



VOLUME ONE

ENCYCLOPEDIA OF QUATERNARY SCIENCE

2ND EDITION

SCOTT A. ELIAS EDITOR-IN-CHIEF
CARY J. MOCK ASSOCIATE EDITOR



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To go to page "123" of Volume 3, type "III - 123"... and so forth.

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ENCYCLOPEDIA OF **QUATERNARY SCIENCE**

SECOND EDITION

Volume 1

A–GLA

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ENCYCLOPEDIA OF QUATERNARY SCIENCE

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VOLUME 1

A–GLA



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DEDICATIONS

The second edition of the Encyclopedia of Quaternary Science is dedicated to the memory of three outstanding scientists who have recently passed away: Russell Coope, Bill Watts, and Aleksis Dreimanis.

Russell Coope

G. Russell Coope (1930–2011) was essentially the founding father of Quaternary Entomology, having initiated the modern era of Quaternary insect fossil research in the late 1950s. Prior to that time, and going back to the nineteenth century, there had been a number of false starts and stops in this field. Previous researchers had assumed that any ice-age insect fossils they found would necessarily represent extinct taxa. This, as it turns out, was a completely false assumption. Beginning with the Late Pleistocene Upton Warren and Chelford sites, Russell was able to determine that fossil insects, especially beetles (Coleoptera) from Pleistocene deposits, represent extant species, albeit species that may live in dramatically different regions today than those where they lived in the past. For many years at the start of this endeavor, Russell was ‘a voice in the wilderness,’ having to convince colleagues in both the Quaternary sciences and modern Entomology that it is possible to identify ‘beetle bits’ (individual sclerites of beetle exoskeleton) to the species level, and that fossil beetle assemblages can yield reliable reconstructions of past environments. One of his greatest battles was to convince skeptical palynologists that the interstadial interval in what is now known as MIS 3, and another one toward the end of the last glaciation, were almost as warm as today, in spite of the fact that the vegetation cover represented in pollen records failed to register this. Likewise, Russell championed the idea that climate oscillations, particularly in the late glacial interval and early Holocene, were extremely rapid and dramatic: switching from full glacial to full interglacial conditions in a matter of decades, at most. Russell’s paleoclimate reconstructions were finally given full credence by the scientific community at large when the oxygen isotope records from Greenland ice cores began to be published, showing nearly the exact same amplitude and timing of climate change he had been publishing since the 1960s.

Russell’s research has had important implications for the field of zoogeography. On the basis of his work, we now know that insects respond to climate change chiefly by shifting their distributional ranges, so that they



Figure 1 Russell Coope at the Chelford quarry fossil site, August 1977. Photograph by Alan V. Morgan, used with permission.

track these changes through space. Through his tenacity and hard work, Russell was able to track down the modern ranges of beetle species he found in Pleistocene deposits to such remote locations as northeastern Siberia and the Tibetan Plateau (cold-adapted species), as well as North Africa and the Mediterranean islands (warm-adapted species).

Russell brought a level of curiosity and enthusiasm for natural history and all things scientific that was infectious. He was a brilliant lecturer and storyteller, holding his audiences practically spellbound. He trained many of the paleoentomologists who took the field forward in recent decades. His seemingly boundless energy kept him working and publishing papers, right up to the end of his life. He published more than 200 papers in his long career. He collaborated with dozens of other Quaternary scientists and was one of the first to appreciate and promote multidisciplinary studies of Pleistocene fossil sites. There is no doubt that his research and publishing legacies will continue to inspire future generations of paleoentomologists.

Bill Watts

William (Bill) A. Watts (1930–2010) was a towering figure in Quaternary paleoecology. He was a skilled botanist and ecologist and this was a secret to his success in unraveling the interglacial, late-glacial, and Holocene vegetation history not only in his native Ireland but also in the midwestern, the southeastern, and eastern USA, and in Italy. A major contribution was his reintroduction and development of quantitative plant macrofossil analysis. Macrofossils can often be identified to lower taxonomic levels than pollen, and Bill took full advantage of the complementarity of both methods. Thus, he was able to determine important features of vegetation history by identifying trees to species level and to illuminate the history of aquatic plants and local vegetational developments.



Figure 2 Photograph of Bill Watts, taken at Laguna de la Roya, Spain. Photograph by Fraser Mitchell, used with permission.

Bill always enjoyed fieldwork and microscope work, carrying out the latter in spare time left over from his job as Provost of Trinity College Dublin and his participation on committees dealing with hospitals and conservation in Ireland. Bill's quiet but intense enthusiasm and his great sense of humor added much to discussions, macrofossil identification sessions, and excursions with him. He and his work are inspirations to those who follow him.



Figure 3 Aleksis Dreimanis, photograph by Stephen Hicock.

Aleksis Dreimanis

Aleksis Dreimanis (1914–2011) was widely considered the world's foremost Quaternary glacial geologist in the 1970s and 80s. Aleksis was born in Latvia. While still an undergraduate at the University of Latvia, he published his first paper on the structural geology of deformed glacial deposits along the banks of the River Daugava – a study that he would return to and revise 60 years later. In 1944, Aleksis was conscripted as a military geologist by the Germans who occupied Latvia. For the next two years, he was separated from his family and forced to do strategic geological mapping in eastern Germany and northern Italy until the end of WWII, followed by internment in prisoner of war and displaced persons camps. In 1948, he escaped to London, England, where he met with the President of the University of Western Ontario who gave him a job in London, Canada (his family was able to join him the following year).

Over the next six decades at the University of Western Ontario, Aleksis ascended to the top of his discipline while gaining international recognition through landmark publications and inspired leadership on committees. His 1970s synthesis and definition of the Wisconsin Stage with Paul Karrow, Jan Terasmae, Dick Goldthwait, and Jane Forsyth became a world standard to which other records were compared. In 1989, he expanded this synthesis and introduced the conceptual till tetrahedron that showed the till continuum with end members occupying the corners of an inverted pyramid. He was among the first geologists to use radiocarbon dating, and he provided samples for inter-lab calibration. He was among the first to quantitatively study the relationship between till lithology and till formation, and to investigate how glacial transport and dynamics determine till texture, deformation, and rheologic states. Aleksis produced over 200 publications spanning eight decades and over a wide range of topics. Largely because of his influence, the study of till genesis and its relation to till properties and past glacial behavior has become a major focus of glacial geologic research in recent decades. The contribution and legacy that forms the body of work of Dreimanis will continue as a major benchmark within the diverse fields of scientific enquiry that are sometimes loosely referred to as 'glacial geology.'

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FOREWORD

As with the publication of the first edition in 2007, the publication of the second edition of the *Encyclopedia of Quaternary Science* represents a landmark in the history of publishing in the field of Quaternary Science. Quaternary Science is a multidisciplinary endeavor which seeks to establish as detailed a picture as possible of the manifold environmental changes that have occurred during the most recent geological period, the Quaternary – an interval of time that spans the past 2.59 million years or the past 0.056% of geological time. It is a period of significant climate and environmental change and witnessed the widespread dispersal of our species, *Homo sapiens*, across the planet.

Since Louis Agassiz and Reverend William Buckland traipsed over parts of the Scottish landscape in 1840 in search of evidence of glaciation, a huge literature has emerged on the science of long-term climate and environmental change. Rapid technological advances in the late twentieth century and the proliferation of scientific journals, particularly in an era of electronic publishing, have resulted in an exponential growth and documentation of knowledge on the climate and environmental changes that have occurred during the Quaternary period. More recently, there has been an increasing public interest in applied Quaternary research as a framework for understanding the basis for recent climate changes and for understanding the nature and frequency of geological hazards and vexing issues such as soil erosion and land degradation, and the adverse effects of ocean water temperature increases and acidification on coral reef environments. In a similar manner, the likely magnitude of future sea-level rise and the associated impacts on coastal landscapes in the twenty-first century have attracted wide public interest. In this sense, Quaternary Science is very much on the political agenda and is a critically important subject to address issues of public concern.

The *Encyclopedia of Quaternary Science* edited by Scott Elias of Royal Holloway, University of London, UK, accordingly represents a particularly welcome addition to the literature. The encyclopedia presents an up-to-date and authoritative overview of Quaternary Science. The encyclopedia should enjoy a wide readership as the entries are presented in a very clear and easily readable style. The text of the articles is written at a level that allows undergraduate students to understand the material, while providing active researchers with a ready reference resource for information in the field. Each entry of up to 4000 words covers the salient points of each topic with very clear illustrations. A central theme that pervades the work is the importance of Quaternary Science in providing an historical context for assessing present environmental changes and as basis for modeling potential future changes.

The encyclopedia consists of four volumes in print form and is available electronically. All the entries have been updated and the text totals around 3,500,000 words. As a major reference work, the encyclopedia has a very wide coverage of topics within the Quaternary sciences reflecting the complex and interdisciplinary nature of the science. Each of the major sections begins with a general overview of the topic prepared by a leading expert in the field. The major sections, for example, examine the analytical methods commonly used in paleoenvironmental reconstructions to unravel in a forensic-like manner the nature of former environments and the tempo of environmental change. Accordingly, a great range of topics such as the former extent of ice cover and nature of Quaternary glaciations, the biological responses and resultant fossil records to fluctuating climate, the expansion and contraction of desert environments, and global and local changes in relative sea levels are examined. Other topics covered include dating techniques, Quaternary stratigraphy, fluvial environments, lake level studies, paleosols, paleobotany, ancient DNA, paleolimnology, vertebrate studies, insect fossil studies, paleoceanography, stable isotope studies, ice core records, and human evolution in the Quaternary. One of the new sections in the encyclopedia examines the application of Quaternary proxy evidence in forensic science. All sections provide a clear summary of the latest advances in the fields of research.

This is an outstanding work and the editors and the publisher are to be congratulated for producing an encyclopedia that cogently summarizes the current state of the science.

A handwritten signature in black ink, appearing to read 'C. Murray-Wallace', with a stylized, flowing script.

Colin V. Murray-Wallace
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INTRODUCTION

It has been my privilege to work with more than 420 leading Quaternary scientists in developing the second edition of the *Encyclopedia of Quaternary Science*. This team of writers and editors represents 28 countries in Europe, Asia, Africa, the Americas, Australia, and New Zealand. Starting with the first edition in 2006, I have had my finger on the pulse of Quaternary science, and this branch of science is truly pulsating! Information now comes from an incredible variety of disciplines: geochemistry, numerical modeling, history, vulcanology, paleobiology, nuclear physics, stratigraphy, sedimentology, climatology, anthropology, archaeology, glacial geology, soil science, ice-stream modeling – the list is staggering. This highly disparate group of people are bound together by one common thread: the desire to know the history of the planet during the last 2.6 million years – the time of the ice ages. For Quaternary scientists, this is a pressing need, not an idle curiosity. Any doubts about this statement can be easily dispelled by a consideration of the lengths to which many of them go to gather the necessary data. Some of them have worked for months in sub-zero temperatures on top of very high mountains, or near the center of polar ice sheets, collecting ice cores. Others have spent many weeks on some of the roughest seas in the world, drilling deep-sea sediment cores. Often the work is more mundane. An oxygen isotope curve for a lengthy marine sediment core represents thousands of hours of patiently picking tiny fossils from layer after layer of sediments, in order to obtain sufficient numbers of calcium carbonate shells to yield samples for isotopic analysis. A map showing proposed ice limits from the last glaciation represents thousands of hours of field mapping of glacial features by dozens of people. Why do all of these people devote their lives to this pursuit of knowledge? Does it really matter so much? The answer becomes clear when you step back and examine the topic of Quaternary science in its proper context.

The world we inhabit has largely been shaped by the events of the Quaternary. All the biological communities that exist today are the end product of a long series of species associations that came together in the past, largely driven by climatic change during the Pleistocene. We cannot properly understand the functioning of modern ecosystems without a solid knowledge of their history, any more than we can understand the plot of a long novel by reading just the final page. We are also living in a time of alarming climate changes. Even though the pace and intensity of some of these changes have not been seen in historical times, there were many rapid, large-scale climatic shifts in the Pleistocene. The best way to predict the effects of global warming on the planet's climate and ecosystems is to look at the history of similarly intense, rapid changes in the prehistoric past. The interval that is most relevant today is the most recent geologic period: the Quaternary. As human populations rise exponentially, increasing numbers of people are exposed to geologic hazards, such as earthquakes (and attendant tsunamis), and volcanic eruptions. These are short-lived events that take place only rarely in any one region. The interval between major events, such as volcanic eruptions, may be centuries or millennia. How do we come to grips with predicting the future likelihood of such erratic phenomena? Again, the answers come from piecing together the ancient history of such events, over many thousands of years.

The Quaternary has been the time when our own species came of age. The beginning of the Quaternary, roughly 2.6 million years ago, was about the time when the earliest member of our genus (*Homo*) first appeared in Africa. Pleistocene environments shaped the course of human evolution, culminating in anatomically modern *Homo sapiens* spreading from Africa throughout most of the world during the last glaciation. Even though human beings largely shape their own environments today, for the vast majority of our species' history, it has been the environment that has shaped us. Our direct ancestors' adaptations to environmental change are deeply ingrained in our genes. Thus, an understanding of the environmental conditions that shaped our species is critical to our understanding of humanity.

Quaternary science is a rapidly changing field, and the articles that appear in this encyclopedia reflect this. New dating techniques, such as cosmogenic nuclide dating, are revolutionizing our understanding of many earth surface processes. The ability to analyze increasingly smaller samples for radiocarbon and stable isotopes of oxygen and hydrogen means that we are gaining a level of precision in the reconstruction of past events that was unheard of just a few years ago. Stable isotope studies of air bubbles trapped in ice cores from Greenland and Antarctica have given Quaternary scientists an entirely new perspective on the rapidity and intensity of climatic change during the last glacial cycle and beyond. Likewise, the discovery of long sequences of annually laminated sediments in both marine and freshwater environments has provided a great leap forward in our ability to resolve the timing of environmental changes in nonpolar regions. The ability to extract and analyze ancient DNA sequences from Pleistocene fossils (both plants and animals) is revolutionizing the field of paleobiology. We are beginning to be able to trace the genetic lineages of a number of different organisms, from beetles to bison. In short, these are very exciting times to be a Quaternary scientist! While it is virtually impossible for any Quaternary researcher or student to keep abreast of all the new discoveries in this multifaceted science, this encyclopedia can be of great help. The articles contained here represent the state of the art in a huge variety of topics, and they offer the opportunity to dig deeper into their respective subjects by providing full citations of the most pertinent literature available. I invite you to come and explore the Quaternary Period in the pages that follow. It is a fascinating story.

Scott A. Elias

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HOW TO USE THE ENCYCLOPEDIA

The material in this encyclopedia is organized into two broad sections. At the start of Volume 1 are four *Introductory Articles* that serve as a broad introduction to the field. It is intended that these articles should serve as the first port of call for readers, to enable them to embark on any of the main entries without the need to consult other material beforehand.

Following the Introductory Articles, the main body of The Encyclopedia is arranged as a series of entries in alphabetical order. These entries fill the remainder of the volumes.

To help realize the full potential of the material in the Encyclopedia we have provided four features to help you find the topic of your choice: a contents list by subject; an alphabetical contents list; cross-references to other articles; and a full subject index.

1. Contents list by subject

Your first point of reference will probably be the contents list by subject. This list appears at the front of each volume, and groups the entries under subject headings describing the broad themes of Quaternary science. This will enable the reader to make quick connections between entries and to locate the entry of interest. Under each main section heading, you will find several subject areas and under each subject area is a list of those entries that covers aspects of that subject, together with the volume and page numbers on which these entries may be found.

2. Alphabetical contents list

The alphabetical contents list, which also appears at the front of each volume, lists the entries in the order in which they appear in the Encyclopedia. This list provides both the volume number and the page number of each entry.

You will find ‘dummy’ entries where obvious synonyms exist for entries or where we have grouped together related topics. Dummy entries appear in both the contents list and the body of the text.

Example

If you were attempting to locate material on *peat studies* via the alphabetical contents list:

PEAT STUDIES *see* PLANT MACROFOSSIL METHODS AND STUDIES: Mire and Peat Macros

The dummy entry directs you to the entry in which *peat* is covered.

If you were trying to locate the material by browsing through the text and you had looked up *Peat Studies*, then the following information would be provided in the dummy entry:

Peat Studies *see* PLANT MACROFOSSIL METHODS AND STUDIES: Mire and Peat Macros

3. Cross-references

All of the entries in the Encyclopedia have been extensively cross-referenced. The cross-references, which appear at the end of the entry, serve three different functions:

- i. To indicate if a topic is discussed in greater detail elsewhere.
- ii. To draw the reader’s attention to parallel discussions in other entries.
- iii. To indicate material that broadens the discussion.

Example

The following list of cross-references appear at the end of the entry GLACIAL LANDFORMS, ICE SHEETS | Growth and Decay

See also: **Introduction:** History of Quaternary Science; **Paleoclimate Reconstruction:** Sub-Milankovitch (DO/Heinrich) Events; **Glacial Climates:** Thermohaline Circulation; **Glaciation, Causes:** Astronomical Theory of Paleoclimates; **Paleoclimate Modeling:** Last Glacial Maximum GCMs; **Glaciations:** Overview; **Sea Level Studies:** Overview; Paleocceanography; **Ice Core Records:** Correlations Between Greenland and Antarctica.

Here you will find examples of all three functions of the cross-reference list: a topic discussed in greater detail elsewhere (e.g., **Paleoclimate Modeling:** Last Glacial Maximum GCMs), parallel discussion in other entries (e.g., **Glacial Climates:** Thermohaline Circulation), and reference to entries that broaden the discussion (e.g., **Glaciations:** Overview).

4. Index

The index provides you with the page number where the material is located. The index entries differentiate between material that is a whole entry, is part of an entry, or is data presented in a figure or a table. Detailed notes are provided on the opening page of the index.

5. Contributors

A full list of contributors is listed at the beginning of each volume.

CONTENTS LIST BY SUBJECT

Location references refer to the volume number and page number (separated by a colon)

Introduction and History of Science

INTRODUCTION | History of Quaternary Science, 1:1
INTRODUCTION | History of Dating Methods, 1:9
INTRODUCTION | Societal Relevance of Quaternary Research, 1:17
INTRODUCTION | Understanding Quaternary Climatic Change, 1:26

Paleoclimatology

Overview and Approaches

PALEOCLIMATE | Introduction, 3:87
PALEOCLIMATE RECONSTRUCTION | Approaches, 3:179
PALEOCLIMATE MODELING | Data-Model Comparisons, 3:135
PALEOCLIMATE | Timescales of Climate Change, 3:93
PALEOCLIMATE | Modern Analog Approaches in Paleoclimatology, 3:102
PALEOCLIMATE | Paleoclimate History of the Arctic, 3:113
PALEOCLIMATE | The Younger Dryas Climate Event, 3:126

Paleoclimate Reconstructions

PALEOCLIMATE RECONSTRUCTION | Pliocene Environments, 3:185
GLACIATION, CAUSES | Tectonic Uplift and Continental Configurations, 2:127
GLACIATION, CAUSES | Astronomical Theory of Paleoclimates, 2:136
PALEOCLIMATE RECONSTRUCTION | Paleodroughts and Society, 3:194
PALEOCLIMATE RECONSTRUCTION | Paleotempestology, 3:209

PALEOCLIMATE RECONSTRUCTION | Sub-Milankovitch (DO/Heinrich) Events, 3:200
PALEOCLIMATE RECONSTRUCTION | Younger Dryas Oscillation, Global Evidence, 3:222
PALEOCLIMATE RECONSTRUCTION | Historical Climatology, 3:237
PALEOCLIMATE RECONSTRUCTION | The Last Millennium, 3:229
PALEOCLIMATE RELEVANCE TO GLOBAL WARMING, 3:244

Climate Forcing Mechanisms

GLACIAL CLIMATES | Thermohaline Circulation, 1:737
GLACIAL CLIMATES | Biosphere Feedbacks, 1:721
GLACIAL CLIMATES | Volcanic and Solar Forcing, 1:748
GLACIAL CLIMATES | Effects of Atmospheric Dust, 1:729

Climate Modelling

PALEOCLIMATE MODELING | Quaternary Environments, 3:147
PALEOCLIMATE MODELING | The Last Interglacial, 3:155
PALEOCLIMATE MODELING | Last Glacial Maximum GCMs, 3:165
PALEOCLIMATE MODELING | Paleo-ENSO, 3:171

Dating Techniques

Radiometric Methods

DATING TECHNIQUES, 1:447
RADIOCARBON DATING | Conventional Method, 4:305
RADIOCARBON DATING | AMS Radiocarbon Dating, 4:316

RADIOCARBON DATING | Sources of Error, 4:324
RADIOCARBON DATING | Variations in Atmospheric ^{14}C , 4:329
RADIOCARBON DATING | Causes of Temporal ^{14}C Variations, 4:336
RADIOCARBON DATING | Calibration of the ^{14}C Record, 4:345
RADIOCARBON DATING | Charcoal, 4:353
RADIOCARBON DATING | ^{14}C of Plant Macrofossils, 4:361
K/Ar AND $^{40}\text{Ar}/^{39}\text{Ar}$ DATING, 2:477
U-SERIES DATING, 4:567
FISSION-TRACK DATING, 1:643
LUMINESCENCE DATING | Thermoluminescence, 2:643
LUMINESCENCE DATING | Optical Dating, 2:653
LUMINESCENCE DATING | Electron Spin Resonance Dating, 2:667
COSMOGENIC NUCLIDE DATING | Overview, 1:407
COSMOGENIC NUCLIDE DATING | Cosmic Ray Interactions in Minerals, 1:418
COSMOGENIC NUCLIDE DATING | Methods, 1:410
COSMOGENIC NUCLIDE DATING | Exposure Geochronology, 1:432
COSMOGENIC NUCLIDE DATING | Landscape Evolution, 1:440

Geomagnetic Excursions and Secular Variations in the Magnetic Field

Geomagnetic Excursions and Secular Variations, 1:705

Biological Methods

LICHENOMETRY, 2:565

Annual Layer Methods

VARVED MARINE SEDIMENTS, 4:582
VARVED LAKE SEDIMENTS, 4:573

Chemical Methods

AMINO-ACID DATING, 1:37

Quaternary Stratigraphy

QUATERNARY STRATIGRAPHY | Tephrochronology, 4:277
QUATERNARY STRATIGRAPHY | Overview, 4:189
QUATERNARY STRATIGRAPHY | Lithostratigraphy, 4:227

QUATERNARY STRATIGRAPHY | Continental Biostratigraphy, 4:206
QUATERNARY STRATIGRAPHY | Pedostratigraphy, 4:250
QUATERNARY STRATIGRAPHY | Morphostratigraphy/Allostratigraphy, 4:243
QUATERNARY STRATIGRAPHY | Climatostratigraphy, 4:222
QUATERNARY STRATIGRAPHY | Sequence Stratigraphy, 4:260
QUATERNARY STRATIGRAPHY | Chronostratigraphy, 4:215

Quaternary Glaciation

Glacial Landforms

GLACIAL LANDFORMS | Introduction, 1:755
GLACIAL LANDFORMS, ICE SHEETS | Growth and Decay, 1:877
GLACIAL LANDFORMS, ICE SHEETS | Evidence of Glacier and Ice Sheet Extent, 1:884
GLACIAL LANDFORMS | Evidence of Glacier Recession, 1:803
GLACIAL LANDFORMS, ICE SHEETS | PaleoeLAS, 1:909
GLACIAL LANDFORMS | Glacitectonic Structures and Landforms, 1:839
GLACIAL LANDFORMS, SEDIMENTS | Glaciogenic Lithofacies, 2:18
GLACIAL LANDFORMS | Moraine Forms and Genesis, 1:769
GLACIAL LANDFORMS, SEDIMENTS | Glaciofluvial Landforms of Deposition, 2:6
GLACIAL LANDFORMS | Glaciofluvial Landforms of Erosion, 1:825
GLACIAL LANDFORMS, SEDIMENTS | Glacial Erratics and Till Dispersal Indicators, 2:81
GLACIAL LANDFORMS, SEDIMENTS | Glaciomarine Sediments and Ice-Rafted Debris, 2:30
GLACIAL LANDFORMS, SEDIMENTS | Tills, 2:62
GLACIAL LANDFORMS, SEDIMENTS | Till Fabric Analysis, 2:76
GLACIAL LANDFORMS | Glacial Land Systems, 1:813
GLACIAL LANDFORMS, EROSIONAL FEATURES | Micro- to Macroscale Forms, 1:865
GLACIAL LANDFORMS, EROSIONAL FEATURES | Major Scale Forms, 1:847
GLACIAL LANDFORMS, ICE SHEETS | Trimlines and Palaeonunataks, 1:918
GLACIAL LANDFORMS | Quaternary Vulcanism: Subglacial Landforms, 1:780
GLACIAL LANDFORMS, SEDIMENTS | Micromorphology of Glacial Sediments, 2:52

GLACIAL LANDFORMS, SEDIMENTS | Clast Form Analysis, 2:1
 GLACIAL LANDFORMS, ICE SHEETS | Evidence of Glacier Flow Directions, 1:895
 GLACIAL LANDFORMS, SEDIMENTS | Glaciolacustrine, 2:43
 GLACIAL LANDFORMS, SEDIMENTS | Glacial Sequence Stratigraphy, 2:85

Permafrost and Periglacial Features

PERMAFROST AND PERIGLACIAL FEATURES | Permafrost, 3:464
 PERMAFROST AND PERIGLACIAL FEATURES | Active Layer Processes, 3:421
 PERMAFROST AND PERIGLACIAL FEATURES | Paraglacial Geomorphology, 3:553
 PERMAFROST AND PERIGLACIAL FEATURES | Slope Deposits and Forms, 3:481
 PERMAFROST AND PERIGLACIAL FEATURES | Frost Mounds: Active and Relict Forms, 3:472
 PERMAFROST AND PERIGLACIAL FEATURES | Patterned Ground, 3:452
 PERMAFROST AND PERIGLACIAL FEATURES | Talus Slopes, 3:566
 PERMAFROST AND PERIGLACIAL FEATURES | Thermokarst Topography, 3:574
 PERMAFROST AND PERIGLACIAL FEATURES | Ice Wedges and Ice-Wedge Casts, 3:436
 PERMAFROST AND PERIGLACIAL FEATURES | Cryoturbation Structures, 3:430
 PERMAFROST AND PERIGLACIAL FEATURES | Blockfields (Felsenmeer), 3:523
 PERMAFROST AND PERIGLACIAL FEATURES | Block/Rock Streams, 3:514
 PERMAFROST AND PERIGLACIAL FEATURES | Rock Weathering, 3:500
 PERMAFROST AND PERIGLACIAL FEATURES | Rock Glaciers and Protalus Forms, 3:535
 PERMAFROST AND PERIGLACIAL FEATURES | Periglacial Fluvial Sediments and Forms, 3:490
 PERMAFROST AND PERIGLACIAL FEATURES | Permafrost and Glacier Interactions, 3:507
 PERMAFROST AND PERIGLACIAL FEATURES | Yedoma: Late Pleistocene Ice-Rich Syngenetic Permafrost of Beringia, 3:542

Tree-ring section

DENDROCHRONOLOGY, 1:453
 DENDROCLIMATOLOGY, 1:459
 GLACIAL LANDFORMS, TREE RINGS | Dendrogeomorphology, 2:91
 GLACIAL LANDFORMS, TREE RINGS | Dendroglaciology, 2:104

Fluvial Environments

FLUVIAL ENVIRONMENTS | Sediments, 1:663
 FLUVIAL ENVIRONMENTS | Terrace Sequences, 1:684
 Glacial–Interglacial Scale Fluvial Responses, 2:112
 FLUVIAL ENVIRONMENTS | Responses to Rapid Environmental Change, 1:676
 FLUVIAL ENVIRONMENTS | Deltaic Environments, 1:693
 PALEOHYDROLOGY, 3:253

History of Quaternary Glaciations

GLACIATIONS | Overview, 2:143
 GLACIATIONS | Transition from Late Neogene to Early Quaternary Environments, 2:151
 GLACIATIONS | Early Quaternary (Pleistocene) and Precursors, 2:167
 GLACIATIONS | Middle Pleistocene in Eurasia, 2:172
 GLACIATIONS | Mid-Quaternary in North America, 2:180
 GLACIATIONS | Middle Pleistocene Glaciations in the Southern Hemisphere, 2:187
 GLACIATIONS | Late Quaternary of Antarctica, 2:216
 GLACIATIONS | Late Quaternary in North America, 2:245
 GLACIATIONS | Late Quaternary in Highland Asia, 2:236
 GLACIATIONS | Late Quaternary of the Southwest Pacific Region, 2:202
 GLACIATIONS | Late Pleistocene in Eurasia, 2:224
 GLACIATIONS | Late Pleistocene in South America, 2:250
 GLACIATIONS | Late Pleistocene Glacial Events in Beringia, 2:191
 GLACIATIONS | Neoglaciation in Europe, 2:257
 GLACIATIONS | Neoglaciation in the American Cordilleras, 2:269

Sea Level Studies

SEA LEVEL STUDIES | Overview, 4:369
 SEA LEVEL STUDIES | Geomorphological Indicators, 4:377
 SEA LEVEL STUDIES | Coral Records of Relative Sea-Level Changes, 4:409
 SEA LEVEL STUDIES | Use of Cave Data in Sea-Level Reconstructions, 4:460
 SEA LEVEL STUDIES | Eustatic Sea-Level Changes – Glacial–Interglacial Cycles, 4:429
 SEA LEVEL STUDIES | Eustatic Sea-Level Changes Since the Last Glacial Maximum, 4:439

SEA LEVEL STUDIES | Isostasy: Glaciation-Induced Sea-Level Change, 4:452
SEA LEVEL STUDIES | Sedimentary Indicators of Relative Sea-Level Changes – High Energy, 4:385
SEA LEVEL STUDIES | Sedimentary Indicators of Relative Sea-Level Changes – Low Energy, 4:396
SEA LEVEL STUDIES | Microfossil-Based Reconstructions of Holocene Relative Sea-Level Change, 4:419
SEA-LEVELS, LATE QUATERNARY | Late Quaternary Relative Sea-Level Changes in High Latitudes, 4:467
SEA-LEVELS, LATE QUATERNARY | Late Quaternary Sea-Level Changes in Greenland, 4:481
SEA-LEVELS, LATE QUATERNARY | Late Quaternary Relative Sea-Level Changes at Mid-Latitudes, 4:489
SEA-LEVELS, LATE QUATERNARY | Late Quaternary Relative Sea-Level Changes in the Tropics, 4:495
SEA-LEVELS, LATE QUATERNARY | Tectonics and Relative Sea-Level Change, 4:503

Paleosols and Wind-Blown Sediments

Paleosol Research

PALEOSOLS AND WIND-BLOWN SEDIMENTS | Overview, 3:357
PALEOSOLS AND WIND-BLOWN SEDIMENTS | Nature of Paleosols, 3:367
LOESS DEPOSITS: ORIGINS AND PROPERTIES, 2:573
PALEOSOLS AND WIND-BLOWN SEDIMENTS | Weathering Profiles, 3:392
PALEOSOLS AND WIND-BLOWN SEDIMENTS | Soil Micromorphology, 3:381
PALEOSOLS AND WIND-BLOWN SEDIMENTS | Soil Morphology in Quaternary Studies, 3:402
PALEOSOLS AND WIND-BLOWN SEDIMENTS | Mineral Magnetic Analysis, 3:375
PALEOSOLS AND WIND-BLOWN SEDIMENTS | Biogeochemical Role of Dust in Quaternary Climate Cycles, 3:412

Dune Studies

DUNE FIELDS | High Latitudes, 1:597
DUNE FIELDS | Mid-Latitudes, 1:606
DUNE FIELDS | Low Latitudes, 1:623

Loess Studies

LOESS RECORDS | Central Asia, 2:585
LOESS RECORDS | China, 2:595
LOESS RECORDS | Europe, 2:606
LOESS RECORDS | North America, 2:620
LOESS RECORDS | South America, 2:629
EOLIAN RECORDS, DEEP-SEA SEDIMENTS, 1:637

Lake Level Studies

LAKE LEVEL STUDIES | Overview, 2:483
LAKE LEVEL STUDIES | Modeling, 2:558
LAKE LEVEL STUDIES | Africa during the Late Quaternary, 2:499
LAKE LEVEL STUDIES | Asia, 2:506
LAKE LEVEL STUDIES | Australia, 2:524
LAKE LEVEL STUDIES | Latin America, 2:531
LAKE LEVEL STUDIES | North America, 2:537
LAKE LEVEL STUDIES | West-Central-Europe, 2:549

Paleobotany

Overview

PALEOBOTANY | Overview of Terrestrial Pollen Data, 2:699

Pollen Studies

POLLEN ANALYSIS, PRINCIPLES, 3:794
POLLEN METHODS AND STUDIES | Use of Pollen as Climate Proxies, 3:805
POLLEN METHODS AND STUDIES | Surface Samples and Trapping, 3:839
POLLEN METHODS AND STUDIES | The Biome Approach to Reconstructing Past Vegetation, 3:861
POLLEN METHODS AND STUDIES | POLLSCAPE Model, 3:871
POLLEN METHODS AND STUDIES | Reconstructing Past Biodiversity Development, 3:816
POLLEN METHODS AND STUDIES | Numerical Analysis Methods, 3:821
POLLEN METHODS AND STUDIES | Changing Plant Distributions and Abundances, 3:854
POLLEN METHODS AND STUDIES | Stand-Scale Palynology, 3:846
PALEOBOTANY | Paleophytogeography, 2:730
POLLEN METHODS AND STUDIES | Databases and Their Application, 3:831

Charred Particle Analysis

PALEOBOTANY | Charred Particle Analyses, 2:716

Quaternary Pollen Records

POLLEN RECORDS, LAST INTERGLACIAL OF EUROPE, 4:1
POLLEN RECORDS, LATE PLEISTOCENE | Africa, 4:9
POLLEN RECORDS, LATE PLEISTOCENE | Australasia, 4:18
POLLEN RECORDS, LATE PLEISTOCENE | Northern Asia, 4:27

POLLEN RECORDS, LATE PLEISTOCENE | Northern North America, 4:39
 POLLEN RECORDS, LATE PLEISTOCENE | Western North America, 4:72
 POLLEN RECORDS, LATE PLEISTOCENE | South America, 4:52
 POLLEN RECORDS, LATE PLEISTOCENE | Middle and Late Pleistocene in Southern Europe, 4:63
 POLLEN RECORDS, POSTGLACIAL | Africa, 4:85
 POLLEN RECORDS, POSTGLACIAL | Australia and New Zealand, 4:104
 POLLEN RECORDS, POSTGLACIAL | Northeastern North America, 4:115
 POLLEN RECORDS, POSTGLACIAL | Northwestern North America, 4:124
 POLLEN RECORDS, POSTGLACIAL | Southeastern North America, 4:133
 POLLEN RECORDS, POSTGLACIAL | Southwestern North America, 4:142
 POLLEN RECORDS, POSTGLACIAL | South America, 4:156
 POLLEN RECORDS, POSTGLACIAL | Northern Asia, 4:164
 POLLEN RECORDS, POSTGLACIAL | Northern Europe, 4:173
 POLLEN RECORDS, POSTGLACIAL | Southern Europe, 4:179
 POLLEN METHODS AND STUDIES | Archaeological Applications, 3:880

Phytoliths

PHYTOLITHS, 3:582

Ancient Plant DNA

PALEOBOTANY | Ancient Plant DNA, 2:705

Plant Macrofossil Studies

PLANT MACROFOSSIL INTRODUCTION, 3:593
 PLANT MACROFOSSIL METHODS AND STUDIES | Surface Samples, Taphonomy, Representation, 3:684
 PLANT MACROFOSSIL METHODS AND STUDIES | Treeline Studies, 3:690
 PLANT MACROFOSSIL METHODS AND STUDIES | Megafossils, 3:621
 PLANT MACROFOSSIL METHODS AND STUDIES | Paleolimnological Applications, 3:657
 PLANT MACROFOSSIL METHODS AND STUDIES | CO₂ Reconstruction from Fossil Leaves, 3:613
 PLANT MACROFOSSIL METHODS AND STUDIES | Rodent Middens, 3:674

PLANT MACROFOSSIL METHODS AND STUDIES | Mire and Peat Macros, 3:637
 PLANT MACROFOSSIL METHODS AND STUDIES | Use in Environmental Archaeology, 3:699
 PLANT MACROFOSSIL METHODS AND STUDIES | Dendroarchaeology, 3:630
 PLANT MACROFOSSIL METHODS AND STUDIES | Validation of Pollen Studies, 3:725

Plant Macrofossil Records

PLANT MACROFOSSIL RECORDS | Arctic Eurasia, 3:733
 PLANT MACROFOSSIL RECORDS | Greenland, 3:760
 PLANT MACROFOSSIL RECORDS | Arctic North America, 3:746
 PLANT MACROFOSSIL RECORDS | Holocene North America, 3:768
 PLANT MACROFOSSIL RECORDS | Late Glacial Multidisciplinary Studies, 3:785

Diatom Studies

DIATOM INTRODUCTION, 1:471
 DIATOM METHODS | Data Interpretation, 1:489
 DIATOM RECORDS | North Atlantic and Arctic, 1:562
 DIATOM RECORDS | Pacific, 1:571
 DIATOM RECORDS | Antarctic Waters, 1:527
 DIATOM RECORDS | Diatom Fossil Records from Marine Laminated Sediments, 1:554
 DIATOM RECORDS | Freshwater Laminated Sequences, 1:540
 DIATOM RECORDS | Large Lakes, 1:546
 DIATOM METHODS | Salinity and Climate Reconstructions from Continental Lakes, 1:507
 DIATOM METHODS | $\delta^{18}\text{O}$ Records, 1:481
 PALEOBOTANY | Silicon Isotopes in Diatoms, 2:734
 DIATOM METHODS | Use in Archaeology, 1:516
 DIATOM METHODS | Diatomites: Their Formation, Distribution, and Uses, 1:501
 Structures and Applications of Biomarkers from Arctic Sea Ice Diatoms, 1:588

Paleolimnology

PALEOLIMNOLOGY | Overview of Paleolimnology, 3:259
 PALEOLIMNOLOGY | Lake Chemistry, 3:292
 PALEOLIMNOLOGY | Physical Properties of Lake Sediments, 3:300
 PALEOLIMNOLOGY | Cladocera, 3:271
 PALEOLIMNOLOGY | Freshwater Mollusks, 3:281

PALEOLIMNOLOGY | Contributions of
Paleolimnological Research to
Biogeography, 3:313
PALEOLIMNOLOGY | Pigment Studies, 3:326
PALEOLIMNOLOGY | Visible and Infrared
Spectroscopical Applications, 3:349
PALEOLIMNOLOGY | Multiproxy Approaches, 3:339

Vertebrate Studies

VERTEBRATE OVERVIEW, 4:590
VERTEBRATE RECORDS | Early Pleistocene, 4:599
VERTEBRATE RECORDS | Early and Middle
Pleistocene of Northern Eurasia, 4:605
VERTEBRATE RECORDS | Mid-Pleistocene of
Southern Asia, 4:651
VERTEBRATE RECORDS | Mid-Pleistocene of
Africa, 4:615
VERTEBRATE RECORDS | Mid-Pleistocene of
Australia, 4:621
VERTEBRATE RECORDS | Mid-Pleistocene of
Europe, 4:639
VERTEBRATE RECORDS | Mid-Pleistocene of
North America, 4:646
VERTEBRATE RECORDS | Late Pleistocene of
Africa, 4:664
VERTEBRATE RECORDS | Late Pleistocene of North
America, 4:673
VERTEBRATE RECORDS | Late Pleistocene of South
America, 4:680
VERTEBRATE RECORDS | Late Pleistocene of
Southeast Asia, 4:693
VERTEBRATE RECORDS | Late Pleistocene
Megafaunal Extinctions, 4:700
VERTEBRATE RECORDS | Late Pleistocene
Mummified Mammals, 4:713
VERTEBRATE STUDIES | Ancient DNA, 4:719
VERTEBRATE STUDIES | Speciation and
Evolutionary Trends in Quaternary
Vertebrates, 4:723
VERTEBRATE STUDIES | Dwarfing and Gigantism in
Quaternary Vertebrates, 4:733
VERTEBRATE STUDIES | Interactions with
Hominids, 4:748

Insect Fossil Studies

Beetle Studies

BEETLE RECORDS | Overview, 1:161
BEETLE RECORDS | Late Tertiary and Early
Quaternary Records, 1:173
BEETLE RECORDS | Middle Pleistocene of Western
Europe, 1:184

BEETLE RECORDS | Late Pleistocene of
Europe, 1:200
BEETLE RECORDS | Late Pleistocene of North
America, 1:221
BEETLE RECORDS | Late Pleistocene of South
America, 1:235
BEETLE RECORDS | Late Pleistocene of
Australia, 1:191
BEETLE RECORDS | Late Pleistocene of Northern
Asia, 1:255
BEETLE RECORDS | Late Pleistocene of Japan, 1:207
BEETLE RECORDS | Late Pleistocene of New
Zealand, 1:244
BEETLE RECORDS | Postglacial Europe, 1:274
BEETLE RECORDS | Postglacial North America, 1:282

Chironomid Studies

CHIRONOMID RECORDS | Chironomid
Overview, 1:355
CHIRONOMID RECORDS | Late Pleistocene of
Europe, 1:373
CHIRONOMID RECORDS | Africa, 1:361
CHIRONOMID RECORDS | Late Pleistocene of
Europe, 1:373
CHIRONOMID RECORDS | Postglacial
Europe, 1:386
CHIRONOMID RECORDS | Postglacial Southern
Hemisphere, 1:398

Oribatid Mite Studies

ORIBATID MITES, 2:680

Paleoceanography

Paleoceanographic Studies

PALEOCEANOGRAPHY | Paleoceanography An
Overview, 2:745
PALEOCEANOGRAPHY, BIOLOGICAL PROXIES |
Corals, Sclerosponges and Mollusks, 2:795
PALEOCEANOGRAPHY, BIOLOGICAL PROXIES |
Benthic Foraminifera, 2:765
PALEOCEANOGRAPHY, BIOLOGICAL PROXIES |
Planktic Foraminifera, 2:825
PALEOCEANOGRAPHY, BIOLOGICAL PROXIES |
Radiolarians and Silicoflagellates, 2:830
PALEOCEANOGRAPHY, BIOLOGICAL PROXIES |
Coccolithophores, 2:783
PALEOCEANOGRAPHY, BIOLOGICAL PROXIES |
Alkenone Paleothermometry Based on the
Haptophyte Algae, 2:755
PALEOCEANOGRAPHY, BIOLOGICAL PROXIES |
Marine Diatoms, 2:816

PALEOCEANOGRAPHY, BIOLOGICAL PROXIES |
Dinoflagellates, 2:800

PALEOCEANOGRAPHY, PHYSICAL AND
CHEMICAL PROXIES | Terrigenous
Sediments, 2:941

PALEOCEANOGRAPHY, PHYSICAL AND
CHEMICAL PROXIES | Oxygen Isotopic
Composition of Seawater, 2:915

PALEOCEANOGRAPHY, PHYSICAL AND
CHEMICAL PROXIES | Oxygen-Isotope
Stratigraphy of the Oceans, 2:907

PALEOCEANOGRAPHY, PHYSICAL AND
CHEMICAL PROXIES | Dissolution of Deep-Sea
Carbonates, 2:859

PALEOCEANOGRAPHY, PHYSICAL AND
CHEMICAL PROXIES | Mg/Ca and Sr/Ca
Paleothermometry from Calcareous Marine
Fossils, 2:871

PALEOCEANOGRAPHY, BIOLOGICAL PROXIES |
Temperature Proxies Census Counts, 2:841

PALEOCEANOGRAPHY, PHYSICAL AND
CHEMICAL PROXIES | Salinity Proxies
 $\delta^{18}\text{O}$, 2:932

PALEOCEANOGRAPHY, PHYSICAL AND
CHEMICAL PROXIES | Nutrient Proxies, 2:899

PALEOCEANOGRAPHY, PHYSICAL AND
CHEMICAL PROXIES | Carbon Cycle
Proxies, 2:849

PALEOCEANOGRAPHY, PHYSICAL AND
CHEMICAL PROXIES | Radioisotope
Proxies, 2:923

PALEOCEANOGRAPHY, BIOLOGICAL PROXIES |
Biomarkers Indicators of Past Climate, 2:775

PALEOCEANOGRAPHY, PHYSICAL AND
CHEMICAL PROXIES | Magnetic Proxies and
Susceptibility, 2:884

Paleoceanographic Records

PALEOCEANOGRAPHY, RECORDS | Early
Pleistocene, 3:1

PALEOCEANOGRAPHY, RECORDS | Late
Pleistocene North Atlantic, 3:9

PALEOCEANOGRAPHY, RECORDS | Late
Pleistocene South Atlantic, 3:18

PALEOCEANOGRAPHY, RECORDS | Late
Pleistocene North Pacific, 3:33

PALEOCEANOGRAPHY, RECORDS | Postglacial
North Atlantic, 3:55

PALEOCEANOGRAPHY, RECORDS | Postglacial
North Pacific, 3:62

PALEOCEANOGRAPHY, RECORDS | Postglacial
South Pacific, 3:73

PALEOCEANOGRAPHY, RECORDS | Postglacial
Indian Ocean, 3:46

Ice Core Studies

Ice Core Studies

ICE CORE METHODS | Overview, 2:278

ICE CORES | History of Research, Greenland and
Antarctica, 2:435

ICE CORE METHODS | Chronologies, 2:303

ICE CORE METHODS | Stable Isotopes, 2:347

ICE CORE METHODS | Conductivity Studies, 2:319

ICE CORE METHODS | Glaciochemistry, 2:326

ICE CORE METHODS | Biological Material, 2:288

ICE CORE METHODS | Microparticle and Trace
Element Studies, 2:342

ICE CORE METHODS | CO_2 Studies, 2:311

ICE CORE METHODS | Methane Studies, 2:334

ICE CORE METHODS | ^{10}Be and Cosmogenic
Radionuclides in Ice Cores, 2:353

ICE CORE METHODS | Studies of Firn Air, 2:361

Ice Core Records

ICE CORES | History of Carbon Monoxide and
Ultra-Trace Gases from Ice Cores, 2:463

ICE CORES | History of Nitrous Oxide from Ice
Cores, 2:471

ICE CORES | Dynamics of the Greenland Ice
Sheet, 2:439

ICE CORES | Dynamics of the East Antarctic Ice
Sheet, 2:456

ICE CORES | Dynamics of the West Antarctic Ice
Sheet, 2:448

ICE CORE METHODS | Borehole Temperature
Records, 2:298

ICE CORE RECORDS | Greenland Stable
Isotopes, 2:403

ICE CORE RECORDS | Antarctic Stable Isotopes, 2:395

ICE CORE RECORDS | Ice Margin Sites, 2:416

ICE CORE RECORDS | Correlations Between
Greenland and Antarctica, 2:410

ICE CORE RECORDS | Africa, 2:373

ICE CORE RECORDS | Chinese, Tibetan
Mountains, 2:379

ICE CORE RECORDS | South America, 2:387

ICE CORE RECORDS | Thermal Diffusion
Paleotemperature Records, 2:431

Stable Isotope Research – Carbonates

CARBONATE STABLE ISOTOPES | Overview, 1:291

CARBONATE STABLE ISOTOPES |
Speleothems, 1:294

CARBONATE STABLE ISOTOPES | Terrestrial Teeth
and Bones, 1:304

CARBONATE STABLE ISOTOPES | Terrestrial
Organic Materials, 1:315
CARBONATE STABLE ISOTOPES | Non-Lacustrine
Terrestrial Studies, 1:322
CARBONATE STABLE ISOTOPES | Lake
Sediments, 1:333
CARBONATE STABLE ISOTOPES | Nonmarine
Biogenic Carbonates, 1:341

Humans in the Quaternary

ARCHAEOLOGICAL RECORDS | Overview, 1:49
ARCHAEOLOGICAL RECORDS | 2.7 Myr–300 000
Years Ago in Africa, 1:59
ARCHAEOLOGICAL RECORDS | 2.7 Myr–300 000
Years Ago in Asia, 1:67
ARCHAEOLOGICAL RECORDS | 1.9 Myr–300 000
Years Ago in Europe, 1:83
ARCHAEOLOGICAL RECORDS | Global Expansion
300 000–8000 Years Ago, Africa, 1:91
ARCHAEOLOGICAL RECORDS | Global Expansion
300 000–8000 Years Ago, Asia, 1:98
ARCHAEOLOGICAL RECORDS | Global
Expansion 300 000–8000 Years Ago,
Australia, 1:108

ARCHAEOLOGICAL RECORDS | Global Expansion
300 000–8000 Years Ago, Americas, 1:119
ARCHAEOLOGICAL RECORDS | Neanderthal
Demise, 1:146
ARCHAEOLOGICAL RECORDS | Human Evolution
in the Quaternary, 1:135
ARCHAEOLOGICAL RECORDS | Postglacial
Adaptations, 1:154

Use of Quaternary proxies in forensic science

USE OF QUATERNARY PROXIES IN FORENSIC
SCIENCE | Use of Geology in Forensic Science:
Search to Locate Burials, 4:522
USE OF QUATERNARY PROXIES IN FORENSIC
SCIENCE | Analysis Techniques in Forensic
Palynology, 4:556
USE OF QUATERNARY PROXIES IN FORENSIC
SCIENCE | Soils and Sediment, 4:535
USE OF QUATERNARY PROXIES IN FORENSIC
SCIENCE | The Use of Macroscopic Plant Remains
in Forensic Science, 4:542
USE OF QUATERNARY PROXIES IN FORENSIC
SCIENCE | Insects, 4:548
Diatoms, 1:522

CONTENTS

<i>Dedications</i>	<i>v</i>
<i>Foreword</i>	<i>ix</i>
<i>Introduction</i>	<i>xi</i>
<i>Editorial Advisory Board</i>	<i>xiii</i>
<i>Contributors</i>	<i>xv</i>
<i>How to Use the Encyclopedia</i>	<i>xxvii</i>
<i>Contents List by Subject</i>	<i>xxix</i>

VOLUME 1

INTRODUCTORY ARTICLES	1
History of Quaternary Science <i>S A Elias</i>	1
History of Dating Methods <i>A G Wintle</i>	9
Societal Relevance of Quaternary Research <i>S A Elias</i>	17
Understanding Quaternary Climatic Change <i>J J Lowe, M J C Walker, and S C Porter</i>	26
A	
ALKENONE STUDIES <i>see</i> PALEOCEANOGRAPHY, BIOLOGICAL PROXIES: Alkenone Paleothermometry Based on the Haptophyte Algae	
ALLOSTRATIGRAPHY <i>see</i> QUATERNARY STRATIGRAPHY: Morphostratigraphy/Allostratigraphy	
Amino Acid Dating <i>G H Miller, D S Kaufman, and S J Clarke</i>	37
ANATOMICALLY MODERN HUMANS <i>see</i> ARCHAEOLOGICAL RECORDS: Global Expansion 300 000–8000 Years Ago, Africa; Global Expansion 300 000–8000 Years Ago, Asia; Global Expansion 300 000–8000 Years Ago, Australia; Global Expansion 300 000–8000 Years Ago, Americas	
ARCHAEOLOGICAL RECORDS	49
Overview <i>C Gamble</i>	49

2.7 Myr–300 000 Years Ago in Africa <i>J W K Harris, D R Braun, and M Pante</i>	59
2.7 Myr–300 000 Years Ago in Asia <i>R Dennell</i>	67
1.9 Myr–300 000 Years Ago in Europe <i>J McNabb</i>	83
Global Expansion 300 000–8000 Years Ago, Africa <i>A E Close and T Minichillo</i>	91
Global Expansion 300 000–8000 Years Ago, Asia <i>M D Petraglia and R Dennell</i>	98
Global Expansion 300 000–8000 Years Ago, Australia <i>R Cosgrove</i>	108
Global Expansion 300 000–8000 Years Ago, Americas <i>T Goebel</i>	119
Human Evolution in the Quaternary <i>L Cashmore</i>	135
Neanderthal Demise <i>W Davies</i>	146
Postglacial Adaptations <i>G Bailey</i>	154

B

BEETLE RECORDS	161
Overview <i>S A Elias</i>	161
Late Tertiary and Early Quaternary Records <i>S A Elias and S Kuzmina</i>	173
Middle Pleistocene of Western Europe <i>G R Coope</i>	184
Late Pleistocene of Australia <i>N Porch</i>	191
Late Pleistocene of Europe <i>G Lemdahl and G R Coope</i>	200
Late Pleistocene of Japan <i>M Hayashi</i>	207
Late Pleistocene of North America <i>S A Elias</i>	221
Late Pleistocene of South America <i>A C Ashworth</i>	235
Late Pleistocene of New Zealand <i>M Marra</i>	244
Late Pleistocene of Northern Asia <i>A Sher and S Kuzmina</i>	255

Postglacial Europe <i>P Ponel</i>	274
Postglacial North America <i>S A Elias</i>	282

BERINGIA *see* ARCHAEOLOGICAL RECORDS: Global Expansion 300 000–8000 Years Ago, Americas;
 BEETLE RECORDS: Late Pleistocene of North America; DUNE FIELDS: High Latitudes; GLACIATIONS:
 Late Pleistocene Glacial Events in Beringia; PALEOCEANOGRAPHY, RECORDS: Postglacial North
 Pacific; PLANT MACROFOSSIL RECORDS: Arctic North America; POLLEN RECORDS, LATE
 PLEISTOCENE: Northern North America; VERTEBRATE RECORDS: Late Pleistocene of North America

BIOGENIC CARBONATE STUDIES *see* CARBONATE STABLE ISOTOPES: Nonmarine Biogenic Carbonates

BOND CYCLES *see* PALEOCLIMATE RECONSTRUCTION: Sub-Milankovitch (DO/Heinrich) Events

C

CARBONATE STABLE ISOTOPES 291

Overview <i>H Schwarcz</i>	291
-------------------------------	-----

Speleothems <i>H Schwarcz</i>	294
----------------------------------	-----

Terrestrial Teeth and Bones <i>H Bocherens and D G Drucker</i>	304
---	-----

Terrestrial Organic Materials <i>D McCarroll and N Loader</i>	315
--	-----

Non-Lacustrine Terrestrial Studies <i>J Quade and T Cerling</i>	322
--	-----

Lake Sediments <i>S M Bernasconi and J A McKenzie</i>	333
--	-----

Nonmarine Biogenic Carbonates <i>S J Carpenter</i>	341
---	-----

CARBON DIOXIDE, ATMOSPHERIC CONCENTRATIONS *see* CARBONATE STABLE ISOTOPES:
 Overview; ICE CORE METHODS: CO₂ Studies; PALEOCEANOGRAPHY, PHYSICAL AND CHEMICAL
 PROXIES: Carbon Cycle Proxies ($\delta^{11}\text{B}$, $\delta^{13}\text{C}_{\text{calcite}}$, $\delta^{13}\text{C}_{\text{organic}}$, Shell Weights, B/Ca, U/Ca, Zn/Ca, Ba/Ca);
 PLANT MACROFOSSIL METHODS AND STUDIES: CO₂ Reconstruction from Fossil Leaves

CAVE ART *see* VERTEBRATE STUDIES: Interactions with Hominids

CHARCOAL STUDIES *see* PALEOBOTANY: Charred Particle Analyses

CHIRONOMID RECORDS 355

Chironomid Overview <i>I R Walker</i>	355
--	-----

Africa <i>H Eggermont and D Verschuren</i>	361
---	-----

Late Pleistocene of Europe <i>S J Brooks</i>	373
---	-----

Postglacial Europe <i>G Velle and O Heiri</i>	386
--	-----

Postglacial Southern Hemisphere <i>J Massferro and M Vandergoes</i>	398
--	-----

CLADOCERA STUDIES *see* PALEOLIMNOLOGY: Cladocera

CLIMATE CHANGE *see* PALEOCLIMATE: Introduction; Timescales of Climate Change; PALEOCLIMATE MODELING: Data–Model Comparisons; Quaternary Environments; The Last Interglacial; Last Glacial Maximum GCMs; Paleo-ENSO; PALEOCLIMATE RECONSTRUCTION: Approaches; Pliocene Environments; Paleodroughts and Society; Sub-Milankovitch (DO/Heinrich) Events; Paleotempestology; Younger Dryas Oscillation, Global Evidence; The Last Millennium; Historical Climatology; Paleoclimate Relevance to Global Warming

CLIMATE MODELING, QUATERNARY *see* PALEOCLIMATE MODELING: Quaternary Environments

COCCOLITHOPHORE STUDIES *see* PALEOCEANOGRAPHY, BIOLOGICAL PROXIES: Alkenone Paleothermometry Based on the Haptophyte Algae; Coccolithophores

COLEOPTERA FOSSIL RECORDS *see* BEETLE RECORDS: Overview; Late Tertiary and Early Quaternary Records; Middle Pleistocene of Western Europe; Late Pleistocene of Australia; Late Pleistocene of Europe; Late Pleistocene of Japan; Late Pleistocene of North America; Late Pleistocene of South America; Late Pleistocene of New Zealand; Late Pleistocene of Northern Asia; Postglacial Europe; Postglacial North America

CORAL STUDIES *see* PALEOCEANOGRAPHY, BIOLOGICAL PROXIES: Corals, Sclerosponges and Mollusks; SEA LEVEL STUDIES: Coral Records of Relative Sea-Level Changes

COSMOGENIC NUCLIDE DATING 407

Overview <i>J C Gosse</i>	407
------------------------------	-----

Methods <i>J M Schaefer and N Lifton</i>	410
---	-----

Cosmic Ray Interactions in Minerals <i>D Lal</i>	418
---	-----

Exposure Geochronology <i>S Ivy-Ochs and F Kober</i>	432
---	-----

Landscape Evolution <i>D E Granger</i>	440
---	-----

CUT-MARKED BONE *see* VERTEBRATE STUDIES: Interactions with Hominids

D

DANSGAARD-OESCHGER EVENTS *see* PALEOCLIMATE RECONSTRUCTION: Paleodroughts and Society

Dating Techniques <i>A J T Jull</i>	447
--	-----

DENDROARCHAEOLOGY *see* PLANT MACROFOSSIL METHODS AND STUDIES: Dendroarchaeology

Dendrochronology <i>B L Coulthard and D J Smith</i>	453
--	-----

Dendroclimatology <i>B H Luckman</i>	459
---	-----

Diatom Introduction	471
<i>V J Jones</i>	
DIATOM METHODS	481
$\delta^{18}\text{O}$ Records	481
<i>M J Leng, P A Barker, G E A Swann, and A M Snelling</i>	
Data Interpretation	489
<i>A Korhola</i>	
Diatomites: Their Formation, Distribution, and Uses	501
<i>R J Flower</i>	
Salinity and Climate Reconstructions from Continental Lakes	507
<i>S C Fritz</i>	
Use in Archaeology	516
<i>N G Cameron</i>	
Diatoms	522
<i>N G Cameron</i>	
DIATOM RECORDS	527
Antarctic Waters	527
<i>C E Stickley, J Pike, and V J Jones</i>	
Freshwater Laminated Sequences	540
<i>H Simola</i>	
Large Lakes	546
<i>A W Mackay</i>	
Diatom Fossil Records from Marine Laminated Sediments	554
<i>J Pike and C E Stickley</i>	
North Atlantic and Arctic	562
<i>N Koç, A Miettinen, and C E Stickley</i>	
Pacific	571
<i>I Koizumi</i>	
Structures and Applications of Biomarkers from Arctic Sea Ice Diatoms	588
<i>S T Belt, G Massé, and M Poulin</i>	
DIATOMS, MARINE <i>see</i> PALEOCEANOGRAPHY, BIOLOGICAL PROXIES: Marine Diatoms	
DINOFLAGELLATES <i>see</i> PALEOCEANOGRAPHY, BIOLOGICAL PROXIES: Dinoflagellates	
DNA, FOSSIL ANIMAL <i>see</i> VERTEBRATE STUDIES: Ancient DNA	
DNA, FOSSIL PLANTS <i>see</i> PALEOBOTANY: Ancient Plant DNA	
DROUGHT HISTORY RECONSTRUCTION <i>see</i> PALEOCLIMATE RECONSTRUCTION: Paleodroughts and Society	
DUNE FIELDS	597
High Latitudes	597
<i>S A Wolfe</i>	
Mid-Latitudes	606
<i>J Sun and D R Muhs</i>	

Low Latitudes <i>N Lancaster</i>	623
-------------------------------------	-----

E

EL NIÑO SOUTHERN OSCILLATION <i>see</i> PALEOCEANOGRAPHY, RECORDS: Postglacial South Pacific; PALEOCLIMATE MODELING: Paleo-ENSO	
ELECTRON SPIN RESONANCE DATING <i>see</i> LUMINESCENCE DATING: Electron Spin Resonance Dating	
EOLIAN SEDIMENTS <i>see</i> PALEOSOLS AND WIND-BLOWN SEDIMENTS: Overview; Nature of Paleosols; Mineral Magnetic Analysis; Soil Micromorphology; Weathering Profiles; Soil Morphology in Quaternary Studies; Biogeochemical Role of Dust in Quaternary Climate Cycles	
Eolian Records, Deep-Sea Sediments <i>D K Rea</i>	637
EQUILIBRIUM LINE ALTITUDE (ELA) RECONSTRUCTION <i>see</i> GLACIAL LANDFORMS, ICE SHEETS: Paleo-ELAs	
ERRATICS, GLACIAL <i>see</i> GLACIAL LANDFORMS, SEDIMENTS: Glacial Erratics and Till Dispersal Indicators	
EXTINCTIONS, QUATERNARY VERTEBRATES <i>see</i> VERTEBRATE RECORDS: Late Pleistocene Megafaunal Extinctions	

F

FELSENMEER (BLOCKFIELDS) <i>see</i> PERMAFROST AND PERIGLACIAL FEATURES: Block/Rock Streams	
Fission-Track Dating <i>J A Westgate, N D Naeser, and B V Alloway</i>	643

FLUVIAL ENVIRONMENTS

Sediments <i>A Aslan</i>	663
Responses to Rapid Environmental Change <i>T E Törnqvist</i>	676
Terrace Sequences <i>D J Merritts</i>	684
Deltaic Environments <i>L Giosan and S L Goodbred</i>	693

FORAMINIFERA STUDIES *see* PALEOCEANOGRAPHY: Paleocyanography An Overview; PALEOCEANOGRAPHY, BIOLOGICAL PROXIES: Temperature Proxies, Census Counts; Benthic Foraminifera; Planktic Foraminifera; PALEOCEANOGRAPHY, PHYSICAL AND CHEMICAL PROXIES: Oxygen-Isotope Stratigraphy of the Oceans

FOSSIL MITES *see* Oribatid Mites

G

Geomagnetic Excursions and Secular Variations <i>A P Roberts and G M Turner</i>	705
--	-----

GLACIAL CLIMATES	721
Biosphere Feedbacks <i>V Brovkin</i>	721
Effects of Atmospheric Dust <i>I Tegen</i>	729
Thermohaline Circulation <i>S Rahmstorf</i>	737
Volcanic and Solar Forcing <i>D T Shindell</i>	748
GLACIAL LANDFORMS	755
Introduction <i>D J A Evans and D I Benn</i>	755
Moraine Forms and Genesis <i>D J A Evans</i>	769
Quaternary Vulcanism: Subglacial Landforms <i>J L Smellie</i>	780
Evidence of Glacier Recession <i>P M Colgan</i>	803
Glacial Landsystems <i>D J A Evans</i>	813
Glaciofluvial Landforms of Erosion <i>A E Kehew, M L Lord, and A L Kozlowski</i>	825
Glacitectonic Structures and Landforms <i>D J A Evans</i>	839
GLACIAL LANDFORMS, EROSIONAL FEATURES	847
Major Scale Forms <i>I S Evans</i>	847
Micro- to Macroscale Forms <i>B R Rea</i>	865
GLACIAL LANDFORMS, ICE SHEETS	877
Growth and Decay <i>M J Siegert</i>	877
Evidence of Glacier and Ice Sheet Extent <i>D M Mickelson and C Winguth</i>	884
Evidence of Glacier Flow Directions <i>C R Stokes</i>	895
Paleo-ELAs <i>A Nesje</i>	909
Trimlines and Paleonunataks <i>C K Ballantyne</i>	918

VOLUME 2**GLACIAL LANDFORMS, SEDIMENTS 1**

Clast Form Analysis 1
D I Benn

Glaciofluvial Landforms of Deposition 6
J L Carrivick and A J Russell

Glaciogenic Lithofacies 18
N Eyles and M Lazorek

Glaciomarine Sediments and Ice-Rafted Debris 30
C Ó Cofaigh

Glaciolacustrine 43
D J A Evans

Micromorphology of Glacial Sediments 52
J F Hiemstra

Tills 62
D J A Evans

Till Fabric Analysis 76
D I Benn

Glacial Erratics and Till Dispersal Indicators 81
D J A Evans

Glacial Sequence Stratigraphy 85
D J A Evans

GLACIAL LANDFORMS, TREE RINGS 91

Dendrogeomorphology 91
H Gärtner and I Heinrich

Dendroglaciology 104
B L Coulthard and D J Smith

Glacial–Interglacial Scale Fluvial Responses 112
M D Blum

GLACIATION, CAUSES 127

Tectonic Uplift and Continental Configurations 127
L A Owen

Astronomical Theory of Paleoclimates 136
A Berger, M-F Loutre, and Q Z Yin

GLACIATIONS 143

Overview 143
J Ehlers and P L Gibbard

Transition from Late Neogene to Early Quaternary Environments 151
M Sarnthein

Early Quaternary (Pleistocene) and Precursors 167
J Ehlers, V Astakhov, P L Gibbard, Ó Ingólfsson, J Mangerud, and J I Svendsen

Middle Pleistocene in Eurasia <i>J Ehlers, V Astakhov, P L Gibbard, J Mangerud, and J I Svendsen</i>	172
Mid-Quaternary in North America <i>C E Jennings, J S Aber, G Balco, R Barendregt, P R Bierman, C W Rovey II, M Roy, L H Thorleifson, and J A Mason</i>	180
Middle Pleistocene Glaciations in the Southern Hemisphere <i>A Coronato and J Rabassa</i>	187
Late Pleistocene Glacial Events in Beringia <i>S A Elias and J Brigham-Grette</i>	191
Late Quaternary of the Southwest Pacific Region <i>D J A Barrell</i>	202
Late Quaternary of Antarctica <i>Ó Ingólfsson</i>	216
Late Pleistocene in Eurasia <i>J Ehlers, V Astakhov, P L Gibbard, J Mangerud, and J I Svendsen</i>	224
Late Quaternary in Highland Asia <i>L A Owen</i>	236
Late Quaternary in North America <i>J T Andrews and A S Dyke</i>	245
Late Pleistocene in South America <i>A Coronato and J Rabassa</i>	250
Neoglaciation in Europe <i>J A Matthews</i>	257
Neoglaciation in the American Cordilleras <i>S C Porter</i>	269

GLOBAL WARMING *see* Paleoclimate Relevance to Global Warming

H

HEINRICH EVENTS *see* PALEOCLIMATE RECONSTRUCTION: Paleodroughts and Society

HISTORICAL CLIMATE RECORDS *see* PALEOCLIMATE RECONSTRUCTION: Historical Climatology; The Last Millennium

HOLOCENE ENVIRONMENTS *see* Dendroclimatology; ARCHAEOLOGICAL RECORDS: Postglacial Adaptations; BEETLE RECORDS: Postglacial Europe; Postglacial North America; CHIRONOMID RECORDS: Postglacial Europe; Postglacial Southern Hemisphere; DIATOM METHODS: Use in Archaeology; GLACIATIONS: Neoglaciation in Europe; Neoglaciation in the American Cordilleras; ICE CORES: Dynamics of the Greenland Ice Sheet; Dynamics of the West Antarctic Ice Sheet; Dynamics of the East Antarctic Ice Sheet; PALEOCEANOGRAPHY, RECORDS: Postglacial Indian Ocean; Postglacial North Atlantic; Postglacial North Pacific; Postglacial South Pacific; PALEOCLIMATE: Timescales of Climate Change; PALEOCLIMATE MODELING: Paleo-ENSO; PALEOCLIMATE RECONSTRUCTION: Historical Climatology; The Last Millennium; PLANT MACROFOSSIL METHODS AND STUDIES: Treeline Studies; PLANT MACROFOSSIL RECORDS: Holocene North America; POLLEN RECORDS, POSTGLACIAL: Africa; Australia and New Zealand; Northeastern North America; Northwestern North America; Southeastern North America; Southwestern North America; South America; Northern Asia; Northern Europe; Southern Europe

HUMAN EVOLUTION *see* ARCHAEOLOGICAL RECORDS: Overview; 2.7 Myr–300 000 Years Ago in Africa; 2.7 Myr–300 000 Years Ago in Asia; 1.9 Myr–300 000 Years Ago in Europe; Human Evolution in the Quaternary

I

ICE CORE METHODS 277

Overview 278
E J Brook

Biological Material 288
J C Priscu, B C Christner, C M Foreman, and G Royston-Bishop

Borehole Temperature Records 298
K M Cuffey

Chronologies 303
J Schwander

CO₂ Studies 311
T Blunier and T M Jenk

Conductivity Studies 319
R Mulvaney

Glaciochemistry 326
K J Kreutz and B G Koffman

Methane Studies 334
J Chappellaz

Microparticle and Trace Element Studies 342
J R McConnell

Stable Isotopes 347
E J Brook

¹⁰Be and Cosmogenic Radionuclides in Ice Cores 353
R Muscheler

Studies of Firn Air 361
C Buizert

ICE CORE RECORDS 373

Africa 373
L G Thompson and M E Davis

Chinese, Tibetan Mountains 379
C P Wake

South America 387
L G Thompson and M E Davis

Antarctic Stable Isotopes 395
E J Brook

Greenland Stable Isotopes 403
B M Vinther and S J Johnsen

Correlations Between Greenland and Antarctica 410
E J Brook

Ice Margin Sites <i>V V Petrenko</i>	416
Thermal Diffusion Paleotemperature Records <i>A M Grachev</i>	431
ICE CORES	435
History of Research, Greenland and Antarctica <i>M Aydin</i>	435
Dynamics of the Greenland Ice Sheet <i>C S Hvidberg, A Svensson, and S L Buchardt</i>	439
Dynamics of the West Antarctic Ice Sheet <i>R Bindshadler</i>	448
Dynamics of the East Antarctic Ice Sheet <i>E D Waddington and C S Lingle</i>	456
History of Carbon Monoxide and Ultra-Trace Gases from Ice Cores <i>M Aydin</i>	463
History of Nitrous Oxide from Ice Cores <i>A Schilt</i>	471
ICE SHEETS, PLEISTOCENE <i>see</i> GLACIAL LANDFORMS, ICE SHEETS: Growth and Decay; Evidence of Glacier and Ice Sheet Extent; Trimlines and Paleonunataks; GLACIATIONS: Early Quaternary (Pleistocene) and Precursors; Middle Pleistocene in Eurasia; Mid-Quaternary in North America; Middle Pleistocene Glaciations in the Southern Hemisphere; Late Pleistocene Glacial Events in Beringia; Late Quaternary of the Southwest Pacific Region; Late Quaternary of Antarctica; Late Pleistocene in Eurasia; Late Quaternary in Highland Asia; Late Quaternary in North America; Late Pleistocene in South America; ICE CORES: Dynamics of the Greenland Ice Sheet; Dynamics of the West Antarctic Ice Sheet; Dynamics of the East Antarctic Ice Sheet	
ICE WEDGES, ICE WEDGE CASTS <i>see</i> PERMAFROST AND PERIGLACIAL FEATURES: Ice Wedges and Ice-Wedge Casts	
K	
K/Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ Dating <i>J R Wijbrans and K F Kuiper</i>	477
L	
LAKE CHEMISTRY RECONSTRUCTION <i>see</i> PALEOLIMNOLOGY: Lake Chemistry	
LAKE LEVEL STUDIES	483
Overview <i>R T Jones and J T Jordan</i>	483
Africa during the Late Quaternary <i>M E Edwards</i>	499
Asia <i>G Yu, B Xue, and Y Li</i>	506
Australia <i>J Magee</i>	524

Latin America <i>S E Metcalfe</i>	531
North America <i>J R Stone and S C Fritz</i>	537
West-Central Europe <i>M Magny</i>	549
Modeling <i>J Vassiljev</i>	558
Lichenometry <i>D P McCarthy</i>	565

LITHOSTRATIGRAPHY *see* QUATERNARY STRATIGRAPHY: Lithostratigraphy

Loess Deposits: Origins and Properties <i>D R Muhs</i>	573
---	-----

LOESS RECORDS 585

Central Asia <i>A E Dodonov</i>	585
China <i>S C Porter</i>	595
Europe <i>D-D Rousseau, E Derbyshire, P Antoine, and C Hatté</i>	606
North America <i>H M Roberts, D R Muhs, and E A Bettis III</i>	620
South America <i>M A Zárate</i>	629

LUMINESCENCE DATING 643

Thermoluminescence <i>O B Lian</i>	643
Optical Dating <i>O B Lian</i>	653
Electron Spin Resonance Dating <i>A J T Jull</i>	667

M

MAGNETIC POLARITY STUDIES *see* Geomagnetic Excursions and Secular Variations

MAMMALIAN EVOLUTION *see* Vertebrate Overview; VERTEBRATE RECORDS: Early Pleistocene; Mid-Pleistocene of Africa; Mid-Pleistocene of Southern Asia; VERTEBRATE STUDIES: Ancient DNA; Speciation and Evolutionary Trends in Quaternary Vertebrates

MARINE ISOTOPE STAGES *see* PALEOCEANOGRAPHY, PHYSICAL AND CHEMICAL PROXIES: Oxygen-Isotope Stratigraphy of the Oceans

MEGAFAUNA, PLEISTOCENE *see* Vertebrate Overview; VERTEBRATE RECORDS: Mid-Pleistocene of Australia; Mid-Pleistocene of North America; Early Pleistocene; Late Pleistocene of Africa;

Late Pleistocene of North America; Late Pleistocene of South America; Late Pleistocene of Southeast Asia; Late Pleistocene Megafaunal Extinctions; Mid-Pleistocene of Africa; Mid-Pleistocene of Europe; Mid-Pleistocene of Southern Asia; VERTEBRATE STUDIES: Interactions with Hominids

MEGAFAUNAL EXTINCTION *see* VERTEBRATE RECORDS: Late Pleistocene Megafaunal Extinctions; VERTEBRATE STUDIES: Interactions with Hominids

METHANE STUDIES, ICE CORES *see* ICE CORE METHODS: Methane Studies

Mg/Ca AND Sr/Ca STUDIES *see* PALEOCEANOGRAPHY, PHYSICAL AND CHEMICAL PROXIES: Mg/Ca and Sr/Ca Paleothermometry from Calcareous Marine Fossils

MICROMORPHOLOGY OF SEDIMENTS *see* GLACIAL LANDFORMS, SEDIMENTS: Micromorphology of Glacial Sediments; PALEOSOLS AND WIND-BLOWN SEDIMENTS: Soil Micromorphology

MIDGES *see* CHIRONOMID RECORDS: Africa; Chironomid Overview; Late Pleistocene of Europe; Postglacial Europe; Postglacial Southern Hemisphere

MILANKOVITCH THEORY *see* GLACIATION, CAUSES: Astronomical Theory of Paleoclimates

MOLLUSKS, MARINE *see* PALEOCEANOGRAPHY, BIOLOGICAL PROXIES: Corals, Sclerosponges and Mollusks

MORAINES, GLACIAL *see* GLACIAL LANDFORMS: Moraine Forms and Genesis; Evidence of Glacier Recession; Glacial Landsystems; Glacitectonic Structures and Landforms; GLACIAL LANDFORMS, SEDIMENTS: Glacial Erratics and Till Dispersal Indicators

MORPHOSTRATIGRAPHY *see* QUATERNARY STRATIGRAPHY: Morphostratigraphy/Allostratigraphy

N

NEANDERTHAL DEMISE *see* ARCHAEOLOGICAL RECORDS: Neanderthal Demise

NEOGLACIATION *see* GLACIATIONS: Neoglaciation in Europe; Neoglaciation in the American Cordilleras

O

OPTICALLY-STIMULATED LUMINESCENCE DATING *see* LUMINESCENCE DATING: Optical Dating

Oribatid Mites

J M Erickson and R B Platt Jr.

680

OXYGEN ISOTOPE STAGES *see* PALEOCEANOGRAPHY, PHYSICAL AND CHEMICAL PROXIES: Oxygen-Isotope Stratigraphy of the Oceans

P

PACKRAT MIDDENS *see* PLANT MACROFOSSIL METHODS AND STUDIES: Rodent Middens

PALEOANTHROPOLOGY *see* ARCHAEOLOGICAL RECORDS: Overview; 2.7 Myr–300 000 Years Ago in Africa; 2.7 Myr–300 000 Years Ago in Asia; 1.9 Myr–300 000 Years Ago in Europe; Global Expansion 300 000–8000 Years Ago, Africa; Global Expansion 300 000–8000 Years Ago, Asia; Global Expansion 300 000–8000 Years Ago, Australia; Global Expansion 300 000–8000 Years Ago, Americas; Human Evolution in the Quaternary; Neanderthal Demise; Postglacial Adaptations

PALEOBOTANY

699

Overview of Terrestrial Pollen Data

699

R H W Bradshaw

Ancient Plant DNA

705

N Wales, R Allaby, E Willerslev, and M T P Gilbert

Charred Particle Analyses <i>K J Brown and M J Power</i>	716
Paleophytogeography <i>A E Bjune</i>	730
Silicon Isotopes in Diatoms <i>J J Tyler</i>	734
PALEOCEANOGRAPHY	745
Paleoceanography An Overview <i>D M Anderson and K E Lee</i>	745
PALEOCEANOGRAPHY, BIOLOGICAL PROXIES	755
Alkenone Paleothermometry Based on the Haptophyte Algae <i>S L Ho, B D A Naafs, and F Lamy</i>	755
Benthic Foraminifera <i>R Saraswat and R Nigam</i>	765
Biomarker Indicators of Past Climate <i>J P Sachs, K Pahnke, R Smittenberg, and Z Zhang</i>	775
Coccolithophores <i>J-A Flores and F J Sierro</i>	783
Corals, Sclerosponges and Mollusks <i>T M Quinn and B R Schöne</i>	795
Dinoflagellates <i>A de Vernal, A Rochon, and T Radi</i>	800
Marine Diatoms <i>F Abrantes and I M Gil</i>	816
Planktic Foraminifera <i>H J Dowsett and M M Robinson</i>	825
Radiolarians and Silicoflagellates <i>D Lazarus</i>	830
Temperature Proxies, Census Counts <i>J D Ortiz</i>	841
PALEOCEANOGRAPHY, PHYSICAL AND CHEMICAL PROXIES	849
Carbon Cycle Proxies ($\delta^{11}\text{B}$, $\delta^{13}\text{C}_{\text{calcite}}$, $\delta^{13}\text{C}_{\text{organic}}$, Shell Weights, B/Ca, U/Ca, Zn/Ca, Ba/Ca) <i>B Hönisch and K A Allen</i>	849
Dissolution of Deep-Sea Carbonates <i>S Barker</i>	859
Mg/Ca and Sr/Ca Paleothermometry from Calcareous Marine Fossils <i>Y Rosenthal and B Linsley</i>	871
Magnetic Proxies and Susceptibility <i>R G Hatfield and J S Stoner</i>	884
Nutrient Proxies <i>T M Marchitto</i>	899

Oxygen-Isotope Stratigraphy of the Oceans <i>F C Bassinot</i>	907
Oxygen Isotope Composition of Seawater <i>E J Rohling</i>	915
Radioisotope Proxies <i>R Francois</i>	923
Salinity Proxies $\delta^{18}\text{O}$ <i>P D Naidu</i>	932
Terrigenous Sediments <i>A M Franzese and S R Hemming</i>	941

VOLUME 3

PALEOCEANOGRAPHY, RECORDS	1
Early Pleistocene <i>T de Garidel-Thoron</i>	1
Late Pleistocene North Atlantic <i>D M Anderson</i>	9
Late Pleistocene South Atlantic <i>S Mulitza, A Paul, and G Wefer</i>	18
Late Pleistocene North Pacific <i>T Kiefer</i>	33
Postglacial Indian Ocean <i>P D Naidu</i>	46
Postglacial North Atlantic <i>M W Kerwin and K A Huguen</i>	55
Postglacial North Pacific <i>J I Martínez</i>	62
Postglacial South Pacific <i>F Lamy and R de Pol-Holz</i>	73
PALEOCLIMATE	87
Introduction <i>C J Mock</i>	87
Timescales of Climate Change <i>P J Bartlein</i>	93
Modern Analog Approaches in Paleoclimatology <i>C J Mock and J J Shinker</i>	102
Paleoclimate History of the Arctic <i>G H Miller, J Brigham-Grette, R B Alley, L Anderson, H A Bauch, M S V Douglas, M E Edwards, S A Elias, B P Finney, J J Fitzpatrick, S V Funder, A Geirsdóttir, T D Herbert, L D Hinzman, D S Kaufman, G M MacDonald, L Polyak, A Robock, M C Serreze, J P Smol, R Spielhagen, J W C White, A P Wolfe, and E W Wolff</i>	113
The Younger Dryas Climate Event <i>A E Carlson</i>	126

PALEOCLIMATE MODELING	135
Data–Model Comparisons <i>S P Harrison</i>	135
Quaternary Environments <i>C S Jackson</i>	147
The Last Interglacial <i>M Montoya</i>	155
Last Glacial Maximum GCMs <i>A J Broccoli</i>	165
Paleo-ENSO <i>C J Mock</i>	171
 PALEOCLIMATE RECONSTRUCTION	 179
Approaches <i>B N Shuman</i>	179
Pliocene Environments <i>R Z Poore</i>	185
Paleodroughts and Society <i>C J Mock</i>	194
Sub-Milankovitch (DO/Heinrich) Events <i>L Labeyrie, L Skinner, and E Cortijo</i>	200
Paleotempestology <i>K-b Liu</i>	209
Younger Dryas Oscillation, Global Evidence <i>S Björck</i>	222
The Last Millennium <i>M E Mann</i>	229
Historical Climatology <i>M Chenoweth</i>	237
 Paleoclimate Relevance to Global Warming <i>S H Schneider and M D Mastrandrea</i>	 244
Paleohydrology <i>V R Thorndycraft</i>	253
 PALEOLIMNOLOGY	 259
Overview of Paleolimnology <i>M S V Douglas</i>	259
Cladocera <i>M Rautio and L Nevalainen</i>	271
Freshwater Mollusks <i>C G De Francesco</i>	281
Lake Chemistry <i>D Antoniadis</i>	292

Physical Properties of Lake Sediments <i>K R Hodder and R Gilbert</i>	300
Contributions of Paleolimnological Research to Biogeography <i>K A Moser</i>	313
Pigment Studies <i>S McGowan</i>	326
Multiproxy Approaches <i>N Michelutti and J P Smol</i>	339
Visible and Infrared Spectroscopical Applications <i>P Rosén and H Vogel</i>	349
PALEOLITHIC <i>see</i> ARCHAEOLOGICAL RECORDS: Overview; 2.7 Myr–300 000 Years Ago in Africa; 2.7 Myr–300 000 Years Ago in Asia; 1.9 Myr–300 000 Years Ago in Europe; Global Expansion 300 000–8000 Years Ago, Africa; Global Expansion 300 000–8000 Years Ago, Asia; Global Expansion 300 000–8000 Years Ago, Australia; Global Expansion 300 000–8000 Years Ago, Americas; Neanderthal Demise	
PALEOSOLS AND WIND-BLOWN SEDIMENTS	357
Overview <i>D R Muhs</i>	357
Nature of Paleosols <i>J A Mason and P M Jacobs</i>	367
Mineral Magnetic Analysis <i>M J Singer and K L Verosub</i>	375
Soil Micromorphology <i>R A Kemp</i>	381
Weathering Profiles <i>E A Bettis III</i>	392
Soil Morphology in Quaternary Studies <i>L McFadden</i>	402
Biogeochemical Role of Dust in Quaternary Climate Cycles <i>K E Kohfeld</i>	412
PATTERNED GROUND <i>see</i> PERMAFROST AND PERIGLACIAL FEATURES: Patterned Ground	
PEAT STUDIES <i>see</i> PLANT MACROFOSSIL METHODS AND STUDIES: Mire and Peat Macros	
PEDOSTRATIGRAPHY <i>see</i> QUATERNARY STRATIGRAPHY: Pedostratigraphy	
PERMAFROST AND PERIGLACIAL FEATURES	421
Active Layer Processes <i>N I Shiklomanov and F E Nelson</i>	421
Cryoturbation Structures <i>J Vandenberghe</i>	430
Ice Wedges and Ice-Wedge Casts <i>J Murton</i>	436
Patterned Ground <i>C K Ballantyne</i>	452

Permafrost <i>C R Burn</i>	464
Frost Mounds: Active and Relict Forms <i>N Ross</i>	472
Slope Deposits and Forms <i>C Harris</i>	481
Periglacial Fluvial Sediments and Forms <i>J van Huissteden, J Vandenberghe, P L Gibbard, and J Lewin</i>	490
Rock Weathering <i>J Murton</i>	500
Permafrost and Glacier Interactions <i>R I Waller</i>	507
Block/Rock Streams <i>P Wilson</i>	514
Blockfields (Felsenmeer) <i>B R Rea</i>	523
Rock Glaciers and Protalus Forms <i>A Kääb</i>	535
Yedoma: Late Pleistocene Ice-Rich Syngenetic Permafrost of Beringia <i>L Schirrmeister, D Froese, V Tumskoy, G Grosse, and S Wetterich</i>	542
Paraglacial Geomorphology <i>C K Ballantyne</i>	553
Talus Slopes <i>B H Luckman</i>	566
Thermokarst Topography <i>C R Burn</i>	574
PERMAFROST HISTORY <i>see</i> PERMAFROST AND PERIGLACIAL FEATURES: Active Layer Processes; Cryoturbation Structures; Patterned Ground; Permafrost	
Phytoliths <i>M S Blinnikov</i>	582
PIGMENTS, FOSSIL <i>see</i> PALEOLIMNOLOGY: Pigment Studies	
PINGOS <i>see</i> PERMAFROST AND PERIGLACIAL FEATURES: Frost Mounds: Active and Relict Forms	
Plant Macrofossil Introduction <i>H H Birks</i>	593
PLANT MACROFOSSIL METHODS AND STUDIES	613
CO ₂ Reconstruction from Fossil Leaves <i>M Rundgren</i>	613
Megafossils <i>G M MacDonald</i>	621
Dendroarchaeology <i>R H Towner</i>	630

Mire and Peat Macros <i>D Mauquoy and B van Geel</i>	637
Paleolimnological Applications <i>M-J Gaillard and H H Birks</i>	657
Rodent Middens <i>S A Elias</i>	674
Surface Samples, Taphonomy, Representation <i>A C Dieffenbacher-Krall</i>	684
Treeline Studies <i>W Tinner</i>	690
Use in Environmental Archaeology <i>S Jacomet</i>	699
Validation of Pollen Studies <i>S T Jackson and R K Booth</i>	725
PLANT MACROFOSSIL RECORDS	733
Arctic Eurasia <i>F Kienast</i>	733
Arctic North America <i>N H Bigelow, G D Zazula, and D E Atkinson</i>	746
Greenland <i>O Bennike</i>	760
Holocene North America <i>R G Baker</i>	768
Late Glacial Multidisciplinary Studies <i>B Ammann, H H Birks, A Walanus, and K Wasylkowa</i>	785
PLATE TECTONICS <i>see</i> GLACIAL LANDFORMS: Glacitectonic Structures and Landforms; GLACIATION, CAUSES: Tectonic Uplift and Continental Configurations; GLACIATIONS: Late Quaternary in North America	
PLIOCENE ENVIRONMENTS <i>see</i> PALEOCLIMATE RECONSTRUCTION: Pliocene Environments	
Pollen Analysis, Principles <i>H Seppä</i>	794
POLLEN METHODS AND STUDIES	805
Use of Pollen as Climate Proxies <i>S Brewer, J Guiot, and D Barboni</i>	805
Reconstructing Past Biodiversity Development <i>B V Odgaard</i>	816
Numerical Analysis Methods <i>H J B Birks</i>	821
Databases and Their Application <i>E C Grimm, R H W Bradshaw, S Brewer, S Flantua, T Giesecke, A-M Lézine, H Takahara, and J W Williams</i>	831
Surface Samples and Trapping <i>A Poska</i>	839

Stand-Scale Palynology <i>R H W Bradshaw</i>	846
Changing Plant Distributions and Abundances <i>T Giesecke</i>	854
The Biome Approach to Reconstructing Past Vegetation <i>M E Edwards</i>	861
POLLSCAPE Model: Simulation Approach for Pollen Representation of Vegetation and Land Cover <i>S Sugita</i>	871
Archaeological Applications <i>M-J Gaillard</i>	880

VOLUME 4

Pollen Records, Last Interglacial of Europe <i>C Tzedakis</i>	1
--	---

POLLEN RECORDS, LATE PLEISTOCENE 9

Africa <i>M E Meadows and B M Chase</i>	9
Australasia <i>P Kershaw and S van der Kaars</i>	18
Northern Asia <i>A V Lozhkin and P M Anderson</i>	27
Northern North America <i>N H Bigelow</i>	39
South America <i>H Hooghiemstra and J C Berrio</i>	52
Middle and Late Pleistocene in Southern Europe <i>J-L de Beaulieu, P C Tzedakis, V Andrieu-Ponel, and F Guiter</i>	63
Western North America <i>R S Thompson</i>	72

POLLEN RECORDS, POSTGLACIAL 85

Africa <i>A-M Lézine</i>	85
Australia and New Zealand <i>J R Dodson</i>	104
Northeastern North America <i>J W Williams and B N Shuman</i>	115
Northwestern North America <i>D G Gavin and F S Hu</i>	124
Southeastern North America <i>D A Willard</i>	133
Southwestern North America <i>P E Wigand</i>	142

South America <i>H Behling</i>	156
Northern Asia <i>A A Andreev and P E Tarasov</i>	164
Northern Europe <i>M J Bunting</i>	173
Southern Europe <i>L Sadori</i>	179

POTASSIUM-ARGON DATING *see* K/Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ Dating

Q

QUATERNARY STRATIGRAPHY 189

Overview <i>B Pillans</i>	189
Continental Biostratigraphy <i>T van Kolfschoten</i>	206
Chronostratigraphy <i>B Pillans</i>	215
Climatostratigraphy <i>P L Gibbard</i>	222
Lithostratigraphy <i>W E Westerhoff and H J T Weerts</i>	227
Morphostratigraphy/Allostratigraphy <i>P D Hughes</i>	243
Pedostratigraphy <i>A Palmer</i>	250
Sequence Stratigraphy <i>T R Naish, S T Abbott, and R M Carter</i>	260
Tephrochronology <i>B V Alloway, D J Lowe, G Larsen, P A R Shane, and J A Westgate</i>	277

R

RADIOLARIAN STUDIES *see* PALEOCEANOGRAPHY, BIOLOGICAL PROXIES: Radiolarians and Silicoflagellates

RADIOCARBON DATING 305

Conventional Method <i>G T Cook and J van der Plicht</i>	305
AMS Radiocarbon Dating <i>A J T Jull</i>	316
Sources of Error <i>E M Scott</i>	324

Variations in Atmospheric ^{14}C <i>J van der Plicht</i>	329
Causes of Temporal ^{14}C Variations <i>G S Burr</i>	336
Calibration of the ^{14}C Record <i>P J Reimer, R W Reimer, and M Blaauw</i>	345
Charcoal <i>M I Bird</i>	353
^{14}C of Plant Macrofossils <i>C Hatté and A J T Jull</i>	361

ROCK GLACIERS *see* PERMAFROST AND PERIGLACIAL FEATURES: Rock Glaciers and Protalus Forms

RODENT MIDDENS *see* PLANT MACROFOSSIL METHODS AND STUDIES: Rodent Middens

S

SEA LEVEL STUDIES 369

Overview <i>I Shennan</i>	369
Geomorphological Indicators <i>P A Pirazzoli</i>	377
Sedimentary Indicators of Relative Sea-Level Changes – High Energy <i>J A G Cooper</i>	385
Sedimentary Indicators of Relative Sea-Level Changes – Low Energy <i>R J Edwards</i>	396
Coral Records of Relative Sea-Level Changes <i>C D Woodroffe</i>	409
Microfossil-Based Reconstructions of Holocene Relative Sea-Level Change <i>W R Gehrels</i>	419
Eustatic Sea-Level Changes – Glacial–Interglacial Cycles <i>C V Murray-Wallace</i>	429
Eustatic Sea-Level Changes Since the Last Glacial Maximum <i>P L Whitehouse and S L Bradley</i>	439
Isostasy: Glaciation-Induced Sea-Level Change <i>G Milne and I Shennan</i>	452
Use of Cave Data in Sea-Level Reconstructions <i>A Dutton</i>	460

SEA-LEVELS, LATE QUATERNARY 467

Late Quaternary Relative Sea-Level Changes in High Latitudes <i>C Ó Cofaigh and M J Bentley</i>	467
Late Quaternary Sea-Level Changes in Greenland <i>S A Woodroffe and A J Long</i>	481
Late Quaternary Relative Sea-Level Changes at Mid-Latitudes <i>A C Kemp, B P Horton, and S E Engelhart</i>	489

Late Quaternary Relative Sea-Level Changes in the Tropics <i>Y Zong</i>	495
Tectonics and Relative Sea-Level Change <i>A R Nelson</i>	503

SEA SURFACE TEMPERATURE RECONSTRUCTION *see* PALEOCEANOGRAPHY: Paleocceanography An Overview; PALEOCEANOGRAPHY, BIOLOGICAL PROXIES: Alkenone Paleothermometry Based on the Haptophyte Algae; Biomarker Indicators of Past Climate; Coccolithophores; Dinoflagellates; Marine Diatoms; Planktic Foraminifera; Temperature Proxies, Census Counts; PALEOCEANOGRAPHY, PHYSICAL AND CHEMICAL PROXIES: Carbon Cycle Proxies ($\delta^{11}\text{B}$, $\delta^{13}\text{C}_{\text{calcite}}$, $\delta^{13}\text{C}_{\text{organic}}$, Shell Weights, B/Ca, U/Ca, Zn/Ca, Ba/Ca); Mg/Ca and Sr/Ca Paleothermometry from Calcareous Marine Fossils; Oxygen-Isotope Stratigraphy of the Oceans; PALEOCEANOGRAPHY, RECORDS: Late Pleistocene North Atlantic; Late Pleistocene North Pacific; Late Pleistocene South Atlantic; Postglacial Indian Ocean; Postglacial North Atlantic; Postglacial North Pacific; Postglacial South Pacific

SILICOFLAGELLATES *see* PALEOCEANOGRAPHY, BIOLOGICAL PROXIES: Radiolarians and Silicoflagellates

SPELEOTHEMS *see* CARBONATE STABLE ISOTOPES: Speleothems

STABLE ISOTOPE STUDIES, CARBONATES *see* CARBONATE STABLE ISOTOPES: Overview; Speleothems; Terrestrial Teeth and Bones; Nonmarine Biogenic Carbonates; Terrestrial Organic Materials; Non-Lacustrine Terrestrial Studies; Lake Sediments

STABLE ISOTOPE STUDIES, DEEP SEA RECORDS *see* PALEOCEANOGRAPHY, PHYSICAL AND CHEMICAL PROXIES: Oxygen-Isotope Stratigraphy of the Oceans; Oxygen Isotope Composition of Seawater; Salinity Proxies $\delta^{18}\text{O}$; PALEOCEANOGRAPHY, RECORDS: Early Pleistocene; Late Pleistocene North Atlantic; Late Pleistocene South Atlantic; Postglacial Indian Ocean; Postglacial North Atlantic; Postglacial North Pacific; Postglacial South Pacific; ICE CORE METHODS: Stable Isotopes; ICE CORE RECORDS: Antarctic Stable Isotopes; Greenland Stable Isotopes; Correlations Between Greenland and Antarctica

STABLE ISOTOPE STUDIES, ICE CORES *see* ICE CORE METHODS: Stable Isotopes; ICE CORE RECORDS: Antarctic Stable Isotopes; Greenland Stable Isotopes; Correlations Between Greenland and Antarctica

SUB-MILANKOVITCH EVENTS *see* PALEOCLIMATE RECONSTRUCTION: Sub-Milankovitch (DO/Heinrich) Events

T

TALUS SLOPES *see* PERMAFROST AND PERIGLACIAL FEATURES: Talus Slopes

TEETH AND BONES, STABLE ISOTOPE STUDIES *see* CARBONATE STABLE ISOTOPES: Terrestrial Teeth and Bones

TEPHROCHRONOLOGY *see* QUATERNARY STRATIGRAPHY: Tephrochronology

THERMOHALINE CIRCULATION OF THE OCEANS *see* GLACIAL CLIMATES: Thermohaline Circulation

THERMOLUMINESCENCE DATING *see* LUMINESCENCE DATING: Thermoluminescence

TILL, GLACIAL *see* GLACIAL LANDFORMS, SEDIMENTS: Till Fabric Analysis; Tills; Glacial Erratics and Till Dispersal Indicators

TREE RINGS *see* Dendrochronology; Dendroclimatology; GLACIAL LANDFORMS, TREE RINGS: Dendrogeomorphology; Dendroglaciology; PLANT MACROFOSSIL METHODS AND STUDIES: Dendroarchaeology

TREELINE RECONSTRUCTION *see* POLLEN METHODS AND STUDIES: Archaeological Applications

U

USE OF QUATERNARY PROXIES IN FORENSIC SCIENCE 522

Use of Geology in Forensic Science: Search to Locate Burials 522
L Donnelly

Soils and Sediment 535
R C Murray

The Use of Macroscopic Plant Remains in Forensic Science 542
J H Bock

Insects 548
D E Gennard

Analytical Techniques in Forensic Palynology 556
V M Bryant

U-Series Dating 567
W G Thompson

V

Varved Lake Sediments 573
B Zolitschka

Varved Marine Sediments 582
K A Hughen and B Zolitschka

Vertebrate Overview 590
D C Schreve

VERTEBRATE RECORDS 599

Early Pleistocene 599
L Rook, M Delfino, M P Ferretti, and L Abbazzi

Early and Middle Pleistocene of Northern Eurasia 605
I Vislobokova and A Tesakov

Mid-Pleistocene of Africa 615
L C Bishop and A Turner

Mid-Pleistocene of Australia 621
G J Prideaux

Mid-Pleistocene of Europe 639
R Sardella

Mid-Pleistocene of North America 646
C J Bell

Mid-Pleistocene of Southern Asia 651
J de Vos

Late Pleistocene of Africa 664
T E Steele

Late Pleistocene of North America 673
J I Mead

Late Pleistocene of South America <i>M Ubilla</i>	680
Late Pleistocene of Southeast Asia <i>Y Chaimanee</i>	693
Late Pleistocene Megafaunal Extinctions <i>S A Elias and D C Schreve</i>	700
Late Pleistocene Mummified Mammals <i>C R Harington</i>	713
VERTEBRATE STUDIES	719
Ancient DNA <i>I Barnes, R Barnett, and B Shapiro</i>	719
Speciation and Evolutionary Trends in Quaternary Vertebrates <i>A M Lister</i>	723
Dwarfing and Gigantism in Quaternary Vertebrates <i>M R Palombo and R Rozzi</i>	733
Interactions with Hominids <i>K V Boyle</i>	748
VOLCANIC ASH <i>see</i> QUATERNARY STRATIGRAPHY: Tephrochronology	
VULCANISM <i>see</i> GLACIAL CLIMATES: Volcanic and Solar Forcing; GLACIAL LANDFORMS: Quaternary Vulcanism: Subglacial Landforms; QUATERNARY STRATIGRAPHY: Tephrochronology	
W	
WIND-BLOWN SEDIMENTS (LOESS, SAND DUNES) <i>see</i> Eolian Records, Deep-Sea Sediments; Loess Deposits: Origins and Properties; DUNE FIELDS: High Latitudes; Mid-Latitudes; Low Latitudes; LOESS RECORDS: Central Asia; China; Europe; North America; South America	
Y	
YOUNGER DRYAS OSCILLATION <i>see</i> PALEOCLIMATE RECONSTRUCTION: Younger Dryas Oscillation, Global Evidence	
<i>Index</i>	757

Introductory Articles

Contents

History of Quaternary Science

History of Dating Methods

Societal Relevance of Quaternary Research

Understanding Quaternary Climatic Change

History of Quaternary Science

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Introduction

The Quaternary sciences represent the systematic study of the Quaternary or the most recent geologic period. This period is generally characterized by a series of glaciations, or Ice Ages, interspersed with relatively warm, interglacial intervals, such as the current interglacial, the Holocene. The study of Quaternary environments began in the late eighteenth century. Quaternary geology and paleontology came of age in the nineteenth century, and other important aspects of Quaternary science, such as paleoceanography, paleoecology, and paleoclimatology, developed to a much greater extent in the twentieth century. As with many branches of science, the pioneers in Quaternary studies had to work hard to overcome many widely held, erroneous ideas from previous generations of scholars.

At the beginning of the nineteenth century, science itself was rapidly changing. Up until that time, university professors and other scholars who performed scientific research had mostly been generalists who dabbled in many different fields. They looked upon themselves as natural historians, studying the workings of the natural world as their whimsy led them. The early nineteenth century saw the beginnings of specialization in science. As the level of scientific knowledge was rapidly increasing, it was no longer possible for individual scholars to keep abreast of all the new discoveries. People began to devote their time and energy to one or just a few lines of research. This new, focused style of scientific study brought great leaps forward for science as a whole, and for Quaternary science in particular, as we shall see in this article.

Establishing the Geologic Framework

The term 'Quaternary' was coined by an Italian mining engineer, Giovanni Arduino (1714–95). He distinguished four

orders of strata comprising all of Earth's history: Primary, Secondary, Tertiary, and Quaternary (Schneer, 1969: 10). Arduino (Figure 1) distinguished four separate stages or 'orders,' which he recognized on the basis of very large strata arranged one above the other.

These four 'orders' were expressed regionally in Italy, as the Atesine Alps, the Alpine foothills, the sub-Alpine hills, and the Po River plain, respectively. The term 'Quaternary' apparently was not used again until the French geologist, Desnoyers (1829), used it to differentiate Tertiary from Younger strata in the Paris basin. It was redefined by another Frenchman, Reboul (1833), to include all strata containing extant flora and fauna.

The Quaternary period, as we now know it, is divided into two epochs: the Pleistocene and the Holocene. The history of these terms is likewise complicated. The term 'Pleistocene' was coined by Scottish geologist, Charles Lyell (1839; Figure 2), to replace his previous term 'Newer Pliocene' (1833).

Lyell defined the Pleistocene as the 'most recent' geologic era and further specified that Pleistocene rocks and sediments are characterized by their fossil content of 90% mollusks that are recognized as living species. As glacial theory began to take shape (see below), Forbes (1846) redefined the Pleistocene as equivalent to the 'Glacial Epoch.' Then, Hörnes (1853) introduced the term 'Neogene' to include Lyell's Miocene and Pliocene. In response, Lyell (1873) specified that the term Pleistocene should be used 'as strictly synonymous with post-Pliocene.' In the same publication, Lyell also separated the Pleistocene (glacial) from the 'Recent' (postglacial) time. The term 'Recent' was later replaced by the term 'Holocene' by Gervais (1867–69).

Thus, by the end of the nineteenth century, the stratigraphic nomenclature of the Quaternary period had been firmly established. However, no one knew when the Tertiary ended and the Quaternary began. In geology, it is standard procedure to



Figure 1 Giovanni Arduino (1714–95).

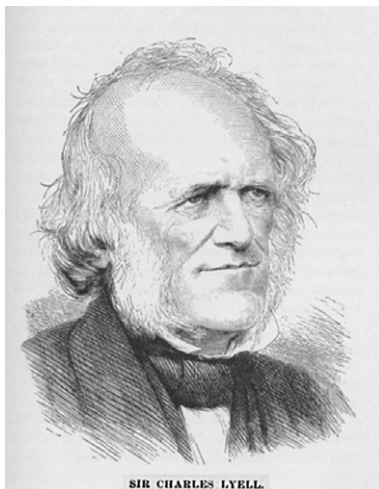


Figure 2 Charles Lyell (1797–1875).

designate a type locality that typifies the boundaries between major stratigraphic units. The 18th International Geological Congress (London, 1948) resolved to select a type locality for the Pliocene–Pleistocene (Tertiary–Quaternary) boundary. After three decades of deliberations, the Vrica section in Calabria, southern Italy, was finally selected. Hence, the Plio–Pleistocene boundary was established at this site, where the boundary falls at approximately 1.64 Ma (Aguirre and Pasini, 1985; Bassett, 1985). Hilgen (1991) calibrated this age, on the basis of an orbital forcing chronology, to an age of 1.81 Ma. These age designations were made possible only through the invention of radiometric dating methods, which came about in the latter half of the twentieth century (see below).

The Discovery of Pleistocene Mammals

The threads of research that eventually led to modern Quaternary science came from a variety of disciplines and were driven



Figure 3 Georges Cuvier (1769–1832).

by scientific observations in a number of fields. One of these was the field of vertebrate paleontology. As with many branches of science, pivotal discoveries often launch major lines of research. One such discovery was made at a Pleistocene site in Kentucky, called Big Bone Lick. The site lies on a tributary of the Ohio River, about 30 km southwest of Cincinnati, Ohio. It was the first major New World fossil locality known to Europeans. Baron Charles de Lougueuil, the commander of a French military expedition, may have been the first European to visit the site in 1739. He collected some mastodon fossils that were later studied by the French naturalists, Daubenton, Buffon, and Cuvier. Cuvier (1825) published a description of the Big Bone Lick mastodon remains. Inspired by this and other Pleistocene fossil discoveries, Cuvier developed his theory of global cooling that led to the extinction of these strange beasts.

In 1807, at the behest of Thomas Jefferson, William Clark conducted a major collecting expedition at Big Bone Lick that yielded about 300 specimens, most of which can still be found at either the National Museum of Natural History in Paris or the Academy of Natural Sciences in Philadelphia. Thus, the fossils from this one site helped to launch Pleistocene vertebrate paleontology in two continents. The discovery of mastodon and other large Pleistocene mammal remains at this site sparked the imagination of scientists and politicians alike. In 1803, the United States purchased the Louisiana Territory from France. This territory included more than 2 million square kilometers of land extending from the Mississippi River to the Rocky Mountains. When President Thomas Jefferson sent Meriwether Lewis and William Clark to explore and map this new American territory, he expected that they might find living specimens of mastodon and other large Pleistocene mammals roaming the uncharted wilderness of the West. Jefferson was an avid naturalist and took great interest in the fossil bones from Big Bone Lick.

Based on discoveries such as these, the field of vertebrate paleontology was starting to take shape during the late eighteenth and early nineteenth centuries. As discussed earlier, one of the most important leaders in this newly emerging field was the French scientist, Georges Cuvier (Figure 3). At the start of the nineteenth century, Cuvier was a professor of animal

anatomy at the Musée National d'Histoire Naturelle (National Museum of Natural History) in Paris.

An opponent of the theory of evolution, Cuvier's most important contribution to science was the establishment of the extinction of ancient species, based on fossil records. Until the nineteenth century, most philosophers and natural historians had rejected the idea that some species had died out and that new species had evolved over time. Most Europeans held to a strict, literal interpretation of the Bible, which dictated that the Earth was created in just 6 days, only a few thousand years ago. But the fossil record that was just beginning to be unearthed by a handful of paleontologists began to challenge this view.

Although Cuvier remained a Creationist, the fossils he was describing were reshaping his views on the nature of that creation. Cuvier believed in the great antiquity of the Earth and held that more or less modern conditions had been in existence for most of Earth's history. However, in order to explain the extinction of species Cuvier had seen in the fossil record, he invoked periodic 'revolutions' in Earth's history. Each 'revolution' was a natural event that had brought about the extinction of a number of species. Unlike others of his time (notably, the Reverend William Buckland, who invoked the Biblical Flood), Cuvier did not equate these 'revolutions' with Biblical or historical events.

Cuvier considered that the last great 'revolution,' the one that had brought about the extinction of such spectacular animals as mammoths and mastodons, might have been a severe and sudden cooling of the planet. Louis Agassiz (Figure 4) took this idea and developed it further, into the concept of a 'Great Ice Age.'

Agassiz was a Swiss naturalist who started his career as Cuvier's assistant. Agassiz thought that mammoths and other extinct mammals must have been adapted to a tropical climate. Here is how he described their demise in the face of the oncoming Ice Age:

The gigantic quadrupeds, the Mastodons, Elephants, Tigers, Lions, Hyenas, Bears, whose remains are found in Europe from its southern promontories to the northernmost limits of Siberia and Scandinavia ... may indeed be said to have possessed the earth in those days. But their reign was over. A sudden intense winter, that was also to last for ages, fell upon our globe; it spread over the very countries where these tropical animals had their homes, and so suddenly did it come upon them that they were embalmed beneath masses of snow and ice, without time even for the decay which follows death (Agassiz, 1866: 208).

The Discovery of Pleistocene Glaciations

Agassiz's theory of the 'Great Ice Age' was first presented to the Swiss Society of Natural Sciences in Neuchâtel in 1837. This was an ideal setting in which to convince geologists and natural historians, because Agassiz could demonstrate the effects of glacial ice in the landscapes of the Alps. He pointed to large boulders that had been transported by ice (glacial erratics), piles of rocks moved by glacial ice (glacial moraines), and scratched surface lines in bedrock, made by the passage of glacial ice and debris. Agassiz published his theory in the books, *Étude sur les glaciers*, in 1840, and *Système glaciaire*, in 1847. These books summarized his findings from Europe. He later found even more evidence of glaciation in North America. Agassiz's theory was initially rejected by many leading geologists, who still held to the idea that the transportation of surficial sediments was due mainly to the effects of the Biblical Flood. Agassiz's ideas on the glaciation eventually won the day, but his ideas about the nature of the Pleistocene megafauna turned out to be largely nonsensical. Far from being tropically adapted animals, the mammoths, mastodons, and other ice-age mammals of Europe were adapted to the very same glacial environments to which Agassiz had ascribed their demise. Most of these animals died out during the transition to warm climate at the end of the last glaciation, not at its beginning.

Evidence for glaciation had been seen by some of Agassiz's contemporaries in other parts of Europe. For instance, Esmark noted the existence of glacial deposits in Norway, Bernhardt found evidence for glaciation in Germany, and de Venetz and Charpentier found evidence for the advance of glacial ice far beyond the limits of modern alpine glaciers in Switzerland. Agassiz himself traveled to Britain and North America and argued that surficial deposits that had previously been considered flood deposits should be reclassified as glacial.

Convinced by Agassiz's ice-age theory, field geologists of the middle and late nineteenth century began looking for evidence to help reconstruct the actual history of glacial events. Agassiz had proposed a single, massive glacial event in which ice sheets covered much of the middle latitudes, as well as the high latitudes of the Earth. Evidence started accumulating that pointed to multiple glaciations, separated by warm periods. By the 1850s, evidence was pointing toward at least two major glaciations in Europe. By 1877, James Geikie (Figure 5) had developed the concept of four or five large glaciations during the Pleistocene on the basis of stratigraphic evidence.

Evidence from North America made it clear that the last glaciation had not been the largest one, because it had not entirely destroyed the evidence for earlier, larger glaciations.

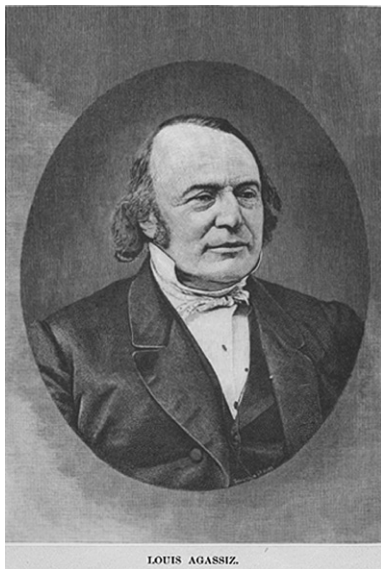


Figure 4 Louis Agassiz (1807–73).

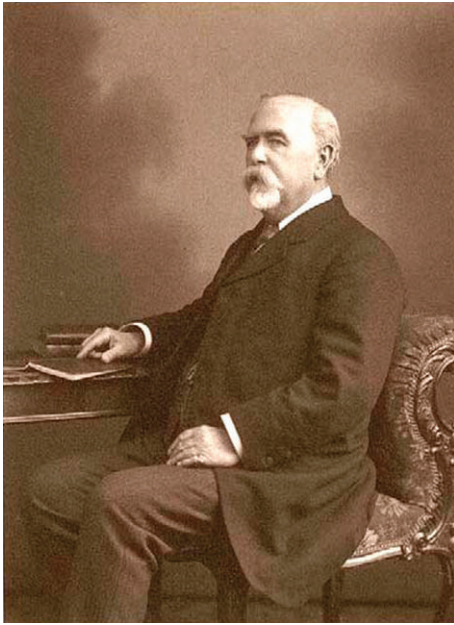


Figure 5 James Geikie (1839–1915).

Geologists coined the terms ‘Nebraskan,’ ‘Kansan,’ ‘Illinoian,’ and ‘Wisconsinan,’ to describe a sequence of four glacial epochs in North America. These were separated by three warm or interglacial periods (the Aftonian, Yarmouthian, and Sangamon) on the basis of the presence of ancient soils buried between glacial deposits.

Pioneering work on establishing the European glacial sequence was carried out by Albrecht Penck and Eduard Brückner (Figure 6), who identified four glaciations, the Günz, Mindel, Riss, and Würm.

These glaciations were named after four rivers in southern Germany. Penck and Brückner’s (1909) work was based on the identification of the stratigraphic sequence of river terraces in the valleys of the northern Alps (Figure 7). In many parts of the world, diligent field studies in the last century have failed to find evidence for more than four glaciations on land.

The ways in which these glaciations were recognized varied from one part of the world to another. In Europe, only the ice advances that reached farther south than the younger ones were recognized as separate glaciations. The traces of any intermediate ice expansion were essentially overridden and destroyed by subsequent larger glacial advances. American glaciations were originally defined as times when the ice sheets extended south to the American Midwest. Interglacials were the times when the Midwest region was free of ice. The classical North American Pleistocene subdivision is one of long interglacials and short glacials, whereas the North European system recognizes short interglacials and long glacials (Kukla, 2005). Penck and Brückner’s Alpine glaciation scheme was the most widely used system of classification in the twentieth century for the correlation of Pleistocene deposits between continents (Flint, 1971).

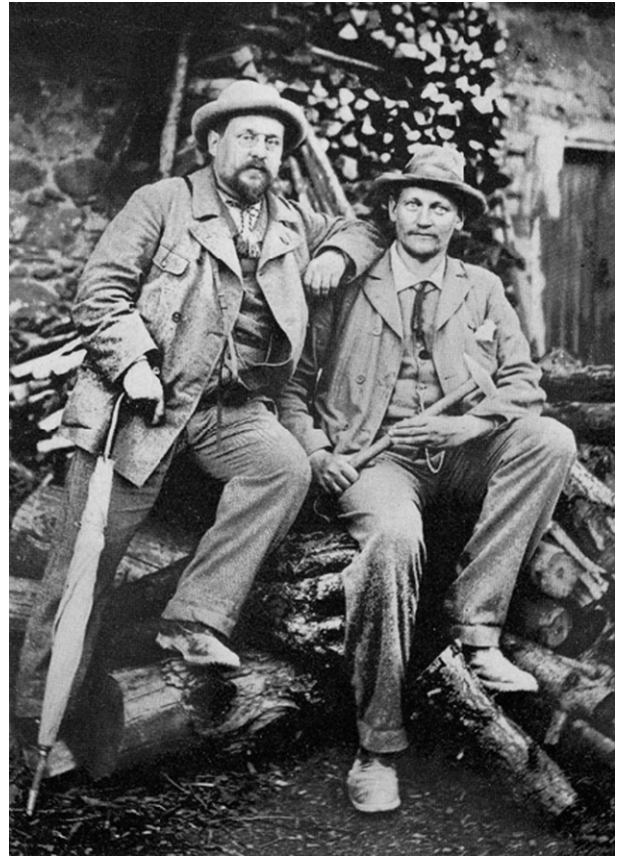


Figure 6 Eduard Brückner (1862–1927) and Albrecht Penck (1858–1945).

Development of Theories on the Causes of Glaciation

As we have seen, by the late nineteenth century, the geologic evidence for repeated, large-scale glaciations of the globe was firmly established. The causes of glaciation, however, remained a mystery. Geikie’s geologic evidence from Scotland showed that warm intervals had developed between glaciations. While the relative length of glacial and interglacial periods remained unknown, it was becoming clear that large-scale climatic oscillations had taken place over many thousands of years of Earth’s recent history. Various suggestions were put forward to explain these cycles. Changes in carbon dioxide levels were proposed, as well as changes in solar intensity.

Croll’s Orbital Theory

One of the earliest theories on the cause of glacial–interglacial cycles was developed by the Scottish scientist, James Croll (Figure 8). Croll had little formal education, but he was a voracious reader.

In 1859, his pursuit of knowledge led him to enter the academia ‘through the back door,’ by becoming a janitor at the museum at Anderson’s Institution in Glasgow. Once there, he began developing a theory about the causes of glaciation. He began writing letters to Charles Lyell, discussing his ideas on the connections between glaciation and variations in the

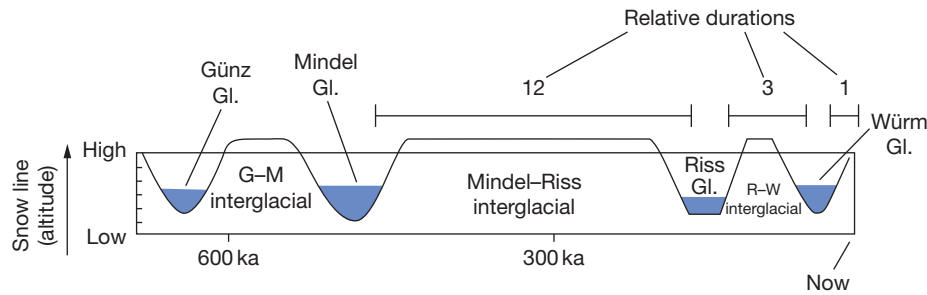


Figure 7 Diagram of European Ice Ages, their relative durations, and the relative snow line during each. Note: During glacial periods, snow fell at lower altitudes than during interglacial periods. Gl, glacial; Intergl, interglacial. Reproduced from Penck A and Brückner E (1909) *Die Alpen im Eiszeitalter*. Leipzig: Tachnitz.



Figure 8 James Croll (1821–90). Photo by J.C. Irons (1896).

Earth's orbit. Lyell was suitably impressed with Croll's scholarship and helped him obtain a clerical position at the Geological Survey of Scotland in 1867. It was here that Croll was encouraged by Archibald Geikie to further develop his theory. Charles Darwin also corresponded regularly with Croll, and both scientists benefited from this exchange of ideas.

Croll started publishing his theories in 1867(a,b), and his major contributions include *Climate and Time, in their Geological Relations* (1875) and *Climate and Cosmology* (1885).

In 1846, French astronomer Urbain Le Verrier published formulas that allow the calculation of changes in the shape of a planet's orbit and its axial precession. In 1864, Croll used these formulas to plot changes in the shape of the Earth's orbit (called orbital eccentricity) over the past 3 million years. He found that a pattern of high eccentricity had persisted for hundreds of thousands of years, followed by a pattern of low eccentricity, as is the case today. The more elliptical the orbit, the greater is the difference in incoming solar radiation (insolation) between the different seasons of the year. Croll realized the importance of calculating the seasonality of insolation, which is one of his major contributions to the science of paleoclimatology. Changes in the Earth's orbit that act to prolong the winter season cause greater amounts of snow to accumulate in the high latitudes (Figure 9).

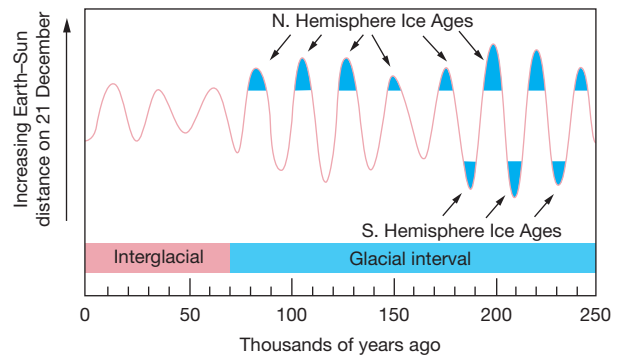


Figure 9 Diagram illustrating Croll's (1887) explanation of Ice Ages, based on changes in the Earth's orbit around the sun.

The extra snow cover reflects more solar energy back out into space, thereby amplifying the orbital effects. Croll argued that this amplification is what triggers the growth of ice sheets.

Croll's theory introduced important new concepts in the field of climatology. Subsequent research has shown that Croll's theory is insufficient to explain the global pattern of Pleistocene glaciations, and his chronology of glaciations has been shown to be in error. Specifically, Croll's scheme made the last Ice Age much older than was inferred from the geologic evidence of Geikie and others. Ultimately, Croll failed to convince most of his contemporaries, and his ideas remained largely ignored by other researchers until the 1940s.

The Milankovitch Theory

Milutin Milankovitch (Figure 10) was a Serbian mathematician who specialized in astronomy and geophysics. In 1909, he became a member of the faculty of applied mathematics at the University of Belgrade.

Imprisoned by the Austro-Hungarian Army in World War I, he recommenced work on his mathematical theory of climate change in 1920, completing this work in 1941. Milankovitch built his theory from previous work done by J.A. Adhemar and James Croll. In 1842, Adhemar explained glacial climate using only precession. Milankovitch used Croll's work to help him develop a mathematical model of climate change. This model incorporates the cyclical variations in the three elements of the Earth's orbit around the sun: eccentricity, obliquity, and precession. Using these three orbital factors, Milankovitch developed

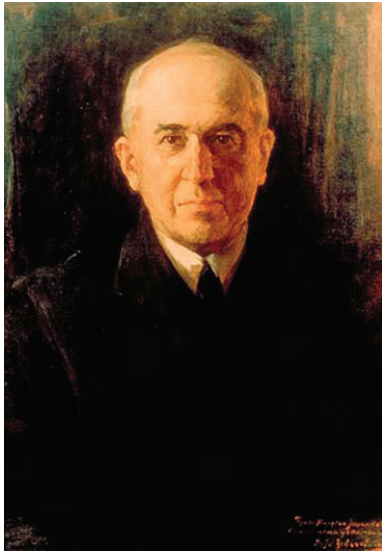


Figure 10 Portrait of Milutin Milankovitch (1879–1958) by Paja Jovanovic (1943). Photo courtesy of Vasko Milankovitch.

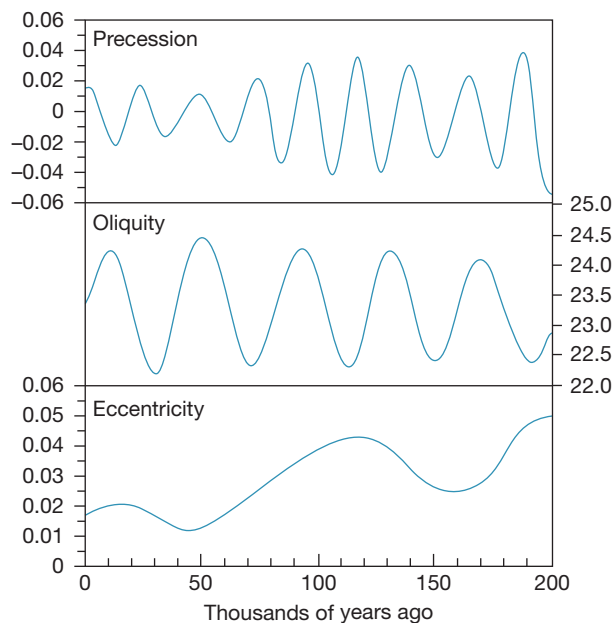


Figure 11 Orbital variations predicted by the Milankovitch theory. Reproduced from Berger A and Loutre MF (1991) Insolation values for the climate of the last 10 million years. *Quaternary Science Reviews* 10: 297–317.

a comprehensive mathematical model that calculated latitudinal differences in insolation and the corresponding surface temperatures during the last 600 000 years (Figure 11).

The next step in Milankovitch's work was an attempt to correlate the orbital variations with glacial–interglacial cycles. Milankovitch worked on the assumption that radiation changes in some latitudes and seasons are key to triggering glaciation and deglaciation. Working with German climatologist Vladimir Koppen, he chose the summer insolation values at 65° N as the critical latitude and season. Their reasoning was

that the continental ice sheets grew near this latitude and that cooler summers might reduce summer snowmelt, leading to a buildup of snow pack and eventually to the growth of ice sheets.

Sadly, Milankovitch's theory was largely ignored for decades. However, Hays et al. (1976) published a study of deep-sea sediment cores and found that Milankovitch's predictions matched their own interpretations of the timing and intensity of climate change during the last 450 000 years. Specifically, they found that major variations in climate were closely associated with changes in the eccentricity, obliquity, and precession of the Earth's orbit.

The Invention of Dating Methods

Without a means of obtaining an absolute age for events in the Quaternary, there would have been no way to test the validity of Milankovitch's orbital variation theory. Until the latter half of the twentieth century, Quaternary scientists lacked the tools to obtain such absolute ages and could only infer the ages of events through relative dating techniques. In other words, they could sometimes establish the *sequence* of events, for instance, by determining the relative stratigraphic position of various kinds of fossils. But they could not tell whether a given sequence of events took place 50 000 or 150 000 years ago, unless they were dealing with long sequences of sedimentary layers that had accumulated in recognizable, annual layers (a very rare phenomenon).

Uranium-Series Dating

Radiometric dating methods were developed in the twentieth century and have now revolutionized Quaternary science. In 1902, physicists Ernest Rutherford and Frederick Soddy discovered that radioactive elements broke down into other elements in a definite sequence or series through the process of nuclear fission. The possibility of using this radioactivity as a means of measuring geologic time was first discussed by Rutherford in 1904. In 1906, Rutherford began calculating the rate of radioactive decay of uranium. This decay process (uranium decaying to lead) has since been discovered to go through multiple steps, with intermediate daughter products. It is now possible to use various uranium-series decay processes to derive age estimates for uranium-bearing fossils and sediments that had existed many millions of years ago.

Radiocarbon Dating

Perhaps the most important breakthrough in the absolute dating of Quaternary fossils and sediments was the invention of radiometric dating methods, especially radiocarbon dating. In 1940, American physicists Martin Kamen and Sam Ruben discovered the long-lived radioactive carbon isotope, carbon-14. Kamen used ^{14}C as a tracer in biological systems. Kamen found that some of the nitrogen in the atmosphere was turned into carbon-14 when bombarded with cosmic rays. The existence of ^{14}C had been postulated since 1934, but it had never been directly observed nor characterized. Kamen succeeded in

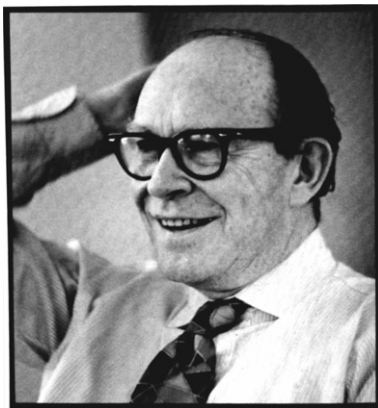


Figure 12 Photograph of Willard F. Libby, inventor of the radiocarbon dating method. Photo courtesy of Geoscience Analytical Inc.

preparing ^{14}C in sufficient amounts to determine its half-life (5700 years), that is, the amount of time it takes for half of a sample of ^{14}C to break down to the stable ^{14}N isotope of nitrogen.

Building on Kamen's discoveries, in 1947, American chemist Willard Libby ([Figure 12](#)) determined that plants absorb traces of ^{14}C during their uptake of carbon in photosynthesis. At death, the plant would stop absorbing carbon, and the ^{14}C it contained would decay at its usual rate without being replaced. By measuring the concentration of ^{14}C left in the remains of a plant, [Libby \(1952\)](#) discovered that it was possible to calculate the amount of time since the plant had died. In addition, it was found that the same concentrations of ^{14}C occur in the tissues of animals as in plants, since animals either directly or indirectly ingest the carbon from plant tissues as their food. Given that it is possible to measure the concentration of the remaining ^{14}C back to nine or ten half-lives, it has thus become possible to obtain absolute age estimates of fossil specimens (both plant and animal) that existed about 45 000–50 000 years. For his work on carbon-14 dating, Libby received the Nobel Prize in chemistry in 1960.

Luminescence Dating

Thermoluminescence (TL) dating is based on the fact that natural minerals can absorb and store energy from ionizing radiation. If a mineral is heated to a sufficiently high temperature, some of the stored energy is released in the form of light called TL. In nature, the energy absorbed by a mineral mainly comes from radiation emitted from radioisotopes within the mineral grain, from its immediate surroundings, and from cosmic rays. This energy is stored in the form of electrons that get trapped at defects in the crystal lattice of some minerals, notably, quartz and feldspar. These trapped electrons build up slowly through time and are released when the mineral is exposed to sunlight or when it is sufficiently heated. TL dating therefore provides a method of dating the time that has elapsed since mineral grains were last exposed to sufficient heat or sunlight. [Daniels et al. \(1953\)](#) first noted the potential of using TL dating in geological and archaeological research, and it was developed in the early 1960s as a means of dating fired pottery ([Aitken et al., 1964](#)). The technique was later modified

so that it also could be used to date the last time that sediments were exposed to sunlight.

Optically stimulated luminescence (OSL) dating is based on the fact that if a mineral is exposed to sufficient light (e.g., sunlight), some or all of this stored energy will be lost. OSL dating developed from TL dating in the mid-1980s ([Huntley et al., 1985](#)). Its main advantage over TL dating is that only the most light-sensitive signal is sampled, which allows for much younger samples to be dated. Moreover, OSL dating usually results in better precision. These two methods have contributed significantly to Quaternary science in that they allow researchers to obtain dates from mineral grains, rather than just from organic compounds (the basis of radiocarbon dating). Furthermore, OSL dating typically provides useful dates (e.g., quartz grains) ranging from a few centuries to about 150 000 years, well beyond the limits of radiocarbon dating.

Conclusion

Other articles in this encyclopedia will highlight the state of the art in the aforementioned fields of Quaternary stratigraphy, vertebrate paleontology, Pleistocene glaciology, paleoclimatology, and dating methods. As with all branches of science, the current generation of researchers have built on the foundations of people such as Agassiz, Lyell, Cuvier, Milankovitch, and Libby. We owe these pioneers an enormous debt of gratitude. Many of these people worked in relative obscurity during their own lifetimes, and their theories were openly ridiculed by their contemporaries. Many survived major political upheavals and wars in the rapidly changing world of the nineteenth and twentieth centuries. The unifying themes of their lives are their intellectual curiosity, their diligence and perseverance, and their breadth of vision. May the same be said of twenty-first century Quaternary scientists by future generations.

See also: [Vertebrate Overview](#). **Glaciation, Causes:** [Astronomical Theory of Paleoclimates](#). **Glaciations:** [Late Pleistocene in Eurasia](#); [Middle Pleistocene in Eurasia](#). **Luminescence Dating:** [Optical Dating](#); [Thermoluminescence](#). **Paleoceanography:** [Paleoceanography An Overview](#). **Paleoclimatology:** [Introduction](#). **Paleoclimatic Reconstruction:** [Sub-Milankovitch \(DO/Heinrich\) Events](#). **Quaternary Stratigraphy:** [Overview](#). **Radiocarbon Dating:** [Conventional Method](#).

References

- Agassiz L (1840) *Etudes sur les glaciers*. Neuchâtel: Jent & Gassmann.
- Agassiz L (1866) *Structure of Animal Life: Six Lectures Delivered at the Brooklyn Academy of Music in January and February, 1866*. New York: Charles Scribner & Co.
- Aguirre E and Pasini G (1985) The Pliocene–Pleistocene boundary. *Episodes* 8: 116–120.
- Aitken MJ, Tite MS, and Reid J (1964) Thermoluminescent dating of ancient ceramics. *Nature* 202: 1032–1033.
- Bassett MG (1985) Towards a 'common language' in stratigraphy. *Episodes* 8: 87–92.
- Berger A and Loutre MF (1991) Insolation values for the climate of the last 10 million years. *Quaternary Science Reviews* 10: 297–317.
- Croll J (1867a) On the eccentricity of the Earth's orbit, and its physical relations to the glacial epoch. *Philosophical Magazine* 33: 119–131.
- Croll J (1867b) On the change in the obliquity of the ecliptic, its influence on the climate of the polar regions and on the level of the sea. *Philosophical Magazine* 33: 426–445.

- Croll J (1875) *Climate and Time in Their Geological Relations*. London: E. Stanford.
- Cuvier G (1825) *Recherches sur les ossements fossiles: où l'on rétablit les caractères de plusieurs animaux dont les révolutions du globe ont détruit les espèces*. Paris: G. Dufour et E. d'Ocagne.
- Daniels F, Boyd CA, and Saunders DF (1953) Thermoluminescence as a research tool. *Science* 117: 343–349.
- Desnoyers J (1829) Observations sur un ensemble de dépôts marins plus récents que les terrains tertiaires du bassin de la Seine, et constituant une formation géologique distincte; précédées d'un aperçu de la non-simultanéité des bassins tertiaires. *Annals Sciences Naturelles (Paris)* 16: 171–214, 402–491.
- Flint RF (1971) *Glacial and Quaternary Geology*. New York: Wiley.
- Forbes E (1846) On the connection between the distribution of existing fauna and flora of the British Isles, and the geological changes which have affected their area, especially during the epoch of the Northern Drift. *Great Britain Geological Survey Memoir* 1: 336–342.
- Gervais P (1867–69) *Zoologie et paléontologie générales. Nouvelles recherches sur les animaux vertébrés et fossils*, vol. 2. Paris: Bertrand.
- Hays JD, Imbrie J, and Shackleton NJ (1976) Variations in the Earth's orbit: Pacemaker of the Ice Ages. *Science* 194: 1121–1132.
- Hilgen FJ (1991) Astronomical calibration of Gauss to Matuyama sapropels in the Mediterranean and implication for the Geomagnetic Polarity Time Scale. *Earth and Planetary Science Letters* 104: 226–244.
- Hörnes M (1853) Mittheilung an Prof. Bronn gerichtet. Wien, 3. Okt., 1853. *Neues Jahrbuch Mineralogie Geologie Geognosie und Petrefaktenkunde* 806–810.
- Huntley DJ, Godfrey-Smith DI, and Thewalt MLW (1985) Optical dating of sediments. *Nature* 313: 105–107.
- Kukla G (2005) Saalian supercycle, Mindel/Riss interglacial and Milankovitch's dating. *Quaternary Science Reviews* 24: 1573–1583.
- Libby WF (1952) *Radiocarbon Dating*. Chicago: University of Chicago Press.
- Lyell C (1833) *Principles of Geology, Being an Attempt to Explain the Former Changes of the Earth's Surface by Reference to Causes Now in Operation*, vol. III. London: John Murray.
- Lyell C (1839) *Eléments de Géologie*. Paris: Pitois-Levrault.
- Lyell C (1873) *The Geological Evidence of the Antiquity of Man*, 4th edn. London: John Murray.
- Milankovitch M (1941) Kanon der Erdbestrahlung und seine Anwendung auf das Eiszeitenproblem. *Académie Royale Serbe Editions Speciales Section des Sciences Mathématiques et Naturelles* 133.
- Penck A and Brückner E (1909) *Die Alpen im Eiszeitalter*. Leipzig: Tachnitz.
- Reboul H (1833) *Géologie de la période Quaternaire et introduction à l'histoire ancienne*. Paris: F.G. Levrault.
- Schneer CJ (1969) *Toward a History of Geology*. Cambridge, MA: MIT Press.

History of Dating Methods

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The first methods for dating Quaternary materials resulted from the discovery of radioactivity 120 years ago. In addition, other dating methods have been developed that rely on other time-dependent changes that occur in natural materials. The different methods can be used to obtain the age of formation of either organic or inorganic materials.

Early Developments

Numerical dating methods based on natural radioactive phenomena were initially developed as a means of determining the age of the Earth (Dalrymple, 1991). Arguments on this topic raged in the second half of the 19th century, a time when the effects of recent glaciation were also being debated. However, it was not until the discovery of radioactivity in the 1880s that consideration was given to radioactive isotopes for use as natural clocks. As early as 1906, Rutherford suggested that ages could be obtained based on the production of helium by the decay of uranium in rocks. The first calculation of the age of the Earth based on the amount of radium in the Earth's crust was made by Russell in 1921. However, it was not until the 1930s, when Nier brought together the understanding of natural isotopes in the uranium and thorium decay chains and the construction of the first mass spectrometers, that it became possible to measure a range of isotopes. Also at this time, the potential of several different decay series that could be used for dating was proposed, based on a better understanding of the atomic structure of elements in the periodic table. The relationship between these early radiometric dates and the evolution of the geological timescale has been covered in a history of the work of Arthur Holmes (Lewis, 2000).

Radiocarbon Dating

The speed at which new geochronological tools based on radioactivity became relevant to Quaternary geology can be seen by examining the prefaces to the 1958 edition of Zeuner's book *Dating the Past*. In 1946, when the first edition was published, geochronology was very much based on the study of superposition of geological deposits and botanical and zoological successions. By the time of the third edition in 1952, theoretical and practical approaches to radiocarbon (^{14}C) dating had been put forward by Libby and the first radiocarbon laboratory was established. Soon after, lists of radiocarbon dates started to appear, although at this time radiocarbon dating was mainly applied to archaeological materials, particularly from Upper Paleolithic sites. Another important result of the application of radiocarbon dating was in providing a chronology for the development of farming in Europe and the Near East. Libby was at the heart of radiocarbon dating development for three

decades, producing the first book on the topic in 1952 and receiving the Nobel Prize in chemistry in 1960 for his work on radiocarbon dating. A comprehensive history of radiocarbon dating, and one that demonstrates the role played by Libby, can be found in Chapter 6 of the book by Taylor (1987) and in the edited volume by Taylor et al. (1992).

The early measurement of beta particles released from solid carbon was rapidly replaced by the use of more efficient counting systems using gases or liquids made from the organic material being used for dating. Both methods enable dating back to 60 ka, although the dating range is usually limited to approximately 40 ka due to sample contamination. In the 1950s, gas counting using carbon dioxide, methane, or acetylene was introduced and continued to be used for many decades. These gases (CO_2 , CH_4 , and C_2H_2) contained molecules with ^{14}C in their structure. Preparing the sample as a gas had the additional advantage that it was possible to enrich the concentration of ^{14}C relative to ^{12}C and ^{13}C in part of the gas volume and thus extend the potential age range back to 75 ka (Grootes et al., 1975). This process was known as isotopic enrichment.

In the 1960s, liquid scintillation counting using benzene (C_6H_6) as the carrier of the ^{14}C was developed. Benzene was easy to prepare directly from acetylene. As a liquid, it had two advantages: It was possible to mix the sample and the organic scintillator chemicals, and it had a far greater density than acetylene and thus took up a smaller volume in the counting equipment. The latter property enabled a reduction in background count rate, and by the 1970s, automated liquid scintillation measurement systems were set up in laboratories worldwide.

The next major breakthrough in measurement techniques came in the 1970s, when methods were put forward in which it was not necessary to measure the beta particles released as the ^{14}C atoms decayed. The objective was to measure the ratio of ^{14}C to the more stable isotope of carbon (^{12}C) using a mass spectrometer. The first results were achieved using a particle accelerator known as a cyclotron, which had the advantage of removing all interfering particles except ^{14}N . In other studies, electrostatic accelerators were used in which molecules with similar charge-to-mass ratios as ^{14}C needed to be removed before being allowed into the mass spectrometer. Destruction of these molecules was accomplished by ionizing the sample and accelerating the ions, prior to passing them through a foil. Subsequently, small accelerators were built and the measurement technique of accelerator mass spectrometry (AMS) was born (Hedges, 1981). Being able to measure the atoms directly and not have to wait for them to decay speeded up measurement time. It also led to the dating of very small samples containing only 1 or 2 mg of elemental carbon. This led to a revolution in applications. A significant development has also been made in sample preparation using acid–base–wet



Figure 1 Long-lived bristlecone pines growing in the White Mountains, California, used to provide a tree-ring record for radiocarbon calibration. Photo by Mike Walker.

oxidation and target production for AMS measurements. This has resulted in finite ^{14}C ages older than 50 ka being obtained for charcoal.

The development of radiocarbon dating was not without minor setbacks. Early tests of the dating technique were carried out on material from the tombs of the ancient Egyptian kings. In the 1960s, extensive sets of measurements were carried out on tree rings for which calendar ages could be determined by counting the rings. This method of dating is known as dendrochronology; the first tree-ring laboratory was set up in Arizona in 1937. Radiocarbon ages were obtained on rings from continuous and overlapping sequences cut from dead and living examples of the bristlecone pine, a tree that grows high up in the White Mountains of eastern California (Figure 1). When the radiocarbon ages were compared with the calendar dates from the counted tree-ring chronology, a significant deviation was observed for samples older than 3000 years old. This deviation was the result of long-term changes in the production rate of ^{14}C in the upper atmosphere.

By 1986, calibration curves from many laboratories were brought together and a combined calibration curve was published. Use of bristlecone pines and fossil trees from Ireland and Germany resulted in a continuous calibration curve (IntCal98) for radiocarbon ages (Stuiver and Van der Plicht, 1998) back to 10 299 years BP. A new replacement calibration curve (IntCal04) has been constructed back to 26 000 years BP using paired radiocarbon and uranium series dates on corals and Foraminifera (Reimer et al., 2004).

Potassium–Argon and Argon–Argon Dating

The potential of dating using the decay of ^{40}K to ^{40}Ar was first reported in 1948. The K–Ar technique was developed in the 1950s for determining the age of rocks (Dalrymple and Lanphere, 1969); the first studies were mainly on pre-Quaternary rocks because of the difficulty in measuring the small amounts of argon that have accumulated in the younger rocks. The $^{40}\text{Ar}/^{39}\text{Ar}$ dating method, in which it is not necessary to measure absolute abundances of either argon or potassium in the sample, was developed in 1966. In the first study, it was

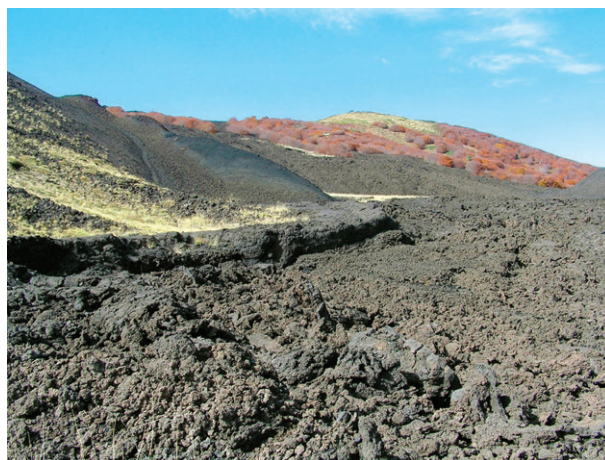


Figure 2 A lava flow on Etna that would provide a record of magnetic field direction and that could be dated by the potassium–argon method.

noted that the new procedure would result in more precise ages, would allow measurements on very small samples, and could provide information on the thermal history of crystalline materials if step heating was used. In the early 1970s, such experiments were carried out on a range of terrestrial rocks, with comparisons being made with conventional K–Ar dating.

An important application of both techniques was the dating of lava flows (Figure 2), particularly since they contained evidence of the direction of the Earth's magnetic field at the time that they cooled. During the Quaternary period, the Earth's magnetic field has changed direction several times and the record of these changes is also found in sedimentary rocks (e.g., deep-sea cores and loess). At approximately the beginning of the Quaternary, the field changed from being in the direction that it is today to being totally reversed, and then at approximately 780 ka it changed back again (Jacobs, 1994). These periods of different direction are referred to as chrons and are named after scientists who have studied magnetic behavior. The change from the Gauss normal chron to the Matuyama reversed chron was dated to 2.47 Ma, and the change from the Matuyama chron to the Brunhes normal chron was first dated by the potassium–argon method to 730 ka. Most recently, an age of 775.6 ± 1.9 ka has been obtained using the $^{40}\text{Ar}/^{39}\text{Ar}$ method. Within the Matuyama reversed chron, there were a number of times when the field direction was normal for a few thousand years; these subchrons, named after places where they were first recorded, occurred at approximately 0.9 Ma (Jaramillo), 1.11 Ma (Cobb Mountain), 1.7 Ma (Olduvai), and 2.1 Ma (Reunion). These reversals have been the subject of extensive K–Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ dating studies during the past 30 years (Jacobs, 1994). This record of reversals has been found in the 195 m of loess that overlies red clay in the Chinese Loess Plateau. This has allowed the transfer of an independent age scale onto the loess and the demonstration of the coincidence of the loess/red clay transition and the beginning of the Quaternary.

This geomagnetic polarity timescale has also been used, often in conjunction with other dating techniques, to obtain ages for sediments containing skeletal remains of early

hominids in Africa. In the Turkana Basin of East Africa, the paleomagnetic history recorded in the sediments has been used along with K–Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ dating to provide a framework for the archaeological material found therein. In exposed sections, layers of volcanic ash (tephra), known as tuff, can be seen, relating to explosive eruptions of volcanoes hundreds of kilometers to the north. The first K–Ar age was for the KBS Tuff, found above the remains of an important skull of *Homo habilis*. However, the age was severely overestimated as a result of the incorporation of detrital minerals during fluvial reworking of the ash after it had been laid down. During the next 20 years, this tuff and approximately 20 others have been dated (and redated) as technological advances have been made (McDougall, 1990; McDougall and Harrison, 1999). The KBS tuff (named after the graduate student who had found stone tools within the ash layer) and the controversy generated by the dating of it and other tuffs were the subject of two chapters in Lewin's book (1987) on human origins in Africa. In the decade that followed the announcement of the first K–Ar date on the KBS tuff, a number of new measurements were carried out, concentrating on the separation of crystals of anorthoclase (a plagioclase feldspar found in the pumice clasts). These were used for both conventional K–Ar dating and $^{40}\text{Ar}/^{39}\text{Ar}$ dating using step heating, a process that would identify the presence of partially heated grains. By 1981, an age of 1.88 ± 0.03 Ma was accepted for the KBS tuff.

At approximately the same time, it was demonstrated that laser light focused onto a small spot on the surface of a grain could release argon. By the end of the 1980s, it became possible to obtain $^{40}\text{Ar}/^{39}\text{Ar}$ ages on individual grains of potassium-rich feldspars of Quaternary age, with the first results for samples of pumice and other volcanic material from a site in Kenya being published in 1990. This method became known as single crystal laser fusion (SCLF) dating (Walter, 1997). In the next few years, this method was employed on crystals from other hominid sites in Kenya and in Ethiopia. The earlier dates that had been obtained for the Turkana Basin by conventional K–Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ techniques were confirmed by the new SCLF technique. On a more recent timescale, SCLF methods were used for sanidine crystals from the Laacher See Tephra in the Eifel region of Germany. Combining the ages for 16 individual crystals, an age of 12.7 ± 0.8 ka was obtained for the eruption.

The ability to use $^{40}\text{Ar}/^{39}\text{Ar}$ for Holocene volcanic samples has been demonstrated using subsamples made up of a few tens of sanidine crystals, feldspars with very high potassium contents. These crystals, up to approximately 1 mm in diameter, were extracted from a pumice clast from the city of Pompeii, buried by the eruption of Vesuvius in 79 AD. The age of 1925 ± 94 years before 1997 AD was obtained not by laser fusion but by using incremental heating of the grains, thus taking methods of analysis to an even higher level. A review of the principles and methodology of K–Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ dating has been published by Renne (2000).

Fission Track Dating

Explosive volcanic eruptions not only produce potassium-rich feldspars but also produce volcanic glass and small zircon and apatite grains. These materials are suitable for yet another

dating method, fission track dating (Westgate et al., 1997). Fission track dating is based on fission of the uranium isotope ^{238}U , a process that has a half-life of 8.2×10^{15} years. The use of fission tracks for dating was born out of the study of fissile material during World War II. Fission is the name given to the process in which an atom spontaneously breaks apart with a tremendous release of energy. In the case of ^{238}U , two nuclei (fission fragments) are formed. They travel through the crystal away from the site of the parent ^{238}U atom in opposite directions, creating a damage trail that is a few micrometers long and approximately one-hundredth of a micrometer in diameter. This damage trail can only be seen as a fission track when the mineral face containing the track is exposed to a chemical etchant that attacks the damage zone along the length of the track. The number of tracks that cross the polished face of the mineral is proportional to the age and ^{238}U content. To obtain an age, this track density is compared with that induced by irradiating the mineral with neutrons, a process that causes the less abundant ^{235}U atoms to undergo fission. The tracks caused in this way are etched and measured as for the spontaneous tracks formed in nature by ^{238}U .

Because the tracks can be removed by heating (annealing), fission track measurements have also been used to determine the time since some thermal event subsequent to the formation of the mineral grains (thermochronology). The historical development of fission track measurement methods was discussed by Fleischer et al. (1975), and a more recent account of fission track dating over the whole of geological time was provided by Wagner and van den Haute (1992). This book also includes discussion of how two different values for the fission half-life were used for many years; this resulted in ages up to 20% different, depending on which half-life was used in the age equation. This problem was resolved by using standard minerals of known age (as obtained by $^{40}\text{Ar}/^{39}\text{Ar}$ dating).

For Quaternary volcanic events, dating has been limited to the use of zircon and apatite grains and volcanic glass. Zircon is the mineral of choice because of its high uranium content; zircon grains will thus have proportionally more tracks than volcanic glass for the same time period. However, very few zircon grains are usually found, perhaps as little as one grain in 1 kg of volcanic sediment (Dumitru, 2000). One of the earliest successes in fission track dating of zircon was the establishment of different ages for a number of volcanic ashes derived from the Quaternary volcanoes in Yellowstone National Park. Two geochemically different examples of the Pearlette Ash, previously used as a time-marker horizon, were found to have ages of 1.9 ± 0.1 and 0.6 ± 0.1 Ma, in agreement with K–Ar ages for two ashes closer to the source.

Even when present, zircon grains tend to be small and thus quite a few are needed to make a reliable measurement. The counting of several grains has the advantage of being able to identify contaminant grains. In the case of a volcanic ash bed in Ethiopia, it was found that the sample contained two populations of zircon grains (Walter, 1989). The ash was dated to 2.3 ± 0.5 Ma by dating the younger grains, but it was shown to contain grains with an age of 13.1 ± 3.2 Ma. If the different track densities had not been noticed, an age of 4.0 ± 0.7 Ma would have resulted.

Because of the scarcity of zircon grains, volcanic glass is usually preferred for Quaternary deposits. Early ages on glass

were found to be too young compared with zircon ages from the same tephra. This was the result of the more rapid annealing of tracks in the case of volcanic glass. To overcome this problem, it was suggested that both irradiated and unirradiated aliquots should be heated to different temperatures for fixed periods of time prior to etching. When a temperature is reached for which the measured track densities are in a fixed ratio, this ratio can be used to obtain the age of the glass.

A more widely used approach has been developed using a lower temperature and longer times. This approach is known as isothermal plateau fission track (ITPFT) dating. More recently, a relationship between the average diameter of the fission tracks and the track density has been found; both are reduced by thermal annealing. ITPFT dating of volcanic glass has proved invaluable for dating ash layers found in the loess deposits of Alaska (Figure 3), the Yukon, and New Zealand (Westgate et al., 1997). A major result of the application of fission track dating of volcanic glass was the dating of the Toba tephra from western India, which, together with geochemical analysis of all known occurrences, firmly establishes that it has a single age, namely 75 ka.

Luminescence Dating

Another method that measures the effect of nuclear processes in minerals is based on the luminescence properties of quartz and feldspars. Unlike fission track dating, it is the effect of ionizing radiation resulting from the decay of ^{238}U , ^{235}U , ^{232}Th , and ^{40}K in the minerals and their surroundings that



Figure 3 Volcanic ash (Old Crow tephra) from Alaska that has been dated by the fission track method. The ash is overlain and underlain by loess that has been dated using luminescence techniques. Photo by Helen Roberts.

results in the age-dependent signal. Unlike most of the other dating methods, luminescence dating techniques do not provide the age of formation of the quartz and feldspar grains to which they are applied. Instead, they determine the point in time at which the grains were either heated or exposed to light. For sedimentary grains, determining the time since the grains were last exposed to light provides the depositional age. This has allowed direct determination of rates of sediment accretion.

Luminescence dating was born out of the need to measure the amounts of ionizing radiation in a laboratory handling radioactivity, or in the environment or as a result of nuclear explosions, and the ability to make the measurements provided by the development in the 1940s of a sensitive light-measuring device known as a photomultiplier tube. The earliest use of thermoluminescence (TL) for dating both archaeological and geological materials was in 1953, but it was not until the 1960s that it was developed as a useful tool. The TL is released from irradiated minerals when they are heated from room temperature to 450 °C in an oxygen-free environment. TL measurements on mineral grains extracted from pottery, which would have had their signal removed by heating during manufacture, were first made in 1964. For the next two decades, TL dating was confined to heated materials, including burnt flint and lava flows. The technical aspects of TL dating are fully covered by Aitken (1985). However, as early as the mid-1960s, TL ages for unheated sediments had been reported in the Russian literature, but they remained unnoticed in the West until the late 1970s when the dating of mineral grains from a deep-sea core was reported. During the next 15 years, TL dating was applied to a wide range of terrestrial deposits ranging from coastal dunes to loess.

The next major breakthrough was in the use of an optically stimulated luminescence (OSL) signal (Huntley et al., 1985). The use of an optically stimulated signal was seen as clearly more appropriate for measuring a signal whose resetting mechanism was exposure to light. Indeed, a light-insensitive TL signal remained when sedimentary grains were exposed to light, either from the Sun or from laboratory light sources. OSL techniques became more widely used when small halogen bulbs, and then light-emitting diodes (LEDs) (Figure 4),



Figure 4 A luminescence dating laboratory showing blue light used for optically stimulated luminescence dating of sand grains. Photo by Risø National Laboratory, Denmark.

replaced the expensive argon ion laser used as the stimulation source.

A peculiar optical stimulation resonance for feldspars, but not quartz, was found in the infrared (IR) region of the spectrum. The availability of IR-emitting diodes soon led to the dating of feldspars using infrared-stimulated luminescence (IRSL). This signal is rapidly zeroed by sunlight, thus making it suitable for dating sediments. The ease of zeroing the IRSL in a relatively short time using such diodes led to the development in 1991 of measurement procedures that could be applied to single portions (aliquots) of K-feldspar grains. This permitted multiple measurements of the equivalent dose (the measure of the past radiation exposure that is proportional to age) to be made.

The stability, power, and wavelength availability of LEDs has led to a rapid growth in methodological developments in the past 10 years (Wintle, 1997; Aitken, 1998). LEDs producing intense blue-green light became available, and a procedure was developed for quartz based on the use of a single aliquot. The single-aliquot regenerative dose procedure is now the basis of most OSL dating applications. IRSL, or OSL, dating of feldspars is less widely applied since for many feldspars the luminescence signals are unstable over geological time (a phenomenon known as anomalous fading). Overcoming this problem for feldspars and getting around the early saturation of the OSL signal for quartz are the current challenges. It should be noted that because of early saturation of the OSL signal found for all quartz, older sediments that are known to be made up of well-bleached grains (e.g., coastal dune ridges) have continued to be dated using a light-sensitive TL signal from quartz.

Also, since the past decade, it has become possible to use a small laser to produce a beam of green light focused onto a 20- μm -diameter spot to obtain OSL from single quartz, or feldspar, grains. This has enabled the measurement of several thousand grains to be carried out and, depending on the brightness of the individual grains, the equivalent dose to be determined for a large number of them (Duller, 2004). Whether these values of equivalent dose form a single population or are scattered enables conclusions to be drawn about whether the grains making up a sediment unit were all well exposed to light at deposition. This is particularly important when sediments of glacial origin are to be dated or when postdepositional reworking is a possibility.

Electron Spin Resonance Dating

Another mineral property that provides a measure of past radiation dose is electron spin resonance (ESR). ESR measurements are made by placing a small aliquot of sample in a magnetic field and observing the amount of microwave energy that is absorbed. This relates to the trapped electrons in the crystalline structure and is proportional to the age since formation (Grün, 1997). The idea of measuring trapped electrons by their behavior in a magnetic field was first put forward in 1936. However, it was not until 1975 that it was used to date a speleothem. In this case, it was used to determine the time that had elapsed since the formation of the calcite that makes up the stalagmite or stalactite. In the 1980s, there was a considerable amount of investigation of ESR as a dating method

for a wide range of Quaternary materials, with the first use for dating corals from raised beaches to establish a terrestrial record of high sea levels. Also in the 1980s, ESR was explored as a method of dating minerals affected by the pressure produced at active faults. It has also been demonstrated that there are particular ESR signals in quartz grains that can be zeroed by sunlight and this would enable sediments to be dated in a way that is analogous to OSL dating. The first measurements of these signals in single grains are also encouraging. Single-grain ESR measurements on quartz grains of volcanic origin have also provided an age for the Toba eruption of 74 ka, albeit with a large amount of scatter.

The potential of using the signal from tooth enamel was investigated in 1987, and by the early 1990s, ESR dating had been applied at a number of sites containing evidence of modern humans. Tooth enamel is made up of hydroxyapatite, a material that provides sufficient ESR sensitivity and remains unaltered during burial. However, the dentine and cementum that make up the rest of the tooth absorb large amounts of uranium from the groundwater that moves through the surrounding sediment, and this has led to the need to model the dose rate through time. Until recently, the upper and lower limits for ESR ages have been calculated using the 'early' or 'linear' uptake models. Such calculations were used in the dating of animal teeth from the Paleolithic site of Pech de l'Azé in France (Grün and Stringer, 1991). Unfortunately, for samples around 150 ka, there is a 40-ka range of values as a result of the uncertainty as to when the uranium was absorbed by the dentine attached to the teeth being dated. In 1993, newly developed thermal ionization mass spectrometers were used for measuring U and Th isotopes when present in low concentrations in very small samples. For teeth beyond the usual range of U-series dating, some assumptions still need to be made about when the uptake of uranium occurred, but the ages obtained with the two extreme hypotheses are much closer, for example, resulting in a concordant age of approximately 400 ka for two teeth from the Hoxne interglacial site. More recent developments in U-series methods using laser ablation to provide atoms for the mass spectrometer have enabled even greater precision, with the isotope contents being obtained spatially across a thin section cut from a tooth (Figure 5).

Uranium Series Methods

The radioactive decay processes of the uranium (^{238}U and ^{235}U) and thorium (^{232}Th) decay chains were investigated in 1938, but at that time the main interest was in the lead isotopes that are at the ends of each decay chain (^{206}Pb , ^{207}Pb , and ^{208}Pb , respectively). In between these lead isotopes and the parent isotopes are a series of daughter isotopes of various elements, each daughter being radioactive. Each of these daughter isotopes will have different chemical properties, with solubility being the most important. ^{230}Th is a daughter product in the ^{238}U decay chain. Thorium (in the form of either ^{230}Th or ^{232}Th) is much less soluble in water than uranium and as a result it is not found in groundwater. Thus, stalagmites and stalactites (Figure 6) formed in caves will take ^{238}U (and its daughter product ^{234}U) into the calcite



Figure 5 Sections of teeth from extinct kangaroos that have been used for combined electron spin resonance and MC-ICP-MS uranium series dating. Photo by Rainer Grün.



Figure 6 Cross section of a stalactite showing growth rings that have been dated using the MC-ICP-MS uranium series dating method. Photo by Miryam Bar-Matthews.

structure as they are formed. ^{230}Th is then produced as the uranium isotopes decay. This in-growth of ^{230}Th increases with time and provides a chronological tool for such materials (Schwarcz, 1997).

The earliest applications of ^{230}Th measurements were made in the 1950s and 1960s and applied to lacustrine carbonates, marine sediments and corals, as well as to cave calcite deposits. Until the 1980s, measurements were made by dissolving samples, depositing the isotopes onto metal disks, and then counting the alpha particles released. The isotopic source of each alpha particle could be recognized by its characteristic energy, a technique known as alpha spectrometry. This approach enabled ages to be obtained back to approximately 350 ka, but it was time-consuming, particularly for older samples, which

required several days of counting. Also, relatively large samples were required to obtain high enough isotope concentrations for counting levels to be sufficiently above background, and this led inevitably to the incorporation of inappropriate material (e.g., recrystallized or reworked material).

In the 1980s, these problems were overcome by the introduction of mass-spectrometric techniques. Using a mass spectrometer, the ^{238}U , ^{234}U , and ^{230}Th atoms could be measured directly, without waiting for their decay. This meant that small samples could be used, thus reducing the possibility of contamination, and counting statistics were improved so that very small errors could be obtained (e.g., a 2σ error of ± 1 ka for a sample with an age of 125 ka). The procedure was termed thermal ionization mass spectrometry (TIMS). Because of its high precision, TIMS has been applied to a wide range of materials, including speleothem (Richards and Dorale, 2003), marine and lacustrine carbonates (Edwards et al., 2003), and materials such as calcitic flowstones under cave art and ostrich eggshell as well as flowstones that seal off sediments containing other evidence related to early human activity (Pike and Pettitt, 2003). Multicollector inductively coupled-plasma mass spectrometry (MC-ICP-MS) is being developed as an even more precise tool for the *in situ* dating of growth rings in stalactites, such as those shown in Figure 6 (Eggins et al., 2005).

Dating the Past 100 Years

^{210}Pb is another isotope further down the ^{238}U decay chain. It has been used for dating younger materials because of its very short lifetime. It was introduced as a dating method in 1963 and is used for dating sediments from the past 100 years. Although several approaches have been developed (Noller, 2000), the current rate of supply model is most widely used, a method that requires the integrated activity of ^{210}Pb to be calculated for the deposit. Counting methods are employed following sample digestion, but it has been suggested that ^{210}Pb could be measured directly using MC-ICP-MS, as used for the longer lived isotopes in the ^{238}U decay chain. An artificial isotope, ^{137}Cs , has also been used as a marker horizon in even younger sediments.

Cosmogenic Nuclide Dating

During the 1990s, a suite of new techniques for dating appeared in the Quaternary literature (Cosse and Phillips, 2001). These methods are based on the production of isotopes in rocks at the Earth's surface and are collectively known as cosmogenic nuclide dating. Although it had been proposed by Davis and Schaeffer as far back as 1955 that cosmic rays would interact with atoms in minerals at the Earth's surface, it was not until the mid-1980s that this was explored. At that time, AMS measurement of cosmogenically produced ^{14}C (produced in the atmosphere) became possible. Using similar AMS techniques, it was possible to measure the small amounts of radioactive isotopes (^{10}Be , ^{26}Al , and ^{36}Cl) produced in rocks (Zreda and Phillips, 2000). In addition, the stable cosmogenic isotopes (^3He and ^{21}Ne) could be measured by mass



Figure 7 Glacial landscape in Patagonia showing granitic erratic for which an exposure age has been obtained using cosmogenic nuclides (^{10}Be). Photo by Neil Glasser.

spectrometry (Niedermann, 2002). The importance of these techniques is that they are able to provide direct information on the exposure history of both rocks and sediments. Thus, for the first time it was possible to obtain rates for geomorphic processes in the Quaternary.

Two types of application were developed. First, the exposure age of a rock surface can be measured directly because the number of atoms will be proportional to time, with due account being taken of both the production rate and radioactive decay. Second, when rock surfaces are removed by geomorphic processes, it is possible to use two isotopes with different lifetimes and obtain information on long-term landscape evolution. Some of the first applications were to glacial deposits, for which the fresh rock surfaces are exposed to cosmic rays as the result of erosion by glaciers (Figure 7), and to the new rock surfaces provided by meteorite impacts. However, any process that has the potential to move previously buried rocks to the surface in a relatively short time can be investigated using cosmogenic isotopes. Such studies have been used to quantify rates of bedrock weathering (Gosse and Phillips, 2001; Cockburn and Summerfield, 2004).

Amino Acid Racemization

During the past 40 years, several other dating methods have been developed that use changes that occur as a result of chemical processes in the environment. Of these, the most widely applicable is amino acid racemization (AAR), which was first put forward as a dating technique in 1969. Amino acids are the basic chemical units from which proteins are formed and several different types (e.g., alanine and isoleucine) are found in organic materials (Hare et al., 1997).

In the natural environment, these amino acids exist in two molecular forms, known as isomers. However, in living materials, they occur only in one form (L-type), and after death they convert to the other form (D-type) with the passage of time until equilibrium is reached. This process is racemization (or epimerization for some amino acids). Racemization occurs

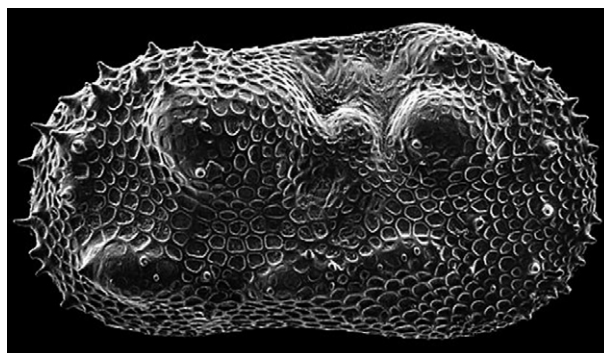


Figure 8 Ostracod valve suitable for amino acid racemization analysis: 0.8-mm-long SEM image of a freshwater cypridoidean ostracod found in British Pleistocene interglacial deposits of Marine Isotope Stage 9. Image by John Whittaker.

more rapidly when the environmental temperature is higher (e.g., during interglacials). As with any process occurring in biological materials, there are additional complications relating to structure and these have led to some early problems.

One of the earliest applications of AAR dating was to bones at a native American site in California. However, due to the development of AMS ^{14}C dating, the early (apparently pre-Clovis) ages were found to be incorrect. Since then, no further dating of bones has been attempted. Instead, the main applications of AAR have been to mollusks, both marine and non-marine, with isoleucine being the amino acid of choice (Wehmiller and Miller, 2000). Although a study has criticized use of the AAR technique for producing absolute ages, D/L ratios have been useful for providing correlations and relative stratigraphies over areas where there is little temperature variation. A more secure approach would be to use more than one amino acid, an approach currently being tested. New technological developments include the use of multiple collector ion gas chromatography and reverse phase chromatography and the analysis of free (rather than bound within the protein) amino acids. These have permitted the measurement of very small samples, such as a single Foraminifera or ostracod valve (Figure 8).

Another application of AAR has been to ostrich and other ratite bird eggshells. These have the advantage of having a mineral structure that is far less susceptible to diagenetic alteration during burial. Fossil eggshells are often found in association with an archaeological material and have provided chronological information at sites in southern Africa and Australia.

Summary

Besides the methods outlined in this brief overview, other time-dependent changes can be used to establish age equivalence, although they do not give rise to numerical ages. These and other methods are well described in Chapter 7 of Walker (2005). In addition, there are edited volumes covering many of the techniques (Taylor and Aitken, 1997; Noller et al., 2000), as well as review papers dealing with one or more of the techniques discussed here (Wintle, 1996).

See also: Dating Techniques; Dendrochronology; Fission-Track Dating; Geomagnetic Excursions and Secular Variations; K/Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ Dating; U-Series Dating. **Cosmogenic Nuclide Dating:** Cosmic Ray Interactions in Minerals; Exposure Geochronology; Landscape Evolution. **Luminescence Dating:** Electron Spin Resonance Dating; Optical Dating; Thermoluminescence. **Radiocarbon Dating:** AMS Radiocarbon Dating; Calibration of the ^{14}C Record; Conventional Method; Variations in Atmospheric ^{14}C .

References

- Aitken MJ (1985) *Thermoluminescence Dating*. London: Academic Press.
- Aitken MJ (1998) *An Introduction to Optical Dating*. Oxford: Oxford University Press.
- Cockburn HAP and Summerfield MA (2004) Geomorphological applications of cosmogenic isotope analysis. *Progress in Physical Geography* 28: 1–42.
- Dalrymple GB (1991) *The Age of the Earth*. Stanford, CA: Stanford University Press.
- Dalrymple GB and Lanphere MA (1969) *Potassium–Argon Dating*. San Francisco: Freeman.
- Davis R and Schaeffer OA (1955) Chlorine-36 in nature. *Annals of the New York Academy of Science* 62: 105–122.
- Duller GAT (2004) Luminescence dating of Quaternary sediments: Recent advances. *Journal of Quaternary Science* 19: 183–192.
- Dumitru TA (2000) Fission-track geochronology. In: Noller JS, Sowers JM, and Lettis WR (eds.) *Quaternary Geochronology: Methods and Applications*, pp. 131–155. Washington, DC: American Geophysical Union.
- Edwards RL, Gallup CD, and Cheng H (2003) Uranium-series dating of marine and lacustrine deposits. *Reviews in Mineralogy and Geochemistry* 52: 363–405.
- Eggins S, Grün R, McCulloch MT, et al. (2005) In situ U-series dating by laser-ablation multi-collector ICPMS: New prospects for Quaternary geochronology. *Quaternary Science Reviews* 24: 2523–2538.
- Fleischer RL, Price PB, and Walker RM (1975) *Nuclear Tracks in Solids*. Berkeley: University of California Press.
- Gosse JC and Phillips FM (2001) Terrestrial in situ cosmogenic nuclides: Theory and applications. *Quaternary Science Reviews* 20: 1475–1560.
- Grootes PM, Mook WG, Vogel JC, et al. (1975) Enrichment of radiocarbon for dating samples up to 75,000 years. *Zeitschrift für Naturforschung A* 30: 1–14.
- Grün R (1997) Electron spin resonance dating. In: Taylor RE and Aitken MJ (eds.) *Chronometric Dating in Archaeology*, pp. 217–260. New York: Plenum.
- Grün R and Stringer CB (1991) Electron spin resonance dating and the evolution of modern humans. *Archaeometry* 33: 153–199.
- Hare PE, Von Endt DW, and Kokis JE (1997) Protein and amino acid diagenesis dating. In: Taylor E and Aitken MJ (eds.) *Chronometric Dating in Archaeology*. New York: Plenum.
- Hedges REM (1981) Radiocarbon dating with an accelerator: Review and preview. *Archaeometry* 23: 3–18.
- Huntley DJ, Godfrey-Smith DI, and Thewalt MLW (1985) Optical dating of sediments. *Nature* 313: 105–107.
- Jacobs JA (1994) *Reversals of the Earth's Magnetic Field*. Cambridge, UK: Cambridge University Press.
- Lewin R (1987) *Bones of Contention*. New York: Simon & Schuster.
- Lewis C (2000) *The Dating Game*. Cambridge, UK: Cambridge University Press.
- Libby WL (1952) *Radiocarbon Dating*. Chicago: University of Chicago Press.
- McDougall I (1990) Potassium–argon dating in archaeology. *Science Progress* 74: 15–30.
- McDougall I and Harrison TM (1999) *Geochronology and Thermochronology by the $^{40}\text{Ar}/^{39}\text{Ar}$ Method*. New York: Oxford University Press.
- Niedermann S (2002) Cosmic-ray-produced noble gases in terrestrial rocks: Dating tools for surface processes. *Reviews in Mineralogy and Geochemistry* 47: 731–784.
- Noller JS (2000) Lead-210 geochronology. In: Noller JS, Sowers JM, and Lettis WR (eds.) *Quaternary Geochronology: Methods and Applications*, pp. 115–120. Washington, DC: American Geophysical Union.
- Noller JS, Sowers JM, and Lettis WR (2000) *Quaternary Geochronology: Methods and Applications*. Washington DC: American Geophysical Union.
- Pike AWG and Pettitt PB (2003) U-series dating and human evolution. *Reviews in Mineralogy and Geochemistry* 52: 607–630.
- Reimer PJ, Baillie MGL, Bard E, et al. (2004) IntCal04 terrestrial radiocarbon age calibration, 0–26 cal kyr BP. *Radiocarbon* 46: 1029–1058.
- Renne PR (2000) K–Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ dating. In: Noller JS, Sowers JM, and Lettis WR (eds.) *Quaternary Geochronology: Methods and Applications*, pp. 77–100. Washington, DC: American Geophysical Union.
- Richards DA and Dorale JA (2003) Uranium-series chronology and environmental applications of speleothems. *Reviews in Mineralogy and Geochemistry* 52: 407–460.
- Schwarz HP (1997) Uranium series dating. In: Taylor RE and Aitken MJ (eds.) *Chronometric Dating in Archaeology*, pp. 159–182. New York: Plenum.
- Stuiver M and Van der Plicht J (1988) IntCal98 calibration issue. *Radiocarbon* 40: 1041–1164.
- Taylor RE (1987) *Radiocarbon Dating*. London: Academic Press.
- Taylor RE and Aitken MJ (1997) *Chronometric dating in archaeology*. Plenum, New York: Advances in Archaeological and Museum Science.
- Taylor RE, Long A, and Kra RS (1992) *Radiocarbon after Four Decades: An Interdisciplinary Perspective*. New York: Springer-Verlag.
- Wagner GA and van den Haute P (1992) *Fission Track Dating*. Kluwer, The Netherlands: Dordrecht.
- Walker M (2005) *Quaternary Dating Methods*. Chichester, UK: Wiley.
- Walter RC (1989) Application and limitation of fission-track geochronology to Quaternary tephra. *Quaternary International* 1: 35–46.
- Walter RC (1997) Potassium–argon/argon–argon dating methods. In: Taylor RE and Aitken MJ (eds.) *Chronometric Dating in Archaeology*, pp. 97–126. New York: Plenum.
- Wehmiller JF and Miller GF (2000) Aminostratigraphic dating methods in Quaternary geology. In: Noller JS, Sowers JM, and Lettis WR (eds.) *Quaternary Geochronology: Methods and Applications*, pp. 187–222. Washington, DC: American Geophysical Union.
- Westgate J, Sandhu A, and Shane P (1997) Fission-track dating. In: Taylor RE and Aitken MJ (eds.) *Chronometric Dating in Archaeology*, pp. 127–158. New York: Plenum.
- Wintle AG (1996) Archaeologically-relevant dating techniques for the next century. *Journal of Archaeological Science* 23: 123–138.
- Wintle AG (1997) Luminescence dating: Procedures and protocols. *Radiation Measurements* 27: 769–817.
- Zeuner FE (1958) *Dating the Past*. London: Methuen.
- Zreda MG and Phillips FM (2000) Cosmogenic nuclide buildup in surficial materials. In: Noller JS, Sowers JM, and Lettis WR (eds.) *Quaternary Geochronology: Methods and Applications*, pp. 61–76. Washington, DC: American Geophysical Union.

Societal Relevance of Quaternary Research

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Introduction

The great nineteenth century geologist and natural historian, Charles Lyell (1830), is generally credited as being the author of the concept of Uniformitarianism (see [History of Quaternary Science](#)). Simply stated, this concept says that 'the present is the key to the past.' In other words, we may interpret the ancient history of the planet through our understanding of modern-day processes. This concept represented a major step forward in scientific thought, greatly influencing contemporary scientists, such as Charles Darwin (see [History of Quaternary Science](#)). However, in this article the author hopes to make it clear that the reverse of Lyell's concept is equally true: the past is the key to the present. In other words, knowledge of the past is key to our understanding of the present world, especially in the multiple disciplines known as the environmental sciences. No matter which aspect of the environment one considers, be it the atmosphere, the oceans, the cryosphere, the lithosphere, or the biosphere, it is impossible to fully comprehend modern processes and conditions without some knowledge of the past. To try to understand present-day environments without the knowledge of their Quaternary history would be like trying to understand the plot of a very long novel by reading only the last page.

Climate History

The Quaternary period saw a large number of climatic oscillations (see [Introduction](#)) on a scale that was probably greater than at any other time in the last 60 million years (Bradley, 1999). Thus, a study of Quaternary climate change provides us with an understanding of climatic variation on a much larger scale than has been recorded in historical records of the last few centuries. An understanding of the magnitudes and rates of climate change during the Quaternary period is necessary to develop our comprehension of modern climate. Much of climate modeling is based on data derived from reconstructions of ancient climates. These models require long-term data, which are available only in ancient climate records, reconstructed from various kinds of proxies, such as geochemical records from polar ice cores, and terrestrial and oceanic fossil records.

The predicted amplitude of global warming in the twenty-first century varies greatly from one scientific study to another, but in the absence of any mitigating policies, Wigley and Raper (2001) predicted with a 90% probability that by the end of the twenty-first century, global mean temperatures would have risen by 1.7–4.9° C. The [Intergovernmental Panel on Climate Change \(2007\)](#) estimated that in the next two decades, global mean temperatures will rise by 0.2° C per decade. Subsequent

levels of warming will depend on the levels of greenhouse gas emissions. The warming to occur in the coming decades is certainly going to be unprecedented within recorded history, but it falls within the boundaries of more than one previous interglacial period, including the last interglacial (ca. 120 000–115 000 years ago), when mean temperatures were substantially warmer than they are today in many parts of the world. Evidence from the Vostok Ice Core in Antarctica (see [Antarctic Stable Isotopes](#)) indicates that at the beginning of the interglacial, between about 130 000 and 127 000 years ago, regional temperatures climbed as much as 3° C higher than today (Figure 1; Kukla et al., 2002). As Velichko et al. (1993) observed, there is abundant proxy evidence of large-scale climatic change during the Quaternary, especially in the northern high latitudes. During the long series of glacial–interglacial cycles, high latitude temperatures appear to have fluctuated within about the same range that is projected for the next century (see [Paleoclimate Relevance to Global Warming](#)).

Predictions of future temperature changes, based on orbital forcing and increased greenhouse gas concentrations in the atmosphere (Loutre and Berger, 2003), suggest that anthropogenic (human-induced) effects may force global climate into an interglacial mode that may persist for 30 000–40 000 years, not unlike an interglacial period that occurred about 405 000–340 000 years ago.

Pace of Climate Change

Global warming is predicted to increase at a much faster rate in the coming decades than the rates of temperature change observed in recent centuries. For instance, Vinnikov and Grody (2003) analyzed upper atmospheric (troposphere) temperatures for the interval 1978–2002, and from these data they inferred a future global warming trend of 0.22–0.26° C per decade. Has the Earth previously experienced temperature changes as rapid as this? Fossil beetle evidence from Britain indicates that even greater rates of warming have occurred in the past. For instance, at the end of the last glaciation, the temperature change during the transition from the glacial climates of the Younger Dryas interval to the postglacial climates at the beginning of the current interglacial was extremely rapid and intense. The beetle evidence (see [Late Pleistocene of Europe](#)) suggests that mean summer temperatures warmed on the order of 0.35° C per decade during this transition (Figure 2; Lowe et al., 1995). This climate change led to wholesale changes in the flora and fauna of many regions. We cannot hope to predict the kinds of changes that may take place in the modern biota without some knowledge of how past biotic communities responded to large-scale climate change.

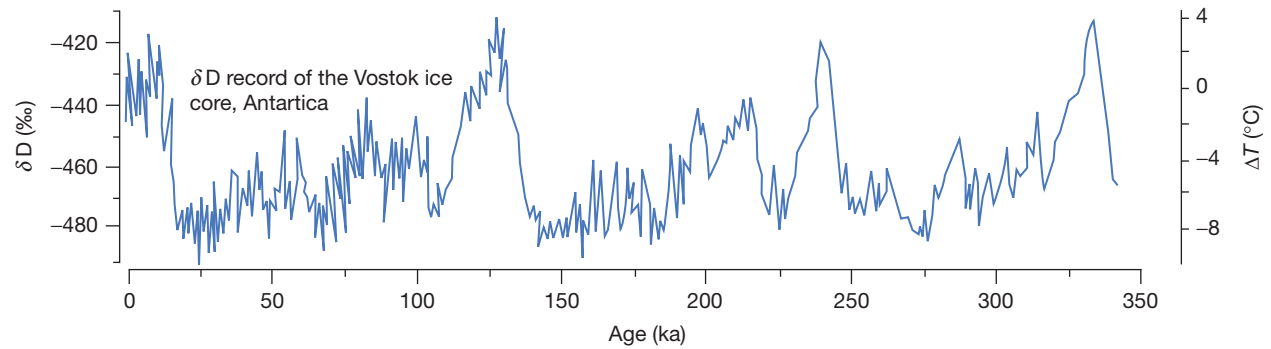


Figure 1 Deuterium isotope record from the Vostok ice core, Antarctica, showing large-scale fluctuations in temperature during the last four glacial–interglacial cycles. Reproduced from Petit JR, Jouzel J, Raynaud D, et al. (1999) Climate and atmospheric history of the past 420,000 years from the Vostok ice core, Antarctica. *Nature* 399: 429–436.

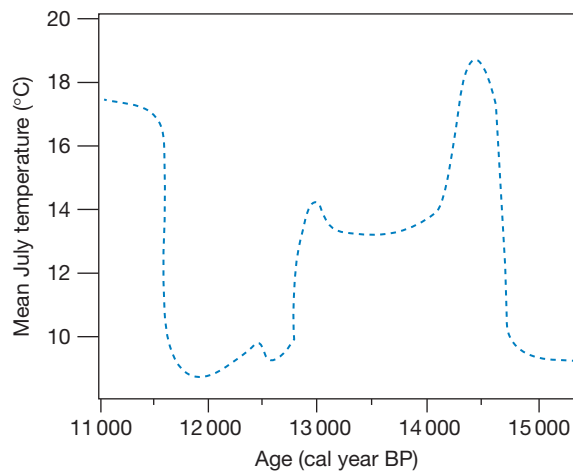


Figure 2 Reconstructed mean July temperature change during the interval 15,000–11,000 years ago in Britain, based on insect fossil analysis. Data from Lowe et al., 1995.

Ocean History

By studying the Quaternary fossil record from deep-sea sediments, paleoceanographers have come to understand the role of oceans in the transfer of heat from the tropics to the high latitudes, via ocean currents (see [Paleoceanography An Overview](#)). Broecker's (1987) model of the oceanic thermohaline conveyor belt demonstrated the manner of functioning of this system, both in the past and in the present. This thermohaline conveyor belt acts as a major conduit for the exchange of thermal energy, nutrients, and dissolved oxygen between the shallow and deep oceans of the world. There is little doubt that the conveyor belt's capability to transfer heat in large volumes and across vast distances profoundly influences global climate. There is some evidence that the current climatic warming, especially in the high latitudes, may be weakening the thermohaline circulation (Häkkinen and Rhines, 2004). Recent oceanographic research (Curry and Mauritzen, 2005) suggests that the northern regions of the North Atlantic Ocean have been diluted by increasing amounts of freshwater since the 1960s. Climate models indicate that, if the conveyor belt were to shut down, much of Europe would

be 5–10° C colder, the equatorial regions would be 4–5° C warmer, and Greenland would be as much as 16° C colder than it is today. Rainfall patterns would radically change, and the atmosphere would become dustier (Broecker, 1999). Again, without the paleontological data from Quaternary marine sediments, we would have little or no understanding of these phenomena (Figure 3).

Sea Level Change

There is a growing body of evidence that the current level of global warming is going to bring about substantial rises in sea level, as polar ice caps melt. The Intergovernmental Panel on Climate Change (IPCC) (2007) has estimated that sea level will rise 50 cm above current levels by the year 2100 (Figure 4). This is a substantially greater rise in sea level than occurred in the last century. The ecological and human impacts of rising oceans would be substantial, including increased flooding, coastal erosion, salination of aquifers, and loss of coastal agricultural land and living space (IPCC, 2001). The loss of living space is particularly critical, given that 50% of the world's population lives within 6 km of the sea, and that 14 of the 15 largest cities are coastal.

Again, one must refer to the Quaternary record to find sea levels that were as high, or even higher than, the levels predicted for the coming century. Reconstructions of global mean sea level during the last interglacial period (see [Eustatic Sea-Level Changes – Glacial–Interglacial Cycles](#)) place it 6–7 m above modern mean sea level at about 125,000 years ago (Bard et al., 1993; Chappell and Shackleton, 1986). By studying the effects that these higher sea levels had on ancient coastal landscapes, we are able to develop much more useful models of what may happen in the future.

Coastal Deposition and Erosion

As sea levels rise and storm intensity increases with global warming, many of the world's coastlines are being eroded (Figure 5). Needless to say, the impacts on society are relatively large and costly. For instance, a report by the Heinz Center for Science, Economics, and the Environment states that in the United States, approximately 1500 homes will be lost to coastal erosion each year for the next several decades at an

The Atlantic thermohaline circulation
a key element of the global oceanic circulation

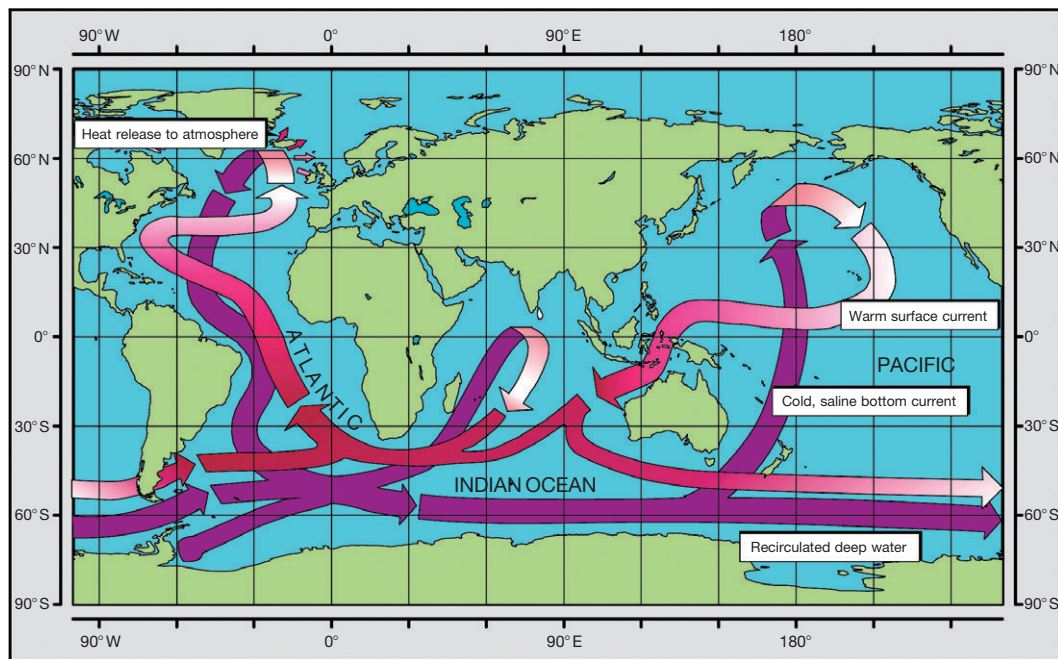


Figure 3 Schematic diagram of the global ocean circulation pathways, the 'conveyor' belt (reproduced from W. Broecker, modified by E. Maier-Reimer).

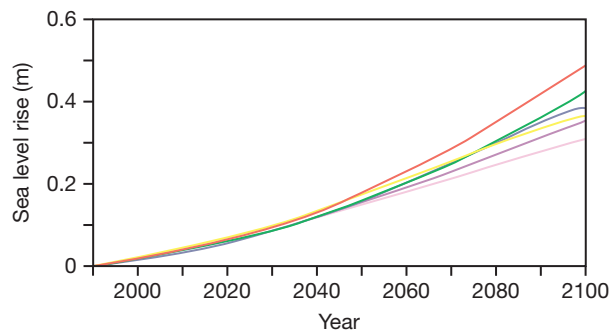


Figure 4 Predicted global sea level rise during the twenty-first century, based on six different models. Data from the Intergovernmental Panel on Climate Change (2001).



Figure 5 Coastal erosion photograph from Peter French.

annual cost to coastal landowners of approximately \$530 million (Heinz Center, 2000). The predictions for coastal change under the scenario of global sea level rise warn of impending disaster for the variety of coastal features and their associated habitats (Psuty and Silveira, 2010).

Of course coastal erosion has been going on for many thousands of years, and has been documented by marine geologists studying Quaternary near-shore deposits. What is happening today is just the most recent episode in an ongoing Holocene marine transgression. The erection of coastal defenses has altered the process, at least temporarily, but ultimately the phenomenon is inexorable. Quaternary geologists have been studying such processes as the supply and sink of coastal sediments, interactions of the retreating shore face with older deposits, and long-term changes in the direction of shoreline changes (see *Sedimentary Indicators of Relative Sea-Level Changes – High Energy*) (McNinch, 2004). They have found that variations in the rates and directions of shoreline change are strongly controlled by changes in ocean wave and tidal energy (van der Molen and van Dijk, 2000). These are highly complex problems that are very difficult to model, but data from Quaternary studies are essential to the process.

Environmental Change in Oceanic Ecosystems

The world's continental shelves are highly productive marine ecosystems (Lohrenz et al., 2002). However, we do not fully understand the role of continental shelves in the global climate system. The science of paleoceanography is providing the kind of information needed to gain a better understanding of

this role. Specifically, we need to know more about changes in continental shelf productivity during times of different sea levels, the storage of nutrients on continental shelves, and the effects of changes in sea level on the total sea surface area. As with the other problems discussed here, these are not just ‘academic’ issues. For instance, they affect the biological productivity and diversity of the oceans, both of which have direct implications for the world’s fisheries.

Cryosphere History

The waxing and waning of ice sheets has been the dominant environmental phenomenon of the Quaternary, either directly or indirectly affecting every region of the globe. In the current climatic conditions, polar ice caps are thinning, mountain glaciers are retreating in most regions, and many are in danger of disappearing altogether. Notable among these are the glaciers and snowfields that are perched on the tops of high mountains near the Equator (see [Africa](#)). A recent study of Mount Kilimanjaro’s summit in Africa (Thompson et al., 2009) showed that 85% of the ice cover that was present in 1912 has disappeared, and that 26% of the ice cover that was present in 2000 has now disappeared. At the current rate of melting, this ice cap will be gone within a few years (Figure 6). This study, based on ice cores from Kilimanjaro, revealed that the processes driving the current shrinking and thinning of the Kilimanjaro ice fields are unique and unrecorded within the last 11 700 years. Similarly, the massive Quelccaya ice cap in the Peruvian Andes is melting (see [South America](#)). The ice cap shrank from 56 km² in 1976 to just 44 km² in 2001, and it may disappear within the next 20 years.



Figure 6 Aerial photograph of the top of Mount Kilimanjaro, showing the remnants of the ice cap. Photograph courtesy of Byrd Polar Center, Ohio State University.

The Antarctic and Greenland ice sheets together hold 33 million cubic km of ice, representing enough water to raise global sea level by 70 m (Rignot and Thomas, 2002). Current research by several different research groups indicates that the Greenland Ice Sheet is thinning, especially near the coasts (see [Dynamics of the Greenland Ice Sheet](#)). The Antarctic story is more complicated. It appears that the West Antarctic Ice Sheet is thinning (Figure 7; see [Dynamics of the West Antarctic Ice Sheet](#)) overall, but it is thickening in some regions and thinning in others (Rignot and Thomas, 2002). Recent modeling of the West Antarctic Ice Sheet by Bamber et al. (2009) indicates that its collapse would raise global sea level by an average of 3.3 meters. Large sectors of ice in south-east Greenland, West Antarctica, and the Antarctic Peninsula are changing quite rapidly, but glaciologists do not yet understand all the processes involved. Modeling is a key element in developing that understanding, and the best sources of data for such models come from paleoenvironmental research. It is clear that just as global climate affects polar ice sheets, so also ice sheets affect global climate. As Clark et al. (1999) have said, given the importance of ice sheets in the climate system, establishing the factors that control their evolution and behavior is necessary for understanding their influence on climate over the long term. This applies equally to questions about the future stability of the polar ice sheets.

Lithosphere History

The study of geomorphology, or the processes that act to shape the surface of the Earth, is rooted firmly in Quaternary science. Understanding Earth’s geomorphic history is the key to understanding current geomorphological problems. Knowledge of contemporary earth surface processes is critical to the study of the hazards and sustainability of the landscape.

Soil Studies

Modern soils in the rich agricultural regions of the middle latitudes of the Northern Hemisphere owe much of their formation to Pleistocene glaciations. Glaciers and ice sheets



Figure 7 Photograph of Pine Island Glacier taken by Tom Kellogg onboard the US Coast Guard icebreaker ‘Glacier,’ 1985, in Pine Island Bay.

pulverized layers of bedrock and organic matter that fell within the ice boundaries, creating the mixtures of sand and silt, and clay that became the best soils in which to grow crops. This is due to their moisture retention characteristics and nutrient storage ability. These soils can be considered part of the world's Pleistocene heritage (see [Overview](#)).

These rich agricultural soils are the product of thousands of years of glacial and postglacial environments. They are quite durable, yet susceptible to degradation through overuse and poor land management practices. Soils in other regions of the world vary in their ability to sustain agriculture. By one estimate ([Oldeman et al., 1991](#)), about 15% of Earth's land surface has been seriously degraded by human activities ([Table 1](#)). Perhaps 25% more of Earth's soils are at risk of serious degradation ([World Resources Institute, 1990](#)). The loss of agricultural land through erosion is estimated at 6 or 7 million hectares (ha) per year, with an additional annual loss of 1.5 million ha as a result of waterlogging, salinization, and alkalization ([Brundtland et al., 1987](#)). Degradation of dry-land soils is now thought to be decreasing plant productivity by 20–40% across the arid and semiarid regions of the globe ([Zika and Erb, 2009](#)). The ever-expanding human population of the world is essentially using up the vital resource of arable soil at an alarming rate. It took many thousands of years for these soils to develop, but only a few centuries for them to degrade.

Geologic Hazards

Our understanding of most geologic hazards has been greatly enhanced by Quaternary studies. The reason we must probe into the past to understand these phenomena is that most of them are highly sporadic and largely unpredictable, despite the best efforts of geoscientists. Catastrophes such as large volcanic eruptions, earth quakes, and tsunamis, take place only rarely in any given region. These phenomena are discussed in a document published by the [United Nations Environment Program \(2005\)](#). The interval between events may be hundreds or even thousands of years – well beyond the span of recorded history. The Quaternary record affords us insights into the timing and intensity of past catastrophic events.

Table 1 1991 soil degradation statistics by region

Region	Human-induced soil degradation (million ha)	Total land surface (million ha)	Percent of land with soil degradation
Africa	494	2966	17
Asia	748	4256	18
South America	243	1768	14
Central America	63	306	21
North America	95	1885	5
Europe	219	950	23
Australasia	103	882	12
World	1964	13013	15

Source: [Oldeman LR, Hakkeling RTA, and Sombroek WG \(1991\) World Map of the Status of Human-Induced Soil Degradation: An Explanatory Note](#), 2nd edn. Wageningen, The Netherlands: International Soil Reference and Information Centre.

Volcanic Eruptions

Sitting astride the boundaries of the world's tectonic plates, volcanoes are a major hazard to human life in many parts of the world. Volcanic eruptions have claimed more than a quarter of a million lives since the destruction of Pompeii in AD 79 ([Table 2](#)). The principal hazard is not lava, which usually flows sufficiently slowly to allow all but the closest inhabitants to the volcano to escape. Volcanic ash, or tephra, is the principal hazard. It is ejected in far larger quantities than lava in most eruptions ([Figure 8](#)), and it can travel thousands of kilometers, often circling the globe in a plume that is visible from space ([Figure 9](#)). In rare cases, ignimbrites are ejected during volcanic eruptions. These are pyroclastic rocks consisting of glass shards, crystals, and lithic fragments. Ignimbrites are formed by the deposition of hot, rapidly expanding, turbulent magmatic gases. This happened in the 1902 eruption of Mt. Pelee on the Caribbean island of Martinique, killing all but two

Table 2 Major historical volcanic eruptions

Locality	Date	Estimated human death toll
Pompeii and Herculaneum, Italy (Vesuvius)	AD 79	Thousands (uncounted)
Iceland (Laki volcano)	AD 1783	9350
Kyushu, Japan	AD 1793	14 300 (triggered avalanche and tsunami)
Sumbawa, Indonesia (Tambora volcano)	AD 1815	92 000 (10 000 directly; 82 000 through starvation and disease)
Central Ecuador (Mt. Cotopaxi)	AD 1877	1000
Krakatau, Indonesia	AD 1883	36 000
St. Vincent, West Indies (Soufrière volcano)	AD 1902	1680
Martinique, West Indies (Mt. Pelée)	AD 1902	40 000
Northern Columbia (Nevada del Ruiz)	AD 1985	25 000
Total loss of life		>250 000



Figure 8 Photograph of volcanic eruption, Kanaga volcano, Aleutian Islands, Alaska, 1994. Photograph courtesy of US Geological Survey.

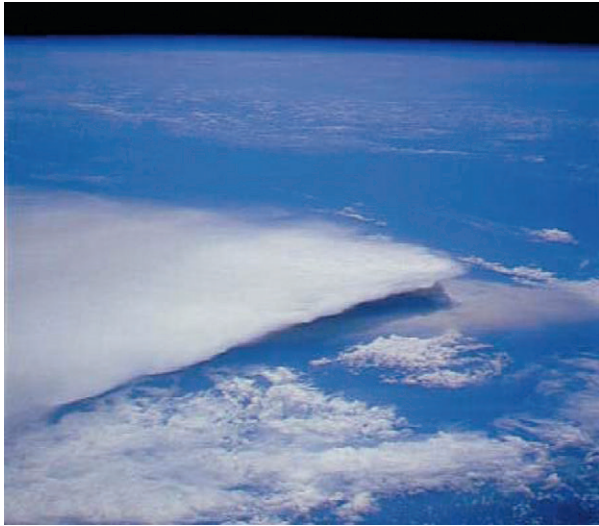


Figure 9 Eruption of Rabaul volcano, Papua New Guinea, 1994, as seen from the space shuttle. Photograph courtesy of NASA.

people on the island. When volcanic ash and rock mix with water, they may form a lahar, which comes sweeping rapidly down the mountain into surrounding valleys. Volcanic tephra layers deposited in Quaternary sediments are the chief means by which scientists reconstruct the history of prehistoric eruptions (see [Tephrochronology](#)). We can say with absolute certainty that some of the volcanic eruptions that took place in the Pleistocene were far greater, and affected larger areas, than anything in recorded history. For instance, 600 000 years ago, the central part of what is now Yellowstone National Park, Wyoming, exploded in an enormous volcanic eruption. The eruption spewed out nearly 1000 km³ of debris. What is now the park's central portion then collapsed, forming a 45- by 75-km caldera (or basin). Tephra from this eruption is known to have blanketed more than 5 million square kilometers of western and central North America ([Smith and Siegel, 2000](#)). If an eruption on this scale were to happen today, it would ecologically devastate large regions, and cause climatic cooling for several years afterwards because of the screening of sunlight by volcanic ash and gases in the atmosphere. To put this Yellowstone eruption into context, the largest volcanic eruption in historic times was in Tambora, in Indonesia, in 1815. This eruption is estimated to have ejected about 150 km³ of ash into the atmosphere ([Oppenheimer, 2003](#)). This is less than one-third the amount estimated for the Yellowstone eruption. Anomalously cold weather hit the northeastern USA, the maritime provinces of Canada, and Europe the following year. 1816 came to be known as the 'Year without a summer' in these regions.

The Yellowstone volcanic eruption story is a cautionary tale. Far greater eruptions have taken place in prehistory than have happened in historic times. Our knowledge of the prehistoric eruptions has come through the analysis of ancient volcanic deposits. Tephrochronology, the study of volcanic ash deposits, has helped scientists determine the source of ash deposits (through chemical fingerprinting). By piecing together regional histories of volcanic activity, Quaternary



Figure 10 Satellite image of a tsunami striking the coast of Sri Lanka following the massive earthquake in the Indian Ocean, on 29 December 2004. The image shows the backwash of the tsunami. Photograph copyright DigitalGlobe.

scientists have been able to reconstruct the size and timing of ancient volcanic eruptions. By working with modern volcanologists, we have begun the difficult task of predicting volcanic eruptions. The science of volcanic eruption prediction is far from exact, but it has enabled the evacuation of some threatened regions *before* major eruptions took place. For instance, in 1991 volcanologists were able to warn the people living near Mount Pinatubo in the Philippines a few days before devastating eruptions took place. In total, 58 000 people were evacuated from a 30-km radius around the volcano ([Wolfe and Hoblitt, 1996](#)).

Earth Quakes and Tsunamis

There is no doubt that earthquakes and the resulting tsunamis ([Figure 10](#)) have killed more people in historic times than any other geologic hazard. As shown in [Table 3](#), major earthquakes in the past 1200 years are known to have killed more than 4 million people. This total includes more than 2 million people in just the twentieth century. As human populations increase, and the number of poorly built multi-story buildings goes up, these numbers are only likely to grow larger. But, like volcanic eruptions, earthquakes are relatively rare events in any one region, and we have yet to develop a meaningful way of predicting them. Geoscientists may be able to predict that there will be a major earthquake somewhere along a given fault system, but as of now they can only say that it will likely happen within the next decade or the next century. In fact, some of the most useful data in the study of earthquakes comes from Quaternary science. Ancient earthquakes have left behind clear indicators in buried sediments. For instance, [Fumai et al. \(1993\)](#) were able to reconstruct a 100-year recurrence interval for the San Andreas Fault zone, 70 km northeast of Los Angeles, based on sedimentary evidence. Earthquakes caused debris-flow deposits, ruptured, tilted, and folded bedding planes in the sediments.

The timing of an earthquake in the year AD 1700 along the coast of Washington state, USA, was able to be accurately dated through analysis of tree-rings (see [Dendrochronology](#)). [Yamaguchi et al. \(1997\)](#) studied Japanese historical evidence for a large tsunami of previously unknown origin and were

Table 3 Major historical earthquakes (>10 000 casualties)

<i>Locality</i>	<i>Date</i>	<i>Estimated human death toll</i>
Damghan, Iran	AD 857	200 000
Ardabil, Iran	AD 893	150 000
Aleppo, Syria	AD 1138	230 000
Chihli, China	AD 1290	100 000
Shaanxi province, China	AD 1556	830 000
Shemakha, Caucasasia	AD 1667	80 000
Sicily, Italy	AD 1693	60 000
Tabriz, Iran	AD 1727	77 000
Lisbon, Portugal	AD 1755	70 000
Calabria, Italy	AD 1783	50 000
Messina, Sicily	AD 1908	70 000–100 000
Avezzano, Italy	AD 1915	29 980
Gansu province, China	AD 1920	200 000
near Tokyo, Japan	AD 1923	>140 000
Xining, China	AD 1927	200 000
Gansu, China	AD 1932	70 000
Quetta, Pakistan	AD 1935	30 000–60 000
Northern Turkey	AD 1939	100 000
Ashgabat, Turkmenistan	AD 1948	110 000
Assam, India	AD 1950	20 000–30 000
Peru	AD 1970	66 000
Tangshan, China	AD 1976	655 000
Tabas, Iran	AD 1978	25 000
Mexico City	AD 1985	25 000
Armenia	AD 1988	25 000
Northwest Iran	AD 1990	>50 000
Izmit, Turkey	AD 1999	>17 000
Bhuj, India	AD 2001	>20 000
Bam, Iran	AD 2003	>30 000
Sumatra, Indonesia (marine earthquake and tsunami)	AD 2004	>226 000
Total loss of life from major historic earthquakes		>4 000 000

able to pin down the source of the earthquake that caused it. They proposed that a magnitude 9 earthquake in the Cascadia subduction zone of the Pacific Northwest region of North America was the trigger for the tsunami that occurred on 26 January 1700. Tree-ring records from the central Cascadia region (**Figure 11**) support this theory. When the earthquake struck, much of the coast between southern British Columbia and northern California was abruptly lowered, submerging some coastal forests in more than a meter of tidewater. Yamaguchi et al. studied the annual growth rings of some Sitka spruce trees that survived this tidal submergence, and the rings in some trees show changes in width and anatomy consistent with disturbance (tilting, increased flooding, or both) in AD 1700 and a few years afterward.

Tsunamis may also occur in tectonically rather stable areas, as shown by the Storegga Slide that struck the coasts of Scotland and Norway, about 7000 years BP. (Dawson et al., 1988).

Biosphere History

Our knowledge of how ecosystems function has been greatly aided by Quaternary paleontological studies. In the absence of

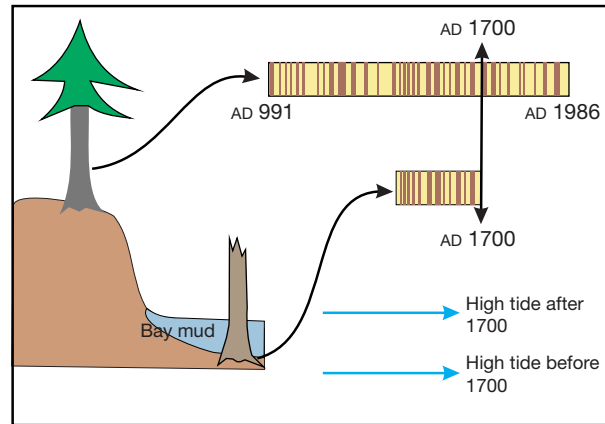


Figure 11 Tree-ring dating of the 1700 Cascadia earthquake. The dead tree on the right was drowned in salt water due to subsidence during the earthquake. The outermost tree-rings from a number of dead snags in regional salt marshes correspond to the year AD 1700 in tree-ring sequences from living (upland) trees in the region.

such studies, modern ecologists would have little or no idea about the longevity of ecosystems and their resistance to change in the face of large-scale environmental variations.

Longevity, Resistance, and Resilience of Ecosystems

Studies of plant and animal remains from the Quaternary have shown that the current ecosystems are not composed of species that are inherently the best suited for their environments. Rather, modern ecosystems represent just the latest reshuffling of species, and the mixture of species in biological communities at any given time has been shaped by many forces. In the middle- and high latitudes, the catalyst for changes in ecosystems has been large-scale climatic fluctuations of the Pleistocene, accompanied by the waxing and waning of continental ice sheets and mountain glaciers (see [Evidence of Glacier and Ice Sheet Extent](#)). There have been hundreds of these large-scale fluctuations in climate during the last 2.4 million years. Amazingly, some groups of organisms (notably beetles) have remained intact throughout this interval, shifting their ranges in response to climatic change (see [Overview](#)). Other groups, such as mammals, have undergone waves of speciation during the Quaternary (see [Vertebrate Overview](#)). This includes, of course, the hominid lineage that gave rise to our species (see [Overview](#)). Towards the end of the last glaciation, a large number of megafaunal mammals (species with adult weight greater than 40 kg) became extinct. During the middle of the last glaciation, more than 150 genera of megafauna were alive on the planet. By 10 000 years ago, this number was reduced to just over 50 genera (see [Late Pleistocene Megafaunal Extinctions](#)). Thus, approximately two-thirds of the large mammal groups became extinct during this interval ([Figure 12](#)). This extinction event has been the subject of hot scientific debate for the last 40 years.

Pleistocene Megafaunal Extinctions

The conclusion of the debate over the megafaunal mammal extinctions at the end of the Pleistocene rests on determining

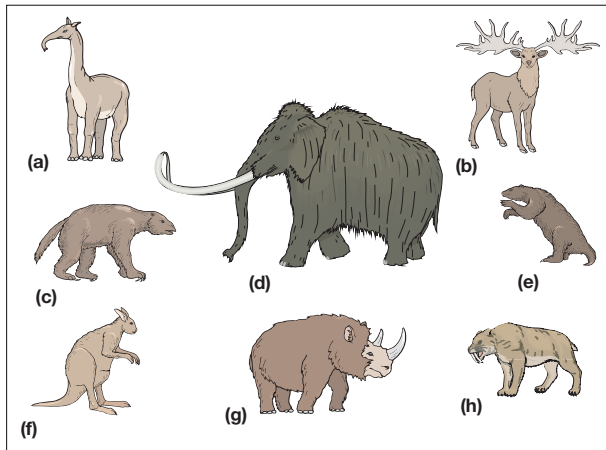


Figure 12 Megafaunal mammals that became extinct towards the end of the last glaciation. (a) *Macrauchenia* (South American browser – no living relatives); (b) *Megaloceros* (Giant deer); (c) *Eremotherium* (Ground sloth); (d) *Mammuthus* (Woolly mammoth); (e) *Megatherium* (Giant ground sloth); (f) *Procoptodon* (Giant short-faced kangaroo); (g) *Coelodonta* (Woolly rhinoceros); (h) *Smilodon* (saber-tooth cat).

its cause. Paleontological evidence has shown that previous glacial-to-interglacial transitions in the Pleistocene did not bring about wholesale extinctions of species. This evidence has led some scientists to conclude that human predation was what dealt the final blow to the Pleistocene megafauna, at the end of the last Ice Age (Martin and Klein, 1989; Flannery, 1994). Others believe that changing environments brought about this extinction, and express the view that humans had little or nothing to do with it (Grayson and Meltzer, 2003). A third group of scientists holds a middle view that perhaps people played a part in megafaunal extinction, but that the degree of human impact has yet to be determined (Barnowsky et al., 2004).

Modern Extinction Rates

We may never find the ‘smoking gun’ for the end-of-Pleistocene extinction event, but the fossil record demonstrates that such large-scale extinction events have been rare in Earth’s history. There have been at least five well-documented mass extinctions in the planet’s history. Some, such as the dinosaur extinction, 65 million years ago, are thought to have been caused by meteor impacts. The causes of other extinction events remain unknown. However, the causes of the modern extinction of species are all too obvious. As human populations swell and human land use alters once-pristine ecosystems, the biodiversity of the planet is more and more degraded. For the past 300 years, recorded extinctions of some groups of organisms have shown rates of extinction at least several hundred times the rate expected on the basis of the geological record (Dirzo and Raven, 2003). Even if this rate of extinction has been exaggerated, as suggested by some authors (Stork, 2010), it cannot be denied that the loss and fragmentation of natural habitats is severely affecting the viability of untold numbers of species around the globe.

Human ‘Coming of Age’ in the Quaternary

As we have seen, the study of Quaternary environments and organisms forms the basis for gauging much of what is happening in the modern world. Quaternary science provides the necessary information for assessing the potential impacts of future global warming, recurrence rates for geologic hazards, modern species extinction rates, and a host of other topics of vital interest to humanity. Our own species, *Homo sapiens*, first appeared only about 160 000 years ago (Clark et al., 2003). Although modern societies are quite capable of shaping their immediate environment to suit their needs, for more than 90% of human history, the environment has been shaping us. Our species, and its immediate ancestors, were all products of the Pleistocene, first shaped by environmental changes in Africa, followed by Eurasia, finally colonizing most of the globe within the last 12 000 years (since the end of the last glaciation; see Overview). *Homo sapiens* means ‘wise man.’ Will our species live up to its scientific name?

References

- Bamber JL, Riva REM, Vermeersen BLA, and LeBrocq AM (2009) Reassessment of the potential sea-level rise from a collapse of the West Antarctic Ice Sheet. *Science* 324: 901–903.
- Bard E, Stuiver M, and Shackleton N (1993) How accurate are our chronologies of the past? In: Eddy JA and Oeschger H (eds.) *Global Changes in the Perspective of the Past*, pp. 109–120. New York: Wiley.
- Barnowsky AD, Koch PL, Feranec RS, Wing SL, and Shabel AB (2004) Assessing the causes of Late Pleistocene extinctions on the continents. *Science* 306: 70–75.
- Bradley R (1999) *Paleoclimatology. Reconstructing Climates of the Quaternary*, Second edition. San Diego: Academic Press.
- Broecker WS (1987) The biggest chill. *Natural History* 96: 74–82.
- Broecker W (1999) What if the conveyor were to shut down? *Geological Society of America Today* 9: 1–6.
- Brundtland GH, Khalid M, et al. (1987) Our common future. *Report of World Commission on Environment and Development Presented to the Chairman of Intergovernmental Intersessional Preparatory Committee, UNEP Governing Council*. Oxford: Oxford University Press.
- Chappell J and Shackleton NJ (1986) Oxygen isotopes and sea level. *Nature* 324: 137–140.
- Clark PU, Alley RB, and Pollard D (1999) Northern hemisphere ice-sheet influences on global climate change. *Science* 286: 1104–1111.
- Clark JD, Beyene Y, Wolde-Gabriel G, et al. (2003) Stratigraphic, chronological and behavioural contexts of Pleistocene *Homo sapiens* from Middle Awash, Ethiopia. *Nature* 423: 747–752.
- Curry R and Mauritzen C (2005) Dilution of the Northern North Atlantic Ocean in recent decades. *Science* 308: 1772–1774.
- Dawson AG, Long D, and Smith DE (1988) The Storegga slides: Evidence from Eastern Scotland for a possible Tsunami. *Marine Geology* 82: 271–276.
- Dirzo R and Raven PH (2003) Global state of biodiversity and loss. *Annual Review of Environment and Resources* 28: 137–167.
- Flannery TF (1994) *The Future Eaters. An Ecological History of Australasian Lands and People*. Sydney: Reed New Holland.
- Fumai TE, Pezzopane SK, Weldon RJ, and Schwartz DP (1993) A 100-year average recurrence interval for the San Andreas Fault at Wrightwood, California. *Science* 259: 199–203.
- Grayson DK and Meltzer DJ (2003) A requiem for North American overkill. *Journal of Archaeological Science* 30: 585–593.
- Häkkinen S and Rhines PB (2004) Decline of subpolar North Atlantic circulation during the 1990s. *Science* 304: 555–559.
- Heinz Center (2000) Evaluation of erosion hazards, a collaborative research project of the H. John Heinz III Center for Science, Economics and the Environment. Website at <http://www.heinzcenter.org/>.
- Intergovernmental Panel on Climate Change (2007) *IPCC Fourth Assessment Report: Climate Change 2007*. Geneva: IPCC.

- Kukla GJ, Bender ML, deBeaulieu J-L, et al. (2002) Last interglacial climates. *Quaternary Research* 58: 2–13.
- Lohrenz SE, Redalje DG, Verity PG, Flagg CN, and Matulewski KV (2002) Primary production on the continental shelf off Cape Hatteras, North Carolina. *Deep Sea Research Part II: Topical Studies in Oceanography* 49: 4479–4509.
- Loutre MF and Berger A (2003) Marine Isotope Stage 11 as an analogue for the present interglacial. *Global and Planetary Change* 36: 209–217.
- Lowe JJ, Coope GR, Sheldrick C, Harkness DD, and Walker MJC (1995) Direct comparison of UK temperatures and Greenland snow accumulation rates, 15000–12000 yr ago. *Journal of Quaternary Science* 10: 175–180.
- Lyell C (1830) *Principles of Geology*. London: John Murray.
- Martin PS and Klein RG (1989) *Quaternary Extinctions: A Prehistoric Revolution*, Second edition. Tucson, AZ: University of Arizona Press.
- McNinch JE (2004) Geologic control in the nearshore: Shore-oblique sandbars and shoreline erosional hotspots, Mid-Atlantic Bight, USA. *Marine Geology* 211: 121–141.
- Oldeman LR, Hakkeling RTA, and Sombroek WG (1991) *World Map of the Status of Human-Induced Soil Degradation: An Explanatory Note*, 2nd edn. Wageningen, The Netherlands: International Soil Reference and Information Centre.
- Oppenheimer C (2003) Climatic, environmental and human consequences of the largest known historic eruption: Tambora volcano (Indonesia) 1815. *Progress in Physical Geography* 27: 230–259.
- Petit JR, Jouzel J, Raynaud D, et al. (1999) Climate and atmospheric history of the past 420,000 years from the Vostok ice core, Antarctica. *Nature* 399: 429–436.
- Psuty NP and Silveira TM (2010) Global climate change: An opportunity for coastal dunes? *Journal of Coastal Conservation* 14: 153–160.
- Rignot E and Thomas RH (2002) Mass balance of polar ice sheets. *Science* 297: 1502–1506.
- Smith RB and Siegel L (2000) *Windows into the Earth: The Geologic Story of Yellowstone and Grand Teton National Park*. New York: Oxford University Press.
- Stork NE (2010) Re-assessing current extinction rates. *Biodiversity and Conservation* 19: 357–371.
- Thompson LG, Brecher HH, Mosley-Thompson E, Hardy DR, and Mark BG (2009) Glacier loss on Kilimanjaro continues unabated. *Proceedings of the National Academy of Sciences* 106: 19770–19775.
- United Nations Environment Program (2005) *One Planet Many People: Atlas of Our Changing Environment*. Nairobi, Kenya: United Nations Environment Program 336 pp.
- van der Molen J and van Dijk B (2000) The evolution of the Dutch and Belgian coasts and the role of sand supply from the North Sea. *Global and Planetary Change* 27: 223–244.
- Velichko AA, Borissova OK, Zelikson EM, et al. (1993) Greenhouse warming and the Eurasian biota: Are there any lessons from the past? *Global and Planetary Change* 7: 51–67.
- Vinnikov KY and Grody NC (2003) Global warming trend of mean tropospheric temperature observed by satellites. *Science* 302: 269–272.
- Wigley TML and Raper SCB (2001) Interpretation of high projections for global-mean warming. *Science* 293: 451–454.
- Wolfe EW and Hoblitt RP (1996) Overview of the eruptions. In: Newhall CG and Punongbayan S (eds.) *Fire and Mud, Eruptions and Lahars of Mount Pinatubo, Philippines*, pp. 1–5. Seattle: University of Washington Press.
- World Resources Institute, World Resources 1990–1991 (1990). *A Guide to the Global Environment. World Resources Institute in collaboration with the United Nations Environment Programme and the United Nations Development Programme*. Oxford: Oxford University Press.
- Yamaguchi D, Atwater BF, Bunker DE, Benson BE, and Reid MS (1997) Tree-ring dating the 1700 Cascadia earthquake. *Nature* 389: 922–923.
- Zika M and Erb K-H (2009) The global loss of net primary production resulting from human-induced soil degradation in drylands. *Ecological Economics* 69: 310–318.

Understanding Quaternary Climatic Change

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Introduction

Quaternary science is an interdisciplinary subject concerned with the history of environmental changes on Earth during the past 2.6 My. It is the goal of Quaternary scientists to interpret the geological record from this period in order to understand the key processes (physical, chemical, biological, atmospheric, human-induced) that trigger and modulate environmental change at all geographic and temporal scales. Given this very wide scope, it is not surprising that the technical and intellectual tools required to achieve this goal have been drawn from almost every scientific discipline. The accurate dating of many Quaternary events, for example, became possible only after major developments took place in atomic physics during the first half of the twentieth century; recent discoveries in evolutionary biology and biomolecular analysis have revolutionized our interpretation of the Quaternary fossil record; engineering breakthroughs just after the Second World War enabled sediment cores to be recovered from the deepest parts of the oceans and polar ice caps, the analysis of which revealed for the first time the highly sensitive nature of the Earth's climate; and space exploration has enabled processes at the Earth's surface to be measured and monitored independently.

Applications of these various approaches to the interpretation of Quaternary geological records over the course of the past 80 years or so has significantly changed the way in which we now interpret recent Earth history (Porter, 1999). The analytical tools available for studying the records grow ever more sophisticated and diverse, and past events can now be reconstructed at a much higher temporal resolution than was possible only a few decades ago. What emerges from these analyses is a highly resolved record of global climate change during glacial and interglacial episodes, with shifts in temperature of 10 °C or more, sometimes occurring on millennial or, in some cases, centennial timescales.

Over the past 2.6 My, the Earth acquired the combination of physical characteristics (e.g., shape and disposition of continents and mountain belts, ocean circulation systems, atmospheric gas content, and major vegetation biomes) that we can observe today, and which differ markedly from those of earlier eras. The Quaternary geological record of this time interval has revealed just how frequently, and occasionally spasmodically, the global climate system can change. It provides the basis for testing outputs of Global Circulation Models (GCMs), numerical models that simulate the workings of the present global climate system. It was also during the last 2.6 My that anatomically modern humans evolved, diffused globally, and became increasingly influential in modifying the Earth's surface and atmosphere.

Quaternary scientists integrate information obtained from the natural science disciplines, such as climatology, ecology,

geology, and oceanography, as well as from subjects usually associated with the humanities, such as archaeology and anthropology. This reflects the belief, long held by Quaternary scientists that a multidisciplinary approach is essential for understanding global environmental changes, how their effects might be controlled or mitigated, and how humans might learn to cope with environmental stress – knowledge which could well have a bearing on long-term human survival. This brief review considers the primary methodological developments that underpin current thinking about ways in which climatic fluctuations regulate Earth surface processes and conditions. We begin with the publication of the Astronomical theory, which represented the first systematic attempt to explain the rhythmic nature of Quaternary climate change. Coincidentally, the formulation of the Astronomical theory broadly coincided with the establishment of INQUA, the International Union for Quaternary Research (Porter, 1999). INQUA held its first international congress in 1928 and since then its diverse, multidisciplinary membership has been consistently at the cutting edge of research into global climate change.

Astronomical Rhythm of Climate Change

It had long been suspected that the Earth has experienced periodic changes in climate. The first tentative suggestions about these matters were made in the later years of the nineteenth and early years of the twentieth centuries, although in those times the geological evidence available for testing the emerging ideas was limited and equivocal. Since then, geological archives have become more robust and diverse. Rhythmical change is clearly reflected, for example, in the oscillatory nature of the stable oxygen isotope signal in deep-marine sediments and polar ice cores, in the regular occurrence of well-developed soil units (paleosols) within the thick accumulations of wind-blown dust (loess) in central China, and in an increasing number of lake-sediment records, which suggest that the margins of forests in Europe and in high mountain regions, such as the Andes, regularly migrated backwards and forwards across considerable tracts of land. These are manifestations of a global response to some rhythmical forcing agent, now widely believed to be variations in the Earth's orbit and axis. This is the 'Astronomical' or 'Milankovitch theory', named after the Serbian geophysicist, Milutin Milankovitch who, in a series of papers between 1920 and 1941 (e.g., Milankovitch, 1930), proposed that cyclical variations in the amount of solar radiation received at the surface of the Earth could induce major changes in global climates. The principal components of the theory are the precession of the equinoxes (apparent movement of the seasons around the sun) with a periodicity of ~19–23 ka, the obliquity of the ecliptic (variations in the tilt of

the Earth's axis) with a periodicity of ~ 41 ka, and the eccentricity of the orbit (changes in the shape of the Earth's orbit) with a periodicity of ~ 100 ka.

Developing a convincing means of testing this theory has been one of the notable achievements of Earth science in the twentieth century. This was achieved initially through the detailed analysis of stable oxygen isotope variations in marine microorganisms (Foraminifera) obtained from deep-ocean sediments (Figure 1). The ratio of the lighter oxygen isotope (^{16}O) to the heavier isotope (^{18}O) is mainly controlled by the volume of global ice cover and hence, is an index of global climatic change. Deviations towards heavier ratios (downward swings in the isotope curve) reflect times of increased land-ice cover ('glaciations'), and upward shifts represent warmer periods ('interglaciations') when it was significantly reduced.

This evidence offers a deep insight into how the Earth responds to the external stimulus of insolation (solar radiation), how the incoming radiation signal may influence processes operating at the surface of the Earth, and how the behavior of the Earth's climate at the 105-year scale may be predictable over the long term. Some of the more important features of the Earth's recent climate include the following:

1. As Figure 1 implies, cyclic change in global climate is the hallmark of the Quaternary. There is no evidence in the records of extended periods of climatic stability; instead, the Earth's climate system appears to have been constantly in transition throughout the past 2.6 My.
2. The frequency of the climate cycle changed during the Quaternary. Milankovitch's 'precession cycle' appears to have been predominant during the late Pliocene, between ~ 2.8 and 2.6 Ma; the 'obliquity signal' dominated in the period 2.6–0.8 ka, and the 'eccentricity signal' is characteristic of the last 800 ka (Raymo and Ruddiman, 1992).
3. Whereas the calculated values for variations in insolation vary symmetrically over time (Berger et al., 1999),

measured oscillations in oceanic stable isotope ratios are nonsymmetrical (Figure 1). Rates of warming at the start of interglacial stages are usually more rapid than the more gradual cooling trends leading to glacial stages. The implication is that climate response to astronomical variations is not linear, and that the radiation signal is modulated in some way by processes operating at the Earth's surface.

4. Figure 1 also shows how the amplitude of climate oscillation increased during the Quaternary, particularly within the last 800 ka or so, when the eccentricity cycle was predominant. Interglaciations became generally warmer during the later part of the Quaternary, while glaciations appear to have been characterized by colder extremes than during earlier times.
5. During later Quaternary time, interglaciations have been relatively short. Periods during which conditions were as warm as the present appear to last, on average, about 15–17 ka, before a decline begins that is the precursor to a new glaciation.

Global climatic signals at orbital frequencies have been detected and confirmed in other contexts, most notably in Antarctic and Greenland ice cores (EPICA Community Members, 2004; Masson-Delmotte et al., 2005), and in Chinese loess–paleosol sequences (Sun et al., 2006). It seems evident that astronomical cycles will continue to exert a strong influence on the Earth's climate and that the present interglacial period will, in due course, be succeeded by the next glaciation. Indeed, some believe that the next glaciation may have already begun (e.g., Ruddiman, 2003; Ruddiman, 2005).

Short-Term ('Sub-Milankovitch') Climatic Variations

Climatic perturbations that occur over much shorter frequencies cannot be explained by the astronomical cycles alone.

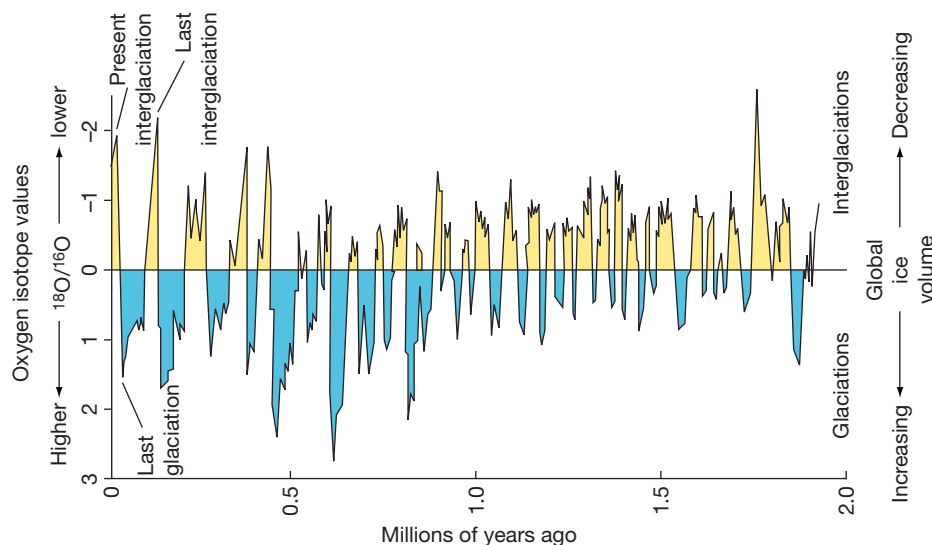


Figure 1 Stable oxygen isotope variations in marine cores spanning the last 2 My. The values are shown as deviations from a long-term mean. Variations in the marine isotope ratio reflect variations in the volume of global land ice, and hence the curve is primarily a 'paleoglaciational' curve, and only indirectly a paleoclimatic curve. Interglacial stages are shown in yellow, and glacial stages in blue. (See Lowe, 2001; Lowe and Walker, 1997 for further details.)

Short-term perturbations in the global climate system are most clearly reflected in cores obtained from the polar ice sheets (Alley, 2000). During the 1960s, several deep boreholes were drilled in Greenland and Antarctica, while a number of smaller ice masses in other parts of the world were also investigated. These pioneering studies showed that important paleoenvironmental information could be obtained from the analysis of stable isotope ratios, gas content, chemicals, and other impurities in ice that had accumulated over many thousands of years. Subsequent work in Greenland in the 1990s, most notably the GRIP and GISP-2 projects (Alley, 2000), and more recently, the NGRIP programme (2003), built on the previous research findings and provided penetrating new insights into recent global environmental change over the last glacial–interglacial cycle (NorthGRIP Members, 2004). In Antarctica, lower rates of snow accumulation mean that the ice-core sequences are generally stratigraphically less well-resolved than in Greenland, but there the record extends over a much longer time. The EPICA core, which is still being drilled, now extends back ~800 ka and may eventually reach 1 My (EPICA Community Members, 2004). The polar ice cores also contain a history of long-term stable isotope variations that generally match those obtained from marine cores that span the same time intervals. This indicates that the oceans and ice sheets are closely linked ('coupled') in both physical state and behavior. Mathematical models explain the exchange of stable isotopes between the oceans, the atmosphere, and the cryosphere, which are the dominant global reservoirs constituting the hydrosphere (Jouzel et al., 1987). The stable isotope and other proxy records reveal patterns of climate change at both Milankovitch and sub-Milankovitch (short-term) frequencies. Millennial and submillennial climate variations are clearly recorded in the Greenland ice-core records. The central sector of the Greenland ice sheet

is ~1500 m thick and provides a continuous environmental sequence extending over more than 100 ka. The precise period represented is uncertain due to deformation and thinning of the basal ice layers, but conservative estimates place the minimum age of the basal ice at ~123 ka (NorthGRIP Members, 2004). Distinctive annual layers occur in the upper levels of the core sequence, which means that past environmental changes can be resolved at a very high resolution, far higher than is normally achievable with marine- or lake-sediment records. An irregular series of peaks and troughs in the isotope signal (referred to as Dansgaard–Oeschger oscillations, after two leading ice-core scientists, Willi Dansgaard and Hans Oeschger) is replicated almost exactly between the GISP-2 and GRIP cores, and is equally well matched in the more northerly NGRIP core (Figure 2). The remarkable degree of conformity between these three records suggests that the ice sheet as a whole responded cyclically to hemispherical climatic forcing, the shifts in the stable isotope record reflecting a combination of changes in ambient temperatures, moisture source and delivery, and wind strength and direction. Notable features of this high-resolution paleoclimatic record include the following:

1. Modern isotope–temperature relationships suggest that the amplitude of isotopic variations represented in the Greenland ice-core records reflects shifts of up to 15 °C in ambient temperature during the last cold stage.
2. During the last cold stage (115–11.5 ka), as many as 27 distinct warming events (interstadials) occurred, some with maximum temperatures that approached, but did not quite reach, modern temperature values over central Greenland.
3. The climatic oscillations throughout the last cold stage, as reflected in the stable isotope records, typically range

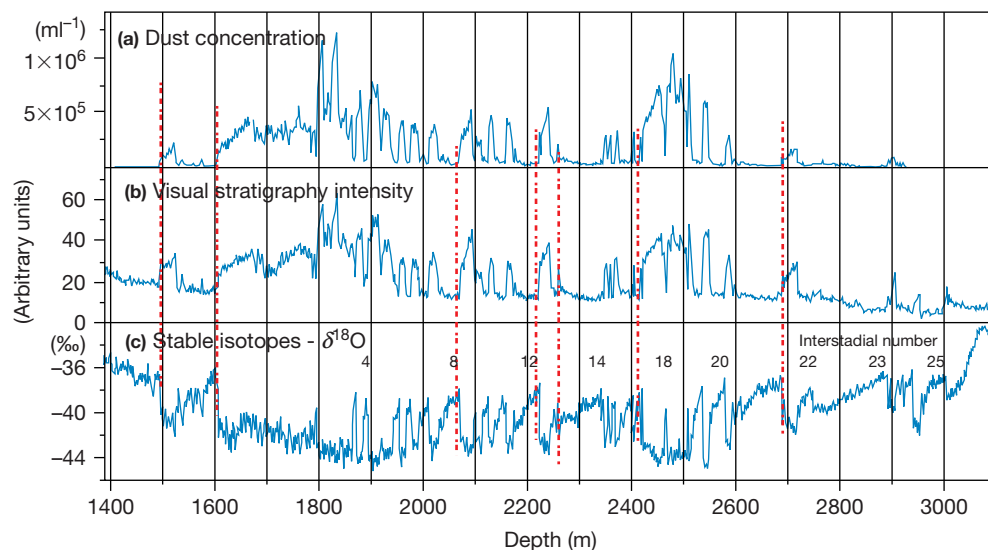


Figure 2 Stratigraphic variations in the NorthGRIP core spanning the period ~9 ka (1400 m depth) to 120 ka (3200 m depth). The lowest curve shows variations in stable oxygen isotope ratios, the key proxy that reflects temperature variations over central Greenland. The middle panel shows variations in visible stratigraphy (ice color and transmissivity), while the top panel shows variations in dust concentration in the ice. The red dotted lines emphasize the antiphase relationships between oxygen isotope variations and dust concentration. Numbers in the lower panel indicate sequence order of abrupt warming events (interstadials). Reproduced from Svensson A, Nielsen SW, and Kipfstuhl S, et al. (2005) Visual stratigraphy of the North Greenland Ice Core Project (NorthGRIP) ice core during the last glacial period. *Journal of Geophysical Research* 110(D02): 108.

between 500 and 2000 years in duration, and the isotopic profiles show characteristic asymmetric shapes reflecting abrupt warmings that are followed by more gradual cooling trends.

4. In the upper layers of the ice cores, the duration of the transitions from cold stadial intervals to maximum warmth can be estimated by ice-layer counting; remarkably, the evidence indicates that some of these marked climatic shifts have occurred within a few decades (i.e., within a modern human lifetime), while the most recent major warming, marking the start of the current interglacial period at ~ 11.7 ka, may have occurred over an even shorter interval.

Collectively, the ice-core data suggest a considerable degree of instability in Greenland climate during the last cold stage, with oscillations between stadial and interstadial modes, a process that Taylor et al. (1993) likened to the operation of a 'flickering switch.' Prior to the publication of these records, climatic shifts of this magnitude and over such restricted timescales were not considered feasible. Two questions arose: (1) Were these climatic oscillations confined to Greenland? (2) What mechanism might cause the signal to flicker in this way? Subsequent research has shown that the ice-core climatic record for the last cold stage is not unique to Greenland, for similar climate histories have been discovered in other areas of the world. The answer to question 2 lies in the deep oceans where changes in ocean circulation appear to be the principal forcing factor in the operation of the climatic switch.

The Role of the Oceans

Shortly before the publication of the GISP-2 and GRIP ice-core results, Heinrich (1988) drew attention to a series of coarse sediment layers (now termed 'Heinrich Layers') in cores obtained from the floor of the North Atlantic. These are believed to be the products of glacier-calving around the margins of the northern ice sheets, which resulted in massive 'armadas' of icebergs in the North Atlantic circulating in a counterclockwise flow (gyre) and releasing sediment (ice-rafted debris) as they melted (Hemming, 2004). Eight such 'Heinrich events' occurred between 70 and 10 ka, indicating periodic and large-scale instability of the ice-sheet margins during times of maximal expansion. A close relationship was detected between episodes of coarse sediment influx into the North Atlantic, abundance of cold-adapted marine microfossils, and isotope variations in the GRIP record (Bond and Lotti, 1995; Figure 3). Hughen et al. (1996) presented compelling evidence for an even closer match between sediment variations in the tropical Atlantic and the GISP-2 record (Figure 4). Since then, geological archives from places as widely scattered as China, the Arabian Sea, and the sub-Antarctic have yielded proxy climate data that closely resemble oceanographic and climatic changes reflected in the North Atlantic records (e.g., Hemming, 2004; Porter and Zhou, 2006; Sachs and Anderson, 2005).

Several important points that have a bearing on our interpretation of the glacial record and of the global climate sequence arise from this work:

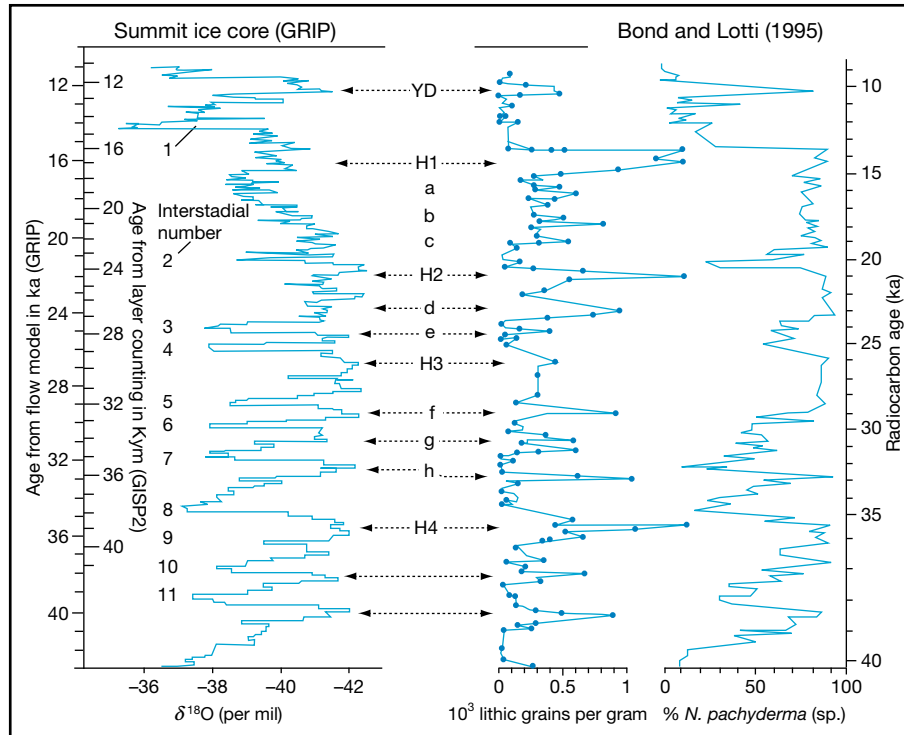


Figure 3 Variations in concentration of coarse mineral grains (lithic particles) and in abundance of the cold-adapted plankton species, *N. pachyderma* in sediment cores recovered from the floor of the North Atlantic, compared with the stable oxygen isotope record from the GRIP ice-core record. Reproduced from Bond GC and Lotti R (1995) Iceberg discharges into the North Atlantic on millennial timescales during the last glaciation. *Science* 267: 1005–1010.

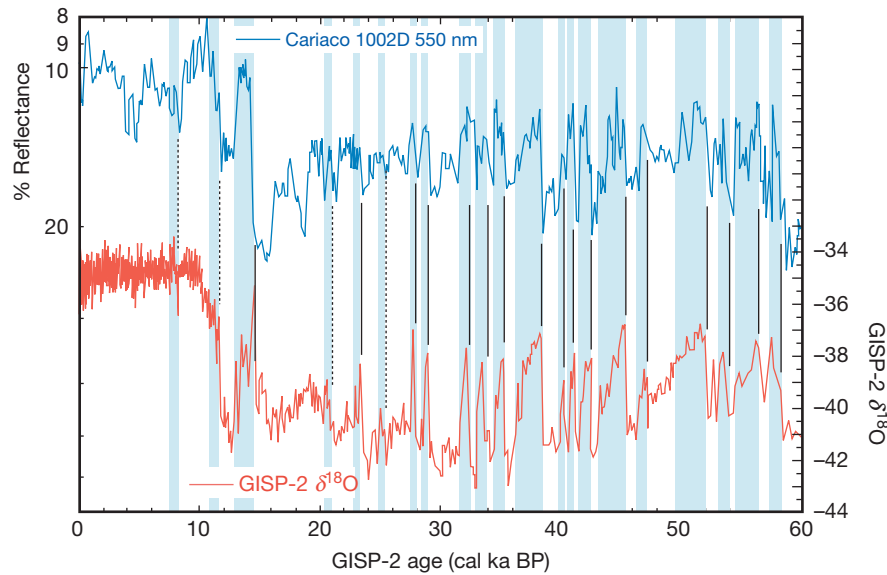


Figure 4 Comparison between variations in sediment optical properties (light reflectance) of a laminated sequence of marine sediments in the Cariaco Basin, north of Venezuela, and the GISP-2 stable oxygen isotope record for the past 90 ka. Gray-shaded vertical bars represent Greenland interstadial events. Reproduced from Hughen KA, Overpeck JT, Peterson LC, and Trumbore S (1996) Rapid climate changes in the tropical Atlantic region during the last deglaciation. *Nature* 380: 51–54.

1. It is now clear that the history of glaciation is far more complex than was previously realized. Traditional models of glacier/ice-sheet advance and retreat have been based on painstaking investigation of continental records, but the evidence on land is partial simply because more recent glacier advances tend to bury or destroy the evidence of earlier glacier activity. The glacial-geological evidence from the last glaciation, for example, does not provide evidence of all the numerous stadial-interstadial oscillations that are reflected in the ice-core records.
2. The close correspondence between the isotope signal from Greenland and that from the sedimentary record in the North Atlantic suggests that the ice sheet and oceans were responding to the same external stimulus during the course of the last cold stage.
3. The often-abrupt climatic oscillations of the last cold stage were not confined to Greenland and the North Atlantic, but appear to have occurred in other parts of the world as well.
4. It is frequently assumed, although it has not yet been tested satisfactorily, that these abrupt climatic changes occurred more-or-less simultaneously on a global scale, and that major shifts in climate somehow operate contemporaneously.

A number of explanations have been put forward to account for the periodic instability of the ice sheets. These include self-regulating processes that came into play as the ice sheets became large enough to influence their immediate environment, by increased melting at the base, by extending into deeper water where marginal calving would be accelerated, or by some complex interplay between these factors and concomitant changes in global sea level. It has also been suggested that the timing of the onset of the Greenland interstadial events during the last cold stage approximates a 1.47 ky climate cycle

and hence ice-sheet instability could, in part, be a response to a more regular forcing mechanism such as solar variations (e.g., Schulz (2002)). Others, however, favor a different explanation, namely changes in the rate or pattern of ocean circulation.

One of the most persuasive advocates of the latter hypothesis is Broecker (1994) and Broecker (1997), who cites in its support the considerable body of paleoceanographic evidence which shows that the oceans, and the North Atlantic in particular, experienced marked changes in salinity, density, and rate of water movement during the last cold stage. These must, in Broecker's view, have had a major impact on the speed and pattern of flow of the thermohaline circulation, the network of surface and deep-water currents that circulates ocean water around the globe (Rahmstorf, 2000). The present arterial flows, such as the Gulf Stream, are maintained by thermal and salinity stratification, which leads to warm salty surface water being transported northward across the North Atlantic where it cools, sinks, and forms a return southward-flowing deep-water current. This circulatory system, or 'conveyor,' keeps NW Europe warmer than other regions in the northern hemisphere at comparable latitudes, and also delivers moisture to the Greenland ice sheet and to the smaller ice bodies in Iceland, Spitsbergen, and Scandinavia. According to this hypothesis, rapid melting of ice sheets would have delivered enormous volumes of cold freshwater onto the ocean surface, and since this would be lighter (less dense), it would form a cap over the ocean, preventing sinking, and thereby reducing the pressure gradient that drives the conveyor. This, in turn would have led to a marked cooling in the North Atlantic and NW Europe, and a reduction in moisture supply to the Greenland ice sheet.

This explanatory mechanism has been widely favored, partly because it is based on the physical parameters that govern ocean-water movement, and partly because numerical

modeling of the North Atlantic shows how sensitive the three-dimensional circulation pattern is to changes in surface water fluxes (Rahmstorf, 1995; Stocker and Wright, 1991). In addition, quantitative estimates of the volumes of water released to the North Atlantic during ice-sheet wastage at the end of the last glaciation appear to provide adequate fluxes for major disruptions of the ocean circulation system (Clark et al., 2003; Teller et al., 2002). It also offers a plausible explanation for why ventilation of the deep ocean (exchange of waters, nutrients, and oxygen with surface waters) was periodically interrupted throughout the last cold stage (Bard, 2002). Finally, it might explain why sudden changes in climate in the North Atlantic were also experienced in other parts of the world.

What cannot yet be adequately explained by these inferred oceanographic changes, however, is the trigger mechanism for ice-sheet instability. In seeking a resolution to this problem, it may prove to be a mistake to focus solely on the ice sheets and the oceans, for there is now a growing body of evidence for the operation of other factors in controlling abrupt climate change.

Global Feedback Mechanisms

The ice-core records constitute an unparalleled archive for showing how the different components of the global environmental system interconnect. The gas bubbles locked in the ice, for example, provide an independent record of variations in atmospheric gas content, which enables a direct comparison to be made with the stable isotope records that reflect ambient temperature. The evidence reveals a very close correlation between atmospheric levels of carbon dioxide (CO_2), deuterium, and global temperature (Figure 5), a relationship that has stimulated wide public debate about the continued rise in greenhouse gases in the atmosphere, and the implications for future global climate (IPCC, 2001).

There is also a close relationship between isotopic variations and the dust content of polar ice, at both glacial-interglacial (Delmas, 1992) and stadial-interstadial scales (Svensson et al., 2005). Figure 2 shows how peaks in the record of dust flux to the Greenland ice sheet during the last cold stage coincided with stadial intervals (lighter isotope ratios), while abrupt warmings at the start of the interstades coincided with marked reductions in dust content. Clearly, the atmosphere was dustier during cold periods. This may partly reflect increased transport of wind-borne dust under a more vigorous atmospheric circulation regime from arid and semiarid areas such as western China. It may also be partly explained by a reduction in stabilizing vegetation cover in such regions, and partly by wind erosion of exposed seafloor during times of reduced sea level.

Many other processes and environmental fluxes contributed to, and were affected by, abrupt climatic change, but the cases of CO_2 and dust serve to illustrate how difficult it is to determine with present knowledge which agents and processes were the key initiators of change. The problems are threefold: (1) The majority of geological records cannot yet be dated with a precision sufficient to determine the order in which different components of the Earth system change. (2) The linkages frequently are mutually dependent. For example, a change in climate might lead to a change in the global carbon cycle, but the latter could equally result from the operation of nonclimatic factors, for example, from an increase in ocean biomass due to a change in nutrient supply. One consequence of this might be an increase in the rate of extraction of CO_2 from the atmosphere, which then becomes locked in the oceans (a process referred to as 'draw-down' of CO_2) and which might, in turn, impact on climate. (3) The same changes may sometimes be induced by the operation of opposing or quite different processes. The amount of dust entrained in the atmosphere could, for example, increase with a shift to colder

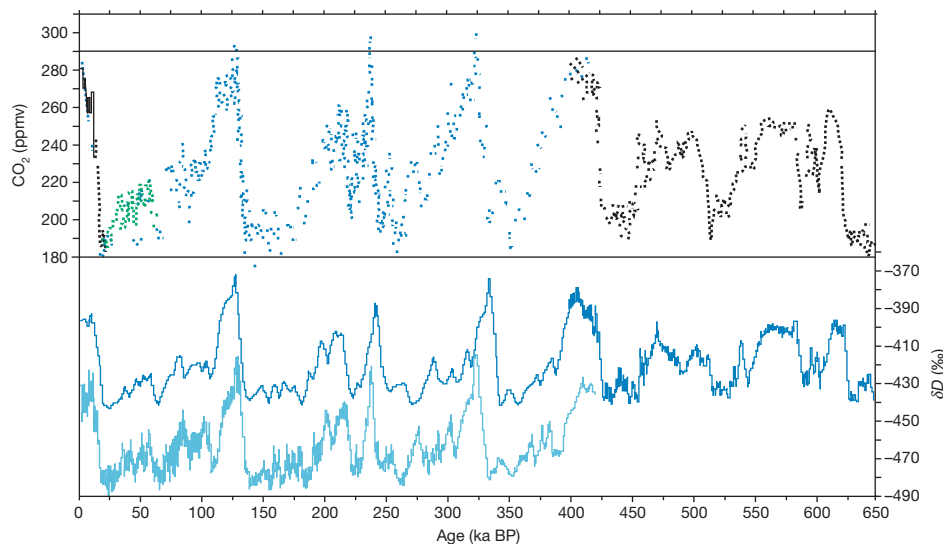


Figure 5 Composite records of variations in atmospheric CO_2 concentrations (upper panel) and deuterium ratios (a paleotemperature index – lower panel) in gas bubbles in Antarctic ice cores spanning the last 650 ka. The upper panel shows combined data from the Dome C (black dots), Vostok (blue) and Taylor Dome (green) ice cores, while the lower panel shows deuterium records for Dome C (upper, dark blue curve) and Vostok (grey-blue). Reproduced from Siegenthaler U, Stocker TF, and Monnin E, et al. (2005) Stable carbon cycle–climate relationship during the late Pleistocene. *Science* 310: 1313–1317.

conditions through a lowering of sea level and exposure of silty areas to wind erosion, or by an increase in the transport of dust to polar regions from arid and semiarid regions. A similar effect could result from a warming trend that led to a reduction in effective precipitation in some regions and the expansion of arid and semiarid zones.

Understanding the complex chain of feedback mechanisms that constitutes the global environmental system is one of the most challenging areas of Quaternary science, but given the problems of interpretation referred to above, such an understanding will be difficult to achieve. However detailed and diverse Quaternary paleoenvironmental records may be, they do not, in themselves, constitute tests of causation between multiple competing variables. What is needed is some basis for experimentation, where key variables can be controlled while others are deliberately varied (and, perhaps, exaggerated) in order to isolate those that are the most influential. This takes us into another area where considerable progress has been made in Quaternary science in recent years: numerical modeling.

Numerical Environmental Modeling

Over the past several decades, increasingly sophisticated numerical models have been developed that can simulate a range of natural processes operating at the surface of the Earth (Schellnhuber, 1999). They are based on mathematical descriptions of the behavior of, and interactions between, specified variables, each component or interaction being mathematically prescribed to obey the laws of physics. Some models are simple, being constrained to focus on a particular process or on the interaction between selected variables within a limited geographical area, as, for example, a hydrological model of water-balance changes in a river catchment (Kutzbach, 1980). More complex models, however, are designed to represent the operation of Earth systems at a global scale. The best known and most sophisticated are general circulation models (GCMs), which simulate the operation of the global climate system (McGuffie and Henderson-Sellers, 2001). The capabilities of numerical models have increased dramatically within the last few decades following major advances in computer power and speed, software design, and the tools used to measure and monitor the natural fluxes between global reservoirs. Increased national and international cooperation in model design and intercomparison has also been a significant factor.

Quaternary scientists have long appreciated the potential of numerical models for providing analogs of the Earth system under different climate states, and for helping to clarify the key mechanisms that drive environmental change. Important integrated research programs employing numerical modeling have included simulations of the conditions in the ocean at the Last Glacial Maximum (CLIMAP Project Members, 1976; CLIMAP Project Members, 1981), and an analysis of the effects of insolation variations and concomitant changes in surface conditions on global climate over the past 18 ka (COHMAP Members, 1988; Kutzbach and Guetter, 1986). These were important milestones in a developing collaboration between Earth scientists who generate the paleodata and the numerical modeling community. The relationship has been reciprocal with, on the one hand, Quaternary scientists learning from

the modelers about the types of input data that are needed to generate the best model results, while on the other, the modelers require snapshots of past environmental conditions (based on paleoenvironmental reconstructions) from the Quaternary science community to test model outputs (van Andel, 2002). The latter are especially important because models are simplifications of the natural world with built-in assumptions. Hence, model outputs invariably have large uncertainties. Nowhere is this more apparent (or relevant) than in the current debate centering around the wide divergence in predictions of the nature and magnitude of future climate change generated by different GCMs (e.g., Allen et al., 2000).

One possible means of testing whether a GCM captures realistic climate situations, therefore, is to use the model to 'predict' an outcome for which an independent estimate already exists: this can be done by generating model evaluations of climate states for times in the recent geological past. In this procedure, a time period is selected for which the global boundary conditions are reasonably well known, and for which a detailed paleorecord can be compiled. The paleodata provide the basis for a reconstruction of the climatic conditions that may have prevailed at that time, and the model can be run to generate a stable climate state when forced by the prescribed (past) boundary conditions (e.g., Wright et al., 1993). The two outcomes can then be compared. One problem, of course, is that any significant differences between the results would be difficult to reconcile, for there is no independent means of determining the extent to which either, or both, are in error. Nevertheless, model data comparison is widely practised (e.g., Paleoclimate Modeling Intercomparison Project (PMIP) – <http://www-lsce.cea.fr/pmip/>), for there is a widespread view that improvements in both model design and in the amount and quality of the paleodata being generated will ultimately lead to greater convergence between the two outputs.

Numerical modeling plays an increasingly important role in contemporary Quaternary science because it permits speculative questions to be posed with some basis for constraining the answers provided. The balance between variables and inputs can be deliberately altered within a model in order to evaluate the nature, magnitude, and timing of environmental response. This approach has a range of applications. One example is the recent use of a GCM, forced by the insolation cycle alone, to examine future climate trends (Hargreaves and Annan, 2002). This particular model suggests that the Earth will experience a significant cooling step, followed by a slower cooling trend that will culminate in the next glacial maximum in about 60 ka. It also predicts that any 'global warming' effect caused by anthropogenic pollution will arrest this trend for about 30 years, but is unlikely to be large enough to prevent long-term cooling thereafter. A second example predicts the mass balance changes of the polar ice sheets and the subsequent effects on global sea level (Vaughan and Spouge, 2002), a topic of considerable concern given that a large percentage of the world's population lives within a few meters of present sea level. A third is the use of a three-dimensional ice-sheet model to examine the hypothesis that periodic large-scale oscillations of the North American Laurentide ice sheet may have been responsible for Heinrich events (Papa et al., 2006).

Although numerical modeling has provided important new insights into the workings of the global climate system, even

the most sophisticated of models have limitations. Some of the main areas of concern are the following:

1. Models invariably (and inevitably) oversimplify environmental complexity, for they cannot represent every component and synergistic interaction that occurs in nature. New discoveries continue to show how small-scale processes, such as feedback mechanisms involving cloud formation, can have far bigger impacts on climate (and also model output) than might have been anticipated (Williams et al., 2006).
2. Because numerical models can only handle a finite number of calculations, all the global processes and interactions represented in the models must be scaled down in order to be manageable. This inevitably leads to a 'scaling-up' problem when it comes to interpreting the output: this is the reverse logic problem of knowing whether the scaled-down, almost artificial microcosm, created in the model truly reflects the real operation of the system at the global scale.
3. Interactions between global subsystems (the atmosphere, hydrosphere, cryosphere, biosphere, geosphere) are invariably nonlinear. Hence, whereas it may be possible to predict responses within a simplified subsystem when only a few variables are involved, the greater the number of variables, the greater the complexity in terms of links between different subsystems. This almost inevitably means that the outcomes are less predictable (Clark et al., 1999; Rind, 1999).
4. Only very small changes in one or more parameters can have significant effects on model output. Running the same model twice using exactly the same parameters and boundary conditions can generate quite different outputs. Therefore, the outcomes may be unpredictable (Sivakumar, 2004), which may help explain why GCMs often tend to produce quite different predictions of future climatic scenarios.

Paradoxically, however, if an element of chaotic behavior is involved in all natural processes, then the only viable way forward in the search for an understanding of the global climate system, and predicting its behavior, is through numerical modeling. If this challenge is to be met, future generations of numerical models must become even more sophisticated. They will have to incorporate a much wider range of variables and linkages to reflect more realistically the complexity of the Earth system.

Climate and Humans

Concern over the modern climate-warming trend and the extent to which it may have been generated by human activity has stimulated widespread debate. The degree of concern over this issue is reflected in the tenor of the policy statements drawn up by almost every national government, as well as the increasing acceptance of the need for agreed protocols to curb greenhouse gas emissions. By far the most compelling evidence has been provided by Antarctic ice cores, which contain a detailed record of atmospheric gas ratios stretching back over the past 650 ka. Figure 6 shows the record of variations in concentration of atmospheric CO₂ and CH₄ over the past

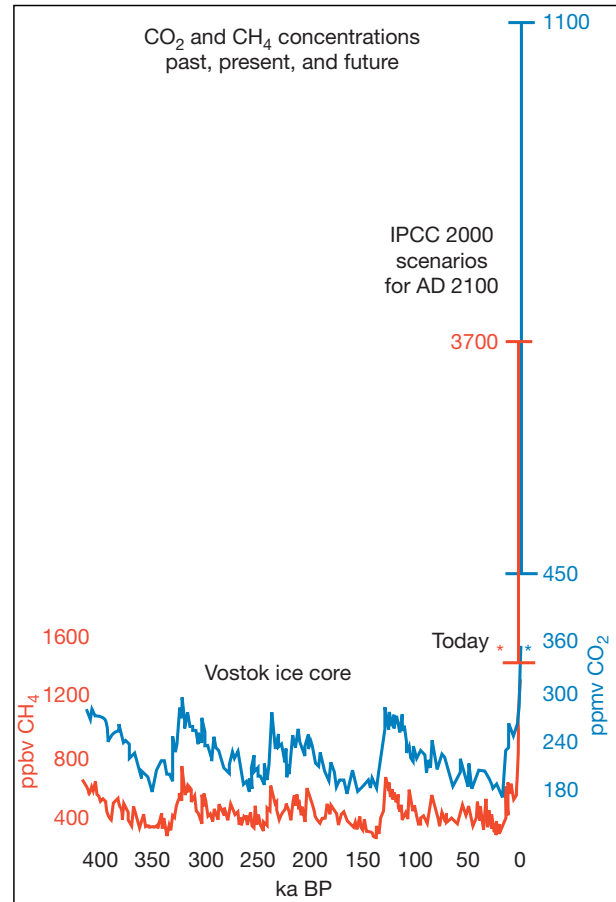


Figure 6 Record of CO₂ (blue curve) and CH₄ (red) variations in the Vostok (Antarctic) ice core over the past 400 ka and projected future levels based on IPCC estimates. PAGES web resources: <http://www.pages.unibe.ch/about/index.html>.

400 ka obtained from the Vostok ice core (Petit et al., 1999). These data, together with the longer records from a second Antarctic borehole, Dome Concordia (Figure 5), indicate that natural atmospheric levels have normally been well below 300 ppm for CO₂ and 800 ppb for CH₄. The concentration of these greenhouse gases in the modern atmosphere is significantly above these levels (double in the case of CH₄), with the evidence pointing to a steady rise in concentrations from the time of the Industrial Revolution. However, it is the abrupt increase over the course of the second half of the twentieth century that gives the greatest cause for concern. Although the causal connections between atmospheric gas concentration and global temperature change continues to be a matter of dispute (e.g., Mann et al., 1998; McIntyre and McKittrick, 2003), there is a widespread acceptance that the late twentieth century warming detected in a range of proxy and instrumental records is anomalous (Osborn and Briffa, 2006), and that this is inextricably linked to increased atmospheric concentrations of greenhouse gases.

This view is reinforced by predictions, based on GCMs, of the likely atmospheric concentrations of these gases in ad 2100, should current fossil fuel consumption be allowed to continue unchecked. These values (Figure 6) are far in excess of

any recorded in the recent geological record. Should these predictions prove to be anywhere near reliable, then it is difficult to imagine that global climate will remain unaffected. Whether this will prove to be as detrimental to humans as is frequently predicted is difficult to assess, for there are no analogs in the Quaternary geological record for a world with climatic boundary conditions comparable to those of the present time having such high concentrations of atmospheric greenhouse gases.

The history of human impact on climate may, however, predate the Industrial Revolution, for it has recently been suggested that anthropogenically enhanced emissions of greenhouse gases (notably CO₂ and CH₄) could have begun as early as 8 ka as a result of the spread of deliberate burning of vegetation and the adoption of animal husbandry, cultivation, and other farming practices (Ruddiman, 2003; Ruddiman, 2005). Ruddiman's analysis of the natural rhythm of climate variations during the Quaternary leads him to conclude that if it were not for the present abnormally high levels of atmospheric CO₂ and CH₄, global climate would be ca. 2 °C cooler than it is today, and the world would be about a third of the way towards full glacial conditions (Ruddiman et al., 2005). Whether or not these conclusions are valid, they are examples of the types of question that are now being posed about the Earth's past and future climatic history, and it is in the rich archive of the Quaternary geological record where answers to such questions perhaps can be found best.

See also: **Glacial Climates:** Thermohaline Circulation. **Glacial Landforms, Ice Sheets:** Growth and Decay. **Glacial Landforms, Sediments:** Glaciomarine Sediments and Ice-Rafted Debris. **Glaciation, Causes:** Astronomical Theory of Paleoclimates. **Ice Core Methods:** Chronologies; CO₂ Studies; Methane Studies. **Ice Core Records:** Antarctic Stable Isotopes; Greenland Stable Isotopes. **Ice Cores:** History of Research, Greenland and Antarctica. **Lake Level Studies:** Latin America; West-Central Europe. **Loess Records:** China. **Paleoceanography, Physical and Chemical Proxies:** Oxygen-Isotope Stratigraphy of the Oceans. **Paleoclimate:** Introduction. **Paleoclimate Modeling:** Data–Model Comparisons; Quaternary Environments. **Paleoclimate Reconstruction:** Sub-Milankovitch (DO/Heinrich) Events.

References

- Allen MR, Stott PA, Mitchell JFB, Schnur R, and Delworth TL (2000) Quantifying the uncertainty in forecasts of anthropogenic climate change. *Nature* 407: 617–620.
- Alley RB (2000) *The Two-Mile Time Machine: Ice Cores, Abrupt Climate Change and Our Future*. Princeton, NJ: Princeton University Press.
- Bard E (2002) Climate shock: Abrupt changes over millennial time scales. *Physics Today* 55: 32–38.
- Berger A, Li XS, and Loutre MF (1999) Modelling northern hemisphere ice volume over the last 3 Ma. *Quaternary Science Reviews* 18: 1–11.
- Bond GC and Lotti R (1995) Iceberg discharges into the North Atlantic on millennial timescales during the last glaciation. *Science* 267: 1005–1010.
- Broecker WS (1994) An unstable superconveyor. *Nature* 367: 414–415.
- Broecker WS (1997) Thermohaline circulation, the Achilles Heel of our climate system: Will man-made CO₂ upset the current balance? *Science* 278: 1582–1588.
- Clark PU, Alley RB, and Pollard D (1999) Northern hemisphere ice sheet influences on global climate change. *Science* 286: 1104–1111.
- Clark G, Leverington D, Teller J, and Dyke A (2003) Superlakes, megafloods, and abrupt climatic change. *Science* 301: 922–923.
- CLIMAP Project Members (1976) The surface of ice-age earth. *Science* 191: 1131–1137.
- CLIMAP Project Members (1981) Seasonal reconstructions of the earth's surface at the last glacial maximum. *Geological Society of America Map and Chart Series MC-36* 1–18.
- COHMAP Members (1988) Climatic changes of the last 18,000 years: Observations and model simulations. *Science* 241: 1043–1052.
- Delmas J (1992) Environmental information from ice cores. *Reviews of Geophysics* 30: 1–21.
- EPICA Community Members (2004) Eight glacial cycles from an Antarctic ice core. *Nature* 429: 623–628.
- Hargreaves J and Annan J (2002) Assimilation of paleo-data in a simple earth system model. *Climate Dynamics* 19: 371–381.
- Heinrich H (1988) Origin and consequences of cyclic ice rafting in the northeast Atlantic ocean during the past 130,000 years. *Quaternary Research* 29: 143–152.
- Hemming SR (2004) Heinrich events: Massive Late Pleistocene detritus layers of the North Atlantic and their global climate imprint. *Review of Geophysics* 42(1): RG1005.
- Hughen KA, Overpeck JT, Peterson LC, and Trumbore S (1996) Rapid climate changes in the tropical Atlantic region during the last deglaciation. *Nature* 380: 51–54.
- Intergovernmental Panel on Climate Change (2001) *Climate Change 2001: The Scientific Basis*. Cambridge: Cambridge University Press.
- Jouzel J, Russell GL, Suozzo RJ, Koster RD, White JWC, and Broecker WS (1987) Simulations of the HDO and H₂18O atmospheric cycles using the NASA/GISS general circulation model: The seasonal cycle for present-day conditions. *Journal of Geophysical Research* 92: 14739–14760.
- Kutzbach J (1980) Estimates of past climate at palaeolake Chad, North Africa, based on a hydrological and energy balance model. *Quaternary Research* 14: 210–223.
- Kutzbach J and Guetter PJ (1986) The influence of changing orbital parameters and surface boundary conditions on climate simulations for the past 18,000 years. *Journal of the Atmospheric Sciences* 43: 1726–1759.
- Lowe JJ (2001) Quaternary geochronological frameworks. In: Brothwell DR and Pollard AM (eds.) *Handbook of Archaeological Sciences*, pp. 9–21. Chichester: Wiley.
- Lowe JJ and Walker MJC (1997) *Reconstructing Quaternary Environments*, 2nd edn. Harlow: Addison, Wesley, Longman Ltd.
- Mann ME, Bradley RS, and Hughes MK (1998) Global-scale temperature patterns and climate forcing over the past six centuries. *Nature* 392: 779–787.
- Masson-Delmotte V, Jouzel J, Landais A, et al. (2005) GRIP deuterium excess reveals rapid and orbital-scale changes in Greenland moisture origin. *Science* 309: 118–121.
- McGuffie K and Henderson-Sellers A (2001) Forty years of numerical climate modelling. *International Journal of Climatology* 21: 1067–1109.
- McIntyre S and McKittrick R (2003) Corrections to the Mann et al. (1998) proxy data base and northern hemispheric average temperature series. *Energy and Environment* 14: 751–771.
- Milankovitch M (1930) Mathematische klimalehre und astronomie theorie der klimaschwankungen. In: Köppen W and Geiger R (eds.) *Handbuch der Klimatologie*, vol. I(A), pp. 1–176. Berlin: Gebrüder Borntraeger.
- North Greenland Ice-Core Project (NorthGRIP) Members (2004) High resolution climate record of the northern hemisphere reaching into the last glacial interglacial period. *Nature* 431: 147–151.
- Osborn TJ and Briffa KR (2006) The spatial extent of 20th-century warmth in the context of the past 1200 years. *Science* 311: 841–845.
- Papa BD, Mysak LA, and Wang Z (2006) Intermittent ice sheet discharge events in northeastern North America during the last glacial period. *Climate Dynamics* 26: 201–216.
- Petit JR, Jouzel J, Raynaud D, et al. (1999) Climate and atmospheric history of the past 420,000 years from the Vostok ice core, Antarctica. *Nature* 399: 429–436.
- Porter SC (1999) INQUA and quaternary science at the millennium: A personal retrospective. *Quaternary International* 62: 111–117.
- Porter SC and Zhou WJ (2006) Synchronism of Holocene East Asian monsoon variations and North Atlantic drift-ice tracers. *Quaternary Research* 65: 443–449.
- Rahmstorf S (1995) Bifurcation of the Atlantic thermohaline circulation in response to changes in the hydrological cycle. *Nature* 378: 145–149.
- Rahmstorf S (2000) The thermohaline ocean circulation: A system with dangerous thresholds? *Climatic Change* 46: 247–256.
- Raymo ME and Ruddiman WF (1992) Tectonic forcing of late Cenozoic climate. *Nature* 359: 117–124.
- Rind D (1999) Complexity and climate. *Science* 284: 105–107.

- Ruddiman WF (2003) The anthropogenic Greenhouse Era began thousands of years ago. *Climatic Change* 61: 261–293.
- Ruddiman WF (2005) *Plows, Plagues and Petroleum: How Humans Took Control of Climate*. Princeton, NJ: Princeton University Press.
- Ruddiman WF, Vavrus SJ, and Kutzbach JE (2005) A test of the overdue-glaciation hypothesis. *Quaternary Science Reviews* 24: 1–10.
- Sachs JP and Anderson RF (2005) Increased productivity in the sub-Antarctic ocean during Heinrich Events. *Nature* 434: 1118–1121.
- Schellnhuber HJ (1999) Earth system analysis and the second Copernican revolution. *Nature* 402: C19–C23.
- Schulz M (2002) On the 1470-year pacing of Dansgaard–Oeschger warm events. *Paleoceanography* 17: 1014–1029.
- Sivakumar B (2004) Chaos theory in geophysics: Past, present and future. *Chaos, Solitons & Fractals* 19: 441–462.
- Stocker TF and Wright DG (1991) Rapid transitions of the ocean's deep circulation induced by changes in surface water fluxes. *Nature* 351: 729–732.
- Sun Y, Clemens SC, An Z, and Yu Z (2006) Astronomical timescale and palaeoclimatic implications of stacked 3.6-Myr monsoon records from the Chinese Loess Plateau. *Quaternary Science Reviews* 25: 33–48.
- Svensson A, Nielsen SW, Kipfstuhl S, et al. (2005) Visual stratigraphy of the North Greenland Ice Core Project (NorthGRIP) ice core during the last glacial period. *Journal of Geophysical Research* 110(D02): 108.
- Taylor KC, Lamorey GW, Doyle GA, et al. (1993) The 'flickering switch' of late Pleistocene climate change. *Nature* 361: 432–436.
- Teller JT, Leverington DW, and Mann JD (2002) Freshwater outbursts to the oceans from glacial Lake Agassiz and their role in climate change during the last deglaciation. *Quaternary Science Reviews* 21: 879–887.
- Van Andel TH (2002) The climate and landscape of the middle part of the Weichselian glaciation in Europe: The stage 3 project. *Quaternary Research* 57: 2–8.
- Vaughan DG and Spouge JR (2002) Risk estimation of collapse of the west Antarctic ice sheet. *Climatic Change* 52: 65–91.
- Williams KD, Ringer MA, Senior CA, et al. (2006) Evaluation of a component of the cloud response to climate change in an intercomparison of climate models. *Climate Dynamics* 26: 145–165.
- Wright HE Jr., Kutzbach JE, Webb T III, Ruddiman WF, Street-Perrott AF, and Bartlein PJ (eds.) (1993) *Global Climates Since the Last Glacial Maximum*. Minneapolis, MN: University of Minnesota Press.

Relevant Website

www.lsce.cea.fr/pmip – Paleoclimate Modeling Intercomparison Project (PMIP).

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Alkenone Studies *see* **Paleoceanography, Biological Proxies:** Alkenone Paleothermometry Based on the Haptophyte Algae

Allostratigraphy *see* **Quaternary Stratigraphy:** Morphostratigraphy/Allostratigraphy

Amino Acid Dating

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Amino acids are simple organic molecules that are the fundamental building blocks of all proteins. Although there are a large number of possible amino acids, only approximately 20 are common constituents of proteins. Amino acids are assembled in long chains, in which they are linked by peptide bonds. Except for glycine, one of the simplest forms, all protein amino acids have at least one asymmetric carbon atom or center of symmetry. Such molecules are considered chiral molecules (derived from the Greek word for ‘hand’). Chiral molecules exhibit ‘handedness’ in that their mirror images are not superimposable (**Figure 1**). Consequently, they can exist in at least two different configurations or enantiomers.

Chiral forms of the same amino acid were originally identified by the way in which they rotated plane-polarized light passed through a solution containing the pure enantiomer. Enantiomers rotating the light to the left are designated L-amino acids (levorotatory), whereas those enantiomers rotating the light to the right are D-amino acid (dextrorotatory). For reasons that remain debated, life is left-handed, and almost all living organisms utilize exclusively L-amino acids to build protein. However, once biological constraints are removed upon the death of an organism, diagenetic reactions begin to degrade proteinaceous residues into their constituent amino acids, which spontaneously interconvert to their D-configurations in a regular and predictable manner. The process whereby an L-amino acid interconverts to its D-configuration is known as racemization.

Philip Abelson, working at the Carnegie Institution of Washington’s Geophysical Laboratory, was the first to recognize that amino acids could survive for millions of years at ambient Earth conditions, and that their unique initial status of 100% left-handed configuration offered the potential of

using amino acid racemization (AAR) as a geological clock or dating technique ([Abelson, 1954](#)). The first application of AAR as a dating tool was by Ed Hare, working with Abelson in the late 1960s ([Hare and Abelson, 1968](#); [Hare and Mitterer, 1967](#)). Early reviews of the application of AAR to geological dating are provided by [Schroeder and Bada \(1976\)](#) and [Williams and Smith \(1977\)](#). Recent reviews include [Johnson and Miller \(1997\)](#), [Kaufman and Miller \(1992\)](#), [Miller and Brigham-Grette \(1989\)](#), [Murray-Wallace and Kimber \(1993\)](#), [Wehmiller \(1990\)](#), and [Wehmiller and Miller \(2000\)](#), as well as two edited volumes focusing on the chemistry of amino acids in geological environments ([Goodfriend et al., 2000](#); [Hare et al., 1980](#)).

Factors Influencing the State of Biomolecules over Geologic Time

Amino acid geochronology belongs to the chemical family of dating methods. Chemical methods differ from radioactive dating techniques in that their reaction rate depends on one or more environmental parameters, whereas radioactive decay remains constant regardless of environmental conditions. The rates at which proteins are degraded and their constituent amino acids altered, the processes fundamental to amino acid geochronology, are primarily dependent on the temperature the samples have experienced. Consequently, the extent of any particular chemical alteration of amino acids depends on the time elapsed since the death of the organism and the integrated thermal history, and to a lesser extent in most settings on environmental pH, and vital effects unique to each taxon.

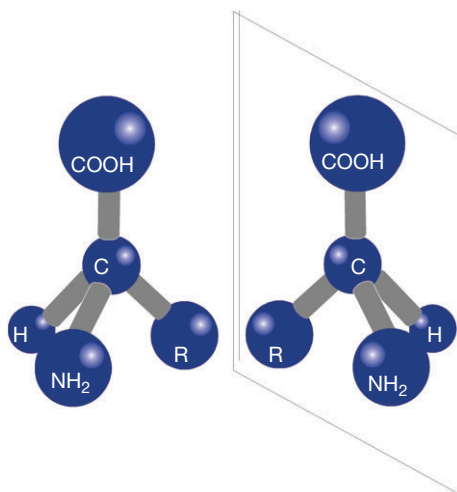


Figure 1 The structure of a generic amino acid and the mirror-image relation of its two stereoisomers or enantiomers.

Preservation of Amino Acids

The idea that amino acids might be preserved in biominerals over long periods of geological time was first proposed by [Abelson \(1954\)](#), who reported that the amino acids most stable at high temperatures in the laboratory were also present in fossils, the oldest of which was a 360 million-year-old Devonian fish. [Hare and Abelson \(1968\)](#) were able to separate L-isoleucine from its epimer, D-alloisoleucine, and they showed a correlation between the ratio of these epimers and time since the death of the organism. Subsequently, [Hare and Mitterer \(1969\)](#) demonstrated that changes in the amino acid composition of mollusk shells over geological time could be induced in the laboratory by heating, implying that the diagenesis of amino acids took place systematically. Supported by this observation, these authors were the first to estimate an age for a fossil based on the extent of AAR.

Amino acids are preserved in the geological record if they fail to enter the biotic nitrogen cycle. The nitrogen of polypeptides represents the major pool of organic nitrogen readily utilized in the biosphere. Without biological attack, the simplest amino acids are stable at ambient Earth surface temperatures for hundreds of millions of years. A common environment for the preservation of amino acids is the carbonate skeleton of invertebrates such as mollusks and Foraminifera. Organic material is incorporated into carbonate exoskeletons either between the mineral crystals (intercrystalline), where it usually has a role in determining the form and shape of the mineral phase during precipitation, or within crystals (intracrystalline), where it may serve to provide resiliency to the mineral phase.

Diagenesis refers to reactions that degrade organic molecules and their constituent amino acids during their residence in geological environments following the death of the host organism or its disuse of the molecules (e.g., avian eggshell proteins). Ultimately, the original complex proteins are degraded into simple hydrocarbons and nitrogen compounds. The organic residues present in fossils are not proteins in a biochemical sense but are commonly referred to as indigenous protein residues, reflecting their biological origin. In nearly all

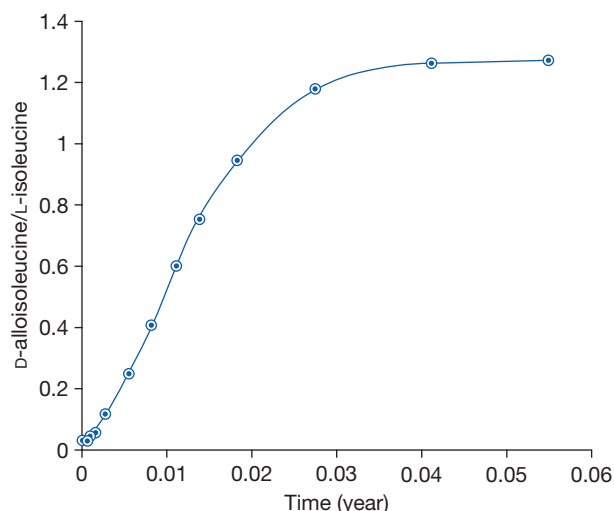


Figure 2 Characteristic trajectory of increasing D-alloisoleucine/L-isoleucine (A/L) with increasing isothermal heating time of a modern biomineral, in this case a modern eggshell of the Australian emu heated at 143 °C. Modified from Miller GH, Hart CP, Roark EB, and Johnson BJ (2000) Isoleucine epimerization in eggshells of the flightless Australian birds, *Genyornis* and *Dromaius*. In: Goodfriend G, Collins M, Fogel M, Macko S, and Wehmiller J (eds.) *Perspectives in Amino Acid and Protein Geochemistry*, pp. 161–181. New York: Oxford University Press.

situations, when a biomineral enters the geological environment, all of its amino acids are in the L-configuration; this defines the starting point of the geological clock.

Amino Acid Racemization

Racemization is the chemical reaction that interconverts an amino acid from one molecular arrangement into its alternative enantiomer or mirror-image form ([Figure 1](#)). The rate at which racemization proceeds varies between amino acids according to the ability of R groups to stabilize a carbanion intermediate. The carbanion intermediate is the molecule formed when the hydrogen side chain is abstracted from an amino acid. In an ideal system, upon readdition of the hydrogen atom to the carbanion intermediate, there is an equal probability of L- and D-amino acid formation, and the forward and reverse reaction rates are equal. Under these conditions, D/L proceeds from values near zero in modern tissues until D/L reaches unity at racemic equilibrium. Because the reaction is reversible, a consequence of these conditions is the slowing of the apparent racemization rate as the extent of racemization increases – the characteristic curvilinear trend produced when D/L ratios are plotted against linear time ([Figure 2](#)).

Not all amino acids are suitable for geochronology. Glycine, the simplest and one of the most common amino acids, is not a chiral molecule. Other simple amino acids, such as alanine, are created in fossils by the decomposition of other more complex amino acids. Thermodynamically unstable amino acids, such as serine and threonine, have half-lives too short for most geological applications. The relatively slow-racemizing amino acids isoleucine, leucine, valine, and

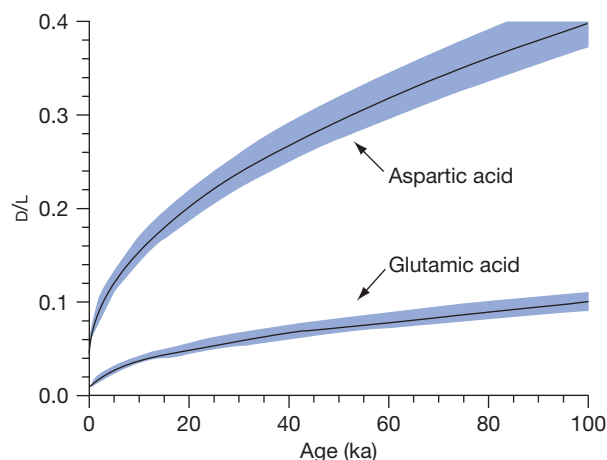


Figure 3 Relation of D/L for aspartic acid (Asp) and glutamic acid (Glu) with respect to calibrated age in the ostracode *Candona* from deep lakes in the western United States. Gray envelopes denote $\pm 1\sigma$ uncertainty. Reproduced from Kaufman DS (2003) Dating deep-lake sediments by using amino acid racemization in fossil ostracodes. *Geology* 31: 1049–1052.

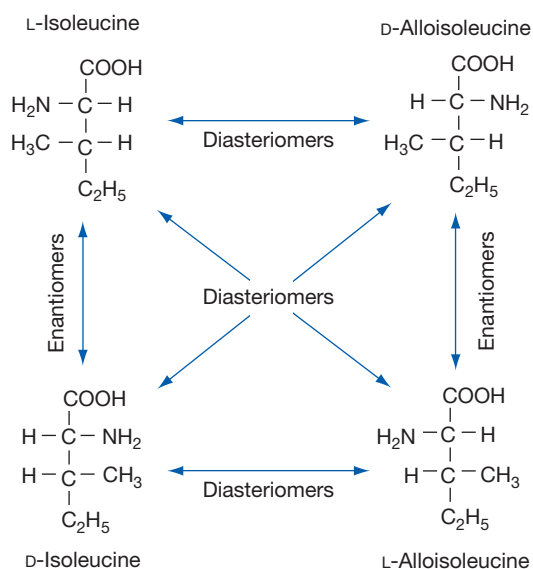


Figure 4 Possible racemization pathways for the protein amino acid L-isoleucine. In nature, nearly all of the L-isoleucine epimerizes to D-alloisoleucine.

glutamic acid are widely used in geochronology, as is the relatively fast-racemizing aspartic acid (Figure 3).

The amino acids commonly utilized for geochronological applications, with the exception of isoleucine, have a single chiral carbon atom and can exist in either their D- or L-forms, which have identical physical and chemical properties except in their interaction with other chiral substances. Isoleucine has two chiral carbon atoms and can racemize about either or both carbon atoms (Figure 4). In nature, L-isoleucine racemizes almost exclusively about the alpha carbon, producing a new molecule, D-alloisoleucine ('allo' meaning other). Because D-alloisoleucine is not a mirror image of L-isoleucine, it has

different chemical and physical characteristics, making it easier to separate in the laboratory. The reaction by which L-isoleucine converts to D-alloisoleucine is known as epimerization, and the extent of isoleucine epimerization is abbreviated A/I or alle/Ile. The equilibrium ratio of D-alloisoleucine to L-isoleucine of approximately 1.30 indicates the reverse reaction (D-allo formation) is 30% more likely than the forward reaction (L-Ile formation; Williams and Smith, 1977).

Temperature

Rates of AAR and most other diagenetic reactions, such as hydrolysis, deamination, decarboxylation, and condensation, are highly sensitive to temperature. The temperature dependency of these reactions is described by the Arrhenius equation:

$$k_1 = Ae^{(-E_a/RT)} \quad [1]$$

$$\ln(k_1) = \ln(A) - E_a/RT \quad [2]$$

where k_1 is the forward rate constant; A is a constant, often referred to as the Arrhenius factor; E_a is the activation energy; R is the universal gas constant; and T is the integrated thermal history T_{eff} in Kelvin.

The Arrhenius equation is derived empirically as the best fit to observed changes in reaction rates with respect to temperature. This is usually achieved by heating modern samples isothermally for specific time intervals in the presence of water vapor and determining the temperature-specific rate constant (k_1) for each temperature. In practice, this can be accomplished for temperatures between 80 and 160 °C. Lower temperature rate constants are usually derived from radiocarbon-dated samples in sites of near-constant temperature. Equation [2] indicates that plotting $\ln(k_1)$ against the reciprocal of its absolute temperature will define a straight line, with the activation energy directly proportional to the slope of the line and the Arrhenius factor defined by the intercept (Figure 5).

The rate constant, k_1 , is related to time (t , measured in years) by the integrated rate equation (Bada and Schroeder, 1972):

$$k_1 = \ln \left[(1 + D/L) / (1 - K'D/L) \right] \times \left[(1 + K')t \right]^{-1} + c \quad [3]$$

where $K' = k_2/k_1$; c is a constant to account for laboratory-induced racemization: $c = \ln[(1 + D/L_0)/(1 - K'D/L_0)]$; and D/L_0 represents the extent of racemization measured in a modern specimen (e.g., live at time of collection, equivalent to time zero, or the start of the geologic clock). For most amino acids the forward and reverse rate constants, k_1 and k_2 , are equal; therefore, $K' = 1$. For the interconversion of L-Ile to D-allo, $k_1/k_2 = 1.3$, and $K' = 0.77$.

The activation energies for racemization of most amino acids preserved in carbonate fossils are between 28 and 30 kcal mol⁻¹. This range implies that the reaction rate approximately doubles for every 4 °C increase in burial temperature at typical ambient temperatures (−10 to +30 °C). The effect of temperature on reaction rate is readily apparent in a plot of racemization levels for last interglacial 'moderate' racemization rate mollusks from the European Arctic to the Mediterranean (Figure 6).

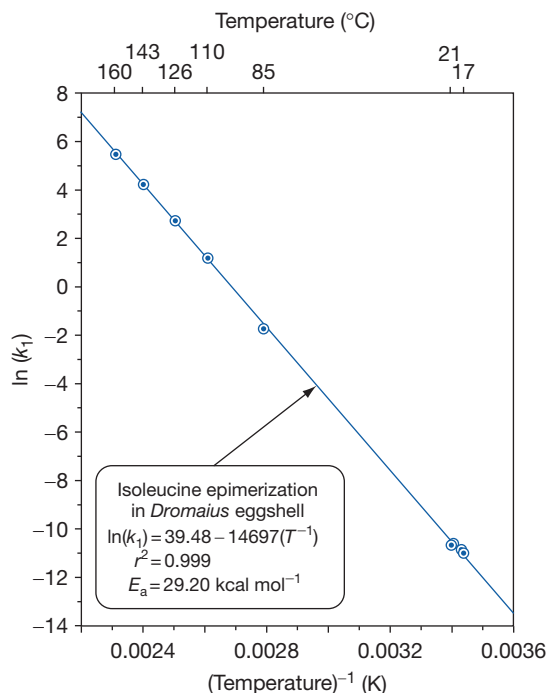


Figure 5 Arrhenius plot for isoleucine epimerization in eggshells of *Dromaius*, the Australian emu, based on isothermal heating of modern eggshell at 160, 143, 126, 110, and 85 °C and radiocarbon-dated early Holocene specimens between 21 and 17 °C. Reproduced from Miller GH, Hart CP, Roark EB, and Johnson BJ (2000) Isoleucine epimerization in eggshells of the flightless Australian birds, *Genyornis* and *Dromaius*. In: Goodfriend G, Collins M, Fogel M, Macko S, and Wehmiller J (eds.) *Perspectives in Amino Acid and Protein Geochemistry*, pp. 161–181. New York: Oxford University Press, with permission.

The exponential dependence of reaction rate on temperature [eqn. [1]] places significant constraints on optimal geological contexts for samples. The T_{eff} experienced by a buried biomineral is strongly influenced by the magnitude of the temperature variations during the postdepositional interval. Consider a simple case. A fossil spends half of its time buried at a temperature of 0 °C and half at 20 °C. The arithmetic mean temperature is 10 °C, but the extent of racemization exhibited by the fossil would be the same as if it had been stored at a constant 16.7 °C for the entire time, dramatically higher than the arithmetic mean temperature (Figure 7). For this reason, racemization temperature is defined as the effective diagenetic temperature (T_{eff}). T_{eff} is the equivalent temperature required to explain the extent of observed racemization if the sample had experienced a constant postburial temperature. It is always equal to or higher than the arithmetic mean temperature.

Sites with high diurnal or seasonal temperature fluctuations will have T_{eff} well above the arithmetic mean temperature. In most environments, samples only need to be buried a few centimeters to escape the diurnal heating cycle, but the seasonal cycle in continental climates, although rapidly damped, can extend to a depth of several meters. Reactions monitored at a continental site in southern New Mexico exhibit a seasonal temperature amplitude over 20 °C in the upper few centimeters, resulting in an T_{eff} more than 10 °C above the mean annual air temperature (MAT; Figure 8). The amplitude of

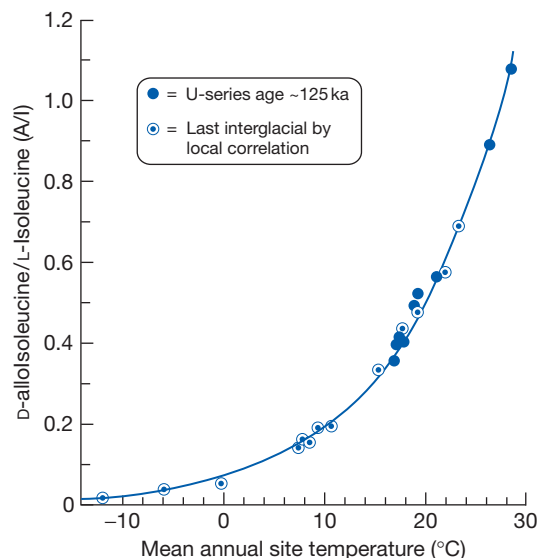


Figure 6 D-alloisoleucine/L-isoleucine (A/I) in moderate racemization rate mollusks from last interglacial (~125 ka) sites across western Europe, ranging from Svalbard and Arctic Russia to the Mediterranean Basin, plotted against current mean annual site temperature. Last interglacial sites are dated by U/Th on corals or correlated to the last interglacial on the basis of diagnostic marine faunal elements. These data document the close correlation to and exponential dependency of racemization rate to site temperature. Unpublished data from G. H. Miller and P. J. Hearty, University of Colorado.

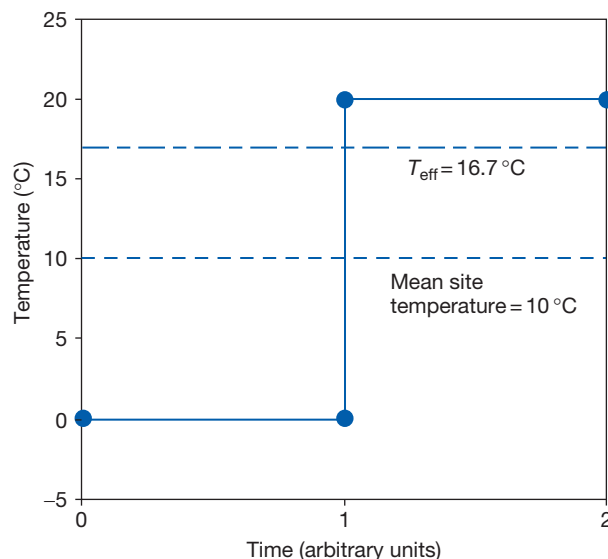


Figure 7 A simplified illustration of the impact of variable burial history. A fossil is buried for equal time at 20 and 0 °C. The arithmetic mean temperature is 10 °C. However, because of the exponential dependency of racemization rate on temperature, the fossil will exhibit the same amount of racemization as a sample maintained for the same total time at a much higher temperature (16.7 °C). This is the effective diagenetic temperature (T_{eff}).

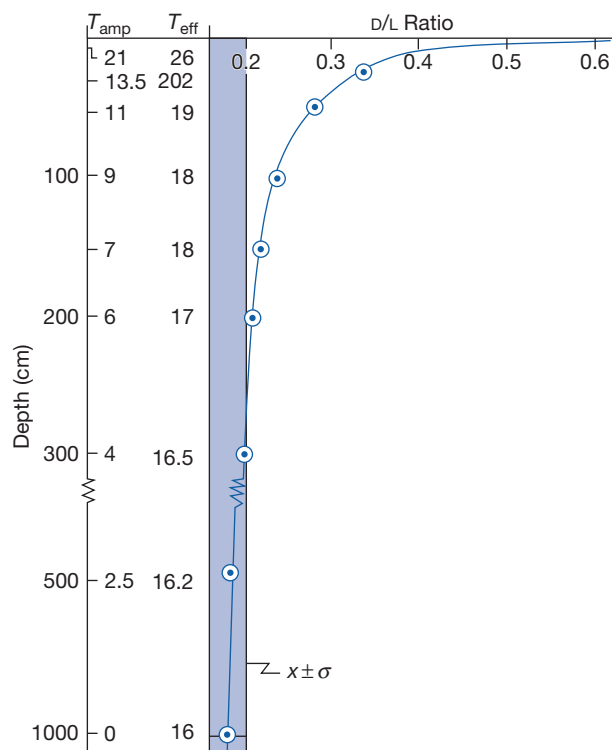


Figure 8 The difference between effective diagenetic temperature (T_{eff}) and mean annual air temperature (MAT; mean, 16.8 °C) with depth based on multiyear instrumentation of a strongly continental site in southern Arizona. The relatively high amplitude of the annual temperature cycle near the surface ($T_{\text{amp}} = 21$ °C) results in an T_{eff} 9 °C above the MAT. The predicted A/I in a gastropod held at specific depths for 14 ka is plotted on the By a depth of approximately 1 m, the difference between predicted and expected A/I is within analytical uncertainties, but at shallow burial depths the predicted A/I deviates sharply from that expected based on the site MAT. Modified from Miller GH and Brigham-Grette J (1989) Amino acid geochronology: Resolution and precision in carbonate fossils. *Quaternary International* 1: 111–128.

the annual temperature cycle decreases rapidly so that by 2 m depth it is <6 °C, at which point the difference between T_{eff} and MAT is negligible (<1 °C; **Figure 8**).

Sampling strategies must recognize the importance of avoiding high-amplitude temperature fluctuations. In exposed continental sites, collecting sites should have been buried 2 m or more for most of their postdepositional history. Shaded sites and sites from more maritime regions need not be so deeply buried.

Environmental pH

Racemization is known to be strongly base catalyzed (e.g., **Smith et al., 1978**), and biominerals heated in solutions above pH 9 racemize more quickly than do the same biominerals heated at neutral pH. For example, Asp D/L values were 29% higher in mollusk shell heated in pH 10 solution compared with pH 7, and Glu D/L values were 134% higher (**Figure 9**; **Orem and Kaufman, 2011**). The proportion of free amino acids in the laboratory-heated shells also increased with increasing pH, as did the rate of removal of amino acids from

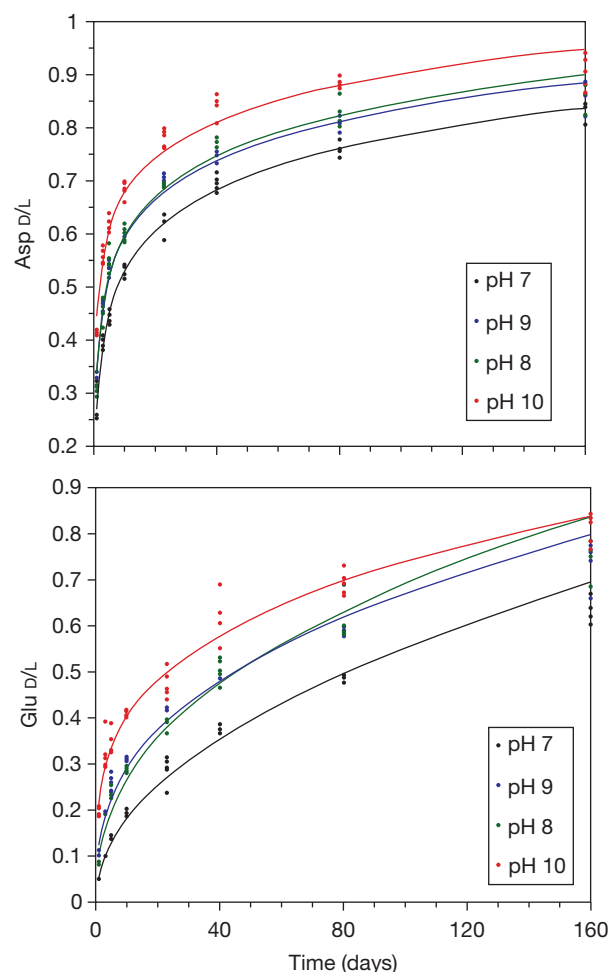


Figure 9 Rate of racemization for aspartic acid, and glutamic acid in *Margaritifera* shell heated in buffered solutions with pH ranging from 7 to 10. Shells were heated at 110 °C for up to 160 h. Rates of racemization increase with increasing pH. Curves are simple logarithmic (Asp) and power (Glu) fits. Data from **Orem and Kaufman (2011)**.

the shell matrix (leaching). A similar experiment using ostracodes showed that the D/L values and amino acid concentrations were higher in valves heated in pH 10 buffer than in valves heated in pH 7 buffer (**Bright and Kaufman, 2011a**). Carbonate fossils have the advantage of buffering ambient pH and maintaining a local environmental pH of ~8. Samples from strongly basic environments (pore fluid pH > 9) should be avoided.

Taxonomic Effects

Position in a Polypeptide Chain

The negative correlation between the molecular weight of polypeptides and the extent of racemization among their amino acids is a product of the relation between the racemization rate of an amino acid and its location within the polypeptide. For example, the extent of isoleucine epimerization in various molecular fractions of three *Genyornis* eggshells is always lowest in the highest molecular-weight fraction, whereas the

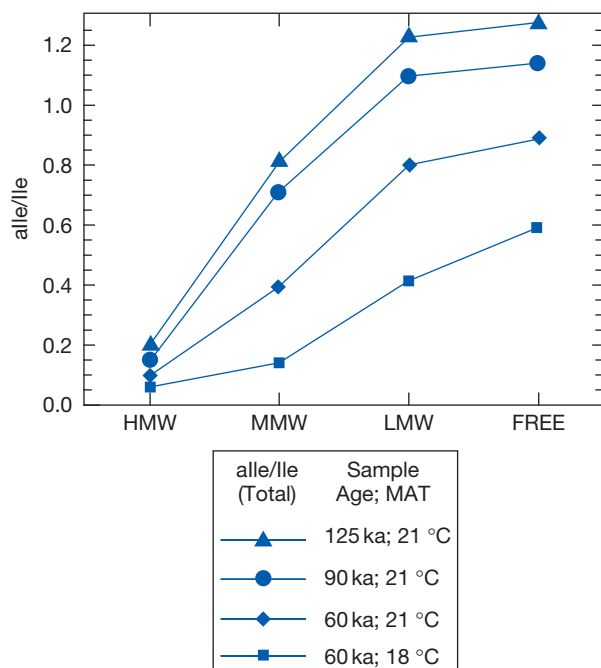


Figure 10 Extent of isoleucine epimerization (alle/Ile) measured in four molecular-weight fractions of four late Pleistocene eggshells of the extinct giant Australian bird, *Genyornis*. Alle/Ile was measured in high-molecular-weight (HMW), moderate-molecular-weight (MMW), low-molecular-weight (LMW), and free (FREE) fractions. In all four samples, alle/Ile increased with decreasing molecular weight. Symbols on the right are alle/Ile in the total acid hydrolysate. Modified from Kaufman DK and Miller GH (1995) Isoleucine epimerization and amino acid composition in molecular-weight separations of Pleistocene *Genyornis* eggshell. *Geochimica et Cosmochimica Acta* 59: 2757–2765.

low-molecular-weight and free amino acid fractions have the highest *A/I* values (Figure 10; Kaufman and Miller, 1995). Laboratory experiments indicate that amino acids located at interior positions within peptide chains feature low rates of racemization, whereas amino acids located at terminal positions and incorporated in diketopiperazines racemize most rapidly. The *D/L* of free amino acids is commonly high but the reaction rate is low when the molecules exist in this state. This potential incongruity is explained by the preferential release of relatively highly racemized amino acids from terminal positions into the free amino acid pool. Relatively intact polypeptides have a higher ratio of interior to terminal amino acids than smaller polypeptides; consequently, high-molecular-weight separations feature a lower extent of racemization.

The differences in racemization rates produce two pools of amino acids with different average *D/L* that are readily separated: 'free' and 'total acid hydrolysate.' Normally, peptide bonds are hydrolyzed prior to amino acid analysis by heating the decalcified residue (typically 22 h at 110 °C) in a strong acid (6 N HCl) to yield the total acid hydrolysate fraction. Alternatively, the unhydrolyzed residue, or free fraction, is dominated by naturally hydrolyzed amino acids. Due to the more advanced level of racemization in this pool, *D/L* in the free fraction may provide a more effective means of separating intervals of time than is attainable with *D/L* in the total acid hydrolysate.

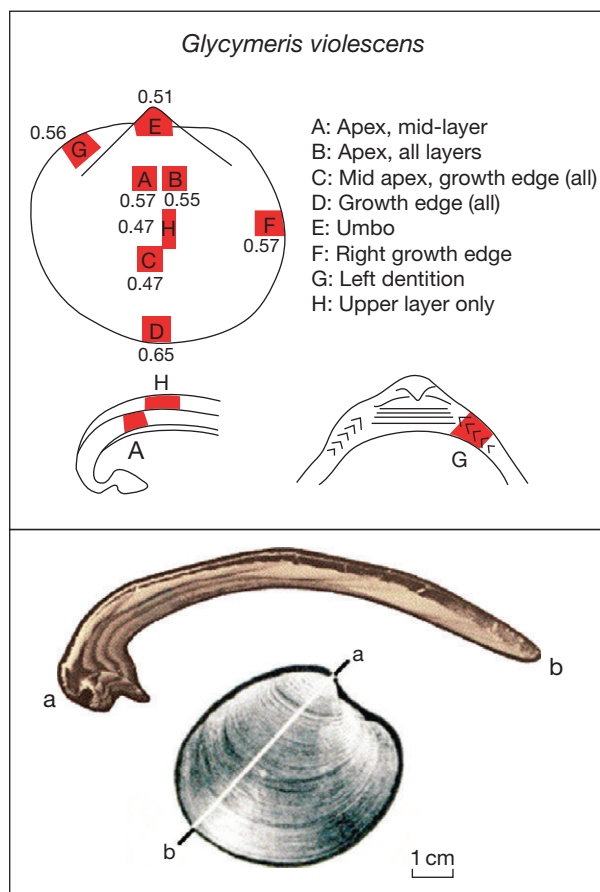


Figure 11 (Top) Typical intrashell variability in *A/I* – in this case, in a last interglacial age marine taxodont *Glycymeris* from the Mediterranean. (Bottom) A cross section through a marine pelecypod, showing how growth bands extending to the surface can serve as pathways that allow leaching of low-molecular-weight organic compounds. The optimal sample site in both examples is an inner layer, near the umbo region ('A' and 'B' in the top panel), where the shell is dense and there are fewer pathways for leaching.

Optimizing Sample Sites Within a Fossil

Many fossils, including mollusks and eggshells, have different structural layers. Usually, each structural layer contains a different suite of amino acids. Because the racemization rate for any specific amino acid depends to some extent on the nature of its bonds with adjacent amino acids in its peptide chain, racemization rates commonly vary in different structural layers. Consequently, sampling should be restricted to a single layer.

Tests of mollusk *A/I* variability indicate values may differ by up to 40% in different parts of even relatively simple bivalves (Figure 11, top). In bivalve shells, most of the protein residues reside along intercrystalline growth, where they act to nucleate incremental shell growth. Leaching of amino acids and small peptides is enhanced where these growth bands intersect the surface of the shell (Figure 11, bottom). Consequently, optimal sampling sites for AAR studies are located at sites 'A' or 'B' in Figure 11 (top).

Other Vital Effects

Differences up to a factor of 2 in racemization rates according to taxonomy have been noted for mollusks, Foraminifera, and avian eggshells. Among mollusks, there are significant differences between genera but broadly comparable reaction rates among species of the same genus. The reason for the taxonomic dependency is not precisely known, but because they are accompanied by differences in amino acid composition, likely explanations are differences either in the amino acid composition of the precursor proteins or in the relative abundance of the numerous proteins involved in the secretion of these carbonate structures. For example, Goodfriend (1992) observed high concentrations and very rapid racemization of Asx residues in the coral *Porites australiensis*. Because peptide bonds at Asx sites are prone to hydrolysis, the authors interpreted these observations as cause and effect. Where multiple species are recovered in association, intergeneric differences in racemization rates provide an internal check on the reliability of D/L values. Similarly, because amino acids feature characteristic rates of racemization, a comparison of the D/L values of different amino acids within a sample provides an additional means of assessing the veracity of data.

Almost all biominerals contain a relative abundance of specific amino acids that are unique at the generic level. This characteristic signature can be used as a taxonomic tool and provides a test for impurities in most geological samples.

Leaching

The physical transport of amino acids out of mineral matrices by an aqueous phase, commonly water, is termed 'leaching.'

The rate of leaching is controlled by the size of the mobile molecules and their sorption to the fossil matrix, the diameter and complexity of the diffusion pathway, and temperature (Collins and Riley, 2000). Molecules most susceptible to leaching are the low-molecular-weight solutes (i.e., free amino acids, small peptides, and diketopiperazines) in which protein residues are situated in pathways that lead to the exterior of the biomineral. Because low-molecular-weight residues tend to be more racemized, leaching decreases the overall level of racemization. Hot, alternately wet and dry environments are most conducive to leaching.

Protein residues preserved in intracrystalline sites are least susceptible to leaching, and have been shown to approximate a closed system (e.g., Miller et al., 2000; Sykes et al., 1995). Isolating and analyzing this fraction of amino acids can improve the reliability of amino acid geochronology. Most studies of the intracrystalline fraction for Quaternary geochronology have focused on mollusks. Exposure of powdered mollusk shell to concentrated sodium hypochlorite (bleach) for 48 h effectively reduces the amino acid content to a residual level, assumed to represent the intracrystalline fraction. Penkman et al. (2008) showed that the shell-to-shell variability in D/L values for the intracrystalline fraction of a single-age population was reduced by 50% compared with conventionally analyzed shells, and the correlations between D/L values between free and hydrolyzed amino acids were increased (Figure 12). The effect of the same bleaching pretreatment on freshwater ostracodes (crustacean), however, is less clear. Bright and Kaufman (2011b) showed that bleaching had virtually no impact on the amino acid concentration or the D/L values in fossil ostracodes valves older than approximately 1 ka. Nor did the pretreatment reduce the intrasample variability. The proteins associated with the chitin-rich organic matrix of ostracode

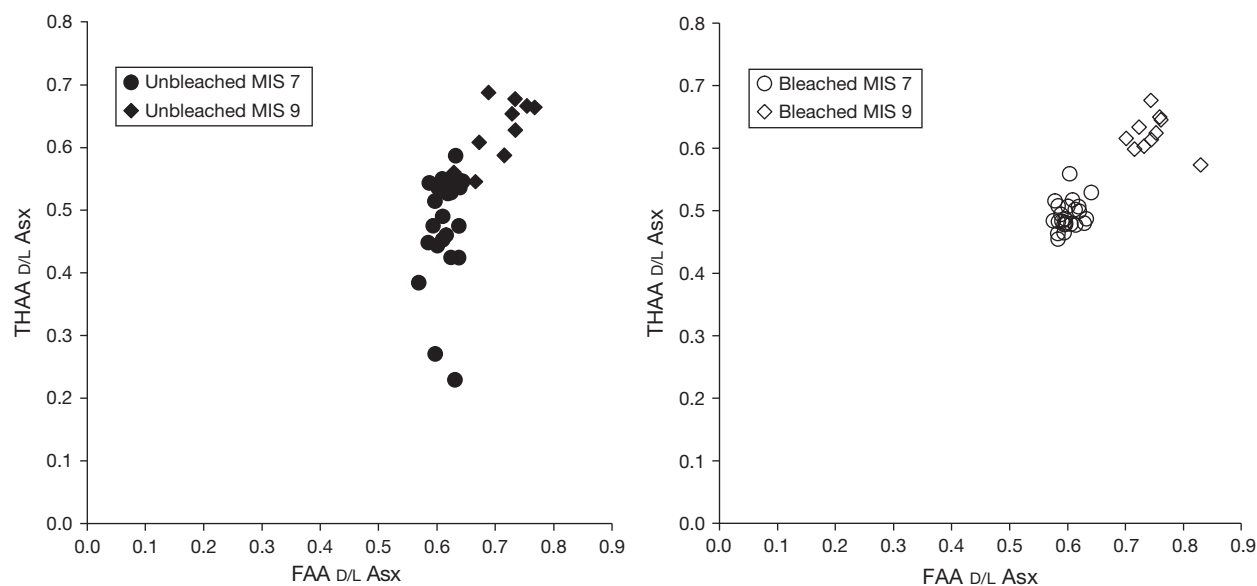


Figure 12 Asx D/L values in the total hydrolysable amino acid (THAA) population versus Asx D/L in the free amino acid (FAA) population of the same shell for unbleached (left) and bleached (right) subsamples of *Bithynia* shells from marine isotope stage 7 and 9 deposits in the United Kingdom. Results of analyses on bleached shells (intracrystalline fraction) yielded greater consistency. Reproduced from Penkman KEH, Kaufman DS, Maddy D, and Collins MJ (2008) Closed-system behaviour of the intra-crystalline fraction of amino acids in mollusc shells. *Quaternary Geochronology* 3: 2–25.

valves appear to behave somewhat differently than in mollusk shells.

Separation and Detection of D- and L-Amino Acids

The separation, detection, and quantification of the D- and L-amino acid enantiomers and diastereomers are accomplished by either gas or high-pressure liquid chromatography (HPLC). L-Isoleucine can be separated from its D-alloisoleucine pair by conventional ion-exchange HPLC, which provides a quantitative assessment of most other protein amino acids and allows an independent assessment of sample quality. Most other D-/L-amino acid pairs are separated by gas chromatography or reverse-phase liquid chromatography using a chiral stationary or mobile phase to separate the enantiomers (Figure 13). An interlaboratory standard (Wehmiller, 1984) provides an independent check on instrumental accuracy.

Applications

AAR has potential application in the Quaternary sciences wherever amino acid residues are preserved over geological time and questions of time or temperature remain. To utilize AAR as a dating tool requires a medium that not only preserves indigenous protein amino acids from bacterial attack and

leaching but also excludes contamination by other amino acids. The more tightly the amino acids are integrated into the biominerals, the closer the medium will be to a closed system. Commonly utilized fossils include marine bivalves, land and aquatic snail shells, eggshells, Foraminifera, and ostracodes. Most are abundant in the geologic record and preserve amino acids well over geologically significant timescales. Other materials that have been analyzed, some with mixed success, include fish otoliths, woodrat middens, wood, bones and teeth, corals, and the carbonate spores of charophyte oögonia. Although most applications utilize amino acids derived from known biological sources, racemization studies in biogenic carbonate breccia (e.g., 'whole rock' eolianites), speleothems, and paleosols have proven successful.

The time interval over which racemization is effective is determined by the reaction rate: it is longer in cold climates (low rate) than in tropical climates (high rate). At room temperature (22 °C), conversion from an optically pure L-amino acid ($D/L=0$) to a racemic mixture lacking optical activity ($D/L=1.0$ for most amino acids) requires several tens of thousands of years for most amino acids. At 160 °C, a racemic mixture is reached within 10–20 h, whereas in Arctic regions (–10 °C) racemic equilibrium requires 1–2 Ma. The useful time range for dating under different mean annual temperatures for isoleucine epimerization (valine, leucine, and glutamic acid racemization have similar time ranges) is illustrated in Figure 14.

Aspartic acid racemizes approximately an order of magnitude faster than other amino acids and is frequently used for very young materials or extremely cold sites (Figure 3). In tropical settings, the rapid racemization of Asx residues provides age resolution for very young samples, whereas the slower epimerization of isoleucine is effective at distinguishing events that are thousands of years apart (Figures 3, 15, and 16).

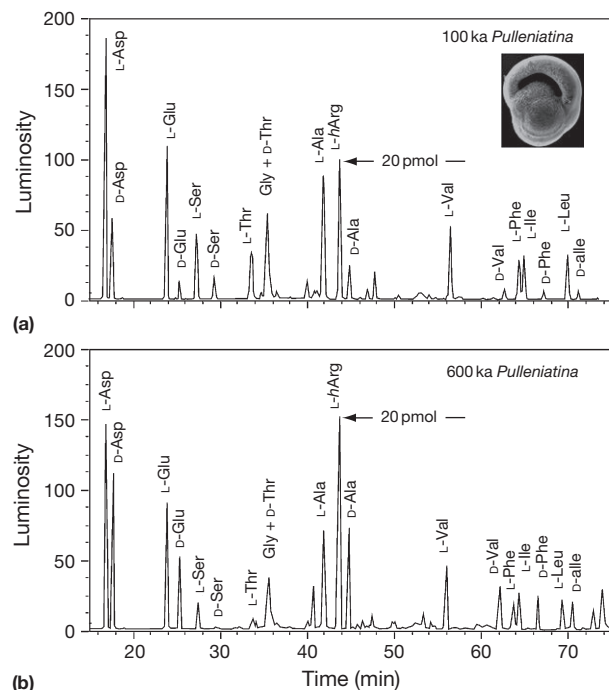


Figure 13 Reverse-phase liquid chromatography chromatograms of single individuals of the large planktonic foraminifer *Pulleniatina* from marine cores recovered off northeastern Australia, with approximate ages of 100 ka (a) and 600 ka (b). Modified from Hearty PJ, O'Leary MJ, Kaufman DS, Page M, and Bright J (2004) Amino acid geochronology of individual foraminifer (*Pulleniatina obliquiloculata*) tests, north Queensland margin, Australia: A new approach to correlating and dating Quaternary tropical marine sediment cores. *Paleoceanography* 19: PA4022.

Aminostratigraphy

AAR can be applied as a geological dating tool in a relative and absolute sense. As a relative dating tool, the technique is

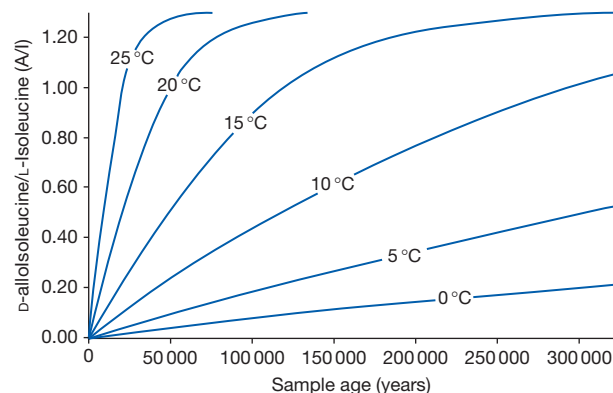


Figure 14 Relation between the extent of isoleucine epimerization (A/I) and sample age for T_{eff} between 0 and 25 °C showing the strong dependence of epimerization rate on temperature. Lines are derived from kinetics in eggshells of the Australian emu (*Dromaius*) shown in Miller et al. (2000), but the relation is generally applicable to most carbonate fossils.

known as aminostratigraphy. Aminozones, a chemostratigraphic unit based on clustering in the extent of AAR in replicate fossils, are the fundamental units of aminostratigraphy. The premise is that over a limited geographic range and elevation, samples experienced a similar thermal history (MAT, $\pm 1^\circ\text{C}$; temperature amplitude at collection site, $< 6^\circ\text{C}$). Over this limited region, samples with similar D/L are of the same age, those sites with higher D/L are older, and those with lower D/L are younger. Aminostratigraphy does not require an

assessment of racemization kinetics or estimates of past thermal histories, and it is the least ambiguous application of amino acid geochronology.

Aminostratigraphy has been applied to mollusks excavated from raised marine terraces in coastal regions throughout the world. One example is a series of uplifted interglacial terraces along the Peruvian coast from which bivalve mollusks exhibit a regular increase in D/L leucine in higher (older) terraces, with corresponding absolute ages derived from electron spin resonance (ESR) dating (Figure 17).

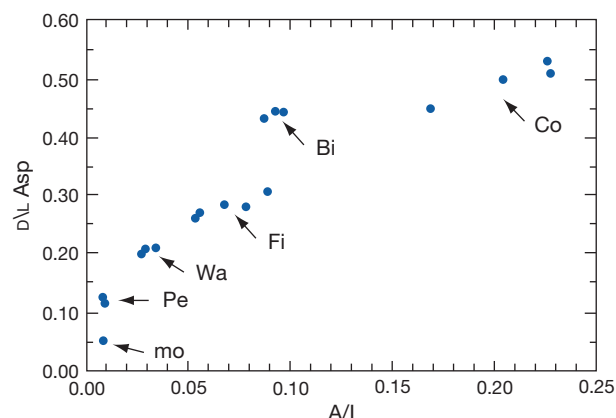


Figure 15 Fast-racemizing amino acids provide superior resolution for cold or very young samples. D/L aspartic acid and D-alloisoleucine/L-isoleucine (A/I) were measured in the marine bivalve *Hiatella arctica* collected from raised marine terraces along the north coast of Alaska (mean annual temperature, ca. -12°C). For the younger shorelines, D/L Asp provides a clearer separation of the sites, but this difference decreases for the older shorelines. Aminozones: mo, modern; Pe, Pelukian shoreline (last interglaciation); Wa, Wainwrightian shoreline; Fi, Fishcreekian shoreline; Bi, Bigbendian shoreline; and Co, Colvillian shoreline (probably late Pliocene). Modified from Goodfriend GA, Brigham-Grette J, and Miller GH (1996) Enhanced age resolution of the marine Quaternary record in the Arctic using aspartic acid racemization dating of bivalve shells. *Quaternary Research* 45: 176–187.

Absolute Age and Paleothermometry

Once the temperature-dependence of racemization in a particular biomineral has been derived, the equation describing racemization contains three unknowns: D/L (the extent of the reaction, which can be measured in the laboratory); t , the time since death; and T , the integrated thermal history. Theoretically, if either time (t) or effective diagenetic temperature (T) is known, the other can be predicted (from eqns. [1] and [3]):

$$Ae^{(-E_a/RT)} = \ln[(1 + D/L)/(1 - K'D/L)] \times [(1 + K')t]^{-1} \quad [4]$$

In theory, deriving absolute ages from the extent of AAR requires only an estimate ($\pm 1^\circ\text{C}$) of the T_{eff} . For samples of late Holocene age, this can be reasonably estimated from the instrumental record, but secure estimates for the time-dependent changes in MAT across glacial–interglacial cycles are rarely possible. Bada and Protsch (1973) presented an alternative approach in which an appropriate *in situ* value for k_1 is determined from the D/L ratio of a specimen of known age. This ‘calibrated’ rate constant may be used to determine the age of nearby specimens from their D/L value, assuming they have experienced a comparable thermal history. For Quaternary samples, a reasonable assumption is that a calibration site of

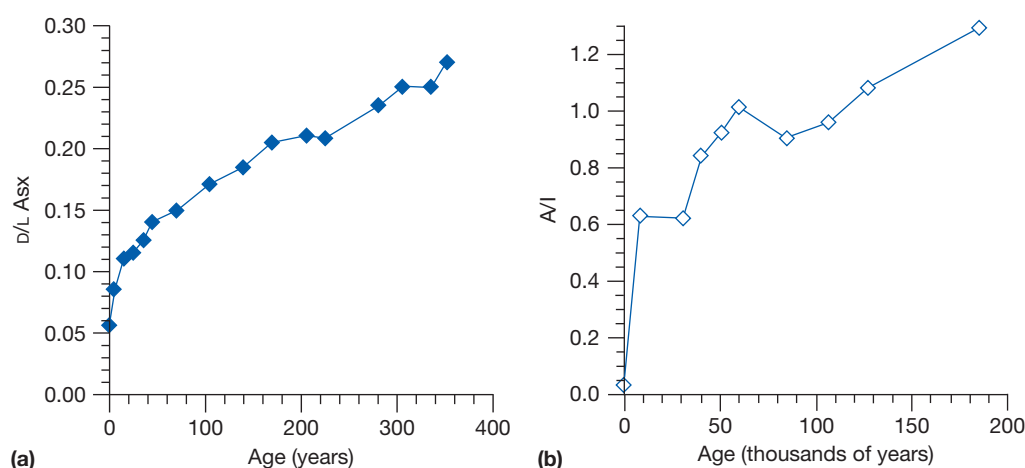


Figure 16 Examples of racemization applied to two very different timescales. D/L Asx in very young corals ((a) data from Goodfriend GA, Hare PE, and Druffel RM (1992) Aspartic acid racemization and protein diagenesis in corals over the last 350 years. *Geochimica et Cosmochimica Acta* 56: 3847–3850) and A/I in the giant clam *Tridacna* from the Huon Peninsula ((b) data from Hearty PJ and Aharon P (1988) Amino acid chronostratigraphy of late Pleistocene coral reef sites: Huon Peninsula, New Guinea and the Great Barrier Reef, Australia. *Geology* 16: 579–583). Note timescales differ by three orders of magnitude.

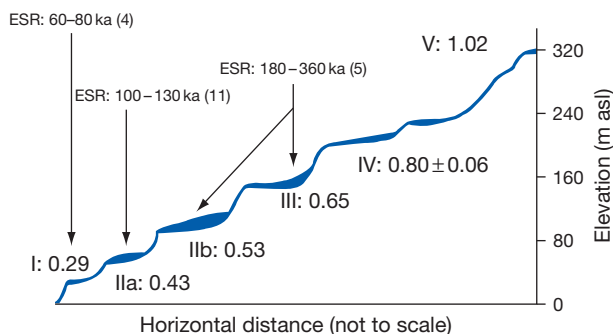


Figure 17 Schematic cross section through a flight of raised marine terraces on the coast of Peru with amino zones (in roman numerals) defined on the extent of leucine racemization. Electron spin resonance (ESR) dates provide additional chronological control. The highest terraces are of early Quaternary age. Modified from Wehmiller JF (1993) Applications of organic geochemistry for Quaternary research: Aminostratigraphy and amino chronology. In: Engel M and Macko S (eds.), *Organic Geochemistry*, pp. 755–783. New York: Plenum.

at least 100 ka has a T_{eff} that is a reliable estimate for most of the Quaternary.

Age estimates for many fossils can be complicated by kinetic uncertainties. An inherent assumption is that racemization follows simple, reversible first-order kinetics. Although this assumption is approximated during the early stages of racemization, most mollusk and Foraminifera shells deviate from predicted linear kinetics in the later stages of diagenesis. This has led several researchers to utilize empirically derived models (Figure 18) that more closely approximate the observed change in D/L with time.

In an evaluation of uncertainties, McCoy (1987) showed that predicting paleotemperatures is inherently more precise than predicting ages. Unlike most biological proxies that record the instantaneous temperature at which an organism lived, racemization-derived paleotemperatures reflect the integrated thermal history (T_{eff}) since the organism died. In the case of deeply buried samples, this equates to the integrated mean annual temperature. If samples with independently derived ages are available from a limited region, then the evolution of temperature across the region can be calculated. For example, the Pleistocene/Holocene temperature change in semiarid Australia was quantified from A/I in emu eggshells, each of which had also been dated by ^{14}C (Figure 19). Not all samples were deeply buried; hence, some samples exhibit more advanced levels of racemization than expected, but the change in slope between samples with the lowest A/I dated <16 ka and those >16 ka is unambiguous and reflects at least 7 °C warming from the cold Last Glacial Maximum to the warmth of the present interglacial.

Time-Averaging

A burgeoning application of amino acid geochronology is dating individual shell fragments that comprise fossil death assemblages. Such assemblages often comprise large collections of mixed-aged shells of one or more species. The temporal resolution of paleoenvironmental information that ultimately can

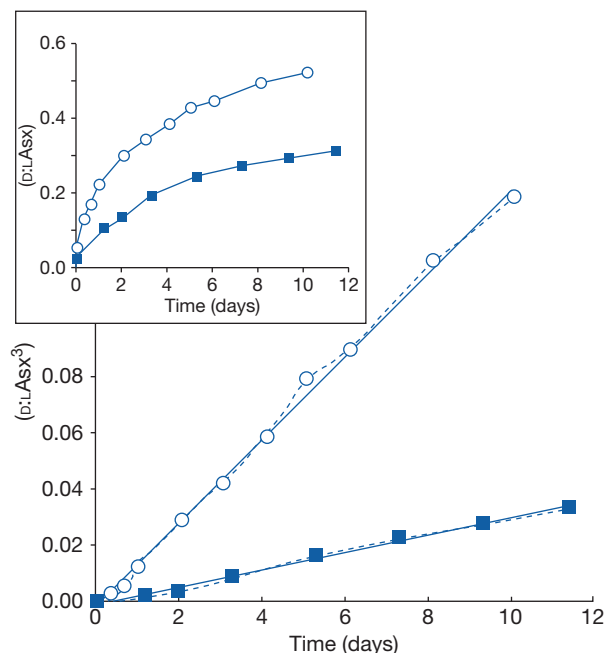


Figure 18 Cubic transformation of D/L Asx to linearize the original kinetics (inset), from which the slope is proportional to sample age. Original data from Goodfriend (1992) (land snails heated at 100 °C, open circles) and Brinton and Bada (1995) (free asparagines heated at 100 °C). Reproduced from Collins MJ, Waite ER, and van Duin ACT (1999) Predicting protein decomposition, the case of aspartic acid racemization kinetics. *Royal Society of London Philosophical Transactions, Series B, Biological Sciences* 354: 51–64, with permission.

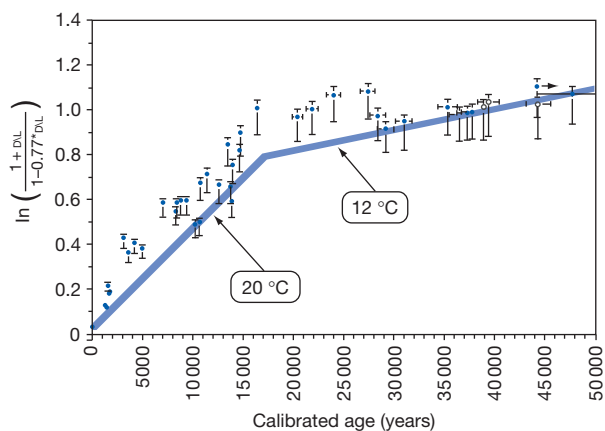


Figure 19 A/I (in its linearized form) plotted against corresponding calibrated radiocarbon ages for 43 samples from central Australia. Asymmetric vertical error bars incorporate the possibility of shallow burial. Horizontal error bars reflect $\pm 1\sigma$ dating errors where the uncertainty exceeds the diameter of the plotted point. Regression lines, which are directly proportional to temperature, go through the lowest A/I . The lower slope for samples older than ~16 ka indicates much lower Pleistocene temperatures in central Australia than during the Holocene. Modified from Miller GH, Magee JW, and Jull AJT (1997) Low-latitude glacial cooling in the Southern Hemisphere from amino acids in emu eggshells. *Nature* 385: 241–244.

be derived from a fossil assemblage is determined by its range of ages (time-averaging). Biogenic carbonate fossil remains are known to survive extensive reworking and prolonged residence times at the sediment–water interface (e.g., Flessa and Kowalewski, 1994). The age distribution of shells provides important information about compositional biases in the fossil record that result from ‘filtering’ processes that are both taxon- and facies-dependent, including shell input and durability (Carroll et al., 2003; Kidwell et al., 2005; Kosnik et al., 2007). In addition, the ages of fossil assemblages provide baseline geohistorical data needed to understand the ecological and evolutionary response of species to environmental changes, the basis of conservation paleobiology (Dietl and Flessa, 2011).

Amino acid geochronology is a cost-effective approach to dating hundreds of shells with accuracy ranging from multi-decadal to centennial, depending on ambient water temperature and on the ability to accurately calibrate the rate of racemization. The basic procedure is to analyze many shells to determine the overall relative-age structure of the death assemblage. Multiple individual fragments distributed across the range of ages indicated by the D/L values are then targeted for accelerator mass spectrometry (AMS) ^{14}C analysis. The resulting relation between ^{14}C age and D/L is used to calibrate the rate of racemization for a particular taxon from a study area (Figure 20). The calibrated rate can then be applied to other shells whose ages were not determined by ^{14}C (e.g., Kosnik et al., 2008).

Because the aim of time-averaging studies is to determine the structure of the age distribution, any data screening must be

done cautiously and systematically. A broad range of screening criteria for identifying outliers and determining the suitability of specimens for numerical dating have been explored based on a large collection of mollusk shells of multiple taxa from a shallow-water core in the central Great Barrier Reef (Kosnik and Kaufman, 2008). The results show that some taxa provide higher-resolution temporal information than others, and highlight the importance of identifying the taxa that are likely to preserve the best geochronological information.

Conclusions

Amino acid geochronology offers a simple but powerful relative dating tool (aminostratigraphy) that provides temporal constraints on biominerals beyond the range of radiocarbon dating. The fast-racemizing Asx can be used to assess differences in age over a few decades to centuries in optimal settings, an interval over which radiocarbon calibration uncertainties are unusually large. In all but the hottest regions, the overall time range includes all or most of the Middle and Late Quaternary, a time window not readily dated by many other methods. When independent calibration is available, the extent of racemization can be used to reconstruct past thermal histories with a precision of $\pm 1^\circ\text{C}$. Also, with suitable calibration the extent of racemization can be converted to reliable absolute ages. Although the measured extent of racemization depends heavily on temperature, the racemization reaction does not involve the loss of reactants; hence, the measurement

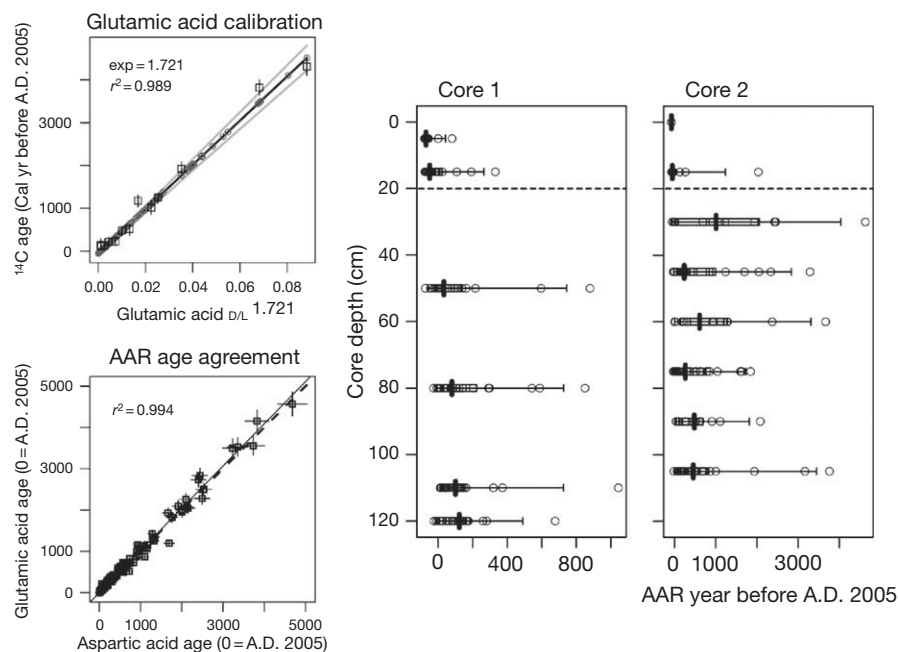


Figure 20 AAR calibration curves for *Tellina* shells from Rib Reef, Australia (left two panels). Open squares represent ^{14}C -dated specimens (vertical line is the 2σ age range). Lines are best-fit linear regression and confidence intervals. Gray circles are AAR ages of shells. D/L values are transformed exponentially to linearize the trend. Right two panels compare racemization age of individual *Tellina* shells to their depth in two different cores. Median (bar), fiftieth percentile (box), and ninety-fifth percentile (whisker) are shown. Samples above 20 cm are younger than samples below the dashed lines. However, there is no clear age versus depth relation below 20 cm, suggesting deep, active bioturbation. Modified from Kosnik MA, Hua Q, Jacobsen GE, Kaufman DS, and Wüst RA (2007) Sediment mixing and stratigraphic disorder revealed by the age-structure of *Tellina* shells in Great Barrier Reef sediment. *Geology* 35: 811–814.

statistics for old samples are similar to those for young samples. For the most stable amino acids, the concentration of the summed D- and L-forms at racemic equilibrium is of a similar magnitude to the concentration of the original L-amino acid, providing there has been no diffusional loss from the biomineral matrix. Not all biominerals are optimal for amino acid geochronology. Those that approximate closed system behavior with respect to indigenous amino acids are optimal. This is usually met in samples for which the amino acids are sequestered within the crystal matrix of a fossil.

See also: Dating Techniques; Fission-Track Dating; K/Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ Dating; U-Series Dating. **Introduction:** History of Dating Methods.

References

- Abelson PH (1954) Organic constituents of fossils. *Carnegie Institution of Washington Year Book* 53: 97–101.
- Bada JL and Protsch R (1973) Racemization reaction of aspartic acid and its use in dating fossil bones. *Proceeding of the National Academy of Sciences USA* 70: 1331–1334.
- Bada JL and Schroeder RA (1972) Racemization of isoleucine in calcareous marine sediments: Kinetics and mechanism. *Earth and Planetary Science Letters* 15: 1–11.
- Bright J and Kaufman DS (2011a) Amino acids in lacustrine ostracodes, part III: Effects of pH and taxonomy on racemization and leaching. *Quaternary Geochronology* 6(6): 574–597.
- Bright J and Kaufman DS (2011b) Amino acid racemization in lacustrine ostracodes, part I: Effect of oxidizing pre-treatments on amino acid composition. *Quaternary Geochronology* 6: 154–173.
- Brinton KL and Bada JL (1995) Aspartic acid racemization and protein diagenesis in corals over the last 350 years – Comment. *Geochimica et Cosmochimica Acta* 59: 415–416.
- Carroll M, Kowalewski M, Simões MG, and Goodfriend GA (2003) Quantitative estimates of time-averaging in terebratulid brachiopod shell accumulations from a modern tropical shelf. *Paleobiology* 29: 381–402.
- Collins MJ and Riley M (2000) Amino acid racemization in biominerals: The impact of protein degradation and loss. In: Goodfriend G, Collins M, Fogel M, Macko S, and Wehmiller J (eds.) *Perspectives in Amino Acid and Protein Geochemistry*, pp. 120–142. New York: Oxford University Press.
- Dietl GP and Flessa KW (2011) Conservation paleobiology: Putting the dead to work. *Trends in Ecology & Evolution* 26: 30–37.
- Flessa KW and Kowalewski M (1994) Shell survival and time averaging in nearshore and shelf environments: Estimates from the radiocarbon literature. *Lethaia* 27: 153–165.
- Goodfriend (1992) Rapid racemization of aspartic acid in mollusc shells and potential for dating over recent centuries. *Nature* 357: 399–401.
- Goodfriend GA, Collins MJ, Fogel ML, Macko SA, and Wehmiller JF (2000) *Perspectives in Amino Acid and Protein Geochemistry*. New York: Oxford University Press.
- Hare PE and Abelson PH (1968) Racemization of amino acids in fossil shells. *Carnegie Institution Washington Year Book* 66: 526–528.
- Hare PE, Hoering TC, and King K (eds.) (1980) *Biogeochemistry of Amino Acids*. New York: John Wiley & Sons.
- Hare PE and Mitterer RM (1967) Nonprotein amino acids in fossil shells. *Carnegie Institution Washington Yearbook* 65: 362–364.
- Hare PE and Mitterer RM (1969) Laboratory simulation of amino acid diagenesis in fossils. *Carnegie Institution Washington Yearbook*. 205–208.
- Johnson BJ and Miller GH (1997) Archaeological applications of amino acid racemization. *Archaeometry* 39: 265–287.
- Kaufman DS and Miller GH (1992) Overview of amino acid geochronology. *Comparative Biochemistry and Physiology* 102B: 199–204.
- Kaufman DS and Miller GH (1995) Isoleucine epimerization and amino acid composition in molecular-weight separations of Pleistocene *Genyornis* eggshell. *Geochimica et Cosmochimica Acta* 59: 2757–2765.
- Kidwell SM, Best MR, and Kaufman DS (2005) Taphonomic tradeoffs in tropical marine death assemblages: Differential time-averaging, shell loss, and probable bias in siliciclastic versus carbonate facies. *Geology* 33: 729–732.
- Kosnik MA, Hua Q, Jacobsen GE, Kaufman DS, and Wüst RA (2007) Sediment mixing and stratigraphic disorder revealed by the age-structure of *Tellina* shells in Great Barrier Reef sediment. *Geology* 35: 811–814.
- Kosnik MA and Kaufman DS (2008) Identifying outliers and assessing the accuracy of amino acid racemization measurements for geochronology: II. Data screening. *Quaternary Geochronology* 3(4): 328–341.
- Kosnik MA, Kaufman DS, and Quan H (2008) Identifying outliers and assessing the accuracy of amino acid racemization measurements for geochronology: I. Age calibration curves. *Quaternary Geochronology* 3(4): 308–327.
- McCoy WD (1987) The precision of amino acid geochronology and paleothermometry. *Quaternary Science Reviews* 6: 43–54.
- Miller GH and Brigham-Grette J (1989) Amino acid geochronology: Resolution and precision in carbonate fossils. *Quaternary International* 1: 111–128.
- Miller GH, Hart CP, Roark EB And, and Johnson BJ (2000) Isoleucine Epimerization in Eggshells of the Flightless Australian Birds, *Genyornis* and *Dromaius*. In: Goodfriend GA, Collins MJ, Fogel ML, Macko SA, and Wehmiller JF (eds.) *Perspectives in Amino Acid and Protein Geochemistry*, pp. 161–181. New York: Oxford University Press.
- Miller GH, Magee JW, and Jull AJT (1997) Low-latitude glacial cooling in the Southern Hemisphere from amino acid racemization in emu eggshells. *Nature* 385: 241–244.
- Murray-Wallace CV and Kimber RWL (1993) Further evidence for apparent 'parabolic' racemization kinetics in Quaternary molluscs. *Australian Journal of Earth Sciences* 40: 313–317.
- Orem CA and Kaufman DS (2011) Effects of basic pH on amino acid racemization and leaching in mollusk shell. *Quaternary Geochronology* 6: 233–245.
- Penkman KEH, Kaufman DS, Maddy D, and Collins MJ (2008) Closed-system behaviour of the intra-crystalline fraction of amino acids in mollusc shells. *Quaternary Geochronology* 3: 2–25.
- Schroeder RA and Bada JL (1976) A review of the geochemical applications of the amino acid racemization reaction. *Earth-Science Reviews* 12: 347–391.
- Smith GG, Williams KM, and Wonnacott DM (1978) Factors affecting the rate of racemization of amino acids and their significance to geochronology. *Organic Chemistry* 43: 1–5.
- Sykes GA, Collins MJ, and Walton DI (1995) The significance of a geochemically isolated intracrystalline organic fraction within biominerals. *Organic Geochemistry* 23: 1059–1065.
- Wehmiller JF (1984) Interlaboratory comparison of amino acid enantiomeric ratios in fossil Pleistocene mollusks. *Quaternary Research* 22: 109–121.
- Wehmiller JF (1990) Amino acid racemization: Applications in chemical taxonomy and chronostratigraphy of Quaternary fossils. In: Carter J (ed.) *Skeletal Biomineralization: Patterns, Processes, and Evolutionary Trends*, pp. 583–608. New York: Van Nostrand Reinhold.
- Wehmiller JF and Miller GH (2000) Aminostratigraphic dating methods in Quaternary geology. In: Noller J, Sowers J, and Lettis W (eds.) *Quaternary Geochronology: Methods and Applications*, pp. 187–222. Washington, DC: American Geophysical Union.
- Williams KM and Smith GG (1977) A critical evaluation of the application of amino acid racemization to geochronology and geothermometry. *Origins of Life* 8: 91–144.

ARCHAEOLOGICAL RECORDS

Contents

Overview

2.7 Myr–300 000 Years Ago in Africa

2.7 Myr–300 000 Years Ago in Asia

1.9 Myr–300 000 Years Ago in Europe

Global Expansion 300 000–8000 Years Ago, Africa

Global Expansion 300 000–8000 Years Ago, Asia

Global Expansion 300 000–8000 Years Ago, Australia

Global Expansion 300 000–8000 Years Ago, Americas

Human Evolution in the Quaternary

Neanderthal Demise

Postglacial Adaptations

Overview

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Introduction

The study of hominids, hominins, and humans ([Table 1](#)) in the Quaternary period is an interdisciplinary endeavour that combines the expertise of the biological, physical, human, and natural sciences ([Figure 1](#)). Scientific advances in this collective enterprise have been most marked in the past 50 years. They can be summarized by the appearance of a new subdiscipline, paleoanthropology, that emerged in the 1970s. The title includes, at a minimum, the following specialists: archaeologists, physical anthropologists, molecular geneticists, geochronologists, and paleoecologists. Paleoanthropologists are not only interested in investigating well-dated sequences that contain environmental and hominin data but also in using these archives to study the behavior of our earliest ancestors. They work in the wider framework of evolutionary models and principles. Their primary interests lie in using the current wealth of paleoenvironmental data to understand changing adaptations in the various hominin lineages, and using the continuous records of climate change, particularly those from the ocean and ice-core archives, to examine whether a forcing mechanism existed that explains

both anatomical and behavioral evolution. More recently, the field has developed an interest in the biogeography of hominins and in particular the timing of major dispersals and their explanation ([Straus and Bar-Yosef, 2001](#)).

The papers in this section address these and other themes in paleoanthropology on a geographical basis. The evidence is presented and assessed and a flavor of the common but distinctive traditions of regional and national research is provided. The date of 300 ka has been used to organize the archaeological evidence. This corresponds to the beginning of MIS8 in the marine record. This cold stage does not in itself mark an abrupt change either in hominin archaeology or anatomy, but after this date significant technical, cultural, and social changes occurred cumulatively, and especially during the Upper Pleistocene (MIS5-2).

The Transition to Paleoanthropology

The development of paleoanthropology within Quaternary science can be illustrated by some selected texts. In 1958,

Frederick Zeuner published the fourth edition of his highly influential synthesis, *Dating the Past: An Introduction to Geochronology* (Zeuner, 1958). A year later, he followed it with *The Pleistocene Period: Its Climate, Chronology and Faunal Successions* (Zeuner, 1959). The two volumes provide a benchmark with which to assess the current work of paleoanthropologists.

Zeuner's syntheses were dominated by the Pleistocene sequences of Europe that had been established in southern Germany 50 years before by Penck and Brückner. In 1959, research had barely begun in Africa, Asia, and Australia, and

the number of reliable sequences was very small. America did not possess a deep Pleistocene record for hominins although the glacial sequences had been established.

Zeuner's discussion of the archaeology was dominated by European typology and taxonomy, particularly in *Dating the Past*. Europe also loomed large in his discussion of the fossil record. There are, for example, very few references to the Australopithecines in either book even though they had been named 30 years before in South Africa. The state of play in 1959 is summarized in Table 2.

However, 1959 was a significant year for paleoanthropology outside of Europe. During their annual field season at Olduvai Gorge, Tanzania, Mary and Louis Leakey discovered a robust Australopithecine (*Paranthropus (Zinjanthropus) boisei*) in their excavations. The development of potassium/argon dating techniques and their application soon after (Leakey et al., 1961) to the volcanic tuffs that segment the long Olduvai sequence achieved a breakthrough that the Europeans could not match. The hominin 'living floor,' as Leakey (1971) referred to it, and where first *Zinjanthropus* and later *Homo habilis* (Leakey et al., 1964) had been found, was now dated in Bed I

Table 1 Three important terms in paleoanthropology. Hominin replaced hominid in the 1990s when the genetic classification of extant apes and humans caused a rethink of their ancestral relations. The use of the term, however, is not yet ubiquitous

Human	All living people and their recent Pleistocene ancestors
Hominin	The above, and all the Pliocene and Pleistocene fossil ancestors (genera and species) of humans
Hominid	The above, and the African great apes: chimpanzee, bonobo, and gorilla, and their fossil ancestors.

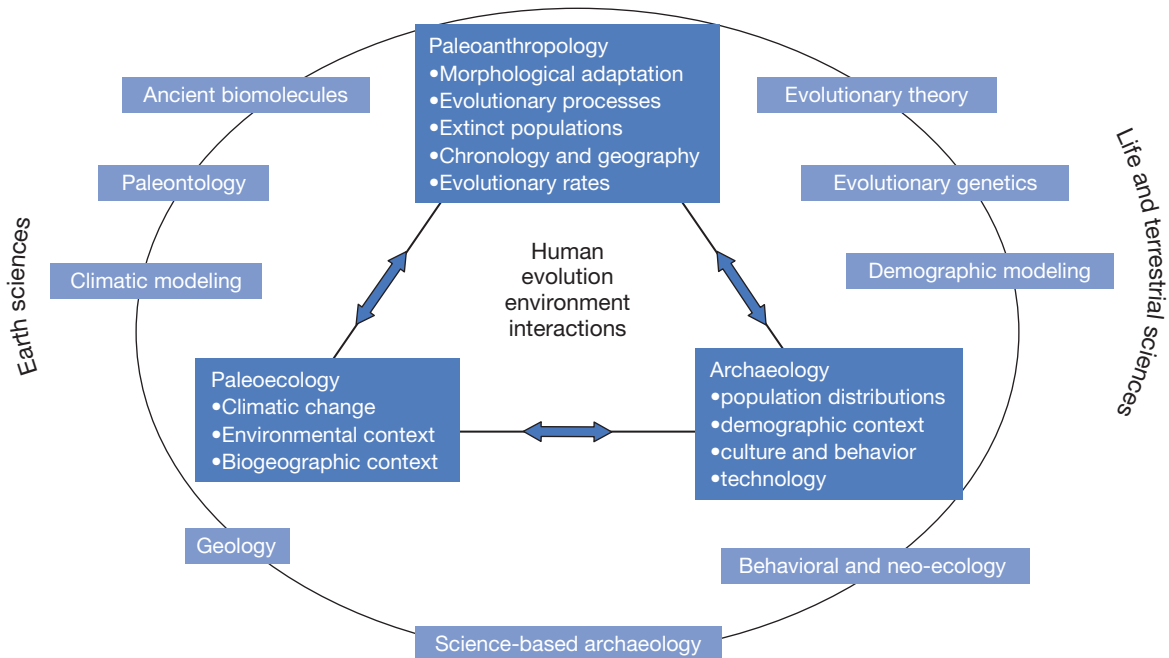


Figure 1 The interdisciplinary relationships involved in paleoanthropology. Foley (nd) with permission. Prepared for the NERC, Environmental Factors and Chronology in Human Evolution and Dispersal programme 2003.

Table 2 The Pleistocene framework for the study of hominins in 1959 (Zeuner 1959: 219). Zeuner was following the four Ice Age model of Penck and Brückner. The table reveals the paucity of fossil finds, mostly from Europe, and the classification of what there was into a very few species. The age estimates were based on a number of sources including the astronomical theory of the Serbian astrophysicist Milutin Milankovitch who died in 1958

Upper Pleistocene	Back to about 180 ka BP	Last glaciation (1) and last interglacial	<i>Homo sapiens</i> <i>Homo neanderthalensis</i>
middle Pleistocene	About 180 to 425 ka BP	Penultimate glaciation (2) and penultimate interglacial	<i>Homo cf. sapiens</i>
Lower Pleistocene	About 425 to 600 ka BP	Antepenultimate glaciation (3), antepenultimate interglacial, and early glaciation (4)	<i>Homo (Pithecanthropus) erectus</i> and <i>pekinensis Homo heidelbergensis</i>

BP = Before Present.

to more than 1.7 Myr old, placing the finds in the Pliocene. Moreover, the association of the fossil hominins with rich faunal remains and stone tools provided an opportunity to conduct a paleoanthropological investigation into the capabilities and adaptive behavior of these remote ancestors. Discussions of these data then ranged around the existence of central place foraging (Isaac, 1978), the contribution of hunting and scavenging to the diet (Binford, 1981; Bunn, 1981), and the importance of using modern analogs to distinguish hominin behavior from that of other accumulators and modifiers of bone and stone such as hyenas and rivers (Behrensmeyer and Hill, 1980; Brain, 1981).

Following this lead similar issues were tackled in Europe. The adaptations of hominins to their environment were studied by means of their stone technology as well as by using evolutionary ecology to analyze the decisions they had taken about where to stay and which parts of the landscape they repeatedly visited. A pioneering project in this emerging paleoanthropology took place in the Ambrona valley, Spain, led by Clark Howell and Emiliano Aguirre. While issues of dating and comparative stone tool typology were uppermost at their excavations of the Lower Paleolithic sites of Torralba and Ambrona (Howell, 1966), their aim was also to turn those bones and stones into information about the process of human evolution. As at Olduvai, the questions dealt with adaptations to the food quest: how did they get their food, what did they eat, and also did they eat well?

The excavations impressed because of their scale (Santonja, 2005) and the association of stone tools with elephant bones. Clark Howell, in his popular book *Early Man* (Howell, 1965), presented this association as evidence for big-game hunting by *Homo erectus* using Acheulean handaxes. The conclusions emphasized their human capacity to plan, cooperate, share, and instruct.

These capacities were summarized by Butzer (1982) who used his geological work to reconstruct the wider landscape to which the hominins were adapting. In his study, the two sites, Torralba and Ambrona, formed a hub in a network of movements that radiated over large distances as hominins adjusted to seasonal fluctuations in their food supply (Figure 2). Hominins would come together to hunt when game was migrating and then disperse in small groups across a much larger territory. An important factor in these seasonal movements was the high elevation (over 1000 m) for the sites suggesting that winters were harsh and that prey species would migrate in search of food. The ability to kill big game and to adjust population to the seasons all indicated modern capabilities for these handaxe makers.

Subsequently, several of these claims have been disputed and revised (Villa, et al., 2005), and the debate about the ability of early hominins either to hunt or to scavenge has seesawed with new and more precise evidence (Dominguez-Rodrigo, 2002; Stiner, 1994). However, whatever the outcome, and currently the hunting hypothesis commands the majority view (Gamble, 1999), these debates illustrate the questions that paleoanthropologists came to ask about Quaternary hominins. At the heart of these questions is the time depth of human rather than hominin behavior in the Quaternary and it is to these issues that the author will devote the rest of this contribution.

Adaptive Radiations and Climate Forcing

While paleoanthropologists work within the framework of evolutionary science, their approach, it has been claimed (Foley (2001), p. 5), is largely untouched by advances in evolutionary theory. New data and new methods in the last 50 years, and in particular the ongoing development of absolute chronologies, has led to the development of more complicated models of hominin ancestry. For example, one of the most significant findings in Bed I at Olduvai Gorge was the presence of two contemporary hominin genera, *Paranthropus* and *Homo*. The wealth of fossil hominins that have since been discovered in the Hadar region of Ethiopia, Koobi Fora in Kenya, and throughout the African Rift Valley, as well as the limestone caves of South Africa (Klein, 1999; Stringer and Andrews, 2005), have confirmed this pattern. At any one moment, there were several species present at a subcontinental scale and often within the same local geographical region. This is well illustrated at Koobi Fora where an African example of *H. erectus* was found in the same geological horizon, dated to 1.3 Ma, as a robust *Australopithecus* (Leakey and Walker, 1976).

These finds changed the basic model of hominin evolution. No longer was there a single species evolving in well-recognized anatomical stages toward modern humans. Instead, at any one time, there were several contenders, like branches on a bush, and the task of the paleoanthropologist was to decide, after a review of all the evidence, which one was indeed the ancestor.

The problem of deciding the path of human ancestry is illustrated by stone tools. For example, which of the two hominin genera at Koobi Fora (see above) had made the stone tools that were excavated from the same geological deposits? Furthermore, the observations of tool making and usage by chimpanzees at Gombe in the 1960s has now been extended to New World monkeys who use stones to dig for roots in Brazil. Such widespread technological competence has demanded a rewrite of the earlier certainties about technology and the evolution of the human brain. This is an ongoing debate currently concerned with the competence of the small-brained hominin *Homo floresiensis* from Indonesia to make the stone tools that have been excavated from Late Glacial levels in the Liang Bua cave (Morwood et al. 2004).

In order to inject a more theoretical perspective into paleoanthropology, Robert Foley (2002, figure 3) has recently proposed seven adaptive radiations in hominid evolution (Table 3; Figure 3). This model not only responds to the empirical evidence of many hominin species but also establishes a behavioral basis to the pattern of change that is evident in both the archaeology and anatomy.

Adaptive radiations, according to Foley, are based on the process of diversification that stems from dispersal, and the adaptive basis of these dispersals, rather than on the process of speciation. In his framework, dispersal is the mark of evolutionary success. It is an indication that speciation has occurred due to a combination of three factors: allopatry (or geographical isolation), local adaptation, and genetic drift.

The six hominin adaptive radiations are each characterized by a large number of species, and we can predict that an even larger number await discovery. The route to modern humans will not therefore be traced in a simple lineal fashion from one fossil species to another. Rather it will be through major

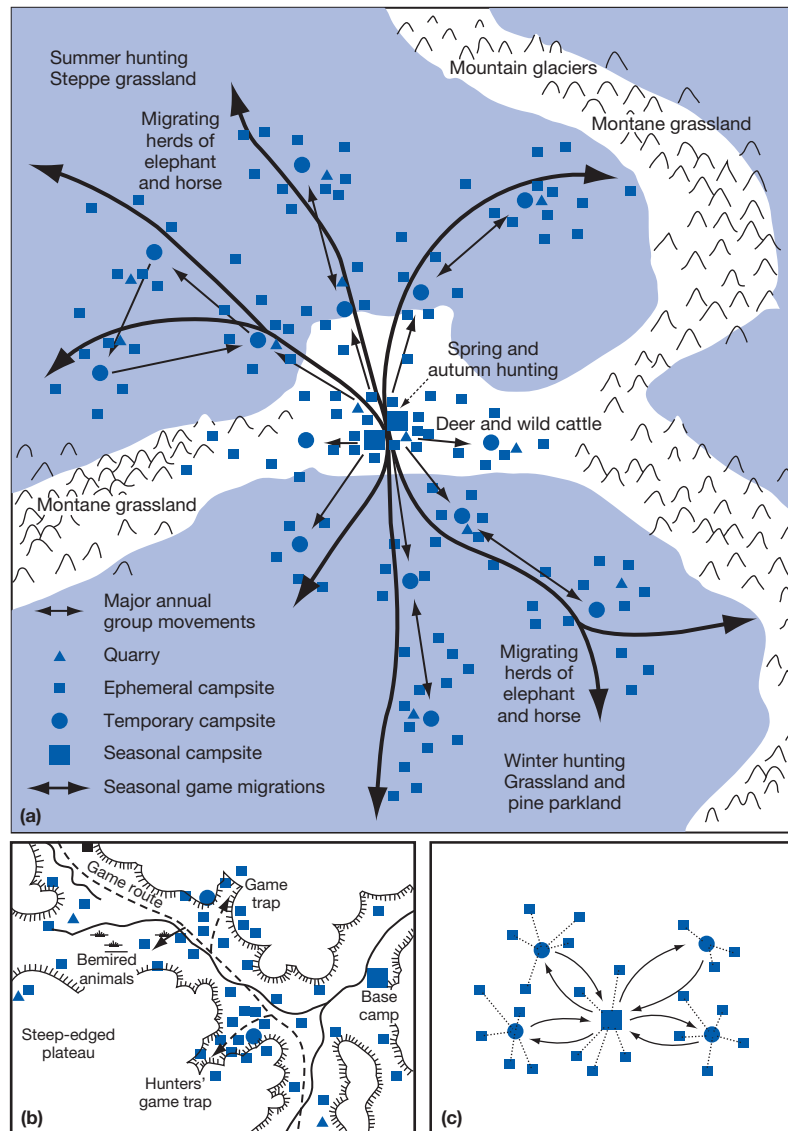


Figure 2 A reconstruction of the movements of Acheulean hunters on the Spanish meseta (after Butzer (1982)). (a) Intercept hunting during the spring and autumn game migrations. (b) The Ambrona valley used as a game trap. (c) Seasonal movement between camps of varying size provides a model of the archaeological evidence.

Table 3 Seven hominid adaptive radiations. The timescale is approximate

Suggested timescale	Adaptive radiation	Adaptive pattern	Grade	Representative taxa
Last 100 ka	7	Aquatic resources	Anatomically modern humans	<i>H. sapiens</i>
500 ka	6	Projectiles ?Fire	Larger-brained <i>Homo</i>	<i>H. neanderthalensis</i> <i>H. heidelbergensis</i>
1 Ma	5	Carnivory	<i>Homo</i>	<i>H. ergaster</i>
2 Ma	4		Earliest <i>Homo</i>	<i>H. rudolfensis</i>
3 Ma	3	Megadonty	<i>Paranthropus</i>	<i>P. boisei</i>
4 Ma	2	Bipedalism	<i>Australopithecus</i>	<i>A. afarensis</i>
5 Ma	1	African apes		

After Foley, R. A. (2002) Adaptive radiations and dispersals in hominin evolutionary ecology. *Evolutionary Anthropology* 11, 32–37.

adaptive changes that involved a suite of regional hominin forms that shared the major adaptive response of that radiation, such as carnivory or a composite weapon technology. The task of the paleoanthropologist, as reviewed by Potts (1998), is therefore to examine the evidence for significant territorial

expansion and longer, sustained occupancy in habitats that experienced variable climatic selection.

One example is the turnover pulse model (Vrba, 1996) that draws an explicit link between the timing of speciation among a range of African mammals, including hominins in the

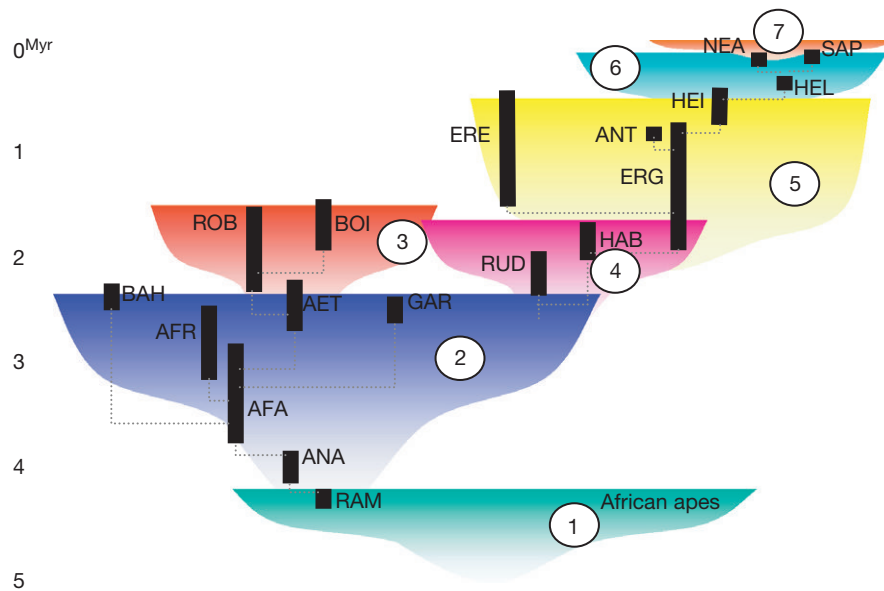


Figure 3 The adaptive radiations of the hominins (after Foley (2002), figure 3). The radiations are described in Table 3. Key – *Ardipithecus*: RAM = *ramidus*. *Australopithecus*: ANA = *anamensis*; AFA = *afarensis*; AFR = *africanus*; BAH = *bahrelghazali*; AET = *aethiopicus*; GAR = *garhi*. *Paranthropus*: ROB = *robustus*; BOI = *boisei*. *Homo*: RUD = *rudolfensis*; HAB = *habilis*; ERG = *ergaster*; ERE = *erectus*; ANT = *antecessor*; HEI = *heidelbergensis*; HEL = *helmei*; NEA = *neanderthalensis*; SAP = *sapiens*.

Plio-Pleistocene, and the possible forcing effect of climate change (deMenocal, 2003). In particular, attention has been drawn to an explosion of new species between 2 and 2.4 Ma that is interpreted by Vrba as the impact of climate change on low-latitude forests and the opening up of new savannah habitats. This change would have favored a radiation of bovid species as wet and dry conditions alternated more rapidly. In Vrba's view, such turnovers are associated with ecological generalists who can adjust their feeding strategies to changing conditions in the environment. Once established, the possibility emerges for a rapid, pulse-like radiation into new niches, so producing a wealth of new taxa. Therefore, as soon as the environment has become 'saturated,' specialist adaptations with a different set of ecological strategies emerge (Gamble, 1993, pp. 87–89).

The argument is persuasive but a statistical examination (Foley, 1994) has shown that species appearance is not causally linked to climatic change as indicated in the marine cores. However, the same statistical study does show that species extinction is more likely to be climatically controlled (Foley (1994), p. 285). Foley's observation is important in establishing the nature of the link between climate change and hominin evolution on the large scale. The implication is that habitats where specialist feeders might normally be found are lost more quickly than they are gained. And if climate change during the Quaternary controls extinction, then we can see that a major biogeographical pattern in hominin evolution has been the repeated contraction and expansion of their geographical ranges. As Dennell (2003, p. 421) has pointed out, the question of 'how successful' these dispersals might have been has received far less attention than the 'when' and 'how often.' It is expected that within radiations 4–6 (Table 3), these contractions were frequent and short lived, especially in the northern areas of the Old World where climate exerted a more controlling effect through the seasonal distribution of food resources.

From Dispersals to Diasporas and the Legacy of the Pleistocene

The first six adaptive radiations (Table 3) all took place within the confines of the Old World (Table 4). Moreover, radiations 2–6, characterized by bipedalism, megadonty (large teeth), carnivory, and projectiles only achieved a restricted distribution within the Old World. For example, Siberia was not settled until the end of the Upper Pleistocene. This pattern helps to explain the late colonization of the Americas. Furthermore, even though stepping-stone islands existed to facilitate dispersal to a large, low-latitude island such as Madagascar, this did not occur during the Pleistocene. It was only adaptive radiation 7, marked by the use of aquatic resources, that saw a global distribution of hominins – a process, as the Australian and American evidence shows, that started no earlier than 60 ka BP and probably more recently in all the three continents.

The legacy of the Quaternary viewed from a hominin perspective is therefore a single species globally distributed (Table 4). The process by which the Old World and subsequently the globe came to be occupied by hominins and humans is now described as a series of first dispersals (Gamble, 1993) and then in adaptive radiation 7, diasporas (LL Cavalli-Sforza and F Cavalli-Sforza, 1995). The length of time it took to colonize the Earth and so realize the Quaternary legacy has been put down to cognitive limitations, a lack of transport technology (especially boats), and environmental barriers (particularly oceans) (Gamble (1993), chapters 2 and 3).

Three examples question these views.

1. The earlier dispersals of 2 Ma BP are characterized as multiple and frequent. Africa is regarded as the source and Asia/Europe as the recipient of population. However, this view has been questioned for the earliest dispersals. Robin

Table 4 Aspects of the two adaptive radiations of *Homo*

Adaptive radiations	1–6	7
Global habitats occupied	25%	100%
Duration	2–2.5 Ma	300–100 ka
Biological diversity	Multiple species, genera	Geographical populations, races
Technological skills	Generic and transferable	Place specific
Social networks	Local	Continental
Release from social proximity	Limited	Achieved
Climate change response	Retreat to refuge	Track from refuge
Representative taxa in <i>Homo</i>	<i>erectus/ergaster/heidelbergensis/helmei/neanderthalensis</i>	<i>sapiens</i>

After Gamble, C. S., Davies, W., Pettitt, P., and Richards, M. (2004). Climate change and evolving human diversity in Europe during the last glacial. *Philosophical Transactions of the Royal Society Biological Sciences* 359, 243–254.

Dennell (Dennell and Roebroeks, 2005) has pointed to evidence from Asia which suggests that one of paleoanthropology's most basic paradigms – dispersal by early *Homo* Out of Africa – needs rethinking. The Sahara was not the barrier to movement as often presented so that an earlier hominid dispersal could have resulted in the origin of hominins, and in particular the genus *Homo*, outside Africa with subsequent dispersal back 'into' that continent. This scenario is by no means proven but serves as an alternative hypothesis to the Out of Africa 1 model for the origin of the genus *Homo*. Consequently, Dennell and Roebroeks (2005, p. 1101) argue that what paleoanthropologists need are spatial units other than modern political boundaries in order to investigate such dispersals and their internal movements. They propose 'Savannahstan' as one such unit to model adaptations and dispersals since it links the Pliocene grasslands from west Africa to northern China.

2. The multiple dispersals of hominins is therefore a parallel to the multiple species that are now expected, apart from the present day, at any one time in the fossil record. The implications of such a pattern have been discussed by Marta Lahr and Robert Foley (Foley and Lahr, 1997) in the context of the origins of modern humans. Rather than a single diaspora of *H. sapiens*, the Out of Africa 2 model, they argue for repeated dispersals after 300 ka BP and that these included the much later diasporas into Australia and the Americas. The multiple dispersal model combines genetic, anatomical, and archaeological data.
3. The diasporas of the last 60 ka can be studied in greater detail as a result of enhanced preservation, greater chronological precision, and the use of molecular genetics to produce phylogeographies. The articles in this Encyclopedia on Australia and the Americas indicate the fine-grained control that is possible while a combination of genetic, archaeological, and radiocarbon evidence is allowing cross-checking on issues of timing and direction in the recolonization of formerly glaciated and periglacial areas of northern Europe (Gamble et al., 2004; Straus et al., 1996). Comparison of the radiocarbon dating of these Late Glacial diaspora and settlement patterns with the GRIP ice-core chronology of Greenland (Figure 4) reveals that another of paleoanthropology's basic paradigms – that climate warming encourages recolonization – is not supported, since populations left the southern glacial refuges in GS-2a, a time of extreme

cold. Following a rise in population numbers during GI-1 (Bocquet-Appel and Demars, 2000), the return to cold conditions in GS-1, also known as the Younger Dryas, did lead to a reduction in settlement numbers confirming, as we saw above, that Quaternary climates controlled extinction (population loss) rather than speciation (population gain) (Gamble et al., 2004).

Social Rather than Environmental Change

The move from dispersals to diaspora is a further example of the continuing debate among paleoanthropologists of where to draw the hominin to human line. Cognitive, technological, and environmental reasons do not seem sufficient to account for the limited Old World distribution of all hominins prior to the late diasporas of modern humans. The novelty of these later diasporas, as opposed to the earlier dispersals, was the release from social proximity which they achieved. This release characterizes a human as compared to a primate community (Rodseth et al., 1991). In other words, human social life was extended to include participants who were not always living together and who might be physically absent from their immediate social networks for lengthy periods (Gamble, 1999). This pattern of social extension is a hallmark of humans that does not appear to have been shared in such a dramatic way with other hominins. The evidence that their social lives were local (Table 4) comes from the limited distances that raw materials were transported (Féblot-Augustins, 1997). These were usually less than 5 km and rarely more than 80 km from source (Gamble (1999), Table 3.4). By comparison, diasporic humans exchanged shells and stone sometimes over distances of 1000 km and regularly over 100 km. This extension in the transfer and exchange of raw materials is supported by the ocean crossing to Australia and the rapid peopling of its hyperarid Pleistocene interior (Smith, 2005), that involved very low population densities and physical separation. But such colonization did not lead to a breakdown in social life as must have been the case when earlier hominins faced such socially challenging environments. With the final, aquatic (7) radiation, the emphasis now shifted to the construction of society over longer timescales and distances. This release from proximity was achieved by connecting people and places in an extended social network.

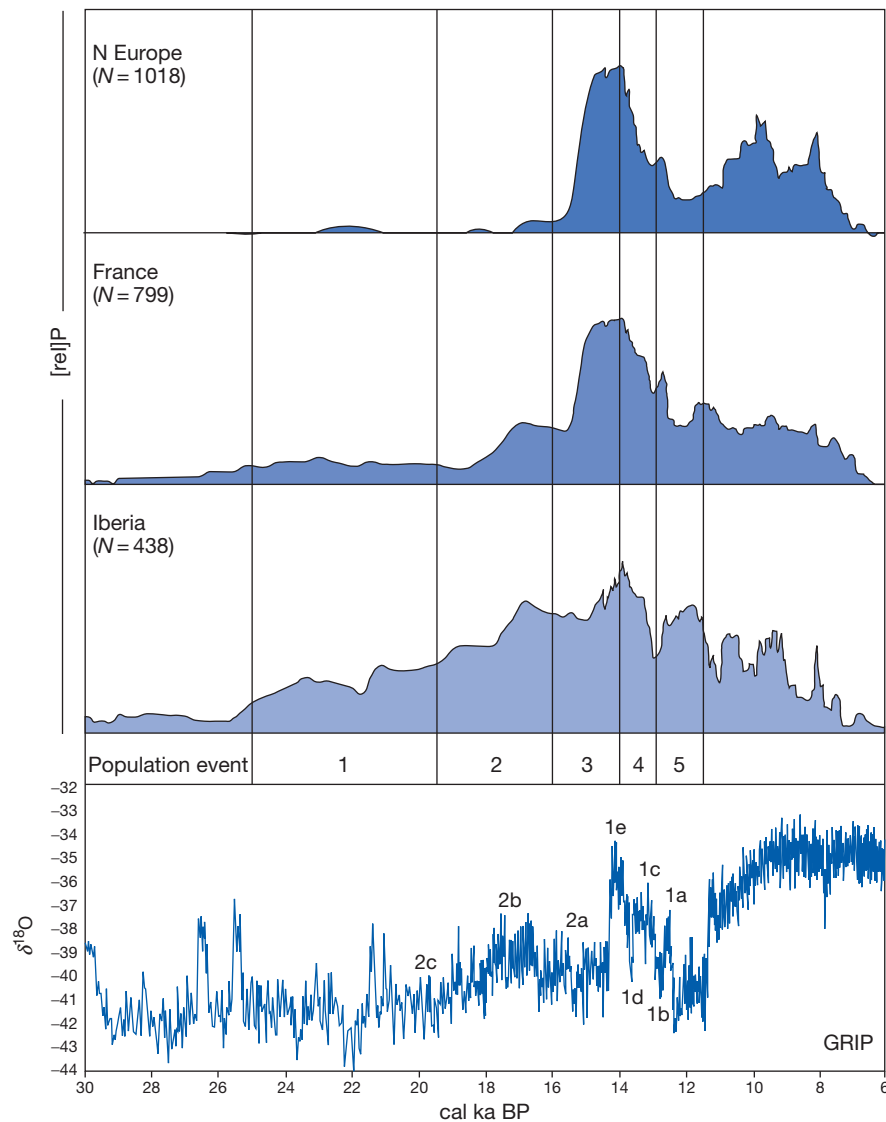


Figure 4 The pattern of recolonization of northern Europe from a southern refuge during the Last Glacial Maximum. The frequency curves show the distribution of radiocarbon dates and their increase points to the timing of population dispersal (after Gamble et al. (2004), figure 2). The radiocarbon determinations are calibrated using CALPAL (Weninger and Jöris, 2000) and presented by three major regions. Five population events associated with the dispersal are indicated. The distribution of radiocarbon dates points to the major dispersal, population event 3, preceding the warming of the Greenland interstadial. *N* = number of calibrated radiocarbon determinations in the frequency curve. Stratotype boundaries in GRIP ice-core years are as follows: GI-2 = 21.8–21.2; GS-2 = 21.2–14.7; GI-1 = 14.7–12.7; GS-1 = 12.7–11.5 ka BP.

Technological Modes and the Evolution of a Social Brain

Stone tools have been the main source for studying the significance of encephalization in hominin evolution (Schick and Toth, 1993; Wynn, 1993). Changes in Paleolithic traditions (Table 5) were once seen as a demonstration of progressive evolution. Each stage in the Lower–Middle–Upper (Zeuner, 1958) sequences was linked to a fossil hominin. However, as we have seen above, such simple correlations (Table 2) are no longer possible and the criteria which define such classification as the African Earlier–Middle–Later Stone Age are not as hard and fast as once thought. For example, the advent of radiometric

dating and the investigation of long stratified sequences, as at Klasies River Mouth, South Africa (Klein, 1999), have shown that stone blades, once considered a diagnostic type fossil of the European Upper Paleolithic and the African Later Stone Age, have a much greater ancestry. Blades are any intentionally struck stone blank that is at least twice as long as it wide. With such a definition, it is not surprising that blades frequently appear as elements in Middle Paleolithic and Middle Stone Age assemblages and the age of their appearance has been extended from some 40 ka BP to over 300 ka BP (Bar-Yosef and Kuhn, 1999).

An alternative scheme, proposed by Clark (1969), recognized five technological modes (Table 5). His approach has been widely followed as studies of Quaternary hominins and

Table 5 Technological modes in the Paleolithic

	<i>Dominant lithic technologies</i>	<i>Conventional divisions in Europe</i>	<i>Conventional divisions in Africa</i>
Mode 5	Micro lithic components of composite artifacts	Mesolithic	Later Stone Age
Mode 4	Punch-struck blades with steep retouch	Upper Paleolithic	
Mode 3	Flake tools from prepared cores	Middle Paleolithic	Middle Stone Age
Mode 2	Bifacially flaked handaxes	Lower Paleolithic	Earlier Stone Age
Mode 1	Chopper tools and flakes		

Clark offered the modes as a homotaxial sequence which at least in Africa and, at the time he was writing, in Eurasia followed the progression from mode 1 to mode 5. However, although he was writing a world prehistory, Clark never claimed his modes were universal stages (1969, p. 30). Instead, they were his attempt to escape from the historical thorn bushes of what to call prehistoric hunters and gatherers in different continents. For example, they certainly were neither Paleolithic nor Stone Age in Australia or the Americas. Reproduced from Clark JGD (1969) *World Prehistory a New Outline*, 2nd edn., p. 31. Cambridge: Cambridge University Press.

their technology became a truly worldwide endeavour and where the European Paleolithic and African Stone Age terms appeared anachronistic. However, one problem with Clark's modes is that they unduly privilege stone tools, and all other aspects of technology and material culture are forgotten. These would include wooden spears, hearths, huts, art, and items of display, all of which have a varied and complex history.

Clark's five technological modes do however highlight two aspects of hominin technology during the Quaternary. First, it was very static, and second, it did not follow the ethnographic pattern found among today's hunters and gatherers, where the local environment exerts a selection pressure on the shape, form, and complexity of technology. For example, mode 2, typified by large bifaces such as Acheulean handaxes, is first found in east Africa at 1.5 Myr BP and persists for more than a million years (Klein, 1999). Moreover, these bifaces are found from the equator to the northern latitudes of Britain and in environments with very different plant and animal resources, shorter growing seasons, and a range of adaptive challenges. As a result, the technological skills of the early hominins can be characterized as generic (Table 4). They could be transferred to environments as different as the tropical savannahs and the northern grasslands. A very different picture is presented by mode 4 technologies that, after the 60 ka diaspora, compare well in their diversity and complexity with those of contemporary hunters and gatherers and where the skills are specific to local conditions and places. However, the picture is further complicated by the evidence from Australia where mode 3 continued to the present day. Technology does not therefore correlate with particular hominin fossils and their assumed levels of intelligence and problem solving. In addition, the role of environmental selection on innovation and adoption of new techniques is restricted.

Technology, especially the appearance of art and ornament, has been used as a guide to major cognitive and behavioral changes, including language and symbolic thought (Mithen, 1996). However, an examination of hominin encephalization (Figure 5) by Aiello and Dunbar (1993) revealed that the most significant increase in brain size occurred half a million years ago by which time hominin brains were more than twice the size, proportionately, than any of the earlier Australopithecines. Using a comparative sample of living primates, they also showed how early hominin brain sizes implied much larger social groupings. They calculated that group size had risen at this time to at least 120, whereas the largest among primates is

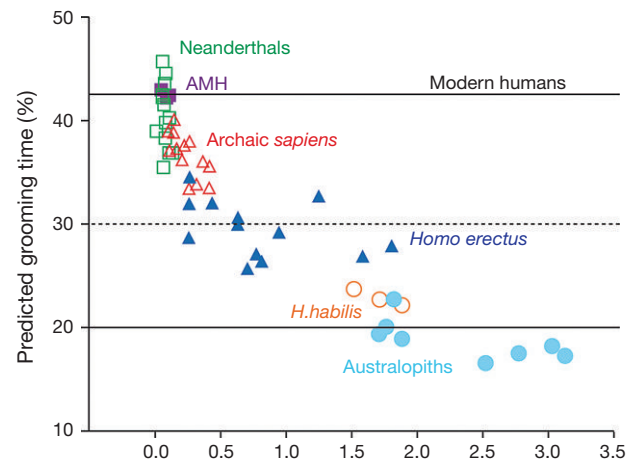


Figure 5 The implications of increases in brain size during 3.5 Ma of hominin evolution (after Dunbar (2003)). At 500 ka, there is a marked encephalization. According to the social brain hypothesis, this resulted in an increase in group sizes and the amount of time that would have been needed daily to undertake physical grooming to maintain social ties. The 20% (lower solid line) of the day spent in social activities such as grooming represents a significant limit for today's great apes since they have to spend much of the day traveling and looking for food. This threshold would have been similar for the small-brained Australopithecines. The 30% of daily activity spent on grooming (dotted line) represents an upper limit beyond which other mechanisms of social integration such as language are needed. The primate-style social model was therefore exceeded at least 500 ka. AMH = anatomically modern human.

60. Dunbar (2003) has since expanded this observation into a social brain hypothesis: our social lives drove encephalization. In explanation, bigger brains have proportionately larger amounts of neocortex, the thinking part of the brain, that also manages relationships and interactions between group members. By 500 ka BP, hominin group size was now so much larger than those of primates that the traditional means of maintaining and organizing groups through physical grooming was no longer an option. Dunbar's suggestion is that language appeared at that time to facilitate interaction and organize larger social groupings. However, such an age is well before the fluorescence of art and ornament that appears in Europe some 40 ka. The social brain hypothesis focuses attention on how exactly technology, and material culture more generally, contributed to those tipping points in hominin evolution such as language.

The Human Revolution: Gradual or Abrupt?

My final theme, the human revolution, became one of the central concerns of paleoanthropologists in the 1980s (Mellars and Stringer, 1989). This was not the first time that the idea of a human revolution had been applied to the fossil and archaeological record. What marked this application out was the claim for recency. As Paul Mellars has shown (Mellars, 2005), the conjunction in Europe of innovation in technology and material culture coincides with the replacement of the native Neanderthals by incoming modern humans *H. sapiens* (Stringer and Gamble, 1993). The model of replacement was initially strongly contested by an ancient origin for modern humans, often described as multiregional evolution (Wolpoff, 1989). Its supporters argue that through a process of convergent evolution the founding populations of Out of Africa 1 in the late Pliocene evolved into the distinctive geographical populations, or races, that are encountered today. The model of a recent African origin for *H. sapiens* is also known as Out of Africa 2. The evidence now available from dating both the anatomical (McDougall et al., 2005) and genetic changes (Cann et al., 1987; Krings, 1997) indicates conclusively that modern humans originated in Africa.

The current interest in the human revolution model centers on its timing. The Upper Paleolithic revolution of Europe has been questioned by paleoanthropologists working in Africa (Wadley, 2001) and characterized as 'the revolution that wasn't' (McBrearty and Brooks, 2000). Excavations at Blombos Cave, South Africa, provide evidence that shell ornaments, engraved ochres, bone tools, and culturally specific projectile points were all present in excess of 80 ka BP (Henshilwood and Marean, 2003). Confirmation comes from Israel, based on the deliberate burial of modern people with items that can be interpreted as intentional grave goods more than 100 ka BP (Grün et al., 2005).

The issue that the human revolution therefore addresses is explaining why such a long delay occurred between the appearance of anatomically modern people in Africa and the suite of cultural changes that led eventually to the Australian and American diasporas. The African record is indeed more complex than once thought and of much greater antiquity. The claims (d'Errico et al., 1998) are also not convincing that Europe's Neanderthals produced their own independent cultural revolution that just happened to coincide with the arrival of modern humans from Africa (Mellars, 2005). As a result, attention has focused on the effect of major environmental events such as the eruption at 71 ka BP of Toba on Sumatra (Ambrose, 1998) and its impact through a prolonged volcanic winter on human populations that favored a subsequent African diaspora.

Revolution, with its accent on abrupt, dramatic change, is not an appropriate term at Quaternary timescales. Neither is the alternative tenable, namely, a gradual addition of innovation in culture and society. Much of the argument revolves around the identification of what constitutes art, symbolic behavior, language, and hence the modern mind (Mithen, 1996) among archaeological data. One person's evidence for art is another's scratched pebble (Marshack, 1972). These are difficult issues not helped by small samples, widely distributed through the Old World.

However, what is apparent is that the major African diaspora that led to modern humans peopling Australia, Europe, and later the Americas depended upon the social ability to extend society across time as well as space. This occurred with, as was the case in Europe, or without, as in America, either much art and ornament or mode 4 technologies, as the Australian record shows. Nobody disputes that these diasporas involved modern humans as defined genetically, anatomically, and socially. This cultural diversity suggests that paleoanthropology can no longer rely on universal processes to construct an understanding of the timing and sequence of changes among hominins. While the evidence from the paleoecological archives is vital to such discussions, the trend toward the detailed history of social and cultural development in selected regions is returning to favor (Straus and Bar-Yosef, 2001).

Future Prospects

These historical accounts will be assisted by a number of recent breakthroughs in Quaternary science, promising an equally productive and exciting next 50 years. Among many, three with the potential to transform the subject can be singled out:

1. adding to the continued precision of dating through such techniques as microtephras permitting independent, long-distance correlation that bring the study of contemporaneity down to below a single human generation within any single MIS;
2. tracing the phylogenies of fossils by analyzing their ancient DNA; and
3. using the direct, measures contained in stable isotopes to characterize hominin diet, and as a proxy for the provenance, and dispersal of hominins.

All of these, and many more techniques, will allow paleoanthropologists to use archaeological evidence in new ways. They will be able to analyze the material culture of our earliest ancestors as historical information in the unfolding story of human evolution during the Quaternary, and provide an examination of how deep-seated, or superficial, our current differences might be.

See also: [Archaeological Records: 1.9 Myr–300 000 Years Ago in Europe; 2.7 Myr–300 000 Years Ago in Africa; 2.7 Myr–300 000 Years Ago in Asia; Global Expansion 300 000–8000 Years Ago, Africa; Global Expansion 300 000–8000 Years Ago, Americas; Global Expansion 300 000–8000 Years Ago, Asia; Global Expansion 300 000–8000 Years Ago, Australia; Neanderthal Demise; Postglacial Adaptations.](#) [Vertebrate Studies: Interactions with Hominids.](#)

References

- Aiello L and Dunbar R (1993) Neocortex size, group size and the evolution of language. *Current Anthropology* 34: 184–193.
- Ambrose SH (1998) Late Pleistocene human population bottlenecks, volcanic winter, and differentiation of modern humans. *Journal of Human Evolution* 34: 623–651.
- Bar-Yosef O and Kuhn S (1999) The big deal about blades: Laminar technologies and human evolution. *American Anthropologist* 101: 322–338.

- Behrensmeier AK and Hill AP (1980) *Fossils in the Making*. Chicago: University of Chicago Press.
- Binford LR (1981) *Bones: Ancient Men and Modern Myths*. New York: Academic Press.
- Bocquet-Appel J-P and Demars P-Y (2000) Population kinetics in the Upper Palaeolithic in western Europe. *Journal of Archaeological Science* 27: 551–570.
- Brain CK (1981) *The Hunters or the Hunted?* Chicago: Chicago University Press.
- Bunn HT (1981) Archaeological evidence for meat-eating by Plio-Pleistocene hominids from Koobi Fora and Olduvai Gorge. *Nature* 291: 574–577.
- Butzer KW (1982) *Archaeology as Human Ecology*. Cambridge: Cambridge University Press.
- Cann R, Stoneking M, and Wilson A (1987) Mitochondrial DNA and human evolution. *Nature* 325: 31–36.
- Cavalli-Sforza LL and Cavalli-Sforza F (1995) *The Great Human Diasporas: The History of Diversity and Evolution*. Reading, MA: Perseus Books.
- Clark JGD (1969) *World Prehistory a New Outline*, 2nd edn. Cambridge: Cambridge University Press.
- d'Errico F, Zilhao J, Julien M, Baffier D, and Pelegrin J (1998) Neanderthal acculturation in Western Europe? A critical review of the evidence and its interpretation. *Current Anthropology* 39 (supplement 1): S1–S44.
- deMenocal PB (2003) African climate and faunal evolution during the Pliocene–Pleistocene. *Earth and Planetary Science Letters* 220: 3–4.
- Dennell R (2003) Dispersal and colonisation, long and short chronologies: How continuous is the Early Pleistocene record for hominids outside East Africa. *Journal of human evolution* 45: 421–440.
- Dennell R and Roebroeks W (2005) An Asian perspective on early human dispersal from Africa. *Nature* 438: 1099–1104.
- Dominguez-Rodrigo M (2002) Hunting and scavenging by early humans: The state of the debate. *Journal of World Prehistory* 16: 1–54.
- Dunbar RIM (2003) The social brain: Mind, language, and society in evolutionary perspective. *Annual Review of Anthropology* 32: 163–181.
- Féblot-Augustins J (1997) *La circulation des Matières Premières au Paléolithique*, vol. 75. Liège: ERAUL.
- Foley R and Lahr MM (1997) Mode 3 technologies and the evolution of modern humans. *Cambridge Archaeological Journal* 7: 3–36.
- Foley RA (1994) Speciation, extinction and climatic change in hominid evolution. *Journal of Human Evolution* 26: 275–289.
- Foley RA (2001) In the shadow of the modern synthesis? Alternative perspectives on the last fifty years of palaeoanthropology. *Evolutionary Anthropology* 10: 5–15.
- Foley RA (2002) Adaptive radiations and dispersals in hominin evolutionary ecology. *Evolutionary Anthropology* 11: 32–37.
- Gamble CS (1993) *Timewalkers: The Prehistory of Global Colonization*. Cambridge, MA: Harvard University Press.
- Gamble CS (1999) *The Palaeolithic societies of Europe*. Cambridge: Cambridge University Press.
- Gamble CS, Davies W, Pettitt P, and Richards M (2004) Climate change and evolving human diversity in Europe during the last glacial. *Philosophical Transactions of the Royal Society Biological Sciences* 359: 243–254.
- Grün R, Stringer C, McDermott F, et al. (2005) U-series and ESR analyses of bones and teeth relating to the human burials from Skhul. *Journal of Human Evolution* 49: 316–334.
- Henshilwood CS and Marean CW (2003) The origin of modern human behaviour: Critique of the models and their test implications. *Current Anthropology* 44: 627–651.
- Howell FC (1965) *Early man*. London: Time Life Books.
- Howell FC (1966) Observations of the earlier phases of the European Lower Palaeolithic. *American Anthropologist* 68: 88–201.
- Isaac G (1978) The food sharing behaviour of proto-human hominids. *Scientific American* 238: 90–108.
- Klein RG (1999) *The Human Career: Human Biological and Cultural Origins*, 2nd edn. Chicago: University of Chicago Press.
- Krings M (1997) Neanderthal DNA sequences and the origins of modern humans. *Cell* 90: 19–30.
- Leakey LSB, Evernden JF, and Curtis GH (1961) Age of Bed I, Olduvai Gorge, Tanganyika. *Nature* 191: 478–479.
- Leakey LSB, Tobias PV, and Napier JR (1964) A new species of the genus *Homo* from Olduvai Gorge. *Nature* 202: 308–312.
- Leakey MD (1971) *Olduvai Gorge: Excavations in Beds I and II 1960–1963*. Cambridge: Cambridge University Press.
- Leakey REF and Walker A (1976) *Australopithecus, Homo erectus and the single species hypothesis*. *Nature* 261: 572–574.
- Marshall A (1972) *The Roots of Civilization*. New York: McGraw-Hill.
- McBrearty S and Brooks AS (2000) The revolution that wasn't: A new interpretation of the origin of modern humans. *Journal of Human Evolution* 39: 453–563.
- McDougall I, Brown FH, and Fleagle JG (2005) Stratigraphic placement and age of modern humans from Kibish, Ethiopia. *Nature* 433: 733–736.
- Mellars P (2005) The impossible coincidence: A single-species model for the origins of modern human behaviour in Europe. *Evolutionary Anthropology* 14: 12–27.
- Mellars PA and Stringer C (eds.) (1989) *The Human Revolution: Behavioural and Biological Perspectives on the Origins of Modern Humans*. Edinburgh: Edinburgh University Press.
- Mithen S (1996) *The Prehistory of the Mind*. London: Thames and Hudson.
- Morwood MJ, Soejono RP, Roberts RG, et al. (2004) Archaeology and age of a new hominid from Flores in eastern Indonesia. *Nature* 431: 1087–1091.
- Potts R (1998) Environmental hypotheses of hominid evolution. *Yearbook of Physical Anthropology* 41: 93–136.
- Rodseth L, Wrangham RW, Harrigan A, and Smuts BB (1991) The human community as a primate society. *Current Anthropology* 32: 221–254.
- Santonja M (ed.) (2005) *Esperando el diluvio: Ambrona y Torralba hace 400,000 años*. Madrid: Museo Arqueológico Regional.
- Schick KD and Toth N (1993) *Making Silent Stones Speak: Human Evolution and the Dawn of Technology*. New York: Simon and Schuster.
- Smith MA (2005) Moving into the southern deserts: An archaeology of dispersal and colonisation. In: Smith MA and Hesse P (eds.) *23° S: Archaeology and Environmental History of the Southern Deserts*, pp. 92–107. Canberra: National Museum of Australia Press.
- Stiner MC (1994) *Honor Among Thieves: A Zooarchaeological Study of Neanderthal Ecology*. New Jersey: Princeton University Press.
- Straus LG and Bar-Yosef O (eds.) (2001) Out of Africa in the Pleistocene. *Quaternary International* 75.
- Straus LG, Eriksen BV, Erlandson JM, and Yesner DR (eds.) (1996) *Humans at the End of the Ice Age: The Archaeology of the Pleistocene–Holocene Transition*. New York: Plenum.
- Stringer C and Andrews P (2005) *The Complete World of Human Evolution*. London: Thames and Hudson.
- Stringer C and Gamble C (1993) *In Search of the Neanderthals: Solving the Puzzle of Human Origins*. London: Thames and Hudson.
- Villa P, Soto E, Santonja M, et al. (2005) New data from Ambrona: Closing the hunting debate. *Quaternary International* 126–128: 223–250.
- Vrba ES (1996) Climate, heterochrony, and human evolution. *Journal of Anthropological Research* 52: 1–28.
- Wadley L (2001) What is cultural modernity? A general view and a South African perspective from Rose Cottage Cave. *Cambridge Archaeological Journal* 11: 201–221.
- Weninger B and Jöris O (2000) "CALPAL", 1.10.00 edition. Köln: Radiocarbon Laboratory Institut Für Urgeschichte Und Frühgeschichte Weyertal 125: D-50923. <http://www.calpal.de>.
- Wolpoff MA (1989) Multiregional evolution: The fossil alternative to Eden. In: Mellars P and Stringer C (eds.) *The Human Revolution: Behavioural and Biological Perspectives on the Origins of Modern Humans*, pp. 62–109. Edinburgh: Edinburgh University Press.
- Wynn T (1993) Two developments in the mind of early *Homo*. *Journal of Anthropological Archaeology* 12: 299–322.
- Zeuner FE (1958) *Dating the Past: An Introduction to Geochronology*, 4th edn. London: Methuen.
- Zeuner FE (1959) *The Pleistocene Period: Its Climate, Chronology and Faunal Successions*. London: Hutchinson.

2.7 Myr–300 000 Years Ago in Africa

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Introduction

The continent of Africa preserves the longest record of archaeological manifestations of hominin behavior beginning with the discoveries in the Kada Gona region of Ethiopia (Semaw et al., 1997). These artifacts, dating to 2.5 Ma, are Oldowan core and flake industries with associated hominin-modified bone. This overview focuses entirely on Africa during the long formative period of human prehistory known as the Early Stone Age (ESA, Oldowan and Acheulean Industries) spanning from about 2.5 Ma to 200–300 ka. All of the assemblages discussed in this section predate technological innovations such as point production, hafting, blade production, and core-reduction methods such as the Levallois technique. These later techniques characterize the Middle and Later Stone Age and will not be part of this review.

The ESA of Africa is divisible into the Oldowan Industry (Industrial Complex) or the Acheulean Industry (Industrial Complex). The Oldowan tradition of craftsmanship is characterized by simple fist-sized cores and whole flakes produced by hand-to-hand hard hammer percussion flaking, where the cobble or nodule from which flakes are to be removed is held in one hand and a dynamic blow is delivered with a stone hammer from the opposing hand. Instances of the bipolar technique, where the cobble is placed on a hard surface or anvil and a flake or detached piece is removed by initiating a wedging fracture with a hard hammer, also exist at some sites. Fixed design core forms are usually lacking and there is minimal secondary retouch on whole flakes. More formalized tools that appear in some industries appear to be the result of raw material shapes and the properties of these stones. Assemblages attributed to the Oldowan Industry *sensu stricto* range in age from 2.5 to 1.8/1.7 Ma. The Oldowan was first described by Mary Leakey from assemblages found at Olduvai Bed I (Leakey, 1971). These industries fall into the mode I technique described by Clark (1969). The transition to the beginnings of the Acheulean Industry dates to between 1.7 and 1.5 Myr and is marked by highly variable assemblages that contain examples of larger cores, larger flakes, rare instances of crude bifaces as well as the previous suite of artifacts usually associated with the Oldowan Industrial Complex. Industries in this time range have collectively been called the Developed Oldowan (see Figure 1).

While there is no total agreement on the actual definition of this term, technological studies of these assemblages have referred to these industries as early Acheulean (Ludwig and Harris, 1998).

The earliest well-documented and unequivocally Acheulean assemblages are found at Konso-Gardula (Ethiopia), FxJj 63 (Kenya), and EF-HR (Olduvai, Tanzania) dating to approximately 1.5 Ma. These assemblages are characterized by large bifacially flaked hand-axes and cleavers made on much larger flakes than is evident in the Oldowan. These flakes are struck

from heavier and larger cores. Smaller but nonetheless significant retouched flake tool forms constitute a conspicuous part of the Acheulean stone tool assemblages. Through the course of the Acheulean Industry, there appears to be a general trend toward thinner cores produced through techniques of bifacial thinning. Bifacial thinning is often conducted using a soft hammer, usually a wood or bone billet. These assemblages are often associated with the onset of the Levallois core technique of core preparation. Sites from the Acheulean industry are found in a much broader geographical range than the previous Developed Oldowan sites (see Figure 2). The Acheulean Industry is succeeded in the geological record by stone assemblages attributed to the Middle Stone Age (MSA) between 200 and 300 ka with the earliest MSA, dating to around 280 ka in the Kapthurin Formation, Kenya (Tryon and McBrearty, 2002). The Acheulean would fall into the mode II of the schema proposed by Clark (1969).

One of the most important events in human evolution was the dietary shift toward meat-eating that, for the first time, put our earliest stone tool-using ancestors in competition with large carnivores that remained dangerous predators. In previous ESA studies, some interpretations of hominin subsistence strategies have relied heavily upon the stratigraphic association of fragmentary animal bones and stone artifacts and overlooked the range of taphonomic processes that may have contributed to the accumulations of artifacts and bones excavated from archaeological sites. Archaeologists have since realized that agents other than hominins, including carnivores, disease, droughts, and water flow, may contribute to the formation of an archaeological site. As a result, identifying the relative contribution of hominins to the accumulation of archaeofaunas has become the focus of zooarchaeological investigations, requiring a theoretically grounded body of research directed at identifying the traces left by various taphonomic agencies (Binford, 1981). Bone-surface modifications including stone tool cut marks, hammerstone percussion marks, and carnivore tooth marks have proved to be the most reliable indicators of hominin and carnivore involvement in the accumulation of bone assemblages and since the first descriptions of cut marks by Bunn (1981) archaeologists are no longer dependent on the association of artifacts and bones to infer hominin involvement in site formation. Therefore, two types of bone assemblages will be discussed here, those that exhibit evidence of hominin-induced bone-surface modifications and those that do not.

Oldowan

The discovery of artifacts in the Kada Gona in sediments that are securely dated to 2.5 Myr sets the base of the archaeological record (Semaw et al., 1997). It is very likely that the earliest tool use has not been found in the record because sites are too

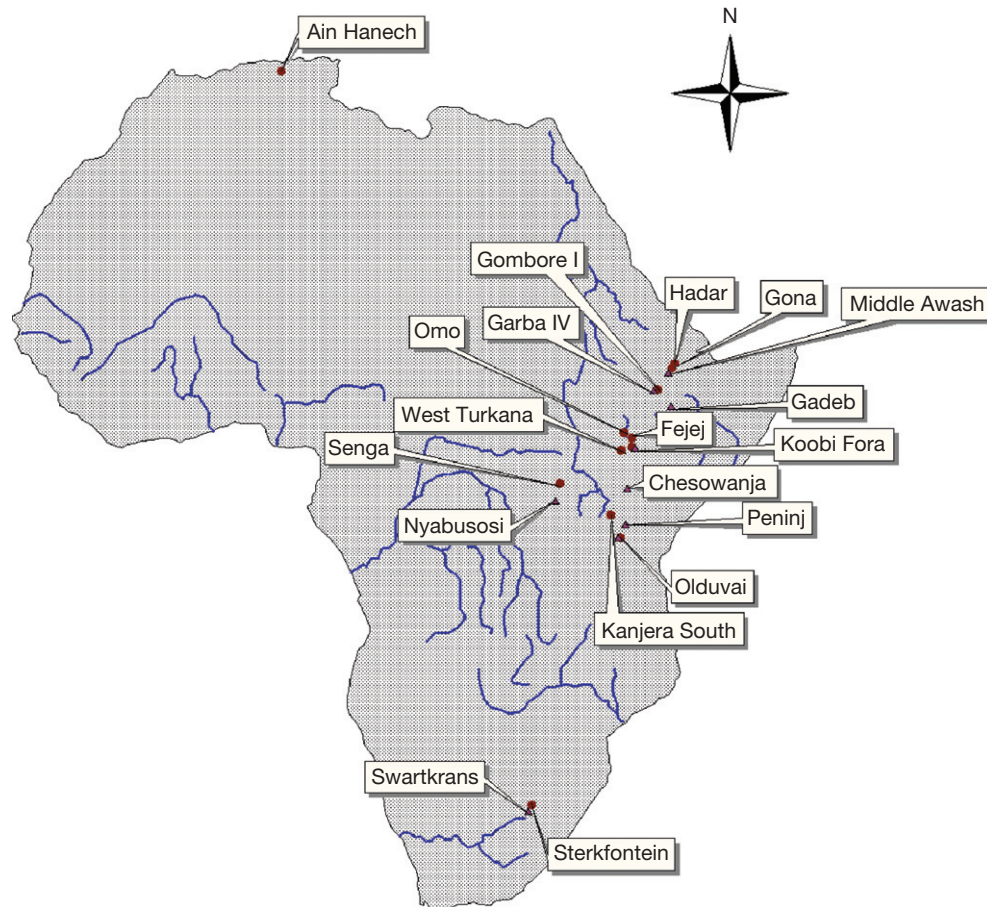


Figure 1 Site localities attributed to the Oldowan Industrial Complex on the African Continent. These include sites that have been collectively called Developed Oldowan and Oldowan.

diffuse or scattered (Panger et al., 2002). However, current archaeological data support the evidence of 2.6 Myr as the onset of technological evolution. Oldowan assemblages that represent the technological system of the earliest tool makers are confined to time periods between 2.6 and 1.5 Myr (see Figure 3).

The majority of these sites are confined to the eastern rift valley of Ethiopia, Kenya, and Tanzania, although sites can be found as far north as Algeria (Ain Hanech) and as far south as South Africa (Sterkfontein) (see Figure 1). Further fieldwork in northeast (e.g., Eritrea) and Northwest Africa (e.g., Morocco) could begin to fill in the current gaps in the geographical range of Oldowan industries. A notable exception to the Africa-centered distribution of Oldowan sites is the remarkable discovery of Oldowan-like artifacts in the Republic of Georgia at the site of Dmanisi (Gabunia et al., 2001). New research in eastern and central Asia suggests hominins began encroaching on these areas in the early Pleistocene.

The classic African Oldowan assemblages are found in sediments that are dated to between 2.6 and 1.8 Myr. The earliest sites are found in the lower Awash River Valley of Ethiopia at three major Oldowan research areas. The oldest of these three localities are found in the Gona region in drainages feeding into the modern Awash River (Semaw et al., 1997). These sites were located near large cobble conglomerates, likely on the

floodplains of a paleo-Awash river. These sites also preserve some of the earliest evidence of stone tool use in the form of cut marked bones at a few localities in the Dana Aouli and Ounda Gona drainage system (Dominguez-Rodrigo et al., 2005). Sites in the adjacent Hadar region also preserve numerous evidence of stone-flaking activity (Kimbel et al., 1996). Further south in the Middle Awash research area, evidence of stone tool use occurs only in the form of hominin-modified bone at the site of Bouri (de Heinzelin et al., 1999). The lack of stone artifacts at this site has been suggested to be the product of early hominin ‘curation’ of stone used for the manufacture of tools. Similar patterns of artifact transport have been suggested for sites in East Turkana where hominin-modified bones are found without associated tools (Harris et al., 2002).

The Turkana Basin in northern Kenya arguably contains the most productive deposits for studying the earliest archaeology. The oldest sites are found in the lower Omo valley in the Shungura Formation. The assemblages are dominated by quartz raw material and are often suggested to be the product of expedient bipolar reduction of pebbles. However, re-analysis suggests that a relatively high degree of planning went into the hard hammer percussion of small pebbles in some of these early Omo assemblages (de la Torre, 2004). This interpretation of the Omo assemblages is particularly interesting given the fact that sites of a similar age in the

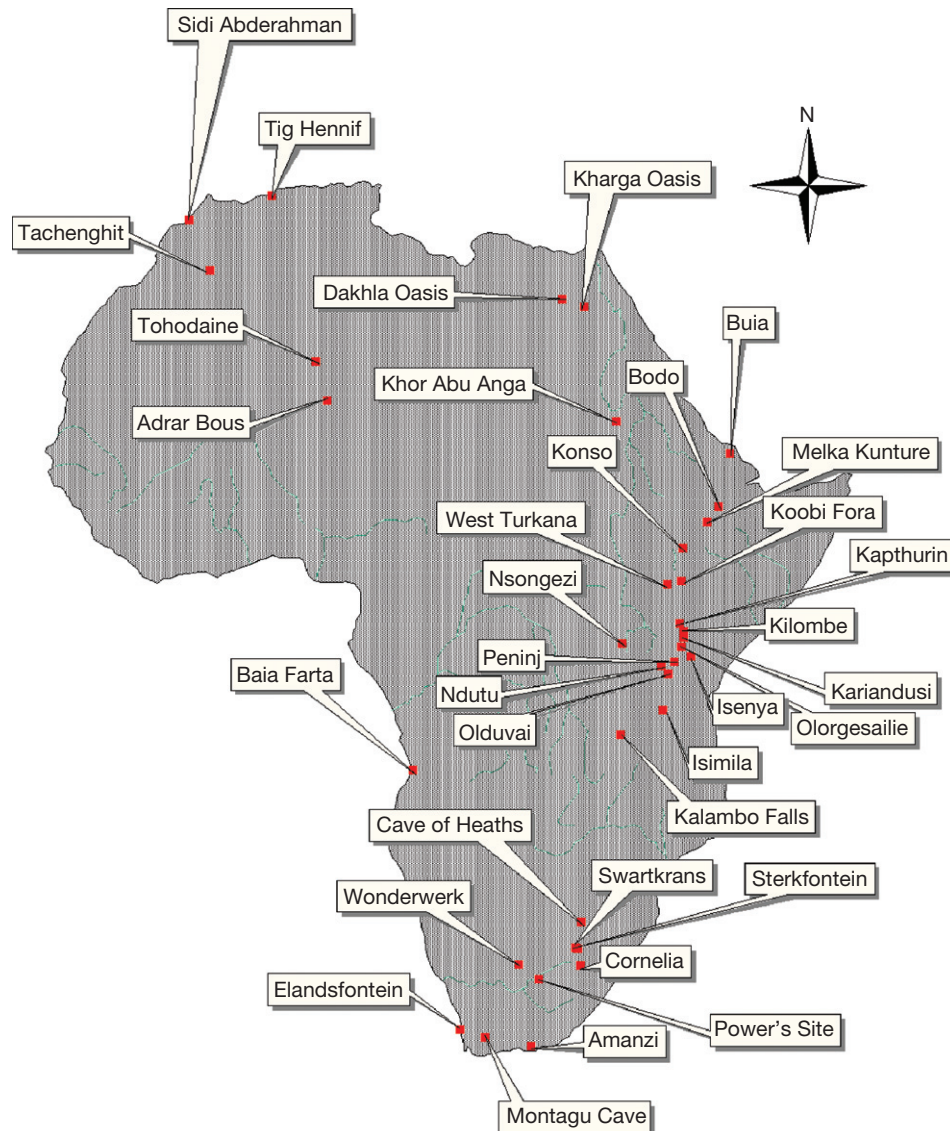


Figure 2 Site localities attributed to the Acheulan Industrial Complex on the African Continent.

Nachukui Formation on the West Side of Lake Turkana show evidence of markedly variable skill levels in artifact production. The Lokalalei 2c site shows evidence of well-defined rules in stone tool production and core reduction (Delagnes and Roche, 2005). The appearance of stone artifacts is presently only known from the KBS Member (1.89–1.65 Myr) in the Koobi Fora Formation on the eastern shores of Lake Turkana (Isaac, 1997). These sites are dominated by freehand percussion of basaltic materials (Isaac, 1997). However, further north on the east side of Lake Turkana at the site of Fejej (de Lumley and Beyene, 2004), bipolar reduction of small quartz pebbles is the most common form of core reduction. This dichotomy in stone tool production is considered to be caused by the differences in available raw material in the different parts of the basin and attests to the limited mobility of the earliest tool makers (Rogers et al., 1994).

However, new evidence from the site of Kanjera South in western Kenya suggests the majority of the raw material used

for artifact manufacture was transported to this site (Plummer, 2004). It is quite likely that the technological system found at the Kanjera sites is unique among Oldowan localities. Transport and selection of specific raw materials seems to be a very tightly constrained part of the technological organization at Kanjera (Braun et al., 2005). Similarly, transport of materials at Olduvai Gorge seems to follow some specific patterns about the use and manufacture of tools. Recent landscape archaeology studies in the Olduvai basin have suggested that hominins practiced selective transport over long distances but movement of raw materials over shorter distances did not seem to abide by the usual relationship between distance from stone source and attrition of a specific raw material in an assemblage.

Although there is general agreement that Oldowan hominins could consistently flake stone, there is considerable disagreement over the relative skill level of the first toolmakers. Research at the Gona sites has suggested that even at the very inception of stone tool technology, Oldowan hominins had



Figure 3 Oldowan core, initially classified by M.D. Leakey (1971) as a chopper in her terminology.

already mastered stone fracture mechanics (Semaw et al., 1997). However, the evidence from West Turkana shows marked variation in the ability of early hominins to produce flakes and reduce cores in a systematic way (Delagnes and Roche, 2005).

Faunal assemblages associated with Oldowan artifact assemblages are dispersed throughout Africa with the earliest of these occurrences in East Africa. Hominin-induced modifications have only been identified from a few of the Oldowan artifact-bone concentrations that have been excavated. However, the absence of hominin-induced bone modifications is not always an indicator of the lack of hominin involvement in the formation of a bone assemblage. In fact, it may be the result of poor bone-surface preservation that obscures bone-surface modifications due to the simple fact that few bone assemblages have been subjected to analyses concerned with the identification of bone-surface modifications. Localities that have produced Oldowan bone assemblages that have been identified as archaeofaunas based on their spatial association with artifacts include Hadar, Middle Awash, Omo (Shungura Formation), and Melka Kunture in Ethiopia; Upper Semliki Valley in the Democratic Republic of Congo; Ain Hanech and El Kherba in Algeria; and Sterkfontein in South Africa (Plummer, 2004). Localities that have produced Oldowan bone assemblages that exhibit evidence of hominin-induced modifications include Gona and Bouri in Ethiopia; Kanjera South, West Turkana, and Koobi Fora in Kenya; and Olduvai Gorge in Tanzania. These sites have the potential to provide insights into the subsistence capabilities of hominins and the interactions between hominins and carnivores who were competing for animal carcasses.

Sites that preserve bone-surface modifications range in composition from a few cut-marked bones found on the surface to thousands of bones that preserve modifications induced by both hominins and carnivores. The earliest examples of cut-marked bones come from the finds at Gona and Bouri in Ethiopia and date between 2.5–2.6 Myr. These sites are important in that they represent the beginning of a dietary shift toward meat eating that coincides with the earliest evidence of stone tool manufacture. However, due to the extremely limited number of bones preserving hominin-induced modifications at these sites, it is difficult to determine the extent of hominin carnivory in this remote time period. As a result our

conception of early tool-using hominins as hunters or scavengers is still open to interpretation.

The majority of attention from studies interested in interpreting early hominin food acquisition has been focused on the bone assemblages excavated from the FLK 22 site, Olduvai Gorge, Tanzania and the co-occurrence of hominin- and carnivore-induced modifications that are associated with carcass consumption. Some authors assert that the abundance of cut marks on bones from Plio–Pleistocene archaeological sites indicates that hominins had primary access to flesh (Dominquez-Rodrigo, et al., 2002), whereas others argue that cut marks result only from the removal of scraps of flesh after large cats had primary access to carcasses (Blumenschine, 1995). While a consensus has still not been reached on the food acquisition and consumption practices of Oldowan hominins at FLK 22, archaeologists generally agree that by 1.8 Myr hominins were depending more on nutrients obtained from carcasses than their predecessors at 2.6 Myr. These findings have extensive implications for the behavior of *Homo erectus*, a species that appears in the fossil record at about this time. It is possible that a major reason for the success and geographic radiation of *H. erectus* is the incorporation of larger amounts of high-quality protein (meat) in the diet.

Transitional Industries: Developed Oldowan–Early Acheulean

Assemblages of artifacts and bones found between 1.7 and 1.5 Myr are often described as Developed Oldowan industries. Many of these sites appear to have a mixture of features associated with both mode I and mode II industries. These ‘transitional’ assemblages include the variability of the earliest Oldowan assemblages but are also precursors of the imposition of form on core morphology. This pattern is manifested in the repeated use of particular forms or core-reduction techniques. The clearest example of this appears in the Karari Industry on the east side of Lake Turkana (Harris and Isaac, 1976) with the introduction of larger ‘scraper’ forms that dominate the Oldowan assemblages after 1.6 Myr. Similar patterns can be found in sites such as Peninj, Chesowanja, and Gadeb (see Figure 4).

The appearance of these ‘transitional’ industries that are usually grouped with the Oldowan but show technological similarities with mode II industries suggests a continuity in the Paleolithic record which may argue against previously suggested models of episodic change in ‘cultural evolution’ (Semaw et al., 1997). These transitional industries are particularly interesting because at about 1.8 Myr (or possibly earlier), fossil remains attributed to *Homo ergaster* (or early African *H. erectus*) began to appear in sediments in East Africa. Well-preserved cranial and postcranial remains from Lake Turkana basin, most notably the Nariokotomie skeleton (KNM-ER 15 000), reflects major changes in the anatomy and physiology of hominins at this time. The concurrent change in the archaeological record represented by transitional (Developed Oldowan and Early Acheulean) assemblages may record a shift in behavioral complexity of this new hominin form (see Figure 5).

Behavioral traces from this time period began to be found in habitats that were not previously exploited by early

hominins, such as highland plateaus (Gadeb, Melka Kunture) and possibly drier savannah habitats, on the floor of the East African Rift Valley at sites such as Koobi Fora and Olduvai. Although it is difficult to make a strict link between these transitional assemblages and the appearance of *H. ergaster*, certain aspects of the biology of a larger hominin, such as



Figure 4 Developed Oldowan or Karari Industry core, initially classified as a Karari core scrapper; Harris and Isaac (1976).

secondary altriciality and larger home range sizes, suggest links between behavior and biology. These anatomical changes complement the archaeological record, which suggests increased efficiency in raw material procurement and use in these transitional industries. This is consistent with possibly complex foraging strategies, which may have been linked to an economic organization which included a highly variable tool-kit adapted to more variable ecological settings.

Acheulean

The Acheulean Industry is often described as mode II industry. **Figure 2** displays the localities of Acheulean sites distributed over large parts of the African continent. All of them include some of the characteristic hand-axes that mark the transition to mode II industry. Often referred to as large cutting tools, these larger implements are the hallmark of the Acheulean Industry (see **Figure 6**).

Acheulean sites are now known to exist in the African interior in particular, in more open, elevated, drier parts of sedimentary basins (e.g., Olduvai Bed III, Koobi Fora-Chari Member), and on the high plateaus of the flanks of the east African rift valley at high altitude (e.g., Gadeb, Isenya). Acheulean sites are found in a wide variety of settings including marine beach localities in northwest Africa (e.g., Sidi Abderrahman) and along the eastern and western coastlines of the southern part of the continent (Baia Foura and Amanzi). The distribution of sites does not necessarily indicate the penetration of the tropical rainforest in the watersheds and main drainages of the Congo River basin or the more forested regions of West Africa. However, the presence of Acheulean toolmakers is indicated in the more closed habitats of east Africa such as the site of Nsongezi, Uganda. Sites located

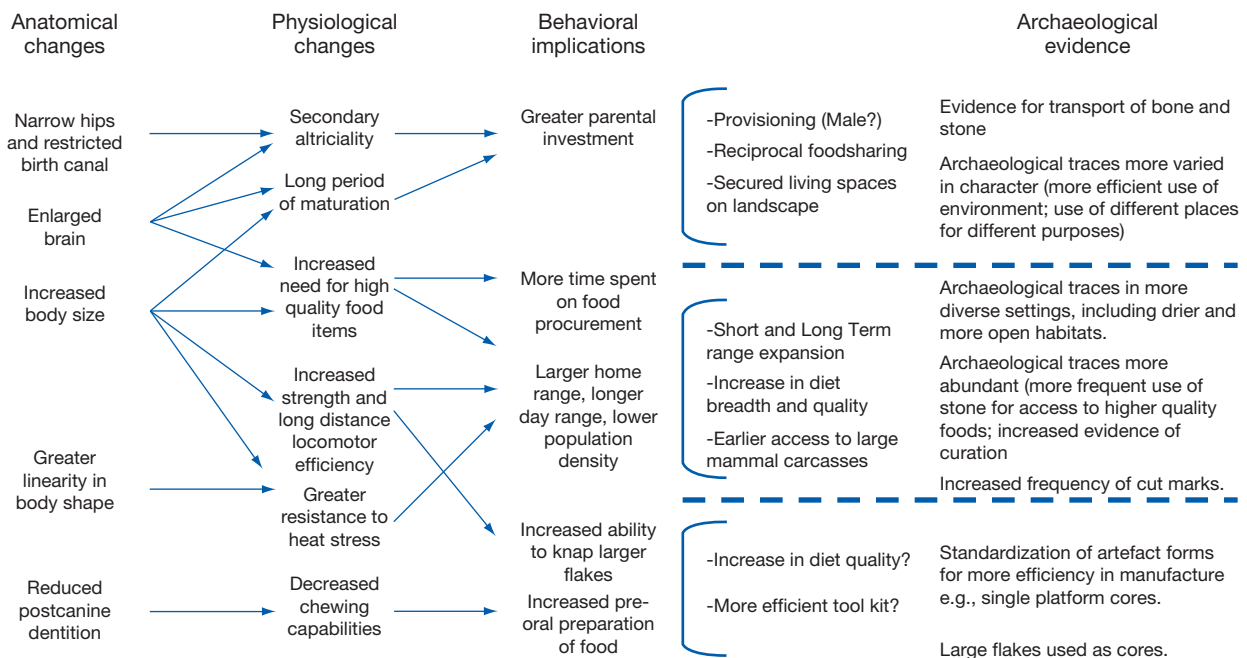


Figure 5 Important shifts in behavioral complexity of early African *Homo erectus* or *Homo ergaster* associated with Developed Oldowan and Early Acheulean archaeological assemblages. (Modified after Rogers, Harris, and Fiebel, 1994)



Figure 6 Acheulean biface or handaxe, which may have had multiple uses, including as a core for the production of flakes.

in this setting commonly have thick and broadly flaked bifaces, often attributed to the Sangoan tradition. Acheulean sites are also found in and around the margins of the Sahara desert. During the Pleistocene, these more arid regions may have undergone dramatic oscillations, which would have been host to less harsh steppe environments (Isaac, 1982).

In addition to the diversity of topographic and ecological settings, Acheulean sites are known from open air and cave settings. In most instances, Acheulean sites are characteristically found near water sources (spring deposits or fluvial systems). Overall, the presence of these sites in these varied habitats suggests that *H. ergaster* and subsequently *H. heidelbergensis* exploited a diversity of habitats. It can be inferred that the increased dietary requirements of these more encephalized hominins forced increased home and day ranges, which led to an ever-increasing overall geographic range resulting in a pan-African distribution. The perils of exploiting open habitats would have been ameliorated by the larger body size of *H. erectus* as well as the possible use of fire. The behavioral package displayed by *H. erectus* (and later *H. heidelbergensis*) would have allowed this hominin to survive in a niche that would have required direct competition with the Pleistocene carnivore guild. Controversial evidence for the control of fire exists as early as 1.5 Myr at Koobi Fora, Chesowanja, and Gadeb. It has yet to be proved whether fire was an adaptive tool to enhance access to high-quality food items in more open grasslands as well as high-altitude settings subject to freezing overnight temperatures. Yet by 300–400 ka, evidence at Kalambo Falls of charred logs, ash, and fire-reddened soil indicates unequivocal use of fire by hominins by the middle Pleistocene (Clark and Harris, 1985). Furthermore, ash deposits at the Acheulean sites of Cave of Hearths and Montagu Cave in southern Africa indicate that this was definitely part of

middle Pleistocene hominin adaptations. The substantial changes in the body size of middle Pleistocene hominins and concurrent encephalization would have required a shift in diet that facilitated increased nutritional intake (Milton, 1987). The exploitation of larger mammal carcasses may have provided a source for this increased nutritional intake. Therefore, assessing changes in the subsistence capabilities between *H. ergaster* and their Oldowan hominin ancestors is essential to determining the significance of meat eating in Plio–Pleistocene hominin evolution. Despite the obvious importance of Acheulean archaeofaunas, very few of these assemblages have enjoyed the full suite of taphonomic analyses. Acheulean sites that have produced extensive faunal assemblages include Ternifine, Sidi Abderrahman, and Bir Tarfawi in north Africa; Bodo, Melka Kunture, Gadeb, Karari (Koobi Fora), Konso, Kapthurin, Kanjera North, Kariandusi, Kilombe, Olorgesailie, Isenya, Lainyamok, Olduvai (Bed III–IV), Ndutu, and Peninj in east Africa; and Kalambo Falls, Wonderwerk, Swartkrans, Sterkfontein, Amanzi, and Elandsfontein in southern Africa. Despite an abundance of sites, meaningful interpretations of Acheulean hominin behavior are limited because several of these sites are from secondary contexts. Winnowed or transported assemblages often do not have the integrity required to make meaningful behavioral interpretations. Subsequently, many of these assemblages are overlooked in discussions of Acheulean paleoenvironmental context or the dietary adaptations of Acheulean hominins. Recent zooarchaeological investigations of Acheulean assemblages are beginning to produce interesting results. New rounds of data analysis, including sites such as Swartkrans (South Africa) and Peninj (Tanzania), are intermediary sites that are associated with ‘Developed Oldowan’ and Acheulean lithic assemblages. The association between the faunal remains and lithic assemblages at Swartkrans is still tenuous. Yet, the faunal assemblages at Peninj in Tanzania demonstrate clear differences between the ecological contexts of accumulations associated with Oldowan tools and those associated with Acheulean bifaces (Dominguez-Rodrigo et al., 2001). At Peninj, Oldowan-like tools are found near the basin margin and are associated with butchery activities. In contrast, the riverine deposits further from the Natron basin axis are found in fluvial contexts and are devoid of archaeofaunal evidence. Phytolith analysis of Acheulean hand-axes at Peninj suggests these bifaces were being used as wood-working tools. The subsequent inference is that the behavioral repertoire represented at these sites also included wooden spears, which subsequently eroded away. At all the sites found in Peninj, hominins are believed to have acquired early access to carcasses based on the anatomical location of cut marks on long bone fragments (Domingo-Rodriguez et al., 2002). This may indicate that Acheulean hominins were hunting or confrontationally scavenging to obtain meat, rather than passively scavenging flesh scraps and marrow from carnivore kills as their Oldowan hominin ancestors are believed to have done.

Excavations at other Acheulean sites, including Olorgesailie and Olduvai Gorge, have recovered Acheulean bifaces in association with extensive faunal remains (Leakey and Roe, 1994; Potts et al., 1999). Olorgesailie preserves an association between the bones of 57 giant gelada baboons and over 4000 stone artifacts, which, along with breakage patterns, lead some researchers to infer that hominins, were hunting baboons

(Shipman et al., 1981). Taphonomic analyses of this assemblage and other assemblages are in progress and will presumably reveal the diversity of hominin dietary adaptations during the middle Pleistocene. Of particular interest is whether the behavioral patterns seen at Peninj are a local response to the ecological contexts of the Natron basin or are indicative of an overall Acheulean pattern. The Middle Awash Valley of Ethiopia has preserved an extensive record of the Acheulean, dating between 1.0 Ma and 0.1 Myr (Clark and Schick, 2000). The artifact-bone concentrations are found in both primary and secondary contexts with several localities preserving bifaces in association with hominin-modified bone. The Herto Member (400–100 ka) hippo butchery sites in the Bouri Formation hold special interest in this regard. Here, bifaces are found in association with several hippo butchery sites; they exhibit evidence of hominin-induced modification to bone including a hippo cranium with large chop marks, suggesting bifaces may have been used to butcher large animal carcasses.

The incorporation of plant foods into the diet of *H. erectus* cannot be underestimated, even though plant remains are sparse in the archaeological record. Edible plant foods such as fruits, nuts, berries, roots, tubers, and seeds may have been important ‘fall-back foods’ in times of environmental stress. Two Acheulean sites, Amanzi Springs and Kalambo Falls, in southern Africa have evidence of preserved plant materials due to the waterlogged conditions of the sediments. At both sites, woods and seeds are preserved. At Kalambo Falls, evidence of fruit underscores the potential of these food items. These discoveries at Acheulean sites underscore the importance of plant foods in the overall diet of *H. erectus*.

The presence of bone tools in early Pleistocene archaeological sites in South Africa and East Africa contributes to our understanding of the diversity of tools incorporated into early hominin’s foraging practices. A recent study of the collection of bone tools from Member 1–3 (1.8–1.0 Myr) at Swartkrans (South Africa) shows that, in some instances, there was an intentional grinding to modify the polished pointed surface of bone tools. It is believed that this practice increased the digging efficiency of these tools, particularly in opening termite mounds (Backwell and D’Errico, 2005). The clear time investment in tool manufacture for termite extraction emphasizes the importance of insect foods as a source of protein in the Acheulean diet. The bone tools from Olduvai Gorge in East Africa (Leakey and Roe, 1994) were deliberately flaked into mostly irregular shapes. Some of the more heavy-duty hammer and anvil tools may have been used in large mammal carcass-processing (Backwell and D’Errico, 2005). The geographical differences may be a reflection of regional adaptations to foraging for variable food resources. This would be taken to reflect the flexibility of *H. erectus* to exploit a great diversity of food resources that might have been only seasonally available in different ecological settings. One interpretation of the variation in bone tool production strategies is that different hominin taxa (*Paranthropus robustus* in South Africa and *Australopithecus boisei* in East Africa) created these different industries.

Highly variable assemblages including more advanced core preparation and flake removal techniques attributed to the Levallois technique also characterize the final stages of the Acheulean. In addition, numerous middle Pleistocene assemblages are associated with more elongated thick core axes,

pick-like bifaces and a variety of small retouched or utilized flakes generally designated as Sangoan. The Sangoan tool assemblages could be viewed as a transitional industry as they continue the tradition of bifacial core axes and pointed forms. These industries could represent early manifestations that are present in younger industries that characterize the MSA.

Conclusion

A major unresolved issue in the ESA is whether the beginnings of Oldowan tool manufacture and use occurs abruptly in a punctuated period of cultural evolution or the origin of tool use has been a gradual event. An alternative hypothesis would be that stone tools were part of a behavioral repertoire of earlier Pliocene hominins that have yet to be recovered in the archaeological record. Based upon the current evidence in the Kada Gona region, the beginnings of stone tool manufacture could be interpreted as an important adaptive threshold or innovation occurring abruptly, perhaps in response to changing environmental conditions brought about by a major global cooling event between 2.8 and 2.4 Myr. This time interval also coincides with an adaptive radiation of the Homininae. It is possible that a more efficient processing of sources of high-quality protein facilitated this adaptive radiation. The incorporation of high-quality protein (presumably meat and marrow) and other nutritious food sources into the diet of one hominin species (most likely early *Homo* or by multiple hominins including the australopithecines and paranthropines) surely changed regional ecological dynamics in the early Pleistocene.

Clearly by 1.5 Myr (if not earlier), there does seem to be a correspondence between anatomical change and technological elaboration characterizing the Developed Oldowan and early Acheulean toolkits. The increasingly widespread nature, geographically and ecologically, of Acheulean sites may be linked to the emergence of a bigger-bodied hominin (early African *H. erectus* or *H. ergaster*). Through the course of the time period associated with Acheulean assemblages, large bifacially flaked cutting tools appear to become more refined and the end of the Acheulean is associated with prepared core technologies.

The growing complexity of the toolkit and evidence of a wide-ranging hominid inhabiting more diverse locations on the landscape may indicate the increasing intelligence, mental mapping, and growing social complexity in Acheulean lifeways. A possible indication of interpersonal group conflict comes from cut marks found on the human cranium recovered at the site of Bodo, in the Middle Awash region of Ethiopia. This specimen, attributed to *H. heidelbergensis* is associated with a large occurrence of Acheulean tools. The extent of the hominin modification found on this specimen suggests it was intentionally defleshed. This rare Bodo example may be one of the earliest recorded instances of cannibalism. This evidence may indicate that symbolic behavior, more characteristic of anatomically modern humans and their associated MSA technologies, had its beginnings in the late Acheulean. It is possible that many of the features Paleolithic archaeologists often associate with the development of modern human behavior had their roots in the deep time of the ESA.

See also: [Archaeological Records: Overview](#); [2.7 Myr–300 000 Years Ago in Asia](#); [1.9 Myr–300 000 Years Ago in Europe](#); [Global Expansion 300 000–8000 Years Ago, Africa](#); [Global Expansion 300 000–8000 Years Ago, Asia](#); [Global Expansion 300 000–8000 Years Ago, Australia](#); [Global Expansion 300 000–8000 Years Ago, Americas](#); [Neanderthal Demise](#); [Postglacial Adaptations](#); [Vertebrate Studies: Interactions with Hominids](#).

References

- Backwell L and D'Errico F (2005) The origin of bone tool technology and the identification of early hominid cultural traditions. In: D'Errico F and Backwell L (eds.) *From Tools to Symbols*, pp. 238–275. Johannesburg: Witwatersrand University Press.
- Binford LR (1981) *Bones, Ancient Men and Modern Myths*. New York: Academic Press.
- Blumenschine RJ (1995) Percussion marks, tooth marks and experimental determinations of the timing of hominid and carnivore access to long bones at FLK Zinjanthropus, Olduvai Gorge, Tanzania. *Journal of Human Evolution* 29: 21–51.
- Braun DR, Plummer T, Ferraro JV, et al. (2005) *Oldowan technology at Kanjera South, Kenya: The context of technological diversity*. PaleoAnthropology.
- Bunn HT (1981) Archaeological evidence for meat-eating by Plio-Pleistocene hominids from Koobi Fora and Olduvai Gorge. *Nature* 291: 547–577.
- Clark G (1969) *World Prehistory: A New Outline*. Cambridge: Cambridge University Press.
- Clark JD and Harris JWK (1985) Fire and its roles in early hominid lifeways. *African Archaeological Review* 3: 3–27.
- Clark JD and Schick K (2000) Acheulean archeology of the eastern middle awash. In: de Heinzelin J, Clark JD, Schick K, and Gilbert WH (eds.) *The Acheulean and the Plio-Pleistocene Deposits of the Middle Awash Valley, Ethiopia, Vol. 104: Annales Sciences Geologiques*. Musee Royal De L'Afrique Centrale, Tervuren, Belgique.
- de Heinzelin J, Clark D, White T, et al. (1999) Environment and behavior of 2.5-million-year-old Bouri hominids. *Science* 284: 625–629.
- Delagnes A and Roche H (2005) Late Pliocene hominid knapping skills: The case of Lokalalei 2C, West Turkana, Kenya. *Journal of Human Evolution* 48: 435–472.
- De la Torre I (2004) Omo revisited: Evaluating the technological skill of Pliocene hominids. *Current Anthropology* 45: 439–465.
- deLumley H and Beyene Y (2004) *Les Sites Préhistoriques de la Région de Fejej, Sud-Omo, Éthiopie, dans leur Contexte Stratigraphique et Paléontologique*. Paris: Editions Recherche sur les Civilisations.
- Dominguez-Rodrigo M (1997) Meat-eating by early hominids at the FLK 22 Zinjanthropus site, Olduvai Gorge (Tanzania): An experimental approach using cut-mark data. *Journal of Human Evolution* 33: 669–690.
- Dominguez-Rodrigo M, de la Torre I, de Luque L, et al. (2002) The ST site complex at Peninj, West Lake Natron, Tanzania: Implications for early hominid behavioural models. *Journal of Archaeological Science* 29: 639–665.
- Dominguez-Rodrigo M, Pickering TR, Semaw S, and Rogers MJ (2005) Cutmarked bones from Pliocene archaeological sites at Gona, afar, Ethiopia: Implications for the function of the world's oldest stone tools. *Journal of Human Evolution* 48: 109–121.
- Dominguez-Rodrigo M, Serrallonga J, Jaun-Tresserras J, Alcalá L, and Luque L (2001) Woodworking activities by early humans: A plant residue analysis on Acheulean stone tools from Peninj (Tanzania). *Journal of Human Evolution* 40: 289–299.
- Gabunia L, Anton SC, Lordkipanidze D, Vekua A, Justus A, and Swisher CC (2001) Dmanisi and dispersal. *Evolutionary Anthropology* 10: 158–170.
- Harris JWK, Braun D, McCoy JT, Pobiner BL, and Rogers M (2002) New directions and preliminary results from a landscape approach to the study of archaeological traces for the behavior of Plio-Pleistocene hominids at Koobi Fora. *Journal of Human Evolution*. 42: A15.
- Harris JWK and Isaac GLL (1976) The Karari Industry: Early Pleistocene archaeological evidence from the terrain east of Lake Turkana, Kenya. *Nature* 262: 102–107.
- Isaac G (1982) The earliest archaeological traces. In: Clark JD (ed.) *Cambridge History of Africa*, vol. I, pp. 157–247. Cambridge: Cambridge University Press.
- Isaac G. L1 (1997) *Koobi Fora Research Project: Plio-Pleistocene Archaeology*. Oxford: Clarendon Press.
- Kimbel WH, Walter RC, Johanson DD, et al. (1996) Late Pliocene *Homo* and Oldowan tools from the Hadar formation (Kada Hadar Member), Ethiopia. *Journal of Human Evolution* 31: 549–561.
- Leakey MD (1971) *Olduvai Gorge: Excavations in Beds I and II, 1960–1963*. Cambridge: Cambridge University Press.
- Leakey MD and Roe D (eds.) (1994) *In Olduvai Gorge Excavations in Beds III, IV, and the Masek Beds 1968–1971*. Chapters 2–7, Cambridge: Cambridge University Press.
- Ludwig B and Harris JWK (1998) Towards a technological reassessment of East African Plio-Pleistocene lithic assemblages. In: Petraglia MD and Korisettar R (eds.) *Early Human Behavior in Global Context*, pp. 84–107. London and New York: Routledge.
- Milton K (1987) Primate diets and gut morphology: Implications for hominid evolution. In *Food and Evolution Toward a Theory of Human Food Habits*. Philadelphia: Temple University Press, pp. 93–115.
- Panger MA, Brooks A, Richmond B, and Wood B (2002) Older than the Oldowan? Rethinking the emergence of hominid tool use. *Evolutionary Anthropology* 11: 235–245.
- Plummer T (2004) Flaked stones and old bones: Biological and cultural evolution at the dawn of technology. *Yearbook of Physical Anthropology* 47: 118–164.
- Potts R, Behrensmeier AK, and Ditchfield P (1999) Paleolandscape variation and Early Pleistocene hominid activities: Members 1 and 7, Olorgesailie formation, Kenya. *Journal of Human Evolution* 37: 747–788.
- Rogers MJ, Harris JWK, and Feibel CS (1994) Changing patterns of land use by Plio-Pleistocene hominids in the Lake Turkana basin. *Journal of Human Evolution* 27: 139–158.
- Semaw S, Renne P, Harris JWK, et al. (1997) 2.5 Million-year-old stone tools from Gona, Ethiopia. *Nature* 385: 333–336.
- Shipman P, Bosler W, and Davis KL (1981) Butchering of giant geladas at an Acheulean site. *Current Anthropology* 22: 257–264.
- Tryon CA and McBrearty S (2002) Tephrostratigraphy and the Acheulean to Middle Stone Age transition in the Kapthurin, Kenya. *Journal of Human Evolution* 42: 211–235.

2.7 Myr–300 000 Years Ago in Asia

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Introduction

In the early twenty-first century, we know as little about early human evolution and settlement in Asia as we knew about that of East Africa in the early 1960s (when *Homo habilis* and *Zinjanthropus* were first discovered), and compared to Europe and much of Africa, there is still a chronic shortage of well-dated sites >300 ka, especially those with associated faunal, fossil hominin, and environmental evidence. Consequently, most statements about Asia's earliest inhabitants are highly vulnerable to new discoveries. Relevant examples from recent years are the wholly unexpected discoveries of very primitive and small-brained hominins dating from 1.75 Ma in Dmanisi, Georgia, the diminutive late Pleistocene hominins from Flores, the Denisovans from Siberia, known only from their ancient DNA, and the re-dating of the earliest archaeological evidence from China to >1.5 Ma. Dennell (2009) provides a comprehensive summary of the Asian evidence before 125 ka. The first issue to consider is when hominins first inhabited Asia.

Early Pleistocene Hominins in Asia

The history of hominins in Asia before 300 ka is often summarized as the outcome of a dispersal event from Africa ca. 1.8 Ma, also known as "Out of Africa 1". However, very little is known about the faunal record of much of Asia before 1.7 Ma, and it remains unclear when hominins first left Africa. Because australopithecines were already in Chad, 2500 km west of the Rift Valley, 3.0–3.5 Ma, some may already have been in Southwest and other parts of Asia by this time. Hominins might also have left Africa shortly after the flaking of stone to make tools became routine ca. 2.6 Ma in East Africa (Dennell and Roebroeks, 2005). Indications of when hominins might have left Africa are provided by the sapropel record of the East Mediterranean (Kroon et al., 1998), and the loess/paleosol record of North China (Huayu Lu et al. (1999); see Figure 1). Sapropels are organic-rich sediments associated with greater inputs of fresh water into the sea; those in the East Mediterranean resulted largely from increased influx from the Nile, presumably during periods of higher rainfall. The Chinese loess and paleosol sequence indicates numerous alternating cool and dry glacial periods, and warmer and moister interglacial ones. Both records indicate several 'windows of opportunity' before 1.7 Ma, during which rainfall and temperatures, and hence biological productivity, were higher, and conditions more favorable for hominins to disperse. In intervening cooler and drier periods, deserts (particularly those between northeast Africa and Southwest Asia) would have encroached onto grasslands, and reduced opportunities for animals such as hominins to move freely between Africa and Asia (Dennell 2013).

At present, the earliest skeletal evidence for hominins in Asia dates from ca. 1.75 Ma from Dmanisi, Georgia and ca. 1.6 Ma from Java (see Figure 2 and Table 1).

Early Pleistocene Hominin Remains from Java and Dmanisi

The hominin remains from Java are classified as *H. erectus sensu stricto*, and were first discovered there in 1891, at Trinil. The earliest hominins from Sangiran date from ca. 1.6 Ma (Larick et al., 2001), and the child's cranium from Mojokerto, which had been dated to 1.81 ± 0.04 Ma, has now been re-dated to <1.49 Ma (Morwood et al., 2003). In The Dmanisi hominins have been precisely dated by paleomagnetism and Ar^{40} to ca. 1.75 Ma. They are remarkably small-brained and were probably 1.5 m or less in height, and their post-cranial skeleton exhibited an interesting mixture of modern and archaic features (Lieberman 2008). The Dmanisi hominins have proved difficult to classify. One mandible has been assigned to a new taxon, *H. georgicus*. The crania are very primitive, and show many resemblances to *H. habilis*. The most authoritative taxonomic assessment is that they are a very primitive form of *H. erectus sensu lato*, and may be ancestral to both early East African populations of *H. erectus* (or *H. ergaster*) as well as the Javan *H. erectus sensu stricto*. That is to say, *H. erectus* may have originated in Asia, and then migrated back into Africa, as well as eastwards across southern Asia to Java and possibly north China. Until more is known of the fossil record of Southwest Asia, this possibility cannot be excluded (Rightmire et al., 2006). The most significant aspect of the Dmanisi evidence is that hominins did not need large brains or bodies in order to venture out of Africa; this clearly implies that earlier, small-bodied and small-brained hominins might also have done the same. Some researchers have, for example, suggested that the late Pleistocene, diminutive *H. floresiensis* (the "hobbit") from Flores, Indonesia (see below), has australopithecine affinities, which may in turn imply a very early dispersal event by the hobbit's ancestors from Africa.

Early Pleistocene Lithic Technology in Asia

Examples of early Pleistocene lithic technology are now known from five regions of Asia. The main evidence for the earliest stone technology of early *H. erectus* in Asia comes from Dmanisi (de Lumley et al. (2005) in the Caucasus; see Figure 3). Here, the earliest tools predate the hominin remains and are ca. 1.85 Ma (Ferring et al., 2011). The artifacts are very simple (in keeping with the primitive nature of their makers), and have been called pre-Oldowan on technological grounds because of the absence of retouched tools. On-going research in Armenia indicates that bifacial Acheulean assemblages are older than 1.4 Ma and possibly ≥ 1.75 Ma (Presnaykov et al., 2012); if the

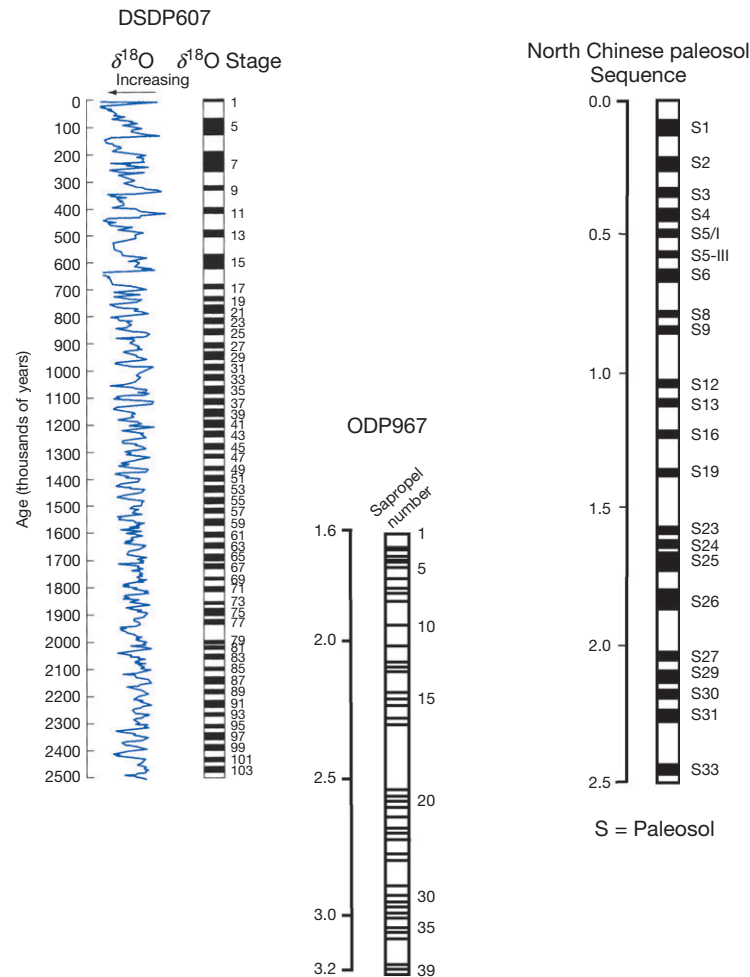


Figure 1 Climatic change and hominin dispersals: the oxygen isotope, East Mediterranean sapropel and Chinese loess records. As can be seen, there were numerous opportunities for hominins to disperse out of, into, and within Asia during the late Pliocene and early Pleistocene. Wilson et al. (2000), figure 4.15; Kroon et al. (1998), figure 3; and Hyayu Lu et al., (1999), figures 2 and 3.

older ages are confirmed, a major re-assessment will be needed of the origins of the Acheulean in Asia. The next oldest sites with stone tools are from the Nihewan Basin in North China, where the sites of such as Majuangou and Xiaochangliang are dated by paleomagnetism to ca. 1.66 and 1.36 Ma, respectively (Zhu et al. (2003, 2004); see Figure 4). These assemblages are broadly similar to Oldowan ones from East Africa that date from 2.6 Ma (see 2.7 Myr–300 000 Years Ago in Africa), and comprise flakes and the cores from which they were struck. The cores are simple in that they were not shaped prior to flaking, and no attempts were made to detach flakes of particular sizes or shapes. The main types recognized are choppers (cores with a few flakes struck from opposing angles), knives and scrapers (i.e., flakes with a sharp edge suitable for cutting or scraping).

The Levant also has an extremely long record of Asian hominins. One of the most informative early Pleistocene sites is 'Ubeidiya, Israel, which comprises a long lacustrine sequence dated faunistically to ca. 1.4–1.0 Ma (Bar-Yosef, 1998). Stone tool assemblages were mainly found in or on swamp deposits and former lake beaches, as well as in fluvial conglomerates.

Assemblages that lacked bifaces are described as Developed Oldowan, and those with bifaces, or cores that have been flaked on both sides to make an ovate or pointed tool and to remove most of the cortex, are classified as early Acheulean (see Figures 5 and 6). It remains unclear whether these were made by different hominins, or are part of the same lithic techno-complex.

South Asia also has evidence of hominins before 1 Ma. The earliest archaeological evidence is a small stone tool assemblage from Riwat, Pakistan, dated to ca. 1.9 Ma. Recently, Acheulean assemblages at Attirampakkam in South India have been dated to at least 1.0 Ma and possibly 1.5 Ma, and this evidence considerably extends eastwards the Early Acheulean in Asia (Pappu et al., 2011). Finally, the recent and unexpected discovery of simple stone tools from the island of Flores, Indonesia shows that this island was colonised by 1 Ma (Brumm et al., 2010), and probably again at 800 ka (Morwood et al., 1998). Because Flores would have been an island even at times of low sea level, the colonists probably arrived accidentally by drifting on natural rafts of vegetation (Smith 2001).

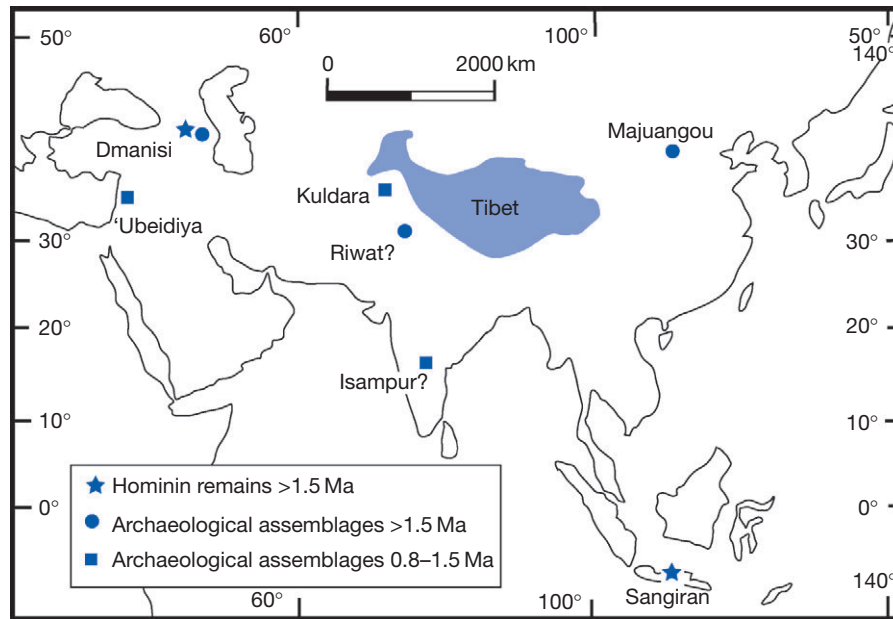


Figure 2 Early Pleistocene hominin and archaeological sites in Asia.

Table 1 Cranial capacities of specimens of early Pleistocene *Homo erectus* specimens from Dmanisi and Java, and middle Pleistocene ones from Zhoukoudian, China. Those from Dmanisi are ca. 1.75 Ma, and those from Java are between 1.0 and 1.6 Ma, except for Mojokerto, which may be slightly older

Origin	Specimen	Description	Cranial capacity (cc ³)
Dmanisi, Georgia	D2280	Adult braincase, possibly male	775
	D2282/D211	Partial cranium of a young adult	650–660
	D2700/D2735	Complete skull of small subadult	600
	D3444/D3900	Edentulous old cranium	625
Java, Indonesia	Mojokerto	Child's cranium	800 ^a
	Trinil 2		940
	Sangiran 2	Calotte	813
	Sangiran 4	Calvarium	908
	Sangiran 10	Calotte	855
	Sangiran 12	Calotte	1059
	Sangiran 17	Calvarium	1004
	Sangiran IX	Calotte	845–
	Sangiran	Average of Sangiran 2, 4, 10, 12, 17, IX and Trinil 2	918
Zhoukoudian, China	Zhoukoudian II	middle Pleistocene, ca. 500 ka	1030
	Zhoukoudian III	middle Pleistocene, ca. 500 ka	915
	Zhoukoudian V	middle Pleistocene, ca. 500 ka	1140
	Zhoukoudian VI	middle Pleistocene, ca. 500 ka	850
	Zhoukoudian X	middle Pleistocene, ca. 500 ka	1225
	Zhoukoudian XI	middle Pleistocene, ca. 500 ka	1015
	Zhoukoudian XII	middle Pleistocene, ca. 500 ka	1030
			1029
Average of crania II, III, V, VI, X–XII			

Rightmire et al. (2005) for Dmanisi, and Antón (2002): Table 1, for Java and Zhoukoudian.

^aEstimated adult size.

Early Pleistocene Hominin Subsistence and Settlement

Because animal bones can be associated with stone tools for a variety of reasons (such as stream action mixing the two together, or hominins discarding stone tools in an area previously used by a carnivore), the main evidence for the

subsistence activities of early Asian hominins depends primarily upon the discovery of bone fragments with cut marks resulting from hominins butchering carcasses with stone tools. (As plant remains are preserved only very exceptionally, we know almost nothing about their use of plant foods.) At present,

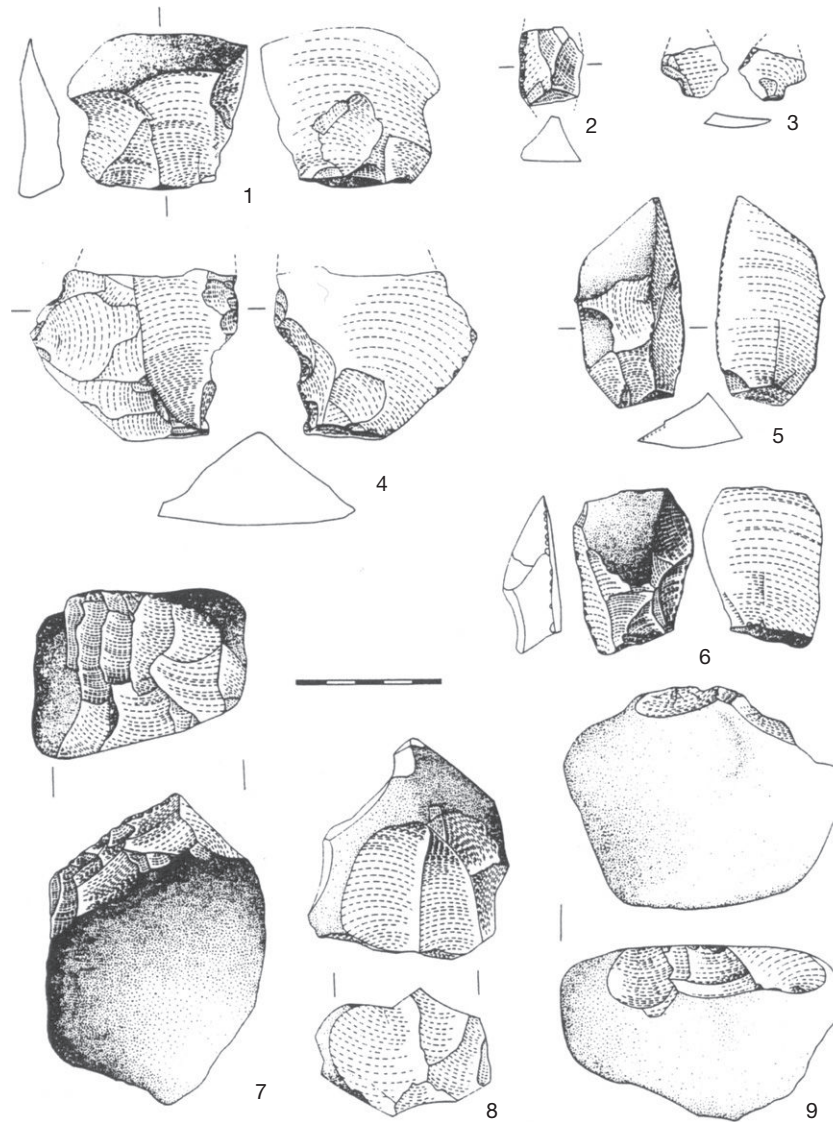


Figure 3 Stone artifacts from Dmanisi. Bar-Yosef (1998), figure 8.4.

little is known about the hunting or scavenging activities of early Pleistocene Asian hominins. There is no evidence for the use of fire, artificial shelters, or projectiles for hunting. Recent faunal evidence from 'Ubeidiya has been interpreted as showing the butchery and hunting of medium-sized game, although the evidence of hominin involvement in either activity is very slight (Gaudzinski, 2004). The nonhominin faunal remains at Dmanisi are still being studied, and those from sites in the Nihewan basin are too fragmented to allow any meaningful insights about subsistence. Inferences from contemporaneous sites in East Africa suggest that early Pleistocene hominins did not hunt, but obtained most of their meat through scavenging, either passively (i.e., taking whatever a carnivore had left) or actively, by capturing carcasses from carnivores as they were feeding (Dominguez-Rodrigo, 2002). Either possibility underlines the importance of predators in the daily life of early *H. erectus* – either to avoid being eaten, or to obtain meat in a world where competition for it was fierce.

The likely size of the home ranges of early *H. erectus* has been estimated from their inferred height and weight as only

330–430 ha per individual (Antón and Swisher, 2004), or an operating radius of 5.1–5.6 km for a group of 25, which is probably the minimum viable group size. They were thus constrained to environments where the basic necessities of water, animal and plant resources, stone for making tools, and shelter from carnivores could be found within a small area. Small lake basins, such as Dmanisi, the Nihewan Basin, or 'Ubeidiya appear to have been the preferred type of hominin location in Asia; such basins are also ones where the chances of finding their remains are also high (see Asia).

Although hominins were distributed across much of Asia by 1 Ma, their populations were probably largely confined to a small number of refugia, where resources were always plentiful. They were also likely discontinuous geographically and temporally (see Figure 7), because climatic fluctuations between glacials and interglacials every 41 ka after 2.5 Ma (see Pliocene Environments) would have greatly influenced the extent of areas suitable for hominins. Many dispersal events into and within Asia are likely to have been short term and unsuccessful in establishing populations that were permanently resident

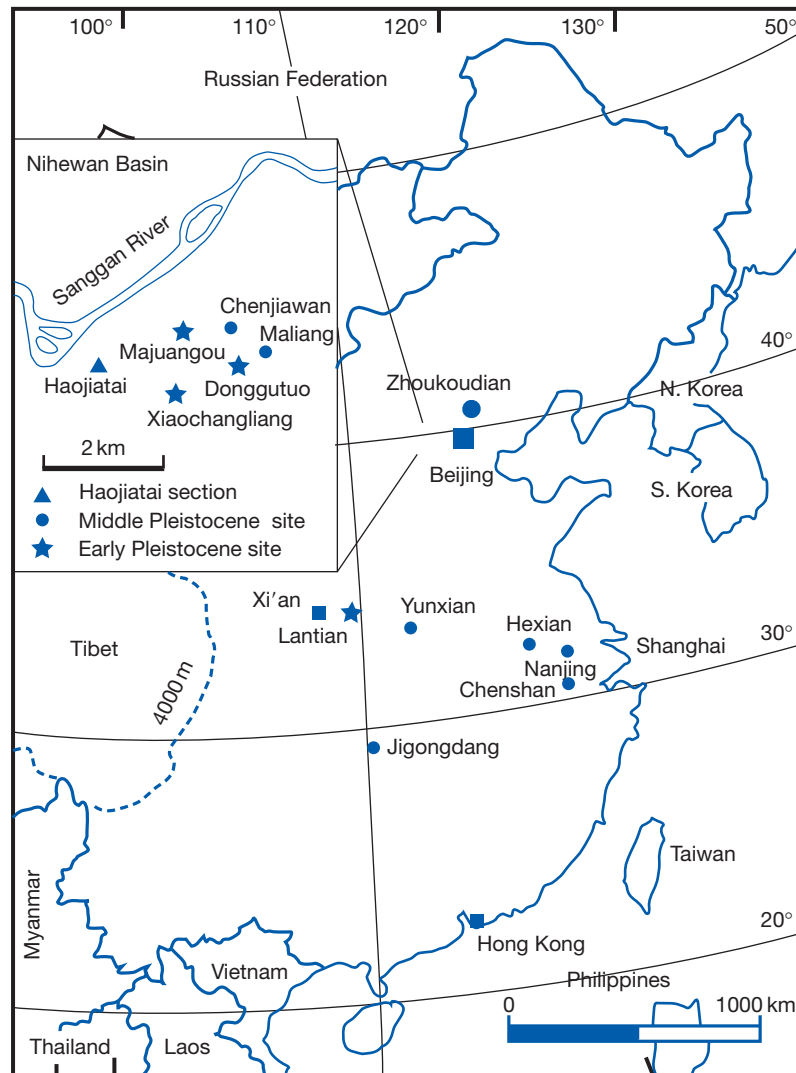


Figure 4 Early and middle Pleistocene sites in China.

(Dennell, 2003). Out of Africa 1 – the first movement of hominins from Africa – was probably a series of dispersal events that were not necessarily one-way from Africa to Asia.

Middle Pleistocene Hominin Archives in Asia, 800–300 ka

In the middle Pleistocene, climatic alternations between warm and cold intervals were more pronounced than before, and cold phases lasted much longer, on a 100 ka cycle instead of 41 ka (see [Middle Pleistocene in Eurasia](#)). Sea level changes were more dramatic, with much of Indonesia joined to mainland Southeast Asia and an extensive coastal shelf exposed off China. Major faunal changes also occurred during what is often referred to as the Galerian Faunal Turnover, which led to the emergence of a modern type of mammalian fauna across Siberia, South Asia, and western Asia (see [Mid-Pleistocene of Southern Asia](#)). The middle Pleistocene also witnessed substantial tectonic activity, especially along the forefront of the Himalayas and Karakorum, and the Qinling Range that now divides

North from South China (see [Tectonic Uplift and Continental Configurations](#)). Hominin regional records thus need to be considered against a background of climatic, faunal, and even tectonic changes (Dennell, 2004).

The middle Pleistocene evidence for hominins in Asia is much more extensive than that for the early Pleistocene (see [Figure 8](#)), but is still largely confined to latitudes south of 40° N, although they may have up dispersed to 50° N in Kazakhstan, where, as yet undated, Acheulean bifaces have been found (Vishnyatsky, 1999). There are several middle Pleistocene regional records from Asia, of which the best are from the Levant ([Figure 9](#)), India ([Figure 11](#)), Tajikistan ([Figure 13](#)), the Caucasus and China.

The Levant

There are two ‘flagship’ sites in this region, Gesher Benot Ya’aqov, Israel (Goren-Inbar et al., 2002, 2004), and Latamne, Syria (Bar-Yosef, 1998). The former lies in the Dead Sea Valley. Its deposits are strongly tilted former river- and lake-shore sediments, dated to ca. 800 ka by paleomagnetism. The site

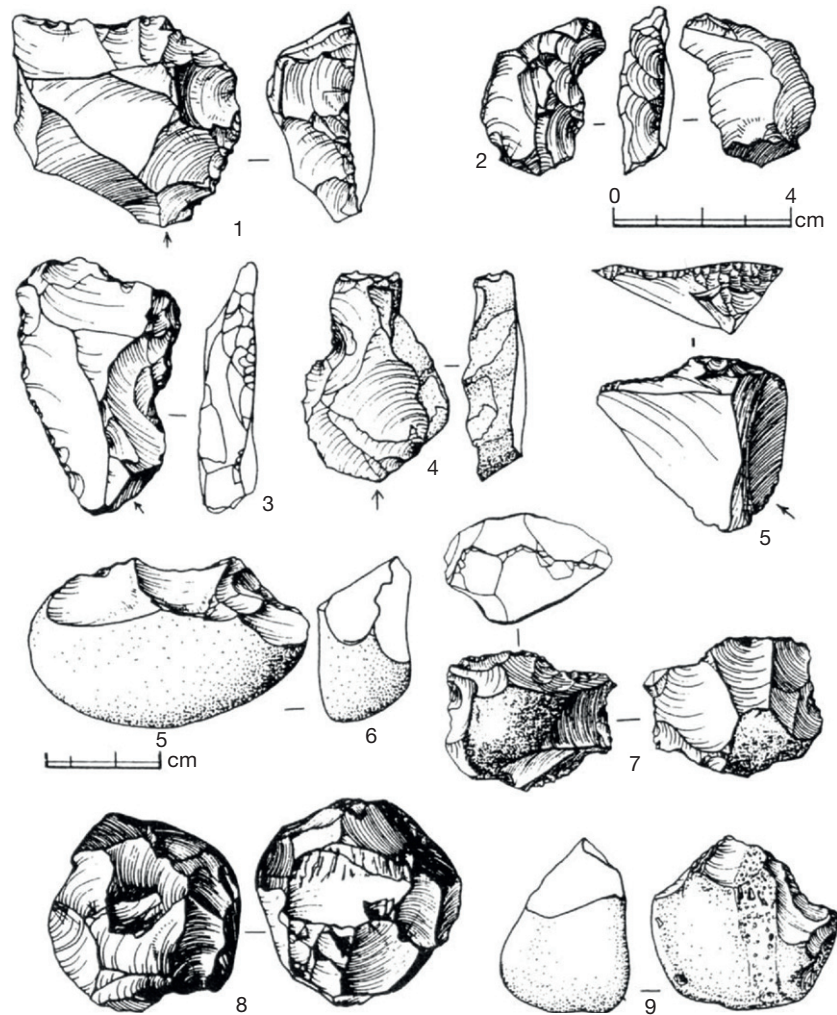


Figure 5 Selected retouched flakes including a scraper, and also core-chopper and a spheroid from 'Ubeidiya. Bar-Yosef (1998), figure 8.9.

contains Acheulean assemblages containing large numbers of basalt cleavers, often made by the Kombewa technique (i.e., from flakes with two opposing bulbs of percussion). The choice of basalt as a raw material, the use of the Kombewa technique, and production of cleavers strongly imply the dispersal of hominins or ideas from Africa, where all these traits occur earlier. In one area of the site, the remains of an elephant skull and a small tool assemblage were found, which has been interpreted as evidence of hunting, rather than the natural death of an elephant in an area where hominins also discarded stone tools. As the site is waterlogged, there is also evidence of plant remains, including charcoal, seed, and nut fragments. According to the excavators, Geshen Benot Ya'aqov contains the earliest evidence for the controlled use of fire, for processing nuts (a highly nutritious resource), and for gathering other plant foods. Faunal evidence indicates that animals were hunted, unlike at 'Ubeidiya, with fallow deer as the main prey (Rabinovitch *et al.*, 2012).

Latamne lies on the banks of the Orontes (see Figure 10). It is probably younger than Geshen Benot Ya'aqov, and is dated to the mid-middle Pleistocene. An area ca. 60 m⁻² was

excavated, within which several hundreds of stone artifacts were found; there was no evidence of structures or fire, and faunal remains were not preserved. Latamne is usually interpreted as a living site, where a group camped while hunting game. Alternatives are that it is a palimpsest of numerous short visits, or a site used seasonally by part of a group.

Another important Levantine Lower Paleolithic site is Berekhat Ram (dated as between 230 and 780 ka) that contains an Acheulean assemblage made from basalt, and an intriguing 'figurine' that is a small piece of volcanic tuff that was intentionally modified. Whether or not this is the earliest example of representational art, it is clearly nonutilitarian, and indicates at least the capacity for some degree of symbolism. Other Levantine sites that are >800 ka are Evron and Bizat Ruhama. The Acheulean at Holon, Revadim, Umm Qatafa, the basal layers (F and G) of Tabun Cave, and Ghamarchi 1b are younger but probably >300 ka (Bar-Yosef, 1998). Although the Acheulean is the most conspicuous part of the Levantine Early Paleolithic, sites such as Bizat Ruhama lack bifaces and contain only flakes and small cores. The final part of the Levantine Early Paleolithic also contains a distinct regional variant known as the Yabrudian.

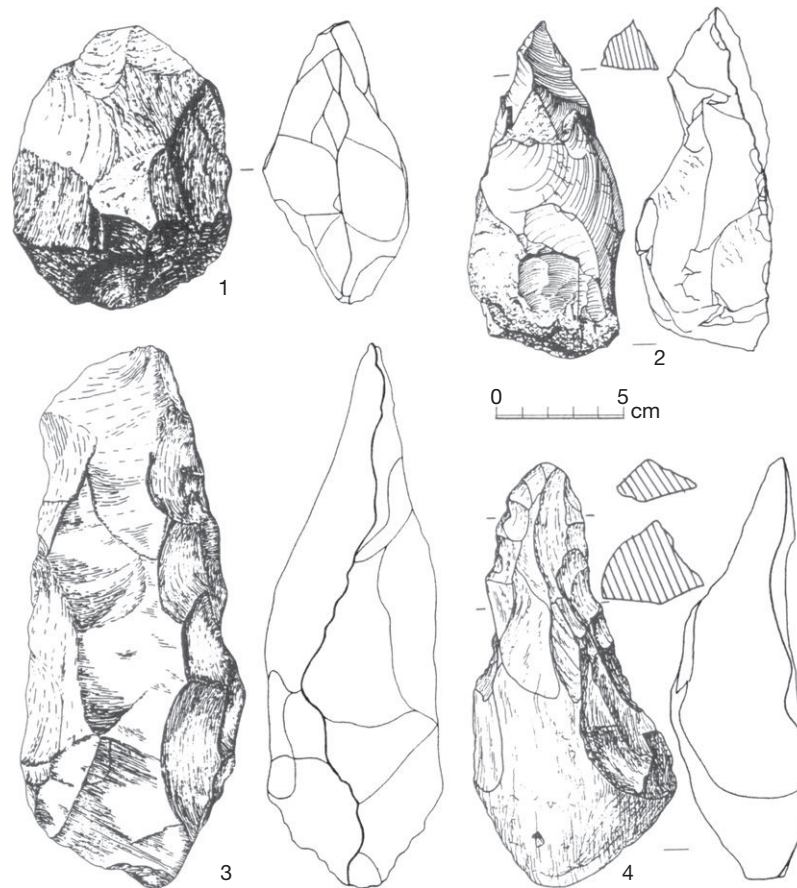


Figure 6 Acheulean bifaces (1,3,4) and a trihedral (2) from ‘Ubeidiya. Bar-Yosef (1998), figure 8.8.

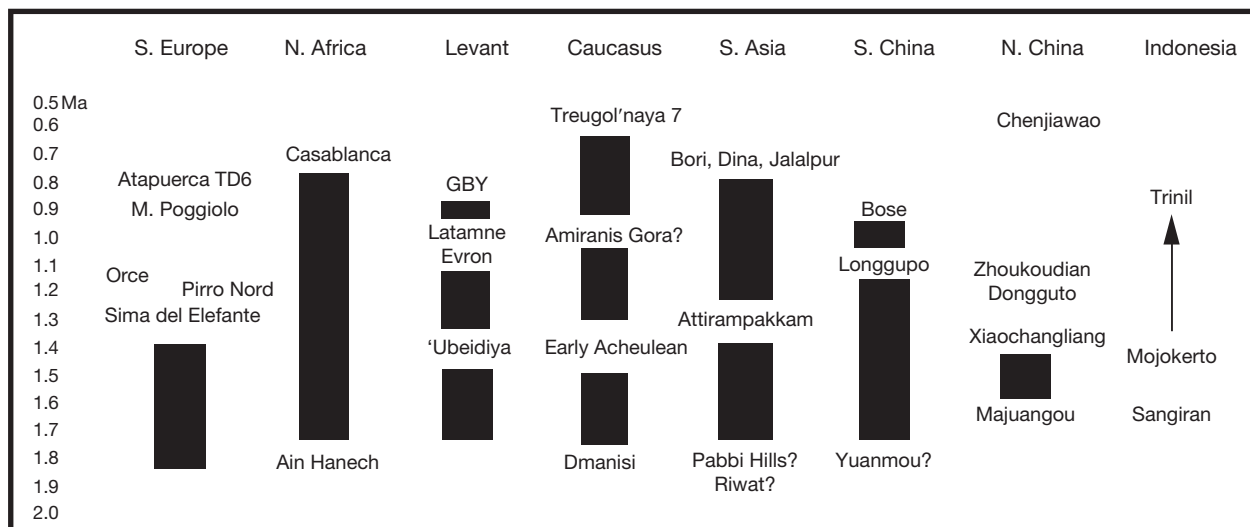


Figure 7 Discontinuities in hominin occupation in the early and middle Pleistocene of Asia.

This is found over Syria, Lebanon, and northern Israel, and comprises a non-Levallois flake industry with a distinctive convergent (*déjeté*) side scraper. Yabrudian assemblages in layer E of Tabun Cave date from ca. 350 to 250 ka, after which they are replaced by Middle Paleolithic, Levallois-Mousterian

assemblages (see [Global Expansion 300 000–8000 Years Ago, Asia](#)). A recent superlative sequence of Yabrudian deposits at the cave of Qesem in Israel, dated to 400–200 ka, shows that animals were hunted, and butchered by several individuals around hearths (Stiner *et al.*, 2010).

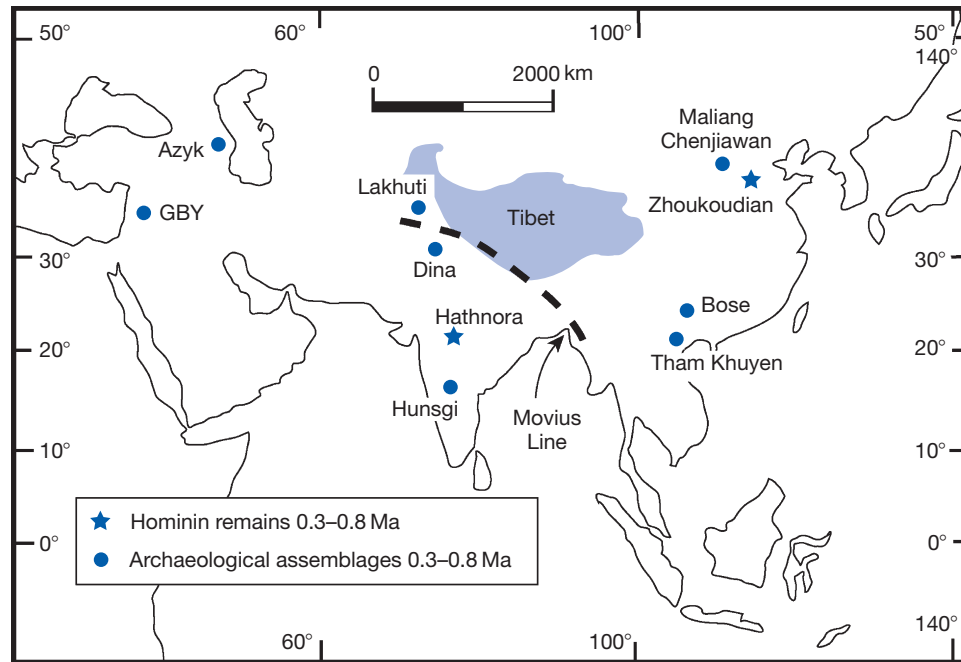


Figure 8 Middle Pleistocene sites in Asia.

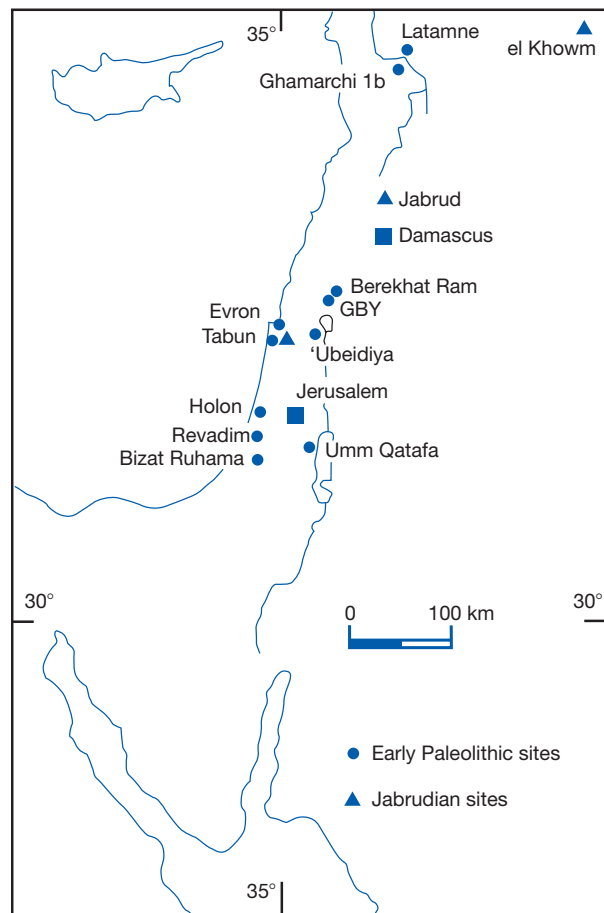


Figure 9 Principal archaeological sites in the Levant.

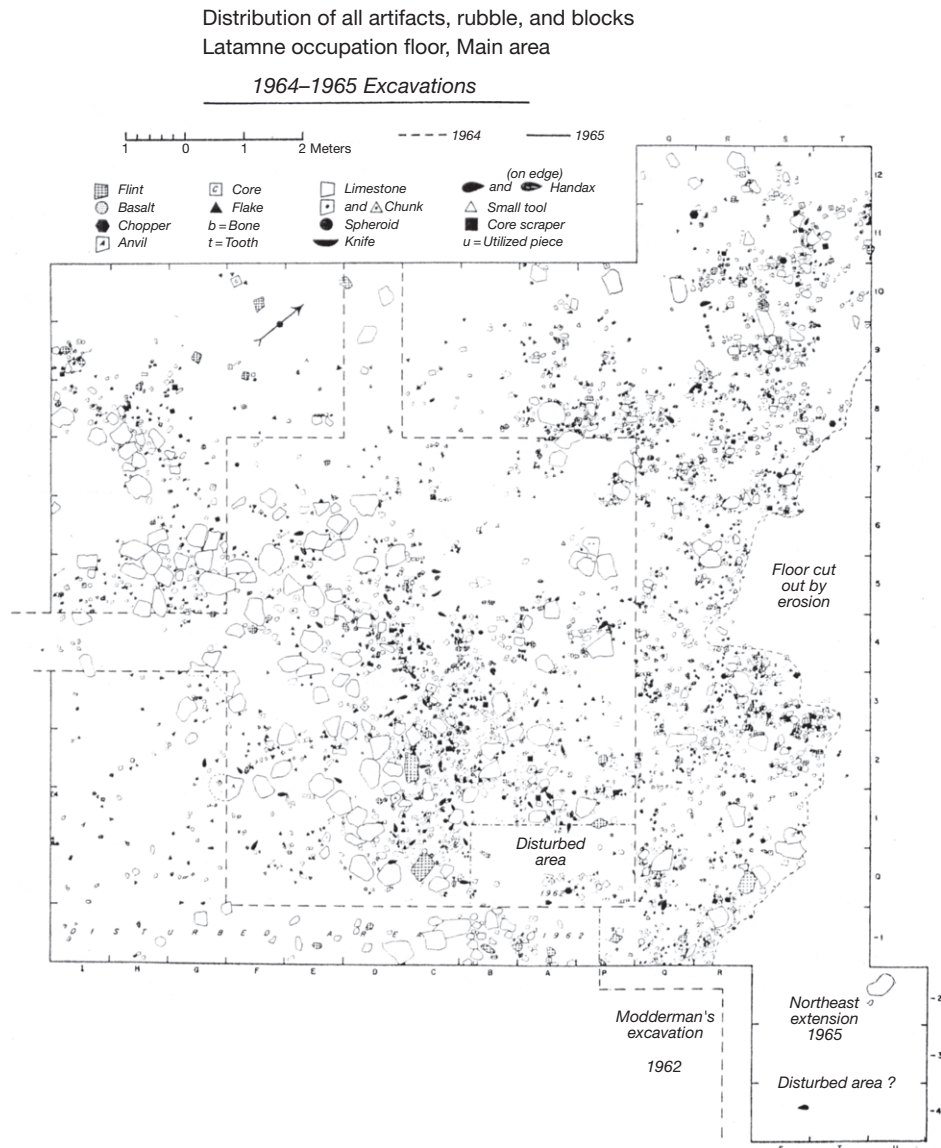


Figure 10 Plan of the 'living floor' at Latamne, Syria. Clark (1969), figure 10.

India

The Indian Lower Paleolithic includes several sites and thousands of bifaces and cleavers, but unfortunately few sites and sequences are dated, and faunal remains are very rarely preserved (Petraglia, 1998). Early middle Pleistocene Acheulean hand axes are known from Dinar and Jalalpur, Pakistan, and Bori, India (Petraglia, 1998) (Figure 11). Apart from a poorly dated but probably Late middle Pleistocene cranial fragment from Hathnora in the Narmada Valley (Cameron et al., 2004), there are no hominin remains from India >30 ka. Nevertheless, a long cave sequence for rock shelter FIII-23 at Bhimbetka from the Acheulean to the Mesolithic shows the persistence of the Acheulean tradition up to ca. 125 ka. Open-air sites are known from Raisen, where more than 90 Early Paleolithic sites were recorded. Best known of all is the remarkable, world-class record from Hunsgi-Baichbal, Karnataka, that contains >200 Acheulean sites (see Figure 12), and a settlement pattern that appears to have been

bipolar, with dry season aggregation camps near water holes, and smaller wet season camps from which plant foods and small game were obtained (Paddayya, 2001). Such is the quality of information from sites such as Isampur that the actions of the hominins that quarried rock to make cleavers can be monitored in detail (Petraglia et al., 2005).

Tajikistan

The loess and paleosol sequence (see Central Asia) of southern Tajikistan, between the Pamirs and Hindu Kush, is another Asian region with an impressive archaeological record (see Figure 13). Sites are always found in interglacial pedocomplexes, a fact that probably reflects the inhospitable nature of the area in glacial periods, as well as the scarcity of stone at times of active loess deposition. The oldest site found so far is Kuldara, in pedocomplex 10–11, and dated by paleomagnetism and correlation with

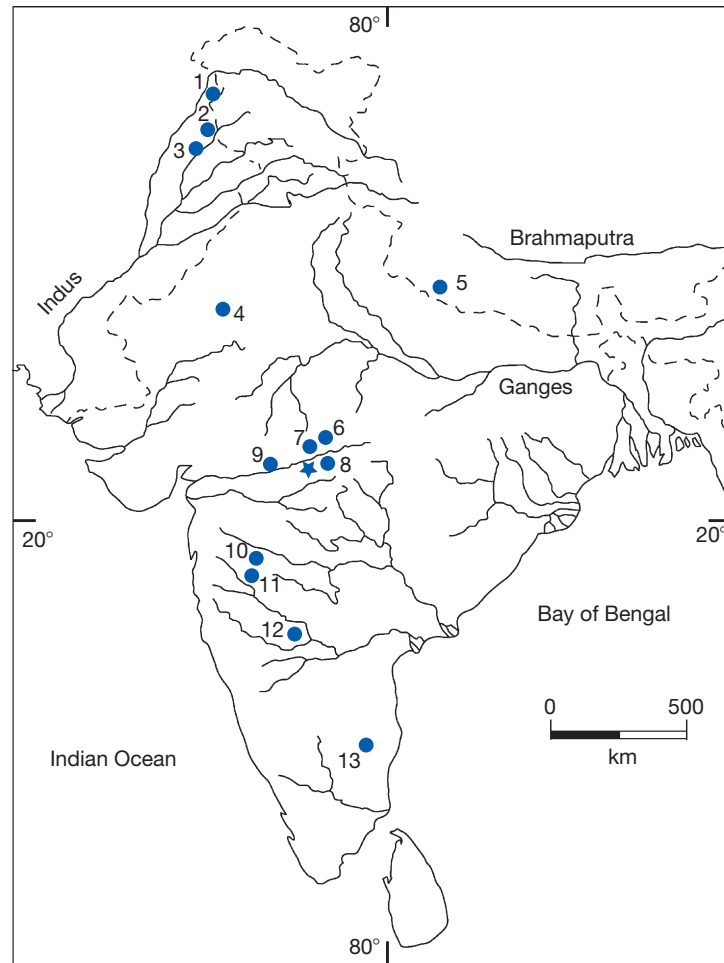


Figure 11 Principal archaeological sites in India. 1, Riwat; 2, Dina; 3, Jalalpur, Singhi Talav; 5, Dang Valley; 6, Raisen Complex; 7, Bhimbetka; 8, Adamgarh; 9, Durkadi; 10, Chirki-Nevasa; 11, Kukdi; 12, Hunsgi-Baichbal Complex; 13, Attirampakkam. The star represents Hathnora.

the marine isotope record to 855–950 ka (see [Figure 14](#)). The stone tools were made from small (typically <5 cm long) chert and metamorphic pebbles, from which small flakes were obtained (see [Figure 15](#)). There was no standardization of tool types, minimal retouch, or core preparation. Given the size of the raw material, it is unsurprising that hand axes were absent. Later sites, also from interglacial pedocomplexes, are known from Karatau, Lakhuti, and Obi-Mazar, and extend the Tajik record into the Upper Pleistocene ([Ranov, 1995](#)).

The Caucasus

In the middle Pleistocene, several caves were formed that were subsequently used for denning by cave bear (the remains of which virtually dominate most faunal sequences) and occasionally by hominins. Examples are Azyk in Azerbaijan, the caves of Kudaro I, III, and Tconia in Georgia, and Treugol'naya in southern Russia. All except Treugol'naya have Acheulean assemblages containing some hand axes and a few cleavers. ([Ljubin and Bosinski, 1995](#)).

China

In eastern Asia, simple (mode 1) core and flake assemblages continued into the middle Pleistocene, and the type of Acheulean bifaces and cleavers that are so widespread in western and southern Asia are rare. Good examples are the sites of Chenjiaowan (970–900 ka) and Maliang (≤ 730 ka) in the Nihewan Basin ([Chen Shen and Wei Qi \(2004\)](#); see also [Figure 4](#)). Here, analysis of refits, wear analysis, and stone reduction sequences show deliberate selection of high-quality stone, and the progressive reduction of cores by centripetal flaking (see [Figures 16 and 17](#)). Although these assemblages can easily be dismissed as 'simple,' their makers nonetheless exhibited considerable skill in their use of flaking techniques and sequences when making tools for cutting, scraping, and processing meat. The best-known site in northern China is the fissure of Zhoukoudian (formerly Choukoutien) near Beijing, where hundreds of fragments of *H. erectus sensu stricto* were discovered before 1939. The 20 m deep section of locality I, where the main discoveries were made, has recently been re-dated to ca. 700 ka ([Guanjun Shen et al., 2009](#)). It is primarily a record of

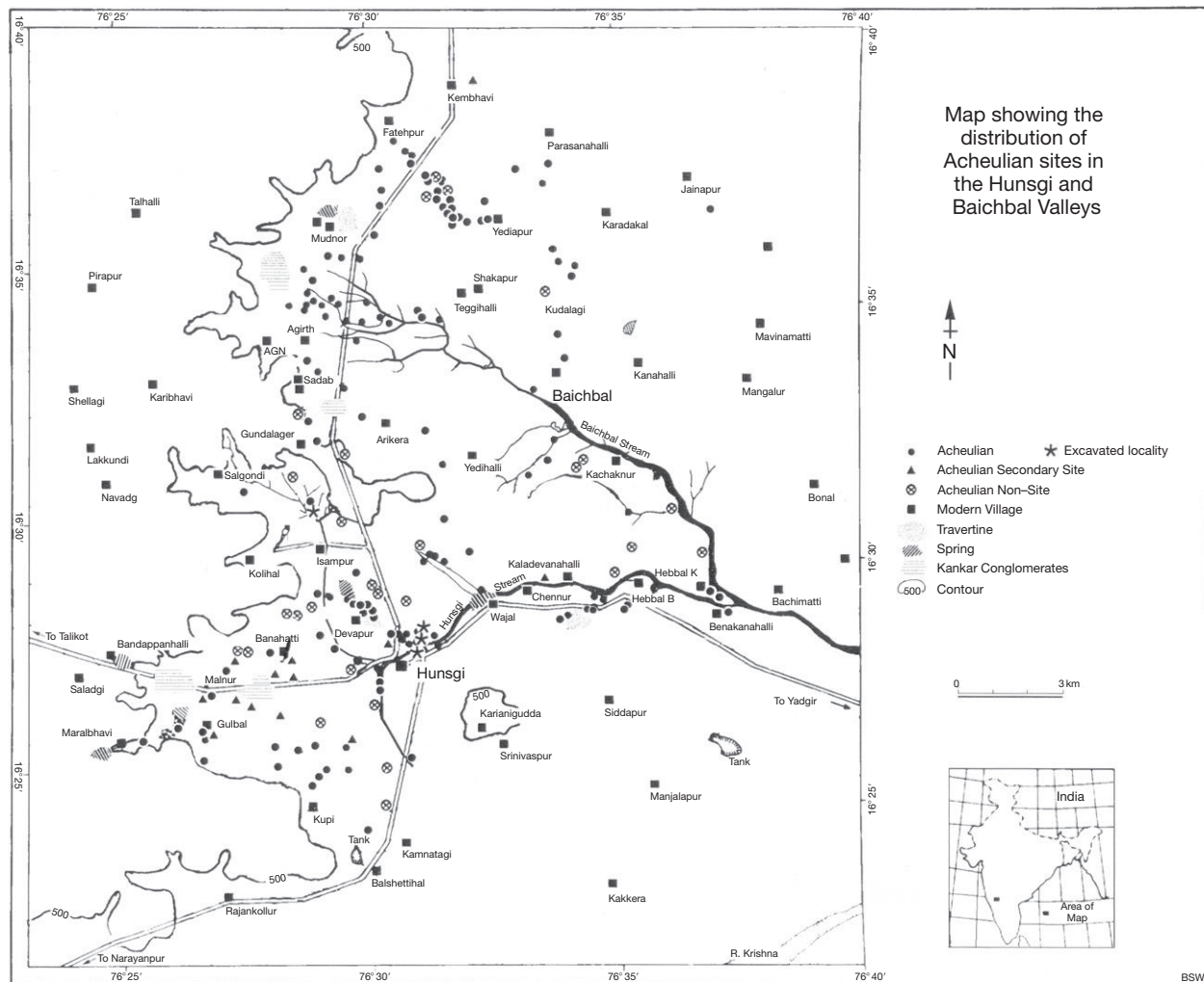


Figure 12 Acheulean sites in the Hunsgi-Baichbal Valleys, India. Paddayya (2001), figure 16.1.

the giant hyena, *Pachycrocuta brevirostris*, which accumulated almost all the other bones, including those of *H. erectus*. Hominins also used the cave (hence their artifacts) but there is no clear evidence that they hunted (Boaz et al., 2000).

The same type of core and flake assemblages are found in South China, but the archaeological record is less clear. Sites such as Chenshan and Jigongshan, and hominin remains from Hexian, Nanjing, and Yunxian show that the Yangtze River south of the Qinling Range was occupied from the Early middle Pleistocene (Wang Youping, 2001). Less is known about the middle Pleistocene record of Indo-China and the southern part of China, which were tropical or subtropical, and had a distinctive *Stegodon-Auluripoda* fauna (see [Mid-Pleistocene of Southern Asia](#)). One of the earliest examples is Bose, South China, where large bifaces have been found, some of which look like Acheulean bifaces; these have been dated to ca. 800 ka based on their association with tektites that were part of the fallout from the Australasian Tektite Shower that occurred at that time (Yamei Hou et al., 2000).

At a continental scale, two issues concerning the middle Pleistocene hominin record of Asia deserve attention.

The Movius Line

One of the most durable aspects of the Asian Paleolithic is the distinction first recognized by Movius (1949) between the Acheulean (mode 2) biface assemblages of Southwest and South Asia, and simpler core and flake (mode 1) assemblages of Southeast Asia and China. (The northern boundary of the Movius Line is less distinct, as the Tajik 'loess Paleolithic' is a simple core and flake industry, but bifaces (currently undated) are known from Iran and Kazakhstan.) Although Acheulean-like bifaces are known from Bose in South China and Korea, they are very rare east of the Ganges, compared with the thousands known from India and the Levant. In South and Southwest Asia, the Acheulean is remarkably homogenous (see [Figure 18](#)). Because it is earlier in Africa, its presence in western Asia is widely regarded as the result of the diffusion of ideas or hominins. If so, it seems that neither succeeded in entering China via a corridor north of the Tibetan Plateau, or the more forested regions of Southeast Asia.

Several reasons have been suggested for this division (Schick, 1994). One suggestion is that hand axes and cleavers

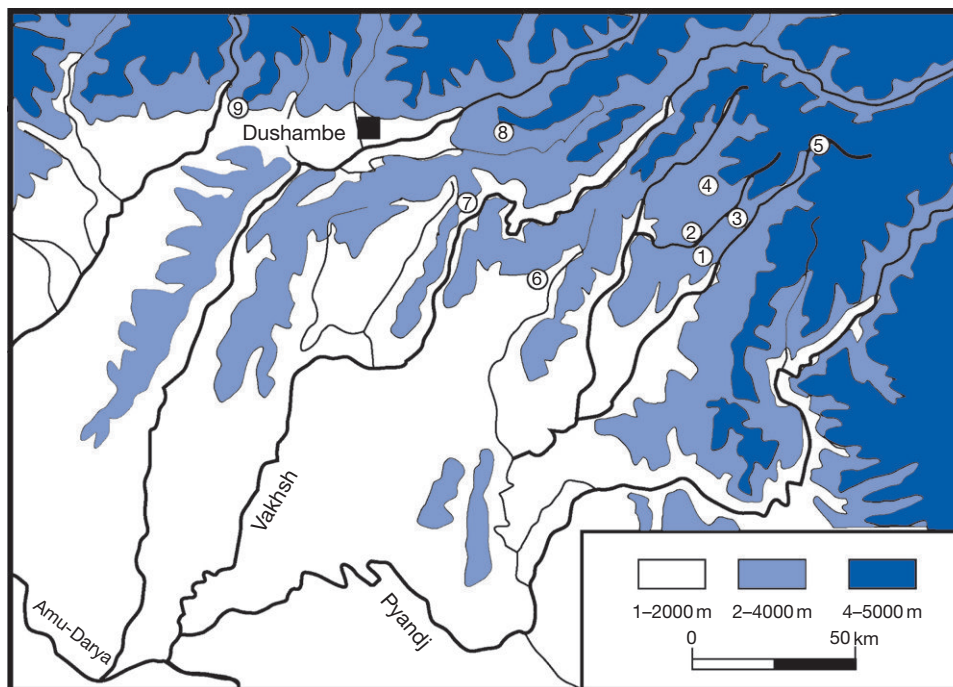


Figure 13 Principal archaeological sites in Tajikistan. Paleolithic sites: 1, Kuldara; 2, Obi-Mazar/Lakhuti; 3, Khanako; 4, Chashmanigar; 5, Shugnou; 6, Ogizkuchlik; 7, Karatau; 8, Karamaidan; 9, Khudji. [Ranov and Dodonov \(2003\)](#), figure 1., pp. 133–147.

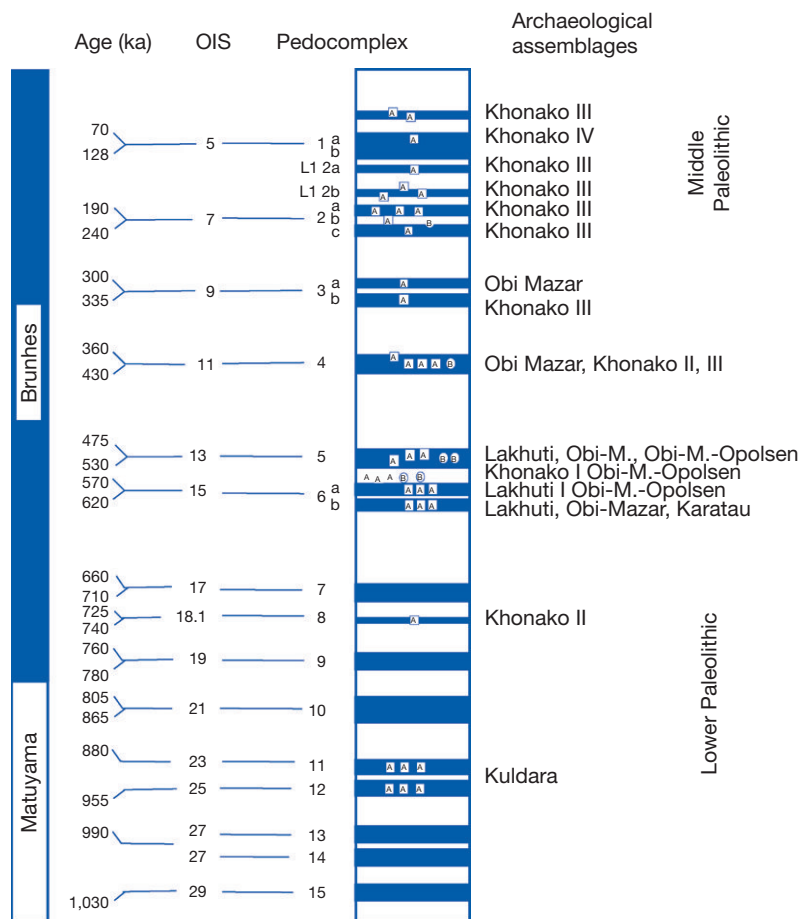


Figure 14 The stratigraphic context and age of sites in the 'loess Paleolithic' of Tajikistan. [Dodonov \(2002\)](#), tables 9 and 14; [Ranov and Dodonov \(2003\)](#), figure 10.

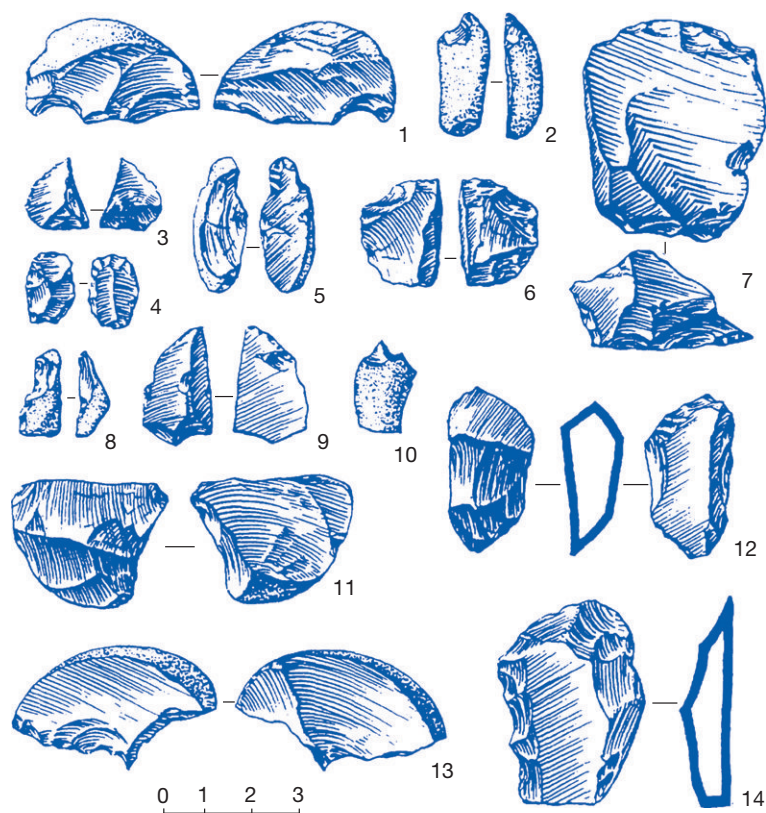


Figure 15 Stone tools from Kuldara, Tajikistan. [Ranov \(1995\)](#), Figure 3.

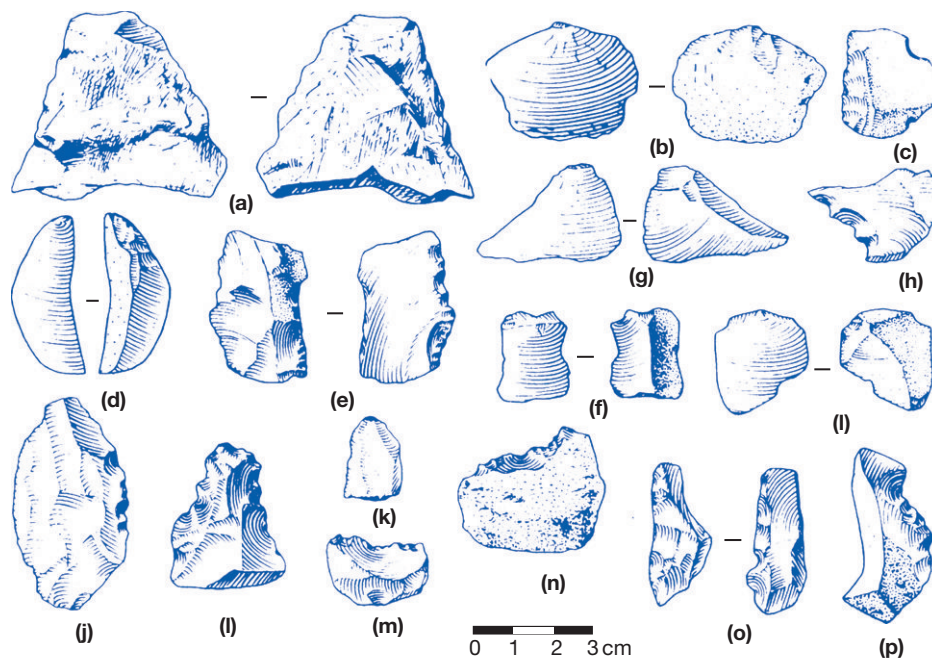


Figure 16 Stone artifacts from Chenjiawan, Nihewan basin, China. C,E,H,I,M, side-scrapers; P, notch; A,B,F,G,I, flake.

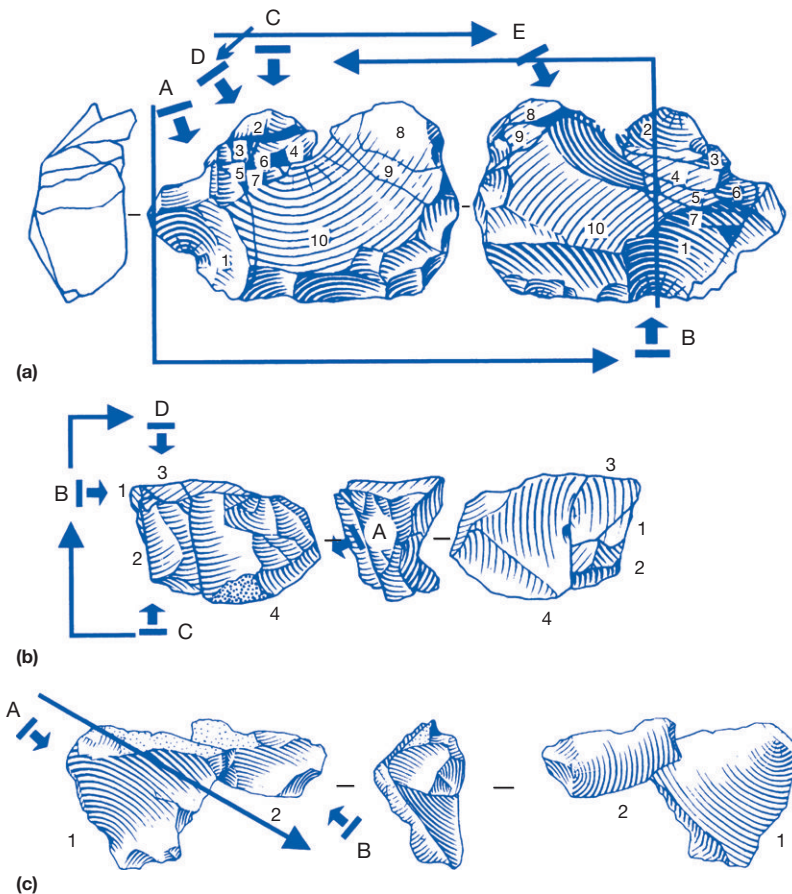


Figure 17 Refitting stone sets from Chenjiawan, Nihewan basin, China. (a) Refitting-set #1-1986; (b) Refitting-set #3-1992; (c) Refitting-set #1-1992. [Chen Shen and Wei Qi \(2004\), figures 2 and 3.](#)

were not needed in SE and East Asia because bamboo was widely available, and highly versatile as a raw material. Others are that the complex rules for making hand axes were not viable in the forested parts of Southeast Asia; or that hominins beyond the Movius Line did not require an overdesigned way of obtaining a cutting or chopping edge. Another reason may be related to the size of foraging groups; if these were small, there would have been fewer hands to carry bulky items such as bifaces as well as any food that had been obtained. However, it is important to point out that none of the non-hand-axe assemblages studied by Movius from Myanmar, mainland and island Southeast Asia are demonstrably contemporaneous with early and middle Pleistocene bifacial assemblages in South and West Asia, so there is in fact little factual evidence underpinning the original concept of the “Movius Line”.

Middle Pleistocene Hominins: Speciation Events and Isolation?

Three, and possibly four types of hominins inhabited Asia in the middle Pleistocene. In eastern Asia, *H. erectus sensu stricto* was the main hominin type (see [Table 1](#)). By ca. 500 ka, its cranial capacity (one of our few direct indicators of cognitive abilities and body size) had increased to within the

modern human range of ca. 1300 cc, and thus it would have had a greater range of skills and abilities with which to survive. Because *H. erectus* is the only type of middle Pleistocene hominin known from China and Indonesia (excluding Flores), it is possible that the increased severity and duration of cold, glacial periods in the middle Pleistocene may have resulted in hominin populations in eastern Asia becoming more isolated than in the early Pleistocene. The same pattern is probably true of western Asia, although its middle Pleistocene fossil hominin record is almost nonexistent. Western Asia may have been inhabited by a different type of hominin from *H. erectus* and possibly by the ancestor of *H. heidelbergensis*, which is best known from Europe and possibly sub-Saharan Africa. A third type are known as Denisovans, and are known primarily from the ancient DNA extracted from a phalanx and tooth from Denisova Cave, Siberia ([Krause et al., 2010](#)). Its origins are unknown and may be African or Asian ([Martín-Torres et al., 2010](#)). Finally, the earliest inhabitants of Flores are unknown. Some researchers regard the late Pleistocene *H. floresiensis* as an extreme example of dwarfing (here, of *H. erectus*) brought about by isolation; others as possibly derived from a very early form of *Homo*. As with so many other aspects of early Asian prehistory, further research is needed.

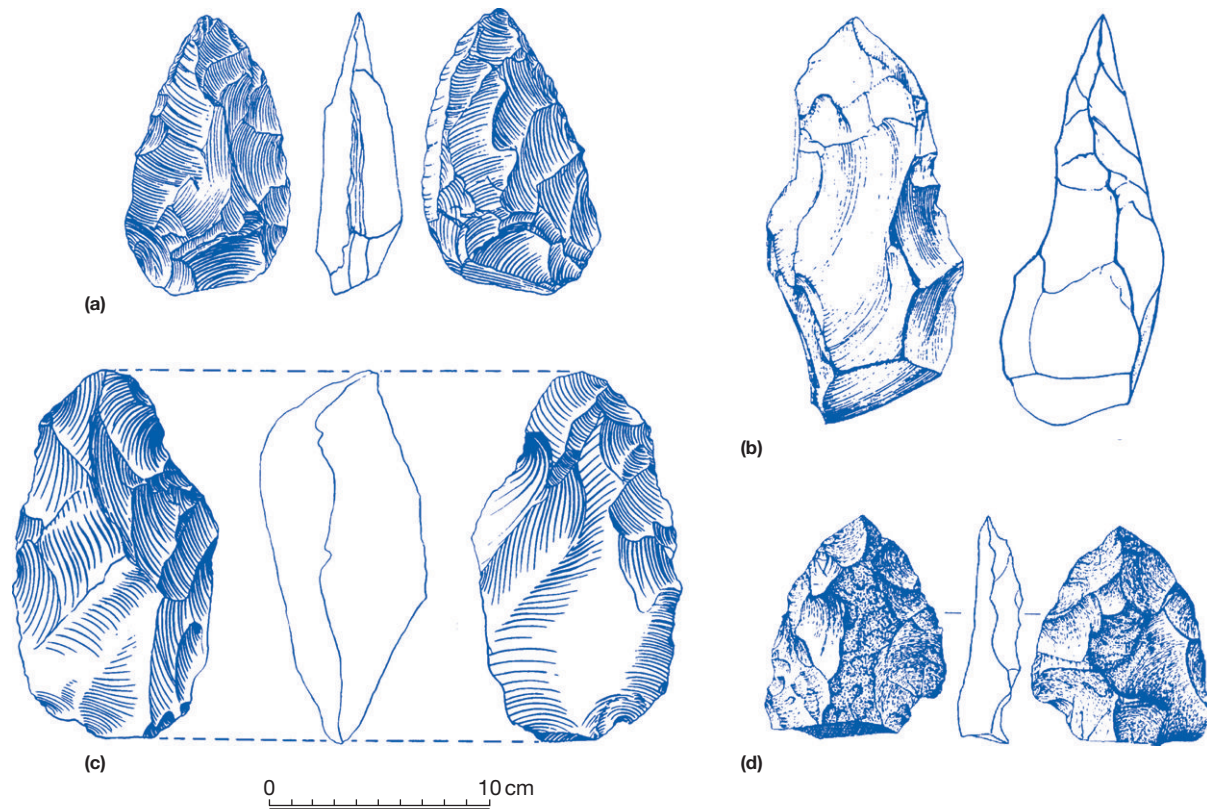


Figure 18 Acheulean bifaces from: (a) Cona cave, Georgia; (b) Evron, Israel; (c) Isampur India; and (d) the Krasnovodsk Plateau, Kazakhstan. Note that biface (d) was broken at the proximal end. Sources: Ljubin and Bosinski (1995), figure 24; Bar-Yosef (1998), figure 8.10; Paddayya (2001), figure 16.12; and Vishnyatsky (1999), figure 2.

See also: **Archaeological Records:** Overview; 2.7 Myr–300 000 Years Ago in Africa; Global Expansion 300 000–8000 Years Ago, Asia; **Glaciation, Causes:** Tectonic Uplift and Continental Configurations; **Glaciations:** Middle Pleistocene in Eurasia; **Lake Level Studies** | Asia; **Loess Records:** Central Asia; **Paleoclimate Reconstruction:** Pliocene Environments; **Vertebrate Records:** Mid-Pleistocene of Southern Asia.

References

- Antón S and Swisher CC III (2004) Early dispersals of *Homo* from Africa. *Annual Review of Anthropology* 33: 271–296.
- Bar-Yosef (1998) Early colonizations and cultural continuities in the Lower Palaeolithic of western Eurasia. In: Petraglia M and Korisettar R (eds.) *Early Human Behavior in Global Context: The Rise and Diversity of the Lower Palaeolithic Record*, pp. 221–279. London and New York: Routledge.
- Boaz NT, Ciochón RL, Qing Xu, and Jinji Liu (2000) Large mammalian carnivores as a taphonomic factor in the bone accumulation at Zhoukoudian. *Acta Anthropologica Sinica* 19(supplement): 224–234.
- Brumm A, et al. (2010) Hominins on Flores, Indonesia, by one million years ago. *Nature* 464: 748–752.
- Cameron D, Patnaik R, and Sahni A (2004) The phylogenetic significance of the Middle Pleistocene Narmada cranium from Central India. *International Journal of Osteoarchaeology* 14: 419–447.
- Chen Shen and Wei Qi (2004) Lithic technological variability of the Middle Pleistocene in the eastern Nihewan Basin, northern China. *Asian Perspectives* 43(2): 281–301.
- Clark JD (1969) The Middle Acheulean site at Latamne, Northern Syria (Second paper). *Quaternaria* 10: 1–60.
- Dennell RW (2003) Dispersal and colonisation, long and short chronologies: How continuous is the Early Pleistocene record for hominids outside East Africa? *Journal of Human Evolution* 45: 421–440.
- Dennell RW (2004) Hominid dispersals and Asian biogeography during the Lower and Early Middle Pleistocene, ca. 2.0–0.5Mya. *Asian Perspectives* 43(2): 205–226.
- Dennell, RW (2009) *The Palaeolithic Settlement of Asia*. Cambridge University Press.
- Dennell RW and Roebroeks W (2005) Out of Africa: An Asian perspective on early human dispersal from Africa. *Nature* 438: 1099–1104.
- Dennell, RW (In Press) Hominins, deserts and the hominin colonization and settlement of continental Asia. *Quaternary International*.
- de Lumley H, de Nioradzé M, Barsky D, et al. (2005) Les industries lithiques préoldowayennes du début du Pléistocène inférieur du site de Dmanissi en Géorgie. *L'Anthropologie* 109(1): 1–182.
- Dominguez-Rodrigo M (2002) Hunting and scavenging by early humans: The state of the debate. *Journal World Prehistory* 16(1): 1–54.
- Dodonov A (2002) *Quaternary of Middle Asia: Stratigraphy, Correlation and Paleogeography*. Moscow: Geos. (In Russian).
- Ferring R, et al. (2011) Earliest human occupations at Dmanisi (Georgian Caucasus) dated to 1.85–1.78 Ma. *Proceedings of the National Academy of Sciences USA* 108: 10432–10436.
- Gaudzinski S (2004) Early hominid subsistence in the Levant: Taphonomic studies of the Plio–Pleistocene 'Ubeidya Formation (Israel). In: Goren-Inbar N and Speth JD (eds.) *Human Paleoeecology in the Levantine Corridor*, pp. 75–88. Oxbow Books.
- Goren-Inbar N, Sharon G, Melamed Y, and Kislev M (2002) Nuts, nut-cracking, and pitted stones at Geshert Benot Ya'aqov, Israel. *Proceedings of the National Academy of Sciences of the United States of America* 99(4): 2455–2460.
- Goren-Inbar N, Alpers N, Kislev ME, et al. (2004) Evidence of hominin control of fire at Geshert Benot Ya'aqov, Israel. *Science* 30: 725–727.
- Guanjun Shen, Xing Gao, Bin Gao, and Granger DE (2009) Age of Zhoukoudian *Homo erectus* determined with $^{26}\text{Al}/^{10}\text{Be}$ burial dating. *Nature* 458: 198–200.
- Krause J, et al. (2010) The complete mitochondrial DNA genome of an unknown hominin from southern Siberia. *Nature* 464: 894–897.

- Kroon D, Alexander I, Little M, et al. (1998) Oxygen isotope and sapropel stratigraphy in the eastern Mediterranean during the last 3.2 million years. *Proceedings of the Ocean Drilling Program, Scientific Results* 160: 181–189.
- Larick R, et al. (2001) Early Pleistocene $^{40}\text{Ar}/^{39}\text{Ar}$ ages for Bapang Formation hominins, Central Java, Indonesia. *Proceedings of the National Academy of Sciences USA* 98: 4866–4871.
- Ljubin VP and Bosinski G (1995) The earliest occupation of the Caucasus region. In: Roebroeks W and van Kolfschoten T (eds.) *The Earliest Occupation of Europe*, pp. 207–253. University of Leiden Press.
- Lu Huayu, Liu X, Zhihang F, An Z, and Dodson J (1999) Astronomical calibration of loess-paleosol deposits at Luochuan, central Chinese Loess Plateau. *Palaeogeography, Palaeoclimatology, Palaeoecology* 154: 237–246.
- Martinón-Torres M, Dennell RW, and Bermudez de Castro JM (2011) The Denisova hominin need not be an African story. *Journal of Human Evolution* 60(2): 251–255.
- Morwood MJ, O'Sullivan PB, Aziz F, and Raza A (1998) Fission-track ages of stone tools and fossils on the east Indonesian island of Flores. *Nature* 392: 173–176.
- Morwood MJ, et al. (2003) Revised age for Mojokerto 1, an early *Homo erectus* cranium from East Java, Indonesia. *Australian Archaeology* 57: 1–4.
- Movius H (1949) The lower Palaeolithic cultures of southern and eastern Asia. *Transaction of the American Philosophical Society* 38(4): 329–420.
- Paddayya K (2001) The Acheulean Culture Project of the Hunsgi and Baichbal Valleys, peninsular India. In: Barham L and Robson-Brown K (eds.) *Human Roots: Africa and Asia in the Middle Pleistocene*, pp. 235–258. Western Academic and Specialist Press.
- Pappu S et al. (2011) Early Pleistocene presence of Acheulian hominins in South India. *Science* 331, 1596–1599.
- Petraglia MD (1998) The Lower Palaeolithic of India and its bearing on the Asian record. In: Petraglia M and Korisettar R (eds.) *Early Human Behavior in Global Context: The Rise and Diversity of the Lower Palaeolithic Record*, pp. 343–390. Routledge.
- Petraglia MD, Shipton C, and Paddayya K (2005) Life and mind in the Acheulean: A case study from India. In: Gamble C and Porr M (eds.) *The Hominid Individual in Context: Archaeological Investigations of Lower and Middle Palaeolithic Landscapes, Locales and Artefacts*, pp. 197–219. Routledge.
- Presnyakov SI, et al. (2012) Age of the earliest Paleolithic sites in the northern part of the Armenian Highland by SHRIMP -II U-Pb geochronology of zircons from volcanic sites. *Gondwana Research* 21: 921–938.
- Rabinovitch R, Gaudzinski-Windheuser, S, Kindler, L and Goren-Inbar, N. (2012) *The Acheulian site of Gesher Benot Ya'aqov. Volume 3: mammalian taphonomy, the assemblages of Layers V-5 and V-6*. Dordrecht, Heidelberg, London & New York: Springer Science+Business Media.
- Ranov VA (1995) The 'Loessic Palaeolithic' in South Tadzhikistan, Central Asia: Its industries, chronology and correlation. *Quaternary Science Review* 14: 731–745.
- Ranov VA and Dodonov AE (2003) Small instruments of the Lower Palaeolithic site Kuldara and their geoarchaeological meaning. In: Burdukiewicz JM and Ronen A (eds.) *British Archaeological Reports (International Series) 1115: Lower Palaeolithic Small Tools in Europe and Asia*, 133–147.
- Rightmire GP, Lordkipanidze D, and Vekua A (2006) Anatomical descriptions, comparative studies and evolutionary significance of the hominin skulls from Dmanisi, Republic of Georgia. *Journal of Human Evolution* 50: 115–141.
- Schick KD (1994) The Movius Line reconsidered: Perspectives on the Earlier Paleolithic of Eastern Asia. In: Corruccini R and Ciochon S (eds.) *Integrative Paths to the Past: Paleoanthropological Advances in Honor of F.Clark Howell*, pp. 569–594. Prentice Hall.
- Smith JMB (2001) In: Smith JMB, Morwood M, and Davidson I (eds.) *Did early hominids cross sea gaps on natural rafts? In Faunal and Floral Migration and Evolution in SE Asia-Australia*, pp. 409–416. Lisse, Netherlands: Swets & Zeitlinger.
- Stiner M, Gopher A, and Barkai R. (2010) Hearthside socioeconomics, hunting and paleoecology during the late lower Paleolithic at Qesem Cave, Israel. *Journal of Human Evolution*.
- Vishnyatsky LB (1999) The paleolithic of Central Asia. *Journal of World Prehistory* 13(1): 69–122.
- Wilson RCL, Drury SA, and Chapman JL (2000) *The Great Ice Age*. Routledge.
- Yamei Hou, Potts R, Baoyin Y, Zhengtang G, Deino A, and Wei W (2000) Mid-Pleistocene Acheulean-like stone technology of the Bose Basin, South China. *Science* 287: 1622–1626.
- Youping Wang (2001) The Middle Pleistocene archaeology of the valley of the Yangtze River. In: Barham L and Robson-Brown K (eds.) *Human Roots: Africa and Asia in the Middle Pleistocene*, pp. 149–157. Western Academic and Specialist Press.
- Zhu R, Zhinsheng An, Potts R, and Hoffman KA (2003) Magnetostratigraphy of early humans in China. *Earth-Science Reviews* 61: 341–359.
- Zhu RX, Potts R, Xie F, et al. (2004) New evidence on the earliest human presence at high northern latitudes in northeast Asia. *Nature* 431: 559–562.

1.9 Myr–300 000 Years Ago in Europe

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Glossary

Acheulean: Name given to an assemblage, or assemblage type, characterized by the presence of handaxes or thinning flakes. First identified at St. Acheul, Abbeville, France.

Assemblage: A collection of artifacts, usually stone tools, but can refer to any preserved material, that is united by a common feature or set of related features. This commonality may be the presence or absence of diagnostic tools. It may, on the other hand, be a particular spatial or temporal boundary that marks the collection of artifacts as discreet. For example, a collection of stone tools from a site/layer which lack a particular type of stone tool might be grouped together and characterized by that absence; alternatively, the layer, or restriction to a part of the site, might impart distinctiveness or discreteness.

Assemblage type: A group of assemblages that share a common feature or features.

Composite tools: Tools made up of two or more different components, often of two different substances. For example, a wooden or antler shaft, with a stone tip.

Cores: A nodule or river cobble of rock that has been shaped by knapping. Flakes have been removed by direct percussion.

Denticulate: A particular type of stone tool usually made on a flake. The edge is retouched to form a ragged and uneven edge when seen in profile. Often described as saw-toothed.

Direct percussion: The process of striking a block of raw material directly with a percussor or hammer (hard or soft).

Fitness indicator: A physical attribute of an animal that stands proxy for a statement on the physical health, genetic quality, social standing, and desirability as a mate of an individual (phenotype).

Flaked flake: A particular type of stone tool made on a flake. This is a flake that has had another flake removed from any edge or off either upper or lower face of the flake. Sometimes the concavity left by the detachment was used, sometimes the spall was used, sometimes both were.

Flakes: The portions of raw material that are detached from a core during knapping by direct percussion.

Formation: A single sedimentary layer, or group of related sedimentary/rock strata having a common feature that unites them. For example, a river terrace may be thought of as a formation, it may be made of one layer of gravel, but it is more likely to be made of a number of different gravel layers interspersed with other layers of sand, silt, or clay.

HandAxe: A stone tool made by shaping a block of raw material, cobble, or big flake. The tool is bifacial – knapped on both upper and lower faces, and the whole is thinned and shaped to a number of different outline shapes. The earliest

are about 1.7 Myr from Africa, and they continued to be in use until the end of the Middle Paleolithic. The process of thinning the handaxe produces a characteristic type of flake – the thinning flake.

HandAx assemblage: An assemblage of stone tools from one layer at a site, as from a number of related layers which contains handaxes or thinning flakes, or both.

Hard hammer or Hammer stone: A pebble or cobble of rock used in knapping by direct percussion, usually cores, but also in the early stages of roughing out a handaxe.

Hominin: A term replacing the older word hominid. It has been adopted because of the genetic work which demonstrates that humans, their ancestors, and related genera, are closer genetically to chimpanzees. But gorillas separated from this group of related genera a long time ago. The old use of the word hominid included gorillas with the human family on the basis that their skeletons showed marked similarities.

Knapping: The name given to the process of controlled breakage of a block of raw material. The process may be employed to shape a block of raw material into a tool. Alternatively, it may be to use the flakes that come off during the process of knapping to make tools from.

Levallois/Prepared core technology (PCT): A way of knapping a block of raw material that carefully shapes and prepares one surface of that block. The resulting core, with its carefully prepared surface, can be used to detach specific types/shapes of stone tool, depending on the particular type of surface preparation applied to the core. The first appearance of Levallois is taken to mark the beginning of the Middle Paleolithic. It may occur at different times in different regions.

Microfauna: Small mammals, such as small rodents.

Middle Pleistocene transition (MPT): The period of time it took for the Earth's orbit to shift from a more circular path around the Sun to a more elliptical one. Prior to the MPT, the glacial–interglacial cycle was dominated by the precessional cycle. The rotation of the Earth around its axis wobbles like a spinning top. The 41-ka precessional cycle is the length of time it takes to complete one rotational wobble. After the MPT, the orbital pattern was more elliptical, and at this time the amplitude and wavelength of the glacial–interglacial cycle changed. The cycle is now more closely tuned to a 100-ka pattern. But whether the orbital eccentricity has been the major factor in causing this change remains unclear.

Non-handAx assemblage: A site, or group of related layers from a site layer from a site, where the stone tool assemblage lacks hand axes or the evidence of their manufacture – the thinning flakes.

Notch: A particular kind of stone tool usually made on a flake. A concavity is made on the edge of a flake by small contiguous retouch removals.

Percussor: A hammer such as a hard hammer stone or a soft hammer such as one made out of bone, antler, or wood.

Retouch or Retouched tools: The process of shaping an edge, usually a flake, in order to make a hand-held tool on a flake. Retouch is usually confined to the flake's edge, and only shapes or sculpts the edge.

Scraper: A particular type of stone tool usually made on a flake. A length of flake's edge is shaped by retouch to make a long continuous retouched edge with a smooth profile.

Soft hammer: Usually made of wood, bone, or antler. These are used in direct percussion in the Lower Paleolithic for the manufacture of handaxes. They are used in thinning and shaping a handax as they take off much less volume of surface raw material during the thinning process than a hammer stone. It is possible that soft rock types can also be used as soft hammers.

Introduction

The middle Pleistocene transition (MPT) represents the period of time between ca. 942 and 641 ka when the Earth's orbit shifted from a more circular to a more elliptical path around the Sun. After this time, the amplitude and duration of Northern Hemisphere glaciations changed markedly when compared to those before the transition (see [Figure 1](#); [Lisiecki, 2005](#); [Lisiecki and Raymo, 2005](#)). Against this climatic backdrop, the hominin colonization and settlement of southern and northern Europe occurred. The MPT makes a useful framework through which this process event can be viewed.

The Pattern of Hominin Colonization of Western Europe

Prior to the MPT

The earliest occupants of Western Europe appear to have arrived either from Africa or from Eurasia. In the former case, they may have represented *Homo ergaster*, or some other small brained species of African *Homo*, moving up the African Rift Valley and its Levantine extension. At the so-called Gates of Europe, the site of Dmanisi, in the Caucasus, represents the earliest Eurasian hominin site with stone tools so far discovered outside of Africa. The site is dated to ca. 1.7 Myr, and the stone tool assemblage was one of cores and flakes – see [Table 1](#). Some of the hominins may have been the prey of hyaena, as a number of the fossils and artifacts were recovered from the infill of their dens. Rightmire and colleagues ([Rightmire et al., 2006](#)) suggest there may be two hominin populations at this site, *H. ergaster* and a large hominin known from a single jaw, *H. georgicus*. Whether this latter was an African hominin, a purely Eurasian species, or a Eurasian descendant of an earlier African migrant remains to be determined. We do not know which hominin was responsible for the manufacture of the stone tools.

South of the Caucasus there are a number of early sites. In the Erq-el-Ahmar Formation, in the Jordan Valley, Israel, sediments are dated to the Olduvai magnetic event at 1.96–1.78 Myr ([Ron and Levi, 2001](#)). Other sediments, some 1.5 km away, but in the same formation, contain cores and flakes. Some researchers are not convinced there are artefacts. Ubeidiya, also in the Jordan Valley, is dated to ca. 1.4 Myr ([Bar-Yosef and Goren-Inbar, 1993](#)). Here a series of handax and non-handax assemblages alternate with each other.

A non-handax assemblage is present at Bizat Ruhama on the southern coastal plain of Israel at about 1.0 Myr.

Another possible crossing point for hominins into western Europe is from the North African coast via the Straits of Gibraltar. Core- and flake-using hominins, presumably *H. ergaster*, were certainly present in North Africa, at sites such as Ain Hanech, at 1.8 Myr ([Sahnouni et al., 2002](#)). The earliest sites in western Europe are the Barranco León 5 and Fuente Nueva 3 localities in the Guadix–Baza basin (previously the Orce sites; [Gibert et al. \(2001\)](#)) in Spain. These have been redated on paleomagnetic, but mostly on faunal, grounds to slightly younger than Dmanisi, perhaps ca. 1.6 Myr in age ([Gibert et al. \(2006\)](#), and references therein). These sites contain core-and-flake assemblages. Also in Spain is a small collection of cores and flakes from levels E9 to E14 at the Sima del Elefante, in the Sierra Atapuerca. They are certainly older than the Brunhes/Matuyama boundary at ca. 780 ka and may be close to ca. 1.0 Myr as indicated by microfauna ([Parés et al., 2006](#)). Contemporary with this, or perhaps slightly younger, are a series of sites at Ceprano, Italy. Small collections of cores and flakes have been recovered and dated between 980 ka and 1.0 Myr.

Whichever species is ultimately shown to be the original colonizer of Western Europe, current evidence does not support an extensive initial migration. [Dennell \(2003\)](#) has argued that first appearances may be deceptive; hominin occupation may represent, in these cases, brief incursions into new areas. Bar-Yosef and Belfer-Cohen have noted that a successful migrant does not always make a successful colonist ([Bar-Yosef and Belfer-Cohen, 2001](#)). The paucity of early sites before the MPT in Western Europe, compared to the stronger hominin signal to the east of the region, remains an enigma (see [Dennell \(2003\)](#) and [McNabb \(2005\)](#) for reviews and references therein). Do the Spanish/Italian data imply a limited, or ultimately unsuccessful, one-off colonization event over the Gibraltar Straits? We do not know.

Across the MPT

Once again a hominin presence on the eastern margins of western Europe is attested to by sites in the Caucasus and in the Levantine Rift Valley. Gesher Benot Ya'akov (GBY) is a handax site on the banks of the Jordan River dated to ca. 780 ka ([Goren-Inbar et al., 2000](#)), and notable for early evidence of the use of fire. The strong similarity between its handax and cleaver component, and those found at contemporary African

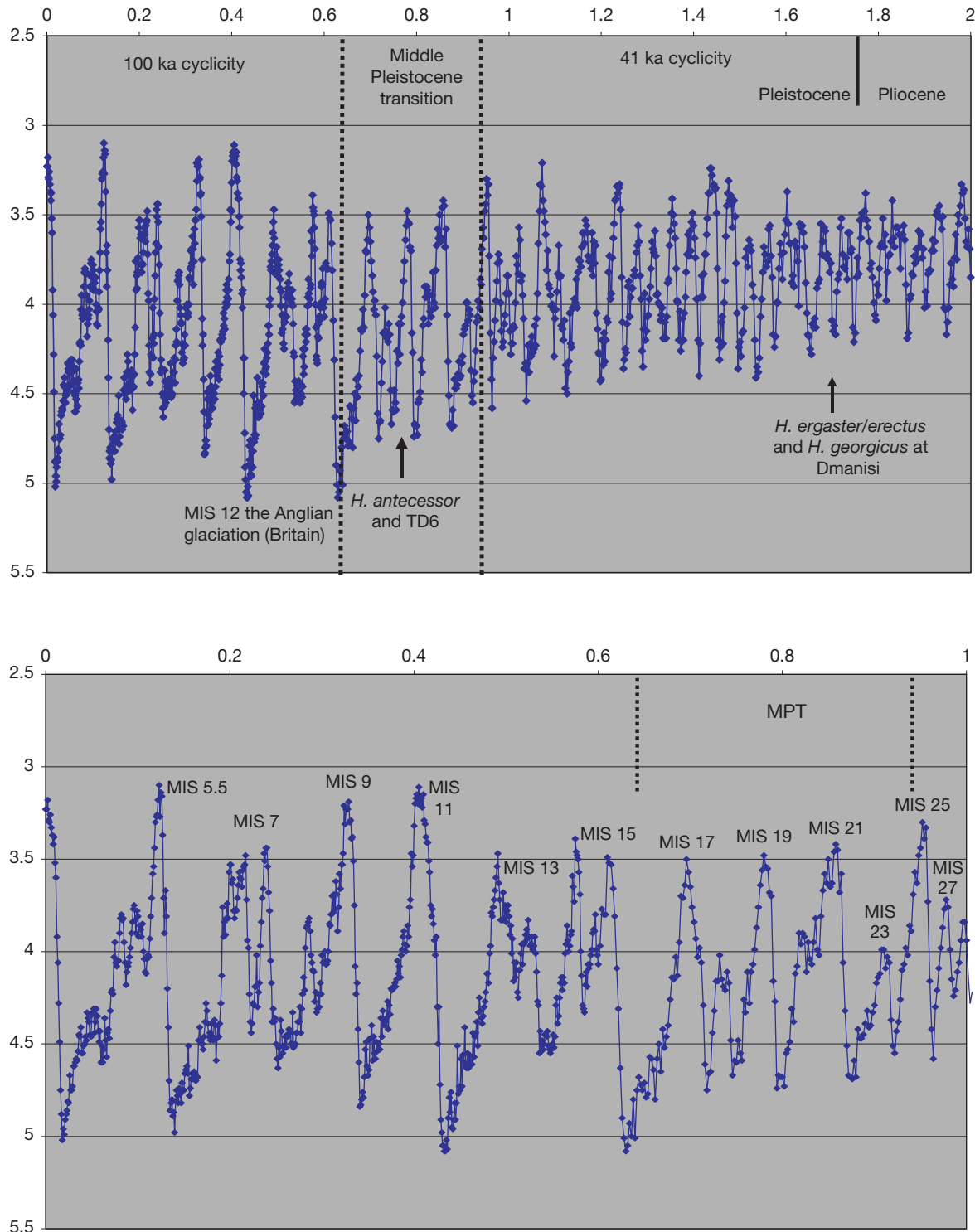
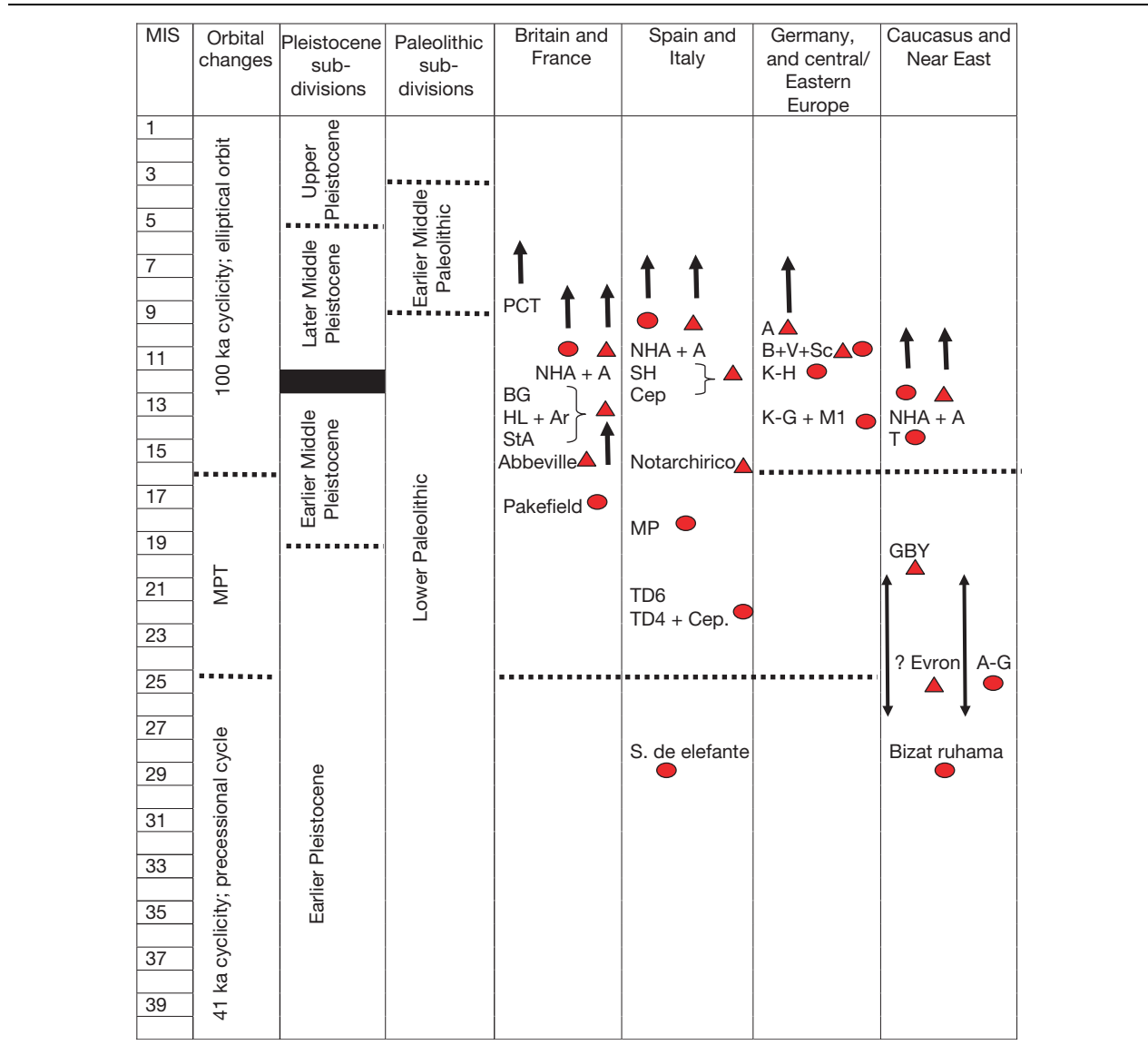


Figure 1 Marine Isotope Stages (MIS) and climate curve from combined data stack (LR04) of 57 different records from a global distribution. Data from Lisiecki 2005, and Lisiecki and Raymo 2005. Data used with permission. Horizontal scale is time measured in millions of years (Myr). Vertical scale is delta ¹⁸O record of benthic Foraminifera. Upper diagram shows overall climate change from Dmanisi (1.7 Myr) until the Anglian glaciation (MIS 12) of Britain, herein taken as the boundary between the earlier middle Pleistocene and the later middle Pleistocene – see table 1. The upper diagram makes the point that after the shift to orbital eccentricity (period of change in orbit = middle Pleistocene Transition - MPT), glaciations are colder, and interglacial peaks are warmer. In the lower diagram the climate curve is depicted for the last 1.0 Myr. MIS have been added.

Table 1 Selected stone tool assemblages, and trends in stone tool assemblage types from Europe between 1.0 mya (MIS 28) and the advent of Levallois/prepared core technology (PCT) marking the beginning of the Middle Palaeolithic in Europe. Key to table: NHA = non-handaxe assemblages; A = Acheulean (handaxe) assemblages; BG = Boxgrove; HL = High Lodge; Ar = Arago; StA = St Acheul; SH = Sima de los Huesos; Cep = Ceprano; MP = Monte Poggiolo; TD6 = Atapuerca Trincheras Dolina 6; TD 4 = Atapuerca Trincheras Dolina 4; S. de Elefante = Atapuerca Sima del Elefante; B = Bilzingsleben; V = Vértesszöllös; Sc = Schöningen; K-H = Kärlich H; K-G = Kärlich G; M1 = Miesenheim 1; T = Treugol'naya Cave; GBY = Gesher Benot Ya'aqov; A-G = Amiranis-Gora (Akhalkalaki). Dotted lines indicate period boundaries and their continuation. Double headed black arrows indicate potential date range for individual sites. Single headed black arrows indicate continuation over time. For sources see text citations and references



Acheulean sites, has led some researchers to suggest GBY was a stopping-off point for African handaxe-making hominins migrating up the rift valley. At the Evron Quarry, on the coastal plain of western Galilee, in this case situated away from the rift valley, there are hand axes (found out of context) which date anywhere between the Brunhes–Matuyama boundary and possibly 1.0 Myr (best estimates suggest the earlier end of the range). To the north there is hominin occupation in the Caucasus at sites such as Amiranis-Gora (Akhalkalaki), an open-air site with a core-and-flake assemblage dated on faunal grounds to the Late-early Pleistocene/earlier-middle

Pleistocene transition (see [Table 1](#); see papers in [Lordkipanidze et al. \(2000\)](#)).

There are three, and possibly four, occurrences in southern western Europe, where archaeological sites dating to the MPT are known: two in Spain and two in Italy. The two Spanish examples represent different layers at the same site, the Trincheras Dolina in the Sierra Atapuerca. The lower of the two levels, TD4, represents a small collection of cores and flakes, made on local quartzite, and dated to MIS 22. The higher layer, TD6, is a more significant occurrence ([Bermúdez de Castro et al., 2004](#)). Over 300 stone tools, but no hand axes,

have been discovered. Many of them have been retouched to form scrapers or denticulates. Associated with these artifacts are specimens of a new hominin type, *H. antecessor*. Just as with the Dmanisi hominins, this may be an African migrant, the descendant of one, or a purely Eurasian species, we do not as yet know. To date, no examples of this species have been found outside of southern Europe. Cut marks on the *H. antecessor* bones suggest that this species was a cannibal. Two Italian sites are dated to the MPT. At Ceprano, already noted to have core-and-flake assemblages near the MPT boundary, a skull cap was discovered higher up in the valley sequence than the level of the archaeological sites (Manzi, 2004). Unaccompanied by stone tools, the skull cap may represent a portion of the cranium of an adult *H. antecessor* (Manzi, 2004). Monte Poggiolo, in Italy, is suggested to date from this period; however, uncertainties concerning its radiometric dating, and the clear association of the artifacts with securely dated deposits, have left doubts hanging over this substantial early core-and-flake assemblage (Villa, 2001).

To the north of the Alps, only one site has been securely dated to the MPT. This is the site of Pakefield, Suffolk, England (Parfitt et al., 2005). Flakes and cores have been discovered in sediments of the Cromer Forest-bed Formation. Sediments from this section of the formation date to at least ca. 700 ka (MIS 17). The site clearly demonstrates that earlier phases of the earliest occupation of Europe were not confined to the south of Europe. Associated fauna suggest a warm Mediterranean-like climate.

After the MPT

One thing is clear from Table 1: it is only after the shift to the 100-ka cycle that hand axes appear in the archaeological record of western Europe with any frequency. From this time on, the only hominin present in Europe is *H. heidelbergensis*, whose evolutionary trajectory ultimately led to the Neanderthals. It is tempting to associate the appearance of *H. heidelbergensis* and hand axes with the shift to the 100-ka periodicity. Was the climatic change sufficient to open up previously closed environmental niches and so initiate the movement of animals, and the hominins that preyed on them, into new habitats? Did *H. heidelbergensis* migrate into Europe in one wave, or was it a series of migrations? How was it that this species seemed better able to survive European mosaic environments than *H. antecessor*? Within its suite of adaptive capabilities, what allowed it to cope with European seasonality, and to overcome the difficulties that accompanied overwintering in European environments? At present, we cannot answer these questions. See Roebroeks (2001) for a summary and review.

Throughout western, central, and eastern Europe, assemblages with and without handaxes have been found across a variety of locally variable open-air and cave sites. At present, there is no reason to doubt these were all made by groups of *H. heidelbergensis*. The handax sites are all subsumed under the label Acheulean. The non-handax sites are given a series of names reflecting either regionally significant cultural patterns, or local variations on the non-handax assemblage type. It is a moot point whether these non-handax sites represent distinct cultural phenomena, the result of extended lineages of intergenerational learning, or whether they represent locally variable responses by hominins who normally made hand axes.

For a good review of the Europe-wide variability in middle Pleistocene stone tool assemblages, see Roebroeks and van Kolfschoten (1995).

The frequency of stone tool assemblages in western, central, and eastern Europe, which date to the Later middle Pleistocene (post-MIS 12 glaciation), suggest an increase in population, though the poor degree of survival of older sites should be factored into such interpretations. From the point of view of the stone tool assemblages, however, there is little change in the character of what was made until the wide-scale adoption of Levallois/prepared core technology (PCT), possibly at the end of MIS 9. This date of ca. 300 ka, and the adoption of PCT, is taken to be the end of the Lower Paleolithic and the beginning of the Middle Paleolithic – see Table 1. In some areas, the beginning of the Middle Paleolithic may have been earlier. Many scholars see a link between the handax phenomenon and the rise of Levallois. French archaeologists in particular can track nascent tendencies toward Levallois-like PCT throughout this period (Orgnac 3 (Ardèche) and Baume Bonne (Quinson, Alpes de Haute-Provence; see papers in Tuffreau (1996). Whether this implies geographically distinct multiple local inventions, or alternatively the widespread diffusion of ideas or Levallois-using hominin groups, remains unclear.

Social Construction

Archaeology reveals very few clues that relate to the construction of middle Pleistocene society, by which I mean demonstrating the nature of relationships between individuals, and between individuals and larger groups at a series of hierarchical levels (e.g., personal core group, hunting coalition, family group, whole group, etc.).

Currently, we can only speculate on the nature of relationships between individuals. Artifact analysis may shed some light on these relationships, given the assumption that social signals are encoded within the making, character, and use of material culture. A number of scholars have argued that very large and symmetrical hand axes have little practical use. Rather, they represent social signaling devices aimed at attracting mates (Kohn and Mithen 1999) – in effect the hominin males' substitute for the peacock's tail; as such, they would represent an individual fitness indicator. Much work remains to be done on this aspect of material culture studies.

Gamble (1999) has argued that hominins during the Lower Paleolithic lived in small groups, moving around familiar territories from which they rarely, if ever, strayed. Their knowledge of that territory was acquired through movement along well-known paths and trails. This world, and the hominins' comprehension of it, was dominated by the local and the familiar. Because of a limited linguistic capacity, relations were maintained with individuals via visual signals and the familiarity of often-repeated socially sanctioned gestures and activities. The local group represented the totality of the social world, and contact with other groups in other territories would have been discouraged. Gamble uses the transport of raw stone materials, and finished artifacts whose rock type can be traced to their source, as evidence of the extent of group territories.

Essential to identifying any form of social construction is the presence of language and of the ability to correctly interpret

the intentions and actions of others. Dunbar (2004) has argued that *H. heidelbergensis* at this time was capable of limited grammatical language, and had sufficient cognitive abilities (see below) to empathize to some degree with other individuals. Such abilities would greatly empower socially cooperative action. How much of this would have encouraged a sense of individuality and personality within and between group members is a subject that will engage much future research. See papers in Gamble and Porr (2005).

Subsistence Behavior

The question of how *H. antecessor* and *H. heidelbergensis* acquired their food, and what they ate, is intimately bound with the character of social relations between individuals, sub-groups, and between the sexes. These relationships would have influenced the pattern of hunting and gathering adopted. It may well be that socially and geographically distant groups had very different subsistence practices. There are no *a priori* grounds why a single species-wide pattern need have prevailed everywhere.

At a site in Schöningen, Germany, dated to MIS 11, a series of wooden spears, made of spruce wood, has been excavated in a context strongly suggestive of their use in deliberate hunting. A herd of horses is interpreted to have been mired along the margins of an ancient lake and then killed. Deliberate fires were set along the shoreline, possibly as part of the hunt, possibly for carcass processing and cooking/eating of the meat. Interestingly, this site contained no handaxes. Another wooden spear point of the same date was recovered from the core-and-flake site at Clacton-on-Sea, Essex, England; this one was made of yew wood. Spears are taken to be definitive hunting equipment, whereas stone tools, whether they be the handaxes of the Acheulean, or the sharp flakes and retouched flake tools of the non-handax assemblages, were hand-held carcass-processing equipment, used on the ground (among other things – see below).

At Boxgrove, an MIS 13 site in Sussex, England, hunting is posited on the basis of the successful defense of the carcass of a large species of extinct horse. The extensive cut marks from stone tools left on the horses' bones indicate that the hominins were in no hurry to leave, and they controlled the carcass long enough to extensively butcher it. Rhinoceros bones from this site may show similar evidence. The basis of this assertion is the primacy of stone tool cut marks on the bones. Here a tool has slipped during the process of filleting the meat and removing sinew, cutting into the bone by accident. These marks are overlain by the teeth marks of carnivores, who must then have only gained access to the carcass after the hominins had finished with it. Whether deliberate hunting was involved at Boxgrove, or whether a party of hominins came upon an injured or sick animal that they were able to dispatch with relatively little effort, remains to be proven. The horse butchery was accomplished with the aid of hand axes. Both Boxgrove and Schöningen very clearly imply sufficient social cooperation between individuals acting as a group in order to secure and then defend the carcass against other predators and scavengers. See papers in Gamble and Porr (2005) for more details, and Gamble (1999) for an overview.

By the time of *H. heidelbergensis*, and *H. antecessor*, genus *Homo* was adapted to a predominantly meat-based diet. Only from such a highly protein-rich food source would hominins have obtained sufficient energy resources to survive winter conditions when edible plant resources would have been at a minimum. But the question of how early *Homo* acquired meat remains frustratingly difficult to resolve; was this animal a predator-hunter, or a scavenger, or something in between? The archaeological record is contradictory and fragmented. At the two Spanish sites of Aridos I and II, to the south of Madrid, there is ample evidence of hominins interacting with elephant carcasses. In the former site, bifaces and a small collection of Levallois pieces were found scattered among the bones of the animal. Butchery is strongly implied. In the latter site, the case is similar, except that no Levallois pieces were found. Both assemblages would class as Acheulean. Unfortunately, it was not possible to discern how the elephants died. Two other Spanish sites, in the mountain heartland of Spain, Torralba and Ambrona, were for many years considered good evidence of big game hunting – elephants driven into swampy ground and subsequently butchered. More recent studies set the two sites against a background of occasional visits by hominins to an area not intensively inhabited, thus explaining the low frequency of cut marks and stone tools from the sites. While it can be proved that hominins butchered occasional animals in the vicinity, proactive big-game hunting can not be clearly demonstrated.

As important as meat itself is a supply of fats, which aid in the breaking down of meat proteins. A diet of animal tissue represents the only effective overwintering solution for hominins surviving the cold and resource-scarce winter months and the early spring. Late winter/early spring may have been a time of dietary stress for hominins. Although animals would still have been present, their fat reserves would have been severely depleted over the winter. For hominins, a surfeit of lean meat would have represented a source of poor nutrition without the fat content to break the meat proteins down. Seasonal dietary stress may have been a frequent part of middle Pleistocene hominin life.

While it is an assumption that edible plant material would have formed an important contribution to the diet, just what was eaten, and when, remains unknown. Presumably, the majority of edible plant foodstuffs would be a seasonal, late spring/summer resource. Fish, marine mammals, and birds, may also have contributed to the diet, but as yet there are no substantive data to support the idea that these resources were exploited frequently.

Tool Behavior

Most tractable rock types are capable of being knapped, and were used to manufacture stone tools in the Lower Paleolithic. In regions dominated by fine-grained quartzites, and by siliceous rocks such as flint, these appear to have been extensively used. Knapping involved either the use of a hard hammer stone, usually a cobble, for working cores to make flakes, and roughing out blanks for hand axes, and/or the use of a soft hammer to thin and shape already-worked rough-outs for

hand axes. Soft hammers may have been made of antler, bone, wood, or even softer rock types.

Although there are regional variations on what was made throughout Europe during the Lower Paleolithic, for the most part these represent variations on a small number of basic tool themes. These were handaxes, and their characteristic debitage, thinning flakes; cores knapped for flakes; the un-retouched flakes which could be used as sharp cutting knives; and a variety of retouched tools. The latter were flakes or pieces of shattered raw material whose edges were trimmed by direct percussion along a length of one or more edges. Commonly this trimming was applied to make a scraper, but other retouched tool types were equally as common: notches, denticulates, flaked flakes, etc. As noted above, stone tools were a hand-held processing technology. They were used for carcass dismemberment and processing; usually only the hand axes were removed from a butchery site for further use. The cores, flakes, and retouched tools were left behind. It is only during the Middle Paleolithic that deliberate stone weapon tips begin to appear in the archaeological record.

Materials other than stone were utilized, probably extensively, but the evidence for organic materials is very sparse, a reflection of chance preservation. Shaped wooden spears from Schöningen and Clacton-on-Sea have already been discussed. Schöningen also preserved lengths of wood that have been split, and the discoverers interpreted these as composite tools, the wooden handles into which worked stone items were inserted (Thieme, 2005). Other wooden tools were interpreted as a throwing stick and even a charred cooking stick, used to hang meat over the fire. At a number of sites there are a small number of hand axes that were flaked on fragments of thick bone. They may represent a regionally specific behavioral response to shortages of suitable raw materials since they almost all come from Italy. There is an excellent German example from Bilzingsleben. From Boxgrove, a series of soft hammers (Pitts and Roberts, 1997) has been discovered in antler and in bone. They show extensive wear from use and were clearly used as knapping tools for some time before being abandoned or lost.

The hominin use of fire, its early discovery and the nature of its subsequent usage, has always been surrounded by controversy. But sites like Schöningen and Beeches Pit, Suffolk, England clearly demonstrate the controlled use of fire. Gowlett (2006) argues that hominins may have used fire, and been able to sustain it, but may not have been able to kindle it. As such, it was more of an ecological tool.

What the Archaeology can tell us about Cognition

The material culture of *H. antecessor* is confined to cores and flakes, but the more varied repertoire of *H. heidelbergensis* allows some insights into the cognitive capabilities of this species. To begin with the ability to retain a complicated sequence of thoughts/actions in the memory, and act them out, is implied by the process of cooperative behavior (see above). This is also implied in the ability to conceive of composite tools (see above), especially since they comprise elements of two different media (stone/organic). The enchainment of separate actions and distinct concepts, as part of a

single activity, is further implied in the use of two different hammer types (hard and soft) to make many Acheulean handaxes. The evidence of curated soft hammers from Boxgrove underscores this and implies the capacity for forward planning (Hallos, 2005).

But the signals are contradictory. The curated soft hammers from Boxgrove may have been abandoned when they were worn out, but the spears from Schöningen could certainly have been used again. While some hand axes appear to have been removed from kill sites for later use, a great many perfectly serviceable examples are scattered round the landscape, seemingly abandoned for no particular reason. Was *H. heidelbergensis*' memory/cognitive capacity sufficient only for the immediate task? Or is the paucity of information merely blurring a situationally variable behavioral pattern? Apart from the bone handaxes cited above, and despite the survival of bones on a great number of middle Pleistocene sites, no other bone tools have ever been found. It is worth noting that *H. heidelbergensis* made only handaxes out of bone – no other tool types appear to have been fashioned. Does this actually imply a rather limited cognitive capacity, or alternatively a lack of cognitive flexibility in developing novel solutions for familiar problems, and adapting old practices to new situations?

The Ideational Realm

We have no direct clues as to what early *Homo* in Europe, or anywhere else for that matter, thought about its world, its place within it, and what its relationship to the other world might be – if hominins thought about it all. There are no clear-cut examples of deliberate burials in the middle Pleistocene of Europe. At the Sima de los Huesos, in the Sierra Atapuerca, Spain, some 28, or possibly more, *H. heidelbergensis* individuals were thrown down a deep vertical shaft at the back of the cave. The excavators interpret this as deliberate interment, and cite the presence of a single handaxe found in and among the remains as an example of an offering left among the dead (Bermúdez de Castro et al., 2004). But the hominin bones were found among carnivore remains, and the lower chamber may have been a bear's den with a lower entrance as yet undiscovered. This implies that the hominins' remains were brought in as prey. Even if they were deliberately thrown down the shaft, it may represent nothing more than the disposal of corpses. The case for deliberate interment, and any associated ritualistic qualities, remains to be proved.

There are no examples of Lower Paleolithic artwork in Europe. There are two examples of hominin figurines, one from Berekhat Ram in Israel, and one from outside the area at Tan-Tan in Morocco, associated with Acheulean handaxes (Bednarik, 2003). But in both cases, the resemblance to a human-like figure is questionable, and their interpretation should remain in a suspense account until validated by unambiguous examples.

A sense of the esthetic is often claimed for hand axes which are exceptionally large and which show a clear effort to produce bilateral symmetry. This is a possibility, and these artifacts may have had a distinct social resonance (see above). In other cases, hand axes which have natural holes in the center, around

which the knapper has carefully flaked a working edge along the margins of the nodule or blank, might imply a sense of visual appreciation, or even humor. Hand axes with fossils in the center of the blank or nodule are also sometimes suggested to have been purposefully knapped to emphasize the decorative character of the fossil. But like hand axes with a hole in the center, these are so few and far between that gifting their knappers with a clear sense of their overall esthetic value is difficult. They may well have failed to recognize fossils as having any form of visually pleasing quality, and a hole in the center of a knappable piece of flint may have been nothing more than that.

Conclusion

The archaeological record from the earliest occupation of Europe until ca. 300 ka and the introduction of PCT is a rich and varied one. The European data provide key insights into the variability present in the adaptation and lifestyles of early species of *Homo*. This record is every bit as rich as that found in other parts of the Old World. Future research directives will need to focus on the nature of social behaviour in the middle Pleistocene, and especially the significance of material culture is its social context. Other critical research agendas include the date and character of the earliest occupation of Europe, the question of the relationship between *Homo antecessar* and *Homo heidelbergensis*, and the explanation of non-handaxe assemblages in Europe.

See also: **Archaeological Records:** 2.7 Myr–300 000 Years Ago in Asia; Global Expansion 300 000–8000 Years Ago, Africa; Global Expansion 300 000–8000 Years Ago, Americas; Global Expansion 300 000–8000 Years Ago, Asia; Global Expansion 300 000–8000 Years Ago, Australia; Neanderthal Demise; Overview; Postglacial Adaptations. **Vertebrate Studies:** Interactions with Hominids.

References

- Bar-Yosef O and Belfer-Cohen A (2001) From Africa to Eurasia – Early dispersals. *Quaternary International* 75: 19–28.
- Bar-Yosef O and Goren-Inbar N (1993) *The lithic assemblages of "Ubeidiya: A Lower Palaeolithic Site in the Jordan Valley"*. Jerusalem: The Hebrew University of Jerusalem.
- Bednarik RG (2003) A figurine from the African Acheulian. *Current Anthropology* 44: 405–413.
- de Castro JM Bermúdez, Martínón-Torres M, Carbonell E, et al. (2004) The Atapuerca sites and their contribution to the knowledge of Human Evolution in Europe. *Evolutionary Anthropology* 13: 25–41.
- Dennell R (2003) Dispersal and colonisation, long and short chronologies: How continuous is the Early Pleistocene record for hominids outside East Africa. *Journal of Human Evolution* 45: 421–440.
- Dunbar R (2004) *The Human Story*. London: Faber and Faber.
- Gamble C (1999) *The Palaeolithic Societies of Europe*. Cambridge University Press.
- Gamble C and Porr M (2005) *The Hominid Individual in Context*. Abingdon: Routledge.
- Gibert J, Gibert L, Ferrández-Canyadell C, Iglesias A, and González F (2001) Venta Micena, Barranco León-5 and Fuentenueva-3: Three archaeological sites in the Early Pleistocene deposits of Orce, South-East Spain. In: Milliken S and Cook J (eds.) *A Very Remote Period Indeed, Papers on the Palaeolithic Presented to Derek Roe*, pp. 144–152. Oxford: Oxbow Books.
- Gibert L, Scott G, and Ferrández-Canadell C (2006) Evaluation of the Olduvai Subchron in the Orce ravine (SE Spain): Implications for Plio-Pleistocene mammal biostratigraphy and the age of the Orce archaeological sites. *Quaternary Science Reviews* 25: 507–525.
- Goren-Inbar N, Feibel CS, Verosub KL, et al. (2000) Pleistocene milestones on the Out of Africa corridor at Geshen Benot Ya'aqov, Israel. *Science* 289: 944–946.
- Gowlett JAJ (2006) The early settlement of Northern Europe: Fire history in the context of climate change and the social brain. *Comptes Rendus Palevol* 5: 299–310.
- Hallors J (2005) "15 Minutes of Fame": Exploring the temporal dimension of Middle Pleistocene lithic technology. *Journal of Human Evolution* 49: 155–179.
- Kohn M and Mithen S (1999) Handaxes: products of sexual selection? *Antiquity* 73: 518–526.
- Lisiecki L (2005) <http://www.lorraine-lisiecki.com>.
- Lisiecki L and Raymo ME (2005) A Pliocene-Pleistocene stack of 57 globally distributed benthic $\delta^{18}\text{O}$ records. *Paleoceanography* 20: 1–17 PA1003. <http://dx.doi.org/10.1029/2004PA001071>.
- Lordkipanidze D, Bar-Yosef O, and Otte M (2000) Early humans at the gates of Europe. In: *Proceedings of the First International Symposium Dmanisi, Tbilisi (Georgia)* September 1998, p. 92. Etudes et Recherches Archéologiques de l'Université de Liège (E.R.A.U.L.).
- Manzi G (2004) Human evolution at the Matuyama-Brunhes Boundary. *Evolutionary Anthropology* 13: 11–24.
- McNabb J (2005) Hominins and the Early-Middle Pleistocene transition: Evolution, culture and climate in Africa and Europe. In: Head MJ and Gibbard PL (eds.) *Early-middle Pleistocene Transitions. The Land-Ocean Evidence*. The Geological Society: Geological Society, Special Publications 247.
- Parés JM, Pérez-González A, Rosas A, et al. (2006) Matuyama-age tools from the Sima del Elefante site, Atapuerca (northern Spain). *Journal of Human Evolution* 50: 163–169.
- Parfitt SA, Barendregt RW, Breda M, et al. (2005) The earliest record of human activity in Northern Europe. *Nature* 438: 1008–1012.
- Pitts M and Roberts M (1997) A Fairweather Eden. *Century Books*.
- Rightmire GP, Lordkipanidze D, and Vekua A (2006) Anatomical description, comparative studies and evolutionary significance of the hominin skulls from Dmanisi, Republic of Georgia. *Journal of Human Evolution* 50: 115–141.
- Roebroeks W (2001) Hominid behaviour and the earliest occupation of Europe: An exploration. *Journal of Human Evolution* 41: 437–461.
- Roebroeks W and van Kolfschoten T (1995) The earliest occupation of Europe. *Proceedings of the European Science Foundation workshop at Tautavel (France), 1993, Analecta Praehistorica Leidensia* 27. Leiden: University of Leiden.
- Ron H and Levi S (2001) When did hominids first leave Africa? New high-resolution paleomagnetic evidence from the Erk El-Ahmar Formation, Israel. *Geology* 29: 887–890.
- Sahnouni M, Hadjouis D, Made JVD, et al. (2002) Further research at the Oldowan site of Ain Hanech, North-Eastern Algeria. *Journal of Human Evolution* 43: 925–937.
- Thieme H (2005) The Lower Palaeolithic art of hunting. In: Gamble C and Porr M (eds.) *The Hominid Individual in Context*, pp. 115–132. Routledge.
- Tuffreau A (1996) L'Acheuléen dans l'ouest de l'Europe. In: *Centre d'Etudes et de Recherches Préhistoriques Université des Sciences et Technologies de Lille*, p. 4. Publications du CERP.
- Villa P (2001) Early Italy and the colonization of Western Europe. *Quaternary International* 75: 113–130.

Global Expansion 300 000–8000 Years Ago, Africa

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Models of the Geographical Origin of Modern Humans

There were, until recently, two major opposing models for the origin of our species, the out-of-Africa and the multiregional model. Of the two, the latter is now largely abandoned except by a small group of adherents. However, in light of their continued existence in the literature, a summary of the multiregional model is appropriate.

The Multiregional Model

The multiregional model currently holds that *Homo sapiens* evolved in Africa, but, by means of gene flow and hybridization, modern peoples in regions of the Old World retain genetic contributions from earlier hominids: specifically from *Homo erectus* in East Asia and *Homo neanderthalensis* in Europe. The model has two forms. The strong form holds that *H. sapiens* has been the only species of *Homo* since at least 1800 ka, and that the modern form evolved in several places at once, with earlier local populations making substantial genetic contributions to their modern counterparts. The weak form holds that a genetic contribution of some kind was made by local 'archaic' populations to the earliest local modern ones. All proponents of the model have shifted to increasingly weaker forms as new evidence has made the stronger forms increasingly unlikely (Pearson, 2004). Since it is agreed that most (if not all) of the genetic origin of *H. sapiens* is African, the debate has become one of statistically remote possibilities and is essentially over.

The Out-of-Africa Model

The out-of-Africa model, also referred to as the out-of-Africa-2 model (since there was at least one earlier *Homo* exodus from Africa), the African Eve model, or the replacement model, holds that *H. sapiens* evolved only in Africa and then migrated to all other parts of the globe with very little or no genetic contribution from other hominid species, which were rapidly replaced.

Evidence for the Geographical Origin of Modern Humans

Three forms of evidence have been brought to bear on each of these two major models: morphological, genetic, and archaeological. Together, they support only the out-of-Africa model for the origin of our species.

Skeletal Morphological Evidence

Supporters of the out-of-Africa model note continuities in the skeletal morphology of African populations that occur nowhere else. They believe that the earliest modern people

in Europe were morphologically more African than later Europeans, and that any morphological similarities to preceding archaic *Homo* populations outside Africa are the result of convergent evolution under similar environmental stresses or of retained traits from a very ancient common ancestor.

The earliest skeletal materials assigned to *H. sapiens* are all African. The oldest of all *H. sapiens* are the Omo finds from near Lake Turkana, Ethiopia, which have recently been dated to 196 ka (McDougall et al., 2005) (Figure 1). The second oldest *H. sapiens* material comes from near Herto, Ethiopia, and has been dated to between 154 and 160 ka (White et al., 2003) (Figure 1). Specimens from southern African Middle Stone Age (MSA) sites identified as *H. sapiens* include material from Border Cave, Klasies River Mouth (Singer and Wymer, 1982), Die Kelders, and Pinnacle Point 13B (Figure 1). The recent application of multiple dating methods has led to a consensus that at least some of the Klasies and Border Cave material is older than 100 ka and the rest of the Klasies material is older than 60 ka.

In northwestern Africa, Jebel Irhoud (Morocco) yielded several well-preserved fossils, which are almost certainly older than OIS 4 (Hublin, 1992, pp. 186–187), and electron spin resonance (ESR) dates place them between 190 and 90 ka. There are no human fossils of comparable age in northeastern Africa, but modern humans at Skhul and Qafzeh (Israel) are dated to 90–100 ka, are associated with African faunal elements, and have African body proportions. Hublin (1992) considers them to be very similar to those from Jebel Irhoud. They represent a temporary expansion by moderns out of northeastern Africa.

General features such as cranial capacity, tooth size and morphology, and stature, when they can be measured, fall within the range of modern (Holocene) African populations. However, the sample of early *H. sapiens* is still small, so that it is difficult to assess population-scale questions of variability and change. The variability shown has been interpreted both as exactly what would be expected from the founding modern population, and also as having no discernible relationship to living populations.

Genetic Evidence

The genetic evidence strongly supports a single African origin for all living people. This was first applied to modern human origins in a germinal study (Cann et al., 1987) which compared the mitochondrial DNA (mtDNA) from living populations. Cann et al. found that living Africans have the greatest diversity and, thus, the greatest age, and they estimated that all living people are descended from a related group of African women who lived between 200 and 100 ka.

The study was initially attacked as being flawed in methodology. However, since then, many unflawed studies of living

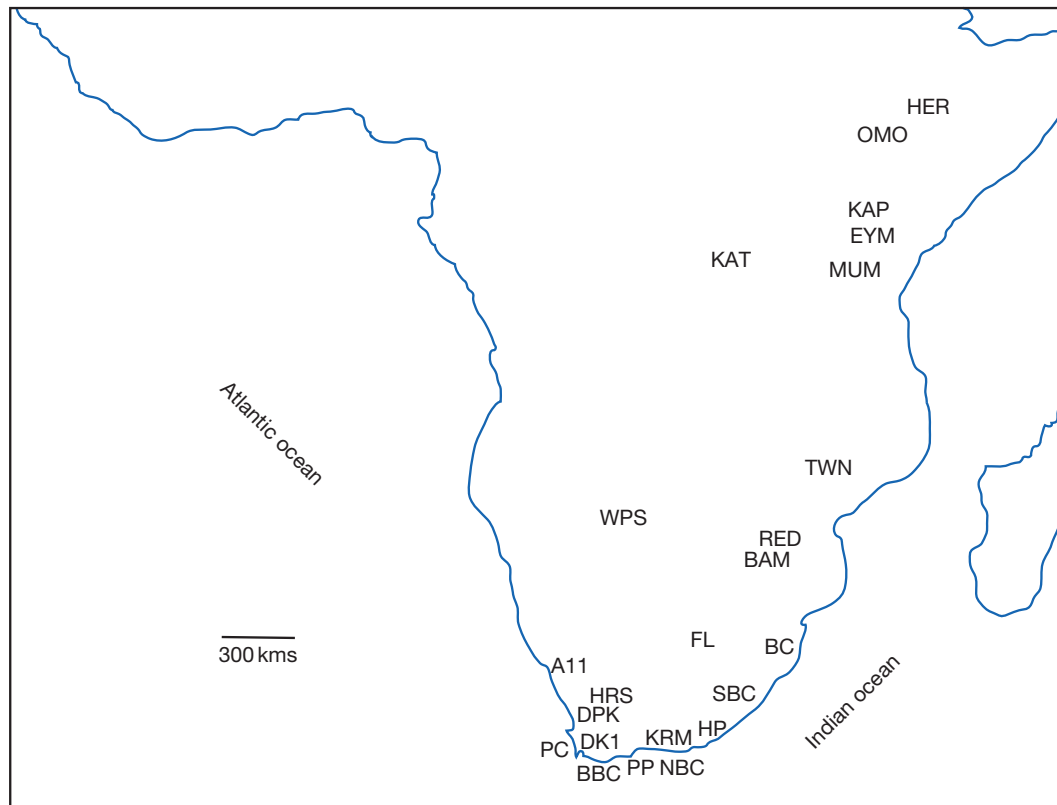


Figure 1 Middle Stone Age sites in sub-Saharan Africa. Key: A11 = Apollo 11, BAM = Bambata, BBC = Blombos Cave, BC = Border Cave, DK1 = Die Kelders, DPK = Diepkloof, HP = Howieson's Poort, EYM = Enkapune ya Muto, FL = Florisbad, HER = Herto, HRS = Hollow Rock Shelter, KAP = Kapthurin, KAT = Katanda, KRM = Klasies River Mouth, MUM = Mumba, NBC = Nelson Bay Cave, OMO = Omo, PC = Peers Cave, PP = Pinnacle Point, RED = Redcliff, SBC = Sibudu Cave, TWN = Twin Rivers, WPS = White Paintings Shelter. McBrearty and Brooks (2000), Mitchell (2002), and Minichillo (2005).

populations (Pearson, 2004) have reached conclusions very similar to the first study: that our species has a single origin in Africa.

Harpending et al. (1998) utilized both microsatellites on the Y chromosome and mtDNA and found that our ancestors underwent a rapid population expansion about 100 ka. They concluded that the population expansion took place after some bottleneck event, that it occurred in Africa, and that it is completely incompatible with any version of the multiregional model.

Archaeological Evidence from Southern Africa

Skeletal and genetic evidence refers to human biology or anatomy. Archaeological evidence refers to behavior: a profoundly vague concept that includes social relations, customs, technology, language, intellectual capacity, and many other aspects. Within evolutionary biology, it is generally held that if anatomical and behavioral changes do not coincide, then it is the behavioral changes that occur first. In the study of modern humans, however, there has been uncritical acceptance of the idea that anatomical change preceded behavioral. This idea has persisted for over three decades, perhaps because it allows the European record some relevance. However, in the words of

Lewin (1998, p. 113), 'the argument must be recognized as special pleading with no empirical basis.'

MSA studies in sub-Saharan Africa have recently undergone a major paradigm shift, involving reinterpretation of previously known sites and new research goals. For convenience, we name the earlier paradigm that prevailed in MSA studies from the late 1970s until recently, the Klasies paradigm, and the newer one, which has emerged since the late 1990s, the Blombos paradigm.

The Klasies paradigm

The MSA cave sites at the Klasies River Mouth (KRM) in southernmost Africa are crucial to research on modern human origins (Figure 1). Excavated in 1966–68, the MSA deposits there were massive and provided tens of thousands of lithic and faunal artifacts from gross-scale, stratigraphic units. There was also evidence that the oldest deposits were least 120 ka old (Singer and Wymer, 1982). An important pattern in the lithic assemblage was what appeared to be long periods (some 80 ka altogether) without much technological change and near-absence of more modern artifacts, such as fully bifacial points and bone tools (Figure 2). Indeed, the last paragraph of the report includes the dismal statement, 'Most significant is the correlation of such a long period of occupation. . . with little if

anything to show any change in subsistence and economy' (Singer and Wymer, 1982, p. 210).

On the basis of the Klasies sequence, Singer and Wymer outlined the MSA sequence as shown in Table 1, with the Howiesons Poort as the only (very brief) break in the monotony. The Howiesons Poort substage, dating to around 60 ka, includes small-backed blades and true-backed bladelets (Figure 3) that we now know were hafted as parts of composite tools (Figure 4) (Minichillo, 2005). Both backing of lithics and the creation of composite tools are considered to be advanced technologies. The increase in mineral pigment use and color variability during this period has been interpreted as increasingly symbolic behavior (Figure 5).

In 1981, when Volman found that the Klasies sequence did not fit that of Border Cave, he devised the sequence shown in the second column of Table 1. Before long, this proved inadequate as well.

The fauna from Klasies has also played a critical role in discussions of the modernity of MSA people. One such discussion was engendered by Binford's (1984) interpretation, which focused on the relative abundance of parts of animals. He proposed that the MSA people were primarily scavengers and had poor organizational and planning skills. This interpretation has been dismissed on the grounds that the materials he analyzed lacked long bone shaft fragments due to excavation and analytical bias, and because there is now direct evidence for MSA hunting of large bovids.

Other patterning in the faunal assemblages of the MSA continues to be used to construct models of human behavior. Klein (1999) compared the patterns of faunal exploitation in local MSA and LSA sites to argue forcefully that MSA people were inefficient hunters compared with subsequent LSA people. This argument has two parts: first, that fewer species were exploited during the MSA, and, second, that aggressive species, such as adult cape buffalo and bush-pig, were avoided. Klein went on to link this directly to his idea that MSA people lacked some modern intellectual capacities until 50–40 ka. However, a statistical analysis of the same faunal data found that the faunal patterns are poor markers of behavioral modernity, good markers of climatic change at the Pleistocene–Holocene transition, and are better explained by sample sizes and the

application of foraging theory models than by changes in human cognition (Minichillo, 2005).

Within the Klasies paradigm, all models of MSA behavior had to take into account the apparent stasis observed there.



Figure 2 Unifacial point (top left), Still Bay bifacial points and worked bone point from Blombos Sands and Kleinjongensfontein, near Still Bay, South Africa. The Heese collection at IZIKO: South African Museum. Bar is 1 cm. Minichillo, 2005.



Figure 3 Backed segments and crescents for the Howiesons Poort layers at Klasies River Mouth, the Singer and Wymer collection at IZIKO: South African Museum. Bar is 1 cm.

Table 1 Proposed substages of the Middle Stone Age in southern Africa

<i>Singer and Wymer (1982)</i>	<i>Volman (1981)</i>	<i>Wurz (2002)</i>	<i>Minichillo (2005)</i>	<i>Sites/Layers</i>	<i>Age estimates (years ago)</i>
MSA III and IV	MSA 3/4	Post-Howiesons Poort	Post-Howiesons Poort	KRM (Upper), Rose CottageCave, Sibudu, Ysterfontein?	55–35 000
Howiesons Poort	Howiesons Poort	Howiesons Poort	Howiesons Poort	KRM (HP), HowiesonsPoort shelter (S&H)	65–55 000
MSAII	MSA 2b		Die Kelders	Die Kelders, Diepkloof? Howiesons Poort shelter (D&D)?	70–60 000
		Still Bay	Still Bay	Blombos Cave (M1), Dale Rose Parlour	79–69 000
		Mossel Bay	Mossel Bay	Blombos Cave (M2)	?
				KRM (SAS)	72–68 000(too young??)
MSA I	MSA 2a MSA 1	Klasies	Klasies	KRM (LBS) Border Cave	>120–105 000 >125 000

Adapted from Wurz (2002) and Minichillo (2005).

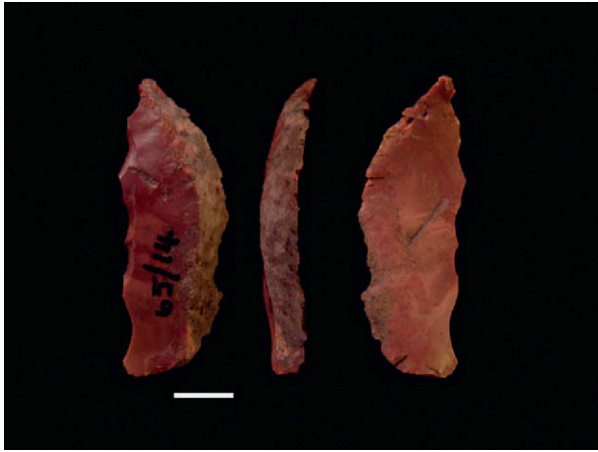


Figure 4 Backed blade with adhesive residue from hafting as a wood-scraper from the Howieson's Poort shelter, near Grahamstown, South Africa. The Deacon and Deacon collection, Albany Museum. Bar is 1 cm. [Minichillo, 2005](#).



Figure 5 Ochre crayons from the Howieson's Poort shelter, near Grahamstown, South Africa. The Deacon and Deacon collection, Albany Museum. Bar is 1 cm.

The Blombos paradigm

Blombos Cave ([Figure 1](#)) was excavated using extremely fine-scaled recovery and stratigraphic techniques, and the MSA has precise age-estimates based on single-grain optically stimulated luminescence. The sequence is quite different from that at Klasies, and [Wurz \(2002\)](#) has thus proposed a new scheme for organizing the southern African MSA ([Table 1](#), column 3).

The MSA at Blombos is sealed below a dune dating to 65–69 ka. The uppermost MSA deposits, M1, have yielded bifacial points of the Still Bay type and are sometimes referred to as the Still Bay layers. These have been dated to 74 ± 5 ka ([Henshilwood et al., 2004](#)). Mineral pigments abound, and some engraved pieces of ochre are seen as the earliest expression of symbolic thought. There are also perforated marine shells interpreted as a necklace and, thus, the oldest known personal adornment.

The underlying M2 is best known for the many pieces of worked bone. Careful excavation revealed that the abundant bifacial points in M1 grade into the abundant bone points in M2; that is, two types of artifacts vary inversely. This is quite different from the transitions at Klasies where there were hiatus in occupation between major stratigraphic units.

Recent work at Klasies confirmed the general chronology of the site, but not the specific interpretations. New dates, the lack of climatic change, and low density of shells suggest that the main stratigraphic unit (the SAS Member) was not deposited over tens of millennia, but over a very few millennia or perhaps even centuries ([Minichillo, 2005](#)). This removes the appearance of stasis; the stability within the SAS Member still represents stylistic unity ([Wurz, 2002](#); [Wurz et al., 2003](#)), but for a much briefer period. Rapid deposition has also been suggested for the large MSA deposit at Die Kelders, and technological change within a few millennia is clearly shown at Blombos.

In sum, the MSA archaeological record in southern Africa shows several profound technological changes occurring in periods of no more than a few millennia ([Table 1](#)); Sophisticated and composite artifacts, fine bone-work, use of color and incised patterns, and perhaps even jewelry — these are surely modern behaviors.

The Expansion of Modern Humans

Since there are the remains of early modern humans from the southernmost Cape to Morocco, we know that they expanded throughout Africa. When modern humans began to expand beyond Africa, it was almost certainly from northern Africa, into either Iberia or the Levant. We do not know enough about Arabia to assess the probability of a route through the Horn of Africa. We will outline the archaeological evidence from northern Africa up to 8 ka and then discuss what this reflects about expansion.

Archaeological Evidence from Northern Africa

The great Sahara Desert is the largest and driest desert in the world; very little of it receives more than 25 mm precipitation per annum and some parts receive none at all. At some periods in the past, the Sahara has seen significant rainfall and human occupation extending from the modern sahel to the Mediterranean. During other periods, it has expanded northwards to meet the sea itself, as it does today. Rainfall has always been the most important factor in North African prehistory.

Middle Paleolithic or Middle Stone Age

As in southern Africa, the middle Paleolithic (or Middle Stone Age) of northern Africa probably began by 300–250 ka. It includes two major units, the Mousterian and the Aterian, which overlapped considerably in time. The Mousterian predominates in northeastern Africa and the Aterian in northwestern, but the Mousterian is known as far west as Morocco and the Aterian as far east as the Nile Valley. Within the Nile Valley, other variants have been defined, such as the Nubian complex and the Taramsan, but these are undated. In fact, the only long Middle Paleolithic sequence is from the Haua Fteah, Libya, which was excavated in the early 1950s and is therefore essentially undated.

In artifacts, the Mousterian is very similar to Charentian and Denticulate Mousterian as known in Europe and southwestern Asia, and has a much higher frequency of retouching than is found in southern Africa. Bifacial foliates occur in the Nubian Complex in the Nile Valley and the Red Sea Hills and

also in the Aterian to the west. The Aterian is best known for the presence of tanged tools (not merely tanged points), which are generally taken to indicate hafting. Wurz et al. (2005) concluded that there is the same kind of variability in the MSA technology from Klasies as there is in the Middle Paleolithic of the Nile Valley. However, the latter remains undated and, at this level of technology, convergence is not surprising.

During a warm period about 100–90 ka, the northern African moderns moved briefly into the Levant, as did several other African animals. This colonization failed and the Africans returned south when the climate became colder. The successful and permanent occupation beyond Africa was closer to 50 ka, or later (Hublin, 2000).

There is very little information about when the middle Paleolithic ended. In the Nile Valley, Vermeersch (2000, table 14.1) estimates that the middle Paleolithic ended about 80–70 ka. The Sahara could be inhabited only during periods of increased rainfall, and the last wet period before the final Pleistocene lies far beyond the range of radiocarbon dating. Thus, the end of the middle Paleolithic could be dated only when techniques were developed that extend beyond 40 ka (luminescence, ESR, uranium-series and others). Age estimates for the last wet phase are available for at least seven regions of the Sahara. They range for a low estimate of ≥ 60 ka in the Tadrart Acacus in central Libya to high estimates of ≥ 80 ka in the southcentral Sahara and the region of the Great Eastern Erg. The middle Paleolithic in the Sahara must therefore all be older than 65–70 ka (Figure 6).

There are few reliable absolute dates for the end of the middle Paleolithic. The date of 45 ka at the Haua Fteah is minimal. At Jebel el Gharbi, western Libya, two sites have radiocarbon dates of about 44–43 ka (Garcea, 2004, pp. 33–34). Dating the end of the Middle Paleolithic is particularly contentious in the Maghreb, where the Mousterian sometimes underlies the Aterian. Recent critical compilations of dates for the Aterian (Cremaschi et al., 1998, table I; Bouzouggar et al., 2002, table 7; Close, 2009, table I) indicate that it may not have lasted much (if at all) beyond about 40 ka.

Upper/Late Paleolithic or Late Stone Age

The earliest Upper Paleolithic (in the European and southwest Asian sense) in northern Africa is the Dabban in Libya. It dates to about 40 ka in Cyrenaica (McBurney (1967, p. 326, 1975,

p. 419)) and to about 30 ka in Jebel el Gharbi (Garcea (2004, pp. 33–34)). It is unknown in the Maghreb, where there is no archaeological record between ≥ 40 ka and the first bladelet industry (Iberomaurusian) not long before 20 ka (Figure 7).

The Dabban persisted at Jebel el Gharbi until no later than about 25 ka (Garcea 2004, pp. 33–34). At the Haua Fteah, the only two dates that definitely refer to the Dabban are about 33.1 and 28.5 ka (McBurney 1967, table III.1). The Dabban may have disappeared from both Cyrenaica and Jebel el Gharbi at about the same time, and both areas then remained unoccupied for several millennia.

In the Nile Valley, chert mines date to 35–30 ka followed by a hiatus until about 25 ka (Vermeersch (1992, pp. 121–122)).

At about 20 ka, there was a profound change in technology, to one directed almost exclusively toward the production of bladelets, some of which were then backed. These appeared in Nubia and Upper Egypt at about 20 ka in Cyrenaica by 19 ka, and in the Maghreb before 20 ka at Tamar Hat and Taforalt (Close and Wendorf, 1989) (Figure 7). At the Haua Fteah, the change from the Dabban was so great McBurney (1967) suggested an actual change of population; this would, indeed, be the case if Cyrenaica had been unoccupied for several millennia. The Iberomaurusian certainly represents a new population in the Maghreb, which had been unoccupied for some 20 ka. The Sahara remained uninhabitable. The origin of the bladelet-makers remains unclear but may have been in sub-Saharan Africa.

Epipaleolithic and Neolithic

About 12 ka, the monsoon-belt moved northward and rain came again to the Sahara. For the next 4 ka, all of northern Africa was occupied by descendants of earlier local populations, as far as we know. In the Nile Valley, along the coast and in the Maghreb, freshwater was relatively copious. There were, in addition, the rich resources of the Nile itself and of the Mediterranean. In these regions, therefore, people were able to continue a gathering–hunting–fishing way of life until after 8 ka.

In the desert, however, particularly in what is now the southern Sahara, water was much less abundant – probably less than 20 mm per annum except in mountainous areas – and subsistence was more precarious. People thus ensured their survival as foragers by also keeping domestic animals.

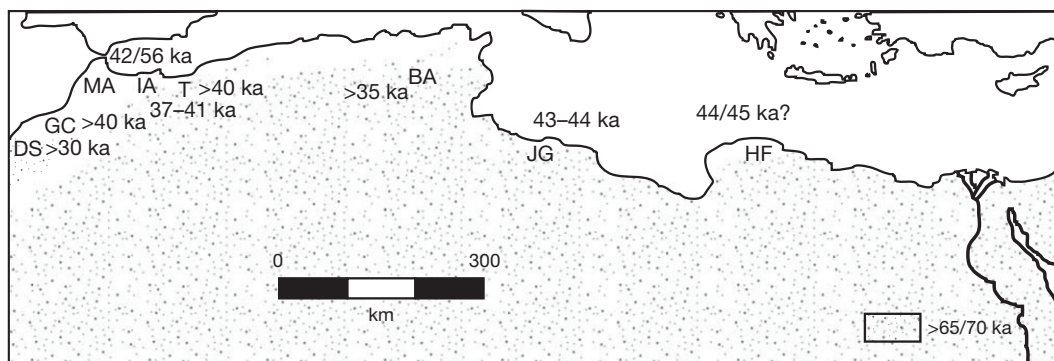


Figure 6 Dates (in thousands of years ago) for the end of the middle Paleolithic in northern Africa. Key: DS = Dar-es-Soltane, GC = Grotte des Contrebandiers (Témara), MA = Mugharet el 'Aliya, IA = Ifri n'Ammar, T = Taforalt (Grotte des Pigeons), BA = Bir el Ater, JG = Jebel el Gharbi, HF = Haua Fteah.

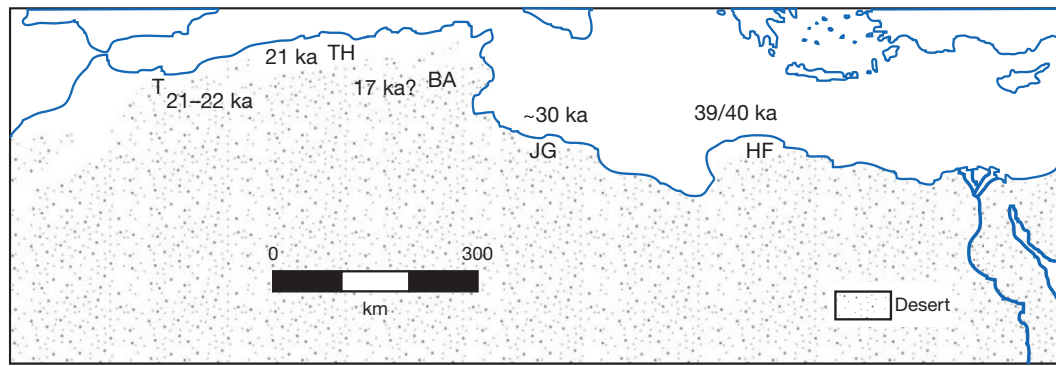


Figure 7 Dates (in thousands of years ago) for the beginning of the Upper or late Paleolithic in northern Africa. Key: T = Taforalt (Grotte des Pigeons), TH = Tamar Hat, BA = Bir Oum Ali, JG = Jebel el Gharbi, HF = Haua Fteah.

Cattle were domesticated in the southern Sahara before 8 ka, but their herders remained essentially gatherer-hunters (Close, 1992). Farming did not become an important factor in northern Africa until the arrival of the southwest Asian domesticated plants and animals, some time after 8 ka.

Routes Out of and into Northern Africa

The archaeological record suggests that between 300 and about 20 ka, northern Africa was sometimes occupied by modern humans and, with the possible exception of the Nile Valley, was sometimes not occupied by anyone. Periods of occupation could be separated by millennia, especially in northwestern Africa. Boats are more recent, so any contact would have to be overland into the Levant, or, perhaps, across the very narrow Strait of Gibraltar.

Although the Strait of Gibraltar is only about 14 km wide at its narrowest point, it is also extremely turbulent and the safest crossing is west of 6° W and ≥ 100 km long. Nonetheless, it is possible to see across the strait today, and, during periods of lowest sea levels, a series of rocky islands would have been exposed, reducing the maximum water crossing to about 4 km (Bouzouggar et al. (2002, figure 23)). The earliest (nonmodern) humans from Iberia almost certainly crossed Gibraltar (Bermudez de Castro et al., 1997), so it is difficult to think that modern humans could not have crossed in the Upper Pleistocene. However, those earliest humans moved into empty territory with no direct competition, whereas until after 30 ka, moderns would have found the hominid niche in Iberia occupied by Neanderthals and their predecessors. In this case, moderns would have been archaeologically visible only if there was a massive invasion of (Hublin 2000, p. 170).

Similarly, only a massive invasion would result in early moderns having a detectable genetic effect in Iberia (Hublin (2000, pp. 170–171)). Today, living Iberians share a common ancestry dating to the Upper Paleolithic, and they are genetically much closer to other West European populations than they are to North Africans. This is the genetic pattern of living populations, so that failure to cross the strait has nothing to do with modernity. Genetic comparison of living people in northern Africa, Europe, and the Levant shows that the oldest European group and the main group in northern Africa (including the Berbers) shared a common Levantine ancestor

until about 50 ka, suggesting that Europe and northern Africa underwent independent colonization by moderns from the Levant. Moderns initially spread from Africa into the Levant by around 50 ka and some of their descendants returned to recolonize northwestern Africa (at least) after it had been empty of people for ≥ 20 ka (40–20 ka).

The Insularity of the Coast of Northern Africa

Modern humans evolved in Africa some 200 ka. Since then and unexpectedly, much of northern Africa has been empty of people for tens of millennia. The discrepancy may lie in our view of northern Africa.

The Arab invaders of the seventh century AD named northwestern Africa *el Jezirat el Maghreb*—the Island of the West (Vaufrey 1955, p. 9) – because it was completely isolated between the Mediterranean and the desert. This isolation is true of almost the entire coast of northern Africa and its prehistory becomes more understandable when it is viewed as a series of islands (Shaw (2003, p. 98)): the Nile Valley, Jebel el Akhdar, Jebel el Gharbi, and the Maghreb (Figure 8). The islands have permanent sources of drinking water and are separated from each other by hundreds of kilometers of coastline, where the Sahara meets the Mediterranean and drinking water is rare to lacking.

During wetter phases, the islands were easily accessible from the rest of northern Africa and most of what is now desert was occupied. During phases of aridity, contact between the islands is very unlikely and each would become isolated. By 40 ka, human occupation had ceased, certainly in the Maghreb, probably in Jebel el Gharbi and perhaps in Cyrenaica. We conclude that, in each area, modern humans may have become locally extinct.

The presence of the Mediterranean and the absence of boats meant that any travel between the islands of northern Africa had to be overland. Historically, such journeys have depended upon the use of domestic animals. There were no domestic animals on the coast (although they existed in the southern Sahara) until after 8 ka and people had to carry everything they would need themselves, especially water.

Nevertheless, inter-island journeys were sometimes made successfully and islands could be recolonized for periods of millennia. Thus, about 40 ka, the makers of the Dabban

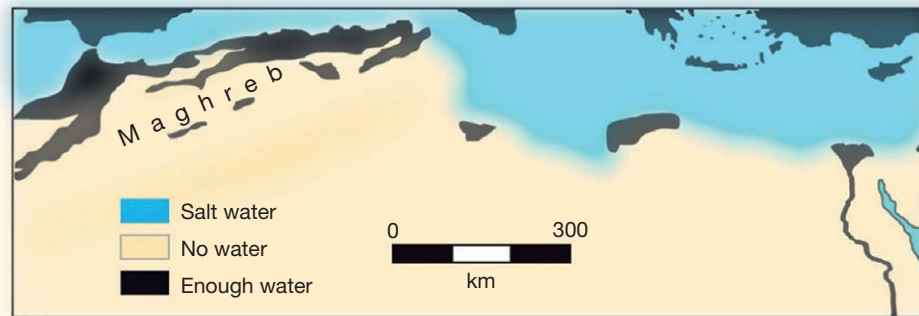


Figure 8 Northern Africa seen as a chain of islands. The boundary between the desert and the sea is deliberately blurred because its precise location is not crucial, and the islands remain insular no matter what the sea level.

reached Cyrenaica from the east and remained there until well after 30 ka. Their descendants (presumably) can be seen as far west as Jebel el Gharbi till about 30 ka; the area was occupied until about 25 ka. But they seem not to have gone farther west into the Maghreb. The final colonization of the islands began somewhat more than 20 ka with the makers of the backed-bladelet industries; modern human beings have occupied all of the islands ever since then.

See also: **Archaeological Records:** Overview; 2.7 Myr–300 000 Years Ago in Asia; 1.9 Myr–300 000 Years Ago in Europe; Global Expansion 300 000–8000 Years Ago, Asia; Global Expansion 300 000–8000 Years Ago, Australia; Global Expansion 300 000–8000 Years Ago, Americas; Neanderthal Demise; Postglacial Adaptations; **Vertebrate Studies:** Interactions with Hominids.

References

- Bermúdez de Castro JM, Arsuaga JL, Carbonell E, Rosas A, Martínez I, and Mosquera M (1997) A hominid from the Lower Pleistocene of Atapuerca, Spain: Possible ancestor to Neanderthals and modern humans. *Science* 276: 1392–1395.
- Binford LR (1984) *Faunal Remains from Klasies River Mouth*. Academic Press.
- Bouzouggar A, Kozłowski JK, and Otte M (2002) Étude des ensembles lithiques atériens de la grotte d'El Aliya à Tanger (Maroc). *L'Anthropologie* 106: 207–248.
- Cann RL, Stoneking M, and Wilson AC (1987) Mitochondrial DNA and human evolution. *Nature* 325: 31–36.
- Close AE (1992) Holocene occupation of the eastern Sahara. In: Klees F and Kuper R (eds.) *New Light on the Northeast African Past: Current Prehistoric Research*, pp. 154–183. Heinrich-Barth-Institut.
- Close AE and Wendorf F (1989) North Africa at 18 000 BP. In: Gamble C and Soffer O (eds.) *The World at 18 000 BP. The Low Latitudes*, vol. 2, pp. 41–57. Unwin Hyman.
- Close AE (2009) On the absence of a Middle–Upper Paleolithic transition in Mediterranean Northwest Africa. In: Camps M and Szmidt C (eds.) *The Mediterranean from 50,000 to 25,000 BP: Turning Points and New Directions*. Oxford: Oxbow Books and David Brown Book Company.
- Cremaschi M, Di Lernia S, and Garcea EAA (1998) Some insights on the Aterian in the Libyan Sahara: Chronology, environment and archaeology. *African Archaeological Review* 15: 261–286.
- Garcea EAA (2004) Crossing deserts and avoiding seas: Aterian North-African–European relations. *Journal of Anthropological Research* 60: 27–53.
- Harpending HC, Batzer MA, Gurven M, Jorde LB, Rogers AR, and Sherry ST (1998) Genetic traces of ancient demography. *Proceedings of the National Academy of Sciences of the United States of America* 95: 1961–1967.
- Henshilwood C, d'Errico F, Vanhaeren M, van Niekerk K, and Jacobs Z (2004) Middle Stone Age shell beads from South Africa. *Science* 304: 404.
- Hublin J-J (1992) Recent human evolution in northwestern Africa. *Philosophical Transactions of the Royal Society of London B* 337: 185–191.
- Hublin J-J (2000) Modern–nonmodern hominid interactions: A Mediterranean perspective. In Bar-Yosef O and Pilbeam D (eds.) *The Geography of Neanderthals and Modern Humans in Europe and the Greater Mediterranean*, pp. 157–182. Peabody Museum of Archaeology and Ethnology.
- Klein RG (1999) *The Human Career: Human Biological and Cultural Origins*, 2nd edn. University of Chicago Press.
- Lewin R (1998) *The Origin of Modern Humans*. Scientific American Library.
- McDougall I, Brown FH, and Fleagle JG (2005) Stratigraphic placement and age of the modern humans from Kibish, Ethiopia. *Nature* 433: 733–736.
- McBurney CBM (1967) *The Haua Fteah (Cyrenaica) and the Stone Age of the South-East Mediterranean*. Cambridge University Press.
- McBurney CBM (1975) Current status of the Lower and Middle Paleolithic in the entire region from the Levant through North Africa. In: Wendorf F and Marks AE (eds.) *Problems in Prehistory: North Africa and the Levant*, pp. 411–426. Southern Methodist University Press.
- Minichillo T (2005) *Middle Stone Age Lithic Study, South Africa: An Examination of Modern Human Origins*. PhD Dissertation, Seattle: University of Washington.
- Pearson OM (2004) Has the combination of genetic and fossil evidence solved the riddle of modern human origins? *Evolutionary Anthropology* 13: 145–159.
- Shaw BD (2003) A peculiar island: Maghrib and Mediterranean. *Mediterranean Historical Review* 18: 93–125.
- Singer R and Wymer J (1982) *The Middle Stone Age at Klasies River Mouth in South Africa*. University of Chicago Press.
- Vaufrey R (1955) *Préhistoire de l'Afrique. Tome I: Le Maghreb*. Masson.
- Vermeersch PM (1992) The Upper and Late Palaeolithic of northern and eastern Africa. In: Klees F and Kuper R (eds.) *New Light on the Northeast African Past: Current Prehistoric Research*, pp. 99–153. Heinrich-Barth-Institut.
- Vermeersch PM (ed.) (2000) *Palaeolithic Living Sites in Upper and Middle Egypt*. Leuven University Press.
- Volman TP (1981) *The Middle Stone Age in the Southern Cape*. PhD Dissertation, University of Chicago.
- White TD, Asfaw B, DeGusta D, et al. (2003) Pleistocene *Homo sapiens* from Middle Awash, Ethiopia. *Nature* 423: 742–747.
- Wurz S (2002) Variability in the Middle Stone Age lithic sequence, 115,000–60,000 years ago at Klasies River, South Africa. *Journal of Archaeological Science* 29: 1001–1015.
- Wurz S, le Roux NJ, Gardner S, and Deacon HJ (2003) Discriminating between the end products of the earlier Middle Stone Age sub-stages at Klasies River using biplot methodology. *Journal of Archaeological Science* 30: 1107–1126.
- Wurz S, Van Peer P, le Roux NJ, Gardner S, and Deacon HJ (2005) Continental patterns in stone tools: A technological and biplot-based comparison of Early Late Pleistocene assemblages from northern and southern Africa. *African Archaeological Review* 22: 1–24.

Global Expansion 300 000–8000 Years Ago, Asia

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Transitions in Lithic Production after 300 ka

In western Asia (including the Caucasus), changes in lithic production resulted in the type of Mousterian industries seen in Europe, whereas changes were more muted in southern Asia, and resulted in a non-Mousterian Middle Paleolithic. In China and the Far East, changes were even more subtle, and stone tool assemblages continued without major industrial innovations. Outside the Levant, details are still sketchy as to when and how these late middle Pleistocene transitions occurred.

The Lower to Middle Paleolithic transition in the Levant occurred ca. 250 ka (Shea, 2003) (Table 1). The immediate precursor of the Levantine Mousterian is the Yabrudian industry, which dates from 350 ka, and is the earliest regional lithic variant in Asia. It is found in Syria, Lebanon, and northern Israel, and features steeply retouched scrapers and relatively thick flakes (Figure 1); the Levallois technique is absent, and bifaces are rare or absent. The principal Middle Paleolithic industry of the Levant is Mousterian. Its most conspicuous attribute is the use of recurrent Levallois core reduction strategies to produce triangular and subtriangular flakes. The earliest Mousterian lithic assemblages are distinctive, with high values for Levallois and laminar indices. The transition from hand-held handaxes in the Lower Paleolithic to the use of hafted tools (i.e., stone points or blades mounted on wooden shafts) in the Middle Paleolithic may be considered a profound technological shift, as later hominids employed stone projectile weaponry to obtain game (e.g., Shea (2003)). Early Mousterian industries in the Levant range from ca. 250 to 128 ka, thus including a transition from warm and humid conditions during oxygen isotope stage (OIS) 7 to colder and more arid conditions in OIS 6. It is not clear which hominin(s) made the earliest Levantine Middle Paleolithic tools because both Neanderthals and early modern humans in the Levant are associated with the same types of Middle Paleolithic assemblages.

The Lower to the Middle Paleolithic transition in South Asia is poorly dated, though there is good stratigraphic evidence to indicate that the transition from large bifacial industries of the Acheulean to prepared core and flake industries of the Middle Paleolithic occurred by ca. 150 ka (Misra, 2001; Petraglia et al., 2003) (Figure 2). Acheulean industries in India may have been made by *Homo heidelbergensis*, based upon the identification of a cranial fragment from the Narmada river valley (Cameron et al., 2004). The hominin(s) that made the stone tools associated with the Indian Middle Paleolithic remain unknown but likely to have been *H. sapiens* by at least 60 ka (Mellars, 2006), if not earlier (Petraglia et al., 2007; Clarkson et al., 2012).

The Early Middle Paleolithic in Central Asia is poorly dated, but those from the last glaciation are Mousterian, or have Mousterian affinities. Several caves in Uzbekistan, notably Obi Rakhmat (ca. 60–90 ka), Anghilak (ca. 36–45 ka), Khudji (ca. 40 ka), and Teschik-Tasch, show that Neanderthals were

present in OIS 4–3 (Glantz et al., 2008). The best-dated Early Middle Paleolithic assemblages are from Tajikistan and Uzbekistan; these date from OISs 5 and 7, and show an increase in Levallois flaking. Their makers are unknown, but likely Neanderthal.

Although a Middle Paleolithic has been recognized in China on the basis of stone tools classifiable as scrapers, points, and choppers, these assemblages are so similar to earlier ones that it is doubtful whether it is useful to regard them as Middle Paleolithic rather than a later variant of the Early Paleolithic (Gao and Norton, 2002). Additionally, very few Chinese sites are sufficiently well-dated and -documented to allow detailed analyses of the changes that occurred between 300 and 50 ka. The best-documented late middle Pleistocene site in China is currently the cave of Panxian Dadong (ca. 100–250 ka) in southern China (Miller-Antonio et al., 2004); although lithic reduction sequences were complex, they were not dependent upon the type of prepared-core technologies so common in western Asia.

The Emergence of the Upper Paleolithic and the Expansion of Modern Humans

The earliest specimens of anatomically modern humans (AMH) in Northeast Africa date from ca. 190 ka, and are assumed to have originated in that continent. By ca. 25–30 ka, AMH is the only type of humanity in mainland Eurasia, and its expansion at the expense of indigenous populations (encapsulated by the 'Out of Africa 2' model) was a complex, lengthy, and still very poorly understood process (Dennell and Petraglia, 2012). The main archaeological change that occurs across much of Asia as part of this process is the replacement of Middle Paleolithic assemblages by Upper Paleolithic ones. The relationship of these changes to the expansion of modern humans is not clear-cut: in the Levant, for example, the Later Middle Paleolithic was made by both Neanderthals and modern humans, and early Upper Paleolithic assemblages are far from being unambiguous indicators of modern humans. Additionally, the quality of regional records of this process generally declines from west to east across Eurasia.

The Levant

The Israeli caves at Mount Carmel have yielded extensive fossil material of Neanderthals and anatomically modern humans (Shea, 2003). The cave sites of Tabun, Kebara, and Amud have yielded Neanderthals, whereas Qafzeh and Skhul contain anatomically modern humans, of which the earliest date to ca. 100 ka. This date is widely regarded as indicating the time when modern humans first entered Asia from Africa; however,

Table 1 Periodization of the Levantine Middle Paleolithic (after [Shea 2003](#): Table V)

Period	Dates (ka)	Marine OIS and Levantine climate	Hominins	Lithic industries
Late Lower Paleolithic	350–250	OIS 8–7. Cold, then warmer	<i>Homo</i> sp.	Late Acheulean and Acheulo-Yabrudian
Early Middle Paleolithic	250–128	OIS 7–6. Warm, then colder	<i>Homo</i> sp.	Phase 1 Levantine Mousterian
Mid Middle Paleolithic	128–71	OIS 5. Initially warm and hyperarid but growing colder, more humid	Neanderthals and early modern humans	Phase 2 Levantine Mousterian
Late Middle Paleolithic	71 to <47	OIS 4–early 3. Cold and dry	Neanderthals	Phase 3 Levantine Mousterian
Initial Upper Paleolithic	47–32	Mid-late OIS 3. Cold and dry	Modern humans	Early Upper Paleolithic (Emiran and Ahmarian), possible late Levantine Mousterian

given that there is no hominin skeletal data in western Eurasia between 100 and 300 ka, an earlier date cannot be excluded. Dated specimens of Neanderthals indicate that they were present in the Levant at and probably between 120 and 60 ka. Although the dating evidence implies that moderns and Neanderthals coexisted in the same region, it is of course possible that they never met, as different populations could have expanded during particular climatic phases, the Neanderthals occupying the region during colder periods, and the moderns during warmer episodes. The presence of both Neanderthals and modern humans in the region does however support the single-origin model of human evolution, as Neanderthals clearly became extinct in the Later Pleistocene ([Stringer, 2002](#)).

Significant technological changes occur at ca. 47–45 ka in the Levant, and early Upper Paleolithic industries occur in Ksar Akil (Lebanon) and Boker Tachtit (Israel) at around 45–38 ka. Two later Upper Paleolithic assemblages, known as the Ahmarian and the Aurignacian, are dated to ca. 38–30 ka. The Ahmarian is characterized by a blade and bladelet industry whereas the Aurignacian consists of a flaked-based industry with lamellar retouch on carinated scrapers and burins. The latter type of assemblage may have originated in Central Asia or in western Iran, and is generally regarded as the earliest unequivocal Upper Paleolithic tradition in Europe. Though these earliest Levantine Upper Paleolithic assemblages are not associated with human fossils, it is generally assumed that they were produced by modern humans ([Bar-Yosef, 2002](#)).

Anatolia and the Taurus-Zagros Mountains

Beyond the Eastern Mediterranean of the Levant, Neanderthals lived in the Taurus-Zagros Mountain zone, the Caucasus, and Central Asia as far east as the Lake Baikal region, Siberia ([Krause et al., 2007](#)). The Middle Paleolithic industries of this region are differentiated from those of the Levant as they often have low Levallois indices, little centripetal core preparation, and few retouched tools. One of the Neanderthals from Shanidar Cave, Iraq, has signs of disease and numerous injuries, perhaps indicating that this individual was cared for by others. The Baradostian industry, dating to ca. 40–35 ka, directly overlies Mousterian assemblages in cave sites of Iran and Iraq. The Baradostian is characterized by high percentages of burins, and rare carinate scrapers and busked burins, reminiscent of European and Levantine Aurignacian assemblages. The Mousterian assemblages in Karain Cave of western Turkey contain high

frequencies of retouched and resharpened tools, including side-scrapers, points, and convergent scrapers of nonlocal materials ([Kuhn, 2002](#)). The assemblages are primarily discoid, though Levallois technology appears in at least one stratigraphic layer. The Karain assemblages are thought to have affinities with the Zagros Mousterian or the Middle Paleolithic of the Balkans and southeastern Europe, rather than the Levant.

Upper Paleolithic sites are generally rare in elevated zones of the Zagros, the Iranian plateau, and the Anatolian plateau, leading some investigators to wonder if gaps in the archaeological records indicated abandonment during OISs 2 and 3. Few have been intensively studied and/or dated, and the Middle to Upper Paleolithic transition in this region is still poorly understood. Üçağızlı Cave is an early Upper Paleolithic sequence in Turkey that has yielded a combination of typical Middle and Upper Paleolithic tools that have been dated to ca. 40 ka ([Kuhn, 2002](#)) ([Figure 3](#)). The forms include end-scrapers, burins, and retouched blades, many pieces classifiable as Levallois flakes, blades, and points. The upper layers contain assemblages that are similar to early Ahmarian ones, including a prismatic blade technology dating to between 33 and 28 ka. Bone tools are common and the assemblages contain shell ornaments. Epipaleolithic or late Upper Paleolithic assemblages with microlithic-backed pieces have been reported throughout the region. The number of sites and the diversity of ecological niches exploited suggest localized specializations involving hunting of particular species.

South Asia

Further eastward, South Asia contains numerous Middle Paleolithic sites, but these are not associated with any human fossils ([Misra, 2001](#); [James and Petraglia, 2005](#)). During the Middle Paleolithic, there does not appear to be a single or favored technique of core preparation throughout the Indian subcontinent, though Levallois and discoidal techniques are most often described. As in Southwest Asia outside the Levant, there is urgent need for more well-dated sequences between 125 and 25 ka. From approximately 45 ka, increases in blade production and variability in assemblage composition characterize the South Asian record, with assemblages divided into flake-blade, blade-based, and blade and burin industries ([Murty, 1979](#)). Microlithic industries are as old as 36 ka, indicating that they were part of a diverse late Paleolithic package ([Clarkson et al., 2009](#); see also [Figure 4](#)). In sites such as Patne,

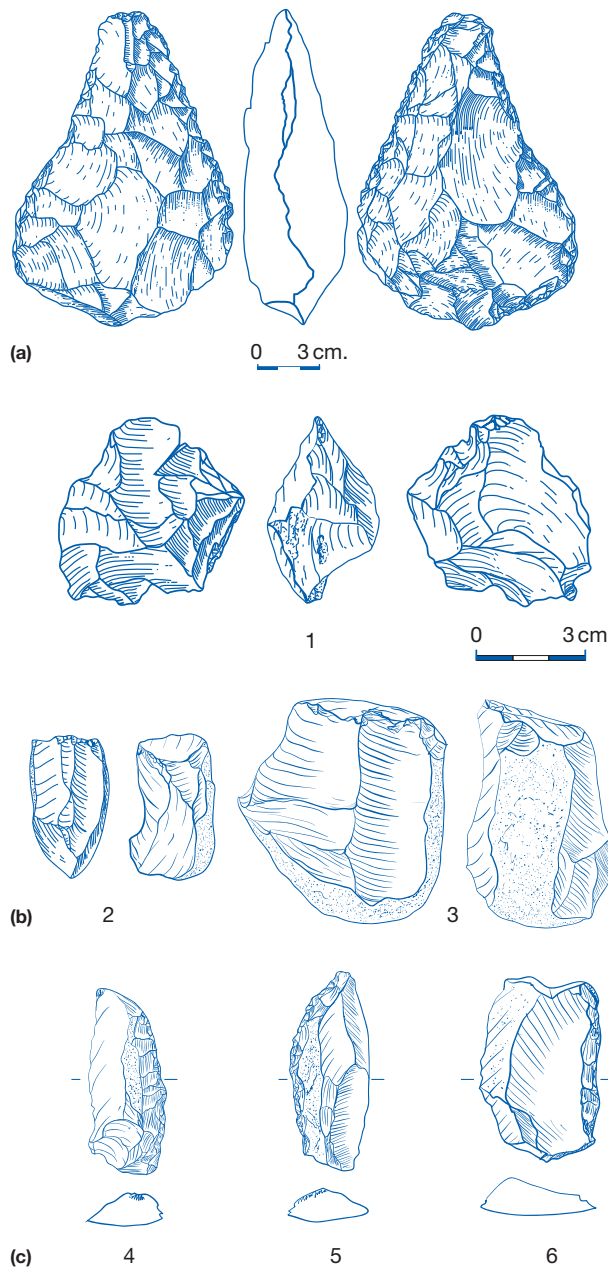


Figure 1 Transitional Acheulo-Yabrudian assemblages from Qesem Cave, Israel, dated to about 300–200 ka. (a) Handaxe; (b) radial core (1), cores on nodules (2, 3); (c) side-scrapers (4–6). Adapted from Gopher A, Barkai R, Shimelmitz R, *et al.* (2005) Qesem Cave: An Amudian site in Central Israel. *Journal of the Israel Prehistoric Society* 35: 69–92.

geometric microlithic technology developed from an industry characterized by backed blades and burins (Sali, 1989) (Figure 5). With the exception of the geometric microliths, the standardization of retouched artifact forms is not comparable to those seen in the Upper Paleolithic of Europe or Western Asia. There does not appear to be a sudden shift to ‘classic’ prismatic cores at the onset of the late Paleolithic, but rather, there is contemporaneity between flake- and blade-based technology, with the introduction of microlithic industries by 36 ka (Petruglia *et al.*, 2009). There is some evidence

for deliberate construction of shelters by about 45 ka and some evidence for symbolism after 28.5 ka, though the evidence is rare (James and Petruglia, 2005).

Eastern Asia

Late middle Pleistocene hominins such as the Dali and Jinniushan specimens have been classified as archaic *Homo sapiens* (Wu and Poirier, 1995), late *H. erectus* (Thorne and Wolpoff, 1992) or even as *H. heidelbergensis* (Rightmire, 2001), on the grounds that they are intermediate (at least chronologically) between *H. erectus* and *H. sapiens*. Discussion is often obfuscated by lack of agreement over what constitutes an ‘archaic’ or ‘modern’ feature, or suite of features. Further evidence is clearly needed to clarify what was probably a complicated set of biogeographic and phylogenetic relationships in eastern Asia (Lahr, 1996). As a further complication, some Chinese specimens may be “Denisovan”: this type of hominin is likely related to Neanderthals but known only from the ancient DNA that was extracted from a phalange and tooth from Denisova Cave, Siberia (Krause *et al.*, 2010).

Late Paleolithic Far Eastern lithic artifact assemblages vary in reduction techniques, and include bipolar flakes, Levallois blades, and microliths (Jia and Huang, 1985; Brantingham *et al.*, 2001) (Figure 6). The Zhoukoudian Upper Cave, which is associated with modern humans, contains an assortment of elaborate artifacts, including bone needles and harpoons, perforated marine shells, stone beads, and grave goods in burials. The association of modern humans with symbolic artifacts may indicate their spread, and the replacement of archaic hominins. Microlithic artifacts, perhaps as old as 28 ka, occur in a variety of ecological settings, from deserts to lakeside settings, indicating a wide range of adaptations.

The Replacement of Indigenous Populations of Neanderthals, *H. erectus* and *H. floresiensis* by Modern Humans

The prevailing view among many geneticists and paleoanthropologists is that all living humans in Asia today are descended from a relatively small population which originally arose in Africa (e.g., Ingram *et al.* (2000)). Based on slight and generally ambiguous fossil and archaeological evidence, it has been argued that modern humans reached the Levant by 100 ka, and later, other populations traveled (perhaps along the Indian Ocean rim) to reach Australasia by 45–60 ka (Lahr and Foley, 1994; Field and Lahr, 2006) (Figure 7). An alternative scenario is that our species entered the Levant and Arabia by OIS 5 (Petruglia *et al.*, 2011). Whichever model is preferred, modern humans replaced Neanderthals in Southwest and Central Asia, indigenous populations (type(s) unknown in South Asia), populations of *H. erectus* in Eastern Asia, and ultimately, *H. floresiensis* in the island of Flores, Indonesia.

The appearance of anatomically modern humans in Africa is accompanied by a variety of economic, social, and symbolic traits that are often associated with modern human culture, although the materialization of such traits is temporally and geographically patchy over the last 250 ka (McBrearty and Brooks, 2000), and few of these traits are both widespread and exclusive to modern humans.

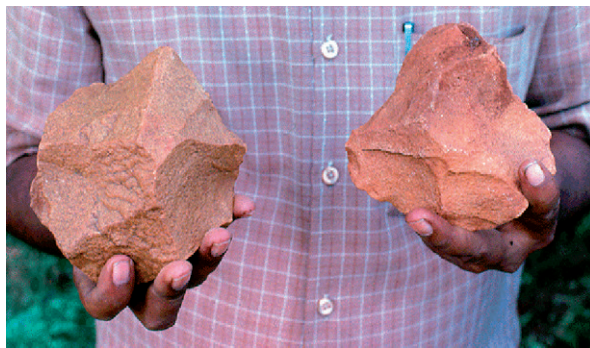
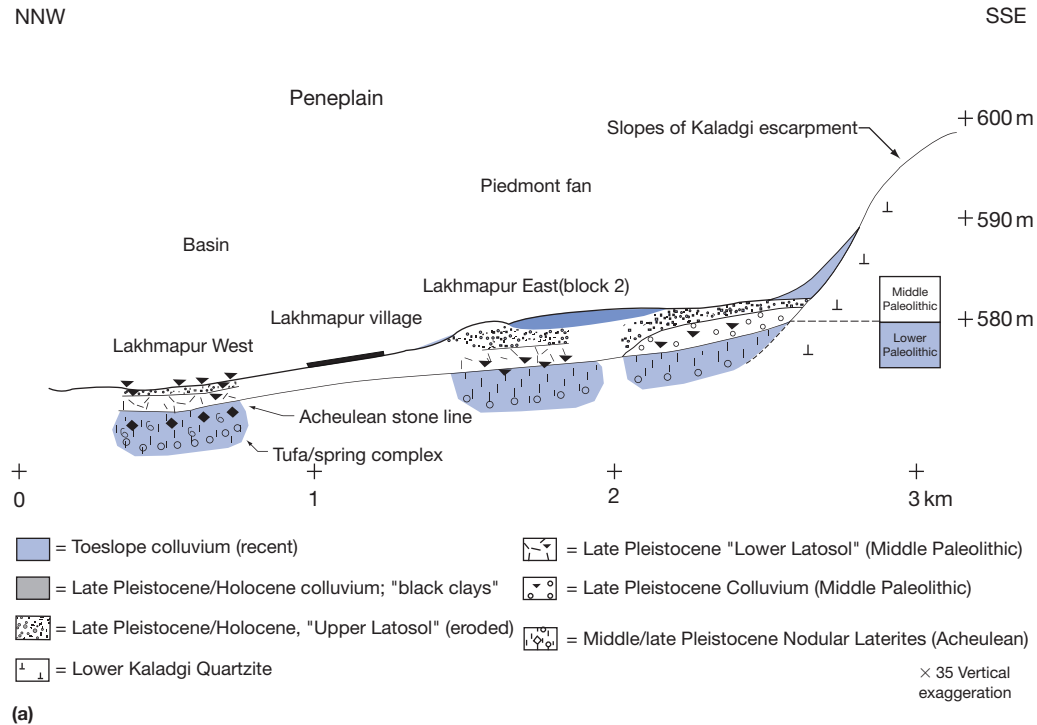


Figure 2 Landscape relations between transitional Lower and Middle Paleolithic assemblages from Lakhmapur, Malaprabha Valley, India. (a) Prepared cores in early Middle Paleolithic assemblages are associated with (b,c) diminutive hand axes and cleavers. (a) Adapted from Petraglia MD, Schuldenrein J, and Korisettar R (2003) Landscapes, activity, and the Acheulean to Middle Paleolithic transition in the Kaladgi Basin, India. *Journal of Eurasian Prehistory* 1(2): 3–24.

The period between 45 and 30 ka in the Levant is marked by the extinction of the Neanderthals and the expansion of modern humans equipped with Upper Paleolithic adaptations (Shea, 2003). Neanderthals throughout western and Central Eurasia appear to have been well-adapted to temperate woodlands and cold steppes, with adaptations geared toward meat acquisition. Economic intensification in later Neanderthal adaptations in the Levant are indicated by multiseasonal site occupations, a broadening of the subsistence base, and production of spear points to procure prey (Shea, 2003). Though Neanderthals may have had complex social and economic behaviors, small behavioral and adaptive advantages on the part of modern humans may have eventually led to their replacement by modern humans.

In addition to the expansion of modern humans into the Levant, populations may have used the Bab al Mandab strait

across the southern end of the Red Sea for expansions during OIS 5, as is exemplified by the presence of Middle Paleolithic sites in southern Arabia with Northeast African affinities (Petraglia and Alsharekh, 2003; Rose, 2004). Though it is clear that modern humans reached Australia by 45–42 ka and possibly earlier (Bowler et al., 2003; O'Connell and Allen, 2004), the spread of modern humans across Asia has not been documented from skeletal evidence or archaeological assemblages, and remains at present a hypothesis based largely on modern genetic evidence. Y-chromosome and MtDNA data suggest the colonization of South Asia by modern humans originating in Africa, sometime between 73 and 55 ka (Kivisild et al., 2003). The genetic data suggest a single, early migration was responsible for the initial settlement toward South Asia and toward the east (Thangaraj et al., 2005). It has been suggested that modern movements into South Asia may have

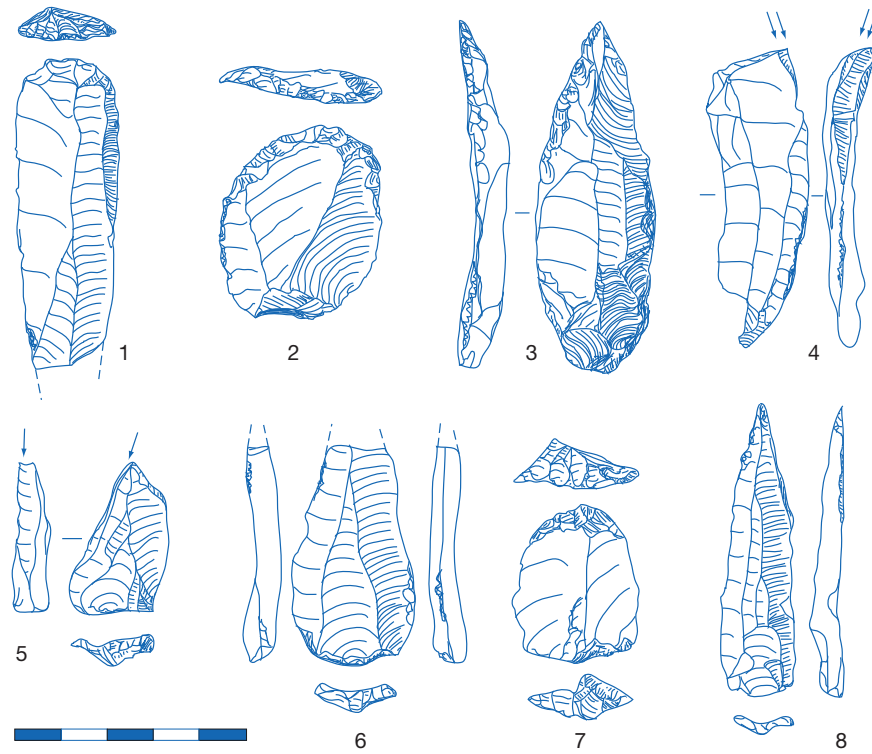


Figure 3 Early Upper Paleolithic artifacts from Üçağızlı Cave, Turkey. Ahmarian (1–4) and initial Upper Paleolithic (5–8) tools. Adapted from Kuhn SL (2004) Upper Paleolithic raw material economies at Üçağızlı cave, Turkey. *Journal of Anthropological Archaeology* 23: 431–448, figure 3.

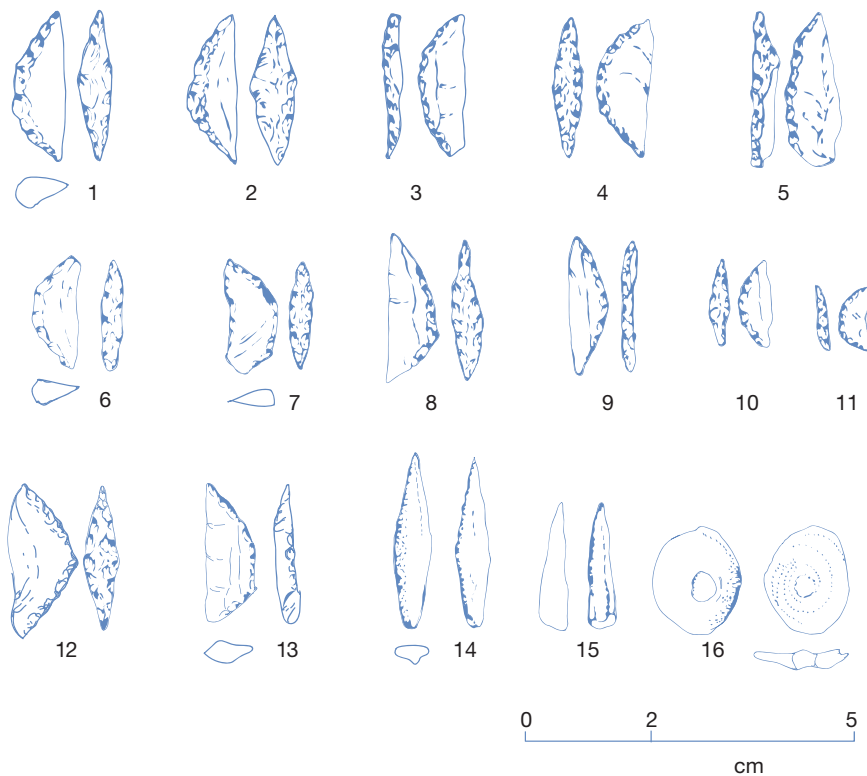


Figure 4 Microlithic artifacts from Sri Lanka. Geometric microliths (1–13), bone points (14, 15) and bead (16). Adapted from Deraniyagala SU (1992) *The Prehistory of Sri Lanka: An Ecological Perspective*. Colombo: Department of the Archaeological Survey, Government of Sri Lanka, figure 59.

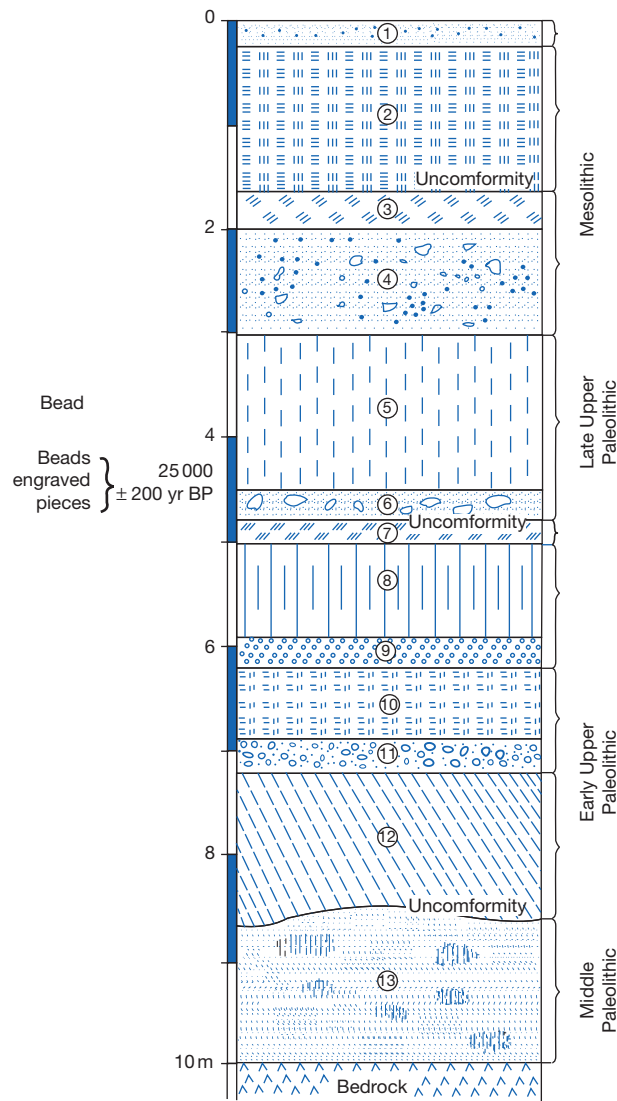


Figure 5 Late Pleistocene stratigraphic profile of Patne, India, showing Middle to Mesolithic sequence and location of symbolic artifacts and radiocarbon date. Adapted from Sali SA (1989) *The Upper Palaeolithic and Mesolithic Cultures of Maharashtra*. Pune: Deccan College Post Graduate and Research Institute, figure 11.

originally been accompanied by a Middle Paleolithic technology, presenting problems in discriminating such assemblages from those being used by resident populations of archaic hominins (James and Petraglia, 2005). Though modern humans were probably present in South Asia sometime after 70 ka, evidence for modern human behavior develops slowly and such signs are temporally and spatially patchy. Because stone tool assemblages in India before the great Toba volcanic eruption of 74 ka are similar to subsequent ones that were likely made by *H. sapiens*, it is possible that our species was already in South Asia before 74 ka (Petraglia et al., 2007). The development of new cultural innovations in the Late Paleolithic of South Asia may be related, in part, to fluctuations in the environment and demographic changes, indicating that it may be difficult to discern movements of modern humans. Modern humans probably reached east and south Eastern

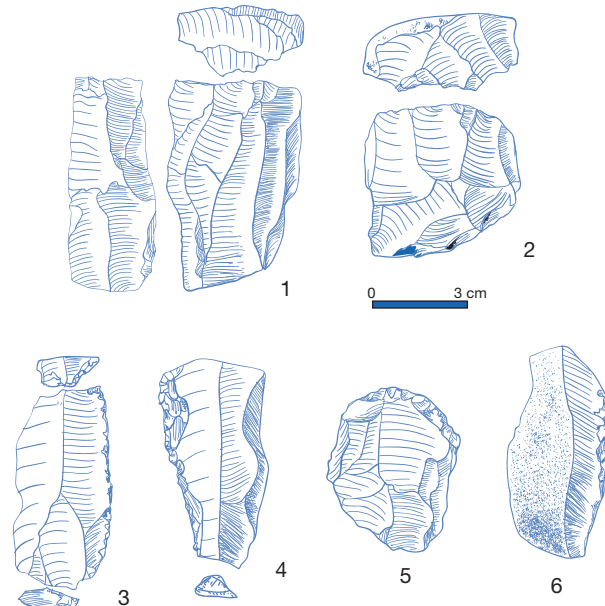


Figure 6 Initial Upper Paleolithic assemblages from Shuidonggou, China (adapted from Brantingham *et al.* (2001), figure 5). Flat-faced 'Levallois' cores (1–2); Levallois blades (3–4); retouched flake-scraper (5); side-scraper (6).

Asia by 40 ka, as indicated by the remains of modern humans at Tianyundong Cave, North China (Shang et al., 2010) and Niah Cave, Borneo (Barker, 2002), both dated at ca. 40–43 ka. Even earlier indications of *H. sapiens* in East Asia include a metatarsal from Callao Cave, the Philippines (Mijares et al., 2010) and a cranium from Tam Pa Ling, Laos, that has been directly dated at ca. 63 ka (Demeter et al., 2012). Though encounters between archaic and modern humans are unclear across most of Asia, the replacement of indigenous archaic populations was probably completed by 18 ka, with the extinction of hominins such as *H. floresiensis* (Brown et al., 2004) in remote island settings. In Java, *H. erectus* may have been extinct before the arrival of *H. sapiens*, as the Ngandong assemblage of *H. erectus* has recently been re-dated to 143–546 ka (Indriati et al., 2011).

The Colonization of New Territories after 125 ka

Before the last interglacial, hominin populations inhabited the same parts of Asia as had been colonized in the middle Pleistocene, that is, up to ca. 40–45° N, with perhaps short forays further north when climatic conditions permitted. As with so much of early Asian prehistory, more dates are needed to indicate when and how the expansion of humans across the whole of Asia occurred. Neanderthals in Central Asia appear to have been the first to expand northward into southern Siberia up to 55° N, probably in the early part of OIS 4 in the last glaciation, and possibly earlier. This process of colonization accelerated after 40 ka, and was most likely accomplished by modern humans, which colonized the Japanese islands by 40 ka (most likely by sea), and began to colonize areas >55° N after 26 ka (Goebel, 1999). Modern humans rather than

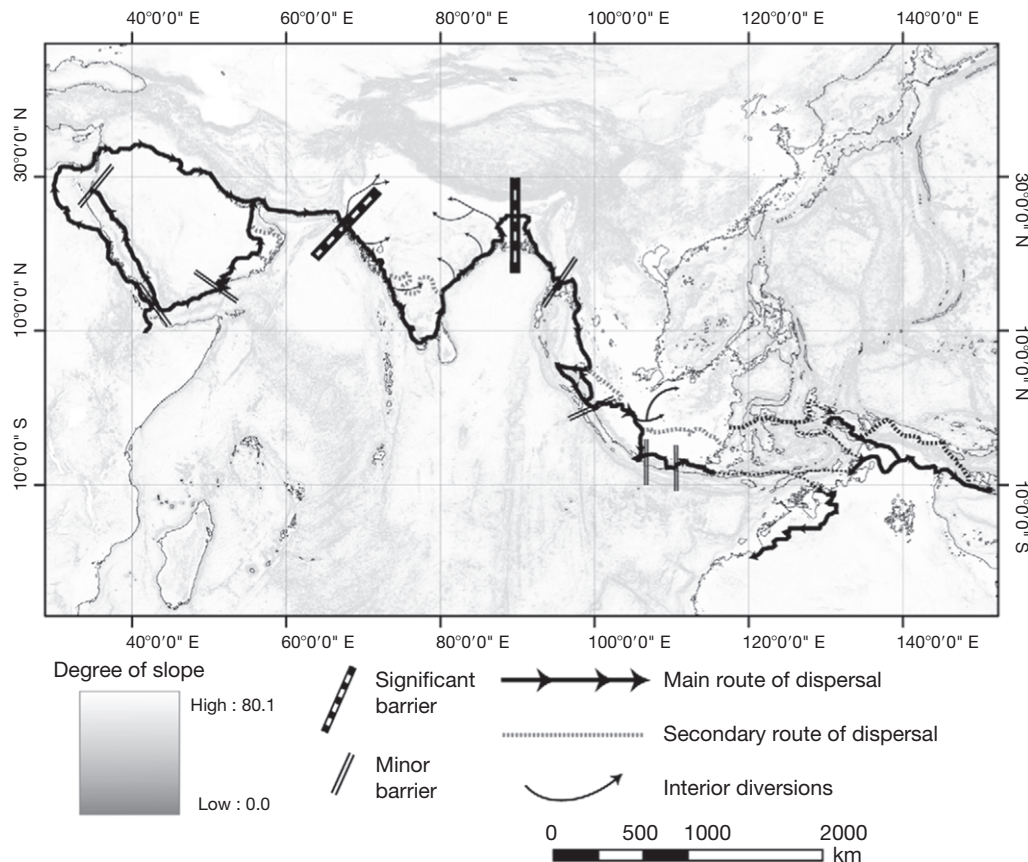


Figure 7 Potential routes of anatomically modern humans during OIS 4, showing routes and barriers. Adapted from Field JS and Lahr MM (2006) Assessment of the southern dispersal: GIS-based analyses of potential routes at oxygen isotopic stage 4. *Journal of World Prehistory*, figure 14.

H. erectus may also have colonized the rainforest regions of SE Asia. Finally, the Tibetan Plateau, 5 million sq km in extent and averaging 5000 m a.s.l., was uninhabited until the Late Glacial, and was one of the last parts of Eurasia to be settled (Brantingham et al., 2001).

Settling and Sedentism: Developments after the Last Glacial Maximum

Major changes in subsistence occurred after the Last Glacial Maximum, ca. 16–18 ka, and resulted in the emergence of agricultural societies across much of Asia by 8 ka BP (Mithen, 2003). The background to this process was the ending of the last Ice Age and onset of modern climatic conditions. In Southeast Asia, the most conspicuous environmental change was the flooding of the Sunda Shelf, which resulted in the loss of >1.5 million sq km of land, or three-quarters the present day area of mainland Southeast Asia (Voris, 2000). The archaeology of foraging populations in the Levant between 20 and 14 ka is relatively well known (Bar-Yosef, 1998). The Kebaran is characterized by obliquely truncated blades, bladelets, and microliths, with the addition of groundstone implements, perhaps used to process nuts and grasses. The Kebaran is thought to reflect seasonal mobility by hunting and gathering groups between highland and lowland settings. Ohalo II, on the

shore of the Sea of Galilee of Israel, is a spectacular waterlogged site dated to ca. 20 ka (Nadel, 2002). A community geared toward fishing and hunting-gathering left behind brush huts, hearths, hundreds of thousands of charred seeds/fruits, and fish and animal bones (Figure 8). During a humid interval at around 14.5–13 ka, an expansion of settlement occurs into the interior and highland zones of the Negev, Sinai, and southern Jordan. Climatic improvement appears to be responsible for population expansions into the steppes and desertic areas.

The date of 11 ka BP is usually given as the beginning of food production based on the domestication of relatively few wild plant and animal species (Diamond and Bellwood, 2003). Until the end of the Pleistocene, all people in Asia lived as hunter-gatherers. The introduction of food production led to major changes in subsistence and settlement practices in some populations, whereas in other populations it led to territorial marginalization or to even to extinction. Some foraging populations continued to live alongside Neolithic peoples, sometimes engaging in trade and exchange with settled groups.

The emergence and subsequent expansion of Natufian populations in the Levant is dated to about 13–10 ka (Bar-Yosef, 1998) (Figure 9). Some large sites, such as Ain Mallaha, hint at prolonged occupations based on substantial semi-subterranean dwellings and multiple superimposed occupation floors. Stone tool assemblages include numerous crescentic microliths and bone, and horncore tools include

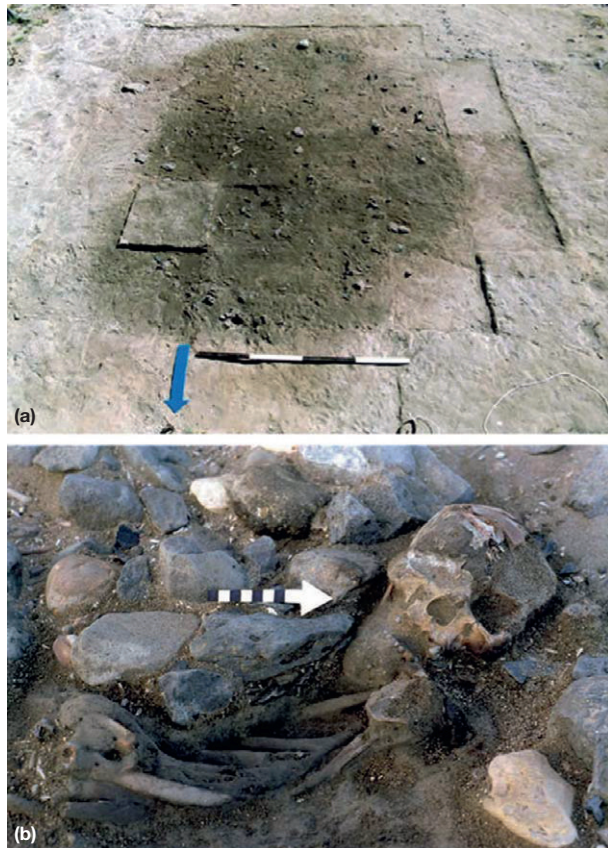


Figure 8 Ohalo II excavations, showing (a) hut remains and (b) human skeleton. Photographs by D. Nadel.

harpoons, fishhooks, spear and arrow points, awls, and needles. Large groundstone querns are often times found and may have functioned in the processing of nuts or cereals. The availability of cereal grasses and semisedentary lifestyles were likely major contributing factors in leading to agriculture. The Pre-Pottery Neolithic witnessed an increase in the number and diversity of sites, and changes in settlement patterns may be partly a reflection of climatic changes occurring between 11 and 10 ka (Bar-Yosef, 1998). Abu Hureyra, located in Syria and occupied from about 11.5 to 7 ka, is a good example of changes in lifestyle and economy that occurred over this period (Moore et al., 2000). The first period of habitation, before agricultural development, included a settlement that appears to represent year-round occupation. The populations gathered a variety of wild seeds, including cereals, and hunted migrating gazelle populations. After about 10 ka, Abu Hureyra was inhabited by permanently settled communities who had cultivated and processed a variety of domesticated seeds, such as oats, barley, chickpeas, emmer, and lentils. Such changes in Pre-Pottery Neolithic communities were precursors to more significant technological, social, and economic changes that were to take place throughout Western Asia.

By 13–8 ka, many parts of Asia were filled with complex hunting and gathering communities. Siberia, which was largely uninhabited during the Last Glacial Maximum, was rapidly recolonized up to the Arctic Ocean by 12.5 ka (Goebel, 1999). After 12.5 ka, Jomon sites in Japan appear to show production of pottery and sedentary lifestyles based on fishing, with supplementation of cereal grains by around 9 ka. Mesolithic sites are well represented in the Indian subcontinent, leading archaeologists to infer that the large number of sites relates to marked growth in human populations (Misra, 2001).

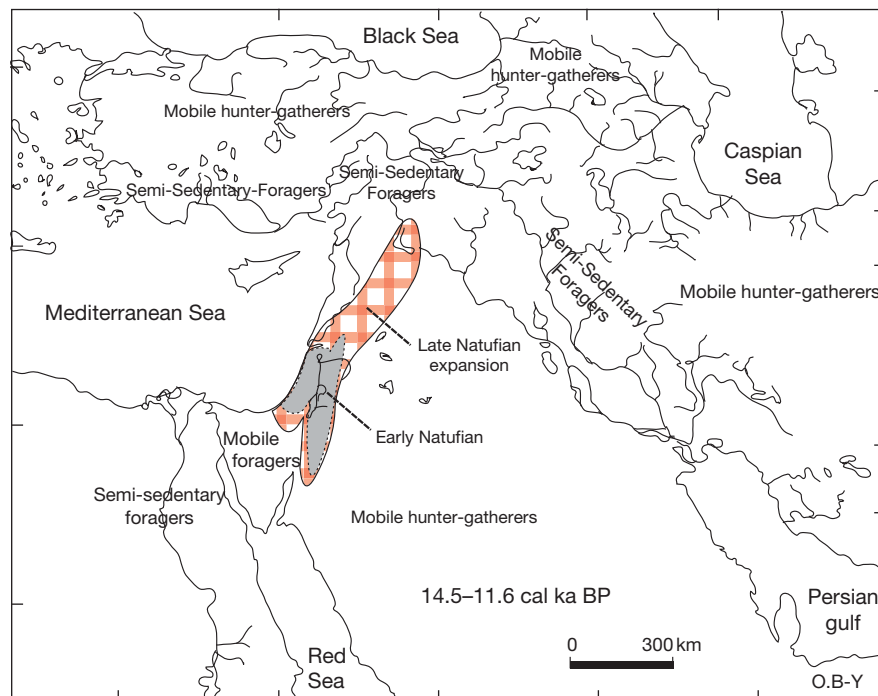


Figure 9 Map of initial Natufian origin and later expansions. Adapted from Bar-Yosef O (1998) The Natufian culture in the Levant, threshold to the origins of agriculture. *Evolutionary Anthropology* 6: 159–177, figure 1.

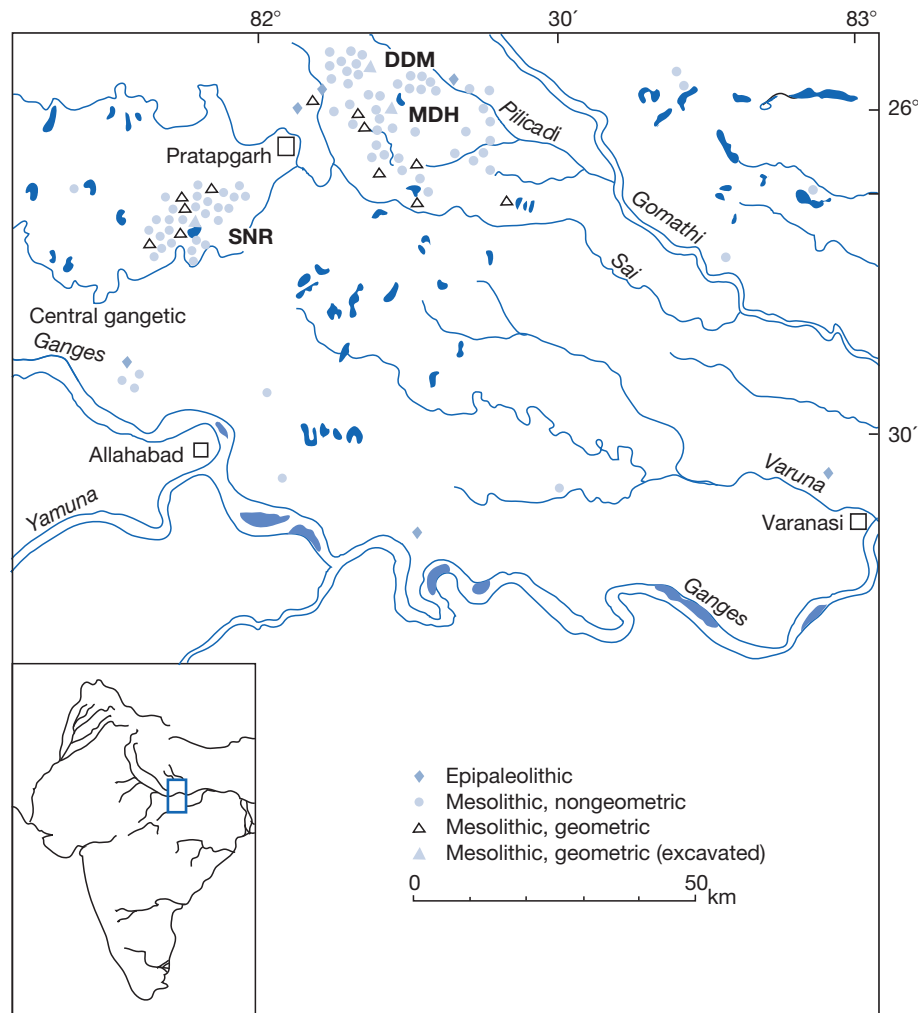


Figure 10 Map showing Epipaleolithic and Mesolithic sites of the central Ganga Plains. Adapted from Chattopadhyaya UC (1996) Settlement pattern and the spatial organization of subsistence and mortuary practices in the Mesolithic Ganges Valley, north-central India. *World Archaeology* 27(3): 461–476, figure 1.

The increased food supply available in the Mesolithic is thought to have led to a reduction in mobility, as reflected in the large size of sites, evidence for huts and communal hearths, the substantial appearance of habitation deposits, and the presence of cemeteries, particularly in the Ganga Valley (Chattopadhyaya, 1996) (Figure 10).

Recent archaeological evidence suggests two distinct centers of early rice cultivation in Asia (Fuller, 2007). In China, it is widely accepted that rice cultivation was underway in the Middle Yangzi, and adjacent South China by 8 ka, and in India, rice cultivation may date back to as early as ca. 9 ka. While further research is needed in South Asia, the recent evidence indicates at the very least that foragers were exploiting wild rice from about 9 ka. In northwestern South Asia, the dominant crops from the time of earliest evidence derived from the Southwest Asian Neolithic founder crops. Representatives of this crop package spread to western Pakistan by the time of Neolithic Mehrgarh after 9 ka (Jarrige et al., 2006), and

Central Asia by ca. 8 ka. Such subsistence and settlement shifts set into motion fundamental changes in the demography, health, and lifestyles of populations inhabiting Asia.

See also: [Archaeological Records: Overview](#); [2.7 Myr–300 000 Years Ago in Asia](#); [1.9 Myr–300 000 Years Ago in Europe](#); [Global Expansion 300 000–8000 Years Ago, Africa](#); [Global Expansion 300 000–8000 Years Ago, Australia](#); [Global Expansion 300 000–8000 Years Ago, Americas](#); [Neanderthal Demise](#); [Postglacial Adaptations](#); [Vertebrate Studies: Interactions with Hominids](#).

References

- Barker GWW (2002) Prehistoric foragers and farmers in South-east Asia: Renewed investigations at Niah Cave, Sarawak. *Proceedings of the Prehistoric Society* 68: 147–164.

- Bar-Yosef O (1998) The Natufian culture in the Levant, threshold to the origins of agriculture. *Evolutionary Anthropology* 6: 159–177.
- Bar-Yosef O (2002) The Upper Paleolithic revolution. *Annual Reviews of Anthropology* 31: 363–393.
- Bowler JM, Johnstone H, Olley J, et al. (2003) New ages for human occupation and climate change at Lake Mungo, Australia. *Nature* 421: 837–840.
- Brantingham PJ, Krivoshapkin AI, Jinzeng L, and Tserendagva Y (2001a) The initial Upper Paleolithic in Northeast Asia. *Current Anthropology* 22: 735–747.
- Brantingham PJ, Olsen JW, and Schaller GB (2001b) Lithic assemblages from the Chang Tang region, northern Tibet. *Antiquity* 75: 319–327.
- Brown P, Sutikna T, Morwood MJ, et al. (2004) A new small bodied hominin from the Late Pleistocene of Flores, Indonesia. *Nature* 431: 1055–1061.
- Cameron D, Patnaik R, and Sahni A (2004) The phylogenetic significance of the Middle Pleistocene Narmada hominin cranium from central India. *International Journal of Osteoarchaeology* 14: 419–447.
- Chattopadhyaya UC (1996) Settlement pattern and the spatial organization of subsistence and mortuary practices in the Mesolithic Ganges Valley, north-central India. *World Archaeology* 27(3): 461–476.
- Clarkson C, Petraglia M, Korisettar R, et al. (2009) The oldest and longest enduring microlithic sequence in India: 35,000 years of modern human occupation and change at the Jwalapuram Locality 9 rockshelter. *Antiquity* 83: 326–348.
- Clarkson C, Jones S, and Harris C (2012) Continuity and change in the lithic industries of the Jurreru Valley, India, before and after the Toba eruption. *Quaternary International* 258: 165–179.
- Demeter F, Shackelford LL, Bacon A-M, et al. (2012) Anatomically modern human in Southeast Asia (Laos) by 46 ka. *Proceedings of the National Academy of Sciences USA* 109(36): 14375–14380.
- Dennell RW and Petraglia MD (2012) The dispersal of *Homo sapiens* across southern Asia: how early, how often, how complex? *Quaternary Science Reviews* 47: 15–22.
- Deraniyagala SU (1992) *The Prehistory of Sri Lanka: An Ecological Perspective*. Colombo: Department of the Archaeological Survey, Government of Sri Lanka.
- Diamond J and Bellwood P (2003) Farmers and their languages: The first expansions. *Science* 300: 597–603.
- Field JS and Lahr MM (2006) Assessment of the southern dispersal: GIS-based analyses of potential routes at oxygen isotopic stage 4. *Journal of World Prehistory*.
- Fuller D (2007) Non-human genetics, agricultural origins and historical linguistics in South Asia. In: Petraglia MD and Allchin B (eds.) *The Evolution and History of Human Populations in South Asia: Inter-disciplinary Studies in Archaeology, Biological Anthropology, Linguistics and Genetics*, pp. 393–443. The Netherlands: Springer.
- Gao X and Norton C (2002) A critique of the Chinese “Middle Palaeolithic” *Antiquity* 76: 397–412.
- Glantz M, Viola B, Wrinn P, et al. (2008) New hominin remains from Uzbekistan. *Journal of Human Evolution* 55: 223–237.
- Goebel T (1999) Pleistocene human colonization of Siberia and peopling of the Americas: An ecological approach. *Evolutionary Anthropology* 8(6): 208–227.
- Gopher A, Barkai R, Shimelmitz R, et al. (2005) Qesem Cave: An Amudian site in Central Israel. *Journal of the Israel Prehistoric Society* 35: 69–92.
- Indriati E, Swisher C III, Lepre C, et al. (2011) The age of the 20 meter Solo River Terrace, Java, Indonesia and the survival of *Homo erectus* in Asia. *PLoS ONE* 6(6): e21562.
- Ingman M, Kaessmann H, Paabo S, and Gyllenstein D (2000) Mitochondrial genome variation and the origin of modern humans. *Nature* 408: 708–713.
- James HVA and Petraglia MD (2005) Modern human origins and the evolution of behavior in the Later Pleistocene record of South Asia. *Current Anthropology* 46: S3–S27.
- Jarrige J-F, Jarrige C, and Quivron G (2006) Mehrgarh Neolithic: The updated sequence. In: Jarrige C and Lefèvre V (eds.) *South Asian Archaeology 2001*, pp. 129–142. Paris: ADFP Éditions Recherche sur les Civilisations.
- Jia L and Huang W (1985) The Late Palaeolithic of China. In: Wu R and Olsen JW (eds.) *Palaeoanthropology and Palaeolithic Archaeology in the People's Republic of China*, pp. 211–223. London: Academic Press.
- Kivisild T, Rootsi S, Metspalu M, et al. (2003) The genetic heritage of the earliest settlers persists both in Indian tribal and caste populations. *American Journal of Human Genetics* 72: 313–332.
- Krause J, Orlando L, Serre D, et al. (2007) Neanderthals in central Asia and Siberia. *Nature* 449: 902–904.
- Krause J, Fu Q, Good JM, et al. (2010) The complete mitochondrial DNA genome of an unknown hominin from southern Siberia. *Nature* 464: 894–897.
- Kuhn SL (2002) Paleolithic archaeology in Turkey. *Evolutionary Anthropology* 11: 198–210.
- Kuhn SL (2004) Upper Paleolithic raw material economies at Üçaizlı cave, Turkey. *Journal of Anthropological Archaeology* 23: 431–448.
- Lahr MM (1996) *The Evolution of Modern Human Diversity*. Cambridge: Cambridge University Press.
- Lahr MM and Foley R (1994) Multiple dispersals and modern human origins. *Evolutionary Anthropology* 3: 48–60.
- McBrearty S and Brooks A (2000) The revolution that wasn't: A new interpretation of the origin of modern human behaviour. *Journal of Human Evolution* 39: 453–563.
- Mellars P (2006) Going east: new genetic and archaeological perspectives on the modern human colonization of Eurasia. *Science* 313: 796–800.
- Mijares AS, Delfino F, Piper P, et al. (2010) New evidence for a 67,000-year-old human presence at Callao Cave, Luzon, Philippines. *Journal of Human Evolution* 59: 123–132.
- Miller-Antonio S, Schepartz L, Karkanas P, et al. (2004) Lithic raw material use at the Late Middle Pleistocene site of Panxian Dadong. *Asian Perspectives* 43(2): 314–332.
- Misra VN (2001) Prehistoric human colonization of India. *Journal of Bioscience* 26(4): 491–531.
- Mithen SJ (2003) *After the Ice: A Human Global History*. London: Weidenfeld and Nicolson.
- Moore AMT, Hillman GC, and Legge AJ (2000) *Village on the Euphrates*. Oxford: Oxford University Press.
- Murty MLK (1979) Recent research on the Upper Palaeolithic Phase in India. *Journal of Field Archaeology* 6: 301–320.
- Nadel D (ed.) (2002) *Ohalo II – a 23000 Year-Old Fisher-Hunter-Gatherers' Camp on the Shore of the Sea of Galilee*. Haifa: Hecht Museum.
- O'Connell JF and Allen J (2004) Dating the colonization of Sahul (Pleistocene Australia-New Guinea): A review of recent research. *Journal of Archaeological Science* 31: 835–853.
- Petraglia MD and Alsharekh A (2003) The Middle Palaeolithic of Arabia: Implications for modern human origins, behaviour and dispersals. *Antiquity* 77(298): 671–684.
- Petraglia MD, Alsharekh A, Crassard R, et al. (2011) Middle Palaeolithic occupation on a Marine Isotope Stage 5 lakeshore in the Nefud Desert, Saudi Arabia. *Quaternary Science Reviews* 30: 1555–1559.
- Petraglia MD, Clarkson C, Boivin N, et al. (2009) Population increase and environmental deterioration correspond with microlithic innovations in South Asia ca. 35,000 years ago. *Proceedings of the National Academy of Sciences USA* 106: 12261–12267.
- Petraglia MD, Korisettar R, Boivin N, et al. (2007) Middle Palaeolithic assemblages from India before and after the Toba super-eruption. *Science* 317: 114–116.
- Petraglia MD, Schuldenrein J, and Korisettar R (2003) Landscapes, activity, and the Acheulean to Middle Paleolithic transition in the Kalladgi Basin, India. *Journal of Eurasian Prehistory* 1(2): 3–24.
- Rightmire GP (2001) Comparison of Middle Pleistocene hominids from Africa and Asia. In: Barham L and Robson-Brown K (eds.) *Human Roots: Africa and Asia in the Middle Pleistocene*, pp. 123–135. Bristol: Western Academic and Specialist Press.
- Rose JI (2004) New evidence for the expansion of an Upper Pleistocene population out of East Africa, from the site of Station One, northern Sudan. *Cambridge Archaeological Journal* 14(2): 205–216.
- Sali SA (1989) *The Upper Palaeolithic and Mesolithic Cultures of Maharashtra*. Pune: Deccan College Post Graduate and Research Institute.
- Shang H, Tong H, Zhang S, et al. (2010) An early modern human from Tianyuan Cave, Zhoukoudian, China. *Proceedings of the National Academy of Sciences USA* 104: 6573–6578.
- Shea JJ (2003) The Middle Paleolithic of the East Mediterranean Levant. *Journal of World Prehistory* 17: 313–394.
- Stringer C (2002) Modern human origins: Progress and prospects. *Philosophical Transactions of the Royal Society, London B* 357: 563–579.
- Thangaraj K, Chaubey G, Kivisild T, et al. (2005) Reconstructing the origin of Andaman Islanders. *Science* 308: 996.
- Thorne AG and Wolpoff MH (1992) The multiregional evolution of humans. *Scientific American* 266(4): 76–83.
- Voris H (2000) Maps of Pleistocene sea levels in Southeast Asia: Shorelines, river systems and time durations. *Journal of Biogeography* 27: 1153–1167.
- Wu XZ and Poirier FE (1995) *Human Evolution in China*. New York: Oxford University Press.

Global Expansion 300 000–8000 Years Ago, Australia

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Glossary

Artifact: Any object shaped and made by humans

BP: Radiocarbon years before the present, conventionally measured from 1950

Obsidian: A natural glass produced from erupting volcanoes. Molten rock ejected from the crater solidifies extremely rapidly prohibiting the growth of crystals and forming glass

Luminescence dating: A method of dating inorganic material like clay and sediments. Electrons

built up and trapped in the crystal lattice structure of quartz grains are artificially stimulated by light that releases them. The amount of energy given off is an indication of how long these sediments have been buried

Taphonomy: Study of the laws of burial, particularly transformation of organics from the biosphere to the lithosphere

Introduction

The human colonization of the Sahul landmass, the continent that joined New Guinea, Australia, and Tasmania after dry land was exposed due to lowered sea level, was a significant event in our species global expansion. It heralded the first sea crossing by peoples whose ancestors spread out of Africa perhaps 70 000 years ago (ya). It would have involved the construction of water craft sophisticated enough to make voyages of up to 100 km from Timor and lesser distances through the Indonesian archipelago. Clearly, organizational skills and language must have been present to make such trips. Crossing this water gap brought people into a world that, in some respects, would have been familiar while in others unique. Many plants would have been recognizable especially rainforest varieties, as well as marine life but most animals encountered would have been alien to them. On passing through the biogeographical transitional zone of Wallacea and across Lydekkers Line, which forms the faunal boundary of the Australian region, people entered a land of giant marsupials, egg-laying mammals, giant flightless birds, and enormous reptiles (Figure 1).

Unlike the Eurasian fauna human groups would have co-existed with, there were no large carnivorous pack animals such as hyenas or wolves. Large cave bears and felids were absent as were smaller carnivores like the fox (until European settlement), wolverine, and lynx. Only three marsupial carnivores were present, the thylacine (*Thylacinus cynocephalus*), Tasmanian devil (*Sarcophilus harrisii*), and the marsupial lion (*Thylacoleo carnifex*), the former two surviving across the continent until about 3000 years before present (BP) and in Tasmania into modern times (Figure 2). Indeed Sahul, like the Americas, also had no earlier hominins. This is despite the recent Indonesian discoveries of *Homo erectus* and their 'Hobbit' descendents, *H. floresiensis*, living between 800 000 and 12 000 ya just 100 km from Sahul's coastline (Morwood et al., 2004). As yet, there is no convincing evidence that premodern humans had colonized Sahul (O'Connell and Allen, 2004). The people that entered the continent were behaviorally modern humans.

Paleoenvironments

Research on the past climate of Sahul (Pleistocene New Guinea–Australia–Tasmania) has identified major changes in sea levels and terrestrial environments. These have had a significant influence on the ecological structure of this continent, affecting the distribution of potential food and the ability of humans to make a living off the land, especially around the coasts, mountains, and arid interior. Human responses were not uniform however and different behavioral patterns are documented in archaeological sites from different regions. These climatic changes were driven by large-scale Milankovitch Cycles that changed the distribution of incoming solar radiation permitting the buildup of ice at the poles. Vast amounts of water were frozen and locked up as glaciers, reducing the sea levels by at least 100 m across the globe. Research on deep-sea cores and uplifted coal terraces on the Huon Peninsula, Papua New Guinea, suggest that sea levels have fluctuated over the last 100 000 years or more. The last glacial cycle began about 130 000 ya with a series of sea-level rises and falls trending toward lower levels until about 18 000 ya after which the sea rose, stabilizing about 6000 years BP (Lambeck and Chappell, 2001).

Previously, it had been thought that the last glacial was a spike of much colder temperatures and lower precipitation between 20 000 and 18 000 ya. New data suggest that it was more active over a longer period with increased magnitude and frequency from about 30 000 ya (Lambeck et al., 2002). The climatic cycles appear to have oscillated rapidly approaching the Last Glacial Maximum (LGM). These changes would have had implications for humans over the medium term with influences on the distribution of prey animals and plant foods. More significantly for people, embedded within these large environmental fluctuations were smaller, more dramatic El Niño–Southern Oscillation (ENSO) events that began at least 45 000 ya (Kershaw et al., 2003). During this period there was an increase in fire-tolerant vegetation and a steady decline in precipitation that in turn increased periods of dust circulation blowing west out over the Indian Ocean and east into the

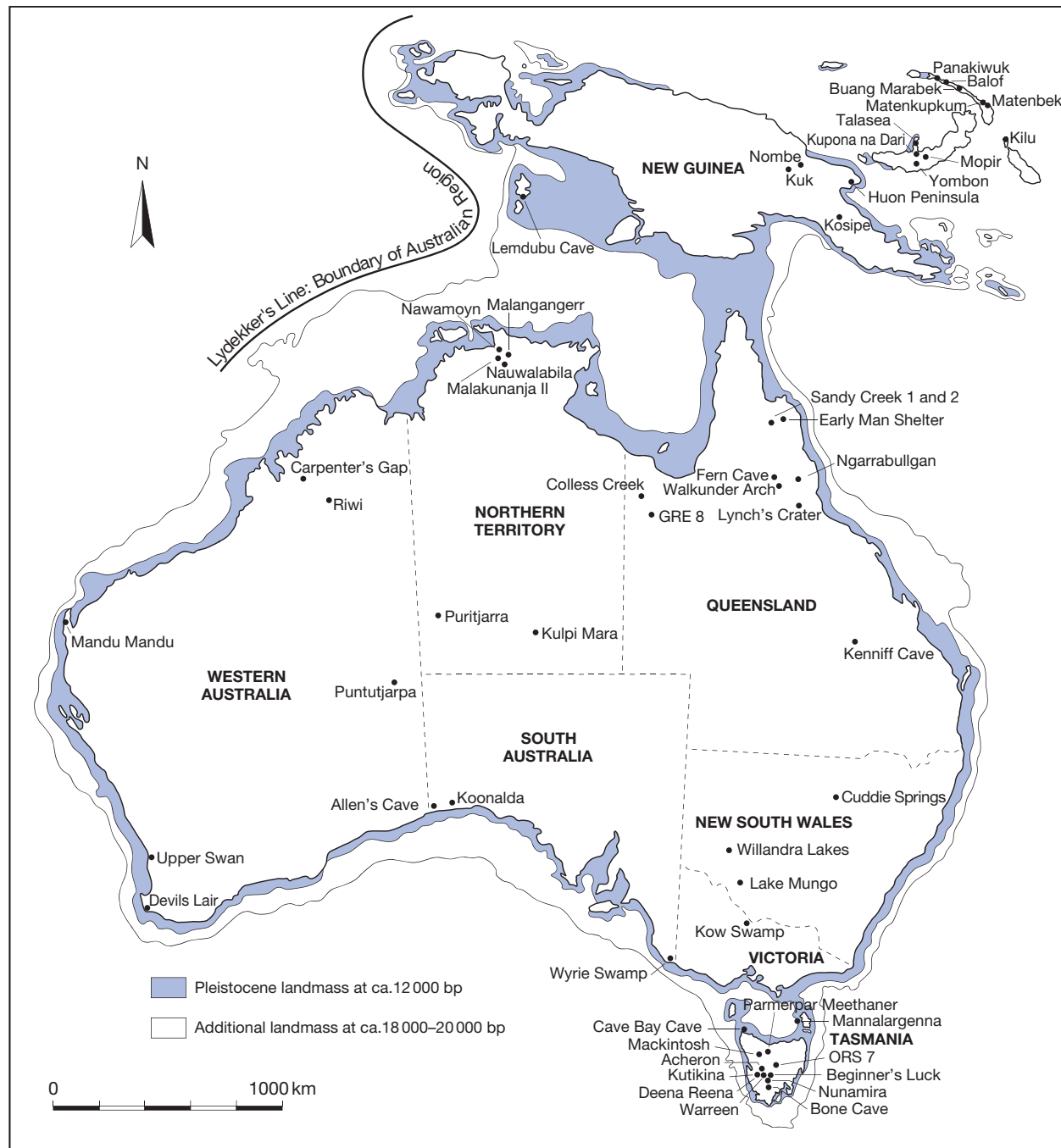


Figure 1 Selected late Pleistocene sites from New Guinea, Australia, and Tasmania. During this time the continent expanded to over 11.5 million square kilometers when sea levels dropped to 130 m below present day levels.

Tasman Sea (Hesse and McTainsh, 2003; Shulmeister et al., 2004). Although some researchers have suggested wind speeds were at least 8 km/hour greater during the LGM than present day, dust particle analysis would indicate that it was more arid than windier.

Terrestrial evidence from the Willandra Lakes system in southeastern Australia suggests alternating wet and dry periods recorded in the lakeside lunettes as sand and clay layers, respectively (Bowler, 1998). These support evidence

from the marine chronology for moister conditions before 43 000 years BP when the lake system was full. A sharp but short-lived decline in water levels was recorded around 40 000 years BP. It is at this time that Australia's oldest human remains were recorded, Mungo 3. From this period the lake levels fluctuated with less severe conditions represented by alternating clay and gravel deposits. Freshwater shell midden and aquatic fauna are abundant in human sites around the lakes at this time. Fishing with gill nets has been

argued based on the remains of fish otoliths in the middens. After 16 000 years BP there was a steady decline in human occupation as the lakes dried up completely.

In the tropical north, pollen cores from Lynch's Crater on the Atherton Tableland have provided detailed sediment records back to at least 130 000 years BP (Kershaw, 1994; Figure 3).

From then until about 78 000 years BP rainforest angiosperms were dominant with some minor dry sclerophyll woodland species. Increasing amounts of *Eucalyptus* pollen, particularly around 35 000 years BP indicates further drying in northern Australia concomitant with changes in southern Australia. Increases in charcoal around 45 000 years BP have been argued to represent human occupation of the Tableland (Turney et al., 2001) although recent archaeological



Figure 2 The Tasmanian Devil (*Sarcophilus harrisi*) remains Australia's sole marsupial carnivore. Unlike *Thylacinus cynocephalus* and *Thylacoleo carnifex* it is primarily a scavenger (photograph by R. Cosgrove).



Figure 3 Lynch's Crater on the Atherton Tablelands, north Queensland has a pollen record stretching back at least 130 000 years. The extinct volcano contains a long continuous record of peat accumulation which records the change from rainforest to fire-adapted species like *Eucalyptus* around 30 000 years BP until 8000 years BP when rainforest again invaded the landscape (photograph by R. Cosgrove).

research suggests the earliest dates for humans here is around 7500 years BP (Cosgrove, 2005). A major period of glacial advance and global cooling after 30 000 years BP saw sclerophyll vegetation become dominant on the Atherton Tablelands, all but replacing rainforest species (Kershaw, 1994). Some tropical rainforest refugia remained that sustained endemic species like the tree kangaroo and cassowary into the Holocene. It was only after 8000 years BP that rainforests began to expand again after increases in rainfall during the climatic optimum.

The arid center of Australia gradually become drier, particularly after about 20 000 years BP with increased dust and lower temperatures enhanced by reduced vegetation cover and a trend towards grass species (Field et al., 2003; Hesse and McTainsh, 2003; Pack et al., 2003). At this time, the complete drying of lake systems around the arid margins by 16 000 years BP led to reactivation of dune and lunette building activity (Bowler, 1998). Other environmental changes to Australia's Pleistocene ecosystems have been said to be a direct result of human firing regimes (Miller et al., 2005) which led to the extinction of *Genyornis newtonii*, a large flightless bird in central Australia 50 000 years BP. Results from other studies on the edge of the arid zone show this bird co-existed with humans until at least 30 000 years BP (Trueman et al., 2005).

New Guinea and Tasmania were finally cut off from Australia by rising seas toward the end of the LGM around 9000 and 14 000 years BP, respectively.

Chronology of Human Arrivals

There has been continuing debate about the timing of human entry into Australia. Several challenges have confronted archaeologists when addressing this question. The first and most contentious is the accuracy of the range of dating techniques, the dynamics of site formation processes, and taphonomy. The protagonists fall into two main groups; those advocating settlement 40 000–45 000 years BP and those arguing for occupation before 60 000 years BP. There have been claims of occupation before 100 000 years BP but so far direct archaeological evidence for this has been lacking.

Much of the debate hinges on demonstrating the clear and unequivocal relationships between dates, artifacts, and sediments (O'Connell and Allen, 1998, 2004). Because the timing of human occupation intersects with the present known limits of radiocarbon, it has been necessary for geochronologists to refine dating methods and to understand archaeologists' concerns regarding site-formation processes. Developments in accelerator mass spectroscopy (AMS) and luminescence dating have provided a better chronological framework to the problem but have exposed weaknesses in the application of the results to archaeological records. Concerns have been raised about the taphonomic effects of termites, contamination by older sediments of younger archaeological deposits, movement of stone tools within sandy deposits, and the inadequate bleaching of the quartz grains to reset the luminescence clock (Figure 4).

O'Connell and Allen (2004) list eight sites that they argue have unequivocal archaeological evidence for earliest human occupation between 40 000 and 45 000 years BP.

Geographically these sites are located across the continent, from the northern tropics (e.g., Riwi, Huon Peninsula, and Carpenter's Gap), to the temperate southern and arid regions of Devils Lair, Allen's Cave and Lake Mungo. More controversial is the claim for artifacts in the oldest Lower Mungo sedimentary unit dated to about 50 000 ya (Bowler, 1998). Other researchers remain unconvinced since no refitting of artifacts from this level has been undertaken (O'Connell and Allen, 2004). In higher levels, stone flakes have been displaced from their cores by up to 60 cm, implying similar mechanisms for reworking the sediments and their contents in the lower layers. If the post-50 000 years BP chronological patterns hold, it suggests that a wide variety of environments were settled relatively swiftly, within 5000 years of initial landfall. O'Connell and Allen dismiss the two sites claimed to be 50 000 years BP of Malakunanja and Nauwalabila as having unconvincing associations of stratigraphy, chronology, and artifacts (Figure 5).

By the time the dry Bass Plain had been exposed (40 000 years BP), people took the opportunity to cross into what was then the toe of Sahul, now Tasmania. Here evidence based on radiocarbon chronology suggests Tasmania was first settled 35 000 years BP. The two earliest dates come from Warreen Cave and Parmerpar Meethaner in the southwest and north of Tasmania, respectively (Cosgrove, 1999; Figures 6 and 7).

In Melanesia, humans had settled in the Bismarck Archipelago and the Solomon Islands by at least 40 000 ya at Buang Marabek and 29 000 ya at Kilu (Leavesley et al., 2002). By 20 000 ya people were using boats to move possums and obsidian across water gaps as a means of subsistence and technological adaptations to new lands (Allen et al., 1989). There is also data to show humans had begun utilizing tropical rainforest on New Britain at the open site of Yombon by 35 000 years BP (Pavildes, 2004). The spread and relative speed of settlement across Sahul suggests that environmental

barriers did not restrain early modern human movement. They appear to have taken the opportunities to expand their range and adapt to the many and varied ecosystems of the Sahul land mass. These achievements point to an established social and economic system, probably already in place before they crossed the sea separating Southeast Asia and Sahul (Figures 8 and 9).

Physical Anthropology

Much of what we know about the physical anthropology of the earliest Australians has come from southeast Australia.



Figure 5 The site of Nauwalabila is claimed to be at least 50 000 years old (photograph by J. Allen).



Figure 4 Dating the earliest human arrivals in Sahul has been contentious. Recent use of optically stimulated luminescence (OSL) techniques has pushed back the ages to between 45 000 and 55 000 ya. Problems arise through contamination and the inadequate resetting of the luminescence clock. Termite activity can be a source of this because they continually circulate sand grains above and below ground. A large termite mound is located within the overhang of this rock shelter and when it collapses may contribute to this problem (photograph by J. Allen).



Figure 6 Parmerpar Meethaner is one of the oldest Tasmanian sites dating back to at least 34 000 years BP. It is located in the Forth River valley in northern Tasmania (photograph by R. Cosgrove).

Kow Swamp and Lake Mungo have provided much of this evidence although there are widely differing interpretations of these human fossil remains. The debate largely stems from the two competing models of the origin and global spread of humans known as the Out-of-Africa and Regional Continuity (Mulvaney and Kamminga, 1999). The former asserts that anatomically and behaviorally modern humans moved out of Africa perhaps 100 000 ya replacing an earlier migration of hominid forms represented by robust *H. erectus* and *H. ergaster*



Figure 7 The archaeological section of Parmerpar Meethaner reveals a series of layers containing stone artifacts, bone, and charcoal. The upper 30 cm is Holocene dating to between 3500 years BP and modern times. From the middle section to the base of the trench ages range from 18 000 to 34 000 years BP. A date of 40 000 years BP in the lowest section does not contain any human artifacts (photograph by R. Cosgrove).



Figure 8 Matemkupkum on New Ireland dates back at least 35 000 years. It contains the oldest evidence of sea fishing anywhere in the world (photograph by R. Frank).

that settled in Eurasia and Southeast Asia. Some archaeologists question the chronology although arguments have been mounted to show anatomically modern humans did not establish beyond the Levant after reaching there 70 000 years BP. It is only after 45 000 years BP that there is unequivocal data to show behaviorally modern humans reaching Southeast Asia and Sahul. Others propose an earlier movement of modern humans out of Africa earlier, around 70 000 years BP along the southern coasts of Eurasia into Southeast Asia, and then perhaps into Sahul. However there is no reliable archaeological evidence to suggest a human landfall of this antiquity apart from problematic dating of the Mungo 3 skeleton at around 60 000 years BP (Bowler and Magee, 2000; Gillespie and Roberts, 2000; Thorne and Curnoe, 2000; Figure 10).

Put simply, the Regional Continuity model proposes that there was only one migration out of Africa by *H. erectus/ergaster* over 1 million ya whose ancestors evolved *in situ* into regional populations of later modern humans. Physical anthropologists who argue for this model stress change through time from robust to gracile morphology through gene flow between geographically separate populations. However, significant gaps in the fossil record and the appearance of similar cranial morphologies on different continents 10 000 km apart continue to hamper the testing of this model (Stringer, 2002). In addition, the new finds from Flores and their late dates suggest that *H. erectus* did not evolve into later modern humans but remained a separate species until only 12 000 ya in the form of *H. floresiensis* (Morwood et al., 2004). Its discovery on the



Figure 9 Yombon has a series of volcanic ash layers seen here as alternating dark and light bands. Dr Christina Pavlides indicates the lowest late Pleistocene layer dated to 35 570 years BP. Over 20 chert stone tools were found here. The raw material for these artifacts was quarried locally while the obsidian found in the layers above was imported from the other side of the island (photograph by C. Pavlides).

island of Flores poses more questions than answers and reflects the complex nature of the human evolutionary process.

The Continuity Model asserts that the earliest Australian arrivals were essentially robust, represented by the incomplete WLH50 skull from the Willandra Lakes, although others argue that its form is the result of pathology, large size, or deformation. The model posits two waves of migration into Australia; an early one represented by WLH50 from Java and a later one represented by more gracile morphology from China. The proponents argue that the present day Aboriginal populations are the result of a mixture of the robust and gracile groups during the Holocene. A second competing view sees the presence of the two morphologies as the result of skeletal variations and sexual dimorphism within a single population. It has also been argued that cultural isolation and artificial head binding also influenced the morphological characteristics found in the Kow Swamp population (Brown, 2005; Pardoe, 1995).

Both models imply an increasing gracilization over time, where robust forms are followed by gracile skeletal morphology, particularly of the skull. The interesting aspect of the Australian data is that the earliest skeletal remains at Lake Mungo are gracile. These date to around 40 000–45 000 years BP. Mungo 3 and Mungo 1 contrast with the robust remains from Kow Swamp, Cohuna, and the Willandra Lakes hominid WLH50. These latter specimens are all dated younger than 20 000 years BP, and in the case of the best preserved at Kow Swamp, date only between 13 000 and 9500 years BP. That the Mungo fossils predate the robust group by a considerable time implies a significant degree of diversity and variation in Aboriginal populations in Sahul. The evidence does not support the notion of separate ancestral lineages (Stringer, 2002). It suggests a more complex situation involving multiple waves of people entering Sahul, affected morphologically by the combined effects of a harsh glacial climatic, generally low population numbers, and cultural and/or geographic isolation. Only one thing is clear, that the samples are small and represent individuals rather than populations. The study of a more representative sample of pre-Holocene skeletal remains will bring a better understanding of the past population dynamics of Sahul.



Figure 10 Lake Mungo and the Walls of China. Here severe gully erosion can have a profound influence on the location of artifacts and bones. Although human settlement is well dated to about 45 000 years BP, there is still some disquiet about the integrity of some of the lowest artifacts found below the burial site of the Mungo 3 skeleton (photograph J. Allen).

Impacts of Humans on Environments

Two issues dominate the debates on human influences on past environments, the extinction of Sahul's megafauna and the effect of fire on the biota (Bowman, 1998; Field, 2004; Roberts et al., 2001; Wroe et al., 2004). More than 50 species of animals became extinct during the late Pleistocene, and while many weighed less than 40 kg, a significant proportion weighed over 100 kg. The most well-known include *Diprotodon*, *Sthenurus*, and the large flightless bird, *Genyornis*. Arguments over the causes and timing of their extinction have been intense. The explanations include the over-kill hypothesis or 'blitzkrieg' whereby humans, shortly upon arrival, hunted these large beasts to extinction, climatic effects of increasing aridity, ecosystem modification, or a combination of climate and human hunting (Figures 11 and 12).



Figure 11 The skull of a *Diprotodon optatum*. Note the large upper and lower incisors as well as the bony structure on top of the maxilla thought to be for the attachment of powerful muscles. These large herbivores weighed more than 1000 kg. They were probably browsers preferring the open, arid lands, using their sharp incisors to crop hard-stemmed vegetation (photograph by R. Cosgrove).



Figure 12 The postcranial skeleton of *Diprotodon* suggests a very heavily built animal with long proximal and short distal limbs. The feet were relatively small and probably had heavy padding (photograph by R. Cosgrove).

Evidence for the swift demise of megafauna due to over-hunting has come from systematic dating of sediments using luminescence techniques. Age estimates on some 12 taxa suggest extinction between 40 000 and 50 000 ya. The dates come from sediments associated with articulated megafauna bones, not from archaeological sites. The results have been questioned by a number of researchers who are critical of the selection of naturally deposited semi-articulated skeletons. They argue that to truly understand the mechanism and age of the extinctions, human sites associated with megafauna must be studied (Wroe et al., 2004).

Other geochronologists have dated eggshell from the extinct *Genyornis* and extant emu (*Dromaius*) and found isotope variation in the two species from about 140 000 ya (Miller et al., 2005). Their work included the analysis of thousands of eggshell pieces of each species from central Australia. They argue that *Genyornis* became extinct about 45 000 years BP under a regime of altered ecological conditions as shown in the changed isotope values beginning about 50 000 years BP. They dismiss climate as a cause and propose human alternation to natural vegetation patterns by fire as an explanation. Drought adapted vegetation, including fertile grasses, gave way to fire adapted desert scrub. Because isotope values in the eggshell demonstrate *Genyornis* was apparently restricted to a narrow range of grasses, it is suggested that it could not adapt to the fire-prone desert vegetation and died out. Other workers reject the stated impact of Aboriginal firing on the Pleistocene landscape, arguing that it has been overemphasized (Bowman, 1998). They see the relationship between past Aboriginal firing practices and the resultant landscape as a highly complex ecological interaction. To understand it requires an apposite temporal resolution of the ecological and archaeological record.

Other data contradict the view that there was widespread extinction of *Genyornis* by 50 000 years BP (Trueman et al., 2005). Evidence from the archaeological site of Cuddie Springs suggests that *Genyornis* survived for another 15 000 years in co-existence with humans. In addition, the site contains thousands of megafaunal bones in layers below the lowest stone tools showing a diminishing species diversity in the absence of humans through time. The implication is that megafauna may have already been in decline due to sustained long-term landscape changes driven by climatic factors (Figures 13 and 14).

The most common species at Cuddie Springs are *Diprotodon optatum* and *Genyornis newtonii* and these are clearly associated with artifacts sandwiched between two hardened pavements dated to between 30 000 and 35 000 years BP (Field, 2004). The most parsimonious explanation for the data is that these species retreated to the better-watered edges of the expanding arid zone, leaving the less specialized emus to survive in the drier interior where they live at present day. *Genyornis*'s prolonged co-existence with humans at Cuddie Springs can be explained this way without recourse to the overdramatization of the human-caused extinction theory. Nevertheless they appear to have finally succumbed to a drying continent and human predation (Figures 15 and 16).

Ultimately the question of megafauna extinction and human influence is undeniably linked to a determination of the arrival time of people on the Sahul continent. Until this is resolved the cause of extinctions remains problematic. In Tasmania there is a reasonably good understanding of the

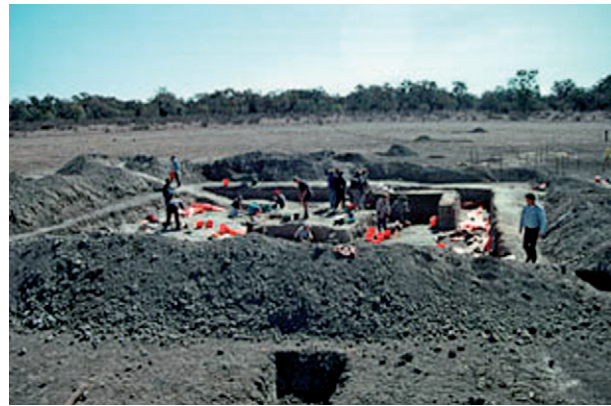


Figure 13 Cuddie Springs is situated in the centre of a shallow depression. Many thousands of megafauna bones have been excavated in association with artifacts (photograph by R. Cosgrove).



Figure 14 Excavation at Cuddies Springs revealed the long bones of *Genyornis* lying next to stone implements (arrowed). The lowest human occupation is dated to 35 000 years BP (photograph by R. Cosgrove).

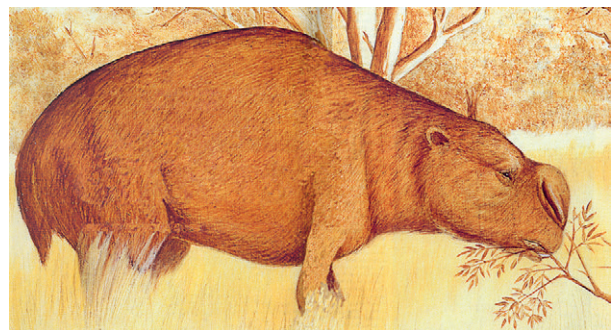


Figure 15 An artistic impression of *Diprotodon optatum*. It probably stood about 1.8 m at the shoulder and had an overall length of 3.4 m (illustration A. Musser and Australian Museum).

human's entry time of about 35 000 years BP. Although megafauna were present in Tasmania, none have been recorded in over 600 000 bones excavated from seven late Pleistocene human cave sites (Cosgrove and Allen, 2001). All are extant animal species. It strongly suggests that megafauna were already locally extinct before humans arrived (Figures 17 and 18).



Figure 16 *Genyornis newtonii* is a distant relative of the emu but was larger. It stood over 2 m and had large robust feet (illustration A. Musser and Australian Museum).



Figure 17 Bone Cave in southwest Tasmania has human occupation extending back at least 29 000 years. It contains two main occupation periods; from 29 000 years BP to about 23 000 years BP then a hiatus until 16 000 years BP when occupation intensified (photograph by R. Cosgrove).

Continent-wide explanations are therefore unlikely to address this issue in the short term. The expectation is that these will be much more complex and focused at local or regional scales. Additional work needs to be done in identifying causal mechanisms and the effects. Ecological modeling and the identification of other archaeological sites to establish regional variability may go some way in solving this vexatious question.



Figure 18 Southwest Tasmanian sites are characterized by there extreme richness of stone tools and bones. It is not uncommon to find 250 000 bones and 35 000 stone tools in less than a cubic meter of deposit (photograph by R. Cosgrove).

Technology and Economy

late Pleistocene Sahul stone technology has been characterized as conservative and for the most part unchanging. In the past, these notions have been used to portray the early colonists as technologically backward, without regional variability. Recent advances in fieldwork, method, and theory have largely debunked this view, recognizing much more behavioral variability across Sahul. It is clear that the first settlers must have had reliable sea-going rafts or canoes. A level of organizational capacity and language must have also been present to carry out these voyages. Obviously the organic material necessary for the construction of these boats has long since decayed although sea faring continued throughout the Bismarck Archipelago during the late Pleistocene as demonstrated by the presence of traceable stone artifact raw material. Significant quantities of obsidian have been recorded in the late Pleistocene sites on New Britain and New Ireland for example. Recent research at Kupona na Dari indicates that it was used soon after human arrival about 40 000 years BP (Torrence et al., 2004). Obsidian was transported by boat up to 32 km and in the case of sites on New Ireland at Matenbek, up to 300 km from sources at Mopir and Talasea by 20 000 ya (Figure 19).

Coupled with this movement is evidence for the economic structuring by human colonizers. It has been argued that the early settlers to the Bismarks were highly mobile, utilizing resource patches, cleared with 'waisted' axes and fire as a means of breaking the tropical canopy. Before 20 000 years BP water crossings were made into previously uninhabited areas where resource levels and interior geography were unknown. Significant marine distances were crossed where the next landfall could not be seen before leaving their origin point, as was the case of Kilu, Solomon Islands 29 000 years BP. Exploration included a degree of deliberate voyaging to locate stone quarries that meant groups were mobile, exploiting a wide range of resources from both the coast and inland. After 20 000 years BP there were transfers of animals as a risk-reducing device, raw materials, and increased sailing to more remote islands.

The earliest and most typologically distinct tools are ground-edge axes that have been found across northern Australia and New Guinea. They only appear in the archaeological record in southern Australia after 5000 years BP. 'Waisted' and ground-edge hatchets are dated to over 40 000 years on the Huon Peninsula and 26 000 years at the New Guinea highlands site of Kosipe. Ground-edge hatchets were recovered from the sites of Malangangerr and Nawamoyin in the Northern Territory that date to 23 000 years BP. Archaeologists have suggested that the typological links indicate a degree of cultural connectedness and perhaps a common origin of human groups operating at this time. They were probably used in forest clearance to create small, disturbed forest patches to encourage useful and productive plant foods (Figures 20 and 21).

Although typological schemes have been put forward in the past to describe the flaked stone technology of Sahul, several objections have been raised that counter the usefulness of typological groupings (Hiscock and Attenbrow, 2003). Indeed with

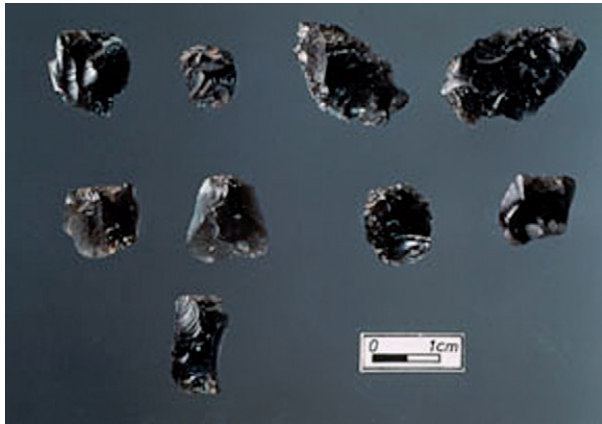


Figure 19 Small obsidian artifacts found in the sites on New Ireland. These implements have been worked down to a small size, indicative of raw-material rationing (photograph by R. Frank).



Figure 20 This 23 000-year-old ground-edge hatchet was found at the northern Australian site of Malangangerr. They are made from porphyritic dolerite. It is one of the oldest such implements in the world (photograph by J. Allen).

the recent discovery of microliths dated to 15 000 years BP at GRE 8 and Walkunder Arch in Queensland aged 13 000 years BP, their usefulness as chronological markers of mid-Holocene technological change is no longer viable (Figure 22).

The appearance of grindstones perhaps 30 000 years BP also suggests that early tools for seed-grinding exploitation were instrumental in opening up opportunities for the settlement of challenging environments such as deserts. The use of organic technology has also been recorded at Wylie Swamp where barbed spears and boomerangs dated to at least 10 000 years



Figure 21 Two main phases of occupation were found at Malangangerr; an early Pleistocene and a later Estuarine phase containing shell midden and smaller-sized artifacts (photograph by J. Allen).



Figure 22 Walkunder Arch is a large limestone cave that contains a series of fine soil lenses dating to the Holocene period. In the darker reddish layers small backed microliths were recovered indicating that these implements have a late Pleistocene origin (photograph by R. Cosgrove).

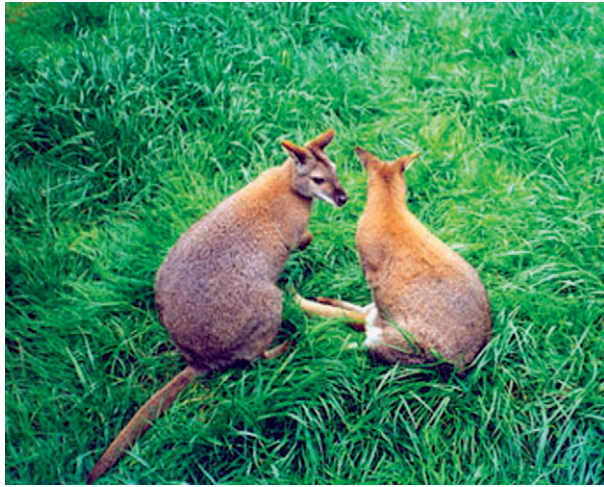


Figure 23 Bennett's wallaby is the most common prey animal in the Tasmanian Ice Age sites. Studies of the annuli in their teeth show that the majority were hunted during the autumn and late spring, with many killed during winter. This was the harshest time of the year but people were able to use the valleys in a planned and systematic manner that indicates organization skill and forward planning (photograph by R. Cosgrove).

BP have been preserved. In Tasmania and southwest Western Australia bone points dating back 30 000 years BP have been identified. These were probably used as awls, needles, and spear points (Mulvaney and Kamminga, 1999). Given these developments the classification of the late Pleistocene stone technology under the rubric, Core Tool and Scraper Tradition is now considered redundant.

In Tasmania, humans were exploiting wallabies as the major prey animals. Hunters used sites in the valleys primarily between autumn and early spring, capturing the macropods for their fat, fur, and sinews. Body-part analysis of the butchered carcasses suggests selection of the lower limb bones processed for marrow. The pattern of exploitation suggests optimal foraging of Bennett's wallaby. They were an excellent prey choice as they took little effort to obtain, energy returns were higher for the effort spent in their search, and killing and processing took place in the patches of grassland on which they fed (Cosgrove and Pike-Tay, 2004; Figure 23).

See also: **Archaeological Records:** 1.9 Myr–300 000 Years Ago in Europe; 2.7 Myr–300 000 Years Ago in Asia; Global Expansion 300 000–8000 Years Ago, Africa; Global Expansion 300 000–8000 Years Ago, Americas; Global Expansion 300 000–8000 Years Ago, Asia; Neanderthal Demise; Overview; Postglacial Adaptations. **Vertebrate Records:** Mid-Pleistocene of Africa. **Vertebrate Studies:** Interactions with Hominids.

References

- Allen J, Gosden C, and White JP (1989) Human Pleistocene adaptations in the tropical island Pacific: Recent evidence from New Island, a Greater Australian outlier. *Antiquity* 63: 548–561.
- Bowler J (1998) Willandra Lakes revisited: Environmental framework for human occupation. *Archaeology in Oceania* 33.
- Bowler J and Magee JW (2000) Redating Australia's oldest human remains: A sceptic's view. *Journal of Human Evolution* 38.
- Bowman DMJS (1998) The impact of Aboriginal landscape burning on the Australian biota. *New Phytologist* 140: 385–410.
- Brown P (2005) Peter Brown's Australian and Asian palaeoanthropology. <http://www-personal.une.edu.au/~pbrown3/palaeo.html> (lasted accessed 3 June 2006).
- Cosgrove R (1999) Forty-two degrees south: The archaeology of late Pleistocene Tasmania. *Journal of World Prehistory* 13: 357–402.
- Cosgrove R (2005) Coping with noxious nuts. *Nature Australia* 28: 46–53.
- Cosgrove R and Allen J (2001) Prey choice and Hunting Strategies in the Late Pleistocene: Evidence from Southwest Tasmania. In: Anderson A, O'Connor S, and Lilley I (eds.) *Histories of Old Ages: Essays in Honour of Rhys Jones*. Canberra: Coombs Academic Publishing, Australian National University.
- Cosgrove R and Pike-Tay A (2004) The Middle Palaeolithic and Late Pleistocene Tasmania hunting behaviour: A reconsideration of the attributes of modern human behaviour. *International Journal of Osteoarchaeology* 14: 321–332.
- Field J (2004) Australian Late Pleistocene faunal extinctions and the archaeological record: A view from Cuddie Springs. In: Murray T (ed.) *Archaeology from Australia*. Melbourne: Australian Scholarly Publishing.
- Field JH, Dodson JR, and Prosser IP (2003) A Late Pleistocene vegetation history from the Australian semi-arid zone. *Quaternary Science Reviews* 21: 1023–1037.
- Gillespie R and Roberts R (2000) On the reliability of age estimates for human remains at Lake Mungo. *Journal of Human Evolution* 38: 727–732.
- Hesse PP and McTainsh GH (2003) Australian dust deposits: Modern processes and the Quaternary record. *Quaternary Science Reviews* 22: 2007–2035.
- Hiscock P and Attenbrow V (2003) Early Australian implement variation: A reduction model. *Journal of Archaeological Science* 30: 239–249.
- Kershaw P (1994) Pleistocene vegetation of the humid tropics of northeastern Queensland, Australia. *Palaeogeography, Palaeoclimatology, Palaeoecology* 109: 399–412.
- Kershaw P, van der Kaars S, and Moss P (2003) Late Quaternary Milankovich-scale climatic change and variability and its impact on monsoonal Australasia. *Marine Geology* 201: 81–95.
- Lambeck K and Chappell J (2001) Sea level change through the last glacial cycle. *Science* 292: 672–686.
- Lambeck K, Esat TM, and Potter EK (2002) Links between climate and sea levels for the past three million years. *Nature* 419: 199–206.
- Leavesley MG, Bird MI, Fifield LK, et al. (2002) Buang Merabak: Early evidence for human occupation in the Bismark Archipelago, Papua New Guinea. *Australian Archaeology* 54: 55–57.
- Miller GH, Fogel ML, Magee JW, Gagan MK, Clarke SJ, and Johnson BJ (2005) Ecosystem Collapse in Pleistocene Australia and a Human Role in Megafaunal Extinction. *Science* 309: 287–290.
- Morwood MJ, Soejono RP, Roberts RG, et al. (2004) Archaeology and age of a new hominin from Flores in eastern Indonesia. *Nature* 431: 1087–1091.
- Mulvaney J and Kamminga J (1999) *Prehistory of Australia*. Sydney: Allen and Unwin.
- O'Connell JF and Allen J (1998) When did humans first arrive in Greater Australia and why is it important to know? *Evolutionary Anthropology* 6: 132–146.
- O'Connell JF and Allen J (2004) Dating the colonization of Sahul (Pleistocene Australia–New Guinea): A review of recent research. *Journal of Archaeological Science* 31: 835–853.
- Pack SM, Miller GH, Fogel ML, and Spooner NA (2003) Carbon isotopic evidence for increased aridity in northwestern Australia through the Quaternary. *Quaternary Science Reviews* 22: 629–643.
- Pardoe C (1995) Riverine, biological and cultural evolution in southeastern Australia. *Antiquity* 69: 696–713.
- Pavides C (2004) From Misisil Cave to Eliva Hamlet: Rediscovering the Pleistocene in interior west New Britain. *Records of the Australian Museum* 29: 97–108.
- Roberts R, Flannery T, Ayliffe LK, et al. (2001) New ages for the last Australian megafauna: Continent-wide extinction about 46,000 years ago. *Science* 292: 1888–1892.
- Shulmeister J, Goodwin I, Renwick J, et al. (2004) The Southern Hemisphere westerlies in the Australasian sector over the last glacial cycle: A synthesis. *Quaternary International* 118–119: 23–53.
- Stringer C (2002) Modern human origins: Progress and prospects. *Philosophical Transactions of the Royal Society London, Series B* 357: 563–579.
- Thorne A and Curnoe D (2000) Sex and significance of Lake Mungo 3: Reply to Brown "Australian Pleistocene variation and the sex of Lake Mungo 3. *Journal of Human Evolution* 39: 587–600.

- Torrence R, Neall V, Dolman T, et al. (2004) Pleistocene colonisation of the Bismark Archipelago: New evidence from West New Britain. *Archaeology in Oceania* 39: 101–130.
- Trueman C, Field JH, Dortch J, Charles B, and Wroe S (2005) Prolonged coexistence of humans and megafauna in Pleistocene Australia. *Proceedings of the National Academy of Sciences* 102: 8381–8385.
- Turney CSM, Kershaw AP, Moss P, et al. (2001) Redating the onset of burning at Lynch's Crater (North Queensland): Implications for human settlement in Australia. *Journal of Quaternary Science* 16: 767–771.
- Wroe S, Field J, Fullagar R, and Jermin LS (2004) Megafaunal extinction in the Late Quaternary and the global overkill hypothesis. *Alcheringa* 28: 291–332.

Global Expansion 300 000–8000 Years Ago, Americas

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Introduction

The peopling of the Americas during the Late Quaternary was one of the most significant events in the history of our species. It initially required adaptation to the far north (Hoffecker, 2005), then further adaptation to a series of new environments – littoral zones, inland deserts, woodlands and grasslands, high mountains, tropical forests, even Sub-Antarctic islands in the Horn of South America (Dillehay, 2000). The earliest Americans made first contact with a variety of plant and animal species, and these surely had a hand at shaping their technology, subsistence, settlement, and ideology. In North America, the first Americans dispersing from the north found familiar large-bodied prey species, for example, mammoth, bison, horse, and a variety of cervids, but as they dispersed further into Mesoamerica and South America, they encountered alien plants and animals, requiring significant foraging and technological adjustments. Dispersal may have begun as early as 30 000 calendar years ago (ka), but more likely the process unfolded during the Late Glacial, closer to 16–13 ka. By 12 ka, modern humans occupied virtually all of ice-free America.

Study of the origins and dispersal of the first Americans is an interdisciplinary endeavor. Archaeologists, biological anthropologists, geneticists, geoscientists, and paleoecologists work together to disclose who the first Americans were, where and when dispersal began, the routes taken and environments encountered, and processes of exploration and settling of new lands. Human colonization of North and South America certainly had a significant impact on plant and animal communities, but equally important were effects of climate change on dispersing Americans. Dispersal occurred amidst significant global climate change – the transition from the last glacial to the present interglacial period.

Old World Origins

From where did the first Americans come? There are two possibilities – from northeast Asia via the Bering Land Bridge or southwest Europe via the North Atlantic Ocean (Figure 1). Current genetic evidence overwhelmingly supports the Beringian model. All contemporary Native American lineages, or haplotypes, of mitochondrial (mt)DNA and the non-recombining portion of the Y chromosome share common ancestors with southern Siberians, not Europeans (Derenko et al., 2001; Karafet et al., 2006; Merriwether, 2006; Perego et al., 2010; Schurr and Sherry, 2004; Starikovskaya et al., 2005; Zegura et al., 2004). Further, segments of ancient DNA amplified from prehistoric Native American remains so far have been attributed only to haplogroups or haplotypes of Asian origin (Gilbert et al., 2008; Kaestle and Smith, 2001; Kemp et al., 2007; Malhi et al., 2007; Perez et al., 2009; Raff et al.,

2011; Smith et al., 2005). Some of these remains are quite ancient: the human coprolites from Paisley Caves (Oregon) date to ~14.9–13.9 ka, while human bone from On Your Knees Cave (southeast Alaska), Wizards Beach (Nevada), and Arroyo Seco 2 (Argentina) date to 10.5–10.2, 10.6–10.2, and 8.8–7 ka, respectively (Figure 1).

Corroborating evidence supporting an Asian origin of the first Americans comes from recent comparative analyses of the First Americans themselves. Skulls from Spirit Cave (Nevada), Kennewick (Washington), Lagoa Santa (Brazil), and other 'Paleo-American' sites bear strong similarities with modern Ainu and prehistoric Jomon populations of Japan as well as southeast Asian populations in general (Brace et al., 2001; Chatters, 2000; González-José et al., 2005; Lahr, 1995; Neves and Hubbe, 2005; Neves et al., 2004, 2007b; Powell and Neves, 1999; Steele and Powell, 2002). All these groups are likely related through some ancient source population that existed across Asia during the late Pleistocene, one that will become better characterized as the sample of Asian Paleolithic specimens increases. Archaeologically, greater northeast Asia (including eastern Russia, Mongolia, north China, Korea, and Japan) is known to have a rich record of Upper Paleolithic sites and complexes; continued research there will ultimately lead to a precise record of Native American origins.

Why, then, do some scientists continue to argue that the first Americans came from Europe? The reasons for this are based on archaeological observations made by proponents of the Iberian model (Bradley and Stanford, 2004). Based on particular similarities in the lithic industries of late Pleistocene European and North American complexes, they argue that prior to 19 ka Upper Paleolithic Iberians of the Solutrean culture carried bifacial and blade technologies in boats across the North Atlantic Ocean to eastern North America. These populations persisted there in sufficient numbers to create a minimal pre-Clovis record, and around 13 ka, they developed bifacial fluting technology that spread geographically across North America by either migration or diffusion among existing populations. Countering the Solutrean origins model are the arguments of Straus et al. (2005) that (1) technological similarities between Solutrean and Clovis are more likely due to evolutionary convergence, not phylogenetic relationship; (2) Solutrean foragers rarely exploited maritime resources, and there is no evidence for Solutrean seafaring technology; (3) the evidence for pre-Clovis in the eastern United States is unconfirmed; and (4) the temporal gap between the youngest Solutrean and oldest Clovis sites is too wide to provide a credible link. Given the current genetic record, the most parsimonious explanation for the origins of the first Americans is the traditional Beringian model.

Determining when humans first dispersed into Beringia and subsequently the Americas is less clear. Two benchmark events, however, anchor the beginning and ending points of this process, one in Asia and the other in North America. First,

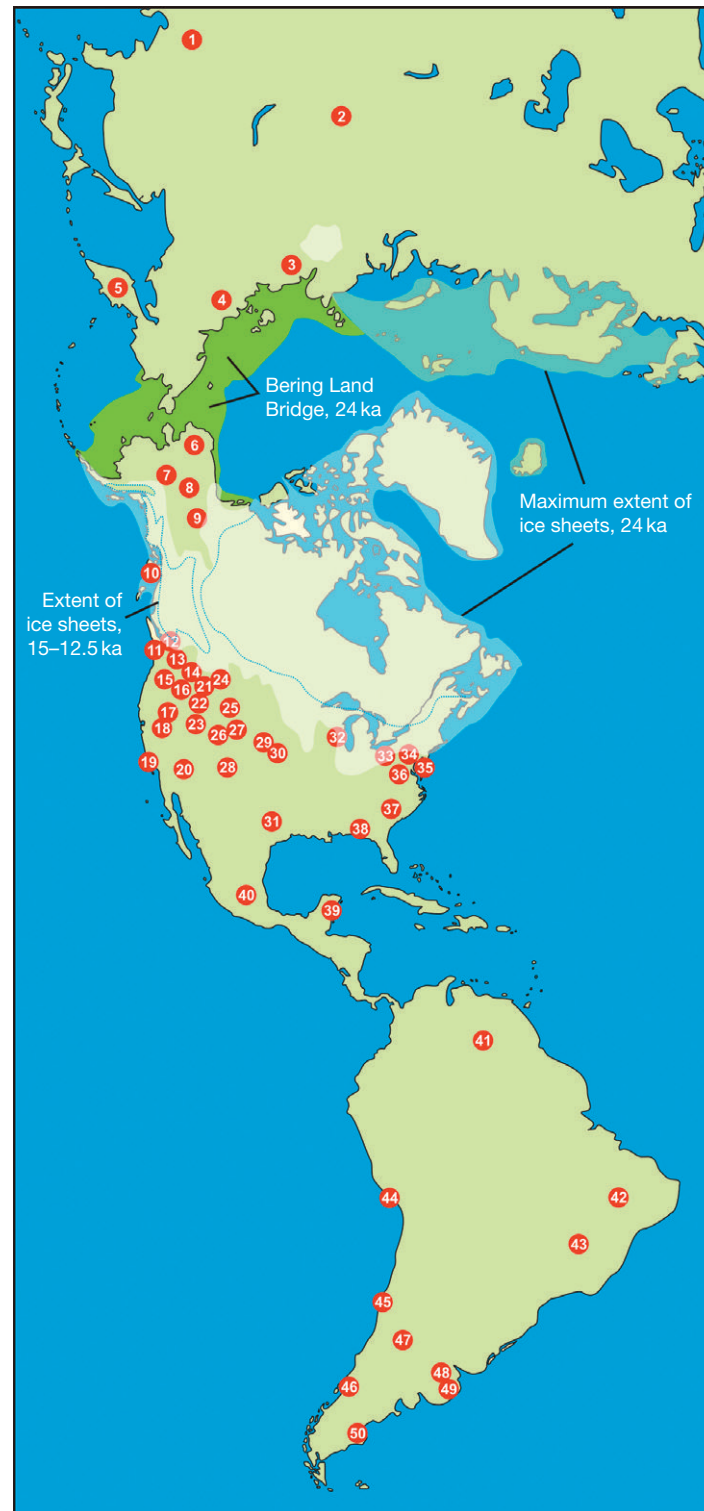


Figure 1 Early archaeological and human paleontological sites mentioned in text (1, Tianyuan; 2, Pokrovka; 3, Yana; 4, Berelekh; 5, Ushki; 6, Tuluaq; 7, Nenana complex sites (Walker Road, Moose Creek, Dry Creek, Owl Ridge); 8, Tanana valley sites (Upward Sun, Swan Point, Broken Mammoth); 9, Bluefish Caves; 10, On Your Knees Cave; 11, Manis; 12, Ayer Pond; 13, East Wenatchee; 14, Kennewick; 15, Paulina Lake; 16, Paisley; 17, Wizards Beach; 18, Spirit Cave; 19, Channel Island sites (Arlington Springs, Cardwell Bluffs, CA-SRI-512); 20, San Pedro valley sites (Lehner, Murray Springs); 21, Simon; 22, Buhl; 23, Bonneville Estates Rockshelter; 24, Anzick; 25, Colby; 26, Mountaineer; 27, Barger Gulch; 28, Arch Lake; 29, La Sena; 30, Lovewell; 31, Buttermilk Creek sites (Friedkin, Gault); 32, Kenosha County sites (Schaefer, Hebior, Mud Lake); 33, Meadowcroft; 34, Shawnee-Minisink; 35, Miles Point; 36, Cactus Hill; 37, Topper; 38, Page-Ladson; 39, Quintana Roo enclaves; 40, Peñon; 41, Pedra Pintada; 42, Pedra Furada; 43, Lagoa Santa; 44, Quebrada Jaguay; 45, Quebrada Santa Julia; 46, Monte Verde; 47, Agua de la Cueva; 48, Arroyo Seco 2; 49, Pasa Otero 5; 50, Piedra Museo).

human paleontological remains from northeast Asia, including the nearly complete modern human skeleton from Tianyuan Cave, China (directly dated to 42–39 ka) (Shang and Trinkaus, 2010) and the front of a modern human calotte from Pokrovka, Siberia (directly dated to 32.5–31.5 ka) (Akimova et al., 2010) together indicate that modern *Homo sapiens* ultimately dispersing from Africa had arrived in the region 40–30 ka. This is corroborated by genetic coalescence estimates for the mtDNA haplotypes of contemporary northeast Asians (Comas et al., 2004; Forster, 2004), as well as radiocarbon estimates for the emergence of Upper Paleolithic technologies in southern Siberia (Goebel, 1999). The second benchmark even relates to the appearance of Clovis, the earliest unequivocal complex of archaeological sites known in the Americas. Bifacial projectile points with distinctive flutes, or grooves, at their bases, blade tools detached from prismatic blade cores, and carefully carved ivory and bone tools characterize Clovis and have been documented across most of temperate North America 13.3–12.8 ka (Walters and Stafford, 2007). These two events – appearance of the first modern humans in northeast Asia and emergence of the Clovis complex in North America – provide lower- and upper-limiting dates for dispersal to the Americas: 40 and 13.3 ka.

Paleoenvironments and Routes of Dispersal

The ‘empty’ environments of late Pleistocene America were highly variable geographically and temporally and created numerous opportunities and challenges for human settlers.

The first challenge was entering the hemisphere. The human genetic evidence discussed above suggests the entry point was the Bering Land Bridge, which connected northeast Siberia and Alaska during glacial periods of the Pleistocene. During the late Pleistocene between about 30 and 20 ka, the land bridge reached its greatest extent in recent history, spanning nearly 1000 km from north to south (Figure 1; Brigham-Grette et al., 2004). Late-glacial melting of the world’s glaciers and ice sheets caused rising sea levels and gradual flooding of the land bridge, in turn leading to a shrinking terrestrial landscape and an ever-changing sea margin (Elias et al., 1996, 1997; Manley, 2002). Current estimates on final flooding of the land bridge suggest this event occurred around 13–12.5 ka (Brigham-Grette et al., 2004; Elias et al., 1996). The land bridge itself remained relatively ice-free through most of the last glacial (Glushkova, 2001; Kaufman and Manley, 2004), and full-glacial vegetation communities in Beringia are generally thought to have been open tundra-steppes reflecting cold, dry climate. Local conditions, however, were variable, with steppe-tundra occurring in inland areas of eastern Beringia (interior Alaska and northern Yukon) and more mesic tundra dominating lowlands of the land bridge itself (the now submerged shelves of the Bering and Chukchi seas) (Bigelow et al., 2003; Elias and Crocker, 2008; Kuzmina et al., 2008). Large mammal fauna characterized the land bridge – mammoth, bison, horse, reindeer, and musk ox during full-glacial times (e.g., Lorenzen et al., 2011), with additions of wapiti and moose during the Late Glacial (Guthrie, 2003). The land bridge connecting Asia and America, therefore, offered opportunities for settlement throughout the late Pleistocene.

Canada during the late Pleistocene was covered by widespread glacial ice sheets. These would have formed a

significant barrier for human dispersal during most of the last glacial cycle; however, during deglaciation, ‘ice-free corridors’ opened first, leading to plant and animal dispersals from the north and south into recently deglaciated landscapes. Two corridors emerged, one along the Pacific Ocean coast and the other along the eastern flank of the Canadian Rocky Mountains (Figure 1; Clague et al., 2004; Dyke, 2004). The coastal corridor may have opened first, by 16–15 ka (Mandryk et al., 2001); however, luminescence dates on wind-blown sands in central Alberta indicate an ice-free interior corridor by about the same time (Munyikwa et al., 2011), nearly 2000 years earlier than previously thought. Confirmation on the timing of the opening of the corridors eventually may come from direct dating and DNA analysis of dispersing bison and other megafauna (e.g., Shapiro et al., 2004; Zazula et al., 2009). The current data suggest that although Beringia was cut off from the rest of the Americas during the Last Glacial Maximum (LGM), by 15 ka two south-bound avenues of dispersal may have been available for humans.

In temperate North America during the height of the last glacial, the Cordilleran ice sheet expanded southward into northern Montana and Washington, and the Laurentide ice sheet spread deep into the American Midwest and Mid-Atlantic states (Dyke, 2004). Ice-sheet recession during the Late Glacial was gradual and at times stalled or even reversed. The northern United States, though, became completely ice-free by about 13 ka, while in northeastern Canada, deglaciation was not complete until well into the Holocene. Mountain glaciers south of Canada expanded and receded during the late Pleistocene as well, reaching their greatest extent around the LGM (Gillespie and Zehfuss, 2004; Pierce, 2003). Late-glacial humans entering temperate western North America encountered relatively cool and wet climates and richer environments than exist there now. Large pluvial lakes filled many valleys of today’s Great Basin, Sonoran, and Chihuahuan deserts (Ballenger et al., 2011; Goebel et al., 2011; Metcalfe et al., 2002), and open woodlands would have been common there as well as in the Great Plains beyond the Rocky Mountains. East of the Mississippi River, coniferous forests dominated landscapes as far south as Tennessee and the Carolinas, and mixed hardwood forests existed in the deep south (Delcourt and Delcourt, 1987; Williams et al., 2004). These conditions gradually changed during the early postglacial period, leading to significant drying in the west and northward displacement of boreal communities in the east. Extinctions of mammoth, mastodon, horse, and other large mammals occurred during this time, some before the Clovis era and others synchronous with it (Grayson and Meltzer, 2002).

In Central and South America, first colonizers encountered environmental conditions that were generally cool and dry, compared to today (Markgraf, 1989; Munyikwa, 2005; Tripaldi et al., 2011). In Central America, forests still likely covered lowland areas (Ranere, 2006). The Panamanian isthmus was not significantly broader during the Late Glacial than it is today; it certainly would have been a ‘bottleneck’ for dispersing human populations. In South America, environmental records in the Andes Mountains chronicle a glacial advance coincident with the LGM in the northern hemisphere, ca. 20–18 ka, but some records also point to a secondary advance during the Late Glacial, 17–14 ka, with deglaciation occurring by 12.5 ka

(Coronato et al., 2004; Harrison, 2004; Hastenrath, 2009; Helmens, 2004; Mark et al., 2005). From the central Mexican highlands to the southern Andes, snow line during the LGM appears to have dropped as much as 1400 m from today, implying maximum summer temperatures as much as 8 °C cooler than now (Harrison, 2004; Helmens, 2004; Kühle, 2004; Mark et al., 2005). Vegetation communities in Mexico were generally characterized by extensive grasslands and open mixed forests (Lozano-García et al., 2005), while in South America, they varied temporally, with open grasslands or dry shrublands during the height of the last glacial, followed by the development of forests during the Late Glacial, as early as 17 ka locally (Burbridge et al., 2004; Hessler et al., 2010; Hodel et al., 2008). In some areas, for example the Peruvian Puna and Argentinian Pampas, open treeless landscapes persisted into the Holocene, despite relatively humid conditions ~14–12 ka (Aldenderfer, 1999; Quattrocchio et al., 2008). Highland Peru, the northern Bolivian Altiplano, and southern Chile, however, appear to have been wetter during the late Pleistocene than today, with Chilean landscapes being characterized by ‘hyperhumid’ forests (Chepstow-Lusty et al., 2005; Hillyer et al., 2009; Kaiser et al., 2008). Southeast Brazil, finally, may have seen moist conditions and a forest/grassland mosaic throughout the last glacial cycle (Pessenda et al., 2009). As in North America, late-glacial environmental change coincided with extinctions of large mammals.

A likely route of dispersal from North to South America was the Pacific coast, a food-rich littoral zone or ‘kelp highway’ (Erlandson, 2007; Erlandson et al., 2008). A variety of food resources including seals, sea otters, sea urchins, abalone, and numerous species of fish could have been regularly collected along this route, and its relatively calm waters could have facilitated rapid boat travel along this corridor, which may have extended across the north Pacific rim from Japan to California. Rising sea level during the early Holocene submerged much of the late Pleistocene coastal strip, however, creating an archaeological record that is difficult to access (Fedje et al., 2004). Nonetheless, early humans are known to have occupied the Channel Islands by Clovis times, implying the use of seafaring technology at this early date (Erlandson et al., 2008). Archaeological survey of submerged coast lines – in the west and the east – will be an important feature in future first Americans research (Bicho et al., 2011).

Chronology of Human Arrivals

As discussed above, dispersal to the Americas had to have occurred between 40 000 and 13 000 cal BP. The precise timing of initial peopling is still debated; some argue that it occurred before the LGM and coalescence of the Canadian ice sheets (e.g., Madsen, 2005), ~28–24 ka (Clague et al., 2004; Dyke et al., 2002), while others contend that it occurred after the LGM upon reopening of the western Canadian ice-free corridors, ~16–14 ka (e.g., Goebel et al., 2008). Further, some archaeologists still hold that dispersal may not have begun until 14–13 ka, with the appearance of Clovis across North America representing the initial founding event (e.g., Haynes, 2005). The difference of opinion is largely based on a less than secure ‘pre-Clovis’ archaeological record.

The Case for Pre-LGM Colonization, >28 ka

To colonize the Americas, modern humans first had to conquer the windswept, treeless tundra-steppe of arctic/subarctic Siberia. They appear to have done so by 32–31 ka, according to the archaeological record of a cluster of localities along the Yana River, located high above the Arctic Circle in extreme northwestern Beringia, 71° N latitude (Pitul’ko and Pavlova, 2010; Pitulko et al., 2004). Occurring in a deeply buried deposit of frozen silt, preservation at the Yana sites is remarkable. Beyond heavily worked stone tools made on cobbles and flakes (and formed predominantly into side scrapers) and a few small unifacial points made on blade-like flakes (Figure 2(g)), the Yana RHS locality’s inventory includes well-preserved osseous tools, including beveled foreshafts made on mammoth ivory and woolly rhinoceros horn. Faunal remains are especially abundant and suggest the hunting of mammoth, horse, rhino, bison, reindeer, and hare. At another Yana locality, more than 1000 mammoth bones representing at least 26 individuals have been recovered; this accumulation is thought to be the product of human activity (Basilyan et al., 2011). Across the present-day Bering Strait in eastern Beringia, no unequivocal evidence for pre-LGM humans has been found, but at Bluefish Caves, Yukon, a mammoth bone ‘core and flake’ thought by a few scientists to represent the work of humans have been directly dated to 29–28 ka (Cinq-Mars and Morlan, 1999; Morlan, 2003). Bluefish has many critics, however, and the worked bones generally have been disregarded (Bever, 2001; Dixon, 2001).

South of the Canadian ice sheets, archaeological evidence of pre-LGM human settlement is just as tenuous. At the Topper site, South Carolina, a possible hearth feature and a stone assemblage interpreted as a smashed core and microflake industry have been found in sands dating to as early as 50 000 ka (Goodyear, 2005a,b), but geological investigations suggest that the feature consists of naturally accumulated uncharred plant remains, and the artifacts are likely the product of thermal fragmentation (either from fires or repeated freezing and thawing) (Waters et al., 2009; see also Goebel et al., 2008). Another important complex of sites is represented by La Sena and Lovewell, located in Nebraska and Kansas, respectively. Both contain mammoth remains that have been dated to ~25–21 ka and show signs of taphonomic alteration, including spirally fractured bone fragments, percussion-flaked bones and bone flakes, possible bone-stacking features, and even a polished bone interpreted to represent a fragment of a beveled bone point or foreshaft (Holen, 2006, 2007). These features suggest to Holen (2006, 2007) that humans quarried and flaked the bones while they were fresh, within a few years of the animals’ deaths; however, Haynes and Krasinski (2010) argue that they could be the result of natural processes (e.g., wolf/bear gnawing, animal trampling, sediment deformation, or ice rafting during spring thaws). Their elephant-bone-breaking experiments with stone hammers and anvils created a pattern of impact scars not seen on the La Sena and Lovewell bones. The localities also lack the stone artifacts commonly found associated with flaked mammoth bones in northern Eurasia (Goebel et al., 2008). Another recently discovered site, Miles Point, Maryland, may contain lithic artifacts in a buried context potentially dating ~33–25 ka

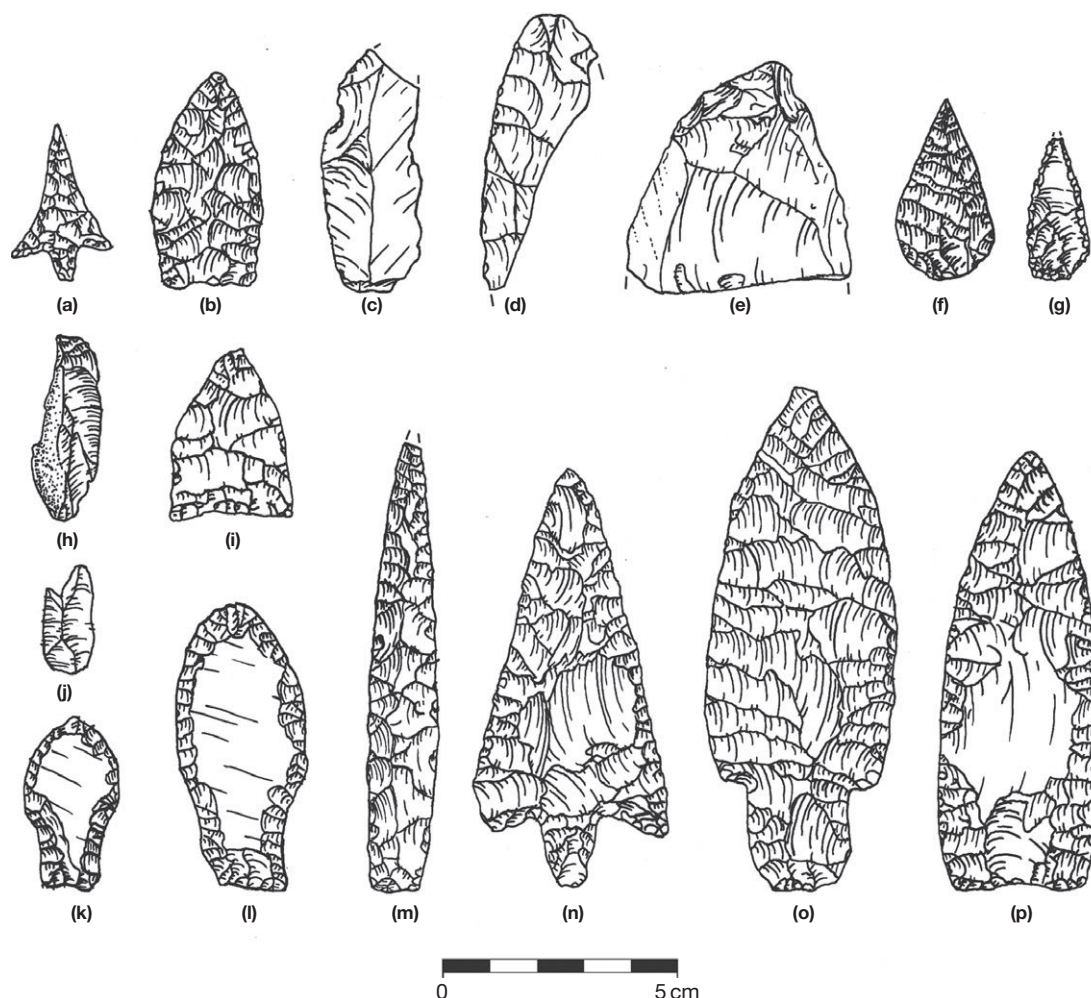


Figure 2 Artifacts from sites mentioned in text representative of some of the earliest archaeological complexes of Beringia and the Americas (a, Channel Island barbed point from CA-SRI-512W site, Santa Rosa Island, California, USA; b, Miller point from Meadowcroft Rockshelter, Pennsylvania, USA; c, retouched blade from Friedkin site, Texas, USA; d–e, biface fragments from Friedkin site; f, Chindadn point from Walker Road, Alaska, USA; g, unifacially worked point from Yana RHS, Russia; h, j, blade-like flakes, Schaefer, Wisconsin, USA; i, basally thinned bifacial point from Cactus Hill, Virginia, USA; k, fishtail point from Cerro El Sombrero, Abrigo 1, Argentina; l, fishtail point from Uruguay; m, El Jobo point from Monte Verde, Chile; n, triangular, stemmed point from Caverna de Pedra Pintada, Brazil; o, Great Basin stemmed point from Buhl, Idaho, USA; p, Clovis fluted point from the Simon cache, Idaho, USA (illustrations after: a, [Erlandson et al., 2011](#); b, [Adovasio et al., 1988](#); c–e, [Waters et al., 2011a](#); f, [Hoffecker et al., 1993](#); g, [Pitul'ko and Pavlova, 2010](#); h, j, [Overstreet, 2005](#); i, [Goodyear, 2005a](#); k–l, [Flegenheimer et al., 2003](#); m, [Collins, 1997](#); n, [Roosevelt et al., 1996](#); o, [Green et al., 1998](#); p, [Kohntopp, 2011](#)) (drawing by Joshua Keene).

([Lowery et al., 2010](#)); however, only test excavations have been conducted there, and the site is being actively eroded by shore-line processes. An extensive excavation and geoarcheological study of the site are needed to confirm the preliminary findings.

In South America, arguments for a pre-LGM occupation have been made based on evidence primarily from one site – Toca do Boqueirão da Pedra Furada – a large, high rockshelter in north-eastern Brazil. Its early contents occur in stratified colluvial deposits and include concentrations of flakes and other stone pieces associated with hearth features, charcoal samples from which span from >50 to 17 ka ([Guidon and Arnaud, 1991](#); [Guidon and Delibrias, 1986](#); [Santos et al., 2003](#)). Fragments of pictographs preserved on the rockshelter's walls also have calcite

concretions covering them that have been luminescence dated in excess of 35 ka ([Watanabe et al., 2003](#)). Criticisms of Pedra Furada relate to the formation of the lithic artifacts and features: detractors have argued that Pedra Furada's lithics were naturally broken from high-energy geological processes (rock fall and moving water), and that the concentrations of charcoal represent redeposited natural charcoal from range fires ([Lynch, 1990](#); [Meltzer et al., 1994](#)). Further, independent radiocarbon dating of the rock art at Pedra Furada and other nearby caves suggests these features are late Holocene in age ([Rowe and Steelman, 2003](#)). Despite these differences in interpretation, the work at Pedra Furada continues, with new reports describing open-air sites with possible artifacts as old as those from the original rockshelter excavation ([Boëda, 2010](#)).

In the past, other sites have been proposed to predate the LGM (e.g., Calico, California; Tule Springs, Nevada; Pendejo Cave, New Mexico; and Tlapacoya, Mexico; Pikimachay Cave, Peru), but these have not held up to scientific scrutiny (e.g., Meltzer, 2009; Waters, 1985). The same is true for recent reports of fossilized human footprints in Mexico (González et al., 2006; Huddart et al., 2008; Renne et al., 2005). As with Topper, La Sena/Lovewell, and Pedra Furada, critiques of these sites have typically been made on taphonomic grounds, repeatedly showing that the burden of evidence is not being carried by investigators of the proposed pre-Clovis sites, as it should be (Haynes, 2007a; Haynes and Krasinski, 2010). Despite the presence of humans in northwestern Beringia by 32–31 ka, unequivocal evidence has yet to be found documenting their further dispersal into Alaska and the Americas.

The Case for an Early Late-Glacial Dispersal, 16–14 ka

Evidence for human dispersal soon after the LGM, 16–14 ka, is strong but still not secure. In Beringia, archaeological sites dating to this time have still not been found, despite genetic models predicting that they should be there (e.g., Tamm et al., 2007). The region's earliest late-glacial site, Swan Point, dates to just before 14 ka (Holmes, 2011), but all other early occupations – Berelekh, Ushki, Tuluq Hill, Broken Mammoth, Mead, Upward Sun, Walker Road, Dry Creek, Moose Creek, and Owl Ridge (Goebel et al., 2010; Graf and Bigelow, 2011; Hamilton and Goebel, 1999; Holmes, 2001; Pearson, 1999; Pitul'ko and Pavlova, 2010; Potter et al., 2011; Rasic et al., 2011) – all seem to date between 14 and 13 ka, perhaps too late to have played a meaningful role in the dispersal of modern humans south of the Canadian ice sheets (Bever, 2006). Nonetheless, the emerging record from these sites indicates that biface-and-blade-producing humans existed in the region a millennium before the Clovis era (Hoffecker and Elias, 2007; Figure 2(f)).

Some of the earliest purported sites in North America are just where they are expected – in the northern United States just south of the late-glacial limits of the Canadian ice sheets. Schaefer and Hebior, two sites in Wisconsin, have yielded a few stone artifacts (Figure 2(h) and 2(j)) in association with disarticulated remains of single mammoths. The bones bear cut-and-pry marks presumably made by stone tools, and these come from a sealed, stratified deposit consistently radiocarbon dated to 14.8–14.2 ka (Joyce, 2006; Overstreet, 2005). At Paisley Caves, Oregon, three coprolites with human DNA have been directly dated to 14.9–13.9 ka (Gilbert et al., 2008; Jenkins, 2007; Figure 3). These appear to have been associated with stone tools and perishable artifacts, but other materials from the same deposit have yielded significantly younger radiocarbon ages (Jenkins, 2007), indicating site formation processes at the site are exceedingly complex and require careful consideration.

Other northern sites are less secure but nonetheless compelling. Meadowcroft Rockshelter, Pennsylvania, yielded a stone-artifact assemblage with a small lanceolate biface (Figure 2(b)) presumably dating 15.2–13.4 ka (Adovasio and Pedler, 2004), but discrepancies between these dates and the character of associated faunal and floral remains (i.e., they suggest much warmer conditions than expected) continue to



Figure 3 View of Paisley Caves, Oregon, USA. Photograph courtesy of the Center for the Study of the First Americans.

hamper its potential significance (Haynes, 2005). Three other possible archaeological sites – Mud Lake (Wisconsin), Manis (Washington), and Ayer Pond (Washington) – have yielded modified bones but no stone tools (Johnson, 2007; Kenady et al., 2011; Overstreet and Kolb, 2003; Waters et al., 2011b). The Mud Lake find, dating to 16.9–16.2 ka, is a mammoth ulna with a dizzying array of marks interpreted to be from stone-tool cutting (Johnson, 2007); however, these marks are far more numerous than expected and anatomically misplaced, according to butchery experiments with modern elephant carcasses in Africa (Haynes and Krasinski, 2010). Manis consisted of a disarticulated mastodon skeleton with spirally fractured bones (Gustafson et al., 1979) directly dated to 13.9–13.7 ka, and into one of its ribs had penetrated another bone object, recently CT scanned and interpreted to represent the tip of a mastodon-bone projectile point (Waters et al., 2011b). The inserted bone, however, appears to have splintered into multiple pieces, so that its original form is no longer evident. The Ayer Pond site consists of a partial skeleton of a bison directly radiocarbon dated to 14–13.7 ka (Kenady et al., 2010). Bone surfaces are well preserved, and spiral fractures, percussion marks, and a few surface abrasions interpreted as cut marks are evident on them. The context of the bison remains – in deep-water pond sediments – is problematic, unless humans killed and butchered the animal carcass upon a frozen lake in the winter (Kenady et al., 2010). Despite the deficiencies in these sites – especially the lack of stone artifacts and equifinality in determining the agents responsible for altering the recovered bones – the emerging record from the northern United States suggests the southern fringe of the Cordilleran/Laurentide ice sheets may have been one of the first areas to be colonized by late-glacial Americans.

In the southern United States, three possible early sites have been found and excavated, but again none are secure. At Cactus Hill, Virginia, a lithic assemblage including small polyhedral cores, blade-like flakes, and basally thinned bifacial points (Figure 2(i)) was recovered from a sand-sheet deposit and dated to 20–18 ka (Goodyear, 2005a; McAvoy and McAvoy, 1997). The early occupation was not stratigraphically sealed, however, and segregation of the pre-Clovis assemblage from

later Clovis and post-Clovis components was based on elevational positions of artifacts. Further, dates as young as 10.6–10.2 ka also came from the pre-Clovis component; these inconsistencies in the radiocarbon chronology suggest postdepositional translocation of materials (Goebel et al., 2008; Haynes, 2005). At Page-Ladson, Florida, nine probable stone artifacts were recovered along with remains of extinct fauna, including a grooved mastodon tusk, from a submerged sinkhole (Webb, 2006). Radiocarbon dates suggest an age of ~14.4 ka for the artifact-bearing deposit; however, how these remains became associated in the sinkhole remains an open question (i.e., Were humans occupying the sinkhole, and if so, why are there so few remains, or did their artifacts become emplaced there secondarily?). The Friedkin site in Texas has produced a large assemblage of lithic artifacts (>15 000 pieces) (Figure 2(c)–2(e)), including small cores, bifaces, bladelet fragments, and a polished stone of hematite, and these are associated with optically stimulated luminescence (OSL) dates ranging 17.5–14 ka, but the excavators of the site conservatively placed the age range at 16–13.2 ka (Waters et al., 2011a). Although deeply buried, the site contains no discernible geological stratigraphy, and the pre-Clovis component, referred to as the Buttermilk Creek complex, was not sealed from younger Clovis and late-Paleoindian components, meaning that cultural-stratigraphic boundaries had to be defined arbitrarily, based on presence of time-sensitive artifacts and OSL dates. Despite this, supporting geoarchaeological and chemical studies suggest a secure pre-Clovis component; only a refitting analysis of artifacts from between levels and components is needed to test the proposed chronology for the site.

Late-glacial sites dating 16–14 ka have also been proposed for South America, but most have not survived peer review and redating (Gruhn, 2005; Steele and Politis, 2009). Two, however, are still considered as evidence of humans in the Southern Cone a millennium before Clovis: Arroyo Seco 2 and Monte Verde. Arroyo Seco 2 is located in the Argentinian Pampas, in a low loess dune. Under an early-mid-Holocene occupation with numerous human burials (Politis et al., 2009), archaeologists uncovered remains of extinct ground sloth (*Megatherium* sp.), toxodon (*Toxodon* sp.), and horses (*Hippidion* sp. and *Equus* sp.) directly AMS ^{14}C dated to 14.2–13.8, 13.7–13.3, and 13.2–12.9 ka, respectively (Steele and Politis, 2009). Some of these remains reportedly show signs of cultural modification and were found in association with quartzite artifacts with unifacial, marginal retouch. A final report on the Arroyo Seco 2 excavations is forthcoming (Politis et al., in press; Figure 4). Excavations at Monte Verde, Chile, occurred from 1977 to 1985, and publication of its monumental site report (Dillehay, 1997) was truly a watershed event in Paleoindian studies (Meltzer et al., 1997). The site's earliest cultural component contained unequivocal artifacts – lithic and nonlithic – from a stratified context sealed by a terminal Pleistocene peat deposit, and it was repeatedly ^{14}C dated to ~14.5 ka. Among the materials recovered *in situ* were stone bifacial points and grooved spheres, remains of nearly 70 species of plants, bones of gomphotheres, some still with adhering soft tissue, wooden artifacts (including lances, digging sticks, and stakes), foundations of small huts, and even a human footprint. Despite the site's preservation, questions about its context have been raised. Critics have suggested



Figure 4 View of Arroyo Seco 2, Argentina. Photograph courtesy of Maria Gutierrez.

that most of the well preserved remains accumulated naturally and the handful of clearly recognizable artifacts represent a later occupation and were intrusive into the 14.5-ka layer (Fiedel, 1999; Haynes, 1999; Lynch, 1990). Dickinson (2011) further points out the inferred age of the Monte Verde occupation suggests it was coeval with local deglaciation and the site thus would have been in a periglacial setting, potentially contradicting its rich paleobotanical record. Nonetheless, Monte Verde still is the strongest candidate for a late-glacial-aged pre-Clovis site in the Americas.

As the above review of sites attests, the evidence for humans in the Americas before the time of Clovis continues to mount. The stone tools associated with proboscideans from Wisconsin, ancient human coprolites from Oregon, extensive Buttermilk Creek complex from Texas, and well-preserved Monte Verde occupation from Chile strongly suggest human presence in both North and South America by 15–14 ka and possibly earlier. Continued research at these plus other important sites like Arroyo Seco 2 may eventually prove an early late-glacial colonization of the Americas, which is seemingly corroborated by current genetic models calculating the timing of human dispersal from Beringia to the rest of the Americas by 15 ka (Perego et al., 2010; Tamm et al., 2007). Evidence from even earlier sites as these, however, may force further revision of this date.

Physical Anthropology

Without doubt, human skeletal evidence across the Americas shows that modern *H. sapiens* colonized the New World. Their remains come from a multitude of archaeological sites dating as early as 13 ka. In North America, among the earliest are Arlington Springs (13.1–12.6 ka), California; Anzick (12.7–12.5 ka), Montana; Buhl (12.4–11.8 ka), Idaho; and Arch Lake (11.8–11.3 ka), Texas (Green et al., 1998; Johnson et al., 1999; Morrow and Fiedel, 2006; Owsley et al., 2010; Stafford et al., 1987). In Mexico, the Peñon remains date to 12.8–12.5 ka, and several nearly complete, well-preserved human skeletons possibly dating as early as 13.7 ka have been recovered from three submerged Yucatan cenotes (underwater

sinkholes) named Naharon, Las Palmas, and Aktun Ha (Gonzalez, 2006; González González et al., 2008). In South America, many remains of early humans have also been recovered, most notably nearly 100 individuals from the Lagoa Santa region of eastern Brazil, most dating around 10 ka, including the Luzia skull (Lapa Vermelha IV) possibly dating as early as 11.4 ka or earlier (Araujo et al., 2008; Feathers et al., 2010; Neves et al., 2007b). These and other human remains spanning the terminal Pleistocene and early Holocene provide an opportunity to examine the origins and adaptations of the earliest Americans.

Recent paleoanthropological studies indicate that the first Americans were physically quite variable, leading some to speculate that they represent two migrations from more than one source (e.g., Neves et al., 1999). Through multivariate comparisons, in the late 1990s, several paleoanthropologists studying the human skeletal remains from Spirit Cave (Nevada) and Kennewick (Washington) found that their crania were 'narrow and long' like modern Europeans, not 'broad and round' like modern Native Americans and northeast Asians (Jantz and Owsley, 1997; Lahr, 1997). They labeled these individuals as 'Caucasoid,' and the media quickly disseminated the concept of a European origin for Kennewick man. Continued craniometric studies reinforced these interpretations of a highly variable 'Paleoamerican' population that was significantly different from later Amerindians, but with an ever-increasing sample it soon became clear that the 'Paleoamericans' were much more similar to modern Ainu and prehistoric Jomon populations of Japan as well as southeast Asian populations in general, suggesting an ancestral-descendant relationship with ancient Asians, not Europeans (Brace et al., 2001; Chatters et al., 2000; González-José et al., 2003, 2005; Lahr, 1995; Neves and Hubbe, 2005; Neves et al., 2004, 2007b; Powell and Neves, 1999; Steele and Powell, 2002; Figure 5). According to

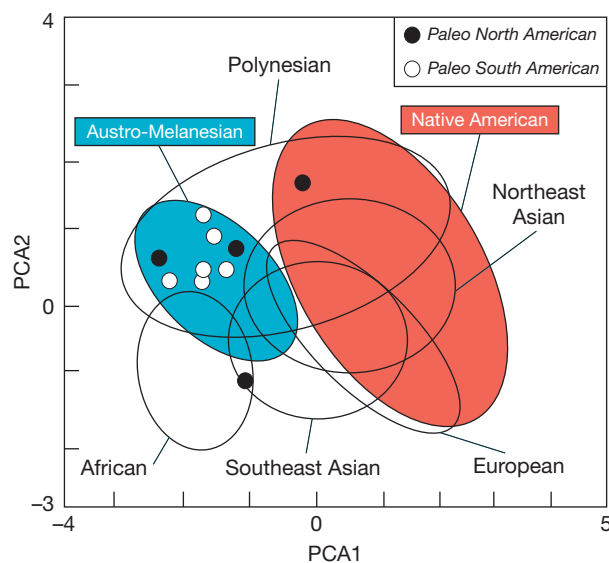


Figure 5 Principal components chart showing variability in craniometrics of early North and South Americans, in context of more recent modern human populations. Reproduced from Powell JF (2005) *The First Americans: Race, Evolution and the Origin of Native Americans*. Cambridge: Cambridge University Press.

some paleoanthropologists, the craniometric differences between Paleoamericans and more recent Amerindians are best explained through a two-migration model, with Paleoamericans being replaced by Amerindians sometime during the terminal Pleistocene or early-mid-Holocene (Hubbe et al., 2010; Neves and Hubbe, 2005; Neves et al., 1999, 2007a). The Kennewick man had the tip of a stone spearpoint imbedded in his right hip (Chatters, 2002), could this be evidence of struggle between the two populations?

Other lines of evidence do not support the replacement theory. Genetic models based on extant human populations as well as empirical evidence from ancient DNA overwhelmingly imply a single dispersal out of Beringia, not two or more (Perego et al., 2010; Raff et al., 2011; Schroeder et al., 2009). According to Hubbe et al. (2010), the genetic studies are misleading, because they have not included mtDNA from ancient Paleoamerican populations and perhaps because the variation in Native American DNA was the product of different microevolutionary events than those that led to the measured differences in cranial shape. However, the few Paleoamericans that have been tested genetically do bear Amerindian mtDNA lineages (Raff et al., 2011), and even in southern South America, a combined study of craniometrics and ancient DNA failed to demonstrate a genetic change across the Paleoamerican–Amerindian transition in cranial morphology (Perez et al., 2009). The archaeological record is also mute regarding a Paleoamerican–Amerindian transition. A rupture in lithic technology, tool forms, or lifeways is expected coincident with this transition. In the intermountain west of the United States, where human skeletons representing both conditions have been dated to 12.4–10.2 ka, there is no evidence for synchronous competing cultures, and no recognized archaeological transition at any time during this period (Goebel et al., 2011). Changes in technology and subsistence did occur before 12.5 ka and after 10 ka (i.e., the Clovis-to-stemmed-point transition and stemmed-to-notched-point transition, respectively), but these did not coincide with changes in skull morphometrics. The same is true for other regions of North America. More likely change between Paleoamericans and Amerindians is the product of drift and selection, not migration and replacement (González-José et al., 2008; Powell, 2005). If multiple migrations did occur, then the mtDNA of the original American population has been eradicated, or the two dispersals occurred simultaneously from the same genetic source in Beringia (Perego et al., 2009).

Impact of Humans on Environments

The role of humans in the extinction of Pleistocene megafaunal species has been debated for decades (Barnosky et al., 2004). Traditionally, polarized scientific positions saw either humans or climate as the driving force behind extinctions (e.g., Grayson, 1989; Grayson and Meltzer, 2002, 2003; Martin, 1967, 1984), but recently other theories, for example, the spread of disease among megafauna (MacPhee and Marx, 1997) and a catastrophic meteor impact (Firestone et al., 2007) have been proposed to account for some large animal disappearances. New studies combining genetic, paleontological, and archaeological data indicate, however, that individual megafaunal species

responded differently to end-of-the-Pleistocene changes in climate, habitat distribution, and human predation (Lorenzen et al., 2011). There is no ‘one-size-fits-all’ model explaining megafaunal extinctions.

In northeast Asia, roles of climate and humans in faunal extinctions varied. While the disappearance of musk ox and woolly rhinoceros was likely driven by climate alone, the extinction of steppe bison and dramatic reduction in the range of horses were affected by both climate and humans (Lorenzen et al., 2011). Late Upper Paleolithic Siberians regularly hunted on horse and bison for thousands of years (Goebel, 2002; Guthrie, 1990), and even in Alaska, remains of these taxa occur in the earliest site, Swan Point (Holmes, 2011). After the LGM, woolly mammoths occur in relatively few Siberian archaeological sites, potentially reflecting a northward-shifting range for this species that could have been caused by breakdown of the tundra-steppe in the south (Lorenzen et al., 2011). In late-glacial Beringia, mammoth remains – primarily tusk fragments – occur at Berelekh, Swan Point, and Broken Mammoth (Holmes, 2011; Mochanov, 1977; Yesner et al., 2011), but this may mean that humans were mining old mammoth ivory and bone for use as tools, not regularly hunting live animals (Yesner, 2007).

In North America, at least 17 of 35 mammal species became extinct ~13 ka, coincident with the time of the Clovis complex (Faith and Surovell, 2009; Haynes, 2007b). The role of humans in many of these extinction events may have been negligible (Grayson and Meltzer, 2002, 2003), and even in Clovis sites like Colby (Wyoming) and Lehner and Murray Springs (Arizona) where extinct fauna (e.g., American camel (*Camelops*) and horse (*Equus*)) have been found, it is difficult to prove that the remains resulted from human hunting (Haynes, 2007b). But the record for repeated mammoth/mastodon hunting in late Pleistocene North America cannot be ignored: “at least 14 kills of proboscideans were made and have been preserved from a period possibly lasting only 300 years, far more in number than on any other continent and far more in such a brief time interval than in all of prehistory” (Haynes, 2007b, p. 89, italics in original). Humans did hunt mammoth and mastodon in late Pleistocene North America, and quite possibly, they did play a significant role in their extinction.

In South America, the timing of most late Pleistocene megafaunal extinctions was coeval with those in the north (Barnosky and Lindsey, 2010; Borrero, 2008). Twelve of 15 species that became extinct did so after the appearance of humans on the continent, and for ten of these extinctions occurred during the first 1000 years of documented human prehistory. So far, remains of extinct fauna including giant ground sloths (*Glossotherium* sp., *Megatherium* sp., and *Mylodon* sp.), horse (*Equus* sp. and *Hippidion saldiasi*), toxodon (*Toxodon* sp.), extinct camelids (*Hemiauchenia* sp. and *Paleolama* sp.), and giant armadillo (*Glyptodon* sp.) have been recovered from Paleoindian sites like Quebrada Santa Julia (Chile) and Pasa Otero 5 (Argentina) (Jackson, 2003; Jackson et al., 2007; Martínez, 2001; Massone, 2003; Miotti and Cattáneo, 2003; Miotti et al., 2003; Paunero, 2003; Vialou, 2003). Many of these remains show signs of having been butchered, but others may represent mining of old bone, like in Beringia (Borrero, 2006). In some areas, for example Brazil, multiple species disappear from the fossil record simultaneously with human

colonization, whereas in other areas, for example, the Southern Cone, some species persist for a thousand years or more (Barnosky and Lindsey, 2010). Given that end-of-the Pleistocene climate changes affected South America’s varied environments in different ways, as in the Northern Hemisphere processes of megafaunal extinctions may have been species-specific and regionally variable, with humans potentially playing variable roles.

Certainly, Paleoindians intentionally transformed other aspects of the New World’s ‘virgin’ environments. Some effects were subtle, for example, the artificial redistribution of lithic materials across landscapes (Borrero, 2011) or the artificial concentration of productive plant species in the Neotropics (Gnecco, 2003a,b; Piperno, 2009), but others may have been dramatic, most notably the intentional setting of range fires to promote productive plant and animal communities (Pinter et al., 2011). Later Native North Americans practiced this resource management strategy (Keeley, 2002), and in the Neotropics, horticulturalists practiced slash-and-burn agriculture as early as 6–5 ka (Arroyo-Kalin, 2012). Vegetation records for the end of the Pleistocene show significant increases in charcoal and fire-driven revegetation, and while most paleobotanists interpret this to have been the result of natural fire-regime changes (Marlon et al., 2009; Power et al., 2008), Pinter et al. (2011) suggest that the increase in fire activity is time-transgressive across the Western Hemisphere and, perhaps, correlative to the appearance of a ‘new and significant ignition source’ – humans.

Technology and Economy

The Schaefer and Hebior sites provide strong evidence that humans existed in North America by 14.8–14.2 ka, synchronous with the early finds from Monte Verde, Chile. If the finds from Paisley Caves, Meadowcroft Rockshelter, Page-Ladson, and Friedkin are confirmed with additional work or clear reporting, together they would suggest that biface-, blade-, and bone-tool-making technologies were established in North America at least 1000 years earlier than Clovis (Waters et al., 2011a). The well preserved remains from Monte Verde indicate that early Americans also used perishable materials like wood and fiber as tools (Dillehay, 1997). Less is known about pre-Clovis subsistence and settlement. Schaefer, Hebior, Monte Verde, and Page-Ladson contained remains of mammoth or mastodon, but whether any of these resulted from human predation currently cannot be proven. At Monte Verde, however, exotic sediment was found embedded in the mastodon bones, suggesting humans had scavenged carcass parts, transporting them back to the site (Borrero, 2008). Monte Verde also contained a rich array of floral remains, including rhizomes, tubers, seeds, fruits, and nuts, presumably consumed by its occupants indicating a broad diet, and preserved features interpreted as dwellings indicate a stationary residence (Dillehay, 1997).

We know much more about early American technology and economy during the Clovis era, 13.3–12.8 ka. Clovis people regularly used stone bifacial technology to prepare weapon tips and blade technology to prepare a variety of unifacial tools like scrapers, knives, and graters (Bradley et al., 2010; Tankersley,

2004). The Clovis fluted point typifies the complex – both its lanceolate, concave-based form, and the particular broad fluting and overpass flaking that characterize its technology (Figure 2(p)). Clovis people also regularly produced osseous tools (made of bone, antler, or ivory), including cylindrical rods beveled at one or both ends and functioning as projectile points or foreshafts. Most Clovis stone tools were manufactured on high-quality cherts and obsidians often procured hundreds of kilometers from where they were last used and discarded (Tankersley, 2004). The long-distance transport of prepared cores and tools, together with the repeated caching of these artifacts at sites like East Wenatchee and Simon (Kilby, 2011), suggest that in some areas of North America, Clovis people were highly mobile (Tankersley, 2004). In Texas and the southeastern United States, however, dense concentrations of Clovis artifacts at sites like Gault and Topper have been interpreted to represent quarry-campsites suggesting significantly lower degrees of mobility (Smallwood, 2010; Waters et al., 2011c). In terms of subsistence, Clovis people appear to have regularly hunted mammoth and mastodon as well as scavenged their remains (Haynes, 2002, 2007a,b); however, they also are known to have taken bison, deer, hares, reptiles, and amphibians (Cannon and Meltzer, 2004). At Shawnee-Minisink, Pennsylvania, carbonized seeds of ten different kinds of plants, including hackberry, blackberry, hawthorn, plum, and grape, as well as calcined fish bones were recovered from hearths associated with Clovis artifacts and dated to 12.9–12.6 ka (Figure 6; Dent, 2007; Gingerich, 2011). Clovis certainly represents a temperate North American phenomenon, but Clovis-like bifaces have been recovered as far south as Mexico, Panama, and possibly even Chile, suggesting the technology and possibly even the people dispersed well south of the modern US border (Gaines et al., 2009; Jackson, 2006; Ranere, 2006).

Other late Pleistocene archaeological complexes are still rich in bifaces, but blade and osseous technologies are less common. In the far west of North America, large stemmed-and-shouldered bifacial points (Figure 2(o)) replaced fluted

points as the primary projectile-point form (Beck and Jones, 2010), whereas in the Great Plains and eastern United States other fluted-point forms (e.g., Folsom, Gainey, Cumberland) appear to have succeeded Clovis (Anderson et al., 2010; Stanford et al., 2005). Long-distance transport of lithic materials and manufacture of formal flake-tool forms continued to characterize these complexes, however (Ellis et al., 2011; Jones et al., 2003). In the desert west, excavations in dry caves and rockshelters like Bonneville Estates Rockshelter and Spirit Cave have revealed rich perishable technologies of bone needles and awls, fiber cordage, and fiber textiles including baskets, bags, and baskets (Connolly and Barker, 2004; Fowler et al., 2000; Goebel et al., 2011; Jenkins, 2007).

By 12.5 ka, human diets across North America had become quite diverse. On Santa Rosa Island off the coast of California, by 12–11.8 ka humans producing small barbed points (Figure 2(a)) had accumulated middens full of remains of waterfowl and sea birds, pinnipeds, and other sea mammals, marine fish, and marine shells (Erlandson et al., 2011). In the Great Basin, humans regularly took artiodactyls like deer, pronghorn antelope, and bighorn sheep but also relied on small mammals (e.g., hares), grouse, waterfowl, and even insects (e.g., grasshoppers), and fish (salmon) (Goebel et al., 2011; Hockett, 2007). At Dust Cave, Alabama, late Paleoindians processed hickory and acorn nuts, collected seeds and fruits, and hunted a variety of animals, chiefly birds, and they even cached 23 Canada goose (*Branta canadensis*) humeri representing at least 12 individuals (Walker, 2010; Walker et al., 2007). Similarly, diverse diets have been documented for other regions of the continent. Residential sites with preserved remains of dwellings are rare but have been found at several Folsom sites in Colorado (e.g., Mountaineer and Barger Gulch) (Stiger, 2006; Surovell and Waguespack, 2007) as well as at a stemmed-point site in Oregon (Paulina Lake) (Connolly and Jenkins, 1999).

In South America, besides the rich inventory of organic artifacts recovered from Monte Verde (described in Section



Figure 6 A hearth feature shown in profile (a) with inset (b) of charred hawthorn seeds from the Shawnee-Minisink Clovis site, Pennsylvania, USA. Photographs courtesy of Joseph Gingerich.

'The Case for an Early Late-Glacial Dispersal, 16–14 ka'), most of our knowledge of Paleoindian technology is based on lithics. Fishtail bifacial points (Figure 2(k) and 2(l)), often fluted, seemingly dominate the Paleoindian record, but other projectile forms also occur including stemmed points of the Paijan complex in Peru, and leaf-shaped El Jobo points of Monte Verde (Figure 2(m)) and other early sites (Bryan et al., 1978; Dillehay et al., 2003, 2004), and barbed points at Pedra Pintada, Brazil (Figure 2(n)) (Roosevelt et al., 1998). Milling stones appear in some regions as early as ~12 ka (Dillehay et al., 2003; Gnecco, 2003a,b), as do sophisticated textiles and cordage (Jolie et al., 2011; Sandweiss et al., 1998). Increasingly broad diets are also evident: by the late Paleoindian period, ~12.5–12 ka, Amazonians at Pedra Pintada, were fishing and collecting tree fruits (Roosevelt et al., 1996; Figure 2(n)), while humans along the Pacific coast of Peru at Quebrada Jaguay had developed a specialized maritime adaptation that focused on net fishing for drums (*Sciaenae*) and collecting wedge clams (*Mesodesma donacium*) (Sandweiss et al., 1998). In the Southern Cone, at sites like Agua de la Cueva and Piedra Museo, late Paleoindians regularly hunted extant camelids and gathered eggs of the large flightless bird *ñandú* (Rheidae) (García, 2003; Miotti and Cattáneo, 2003). These Patagonian Paleoindians habitually transported tool stones long distances (Flegenheimer et al., 2003); however, in Peru, late Paleoindians increasingly turned toward local raw materials, presumably as territories shrunk and more permanent camps were developed (Dillehay et al., 2003). As in North America, the variety of archaeological complexes and presence of Paleoindian sites across the continent tell us that humans rapidly settled into South America's varied environments, by 13–12 ka.

Summary

Genetic evidence strongly suggests all Native Americans – past and present – originated in greater northeast Asia. Modern humans appeared in Siberia and eastern Asia by 40 ka, and dispersal from this homeland to the Americas may have begun before the LGM. The colonization process, however, appears to have stalled in Beringia for 10 000 years or more. There the earliest archaeological site, Yana RHS, dates to ~32–31 ka, but after this, the next earliest site dates to only ~14 ka. During the Late Glacial, 16–14 ka, Beringians seemingly dispersed south through two corridors that opened as the great ice sheets of western Canada retreated. One route of colonization may have been the northwest Pacific coast, while another was along the eastern flank of the Canadian Rocky Mountains. A few archaeological sites like Paisley Caves, Schaefer/Hebior, and Friedkin in North America and Monte Verde in South America suggest the presence of humans south of the ice sheets during this time, but it is not until the Clovis era, ~13 ka, that a consistent pattern of occupation emerges. Sites with Clovis artifacts occur across temperate North America and Mexico, possibly even as far south as Panama or Chile.

Genetic evidence combined with studies of early human skeletal remains suggest that only a single dispersal event out of Beringia occurred during the late Pleistocene. Humans spreading via both the coastal and interior corridors may

have done so from the same northern source, at roughly the same time. Although early Americans were craniometrically diverse, this variability is not surprising given the genetic mechanism of founder effect. Apparent similarities with modern southeast Asians more likely reflect the late Pleistocene relationship early Americans had with most early modern Asians, a relationship obscured by later Holocene craniofacial evolution resulting from dietary changes and other selective forces.

The first Americans dispersed rapidly, and by 13–12 ka they had successfully adapted to most of the New World's varied environments, including the Neotropics and Southern Cone of South America. In some settings, humans may have contributed to extinctions of Pleistocene megafauna. Clovis hunters, for example, regularly preyed on mammoth and mastodon, and in southern South America, there are repeated associations of artifacts and remains of extinct animals like giant ground sloth and giant armadillo. In other contexts, though, these animals were not the basis of Paleoindian subsistence, and it is clear that some early Americans subsisted on a variety of animal and plant species, most of which are still in existence today.

Although much is known about the late Quaternary dispersal of humans to the Americas, most details about the process remains to be discovered. In Beringia, for example, geologists and archaeologists from two continents still must work together to reveal the initial episode of colonization. New discoveries elsewhere in the Americas, as well as the development of new methods in genetics, archaeology, and Quaternary paleoecology, will undoubtedly lead to a more comprehensive and detailed portrayal of this important event in the history of the human species.

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See also: **Archaeological Records:** Overview; Postglacial Adaptations. **Vertebrate Studies:** Interactions with Hominids.

References

- Adovasio JM, Boldurian AT, and Carlisle RC (1988) Who are those guys? Some biased thoughts on the initial peopling of the New World. In: Carlisle RC (ed.) *Americans Before Columbus: Ice-Age Origins. Ethnology Monographs*, vol. 12, pp. 45–62. Pittsburgh: Department of Anthropology, University of Pittsburgh.
- Adovasio JM and Pedler DR (2004) Pre-Clovis sites and their implications for human occupation before the last glacial maximum. In: Madsen DB (ed.) *Entering America: Northeast Asia and Beringia Before the Last Glacial Maximum*, pp. 139–158. Salt Lake City: University of Utah Press.
- Akimova E, Higham T, Stasyuk I, Buzhilova A, Dobrovolskaya M, and Mednikova M (2010) A new direct radiocarbon AMS date for an Upper Palaeolithic human bone from Siberia. *Archaeometry* 52: 1122–1130.
- Aldenderfer M (1999) The Pleistocene/Holocene transition in Peru and its effects upon human use of the landscape. *Quaternary International* 53/54: 11–19.

- Anderson DG, Miller DS, Yerka SJ, et al. (2010) PIDBA (Paleoindian Database of the Americas) 2010: Current status and findings. *Archaeology of Eastern North America* 38: 63–90.
- Araujo AGM, Feathers JK, Arroyo-Kalin M, and Tizuka MM (2008) Lapa dos Boieiras Rockshelter: Stratigraphy and formation processes at a Paleoindian site in central Brazil. *Journal of Archaeological Science* 35: 3186–3202.
- Arroyo-Kalin M (2012) Slash-burn-and-churn: Landscape history and crop cultivation in pre-Columbian Amazonia. *Quaternary International* 249: 4–18.
- Ballenger JAM, Holliday VT, Kowler AL, et al. (2011) Evidence for Younger Dryas global climate oscillation and human response in the American Southwest. *Quaternary International* 242: 502–519.
- Barnosky AD, Koch PL, Feranec RS, Wing SL, and Shabel AB (2004) Assessing the causes of late Pleistocene extinctions on the continents. *Science* 306: 70–75.
- Barnosky AD and Lindsey EL (2010) Timing of Quaternary megafaunal extinction in South America in relation to human arrival and climate change. *Quaternary International* 217: 10–29.
- Basilyan AE, Anisimov MA, Nikolskiy PA, and Pitulko VV (2011) Woolly mammoth mass accumulation next to the Paleolithic Yana RHS site, Arctic Siberia: Its geology, age, and relation to past human activity. *Journal of Archaeological Science* 38: 2461–2474.
- Beck C and Jones GT (2010) Clovis and western stemmed: Population migration and the meeting of two technologies in the intermountain west. *American Antiquity* 75(1): 81–116.
- Bever MR (2001) An overview of Alaskan late Pleistocene archaeology: Historical themes and current perspectives. *Journal of World Prehistory* 15: 125–191.
- Bever MR (2006) Too little, too late? The radiocarbon chronology of Alaska and the peopling of the New World. *American Antiquity* 71(4): 595–620.
- Bicho NF, Haws JA, and Davis LG (eds.) (2011) *Trekking the shore: Changing coastlines and the antiquity of coastal settlement*. New York: Springer.
- Bigelow NH, Brubaker LB, Edwards ME, et al. (2003) Climate change and Arctic ecosystems: 1. Vegetation changes north of 55° N between the last glacial maximum, mid-Holocene, and present. *Journal of Geophysical Research* 108: 8170–8170-25. <http://dx.doi.org/10.1029/2002JD002558>.
- Boëda E (2010) Pedra Furada (Brasil, Piauí), una relectura tafonómica y tecnológica de las industrias líticas: o porque lo inaceptable puede ser aceptado? Paper presented at V Simposio Internacional *El Hombre Temprano en América*, La Plata, Argentina, 22–26 November.
- Borrero LA (2006) Paleoindians without mammoths and archaeologists without projectile points? The archaeology of the first inhabitants of the Americas. In: Morrow JE and Gnecco C (eds.) *Paleoindian Archaeology: A Hemispheric Perspective*, pp. 9–20. Gainesville, FL: University Press of Florida.
- Borrero LA (2008) Extinction of Pleistocene megamammals in South America: The lost evidence. *Quaternary International* 185: 69–74.
- Borrero LA (2011) The archaeology of transformation. *Quaternary International* 245: 178–181.
- Brace CL, Nelson AR, Seguchi N, et al. (2001) Old World sources of the first New World human inhabitants: A comparative craniofacial view. *Proceedings of the National Academy of Sciences of the United States of America* 98: 10017–10022.
- Bradley BA, Collins MB, and Hemmings A (2010) *Clovis Technology*. Ann Arbor, MI: International Monographs in Prehistory.
- Bradley B and Stanford D (2004) The North Atlantic ice-edge corridor: A possible Palaeolithic route to the New World. *World Archaeology* 36: 459–478.
- Brigham-Grette J, Lozhkin AV, Anderson PM, and Glushkova OY (2004) Paleoenvironmental conditions in western Beringia before and during the last glacial maximum. In: Madsen DB (ed.) *Entering America: Northeast Asia and Beringia Before the Last Glacial Maximum*, pp. 29–62. Salt Lake City: University of Utah Press.
- Bryan AL, Casimiquiela RM, Cruxent JM, Gruhn R, and Ochsenius C (1978) An El Jobo mastodon kill at Taima-tamia. *Science* 200: 1275–1277.
- Burbridge RE, Mayle FE, and Killen TJ (2004) Fifty-thousand-year vegetation and climate history of Noel Kempff Mercado National Park, Bolivian Amazon. *Quaternary Research* 61(2): 215–230.
- Cannon MD and Meltzer DJ (2004) Early Paleoindian foraging: Examining the faunal evidence for large mammal specialization and regional variability in prey choice. *Quaternary Science Reviews* 23: 1955–1987.
- Chatters JC (2000) The recovery and first analysis of an early Holocene human skeleton from Kennewick, Washington. *American Antiquity* 65(2): 291–316.
- Chatters JC (2002) *Ancient Encounters: Kennewick Man and the First Americans*. New York: Simon & Schuster.
- Chepstow-Lusty A, Bush MB, Frogley MR, Baker PA, Fritz SC, and Aronson J (2005) Vegetation and climate change on the Bolivian Altiplano between 108,000 and 18,000 yr ago. *Quaternary Research* 63(1): 90–98.
- Cinq-Mars J and Morlan RE (1999) Bluefish Caves and Old Crow basin: A new rapport. In: Bonnicksen R and Turnmire KL (eds.) *Ice Age Peoples of North America: Environments, Origins, and Adaptations of the First Americans*, pp. 200–212. Corvallis, OR: Oregon State University Press.
- Clague JJ, Mathewes RW, and Ager TA (2004) Environments of northwestern North America before the last glacial maximum. In: Madsen DB (ed.) *Entering America: Northeast Asia and Beringia Before the Last Glacial Maximum*, pp. 63–94. Salt Lake City: University of Utah Press.
- Collins MB (1997) The lithics from Monte Verde, a descriptive-morphological analysis. In: Dillehay TD (ed.) *Monte Verde: A Late Pleistocene Settlement in Chile*, pp. 383–506. Washington, DC: Smithsonian Institution Press.
- Comas D, Plaza S, Wells RS, et al. (2004) Admixture, migrations, and dispersals in central Asia: Evidence from maternal DNA lineages. *European Journal of Human Genetics* 12: 495–504.
- Connolly TJ and Barker P (2004) Basketry chronology of the Early Holocene in the northern Great Basin. In: Connolly TJ (ed.) *Early and middle Holocene Archaeology of the Northern Great Basin*, pp. 86–130. Eugene, OR: University of Oregon. Anthropological Papers 62.
- Connolly TJ and Jenkins DL (1999) The Paulina Lake site (35DS34). In: Connolly TJ (ed.) *Newberry Crater: A Ten-Thousand-Year Record of Human Occupation and Environmental Change in the Basin-Plateau Borderlands*, pp. 86–127. Salt Lake City, UT: University of Utah. Anthropological Paper No. 121.
- Coronato A, Martinez O, and Rabassa J (2004) Glaciations in Argentine Patagonia, southern South America. In: Ehlers J and Gibbard PL (eds.) *Quaternary Glaciations – Extent and Chronology*, Part III: pp. 49–67. Amsterdam: Elsevier.
- Delcourt PA and Delcourt HR (1987) *Long Term Forest Dynamics of the Temperate Zone: A Case Study of Late-Quaternary Forests in Eastern North America*. New York: Springer-Verlag.
- Dent RJ (2007) Seed collecting and fishing at the Shawnee Minisink Paleoindian site: Everyday life in the late Pleistocene. In: Walker RB and Driskell BN (eds.) *Foragers of the Terminal Pleistocene*, pp. 116–131. Lincoln: University of Nebraska Press.
- Derenko MV, Grzybowski T, Malyarchuk BA, Czarny J, Miścička-Śliwka D, and Zakharov IA (2001) The presence of mitochondrial haplogroup Z in Altaians from south Siberia. *American Journal of Human Genetics* 69: 237–241.
- Dickinson WR (2011) Geological perspectives on the Monte Verde archaeological site in Chile and pre-Clovis coastal migration in the Americas. *Quaternary Research* 76(2): 201–210.
- Dillehay TD (1997) *Monte Verde: A Late Pleistocene Settlement in Chile: The Archaeological Context*, vol. II. Washington, DC: Smithsonian Institution Press.
- Dillehay TD (2000) *The Settlement of the Americas: A New Prehistory*. New York: Basic Books.
- Dillehay T, Bonavia D, and Kaulicke P (2004) The first settlers. In: Silverman H (ed.) *Andean Archaeology*, pp. 16–34. Malden, MA: Blackwell.
- Dillehay TD, Rossen J, Maggard G, Stackelbeck K, and Netherly P (2003) Localization and possible social aggregation in the late Pleistocene and early Holocene on the north coast of Peru. *Quaternary International* 109–110: 3–11.
- Dixon EJ (2001) Human colonization of the Americas: Timing, technology and process. *Quaternary Science Reviews* 20: 277–299.
- Dyke AS (2004) An outline of North American deglaciation with emphasis on central and northern Canada. In: Ehlers J and Gibbard PL (eds.) *Quaternary Glaciations – Extent and Chronology*, Part II: pp. 373–424. Amsterdam: Elsevier.
- Dyke AS, Andrews JT, Clark PU, et al. (2002) The Laurentide and Innuitian ice sheets during the Last Glacial Maximum. *Quaternary Science Reviews* 21: 9–31.
- Elias SA and Crocker B (2008) The Bering land bridge: A moisture barrier to the dispersal of steppe-tundra biota? *Quaternary Science Reviews* 27: 2473–2483.
- Elias SA, Short SK, and Birks HH (1997) Late Wisconsin environments of the Bering land bridge. *Palaeogeography, Palaeoclimatology, Palaeoecology* 136: 293–308.
- Elias SA, Short SK, Nelson CH, and Birks HH (1996) Life and times of the Bering land bridge. *Nature* 382: 60–63.
- Ellis CJ, Carr DH, and Loebe TJ (2011) The Younger Dryas and late Pleistocene peoples of the Great Lakes region. *Quaternary International* 242: 534–545.
- Erlanson JM (2007) Sea change: The Paleocoastal occupations of Daisy Cave. In: Neusius SW and Gross GT (eds.) *Seeking Our Past: An Introduction to North American Archaeology*, pp. 135–143. Oxford: Oxford University Press.
- Erlanson JM, Moss ML, and Des Lauriers M (2008) Life on the edge: Early maritime cultures of the Pacific Coast of North America. *Quaternary Science Reviews* 27: 2232–2245.
- Erlanson JM, Rick TC, Braje TJ, et al. (2011) Paleoindian seafaring, maritime technologies, and coastal foraging on California's Channel Islands. *Science* 331: 1181–1185.

- Faith JT and Surovell TA (2009) Synchronous extinction of North America's Pleistocene mammals. *Proceedings of the National Academy of Sciences of the United States of America* 106: 20641–20645.
- Feathers J, Kipnis R, Piló L, Arroyo-Kalin M, and Coblenz D (2010) How old is Luzia? Luminescence dating and stratigraphic integrity at Lapa Vermelha, Lagoa Santa, Brazil. *Geoarchaeology* 25(4): 395–436.
- Fedje DW, Mackie Q, Dixon EJ, and Heaton TH (2004) Late Wisconsin environments and archaeological visibility on the northern Northwest Coast. In: Madsen DB (ed.) *Entering America: Northeast Asia and Beringia Before the Last Glacial Maximum*, pp. 97–138. Salt Lake City: University of Utah Press.
- Fiedel SJ (1999) Artifact provenience at Monte Verde: Confusion and contradictions. *Monte Verde Revisited: Discovering Archaeology Special Report*, pp. 1–12.
- Firestone RB, West A, Kennett JP, et al. (2007) Evidence for an extraterrestrial impact 12,900 years ago that contributed to the megafaunal extinctions and the Younger Dryas cooling. *Proceedings of the National Academy of Sciences of the United States of America* 104(41): 16016–16021.
- Flegenheimer N, Bayón C, Valente M, Baeza J, and Femení J (2003) Long distance tool stone transport in the Argentine Pampas. *Quaternary International* 109–110: 49–64.
- Forster P (2004) Ice Ages and the mitochondrial DNA chronology of human dispersals: A review. *Philosophical Transactions of the Royal Society of London, Series B: Biological Sciences* 359: 255–264.
- Fowler CS, Hattori EM, and Dansie AJ (2000) Ancient matting from Spirit Cave, Nevada: Technical implications. In: Drooker PB and Webster LD (eds.) *Beyond Cloth and Cordage: Archaeological Textile Research in the Americas*, pp. 119–139. Salt Lake City: University of Utah Press.
- Gaines EP, Sanchez-Miranda G, and Holliday VT (2009) Paleoindian archaeology in northern and central Sonora, Mexico: A review and update. *Kiva* 74: 305–335.
- García A (2003) Exploitation territory at Agua de la Cueva (southern area) site (11,000–9000 RCYBP). In: Miotti L, Salemme M, and Flegenheimer N (eds.) *Where the South Winds Blow: Ancient Evidence of Paleo South Americans*, pp. 83–86. College Station, TX: Center for the Study of the First Americans, Texas A&M University.
- Gilbert MTP, Jenkins DL, Götherstrom A, et al. (2008) DNA from pre-Clovis human coprolites in Oregon, North America. *Science* 320: 786–789.
- Gillespie AR and Zehfuss PH (2004) Glaciations of the Sierra Nevada, California, USA. In: Ehlers J and Gibbard PL (eds.) *Quaternary Glaciations – Extent and Chronology*, Part II: pp. 51–62. Amsterdam: Elsevier.
- Gingerich JAM (2011) Down to seeds and stones: A new look at the subsistence remains from Shawnee-Minisink. *American Antiquity* 76(1): 127–144.
- Glushkova OY (2001) Geomorphological correlation of late Pleistocene glacial complexes of western and eastern Beringia. *Quaternary Science Reviews* 20: 405–417.
- Gnecco C (2003a) Against ecological reductionism: Late Pleistocene hunter-gatherers in the tropical forests of northern South America. *Quaternary International* 109–110: 13–21.
- Gnecco C (2003b) Agrilocalities during the Pleistocene/Holocene transition in northern South America. In: Miotti L, Salemme M, and Flegenheimer N (eds.) *Where the South Wind Blows: Ancient Evidence of Paleo South Americans*, pp. 7–11. College Station, TX: Center for the Study of the First Americans, Texas A&M University.
- Goebel T (1999) Pleistocene colonization of Siberia and peopling of the Americas: An ecological approach. *Evolutionary Anthropology* 8: 208–227.
- Goebel T (2002) The 'microblade adaptation' and recolonization of Siberia during the late Upper Pleistocene. In: Elston RG and Kuhn SL (eds.) *Thinking Small: Global Perspectives on Microlithization*, pp. 117–132. Arlington, VA: American Anthropological Association. Archeological Paper No. 12.
- Goebel T, Hockett B, Adams KD, Rhode D, and Graf K (2011) Climate, environment, and humans in North America's Great Basin during the Younger Dryas, 12,900–11,600 calendar years ago. *Quaternary International* 242: 479–501.
- Goebel T, Slobodin SB, and Waters MR (2010) New dates from Ushki-1, Kamchatka, confirm 13,000 cal BP age for earliest Paleolithic occupation. *Journal of Archaeological Science* 37: 2640–2649.
- Goebel T, Waters MR, and O'Rourke DH (2008) The Late Pleistocene dispersal of modern humans in the Americas. *Science* 319: 1497–1502.
- Gonzalez S (2006) Global expansion 300,000–8000 years ago, Americas. In: Elias SA (ed.) *Encyclopedia of Quaternary Science*, pp. 129–135. Amsterdam: Elsevier.
- González González AH, Sandoval CR, Mata AT, et al. (2008) The arrival of humans on the Yucatan Peninsula: Evidence from submerged caves in the state of Quintana Roo, Mexico. *Current Research in the Pleistocene* 25: 1–24.
- González S, Huddart D, Bennett MR, and González-Huseca A (2006) Human footprint in central Mexico older than 40,000 years. *Quaternary Science Reviews* 25: 201–222.
- González-José R, Bortolini MC, Santos SR, and Bonatto SL (2008) The peopling of America: Craniofacial shape variation on a continental scale and its interpretation from an interdisciplinary view. *American Journal of Physical Anthropology* 137: 175–187.
- González-José R, González-Martín A, Hernández M, et al. (2003) Craniometric evidence for Palaeoamerican survival in Baja California. *Nature* 425: 62–64.
- González-José R, Neves W, Lahr MM, et al. (2005) Late Pleistocene/Holocene craniofacial morphology in Mesoamerican Paleoindians: Implications for the peopling of the New World. *American Journal of Physical Anthropology* 128: 772–780.
- Goodyear AC (2005a) Summary of the Allendale Paleoindian expedition – 2003 and 2004 field seasons. *Legacy: South Carolina Institute of Archaeology and Anthropology* 9(1–2): 1.
- Goodyear AC (2005b) Evidence for pre-Clovis sites in the eastern United States. In: Bonnicksen R, Lepper BT, Stanford D, and Waters MR (eds.) *Paleoamerican Origins: Beyond Clovis*, pp. 103–112. College Station, TX: Center for the Study of the First Americans, Texas A&M University.
- Graf KE and Bigelow NH (2011) Human response to climate during the Younger Dryas chronozone in central Alaska. *Quaternary International* 242: 434–451.
- Grayson DK (1989) The chronology of North American late Pleistocene extinctions. *Journal of Archaeological Science* 16: 153–165.
- Grayson DK and Meltzer DJ (2002) Clovis hunting and large mammal extinction: A critical review of the evidence. *Journal of World Prehistory* 16(4): 313–359.
- Grayson DK and Meltzer DJ (2003) Requiem for North American overkill. *Journal of Archaeological Science* 30: 585–593.
- Green TJ, Cochran B, Fenton TW, et al. (1998) The Buhl burial: A Paleoindian woman from southern Idaho. *American Antiquity* 63: 437–456.
- Gruhn R (2005) The ignored continent: South America in models of earliest American prehistory. In: Bonnicksen R, Lepper BT, Stanford D, and Waters MR (eds.) *Paleoamerican Origins: Beyond Clovis*, pp. 199–208. College Station, TX: Center for the Study of the First Americans, Department of Anthropology, Texas A&M University.
- Guidon N and Arnaud B (1991) The chronology of the New World: Two faces of one reality. *World Archaeology* 23(2): 167–178.
- Guidon N and Delibrias G (1986) Carbon-14 dates point to man in the Americas 32,000 years ago. *Nature* 321: 769–771.
- Gustafson CE, Gilbow D, and Daugherty R (1979) The Manis Mastodon site: Early man on the Olympic Peninsula. *Canadian Journal of Archaeology* 3: 157–164.
- Guthrie RD (1990) *Frozen Fauna of the Mammoth Steppe: The Story of Blue Babe*. Chicago: University of Chicago Press.
- Guthrie RD (2003) Rapid body size decline in Alaskan Pleistocene horses before extinction. *Nature* 426: 169–171.
- Hamilton TD and Goebel T (1999) Late Pleistocene peopling of Alaska. In: Bonnicksen R and Turnmire K (eds.) *Ice Age Peoples of North America: Environments, Origins, and Adaptations of the First Americans*, pp. 156–199. Corvallis, OR: Oregon State University Press.
- Harrison S (2004) The Pleistocene glaciations of Chile. In: Ehlers J and Gibbard PL (eds.) *Quaternary Glaciations – Extent and Chronology*, Part III: pp. 89–103. Amsterdam: Elsevier.
- Hastenrath S (2009) Past glaciation in the tropics. *Quaternary Science Reviews* 28: 790–798.
- Haynes CV Jr (1999) Monte Verde and the Pre-Clovis situation in America. *Monte Verde Revisited: Discovering Archaeology Special Report*, pp. 17–19.
- Haynes GA (2002) *The Early Settlement of North America: The Clovis Era*. Cambridge: Cambridge University Press.
- Haynes CV Jr. (2005) Clovis, Pre-Clovis, climate change, and extinction. In: Bonnicksen R, Lepper BT, Stanford D, and Waters MR (eds.) *Paleoamerican Origins: Beyond Clovis*, pp. 113–132. College Station, TX: Center for the Study of the First Americans, Department of Anthropology, Texas A&M University.
- Haynes G (2007a) A review of some attacks on the overkill hypothesis, with special attention to misrepresentations and doubletalk. *Quaternary International* 169–170: 84–94.
- Haynes GA (2007b) Rather odd detective stories: A view of some actualistic and taphonomic trends in Paleoindian studies. In: Pickering T, Schick K, and Toth N (eds.) *African Taphonomy: A Tribute to the Career of C. K. 'Bob' Brain*, pp. 25–35. Bloomington, IN: CRAFT Center, Indiana University. Stone Age Institute Publications Series, No. 2.
- Haynes GA and Krasinski KE (2010) Taphonomic fieldwork in southern Africa and its application in studies of the earliest peopling of North America. *Journal of Taphonomy* 8: 181–202.
- Helmens KF (2004) The Quaternary glacial record of the Colombian Andes. In: Ehlers J and Gibbard PL (eds.) *Quaternary Glaciations – Extent and Chronology*, Part III: pp. 115–134. Amsterdam: Elsevier.
- Hessler I, Dupont L, Bonnefille R, et al. (2010) Millennial-scale changes in vegetation records from tropical Africa and South America during the last glacial. *Quaternary Science Reviews* 29: 2882–2899.

- Hillyer R, Valencia BG, Bush MB, Silman MR, and Steinitz-Kannan M (2009) A 24,700-yr paleolimnological history from the Peruvian Andes. *Quaternary Research* 71(1): 71–82.
- Hockett B (2007) Nutritional ecology of late Pleistocene to middle Holocene subsistence in the Great Basin: Zooarchaeological evidence from Bonneville Estates Rockshelter. In: Graf KE and Schmitt DN (eds.) *Paleoindian or Paleoarchaic? Great Basin Human Ecology at the Pleistocene-Holocene Transition*, pp. 204–230. Salt Lake City: University of Utah Press.
- Hodell DA, Anselmetti FS, Ariztegui D, et al. (2008) An 85-ka record of climate change in lowland Central America. *Quaternary Science Reviews* 27: 1152–1165.
- Hoffecker JF (2005) *A Prehistory of the North: Human Settlement of the Higher Latitudes*. New Brunswick: Rutgers University Press.
- Hoffecker JF and Elias SA (2007) *Human Ecology of Beringia*. New York: Columbia University Press.
- Hoffecker JF, Powers WR, and Goebel T (1993) The colonization of Beringia and the peopling of the New World. *Science* 259: 46–53.
- Holen SR (2006) Taphonomy of two last glacial maximum mammoth sites in the central Great Plains of North America: A preliminary report on La Sena and Lovewell. *Quaternary International* 142–143: 30–43.
- Holen SR (2007) The age and taphonomy of mammoths at Lovewell Reservoir, Jewell County, Kansas, USA. *Quaternary International* 169–170: 51–63.
- Holmes CE (2001) Tanana valley archaeology circa 14,000 to 9000 B.P. *Arctic Anthropology* 38(2): 154–170.
- Holmes CE (2011) The Beringian and Transitional Periods in Alaska: Technology of the East Beringian tradition as viewed from Swan Point. In: Goebel T and Buvit I (eds.) *From the Yenisei to the Yukon: Interpreting Lithic Assemblage Variability in Late Pleistocene/Early Holocene Beringia*, pp. 179–191. College Station, TX: Texas A&M University Press.
- Hubbe M, Neves WA, and Harvati K (2010) Testing evolutionary and dispersion scenarios for the settlement of the New World. *PLoS One* 5(6): e11105. <http://dx.doi.org/10.1371/journal.pone.0011105>.
- Huddart D, Bennett MR, González S, and Velay X (2008) Analysis and preservation of Pleistocene human and animal footprints: An example from Toluquilla, Valsequillo Basin (central Mexico). *Ichnos* 15: 232–245.
- Jackson SD (2003) Evaluating evidence of cultural associations of Mylodon in the semiarid region of Chile. In: Miotto L, Salemme M, and Flegenheimer N (eds.) *Where the South Winds Blow*, pp. 77–82. College Station, TX: Center for the Study of the First Americans, Texas A&M University.
- Jackson LJ (2006) Fluted and fishtail points from southern coastal Chile. In: Morrow JE and Gnecco C (eds.) *Paleoindian Archaeology: A Hemispheric Perspective*, pp. 105–120. Gainesville, FL: University Press of Florida.
- Jackson D, Méndez C, Seguel R, Maldonado A, and Vargas G (2007) Initial occupation of the Pacific coast of Chile during late Pleistocene times. *Current Anthropology* 48(5): 725–731.
- Jantz RL and Owsley DW (1997) Pathology, taphonomy, and cranial morphometrics of the Spirit Cave Mummy. *Nevada Historical Society Quarterly* 40: 62–84.
- Jenkins DL (2007) Distribution and dating of cultural and paleontological remains at the Paisley Five Mile Point Caves in the northern Great Basin: An early assessment. In: Graf KE and Schmitt DN (eds.) *Paleoindian or Paleoarchaic? Great Basin Human Ecology at the Pleistocene-Holocene Transition*, pp. 57–81. Salt Lake City: University of Utah Press.
- Johnson E (2007) Along the ice margin – The cultural taphonomy of late Pleistocene mammoth in southeastern Wisconsin (USA). *Quaternary International* 169–170: 64–83.
- Johnson JR, Stafford TW Jr., Ajie HO, and Morris DP (1999) Arlington Springs revisited. *Proceedings of the Fifth California Islands Symposium*, 23 March to 1 April 1999, pp. 541–545. Washington, DC: US Department of the Interior, Minerals Management Service, Pacific Outer Continental Shelf Region.
- Jolie E, Lynch TF, Geib PR, and Adovasio JM (2011) Cordage, textiles, and the late Pleistocene peopling of the Andes. *Current Anthropology* 52(2): 285–296.
- Jones GT, Beck C, Jones EE, and Hughes RE (2003) Lithic source use and Paleoarchaic foraging territories in the Great Basin. *American Antiquity* 68: 5–38.
- Joyce DJ (2006) Chronology and new research on the Schaefer mammoth (*Mammuthus primigenius*) site, Kenosha County, Wisconsin, USA. *Quaternary International* 142–143: 44–57.
- Kaestle F and Smith DG (2001) Ancient Native American DNA from western Nevada: Implications for the Numic expansion hypothesis. *American Journal of Physical Anthropology* 115: 1–12.
- Kaiser J, Schefuß E, Lamy F, Mohtadi M, and Hebbeln D (2008) Glacial to Holocene changes in sea surface temperature and coastal vegetation in north central Chile: High versus low latitude forcing. *Quaternary Science Reviews* 27: 2064–2075.
- Karafet TM, Zegura SL, and Hammer MF (2006) Y chromosomes. In: Ubelaker DH (ed.) *Handbook of North American Indians. Environment, Origins, and Population*, vol. 3, pp. 831–839. Washington DC: Smithsonian Institution Press.
- Kaufman DS and Manley WF (2004) Pleistocene maximum and late Wisconsinan glacier extents across Alaska, U.S.A. In: Ehlers J and Gibbard PL (eds.) *Quaternary Glaciations – Extent and Chronology*, Part II: pp. 9–27. Amsterdam: Elsevier.
- Keeley JE (2002) Native American impacts on fire regimes of the California coastal ranges. *Journal of Biogeography* 29: 303–320.
- Kemp BM, Malhi RS, McDonough J, et al. (2007) Genetic analysis of early Holocene skeletal remains from Alaska and its implications for the settlement of the Americas. *American Journal of Physical Anthropology* 132: 605–621.
- Kenady SM, Wilson MC, Schalk RF, and Mierendorf RR (2011) Late Pleistocene butchered Bison antiquus from Ayer Pond, Orcas Island, Pacific Northwest: Age confirmation and taphonomy. *Quaternary International* 233: 130–141.
- Kilby JD (2011) Les caches Clovis dans le cadre du Paléolindien ancien en Amérique du Nord. In: Vialous D (ed.) *Peuplements et Préhistoire en Amériques*. Paris, France: Muséum National d'Histoire Naturelle.
- Kohntopp SW (2011) *The Simon Clovis Cache: One of the Oldest Archaeological Sites in Idaho*. College Station, TX: Center for the Study of the First Americans, Department of Anthropology, Texas A&M University.
- Kuhle M (2004) The last glacial maximum (LGM) glacier cover of the Aconcagua group and adjacent massifs in the Mendoza Andes (South America). In: Ehlers J and Gibbard PL (eds.) *Quaternary Glaciations – Extent and Chronology*, Part III: pp. 75–88. Amsterdam: Elsevier.
- Kuzmina S, Elias S, Matheus P, Storer JE, and Sher A (2008) Paleoenvironmental reconstruction of the last glacial maximum, inferred from insect fossils from a tephra buried soil at Tempest Lake, Seward Peninsula, Alaska. *Palaeogeography, Palaeoclimatology, Palaeoecology* 267: 245–255.
- Lahr MM (1995) Patterns of modern human diversification: Implications for Amerindian origins. *Yearbook of Physical Anthropology* 38: 163–198.
- Lahr MM (1997) History in the bones. *Evolutionary Anthropology* 6: 2–6.
- Lorenzen ED, Nogués-Bravo D, Orlando L, et al. (2011) Species-specific responses of Late Quaternary megafauna to climate and humans. *Nature* 479: 359–364.
- Lowery DL, O'Neil MA, Wah JS, Wagner DP, and Stanford DJ (2010) Late Pleistocene upland stratigraphy of the western Delmarva Peninsula, USA. *Quaternary Science Reviews* 29(11–12): 1472–1480.
- Lozano-García S, Sosa-Nájera S, Sugiura Y, and Caballero M (2005) 23,000 yr of vegetation history of the Upper Lerma, a tropical high-altitude basin in central Mexico. *Quaternary Research* 64(1): 70–82.
- Lynch TE (1990) Glacial-age man in South America: A critical review. *American Antiquity* 55(1): 12–36.
- MacPhee RDE and Marx PA (1997) The 40,000 year plague: Humans, hyperdisease and first-contact extinctions. In: Goodman S and Patterson B (eds.) *Natural Change and Human Impact in Madagascar*, pp. 169–217. Washington, DC: Smithsonian Institution Press.
- Madsen D (2005) Recapitulation: The relative probabilities of late pre-LGM or early post-LGM ages for the initial occupation of the Americas. In: Madsen DB (ed.) *Entering America: Northeast Asia and Beringia Before the Last Glacial Maximum*, pp. 389–396. Salt Lake City: University of Utah Press.
- Malhi RS, Kemp BM, Eshleman JA, et al. (2007) Haplogroup M discovered in prehistoric North America. *Journal of Archaeological Science* 34: 642–648.
- Mandryk CAS, Rosenhans H, Fedje DW, and Mathews RW (2001) Late Quaternary paleoenvironments of northwestern North America: Implications for inland versus coastal migration routes. *Quaternary Science Reviews* 20: 301–314.
- Manley WF (2002) *Post-glacial flooding of the Bering land bridge: A geospatial animation*. INSTAAR, University of Colorado. http://instaar.colorado.edu/qgis/bering_land_bridge.
- Mark BG, Harrison SP, Spessa A, New M, Evans DJA, and Helmens KF (2005) Tropical snowline changes at the last glacial maximum: A global assessment. *Quaternary International* 138–139: 168–201.
- Markgraf V (1989) Paleoclimates in central and south America since 18,000 BP based on pollen and lake-level records. *Quaternary Science Reviews* 8: 1–24.
- Marlon JR, Bartlein PJ, Walsh MK, et al. (2009) Wildfire responses to abrupt climate change in North America. *Proceedings of the National Academy of Sciences of the United States of America* 106(8): 2519–2524.
- Martin PS (1967) Prehistoric overkill. In: Martin PS and Wright HE Jr. (eds.) *Pleistocene Extinctions: The Search for a Cause*, pp. 75–120. New Haven, CT: Yale University Press.
- Martin PS (1984) Prehistoric overkill: The global model. In: Martin PS and Klein RG (eds.) *Quaternary Extinctions: A Prehistoric Revolution*, pp. 354–403. Tucson, AZ: University of Arizona Press.

- Martínez GA (2001) Fish-tail projectile points and megamammals: New evidence from Paso Otero 5 (Argentina). *Antiquity* 75: 523–528.
- Massone M (2003) Fell 1 hunters' hearths in the Magallanes region by the end of the Pleistocene. In: Miotto L, Salemme M, and Flegenheimer N (eds.) *Where the South Winds Blow*, pp. 153–159. College Station, TX: Center for the Study of the First Americans, Texas A&M University.
- McAvoy JM and McAvoy LD (1997) Archaeological investigations of site 44SX202. Research Report Series No. 8. Virginia Department of Historic Resources, Richmond, VA.
- Meltzer DJ (2009) *First Peoples in a New World: Colonizing Ice Age America*. Berkeley, CA: University of California Press.
- Meltzer DJ, Adovasio JM, and Dillehay TD (1994) On a Pleistocene human occupation at Pedra Furada, Brazil. *Antiquity* 68: 695–714.
- Meltzer DJ, Grayson DK, Ardila G, et al. (1997) On the Pleistocene antiquity of Monte Verde, southern Chile. *American Antiquity* 62(4): 659–663.
- Merriwether DA (2006) Mitochondrial DNA. In: Ubelaker DH (ed.) *Handbook of North American Indians: Environment, Origins, and Population*, vol. 3, pp. 817–830. Washington, DC: Smithsonian Institution Press.
- Metcalfe S, Say A, Black S, McCulloch R, and O'Hara S (2002) Wet conditions during the last glaciation in the Chihuahuan Desert, Alta Babicora Basin, Mexico. *Quaternary Research* 57(1): 91–101.
- Miotto L and Cattáneo R (2003) Variations in strategies of lithic production and faunal exploitation in the Pleistocene/Holocene transition at Piedra Museo and surrounding region. In: Miotto L, Salemme M, and Flegenheimer N (eds.) *Where the South Winds Blow*, pp. 105–112. College Station, TX: Center for the Study of the First Americans, Texas A&M University.
- Miotto L, Salemme M, and Rabassa J (2003) Radiocarbon chronology at Piedra Museo locality. In: Miotto L, Salemme M, and Flegenheimer N (eds.) *Where the South Winds Blow*, pp. 99–104. College Station, TX: Center for the Study of the First Americans, Texas A&M University.
- Mochanov IuA (1977) *Drevneishie Etapy Zaseleniia Chelovekom Severo-Vostochnoi Azii*. Novosibirsk: Nauka.
- Morlan RE (2003) Current perspectives on the Pleistocene archaeology of eastern Beringia. *Quaternary Research* 60: 123–132.
- Morrow JE and Fiedel SJ (2006) New radiocarbon dates for the Clovis component of the Anzick site, Park County, Montana. In: Morrow JE and Gnecco C (eds.) *Paleoindian Archaeology: A Hemispheric Perspective*, pp. 86–104. Gainesville, FL: University Press of Florida.
- Munyikwa K (2005) Synchrony of Southern Hemisphere late Pleistocene arid episodes: A review of luminescence chronologies from arid Aeolian landscapes south of the Equator. *Quaternary Science Reviews* 24: 2555–2583.
- Munyikwa K, Feathers J, Rittenour T, and Shrimpton H (2011) Constraining the late Wisconsinan retreat of the Laurentide ice sheet from western Canada using luminescence ages from postglacial aeolian dunes. *Quaternary Geochronology* 6: 407–422.
- Neves WA, González-José R, Hubbe M, Kipnis R, Araujo AGM, and Blasi O (2004) Early human skeletal remains from Cerca Grande, Lagoa Santa, central Brazil, and the origins of the first Americans. *World Archaeology* 36: 479–501.
- Neves WA and Hubbe M (2005) Cranial morphology of early Americans from Lagoa Santa, Brazil: Implications for the settlement of the New World. *Proceedings of the National Academy of Sciences of the United States of America* 102: 18309–18314.
- Neves WA, Hubbe M, and Correal G (2007a) Human skeletal remains from Sabana de Bogotá, Colombia: A case for Paleoamerican morphology late survival in South America? *American Journal of Physical Anthropology* 133: 1080–1098.
- Neves WA, Hubbe M, and Piló LB (2007b) Early Holocene human skeletal remains from Sumidouro Cave, Lagoa Santa, Brazil: History of discoveries, geological and chronological context, and comparative cranial morphology. *Journal of Human Evolution* 52: 16–30.
- Neves WA, Powell JF, and Ozolins EG (1999) Extra-continental morphological affinities of Lapa Vermelha IV, Hominid 1: A multivariate analysis with progressive numbers of variables. *Homo* 50: 263–282.
- Overstreet DF (2005) Late-glacial ice-marginal adaptation in southeastern Wisconsin. In: Bonnichsen R, Lepper BT, Stanford D, and Waters MR (eds.) *Paleoamerican Origins: Beyond Clovis*, pp. 183–195. College Station, TX: Center for the Study of the First Americans, Department of Anthropology, Texas A&M University Press.
- Overstreet DF and Kolb MF (2003) Geoarchaeological contexts for Late Pleistocene archaeological sites with human-modified woolly mammoth remains in southeastern Wisconsin. *Geoarchaeology: An International Journal* 18(1): 91–114.
- Owsley DW, Jodry MA, Stafford TW Jr., Haynes CV Jr., and Stanford DJ (2010) *Arch Lake Woman: Physical Anthropology and Geoarchaeology*. College Station, TX: Texas A&M University Press.
- Paunero RS (2003) The Cerro Tres Tetras (C3T) locality in the Central Plateau of Santa Cruz, Argentina. In: Miotto L, Salemme M, and Flegenheimer N (eds.) *Where the South Winds Blow*, pp. 133–140. College Station, TX: Center for the Study of the First Americans, Texas A&M University.
- Pearson GA (1999) Early occupations and cultural sequences at Moose Creek: A late Pleistocene site in central Alaska. *Arctic* 52(4): 332–345.
- Perego UA, Achilli A, Angerhofer N, et al. (2009) Distinctive but concomitant Palaeo-Indian migration routes from Beringia marked by two rare mtDNA haplogroups. *Current Biology* 19: 1–8.
- Perego UA, Angerhofer N, Pala M, et al. (2010) The initial peopling of the Americas: A growing number of founding mitochondrial genomes from Beringia. *Genome Research* 20: 1174–1179.
- Perez SI, Bernal V, Gonzalez PN, Sardi M, and Politis GG (2009) Discrepancy between cranial and DNA data of early Americans: Implications for American peopling. *PLoS One* 4(5): e5746. <http://dx.doi.org/10.1371/journal.pone.0005746>.
- Pessenda LCR, De Oliveira PE, Molatto M, et al. (2009) The evolution of a tropical rainforest/grassland mosaic in southeastern Brazil since 28,000 14C yr BP based on carbon isotopes and pollen records. *Quaternary Research* 71(3): 437–452.
- Pierce KL (2003) Pleistocene glaciations of the Rocky Mountains. *Development in Quaternary Science*, vol. 1, pp. 63–76. Amsterdam: Elsevier. [http://dx.doi.org/10.1016/S1571-0866\(03\)01004-2](http://dx.doi.org/10.1016/S1571-0866(03)01004-2).
- Pinter N, Fiedel S, and Keeley JE (2011) Fire and vegetation shifts in the Americas at the vanguard of Paleoindian migration. *Quaternary Science Reviews* 30: 269–272.
- Piperno DR (2009) The origins of plant cultivation and domestication in the new world tropics. *Current Anthropology* 52(S4): S453–S470.
- Pitul'ko VV and Pavlova EIU (2010) *Geoarkheologiya i Radiouglerodnaia Khronologiya Kamennogo Veka Severo-Vostochnoi Azii*. Sankt-Peterburg: Nauka.
- Pitulko VV, Nikolsky PA, Giryay EY, et al. (2004) The Yana RHS site: Humans in the Arctic before the last glaciation. *Science* 303: 52–56.
- Politis G, Guitierrez MA, and Scabuzzon C (in press) *Estado Actual de la Investigaciones en el Sitio Arroyo Seco 2 (Región Pampeana, Argentina)*. Serie Monográfica, INCUAPA 6, Facultad de Ciencias Sociales, UNCPBA, Olavarría.
- Politis GG, Scabuzzon C, and Tykot RH (2009) An approach to pre-Hispanic diets in the Pampas during the early/middle Holocene. *International Journal of Osteoarchaeology* 19: 266–280.
- Potter BA, Irish JD, Reuther JD, Gelvin-Reymiller C, and Holliday VT (2011) A terminal Pleistocene child cremation and residential structure from eastern Beringia. *Science* 331: 1058–1062.
- Powell JF (2005) *The First Americans: Race, Evolution and the Origin of Native Americans*. Cambridge: Cambridge University Press.
- Powell JF and Neves WA (1999) Craniofacial morphology of the first Americans: Pattern and process in the peopling of the New World. *American Journal of Physical Anthropology* 110(S29): 153–188.
- Power MJ, Marlon J, Ortiz N, et al. (2008) Changes in fire regimes since the last glacial maximum: An assessment based on a global synthesis and analysis of charcoal data. *Climate Dynamics* 30: 887–907.
- Quattrocchio MA, Borromei AM, Deschamps CM, Grill SC, and Zavala CA (2008) Landscape evolution and climate changes in the late Pleistocene-Holocene, southern Pampa (Argentina): Evidence from palynology, mammals and sedimentology. *Quaternary International* 181: 123–138.
- Raff JA, Bolnick DA, Tackney J, and O'Rourke DH (2011) Ancient DNA perspectives on American colonization and population history. *American Journal of Physical Anthropology* 146: 503–514.
- Ranere TJ (2006) The Clovis colonization of Central America. In: Morrow JE and Gnecco C (eds.) *Paleoindian Archaeology: A Hemispheric Perspective*, pp. 69–85. Gainesville, FL: University Press of Florida.
- Rasic JT (2011) Functional variability in the late Pleistocene archaeological record of eastern Beringia: A model of late Pleistocene land use and technology from northwest Alaska. In: Goebel T and Buvit I (eds.) *From the Yenisei to the Yukon: Interpreting Lithic Assemblage Variability in Late Pleistocene/Early Holocene Beringia*, pp. 128–164. College Station, TX: Texas A&M University Press.
- Renne PR, Feinberg JM, Waters MR, et al. (2005) Geochronology: Age of Mexican ash with alleged 'footprints'. *Nature* 438: E7–E8. <http://dx.doi.org/10.1038/nature04425>.
- Roosevelt AC, Lima da Costa M, Lopes Machado C, et al. (1996) Paleoindian cave dwellers in the Amazon: The peopling of the Americas. *Science* 272: 373–384.
- Rowe MW and Steelman KL (2003) Comment on 'some evidence of a date of first humans to arrive in Brazil'. *Journal of Archaeological Science* 30: 1349–1351.
- Sandweiss DH, McInnis H, Burger RL, et al. (1998) Quebrada Jaguay: Early South American maritime adaptations. *Science* 281: 1830–1832.
- Santos GM, Bird MI, Parenti F, Fifield LK, Guidon N, and Hausladen PA (2003) A revised chronology of the lowest occupation layer of Pedra Furada Rock Shelter, Piauí, Brazil: The Pleistocene peopling of the Americas. *Quaternary Science Reviews* 22: 2303–2310.

- Schroeder KB, Jakobsson M, Crawford MH, et al. (2009) Haplotypic background of a private allele at high frequency in the Americas. *Molecular Biological Evolution* 26: 995–1016.
- Schurr TG and Sherry ST (2004) Mitochondrial DNA and Y chromosome diversity and the peopling of the Americas: Evolutionary and demographic evidence. *American Journal of Human Biology* 16: 420–439.
- Shang H and Trinkaus E (2010) *The Early Modern Human from Tianyuan Cave*. College Station, TX: China. Texas A&M University Press.
- Shapiro B, Drummond AJ, Rambaut A, et al. (2004) Rise and fall of the Beringian steppe bison. *Science* 306: 1561–1564.
- Smallwood AM (2010) Clovis biface technology at the Topper site, South Carolina: Evidence for variation and technological flexibility. *Journal of Archaeological Science* 37: 2413–2425.
- Smith DG, Malhi RS, Eshleman JA, Kaestle FA, and Kemp BM (2005) Mitochondrial DNA haplogroups of Paleoamericans in North America. In: Bonnicksen R, Lepper BT, Stanford D, and Waters MR (eds.) *Paleoamerican Origins: Beyond Clovis*, pp. 243–254. College Station, TX: Center for the Study of the First Americans, Department of Anthropology, Texas A&M University Press.
- Stafford TW Jr., Jull AJT, Brendel K, Duhamel RC, and Donahue D (1987) Study of bone radiocarbon dating accuracy at the University of Arizona NSF Accelerator Facility for Radioisotope Analysis. *Radiocarbon* 29(1): 24–44.
- Stanford D, Bonnicksen R, Meggers B, and Steele DG (2005) Paleoamerican origins: Models, evidence, and future directions. In: Bonnicksen R, Lepper BT, Stanford D, and Waters MR (eds.) *Paleoamerican Origins: Beyond Clovis*, pp. 313–354. College Station, TX: Center for the Study of the First Americans, Department of Anthropology, Texas A&M University Press.
- Starikovskaya EB, Sukernik RI, Derbeneva OA, et al. (2005) Mitochondrial DNA diversity in indigenous populations of the southern extent of Siberia, and the origins of Native American haplogroups. *Annals of Human Genetics* 69: 67–89.
- Steele J and Politis G (2009) AMS ¹⁴C dating of early human occupation of southern South America. *Journal of Archaeological Science* 36: 419–429.
- Steele DG and Powell JF (2002) Facing the past: A view of the North American human fossil record. In: Jablonski NG (ed.) *The First Americans, Memoirs of the California Academy of Sciences*, vol. 27, pp. 93–122. San Francisco, CA: California Academy of Sciences.
- Stiger M (2006) A Folsom structure in the Colorado mountains. *American Antiquity* 71(2): 321–351.
- Straus LG, Meltzer DJ, and Goebel T (2005) Ice Age Atlantis? Exploring the Solutrean–Clovis ‘connection’. *World Archaeology* 37(4): 507–532.
- Surovell TA and Waguespack NM (2007) Folsom hearth-centered use of space at Barger Gulch, Locality B. In: Brunswig RS and Pitblado B (eds.) *Emerging Frontiers in Colorado Paleoindian Archaeology*, pp. 219–259. Boulder, CO: University of Colorado Press.
- Tamm E, Kivisild T, Reidla M, et al. (2007) Beringian standstill and spread of Native American founders. *PLoS One* 2: 3829. <http://dx.doi.org/10.1371/journal.pone.0000829>.
- Tankersley KB (2004) The concept of Clovis and the peopling of North America. In: Barton CM, Clark GA, Yesner DR, and Pearson GA (eds.) *The Settlement of the American Continents: A Multidisciplinary Approach to Human Biogeography*, pp. 49–63. Tucson, AZ: University of Arizona Press.
- Tripaldi A, Zárate MA, Brook GA, and Li G-Q (2011) Late Quaternary paleoenvironments and paleoclimatic conditions in the distal Andean piedmont, southern Mendoza, Argentina. *Quaternary Research* 76(2): 253–263.
- Vialou ÁV (2003) Santa Elina Rockshelter, Brazil: Evidence of the coexistence of man and *Glossotherium*. In: Miotto L, Salemme M, and Flegenheimer N (eds.) *Where the South Winds Blow*, pp. 21–28. College Station, TX: Center for the Study of the First Americans, Texas A&M University.
- Walker RB (2007) Hunting in the late Paleoindian period: Faunal remains from Dust Cave, Alabama. In: Walker RB and Driskell BN (eds.) *Foragers of the Terminal Pleistocene in North America*, pp. 99–115. Lincoln: University of Nebraska Press.
- Walker RB (2010) Paleoindian and Archaic activities at Dust Cave, Alabama: The secular and the sacred. *North American Archaeologist* 31(3–4): 427–445.
- Walters MR and Stafford TW Jr. (2007) Redefining the age of Clovis: Implications for the peopling of the Americas. *Science* 315: 1122–1126.
- Watanabe S, Ayta WEF, Hamaguchi H, et al. (2003) Some evidence of a date of first humans to arrive in Brazil. *Journal of Archaeological Science* 30: 351–354.
- Waters MR (1985) Early man in the New World: An evaluation of the radiocarbon dated pre-Clovis sites in the Americas. In: Mead JI and Meltzer DJ (eds.) *Environments and Extinctions: Man in Late Glacial North America*, pp. 125–143. Orono, ME: Center for the Study of Early Man, University of Maine.
- Waters MR, Forman SL, Stafford TW Jr., and Foss J (2009) Geoarchaeological investigations at the Topper and Big Pine Tree sites, Allendale County, South Carolina. *Journal of Archaeological Science* 36: 1300–1311.
- Waters MR, Forman SL, Jennings TA, et al. (2011a) The Buttermilk Creek complex and the origins of Clovis at the Debra L. Friedkin site, Texas. *Science* 331: 1599–1603.
- Waters MR, Pevny CD, Carlson DL, et al. (2011b) *Clovis Lithic Technology: Investigation of a Stratified Workshop at the Gault Site, Texas*. College Station, TX: Texas A&M University Press.
- Waters MR, Stafford TW Jr., McDonald HG, et al. (2011c) Pre-Clovis mastodon hunting 13,800 years ago at the Manis site, Washington. *Science* 334: 351–353.
- Webb SD (ed.) (2006) *First Floridians and last mastodons: The Page-Ladson site in the Aucilla River*. Dordrecht, The Netherlands: Springer.
- Williams JW, Shuman BN, Webb T III, Bartlein PJ, and Leduc PL (2004) Late-Quaternary vegetation dynamics in North America: Scaling from taxa to biomes. *Ecological Monographs* 74(2): 309–334.
- Yesner DR (2007) Faunal extinction, dietary diversity, and hunter-gatherer strategies in eastern Beringia. In: Walker RB and Driskell BN (eds.) *Foragers of the Terminal Pleistocene in North America*, pp. 15–31. Omaha, NE: University of Nebraska Press.
- Yesner DR, Crossen KJ, and Easton NA (2011) Geoarchaeological and zooarchaeological correlates of early Beringian artifact assemblages: Insights from the Little John site, Yukon. In: Goebel T and Buvit I (eds.) *From the Yenisei to the Yukon: Interpreting Lithic Assemblage Variability in Late Pleistocene/Early Holocene Beringia*, pp. 308–322. College Station, TX: Texas A&M University Press.
- Zazula GD, MacKay G, Andrews TD, Shapiro B, Letts B, and Brock F (2009) A late Pleistocene steppe bison (*Bison priscus*) partial carcass from Tsiigehtchic, Northwest Territories, Canada. *Quaternary Science Reviews* 28: 2734–2742.
- Zegura SL, Karafet TM, Zhivotovsky LA, and Hammer MF (2004) High-resolution SNPs and microsatellite haplotypes point to a single, recent entry of Native American Y chromosomes into the Americas. *Molecular Biology and Evolution* 21(1): 164–175.

Human Evolution in the Quaternary

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Introduction

The last 2 My have seen the evolution and development of one of the most unique genera to have ever existed, *Homo*. Despite the presence in the fossil record of a wide range of hominin species, *Homo sapiens* is now the only living representative of this genus. It can also be considered one of the most successful, being the most populous and geographically diverse of all the hominins. Over a relatively short space of time, *Homo* has evolved from ape-like origins to become the most technologically sophisticated of all animals, and unique within the primate order in terms of its complex symbolic cognition. The evolutionary processes from which the unique human species emerged are necessarily complex and often frustratingly inscrutable. Only by studying the available fossil, lithic, faunal, and environmental evidence will it be possible to start to tease apart the hows, wheres, whys, and whens of hominin evolution and begin to understand the unique position of the genus *Homo* in the Primate order and the reasons why *H. sapiens* ultimately out-survived all other hominin species.

The article is split into four sections, covering the major evolutionary milestones in the development of the human genus. Section 'The Emergence of *Homo*' will deal with the emergence of *Homo* at the beginning of the Quaternary Period, from a landscape that was unique in containing a number of competing species, only one of which would be the ancestor of all other members of *Homo*. Section '*H. erectus* and the First Human Expansions' will discuss a key watershed in the evolution of *Homo*, the appearance of *H. erectus*, and cover the first expansions of *Homo* outside Africa, an event which appears to have taken place almost as soon as the genus first appeared. Section 'The "Transitional" Hominins' will focus on what can be considered 'transitional' hominins, those species who succeeded *H. erectus* and paved the way for later *H. sapiens*. This section will also deal with the most well known of all the hominin cousins, the Neanderthals. Finally, section 'The Appearance of *H. sapiens*' will look at the emergence of *H. sapiens* and the factors that resulted in modern humans becoming the last surviving member of the human genus. It will also look at the recent surprising discovery of *H. floresiensis*, which suggests that *H. sapiens* has not been alone for as long as first thought.

The Emergence of *Homo*

Around the start of the Quaternary Period, approximately 2Ma, it is believed that there were at least five species of hominins living in southern and eastern Africa: *Paranthropus robustus*, *P. boisei*, *Australopithecus sebidia*, *H. habilis*, and *H. rudolfensis* (and possibly also *H. ergaster*; see Figure 1; Wood and Lonergan, 2008). Although most of these species were ultimately evolutionary dead ends, one of these

species became the ancestor of all other members of the genus *Homo*. The exact identity of this ancestral species is not certain, but many believe the most likely candidate to be *H. habilis*, or the closely related *H. rudolfensis*. *H. habilis* is widely thought to be the first member of *Homo* and was first identified from material found at Olduvai Gorge, Tanzania (Leakey, 1961; Leakey et al., 1964). The discoverers of the OH 7 (Olduvai Hominid 7) specimens noted its 'modern' suite of traits when compared to *Australopithecus*, such as increased cranial capacity (estimated between 642 and 732 cc by Tobias (1964)), more gracile mandibular teeth, and modern-like morphology of the hand bones. It was this modern morphology (particularly in the hands) and the discovery of OH 7 in an archaeological context associated with Oldowan stone tools (the earliest identified hominin stone tool industry, discovered at Olduvai Gorge and named after this area) that lead to this new species being placed at the root of the human family tree. This name given to this species, *H. habilis* ('Handy Man' in Latin), highlights the importance that was placed on stone tool use and manufacture as a defining characteristic of being 'human.'

As subsequent fossil finds were assigned to this new species it became increasingly apparent that *H. habilis* contained a larger amount of variation than might be considered reasonable, typified by the differences seen in cranial size and facial morphology between skulls from the Kenyan site of Koobi Fora, KNM-ER 1470 and KNM-ER 1813 (see Holloway et al., 2004). These differences have led to the KNM-ER 1470 specimen and others being assigned to *H. rudolfensis* (Wood, 1992, 1999). The precise nature of the relationship between these two species is unclear, complicated by a general lack of post-cranial material (i.e., parts of the skeleton other than the skull) attributed to either species and a lack of consensus on the amount of variation these species should accommodate. How *H. habilis* and *H. rudolfensis* interacted with both their contemporaries and their likely successor, *H. erectus* is not known. It is likely that there would have been competition for resources within a given area and therefore pressure to find new avenues through which to exploit their environments. This is certainly suggested in the range of morphology displayed across these species, in particular in the megadont (large-toothed) *Paranthropus* species. It may also have been this resource competition that encouraged early hominins such as *H. habilis* and *H. rudolfensis* to develop a specialization for the use and manufacture of stone tool (lithic) technology, to gain an advantage which ultimately allowed them to expand into and dominate new niches. There is also evidence to suggest that climatic change in Africa between 3 and 2Ma would have further impacted on the interspecies competition (Klein, 2009).

The timings of the first appearance datum (FAD) and last appearance datum (LAD) of *H. habilis* and *H. rudolfensis* are important for understanding the ancestry of *Homo*

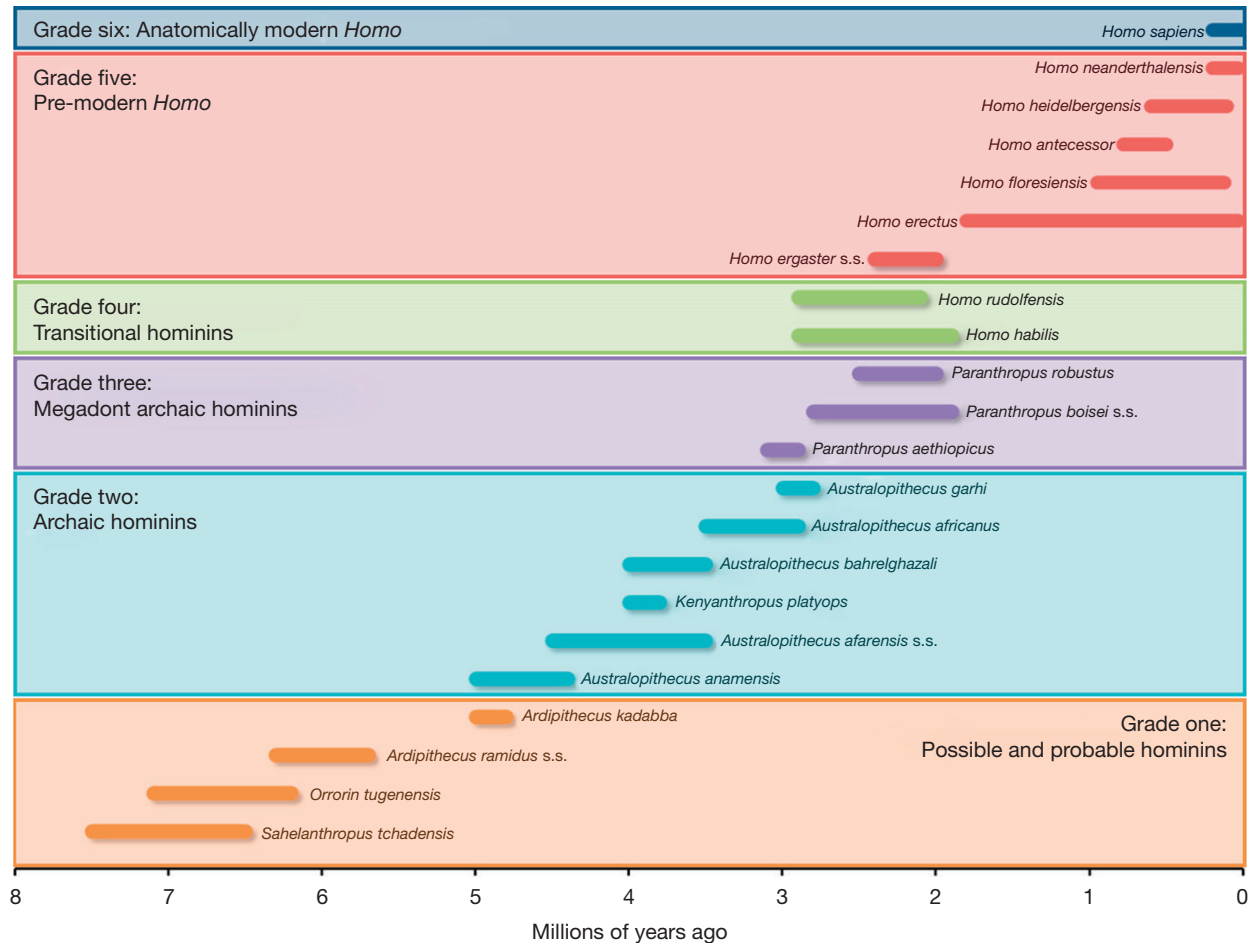


Figure 1 Timeline of hominin evolution. Based on data from Wood and Lonergan (2009). Used with permission from M. Grove.

(see Figure 1). It has been proposed by some that *H. habilis* is the most likely ancestor of *H. erectus* (e.g., Lieberman et al., 1996; Wood, 1991). If ancestry is based on cranial size, however, then the larger inferred brain size of *H. rudolfensis* places it on the path to *H. erectus*, to the exclusion of *H. habilis* (Klein, 2009). An alternative view sees both *H. habilis* and *H. rudolfensis* as a side branch of the hominin family tree, which evolved in parallel with *H. erectus*, but did not evolve into *H. erectus* (Wolpoff, 1999). This view holds that the variation shown by *H. habilis* and *H. rudolfensis* falls within that seen in the Australopithecines (Wood and Collard, 1999).

The exclusive association made by Leakey et al. (1964) between *H. habilis* and Oldowan tools has been questioned since the species was first described. The discovery of lithic artifacts dating back to around 2.6 Ma (Semaw et al., 1997) implies a much earlier emergence of *Homo* that is supported by the fossil evidence. Despite this, there has been little evidence linking *Australopithecus* or *Paranthropus* to the manufacture of these first stone tools, despite the inference that these early Oldowan tools look too 'well made' to represent the first appearance of lithic flaking technology and that the first tool makers are likely to have preceded *Homo* (Braun, 2010). Recently, animal bones bearing scratches attributed to tool use have been found in Dikika, Ethiopia and dated to at least 3.2 Ma (McPherron et al., 2010). The location of these bones, ~300 m from the site of the

discovery of a juvenile *A. afarensis* skeleton (Alemseged et al., 2006) strongly suggests a non-*Homo* manufacturer. While persuasive, the interpretation of these scratch marks as being made by tools (rather than natural causes) has been questioned (Dominguez-Rodrigo et al., 2010).

At the time of the discovery of *H. habilis*, the notion that it represented the first member of *Homo* was disputed. Some believed that the material attributed to *H. habilis* could be reasonably included inside either *Australopithecus* or *H. erectus* (Brace et al., 1973; Robinson, 1960, 1965, 1966). More recently, it has been again proposed that *H. habilis* (and *H. rudolfensis*) are not reliable species and that both should be reclassified within *Australopithecus* (Wood and Collard, 1999). Wood and Collard propose that species should only be included in *Homo* if there are found to be more similar to *Homo* than *Australopithecus* in five key regards: (1) estimated body mass, (2) body proportions, (3) a postcranial morphology adapted to obligate bipedalism with limited capacity for climbing, (4) mandibular and dental morphology, and (5) period of growth and development (i.e., childhood). Based on this definition, *H. habilis* and *H. rudolfensis* do not match the criteria for inclusion in *Homo*, and if they were to be members of the genus then *Homo* would no longer be a 'good' genus, understood by the authors to be "a monophylum whose members occupy a single adaptive zone" (Wood and Collard, 1999,

p. 70). For that reason, these species are reassigned as *A. habilis* and *A. rudolfensis*. Based on this reclassification, the earliest bona fide species of the genus *Homo* would be *H. erectus*, which can be indisputably defined as *Homo* based on Wood and Collard's criteria.

H. erectus and the First Human Expansions

The appearance of *Homo* in Africa saw an almost immediate migratory move toward new environments in Asia. This migration and expansion appears to have been carried out by a new member of the human genus, *H. erectus*. This species differed from *H. habilis* and *H. rudolfensis* in that it was the first member of *Homo* that can be considered to have a truly modern-like morphology (Wood and Lonergan, 2008) and it may have been this morphological 'modernity' that facilitated this dispersal.

The species known as *H. erectus* was first identified from fossil material found by Eugene Dubois at Trinil in Java, Indonesia toward the end of the nineteenth century. These fragmentary remains (consisting of a left femur and calvaria (skull cap)) were named *Pithecanthropus erectus*, reflecting the modern morphology of the femur ('erectus' meaning 'upright') but the primitive nature of the skull cap. Subsequent fossil finds from this region were attributed to *Pithecanthropus*, with specimens found in China attributed to *Sinanthropus pekinensis* (Klein, 2009). At the same time, as fossils of early *Homo* were being discovered in eastern Asia, similar specimens were starting to be found in Africa (Cartmill and Smith, 2009). By the 1950s, the idea had been presented that both *Pithecanthropus* and *Sinanthropus* should be subsumed into *Homo* (e.g., Mayr, 1950), and by the 1960s, the classification *H. erectus* became accepted terminology for both the Asian and African material (e.g., Howell, 1960; Howells, 1966).

Despite first being found in Asia, radiometric dating now suggests that the *H. erectus* originally emerged within Africa. Of the African specimens, those that can be definitively be assigned to *H. erectus* come from the Koobi Fora site in Kenya and are dated to around 1.8–1.7 Ma (Cartmill and Smith, 2009). Some fragmentary pelvic and occipital bone material, also from Koobi Fora, has been dated to between 2 and 1.95 Ma, but there is some debate as to whether this material can be reliably assigned to *H. erectus* (Cartmill and Smith, 2009; Klein, 2009). The dating of the Asian *H. erectus* specimens is trickier, due to the nature of the discovery of some of the earlier finds, which were collected rather than excavated, and due to the complicated stratigraphic context in which the fossils have been found (Cartmill and Smith, 2009). Addressing the debate around dating of the Javanese samples, Langbroek and Roebroeks (2000) question the older FAD of between 1.8 and 1.6 Ma proposed by Swisher et al. (1994) and reattribute the fossils from the Javanese site of Sangiran to between 1.1 Ma and 780 ka.

Claims have been made that *H. erectus* was present in China as early as 1.9 Ma, at the site of Longgupo (Huang et al., 1995). That the limited mandibular and dental remains found at this site can be attributed to *Homo* has been disputed (e.g., Wolpoff, 1999), and more reliable evidence for the first human occupations of modern-day China comes from the Nihewan Basin,

although this is based only on lithic material dating to 1.36 Ma (Zhu et al., 2001). In addition, the major site of Zhoukoudian, while challenging to date, has providing dates of between 669 and 280 ka (based on electron spin resonance) and around 400 ka (based on thermal-ionization mass spectrometry) for its fossil-bearing layers (Grün et al., 1997; Huang et al., 1991). While dating of these eastern Asia sites remains problematic, together these data support the notion that *H. erectus* first appeared in Africa before migrating toward eastern Asia (see Figure 1).

The appearance of *H. erectus* in the fossil record marks a significant advance in hominin evolution, as they are the first species that unequivocally display modern-like body proportions and a commitment to bipedalism with an upright posture (although their brain capacities were considerably less than in modern *H. sapiens*; Wood and Lonergan, 2008). Table 1 summarizes some of the key traits of *H. erectus* (both African and Asian) which distinguishes it from both earlier hominins and later *H. sapiens*.

While *H. erectus* displays a degree of expansion in cranial size, it is worth noting that much of this expansion can be explained by the clear increase in body size in this species, suggesting that *H. erectus* may have exhibited only a limited leap in cognitive capabilities, albeit a leap that appears to have prompted the development of a stone tool industry, the Acheulean (Klein, 2009).

The postcranial skeleton of African *H. erectus* is best known from the relatively complete skeleton of KNM-ER WT 15000, also known as the Nariokotome Boy (Walker and Leakey, 1993). This skeleton is thought to be from a juvenile between 10 and 15 years old at time of death (Smith, 1993), although more recently his age has been estimated at as young as 8 years old (Dean and Lucas, 2009). His morphology was found to be very modern-like and if he had reached maturity, he would have attained an adult height of 185 cm (6 ft) tall (Ruff and Walker, 1993). The long limb bone proportions and adaptations for obligate bipedalism in this individual suggest that *H. erectus* was better adapted to long-distance walking, and perhaps endurance running, than any other previous hominin (Bramble and Lieberman, 2004). By comparison, the postcranial morphology of Asian *H. erectus* is not well known, with

Table 1 Key diagnostic features of *Homo erectus* (African and Asian)

<i>Craniofacial characteristics</i>
Average cranial capacity of ~900 cm ³
Low cranial vault height, with thick walls
Continuous supraorbital torus
Occipital bun
Reduced postcanine dentition and jawbone robusticity, relative to earlier hominins, but larger than seen in later <i>Homo</i>
Reduced number of roots on maxillary premolars
More modern dental eruption schedule
Vertical shortening of face
Lack of 'chin' (compared to <i>H. sapiens</i>)
<i>Postcranial characteristics</i>
Modern-like postcranial proportions
Increased cortical bone thickness
Anterior–posterior femoral flattening, with mediolateral tibial flattening
Absence of pilaster muscle insertion on femur

Based on Cartmill and Smith (2009), Klein (2009), and Wood and Lonergan (2008).

only limited remains having been found to date. Only those remains found in Zhoukoudian can be firmly associated with *H. erectus* and their morphology appears to be similar to that seen in African *H. erectus* (Cartmill and Smith, 2009).

A degree of variation has been identified between the African and Asian material, and this has led some researchers to propose that the individuals in these regions in fact represent two separate species, *H. erectus* in Asia and *H. ergaster* in Africa (Andrews, 1984; Stringer, 1984; Wood, 1984). The reasons behind the separation of *H. ergaster* from *H. erectus* are based not only on the greater age of the African material, but also on the features of *H. ergaster* cranial morphology that are considered less derived than those seen in Asian *H. erectus*, particularly in the mandibular premolars and features of the cranial vault and base (Wood and Lonergan, 2008). The splitting of these species leaves only a small number of individuals within *H. ergaster*, dating to between 1.9 and 1.5 Ma (Klein, 2009). Such splitting has also been criticized on the basis that the level of variation seen between the fossils can reasonably be considered within-species variation (Gabunia et al., 2001), a view supported by morphological analysis (Bräuer, 1994; Rightmire, 1990, 1994; Wu et al., 2005).

The traditional view of the migration of African *H. erectus* into Asia sometime around 1 Ma has been challenged by the discovery of fossil remains, purportedly representing *H. erectus*, at the site of Dmanisi, Georgia. The collection of cranial and postcranial material recovered from the site (including four relatively complete crania), found in context with faunal material and Oldowan-type tools, has been dated to between 1.8 and 1.1 Ma, with stratigraphic and faunal analysis supporting the older end of these dates (Gabunia et al., 2000, 2001).

This early date for the Dmanisi specimens, in addition to both their 'primitive' anatomy and the absence of any hand axes from this context (suggesting an Oldowan-like, pre-Acheulean

industry), indicates a much earlier dispersal of *H. erectus* from Africa than previously considered. Definitions of *H. erectus* have also been challenged by the Dmanisi material. The four fossil skulls show very small cranial capacities, ranging between 600 cm³ for specimen D 2700 and 775 cm³ for specimen D 2280 (Gabunia et al., 2000; Vekua et al., 2002), which aligns them more closely with *H. habilis* than with *H. erectus* (Rightmire et al., 2006), which has a mean cranial capacity (for both African and Asian specimens) of 902 cm³ (Cartmill and Smith, 2009). The body size of the Dmanisi individuals, estimated from postcranial measurements, has also been found to fall within the range seen in *H. habilis*, and the degree of humeral torsion (curvature in the shape of the upper arm humerus bone) is similar to that seen in *Australopithecus* and *H. floresiensis* (Lordkipanidze et al., 2007).

Other aspects of the skull and face anatomy of the Dmanisi sample appear to share characteristics more diagnostic of *H. erectus*, for example, the presence of a 'bulge' at the bregma (bregma eminence), flexion in the occipital bone, and a thickening of the bone along the sagittal suture of the parietal bones (sagittal keeling; see Figure 2; Rightmire et al., 2006). The general limb morphology and proportions of the Dmanisi individuals also appear to be more like those of *H. erectus*, being similar to the modern human pattern (Lordkipanidze et al., 2007).

The discovery of a mandible (specimen D 2600) in 2000, with its large size and general morphology differentiating it from other Dmanisi and *H. erectus* material, has led some to designate D 2600 as belonging to a new species, named as *H. georgicus* (Gabunia et al., 2002). This taxonomic assignment has not been well supported to date, with some researchers suggesting that the morphological variation seen at Dmanisi can be incorporated within existing descriptions of *H. erectus* (or into a subspecies of *H. erectus*; Rightmire et al., 2006; Vekua et al., 2002). It is clear that the Dmanisi group has a

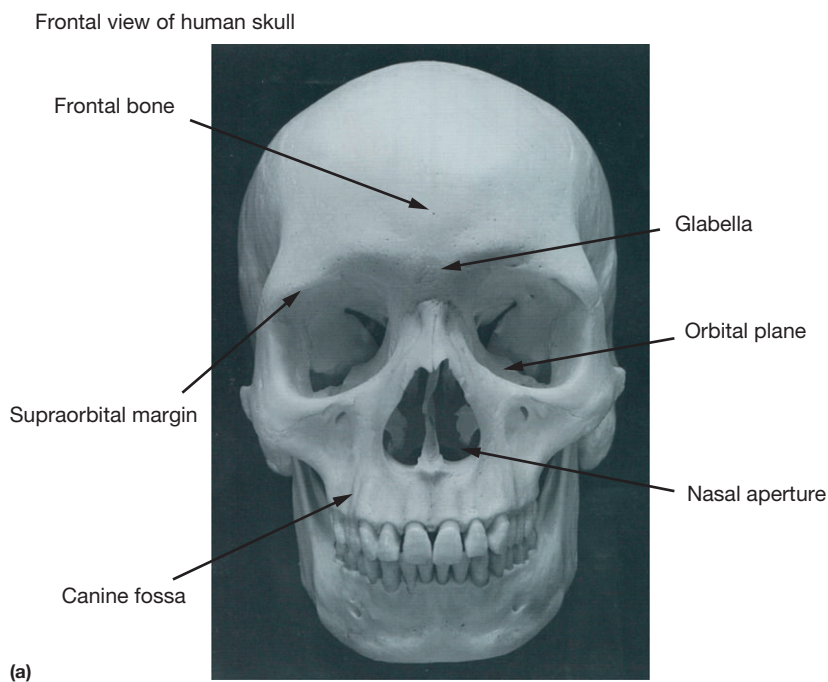


Figure 2 Osteological landmarks of the human skull, as seen in (a) frontal view, (b) lateral view, and (c) superior view.

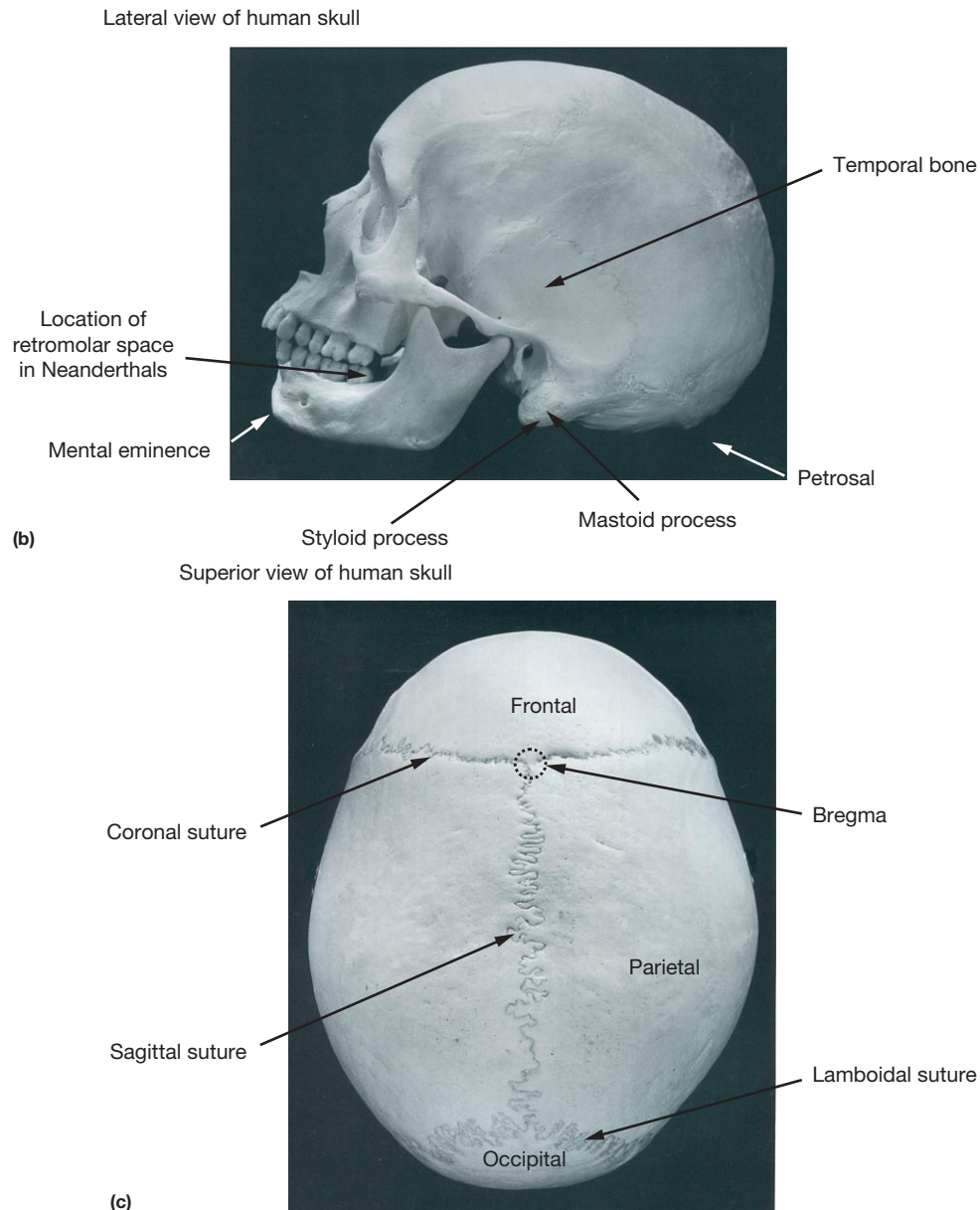


Figure 2 (Continued)

morphological mosaic unlike other contemporaneous species. Although some of these morphological features can be interpreted as symplesiomorphies (i.e., ancestral traits which are shared by two or more groups), there are also traits which appear unique to this group (Rightmire et al., 2006). Taken in combination with their age and location, this population can be interpreted as an early stage in the evolution of *H. erectus*, who left Africa before the emergence of *H. erectus* (sensu stricto; e.g., Rightmire et al., 2006; Vekua et al., 2002).

The 'Transitional' Hominins

The appearance of *H. erectus* in the fossil record around 1.8Ma and the subsequent appearance of *H. sapiens* (see

'The Appearance of *H. sapiens*') can be considered watersheds in the evolution of the human genus. Between these species, however, are found a number of 'transitional' hominins, who are crucial to the understanding of hominin evolution. The most well known (and perhaps least 'transitional') of these hominins are the Neanderthals (*H. neanderthalensis*). Less well understood are their African precursors, *H. heidelbergensis* and *H. antecessor*.

The name *H. heidelbergensis* was first assigned to a mandible discovered in the early 1900s in a quarry in the village of Mauer, near Heidelberg, Germany. The species definition, however, is based upon more complete cranial material from sites across Europe and Africa. Dating of these sites ranges between 600 and 400 ka which, in tandem with genetic evidence (e.g., Krings et al., 1997), support the notion that

H. heidelbergensis is a likely common ancestor of both *H. neanderthalensis* and *H. sapiens*. This is further supported by the anatomy of *H. heidelbergensis*, which can be considered 'transitional' between that of *H. erectus* and *H. neanderthalensis*.

Certain features of the cranium, such as a low frontal bone, large supraorbital torus (brow ridge), thick cranial vault walls, a raised and thickened ridge on the parietal bones (angular torus) and a lack of mental eminence (chin) are similar to those seen in *H. erectus* (see [Figure 2](#)). Other features, however, like the division of the supraorbital torus into medial (toward the midline of the body) and lateral (away from the midline) sections, the vertical (rather than sloping) margin of the nose, an incisive canal which runs vertically through the roof of the mouth, the general cranial base morphology, and a cranial capacity of around 1250 cm³ are more reminiscent of later *Homo* ([Rightmire, 2008](#)).

It can be seen, therefore, that it is the combination of these more primitive and more derived features that define *H. heidelbergensis* ([Rightmire, 1998](#)). It is precisely this 'transitional' morphology that can make it difficult to determine which specimens should be attributed to *H. heidelbergensis*, which has led to the species often being used as an umbrella term for individuals who are difficult to place elsewhere ([Bae, 2010](#); [Klein, 2009](#)). This has particularly been the case in the Asian fossil record, where taxonomic uncertainty has led to a number of 'transitional' hominins being assigned to *H. heidelbergensis* if they are not clearly *H. erectus* or 'archaic' *H. sapiens* ([Groves and Lahr, 1994](#); [Lahr, 1996](#); [Stringer, 2002](#)).

H. antecessor has been identified as an earlier potential common ancestor of modern humans and Neanderthals ([Bermúdez de Castro et al., 1997](#); [Carbonell et al., 2008](#)). This species is represented primarily by material found at the prolific Sierra de Atapuerca site of Gran Dolina (TD6 level) and Sima del Elefante (TD9 level) in Northern Spain and dated at around 780 ka (for TD6) ([Carbonell et al., 1995](#); [Parés and Pérez González, 1995](#)) and 1.2 Ma for TD9 ([Carbonell et al., 2008](#)). In this chronology, *H. antecessor* gave rise to both *H. heidelbergensis* in Europe (and ultimately Neanderthals) and to another species in Africa (potentially *H. rhodesiensis* or *H. helmei*) which was the forerunner of *H. sapiens*.

H. antecessor is known from a relatively large sample of cranial (see [Bermúdez de Castro et al., 1997](#); [Carbonell et al., 1995, 2008](#)) and postcranial (see [Carretero et al., 1999](#); [Lorenzo et al., 1999](#)) material. Some traits identified by Bermúdez de Castro and coauthors, such as a 'modern' mid-facial morphology with canine fossa (a depression in the facial bone above the canine teeth) align the specimens with later *H. sapiens*, whereas many aspects of the dental morphology can be considered more akin to the primitive condition exhibited by *H. erectus*/*H. ergaster*. This material is both fragmentary, however, and derived predominantly from subadult and juvenile individuals ([Cartmill and Smith, 2009](#)). This makes comparisons between this material and other specimens of adult status problematic (e.g., the calvaria from Ceprano, Italy). The taxonomic status of *H. antecessor* has been questioned ([Stringer, 2002](#)). It is problematic because it is known primarily from only one site (but see [Manzi et al. \(2001\)](#) for a description of the calvaria from Ceprano, and [Gilbert et al. \(2003\)](#) for a critique). Recently, lithic and faunal evidence from the site of Happisburgh in Norfolk, England has been

dated to at least 780 ka, but perhaps as old as 1 Ma ([Parfitt et al., 2010](#)). While no skeletal remains have been found at Happisburgh, it can be inferred that the site represents the activities of *H. antecessor*, the only hominin so far identified in Europe during this period.

Until such time as more, preferably adult, skeletal material is found that can be reliably dated and attributed to *H. antecessor*, it will remain difficult to understand the true range of variation in this species and its anatomical relationship with both *H. erectus* and *H. heidelbergensis*/*H. neanderthalensis*. This is necessary to determine the validity of *H. antecessor* as a species and its place in the fossil record as potentially the last common ancestor of Neanderthals and *H. sapiens*.

The Neanderthals are perhaps the best known of all the extinct hominin relatives, but in many ways they remain enigmatic. The history of the discovery and understanding of the species is also the story of the paleoanthropological discipline. Neanderthals have become such a part of popular human culture that their name is now used as a euphemism for primitive or uncivilized behavior. Increasingly, however, it is found that there was little in their behavior that separated them from *H. sapiens*. This understanding comes from the fossil and lithic record attributed to the Neanderthals, which is more extensive than for any other hominin species, with the unique addition of preserved DNA for some individuals. Yet despite this seeming abundance of data, there are still a large number of questions regarding their life history that remain unanswered, particularly regarding the 'modernity' of their behavior, the degree of interaction with *H. sapiens* in Europe, and the nature of their extinction.

The first Neanderthal specimen was found in 1830 in Gibraltar, although it was not recognized as such until much later. The first specimen recognized as Neanderthal was discovered in 1856 in the Neander Valley, Germany. The 'transitional' nature of preceding species like *H. heidelbergensis* make it difficult to pinpoint the exact FAD and LAD of *H. neanderthalensis*, but it is likely to be around 200 and 30 ka, respectively ([Wood and Lonergan, 2008](#); see [Figure 1](#)).

During their tenure in Europe, the Neanderthals appear to have expanded through most of the continent. There is lithic and skeletal evidence for their expansion into modern-day Israel (the sites of Skhul and Qafzeh) and Siberia (Okladnikov) to the East and into Northern Africa (the sites of Jebel Irhoud in Morocco and Haua Fteah in Libya) to the South. Lithic evidence should be interpreted with caution in the absence of associated Neanderthal skeletal remains, however, as Mousterian tools (the stone tool industry primarily associated with Neanderthals) in these Middle Eastern and African contexts have also been found with skeletal material believed to be *H. sapiens* ([Cartmill and Smith, 2009](#)).

In general terms, the cranial anatomy of Neanderthals can be distinguished from that of *H. sapiens* through examination of the cranial vault, which is much lower, longer, and wider than in *H. sapiens*, giving it a distinctive oval shape ([Hublin, 1980](#)). In addition, many European Neanderthals can be distinguished by the presence of an occipital 'bun,' a protrusion at the back of the skull ([Cartmill and Smith, 2009](#)). Interestingly, it has been found that Neanderthals had a unique inner ear bony labyrinth morphology, although the reasons for this are unclear ([Spoor et al., 2003](#)). Neanderthals are also known for

having significant midfacial prognathism (forward projection of the face), a broad supraorbital torus (brow ridges), a wide nasal aperture (nasal cavity), and a retromolar space (a gap between the last molars and the ascending arm of the jaw bone) (although this feature does not appear to be unique to Neanderthals, e.g., Trinkaus et al., 2003; see Figure 2). Certain aspects of the Neanderthal face are considered to be adaptations to living in cold environments, for example, the internal morphology of the nasal aperture (e.g., Churchill, 1998). This association has recently been challenged on the grounds that Neanderthal nasal sinus morphology does not differ significantly from *H. sapiens* when scaled for size. Alternative explanations for the shape of the Neanderthal face, such as biomechanical adaptation to masticatory stress and genetic drift have been proposed (Rae et al., 2011).

More support for cold adaptation can be found in Neanderthal postcranial morphology, particularly in their limb bone proportions. A short lower arm to upper arm ratio suggests conformation to Allen's rule, which states that animals living in cold environments develop shortened forelimbs in order to reduce their surface-to-volume ratio and therefore their heat loss (Holliday, 1997; Pearson, 2000; Ruff, 2002). Neanderthals are also known to have broad thoraxes (chest), which is interpreted as following Bergmann's rule, which is similar to Allen's rule in stating that increasing body mass reduces the surface-to-volume ratio and further conserving heat. This large thorax would also facilitate increased lung capacity which would allow Neanderthals to better deal with activity in a cold climate (Churchill, 2006). In addition, the Neanderthal skeleton is extremely robust in the size and shape of the long bones, with developed sites of muscle attachment on the bone, suggesting that the Neanderthals lived in a challenging physical environment (Pearson et al., 2006).

The discovery of ancient DNA in Neanderthal bones was a pivotal moment in the understanding of hominin evolution (Krings et al., 1997). Since that time, what has been learnt about Neanderthal genetics has proved difficult to interpret and has challenged our understanding of population genetics. It was originally believed that there was no genetic admixture between Neanderthal and contemporary modern human populations in Europe, suggesting little or no gene flow between the groups (e.g., Stringer and Andrews, 1988). More recently, this picture has been challenged, with analyses indicating that in fact as much as 4% of the modern human genome of populations outside of Africa is shared with that of Neanderthals. To obtain this level of shared variation suggests a high level of gene flow between the groups, and a level of interbreeding far in excess of what was previously assumed (Green et al., 2010; Plagnol and Wall, 2006).

This picture of Neanderthal and *H. sapiens* interaction during the Paleolithic has been further complicated by recent genetic evidence from Denisova, Siberia, which suggests that there was another, hitherto unknown, species of hominin living in this area of Eurasia between 50 and 30 ka (Reich et al., 2010). This species of hominin, known only as 'Denisovians,' is noteworthy for a number of reasons: their mitochondrial DNA (mtDNA – inherited solely through the maternal line) appears to diverge from the line leading to Neanderthals and *H. sapiens* approximately 1 Ma, twice the temporal distance posited for the mtDNA split between

Neanderthals and *H. sapiens*. The morphology of the associated skeletal remains shares no derived features that would reasonably link it to Neanderthals or *H. sapiens*, suggesting that this group continued to evolve in parallel after their split from the common Neanderthal/*H. sapiens* lineage. Reich and colleagues conclude that the Denisovian group sampled in their study share 4–6% of their genome with existing modern humans, specifically modern Melanesians. This finding clearly indicates that there was also a significant amount of gene flow between the Denisovians and *H. sapiens*, more than has been suggested for *H. sapiens* and Neanderthals, although the precise nature of this relationship and the identity of the Denisovians are currently unknown.

The Appearance of *H. sapiens*

The specter of *H. sapiens* has loomed over the discussions in the previous sections, used as a reference point for the 'modernity' or otherwise for all aspects of early *Homo* morphology and behavior. In order to be able to use modern *H. sapiens* as a lens through which to interpret other hominin species, however, one must first understand the history and development of this species, of which perhaps less is known than expected.

The understanding of the date of the first appearance of *H. sapiens* in the fossil record depends in part on how '*H. sapiens*' has been classified. Until relatively recently, it was thought that modern *H. sapiens* appeared across the Old World (i.e., Africa, Europe, and Asia) around 40 ka (e.g., Trinkaus and Smith, 1985; Wolpoff, 1980). This appearance was associated with the apparently abrupt emergence of 'modern' symbolic material culture, particularly prevalent in Europe. Since that time, this view has mostly given way to the notion that *H. sapiens* in fact first appeared in Africa as early as 200 ka. This has been supported by data on modern human genetic diversity, particularly mtDNA data, which indicate a common female ancestor of all modern groups, found in Africa ~200 ka, popularly known as 'mitochondrial Eve' (Cann et al., 1987; Ingman et al., 2000; Penny et al., 1995; Vigilant et al., 1991). The fossil record in Africa also lends support to this view, and suggests an even older FAD for *H. sapiens*, with the Florisbad (South Africa) skull being dated to around 260 ka (Grün et al., 1996). Other specimens from across Africa have been dated to more than 100 ka, from Herto (Ethiopia) at 160 ka (White et al., 2003), Jebel Irhoud (Morocco) at between 190 and 130 ka (Grün and Stringer, 1991), and Klasies River Mouth (South Africa) at about 100 ka (Rightmire and Deacon, 1991). It can be problematic to date sites of this age, which often provide only fragmentary skeletal remains, but the wealth of data that so far exists, both genetic and skeletal, supports the early appearance of *H. sapiens* in Africa.

This older date for the origins of *H. sapiens* has led to the inclusion of 'archaic' human specimens within the *H. sapiens* hypodigm (all of the fossil material of this species available for study). This has led the species to become extremely broad (when modern human variation is included), making a morphological definition of *H. sapiens* particularly challenging (Stringer, 2002). While there appears to be no formal definition of the morphology of *H. sapiens* (Schwartz and Tattersall, 2010), the species can be distinguished from its predecessors

by the presence of a two-part brow ridge with a glabellar 'butterfly' (the space in between the eyebrows and above the nose, which can protrude to resemble a butterfly) and a clear 'chin' (Schwartz and Tattersall, 2010). To this very short list can also potentially be added a number of other cranial features related to the morphology of the ectotympanic bone (structure within the middle ear that supports the eardrum), the styloid process of the temporal bone (pointed bone below the ear which anchors tongue and larynx muscles), occipital plane (outer surface of the occipital bone where the *occipitalis* muscle attaches), interdigitation (interlocking) of the cranial sutures (joints between cranial bones), the petrous (stone-like) part of the temporal bone (petrosal), subarcuate fossa (depression on the inner surface of the petrosal for the subarcuate artery), and inferior orbital plane (the inner surface of the eye orbit that is part of the maxilla (upper jaw); see Figure 2; Schwartz and Tattersall, 2010). More 'traditional' modern human traits, such as a high cranial vault with low frontal angle, a projecting mastoid process (protrusion of the temporal bone behind the ear, which is a point of attachment for several muscles), a canine fossa, an orthognathic face (i.e., located under the brain case) and absence of a retromolar space show a lot of variability in archaic and modern *Homo* and therefore highlight the difficulty in identifying those traits which are unique to *H. sapiens* but also shared by all members of the species (both archaic and modern; Figure 2) (Cartmill and Smith, 2009).

While an early origin of *H. sapiens* in Africa appears indisputable, the first appearance of 'modern' human material culture continues to be debated, due to its seemingly abrupt appearance in Europe around 40 ka. This symbolic material culture, in the form of rock art, carvings, and personal ornaments, has been assigned to the Aurignacian technological complex. The Aurignacian derives its name from the site of Aurignac in France, where the first stone tools assigned to this culture were found. This industry differs from earlier stone tool technologies (e.g., the Mousterian, also centered in Europe) by displaying a greater range of tool types, types which can be considered more 'finely crafted,' and the inclusion of standardized stone tools (Cartmill and Smith, 2009, p. 414). The abruptness of the appearance of the Aurignacian culture, the increased levels of symbolism relative to earlier technologies, and the association with modern *H. sapiens* archaeological contexts led to the period from 40 ka onward to be referred to as the 'Upper Paleolithic Revolution' (e.g., Bar-Yosef, 2002).

The notion of a European cultural 'revolution' was once prevalent in the understanding of modern human origins, but has been challenged on a number of fronts. The 'Eurocentric' nature of the 'Upper Paleolithic Revolution,' with its focus on a European origin for cultural modernity has been criticized by some (e.g., McBready and Brooks, 2000) for overlooking evidence for cultural sophistication in the African archaeological record dating back to 100 ka and developing complexity over time. For example, perforated shell beads and red ochre engraved with geometric patterns, found in the Blombos Caves in South Africa, have been dated to 75 ka and represent some of the earliest evidence for personal adornment and abstract image creation (d'Errico et al., 2005; Henshilwood et al., 2002).

Another area of debate has been the mechanism by which *H. sapiens* came to be so widely dispersed. Currently, the most

widely held viewpoint regarding the dispersal of humans through the Old World is the 'Out of Africa hypothesis' (or Recent African Origin hypothesis; e.g., Stringer and Andrews, 1988; Stringer et al., 1984). This states that *H. sapiens* evolved exclusively within Africa between 200 and 100 ka and that all humans alive today ultimately descend from these individuals, with the groups who left Africa replacing the last remnants of *H. erectus* in eastern Asia and Neanderthals in Europe. An opposing view, known as the 'multiregional hypothesis' (e.g., Wolpoff et al., 1984, 1988) proposes the concept of regional continuity, with *H. erectus* leaving Africa and evolving into *H. sapiens* evolving *in situ* across the world. This hypothesis sees the species definition of *H. sapiens* expanded to include species such as *H. erectus* and *H. neanderthalensis* and that a variety of factors such as local adaptation, interbreeding, genetic selection and drift resulted in local evolution that converged on the regional variation seen in *H. sapiens* today. An extreme interpretation of the Out of Africa hypothesis does not allow for any interbreeding or gene flow between *H. sapiens* and the populations they replaced. This view has been challenged by genetic analyses which suggest at least a certain amount of admixture between *H. sapiens* and Neanderthals (Green et al., 2010; Plagnol and Wall, 2006) and therefore the true story of evolution outside Africa is likely to be a less extreme form of the Out of Africa hypothesis, where some amount of gene flow took place between species.

What is becoming increasingly clear, particularly in light of the increasing amount of ancient DNA evidence that is becoming available, is that the story of *H. sapiens* migration across the Old World (but also into the New World) is incredibly complex. Recent DNA analysis of hominin material from Denisova, Siberia (Reich et al., 2010) has shown that archaeological evidence alone will not be able to explain everything that is needed to be known about the history of human expansion. A more comprehensive view of modern and ancient human population genetics, in combination with the rich anatomical and archaeological records from these periods, is ultimately the best hope for understanding the true nature of the human species' history and development.

It has long been assumed that, following the extinction of *H. erectus* in Asia and *H. neanderthalensis* in Europe by 30 ka (Wood and Lonergan, 2008), *H. sapiens* was left as the only surviving member of the genus *Homo*. This accepted picture was turned on its head with the discovery of skeletal and lithic material dating to between 95 and 18 ka on the Indonesian island of Flores (Brown et al., 2004; Morwood et al., 2005). What was revolutionary about these finds was the distinctive, and unexpected, morphology of the Flores individuals, which appeared to be a combination of primitive (e.g., small cranial capacity, wrist morphology) and more modern (e.g., obligate bipedalism, reduced dentition size and facial prognathism) traits which are unique to these individuals (Brown et al., 2004). The Flores individuals (and the LB1 holotype in particular) also display some unexpected traits, such as a very short stature and unusual limb proportions, represented in LB1 having very long feet relative to its leg length (Jungers et al., 2009).

Such was the uniqueness of the skeletal remains that they were assigned to a new species, *H. floresiensis*. Unfortunately, the preservation of DNA material in the *H. floresiensis* bones was poor, making phylogenetic analysis impossible. The

morphology of *H. floresiensis*, in combination with what are considered relatively advanced stone tools (Morwood et al., 2005), makes an understanding of the relationship of this species to other hominin species challenging. Fierce, and often acrimonious, debate has raged between researchers since the Flores material was first published. A range of species assignments have been made, but they can be split into two broad groups: those who believe that *H. floresiensis* is a derived descendent of early *Homo* and those who believe that the Flores individuals are pathological specimens of modern *H. sapiens*.

Those who believe that *H. floresiensis* is a new species derived from early *Homo* base their view on the primitive morphology. The morphology of the wrist and many aspects of the crania have suggested to researchers that the most likely ancestral candidate is *H. erectus* (Wood and Lonergan, 2008). If this is the case, this allows one to draw certain conclusions about the technological sophistication of the species that first arrived on the island. Even at the lowest sea levels, the island of Flores would have been separated from the nearest land mass by about 20 km of water; therefore reaching Flores would have involved a significant water crossing. While it is known that some form of water-crossing technology was necessary for *Homo* to reach the Australian landmass, around 60 ka (Thorne et al., 1999), the discoveries on Flores suggest a much earlier development of this behavior.

Some researchers have suggested an even more ancient origin for *H. floresiensis*, believing them to have descended possibly from *Australopithecus* or another early hominin species (Argue et al., 2006; Falk et al., 2005; Jungers et al., 2009). While further data are necessary to determine the veracity of this hypothesis, it forces one to consider the potential technological sophistication and capabilities of the earliest hominins.

Critics of the *H. floresiensis* concept believe that the material on Flores represents pathological examples of dwarf modern *H. sapiens*. The most commonly proposed idea is that the LB1 cranium is from an individual suffering from a form of microcephaly, which results from compromised neural development and leads to a small brain size relative to face size (Henneberg and Thorne, 2004; Jacob et al., 2006; Martin et al., 2006; Weber et al., 2005). Modern humans with this condition often suffer from cognitive impairments and a certain amount of dwarfism.

This view has been strongly criticized, on the grounds that LB1's suite of features is primitive and cannot be explained by any developmental disorder (Argue et al., 2006; Falk et al., 2005, 2006, 2007; Larson et al., 2007; Tocheri et al., 2007). Another blow to this hypothesis came when additional skulls were found in the Liang Bua excavations (e.g., LB6) – all exhibiting small cranial capacity. Microcephaly is a rare genetic abnormality in humans, so the odds that multiple individuals from a small group all exhibited this same genetic disorder seem quite small. Undeterred by this, other possible disorders have been suggested to explain the morphology of *H. floresiensis*, including Laron Syndrome (Hershkovitz et al., 2007) and myxoedematous endemic cretinism (Obendorf et al., 2008), again strongly refuted (Falk et al., 2009).

With the data currently available, the most parsimonious explanation for the Flores material seems to be that they represent some descendent of an earlier species of *Homo* that underwent island dwarfism once reaching Flores. Such is the unique and incomparable nature of these individuals; however, it is unlikely that the disagreements will be settled until

more undisputed material is found. Whatever future evidence is obtained, it will no doubt shed light on one of the most intriguing issues in paleoanthropology and force the reevaluation of the key aspects of hominin evolution such as brain expansion and cognitive organization and complexity.

Conclusions

The complexity and scope of human evolution cannot be done justice in so few words, but what this article has attempted to show is that, while the human genus has been subject to the same selection pressures as other animals, it is how hominins have responded to those pressures that has ultimately made the human genus distinct. It is the unique behavioral plasticity of *Homo* that has allowed hominins to respond to environmental pressures through the development of technologies. This encouraged a more diverse exploitation of resources, and ultimately led to the development of symbolic material culture. Significant changes in the locomotor repertoire of humans has facilitated an expansion of the geographic range of the human genus that is unknown in any other primate, thus taking *Homo* into new and challenging environments, further promoting adaptability of human beings. What processes led to the behavioral flexibility and adaptability of humans are unclear, and given the somewhat incomplete fossil record, it is understandable that passionate debate on these questions continues and contentious issues remain. Ongoing research will enable more pieces to be added to this fascinating evolutionary jigsaw and increasingly shed more light on the enigmatic origins of this unique primate genus.

See also: [Archaeological Records: 1.9 Myr–300 000 Years Ago in Europe](#); [2.7 Myr–300 000 Years Ago in Africa](#); [2.7 Myr–300 000 Years Ago in Asia](#); [Global Expansion 300 000–8000 Years Ago, Africa](#); [Global Expansion 300 000–8000 Years Ago, Americas](#); [Global Expansion 300 000–8000 Years Ago, Asia](#); [Global Expansion 300 000–8000 Years Ago, Australia](#); [Neanderthal Demise](#); [Overview](#); [Postglacial Adaptations](#).

References

- Alemseged Z, Spoor F, Kimbel WH, et al. (2006) A juvenile early hominin skeleton from Dikika, Ethiopia. *Nature* 443: 296–301.
- Andrews P (1984) An alternative explanation of the characters used to define *Homo erectus*. *Courier Forschungsinstitut Senckenberg* 69: 167–178.
- Argue D, Donlon D, Groves C, and Wright R (2006) *Homo floresiensis*: Microcephalic, pygmoid, *Australopithecus*, or *Homo*? *Journal of Human Evolution* 51: 360–374.
- Bae CJ (2010) The Late Middle Pleistocene hominin fossil record of Eastern Asia: Synthesis and review. *Yearbook of Physical Anthropology* 53: 75–93.
- Bar-Yosef O (2002) The Upper Paleolithic revolution. *Annual Review of Anthropology* 31: 363–393.
- Bermúdez de Castro JM, Arsuaga JL, Carbonell E, Rosas A, Martínez I, and Mosquera M (1997) A hominid from the Lower Pleistocene of Atapuerca, Spain: Possible ancestor to Neanderthals and modern humans. *Science* 276: 1392–1395.
- Brace CL, Mahler P, and Rosen R (1973) Tooth measurements and the rejection of the taxon '*Homo habilis*'. *Yearbook of Physical Anthropology* 16: 50–68.
- Bramble DM and Lieberman DE (2004) Endurance running and the evolution of *Homo*. *Nature* 432: 345–352.

- Bräuer G (1994) How different are Asian and African *Homo erectus*? In: Franzen J (ed.) 100 Years of *Pithecanthropus*. The *Homo erectus* Problem. *Courier Forschungsinstitut Senckenberg* 38–46.
- Braun DR (2010) Australopithecine butchers. *Nature* 466: 828.
- Brown P, Sutikna T, Morwood MJ, et al. (2004) A new small-bodied hominin from the Late Pleistocene of Flores, Indonesia. *Nature* 431: 1055–1061.
- Cann RL, Stoneking M, and Wilson AC (1987) Mitochondrial DNA and human evolution. *Nature* 325: 31–36.
- Carbonell E, Bermúdez de Castro JM, Arsuaga J, et al. (1995) Lower Pleistocene hominids and artefacts from Atapuerca-TD6 (Spain). *Science* 269: 826–830.
- Carbonell E, Bermúdez de Castro JM, Parés JM, et al. (2008) The first hominin of Europe. *Nature* 452: 465–469.
- Carretero JM, Lorenzo C, and Arsuaga JL (1999) Axial and appendicular skeleton of *Homo antecessor*. *Journal of Human Evolution* 37: 459–499.
- Cartmill M and Smith FH (2009) *The Human Lineage*. Hoboken, NJ: Wiley.
- Churchill SE (1998) Cold adaptation, heterochrony and Neandertals. *Evolutionary Anthropology* 7: 46–60.
- Churchill S (2006) Bioenergetic perspectives on Neanderthal thermoregulatory and activity budgets. In: Harvati K and Harrison T (eds.) *Neanderthals Revisited: New Approaches and Perspectives*, pp. 113–133. Dordrecht: Springer.
- d'Errico F, Henshilwood C, Vanhaeren M, and van Niekerk K (2005) *Nassarius kraussianus* shell beads from Blombos Cave: Evidence for symbolic behaviour in the Middle Stone Age. *Journal of Human Evolution* 48: 3–24.
- Dean MC and Lucas VS (2009) Dental and skeletal growth in early fossil hominins. *Annals of Human Biology* 36: 545–561.
- Dominguez-Rodrigo M, Pickering TR, and Bunn HT (2010) Configurational approach to identifying the earliest hominin butchers. *Proceedings of the National Academy of Sciences of the United States of America* 1–6, 107: 20929–20934.
- Falk D, Hildebolt C, Smith K, et al. (2005) The brain of LB1, *Homo floresiensis*. *Science* 308: 242–245.
- Falk D, Hildebolt C, Smith K, et al. (2006) Response to comment on 'The Brain of LB1, *Homo floresiensis*'. *Science* 312: 999c.
- Falk D, Hildebolt C, Smith K, et al. (2007) Brain shape in human microcephalics and *Homo floresiensis*. *Proceedings of the National Academy of Sciences of the United States of America* 104: 2513–2518.
- Falk D, Hildebolt C, Smith K, et al. (2009) The type specimen (LB1) of *Homo floresiensis* did not have Laron syndrome. *American Journal of Physical Anthropology* 140: 52–63.
- Gabounia L, de Lumley MA, Vekua A, Lordkipanidze D, and de Lumley H (2002) Discovery of a new hominid at Dmanisi (Transcaucasia, Georgia). *Comptes Rendus Palevol* 1: 243–253.
- Gabunia L, Antón SC, Lordkipanidze D, Vekua A, Justus A, and Swisher CC (2001) Dmanisi and dispersal. *Evolutionary Anthropology* 10: 158–170.
- Gabunia L, Vekua A, Lordkipanidze D, et al. (2000) Earliest Pleistocene hominid cranial remains from Dmanisi, Republic of Georgia: Taxonomy, geological setting, and age. *Science* 288: 1019–1025.
- Gilbert W, White T, and Asfaw B (2003) *Homo erectus*, *Homo ergaster*, *Homo 'cepranensis'*, and the Daka cranium. *Journal of Human Evolution* 45: 255–259.
- Green RE, Krause J, Briggs AW, et al. (2010) A draft sequence of the Neanderthal genome. *Science* 328: 710–722.
- Groves CP and Lahr MM (1994) A bush not a ladder: Speciation and replacement in human evolution. *Perspectives in Human Biology* 4: 1–11.
- Grün R, Brink JS, Spooner NA, et al. (1996) Direct dating of Florisbad hominid. *Nature* 382: 500–501.
- Grün R, Huang PH, Wu X, Stringer CB, Thorne AG, and McCulloch M (1997) ESR analysis of teeth from the palaeoanthropological site of Zhoukoudian, China. *Journal of Human Evolution* 32: 83–91.
- Grün R and Stringer CB (1991) Electron spin resonance dating and the evolution of modern humans. *Archaeometry* 33: 153–199.
- Henneberg M and Thorne A (2004) Flores human may be pathological *Homo sapiens*. *Before Farming* 4: 2–4.
- Henshilwood C, D'Errico F, Yates R, et al. (2002) Emergence of modern human behavior: Middle Stone Age engravings from South Africa. *Science* 295: 1278–1279.
- Hershkovitz I, Kornreich L, and Laron Z (2007) Comparative skeletal features between *Homo floresiensis* and patients with primary growth hormone insensitivity (Laron syndrome). *American Journal of Physical Anthropology* 134: 198–208.
- Holliday TW (1997) Postcranial evidence of cold adaptation in European Neanderthals. *American Journal of Physical Anthropology* 104: 245–258.
- Holloway R, Broadfield D, and Yuan M (2004) *The Human Fossil Record. Brain Endocasts – The Paleoneurological Evidence*, vol. 3. New York: Wiley-Liss.
- Howell FC (1960) European and northwest African Middle Pleistocene hominids. *Current Anthropology* 1: 195–232.
- Howells WW (1966) *Homo erectus*. *Scientific American* 215: 46–53.
- Huang W, Ciochón R, Gu Y, et al. (1995) Early *Homo* and associated artefacts from Asia. *Nature* 378: 275–278.
- Huang P, Jin S, Liang R, et al. (1991) Study of ESR dating for burying age of the first skull of Peking Man and chronological scale of the cave deposit in Zhoukoudian site. *Acta Anthropologica Sinica* 10: 107–115.
- Hublin J-J (1980) La Chaise Suard, Engis 2 et La Quina H 18: développement de la morphologie occipitale externe chez l'enfant préneandertalien et néandertalien. *Comptes Rendus de l'Académie des Sciences Paris* 291: 669–672.
- Ingman M, Kaessmann H, Pääbo S, and Gyllenstein U (2000) Mitochondrial genome variation and the origin of modern humans. *Nature* 408: 708–713.
- Jacob T, Indriati E, Soejono RP, et al. (2006) Pygmoid Australomelanesian *Homo sapiens* skeletal remains from Liang Bua, Flores: Population affinities and pathological abnormalities. *Proceedings of the National Academy of Sciences of the United States of America* 103: 13421–13426.
- Jungers WL, Harcourt-Smith WEH, Wunderlich RE, et al. (2009) The foot of *Homo floresiensis*. *Nature* 459: 81–84.
- Klein RG (2009) *The Human Career: Human Biological and Cultural Origins*, 3rd edn. Chicago: University of Chicago Press.
- Krings M, Stone A, Schmitz RW, Krainitzki H, Stoneking M, and Pääbo S (1997) Neandertal DNA sequences and the origin of modern humans. *Cell* 90: 19–30.
- Lahr MM (1996) *The Evolution of Modern Human Diversity: A Study of Cranial Variation*. Cambridge: Cambridge University Press.
- Langbroek M and Roebroeks W (2000) Extraterrestrial evidence on the age of the hominids from Java. *Journal of Human Evolution* 38: 595–600.
- Larson SG, Jungers WL, Morwood MJ, et al. (2007) *Homo floresiensis* and the evolution of the hominin shoulder. *Journal of Human Evolution* 53: 718–731.
- Leakey L (1961) New finds at Olduvai Gorge. *Nature* 189: 649–650.
- Leakey LSB, Tobias PV, and Napier J (1964) A new species of the genus *Homo* from Olduvai Gorge. *Nature* 202: 7–9.
- Lieberman D, Wood B, and Pilbeam D (1996) Homoplasy and early *Homo*: An analysis of the evolutionary relationships of *H. habilis* sensu stricto and *H. rudolfensis*. *Journal of Human Evolution* 30: 97–120.
- Lordkipanidze D, Jashashvili T, Vekua A, et al. (2007) Postcranial evidence from early *Homo* from Dmanisi, Georgia. *Nature* 449: 305–310.
- Lorenzo C, Arsuaga JL, and Carretero JM (1999) Hand and foot remains from the Gran Dolina Early Pleistocene site (Sierra de Atapuerca, Spain). *Journal of Human Evolution* 37: 501–522.
- Manzi G, Mallegni F, and Ascenzi A (2001) A cranium for the earliest Europeans: Phylogenetic position of the hominid from Ceprano, Italy. *Proceedings of the National Academy of Sciences of the United States of America* 98: 10011–10016.
- Martin RD, MacLarnon AM, Phillips JL, Dussubieux L, Williams PR, and Dobyns WB (2006) Comment on 'The Brain of LB1, *Homo floresiensis*'. *Science* 312: 999b.
- Mayr E (1950) Taxonomic categories in fossil hominids. *Cold Spring Harbor Symposia on Quantitative Biology* 15: 109–118.
- McBrearty S and Brooks AS (2000) The revolution that wasn't: A new interpretation of the origin of modern human behavior. *Journal of Human Evolution* 39: 453–563.
- McPherron SP, Alemseged Z, Marean CW, et al. (2010) Evidence for stone-tool-assisted consumption of animal tissues before 3.39 million years ago at Dikika, Ethiopia. *Nature* 466: 857–860.
- Morwood MJ, Brown P, Jatmiko W, et al. (2005) Further evidence for small-bodied hominins from the Late Pleistocene of Flores, Indonesia. *Nature* 437: 1012–1017.
- Obendorf PJ, Oxnard CE, and Kefford BJ (2008) Are the small human-like fossils found on Flores human endemic cretins? *Proceedings of the Royal Society B* 275: 1287–1296.
- Parés J and Pérez González A (1995) Paleomagnetic age for hominid fossils at Atapuerca archaeological site, Spain. *Science* 269: 830–832.
- Parfitt SA, Ashton NM, Lewis SG, et al. (2010) Early Pleistocene human occupation at the edge of the boreal zone in northwest Europe. *Nature* 466: 229–233.
- Pearson OM (2000) Postcranial remains and the origin of modern humans. *Evolutionary Anthropology* 9: 229–247.
- Pearson OM, Cordero RM, and Busby AM (2006) How different were Neanderthals' habitual activities? A comparative analysis with diverse groups of recent humans. In: Harvati K and Harrison T (eds.) *Neanderthals Revisited: New Approaches and Perspectives*, pp. 135–156. Dordrecht: Springer.
- Penny D, Steel M, Waddell PJ, and Hendy MD (1995) Improved analyses of human mtDNA sequences support a recent African origin for *Homo sapiens*. *Molecular Biology and Evolution* 12: 863–882.
- Plagnol V and Wall JD (2006) Possible ancestral structure in human populations. *PLoS Genetics* 2: e105.

- Rae TC, Koppe T, and Stringer CB (2011) The Neanderthal face is not cold adapted. *Journal of Human Evolution* 60: 234–239.
- Reich D, Green RE, Kircher M, et al. (2010) Genetic history of an archaic hominin group from Denisova Cave in Siberia. *Nature* 468: 1053–1060.
- Rightmire GP (1990) *The Evolution of Homo erectus*. Cambridge: Cambridge University Press.
- Rightmire GP (1994) The relationship of *Homo erectus* to later Middle Pleistocene hominids. In: Franzen J (ed.) *100 Years of Pithecanthropus: The Homo erectus Problem* Courier Forschungsinstitut Senckenberg 171: 1–361.
- Rightmire GP (1998) Human evolution in the Middle Pleistocene: The role of *Homo heidelbergensis*. *Evolutionary Anthropology* 6: 218–227.
- Rightmire GP (2008) *Homo* in the Middle Pleistocene: Hypodigms, variation, and species recognition. *Evolutionary Anthropology* 17: 8–21.
- Rightmire GP and Deacon H (1991) Comparative studies of Late Pleistocene human remains from Klasies River Mouth, South Africa. *Journal of Human Evolution* 20: 131–156.
- Rightmire GP, Lordkipanidze D, and Vekua A (2006) Anatomical descriptions, comparative studies and evolutionary significance of the hominin skulls from Dmanisi, Republic of Georgia. *Journal of Human Evolution* 50: 115–141.
- Robinson JT (1960) The affinities of the new Olduvai australopithecine. *Nature* 186: 456–458.
- Robinson JT (1965) *Homo 'habilis'* and the australopithecines. *Nature* 205: 121–124.
- Robinson JT (1966) The distinctiveness of *Homo habilis*. *Nature* 209: 957–960.
- Ruff C (2002) Variation in human body size and shape. *Annual Review of Anthropology* 31: 211–232.
- Ruff CB and Walker A (1993) Body size and body shape. In: Walker A and Leakey RE (eds.) *The Nariokotome Homo erectus Skeleton*, pp. 234–265. Cambridge, MA: Harvard University Press.
- Schwartz JH and Tattersall I (2010) Fossil evidence for the origin of *Homo sapiens*. *Yearbook of Physical Anthropology* 53: 94–121.
- Semaw S, Renne P, Harris JWK, et al. (1997) 2.5-million-year-old stone tools from Gona, Ethiopia. *Nature* 385: 333–336.
- Smith BH (1993) The physiological age of KNM-WT 15000. In: Walker A and Leakey R (eds.) *The Nariokotome Homo erectus Skeleton*, pp. 195–220. Cambridge, MA: Harvard University Press.
- Spoor F, Hublin J-J, Braun M, and Zonneveld F (2003) The bony labyrinth of Neanderthals. *Journal of Human Evolution* 44: 141–165.
- Stringer CB (1984) The definition of *Homo erectus* and the existence of the species in Africa and Europe. *Courier Forschungsinstitut Senckenberg* 69: 131–143.
- Stringer CB (2002) Modern human origins: Progress and prospects. *Philosophical Transactions of the Royal Society B* 357: 563–579.
- Stringer CB and Andrews P (1988) Genetic and fossil evidence for the origin of modern humans. *Science* 239: 1263–1268.
- Stringer CB, Hublin J-J, and Vandermeersch B (1984) The origin of anatomically modern humans in Western Europe. In: Smith F and Spencer F (eds.) *The Origins of Modern Humans: A World Survey of the Fossil Evidence*, pp. 51–135. New York: Liss.
- Swisher C, Curtis G, Jacob T, Getty A, and Suprijo A (1994) Age of the earliest known hominids in Java, Indonesia. *Science* 263: 1118–1121.
- Thorne A, Grün R, Mortimer G, et al. (1999) Australia's oldest human remains: Age of the Lake Mungo 3 skeleton. *Journal of Human Evolution* 36: 591–612.
- Tobias PV (1964) The Olduvai Bed I hominine with special reference to its cranial capacity. *Nature* 202: 3–4.
- Tocheri MW, Orr CM, Larson SG, et al. (2007) The primitive wrist of *Homo floresiensis* and its implications for hominin evolution. *Science* 317: 1743–1745.
- Trinkaus E, Moldovan O, Milota S, et al. (2003) An early modern human from the Peștera cu Oase, Romania. *Proceedings of the National Academy of Sciences of the United States of America* 100: 11231–11236.
- Trinkaus E and Smith F (1985) The fate of the Neanderthals. In: Delson E (ed.) *Ancestors: The Hard Evidence*, pp. 325–333. New York: Liss.
- Vekua A, Lordkipanidze D, Rightmire GP, et al. (2002) A new skull of early *Homo* from Dmanisi, Georgia. *Science* 297: 85–89.
- Vigilant L, Stoneking M, Harpending H, Hawkes K, and Wilson AC (1991) African populations and the evolution of human mitochondrial DNA. *Science* 253: 1503–1507.
- Walker A and Leakey R (1993) The postcranial bones. In: Walker A and Leakey R (eds.) *The Nariokotome Homo erectus Skeleton*, pp. 95–160. Cambridge: Harvard University Press.
- Weber J, Czarnetzki A, and Pusch C (2005) Comment on: The brain of LB1, *Homo floresiensis*. *Science* 310: 236.
- White TD, Asfaw B, DeGusta D, et al. (2003) Pleistocene *Homo sapiens* from Middle Awash, Ethiopia. *Nature* 423: 742–747.
- Wolpoff M (1980) Cranial remains of Middle Pleistocene European hominids. *Journal of Human Evolution* 9: 339–358.
- Wolpoff M (1999) *Paleoanthropology*, 2nd edn. New York: McGraw-Hill.
- Wolpoff MH, Spuhler JN, Smith FH, et al. (1988) Modern human origins. *Science* 241: 772–774.
- Wolpoff M, Wu X, and Thorne A (1984) Modern *Homo sapiens* origins: A general theory of hominid evolution involving the fossil evidence from East Asia. In: Smith F and Spencer F (eds.) *The Origins of Modern Humans: A World Survey of the Fossil Evidence*, pp. 411–483. New York: Liss.
- Wood B (1984) The origin of *Homo erectus*. *Courier Forschungsinstitut Senckenberg* 69: 99–111.
- Wood B (1991) *Koobi Fora Research Project. Hominid Cranial Remains*, vol. 4. Oxford: Clarendon Press.
- Wood B (1992) Origin and evolution of the genus *Homo*. *Nature* 355: 783–790.
- Wood B (1999) *Homo rudolfensis* Alexeev, 1986 – Fact or phantom? *Journal of Human Evolution* 36: 115–118.
- Wood B and Collard M (1999) The human genus. *Science* 284: 65–71.
- Wood B and Loneragan N (2008) The hominin fossil record: Taxa, grades and clades. *Journal of Anatomy* 212: 354–376.
- Wu L, Zhang Y, and Wu X (2005) Middle Pleistocene human cranium from Tangshan (Nanjing), southeast China: A new reconstruction and comparisons with *Homo erectus* from Eurasia and Africa. *American Journal of Physical Anthropology* 127: 253–262.
- Zhu RX, Hoffman KA, Potts R, et al. (2001) Earliest presence of humans in northeast Asia. *Nature* 413: 413–417.

Relevant Websites

- <http://johnhawks.net/weblog> – A comprehensive blog dealing with current paleoanthropological issues, written from an academic perspective by Dr John Hawks at the University of Wisconsin.
- <http://www.archaeologyinfo.com/glossary.htm> – A useful online glossary of commonly used osteological, paleanthropological and archaeological terms.
- <http://humanorigins.si.edu/> – A website run by the Smithsonian National Museum of Natural History that provides multimedia information resources and up-to-date news on human origins issues.
- <http://www.eskeletons.org/> – An interactive website, developed by the University of Texas, that facilitates learning of human and primate anatomy through their digital osteological database.

Neanderthal Demise

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Introduction

The extinction of the Neanderthals (*Homo neanderthalensis*), and the arrival of modern humans (*Homo sapiens*) in Europe between 45 and 30 thousand years ago ('ka BP') is one of the most hotly debated topics in Paleolithic archaeology and paleoanthropology. Why? What is the evolutionary and behavioral importance of this transition?

Most researchers in this period of human evolution incline more toward a single-origin model for the evolution of modern humans, with a subsequent expansion out of Africa into Asia, Greater Australasia, and Europe, ultimately replacing the Neanderthals in the latter subcontinent (e.g., Mellars (1989), Stringer (1989), and d'Errico et al. (1998)). However, a substantial minority of researchers believe the multiregional origin hypothesis to be more appropriate, with modern humans evolving independently in many regions of the Old World, yet sufficiently connected to each other to facilitate the exchange of genes, allowing *Homo sapiens* to develop at a global scale as a single species for the last two million years (e.g., Wolpoff and Caspari (1997)). The models have distinctive temporal and spatial scales: single-origin models tend to think in terms of speciation locations (with subsequent dispersals) and events, while multiregional ones think in terms of global population interactions and broad (sympatric) speciation processes. Both models can conceive of population extinctions and replacements, although these are more geographically restricted if one is a multiregionalist (Wolpoff and Caspari, 1997). Because this author has difficulties in believing that multiregionalism can operate at the large spatio-temporal scales proposed by its adherents, the single-origin ('Out-of-Africa 2') model will be given stronger preference for much of the discussion in this article.

Species Definitions

Neanderthals

Although specimens had been discovered in 1830 (Engis, Belgium) and 1848 (Gibraltar), it was not until a third one was recovered from the Neander valley ('Neanderthal') in Germany in 1856, that species-specific traits were identified (King, 1864) and a new hominid species was classified, indeed the first human species to be distinguished since the publication of Darwin's *On the Origin of Species* in 1859. Neanderthals were notable for their stocky build, but it is their cranial characteristics that are the most diagnostic: a protruding mid-facial region, emphasized by the lack of a distinct chin, and thick, well-developed brow-ridges over the eyes, which grade backward into a weakly developed forehead. Seen from above, it is as if someone has taken the modern human skull shape we are used to, and stretched it longitudinally. Nevertheless, the

cranial capacity (brain size) of Neanderthals is not notably different from that of modern humans, averaging some 1400 cc. Perhaps because of our simplistic views of what characterized the climatic conditions of Ice Ages, it has often been claimed that Neanderthals were adapted to exceptionally cold ('hyperarctic') conditions (e.g., Holliday (1997)). However, if Neanderthals preferred cold conditions, then how can we explain their presence in the Near East or Iberia: two regions not known for their climatic affinity with the Arctic (see Aiello and Wheeler (2003))?

Two enlightenment-derived perceptions of Neanderthals have tended to dominate, both crystallized from mythology: men-as-monsters and men-as-gods. An example of the former is to see Neanderthals as something akin to Shakespeare's Caliban: a distillation of all that is monstrous in humans; an example of the latter is surely the concept of the 'Noble Savage.' Presumably owing to their 'coarse' physical features, Neanderthals have tended to be seen as more monstrous than noble, but recently the pendulum has begun to swing the other way (e.g., d'Errico et al. (1998)). It will be seen, however, that such Cartesian dualities are difficult to test or falsify against evidence that is largely ambiguous.

The origins and extent of Neanderthal symbolic behavior is one of the most problematic debates in Paleolithic archaeology, and is dependent upon the vagaries and chronological precision of the extant archaeological record. It was frequently thought that Neanderthals were incapable of complex spoken language, not just on the grounds of mental ability but also of physical capability. The claims of, among others, Lieberman (1989), that the configuration and morphology of the Neanderthal vocal tract ensured that they could not vocalize like ourselves, have been largely discredited by the morphological similarity of the hyoid bone from a Near Eastern Neanderthal (at Kebara, Israel) to those of modern humans (Arensburg, 1989).

Modern Humans

Homo sapiens are one of the most poorly defined species in the animal kingdom, with relatively few diagnostic physiological features. In comparison to Neanderthals, we have a much flatter mid-facial region, with a pronounced chin and steep forehead (with weak brow-ridges). The elongated body shapes of many modern humans, today and in the past, are congruent with our proposed origins in the tropics (of Africa). Behaviorally, modern humans have been defined by the universal use of complex spoken language, with a strong capacity for abstract and metaphorical thought (e.g., Mellars (1989)). In recent decades, this 'symbolism' has leapt to the forefront of the debate on the origins of modern humans, with positions on the assembly of the 'modern human behavioral package' being dichotomized as either 'long chronology' (the gradual accretion of modern characteristics) or 'short chronology' (a sudden

explosion of unprecedented modern behavior) (e.g., McBrearty and Brooks (2000) vs. Klein (1995)).

Neanderthal and modern human technologies

Tables 1 and 2 give brief summaries of the archaeological groupings we shall consider, and their skeletal associations (where known). Some terms, especially in Table 1, need further explanation, and the archaeological groupings, henceforth called technocomplexes, need to be set into their spatial and temporal contexts. A technocomplex (following Clarke (1968)) is a group of cultures that share a range of common characteristics, but differ – perhaps by region or time period – in the presence/absence of specific artifact types. The term ‘culture,’ while widely used, is seldom defined, and carries much interpretative baggage. Hence, technocomplex will be preferred here; in any case, its broad spatio-temporal scaling probably better matches most of the archaeological groupings we are discussing.

The technocomplexes listed in Tables 1 and 2 are not exhaustive, but represent a fair selection of those thought to be significant, and/or associated with human remains. They can be grouped into two larger conceptual units, defined largely by how the stone tools were made. Middle Paleolithic assemblages (‘Mousterian,’ if they contain Mousterian points) are generally flake based, that is, their ‘blanks’ for tool production are generally wide and relatively short, and were frequently obtained by a knapping technique called ‘Levallois.’ Classic Levallois entails the careful preparation of a flat stone nodule by first regularizing its entire margin through the removal of overlapping flakes from one (normally the upper) surface, making the core ovoid in shape. The flat surfaces created by these peripheral flake scars are then used as a platform to remove overlapping flakes centripetally from the whole upper surface of the core, creating a ‘tortoise-shell’ gently domed surface that will yield an oval, gently domed flake when struck from one end. Control of the convexity and regularity of this surface is critical. Variants of this Levallois technique exist, producing triangular flakes (‘Levallois points’) and blades (blanks whose lengths are at least twice their widths): the different shapes are created by changes in the orientation of the upper surface’s preparatory removals rather than in the convexity of the surface. Upper Paleolithic assemblages, conversely, frequently used blades produced from prismatic – often conical – cores: not only are these blades at least twice as long as they are wide, but they also carry parallel scars of blades previously removed from the core, whose scar edges form longitudinal ridges on their upper (‘dorsal’) surfaces, typically giving them a trapezoidal cross section. Bladelets are small blades, generally less than 12 mm wide, and sometimes attaining lengths of 50–60 mm (Hassan, 1972).

Having generalized that Middle Paleolithic (flake-based) technologies are associated with Neanderthals, and Upper Paleolithic (blade-based) ones with modern humans, we should be aware that there are important exceptions. In the Levant of southwest Asia, modern human remains are associated with ‘Tabūn C’-type Mousterian assemblages at sites such as Qafzeh and Skhūl at about 100–90 cal ka BP. Unlike thermoluminescence (TL), optically stimulated luminescence (OSL), uranium series, and electron-spin resonance (ESR), radiocarbon dates need to be converted into calendar years (normally designated

as ‘cal BP’). Only recently, however, have calibration curves been applied to radiocarbon dates in our study period (van Andel et al. (2003), etc.), and readers should be aware that there is not yet any universally accepted calibration curve, and probably never will be, as such curves tend to be revised regularly to accommodate new data. The calibrated radiocarbon dates given here (using the CalPal 2005 default curve; Weninger and Jöris (2005) – <http://www.calpal.de>) should be treated broadly and with a degree of caution, but they have the advantage of giving more accurate age estimates than uncalibrated radiocarbon dates, and they allow comparison with the other, aforementioned dating techniques. There appears to be little in their behavioral repertoires to distinguish them from the contemporary and subsequent Neanderthals in that area (Bar-Yosef (1993); Mellars (1989): p. 354). At about the same time in northern Europe, Neanderthals appear to be making blades in exactly the same way as would become the norm in the Upper Paleolithic, before apparently reverting to more ‘typical’ flake-based technologies (e.g., Révillion and Tuffreau (1994)). Therefore, our debate should be less about technological capacity than about preferred technological modes, and about the frequencies of particular core-reduction strategies employed to produce tool blanks.

In most parts of western Eurasia, the Middle Paleolithic technocomplexes listed in Table 1 have disappeared by ca. 35 cal ka BP, although some persisted, particularly in southern Iberia, Croatia, the Crimea, and the Caucasus, until perhaps 30 cal ka BP (e.g., van Andel et al. (2003)). Outside these apparently isolated ‘pockets’ of survival, it becomes much more difficult to identify surviving Neanderthals. The Châtelperronian of France (and perhaps northern Iberia) is the only Upper Paleolithic technocomplex attributed to Neanderthals with any confidence, lasting between ca. 43 and 30 cal ka BP. In particular, this association is based upon the finds from two sites in northern central France: Roche à Pierrot (St.-Césaire) and Grotte du Renne (Arcy-sur-Cure) (e.g., Hublin et al. (1996)). Thus, toward the end of their existence we can see Neanderthals’ behavior changing rapidly in some environments, and utilizing, for apparently the first time, bone and ivory to produce not just tools (awls, etc.) but also elements of personal adornment (beads, pendants, etc.), apparently providing evidence for the application of social meaning to particular artifacts by this species. The big debate in archaeology at the moment is whether this apparent increase in symbolic activity among some Neanderthals was the result of socially internal developmental trajectories, or the result of contact and social interactions with incoming modern human groups – with the behavior of both groups being changed (e.g., d’Errico et al. (1998)). Perhaps the deteriorating conditions in the northern part of the Neanderthal range in France required a change in their social and symbolic behavior: beads, etc., might have been needed to reinforce links between groups in adverse conditions. Certainly, there is little evidence of identifiable symbolic behavior in Neanderthals prior to the Châtelperronian; there are very few isolated and unique finds from some Middle Paleolithic sites, but nothing that can be compared stylistically at the intersite scale. Paradoxically, it seems that Neanderthals were adopting explicitly symbolic behavior (and thus hinting to many archaeologists at the capacity for modern human behavior) just at the time of their extinction (Mellars et al., 1999).

Table 1 Characteristics of selected western Eurasian archaeological technocomplexes at the transition between Neanderthals and modern humans

<i>Technocomplexes</i>			<i>Diagnostic characteristics</i>
Middle Paleolithic	SW Asia	Tabun C-type Mousterian	Squat, thin triangular flakes (Levallois points) are common. Succeeds the Tabun-C variant, and has higher quantities of more elongate Levallois points and retouched tools.
		Tabun B-type Mousterian	
	East Europe	Caucasian Mousterian	Several variants are present: 'para-Micoquian' in the north Caucasus. Typical Mousterian (Levallois <i>and</i> non-Levallois, sometimes with bifaces), (Levallois) Denticulate Mousterian on the NE Black Sea coast, and a Charentian Mousterian site with bifaces as well as sidescrapers.
		Crimean Mousterian	Several variants: the 'Ak-Kaya ('para-Micoquian') tradition,' with bifacial (foliate) knives and points, sidescrapers and some Charentian characteristics (e.g., slug-like highly retouched stone tools called 'limaces'), also encompasses the Kiik-Koba and Staroselian variants. Typical Mousterian assemblages (see below) are concentrated in the nearby Dnestr basin, and they are founded upon Levallois technology.
		East European Micoquian	Perhaps better called <i>Keilmessergruppen</i> , named after the characteristic large, bifacially retouched knife-like tool with an acute cutting edge on one margin. Also contains elongate bifaces with asymmetric cross sections (<i>Halbkeile</i>), small pointed bifaces (<i>Faüstel</i>) and foliate bifaces (<i>Faustkeilblätter</i>). Many of these bifacial forms were made on flat stone nodules; the Levallois technique is sometimes present.
	West Europe	Typical Mousterian	Relatively undifferentiated from other western European Mousterian variants, except that bifaces and backed knives are absent or very scarce. Contains sidescrapers (some of Quina type), Mousterian points, some denticulates and notches.
		Mousterian of Acheulean Tradition ("MTA")	Two phases: (1) MTA-A, with triangular and pointed ovate ('cordiform') bifaces and scarce backed knives, then (2) MTA-B, with less common, exclusively cordiform bifaces and numerous backed knives (some of which anticipate the Châtelperron knife in their morphology). Both phases also contain Mousterian points, sidescrapers, notches and denticulates.
		Charentian Mousterian	Two variants, in roughly chronological order: (1) Ferrassie type, with presence of the Levallois knapping technique and numerous Mousterian points, then (2) Quina type, whose prevalent 'salami-slice' knapping method ('slicing off' flakes from a sausage-shaped stone nodule) was primarily used to make thick, distinctively retouched sidescrapers. Both variants are dominated by sidescrapers (some with bifacially worked convex edges), with fewer notches, denticulates, burins, and endscrapers.
Upper Paleolithic	SW Asia	Denticulate Mousterian	Slightly more distinctive than the typical Mousterian, with high ($\leq 80\%$) levels of notches and denticulates, generally $\leq 10\%$ sidescrapers, scarce endscrapers, borers, and burins; other Mousterian tools are (near-)absent.
		Emiran	Mousterian-like assemblages, characterized by Levallois points carrying bifacial thinning on their bulbous (proximal) ends.
		Baradostian	Production of bladelets from carinated cores, and their transformation into Dufour bladelets and points resembling Font-Yves ones has led some to call it 'Zagros Aurignacian' (Olszewski and Dibble, 1994), but other elements, such as backed blades and 'retouched rods' are not so easy to fit within the Aurignacian.
		Ahmarian	Blade/bladelet dominated assemblages, with diagnostic 'chamfered' pieces, el-Wad points, backed blades/bladelets. Surviving symbolic activity is rare, and bone-working is limited.
	East Europe	Bohunician	Contains a mixture of Middle and Upper Paleolithic elements: sidescrapers, Mousterian points, and a few bifacial leafpoints in association with endscrapers (some Aurignacian-type) and

(Continued)

Table 1 (Continued)

	<i>Technocomplexes</i>	<i>Diagnostic characteristics</i>
West Europe	Szeletian	burins. Both (Upper Paleolithic) blade reduction and Levallois techniques were applied to the same cores. Defined by bifacial leafpoints, with endscrapers (a few Aurignacian type), sidescrapers, Mousterian points and pointed retouched blades; more flake orientated than the Bohunician.
	Jerzmanowician	Defined by laminar leafpoints retouched mainly on the bulbed end of one (the ventral) surface. Splintered pieces and some Mousterian tool types (no sidescrapers) also present.
	Lincombian	Characterized by 'blade-points' (a type of unifacial leafpoint: cf. Jerzmanowician), with some bifacial points. Otherwise very heterogeneous (mixed assemblages?), with Aurignacian tools sometimes present, e.g., <i>busqué</i> burins.
	Châtelperronian	Defined by the backed Châtelperron knife/point; associated with Mousterian-type tools (sidescrapers, etc.). A mixture of flake and blade-orientated strategies is present. Beads, pendants and some bone-working found in some northerly sites.
	Uluzzian	Defined by generally scarce small backed lunates, mostly on flake fragments, although apparently on blades in Greece; also Mousterian tools, splintered pieces, endscrapers, and some bone-working ("awls").
	Aurignacian	Characterized by osseous projectiles, notably the split-based antler point, but also simple-based/lozangic/Mladečc and biconical ones; lithic tools include Dufour bladelets, Font-Yves/Krems/el-Wad points, Vachons and <i>busqué</i> burins, nosed and carinated scrapers, and marginally retouched Aurignacian blades. Beads/pendants are common; some anthropomorphic and zoomorphic figurines.
	Gravettian	Defined by the Gravette point, associated variously with Noailles and Bassaler burins, <i>fléchettes</i> , Font-Robert points and Kostenki points. Prismatic blade(let) technologies are strongly present. Symbolic activity is especially characterized by the female 'Venus' figurines.

Allsworth-Jones, 1986; Bar-Yosef, 1993; Bietti, 1997; Bordes, 1968, 1977; Bosinski, 1967; Campbell, 1980; Chabai, 1996; Cohen and Stepanchuk, 1999; Davies, 2001; Garrod, 1955; Gilead, 1991; Hole and Flannery, 1967; Hublin et al., 1996; Jacobi, 1990; Kozłowski, 1990, 2000; Otte, 1981; Richter, 1997; Svoboda et al., 1996; Veil et al., 1994.

The other European early Upper Paleolithic technocomplexes that are often assumed to have been made by Neanderthals, by analogy with the 'transitional' characteristics of the Châtelperronian, are actually blank slates on which archaeologists can impose their prejudices. The Uluzzian, Szeletian, Jerzmanowician-Lincombian, and Bohunician have not yet yielded any (diagnostic) human remains, although this has not stopped them being attributed to Neanderthals (e.g., d'Errico et al. (1998)). The Levantine Emiran, perhaps dated to about 50 cal ka BP (or earlier), appears to have developed into the Ahmari in situ within the Levant (Marks and Volkman, 1983); the latter technocomplex was probably associated with modern humans, if the skeletal remains from the Lebanese site of Ksar 'Akil are representative and reliable (Ohnuma and Bergman, 1990). The Emiran has also been connected with the Bohunician of central Europe – they share a similar use of Levallois technology – and it has been postulated that both might have been made by modern humans, who dispersed from the Levant to Europe at around 45 cal ka BP (e.g., Tostevin (2003)). At present, it is difficult to test this hypothesis, although the Bohunician does seem to represent a technological break with the local Middle Paleolithic of central Europe. This is different from the western European situation, where the

Châtelperronian shares several characteristics with the MTA-B (Table 1), for example, backed knives/points, and probably was derived from it.

The pan-European Aurignacian and Gravettian technocomplexes have been extensively studied, and are presumed to be 'true' Upper Paleolithic entities. In addition to the working of bone, antler, and ivory (not just to produce awls), the use of bladelets in both was extensive, and is not really seen in preceding (Neanderthal) European technocomplexes (Bar-Yosef and Kuhn, 1999). Bladelets permit flexibility, in that they can be arrayed in a shaft in different combinations to perform different tasks; any component unit ('microlith') can be replaced easily if it breaks.

It has often been argued that the Aurignacian was carried out of the Levant into eastern Europe, and then continued to move westward (e.g., Mellars (1989)). However, the earliest Aurignacian-type assemblages appear to be eastern European, dated at about 43 cal ka BP, and slightly later Aurignacian assemblages have been found further east in the Levant, as well as further west in Germany, Franco-Cantabria, and Belgium. At face value, the relative lateness and restricted spatial distribution of Levantine Aurignacian sites might appear to contradict hypotheses arguing for a spread of modern humans

Table 2 Skeletal associations of archaeological technocomplexes in western Eurasia

		<i>Homo neanderthalensis</i>	<i>Homo sapiens</i>	<i>Unknown (Homo sp.)</i>
Middle Paleolithic		European Middle Paleolithic and Mousterian SW Asian Middle Paleolithic and Mousterian		
Upper Paleolithic	SW Asia Europe	Szeletian? Châtelperronian	Ahmarian Aurignacian ^a Gravettian	Emiran Baradostian Bohunician Uluzzian Jerzmanowician Lincombian

^aAurignacian in the broad sense, including the 'Bachokirian' of eastern Europe.

out of Africa into Europe via the Levant, but it can easily be argued that what we know now as the 'Aurignacian' developed as people had to respond to different environmental conditions and spatial distributions of raw materials and resources. The Levantine Aurignacian might thus represent a back-migration of the European Aurignacian into the Near East. The Aurignacian seems to have been made by modern humans, as far as we can judge, although in contrast to the preceding Middle Paleolithic and succeeding Gravettian technocomplexes, it has tended to yield fragmentary associated human remains (e.g., [Henry-Gambier et al. \(2004\)](#)). Perhaps the desire to break up bodies was related to the apparent adaptations to high mobility shown by the Aurignacian. The Gravettian can be more confidently ascribed to modern humans, as whole human bodies were interred in identifiably symbolic burials. Some of the earliest modern human remains in Europe have no associated archaeology ([Table 3](#)): they are certainly contemporary with the Aurignacian, although we cannot be certain that they themselves were its authors.

How and why did Neanderthals become extinct?

We shall explore Neanderthal responses to changing climate in this section, which of course means we must consider elements of chronology and how they exploited food and other resources. Intercomparability of dating techniques is important, as it allows us to compare our results with the Greenland ice-core climate records (given in cal BP). Taking the dated sites we have at face value, it would appear that Neanderthals were most successful between 60 and 40 cal ka BP, which was the stable warm phase of (marine) oxygen isotope stage (OIS) 3, immediately after the cold maximum of OIS-4 (72–60 cal ka BP) ([van Andel et al. 2003](#)). After 40 cal ka BP, the climate began to cool, and by 30 cal ka BP Neanderthals appear to be extinct.

Naturally, the dating methods we use, and how we apply them, are crucial if we are to be confident in the chronology we have (see [Dating Techniques](#); [AMS Radiocarbon Dating](#); [U-Series Dating](#); [Thermoluminescence](#); [Electron Spin Resonance Dating](#)). Radiocarbon is still the main dating technique for our study period – owing to reasons of its affordability, the prevalence of organic remains at many archaeological sites, and its availability over more than 55 years – even though its reliability decreases beyond 30 ka BP and only a few laboratories produce reliable age estimates beyond 40 ka BP. 50 ka BP is essentially the limit of the method. Other dating methods, such as TL, OSL, uranium series, and ESR, are needed

to extend our (reliable) chronology further back in time, certainly beyond 30 ka BP.

If we believe that replacement of Neanderthals by modern humans occurred in Eurasia, then we have recourse to two competing hypotheses: either Neanderthals were outcompeted by modern humans (what one might call 'confrontational replacement'), or they had already become extinct before modern humans arrived in an area ('successional replacement'). To test these models, archaeologists have traditionally used two lines of chronological/associative evidence: the relative stratigraphic positions of assemblages attributed to both human species, and/or absolute dating of the latter. While relative stratigraphic positions can give us an idea of succession, or perhaps indicate contemporaneity if interstratification can be proved, they give no real sense of chronology: without absolute dating methods, we do not know the duration of each depositional (or erosional) phase represented by a geological layer. In addition, it is by no means evident that a geological layer is equivalent to a phase of human occupation at a given site. It is instead obvious that in many cases a geological layer contains the remnants of more than one occupation episode; taphonomic disturbance between levels is not necessarily the only explanation for wide age estimate ranges for a given layer. Obtaining direct dates on anthropogenic material, ideally diagnostic artifacts, is obviously the way to identify such occupational phases, but has only recently begun to be practiced.

Current evidence does not tend to support arguments of interstratification, with the claimed occurrences at the French sites of Roc de Combe and Le Pige being now generally attributed to taphonomic disturbance and redeposition of sediments (see [d'Errico et al. \(1998\)](#)). Nevertheless, a recent claim has been made for interstratification of Châtelperronian (assumed to be Neanderthal) and Aurignacian (modern human) levels at the French site of Châtelperron (see [Mellars et al. \(1999\)](#), p. 346). Notwithstanding the stratigraphic evidence, current absolute dates for both Neanderthal and modern human technocomplexes do overlap temporally at the regional scale, especially in western Europe (France and Iberia), between ca. 43–33 ka BP (ca. 47–38 cal ka BP).

Stratigraphic information, then, would suggest that Neanderthals almost always preceded modern humans at sites where technocomplexes from both species are found, while the absolute ages we have (including those from sites with technocomplexes attributed to just one species) suggest a broader contemporaneity. Both lines of evidence are actually telling us subtly different things: the stratigraphic evidence is telling us about the chronology of occupation in a site, and,

Table 3 Directly dated late Neanderthal remains and those of early modern humans in Europe between ca. 50 and 30 cal ka BP

	<i>Site/specimen</i>	<i>Lab. No.</i>	<i>Date (uncal BP)</i>	<i>Date (cal BP)</i>	<i>Comments</i>
Neanderthal	Mezmaiskaya: layer 3d	Ua-14512	29 195 ± 965	33 855 ± 1,318	A late Neanderthal from the Caucasus mountains.
	Vindija Cave: layer G1	OxA-8296	29 080 ± 400	34 180 ± 640	Both C14 dates (OxA-) are much younger than the 2 uranium series ones.
		**	51 000 ± 8000	51 000 ± 8000	
		**	46 000 ± 7000	46 000 ± 7000	
	Feldhöfer-kirche (Neandertal)	OxA-8295	28 020 ± 360	32 560 ± 728	Both dates are from the Neanderthal type-specimen(s).
		[ETH-??]	39 240 ± 670	43 468 ± 594	
Modern human	Banyoles: lake travertine (5 m depth) around mandible	[ETH-??]	39 900 ± 620	43 806 ± 564	A striking difference between the very young C14 date (UCLA-) and the uranium series one.
		**	45 000 ± 4000	45 000 ± 4000	
	Pestera cu Oase: mandible	UCLA-930	17 600 ± 1000	21 098 ± 1151	This site appears to have the earliest-known modern human in Europe; no associated archaeology however.
		OxA-11711	>35 200	N/A	
	Oblazowa: layer VIII: phalange	GrA-22810	34 290 + 970/-870	39 823 ± 1,301	
		OxA-4586	31 000 ± 550	36 110 ± 496	
	Mladeč: ulna & teeth	VERA-2736	26 330 ± 170	30 997 ± 130	Some associated (Aurignacian) artefacts, but the dates are quite variable: the ulna (VERA-2736) and the 'brown-colored collagen fraction' (VERA-3075) are notably younger than the other two dates (on teeth).
		VERA-3073	31 190 + 400/-390	36 263 ± 381	
		VERA-3074	31 320 + 410/-390	36 367 ± 382	
		VERA-3075	30 680 + 380/-360	35 873 ± 385	
		VERA-3076A	31 500 + 420/-400	36 512 ± 408	
		VERA-3076B	27 370 ± 230	31 646 ± 290	
	Grotte de Cussac, loc.1 : rib	Beta-156643	25 120 ± 120	30 103 ± 193	A Gravettian, not Aurignacian, age.
	Kent's Cavern: maxilla	OxA-1621	30 900 ± 900	36 065 ± 851	The earliest modern human in western Europe?
	Paviland (Goat's Hole)	OxA-1815	26 350 ± 550	30 896 ± 339	A Gravettian, not Aurignacian, age from a complete burial.
		OxA-8025	25 840 ± 280	30 684 ± 232	

Trinkaus E et al. (2003) *Proceedings of the National Academy of Sciences* 100: 11231–11236; Wild EM et al. (2005) *Nature* 435: 332–335; other references in Davies W (2001) Stage Three Project Archaeological Database: <http://www.esc.cam.ac.uk/oistage3/details/homepage.html>

when taken at the regional scale, is telling us something about how groups of each human species used their landscape and its resources. The absolute ages are telling us about when sites were occupied, and by incorporating ages from sites with, say, only Aurignacian remains (and none attributed to Neanderthals), we can plot the distributions of Neanderthals over time and space: at certain periods, were some areas 'Neanderthal' or 'modern human?' Any sites with interstratified Neanderthal and modern human archaeological assemblages would thus represent shifting boundaries between the two species.

The desire of d'Errico et al. (1998) to see Neanderthals never overlap with modern humans, based on site stratigraphic successions, favors independent invention by Neanderthals of blade technology and bone working without any influence (known as 'acculturation') by incoming modern humans. Having found new solutions to (new) problems, they then suffer the misfortune of local extinction. Acculturation is very much in the thoughts of those who favor a contemporaneous existence for both species prior to Neanderthal extinction, with social connections between Neanderthal and modern human groups catalyzing behavioral changes, either equally or unequally distributed between each species. At present, our data are not good enough to identify any possible acculturation in either direction between modern humans and Neanderthals, but a useful start to addressing this problem would be made if the bones of contention (diagnostic artifacts) were themselves dated, and the chronologies then compared directly.

Archaeologists like to identify boundaries in the Paleolithic world, and a major one that has arisen in the literature over the last 15 years has been the so-called 'Ebro frontier' (e.g., d'Errico et al. (1998)). It is claimed that modern humans were either unable or unwilling to disperse into the parts of Iberia south of the river Ebro until about 30 cal ka BP; until that time, the region remained a Neanderthal bastion. It is of course doubtful that a river could form a boundary to a species that had already dispersed over huge distances, and even reached Australia, and the frontier's proponents thus claim that it was ecological factors (notably the structure of habitats, and the resources available) that held the marauding moderns at bay. When the Mousterian sites in Iberia are plotted on the map, it is evident that the sites we have are concentrated around the margins of this peninsula: the central, upland Meseta region is more or less uninhabited, and perhaps it formed a 'buffer zone' between the largely modern human occupied north, and the late Neanderthals in the southerly coastal regions.

The oddity is that these last Neanderthals in southern Iberia were relatively conservative in behavioral terms: they did not make Châtelperronian assemblages, and continued with the Mousterian. Perhaps they could afford to behave in this fashion because environmental change in this region was lesser than elsewhere in western Europe; the tightly packed mosaic of different habitats ensuring that Neanderthal groups did not have to range too widely to obtain the resources they needed to

survive. Modern humans, in contrast, tended to range more widely, certainly in terms of acquiring stone raw materials for their tools, and to make greater use of these 'exotic' raw materials; they also obtained marine shells over long distances for use in personal adornment (see Mellars (1989), p. 361). However, in terms of animal species exploited, there are fewer clear-cut differences between the two human species: both tended to hunt medium-to-large mammals (reindeer, red deer, wild boar, horse, mammoth), although at some sites we can see that modern humans exploited an impressive range of animals, ranging from mammoths at one end of the scale to hares, fish, and birds at the other. Only in the last two decades or so have archaeologists begun to ask how Neanderthals and modern humans exploited the same food species, and whether their exploitation strategies show any clear differences (Stewart, 2004). It has been observed that Neanderthal skeletons show high levels of trauma and injury: probably their hunting methods led to frequent close-contact encounters with their prey, but in the absence of comparable early modern human remains, we cannot say if these injuries were unique to Neanderthals (Berger and Trinkaus, 1995). Nevertheless, we are still a long way from identifying seasonal differences in the occupation periods at sites: where Neanderthals and modern humans were at particular times of year is not a question we can yet answer.

As fine-grained mosaics of different habitats became increasingly rare in Europe after 37 cal ka BP, replaced by swathes of more homogeneous environments, Neanderthals would have found it more difficult to obtain the resources they needed and to move to surviving refuges. Fragmentation into small pockets of suitable habitats would render them much more susceptible to extinction, especially with modern humans 'playing cuckoo' in the interstitial areas they once inhabited. Whether or not modern humans had a significant behavioral advantage (they at least managed to survive the last cold maximum of the Ice Age at ca. 21 cal ka BP), they could have administered the fatal blow to Neanderthals simply because they were not restricted to Europe: waves of modern humans coming in to Europe could always replenish groups that had become extinct, whereas Neanderthals could never call upon reinforcements from outside Europe.

We stand today at the threshold of a new dating revolution in the transition from Neanderthals to modern humans: being able to date diagnostic organic artifacts, for example, the bone industries from some Châtelperronian sites, will allow us to state with more confidence whether they can be attributed to Neanderthal authorship or not. If they are not of the same age, then we can break the presumed association. Similarly, we can apply the same dating techniques to the skeletal remains and artifacts presumed to belong to the earliest modern humans: AMS radiocarbon dating of small (=0.3 g) organic samples (diagnostic pieces) and TL dating of heated (diagnostic) stone tools. The direct dating of human remains and particular classes of tools will allow archaeologists to compare like with like across regions, and to assess the validity and spatio-temporal coherence of our presumed technocomplexes, rather than quibble over the quality of association on a site-by-site basis. Only when we have placed our chronology on a firmer footing can the transition from Neanderthals to modern humans be reassessed.

See also: Dating Techniques; **Luminescence Dating:** Thermoluminescence; Electron Spin Resonance Dating; **Radiocarbon Dating:** AMS Radiocarbon Dating; U-Series Dating.

References

- Aiello LC and Wheeler P (2003) Neanderthal thermoregulation and the glacial climate. In: van Andel TH and Davies W (eds.) *Neanderthals and Modern Humans in the European Landscape During the Last Glaciation*, pp. 147–166. Cambridge: McDonald Institute Monographs.
- Allsworth-Jones P (1986) *The Szeletian and the Transition from Middle to Upper Palaeolithic in Central Europe*. Oxford: Clarendon.
- Arensburg B (1989) New skeletal evidence concerning the anatomy of Middle Palaeolithic populations in the Middle East. In: Mellars P and Stringer C (eds.) *The Human Revolution: Behavioural and Biological Perspectives in the Origins of Modern Humans*, pp. 165–171. Edinburgh: Edinburgh University Press.
- Bar-Yosef O (1993) The Role of Western Asia in Modern Human Origins. In: Aitken MJ, Stringer CB, and Mellars PA (eds.) *The Origin of Modern Humans and the Impact of Chronometric Dating*, pp. 132–147. Princeton: Princeton University Press.
- Bar-Yosef O and Kuhn SL (1999) The big deal about blades: Laminar technologies and human evolution. *American Anthropologist* 101: 322–338.
- Berger TD and Trinkaus E (1995) Patterns of trauma among the Neandertals. *Journal of Archaeological Science* 22: 841–852.
- Bietti A (1997) The transition to anatomically modern humans: The case of peninsular Italy. In: Clark GA and Willermet CM (eds.) *Conceptual Issues in Modern Human Origins Research*, pp. 132–147. New York: Aldine de Gruyter.
- Bordes F (1968) *The Old Stone Age*. London: Weidenfeld and Nicolson (World University Library).
- Bordes F (1977) Time and space limits in the Mousterian. In: Wright RVS (ed.) *Stone Tools as Cultural Markers*, pp. 37–39. Canberra: Australian Institute of Aboriginal Studies.
- Bosinski G (1967) *Die Mittelpaläolithischen Funde im Westlichen Mitteleuropa*. Köln: Fundamenta, Reihe A Bd. 4.
- Campbell JB (1980) Le problème des subdivisions du Paléolithique supérieur britannique dans son cadre européen. *Bulletin de la Société Royale Belge d'Anthropologie et de Préhistoire* 91: 39–77.
- Chabai VP (1996) Kabazi-II in the context of the Crimean Middle Paleolithic. *Préhistoire Européenne* 9: 31–48.
- Clarke DL (1968) *Analytical Archaeology*. London: Methuen.
- Cohen VY and Stepanchuk VN (1999) Late Middle and Early Upper Paleolithic evidence from the East European Plain and Caucasus: A new look at variability, interactions, and transitions. *Journal of World Prehistory* 13: 265–319.
- Davies W (2001) A very model of a modern human industry: New perspectives on the origins and spread of the Aurignacian in Europe. *Proceedings of the Prehistoric Society* 67: 195–217.
- D'Errico F, Zilhão J, Julien M, Baffier D, and Pélegrin J (1998) Neanderthal acculturation in western Europe? A critical review of the evidence and its interpretation. *Current Anthropology* 39(supplement): S1–S44.
- Garrod DAE (1955) The Mughareh el Emireh in lower Galilee: Type station of the Emiran industry. *Journal of the Royal Anthropological Institute* 85: 141–162.
- Gilead I (1991) The Upper Paleolithic period in the Levant. *Journal of World Prehistory* 5: 105–154.
- Hassan FA (1972) Toward a definition of "Lames", "Lamelles", and "Microlamelles". *Libya* 20: 163–267.
- Henry-Gambier D, Maureille B, and White R (2004) Vestiges humains des niveaux de l'aurignacien ancien du site de Brassempouy (Landes). *Bulletins et Mémoires de la Société d'Anthropologie de Paris (n.s.)* 16: 49–87.
- Hole F and Flannery KV (1967) The Prehistory of Southwestern Iran: A preliminary report. *Proceedings of the Prehistoric Society* 33: 147–206.
- Holliday TW (1997) Postcranial evidence of cold adaptation in European Neandertals. *American Journal of Physical Anthropology* 104: 245–258.
- Hublin J-J, Spoor F, Braun M, Zonneveld F, and Condemi S (1996) A late Neanderthal associated with Upper Palaeolithic artefacts. *Nature* 381: 224–226.
- Jacobi RM (1990) Leaf-points and the British early Upper Palaeolithic. In: Kozłowski JK (ed.) *Feuilles de Pierre: Les Industries à Pointes Foliacées du Paléolithique Supérieur Européen*, Liège: Études et Recherches Archéologiques de l'Université de Liège, 42: (Actes du Colloque de Cracovie, 1989).
- King W (1864) The reputed fossil man of the Neanderthal. *Quarterly Journal of Science* 1: 88–97.

- Klein RG (1995) Anatomy Behavior, and Modern Human origins. *Journal of world prehistory* 9: 167–198.
- Kozłowski JK (1990) Certains aspects techno-morphologiques des pointes foliacées de la fin du Paléolithique moyen et du début du Paléolithique supérieur en Europe centrale. In: Farizy C (ed.) *Paléolithique moyen récent et paléolithique supérieur ancien en Europe. Ruptures et transitions: Examen critique des documents archéologiques* 3: pp. 125–133. Nemours: Mémoires d'Île-de-France.
- Kozłowski JK (2000) Châtelperronien, Uluzzien et quoi plus à l'est? *Anthropologie* 38: 249–259.
- Lieberman P (1989) The origins of some aspects of human language and cognition. In: Mellars P and Stringer C (eds.) *The Human Revolution: Behavioural and Biological Perspectives in the Origin of Modern Humans*, pp. 391–414. Edinburgh: Edinburgh University Press.
- Marks AE and Volkman PW (1983) Changing core reduction strategies: A technological shift from the Upper Paleolithic in the Southern Levant. In: Trinkaus E (ed.) *The Mousterian Legacy: Human Biocultural Change in the Upper Pleistocene*, Oxford: British Archaeological Reports, International Series 164.
- Mellars P (1989) Major issues in the emergence of modern humans. *Current Anthropology* 30: 349–385.
- Mellars PA, Otte M, and Straus LG (1999) The Neanderthal problem continued [with reply from Zilhão J and d'Errico F]. *Current Anthropology* 40: 341–364.
- Ohnuma K and Bergman CA (1990) A technological analysis of the Upper Palaeolithic levels (XXV–XXVI) of Ksar Akil, Lebanon. In: Mellars P (ed.) *The Emergence of Modern Humans: An Archaeological Perspective*, pp. 91–138. Edinburgh: Edinburgh University Press.
- Olszewski DI and Dibble HL (1994) The Zagros Aurignacian. *Current Anthropology* 35: 68–75.
- Otte, M. (1981). *Le Gravettien en Europe Centrale* (Dissertationes Archaeologicae Gandenses, vol. 20). 2 vols, De Tempel, Brugge. (Dissertationes Archaeologicae Gandenses, vol. 20).
- Révillion, S., and Tuffreau, A. (Eds.) (1994). *Les industries laminaires au Paléolithique moyen* (Dossier de Documentation Archéologique 18), CNRS Editions, Paris.
- Richter, J. (1997). *Sesselfelsgrötte III—Der G-Schichten-Komplex der Sesselfelsgrötte: Zum Verständnis des Micoquien (Forschungsprojekt "Das Paläolithikum und Mesolithikum des Unteren Altmühltals II" Teil III)*. Saarbrücker Druckerei und Verlag, Saarbrücken. (Quartär-Bibliothek, Bd. 7).
- Stewart JR (2004) Neanderthal—modern human competition? A comparison between the mammals associated with Middle and Upper Palaeolithic industries in Europe during OIS 3. *International Journal of Osteoarchaeology* 14: 178–189.
- Stringer CB (1989) The origin of early modern humans: A comparison of the European and non-European evidence. In: Mellars P and Stringer C (eds.) *The Human Revolution: Behavioural and Biological Perspectives in the Origin of Modern Humans*, pp. 232–244. Edinburgh: Edinburgh University Press.
- Svoboda J, Ložek V, and Vlček E (1996) *Hunters between East and West: The Paleolithic of Moravia*. New York: Plenum.
- Trinkaus E, Mowovan O, Milota S, et al. (2003) An early modern human from the peștera ca Oase, Romania. *Proceedings of the National Academy of Sciences* 100: 11,231–11,236.
- Tostevin GB (2003) A quest for antecedents: A comparison of the Terminal Middle Palaeolithic and early Upper Palaeolithic of the Levant. In: Goring-Morris AN and Belfer-Cohen A (eds.) *More than Meets the Eye: Studies on Upper Palaeolithic diversity in the Near East*, pp. 54–67. Oxford: Oxbow Monographs.
- van Andel TH, Davies W, and Weninger B (2003) The human presence in Europe during the last glacial period. I: Human migrations and the changing climate. In: van Andel TH and Davies W (eds.) *Neanderthals and Modern Humans in the European Landscape During the Last Glaciation*, pp. 31–56. Cambridge: McDonald Institute Monographs.
- Veil S, Breest K, Höfle H-C, et al. (1994) Ein mittelpaläolithischer Fundplatz aus der Weichsel-Kaltzeit bei Lichtenberg, Ldkr. Lüchow-Dannenberg: *Zwischenbericht über die archäologischen und geowissenschaftlichen Untersuchungen 1987–1992*, Germania 72.
- Wild EM, Teschler-Nicola M, Kutschera W, Steies P, Trinkaus E, and Wanck B (2005) Direct dating of Early upper Palaeolithic human remains from Mladeč. *Nature* 435: 332–335.
- Wolpoff M and Caspari R (1997) *Race and Human Evolution*. New York: Simon and Schuster.

Relevant Websites

- Davies W (2001b) Stage Three Project Archaeological Database. <http://www.esc.cam.ac.uk/oistage3/Details/Homepage.html>.
- Weninger Jöris O (2005) Cologne Radiocarbon Calibration & Paleoclimate Research Package. <http://www.calpal.de>.

Postglacial Adaptations

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World Overview

By archaeological convention, the Postglacial is defined as the period beginning with the final melting of the Scandinavian ice sheet at about 10 radiocarbon ka BP. This date also marks the conventional beginning of the Mesolithic period, seen as the continuation of a Paleolithic hunter-gatherer way of life adapted to the changed environmental conditions of the Postglacial – in particular, the change from the glacial steppe with its migrating herds of large mammals, to heavily forested environments with fewer resources. This is a view shaped by European data and intellectual history, in which the Mesolithic has traditionally been viewed as a period of cultural and demographic decline or stasis, marking time until Neolithic farmers from the Near East entered Europe, some 8 radiocarbon ka BP, replacing the hunter-gatherer way of life across the continent over the following three millennia.

Archaeological investigations now demonstrate that this notion of progressive evolution through uniform stages of development is greatly oversimplified. So far from being a period of cultural stagnation, these ‘Mesolithic’ millennia that preceded the development of full-scale village farming are now seen as representing a period of radical change and innovation, while Mesolithic achievements in their turn were built on developments, already well underway, in the closing millennia of the Last Glacial period (Mithen, 2003).

From an archaeological point of view, there are good reasons to extend the period of ‘Postglacial’ adaptations back to the beginning of the warming trend initiated at about 13 radiocarbon ka BP with the Late Glacial interstadials, and perhaps back to 16 radiocarbon ka BP, when sea levels began their sustained rise from the low point of the Last Glacial Maximum (see [Last Glacial Maximum GCMs](#)). As sea level rose to reach its present position about 6 radiocarbon ka BP, progressive inundation of the continental shelf removed extensive areas of lowland territory, breached inter-continental land connections, brought existing hinterlands within reach of milder ‘oceanic’ climates, and culminated in the creation of entirely new coastal landscapes. The impact of sea-level rise on paleoecology and paleogeography would have differed in different areas, but the effects would have been experienced worldwide. They were especially marked in regions with shallow continental shelves, such as northwest Europe, where large areas of hunting territory were lost, only to be replaced by the creation of shallow inshore waters, indented coastlines, and offshore islands and archipelagos, with more productive and more easily accessible supplies of fish, sea mammals, and intertidal mollusks.

It is during this Late Glacial period of changing environmental conditions, and especially between 13 and 12 radiocarbon ka BP, that we see the first stages of population expansion into new territory exposed by glacial retreat in northern Eurasia, the first

unequivocal evidence for the expansion of human populations into the Americas (Dillehay, 2000), the extensive use of micro-blade technologies with miniature bladelets and points blunted along one edge to provide inserts for hafted hunting weapons throughout large areas of the inhabited world, and the development of pottery in China, the Russian Far East, and Japan (Habu, 2004, Kuzmin et al., 2004).

Regions strongly affected by glaciation saw the rapid spread of populations into deglaciated territory, and the development of coastal and inland economies with significant dependence on aquatic resources along the shorelines of newly created lakes and coastlines throughout the northerly regions of Eurasia and North America. In subsequent millennia, subsistence economies throughout the world diversified into new niches exploiting a wider range of animal and plant food resources, including the first evidence for sustained and intensive use of marine resources on many of the world’s coastlines. In semi-arid environments at lower latitudes, stands of seed-bearing plants began to expand as climatic conditions improved, creating the basis for intensive harvesting of plant foods. In the Near East, this process was underway from as early as 12 radiocarbon ka BP, ultimately leading to agriculture based on cereal cultivation and domestic animals.

In temperate latitudes, the same climatic trends favored the expansion of woodland cover and reimmigration of deciduous trees, resulting in a complex and rapidly shifting geographical mosaic of vegetation and faunal patterns, in which the large herds of steppe animals such as reindeer, horse, and bison were replaced by woodland species of deer and wild boar. At the same time, these changes in vegetation brought with them new opportunities for food – nut-bearing shrubs and trees such as hazel, oak, and chestnut, and edible roots, leafy greens, and fruits. Hunting parties found new opportunities for seeking out prey such as chamois and ibex on the higher slopes of mountains above the treeline made accessible by deglaciation. Other resources such as birds and small mammals played a larger role, facilitating greater residential stability, larger communities of people, and smaller territories.

High-latitude regions with sub-Arctic conditions were first occupied from at least 10 radiocarbon ka BP on the coastlines of Norway and Alaska, and the final retreat of the North American ice sheets and the opening up of tundra environments in northern Canada and Greenland was closely followed by the expansion of ‘paleoeskimo’ populations into the high Arctic regions of North America between 5 and 4 radiocarbon ka BP. The occupation of this new niche was accompanied by exploitation of sea mammals alongside hunting of animals on land, and by dramatic and rapid expansions of human range. In Norway, 2,000 km of coastline was colonized in a matter of generations (Bjerck, 2008). In North America, 4,800 km of new territory from the Alaskan border to northeast Greenland was opened up in a few centuries (Fiedel, 1987).

Europe

The European landmass, broadly defined, provides an appropriate microcosm of these global trends (Figure 1). With a wide range of environments from the Eastern Mediterranean to the sub-Arctic shores of northern Scandinavia, and a long history of Mesolithic research stimulated by regular international meetings and periodic overviews (most recently Bailey and Spikins (2008), Larsson et al. (2003)), it provides an unparalleled range and richness of archaeological material. We can subdivide developments into three broadly defined periods: (1) Late Glacial, ca. 13–10 radiocarbon ka BP, characterized by expansion into newly available territory; (2) Early Mesolithic, ca. 10–8 radiocarbon ka BP, with consolidation of settlement and adaptation of hunting and gathering to new vegetational conditions, and a renewed phase of expansion into high-latitude maritime environments; (3) Late Mesolithic, ca. 8–5 radiocarbon ka BP, diversification of settlement and economy and greater regional differentiation, development of sedentary villages in productive aquatic environments on coastlines, rivers, and lakes, the husbandry of naturally occurring plant and animal resources, including evidence of environmental modification, and interaction with farmers and selective adoption of agricultural practices. These chronological

boundaries vary according to region, but nevertheless appear to coincide with major climatic changes, particularly in more northerly regions most sensitive to the effects of deglaciation.

Glacial Prelude

At the height of the Last Glacial, cold and aridity resulted in environments dominated by open steppe or forest-steppe vegetation that extended from the periglacial margins of the ice sheets in the north to the shores of the Mediterranean. Trees were confined to localized stands of coniferous trees in sheltered localities, whereas deciduous species were restricted to pockets of warmer and wetter conditions in refugia at intermediate altitude in the Pyrenees and the circum-Mediterranean mountain ranges. Human populations were largely dependent on migratory herd animals such as reindeer, horse, and bison in colder regions and wild cattle, red deer, gazelle, and the steppe ass in the south and southeast. Woolly mammoth were abundant on the more extensive steppes of eastern Europe and were an important source of meat, bone for artifacts and fuel, and ivory for decorative items. Plant foods were scarce nearly everywhere, though probably exploited where available.

Human populations were distributed in small and mobile bands with large annual territories and social contacts with

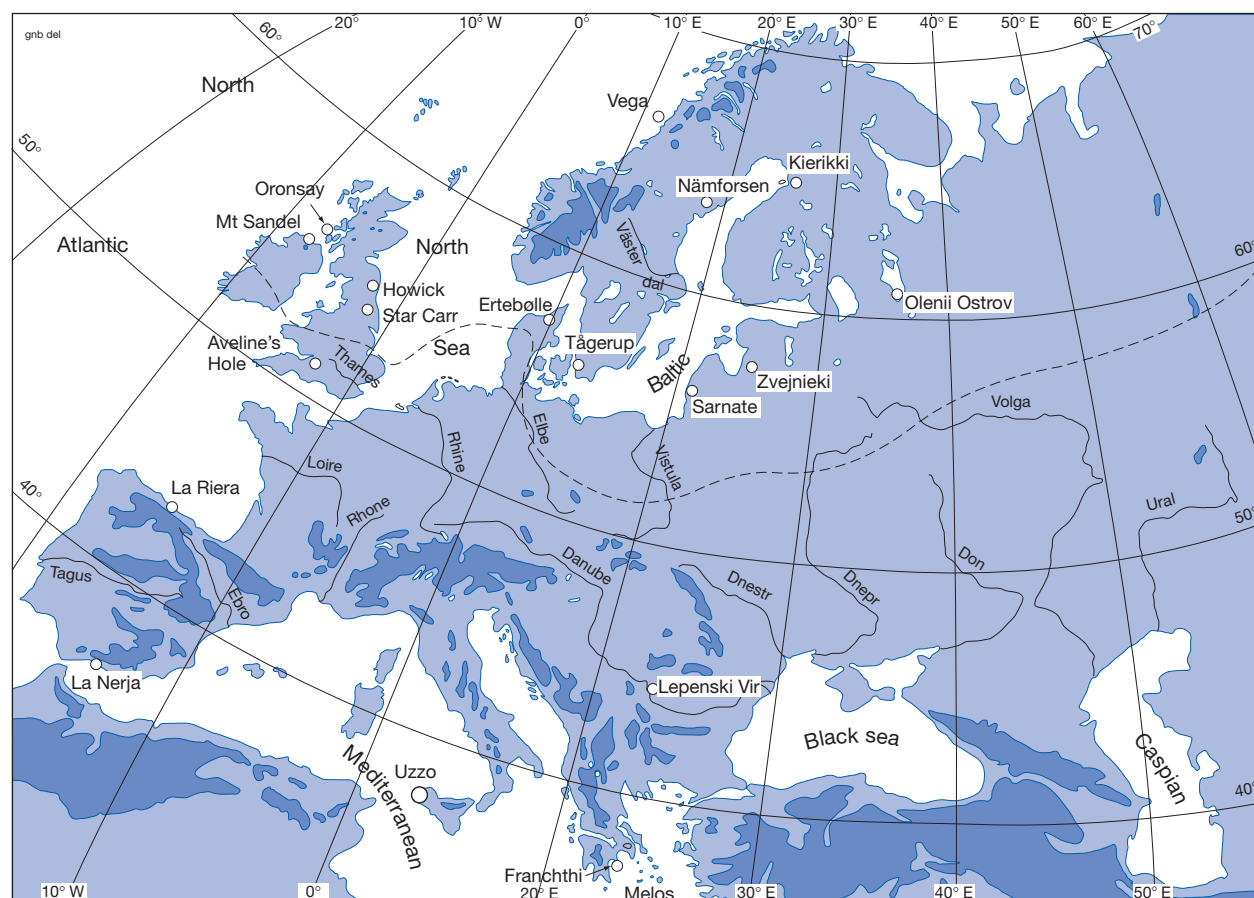


Figure 1 Map of Europe showing principal topographic features and sites (see text). Shaded areas indicate land above about 1000 m, and the dashed line indicates the maximum extent of the Scandinavian ice sheet at the glacial maximum.

neighboring bands over long distances, signaled in the archaeological record by the movement of exotic materials such as marine shells and rare and fine-grained flint many hundreds of kilometers from their sources. The most favorable environments, the largest concentrations of archaeological sites, and the highest population densities were in southwest France and northern Spain in the west and the riverine basins of eastern Europe and the Ukraine in the east. These were also the centers of cultural innovation and of social aggregation and ritual, associated in the west with the flowering of Franco-Cantabrian rock art, and in the east with the production of a rich variety of portable art and the construction of dwelling structures built from mammoth bones (see [Interactions with Hominids](#)).

Late Glacial Expansion

It was from these centers that human populations first expanded into newly accessible territory as the ice sheets retreated and climate ameliorated. Many of the animal species of former significance moved their ranges northward, as with reindeer and horse, or became extinct, as with mammoth and bison, and were replaced by forest or forest-edge species such as elk, red deer, roe deer, cattle, and wild boar, according to regional and local circumstances.

From about 13 radiocarbon ka BP onward, settlements associated with hunting of reindeer and, later, horse, cattle, and elk, became established across the north European plain in a broad swathe from the British lowlands in the west to the Crimea in the east ([Dolukhanov, 2008](#), [Tolan-Smith, 2008](#), [Zvelebil, 2008](#)). Stone industries were characterized by tanged points for hafting on the ends of spear or arrow shafts, and by 'knife' blades with a curved and blunted edge. The southern end of the North Sea was still dry land at this period and the recovery of an antler harpoon dredged up from the Leman and Ower Bank and dated to 11.7 radiocarbon ka BP shows that this territory was used from an early period ([Tolan-Smith and Bonsall, 1999](#)). Variations in the proportion of these artifacts and minor differences in morphology of the key artifact types have given rise to a variety of cultural labels, Swiderian in Poland, Hamburgian, Bromme, and Ahrensburgian in northern Germany and Denmark, and Federmesser and Creswellian in the Low Countries and Britain. Whether these represent stylistic markers of distinct social groupings or functional adaptations to different types of weapons and prey species is unclear. The smaller and lighter tanged points of the Ahrensburgian, associated with the final cold phase of the Younger Dryas period between 11 and 10 radiocarbon ka BP (see [Younger Dryas Oscillation, Global Evidence](#)) could well have been used as arrow points, since actual bows have been recovered from this period. At any rate, the similarities across this vast territory are as impressive as any differences, with relatively small settlements, specialized toolkits of stone and antler, and a considerable degree of mobility. Importation of raw materials for making artifacts from favored sources over distances of 150 km or more reflects both the extended lines of communication in small mobile populations engaged in the early stages of colonizing new territory and lack of familiarity with less suitable materials available nearer to hand or the absence of any need to use them.

The Early Mesolithic

Technologically, there is no obvious break with the preceding period, although there is a greater diversity of raw materials, techniques of stone working, and tool types. Microlithic stone tools made from bladelet segments proliferated into new geometric forms, particularly rectangular pieces, triangles, and crescents. Many of these were most probably used to provide replacements for arrow tips, and the bow and arrow were widely in use throughout Europe during this period. Some may also have been used as replaceable components in other multipart tools such as knives with bone or wooden handles. There was also a general diversification of artifact types made from stone, bone, antler, and wood throughout the Mesolithic period, including macrolithic tools of worked antler, flaked flint, or other stone, and polished slate, used for working wood, digging the ground, or processing sea-mammal carcasses and fish. Harpoons of bone or antler and bone fishhooks are evidence of new technologies for exploiting marine and aquatic resources. Wooden artifacts include dugouts, paddles, fish weirs, spear and arrow shafts, and timber-framed houses. Artificial dwelling structures are more widely in evidence and range from lightly built tent-like structures to substantial stone-built constructions, depending on the available raw materials, climatic conditions, the function of the structure, the nature of the settlement, and the size and permanence of the resident population.

In areas of the north European plain, first occupied in the preceding Late Glacial, there was a consolidation of settlement across this extensive lowland region, a trend already apparent toward the end of the Late Glacial, but one that became more marked with the onset of the Preboreal period after 10 radiocarbon ka BP (see [The Last Interglacial](#)). The archaeological record indicates more sites, greater use of local raw materials, a wider range of game animals, and a marked preference in site locations for lake-edge settings, with evidence for fishing and fowling as well as hunting. Star Carr in Britain, dated at about 9.5 radiocarbon ka BP is an early example of this pattern, with broad-blade industries typical of the early Mesolithic period, flaked stone axes, barbed harpoons of red deer antler, antler mattocks, and faunal remains of wild cattle, red deer, roe deer, elk, and boar, though fish are absent here ([Mellars and Dark, 1998](#)). The presence of a wooden paddle indicates the use of water transport, most probably dugouts. Similar sites are well documented in Denmark in association with the Maglemosean culture ([Blankholm, 2008](#)), and similar groupings are found in Poland, southern Sweden, the eastern Baltic and Russia ([Dolukhanov, 2008](#), [Zvelebil, 2008](#)). At the very end of this period, about 7.8 radiocarbon ka BP, the Arctic island of Zhokov in the East Siberian Sea was first visited when still accessible from the Siberian mainland, with a fauna dominated by reindeer and polar bear ([Pitul'ko, 1995](#)). Nuts, greens, roots, seeds, and fruits became more widely available and other resources such as birds and small mammals created more varied and diverse paleodiets, but in all cases seasonality evidence suggests seasonal occupations and relatively small social units of one or a few families, implying a large degree of residential mobility.

Maritime adaptations

The earliest use of marine resources is obscured by the lowered sea levels that persisted throughout the glacial period and the

removal or submergence of relevant evidence before sea level approximated its present position about 6 radiocarbon ka BP. This problem of differential visibility of evidence is especially acute on the shallower and broader stretches of continental shelf around the coastlines of the North Sea basin and northern France, but even on the steeply shelving coastlines of southern Europe, the shorelines of the glacial maximum would have been at least 5 km away, far enough to take most marine resources out of reach of sites located on the present-day coastline. Underwater survey and excavation are beginning to reveal new evidence, and it seems plausible to argue that late Paleolithic hunters were hunting seals and taking other marine resources along these now submerged coastlines from an early period (Fischer 1995, 1996; Flemming 2004).

Quantities of marine mollusk shells appear in the coastal cave deposits of northern Spain, such as La Riera and El Juyo, from as far back as the glacial maximum, with evidence for increased quantities in the Late Glacial period (Straus, 2008). Cueva de la Nerja in southeast Spain contains evidence of marine mollusks, fish, seal, and birds, associated with an early stage of the Late Glacial dating between about 14 and 12 radiocarbon ka BP (Morales et al. 1998). Given that shorelines were still at some distance from the present coastline, this is a significant indication of interest in marine resources and most probably only the visible tip of a more extensive pattern of marine exploitation (Bailey and Craighead, 2003). As shorelines approached their present position, thick shell middens with many millions of discarded mollusk shells begin to appear. The earliest group of such sites in the Iberian Peninsula is the Asturian shell middens, accumulated in rockshelter mouths on the north Spanish coast between about 9 and 6 radiocarbon ka BP. Here, as with the larger and more numerous shell middens that abound during the Late Mesolithic, hunting and gathering of resources on land continued alongside the collection of mollusks and other marine resources.

In the eastern Mediterranean, the occupants of the Franchthi Cave on the Argolid Peninsula of southern Greece were importing obsidian from the island of Melos 11 radiocarbon ka BP, which would have required a sea journey of 100 km by the direct route, or a series of sea crossings of at least 20 km by a more circuitous route. However, apart from a few limpet shells, evidence for the exploitation of marine resources does not appear in the Franchthi deposits until somewhat later in the sequence, most probably because the coastline was still too far away (Pluciennik, 2008).

A notable exception to this general picture of submergence and loss of coastal archaeology is the Norwegian coast, which has undergone substantial isostatic uplift following the removal of the Scandinavian ice sheet, so that 13-ka-old shorelines formed when the eustatic level of the sea was still about 50 m below the present are now elevated above present sea level along much of the Norwegian coastline.

Despite the opportunities for preservation of early coastal sites and the fact that the Norwegian coastline was ice free from an early period, there is no convincing evidence of human occupation, apart from stray and ambiguous finds, until about 10 radiocarbon ka BP, when numerous sites appear all along the Norwegian coastline in association with locally named Komsa and Fosna cultures characterized by large numbers of flake adzes and tanged points of Ahrensburgian type.

Preservation of organic materials is almost nonexistent, so that there are few indications of the animals exploited, but the location of the sites (some of them on offshore islands), the nature of the resources available, and the need for substantial intake of animal fats for human survival in these northerly regions indicate that marine resources and especially sea mammals must have been a major source of subsistence, while the flake adzes would have been well suited for scraping blubber from seal carcasses (Bjerck, 2008). The sites on the three large islands of Vega, some 20 km offshore, provide a good example. These were occupied from the earliest period, and by about 8.5 radiocarbon ka BP comprised an interrelated network of some 12 settlements scattered across the islands. The largest site of Åsgarden was associated with the best natural harbor and had evidence of solid circular dwelling structures with foundations partially excavated into the subsoil ('pit dwellings'), stone or turf walls, and a wide range of artifacts suggesting a residential base. The other sites were smaller with a narrower range of tool types indicating shorter-lived occupations for specialist activities.

The absence of earlier sites on the Norwegian coast may be due to the lack of technology or skills to make the seaworthy boats that would have been essential for sea travel in the region. Or it may be that the resource options available for human survival were too narrow and too risky to encourage settlement until the retreat of the Scandinavian ice sheet and the expansion of the Gulf Stream created more productive conditions on land and at sea.

At about the same time, during the Preboreal period, the British Isles also witnessed a new phase of expansion northward and westward beyond the lowland regions of England, with a number of new settlements appearing around the coastlines of Scotland and in the Western Isles, dating from about 9.5 radiocarbon ka BP (Tolan-Smith, 2008). Ireland, which had never had a passable land connection to mainland Britain, was also occupied for the first time. As in Norway, the appearance of the Gulf Stream would have been a significant attraction to the new settlement. Seaworthy boats would have been essential for navigating the Western Isles and especially for the crossing to Ireland, regarded as one of the most treacherous sea crossings in the British Isles. Sturdy dwelling structures would also have been essential in the harsher winter climates of the north. Faunal preservation on many of these sites is poor, but Mt. Sandel in Northern Ireland has preserved evidence of salmon fishing and boar hunting, postholes indicating substantial dwelling structures and pits that could have been used for food storage. The coastal site of Howick on the North Sea coast of Northumberland in northern England belongs to this phase of expansion, and has yielded traces of a substantial timber-framed dwelling and evidence of sealing and boar hunting.

The Late Mesolithic

During this period, sea levels stabilized in most parts of Europe at about their present position, resulting in the development of estuarine and inshore mudflats, and climate became slightly warmer and wetter – the Climatic Optimum (see Overview) – with temperatures 2–3 °C higher on the average than today's. These changes would have had a number of generally

beneficial effects on the productivity of coastlines, lakes, and rivers, and during this period we see the proliferation of large numbers of sites and settlements in these settings, many of them associated with permanent villages, food storage, cemeteries, and indications of social ranking. More specialized sites for hunting or gathering of plant foods associated with mobile settlement patterns continued in the less productive regions of Europe, or as outlying locations visited by hunting parties from residential bases elsewhere. Pollen evidence of woodland clearance in southern Sweden, southern Finland, and eastern Latvia (Zvelebil, 2008), and in upland regions of Britain (Tolan-Smith, 2008) is associated with deliberate clearance to improve the productivity of food-bearing shrubs and plants, either for direct human consumption or to concentrate animal resources. Freshwater or saltwater fish were a major source of protein, the domestic dog was widely in evidence, and pigs were commonly hunted, and may have been husbanded without necessarily involving biological domestication, the combination of fish, pigs, and nuts creating a powerful 'package' of resources capable of sustaining large communities.

The technological characteristics of this period represent a continuation or elaboration of existing trends. Among the geometric microliths, trapezes and rhomboid-shaped pieces replaced the earlier shapes across extensive territories. Pottery also came into regular use in some areas. The earliest examples occur between 9 and 7.3 radiocarbon ka BP in European Russia and ceramic containers were widely used from 6.5 radiocarbon ka BP in the Baltic and southern Scandinavia in late Mesolithic contexts.

Shell mounds

Coastlines are associated with shell mounds of varying size and number, the largest and best known grouping being the Ertebølle in Denmark, dated between about 6.5 and 5 radiocarbon ka BP, the largest mound, as at the type site of Ertebølle itself, being several hundred meters long and several meters thick. These sites were built up slowly over many centuries of repeated use and discard of oyster shells and other refuse, and over 400 shell mounds have been recorded. The larger sites were residential bases with a full range of artifacts including microlithic arrow barbs and flaked axes, an antler and bone industry, and pottery jars and bowls. Food remains include sea and land mammals, fish, and fowl (Blankholm, 2008). Marine resources were most probably the major source of protein, among which fish rather than shellfish were the major food item, a conclusion reinforced by stable-isotope analyses of human bones recovered from these sites. Coastal sites without mollusk shells are just as common and probably played the same role in the settlement system, occurring on shorelines lacking local supplies of oysters. Some of the coastal settlements of this period have been submerged in shallow water because of isostatic submergence, resulting in spectacular conditions of organic preservation, notably at the underwater settlement of Tybrind Vig, which produced a dugout, a paddle richly incised with geometric designs, and large numbers of wooden stakes associated with a landing stage and a fish weir. Many other submerged sites have been discovered in Danish waters, including other examples of fish weirs, suggesting a major investment in communal facilities.

Other well-studied concentrations of shell mounds are those on the small island of Oronsay in the Western Islands of Scotland, used between about 6 and 5.5 radiocarbon ka BP for seal hunting, fishing, and the collection of limpets – perhaps for fish bait (Tolan-Smith, 2008), and the Portuguese shell mounds on the Muge and Sado tributaries of the Tagus estuary (Straus, 2008). Around the Mediterranean coastline, fishing and shellgathering were combined with gathering of plant foods and hunting of mammals, supporting year-round occupations in some cases, notably at Franchthi Cave in Greece and the Uzzo Cave in Sicily, while the offshore islands of Mallorca, Corsica, Sardinia, and Cyprus were occupied or visited (Pluciennik, 2008). The apparent 'explosion' in the numbers of these coastal sites may be misleading, given that earlier coastlines are now mostly submerged in much deeper water.

Lake and river settlements

Elsewhere in Europe, large settlements have been recorded on the edges of lakes and rivers. Notable examples are Tågerup in Sweden (8–6.5 radiocarbon ka BP), with timber dwellings, wooden jetties for boats, and basketry used in trapping and carrying fish. On the Latvian coast, Sārņate (7–4 radiocarbon ka BP) has wooden houses with evidence of seal hunting, fishing, and large-scale processing of water chestnuts. In N. Finland, Kierikki (4.8–3.8 radiocarbon ka BP) on the estuary of the River Li revealed more than 300 house floors of pit dwellings, very large pottery jars used for storing seal oil, a large stone structure 50 m × 30 m suggesting a communal or ritual purpose, and evidence of long-distance trade in flint, greenstone, amber, and copper (Zvelebil, 2008). In the Iron Gates region on the middle reaches of the Danube River is a group of sites dated between about 8 and 7 radiocarbon ka BP, such as Lepenski Vir, Vlasac, and Schela Cladovei, with trapezoidal dwelling structures and faunal remains including deer, dogs, fish, freshwater mollusks, and land snails (Bonsall, 2008). Fish include the migratory sturgeon, which can grow to an individual weight of 150 kg, and stable isotope analyses of human bones indicate a major dependence on fish protein.

Burials and rock art

Burials are common, either in intimate association with domestic dwellings or as separately demarcated cemetery areas. The earliest example is at Aveline's Hole, a cave in the Cheddar Gorge of southwest England, with a date of ca. 9.1 radiocarbon ka BP, with more than 70 human burials, interpreted as the cemetery of a larger social group (Tolan-Smith, 2008). Olenii Ostrov on an island in Lake Onega in Karelia, dated at about 7.5 radiocarbon ka BP, had over 300 burials. At Zvejnieki in Latvia, 315 burials have been excavated indicating repeated use between 8.2 and 4.2 radiocarbon yr BP. Grave goods and the position of the skeletons indicate a variety of burial rites and evidence of social ranking based on age, gender, personal wealth, clan membership, and the presence of shamans (Zvelebil, 2008). The Portuguese shell mounds of the Muge River are also estimated to have contained at least 300 burials (Straus, 2008). In Denmark, human remains are rare in the Ertebølle shell mounds but 18 burials were found at the coastal settlement of Vedbaek. At Skateholm in southern Sweden, also belonging to the Ertebølle culture, a total of 85 burials

were recovered including evidence of a wooden mortuary structure (Blankholm, 2006). There are many instances of violent deaths, suggesting conflict associated with increased competition or territoriality.

Rock art is commonly regarded as absent in the Mesolithic period, an absence taken as evidence of cultural impoverishment compared with the Upper Paleolithic. However, rock art is in fact quite common, as are decorative artifacts, but concentrated in the new centers of population growth located in different regions from their Upper Paleolithic predecessors, providing additional insights into ritual and symbolism. Well-dated rock art is present in northern Norway by about 9 ka BP with naturalistic representations of elk, reindeer, bear, whale, and seal. Many hundreds of sites have been recorded in Norway and Sweden with images of anthropomorphs, deer, boats, sea mammals, bear, water birds, fish, reptiles, tracks, hunting and fishing implements, and abstract designs. One of the largest such concentrations is the site of Nämforsen (5.5–3.5 radiocarbon ka BP) in Swedish Norland, which was a major focus of ritual and seasonal aggregation for settlements in the surrounding region. At the other end of Europe is the Levantine rock art of eastern Spain, with schematic and naturalistic representations including hunting scenes and humans, and the Addaura Cave in Sicily (Pluciennik, 2008).

Conclusion

The major driving force behind these postglacial adaptations was the changing environmental conditions associated with deglaciation and climatic amelioration, which created new territories for human settlement, new supplies of food and raw materials for subsistence, new technological demands for tools, shelter, and transport, and new opportunities for population growth. These changes resulted in major reconfigurations of social geography, with expansion of human populations into new territory and the growth of new centers of population density, often associated with permanent ‘villages’ repeatedly occupied on a year-round basis by large communities, or with aggregation sites that formed a seasonal focus for more widely dispersed populations. The more ecologically productive environments capable of supporting higher population densities were often associated with a rich ritual and symbolic life and evidence of social differentiation. The contrasts often drawn with the Upper Paleolithic period thus reflect changes in the geographical distribution of population centers and the occupation of ecologically productive regions that were unavailable during the Last Glacial period, rather than an overall advance in social or economic complexity. It was these closing millennia of the hunting, fishing, and gathering way of life, in their turn, that witnessed the emergence and synthesis of the technological, economic, and social capabilities on which the later developments of agriculture, urban civilization, and long-distance trade were founded.

References

- Bailey G and Craighead A (2003) Late Pleistocene and Early Holocene coastal palaeoeconomies: A reconsideration of the molluscan evidence from Northern Spain. *Geoarchaeology: An International Journal* 18(2): 175–204.
- Bailey G and Spikins P (eds.) (2008) *Mesolithic Europe*. Cambridge: Cambridge University Press.
- Bjerck H (2008) Norwegian Mesolithic trends: A review. In: Bailey G and Spikins P (eds.) *Mesolithic Europe*. Cambridge University Press, Cambridge.
- Blankholm HP (2008) Southern Scandinavia. In: Bailey G and Spikins P (eds.) *Mesolithic Europe*. Cambridge University Press, Cambridge.
- Bonsall C (2008) The Mesolithic of the iron gates. In: Bailey G and Spikins P (eds.) *Mesolithic Europe*. Cambridge University Press, Cambridge.
- Dillehay TD (2000) *The Settlement of the Americas. A New Prehistory*. New York: Basic Books.
- Dolukhanov P (2008) The Mesolithic of European Russia, Byelorussia and the Ukraine. In: Bailey G and Spikins P (eds.) *Mesolithic Europe*. Cambridge University Press, Cambridge.
- Fiedel SJ (1987) *Prehistory of the Americas*. Cambridge: Cambridge University Press.
- Fischer A (ed.) (1995) *Man and Sea in the Mesolithic: Coastal Settlement Above and Below Present Sea Level*. Oxford: Oxbow.
- Fischer A (1996) At the border of human habitat: The late Palaeolithic and early Mesolithic in Scandinavia. In: Larsson L (ed.) *The Earliest Settlement of Scandinavia and Its Relationship with Neighbouring Areas, Acta Archaeologica Lundensia, Series In 8, 24*, pp. 157–176. Stockholm: Almqvist and Wiksell.
- Flemming N (ed.) (2004) *Submarine Prehistoric Archaeology of the North Sea: Research Priorities and Collaboration with Industry*, CBA Research Report 141, London.
- Habu J (2004) *Ancient Jomon Japan*. Cambridge: Cambridge University Press.
- Kuzmin YV, Jull AJT, Burr GS, O'Malley JM (2004) The timing of pottery origins in the Russian Far East: ¹⁴C chronology of the earliest Neolithic complexes. In: Higham T, Bronk-Ramsey C, and Owen C (eds.) *Radiocarbon and Archaeology: Proceedings of the Fourth International Symposium, Oxford 2002, Monograph*, pp. 153–159. Oxford: Oxford University School of Archaeology.
- Larsson L, Kindgren H, Knutsson K, Loeffler D, and Åkerlund A (eds.) (2003) Mesolithic on the move: papers presented at the Sixth International Conference on the Mesolithic in Europe, Stockholm 2000, Oxford: Oxbow.
- Mellars P and Dark P (1998) *Star Carr in Context*. Cambridge: McDonald Institute for Archaeological Research.
- Mithen S (2003) *After the Ice*. London: Weidenfeld and Nicolson.
- Morales A, Roselló E, and Hernández F (1998) Late upper Palaeolithic subsistence strategies in southern Iberia: Tardiglacial faunas from Cueva de Nerja (Málaga, Spain). *European Journal of Archaeology* 1(1): 9–50.
- Pitul'ko V (1995) High arctic Mesolithic culture: Man and Environment. In: Fischer A (ed.) *Man and Sea in the Mesolithic: Coastal Settlement Above and Below Present Sea Level*, pp. 351–359. Oxford: Oxbow.
- Pluciennik M (2008) The coastal Mesolithic of the European Mediterranean. In: Bailey G and Spikins P (eds.) *Mesolithic Europe*. Cambridge University Press, Cambridge.
- Straus LG (2008) The Mesolithic of Atlantic Iberia. In: Bailey G and Spikins P (eds.) *Mesolithic Europe*. Cambridge University Press, Cambridge.
- Tolan-Smith C (2008) Mesolithic Britain. In: Bailey G and Spikins P (eds.) *Mesolithic Europe*. Cambridge University Press, Cambridge.
- Tolan-Smith C and Bonsall C (1999) Stone Age studies in the British Isles: The impact of accelerator dating. In: Evin J, Oberlin C, Daugas J-P, and Salles J-F (eds.) *14C et Archéologie; Actes du 3ème congrès international (Lyon, 6–10 avril 1998)*. Paris: Mémoires de la Société Préhistorique Française 26, 1999 et Supplément 1999 de la Revue d'Archéométrie, pp. 249–257.
- Vigne J-D and Desse-Berset N (1995) The exploitation of animal resources in the Mediterranean Islands during the Pre-Neolithic: The example of Corsica. In: Fischer A (ed.) *Man and Sea in the Mesolithic: Coastal Settlement Above and Below Present Sea Level*, pp. 309–318. Oxford: Oxbow.
- Zvelebil M (2008) Innovative hunter-gatherers: The Mesolithic in the Baltic. In: Bailey G and Spikins P (eds.) *Mesolithic Europe*. Cambridge University Press, Cambridge.

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B

BEETLE RECORDS

Contents

Overview

Late Tertiary and Early Quaternary Records

Middle Pleistocene of Western Europe

Late Pleistocene of Australia

Late Pleistocene of Europe

Late Pleistocene of Japan

Late Pleistocene of North America

Late Pleistocene of South America

Late Pleistocene of New Zealand

Late Pleistocene of Northern Asia

Postglacial Europe

Postglacial North America

Overview

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Introduction

The study of fossil beetles (Coleoptera) from Quaternary deposits began in earnest in the 1950s in Britain, and has since spread to many parts of the world, most recently to Australia, New Zealand, and Japan. Fossil beetle research has led to many exciting breakthroughs in the understanding of the pace and intensity of climate change in terrestrial landscapes. In many ways, beetles are the ideal proxy for tracking terrestrial environmental change.

The Perfect Proxy?

If one were to design the ideal source of biological proxy for the reconstruction of Quaternary terrestrial environments, certain aspects of the nature of this ideal organism would have to conform to the following guidelines:

1. It must be sensitive to environmental change. Organisms that are able to withstand decades or centuries of adverse climate *in situ* before eventually dying out would be of far

less use than organisms that simply could not survive such adversity.

2. There should be great species diversity within the group of organisms to be studied. Greater diversity yields more precise and diverse information on paleoenvironments. Taxonomic groups with only a few dozen species necessarily offer less precise information about past environments than groups with many thousands of species. Each species has its own ecological niche, so by studying the fossil record of a group that represents thousands of ecological niches, one is able to piece together a broad spectrum of information about past environments (climate, soils, water quality, and vegetation).
3. It should be highly mobile, and thereby able to respond rapidly to environmental change by rapidly shifting its distribution into new areas where the environment is suitable. This quality ensures little or no problem with 'migration lag,' wherein an organism requires decades, centuries, or even millennia to become established in regions with potentially suitable environments.
4. It should have an abundant, well-preserved fossil record. This allows regional reconstructions to be built up for given

time intervals, based on multiple study sites containing assemblages of the fossil type in question. If the fossil remains of an organism are too scarce, or too poorly preserved to be identified, then their fossil record will contribute little to regional paleoenvironmental reconstructions.

5. It should be readily identifiable to the species level. Even if a group of organisms satisfy the first four criteria, their fossil record will be of limited value if their remains cannot be specifically identified. There is often a great deal of ecological variation among the species in a given genus, so a generic identification provides little that is useful to paleoenvironmental reconstruction.
6. It should have many species with narrow ecological or climatic tolerances. Again, this harkens back to the issue of specificity. The more specific the environmental requirements of the organism, the more specific the reconstruction that can be made, based on its presence in the fossil record. This criterion works together with criterion 2. If taxonomic groups with a great deal of species diversity also contain many species with very narrow ecological or climatic tolerances, then a complex mosaic image of the paleoenvironment can be constructed.
7. None of the above criteria are of much use unless the taxonomic group has remained extant throughout the study period one is interested in. If the group of organisms has gone extinct, no matter how diverse, abundant, or well-preserved, then only educated guesses can be made about the ecological or climatic requirements of the ancestral forms. The modern ecological and climatic data available for the surviving species in such a group of organisms can never be fully applied to extinct species.
8. Finally, it is a great help to the paleoecologist if the proxy group of organisms is well known to science. In other words, if a group of organisms has been well studied in recent times, then paleoecologists have at their disposal a great body of data, built up through many generations of scientific study, concerning the modern ranges and ecological requirements of the species.

The reason that beetles (Coleoptera) are so useful as proxy indicators of past terrestrial environments is that they fulfill all of these requirements. Beetles are the most diverse group of organisms on Earth, with more than 1 million species known to science. A large proportion of beetle species are known to be quite sensitive to environmental change, and to shift their distributions across continents in order to become established in regions of suitable environment (Elias, 1994). As discussed in this section of the encyclopedia, beetles have a well-preserved, abundant fossil record in many regions of the world. Beetle exoskeletons are reinforced with chitin, a highly durable compound that resists decay in waterlogged and other anoxic sedimentary environments. That chitinous exoskeleton is covered with taxonomically useful characters on the main body parts, including the head capsule, thoracic shield (pronotum), and wing covers (elytra). In many study regions, paleoentomologists are able to specifically identify upwards of 50% of these fossil beetle sclerites; in some well-studied regions, this percentage climbs closer to 75%. Quaternary beetle faunal assemblages frequently contain hundreds of identified species, more than any other group of organisms preserved in Quaternary deposits.

Many beetle species are stenothermic, adapted to a narrow range of temperatures (Figure 1). Predatory and scavenging families of beetles comprise less than half the total number of families, but their species' numbers are often proportionally high in fossil assemblages. These are the two groups that receive the most attention in paleoclimatic reconstructions. The rationale behind this decision is that predators and scavengers are not tied to particular species of host plants, and so are able to become established most rapidly in new regions in response to changing climates. For instance, the Pleistocene fossil record indicates that these groups have colonized newly deglaciated landscapes within a few years of ice-margin retreat (Coope, 1977). Perhaps because of beetle mobility in the face of changing environments, the fossil evidence strongly indicates that nearly all species found in Quaternary assemblages remain extant today. It appears likely that beetle populations never remained genetically isolated for sufficient lengths of time in the last 2.6 My to develop new species. The constant mixing of gene pools through shifting distributions, in response to 50+ glacial-interglacial cycles, prevented this from happening (Coope, 1978). Finally, beetles are some of the best-studied groups of insects, with modern ecological requirements and ranges well understood for many species living in the temperate and high latitude regions (where the vast majority of fossil investigations have taken place).

Where Fossil Beetles Are Found

Fossil beetle assemblages have been recovered from a wide variety of sedimentary environments, especially anoxic waterlain sediments that concentrate the remains in layers of organic detritus. Lacustrine (lake and pond) sediments have yielded abundant, diverse assemblages of fossil beetles, especially in deposits from the littoral zone and where a stream enters the lake or pond (Figure 2). Such deltaic deposits can yield astonishing numbers of fossil beetle specimens. Fluvial sediments also yield fossil beetles. The most productive type of fluvial deposit is an accumulation of organic detritus, laid down in secondary channel bends, backflows, and pools between riffles (Figure 3). Bogs, fens, and mires can also yield good accumulations of fossil beetles, especially in locations representing the edge of such bodies, where there is increased input from the upland beetle fauna (Figure 4). One of the more unusual types of deposit yielding beetle remains are the middens of certain rodents, including packrats or woodrats (*Neotoma*) – the native group of North American rats. Their middens (Figure 5), intermittently wet and sticky from rat urine and feces, serve as traps for beetles entering the shallow caves and rock shelters of desert and semiarid regions where these rats thrive (Elias, 1990). Unlike the waterlogged sediments of lakes, rivers, and bogs, packrat middens preserve beetle (and other arthropod) exoskeletons in extremely dry conditions, sometimes mummifying their soft tissues.

Methods of Sampling and Extraction

When sampling the kinds of sediments discussed above, the aim is to obtain a sufficient quantity of material to yield at least a liter of organic detritus. The details of sampling and

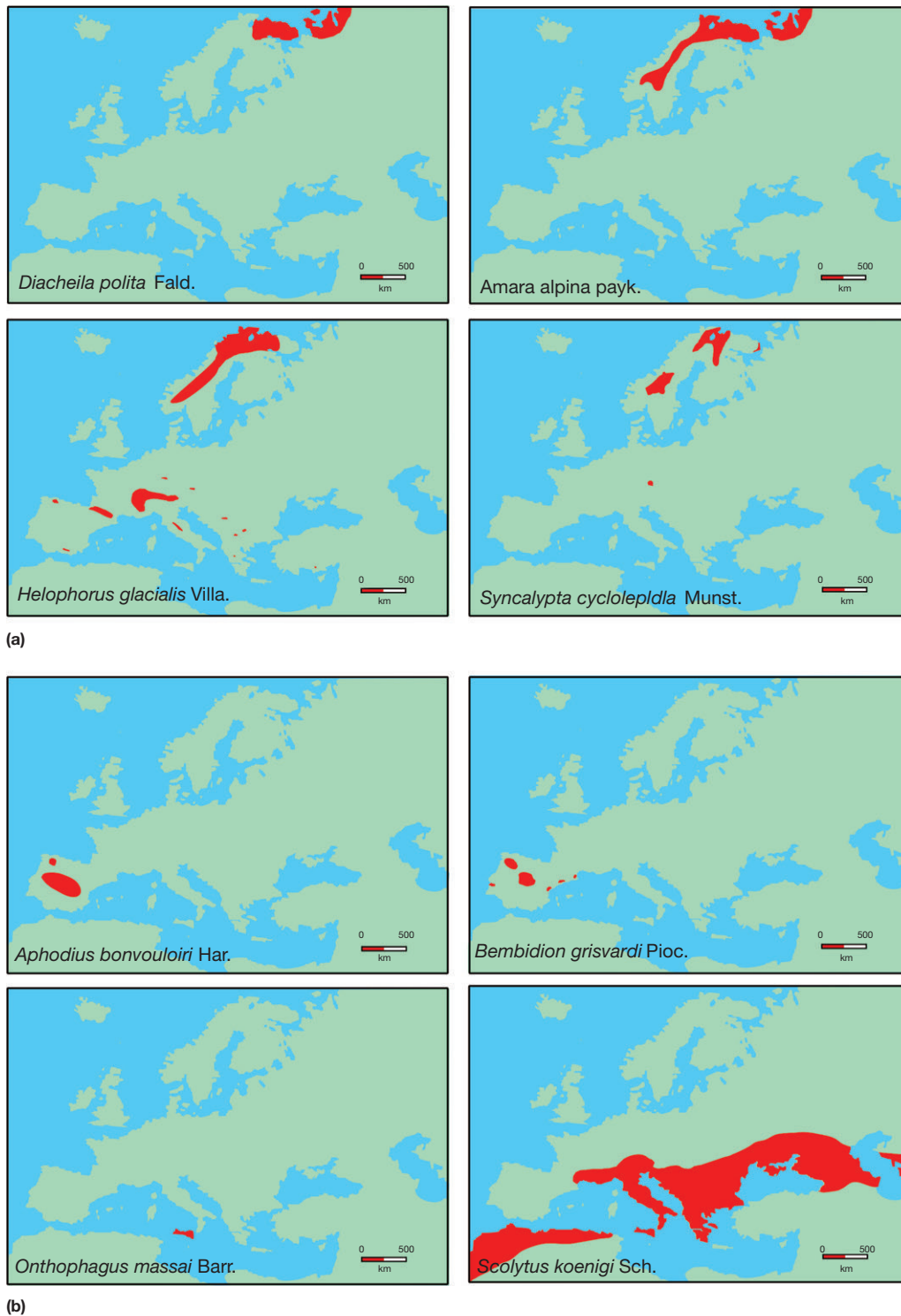


Figure 1 Modern European distributions of stenothermic beetle species found in British Pleistocene deposits. (a) Cold-adapted species and (b) warm-adapted species.

extraction procedures are discussed in [Elias \(1994, 2010\)](#). In lacustrine sediments that are only 3–5% organic detritus by volume, this requires the extraction of many kilograms of sediment, followed by wet-screening to isolate the organic fraction. Generally, the quantity of sediment obtainable by

piston coring of lake and pond sediments is insufficient to yield adequate amounts of detritus. While some success has been achieved by taking multiple, large diameter cores, this is not often practical. The best results are obtained by sampling exposures of sediments, either natural or man-made. Natural



Figure 2 Deltaic deposit exposed at Lake Isabelle, Colorado. The layers of organic detritus yielded thousands of fossil beetle specimens. Photograph by the author.

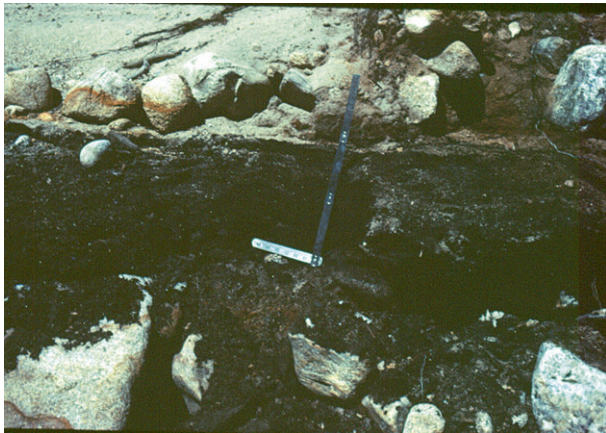


Figure 3 Lens of organic detritus exposed after a flood on the Roaring River, Rocky Mountain National Park, Colorado. This late Holocene deposit contained a rich, diverse beetle fauna. Photograph by the author.

exposures include cutbanks of streams and ocean and lake bluffs. Man-made exposures include gravel and clay pits, irrigation ditches, trenches, and building sites.

The sequence of steps in extraction and mounting of insect fossils is summarized in [Figure 6](#). Fossil insect extraction is



Figure 4 Holocene peat from a high mountain bog at La Poudre Pass, Colorado, exposed along an irrigation canal. Photograph by the author.



Figure 5 Packrat (*Neotoma*) midden sample from a rock shelter at Emery Falls, Grand Canyon, Arizona. Photograph by Thomas Van Devender.

relatively safe, cheap, and easy. The only lengthy process that may be involved is the pretreatment of samples to disaggregate the organic detritus from inorganic matrix (such as calcareous sediments and clays), or to soften and then disperse felted peats or lignite. Processing procedures are largely a matter of personal preference and availability of equipment, and some workers use a slightly different procedure than the one discussed below. For instance, Russian workers sometimes prefer to skip the kerosene flotation step discussed below, and sort through entire samples of washed organic detritus, in order to

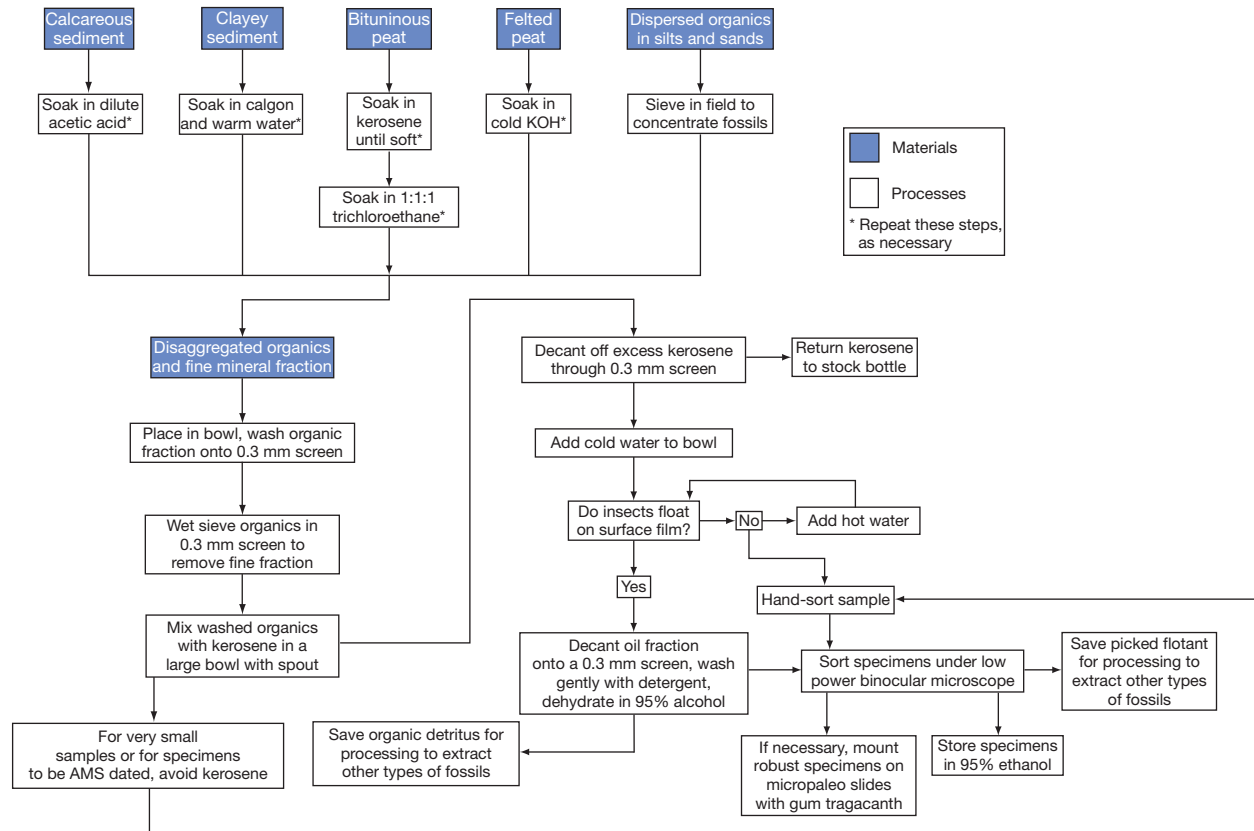


Figure 6 Flow chart of fossil beetle sample processing for kerosene flotation. Reproduced from Elias SA (1994) *Quaternary Insects and Their Environments*. Washington, DC: Smithsonian Institution Press.

maximize the recovery of such heavy-bodied groups as weevils (Curculionidae) that sometimes do not rise to the top during flotation. Japanese workers studying peat deposits prefer to split the individual layers of peat and examine each layer under a microscope.

Once disaggregated organic detritus is obtained, the next step is to wet screen the sample in a 300 μ m sieve. This process removes fine particles, such as silt, that may fill the concavities of rounded insect sclerites. Once the detritus has been screened, the residual material is placed, still damp, in a large bowl with a spout, or a rectangular dishpan, and processed by kerosene flotation to isolate and concentrate insect fossils. Kerosene or other lightweight oil is added to cover the sample, and gently worked into the sample by hand for several minutes. The oil adheres to the insect sclerites but not to plant detritus. The remaining kerosene is decanted from the bowl and cold water is vigorously added to the oily detritus. In most samples, nearly all of the insect sclerites will rise to the top, and float at the oil–water interface. Within 15–60 min, most plant residue sinks to the bottom of the bowl, and the now-concentrated insects may be decanted onto a screen, to be washed gently in detergent and dehydrated in 95% ethanol before microscopic sorting. Small samples should be sorted completely, rather than processed with kerosene. Also, samples that may need to be submitted for radiocarbon dating should not be exposed to kerosene.

Specimen sorting is done under low power (10 $\times$) binocular microscope in alcohol or water. Specimens may be stored in

vials of ethanol, or more robust specimens may be mounted with gum tragacanth (a water-soluble glue) onto micropaleontological slides. Identification of fossil specimens is usually done through comparison with modern identified specimens (often from museum collections). This is another potentially lengthy process.

Exoskeletal Parts Preserved in Quaternary Deposits

The chitinous remains of beetles often preserve extremely well in Quaternary sediments, and can be matched exactly with modern specimens. The principal sclerites that are used in the identification of Quaternary beetles are the head, pronotum, and elytra (Figure 7). The male genitalia (aedeagi) are likewise reinforced with chitin, and are sometimes recovered from compressed groups of abdominal segments found in fossil assemblages. The aedeagus is often the most diagnostic sclerite in the identification of species, especially in genera that contain dozens or hundreds of species.

Useful Characters

The process of identification begins with the shape and size of the main exoskeletal parts. Often times colors derived from pigments are preserved, and color patterns can be quite useful in the identification of groups that have spots, stripes, and other noticeable color variations on the elytra or pronota.

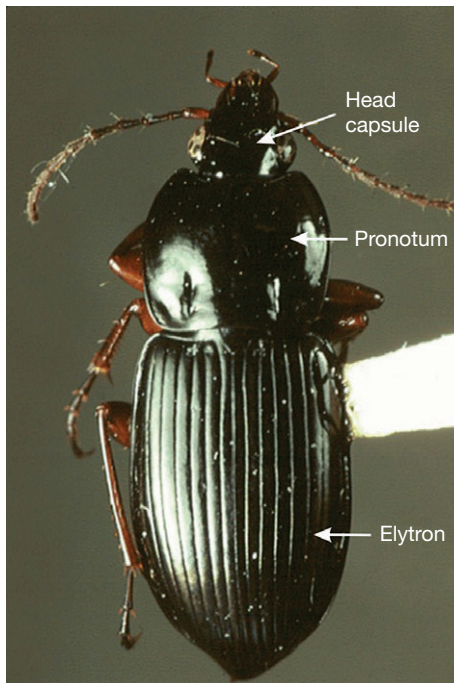


Figure 7 The ground beetle, *Pterostichus leconteianus*, with principal body parts found in fossil assemblages identified. Photograph by the author.

Metallic coloration preserves extremely well because it is based on surface microsculpture, rather than pigments. The depth, shape, and distribution of such surface features as striae, punctures, and microsculpture are all potentially diagnostic characters, as are the presence and placement of scales and setae in some groups. As discussed above, the size and shape of the aedeagus is one of the most diagnostic characters for beetles, although these are not as commonly found as heads, pronota, and elytra.

Basic Assumptions for Use in Paleoecology

There are several basic assumptions made in the reconstruction of paleoenvironments, based on fossil beetle assemblages. These assumptions fall within the framework of uniformitarianism in the sense that it is assumed that ancient beetle populations had the same environmental tolerances and behavior patterns as their modern counterparts.

Species Constancy Throughout the Quaternary

There are three lines of evidence that support the assumption that the beetle species found in Quaternary deposits are the same species that live today. The first is that the morphology of their exoskeletons has remained constant. In other words, no differences can be found between the fossil specimens and their modern counterparts that fall outside of the boundaries of known variation within the modern species. The youngest beetle fossils that exhibit morphological variations outside the bounds of modern species come from fossil assemblages that date to the Pliocene/Pleistocene boundary, approximately 2.6 Ma. However, there are many reliably identified beetle

species from Pliocene assemblages that exactly match modern species. A full discussion of this topic can be found elsewhere in this encyclopedia. The second line of evidence for Quaternary species constancy comes from the stability of shape and size of fossil aedeagae, as discussed in section 'Exoskeletal Parts Preserved in Quaternary Deposits,' above.

The third line of evidence deals with the issue of constancy of physiology, as expressed in environmental tolerances. This issue can only be dealt with indirectly, as the exoskeletal evidence does not speak to it. However, what has been repeatedly observed in fossil beetle assemblages from sites throughout the world and ranging in age from early to late Quaternary times, is that the species found together in fossil assemblages remain ecologically compatible today, even if they no longer live in the same geographic regions. For instance, a fossil assemblage associated with Pleistocene periglacial environments may contain a mixture of species found today in arctic and alpine habitats, in various parts of the Northern Hemisphere. Conversely, an interglacial fauna may be found to contain large numbers of warm-adapted beetle species that today are found in a variety of temperate or subtropical habitats in various regions. In order for this persistence of ecological compatibility to occur, one of two scenarios must be invoked. Either whole suites of species have evolved unidirectionally during the Quaternary, so that their past environmental tolerances have changed together, or their environmental tolerances have not changed at all. The latter scenario seems by far the likeliest, as evolutionary theory indicates that each species evolves separately from others, in response to its own unique set of selective pressures (Elias, 1994).

Usefulness of Beetles as Paleoclimate Proxies

Insect ecologists support the idea that insect abundance and diversity may be controlled by biotic factors (e.g., predators, competition) in the central part of a species' range, but that climatic factors probably limit populations toward the edges of these ranges. Oxygen isotope records from deep sea sediments and polar ice cores have shown that large-scale, climatic change was practically a constant feature of Pleistocene environments, it would appear that the shifting of insect distributions in response to these changes would have placed a large proportion of populations in this 'abiotic factor' zone of their respective ranges, during much of the Pleistocene.

Thermal Tolerances of Beetles

The modern distribution patterns of many beetle species in the temperate and high latitudes have been shown to coincide closely with climatic zones. The distribution of individual species reflects temperature regime, especially summer warmth and degree of seasonality (i.e., annual temperature range). Stenothermic species with limited thermal tolerances are found in most fossil assemblages (Figure 1). The physiology of stenothermic species is closely tied to their thermal requirements. Experiments on metabolic rates of ground beetles (Carabidae; Thiele, 1977) showed that beetles' optimal metabolisms correspond with their thermal tolerance zones. This phenomenon is strengthened during the process of cold-

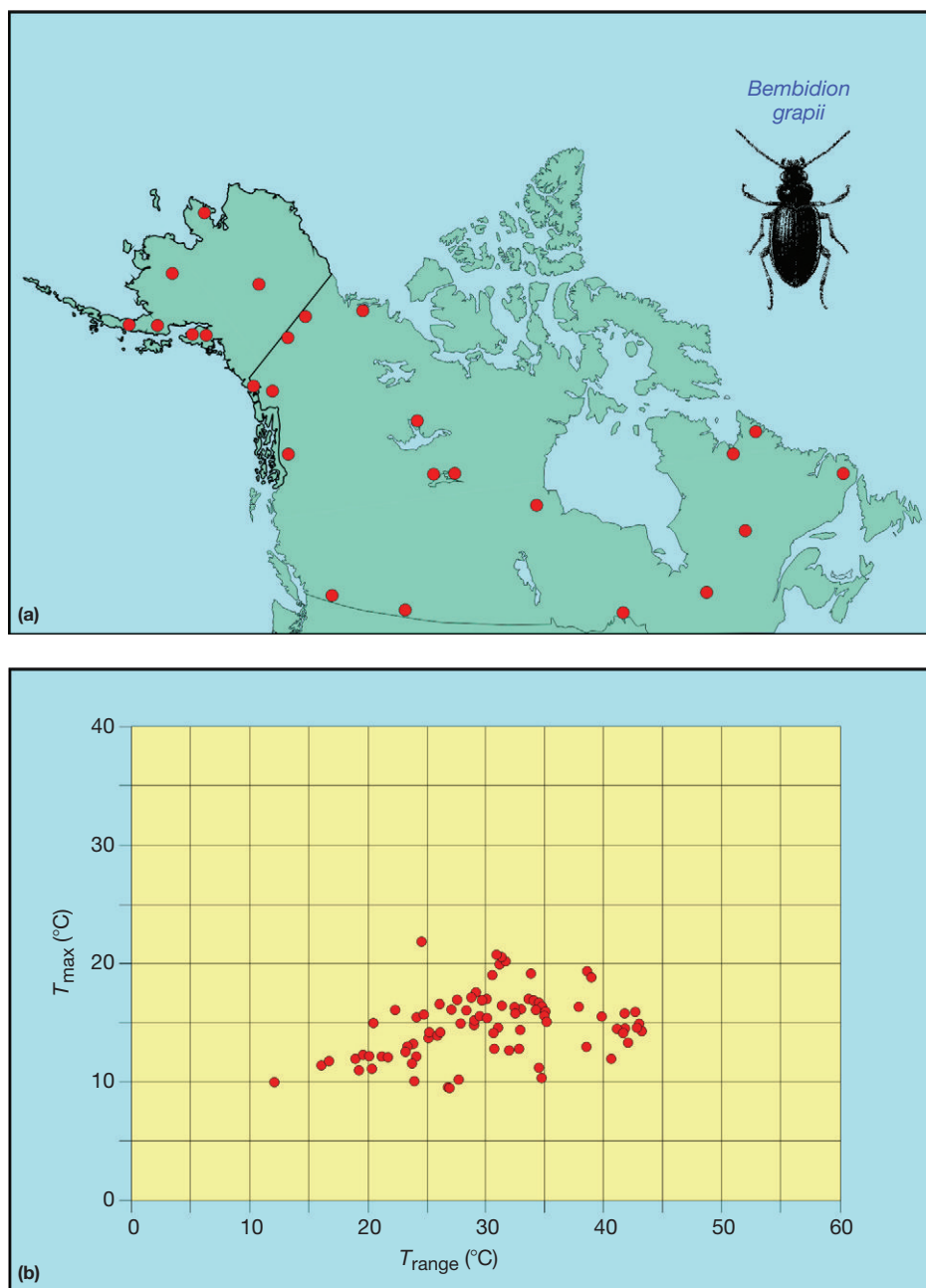


Figure 8 (a) Modern North American distribution of the ground beetle, *Bembidion grapii*, and (b) species climate envelope for *B. grapii*, based on climatic parameters (T_{max} and T_{range}) of collecting localities shown in (a).

hardening, which takes place in late summer or fall. In preparation for winter cold, ground beetles in northern regions become progressively cold-adapted (able to function at cold temperatures). Under these conditions, the beetles' metabolic rates peak at low temperatures, and they may become paralyzed or die if exposed to the temperatures they experience in mid-summer.

Mutual Climatic Range Analysis of Fossil Assemblages

The mutual climatic range (MCR) method of paleoclimate reconstruction is based on the concept that the proxy

organisms (in this case, beetles) live within the bounds of a set of climatic parameters, as defined by the climatic conditions within their geographic range. Thus beetles live in a quantifiable piece of climate space. In the MCR method, the first step is to develop a climate envelope for the species of predators and scavengers that are found in the fossil assemblage (Figure 8). The full method is described in Atkinson et al. (1986) for European studies, and by Elias et al. (1996) for North American studies. These envelopes typically are defined on the basis of the mean July and mean January temperatures of the locations where the species are known to occur today. The climatic parameters of each of these localities were plotted

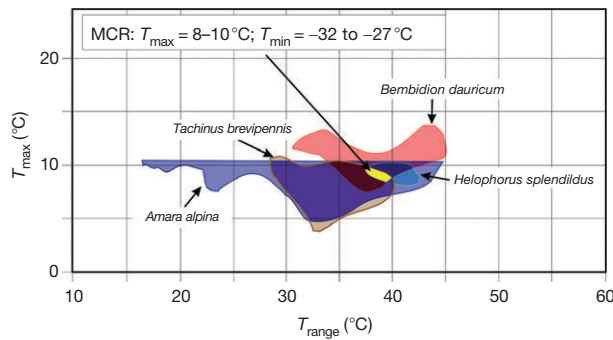


Figure 9 An example of the application of the mutual climatic range (MCR) method, showing the overlap of the climate envelopes of four species of cold-adapted beetles. In nearly all cases, far more species are involved in MCR reconstructions. The number is limited here, for the sake of visual clarity.

on diagrams of mean July temperature ($T_{\max}$) versus the difference between the mean July and mean January temperature (T_{range}). Once the individual species' climate envelopes are created, then the envelopes for the species found together in a fossil assemblage are put together to find the area of overlap in climate space. This area is the MCR of the faunal assemblage. In the example shown in [Figure 9](#), the climate envelopes of just four cold-adapted species are combined to produce the MCR, shown in bright yellow.

European and North American MCR studies have focused on species that are predators or scavengers. Plant-feeding species have been avoided, because of the possibility that the migration rate of the host plants of these beetles may lag behind climatic change. For instance, in recently deglaciated landscapes, the process of ecological succession from bare mineral soil to mature forest stands may take many centuries to achieve. Predators and scavengers are able to exploit such landscapes as part of the earliest pioneering biological communities, as soon as the climate is suitable ([Elias, 1994](#)).

Modern tests of the reliability of the MCR method have been performed for the European and North American beetle faunas. These tests compared predicted with observed temperatures for a broad spectrum of sites, using the MCR of species found today at localities to predict modern temperatures of those localities, and comparing these predictions with modern climatological data for the same sites. For 35 sites in North America, a linear regression of actual versus predicted $T_{\max}$ values yielded an r^2 value of 0.94. A regression of observed versus predicted $T_{\min}$ values yielded an r^2 value of 0.82. The standard deviation for $T_{\max}$ was 0.7 °C and for $T_{\min}$ was 10 °C. Similarly for Europe, a linear regression of actual versus predicted $T_{\max}$ values yielded an r^2 value of 0.88, and a regression of observed versus predicted $T_{\min}$ values yielded an r^2 value of 0.88. The standard deviation for $T_{\max}$ was 0.83 °C, and for $T_{\min}$ was 2.42 °C. Beetles living in regions with cold winters are poor indicators of winter air temperatures because they seek shelter from exposure to winter air through various means (e.g., burrowing in mud, leaf litter, under bark or stones).

The MCR method has allowed paleoentomologists to produce quantified paleotemperature estimates for both summer and winter seasons for hundreds of fossil assemblages, ranging in age from late Tertiary ([Elias and Matthews, 2002](#)) to the late

Holocene ([Lavoie and Arseneault, 2001](#)). Regional summaries of MCR reconstructions have been prepared for northwest Europe ([Coope, 1987](#); [Coope et al., 1998](#)), eastern and central North America ([Elias et al., 1996](#)), the Rocky Mountain region ([Elias, 1996](#)), and Alaska and the Yukon Territory of Canada ([Elias, 2001](#)). The main results of these studies are discussed elsewhere in this encyclopedia. The principal findings in both Europe and North America are that late Pleistocene climatic changes were rapid and often of large-scale. Interglacial climates were at least 3 °C warmer than today in some regions, but during the Last Glacial Maximum (LGM), $T_{\max}$ was depressed by 8–10 °C in most temperate regions. The LGM summer temperatures of the Arctic regions were only depressed by 3–5 °C in Beringia. $T_{\min}$ was depressed by as much as 25 °C in some mid-latitude regions, but $T_{\min}$ was within 1–2 °C of modern levels in most regions of eastern Beringia.

In New Zealand, collections of the modern beetle fauna are not sufficiently complete to allow species' climate envelopes to be constructed. As discussed elsewhere in this encyclopedia, Marra and colleagues have developed an alternative method, called the maximum likelihood envelope, that has allowed them to successfully reconstruct late Pleistocene climates. In Australia, [Porch \(2010\)](#) has developed another alternative approach, using bioclimatic envelopes to model both temperature and precipitation history. Another regional variation on the MCR method has been developed by Russian scientists, working on northeastern Siberian fossil beetle assemblages. They argue that certain plant-feeding beetle species should be included in MCR reconstructions, because they feed on plants such as sedges and mosses that remained ubiquitous in Siberia throughout the Pleistocene. Hence there should be no problem with plant migration lag under these circumstances ([Sher et al., 2005](#)).

Multifaceted Paleoenvironmental Reconstructions

The order Coleoptera is the most diverse group of organisms on Earth, with a known modern fauna exceeding 1 million species. This number is roughly four times that of all the flowering plant species (angiosperms), and more than 200 times the number of all mammal species. This incredible diversity carries through to Quaternary beetle faunal assemblages, which often contain between 100 and 200 species from dozens of families. While predators and scavengers provide reliable information on past climates, plant-feeding species serve as proxies for the reconstruction of past plant communities, and can provide such information as the relative maturity and health of forest stands, and the presence of downed, rotting trees and associated fungi. In some cases, fossil beetle evidence may provide 'smoking gun' evidence for an environmental change. Such was the case when fossil remains of the elm bark beetle, *Scolytus scolytus*, were found in British sediments that date to the time of the elm decline in the mid-Holocene. The decline may have been due to Dutch elm disease Dutch elm disease, which is transmitted by the beetle ([Girling, 1988](#)).

Lacustrine aquatic beetles provide useful information on the temperature, trophic status, substrates, and size of lakes and ponds. Stream-dwelling beetles provide information on

the size and velocity of streams, as well as temperature, trophic status, substrates, and degree of water clarity.

As in other macrofossil studies, the 'story' obtained from fossil beetle assemblages is a local one, and needs substantiating with numerous regional replicates to produce a valid paleoenvironmental synthesis. But because of the ubiquity of fossil beetle preservation in water-lain deposits, such regional syntheses are relatively easy (if time consuming) to perform.

Other Types of Studies

In addition to straightforward paleoenvironmental reconstructions, Quaternary beetle assemblages have also been used to shed light on various aspects of biological and environmental history. Beetle species longevity, combined with our ability to identify the species of many fossil specimens, allows us greater understanding of how these animals have dealt with the vicissitudes of Pleistocene environmental change.

Zoogeography

One application of fossil beetle data has been for zoogeographical studies. By plotting the shifts of beetle distributions through time and space, researchers have been able to reconstruct at least partial zoogeographic histories of various species. These studies have reinforced the theory that beetles have shifted their ranges repeatedly, and sometimes on very large spatial scales, in order to find regions of suitable climate. Such was the case for the tiny rove beetle, *Micropeplus hoogendorni*. As discussed elsewhere in this encyclopedia, this species was initially thought to have become extinct in the Pliocene or early Pleistocene, because it was only known from a 5.7 My-old deposit in Alaska. However, Coope found middle Pleistocene specimens of this beetle in Britain, and it has recently been found to be synonymous with a Siberian species, *M. dokuchaevi* (Elias, 1994).

The history of comings and goings of various groups of beetles across the Bering land bridge have been discussed by Elias et al. (2000), and the late Pleistocene and Holocene zoogeographic history of Chihuahuan Desert beetles was analyzed by Elias (1992). The fossil evidence used to develop these histories frequently disproves zoogeographic theories based solely on modern distributions (Elias, 1994). Fossil beetle evidence documents the importance of glacial refugia in maintaining high latitude faunas through glacial periods. Beringia was the most important refugium of the cold-adapted fauna in the Arctic. At the other end of the world, the beetle evidence suggests that ice-free regions of the Chilean coast formed a refuge for the cold-adapted fauna of South America.

Environmental Archaeology

There are two types of fossil beetle studies done in aid of environmental archaeology. The first is the reconstruction of natural environments associated with archaeological sites. This kind of work has mainly been done in North America, where prehistoric human populations were mainly hunter-gatherers who had relatively light impacts on the landscape. With a few

notable exceptions, these peoples built no permanent buildings, so they did not create any anthropogenic environments that attract warmth-loving insects incapable of surviving winter climate out-of-doors. This line of research is discussed elsewhere in this encyclopedia. This kind of research has also been applied to Paleolithic sites in Britain, where fossil beetle assemblages have shed light on middle and late Pleistocene environments were experienced by *Homo erectus* and *Homo neanderthalensis*.

The other type of environmental archaeological studies done with beetle (and other arthropod) remains concerns the reconstruction of anthropogenic environments, the houses, barns, and other man-made buildings where warmth-loving insects live as uninvited guests. This type of research is most common in Europe, where large-scale human disturbance of landscapes began by at least the beginning of the Bronze Age. In fact one of the difficulties of middle and late Holocene beetle studies in Europe is the inability to tease apart natural environmental change from anthropogenic change. This line of research is discussed more thoroughly elsewhere in this encyclopedia.

Emerging Research Areas

Great strides have recently been made in the study of the ratios of stable isotopes of hydrogen and oxygen in fossil insect chitin. As discussed above, chitin gives strength and rigidity to insect exoskeletons. In beetles, the exoskeleton forms as the adult insect emerges from the larva. As a long-chain nitrogenous polysaccharide molecule (a complex of sugar molecules and nitrogen), chitin incorporates both hydrogen and oxygen atoms in its matrix during formation. These atoms come from the environment – either through food eaten by the larva, or, in the case of water beetles, directly from the water in which the beetles live. As discussed in Gröcke et al. (2010), a series of modern studies have shown that the isotopic signature of hydrogen and oxygen in chitin (i.e., the deuterium-hydrogen (D/H) ratio and the ratio of ^{18}O - ^{16}O) closely matches the isotopic signature of precipitation in the regions where the beetles live. Furthermore, there is good evidence that once chitin molecules form, they are chemically inert. They do not exchange atoms (including hydrogen and oxygen) with the surrounding environment, either during the insect's lifetime, or after death.

The relationship between the D/H ratio of chitin and the D/H ratio of meteoric water in the organism's habitat has been demonstrated by several studies, using ladybird beetles (Coccinellidae) and beetles from various families (Miller et al., 1988). The hydrogen isotope composition of environmental water varies widely and systematically across the globe, and a strong and well-understood correlation exists between temperature and D/H ratios in precipitation (Rozanski et al., 1993). Because water is a key constituent of many chemical reactions in organisms, the isotope signature present in local environmental water is transferred to organisms that live in it. Because of the chemical stability of chitin, aquatic insects such as water beetles 'lock' this isotope signature in their exoskeletons. In lakes where evaporation is limited and water is in isotopic equilibrium with precipitation, water beetles thus record the isotopic signature of precipitation at the time of chitin synthesis (Gröcke et al., 2006), which makes water beetle chitin a

valuable source of climatic information. Because of its longevity in the fossil record, chitin can therefore provide scientists with an extremely powerful proxy for the reconstruction of ancient terrestrial climates.

Midge (Chironomidae) fly larvae are also aquatic, and recent studies have used stable isotope ratios from fossil midge larval remains to reconstruct past temperatures in Arctic lakes. Wang et al. (2009) demonstrated that about one third of hydrogen in chironomid larvae chitin is derived from the water in which they live. These results were obtained from laboratory studies using cultured specimens. Comparable culturing studies will need to be done to test this relationship in water beetles. Thus far, only a handful of studies have attempted to use isotopic data from insect exoskeletons to reconstruct climate (Gröcke et al., 2006; Miller et al., 1988; Wooler et al., 2004, 2008). Interest in this topic is likely to increase as more detailed knowledge of the geographic distribution of D/H from precipitation becomes available, and as advances in mass-spectrometry facilitate the rapid analysis of small samples. Much work still needs to be done to improve the methodology. For instance, more needs to be known about the influence of trophic levels, water loss, and food-derived water on D/H values of beetle chitin. Such studies have been done for laboratory-raised chironomids (Wang et al., 2009); these should also be undertaken on beetles.

Ancient DNA

In recent decades the study of ancient DNA, or aDNA, has been revolutionizing our view of the genetic history of many species of plant and animals. Ancient DNA is recovered using the polymerase chain reaction (PCR). This is a key technique in molecular genetics that permits the analysis of any short sequence of DNA without having to clone it. PCR is used to reproduce (amplify) selected sections of DNA. PCR makes a large number of copies of a gene. This is necessary to have enough starting template for sequencing. Sequencing is the process of determining the order of DNA nucleotides in a gene.

As discussed more fully in Elias (2010), the study of ancient DNA from Quaternary beetle remains began with Reiss' (2006) study of fossil insect DNA from the American Southwest. Thus far, aDNA studies from insects have focused either on late Holocene fossils, or on fossils preserved in permanently frozen sediments, or in extremely arid conditions, such as packrat (*Neotoma*) middens from the American Southwest. Both of these extreme environments tend to preserve DNA strands over long periods of time. Molecular genetic studies of fossil assemblages have the potential to test the widely held assumption (based on external morphology) that most, if not all beetle species have remained intact for hundreds of thousands (and in some cases, millions) of years (Coope, 1978; Elias, 1994).

In the study by Reiss (2006), mitochondrial DNA (mtDNA) was extracted from fossil beetle specimens identified by the author from his work in the Big Bend region of Texas. The fossils came from a packrat midden dated 23 900 cal years BP, and included fossil elytra from the weevil genus *Ophryastes*, the ground beetle genus *Harpalus*, and the dung beetle genus *Onthophagus*. After an initial failure to obtain aDNA from these specimens, Reiss eventually succeeded in getting a 316

base pair (bp) fragment from the fossil ground beetle (*Harpalus*) specimens.

Willerslev et al. (2007) demonstrated that aDNA preserves well in permanently frozen conditions. They investigated aDNA from samples of silt-rich ice taken from the Dye 3 and GRIP ice cores from Greenland, as well as from sediments sampled from the Kap København formation in northernmost Greenland. They obtained small fragments (97 bp) of mtDNA from the Dye 3 core samples, associated with the cytochrome oxidase I (COI) gene. The most reliable indication of insect mtDNA in the study was associated with Lepidoptera (moths and butterflies). Less substantial evidence was found for the recovery of COI gene sequences associated with beetles, flies, and spiders. Interestingly, no macroscopic remains of any of these arthropods were recovered from the ice-core samples, only their ancient DNA. Willerslev et al. estimated that their samples are between 800 000 and 450 000 years old, although they could not rule out a much younger age from the last interglacial interval (marine isotope stage (MIS) 5e). The Kap København sediments, thought to date from 2.5 to 2 Ma, failed to yield any identifiable fragments of DNA.

King et al. (2009) extracted aDNA from grain weevil (*Sitophilus granarius*) fossils from waterlogged archaeological sites in England. They selected six fossil specimens for aDNA extraction, and reported successful extraction of mtDNA from five of the six specimens. The modern and medieval specimens yielded fragments of various lengths from the COI gene. Two out of three Roman-age specimens yielded aDNA, but only shorter (98 bp) fragments of the COI gene.

The author is currently collaborating with Ian Barnes and their graduate student, Peter Heintzman (School of Biological Sciences, Royal Holloway, University of London), in a study to extract and analyze aDNA from fossil beetles. This will allow for a detailed assessment of phylogeographic and taxonomic questions, which cannot be elucidated by analysis of morphology alone. This work focuses on fossil specimens preserved in late Pleistocene and Holocene permafrost deposits in Alaska and the Yukon, in addition to recent (mostly twentieth century) North American museum and wild-caught specimens. Thus far, work has focused on the ground beetle *Amara alpina*, with fossil specimens originating from deposits dated from >60 000 to <7000 cal years BP. A 569 bp sequence fragment of DNA from eight overlapping fragments within the mitochondrial COI gene was targeted. Using existing sequence data from previous experiments and the literature, novel species-specific primer pairs were developed to ensure an accurate fit to primer-binding sites. Following this optimization, amplification has been successful for all of the wild-caught ($n = 13$) and 28 of the 30 (93%) museum samples tested (Figure 10). The rate of DNA recovery from late Pleistocene and Holocene samples, determined by the retrieval of at least one sequence-verified fragment, has been lower at 20% (10 of the 51 samples tested). Although this success rate is far lower than the majority of late Pleistocene, permafrost-preserved mammal bone, it is comparable to nonpermafrost late Pleistocene material. Interestingly, the success rate for aDNA extraction in these fossil specimens does not appear to be related to the age of the fossil (Table 1). One might expect that younger fossil specimens would yield more aDNA than older fossils, but successful aDNA extraction (Figure 11) was accomplished using specimens ranging from



Figure 10 Modern specimen of the ground beetle *Amara alpina*, from the Canadian National Collection of Insects, Ottawa, Ontario. One hind leg per specimen was removed for the extraction of DNA. Photograph by Peter Heintzman, used with permission.

Table 1 aDNA extraction success rates for fossil specimens of *Amara alpina* from Alaska and the Yukon, compared with fossil age

Fossil age (years BP)	Specimen	Successful aDNA extraction?
6000	A	Yes
6000	B	No
8000	A	No
20 000	A	Yes
20 000	B	No
30 000	A	No
35 000	A	No
42 000	A	No
42 000	B	Yes
55 000	A	No
>60 000	A	Yes
>60 000	B	No



Figure 11 Extraction of ancient DNA from fossil beetle specimen, School of Biological Sciences, Royal Holloway, University of London. Photograph by Peter Heintzman, used with permission.

>60 000 years BP to the mid-Holocene. Perhaps the post-mortem history of each beetle specimen plays the most important role in DNA preservation. For instance, if a beetle dies during the peak of the summer season and its body spends some weeks exposed to the air before it becomes buried and incorporated into permanently frozen sediments, then decomposition will have progressed to the point where the beetle's DNA has seriously degraded before burial. If, on the other hand, a beetle dies in autumn and its body is almost immediately frozen, then the odds of DNA preservation are greatly improved.

Given their considerable abundance, there is great potential for the use of aDNA to address phylogeographic and taxonomic questions using the permafrost-preserved remains of beetles. This aspect of Quaternary Entomology promises to shed light on the population structure and past movements of populations. Thus, it may be required to begin to study the beetles themselves, rather than only employing their presence in fossil assemblages as a proxy for reconstructing environmental change.

See also: [Beetle Records: Late Pleistocene of Australia](#); [Late Pleistocene of Japan](#); [Late Pleistocene of New Zealand](#); [Late Pleistocene of North America](#); [Late Pleistocene of Northern Asia](#); [Late Pleistocene of South America](#); [Late Tertiary and Early Quaternary Records](#); [Middle Pleistocene of Western Europe](#); [Postglacial Europe](#); [Postglacial North America](#). [Glacial Climates: Biosphere Feedbacks](#); [Thermohaline Circulation](#). [Paleoclimate: Modern Analog Approaches in Paleoclimatology](#). [Paleoclimate Modeling: Last Glacial Maximum GCMs](#); [Quaternary Environments](#); [The Last Interglacial](#).

References

Atkinson TC, Briffa KR, Coope GR, Joachim M, and Perry D (1986) Climatic calibration of coleopteran data. In: Berglund BE (ed.) *Handbook of Holocene Palaeoecology and Palaeohydrology*, pp. 851–858. New York: Wiley.

Coope GR (1977) Fossil Coleopteran assemblages as sensitive indicators of climatic changes during the Devensian (Last) cold stage. *Philosophical Transactions of the Royal Society of London, Series B* 280: 313–340.

Coope GR (1978) Constancy of insect species versus inconstancy of Quaternary environments. In: Mound LA and Waloff N (eds.) *Diversity of Insect Faunas Symposium*, vol. 9, pp. 176–187. London: Royal Entomological Society of London.

Coope GR (1987) Fossil beetle assemblages as evidence for sudden and intense climatic change in the British Isles during the last 45,000 years. In: Berger WH and Labeyrie LD (eds.) *Abrupt Climate Change*, pp. 147–150. Dordrecht: D. Reidel Publishing Company.

Coope GR, Lerdahl G, Lowe JJ, and Walkling A (1998) Temperature gradients in northern Europe during the last glacial-Holocene transition (14–9 ¹⁴Ckyr BP) interpreted from coleopteran assemblages. *Journal of Quaternary Science* 13: 419–433.

Elias SA (1990) Observations on the taphonomy of late Quaternary insect fossil remains in packrat middens of the Chihuahuan Desert. *Palaos* 5: 356–363.

Elias SA (1992) Late Quaternary zoogeography of the Chihuahuan Desert insect fauna, based on fossil records from packrat middens. *Journal of Biogeography* 19: 285–297.

Elias SA (1994) *Quaternary Insects and Their Environments*. Washington, DC: Smithsonian Institution Press.

Elias SA (1996) Late Pinedale and Holocene seasonal temperatures reconstructed from fossil beetle assemblages in the Rocky Mountains. *Quaternary Research* 46: 311–318.

Elias SA (2001) Mutual climatic range reconstructions of seasonal temperatures based on late Pleistocene fossil beetle assemblages in eastern Beringia. *Quaternary Science Reviews* 20: 77–91.

Elias SA (2010) *Advances in Quaternary Entomology. Developments in Quaternary Sciences*, vol. 12. Amsterdam: Elsevier.

Elias SA, Anderson KH, and Andrews JT (1996) Late Wisconsin climate in northeastern USA and southeastern Canada, reconstructed from fossil beetle assemblages. *Journal of Quaternary Science* 11: 417–421.

Elias SA, Berman D, and Alfimov A (2000) Late Pleistocene beetle faunas of Beringia: Where east met west. *Journal of Biogeography* 27: 1349–1363.

Elias SA and Matthews JV Jr. (2002) Arctic North American seasonal temperatures in the Pliocene and early Pleistocene, based on mutual climatic range analysis of fossil beetle assemblages. *Canadian Journal of Earth Sciences* 39: 911–920.

Girling MA (1988) The bark beetle *Scolytus scolytus* (Fabricius) and the possible role of elm disease in the early Neolithic. In: Jones M (ed.) *Archaeology and the Flora*

- of the British Isles. *Archaeology Monograph*, vol. 14, pp. 34–38. Oxford: Oxford University Committee.
- Gröcke DR, Elias SA, Miller RM, and Schimmelmann A (2006) Hydrogen isotopes in chitin. *Quaternary Science Reviews* 25: 1850–1864.
- Gröcke DR, Hardenbroek M, Sauer P, and Elias S (2010) Hydrogen isotopes in beetle chitin. In: Gupta SS (ed.) *Chitin: Formation and Diagenesis*, pp. 105–115. New York, NY: Springer.
- King GA, Gilbert TP, Willerslev E, Collins MJ, and Kenward H (2009) Recovery of DNA from archaeological insect remains: First results, problems and potential. *Journal of Archaeological Science* 36: 1179–1183.
- Lavoie C and Arseneault D (2001) Late Holocene climate of the James Bay area, Québec, Canada, reconstructed using fossil beetles. *Arctic, Antarctic, and Alpine Research* 33: 13–18.
- Miller RF, Fritz P, and Morgan AV (1988) Climatic implications of D/H ratios in beetle chitin. *Palaeogeography, Palaeoclimatology, Palaeoecology* 66: 277–288.
- Porch N (2010) Climate space, bioclimatic envelopes and coexistence methods for the reconstruction of past climates: A method using Australian beetles and significance for Quaternary reconstruction. *Quaternary Science Reviews* 29: 633–647.
- Reiss RA (2006) Ancient DNA from ice age insects: Proceed with caution. *Quaternary Science Reviews* 25: 1877–1893.
- Rozanski K, Araguas-Araguas L, and Gonfiantini R (1993) Isotopic patterns in modern global precipitation. In: Swart PK, Lohmann KC, McKenzie J, and Savin S (eds.) *Climate Change in Continental Isotopic Records. Geophysical Monograph*, vol. 78, pp. 1–36. Washington, DC: American Geophysical Union.
- Sher AV, Kuzmina SA, Kuznetsova TV, and Sulerzhitsky LD (2005) New insights into the Weichselian environment and climate of the East Siberian Arctic, derived from fossil insects, plants, and mammals. *Quaternary Science Reviews* 24: 533–569.
- Thiele HU (1977) *Carabid Beetles in Their Environments*. Berlin: Springer.
- Wang Y, O'Brien DM, Jenson J, Francis D, and Wooller MJ (2009) The influence of diet and water on the stable oxygen and hydrogen isotope composition of Chironomidae (Diptera) with paleoecological implications. *Oecologia* 160: 225–233.
- Willerslev E, Cappellini E, Boomsma W, et al. (2007) Ancient biomolecules from deep ice cores reveal a forested southern Greenland. *Science* 317: 111–114.
- Wooller MJ, Axford Y, and Wang Y (2008) A multiple stable isotope record of Late Quaternary limnological changes and chironomid paleoecology from northeastern Iceland. *Journal of Paleolimnology* 40: 63–77.
- Wooller MJ, Francis D, Fogel ML, Miller GH, Walker IR, and Wolfe AP (2004) Quantitative paleotemperature estimates from $\delta^{18}\text{O}$ of chironomid head capsules preserved in arctic lake sediments. *Journal of Paleolimnology* 31: 267–274.

Late Tertiary and Early Quaternary Records

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Introduction

During the past 30 years, Quaternary insect paleontologists working in the northern high latitudes have had the opportunity to study extremely rare deposits of fossil insects that date back millions of years. In some cases, these fossil assemblages represent Late Tertiary environments that preceded the earliest glaciations of the Quaternary. Other assemblages represent Early Quaternary environments, when glacial–interglacial cycles were beginning. These fossil assemblages afford us rare glimpses into the history of the Arctic and subarctic regions, during periods when regional environments were startlingly different from today. In many cases, there are no modern analogs for the fossil assemblages. The challenge for insect paleontologists has been to make ecological sense of these fossil assemblages, to reconstruct the environments in which they lived. It has been difficult, painstaking work that, in some cases, required decades of research to bring to completion.

These fossil insect faunas bridge a paleontological gap between the actual exoskeletal remains preserved in the unconsolidated sediments of the Pleistocene, and the mineral replacements and trace fossils preserved in deposits of bedrock, dating from the Tertiary back to the Paleozoic. What makes these Late Tertiary and Early Quaternary fossils unique is that they are millions of years old, yet they are essentially the same kind of exoskeletal remains found in much younger (Pleistocene and Holocene) sediments. In other words, these fossils are not mineralized replacements of insect exoskeletons, but the exoskeletons themselves, preserved in frozen deposits since their death. When removed from their frozen sedimentary matrix, these fossils have the same freshness of appearance as much younger specimens from other parts of the world. To the insect paleontologist studying these specimens under the microscope, it appears as if these insect specimens have spent up to several million years in a state of suspended animation. Together with frozen plant remains and vertebrate animal bones from these deposits, the Arctic insect fossils preserved from Late Tertiary and Early Quaternary deposits provide a treasure trove of information on the deep history of the Arctic regions.

History of Late Tertiary/Early Quaternary Insect Fossil Studies

The existence of these fossil deposits was unknown until quite recently. Paleontological investigation of remote Arctic and subarctic regions only came about in the latter half of the twentieth century. These studies require lengthy field expeditions that are logistically difficult and expensive. The possibility of fieldwork in such regions is generally limited to a brief summer season. Much of the research has been carried out under the auspices of governmental organizations, such as the geological surveys of Canada and Denmark, and the Academy of Sciences of the former Soviet Union. However, some of

the work has been done by governmental organizations in cooperation with universities, such as the research done by Sergei Kiselev (Moscow State University, Russia) and Svetlana Kuzmina (now at University of Alberta, Canada).

One of the most difficult aspects of this research has been establishing the chronologies of the fossil assemblages. While radiometric dating has been used to establish the age of a few samples (notably the Lava Camp assemblages from Alaska), most of the assemblages discussed below have only been dated on the basis of site stratigraphy. Many sites in Siberia have been reliably dated on the basis of the biostratigraphy of their large mammal and rodent fossils; the great antiquity of many sites provides the possibility of assessing evolutionary changes in the insect faunas. Nevertheless, the insect assemblages discussed here necessarily have only approximate age estimates. The ages discussed below represent the most up-to-date interpretations of the site investigators, but these chronologies are very likely to change in future, as more geologic research is done in the study regions.

Preservation of Late Tertiary/Early Quaternary Insect Fossils

Exoskeletal remains of insects from the Late Tertiary and Early Quaternary have been found almost exclusively in Arctic and subarctic regions. Chitin, the main structural constituent of insect exoskeletons, is highly resistant to decay, but eventually it decomposes, especially when exposed to air. The long-term preservation of chitinous exoskeletons in the high latitudes is greatly facilitated by permanently frozen ground, or permafrost.

Importance of Permafrost

Permafrost is key to the preservation of insect exoskeletons in the Arctic and subarctic, for several reasons. First, it greatly retards the pace of chemical reactions and bacterial decomposition, allowing chitin to remain virtually unaltered for as much as several million years. Second, permafrost protects insect exoskeletal remains from the physical abrasion that often takes place as sediments are reworked by wind, water, and gravity. Layers of frozen sediments tend to remain intact, or only are moved as large blocks, in response to thermokarst action (the melting of frozen sediments, leading to erosion and slumping). Thus, the frozen sediments of the North have entombed insect exoskeletons, and have protected them from the physical and chemical degradation that breaks down chitin in other types of environments.

One topic currently being explored is the recovery of frozen DNA samples from insect fossils preserved in frozen sediments. It has recently been demonstrated that frozen sediments in Siberia, Alaska, and Canada can yield well-preserved fossil

DNA from plants and mammals (Shapiro et al., 2004; Willerslev et al., 2003). It would appear likely that the frozen soft tissues of fossil insects or the frozen sediments themselves will also yield ancient DNA. If it can be obtained, this DNA will be able to help entomologists work out the genetic history of insect species in the past few million years.

Rarity of Fossil Localities

Although fossil insect impressions in bedrock and insect remains trapped in fossil amber dating back many millions of years, fossil sites containing actual exoskeletal insect remains more than 1 My old are quite rare. Permafrost environments only occur in the high latitudes, and many high latitude regions in America and Europe were repeatedly glaciated in the Quaternary. Siberia had a different situation: many ice-free regions in the high latitudes, but a lack of deposits older than the middle Pleistocene. Glaciers and ice sheets have obliterated underlying (Late Tertiary and Quaternary) deposits, stripping the land surface down to bedrock in many Arctic regions, and then leaving a mantle of reworked debris as their margins retreated. Thus, one of the main reasons why unconsolidated, organic-rich deposits of Late Tertiary and Early Quaternary age are rare is that they could only ever occur in regions that were not subject to repeated Quaternary glaciations. However, there were some regions in the Arctic that remained unglaciated through much if not all of the Pleistocene. Chief among these was the region known as Beringia, which included the unglaciated lowlands of northeastern Russia, Alaska, and the Yukon Territory, linked together by the Bering Land Bridge. This land bridge formed during glaciations, as the growth of continental ice sheets trapped huge quantities of water, causing global sea level to drop. The continental shelf regions of the Bering and Chukchi seas are relatively shallow (most of the shelf regions are less than 90 m below modern sea level), so these shelves became dry land during glaciations, connecting northeastern Siberia with western Alaska.

Types of Insects Preserved in Frozen Sediments

By far the most abundant and diverse group of insects preserved in these deposits are beetles (Coleoptera). The beetle families most often found in these ancient assemblages are ground beetles (Carabidae), water beetles (Dytiscidae and Hydrophilidae), rove beetles (Staphylinidae), leaf beetles (Chrysomelidae), and weevils (Curculionidae). Some small families are represented by very numerous remains, including pill beetles (Byrrhidae) of the genus *Morychus*, and small carrion beetles (Leiodidae). Various kinds of flies (Diptera) have been found, especially nonbiting midges (Chironomidae), but the traditional method of laboratory screening misses most of these small fossils. The pupal cases of flies are sometimes preserved. Parasitic wasps and ants (Hymenoptera) are also commonly found. Finally, several families of other kinds of arthropods, such as oribatid mites (Arachnida: Oribatida) are found in these fossil assemblages. Oribatids are soil-dwelling mites that are the chief agents of organic decomposition in Arctic soils.

The fossil insect faunas of Late Tertiary and Early Quaternary age from North America and Greenland comprise 437 taxa of insects and arachnids. Of these, 180 have been

identified to the species level (Matthews and Telka, 1997). Siberian faunas were studied without arachnids, parasitic wasps and flies (Kiselev, 1981; Kuzmina, 1989). Insect faunas of this age have also been published from northeastern Siberia by Kiselev (in Kiselev and Nazarov, 2009). These include 139 beetle, 3 ant, and 2 bug taxa, of which 119 species have been identified. Surprisingly, there are only ten species in common between the Siberian and North American/Greenland faunal lists for these ages of fossil assemblages.

Research in Alaska and Canada

John Matthews, Geological Survey of Canada, has studied a number of fossil insect assemblages of Late Tertiary and Early Quaternary age in Alaska and northern Canada, and his work has undoubtedly made the greatest contribution to our knowledge of insect faunas of this age. His work in this topic spans 25 years and 19 publications. Nine faunal assemblages dating from about 5.7 to 1.8 Ma have thus far been analyzed (Table 1). The sites range from just south of the Arctic Circle at the Lost Chicken site, to north of 80° latitude at the Meighen Island and Wolf Valley sites. Several new sites in Alaska and the Yukon that are of early Pleistocene age have been studied by Kuzmina, but have not yet been published.

The series of fossil insect assemblages (Figure 1) from these sites document very warm, generally maritime Late Tertiary climates. The insect faunas of this age are quite mixed, in terms of their modern ranges. Some of these species live today only in Asia; others live today in the temperate regions of eastern North America. As might be expected for such warm-adapted insect faunas, there is considerable taxonomic diversity in these assemblages. The level of diversity declined during the onset of climatic cooling associated with Tertiary–Quaternary transition (Matthews and Telka, 1997). The Late Tertiary faunas of the Arctic regions of North America lived in coniferous forests that were likewise more diverse than the modern boreal forests of Canada and Alaska. The ancient forests had some taxa that are less tolerant of cold climate, such as hemlocks and elderberry.

By about 2.5–2 Ma, there is reliable evidence for climatic cooling from both the insect and plant fossil records. Nevertheless, temperatures remained well above their modern levels in the high Arctic. Sites such as Wolf Valley, Ellesmere Island (2.1–1.7 Ma) and Cape Deceit, Alaska (1.8 Ma) contain mixtures of Arctic tundra and coniferous forest insect fauna and flora. The Cape Deceit fauna includes some elements that typify the mesic tundra environments that existed throughout much of the late Pleistocene, including a wide variety of ground beetle species in the *Cryobius* group of the genus *Pterostichus*. The same reconstruction has been made from the 2 Ma insect fauna from Dawson Cut Bed, Palisade site on the Yukon River, Alaska (Elias et al., unpublished).

Fossil insect assemblages of probable Early Quaternary are known from the Old Crow and Bluefish River Basins in the northern Yukon, from the Klondike gold mining district of the central Yukon, and from the Palisades site in central Alaska (Figure 2). These faunas, unlike their late Pliocene counterparts, do not contain extinct species, with the possible exception of a pill beetle (family Byrrhidae) in the genus *Morychus*

Table 1 Late Tertiary and Early Quaternary insect fossil sites

Site	Age (millions of years ago)	Insect fauna	Environmental reconstruction	References
(1) Lava Camp Mine, Alaska	5.7 ± 0.2 Ma (Late Miocene)	83 insect and arachnid taxa, including several extinct species; fauna indicative of coniferous forest	Warm, maritime climate; fossil flora includes species found today in the forests of the Pacific Northwest region	Hopkins et al. (1971), White et al. (1997, 1999), and Elias and Matthews (2002)
(2) Ary-Mas, NE Siberia, Russia	Late Miocene	31 insect taxa, including at least five extinct species; includes several North American species not found in any other Asian faunas	Fauna includes mixture of coniferous forest and open ground taxa, probably indicating open coniferous woodland or forest tundra	Elias et al. (2006) and Kiselev and Nazarov (2009)
(3) Niguanak, Alaska	Early Pliocene?	47 insect and arachnid taxa, including extinct species, <i>Diacheila matthewsi</i> . Fauna not yet fully analyzed, but has some Asian affinities	Plant fossils indicate coniferous forest, including hemlocks; climate was far warmer and less continental than today	Matthews and Telka (1997) and Elias and Matthews (2002)
(4) Ballast Brook, Banks Island, NWT, Canada	5–3 Ma	39 insect and arachnid taxa, indicative of boreal environment	Coniferous forest, similar to modern forests in the subarctic regions of Canada. There are indications that the site was near northern treeline at the time of fossil deposition. Climate substantially warmer than today	Fyles et al. (1994) and Elias and Matthews (2002)
(5) Strathcona Beaver Peat, Ellesmere Island, NWT, Canada	>3.3 Ma	86 insect and arachnid taxa, including both Arctic and boreal elements; some species with Asian affinities	Fauna and flora indicative of northern treeline setting; climate substantially warmer than today	Matthews and Telka (1997) and Elias and Matthews (2002)
(6) Meighen Island sites, NWT, Canada	3 Ma	198 insect and arachnid taxa, including both boreal and Arctic tundra species. Faunal affinities with eastern North America and Asia	Flora and fauna indicative of northern treeline environment; climate substantially warmer than today	Matthews and Telka (1997) and Elias and Matthews (2002)
(7) Lost Chicken, Alaska	3 Ma	83 insect and arachnid taxa, including blind weevil, <i>Alaocybites</i> , found today in California and the Russian Far East; boreal fauna with several extinct species	Flora and fauna indicative of coniferous forest; several warmth-loving plant species; climate with warmer winters than today	Matthews and Telka (1997) and Elias and Matthews (2002)
(8) Bluefish Basin, Yukon Territory	<3 Ma (Late Pliocene)	26 insect and arachnid taxa, including <i>Notiophilus aeneus</i> , now confined to eastern N. America, and extinct species, <i>Helophorus meighensis</i>	Bluefish plant fossil assemblages include species indicative of substantially warmer, less continental climate than exists in the Yukon today; MCR results indicate TMAX up to 3 °C warmer than today	Matthews and Ovenden (1990), Matthews and Telka (1997), and Elias and Matthews (2002)
(9) Krestovka, NE Siberia, Russia	Late Pliocene (Kutuyakh Fm.)	22 beetle taxa; mixture of boreal and Arctic fauna; some elements of steppe tundra fauna already present	Larch forest tundra landscape with steppe patches	Kiselev (1981)
	Early Quaternary (Olyorian Suite)	51 beetle taxa; most found today in Arctic tundra regions; some taxa with steppe affinities	Steppe-tundra landscape with patches of shrub tundra and coniferous woodlands	Kiselev (1981) and Kuzmina (1989)
(10) Chukochya River, NE Siberia, Russia	Late Pliocene to Early Quaternary (Olyorian Suite)	80 beetle species; most found today in Arctic tundra regions; some taxa with steppe affinities	Steppe-tundra landscape with patches of shrub tundra and coniferous woodlands	Kiselev (1981) and Kuzmina (1989)
(11) Alazea, NE Siberia, Russia	Late Pliocene to Early Quaternary (Olyorian Suite)	56 beetle species; most found today in Arctic tundra regions; some taxa with steppe affinities	Steppe-tundra landscape with patches of shrub tundra and coniferous woodlands	Kiselev (1981) and Kuzmina (1989)
(12) Sededema, NE Siberia, Russia	Late Pliocene to Early Quaternary (Olyorian Suite)	55 beetle species; most found today in Arctic tundra regions; some taxa with steppe affinities	Steppe-tundra landscape with patches of shrub tundra and coniferous woodlands	Kiselev (1981) and Kuzmina (1989)

(Continued)

Table 1 (Continued)

Site	Age (millions of years ago)	Insect fauna	Environmental reconstruction	References
(13) Palisades, Alaska	Late Pliocene	26 insect taxa including steppe-tundra and mesic tundra species; no extinct taxa identified	Both dry and mesic tundra habitats represented with TMAX 3–5 °C cooler than today	Elias et al. (unpublished)
	Early Pleistocene	27 insect taxa, all extant, indicative of northern treeline environments	Both dry and mesic tundra with some boreal elements present	Elias et al. (unpublished)
(14) Kap København, Greenland	2.5–2 Ma (Early Quaternary)	154 insect and arachnid taxa, indicative of northern treeline environments	Probably deposited after first glaciation; less faunal diversity than Meighen Island faunas, indicating climatic cooling; climates still substantially warmer than today. TMAX about 8 °C warmer than today; TMIN about 10 °C warmer than today	Böcher (1995) and Elias and Matthews (2002)
(15) Wolf Valley, Ellesmere Island, Nunavut, Canada	2–1.7 Ma	21 insect and arachnid taxa, indicative of northern treeline environments	Plant and insect fossils suggest northern treeline near site at time of deposition; climate cooler than that of Beaufort Formation faunas; TMAX 5–7 °C warmer than today	Matthews and Fyles (2000) and Elias and Matthews (2002)
(16) Cape Deceit, Alaska	1.8 Ma	86 insect and arachnid taxa, indicative of northern treeline environments; some elements of tundra fauna seen for first time in Eastern Beringian region	The vegetation record suggests that temperatures were slightly warmer than they are today. MCR results indicate TMAX about 2 °C warmer than today; TMIN similar to modern winter temperatures	Matthews (1974), McDougall (1995), Matthews and Telka (1997), and Elias and Matthews (2002)
(17) Old Crow Hidden Bluff, Bluefish Basin, Yukon	Early Pleistocene	43 insect taxa, indicative of forest but with a more significant contribution of xerophilous species (tundra and steppe) than are found here today	Tree remains suggest well-developed forest, but insect assemblage indicates high proportion of open disturbed ground and steppe patches	Kuzmina (unpublished)
(18) Klondike, Yukon	Early Pleistocene	61 insect taxa indicative of forest and boggy environment, significant numbers of mesic tundra species and some indicators of steppe-tundra	The beds yield branches and stumps of trees, smaller than recent in this area. Insect assemblage indicates climate similar to modern, but somewhat cooler and dryer	Kuzmina (unpublished)

that has not been matched with any extant species (Kuzmina et al., 2008). This beetle, and the sage-feeding weevil *Connatichela artemisiae* are not commonly preserved in these faunas. They are indicators of steppe-tundra environments that were apparently much more common in the late Pleistocene of Eastern Beringia. However, these species have been found in the Klondike fauna that is of early Pleistocene age.

Research in Siberia

The Siberian faunal assemblages of Late Tertiary and Early Quaternary age have been described from four sites, including just two of pre-Pleistocene age (Figure 3; Table 1). The sites are along the banks of large rivers (Figure 4). The oldest Western Beringian insect fauna was described from the Ary-Mas site (Figure 3). This Miocene-age fossil fauna indicates mostly riparian environments. While preservation of these fossils is generally poor, the fauna likely includes several extinct species (Elias et al., 2006; Kiselev and Nazarov, 2009).

In contrast to the fossil record of the North American Arctic, the Kutuyakh Formation Beds (Krestovka site) of late Pliocene age in northeastern Siberia already have a well-developed cold-adapted insect fauna. These assemblages include species of *Pterostichus* (*Cryobius*) spp. and *Blethisa catenaria* that live on mesic to wet tundra and forest tundra, as well as woodland species, especially weevils associated with conifers (Kiselev, 1981). Faunal assemblages from stratigraphic horizons between the Kutuyakh and Olyorian Suite Formations contain mostly tundra species, including species of *Pterostichus* (*Cryobius*) spp., as well as the more dry-adapted arctic ground beetle, *Amara alpina* (*Curtonotus alpinus* in the Russian literature). However, elements of the taiga fauna persist in these faunas, especially tree-feeding weevils. Some steppe-associated beetles made their first appearance in northern Siberian faunal assemblages of this age (late Pliocene). The combined faunal and floral evidence suggests a larch forest-tundra landscape with patches of steppe.

The Olyorian Suite is now thought to represent the early Pleistocene and part of the middle Pleistocene. Previous

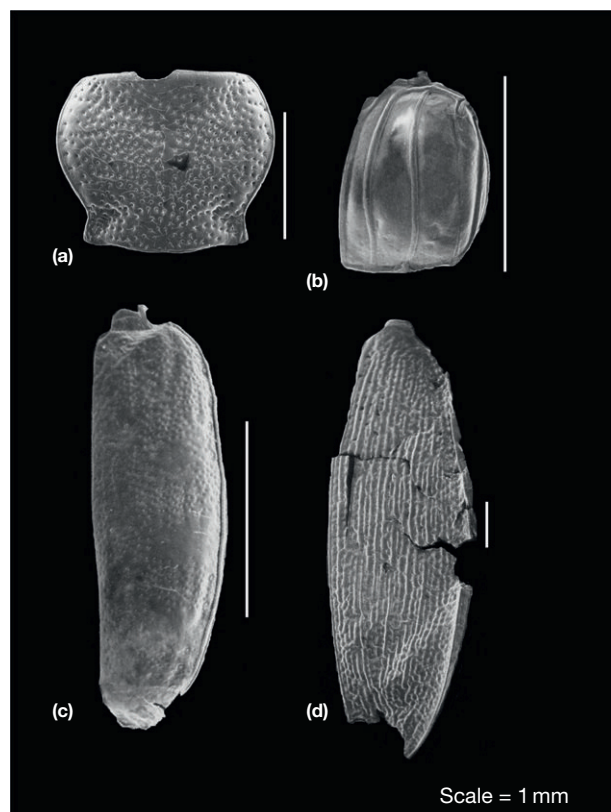


Figure 1 Scanning electron micrographs of fossil beetle specimens from Canada and Alaska. (a) Pronotum of *Diacheila matthewsi* from the Beaufort Formation, Meighen Island; (b) elytron of extinct species of *Kalissus* from the Beaufort Formation, Meighen Island; (c) elytron of extinct species of *Asaphidion* from the Lost Chicken site, Alaska; (d) elytron of extinct species of *Carabus* from the Beaufort Formation, Meighen Island. Photos by Alice Telka and John Matthews, Geological Survey of Canada.

attributions of this complex to the late Pliocene–early Pleistocene (Kiselev, 1981; Sher, 1974) were made on the basis of Soviet stratigraphic boundaries, in which the Pliocene–Pleistocene boundary was set at the Brunnhes–Matuyama transition of 781 000 years BP (Sher, 1997; Minyuk, 2004). Some deposits of this age are found in north eastern Siberia, especially in the Kolyma lowland region. The dominant feature of these beetle assemblages (Figure 5(a)–5(d)) is that they are virtually identical in composition to the subsequent regional faunas from Middle and late Pleistocene times. The only difference recognized thus far (Kuzmina, 1989) is that the steppe component of Olyorian Suite faunas is less important than it is in most of the middle and late Pleistocene insect assemblages. The typical Olyorian insect faunas consist mostly of tundra species, with dry tundra insects dominating over mesic and wet tundra taxa.

One very interesting site is Chomus Yuryakh is north eastern Siberia. One very large sample containing the remains of more than 5000 insects was recovered from the Olyorian Suite at this site. The fauna contained at least four extinct species: two weevils of the genus *Phyllobius* that were probably members of steppe fauna, and two species of the willow-feeding

weevil genus *Lepyryus* (Korotyaev and Kuzmina, unpublished). This is a very rare example of species extinction within the Quaternary.

Research in Greenland

Thus far, only one Greenland site has been studied that contains insect fossil assemblages in the age range being considered here. Most of Greenland has been covered by ice during the many glaciations of the Quaternary, making the fossil site all the more remarkable. Kap København (Figure 6) is the northernmost site yet studied for insect fossils of this age. It is situated near the northern tip of Greenland (Figure 3, No. 12). The fossil insect assemblages (Figure 5(e)–5(h)) have been described by Böcher (1995). The age of the deposits is estimated between 2.5 and 2 Ma, at approximately the boundary between the Pliocene and the Pleistocene. At that time, it is clear that northern high latitude climates were remarkably warm. The Kab København flora and insect fauna are indicative of northern treeline environments. It appears that the Kab København fossil beds were deposited soon after the first glacial period, near the end of the Pliocene. The Meighen Island fauna from northern Canada is more diverse, and contains more warm-adapted species than the Kab København fauna. The former fauna is thought to be 500 000–700 000 years older than the latter (Matthews and Telka, 1997).

Potential Importance of Research

The research outlined above is important to a number of disciplines. It has helped define paleoclimate reconstructions for regions that formerly had little or no reconstructions for the critical climatic transition from the relatively warm climates of the Pliocene to the oscillating glacial and interglacial climates of the Quaternary. The wealth of paleoenvironmental data extracted from the fossil insect assemblages has also added greatly to our understanding of soils, moisture conditions, and vegetation cover of the Arctic and subarctic regions during the late Pliocene and Early Quaternary. The fossil insect evidence also provides important evidence concerning the longevity of many insect species, especially beetles. Finally, the combination of plant and insect fossil assemblages has given us important new insights into the evolution, functioning, and longevity of northern biological communities.

Insects as Paleoclimate Indicators

The fossil insect research described above has demonstrated the morphological constancy of exoskeletal features in beetles and some other insects, over several million years for some species. This, alone, would seem a strong indicator that the species in question have remained constant throughout this time interval. However, insect exoskeletons provide little or no data on the internal physiology or ecological requirements of these organisms. In order to test these aspects of species' constancy, another, indirect method had to be employed. Fortunately, we do have a method of examining the ecological requirements of apparently extant species in fossil assemblages – through

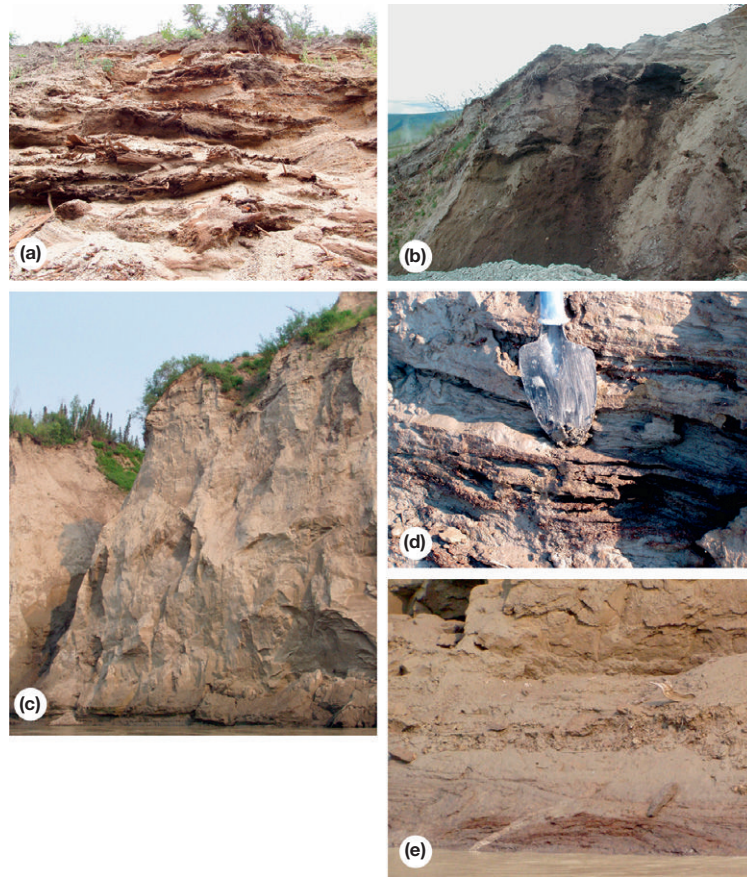


Figure 2 Fossil sites in Alaska and the Yukon. (a) Early Pleistocene (?) exposure at Hidden Bluff in the Bluefish Basin, Yukon; (b) Late Pliocene exposure at the Lost Chicken Mine site, Alaska; (c) Quaternary deposits exposed at the Palisades of the Yukon, Alaska; (d) Late Pliocene Dawson Cut Forest Bed layers at Palisades site; (e) base of the Palisades exposure, showing Dawson Cut Forest Bed sediments. Photos by Svetlana Kuzmina.

the analysis of their modern counterparts. The modern distributions of species found in fossil assemblages fall within certain climatic parameters. When these parameters (such as mean temperature of the warmest or the coldest month of the year) are plotted for all the known modern collecting localities of a species, it is possible to generate a 'climate envelope' for that species (Atkinson et al., 1986). These climate envelopes vary from broad to narrow, depending on the range of thermal tolerances of the species. Elias and Matthews (2002) analyzed the climatic compatibility of the Late Tertiary and early Pleistocene fossil assemblages from Greenland, Canada, and Alaska, and in all cases the species in the fossil assemblages were found to be climatically compatible. Taking this process a step further, the climate envelopes of the species in the fossil assemblages overlapped to yield a mutual climatic range (MCR), a set of climatic parameters within which all the species in the assemblage can live.

Longevity of Insect Species

The MCR study indicates that the species in question have not evolved new sets of climatic tolerances since the Late Tertiary. Dozens of species were used in the MCR reconstructions. If this kind of physiological evolution had taken place, then it would be extremely unlikely that the climate envelopes based

on the modern distribution of these beetles would overlap to produce an MCR.

Given the fact that the modern representatives of the species in the fossil assemblages remain climatically compatible with each other, then there has either been no evolution of their thermal tolerances since the Late Tertiary, or whole suites of species have evolved new tolerances in concert with each other. The latter scenario appears extremely unlikely, as evolutionary theory holds that each species evolves independently from others.

However, as Matthews and Telka (1997) point out, the Late Tertiary beetle faunas from the Arctic also contain 'clear examples of extinct species.' It appears likely that most of these species were relatively warm-adapted, and that they failed to become established in more southerly regions during the climatic cooling that signaled the beginning of glaciations in the Quaternary. The fossil evidence demonstrates that the evolutionary 'clock' of beetles has not stopped, it is just moving extremely slowly. Some Late Tertiary species apparently were precursors to modern species that evolved during the Quaternary. For instance, Matthews and Telka (1997) discussed the possible evolutionary history of a ground beetle in the genus *Asaphidion*, an extinct species closely related to the contemporary subarctic carabid *Asaphidion yukonense*. A fossil elytron from the Lost Chicken site (Figures 1(c) and 2) clearly shows



Figure 3 Map of the Arctic, showing fossil sites discussed in text.

bare (impunctate) patches that characterize the modern species, but the patches are less well developed and elytral micro-sculpture better developed than in the extant species. Fossils that are similar to *A. yukonense* have been found in other Late Tertiary assemblages from the Arctic, but some of those have even more primitive elytral sculpture than the Lost Chicken specimens. These fossils could represent various stages in the evolution of the modern species. Matthews (1976) also described Late Tertiary fossil specimens of the water-scavenger beetle genus, *Helophorus*, which appear to be precursors to the modern species, *H. tuberculatus*. The fossils, from Lava Camp Mine and Meighen Island, exhibit more primitive elytral sculpture. Matthews traced the possible evolution of these fossil types into the extant species. Matthews (1970) also discussed the possible evolution of the more primitive rove beetles, *Micropeplus hoogendorni* and *M. hopkinsi*, leading to the modern species, *M. cribratus* and *M. punctatus*. Matthews assumed that the two former species had become extinct in the Quaternary,

but *M. hoogendorni* has subsequently been found to be extant. Modern specimens were collected from Siberia, by the Russian entomologist, Rjabukhin. Unaware of the fossil evidence from Alaska, he described the Siberian specimens as a new species, *M. dokuchaevi* (Elias, 1994). The unique example of rapid evolution in the rove beetle *Tachinus apterus* was documented by Matthews (1974) from the Cape Deceit fossil assemblages in western Alaska. The elytral proportions changed with time, together with a reduction in wing size to the point where the species became flightless. This evolutionary trend may have served as an adaptation to life in the increasing cold environments of the early Pleistocene.

Insects as Paleoecological Indicators

Some of the environments reconstructed on the basis of these insect faunal assemblages have no modern analogs, in fact, most of them cannot be found today. This does not mean



Figure 4 Fossil sites in northeastern Siberia. (a) Exposure of Olyorian Suite Formation sediments at Chukochya River; (b) Begunovskaya Suite at the Krestovka site. Photos by David Hopkins and Andrei Sher.

that the insect evidence is contradictory, it simply means that some of the ancient environments in which they lived have ceased to exist. This is especially true for the Arctic Pliocene environments. From 3 to 4 Ma, and up to parts of the Early Quaternary at Kap København, coniferous forests were growing at very high latitudes. These regions are polar deserts today, with a minimum of tundra vegetation interspersed with areas of bare rock, sand, and gravel. Paleoclimatologists have yet to develop paleoclimate models that can satisfactorily explain the existence of very warm climates in polar regions during the transition from Late Tertiary to Quaternary times, but the fossil data (plant and animal) clearly demonstrate its existence.

Origins and Longevity of Biological Communities

The fossil beetle evidence, in agreement with the fossil plant evidence from the same or similar deposits, demonstrates that biological communities are ephemeral associations of species. Certain *functional types* of communities may persist through great lengths of time (for instance, coniferous forests or Arctic tundra), but the constituent species within these communities shift with great regularity, in response to environmental changes. The Late Tertiary and Early Quaternary biota of the Arctic constituted boreal-style coniferous forests,

but these were unlike any forests that exist today, in any particular region.

One of the dominant trends in this sequence of Arctic faunas is the gradual depletion of the number of species through time. The earlier, warm-adapted faunas were relatively rich in species. The later, cold-adapted faunas contain fewer species. This makes sense from an ecological standpoint, as few insects are adapted to life in extremely cold climates. Since they do not maintain their own internal temperature through metabolism, as do mammals, insects are at a distinct disadvantage in cold climates. The modern statistics clearly reveal this pattern. Danks (1981) listed just 1800 named species of arthropods living today in the North American Arctic. The number of insect species living today in North America as a whole exceeds 90 000 species (Borror and White, 1970).

MCR Reconstructions

As discussed above, Elias and Matthews (2002) employed the predaceous and scavenging beetle species found in the Late Tertiary and Early Quaternary fossil assemblages to produce their MCR estimates (for a discussion of MCR methods, see Elias, 1994). These beetles are in various families, including ground beetles (Trachypachidae and Carabidae), predaceous diving beetles (Dytiscidae), water-scavenger beetles (Hydrophilidae), rove beetles (Staphylinidae), carrion beetles (Silphidae), and dung beetles (Scarabaeidae). Eleven fossil assemblages associated with this time interval contained sufficient numbers of identified species to yield MCR estimates. Elias and Matthews (2002) found that average winter temperatures (TMIN) in the Arctic during the latest Miocene and Pliocene were substantially warmer than they are today. The MCR estimates agree with the evidence from paleobotanical data, that Arctic Pliocene climates were far less continental than they are today. Several Pliocene-age assemblages from the high Arctic yielded mean summer temperature (TMAX) estimates 9–10 °C warmer than modern values at the sites. This is the same degree of warming required to allow coniferous forests to grow in the high Arctic. By 3 My, a cooling trend is marked in both the paleobotanical and fossil beetle evidence from Alaska. All assemblages dating between 5.7 and 2 Ma yielded TMAX values between about 12 and 14 °C, regardless of location. Thus, the insect fossil data support the theory that there was far less latitudinal gradation in temperatures during the late Pliocene than there is today. These results agree well with a study of global environments during the mid-Pliocene (Dowsett et al., 1999), which showed greatly reduced continental ice volume, greatly reduced sea-ice with the Arctic being seasonally ice free, and an expansion of evergreen forests to the margins of the Arctic Ocean. The MCR reconstructions suggest that regional climatic cooling (especially winter temperatures) began by at least 2 Ma.

The fossil insect history of the Arctic provides fascinating glimpses into a world that ceased to exist about 2 Ma – a world in which a substantially warmer Arctic was home to coniferous forests, even at 80° N, near the northernmost point of land in the Northern Hemisphere. The ecosystems of this warm Arctic zone collapsed as the Ice Ages began in earnest, but,

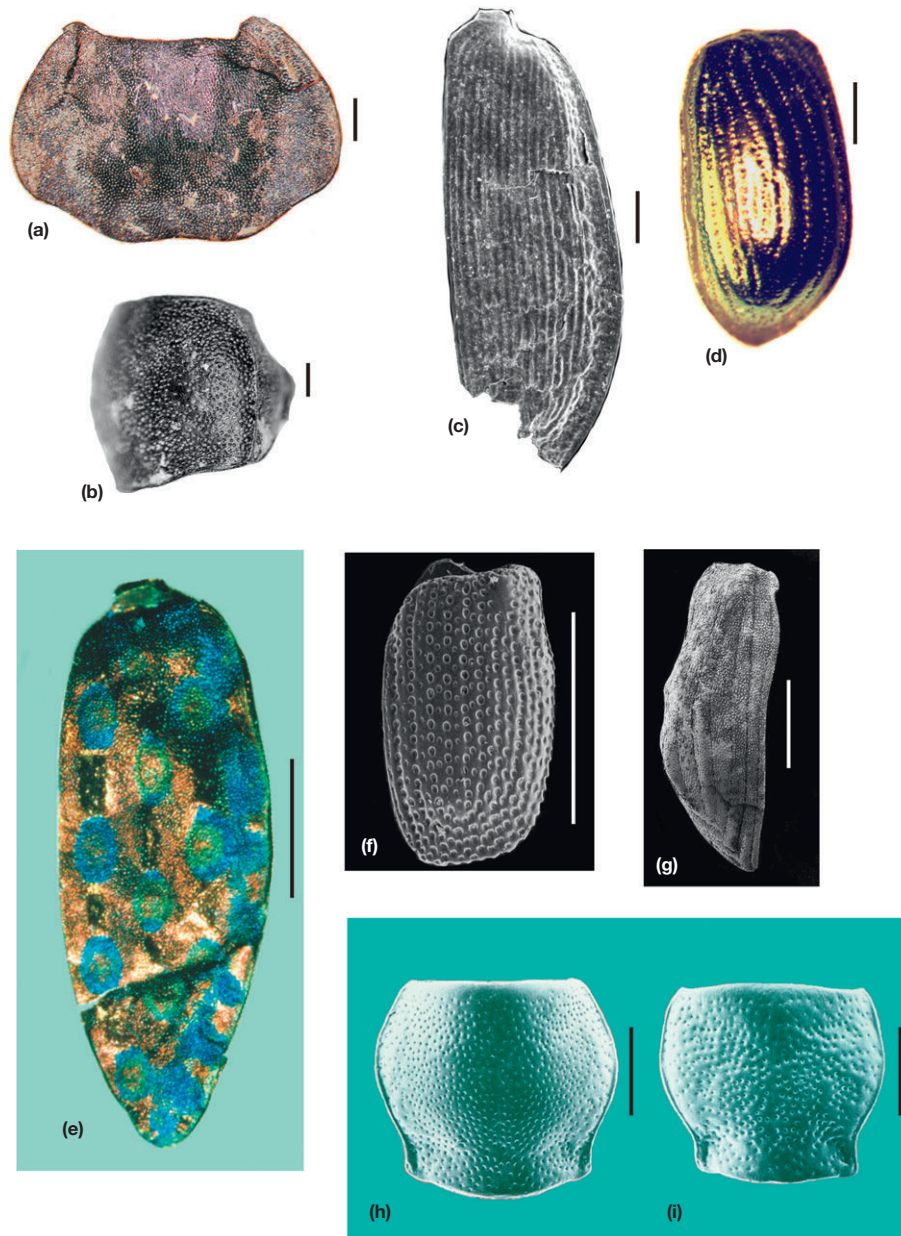


Figure 5 Light microscope photographs and scanning electron micrographs of fossil beetle specimens from Olyorian faunal assemblages at the Alazea site, northeastern Siberia ((a)–(d)) and the Kap København formation in Greenland ((e)–(h)). (a) Pronotum of *Thanatophilus dispar*; (b) pronotum of *Upis ceramoides*; (c) elytron of *Carabus odoratus*; (d) elytron of *Chrysolina brunnicornis*; (e) elytron of *Elaphrus tuberculatus*; (f) elytron of *Scolytus piceae*; (g) elytron of *Grypus equiseti*; (h) pronotum of *Diacheila polita* from Kap København; (i) pronotum of *Diacheila matthewsi*. Scale bars equal 1 mm. Photos (a)–(d) by Svetlana Kuzmina; photos (e)–(h) by Jens Böcher.

surprisingly, most of the insect fauna did not die out. The many species went their separate ways, some into eastern North America or southern parts of western North America, some to unglaciated regions of Beringia, others to other parts of Asia. Each species has its own unique history, its own remarkable story of survival in the face of shifting climates. Before the fossil data were published, few biologists would have guessed that 90% of the Arctic fauna of the Pliocene has remained extant to the present day. Thus, the trend of beetles shifting their ranges in response to climate change, first noted

for Pleistocene fossil assemblages (Coope, 1978), began far earlier than we might have expected.

See also: **Sea Level Studies:** Eustatic Sea-Level Changes – Glacial–Interglacial Cycles; **Plant Macrofossil Records:** Arctic Eurasia; **Beetle Records:** Postglacial North America; Late Pleistocene of Northern Asia; **Chironomid Records:** Postglacial Southern Hemisphere.

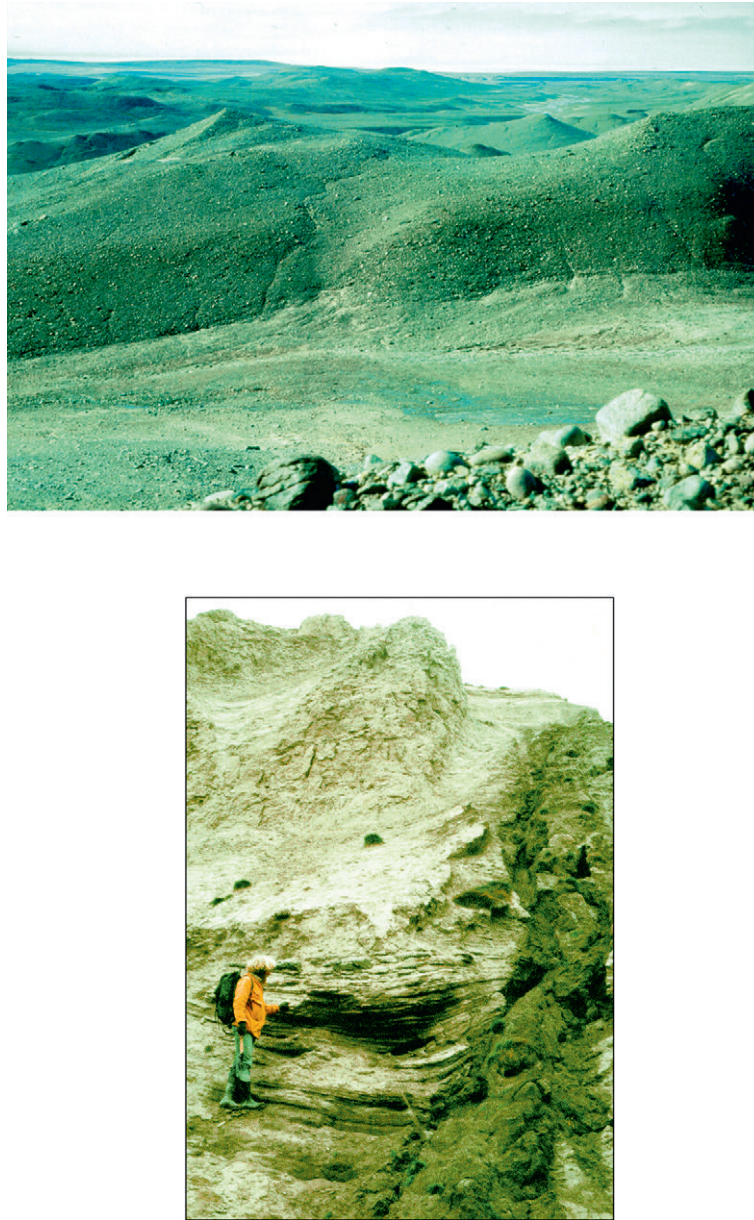


Figure 6 Kap København. Upper: view of the modern polar desert landscape; lower: view of the Early Quaternary fossil beds. Photographs by Jens Böcher.

References

- Atkinson TC, Briffa KR, Coope GR, Joachim M, and Perry D (1986) Climatic calibration of coleopteran data. In: Berglund BE (ed.) *Handbook of Holocene Palaeoecology and Palaeohydrology*, pp. 851–858. New York: Wiley.
- Böcher J (1995) Paleontomology of the Kap København Formation, a Plio-Pleistocene sequence in Peary Land, North Greenland. *Meddelelser om Grønland, Geoscience* 33: 82.
- Borror DJ and White RE (1970) *A Field Guide to the Insects of America North of Mexico*. Boston: Houghton Mifflin.
- Coope GR (1978) Constancy of insect species versus inconstancy of Quaternary environments. In: Mound LA and Waloff N (eds.) *Diversity of Insect Faunas, Symposium*, vol. 9, pp. 176–187. London: Royal Entomological Society of London.
- Danks HV (ed.) (1981) *Arctic Arthropods*. Ottawa: Entomological Society of Canada.
- Dowsett HJ, Barron JA, Poore RZ, et al. (1999) Middle Pliocene Palaeoenvironmental Reconstruction: PRISM2. U.S. Geological Survey Open File Report, 99–535.
- Elias SA (1994) *Quaternary Insects and Their Environments*. Washington, DC: Smithsonian Institution Press.
- Elias SA, Kuzmina S, and Kiselev S (2006) Late Tertiary origins of the Arctic beetle fauna. *Palaeogeography, Palaeoclimatology, Palaeoecology* 241: 373–392.
- Elias SA and Matthews JV Jr. (2002) Arctic North American seasonal temperatures in the Pliocene and Early Pleistocene, based on mutual climatic range analysis of fossil beetle assemblages. *Canadian Journal of Earth Sciences* 39: 911–920.
- Fyles JG, Hills LV, Matthews JV Jr., et al. (1994) Ballast Brook and Beaufort formations (Late Tertiary) on Northern Banks Island, Arctic Canada. *Quaternary International* 22/23: 141–172.
- Hopkins DM, Matthews JV, Wolfe JA, and Silberman ML (1971) A Pliocene flora and insect fauna from the Bering Strait region. *Palaeogeography, Palaeoclimatology, Palaeoecology* 9: 211–231.
- Kiselev SV (1981) *Late Cenozoic Coleoptera of Northeastern Siberia*. Moscow: U.S.S.R. Academy of Sciences, Paleontological Institute, Izdatel'stvo, Nauka Press (in Russian).

- Kiselev SV and Nazarov VI (2009) Late Cenozoic insects of Northern Eurasia. *Paleontological Journal* 43(supplement): 1–128.
- Kuzmina S (1989) Late Pleistocene insects from the Alazeya River (Kolyma lowland). *Bulletin MOIP, Geology Series* 64: 42–55 (in Russian).
- Kuzmina S, Elias SA, Matheus P, Storer JE, and Sher A (2008) Paleoenvironmental reconstruction of the Last Glacial Maximum, inferred from insect fossils from a tephra buried soil at Tempest Lake, Seward Peninsula, Alaska. *Palaeogeography, Palaeoclimatology, Palaeoecology* 267: 245–255.
- Matthews JV Jr. (1970) Two new species of *Micropeplus* from the Pliocene of western Alaska, with remarks on the evolution of Micropeplinae (Coleoptera: Staphylinidae). *Canadian Journal of Zoology* 48: 779–788.
- Matthews JV Jr. (1974) Quaternary environments at Cape Deceit (Seward Peninsula, Alaska): Evolution of a tundra ecosystem. *Geological Society of America Bulletin* 85: 1353–1384.
- Matthews JV Jr. (1976) Evolution of the subgenus *Cyphelophorus* (Genus *Helophorus*, Hydrophilidae): Description of two new fossil species and discussion of *Helophorus tuberculatus* Gyll. *Canadian Journal of Zoology* 54: 653–673.
- Matthews JV Jr. and Fyles JG (2000) Late Tertiary plant and arthropod fossils from the High Terrace Sediments on the Fosheim Peninsula of Ellesmere Island (Northwest Territories, District of Franklin). *Geological Survey of Canada, Bulletin* 529: 295–317.
- Matthews JV Jr. and Ovenden LE (1990) Late Tertiary plant macrofossils from localities in arctic/subarctic North America: A review of the data. *Arctic* 43: 364–392.
- Matthews JV Jr. and Telka A (1997) Insect fossils from the Yukon. In: Danks HV and Downes JA (eds.) *Insects of the Yukon*, pp. 911–962. Ottawa: Biological Survey of Canada (Terrestrial Arthropods).
- McDougall K (1995) Age of the Fishcreekian transgression. *Palaio* 10: 199–220.
- Minyuk PS (2004) *Magnetostratigraphy of the Cenozoic of North-East Russia*, p. 198. Magadan: NEISI Feb. RAS Press (in Russian).
- Shapiro B, Drummond AJ, Rambaut A, et al. (2004) Rise and fall of the Beringian steppe bison. *Science* 306: 1561–1565.
- Sher AV (1974) Pleistocene mammals and stratigraphy of the Far Northeast USSR and North America. *International Geology Review* 16: 7–10.
- Sher AV (1997) A brief overview of the Late-Cenozoic history of the Western Beringian lowlands. In: Edwards ME, Sher AV, and Guthrie RD (eds.) *Terrestrial Paleoenvironmental Studies in Beringia*, pp. 3–6. Fairbanks: University of Alaska Museum.
- White JM, Ager TA, Adam DP, et al. (1997) An 18 million year record of vegetation and climate change in northwestern Canada and Alaska: Tectonic and global climatic correlates. *Palaeogeography, Palaeoclimatology, Palaeoecology* 130: 293–306.
- White JM, Ager TA, Adam DP, et al. (1999) Neogene and Quaternary quantitative palynostratigraphy and paleoclimatology from sections in Yukon and adjacent Northwest Territories and Alaska. *Geological Survey of Canada, Bulletin* 543: 30.
- Willerslev E, Hansen AJ, Binladen J, et al. (2003) Diverse plant and animal genetic records from Holocene and Pleistocene sediments. *Science* 300: 791–794.

Middle Pleistocene of Western Europe

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Introduction

Insect fossils are abundant in Quaternary deposits that have remained unoxidized since their deposition, such as ancient pond-bottom sediments, infillings of paleochannels or fen peats. Accumulations of debris in archaeological contexts also yield abundant insect remains. The most conspicuous of these fossils are fragments of Coleoptera (beetles) whose skeletons are robust and resist decomposition. Coleopteran assemblages are now known from all Quaternary periods from widespread localities in most parts of the world. Almost all the fossils show a remarkable degree of evolutionary stability so that they can be exactly matched with living species. Furthermore, there is very little evidence of global extinction of any beetle species during this period. They do, however, show abundant evidence of widespread local extinctions as the whereabouts of acceptable habitats change with time. At least on the continental masses insect species frequently responded to climatic oscillations by changing their geographical ranges, often on an enormous scale; tracking the changing locations of suitable environments sometimes for thousands of kilometers. The fact of their evolutionary constancy and their geographical mobility makes the order Coleoptera a valuable tool in reconstructing past environments and, in particular, past climates.

In recent years, the complexity of the Quaternary stratigraphy of Europe has been partially resolved by the application of a notional framework based on the marine isotope stages (MIS). In these, global fluctuations in climate, oscillating between cold and warm periods, can be recognized. It has been possible to correlate, albeit provisionally, the climatic events recorded in terrestrial sequences with the MIS sequence. However, the climatic events on land, recorded in the biotic response, may differ significantly from the marine signal. This is shown in particular by the occurrence of thermophilous taxa of both plants and animals from middle Pleistocene interglacial deposits in Britain. Using the mutual climatic range (MCR) method (Atkinson et al., 1987) on carnivorous or scavenging beetle species, it has been possible to make quantified paleotemperature estimates that are independent of those based on the macroflora. These show that the middle Pleistocene climate in Britain was sometimes warmer than that of the present day in contrast to the marine record which suggests that it was cooler globally (Candy et al., 2010).

In this article, the base of the middle Pleistocene is taken as the Brunhes/Matuyama geomagnetic boundary during MIS 19 and its upper limit as the base of MIS 5e, at the start of the Last (Ipswichian, Eemian, Wisconsinan) Interglacial. The beetle assemblages will, wherever possible, be allocated to the various MIS events. This means that several of the earlier descriptions of beetle assemblages which may well be of middle Pleistocene age will not be mentioned here because it is difficult to relate these

to the modern Quaternary stratigraphy. However, most of these early papers have been included in *A Bibliography and Literature Review of Quaternary Entomology* (Buckland and Coope, 1991).

Middle Pleistocene Beetle Assemblages

It is convenient to subdivide these middle Pleistocene interglacial beetle faunas into two groups: those that predate the extensive Anglian (Saalian) Glaciation (MIS 12) and those which postdate this major glacial event.

Pre-Anglian Middle Pleistocene Beetle Localities

Several organic deposits at the foot of the Norfolk coastal cliffs have in the past been grouped together as the 'The Cromer Forest Bed Formation.' These are now known to represent a complex sequence of interglacials that can only be provisionally attributed to any particular MIS. The deposits are highly fossiliferous and include abundant insect remains, which clearly indicate very different climates. All these deposits predate the Anglian (Saalian) Glaciation (=MIS 12). The occurrence of numerous temporary exposures of organic deposits in active gravel workings at inland localities means that the understanding of their biostratigraphy based largely on vertebrate fauna is rapidly developing. A brief summary of the more important sites will be given here but it must be borne in mind that others are currently under investigation and will be published in the near future.

Beeston

The earliest beetle assemblage from this period was obtained from deposits at Beeston, north Norfolk and can be correlated with Pre-Pastonian (West, 1980). Because they rest directly on the Chalk it is difficult to give them a precise age but they probably are equivalent to MIS 18 or MIS 20. They are possibly equivalent to the Tiglian deposits of the Netherlands. They are included here because the insect fauna is the earliest Pleistocene assemblage from the British Isles (Coope, 2000). This fauna consists of 15 taxa of which 11 could be determined to species. All are of extant species with the exception of a species of the water beetle *Agabus*, which has not yet been identified. MCR paleotemperature estimates were equivalent to northern Britain at the present day.

Broomfield

Details of the stratigraphy and paleontology of this site are given in Gibbard et al. (1996). At the moment, it is not possible to correlate these deposits precisely with any MIS event. Altogether 69 beetle taxa were recovered from the organic horizon of which 52 could be identified to species. The assemblage included a large number of species characteristic of running water and marginal riverine habitats. The discovery here

[†]Deceased.

of beetle species with southern European distributions at the present day indicates that the mean July temperature was at least 19 °C and probably a degree or so warmer than this, in other words somewhat warmer than southern England today.

Pakefield

This site has for long been known for its fossils of large vertebrates. An account of the rich beetle fauna has been recently published (Parfitt et al., 2005). The most likely correlation of these deposits is with MIS 17. Here, as at Broomfield, the fauna was characterized by numerous riverine and marsh species indicating that this deposit originated in a mature river with both shallow riffles and quieter pools in which there were submerged trunks of decomposing oak trees. The beetles indicate the presence of dried out carcasses that were partially consumed by maggots. The beetle assemblage included a number of southern European species suggesting that the climate was warmer than that of the present day. Quantified estimates of the paleoclimate using the MCR method on this beetle assemblage (Atkinson et al., 1987) indicate that the mean July temperature lay between 17 and 23 °C and that the mean January/February temperature lay between –6 and 4 °C.

Sugworth

A coleopteran assemblage has been described from a preglacial channel deposit at Sugworth near Oxford (Osborne, 1979). Again, this fauna indicates that the deposit was largely riverine. It also included several exclusively southern European species. The exclusive presence at both Sugworth and at Pakefield, of fossils of the exotic species *Oxytelus* (= *Anotylus*) *opacus*, a south eastern European species with a distribution centered on the Danube basin, suggests that both sites may be of the same age. MCR estimates on this beetle fauna indicate that the mean July temperature also lay between 17 and 22 °C and the mean January/February temperature lay between –6 and 4 °C.

West Runton Freshwater Bed

In 1995, excavations were carried out of a partial mammoth skeleton at West Runton near Cromer, north Norfolk: the type section for the Cromer Forest Bed Formation. Samples of the associated sediment were analyzed for beetle remains (Coope, 2000, 2010). The upper parts of the sedimentary sequence were devoid of any insect fossils probably due to their removal by oxidation by contemporaneous weathering. The beetles indicate a reedy marsh with any open water overgrown with the duckweed *Lemna*. Oak, elm and coniferous trees grew nearby. Of particular interest is the scarcity of dung beetles in this assemblage suggesting that the carcass may have been transported *postmortem* into its present location. MCR estimates of the climate, based on the beetle assemblage, indicate that mean July temperatures lay between 16 and 19 °C and mean January/February temperatures lay between –6 and 5 °C.

Sidestrand

About 15 km southeast of West Runton, exposures of the Cromer Forest Bed Formation occasionally outcrop at the foot of the cliff. The coleopteran assemblage includes a number of thermophilous species that were absent from West Runton indicating that the two deposits were not of the same age

(Preece et al., 2009). Here again the paleotemperatures were rather higher than in present day Norfolk.

Happisburgh 1

An organic deposit is exposed from time to time below high tide level at Happisburgh. It predates the earliest evidence of glaciation in Britain. This site also yielded a flint hand axe. The geological and archaeological contexts are given by Parfitt (2005). The insect fauna from the organic bed indicates stagnant water surrounded by a reedy marsh. It includes several species that have relatively northern distributions at the present day. MCR estimates on the beetle assemblage indicate that the mean July temperature lay between 12 and 15 °C and the mean January/February temperatures lay between –11 and –3 °C (Coope, unpublished data).

Happisburgh 3

It is important to distinguish the Happisburgh 1 locality from the nearby site of Happisburgh 3 which is older than the Brunhes/Matuyama geomagnetic boundary and therefore technically Lower Pleistocene in age. The beetle fauna from Happisburgh 3 is very rich and indicates climatic cooling from temperatures as warm as southern England today to conditions rather cooler than modern. Happisburgh 3 has produced sharp, unrolled flint flakes indicating the earliest recorded presence, so far known, of human occupation of northern Europe (Parfitt et al., 2010a).

Ardleigh

An interglacial beetle fauna has been described from middle Pleistocene deposits at Ardleigh, Essex, which is possibly equivalent to MIS 17 or 15 (Gibbard et al., unpublished). The fauna is indicative of a cool-temperate climate not much different from that in England at the present day. However, insufficient species were present to permit MCR paleotemperature reconstructions.

High Lodge

This site near Mildenhall, Suffolk, has for long been famous as a source of Lower Paleolithic flint implements. Beetle assemblages have been recovered from the organic silts which yielded a beetle fauna of 47 taxa of which 32 could be identified to species (Coope, 1992). The local environment was dominated by reedy swamp conditions. MCR estimates indicate that mean July temperatures lay between 15 and 16 °C and the mean January/February temperatures lay between –4 and 1 °C.

Waverley Wood

This site is situated south of Coventry, Warwickshire. The organic deposits were located in a series of channels directly underneath glacial deposits now attributed to the Anglian (Saalian) Glaciation. The precise age of these channels is difficult to ascertain other than that they are pre-Anglian. A large beetle assemblage was obtained from one of the lowermost of these channels (Shotton et al., 1993). The beetles indicate that sedimentation commenced in running water which later developed into a pond surrounded by a marsh. The presence of *Agonum quadripunctatus* is important because this species is attracted by forest fires and it is thus significant that charcoal

fragments occurred throughout the deposit. The climatic history during the channel filling is complex (see [Shotton et al., 1993](#), Figure 9). At the base of the Channel 2, MCR estimates indicate that the mean July temperatures lay between 15 and 18 °C and the mean January/February temperatures lay between –13 and 2 °C. However, at the top of the Channel 2 the mean July temperatures have deteriorated to between 8 and 11 °C with mean January/February values to between –25 and –10 °C. At the base of the succeeding Channel 3, temperatures had returned once again to the same cool-temperate conditions that prevailed at the base of Channel 2; evidence of a considerable climatic oscillation. A similar intense cold interlude in an otherwise temperate episode is indicated by the beetle fauna from the MIS 11 interglacial at Quinton ([Horton, 1989](#)). A struck quartzite flake in mint condition was found *in situ* within the silts of the lowest channel, indicating the presence of humans in the English Midlands living under similar climatic conditions to those outlined above.

Pool's Farm Pit

A similar beetle assemblage to that at Waverley Wood Pit has been recorded from a nearby gravel pit at Pools Farm near Brandon, Warwickshire ([Maddy et al., 1994](#)).

Mathon

This site was in the old gravel workings at The Brays Pit, near Mathon, Herefordshire, in the western English Midlands. The beetle assemblage was obtained from organic sediments at the base of the pit, underlying glacial deposits which in turn were cut into by a channel filled with interglacial deposits attributed, on palynological grounds, to the MIS 12 interglacial. The beetle fauna from the basal sediments was dominated by running water species and others that are dependent on coniferous trees ([Coope et al., 2002](#)). The climatic implications of this fauna are interesting in that one of the most abundant of these running water species, *Dupophilus brevis*, is not found today anywhere in northern regions but is confined to south-western Europe where it lives in streams that are relatively cool in summer. The conifer forest at Mathon was not therefore the strict analog of the modern boreal forests but rather resembles those of central Europe at the present day.

Coleopteran Faunas and the Possible Correlation of Middle Pleistocene Deposits

The four sites Happisburgh 1, Waverley Wood, Pool's Farm Pit, and Mathon present an interesting problem in correlation. In the absence of any evolutionary change among beetles and little evidence of global extinctions during the Quaternary, it is not possible to use the classical methodologies for paleontological correlation: it is always possible for similar assemblages to come together at different times whenever similar environments arise. Thus, gross faunal similarities do not necessarily indicate contemporaneity. However, the presence of exotic species carries more stratigraphical weight: the more unexpected the fossil occurrence of a species, the more likely it is to have stratigraphic value. The faunas from the above four sites have several exotic species in common ([Coope, 2006](#)). Perhaps the most unexpected of these is the small staphylinid species *Micropeplus hoogendorni*, a distinctive species originally

described from Pliocene deposits of Alaska that were at least 5.7 million years old ([Matthews, 1970](#)). The species was believed by Matthews to be extinct. However, this species has now been recognized as identical with the living Siberian species *M. dokuchaevi*. It is therefore interesting to note the presence in Britain of this exotic species at all four of these sites where it must have been fairly abundant. So far, this species has been found as a fossil nowhere else. The presence of unexpected species such as the above example strongly suggests that all four of these deposits are of the same age.

Coleopteran Faunas that Probably Date from Early in the Anglian Glaciation (MIS 12)

Ardleigh J, Essex

Directly overlying the Ardleigh interglacial deposits described above, was a lens of organic sediment interbedded within the gravels and named Ardleigh J. This horizon can probably be attributed to an early climatic stage of the Anglian (Saalian) Glaciation (MIS 1; [Bridgland and Allen, 1996](#)). A beetle assemblage from this deposit was dominated by high arctic species which included *Diacheila polita*, *Bembidion dauricum*, *Helophorus obscurus*, and *Holoboreaphilus nordenskiöldi*. Of particular interest was the presence also of *H. arcticus* whose modern range is in northernmost North America and the extreme northeast of Siberia in Kamchatka. This fossil occurrence is the only record of this species outside its present day range. MCR estimates based on this assemblage indicate that mean July temperatures lay somewhere between 10 and 11 °C and mean January/February temperatures lay between –14 and –10 °C.

Ostend, Norfolk

The 'Arctic Fresh-water Bed' which overlies the Cromer Forest Bed Formation was exposed for a short period at the bottom of the cliff at Ostend, Norfolk. The organic deposit was highly fossiliferous and included several arctic species of beetle including *Pterostichus brevicornis*, *P. middendorfi*, and *H. obscurus* ([Parfitt et al., 2010b](#)). MCR estimates based on the coleopteran assemblage indicated that the mean July temperature lay somewhere between 9 and 11 °C and mean January/February temperatures somewhere between –36 and –10 °C.

Brooksby, Leicestershire

Recent gravel extraction near Brooksby has exposed organic deposits filling a paleochannel, which has yielded a large coleopteran fauna that includes many cold-adapted species such as *H. obscurus*, *H. glacialis*, *Tachinus caelatus*, and *Hippodamia arctica* (Coope, unpublished data). MCR estimates based on the beetle assemblage suggest that mean July temperatures were somewhere between 11 and 13 °C and mean January/February temperatures somewhere between –17 and –11 °C. The paleochannel underlies glacial deposits of Anglian (MIS 12) age and this arctic fauna may represent an early phase of this glaciation. However, since at Waverley Wood, Channel 2 showed the gradual development of a similar cold-adapted beetle assemblage as part of a larger climatic oscillation, it is possible that the channel deposits recently exposed at Brooksby should be correlated with the terminal phase of some earlier interglacial. Of particular interest at this site is the occurrence of Paleolithic hand axes in the gravels

directly overlying the channel deposits, though as yet their exact provenance has not been determined because they were all recovered from the heap of stones rejected by the gravel company. However, as extraction progresses, definitive sections are likely to be exposed which will resolve the question of their stratigraphical context.

Early Middle Pleistocene Coleopteran Faunas and Early Lower Paleolithic Archaeology

Several of the insect faunas mentioned above are associated with evidence of early human occupation of Britain; biface hand axes and/or characteristically struck flint flakes. At some sites there were also bones that had apparently been deliberately broken open probably to extract the marrow. The beetle faunas provide intimate information about the living conditions at these times.

The fossiliferous deposits at Happisburgh 3 contain flint flakes that are the earliest evidence of human presence in northern Europe (Parfitt et al., 2010a). The beetles indicate that the people lived beside a large meandering river which occasionally inundated its floodplain. The beetle evidence also shows that the climate at that time was as warm as southern England today but, during this period of human occupation, it deteriorated to conditions at least as cold as northern Britain at the present time.

At Pakefield, on the other hand, humans occupied Britain at a time when the climate was substantially warmer than it is in Suffolk today (Parfitt et al., 2005).

At Waverley Wood at least one implement was found in place within the sediments of the lowest channel in climatic conditions not much different from those of the present day.

The sites at Happisburgh 1, and High Lodge yielded beautifully made hand axes in mint condition. The associated beetle assemblages indicate open marshland with nearby woodlands made up of both deciduous and coniferous trees, but at a time when the climate was rather cooler than that of the present day.

Post-Anglian Middle Pleistocene Beetle Assemblages

The ice sheets of the Anglian (Saalian) glacial stage covered most of the British Isles and much of north-west Europe and their deposits are relatively easy to recognize and correlate over a wide area. This event is widely attributed to MIS 12. It is convenient to use this glacial event to subdivide the middle Pleistocene insect faunas. For the most part these insect assemblages have been obtained from temperate interglacial deposits although the presence of interbedded cold-adapted coleopteran faunas at two sites toward the end of the interglacial, indicate climatic complexity.

Interglacials Correlated with MIS 11

Large coleopteran assemblages have been recorded from five sites in Britain that can be attributed to MIS 11.

Hoxne, Suffolk

This site is near the village of Hoxne in an abandoned brick pit that was, for many years famous as a source of beautifully made flint hand axes. The fossiliferous deposit is largely lake

clay. Extensive excavations have been made to investigate the archaeology, vertebrate and invertebrate paleontology and also the paleobotany of this site. At the base of the sequence, insect fossils have been obtained from the early cold phase that preceded the interglacial proper, from what may be considered as the final stage of the preceding glaciations (MIS 12) or the beginning of the Hoxnian Interglaciation (MIS 11). Beetles from a small sample of this deposit included such cold-adapted species as *D. polita*, *B. hasti*, *Oreodytes alpinus*, *T. caelatus*, and *H. nordenskiöldi*, indicating that the climate was arctic and continental (Coope, 1993).

The beetle fauna from the overlying interglacial deposits indicates still water surrounded by lush, marshy, meadow-like habitats in a mature forest of various deciduous trees, many of which were dead and decomposing. Recent investigations of the beetle faunas show that, after this temperate phase there was a return to arctic-continental conditions with the presence of exclusively arctic and Siberian species such as *H. obscurus* and *H. nordenskiöldi* (Ashton et al., 2008).

Quinton, West Midlands

Interglacial lake deposits were encountered in the construction of the M5 motorway 10 km south of Birmingham. On palynological grounds, these deposits have been equated with those from Hoxne. The insect fauna was sparse in the lower part of the sequence, but indicated warm temperate conditions somewhat warmer than those of the present day. In the upper part, where the insect fauna was richer, it showed an episode of sudden and intense climatic deterioration with the appearance of exclusively arctic-continental species such as *D. arctica*, *H. obscurus*, and *H. nordenskiöldi*. Immediately after this cold episode, a cool-temperate beetle fauna returned. This recurrence cannot be attributed to the redeposition of fossils eroded out from the earlier temperate sediments, since the earlier suite of temperate species differs considerably from those in the later temperate deposit (Coope and Kenward, 2007). This sudden and intense cold oscillation has much in common with the climatic fluctuation recognized during the later part of the same interglacial at Hoxne (see Section 'Hoxne, Suffolk').

Nechells, Birmingham, West Midlands

A sequence of beetle assemblages has been described from borehole samples through interglacial lake deposits correlated with MIS12 (Shotton and Osborne, 1965). Unfortunately, only the first part of the interglacial was represented at Nechells. Interesting members of this assemblage are southern European species that depend on trees; the weevil *Stenoscelis* (= *Brachytemnus*) *submuricatus* and the bark beetle *Platypus oxyurus* which lives exclusively on the silver fir, *Abies*.

Woodston Beds, Peterborough, Cambridgeshire

Quaternary deposits have been exposed for many years as channel-like deposits in the overburden being cleared from the Oxford clay used for brick making (Horton et al., 1992). The insect assemblage was typically interglacial and was dominated by running water species as well as indications of still or slowly flowing water, suggesting a mature river meandering across its floodplain. The likely age of these interglacial deposits is MIS 11 but here again, only the earlier part of the interglacial was represented.

Frog Hall Pit, Warwickshire

Details of the stratigraphy and paleontology of the organic deposits at Frog Hall Pit have been described by Keen et al. (1997). The beetle assemblage indicates a river meandering across a marshy floodplain flanked by oak woodland. The climate was temperate. MCR estimates of the paleoclimate indicate that mean July temperatures lay between 16 and 18 °C and mean January/February temperatures lay between –8 and 12 °C. Precipitation was sufficient to maintain a year-round flow in the river. Unfortunately, here again only the early part of the interglacial was represented at Frog Hall.

Interglacial Beetle Assemblages Correlated with MIS 9

Only three beetle faunas can be attributed to this interglacial.

Hackney, North London

A sequence of organic silts was exposed during building work at the Nightingale Estate. They yielded the only large beetle fauna that can be attributed to MIS 9. It comprised 254 taxa of which 181 could be identified to species (Green et al., 2005). These provide a complex mosaic picture of the local environment. Much of the deposit originated in a mature meandering river with shallow riffles and deeper pools. It is possible that some of the beetle species in this assemblage inhabited oxbow lakes on the floodplain. Nearby were marshes and damp meadow-like habitats frequented by large herbivorous mammals. The adjacent woodland consisted of both coniferous and deciduous trees. This assemblage includes many thermophilous beetle species. MCR estimates of the thermal climate gave mean July temperatures of about 18 °C or somewhat warmer mean January/February temperatures were between –4 and 1 °C. Here again adequate precipitation must have been available to keep the river flowing throughout the year.

Barling, Essex

During a multidisciplinary investigation, insect fossils were obtained from organic sediments filling a channel in a gravel pit on the north side of the Thames estuary (Bridgland et al., 2001). The beetle assemblage comprised 29 taxa of which 15 could be identified to species. Here again the Coleoptera indicate a riverine and swampy grassland grazed by large herbivorous mammals. Abundant Paleolithic implements have been found in the immediate vicinity. Thermophilous beetle species include the dung beetle *Caccobius schreberi*, and rotting wood beetle *Valgus hemipterus*. MCR estimates based on the beetle fauna indicate that mean July temperatures lay between 17 and 26 °C and that mean January/February temperatures lay between –11 and 13 °C.

Cudmore Grove, Mersea Island, Essex

Organic sediments occupying a broad channel-like depression are exposed at the foot of the sea cliff. A rich beetle assemblage was obtained from near the base of this sequence (Roe et al., 2009) comprising 52 taxa of which 37 could be named to species. Here, the beetles indicate slowly moving or stationary water with reedy marshes and grasslands where large mammals were feeding. Of outstanding interest in this assemblage, are beetles that depend on dead timber, such as *Osmoderma eremite*

and *Prionus coriarius*. The presence of exotic thermophilous species indicates that climate was warmer than that of the present day. MCR estimates show that mean July temperatures lay between 16 and 22 °C and mean January/February temperatures lay between –7 and 4 °C.

Interglacials Correlated with MIS 7

Five large insect faunas are known that can be attributed to this climatically complex period.

Stanton Harcourt, Oxfordshire

An account of the stratigraphy and paleontology of this site have been given by Briggs et al. (1985). The fossiliferous deposits were in a channel cut into the Oxford clay. A single 10 kg sample yielded 122 beetle taxa of which 88 could be identified to species. This fauna indicates a moderately large river with riffles and pools with bare clay banks in places and lush marshes alongside. This habitat attracted many large herbivorous mammals. Drier ground nearby was thinly vegetated with heathers and oaks. MCR estimates of the climate indicated mean July temperatures between 16 and 18 °C with mean January/February temperatures close to 0 °C. Numerous Paleolithic implements were found at this site.

Marsworth, Buckinghamshire

The complicated stratigraphy and paleontology of this site has been described by Murton et al. (2001). A beetle fauna from the Lower Channel comprised 92 taxa of which 57 could be named to species. There is a curious rarity of water beetles and a complete absence of the species indicative of running water suggesting that when the sediment was laid down the area had become a marsh surrounded by drier sandy soil. MCR estimates indicate that the mean July temperature lay between 15 and 17 °C and mean January/February temperatures lay between –9 and 1 °C.

Stoke Goldington, Bedfordshire

The stratigraphy and paleontology of this site has been described by Green et al. (1996). A beetle assemblage of 91 taxa, of which 60 could be identified to species, was obtained from the deposits in the 'lower channel.' This assemblage indicated very slowly moving or stationary water with much swamp vegetation, notably the sweet grass *Glyceria*, which is attractive to herbivorous mammals, notably abundant mammoths. The beetle assemblage indicates similar climatic conditions to those at Marsworth.

Upper Strensham, Worcestershire

Organic deposits were encountered in a temporary exposure together with numerous mammoth bones (De Rouffignac et al., 1995). The associated beetle assemblage comprised 119 taxa of which 79 could be identified to species. They indicate that the sediment accumulated in a pond surrounded by meadow-like habitats. In spite of the apparent rarity of trees, the beetles were entirely temperate, suggesting that the mean July temperature was about 16 °C and mean January/February temperatures were about 0 °C.

Aveley, Essex

Temporary exposures near Aveley, Essex, have attracted attention because of the occurrence of mammalian bones. These deposits are now generally accepted as being of MIS 7 age. Associated organic sediments have yielded large insect faunas (Schreve et al., in press). They indicate an extensive marsh with adjacent grassland and oak woodland. The climate was temperate. MCR estimations based on the beetle fauna indicated that mean July temperatures lay between 17 and 18 °C and mean January/February temperatures lay between –2 and 0 °C.

Cold Stage Correlated with MIS 6

Balderton Sand and Gravel Formation

Terrace deposits of the river Trent between Newark and Lincoln have, on geological grounds, been equated with MIS 6 and are associated with a cold stage mammalian fauna. At several sites, organic sediments near to the base of these gravels have yielded a beetle assemblage that included numerous exclusively arctic-continental species (Brandon and Sumbler, 1991; Coope and Taylor, 1991). These included *D. polita*, *D. arctica*, *H. obscurus*, *H. sibiricus*, *H. jacutus*, *Boreaphilus henningianus*, and *T. caelatus*. The local environment was reminiscent of tundra. MCR estimates based on the beetle faunas indicated that the mean July temperature lay between 10 and 13 °C and the mean January/February temperatures lay between –25 and –12 °C.

Middle Pleistocene Coleopteran Studies in Western Europe

Two important middle Pleistocene coleopteran assemblages have been described from Western Europe. Firstly, from interglacial deposits at Dingé, in northwest France, a large beetle assemblage has been described as part of a multidisciplinary study (Andrieu et al., 1997). This fauna indicative of a warm temperate climate, is believed to date from MIS 7 or 9. Secondly, an extensive insect assemblage has been described as part of a multidisciplinary study from interglacial deposits at La Côte in the western French Alps that probably date from MIS 11 (Field et al., 2000). The wealth of paleoecological data that is emerging from these investigations of coleopteran assemblages points to exciting new opportunities throughout western Europe for reconstructing past local environments and also for giving quantified estimates of past climatic conditions.

Conclusion

The sequence of coleopteran assemblages outlined above provides an ecological picture that is consistent with the other physical and biological environmental indicators. Thus, the fossil beetle assemblages show constancy in their environmental requirements as well as stability in the morphology of their component species. Using the MCR method, it has been possible to quantify climatic parameters for both interglacial and glacial events. In particular, many middle Pleistocene

interglacial faunas from Britain include southern European species indicating that the climate was substantially warmer at those times than it is at the present day. During the intervening cold periods, the presence of exotic cold-adapted beetle species in Britain showed that the climate was of arctic and extreme continental severity. In stratigraphical correlation, the presence of unexpected exotic species in assemblages may provide better indicators of similarity of age than their gross faunal (or floral) similarity.

See also: [Beetle Records: Late Pleistocene of Europe; Overview; Postglacial Europe.](#)

References

- Andrieu V, Field MH, Ponel P, et al. (1997) Middle Pleistocene temperate deposits at Dingé, Ille-et-Vilaine, northwest France. *Journal of Quaternary Science* 12(4): 309–331.
- Ashton N, Lewis SG, Parfitt SA, Penkman KEH, and Coope GR (2008) New evidence for complex changes in MIS 11 from Hoxne Suffolk, UK. *Quaternary Science Reviews* 27: 652–668.
- Atkinson TC, Briffa KR, and Coope GR (1987) Seasonal temperatures in Britain during the past 22,000 years, reconstructed using beetle remains. *Nature (London)* 325: 587–592.
- Brandon A and Sumbler MG (1991) The Balderton Sand and Gravel; pre-Ipswichian cold stage fluvial deposits near Lincoln. *Journal of Quaternary Science* 6: 117–138.
- Bridgland DR and Allen P (1996) A revised model for terrace formation and its significance for the early Middle Pleistocene terrace aggradations of north-east Essex, England. In: Turner C (ed.) *The Early Middle Pleistocene in Europe*, pp. 121–134. Rotterdam: Balkema.
- Bridgland DR, Preece RC, Roe HM, et al. (2001) Middle Pleistocene interglacial deposits at Barling, Essex, England: Evidence for a longer chronology for the Thames terrace sequence. *Journal of Quaternary Science* 16: 813–840.
- Briggs DJ, Coope GR, and Gilbertson DD (1985) The Chronology and Environmental Framework of Early Man in the Upper Thames Valley. British Archaeological Reports, 137. Oxford: 1–176.
- Buckland PC and Coope GR (1991) *A Bibliography and Literature Review of Quaternary Entomology*. Sheffield: John Collis Publications, University of Sheffield.
- Candy I, Coope GR, Lee JR, et al. (2010) Pronounced warmth during early Middle Pleistocene interglacials: Investigating the Mid-Brunhes Event in the British terrestrial sequence. *Earth-Science Reviews* 103: 183–196.
- Coope GR (1992) The High Lodge insect fauna. In: Ashton NM, Cook J, Lewis SG, and Rose J (eds.) *High Lodge, Excavations by G.de.G. Sieveking, 1962–8, and J. Cook, 1988*, pp. 117–119. London: British Museum Press.
- Coope GR (1993) Late-Glacial (Anglian) and Late-Temperate (Hoxnian) Coleoptera. In: Singer R, Gladfelter BG, and Wymer JJ (eds.) *The Lower Palaeolithic Site at Hoxne, England*, pp. 156–162. Chicago: University of Chicago Press.
- Coope GR (2000) Coleoptera from Beeston and West Runton, Norfolk. In: Lewis SG, Whiteman CA, and Preece RC (eds.) *The Quaternary of Norfolk and Suffolk: Field Guide*, pp. 73–75. London: Quaternary Research Association.
- Coope, GR (2006) Insect Faunas associated with Palaeolithic industries from five sites of pre-Anglian age in Central England. *Quaternary Science Reviews* 25: 1738–1754.
- Coope GR (2010) Coleoptera from the Cromerian type site at West Runton, Norfolk, England. *Quaternary International* 228: 46–52.
- Coope GR, Field MH, Gibbard PL, Greenwood M, and Richards AE (2002) Palaeontology and biostratigraphy of Middle Pleistocene river sediment in the Mathon Member, at Mathon, Herefordshire, England. *Proceedings of the Geologists Association* 113: 237–258.
- Coope GR and Kenward HK (2007) Evidence for a short but intense cold interlude during the late-temperate stage of the Hoxnian Interglacial (MIS11) based on coleopteran assemblages from Quinton, West Midlands, UK. *Quaternary Science Reviews* 26: 3276–3285.
- Coope GR and Taylor BJ (1991) Coleoptera from the Balderton Sand and Gravel. In: Brandon A and Sumbler MG (eds.) *The Balderton Sand and Gravel: Pre-Ipswichian Cold Stage Fluvial Deposits Near Lincoln, England*, *Journal of Quaternary Science* 6: 130–133.

- De Rouffignac C, Bowen DQ, Coope GR, et al. (1995) Late Middle Pleistocene interglacial deposits at Upper Strensham, Worcestershire, England. *Journal of Quaternary Science* 19: 15–31.
- Field MH, de Beaulieu J-L, Guiot J, and Ponel P (2000) Middle Pleistocene deposits at La Côte, Val-de-Lans, Isère department, France: Plant macrofossils, palynological and fossil insect investigations. *Palaeogeography, Palaeoclimatology, Palaeoecology* 159: 53–83.
- Gibbard PL, Aalto MM, Coope GR, et al. (1996) In: Turner C (ed.) *The Early Middle Pleistocene in Europe*, pp. 83–119. Rotterdam: Balkema.
- Green CP, Branch NP, Coope GR, et al. (2005) Marine Isotope Stage 9 environments of fluvial deposits at Hackney, North London, UK. *Quaternary Science Reviews* 25(1–2): 89–113.
- Green CP, Coope GR, Jones RL, et al. (1996) Pleistocene deposits at Stoke Goldington, in the valley of the Great Ouse, UK. *Journal of Quaternary Science* 11: 59–87.
- Horton A (1989) Quinton. In: Keen DH (ed.) *West Midlands Field Guide*, p. 69. Cambridge: Quaternary Research Association.
- Horton A, Keen DH, Field MH, et al. (1992) The Hoxnian Interglacial deposits at Woodston, Peterborough. *Philosophical Transactions of the Royal Society of London* 338: 131–164.
- Keen DH, Coope GR, Jones RL, et al. (1997) Middle Pleistocene deposits at Frog Hall Pit, Stretton-on-Dunsmore, Warwickshire, English Midlands, and their implications for the age of the type Wolstonian. *Journal of Quaternary Science* 12(3): 183–208.
- Maddy D, Coope GR, Gibbard PL, Green CP, and Lewis SG (1994) Reappraisal of Middle Pleistocene fluvial deposits near Brandon, Warwickshire and their significance for the Wolston glacial sequence. *Journal of the Geological Society* 151: 221–233.
- Matthews JV Jr. (1970) Two new species of *Micropeplus* from the Pliocene of western Alaska with remarks on the evolution of Micropepliinae (Coleoptera, Staphylinidae). *Canadian Journal of Zoology* 48: 779–788.
- Murton JB, Baker A, Bowen DQ, et al. (2001) A late Middle Pleistocene temperate–periglacial temperate sequence (Oxygen Isotope Stages 7–5e) near Marsworth, Buckingham, UK. *Quaternary Science Reviews* 20: 1787–1825.
- Osborne PJ (1979) The insect fauna of the organic deposits at Sugworth and its environmental and stratigraphical implications. *Philosophical Transactions of the Royal Society of London* B289: 119–133.
- Parfitt SA (2005) A butchered bone from Norfolk: Evidence for early human presence in Britain. *Archaeology International* 8: 14–17.
- Parfitt SA, Ashton NM, Lewis SG, et al. (2010a) Early Pleistocene human occupation at the edge of the boreal zone in northwest Europe. *Nature (London)* 466(8): 229–233.
- Parfitt SA, Barendregt RW, Breda M, et al. (2005) The earliest record of human activity in northern Europe. *Nature (London)* 438/15: 1008–1012.
- Parfitt SA, Coope GR, Field MH, Pegler SM, Preece RC, and Whittaker JE (2010b) Middle Pleistocene biota of the early Anglian ‘Arctic Fresh-water Bed’ at Ostend, Norfolk, UK. *Proceedings of the Geologists Association* 121: 55–65.
- Preece RC, Parfitt SA, Coope GR, Penkman KEH, Ponel P and Whittaker JE (2009) Biostratigraphic and aminostratigraphic constraints on the age of the Middle Pleistocene glacial succession in north Norfolk, UK. *Journal of Quaternary Science* 24: 557–580.
- Roe HM, Coope GR, Devoy RJN, et al. (2009) Differentiation of MIS 9 and MIS 11 in the continental record: Faunal, aminostratigraphical and sea-level evidence from coastal sites in Essex, UK. *Quaternary Science Reviews* 28: 2342–2373.
- Schreve DC, Bridgland DR, Allen P, et al. (in press) Late Middle Pleistocene River Thames terrace deposits at Aveley, Essex, UK: A multiproxy framework for the penultimate (MIS 7) interglacial. *Quaternary Science Reviews*.
- Shotton FW, Keen DH, Coope GR, et al. (1993) The Middle Pleistocene deposits at Waverley Wood Pit, Warwickshire, England. *Journal of Quaternary Science* 8(4): 293–325.
- Shotton FW and Osborne PJ (1965) The fauna of the Hoxnian Interglacial deposits of Nechells Birmingham. *Philosophical Transactions of the Royal Society of London* 248B: 353–378.
- West RG (1980) *The Pre-glacial Pleistocene of the Norfolk and Suffolk Coasts*, pp. 1–201. Cambridge: Cambridge University Press.

Late Pleistocene of Australia

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Introduction

Like other parts of the Southern Hemisphere, the history of research into late Quaternary beetle records in Australia is relatively short. This stems from a range of factors including the dominance of pollen analysis in exploration of Quaternary climates and environments, a longstanding lack of interest in alternative proxies, the difficulty of developing new proxies to modern standards and, perhaps especially, the relative lack of knowledge regarding the modern Australian beetle fauna (see below). This last factor applies to other regions of the Southern Hemisphere but is extreme in Australia because of the massive size of the fauna. Nonetheless, within this context significant advances have been made in terms of methods for the reconstruction of climatic parameters (Porch, 2010) and in the application of these methods to the Quaternary record (Porch, 2006; Porch et al., 2009; Sniderman et al., 2009). This article presents an overview of the current state of Quaternary beetle analysis in the Australian context, focusing on records from the late Quaternary.

The Australian Beetle Fauna

With a described beetle fauna of about 24 000 species (Lawrence and Britton, 1994) and an estimated fauna that may include 80–100 000 species, the Australian beetle fauna is large by world standards (Yeates et al., 2003). Indeed, it is likely that the Australian fauna contains at a minimum of 10 times, and potentially 20 times, as many species as Great Britain. In this context, it is sobering to note that there are almost as many *genera* of beetles in Australia as there are *species* in Great Britain and that some large Australian genus level radiations (such as the genus *Castiarina* Gory & Laporte (Buprestidae) and the genus *Heteronyx* Guérin-Ménéville (Scarabaeidae)) may individually contain the equivalent of around 10% of the British fauna. Special aspects of the Australian fauna include a number of large radiations associated with the dominant sclerophyll flora and many examples of taxa exhibiting 'southern connections.' These are taxa, usually at the level of genus or above, with relatives in parts of the former Gondwanaland (New Zealand, South America, New Caledonia and to a lesser extent, Southern Africa). These issues are reviewed for the Australian invertebrate fauna by Austin et al. (2004).

Given these difficulties in understanding the Australian fauna, it may seem that analysis of Quaternary assemblages would be intractable, but fortunately there are a number of factors that work in favor of a Quaternary entomologist. First, there are a number of important groups that are, in contrast to much of the fauna, well to very well known taxonomically and to a lesser extent ecologically. These coincidentally include

some of the key taxa utilized in Quaternary entomology in regions where it is well established (e.g., Scarabaeidae: Aphodiinae, Scarabaeinae; Dytiscidae; Hydrophilidae; Hydraenidae; and, some groups of Carabidae). Second, many groups that may not be well known systematically are well collected and represented in museum collections, reflecting the high level of interest in the nature and composition of the Australian fauna. Finally, there are methods available that can extract high-quality climatic data from the distributional data stored in collections and the literature. These methods are explored in the following section.

Paleoclimatic Reconstruction

Methodological Context

Bearing in mind the limitations noted above, it is obvious that an approach equivalent to that taken in regions of the Northern Hemisphere, where Quaternary beetle assemblages are typically examined, is untenable in the Australian context. The lack of specific ecological data for many taxa limits the detail of environmental reconstruction that may be attempted and the smaller amount of collection data (e.g., number of locality records) has meant that an approach like the mutual climatic range (MCR) method so commonly used in Europe and North America was unlikely to be of much use here. This is because the MCR method implicitly assumes that the climatic distribution of each taxon is well known, something reasonable in regions with a long and intensive history of taxonomic, ecological, and distributional research. In the Australian region, alternative methods have been developed that take into account how the climatic significance of an individual taxon is affected by the quality of its collection data. Marra et al. (2004) came to similar conclusions in New Zealand and developed the maximum likelihood envelope (MLE) method for estimating the likely climatic range for each taxon in fossil assemblages from that country.

The Coexistence Approach to Paleoclimatic Reconstruction

Details regarding the approach for quantification of past climatic parameter values can be found in Porch (2010). A summary of the approach and comment on potential limitations of the method are presented here. All recent methods of reconstructing Quaternary paleoclimates using fossil beetle assemblages utilize methods that rely on climatic coexistence, that is, the overlap of climatic ranges of the species in the assemblage (Kershaw and Nix, 1988; Mosbrugger and Utescher, 1997). This approach does not rely on the actual coexistence of species in geographic space but on their potential coexistence based on the climatic requirements of individual species (Figure 1) – it takes no account of other potential factors that

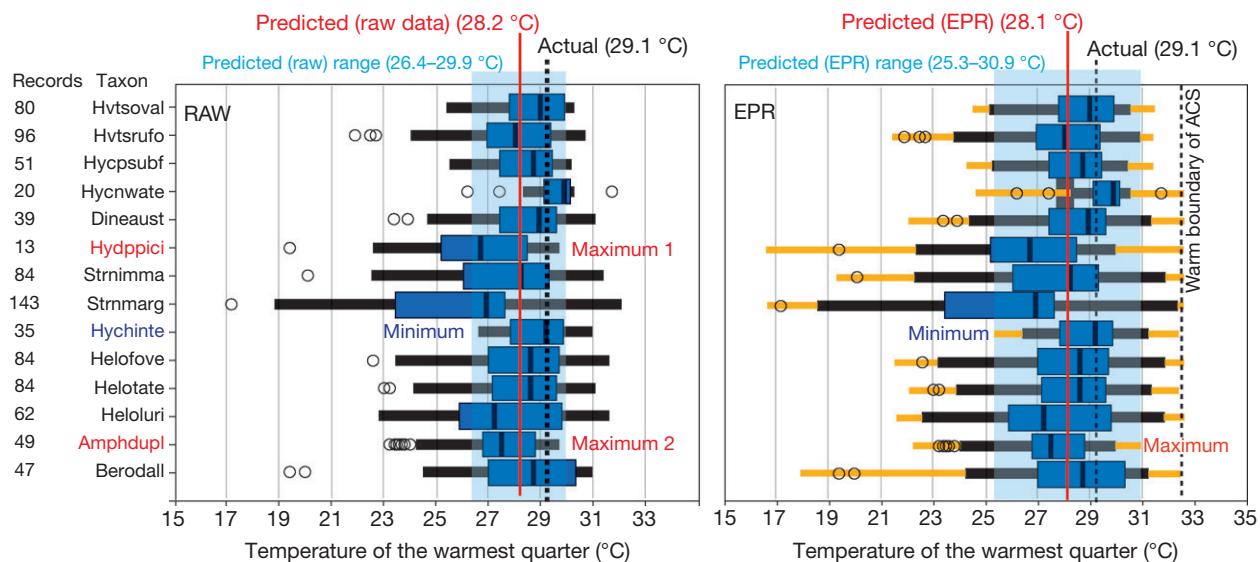


Figure 1 Reconstructing parameter values: an example using ‘temperature of the warmest quarter’ for a modern assemblage of 14 aquatic beetles from northern Australia (after [Porch 2010](#)). Species in the assemblage are listed (taxon) at the left side of the figure along with the number of geocoded localities (records) that contribute to their bioclimatic profile. Box-plots summarize the range of values for each taxon. The left figure (RAW) shows the raw values based exclusively on the range of values in the bioclimatic profile whilst the figure on the right (EPR) shows this range supplemented by the estimate parameter range ‘error’ (see [Porch 2010](#)). For both cases the envelope of overlap determines the range of potential values for the assemblage (26.4–29.9 °C for RAW and 25.3–30.9 °C for EPR) and the mid-point the predicted value for comparison with the observed value at the collection site (see [Figure 2](#)). The EPR figure illustrates that in northern Australian assemblages the edge of Australian climate space (ACS) is reached when the error values are included limiting the ability to predict climates warmer than the present value in ACS. For this reason coexistence methods always underestimate predicted values at the warm end of ACS and overestimate values at the cold end of ACS.

determine or limit a species’ range. The dominant approach in the Northern Hemisphere MCR method relies on climatic coexistence in two dimensional thermal space ([Atkinson et al., 1986](#)), but has also been applied infrequently to a climatic space that combines thermal and precipitation values ([Elias, 1997](#)). The MLE method developed in New Zealand reconstructs single parameters individually and has been used to reconstruct both temperature ([Marra et al., 2004, 2006](#)) and precipitation values ([Burge and Schulmeister, 2007a](#)). The approach adopted in Australia ([Figure 1](#)) has many similarities to the MLE method but with some significant differences, especially in relation to the estimation of parameter errors associated with relatively poorly known taxa and validation of the approach using jack-knifed parameters based on modern assemblages ([Porch, 2010](#)).

Thus far, bioclimatic profiles have been assembled for over 1000 species of Australian beetles using data from the literature, from museum databases and collections, and from the author’s own focused collection on groups that are common in the fossil record. These bioclimatic profiles are based upon three-dimensionally geocoded locality data and produced using the ANUCLIM 5.1 computer program ([Houlder et al., 2000](#)). BIOCLIM, a program within ANUCLIM, generates estimates of up to 35 climatic parameters (temperature, precipitation, radiation, and moisture index) for each locality from which a taxon is recorded. This *bioclimatic profile* forms the basic climatic summary for each taxon, mostly species, but in some cases species groups or even genera or families. Although it is theoretically possible to estimate the value for any climatic parameter, a study of these factors has

determined that only a relatively small proportion are actually likely to be ecologically informative (see [Porch, 2010](#), for discussion).

Because the amount of geocoded collection data varies between taxa, and few taxa are extremely well known, an approach for estimating the actual bioclimatic limits (parameter range) is necessary. The approach taken in Australia utilizes the relationship between the range of values for a parameter, the number of geocoded records for a taxon, and the effect of increasing data quality on range estimation (see [Porch, 2010](#)). The efficacy of the approach was then tested by using jack-knifed data from modern assemblages to ensure that the error assessments were sufficiently broad to enable accurate climatic reconstruction under modern circumstances, a critical test of the limitations of the method (e.g., [Figure 2](#)). Results of the jack-knifed modern tests indicated that critical climatic parameters including summer temperatures and precipitation could be accurately reconstructed. [Figure 3](#) illustrates the relationship between increasing numbers of taxa contributing to an assemblage and the precision of the reconstruction. Limitations in reconstructing past climates include the potential complexity of interpreting seasonal temperature regimes under climatic conditions that do not presently exist ([Jackson and Overpeck, 2000](#)): for example, some late Pleistocene intervals in Australia saw warm summers and enhanced thermal seasonality with winter climates colder than presently exist here. This is not an issue limited to Australia or to the BIOCLIM method, but rather one that exists anywhere such no modern-analog conditions may be encountered.

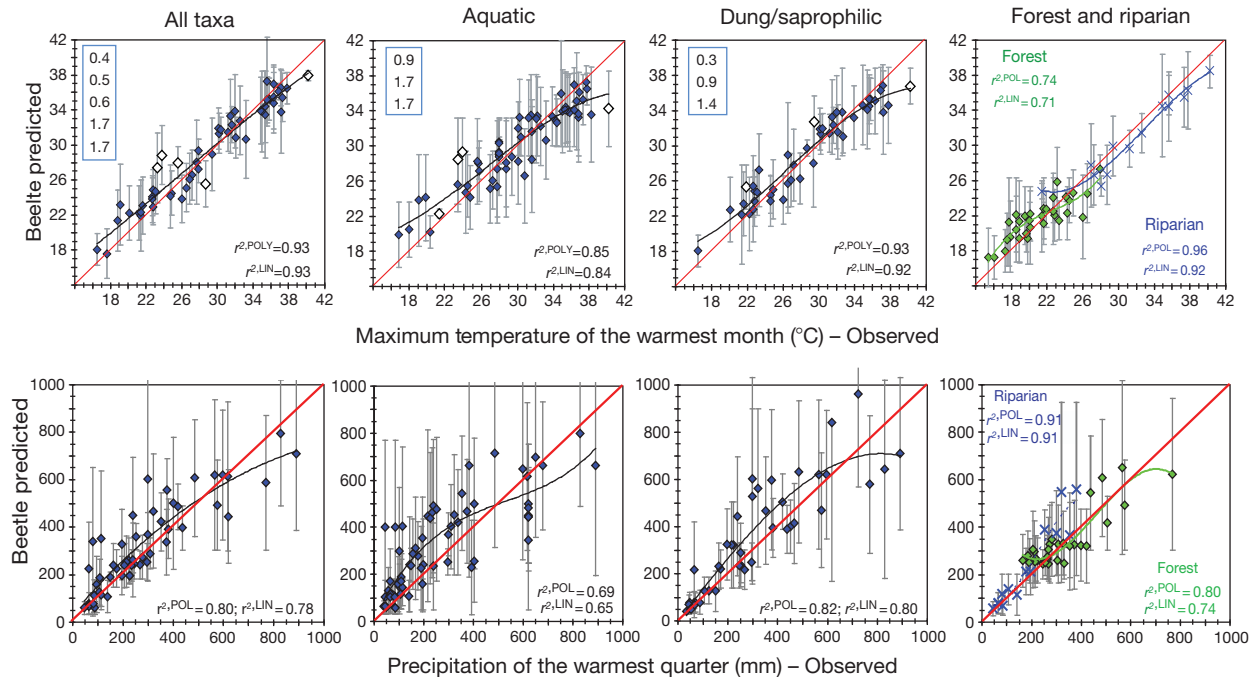


Figure 2 Summary of the observed (x-axis) versus beetle-predicted (y-axis) values for the parameters ‘temperature of the warmest quarter’ and ‘precipitation of the warmest quarter’ using four ‘ecological’ assemblage types (aquatic (63), dung/saprophage (46), forest (32) and riparian (14)) and the complete assemblage – all taxa (53); (figures refer to number of modern assemblages that contribute to the observed beetle-predicted assessment). Red line is the line of equilibrium. The best fitting third order polynomial is shown for each assemblage and values of the r^2 for both linear and these third order polynomial regressions are presented. Values in boxes are for represent the minimum difference between the predicted range and observed value for assemblages that fail to overlap the observed value. Note in particular the excellent r^2 values for the temperature parameter and the sigmoidal shape of the polynomial regression representing the over and underestimated noted in reference to Figure 1, and the fact that the observed value falls within predicted range for the precipitation estimates for all assemblages.

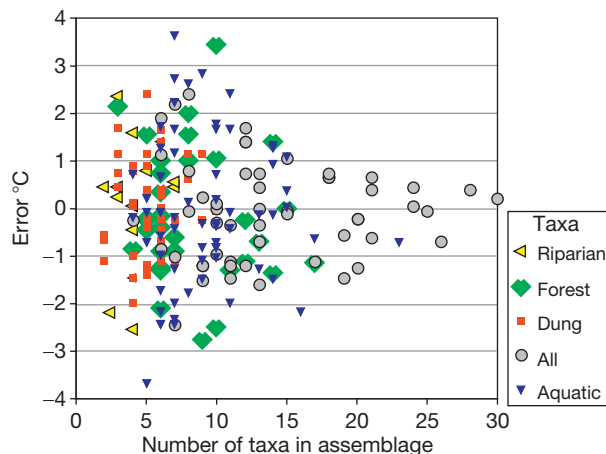


Figure 3 The relationship between the number of taxa in an assemblage (by ecological category as per Figure 2 – see caption) and the prediction error for the parameter temperature of the warmest quarter (difference between observed and predicted value). The accuracy of the prediction increases markedly with increasing numbers of taxa represented in the assemblage to the extent that with more than 20 taxa in an assemblage the predicted value is likely to be within about 1 °C of the observed value in the majority of cases (after Porch 2010).

A Longer Quaternary Context: The early Pleistocene Stony Creek Basin Site

The remarkable site of Stony Creek Basin in southeastern Australia (Figure 4) contains 40 m of organic clay that dates to the period between approximately 1.85 and 1.55 Ma (Sniderman et al., 2007). It is incredibly rich and generally well-preserved insect assemblages (e.g., Figure 5) are associated with diverse rainforest and wet sclerophyll forest, vegetation types that no longer occur in the region (Sniderman et al., 2007). The beetle record also includes a wide range of regionally extirpated taxa, with many restricted to either warmer wet forests of the east coast of Australia or taxa that are today found in the cool wet forests of southeastern Australia. Although the Stony Creek Basin assemblages have no modern analog, paleoclimatic reconstruction implies they were deposited under a warmer climatic regime with substantially enhanced summer rainfall conditions (Sniderman et al., 2009), demonstrating that Pliocene-like climates were maintained until at least the early Pleistocene. Enhanced summer rainfall would be necessary to allow the development of wet forest and rainforest on the site today, because the modern regional climate is too dry to support either of these vegetation types. To date, only a fraction of the Stony Creek Basin insect faunal record has

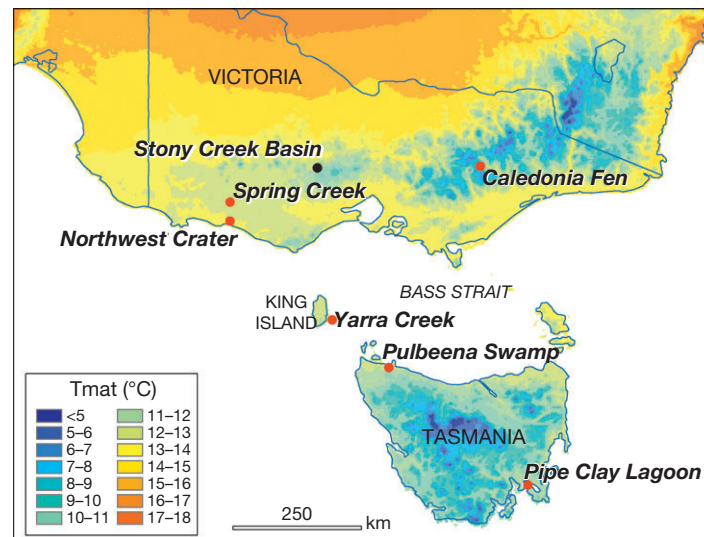


Figure 4 Sites mentioned in text. late Pleistocene beetle fossil sites indicated by red dots.

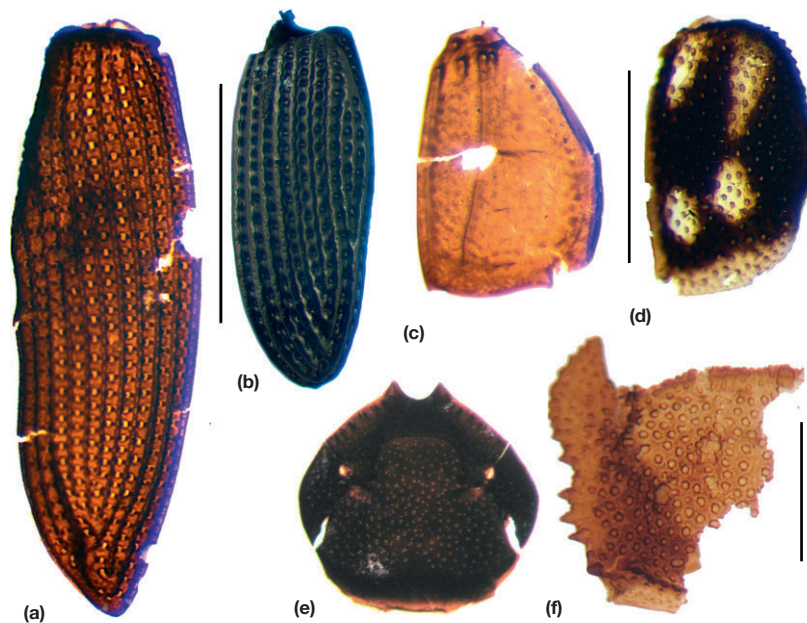


Figure 5 Typical beetle specimens from sclerophyll forest assemblages of the early Pleistocene site of Stony Creek Basin. (a) Left elytron of *Pycnomerus* cf. *fuliginosus* Erichson (Zopheridae); (b) right elytron of *Saprosites* Redtenbacher (Scarabaeidae: Aphodiinae); (c) right elytron of a pselaphine cf. *Anabaxis* Raffray (Staphylinidae: Pselaphinae); (d) right elytron of *Austrorhysus* Steel (Staphylinidae: Proteininae); (e) head of *Lepanus* Balthasar (Scarabaeidae); (f) pronotum fragment of *Pristoderus* cf. *pustulosus* (Blackburn) (Colydiidae). Scale bars: (a, b) 1 mm, (c, d) 0.5 mm, (e, f) 0.5 mm.

been examined and there remains much to be learned from the unique beetle fossil assemblages from this site. The presence of a range of potentially extinct species in genera that would be considered well-known (e.g., *Stenus* (Staphylinidae); *Lepanus* (Scarabaeinae)) hints that the demise of the warm wet climates of the early Pleistocene may have resulted in significant levels of extinction as rainforest and wet forest faunas contracted to their Quaternary refuges. Given there is substantial evidence of plant extinction in early to middle Quaternary records in southern Australia (Jordan, 1997), and in the Stony Creek

Basin record specifically (Sniderman et al., 2007), this may not be surprising.

Late Pleistocene Records

Yarra Creek, King Island, Tasmania

The Yarra Creek assemblage is derived from peat and peaty sand outcropping in coastal cliffs on the southeastern side of King Island in Bass Strait between Tasmania and the Australian

mainland (Figure 4). The assemblage is considered to represent the Last Interglacial interval, and was deposited by a fast-flowing stream in a back-dune swamp (Porch et al., 2009). A less likely alternative interpretation of the thermoluminescence-based chronology may suggest the assemblage is younger than marine isotope stage (MIS) 5e but that it is certainly from MIS 5. A number of pieces of evidence, however, including the presence of coastal plants in the fossil assemblage, imply that the MIS 5a interpretation is likely to be correct.

The Yarra Creek beetle assemblage is associated with rich pollen and plant macrofossil records that include several cool temperate rainforest species that no longer occur on King Island: some like the Myrtle Beech (*Nothofagus cunninghamii*) have been absent from King Island through the Holocene. Interpretation of the paleoclimatic significance of the plant assemblage implies it was deposited under conditions of enhanced summer rainfall, one of the principal limiting factors of *N. cunninghamii*. Interpretation of the paleoclimatic significance of the beetle assemblage similarly implies that summer rainfall was higher, although estimates of thermal regimes suggest they may have been similar to present for most parameters. The only significant departure from this pattern of similar thermal regimes is for measures of temperature seasonality. Porch et al. (2009) argued this departure was an artifact of the reconstruction methodology relating to the unique, highly equable climate of King Island. The lack of collection records for many of the plants and insects from highly equable climates necessitates a reconstruction of much more seasonal climates from the assemblage-based data. There are three possible explanations. First, the modern absence of plants like *N. cunninghamii* and some of the beetle taxa found in the fossil assemblages is related to the lack of suitable summer rainfall on King Island. Second, the absence of these taxa is related to

their extinction subsequent to the time of fossil deposition (e.g., during the Last Glacial Maximum (LGM)) and their lack of ability to recolonize. Or, third, and relating to the less well-known insects rather than the plants, some or many of the taxa found in the fossil assemblages may still be present on King Island today, but their presence remains undetected.

Biogeographically, there are a number of interesting patterns in the Yarra Creek insect assemblage. These include the presence of mainland species such as *Archaeoglenes australis* Lawrence (Tenebrionidae) and *Smicrophylax* voucher species AV1 (Trichoptera) (Figure 6) that are presently unknown from anywhere in Tasmania including the Bass Strait islands. In addition, mainland taxa such as *Mecyclothorax 'cordicollis'* (Carabidae) and *Prostomis cornuta* Waterhouse (Prostomidae) have modern ranges that reach as far as the Bass Strait islands but no further. Finally, there are Tasmanian species such as *Ceratognathus westwoodi* Thomson (Lucanidae), whose modern ranges do not reach the mainland or the Bass Strait islands.

Spring Creek, Southeastern Australia

The Spring Creek site in western Victoria yielded a rich insect assemblage dominated by beetles and caddisflies. It is associated with an assemblage of megafaunal mammal species including *Diprotodon*, *Zygomaturus*, and several large extinct kangaroos (Flannery and Gott, 1984). The site has a controversial history concerning its chronology and the interpretation of its depositional history (summarized in White and Flannery, 1995). The insect assemblage is derived from bedded organic fine sands that outcrop at the margins of the creek. Radiocarbon dating of both plant macrofossils and insect remains suggest the material was deposited prior to about 45 000 radiocarbon years ago, confirming White and Flannery's

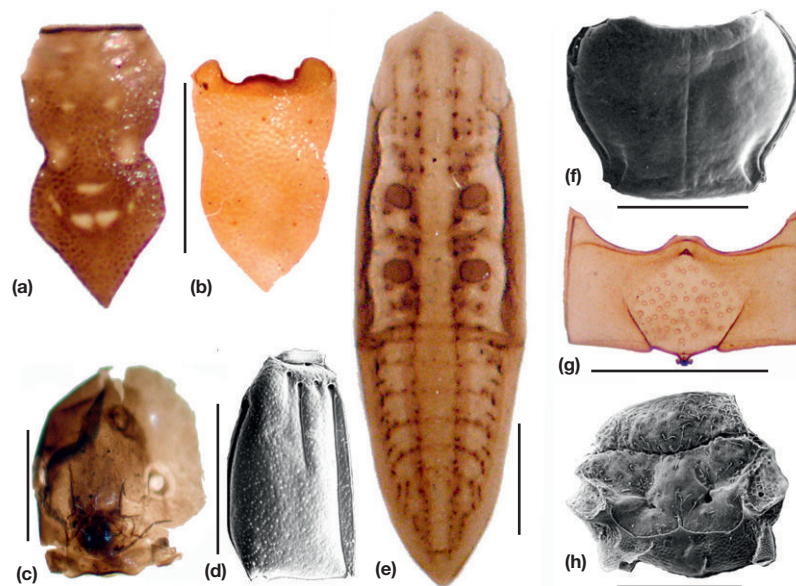


Figure 6 Insect specimens from late Pleistocene sites. (a) *Triplectides Kolenati* (Trichoptera) frontoclypeus, Pipe Clay Lagoon, Tasmania. (b) *Smicrophylax* AV1 (Trichoptera) frontoclypeus, Yarra Creek, King Island. (c) Ant (*Iridomyrmex*?), Northwest Crater, Victoria. (d) Unidentified Pselaphine genus (Staphylinidae), Pulbeena Swamp, Tasmania. (e) *Casuarina* scale insect, Northwest Crater, Victoria. (f) *Bembidion ateradustum* Liebherr (Carabidae), Pulbeena Swamp, Tasmania. (g) *Notocercyon Hansen metasternum* (Hydrophilidae: Sphaeridiinae), Northwest Crater, Victoria. (h) *Gymnochthebius australis* or *setosus* (Hydraenidae) head, Pulbeena Swamp, Tasmania. Scale bars: a,b,c 1 mm; d 0.4 mm; e,h 0.5 mm; f,g 0.8 mm.

(1995) view that the fauna dates to older than 35 000 years, based on contaminated accelerator mass spectrometry (AMS) ages on bone. Recent OSL dating of the sequence have yielded significantly younger ages (Fitzsimmons, personal communication, 2010). Three radiocarbon dates on plant macrofossils and two dates from insect remains resulted in ages consistent with an age older than the limit of the radiocarbon method, so it is likely the insect assemblage is older than 45 000 years (Porch and Kershaw 2010).

The Spring Creek assemblage is rich in aquatic insects, including a range of caddisfly taxa characteristic of warm lowland streams. Associated beetles include several species of the riffle beetle (Elmidae) *Austrolimnius* Carter & Zeck, a group restricted to well-oxygenated flowing water. Other water beetles include a range of generalist predaceous diving beetles (Dytiscidae) that are commonly found in lowland streams. The terrestrial assemblage is notable in the absence of taxa that are specifically or even frequently associated with trees. This absence parallels the absence of trees inferred from the pollen record that consistently has less than 5% tree pollen. Grass-feeding taxa such as the fungus weevil (Anthribidae) *Euciodus suturalis* Pascoe are present, but not in the numbers that occur in Holocene western Victorian basalt plains sites. The presence of the tiny metallic wood-boring beetle *Germarica* Blackburn (Buprestidae) implies the local presence of *Casuarina*/*Allocasuarina*. These genera are frequently trees, but the occurrence of abundant branch fragments belonging to a low growing *Allocasuarina* species suggests the beetle was feeding on these shrubs rather than trees of the same genus/genera.

The apparent lack of trees indicated by both the pollen and beetle records implies cold and dry conditions, at least based on the conventional interpretation of treeless vegetation in southern Australian pollen records (Kershaw et al., 1991). However, the composition of both the aquatic and terrestrial insect records completely contradicts this climatic interpretation. As noted above the caddisfly assemblage contains species such as *Archaeophylax canarus* Neboiss that are restricted to lowland habitats, especially to the east in the slightly more continental inland regions. Similarly, the terrestrial beetle assemblage is characteristic of low montane to lowland habitats in eastern Australia. Species such as the moisture-loving ground beetles *Pericompsus semistriatus* (Blackburn) and *Bembidion ateradustum* Liebherr (formerly *Bembidion blackburni* Csiki) imply conditions similar to those that occur on the site today. In terms of moisture regimes, there are several species that would require at least as much summer rainfall as today. The water scavenger beetle (Hydrophilidae: Sphaeridiinae) *Notoceryon* Hansen, although poorly known at the species level, is well collected and found in dung and moist leaf litter in forests throughout eastern Australia. Both the aphodiine dung beetles *Proctophanes sculptus* (Hope) and *Ataenius basiceps* Lea require relatively moist summer environments similar to those of *Notoceryon* Hansen. Together these and other species, and the evidence from the caddisfly assemblage, demonstrate that summer precipitation was near the present level. Thus, the lack of trees must be explained by some phenomenon other than lack of available moisture.

Pulbeena Swamp, Northwestern Tasmania

The record from Pulbeena Swamp (Colhoun et al., 1982), was the first to be examined for fossil insects in Australia. Material

from the site is unfortunately relatively poorly preserved, reflecting the combination of relatively slow sediment accumulation and the highly alkaline nature of the site sediments. Insect remains are present in several moderately humified and, in places, wood-rich peats that date to MIS 3. Younger peats that date to MIS 2 are well-humified and contain very poorly preserved insects or no insects at all. Assemblages from the MIS 3 peats include typical Tasmanian wet forest inhabitants such as the ground beetle *Promecoderus viridianeus* Sloane, abundant short-winged mold beetle remains (Staphylinidae: Pselaphinae), as well as rove beetles (Staphylinidae) in the Aleocharinae group and moisture-loving ground beetles such as *B. ateradustum* Liebherr, *Cyphotrechodes gibbipennis* (Blackburn) and a wide range of marsh beetles (Scirtidae). These assemblages occur with a range of wet forest land snail species; collectively they indicate wet conditions with temperatures that range from modern parameters to several degrees colder and at least as wet as present during the summer.

Caledonia Fen, Montane Southeastern Australia

Small numbers of beetles have been recovered from the site of Caledonia Fen at 1200 meters elevation in the Victorian alpine region (Figure 4). A continuous pollen record for the site has been published (Kershaw et al. 2007), which demonstrates that for all of MIS 3 and MIS 2 the site was surrounded by grass and herb dominated habitats. Intermittent deposits containing beetles were recovered from the 200–500 cm interval which dates between 50 and 25 ka radiocarbon years BP, based on AMS radiocarbon dates on twigs and seeds (Kershaw et al., 2007). The most common beetles in the record are the riffle beetle *Austrolimnius* Carter & Zeck and the minute moss beetle (Hydraenidae) *Tympanogaster* subgenus *Hygrotympnogaster* Perkins which occur from ~50–32 ka BP. These taxa do not occur in the immediate vicinity of the site today (a fen), but are restricted to gravel in flowing water and water-soaked rock faces, respectively. This implies that throughout the deposition of at least much of MIS 3, the site was fed by flowing streams rather than being isolated from the fluvial system of the catchment. This accords with the dominance of peat accumulation during the Holocene and last interglacial and clastic sedimentation between these periods.

Northwest Crater, Tower Hill, Southeastern Australia

Preliminary examination of LGM sections of Northwest Crater pollen cores has demonstrated the preservation of insects within the LGM part of the sequence. Samples examined (50 ml) contain very small assemblages consisting of both aquatic and terrestrial taxa. Aquatics include chironomids, caddisflies, predaceous diving beetles (*Necterosoma* Macleay, *Sternopriscus* Shapr), and water scavenger beetles (*Hydrochus* Leach, *Enochrus* Thomson, *Paracymus* Thomson). The few terrestrial taxa thus far found include rove beetles in the Quediinae group, the water scavenger beetle genus *Notoceryon* Hansen, minute brown scavenger beetles (Latridiidae) and ants. The presence of *Notoceryon* Hansen in a sample from 12.85 to 12.88 m in the sequence is significant in terms of moisture regimes. All modern records of the genus are from the moist forest habitats of eastern Australia: it is absent from

dry inland habitats that have suitable temperature regimes. Further work, using larger samples, is required to fully assess the nature of LGM climates in southeastern Australia.

Pipe Clay Lagoon, Southeastern Tasmania

The Pipe Clay Lagoon sequence outcrops in a wave-cut cliff at the edge of Pipe Clay Lagoon, a marine embayment in southern Tasmania. Colhoun (1977) described the pollen record from the site and published several radiocarbon dates that demonstrated the sequence was deposited in late MIS 3 to very early MIS 2. The sequence consists of two sandy peat beds varying from 5 to 15 cm in thickness, separated by about 5–8 cm of inorganic sand. The organic beds are set within a larger inorganic sand unit that Colhoun (1977) interpreted as wind-blown (aeolian). Examination of the two organic-rich sand beds at a single location revealed the presence of moderately well-preserved insects in the uppermost bed, and very little fossil material in the lower bed. A single AMS determination of seeds was obtained to test the original chronology of Colhoun (1977). The result ($21\,070 \pm 100$ (OZF799) on *Neopaxia* seeds) demonstrates that the upper peat bed clearly dates to the last millennium of MIS 3, about 26 000 cal BP (Porch, 2006).

The insect assemblage from the upper organic bed was relatively small and dominated by aquatic and riparian species, rather than terrestrial taxa. The species present suggest that the assemblage was deposited in relatively shallow water, perhaps with seasonally or locally variability, allowing the site to be occupied by hydrophilic taxa such as the ground beetles *Pseudoceneus* sp., *Bembidion aderustum* Lieberr and *C. gibbipennis* (Blackburn). The aquatic species in the assemblage are wide ranging taxa, except the predaceous diving beetle *Sternopriscus* spp. which may be *S. tasmanicus* Sharp, a Tasmanian endemic. Terrestrial species include specifically unidentified short-winged mold beetles and other rove beetles, the aphodiine dung beetles *A. basiceps* Lea and *Podotenus erosus* (Erichson), a Tasmanian endemic and the saproxylic (living in dead or decaying wood) cylindrical bark beetle *Pycnomerus secutus* (Pascoe).

Paleoclimatic interpretation of the assemblage implies warm summers with a range from several degrees cooler than present to the present value and increased continentality with increased seasonality, or cooler winters, at least in terms of maximum and minimum values. The presence of several taxa including *A. basiceps*, *P. erosus*, and *P. secutus* implies the precipitation estimates exceed present values for summer. *A. basiceps* is much more abundant on the mainland and has only been collected once in Tasmania, in the northeast part of the island, although the habitat of this species (wet grasslands/woodlands) has not been examined as well as wet forest habitats, so its distribution may be wider than currently recognized. On the other hand, both *P. erosus* and *P. secutus* are widespread in forest habitats today, and are commonly picked up using mass collecting techniques, meaning they are probably better known and thus provide more information regarding paleoclimate. It is clear, however, based on the nature of the beetle assemblage and the sedimentary context (the occurrence of organic sedimentation in a lowland setting), that the late MIS 3 climate was moist and cool rather than dry and cold.

Discussion

Late Quaternary Environments and Paleoclimates: Reconciling Plants and Beetles

A theme that emerges from the above review is that it is difficult to reconcile the pollen-based interpretation of treelessness as implying cold and dry conditions and the nature of insect assemblages from these sites. This aspect parallels somewhat the results of similar studies in other regions of the globe where at times there has been significant differences in the interpretation of paleoclimate based on pollen and beetle assemblages. In England, Coope and Angus (1975) noted this issue with reference to the Isleworth site where a treeless pollen assemblage was inferred to indicate tundra and cold conditions, whereas the beetle assemblage contained species indicating warm conditions. Coope (1977) argued that this reflected a lag in the vegetation response to warming conditions relative to the more vagile insects which were able to rapidly colonize suitable climatic habitat. The situation in Australia is quite different, partly as a result of the fact there was no continental-scale migration of tree species, a situation that parallels New Zealand where similar arguments about the nature of treelessness and distribution of forest during the last glacial are ongoing (Burge and Schulmeister, 2007b; McGlone et al., 2010). These arguments, however, tend to revolve around the extent of woody vegetation rather than the climatic significance of dominantly treeless vegetation.

The juxtaposition of pollen assemblages containing little evidence of local trees with insect assemblages that suggest warm wet conditions similar to today is difficult to reconcile, especially, given the traditional view of the ecological conditions associated with treeless vegetation. In Australia, as in New Zealand, there are several potential explanations. The most likely may relate to the impact of winter climates on trees. Under this scenario, during the last glacial interval trees were not limited by summer temperatures as they are presently, but rather by extreme winter temperatures resulting in lethal frosts that removed trees from all but the most protected landscapes (Harle et al., 2004; McGlone et al., 1993). The western plains of Victoria, for example, with negligible relief and no protection from western- or southerly derived air masses, would be particularly prone to the incursion of polar air masses, resulting in mass death of trees. This explanation reconciles the near total absence of trees in regions with a moderate and moist summer climate that would otherwise seem to be perfectly suitable for the establishment of woodland or forest. Alternatively, as suggested for New Zealand by Marra et al. (2007) and Burge and Schulmeister (2007b), trees may have been more widespread than has been detected in fossil pollen records. In the Australian context, there is little doubt that trees survived *in situ* in pockets of suitable habitat throughout the last glacial interval. Recent molecular evidence demonstrates the extent of *in situ* survival of key forest species in multiple refugia in mainland southeastern Australia and Tasmania (Nevill et al., 2010; Worth et al., 2009, 2010). These sites, incidentally, have long been recognized as key areas of forest refuge because they contain rich and endemic invertebrate taxa (species and some genera). The final explanation is that we may be overinterpreting the climatic significance of our beetle assemblages. Given that similar issues are cropping up in different parts of the globe and using

alternative methodologies, this seems like the least likely explanation. The resolution of this issue will depend on the development of further records, especially LGM records, in order to attempt to reconcile the maintenance of diversity in refugia throughout the late Quaternary, the pollen evidence of massive contraction of forest, and the geomorphic evidence of temperature reduction (Hesse et al., 2004; Williams et al., 2009).

Quaternary Contraction of Wet Forest Biota

It is clear from the southeastern Australian Quaternary beetle record that the range of wet forest biota became increasingly constrained through the Quaternary. The early Pleistocene Stony Creek Basin record demonstrates for both insects and flora that the range of both mesotherm rainforest/wet forest taxa and microtherm cool temperate rainforest/wet sclerophyll taxa (see Sniderman et al., 2009) has been substantially reduced since that time. Although the exact timing of this trend is uncertain, the major part of the contraction almost certainly coincided with the middle Pleistocene transition to drier conditions that is recorded throughout Australia. That this process was continuing in the late Quaternary is demonstrated by the last interglacial Yarra Creek assemblage. This assemblage contains a variety of plant and insect taxa that were living beyond their current range in areas that became too dry for them during the Holocene.

This history of contraction has highly significant implications for understanding the evolution and regional extinction of the Australian biota through the Quaternary. Scientists at the cutting edge of molecular research into the history of biota, whether attempting to reconstruct the relationships of extant taxa or looking at the phylogeography of species, are prone to forget Quaternary history. This history demonstrates that many populations are missing from areas where they once occurred, many species and even genera are regionally or even continentally extinct (better evidenced by the plants but certainly evident in the beetles), and that without Quaternary fossil records we could not begin to reconstruct these histories.

Summary

The development of methods for the reconstruction of past climates in Australia has allowed the exploration of a range of issues that have remained unresolved using traditional approaches like pollen analysis. Foremost is the quantification of past temperature and precipitation values for Quaternary sites ranging from the early Pleistocene to the last glacial, although sites from the latter period, especially the LGM, are still poorly characterized. These quantitative reconstructions provide evidence that in order to more completely understand Quaternary climates and environments, we need an approach that integrates and compares evidence from a range of disciplines, including Quaternary stratigraphy, geochronology, pollen analysis, plant macrofossil analysis, and insect analysis. Further beetle-based research and the development of chironomids as a tool of Quaternary paleoclimatology in Australia (Dimitriadis and Cranston, 2001; Rees and Cwynar, 2010) will enhance our understanding of the nature and limitations of

individual proxies and by so doing enable better understanding of the past climates of the Australian region.

See also: Pollen Records, Late Pleistocene: Australasia; Beetle Records: Late Pleistocene of New Zealand; Late Pleistocene of South America; Overview.

References

- Atkinson TC, Briffa KR, Coope GR, Joachim M, and Perry D (1986) Climatic calibration of coleopteran data. In: Berglund BE (ed.) *Handbook of Holocene Palaeoecology and Palaeohydrology*, pp. 851–858. New York: John Wiley & Sons.
- Austin AD, Yeates DK, Cassis G, et al. (2004) Insects 'Down Under' – Diversity, endemism and evolution of the Australian insect fauna: Examples from select orders. *Australian Journal of Entomology* 43: 216–234.
- Burge PI and Schulmeister J (2007a) An MIS 5a to MIS 4 (or early MIS 3) environmental and climatic reconstruction from the northwest South Island, New Zealand, using beetle fossils. *Journal of Quaternary Science* 22: 501–516.
- Burge PI and Schulmeister J (2007b) Re-envisioning the structure of last glacial vegetation in New Zealand using fossil beetles. *Quaternary Research* 68: 121–132.
- Colhoun EA (1977) A sequence of late Quaternary deposits at Pipe Clay Lagoon, southeastern Tasmania. *Papers and Proceedings of the Royal Society of Tasmania* 111: 1–12.
- Colhoun EA, van de Geer G, and Mook WG (1982) Stratigraphy, pollen analysis and paleoclimatic interpretation of Pulbeena Swamp, northwestern Tasmania. *Quaternary Research* 18: 108–126.
- Coope GR (1977) Fossil coleopteran assemblages as sensitive indicators of climatic changes during the Devensian (Last) cold stage. *Philosophical Transactions of the Royal Society of London, Series B* 280: 313–340.
- Coope GR and Angus RB (1975) An ecological study of a temperate interlude in the middle of the last glaciation, based on fossil Coleoptera from Isleworth, Middlesex. *Journal of Animal Ecology* 44: 365–391.
- Dimitriadis S and Cranston PS (2001) An Australian Holocene climate reconstruction using Chironomidae from a tropical volcanic maar lake. *Palaeogeography, Palaeoclimatology, Palaeoecology* 176: 109–131.
- Elias SA (1997) The mutual climatic range method of palaeoclimate reconstruction based on insect fossils: New applications and interhemispheric comparisons. *Quaternary Science Reviews* 16: 1217–1225.
- Flannery TF and Gott B (1984) The Spring Creek locality, southwestern Victoria, a late surviving megafaunal assemblage. *Australian Zoologist* 22: 385–422.
- Harle KJ, Kershaw AP, and Clayton E (2004) Patterns of vegetation change in southwest Victoria (Australia) over the last two glacial/interglacial cycles: Evidence from Lake Wangoom. *Proceedings of the Royal Society of Victoria* 116: 105–137.
- Hesse PP, Magee JW, and van der Kaars S (2004) Late Quaternary climates of the Australian arid zone: A review. *Quaternary International* 118–119: 87–102.
- Houlder DJ, Hutchinson MF, Nix HA, and McMahon JP (2000) *Anuclim v. 5.1. Centre for Resource and Environmental Studies*. Canberra: Australian National University.
- Jackson ST and Overpeck JT (2000) Responses of plant populations and communities to environmental changes of the late Quaternary. *Paleobiology* 26: 194–220.
- Jordan GJ (1997) Evidence of Pleistocene plant extinction and diversity from Regatta Point, western Tasmania, Australia. *Botanical Journal of the Linnean Society* 123: 45–71.
- Kershaw AP, D'Costa DM, McEwen Mason JRC, and Wagstaff BE (1991) Palynological evidence for Quaternary vegetation and environments of mainland southeastern Australia. *Quaternary Science Reviews* 10: 391–404.
- Kershaw AP, McKenzie GM, Porch N, et al. (2007) A high resolution record of vegetation and climate through the last glacial cycle from Caledonia Fen, southeastern highlands of Australia. *Journal of Quaternary Science* 22: 481–500.
- Kershaw AP and Nix HA (1988) Quantitative palaeoclimatic estimates from pollen data using bioclimatic profiles of extant taxa. *Journal of Biogeography* 15: 589–602.
- Lawrence JF and Britton EB (1994) *Australian Beetles*. Melbourne: Melbourne University Press.
- Marra MJ, Crozier M, and Goff M (2007) Palaeoenvironment and biogeography of a late MIS 3 fossil beetle fauna from South Taranaki, New Zealand. *Journal of Quaternary Science* 24: 97–107.

- Marra MJ, Shulmeister J, and Smith EGC (2006) Reconstructing temperature during the last glacial maximum from Lyndon Stream, South Island, New Zealand using beetle fossils and maximum likelihood envelopes. *Quaternary Science Reviews* 25: 1841–1849.
- Marra MJ, Smith EGC, Shulmeister J, and Leschen R (2004) Late Quaternary climate change in the Awatere Valley, South Island, New Zealand using a sine model with a maximum likelihood envelope on fossil beetle data. *Quaternary Science Reviews* 23: 1637–1650.
- McGlone MS, Newnham RM, and Moar NT (2010) The vegetation cover of New Zealand during the last glacial maximum: Do pollen records under-represent woody vegetation? In: Haberle S, Stevenson J, and Prebble M (eds.) *Altered ecologies: Fire, Climate and Human Influence on Terrestrial Landscapes. Terra Australis* 32, pp. 49–68. Canberra: ANU E-Press.
- McGlone MS, Salinger MJ, and Moar NT (1993) Palaeovegetation studies of New Zealand's climate since the last glacial maximum. In: Wright HE, Kutzbach JE, Webb T III, Ruddiman WF, Street-Perrott FA, and Bartlein PJ (eds.) *Global Climates since the Last Glacial Maximum*, pp. 294–317. Minneapolis: University of Minnesota Press.
- Mosbrugger V and Utescher T (1997) The coexistence approach – a method for quantitative reconstructions of Tertiary terrestrial palaeoclimate data using plant fossils. *Palaeogeography, Palaeoclimatology, Palaeoecology* 134: 61–86.
- Nevill PG, Bossinger G, and Ades PK (2010) Phylogeography of the world's tallest angiosperm, *Eucalyptus regnans*: Evidence for multiple isolated Quaternary refugia. *Journal of Biogeography* 37: 179–192.
- Porch N (2006) *A Method for Reconstructing the Quaternary Climates of Australia Using Fossil Beetles*. Melbourne: School of Geography and Environmental Science, Monash University.
- Porch N (2010) Climate space, bioclimatic envelopes and coexistence methods for the reconstruction of past climates: A method using Australian beetles and significance for Quaternary reconstruction. *Quaternary Science Reviews* 29: 633–647.
- Porch N, Jordan GJ, Price DM, Barnes RW, Macphail MK, and Pemberton M (2009) Last interglacial climates of south-eastern Australia: Plant and beetle-based reconstructions from Yarra Creek, King Island, Tasmania. *Quaternary Science Reviews* 28: 3197–3210.
- Porch N and Kershaw AP (2010) Comparative AMS 14C dating of plant macrofossils, beetles and pollen preparations from two Late Pleistocene sites in southeastern Australia. In: Haberle S, Stevenson J, and Prebble M (eds.) *Altered Ecologies: Fire, Climate and Human Influence on Terrestrial Landscapes. Terra Australis* 32, pp. 395–403. Canberra: ANU E-Press.
- Rees ABH and Cwynar LC (2010) Evidence for early postglacial warming in Mount Field National Park. *Tasmania Quaternary Science Reviews* 29: 443–454.
- Sniderman JMK, Pillans B, O'Sullivan PB, and Kershaw AP (2007) Climate and vegetation in southeastern Australia respond to Southern Hemisphere insolation forcing in the late Pliocene–early Pleistocene. *Geology* 35: 41–44.
- Sniderman JMK, Porch N, and Kershaw AP (2009) Quantitative reconstruction of Early Pleistocene climate in southeastern Australia and implications for atmospheric circulation. *Quaternary Science Reviews* 28: 3185–3196.
- White JP and Flannery T (1995) Late Pleistocene fauna at Spring Creek, Victoria: A re-evaluation. *Australian Archaeology* 40: 13–17.
- Williams M, Cook E, van der Kaars S, Barrows T, Shulmeister J, and Kershaw P (2009) Glacial and deglacial climatic patterns in Australia and surrounding regions from 35 000 to 10 000 years ago reconstructed from terrestrial and near-shore proxy data. *Quaternary Science Reviews* 28: 2398–2419.
- Worth JRP, Jordan GJ, Marthick GE, and Vaillancourt RE (2010) Chloroplast evidence for geographic stasis of the Australian bird-dispersed shrub *Tasmannia lanceolata* (Winteraceae). *Molecular Ecology* 19: 2949–2963.
- Worth JRP, Jordan GJ, McKinnon GE, and Vaillancourt RE (2009) The major Australian cool temperate rainforest tree *Nothofagus cunninghamii* withstood Pleistocene glacial aridity within multiple regions: Evidence from the chloroplast. *New Phytologist* 182: 519–532.
- Yeates DK, Harvey MS, and Austin AD (2003) New estimates for terrestrial arthropod species-richness in Australia. *Records of the Australian Museum Monograph Series* 7: 231–241.

Late Pleistocene of Europe

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Introduction

In this article, the response of Coleoptera (beetles) to Upper Pleistocene climatic changes are summarized. This should cast light upon our understanding of the nature of these changes and thus enable us to predict the ways in which other animal and plant species might respond to any future climatic events. By using the changes in the range of beetle species during the Quaternary, it has been possible to quantify thermal climatic conditions and to estimate both the amounts and the rates of change involved. Furthermore, it has been possible to map the regional differences in thermal climates within northwestern Europe.

The beetle assemblages described here include those from the Last (Eemian, Ipswichian) Interglacial, the Last (Weichselian, Devensian) Glaciation, and the transition to the present (Holocene) Interglacial. Most of the examples have been drawn from sites in the British Isles and northern Europe because it is there that these faunas have been most intensively investigated. The beetle record shows that the climate oscillated rapidly between temperate–oceanic and arctic–continental conditions. Where possible, the fossil insect assemblages have been correlated with marine isotope stages (MIS) or with the event stratigraphy of the Greenland ice cores. In the account that follows, paleotemperature estimates are given using the mutual climatic range (MCR) method (Atkinson et al., 1987) where $T_{\max}$ indicates the mean temperature of the warmest month (July) and $T_{\min}$ indicates the mean temperature of the coldest months (January and February; [Figure 1](#)).

Upper Pleistocene Beetle Assemblages

Eemian Interglacial (MIS 5e)

Extensive beetle assemblages of this age are known from seven localities in southeast England ([Coope, 2000a](#)). These faunas include a number of southern European species, often occurring in abundance ([Coope, 1990](#)). Quantitative estimates of the thermal climate of these sites have given the following figures: $T_{\max}$ 18–24 °C, $T_{\min}$ –6 to 6 °C. It is likely that the actual mean July temperatures may have been about 21 °C and that mean January/February temperatures may have been about 4 °C. Thus, the mean temperature of the warmest month was of the order of 3 °C warmer than it is in southern England today. Winter figures are more uncertain, but they may not have been much different from those of the present day. Precipitation at that time was sufficient to maintain a vigorous flow in the rivers throughout the year.

The termination of the Eemian Interglacial is difficult to investigate because this stage is apparently missing from all of

the British sequences. However, a borehole at Elsing in the valley of the River Wensum, about 22 km west of Norwich, penetrated Eemian deposits immediately overlain by sediments with arthropod faunas indicative of rapidly alternating climates ranging from temperate to arctic conditions ([Taylor and Coope, 1985](#)).

At the Oerel site, in northwest Germany ([Behre et al., 2005](#)), beetle assemblages were recovered from sediments representing the late part of the Eemian interglacial. They were divided into two different faunal units, in which the older is rather species-rich and indicative of temperate conditions, compared with the younger sparser fauna, which lacked thermophiles. MCR estimates based on the older assemblages are: $T_{\max}$ 14–24 °C, $T_{\min}$ –11 to 6 °C.

Further southwest, at the La Grande Pile site, Vosges, France, beetle remains were analyzed from ten samples assigned to Eemian deposits ([Ponel, 1995](#)). The beetle assemblages were divided into an early and a late faunal unit. The early unit is characterized by an abundance of beetles dependent on deciduous trees, and thermophilous predators and scavengers. MCR estimates based on this assemblage are: $T_{\max}$ 16–22 °C, $T_{\min}$ 0–10 °C. The faunal unit corresponding to the later part of the interglacial includes species feeding on both deciduous and coniferous trees. A number of species indicate a mild and humid climate: $T_{\max}$ 17–26 °C, $T_{\min}$ –6 to 12 °C.

Herning Stadial (MIS 5d)

The insect assemblages from the Oerel and Gross Todtshorn sites in Germany ([Walking, 1997](#)), attributed to this period, are made up of only a few taxa. At Oerel, a few species confined to a marshy environment were identified. At Gross Todtshorn typical heath taxa were recovered. However, the faunas were too small to permit MCR reconstructions to be made. At La Grande Pile one sample included a number of species living on open ground, such as open grassland. However, the MCR estimates do not indicate arctic conditions during this time: $T_{\max}$ 15–19 °C, $T_{\min}$ –15 to 9 °C.

Brørup Interstadial (MIS 5c)

Beetle faunas were described from four sites in central and northern Sweden that were originally thought to be interglacial in age ([Lindroth, 1948](#)) but which are now attributed to an early Weichselian interstadial. These faunas included several exclusively high northern and Asiatic species. More recently, Coleoptera and other insect remains were found and identified from deposits of Brørup age (Peräpohjola) at ten different sites in northernmost Sweden ([Lemdahl, 1997](#)). All the recorded taxa give a very consistent picture of a xeric to mesic tundra environment. A number of the identified species are not found within Fennoscandia today, but in northern Russia and

[†]Deceased.

Eemian and Weichselian sites



Figure 1 Map showing Eemian to full-glacial Weichselian sites where fossil beetle assemblages were studied. (1) Early Weichselian sites in northern Sweden, (2) Dimlington, (3) Chelford, (4) Elsing near Norwich, (5) Latton, (6) Cassington near Oxford, (7) South Kensington, (8) Isleworth, (9) Peelo, (10) Oerel, (11) Gross Todtshorn, (12) Belchatów, (13) La Grande Pile, (14) Gossau, (15) full-glacial site near Verona.

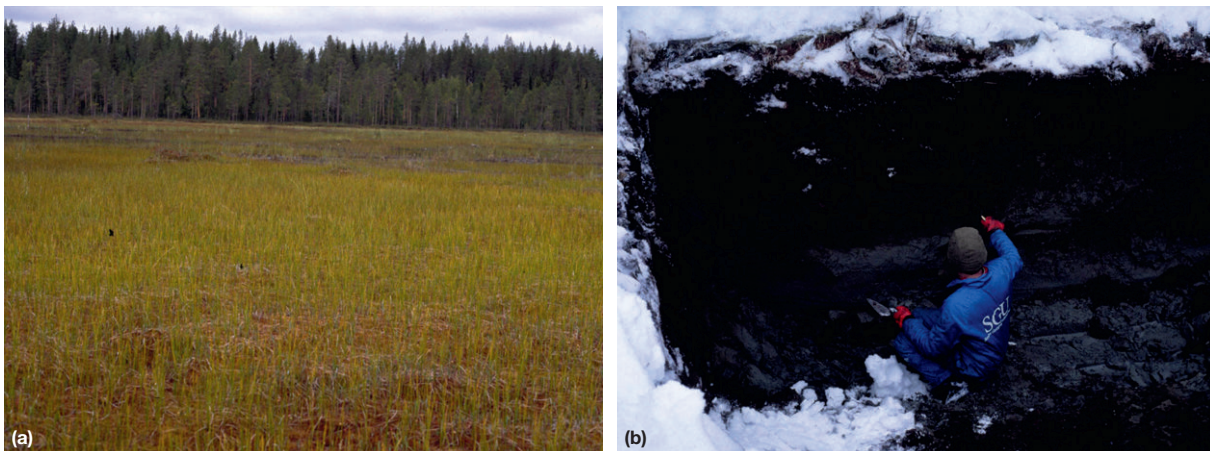


Figure 2 (a) The Veiki moraine plateau at Outöjärvi, north easternmost Sweden. Here organic silty sediments dating from the Early Weichselian Brörup interstadial (MIS 5c) were found between 1.5 and 5.5 m below the present soil surface. The sediments are covered by 1.5 m Holocene peat. (b) Sampling of the Early Weichselian sediments for plant macrofossil and insect analyses at Outöjärvi in an open section made by a dredger, during winter time when the upper meter of the ground was frozen.

Mongolia. MCR estimates indicate cold, harsh conditions: $T_{\max}$ 9–11 °C, $T_{\min}$ –20 to –9 °C (**Figure 2(a)** and **2(b)**).

Beetle faunas of this age indicate mixed conifer and birch woodland and are known from several sites in the English Midlands. The most extensively studied is from Chelford, Cheshire (Coope, 1959). MCR analysis of this fauna gives the following figures: $T_{\max}$ 15–18 °C, $T_{\min}$ –11 to 1 °C. It is likely

that the actual figure for $T_{\max}$ was closer to 15°. Thus, the summer temperatures were similar to those in central England today, but the winters were substantially colder.

Results from the Oerel site present two contrasting faunas from this interstadial. The earliest part, of low diversity, included northern species. This was followed by a striking increase in number of species, with terrestrial and aquatic

beetles that today are found only in temperate regions where they inhabit fens and coniferous woodlands. Similar patterns are found in the results from Gross Todtshorn. MCR estimates from Oerel are as follows: $T_{\max}$ 12–19 °C, $T_{\min}$ –21 to 6 °C. At La Grande Pile the assemblage from this time is characterized by a rich fauna of tree-dependent species living on both deciduous and coniferous trees. Climate conditions were likely to have been less temperate than during the Eemian: $T_{\max}$ 13–24 °C, $T_{\min}$ –13 to 10 °C.

Rederstal Stadial (MIS 5b) and Odderade Interstadial (MIS 5a)

Insect assemblages dating to the Odderade Interstadial (Tärendö) were recovered from four Swedish sites (Lem Dahl, 1997). They are characterized by arctic tundra faunas of low diversity. The climatic conditions seem to have been even more severe than those of the Brørup: $T_{\max}$ 10–11 °C, $T_{\min}$ –20 to –17 °C. There are no beetle faunas from Britain that can be unequivocally attributed to MIS 5b and 5a.

The Rederstal assemblages from Oerel and Gross Todtshorn are made up of only a few beetle taxa and indicate the presence of shrub tundra. $T_{\max}$ estimates from Gross Todtshorn are around 11 °C. The Odderade Interstadial is well represented at Oerel where two different insect assemblages reflect two different climate regimes during the period. The first part of the interstadial is characterized by a thermophilous marsh fauna ($T_{\max}$ 13–25 °C, $T_{\min}$ –8 to 10 °C). During the latter part, there is a rather rapid change in which cold stenotherm species replace the temperate fauna ($T_{\max}$ 13–14 °C, $T_{\min}$ –14 to –10 °C). At Gröbern in eastern Germany, the beetle succession showed large scale post-Eemian climatic oscillations that probably belong to this period. These involved changes of at least 7 °C in mean July temperatures (Walking and Coope, 1996). Beetle assemblages dating to the Rederstal from La Grande Pile show a fall in the number of tree-dependent taxa and a cold-adapted species was also recorded ($T_{\max}$ 13–20 °C, $T_{\min}$ –16 to 8 °C). During the Odderade, the number of tree-dependent beetles increased again. The climate was probably slightly cooler than during the Brørup ($T_{\max}$ 12–14 °C, $T_{\min}$ 0–6 °C).

Middle Weichselian Stadials and Interstadials (MIS 4)

This period was climatically complex. There are at least two separate middle Weichselian temperate interstadials in central and southern England separated from one another by a severely cold interval. Both of these temperate episodes postdate MIS 5 (Coope, 2000b) and yielded pollen indicating a treeless landscape. Beetle assemblages are known from three sites in England that date from the earlier of these two interstadials, namely that from Islesworth (Coope and Angus, 1975), the post interglacial deposits at Cassington (Maddy et al., 1998) and from Latton (Lewis et al., 2006). The interstadial beetle assemblages from these sites are rich and diverse and are made up wholly of temperate–oceanic species. Average MCR analyses based on the beetle faunas from these three sites gave the following estimates: $T_{\max}$ 17–18°, $T_{\min}$ –4 to 4 °C.

At Cassington, these temperate interstadial deposits were overlain by gravels in the middle of which organic silts filled a small paleochannel with a fauna made up of an unusual group

of exclusively arctic and continental beetle species. MCR analysis of the beetles from this deposit yielded the following estimates: $T_{\max}$ 7–11 °C, $T_{\min}$ –30 to –10 °C. An almost identical beetle assemblage was also found at South Kensington, London, suggesting that these two deposits are of the same age (Coope et al., 1997). At South Kensington the level containing this arctic fauna was overlain almost immediately by organic silty gravels which yielded a rich interstadial fauna composed entirely of temperate–oceanic beetle species. MCR climatic calculations of this upper temperate interlude gave the following estimates: $T_{\max}$ 16–18 °C, $T_{\min}$ –6 to –1 °C. This temperate period follows so closely on the arctic bed, both of them occurring within the same small channel deposit within the gravel sequence, that it seems likely that the climatic warming took place very rapidly. However, in the absence of reliable dates, the actual rate of climatic change cannot be determined.

It is environmentally significant that, during both these temperate interludes, pollen analysis shows that the landscape in Britain was totally devoid of trees in spite of the climate being warm enough at these times to support mixed oak forest.

Middle Weichselian (MIS 3)

Numerous organic deposits are known from England which have yielded beetle assemblages that can be attributed to the Middle Weichselian on stratigraphical grounds (Coope, 1987a,b), but it is not yet possible to allocate them precisely to their equivalent positions in the Greenland ice-cores chronology.

Several sites that have been radiocarbon dated to about 42.0 ky BP have yielded beetle assemblages that include relatively southern species indicating warmer climates (Coope et al., 1961; Coope 1968, 2000b, 2002). Because of the unreliability of radiocarbon dates of this age, it is not possible at the moment to be sure if these isolated deposits represent just one warm episode; or several of them (i.e., the Upton Warren Interstadial Complex of Coope and Sands (1966)).

A large number of beetle faunas from deposits with radiocarbon dates between 40.0 ky BP and 20.0 ky BP have remarkably consistent assemblages and include numerous species with exclusively arctic and/or Asiatic distributions today (Coope, 1995). Many of these exotic species are represented by large numbers of individuals. MCR estimates have been made on over 20 of these faunas (Coope, 1987b) and an approximate average may be summarized as follows: $T_{\max}$ 9–11 °C, $T_{\min}$ –25 to –10 °C. It is possible that these faunas represent several cold interstadials interspersed with even colder intervals as the Greenland Ice cores suggest (Johnsen et al., 1992) in which case biological activity in Britain may have fallen to a paleontologically invisible minimum.

Although Scotland was totally overridden by ice sheets at the Last Glacial Maximum (LGM), insect-bearing sediments are occasionally found underneath the veneer of glacial deposits. Notable among these is the prolific site at Sourlie, near Irvine, western Scotland, which yielded a rich assortment of plant and animal fossils, including reindeer and woolly rhinoceros as well as abundant insect remains. Radiocarbon dates on plant remains and the collagen from a reindeer antler found *in situ* grouped around 30.0 ky BP (Bos et al., 2004). MCR paleoclimatic estimates based on the beetle fauna gave mean July temperatures of between 6 and 9 °C and mean January/

February temperatures of -34°C and -11°C . A more recent study of a glaciectonized organic deposit at Balglass Burn, high on the slopes of the Campsie Fells east of Glasgow, gave ages between 34.5 and 28.0 ky BP based on six radiocarbon dates on sedge fruits and beetle remains (Brown et al., 2007). Besides insects, pollen and plant macrofossil remains were also studied. All these indicate a small pond situated in an open tundra landscape. MCR estimates based on the beetle assemblage give T_{max} $8-10^{\circ}\text{C}$ and T_{min} -26 to -10°C . The climatic implications of these beetle faunas demonstrate that much of central Scotland was not covered by glaciers during the middle pleniglacial at a time when thermal conditions were apparently sufficiently cold to permit ice accumulation in the highlands. Some other factor such as inadequate precipitation must also have been involved in maintaining ice-free conditions in northern Britain at this time.

In the Netherlands pleniglacial, two extensive Middle Weichselian insect assemblages are known. Thus, a beetle fauna from Peelo included many of the exclusively arctic and/or Asiatic species that were also present in the British Isles at the same time (Coope, 1969). Excavations at Orvelte, near Drenthe, to recover mammoth and woolly rhinoceros remains, enabled a large beetle fauna to be investigated that included an assemblage of cold-adapted species similar to the ones discussed above (Cappers et al., 1993). Both of these faunas yield MCR T_{max} temperature estimates at or below 10°C and T_{min} values below -20°C .

A Middle Weichselian insect fauna from Poland included a very similar group of high arctic and Asiatic species (Kasse et al., 1998). MCR estimations based on this assemblage gave the following estimates: T_{max} $8-12^{\circ}\text{C}$, T_{min} -27 to -20°C .

Insect remains were found in deposits referring to the Oerel interstadial both at the Oerel site and at Gross Todtshorn. The beetle assemblages are composed of northern Palaearctic species, which indicate a treeless environment and cold climate (T_{max} $12-13^{\circ}\text{C}$, T_{min} -9 to -2°C). At the Oerel site, insect remains were also recovered from a peat layer referred to the younger Glinde Interstadial. The beetle assemblage from this horizon is similar in composition to those from the earlier interstadial. Climate conditions were also likely to have been very similar (T_{max} $9-14^{\circ}$, T_{min} -17 to -2°C).

In Switzerland, two insect assemblages of middle Weichselian age have been recorded from Gossau (Jost-Stauffer et al., 2005). They include several exclusively arctic and Asiatic species that do not occur anywhere in central Europe at the present day. MCR estimates gave the following figures: T_{max} $8-13^{\circ}\text{C}$, T_{min} -21 to -7°C . A second beetle assemblage from the Swiss plateau was obtained from samples taken in a peaty matrix of the mammoth skeleton at Niederweningen that was radiocarbon dated to around 45.0 ky BP (Coope, 2007). This fauna also includes many exclusively arctic/Asiatic species. The mammoths were probably mired in a reedy swamp in a landscape with scattered willows, birches, and conifers. An MCR estimate indicates T_{max} around 10°C and T_{min} below -10°C .

Middle Weichselian beetle assemblages from La Grande Pile are characterized by the presence of almost entirely arctic and continental species. Obligate tree-dependent species were not recorded. The climate was cold in general through the period (T_{max} around 12°C , T_{min} approximately -10°C or below).

A site of outstanding archaeological importance has recently been recorded at Lynford, Norfolk, where an extensive

beetle fauna was found in close association with abundant evidence of Neanderthal exploitation of mammoths: smashed bones and numerous flint tools (Boismier et al., 2011). There can be little doubt that the beetle assemblage is contemporary with the human activities and provides an intimate view of the local environment (Coope, 2011). However, the beetle assemblage was unusual. It included abundant high northern and Asiatic species suggesting cold continental climatic conditions. But this fauna lacks several of the extreme cold-adapted species present in the typical full-glacial assemblages and includes rare species whose distributions do not extend into arctic regions today. These make MCR estimates difficult. Nevertheless, it is likely that the mean July temperatures were between 12 and 14°C and January/February figures at or below -10°C . The age of this important site is probably sometime late in MIS 4 or early MIS 3.

Last Glacial Maximum (MIS 2)

Conditions during the LGM were extremely harsh and insect faunas from this period are consequently very rare. A small beetle assemblage obtained from Dimlington, East Yorkshire, has been radiocarbon dated to 18.5 ky BP (Penny et al., 1969). Given the smallness of this insect fauna, only a provisional MCR estimate is possible: T_{max} $9-11^{\circ}\text{C}$, T_{min} -27 to -7°C . These figures do not necessarily represent the typical climate at the LGM for England because the insects came from a single fossiliferous horizon in an otherwise barren sequence of laminated silts and may thus indicate a short period of atypically warm conditions in an otherwise even colder episode.

In continental Europe, two beetle assemblages are known from the LGM. Beetle assemblages from La Grande Pile dating to this period indicate very cold and harsh conditions, with T_{max} mostly around 10°C and T_{min} well below -10°C . Fossil remains of beetles and oribatid mites, which are dated to around 18 800 BP, were recovered from a site close to Verona, northeastern Italy (Foddai and Minelli, 1994). They indicate climate conditions colder and wetter than today. T_{max} may have been $8-9^{\circ}\text{C}$ lower than present day.

Late Glacial Period

This period falls well within the radiocarbon dating techniques. Dates are cited here in radiocarbon years BP and calibrated calendar years are shown in parentheses as cal years BP. Correlations are made with the climatic event stratigraphy of the Greenland Ice cores (Walker et al., 1999). As a large number of studies have been carried out on beetle assemblages from sites scattered all over Europe (cf. Coope et al., 1998), only a representative selection is discussed here (Figure 3).

Late Glacial stadial (Greenland stadial 2)

Faunas that date from the cold period (Greenland stadial 2, GS-2) that immediately preceded the Late Glacial Interstadial, namely from just before 13.0 ky BP (ca. 15.4 cal ky BP), are rare. In Britain, the most stratigraphically secure faunal assemblage of this age is from the North Wales coast at Glanllynau, where a sequence of insect-bearing deposits spans almost the whole of the early part of the Late Glacial Interstadial (Coope and Brophy, 1972, cited in Coope et al., 1998). The stadial

Late Glacial sites

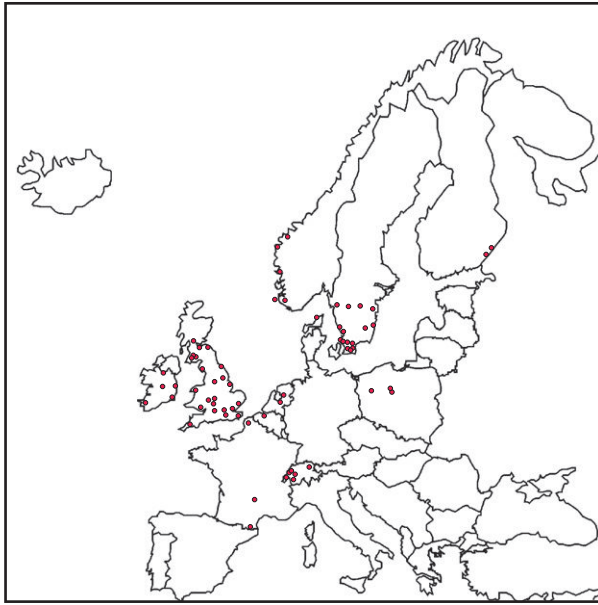


Figure 3 Map showing Late Glacial sites where fossil beetle assemblages were studied.

deposits yielded exclusively arctic and Siberian species. MCR calculations on this assemblage gave the following estimates: $T_{\max}$ 10–11 °C, $T_{\min}$ –26 to –23 °C.

A second beetle assemblage of this age was associated with the recovery of the skeletons of an adult mammoth plus at least three juveniles from silty clay in a kettle hole at Condover, near Shrewsbury (Allen et al., 2009). The GS-2 stadial beetle fauna included many exclusively high arctic species, but unfortunately the wallowing of the mammoths in their efforts to extricate themselves from the adhesive clay had churned up the sediment, incorporating some of the more recent Late Glacial sediment into the stadial deposits, making the precise interpretation of the environment difficult. It is, however, clear that the stadial assemblage of beetles lacked the eastern Asiatic component so characteristic of the full-glacial faunas, suggesting that although the climate must have been of arctic severity at this time, it was probably less continental than during the mid-glacial (pleniglacial) episode.

Only two beetle assemblages of this age are known from northern Europe. Both of these faunas are from Kullaberg, southwestern Sweden, and can be dated with confidence to this period (Lemdahl, 1988, cited in Coope et al., 1998). Although these include only a few species, they resemble those from Britain. Similarly, cold-adapted suites of beetle species are also known from Poland, and from the Swiss Plateau (Coope and Elias, 2000; Coope and Lemdahl, 2009; Elias and Wilkinson, 1983; Gaillard and Lemdahl, 1994; Hardon et al., 2000). MCR estimates gave the following figures: $T_{\max}$ 10–13 °C, $T_{\min}$ –17 to 0 °C.

Late Glacial interstadial

Beetle faunas from the Late Glacial interstadial 1 (GI-1) are widespread across the whole of northwestern Europe. In Britain, important faunas have been described from Glanllynau, North

Wales (Coope and Brophy, 1972, cited in Coope et al., 1998), St Bees, Cumbria (Coope and Joachim, 1980, cited in Coope et al., 1998), Gransmoor, East Yorkshire (Walker et al., 1993, cited in Coope et al., 1998), Hollywell Coombe, Kent (Coope, 1998) and from Llanilid, South Wales (Walker et al., 2003).

Faunas from the early part of the GI-1e interstadial (the Bølling phase) from about 12.8 ky BP (15.1 cal ky BP) or slightly earlier, include an ecologically diverse variety of relatively southern species. At this time the vegetation was thin and made up chiefly of pioneer plant species. The only trees that were present were birches and willows and these were probably confined to the damp valley bottoms. At this time northern Europe as a whole was largely without a woodland cover in spite of the fact that for much of the region, the climate was warm enough to have supported mixed oak forest. Edaphic factors are thought to have been largely responsible for this open landscape. Average MCR estimates from these sites give approximate figures: $T_{\max}$ 17–18 °C, $T_{\min}$ –7 to –1 °C.

Beetle faunas indicate that the initial warming at the start of the interstadial involved a rapid rise in mean July temperatures of at least 7 °C and a corresponding rise in winter temperatures substantially greater than this. The rate at which the mean July temperature rose was about 1 °C per decade and even higher (Coope and Brophy, 1972, cited in Coope et al., 1998). Because this change also involved a switch from continental to oceanic climatic conditions, the rise in *mean annual* temperature was considerably greater than these figures suggest, though the precise number of degrees involved is impossible to estimate with confidence. The beetle reconstruction that yielded this unexpectedly high rate of climatic change has subsequently been supported by evidence from the Greenland Ice cores, which show a dramatic climatic amelioration at about this time (Johnsen et al., 1992).

It is interesting to observe that this episode of rapid climate warming at the start of the Late Glacial Interstadial is not recorded by beetle assemblages from southern Sweden (Lemdahl, 1988, cited in Coope et al., 1998) or from the coast of western Norway. During this period, the Scandinavian beetle assemblages are of an arctic or boreal character. It is likely that this reflects the local cooling effect provided by the close proximity of the waning Fennoscandian ice sheet.

In contrast to the Scandinavian situation, faunas from Poland and the Netherlands (Geel et al., 1989, cited in Coope et al., 1998) indicate temperatures similar to those of the British Isles. At about the same time, warmth demanding beetle faunas are recorded from the Swiss Plateau (Coope and Elias, 2000; Coope and Lemdahl, 2009; Elias and Wilkinson, 1983; Gaillard and Lemdahl, 1994; Hardon et al., 2000). Similar faunal changes are traced in southern France (Ponel and Coope, 1990; Ponel et al., 2001).

Shortly before 12.0 ky BP (ca. 14.0 cal ky BP), there was a sudden climatic cooling. $T_{\max}$ figures suddenly fell to around 15 °C or even lower and never recovered again during the rest of the interstadial; in other words for much of Western Europe the whole of the Allerød phase was substantially cooler than the preceding Bølling phase. The beetles indicate that during the Allerød there appears to have been several relatively minor climatic oscillations, well illustrated by the climatic curves from Gransmoor (Walker et al., 1993, cited in Coope et al., 1998) and Llanilid (Walker et al., 2003, and in summary by



Figure 4 Fieldwork at the small marsh ($\sim 80 \times 30$ m) at Hérémence situated at 2290 m.a.s.l. in the Swiss Alps. Four sites at different altitudes, including this site, were studied in a multiproxy project in order to record biotic responses to rapid climatic changes around the Younger Dryas in the Swiss Alps. Samples were recovered by using a Streif-piston corer of 8 cm diameter.

Lowe and Walker, 1997, cited in Coope et al., 1998). The frequency of these oscillations makes it difficult to give useful MCR estimates for the latter half of the Late Glacial Interstadial.

Again, in contrast to the British Isles and much of Europe, the beetle assemblages from southern Sweden indicate that the highest temperatures were reached during the latter part of the Late Glacial Interstadial. However, it must be emphasized that the actual MCR temperature estimates were similar to the British figures at this time. In Norway arctic faunas are recorded for the Allerød phase (Birks et al., 1993, cited in Coope et al., 1998; Lemdahl, 2000a). In Poland species with a boreal distribution appear during this period (Lemdahl, 1991a, cited in Coope et al., 1998), which also seems to be the case in southern France. These beetle assemblages show that during the GI-1 there were distinct regional differences in climate and steep temperature gradients across much of northern Europe (Coope and Lemdahl, 1995; Coope et al., 1998).

Late Glacial stadial (GS-1) = Younger Dryas

The beetle evidence from the British Isles suggests that the beginning of the GS-1 (Younger Dryas) cold period appears as a culmination of the sequence of Allerød cooling events outlined above. Its beginning is conveniently dated at about 11.0 ky BP (ca. 13.0 cal ky BP). Many beetle assemblages are known from the Younger Dryas period (e.g., Coope et al., 1979, Coope and Joachim, 1980, cited in Coope et al., 1998; Coope, 1998; Walker et al., 1993, 2003) and include a large number of exclusively arctic species, many of which were previously encountered in the assemblages from full-glacial times. However, exclusively Asiatic species are very rare at this time. An average of MCR estimates based on faunas from this period indicates the following figures: $T_{\max}$ 9–11 °C, $T_{\min}$ –10 to –20 °C (Figures 4 and 5).

In northern France, insect assemblages from Younger Dryas fluvial deposits recorded at St.-Momelin, yielded these MCR estimates: $T_{\max}$ around 10 °C and $T_{\min}$ close to –12 °C (Poné



Figure 5 Fossil head and thorax matching a modern specimen of the carrion beetle *Thanatophilus dispar*. Remains of this species were abundant in lake sediments dating to the Oldest Dryas (GS-2) at the site Hauterive Rouges-Terres, at the shore of the Lake Neuchâtel, Switzerland. A Paleolithic settlement was excavated adjacent to the site. Today *T. dispar* is mainly found on fish carrion where it probably feeds on fly larvae.

et al., 2007). MCR reconstructions show that, in contrast to the Late Glacial Interstadial, relatively similar climatic conditions seem to have prevailed in most of northern Europe at this time. In eastern Finland high arctic beetle faunas of Younger Dryas and early post glacial age colonized recently deglaciated areas (Bondestam et al., 1994). Although Younger Dryas beetle assemblages are very rare from the Swiss Plateau, they suggest cold, arctic conditions (Lemdahl, 2000b).

Many sites that include the transition from the Late Glacial to the Holocene show signs of a hiatus, making it difficult to estimate the rate of change involved. However, all the available beetle evidence suggests that the change was abrupt and intense and probably similar in magnitude to the one at the start of the Late Glacial Interstadial (e.g., Ashworth, 1973; Lemdahl, 1991b, cited in Coope et al., 1998; Lemdahl, 2000; Poné et al., 2007). The beetles indicate that throughout most of Europe at the start of the Holocene, temperatures reached present day levels in a geologically immeasurably short period of time.

See also: **Beetle Records: Middle Pleistocene of Western Europe; Overview; Postglacial Europe.**

References

- Allen JRM, Scourse JD, Hall AR, and Coope GR (2009) Palaeoenvironmental context of the Late-Glacial woolly mammoth (*Mammoth primigenius*) discoveries at Condover, Shropshire, UK. *Geological Journal* 44: 414–446.
- Behre K-E, Hölzer A, and Lemdahl G (2005) Botanical macro-remains and insects from the Eemian and Weichselian site at Oerel (NW-Germany) and their evidence for the history of climate. *Vegetation History and Archaeobotany* 14: 31–53.
- Boismier WA, Gamble CS, and Coward F (2011) *Neanderthals Amongst Mammoths: Excavations at Lynford Quarry, Norfolk, UK*. London: English Heritage.
- Bondestam K, Vasari A, Vasari Y, Lemdahl G, and Eskonen K (1994) Younger Dryas and Preboreal in Salpausselkä foreland, Finnish Karelia. *Dissertationes Botanicae* 234: 161–206.
- Bos JAA, Dickson JH, Coope GR, and Jardine WG (2004) Flora, fauna and climate of Scotland during the Weichselian Middle Pleniglacial – Palynological, macrofossil

- and coleopteran investigations. *Palaeogeography, Palaeoclimatology, Palaeoecology* 204: 65–100.
- Brown EJ, Rose J, Coope GR, and Lowe JJ (2007) An MIS 3 age organic deposit from Balglass Burn, central Scotland: Palaeoenvironmental significance and implications for the timing of the onset of the LGM ice sheet in the vicinity of the British Isles. *Journal of Quaternary Science* 22: 295–308.
- Cappers RTJ, Bosch JHA, Bottema S, et al. (1993) De reconstructie van het landschap. In: Sanden WAB, Cappers RTJ, Beuker JR, and Mol D (eds.) *Mens en Mammoet*, Archaeologische Monografieën van Drentes Museum Deel 5, pp. 27–41. Assen: Drentes Museum.
- Coope GR (1959) A Late Pleistocene insect fauna from Chelford, Cheshire. *Proceedings of the Royal Society of London B* 151: 70–86.
- Coope GR (1968) An insect fauna from Mid-Weichselian deposits at Brandon, Warwickshire. *Philosophical Transactions of the Royal Society of London B* 254: 425–456.
- Coope GR (1969) Insect remains from Mid Weichselian Deposits at Peelo, The Netherlands. *Mededelingen Rijk Geologische Dienst* 20: 79–83.
- Coope GR (1987a) The response of late Quaternary insect communities to sudden climatic changes. In: Gee JHR and Giller PS (eds.) *Organisation of Communities – Past and Present*, pp. 421–438. Oxford: Blackwell Scientific.
- Coope GR (1987b) Fossil beetle assemblages as evidence of sudden and intense climatic changes in the British Isles during the last 45,000 years. In: Berger WH and Labeyrie LD (eds.) *Abrupt Climatic Change, Environment and Implications*, pp. 147–150. Dordrecht: Reidel.
- Coope GR (1990) The invasion of Northern Europe during the Pleistocene by Mediterranean species of Coleoptera. In: di Castri F, Hansen AJ, and Debussche M (eds.) *Biological Invasions in Europe and the Mediterranean Basin*, pp. 203–215. Dordrecht: Kluwer.
- Coope GR (1995) Insect faunas in ice age environments: Why so little extinction? In: Lawton JH and May RM (eds.) *Extinction Rates*, pp. 55–74. Oxford: Oxford University Press.
- Coope GR (1998) Insects. In: Prece RC and Bridgland D (eds.) *Late Quaternary Environmental Change in North-west Europe: Excavations at Hollywell Coombe, South-East England*, pp. 213–232. London: Chapman & Hall.
- Coope GR (2000a) The climatic significance of coleopteran assemblages from the Eemian deposits of southern England. *Geologie en Mijnbouw* 79: 257–267.
- Coope GR (2000b) Middle Devensian (Weichselian) coleopteran assemblages from Earith, Cambridgeshire (UK) and their bearing on the interpretation of a 'Full glacial' floras and faunas. *Journal of Quaternary Science* 15: 779–788.
- Coope GR (2002) Changes in the thermal climate in Northwestern Europe during marine oxygen isotope stage 3, estimated from fossil insect assemblages. *Quaternary Research* 57: 401–408.
- Coope GR (2007) Coleoptera from the 2003 excavations of the mammoth skeleton at Niederweningen, Switzerland. *Quaternary International* 164–165: 130–138.
- Coope GR (2011) The insect remains from the mammoth channel at Lynford. In: Boismier WA, Gamble CS, and Coward F (eds.) *Neanderthals Amongst Mammoths: Excavations at Lynford Quarry, Norfolk, UK*. London: English Heritage.
- Coope GR and Angus RB (1975) An ecological study of a temperate interlude in the middle of the last glaciation, based on fossil Coleoptera from Isleworth, Middlesex. *Journal of Animal Ecology* 44: 365–391.
- Coope GR, Dickson JH, McCutcheon JA, and Mitchell GF (1979) The lateglacial and early postglacial deposit at Drumurcher, Co. Monaghan. *Proceedings of the Royal Irish Academy B* 79: 63–85.
- Coope GR and Elias SA (2000) The environment of Upper Palaeolithic (Magdalenian and Azilian) hunters at Hauterive-Champréveyres, Neuchâtel, Switzerland, interpreted from coleopteran remains. *Journal of Quaternary Science* 15: 157–175.
- Coope GR, Gibbard PL, Hall AR, Preece RC, Robinson JE, and Sutcliffe AJ (1997) Climatic and environmental reconstructions based on fossil assemblages from Middle Devensian (Weichselian) deposits of the River Thames at South Kensington, Central London UK. *Quaternary Science Reviews* 16: 1163–1195.
- Coope GR and Lemdahl G (1995) Regional differences in the Lateglacial climate of northern Europe based on coleopteran analysis. *Journal of Quaternary Science* 10: 391–395.
- Coope GR and Lemdahl G (2009) Insect analyses. In: Thew N, Hadorn P, and Coope GR (eds.) *Hauterive/Rouges-Terres. Reconstruction of Upper Palaeolithic and Early Mesolithic Natural Environments*, pp. 125–159. Archéologie neuchâteloise 44.
- Coope GR, Lemdahl G, Lowe JJ, and Walking AP (1998) Temperature gradients in northern Europe during the last glacial–interglacial transition (14–9 ¹⁴C kyr BP) interpreted from coleopteran assemblages. *Journal of Quaternary Science* 13: 419–433.
- Coope GR and Sands CHS (1966) Insect faunas of the last glaciations from the Tame Valley, Warwickshire. *Proceedings of the Royal Society B* 165: 389–412.
- Coope GR, Shotton FW, and Strachan I (1961) A Late Pleistocene fauna and flora from Upton Warren, Worcestershire. *Philosophical Transactions of the Royal Society of London B* 244: 379–421.
- Elias SA and Wilkinson B (1983) Lateglacial insect fossil assemblages from Lobsigensee (Swiss Plateau). *Studies in the Late Quaternary of Lobsigensee* 3. *Revue de Paléobiologie* 2: 189–204.
- Foddai D and Minelli A (1994) Fossil arthropods from a full-glacial site in northeastern Italy. *Quaternary Research* 41: 336–342.
- Gaillard M-J and Lemdahl G (1994) Lateglacial insect assemblages from Grand-Maraïs, South-western Switzerland – Climatic implications and comparison with pollen and plant macrofossil data. *Dissertationes Botanicae* 234: 287–308.
- Hardon P, Thew N, Coope GR, Lemdahl G, Hajdas I, and Bonani G (2000) A Late-Glacial and Early Holocene environment and climate history for the Neuchâtel region (CH). In: Richard H and Vignot A (eds.) *Équilibres et ruptures dans les écosystèmes Durant les 20 derniers millénaires en Europe de l'Ouest, Actes du colloque international de Besançon, septembre 2000*, pp. 75–90. Besançon: Presses Universitaires Franc-Comtoises.
- Johnsen SJ, Clausen HB, Dansgaard W, et al. (1992) Irregular glacial interstadials recorded in a new Greenland Ice core. *Nature (London)* 359: 311–313.
- Jost-Stauffer M, Coope GR, and Schlüchter C (2005) Environmental and climatic reconstructions during Marine Oxygen Isotope Stage 3 from Gossau, Swiss Midlands, based on coleopteran assemblages. *Boreas* 34: 53–60.
- Kasse C, Huitzer AS, Krzyszowski D, Bohncke SJ, and Coope GR (1998) Weichselian Late-Pleniglacial and Late-glacial depositional environment, Coleoptera and periglacial climate records from central Poland (Belchatów). *Journal of Quaternary Science* 13: 455–469.
- Lemdahl G (1997) Early Weichselian insect faunas from northern Sweden: Climatic and environmental implications. *Arctic and Alpine Research* 29: 63–74.
- Lemdahl G (2000a) Late-glacial and Early Holocene Coleoptera assemblages as indicators of local environment and climate at Kråkenes Lake, western Norway. *Journal of Paleolimnology* 23: 57–66.
- Lemdahl G (2000b) Lateglacial and Early Holocene insect assemblages from sites at different altitudes in the Swiss Alps – implications on climate and environment. *Palaeogeography, Palaeoclimatology, Palaeoecology* 159: 293–312.
- Lewis SG, Maddy D, Buckingham C, et al. (2006) Pleistocene fluvial sediments, paleontology and archaeology of the upper River Thames at Latton, Wiltshire, England. *Journal of Quaternary Science* 21: 181–205.
- Lindroth CH (1948) Interglacial insect remains from Sweden. *Årsbok Sveriges Geologiska Undersökning C42*: 1–29.
- Maddy D, Lewis SG, Scaife RG, et al. (1998) The Upper Pleistocene deposits at Cassington, near Oxford, England. *Journal of Quaternary Science* 13: 205–231.
- Penny LF, Coope GR, and Catt JA (1969) Age and insect fauna of the Dimlington Silts, East Yorkshire. *Nature* 224: 65–67.
- Ponel P (1995) Rissian, Eemian and Würmian Coleoptera assemblages from La Grande Pile (Vosges, France). *Palaeogeography, Palaeoclimatology, Palaeoecology* 114: 1–41.
- Ponel P and Coope GR (1990) Lateglacial and early Flandrian from La Taphanel, Massif Central, France: climatic and ecological implications. *Journal of Quaternary Science* 5: 235–249.
- Ponel P, Andrieu-Ponel V, Parchoux F, Juhasz I, and de Beaulieu J-L (2001) Late-glacial and Holocene high-altitude environmental changes in Vallée des Merveilles (Alpes Maritimes, France): Insect evidence. *Journal of Quaternary Science* 16: 795–812.
- Ponel P, Gaudouin E, Coope GR, et al. (2007) Insect evidence for environmental and climate changes from Younger Dryas to Sub-Boreal in a river floodplain at St-Momelin (St-Omer basin, northern France), Coleoptera and Trichoptera. *Palaeogeography, Palaeoclimatology, Palaeoecology* 245: 483–504.
- Taylor BJ and Coope GR (1985) Arthropods in the Quaternary of East Anglia – Their role as indices of local Palaeoenvironments and regional Palaeoclimates. *Modern Geology* 9: 159–185.
- Walker MJC, Björck S, Lowe JJ, et al. (1999) Isotopic 'events' in the GRIP ice-core: A stratotype for the Late Pleistocene. *Quaternary Science Reviews* 18: 1143–1150.
- Walker MJC, Coope GR, and Lowe JJ (1993) The Devensian (Weichselian) Lateglacial palaeoenvironmental record from Gransmoor, East Yorkshire, England. *Quaternary Science Reviews* 12: 659–680.
- Walker MJC, Coope GR, Sheldrick C, et al. (2003) Devensian Lateglacial environmental changes in Britain: A multi-proxy environmental record from Llanilid, South Wales, UK. *Quaternary Science Reviews* 22: 475–520.
- Walking AP (1997) Käferkundliche Untersuchungen an weiselzeitlichen Ablagerungen der Bohrung Gross Todtshorn (Kr. Harburg; Niedersachsen). *Schriftenreihe der Deutschen Geologischen Gesellschaft* 4: 87–102.
- Walking AP and Coope GR (1996) Climatic reconstructions from the Eemian/early Weichselian transition in Central Europe based on the coleopteran record from Göbern, Germany. *Boreas* 25: 145–159.

Late Pleistocene of Japan

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Introduction

The study of Japanese Quaternary insects began in 1978 at the Tategahana Site of the Nojiri-ko Site Group, Nagano Prefecture. This study was carried out by the Fossil Insect Research Group for Nojiri-ko Excavation (FIRGNE). Tategahana is a Paleolithic site, yielding bone artifacts and large-mammal fossils, formed on the shore of Lake Nojiri (Japanese name, Nojiri-ko) during the last glaciation. In total, the research group has so far identified 85 species of beetles from the site (see discussion of the Nojiri-ko insect fauna, below). This research group also studied the methodology of Quaternary Entomology and published a Japanese handbook on fossil insects (FIRGNE, 1988).

In Japan, Quaternary peat beds have yielded abundant fossil beetles, especially ground beetles (Carabidae) and donaciine leaf beetles (Donaciinae). Last glacial peat beds are known from many Japanese regions but only a few studies on fossil beetles have so far been performed. However, Japanese Quaternary entomologists have recorded 115 species of beetles from 16 sites (Tables 1, 2). These beetle records are reviewed here.

Fossil Sites

Sixteen fossil sites are shown in Table 1 and Figure 1. Most sites are on Honshu, the main island of Japan, and only one site is on Kyushu, in western Japan (Figure 1). The late Pleistocene fossil beds are mainly discovered in archaeological excavations, terrace deposits, and lake deposits.

The Owatari II Site is on the lower terrace of Waga River, Yuda-machi, Iwate Prefecture. The terrace deposit is intercalated with three peat beds, identified as the “1st” peat (ca. 14–12 ka), the “2nd” peat (ca. 23–19 ka), and the “3rd” peat (ca. 25–24 ka), respectively (Mori, 1995). Twenty-eight taxa of beetles from these deposits have been recorded by Mori (1995). These include *Elaphrus japonicus* (Carabidae), *Ilybius poppiusi* (Dytiscidae), *Dineutus orientalis* (Gyrinidae), *Enochrus* sp. (Hydrophilidae), *Plesiophthalmus* sp. (Tenebrionidae), and *Plateumaris sericea* (Chrysomelidae). Mori recognized two unidentified species of the genus *Agabus* (Dytiscidae) as being extinct species in Japan.

The Tomizawa Site is on Sendai Plain, Sendai City, Miyagi Prefecture. Late Pleistocene beetles were found from the ‘27th’ to ‘25th’ beds of the site. Mori and Itoh (1992) recorded 43 taxa of beetles from the beds, such as *Pterostichus* sp. (Carabidae), *Noterus japonicus* (Noteridae), *Ilybius poppiusi* (Dytiscidae), *Laccobius bedeli* (Hydrophilidae), *Geotrupes auratus* (Scarabaeidae), *Donacia splendens* (Chrysomelidae), and *Limnobaris* sp. (Curculionidae).

The Maebashi Peat Formation (Figure 2) is found in the Maebashi Terrace, Takasaki City and Maebashi City, Gunma

Prefecture. The formation is characterized by peat beds intercalated with silt, sand, gravel beds, and volcanic ash layers. Hayashi (1996, 2005) reported fossil beetles from peat beds at two sites, Sojya in Maebashi City and Shimowada in Takasaki City. The base of the peat at the Shimowada site is older than that of the Sojya site. The age of the Shimowada peat is ca. 2.0 ka (Nirei and Hayashi, 2004). Hayashi (1996) recorded 10 taxa from the Sojya Site and Hayashi (2005) recorded 14 taxa from the Shimowada Site.

The Tsukumogawa Peat Bed is in the lower terrace deposit of the Tsukumogawa River, Annaka City, Gunma Prefecture. The Aira-Tn tephra, dated ca. 25–24 ka, overlies the peat bed. Hayashi and Shimadu (2005) recorded 12 taxa of beetles from the bed, such as *Chlaenius* sp. (Carabidae), *Agabus* sp. *Ilybius* sp., *Hydrochara libera* (Hydrophilidae), and *Donacia splendens*.

The Nukui Peat Bed is found in Nukui, Agastuma-machi, Gunma Prefecture. The peat bed is intercalated with the Aira-Tn tephra. *Damaster blaptoides* (Carabidae) and *Plateumaris* sp. (Chrysomelidae) were found in this bed.

The Taguro Peat Bed is distributed in Taguro, Tamagawamura, Saitama Prefecture. A radiocarbon date from the wood was more than 38 020 yr BP (Gak-19303). Nirei and Hayashi (1998) recorded six taxa of beetles from this deposit: *Elaphrus japonicus*, *Tachyura* sp., *Agabus conspicuus*, *Plateumaris constricticollis*, *Plateumaris sericea*, and *Donacia splendens*.

The Nojiri-ko Site Group in Shinano-machi, Nagano Prefecture includes Tategahana, Nakamachi, Ikejirigawa, and other sites (Figure 3). Fossil beetles were found from three of the sites. The Upper Pleistocene in the site groups is divided into three strata: the Biwajima-oki Peat Formation (ca. 100–70 ka), the Kannoki Formation (ca. 60 ka) and the Nojiri-ko Formation (ca. 50–10 ka), respectively. The Nojiri-ko Excavation Research Group has excavated the Nojiri-ko Formation from these three sites. The Biwajima-oki Peat Formation and the Kannoki Formation have been excavated at the Nakamachi Site. The last glacial beetles from the group of sites have been reported by FIRGNE (1980, 1984, 1987, 1990, 1993a,b, 1996a,b, 2000a,b, 2003).

The Ono Peat bed is located in Ono, Tatsuno-machi, Nagano Prefecture. Its age is ca. 60 ka. Hayashi (1998) recorded eight species of beetles from the bed, including well-preserved fossils of *Agabus japonicus* (Dytiscidae) and *Plateumaris sericea*.

The Miomote Site Group is on river terraces of the Miomote River, Asahi-mura Niigata Prefecture. This group includes Tarukuchi (Figure 4), Garahagi, and other sites. Two peat beds were found from the Tarukuchi Site. The upper peat bed overlies the Aira-Tn tephra. These peat beds yielded *Carabus vanvolxemi* (Carabidae), *Agabus miyamotoi* (Dytiscidae), *Silpha longicornis* (Silphidae), *Plateumaris sericea* and some undetermined beetles (Hayashi and Miyatake, 1996). A radiocarbon age from the peat bed of the Garahagi Site was more than

Table 1 Late Pleistocene fossil sites in Japan

No.	Site	Strata	Age	References
1	Iwate Pref. Owatari II Site, Yuda-machi Miyagi Pref.	3rd to 1st Peat Beds	ca. 25-12 ka	Mori, 1995
2	Tomizawa Site, Sendai City Gunma Pref.	27-25th Beds	ca. 24-19 ka	Mori and Itoh, 1992
3	Sojya, Maebashi City	Maebashi Peat Bed	ca. 14-13 ka	Hayashi, 1996
4	Shimowada, Takasaki City	Maebashi Peat Bed	ca. 24-13 ka	Hayashi, 2005
5	Taira, Annaka City	Tsukumogawa Peat Bed	ca. 25-24 ka	Hayashi and Shimadu, 2005
6	Nukui, Agatsuma-machi Saitama Pref.	Nukui Peat Bed	ca. 25-24 ka	Hayashi, 1999
7	Taguro, Tamagawa-mura Nagano Pref.	Taguro Peat Bed	Last glacial age	Nirei and Hayashi, 1998
8	Ikejirigawa Site, Shinano-machi	Nojiri-ko Formation	ca. 50-10 ka	FIRGNE, 1987, 1990
9a	Nakamachi Site, Shinano-machi	Nojiri-ko Formation	ca. 50-10 ka	FIRGNE, 1993b, 1996b, 2000b
9b	Nakamachi Site, Shinano-machi	Kannoki Formation	ca. 60 ka	FIRGNE, 1993b, 1996b, 2000b
9c	Nakamachi Site, Shinano-machi	Biwajimaoki Peat F.	ca. 70 ka	FIRGNE, 1993b, 1996b, 2000b
10	Tategahana Site, Shinano-machi	Nojiri-ko Formation	ca. 50-10 ka	FIRGNE, 1980, 1984, 1987, 1990, 1993a, 1996a, 2000a, 2003
11	Ono, Tatsuno-machi Niigata Pref.	Ono Peat Bed	ca. 70-60 ka	Hayashi, 1998
12	Tarukuchi Site, Asahi-mura	Terrace deposit	ca. 25-24 ka	Hayashi and Miyatake, 1996
13	Garahagi Site, Asahi-mura	Terrace deposit	ca. 40-30 ka	Hayashi and Miyatake, 1996
14	Kujiranami, Kashiwazaki City	Yasuda Formation	ca. 100 ka	Kashiwazaki Naumann's Elephant Research Group, 1991
15	Gifu Pref. Miyanomae Site, Miyagawa-mura Kagoshima Pref.	Terrace deposit	ca. 14-12 ka	Mori et al. 1997
16	Shimogose, Yoshimatsu-cho	Mizozono Formation	unknown	Hayashi et al. 2002

Table 2 Late Pleistocene beetle records in Japan. Site number refers to [Table 1](#)

Scientific name/Fossil site	1	2	3	4	5	6	7	8	9a	9b	9c	10	11	12	13	14	15	16
Carabidae																		
<i>Calosoma inquisitor</i> (Linnaeus)												+						
<i>Carabus albrechti</i> Morawitz												+						
<i>Carabus granulatus</i> Linne								+	+		+	+						
<i>Carabus insulicola</i> Chaudoir								+			+	+				+		
<i>Carabus vanvolxemi</i> Putzeys								+			+			+				
<i>Damaster blaptoides</i> Kollar						+						+						
<i>Elaphrus japonicus</i> S.Ueno	+						+	+			+				+			
<i>Scarites terricola pacificus</i> Bates																+		
<i>Trigonognatha aurescens</i> Bates														+				
<i>Lesticus magnus</i> (Motschulsky)																+		
<i>Pterostichus leptis</i> Bates																	+	
<i>Pterostichus longinquus</i> Bates											+							
<i>Pterostichus microcephalus</i> (Motschulsky)											+							
<i>Pterostichus planicollis</i> (Motschulsky)									+							+		
<i>Pterostichus prolongatus</i> Morawitz											+							
<i>Colpodes japonicus</i> (Motschulsky)											+							
<i>Epomis nigricans</i> (Wiedemann)																+		
<i>Chlaenius circumdatus</i> Brulle																+		
<i>Chlaenius gebleri</i> Ganglbauer										+	+							
<i>Chlaenius pallipes</i> Gebler											+							
<i>Calleida lepida</i> Redtenbacher											+							
Dytiscidae																		

(Continued)

Table 2 (Continued)

Scientific name/Fossil site	1	2	3	4	5	6	7	8	9a	9b	9c	10	11	12	13	14	15	16
<i>Platambus pictipennis</i> (Sharp)									+			+						
<i>Agabus conspicuus</i> Sharp							+		+									
<i>Agabus japonicus</i> Sharp				+							+		+					
<i>Agabus miyamotoi</i> Nakane														+				
<i>Agabus optatus</i> Sharp												+						
<i>Ilybius apicalis</i> Sharp								+	+									
<i>Ilybius poppiusi</i> Zaitsev	+	+		+														
<i>Ilybius weymarni</i> Balfour-Browne			+															
<i>Rhantus erraticus</i> Sharp											+							
<i>Dytiscus marginalis czerskii</i> Zaitzev												+						
<i>Dytiscus sharpi</i> Wehncke										+								
<i>Cybister brevis</i> Aube								+			+							
<i>Cybister japonicus</i> Sharp																+		
Hydrophilidae																		
<i>Coelostoma orbiculare</i> (Fabricius)	+			+	+			+	+		+	+			+	+		+
<i>Coelostoma stultum</i> (Walker)		+						+				+						
<i>Pachysternum haemorrhoum</i> Motschulsky												+						
<i>Anacaena asahinai</i> M.Sato	+								+									
<i>Enochrus japonicus</i> (Sharp)											+					+		
<i>Hydrophilus acuminatus</i> Motschulsky								+			+	+				+		
<i>Sternolophus rufipes</i> (Fabricius)												+				+		
<i>Hydrochara affinis</i> (Sharp)								+				+						
<i>Hydrochara libera</i> (Sharp)				+	+			+			+	+						+
<i>Regimbartia attenuata</i> (Fabricius)	+																	
Gyrinidae																		
<i>Dineutus orientalis</i> (Modeer)	+							+				+						
Histeridae																		
<i>Atholus duodecimstriatus</i> (Gyllenhal)											+							
<i>Hister concolor</i> Lewis												+						
<i>Hister simplicisternus</i> Lewis																	+	
<i>Margarinotus niponicus</i> (Lewis)											+							
Silphidae																		
<i>Nicrophorus vespilloides</i> (Herbst)											+							
<i>Oiceoptoma thoracicum</i> (Linne)												+						
<i>Silpha longicornis</i> Portevin								+				+		+				
<i>Eusilpha japonica</i> (Motschulsky)												+						+
<i>Phosphuga atrata</i> (Linnaeus)												+						
Staphylinidae																		
<i>Paederus parallelus</i> Weise												+						
Scaphidiidae																		
<i>Scaphidium rufopygum</i> Lewis												+						
Lucanidae																		
<i>Nipponodorcus rubrofemoratus</i> (Snellen van Vollenhoven)												+						
<i>Platycerus acuticollis</i> Y. Kurosawa								+										
Geotrupidae																		
<i>Geotrupes auratus</i> Motschulsky								+			+	+			+			
Scarabaeidae																		
<i>Copris pecuarius</i> Lewis												+						
<i>Copris tripartitus</i> Waterhouse																+		
<i>Caccobius jessoensis</i> Harold												+						
<i>Caccobius nikkoensis</i> Lewis											+							
<i>Onthophagus lenzii</i> Harold											+							+
<i>Aphodius brachysomus</i> Solsky								+				+						
<i>Aphodius brevisculus</i> (Motschulsky)												+						
<i>Aphodius elegans</i> Allibert												+						
<i>Aphodius igai</i> Nakane												+						
<i>Aphodius quadratus</i> Reiche								+			+							
<i>Aphodius rectus</i>											+	+						

(Continued)

Table 2 (Continued)

Scientific name/Fossil site	1	2	3	4	5	6	7	8	9a	9b	9c	10	11	12	13	14	15	16
<i>Aphodius rufipes</i> (Linne)											+	+						
<i>Aphodius yamato</i> Nakane												+						
<i>Phyllopertha diversa</i> Waterhouse		+																
<i>Mimela holosericea</i> (Fabricius)		+																
<i>Mimela testaceipes</i> Motschulsky												+						+
<i>Mimela cuprea</i> Hope												+						
<i>Anomala octiescostata</i> (Burmeister)																+		
<i>Rhomborrhina japonica</i> Hope																	+	
<i>Rhomborrhina unicolor</i> Motschulsky											+	+				+		
<i>Eucetonia roelofsi</i> (Harold)										+		+						
Byrrhidae																		
<i>Cytilus sericeus</i> (Forster)										+	+		+					
Elateridae																		
<i>Selatosomus puncticollis</i> (Motschulsky)											+							
Lagriidae																		
<i>Luprops cribrifrons</i> Marseul																	+	
Trogossitidae																		
<i>Leperina squamulosa</i> (Gebler)												+						
Coccinellidae																		
<i>Harmonia axyridis</i> (Pallas)												+						
Cerambycidae																		
<i>Pterolophia caudata</i> Bates												+						
<i>Spondylis buprestoides</i> Linnaeus																	+	
Chrysomelidae																		
<i>Plateumaris constricticollis</i> (Jacoby)		+	+	+	+		+	+	+		+	+			+			
<i>Plateumaris sericea</i> (Linne)	+						+				+		+	+				
<i>Plateumaris shirahatai</i> Kimoto		+																
<i>Donacia flemora</i> Goecke																		+
<i>Donacia japana</i> Chujo et Goecke								+			+							+
<i>Donacia lenzi</i> (Schonfeldt)									+		+					+		+
<i>Donacia ozensis</i> Nakane											+	+				+		
<i>Donacia provostii</i> Fairmaire																+		
<i>Donacia sparganii</i> Ahrens								+			+							
<i>Donacia splendens</i> Jacobson				+	+		+	+	+		+	+			+			
<i>Donacia vulgaris</i> Zschach																+		
<i>Basilepta fulvipes</i> (Motschulsky)																+		
<i>Chrysolina aurichalcea</i> (Mannerheim)												+		+		+		
<i>Chrysolina exanthematica</i> (Wiedemann)				+														
<i>Gastrolina depressa</i> Baly								+										
<i>Gastrolina peltoides</i> (Gebler)												+						
<i>Linnaea aenea</i> (Linne)											+	+						
<i>Goniocetena japonica</i> Chujo et Kimoto												+						
<i>Cneorane elegans</i> Baly																	+	
<i>Liroetis coeruleipennis</i> Weise		+																
<i>Agelastica coerulea</i> Baly												+						
Attelabidae																		
<i>Byctiscus puberulus</i> (Motschulsky)	+											+						
<i>Byctiscus venustus</i> (Pascoe)												+						
<i>Euops punctatostriatus</i> (Motschulsky)												+						
Curculionidae																		
<i>Limnobaris japonica</i> Yoshihara et Morimoto					+						+							
<i>Hylobitelus pinastri</i> (Gyllenhal)	+																	
<i>Dyscerus roelofsi</i> (Harold)								+				+						
<i>Niphades variegatus</i> (Roelofs)	+																	

35 900 yr BP (Gak-18168). Hayashi and Miyatake (1996) recorded nine taxa of beetles from the bed, including *Elaphrus japonicus*, *Agabus* sp., *Coelostoma orbiculare*, *Donacia splendens*, and *Plateumaris constricticollis*.

The Yasuda Formation is a deposit from the last interglacial (MIS 5e) in Kashiwazaki City, Niigata Prefecture. Abundant fossil beetles comprising 24 taxa were obtained from the formation during the excavation of the Naumann's Elephant



Figure 1 Late Pleistocene fossil sites in Japan. Site numbers match those in Table 1.



Figure 2 The Maebashi Peat Formation in Shimowada, Takasaki City, Gunma Pref. © 2005, Photograph, M. Hayashi.

(*Palaeoloxodon naumanni*) in 1986 (Kashiwazaki Naumann's Elephant Research Group, 1991). Beetle faunas from the last interglacial are poorly studied in Japan. The fauna is dominated by water beetles, ground beetles, plant-feeding beetles, and dung beetles. Most species are still found today at the fossil site and in the surrounding regions, but a dung beetle, *Copris tripartitus* is no longer living on the island of Honshu. The ground beetles include *Carabus insulicola*, *Scarites terricola pacificus*, *Lesticus magnus*, *Pterostichus planicollis*, *Epomis nigricans* and *Chlaenius circumdatus*. The plant-feeding beetles include four donaciine beetles of *Donacia ozensis*, *Donacia lenzi*, *Donacia provostii* and *Donacia vulgaris*. The paleoenvironments of the Yasuda Formation have been reconstructed as ranging from

forest edges to grassland, surrounding an area of still water with emergent and floating vegetation.

The Miyanomae Site is on a terrace of the Miyagawa River, Miyagawa-mura, Gifu Prefecture. Twenty-four taxa of beetles were found from the late Paleolithic horizon, radiocarbon dated between 14–12 ka (Mori et al., 1997).

The Mizozono Formation (Figure 5) is a late Pleistocene deposit in Yoshimatsu-cho, Kagoshima Prefecture, but its exact age is unknown. Hayashi et al. (2002) recorded 36 taxa of beetles, such as *Hister simplicisternus* (Histeridae), *Eusilpha japonica* (Silphidae), and *Onthophagus lenzii* (Scarabaeidae), and *Donacia flemora* (Chrysomelidae) from this site.

Fossil Records

The fossil beetles identified from the various Upper Pleistocene fossil assemblages in Japan are shown in Table 2.

Carabidae

Abundant ground beetle fossils are typically found in Japanese peat beds, but their specific identification has been difficult for Japanese Quaternary entomologists. Therefore, most fossils have been identified only to the subfamily or tribe level. Twenty-one species are recorded from ten sites (Table 2): six species of Carabinae, one species of Elaphrinae, one species of Scaritidae, eight species of Pterostichinae, four species of Callistinae, and one species of Lebiinae. *Elaphrus japonicus* (Figure 6) is recorded from five sites and *Carabus granulatus* (Figure 7) is recorded from four sites. Today these species inhabit marshes.

Dytiscidae

Thirteen species of predaceous diving beetles are recorded from eleven Japanese sites (Table 2). These comprise nine species of Colymbetinae and four species of Dytiscinae. *Agabus japonicus* (Figure 8) and *Ilybius poppiusi* (Figure 9) are each known from three sites. Two species of *Dytiscus* are recorded from the Nojiri-ko Site Group: *D. marginalis czerskii* (Figure 10) from the Nojiri-ko Formation in the Tategahana, and *D. sharpi* (Figure 11) from the Kannoki Formation in the Nakamachi Site.

Hydrophilidae

Ten species of water scavenger beetles have been recorded from ten sites (Table 2): three species of Sphaeridiinae, two species of Hydrobiinae, four species of Hydrophilinae, and one species of Chaetritrinae. *Coelostoma orbiculare* (Figure 12) is known from nine sites and *Hydrochara libera* is known from six sites. Most of these beetles are aquatic species but *Pachysternum haemorrhoum* is a dung-feeder.

Gyrinidae

One species of whirligig beetle, *Dineutus orientalis*, is recorded from the Owatari II, Ikejirigawa, and Tategahana Sites. Undetermined specimens of *Gyrinus* spp. are known from several sites.



Figure 3 The Tategahana Site in Shinano-machi, Nagano Pref. At 12th Excavation in 1993. © 2005, Photograph, FIRGNE.



Figure 4 (a), The Tarukuchi Site in Asahi-mura, Niigata Pref.; (b), peat bed in the site. © 2005, Photograph, M. Hayashi.

Histeridae

Four species of hister beetles are recorded from three sites (Table 2): *Atholus duodecimstriatus* (Figure 13) and



Figure 5 The Mizozono Formation in Shimogose, Yoshimatsu-cho, Kagoshima Pref. © 2005, Photograph, E. Kitabayashi.

Margarinotus niponicus from the Nakamachi Site, *Hister concolor* (Figure 14) from the Tategahana Site, and *Hister simplicisternus* from Shimogose, Kagoshima Prefecture.

Silphidae

Five species of carrion beetles are recorded from five sites: one species of Nicrophorini and four species of Silphini. *Silpha longicornis* (Figure 15) is known from three sites. This species is currently found in the cool-temperate zone forests in Japan.

Staphylinidae

Indetermined fossils of the rove beetle family are commonly found from the Pleistocene peat beds but one species, *Paederus parallelus* is recorded from the Tategahana Site.

Scaphidiidae

Fossils of this family (shining fungus beetles) are poorly known in Japan. One species, *Scaphidium rufopygum* (Figure 16) is recorded from the Tategahana Site.



Figure 6 *Elaphrus japonicus*, mesosternum and right elytron, from the Nojiri-ko Formation in the Ikejirigawa Site. © 2005, Photograph, FIRGNE. Scale bar = 1 mm.

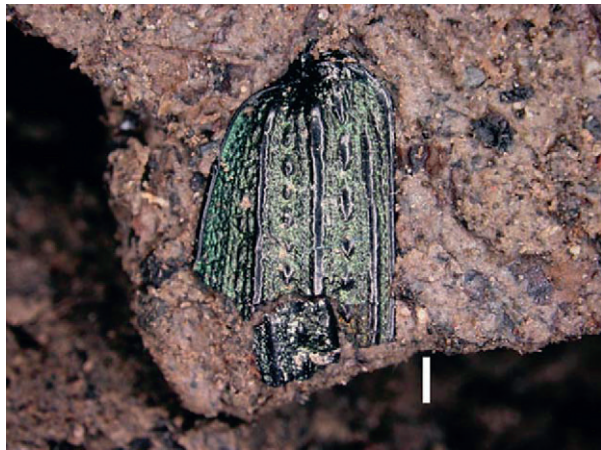


Figure 7 *Carabus granulatus*, left elytron, from the Nojiri-ko Formation in the Tategahana Site. © 2005, Photograph, FIRGNE. Scale bar = 1 mm.

Lucanidae

Two species of stag beetles are recorded (Table 2): *Nipponodorcus rubrofemoratus* from the Tategahana Site and *Platycerus acuticollis* (Figure 17) from the Ikejirigawa Site. *P. acuticollis* currently lives in the cool-temperate zone forest in Japan.

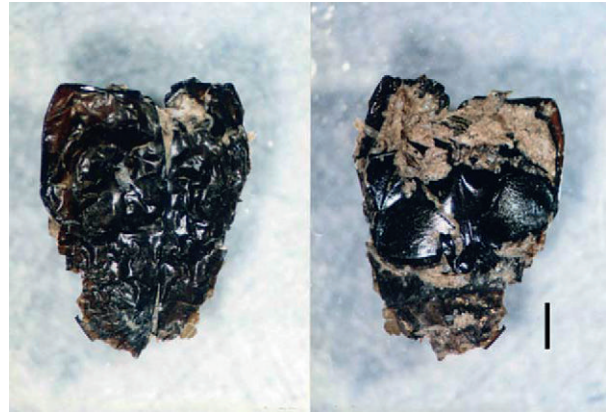


Figure 8 *Agabus japonicus*, elytra and sternum, from the Ono Peat Bed. © 2005, Photograph, M. Hayashi. Scale bar = 1 mm.



Figure 9 *Ilybius poppiusi*, left elytron, from the Maebashi Peat Formation. © 2005, Photograph, M Hayashi. Scale bar = 1 mm.

Geotrupidae

One species of earth-boring dung beetle, *Geotrupes auratus* (Figure 18), is recorded from four sites. It is found today in broad-leaved forests of mountain regions, where it feeds on animal dung.

Scarabaeidae

Twenty-one scarab beetle species are recorded from seven sites (Table 2): five species of Scarabaeinae, eight species of Aphodiinae, five species of Rutelinae, three species of Cetoniinae.



Figure 10 *Dytiscus marginalis czerskii*, right elytron, from the Nojiri-ko Formation in the Tategahana Site. © 2005, Photograph, FIRGNE. Scale bar = 1 mm.



Figure 11 *Dytiscus sharpi*, head, from the Kannoki Formation in the Nakamachi Site. © 2005, Photograph, FIRGNE. Scale bar = 1 mm.

Thirteen species of Scarabaeinae and Aphodiinae are dung-feeders and the other species of subfamilies are plant-feeders.

Byrrhidae

One species of pill beetle, *Cytilus sericeus* (Figure 19), is recorded from the Nakamachi Site and Ono, Nagano Prefecture.

Elateridae

Undetermined fossils of click beetles are commonly found from the Pleistocene peat beds. One species, *Selatosomus puncticollis* is recorded from the Nakamachi Site.



Figure 12 *Coelostoma orbiculare*, left elytron, from the Nojiri-ko Formation in the Ikejirigawa Site. © 2005, Photograph, FIRGNE. Scale bar = 1 mm.



Figure 13 *Atholus duodecimstriatus*, right elytron, from the Biwajima-oki Peat Formation the Nakamachi Site. © 2005, Photograph, FIRGNE. Scale bar = 1 mm.

Lagriidae

Fossils of the long-jointed beetle family are poorly known in Japan. One species, *Luprops cribrifrons* is recorded from the Miyanomae Site.

Trogossitidae

Fossils of the bark-gnawing beetle family are poorly known in Japan. One species, *Leperina squamulosa* (Figure 20) is recorded from the Tategahana Site.

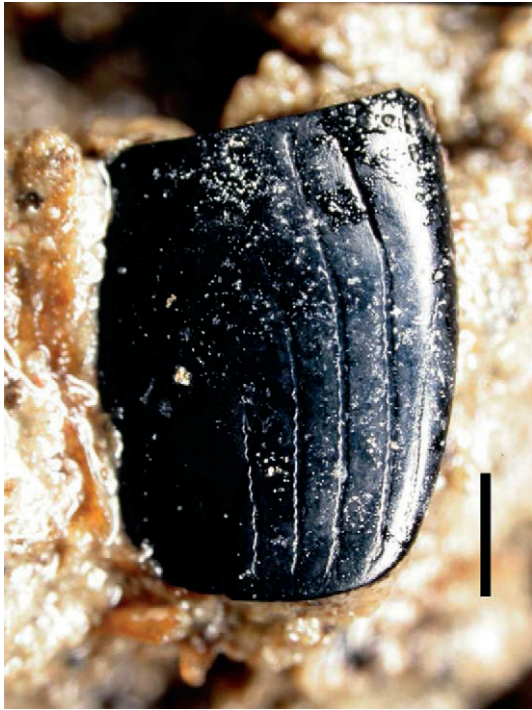


Figure 14 *Hister concolor*, right elytron, from the Nojiri-ko Formation in the Tategahana Site. © 2005, Photograph, FIRGNE. Scale bar = 1 mm.



Figure 16 *Scaphidium rufopygum*, left elytron, from the Nojiri-ko Formation in the Tategahana Site. © 2005, Photograph, FIRGNE. Scale bar = 1 mm.

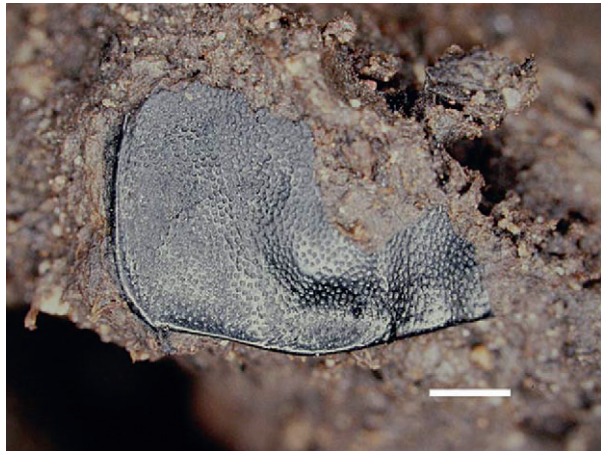


Figure 15 *Silpha longicornis*, pronotum, from the Nojiri-ko Formation in the Ikejirigawa Site. © 2005, Photograph, FIRGNE. Scale bar = 1 mm.



Figure 17 *Platycerus acuticollis*, pronotum, from the Nojiri-ko Formation in the Ikejirigawa Site. © 2005, Photograph, FIRGNE. Scale bar = 1 mm.

Coccinellidae

Fossils of the ladybird family are poorly known in Japan. One species, *Harmonia axyridis* is recorded from the Tategahana Site.

Cerambycidae

Two species of long-horn beetles are recorded, *Pterolophia caudata* (Figure 21) from the Tategahana Site and *Spondylis buprestoides* from the Miyanomae Site.

Chrysomelidae

Twenty-one species of leaf beetles are recorded from fifteen sites (Table 2): eleven species of Donaciinae, one species of Eumolpinae, six species of Chrysomelinae, and three species of Galerucinae. *Plateumaris constricticollis* (Figure 22) is known from nine sites and *Donacia splendens* (Figure 23) is known from seven sites. These two species of Donaciinae are commonly found from the Pleistocene peat beds in Japan.



Figure 18 *Geotrupes auratus*, protibia, from the Nojiri-ko Formation in the Ikejirigawa Site. © 2005, Photograph, FIRGNE. Scale bar = 1 mm.



Figure 19 *Cytilus sericeus*, right elytron, from the Biwajima-oki Peat Formation in the Nakamachi Site. © 2005, Photograph, FIRGNE. Scale bar = 1 mm.



Figure 20 *Leperina squamulosa*, pronotum, from the Nojiri-ko Formation in the Tategahana Site. © 2005, Photograph, FIRGNE. Scale bar = 1 mm.



Figure 21 *Pterolophia caudata*, right elytron, from the Nojiri-ko Formation in the Tategahana Site. © 2005, Photograph, FIRGNE. Scale bar = 1 mm.

Attelabidae

Three species of leaf-rolling beetles are recorded from two sites (Table 2): *Byctiscus puberulus* from the Owatari II Site and the Tategahana Site, *Byctiscus venustus* and *Euops punctatostriatus* from the Tategahana Site.

Curculionidae

Four weevil species are recorded from five sites (Table 2): one species of Baridinae and three species of Hylobiinae. *Limnobaris japonica* (Figure 24) is known from two sites that lives in marsh and eats leaves of *Carex* spp.



Figure 22 *Plateumaris constricticollis*, pronotum, from the Nojiri-ko Formation in the Ikejirigawa Site. © 2005, Photograph, FIRGNE. Scale bar = 1 mm.



Figure 24 *Limnobaris japonica*, head and pronotum, from the Biwajima-oki Peat Formation in the Nakamachi Site. © 2005, Photograph, FIRGNE. Scale bar = 1 mm.

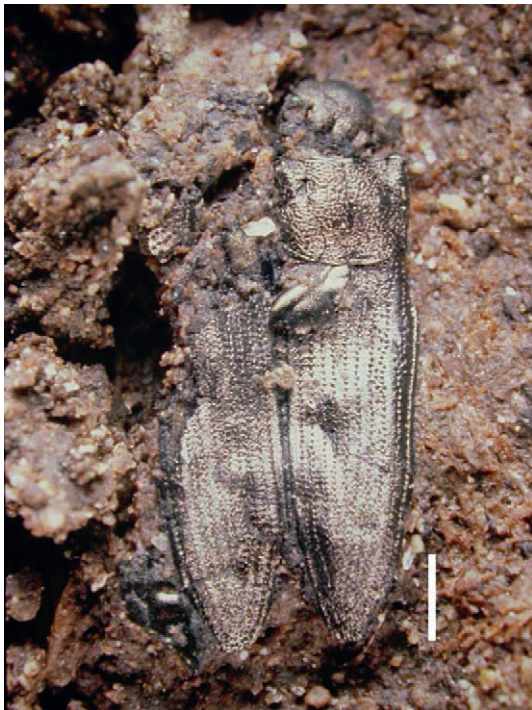


Figure 23 *Donacia splendens*, pronotum and elytra, from the Biwajima-oki Peat Formation in the Nakamachi Site. © 2005, Photograph, FIRGNE. Scale bar = 1 mm.

Nojiri-Ko Insect Fauna

The Nojiri-ko insect fauna comprises the last glacial insects from the Nojiri-ko Site Group, including 85 species of beetles from the Nojiri-ko, the Kannoki, and the Biwajima-oki Peat Formations. The fauna is dominated by water beetles, ground beetles, plant-feeding beetles, and dung beetles. Most species are found today in the cool-temperate zone of Japan.

The aquatic beetles include Dytiscidae, Hydrophilidae and Gyrinidae. Species of *Agabus*, *Ilybius*, *Rhantus*, *Dytiscus*, *Cybister*

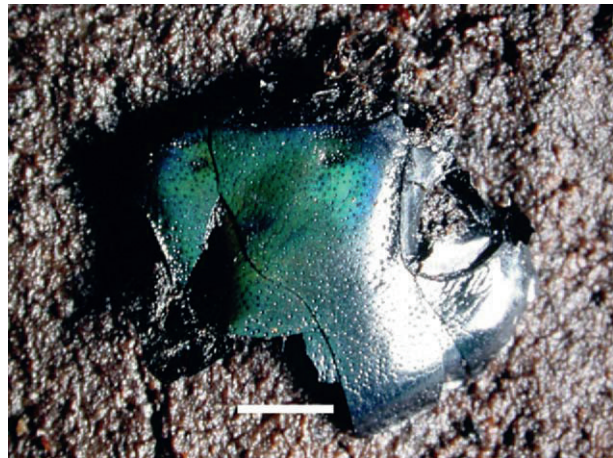


Figure 25 *Cybister brevis*, head, from the Biwajima-oki Peat Formation in the Nakamachi Site. © 2005, Photograph, FIRGNE. Scale bar = 1 mm.



Figure 26 *Hydrophilus acuminatus*, head, from the Nojiri-ko Formation in the Tategahana Site. © 2005, Photograph, FIRGNE. Scale bar = 1 mm.



Figure 27 *Platambus pictipennis*, left elytron, from the Nojiri-ko Formation in the Tategahana Site. © 2005, Photograph, FIRGNE. Scale bar = 1 mm.



Figure 28 *Pterostichus prolongatus*, pronotum, from the Biwajima-oki Peat Formation in the Nakamachi Site. © 2005, Photograph, FIRGNE. Scale bar = 1 mm.

(Figure 25), *Coelostoma*, *Anacaena*, *Enochrus*, *Hydrophilus* (Figure 26), *Sternolophus*, *Hydrochara*, *Regimbartia*, and *Dineutus* live in areas of still water with emergent and floating vegetation. *Platambus pictipennis* (Figure 27) lives in streams. This species is found in the Nojiri-ko Formation at the Tategahana Site.

The terrestrial beetles include Carabidae, Histeridae and Silphidae. Most of them live on the forest floor and/or its edge. Five species live in wet ground with vegetation, including *Carabus granulatus* (Figure 7), *Elaphrus japonicus* (Figure 6), *Pterostichus prolongatus* (Figure 28), *Epomis nigricans*, and *Chlaenius gebleri* (Figure 29). Two species, *Calosoma inquisitor* (Figure 30) and *Calleida lepida* live in trees, not on the ground.



Figure 29 *Chlaenius gebleri*, pronotum, from the Biwajima-oki Peat Formation in the Nakamachi Site. © 2005, Photograph, FIRGNE. Scale bar = 1 mm.



Figure 30 *Calosoma inquisitor*, elytron, from the Nojiri-ko Formation in the Tategahana Site. © 2005, Photograph, FIRGNE. Scale bar = 1 mm.

The plant-feeding beetles include Lucanidae, Scarabaeidae (the genera *Mimela*, *Anomala*, *Rhomborrhina* and *Eucetonia*), Cerambycidae, Chrysomelidae, Attelabidae and Curculionidae. Most of these species live in forests and/or the forest edge. Species of *Plateumaris*, *Donacia*, and *Limnobaris* are tied to aquatic host-plants, such as Cyperaceae, Potamogetonaceae, Sparganiaceae, Typhaceae, and Nymphaeaceae.

The dung-feeding beetles include *Pachysternum haemorrhoum* (Figure 31) (Hydrophilidae) and dung beetles (Scarabaeidae) in the genera *Copris* (Figure 32), *Caccobius*, *Onthophagus* and *Aphodius* (Figure 33). Two extinct mammals,



Figure 31 *Pachysternum haemorrhoum*, left elytron, from the Nojiri-ko Formation in the Tategahana Site. © 2005, Photograph, FIRGNE. Scale bar = 1 mm.



Figure 33 *Aphodius brachysomus*, left elytron, from the Nojiri-ko Formation in the Tategahana Site. © 2005, Photograph, FIRGNE. Scale bar = 1 mm.



Figure 32 *Copris pecuarius*, pronotum, from the Nojiri-ko Formation in the Tategahana Site. © 2005, Photograph, FIRGNE. Scale bar = 1 mm.

Naumann's elephant (*Palaeoloxodon naumanni*) and Yabe's giant fallow deer (*Sinomegaceros yabei*) have been excavated from the Nojiri-ko Formation in the Tategahana Site. The dung beetles from the Tategahana Site may have fed on the dung of these mammals.

Biogeography

The Japanese Islands are surrounded by sea, therefore one might expect many endemic species to be found on the islands. However, during the Pleistocene, these islands were connected

to continental Asia during intervals of lowered sea level. Modern entomologists have speculated that many recent species immigrated to the Japanese Islands during the last glaciation (e.g. Ishikawa, 1988). However, late Pleistocene fossil records indicate that the recent fauna was formed before the last glaciation.

Distributional ranges of beetles have shifted in response to environmental changes in geologic time. The transition from glacial to interglacial intervals exerted a particularly strong influence on these shifts. For example, *Copris tripartitus*, recorded from the Yasuda Formation in Niigata Prefecture, is found today on Tsushima Island, Chejido Island, Korea, and China. The fossil record indicates that *C. tripartitus* lived on Honshu during the last interglaciation, and died out there during the last glacial age. On the other hand, two dytiscid beetles, *Ilybius poppiusi* (Figure 9) and *Ilybius weymarni* (Figure 34), were recorded from several fossil assemblages from the last glaciation on eastern Honshu. These are currently found on Hokkaido and in northern Asia. The fossil records indicate that they had been living on Honshu during the last glaciation and died out there during the Holocene. *Cytilus sericeus* (Figure 19) was also recorded from several last glacial deposits on eastern Honshu. It is found today on Hokkaido and across Asia to Europe. The distribution of these species may have been influenced by the climatic change of the Pleistocene-Holocene transition.

Future of Japanese Quaternary Entomology

Japanese Quaternary entomology will continue to supply vital information on the evolution of Japanese beetle fauna. The



Figure 34 *Ilybius weymarni*, left elytron, from the Maebashi Peat Bed. © 2005, Photograph, M. Hayashi. Scale bar = 1 mm.

framework of speciation in the islands differs from that in continental regions. The formation of land bridges between island and continent or island and island are the most important element of the regional zoogeography. In the island arcs, the formation of land-bridges was caused by a combination of lowered sea level and/or uplifting of the sea floor. Japanese evolutionary scientists try to reconstruct biogeographic histories based on molecular phylogenetic trees and paleogeographic maps. However, they lack useable timescales for their theories, because the land bridge was repeatedly formed during geologic time. The fossil beetle record from Quaternary deposits provides concrete evidence that species were living at the fossil locality in times past. Therefore, evolutionary scientists are seeking information from Quaternary entomologists about the fossil records. To date, the Japanese Quaternary entomologists have insufficient data to meet these requests. However, our studies will continue, and new Quaternary entomologists will be trained.

See also: **Beetle Records: Late Pleistocene of Northern Asia; Overview.**

References

- Fossil Insect Research Group for Nojiri-ko Excavation (1980) Fossil insects from the Nojiri-ko Formation. *The Memoirs of the geological Society of Japan* 19: 147–159.
- Fossil Insect Research Group for Nojiri-ko Excavation (1984) Fossil insects obtained from the Nojiri-ko Excavations in 1978 to 1982. *Monograph of the Association for the Geological Collaboration in Japan* 27: 137–156.
- Fossil Insect Research Group for Nojiri-ko Excavation (1987) Fossil insects obtained from the Nojiri-ko Excavations in the 9th Nojiri-ko Excavation and the 4th Hill Site Excavation. *Monograph of the Association for the Geological Collaboration in Japan* 32: 117–136.
- Fossil Insect Research Group for Nojiri-ko Excavation (1988) *Handbook for Fossil Insects*. p. 129. Tokyo: New Science Co.
- Fossil Insect Research Group for Nojiri-ko Excavation (1990) Fossil insects obtained from the Nojiri-ko Formation during the 10th Nojiri-ko and the 5th Hill Site Excavations. *Monograph of the Association for the Geological Collaboration in Japan* 37: 93–110.
- Fossil Insect Research Group for Nojiri-ko Excavation (1993a) Insect fossils obtained from the Nojiri-ko Formation during the 11th Nojiri-ko Excavation. *Bulletin of the Nojiri-ko Museum* 4: 69–81.
- Fossil Insect Research Group for Nojiri-ko Excavation (1993b) Insect fossils obtained from the Nojiri-ko Formation during the 6th Hill Site Excavation. *Bulletin of the Nojiri-ko Museum* 4: 203–214.
- Fossil Insect Research Group for Nojiri-ko Excavation (1996a) Insect fossils obtained from the Nojiri-ko Formation during the 12th Nojiri-ko Excavation. *Bulletin of the Nojiri-ko Museum* 4: 71–80.
- Fossil Insect Research Group for Nojiri-ko Excavation (1996b) Insect fossils obtained from the Nojiri-ko Formation during the 7th Nojiri-ko Hill Site Excavation. *Bulletin of the Nojiri-ko Museum* 4: 199–206.
- Fossil Insect Research Group for Nojiri-ko Excavation (2000a) Insect fossils obtained from the Nojiri-ko Formation during the 13th Nojiri-ko Excavation. *Bulletin of the Nojiri-ko Museum* 8: 53–58.
- Fossil Insect Research Group for Nojiri-ko Excavation (2000b) Insect fossils obtained from the Upper Pleistocene during the 8th Nojiri-ko Hill Site Excavation. *Bulletin of the Nojiri-ko Museum* 8: 139–147.
- Fossil Insect Research Group for Nojiri-ko Excavation (2003) Insect fossils obtained from the Nojiri-ko Formation during the 14th Nojiri-ko Excavation. *Bulletin of the Nojiri-ko Museum* 11: 83–88.
- Hayashi M (1996) Fossils of a Dytiscid beetle, *Ilybius weymarni*, from Late Pleistocene Maebashi Peat in Maebashi City, central Japan. *Daiyonki-kenkyu* 35: 305–312.
- Hayashi M (1998) Upper Pleistocene fossil beetles from the Ono Peat Bed in Tatsunomachi, Nagano Prefecture, Japan. *Bulletin of the Mizunami Fossil Museum* 25: 117–125.
- Hayashi M (1999) Fossil donaciine beetles (Coleoptera: Chrysomelidae: Donaciinae) from the Pleistocene around the Haruna Volcano, Gunma Prefecture, central Japan. *Bulletin of the Mizunami Fossil Museum* 26: 149–162.
- Hayashi M (2005) Fossil Insects and reconstructed paleoenvironments of the Upper Pleistocene Maebashi Peat in Takasaki, Gunma Prefecture, Central Japan. *Bulletin of the Gunma Museum of Natural History* 9: 93–99.
- Hayashi M and Shimadu Y (2005) Stratigraphy and fossil insects of the Upper Pleistocene Tsukumogawa Peat Bed in Annaka, Gunma Prefecture, Japan. *Bulletin of the Gunma Museum of Natural History* 9: 101–107.
- Hayashi M and Miyatake Y (1996) Last Glacial insect fossils from the Oku Miomote site group, Niigata Prefecture, Japan. *Chikyū Kagaku* 50: 188–193.
- Hayashi M, Yahiro K, and Kitabayashi E (2002) Late Pleistocene insects from the Mizozono Formation in Yoshimatsu-cho, Kagoshima Prefecture, Japan. *Bulletin of the Mizunami Fossil Museum* 29: 161–168.
- Ishikawa R (1988) Origin and speciation of the Japanese Caraninae. In: Sato M (ed.) *The Japanese Coleoptera: its origin and speciation*, pp. 23–32. Tokai: Tokai University Press.
- Kashiwazaki Naumann's Elephant Research Group (1991) *Palaeoloxodon naumanni* (Makiyama) found in the upper Pleistocene, Yasuda Formation in the Kashiwazaki City, Niigata Prefecture, Central Japan, and its palaeoenvironment. *Chikyū-Kagaku* 45: 161–176.
- Mori Y (1995) Fossil insects and the paleoenvironments. In: *Reports on Owatari II Site Excavation*, pp. 351–370. Iwate: Archaeological Research Center.
- Mori Y and Itoh T (1992) Paleoenvironments of the Tomizawa Site reconstructed by fossil insects and diatoms. In: *Reports on Tomizawa Site (Paleolithic)*, pp. 126–141. Sendai City: Broad of Education.
- Mori Y, Nakamura T, and Yoshii R (1997) Insect and macroplant fossils found from the Miyanomae Site. *Abstract of the Japan Association for Quaternary Research* 27: 130–131.
- Nirei T and Hayashi M (1998) Fossil pollen and insect assemblage from Late Pleistocene peat deposits at Taguro, Tamagawa Village, western margin of Kanto Plain, Central Japan. *Bulletin of the Saitama Museum of Natural History* 16: 15–21.
- Nirei T and Hayashi M (2004) Fossil pollen assemblages of the Late Pleistocene Maebashi Peat Bed and palaeoclimatic oscillation at Takasaki, Gunma, central Japan. *The Natural Environmental Science Research* 17: 43–49.

Late Pleistocene of North America

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Beetles are the most diverse group of organisms on Earth, and their exoskeletal remains preserve well in water-logged sediments of Pleistocene age. Studies of late Pleistocene fossil beetle assemblages have been carried out throughout many regions of North America in recent decades. This article reports on the major findings of these studies in terms of paleoenvironmental reconstructions and shifting distribution patterns (biogeography) of the species found in the fossil assemblages, all of which remain extant today.

The paleotemperature estimates reported here are derived from the mutual climatic range (MCR) of paleoclimate analysis, as discussed in the overview article for this section. For the sites in eastern and central North America, the MCR method was used to provide estimates of mean temperatures of the warmest ($T_{\max}$) and coldest ($T_{\min}$) months of the year. $T_{\max}$ and $T_{\min}$ estimates are discussed in terms of departures from the modern $T_{\max}$ and $T_{\min}$ of the study sites in question.

Research in Eastern North America

Pleistocene insect fossil studies began in earnest in the 1970s, with the most intensively studied region of North America being the central and eastern United States and southeastern Canada. Several fossil beetle assemblages in eastern North America are thought to represent Marine Isotope Stage (MIS) 5e, the last interglaciation. However, the ages of these fossil assemblages are tentative. One of the richest beetle faunas in this group is from Innerkip, Ontario (Figure 1, No. 1). This fauna includes several warm-adapted species, and MCR analysis of the fauna (Table 1) yielded an estimate of $T_{\max}$ that is essentially the same as modern parameters. The fossil assemblage from the Pointe-Fortune site on the southern Ontario–Quebec border (Figure 1, No. 2) is also thought to derive from MIS 5 (Anderson et al., 1990). This fauna contains boreal species that indicate colder than modern climate (Table 1), so perhaps it reflects a late MIS 5 environment. Likewise, faunal assemblages from the Henday site in northern Manitoba (Figure 1, No. 3) and the Owl Creek site in northeastern Ontario (Figure 1, No. 4) indicate climates that were colder than today (Table 1), based on the presence of Arctic ground beetles (Carabidae).

Assemblages from MIS 4 and 3 have been described from several localities in the Great Lakes region, as summarized by Elias (1994) and Ashworth (2004). MIS 4 assemblages from southeastern Canada contain tundra and tree-line beetle faunas. Faunal assemblages containing the same species also occur in regional MIS 2 assemblages, so it appears that tundra and tree-line species inhabited the margins of the Laurentide ice sheet continually during Wisconsinan time.

A series of insect faunas indicate changing conditions during the last (Wisconsin) glaciation. To date, late Pleistocene beetle assemblages have been published from 26 sites in this

region (Table 1). These range in age from >52,000 calibrated years before present (>52 ka) to 11.5 ka. The mid-Wisconsin interstadial, MIS 3, is thought to have lasted from approximately 65–25 ka. Isotopic data from Greenland ice cores and North Atlantic oceanic cores show multiple, abrupt climate changes during marine isotope stage 3 (Groote et al., 1993). Events within the first half of this interval cannot be dated by the radiocarbon method, and organic deposits that are stratigraphically correlated to the mid-Wisconsin interval are poorly dated in many regions of North America. The analysis of fossil beetle assemblages has progressed sufficiently in North America to allow reconstructions of the timing and intensity of MIS 3 climatic change (Elias, 1999).

Paleoenvironments of Central and Eastern North America during the Last Glaciation

The oldest MIS 3 beetle fauna considered here comes from the Chaudière valley, southeastern Quebec (Figure 1, No. 5). The fauna is associated with a mid-Wisconsin cold interval. Radiocarbon ages of this formation include a date of >54 ka. The sediments that yielded the fossils were underlain by a glacial till and overlain by late Wisconsin till. The beetle fauna is indicative of arctic tundra environments and includes many cold-adapted species. $T_{\max}$ was depressed by approximately 6 °C, and $T_{\min}$ was approximately 13 °C colder than today (Table 1).

The clearest indication of the strength of interstadial warming in eastern North America comes from the series of beetle faunal assemblages from Titusville, Pennsylvania (Figure 1, No. 6). The six fossil assemblages were deposited between approximately 46 and 43 ka. Within this interval, the Titusville faunas reflect climatic amelioration, followed by cooling. The oldest fauna (46.5 ka) indicates cooling of both $T_{\max}$ and $T_{\min}$ to approximately the same level as at the Chaudière valley site (Table 1). At the height of regional warming (ca. 45.7 ka), $T_{\max}$ was only 2 °C less than modern levels, and $T_{\min}$ was approximately 6 °C colder than modern. This warm interval occurred at approximately the same time as the Upper Warren Interstadial in Britain and the warm sea-surface temperature interval between Heinrich events 4 and 5 (Elias, 1999). By the end of the interstadial event (43.2 ka), regional climates had cooled to even lower levels than previously. Thus, in the space of approximately 3000 years, regional climates oscillated from subarctic to boreal and back again.

A beetle assemblage dated 38.5 ka from St. Charles, Iowa (Figure 1, No. 8), reflects prairie or savanna environments with patches of conifer–hardwood forest (Baker et al., 1991). The St. Charles beetle fauna was indicative of somewhat milder climate than during full glacial conditions but well below the levels of warming suggested by the interstadial fauna from Titusville. Later in the MIS 3 interval (32.9–28.9 ka), fossil

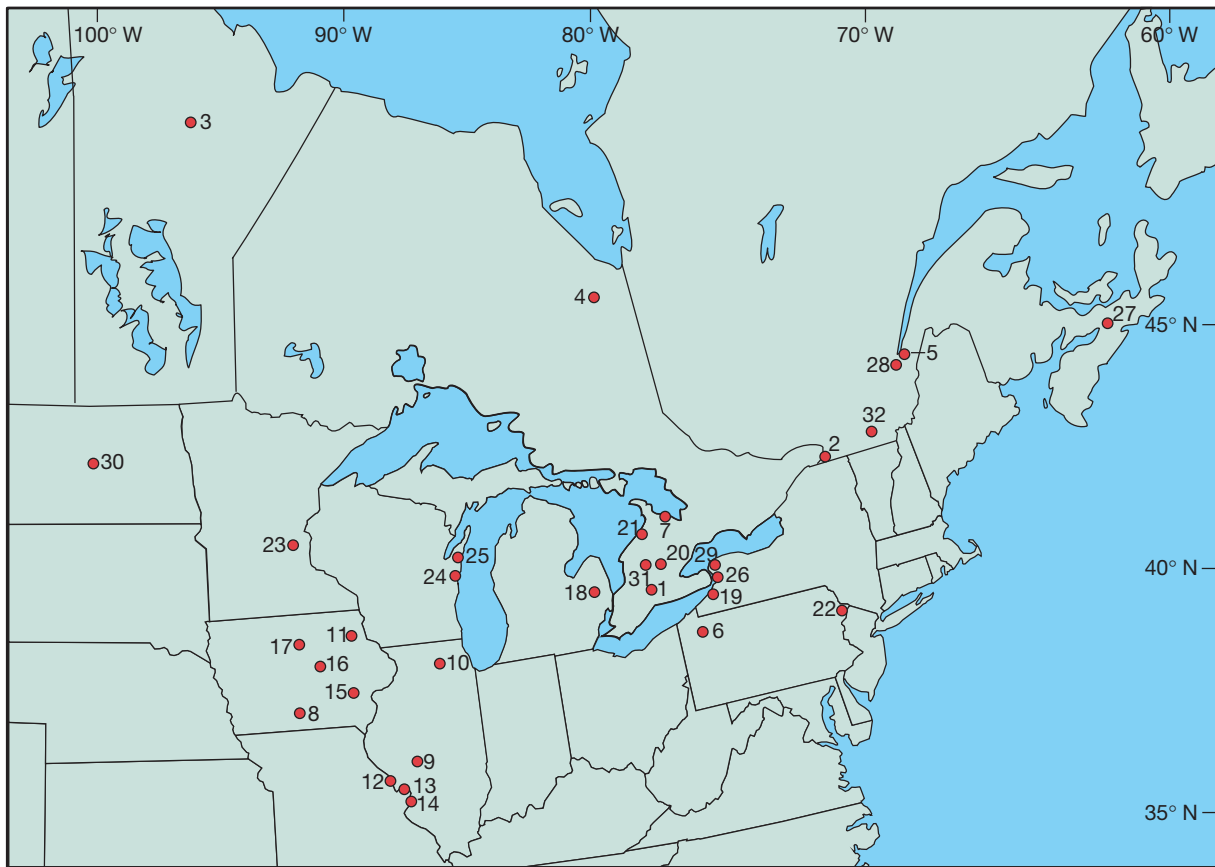


Figure 1 Map of eastern and central North America showing locations of fossil sites discussed in text. Site numbers match those in [Table 1](#).

assemblages from Illinois and Iowa indicate closed spruce forests. The sketchy picture that emerges for the early and mid-Wisconsinan is one of spatial and temporal heterogeneity in which populations of plants and insects were responding more dynamically than at any time during the Holocene ([Ashworth, 2004](#)).

By approximately 30 ka, the beetle evidence suggests that central North American climates had cooled to full-glacial levels ([Table 1](#)). $T_{\max}$ was approximately 10 °C colder than today at sites such as Athens and Wedron in Illinois ([Figure 1](#), Nos. 9 and 10). Regional faunas contained mixtures of boreal and arctic species, depending on latitude. Southerly sites, such as Salt Spring Hollow, Illinois ([Figure 1](#), No. 13), contained boreal beetle assemblages just before and during the Last Glacial Maximum (LGM). More northerly sites, such as Conklin Quarry, Iowa ([Figure 1](#), No. 15), contained beetles with Arctic and subArctic affinities.

In the mid-western region, Arctic beetles replaced the boreal forest fauna along the southern margin of the Laurentide ice sheet at approximately 25.8 ka. Colonization was probably from populations that dispersed southward in front of the growing ice sheet and from populations that dispersed westward and eastward from montane refugia in the Appalachian and Rocky Mountains, respectively ([Schwert and Ashworth, 1988](#)). From 25.8–17.7 ka, Arctic and subArctic species inhabited a discontinuous tundra zone along the margin of the ice sheet. Fossil assemblages typical of this time have been

reported from Iowa to New York ([Ashworth, 2004](#)). At the end of the glaciation, these cold-adapted faunas were forced to migrate in order to survive. Some species found refuge in high mountains, such as the Rocky Mountains in the west and the tops of the Appalachian Mountains in the east. Other species died out in the middle latitudes of the continent, and only their Beringian populations in ice-free regions of northwestern North America survived beyond the end of the Pleistocene.

The earliest sign of warming following the LGM comes from Weaver Drain, Michigan ([Figure 1](#), No. 18). This site was ice proximal at 17 ka, but MCR analysis ([Table 1](#)) shows that $T_{\max}$ had started to rise above the levels reconstructed from a 17.3-ka assemblage at Fort Dodge, Iowa ([Figure 1](#), No. 17). Ice proximal sites showed the effects of local cooling until the ice margin retreated and pockets of stranded dead ice melted, after approximately 12.6 ka.

By approximately 15.4 ka, substantial warming had occurred at Winter Gulf, New York ([Figure 1](#), No. 19). $T_{\max}$ values were less than 3 °C cooler than modern, and $T_{\min}$ values were less than 5 °C cooler than modern. This degree of warming was not seen in the oldest assemblages from Gage St., Ontario ([Figure 1](#), No. 20). However, during the interval 14.5–12.8 ka, $T_{\max}$ and $T_{\min}$ at Gage St. rose above modern levels ([Table 1](#)). This was also the case at Nichols Brook, New York ([Figure 1](#), No. 26) and Rostock, Ontario ([Figure 1](#), No. 21) between 13.3 and 11.5 ka.

Table 1 Summary of fossil beetle assemblage data from sites in central and eastern North America

Site	Age (cal yr BP × 1000)	late Pleistocene		Modern		Change in temperature		References ^a
		T _{max} (°C)	T _{min} (°C)	T _{max} (°C)	T _{min} (°C)	July ΔT (°C)	January ΔT (°C)	
1. Innerkip, Ont.	MIS 5e	16–19	–16 to –8	20.4	–6.3	–4.4 to –1.4	–9.7 to –1.7	Pilny and Morgan (1987), Churcher et al. (1990)
2. Pointe Fortune, Que	MIS 5	14–17	–22.5 to –6	19.3	–9.8	–5.3 to –2.3	–12.7 to +3.8	Anderson et al. (1990)
3. Henday, Man.	MIS 5	8.5–10	–32.5 to –27.5	15.3	–25.8	–6.8 to –5.3	–6.7 to –1.7	Nielsen et al. (1986)
4. Owl Creek, Ont.	MIS 5	13–15	–24 to –19	17.4	–17.5	–4.4 to –2.4	–6.5 to –1.5	Mott and DiLabio (1990)
5. Chaudiere Valley, Que	>52	10–10.5	–30 to –28	18	–13	–8 to –7.5	–17 to –10	Matthews and Mott (1987), Elias (1999)
6. Titusville, PA #6	46.5	12.75–13.5	–19.3 to –17.5	20.3	–5.1	–7.4 to –6.8	–14.2 to –12.4	Cong et al. (1996), Elias (1999)
6. Titusville, PA #5	45.7	15–15.5	–24 to –23	20.3	–5.1	–5.3 to –4.8	–18.9 to –17.9	Cong et al. (1996), Elias (1999)
6. Titusville, PA #4	45	18.5–19.3	–10.5 to –8.25	20.3	–5.1	–1.8 to –1	–5.4 to –3.2	Cong et al. (1996), Elias (1999)
6. Titusville, PA #3	44.3	13.5–14.5	–30.5 to –27.5	20.3	–5.1	–6.8 to –5.8	–25.4 to –22.4	Cong et al. (1996), Elias (1999)
6. Titusville, PA #2	43.7	12–13	–31 to –28	20.3	–5.1	–8.3 to –7.3	–25.9 to –17.5	Cong et al. (1996), Elias (1999)
6. Titusville, PA #1	43.2	11.8–12.5	–31.3 to –22.5	20.3	–5.1	–8.5 to –7.8	–26.2 to –17.4	Cong et al. (1996), Elias (1999)
7. Clarksburg, Ont.	>40.6	10–12	–31 to –17	19.1	–8.9	–9.1 to –7.1	–22.1 to –8.1	Warner et al. (1988), Elias (1999)
8. St. Charles, IA	40.1	17–20.5	–16.5 to –6.5	23.8	–6.9	–6.8 to –3.3	–9.6 to +0.4	Baker et al. (1991), Elias (1999)
9. Athens, IL	29.9–26.7	13–15	–19.5 to –13	24.6	–4.1	–11.6 to –9.6	–15.4 to –8.9	Morgan (1987), Elias (1999)
10. Wedron, IL	25.8	12–12.5	–30 to –26.5	23.6	–5.9	–11.6 to –11.1	–24.1 to –20.6	Garry et al. (1990), Elias (1999)
11. Elkader, IA	24.5	13–15	–27 to –22	22.6	–9.2	–9.6 to –7.6	–17.8 to –12.8	Schwert (1992), Elias (1999)
12. Salt River, MO	23.3	14.5–16	–22.3 to –15.5	22	–4.4	–7.5 to –6	–17.9 to –11.1	Schwert et al. (1997), Elias (1999)
13. Salt Spring Hollow, IL	22.5	12.8–14.8	–27.3 to –19.3	24.6	–3.5	–11.8 to –9.8	–23.8 to –15.8	Schwert et al. (1997), Elias (1999)
14. Bonfils Quarry, MO	20.8	14–16.3	–26 to –17	26.6	–1.5	–12.6 to –10.3	–24.5 to –15.5	Schwert et al. (1997), Elias (1999)
15. Conklin Quarry, IA	20.7	11.5–12.5	–29 to –27	22.3	–9	–10.8 to –9.8	–20 to –18	Baker et al. (1986), Elias et al. (1996a)
16. Saylorville, IA	19.4	12–16	–30 to –10	22.7	–9.1	–10.7 to –6.7	–20.9 to –0.9	Schwert (1992), Elias et al. (1996a)
17. Ft. Dodge, IA/I	18	12–13	–26 to –20	23.1	–9.4	–11.9 to –10.9	–16.6 to –10.6	Schwert (1992), Elias et al. (1996a)
17. Ft. Dodge, IA/II	17.3	11.5–12.5	–28.5 to –19.5	23.1	–9.4	–11.4 to –12.4	–19.1 to –10.1	Schwert (1992), Elias et al. (1996a)
18. Weaver Drain, MI	17	12.5–16.5	–29.5 to –6.5	21.1	–6.2	–8.6 to –4.6	–23.3 to –0.3	Morgan et al. (1981), Elias et al. (1996a)
19. Winter Gulf, NY/W08	15.4	17.7–20.5	–10 to –4.5	20.4	–5.2	–2.9 to +0.1	–4.8 to +0.7	Schwert and Morgan (1980), Elias et al. (1996a)
20. Gage St., Ont./G8	15.0–14.5	11.5–13.5	–25.5 to –11.5	19.9	–7.5	–8.4 to –6.4	–18 to –4	Schwert et al. (1985), Elias et al. (1996a)
19. Winter Gulf, NY/W06	14.8	16.5–21	–11.5 to –4	20.4	–5.2	–3.9 to +0.6	–8.6 to +1.2	Schwert and Morgan (1980), Elias et al. (1996a)
21. Rostock, Ont.	14.7	12–15	–28 to –20	19.4	–7.8	–7 to –4.4	–20.2 to –12.2	Pilny and Morgan (1987), Elias et al. (1996a)
22. Newton, PA	14.6	10–15	–31 to –9	19.7	–5.6	–9.7 to –4.7	–25.4 to –3.4	Barnowsky et al. (1988), Elias et al. (1996a)
23. Norwood, MN/Peat	14.5	16.5–21.5	–17.5 to –3	22.1	–12.3	–5.6 to –0.6	–5.2 to +9.3	Ashworth et al. (1981), Elias et al. (1996a)
19. Winter Gulf, NY/W04	14.5	15–21.5	–14 to 0	20.4	–5.2	–5.4 to +1.1	–9.8 to +5.2	Schwert and Morgan (1980), Elias et al. (1996a)
20. Gage St., Ont./G7	14.5–12.8	20–21	–5.5 to –2.5	19.9	–7.5	0 to +1.1	+2 to +5	Schwert et al. (1985), Elias et al. (1996a)
23. Norwood, MN/M.S.	13.8	15–20	–25 to –12	22.1	–12.3	–7.1 to –2.1	–12.7 to –0.3	Ashworth et al. (1981), Elias et al. (1996a)
21. Rostock, Ont/65–75	13.6	15–20.5	–22.5 to –4.5	19.4	–7.8	–4 to +0.6	–14.7 to +3.3	Pilny and Morgan (1987), Elias et al. (1996a)
24. Two Creeks, WI	13.6	12.5–13.5	–16.5 to –11.5	21.1	–7.3	–8.8 to –7.8	–10.2 to –4.2	Morgan and Morgan (1979), Elias et al. (1996a)

(Continued)

Table 1 (Continued)

<i>Site</i>	<i>Age (cal yr BP × 1000)</i>	<i>late Pleistocene</i>		<i>Modern</i>		<i>Change in temperature</i>		<i>References^a</i>
		<i>T_{max} (°C)</i>	<i>T_{min} (°C)</i>	<i>T_{max} (°C)</i>	<i>T_{min} (°C)</i>	<i>July ΔT (°C)</i>	<i>January ΔT (°C)</i>	
25. Kewaunee, WI	13.5	12.5–17.5	–28 to –5	20.8	–7.8	–8.3 to –3.3	–20.2 to +2.8	Garry et al. (1990), Elias et al. (1996a)
23. Norwood, MN/U.S.	13.4	17–18	–11 to –7	22.1	–12.3	–5.1 to –4.1	+1.3 to +5.3	Ashworth et al. (1981), Elias et al. (1996a)
21. Rostock, Ont/55–65	13.3	17–21	–15.5 to –2.5	19.4	–7.8	–2.4 to +1.6	–7.7 to +5.3	Pilny and Morgan (1987), Elias et al. (1996a)
27. Brookside, NS	13	12.5–20	–31.5 to 5	18.4	–7.4	–5.9 to +1.6	–24.1 to +12.4	Mott et al. (1986)
28. St. Eugene, Que	12.9	10–12.5	–31 to –18.5	17.2	–13.1	–7.2 to –4.7	–17.9 to –5.4	Mott et al. (1981)
29. Lockport Gulf, NY L01	12.8	17.5–21.5	–14 to –3	21.5	–4.8	–4 to 0	–9.2 to +1.8	Miller and Morgan (1982)
30. Johns Lake, ND	12.7	13.5–15.5	–26.5 to –20	20	–17	–6.5 to –4.5	–6.5 to 0	Ashworth and Schwert (1992)
31. Eighteen Mile R., Ont.	12.4	14–15	–16.5 to –10	19.8	–5.8	–5.8 to –4.8	–10.7 to –4.2	Ashworth (1977)
29. Lockport Gulf, NY L02	12	12.5–17.5	–25 to –5	21.5	–4.8	–9 to –4	–20.2 to +0.2	Miller and Morgan (1982)
32. St. Hilaire, Que	11.7	16.5–17.5	–15 to –10	20.7	–10.2	–4.2 to –3.2	–4.8 to +0.2	Mott et al. (1981)
26. Nichols Brook, NY–C	11.5–10.1	14.5–18	–21 to –12.5	18.1	–10	–3.5 to –0.1	–11 to –2.5	Fritz et al. (1987)

^aReferences available from the QBib Web site, www.bugs2000.org/qbib.html

Between approximately 12.9 and 11.5 ka, a third set of sites (St. Eugene and St. Hilaire, Quebec, Eighteen Mile River, Ontario, and Lockport Gulf, New York) were situated near large bodies of glacial meltwater (proglacial lakes and the Champlain Sea). The large bodies of chilled water adjacent to these sites strongly affected local climates (Ashworth, 2004). After 11.5 ka the ice sheet margin had retreated well north of the study sites, and all the fossil assemblages reflect $T_{\max}$ values near modern parameters (Table 1).

Beetle MCR reconstructions from northwest Europe show a sudden rise in seasonal temperatures by approximately 15.9 ka, followed by a general cooling through the Late Glacial interstadial, before dramatic cooling at 12.9 ka (the Younger Dryas chronozone), followed by a sudden rise in temperatures at approximately 11.5 ka (Coope and Lemdahl, 1995). The eastern North American results show almost a plateau of summer temperatures from 13–10 ka. The vegetation history, as reconstructed from fossil pollen assemblages, also suggests no climatic reversals during this interval (Grimm and Jacobson, 2004). The lack of oscillations in the North American assemblages may be due to the effects of continental climate, whereas Late Glacial climates in northwest Europe were strongly affected by shifts in North Atlantic surface water (Coope and Lemdahl, 1995). The only insect evidence for cooling between 12.9 and 11.5 ka in North America comes from the maritime provinces of Canada (Walker, et al., 1991) and from Arctic Alaska.

Research in Western North America

Late Pleistocene beetle assemblages have also been studied from a number of sites in western North America (Table 2). Insect fossils of this age from the Rocky Mountain region are discussed by Elias (1990, 1991). Full glacial age faunas, such as the assemblage dated 17.7 ka from Lamb Spring, Colorado (Figure 2, No. 1), indicate that $T_{\max}$ was 10–11 °C colder than today. The earliest evidence for the start of postglacial warming is a 16.9-ka assemblage at the Mary Jane site in Colorado (Figure 2, No. 2). Rapid warming occurred after 13.3 ka, with summer temperatures becoming warmer than present by 11.5 ka at False Cougar Cave, Montana (Figure 2, No. 4). Many of the cold-adapted species in the late Pleistocene Rocky Mountain fossil assemblages have since retreated upslope to the Alpine tundra, but not all species were able to do this. For instance, the cold-adapted rove beetle, *Holoboreaphilus nordenskiöldi*, was found in a Late Glacial age assemblage from Marias Pass, Montana (Figure 2, No. 3). This species apparently died out in the Mid-Latitudes at the end of the last glaciation. It is found today only in Arctic North America (Elias, 1991).

In the past 20 years, insect fossils from arid regions of the southwestern United States and northern Mexico have received intensive study. These studies are based on fossil remains extracted from packrat (*Neotoma*) middens. Packrat middens are caches of objects, including edible plants, cactus spines, insect and vertebrate remains, small pebbles and feces, brought to the rat's den site for a variety of reasons, including food, curiosity, and den protection; then they are cemented into black tarry masses by packrat urine. Late Pleistocene insect fossils have been studied from the Great Basin, but the most

intensively studied region is the Chihuahuan Desert (Figure 3). Most packrat midden beetle assemblages from the Sonoran Desert date to the Holocene.

Late Pleistocene faunas from the arid southwest were mixtures of temperate and desert species not seen in any one region today. Since the end of the Pleistocene, some of these species have become established in different desert regions (Elias, 1992). Others now live outside the desert zone, in temperate regions. The fossil insect record indicates that even sedentary, flightless beetles (e.g., the heavy-bodied weevils) have undergone marked distributional shifts in the American southwest within the space of a few centuries. Moreover, even highly specialized cave dwellers have somehow managed to move from one cave system to another in response to changes in late Wisconsin and Holocene environments.

Late Pleistocene beetle assemblages from the southern Chihuahuan Desert of Mexico (Figure 2, Nos. 21 and 22) contain mixtures of desert and temperate zone species (Elias et al., 1995). Midden assemblages from locations farther north in the Chihuahuan Desert are generally separated into glacial age faunas with temperate zone affinities and postglacial faunas with desert zone affinities. Faunal assemblages with no modern analog are indicative of the late Quaternary environments unlike any that exist today. This conclusion is also borne out by the plant macrofossil record.

The insect faunas from the Big Bend region of southwest Texas (Figure 2, Nos. 8, 10, 12, 17, and 19) suggest that there was greater effective moisture from 35–14 ka (Elias and Van Devender, 1990). During the late Wisconsin (MIS 2), many temperate grassland species lived in the Big Bend region. After 14 ka, most of these species were replaced by either desert species or more cosmopolitan taxa. The faunal change suggests a climatic shift from cool, moist conditions of full glacial times to hotter, drier conditions during the Late Glacial and early Holocene.

In the northern Chihuahuan Desert, full glacial (26.5–21.6 ka) faunal assemblages (Figure 2, Nos. 5, 6, 14–16, 23, and 24) suggest widespread coniferous woodland at elevations as low as 1200–1400 m above sea level (Elias and Van Devender, 1992). These woodland environments persisted until 12.9 ka, but the insect data suggest considerable open ground, with grasses at least locally important at the midden sites. The grassland nature of the arthropod fauna was also suggested in the regional vertebrate record. The transition from the temperate Wisconsin fauna to the more xeric postglacial fauna started by 14.8 ka. The timing of this faunal change was essentially synchronous throughout the region. A major difference between the Big Bend and northern Chihuahuan Desert scenarios is the nature of this faunal change. In the Big Bend region, the transition was characterized by the disappearance from the record of all but one of the temperate insect species by approximately 14 ka. However, the xeric-adapted fauna did not appear in the Big Bend records until approximately 8.3 ka. In the northern Chihuahuan Desert assemblages, the xeric species first appeared at 14.8 ka, and several of the temperate grassland species from the Wisconsin interval persisted well into the Holocene. This mixture of xeric and temperate elements makes ecological sense because these northern faunas were living close to the edge of the Chihuahuan Desert. The gradual shifting of northern desert boundaries in postglacial times

Table 2 Summary of fossil beetle assemblage data from sites in western North America

Site	Age (cal yr BP $\times$ 1000)	Late quaternary		Modern		Change in temperature		References ^a
		T _{max} (°C)	T _{min} (°C)	T _{max} (°C)	T _{min} (°C)	July Δ T (°C)	January Δ T (°C)	
1. Lamb Spring, CO	17.7	10–11	–31 to –27	21.4	–1.3	–11.4 to –10.4	–29.7 to –25.7	Elias and Toolin, 1990; Elias, 1996
2. Mary Jane, CO	16.2	9.8–10.2	–29.3 to –27.6	13.4	–8.6	–3.6 to –3.2	–20.7 to –19	Short and Elias, 1987; Elias, 1996
2. Mary Jane, CO	15.4	10–10.2	–29.1 to –27.6	13.4	–8.6	–3.4 to –3.2	–20.5 to –19	Short and Elias, 1987; Elias, 1996
3. Marias Pass, MT	14.2	11–14	–30 to –16	13.9	–9	–2.9 to +0.1	–21 to –7	Elias, 1988, 1996
4. False Cougar Cave, MT	11.5	15.5–17.7	–4.5 to +0.25	13.5	–10.1	+2 to +4.25	+5.6 to +10.4	Elias, 1990, 1996
<i>Chihuahuan Desert and Great Basin packrat midden faunas</i>								
5. S. Organ Mountains, NM	>46.3–41.2	NA	NA	NA	NA	NA	NA	Elias and Van Devender (1992)
6. Hueco Mountains, TX	44.9–12.4	NA	NA	NA	NA	NA	NA	Elias and Van Devender (1992)
7. Kaetan Cave, AZ	35.8–17.5	NA	NA	NA	NA	NA	NA	Elias et al. (1992)
8. Maravillas Canyon, TX	32.7–13.8	NA	NA	NA	NA	NA	NA	Elias and Van Devender (1990)
9. Salt Creek Canyon, UT	32.4	NA	NA	NA	NA	NA	NA	Elias et al. (1992)
10. Baby Vulture Den, TX	31–17.5	NA	NA	NA	NA	NA	NA	Elias and Van Devender (1990)
11. Bida Cave, AZ	29.2–19.5	NA	NA	NA	NA	NA	NA	Elias et al. (1992)
12. Tunnel View, TX	28.9–12.3	NA	NA	NA	NA	NA	NA	Elias and Van Devender (1990)
13. Owl Roost, AZ	25.7	NA	NA	NA	NA	NA	NA	Elias et al. (1992)
14. Sacramento Mountains, NM	22.1–12.4	NA	NA	NA	NA	NA	NA	Elias and Van Devender (1992)
15. Bennett Ranch, TX	21.9–12.9	NA	NA	NA	NA	NA	NA	Elias and Van Devender (1992)
16. Streeruwitz Hills, TX	21.7–12.7	NA	NA	NA	NA	NA	NA	Elias and Van Devender (1992)
17. Ernst Tinaja, TX	20.6–13.3	NA	NA	NA	NA	NA	NA	Elias and Van Devender (1990)
18. Shafter, TX	19.2–13.3	NA	NA	NA	NA	NA	NA	Elias and Van Devender (1992)
19. Terlingua, TX	18.3	NA	NA	NA	NA	NA	NA	Elias and Van Devender (1990)
20. Bechan Cave, UT	16.2–13.4	NA	NA	NA	NA	NA	NA	Elias et al. (1992)
21. Puerto de Ventanillas, Coahuila, Mexico	16.2–15.2	NA	NA	NA	NA	NA	NA	Elias et al. (1993)
22. Cañon de la Fragua, Coahuila, Mexico	16.1–14.5	NA	NA	NA	NA	NA	NA	Elias et al. (1993)
23. Guadalupe Mountains, TX	15.9–13.8	NA	NA	NA	NA	NA	NA	Elias and Van Devender (1992)
24. Bonneville Estates, UT	13.6	17.5–19	NA	23.9	NA	–6.4 to –4.9	NA	Rhode (2000)
25. Guadalupe Mountains, NM	12.4–11.6	NA	NA	NA	NA	NA	NA	Elias and Van Devender (1992)
<i>Pacific regional sites</i>								
26. Kalaloch, WA	24.1–20.5	14–17	–19 to –11	18.9	5.2	–4.9 to –1.9	–24.2 to –16.2	Cong and Ashworth (1986)
27. Mary Hill, BC	22.6–22.5	11–18	–29 to –4	18.1	4.8	–7.1 to –0.1	–24.2 to –8.8	Miller et al. (1985)
28. Port Moody, BC	22.1	11–18	–29 to –4	18.1	4.8	–7.1 to –0.1	–24.2 to –8.8	Miller et al. (1985)
29. Discovery Park, WA	18	15–17	–16 to –10	18.1	4.8	–3.1 to –1.1	–24.8 to –14.8	Nelson and Coope (1982)
30. McKittrick, CA	14.3	NA	NA	NA	NA	NA	NA	Miller (1983)

^aReferences available from the QBib Web site, www.bugs2000.org/qbib.html

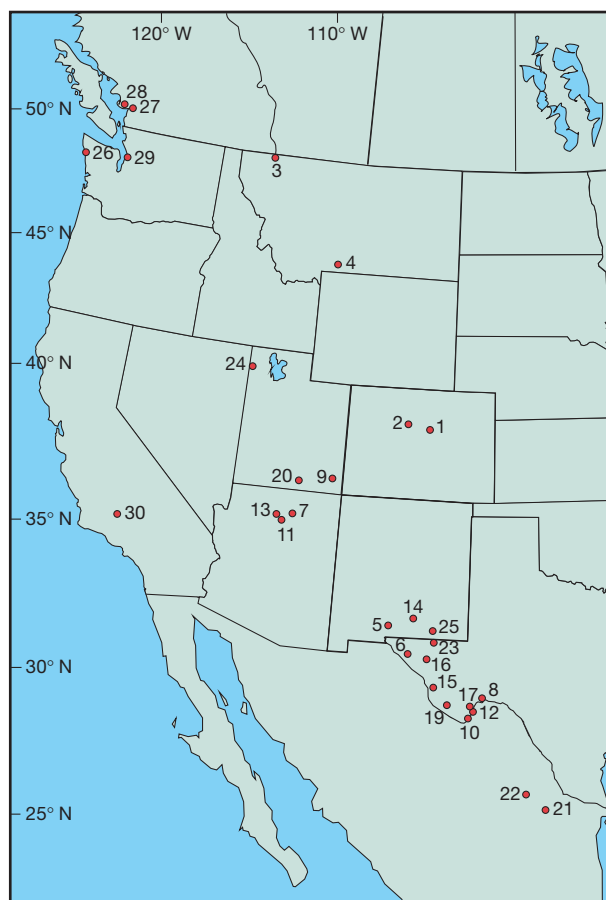


Figure 2 Map of western North America showing locations of fossil sites discussed in text. Site numbers match those in [Table 2](#).

probably created many marginal habitats for temperate species in ecotones between grassland and desert scrub.

Work on late Quaternary insect faunas from the Colorado Plateau region ([Figure 2](#), Nos 7, 9, 11, 13, and 20) suggests that late Wisconsin climatic conditions there were cooler and moister than present, and that the plateau supported a mosaic of grassland and shrub communities without modern analog ([Elias et al., 1992](#)).

MCR analysis of arid southwestern faunas ([Elias, 1998](#)) suggests that effective moisture exerts greater control on desert beetle distributions than does temperature. The author's MCR analysis of late Pleistocene beetle assemblages from the Mojave Desert ([Elias](#), unpublished data) indicates that $T_{\max}$ was 7–8 °C cooler than it is today from 38–36 ka, and mean annual precipitation (MAP) was 3.5–4.4 times greater than it is today. In the eastern Great Basin, the Bonneville Estates site yielded a beetle fauna dated 13.6 ka. Several of the species in this fauna are found today in the Pacific Northwest region. MCR analysis of this fauna indicated that $T_{\max}$ was 5–6 °C cooler than today, and MAP was 525–675 mm, compared to modern MAP at the site of approximately 190–210 mm. These results agree well with paleoclimate reconstructions based on ancient pluvial lake levels ([Benson, 2004](#)) and regional paleobotany ([Thompson et al., 2004](#)).

[Miller \(1983\)](#) discussed insect fossil assemblages from California asphalt deposits at McKittrick ([Figure 2](#), No. 29)

and at Rancho La Brea in Los Angeles. The Rancho La Brea assemblages remain undated; the late Pleistocene McKittrick fauna dates to approximately 14.3 ka. Unlike the faunas of most other regions discussed here, the McKittrick and Rancho La Brea assemblages reflect conditions similar to those in southern California today.

[Nelson and Coope \(1982\)](#) described a diverse insect fauna dating from 18 ka from Seattle, Washington ([Figure 2](#), No. 28). Although the pollen spectra associated with this assemblage suggest conditions substantially colder than present, the insects are characteristic of the modern Puget lowland. Nelson and Coope suggest that the discrepancy between the flora and insect fauna may be due to increased climatic continentality just before the last glacial advance. MCR analysis of this assemblage ([Table 2](#)) suggests that $T_{\max}$ was 1–3 °C colder than modern at the site, and that $T_{\min}$ was considerably colder than modern.

[Cong and Ashworth \(1996\)](#) described a beetle fauna from the Kalaloch site in northwest Washington ([Figure 2](#), No. 25). This fauna dates from 24.1–20.5 ka, and MCR analysis indicates that summer temperatures were only slightly cooler than today ([Table 2](#)). Most of this fauna can be found today in the lowlands of the Pacific Northwest, but a few species are restricted today to the northern boreal forest regions ([Ashworth, 2004](#)).

In southern British Columbia, [Miller et al. \(1985\)](#) studied insect fossils from 22.6–22.1 ka at Mary Hill and Port Moody

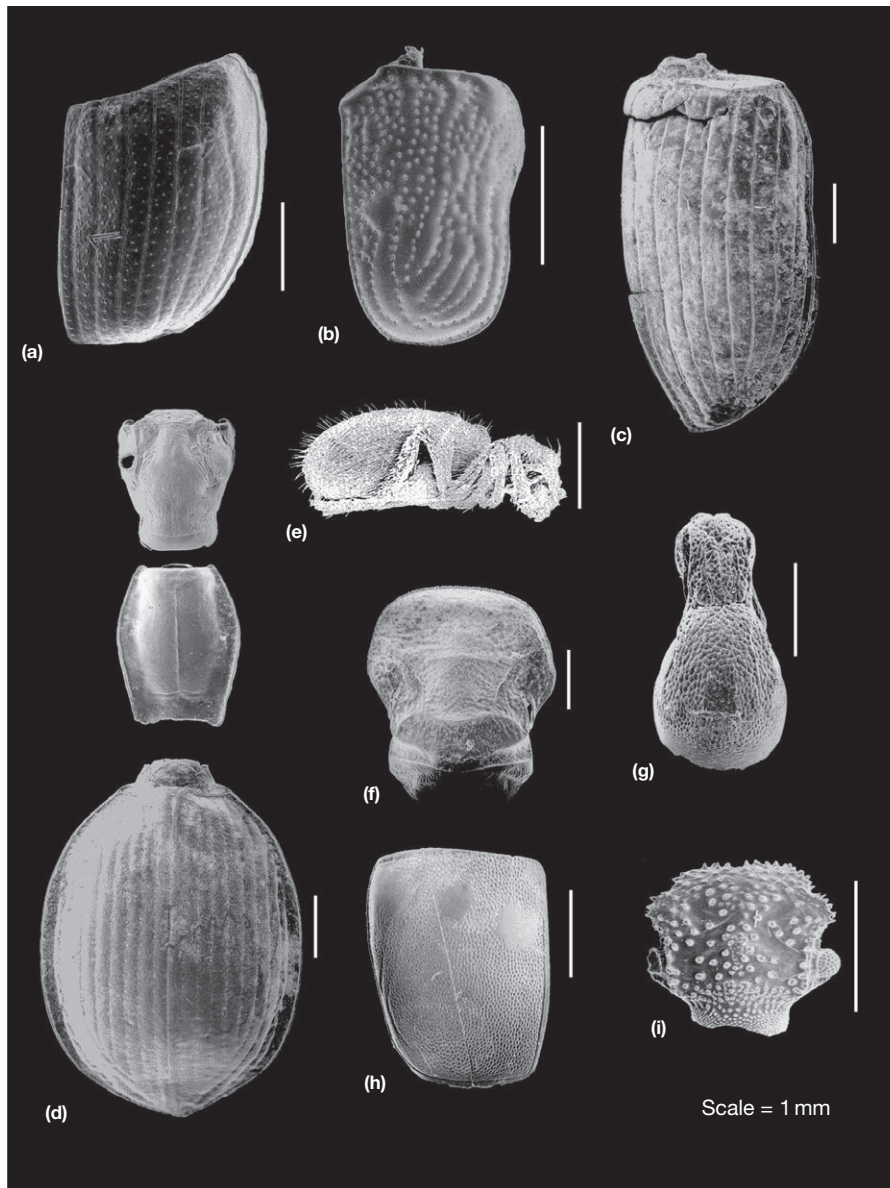


Figure 3 Scanning electron micrographs of late Pleistocene fossil beetle specimens from the Chihuahuan Desert. (a) Elytron of *Onthophagus lecontei*; (b) elytron of *Pachybrachis mitis*; (c) elytron of *Piosoma setosum*; (d) head, pronotum, and elytra of *Rhadine longicollis*; (e) intact exoskeleton of *Ptinus*; (f) head of *Onthophagus cochisus*; (g) head of *Sapotes longipillis*; (h) elytron of *Hypocaccus estriatus*; (i) head of *Rhagoderma costata*.

(Figure 2, Nos. 26 and 27). These assemblages represent an open coniferous forest floor community, developed in cool, dry climatic conditions between two advances of the Cordilleran ice sheet.

Throughout western North America, both temperature and precipitation regimes changed dramatically throughout the late Pleistocene. These climatic changes brought about wholesale changes in regional ecosystems, as reflected in the fossil beetle record. Unlike eastern North America, the western regions are topographically diverse, with dozens of mountain ranges interspersed with valleys, canyons, and flatlands. Cold-adapted beetles dominated many late Pleistocene faunal assemblages, expanding their ranges downslope and southward

in response to climatic cooling, then retreating upslope and northward in response to warming.

Research in Eastern Beringia

During the late Pleistocene, eastern Beringia comprised the unglaciated regions of Alaska and the Yukon Territory. This region was an important refuge for arctic biota, including insects. Because of preservation in permafrost, the fossil beetle record of this region extends back into the late Tertiary. The nature of Pleistocene environments in eastern Beringia is a topic of considerable debate. Some patterns concerning

Table 3 Summary of fossil beetle assemblage data from sites in eastern Beringia

Site	Age (cal yr BP × 1000)	Late quaternary		Modern		Change in temperature		References ^a
		T _{max} (°C)	T _{min} (°C)	T _{max} (°C)	T _{min} (°C)	July ΔT (°C)	Jan ΔT (°C)	
1. Old Crow CRH15 78–91	MIS 6	8.75–10	–29 to –31.75	16	–29	–7.25 to –6	–2.75 to 0	Morlan and Matthews (1983)
2. Chi'jee's Bluff Stn 1A, 4	MIS 6	12.5–13.25	–20.5 to –18.5	16	–29	–3.5 to –2.75	+8.5 to +10.5	Matthews et al. (1990a)
3. Kulukbuk Bluff 25A	MIS 6	7.25–9.25	–32.75 to –28.75	14.8	–22.6	–7.55 to –5.55	–10.15 to –6.15	Elias (1992a)
4. Noatak NK–26 Unit 4	140	3.75–10.5	–17.5 to –36.25	11	–25.8	–7.25 to –0.5	–10.45 to +8.3	Elias et al. (1998a)
5. Noatak NK29a unit 3, lower	140	5–9	–29 to –21.5	11	–29	–6.0 to –2.0	0 to +7.5	Elias et al. (1998a)
2. Chi'jee's Bluff Stn 4, I&J	140	3.5–9.5	–36 to –21	16	–29	–12.5 to –5.5	–7.0 to +8.0	Matthews et al. (1990a)
2. Chi'jee's Bluff 87–2	MIS 6–5 transition	11.75–12.5	–28 to –27.5	16	–29	–4.25 to –3.5	+1.0 to +1.5	Matthews et al. (1990a)
5. Noatak 29A unit 3, upper	MIS 6–5 transition	12–12.5	–29 to –27	11	–25.8	+1.0 to +1.5	–3.2 to –1.2	Elias et al. (1998a)
6. Noatak 37 MEE 0.13–0.22	MIS 6–5 transition	9–11	–25.5 to –22	11	–25.8	–2.0 to 0	+0.3 to +3.8	Edwards et al. (2003)
7. Nuyakuk 13-1	MIS 6–5 transition	12.5–13	–29.25 to –28	12.7	–9.7*	–0.2 to +0.3	NA	Elias and Short (1992)
6. Noatak 37 MEE2 140–153	MIS 5e	12.5–15	–28.5 to –19	11	–25.8	+1.5 to +4	–2.7 to +6.8	Edwards et al. (2002)
6. Noatak 37 MEE2 0.97–1.06	MIS 5e	12–16	–27 to –16	11	–25.8	+1 to +5	–1.2 to +9.8	Edwards et al. (2002)
8. Eva Creek 3-1A	MIS 5e	12.5–13	–22.5 to –20	16.9	–23.4	–4.5 to –3.9	+0.9 to +3.4	Matthews (1968)
3. Kulukbuk Bluff 25Z	MIS 5e	11.75–12	–22.5 to –20.5	14.8	–22.6	–3.05 to –2.8	+0.1 to +2.1	Elias (1992a)
2. Chi'jee's Bluff Stn 4, N-A	MIS 5e	12–13.25	–23 to –20.25	16	–29	–4.0 to –2.75	+6.0 to +8.75	Matthews et al. (1990a)
2. Chi'jee's Bluff Stn 4, N-B	MIS 5e	14.5–15.5	–25.25 to –21.5	16	–29	–1.5 to –0.5	+3.75 to +7.5	Matthews et al. (1990a)
7. Nuyakuk 13-2	MIS 5e	13–16.25	–30 to –21.75	12.7	–9.7*	+0.3 to +3.55	NA	Elias and Short (1992)
7. Nuyakuk 13-4	Late MIS 5	12.5–12.75	–12 to –11.25	12.7	–9.7*	–0.2 to +0.05	NA	Elias and Short (1992)
7. Nuyakuk 13-7	Late MIS 5	10.5–12	–20 to –19.5	12.7	–9.7*	–2.2 to –0.7	NA	Elias and Short (1992)
6. Noatak 37 94-25	Late MIS 5	7.5–9	–26.5 to –22.5	11	–25.8	–3.5 to –2	–0.7 to +3.3	Elias (1997)
9. Hungry Creek 52&53	Early MIS 4	6.5–10.25	–34.5 to –24.75	16	–29	–9.5 to –5.75	–5.5 to +4.25	Hughes et al. (1981)
10. Igushik 9-M-2	Early MIS 4	9–10	–30 to –23	12.7	–9.7*	–3.7 to –2.7	NA	Lea et al. (1991)
7. Chi'jee's Bluff Sample A	52 K	8.25–10	–32.75 to –27.25	16	–29	–7.25 to –6	–3.75 to +1.75	Matthews and Telka (1997)
9. Hungry Creek 27	Early MIS 3	11.5–12.5	–30 to –28	16	–29	–4.5 to –3.5	–1.0 to +1.0	Hughes et al. (1981)
10. Igushik 1-4	Early MIS 3	8.5–10	–30.5 to –22	12.7	–9.7*	–4.2 to –2.7	NA	Lea et al. (1991)
10. Igushik 4-M-1	Early MIS 3	12.5–15.5	–28 to –19	12.7	–9.7*	–0.2 to +2.8	NA	Lea et al. (1991)
11. Kuskokwim KBR/24	Early MIS 3	11–12.75	–24.25 to –19.75	14.8	–22.6	–3.8 to –2.05	–1.65 to +2.85	Elias (1992a)
11. Kulukbuk Bluff Unit C, A	Early MIS 3	9.75–11.5	–31.25 to –26.5	14.8	–22.6	–5.05 to –3.3	–8.65 to –3.9	Elias (1992a)
9. Hungry Creek 35	Mid-MIS 3	10–12.5	–19.5 to –13	16	–29	–6 to –3.5	+9.5 to +16.0	Hughes et al. (1981)
1. Old Crow 77-51	MIS 3	14.5–15.5	–28.5 to –25	16	–29.0	–1.5 to –0.5	+0.5 to +4.0	Morlan and Matthews (1983)
12. Titaluk River 17A	46.4	9.5–10	–31 to –28	10.3	–25.2	–0.8 to –0.3	–5.8 to –2.8	Elias (1997)
13. Cutler River	46.3	8–9.5	–26 to –22	11	–25.8	–3.0 to –1.5	–0.2 to +3.8	Elias (1997)
14. Cape Deceit Peat 5	43.3	6–10.5	–35.5 to –18	12.1	–20.4	–6.1 to –1.6	–15.1 to +2.4	Matthews (1968)
15. Upper Porcupine River	42.2	9.5–10	–29 to –27.5	16	–29	–6.5 to –6.0	0 to +1.5	Matthews and Telka (1997)
14. Cape Deceit S-5	42.2	11–11.75	–20.25 to –19.75	12.1	–20.4	–1.1 to –0.35	+0.15 to +0.65	Matthews (1968)
16. Blue Babe Site	41.8	6–10.5	–32.5 to –18.5	16	–24.4	–10 to –5.5	–8.1 to +5.9	Guthrie (1990)
8. Eva Creek 3-1A	38.5	11–13.5	–24 to –17	16.9	–23.4	–5.9 to –3.4	–0.6 to +6.4	Matthews (1968)
7. Chi'jee's Bluff Stn 1, B	37.8	10–11	–27.5 to –23	16	–29	–6.0 to –5.0	+1.5 to +6.0	Matthews et al. (1990a)
17. Isabella Basin IS-69	36.6	12–14.25	–32 to –22.25	16.9	–23.4	–4.9 to –2.65	–8.6 to +1.15	Matthews (1974)
12. Titaluk River 7D15	36	8–9.5	–31.5 to –25.5	10.3	–25.2	–2.3 to –0.8	–6.3 to –0.3	Nelson and Carter (1987)

(Continued)

Table 3 (Continued)

Site	Age (cal yr BP $\times$ 1000)	Late quaternary		Modern		Change in temperature		References ^a
		T _{max} (°C)	T _{min} (°C)	T _{max} (°C)	T _{min} (°C)	July Δ T (°C)	Jan Δ T (°C)	
8. Eva Creek 3-3B	35.2	8.5–9.5	–31.5 to –26	16.9	–23.4	–8.4 to –7.4	–8.1 to –2.6	Matthews (1968)
11. Kulukbuk Bluffs Unit C, B	35	12.25–13	–30.75 to –28	14.8	–27.6	–2.55 to –1.8	–3.15 to –0.4	Elias (1992a)
12. Titaluk River 20A	33.6	10.75–12.25	–20 to –19.25	10.3	–25.2	+0.45 to +1.95	+5.2 to +5.95	Nelson and Carter (1987)
18. Mayo Village Unit 1	31.5	9–9.75	–29.75 to –28	15.2	–29	–6.2 to –5.45	–0.75 to +1.0	Matthews et al. (1990b)
19. Baldwin Peninsula	31.5	8.5–9.5	–31 to –28	12.1	–18.3*	–3.6 to –2.6	NA	Hopkins et al. (1976)
20. Rock River	30.1	9.5–10.5	–28 to –22.5	16	–29	–6.5 to –5.5	+1.0 to +6.5	Matthews and Telka (1997)
9. Hungry Creek 49	26.1	11–12.25	–29.5 to –25.5	16	–29	–5.0 to –3.75	–0.5 to +3.5	Hughes et al. (1981)
10. Bering Shelf 78-15	24.9	5.5–10.5	–34.5 to –18.75	10.8	–15.6*	–5.3 to –0.3	NA	Elias et al. (1996, 1998b)
21. Bluefish	24.1	8.5–9.5	–31.5 to –28	16	–27.8	–7.5 to –6.5	–3.7 to –0.2	Matthews and Telka (1997)
22. Bering Land Bridge Park	21.7	5.5–9.5	–34 to –21	8.4	–18.3*	–2.5 to +1.1	NA	Elias (1997)
14. Cape Deceit S-1	ca. 21.6	9.5–10.5	–25 to –22	16	–20.4*	–6.5 to –5.5	NA	Matthews (1968)
23. Colorado Creek	18.2	11.25–12.5	–21.25 to –18.25	14.8	–22.6	–3.55 to –2.3	+1.35 to +4.35	Elias (1992b)
24. Clarence Lagoon	15.3	9.25–15.25	–35.25 to –19.25	7.5	–26	+1.75 to +7.75	–9.25 to +6.75	Matthews (1975)
25. Flounder Flat 3-M-8	15	12.5–13	–30.5 to –28	12.7	–9.4*	–0.2 to +0.3	NA	Elias (1992b)
26. Bering Shelf 76–121	13.4	11.5–13	–28.5 to –23.5	10.8	–15.6*	+0.7 to +2.2	NA	Elias et al. (1996, 1998b)
6. Noatak NK35-95AHA-60	13.2	15–16	–26 to –20.5	11	–25.8	+4.0 to +5.0	–0.2 to +5.3	Elias (1997)
27. Chukchi Shelf 85-69	12.9	9.25–10.25	–32.25 to –27.25	4	–27.7*	+5.25 to +6.25	NA	Elias et al. (1996, 1998b)
6. Noatak NK35-94AHA-24	12.3	12.5–13.5	–22 to –18	11	–25.8	+1.5 to +2.5	+3.8 to +7.8	Elias (1997)
6. Noatak NK32A-94AHA-17	12.3	10–10.5	–30 to –28.5	11	–25.8	–1.0 to –0.5	–4.2 to –2.7	Elias (1997)
28. Eagle River	11.5	12–13.5	–31.25 to –27	14.9	–30.5	–2.9 to –1.4	–0.75 to +3.5	Matthews and Telka (1997)

^aReferences available from the QBib Web site, www.bugs2000.org/qbib.html

*Coastal sites with unreliable reconstructions of TMIN. See text for discussion.

Beringian paleoenvironments are beginning to emerge, but our reconstructions are far from complete. When I reported on the state of Pleistocene insect fossil research in this region in 1994, I noted that ‘an enormous amount of work is yet to be done.’ Although our knowledge has increased in the past decade, the statement remains equally true today. The following discussion deals with published fossil data from 28 sites (Table 3, Figure 4). The evidence has been reviewed most recently by Elias (2000, 2001) and Ashworth (2004).

Based on their stratigraphic position, a number of fossil beetle assemblages can be confidently placed in MIS 6 (Table 3), including assemblages from the Old Crow site (Figure 4, No. 1), Chi’jee’s Bluff (Figure 4, No. 2), and Kulukbuk Bluffs (Figure 4, No. 3). These faunas reflect climates with depressed summer temperatures, but some sites yielded MCR estimates of $T_{\min}$ that are near modern levels. Based on stratigraphic relationships with sediments containing the Old Crow tephra (dated 140 ka), several beetle assemblages from Alaska and the Yukon indicate that prior to the onset

of full interglacial conditions in MIS 5e, temperatures were beginning to warm. This evidence comes from several sites on the Noatak River in northwestern Alaska (Figure 4, Nos. 4–6) and the Nuyakuk site in southwestern Alaska (Figure 4, No. 7). Full interglacial warming peaked at levels that varied from region to region. In southwestern Alaska, the height of MIS 5e amelioration was as much as 3.5 °C warmer than modern at the Nuyakuk site. At the NK-37 site in northwestern Alaska, $T_{\max}$ climbed to as much as 4.5 °C above modern levels. Further east, summer temperatures were probably closer to modern levels. The best constrained estimate of average winter temperatures during MIS 5e comes from Chi’jee’s Bluff. This estimate suggests that $T_{\min}$ was approximately 4–7 °C warmer than today, even though average summer temperatures at this site may have been slightly cooler than modern. Boreal insects and paleobotanical evidence combine to document the presence of coniferous forests in much of eastern Beringia during MIS 5e.

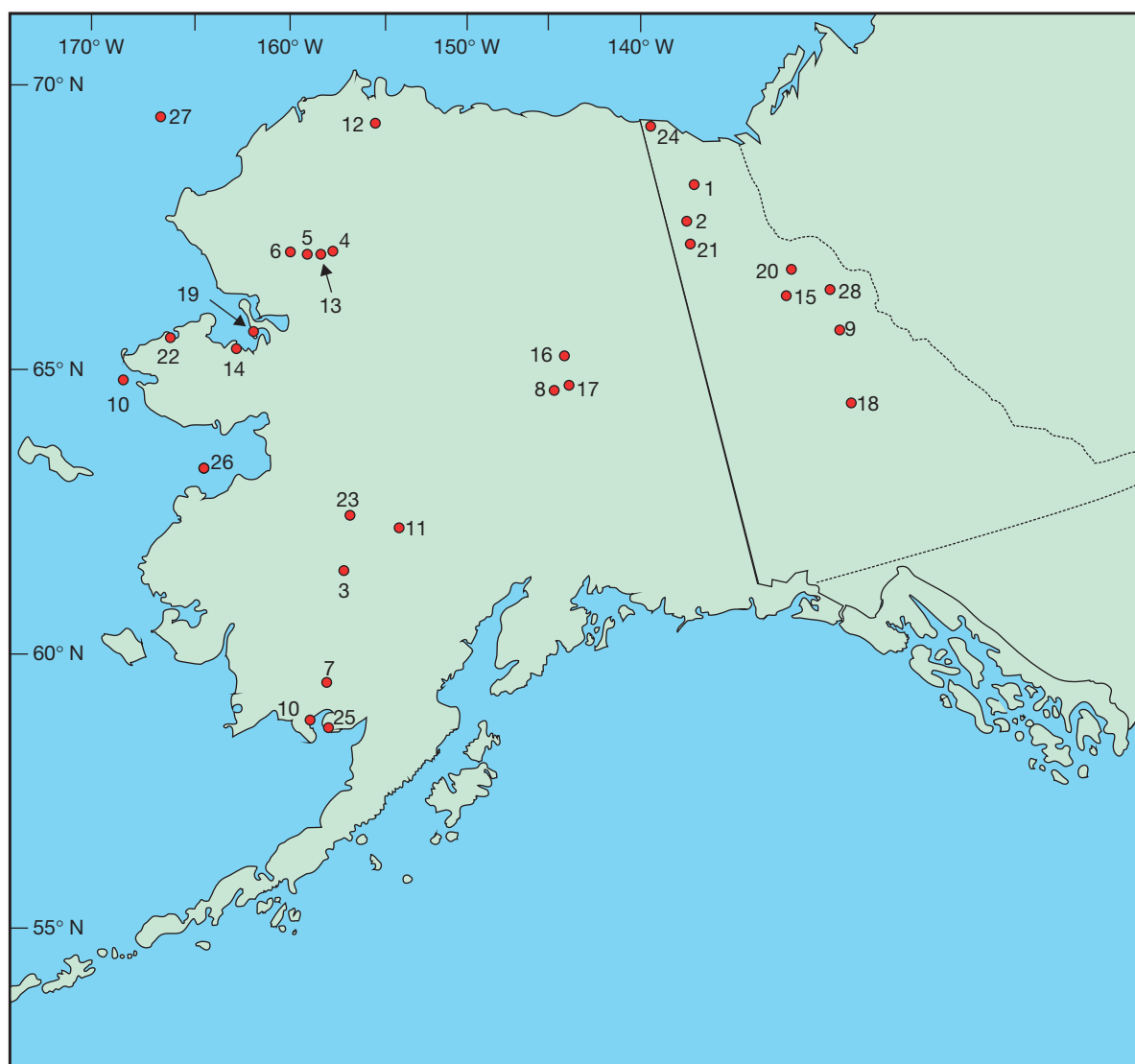


Figure 4 Map of northwestern North America showing locations of fossil sites discussed in text. Site numbers match those in Table 3.

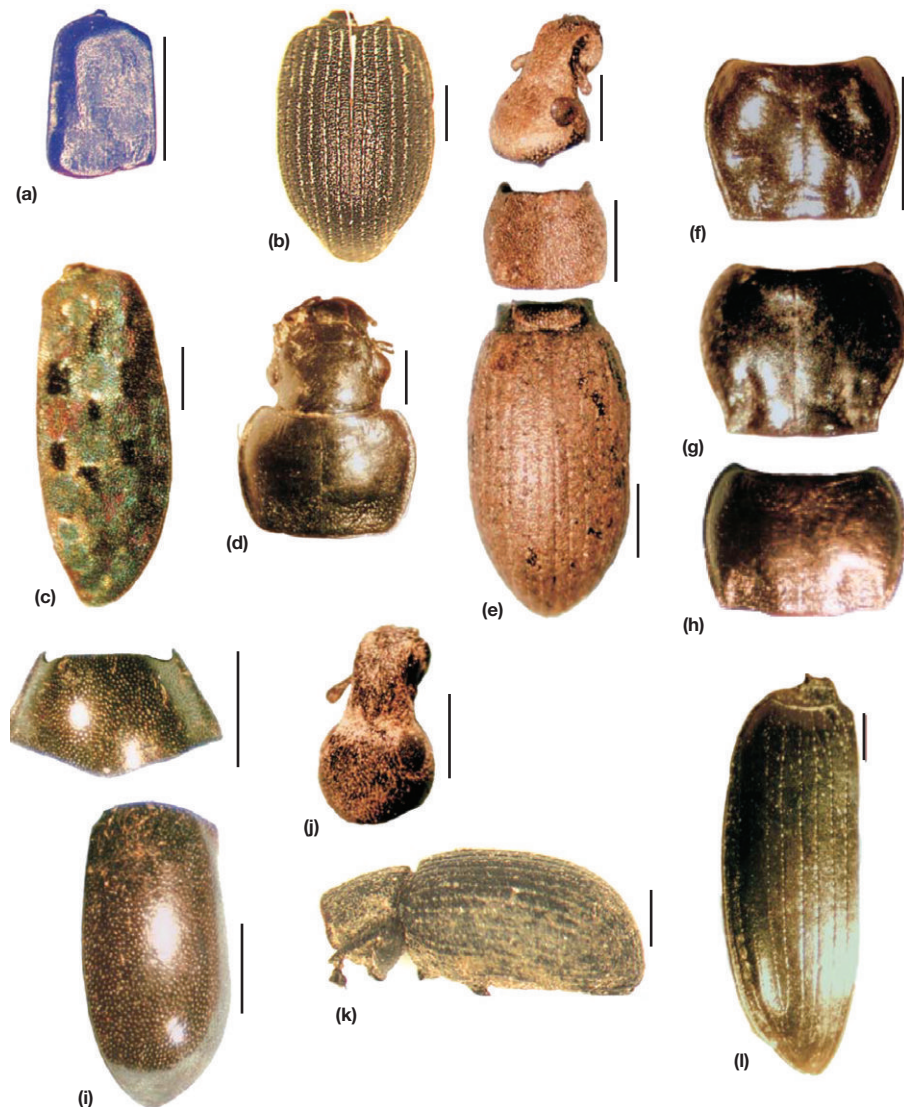


Figure 5 Light microscope photographs of late Pleistocene fossil beetle specimens from eastern Beringia. (a) Left elytron of *Tachinus brevipennis*; (b) elytra of *Vitavitus thulius*; (c) right elytron of *Elaphrus parviceps*; (d) head and pronotum of *Harpalus amputatus*; (e) head, pronotum, and elytra of *Connatichela artemisiae*; (f) pronotum of *Pterostichus brevicornis*; (g) pronotum of *Pterostichus pinguedineus*; (h) pronotum of *Amara alpina*; (i) pronotum and right elytron of *Morychus* sp.; (j) head of *Lepidophorus linneaticollis*; (k) pronotum and elytron of *L. linneaticollis*; (l) left elytron of *Pterostichus nearcticus*. Scale bars = 1 mm.

Beetle faunas associated with younger intervals of MIS 5 have yielded MCR estimates showing climatic cooling, with $T_{\max}$ levels declining 2–3 °C below modern parameters (Table 3). Two faunas thought to derive from MIS 4 indicate that summer temperatures continued to cool. The MIS 4 faunal assemblage from Hungry Creek, Yukon (Figure 4, No. 9), yielded a $T_{\max}$ estimate 6–9 °C cooler than modern.

The long MIS 3 interstadial complex in eastern Beringia is represented by 25 faunal assemblages from 18 different sites. Warming intervals are indicated by faunas from the Titaluk River in northern Alaska (Figure 4, No. 12) (46.4, 36, and 33.6 ka), from Cape Deceit on the Seward Peninsula (Figure 4, No. 14) (42.2 ka), and from Kulukbuk Bluffs in southwestern Alaska (35 ka). The strongest indication of interstadial warming comes from the Titaluk River fauna dated 33.6 ka. This fauna

yielded a $T_{\max}$ estimate 0.5–2 °C warmer than modern. The other faunas discussed previously yielded $T_{\max}$ estimates that were 0.5–2 °C cooler than modern. Interestingly, a fauna dated 31.5 ka from Mayo Village, Yukon (Figure 4, No. 18), indicates that regional $T_{\max}$ had fallen to 5–6 °C colder than modern levels. Likewise, a fauna dated 35.2 ka from Eva Creek, interior Alaska (Figure 4, No. 8), indicated $T_{\max}$ levels 7–8 °C colder than modern. Thus, within the space of 2,000 years, temperatures appear to have oscillated dramatically in eastern Beringia. The beetle faunas that yielded the indications of interstadial warming are composed of species found today in open-ground habitats within the boreal forest. They do not rely on the presence of trees. The paleobotanical evidence suggests that most if not all of eastern Beringia remained open-ground tundra or steppe–tundra throughout MIS 3.

During the LGM, the beetle evidence indicates that $T_{\max}$ was depressed in eastern Beringia, but that $T_{\min}$ was within 1–2 °C of modern levels in most regions. Elias et al. (1999) discussed difficulties in reconstructing $T_{\min}$ from faunal assemblages sampled from coastal sites. First, the ancient coastline of Beringia was removed from its current position by more than 1000 km in some regions. Therefore, the modern $T_{\min}$ of a coastal site, greatly influenced by maritime climate, bears little relation to the ancient climate of the same site when it was hundreds of kilometers inland because of lowered sea level. Also, even the modern $T_{\min}$ of coastal sites is a poor reflection of winter temperatures at these sites because of the incursion of Arctic air masses during the winter season. These periods of very low temperature are sufficiently long, and sufficiently frequent, to virtually eliminate the less cold-resistant beetle fauna of coastal Alaska.

Following the LGM, beetle faunas from throughout eastern Beringia are indicative of climatic amelioration, beginning by at least 15.3 ka. This warming is especially noteworthy in the Arctic zone, where $T_{\max}$ rose above modern levels during the Late Glacial interval. The warming indicated by faunas from the subArctic zone is less marked during this interval. There is also evidence, again from the Arctic assemblages, for a brief climatic oscillation during the Younger Dryas chronozone (Elias, 2000).

Two principal ecosystems appear to have dominated eastern Beringia during late Pleistocene glacial periods: steppe–tundra and mesic tundra. Steppe–tundra, or ‘mammoth steppe’ as Guthrie (1990) described it, was rich in grasses and sagebrush (*Artemisia*). It supported a diverse, abundant megafauna of grazers and their predators. In Alaska and the Yukon, this ecosystem also had a characteristic beetle fauna, dominated by the ground beetles *Amara alpina* (Figure 5(h)) and several species of *Harpalus* (Figure 5(d)) pill beetles (Byrrhidae) in the genus *Morychus* (Figure 5(i)), dung beetles in the genus *Aphodius*, and the weevils *Lepidophorus lineaticollis* (Figures 5(j) and 5(k)) and *Connatichela artemisiae* (Figure 5(e)). This steppe–tundra fauna, first recognized by Matthews (1983), is commonly found in fossil assemblages throughout the Yukon and interior Alaska. However, in other regions of Alaska, a different suite of beetle taxa dominate late Pleistocene assemblages. This group includes the *Cryobius* group of ground beetles in the genus *Pterostichus* (Figures 5(f) and 5(g)), several other ground beetles associated with mesic and moist habitats, rove beetles in the subfamily Omaliinae, and weevils that feed on dwarf birch and dwarf willow. This mesic tundra fauna has been found in fossil assemblages throughout southwestern and northwestern Alaska, as well as in samples taken from the ancient Bering Land Bridge that connected Alaska to Siberia during intervals of lowered sea level (Elias et al., 2000). The land bridge may have served as a biological filter, keeping dry-adapted species, such as the woolly rhinoceros, from entering North America (Guthrie, 2001).

Conclusions

Late Pleistocene beetle assemblages from sites throughout North America have documented large-scale, rapid climatic changes. During MIS 3, beetle assemblages have demonstrated

oscillating interstadial warming and cooling events that coincide with climatic patterns inferred from oxygen isotope records in Greenland ice cores. The pace and severity of such changes have often gone undetected in paleoenvironmental reconstructions based on fossil pollen, although new analytical methods are making breakthroughs (Grimm and Jacobson, 2004). As has been noted in the history of the European beetle fauna (Coope, 1978), North American beetles appear to have responded to the vagaries of Pleistocene climates by shifting their distributions in order to stay within climatic conditions to which they are best adapted. Thus, the Pleistocene beetle fauna of North America remains extant today, albeit living in different regions. Each new climatic regime of the Pleistocene brought new associations of species together in a given region. Many of the species in given types of associations remained constant through long stretches of time, but different species were added or subtracted from the fauna of any specific region as ecosystems waxed and waned.

See also: [Beetle Records: Late Pleistocene of Europe](#); [Late Tertiary and Early Quaternary Records](#); [Overview](#); [Postglacial North America](#).
Ice Core Records: [Greenland Stable Isotopes](#). **Paleoceanography, Physical and Chemical Proxies:** [Oxygen-Isotope Stratigraphy of the Oceans](#). **Plant Macrofossil Methods and Studies:** [Rodent Middens](#).

References

- Anderson TW, Matthews JV Jr., Mott RJ, and Richard SH (1990) The Sangamonian Pointe-Fortune site, Ontario–Quebec border. *Géographie Physique et Quaternaire* 44: 271–287.
- Ashworth AC (2004) Quaternary Coleoptera of the United States and Canada. In: Gillespie AR, Porter SC, and Atwater BF (eds.) *The Quaternary Period in the United States*, pp. 505–517. New York: Elsevier.
- Baker RG, Schwert DP, Bettis EA III, Kemmis TJ, Horton DG, and Semken HA (1991) Mid-Wisconsinan stratigraphy and paleoenvironments at the St. Charles site in south-central Iowa. *Geological Society of America Bulletin* 103: 210–220.
- Benson L (2004) Western lakes. In: Gillespie AR, Porter SC, and Atwater BF (eds.) *The Quaternary Period in the United States*, pp. 185–204. New York: Elsevier.
- Cong S and Ashworth AC (1996) Palaeoenvironmental interpretation of middle and late Wisconsinan fossil coleopteran assemblages from western Olympic Peninsula, Washington, USA. *Journal of Quaternary Science* 11: 345–356.
- Coope GR (1978) Constancy of insect species versus inconstancy of Quaternary environments. In: Mound LA and Waloff N (eds.) *Diversity of Insect Faunas*, pp. 176–187. London: Royal Entomological Society of London. Symposium No. 9.
- Coope GR and Lemdahl G (1995) Regional differences in the late glacial climate of northern Europe based on coleopteran analysis. *Journal of Quaternary Science* 10: 391–395.
- Elias SA (1990) The timing and intensity of environmental changes during the Paleoindian period in western North America: Evidence from the fossil insect record. In: Agenbroad LD, Mead JJ, and Nelson LW (eds.) *Megafauna and Man*, pp. 11–14. Flagstaff: Northern Arizona Press.
- Elias SA (1991) Insects and climate change: Fossil evidence from the Rocky Mountains. *BioScience* 41: 552–559.
- Elias SA (1992) Late Quaternary zoogeography of the Chihuahuan Desert insect fauna, based on fossil records from packrat middens. *Journal of Biogeography* 19: 285–297.
- Elias SA (1994) *Quaternary Insects and Their Environments*. Washington, DC: Smithsonian Institution Press.
- Elias SA (1998) The mutual climatic range method of paleoclimate reconstruction based on insect fossils: New applications and interhemispheric comparisons. *Quaternary Science Reviews* 16: 1217–1225.

- Elias SA (1999) Mid-Wisconsin seasonal temperatures reconstructed from fossil beetle assemblages in eastern North America: Comparisons with other proxy records from the Northern Hemisphere. *Journal of Quaternary Science* 14: 255–262.
- Elias SA (2000) Late Pleistocene climates of Beringia, based on fossil beetle analysis. *Quaternary Research* 53: 229–235.
- Elias SA (2001) Mutual climatic range reconstructions of seasonal temperatures based on late Pleistocene fossil beetle assemblages in eastern Beringia. *Quaternary Science Reviews* 20: 77–91.
- Elias SA and Van Devender TR (1990) Fossil insect evidence for late Quaternary climatic change in the Big Bend region, Chihuahuan Desert, Texas. *Quaternary Research* 34: 249–261.
- Elias SA and Van Devender TR (1992) Insect fossil evidence of late Quaternary environments in the northern Chihuahuan Desert of Texas and New Mexico: Comparisons with the paleobotanical record. *Southwestern Nature* 37: 101–116.
- Elias SA, Mead JI, and Agenbroad LD (1992) Late Quaternary arthropods from the Colorado Plateau, Arizona and Utah. *Great Basin Naturalist* 52: 59–67.
- Elias SA, Van Devender TR, and De Baca R (1995) Insect fossil evidence of late glacial and Holocene environments in the Bolson de Mapimi, Chihuahuan Desert, Mexico: Comparisons with the paleobotanical record. *Palaios* 10: 454–464.
- Elias SA, Andrews JT, and Anderson KH (1999) New insights on the climatic constraints on the beetle fauna of coastal Alaska derived from the mutual climatic range method of paleoclimate reconstruction. *Arctic, Antarctic, and Alpine Research* 31: 94–98.
- Elias SA, Berman D, and Alifimov A (2000) Late Pleistocene beetle faunas of Beringia: Where east met west. *Journal of Biogeography* 27: 1349–1363.
- Grimm EC and Jacobson GL Jr. (2004) Late Quaternary vegetation history of the eastern United States. In: Gillespie AR, Porter SC, and Atwater BF (eds.) *The Quaternary Period in the United States*, pp. 381–402. New York: Elsevier.
- Grootes PM, Stuiver M, White JWC, Johnsen S, and Jouzel J (1993) Comparison of oxygen isotope records from the GISP2 and GRIP Greenland ice cores. *Nature* 366: 552–554.
- Guthrie RD (1990) *Frozen Fauna of the Mammoth Steppe. The Story of Blue Babe*. Chicago: University of Chicago Press.
- Guthrie RD (2001) Origins and causes of the mammoth steppe: A story of cloud cover, woolly mammoth tooth pits, buckles, and inside-out Beringia. *Quaternary Science Reviews* 20: 549–574.
- Matthews JV Jr. (1983) A method for comparison of northern fossil insect assemblages. *Géographie Physique et Quaternaire* 37: 297–306.
- Miller RF, Morgan AV, and Hicock SR (1985) Pre-Vashon fossil Coleoptera of Fraser age from the Fraser Lowland, British Columbia. *Canadian Journal of Earth Science* 22: 498–505.
- Miller SE (1983) Late Quaternary insects of Rancho La Brea and McKittrick, California. *Quaternary Research* 20: 90–104.
- Nelson RE and Coope GR (1982) A late Pleistocene insect fauna from Seattle, Washington. American Quaternary Association seventh biennial meeting, program and abstracts 146.
- Schwert DP and Ashworth AC (1988) Late Quaternary history of the northern beetle fauna of North America: A synthesis of fossil distributional evidence. *Memoirs of the Entomological Society of Canada* 144: 93–107.
- Thompson RS, Shafer SL, Strickland LE, Van de Water PK, and Anderson KH (2004) Quaternary vegetation and climate change in the western United States: Developments, perspectives, and prospects. In: Gillespie AR, Porter SC, and Atwater BF (eds.) *The Quaternary Period in the United States*, pp. 403–426. New York: Elsevier.
- Walker IR, Mott RJ, and Smol JP (1991) Allerød-Younger Dryas lake temperatures from midge fossils in Atlantic Canada. *Science* 253: 1010–1012.

Late Pleistocene of South America

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Background

Studies of fossil beetle assemblages in South America have been directed at coming to a better understanding of how beetles respond to climate change and to the use of that information for interpreting paleoclimate. One of the primary challenges of the research has been to demonstrate that the results would be sufficiently robust for other researchers in archaeology, plant paleoecology, glacial geology, and paleoclimatology to use with confidence in their studies. Of particular interest has been the question of whether a climatic reversal occurred in southern South America that was of similar magnitude and timing to the Younger Dryas interval defined from the North Atlantic region. The sites from which Pleistocene fossil assemblages have been studied in South America are located primarily in the Chilean lake region and the Chilean channels, between 40 and 50° S (Figure 1). A single fossil assemblage of early to mid-Holocene age has also been examined from the Falkland Islands in the Atlantic Ocean between 51 and 52° S (Buckland and Hammond, 1997).

The Landscape of Southern Chile

In the Región de Los Lagos (lake region), the terrain is mountainous with the Cordillera de la Costa (coastal ranges) separated from the stratovolcanoes of the Cordillera de los Andes (Andes) by a wide north–south trending lowland, the Valle Longitudinal (central valley). Dense forests once cloaked the entire landscape of the lower elevations but have mostly been cleared for agriculture. Today, extensive tracts of native Valdivian and North Patagonian rain forests survive only at higher elevations in the lake region (Figure 2(a)). Above treeline, the forests are replaced by an Andean tundra (Figure 2(b)) that extends to the fell fields, and permanent snow and ice capping the higher volcanic peaks. Further south in the Chilean channels, the landscape is more fragmented. Island archipelagoes, crisscrossed by deep glacially excavated fjords, extend to the tip of the continent. The permanent snow line is lower, and outlet glaciers of the North and South Patagonian ice fields calve icebergs into the fjords (Figure 3(a)). The lowland vegetation is a dense cold-temperate rainforest (Figure 3(b)) which, at higher elevations, is replaced by the saturated soils and blanket bogs of the Magellanic moorland (Figure 3(c)). The climate in southern South America, from the mid- to the high latitudes, is dominated by a westerly circulation stronger in the south than in the north. Whitlock et al. (2001) discuss the relationship between climate, vegetation, and the insect fauna.

The Existing Beetle Fauna

The beetle fauna of the forests, moorlands, and steppes of southern South America is unusually rich in endemic taxa.

Endemism implies ancient evolutionary relationships and Patagonia is the home of many surviving lineages of ancient clades. The Migadopini, for example, are an exclusively Southern Hemisphere tribe of Carabidae (ground beetles) that have a typically Gondwanan distribution. They occur throughout southern South America, including the Falkland Islands, Tasmania and southeastern Australia, and New Zealand and the Auckland Islands. Similar restricted geographic distributions occur within several families of beetles in the southern South American fauna, notably the Protocujidae, Chalcodryidae, Nemonychidae, and Belidae. These groups are either monotypic or represented by a small number of genera that are associated with *Nothofagus* forest. The taxa in many of these families evolved during the late Cretaceous when southern South America was still connected to Antarctica. The connection was severed by seafloor spreading to form the Drake Passage between 34 and 30 Ma. The recent discovery of fossils of listroderine weevils in Neogene deposits, 500 km from the South Pole, confirms the evolutionary ties that South America had with Antarctica (Ashworth and Kuschel, 2003). Listroderine weevils are presently widely distributed in the forests, steppes, and moorlands of southern South America.

Joseph Banks, naturalist on the HMS Endeavour, Captain James Cook's first voyage of exploration, collected beetles from southern South America in 1769. One of those specimens, the ground beetle *Ceroglossus suturalis*, was described by Johann Fabricius in 1775. In the nineteenth century, Charles Darwin and other naturalists, who followed, sent their specimens back to Europe to be described by specialists in the natural history museums in London and Paris. Darwin was responsible for collecting one of the most distinctive beetles of the Chilean rain forest, the lucanid *Chiasognathus grantii* Stephens. The male is characterized by mandibles that are almost twice as long as its body (Figure 4(a)) giving rise to its common name in southern Chile of 'ciervo volante' or 'flying deer.' In *The Descent of Man*, Darwin wrote 'But the mandibles were not strong enough so as to cause actual pain.' The specimens are preserved in the Natural History Museum, London, where the male is known as 'the beetle that bit Darwin.' The nineteenth century naturalist who made the greatest contribution to the knowledge of the Chilean fauna was Claudio Gay. He made extensive collections of the fauna during travels between 1828 and 1842, and those specimens are described and illustrated in color in his 1849 and 1851 volumes of *La Historia Física y Política de Chile* (Figure 4(a)). More recently, Elizabeth Arias (2000) has described the most distinctive species of the fauna in *Coleópteros de Chile*.

There are numerous excellent taxonomic treatments of the existing beetle fauna which are invaluable as aids in the identification of Pleistocene fossils. There are few faunistic and ecological studies, however, on which to base paleoenvironmental and paleoclimatic interpretations. Ashworth and Hoganson (1987) made extensive collections in the Parque

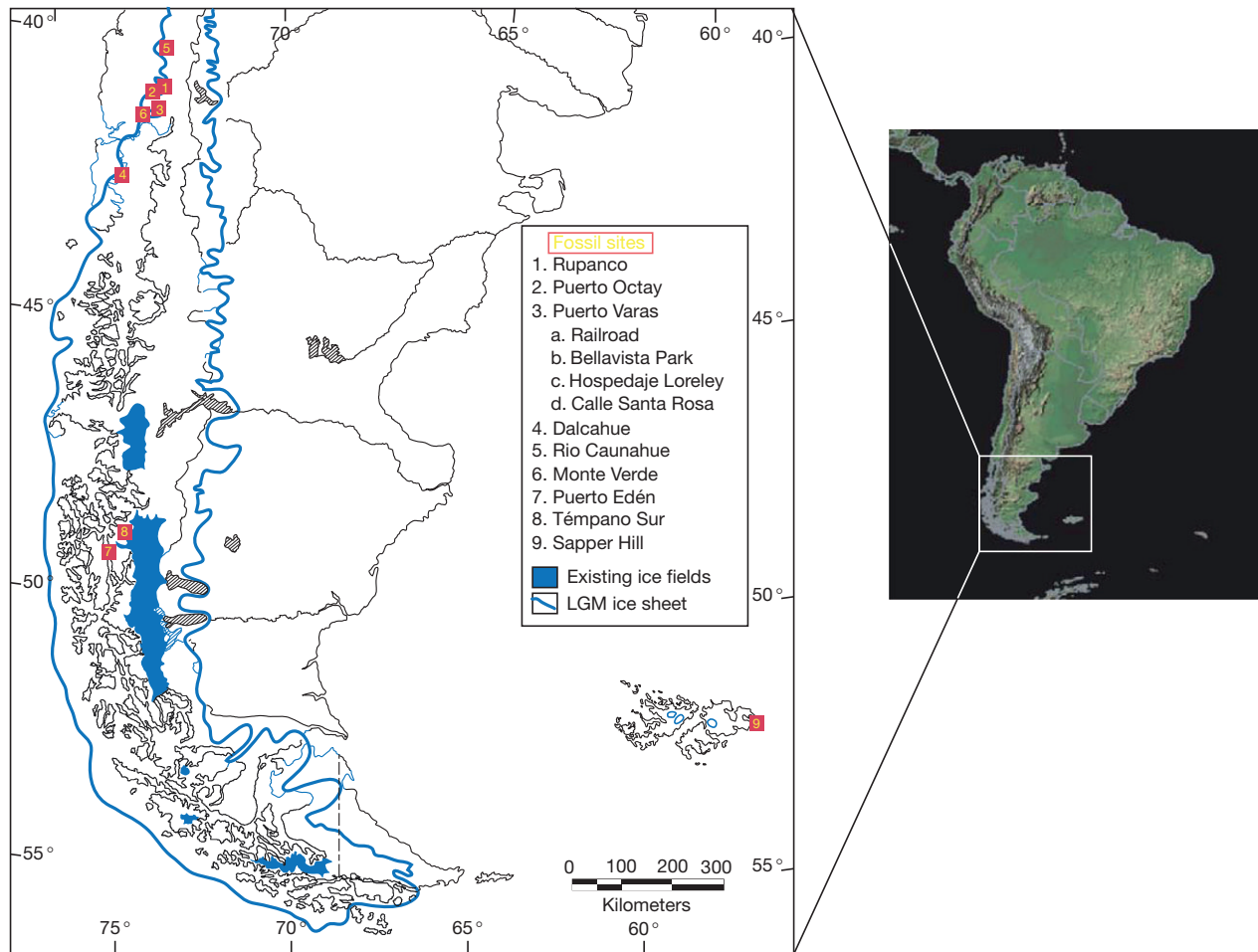


Figure 1 Location of fossil beetle assemblages from southern South America. The representation of the Patagonian ice sheet during the last glaciation is from [Hollin and Schilling \(1981\)](#).

Nacional Puyehue and other undisturbed habitats in the Chilean lake region ([Figure 4\(b\)](#)). They described a diverse fauna of 462 species from habitats collected systematically along an elevational transect from sea level to 1200 m. Their ordination analysis showed changes in beetle communities along the transect which generally correlated with changes in vegetation.

Fossil Studies

Chilean Lake Region

In the Chilean lake region, fossil assemblages ranging in age from 24,000 to 4,525 ka BP have been examined. Assemblages from sites at Puerto Octay, Puerto Varas (Park and Railroad), Rio Caunahue, and Monte Verde have been described and their significance discussed in a series of papers ([Hoganson et al., 1989](#); [Hoganson and Ashworth, 1992](#); [Ashworth and Hoganson, 1993](#)). Additionally, unpublished fossil assemblages have been studied by Ashworth from sites at Rupanco, Dalcahue, and Puerto Varas (Loreley and Calle Santa Rosa) ([Figure 1](#)). All of the sites, with the exception of Monte Verde, are from locations in the lowlands of the central valley that were glaciated during the Last Glacial Maximum. The Monte Verde site is

located west of the terminal moraines of the last glaciation and is associated with a drainage that developed on glacial outwash.

The oldest fossil beetle assemblages to have been examined are from peat deposits near the west end of Lago Rupanco ([Figure 1](#)). Fossils were extracted from an *in situ* peat that is overlain by gravel, till, and volcanic deposits ([Mercer, 1976](#); [Denton et al., 1999](#)). The assemblages range in age from 19 500 to 24 000 yr BP and are representative of the beetle fauna of the lake region immediately prior to the Last Glacial Maximum. The assemblages are the least diverse of any examined from the lake region consisting of only 9–10 species. Exceptionally well-preserved skeletal parts of the listroderine weevil *Germaniellus dentipennis* (Germain) ([Figure 5](#)), including heads, pronota, elytra, sternites, leg segments, and genitalia, make up 90% of the fossils.

Similar assemblages dominated by *G. dentipennis* occur within the lake region until about 14 000 to 14 500 yr BP. The youngest of these assemblages, dated at 14 340 yr BP, is from near Dalcahue on La Isla Grande de Chiloé (Chiloé). The peaty deposits are overlain by outwash gravels and till of the final major glacial advance in the mid-latitudes of South America ([Figure 6](#)). Included in the 22 taxa from this



Figure 2 Habitats in the Chilean lake region: (a) The lower Valdivian rain forest in the lake region near Aguas Calientes, Parque Nacional Puyehue; (b) the Andean tundra and krummholz at treeline east of Antillanca, Parque Nacional Puyehue. Volcán Casablanca is in the background. Photographs by A. C. Ashworth.

assemblage is the weevil *Paulsenius carinicolis* (Blanchard) and the ground beetle *Cascellius septentrionalis* Roig-Juñent that today both inhabit higher elevations in the Andes. Villagrán (1990) made the observation that Magellanic moorland plants, which currently inhabit relict patches at 700 m elevation in the coastal ranges of Chiloé, grew near sea level during the last glaciation Villagrán (1990). Using an adiabatic lapse rate of $0.65^{\circ}\text{C}/100\text{ m}$, Ashworth and Hoganson (1993) estimated that mean summer temperature during the glacial interval was $4\text{--}5^{\circ}\text{C}$ lower than today based on the species composition of fossil beetle assemblages. Denton et al. (1999) estimated that snow and treeline were lowered by 1000 m between 29 400 and 14 400 yr BP representing mean summer temperatures $6\text{--}8^{\circ}\text{C}$ lower than today.

Fossil assemblages postdating 14 000 yr BP are fundamentally different in their composition. Not only are they considerably more diverse but also contain higher percentages of taxa that are obligate forest inhabitants. Most of the assemblages that have been examined are from exposed cut bank sections on the Rio Caunahue, a large river that originates in the Andes and enters Lago Ranco on the east side. Rhythmically laminated deposits containing abundant organics and volcanic ashes were deposited in a narrow arm of the lake immediately following deglaciation (Figure 7(a)). The basal sediments are inorganic



Figure 3 Habitats in the Chilean channels: (a) The Glaciar Pio XI in the Chilean channels; (b) dense rainforest of *Nothofagus betuloides* in the Fiordo Bernardo. (c) Magellanic moorland in the coastal mountains west of Puerto Edén. Photographs by A. C. Ashworth.

clays representing the deposition of rock flour. The 7-m sequence, with a basal age of 13 900 yr BP (Ashworth and Hoganson, 1993), contains exceptionally well-preserved fossils (Figure 7(b)). The species compositions of 30 intervals pre-dating 10 000 yr BP were reported in Hoganson and Ashworth (1992); the number of taxa identified was 154, of which 41–48% consisted of obligate forest taxa. All of the taxa currently inhabit the forests of the lake region, mostly the Valdivian rain forest.

The only other fossil beetle assemblage post-dating 14 000 yr BP that has been examined from the lake region is



Figure 4 Modern beetle fauna: (a) Enlargement of one of the plates from Gay's nineteenth-century *Historia Física y Política de Chile*. The beetle with the large mandibles is the distinctive Valdivian rain forest lucanid *Chiasognathus grantii*; (b) studies of the modern beetle fauna were essential to establishing the base for paleoclimatic interpretation. Photograph by A. C. Ashworth.

from the archaeological site at Monte Verde, near Puerto Montt (Figure 8). At Monte Verde, humans are believed to have constructed primitive shelters and hunted mastodon. Fossil beetles were retrieved from horizons considered to be contemporaneous with the human habitation (MV-6) dated between 13 565 and 11 790 yr BP, and also from a peat bed (MV-5) that

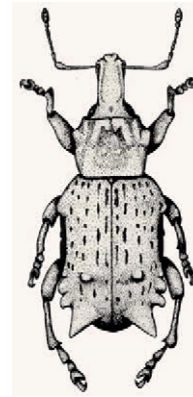


Figure 5 The weevil *Germainiellus dentipennis* (Morrone, 1993) whose exoskeletal remains are the most abundant in fossil assemblages dating from 24 000 to 14 000 yr BP. Reproduced from Morrone JJ (1993) Revisión sistemática de un nuevo género de Rhytirrhini (Coleoptera, Curculionid), con un análisis biogeográfico del dominio subantártico. *Boletín de la sociedad de Biología de Concepción, Chile*, 64: 121–145.

sealed the cultural horizon and is dated between 11 810 and 10 860 yr BP (Dillehay and Pino, 1989). Statistical comparison of the 95 taxa occurring in MV-6 and MV-5 demonstrated that the assemblages were similar to one another (Hoganson et al., 1989).

Using cluster analysis, the species compositions of the fossil assemblages were compared to modern communities from 41 locations within the Chilean lake region. The Monte Verde assemblage was most similar to the communities from slightly disturbed coastal forests and from the Valdivian Rain Forest at about 500 m elevation (Hoganson et al., 1989). The interpretation that the paleoclimate at Monte Verde was essentially similar to that of today during the human occupation is consistent with interpretations from similar-aged fossil beetle assemblages at the Rio Caunahue site.

One of the intriguing puzzles of the Monte Verde beetle assemblage is that no synanthropic beetles were identified. In Europe, fossil beetle assemblages from archaeological sites ranging in age from Mesolithic to historical times show evidence of modification, especially those associated with agricultural. Why there would be no synanthropic insects at Monte Verde is unknown, but it is possible that hunter-gatherer groups did not disturb the natural environment to such an extent that it caused an ecological response in the beetle fauna. A similar observation was made for early Postglacial beetle faunas associated with human occupation in North America (Elias, 1994).

To examine longer-term trends represented by the fossil beetle assemblages, the total taxa and the ratio of taxa dependent on trees are examined for 1 ka intervals between 24 000 and 9 000 yr BP (Figure 9(a)). The beetle fauna from 24 000 yr BP to the time of the last glacial advance about 14 000 yr BP was composed of species of wet moorland habitats. The assemblages are characterized by low diversity and low numbers of taxa dependent on shrubs and trees. Between 14 000 and 13 000 yr BP the diversity is low but the percentage of arboreal taxa is higher. Between 13 000 and 12 000 yr BP, the assemblages are at least 5 times as diverse as those of the glacial interval. The changes indicate that by 13 000 yr BP, moorland



Figure 6 (a) At Dalcahue, Isla Grande de Chiloé, organic sediments (brown) contain the youngest fossil beetle assemblages representing the glacial beetle fauna. Fossil wood at the top of the section (by the knife) has an age of 14.34 ka BP; (b) organic sediments accumulated in a small depression on the surface of a weathered till. The surface was buried by outwash gravel (top of the section) during the last glacial advance in the lake region. Scale = cm divisions. Photographs by A. C. Ashworth.



Figure 7 (a) The Rio Caunahue assemblages are from laminated silts interbedded with volcanic ashes. The bench on which the person is standing has an age of 12.81 ka BP and the prominent white volcanic ash above his head has an age of about 10.8 ka BP. (b) The fused fossil elytra (wing cases) of a specimen of the ground beetle *Ceroglossus*, that is from the horizon dated at 10 ka BP. (c) A modern specimen of *Ceroglossus* from the Valdivian rain forest. Scale bars = 5 mm. Photographs by A. C. Ashworth.

habitats, which had existed in the central valley for more than 10 000 yr, had been replaced by forested habitats similar to those that exist in the region today.

The numerical data are also used to test two null hypotheses (Figure 9(b)). The first is that no relationship exists between

total taxa and ratio of arboreal taxa. This hypothesis was tested using linear regression ($\alpha = 0.05$). The null hypothesis was shown not to be true, and the relationship between total taxa and ratio of arboreal taxa was found to be highly significant ($P = 0.001$, $r^2 = 0.62$). The interval from 14 000 to 13 000 yr



Figure 8 The Monte Verde archaeological excavation. An excavated hearth-like structure is visible in the foreground. The peat MV-5 which contains abundant fossil beetles and is dated between 11.81 and 10.86 ka BP is visible at the base of the cut on the far wall of the excavation. Photograph by A. C. Ashworth.

BP has a studentized residual value of 11.5 making it very different from all other samples. The difference is explained because the sample represents most of the change in the transition between moorland and forested biomes.

The second null hypothesis is that no significant difference exists between the ratios of arboreal taxa for two subsets of samples between 24 000–15 000 yr BP and 13 000–9 000 yr BP. The sample representing the 13 000–14 000 yr BP interval was not included in the test because it is a statistical outlier. The null hypothesis was tested using a two-sample T-test ($\alpha = 0.05$). The result was that the null hypothesis was shown not to be true. The ratios of arboreal taxa to total taxa for the glacial interval 24 000–15 000 yr BP are significantly different than for the postglacial interval between 13 000 and 9 000 yr BP ($t = 9.842$, $df = 8.5$, $P < 0.001$).

Chilean Channels

Fossil beetle assemblages have been examined from two locations in the Chilean channels. The Puerto Edén site on the Canal Messier is from an excavation in a bog (Figure 10), and the Témpano Sur site is from an outcrop in the wall of an abandoned meltwater channel about 1 km from the margin of the Glaciar Témpano, an outlet glacier of the South

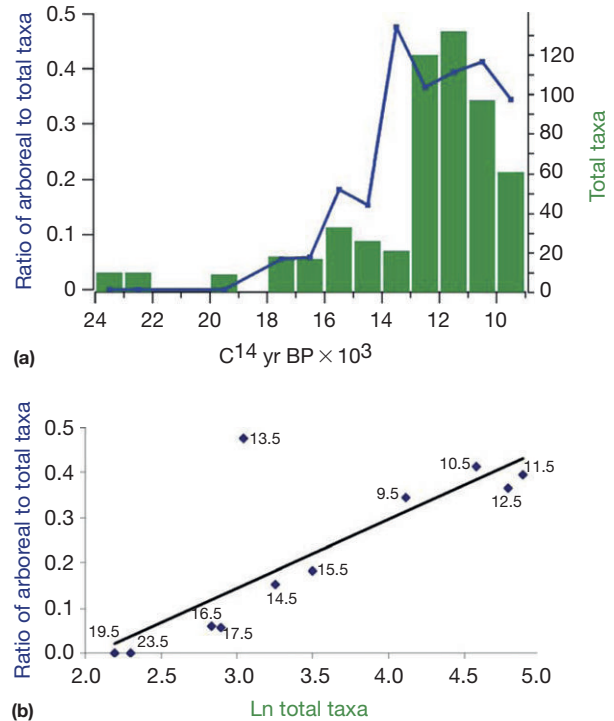


Figure 9 Longer-term trends for the interval 24–9 ka BP represented by a compilation of data from all of the fossil assemblages: (a) plots of the total number of taxa (bar graph) and the ratio of obligate arboreal taxa (line graph); (b) the relationship between total and ratio of arboreal taxa. The x-axis is natural log transformed. The labels shown are mid-ranges for samples, each of which represents a 1 ka interval of time. The line represents a significant linear regression. The data point at 13.5 (14–13 ka BP) is a significant outlier representing the time of major change between moorland and forest assemblages.



Figure 10 The bog at Puerto Edén is located between morainic arcs deposited by a glacier that retreated into the coastal ranges. Glacial ice also filled the Paso del Indio, the marine channel that leads northward to the Messier Canal. Photograph by A. C. Ashworth.

Patagonian ice field (Figure 11). At both sites, pollen and fossil beetles were examined from the same samples (Ashworth and Markgraf, 1989; Ashworth et al., 1991). As the glacier at Puerto Edén receded westward to the coastal mountains, peat began to accumulate at about 13 000 yr BP in a shallow depression

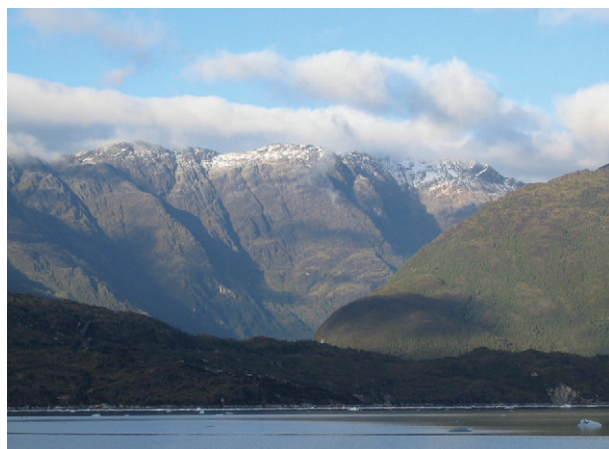


Figure 11 The fossil site at Témpano Sur is located behind the bedrock ridge in the foreground. It is located in an abandoned meltwater channel that has an outlet to the fjord at the extreme right of the photograph. The floating ice is from melting of the Glaciar Témpano about 1 km to the left of the image. Photograph by A. C. Ashworth.

between morainal ridges. Throughout the peat profile there is excellent agreement between the interpretations based on fossil beetles and those based on pollen (Ashworth et al., 1991). Immediately following deglaciation, the colonizing plants and beetles were those of open ground communities, similar to those inhabiting active glacial outwash plains about 45 km east of the site. From 13 000 to about 9 500 yr BP the biota was that of a *Nothofagus* woodland which persisted until 5 500 yr BP, when it was replaced by a rainforest-moorland mosaic similar to that existing in the region today. Sometime between 5 500 and 3 000 yr BP, the biota changed dramatically on the moraines surrounding the bog, and the bog itself dried up. After 3 000 yr BP, a biota similar to that which had existed between 9 500 and 5 500 yr BP returned. Puerto Edén has a precipitation of about 3000 mm yr⁻¹. For the bog to completely dry up requires a very significant climatic change. Markgraf et al. (2003) reported changes in the sedimentology, paleontology, and geomorphology of Lago Cardiel (49° S) on the Patagonian steppe, east of Puerto Edén indicating severe droughts between 6000 and 5000 ka BP. They point out that widespread mid-Holocene aridity has been reported in records from the Patagonian steppe to the North American plains, but for which, as yet, there is no compelling climatological explanation.

At the head of the Témpano fjord, interbedded sands and peats infilled a bedrock-floored channel that was later eroded by meltwater. The basal peat is dated between 11 180 and 10 130 yr BP. Fossil beetle and pollen assemblages from the peat indicate a fauna and flora very similar to that inhabiting stable substrates within the area today. The implication is that the South Patagonian ice field during the Younger Dryas interval was no larger than it is today and could even have been smaller (Ashworth and Markgraf, 1989).

Biogeographic Considerations

In the montane and island-fragmented landscape of southern Chile, the onset of the last phase of the glaciation is dated at 29 400 yr BP (Denton, et al., 1999). Nothing is known of the

beetle fauna that existed before the final phase of glaciation; however, it is reasonable to assume that it was a forest fauna based on available pollen analyses. Further, it is reasonable to speculate that as the climate cooled, species became regionally extinct. It is assumed that these species survived in more northerly locations and were available for recolonization when the climate warmed between 14 000 and 13 000 yr BP. The fauna that survived glaciation in the moorland habitat of the lake region has no modern analog. It consisted of: (1) species that had dispersed downslope from habitats high in the coastal ranges and the Andes and (2) species that were 'climatic generalists' that could survive the lower temperatures. There is no evidence that cold-adapted species from more southerly latitudes migrated to the lake region.

The absence of evidence for latitudinal movement of beetle taxa in response to climate change during the last glaciation makes southern South America very different than either Europe or North America where species were displaced more than 1000 km from their existing distributions (Ashworth, 2001). The prevailing hypothesis is that the biota south of Chiloé was completely extirpated during the last glaciation and consequently would not have been available to colonize more northerly regions. The hypothesis has serious flaws (see below), but even if pockets of biota did survive to the south, northward dispersal would have been blocked by barriers imposed by marine inlets and large glaciers.

Reconstructions of ice cover for southern South America during the Last Glacial Maximum portray an ice sheet that was continuous from latitude 43° S in the lake region to 56° S in Tierra del Fuego (Hollin and Schilling, 1981) (Figure 1). At the northern end, parts of Chiloé, the coastal ranges, and the central valley were ice free. South of Chiloé, however, ice is shown as covering the coastal mountains and extending to the Pacific Ocean. An equilibrium line altitude lowering of 900 m was used by Hubbard et al. (2005) to model the central regions of the Patagonian ice sheet between 45° and 48° S. Their model shows that by 21 000 yr BP ice covered the Taitao Peninsula extending to the coastal shelf, comparable to the Hollin and Schilling (1981) reconstruction. Using climatic forcing modified from the Vostok ice core, their model shows an ice sheet that slowly retreated until 14 500 yr BP, after which it collapsed rapidly reaching its present-day configuration by 11 000 yr BP after briefly stabilizing during the Antarctic Cold Reversal.

The effect of glaciation is considered to have been devastating to the flora and fauna, resulting in the total extirpation of species south of Chiloé. Two pieces of evidence, however, suggest that the extent of glaciation and extirpation of the biota may have been overestimated. The first is that there are several species of large- to moderate-sized flightless beetles in the basal layer of the bog at Puerto Edén with an age of about 13 000 yr BP. The fossils include the ground beetles *Ceroglossus suturalis* F., *Creobius eydouxi* Guérin, *Cascellius gravesii* Curtis, and *C. septentrionalis* Roig-Juñent, and the weevils, *Aegorhinus kuscheli* Elgueta and *Germainiellus rugipennis* (Blanchard). For flightless species to be among the earliest colonizers of the bog implies that they had to have survived glaciation locally, either in the coastal mountains or more probably along the coast. The second piece of evidence supporting survival in local refugia is the high genetic diversity measured in isolated populations of the tree species *Pilgerodendron uvifera* (D. Don) Florin (Premoli

et al., 2002). If these populations were Holocene immigrants derived by long-distance dispersal from northern refugia, then it would be expected that the severity of the genetic bottleneck they have been through would have resulted in the populations having much lower genetic diversity.

The Younger Dryas Conundrum

One of the enduring problems associated with paleoclimatic interpretation in the Chilean lake region is whether the pattern of climatic changes at the end of the last glaciation was identical to that of the North Atlantic region. In particular, the problems focused on the existence of a Younger Dryas equivalent between 11 000 and 10 000 yr BP. Currently, the question involves whether there is a climatic cooling, the Huelmo-Mascardi cold reversal (Hajdas et al., 2003), between 12 400 and 10 000 yr BP. The initial disagreement centered on several interpretations based on pollen studies for a climatic reversal of 5–10 °C precisely correlated with the Younger Dryas (Heusser, 1974). The studies were considered to be strong evidence for the Younger Dryas being a global event. Ashworth and Hoganson (1993), however, argued that there was no evidence for any significant amount of cooling indicated by fossil beetle assemblages after 14 000 yr BP precipitating a disagreement that continues today.

In southern South America, outside of the Chilean lake region, the evidence for climatic cooling coeval with the Younger Dryas is equally problematical. Ariztegui et al. (1997) reported indirect evidence for a Younger Dryas glacial advance on Monte Tronador, based on studies of sediments, diatoms, and pollen from Lago Mascardi. Redating of that event has demonstrated that the proposed cooling includes the Younger Dryas but pre-dates it by 550 yr (Hajdas et al., 2003). Earlier, pollen studies in the region showed that there had been no response of vegetation to climate change between 11 000 and 10 000 yr BP (Markgraf, 1984). In pollen records from the Chonos (44° S) and Taitao peninsulas (46°), the vegetation shows a unidirectional shift from heath to forest, with essentially the modern forest developed by 12 400 yr BP (Bennett et al., 2000; Haberle and Bennett, 2004). No evidence was found for cooling during either the Younger Dryas or the earlier Antarctic Cold Reversal. However, one of the possible explanations for changes in chironomid (midge) fossil assemblages from a lake studied by Bennett et al. (2000) was for climatic cooling during the Younger Dryas (Massaferro and Brooks, 2002). Further south in the Chilean Channels (49° S) combined analyses of fossil beetle assemblages and pollen also showed no evidence of a Younger Dryas climatic cooling (Ashworth et al., 1991). At the Témpano glacier, Ashworth and Markgraf (1989) reported that the North Patagonian ice field was within its present boundaries during the Younger Dryas. In Tierra del Fuego, Heusser and Rabassa (1987) reported a glacial and vegetational response consistent with cooling during the Younger Dryas. From the same region, Markgraf (1993) reported the occurrence of charcoal during the Late Glacial and preferred to interpret the pollen in terms of local environmental instability rather than climatic cooling.

Moreno et al. (1999) reported a climatic cooling of 2–3 °C between 12 200 and 9 800 yr BP in the Chilean lake region based mostly on the occurrence of the pollen of *Podocarpus*

nubigena Lindl., a conifer which occurs in the North Patagonian and Valdivian rain forests. They discussed how the proposed cooling produced only minor changes in forest structure but still considered climate change to be the probable cause rather than disturbance from a nonclimatic cause. Ashworth and Hoganson (1987) were unable to discriminate statistically between the modern beetle communities of the North Patagonian and upper Valdivian rain forest which, in terms of paleoclimatic interpretation, would translate to ± 1 °C. The difference between interpretations based on pollen and those on fossil beetle assemblages is now within ± 1 °C or what many paleoecologists would consider noise within the records and irresolvable between climatic and nonclimatic causes.

The situation now is very different than when disagreements between interpretations were initially stated. Then, the pollen evidence was unequivocally for a climatic cooling of similar magnitude and timing to the Younger Dryas in the North Atlantic. Currently, even proponents for climate change argue that it is subtle and that it begins 500–1 200 yr before the Younger Dryas. Moreno (1997), Moreno et al. (1999), and Hajdas et al. (2003) still consider the cooling to be part of a globally synchronous event but Ashworth and Hoganson (1993), Markgraf (1993), and Bennett et al. (2000) find no evidence to support climatic reversals at the time of the Younger Dryas (11 000–10 000 yr BP) or during the Antarctic Cold Reversal (14 500–12 900 yr BP).

See also: **Beetle Records: Late Tertiary and Early Quaternary Records; Overview. Pollen Records, Late Pleistocene: South America.**

References

- Arias ET (2000) *Coleópteros de Chile*. Santiago de Chile: Fototeknika.
- Ariztegui D, Bianchi MM, Massaferro J, Lafargue E, and Niessan F (1997) Interhemispheric synchrony of late-glacial climatic instability as recorded in proglacial Lake Mascardi, Argentina. *Journal of Quaternary Science* 12: 133–138.
- Ashworth AC (2001) Perspectives on Quaternary Beetles and Climate Change. In: Gerhard L, Harrison W, and Hanson B (eds.) *American Association of Petroleum Geologists Studies in Geology, Geological Perspectives of Global Climate Change*, vol. 47, pp. 153–168. American Association of Petroleum Geologists, Tulsa, Oklahoma.
- Ashworth AC and Hoganson JW (1987) Coleoptera bioassociations along an elevational gradient in the lake region of southern Chile and comments on the postglacial development of the fauna. *Annals of the Entomological Society of America* 80: 865–895.
- Ashworth AC and Hoganson JW (1993) Magnitude and rapidity of the climate change marking the end of the Pleistocene in the mid-latitudes of South America. *Palaeogeography, Palaeoclimatology, Palaeoecology* 101: 263–270.
- Ashworth AC and Kuschel G (2003) Fossil weevils (Coleoptera:Curculionidae) from latitude 85° S Antarctica. *Palaeogeography, Palaeoclimatology, Palaeoecology* 191: 191–202.
- Ashworth AC and Markgraf V (1989) Climate of the Chilean channels between 11 000 and 10,000 yr BP based on fossil beetle and pollen analyses. *Revista Chilena de Historia Natural* 62: 61–74.
- Ashworth AC, Markgraf V, and Villagrán C (1991) Late Quaternary climatic history of the Chilean channels based on fossil pollen and beetle analysis, with an analysis of the modern vegetation and pollen rain. *Journal of Quaternary Science* 6: 279–291.
- Bennett KD, Haberle SG, and Lumley SH (2000) The last Glacial–Holocene transition in southern Chile. *Science* 290: 325–328.
- Buckland PC and Hammond PM (1997) The origins of the biota of the Falkland Islands and South Georgia. In: Ashworth AC, Buckland PC, and Sadler JP (eds.) *Quaternary Proceedings 5: Studies in Quaternary Entomology—An Inordinate Fondness for Insects*, pp. 59–66. Chichester: Wiley.

- Denton GH, Lowell TV, Heusser CJ, et al. (1999) Geomorphology, stratigraphy, and radiocarbon chronology of Llanquihue drift in the area of the southern Lake District, Seno Reloncavi, and Isla Grandé de Chilo., Chile. *Geografiska Annaler* A81: 167–229.
- Dillehay TD and Pino M (1989) Stratigraphy and chronology. In: Dillehay TD (ed.) *Monte Verde – A late Pleistocene Settlement in Chile*, pp. 133–145. Washington: Smithsonian Institution Press.
- Elias SA (1994) *Quaternary insects and their environments*. Washington: Smithsonian Institution Press.
- Haberle SG and Bennett KD (2004) Postglacial formation and dynamics of North Patagonian rainforest in the Chonos Archipelago, southern Chile. *Quaternary Science Reviews* 23: 2433–2452.
- Hajdas I, Bonani G, Moreno PI, and Ariztegui D (2003) Precise radiocarbon dating of late-glacial cooling in mid-latitude South America. *Quaternary Research* 59: 70–78.
- Heusser CJ (1974) Vegetation and climate of the southern Chilean lake district during and since the last interglaciation. *Quaternary Research* 4: 290–315.
- Heusser CJ and Rabassa J (1987) Cold climatic episode of Younger Dryas in South America. *Nature* 328: 609–611.
- Hoganson J, Gunderson M, and Ashworth A (1989) Fossil-beetle analysis. In: Dillehay TD (ed.) *Monte Verde – A late Pleistocene settlement in Chile*, pp. 211–226. Washington: Smithsonian Institution Press.
- Hoganson JW and Ashworth AC (1992) Fossil beetle evidence for climatic change 18,000–10,000 years BP in South-Central Chile. *Quaternary Research* 37: 101–116.
- Hollin JT and Schilling DH (1981) Late-Wisconsin-Weichselian mountain glaciers and small ice caps. In: Denton GH and Hughes TJ (eds.) *The Last Great Ice Sheets*, pp. 179–206. New York: Wiley.
- Hubbard A, Hein AS, Kaplan MR, Hulton NRJ, and Glasser N (2005) A modelling reconstruction of the late glacial maximum ice sheet and its deglaciation in the vicinity of the Northern Patagonian Icefield, South America. *Geografiska Annaler* A87: 375–391.
- Markgraf V (1984) Late Pleistocene and Holocene vegetation history of temperate Argentina: Lago Morenito, Barailoche. *Dissertationes Botanicae* 72: 235–254.
- Markgraf V (1993) Younger Dryas in southernmost South America – An update. *Quaternary Science Reviews* 12: 351–355.
- Markgraf V, Bradbury JP, Schwalb A, et al. (2003) Holocene palaeoclimates of southern Patagonia: Limnological and environmental history of Lago Cardiel, Argentina. *The Holocene* 13: 597–607.
- Massaferro J and Brooks SJ (2002) Response of chironomids to late Quaternary environmental change in the Taitao peninsula, southern Chile. *Journal of Quaternary Science* 17: 101–111.
- Mercer JH (1976) Glacial history of southernmost South America. *Quaternary Research* 6: 125–166.
- Moreno PI (1997) Vegetation and climate near Lago Llanquihue in the Chilean Lake district between 20,200 and 9500 14C yr BP. *Journal of Quaternary Science* 12: 485–500.
- Moreno PI, Lowell TV, Jacobson GL Jr., and Denton GH (1999) Abrupt vegetation and climate changes during the last glacial maximum and last termination in the Chilean Lake District: A case study from Canal de la Puntilla (41° S). *Geografiska Annaler* A81: 285–311.
- Morrone JJ (1993) Revisión sistemática de un nuevo género de Rhytirrhini (Coleoptera, Curculionidae), con un análisis biogeográfico del dominio Subantártico. *Boletín de la Sociedad de Biología de Concepción, Chile* 64: 121–145.
- Premoli AC, Souto CP, Rovere AE, Allnut TR, and Newton A (2002) Patterns of isozyme variation as indicators of biogeographic history in *Pilgerodendron uviferum* (D. Don) Florin. *Diversity and Distributions* 8: 57–66.
- Villagrán C (1990) Glacial climates and their effects on the history of the vegetation of Chile: A synthesis based on palynological evidence from Isla de Chiloé. *Review of Palaeobotany and Palynology* 65: 17–24.
- Whitlock C, Bartlein PJ, Markgraf V, and Ashworth AC (2001) The mid-latitudes of North and South America during the last glacial maximum and early Holocene: Similar paleoclimatic sequences despite differing large-scale controls. In: Markgraf V (ed.) *Interhemispheric Climate Linkages*, pp. 391–416. Cambridge: Cambridge University Press.

Late Pleistocene of New Zealand

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Introduction

Fossil beetle reconstructions are relatively new in New Zealand (NZ) with the first published work dating only to 2000. Before 2000, terrestrial paleoenvironmental reconstructions relied mainly on pollen records, which defined regional vegetation patterns for the late Quaternary, from which broad climate inferences were drawn. Quantitative estimates for past climate remained less well defined because there is no clear relationship between temperature and vegetation. Hence, there was a need for additional proxies and particularly for proxies capable of quantifiable paleoclimate reconstruction.

To date, Quaternary beetle assemblages have been recorded from 14 NZ sites (Table 1; Figure 1) ranging in age from 1.3 Ma to late Holocene. This literature includes examining the persistence of beetle species in NZ since the early-mid-Quaternary, the development of a beetle-based climate estimation model, and its application to sites dating from the last glaciation. Paleoecological and biogeographical reconstructions have been made for each site, detailing ecosystems at a local scale and at community and species levels. The following is a review of this literature, prefaced by a summary of present knowledge of the modern beetle fauna and a brief description of the NZ physical environment, identifying factors that affect beetle distribution.

Reference to the Last Glacial Maximum (LGM) follows the definition of Newnham et al. (2007) for an extended Last Glacial Maximum (eLGM) in NZ, 29–19 kcal years BP, which is longer and earlier than the LGM *sensu stricto*.

Background

Modern Fauna

Knowledge of the modern fauna underpins both fossil identification and the ecological controls on distribution that are the basis for paleoclimate reconstruction. The NZ beetle fauna is relatively rich: approximately 5223 native and 356 introduced species have been described, comprising approximately 1100 genera in 82 families (Klimaszewski and Watt, 1997). The actual number of species is unknown but is certain to be significantly higher. The Lynfield survey (Kuschel, 1990), regarded as the most comprehensive survey so far in NZ, identified approximately 1000 species. As this is thought to be about 8% of the total NZ beetle fauna, it suggests the total number is around 10000–10500 species (Klimaszewski and Watt, 1997). Consequently, the 5223 species held in NZ collections represent about half the total NZ beetle fauna but several groups need revision. Endemism is high at most levels, with one endemic family, several endemic subfamilies, high but unquantified endemism of genera and about 90% species endemism. A summary of the dominant families and the number of described species in each of these families is listed in Table 2.

NZ Setting

NZ geography imposes many barriers to the dispersal of insects and the tracking of ecosystems across the landscape in response to climate change. It is an island archipelago focussed on two main islands (North Island, South Island), and has a land area of ~270 550 km². The South Island is dominated by an axial mountain range, the Southern Alps, extending about 500 km along the western side of the island, with peaks reaching over 3000 m. During the eLGM the Southern Alps developed a piedmont icecap that extended the length of the range. Glaciers advanced to the south-western coast of the South Island, while to the east they extended beyond the range fronts onto inland basins and coastal outwash fans. A barrier to biotic dispersal during interglacials is the separation of the North and South Islands by the 32 km-wide Cook Strait. However, during glaciations when sea level was lower, a land bridge existed.

Quaternary Stasis in NZ Beetles

Early and Mid-Pleistocene Beetle Assemblages

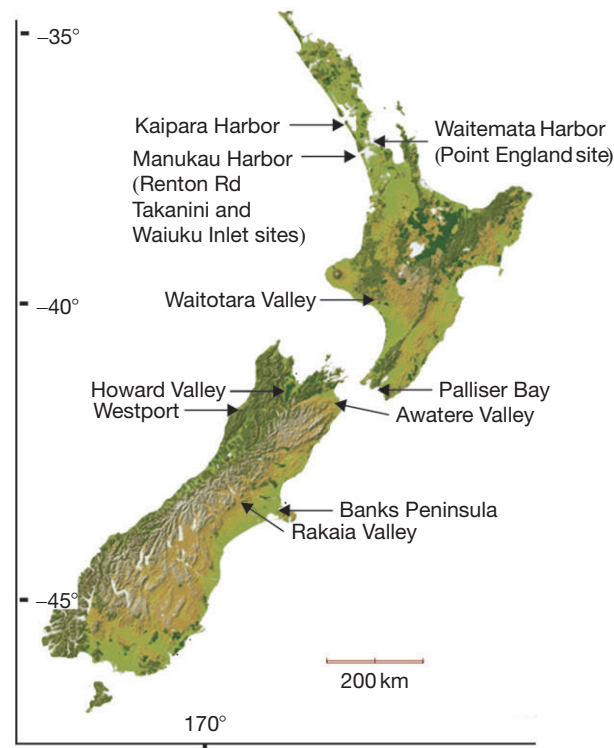
Background

Quaternary stasis is the foundation of fossil beetle research because if the fossil specimens represent extant species, their biology and habitat requirements can be observed from their modern counterparts. It is thought that this morphological stasis is a result of Milankovich oscillations where glacial–interglacial climate change continually changed directions of selection pressures and speciation only occasionally occurred through isolation (Bennett, 2004). This is referred to by Coope (2004) as ‘stirring of the gene pool.’ At least this is the case in continental locations where there was high amplitude of climate change and where species had very good dispersal opportunities. NZ provided the opportunity to test this theory of stasis in an island archipelago situation where Quaternary climate fluctuations were relatively subdued and where dispersal options were very limited.

The early-mid-Pleistocene fossils were obtained from a series of coastal outcrops containing ancient forests each buried by volcanic deposits (Figure 2). The stratigraphies and tephra chronologies are described (Alloway et al., 2004) and sites range in age from 1.3 My to 170 000 years old. They are all located in the Auckland Region (Figure 1). This location was ideal to test Quaternary stasis in NZ beetles because the area remained forested throughout glacial cycles and the temperature change between glacial and interglacial conditions was only about 4–5 °C (Sandiford et al., 2003). This location was also ideal because the modern fauna is very well understood and documented (Kuschel, 1990), providing the background comparisons for fossil assemblages for the early-mid-Quaternary assemblages. The investigation was an important step for fossil beetle research in NZ because there is also some evidence for extinctions in some locations (Marra and Leschen, 2004).

Table 1 Summary of the New Zealand fossil beetle sites

Site location	Age of site	Number of fossil taxa	Reference
Awatere Valley	Mid-late MIS 1	80	Marra and Leschen (2004) and Marra et al. (2004)
Awatere Valley	MIS 2	26	Marra and Leschen (2004) and Marra et al. (2004)
Howard Valley	MIS 3–2	54	Marra and Thackray (2010)
Rakaia Valley	MIS 2	29	Marra et al. (2006a)
Waitotara Valley	Late MIS 3	29	Marra et al. (2009)
Kaipara Harbor	Early MIS 3	14	Marra (unpublished data)
Westport	MIS 3–2	76	Burge and Shulmeister (2007a)
Westport	MIS 5–4	42	Burge and Shulmeister (2007b)
Palliser Bay	Last Interglacial	64	Marra (2003a)
Banks Peninsula	Late MIS 6	19	Marra (2003b)
Manukau Harbor (Renton Rd)	Late MIS 7	98	Marra et al. (2006b)
Takanini (Manukau Harbor)	~0.5 Ma	5	Marra and Leschen (2011)
Point England (Waitemata Harbor)	~1 Ma	6	Marra and Leschen (2011)
Waiuku (Manukau Harbor)	~1.3 Ma	31	Marra and Leschen (2011)
	Total	573	

**Figure 1** Map of New Zealand showing location of research sites mentioned in the text.

The fossil evidence

The fossil beetles from the early–mid-Pleistocene all represent extant species found in the modern regional fauna. They indicate that stasis was the dominant response of NZ beetles during the Quaternary. For example, fossils from the Waiuku Inlet site in the Auckland Region, aged about 1.3 Ma, indicate a coastal swamp forest and include species such as *Cafius algophilus* (Staphylinidae) and *Pericoptus* (either *P. stupidus* or *P. frontalis* (Scarabaeidae)) which are found today in sandy coastal habitats. In addition, three weevil species were

Table 2 List of eight dominant families of beetles in New Zealand and the number of described species in each family (Klimaszewski and Watt, 1997)

Family in order of dominance	Number of described NZ species
Curculionidae (incl. Scolytidae)	1542
Staphylinidae (incl. Pselaphidae)	1021
Carabidae	445
Scydmaenidae	202
Colydiidae	196
Cerambycidae	188
Chrysomelidae	153
Scarabaeidae	144

found: *Microcryptorhynchus perpusillus* (Curculionidae) that lives in the dead branches of trees, *Stenotrupis debilis* (Curculionidae) that lives in tree ferns in swamps, and *Dysnocryptus inflatus* (Anthribidae) that inhabits and feeds on grasses (Marra and Leschen, 2011).

The best example of stasis is from a fossil assemblage aged about 170 000 years, also located in the Auckland Region, which also shows stasis at the community level. This assemblage is from a paleo-forest that was instantaneously buried by an explosive phreatomagmatic eruption (Marra et al., 2006b). Ninety-eight fossil beetle taxa belonging to 28 families were identified, all of which belong to the modern fauna. Fossils were very well preserved and include species such as *Inosomus rufopiceus* (Curculionidae; Figure 3(a)) which inhabits *Agathis australis* (Araucariaceae) and *Dacrydium cupressinum* (Podocarpaceae), and *Heterotyles argentatus* (Curculionidae; Figure 3(b)), a species that inhabits various podocarp trees. The assemblage indicates that the same fauna remained in place or reassembled in the same place across two glacial cycles.

Despite this faunal stability in northern NZ, extinction rates in other locations in NZ may have been higher than in Holarctic locations. Of the ~570 NZ fossil taxa thus far identified (Table 1), three are possible extinctions. These fossils are distinctive and well preserved, but given the number of undescribed species in the modern fauna, it is difficult to say

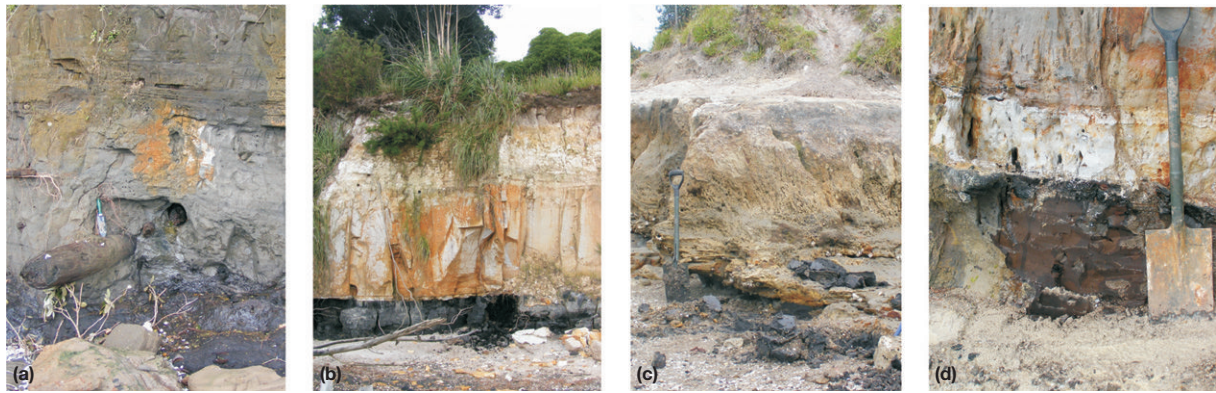


Figure 2 Outcrops of ancient forests that are buried by volcanic ash deposits that provided the early–mid-Pleistocene fossil beetles: (a) Renton Rd (ca. 170 ka BP); (b) Takanini (0.55 My BP); (c) Point England (ca. 1 My BP); and (d) Waiuku Inlet (1.3 My BP).

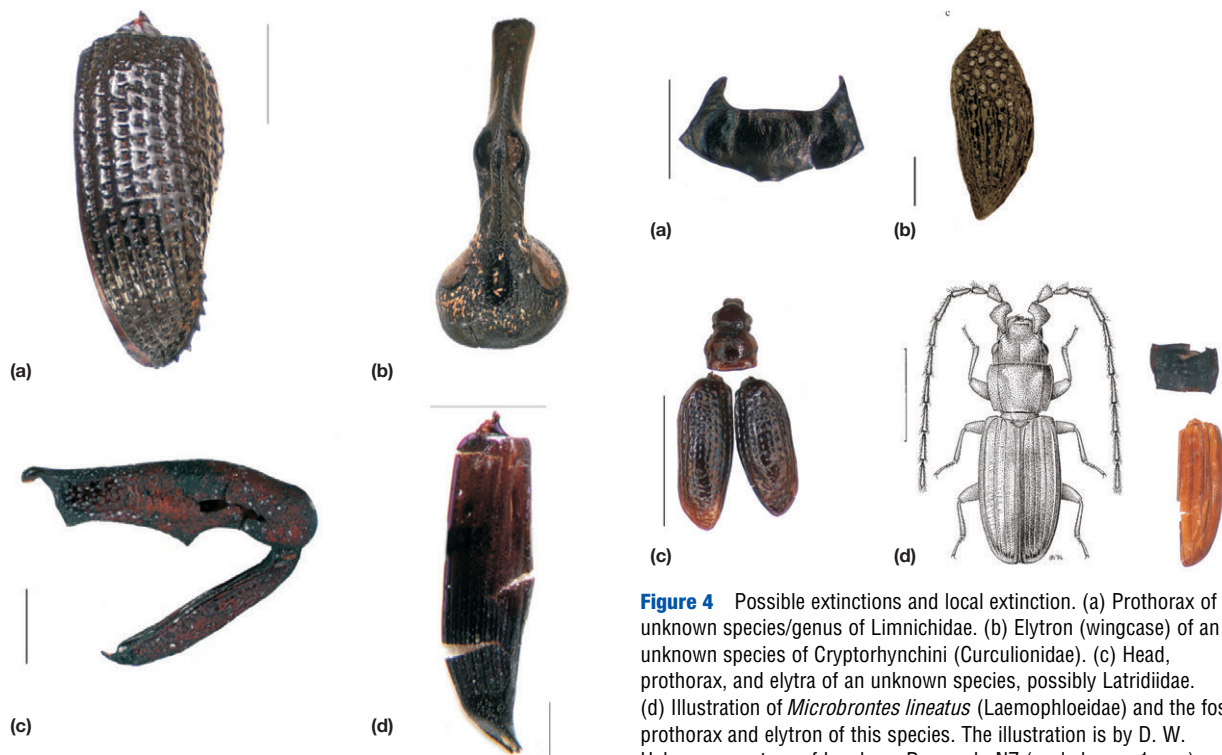


Figure 3 Examples of fossil beetles from New Zealand: (a) *Inosomus rufopiceus* (Curculionidae); (b) *Heterotyles argentatus* (Curculionidae); (c) *Ectopsis ferrugalis* (Curculionidae); and (d) *Platypus caviceps* (Curculionidae).

Figure 4 Possible extinctions and local extinction. (a) Prothorax of an unknown species/genus of Limnichidae. (b) Elytron (wingcase) of an unknown species of Cryptorhynchini (Curculionidae). (c) Head, prothorax, and elytra of an unknown species, possibly Latridiidae. (d) Illustration of *Microbrontes lineatus* (Laemophloeidae) and the fossil prothorax and elytron of this species. The illustration is by D. W. Helmore, courtesy of Landcare Research, NZ (scale bars = 1 mm). Images by B. Rhode of Landcare Research (scale bars = 1 mm).

whether the unidentified fossils belong to extinct species or represent extant taxa that have yet to be collected. One of these is a fossil prothorax from a last Interglacial assemblage that is clearly in the family Limnichidae. The prothorax has a distinctive oblique ridge cutting across each posterior pronotal angle (Figure 4(a)), a feature that is not found on any of the known modern Limnichidae in NZ or Australia. Similarly, four distinctively ornamented weevil elytra (Figure 4(b)) from a marine isotope stage (MIS) 6 site in the eastern South Island are also unknown in the modern fauna.

The third possible extinction is a fossil specimen in the family Latridiidae (Figure 4(c)). It is abundant as a fossil in glacial assemblages and is probably a subalpine species. Although all the major fossil sclerites have been collected, providing good detail for its identification, it has no known modern equivalent.

Extinctions of NZ beetles are well documented for the time since the arrival of predators within the past 2000 years and human settlement ~750 years ago, and the subsequent destruction of forest and shrubland habitats. For example, the large weevil *Ectopsis ferrugalis* (Curculionidae; Figure 3(c)) occurs as fossils in South Island assemblages but it is not found today in the South Island. Its recent absence has been

attributed to introduced predators (Kuschel and Worthy, 1996). This species is important as a fossil because it has a specific host tree and its larvae have specific tree-size burrowing requirements, which not only identify the presence of the host tree, but document its size as well. In some cases, the fossil record is the only evidence of some taxa. For example, a large (24.5 mm) extinct beetle species in the family Ulodidae, found in cave deposits, has been described solely from fossil specimens (Leschen and Rhode, 2002). Its extinction is also attributed to introduced predators.

Prior to human arrival, the fossil data from northern NZ indicate stability in beetle communities over much of the Quaternary. However, the stability of these northern regional assemblages is not reflected in assemblages for the remainder of NZ where, during ice ages, the landscape was dominated by glaciation and vegetation changes were severe. Fossil data from central and southern NZ show that very different communities occurred in these locations in response to changes in climate and vegetation. A very good example is *Microbrontes lineatus* (Laemophloeidae; Figure 4(d)). This species is very rare in modern collections, yet it occurs as a fossil in two locations outside of its modern distribution, which may indicate it may have been more abundant in the past.

Late Pleistocene Beetle Distributions

Changing Distributions

The penultimate glaciation (MIS 6)

The modern beetle fauna of Banks Peninsula (Figure 1) is well known from several published natural histories and inventories (e.g., Johns, 1986). These inventories provide a basis for comparison with a fossil assemblage from Gebbies Valley, Banks Peninsula.

Nineteen beetle fossils extracted from core material are reported in Marra (2003b). The fossil assemblage comprised montane forest taxa in a coastal setting. Based on the modern distributions and upper altitudinal limits of the fossil fauna, Marra (2003b) interpreted the climate as interstadial rather than pleniglacial, and by using a simple environmental lapse rate (0.55 ± 0.05 °C/100 m), she estimated the lowering of the montane zone by ~450 m, from its present elevation down to sea level. This equates to a maximum temperature depression of around 2–3 °C.

Originally, the site was described from pollen evidence as a back barrier, freshwater setting (Shulmeister et al., 1999; Soons et al., 2002). However, the beetle fossils indicate the paleoenvironment was estuarine (Marra, 2003b). The pollen-based description of the surrounding vegetation is of low-forest shrubs (Soons et al., 2002). The fossil beetle *Pachyrinus metallicus* (Belidae), that is host specific to *Podocarpus totara* and *Podocarpus hallii* (Podocarpaceae) provided confirmation of this forest and identified the forest type. Both *P. totara* and *P. hallii* grow on Banks Peninsula but *P. hallii* dominates the higher elevations, so it was probably this species that occupied the lowland site during the cooler MIS 6, as part of the low forest described from the pollen spectra.

The MIS 6 fossil fauna is very different from the modern Banks Peninsula fauna. Of the nineteen beetle species identified, only three weevils (*P. metallicus*, *Eucossonus setiger*,

Cecyropa modesta) are found in the modern Banks Peninsula fauna, although undetermined carabids and species of *Nestrius* and *Bledius* could also be present. Species of *Epichorius* (Byrrhidae) *Spinditeles* (Anobiidae) *Mandalotus*, *Baeosomus*, *Microcryptorhynchus*, or *Scelodolichus* (Curculionidae) in the fossil assemblage are not part of the modern Banks Peninsula fauna. The different faunal composition of MIS 6 fauna is a result of the montane environmental conditions at the site at that time.

The last interglaciation (MIS 5)

The fossil beetle fauna from Palliser Bay on the southern coast of the North Island (Figure 1) (Marra, 2003a) is markedly different from the modern assemblage. The site dates from the last interglaciation, MIS 5. Two fossil beetle assemblages totalling 63 taxa were identified from two organic silt beds occurring in a sequence of fluvial and lacustrine sediments. Both assemblages are similar in that they indicate ecological successions from lake margins to forest but they comprise different taxa. Of the two fossiliferous layers, the lower contained beetles that are found today some 600–700 km north of the study site (Figure 5). This fauna represents a significant southward expansion of northern species compared to modern distributions, despite the physical barrier of the central North Island volcanic plateau. This is a minimum southward dispersal because Cook Strait would have impeded further movement to the south.

Pre-eLGM (late MIS 3)

Fossil beetles from a last glacial landslide-generated lake-forest sequence aged 33 480–34 410 cal years BP (late MIS 3) were identified from the Waitotara Valley (Marra et al., 2009). Information on ancient ecosystems from this site is important because it is located in a region that is characterized by low endemism, low diversity, and disjunctive distributions. The region is described as an ecological gap, and referred to as the Manawatu Gap that is located within a wider ecological gap, the Northern Gap. Many taxa that are otherwise widespread in NZ are absent from this gap. There is much debate about the cause and timing of the Northern Gap. One theory is that it is a result of long term landform evolution dating back to the Oligocene; another theory is that the Gap in distributions is a result of the last glaciation. Therefore, a fossil beetle site from early MIS 3 within the Gap provided an excellent opportunity to examine the antiquity of the Gap.

The Waitotara Valley is characterized by numerous large-scale deep-seated landslides, some of sufficient magnitude to block rivers and form lakes. The fossils were obtained from one of these lakes where the stratigraphy records a lake-forest sequence. Twenty-nine fossil beetle taxa were identified and include such species as *Platypus caviceps* (Curculionidae; Figure 3(d)), which is a timber-boring weevil that inhabits stressed or freshly felled *Nothofagus* (Nothofagaceae) trees. The beetles tunnel into the wood and cultivate the fungi on which they feed (May, 1993).

Of major biogeographical interest is that none of the fossil beetle species are present in the modern regional fauna and many taxa such as *P. caviceps*, *Alema paradoxa*, *Rhyzobius consors* (Coccinellidae), *Syrphetodes ater* (Ulodidae), *Cyclaxyra impressa* (Phalacridae), and species of *Grynoma* (Trogossitidae) and

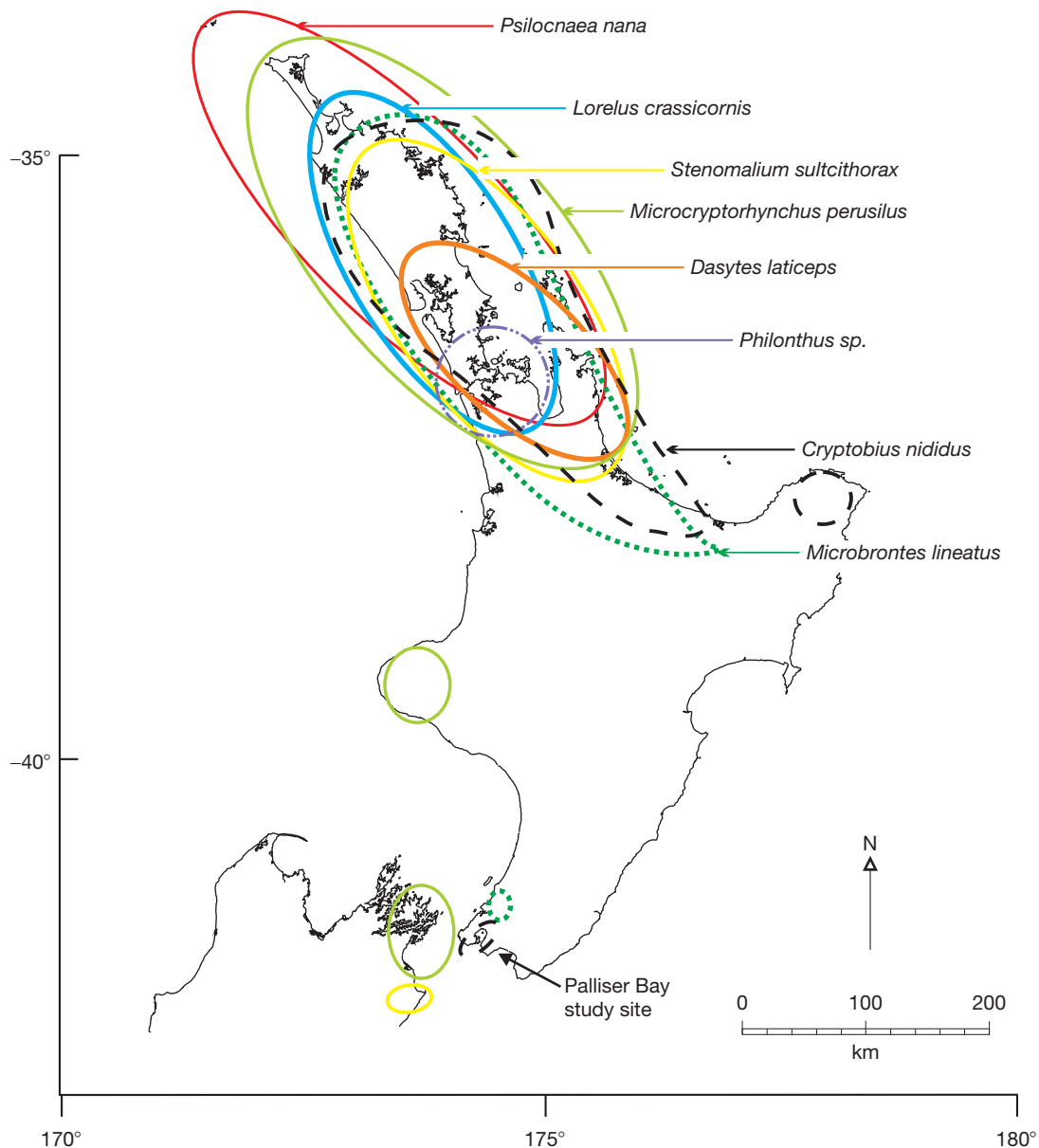


Figure 5 The modern-day distribution of species as they relate to the fossil site at Palliser Bay. Adapted from Marra MJ (2003) A last interglacial beetle fauna from New Zealand. *Quaternary Research* 59: 122–131.

Pycnomerus (Zopheridae) are now absent from part or all of the lower North Island. These results suggest that the modern biogeographical pattern stems from the last glaciation. There is no obvious ecological or habitat reason for these taxa to be absent in the modern fauna. **Figure 6** illustrates the Northern Gap and the modern distribution of the late MIS 3 fossil species.

eLGM (MIS 2)

Fossil beetle assemblages from the eLGM at Nina Stream in the Awatere Valley (**Figure 1**) in the north-eastern South Island have been reported in **Marra and Leschen (2004)**. The Awatere Valley flows northeast along the northern side of the Inland Kaikoura Range (2880 m). At the site, the Awatere fault transects the stream. This is one of the series of active faults that continue the NZ plate boundary into the North Island.

Movement on the fault impedes stream drainage, which led to the accumulation of a sequence of fan gravels intersected by organic silts. Radiocarbon ages for organic beds at the base of the section are 25 550–23 950 and 24 650–22 050 cal years BP.

The fossil fauna includes subalpine beetles from habitats that include stream side, tussocks, and shrubs. The fauna includes the ground beetles (Carabidae), *Notagonum feredayi* and *Bembidion hokitikense* which today are found near water in tussock grasslands. Other subalpine species include *Peristoreus fusconotatus* (Curculionidae) that lives on *Olearia* (a genus of tree daisies), and a species of *Lyperopais* (Curculionidae) that so far has only been collected above treeline. Both *Acritelater reversus* (Elateridae) and *Aleochara hammondi* (Staphylinidae) also inhabit subalpine tussock grasslands. The fauna suggests a downward shift in the regional vegetation zones of several

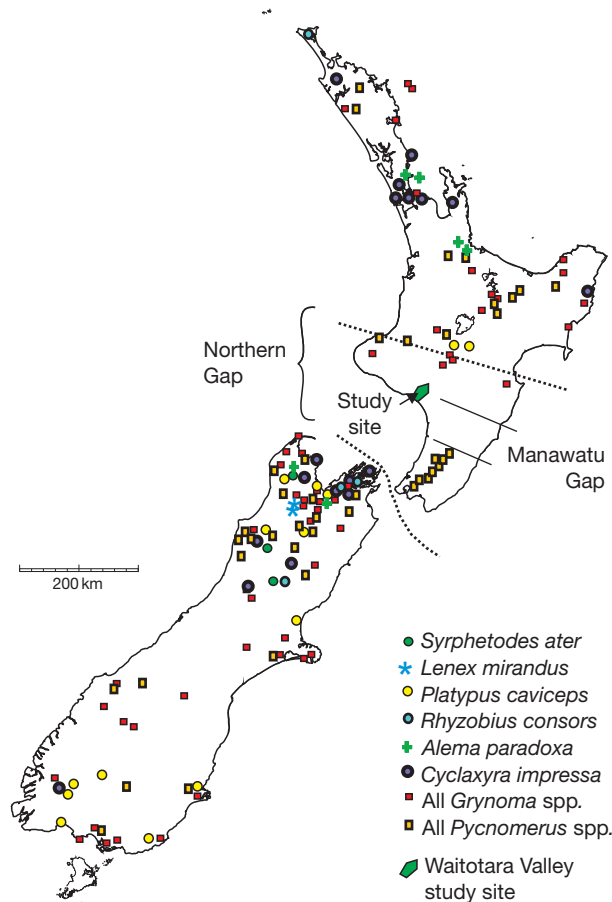


Figure 6 The modern distribution of fossil taxa from the Waitotara Valley site showing the modern disjunctive distributions in the Northern Gap and particularly within the Manawatu area. Adapted from Marra M, Crozier M, and Goff J (2009) Paleoenvironment and biogeography of a late MIS 3 fossil beetle fauna from South Taranaki, New Zealand. *Journal of Quaternary Science* 24: 97–107.

hundred meters, presumably in response to the cold climate conditions (see climate reconstruction given below).

Quantifying Past Climates

Method for Climate Quantification

Method development

A maximum likelihood envelope (MLE) method has been developed (Marra et al., 2004) that is suitable for quantifying past climates from beetle fossils in NZ. MLE was developed to deal with problems associated with the incomplete knowledge of the NZ fauna. Incomplete knowledge limits robust climatic reconstruction because if the observed distribution of species is poorly known, the actual distribution of the ecological tolerances may be underestimated.

The MLE method is based on the premise that beetle distributions are climate driven and that the distribution of each taxon is controlled by finite physiological requirements. It therefore uses a sine distribution model rather than a Gaussian distribution model because a Gaussian model does not

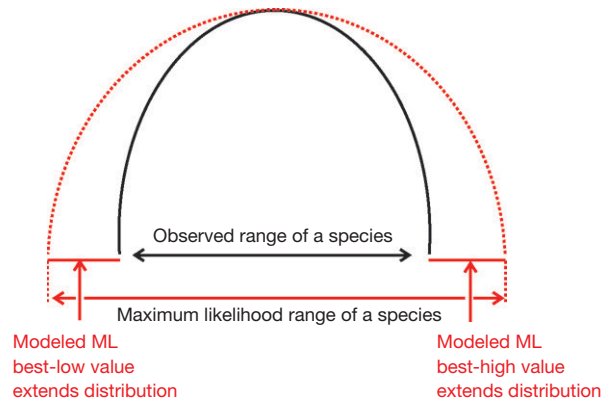


Figure 7 Maximum likelihood (ML) best-high and best-low endpoints attached to a sine distribution model, effectively increasing the climate range of a species.

prescribe absolute limits to distributions needed for the presence–absence approach intrinsic to beetle climate modeling (e.g., mutual climate range – MCR; Atkinson et al., 1987). Maximum likelihood (ML) best-high and best-low values are determined and fitted to sine distributions to allow for the uncertainty that the samples are realistic representations of the actual climatic distributions. Figure 7 illustrates how the estimated ML endpoints effectively increase the distribution of a taxon. The endpoints define a 95% confidence distribution.

Climate variables for reconstruction

The climate reconstructions are based on mean February and mean minimum July temperatures. Although MCR is based on summer temperature and seasonality, seasonality is much more subdued in NZ than in continental locations, so it is unlikely that this is a major climatic consideration. Anecdotal evidence suggests that length of summer and extreme winter temperatures may drive beetle distributions and hence, mean February and minimum July temperatures.

These climate variables were tested (Marra et al., 2004) by pooling the distribution data points of all the species. Each datum (x_i) was transformed to $(x_i - a)/(b - a)$ to give each species' actual distribution and relative position on a standard range (0,1) of all the data points (all the distribution points of all the species; Figure 8).

The test showed that the sine model fitted the July minimum temperature rather well, but the February mean temperature pooled data were significantly positively skewed. As the same skewness is not apparent in the July minimum temperature data as expected, it suggests that, aside from temperature, some other climate variable was affecting beetles significantly in summer but not in winter, and this may be related to moisture.

eLGM Climate Reconstruction

Lyndon Stream

Using MLE, the temperature for an interval during the eLGM was estimated from fossil beetles from the Lyndon Stream (Marra et al., 2006a). Lyndon Stream is located in the eastern foothills of the Southern Alps (alt. ~700 m; Figure 1) proximal

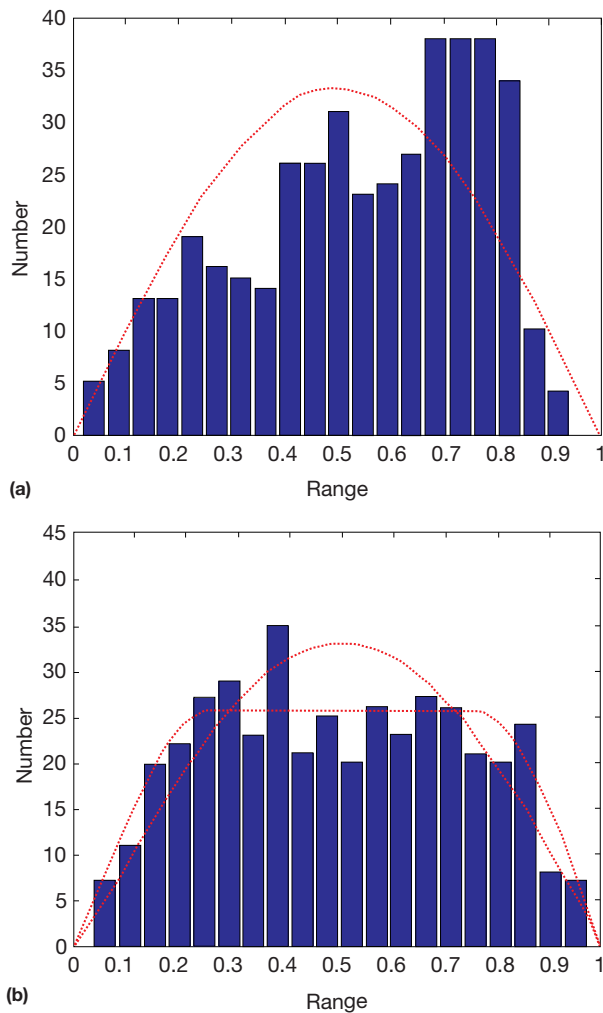


Figure 8 Graphs showing the fit of sine models to the climate variables for mean February temperature (a) and mean July minimum temperature (b). Pooled mean February data and best-fitting sine model and bi-sine model (a) and pooled mean July minimum temperature data and two sine models; sine and sine-tapered boxcar (b). Adapted from Marra MJ, Smith EGC, Shulmeister J, and Leschen R (2004) Late Quaternary climate change in the Awatere Valley, South Island, New Zealand using a sine model with a maximum likelihood envelope on fossil beetle data. *Quaternary Science Reviews* 23: 1637–1650.

to, and a tributary of the Rakaia river valley, which contained a major glacier during the eLGM. The site comprises lake beds bounded by eLGM proximal outwash, confining the lake within the eLGM chron (Figure 9). The site was first examined by Soons and Burrows (1978) who dated the paleolake to between 24 840–26 100 cal years BP (base of the section) and 21 980–23 280 cal years BP (top of the section). Using pollen and plant macrofossils, they described an environment during the eLGM that was nearly identical to modern conditions.

Nine fossil beetle taxa were used to reconstruct mean February (summer) and mean minimum temperatures for July (the coldest month; Marra et al., 2006a).

The February temperature reconstruction is illustrated in Figure 10. The broken line is the modern temperature at the site, and horizontal bars are the temperature ranges for the



Figure 9 The Lyndon Stream site lake beds bounded by extended Last Glacial Maximum (eLGM) proximal outwash.

taxa. The shadings at the end of each bar are the ML best-high and best-low statistical endpoints. The envelopes (yellow or shaded) are the mutual temperature overlaps of all the taxa, which are defined at the lower limit by the species with the highest minimum, and the upper limit by species with the lowest maximum. The yellow shows the overlap of the known temperature ranges, whereas the shaded envelope is the MLE estimate that has statistically extended the known ranges. The estimates indicate that the mean temperature at the site at eLGM was between 0.5 °C warmer and 1.9 °C cooler in February than at the present day. The modeled mean minimum winter temperature was similarly close to present values, being between 1 °C warmer and 2.2 °C cooler than present day.

Even at the cooler extremes of these modeled estimates, the climate during the eLGM appears to have been milder than that of the traditional view, for which the temperature depression has been estimated at 4–7 °C (e.g., Porter, 1975; Soons, 1979). However, the predictions from beetle environmental requirements fit remarkably well with the reconstructions from plant macrofossils and are consistent with new evidence of relatively mild conditions (Hormes et al., 2003) during at least some periods within the eLGM in NZ.

Awatere Valley

In contrast, the MLE climate reconstruction for the Awatere Valley, ~300 km to the north (Marra et al., 2004), shows a very much cooler climate during the same time (24 650–22 050 and 25 550–23 950 cal years BP) during eLGM. At this site, summer (February) mean temperature was ~3.5–4 °C cooler, and July (winter) mean daily minimum temperature was ~4–5 °C cooler than present day temperatures. The ML estimates broadened the reconstructed temperature ranges to 2.5–5 °C cooler for February temperatures and 3.5–6.0 °C cooler for mean minimum daily temperature of the coldest month (July). These estimates are more consistent with the traditional cooler eLGM model of 4–7 °C cooler.

The conflict between sites of apparently similar age needs to be resolved. It is most likely to be a result of poorly constrained ages. Tighter age control may reveal that these sites are close but not identical in age, and that the different environmental estimates reflect rapid climate change from cold glacial to warm interstadial such as described above at the Lyndon

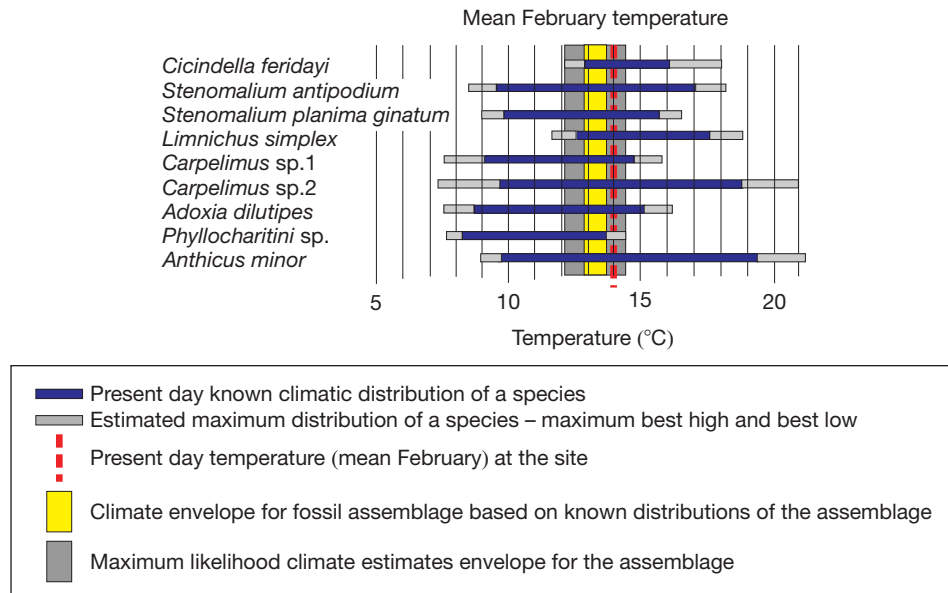


Figure 10 Application of maximum likelihood envelope to reconstruct the mean February temperature for the eLGM at Lyndon Stream showing the difference between the present-day temperature at the site to the estimated temperature at the eLGM. Adapted from Marra MJ, Shulmeister J, and Smith E (2006) Last glacial maximum beetle fauna, Lyndon Stream, Rakaia River Valley, South Island, New Zealand. *Quaternary Science Reviews* 25: 1841–1849.

Stream site. Another possibility is that the Lyndon Stream site plots are warmer because they are sheltered from cold southerly and easterly winds.

Reconstructions of rates of climate change have not yet been attempted. The ability to determine rates of climate change and to identify short-term climatic oscillations is the real strength of fossil beetle analyses and these should be investigated. There is also no literature on the Postglacial warming, an interval of rapid climate change about which there are many conflicting proxy reconstructions.

Forest Refugia in NZ During the eLGM

Fossil Beetle Evidence for Forest Refugia

Awatere Valley

Late Pleistocene beetle research has highlighted the importance of scale in reconstructions (e.g., Marra and Leschen, 2004). It is evident that pollen provides regional-scaled vegetation information, but the beetles provide local-scale insights. For example, pollen data indicate that the vegetation in the eastern South Island was predominantly grassland from 26 000 to 12 000 years BP (Moar, 1980). Low values of tree pollen indicated that trees were present throughout this period, although the actual source of the pollen remains a mystery. The fossil beetle assemblages from several sites precisely indicate the location of one of these glacial-aged forest patches/refugia. The first and most exciting of these discoveries was that of the forest-dwelling beetles in the Awatere Valley fauna, the most significant of which was the weevil *Hypotagea lewisi* (Curculionidae; Figure 11(a)) which is host-specific on living *Nothofagus* trees, both the larvae and adults feeding on *Nothofagus* parenchyma tissue.

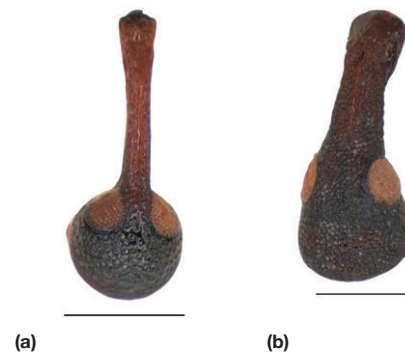


Figure 11 (a) The head capsule of *Hypotagea lewisi* (Curculionidae). (b) Head capsule of *Rhopalomerus tenuirostris* (Curculionidae). Images by B. Rhode of Landcare Research (scale bars = 1 mm).

The fossil of *H. lewisi* pinpoints a stand of *Nothofagus* forest in Marlborough during the eLGM, the first confirmed eastern forest refugium for this time. No *Nothofagus* pollen was found at the site but recent additional (unpublished) data include additional beetle taxa associated with *Nothofagus* plus fossil leaves of *Nothofagus solandri*, both of which confirm the original interpretation of this forest refugium.

The Lyndon Stream

Similarly, although the data are not as conclusive as those from the Awatere Valley, the Lyndon Stream eLGM fossil beetle fauna also contains at least two forest taxa. The pollen and plant macrofossils at the site indicate that the local vegetation at the eLGM consisted of grasses and shrubs. However, the beetles suggest a more diverse ecosystem at that time. The 25 fossil beetle taxa identified from Lyndon Stream represent a

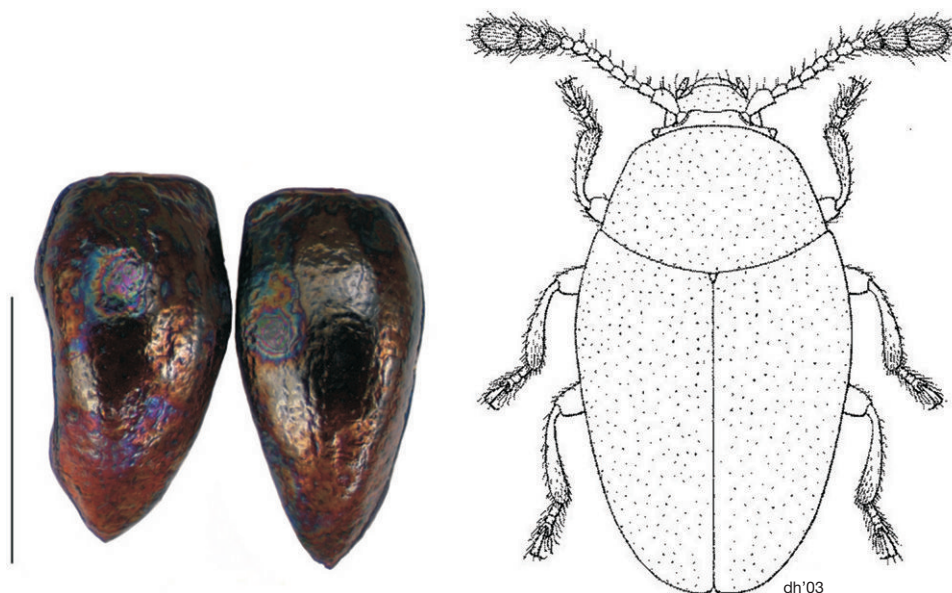


Figure 12 Illustration of new species of *Phyllocharitini* (Chrysomelidae) and the fossil elytra. The illustration is by D. W. Helmore, courtesy of Landcare Research, NZ. Image by B. Rhode of Landcare Research (scale bar = 1 mm).

rich and diverse ecosystem at the paleo-Lyndon lake at some period during the eLGM (Marra et al., 2006a). The beetles are typical of those found in high-country Canterbury lakes, which is consistent with the inferred interstadial climate discussed above, though it must be noted that the modern landscape is partially deforested by human activity. Although most of the fossil beetles are aquatic and riparian taxa, the assemblage also includes the weevil species *Rhopalomerus tenuirostris* (Figure 11(b)). Adults of *R. tenuirostris* feed on pollen, but the larvae inhabit very rotten wood, typically on the forest floor. In addition the fauna included a species of *Phyllocharitini* (Chrysomelidae; Figure 12). This is a newly-discovered taxon that is found in wet forests. Although not conclusive, the presence of these two taxa supports the presence of forest habitat at the site.

Westport

Burge and Shulmeister (2007a) characterized the fossil beetles from the Westport area (Figure 1) from a paleo-dunefield environment that spans 37 000–21 300 cal years BP. The palynology of sediments from the site indicates subalpine shrubland prior to glacial onset (34 000 cal years BP) and widespread grassland with little woody vegetation during the period of maximum glacial cooling. Fossils of forest beetles indicate forest vegetation prior to glacial onset, and a mosaic of closed canopy patches and open tussock grassland during full-glacial conditions.

The Howard Valley

Howard Valley is located in southwest Nelson in the Nelson Lakes district, at the northern end of the Southern Alps (Figure 1), between the Rotoiti and Rotoroa valleys. The area was extensively glaciated at the height of the last glaciation when glaciers flowed from surrounding ranges down the Rotoiti and Rotoroa valleys. The Howard Valley was not

glaciated because its headwaters do not extend into the Southern Alps as they do in the adjacent valleys.

Extensive terraces can be located in the Howard Valley and outcrops reveal stratigraphic sequences of alternating fluvial gravels and organic rich silts dated between ca. 34 000 and ca. 18 500 cal years BP. The paleoecological investigation of three sites along the terraces provided very good evidence for a forest refugium at times during the eLGM. Fifty-four beetle taxa represent seven habitat types: forest, forest or scrub, riparian and aquatic, litter, grass/tussock, marshland, and moss habitats. Fossils are supported by leaves of *Nothofagus menziesii*, which provide evidence of a forest refugium. Tree-line taxa such as *Taenarthrus* sp. 1 (Carabidae) and *Podocarpus* sp. (Podocarpaceae) indicate that the site may represent the upper tree limit for full-glacial time. Fossil forest taxa at this elevated site add to the growing fossil evidence for multiple forest refugia in NZ during the last glaciation and are consistent with the pollen records that have consistently indicated the presence of forest species during the last glaciation (Marra and Thackray, 2010).

Thus, the fossil beetle assemblages from several sites provide information on forest refugia during the eLGM. These results challenge the model of open vegetation at low elevations during glacial periods and a new model has emerged that shows forest persisted in areas previously considered grasslands. Figure 13 shows the altitude of eLGM sites where forest fossil beetles occur. They are illustrated against the modern altitudinal vegetation zones. The altitudes at which eLGM forest occurred are within the expected down slope adjustment (environmental lapse rate of 0.55 ± 0.05 °C/100 m) associated with the eLGM cooling of 4–5 °C estimated for NZ.

In addition, the eLGM forest stands/refugia have similar environmental conditions in that they are all in interior, sheltered valleys. They are also sites where moisture was not limited because at each location the local hydrology was disrupted by a geomorphic factor (such as landslide, fault movement, etc.)

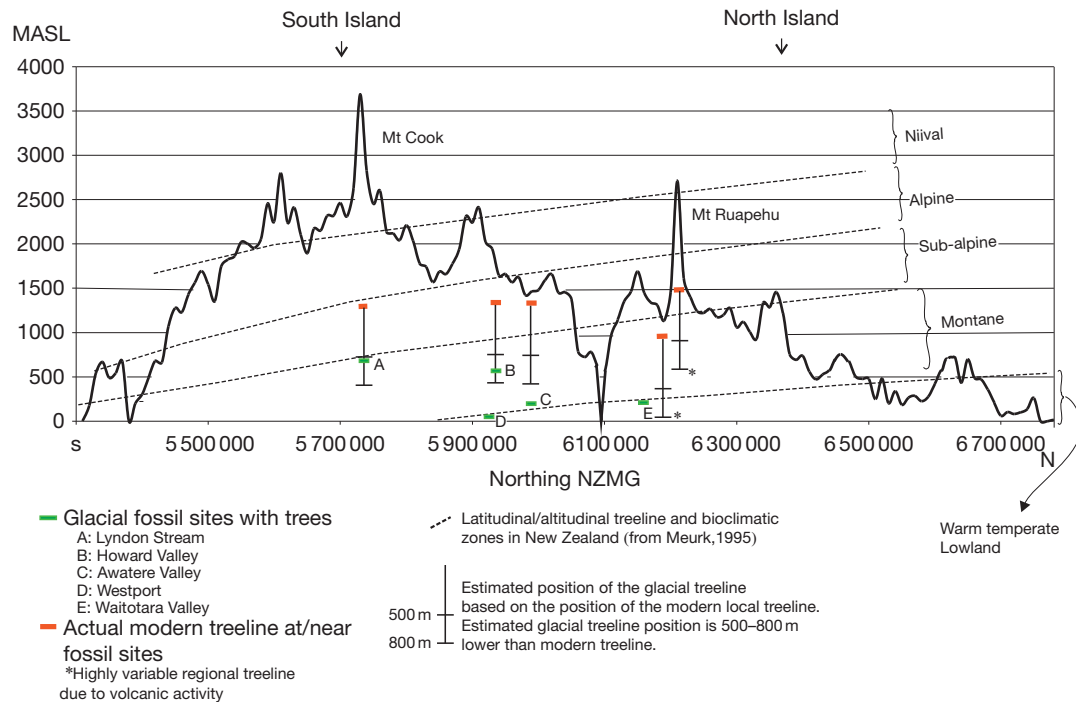


Figure 13 North-south profile of New Zealand showing eLGM forest sites against modern altitudinal vegetation zones. Adapted from Marra MJ and Thackray GD (2010) Glacial forest refugium in Howard Valley, South Island, New Zealand. *Journal of Quaternary Science* 25: 309–310.

that impeded local drainage, allowing lakes and swampy environments to be formed (Marra and Thackray, 2010). The combination of moisture availability and protection from wind appears to have increased the survival of forest species in sheltered inland areas.

See also: Phytoliths. **Beetle Records:** Late Pleistocene of Australia; Late Pleistocene of South America. **Pollen Records, Late Pleistocene:** Australasia.

References

- Alloway B, Westgate J, Pillans B, et al. (2004) Stratigraphy, age and correlation of middle Pleistocene silicic tephra in the Auckland region, New Zealand: A prolific distal record of Taupo Volcanic Zone volcanism. *New Zealand Journal of Geology and Geophysics* 47: 447–479.
- Atkinson TC, Briffa KR, and Coope GR (1987) Seasonal temperatures in Britain during the last 22,000 years, reconstruction using beetle remains. *Nature* 325: 587–592.
- Bennett KD (2004) Continuing the debate on the role of Quaternary environmental change for macroevolution. *Philosophical Transactions of the Royal Society B: Biological Sciences* 359: 295–303.
- Burge PI and Shulmeister J (2007a) Re-envisioning the structure of New Zealand vegetation during the last glaciation using beetle fossils. *Quaternary Research* 68: 121–132.
- Burge PI and Shulmeister J (2007b) An OIS 5a to OIS 4 (or early OIS 3) environmental and climatic reconstruction from the Northwest, South Island, New Zealand, using beetle fossils. *Journal of Quaternary Science* 22: 501–516.
- Coope GR (2004) Several million years of stability among insect species because of, or in spite of, Ice Age climatic instability? *Philosophical Transactions of the Royal Society B: Biological Sciences* 359(1442): 209–214.
- Hormes A, Preusser F, Denton G, et al. (2003) Radiocarbon and luminescence dating of overbank deposits in outwash sediments of the Last Glacial Maximum in North Westland, New Zealand. *New Zealand Journal of Geology and Geophysics* 46: 95–106.
- Johns PM (1986) Arthropods of Banks Peninsula. Report to the Commissioner of Crown Lands, Christchurch.
- Klimaszewski J and Watt JC (1997) *Coleoptera: Family-Group Review and Keys to Identification. Fauna of New Zealand, No. 37*. Canterbury, New Zealand: Manaaki Whenua Press.
- Kuschel G (1990) *Beetles in a suburban environment: A New Zealand case study. DSIR Plant Protection Report No. 3*. New Zealand: Mt Albert Research Centre.
- Kuschel G and Worthy TH (1996) Past distribution of large weevils (Coleoptera: Curculionidae) in the South Island of New Zealand, based on Holocene fossil remains. *New Zealand Entomologist* 19: 15–21.
- Leschen RAB and Rhode BE (2002) A new genus and species of large extinct Ulodidae (Coleoptera) from New Zealand. *New Zealand Entomologist* 25: 57–64.
- Marra MJ (2003a) A last interglacial beetle fauna from New Zealand. *Quaternary Research* 59: 122–131.
- Marra MJ (2003b) A description and interpretation of a fossil beetles assemblage from marine isotope stage 6 from Banks Peninsula, New Zealand. *New Zealand Journal of Geology and Geophysics* 46: 523–528.
- Marra MJ, Alloway BV, and Newnham RM (2006a) Paleoenvironmental reconstruction of a well-preserved Stage 7 sequence catastrophically buried by basaltic eruptive deposits, northern New Zealand. *Quaternary Science Reviews* 25: 2143–2161.
- Marra MJ, Crozier M, and Goff J (2009) Paleoenvironment and biogeography of a late MIS 3 fossil beetle fauna from South Taranaki, New Zealand. *Journal of Quaternary Science* 24: 97–107.
- Marra MJ and Leschen RAB (2004) Late Quaternary paleoecology from fossil beetle communities in the Awatere Valley, South Island, New Zealand. *Journal of Biogeography* 31: 571–586.
- Marra MJ and Leschen RAB (2011) Persistence of New Zealand Quaternary beetles. *New Zealand Journal of Geology and Geophysics* 54(4): 403–413.
- Marra MJ, Shulmeister J, and Smith E (2006b) Last glacial maximum beetle fauna, Lyndon Stream, Rakaia River Valley, South Island, New Zealand. *Quaternary Science Reviews* 25: 1841–1849.
- Marra MJ, Smith EGC, Shulmeister J, and Leschen R (2004) Late Quaternary climate change in the Awatere Valley, South Island, New Zealand using a sine model with a maximum likelihood envelope on fossil beetle data. *Quaternary Science Reviews* 23: 1637–1650.
- Marra MJ and Thackray GD (2010) Glacial forest refugium in Howard Valley, South Island, New Zealand. *Journal of Quaternary Science* 25: 309–310.

- May BM (1993) *Larvae of Curculionoidea (Insecta: Coleoptera): A Systematic Overview. Fauna of New Zealand, No. 28*. Canterbury, New Zealand: Manaaki Whenua Press.
- Meurk CD (1995) Evergreen broadleaved forests of New Zealand and their bioclimatic definition. In: Box EO, Peet RK, Masuzawa T, Yamada I, Fujiwara K, and Maycock PF (eds.) *Vegetation Science in Forestry*, pp. 151–197. The Netherlands: Kluwer Academic.
- Moar NT (1980) Late Otiran and early Aranuian grassland in central South Island. *New Zealand Journal of Ecology* 3: 4–12.
- Newnham RM, Lowe DJ, Giles TM, and Alloway BV (2007) Vegetation and climate of Auckland, New Zealand, since ca 32,000 cal yr ago: Support for an extended LGM. *Journal of Quaternary Science* 22: 517–534.
- Porter SC (1975) Equilibrium-line altitude of late Quaternary glacier in the Southern Alps, New Zealand. *Quaternary Research* 5: 27–48.
- Sandiford A, Newnham R, Alloway B, and Ogden J (2003) A 28 000–7600 cal yr BP pollen record of vegetation and climate change from Pukaki Crater, northern New Zealand. *Palaeogeography, Palaeoclimatology, Palaeoecology* 3–4: 235–247.
- Shulmeister J, Soons JM, Berger GW, et al. (1999) Environmental and sea-level changes on Banks Peninsula (Canterbury, New Zealand) through three glaciation–interglaciation cycles. *Palaeogeography, Palaeoclimatology, Palaeoecology* 152: 101–127.
- Soons JM (1979) Late Quaternary environments in the central South Island of New Zealand. *New Zealand Geographer* 35: 16–23.
- Soons JM and Burrows CJ (1978) Dates for Otiran deposits, including plant microfossils and macrofossils, from Rakaia Valley. *New Zealand Journal of Geology and Geophysics* 21: 607–615.
- Soons JM, Moar N, Shulmeister J, Wilson HD, and Carter JA (2002) Quaternary vegetation and climate changes on Banks Peninsula, South Island, New Zealand. *Global and Planetary Change* 33: 301–314.

Late Pleistocene of Northern Asia

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Glossary

Dendrophilic Species that live and feed on trees.

Thermophilic Species that live in relatively warm climates.

Xerophilic Species that prefer dry and relatively warm habitats, often on open ground, with or without sparse shrubs and trees.

Xylophages Species that feed on wood.

Late Pleistocene Beetle Records from Northern Asia

Preservation, Abundance, and Collecting Fossil Beetles in the Siberian Arctic

Insect fossils are common in various types of Quaternary deposits in Northern Asia. They are most abundant and well preserved in silt or sandy silt with plant detritus. Two main factors – the predominantly fine-grained composition of sediment matrix and ubiquitous permafrost – allow excellent preservation of the Late Pleistocene (LP) insect fossils in this region (Figures 1 and 2). Massive peat layers in the Arctic usually contain few insect remains, but thin-bedded peat intercalating with silt may contain rich insect faunas. The coarser sediment, such as sand with fine gravel deposited in rapidly running water, usually preserves only the most durable insect remains (Figure 3).

The average abundance of insect fossils in the sediment is not very high (Table 1). This means that in order to obtain a statistically representative sample with a minimum number of 100 individuals, it is necessary to process 50–60 kg of moist sediment with an average content of fossils, and sometimes up to 200 kg of sediment with a poor content of fossils. Taking into account the difficulties and expense of working in remote arctic regions, usually several weeks are spent at one site and 20–30 samples are collected there. The enormous cost of transportation makes it mandatory for one to reduce the weight of samples to be shipped, so samples are screened in the field (Figures 4–5) and after that pick up insects in the laboratory (Figure 6).

The identification of insect fossils in Northern Asia is difficult because of insufficient knowledge of both taxonomy and distribution of modern species. That is why modern insects are always collected during the authors' field season in these remote regions. This only marginally advances the cause of regional beetle taxonomy, however. Many groups are still poorly studied.

Localities of Fossil Beetles in Northern Asia

The first studies of Pleistocene insects in Northern Asia took place in the 1960s (Grushevskiy and Medvedev, 1962). These studies were associated with famous discoveries of complete mammal carcasses in permafrost or rich accumulations of mammoth bones, the so-called 'mammoth cemeteries' (Medvedev and Voronova, 1977). The regular sampling of Quaternary sediments for insect fossils in Northeastern Siberia and their

systematic research was started in the 1970s by Sergey Kiselev (1981). Until 2004 Svetlana Kuzmina continued Kiselev's work in that region, and Evgeny Zinovyev studied Quaternary insects in northwestern Siberia.

More than 30 years of regular paleoentomological research in Northern Asia has resulted in a large number of studied sites and analyzed assemblages, but their distribution over such a vast territory is uneven and discontinuous. For example, Zinoviev (1997) studied about 36 sites in northwestern Siberia, most of them of LP age. However, further east, the huge region between the mouth of the Ob' River and almost to the Lena River Delta is virtually unstudied. From the Lena Delta to Chukotka (Northeastern Siberia), the number of sites is much higher, but they are unevenly distributed (Figure 7). More than 200 LP beetle assemblages have been studied from 44 localities in this region (Table 2). The most important are discussed below.

Geological Background of the LP Beetle Record in Northern Asia

Northern Asia from west to east

While northwestern Siberia and the Taimyr Peninsula were strongly affected by the advances of the Barents-Kara Ice Sheet (Svendsen et al., 2004), no ice sheets existed east of the Taimyr. Only some high mountain ridges received enough moisture to build valley glaciers. Thus, the huge territory of Northern Asia between the Taimyr and Chukotka (i.e., between about 112° and 175° E) has never experienced major glaciations. Instead, this area was a broad arena for permafrost and periglacial processes. At present, the environment of the vast lowlands bordering the East Siberian and Laptev Seas is strongly influenced by the Arctic Ocean, but in the LP these flatlands had a much more continental climate, as they extended much further north during long periods when the shallow continental shelf was above sea level.

What is Yedoma?

The most remarkable and widespread Pleistocene sedimentary unit on these flatlands is ice-rich silt and sandy silt with thick ice wedges (Figure 8). It usually builds the main surface of the lowland (up to 60–100 m above the rivers) or its remaining hills when it is destroyed by thermokarst and erosion. Local people call these hills 'Yedoma,' and geologists used that name for this kind of sediment (Figure 9). The high content of ice in Yedoma sediments (up to 80%; Figures 10 and 11) explains

[†]Deceased.

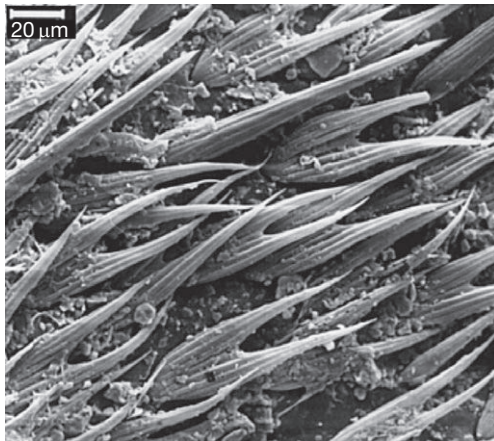


Figure 1 Fossil scales on the surface of the weevil, *Hypera ornata*, Bykovsky Peninsula, late Pleistocene. The condition of fossil chitin in frozen silts is usually as good as in modern beetles. Even such delicate and fragile elements as scales, hairs, and antennae are sometimes preserved.



Figure 2 Fossil head, pronotum, and leg of a dung beetle *Aphodius* sp., from the Duvanny Yar site, late Pleistocene. Excellent preservation is common for beetle remains recovered from buried ground squirrel nests, where some specimens have been found with legs preserved in their original position.



Figure 3 Fossil head capsules of weevils (above and right) and a predaceous diving beetle (below, left) from a sandy gravel unit, Keremesit River.

another common name for this formation – Ice Complex. Some important features of Yedoma, or Ice Complex, are shown in [Figures 12–14](#).

The same kind of ice-rich silt builds not only thick sequences of high Yedoma, but also lower terraces ([Figure 15](#)). It is also found on the slopes of rock hills and fills the valleys of small creeks in the highlands. In other words, a similar kind of sediment can occur in many different depositional environments. That is why earlier heated debates about the origin of Yedoma silts (fluvial, eolian, eluvial–colluvial, etc.) gradually gave way to the understanding that the ice-rich Yedoma silts are of polygenetic origin. They include sediments of different facies, which appear similar due to the homogeneity of the original material generated by cryogenic weathering during the Pleistocene and due to the activity of permafrost processes that universally affected sedimentation in various environments.

In closer proximity to mountain areas, active denudation yielded mostly coarser sediments, especially sands. These are the so-called Sand Yedoma sections known in various regions of Northeastern Siberia. Usually, but not always, they include much smaller ice wedges ([Figures 16 and 17](#)). A similar situation can occur along large rivers, building sand terraces with scattered ice wedges.

Thus Yedoma is not a facies, but a type of sedimentation reflecting a specific climate and environment, and in this sense the synonym ‘Ice Complex’ seems more correct. Interestingly, similar sedimentation took place in northwestern Siberia, when the Barents-Kara ice sheet retreated north in the Late Weichselian (LW) ([Svendsen et al., 2004](#)). The widest distribution of the Ice Complex sediments and their overall similarity make their stratigraphic characterization difficult.

Other types of sections

Numerous sections of ‘alasses’ – thermokarst depressions formed by the thawing of the Ice Complex – are another widespread element of regional geology. The majority of them were formed at the Pleistocene–Holocene transition, and are composed of reworked Yedoma silt or sand with low ice content ([Figure 18](#)). They may include a lacustrine unit of former thermokarst lakes and are overlain by a more or less thick peat layer, usually of early Holocene age. Earlier thermokarst events are recorded in buried alas units. They have a similar composition, but are overlain by an Ice Complex unit, and so mark periods of regional, or at least local, degradation of permafrost in the past. In several sites where ancient alas sediments underlie the Weichselian Ice Complex, and overlie supposedly middle Pleistocene units, they probably represent the Last Interglacial (Eemian). Only one such case has been documented by direct dating (infrared stimulated luminescence (IRSL), Krest-Yuryakh at Bolshoy Lyakhovsky Island; [Andreev et al., 2004](#)). If a thermokarst lake reappeared several times in one place, then composite sections formed, consisting of a few superimposed alas complexes ([Figure 19](#)).

The LP stratigraphy in Northeastern Siberia

The stratigraphic subdivision of Ice Complex (Yedoma) sequences is not easy because of the high homogeneity of the sediments. It is mostly based on radiocarbon dating, which makes it possible to reliably distinguish the LW units, but problems were encountered with the Middle Weichselian ([Sher and Plakht, 1988](#)). Ice Complex sections are referred to

Table 1 Concentration of insect fossils in the late Pleistocene frozen silts. The Lena Delta, Late Weichselian, Mamontouy Khayata section

<i>Sample No.</i>	<i>Weight of the screened wet sediment (kg)</i>	<i>Number of insect fossils extracted</i>	<i>MNI (minimum number of individuals)</i>	<i>The number of fossils per kilo</i>	<i>The number of insect individuals per kilo</i>	<i>Weight of matrix to be screened to get 100 insect individuals (kg)</i>
MKh99-18	9	52	30	6.1	3.5	28
MKh99-16	20	32	25	1.6	1.2	82
MKh99-14	20	10	9	0.5	0.4	227
MKh99-12	9	17	12	2.0	1.4	71
MKh99-22	17	160	88	9.4	5.2	19
MKh99-6	26	109	56	4.3	2.2	46
MKh99-5a	9	42	23	4.9	2.7	37
MKh99-5o	14	111	47	8.2	3.5	29
MKh99-5s	14	146	57	10.7	4.2	24
MKh99-3a	7	85	43	12.5	6.3	16
Average				6.0	3.1	62

Note: The volume of screened sediments in these samples is small, as they were screened by geologists. However, the weight of samples was measured, and this case can be used to estimate concentration of fossils.

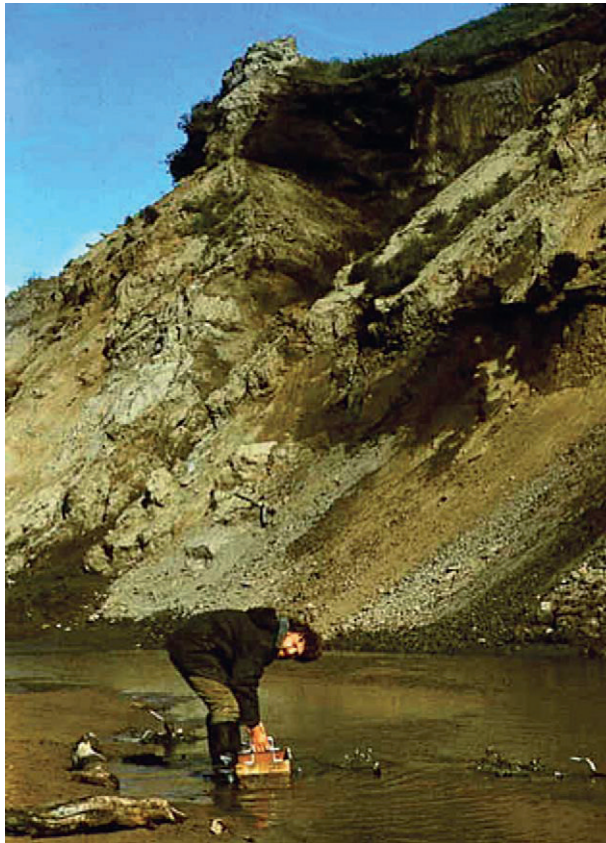


Figure 4 Sediments in the field are screened, using a large hand-made box with 0.5 mm mesh sieve, letting river, lake, or pond water in and out through the sieve only. The screening is aimed at removing most, but not all of the mineral component, so some amount of silt remains in the box, helping preserve the shape of many fossils smaller than 0.5 mm that might otherwise be lost. These specimens usually float on top of the silt and remain in the sample.

the Early Weichselian if they have an infinite ^{14}C age and are overlain by the dated middle to LW sediments. Strictly speaking, such arguments are not fully conclusive. There are Ice Complex sections dated from the late middle Pleistocene (Kaplina et al., 1980a,b), or even earlier. On the other hand,



Figure 5 The screen residues are dried as much as the weather allows, at first openly, later in fabric bags.

the Ice Complex units with 'finite' ^{14}C dates in the range 35–45 ka do not necessarily belong to the Middle Weichselian, as there were many cases when such units turned out to be much older (Sher et al., 2005).

Table 3 summarizes the geological position and the age of the LP sites from which the insect assemblages have been studied.

Peculiar Features of Fossil Beetle Assemblages in Northeastern Asia

General composition

The species composition of the Pleistocene beetle fauna of Northeastern Asia is strange, compared with the other arctic regions:

1. It has higher diversity and abundance of plant-feeding insects, mostly weevils (Curculionidae) and leaf beetles (Chrysomelidae). Even among ground beetles (Carabidae), mostly a predatory family, one of the commonly found species is a plant-feeder *Curtonotus alpinus*.



Figure 6 Quite often, silt sticks to the fossils, or fills the insides of pronota, head capsules, or joined elytra. This limits the effectiveness of the laboratory flotation method. Instead, dry residuals are separated into the size fractions through regular soil sieves, then the insect and other fossils are picked by hand, under binocular microscope for smaller fractions.

2. Fossil assemblages usually include many species that currently live in tundra and forest tundra, but the diversity of some groups, such as litter inhabitants (rove beetles), carion beetles, and fungus beetles, is notably lower than in modern tundra communities. Also, the diversity of water and riparian species in the fossil assemblages is less than today in the Arctic.
3. Along with tundra species, most Pleistocene assemblages show significant numbers of steppe, meadow-steppe, and other xerophilic insects. Especially notable is the abundance of the pill beetle *Morychus viridis*, an inhabitant of cold steppe-like biotopes in Northeastern Siberia today (see [Tables 4, 5 and 6](#)). This species makes up as much as 90% of some LP insect assemblages.
4. Xylophages and other dendrophilic species are very rare in the LP insect assemblages in this region, so no true forest environment can be reconstructed here for any part of the LP.

The above features characterize many LP insect assemblages in Northeastern Asia, which combine species presently inhabiting a variety of different regions and ecosystems. For example, such arctic species as the ground beetle *Pterostichus costatus*, the leaf beetles *Chrysolina tolli* and *Ch. subsulcata*, and the willow weevil *Isochnus arcticus*, are often found in one fossil assemblage with such steppe leaf beetles as *Ch. perforata*, and the weevils *Stephanocleonus eruditus*, *S. fossulatus*, and *S. incertus*, which now live in southern Siberian and Mongolian steppe. On the other hand, these assemblages sometimes include a few species of insects whose current distribution is restricted mostly to the taiga zone, such as the boreal ant species *Camponotus*

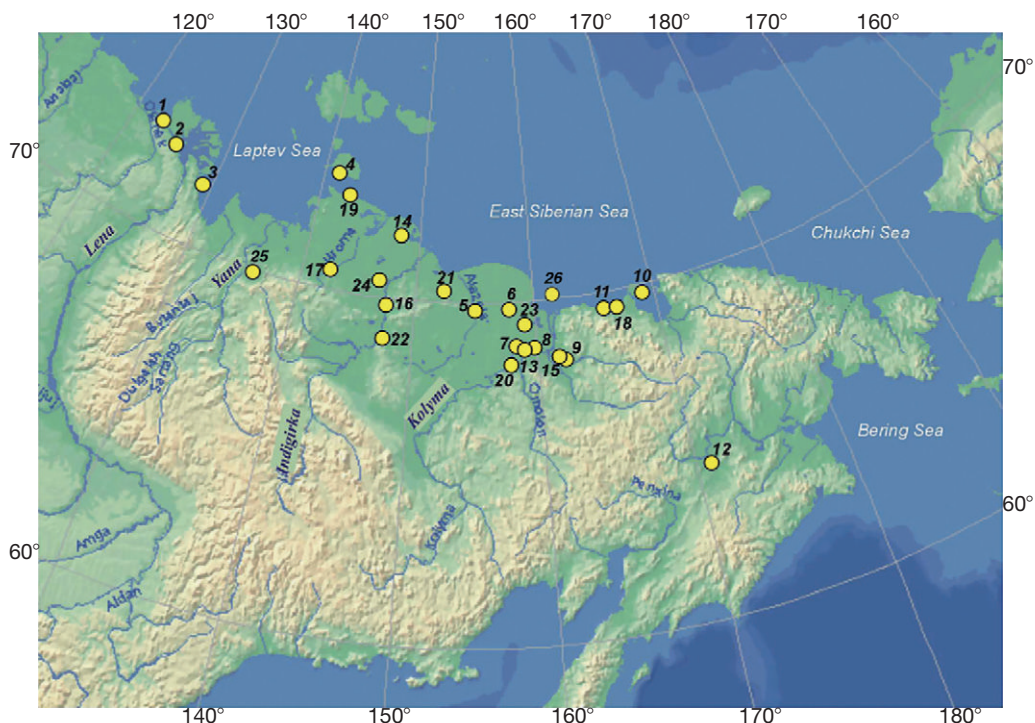


Figure 7 Map of late Pleistocene fossil beetle localities in Northeastern Siberia. 1–3 Lena Delta region: 1, Olenyok Channel, Nagym; 2, Byor-Khaya; 3, Bykovsky Peninsula; 4, Bolshoy Lyakhovskiy Island; 5, Alazeya; 6, Chukochya; 7, Alyoshkina Zaimka; 8, Duvanny Yar; 9, Stanchikovskiy Yar; 10, Ayon Island; 11, Milkera; 12, Main River, Ledovy Obryv (Ice Bluff); 13, Omolon; 14, Yana-Indigirka; 15, Krasivoye; 16, Shamanovo; 17, Khroma River; 18, Primorsky; 19, Oyagosskiy Yar; 20, Krestovka; 21, Khomus-Yuryakh; 22, Synpoy Yar; 23, Yakutskoe Lake; 24, Achchaggy-Allaikha; 25, Yana; 26, Medvezhiy Islands.

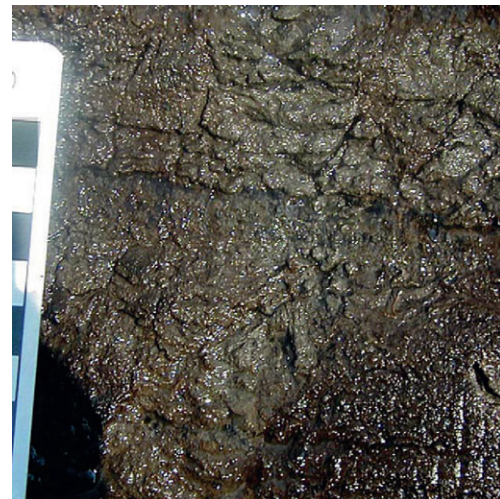
Table 2 The distribution of studied insect assemblages of the late Pleistocene age in Northeastern Siberia

Region	Number of studied sections	Number of insect samples with MNI > 30
Lena Delta (1–3)	3	63
Yana-Indigirka Lowland (14,16,17,19,22,24,25)	11	42
New Siberian Islands (4)	4	11
Kolyma Lowland (5–9,13,15, 20,21,23)	16	50
Medvezhyi Islands (26)	2	4
West Chukotka (Chaun) (11,18)	3	16
Ayon Island (10)	4	38
South Chukotka (Main R.) (12)	2	7
Total	45	231

**Figure 8** One of the famous exposures of late Pleistocene icy silts – Vorontsovskiy Yar on the Indigirka River. The top of this bluff, which is about 40 m high, lies about 90 m above the river level.**Figure 9** The Duvanny Yar site, in the lower course of the Kolyma River, is a cut bank exposure of the so-called Omolon-Anyuy Yedoma – the high-eroded surface between the Omolon and Anyuy Rivers – large tributaries of the Kolyma.

herculeanus and *Formica gagatoides*, the weevils *Hylobius piceus* and *Pissodes irroratus*, and others.

Table 4 lists the most common species of LP beetles found in more than 40% of all LP assemblages and, additionally, a few less abundant, but rather indicative species. Figure 20 shows the fossils of some beetle species found in the LP sediments in Northeastern Siberia.

**Figure 10** Mamontovyy Khayata Cliff, Bykovsky Peninsula, eastern side of the Lena R. Delta. Extensive ice bodies such as this, which in earlier times were interpreted as buried glaciers, in fact represent polygonal systems of syngenetic (growing simultaneously with continued sedimentation) ice wedges.**Figure 11** The ground pillars between the ice wedges also contain lots of ice, but as small layers and microscopic lenses segregated when the silt was freezing.

Non-analog assemblages. What is tundra-steppe?

The fossil insect assemblages of the kind discussed above have no modern analog. They resemble non-analog Pleistocene mammal communities, in which, for example, saiga antelope, a modern inhabitant of dry steppe and semidesert, existed



Figure 12 One of the typical features of Yedoma silts is the abundance of very thin plant roots, most likely belonging to grasses. They occur *in situ* and usually form complicated branching systems, permeating frozen silt.



Figure 13 Some frozen silt layers include woody roots of small shrubs.

together with arctic fox and musk ox. It has been shown that such assemblages are not a random mixture of fossils, but a natural combination of species related to a peculiar type of Pleistocene environment (Sher, 1988, 1990; Vangengeim, 1977). In Russia, this environment and associated plant and animal communities are traditionally called ‘tundra-steppe’ after Tugarinov (1929). In Europe, tundra-steppe communities were related mostly to the periglacial zone south of the ice sheets, but in Siberia, under a more continental climate, they were distributed much more widely and existed more permanently (Vangengeim and Ravsky, 1965). Kiselyov (1981) has shown that insect assemblages of the tundra-steppe type were very common in Northeastern Siberia during the Pleistocene. Moreover, he demonstrated that such insect communities developed in Siberia as early as the late Pliocene. Steppe elements

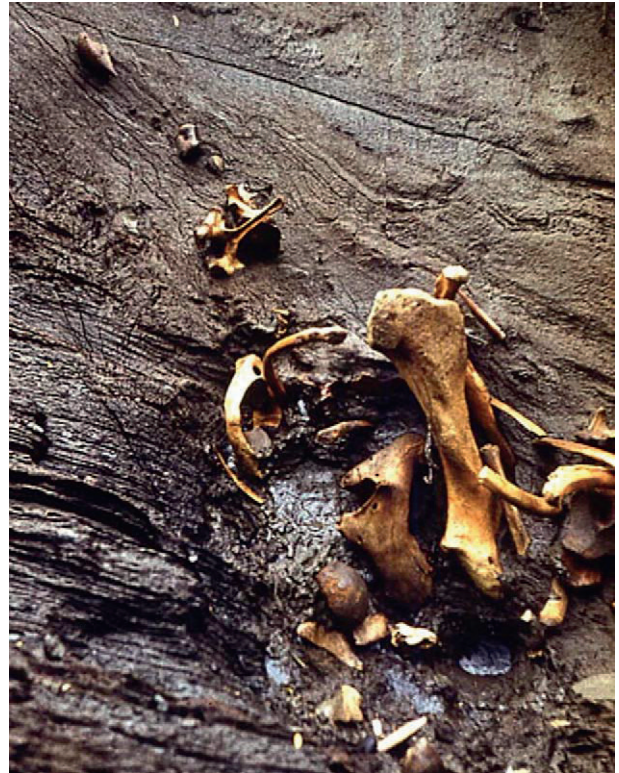


Figure 14 The Yedoma silts are very rich in various kinds of fossils – plants, freshwater shells, and insects. The most famous remains are the large number of mammal bones and complete or partial frozen carcasses of Pleistocene mammals.



Figure 15 The same kind of ice-rich silt builds not only thick sequences of high Yedoma, but also terrace-like bodies of lower elevation (Low Yedoma, 15–25 m above the river), more clearly positioned in river valleys (Omolon River, Kolyma River basin).

were especially abundant in tundra faunas during relatively cold (glacial) periods, but they survived even during warmer (interglacial) times (Kiselyov, 1981, 1994). The significant destruction of tundra-steppe communities occurred only at the Pleistocene–Holocene boundary, when the dry Pleistocene grassland mostly disappeared, giving way to wet tundra and bogs. This brought about the extinction of the most remarkable large mammalian herbivores, such as mammoth and woolly



Figure 16 Impressive sand sequences up to 60 m high are usually exposed where large rivers run in low mountain areas. The best-known examples are Sychnoy Yar on the Indigirka River and in the city of Verkhoyansk area (Yana and Adycha River basins). A coarser sediment (sands) limited the scale of frost cracks, and the ice wedges in these fluvial sequences are much smaller than in classical Yedoma. They are restricted to the lenses of silty sand (floodplain and ox-bow deposits).



Figure 17 The Ice Bluff section, Main River, southern Chukotka. Sand sections such as this, with a higher content of silt, allowed the continuous growth of ice wedges.

rhino, and the dramatic shrinkage of ranges of others, including horses, bison, and saiga antelope. However, the plant communities of the tundra-steppe type persisted in relatively small relic habitats (Yurtsev, 1981). The ranges of steppe insects also shrank, but many of them survived in relic steppe habitats. Such habitats are widespread in the inner mountains of Northeastern Siberia that have an extremely continental climate. Many thermophilic steppe species found refugia there on south-facing slopes (Berman et al., 2001). Less thermophilic steppe beetles were able to survive in the warmest localities in the Arctic, these habitats being especially abundant on Wrangel Island (Berman, 1986; Khruleva, 2004). The persistence of these beetle species – relics of the Pleistocene tundra-steppe – is one of

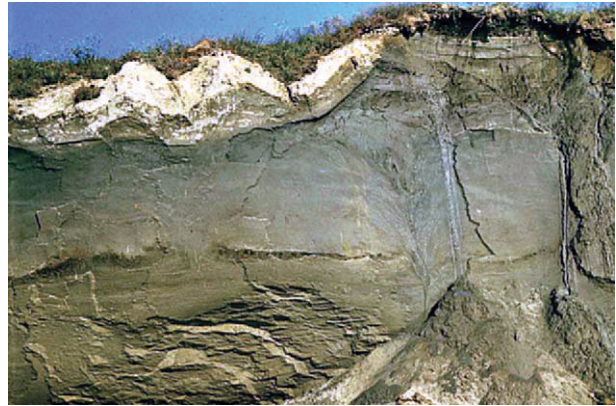


Figure 18 In the thermokarst (alas) sections, the Ice Complex sediments have thawed through, that is, reworked in place. As a result, the ice content is much lower here, as former ice wedges have been replaced by ice-wedge casts, or pseudomorphs.



Figure 19 One such reworked section, Molotkovsky Kamen' at the Malyy Anyuy River, had been selected as the regional type of Middle Weichselian Interstadial (Molotkov Horizon) with alternating warmings and coolings. Subsequent redating of the buried peat layers in this section, however, showed much earlier ^{14}C dates, so the age of the Molotkov Horizon should be revised (Sher et al., 2005).

the most remarkable features of the present insect fauna of Northeastern Siberia.

Comparison of Insect Assemblages of Different Ages Using Ecological Grouping

The high frequency of occurrence of a restricted number of species in fossil assemblages (Table 4), and the permanent presence of insects associated with tundra, steppe, and other xerophilic habitats makes the species composition of various assemblages from the Pleistocene of Northeastern Siberia rather monotonous. They differ from each other mostly in the relative abundance of fossils of certain taxa. That is why Kiselyov (Kiselyov, 1981; Sher et al., 1979) used a quantitative approach to trace faunal and environmental changes through time. He classified all species into a number of ecological groups according to their preferred modern habitat, and then he counted the total number of individuals belonging to all species in each group. It was much easier to compare

Table 3 Geological position and the age of the late Pleistocene sites in Northeastern Siberia, from which the insect assemblages have been studied

<i>The section type</i>	<i>Sediment</i>	<i>Geocryology</i>	<i>Sections and map no., figure</i>	<i>Age range</i>	<i>Reference</i>
High Yedoma section (40–60 m and more above the river or sea level)	Mostly silt with sandy silt	One penetrating system of large ice wedges	Mamontovyy Khayata (Figure 7 , No. 3; Figure 10)	EW (?)–MW-LW (>50–12 ka)	Sher et al. (2005)
			Duvanny Yar (Figure 7 , No. 8; Figure 9)	Mostly EW (?)–MW?. Probably LW at the top (>50–18 ka)	Sher and Plakht (1988)
		Multiple (2–3) overlying systems of ice wedges	Khomus-Yuryakh, Loc. 83 (Figure 7 , No. 21)	Late middle Pleistocene	Sher (unpublished)
High Sand section (>30, up to 60 m and more above the river or sea level)	Sand, silty sand	Relatively small ice wedges are restricted to more silty lenses	Sypnoy Yar (Figure 7 , No. 22; Figure 16)	LP undivided (>45 000 years BP in the lower third of the section)	Kaplina (1981)
			Utatgyr Suite, Ayon Island (Figure 7 , No. 10)	LP undivided	Svitoch (1980) and Kiselyov (1994)
		One penetrating system of large ice wedges	Ledovyy Obryv (Ice Bluff) (Figure 7 , No. 12; Figure 17)	MW-LW (42–15 000)	Anderson, et al. (2002) and Sher (unpublished)
Topping Yedoma – Ice Complex up to 20–25 m thick on the top of composite sections	Mostly silt with sandy silt	One penetrating system of large ice wedges	Buor-Khaya (Figure 7 , No. 2)	EW-LW (>50–17 ka)	Schirrmeister et al. (2003)
			Nagym (Figure 7 , No. 1)	MW ? (44–45 ka)	Schirrmeister et al. (2003)
			Zimovye R. Mouth, Bol. Lyakhovsky Island (Figure 7 , No. 4)	EW-MW (100–28 ka)	Schirrmeister et al. (2004)
			Yakutskoe Lake (Figure 7 , No. 23)	LW?	Kiselyov (1994)
			Khroma, Alazeya, Chukochya, Krestovka (Figure 7 , No. 17, 5, 6, 20)	EW-MW? (Alazeya, the date of 38 ka in the lower part of the unit)	Kaplina et al. (1981)
			Achchagyy-Allaikha (Figure 7 , No. 24)	Late middle Pleistocene	Kaplina et al. (1980a,b)
			Oyagosskiy Yar (Figure 7 , No. 19)	Late middle Pleistocene–LW	Nikolsky and Basilyan (2003)
			Omolon-I (Figure 7 , No. 13; Figure 15)	LW (24–15 ka)	Sher (unpublished)
Low Yedoma terraces (20–30 m high above the river level)	Mostly silt with sandy silt	One penetrating system of large ice wedges	Krasivoye Bluff (Figure 7 , No. 15)	LW (18 700 years BP)	Kiselyov (1994)
			Shamanovo (Figure 7 , No. 16)	MW-LW (33 ka and younger)	Kaplina et al. (1980a,b)
			Alyoshkina Zaimka (Figure 7 , No. 7)	LW (17.2–13.2 ka) from >28 ka to early Holocene at the top	Sher et al., 1979; Alfimov et al. (2003); Pitul'ko et al., 2007
Low Sand terraces (20–25 m high above the river level)	Sand, silty sand	Relatively small ice wedges are restricted to more silty lenses.	Yana archaeological site		
Multilayer alasses	Silt, peat	Ice-wedge pseudomorphs	Primorskiy, coast near Rauchua R. mouth (Figure 7 , No. 18)	>28 000 years BP in the lower part of the sequence, 24 800 years BP in the middle part, 10.4 ka at the top	Kiselyov (1994)
Buried alass complexes	Silt, peat	Ice-wedge pseudomorphs	Khroma, Kyl-Bastakh Layers (Figure 7 , No. 17)	LI? (overiles mid-Pleistocene Khroma	Kaplina et al. (1983)

(Continued)

Table 3 (Continued)

The section type	Sediment	Geocryology	Sections and map no., figure	Age range	Reference
			Alazeya, Kuobakh Unit (Figure 7, No. 5)	Suite, overlain by Ice Complex) LI? (overlies mid-Pleistocene (?) Maastakh Suite, overlain by Ice Complex)	Kaplina et al. (1981)
			Zimovye R. Mouth, Bol. Lyakhovsky Island (Figure 7, No. 4)	LI ? (overlies mid-Pleistocene (?)) Kuchchuguy Suite, overlain by Ice Complex)	Andreev et al. (2004)
Low terraces		16 m terrace	Mil'kera (Figure 7, No. 11)	MW (32800) ?	Kiselyov (1994)

LI, Last Interglacial; LP, Late Pleistocene; EW, Early Weichselian; MW, Middle Weichselian; LW, Late Weichselian.

Sources:

Alfimov AV, Berman DI, and Sher AV (2003) Tundra-steppe insect assemblages and reconstruction of Late Pleistocene climate in the lower reaches of the Kolyma River. *Zoologicheskii Zhurnal* 82(2): 281–300 (in Russian).

Anderson PM, Kotov AN, Lozhkin AV, and Trumpe MA (2002) Palynological and radiocarbon data from Late Quaternary deposits of Chukotka. In: Anderson PM and Lozhkin AV (eds.) *Late Quaternary Vegetation and Climate of Siberia and Russian Far East (Palynological and Radiocarbon Database)*, pp. 35–79. Magadan: NESR FEB RAS.

Kaplina TN (1981) History of frozen sequences of northern Yakutia in the late Cenozoic. In: Dubikov GI and Baulin VV (eds.) *Istoriya razvitiya mnogoletnemerzlykh porod Evrazii., M.*, pp. 153–181. Novosibirsk: Nauka (in Russian).

Kaplina TN, Lakhtina OV, and Rybakova NO (1981) Cenozoic deposits in the middle course of the Alazeya R. (Kolyma Lowland). *Izvestiya Akademii Nauk SSSR. Ser. Geologicheskaya* 8: 51–63 (in Russian).

Kaplina TN, Ovander MG, Lozhkin AV et al. (1983) Quaternary deposits in the middle current of the Khroma River (Yana-Indigirka Lowland). In: Shilo NA (ed.) *Stratigrafiya i paleogeografiya pozdnego kaynozoya Vostoka SSSR*, pp. 80–95. Magadan: SV KNII DVNTs AN SSSR (in Russian).

Nikolsky PA and Basilyan AE (2003) Svyatoy Nos Cape – The key section of the Pleistocene in the North of Yana-Indigirka Lowland. In: *The natural history of the Russian East Arctic in the Pleistocene and Holocene*, pp. 5–13. Moscow: GEOS Publishers (in Russian).

Pitul'ko VV, Pavlova EYu., Kuz'mina SA, et al. (2007) Natural climatic changes in the Yana-Indigirka Lowland during the terminal kargino time and habitat of Late Paleolithic man in northern part of East Siberia. *Doklady Akademii Nauk* 417(1): 103–108.

Schirmer L, Andreev A, Grosse G, and Meyer H (2004) Saalian to Holocene Paleoenvironmental History Documented in Permafrost Sequences of Arctic Siberia (New Siberian Archipelago, Bolshoy Lyakhovsky Island), AGU Fall Meeting, 13–17 December 2004, San Francisco, California C13B-0286

Schirmer L, Kunitsky VV, Grosse G, et al. (2003) Late Quaternary history of the accumulation plain north of the Chekanovsky Ridge (Lena Delta, Russia) – A multidisciplinary approach. *Polar Geography* 27(4): 277–319.

Svitoch AA (1980) The description of the latest deposits [of the Ayon Island]. In: Kaplin PA (ed.) *The latest deposits and paleogeography of the Pleistocene in Chukotka*, pp. 180–202. Moscow: Nauka (in Russian).

assemblages from different horizons or even sites by the percentage of ecological groups. The authors use the same method with minor variations (Table 5).

For large sections with many insect samples, percentage diagrams are used for better presentation. Recently, the authors suggested the establishment of Insect Faunal Zones (IFZs) in such sections, based on the ecological structure of assemblages. This zonation scheme is similar in some ways to pollen zones (Sher et al., 2005). In Figure 21, one can see how the composition of insect assemblages changed in the Lena Delta during the last 50 ka (this sequence will be considered in more detail below).

The LP Beetle Faunas of Certain Periods and Areas

Below the most important LP beetle assemblages are described, starting with the youngest because they are the most reliably dated (Table 3).

LW insect assemblages

The fossil insect assemblages dated from about 24 to 12 ka were first studied in the Kolyma Lowland, which later proved to be almost the center of continentality and formation of

steppe habitats in the whole of Northern Asia during the Pleistocene. The most interesting of these sites is Alyoshkina Zaimka on the lower course of the Kolyma River (Figure 7, no. 7; Table 3). This site yielded four rich insect faunas within the range of 17–13 ka (all dates in this article are uncalibrated radiocarbon ages), almost completely (up to 80–95%) dominated by thermophilic xerophiles, including true steppe species (10–20% of the whole assemblage), with a very low percentage of tundra beetles (Kiselyov, 1981). The assemblages show high diversity and abundance of steppe and meadow-steppe species (Table 6). Either meadow-steppe beetles, or a pill beetle *M. viridis* dominate all samples. These faunas represent the most 'steppic' dominated type of non-analog tundra-steppe assemblages.

Scattered similar assemblages of the same age are known elsewhere, but mostly within the Kolyma Lowland. For example, they are found in the Omolon-I section on the Kolyma River (Figure 7, no. 13; Figure 15; Table 3), Assemblages with more than 40% of meadow-steppe and steppe species are present here in the interval between 7.0 and 12 m depth (~23–18 ka) (Kiselyov et al., 1987). Tundra-steppe assemblages with extremely high numbers of *M. viridis* (Table 7), but less diverse and abundant steppe species occur below and above this interval.

Table 4 Characteristics of the most common and some indicative beetle species in the late Pleistocene assemblages in Northeastern Siberia

Family and species	% of all LP assemblages, where the species is found	Description	Taxonomic notes
<i>Ground beetles, Carabidae</i>			
<i>Pterostichus</i> (<i>Cryobius</i>) spp.	78	Most <i>Cryobius</i> are common in modern tundra and northern taiga and prefer boggy and wet habitats. <i>P. (C.) nigripalpis</i> Popp. is abundant today on Wrangel Island on dry patches with xerophilic vegetation (O. Khruleva, personal comm.)	Of numerous species in this subgenus, fossils of <i>P. (C.) ventricosus</i> Esch., <i>P. (C.) pinguedineus</i> Esch., <i>P. (C.) brevicornis</i> (Kirby), and probably <i>P. (C.) nigripalpis</i> Popp can be identified
<i>Curtonotus alpinus</i> (Payk)	92	Very widespread species from taiga to arctic tundra; prefers relatively dry and warm patches; feed on grasses. Hardly distinguishable fossils of a close species, <i>C. bokori</i> , can be present in some assemblages, but <i>C. alpinus</i> is much more common	In Europe and America this species is usually called <i>Amara alpina</i> Payk. In Russia, the subgenus <i>Curtonotus</i> , to which this species belongs, was raised to the generic level (Kryzhanovskiy, 1965; Kryzhanovskiy et al, 1995)
<i>Poecilus</i> (<i>Derus</i>) <i>nearcticus</i> Lth.	49	Modern specimens of <i>P. nearcticus</i> were collected on dry sandy slopes at a few sites in the Siberian Arctic and Alaska. In the Pleistocene, it was much more common, though not very abundant. It is a typical relict of the tundra-steppe beetle complex	
<i>Pterostichus</i> (<i>Tundraphilus</i>) <i>sublaevis</i> Sahlb.	35	This beetle lives now in tundra and prefers dry and warm habitats, but it is relatively rare. It was more widespread and abundant during the Pleistocene	
<i>Small carrion and round fungus beetles, Leiodidae</i>			
<i>Cholevinus sibiricus</i> (Jean.)	47	This is a rare species in modern tundra, preferring wet habitats. It is probably necro-, copro-, or/and detritophagous. <i>Ch. sibiricus</i> was more important in the Pleistocene tundra-steppe communities (Perkovsky and Kuzmina, 2001). Despite rather large populations of grazing mammals in tundra-steppe ecosystem, the revealed diversity of necro- and coprophagous beetles is very low. <i>Ch. sibiricus</i> might use this kind of food resources	Prior to 1999, this species was called <i>Cryocatops poppiusi</i> Jean., family Catopidae (Perkovsky, 1999)
<i>Rove beetles, Staphylinidae</i>			
<i>Tachinus arcticus</i> Motsch. and <i>T. brevipennis</i> Sahlb.	45	Both species live in wet tundra habitats, but <i>T. arcticus</i> is more abundant in the north, while <i>T. brevipennis</i> – in the south of the zone, and is probably more thermophilic	Earlier these fossils were referred to <i>T. apterus</i> Maekl (Kiselyov, 1981). Later, the identification was changed to <i>T. arcticus</i> . Now it is suspected that it can be a mixture of fossils of two different species
<i>Dung beetles, Scarabaeidae</i>			
<i>Aphodius</i> sp.	44	This dung beetle occurs in many Pleistocene sites and is clearly associated with tundra-steppe assemblages. Its well-preserved fossils have been found in buried ground squirrel nests	Probably an extinct species
<i>Pill beetles, Byrrhidae</i>			
<i>Morychus viridis</i> Kuzm. et Kor.	86	This beetle is a real symbol of the Pleistocene biota in Northeastern Siberia. Found in the majority of fossil assemblages, often in very high numbers, it is clearly associated with thermophilic steppe. At present, it lives in cold steppe-like habitats in the Kolyma Highlands, Chukotka, and Wrangel Island, called 'hemicycophytic steppe' after Yurtsev (1981). Modern habitats of <i>M. viridis</i> have extremely contrasting microclimatic conditions – very hot and dry in summer, and snowless and very cold in winter (Berman et al., 2001)	Before this species was described (Kuzmina and Korotyaev, 1987), its fossils were referred to <i>Chrysobyrrhulus rutilans</i> Motsh. (Kiselyov, 1981; Medvedev and Voronova, 1977)
<i>Leaf beetles, Chrysomelidae</i>			
<i>Chrysolina tolli</i> Jac.	25	This is one of the most cold-resistant species of beetles in the world. <i>Ch. tolli</i> was found in polar desert on Novaya Zemlya, Severnaya Zemlya and Novosibirsk Islands. In	The fossils were referred to <i>Ch. cavigera</i> Sahlb. until the recent revision by Bienkowski (1999), who showed that the modern arctic populations

(Continued)

Table 4 (Continued)

Family and species	% of all LP assemblages, where the species is found	Description	Taxonomic notes
		the tundra zone, it occupies different habitats, and feeds on Crucifera. <i>Ch. tolli</i> is found in many Pleistocene assemblages, though in low numbers, and is most common in northern sites	belong to <i>Ch. tolli</i> , while <i>Ch cavigera</i> is restricted to the Kamchatka Peninsula and the Okhotsk Sea coast
<i>Weevils, Curculionidae</i>			
<i>Sitona borealis</i> Kor.	46	This weevil feeds on legumes. Currently it lives mostly in meadow-steppe habitats in the taiga zone, but it also occurs on dry warm patches in tundra. It was more common in the Pleistocene than it is now	Formerly named as <i>S. ovipennis</i> Hochh. (Kiselyov, 1981), then <i>S. ovipennis borealis</i> Kor., this weevil is now considered as a separate species <i>S. borealis</i> Kor. (Korotyaev, 1996)
<i>Stephanocleonus eruditus</i> Faust	45	The main modern range of this weevil is Southern Siberia and Mongolia (mountain steppe). It is the most common steppe element in the Pleistocene assemblages. Having been very widely distributed in the Pleistocene, it survived in scattered relic steppe communities in the Kolyma-Indigirka Highlands (Berman et al., 2001). This species is a good indicator of relatively high summer temperature (Berman and Alfimov, 1993)	
<i>Lepyrus nordenskiöldi</i> Faust	46	This weevil lives on willow, rarely on birch, and is common in tundra, taiga, and relic steppe modern environments, but prefers drier habitats with sandy soil. Thus, being very often a single representative of shrub insect group in the Pleistocene assemblages, <i>L. nordenskiöldi</i> can be considered as an indicator of relatively dry environment, although the authors count this species among the forest-tundra species	
<i>Hypera ornata</i> (Cap.)	48	The distribution and ecology of this weevil are similar to those of <i>Sitona borealis</i> (see above)	Former name <i>Phytonomus omatus</i> Cap.
<i>Isochnus arcticus</i> (Kor.)	44	This species lives on willows, but its distribution is restricted to the typical tundra subzone. It is most abundant on Wrangel Island, but was also found on the Taimyr Peninsula, in western Chukotka and northern Alaska. It can be found in the Pleistocene assemblages of different age in small numbers, but became the dominant species during the Last Glacial Maximum at some (mostly northern) sites	Former name <i>Rhynchaeus arcticus</i> Kor.

Source: Alfimov AV, Berman DI, and Sher AV (2003) Tundra-steppe insect assemblages and reconstruction of Late Pleistocene climate in the lower reaches of the Kolyma River. *Zoologicheskii Zhurnal* 82(2): 281–300 (in Russian); Berman DI and Alfimov AV (1993) Microclimatic conditions of existence of steppe ecosystems in the Subarctic of north-east Asia. *Bulletin of Moscow Society of Naturalists, Biology Section* 98 (3): 118–128 (in Russian); Bienkowski AO (1999) A review of the species of the subgenus *Chrysolina* (Pleurosticha) Motschulsky (Insecta: Coleoptera, Chrysomelidae). *Entomologische Abhandlungen*. 58(10): 165–179; Korotyaev BA (1996) 165. *Sitona* Germ. In: Egorov AB, Zherichin VV, and Korotyaev BA *Sem. Curculionidae – Dolgonosiki. Key to the insects of the Russian Far East*, vol. 3, Coleoptera, Pt. 3, Supplement, p. 499–503. Vladivostok: Dal'nauka. (in Russian); Kryzhanovskiy OL (1965) Fam. Carabidae – Ground beetles. *Guide book to the identification of the insects of European part of USSR*, vol. 2, p. 29–77. Moscow-Leningrad: Nauka (in Russian); Kryzhanovskiy OL, Belousov IA, Kabak II, Kataev BM, Makarov KV, and Shilenkov VG (1995) *A checklist of the ground-beetles of Russia and adjacent lands (Insecta, Coleoptera, Carabidae)*, 272 pp. Sofia – Moscow: Pensoft Publishers; Kuzmina SA and Korotyaev BA (1987) New species of the pill beetle genus *Morychus* Er. (Coleoptera, Byrrhidae) from the North-East of the USSR. *Entomological Review* 66(2): 342–344 (in Russian); Perkovskiy EE (1999) Species of *Cholevinus* (Coleoptera, Leiodidae, Cholevini) from North of Eastern Siberia and Chukotka. *Vestnik zoologii* 33(3): 66 (in Russian); Perkovskiy EE and Kuzmina SA (2001) Beetles of the genus *Cholevinus* (Coleoptera, Leiodidae, Cholevinae) from the Nord-West of Asia from the Pleistocene to today. *Vestnik Zoologii* 35: 31–38 (in Russian).

The *M. viridis*-dominated LW assemblages were distributed much more widely than just in the Kolyma Lowland. They are also known from the Yedomas sections in the Indigirka and Khroma valleys (Table 3), and much further north, in sites from the Lena Delta. In the Mamontovy Khayata (MKh) section (Figure 7, no. 3), they form a series of samples associated with the termination of the LW (IFZ 4) and dated about 14–12

¹⁴C ka (Figure 21). *M. viridis* exclusively dominates four samples (50–60%). In the Lena Delta, these samples represent the highest content of *Morychus*, steppe and other xerophilic species during the whole Weichselian, so this stage has been recognized as the Late Glacial event, a brief interval of higher summer temperatures and drier environments in this Arctic region (Sher et al., 2005).

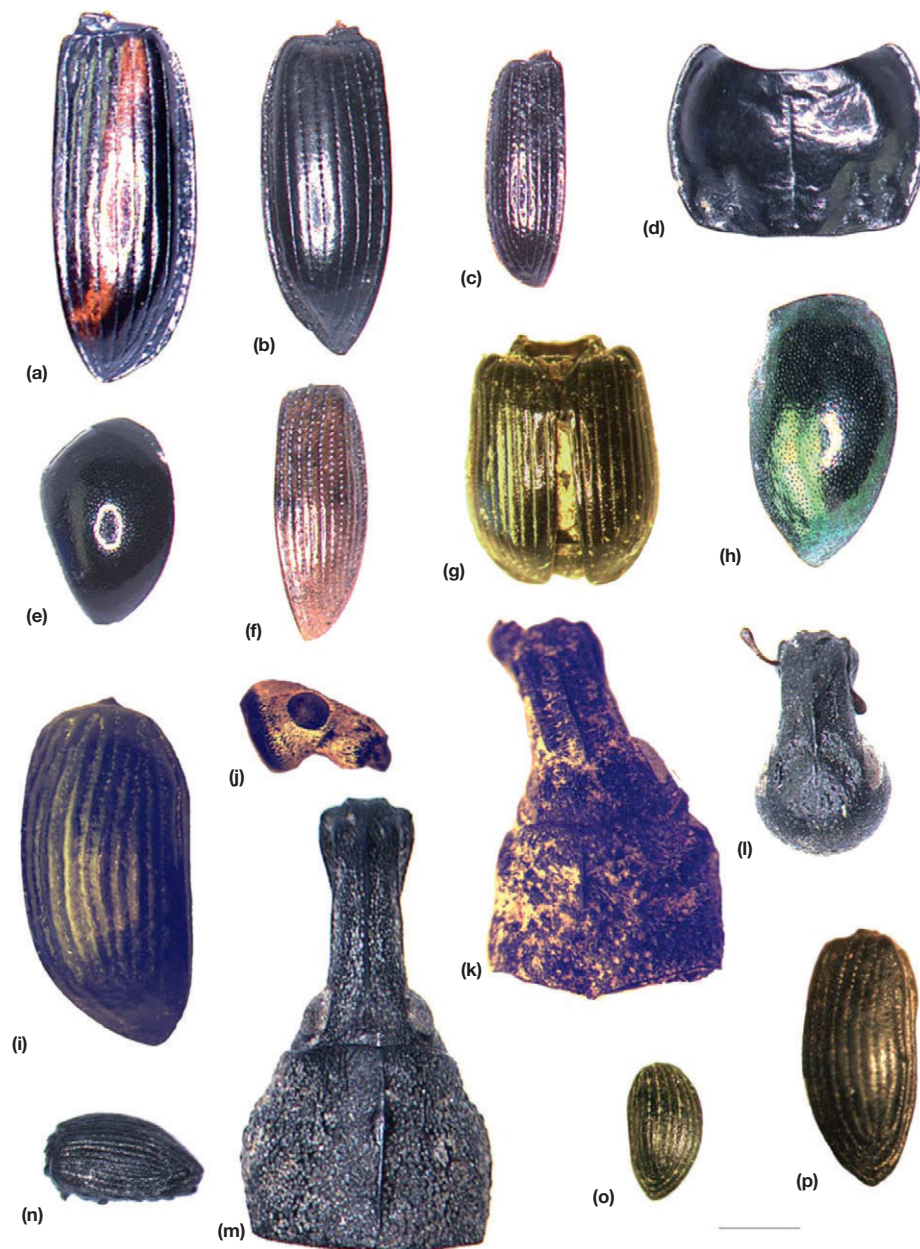


Figure 20 Late Pleistocene fossil beetles from Western Beringia: (a) *Harpalus vittatus kiselevi* Kat. et Shil., right elytron; (b) *Amara glacialis* Mnnh., left elytron; (c) *Dicheirotichus mannerheimi* Sahlb., left elytron; (d) *Pterostichus agonus* Horn., pronotum; (e) *Cyrtoplastus irregularis* Rtt., right elytron; (f) *Helophorus splendidus* Sahlb., right elytron; (g) *Aphodius* sp., body without head and pronotum; (h) *Morychus viridis* Kuzm. et Kor., left elytron; (i) *Chrysolina tolli* Jac., left elytron; (j) *Sitona borealis* Kor., head; (k) *Coniocleonus cinerascens* Hochh., head and pronotum; (l) *Stephanocleonus eruditus* Faust, head; (m) *Lepyrus nordenskiöldi* Faust, head and pronotum; (n) *Isochnus arcticus* Kor., body without head and pronotum; (o) *Mesotrichapion wrangelianum* Kor., left elytron; (p) *Hypera diversipunctata* Schrank., right elytron. (a–f), (h–l), (n) Lena Delta region; (m, o, p) Chukotka. Light microscope photographs, scale bar equals 1 mm.

The LGM insect assemblages in the Lena Delta are strikingly different from those in the Kolyma Lowland. In the MKh section, they are recognized as IFZ 3 (Figure 21). Of nine samples within the range of 24–14 ka, six are strongly dominated by Arctic tundra insects (40–67%), mainly *I. arcticus*. Xerophilic insects constitute only 1.5–3%, with a maximum of 5% of the fauna – the lowest values in the whole Weichselian. Even the ubiquitous *M. viridis* was encountered in only

four of nine samples (below 2%). True steppe insects were totally absent between ~23 and 20 ka and comprised just 1–3% of the faunas in the other samples. These peculiar assemblages (Arctic type of tundra-steppe) are interpreted as evidence of a very cold climate with dry summers (Sher et al., 2005). This reconstruction is corroborated by the plant macrofossil assemblages from the same sites (Kienast et al., 2005).

Table 5 Ecological groups of insects used for quantitative comparison of fossil assemblages

Ecological group and symbol	Group description	Main species
Steppe (st)	The members of this group live today in the steppe zone, some in the mountain belt of southern Siberia and Mongolia. Some of them survive under special microclimatic conditions in the relic steppe patches in NE Asia. However, the presence of any of these species can be considered evidence of unusually high summer temperatures (Berman et al., 2001)	Ground beetles <i>Harpalus pusillus</i> Motsch., <i>Cymindis arctica</i> Kryzh. et Em., leaf beetles <i>Chrysolina perforata</i> Gebl., <i>Ch. brunnicornis bermani</i> Medv., <i>Galeruca interrupta circumdata</i> Duft. weevils <i>Stephanocleonus eruditus</i> Faust, <i>S. fossulatus</i> F-W. <i>S. incertus</i> T-M.
Meadow-steppe (ms)	This group consists of species common in dry (steppe-like) meadows in various ecosystems. They are more cold-tolerant than the previous group. They survive today even in the tundra zone, on the warmest localities	Ground beetles <i>Harpalus vittatus kiselevi</i> Kat. et Shil, soft-winged beetle <i>Troglocollops arcticus</i> L. Medv., the leaf beetle <i>Chrysolina arctica</i> Medv., weevils <i>Phyllobius kolymensis</i> Kor. et Egorov, <i>Coniocleonus cinerascens</i> Hochh., <i>C. ferrugineus</i> Fahr., <i>C. astragali</i> T.-M. et Kor.
Insects of hemicyrophytic steppe (ss)	Hemicyrophytic steppe (Yurtsev, 1981) is a cold steppe-like environment, dominated by a few species of xerophilic sedges and mosses. It has extremely contrasting microclimatic conditions – very hot and dry in summer, and snowless and very cold in winter. It occurs on very dry, snowless hill tops in the taiga or, rarely in tundra zone; (Berman et al., 2001). This environment is currently found in the Kolyma Highlands, Chukotka, and Wrangel Island. It is believed to have many features in common with some kinds of Pleistocene tundra-steppe	Pill beetle <i>Morychus viridis</i> Kuzm. et Kor. is the only representative of this group
Xerophilic insects of various habitats (ks)	Insects of xerophilic habitats of tundra, taiga, and steppe	Ground beetle <i>Notiophilus aquaticus</i> L., dung beetle <i>Aphodius</i> sp.
Insects of dry tundra habitats (dt)	The insects of this group usually occupy the warmest sites in the tundra zone, most commonly well-drained sites with diverse grasses and herbs. Most of them do not reach the Arctic tundra today, but they quite often occupy parts of the taiga zone, in suitable biotopes. Some species can also be found today on relict steppe patches	Ground beetles <i>Bembidion dauricum</i> Motsch., <i>Poecilus (Derus) nearcticus</i> Lth., <i>Pterostichus (Petrophilus) abnormis</i> Sahlb., <i>P. (Tundraphilus) sublaevis</i> Sahlb., <i>Stereocerus haematopus</i> (Dej.), <i>Curtonotus alpinus</i> Payk., <i>Amara interstitialis</i> Dej., <i>A. glacialis</i> Mnnh., <i>Dicheirotichus mannerheimi</i> Sahlb., leaf beetle <i>Chrysolina marginata borealis</i> Medv., weevils <i>Mesotrichapion wrangelianum</i> Kor., <i>Hemitrichapion tschernovi</i> T.-M., <i>Sitona borealis</i> Kor., <i>Hypera ornata</i> Cap., <i>H. diversipunctata</i> Schrank., <i>Lepidophorus thulius</i> (Kiss.)
Insects of typical and arctic tundra (tt)	Most cold-resistant species from different habitats in the Arctic	Leaf beetles <i>Chrysolina tolli</i> Jac., <i>Ch. subsulcata</i> Mnnh., <i>Ch. bungei</i> Jac., weevil <i>Isochnus arcticus</i> Kor.
Insects of mesic tundra habitats (mt)	The insects of this group are now found in different, mostly wet and relatively cold habitats in tundra; some occur on boggy patches in forest tundra	Ground beetles <i>Carabus truncaticollis</i> Esch., <i>Blethisa catenaria</i> Brown. <i>Diacheila polita</i> Fald., <i>Pterostichus (Cryobius) ventricosus</i> Esch., <i>P. (Cryobius) pinguedineus</i> Esch., <i>P. (Cryobius) brevicornis</i> (Kirby), <i>P. (Lenapterus) vermiculosus</i> Men., <i>P. (Lenapterus) costatus</i> Men., <i>P. (Lenapterus) agonus</i> Horn., small carrion beetle <i>Cholevinus sibiricus</i> (Jean.), rove beetles <i>Olophrum consimile</i> Gyll., <i>Tachinus arcticus</i> Motsch., <i>T. brevipennis</i> Sahlb., leaf beetle <i>Chrysolina septentrionalis</i> Men
Shrub (sh)	These beetles feed on shrubs, mostly willows, except dwarf shrubs of typical and Arctic tundra	Leaf beetles <i>Chrysomela blaisdelli</i> Van Dyke, <i>Phratora polaris</i> Schn., <i>Ph. vulgatissima</i> L., weevils <i>Lepyrus nordenskioldi</i> Faust, <i>L. gemellus</i> Kirby, <i>Dorytomus rufulus amplipennis</i> Tourn., <i>Isochnus flagellum</i> Erics
Meadows (me)	Mostly in the forest zone, although some species can occur on tundra meadows	Pill beetles <i>Byrrhus fasciatus</i> Forst., <i>Cytilus sericeus</i> Forst., click beetle <i>Hypnoidus hyperboreus</i> Gyll., leaf beetles <i>Bromius obscurus</i> L., <i>Phaedon concinnus</i> Steph., <i>Ph. armoraciae</i> L., weevils <i>Phyllobius virideaeris</i> Laich., <i>Sitona lineellus</i> Bonnd
Taiga (ta)	Tree-associated insects and insects whose range is restricted to the taiga zone	Ground beetle <i>Pterostichus (Petrophilus) magus</i> Man., click beetle <i>Denticollis varians</i> Germ., weevils <i>Hyllobius piceus</i> DeG., <i>Pissodes insignatus</i> Boh., <i>P. irroratus</i>

(Continued)

Table 5 (Continued)

Ecological group and symbol	Group description	Main species
Riparian (ri)	These insects live near water in several zones, but most of them are restricted to the taiga zone	Reitt., bark beetles <i>Ips cembrae</i> Heer, <i>Polygraphus</i> sp., ants <i>Leptothorax acervorum</i> Fabr., <i>Formica gagatoides</i> Ruzs., <i>Camponotus herculeanus</i> L. Ground beetles <i>Pelophila borealis</i> Payk., <i>Nebria frigida</i> Sahlb., <i>Elaphrus lapponicus</i> Gyll., <i>E. riparius</i> L., <i>Dyschiriodes nigricornis</i> Motsch., <i>Bembidion umiatense</i> Lindrt., <i>Agonum quinquepunctatum</i> Motsch., rove beetles <i>Stenus</i> sp., lady-bird <i>Hippodamia arctica</i> Schneid., leaf beetles <i>Donacia</i> sp., <i>Hydrothassa hannoverana</i> F., weevils <i>Phytobius leucogaster</i> Marsh., <i>Tournotaris bimaculatus</i> F., bugs of Saldidae family
Aquatic (aq)	These insects live in various tundra and taiga water bodies, mostly small ponds	Predaceous diving beetles <i>Hydroporus acutangulus</i> Thoms., <i>Agabus moestus</i> (Curt.), <i>Colymbetes dolabratus</i> (Payk.), whirligig beetle <i>Gyrinus opacus</i> Salb. water scavenger beetles <i>Helophorus splendidus</i> Sahlb., <i>Hydrobius fuscipes</i> F., water bug <i>Sigara</i> sp.
Others (oth)	Insects of intrazonal habitats (such as plant litter, or fungi) and fossils unresolved to a species (of genera with ecologically various members)	Small carrion beetle <i>Cyrtoplastus irregularis</i> Rtt., carrion beetle <i>Thanatophilus dispar</i> Hbst., rove beetles <i>Eucnecus</i> sp., <i>Lathrobium</i> sp., <i>Quedius</i> sp., lady-bird <i>Coccinella</i> sp., minute brown scavenger beetle <i>Corticaria</i> sp.

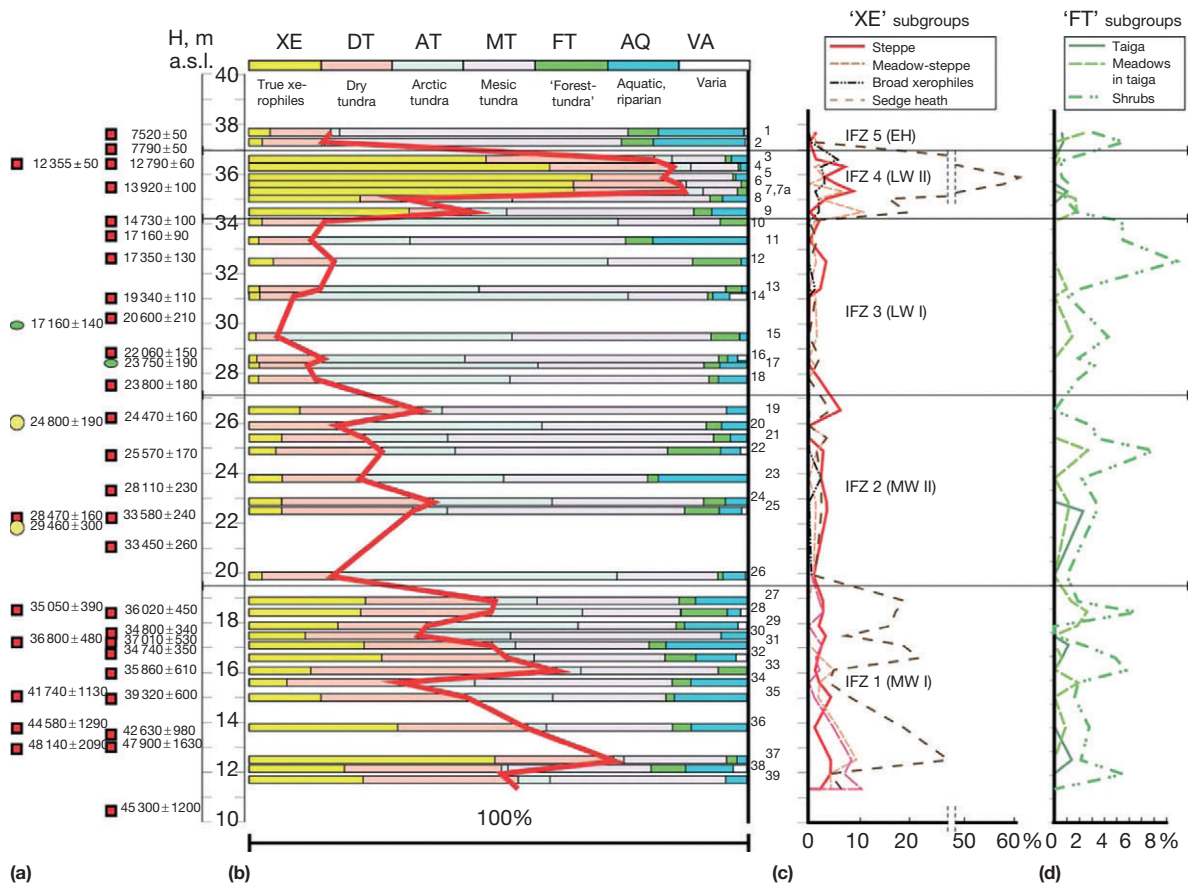


Figure 21 The sequence of fossil beetle assemblages in the Mamontovy Khayata key section, Bykovsky Peninsula, Lena Delta, ranging in age from greater than 50 to 8 ka. Reproduced from Sher AV, Kuzmina SA, Kuznetsova TV, and Sulerzhitsky LD (2005) New insights into the Weichselian environment and climate of the Eastern-Siberian Arctic, derived from fossil insects, plants, and mammals. *Quaternary Science Reviews* 24(5–6): 533–569.

Table 6 Occurrence of xerophilic beetles in the Alyoshkina Zaimka Late Weichselian sequence of samples

Family	Taxon	Eco code	% in the sample	
			Min	Max
Carabidae	<i>Curtonotus fodinae</i> Mnnh.	st	0.1	1.7
Carabidae	<i>Harpalus amputatus amputatoides</i> Mlynar	ms	0.4	1.4
Carabidae	<i>Harpalus vittatus kiselevi</i> Kat. et Shil.	ms	0.4	3.9
Carabidae	<i>Harpalus pusillus</i> Motsch.	st	0.73	0.73
Carabidae	<i>Cymindis arctica</i> Kryzh. et Em.	st	0.7	3.9
Silphidae	<i>Aclypea</i> cf. <i>jacutica</i> (Rjab.)	ms	0.24	0.24
Scarabaeidae	<i>Aphodius</i> sp. nov. ?	ks	0.98	8.92
Byrrhidae	<i>Morychus viridis</i> Kuzm. et Kor.	ss	7.51	64.0
Melyridae	<i>Troglocollops</i> cf. <i>arcticus</i> L. Medv.	ms	0.7	5.9
Chrysomelidae	<i>Chrysolina aurichalcea</i> Mnnh.	ms	2.44	2.44
Chrysomelidae	<i>Chrysolina perforata</i> Gebl.	st	0.65	0.85
Chrysomelidae	<i>Chrysolina aeruginosa</i> Fald.	st	0.48	0.65
Chrysomelidae	<i>Chrysolina brunnicornis bermani</i> Medv.	st	1.45	4.89
Chrysomelidae	<i>Chrysolina rufilabris</i> Fald.	st	0.47	3.91
Chrysomelidae	<i>Colaphellus alpinus</i> Gebl.	ms	0.24	0.24
Chrysomelidae	<i>Galeruca daurica</i> Jeann.	st	0.49	0.49
Apionidae	<i>Loborhynchapion</i> cf. <i>amethystinum</i> Miller	ms	0.5	29.6
Curculionidae	<i>Phyllobius</i> cf. <i>kolymensis</i> Kor. et. Egorov	ms	5.1	29.1
Curculionidae	<i>Phyllobius</i> (<i>Angarophyllobius</i>) sp.	ms	0.72	0.72
Curculionidae	<i>Coniocleonus cinerascens</i> Hochh.	ms	0.97	0.97
Curculionidae	<i>Coniocleonus ferrugineus</i> Fahr.	ms	0.12	0.98
Curculionidae	<i>Coniocleonus astragali</i> T.-M. et Kor.	ms	0.22	0.22
Curculionidae	<i>Stephanocleonus eruditus</i> Faust	st	0.43	4.16
Curculionidae	<i>Stephanocleonus fossulatus</i> F.-W.	st	0.65	1.22
Curculionidae	<i>Stephanocleonus foveifrons</i> Chevr.	st	2.42	13.9
Curculionidae	<i>Stephanocleonus</i> cf. <i>anceps</i> Chevr.	st	0.22	0.22

Table 7 Change of percentage of xerophilic insects along the Late Weichselian section Omolon-I

Estimated age, years BP	Steppe species (st)	<i>Morychus viridis</i> (ss)	Meadow-steppe species (ms)
15 000	6.44	77.65	8.14
17 000	5.49	83.65	3.64
19 000	16.13	41.40	26.88
21 000	1.73	49.14	41.29
23 000	4.29	39.64	41.79
24 000	11.45	76.72	5.34

A similar assemblage is known from the Buor-Khaya exposure on the Central Lena Delta (from the upper part of the Ice Complex), where it is dated about 17 ka (Kuzmina, 2001). Outside the Lena Delta, the Arctic-type tundra-steppe fauna was found in a single, poor sample from the LW Ice Complex in the northern Yana-Indigirka Lowland (Volchya River, 71.7° N).

Thus, during the LGM, tundra-steppe insect faunas of the Arctic type existed only in the northernmost parts of the continent. Further south, they included increasingly thermoxerophilic and steppe elements. For example, at the Krasivoye site on the Malyy Anyuy River (68.3° N; Figure 7, no. 15; Table 3), the Ice Complex assemblage dated about 19 ka was dominated by dry tundra (29%) and xerophilic beetles (53%), with sedge-steppe, meadow-steppe, and true-steppe components each comprising over 15%. Assemblages with an even more pronounced xerophilic component are common in sections (or their members) that are tentatively referred to the LW, all over

Northeastern Siberia. Many exposures in the Kolyma Lowland have yielded faunas with 65–90% xerophiles (alternatively dominated by sedge-steppe or meadow-steppe beetles) in the upper part of the Ice Complex (Duvanny Yar, Lakes Yakutskoye, and others – Figure 7, no. 8; Figure 9; Table 3). In the Indigirka Lowland, the LW sites on the Khroma, Indigirka (Suturuokha), and Yana Rivers are also dominated by xerophiles (60–80%), with *M. viridis* dominating the whole spectrum (30–60%). Similar kinds of insect assemblages (with xerophiles forming 50–95% of the whole spectrum) have been described from the upper part of the Ice Complex of Ayon Island (Utatgyr Suite, Northwestern Chukotka, ~69.5° N – Figure 7, no. 10; Table 3), but here they have an even stronger contribution of meadow-steppe or/and true-steppe species (30–90%). In contrast to this, at the eastern edge of Asia, closer to the Pacific, the steppe component in the LGM faunas is much less pronounced, or absent altogether. In southern Chukotka, for example (the Ice Bluff section on the

Table 8 The insect assemblages of the presumed 'Last Interglacial' age in Northeastern Siberia

Unit	Location	Sediment description	Insect assemblages
Kyl-Bastakh layers	Khroma River (Figure 7, No. 17)	Organic-rich filling of ice-wedge pseudomorphs (the casts of former ice wedges), intruding the top of the Khroma Suite, presumably referred to the late middle Pleistocene, and overlain by the Yedoma Suite (Ice Complex)	Three fossil assemblages are dominated by tundra species (>30%), both mesic and xeric, and the percentage of forest-tundra species is probably the highest elsewhere in the regional record (about 10%). Among the latter, such forest zone species as <i>Phyllobius virideaeris</i> Laich. (a meadow weevil) and <i>Carabus maeander</i> Fisch. (a taiga ground beetle), shrub leaf beetles <i>Chrysomela lapponica</i> L. and <i>Phratora polaris</i> Schn. are recorded. At the same time, the thermophilic xerophiles comprise 15–20% of the whole fauna, and among them are not only the 'sedge-steppe' species <i>Morychus viridis</i> Kuzm. et Kor. and the dung beetles <i>Aphodius</i> sp. nov., but true steppe insects, such as the ground beetles <i>Harpalus pusillus</i> Motsch. and <i>Cymindis arctica</i> Kryzh. et Em., a leaf beetle <i>Chrysolina perforata</i> Gebl. and a weevil <i>Stephanocleonus foveifrons</i> Chev. Remarkably, the assemblages include a few specimens of Arctic tundra leaf beetle <i>Chrysolina subsulcata</i> Mnnh. and an arctic weevil <i>Isochnus arcticus</i> Kor. They also contain a large number of hydrophilic beetles, which is commonly found in peat deposits
Kuobakh unit	Alazeya River (Figure 7, No. 5)	A peat layer and organic-rich ice-wedge pseudomorphs are observed between the late middle Pleistocene Maastakh Suite and the Yedoma Ice Complex	Similar to the Kyl-Bastakh, the assemblages combine a high percentage of hydrophilic and tundra species with a large number (up to 33%) of forest-tundra insects. The latter include mostly the ant <i>Formica gagatoides</i> Ruzs. and various meadow species from the forest zone, such as a click beetle <i>Hypnoidus hyperboreus</i> Gyll., an ant-like beetle <i>Anthicus ater</i> Pz., a leaf beetle <i>Phaedon concinnus</i> Steph., and a weevil <i>Phyllobius virideaeris</i> Laich. Xerophiles comprise up to 8–10% of the faunal assemblages, but do not include steppe species
Krest-Yuryakh unit	Bolshoy Lyakhovsky Island (Figure 7, No. 4)	Two layers of organic-rich ice-wedge pseudomorphs	The unit is in the same stratigraphic position, but the locality is much further north, so it is dominated by mesic tundra species (35–45%) and dry tundra insects (20–30%). However, the Krest-Yuryakh assemblages include both forest-tundra species (2–5%) and up to 10–20% of thermophilic xerophiles, including a few occurrences of steppe species, such as the ground beetle <i>Cymindis arctica</i> Kryzh. et Em., and the weevils <i>Stephanocleonus eruditus</i> Faust and <i>S. fossulatus</i> F.-W.

Main River, [Figure 7](#), no. 12; [Table 3](#)), throughout the interval from 25 to 15 ka, all groups of xerophilic insects rarely reach 30% (usually below 10%) and are mostly represented by *M. viridis*. On the other hand, arctic beetles are abundant in assemblages dated around 21 ka (up to 40%). This is probably indicative of a milder (Pacific) version of the cold-adapted (Arctic-type) LGM faunas.

Middle Weichselian insect assemblages

The Middle Weichselian interval (from about 50 to 25 ka) was traditionally considered as a 'warmer climate interstadial' interval in Northern Asia (called 'Karginsky'), including several cycles of cooling ([Kind, 1974](#)). In many natural sections, where organic-rich deposits (usually peat) occurred under the Ice Complex, or inside it, or separating two 'Ice Complexes' (two members with their own system of polygonal ice wedges), these peats have been ^{14}C dated to 35–40 ka. The pollen spectra of these layers usually show a high percentage of arboreal (mostly shrub) pollen, in contrast to the completely grass-herb-dominated spectra of the Ice Complex itself. In recent years, however, several important new discoveries have been made about this interval. First, repeated ^{14}C dating showed that most of the samples that yielded ^{14}C ages near 35 ka and older, were indeed much older and could actually represent ages beyond the ^{14}C age range ([Sher and Plakht, 1988](#)). Second, it was shown that the 'Karginsky' sections in the type area (Lower Yenisey) had been misinterpreted, and their dating was erroneous. In fact, they turned out to be much older, probably as old as the last interglacial ([Astakhov, 2001](#)). Third, new radiocarbon ages from the regional type sections of the 'Karginsky Interstadial' in Northeastern Siberia also yielded much older dates ([Sulerzhitsky, 1998](#)). Finally, the multidisciplinary study of a continuous section through most of the LW in the Lena Delta, well dated by ^{14}C , revealed a completely different regional history of this period ([Andreev et al., 2002](#); [Schirmermeister et al., 2002](#)). The study of the detailed record of fossil insect assemblages through this sequence allowed us ([Sher et al., 2005](#)) to suggest a very different reconstruction of Middle Weichselian environments.

It has been shown that tundra-type insect assemblages persisted in the Lena Delta throughout the Middle Weichselian interval, with a variable, but almost consistent presence of xerophilic species. Although today the northern limit of taiga (sparse larch forest) is not far south of the site, no fossil assemblages of the forest or even forest-tundra type have ever been found here. The contribution of xerophilic species was more noticeable in the first half of the Middle Weichselian interval, especially in the assemblages ^{14}C dated about 45–48 ka, when they reached 30–50%. The possibility cannot be excluded, however, that these assemblages are actually Early Weichselian in age. Taken together with the dry tundra species contribution, the percentage of relatively thermophilic forms gradually decreased from 48 to 34 ka (IFZ 1; [Figure 21](#)). At the same time, the contribution of arctic species increased, and by 34 ka it almost reached LGM levels (more than 50%). The second half of the Middle Weichselian (34–24 ka, IFZ 2 on [Figure 21](#)) is marked by the lowest levels of xerophilic species (usually below 7%) and the predominance of mesic tundra insects, sometimes with a high contribution of arctic species. This sequence, the most complete in Northern Asia, is

interpreted as a gradual decrease in summer temperatures toward the LGM, probably accompanied by some increased moisture. However, these moisture levels did not reach the levels observed in modern tundra environments. This indicates that during the Middle Weichselian, the climate was more continental than today in this coastal area. This conclusion is in agreement with various observations that sea level in the Eastern Arctic remained low at that time.

Several sites in more southern areas of Northeastern Siberia yielded representative insect assemblages obtained from the Ice Complex and associated with late Middle Weichselian ^{14}C dates. For example, in the Shamanovo Yedomia section on the Indigirka River ([Figure 7](#), no. 16; [Table 3](#)) a rich insect sample came from the unit dated by *in situ* grass roots at 29 300–32 900 years BP ([Kaplina et al., 1980a,b](#)). The assemblage is dominated by *M. viridis* and other xerophilic insects (66%) and dry habitat species prevail in the tundra group ([Kiselyov, 1994](#)). On the Khroma River ([Figure 7](#), no. 17; [Table 3](#)), the assemblage dated 28 800 years BP ([Kaplina et al., 1983](#)) has the same percentage of xerophilic species, but *M. viridis* is much more abundant, while the contribution of steppe species is low.

Interesting assemblages, presumably associated with the Middle Weichselian, were found in the Primorsky alas section in western Chukotka ([Figure 7](#), no. 18; [Table 3](#)). The sampled unit contains tree birch wood, and is dated more than 27 870 years BP ([Kiselyov, 1994](#)). The assemblages ([Table 7](#)) include unusually high percentages of forest-tundra species (7%), comprising such forest inhabitants as the rove beetles *Acidota* sp., *Arpedium* sp., *Xantholinus* sp., *Anthobium* sp., and *Bolitobius* sp., as well as the birch-associated weevil *Betulapion simile*. On the other hand, these assemblages contain high percentages of *M. viridis* and a few true steppe beetles. A peat layer from about 12 m higher in the section is dated at 24 800 ± 400 years BP. Two insect assemblages from this unit are similar to the lower ones, but show even higher percentages of forest-tundra beetles (up to 12%) represented by the same species. The steppe, meadow-steppe beetles and *M. viridis* are still present, although in lower numbers.

However, in southern Chukotka, insect assemblages from the same age interval (34–24 ka) (Ice Bluff locality; [Figure 7](#), no. 12; [Figure 17](#); [Table 3](#)) do not contain any forest beetles ([Kiselyov, 1994](#)). The percentage of xerophilic species is high (38%), mostly at the expense of dung beetles (*Aphodius* sp. nov.?) and *M. viridis*. True steppe and meadow-steppe species are represented only by single specimens.

Early Weichselian insect assemblages

Taking into account the dating problems, it is difficult to confidently assign any fossil insect assemblage to the Early Weichselian interval. On the other hand, the lower parts of many thick Ice Complex sections may belong to this time interval. For example, in the key section of the Ice Complex of Duvannyy Yar ([Figure 7](#), no. 8; [Figure 9](#); [Table 3](#)), a large number of infinite ^{14}C dates (>45 ka) in the lower part of the section (mixed with dates in the range of 33–45 ka) may suggest an Early Weichselian age ([Sher, 1991](#)). The same is possible for the Oyagosskiy Yar assemblage ([Figure 7](#), no. 19; [Table 3](#)) and the sections on the southern coast of Bol. Lyakhovsky Island ([Figure 7](#), no. 4; [Table 3](#)). In the Kolyma

Lowland, the insect assemblages from the lower parts of Ice Complex sections are dominated by *M. viridis* with a small admixture of other xerophiles (usually below 10%) (Duvanny Yar; Sher et al., 1979). However, the northern localities, such as Oyagos, Lyakhovsky, or Nagym in the Lena Delta (Figure 7, no. 1; Table 3), are mostly dominated by tundra species, sometimes with a high percentage of Arctic tundra species or mesic tundra inhabitants. The total percentage of all xerophiles is usually below 15%, and the main role in this group is played by *M. viridis*, while true steppe species are either extremely rare or completely absent (Kuzmina, 2001).

Thus, it may be concluded that the same regularities in spatial distribution of insect assemblages in the LW existed during Early Weichselian time.

'Last Interglacial' insect assemblages

The problems of recognition of Last Interglacial sediments were discussed above (cf. also Sher, 1991). Tentatively, the authors refer some parts of particular complex sections to this interval. These sediments overlie sediments of presumed middle Pleistocene age and they underlie the LP Ice Complex. Also, they contain abundant organic remains. Such localities usually represent buried alas complexes (Table 3). The characteristics of insect assemblages forming three of them are shown in Table 8.

The most remarkable character of all these assemblages is the extreme heterogeneity of their ecological composition, combining representatives from practically all ecological groups. Many species comprising those assemblages do not occur together today. At present, their ranges are often separated by many hundreds of kilometers. This is especially evident for the northernmost assemblages (Bolshoy Lyakhovsky Island), where about 80% of identified insect species currently live much further south. This indicates the non-analog character of the assemblage, which certainly existed in a warmer climate than today, with higher summer and probably somewhat higher winter temperatures. According to the paleobotanical data (Andreev et al., 2004), summer temperatures could have been 4–5 °C higher than present. In general, the climate was more continental than today on Lyakhovsky Island, but less so than during the Weichselian interval. Compared with the modern environment of the island, the humidity was certainly lower in summer, but probably somewhat higher in winter. A thicker snow cover would have allowed the growth of tall shrubs at this high latitude.

Summary

This review demonstrates the large scope of the Pleistocene beetle research in Northeastern Siberia during the recent years. However, many gaps remain in the knowledge shared in this article and some significant problems are still unsolved. Evidently, the characterized beetle faunas are peculiar, but the authors have a poor idea of how far south and west they were distributed. Very few fossil assemblages are known south of the Arctic Circle in this region, and none to the west of the Lena Delta. They also have serious problems with the dating of fossil assemblages older than 30 ka. The authors believe that future research should be concentrated on continuous sedimentary

sequences, such as the Bykovsky section, and this research must be a part of broad multidisciplinary projects. Recent experience shows that successful projects of such kinds must be based on international collaboration.

See also: Beetle Records: Overview. Permafrost and Periglacial Features: Ice Wedges and Ice-Wedge Casts.

References

- Alfimov AV, Berman DI, and Sher AV (2003) Tundra-steppe insect assemblages and reconstruction of late Pleistocene climate in the lower reaches of the Kolyma River. *Zoologicheskii Zhurnal* 82(2): 281–300 (in Russian).
- Anderson PM, Kotov AN, Lozhkin AV, and Trumpe MA (2002) Palynological and radiocarbon data from Late Quaternary deposits of Chukotka. In: Anderson PM and Lozhkin AV (eds.) *Late Quaternary Vegetation and Climate of Siberia and Russian Far East (Palynological and Radiocarbon Database)*, pp. 35–79. Magadan: NESF FEB RAS.
- Andreev AA, Grosse G, Schirmermeister L, et al. (2004) Late Saalian and Eemian palaeoenvironmental history of the Bol'shoy Lyakhovsky, Island Laptev Sea region. Arctic Siberia. *Boreas* 33(4): 319–348.
- Andreev AA, Schirmermeister L, Siegert CH, et al. (2002) Paleoenvironmental changes in Northeastern Siberia during the Late Quaternary – Evidences from pollen records of the Bykovsky Peninsula. *Polarforschung* 70: 13–25.
- Astakhov V (2001) The stratigraphic framework for the Upper Pleistocene of the glaciated Russian Arctic: Changing paradigms. *Global and Planetary Change* 31(1–4): 283–295.
- Berman DI (1986) Invertebrate fauna and population in tundra-steppe communities of Wrangel Island. *Biogeography of the Subarctic Beringian Zone, Vladivostok*, 146–160 (in Russian).
- Berman DI and Alfimov AV (1993) Microclimatic conditions of existence of steppe ecosystems in the Subarctic of north-east Asia. *Bulletin of Moscow Society of Naturalists, Biology Section* 98(3): 118–128 (in Russian).
- Berman DI, Alfimov AV, Mazhitova GG, Grishkan IB, and Yurtsev BA (2001) *Cold Steppe in North-Eastern Asia*, p. 183. Vladivostok: Dal'nauka (in Russian).
- Bienkowski AO (1999) A review of the species of the subgenus *Chrysolina* (Pleurosticha) Motschulsky (Insecta: Coleoptera, Chrysomelidae). *Entomologische Abhandlungen* 58(10): 165–179.
- Grushevskiy II and Medvedev LN (1962) Preliminary data on the application of coleopterological analysis to the study of Quaternary deposits of Northern Yakutiya. In: Shvedov NA (ed.) *Collected Articles on Paleontology and Biostratigraphy of Scientific Research Institute of Arctic Geology*, pp. 38–72. Leningrad: Scientific Research Institute of Arctic Geology (in Russian).
- Kaplina TN (1981) History of frozen sequences of northern Yakutia in the late Cenozoic. In: Dubikov GI and Baulin VV (eds.) *Istoriya razvitiya mnogoletnemerzlykh porod Evrazii*, pp. 153–181. Novosibirsk: Nauka (in Russian).
- Kaplina TN, Lakhtina OV, and Rybakova NO (1981) Cenozoic deposits in the middle course of the Alazeya R. (Kolyma Lowland). *Izvestiya Akademii Nauk SSSR Seriya Geologicheskaya* 8: 51–63 (in Russian).
- Kaplina TN, Ovander MG, Lozhkin AV, et al. (1983) Quaternary deposits in the middle current of the Khroma River (Yana-Indigirka Lowland). In: Shilo NA (ed.) *Stratigrafiya i paleogeografiya pozdnego kaynozoya Vostoka SSSR*, pp. 80–95. Magadan: SV KNII DVNTs AN SSSR (in Russian).
- Kaplina TN, Sher AV, Giterman RE, et al. (1980a) Key section of Pleistocene deposits on the Allaikha river (lower reaches of the Indigirka). Bulletin of the Commission on Quaternary Period Research. *USSR Academy of Sciences* 50: 73–95 (in Russian).
- Kaplina TN, Shilova G, and Pirumova LG (1980b) Shamanovo key section of late Pleistocene and Holocene deposits on the Indigirka. *Izvestiya Akademii Nauk SSSR Seriya Geologicheskaya* 9: 74–171 (in Russian).
- Khruleva OA (2004) Tundra-steppe leaf beetle *Chrysolina brunnicornis wrangeliani* (Coleoptera, Chrysomelidae): Distribution, life history and habitats. In: Jolivet P, Santiago-Blay JA, and Schmitt M (eds.) *New Developments in the Biology of Chrysomelidae*, pp. 541–550. The Hague: SPB Academic.
- Kienast F, Schirmermeister L, Siegert CH, and Tarasov P (2005) Palaeobotanical evidence for warm summers in the East Siberian Arctic during the last cold stage. *Quaternary Research* 63(3): 283–300.

- Kind NV (1974) *Late Quaternary Geochronology According to Isotope Data (Geokhronologiya Pozdnego Antropogena Po Izotopnym Dannym)*, p. 255. Moscow: Nauka (in Russian).
- Kiselyov SV (1981) *Late Cenozoic Coleoptera of North-East Siberia*. Moscow: Nauka (in Russian).
- Kiselyov SV (1994) *Northern Eurasia environment in the Pleistocene and Holocene (according to Coleoptera researches)*. Thesis in Geology, p. 32. Moscow: MSU (in Russian).
- Kiselyov SV, Kolesnikov S, and Rybakova NO (1987) Climate of the vegetation period during the 'Ice Complex' deposits forming from the Omolon River Site. *Byulleten Moskovskogo obshchestva ispytatelei prirody* 62: 113–119 (in Russian).
- Korotyaev BA (1996) 165. *Sitona* Germ. In: Egorov AB, Zherichin VV, and Korotyaev BA (eds.) *Sem. Curculionidae – Dolgonosiki. Key to the insects of the Russian Far East*, vol. 3, Coleoptera, Pt. 3, Supplement, pp. 499–503. Vladivostok: Dal'nauka (in Russian).
- Kryzhanovskiy OL (1965) Fam. Carabidae – Ground beetles. *Guide Book to the Identification of the Insects of European Part of USSR*, vol. 2, pp. 29–77. Moscow-Leningrad: Nauka (in Russian).
- Kryzhanovskiy OL, Belousov IA, Kabak II, Kataev BM, Makarov KV, and Shilenkov VG (1995) *A checklist of the ground-beetles of Russia and adjacent lands (Insecta, Coleoptera, Carabidae)*, p. 272. Sofia, Moscow: Pensoft Publishers.
- Kuzmina SA (2001) *Quaternary Insects of the Coastal Lowlands of Yakutia. Thesis in Biology*. Moscow: Paleontological Institute. RAS (in Russian).
- Kuzmina SA and Korotyaev BA (1987) New species of the pill beetle genus *Morychus* Er. (Coleoptera, Byrrhidae) from the North-East of the USSR. *Entomological Review* 66(2): 342–344 (in Russian).
- Medvedev LN and Voronova NN (1977) Coleopterological analysis of geological sections of mammoth cemeteries in northern Yakutia. In: Scarlato O (ed.) *The mammoth fauna of the Russian Plain and Eastern Siberia*, pp. 72–76. Leningrad: Nauka.
- Nikolsky PA and Basilyan AE (2003) Svyatoy Nos Cape – The key section of the Pleistocene in the North of Yana-Indigirka Lowland. *The Natural History of the Russian East Arctic in the Pleistocene and Holocene*, pp. 5–13. Moscow: GEOS Publishers (in Russian).
- Perkovskiy EE (1999) Species of *Cholevinus* (Coleoptera, Leiodidae, Cholevini) from North of Eastern Siberia and Chukotka. *Vestnik zoologii* 33(3): 66 (in Russian).
- Perkovskiy EE and Kuzmina SA (2001) Beetles of the genus *Cholevinus* (Coleoptera, Leiodidae, Cholevinae) from the Nord-West of Asia from the Pleistocene to today. *Vestnik Zoologii* 35: 31–38 (in Russian).
- Pitul'ko VV, Pavlova EYu, Kuz'mina SA, et al. (2007) Natural climatic changes in the Yana-Indigirka lowland during the terminal Kargino time and habitat of late Paleolithic man in northern part of East Siberia. *Doklady Akademii Nauk* 417(1): 103–108.
- Schirmermeister L, Andreiev A, Grosse G, and Meyer H (2004) *Saalian to Holocene Paleoenvironmental History Documented in Permafrost Sequences of Arctic Siberia (New Siberian Archipelago, Bolshoy Lyakhovsky Island)* AGU Fall Meeting, 13–17 December 2004, San Francisco, California C13B-0286.
- Schirmermeister L, Kunitsky VV, Grosse G, et al. (2003) Late Quaternary history of the accumulation plain north of the Chekanovsky Ridge (Lena Delta, Russia) – A multidisciplinary approach. *Polar Geography* 27(4): 277–319.
- Schirmermeister L, Siegert C, Kuznetsova T, et al. (2002) Paleoenvironmental and paleoclimatic records from permafrost deposits in the Arctic region of northern Siberia. *Quaternary International* 89: 97–118.
- Sher AV (1988) Problems of ecology of Pleistocene mammals. *Aktual'nye problemy morfologii i ekologii vysshikh pozvonochnykh*, pp. 432–464. Moskva: Izdatel'stvo AN SSSR, Chast' II (in Russian).
- Sher AV (1990) Actualism and disconformism in the studies of Pleistocene mammals ecology. *Journal of General Biology* 51(2): 163–177.
- Sher AV (1991) Problems of the last interglacial in Arctic Siberia. *Quaternary International* 10–12: 215–222.
- Sher AV, Kaplina TN, and Giterman RE, et al. (eds.) (1979) *Late Cenozoic of the Kolyma Lowland: XIV Pacific Science Congress, Tour Guide XI, Khabarovsk August 1979*, pp. 1–116. Moscow: Academy of Sciences of the USSR.
- Sher AV, Kuzmina SA, Kuznetsova TV, and Sulerzhitsky LD (2005) New insights into the Weichselian environment and climate of the Eastern-Siberian Arctic, derived from fossil insects, plants, and mammals. *Quaternary Science Reviews* 24(5–6): 533–569.
- Sher AV and Plakht IR (1988) Radiocarbon dating and problems of the Pleistocene stratigraphy in lowlands of the Northeast USSR. *International Geology Review* 30(8): 853–867. *Izvestiya AN SSSR, ser. Geol.* 8, 17–31 (in Russian).
- Sulerzhitsky LD (1998) On the microbial contamination organic matter from permafrost observed during radiocarbon dating. *Kriosfera Zemli (Earth Cryosphere)* 2(2): 76–80 (in Russian).
- Svensden JJ, Alexanderson H, Astakhov VI, et al. (2004) Late Quaternary ice-sheet history of Northern Eurasia. *Quaternary Science Reviews* 23: 1229–1271.
- Svitoch AA (1980) The description of the lates deposits [of the Ayon Island]. In: Kaplin PA (ed.) *The latest deposits and paleogeography of the Pleistocene in Chukotka*, pp. 180–202. Moscow: Nauka.
- Tugarinov AYa (1929) On the origin of arctic fauna. *Priroda* 7–8: 653–680 (in Russian).
- Vangengeim EA (ed.) (1977) *Paleontological foundation of Anthropogene stratigraphy in Northern Asia (by mammals)*, pp. 1–172. Moscow: Nauka (in Russian).
- Vangengeim EA and Ravsky EI (1965) On the intra-continental type of natural zonality in Eurasia during the Quaternary Period (Anthropogene). In: Menner VV (ed.) *The problems of Cenozoic stratigraphy*, pp. 128–141. Moscow: Nedra Publishers.
- Yurtsev BA (1981) *Relict steppe vegetation complexes of Northern Asia (Problems of reconstruction of the cryoxerotic landscapes of Beringia)*. Novosibirsk: Nauka (in Russian).
- Zinoviev EV (1997) Quaternary insects of West Siberian Plain. In: Olynvng VN (ed.) *Successes of Entomology on Ural*, pp. 153–157. Ekaterinburg: UrO REO (in Russian).

Introduction

Insect analyses carried out from Holocene sequences are relatively rare in Britain and Northern Europe, in contrast to late Pleistocene and Late Glacial sequences which have been thoroughly investigated during the last 50 years. Therefore, the early to mid-Holocene history of the Coleopteran (beetle) fauna in Western Europe relies for the most part on the analysis of scattered assemblages, obtained from natural or seminatural deposits. Late Holocene studies, especially from the historical period, are much more abundant, but the vast majority of data concerning this period derive from analyses carried out in archaeological sites from anthropogenic contexts and are dominated by synanthropic insects. These studies provide a wealth of data, mainly concerning the living conditions of man. However, they offer relatively restricted information on past environments, although some archaeological deposits have yielded beetle assemblages providing a detailed picture of the landscape surrounding human settlements (see, e.g., Ponel et al., 2000). Nonanthropogenic organic deposits contemporaneous with human occupation are usually more effective to reconstruct the environmental history in a wider area (see, e.g., the reconstruction of the landscapes along the Thames Valley from the Mesolithic to the Late Bronze Age by Elias et al., 2009). A further complication hampering the paleoenvironmental interpretation of Postglacial beetle records is the growing influence of man on natural ecosystems from the mid-Holocene onwards, through forest clearance, development of agriculture, pastoralism, and fragmentation of habitats. These landscape modifications make it difficult to disentangle the respective role of climate and human impact on beetle faunas.

Considering the wide range of subjects treated by Quaternary entomologists and archaeoentomologists from Postglacial beetle records, this article is necessarily focused on a selection of issues, such as changes in beetle faunal assemblages at the onset of the Holocene and their paleoclimate and paleoenvironmental implications, impact of human activities on landscapes and beetles (especially the effect of forest clearance on saproxylophagous (sap- and wood-feeding) species), and reconstruction of paleoenvironmental changes in Mediterranean mountains from continuous sedimentary sequences.

Climatic Context

The Quaternary is marked by a climatic cyclicity of cold and warm periods (glacial–interglacial), and by the appearance and evolution of humans. The Postglacial or Holocene Epoch that started about 11 000 years ago is the last warm period of the Quaternary (the current interglacial) and can be considered an approximate replica of the penultimate interglacial, the Eemian (130 000–110 000 BP). The

Holocene interglacial has been deeply affected by human impact on natural ecosystems, especially in its second half. The glaciation that immediately preceded the Holocene is biologically important since glacial conditions that prevailed at that time (60 000–15 000 BP) eradicated most of temperate biota from Central and Northern Europe, including insects, which were relegated to temperate refugia (probably located in Mediterranean peninsulas). The insect recolonization of Europe induced by the Holocene warming is the result of the northwards shift of biota from these southern ‘reservoirs.’ Far from being a steady process, the transition from the last glaciation to the current interglacial, approximately 15 000 and 10 000 BP, was marked by a complex series of sharp warm and cold events. This period of climatic instability, commonly named the Late Glacial, includes a first climatic warming between 15 000 and 13 000 BP (the Oldest Dryas), a predominantly warm period between 13 000 and 10 700 BP, the Late Glacial interstadial, and a short but drastic climatic cooling between 10 700 and 10 300 BP (the Younger Dryas). The Late Glacial is particularly well documented from beetle records (Coope, 1994), and one of the major achievements of Quaternary Entomology is the demonstration of the amazing rapidity of the Late Glacial warming at 13 000 BP, since mean July temperatures in Britain rose by about 1 °C per decade (for a total of about 7 °C), and corresponding winter temperatures rose by about 20 °C (Atkinson et al., 1987). Following the second episode of rapid warming (by 10 000 BP), temperate climate conditions prevailed during most of the Holocene.

Vegetation History

Deciduous forest (with hazel, oak, elm, and linden) spread throughout much of Europe by 10 000 BP. Based on paleobotanical reconstructions, a thermal optimum is identified between 8000 and 6000 BP, and it is at that time that an unprecedented event, the Neolithic ‘revolution,’ arose in the Near East and expanded quickly into Europe, inducing forest clearance and agriculture. By 5000 BP, the ‘Urwald,’ or primitive forest, started to shrink in most of lowland Europe, including Britain. ‘Slash and burn’ cultivation practices induced a deep change in plant communities, as indicated by pollen analysis: there was a decline of many forest taxa, and a rise of heliophilous plants (plants adapted to full sunlight in open ground situations) and many weeds (*Plantago*, *Urtica*, plants associated with disturbance). The Iron Age (around 3500–2500 BP) corresponds with a time of marked human population growth, furthering the development of agricultural landscapes. By 2500 BP the intensification of agricultural disturbance on natural vegetation structures is superimposed with a climate change at the Subboreal (warm/dry climate)–Subatlantic (cool/wet climate) transition.

The Onset of the Holocene: Beetles at the Younger Dryas–Preboreal Transition

Shortly before the climatic and environmental changes that occurred in Europe during the early Holocene (Preboreal interval), an arctic/alpine beetle fauna, including many species restricted today to very high latitudes in arctic regions of Scandinavia and northern Russia, thrived in Britain and in most of Europe. This cold-adapted fauna is associated with the Younger Dryas (from 10 700 to about 10 300 BP), a short but intense cold event. Most of these species were already present at the end of the last glaciation, about 15 000 BP, but were temporarily eradicated from mid-latitudes by the warm episode of the Late Glacial interstadial. This extremely cold-adapted fauna (Figure 1) includes many characteristic species such as:

<i>Nebria nivalis</i>	<i>Amara alpina</i>
<i>Diacheila arctica</i>	<i>Pterostichus middendorffi</i>
<i>Elaphrus lapponicus</i>	<i>Agonum consimile</i>
<i>Dyschirius septentrionum</i>	<i>Agonum sahlbergi</i>
<i>Bembidion dauricum</i>	<i>Helophorus fennicus</i>
<i>Bembidion fellmanni</i>	<i>Helophorus glacialis</i>
<i>Bembidion hasti</i>	<i>Helophorus obscurus</i>
<i>Bembidion hyperboreaum</i>	<i>Arpedium brachypterum</i>
<i>Bembidion lapponicum</i>	<i>Boreaphilus henningianus</i>
<i>Bembidion mckinleyi</i>	<i>Olophrum boreale</i>

Some Siberian and eastern Asiatic species (*Pterostichus magus*, *Helophorus jacutus*, *Tachinus jacuticus*) suggest a climate more continental than today. At the time of the Younger Dryas the landscape in most of Europe was comparable to an arctic tundra, and the average July temperature was close to 10 °C.

At the close of the Late Glacial, by 9500 BP, this very cold-adapted fauna was completely and rapidly eradicated. In the British Isles and certainly also in most of Central and Northern

Europe, the representatives of the arctic–alpine fauna disappeared suddenly, to be rapidly replaced by a thermophilous (warm-adapted) fauna, including various saproxylophagous species (beetles associated with mature old trees). The early Holocene horizons from the sedimentary sequence at Holywell Coombe (Coope, 1998) yielded such Postglacial assemblages, with many rather thermophilous Coleoptera, such as:

<i>Clivina fossor</i>	<i>Prosternon tessellatum</i>
<i>Bembidion octomaculatum</i>	<i>Denticollis linearis</i>
<i>Bembidion biguttatum</i>	<i>Melasis buprestoides</i>
<i>Patrobus atrorufus</i>	<i>Eucnemis capucina</i>
<i>Pterostichus strenuus</i>	<i>Dirhagus pygmaeus</i>
<i>Pterostichus minor</i>	<i>Anaglyptus mysticus</i>
<i>Platynus assimilis</i>	<i>Pogonochaerus hispidulus</i>
<i>Licinus depressus</i>	<i>Taphrorychus villifrons</i>
<i>Megarthritis hemipterus</i>	<i>Coenorhinus aeneovirens</i>
<i>Olophrum piceum</i>	<i>Rhyncholus elongatus</i>
<i>Lesteva heeri</i>	

This faunal change clearly suggests the rapid establishment of a climate marked by temperatures as high, or even higher, than those of the present day. Elsewhere in Europe the same pattern is also recorded from a variety of sites scattered from Scandinavia (Lemdahl, 1997) to Central France (Ponel and Coope, 1990), Switzerland (Lemdahl, 2000), and Italy (Ponel, 1997).

The very rich fauna from Lea Marston, analyzed by Osborne (1974) and dated to about 9500 BP, is especially interesting since most of the beetles from the assemblages are rather thermophilous woodland species (such as *Calosoma inquisitor* and *Xylodrepa quadripunctata*, predators on tree-feeding caterpillars). This fauna is compatible with a mixed oak forest, and therefore suggests a warm climate with summer temperatures on the order of 16–17 °C, comparable with that of southern England or southern Sweden today. However, pollen evidence shows that the landscape was dominated at that time by willow and birch woodland, not by thermophilous trees. This time lag between the colonization of warm-adapted beetles during a period of very rapid climate amelioration and the northwards thermophilous tree expansion is analogous to that described by Coope and Brophy (1972) at the onset of the Late Glacial interstadial and should be attributed to the interaction of various factors, such as tree dispersal rate, location of refuges, soil maturation, competition, and ecological succession. A similar observation is made by Coope (1998) at Holywell Coombe: the lithological boundary between clayey Late Glacial ‘cold’ deposits and more organic Postglacial ‘warm’ deposits is not synchronous with the faunal change, since the first occurrences of thermophilous beetles are recorded earlier, that is to say below the lithological boundary. Again, beetles seem to respond more rapidly than plants to the warming and appear in the area before the rise in organic sedimentation triggered by the expansion of vegetation.



Figure 1 *Helophorus glacialis* (Coleoptera: Helophoridae). This cold-adapted boreomontane water beetle lives at the edge of melting snow patches at sea level in the north of its distribution and at high altitude in Mediterranean mountains. It is abundant in cold periods during the Late Glacial and becomes increasingly scarce in most of European sedimentary sequences at the onset of the Holocene.

Climatic Indications from Coleoptera During the Holocene

The recent synthesis by Mayewski et al. (2004) provides evidence for a highly variable climate during the Holocene and reveals at least six periods of rapid climate change at a global

scale. Concerning beetles however, as clearly stated by Coope (1986), "it is easy to demonstrate that many species of coleoptera have had their geographical ranges drastically altered during this time, but it is much more difficult to sort out the contributions made by changes in climate from those imposed by human interference." The quantification of Holocene climate changes reconstructed from beetle data is therefore a relatively neglected issue in spite of the high number of sites investigated, especially in the British Isles. Computerized climatic reconstruction methods such as mutual climatic range have rarely been used.

Despite these limitations, several attempts to retrace Holocene climate changes from beetle records have been made in Britain (Buckland and Wagner, 2001; Kenward, 2004; Osborne, 1976, 1997). Some useful inferences may be drawn from beetles not associated with dead wood habitat, since saproxylophagous taxa are more likely to respond to forest clearance by man than to climate changes. After the initial rapid rise in temperature 10 000 years ago, the following period is rather poorly known and beetle assemblages do not show any obvious evidence of significant climate fluctuations until about 4830 BP. At that time a climatic deterioration is suggested by the occurrence in the British Midlands (Kelly and Osborne, 1965) of several 'northern' Coleoptera (*Bembidion atroviolaceum*, *Sphaerites glabratus*, *Tropiphorus obtusus*, *Barynotus squamosus*). This cool phase seems relatively short since Girling (1984) described an early Neolithic beetle assemblage (5000 BP) suggesting temperatures slightly warmer than today. This fauna contained the thermophilous and continental ground beetle species *Oodes gracilis*.

Younger assemblages do not indicate thermal conditions very different from those of the present time. A marked change is recorded in Bronze Age assemblages (Osborne, 1982, 1988), with the appearance of an apparently warm fauna including several dung beetles (mainly *Onthophagus*) that today are extremely rare or even extinct in Britain. These assemblages probably suggest a thermal climate warmer than today during the Bronze Age around 4000–3000 BP. The late Iron Age site of Fisherwick, at about 2130 BP (Osborne, 1979) yielded fossils of the dung beetle *Heptaulacus testudinarius*, again a very southern species in Britain, and is interpreted as evidence for a warm climate. However, the climatic significance of these dung beetle occurrences remains a matter of debate since it is possible that pastures may have suffered of human perturbations, leading to the retraction or eradication of beetles associated with old grasslands (Buckland and Dinnin, 1993).

The vast majority of records for the historic period have been obtained from urban anthropogenic contexts such as heated buildings, allowing many thermophilous and synanthropic species to survive outside their natural range and blurring the climatic signal. Nevertheless, Kenward (2004) suggests that a careful interpretation of data concerning the nonsynanthropic elements included in insect assemblages from archaeological occupation deposits may indeed provide useful paleoclimatic data.

Climatic reconstructions may be drawn from the comparison of past and modern ranges of *Heterogaster urticae*, a bug mainly dependent on *Urtica dioica*, a very common nettle present everywhere in Britain. The nettlebug is now restricted to southern England but numerous fossil records from the first to sixteenth century (i.e., before the Little Ice Age) in northern England suggest that in this region temperatures were probably 1

or 2 °C higher than today (Kenward, 2004). The fossil occurrence in Roman sites of several other bugs (*Sehirus luctuosus*, *Syromastus rhombeus*, *Aphanus rolandri*), and also beetles (*Airaphilus elongatus*, *Heptaulacus sus*, *Heptaulacus testudinarius*, *Odacantha melanura*, *Plateumaris braccata*, *Bagous czwalinai*), to the north of their modern range, also suggest that temperatures during second to fourth century were 1 or 2 °C higher than today (Kenward, 2004). This corresponds with the Roman climate optimum. However, Holocene distribution shifts of semiaquatic Coleoptera should be interpreted with caution in terms of climate changes, since it is suggested by Andrieu-Ponel and Ponel (1999) that in the French region of Provence (and certainly elsewhere in Europe), the local extinction of fen species such as *Odacantha melanura* may have been a consequence of fen drainage and other human activities.

During the ninth to eleventh centuries, there is also a wealth of data from Heteroptera and Coleoptera supporting the hypothesis of summer temperatures higher than today by 1 or 2 °C (in agreement with the climate optimum around AD 1000): in northern England, examples of this include the presence of such species as *Eurydema oleracea*, *Coreus marginatus*, *Ischnodemus sabuleti* (Heteroptera), *Odacantha melanura*, *Cryptolestes duplicatus*, *Phymatodes testaceus*, *Pediacus dermestoides*, *Anthicus antherinus*, *Anthicus bifasciatus*, *Acritus homoeopathicus*, *Apion difficile*, several *Platysthetus* species, and *Xestobium rufovillosum* (Coleoptera).

There are few beetle records dated to the later Medieval and post Medieval period (Kenward, 2004), and therefore the effect of the Little Ice Age (roughly about AD 1450–1880) on insect ranges is still poorly documented. A recent review of this topic by Buckland and Wagner (2001) emphasizes the difficulty of extracting a climatic signal from beetle assemblages deposited in an environment increasingly perturbed by human activities and fragmentation of habitat, but it seems likely that some species were indeed affected by Little Ice Age cooling. These beetles became extinct in the northern parts of their range or even eradicated from Britain. This may be what happened to the beetles, *Airaphilus elongatus* (Osborne, 1997), *Gyrinus colymbus* (Girling, 1984), *Aphodius quadrimaculatus* (Osborne, 1969), *Xestobium rufovillosum*, *Odacantha melanura*, *Hydrophilus piceus* (Buckland and Dinnin, 1993) and the bug *Heterogaster urticae* (Kenward, 2004).

The present northward expansion of many insects may be interpreted as a consequence of a probable global warming that has taken place since the Little Ice Age. This has been particularly well documented for the Lepidoptera, but there is also evidence from many other insect orders, especially Heteroptera (Kenward, 2004). The discovery of several beetle species in southern France, well beyond their northern limit, may also be attributed to an increase in temperature. This is probably the case of the darkling beetles (Tenebrionidae) *Gonocephalum yelamosi* and *Pseudoseriscius pruinosis*, and the ground beetles *Paracelia simplex* and *Syntomus fuscomaculatus* (Ponel and Jeanne, 1997).

Impact of Forest Clearance on Saproxylophagous and Grassland Coleoptera

Again the overwhelming majority of data have been obtained from British sites, and a recent review of this issue was provided

by Whitehouse (2006). The impact of forest clearance on 'Urwald' insects (associated with old primeval forest) from the mid-Holocene onwards was perceived early in the history of Quaternary entomology, but more recent detailed studies have provided a very accurate picture of the profound changes that affected the beetle fauna in response to the progressive destruction by man of the natural oak and linden forest that dominated the British landscape from the early Holocene to about 5000 BP. Of particular interest is the site of Thorne Moors, thoroughly analyzed by Buckland (1979). This Bronze Age site, dated to about 3000 years BP, yielded a very rich Urwald beetle assemblage. This fauna included many saproxylophagous species now extinct in Britain, such as *Rhysodes sulcatus*, *Isorhipis melasoides*, *Prostomis mandibularis*, *Mycetina cruciata*, *Peltis grossa* (Figure 2), or with an extremely limited distribution in Britain, such as *Dryophthorus corticalis* or *Eucnemis capucina*. According to Whitehouse (2006), at least 40 beetle taxa found in Holocene deposits in Britain and 15 in Ireland are now apparently extinct in the British Isles, and this list is destined to expand in the future with the analysis of more sites (Buckland and Dinnin, 1993).

A similar pattern emerges from the study of continental sites, especially in France, since several records of Urwald beetles were obtained from seminatural archaeological contexts dated to about 5000 BP in the lacustrine village of Chalain (Ponel, unpublished data). Examples include *Peltis grossa*, *Lichenophanes varius*, *Bothrideres contractus*, and *Dryophthorus corticalis*. Unlike Britain, all these species are still present in France today; however, some of them (e.g., *Peltis grossa*) are now relegated to old-growth mountain forests, especially in the Pyrénées. This is probably because the steep slopes of their last strongholds have not been disturbed by human action.



Figure 2 *Peltis grossa* (Coleoptera: Trogositidae), a saproxylic beetle now extinct in Britain and associated with primeval forests.

Their inability to recolonize suitable habitats from such refuges may be a consequence of a low dispersal potential, but it is also possible that suitable old-growth forests are not available elsewhere in France.

Interestingly, the removal of the original forest cover did not affect only saproxylophagous beetles, but also other species not directly associated with woodland. This is probably the case for the water beetle *Agabus wasastjernae* found today in Scandinavian native pine forest where it lives in acid pools (Buckland, 1979) and almost eradicated from Britain along with the coniferous cover.

Wholesale forest clearance during the Neolithic period not only resulted in the retraction and eradication of saproxylophagous beetles but, as might be expected, it led also to the parallel expansion of open environment and grassland beetles. This faunal change is well documented in many British sites, particularly from the Bronze Age, and beetle assemblages from that time are progressively dominated by species associated with dung, such as the representatives of the large coprophagous genera *Onthophagus* and *Aphodius*, and the coprophilous rove beetles *Oxytelus* and *Platysthetus*. Many grass root feeders such as *Adelocera murina* and *Phyllopertha horticola* are also found in significant numbers. Such assemblages dominated by dung and grassland beetles suggest that grazing animals were abundant, in a predominantly open environment. Most of the studied sites show that beetle assemblages are almost totally devoid of tree-dependent species at that time and suggest an increasing representation of open environments in the British landscape, but it is probable that remnants of forests and extensive grasslands coexisted until the late Bronze Age (Osborne, 1978). During the Roman period the landscape in the British Isles was probably very similar to the present. In Southern France, the analysis of beetle assemblages sampled near the Rhône valley also suggests that the landscape was already completely open 2000 years ago and has remained so until now (Andrieu-Ponel et al., 2000).

More recently, Whitehouse and Smith (2010) have reanalyzed a range of British sites and beetle assemblages dating from the first half of the Holocene and a slightly different picture emerges from this synthesis: the proportion of beetle species indicative of open and forested environment is highly variable at the study sites, suggesting that open areas persisted and were locally important during the Mesolithic. The landscape at that time was probably a mosaic of woodlands and open clearings. Beetle assemblages from the later Mesolithic suggest a closing of the forest, although open environments were still present. The early Neolithic period is marked by an increased complexity of the landscape: dung beetles suggest a progressive opening whereas saproxylophagous beetles persist in a context of landscape heterogeneity and high regional variability. According to Whitehouse and Smith (2010), the low representation of the dung beetle community do not clearly suggest that open areas were a consequence of grazing activities of large herbivore herds, prior to the Neolithic period. In this context, it is worth noting that Smith et al. (2010) have shown from modern analog studies that the representation in fossil assemblages of tree-dependent beetles and dung beetles could be an indicator of openness of the landscape and presence of grazing mammals, but further analyses are required to precisely characterize this trend.

Other Consequences of Human Action

As indicated above, there is increasing evidence concerning the role of man, rather than that of climate, on the retraction of the range of hygrophilous or semiaquatic beetle species. This phenomenon has been described from the ancient marsh of Tourves (southern France) in a Holocene sequence spanning the Neolithic transition. At about 4310 BP a marked change is recorded by the beetles as a rich marshy fauna is suddenly replaced by an impoverished fauna in which running-water elements suggest a change in the hydrological regime; *Odocantha melanura* and *Cerapheles lateplagiatus* disappear from the assemblages. Both species are now restricted to Camargue, the last extensive marshy area in Provence, about 100 km westward (Andrieu-Ponel and Ponel, 1999). It is suggested that hydrologic changes were induced by drainage, associated with the intensification of cultivation practices in a region densely populated by Neolithic people.

The disappearance of riffle beetles (Elmidae) from rivers in Britain also seems to be attributed to forest clearance. These running-water beetles, adapted to highly oxygenated rivers, are found only on the gravelly bottom of streams, where they hang on the lower part of shingle by means of their oversized claws. It is likely that a secondary impact of the removal of the forest cover was an increase of soil erosion leading to a rise of alluviation processes in lowland areas, muddying the waters and thus making them unsuitable for Elmidae (Smith, 2000).

A further consequence of the development of agriculture is the range extension of opportunist beetle species now abundant in cultivated areas in Britain but unknown before Roman times. This is the case not only of the ground beetles *Pterostichus madidus* and *Zabrus tenebrioides*, but also of thermophilous species now very rarely found in the wild but usually associated with human habitations and livestock shelters, such as *Aglenus brunneus*, *Tenebroides mauritanicus*, *Oryzaephilus surinamensis*, and many others. The case of the dissemination of several species such as the grain weevil *Sitophilus granarius* is especially interesting since they are strictly associated with stored food products and are unable to live outside human habitations (Figure 3).



Figure 3 A beetle assemblage containing several synanthropic species from a Medieval archaeological site near the Vieux-Port of Marseille (France).

Their geographic origin remains obscure and it is likely that some of them originate from tropical regions, but this issue is still a matter of debate (Buckland, 1981).

Holocene History of the Beetle Fauna of Mediterranean Mountains

During the last 15 years, particular attention has been paid to the Holocene history of insect faunas in southern France and especially in the Mediterranean mountains (Southern Alps, Eastern Pyrénées). Coring of long continuous Holocene sequences was performed with either multiple coring techniques, or with large-volume corers (Figure 4), allowing sufficient quantities of sediment to be obtained. With such sampling procedures it is possible to obtain enough material to conduct multidisciplinary studies including, for example, Coleoptera, chironomids, pollen, and plant macrofossil analyses, leading to increased precision in the reconstruction of Human activities (such as pastoralism, agriculture, and mining) in mountains and in the reconstruction of past fluctuations of altitudinal treeline. A recent study by Ponel et al. (2011) at Lac des Lauzons in the Southern Alps illustrates this combined approach and provides an original insight into the forest history in this region, from the retreat of the ice sheet to the present.

In the Eastern Pyrénées, the key site of La Restanque, located at 1620 m above sea level, provides an excellent example of climatic and paleoenvironmental reconstructions based on a succession of Late Glacial–Holocene insect assemblages. A 5-cm resolution analysis of this 4 m profile yielded one of the most detailed (ongoing) analyses of the last 13 000 years. This record (Ponel et al., 2004) provides original data concerning climatic and environmental changes during the Late Glacial–Holocene transition. This history documents changes in insect faunas, in conjunction with vegetational changes, and human impact on the environment (Figure 5).

The lower part of the La Restanque insect sequence is characterized by the abundance of cold-adapted taxa, particularly of *Helophorus glacialis*, a beetle restricted to the margin of melting snow patches. This species lives at sea level in northern Scandinavia but well above 2000 m in southern Europe. It is



Figure 4 Large-diameter corer (ø 20 cm) used for sampling at La Restanque, French Pyrénées.

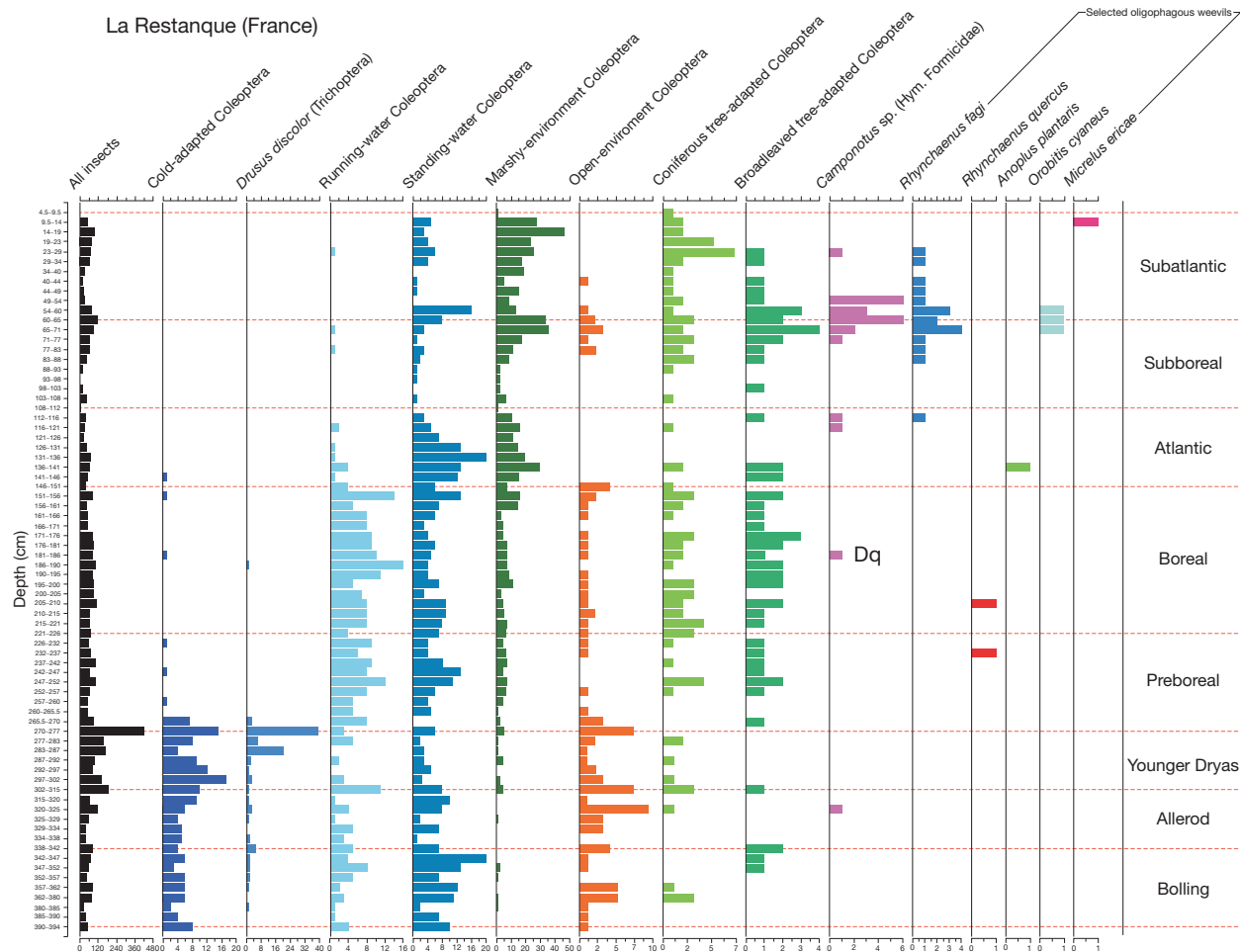


Figure 5 Succession of insect assemblages at La Restanque, French Pyrénées (1620 m), compared with pollen chronozones. The diagram shows the minimum number of individuals. Dq, isolated occurrence of the thermophilous ant *Dolichoderus quadripunctatus*. Preferred host plants of a selection of monophagous or oligophagous weevils: *Rhynchaenus fagi* associated with beech (*Fagus sylvatica*), *Rhynchaenus quercus* associated with oak (*Quercus*), *Anoplus plantaris* associated with birch (*Betula*), *Orobitis cyaneus* associated with violet (*Viola*), *Micrelus ericae* associated with heather (*Calluna vulgaris* and *Erica tetralix*).

a good climate indicator since it requires mean July temperatures below 10 °C to complete its life cycle. The dominance of this cold-adapted insect fauna clearly indicates that no forest cover was present at this time. This part of the sequence may be attributed to the Late Glacial interstadial, since it is likely that at such high altitudes the Late Glacial warming did not reach a sufficient level to allow warm-adapted insects to survive there, but allowed cold-adapted taxa to settle shortly after the Würmian deglaciation. Comparison with pollen analyses carried out on the same sequence confirms this reconstruction.

The perceptible rise in cold-adapted beetles in the next faunal unit suggests that this unit corresponds to the Younger Dryas. Immediately after this cold interval, at the onset of the Holocene, the disappearance of cold-adapted beetle species was dramatic, signaling the beginning of Holocene warming. This event is synchronous with a sharp peak of *Drusus discolor*, a caddisfly (Trichoptera) that lives today in cold and rapid mountain streams. This caddisfly is especially interesting since it suggests a short period of high-energy streams, certainly linked with large-scale snow melting induced by climatic

warming. In its turn this taxon disappears quickly, suggesting rapid environmental change, the disappearance of most of the snow cover and a concurrent decrease in stream velocities.

This initial warming phase was followed by a period characterized by regular occurrences of arboreal beetles. Beetles associated both with broadleaved and coniferous trees were present, suggesting that forest cover was established in the vicinity of the site. This period corresponds with the Boreal chronozone (9000–8000 BP). Surprisingly, few forest-dependent terrestrial beetles (associated with shaded environments) were identified, suggesting that the forest cover was not dense. At this period, temperatures much higher than today are indicated by several taxa, including the arboreal ant *Dolichoderus quadripunctatus*. This is a predominantly lowland species that is rather thermophilous and unable to live today at such high altitudes.

The next unit, between 8000 and 4700 BP, is marked by the disappearance of running-water beetles and by the rise of standing water and hygrophilous marshy species. This period corresponds with the Atlantic chronozone. It is possible that

these events could be related to changes in local hydrology, such as a shift from a fluvial regime to a lacustrine environment, associated with the infilling of the depression.

During the Subboreal (4700–2700 BP) and the Subatlantic (2700 BP onward), the human impact on the fauna and the forest cover is suggested by a complex sequence of events, including a decrease in conifer-dependent beetles, a reappearance of open environment beetles, and a peak of the carpenter ant, *Camponotus*. Most of the representatives of this genus are dependent on dead wood, since they dig galleries in old stumps and trunks to establish their nest. This peak of *Camponotus* is likely attributed to the increased availability of dead wood after forest clearance (fires, tree felling) and not to climatic influence, since no cold-adapted taxa are recorded at that time. Simultaneously, the beech forest became established nearby, as attested by the occurrences of the weevil *Rhynchaenus fagi* that feeds exclusively on beech. Another weevil, *Orobitis cyanea* that feeds exclusively on *Viola* in shaded underwoods, suggests that a true forest environment prevailed at this period. At that time it appears that the Restanque site was situated in an ecotone, at the boundary between a decimated conifer forest and an expanding beech forest. The uppermost samples of the sequence, corresponding to the end of the Subatlantic interval, are marked by an extremely impoverished insect fauna, partly for taphonomic reasons, but certainly also because heathlands expanded, at the expense of forests. This is suggested by the isolated occurrence of the beetle *Micrelus ericae*, a weevil dependent on heaths (Ericaceae).

Despite some possible events of local significance, it is likely that the broad pattern of faunal successions described at La Restanque is valid at a regional scale, and accurately reflects the progression of beetle assemblages and environments at high altitude regions in the eastern Pyrénées.

Conclusion

Beetle are recognized as excellent biological markers of ancient environments and they display a great sensitivity to climate change. The analysis of beetle fauna successions during the Postglacial period in Europe provides ample evidence for climate and environmental changes at the onset of the Holocene, and independent proxy data concerning the disruption of landscapes and vegetation under human activity (forest clearance, habitat fragmentation, agriculture, pastoralism) from the Neolithic period onwards. However, a detailed reconstruction of subtle Holocene climate evolution from beetle data is still not available, precisely because human action and climate-induced faunal changes during this period are extremely difficult to disentangle. A special effort devoted to this problem should be made in the future.

See also: **Beetle Records: Late Pleistocene of Europe; Postglacial North America.**

References

- Andrieu-Ponel V and Ponel P (1999) Human impact on mediterranean wetland coleoptera: An historical perspective at Tourves (Var, France). *Biodiversity and Conservation* 8: 391–407.
- Andrieu-Ponel V, Ponel P, Bruneton H, and Leveau P (2000) Palaeoenvironments and cultural landscape of the last 2000 years reconstructed from pollen and coleopteran records in the Lower Rhône Valley, southern France. *The Holocene* 10: 341–355.
- Atkinson TC, Briffa KR, and Coope GR (1987) Seasonal temperatures in Britain during the past 22,000 years, reconstructed using beetles remains. *Nature* 325: 587–592.
- Buckland PC (1979) *Thorne Moors: A palaeoecological study of a Bronze Age site (A Contribution to the History of the British Insect Fauna)*. Birmingham: Department of Geography, University of Birmingham. Occasional Publication 8.
- Buckland PC (1981) The early dispersal of insect pests of stored products as indicated by archaeological records. *Journal of Stored Product Research* 17: 1–12.
- Buckland PC and Dinnin MJ (1993) Holocene woodlands: The fossil insect evidence. In: Kirby K and Drake CM (eds.) *Dead Wood Matters: The Ecology and Conservation of Saprophytic Invertebrates in Britain*. *English Nature Science* 7: 6–20.
- Buckland PC and Wagner PE (2001) Is there an insect signal from the 'Little Ice Age'? *Climatic Change* 48: 137–149.
- Coope GR (1986) Coleoptera analysis. In: Berglund BE (ed.) *Handbook of Holocene Palaeoecology and Palaeohydrology*, pp. 703–713. Chichester: Wiley.
- Coope GR (1994) The response of insect faunas to glacial–interglacial climatic fluctuations. *Philosophical Transactions of the Royal Society of London B* 344: 19–26.
- Coope GR (1998) Insects. In: Preece RC and Bridgland DR (eds.) *Late Quaternary Environment Change in North-west Europe: Excavations at Holwell Coombe, South-east England*, pp. 213–233. London: Chapman & Hall.
- Coope GR and Brophy JA (1972) Late Glacial environmental changes indicated by a coleopteran succession from North Wales. *Boreas* 1: 97–142.
- Elias S, Webster L, and Amer M (2009) A beetle's eye view of London from the Mesolithic to Late Bronze Age. *Geological Journal* 44: 537–567.
- Girling MA (1984) A Little Ice Age extinction of a water beetle from Britain. *Boreas* 13: 1–4.
- Kelly M and Osborne PJ (1965) Two faunas and floras from the alluvium at Shustoke, Warwickshire. *Proceedings of the Linnean Society of London* 176: 37–65.
- Kenward H (2004) Do insect remains from historic-period archaeological occupation sites track climate change in Northern England? *Environmental Archaeology* 9: 47–59.
- Lemdahl G (1997) Late Weichselian and Early Holocene colonisation of beetle faunas in S Sweden. *Quaternary Proceedings* 5: 153–164.
- Lemdahl G (2000) Lateglacial and Early Holocene insect assemblages from sites at different altitudes in the Swiss Alps – Implications on climate and environment. *Palaeogeography, Palaeoclimatology, Palaeoecology* 159: 293–312.
- Mayewski PA, Rohling EE, Stager JC, et al. (2004) Holocene climate variability. *Quaternary Research* 62: 243–255.
- Osborne PJ (1969) An insect fauna of Late Bronze Age date from Wilsford, Wiltshire. *Journal of Animal Ecology* 38: 555–566.
- Osborne PJ (1974) An insect assemblage of Early Flandrian age from Lea Marston, Warwickshire and its bearing on the contemporary climate and ecology. *Quaternary Research* 4: 471–486.
- Osborne PJ (1976) Evidence from the insects of climatic variations during the Flandrian period: A preliminary note. *World Archaeology* 8: 150–158.
- Osborne PJ (1978) Insect evidence for the effect of man on the lowland landscape. In: Limbrey S and Evans JG (eds.) *The Effects of Man on the Landscape: The Lowland Zone*, pp. 32–34. London: Council for British Archaeology. Research Report 21.
- Osborne PJ (1979) Insect remains. In: Smith C (ed.) *Fisherwick: The Reconstruction of an Iron Age Landscape*, *British Archaeological Reports* 61: 85–87 and 189–193.
- Osborne PJ (1982) Some British later prehistoric insect faunas and their climatic implications. In: Harding A (ed.) *Climatic Change in Later Prehistory*, pp. 68–74. Edinburgh: Edinburgh University Press.
- Osborne PJ (1988) A Late Bronze Age insect fauna from the River Avon, Warwickshire, England: Its implications for the terrestrial and fluvial environment and for climate. *Journal of Archaeological Science* 15: 715–727.
- Osborne PJ (1997) Insects, man and climate in the British Holocene. *Quaternary Proceedings* 5: 193–198.
- Ponel P (1997) Late Pleistocene coleopteran fossil assemblages in high altitude sites: A case study from Prato Spilla (Northern Italy). *Quaternary Proceedings* 5: 207–218.
- Ponel P, Andrieu-Ponel V, Reille M, and Guiter F (2004) A cross-comparison of insect and pollen data of the last 13 ka in the French Pyrénées. XI International Palynological Congress, Granada (Spain), 4–9 July 2004. *Polen* 14: 234.
- Ponel P and Coope GR (1990) Lateglacial and Early Flandrian Coleoptera from La Taphanel, Massif Central, France: Climatic and ecological implications. *Journal of Quaternary Science* 5: 235–249.
- Ponel P, Court-Picon M, Badura M, et al. (2011) Holocene history of Lac des Lauzons (2180 m a.s.l.), reconstructed from multiproxy analyses of Coleoptera, plant macroremains and pollen (Hautes-Alpes, France). *The Holocene* 21(4): 571–588.

- Ponel P and Jeanne C (1997) Un carabique nouveau pour la faune française: *Syntomus fuscumaculatus* (Motschoulsky, 1844) dans les Bouches-du-Rhône [Coleoptera Lebiidae]. *Nouvelle Revue d'Entomologie* 14: 353–357.
- Ponel P, Matterné V, Coulthard N, and Yvinec J-H (2000) La Tène and Gallo-Roman natural environments and human impact at the Touffréville Rural Settlement, Reconstructed from Coleoptera and Plant Macroremains (Calvados, France). *Journal of Archaeological Science* 27: 1055–1072.
- Smith DN (2000) Disappearance of elmids 'riffles beetles' from lowland river systems – The impact of alluviation. In: Nicholson RA and O'Connor TP (eds.) *People as an Agent of Environmental Change*, pp. 75–80. Oxford: Oxbow Books.
- Smith D, Whitehouse N, Bunting MJ, and Chapman H (2010) Can we characterise 'openness' in the Holocene palaeoenvironmental record? Modern analogue studies of insect faunas and pollen spectra from Dunham Massey deer park and Epping Forest, England. *The Holocene* 20: 215–229.
- Whitehouse N (2006) The Holocene British and Irish ancient forest fossil beetle fauna: Implications for forest history, biodiversity and faunal colonisation. *Quaternary Science Reviews* 25: 1755–1789.
- Whitehouse N and Smith D (2010) How fragmented was the British Holocene wildwood? Perspectives on the 'Vera' grazing debate from the fossil beetle record. *Quaternary Science Reviews* 29: 539–553.

Postglacial North America

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Introduction

The exoskeletal remains of beetles (Coleoptera) preserve well in waterlogged sediments of Pleistocene and Holocene age. Although the transition from the last glaciation to postglacial environments was a time-transgressive phenomenon, in most regions of North America this climatic transition took place between 14 000 and 11 550 years ago (12 000–10 000 ^{14}C yr BP). Most of the North American fossil beetle assemblages greater than 11 500 years in age are discussed in the article entitled 'late Pleistocene Beetle Records from North America.' In this article, I discuss the Late Glacial fossil beetle record of just the Maritime region of Canada, as well as Holocene records from across the continent.

In recent decades, studies of postglacial fossil beetle assemblages have been carried out in many regions of North America. [Table 1](#) lists 24 postglacial faunal sites from eastern and central North America, and [Table 2](#) lists 20 postglacial sites from western North America. This article reports on the major findings of these studies, in terms of paleoenvironmental reconstructions and shifting distribution patterns (biogeography) of the species found in the fossil assemblages, all of which remain extant today. Some of the fossil assemblages discussed here formed part of purely paleontological research projects. Others formed part of archaeological research, as noted in the text.

The paleotemperature estimates discussed are derived from the Mutual Climatic Range (MCR) method of paleoclimate analysis, as discussed in the Overview for this section. For sites in eastern and central North America, the MCR method was used to provide estimates of mean temperatures of the warmest (TMAX) and coldest (TMIN) months of the year. TMAX and TMIN estimates are discussed in terms of departures from the modern TMAX and TMIN of the study sites in question.

Research in Eastern North America

Fossil beetle assemblages of postglacial age have been studied in eastern Canada, New England, along the Mississippi drainage in the mid-western United States, and in the Canadian Territory of Nunavut ([Figure 1](#)). Many of these assemblages represent early postglacial environments ([Table 1](#), sites 1–14). Only a few mid-Holocene assemblages have been described from these regions ([Table 1](#), sites 15–18). Late Holocene assemblages, including three from archaeological sites, span the interval 3.5 to 0.1 ka ([Table 1](#), sites 19–24).

Late Pleistocene–Holocene Transition

The fossil beetle record from many parts of North America indicates a relatively smooth transition from glacial to

interglacial climates, beginning about 17 ka (calibrated years ago $\times$ 1000) ([Elias et al., 1996](#)). Throughout most of North America, paleotemperature reconstructions based on beetle assemblage data do not show a climatic reversal during the Late Glacial interval (17–11 ka). The one prominent exception to this comes from the Maritime region of Canada. [Miller and Elias \(2000\)](#) reconstructed mean July temperatures (TMAX) and mean January temperatures (TMIN) from radiocarbon-dated fossil beetle assemblages from a suite of late-glacial sites in this region. These assemblages date from 14.9 ka to 12.5 ka. In the Maritime region, multiple lines of evidence, including stratigraphy, palynology, chironomid fossils, and beetles, suggest a climatic deterioration during the Younger Dryas chronozone. MCR estimates suggest several 'events' in the regional fossil beetle record that appear similar to climate events recorded in the GRIP ice core record, including the Younger Dryas cooling event from GI-1a to GS-1, beginning ca. 12 650 GRIP yr BP. The results of MCR-based temperature reconstruction suggests a drop in TMAX of about 5 °C between 13 ka (at Brookside, Nova Scotia) and 12.5 ka (at West Mabou, Nova Scotia). These results compare well with the temperature reconstruction of summer surface waters based on chironomids ([Levesque et al., 1997](#); [Lotter et al., 1999](#)). The fact that there is little fossil beetle evidence for a Younger Dryas cooling event elsewhere in North America lends support for the paleoceanographic model ([Duplessy et al., 1996](#)) that links this cooling event with changes in circulation patterns in the North Atlantic. The only other North American region from which beetle records indicate a Younger Dryas cooling is arctic Alaska ([Elias, 2000](#)). However, few continuous series of beetle assemblages have been studied from this time interval in most regions of North America.

Most early postglacial beetle faunas from eastern and central North America are indicative of climatic regimes close to modern parameters ([Table 1](#)). The exceptions to this include sites that were located in close proximity to large bodies of glacial meltwater, such as Lockport Gulf, New York. Here, at 11.45 ka, TMAX was 1–4.5 °C colder than today ([Elias et al., 1996](#)). Climatic amelioration through the early Holocene is indicated by the mid-western beetle assemblages. [Schwert and Ashworth \(1988\)](#) discussed the discrepancy between early postglacial environments deduced from insect assemblages and paleobotanical reconstructions in the mid-western United States as follows. Pollen and plant macrofossils that accumulated in ice-marginal deposits just after deglaciation reflect tundra-like communities with sedges and *Dryas*, and are lacking trees and shrubs. However, the insect faunas are consistently different, being analogous to modern faunas from the middle of the boreal forest zone. At each study site, the discrepancy between the plant and insect evidence has been accounted for by invoking a substantial lag in the arrival and establishment of woody plants on deglaciated landscapes.

Table 1 Summary of fossil beetle assemblage data from sites in central and eastern North America

Site	Age (cal yr BP $\times$ 1000)	Holocene		Modern		Change in Temperature		References ¹
		TMAX (°C)	TMIN (°C)	TMAX (°C)	TMIN (°C)	July ΔT	January ΔT	
1) Benacadie, Nova Scotia	13.73	9–13	–31 to –14	18.1	–5.5	–9.1 to –5.1	–25.5 to –8.5	Miller and Elias, 2000
2) Collins Pond, Nova Scotia/Lower	13.73	10–12.5	–20.5 to –11	16.3	–3.9	–6.3 to –3.8	–16.6 to –7.1	Miller and Elias, 2000
3) Todd Mountain, New Brunswick	13.42	10–13	–30.5 to –18	19.1	–9.1	–9.1 to –6.1	–21.4 to –8.9	Miller and Elias, 2000
4) Campbell, Nova Scotia	13.16	11.5–14.5	–27.5 to –8	18.6	–8	–7.1 to –4.1	–19.5 to 0	Miller and Elias, 2000
2) Collins Pond, Nova Scotia/Upper	12.82	11–14	–21.5 to –7.5	16.3	–3.9	–5.3 to –2.3	–17.4 to –3.6	Miller and Elias, 2000
5) Amaguadees, Nova Scotia	12.73	14.5–15.5	–28 to –23.5	18.1	–5.5	–3.6 to –2.6	–22.5 to –18	Miller and Elias, 2000
6) West Mabou, Nova Scotia	12.53	11–14	–18 to –5	18.6	–8	–7.6 to –4.6	–10 to +3	Miller and Elias, 2000
7) Rainy River, Ontario/Site E-13	12.24–11.75	13–15	–27 to –22	18.8	–16.2	–5.8 to –3.8	–10.8 to –5.8	Bajc et al. 2000
8) Wales, Ontario	12.1	N/A	N/A	20.5	–8.1	N/A	N/A	Miller et al. 1987
9) Rainy River, Ontario/Site RR54	11.68	14.5–15.5	–27 to –23.5	18.8	–16.2	–4.3 to –3.3	–10.8 to –7.3	Bajc et al. 2000
10) Mosbeck, Minnesota/Unit E	11.50	15–20	–21 to –5	20.2	–16.9	–5.2 to –0.2	–4.1 to +11.9	Ashworth et al. 1972
11) Lockport Gulf, New York/L03	11.45	17–20.5	–15.5 to –3.5	21.5	–4.8	–4.5 to –1	–10.7 to +0.3	Miller and Morgan, 1982
12) Seibold, North Dakota	11.08	17–21.5	–16 to –5	20.5	–14.9	–3.5 to +1	–0.6 to +9.9	Ashworth and Brophy, 1972
10) Mosbeck, Minnesota Unit D	10.91	16.5–20	–18 to –6.5	20.2	–16.9	–3.7 to –0.2	–1.1 to +10.4	Ashworth et al. 1972
11) Lockport Gulk, New York/L04	10.87	15–22	–22.5 to –1	21.5	–4.8	–6.5 to +0.5	–17.7 to +4.3	Miller and Morgan, 1982
11) Lockport Gulk, NY/L05	10.36	18.5–22	–19.5 to –10.5	21.5	–4.8	–3 to +0.5	–14.2 to –5.2	Miller and Morgan, 1982
10) Mosbeck, Minnesota Unit C	9.44	15–21	–24 to –1.5	20.2	–16.9	–5.2 to +0.8	–7.1 to +15.4	Ashworth et al. 1972
13) Saint Flavien, Quebec Unit 1	7.83	16.75–20	–12.25 to –7	18.7	–13	–1.95 to +1.3	+0.75 to +6	Lavoie et al. 1997a
14) Ennadai Lake, Nunavut Unit D	7.2–6.86	10–14.5	–28.5 to –23	13	–32	–3 to +1.5	+3.5 to +9	Elias, 1982a
14) Ennadai Lake, Nunavut Unit C	6.86–2.92	12–15	–25 to –19	13	–32	–1 to +2	+7 to +13	Elias, 1982
15) Boniface River, Quebec	6.66–1.91	11.8–14.5	N/A	9.4	–24.8	+2.4 to +5.1	N/A	Lavoie et al. 1997
16) Radisson, Quebec	5.84–0	14.5–17.5	N/A	13.4	–3.4	+1.1 to +4.1	N/A	Lavoie, 2001
17) Charlevoix, Quebec	5.25–0	N/A	N/A	19.2	–12.8	N/A	N/A	Lavoie, 2001
18) Au Sable River, Michigan	4.66–4.55	17–18	–14 to –12	20	–5	–3 to –2	–9 to –7	Morgan et al. 1985
13) St Flavien, Quebec Unit 2	4.49–3.55	15–20	–14 to –5	18.7	–13	–3.7 to +1.3	–1 to +9	Lavoie et al. 1997
19) Bongards, Minnesota	3.55–0	N/A	N/A	23.1	–11.2	N/A	N/A	Schwert and Ashworth, 1985
20) Roberts Creek, Iowa	3.19–0	N/A	N/A	22.4	–8.9	N/A	N/A	Schwert, 1996
13) St Flavien, Quebec/Unit 4	3.19–0	16–20	–23.5 to –5	18.7	–13	–2.7 to +1.3	–10.5 to +8	Lavoie et al. 1997
21) Umiakoviarsuk, Labrador	2.21–0.89	12–13.5	–23 to –12	10.7	–18.5	+1.3 to +2.8	–5.5 to +6.5	Elias, 1982
22) Cahokia, Illinois	AD 1050	N/A	N/A	26.6	–1.5	N/A	N/A	Pauketat et al. 2002
23) Boston, Massachusetts	17th century	N/A	N/A	23.1	–1.9	N/A	N/A	Bain, 1998
24) Quebec City, Quebec	19th century	N/A	N/A	19.2	–12.8	N/A	N/A	Bain, 2001

¹References available from the 'QBib' web site: <http://www.bugs2000.org/qbib.html>

Table 2 Summary of fossil beetle assemblage data from sites in western North America

Site	Age (cal yr BP $\times 1000$)	Holocene		Modern		Change in Temperature		References ¹
		TMAX (°C)	TMIN (°C)	TMAX (°C)	TMIN (°C)	July ΔT	January ΔT	
1) False Cougar Cave, Montana	11.59	15.5–17.75	–4.5 to +0.25	13.5	–10.1	+2 to +4.25	+5.6 to +10.4	Elias, 1996
2) Huntington Canyon, Utah	11.46	15–17	–6 to 0	16	–9.1	–1 to +1	+3.1 to +9.1	Elias, 1996
3) La Poudre Pass, Colorado	11.39	15–18	–17.5 to –7	11.3	–7.7	+3.7 to +6.7	–9.8 to +0.7	Elias, 1996
4) Lake Isabelle Delta, Colorado	10.12	11.75–14.5	–31.25 to –15	10.8	–8.2	+1 to +3.7	–23 to –6.8	Elias, 1996
3) La Poudre Pass, Colorado	9.89	13.5–16.5	–19.5 to –9	11.3	–7.7	+2.2 to +5.2	–13.6 to –1.3	Elias, 1996
4) Lake Isabelle Delta, Colorado	9.48	10.5–13	–23.5 to –16	10.8	–8.2	–0.3 to +2.2	–15.3 to –7.8	Elias, 1996
4) Lake Isabelle Delta, Colorado	8.70	11–13	–14 to –9	10.8	–8.2	+0.2 to +2.2	–5.8 to –0.8	Elias, 1996
4) Lake Isabelle Fen, Colorado	7.90	10.25–13	–14.75 to –7.5	10.8	–8.2	–0.6 to +2.2	–6.6 to +0.7	Elias, 1996
3) La Poudre Pass, Colorado	6.14	12.5–13.5	–12.5 to –11.5	10.8	–8.2	+1.7 to +2.7	–4.3 to –3.3	Elias, 1996
3) La Poudre Pass, Colorado	3.79	11.75–15	–21.25 to –14.5	11.3	–7.7	+0.4 to +3.7	–11.8 to –6.8	Elias, 1996
4) Lake Isabelle Fen, Colorado	3.17	10.25–13	–14.5 to –7.5	10.8	–8.2	–0.6 to +2.2	–6.3 to +0.7	Elias, 1996
5) Longs Peak Inn, Colorado	3.14	12–15.5	–26.5 to –14	14.5	–5	–2.5 to +1	–21.5 to –9	Elias, 1996
5) Longs Peak Inn, Colorado	2.82	13.5–15.5	–24 to –15	14.5	–5	–1 to +1	–19 to –10	Elias, 1996
6) Roaring River, Colorado	2.49	14.25–14.75	–18.25 to –16.5	14.3	–5.5	0 to +0.5	–12.75 to –11	Elias, 1996
7) Rat's Nest Cave, Alberta	1.97–0	N/A	N/A	16.2	–8.9	N/A	N/A	Bain et al. 1997
8) Mount Ida Bog, Colorado	0.85	10.25–12	–14.75 to –9	9.3	–11.5	+1 to +2.7	–3.25 to +2.5	Elias, 1996
5) Longs Peak Inn, Colorado	0.43	13.5–15.5	–24.5 to –15	14.5	–5	–1 to +1	–19.5 to –10	Elias, 1996
Chihuahuan Desert Packrat Midden Faunas								
9) Sacramento Mountains, NM	11.91–0.44	Not available						Elias and Van Devender, 1992
10) Tunnel View, BBNP, Texas	11.39–10.05	Not available						Elias and Van Devender, 1992
11) Hueco Mountains, Texas	11.18–2.60	Not available						Elias and Van Devender, 1992
12) Sierra de la Misericordia, Coahuila	10.57–0.720	Not available						Elias et al. 1995
13) Puerto de Ventanillas, Coahuila	9.75–.59	Not available						Elias et al. 1995
14) Maravillas Canyon Cave, Texas	9.22–.09	Not available						Elias and Van Devender, 1990
15) Cañon de la Fragua, Coahuila	6.52–.51	Not available						Elias et al. 1995
Great Basin Packrat Middens								
16) Glen Canyon region, Utah	9.69–1.42	Not available						Elias et al. 1992
17) Canyonlands National Park, Utah	7.82–4.24	Not available						Elias et al. 1992
Sonoran Desert Packrat Mideens								
18) Puerto Blanco Mountains, Arizona	12.40–0	Not available						Hall et al. 1990
19) Ajo Mountains, Arizona	11.47–1.09	Not available						Hall et al. 1989
20) Cataviña, Baja California	1.1	Not available						Clark and Sankey, 1999

¹References available from the 'QBib' web site: <http://www.bugs2000.org/qbib.html>

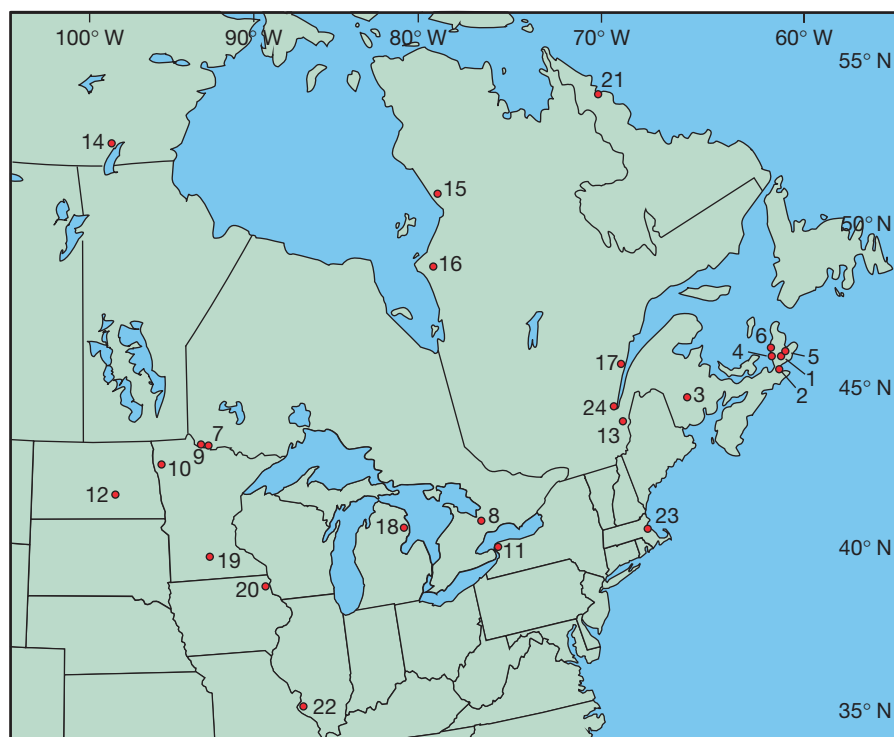


Figure 1 Map of eastern and central North America, showing location of fossil sites discussed in text. Site numbers match those in [Table 1](#).

As in central Europe, this lag may represent the time needed for ecological succession from open ground to herbaceous cover, shrub cover, and finally forest cover. The most recent paleobotanical reconstructions for the eastern and central United States (e.g., [Grimm and Jacobson, 2004](#)) suggest that such lags may have lasted for centuries, not for millennia. The paleoclimatic interpretations from the two disciplines (beetles and pollen) seem to be coming into closer agreement in recent years. For instance, a syntheses of palynological data ([Grimm and Jacobson, 2004](#)) concluded that “the climate of the Northeast was warmest and driest during the early Holocene ... from about 10 200–6000 cal yr BP—a time of enhanced summer insolation.” This conclusion agrees well with fossil beetle reconstructions.

[Elias \(1982\)](#), [Lavoie et al. \(1997\)](#), and [Lavoie and Arsenault \(2001\)](#) studied late Holocene beetle assemblages from sites near arctic treeline in northern Quebec and Labrador, Canada. Based on MCR analysis, [Lavoie et al. \(1997\)](#) found that summer temperatures in northern Quebec were 2.8–5.5 °C higher than modern from 6.7–1.9 ka. [Elias \(1982\)](#) found evidence of climatic cooling in northeastern Labrador from 2.7–0.9 ka. However, [Lavoie and Arsenault \(2001\)](#) found no evidence of cooling in the James Bay region during the last 5.7 ka. This suggests either that climates varied in different parts of Quebec–Labrador during the late Holocene, or that the northern tree-line faunas were more sensitive to small-scale changes than faunas that lived nearer the center of the boreal forest.

Environmental Archaeology Studies

One of the principal differences between European and North American postglacial environments has been the relatively

strong human influence on the landscapes of Europe during at least the last 5000 years, compared with the relatively weak human disturbance of many North American landscapes until at least late prehistoric times. The difficulties experienced by European researchers in disentangling the effects of human perturbation from the effects of climate change on postglacial beetle faunas (see [Postglacial Europe](#)) have not hindered North American workers to the same extent. The hunter-gatherer societies of the Americas did not clear forests and plant crops until the last few centuries.

Fossil beetle assemblages taken from archaeological sites in eastern and central North America have documented physical environments and human living conditions at several sites. Fossil insects were studied from disposal pits at Cahokia, Illinois ([Pauketat et al., 2002](#)). One of the aims of this study was to determine the timing of the disposal events represented by deposition of refuse in barrow pits. All but one of the studied deposits contained sizable quantities of insects active during the warm months between April and October. The fossil assemblages were dominated by beetles (58%), fly pupae (28%), and ants (14%). Some assemblages were indicative of rich mounds of discarded plant and animal wastes that were exposed for sufficient time, several days to several weeks, to allow insect colonization.

[Bain \(1998\)](#) studied beetle remains from a colonial-era archaeological site in Boston, Massachusetts. The fossil assemblages were extracted from sediments that accumulated in a latrine. Bain identified 42 beetle species, including 22 European taxa, and one that was possibly of Caribbean origin. Five beetle species in the fossil assemblage normally attack wood, suggesting that woodworking took place nearby, such as a sawmill, a furniture shop, house building, or stacks of

firewood. A large number of granary weevils (*Sitophilus granarius*) were found, indicating weevil infestation of stored grain or flour.

Bain's (2001) study of insects from the Îlot Hunt site in Quebec City, Canada indicates that the late nineteenth century was a time of great change in urban sanitary regulations and awareness of public health problems. Bain analyzed insect fossil samples from two latrines, a drain, and an abandoned well, representing both domestic and commercial establishments.

Research in Western North America

Postglacial sites in western North America include eight sites in the Rocky Mountain region (Table 2, sites 1–8) and twelve sites in the arid southwest (Table 2, sites 9–20). The Rocky Mountain sites represent organic deposits sampled from lake sediments, peat bogs, and one cave deposit (Rat's Nest Cave, Alberta). The southwestern sites are all woodrat middens.

Studies in the Rocky Mountains

In many ways, the Rocky Mountain region is ideally suited for Quaternary insect studies. The vertical relief of the mountain chain has provided regional biota with a wide variety of physical environments, expressed in the development of ecological communities ranging from grasslands in the eastern foothills zone through several types of montane and subalpine forest communities, to the alpine tundra at or near mountaintops. Furthermore, the ecotones, or boundaries, between these communities have undergone noticeable shifts in elevation in the late Quaternary, facilitating the study of environmental change, as a given site documents different biological communities through time.

Fossil insects from the Rocky Mountains have left us important records of past events and biotic responses. The atmospheric moisture caught by the Rockies makes this mountain chain wetter than the adjacent plains, and this in turn has led to the accumulation and preservation of organic deposits.

Rocky Mountain study sites (Figure 2) span an elevational range from the foothills zone to the Alpine tundra. The general trend in regional faunas was that cold-adapted species lived at high elevations during the postglacial warm interval, but during the last glaciation, these beetles shifted their ranges downslope by 300–1500 m (Short and Elias, 1987; Elias and Toolin, 1989).

MCR analysis of fossil beetle assemblages from the Rockies indicates that following deglaciation, postglacial summer warming was rapid. This is indicated by an assemblage dating 16.2 ka, with TMAX only 2.1 °C cooler than modern. By 11.6 ka, several assemblages indicate warmer than modern TMAX and TMIN values. Pollen evidence suggests that coniferous forests did not reach their modern elevations until about 10 ka, more than 1000 years after climatic amelioration had reached modern levels, based on the beetle evidence. By that time, beetle MCR reconstructions indicate that both summer and winter temperatures were already beginning to decline from early postglacial maxima.

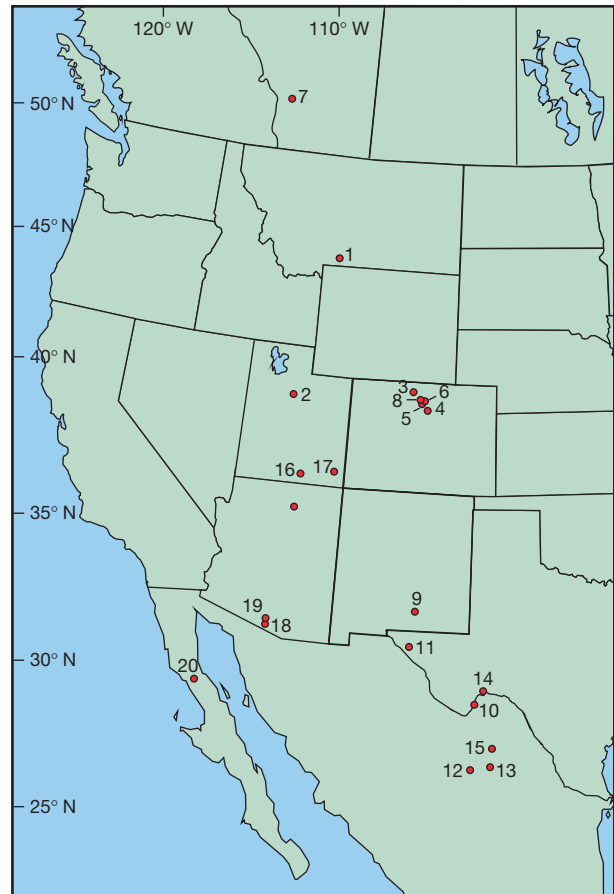


Figure 2 Map of western North America, showing location of fossil sites discussed in text. Site numbers match those in Table 2.

TMAX values remained above modern levels and TMIN values remained below modern levels until about 3.2 ka. These long-term trends were followed by a series of small-scale oscillations in the late Holocene. Pollen spectra from sites in the Colorado Front Range show substantial changes after 3.8 ka, interpreted as either a lowering of treeline, or a thinning of subalpine forests, or both (Short, 1985). Pine pollen also became less abundant in subalpine pollen spectra at this time. These changes were inferred to be a response by the vegetation to climatic cooling.

The site of Rat's Nest Cave in Alberta yielded organic detritus that accumulated at the bottom of a deep natural shaft inside the mouth of a cave (Bain et al., 1997). The beetle faunas span the last 2000 years, and their composition suggests environmental conditions similar to modern parameters at the site. This is in contrast to beetle assemblages from this interval from farther south, in the Central Rocky Mountains. The latter are indicative of a series of climatic oscillations during this time, with a relatively warm period about 1 ka, followed by a relatively cold period in the last 500 years.

Biogeography of the Rocky Mountain Beetle Fauna

Because of the persistently cold climates at high elevations, the Rockies have been a refuge for many cold-adapted species of

both plants and animals, even during the warmest interglacial intervals. Species that shifted southward in front of glacial ice advances in the interior of North America had only two options for survival when the climate began warming and the ice receded at the end of the last glaciation. They had to migrate either northward or up the slopes of mountains, in search of suitably cold climate. Thus, cold-adapted beetles that lived on the plains south of Denver toward the end of the last glaciation have retreated to the Alpine tundra in the Colorado Front Range (Elias, 1986). In contrast to the Rocky Mountain fauna, cold-adapted species living in the central Great Plains region were not able to migrate north at the end of the last glaciation. At that time, climatic warming caused the extirpation (regional extinction) of much of the cold-adapted beetle fauna in this region, with small populations of Arctic and subArctic species surviving in Alpine habitats of the Cordillera and Appalachians, or in short-lived habitats associated with stagnant ice (Schwert and Ashworth, 1988).

There have been a number of regional extirpations from the Rocky Mountain beetle fauna during the last 17 ka, probably as a result of large-scale climatic change (Figure 3). Six of the eight species extirpated before the Holocene have shifted their distributions to the north. These are all cold-adapted animals, either Arctic or boreo-Arctic species. The most extreme displacement of distribution is that of a rove beetle (Staphylinidae), *Holoboreaphilus nordenskiöldi*, which lives today in Arctic Canada, 3000 km north of its Late Glacial locality at Marias Pass, Montana (Figure 2). Two additional species made their last Rocky Mountain appearance in an assemblage from Marias Pass, dated at just over 13.3 ka. These are a ground beetle (Carabidae), *Bembidion sulcipenne hyperboroides*, and a carrion beetle (Silphidae), *Thanatophilus trituberculatus*. *B. sulcipenne hyperboroides* is a northwestern North American species, living today in northern British Columbia, the Yukon Territory, and Alaska. *T. trituberculatus* is a northern, cold-adapted carrion-feeder, which probably also feeds on other insects inhabiting carrion. Its modern range extends north to the Arctic in Canada

and Alaska, but its southern limit does not reach the lower 48 states of the United States.

Two species make their last appearance in the Rocky Mountain fossil record at approximately 14.7 ka, from the Mary Jane site in Colorado. One of these is the water scavenger beetle (Hydrophilidae), *Helophorus splendidus*. This beetle is found today in the northern Holarctic regions, including northeastern Siberia and the arctic coast of the Canadian Northwest Territories. It was found also in the 17.8 ka-assemblage from Lamb Spring, and is indicative of tundra conditions at this site.

The Huntington Canyon site, in the subalpine zone of the Wasatch Mountains of Utah, yielded one species that has been regionally extirpated. The assemblage dates to 13.2 ka. *Bembidion gordonii* is a riparian ground beetle that lives on the banks of fast-running mountain streams. Today, however, its distribution is restricted to the mountains of the Pacific Northwest, from British Columbia to Oregon. Based solely on its modern distribution, a biogeographer might describe *B. gordonii* as endemic to the Pacific Northwest, but the Huntington fossil record demonstrates that it was more widespread in the recent past (i.e., 'recent' in terms of the longevity of the species).

Of the species extirpated during the Holocene, only one has shifted its distribution to the north. The riparian ground beetle, *Bembidion rusticum*, now lives only in British Columbia and northward, a distance of more than 2600 km from Lake Emma, Colorado, where it was found as an early Holocene fossil. Lake Emma also yielded *Bembidion sierricola*, which is restricted today to the mountains of the Pacific Northwest, from California to southern British Columbia (Lindroth, 1963). The dung beetle (Scarabaeidae) *Dialytes ulkei*, was extirpated from the Rockies sometime after 9600 yr BP. Its last occurrence in the regional fossil record was at La Poudre Pass, Colorado (Figure 2). *D. ulkei* is an eastern North American species today, ranging from southern Quebec, southward in the Appalachian Mountain chain to South Carolina. Finally, the ground beetle, *Selenophorus gagatinus*, was last recorded from Colorado in a

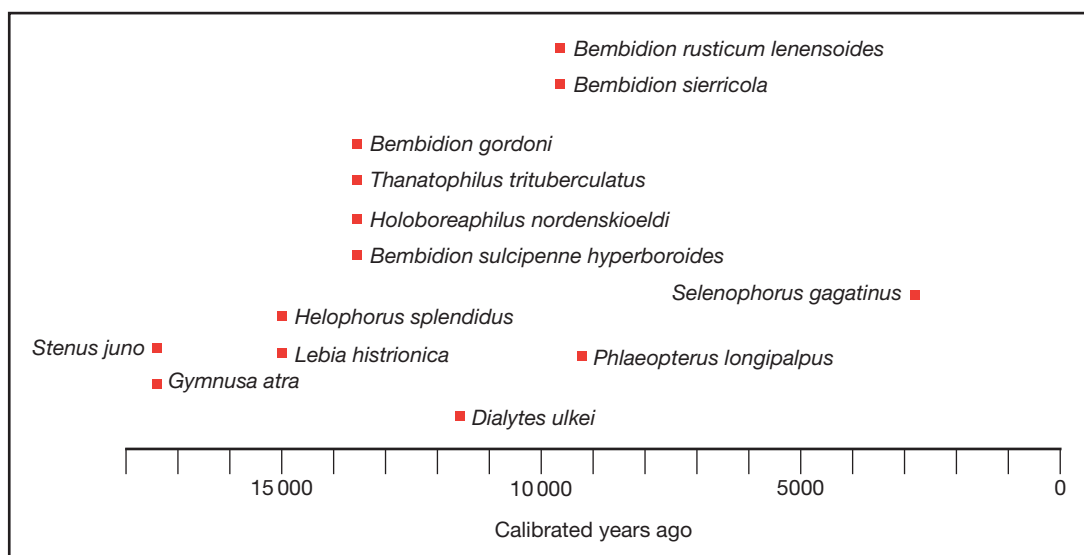


Figure 3 Last recorded fossil occurrence of various beetle species in Rocky Mountain region postglacial fossil assemblages. After Elias, 1981.

2400 yr BP assemblage from the Roaring River, in Rocky Mountain National Park (Figure 2). Today *S. gagatinus* is an eastern North American species, ranging as far west as Indiana. In the late Quaternary interval, it was widespread in western North America, as it was also found in ancient packrat middens from the northern Chihuahuan Desert (Figure 4), in assemblages ranging from 24 to 4.7 ka (Elias and Van Devender, 1990).

Based on the fossil evidence discussed above, it is clear that the current biological communities of the Rocky Mountain region developed only since the end of the last glaciation. They are the product of changing Quaternary environments. Changes in distribution have been large-scale and rapid. These changes demonstrate the great mobility of many beetle species, and their ability to move great distances in order to satisfy their ecological requirements in the face of changing climates. These changes also suggest that the current insect communities of the Rocky Mountain region are simply the latest reshuffling of

species, and that species composition is probably in a more or less continuous state of flux (Elias, 1991).

Since the 1980s, insect fossils from packrat middens have received intensive study. Fossil insect research has been done in the Sonoran Desert and the Great Basin, but the most intensively studied region (191 midden insect fossil samples from 27 sites) is the Chihuahuan Desert (Elias and Van Devender, 1990, 1992; Elias et al., 1995). The fossil insect assemblages from the southern Chihuahuan Desert in Mexico contain mixtures of desert- and temperate-zone species (Figure 4) in almost every interval from the Late Glacial through to the late Holocene. Midden assemblages from locations farther north in the Chihuahuan Desert are generally separated into glacial-age faunas with temperate-zone affinities and postglacial faunas with desert-zone affinities. The 'no modern analog' faunal assemblages indicate that the late Quaternary environments in this region were unlike any that exist today. After 14 ka, most temperate-zone beetle species in the central and northern

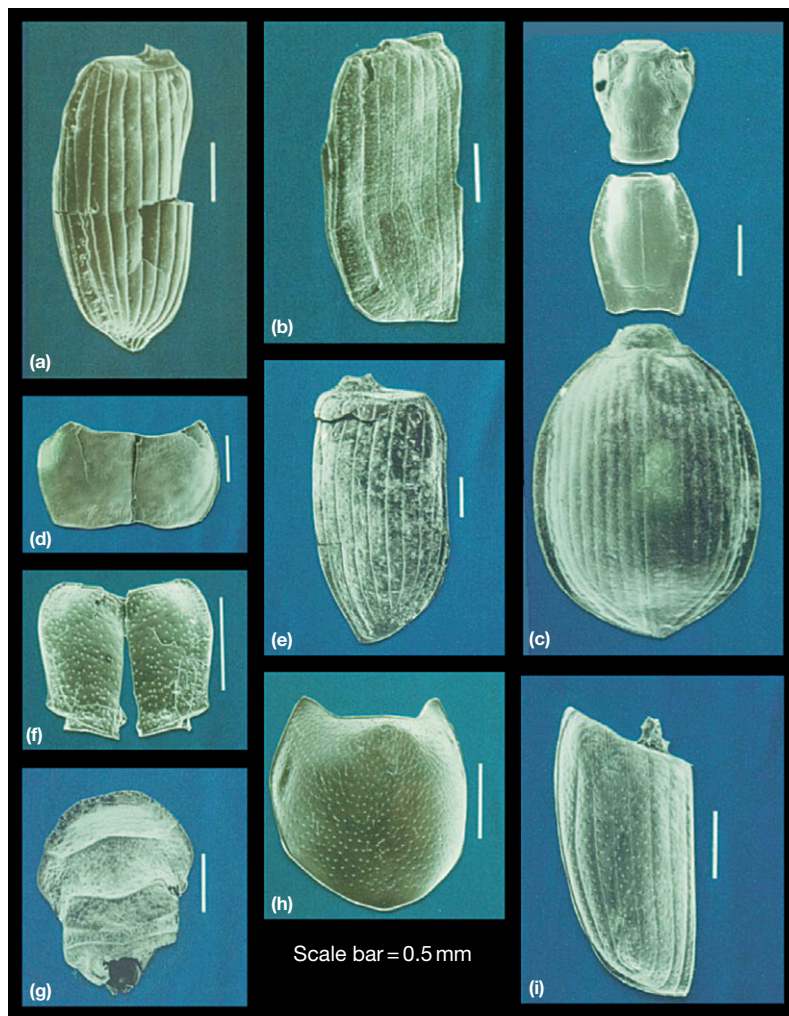


Figure 4 Scanning electron micrographs of fossil beetles from Chihuahuan Desert packrat midden assemblages. (a) Left elytron of the ground beetle *Selenophorus gagatinus*; (b) left elytron of the ground beetle *Lebia lecontei*; (c) head, pronotum, and elytra of the ground beetle *Agonum (Rhadine) longicollis*; (d) pronotum of the ground beetle *Harpalus herbivagus*; (e) right elytron of the ground beetle *Piosoma setosum*; (f) pronotum of the ground beetle *Cymindis* cf. *borealis*; (g) head capsule of the dung beetle *Onthophagus brevifrons*; (h) pronotum of the dung beetle *Onthophagus cavernicollis*; (i) left elytron of the dung beetle *Onthophagus brevifrons*.

regions of the Chihuahuan Desert were replaced by either desert species or by more cosmopolitan taxa. The faunal change suggests a climatic shift from cool, moist conditions of Wisconsin glacial times to hotter, drier conditions in the early Holocene.

In the northern Chihuahuan Desert, woodland environments persisted until 13 ka, but the insect data suggest considerable open ground, with grasses at least locally important at the midden sites. The transition from the temperate (glacial) fauna to the more xeric postglacial fauna started by 14.7 ka. The timing of this faunal change was essentially synchronous throughout the northern Chihuahuan Desert. A major difference between the Big Bend and northern Chihuahuan Desert scenarios is the nature of this faunal change. In the Big Bend region, the transition was characterized by the disappearance of nearly all temperate species at about 14 ka. However, the xeric-adapted fauna did not appear in the Big Bend records until about 8.4 ka. In the northern Chihuahuan Desert assemblages, the xeric species first appeared at 14.7 ka, and several of the temperate grassland species that had lived there during the last glaciation persisted well into the Holocene. This mixture of xeric and temperate elements makes ecological sense, because these northern faunas were living close to the edge of the Chihuahuan Desert. The gradual shifting of northern desert boundaries in the Holocene probably created many marginal habitats for temperate species. By about 8.4 ka, the dominance of xeric species indicates establishment of desert environments, including desert grasslands, throughout the Chihuahuan Desert region. After 2.6 ka, the last of the temperate species was replaced by species associated with desert-scrub communities.

Work on late Quaternary insect faunas from the Colorado Plateau region (Elias et al., 1992) suggests that late Wisconsin climatic conditions were cooler and moister than present, and that the plateau supported a mosaic of grassland and shrub communities without modern analog. The temporal and spatial coverage of midden sites is relatively poor for this enormous region, especially in comparison with the work done on middens from the Chihuahuan Desert. However, it is possible to say that the transition from glacial to postglacial climatic regimes took place after 17 ka. Paleobotanical data suggest that major regional warming was well underway by 13 ka (Cole, 1990). The fossil insect data indicate that the most arid conditions in the Great Basin region occurred within the last 1500 years.

Insect faunas have also been studied from several sites in the Sonoran Desert (Hall et al., 1989, 1990; Clark and Sankey, 1999). Unlike the Chihuahuan Desert insect fauna, the Sonoran Desert fauna indicates little change from the late Pleistocene through to modern times. All taxa found in the Sonoran middens probably live within a few kilometers of the midden sites today. The Sonoran arthropod fauna showed a marked increase in diversity during the late Holocene, as more subtropical plants and warmer climates became established about 4.5 ka. Many warm-adapted beetles probably dispersed into southern Arizona from Sonora, Mexico, during the last 4 kyr. The increase in diversity in fossil arthropod assemblages has been correlated with the frequency of winter freezes, and secondarily with the amount of summer precipitation. This late Holocene peak in species richness is in sharp contrast to the

Chihuahuan Desert insect record, which showed the least number of species in late Holocene samples. If the Sonoran Desert insect fauna truly was stable through the late Quaternary, then this represents a significant difference between the Sonoran and Chihuahuan desert insect faunal histories.

On the whole, the results of North American postglacial beetle studies tend to agree well with reconstructions based on other proxy data, such as pollen and plant macrofossils. The only important discrepancies concern early postglacial environments. The fossil beetle records of all studied North American regions show unequivocally that at least summer temperatures reached modern levels very early in the postglacial period. Warm-adapted beetles were colonizing the recently deglaciated landscapes in the midwest and the Rockies, providing evidence of this warming, even though low herbaceous vegetation persisted in these regions for up to 1000 years after deglaciation. Likewise, in the Chihuahuan Desert, the shift from cool-temperate to warm-xeric beetle faunas anticipated the major changes in regional vegetation (notably, the upslope shift in lower treelines) by up to 1000 years (Elias and Van Devender, 1992). The rapidity and magnitude of postglacial warming identified in the fossil beetle records from both North America and Europe has been verified by rapid changes seen in oxygen isotope ratios of Greenland ice cores (Dansgaard et al., 1989).

See also: [Beetle Records: Overview](#); [Postglacial Europe](#).

References

- Bain A (2001) Archaeoentomological and Archaeoparasitological Reconstructions at Îlot Hunt (CeEt-110): New Perspectives in Historical Archaeology (1850–1900). *British Archaeological Reports* S973.
- Bain AL (1998) A seventeenth century beetle fauna from colonial Boston. *Historical Archaeology* 32: 38–48.
- Bain AL, Morgan AV, Burns JA, and Morgan A (1997) The paleoentomology of Rat's Nest Cave, Grotto Mountain, Alberta, Canada. *Quaternary Proceedings* 5: 23–34.
- Clark WH and Sankey JT (1999) Late Holocene Sonoran Desert arthropod remains from a packrat midden, Cataviña, Baja California Norte, México. *Pan-Pacific Entomologist* 75: 183–199.
- Cole KL (1990) Reconstruction of past desert vegetation along the Colorado River using packrat middens. *Palaeogeography, Palaeoclimatology, Palaeoecology* 76: 349–366.
- Dansgaard W, White JWC, and Johnson SJ (1989) The abrupt termination of the Younger Dryas climate event. *Nature* 339: 532–534.
- Duplessy JC, Labeyrie LD, and Paterne M (1996) North Atlantic sea surface conditions during the Younger Dryas cold event. In: Andrews JT, Austin WEN, Bergsten H, and Jennings AE (eds.) *Late Quaternary Palaeoceanography of the North Atlantic Margins* 111: 167–175. Geological Society Special Publication.
- Elias SA (1982) Holocene insect fossil assemblages from northeastern Labrador. *Arctic Alpine Research* 14: 311–319.
- Elias SA (1986) Fossil insect evidence for late Pleistocene paleoenvironments of the Lamb Spring site, Colorado. *Geoarchaeology* 1: 381–386.
- Elias SA (1991) Insects and climate change: Fossil evidence from the Rocky Mountains. *Bioscience* 41: 552–559.
- Elias SA (2000) Late Pleistocene climates of Beringia, based on fossil beetle analysis. *Quaternary Research* 53: 229–235.
- Elias SA, Anderson K, and Andrews JT (1996) Late Wisconsin climate in the northeastern United States and southeastern Canada, reconstructed from fossil beetle assemblages. *Journal of Quaternary Science* 11: 417–421.
- Elias SA, Mead JI, and Agenbroad LD (1992) Late Quaternary arthropods from the Colorado Plateau, Arizona and Utah. *Great Basin Naturalist* 52: 59–67.

- Elias SA and Toolin LJ (1989) Accelerator dating of a mixed assemblage of late Pleistocene insect fossils from the Lamb Spring site, Colorado. *Quaternary Research* 33: 122–126.
- Elias SA and Van Devender TR (1990) Fossil insect evidence for Late Quaternary climatic change in the Big Bend region, Chihuahuan Desert, Texas. *Quaternary Research* 34: 249–261.
- Elias SA and Van Devender TR (1992) Insect fossil evidence of late Quaternary environments in the northern Chihuahuan Desert of Texas and New Mexico: Comparisons with the paleobotanical record. *Southwest. Naturalist* 37: 101–116.
- Elias SA, Van Devender TR, and De Baca R (1995) Insect fossil evidence of Late Glacial and Holocene environments in the Bolson de Mapimi, Chihuahuan Desert, Mexico: Comparisons with the paleobotanical record. *Palaios* 10: 454–464.
- Grimm EC and Jacobson G Jr. (2004) Late-Quaternary vegetation history of the eastern United States. In: Gillespie AR, Porter SC, and Atwater BF (eds.) *The Quaternary Period in the United States*, pp. 381–402. Amsterdam: Elsevier.
- Hall WE, Olson CA, and Van Devender TR (1989) Late Quaternary and modern arthropods from the Ajo Mountains of southwestern Arizona. *Pan-Pacific Entomologist* 65: 322–347.
- Hall WE, Olson CA, and Van Devender TR (1990) Late Quaternary and modern arthropods from the Puerto Blanco Mountains, Organ Pipe Cactus National Monument, southwestern Arizona. In: Betancourt JL, Van Devender TR, and Martin PS (eds.) *Packrat Middens, the last 40,000 years of biotic change*, pp. 363–379. Tucson: University of Arizona Press.
- Lavoie C, Elias SA, and Payette S (1997) Holocene fossil beetles from a treeline peatland in subarctic Québec. *Canadian Journal of Zoology* 75: 227–236.
- Lavoie C and Arseneault D (2001) Late Holocene climate of the James Bay area, Québec, Canada, reconstructed using fossil beetles. *Arctic, Antarctic, and Alpine Research* 33: 13–18.
- Levesque AJ, Cwynar LC, and Walker IR (1997) Exceptionally steep north-south gradients in lake temperatures during the last deglaciation. *Nature* 385: 423–426.
- Lotter AF, Walker IR, Brooks SJ, and Hofmann W (1999) An intercontinental comparison of chironomid palaeotemperature inference models: Europe vs North America. *Quaternary Science Reviews* 18: 717–735.
- Miller RF and Elias SA (2000) Late-glacial climate in Maritime Canada, reconstructed from Mutual Climatic Range analysis of fossil Coleoptera. *Boreas* 29: 79–88.
- Pauketat TR, Kelly LS, Fritz GJ, Lopinot NH, Elias SA, and Hargrave E (2002) The residues of feasting and public ritual at early Cahokia. *American Antiquity* 67: 52–65.
- Schwert DP and Ashworth AC (1988) Late Quaternary history of the northern beetle fauna of North America: A synthesis of fossil distributional evidence. *Memoirs of the Entomological Society of Canada* 144: 93–107.
- Short SK (1985) Palynology of Holocene sediments Colorado Front Range: Vegetation and treeline changes in the subalpine forest. *American Association of Stratigraphic Palynologists Contribution Series* 16: 7–30.
- Short SK and Elias SA (1987) New pollen and beetle analysis at the Mary Jane site, Colorado: Evidence for Late-Glacial tundra conditions. *Geological Society of American Bulletin* 98: 540–548.

Beringia *see* **Archaeological Records**: Global Expansion 300 000–8000 Years Ago, Americas; **Beetle Records**: Late Pleistocene of North America; **Dune Fields**: High Latitudes; **Glaciations**: Late Pleistocene Glacial Events in Beringia; **Paleoceanography, Records**: Postglacial North Pacific; **Plant Macrofossil Records**: Arctic North America; **Pollen Records, Late Pleistocene**: Northern North America; **Vertebrate Records**: Late Pleistocene of North America

Biogenic Carbonate Studies *see* **Carbonate Stable Isotopes**: Nonmarine Biogenic Carbonates

Bond Cycles *see* **Paleoclimate Reconstruction**: Sub-Milankovitch (DO/Heinrich) Events

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CARBONATE STABLE ISOTOPES

Contents

Overview

Speleothems

Terrestrial Teeth and Bones

Terrestrial Organic Materials

Non-Lacustrine Terrestrial Studies

Lake Sediments

Nonmarine Biogenic Carbonates

Overview

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Concepts in Stable Isotope Geochemistry

The light elements oxygen (O), hydrogen (H), carbon (C), and nitrogen (N) have two or more stable isotopes whose relative abundances vary slightly in nature. Since the early 1950s, geochemists have used isotopic variations in sedimentary materials to reconstruct past changes in climate. [Emiliani \(1955\)](#) first showed cyclical variations in the $^{18}\text{O}/^{16}\text{O}$ ratios of Foraminifera from deep-sea sediments on a 10^5 year timescale. He inferred that these changes reflected cyclical advances and

retreats of continental ice sheets causing coupled changes in both the temperature and $^{18}\text{O}/^{16}\text{O}$ ratio of seawater. While some of the best-known and most continuous isotopic records of Quaternary paleoclimate are arguably from deep-sea sediments ([Shackleton and Opdyke, 1973](#)) and continental ice cores, isotopic records from continental loci also provide important archives of past climate change.

The physical principles behind the use of isotope records can be summarized as follows. When a material is forming either as a result of chemical precipitation or biological activity,

the isotope ratios of the constituent atoms can reflect the temperature of the depositional process. As well, the isotope ratio of the newly formed material reflects the ratio in the medium from which it is formed (usually water) and this in turn can be controlled by climate change. When crystals of calcium carbonate (CaCO_3 ; calcite or aragonite) precipitate at equilibrium, their $^{18}\text{O}/^{16}\text{O}$ ratio is offset from that of the aqueous medium by a fractionation factor $\alpha_{\text{ct-w}}$ that depends on temperature, generally decreasing (causing $^{18}\text{O}/^{16}\text{O}$ of calcite to increase) by $\sim 0.24\text{‰}/^\circ\text{C}$. Likewise, the $^{18}\text{O}/^{16}\text{O}$ and deuterium/hydrogen (D/H) ratios of meteoric precipitation increase with temperature (by $\sim 0.7\text{‰}/^\circ\text{C}$ for $^{18}\text{O}/^{16}\text{O}$) except in the tropics where $^{18}\text{O}/^{16}\text{O}$ is a decreasing function of the amount of precipitation. Isotopes in surface waters derived from rain and snow-melt will likewise reflect changes in average temperature and rainfall. The $^{18}\text{O}/^{16}\text{O}$ ratios of minerals precipitated from water (including apatite in bones and teeth) reflect the ratio in environmental water.

Isotope ratios are reported using the delta (δ) notation. For example:

$$\delta^{13}\text{C} = \{[(^{13}\text{C}/^{12}\text{C})_{\text{sample}} / (^{13}\text{C}/^{12}\text{C})_{\text{std}}] - 1\} \times 1000$$

where 'std' is the standard: VPDB for $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ in calcite, VSMOW for $\delta^{18}\text{O}$ in water, and for δD (derived from $^2\text{H}/^1\text{H} \equiv \text{D}/\text{H}$).

There is a relative paucity of continuously deposited Quaternary sediments from which to extract stable isotopic records in the terrestrial realm compared to the marine record. Most terrestrial isotopic records for this period relate to specific, short-lived sedimentary loci or to collective records of disparate events such as soil profiles. The sections of the Encyclopedia on this topic largely present methodological treatments of the use of isotopes in paleoclimate rather than summarizing detailed climatic records (except for the Holocene, for which quasi-continuous records are more common). The highlights of these sections are summarized below.

Speleothems

Some of the longest, most accurately dated, and continuous isotopic terrestrial isotopic records are taken from deposits of CaCO_3 formed in caves, principally stalagmites and horizontally laminated flowstone. These deposits form as a result of supersaturation of calcium-rich solutions dripping from the cave roof. When deposited deep inside poorly ventilated caves, they can be formed in O isotopic equilibrium with drip water and therefore have the potential to record paleotemperatures. Drip waters isotopically resemble average meteoric precipitation above the cave; therefore $\delta^{18}\text{O}$ of the calcite ($\delta^{18}\text{O}_{\text{ct}}$) depends on both $\delta^{18}\text{O}$ of precipitation (controlled by temperature and rainfall amount) as well as the fractionation between calcite and water, $\alpha_{\text{ct-w}}$. The net effect can be either an increase or a decrease in $\delta^{18}\text{O}_{\text{ct}}$ with increasing atmospheric T (both are observed). Serial records drilled out of speleothems have been used to reconstruct the history of local and regional climate. Their precise U-series chronologies allow correlation with isotopic records in ice and ocean cores. Variation in $\delta^{13}\text{C}_{\text{ct}}$

records changes in organic activity above the cave including changes in amount of C_3 and C_4 plants.

Absolute temperatures of formation of speleothem calcite may be obtainable by analysis of fluid inclusions in the speleothem which represent trapped samples of original drip water. While the $^{18}\text{O}/^{16}\text{O}$ ratio may have changed due to post-trapping exchange, the D/H ratio of the included water can be used to reconstruct initial $\delta^{18}\text{O}$ using the meteoric water relationship $\delta\text{D} = 8 \delta^{18}\text{O} + \delta_o$ where $\delta_o \sim 10\text{‰}$. Thus, speleothems have the potential to provide archives of surface temperatures over the last 0.4 million years (My).

Teeth and Bones

The teeth and bones of mammals consist of hydroxyapatite (HA) and collagen. They contain records of isotopes of O and C (in HA) and C and N (in collagen). The $^{18}\text{O}/^{16}\text{O}$ of HA is mainly determined by the $^{18}\text{O}/^{16}\text{O}$ of consumed water including leaf water (for herbivores). $\delta^{18}\text{O}$ of HA reflects temperature change as it affects $\delta^{18}\text{O}$ in precipitation, and also relative humidity (RH) which causes leaf water to increase in ^{18}O . Serial sections of enamel in single teeth record seasonal changes in temperature or rainfall amount, while HA in bone (which is continuously remodeled) records long-term average climate.

The C and N isotopic records of bone have also been shown to vary with climate. In arid-to-semiarid zones, $\delta^{15}\text{N}$ increases with decreasing rainfall (Heaton et al., 1986) while $\delta^{13}\text{C}$ can increase as RH falls, due both to change in water balance and to increase in the proportion of C_4 plants in the diet.

Bones and teeth represent a highly discontinuous isotopic record, because they do not generally occur in long stratified sequences. An exception is at archaeological sites in Eurasia and Africa where more or less continuous sequences can be assembled from faunal remains at these sites.

Biogenic Carbonates

The shells of mollusks and other CaCO_3 -precipitating fauna are formed approximately in O isotopic equilibrium with the water in which they live. Therefore their $\delta^{18}\text{O}$ values reflect the temperature of deposition, partly because of the temperature-dependence of $\alpha_{\text{ct-w}}$ which decreases by $0.24\text{‰}/^\circ\text{C}$ and also because the $\delta^{18}\text{O}$ of the meteoric source of most surficial waters increases by ca. $0.7\text{‰}/^\circ\text{C}$. In larger lakes and rivers $\delta^{18}\text{O}$ of water is less variable and the effect of temperature on $\alpha_{\text{ct-w}}$ is dominant. There may be a *vital effect* leading to an offset of $\delta^{18}\text{O}$ from equilibrium, but usually the T-slope remains close to $-0.24\text{‰}/^\circ\text{C}$. Seasonal changes in $\delta^{18}\text{O}_{\text{ct}}$ of shells are recorded in annual layers of the shells and can be used to reconstruct seasonality. $\delta^{13}\text{C}$ values of these shells mainly reflect the $\delta^{13}\text{C}$ of dissolved inorganic carbon (DIC) in the water mass. This in turn is influenced by photosynthetic activity (which raises $\delta^{13}\text{C}$ of DIC). The $\delta^{18}\text{O}_{\text{ct}}$ of terrestrial snail shells is controlled by the $\delta^{18}\text{O}$ of water in the leaves of the snail's diet, and mainly reflects changes in RH (Goodfriend et al., 1989) while $\delta^{13}\text{C}$ reflects that of the leafy diet (Goodfriend, 1990).

The aragonitic otoliths of fish are also precipitated in O isotopic equilibrium while $\delta^{13}\text{C}$ reflects DIC of blood; the

latter is controlled by metabolic activity which can change with age and also with the level of activity of the fish.

Long sequential isotopic records of fossil shells in terrestrial sequences are rare due to the lack of extensive exposures of lacustrine sediments, but these can be used effectively to monitor past climate change. Aragonitic ostracod tests have also been used for this purpose in proglacial lake sediments (Schwarcz and Eyles, 1991).

Lacustrine Sediments

In addition to the shells that are enclosed in these sediments, part or most of the sediment itself may consist of chemically or biochemically precipitated CaCO_3 in the form of marl. The $\delta^{18}\text{O}$ of these sediments (generally formed close to isotopic equilibrium) varies in response to temperature, but may also reflect RH in evaporative lakes. Carbon isotopes are controlled by $\delta^{13}\text{C}$ of DIC, and strongly reflect changes in photosynthetic activity in the lake.

Nonlacustrine Terrestrial Carbonates

CaCO_3 is deposited in some arid-region soils as a result of evaporation of Ca-rich water in the soil in the presence of biogenic carbon dioxide (CO_2) and bicarbonate (HCO_3^-). The $\delta^{18}\text{O}$ of these carbonates is strongly influenced by evaporative enrichment of the soil water in ^{18}O with the result that there is no simple relation between $\delta^{18}\text{O}_{\text{ct}}$ and temperature. A regional trend with a slope of $+0.4\text{‰}/^\circ\text{C}$ is observed for soils experiencing low evaporation, close to the value expected for the balance between the meteoric water and $\alpha_{\text{ct-w}}$ effects discussed above.

Diffusive loss of soil CO_2 to the atmosphere leads to a dependence of $\delta^{13}\text{C}$ in soil carbonates on partial pressure of CO_2 in the atmosphere. More important for Quaternary soils, $\delta^{13}\text{C}_{\text{ct}}$ reflects the balance between C_3 and C_4 plants growing on the soil. Because of the preference of C_4 plants for warm, dry climate, this allows us to use variations in $\delta^{13}\text{C}_{\text{ct}}$ in stacked sequences of paleosols as an archive of paleoclimate.

A unique terrestrial carbonate deposit, which could also be considered a speleothem, is the buildup of calcite on the walls of Devil's Hole, a warm spring in Nevada, USA (Winograd et al., 1992). A well-dated record of secular variation in $\delta^{18}\text{O}$ resembles in many precise details the corresponding $\delta^{18}\text{O}$ record for deep-sea Foraminifera. There is, however, a striking offset in the ages of terminations (interglacial-glacial transitions) between the U-series-dated Devil's Hole deposits and the astronomically dated marine record. This discrepancy remains unresolved.

Terrestrial Organic Materials

Plant remains are found in terrestrial sediments both as organic rich peaty deposits and as traces of plant materials in

detrital sediments. Photosynthates (largely cellulose) are significantly depleted in ^{13}C with respect to atmospheric CO_2 as a result of multistage isotopic fractionations in leaves. The degree of depletion varies as a result of change in stomatal opening and is therefore an indicator of water availability. O and H isotopes reflect both the respective isotope ratios in meteoric water (and soil or ground water derived from it), as well as evaporative effects in leaves that enrich leaf-water in heavy isotopes. Thus, a combination of C, O, and H analyses of plant materials can provide a complex image of climatic change.

Serial samples of tree rings provide the most widely cited continuous isotopic records of plant materials. Peat deposits can provide sequences of organic matter mainly derived from mosses. Advances in mass spectrometry have opened up the possibility of isotopic analysis of pollen grains which form a minor component of many terrestrial sediments.

Conclusions

In general, stable isotope ratios in terrestrial materials are affected to approximately equal extents by two or more environmental parameters (including temperature, RH, rainfall, and $\delta^{18}\text{O}$ of meteoric or surface water). It is seldom possible to give a simple and unambiguous interpretation of a serial isotopic record in a sedimentary sequence. Nevertheless, the conflation of multiple isotopic indicators, sometimes from the same material (using different isotopes) can narrow down the possible interpretations assignable to a given series of wiggles on an isotopic graph. Our ability to analyze successively smaller samples opens up the possibility of extracting individual biomolecules for analysis, as well as higher spatial resolution in analyses of serial samples from single deposits such as speleothems or shells.

See also: Carbonate Stable Isotopes: Non-Lacustrine Terrestrial Studies; Nonmarine Biogenic Carbonates; Speleothems; Terrestrial Organic Materials; Terrestrial Teeth and Bones. Ice Core Records: Chinese, Tibetan Mountains.

References

- Emiliani C (1955) Pleistocene paleotemperatures. *Journal of Geology* 63: 538–578.
- Goodfriend GA (1990) Rainfall in the Negev Desert during the Middle Holocene, based on ^{13}C of organic matter in land snail shells. *Quaternary Research* 34: 186–197.
- Goodfriend GA, Magaritz M, and Gat JR (1989) Stable isotope composition of land snail body water and its relation to environmental waters and shell carbonate. *Geochimica et Cosmochimica Acta* 53: 3215–3221.
- Heaton JHE, Vogel JC, von la Chevallerie G, and Collet G (1986) Climatic influence on the isotopic composition of bone nitrogen. *Nature* 322: 822–823.
- Schwarcz HP and Eyles N (1991) Laurentide ice sheet extent inferred from stable isotopic composition (O, C) of ostracodes at Toronto, Canada. *Quaternary Research* 35: 305–320.
- Shackleton NJ and Opdyke ND (1973) Oxygen isotope and paleomagnetic stratigraphy of equatorial Pacific core V28-238: Oxygen isotope temperatures and ice volumes on a 10^5 and 10^6 year time scale. *Quaternary Research* 3: 39–53.
- Winograd IJ, Coplen TB, Landwehr JM, et al. (1992) Continuous 500,000-year climate record from vein calcite in Devils Hole, Nevada. *Science* 258: 255–260.

Speleothems

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Speleothems are crystalline deposits of calcium carbonate (CaCO_3) formed in karstic caves as a result of precipitation from dilute aqueous solutions entering the cave, typically as drips from the roof. Karstic caves (q.v.) are formed by dissolution of soluble bedrock, usually a carbonate rock, by the action of meteoric water percolating from overlying soil or bedrock surfaces. This is partly due to organic activity in the soil that generates carbonic acid, which reacts with calcite or dolomite. Karstic dissolution forms cavities varying in size from less than 1 mm to passages many kilometers in length and tens of meters high (Ford and Williams, 1989). Karst can form in other kinds of rock (e.g., evaporitic sulfates), and speleothems can be formed of minerals other than CaCO_3 . However, this article will deal only with Ca carbonates.

Speleothems can be seen forming today and we can infer that some have formed throughout the Quaternary epoch. They occur in many geometric forms, including stalagmites (which grow upward from the floor of caves beneath a drip site), stalactites (growing downward from the roof of a cave), and flowstone—sheet-like horizontally or inclined layers of calcite produced where water from one or more sources has spread as a film across a cave floor or wall (Figure 1). When viewed in cross section (axially or longitudinally), many speleothems display a prominent laminated structure (Figure 2). These preserve records of secular variation in the properties of the speleothem material that can be interpreted as changes in conditions of deposition in the cave and thus, indirectly, on the Earth's surface above. These properties include the following:

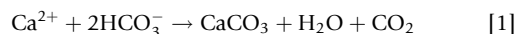
- Chronometric age and rate of growth
- Isotopic composition of the CaCO_3
- Abundance and isotopic composition of fluid inclusions of water trapped during growth
- Trace element concentrations in CaCO_3
- Abundance and composition of organic impurities (especially humic substances)
- Mineralogy (calcite, aragonite, etc.)

Here, we focus on the first three items.

Stable Isotope Systematics in Speleothems

Before percolating through the bedrock, surface water is exposed first to atmospheric CO_2 and then to soil gases enriched in biogenic CO_2 . As water percolates through the bedrock overlying a cave, it can react extensively with the rock or may pass through with relatively little chemical interaction, depending on the nature of the openings through which it passes. Under conditions of diffuse flow through relatively tight fractures and joints, the water can approach chemical and isotopic equilibrium with the rock, whereas in larger, open channels entering caves, surface water may enter essentially unchanged.

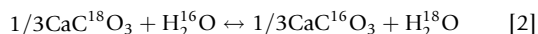
On entering the cave, drip waters can begin to precipitate CaCO_3 as a result of two discrete processes: evaporation, which increases the concentration (activity) of Ca ions until the solubility of CaCO_3 is exceeded, and outgassing of CO_2 from the water, which results in the slow precipitation of CaCO_3 :



CaCO_3 potentially contains a number of distinct isotopic signals, recorded as variations in the relative abundance of the isotopes of Ca, O, and C. Only the latter two have been extensively studied.

Oxygen Isotopes

Variations in the $^{18}\text{O}/^{16}\text{O}$ ratio in water and O-bearing minerals have been widely used as proxies for various environmental records, especially temperature. CaCO_3 can isotopically exchange with water:



Epstein et al. (1953) showed that the equilibrium constant for this reaction, $K_{\text{ct-w}} = (^{18}\text{O}/^{16}\text{O})_{\text{ct}} / (^{18}\text{O}/^{16}\text{O})_{\text{H}_2\text{O}}$, depended only on temperature. Therefore, if a sample of CaCO_3 had precipitated in isotopic equilibrium with water of known $^{18}\text{O}/^{16}\text{O}$ ratio, we could determine the temperature of precipitation. Hendy (1971) showed that under conditions widely encountered in deep caves, speleothems formed at oxygen isotopic equilibrium with drip waters. This happened if the mode of deposition was by relatively slow outgassing of CO_2 from the drip water, whereas evaporative deposition (or rapid outgassing) led to disequilibrium precipitation. That is, under the latter conditions the oxygen isotope fractionation was not equal to the equilibrium constant at the temperature of deposition, and deviation from equilibrium increased as drip water traversed the surface of the speleothem. Hendy showed that the calcite precipitated along this path became progressively enriched in ^{18}O . Therefore, a convenient test for equilibrium was to measure the $^{18}\text{O}/^{16}\text{O}$ ratio of calcite along several points spaced some centimeters apart on a single growth layer. For equilibrium deposits it was expected that $(^{18}\text{O}/^{16}\text{O})_{\text{ct}}$ would remain constant, whereas disequilibrium growth layers would display (1) continuously varying $^{18}\text{O}/^{16}\text{O}$ and (2) a correlation between $^{18}\text{O}/^{16}\text{O}$ and $^{13}\text{C}/^{12}\text{C}$. The latter reflects the fact that rapid outgassing or evaporation were usually coupled with variations in the $^{13}\text{C}/^{12}\text{C}$ ratio of dissolved inorganic carbon (DIC) in the drip water. Speleothems satisfying the criteria of equilibrium deposition are invariably found deep inside long caves (>1 km) or in recently opened, formerly sealed cavities, where relative humidity was ~100% and where lack of significant movement of air resulted in high partial pressure of CO_2 with the result that outgassing of drip water was slow. Recent

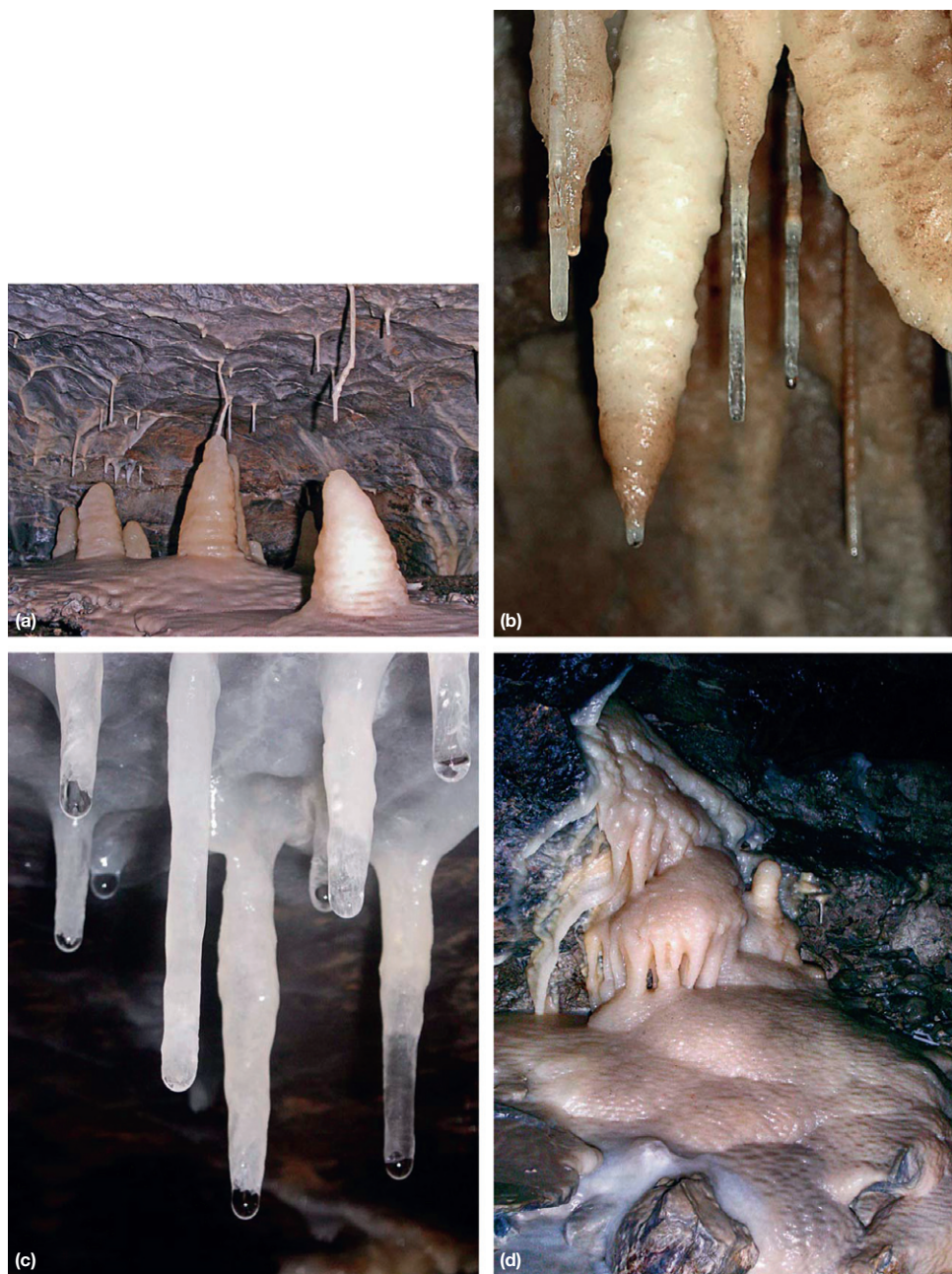


Figure 1 Typical forms of speleothems: (a) stalagmites (middle stalagmite is ~40 cm high); (b) stalactite, ~4 cm in diameter at top, flanked by 'soda straws'; (c) thin stalactites (soda straws), ~5 mm in diameter, showing drip water emerging; and (d) flowstone, ~1 m wide. Photos by P. Beddows.

studies (Mickler et al., 2004) suggest that kinetic effects may prevent the attainment of isotopic equilibrium between calcite and water. However, actively growing stalactites ('soda straws') from deep, high-humidity caves have been shown by other authors to yield isotopic temperatures similar to the temperature in the cave.

The discovery of this means of recognizing speleothems deposited at equilibrium opened up the possibility of using $(^{18}\text{O}/^{16}\text{O})_{\text{ct}}$ variations in speleothems as a paleothermometer. Far from the entrances to most caves (>100–500 m), the temperature remains constant seasonally and approaches the mean annual surface temperature above the cave. Therefore, a

paleotemperature record from an equilibrium speleothem would provide a record of past temperatures on the surface for the duration of growth of the speleothem.

For a given sample of calcite extracted from such an equilibrium deposit, the temperature is given by the paleotemperature equation (Epstein et al. 1953):

$$T(^{\circ}\text{C}) = 16.5 - 4.3(\delta^{18}\text{O}_{\text{ct}} - \delta^{18}\text{O}_{\text{w}}) + 0.14(\delta^{18}\text{O}_{\text{ct}} - \delta^{18}\text{O}_{\text{w}})^2 \quad [3]$$

where $\delta^{18}\text{O}_x$ is defined as

$$\delta^{18}\text{O} = \left\{ \left[(^{18}\text{O}/^{16}\text{O})_x / (^{18}\text{O}/^{16}\text{O})_{\text{std}} \right] - 1 \right\} \times 1000 \quad [4]$$

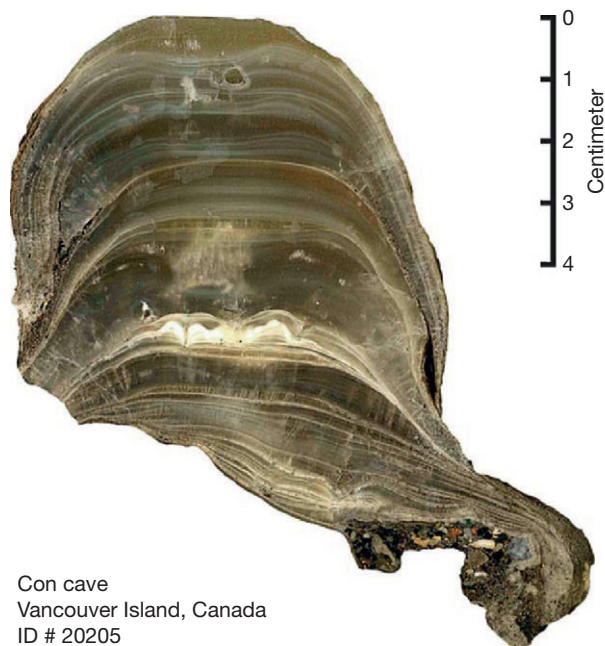


Figure 2 Section through a stalagmite showing laminated structures (growth layers). Lighter zones are richer in fluid inclusions; note the detritus-rich layer at base. The color is due to the presence of humic matter in calcite.

where w is drip water, and std refers to one of the two international standards for $^{18}\text{O}/^{16}\text{O}$ ratios: Vienna Pee Dee Belemnite (VPDB) or Vienna Standard Mean Ocean Water. The units of $\delta^{18}\text{O}$ are ‰ (per mil). The precision of analyses of $\delta^{18}\text{O}$ is typically approximately $\pm 0.1\text{‰}$. Note, however, that in order to determine T , it is necessary to have values for $\delta^{18}\text{O}$ of both the calcite and the water from which it was precipitated.

$\delta^{18}\text{O}_{\text{ct}}$ is usually determined by reacting samples $\geq 100\text{ }\mu\text{g}$ with 100% phosphoric acid and analyzing the resultant CO_2 on a gas-source, isotope ratio-detecting mass spectrometer. Most studies of long-term variation in $\delta^{18}\text{O}_{\text{ct}}$ of speleothems have been done on stalagmites using axial sections (Figure 3) oriented more or less in a diametric section of the stalagmite. Samples are drilled out of the polished surface of the section. High-resolution records (with micrometer spacing) have been obtained using computer-driven milling machines (Mangini et al., 2005) and laser heating to decompose CaCO_3 to $\text{CaO} + \text{CO}_2$ (McDermott et al., 2001). In most studies of secular variation in $\delta^{18}\text{O}_{\text{ct}}$, the authors use the so-called Hendy test to ensure that the deposits were formed in oxygen isotopic equilibrium, reporting $\delta^{18}\text{O}_{\text{ct}}$ and $\delta^{13}\text{C}_{\text{ct}}$ values for samples taken along single growth layers.

It has proven much more difficult to determine values for $\delta^{18}\text{O}_{\text{w}}$. Various approaches to this problem are discussed next.

Carbon Isotopes

The $^{13}\text{C}/^{12}\text{C}$ ratio of the carbonate in calcite is always measured at the same time as $\delta^{18}\text{O}_{\text{ct}}$ during analysis of CO_2 gas on isotope ratio mass spectrometers. Data are reported as $\delta^{13}\text{C}_{\text{ct}}$ with respect to the VPDB standard. The precision of analysis is $\pm 0.1\text{‰}$. Variations in $\delta^{13}\text{C}_{\text{ct}}$ are not significantly temperature

dependent but are rather a consequence of various mixing processes during the formation of DIC.

Dating

Ages of speleothems have been obtained principally by ^{14}C and uranium (U) series dating. Electron spin resonance has been applied with limited success, possibly due to the short lifetime of trapped charges in calcite.

Radiocarbon

^{14}C is introduced into DIC and thus into speleothems through oxidation of plant-derived organic matter in the soil. Ages must be corrected for the presence of dead carbon, nonradioactive C introduced into DIC during dissolution of calcite in the soil zone. From Eq. (1), it appears that approximately half of the C atoms in HCO_3^- ions should be derived from the bedrock calcite, which in most circumstances contains no ^{14}C . Therefore, modern cave-deposited calcite should have a $^{14}\text{C}/^{12}\text{C}$ ratio approximately half that of modern carbon. The initial percentage of dead carbon (dcp) is in fact observed to range from ca. 5 to $<40\%$ (Genty et al., 2001). Lower values are presumably a result of continued exchange between soil-derived CO_2 and recharge water as it percolates through air-filled passages on the way to the cave. The initial dcp cannot be assumed to be constant at a given site but may vary as a result of changing soil activity. Errors due to uncertainty in dcp decrease with age. Speleothems dated by U series have also been used to reveal large oscillations in atmospheric ^{14}C activity (Beck et al., 2001).

U Series

^{238}U decays to a series of short-lived daughter isotopes; after a long time ($>2\text{ My}$), the activity of each of these isotopes (in disintegrations per minute in a given sample) should equal the activity of the parent ^{238}U . However, the daughter isotopes can be geochemically separated from the parent U prior to deposition of calcite because uranium is soluble in groundwater, whereas daughter Th and Pa are insoluble. The age of calcite is determined from the extent of growth of the daughter isotopes thorium-230 (^{230}Th) and ^{234}U into secular equilibrium with ^{238}U and that of protactinium-231 (^{231}Pa) into equilibrium with ^{235}U (Richards and Dorale, 2003). The activity ratio $^{230}\text{Th}/^{234}\text{U}$ increases from zero at the time of deposition to a maximum value of 1 at infinite age. The ratio of ^{234}U to ^{238}U also varies because the radiogenic ^{234}U is a 'hot atom' whose chemical properties during weathering are different from those of its parent ^{238}U .

Speleothems are well suited to U-series dating because their calcite crystals are usually large and have little tendency to recrystallize after deposition. Speleothem calcite typically contains between 100 and 1,000 parts per billion (ppb) U, whereas aragonitic speleothem contains up to 100 ppm U. Some speleothems contain detritus, which is trapped in the calcite during flooding events in caves and which contaminates the samples with ^{230}Th so that the apparent age of the freshly deposited calcite is greater than zero. Analyses of such 'dirty' calcites can be obtained by use of isochrons, in which a non-radiogenic isotope, ^{232}Th , is used to estimate the amount of

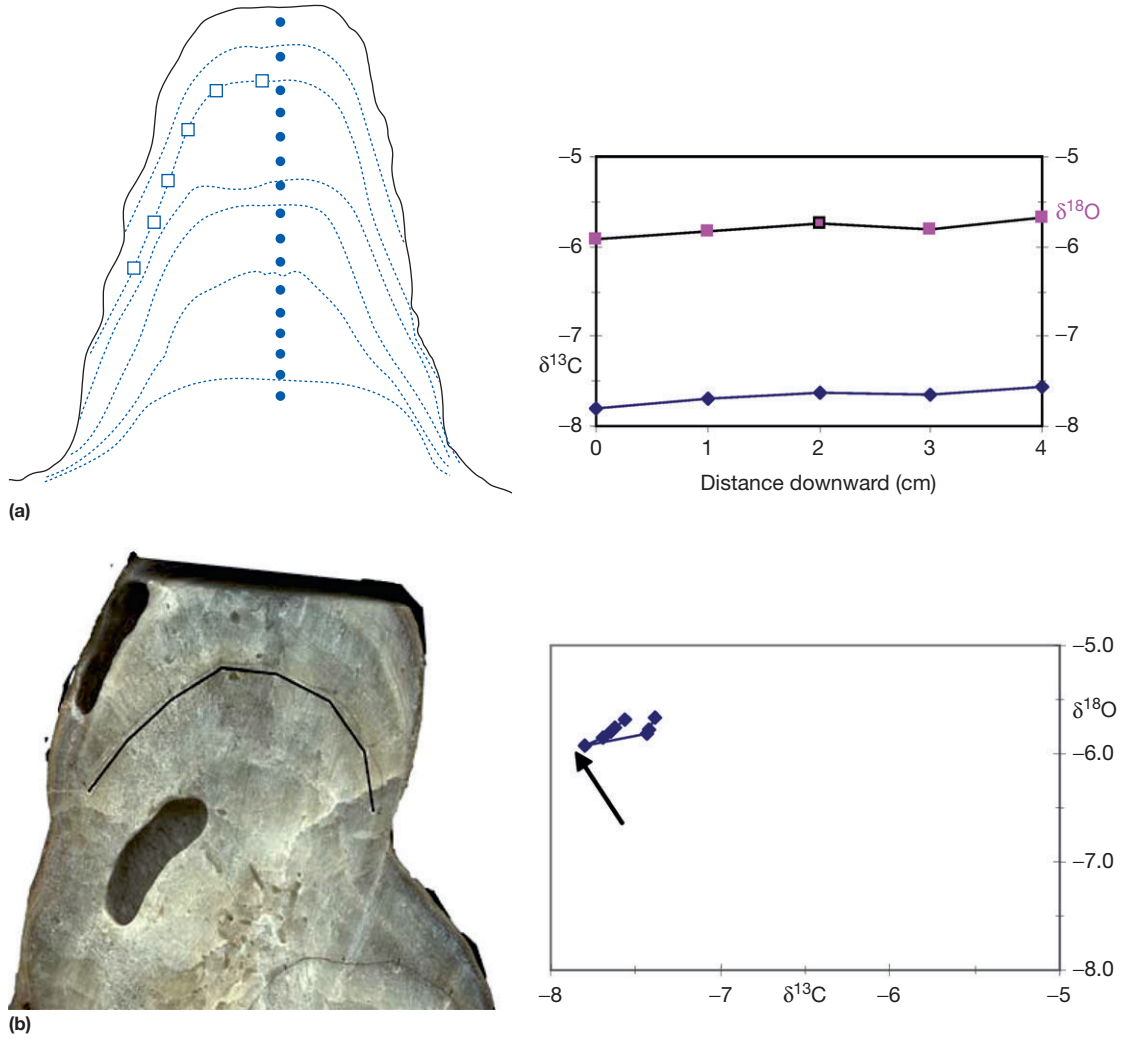


Figure 3 Analysis of layers: (a) Drawing of stalagmite showing sampling localities for Hendy test (squares) and climate records (circles); (b) Hendy test of a growth layer in a Holocene stalagmite from Marengo Cave, Indiana (R. Zhang, unpublished data). (Top graph) $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ along the left side of the growth layer are indicated by the curved line. (Bottom graph) Lack of correlation between $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$; arrow marks the top of the growth layer.

detritus (Bischoff and Fitzpatrick, 1991). Individual analyses can be corrected by using the $^{230}\text{Th}/^{232}\text{Th}$ ratio to estimate the amount of 'detrital' ^{230}Th present in the sample.

U-series isotopes are analyzed using thermal ionization mass spectrometry (TIMS; Edwards et al., 2003) or multicollector inductively coupled plasma mass spectrometry. The precision of analyses of speleothems is better than 1% of the age. Older analyses done by counting of α particles were of lower precision (5–10% of age). The practical age limit for U-series dating is approximately 500 000 years. U-Pb dating by TIMS can be used for older speleothems, although the method requires moderately high U concentrations and analysis of coeval samples with varying U/Pb ratios (Richards et al., 1998).

Paleoclimate Inferred from Oxygen Isotope Records in Speleothems

Eq. (1) shows that for speleothems formed at oxygen isotopic equilibrium, we can infer the temperature of deposition

from $\delta^{18}\text{O}_{\text{Oct}}$ and $\delta^{18}\text{O}_{\text{ppt}}$. In general, we can express a change $\delta^{18}\text{O}_{\text{Oct}}$ as

$$\Delta\delta^{18}\text{O}_{\text{Oct}} = [(d\varepsilon_{\text{ct-w}}/dT) + (d\delta^{18}\text{O}_{\text{ppt}}/dT)]\Delta T + \Delta\delta^{18}\text{O}_{\text{sw}} + \Delta\delta^{18}\text{O}_{\text{s}}$$

where sw is seawater: All meteoric water is derived from the ocean; thus, changes in $\delta^{18}\text{O}$ of seawater during the Pleistocene as a result of changes in ice volume would be reflected in $\delta^{18}\text{O}_{\text{ppt}}$. The last term is a correction to represent changes in the source of water as a result of secular shift in the average path of storm tracks. $\varepsilon_{\text{ct-w}} = (\alpha_{\text{ct-w}} - 1) \times 1000$, and $d\varepsilon_{\text{ct-w}}/dT$ is the temperature (T) dependence of the calcite–water fractionation, approximately $-0.25\text{‰}/^\circ\text{C}$. The second term, $d\delta^{18}\text{O}_{\text{ppt}}/dT$, refers to the long-term (secular) dependence of $\delta^{18}\text{O}_{\text{ppt}}$ on temperature, which is not necessarily equivalent to short-term temperature dependence as observed, for example, in annual records at a single site. The latter T dependence is typically equal to approximately $0.7\text{‰}/^\circ\text{C}$ (Rozanski et al., 1993). If the secular T dependence were also $\sim 0.7\text{‰}/^\circ\text{C}$, then the

net T -dependent change in $\delta^{18}\text{O}_{\text{ct}}$ would be approximately $0.4\text{‰}/^{\circ}\text{C}$. In tropical and subtropical regions, $\delta^{18}\text{O}$ of drip waters can also change as a result of the so-called 'amount effect,' in which $\delta^{18}\text{O}_{\text{ppt}}$ decreases with increasing volume of individual rain events (Rozanski et al., 1993). From the previous relationship, we would expect to find a positive correlation between temperature of deposition and $\delta^{18}\text{O}_{\text{ct}}$. We can test the nature of this correlation in a number of ways:

- Comparison of $\delta^{18}\text{O}_{\text{ct}}$ of modern (or interglacial) speleothems with glacial age samples
- Comparison of $\delta^{18}\text{O}_{\text{ct}}$ record with other coeval, dated paleoclimate records, such as $\delta^{18}\text{O}$ of Foraminifera from deep-sea cores or ice cores or temperatures obtained from Mg/Ca or alkenone ratios in Foraminifera
- Association between a deposit and sedimentary or paleontological deposits of known climatic signature

In published studies of $\delta^{18}\text{O}_{\text{ct}}$ vs age, the correlation between inferred paleotemperature (i.e., paleoclimate, ranging between glacial (cold) and interglacial (warm)) is either positive or negative. We can express this fact in terms of the sign of $\gamma = d(\delta^{18}\text{O}_{\text{ct}})/dT$, which can be positive or negative. This represents the tendency of $\delta^{18}\text{O}_{\text{ct}}$ to rise or fall as local climate becomes warmer or cooler. In principle, we can also have sites with $\gamma = 0$, where the positive and negative climatic effects on $\delta^{18}\text{O}_{\text{ct}}$ exactly cancel out.

Determination of $\delta^{18}\text{O}$ of Paleowaters

In order to determine paleotemperatures from speleothems deposited at isotopic equilibrium, it is necessary to determine $\delta^{18}\text{O}_{\text{w}}$ of the drip water from which the speleothem formed (see Eq. 3). $\delta^{18}\text{O}$ of drip water is close to the annual average of $\delta^{18}\text{O}_{\text{ppt}}$, the $\delta^{18}\text{O}$ value of precipitation falling on land surface above the cave (Yonge et al., 1985; Spötl et al., 2005), but may be locally affected by factors such as surface evaporation and transpiration or selective recharge of seasonal precipitation (e.g., snow meltwater). Most papers reporting $\delta^{18}\text{O}_{\text{ct}}$ data do not explicitly attempt to evaluate variation in $\delta^{18}\text{O}_{\text{ppt}}$ but merely show comparisons between excursions in $\delta^{18}\text{O}_{\text{ct}}$ and other paleoclimate proxies as noted previously. However, at least three distinct approaches have been used to estimate isotopic composition of paleowaters:

- Assuming that secular variation in $\delta^{18}\text{O}_{\text{ppt}}$ is the same as annual (seasonal) variation (Dorale et al., 1992). This assumes a particular value for $d(\delta^{18}\text{O}_{\text{ppt}})/dT$.
- Analysis of fluid inclusions in speleothem, which are presumed to represent the drip water present during growth (Schwarcz et al., 1976). This water can be extracted by crushing the sample in vacuum, and $\delta^{18}\text{O}$ and δD can be determined (Dennis et al., 2001). Although oxygen isotopes may have been exchanged postdepositionally with the calcite, the δD value of the water should be unchanged and can be used to reconstruct $\delta^{18}\text{O}_{\text{ppt}}$ from the meteoric water relationship $\delta\text{D}_{\text{ppt}} = 8 \delta^{18}\text{O}_{\text{ppt}} + \delta_0$, where δ_0 is the so-called deuterium excess. For most continental regions of the world, $\delta_0 = 10\text{‰}$, but this value may have changed through the Pleistocene. For a detailed discussion of the use of fluid inclusions, see McDermott et al. (2006).

- Specific models for variation in $\delta^{18}\text{O}_{\text{ppt}}$, such as derivation of water vapor from seawater whose $\delta^{18}\text{O}$ varied in a known fashion (Frumkin et al., 1999).

Application to Paleoclimate Studies

$\delta^{18}\text{O}_{\text{ct}}$

More than 100 papers report secular variation in $\delta^{18}\text{O}_{\text{ct}}$ of U-series dated speleothems deposited during the past 200 000 years. Most use $\delta^{18}\text{O}_{\text{ct}}$ as a proxy for paleoclimate, an approach that is analogous to the use of $\delta^{18}\text{O}$ analyses of Foraminifera from deep-sea sediments or $\delta^{18}\text{O}$ (and δD) measurements of ice cores. Most studies identify features (excursions or trends) in the speleothem records that match coeval isotopic excursions in the better known proxy records as well as records of pollen, fauna, trace elements in sediments, etc. Thus, we can describe the use of most speleothem $\delta^{18}\text{O}_{\text{ct}}$ data as providing a formal (but not quantitative) proxy for Quaternary paleoclimate. Correlations with $\delta^{18}\text{O}_{\text{ct}}$ records from U-series dated speleothems can be used to give precise dates for critical climatic transitions in other proxy records; they also have been interpreted as indicating how local climate was coupled to global climate change.

An example of a formal proxy is the study of subaqueous calcite precipitated from thermal water in a fissure at Devil's Hole, Nevada (Winograd et al., 1992). In all significant respects, this is identical to studies of speleothems, specifically in that it presents a detailed U-series dated sequence of analyses of $\delta^{18}\text{O}_{\text{ct}}$ for a laminated deposit formed at relatively low temperatures. The $\delta^{18}\text{O}_{\text{ct}}$ record (Figure 4) shows signals corresponding to all the major climatic cycles observed in deep-sea cores from 500 to 70 ka. The timing of the $\delta^{18}\text{O}_{\text{ct}}$ cycles at Devil's Hole is distinctly discordant to that inferred from the Milankovitch theory (e.g., Termination II is approximately 15 000 years earlier than expected).

$\delta^{18}\text{O}_{\text{ct}}$ records from speleothems found in the Soreq Cave and other caves in Israel (Bar-Matthews and Ayalon, 2003; Frumkin et al., 2000) show striking formal agreement between the $\delta^{18}\text{O}_{\text{ct}}$ record and data from ice cores and deep-sea sediments (Figure 5). The correspondence persists over eight oxygen isotope stages, back to 250 000 BP. $\gamma < 0$ for these sites. McGarry et al. (2004) measured δD in fluid inclusions at Soreq. Noting that $\delta_0 = 20\text{‰}$ for Israel today, they used comparisons between alkenone-based sea-surface temperatures and $\delta^{18}\text{O}_{\text{ct}}$ values to infer secular change in δ_0 .

The latter two studies made use of records that spanned glacial–interglacial transitions marked by large shifts in $\delta^{18}\text{O}_{\text{ct}}$. Studies at other sites where the dated speleothem record did not cross a termination generally display smaller scale variations in $\delta^{18}\text{O}_{\text{ct}}$, which nevertheless have been correlated with shorter term cycles in the global paleoclimate record, such as Heinrich events and Dansgaard–Oeschger (D–O) cycles. Examples follow:

Hulu Cave, China: Several speleothems give concordant, partially overlapping records that span a 70 000-year period from ca. 80 to 10 ka (Figure 6). Cycles with amplitudes of $0.5\text{–}1\text{‰}$ ($\gamma < 0$) are correlated by these authors with coeval Heinrich events (cooler periods lasting a few kiloyears

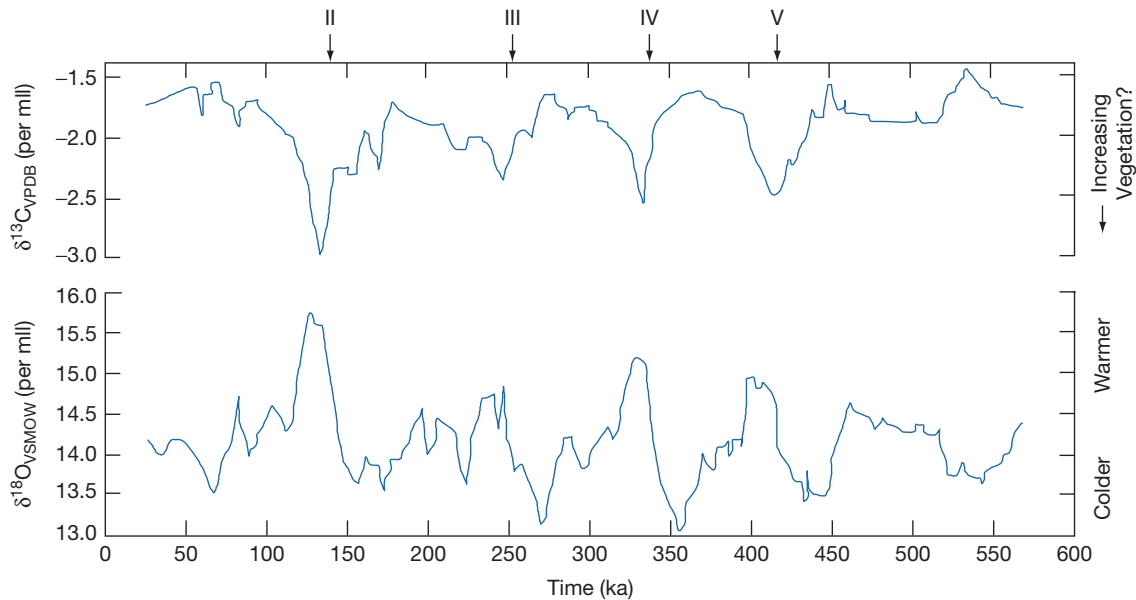


Figure 4 $\delta^{18}\text{O}$ record for Devil's Hole, Nevada. U-series dates are shown at top. From Winograd I, Landwehr J, Ludwig KR, Coplen T, and Riggs AC (1997). Duration and structure of the past four glaciations. *Quaternary Research* 48: 141–154. Reproduced with permission from the National Speleological Society.

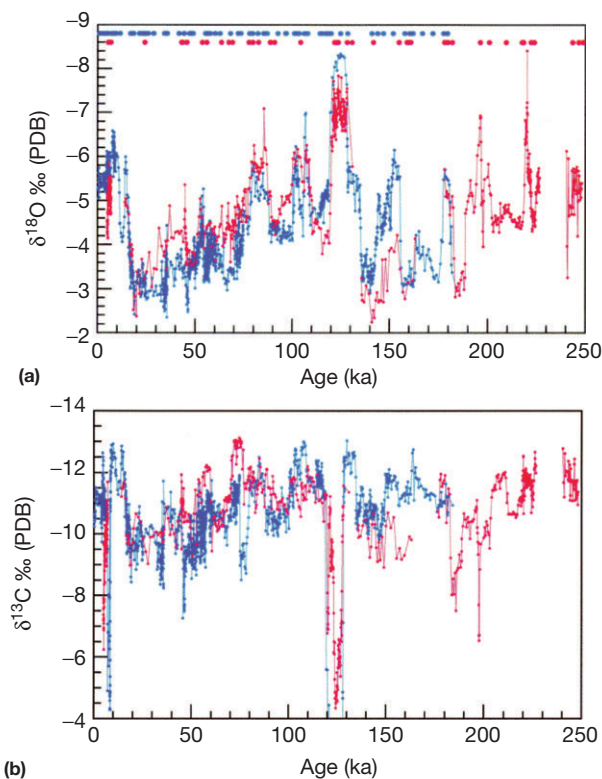


Figure 5 $\delta^{18}\text{O}$ (a) and $\delta^{13}\text{C}$ (b) records for speleothems from Soreq (blue) and Peqiin Cave (red), Israel. Circles in A show U-series ages. From Bar-Matthews M, and Ayalon A (2003) Sea-land oxygen isotopic relationships from planktonic Foraminifera and speleothems in the eastern Mediterranean region and their implication for paleorainfall during interglacial intervals. *Geochimica et Cosmochimica Acta* 67: 3181–3199. Reproduced with permission.

or less). The authors infer that these cycles record variations in monsoon activity (Wang et al., 2001).

Oman: Speleothems from caves in Oman record climatic cycles on a time scale ranging from hundreds of years to 330 ka (Fleitmann et al., 2004). $\delta^{18}\text{O}_{\text{ct}}$ in series of relatively short-lived stalagmites ranges over 8‰ ($\gamma > 0$?). These represent pluvial periods during which the source of moisture has shifted between a continental and monsoonal pattern, as shown by changing δD of fluid inclusions.

Austrian Alps: At the Klee gruben Cave, oscillations in $\delta^{18}\text{O}_{\text{ct}}$ appear to match D-O cycles seen in Greenland ice cores (Spötl and Mangini, 2002; Figure 7) and were used to provide an improved chronology for these cycles; from the correlation with D-O cycles, we infer that $\gamma > 0$ at this site.

North America: At Crevice Cave, Iowa (Dorale et al., 1998), the form of the $\delta^{18}\text{O}_{\text{ct}}$ record is correlated to that of the deep-sea record from 75 to 25 ka BP (with $\gamma > 0$). At Reed's Cave, South Dakota (Serefidin et al., 2003), oscillations in $\delta^{18}\text{O}_{\text{ct}}$ are matched to cycles in Greenland ice cores. Two speleothems from this site have opposite polarities of γ , which the authors attribute to differences in the seasonality of the supply of recharge water. Significant differences in $\delta^{18}\text{O}$ of coeval calcite deposits have been observed in speleothems from other caves.

Holocene records: Records of $\delta^{18}\text{O}_{\text{ct}}$ from Holocene speleothems have shed light on short-term, recent changes in climate in many regions. For example, Burns et al. (1998), who studied the Hoti Cave in Oman, identified pluvial periods in the early Holocene as well as the last interglacial. In Coldwater Cave, Iowa, early Holocene speleothems gave significant differences in $\delta^{18}\text{O}_{\text{ct}}$ between coeval deposits, attributable to evaporative effects on the recharge waters (Denniston et al., 1999); secular changes in $\delta^{18}\text{O}_{\text{ppt}}$ are

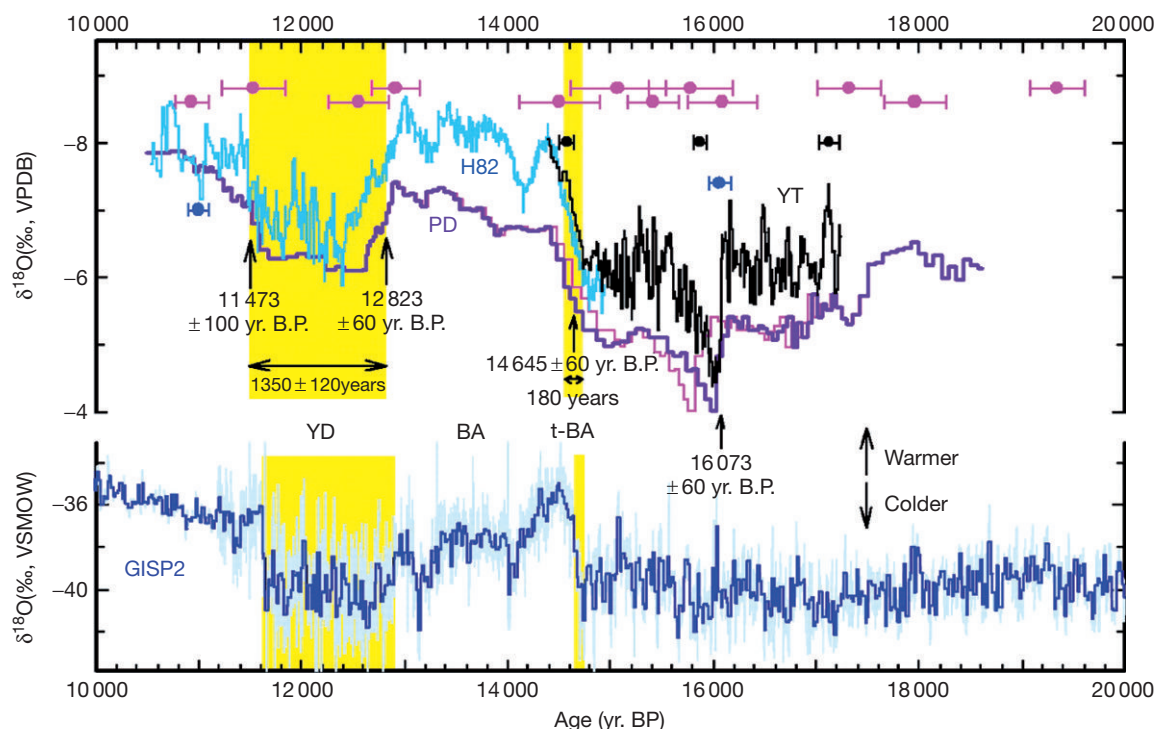


Figure 6 $\delta^{18}\text{O}$ of stalagmites from Hulu Cave, China (purple, black, and blue) and Greenland Ice (dark blue, 20-year averages; gray, 3-year averages) vs. age. YD, Younger Dryas; BA, Bolling-Allerod; t-BA, transition into B. From Wang YJ, Cheng H, Edwards RL, An ZS, Wu JY, Shen CC, and Dorale JA (2001). A high-resolution absolute-dated late Pleistocene monsoon record from Hulu Cave, China. *Science* 294: 2345–2348. Reproduced with permission.

attributed to shifts in storm tracks over the site. Niggeman et al. (2003) showed a high-resolution $\delta^{18}\text{O}_{\text{ct}}$ record from a cave in south Germany, in which power spectrum analyses revealed the influence of solar cycles on $\delta^{18}\text{O}_{\text{ppt}}$.

The vast literature affords many other records of interest. For example, Bard et al. (2002) accounted for a drop in $\delta^{18}\text{O}_{\text{ct}}$ in an Italian speleothem approximately 170 ka BP as a result of the amount effect. This effect increased during a wet period identified by the occurrence of a sapropel layer in coeval sediments of the Mediterranean Sea (Figure 8).

Paleotemperature Records

Relatively few studies have attempted to interpret variations in $\delta^{18}\text{O}_{\text{ct}}$ as paleotemperatures, even where Hendy tests gave positive results. Dorale et al. (1992) and van Beynen et al. (2004) assumed that the secular T dependence of $\delta^{18}\text{O}_{\text{ppt}}$ was equal or similar to the seasonal value and used that relationship to derive temperatures for speleothems in Iowa and New York state (Figure 9). Although Dennis et al. (2001) analyzed fluid inclusions in speleothems, they did not use the data to calculate temperatures; Sereffiddin et al. (2005) did this for three stalagmites and showed that coeval stalagmites with differing $\delta^{18}\text{O}_{\text{ct}}$ also differed in δD of inclusions so that the calculated temperatures of deposition were the same. Mangini et al. (2005) calibrated the relationship between $\delta^{18}\text{O}_{\text{ct}}$ and T by correlating historically reconstructed average annual

temperature with $\delta^{18}\text{O}_{\text{ct}}$ values for speleothem SPA from an Alpine cave.

Carbon Isotopes

Values of $\delta^{13}\text{C}_{\text{ct}}$ directly reflect the isotopic composition of the drip water from which the speleothem grew because the temperature dependence of the carbon isotopic fractionation between calcite and DIC is relatively small. The $\delta^{13}\text{C}$ value of DIC depends critically on the sources of carbon: atmospheric CO_2 ($\delta^{13}\text{C} \sim -7\text{‰}$), carbonate derived from the bedrock ($\delta^{13}\text{C} \sim -5$ to 5‰), and oxidation of soil organic matter ($\delta^{13}\text{C} \sim -26$ to -9‰). Large variations in $\delta^{13}\text{C}_{\text{ct}}$ are observed on various timescales ranging from years (Frappier et al., 2002) to tens of kiloyears. Variations in $\delta^{13}\text{C}_{\text{ct}}$ (the $\delta^{13}\text{C}$ value of the CaCO_3) can be caused by two distinct effects, both of which are potentially influenced by climate change:

- Change in proportion of C_3 to C_4 plants growing above the cave: $\delta^{13}\text{C}$ values of C_4 plants are 17‰ higher than those of C_3 plants; most C_4 plants are grasses; shifts from forest to grassland can cause an increase in $\delta^{13}\text{C}_{\text{ct}}$ (Dorale et al., 1998).
- Change in the density of vegetation above the cave: Dissolution of limestone in the absence of soil cover occurs by reaction with atmospheric CO_2 ($\delta^{13}\text{C} \sim -7\text{‰}$), leading to higher $\delta^{13}\text{C}$ values of DIC in recharge water compared to that produced under C_3 - and even C_4 -based soils (Frumkin et al., 2000).

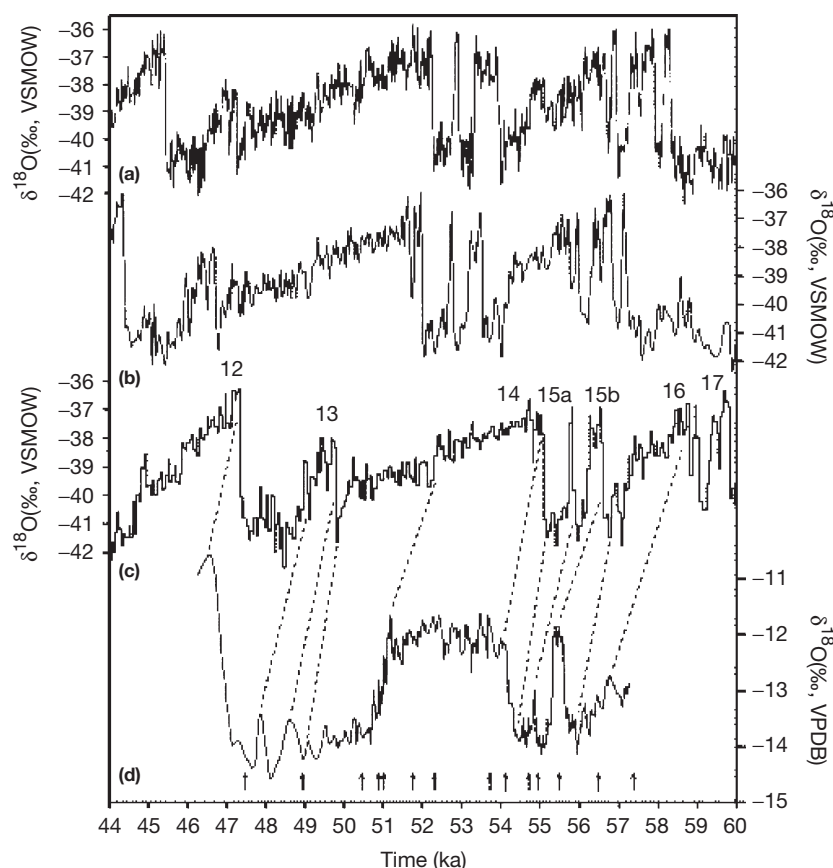


Figure 7 Comparison of oxygen isotope records of the Greenland ice cores and stalagmite SPA 49 from Kleegruben Cave, Austrian Alps: (a) GISP2 record; (b) GRIP, 1995 time scale; (c) GRIP, 2001 chronology; and (d) stalagmite SPA 49. Numbers refer to Greenland Interstadials. Upward arrows indicate positions of U-series samples. Dashed lines indicate major marker events used for cross-correlation. From Spötl C, and Mangini A (2002) Stalagmite from the Austrian Alps reveals Dansgaard–Oeschger events during isotope stage 3; Implications for the absolute chronology of Greenland ice cores. *Earth and Planetary Science Letters* 203: 507–518. Reproduced with permission.

In addition, the extent of exchange of the recharge water with bedrock (limestone) may affect $\delta^{13}\text{C}$ of DIC of the emerging drip waters, depending on whether or not continued air exchange is possible in the passages (fissures and joints) through which the water flows (Hendy, 1971). Finally, during progressive outgassing of CO_2 into the cave atmosphere, both equilibrium and kinetic effects can affect the $\delta^{13}\text{C}$ of the precipitated calcite (Mickler et al., 2004). These effects might be detectable as gradients in $\delta^{13}\text{C}_{\text{ct}}$ on growth layers.

Some studies have reported correlated shifts in $\delta^{18}\text{O}_{\text{ct}}$ and $\delta^{13}\text{C}_{\text{ct}}$ reflecting coupled changes in vegetation (e.g., C_3/C_4 ratios) and temperature (as reflected in $\delta^{18}\text{O}_{\text{ct}}$), whereas other studies show distinctly independent trends for each isotopic signal. Frumkin et al. (2000) showed that the trajectory of $\delta^{18}\text{O}_{\text{ct}}$ and $\delta^{13}\text{C}_{\text{ct}}$ defines simultaneous coupled changes in rainfall and temperature (Figure 10).

Future Directions and Problems

Instrumental advances in the analysis of speleothems may allow further paleoclimatic insights. Higher resolution

sampling of $\delta^{18}\text{O}_{\text{ct}}$ (and $\delta^{13}\text{C}_{\text{ct}}$) may be obtained using lasers or micromilling methods. Analyses of fluid inclusions for $\delta^{18}\text{O}$ and δD may provide better paleotemperatures, but these kinds of studies are challenged by problems in extraction and analysis of sub-microliter volumes of water. There may also be isotopic heterogeneity within the population of inclusions (Schwarcz and Yonge, 1983), partly revealed by infrared analysis (Fourier transform infrared spectroscopy) of calcite (unpublished data).

Future studies must take account of two concerns—one ethical and the other practical. Speleothems are admired for their natural beauty and should be disturbed as little as possible. Researchers must collect samples in a way that does not alter the appearance of the cave, either by using already broken stalagmites or working in caves that are in risk of destruction by erosion or human exploitation of the landscape. Second, we must be aware that, with the exception of actively growing speleothems, it is currently impossible to estimate the age of a speleothem *a priori*. Therefore, in general we cannot target a specific pre-Holocene period for study. As speleothem studies progress, previously analyzed, dated speleothems should be carefully archived for possible future research.

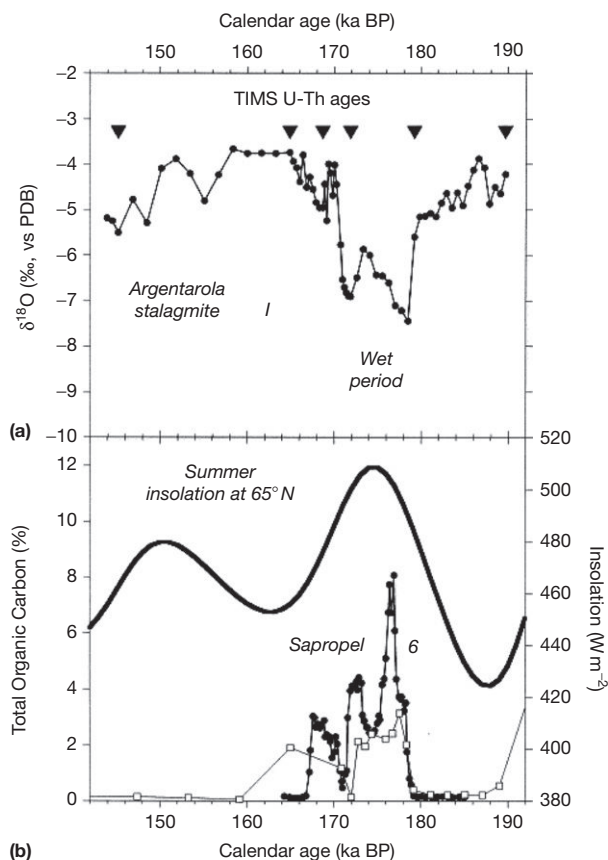


Figure 8 (a) $\delta^{18}\text{O}$ vs age for stalagmite from Argentarola Cave, Italy. Triangles: U-series dated points. (b) Upper curve: summer insolation at 65°N; lower curves: wt% total organic carbon in two nearby marine cores. From Bard, E., Delaygue, G., Rostek, F., Antonioli, F., Silenzi, S., and Schrag, D. P. (2002). Hydrological conditions over the western Mediterranean basin during the deposition of the cold sapropel 6 (ca. 175 ka BP). *Earth and Planetary Science Letters* 202: 481–494. Reproduced with permission.

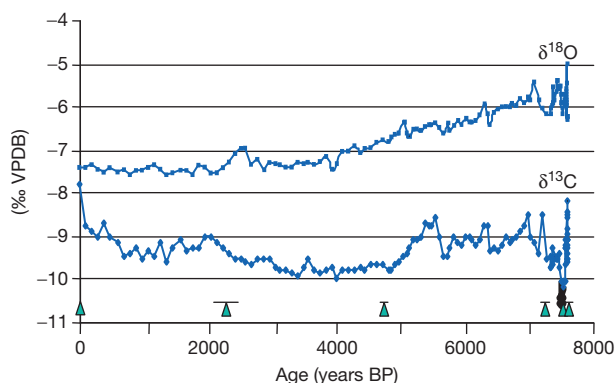


Figure 9 $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ record for stalagmite MF 1, McFail's Cave, New York state; gradient in $\delta^{18}\text{O}$ following hypsithermal at 7500 BP represents temperature decrease of ca. 5 °C. From van Beynen PE, Schwarcz HP, Ford DC, and Morrison J (2004) Holocene climatic variability measured by speleothems in McFail's Cave, New York. *Journal of Karst and Cave Studies* 66: 20–27. Reproduced with permission.

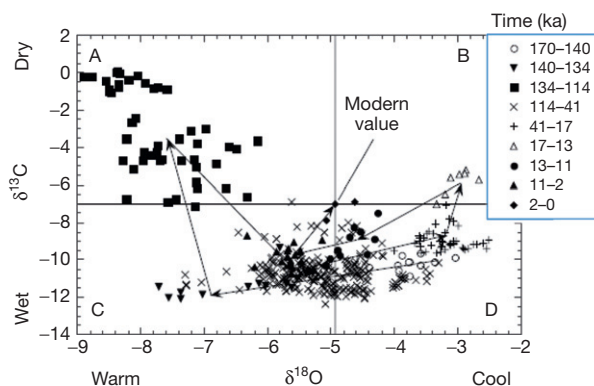


Figure 10 Isotopic and age data for stalagmite AF12 from Jerusalem, Israel, showing the relationship with climate. $\delta^{13}\text{C}$ increases with decreasing rainfall due to a decrease in the contribution of soil HCO_3^- to drip water. $\delta^{18}\text{O}$ increases with decreasing temperature (Frumkin et al., 1999), inferred from correlation with marine record. From Frumkin A, Ford DC, and Schwarcz HP (2000). Paleoclimate and vegetation of the last glacial cycles in Jerusalem from a speleothem record. *Global Biogeochemical Cycles* 14: 863–870. Reproduced with permission.

See also: U-Series Dating. Carbonate Stable Isotopes: Non-Lacustrine Terrestrial Studies. Ice Core Records: Correlations Between Greenland and Antarctica. Radiocarbon Dating: Calibration of the ^{14}C Record.

References

- Bar-Matthews M and Ayalon A (2003) Sea–land oxygen isotopic relationships from planktonic Foraminifera and speleothems in the eastern Mediterranean region and their implication for paleorainfall during interglacial intervals. *Geochimica et Cosmochimica Acta* 67: 3181–3199.
- Bard E, Delaygue G, Rostek F, Antonioli F, Silenzi S, and Schrag DP (2002) Hydrological conditions over the western Mediterranean basin during the deposition of the cold sapropel 6 (ca. 175 kyr BP). *Earth and Planetary Science Letters* 202: 481–494.
- Beck JW, Richards DA, Edwards RL, et al. (2001) Extremely large variations of atmospheric ^{14}C concentration during the last glacial period. *Science* 292: 2453–2458.
- Bischoff J and Fitzpatrick J (1991) U-series dating of impure carbonates: An isochron technique using total-sample dissolution. *Geochimica et Cosmochimica Acta* 55: 543–554.
- Burns SJ, Matter A, Frank N, and Mangini A (1998) Speleothem-based paleoclimate record from northern Oman. *Geology* 26: 499–502.
- Dennis PF, Rowe PJ, and Atkinson TC (2001) The recovery and isotopic measurement of water from fluid inclusions in speleothems. *Geochimica et Cosmochimica Acta* 65: 871–884.
- Denniston RF, Gonzalez LA, Asmerom Y, Baker RG, Reagan MK, and Bettis EA III (1999) Evidence for increased cool season moisture during the middle Holocene. *Geology* 27: 815–818.
- Dorale JA, Gonzalez LA, Reagan MK, Pickett DA, Murrell MT, and Baker RG (1992) A high-resolution record of Holocene climate change in speleothem calcite from Cold Water Cave, northeast Iowa. *Science* 258: 1626–1630.
- Dorale JA, Edwards RL, Ito E, and Gonzalez LA (1998) Climate and vegetation history of the midcontinent from 75 to 25 ka: A speleothem record from Crevice Cave, Missouri, USA. *Science* 282: 1871–1874.
- Edwards RL, Gallup CD, and Cheng H (2003) Uranium-series dating of marine and lacustrine carbonates. *Reviews in Mineralogy and Geochemistry* 52: 363–405.
- Epstein S, Buchsbaum R, Lowenstam HA, and Urey HC (1953) Revised carbonate–water isotopic temperature scale. *Bull. Geol. Soc. Amer* 64: 1315–1326.
- Fleitmann D, Burns SJ, Neff U, Mudelsee M, and Mangini A (2004) Palaeoclimatic interpretation of high-resolution oxygen isotope profiles derived from annually

- laminated speleothems from southern Oman. *Quaternary Science Reviews* 23: 935–945.
- Ford DC and Williams PW (1989) *Karst Geomorphology and Hydrology*. London: Unwin Hyman.
- Frappier A, Sahagian D, Gonzalez LA, and Carpenter SJ (2002) El Niño events recorded by stalagmite carbon isotopes. *Science* 298: 565.
- Frumkin A, Ford DC, and Schwarcz HP (2000) Paleoclimate and vegetation of the last glacial cycles in Jerusalem from a speleothem record. *Global Biogeochemical Cycles* 14: 863–870.
- Frumkin A, Ford DC, and Schwarcz HP (1999) Continental paleoclimatic record of the last 170,000 years in Jerusalem. *Quaternary Research* 51: 317–327.
- Genty D, Baker A, Massault M, et al. (2001) Dead carbon in stalagmites: Carbonate bedrock paleodissolution vs. ageing of soil organic matter. Implications for ^{13}C variations in speleothems. *Geochimica et Cosmochimica Acta* 65: 3443–3457.
- Hendy CH (1971) The isotopic geochemistry of speleothems—The calculation of the effects of different modes of formation on the isotopic composition of speleothems and their applicability as palaeoclimatic indicators. *Geochimica et Cosmochimica Acta* 35: 801–824.
- Mangini A, Spötl C, and Verdes P (2005) Reconstruction of temperature in the Central Alps during the past 2000 yr from a $\delta^{18}\text{O}$ stalagmite record. *Earth and Planetary Science Letters* 235: 741–751.
- McDermott FC, Mathey DP, and Hawkesworth CJ (2001) Holocene climate variation insights from an exceptionally high-resolution speleothem $\delta^{18}\text{O}$ record from SW Ireland. *Science* 294: 1328–1331.
- McDermott F, Schwarcz HP, and Rowe P (2006) Isotopes in speleothems. In: Leng M (ed.) *Developments in Paleoenviromental Research*. Vol. 10. *Isotopes in Paleoenviromental Research*, pp. 185–226. Dordrecht: Kluwer.
- McGarry S, Bar-Matthews M, Matthews A, Vaks A, Schilman B, and Ayalon A (2004) Constraints on hydrological and paleotemperature variations in the eastern Mediterranean region in the last 140 ka given by the δD values of speleothem fluid inclusions. *Quaternary Science Reviews* 23: 919–934.
- Mickler PJ, Banner JL, Stern L, Asmerom Y, Edwards RL, and Ito E (2004) Stable isotope variations in modern tropical speleothems; Evaluating equilibrium vs. kinetic. *Geochimica et Cosmochimica Acta* 68: 4381–4393.
- Niggemann S, Mangini A, Mudelsee M, Richter DK, and Wurth G (2003) Sub-Milankovitch climatic cycles in Holocene stalagmites from Sauerland, Germany. *Earth and Planetary Science Letters* 216: 539–547.
- Richards DA and Dorale JA (2003) Uranium-series chronology and environmental applications of speleothems. *Reviews in Mineralogy and Geochemistry* 52: 407–460.
- Richards DA, Bottrell S, Cliff RA, Stroehle K, and Rowe PJ (1998) U-Pb dating of a speleothem of Quaternary age. *Geochimica et Cosmochimica Acta* 62: 3683–3688.
- Rozanski K, Araguás-Araguás L, and Gonfiantini R (1993) Isotopic patterns in modern global precipitation. In: Swart PK, et al., (eds.) *Climate Change in Continental Isotopic Records*, pp. 1–36. Washington, DC: American Geophysical Union. Geophysical Monograph No. 78.
- Schwarcz HP and Yonge C (1983) Isotopic composition of paleowaters as inferred from speleothem and its fluid inclusions. In: Gonfiantini R (ed.) *Paleoclimates and Paleowaters: A Collection of Environmental Isotope Studies*, pp. 115–133. Vienna, Austria: International Atomic Energy Agency. Proceedings of Advisory Group Meeting STI/PUB/621.
- Schwarcz HP, Harmon RS, Thompson P, and Ford DC (1976) Stable isotope studies of fluid inclusions in speleothems and their paleoclimate significance. *Geochimica et Cosmochimica Acta* 40: 657–665.
- Serefidin F, Schwarcz HP, Ford DC, and Baldwin S (2003) Late Pleistocene paleoclimate in the Black Hills of South Dakota from oxygen isotope records in speleothems. *Palaeogeography, Palaeoclimatology, Palaeoecology* 203: 1–17.
- Serefidin F, Schwarcz H, and Ford D (2005) Use of hydrogen isotope variations in speleothem fluid inclusions as an independent measure of paleoclimate. In: Mora G and Surge D (eds.) *Isotopic and Elemental Tracers of Cenozoic Climate Change*, *Geological Society of America* 395: pp. 43–53 Special Publication.
- Spötl C and Mangini A (2002) Stalagmite from the Austrian Alps reveals Dansgaard-Oeschger events during isotope stage 3; Implications for the absolute chronology of Greenland ice cores. *Earth and Planetary Science Letters* 203: 507–518.
- Spötl C, Fairchild IJ, and Tooth A (2005) Cave air control on dripwater geochemistry, Obir Caves (Austria): Implications for speleothem deposition in dynamically ventilated caves. *Geochimica et Cosmochimica Acta* 69: 2451–2468.
- van Beynen PE, Schwarcz HP, Ford DC, and Morrison J (2004) Holocene climatic variability measured by speleothems in McFail's Cave, New York. *Journal of Karst and Cave Studies* 66: 20–27.
- Wang YJ, Cheng H, Edwards RL, et al. (2001) A high-resolution absolute-dated late Pleistocene monsoon record from Hulu Cave, China. *Science* 294: 2345–2348.
- Winograd IJ, Coplen TB, Landwehr JM, et al. (1992) Continuous 500,000-year climate record from vein calcite in Devils Hole, Nevada. *Science* 258: 255–260.
- Winograd I, Landwehr J, Ludwig KR, Coplen T, and Riggs AC (1997) Duration and structure of the past four glaciations. *Quaternary Research* 48: 141–154.
- Yonge C, Ford DC, Gray J, and Schwarcz HP (1985) Stable isotope studies of cave seepage water. *Isotope Geoscience* 58: 97–105.

Terrestrial Teeth and Bones

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Introduction

Since the 1970s, stable isotopes in terrestrial teeth and bones have been providing paleoenvironmental and paleobiological information on the Quaternary period. Indeed, the skeletal tissues of an animal through its diet and drinking water yield paleoenvironmental information in the form of isotopic ratios. This information is preserved in molecules and minerals found in fossil teeth and bones.

The most commonly used pairs of isotopes are $^{13}\text{C}/^{12}\text{C}$, $^{15}\text{N}/^{14}\text{N}$, and $^{18}\text{O}/^{16}\text{O}$. Isotopes correspond to different types of atoms for a given chemical element, meaning that two isotopes of the same element have the same number of protons, but differ in the number of neutrons. For instance, an atom of ^{12}C (carbon-12) has six protons and six neutrons in its core, while an atom of ^{13}C (carbon-13) has six protons and seven neutrons. Therefore, ^{13}C exhibits a higher atomic weight than ^{12}C and is thus referred to as the heavy isotope (contrary to ^{12}C , which is referred to as the light isotope). Nevertheless, both isotopes enter in the same molecules with the same types of chemical bonds. The relative proportion of ^{13}C and ^{12}C in a given molecule is not random but depends on the relative content of both isotopes in the source of carbon used to synthesize this molecule and on the differences of behavior between both isotopes during chemical reactions.

The relative abundances of the different isotopes are much higher for light isotopes than for heavy isotopes (e.g., Koch, 2007). The variations of isotopic abundances in natural samples are very small. They are measured using an isotopic ratio mass spectrometer, which separates and quantifies the number of heavy and light isotopes of a given element. In order to ensure accuracy, measurements are performed simultaneously on the test sample and on a standard, which allows corrections that make all results comparable to one another. Due to this measurement strategy, the results are expressed as relative abundances, known as 'delta' (δ) values, which are defined as follows:

$$\delta^{\text{E}}\text{X} = \frac{^{\text{E}}\text{R}_{\text{sample}} - ^{\text{E}}\text{R}_{\text{sample}}}{^{\text{E}}\text{R}_{\text{sample}}} \times 1000\%$$

where $^{\text{E}}\text{X}$ stands for ^{13}C , ^{15}N , or ^{18}O , respectively. International reference standards have been established for $\delta^{13}\text{C}$ values (marine carbon Pee Dee Belemnite (PDB)), for $\delta^{15}\text{N}$ values (atmospheric dinitrogen), and for $\delta^{18}\text{O}$ values (an average mixture of oceanic waters called Standard Mean Ocean Water (SMOW)).

Paleobiological Tracking by Isotopes in Teeth and Bones

Bones and teeth are composed of both organic and mineral fractions which are synthesized during the lifetime of a vertebrate (Figure 1). The organic fraction of bone and dentine is

mainly formed of a protein, collagen, which contains carbon (around 40%) and nitrogen (around 15%), while the mineral fraction contains carbon and oxygen. Enamel is formed of a very small organic fraction devoid of collagen. In bone, dentine, and enamel, the mineral fraction is composed of a calcium phosphate (apatite) with many impurities, including 3–5% carbonate. The crystal size of apatite is much larger in enamel than in bone and dentine. The isotopic signatures of nitrogen are measured in the organic fraction, while oxygen isotopic signatures are measured in the mineral fraction, both in phosphate and carbonate. The isotopic signatures of carbon can be measured in the organic and the mineral phases of bone and dentine, and in the mineral fraction of enamel.

Carbon

All the carbon of an organism comes from its dietary intake, in the form of proteins, carbohydrates, and lipids. Some of these nutrients are incorporated directly by the organism and sequestered in different tissues, while other molecules are synthesized by the organism from dietary nutrients. Due to these different biochemical characteristics and isotopic fractionations, the average carbon isotopic abundance of a vertebrate is close to that of its average diet, but those recorded in a given tissue or molecule exhibit specific differences (Figure 2). The $\delta^{13}\text{C}$ values of collagen are typically 5‰ more positive than those of the average diet, while the $\delta^{13}\text{C}$ values of the carbonate fraction of bone and tooth mineral fraction is 9–14‰ more positive than that of the average diet (Cerling and Harris, 1999; Passey et al., 2005). Therefore, the $\delta^{13}\text{C}$ values of collagen and carbonate apatite are typically used to track the type of plant food consumed by herbivores and the type of plants at the base of the food web to which predators belong.

The $\delta^{13}\text{C}$ values discriminate between different types of plants, principally between marine and terrestrial plants. Among terrestrial plants, the carbon isotope signature varies between plants using the two main photosynthetic pathways, called C_3 and C_4 (' C_3 plants' and ' C_4 plants'). C_4 plants are absent or very limited in environments with a mild or cold growing season, as in Europe, Northern latitudes, and high altitudes (Ehleringer et al., 1997). When present, C_4 plants are grasses and forbs. In environments where all plants use the C_3 photosynthetic pathway, an isotopic distinction can be seen between plants growing under a closed canopy and those at the top of the canopy or growing in an open environment. The possible causes of the so-called canopy effect are the concentration of recycled CO_2 due to poor ventilation, the light attenuation, and the relative high water availability in closed canopy forest (e.g., Broadmeadow et al., 1992; Gebauer and Schultze, 1991; van der Merwe and Medina, 1991). As herbivore teeth and bones record the $\delta^{13}\text{C}$ values of their plant food, it is possible to identify which kind of plant was consumed by an herbivore, and therefore the type of environment in which it lived (Figure 3).

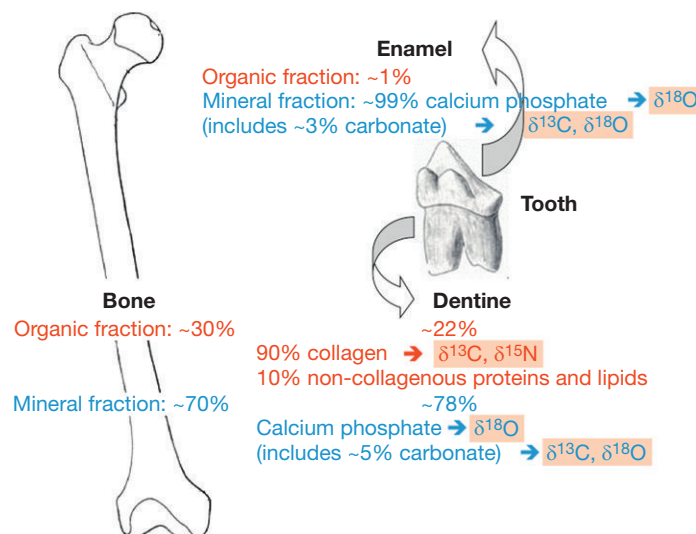


Figure 1 Summary of the composition of bones and teeth, with emphasis on the isotopic signatures that can be retrieved from bone, dentine, and enamel.

	Inorganic precursors	Producers (Plants)	Consumers (body average)	Consumers (bone/tooth)
Carbon ($\delta^{13}\text{C}$)	CO_2 (~ -8‰) HCO_3^- (~ 0‰)	C_3 Plants (~ -28‰) C_4 Plants (~ -13‰) Marine Plants (~ -20‰)	+ 0–1‰	+5‰ in collagen +9 to +14‰ in carbonate apatite
Nitrogen ($\delta^{15}\text{N}$)	NO_2 NO_3 NH_4 (variable) N_2 (~ 0‰)	Non N_2 fixing Plants (~ -5 to +5‰) N_2 fixing Plants (~ 0‰)	+3 to +5‰	Collagen
Oxygen ($\delta^{18}\text{O}$)	H_2O (variable according to environment) O_2	(variable according to taxon) H_2O in leaves O in nutrients O_2	H_2O CO_2	+18‰ in PO_4 +26‰ in CO_3

Figure 2 Summary of the isotopic fractionation factors associated with the transfer of carbon, nitrogen, and oxygen within food webs and the hydrological cycle, with emphasis on bone and tooth tissues. Bold arrows indicate steps where significant isotopic fractionation occurs.

Nitrogen

Nitrogen is incorporated through dietary intake in the organic molecules of an organism's tissues. It is usually measured in the collagen preserved in fossil bones. Contrarily to carbon, the isotopic signature of nitrogen is significantly enriched in vertebrate tissues relative to its average diet, typically by 3–5‰ (Figure 2; review in Robbins et al., 2005). Therefore, the nitrogen isotopic signature of a given individual depends on the isotopic signature at the base of the food web to which it belongs (in the plants), and on the position of the specimen within the food web (herbivore or predator). Not all herbivores present the same $\delta^{15}\text{N}$ values in a given ecosystem because different plants may use nitrogen under different forms, leading to varying isotopic fractionation and $\delta^{15}\text{N}$ values. For

instance, grass and graminoids typically exhibit more positive $\delta^{15}\text{N}$ values than shrubs and trees, as the latter obtain their nitrogen through symbiotic fungal mycorrhizae (e.g., Michelsen et al., 1998; Schulze et al., 1994). Global climatic factors such as aridity and temperature lead to increased $\delta^{15}\text{N}$ values of plants (Amundson et al., 2003), while $\delta^{15}\text{N}$ of plants tends to decrease with increasing altitude (e.g., Männel et al., 2007). Fire also tends to increase $\delta^{15}\text{N}$ values of plants growing after such an event (Högberg, 1997).

Oxygen

Oxygen isotopic signatures are measured in the mineral phase of enamel, dentine, or bone, in the phosphate or the carbonate

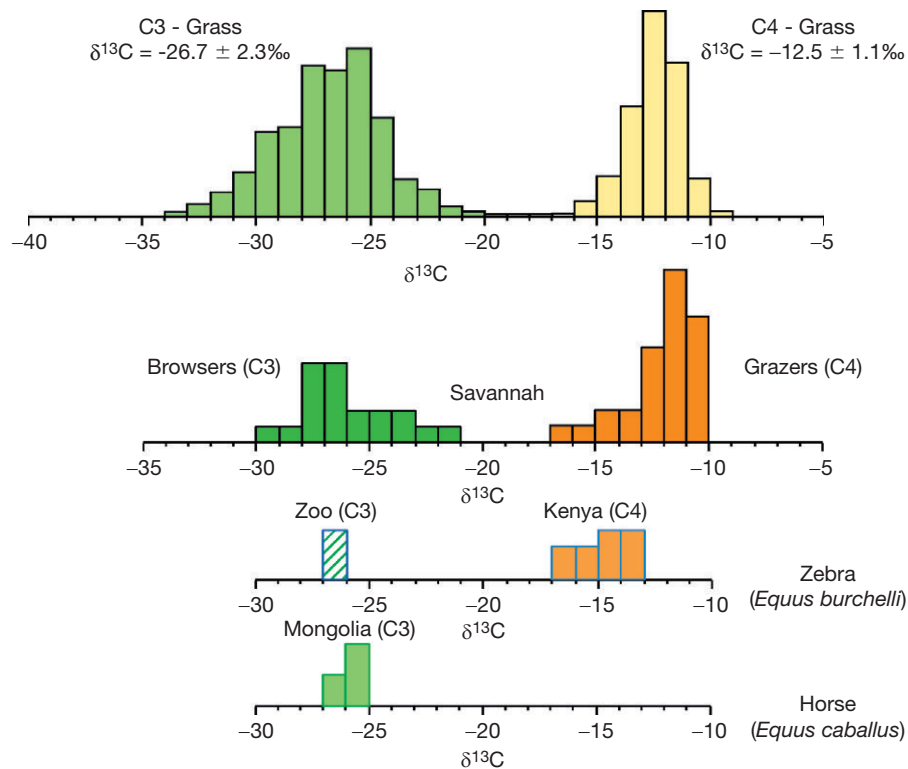


Figure 3 An example of the reconstruction of the carbon isotopic composition of the diet consumed by modern equids (horses and allies) based on their habitat. The $\delta^{13}\text{C}$ of diet has been calculated by adding 14‰ to the $\delta^{13}\text{C}$ values measured in carbonate hydroxylapatite (data from Cerling and Harris, 1999). The reconstructed diet of zebras (*Equus burchelli*) living in a Kenyan savanna exhibit $\delta^{13}\text{C}$ values similar to those of C_4 grasses, while those living in a zoo and fed with C_3 plants track this different diet, and exhibit $\delta^{13}\text{C}$ values similar to those of horse (*Equus caballus*) eating C_3 plants in a Mongolian grassland.

fractions (Kohn and Cerling, 2002). For both fractions, the $\delta^{18}\text{O}$ values are related to those of environmental water, through drinking water. However, there are complications due to the incorporation of water from food, and mixture with oxygen from respiration (Figure 2). This leads to varying species-dependent relationships between $\delta^{18}\text{O}$ values in carbonate or phosphate and $\delta^{18}\text{O}$ values in environmental water taken by the organisms. However, both $\delta^{18}\text{O}$ values are linked by a clear relationship, independent of the organism.

Variation in oxygen isotope ratios studied from terrestrial environments is ascribed to environmental temperature changes, with warmer weather resulting in more positive $\delta^{18}\text{O}$ values and cooler weather resulting in more negative $\delta^{18}\text{O}$ values in meteoric waters. However, the pattern is different in warm environments, where the temperature is higher than 20 °C. In such contexts, the $\delta^{18}\text{O}$ values decrease when the amount of precipitation increases. Local parameters, such as evaporation in ponds or streams originating at high altitudes, may provide drinking water to terrestrial vertebrates with $\delta^{18}\text{O}$ values different from those of local precipitation, and thus complicate the interpretation of the oxygen isotopic composition of fossil teeth and bones.

Chronological Resolution of Skeletal Tissues

Different tissues record the isotopic signature of food or drinking water during the period of their synthesis. Typically, bone grows during the first stages of ontogeny, but in higher vertebrates such

as mammals and birds, it continues to remodel and therefore incorporates newer carbon, nitrogen, and oxygen atoms. An adult mammal thus dies with its bone isotopic signature reflecting the last years of its lifetime. In contrast to this, mammalian dental tissues such as dentine and enamel form during a limited period of an individual's lifetime, and once formed, do not remodel. In an adult mammal, teeth exhibit isotopic signatures corresponding to the early years of the individual, depending on the chronology of tooth development for a given species. Cross sections of dentine and enamel retain their isotopic signature in a chronological order reflecting time of assimilation (e.g., Drucker et al., 2010; Fraser et al., 2008). Tooth enamel may thus record seasonal variations of $\delta^{18}\text{O}$ values (Figure 4). The level of resolution in such tissues, especially tooth enamel, is still under debate as each volume of tissue forms over a significant amount of time (e.g., Kohn and Cerling, 2002).

Physiological Effects (Suckling, Hibernation, Starvation, and Water Stress)

Some events in the life of a vertebrate can complicate the record of dietary isotopic signatures. In mammals, diet changes dramatically between the nursing and adult stages, following weaning. Maternal milk exhibits isotopic signatures different from those of the average adult diet, especially with more positive $\delta^{15}\text{N}$ values and more positive $\delta^{18}\text{O}$ values (Wright and Schwarcz, 1998). Therefore, suckling mammals exhibit

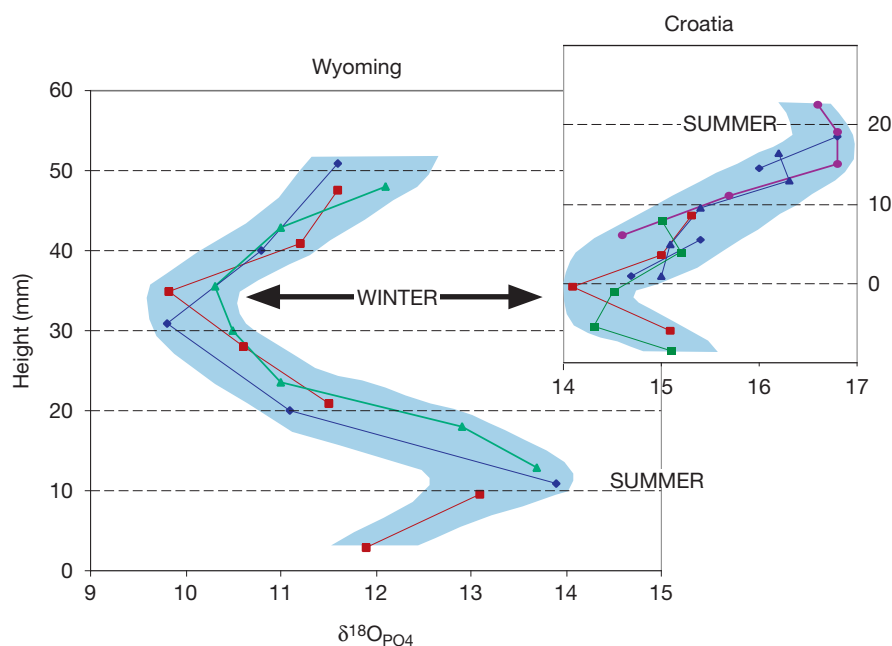


Figure 4 Variations of $\delta^{18}\text{O}$ in deer enamel reflect seasonal variations, in specimens from Wyoming and Croatia (based on Fricke et al., 1998). The differences in amplitudes and absolute values of $\delta^{18}\text{O}$ values are due to different climatic regimen in both areas.

isotopic differences relative to their adult counterparts, and tissues formed during this formative period of life, such as deciduous teeth and first molars, usually exhibit isotopic shifts relative to other teeth and bone tissues formed after weaning.

In hibernating mammals, such as bears, carbon and nitrogen isotopic signatures shift in blood during the winter sleep and these shifts are recorded in tooth dentine (Bocherens, 2004). Starvation and water stress is also a phenomenon that has been suggested to induce ^{15}N shifts in vertebrate tissue (Ambrose, 1991), but recent research on captive and wild mammals suggests that the changes in ^{15}N abundances coincident with harsh environments are most likely due to isotopic changes in the plants, while the fractionation between diet and animals remains relatively constant (e.g., Hartman, 2010; Murphy and Bowman, 2006).

Diagenetic Alteration

The chemical composition of bone and tooth is modified after an individual's death, through physical, chemical, and biochemical mechanisms called diagenesis. The intensity of diagenetic alteration depends on the time elapsed since death, but also on sedimentary conditions. High temperatures and humidity as well as high levels of microbial activity will increase the intensity of diagenetic transformation, increasing the chances of alteration of the biogenic isotopic signatures.

Organic Fraction

Collagen in fossil bones and dentine can survive for tens of thousands of years under favorable conditions. Cold and dry climatic conditions are more favorable than warm and humid ones, while cave deposits are more favorable than open-air sites. Collagen preservation in fossil skeletal remains can be

assessed by the nitrogen content of bulk sample as a proxy (e.g., Bocherens et al., 2005a).

Fossil collagen can be purified through several methods. The reliability of the isotopic signatures of ancient collagen is assessed through its chemical similarity to collagen from fresh bone. Fossil organic extracts with atomic C/N ratios lower than 2.9 or higher than 3.6 are considered unreliable, as well as those containing less than 5% nitrogen (e.g., Ambrose, 1990).

Mineral Fraction

While it is almost always possible to measure carbon and oxygen signatures from fossil bones and teeth, the key question is to determine whether the measured values correspond to those recorded in the organism's tissues when it was alive. Evaluating the extent of diagenetic alteration can be done either using the pattern of isotopic variations observed in the fossil samples as compared to equivalent modern ones, or using indirect tracers of modifications, such as crystallinity indexes and uptake of trace elements (e.g., Kohn and Cerling, 2002). It is commonly assumed that the isotopic signatures of enamel are much more stable through time than those of dentine and bone, while the phosphate fraction is more stable than the carbonate fraction. However, the phosphate fraction can be also affected under special circumstances. The best approach is to assess the extent of diagenetic alteration in each studied site using as many different proxies as possible (e.g., Kohn et al., 1999).

Reconstruction of Terrestrial Paleoenvironments

Tropical Environments

Paleoenvironments where C_3 and C_4 plants coexist are particularly favorable for isotopic tracking with carbon in teeth and

bones. Indeed, C_4 plants increase in proportion as climatic conditions become drier and warmer, and C_3 plants increase in proportion as moisture increases. As the difference between $\delta^{13}\text{C}$ values of these plants relative to C_3 plants is around 12‰, the $\delta^{13}\text{C}$ value recorded in the skeletal tissues of fossil vertebrates document these changes in the proportions of consumed plants. However, other factors such as the atmospheric pressure of CO_2 also play a role in this equation (Ehleringer et al., 1997; Koch et al., 2004). The main climatic parameter in these environments is the amount of annual rainfall, and nitrogen isotopic signatures of herbivores can be used to track past variations in aridity. Some examples are given in the following section.

Development of C_4 biomes during the late Quaternary

The open environments of the southern half of the North American continent are a place of competition between C_3 and C_4 herbaceous plants. Today, there is a clear gradient of decreasing proportions of C_4 plants toward the north, with about 10% of C_4 plants at 48° N. Carbon and oxygen isotopic studies of late Pleistocene herbivore tooth enamel from southwestern United States demonstrated the role of low atmospheric CO_2 in the expansion of C_4 plants during the Last Glacial Maximum, an expansion that could not be predicted based on temperature and precipitation changes alone (e.g., Koch et al., 2004).

Studies based on late Quaternary herbivore tooth enamel showed that the proportions of C_3 and C_4 plants in the Northern Cape Province, South Africa, did not reflect the predictions of climatic models about winter and summer rainfall regimes (Lee-Thorp and Beaumont, 1995) and documented episodes of wetter and drier conditions from 16 000 years ago (Smith et al., 2002).

Aridity

In tropical environments, aridity has been recognized as a factor influencing the $\delta^{15}\text{N}$ of herbivore collagen, with more positive $\delta^{15}\text{N}$ values under more arid conditions (e.g., Ambrose, 1991; Hartman, 2010; Murphy and Bowman, 2006; Schwarcz et al., 1999). This led to the possibility of quantifying past annual rainfall based on the $\delta^{15}\text{N}$ of herbivorous taxa, for instance, macropods (kangaroos) in Australia (Gröcke et al., 1997).

Decreasing humidity has an effect on bone because of increasing evapotranspiration in leaves, leading to $\delta^{18}\text{O}$ increase in leaf water, which is recorded in bone phosphate (e.g., Ayliffe and Chivas, 1990; Luz et al., 1990). In combination with high $\delta^{13}\text{C}$ values, the high $\delta^{18}\text{O}$ values exhibited by middle Pleistocene representatives of marsupial species that became extinct in the late Pleistocene demonstrate that these fauna could survive arid episodes, which weakens the hypothesis that increasing aridity may have led to their final extinction around 50 000 years ago (Prideaux et al., 2007).

Ecological Flexibility of Large Mammals

Isotopic tracking of habitat through the plants consumed by ancient herbivores is a taxon-free approach, and therefore can be used to document possible diet and habitat change of

some species in relationship with climate change or anthropogenic pressure. For instance, some large herbivores from the Pleistocene in Florida exhibit diet and presumably habitat stability between glacial and interglacial periods, such as tapir and mastodon that remain browsers (C_3 diet), while horses remain grazers (C_4 diet). In contrast, other species such as white-tailed deer exhibit less negative $\delta^{13}\text{C}$ values during the interglacial period, but still within the C_3 diet values, while extinct camel *Hemiauchenia* and extinct peccary *Platygonus* exhibit a clear shift from C_3 to C_4 diet between glacial and interglacial periods (DeSantis et al., 2009; Figure 5). By comparing the $\delta^{13}\text{C}$ values of modern mammals in southeastern Asia with those of the same species or related taxa in a middle Pleistocene glacial site in Thailand, it was possible to establish that some species were real forest dwellers, such as Java rhinoceros and orangutan, while others that are nowadays restricted to forest environments were dwelling in C_4 savanna, such as the small bovid *Capricornis*, a likely consequence of increasing anthropogenic pressure (Pushkina et al., 2010; Figure 5). These examples illustrate how isotopic tracking can be used to document possible habitat changes, even for species with modern relatives living in a restricted environment or with a morphology seemingly adapted to a given diet and habitat.

Temperate and Boreal Environments

In terrestrial environments, where all plants use the C_3 photosynthetic pathway, plant $\delta^{13}\text{C}$ values exhibit some differences related to environmental parameters (e.g., Heaton, 1999). In particular, plants growing under a closed canopy exhibit significantly more negative $\delta^{13}\text{C}$ values relative to those growing at the top of the canopy or in open environments. This carbon isotopic fractionation is transferred to herbivores consuming such plants in dense canopy forests. It is thus possible to track the changes of habitat for large herbivores at the beginning of the Holocene in Europe (Figure 6; e.g., Drucker et al., 2003a, 2008).

Variations in $\delta^{15}\text{N}$ values in ungulate bone collagen seem to relate to changes in soil microbial activity during the climatic oscillations that occurred in Europe since 30 000 years ago (Drucker et al., 2003a,b; Hedges et al., 2004; Richards and Hedges, 2003): low temperatures led to decreased microbial activity and thus reduced nitrogen isotopic fractionation that is transferred in plants consumed by herbivores. Indeed, a decrease in $\delta^{15}\text{N}$ values is recorded during the cold peaks of the Last Glacial Maximum and of the Younger Dryas, while an increase in $\delta^{15}\text{N}$ values is documented during the warming of the Bölling–Alleröd Interstadial and that of the early Holocene (Preboreal and Boreal). The role of climate fluctuations to explain these trends in $\delta^{15}\text{N}$ values was confirmed by the $\delta^{18}\text{O}$ values of the phosphate of the same bones that tracks temperature changes, and the observed correlation between increasing $\delta^{15}\text{N}$ and $\delta^{18}\text{O}$ values, and hence temperature, demonstrates the relationship between increasing $\delta^{15}\text{N}$ values in red deer bone collagen and warming in this context (Drucker et al., 2009). These nitrogen isotopic excursions present geographic variations according to the intensity of the temperature changes.

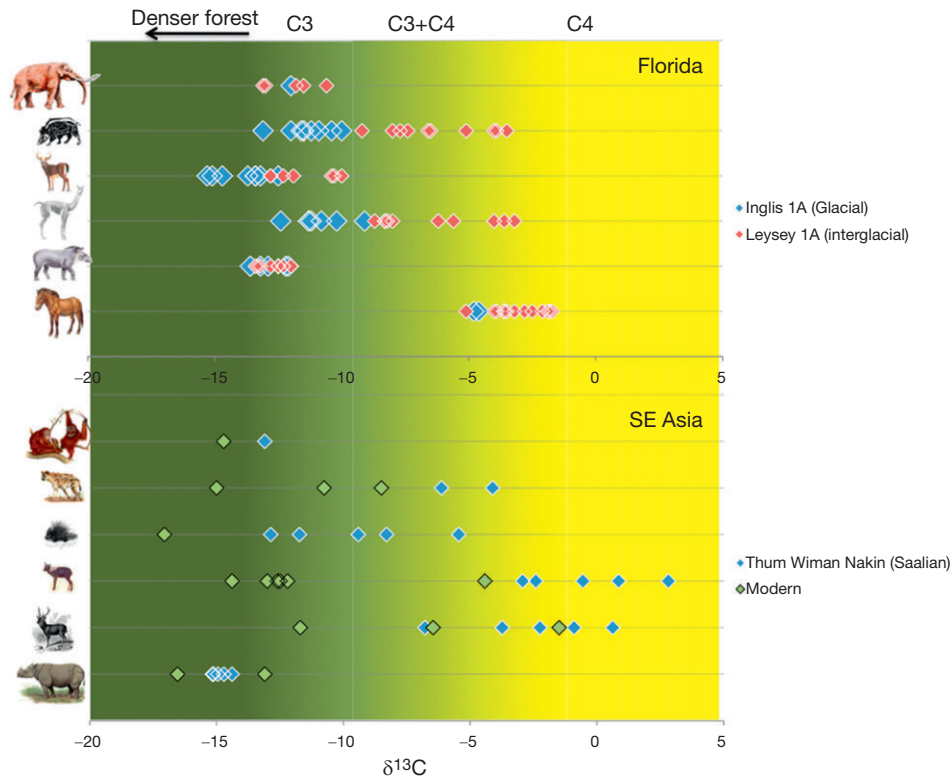


Figure 5 $\delta^{13}\text{C}$ values of tooth enamel illustrate stability or changes in diet between glacial and interglacial periods in Florida (top, data from DeSantis et al., 2009) and between Saalian (glacial) and modern (interglacial) periods in southeastern Asia (bottom, data from Pushkina et al., 2010).

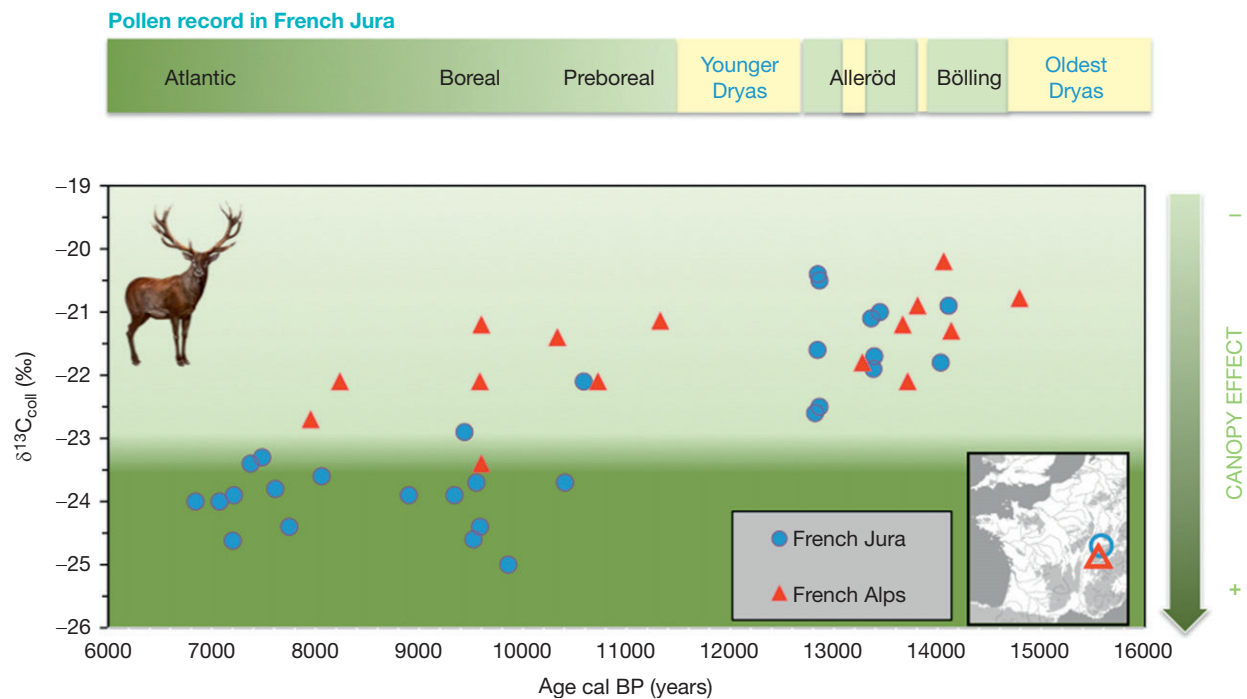


Figure 6 Evolution of collagen $\delta^{13}\text{C}$ values in red deer (*Cervus elaphus*) bones from French Jura and French Alps (data from Drucker et al., 2003a, 2008). The late Pleistocene samples exhibit $\delta^{13}\text{C}$ values indicative of open environments, while Holocene specimens present more negative $\delta^{13}\text{C}$ values in Jura, similar to those exhibited by modern large herbivores dwelling in a closed canopy forest. In the Alps, red deer remain in open environments, probably at higher altitudes.

Seasonality

Using carbon and oxygen isotopic variations in hypsodont tooth enamel, as in horse and bison, allowed the reconstruction of seasonal changes in consumed plant food and precipitation isotopic signatures, thus leading to a better understanding of past climatic regimen and events preceding the death of ancient animals (e.g., [Gadbury et al., 2000](#); [Higgins and MacFadden, 2004, 2009](#); [Kohn and Cerling, 2002](#); [Passey and Cerling, 2002](#)).

Reconstruction of the Paleobiology of Extinct Species

Using skeletal material from which collagen can be extracted, it is possible to reconstruct some aspects of animal diet using carbon and nitrogen isotopic signatures, because different dietary items exhibit different isotopic signatures, such as C_3 or C_4 plants, animal flesh, or food resources of marine origin. With material too old or too altered to yield collagen, the carbon isotopic signatures of tooth enamel can be used to test dietary hypotheses in contexts where different food webs start with plants exhibiting different $\delta^{13}C$ values, such as tropical environments where C_4 grasses coexist with C_3 tree leaves.

Subsistence Patterns of Ancient Hominids

Environmental and dietary changes are often linked to key stages of the evolution of hominids. In some cases, the possible impact of such changes can be tested using the isotopic signature of fossil bones and teeth.

The importance of savanna for African Pliocene–Pleistocene hominids

A hotly debated issue in the study of Quaternary terrestrial paleoenvironments is the role of savanna in the evolution of hominids in Africa. Indeed, modern African apes are restricted to forested environments, and savanna environments require special skills for exploitation by primates, such as a bipedal gait and strong social bonds. The $\delta^{13}C$ values of tooth enamel of fossil mammals, including hominids, from South Africa, ranging in age from 1.8 to 1.5 million years old, indicate a significant C_4 component in the environment, as well as significant C_4 dietary inputs for hominids of the species *Paranthropus robustus*, and *Homo ergaster* ([Figure 7](#); [Lee-Thorp et al., 2000, 2003, 2010](#)). This C_4 dietary input could be linked to the consumption of C_4 plant parts, but also to the consumption of animals feeding on C_4 plants, such as invertebrates, small vertebrates, or grazer meat. Based on other evidence such as tooth microwear and abundance of C_4 plants in the surrounding environment, it seems that the most likely explanation for this C_4 diet should include a mixture of underground plant storage organs and animals resources ([Lee-Thorp et al., 2010](#)). In East Africa, high $\delta^{13}C$ measured in some *Paranthropus boisei* seems to reflect the consumption of abundant C_4 sedges ([Figure 7](#); [Lee-Thorp et al., 2010](#)). In contrast, an older hominid from the Pliocene, *Ardipithecus ramidus*, did not incorporate C_4 resources in its diet although these resources were present in the environment ([Lee-Thorp et al., 2010](#)). This supports the view that hominid evolution is linked to savanna,

at least since about 3 Ma, but also contradicts the hypothesis that different hominids used savanna resources in differing amounts.

The Diet of Neanderthals and Early Anatomically Modern Humans

During the Upper Pleistocene, Europe was populated by a distinctive hominid form, the Neanderthals. The demise of this hominid around 30 000 years ago coincides with the appearance in Europe of anatomically modern humans, seemingly using a more sophisticated culture. Several Neanderthal fossils have yielded preserved collagen that could be analyzed for carbon- and nitrogen-stable isotopic compositions, and compared to those of Upper Paleolithic anatomically modern humans (e.g., [Bocherens et al., 1999, 2005b, in press](#); [Richards et al., 2000, 2001, 2008](#)). Comparing Neanderthals with predators such as hyenas suggests that Neanderthals consumed mainly large herbivore meat, including a large proportion of megaherbivores such as mammoth and woolly rhinoceros ([Bocherens et al., 2005b](#)). It appears also that Neanderthals living under different environmental conditions and at different periods had similar diets. Although the absolute isotopic nitrogen values of early modern humans are higher than those of the last Neanderthals ([Richards and Trinkaus, 2009](#)), possible variations in the $\delta^{15}N$ values of the whole terrestrial food webs at the same time suggest that, finally, the last Neanderthals had diets that did not differ significantly from those of early anatomically modern humans ([Drucker and Bocherens, 2004](#)). This supports the hypothesis of dietary competition between Neanderthals and anatomically modern humans ([Bocherens and Drucker, 2006](#)).

Paleodiet of Ancient Bears

Bears, as omnivorous carnivores, form an interesting group to study using the isotopic approach as their diet can be quite variable, and sometimes difficult to determine based on their morphological features. They are all the more interesting because they have an abundant fossil record, for instance, for cave bears, and cover a dietary spectrum similar to that of humans. Detecting changes in bear diet may yield direct information on the availability of food resources relevant for human diet. Moreover, the paleogenetics of ancient bears is intensively studied and combining genetic and paleodietary data through the evolution of bear lineages provides direct evidence about their evolutionary biology. Using carbon- and nitrogen-stable isotopic signatures, cave bears (*Ursus spelaeus*) were shown to be essentially vegetarian animals ([Bocherens et al., 1994, 2006, 2011a,b](#)), while the extinct giant short-faced bear (*Artodus simus*) from North America has been demonstrated to be a meat eater ([Figure 8](#); [Barnes et al., 2002](#)). Interestingly, the diet of ancient brown bears living at the same time as extinct bears from specialized species shifts when the dietary competition stops with the extinction of their competitor. For instance, brown bears from North America were vegetarian when they coexisted with the carnivorous short-faced bears, but they became much more carnivorous after the extinction of these meat-eating bears ([Barnes et al., 2002](#)).

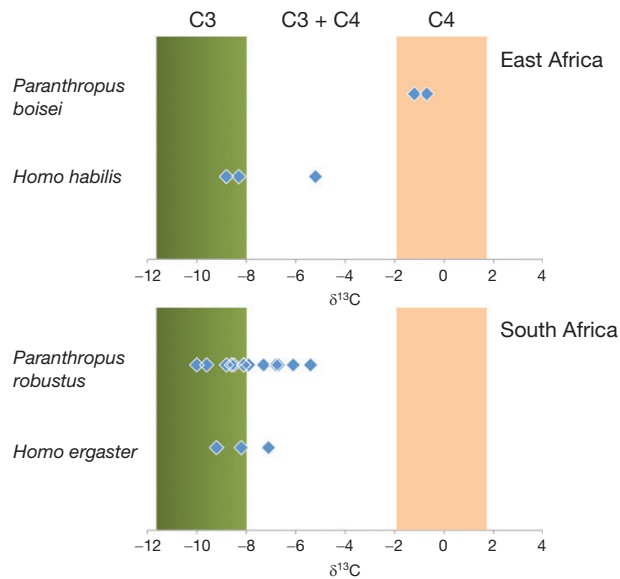


Figure 7 Reconstruction of the paleoenvironment and paleodiet of fossil hominids from South Africa (Swartkrans Members 1 and 2, ~1.5 to 1.8 Ma) and East Africa (Olduvai East and Pening, ~1.5 to 1.8 Ma). The $\delta^{13}\text{C}$ values of *Paranthropus robustus*, and *Homo ergaster* in South Africa show that a significant amount of food resources coming from C_4 environments (most likely savanna) were consumed by these three hominid species (data from Lee-Thorp et al., 2003). In contrast to *Paranthropus robustus* in South Africa, *Paranthropus boisei* from East Africa exhibits a diet almost completely composed of C_4 food resources, while *Homo habilis* had a diet with $\delta^{13}\text{C}$ values similar to *Homo ergaster* in South Africa.

Paleodiet of Ancient Ungulates

Large herbivorous mammals often exhibit specialized diets, in connection with their tooth morphology and digestive physiology. The occurrence of some herbivorous mammals in Quaternary fossil and archaeological localities can be used to infer the paleoenvironments around the site. Isotopic analyses of fossil herbivores have sometimes yielded additional information about the ancient environments in which they used to dwell, and thus allow more precise reconstructions of the actual diet of extinct species. For instance, isotopic investigations of tooth enamel have shown that in southern North America, horses were mixed feeders consuming C_4 grasses and C_3 shrubs, while bison were grazers, eating only C_4 grasses (Koch et al., 2004). Horses have hypsodont (high-crowned) teeth and are traditionally thought to be specialized grazers. Another example of hypsodont ungulates with diverse diets are extinct camelids from North America (Feranec, 2003). Therefore, the specialized morphology of some herbivore teeth is not always indicative of specialized diet, but rather indicates dietary flexibility.

Implications for Late Pleistocene Extinctions

The possibility to reconstruct dietary and/or habitat changes through time for a given species opens the possibility to document possible ecological disruption linked to extinctions,

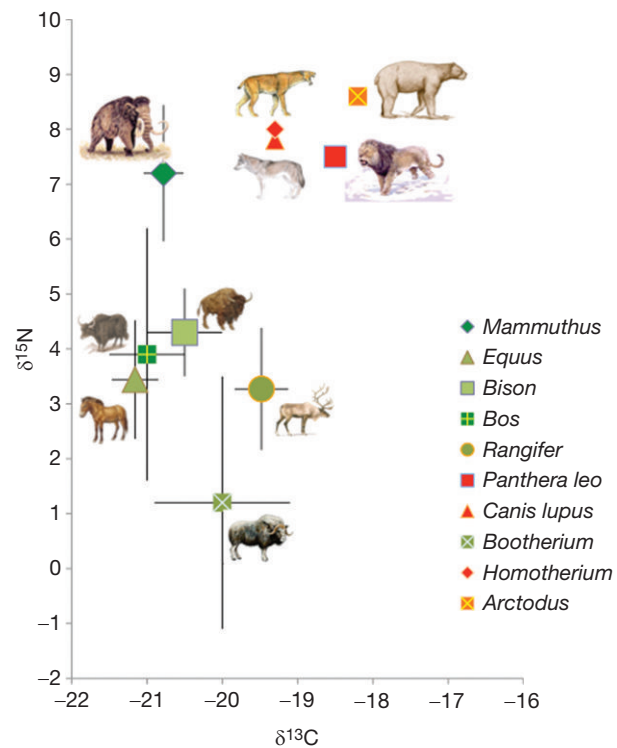


Figure 8 $\delta^{13}\text{C}$ and $\delta^{15}\text{N}$ values of short-faced bears (*Arctodus simus*) collagen from Alaska and Yukon, compared to those of coeval herbivorous and carnivorous species (data from Fox-Dobbs et al., 2008). These isotopic data of short-faced bears plot together with carnivorous species, such as lion, scimitar-toothed cat, and wolf, clearly indicating these that short-faced bears were carnivorous.

especially in the case of the late Pleistocene megafaunal extinctions. Two main explanations are usually given for these extinctions: climate change and human impact (e.g., Koch and Barnosky, 2006). Using the isotopic signatures of extinct species until the moment they become extinct and those of surviving species before, during, and after the extinction event may provide information about the factors that could have changed at this time and therefore help to test different hypotheses. In the case of Australia, the stable isotopic tracking of megafauna paleoecology has yielded very interesting results. For instance, aridity could be ruled out as a significant factor in the extinction of megafauna as the extinct species exhibited carbon and oxygen isotopic signatures that demonstrated that they did live under arid conditions well before the time of their extinction (Prideaux et al., 2007). If humans were involved in the megafaunal extinction in Australia, the question remains whether their action was mostly direct, through overhunting, or rather indirect, through environmental disruption, in increasing wildfire frequency for instance. Isotopic tracking, together with tooth microwear analysis, did provide some answers to this question. In the case of an extinct giant short-faced kangaroo *Procoptodon*, these data suggest a diet including C_4 chenopodiaceae from dry areas and obligate drinking (Prideaux et al., 2009). Aridity would therefore not be a problem for this species, but having to go to water holes would make this species vulnerable to human predation, in contrast with other kangaroos which survived the extinction

event, and which can sustain themselves with the water contained in their plant food and do not need to drink at water holes. An additional contribution of stable isotopic tracking in the debate on megafaunal extinctions in Australia was to demonstrate that species with dietary specialization were much more affected by a collapse of ecosystem and that a critical factor for surviving was the possibility to shift diet (Miller et al., 2005). A more widespread use of this approach to megafaunal extinction on other continents should also yield valuable information. For instance, the carbon- and nitrogen-stable isotopes of the last cave lions in Western Europe dated to around 12 000 years BP suggest that they have relied mainly on reindeer as their preferred prey (Bocherens et al., 2011a). The extirpation of this predatory species in this region could then correspond to the coeval extirpation of its main prey.

Perspectives

Stable isotopes in terrestrial teeth and bones have already yielded invaluable information on Quaternary ecosystems, and this approach is expected to develop further during the coming years.

A better knowledge of isotopic variations in modern ecosystems and modern animals will most probably allow more accurate paleoenvironmental and paleodietary reconstructions to be performed based on the stable isotopic signatures of teeth and bones of Quaternary terrestrial vertebrates. Our understanding of the fractionation factors in tissues prone to fossilization still needs improvement, based on controlled diet experiments of species closely related to the fossil taxa under study, and on investigations of large mammals in monitored wild contexts. Also the chronology of isotopic records in incrementally grown tissues, such as tooth enamel, needs more accurate quantification in the species found in Quaternary fossil assemblages.

The impact of diagenetic alteration on the isotopic signatures of teeth and bones needs to be better understood, especially when dealing with the mineral fraction of fossil vertebrates.

Collagen is currently used for its carbon and nitrogen isotopic signatures, but it could yield the isotopic signatures of additional chemical elements. For instance, the amount of deuterium, the stable heavy isotope of hydrogen, could be used to study paleoclimates and possible migrations, as it varies in a way similar to oxygen in relationship with temperatures (Bowen et al., 2005; Cormie et al., 1994). The isotopic signatures of sulfur in collagen may be used to improve paleodietary reconstructions and to provide identification of geographic origin (Richards et al., 2003).

Improving biomolecular technologies will allow the analysis of isotopic signatures of other organic compounds, such as osteocalcin, single amino acids, cholesterol, and fatty acids, in fossil bones and teeth (e.g., Corr et al., 2005; Smith et al., 2005, 2009). This is expected to lead to breakthroughs in retrieving paleobiological information at different timescales within a single individual, and in obtaining isotopic paleobiological signals in fossil material older than around 100 000 years.

See also: Vertebrate Overview. **Archaeological Records:** Human Evolution in the Quaternary; Neanderthal Demise. **Carbonate Stable Isotopes:** Lake Sediments; Nonmarine Biogenic Carbonates; Overview; Speleothems; Terrestrial Organic Materials. **Ice Core Records:** Antarctic Stable Isotopes; Greenland Stable Isotopes. **Vertebrate Records:** Late Pleistocene Megafaunal Extinctions; Late Pleistocene of Southeast Asia. **Vertebrate Studies:** Ancient DNA; Interactions with Hominids; Speciation and Evolutionary Trends in Quaternary Vertebrates.

References

- Ambrose SH (1990) Preparation and characterization of bone and tooth collagen for isotopic analysis. *Journal of Archaeological Science* 17: 431–451.
- Ambrose SH (1991) Effects of diet, climate and physiology on nitrogen isotope abundances in terrestrial foodwebs. *Journal of Archaeological Science* 18: 293–317.
- Amundson R, Austin AT, Scuur EAG, et al. (2003) Global patterns of the isotopic composition of soil and plant nitrogen. *Global Biogeochemical Cycles* 17(1): 1031. <http://dx.doi.org/10.1029/2002GB001903>.
- Ayliffe LK and Chivas AR (1990) Oxygen isotope composition of the bone phosphate of Australian kangaroos: Potential as a palaeoenvironmental recorder. *Geochimica et Cosmochimica Acta* 54: 2603–2609.
- Barnes I, Mattheus P, Shapiro B, Jensen D, and Cooper A (2002) Dynamics of Pleistocene population. Extinctions in Beringian brown bears. *Science* 295: 2267–2270.
- Bocherens H (2004) Cave bear palaeoecology and stable isotopes: checking the rules of the game. In: Philippe M, Argant A, and Argant J (eds.) *Proceedings of the 9th International Cave Bear Conference*, Cahiers scientifiques du Centre de Conservation et d'Etude des Collections (Muséum d'Histoire naturelle de Lyon) Hors Série No. 2, pp. 183–188.
- Bocherens H and Drucker DG (2006) Dietary competition between Neanderthals and modern humans: Insights from stable isotopes. In: Conard N (ed.) *When Neanderthals and Modern Humans Met*, pp. 129–143. Tübingen: Tübingen Publications in Prehistory, Kerns Verlag.
- Bocherens H, Billiou D, Patou-Mathis M, Otte M, Bonjean D, Toussaint M, and Mariotti A (1999) Palaeoenvironmental and palaeodietary implications of isotopic biogeochemistry of late interglacial Neanderthal and mammal bones in Sceldina Cave (Belgium). *Journal of Archaeological Science* 26: 599–607.
- Bocherens H, Drucker DG, Billiou D, Geneste J-M, and van der Plicht J (2006) Bears and Humans in Chauvet Cave (Vallon-Pont-d'Arc, Ardèche, France): Insights from stable isotopes and radiocarbon dating of bone collagen. *Journal of Human Evolution* 50: 370–376.
- Bocherens H, Drucker DG, Billiou D, and Moussa I (2005a) Une nouvelle approche pour évaluer l'état de conservation de l'os et du collagène pour les mesures isotopiques (datation au radiocarbone, isotopes stables du carbone et de l'azote). *L'Anthropologie* 109: 557–567.
- Bocherens H, Drucker DG, Billiou D, Patou-Mathis M, and Vandermeersch B (2005b) Isotopic evidence for diet and subsistence pattern of the Saint-Césaire I Neanderthal: Review and use of a multi-source mixing model. *Journal of Human Evolution* 49: 71–87.
- Bocherens H, Drucker DG, Bonjean D, et al. (2011a) Isotopic evidence for dietary ecology of cave lion (*Panthera spelaea*) in North-western Europe: Prey choice, competition and implications for extinction. *Quaternary International* 245: 249–261.
- Bocherens H, Fizet M, and Mariotti A (1994) Diet, physiology and ecology of fossil mammals as inferred by stable carbon and nitrogen isotopes biogeochemistry: Implications for Pleistocene bears. *Palaeogeography, Palaeoclimatology, Palaeoecology* 107: 213–225.
- Bocherens H, Germonpré M, Toussaint M, and Semal P (in press) Stable isotopes. In: Semal P and Hauzeur A (eds.) *Spy Cave: State of 120 years of Pluridisciplinary Research on the Bêche-aux-Rotches from Spy*.
- Bocherens H, Stiller M, Hobson KA, et al. (2011b) Niche partitioning between two sympatric genetically distinct cave bears (*Ursus spelaeus* and *Ursus ingressus*) and brown bear (*Ursus arctos*) from Austria: Isotopic evidence from fossil bones. *Quaternary International* 245: 238–248.
- Bowen GJ, Wassenaar LI, and Hobson KA (2005) Global application of stable hydrogen and oxygen isotopes to wildlife forensics. *Oecologia* 143: 337–348.

- Broadmeadow MSJ, Griffiths H, Maxwell C, and Borland AM (1992) The carbon isotope ratio of plant organic material reflects temporal and spatial variations in CO₂ within tropical forest formations in Trinidad. *Oecologia* 89: 435–441.
- Cerling TE and Harris JM (1999) Carbon isotope fractionation between diet and bioapatite in ungulate mammals and implications for ecological and paleoecological studies. *Oecologia* 120: 347–363.
- Cormie AB, Schwarcz HP, and Gray J (1994) Relation between hydrogen isotopic ratios of bone collagen and rain. *Geochimica et Cosmochimica Acta* 58: 377–391.
- Corr LT, Sealy JC, Horton MC, and Evershed RP (2005) A novel marine dietary indicator utilising compound-specific bone collagen amino acid $\delta^{13}\text{C}$ values of ancient humans. *Journal of Archaeological Science* 32: 321–330.
- DeSantis LR, Feranec RS, and MacFadden BJ (2009) Effects of global warming on ancient mammalian communities and their environments. *PLoS One* 4(6): e5750.
- Drucker DG and Bocherens H (2004) Carbon and nitrogen stable isotopes as tracers of diet breadth evolution during Middle and Upper Palaeolithic in Europe. *International Journal of Osteoarchaeology* 14: 162–177.
- Drucker DG, Bocherens H, and Billiou D (2003) Evidence for shifting environmental conditions in Southwestern France from 33 000 to 15 000 years ago derived from carbon-13 and nitrogen-15 natural abundances in collagen of large herbivores. *Earth and Planetary Science Letters* 216: 163–173.
- Drucker D, Bocherens H, Bridault A, and Billiou D (2003) Carbon and nitrogen isotopic composition of Red Deer (*Cervus elaphus*) collagen as a tool for tracking palaeoenvironmental change during Lateglacial and Early Holocene in Northern Jura (France). *Palaeogeography, Palaeoclimatology, Palaeoecology* 195: 375–388.
- Drucker DG, Bridault A, Hobson KA, Szuma E, and Bocherens H (2008) Can carbon-13 abundances in large herbivores track canopy effect in temperate and boreal ecosystems? Evidence from modern and ancient ungulates. *Palaeogeography, Palaeoclimatology, Palaeoecology* 266: 69–82.
- Drucker DG, Bridault A, Iacumin P, and Bocherens H (2009) Bone stable isotopic signatures (^{15}N , ^{18}O) as tracer of temperature variation in Lateglacial and early Holocene: Case study of red deer from Rochedane site in French Jura. *Geological Journal* 44: 593–604.
- Drucker DG, Hobson KA, Münzel SC, and Pike-Tay A (2010) Intra-individual variation in stable carbon ($\delta^{13}\text{C}$) and nitrogen ($\delta^{15}\text{N}$) isotopes in mandibles of modern caribou of Qamanirjuaq (*Rangifer tarandus groenlandicus*) and Banks Island (*Rangifer tarandus pearyi*): Implications for tracing seasonal and temporal changes in diet. *International Journal of Osteoarchaeology* 22: 494–504. <http://dx.doi.org/10.1002/oa.1220>.
- Ehleringer JR, Cerling TE, and Helliker BR (1997) C₄ photosynthesis, atmospheric CO₂, and climate. *Oecologia* 112: 285–299.
- Feranec RS (2003) Stable isotopes, hypsodonty, and the paleodiet of *Hemiauchenia* (Mammalia, Camelidae): A morphological specialization creating ecological generalization. *Paleobiology* 29: 230–242.
- Fox-Dobbs K, Leonard JA, and Koch PL (2008) Pleistocene megafauna from eastern Beringia: Paleoeological and paleoenvironmental interpretations of stable carbon and nitrogen isotope and radiocarbon records. *Palaeogeography, Palaeoclimatology, Palaeoecology* 261: 30–46.
- Fraser RA, Grün R, Privat K, and Gagan MK (2008) Stable-isotope microprofiling of wombat tooth enamel records seasonal changes in vegetation and environmental conditions in Eastern Australia. *Palaeogeography, Palaeoclimatology, Palaeoecology* 269: 66–77.
- Fricke HC, Clyde WC, and O'Neil JR (1998) Intra-tooth variations in $\delta^{18}\text{O}$ (PO₄) of mammalian tooth enamel as a record of seasonal variations in continental climate variables. *Geochimica et Cosmochimica Acta* 62: 1839–1950.
- Gadbury C, Todd L, Jahren AH, and Amundson R (2000) Spatial and temporal variations in the isotopic composition of bison tooth enamel from the Early Holocene Hudson-Meng Bone Bed, Nebraska. *Palaeogeography, Palaeoclimatology, Palaeoecology* 157: 79–93.
- Gebauer G and Schultze E-D (1991) Carbon and nitrogen isotope ratios in different compartments of a healthy and a declining Picea forest in the Fichtelgebirge, NE Bavaria. *Oecologia* 87: 198–207.
- Gröcke DR, Bocherens H, and Mariotti A (1997) Annual rainfall and nitrogen-isotope correlation in Macropod collagen: Application as a paleoprecipitation indicator. *Earth and Planetary Science Letters* 153: 279–285.
- Hartman G (2010) Are elevated $\delta^{15}\text{N}$ values in herbivores in hot and arid environments caused by diet or animal physiology? *Functional Ecology* 25: 122–131.
- Heaton THE (1999) Spatial, species, and temporal variations in the $^{13}\text{C}/^{12}\text{C}$ ratios of C₃ plants: Implications for palaeodiet studies. *Journal of Archaeological Science* 26: 637–649.
- Hedges REM, Stevens RE, and Richards MP (2004) Bone as a stable isotope archive for local climatic information. *Quaternary Science Reviews* 23: 959–965.
- Higgins P and MacFadden BJ (2004) 'Amount effect' recorded in oxygen isotopes of Late Glacial horse (*Equus*) and bison (*Bison*) teeth from the Sonoran and Chihuahuan deserts, Southwestern United States. *Palaeogeography, Palaeoclimatology, Palaeoecology* 206: 337–353.
- Higgins P and MacFadden BJ (2009) Seasonal and geographic climate variabilities during the Last Glacial Maximum in North America: Applying isotopic analysis and macrophysical climate models. *Palaeogeography, Palaeoclimatology, Palaeoecology* 263: 15–27.
- Högberg P (1997) ^{15}N natural abundance in soil-plant systems. *New Phytologist* 137: 179–203.
- Koch PL (2007) Isotopic study of the biology of modern and fossil vertebrates. In: Michener R and Lajtha K (eds.) *Stable Isotopes in Ecology and Environmental Science*, 2nd edn., pp. 99–154. Boston: Blackwell Publishing.
- Koch PL and Barnosky AD (2006) Late Quaternary extinctions: State of the debate. *Annual Review of Ecology, Evolution, and Systematics* 37: 215–250.
- Koch PL, Diffenbaugh NS, and Hoppe KA (2004) The effects of late Quaternary climate and pCO₂ change on C₄ plant abundance in the South-Central United States. *Palaeogeography, Palaeoclimatology, Palaeoecology* 207: 331–357.
- Kohn MJ and Cerling TE (2002) Stable isotope compositions of biological apatite. In: Kohn MJ, Rakovan J, and Hughes JM (eds.) *Phosphates, Geochemical, Geobiological, and Materials Importance. Reviews in Mineralogy and Geochemistry*, vol. 48, pp. 455–488. Washington, DC: Mineralogical Society of America.
- Kohn MJ, Schoeninger MJ, and Barker WW (1999) Altered states: Effects of diagenesis on fossil tooth chemistry. *Geochimica et Cosmochimica Acta* 63: 2737–2747.
- Lee-Thorp JA and Beaumont PB (1995) Vegetation and seasonality shifts during the Late Quaternary deduced from $^{13}\text{C}/^{12}\text{C}$ ratios of grazers at Equus Cave, South Africa. *Quaternary Research* 43: 426–432.
- Lee-Thorp JA, Sponheimer M, Passey BH, de Ruiter DJ, and Cerling TE (2010) Stable isotopes in fossil hominid tooth enamel suggest a fundamental dietary shift in the Pliocene. *Philosophical Transactions of the Royal Society B* 365: 3389–3396.
- Lee-Thorp JA, Sponheimer M, and Van der Merwe NJ (2003) What do stable isotopes tell us about Hominid dietary and ecological niches in the Pliocene? *International Journal of Osteoarchaeology* 13: 104–113.
- Lee-Thorp JA, Thackeray JF, and van der Merwe NJ (2000) The hunters and the hunted revisited. *Journal of Human Evolution* 39: 565–576.
- Luz B, Cormie A, and Schwarcz HP (1990) Oxygen isotope variations in phosphate of deer bones. *Geochimica et Cosmochimica Acta* 54: 1723–1728.
- Männel TT, Auerswald K, and Schnyder H (2007) Altitudinal gradients of grassland carbon and nitrogen isotope compositions are recorded in the hair of grazers. *Global Ecology and Biogeography* 16: 583–592.
- Michelsen A, Quarmby C, Sleep D, and Jonasson S (1998) Vascular plant ^{15}N natural abundance in heath and forest tundra ecosystems is closely correlated with presence and type of mycorrhizal fungi in roots. *Oecologia* 115: 406–418.
- Miller GH, Fogel ML, Magee JW, Gagan MK, Clarke SJ, and Johnson BJ (2005) Ecosystem collapse in Pleistocene Australia and a human role in megafaunal extinction. *Science* 309: 287–290.
- Murphy BP and Bowman DMJS (2006) Kangaroo metabolism does not cause the relationship between bone collagen $\delta^{15}\text{N}$ and water availability. *Functional Ecology* 20: 1062–1069.
- Passey BH and Cerling TE (2002) Tooth enamel mineralization in ungulates: Implications for recovering a primary isotopic time series. *Geochimica et Cosmochimica Acta* 18: 3225–3234.
- Passey BH, Robinson TF, Ayliffe LK, et al. (2005) Carbon isotope fractionation between diet, breath CO₂, and bioapatite in different mammals. *Journal of Archaeological Science* 32: 1459–1470.
- Prideaux GJ, Ayliffe LK, DeSantis LR, et al. (2009) Extinction implications of a chenopod browse diet for a giant Pleistocene kangaroo. *Proceedings of the National Academy of Sciences USA* 106(28):11646–11650.
- Prideaux GV, Long JA, Ayliffe LK, et al. (2007) An arid-adapted middle Pleistocene vertebrate fauna from South-Central Australia. *Nature* 445: 422–425.
- Pushkina D, Bocherens H, Chaimanee Y, and Jaeger J-J (2010) First dietary and paleoenvironmental reconstructions from the late Middle Pleistocene Snake cave in Northeastern Thailand using stable carbon isotopes. *Naturwissenschaften* 97: 299–309.
- Richards MP, Fuller BT, Sponheimer M, Robinson T, and Ayliffe L (2003) Sulphur isotopes in palaeodietary studies: A review and results from a controlled feeding experiment. *International Journal of Osteoarchaeology* 13: 37–45.
- Richards MP and Hedges REM (2003) Variations in bone collagen $\delta^{13}\text{C}$ and $\delta^{15}\text{N}$ values of fauna from Northwest Europe over the last 40 000 years. *Palaeogeography, Palaeoclimatology, Palaeoecology* 193: 261–267.

- Richards MP and Trinkaus E (2009) Isotopic evidence for the diets of European Neanderthals and early modern humans. *Proceedings of the National Academy of Sciences USA* 106: 16034–16039.
- Richards MP, Pettitt PB, Stiner MC, and Trinkaus E (2001) Stable isotope evidence for increasing dietary breadth in the European mid-Upper Paleolithic. *Proceedings of the National Academy of Sciences of the United States of America* 98: 6528–6532.
- Richards MP, Pettitt PB, Trinkaus E, Smith FH, Paunovic M, and Karavanic I (2000) Neandertal diet at Vindija and Neandertal predation: The evidence from stable isotopes. *Proceedings of the National Academy of Sciences of the United States of America* 97: 7663–7666.
- Richards MP, Taylor G, Steele T, et al. (2008) Isotopic dietary analysis of a Neandertal and associated fauna from the site of Jonzac (Charente-Maritime), France. *Journal of Human Evolution* 55: 179–185.
- Robbins CT, Felicetti LA, and Sponheimer M (2005) The effect of dietary protein quality on nitrogen isotope discrimination in mammals and birds. *Oecologia* 144: 534–540.
- Schulze E-D, Chapin FS III, and Gebauer G (1994) Nitrogen nutrition and isotope differences among life forms at the Northern treeline of Alaska. *Oecologia* 100: 406–412.
- Schwarcz HP, Dupras TL, and Fairgrieve SI (1999) ^{15}N enrichment in the Sahara: In search of a global relationship. *Journal of Archaeological Science* 26: 629–636.
- Smith CI, Craig OE, Prigodich RV, et al. (2005) Diagenesis and survival of osteocalcin in archaeological bone. *Journal of Archaeological Science* 32: 105–113.
- Smith CI, Fuller BT, Choy K, and Richards MP (2009) A three-phase liquid chromatographic method for $\delta^{13}\text{C}$ analysis of amino acids from biological protein hydrolysates using liquid chromatography–isotope ratio mass spectrometry. *Analytical Biochemistry* 390: 165–172.
- Smith JM, Lee-Thorp JA, and Sealy JC (2002) Stable carbon and oxygen isotopic evidence for late Pleistocene to middle Holocene climatic fluctuations in the interior of Southern Africa. *Journal of Quaternary Science* 17: 683–695.
- van der Merwe NJ and Medina E (1991) The canopy effect, carbon isotope ratios and foodwebs in Amazonia. *Journal of Archaeological Science* 18: 249–259.
- Wright LE and Schwarcz HP (1998) Stable carbon and oxygen isotopes in Human tooth enamel: Identifying breastfeeding and weaning in Prehistory. *American Journal of Physical Anthropology* 106: 1–18.

Terrestrial Organic Materials

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Introduction

The elements carbon, hydrogen, and oxygen, which are the main constituents of terrestrial organic material, each have more than one stable (nonradioactive) isotope with nearly identical chemical properties but a difference in mass. By convention, the ratio of two stable isotopes is expressed relative to a standard using the delta notation in parts per mil (‰). In terrestrial organic material, the relevant isotope ratios are $\delta^{13}\text{C}$, $\delta^{18}\text{O}$, and $\delta^2\text{H}$ (or δD for deuterium). They are measured by first producing a suitable target gas, by either combustion or pyrolysis, and subsequent analysis by isotope ratio mass spectrometry (McCarroll and Loader, 2005). Since physical and biochemical processes are often sensitive to differences in mass, the ratios of the stable isotopes in organic material are not the same as those in the source (usually air or water). The degree of difference between the source and the organic product, or the total fractionation, may contain information not only on the source material, including the carbon isotopic ratios of atmospheric CO_2 , and the isotopic ratios of source water, but also on the degree of change that occurred as a result of environmental influences on fractionation. The isotopic signatures of the sources and changes due to environmental influences both may contain paleoenvironmental information. The most important continuous archives of terrestrial organic material suitable for stable isotopic analyses are tree rings and peat profiles.

Stable Isotopes in Tree Rings

As natural archives of paleoclimate information, trees have the great advantage that, under many conditions, they produce annual rings. A few tree species can live for millennia, but even for shorter-lived species, counting, measuring, and cross-matching the rings of many overlapping generations allow very long, well-replicated, annually resolved, and precisely dated chronologies to be constructed. Variations in the width of the annual rings in these long chronologies provide one of the most commonly used high-resolution 'proxy' sources of paleoclimate information. Suitable trees are widespread, and sampling and measuring are relatively inexpensive. However, the time series of ring widths need to be analyzed very carefully because they contain trends that relate to the age of the trees, and removing these trends without degrading the climate information is challenging. It is particularly difficult to retain long-term, low-frequency climate changes. One attraction of stable isotopes in tree rings is that the time series probably does not contain long-term trends related to tree age, and may thus retain climate information at all temporal frequencies. Also, the growth of trees, and therefore the annual ring width, is influenced by a myriad of interconnected factors,

and so is difficult to model mechanistically. The interpretation of ring-width series, therefore, generally relies on statistical inference and in particular on regression techniques. In contrast, the factors that control the fractionation of isotopes in trees are quite well understood and relatively simple in comparison, so that a mechanistic understanding of the processes involved, and of the ecophysiology of particular tree species, can be used to guide paleoenvironmental interpretations. Early attempts at 'stable isotope dendroclimatology' used blocks of wood covering several years and/or pooled the wood of several trees prior to analysis. However, recent advances in sample preparation and mass spectrometry have greatly accelerated the rate at which organic samples can be analyzed, and the use of stable isotopes in tree rings to reconstruct past climate is a rapidly expanding field of research (McCarroll and Loader, 2004, 2005).

Isotopic Fractionation by Trees

The main constituents of wood are cellulose, which forms the framework of cells, and lignin, which adds strength. These components contain carbon that is taken from CO_2 that enters the leaves via the stomata, whereas the oxygen and hydrogen are derived from water taken up through the roots. As a consequence of this composite structure, in ancient samples there is a potential for preferential decomposition of lignins or cellulose. To avoid these effects, a single compound is isolated from the wholewood prior to isotopic analysis. Typically, the fraction analyzed is the cellulose, as this signal is derived exclusively from the dated rings of the tree. However, studies with lignin have also been demonstrated to preserve climatic information at high (annual) resolution (Loader et al., 2003).

The stable isotopes of carbon in CO_2 are fractionated first as the gas enters the stomata and again as it is taken up by the photosynthetic enzyme, in both cases leading to an increase in the relative abundance of the lighter isotope and therefore a reduction in $\delta^{13}\text{C}$ (more negative values). The first step occurs because lighter molecules diffuse through a gap more easily than heavier ones. As gas molecules bounce off each other, the lighter ones bounce furthest, so the fractionation due to diffusion is controlled by the difference in mass. This means that the fractionation of carbon as it diffuses into the stomatal chambers remains at about -4.4‰ , almost irrespective of the environmental conditions or the size of the stomatal opening. It only increases under the rare conditions when the stomatal opening is so small that molecules bouncing off the sides, rather than each other, become important. The second fractionation, when the CO_2 is taken up by the photosynthetic enzyme, is similarly insensitive to environmental conditions, including temperature, and remains at about -27‰ (Farquhar et al., 1982) (Figure 1).

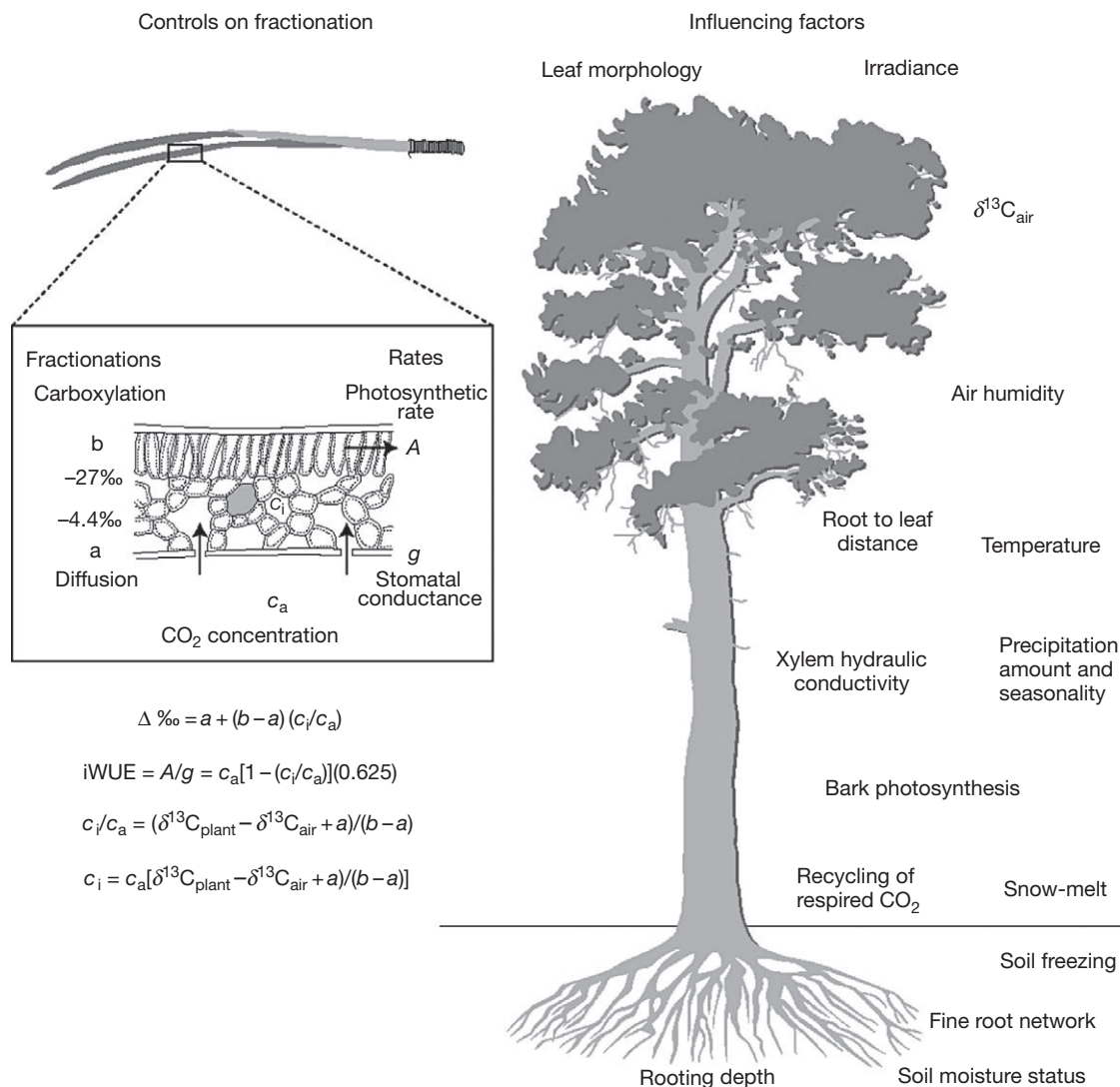


Figure 1 Diagram of a needle-leaf tree showing the main controls on fractionation of carbon isotopes and the environmental factors that influence them. Discrimination against ^{13}C (Δ) is expressed in per mil (‰) relative to a standard (Vienna Pee Dee Belemnite (VPDB)). Discrimination occurs during diffusion through the stomata (a) and again during carboxylation (b). Where the ambient concentration of CO₂ (c_a) is known, the key variable is the internal concentration of CO₂ (c_i), which is controlled by the balance between stomatal conductance (g) and photosynthetic rate (A). This relationship can be used to calculate intrinsic water use efficiency (iWUE). After McCarroll D and Loader NJ (2004) Stable isotopes in tree rings. *Quaternary Science Reviews* 23: 771–801.

The insensitivity of the fractionations to environmental variables means that the isotopic ratios of the gas that enters the leaf is controlled by that of the ambient air, so that tree-ring $\delta^{13}\text{C}$ must be affected by changes in the isotopic composition of the atmosphere. The second fractionation, however, does not act on ambient air but on the CO₂ in air that is within the stomatal chambers. As CO₂ is removed and assimilated, the relative abundance of the heavier isotopes remaining in this intercellular CO₂ must increase, and as more of this enriched CO₂ is assimilated, the $\delta^{13}\text{C}$ of the resulting sugars must also increase, even though the fractionation remains at about -27‰.

The fractionation of carbon that occurs in tree leaves is not a direct response to environmental variables, but rather a measure of the relative rates at which CO₂ enters and is removed

from the stomatal chambers. The controls are therefore stomatal conductance and photosynthetic rate. Where stomatal conductance is high relative to photosynthetic rate, the assimilates will have low $\delta^{13}\text{C}$ values, but where stomatal openings become restricted in response to moisture stress, or when photosynthetic rates are very high, the internal concentration of CO₂ will drop, becoming enriched in ^{13}C and yielding assimilates with higher $\delta^{13}\text{C}$ values. The sugars produced in the leaf may be used directly to produce new wood, in which case, the climatic conditions that influenced the signal imparted in the leaf will be transferred to the wood cells produced in the same year, or they may be stored and remobilized later, leading to further fractionation and a lag between the climatic forcing and its expression in the isotopic composition of the tree ring (Helle and Schleser, 2004). This effect is most pronounced in

the earlywood fraction of the tree ring; consequently, to avoid remobilized carbon, and thus maintain the annual resolution of the signal, many recent studies have used only the latewood portion of tree rings, which forms during the summer.

The ratios of the stable isotopes of oxygen and hydrogen in tree rings are more complex than those of carbon because they conflate two separate signals: a source water signal and an isotopic enrichment due to leaf transpiration. The isotopic ratios in precipitation are controlled by a number of factors (Darling et al., 2005), including (but not always dominated by) temperature. However, to regard water-isotope ratios in tree rings as a simple paleothermometer is overly simplistic. The rainfall must first reach the tree roots, so the depth and structure of the root system of different species can mean that water of different ages is utilized. Wood formed in a given summer may, therefore, utilize water molecules from summer rainfall, precipitation from the previous winter (Robertson et al., 2001), or over a much longer time period. There is no fractionation when roots take up the soil water, but on reaching the leaves, evaporation results in a preferential loss of the lighter isotopes. The controls are very similar to those that govern fractionation due to evaporation from a lake, though there is an additional fractionation as water vapor diffuses through the stomata and leaf boundary layer. Although there is a small temperature effect, the dominant controls are the isotopic composition of water in the atmosphere, relative to the source water, and the difference between the ambient and intercellular vapor pressures. Low air humidity thus leads to greater enrichment of the heavier isotopes of both oxygen and hydrogen. Backward movement of isotopically enriched water from the sites of evaporation is opposed by the supply of unenriched source water, so the isotopic values of total leaf water are lower than those at the sites of evaporation (Barbour et al., 2001). There is a further fractionation when the water is incorporated in sucrose formed in the leaf, leading to a further enrichment of ^{18}O (+27‰) but a larger and less certain depletion of deuterium (−100‰ to −171‰) (Figure 2).

The situation is further complicated because a proportion of the oxygen atoms in the sugars that form in the leaf are able to exchange with (unenriched) atoms from xylem (source) water moving up the stem at the site where the wood is formed. The amount of exchange is difficult to predict, but controls the relative importance of the source water and leaf enrichment signals. The potential for exchange is probably larger for lignin than for cellulose. When sugars are used to form wood cells, there is a large fractionation of hydrogen in the opposite direction and similar in magnitude to that which occurs when the sugar is formed (+144‰ to +166‰), and as with oxygen, there is potential for exchange with xylem water and thus for altering the balance of the two signals (McCarroll and Loader, 2004, 2005).

Interpreting Stable Isotopes from Tree Rings

Early work on stable isotopes in tree rings interpreted changes in both the carbon- and the water-isotope ratios in terms of changes in temperature, treating them as paleothermometers. Unfortunately, this is a gross oversimplification and in reality none of them is controlled in a simple way by a single climatic

variable. Carbon-isotope ratios record the balance between stomatal conductance and photosynthetic rate, whereas the water isotopes take part of their signal from the isotopic ratios of the source water but then enrich it by evaporation in the leaf. Although each of these isotopes may, under certain circumstances, be strongly correlated with temperature, it is unlikely that temperature is actually the controlling factor. Temperature may correlate with carbon-isotopic ratios through covariance with changes in sunlight, which drives photosynthesis, or through covariance with relative humidity and rainfall, which control stomatal conductance. Temperature is only one of the factors that influence the isotopic ratios in source water, and although there is a small direct influence of temperature on fractionation at the sites of evaporation in the leaf, the dominant control is the difference in vapor pressure inside and outside of the leaf, which is strongly dependent on air relative humidity. Thus, although the correlation with temperature during the period when meteorological records are available can be used to reconstruct past temperatures using regression-based techniques, it is important to remember that correlation is not causation and that any change in one of the real controlling factors will change the isotopic ratios, irrespective of whether there was a change in temperature. Since the instrumental period coincides with the period of greatest anthropogenic disturbance, to both forests and the atmosphere, changes in the relationship between temperature and the controlling factors may well have occurred. Of course, this logic applies to most natural archives used as proxy sources of paleoclimate information, it is simply more apparent for tree ring stable isotopes. This is because trees provide absolute temporal control and a clearer (mechanistic) understanding of signal preservation than is often observed in other terrestrial paleorecords.

Rather than relying on the correlation between temperature and isotopic ratios to reconstruct paleotemperatures using 'black box' methods based on regression, it might be more enlightening to take advantage of the mechanistic understanding of fractionation and attempt to tease out the real environmental controls. This can be achieved in part by careful selection of trees and sites so that one control is likely to dominate (McCarroll and Loader, 2005). Shallow-rooted trees on sites with very freely draining soils, for example, are likely to be sensitive to the stomatal conductance control on carbon-isotope ratios and to respond rapidly to changes in source water-isotope ratios. Deeply rooted trees with access to a plentiful supply of groundwater, on the other hand, are likely to yield carbon-isotope ratios that are more responsive to changes in photosynthetic rate and water-isotope ratios, in which the source water signal is smoothed so that the evaporative enrichment signal is clearer. Perhaps even greater potential lies in combining different isotopes and other paleoclimate proxies that can be extracted from trees in a multiproxy approach that is guided by theory rather than purely by statistical inference (McCarroll et al., 2003).

Application of isotopic analysis is not limited to annual increments. Through high-resolution sampling of individual rings, rhythmic trends and periodicities can be identified that can be related to plant physiological processes occurring within the growth cycle (Helle and Schleser, 2003). This approach can be taken forward further to tropical species that do

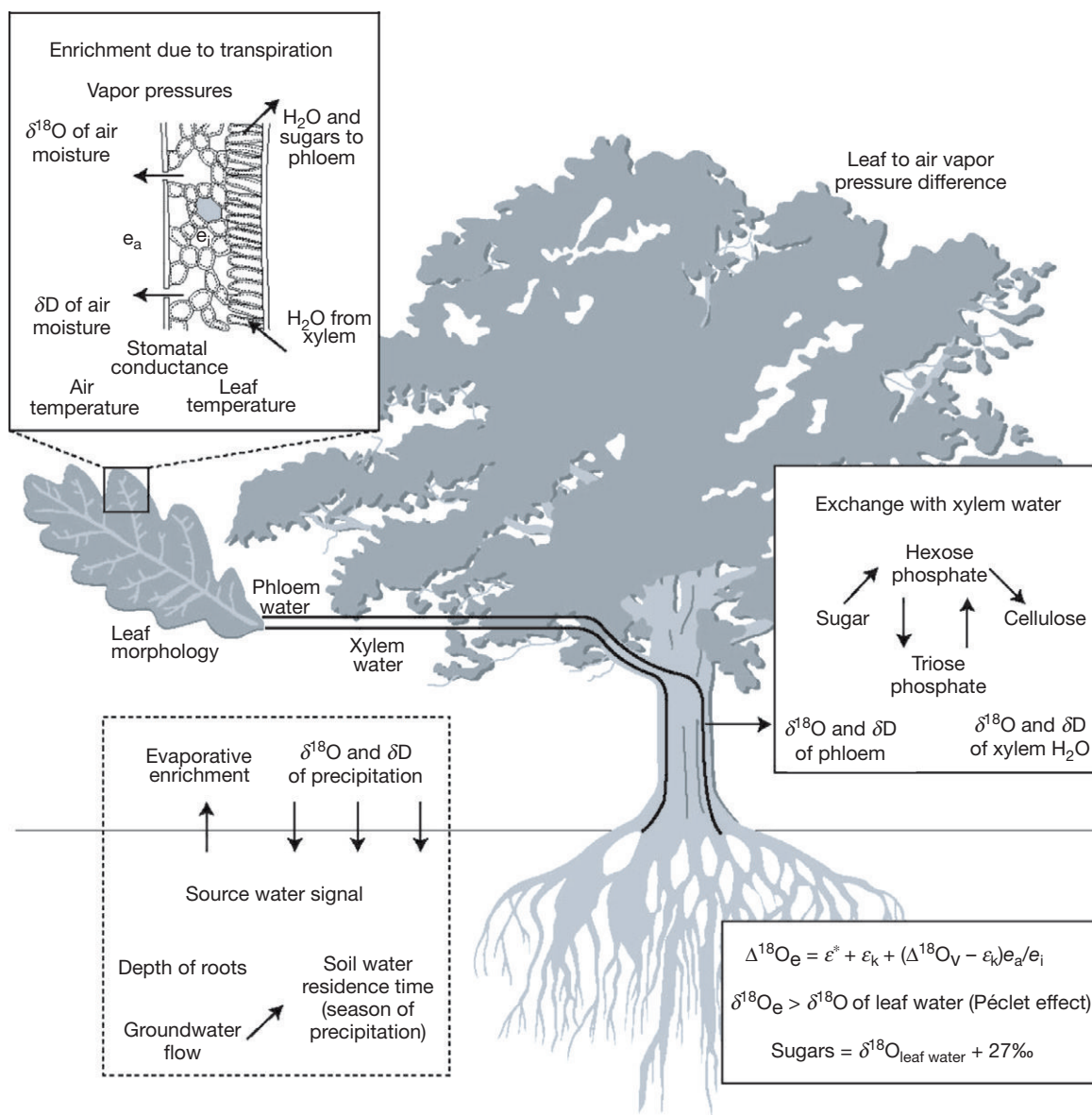


Figure 2 Diagram of a broad-leaved tree showing the main controls on the fractionation of oxygen isotopes and the environmental factors that influence them. There is no fractionation when source water is taken up by the roots. The level of enrichment of leaf water above source water, at the sites of evaporation ($\delta^{18}\text{O}_e$), depends upon the proportional depression of water vapor pressure by the heavier H_2^{18}O (ε^*), fractionation as water diffuses through the stomata and leaf boundary layer (ε_k), the isotopic composition of water vapor in the atmosphere, relative to source water ($\Delta^{18}\text{O}_v$), and the ambient and intercellular vapor pressures (e_a and e_i). The same equations control the fractionation of hydrogen isotopes, though the fractionation factors are different. After McCarroll D and Loader NJ (2004) Stable isotopes in tree rings. *Quaternary Science Reviews* 23: 771–801.

not exhibit regular annual banding, yet where an apparent (possibly seasonal/annual) cyclicity may be identified in their isotopic composition.

Stable Isotopes in Peat Profiles

Peat deposits provide one of the best-known archives of paleoclimate information for the Holocene. Peat accumulation can be continuous and the acidic conditions are suitable for the preservation of a wide range of organic materials including leaves, other plant macrofossils, and pollen. The preserved

macrofossils, or identifiable remains of the peat-forming plants, can be radiocarbon-dated to provide a chronology. Stable isotopic analysis of leaves provides a potential source of information on the stable isotope ratios in the vegetation surrounding the peat at different times (Rundgren et al., 2000). More attention has been paid, however, to isotopic analysis of the mosses and other peat-forming plants.

Some mosses, such as *Sphagnum*, differ from the higher plants in that, although photosynthesis appears to follow the C3 pathway, they do not possess functioning guard cells around their stomata and hence do not regulate water loss. The dominant controls on the fractionation of carbon isotopes,

therefore, are likely to be the isotope ratios of the source gas, photosynthetic rate, and the resistance to diffusion of CO₂. Resistance to diffusion is likely related to the thickness of the water layer, so that at higher moisture contents mosses would be expected to yield less negative values. This relationship with wetness is the opposite to that seen in the higher plants, including trees, where dry conditions lead to a reduction in stomatal conductance, a decline in internal concentration of CO₂, and thus to less negative $\delta^{13}\text{C}$ values. In the mosses, resistance to diffusion is equivalent to stomatal conductance, but a reduction in moisture content decreases resistance to diffusion, leading to more negative $\delta^{13}\text{C}$ values. This is supported by results from several species of *Sphagnum* from two peat bogs in Ireland, where plants growing on relatively dry hummocks yielded higher $\delta^{13}\text{C}$ values than those in wetter hollows (Price et al., 1997).

As with other plants, the isotopic ratio of the source gas used by mosses must be reflected in the isotopic ratios of the photosynthetic products. However, whereas mature trees have access to well-mixed atmospheric gases, mosses on the surface of a peat mire may be in contact with significant amounts of recycled CO₂, which will produce relatively light $\delta^{13}\text{C}$ values. Price et al. (1997) have argued that this effect will be most marked where the mosses are submerged, and particularly where there is a low density of living plants in pools. For this reason, they argue that hummock-top species of *Sphagnum* are likely to be the most useful for stable carbon-isotopic analysis.

The differences in carbon-isotope fractionation between mosses and sedges in peat have been proposed as a relatively high-resolution method for reconstructing past changes in the concentration of atmospheric CO₂, perhaps linked to changes in thermohaline circulation (Figge and White, 1995). The efficacy of this technique is contested, however, since some of the differences could be due to the effects of changes in moisture regime on fractionation by the mosses.

The fractionation of the water isotopes in mosses should be the same as for higher plants, in that isotope ratios should reflect the isotopic values of the source water and isotopic enrichment due to evaporation. However, where the plants in question are fully submerged, evaporation does not take place and so the isotopic values in the plant tissue should mirror the isotopic values of the source water, with a consistent offset due to fractionation during photosynthesis. Oxygen isotope results from *Sphagnum* cellulose confirm that submerged species yield values that mirror those of the water, whereas emergent species give a wider range of values, reflecting differences in the amount of evaporation (Zanazzi and Mora, 2005). Stable hydrogen and oxygen-isotopic analysis of submerged moss species from peat profiles may provide a powerful proxy for past changes in the isotopic composition of the bog water.

Stable Isotope Analysis of Pollen

The physical (numerical) analysis of pollen assemblages preserved in terrestrial and marine sedimentary archives provides one of the most established and powerful tools for the study of paleoecology and vegetation dynamics throughout the Quaternary.

More recently, significant interest has developed into the (CHO) isotopic analysis of pollen grains. This approach has only been made feasible through the development of continuous-flow mass spectrometry methods, which require smaller sample sizes than traditional off-line preparation and dual inlet techniques. A series of pilot studies has emerged within the last decade, which serves to identify the potential of this approach for paleoecological research. Loader and Hemming (2004) provide a review of the current state-of-the-art and discuss the current limitations and research opportunities for this emerging proxy.

Like many terrestrial paleoarchives, pollen grains comprise a range of chemical components. Typically, a single grain may be considered to consist of three principal structural elements: an inner nucleus which contains the reproductive and genetic materials, a cellulose inner layer or 'intine,' and a β -carotenoid ester outer layer or 'exine,' which represents ca. 65–80% of the grain and together may be further bound with lipids or coated with a thin polysaccharide layer. There is still debate as to the exact chemical structure of the exine, although it is believed to approximate C₉₀H₁₅₀O₃₃. Although incredibly resistant, being of a composite origin, pollen is susceptible to differential diagenesis. Typically, it is the exine that is selected for isotopic analysis, as it is this component that preserves in its morphology, the key to the species/genus of plant from which it originated (Figure 3).

Preparation of Pollen Samples

For conventional palynology, exines are purified from the sediment matrix through a suite of aggressive chemical processes. Isotopic analysis of raw and extracted modern pollen exines exhibit a species-specific isotopic offset of between ca. 2 and 4 per mil $\delta^{13}\text{C}$. Consequently, to avoid the effects of differential diagenesis, a chemical preparatory approach should be considered when preparing pollen extracts from the paleoarchive for isotopic analysis. Direct transfer of methodologies developed for conventional pollen preparation is likely to be inappropriate, as acetylation, for example, has been shown to contaminate isotopically the end product.

Carbon-isotopic analysis of modern pollen samples from different species have been shown to record different signals. It is therefore likely that in order to achieve a robust paleo-record, selection of a single species or genus may be required. At present, the greatest limitation to the wider application of this method lies in the extraction and purification of sufficient pollen grains for isotopic analysis. Sample concentration by sieving, dense media separation, and centrifugation, followed by manual picking, provide one of the most accessible methods currently available. Less time-consuming sample-preparation methods are urgently required. Application of fluid dynamic separation (split flow lateral transport thin separation (SPLIT)) methods or automated image recognition and micromanipulation exhibit the greatest potential for future advances in this field. Nevertheless, even using existing labor-intensive picking methods, it is possible to extract sufficient material for either radiocarbon dating or stable isotope analyses.

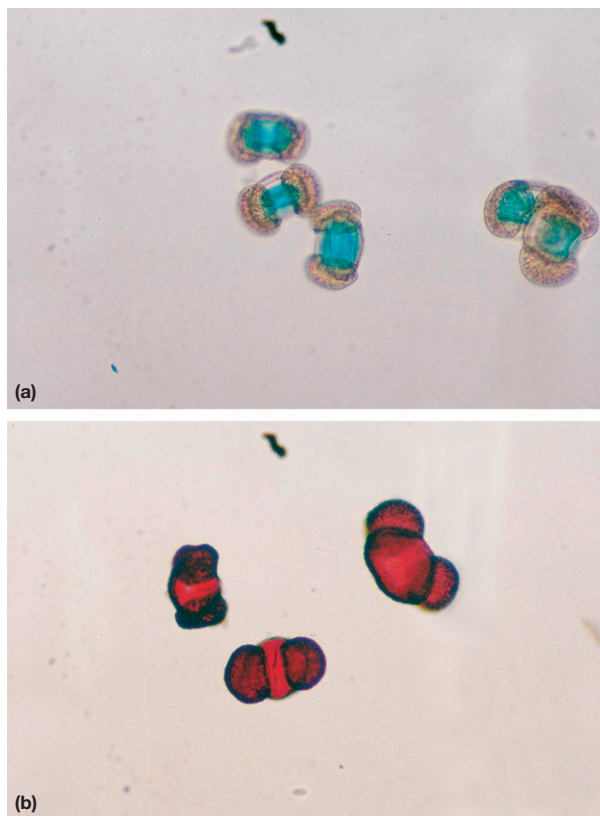


Figure 3 (a) Untreated *Pinus sylvestris* pollen composite-stained with safranin-fast green, highlighting the cellulose intine of the grains (green). Note: No cellulose is present in the saccae and the exine is only partially stained (pink) due to the presence of a polysaccharide layer. (b) Acid treated *Pinus sylvestris* pollen composite-stained with safranin-fast green. The strong safranin staining of the exine (pink) demonstrates removal of the outer polysaccharide layer. The cellulose intine has been completely removed. After Loader NJ and Hemming DL (2000) Preparation of pollen for stable carbon isotope analyses. *Chemical Geology* 165: 339–344.

Large, easily extractable genera such as *Picea*, *Abies*, and, to a lesser extent, *Pinus* represent likely candidate species for paleo-ecological reconstruction based upon isotopic analysis (500 spruce grains $\approx 8 \mu\text{g}$ of carbon). Species differences are less important in applications where Poaceae pollen have been analyzed isotopically to establish the relative proportion of grassland community species which use either C3 or C4 photosynthetic mechanisms. Changes in the isotopic ratio of the bulk Poaceae signal reflect within it changes in (paleo)grassland composition from which paleoclimate may be inferred (Amundson et al., 1997). In all cases, representative sampling and thorough replication are essential, as is an understanding of pollen production, dispersal, and preservation. Interpretation of the isotopic data should also include consideration of the physical pollen record.

Environmental Signals

Unlike the formation of cellulose in tree rings, the imprinting of an environmental signal in pollen grains is still a subject

of some debate. It is likely that many of the mechanisms described for tree rings are relevant to those responsible for pollen formation; however, the subsequent fate of the photo-assimilates and the mechanisms and pathways they follow during their remobilization/use in the production of pollen remains largely unknown. Empirical studies provide some insight into these processes. Carbon isotopes are likely to reflect the composition of the photosynthates occurring during the formation of the pollen initials (year $t - 1$) and the conditions prevailing during pollen maturation and photosynthesis prior to pollen fly (year t). Mechanistically, the oxygen- and hydrogen-isotope ratios of the photosynthates used to form the pollen should reflect the composition of the source water, with evaporative enrichment (and potentially subsequent re-exchange with source water). Initial investigations have broadly demonstrated this relationship for hydrogen, but it would appear that the signal preserved in the oxygen-isotope ratio of pollen exines is more complex and may reflect the relative degree to which the source water signal is transferred to the pollen grain or differences in biochemical fractionations occurring along the respective photosynthetic pathways. Further work is needed to evaluate and better quantify these processes.

See also: [Dendrochronology](#); [Dendroclimatology](#); [Pollen Analysis, Principles](#).

References

- Amundson R, Evett RR, Jahren AH, and Bartolome J (1997) Stable carbon isotope composition of Poaceae pollen and its potential in paleovegetational reconstructions. *Review of Palaeobotany and Palynology* 99: 17–24.
- Barbour MM, Andrews TJ, and Farquhar GD (2001) Correlations between oxygen isotope ratios of wood constituents of *Quercus* and *Pinus* samples from around the world. *Australian Journal of Plant Physiology* 28: 335–348.
- Darling WG, Bath AH, Gibson JJ, and Rozanski K (2005) Isotopes in water. In: Leng MJ (ed.) *Isotopes in Palaeoenvironmental Research*, pp. 1–66. The Netherlands: Springer.
- Farquhar GD, O'Leary MH, and Berry JA (1982) On the relationship between carbon isotope discrimination and intercellular carbon dioxide concentration in leaves. *Australian Journal of Plant Physiology* 9: 121–137.
- Figge RA and White JWC (1995) High resolution Holocene and late glacial atmospheric CO_2 record: Variability tied to changes in thermohaline circulation. *Global Biogeochemical Cycles* 9: 391–403.
- Helle G and Schleser GH (2004) Beyond CO_2 -fixation by Rubisco – An interpretation of C-13/C-12 variations in tree rings from novel intra-seasonal studies on broad-leaf trees. *Plant, Cell Environment* 27: 367–380.
- Loader NJ and Hemming DL (2000) Preparation of pollen for stable carbon isotope analyses. *Chemical Geology* 165: 339–344.
- Loader NJ and Hemming DL (2004) The stable isotope analysis of pollen as an indicator of terrestrial palaeoenvironmental change: A review of progress and recent developments. *Quaternary Science Reviews* 23: 893–900.
- Loader NJ, Robertson I, and McCarroll D (2003) Comparison of stable carbon isotope ratios in the whole wood, cellulose and lignin of oak tree rings. *Palaeogeography, Palaeoclimatology, Palaeoecology* 196: 395–407.
- McCarroll D, Jalkanen R, Hicks S, et al. (2003) Multi-proxy dendroclimatology: A pilot study in northern Finland. *The Holocene* 13: 829–838.
- McCarroll D and Loader NJ (2004a) Stable isotopes in tree rings. *Quaternary Science Reviews* 23: 771–801.
- McCarroll D and Loader NJ (2004b) Isotopes in tree rings. In: Leng MJ (ed.) *Isotopes in Palaeoenvironmental Research*, pp. 67–116. The Netherlands: Springer.
- Price GD, McKenzie JE, Pilcher JR, and Hoper ST (1997) Carbon-isotope variation in Sphagnum from hummocks-hollow complexes: Implications for Holocene climate reconstruction. *The Holocene* 7: 229–233.

- Robertson I, Waterhouse JS, Barker AC, Carter AHC, and Switsur VR (2001) Oxygen isotope ratios of oak in east England: Implications for reconstructing the isotopic composition of precipitation. *Earth and Planetary Science Letters* 191: 21–31.
- Rundgren M, Loader NJ, and Beerling DJ (2000) Variations in the carbon isotope composition of late Holocene plant macrofossils: A comparison of whole-leaf and cellulose trends. *The Holocene* 10(1): 149–154.
- Zanazzi A and Mora G (2005) Palaeoclimate implications of the relationship between oxygen isotope ratios of moss cellulose and source water in wetlands of Lake Superior. *Chemical Geology* 222: 281–291.

Non-Lacustrine Terrestrial Studies

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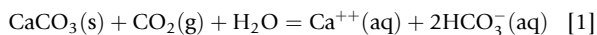
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Introduction

The carbon- (expressed in ‰ as $\delta^{13}\text{C}$ (VPDB)) and oxygen- ($\delta^{18}\text{O}$ (VPDB)) isotopic composition of surficial carbonate has been widely used in paleoclimatic and paleoenvironmental reconstruction during the Quaternary. In surficial carbonates we include: soil carbonates and sphaerosiderites, spring travertines, and a variety of early meteoric cements, usually formed in the saturated zone. Other important surficial meteoric carbonates covered by other articles in this Encyclopedia include lacustrine (see [Lake Sediments](#)) and speleothem (see [Speleothems](#)) carbonates. There is a large literature on this topic that cannot be comprehensively reviewed in this short article. Instead, we examine a few representative soil carbonate records of environmental change over the Neogene and Pleistocene, and groundwater carbonate records from Devils Hole, Nevada.

Formation of soil carbonates and groundwater carbonates can be viewed as expressions of a single groundwater flow continuum, summarized in [Figure 1](#). The general equation for weathering (to the right) and calcite precipitation (to the left) in this continuum is:



Here we use calcite as the parent material but we could just as easily have used most silicate minerals in its place. Calcite or silicate dissolution begins in the soil (vadose) zone in the presence of elevated soil pCO_2 (>350 ppmV) produced by plant respiration and decay. This dissolution process can continue below the vadose zone and into the saturated zone until significant levels of supersaturation with respect to calcite are reached, whereupon carbonate precipitates. In dry soil settings, supersaturation often develops prior to the next soil flushing event, due largely to soil dewatering by plants and by evaporation. In the shallow saturated zone, such in some caves and springs, calcite precipitation is driven by off-gassing of dissolved CO_2 . In deeper systems, such as Devils Hole in southern Nevada, decompression during ascent of deeply circulated groundwater may induce carbonate formation.

Carbonate is among the most abundant of secondary minerals in the geologic record, and can form in a large range of burial depths and conditions. The use of the soil diffusion model (see below) as an interpretive framework for carbon-isotope values in soils assumes that the material being analyzed is a well-preserved soil carbonate. One should rely on a suite of features for accurate identification of soil carbonate over other secondary forms of calcite cementation. These criteria include (1) firm associations with other soil features ([Figure 2](#)), (2) nodule form ([Figures 2\(b\)](#) and [\(c\)](#)) and micritic texture ([Figures 2\(i\)](#) and [\(j\)](#)), (3) reworking of nodules into contemporaneous channels ([Figures 2\(h\)](#) and [\(i\)](#)), and (4) the absence of detrital carbonate contamination. (see [Soil Morphology in Quaternary Studies](#)).

The Devils Hole calcites in southern Nevada lie toward the distal end of a large ($\sim 12\,000\text{ km}^2$) flow system hosted largely by fractured carbonate rocks. Coarse-grained, laminated calcites coat fractures below the top of the water table ([Figure 3](#)). Access to these fracture coatings is possible at Devils Hole, where collapse of the roof of one large fracture has exposed the aquifer and permitted divers to enter, as well as at various locations in nearby Death Valley, where recent tectonism has exhumed other fracture coatings ([Figure 3](#)).

The Oxygen-Isotopic Composition of Surficial Meteoric Carbonates

Background

The oxygen-isotopic composition of surficial meteoric carbonate ($\delta^{18}\text{O}_{\text{cc}}$) (VPDB) is determined by both air and local vadose zone temperature, by the $\delta^{18}\text{O}$ value of rainfall ($\delta^{18}\text{O}_{\text{p}}$), and by soil evaporation.

Temperature effects

Temperature can potentially influence $\delta^{18}\text{O}_{\text{cc}}$ values in two ways, through the influence of (1) soil temperature on fractionation between water and calcite during soil carbonate formation, and (2) air temperature on $\delta^{18}\text{O}_{\text{p}}$.

The temperature dependence of the fractionation factor ($\alpha_{\text{c-w}}$) between calcite and water is:

$$\alpha_{\text{c-w}} = \exp\left(\frac{(18,030/T) - 32.42}{1000}\right) \quad [2]$$

with temperature (T) in K ([Kim and O'Neil, 1997](#)). This equates to an approx. -0.22 to -0.24‰ C^{-1} change in $\delta^{18}\text{O}_{\text{cc}}$ values with soil temperature at constant $\delta^{18}\text{O}_{\text{p}}$. Temperature in the shallow vadose zone (2–10 m) approximates mean annual temperature (MAT), reducing potentially large seasonal temperature extremes. At depths of >100 m, the effects of local geothermal gradients ($2\text{--}3\text{ °C per }100\text{ m}$) will develop, as in the case of the mildly geothermal waters at Devils Hole.

The oxygen-isotope value of precipitation ($\delta^{18}\text{O}_{\text{p}}$) is correlated with air temperature at mid- to high latitudes. The slope of this dependence (the 'Dansgaard relationship') ranges between $+0.2$ and 0.9‰ C^{-1} but averages $+0.6 \pm 0.1\text{‰ C}^{-1}$ in most mid-latitude settings of soil carbonate formation.

These two influences of temperature on $\delta^{18}\text{O}_{\text{cc}}$ values are opposite in sign ($-0.23 \pm 0.1\text{‰ C}^{-1}$ versus $+0.6 \pm 0.1\text{‰ C}^{-1}$). They should combine to increase $\delta^{18}\text{O}_{\text{cc}}$ values by $+0.25$ to $+0.5\text{‰ C}^{-1}$. In agreement with this prediction, in carbonates from modern soils with rainfall exceeding $\sim 10\text{ cm yr}^{-1}$ (low evaporation), the observed slope is $+0.44\text{‰ C}^{-1}$ ($r^2=0.56$) ([Figure 4](#)).

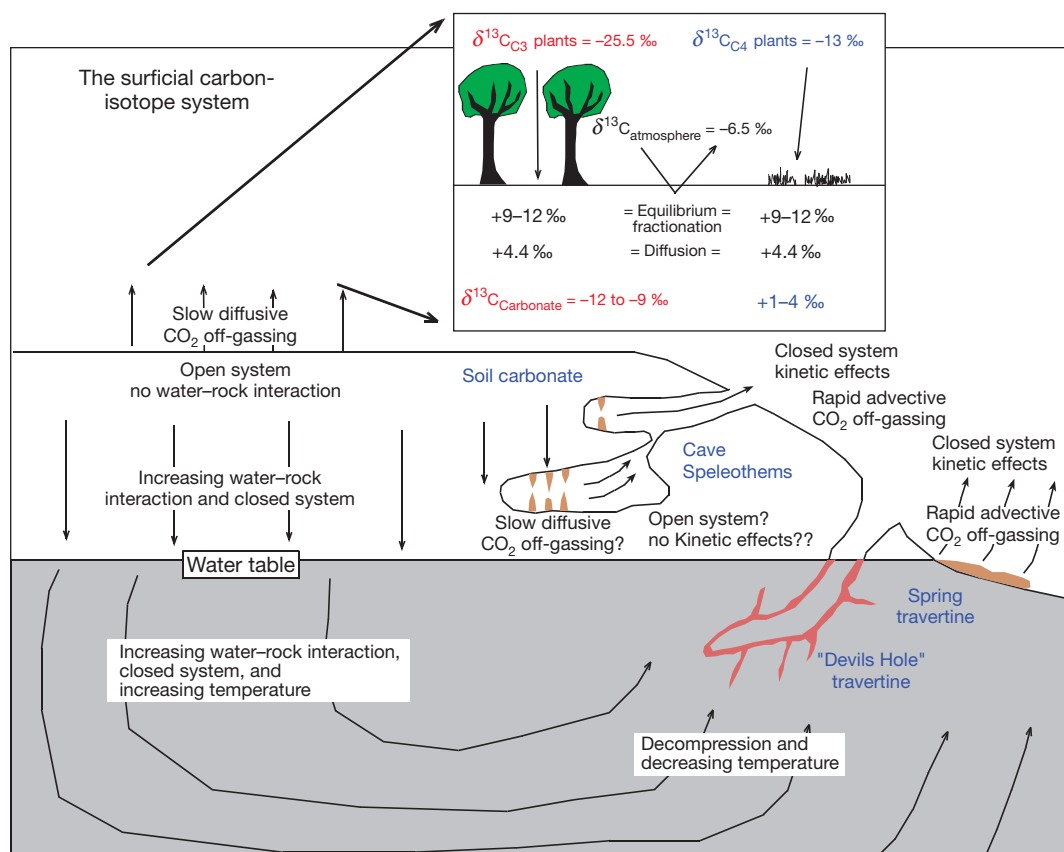


Figure 1 Summary diagram of formation processes for soil carbonates (inset), spring, and deep groundwater (Devils Hole) carbonates.

The $+0.25$ to $+0.5\text{‰}$ C^{-1} dependence means that the large uncertainties in the $\delta^{18}\text{O}_\text{p}$ value produce major uncertainties in the reconstructed paleotemperature, but conversely, that large uncertainties in paleotemperature produce relatively small uncertainties in the reconstructed $\delta^{18}\text{O}_\text{p}$. This is why the $\delta^{18}\text{O}_\text{cc}$ values are poor paleothermometers but can be useful in constraining paleo- $\delta^{18}\text{O}_\text{p}$ values.

$\delta^{18}\text{O}$ composition of precipitation ($\delta^{18}\text{O}_\text{p}$)

$\delta^{18}\text{O}_\text{p}$ is a strong predictor of $\delta^{18}\text{O}_\text{cc}$ values, especially in temperature soils where evaporation rates are minimized (Figure 5). The oxygen value of precipitation is influenced by several factors, including air temperature, changes in the $\delta^{18}\text{O}$ value of seawater ($\delta^{18}\text{O}_\text{seawater}$) and, at low latitudes, by changes in rainfall amount, referred to as the amount effect. The amount effect is the dominant control on $\delta^{18}\text{O}_\text{p}$ at low latitudes and is $\sim -1.5\text{‰}$ per 10 cm rainfall globally (Rozanski et al., 1993). This effect is not only visible in rainfall records from island stations, but also in monsoonal systems, where the $\delta^{18}\text{O}_\text{p}$ values during the summer rainy season are much lower than in the winter nonrainy season.

The ocean is the source of most atmospheric moisture, hence changes in the $\delta^{18}\text{O}_\text{seawater}$ values should translate directly into changes in $\delta^{18}\text{O}_\text{p}$ values. During glacial periods in the Quaternary, $\delta^{18}\text{O}_\text{seawater}$ values increase by $\sim 1.0\text{‰}$. At mid- to high latitudes, increases in $\delta^{18}\text{O}_\text{seawater}$ during glacial periods would dampen decreases in $\delta^{18}\text{O}_\text{cc}$ values due to lower temperatures.

Soil water evaporation

When RH is high, $\delta^{18}\text{O}_\text{soilwater} = \delta^{18}\text{O}_\text{p}$, as discussed above. In sphaerosiderites and groundwater carbonates, the evaporation term is probably unimportant. In soils from dry regions, evaporation can strongly increase $\delta^{18}\text{O}_\text{soilwater}$ values with respect to $\delta^{18}\text{O}_\text{p}$ values. In general, evaporative enrichment in ^{18}O of water increases linearly with decreased relative humidity. Results from the Atacama Desert and the Death Valley portion of the Mojave Desert (Figure 5) illustrate the strong effects of evaporation on $\delta^{18}\text{O}_\text{cc}$ values.

The Carbon-Isotopic Composition of Surficial Meteoric Carbonates

Organic processes tend to dominate the $\delta^{13}\text{C}$ value of surficial carbonates. In this article, we focus on the carbon isotopes in soil carbonates, since the $\delta^{13}\text{C}$ values of other types of carbonates, although of interest from a process point of view, do not contribute much to paleoenvironmental reconstruction. For example, spring travertines and many cave speleothems (Fantidas and Ehhalt, 1970; Hendy, 1971; reviewed in Quade (2003)) undergo kinetic enrichment in ^{13}C due to CO_2 off-gassing and evaporation, greatly reducing their usefulness in paleoclimatic reconstruction. The $\delta^{13}\text{C}$ value of sphaerosiderites (Ludvigson et al., 2002) and other meteoric calcites can be strongly influenced by methanogenesis, especially in wet-logged, organic-rich settings. In other instances, water-

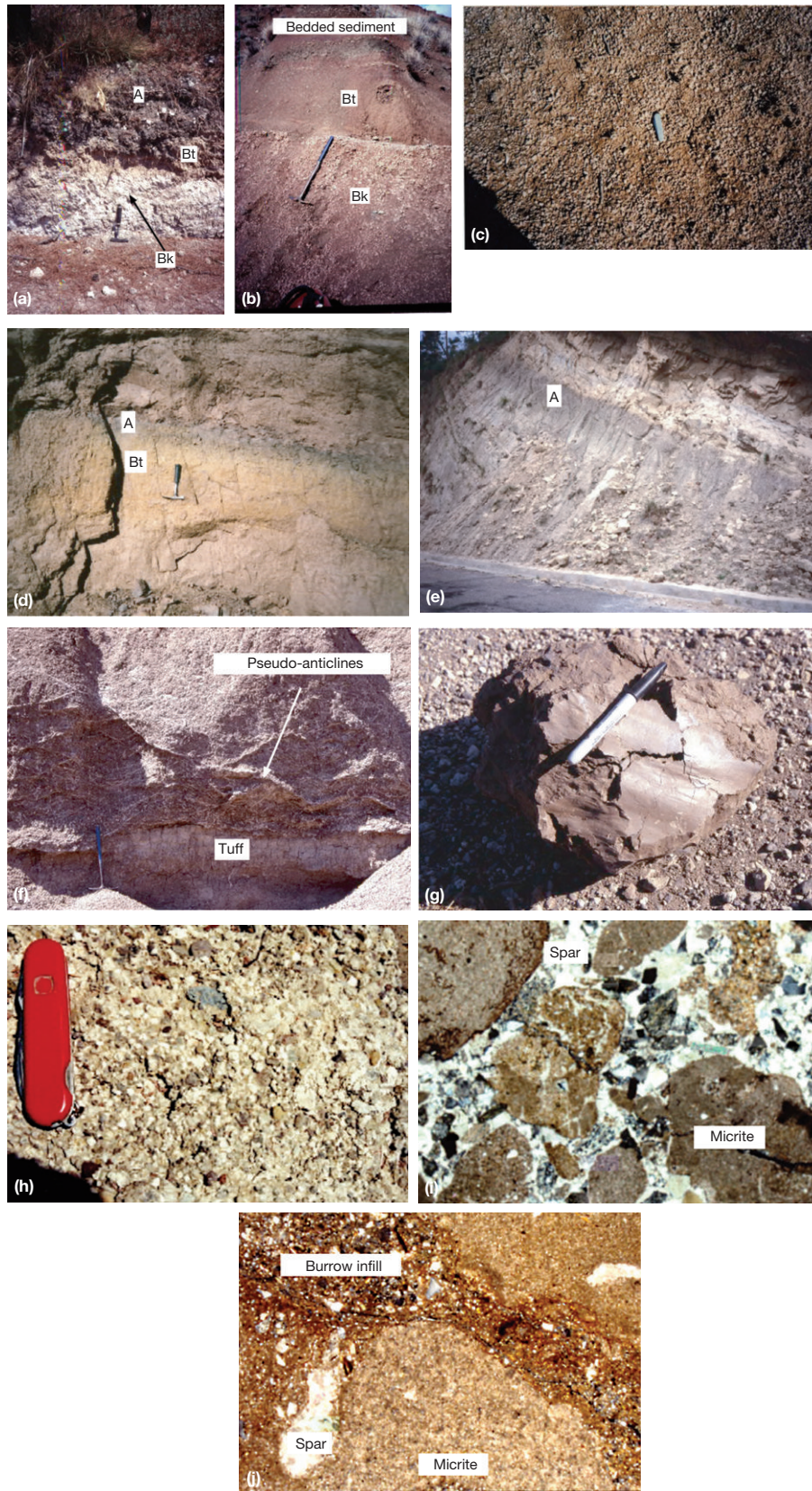


Figure 2 Diagnostic features of soil and paleosols. A = organic-rich A horizon, Bt = clay-rich argillic horizon, and Bk = calcic horizon. (a) Modern soil, Samos, Greece (hammer ~28 cm); (b) buried paleosol, late Miocene, Chinji Formation, Pakistan, (hammer ~66 cm); (c) soil carbonate nodules, late Miocene, Dhok Pathan Formation, Pakistan, (knife ~8 cm); (d) buried paleosol, late Pliocene, Pakistan (hammer ~28 cm); (e) buried organic-rich paleosol, Pliocene, Surai Khola, Nepal; (f) paleovertilsol (hammer ~66 cm) with pseudo-anticlines and underlying airfall tuff; (g) detail of slickensides on ped (pen ~14 cm); (h) soil carbonate nodules reworked into contemporaneous channel lag (knife ~8 cm), middle Pleistocene, Busidima Formation, Gona, Ethiopia; (i) photomicrograph of soil nodules reworked into conglomerate and cemented by secondary spar, Pliocene, Pakistan; (j) photomicrograph of soil micrite with burrow infill and secondary spar cementing fractures, Pakistan, late Miocene.



Figure 3 Groundwater calcite, Death Valley, USA. (a) groundwater calcite veins cutting the Plio-Pleistocene Furnace Creek Formation; (b) core site of DH-11 (Devils Hole), photo by Ray Hoffman, courtesy of Alan Riggs (USGS); (c) vein calcite cutting the lower Furnace Creek Formation; (d) detail of laminated calcite fracture fill (pencil ~15 cm).

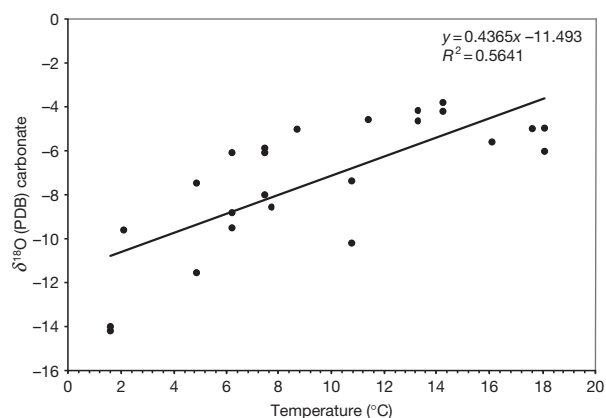


Figure 4 $\delta^{18}\text{O}_{\text{cc}}$ values from modern soils vs. mean annual temperature. The slope (+0.44 ‰/°C) of the best fit line matches the slope predicted by temperature effects of +0.25 ‰/°C to +0.5 ‰/°C.

limestone interactions can alter the primary (surficial) $\delta^{13}\text{C}$ value of dissolved bicarbonate, such as in speleothems (Baker et al., 1997) and groundwater calcites (Devils Hole: Reviewed in Quade (2003)).

Soil Carbonate, Climate, and Vegetation

The $\delta^{13}\text{C}$ value of soil carbonate is largely determined by the proportion of C_3 and C_4 plants growing in the immediate vicinity. Globally, C_3 plants include virtually all trees, almost all shrubs, and those grasses growing under low-light and cool conditions. They average -27‰ in $\delta^{13}\text{C}$ (PDB). C_4 plants occur in 18 plant families; in terms of biomass they comprise mostly grasses and sedges. Of all the grass species, 21% are C_4 , and compared to C_3 grasses, C_4 grasses tend to favor high-light and higher temperature conditions (i.e., warm growing seasons) (Sage, 1999). Globally, C_4 photosynthesis makes up ca. 25% of terrestrial net primary productivity. CAM plants can be found in some ecosystems, but they are not abundant in many soil carbonate-forming settings.

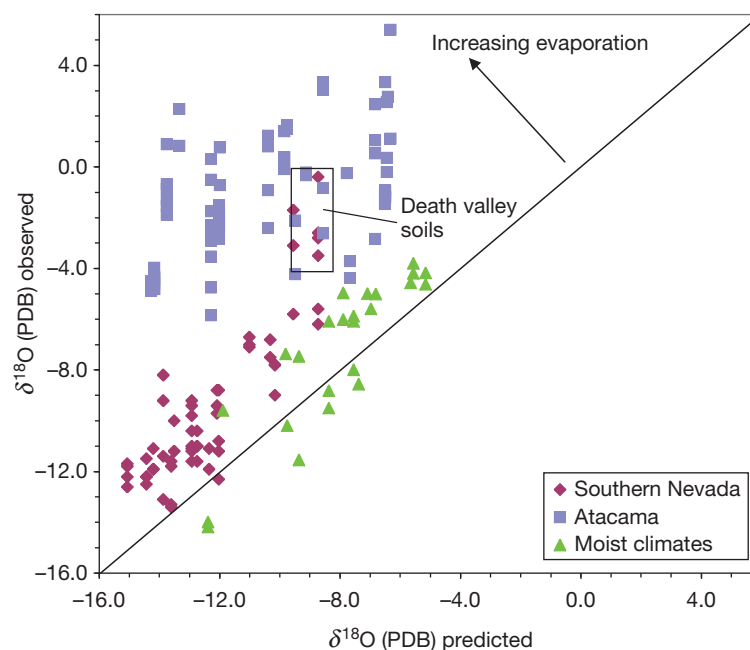


Figure 5 The $\delta^{18}\text{O}_{\text{cc}}$ (observed) (data from Cerling and Quade (1993), Quade et al. (1989b), and unpublished data) vs. predicted $\delta^{18}\text{O}_{\text{cc}}$ values based on local mean annual temperature and $\delta^{18}\text{O}_{\text{p}}$ of local rainfall.

C_3 plants dominate anywhere, forest or shrubs dominate over grass, irrespective of setting, and C_3 grasses dominate over C_4 grasses above mean monthly temperatures for the warmest month of $\sim 22^\circ\text{C}$ (Collatz et al., 1998). The temperature threshold favoring C_4 grasses may have been higher in the past if pCO_2 was higher (Collatz et al., 1998) and may have been lower during glacial episodes when pCO_2 was lower. Other C_4 plants include a few herbaceous species and saltbush, very common in high desert settings. C_4 grasses include several biochemical variants, termed NAD malic enzyme (NAD-ME), NADP malic enzyme (NADP-ME), and phosphoenolpyruvate (PCK). All these variants have distinct $\delta^{13}\text{C}$ values. The proportion of NAD-ME and PCK variant grasses increases with decreasing rainfall and tends to dominate areas where most soil carbonate forms. $\delta^{13}\text{C}$ values of NAD-ME and PCK variant grasses average $-13.8 \pm 1.4\text{‰}$ today, based on one compilation of data ($n=105$) presented in Cerling et al. (2003) for East Africa. To reconstruct past vegetation, these isotopic averages must be increased by 1.5‰ due to fossil fuel burning (the Suess effect; Friedli et al. (1986)). As a result, prior to 1850, pure C_3 biomass in the drier areas is assumed to have averaged $-25.5 \pm 1.3\text{‰}$ ($-27.0 \pm 1.5\text{‰}$) in $\delta^{13}\text{C}$ (VPDB) prior to 1850, and $-11.8 \pm 1.4\text{‰}$ for pure C_4 plants.

Soil Carbonate Formation and the Soil Diffusion Model

Some early models (e.g., Salomans et al. (1978)) of soil carbonate formation invoked the simple stoichiometry of Eqn [1] to suggest that half the C in solution as HCO_3^- derives from plant-derived CO_2 and half from dissolution of antecedent CaCO_3 such as marine limestone. We now know this view to

be incorrect for soils because it fails to take into account isotopic equilibrium considerations in addition to chemical ones. Two key features of soils – abundant CO_2 produced by plants compared to the small amount of C in soil carbonate, and slow dewatering rates in soils – ensure that soil carbonate undergoes continuous, open-system exchange with a large, isotopically unchanging plant CO_2 reservoir. The pore space of soils contains high concentrations (up to 10 000 ppm V) of CO_2 , the result of root respiration and tissue decay of plants. At deep soil levels, this plant CO_2 reservoir is much larger than any potential contribution from the atmosphere (270 ppm V) or from dissolution of limestone from parent material, typically added at the rate of $\sim 0.005 \text{ mmol m}^{-2} \text{ h}^{-1}$. Another critical feature of soils, particularly drier soils, is that they tend to dewater slowly in the days to weeks after a rainfall event. The long dewatering times ($> \text{day}$) of most soils greatly exceed the chemical rate constants of carbonate reactions (seconds to minutes) between carbon species (e.g., Dulinski and Rozanski (1990)). These considerations make it likely that carbon-isotopic equilibrium is maintained between carbon species during soil carbonate formation.

Plant-derived CO_2 – and hence the local proportions of C_3 to C_4 plants growing at the soil surface – can therefore be predicted to determine the $\delta^{13}\text{C}$ value ($\delta^{13}\text{C}_{\text{cc}}$) of carbonate found deep ($> 50 \text{ cm}$) in soils, where plant-derived soil CO_2 is more concentrated. Shallow in soils, plant-derived CO_2 mixes with atmospheric CO_2 , and the proportion of C_3 to C_4 plants cannot be calculated from soil $\delta^{13}\text{C}_{\text{cc}}$ values. For this reason, it is essential to obtain soil carbonate at depth, generally $> 50 \text{ cm}$, as discussed below.

The one-dimensional soil diffusion model presented in Cerling and Quade (1993), and modified in Davidson (1995), provides the quantitative framework for interpretation

of soil-carbonate-isotopic results. These studies show that at steady state, soil CO₂ concentration varies with depth according to:

$$C_s = \frac{k^2 \phi_s^0}{D_s} (1 - e^{-z/k}) + C_a \quad [3]$$

where D_s = the diffusion coefficient of CO₂ in soils (in cm² s⁻¹); C_s = soil CO₂ concentration (in mmol cm⁻³); C_a = atmospheric CO₂ concentration (in mmol cm⁻³); k = the characteristic production depth (cm) of CO₂ in soils, typically 20–25 cm; z = depth (cm); ϕ_s^0 = CO₂ is the production rate at the soil surface ($z=0$) (in mmol cm⁻² s⁻¹); and assuming several boundary conditions, including $C_s = C_a$ at $z=0$ and $dC/dz=0$ at some specified depth L .

Determination of C_s contributes to estimation of the average $\delta^{13}\text{C}$ value of some mixture of C_3 and C_4 plants growing at a soil site ($\delta^{13}\text{C}_{\text{plant}}$), using the approximation presented in Davidson (1995) but solved in terms of $\delta^{13}\text{C}_{\text{plant}}$:

$$\delta^{13}\text{C}_{\text{plant}} = \frac{\delta^{13}\text{C}_s - 4.4 - \delta^{13}\text{C}_a R(z) + 4.4 R(z)}{1.0044(1 - R(z))} \quad [4]$$

where $\delta^{13}\text{C}_s$ and $\delta^{13}\text{C}_a$ are the $\delta^{13}\text{C}$ values of soil CO₂ and air, respectively, and $R(z) = C_a/C_s$ at the depth of sampling z . This equation subsumes the effects of kinetic enrichment in ¹³C of soil CO₂ during diffusion. This sets up a temperature-independent enrichment of 4.2–4.4‰ CO₂ in the soil with respect to the CO₂ flux leaving the soil (Davidson, 1995). This enrichment factor has been observed empirically (Cerling and Quade (1993) and can be calculated using the Stefan–Maxwell equation applied to CO₂ diffusing through air. At isotopic equilibrium, soil carbonate will be enriched in ¹³C with respect to soil CO₂ by 9–12‰, depending on temperature (25–0 °C). The $\delta^{13}\text{C}_s$ values required to solve for $\delta^{13}\text{C}_{\text{plant}}$ in Eqn [4] can be obtained from the $\delta^{13}\text{C}$ value of soil carbonate ($\delta^{13}\text{C}_{\text{cc}}$):

$$\delta^{13}\text{C}_s = \frac{(\delta^{13}\text{C}_{\text{cc}} + 1000)}{\left(\frac{11.98(\pm 0.13) - 0.12(\pm 0.1)T}{1,000} + 1\right)} - 1000 \quad [5]$$

where $11.98(\pm 0.13) - 0.12(\pm 0.1)T$ expresses the temperature-dependent (T = temperature in °C) enrichment factor between calcite and gaseous CO₂ (Romanek et al., 1992). Inspection of this equation shows that the fractionation factor is not strongly sensitive temperature, with a slope of $\sim 0.12\text{‰ } ^\circ\text{C}^{-1}$.

Substitution of Eqn [5] into Eqn [4] allows $\delta^{13}\text{C}_{\text{plant}}$ to be calculated from $\delta^{13}\text{C}_{\text{cc}}$. The fraction C_4 biomass that once grew on the landscape can in turn be calculated from:

$$\text{fraction } C_4 \text{ biomass} = (\delta^{13}\text{C}_{\text{plant}} - \delta^{13}\text{C}_{\text{C3}}) / (\delta^{13}\text{C}_{\text{C4}} - \delta^{13}\text{C}_{\text{C3}}) \quad [6]$$

where $\delta^{13}\text{C}_{\text{C3}}$ and $\delta^{13}\text{C}_{\text{C4}}$ refer to the average, preindustrial carbon isotopic values of C_3 and C_4 plants, respectively.

In summary, where soil respiration rates are high ($>3\text{--}5 \text{ mmol m}^{-2} \text{ hr}^{-1}$) and for low atmospheric CO₂ concentrations ($<500 \text{ ppmV}$), the isotopic composition of soil carbonate should differ from that of coexisting plant cover by about 13–16‰ (gray-shaded zone, Figure 6). This difference ($\Delta_{\text{cc-om}} = \delta^{13}\text{C}_{\text{cc}} - \delta^{13}\text{C}_{\text{om}}$, where the subscript ‘om’ denotes

soil organic matter) involves both equilibrium fractionation between CaCO₃ and CO₂ of 9–12‰, and kinetic diffusion effects of $\sim 4.4\text{‰}$ (Figure 1). This difference is observed in modern soils with moderate to high respiration rates, verifying the general validity of the model (Figure 6). Where respiration rates are lower, such as in deserts, $\Delta_{\text{cc-om}}$ 13–16‰, as observed from a sampling of soils from the Atacama Desert (Figure 6).

Long-Term Records of Climate Change from Nonlacustrine Calcites

Devils Hole

The long isotopic records from Devils Hole archive $\sim 500 \text{ ka}$ of climate change in southern Nevada. The best-known groundwater calcite records in the region come from two cores obtained by scuba divers at Devils Hole. DH-2 spans 50–310 ka BP and DH-11 50–550 ka BP (Figure 7(a); Winograd et al., 1992), based on multiple alpha and TIMS U-series dates from both cores. Neither record has any apparent depositional hiatuses. Dates from the cores lag climatic events by the time required for water to travel from the recharge area to Devils Hole. Thus, dates on laminated calcite (Figure 3) from Devils Hole represent minimum ages (by $<10 \text{ ka}$) for the recorded climate events.

$\delta^{18}\text{O}_{\text{cc}}$ (SMOW) values from Devils Hole fluctuate between +13.05 and +15.75‰ and display the same sawtooth pattern as marine benthic foram $\delta^{18}\text{O}_{\text{c}}$ records over the same timespan (Figure 7(b)). Overall, the similarity in the form of the $\delta^{18}\text{O}_{\text{c}}$ curve from the Devils Hole record to marine (viewed as a proxy for ice volume changes) and ice records such as Vostok (viewed as a proxy for Antarctic air temperature) is striking.

How is the $\delta^{18}\text{O}_{\text{c}}$ record from Devils Hole being interpreted? The Devils Hole flow system is large ($\sim 12,000 \text{ km}^2$) and receives most of its recharge as winter/spring moisture in the Spring and Sheep mountains $>30 \text{ km}$ away. Mixing and additions of water from such a large recharge area dampen out short-term variations, but preserve them at a longer ($\geq 1 \text{ yr}$) scale (Winograd et al., 1992). The temperature of formation of the calcite has likely remained constant due to the deep circulation of water and the large volume of the flow system. The temperature of the water (25–35 °C) is low enough and the volume of water large enough that changes in primary $\delta^{18}\text{O}_{\text{p}}$ due to water–rock interaction are negligible. Thus, Winograd et al. (1992) have suggested that the main determinant of $\delta^{18}\text{O}_{\text{cc}}$ in the Devils Hole records is the value of $\delta^{18}\text{O}_{\text{p}}$ falling mainly in winter and spring on the recharge area to the Devils Hole. The $\delta^{18}\text{O}_{\text{p}}$ value is in turn determined by $\Delta\delta^{18}\text{O}_{\text{seawater}}$ and by changes in the average winter/spring temperature in cloud masses arriving over the recharge area.

The one feature of the Devils Hole record that has sparked great controversy is the dating of glacial terminations II and III at 140 ± 3 and $253 \pm 3 \times 10^3 \text{ yr BP}$ (2σ), respectively. These are minimum age estimates because of the time lag in the Devils Hole record mentioned before. In the orbitally tuned SPECMAP records, these same terminations date to 128 ± 3 and $244 \pm 3 \times 10^3 \text{ yr BP}$, respectively (Figure 7(c)). The offset of $\geq 9 \text{ ka}$ between age estimates for the terminations is important and challenges the widely held view that changes in the Earth’s orbital parameters drive the Ice Ages. For example, the

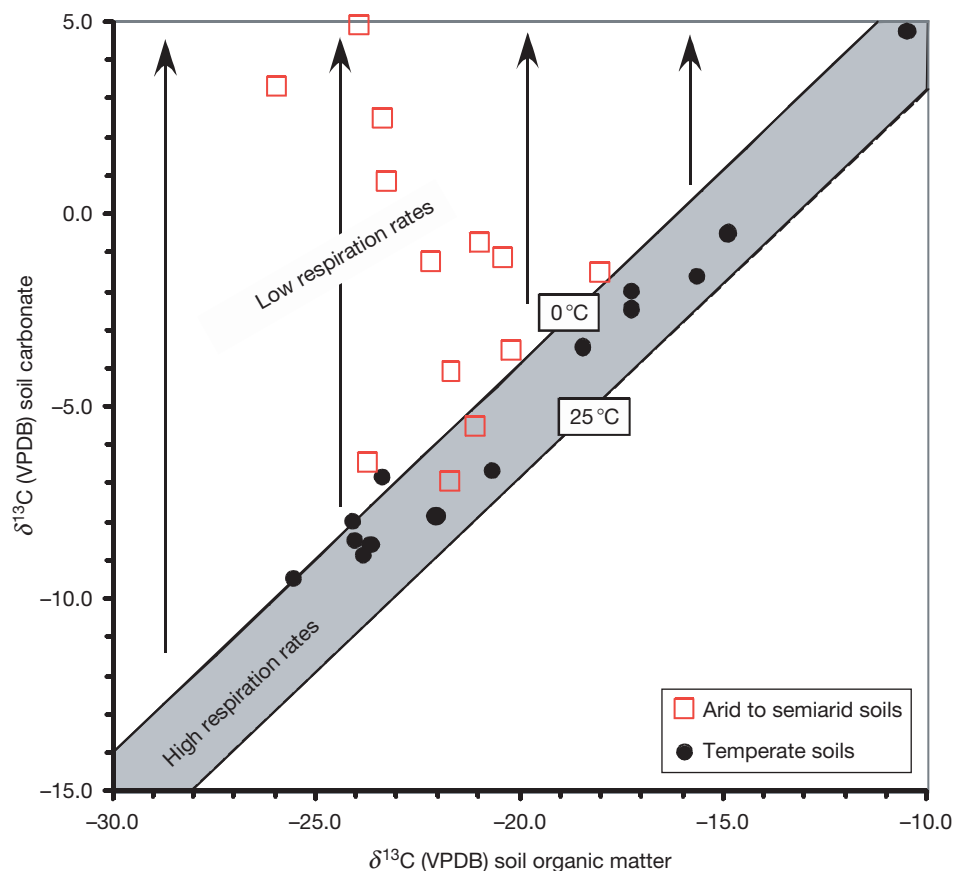


Figure 6 The $\delta^{13}\text{C}$ (VPDB) value of soil organic matter vs. that of coexisting soil carbonate. The shaded zone denotes $\Delta_{\text{cc-om}}$ ($= \delta^{13}\text{C}_{\text{cc}} - \delta^{13}\text{C}_{\text{om}}$) values of 13.4–16.4‰ predicted by the soil diffusion model of Cerling and Quade (1993), and observed in most modern soils with moderate to high respiration rates (solid circles). Most values in which $\Delta_{\text{cc-om}}$ 13–16‰ are produced in soils with low respiration rates, as exemplified by sampling of soils from the Atacama Desert (open squares).

orbitally tuned date of $128 \pm 3 \times 10^3$ yr BP for termination II from SPECMAP coincides with a pronounced summer insolation maximum at $128 \pm 1 \times 10^3$ yr BP for the Northern Hemisphere (Figures 7(b) and (c)), strongly supporting a causal link between maximum summer heating in the Northern Hemisphere and ice-sheet retreat. If, however, the Devils Hole chronology is correct, then the linkage between orbital forcing and waxing and waning of glaciation is not so straightforward.

The central question is whether the Devils Hole $\delta^{18}\text{O}_\text{c}$ record represents a regional versus more global record of temperature change. Winograd et al. (1992) have convincingly argued that local air temperature in the recharge area is the dominant control on the $\delta^{18}\text{O}_\text{c}$ value on the Devils Hole record. However, Herbert et al. (2001) contend that changes in air temperature in the region have not changed in phase with average global air temperature. The basis for this is their paleotemperature estimates spanning multiple glacial and interglacial cycles using unsaturated alkenones from marine sediments off the west coast of North America. The whole issue of how globally representative the Devils Hole record is remains unclear, and awaits, in part, development of well-dated groundwater calcite and speleothem records outside of southwestern North America.

The Global Neogene and Pleistocene Paleosol Record

A number of paleosol isotopic records spanning the Neogene have been developed over the past 15 yr that show dramatic environmental change globally. The shift in $\delta^{18}\text{O}$ values must be related to a global climate change, probably aridification. The shift in $\delta^{13}\text{C}$ values reflects a largely concurrent expansion in C_4 grasses at the expense of C_3 grasses, shrubs, and trees. Within the Pleistocene, however, change in C_4 biomass has been conservative globally except in East Africa, where grasslands expanded significantly, possibly influencing the evolution of the genus *Homo*.

$\delta^{18}\text{O}$ record

Most of the oxygen-isotope records from soil carbonates show gradual (Argentina, Latorre et al. (1997); Greece/Turkey, Quade et al. (1994)) to sharp (Pakistan, Quade et al. (1989a); Nepal, Quade et al. (1995); Africa, Cerling (1992), Quade and Levin (in review)) increases in $\delta^{18}\text{O}_{\text{cc}}$ values starting in the late Miocene to early Pliocene (8–4 Ma) (Figure 8). Only the North American (Fox and Koch, 2004) soil carbonate record shows no increases over the late Neogene.

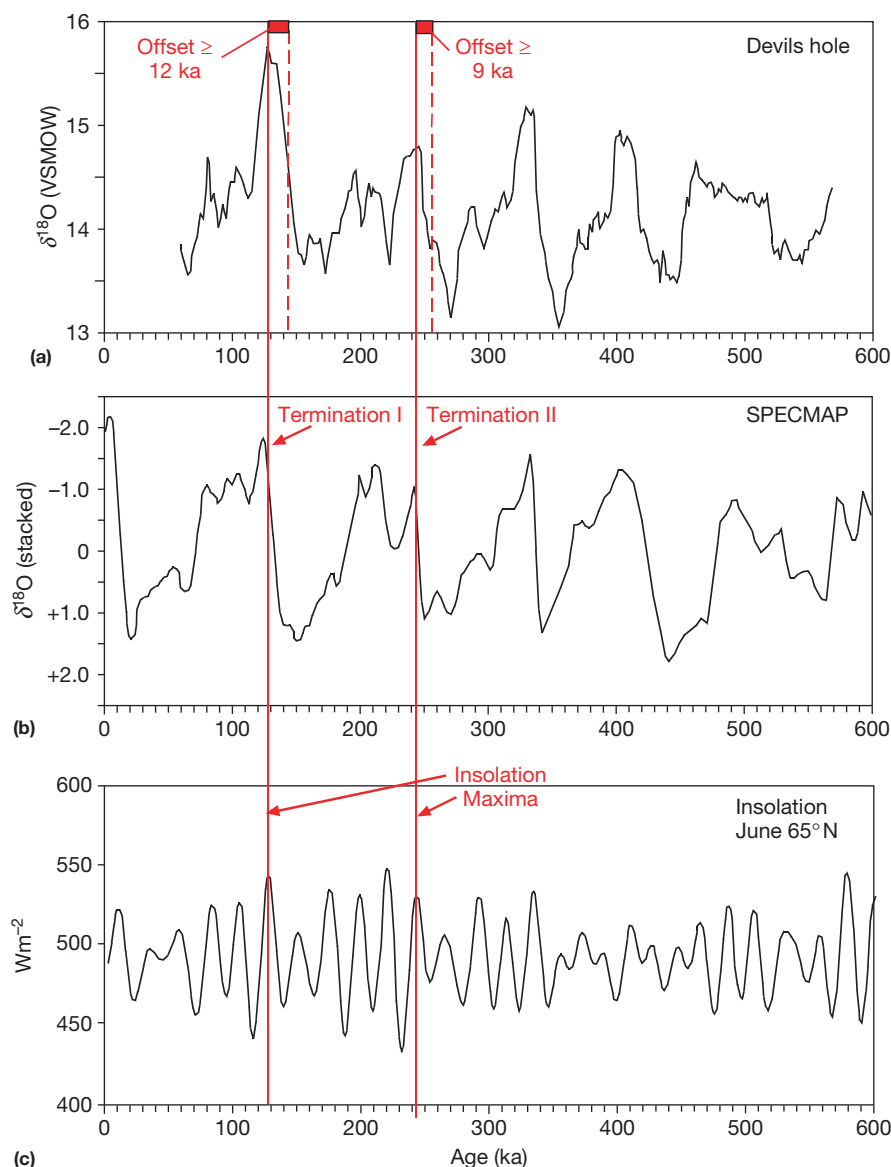


Figure 7 (a) $\delta^{18}\text{O}$ (SMOW) values of a calcite from Devils Hole (Winograd et al., 1992); (b) stacked benthic forams from SPECMAP (Imbrie et al., 1984); and (c) June insolation in W m^{-2} for 60°N (Berger and Loutre, 1991).

The largely global increase in $\delta^{18}\text{O}_{\text{cc}}$ values must reflect important changes in global climate. The shift is too large (3–5‰) and the wrong sign to be explained by temperature declines through the Neogene. As was discussed previously the slope of the theoretical and observed temperature– $\delta^{18}\text{O}_{\text{cc}}$ relationship in modern ‘temperate’ soil carbonate is $+0.44\text{‰ C}^{-1}$. The increase in $\delta^{18}\text{O}_{\text{seawater}}$ is the right sign but far too small ($<1.0\text{‰}$) to explain the increases in $\delta^{18}\text{O}_{\text{cc}}$ values. At mid- to high latitudes, this leaves increased soil water evaporation as the dominant cause for increases in $\delta^{18}\text{O}_{\text{cc}}$ values. At lower latitudes, decreases in rainfall amount also contributed to the increase in $\delta^{18}\text{O}_{\text{cc}}$ values, as well as increased soil water evaporation (Quade and Levin, in review).

$\delta^{13}\text{C}$ record

Most of the carbon-isotope records from soil carbonate show gradual (Argentina, Latorre et al. (1997)) to sharp (Pakistan,

Quade et al. (1989a); Nepal, Quade et al. (1995); Africa, Cerling et al. (1992), Quade and Levin (in review)) increases in $\delta^{13}\text{C}$ values starting in the late Miocene (8–6 Ma) (Figure 9). Others show more modest increases (North America, Fox and Koch (2004)), starting later, in the early Pliocene. One record – the only long record from a Mediterranean (winter rainfall-dominated) climate – shows no appreciable shift through the Neogene (Quade et al., 1994) (Figure 9).

Most of the increase in $\delta^{13}\text{C}$ values can be attributed to increases in C_4 grasses, although the contribution by lower respiration rates to higher $\delta^{13}\text{C}$ values has not been unequivocally excluded in some records (Latorre et al., 1997; Fox and Koch, 2004). The lack of change in $\delta^{13}\text{C}$ values from the eastern Mediterranean record accords with the lack of C_4 plants in that region today. In some areas, the ecologic shift involved a change in life form: C_4 grasses replaced C_3 forest (e.g., Nepal, Quade et al. (1995)). In others, photosynthetic pathway but not life form

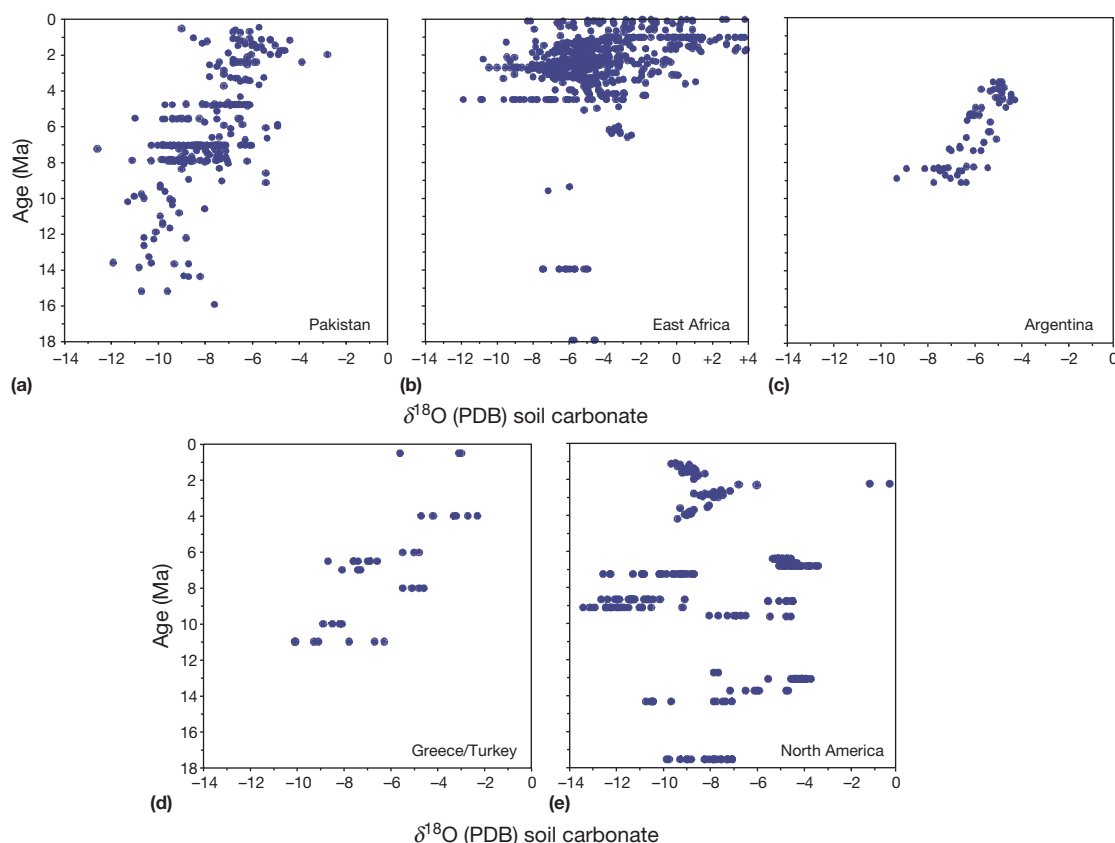


Figure 8 Global Neogene trends in $\delta^{18}\text{O}_{\text{oc}}$ values from (a) Pakistan (Quade et al., 1989a and unpublished data); (b) East Africa (Quade and Levin, in review); (c) Argentina (Latorre et al., 1997); (d) Greece and Turkey (Quade et al., 1994); (e) North America (Fox and Koch, 2004; data courtesy of D. Fox).

changed: C_4 grasses apparently replaced a substantial fraction of C_3 grasses (e.g., North America: Fox and Koch (2004)). Cerling et al. (1997) observed similar increases over the late Neogene in the diet of large herbivores, attributed to increases in the dietary intake of C_4 grasses.

The cause of this dramatic and largely global change in plant cover remains unclear. The C_4 expansion might be explained by a decline in $p\text{CO}_2$ globally, one of the environmental conditions that favors C_4 metabolism (Cerling et al., 1997). However, there is no evidence of decline in $p\text{CO}_2$ over this period (Pagani et al., 1999). Increased moisture stress, like $p\text{CO}_2$ decline, would also favor C_4 plant expansion. This explanation is supported by the clear increases over the same timespan in $\delta^{18}\text{O}_{\text{oc}}$ values, a trend we have already linked to aridification. The main weakness of this explanation is that it does not explain why C_4 metabolism did not expand earlier. Although more prevalent in the late Neogene, semiarid environments must have been present in the pre-late Miocene. C_4 plants are favored by fire, and an increase in fire frequency would support the expansion of C_4 biomass. This, too, is a hypothesis worth exploring.

The global carbon-isotope record from paleosols shows little change over the Pleistocene. A significant exception to this is the sharp increase in $\delta^{13}\text{C}_{\text{cc}}$ values observed from East Africa (Figure 9(b); summarized in Quade and Levin (in review)). At many study sites, $\delta^{13}\text{C}_{\text{cc}}$ values increase for the first time above -2‰ during the Pleistocene, denoting the

appearance of savanna grasslands ($>80\%$ C_4 biomass) so characteristic of the region today. The pattern of expansion is uneven temporally and geographically, commencing as early as the late Miocene at Lothagam, but much later, in the 2.7–1.0 Ma timespan, at most other sites. Many researchers draw the possible connection between this expansion of savanna ecosystems and the development of the genus *Homo* in Africa starting sometime in the latest Pliocene (see review of Klein (1995)). The Pleistocene-age increase in $\delta^{13}\text{C}$ values in East Africa is embedded in a local, longer-term trend toward higher $\delta^{18}\text{O}$ values, again a pattern we attribute to aridification.

The impact on regional vegetation and climate during shifts from glaciations to interglaciations is also apparent in a number of isotopic records from paleosols from the Pleistocene. As just two examples, glacial-age soil carbonates from the Wind River Mountains of Wyoming, USA (Amundson et al., 1996) yielded $\delta^{18}\text{O}_{\text{oc}}$ values significantly higher than modern soil carbonate values. This was interpreted to be the result of a combination of colder temperatures and the increased contribution of high $\delta^{18}\text{O}_{\text{oc}}$ waters from the Gulf of Mexico. In the Great Basin, western USA, $\delta^{13}\text{C}_{\text{cc}}$ values decreased during glaciations, as C_3 -dominated vegetation at higher elevations today was displaced downward (Quade et al., 1990). Recent advances in $^{230}\text{Th}/\text{U}$ dating of individual laminae in desert soil carbonates (e.g. Sharp et al., 2003) should lead to much more extensive use of soil isotopic archives in shorter-term (10^4 – 10^5 yr) paleoclimate reconstruction.

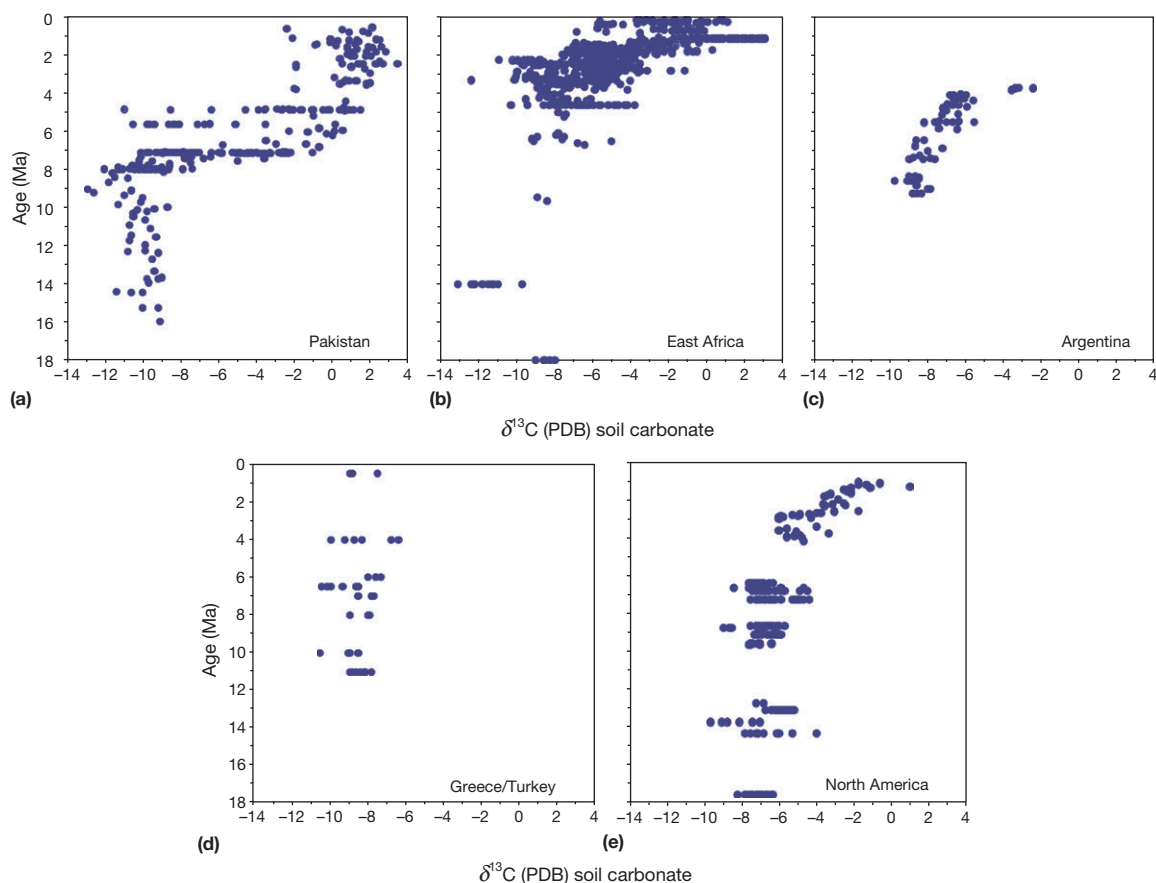


Figure 9 Global Neogene trends in $\delta^{13}\text{C}_{\text{cc}}$ values from (a) Pakistan (Quade et al., 1989a, and unpublished data); (b) East Africa (Quade and Levin, in review); (c) Argentina (Latorre et al., 1997); (d) Greece and Turkey (Quade et al., 1994); and (e) North America (Fox and Koch, 2004; data courtesy of D. Fox).

See also: Carbonate Stable Isotopes: Lake Sediments; Speleothems. Paleosols and Wind-Blown Sediments: Soil Morphology in Quaternary Studies.

References

- Amundson R, Chadwick O, Kendall C, Wang Y, and DeNiro M (1996) Isotopic evidence for shifts in atmospheric circulation during the Quaternary in mid-North America. *Geology* 24: 23–26.
- Baker A, Ito E, Smart PL, and McEwan RG (1997) Elevated and variable values of ^{13}C in speleothems in a British cave system. *Chemical Geology* 136: 263–270.
- Berger A and Loutre MF (1991) Insolation values for the climate of the last 10 million years. *Quaternary Science Reviews* 10(4): 297–317.
- Cerling TE (1992) Development of grasslands and savannas in east Africa during the Neogene. *Palaeogeography, Palaeoclimatology, Palaeoecology* 97: 241–247.
- Cerling TE, Harris JM, MacFadden BJ, et al. (1997) Global vegetation change through the Miocene/Pliocene boundary. *Nature* 389: 153–157.
- Cerling TE, Harris JM, and Passey BH (2003) Diets of east African Bovidae based on stable isotope analysis. *Journal of Mammalogy* 84(2): 456–470.
- Cerling TE and Quade J (1993) Stable carbon and oxygen isotopes in soil carbonates. In: Swart P, McKenzie JA, and Lohman KC (eds.) *Continental Indicators of Climate, Proceedings of Chapman Conference, Jackson Hole, Wyoming, American Geophysical Union Monograph* 78: 217–231.
- Collatz G, Berry J, and Clark J (1998) Effects of climate and atmospheric CO_2 partial pressure on the global distribution of C_4 grasses: Present, past, and future. *Oecologia* 114: 441–454.
- Davidson GR (1995) The stable isotopic composition and measurement of carbon in soil CO_2 . *Geochimica et Cosmochimica Acta* 59(12): 2485–2489.
- Dulinski M and Rozanski K (1990) Formation of $^{13}\text{C}/^{12}\text{C}$ ratios in speleothems: A semi-dynamic model. *Radiocarbon* 32: 7–16.
- Fantidas J and Ehrlert DH (1970) Variations of the carbon and oxygen isotopic composition in stalagmites and stalactites; Evidence of non-equilibrium isotopic effects. *Earth and Planetary Science Letters* 10: 136–144.
- Fox D and Koch PL (2004) Carbon and oxygen isotopic variability in Neogene paleosol carbonates: Constraints on the evolution of C_4 grasslands. *Palaeogeography, Palaeoclimatology, Palaeoecology* 207: 305–329.
- Freidli H, Lotscher H, Oeschger H, Siegenthaler U, and Stauffer B (1986) Ice core record of $^{13}\text{C}/^{12}\text{C}$ ratio of atmospheric CO_2 in the past two centuries. *Nature* 324: 237–238.
- Hendy CH (1971) The isotope geochemistry of speleothems. In: *The calculation of different modes of formation on the isotopic composition of speleothems and their applicability as paleoclimatic indicators*, *Geochimica et Cosmochimica Acta* 35: 801–824.
- Herbert TD, Schuffert JD, Andreasen D, et al. (2001) Collapse of the California current during glacial maxima linked to climate change in land. *Science* 293: 71–76.
- Imbrie J, Hays JD, Martinson DG, et al. (1984) In: Berger AL, Imbrie J, Hays J, Kukla G, and Saltzman B (eds.) *Milankovitch and Climate, Part 1*, pp. 269–305. Dordrecht: Reidel.
- Kim S and O'Neil JR (1997) Equilibrium and nonequilibrium oxygen isotope effects in synthetic carbonates. *Geochimica et Cosmochimica Acta* 61(16): 3461–3475.
- Klein R (1995) *The Human Career*. Chicago: University of Chicago Press.
- Latorre C, Quade J, and McIntosh WC (1997) The expansion of C_4 grasses and global change in the late Miocene: Stable isotope evidence from the Americas. *Earth and Planetary Science Letters* 146: 83–96.

- Ludvigson GA, Gonzalez LA, Metzger RA, et al. (2002) Meteoric sphaeroiderite lines and their use for paleohydrology and paleoclimatology. *Geology* 26: 1039–1042.
- Pagani M, Freeman KH, and Arthur MA (1999) Late Miocene atmospheric CO₂ concentrations and expansion of C₄ grasses. *Nature* 285: 876–879.
- Quade J (2003) Isotopic records from ground-water and cave speleothem calcite in North America. In: Gillespie S, Porter C, and Atwater BF (eds.) *The Quaternary Period in the United States*, pp. 205–220. New York: Elsevier Science.
- Quade J, Cater JML, Ojha TP, Adam J, and Harrison TM (1995) Dramatic carbon and oxygen isotopic shift in paleosols from Nepal and late Miocene environmental change across the northern Indian sub-continent. *Geological Society of America Bulletin* 107: 1381–1397.
- Quade J, Cerling TE, and Bowman JR (1989a) Development of the Asian Monsoon revealed by marked ecological shift in the latest Miocene in northern Pakistan. *Nature* 342: 163–166.
- Quade J, Cerling TE, and Bowman JR (1989b) Systematic variation in the carbon and oxygen isotopic composition of Holocene soil carbonate along elevation transects in the southern Great Basin, USA. *Geological Society of America Bulletin* 101: 464–475.
- Quade J, Cerling TE, and Bowman JR (1990) Stable isotopic evidence for a pedogenic origin of carbonates in Trench 14 near Yucca Mountain, Nevada. *Science* 250: 1549–1552.
- Quade J, Solounias N, and Cerling TE (1994) Stable isotopic evidence from paleosol carbonates and fossil teeth in Greece for forest or woodlands over the past 11Ma. *Palaeogeography, Palaeoclimatology, Palaeoecology* 108: 41–53.
- Romanek CS, Grossman ET, and Morse JW (1992) Carbon isotopic fractionation in synthetic aragonite and calcite: Effects of temperature and precipitation rate. *Geochimica et Cosmochimica Acta* 56: 419–430.
- Rozanski K, Araguas-Araguas L, and Gonfiantini R (1993) Isotopic patterns in modern global precipitation. In: Swart P, McKenzie JA, and Lohman KC (eds.) *Continental Indicators of Climate, Proceedings of Chapman Conference, Jackson Hole, Wyoming, American Geophysical Union Monograph* 78, pp. 1–36.
- Sage RF (1999) Why C₄ photosynthesis? In: Sage RF and Monson RK (eds.) *C4 Plant Biology*, pp. 3–16. San Diego: Academic Press.
- Salomans W, Goudie A, and Mook WG (1978) Isotopic composition of calcrete deposits from Europe, Africa and India. *Earth Surface Processes* 3: 43–57.
- Sharp WD, Ludwig KR, Chadwick O, Amundson R, and Glaser LL (2003) Dating fluvial terraces by ²³⁰Th/U on pedogenic carbonate, Wind River Basin, Wyoming. *Quaternary Research* 59: 139–150.
- Winograd IJ, Coplen TB, Landwehr JM, et al. (1992) Continuous 500,000-year climate record from Devils Hole, Nevada. *Science* 258: 255–260.

Lake Sediments

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Introduction

Lakes are widely distributed across the continents and are found in climatic zones ranging from Antarctica to the tropics and from moist to relatively arid environments. Lacustrine sediments are important archives of climatic and environmental change. Their wide spatial distribution enables the study of the regional impact of global climate change in a variety of climatic regimes with high geographic and temporal resolution. Lakes precipitating authigenic carbonate or containing carbonate in the form of biogenic shells have been subjects of considerable research since the pioneering studies in the early 1970s (e.g., [Stuiver, 1970](#)). Because of the relative ease of analysis and the development of automated analytical systems, isotope studies of bulk lacustrine carbonates have become common. However, the extraction of quantitative climatic and paleoenvironmental information from bulk carbonate records is not straightforward or simple. The main difficulties in interpreting oxygen isotope records are related to three main factors:

1. Bulk carbonates in lakes represent a mixture of particles with different origins;
2. The isotopic composition of the precipitated calcium carbonate is dependent on the water temperature and the isotopic composition of the lake water, and, thus, on the hydrological balance of a lake; and
3. Biological activity in the lake, by influencing the carbonate chemistry, may influence the mechanisms, water depth and timing of carbonate precipitation. In particular, if the hydrology of the lake is not well understood, a quantitative reconstruction of climate is prone to large errors.

Global Spatial- and Age-Distribution of Carbonate-Precipitating Lakes

Carbonate-precipitating lakes and paleolakes are found on all continents, at various latitudes and altitudes. The main control on the presence of carbonate in a lake is the composition of the bedrock. Significant calcium carbonate precipitation in the water column is observed only in lakes that have a large part of the catchment consisting of calcium carbonate-rich rocks, which supply most of the calcium and carbonate ions through weathering reactions. Lakes within catchments dominated by silicate rocks are normally devoid of carbonates. Calcium carbonate supersaturation in the bottom waters and the sediments is also necessary to avoid postdepositional dissolution of calcite, which is promoted by CO₂ production related to organic matter decomposition.

The spatial distribution and age of lakes is dependent on their origin. Lakes of glacial origin are very common in formerly glaciated areas of Europe, Asia, and the Americas.

Although some lakes may have originated in the Tertiary (e.g., the lakes of the insubric region in northern Italy and Switzerland) most of the sediments preserved, or accessible to sampling, in these lakes are younger than the Last Glacial Maximum, about 18 000 years before present. Older sediments were either removed by glacial erosion during the glacial advances or are compacted and present at sediment depths that are reachable only by deep drilling. Lakes of tectonic origin are globally less numerous than glacial lakes, but they are a prominent feature for their size and permanence, and have provided some of the longest records of continental climate. Lakes of tectonic origin include all of the major lakes in Africa; the Aral and Caspian seas, Lake Baikal, Lake Van and the Dead Sea (and its paleolake equivalent Lake Lisan) in Eurasia; Lake Titicaca in South America; the Great Salt Lake, Utah, and its paleolake equivalent Lake Bonneville and Pyramid Lake, Nevada in North America. Crater lakes, which also have provided important long paleoclimatic records, are often devoid of carbonates due to the chemistry of the waters.

Sampling of Lacustrine Sediments

Paleoclimatic studies on lacustrine sediments are mostly carried out on cores obtained by various platforms and coring devices ranging from handheld gravity cores to complex drilling platforms. For a review of coring devices, see [Glew et al. \(2001\)](#). Sedimentation rates, and thus the length of cores required for long paleoclimatic records and the sampling resolution achievable, vary dramatically in lakes. In smaller peri-alpine lakes of glacial origin, sediment accumulation rates are in the range of 0.5–5 mm year⁻¹. Typical length of cores are 1–2 m for simple handheld gravity cores, which allow studies of the last few hundred to a few thousand years in very small lakes with low sedimentation rates. For the recovery of longer sediment sequences of up to 10–20 m, larger platform systems have to be used. A large drilling platform that is managed by the International Continental Drilling Program (<http://www.icdp-online.org>) has been operational since 1998 and has allowed the recovery of climate records from lakes reaching back 1 My.

Nature of Carbonate in Lacustrine Sediments

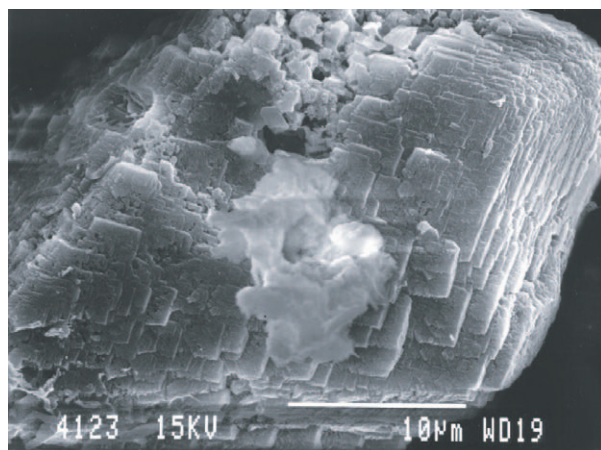
The carbonate fraction of lacustrine sediments can contain particles of multiple *in situ* origins:

1. an inorganically precipitated carbonate,
2. a biogenic fraction formed in the water column, and
3. a fraction which can be formed by diagenesis after the deposition of the sediments.

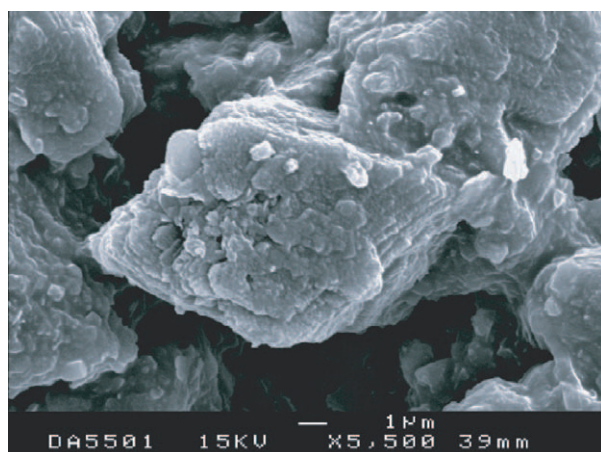
In addition, because carbonate-precipitating, hard-water lakes are found in catchments composed, at least partially, of carbonate bedrocks, some fraction of the carbonate will be of detrital origin. For a quantitative interpretation of the isotopic signal, the origin of the carbonate particles has to be determined by X-ray diffractometry and by petrographic examination with light- or electron microscopy. Only the authigenic particles (Figure 1) formed in the surface waters carry relevant isotopic information for paleoclimatic reconstruction, and, thus, they should be the only or at least the dominant fraction of the sediments used to obtain meaningful information from bulk carbonate isotope analyses.

Carbonate Precipitation in Mid- to High-Latitude, Hard-Water Lakes

Carbonate precipitation in deep, hydrologically open, hard-water lakes is induced by increases in pH and carbonate ion concentration related to the utilization of aqueous CO_2 by photosynthetic organisms, which increases calcium carbonate



Lake Albano, Italy



Lake Butrinti, Albania

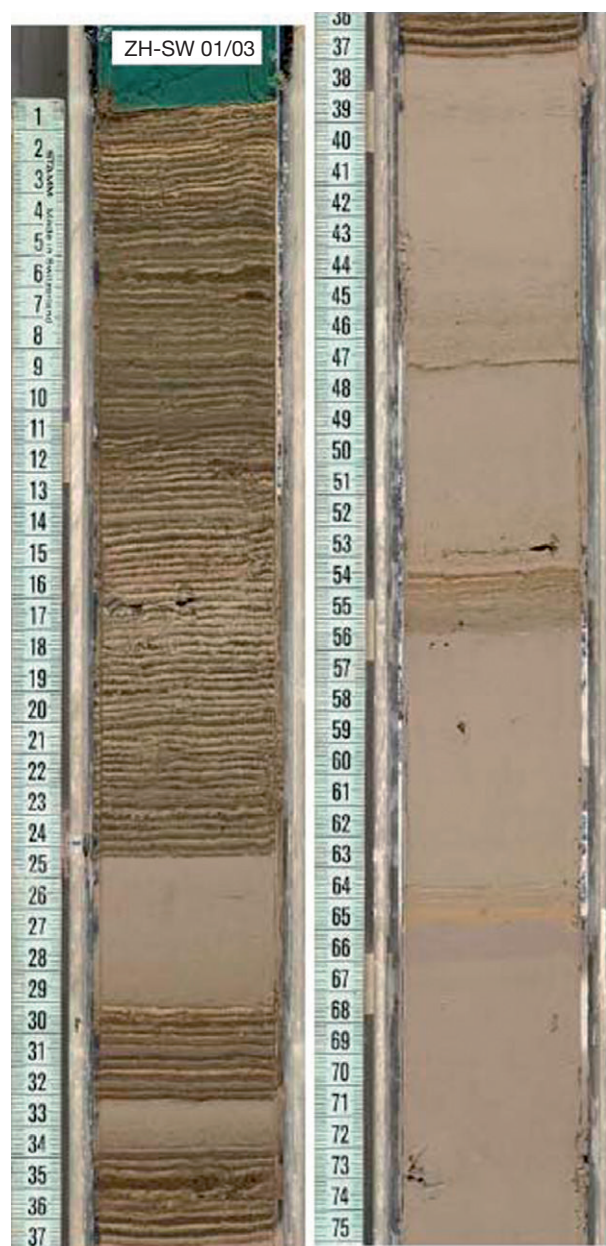
Figure 1 Scanning electron photomicrographs of authigenic calcite crystals from Lake Albano, Italy, and Lake Butrinti, Albania. Note the rhombohedral faces and the crystal-growth steps. (Photos Courtesy of D. Ariztegui, University of Geneva).

supersaturation (Kelts and Hsu, 1978). An active role of cyanobacterial picoplankton in carbonate precipitation has been postulated by Thompson et al. (1997) and Ditttrich et al. (2004), and also has to be considered for the understanding of carbon and oxygen isotope geochemistry of carbonates. Carbonate precipitation in mid- to high-latitude lakes is commonly limited to a period comprised between early spring and late autumn, when higher temperatures and light levels stimulate algal production. This period follows the annual recharging of the surface waters with essential nutrients as a result of the winter overturn of the water column. The isotopic composition of carbonate precipitate, therefore, predominantly reflects spring and summer conditions, whereas the sedimentation in the late fall and winter months consists predominantly of clastic material and organic matter. If the bottom waters of a lake are at least temporarily anoxic, the rhythmic sedimentation induced by the annual productivity cycle can be preserved in the sediment in the form of chemical varves of alternating white summer and dark winter layers (Figure 2). The presence of an annual lamination allows the accurate dating of sedimentary sequences by varve-counting. In addition, it facilitates the acquisition of paleoenvironmental records at annual or even at semiannual resolution (e.g., Teranes et al., 1999b). Even in this case, however, it has to be kept in mind that the residence time of the water in the catchment area and the lake itself is the limiting factor for the true temporal resolution of a sedimentary record.

The precipitation of calcium carbonate occurs close to the surface of the lake, where the biological activity is most intense. The time in the season and the water depth where calcium carbonate precipitates may vary from year to year, potentially creating spurious variations in the isotopic signatures of the carbonates. Large temperature gradients of up to 15°C in the top 20 m of the water column are common, and the temperature range at the lake surface between spring and summer can reach $10\text{--}15^\circ$ (e.g., Livingstone and Lotter, 1999; Teranes et al., 1999a). Because the precipitation of carbonate is dependent on the biological activity, oxygen isotope records from lakes, which have undergone strong changes in trophic level and productivity, have to be considered with caution. Possible effects of eutrophication on oxygen isotope records have been critically discussed, for example, Gat and Lister (1995). Fronval et al. (1995) have shown that carbonate precipitation can occur in isotopic disequilibrium with the water under strongly eutrophic conditions, leading to a negative isotope shift in $\delta^{18}\text{O}$. Teranes et al. (1999a), on the other hand, have shown in eutrophic Baldeggersee, Switzerland, that, although isotopic disequilibrium is observed during events of rapid calcite precipitation, a large part of the calcite can precipitate close to equilibrium.

Carbonate Precipitation in Saline and Closed-Basin Lakes

In hydrologically closed lakes and in arid regions, carbonates are precipitated mainly as a consequence of a negative water balance. In these lakes, the effects of biological activity are less important because the main driving force for carbonate precipitation is the increase in Ca^{2+} and CO_3^{2-} concentrations related to evaporation. In contrast to hard-water lakes where the mineralogy of the authigenic carbonates is dominantly low



Bernasconi & McKenzie

Figure 2 Photo of a gravity core collected from 130 m depth in Lake Zürich, Switzerland. For the last 100 years, a clear annual lamination is visible in the sediments. The light layers are composed mainly of authigenic calcite formed in the water column in spring to late summer. The dark layers are rich in diatoms, organic matter and clays and represent autumn and winter sedimentation. The thick layers at 25–29 cm and 33–34 cm are fine grained redeposited sediments derived from slope instabilities. The bottom of the core shows homogeneous carbonate-rich sediment deposited before significant eutrophication led to the preservation of the annual lamination.

magnesium calcite, a larger variety of carbonate minerals are found in saline lakes, including high magnesium calcite, aragonite, and dolomite, as well as more soluble Na–Ca carbonates such as gaylussite and trona (see e.g., Talbot, 1990). The mineralogy is related to the degree of salinity of the water

and provides an independent way to estimate evaporation levels. For example, in Lake Van, Turkey, the Mg content of calcite increases with increasing salinity and can be used in conjunction with isotopes to constrain evaporation/precipitation ratios (Wick et al., 2003). Isotope records from strongly evaporative soda lakes are relatively rare, due mainly to concerns related to recrystallization of the very soluble carbonates. However, there is potential to extract useful climatic information even from such highly soluble carbonates (Bernasconi et al., 1997).

Isotopic Composition of Lacustrine Carbonate

In paleoclimatic studies, the stratigraphic variation of the oxygen isotope composition of carbonates is used to reconstruct changes in temperature and/or oxygen isotope composition of the lake water. Both these parameters depend on local climate variations and on the hydrologic characteristics of the lake. Before quantitative inferences on climate can be made, the main controls on the oxygen isotopic composition of the lake water have to be evaluated, as each lake shows a different response to climatic forcing. When the hydrology of the lake is known, more precise climate information can be extracted from the oxygen isotope record.

The carbon isotope composition of carbonates does not contain direct climatic information, mainly because carbon isotope fractionation between dissolved inorganic carbon and calcite is only very weakly temperature dependent. The $\delta^{13}\text{C}$ value of carbonates mainly reflects changes in the inputs of dissolved inorganic carbon from the catchment area and exchange with the atmosphere. Moreover, it is strongly influenced by biological activity within the lake, leading to the development of surface-to-bottom water carbon-isotope gradient, similar to that found in the marine environment (McKenzie, 1985). Essentially, there is a transfer of ^{12}C from the surface to the deeper waters by a photosynthesis–respiration pathway, and the degree of this photosynthetic activity is reflected in the $\delta^{13}\text{C}$ value of the dissolved inorganic carbon and incorporated into the carbonate precipitate (Figure 3). Thus, the carbon isotope record archived in lacustrine carbonates documents changes in the paleoproductivity of lakes and, because of the intrinsic link between the inorganic carbonate precipitate and the associated organic carbon, information about the dominant photosynthetic mechanism (CO_2 vs. HCO_3^- utilization) can be gained (Hollander and McKenzie, 1991; Teranes and Bernasconi, 2005).

In addition, the carbon isotope composition of the carbonates provides important information on changes in the catchment and within the lakes, which can be used to better interpret the oxygen isotope signal. For example, the $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values for lacustrine carbonates from many closed-basin lakes have been found to covary (Talbot, 1990 and references therein). As this $\delta^{13}\text{C}$ – $\delta^{18}\text{O}$ covariance is absent or weak in hydrological open lakes, it can be a useful indicator to interpret the open/closed history of a paleo-lake system. Li and Ku (1997) have evaluated the causal mechanism for the observed $\delta^{13}\text{C}$ – $\delta^{18}\text{O}$ covariance in the closed-basin lake setting and attribute its occurrence and varying intensity to a number

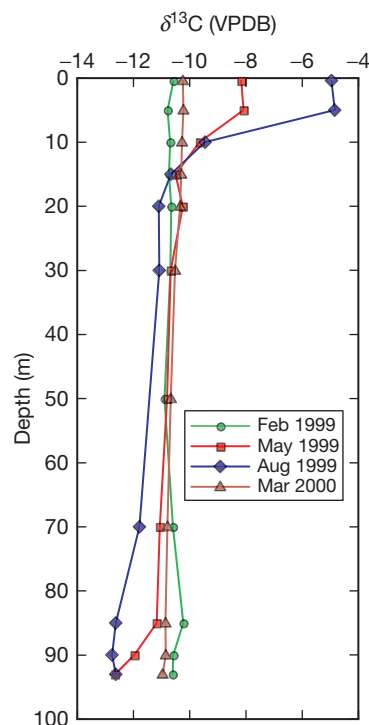


Figure 3 Carbon isotope profiles in the water column of lake Lugano in different months. Note the constant composition in winter after the water column has mixed during spring and summer there is a progressive enrichment in ^{13}C in the photic zone due to the preferential assimilation of ^{12}C by the algae. In contrast, in the bottom waters the oxidation of organic matter leads to a decrease in $\delta^{13}\text{C}$ of the dissolved inorganic carbon. Data from Lehmann et al. (2004).

of factors, including size of lake, duration of hydrological closure, vapor exchange, and lake productivity and alkalinity.

Oxygen Isotope Composition of Lake Water and Its Relation to Climate and Local Hydrology

Lakes are dynamic systems linked to the hydrological cycle via surface and subsurface inflow and outflow and via direct precipitation and evaporation from the lake surface. The factors controlling the oxygen isotope composition of lake waters are schematically summarized in Figure 4. Mathematically, the isotopic water balance of a lake can be expressed as

$$V_{\text{lake}} \delta^{18}\text{O}_{\text{lake}} = \sum I \delta^{18}\text{O}_I + P \delta^{18}\text{O}_P - \sum O \delta^{18}\text{O}_{\text{lake}} - E \delta^{18}\text{O}_E$$

where V is the volume of the lake, $\delta^{18}\text{O}$ is the oxygen isotopic composition of the various components, I is the inputs (rivers and surface runoff, groundwater input), P is direct precipitation on the lake, O is the output (surface and subsurface output), and E is the evaporation. Note that the surface and groundwater outputs have the same isotopic composition as the lake water. The only output component which imparts a fractionation on the isotopic composition of waters is evaporation. Thus, this factor is the most important parameter modifying the oxygen isotope composition of a lake besides the mixing of the various inputs. In hydrologically open basins, the $\delta^{18}\text{O}$ of lake water depends predominantly on the relative

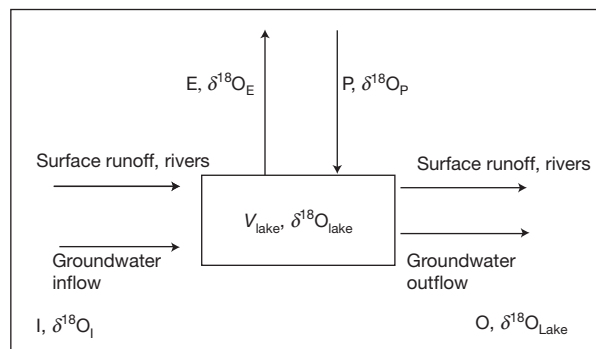


Figure 4 Schematic diagram of the main factors controlling the isotope composition of lake waters.

fluxes and $\delta^{18}\text{O}$ values of the inputs with more or less important modification induced by evaporation processes. In closed lake basins, where there is no surface or subsurface output ($O=0$), the water balance of the lake and its isotopic composition are determined by the difference between evaporation and precipitation in the lake and the inputs in the catchment.

The $\delta^{18}\text{O}$ of the lake water, as it is recorded by sedimentary carbonates formed in the surface waters, is additionally influenced by the pattern of vertical mixing of the lake. In monomictic lakes during the stratification period in the summer when most of the carbonate precipitates, the surface waters becomes enriched in ^{18}O due to the preferential removal of ^{16}O with evaporation. Thus, the carbonate will not necessarily reflect the bulk oxygen isotope composition of the lake but will show a variably enriched composition which is dependent on the evaporation intensity (Figure 5). This enrichment can be significant (up to a few ‰) and introduces an additional uncertainty factor in the interpretation of carbonate oxygen isotope records.

The oxygen isotopic composition of precipitation is dependent on the air temperature (Dansgaard, 1964; Rozanski et al., 1993), the source of moisture (i.e., large scale atmospheric circulation), and the seasonal distribution of precipitation (Rozanski et al., 1993). Dansgaard (1964) first described the temperature dependence of the oxygen isotope composition of precipitation. He noted a global relationship of $0.69\text{‰}^\circ\text{C}^{-1}$ for mid- and high northern latitudes, with stronger variations depending on the location. The more recent compilation of Rozanski et al. (1993) provides an excellent updated analysis of the IAEA/GNIP database (International Atomic Energy Agency Global Network of Isotopes in Precipitation <http://www-naweb.iaea.org/napc/ih>). This database is an irreplaceable tool for the calibration of the modern relationship at many locations around the world, which, in turn, can be used for the calibration and interpretation of the paleo-records of precipitation.

In a recent study, Kohn and Welker (2005) have reanalyzed data from four stations in the United States and concluded that the oxygen isotope-mean temperature correlation can be improved, showing a slope always close to $0.55\text{‰}^\circ\text{C}^{-1}$, if the isotope data are correlated with the average temperature on the days when precipitation occurs, rather than the overall average temperature. This effect can be easily understood if one considers that, in summer, the average temperature on

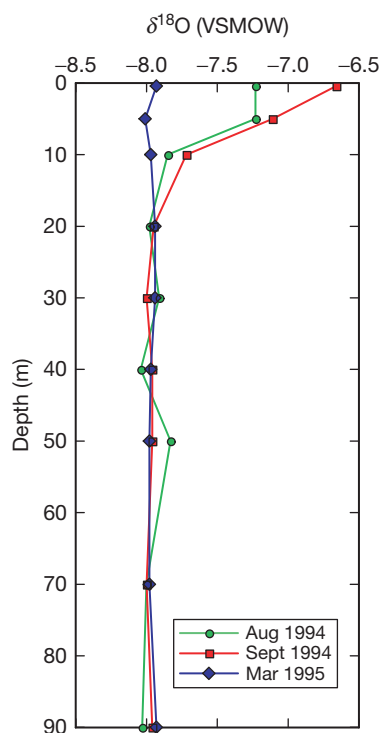


Figure 5 Oxygen isotope composition of lake waters in a water-column profile from Lake Lugano, Switzerland for three different months in 1994–1995. Note the increase in the $\delta^{18}\text{O}$ values of up to 1.5‰ during summer stratification and the homogeneous composition of the water column in March after the lake has mixed.

the days with rain are systematically cooler than days without precipitation, whereas, in winter, the opposite effect is seen (Kohn and Welker, 2005). Thus, the $\delta^{18}\text{O}$ of precipitation most closely reflects a ‘precipitation temperature’ rather than the average yearly or monthly temperature. This adds an additional uncertainty in paleoclimatic reconstructions because, unfortunately, these two temperatures are not well correlated. Changes in moisture source, in the contribution of re-evaporated moisture and seasonal distribution of precipitation also need to be considered. Changes in moisture source or changes in reevaporation along the trajectory of the air mass can change the local temperature/oxygen isotope relationships. Air masses with different sources may have different isotope composition due to the different evaporation source and rain-out history (e.g., Clark and Fritz, 1997, p. 53). An additional important factor is the seasonal distribution of precipitation. In summer, the $\delta^{18}\text{O}$ of precipitation is higher than in winter (e.g., Rozanski et al., 1993). Thus, a switch from more dominant winter to summer precipitation can lead to changes in the $\delta^{18}\text{O}$ of lake water, which could be erroneously interpreted as a temperature change. The effect of changes in atmospheric circulation on lacustrine isotope records has been discussed and holds important potential for developing regional paleoclimate archives (e.g., Hammarlund et al., 2002; Henderson et al. 2010; Hodell et al., 1998; Kirby et al., 2002; McKenzie and Hollander, 1993; Smith and Hollander, 1999; Teranes and McKenzie, 2001).

The above discussion illustrates that, because of the complex nature of the signal, it is virtually impossible to quantitatively reconstruct temperature variations or relative temperature changes without making some important assumptions. For this reason, oxygen isotope studies on lacustrine carbonates should be coupled with other methods and proxies to achieve more precise climate reconstructions. An interesting upcoming new method for temperature reconstructions is the clumped-isotope thermometer (Eiler, 2007; Ghosh et al., 2006). This method is based on the abundance of the ^{13}C – ^{18}O bond in carbonate and allows to determine the temperature of formation of the carbonate independently of the $\delta^{18}\text{O}$ of water. After the temperature is known, it is subsequently possible to determine the $\delta^{18}\text{O}$ of the lake water from the oxygen isotope composition of the carbonate. This method has not been applied to lakes yet. Additional methodologies are palynological or micro-paleontological studies, as well as other novel geochemical proxies such as compound-specific hydrogen isotopes in organic matter (e.g., Huang et al., 2004; Sachs et al., 2009; Sauer et al., 2001). In addition, as suggested by Kohn and Welker (2005), air temperature reconstructions based on the reconstruction of the $\delta^{18}\text{O}$ of precipitation should be validated with modeling efforts conducted using global climate models (GCMs).

Oxygen Isotope Temperature Calculations

Following the discussion of the factors controlling the oxygen isotope composition of lake water, we next consider the factors controlling the oxygen isotope composition of the associated carbonates. These are an indirect archive for the oxygen isotope composition of precipitation and, as such, are an additional filter and source of noise, which has to be understood if quantitative reconstructions of climate change are to be obtained. In particular, the primary question that has to be asked when using oxygen isotope data derived from bulk carbonate in lacustrine sediments is: do they record water temperature or rather the oxygen isotope composition of the water?

The oxygen isotope fractionation between water and calcite has been extensively studied, and a number of paleotemperature equations have been published. In a recent review, Leng and Marshall (2004) proposed to adopt the following modified form of the Kim and O’Neil (1997) calibration:

$$T(^{\circ}\text{C}) = 13.8 - 4.58(\delta_c - \delta_w) + 0.08(\delta_c - \delta_w)^2$$

where δ_c is the oxygen isotope composition of the carbonate with respect to the VPDB standard and δ_w is the oxygen isotope composition of the water with respect to the VSMOW standard (Coplen 1994). The examination of this equation shows that there are two unknowns, the temperature of precipitation and the $\delta^{18}\text{O}$ of the lake water. Thus, for paleoclimate reconstructions, assumptions are necessary to separate these two factors. In the case of tropical to subtropical lakes, even at the scale of glacial to interglacial cycles, the temperature variations at a location are relatively minor, on the order of a couple of degrees centigrade. Thus, the oxygen isotope composition of the carbonate can be interpreted as predominantly a reflection of the $\delta^{18}\text{O}$ of the lake water. Neglecting possible temperature changes will introduce only a minor error in the estimates. In this case, the stratigraphic variations will be dependent

on changes in the water balance, origin of air masses and/or changes in the catchment area (Rosenmeier et al., 2002). As discussed above, in hydrologically open mid- to high-latitude hard-water lakes, carbonate precipitation occurs mainly between April and October, as a consequence of biological activity and increasing temperatures. Thus, possible temperature effects have to be taken into consideration. As shown by Livingstone and Lotter (1999), lake water temperature in the upper epilimnion closely correlates with mean air temperature, therefore, in this case, temperature will influence the oxygen isotope signal in carbonates. An interesting aspect is that increasing average air temperatures would lead to an increase in the $\delta^{18}\text{O}$ of lake water due to the positive correlation between air temperature and $\delta^{18}\text{O}$ of precipitation which is on the order of $0.55\text{‰}\text{°C}^{-1}$ (Kohn and Welker, 2005). At the same time, increasing water temperatures would lead to a decrease in oxygen isotope composition of the calcite because the fractionation between water and calcite decreases with increasing temperature (Kim and O'Neil, 1997), at a rate of approximately $0.2\text{‰}\text{°C}^{-1}$. Thus, the net effect of temperature increase on carbonate isotope composition would be on the order of $0.35\text{‰}\text{°C}^{-1}$. Unfortunately, other factors discussed below, may also play a role in determining the oxygen isotope composition of the carbonate, and only rarely a direct inference of average annual temperature can be made. This approximation has been confirmed in case studies, such as one for Baldeggersee, Switzerland, where, based on an isotope mass-balance model, the $\delta^{18}\text{O}$ /temperature relationship was calculated to be $0.39\text{‰}\text{°C}^{-1}$ (Teranes and McKenzie, 2001).

Environmental Effects (e.g., Vital Effects, pH) on the Oxygen Isotope Composition of Carbonate

Calcite precipitation occurs in lakes only after relatively high levels of supersaturation have been reached (Kelts and Hsu, 1978). Thus, strictly speaking, carbonate precipitation is not an equilibrium process, and the possibility of isotopic disequilibrium has to be taken into consideration. Although most paleoclimatological studies tacitly assume that carbonate precipitation occurs in equilibrium with the water, this has rarely been demonstrated with water column studies. Sediment trap studies in eutrophic lakes indicate that most of the carbonate does precipitate in isotopic equilibrium (Fisher 1996; Teranes et al., 1999a). However, it is not always possible to demonstrate this unambiguously because the water depth at which carbonate precipitation occurs may vary with the trophic level of a lake or with the type of organisms responsible for inducing carbonate precipitation.

The mechanism and water depth of carbonate precipitation are important because the location and timing in the annual cycle of carbonate precipitation influences its isotopic composition. This is a particularly important factor in lakes, as compared to the oceans, because the water column is characterized by strong vertical gradients in temperature and in summer, also, in oxygen isotope composition. In addition, the annual productivity cycle can be variable and produce shifts in the timing of precipitation of carbonates, which can introduce spurious variations in their oxygen isotope composition.

The pH of the lake water is also an important parameter to consider in the interpretation of carbonate records. Recently,

studies pioneered by Zeebe (1999, 2007) have proposed a model of pH dependence of oxygen isotope composition of carbonates, which can be used to explain effects previously attributed to disequilibrium, kinetic fractionations and/or to not always clearly understood 'vital effects.' These studies show that the observed decrease in oxygen isotope composition with the concentration of CO_3^{2-} , first observed by Spero et al. (1996) and commonly considered a 'vital effect' or disequilibrium, can be explained by carbonate precipitation at different pH values. This model is based on the fact that dissolved HCO_3^- has an approximately 7‰ higher $\delta^{18}\text{O}$ than CO_3^{2-} (at 25 °C, Zeebe, 2007). Thus, if carbonate precipitating as calcite contains a mixture of bicarbonate and carbonate, precipitation at higher pH will lead to a progressively lower oxygen isotope composition. This is caused by the higher abundance of CO_3^{2-} with increasing pH. This process is potentially particularly important in lakes because the pH of the surface waters can vary over 2–3 pH units throughout the year. It is also significant that vertical pH gradients are observed in the water column. These gradients are particularly strong during summer when primary productivity is at maximum in concomitance with the stable thermal stratification in the mixolimnion. A pH effect may explain the decrease in $\delta^{18}\text{O}$ observed by Teranes et al. (1999b) in their sediment trap study of Lake Baldegger, Switzerland. A disturbance of the oxygen isotope records of calcite may also be expected when a lake undergoes eutrophication. At higher levels of primary productivity, the pH in the surface waters can be expected to reach higher values. Therefore, some of the negative $\delta^{18}\text{O}$ shifts observed in highly productive lakes may be due to higher average pH related to increased productivity rather than to changes in the $\delta^{18}\text{O}$ of the waters or to a shift to carbonate precipitation later in the summer, when water temperatures are higher.

Quaternary Lacustrine Isotopic Records

Quaternary marine oxygen isotope records show a dominant pattern of change modulated by glacial–interglacial changes in ice volume and temperature, which can be found at most latitudes and longitudes. This clear and reproducible pattern of variation has been used to define a robust chronology, the marine isotope stages, which can be used as an accurate dating tool. In contrast, lacustrine oxygen isotope records are much more dominated by local effects. Therefore, it is not possible to establish a general lacustrine quaternary isotope stratigraphy equivalent to the marine one. Nevertheless, as discussed above, lacustrine isotope records, when compared with marine and other terrestrial archives, provide important climatic constraints. Depending on the type of lake, isotope composition of lacustrine carbonates may reflect changes in average air temperature, seasonal distribution of precipitation, evaporation/precipitation ratio (aridity) or for example changes in atmospheric circulation patterns.

Summary and Conclusions

In spite of the complexities in the interpretation of bulk lacustrine carbonate oxygen isotope values, lacustrine oxygen isotope records are arguably an invaluable source of paleoclimatic

data. The last years have seen major progress in our understanding of the hydrological effects on the oxygen isotope composition of lakes and on the significance and problems related to the interpretation of bulk lacustrine isotope records. The quantitative interpretation of lacustrine oxygen isotope records requires, first of all, a good understanding of the hydrology of the studied lake. Important parameters include a good characterization of the catchment area, which may not be easily determined in karstic areas, where significant underground sources and outflow can occur, and information on the residence time of the water in both the catchment and the lake itself, which controls the temporal resolution of the sedimentary record. In addition, the relationship between $\delta^{18}\text{O}$ of the precipitation in the catchment area and the lake water has to be understood and calibrated. Possible changes in the catchment area through time and its effect on the hydrological and isotopic balance have also to be considered. Soil and vegetation development may change the residence time of the water in the catchment and diversion of tributaries, by natural or anthropogenic causes, may change the size and characteristics of the catchment area and, hence, the isotope composition of the inputs.

Finally, the sediment has to be carefully examined to determine possible disturbances of the primary signal by detrital or authigenic carbonate components of variable origin, which may lead to spurious variations in $\delta^{18}\text{O}$ of the bulk carbonate. Possible effects due to changes in trophic state have to be considered also as a possible complication. This is especially important for lakes strongly affected by anthropogenic or natural eutrophication. Future, in-progress developments include the implementation of isotopes in precipitation in global circulation models, which can be used to evaluate the relative importance of temperature versus changes in atmospheric circulation on the oxygen isotope records. In addition, the robustness of oxygen isotope records needs to be validated with independent records, such as those derived from clumped isotopes, tree rings or compound-specific hydrogen isotopes. In conclusion, there are ongoing efforts to establish stable isotope maps of precipitation for different time slices, using a combination of terrestrial isotope records including lakes. These maps can be used to test and calibrate GCM's to improve our quantitative understanding of paleoclimatic variations.

See also: **Glacial Landforms, Tree Rings:** Dendrogeomorphology. **Quaternary Stratigraphy:** Pedostratigraphy. **Vertebrate Records:** Mid-Pleistocene of Australia; Mid-Pleistocene of North America.

References

- Bernasconi SM, Dobson J, McKenzie JA, et al. (1997) Preliminary isotopic and palaeomagnetic evidence for Younger Dryas and Holocene climate evolution in NE Asia. *Terra Nova* 9: 246–250.
- Clark ID and Fritz P (1997) *Environmental Isotopes in Hydrogeology*, 328 pp. Boca Raton: CRC Press.
- Coplen TB (1994) Reporting of stable hydrogen, carbon, and oxygen isotopic abundances. *Pure and Applied Chemistry* 66: 273–276.
- Dansgaard W (1964) Stable isotopes in precipitation. *Tellus* 16: 436–468.
- Dittrich M, Kurz P, and Wehrli B (2004) The role of autotrophic Picocyanobacteria in calcite precipitation in an oligotrophic lake. *Geomicrobiology Journal* 21: 45–53.
- Eiler JM (2007) "Clumped-isotope" geochemistry – The study of naturally-occurring, multiply-substituted isotopologues. *Earth and Planetary Science Letters* 262: 309–327.
- Fisher A (1996) Isotopengechemische Untersuchungen im Wasser und in den Sedimenten des Soppensees (Kt. Luzern, Schweiz) Ph.D Dissertation ETH Zürich No. 11924.
- Fronval T, Jensen NB, and Buchardt B (1995) Oxygen isotope disequilibrium precipitation of calcite in Lake Arreso, Denmark. *Geology* 23: 463–466.
- Gat JR and Lister GS (1995) The "catchment effect" on the isotopic composition of lake waters; its importance in palaeolimnological interpretations. In: Frenzel B (ed.) *Problems of Stable Isotopes in Tree-Rings, Lake Sediments and Peat-Bogs as Climatic Evidence for the Holocene*, vol. 15, ch. 1, pp. 1–16. Stuttgart, Germany: Akademie der Wissenschaften und der Literatur.
- Ghosh P, Adkins J, Affek H, et al. (2006) C-13-O-18 bonds in carbonate minerals: A new kind of paleothermometer. *Geochimica et Cosmochimica Acta* 70: 1439–1456.
- Glew JR, Smol JP, and Last WM (2001) Sediment core collection and extrusion. In: Last WM and Smol JP (eds.) *Tracking Environmental Change Using Lake Sediments*, 1, vol. 1, ch. 5, pp. 73–106. Dordrecht: Kluwer.
- Hammarlund D, Barnekov L, Birks HJB, Buchardt B, and Edwards TWD (2002) Holocene changes in atmospheric circulation recorded in the oxygen-isotope stratigraphy of lacustrine carbonates from northern Sweden. *The Holocene* 12: 339–351.
- Henderson ACG, Holmes JA, and Leng MJ (2010) Late Holocene isotope hydrology of Lake Qinghai, NE Tibetan Plateau: Effective moisture variability and atmospheric circulation changes. *Quaternary Science Reviews* 29: 2215–2223.
- Hodell DA, Schelske CL, Fahnenstiel GL, and Robbins LL (1998) Biologically induced calcite precipitation and its isotopic composition in Lake Ontario. *Limnology and Oceanography* 43: 187–199.
- Hollander DJ and McKenzie JA (1991) CO₂ control on carbon-isotope fractionation during aqueous photosynthesis: A paleo-p-CO₂ barometer. *Geology* 19: 929–932.
- Huang Y, Shuman B, Wang Y, and Webb T (2004) Hydrogen isotope ratios of individual lipids in lake sediments as novel tracers of climatic and environmental change: A surface sediment test. *Journal of Paleolimnology* 31: 363–375.
- Kelts K and Hsu KJ (1978) Freshwater carbonate sedimentation. In: Lerman A (ed.) *Lakes: Geology, Chemistry, Physics*, pp. 295–323. New York: Springer Verlag.
- Kim ST and O'Neil JR (1997) Equilibrium and nonequilibrium oxygen isotope effects in synthetic carbonates. *Geochimica et Cosmochimica Acta* 61: 3461–3475.
- Kirby ME, Patterson WP, Mullins HT, and Burnett AW (2002) Post-Younger Dryas climate interval linked to circumpolar vortex variability: Isotopic evidence from Fayetteville Green Lake, New York. *Climate Dynamics* 19: 321–330.
- Kohn MJ and Welker JM (2005) On the temperature correlation of delta O-18 in modern precipitation. *Earth and Planetary Science Letters* 231: 87–96.
- Lehmann MF, Bernasconi SM, Barbieri A, Simona M, and McKenzie JA (2004) Seasonal variation of the $\delta^{13}\text{C}$ and $\delta^{15}\text{N}$ of particulate and dissolved carbon and nitrogen in lake Lugano: Constraints on biogeochemical cycling in a eutrophic Lake. *Limnology and Oceanography* 49: 415–429.
- Leng MJ and Marshall JD (2004) Palaeoclimate interpretation of stable isotope data from lake sediment archives. *Quaternary Science Reviews* 23: 811–831.
- Li C and Ku TL (1997) $\delta^{13}\text{C}$ – $\delta^{18}\text{O}$ covariance as a paleohydrological indicator for closed-basin lakes. *Palaeogeography, Palaeoclimatology, Palaeoecology* 133: 69–80.
- Livingstone DM and Lotter AF (1999) The relationship between air and water temperatures in lakes of the Swiss Plateau: A case study with palaeolimnological implications. *Journal of Paleolimnology* 19: 181–198.
- McKenzie JA (1985) Carbon isotopes and productivity in the lacustrine and marine environment. In: Stumm W (ed.) *Chemical Processes in Lakes*, pp. 99–118. New York: Wiley-Interscience, John Wiley & Sons.
- McKenzie JA and Hollander DJ (1993) Oxygen isotope record in recent carbonate sediments from Lake Griefen, Switzerland (1750–1986): Application of continental isotopic indicator for evaluation of changes in climate and atmospheric circulation patterns. In: Swart PK, Lohmann KC, McKenzie JA, and Savin S (eds.) *Climate Change in Continental Isotopic Records. Geophysical Monographs* 78, pp. 101–112. Washington, DC: American Geophysical Union.
- Rosenmeier MF, Hodell DA, Brenner M, Curtis JH, and Guilderson TPA (2002) A 4000-year lacustrine record of environmental change in the southern Maya lowlands, Peten, Guatemala. *Quaternary Research* 57: 183–190.
- Rozanski K, Araguas-Araguas L, and Gonfiantini R (1993) Isotopic patterns in modern precipitation. In: Swart PK, Lohmann KC, McKenzie JA, and Savin S (eds.) *Climate Change in Continental Isotopic Records. Geophysical Monographs* 78, pp. 1–36. Washington, DC: American Geophysical Union.
- Sachs JP, Sachse D, Smittenberg RH, Zhang ZH, Battisti DS, and Golubic S (2009) Southward movement of the Pacific intertropical convergence zone AD 1400–1850. *Nature Geoscience* 2: 519–525.

- Sauer PE, Eglinton TI, Hayes JM, Schimmelmann A, and Sessions A (2001) Compound-specific D/H ratios of lipid biomarkers from sediments as a proxy for environmental and climatic conditions. *Geochimica et Cosmochimica Acta* 65: 213–222.
- Smith MA and Hollander DJ (1999) Historical linkage between atmospheric circulation patterns and the oxygen isotopic record of sedimentary carbonates from Lake Mendota, Wisconsin, USA. *Geology* 27: 589–592.
- Spero HJ, Bijma J, Lea DW, and Bemis BE (1996) Effect of seawater carbonate concentration on foraminiferal carbon and oxygen isotopes. *Nature* 390: 497–500.
- Stuiver M (1970) Oxygen and carbon isotope ratios of fresh-water carbonates as climatic indicators. *Journal of Geophysical Research* 75: 5247–5257.
- Talbot MR (1990) A review of the paleohydrological interpretation of carbon and oxygen isotopic-ratios in primary lacustrine carbonates. *Chemical Geology* 80: 261–279.
- Teranes JL and Bernasconi SM (2005) Factors controlling delta C-13 values of sedimentary carbon in hypertrophic Baldeggersee, Switzerland, and implications for interpreting isotope excursions in lake sedimentary records. *Limnology and Oceanography* 50: 914–922.
- Teranes JL and McKenzie JA (2001) Lacustrine oxygen isotope record of 20th-century climate change in central Europe: Evaluation of climatic controls on oxygen isotopes in precipitation. *Journal of Paleolimnology* 26: 131–146.
- Teranes JL, McKenzie JA, Bernasconi SM, Lotter AF, and Sturm M (1999a) A study of oxygen isotopic fractionation during bio-induced calcite precipitation in eutrophic Baldeggersee, Switzerland. *Geochimica et Cosmochimica Acta* 63: 1981–1990.
- Teranes JL, McKenzie JA, Lotter AF, and Sturm M (1999b) Stable isotope response to lake eutrophication: Calibration of a high-resolution lacustrine sequence from Baldeggersee, Switzerland. *Limnology and Oceanography* 44: 320–333.
- Thompson JB, Schultze-Lam S, Beveridge TJ, and Des-Marais DJ (1997) Whiting events: Biogenic origin due to the photosynthetic activity of cyanobacterial picoplankton. *Limnology and Oceanography* 42: 133–141.
- Wick L, Lemcke G, and Sturm M (2003) Evidence of Late glacial and Holocene climatic change and human impact in eastern Anatolia: High-resolution pollen, charcoal, isotopic and geochemical records from the laminated sediments of Lake Van, Turkey. *The Holocene* 13: 665–675.
- Zeebe RE (1999) An explanation of the effect of seawater carbonate concentration on foraminiferal oxygen isotopes. *Geochimica et Cosmochimica Acta* 63: 2001–2007.
- Zeebe RE (2007) An expression for the overall oxygen isotope fractionation between the sum of dissolved inorganic carbon and water. *Geochemistry, Geophysics, Geosystems* 8. <http://dx.doi.org/10.1029/2007gc001663> Q09002, 1–7.

Nonmarine Biogenic Carbonates

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Introduction

Biogenic carbonates – predominantly calcite and aragonite precipitated by plants and animals in non-marine surface waters – are ubiquitous in Quaternary age deposits. Carbon and oxygen isotope ratios ($\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values) of these materials are routinely used as proxies for climate change. The ultimate utility of these data rests on scientists' ability to calibrate modern analogs, understand biological fractionations or 'vital effects', and use appropriate sampling methods. These materials have the potential for providing detailed data (sub-monthly resolution) for time periods that range from 1 to 50 years in length. This temporal resolution can be used in conjunction with longer term, yet lower temporal resolution, terrestrial isotope records (e.g., ice cores, stalagmites, soil profiles, and lake sediment cores) to examine seasonal variation in temperature, precipitation, vegetation, productivity, and water column stratification.

Equilibrium Precipitation

Isotope geochemists often seek to determine whether CaCO_3 is precipitated in isotopic equilibrium with ambient waters (i.e., the $\delta^{18}\text{O}_{\text{water}}$ and $\delta^{13}\text{C}_{\text{DIC}}$ values). By knowing this relation, it may be possible to determine the temperature or the isotopic composition of waters in which ancient carbonates formed. As with most estimates of this sort, assumptions must be made. Unfortunately, carbonate oxygen isotope thermometry requires knowledge of the $\delta^{18}\text{O}$ value of ambient waters. Conversely, estimates of $\delta^{18}\text{O}_{\text{water}}$ values require knowledge of the temperature at which the carbonate precipitated. To evaluate equilibrium precipitation, the results of numerous experimental and empirical studies must be considered.

Equilibrium $\delta^{18}\text{O}$ values of calcite are most accurately estimated using the $\delta^{18}\text{O}$ value of ambient water, water temperature (in kelvins), and a calcite (low-Mg calcite), the calcite-water fractionation factor (α), and an expression that relates these variables (Friedman and O'Neil, 1977; Kim and O'Neil, 1997).

$$\alpha(\text{CaCO}_3\text{-water}) = \frac{(1000 + \delta^{18}\text{O}_{\text{CaCO}_3(\text{SMOW})})}{(1000 + \delta^{18}\text{O}_{\text{Water}(\text{SMOW})})} \quad [1]$$

$$10^3 \ln \alpha(\text{calcite-water}) = 2.78(10^6/T^2) - 2.89 \quad [2]$$

(Friedman and O'Neil, 1977)

$$10^3 \ln \alpha(\text{calcite-water}) = 18.03(10^3/T) - 32.42 \quad [3]$$

(Kim and O'Neil, 1997)

These values can be corrected for the Mg-content of the calcite (+0.06‰ per mol% MgCO_3 ; Tarutani et al., 1969).

To examine carbonate equilibrium $\delta^{13}\text{C}$ values, it is necessary to know the $\delta^{13}\text{C}$ value of water bicarbonate (HCO_3^-) or the dissolved inorganic carbon ($\delta^{13}\text{C}_{\text{DIC}}$). The calcite- HCO_3^- carbon isotope enrichment factor is +1.0‰ (and is constant from 10 to 40 °C). The aragonite- HCO_3^- enrichment factor is +2.7‰ throughout the same temperatures range (Romanek et al., 1992). Assuming equilibrium precipitation, one can estimate ambient $\delta^{13}\text{C}_{\text{DIC}}$ values by subtracting 1.0 or 2.7‰ from measured calcite and aragonite values. Likewise, aragonite equilibrium $\delta^{13}\text{C}$ values can be estimated by adding the aragonite-calcite enrichment factor to measured calcite $\delta^{13}\text{C}$ values (+1.6‰ of Rubinson and Clayton, 1969; +1.7‰ of Romanek et al., 1992).

The only empirically derived oxygen isotope fractionation factor for freshwater fish otoliths is:

$$10^3 \ln \alpha(\text{aragonite-water}) = 18.56(10^3/T) - 33.49 \quad [4]$$

(Patterson et al., 1993)

The only empirically derived expression describing the oxygen isotope fractionation factor in Unionid aragonite is:

$$10^3 \ln \alpha(\text{aragonite-water}) = 2.559(10^6/T^2) + 0.715 \quad [5]$$

(Dettman et al., 1999)

Equations 4 and 5 have slopes similar to that of Grossman and Ku (1986) for biogenic aragonite:

$$10^3 \ln \alpha(\text{aragonite-water}) = 18.07(10^3/T) - 31.079 \quad [6]$$

(Grossman and Ku, 1986)

'Vital Effects' and Modern Calibration

The most significant impediment to effectively using freshwater biogenic carbonates as climate or environmental proxies is a generally poor understanding of biological fractionation of stable isotopes during formation. 'Vital effects' (Urey et al., 1951) are in many ways a means of categorizing processes that are poorly understood. Numerous studies have examined 'vital effects' in a variety of marine organisms (see these and citations within; corals, Adkins et al., 2003; calcareous algae, Lee and Carpenter, 2001; brachiopods, Carpenter and Lohmann, 1995; Foraminifera, Zeebe, 1999). Common among these studies is a less-than-satisfactory explanation of mechanisms and rates of carbonate precipitation, the incorporation of metabolic CO_2 , and the positive covariance of $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values in many biogenic carbonates. Far less work has been conducted on "vital effects" in freshwater biogenic carbonates.

More work is needed to examine the relation between the $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ value of ambient waters and their biogenic carbonates. These efforts require relatively long term

monitoring of waters and related carbonates (a minimum of one year). In addition to monitoring the $\delta^{13}\text{C}_{\text{DIC}}$, $\delta^{18}\text{O}_{\text{water}}$, and temperature, measurements of pH, $[\text{HCO}_3^-]$, overall stream/lake productivity, the $\delta^{13}\text{C}$ values of potential food sources are needed to examine the seasonal variation in the $\delta^{13}\text{C}_{\text{DIC}}$ and the potential physico-chemical and biological causes of 'vital effects' (e.g., Dettman et al., 1999). In addition to environmental monitoring, detailed examination of the mechanism of calcification and physiological variation during the life history of individual organisms is needed.

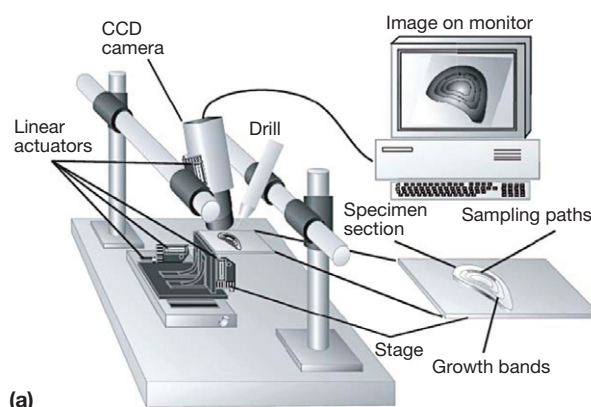
Once such data are collected, empirical fractionation factors can be established for geologically significant species of plants and animals. A concerted effort by funding agencies is needed to establish these baseline data for modern organisms. Until these data are in hand, the utility of stable isotope data from archaeological and geological specimens will be in question.

Micro-Sampling Techniques

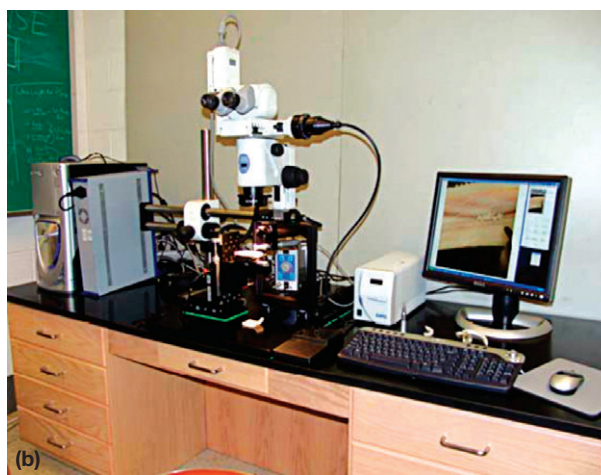
The development of micro-sampling or micro-milling techniques (the work of K.C. Lohmann and associates) to extract small amounts of powdered carbonate from ancient rocks, paleontological specimens, modern shells or otoliths, etc., together with the improved sensitivity of isotope ratio mass spectrometers (IRMS) has created a means of collecting very high temporal resolution data from such materials (Dettman and Lohmann, 1995; Wurster et al., 1999; Frappier et al., 2002). These automated micro-sampling devices are capable of routinely milling 15 to 20 μm -wide paths from biogenic carbonates – permitting sub-monthly sampling resolution in most specimens (Figure 1). These devices employ low eccentricity dental hand pieces equipped with precision drill bits, precision X-Y-Z stages and actuators, microscopes equipped with digital imaging devices, and custom software and motion controllers. Currently, these devices can be purchased commercially from two manufacturers.

Micro-sampling of biogenic carbonates with well-defined growth increments (mollusk shells, otoliths) is critical in deciphering ontogenetic variation in these materials (Figure 2). The importance of collecting sub-monthly resolution records from biogenic carbonates is the understanding of ontogenetic variations (associated with physiological changes such as sexual maturation and overall metabolic activity) and seasonal isotopic variation (induced by temperature and ambient conditions such as photosynthesis, respiration, and water column stratification). By collecting data at approximately bi-weekly intervals, one can examine short-term variation that would be overlooked by bulk analyses or lower-resolution sampling. Weekly to monthly sampling resolution is preferred as this is the temporal scale at which weather events occur (storms and surface water changes, cold front migrations, etc.) and to which the $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values of biogenic carbonates react.

Where shell growth rates are sufficiently rapid, high-resolution micro-sampling of biogenic carbonates is not needed. In such cases, more routine, manual sampling can be conducted to collect the appropriate temporal resolution for the project. The non-linear growth of unionid shells, both during an annual cycle (rapid extension rates in the Summer and slow growth rates in the Winter) and through



(a)



(b)

Figure 1 (a) Schematic diagram of an automated micro-sampler using a video camera for image capture (From Patterson and Smith, 2001, after Wurster et al., 1999). (b) Photograph of the Carpenter Microsystems CM-2 Micro-sampling System at the University of North Carolina. This system is equipped with a precision X-Y-Z stage with linear actuators permitting two inches of travel, a horizontally mounted precision handpiece, 15:1 stereoscopic zoom microscope equipped with both transmitted light and an epifluorescence illumination systems and a 12 mega-pixel digital camera. These systems allow for real time/on-screen digitization of sampling paths.

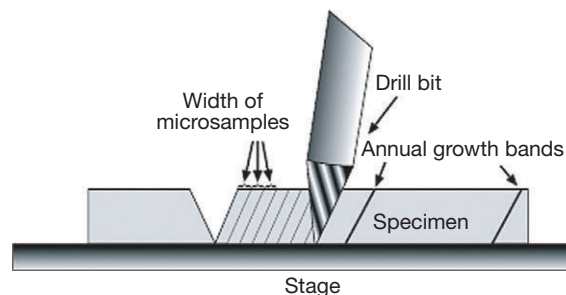


Figure 2 Close up schematic illustration of how individual micro-samples are milled from biogenic carbonate specimens using a rotating, faceted tungsten carbide drill bit. A first pass coarsely mills a trough parallel to the sample paths that allows the bit to shave small increments off one wall of the trough. These subsequent high-resolution paths are approximately 15 to 20 μm wide. Sample powders are manually removed from the specimen using a surgical scalpel. (From Patterson and Smith, 2002, after Wurster et al., 1999).

ontogeny (diminished growth rates with age) requires adjustments in sampling strategies.

Mid-latitude unionids from the Northern Hemisphere are thought to greatly reduce or cease carbonate precipitation below $\sim 12^\circ\text{C}$ (e.g., Howard, 1921 (*Lampsilis luteola*); Negus, 1966 (*Anodonta anatine*); Dettman et al., 1999). Both *Dreissena* sp. (Neumann et al., 1993) and *Corbicula* sp. (Joy, 1985) display little growth below 10°C . Interestingly, *Lampsilis radiata siliquoidea* from the Mackenzie River, NWT, Canada precipitates carbonate in water temperatures below 12°C (Carpenter et al., in prep.). The same species from more southerly latitudes in the United States (southeastern Michigan; e.g., Dettman et al., 1999) does slow shell production at this temperature. The occurrence of non-winter growth bands (pseudoannual bands) in unionids (slowing of shell carbonate production) seems to correspond to abrupt changes in ambient conditions (decreases in water temperature and O_2 levels; Carpenter unpublished data). Clearly, cessation of shell growth is not fully understood and requires examination of genera or species that are geographically widespread.

Cessation of shell growth in unionids results in a lack of data corresponding to the late Fall through early Spring in mid-latitudes. This lack of growth diminishes the ability to examine annual temperature minima, the $\delta^{18}\text{O}$ value of winter precipitation, and the oxidation of organic matter in lakes and streams. However, a uniform, Winter “shut-down” temperature allows calculation of ambient water $\delta^{18}\text{O}$ values by fixing temperature (assuming a known fractionation factor; e.g., Dettman et al., 1999). Given variation in shell growth rates, it's important to sample materials in a way that maximizes sampling efficiency with increased resolution during the Fall through Spring and perhaps a decreased resolution in the summer when shells are growing rapidly. This temporal “accordion effect” and the occurrence of Winter “shut downs” require special sampling strategies and data interpretations (see shell structure of pseudoannual and annual bands of Tevesz and Carter, 1980).

Many researchers have chosen to mill samples from the interior of unionid shells to avoid potential diagenetic alteration of the thin, outer prismatic layer of the shell. Although one should always assess diagenetic alteration, Quaternary age materials often do not exhibit extensive diagenetic alteration. In such cases, the outer surface of shells can be micro-sampled. One must use care to avoid drilling too deeply into the shell surface as these samples may sample carbonate that is older than that expressed at the shell surface. Dettman et al. (1999) found no difference in the $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values of the prismatic and nacreous layers of a modern *Lampsilis radiata siliquoidea* shell (Figure 3).

Unionid Bivalves and Freshwater Mollusks

Seasonal variation of $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values in freshwater mollusk shells from temperate settings produces sigmoidal or sinusoidal ontogenetic trends. In theory, the carbon and oxygen isotope ratios of shell carbonate will be negatively correlated with lower $\delta^{18}\text{O}$ value and higher $\delta^{13}\text{C}$ values occurring during the summer months when temperatures and primary productivity are highest (Figure 4). During the winter months (characterized by low temperatures and minimal productivity), $\delta^{18}\text{O}$

values reach their maximum and $\delta^{13}\text{C}$ values reach their minimum. The $\delta^{18}\text{O}_{\text{water}}$, temperature, and the $\delta^{13}\text{C}$ value of DIC (dissolved inorganic carbon or $\sum\text{CO}_2$) should play important roles in determining shell isotope ratios. There are several complicating factors that modify these theoretical parameters (e.g., specialized food sources, carbon cycling lag times, variable effects of chemical change on food-web dynamics, and the buffering capacity of the ambient water mass (the mass of carbon and oxygen upon which physico-chemical processes work)).

Lampsilis sp. – Huron River, Michigan

Dettman et al.'s (1999) examination of unionids from southeastern Michigan streams provided an introduction to how high-resolution data can be extracted and interpreted. Their data from *Lampsilis* (Huron River) are worthy of mention (Figure 3). There is a strong correlation between the $\delta^{18}\text{O}$ values collected from *L. o. ventricosa* and *L. r. siliquoidea*. Shell $\delta^{18}\text{O}$ values also are in isotopic equilibrium with ambient river

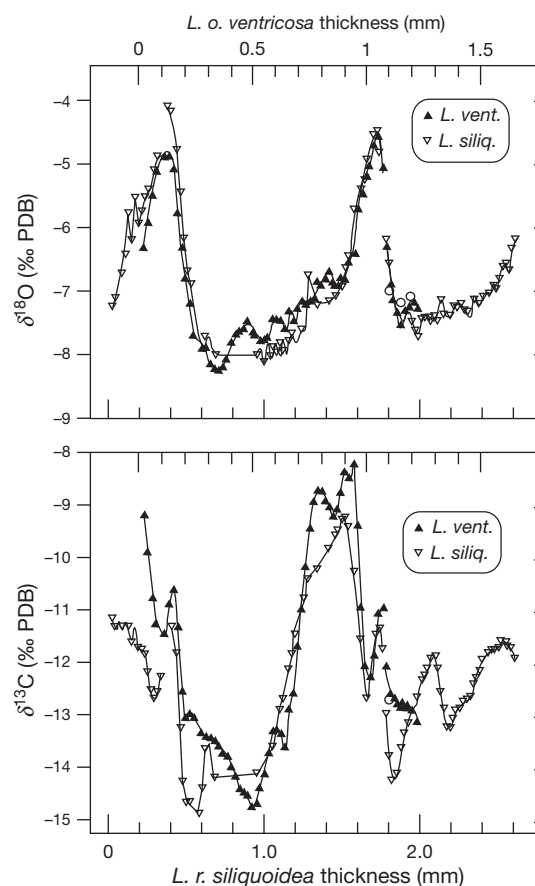


Figure 3 Shell $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values of two unionids from the Huron River, SE Michigan (*Lampsilis ovata ventricosa* and *Lampsilis radiata siliquoidea*). Shell thickness is measured perpendicular to nacreous plates, beginning at an arbitrary point before the 1990–1991 hibernation mark. The isotope records are aligned at $\delta^{18}\text{O}$ maxima. *L. o. ventricosa* data have been stretched by 45% to compensate for its slower growth. Lines connecting points are separated at visible “hibernation” or “shut down” features. Open circles are prismatic layer measurements of *L. o. ventricosa*. From Dettman et al. (1999).

water (Eqn. 5, Figures 3 and 5). The timing of variation in shell $\delta^{13}\text{C}$ values also correlates, yet maximum and minimum $\delta^{13}\text{C}$ values vary significantly between co-occurring species (Figure 3). Similarities between prismatic and nacreous layer isotope ratios confirm the utility of sampling the outer prismatic layer.

One can conclude that both species of *Lampsilis* are responding to similar environmental changes. The seasonal variation in $\delta^{13}\text{C}$ values is difficult to interpret in terms of measured variation in river temperature and $\delta^{13}\text{C}_{\text{DIC}}$ values and the relative timing of $\delta^{13}\text{C}$ value variation in the shells. It appears that there is a significant lag between stream productivity increases (marked by an increase in river $\delta^{13}\text{C}_{\text{DIC}}$ values) and shell $\delta^{13}\text{C}$ values.

Ricken et al. (2003) have found similar isotopic equilibrium precipitation in *Anodonta* and *Unio* from the Rhine River. This study examines the relation between isotopically distinct discharge from alpine and non-alpine catchments, using well documented surface water data. Their *Anodonta* specimens exhibit large decreases in $\delta^{13}\text{C}$ values during winter months (corresponding with high $\delta^{18}\text{O}$ values). Again, these values do not appear to be in isotopic equilibrium with ambient water DIC.

Pyganodon lacustris - Glovers Pond, New Jersey

The lacustrine unionid bivalve, *Pyganodon lacustris* occurs in *Chara*-rich sediment within the shallow water, carbonate bank (~1 m water depth) of Glovers Pond, NJ (Figures 6 and 7). Glovers Pond is a temperate, dimictic lake, in northwestern New Jersey and is spring-fed (through Ordovician dolomite). Today,

Glovers Pond is thermally and chemically stratified with a maximum depth of 9.5 meters and occupies a glacially modified and dammed fault valley currently being infilled concentrically by four types of sediment (peat, marl, calcareous-organic-rich silt, and gyttja, Erickson, 1968). The hypolimnion is undersaturated with respect to CaCO_3 which presently prohibits deposition of calcium carbonate in the deepest portions of the lake (Erickson, 1968). CaCO_3 (calcite) is deposited on a shallow shelf both on the surfaces of *Potamogeton* sp. and to a greater extent by blanketing growths of the green alga *Chara* sp. Marl and peat deposits form a platform that is prograding into the gyttja lithotope in the center of the lake basin (Figure 7).

Shell $\delta^{18}\text{O}$ values of *Pyganodon lacustris* record predicted seasonal changes in oxygen isotope ratios associated with variation in ambient temperature, rainfall, and evaporation. Shell $\delta^{13}\text{C}$ values record seasonal changes in lake productivity and stratification/overturn events (Figure 6). Warming of lake waters in the Spring yields a marked decrease in shell $\delta^{18}\text{O}$ values and the concomitant increase in productivity causes an increase in shell $\delta^{13}\text{C}$ values. Fall overturns (Oct.–Nov.) mix the cold, ^{13}C -depleted deep water (where organic matter is oxidized) with shallow waters. This produces a marked increase in shell $\delta^{18}\text{O}$ values and can produce as much as a 5‰ decrease in shell $\delta^{13}\text{C}$ values. Following this overturn, shell $\delta^{13}\text{C}$ values rise briefly (due to continued productivity in the lake) and then decrease abruptly (~7‰) as low temperatures diminish productivity in the lake and organic matter is oxidized (returning ^{13}C -depleted carbon from plant material back to lake waters). As stated previously, winter “shut downs” represent an unconformity in the shell isotope record when water

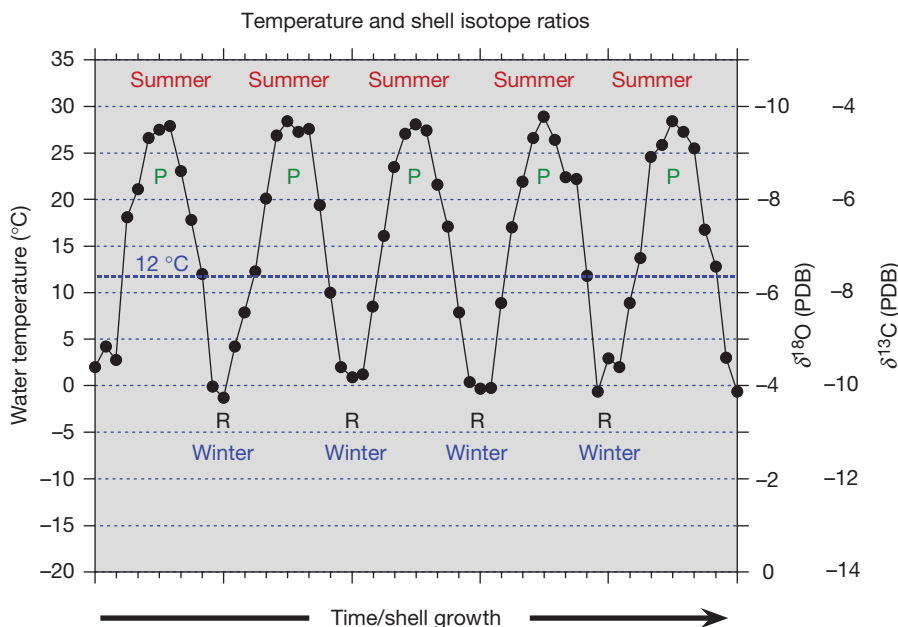


Figure 4 Idealized plot of unionid bivalve shell $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values and water temperature. Sinusoidal trends represent both $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values (scales for each at right). Winters (cold temperatures) are characterized by relatively high $\delta^{18}\text{O}$ values and low $\delta^{13}\text{C}$ values (due to oxidation or respiration (R) of plant material). Summers (warm temperatures) are characterized by relatively low $\delta^{18}\text{O}$ values and high $\delta^{13}\text{C}$ values (due to the removal of ^{12}C by photosynthesis (P) leaving DIC enriched in ^{13}C). Natural data will likely have seasonal changes in $\delta^{13}\text{C}$ values lagging behind $\delta^{18}\text{O}$ values. The 12 °C line represents the temperature at which shell carbonate may stop being precipitated. As a result, all data below this line may be missing in the unionid shell record. The resulting sinusoidal trends will likely have diminished seasonal amplitude and significant missing time associated with annual winter growth bands.

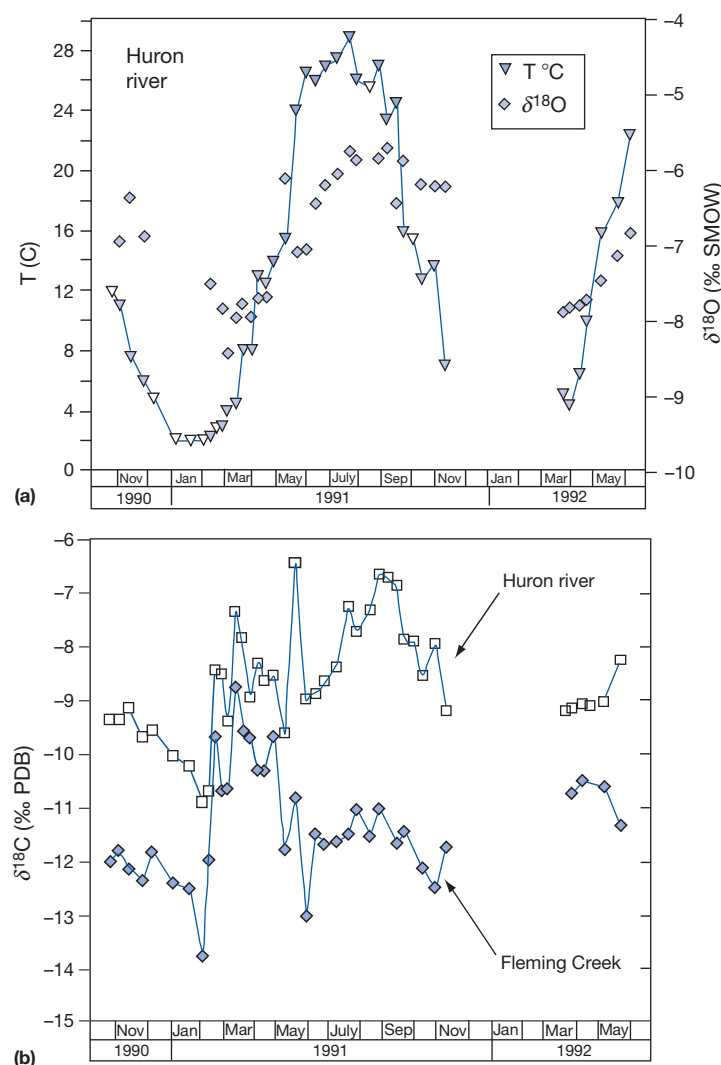


Figure 5 (a) Water temperatures and $\delta^{18}\text{O}$ values for the Huron River from October, 1990 to June, 1992. Filled triangles are temperatures that have associated water $\delta^{18}\text{O}$ measurements. (b) The $\delta^{13}\text{C}$ values of DIC from the Huron River and Fleming Creek, Michigan. These values correspond with shell data in Figure 3. From Dettman et al. (1999).

temperatures fall below $\sim 12^{\circ}\text{C}$. Spring overturns (Feb.–March) cannot be separated from the annual $\delta^{13}\text{C}$ minimum established by the cessation of productivity during the winter and the missing shell record. Return of lake productivity in the spring and summer produces a gradual increase in $\delta^{13}\text{C}$ values (by $\sim 6\text{‰}$) until this trend is modified by the next fall overturn.

Although these data suggest that lacustrine bivalves can provide a detailed seasonal record of temperature, stratification, and productivity in small lake basins, the $\delta^{13}\text{C}$ values of *Pyganodon lacustris* are lower than predicted equilibrium $\delta^{13}\text{C}$ values for shell aragonite. Grovers Pond $\delta^{13}\text{C}_{\text{DIC}}$ values are $\sim -6\text{‰}$ (Quevedo, 2003). Therefore, equilibrium $\delta^{13}\text{C}$ values of shell aragonite should be $\sim -3.3\text{‰}$. This finding, like those of other studies of unionids, is disappointing (e.g., Dettman et al., 1999). However, unlike other studies, data from *Pyganodon lacustris* indicate that it is responding directly to processes at work in this lake. Its growth within the *Chara* mat ringing the lake margin may provide a food source that quickly responds to changes in photosynthesis and respiration in the lake.

The work of Nichols and Garling (2000) and Christian et al. (2004) suggests that the food source of unionids is fine particulate organic matter that has a bacterial origin (not algal) and whose $\delta^{13}\text{C}$ and $\delta^{15}\text{N}$ values vary with season (summer FPOM has lower $\delta^{13}\text{C}$ values than fall FPOM). Further examination of unionid food sources and growth environments is critical in understanding the complicated record of shell $\delta^{13}\text{C}$ values. Among the carbon fluxes in lakes and rivers are: carbonate precipitation, formation and oxidation of organic matter (algae, bacteria, mosses, terrestrial plant material, etc.), methane generation (in lacustrine peats), and bicarbonate-rich ground water.

Corbicula fluminea - Indonesia

Corbicula fluminea and *Dreissena polymorpha* are two invasive species that recently have grown in number in the St. Lawrence Seaway, the Great Lakes, and the Mississippi River due to transport by ships that carry these species in ballast water (see Drake

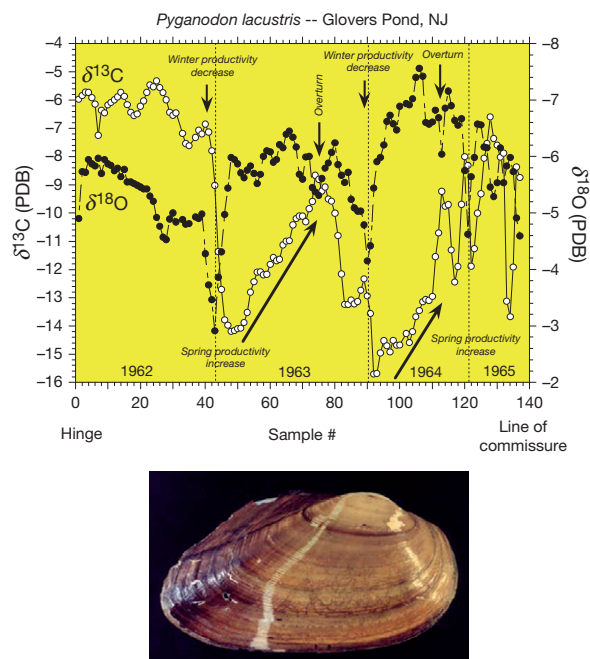


Figure 6 Plot of shell $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values from *Pyganodon lacustris* collected from Glovers Pond, New Jersey. Shell height is 43 mm. Data are plotted relative to sequential sample number to avoid compression of data. Samples were collected from the exterior of the shell (note white sampling swath) using manual micro-sampling techniques (from Carpenter and Erickson, in preparation). Winters represent a missing time/isotopic record.

and Bossenbroek, 2004; Holeck et al., 2004). As a result, these rather adaptive species have been studied extensively. Isotopic examination of their shells also has been conducted (Wurster and Patterson, 2001; Carpenter et al., 2005; Figures 8 and 9).

Corbicula sp. occurs naturally in streams and rivers in southeast Asia. A modern *Corbicula* sp. shell collected alive from a small stream in central Java has $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values that reflect a monsoonal moisture pattern (rainy and dry seasons; Figure 8(a)). The source of waters that feed these streams is rainfall on nearby volcanic peaks (~3000 meters high). Shell $\delta^{13}\text{C}$ values range from -14.9 to -12.3‰ and reflect a riparian vegetation of C3 plants (trees and shrubs).

The Upper Sangiran Formation in Central Java, Indonesia (~1.66 to 1.5 Ma) contains abundant *Corbicula* sp. shells in a paludal to lacustrine setting. This locality is of importance as it has yielded skeletal remains of *Homo erectus*. The first occurrence of *H. erectus* in the Upper Sangiran coincides with wet conditions and soil organic matter and fossil grasses that have high $\delta^{13}\text{C}$ values (~ -12‰ ; C4 Grasses). Grasses of this sort occur in poorly drained, wet conditions in southeast Asia (e.g., Johnston, 1996).

Corbicula sp. from this portion of the Upper Sangiran has $\delta^{13}\text{C}$ values that range from -7 to -1.8‰ (Figure 8(b)). Carbonate soil nodules from this unit have similarly high $\delta^{13}\text{C}$ values. Both soil and molluscan carbonates have $\delta^{13}\text{C}$ values that are ~10 to 11‰ higher than associated plant organic matter. This offset is consistent with the carbon isotope fractionation between CO_2 and carbonate (Emrich et al., 1970). Shell $\delta^{13}\text{C}$ values from modern *Corbicula* from this region are ~10‰ higher than their Pleistocene counterparts (Figure 8).

This suggests that the oxidation of plant material in soils and near streams is controlling the $\delta^{13}\text{C}$ value of shell carbonate (presumably via the $\delta^{13}\text{C}_{\text{DIC}}$). Given the tropical setting, cycling of carbon in these systems is likely very rapid, thus efficiently delivering CO_2 from organic matter decomposition to surface waters. This suggests that fossil mollusks from tropical regions may accurately record local vegetation.

In addition, the $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values of the Pleistocene *Corbicula* display no similarities with the modern monsoonal moisture pattern, but instead exhibit a sinusoidal trend reminiscent of temperate bivalves. In addition, Pleistocene $\delta^{18}\text{O}$ values are significantly higher than analogous modern data (Figure 8) suggesting either greater seasonal temperature variation and/or a markedly different source of precipitation that feeds local surface waters. Modern conditions are controlled by monsoonal rainfall with pronounced dry and rainy seasons. The dry season (May–Oct.) occurs when the ITCZ shifts to the north and dry air masses off Australia dominate. The rainy season (Nov. to April) occurs when the ITCZ shifts southward and humid air masses off SE Asia dominate. Paleosol and isotope data from the *Corbicula* in Figure 8(b) suggest that wet conditions persisted throughout the year without a dry season – perhaps during a glacial event.

Dreissena polymorpha - Keuka Lake, New York

Wurster and Patterson (2001) have conducted high-resolution sampling of the Zebra Mussel (*Dreissena polymorpha*) from Keuka Lake, New York (Figure 9). Shell $\delta^{18}\text{O}$ values closely match the equilibrium aragonite $\delta^{18}\text{O}$ values predicted by using measured water $\delta^{18}\text{O}$ values, the fractionation factor of Dettman et al. (1999) (Equation 5) and both water and air temperatures. Micro-sampling yielded 277 carbonate samples over 14 mm of shell growth (206 were analyzed). While one might debate the utility of sampling at this density, this work indicates that sampling methods can begin to address short-term events (i.e., storms, floods) in molluscan shell records.

Calcareous Algae and Aquatic Plants

Perhaps the most important producer of biogenic carbonate formed in freshwater is calcareous green algae. The family *Characea* (often referred to as Charophytes and *Chara*) is composed of macrophytic green algae that grow in oligotrophic ponds and lakes with relatively high bicarbonate contents. Most species of *Chara* sp. occur in the upper few meters of the water column and under elevated pH and $[\text{Ca}^{2+}]$ conditions, calcify heavily.

Chara, together with carbonate precipitation on aquatic plants such as *Potamogeton* are, to a large extent, responsible for the relatively rapid accumulation of carbonate sediments in marl lakes (Figure 7). This type of lacustrine calcification occurs predominantly during the late Spring to early Fall in temperate latitudes (due to elevated temperatures, photosynthesis, pH, and $[\text{HCO}_3^-]$; McConnaughey et al., 1994). Therefore, carbonates analyzed from lake sediment cores (marl lakes) must exhibit a strong bias toward Summer-time precipitation (e.g., Jones et al., 1996). This requires an explanation of stable isotope ratios that is consistent with physico-chemical conditions within the higher range of seasonal temperature variation (from ~20 to 35 °C for temperate settings). Another

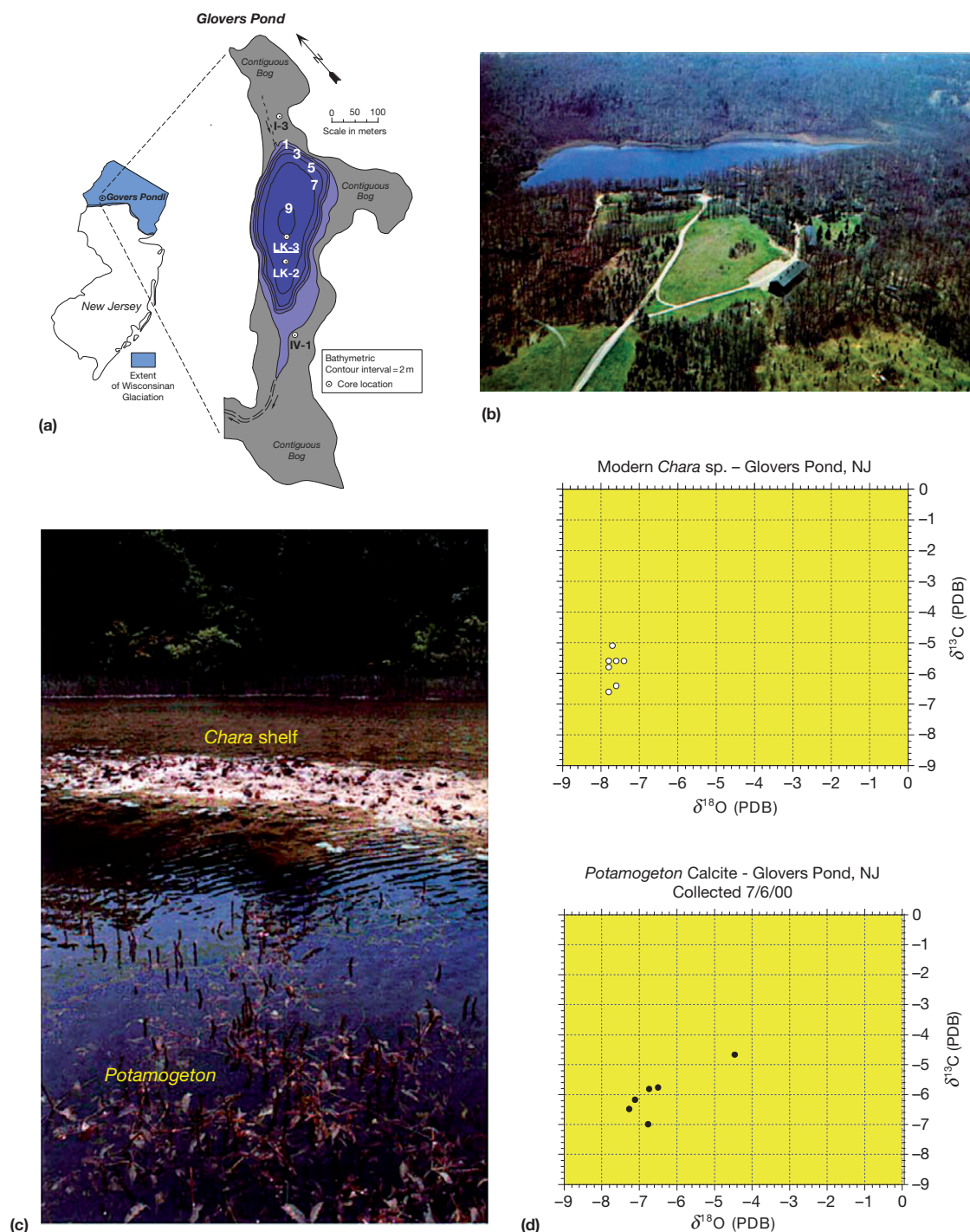


Figure 7 (a) Location and bathymetric map of Glovers Pond, New Jersey. (b) Aerial photograph of the Glovers Pond Basin with *Chara* shelf and bog areas visible. (c) Photograph of the margin of Glovers Pond with the well developed *Chara* shelf and actively calcifying *Potamogeton* growing on the shelf slope. (d) Plots of $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values of modern *Chara* and *Potamogeton* carbonate. $\delta^{13}\text{C}_{\text{DIC}}$ is -6.5‰ and $\delta^{18}\text{O}_{\text{water}}$ is -4.8‰ (SMOW) following overturns. Precipitation of these carbonates under summer temperatures finds that these are precipitated in isotopic equilibrium with ambient waters.

factor in the interpretation of the $\delta^{18}\text{O}$ values of *Chara* calcite is its Mg content (e.g., Tarutani et al., 1969; Anadón et al., 2002).

Interpretation of $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values collected from cores of lacustrine carbonate sediment requires an understanding of the calcification method and the degree to which it is precipitated in equilibrium with ambient waters. There is some debate

over the degree to which Charaphyte carbonate is precipitated in isotopic equilibrium with ambient water (Huon and Mojon, 1993; Jones et al., 1996; Coletta et al., 2001). Both *Chara* sp. and *Potamogeton* sp. from Glovers Pond, New Jersey precipitate carbonate in carbon and oxygen isotopic equilibrium with ambient lake waters (Carpenter and Erickson, in prep.).

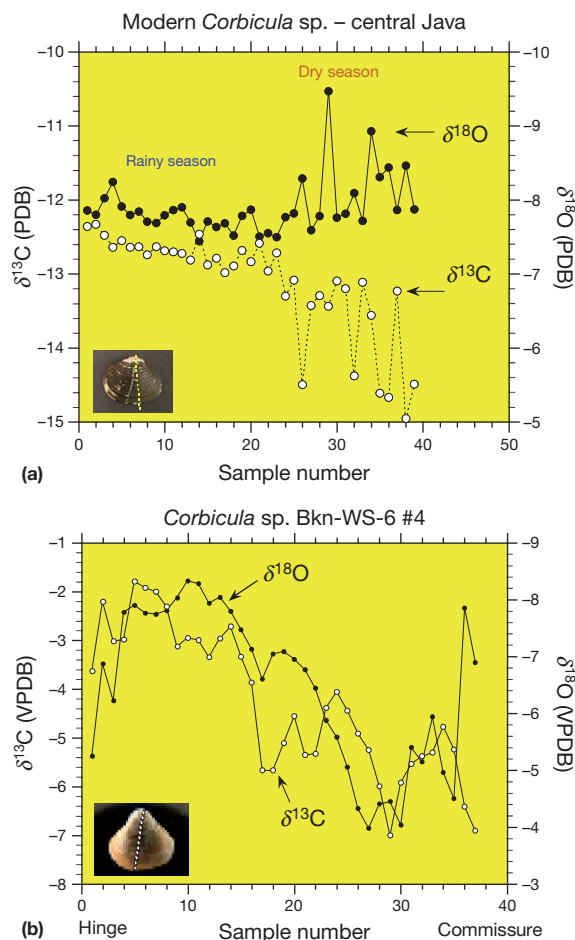


Figure 8 (a) Seasonal variation in $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values in a modern *Corbicula* sp. shell from Central Java. This plot depicts the differences between rainy and dry seasons. (b) Seasonal variation in $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values of a Pleistocene *Corbicula* sp. (Sangiran Fm) – note the higher $\delta^{13}\text{C}$ values and the presence of a sinusoidal trend – suggesting a lack of a monsoonal moisture pattern.

McConnaughey and Falk (1991) and McConnaughey (1991) describe possible calcification methods in *Chara* sp. where calcification occurs along alkaline bands on the exterior of the plant. Their data indicate a 1:1 ratio of calcification to photosynthesis. The extracellular precipitation of calcite by *Chara* as described by McConnaughey (1991) argues for minimal control of the calcification process by the plant and a greater likelihood of equilibrium precipitation (e.g., *Acetabularia* sp., Lee and Carpenter, 2001). If CO_2 hydroxylation plays a significant role in *Chara* calcification then one would anticipate a linear relation between $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values of its carbonate (McConnaughey, 1989a, b). In contrast, $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values collected from modern *Chara* exhibit minimal covariation in these values (Jones et al., 1996; Carpenter and Erickson, in prep.). Positive, linear covariation of $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values collected from carbonate-rich sediment cores (composed of predominantly Charophytes) from temperate, marl lakes (Drummond et al., 1995; Jones et al., 1996) is inconsistent with data from modern *Chara*. This linear covariation has been ascribed to a variety of lake physico-chemical

processes (stratification, CO_2 degassing, temperature variation, evaporation, seasonal changes in precipitation, etc.; Drummond et al., 1995; Jones et al., 1996). The time scales on which these data are formed (months for modern *Chara* and 10 to 100 years for sediment samples) likely account for some of this difference. As lake basins evolve over thousands of years, climate, reservoir size, vegetation, and various other parameters change. It is likely that this variation will impact the $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values of carbonate sediments. Clearly, more detailed examination of these processes is warranted.

Fish Otoliths

Otoliths (carbonate ear stones) of freshwater fish can be used to examine multi-year seasonal variation in $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values. The well-defined sclerochronology of sagittal otoliths (sagittae) provides a record that can be exploited using micro-sampling techniques (Figure 10). Typically aragonite, sagittae can be found in Quaternary age midden sites where fish were caught, eaten, and discarded (e.g., Wurster and Patterson, 2003). Preservation of otoliths is typically good in Quaternary deposits where materials have not been exposed to ground waters for prolonged periods of time. Stable isotope data have been collected from aragonite otoliths from Mesozoic age deposits (e.g., Patterson, 1999; Carpenter et al., 2003).

Two studies of the stable isotope ratios of freshwater fish otoliths are worthy of note. The first is an analysis of fossil freshwater drum (*Aplodinotus grunniens*) otoliths from an archaeological site in northeastern Tennessee (Figure 11; Wurster and Patterson, 2003). They examine intra-otolith variation in $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values from specimens that span the last 5500 years. This study contrasts data from high- and low-resolution sampling (Figure 11). This comparison points out how high-resolution sampling provides significantly more information about fish life histories (Figure 11(a)–(h)). Wurster and Patterson (2003) use the $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values of otoliths from different periods in the Holocene to estimate changes in bioenergetics induced by climate (a combination of temperature, metabolic activity, and other factors).

Wurster et al. (2005) examine intra-otolith variation of $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ value in chinook salmon (*Oncorhynchus tshawytscha*) from Lake Ontario (Figure 12). They study thermal histories (otolith $\delta^{18}\text{O}$ thermometry), stress, and metabolism (otolith $\delta^{13}\text{C}$ values) in this migratory fish. During the summer these fish inhabit epilimnetic waters with temperatures of ~ 19 – 20°C (Figure 12). Chinook would approach but rarely exceed their reported upper incipient lethal limit of approximately 22°C , which suggests that these fish were seeking water with temperatures as high as was tolerable while otolith growth occurred. These results contrast with expected midsummer temperatures for this typically cold-water fish. Bioenergetic simulations find that these salmon undergo significant stress at these temperatures. Increased catabolism and decreased gross conversion efficiency relative to nominal simulations where chinook salmon are modeled nearer their preferred temperature, confirms previous inferences that the chinook salmon population may be near the limits of sustainability (Wurster et al., 2005).

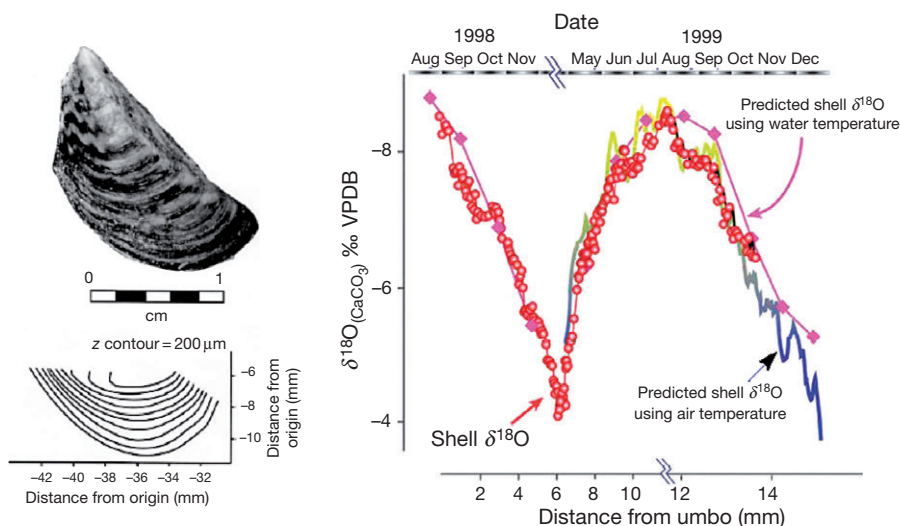


Figure 9 Example of high-resolution microsampling from freshwater bivalve *Dreissena polymorpha* from Keuka Lake, central New York (from Wurster and Patterson, 2001b). Digital image of *D. polymorpha* and illustration of a portion of the morphology of *D. polymorpha* determined using digital path coordinates; Comparison between predicted shell $\delta^{18}\text{O}$ values and measured shell $\delta^{18}\text{O}$ values for *D. polymorpha*. In order to satisfactorily compare the data and account for seasonal changes in the deposition of carbonate, different growth scales were used. A different growth scale was used from the umbo to the minimum shell $\delta^{18}\text{O}$ value and from the minimum shell $\delta^{18}\text{O}$ value to the last sample. In order to show a more detailed comparison, predicted shell $\delta^{18}\text{O}$ values determined using air temperatures from Rochester, New York and the fractionation factor of Dettman et al. (1999) (Eqn [5]) are also shown. Note that the scale was not changed for the date axis. However, Nov. through Mar. are omitted because predicted shell $\delta^{18}\text{O}$ values could not be determined for these months. Modified from Wurster and Patterson (2001b).

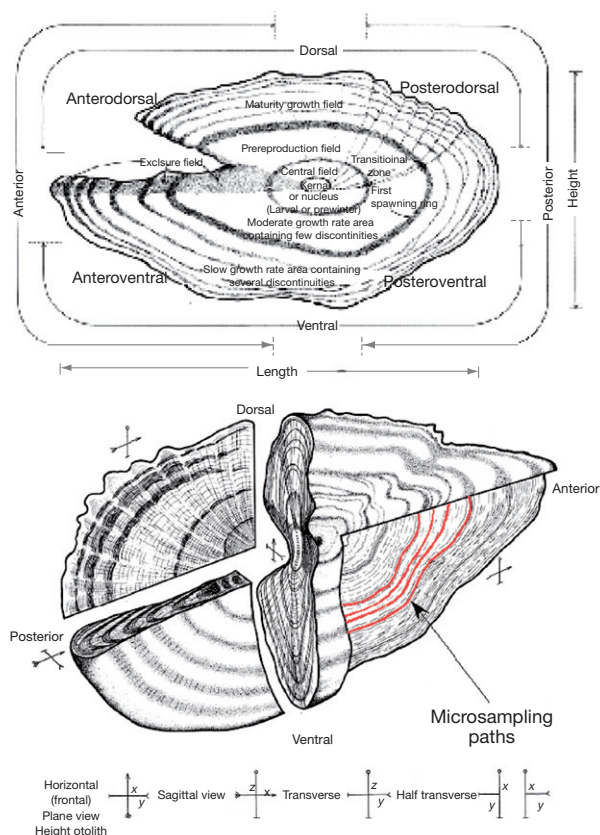


Figure 10 Schematic representation of a sagittal otolith depicting idealized growth bands (timelines). Microsampling paths follow these bands to extract intraotolith variation in $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values. Modified from Pannella (1980).

Discontinuous Records of Climate Change

The occurrence of freshwater biogenic carbonates during the Quaternary is quite variable. The most continuous carbonate records are of *Chara* sp. in lacustrine settings (i.e., lake sediment cores). Where rivers transport mollusks into lake basins, there is a potential to preserve more continuous records of mollusk shells. However, riverine deposits are by nature, discontinuous due to a variety of geological processes (e.g., meandering and channel migration, base level change and erosion, etc.). *In situ* riverine deposits containing mollusks and fish otoliths typically permit brief, episodic glimpses of seasonality and climate during periods of rapid deposition (e.g., flood events). Somewhat more continuous records of mollusk shells can be found in midden deposits. Where such deposits occur on flood plains, their continuous nature relies on the depositional and/or erosional qualities of the river and the extent to which humans or other organisms occupy a given site. Careful attention must be paid to the stratigraphic position, dating techniques used, position of paleosols, and bioturbation of midden deposits. In deposits where fire was used, researchers must use caution when analyzing charred biogenic carbonates.

Collection of long-term, continuous climate records from Quaternary age freshwater mussels and fish otoliths is not likely. Instead, isotopic records can be collected from relatively short intervals of time (typically a few thousand years) dictated by climatic conditions that are conducive to midden production. On a brighter note, well preserved specimens can yield a great deal of information regarding changes in seasonal temperature and water composition. Likewise, season of collection can be obtained in archaeological materials – a useful tool in understanding the timing of human habitation. For long-term climate interpretation, the best-case scenario is one in which well preserved materials occur episodically throughout a deposit where co-occurring dateable materials can provide precise limits on the timing of carbonate deposition (in particular, climatically important time intervals during the late Pleistocene and Holocene).

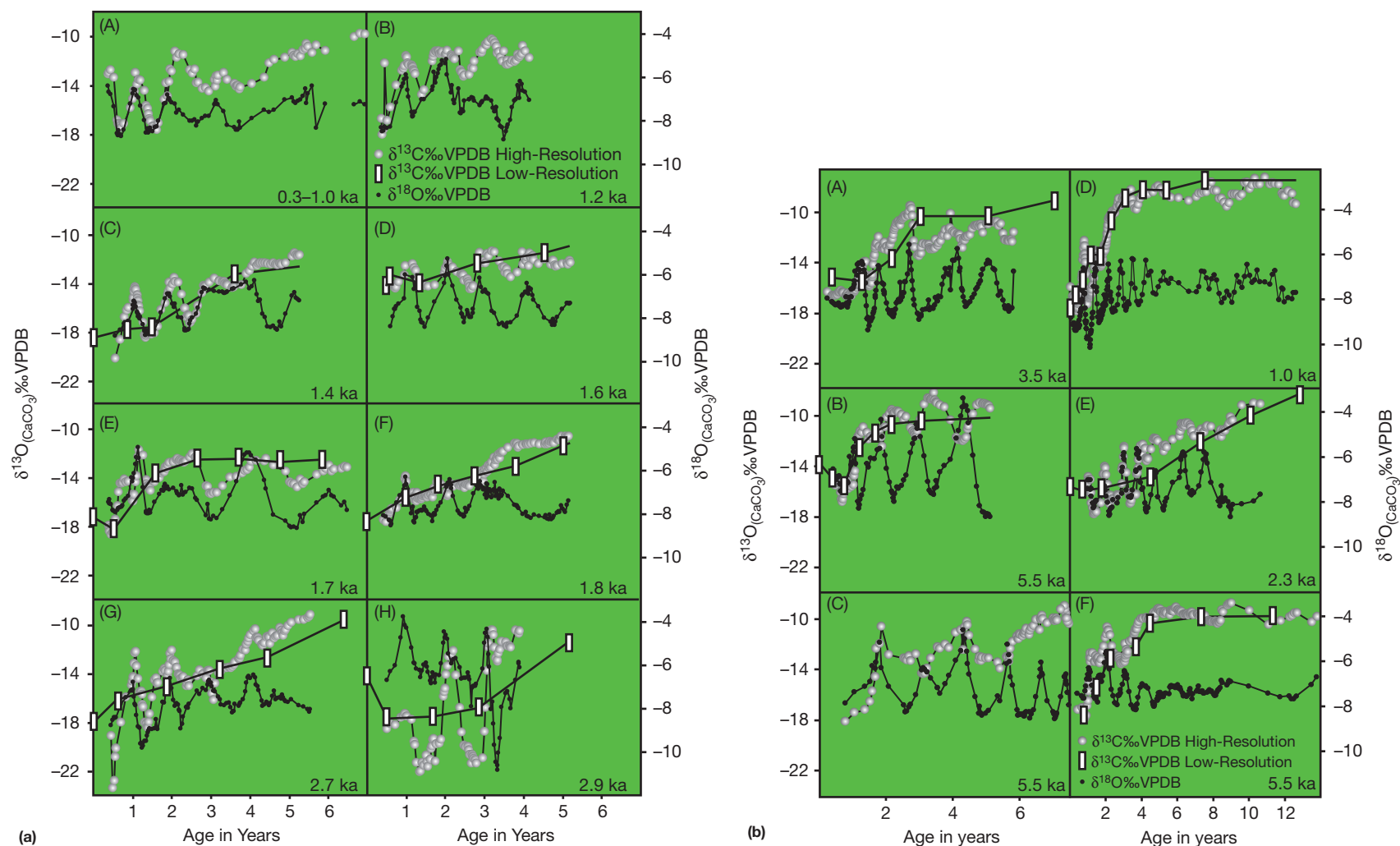


Figure 11 Otolith $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values of the freshwater drum *Aplodinotus grunniens* from northeastern Tennessee (Wurster and Patterson, 2003). (a) Intraotolith variation in high-resolution $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values of micromilled sagittae from time periods 0.3–2.9 ka. Samples were taken only from the first 8 yr of the sagittae or less. Each subplot represents an individual otolith. Each high-resolution sample represents an individual carbonate microsample. Low-resolution samples that approximate the average values of microsamples are plotted for comparison. Represented time periods are 0.3–1.0, 1.2, 1.4, 1.6, 1.7, 1.8, 2.7, and 2.9 ka. Specimens from A and B were destroyed before low-resolution sampling could be performed. (b) Intra-otolith variation in high-resolution $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values of micro-sampled sagittae from time periods 1.0 Ka, 2.3 Ka, 3.5 Ka, and 5.5 Ka. Each subplot represents an individual otolith. Each high-resolution sample represents an individual carbonate aliquot removed by micro-sampling. Low-resolution samples that approximate the average values of micro-sampled aliquots are plotted for comparison. Note different age axes on subplots. Subplots on the left side include samples only from the first eight years of the sagittae or less, whereas subplots on the right include samples from beyond the first eight years. The specimen from C was destroyed before low-resolution sampling could be performed. From Wurster and Patterson (2003).

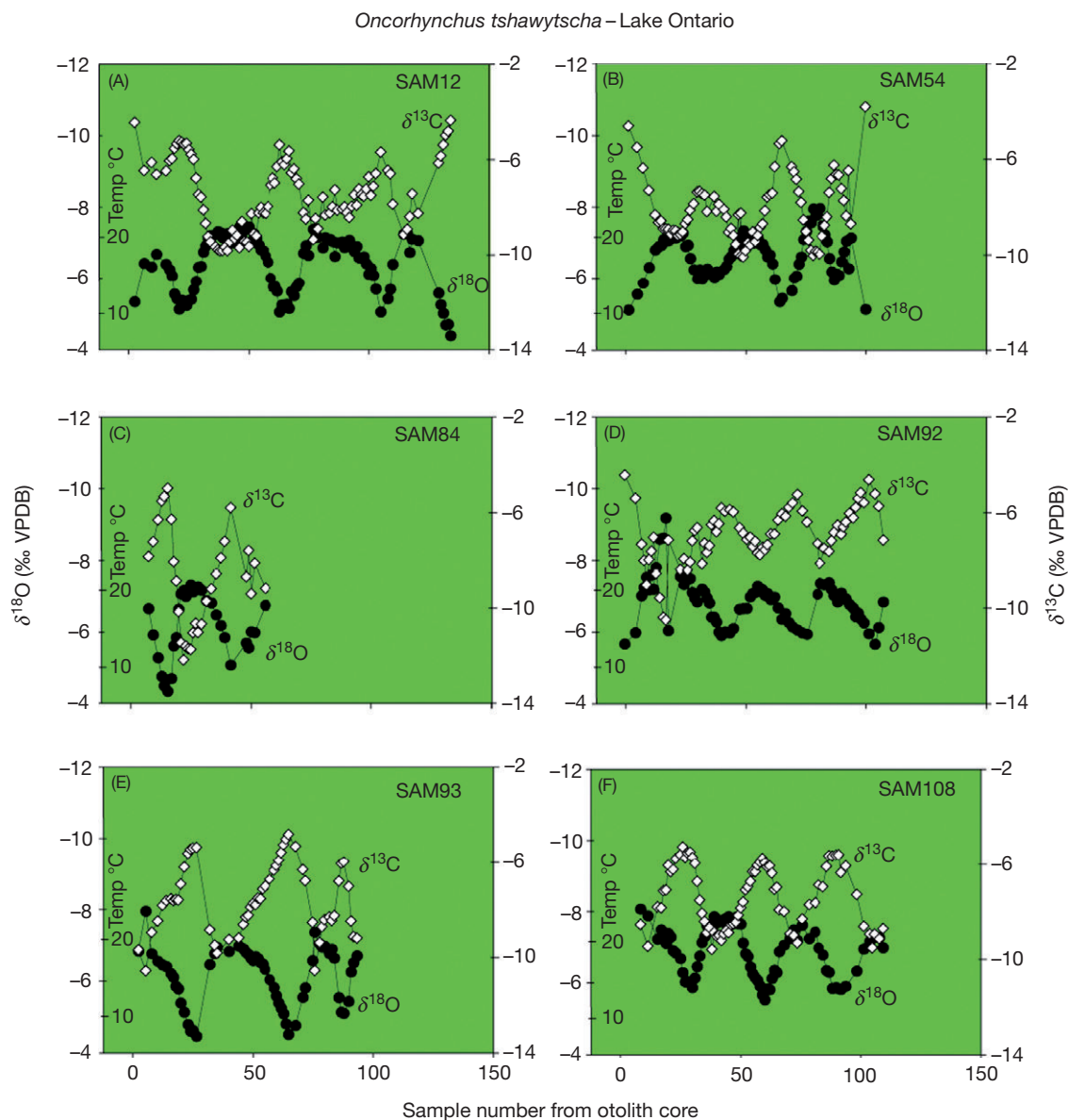


Figure 12 Intraotolith $\delta^{18}\text{O}$ values and calculated temperatures (●) and $\delta^{13}\text{C}$ values (◊) relative to VPDB for the archived group of chinook salmon (*Oncorhynchus tshawytscha*) in Lake Ontario. Archived specimens were collected in 1991 from offshore waters of Lake Ontario. Each subplot represents one otolith from an individual fish. The $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values are plotted according to the number of the samples from the core of the otolith; note that $\delta^{18}\text{O}$ values are plotted with more negative values toward the top. The temperature scale (inset on y-axis) is calculated using a Lake Ontario $\delta^{18}\text{O}$ value of -6.8‰ (VSMOW) and Patterson et al.'s (1993) temperature–fractionation relationship for otolith aragonite of freshwater fish. Modified from Wurster CM, Patterson WP, Stewart DJ, Stewart TJ, and Bowlby JN (2005). Thermal histories, stress and metabolic rates of chinook salmon in Lake Ontario: Evidence from intra-otolith $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values and energetics modeling. *Canadian Journal of Fisheries and Aquatic Sciences* 62: 700–713.

Conclusions

Stable isotope data from freshwater biogenic carbonates have great potential for use in evaluating a variety of environmental conditions in the Quaternary record. Among these are: temperature, $\delta^{18}\text{O}$ value of surface waters, vegetation, timing of seasonal surface water productivity changes, and lake stratification. To enhance the utility of these measurements several advances need to be made: 1) Additional calibration of

geologically significant organisms, 2) Additional examination of the mechanisms responsible for biological fractionations or 'vital effects', and 3) Simultaneous analysis of carbonate minor element chemistry to develop an independent paleothermometer (e.g., Sr/Ca and Mg/Ca ratios). Elemental analyses come with their own caveats – such as calibration of modern materials, and understanding carbonate precipitation rates. Even with such advances, examination of multiple proxies is highly recommended (e.g., mollusks, otoliths, soil

organic matter, stalagmites, lacustrine sediments (carbonate, organic matter, pollen, mites, ostracodes, etc.).

See also: Carbonate Stable Isotopes: Speleothems. Ice Core
Methods: Methane Studies; Overview.

References

- Adkins JF, Boyle EA, Curry WB, and Lutringer A (2003) Stable isotopes in deep-sea corals and a new mechanism for "vital effects" *Geochim. Cosmochim. Acta* 67: 1129–1143.
- Anadón P, Utrilla R, and Vazquez A (2002) Mineralogy and Sr-Mg geochemistry of charophyte carbonates: A new tool for paleolimnological research. *Earth and Planetary Science Letters* 197: 205–214.
- Carpenter SJ, Erickson JM, and Holland FD (2003) Migration of a Late Cretaceous fish. *Nature* 423: 70–74.
- Carpenter SJ and Lohmann KC (1995) $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values of modern brachiopod shells. *Geochim. Cosmochim. Acta* 59: 3749–3764.
- Carpenter SJ and Erickson JM (in preparation) $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ values of shell carbonate from the freshwater mussel *Pyganodon lacustris*: Temperature, stratification, and productivity records of Glovers Pond, New Jersey. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 21ms. pgs.
- Carpenter SJ, Weirich FH, Blesius LJ, and Erickson JM (in preparation) A record of discharge and turbidity in the Mackenzie River, NWT, Canada: A coordinated study of stable isotope data from *Lampsilis radiata siliquoides* and remote sensing. *Geology*, 12ms. pgs.
- Carpenter SJ, Bettis EA III, Milius AK, and Ciochon RL (2005) Carbon isotope ratios of tropical freshwater mollusk shells as indicators of ancient vegetation – reconstructing paleoenvironments at a Homo erectus locality in the Solo Basin, Central Java, Indonesia. *Geological Society of America, Abstracts with Program* 37(7): 363.
- Christian AD, Smith BN, Berg DJ, Smoot JC, and Findlay RH (2004) Trophic position and potential food sources of 2 species of unionid bivalves (Mollusca: Unionidae) in 2 small Ohio streams. *Journal of the North American Benthological Society* 23(1): 101–113.
- Coletta P, Pentecost A, and Spiro B (2001) Stable isotopes in charophyte incrustations: Relationships with climate and water chemistry. *Palaeogeography, Palaeoclimatology, Palaeoecology* 173: 9–19.
- Dettman DL and Lohmann KC (1995) Approaches to microsampling carbonates for stable isotope and minor element analysis: Physical separation of samples on a 20 micrometer scale. *Journal of Sedimentary Petrology* v. A65: 566–569.
- Dettman DL, Reische AK, and Lohmann KC (1999) Controls on the stable isotope composition of seasonal growth bands in aragonitic fresh-water bivalves (Unionidae). *Geochimica et Cosmochimica Acta* v. 63: 1049–1057.
- Drake JM and Bossenbroek JM (2004) The potential distribution of Zebra Mussels in the United States. *BioScience* 54: 931–941.
- Drummond CN, Patterson WP, and Walker JCG (1995) Climate forcing of carbon-oxygen isotopic covariance in temperate region marl lakes. *Geology* 23: 1031–1034.
- Emrich K, Ehrlert DH, and Vogel JC (1970) Carbon isotope fractionation during the precipitation of calcium carbonate. *Earth and Planetary Science Letters* 8: 363–371.
- Erickson JM (1968) *The geologic and limnologic history of Glovers Pond, northwestern New Jersey*, p. 149. Unpublished M.S. Thesis, University of North Dakota, Grand Forks, North Dakota.
- Frappier A, Sahagian D, González LA, and Carpenter SJ (2002) El Niño events recorded by stalagmite carbon isotopes. *Science* 298: 565.
- Grossman EL and Ku T-L (1986) Oxygen and carbon isotope fractionation in biogenic aragonite: Temperature effects. *Chem. Geol.* 59: 59–74.
- Holeck KT, Mills EL, Maciassac HJ, Dochoda MR, Colautti RI, and Ricciardi A (2004) Bridging Troubled Waters: Biological Invasions, Transoceanic Shipping, and the Laurentian Great Lakes. *BioScience* 54: 919–929.
- Howard AD (1921) Experiments in the culture of fresh-water mussels. *Bulletin of the U.S. Bureau of Fisheries* 38: 63–90.
- Huon S and Mojon P-O (1994) Déséquilibre isotopique entre eau et parties calcifiées de Charophytes actuels (*Chara globularis*): résultats préliminaires. *C.R. Acad. Sci. Paris série II* 318: 205–210.
- Johnston WH (1996) The place of C4 grasses in temperate pastures in Australia. *New Zealand Journal of Agricultural Research* 39: 527–540.
- Jones TJ, Fortier SM, Pentecost A, and Collinson ME (1996) Stable carbon and oxygen isotopic compositions of recent charophyte oosporangia and water from Malham Tarn, U.K.: Paleontological implications. *Biogeochemistry* 34: 99–112.
- Joy JE (1985) A 40-week study on growth of the Asian Clam, *Corbicula fluminea* (Müller) in the Kanawha River, West Virginia. *Nautilus* 99: 110–116.
- Kim S-T and O'Neil JR (1997) Equilibrium and nonequilibrium oxygen isotope effects in synthetic carbonates. *Geochimica et Cosmochimica Acta* 61: 3461–3476.
- Lee D and Carpenter SJ (2001) Isotopic Disequilibrium in Marine Calcareous Algae. *Chemical Geology – Isotope Geoscience Section* 172: 307–329.
- McConnaughey T (1989a) ^{13}C and ^{18}O isotopic disequilibrium in biological carbonates. I: Patterns. *Geochimica et Cosmochimica Acta* 53: 151–162.
- McConnaughey T (1989b) ^{13}C and ^{18}O isotopic disequilibrium in biological carbonates. II: In vitro simulation of kinetic isotope effects. *Geochimica et Cosmochimica Acta* 53: 163–171.
- McConnaughey T (1991) Calcification in *Chara corallina*: CO_2 hydroxylation generates protons for bicarbonate assimilation. *Limnol. Oceanogr.* 36: 619–628.
- McConnaughey T and Falk RH (1991) Calcium-proton exchange during algal calcification. *Biol. Bull.* 180: 185–195.
- McConnaughey TA, LaBaugh JW, Rosenberry DO, Reddy MM, Schuster PF, and Carter V (1994) Carbon budget for a groundwater-fed lake: Calcification supports summer photosynthesis. *Limnol. Oceanogr.* 39: 1319–1332.
- Negus CL (1966) A quantitative study of growth and production of unionid mussels in the River Thames at Reading. *J. Anim. Ecol.* 35: 513–532.
- Neumann D, Borcherting J, and Jantz B (1993) Growth and seasonal reproduction of *Dreissena polymorpha* in the Rhine River and adjacent waters. In: Nalepa TF and Schloesser DW (eds.) *Zebra Mussels: Biology, Impacts, and Control*, pp. 95–110. Lewis Publishers.
- Nichols SJ and Garling D (2000) Food-web dynamics and trophic-level interactions in a multispecies community of freshwater unionids. *Canadian Journal of Zoology* 78: 871–882.
- Pannella G (1980) Growth patterns in fish sagittae. In: Rhoads DC and Lutz RA (eds.) *Skeletal Growth of Aquatic Organisms – Biological records of environmental Change*, pp. 519–560.
- Patterson WP (1998) North American continental seasonality during the last millennium: High-resolution analysis of sagittal otoliths. *Palaeogeography, Palaeoclimatology, Palaeoecology* 138: 271–303.
- Patterson WP (1999) Oldest isotopically characterized fish otoliths provide insight to Jurassic continental climate of Europe. *Geology* 27: 199–202.
- Patterson WP and Smith GR (2002) Uses of fishes in analysis of paleolimnology in Tracking Environmental Change Using Lake Sediments: Zoological Indicators. In: Smol JP, Birks HJB, and Last WM (eds.) *Paleolimnology*, Special vol. 4, ch. 9, pp. 173–187.
- Patterson WP, Smith GR, and Lohmann KC (1993) Continental paleothermometry and seasonality using the isotopic composition of aragonitic otoliths of freshwater fishes. In: Swart P, Lohmann KC, McKenzie J, and Savin S (eds.) *Amer. Geophys. Union Monogr. Continental Climate Change from Isotopic Indicators*, pp. 191–202.
- Quevedo HL (2003) A high resolution stable isotope record of climate change from NW New Jersey, p.159. Unpublished Ph.D. Dissertation, University of Iowa: Iowa City, Iowa.
- Ricken W, Steuber T, Freitag H, Hirschfeld M, and Niedenzu B (2003) Recent and historical discharge of a large European river system – oxygen isotopic composition of river water and skeletal aragonite of Unionidae in the Rhine. *Palaeogeography, Palaeoclimatology, Palaeoecology* 193: 73–86.
- Romanek CS, Grossman EL, and Morse JW (1992) Carbon isotopic fractionation in synthetic aragonite and calcite: Effects of temperature and precipitation rate. *Geochimica et Cosmochimica Acta* 56: 419–430.
- Rubinson M and Clayton RN (1969) Carbon-13 fractionation between aragonite and calcite. *Geochimica et Cosmochimica Acta* 33: 997–1002.
- Tarutani T, Clayton RN, and Mayeda TK (1969) The effect of polymorphism and magnesium substitution on oxygen isotope fractionation between calcium carbonate and water. *Geochimica et Cosmochimica Acta* 33: 987–996.
- Tevesz MJS and Carter JG (1980) Environmental relationships of shell form and structure of Unionacean bivalves. In: Rhoads DC and Lutz RA (eds.) *Skeletal Growth of Aquatic Organisms – Biological records of environmental Change*, pp. 295–322.
- Urey HC, Lowenstam HA, Epstein S, and McKinney CR (1951) Measurement of paleotemperatures and temperatures of the upper Cretaceous of England, Denmark, and the southeastern United States. *Bull. Geol. Soc. Am.* 62: 399–416.
- Wurster CH, Patterson WP, and Cheatham MM (1999) Advances in computer-based microsampling of biogenic carbonates. *Computers in Geosciences* 25: 1155–1162.

- Wurster CM and Patterson WP (2001) Stable oxygen and carbon isotope values recovered from lacustrine freshwater mollusks: Paleoclimatic implications for sub-weekly temperature records. *Journal of Paleolimnology* 26: 205–218.
- Wurster CM and Patterson WP (2003) Late Holocene metabolic rate changes of freshwater drum (*Aplodinotus grunniens*): Evidence from high-resolution sagittal otolith stable isotope ratios of carbon. *Paleobiology* 29: 492–505.
- Wurster CM, Patterson WP, Stewart DJ, Stewart TJ, and Bowlby JN (2005) Thermal histories, stress, and metabolic rates of chinook salmon in Lake Ontario: Evidence from intra-otolith $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values and energetics modeling. *Canadian Journal of Fisheries and Aquatic Sciences* 62: 700–713.
- Zeebe RE (1999) An explanation of the effect of seawater carbonate concentration on foraminiferal oxygen isotopes. *Geochimica et Cosmochimica Acta* 63: 2001–2007.

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Carbon Dioxide, Atmospheric Concentrations *see* **Carbonate Stable Isotopes:** Overview; **Ice Core Methods:** CO₂ Studies; **Paleoceanography, Physical and Chemical Proxies:** Carbon Cycle Proxies ($\delta^{11}\text{B}$, $\delta^{13}\text{C}_{\text{calcite}}$, $\delta^{13}\text{C}_{\text{organic}}$, Shell Weights, B/Ca, U/Ca, Zn/Ca, Ba/Ca); **Plant Macrofossil Methods and Studies:** CO₂ Reconstruction from Fossil Leaves

Cave Art *see* **Vertebrate Studies:** Interactions with Hominids

Charcoal Studies *see* **Paleobotany:** Charred Particle Analyses

CHIRONOMID RECORDS

Contents

Chironomid Overview

Africa

Late Pleistocene of Europe

Postglacial Europe

Postglacial Southern Hemisphere

Chironomid Overview

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Introduction

The Chironomidae constitute a family of true flies (Diptera) prominent as living larvae in the bottom communities of almost all freshwater habitats. They are abundantly preserved in Quaternary lake sediments, and are widely used for paleoenvironmental reconstruction. The biology and applications of Chironomidae in Quaternary science have been extensively reviewed by [Armitage et al. \(1995\)](#), [Frey \(1976, 1988\)](#), [Hofmann \(1986, 1988\)](#), [Porinchu and MacDonald \(2003\)](#), and [Walker \(1987, 1993, 1995, 2001\)](#). This entry provides a summary of these, and other important and relevant works.

Adult Chironomidae seldom bear a proboscis, and although they and their phantom midge cousins (family Chaoboridae) are mosquito-like, neither family shares the biting habit of mosquitoes (family Culicidae). Consequently, chironomids are commonly referred to as non-biting midges.

Three related families of aquatic midges do bite as adults and are also preserved as head capsules in lake sediments: Ceratopogonidae ('no-see-ums' or biting midges), Culicidae

(mosquitoes), and Simuliidae (black flies or buffalo gnats). Members of these families are vectors for many diseases of animals, including humans. Although Chaoboridae do not have a head capsule record, their mandibles are common fossils. Collectively, the preserved remains of Chironomidae, Chaoboridae, Culicidae, Ceratopogonidae, and Simuliidae are often referred to as fossil aquatic midges. The earliest known fossils of Chironomidae are of Triassic age.

The larvae, pupae and adults of all of these groups are important food items in the diets of fish. The adults (imagoes) are an important food for swallows and martins.

Biology

Chironomidae are amongst the most ubiquitous of insects, found on islands distributed throughout the world's oceans, and on all continents from the lowland tropics to the high arctic, Antarctic, and alpine. Chironomidae have a life cycle consisting of four stages: egg, larva, pupa, and imago.

Upon hatching, the first instar larva begins a period of growth, which eventually necessitates the shedding of the chitinous exoskeleton. Upon replacing the exoskeleton, the larva then continues growth in its second instar. Two further episodes of ecdysis (replacement of the exoskeleton) and growth follow, defining the third and fourth larval instars.

Although bearing a well developed, strongly sclerotized head capsule, the elongate soft-bodied larvae otherwise resemble maggots. As mature fourth instar larvae, they vary from 1 to 30 mm in length. Lacustrine chironomid larvae inhabit the uppermost sediments of both littoral and profundal (deepwater) environments, cling or burrow into aquatic plants, tunnel in moist wood, or live as parasites or commensal partners on other invertebrates. Similar habitats are occupied in streams, and chironomid taxa are known to inhabit littoral marine environments, water-logged soils, peats, and dung.

The first instar larva of lacustrine midges is generally planktonic, allowing for dispersal of larvae to suitable substrates. Larvae from habitats other than standing water, and the later instars of lacustrine midges, are largely sedentary and benthic (bottom dwelling). The first instar is typically brief, with each later instar usually of progressively longer duration. Although the first instar larvae may derive some nourishment from the egg yolk, they feed on a variety of materials including algae, organic detritus, aquatic plants, and other invertebrates. Some are predaceous, whereas most other chironomids combine algae and organic detritus to form their diet.

During the latter part of the fourth larval instar, the thoracic segments expand, achieving a pre-pupal condition. Following ecdysis, the chironomid enters the pupal stage in which morphology is re-organized to that of the imago (adult). Apart from the Tanypodinae, almost all pupae are sedentary. The pupal stage is brief, persisting for a few days at most. At maturity, the pupa moves to the water surface, allowing emergence of the winged adult.

Lasting for a few weeks only, the adult stage permits dispersal and reproduction. The duration of the midge life cycle varies greatly among taxa, and is also related to habitat. Long life cycles typify arctic taxa, occasionally extending to several years. In warmer climates, several generations are common in a single season.

Classification

Eleven subfamilies constitute the Chironomidae. One subfamily, the Telmatogetoninae, is almost completely restricted to marine environments. However, in Hawaii this family also inhabits fresh water. Three subfamilies are very rare, or known only from a small segment of the world. For example, the Chilenomyiinae, represented by a single species, is known only from southern Chile. The Buchonomyiinae includes three species native to Europe, Asia (Burma), and Central America (Costa Rica). The Usumbaromyiinae is a new subfamily recently described from Tanzania (Andersen and Sæther, 1994).

The subfamily Aphroteniinae, including four genera, is known only from South America, South Africa, and Australia. Such a distribution suggests that this primitive group originated before the Mesozoic disintegration of Gondwanaland. An extinct Aphroteniinae tribe, the Electroteniini, records this subfamily's former, more widespread Cretaceous distribution, extending into Siberia.

Three subfamilies, the Orthoclaadiinae, Tanypodinae, and Chironominae, are widely distributed and constitute the majority of taxa encountered in lake sediments. The remaining subfamilies, Podonominae, Diamesinae, and Prodiamesinae, are predominantly cold-stenothermous taxa inhabiting temperate and montane, to polar and alpine climatic regions.

Fossils and their Identification

Taphonomy

Given an appropriate sedimentary environment, remains of larvae, pupae, and adults (imagoes) may all be preserved. Imagoes are well preserved in Cretaceous amber, but in lake sediments their bodies are commonly disarticulated. Occasionally the hypopygium, a part of the male imago's genitalia, is preserved and allows species-level identifications. Thoracic horns of Tanypodinae pupae have rarely been found in lacustrine deposits, and the thin, delicate bodies of larvae and pupae are too fragile to normally preserve well. In contrast, Quaternary interglacial lake deposits typically contain 50 to 200 well-preserved larval head capsules per cubic centimeter. Thus, these well sclerotized head capsules are far more abundant in lakebeds than the remains of all other insects combined. They provide the basis for almost all Quaternary paleoenvironmental work involving the Chironomidae.

One head capsule is produced by each chironomid, during each larval instar. Thus, a chironomid surviving to the fourth larval instar will have produced four head capsules. A chironomid expiring prior to attaining its fourth instar will produce fewer head capsules. The head capsules subsequently preserved could include both those shed with exuviae during ecdysis, and the remains of dead larvae. However, there exist numerous reasons why the abundance of fossil midge remains might not reflect the abundance of the living larvae from which they were derived.

Quality of preservation may vary among taxa, among instars, and between head capsules derived from expired insects and those shed as exuviae. Bivoltine and multivoltine taxa produce more head capsules per season than univoltine taxa of the same abundance. Re-distribution of head capsules will influence the concentration and composition of fossil assemblages. Remains of stream Chironomidae are often deposited in lakes, and remains of littoral species may be transported offshore to deepwater sites. Erosion along lakes and streams facilitates the redeposition of semi-terrestrial species into lacustrine settings.

Iovino (1975) determined that first and second instars of *Chironomus attenuatus* usually dissolve the procuticle completely prior to ecdysis, leaving only a thin, non-chitinous, epicuticular exuviae. In contrast, third instar exuviae usually include chitin, and fourth instars of *Chironomus attenuatus* were chitinous almost without exception. Thus, exuviae of early instars are much less durable than both those of later instars and remains of expired larvae.

These experiments have only been conducted on *Chironomus attenuatus*. The results possibly apply to all chironomid species. This, in part, would explain the great under-representation of early instar head capsules in lake sediments. Early instar remains may also be lost through the use of coarse sieves during processing.

Tanypodinae head capsules are more delicate than those of other taxa, and may be poorly preserved. First instar head

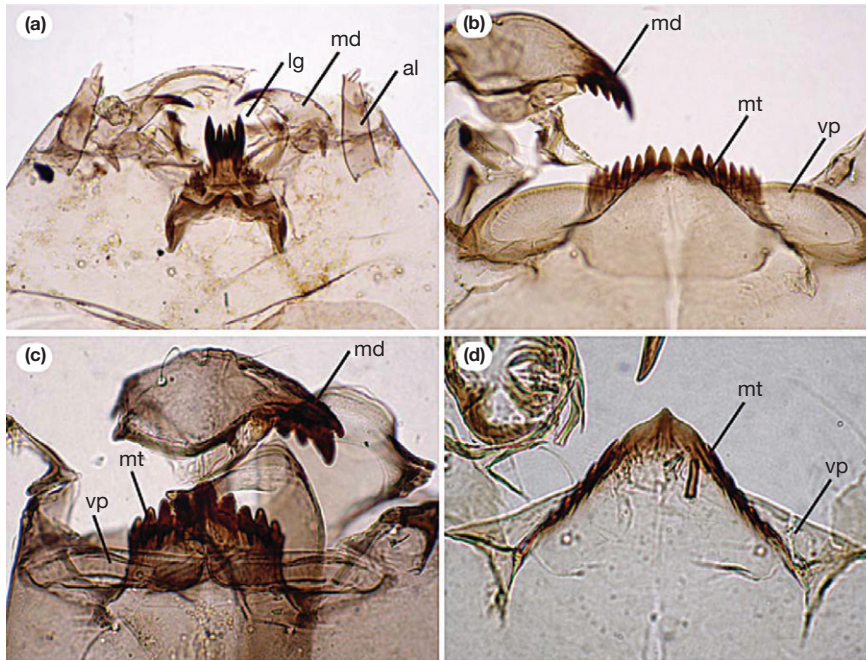


Figure 1 Representative fossils of Chironomidae, illustrating the principal structures used in their identification: (a) *Procladius* (subfamily Tanypodinae), (b) *Sergentia* (tribe Chironomini, subfamily Chironominae), (c) Tribe Tanytarsini (subfamily Chironominae), (d) *Psectrocladius* cf. *calcaratus* (Subfamily Orthocladiinae). Abbreviations: a1, 1st antennal segment; lg, ligula; md, mandible; mt, mentum; vp, ventromental plate.

capsules of all chironomid taxa are normally uncommon to rare as fossils, but may dominate records preserved in the anoxic, deepwater regions of meromictic basins.

Fossil Morphology

Identification of Quaternary fossils relies principally on the morphology of the head capsule, although other larval body parts (e.g., antennae and various mouthparts) are useful when they are still attached to the head. The structure of the head capsule varies, but the greatest part of this variation may be illustrated with reference to fossils of the subfamilies Chironominae, Orthocladiinae, and Tanypodinae (Figure 1).

The head capsules of Tanypodinae (Figure 1(a)) are generally weakly pigmented and somewhat elongate. The retracted first segment of each antenna is often visible within fossil head capsules. The proportions (ratio of length to width) of the head capsule and first antennal segment can be important characters for fossil identification. A fork-shaped ligula is centrally located within the head capsule. The ligula may become separated from head capsules during burial or processing of fossil material. Variations in the shape, number, and color of ligula teeth also aid identification. The dorsomental teeth, lacking in the tribe Pentaneurini, are otherwise situated laterally adjacent to the ligula and will usually remain with the head capsule. In well preserved material, one or both mandibles may be retained. The arrangement of minute hairs (cephalic setae) on the fossil head capsule is also important for identification of larval fossils (Rieradevall and Brooks, 2001).

The principal features facilitating identification of fossil Chironominae (Figures 1(b) and (c)) and Orthocladiinae (Figure 1(d)) are the ventromental plates and the teeth of the mentum. The mentum is situated in an anterior median ventral position. The pair of ventromental plates are situated laterally,

adjacent to the mentum on either side. In the Chironominae, the ventromental plates are very large and usually conspicuously striated. Although fan-shaped in the tribe Chironomini (Figure 1(b)), laterally elongated, strap-shaped plates occur on members of the Tanytarsina (a subtribe of the tribe Tanytarsini) (Figure 1(c)), and in the tribe Pseudochironomini.

In the Orthocladiinae (Figure 1(d)), the ventromental plates are typically much less conspicuous, and never striated. In many Orthocladiinae, these plates are greatly reduced, vestigial structures. In all Chironominae and Orthocladiinae, the shape, number, and arrangement of the mental teeth and characteristics of the ventromental plates are critical diagnostic features. Most other structures are normally lost from the head capsule.

Although taxonomic keys for living larvae rarely permit identification of fossil larval head capsules, reference to associated illustrations and diagnoses for each genus usually permit classification. Fortunately, most of the characters employed by systematists are borne on the head capsule and consequently may be available for fossil work. Few keys have been designed specifically for fossil identifications.

Sample Preparation

Meticulous preparation, sorting, and identification of midge remains are requisites for analysis of past communities. Small lakes in forested watersheds usually contain sediments with a high organic content. In such lakes, 1 or 2 ml of wet sediment typically contains 50 to 200 head capsules, sufficient for analysis. Inorganic sediments, such as those frequently encountered in Late Glacial deposits, may yield much lower concentrations. However, head capsule concentrations vary greatly among sites.

Samples are usually deflocculated in 5 to 10% KOH prior to analysis. In calcareous sediments, an acid wash (10% HCl) may also be necessary. If excessive siliceous mineral matter is present, HF may be used to further concentrate and clean the head capsules. Following these chemical treatments, head capsules are generally separated from finer debris by means of sieving. Methodology varies, but a 100 µm or finer sieve will retain most head capsules and is recommended. Subsequently, the residue is back-washed from the sieve into a container and stored wet until the material is sorted. Unless sorted shortly after sieving, preservation in 99% ethanol is advisable. Sorting of wet residues is best accomplished in a Bogorov counting tray at 40 to 50X. As remains are located, each is transferred with forceps to coverslips, for mounting in an appropriate medium (e.g., Canada balsam, Euparal®, or Entellan®) and subsequent identification.

Ecology and Distribution

Early work on the ecology and distribution of Chironomidae was conducted by biological limnologists (e.g., August Thienemann, Lars Brundin, Ole Sæther) who sought to classify lakes principally with regard to biological productivity. They also emphasized the importance of Chironomidae as bio-indicators of water quality. The importance of water quality parameters, especially food quantity, food quality, and the oxygen concentration at the mud–water interface were considered of paramount importance in regulating the distributions of larval Chironomidae.

Large-bodied larvae, which contained hemoglobin (especially *Chironomus*), were noted as best adapted to the oxygen-deficient bottom waters typical of eutrophic lakes (nutrient-enriched lakes with high algal production). Hemoglobin facilitates the capture and storage of oxygen, whereas large body size allows the larvae to disrupt, and extend their bodies above, the most oxygen-deficient micro-layer at the mud–water interface. In contrast, small-bodied larvae, which lacked hemoglobin (several Tanytarsina and Orthocladiinae), were better adapted to low food availability in the well-oxygenated bottom waters of oligotrophic lakes (nutrient-poor lakes with low algal productivity). Because of these relationships among chironomids, water quality, and oxygen concentrations, extant chironomid communities are sampled extensively for environmental biomonitoring purposes. The Benthic Quality Index, for example, was developed as a means for environmental assessment in Nordic freshwater lakes (Wiederholm, 1980).

Detailed surveys of chironomid distributions in North America and Europe have revealed the importance of other environmental variables. Recent studies continue to recognize the importance of oxygen and biological productivity as controls on species distributions, but greater importance is commonly extended to climatic variables, especially summer air and water temperatures. The Chironomini and Tanypodinae are abundant and widely-distributed in temperate to tropical climates, but the subfamilies Orthocladiinae and Diamesinae, and some members of the subtribe Tanytarsina, are more abundant where colder, alpine and polar climates prevail. In addition to temperature and oxygen, salinity, pH, dissolved organic carbon, nutrient concentrations, and lake depth are

considered important variables defining the distributions of individual chironomid species.

Quaternary Paleoecology

Lake Ontogeny

Early chironomid paleoecological research was primarily directed to answering questions pertaining to lake ontogeny (i.e., those physical, chemical, and biological processes that influence how lakes intrinsically tend to develop through time). Einar Naumann (1919) speculated that lakes should gradually become less productive as a consequence of constant leaching of catchment soils. Many other limnologists, including Naumann's contemporary August Thienemann (1926), perceived eutrophication (a gradual increase in lake productivity) as the dominant, if not universal process. Whiteside (1983) traces this perception to early works published by Lindeman (1942) and Deevey (1955). Deevey emphasized that the gradual infilling of lakes, by reducing the volume of the hypolimnion (the cold-water region at the bottom of the lake) could generate an O₂-poor hypolimnion. Release of phosphorus, an important nutrient, from the sediments beneath the hypolimnion could then produce an increase in lake productivity. This process was dubbed morphometric eutrophication.

This appealing theory was broadly supported by early chironomid paleoecological studies, which suggested a shift from midge communities typical of well-oxygenated, nutrient-poor (oligotrophic) lakes, to those prevalent in the deep water of oxygen-deficient, nutrient-enriched (eutrophic) lakes. However, the theory was developed from a relatively small subset of the world's lakes (i.e., temperate lakes, formed during the relatively recent deglaciation of eastern North America and northwestern Europe), and was never tested more generally on lakes with different origins, or in other climatic regions.

These early investigators also failed to recognize or acknowledge the potential importance of climate in driving the observed changes in both midge fauna and lake productivity. Major changes in the midge faunas typically occurred at each site with the development of forests and rapid deglacial warming approximately 10,000 ¹⁴C yr BP.

Today, these changes in productivity and midge communities are more likely to be interpreted as a consequence of climatic change, not as an intrinsic tendency of lakes to become more productive over time. The importance of climate as a driving force in lake productivity and midge community change is especially evident in late-glacial records, where increases and decreases in the apparent productivity of lakes and changes in midge faunas, coincide with independent evidence of climatic changes. The apparent productivity of lakes abruptly decreases and oligotrophic midge communities return, with the onset of colder conditions.

Human Impact Studies

Although the natural productivity of lakes and their midge faunas may be regulated more by climate than initially recognized, human activities have also had major impacts. Given the long history of environmental biomonitoring using living Chironomidae, there has also been long interest in using the fossils in paleoenvironmental assessments of human impacts.

Some researchers have used mouthpart deformities as indicators of environmental stress (e.g., in the Rhine), but most results are derived from stratigraphic studies of midge community composition, as preserved in lake sediments.

Eutrophication

Increased nutrient loading, for example, from urban and agricultural land use may dramatically increase the productivity of some lakes, to the point where algal blooms foul shorelines and their decomposition produces anaerobic conditions within the hypolimnion. In addition to this increased organic production from nutrient enrichment, external inputs of sewage and other organic wastes compound the de-oxygenation problem.

Since chironomid communities respond dramatically to such trophic alterations, their remains provide a detailed record of eutrophication events. Most studies (e.g., at Esthwaite Water and Lough Neagh (UK), Lake Washington (US), Wood Lake, and Bay of Quinte, Lake Ontario (Canada), Lake Mälaren and Lake Vättern (Sweden), Lake Vanajavesi (Finland)) have relied on qualitative interpretation of midge stratigraphies for impact assessment. In recent years, researchers have instead attempted to develop quantitative reconstructions of productivity related variables.

Most researchers would argue that hypolimnetic oxygen concentrations have the most important and most direct impact on living larvae; thus a quantitative model that infers paleo-oxygen conditions, should ideally be developed. Oxygen conditions vary continuously throughout the year, and also at different depths in the water column, so it is difficult to characterize a lake's oxygen level with a single oxygen measurement. Furthermore, the duration of anoxic conditions and the build-up of toxic H_2S in anoxic hypolimnia, are probably equally important variables. The relationship between midge communities and oxygen concentrations is even more tenuous in shallow lakes (lacking a hypolimnion) and in near-shore (littoral) habitats, than in deep, thermally stratified hypolimnia.

As an alternative to reconstructing oxygen levels, researchers have developed models to infer phosphorus or chlorophyll concentrations. The connection between inorganic nutrients, such as phosphorus, and their impacts on chironomid larvae is indirect, but nutrient concentrations are more easily measured and more widely used to characterize lake productivity. These midge-productivity models are now being used in North America and Europe to better quantify human impacts on the environment. Brodersen and Quinlan (2006) review recent pertinent literature.

Acidification

During the 1980s, lake acidification and the long-range transport of air pollutants were amongst the most controversial and internationally important environmental issues. Much environmental research was devoted to tracing the paths of pollutants, and assessing their damage to ecosystems in both eastern North America and northern Europe. Little was known about the distributions of chironomids with respect to pH, alkalinity, or related environmental variables, but they were used as biomonitoring tools in several studies of lake acidification.

pH is a master variable in aquatic chemistry. Many other variables, for example, lake alkalinity, dissolved organic carbon, and metal concentrations, tend to co-vary with pH, thus it

is difficult to separate the direct effects of pH from the effects of these covariables. At low pH, metals (e.g., Fe, Al) are more soluble, and some (e.g., Al) are known to be toxic to aquatic organisms. However, many acidic lakes are also humic lakes, containing high concentrations of dissolved organic acids.

The organic acids are derived from the incomplete decomposition of terrestrially-derived organic matter, especially from adjacent bogs. These organic acids are natural products, which may contribute to the acidity of lakes, but may also slow the rate of decomposition and bind to nutrients (e.g., P) and metals (e.g., Al), reducing the impact of these pollutants on aquatic life.

Chironomidae are well represented even in strongly-acidic waters (pH 3.5 to 5.5). *Chironomus*, *Phaenopsectra*, *Psectrocladius*, and *Zalutschia* are reported at greater relative abundance at low pH. Favored taxa tend to be large-bodied, having smaller surface areas relative to volume. It may be easier for large larvae to maintain their internal pH balance. Those midges abundant in acidified lakes include taxa reported to be more common in fish diets. Thus acidification, by eliminating fish, could also favor these taxa. Furthermore, the increased buffering capacity afforded by hemoglobin may also favor hemoglobin-bearing larvae.

Although several paleolimnological studies have documented changes in midge populations that accompanied lake acidification, these insects have never been used to quantitatively reconstruct pH. Although technically feasible, the chironomid-pH relationships have never been properly quantified. Once the pH optima and tolerances of chironomid taxa are quantified, they should be useful bio-indicators of lake acidification processes. This is an obvious area to target in future research.

Climate Change

Paleotemperature

Although researchers had previously pondered the possible climatic significance of midge stratigraphies, it was not until the late 1980s and early 1990s that midge-climate relationships were explored in earnest. At first, the possibility of using chironomids for paleoclimatic inference was highly controversial, stimulating a lively debate in Quaternary Research, the Journal of Paleolimnology, and the Canadian Journal of Fisheries and Aquatic Sciences.

Key to resolving the issue was an international effort to map the distribution and abundance of chironomid taxa across long climatic gradients in both North America and Europe. Statistical analyses of these data repeatedly demonstrated that summer air and water temperatures were highly correlated with midge distributions.

Summer climate is now recognized as potentially the single most important variable regulating midge distributions. Fossil midges are also recognized as one of the most reliable indicators of past climatic change. Through the use of statistical inference procedures (e.g., weighted averaging, partial least squares, analog matching, and Bayesian methods) midge-climate models are routinely used in the quantitative reconstruction of late-glacial and Holocene paleotemperatures. Details of the midge-climate relationship, however, are still widely debated. At least in part, midges may be responding to climate indirectly via the influence of climate on lake productivity.

Surveys of midge distributions spanning the treeline commonly reveal distinct forest and tundra communities, with species turnover being most rapid in the vicinity of the treeline.

This observation suggests that treeline communities may be highly sensitive to climate change, and these sensitive sites are targeted by paleoecologists for paleotemperature reconstructions. Conversely, these transects suggest that the fauna within the forest region is more uniform, and insensitive, so such insensitive sites tend to be ignored by paleoecologists. This bias is also likely to impact on midge-derived paleotemperature reconstructions. From the perspective of a researcher wishing to accurately reconstruct paleotemperatures, treeline sites may be overly sensitive, revealing paleotemperature signals larger than the changes that actually occurred, whereas forested sites may pose the opposite dilemma. This issue has implications for accurately reconstructing: 1) the temporal magnitude of paleoclimatic changes, and 2) spatial temperature gradients, for example, as they existed at the margin of ice sheets (Levesque et al., 1997).

Brooks (2006), and Walker and Cwynar (2006) review recent progress in paleotemperature reconstruction using fossil midges.

Lake levels

Statistical analyses have revealed a number of other environmental variables potentially critical for understanding midge distributions. Lake depth, for example, is often just as important as temperature. Since lake levels have often been used as indicators of hydrological balance, lake depth inferences might provide another useful tool for paleoclimatologists. During warm, dry intervals lake levels may drop. During cool, wet intervals lake levels will rise.

Midge–depth relationships are easily modeled for use in lake level reconstructions. Despite this opportunity, midges are seldom used for lake level reconstructions. Model errors indicate that satisfactory reconstructions will only be possible where the amplitude of lake level changes has been very large. Since the surface discharge from lakes can change enormously with small changes in lake level, inferences obtained from open basin lakes are unlikely to provide useful information. The magnitude of lake level changes in closed basin lakes is potentially much larger. These lakes are common in semi-arid regions and potentially contain more useful paleohydrological records.

Paleosalinity

Although not extensively studied, chironomid communities are known to be salinity sensitive. This is never evident in freshwater regions, but saline lakes are common landscape features, for example, in Australia, east Africa, central Asia, and western North America. As in other aquatic organisms, the osmotic stress imposed by high salinities tends to dehydrate larval midges. Many species are intolerant, restricted entirely to fresh water, but a few (e.g., *Cricotopus ornatus* and *Tanytus nubifer*) thrive in highly saline lakes or in marine intertidal habitats.

Since hydrological balance (evaporation/precipitation) potentially regulates the salinity of these lakes, midges and other paleosalinity indicators (e.g., diatoms) may provide important paleoclimate proxy data from these sites. During warm, dry intervals salinity is expected to be high, and during cool, wet intervals salinity should be low. Midge paleosalinity reconstructions are now available for several sites in Africa and western North America. Recent evidence suggests that the paleohydrology of these lakes may be very complex. Paleosalinity reconstructions may be more difficult to interpret climatically than previously assumed.

In addition to their potential climatic significance, midge paleosalinity records may be useful for sea-level studies. Heinrichs and Walker (2006), and Verschuren and Eggemont (2006) have reviewed the potential of midge paleosalinity records for both paleohydrological and sea-level studies.

See also: **Chironomid Records: Africa; Late Pleistocene of Europe; Postglacial Europe; Postglacial Southern Hemisphere. Diatom Methods: Salinity and Climate Reconstructions from Continental Lakes.**

References

- Andersen T and Sæther OA (1994) *Usambaromyia nigrata* gen. n., sp. n., and *Usambaromyiinae*, a new subfamily among the Chironomidae (Diptera). *Aquatic Insects* 16: 21–29.
- Armitage PD, Cranston PS, and Pinder LCV (eds.) (1995) *The Chironomidae: Biology and ecology of non-biting midges*. London: Chapman & Hall.
- Brodersen KP and Quinlan R (2006) Midges as paleoindicators of lake productivity, eutrophication and hypolimnetic oxygen. *Quaternary Science Reviews* 25: 1995–2012.
- Brooks SJ (2006) Fossil midges (Diptera: Chironomidae) as palaeoclimatic indicators for the Eurasian region. *Quaternary Science Reviews* 25: 1894–1910.
- Deevey ES Jr. (1955) The obliteration of the hypolimnion. *Mem. Ist. Ital. Idrobiol.* Suppl 8, 9–38.
- Frey DG (1976) Interpretation of Quaternary paleoecology from Cladocera and midges, and prognosis regarding usability of other organisms. *Canadian Journal of Zoology* 54: 2208–2226.
- Frey DG (1988) Littoral and offshore communities of diatoms, cladocerans and dipterous larvae, and their interpretation in paleolimnology. *Journal of Paleolimnology* 1: 179–191.
- Heinrichs ML and Walker IR (2006) Fossil midges and palaeosalinity: Potential as indicators of hydrological balance and sea level change. *Quaternary Science Reviews* 25: 1948–1965.
- Hofmann W (1986) Chironomid analysis. In: Berglund BE (ed.) *Handbook of Holocene Palaeoecology and Palaeohydrology*, pp. 715–727. Chichester: John Wiley & Sons.
- Hofmann W (1988) The significance of chironomid analysis (Insecta: Diptera) for paleolimnological research. *Palaeogeography, Palaeoclimatology, Palaeoecology* 62: 501–509.
- Iovino AJ (1975) *Extant Chironomid Larval Populations and the Representativeness and Nature of their Remains in Lake Sediments*. Unpublished PhD. thesis, Indiana University, Bloomington.
- Levesque AJ, Cwynar LC, and Walker IR (1997) Exceptionally steep north-south gradients in lake temperatures during the last deglaciation. *Nature* 385: 423–426.
- Lindeman RL (1942) The trophic-dynamic aspect of ecology. *Ecology* 23: 399–418.
- Porinchu DF and MacDonald GM (2003) The use and application of freshwater midges (Chironomidae: Insecta: Diptera) in geographical research. *Progress in Physical Geography* 27: 378–422.
- Rieradevall M and Brooks SJ (2001) An identification guide to subfossil Tanypodinae larvae (Insecta: Diptera: Chironomidae) based on cephalic setation. *Journal of Paleolimnology* 25: 81–99.
- Thienemann A (1926) Der Nahrungskreislauf im Wasser. *Verhandlungen der Deutschen zoologischen Gesellschaft* 31: 29–79.
- Verschuren D and Eggemont H (2006) Quaternary paleoecology of aquatic Diptera in tropical and southern hemisphere regions, with special reference to the Chironomidae. *Quaternary Science Reviews* 25: 1926–1947.
- Walker I (1995) Chironomids as indicators of past environmental change. In: Armitage PD, Cranston PS, and Pinder LCV (eds.) *The Chironomidae: Biology and ecology of non-biting midges*, pp. 405–422. London: Chapman & Hall.
- Walker IR (1987) Chironomidae (Diptera) in paleoecology. *Quaternary Science Reviews* 6: 29–40.
- Walker IR (1993) Paleolimnological biomonitoring using freshwater benthic macroinvertebrates. In: Rosenberg DM and Resh VH (eds.) *Freshwater Biomonitoring and Benthic Macroinvertebrates*, pp. 306–343. New York: Chapman & Hall.
- Walker IR (2001) Midges: Chironomidae and related Diptera. In: Smol JP, Birks HJB, and Last WM (eds.) *Tracking Environmental Change Using Lake Sediments. Zoological Indicators*, Developments in Paleoenvironmental Research, Dordrecht: Kluwer, vol. 4, pp. 43–66.
- Walker IR and Cwynar LC (2006) Midges and palaeotemperature reconstruction – the North American experience. *Quaternary Science Reviews* 25: 1911–1925.
- Whiteside MC (1983) The mythical concept of eutrophication. *Hydrobiologia* 103: 107–111.
- Wiederholm T (1980) Use of benthos in lake monitoring. *Journal of the Water Pollution Control Federation* 52: 537–547.

Africa

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Glossary

Alpha-diversity: Refers to the diversity within a particular area or ecosystem, and is usually expressed by the number of species (i.e., species richness) in that ecosystem.

Anoxia: The absence of oxygen necessary for sustaining most life.

Autoecology: The ecology of an individual species in its natural environment, without consideration of its interactions with other species.

Benthos: Collectively, all organisms living on or in the bottom of a lake, stream or other water body.

Beta-diversity: A comparison of diversity between ecosystems or biomes, usually measured as the amount of species change between the ecosystems.

Calibration data set: A set of species collected from modern sites along with relevant environmental data, often related to one another using a series of quantitative inference models. Calibration data sets form the basis for quantitative paleoecological reconstruction when used in conjunction with fossil assemblages. Used to generate inference models. See inference model.

Canonical correspondence analysis (CCA): A constrained ordination technique based on correspondence analysis, but where the ordination axes are constrained to be linear combinations of the supplied environmental (predictor) variables.

Conductivity: Here a measure of the ability of water to conduct an electric current due to the presence of dissolved solids, normally expressed as microsiemens/cm ($\mu\text{S cm}^{-1}$) or micromhos/cm ($\mu\text{mhos cm}^{-1}$).

Correspondence analysis (CA): An ordination method that simultaneously ordines samples and variables (species), and maximizes the correlation between sample and variable scores. It is widely used in ecology because it assumes a unimodal response of species variables to underlying gradients.

Direct ordination: An ordination in which the important gradients are known and measured. See ordination.

Ecotone: The boundary or transition area between two or more habitats or communities.

Halobiont: Living or growing in a salty environment.

Holomictic: Used for lakes which have complete circulation of the water column.

Hypersaline: Having a total dissolved solid content (salinity) greater than about 40 ppt (parts per thousand) or $60\,000\ \mu\text{S cm}^{-1}$.

Hypolimnion: Cold-water region at the bottom of a thermally stratified lake.

Indicator species: A species whose presence is a marker for a specific habitat condition or environmental value.

Indirect ordination: An ordination in which gradients are unknown *a priori*, and are inferred from species composition data.

Inference model: A mathematical function that describes the relationship between biological species and environmental variables that allow the past values of an environmental variable to be inferred from the composition of a fossil assemblage.

Littoral zone: The zone in a lake or pond extending from the shore to the greatest depth at which rooted plants occur.

Maximum likelihood: A method of environmental reconstruction based on the idea that the relation between the abundance of a taxon and the environmental variable of interest can be modeled by an ecological response curve consisting of a systematic and a random (error) component. Together, the curves of all taxa determine what taxa composition is expected at a given conductivity value. This model of responses and their error structure is used to calculate the probability that a particular value would occur with a given assemblage. The value that gives the highest probability is the maximum likelihood estimate.

Meromictic: Used for lakes which do not undergo complete mixing of the water column at least once each year.

Mesosaline: Degree of salinity in brackish to marine waters; water having a total dissolved solid content (salinity) of 5.0–18.0 ppt or $8000\text{--}30\,000\ \mu\text{S cm}^{-1}$.

Modern analog technique: A method of environmental reconstruction based on matching fossil samples to modern surface-sediment assemblages of similar taxonomic composition.

Morphotype: In this context, group of fossil specimens which are considered to belong to the same species, group of species or a higher unit based on shared morphological characters alone.

Oligosaline: Having a total dissolved solid content salinity between 0.5–5 ppt.

Ordination: The collective term for multivariate techniques that arrange sites along axes on the basis of data on species composition; the usual objective of ordination is to help generate hypotheses about the relationship between the species composition at a site and the underlying environmental gradients.

Osmotic stress: Physiological stress caused when the salinity of an animal's environment changes drastically.

Oxycline: A horizontal boundary layer in the water column, at which dissolved oxygen content changes sharply with depth.

Polysaline: Water having a salinity of 18.0–30.0 ppt or $30\,000\text{--}45\,000\ \mu\text{S cm}^{-1}$.

Profundal: Pertaining to deep-water bottom habitats in lakes.
Proxy data: In this context, data obtained from natural environmental archives and used as surrogates (proxies) for the past condition of another variable, such as past temperature or precipitation, that cannot be measured directly.

Weighted-averaging: In environmental reconstruction, a numerical technique to derive inference models by estimating species optima from the calibration data set (i.e., regression), and then using these optima to estimate the past value of an environmental variable from a fossil assemblage (i.e., calibration). The fundamental assumption of the

technique is that the weighted average of a taxon represents the conditions for which this taxon is most abundant. The species optima and tolerances are estimated from the weighted averages and weighted standard deviations, respectively.

Weighted-averaging partial least squares: An extension of weighted-averaging and calibration where two or more components, that utilize the residual structure in the modern biological data, are used to improve the predictive power of the inference model.

Zoobenthos: Invertebrate animals that live in or on the bottom of a lake, stream or other water body.

Introduction

Building on the tradition to exploit chironomid larvae (Insecta: Diptera: Chironomidae) as biological indicators in contemporary water-quality monitoring of lakes and rivers (Rosenberg, 1993), paleolimnologists are making increasingly profitable use of their fossil remains preserved in lake sediments to reconstruct long-term environmental change affecting lakes and adjacent landscapes (Walker, 1995). In north-temperate Europe and North America, chironomid paleoecology (and particularly chironomid paleoclimatology) is now a mature research discipline (Walker, 2001), involving statistically robust inference models for quantitative reconstruction of past changes in temperature, salinity, lake trophic status, and oxygen regime. In tropical regions including Africa, the growth of chironomid paleoecology as a rigorous, quantitative research discipline has been decidedly slower. The principal reasons for this geographical imbalance are incomplete taxonomic knowledge of tropical Chironomidae (and in particular their larval stage), the scarcity of autecological information that would confer bio-indicator value to local taxa, and the logistic (sometimes also political) difficulties hampering the comprehensive ecological lake surveying needed to obtain this information. Although continuous high-resolution chironomid records spanning the entire postglacial period in Africa are thus far lacking, African chironomid paleoecology has made substantial progress over the past decade (Verschuren and Eggermont, 2006). In this article, we provide an overview of its present status, dealing successively with chironomid records of past environmental change, advances in methodology, and taxonomy.

Chironomid Records of Past Environmental Change in Africa

Fossil Chironomidae have so far contributed to eight paleoenvironmental reconstructions from tropical Africa (Eggermont et al., 2008b; Mees et al., 1991; Ryner et al., 2007; Ryves et al., 2011; Verschuren et al., 1996, 1999b, 2000a, 2002). Mees et al. (1991) reconstructed Holocene lake-level and salinity changes at Malha crater lake in Sudan (15° N 26° E), one of very few permanent waters in the eastern Sahara desert. The lithology and mineralogy of its sediment sequence indicated that six successive lake phases had occurred over the past 8300 ¹⁴C

years, separated by severe lowstands or complete desiccation. The timing of these hydrological changes is poorly constrained, due to lack of high-quality radiocarbon dating targets, but their pattern is consistent with the regional history of millennial-scale fluctuations in rainfall and drought (Hoelzmann et al., 2004). In the early- to mid-Holocene, the lake was deep and meromictic, with surface waters evolving from fresh (as indicated by the presence of *Dicoretendipes fusconotatus*, *Dicoretendipes sudanicus* and *Nilodorum brevibucca*) to mesosaline (indicated by presence of the salt-loving *Kiefferulus disparilis*). The authors interpreted this evolution as a reflection of decreasing rainfall and shrinking groundwater input, which both reduced new freshwater inflows and caused wind-driven mixing to bring increasing amounts of deep salty water to the surface. During the last ~4000 years, Malha Crater Lake was most often shallow and holomictic, and poly- to hypersaline. Chironomidae became locally extinct, and were replaced only by scattered occurrences of other salt-loving midges, such as the brine fly *Ephydra* (Ephydriidae), the biting midge cf. *Culicoides* (Ceratopogonidae), and an unidentified soldier fly larva (Stratiomyidae). Clearly, the late Holocene aquatic environment in Malha Crater approached the limit of what African aquatic invertebrate communities can tolerate.

Verschuren et al. (1996) used the relative abundance of (semi)terrestrial, lake, and stream chironomids extracted from a peat deposit at Kashiru, Burundi (2100 m a.s.l., 3° S 30° E) to describe the general late-Quaternary evolution (last ~40 000 years) of this cold aquatic habitat in the mountains northeast of Lake Tanganyika. Peat sequences are the traditional source of African pollen records, but this is the only study so far which exploited them for chironomid paleoecology. Chironomid genera that are widely distributed in lowland African lakes (*Chironomus*, *Cladotanytarsus*, *Dicoretendipes*, *Kiefferulus*, *Nilodorum*, *Polypedilum*, and *Procladius*) were absent or scarce in the fossil assemblages found at this site; consequently, the ecological literature on African lakes was of little use to their interpretation. Instead, the authors made inferences on past habitat conditions mostly by transferring Holarctic ecological typology (Merritt and Cummins, 1984), assuming that cold-adapted African representatives of Holarctic genera would have similar habitat preferences. Chironomid fossils were markedly concentrated in core sections consisting of decomposed or only slightly fibrous peat, which presumably had been deposited in peat hollows. Strongly fibrous peat (from hummocks)

and sandy horizons (indicating an episode when the local environment was too dynamic to allow peat accumulation) were typically devoid of fossils. Overall, the fossil chironomid fauna at Kashiru indicated long-term persistence of a cold montane peat bog with a fragmented water surface through which small streams occasionally flowed. Stratigraphic trends in the species assemblages support pollen evidence for drier conditions in equatorial East Africa during the Last Glacial Maximum (~25 000–15 000 years ago) and a fairly humid climate before that time (Bonnefille et al., 1990). *Chironomus formosipennis*, another *Chironomus* sp. and *Harnischia*, which all require mud bottoms and open standing water, are limited to the period before 25 000 years ago. Also concentrated in this section are the remains of *Chaetocladius*, which typically occur in swampy lakes, and *Rheocricotopus*, indicating some stream flow through the fen. Lowering of the water table during the Last Glacial Maximum is indicated by the prevalence of taxa resembling Holarctic genera with semiterrestrial larvae, such as cf. *Pseudosmittia* and *Bryophaenocladius*. The bog surface did not desiccate completely, however, as attested by the persistence of benthic water fleas (chydorid Cladocera) and a minority component of aquatic Chironomidae (*Polypedilum*, *Tanytarsus*, Tanypodinae indet). The onset of the postglacial Pluvial period around 15 000 years ago resulted in the disappearance of *Bryophaenocladius* and cf. *Pseudosmittia*, and a return to higher chironomid faunal diversity.

One of the best-resolved African climate-proxy records based on fossil chironomid data is that of Crescent Island Crater (1°S 36°E), a satellite basin of Lake Naivasha in the central Kenya Rift Valley (Kenya). This fossil record documents water-balance variations over the past 1100 years with decade-scale time resolution (Verschuren et al., 2000a). For this study an exploratory salinity-inference model was developed (Verschuren et al., 2004, see also below), calibrated using a compilation of published distribution data on 20 common chironomid species in 32 African surface waters (Verschuren, 1997). The resulting salinity reconstruction (Figure 1) is fully consistent with a parallel reconstruction based on fossil diatoms, and with a lake-level reconstruction based on changes in the lithological characteristics of the sediment sequence. The combined data show that equatorial East Africa was significantly drier than today during the Medieval Warm Period (~AD 1000–1270) and that fairly wet conditions during the Little Ice Age (~AD 1270–1850) were interrupted around AD 1380–1420, 1560–1620, and 1760–1840 by three decade-scale episodes of severe aridity. From ~AD 1670 to 1770, that is, the period which encompasses the Maunder Minimum in solar radiation, Crescent Island Crater stood higher than at any other time in the past millennium. The local chironomid fauna of that time was dominated by *Dicrotendipes septemmaculatus*, in line with the positive relationship between this species and lake depth in the historical (post-1880) portion of the sediment record. Presumably, this reflects expansion of its preferred habitat of submerged macrophyte beds during lake high stands. Superimposed on pronounced lake-level fluctuation during the past 1100 years is a long-term freshening trend, beginning with the transition to a more frequently positive water balance (rainfall and river inflow exceeding evaporation and loss to groundwater) after AD 1270. In the chironomid data this is reflected in the punctuated disappearance of salt-loving species

(*K. disparilis*, *Microchironomus deribae*) after each new high stand episode, to the benefit of a diverse freshwater assemblage including some salt-intolerant taxa (e.g., *Psectrocladius viridescens*).

The observation that recurrent drought is a natural characteristic of Kenya's semiarid Rift Valley climate also followed from a 200-year paleoenvironmental reconstruction at Lake Oloidien, another satellite basin of Lake Naivasha (Verschuren et al., 1999b). This study was intended to provide a long-term historical perspective on the problem of rising salinity during the late 1980s and early 1990s, and the possible role of rapidly increasing water extractions by the sprawling horticulture industry around Lake Naivasha. Fossil evidence for the historical alternation of freshwater and salt-tolerant aquatic biota (including Chironomidae) throughout the nineteenth and twentieth centuries revealed that soda-lake conditions are a natural recurrent phenomenon in Lake Oloidien, due to major natural rainfall variability combined with the dependence of Lake Oloidien on episodes of confluence with Lake Naivasha to remain fresh. The recent salinity rise was temporarily halted in early 1998, thanks to a 2.6-m lake-level rise following high rainfall driven by El Niño. However, the authors noted that diversion of river inflows for cropland irrigation and direct water extractions for the horticulture industry will increasingly limit the ability of such periodic high-rainfall events to restore confluence with Lake Naivasha, and eventually result in Lake Oloidien becoming permanently saline.

Ryner et al. (2007) reconstructed the aquatic ecosystem response of Lake Emakat (Empakaai Crater) in northern Tanzania to hydrological change associated with postglacial warming using microfossil assemblages (chironomids, diatoms, and ostracods) alongside previously published fossil pollen and carbon and nitrogen isotopic data for the same sediment sequence. Conductivity estimates for selected intervals were derived from subfossil chironomid faunas in the surface sediments of 67 lakes in tropical Africa (Eggermont et al., 2006; see also below), and suggest a slight trend of increasing salinity from ~350 $\mu\text{S cm}^{-1}$ at the start of the sequence (ca. 14 800 cal years BP) to ~530–550 $\mu\text{S cm}^{-1}$ from 13 200 to 12 600 cal years BP, to ~610 $\mu\text{S cm}^{-1}$ around 11 000 cal years BP. All but one of the freshwater Chironomidae inhabiting the lake during this period are common eurytopic taxa. Between ca. 14 800 and 14 600 cal years BP, mud-dwelling Chironomidae, mainly *Chironomus* near *alluaudi* indicate poorly developed aquatic vegetation, consistent with high lake level in a steep-sided crater basin. Reduced lake levels and an expansion of macrophytes between ca. 14 400 and 13 200 cal years BP is indicated by chironomid remains such as *Dicrotendipes fusconotatus* and *Tanytarsus formosanus*, which are both inhabitants of littoral vegetation and shredders of coarse plant detritus. Lake transgression between ca. 13 200 and 11 800 cal years BP, then again, is reflected by peak abundances of the chironomid *Cricotopus* type East Africa suggesting well-oxygenated conditions of relatively low algal production and a dynamic unvegetated shoreline. The only midge remains recovered from sediments younger than ca. 10 000 cal years BP are the halobiont species *K. disparilis* and *M. deribae*, indicating a relatively sudden transition to a strongly alkaline soda-lake environment with conductivity exceeding 6000 $\mu\text{S cm}^{-1}$. Some paleohydrological changes reconstructed for Lake Emakat are

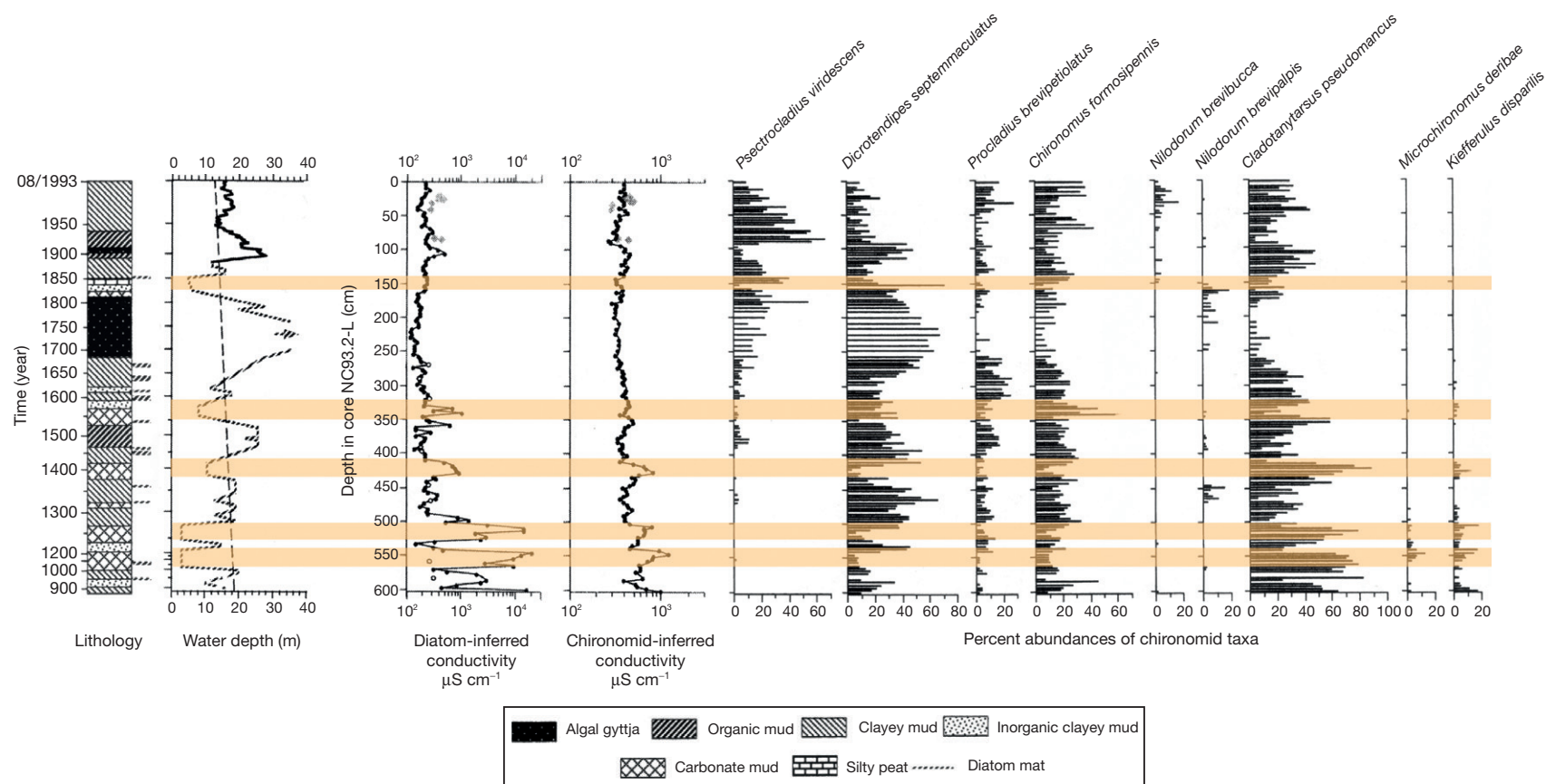


Figure 1 Sedimentological and biological evidence for past lake-depth and salinity fluctuations in the Crescent Island Crater basin of Lake Naivasha, with episodes of pronounced drought highlighted in yellow. From left to right: lithology-inferred lake-level change over the past ~1100 years (solid and stippled line) relative to the elevation of the sill separating Crescent Island Crater from Naivasha (broken line); diatom-inferred conductivity reconstruction using the inference model of Gasse et al. (1995); chironomid-inferred conductivity reconstruction using the hybrid model of Verschuren et al. (2004); and stratigraphic distribution of selected fossil midge taxa with species sorted according to increasing conductivity optima. Orange zones indicate periods of severe aridity. Modified from Verschuren D, Cumming BF, and Laird KR (2004) Quantitative reconstruction of past salinity variations in African lakes using fossil midges (Diptera: Chironomidae): Assessment of inference models in space and time. *Canadian Journal of Fisheries and Aquatic Sciences* 61: 986–998.

in phase with lake evolution elsewhere in the region and thus apparently track broad-scale climate changes, but others reflect a more complex adjustment of the local aquatic ecosystem to temporal variations both in total annual effective precipitation and its seasonal distribution.

Eggermont et al. (2008b) used both chironomid and non-chironomid invertebrate remains to reconstruct the aquatic community response in a groundwater-fed desert lake (Lake Yoa, northeastern Chad) to Holocene desiccation of the Sahara (Kroepelin et al., 2008; Figure 2). Until at least 4800 cal years BP, Lake Yoa was a hydrologically open, deep, and dilute ($<500 \mu\text{S cm}^{-1}$) freshwater environment with diversified near-shore habitat, including submerged macrophyte beds as indicated by a species-rich chironomid fauna mainly consisting of steno- and eurytopic freshwater taxa. A distinct increase of a presumed salt-tolerant *Dicoretendipes* species around 4800 cal years BP suggests that an increasingly more negative water balance resulted in measurable evaporative concentration. By 3900 cal years BP, virtually all steno- and eurytopic freshwater chironomid taxa (*Cryptochironomus* type Tanganyika, *Procladius brevipetiolatus*, *Microchironomus* type East Africa, *Polypedilum* near *deletum*, *Cricotopus* type East Africa) disappeared. Only known salt-tolerant freshwater taxa (*Cladotanytarsus pseudomancus*) persisted, and known salt-loving taxa (*K. disparilis*) appeared for the first time, indicating that the important physiological threshold of $3000 \mu\text{S cm}^{-1}$ was crossed. The transition to a true salt-lake ecosystem ($>25000 \mu\text{S cm}^{-1}$) seems to have occurred rather abruptly around 3400 cal years BP, when the remaining salt-tolerant freshwater taxa were exterminated and new salt-loving taxa appeared (*M. deribae*, *Cricotopus* type Ounianga Kebir) to complement a strongly expanding *K. disparilis*. A fourth ecological transition seems to have occurred around 2700 cal years BP with the disappearance of the halophilic *M. deribae*, suggesting that the limit of $\sim 45000 \mu\text{S cm}^{-1}$ was crossed. Finally, the last ecological transition seems to have occurred around 1500 cal years BP (AD 600), when *Chironomus* cf. *salinarius* disappeared together with a strong expansion of *Cricotopus* type Ounianga Kebir, testifying to a renewed trend of rising salinity. Comparison with proxies tracing the evolution of the terrestrial ecosystem led the authors to conclude that although the fresh-to-saline transition of Lake Yoa occurred fairly abruptly between ~ 4100 and 3400 cal years BP, continuous inflow of fossil groundwater buffered the ecosystem against shorter-term climate fluctuations. The stratigraphic distribution of other aquatic invertebrate remains seems to support this interpretation (Figure 2).

Ryves et al. (2011) analyzed sedimentary parameters (including C/N ratios, $\delta^{13}\text{C}$ of bulk organic matter and $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ of authigenic carbonates) and biotic remains (diatoms, aquatic macrofossils, chironomids) in the sediment records from two small, nearby closed crater lakes in western Uganda (Lake Kasenda and Lake Wandakara) spanning the last 700 (Wandakara) and 1200 years (Kasenda), respectively. Subfossil head capsule concentrations in samples from Lake Kasenda were too low to permit reliable stratigraphic interpretation of the chironomid community. However, major differences in the chironomid profile of Wandakara were controlled by the dominance of *N. fractilobus* and *Dicoretendipes fusconotatus*, which have a distribution restricted to the less saline ($<1200 \mu\text{S cm}^{-1}$) African lakes (Eggermont et al., 2006).

Downcore salinity inferences were made by applying weighted-average optima from Eggermont et al. (2006; see below) to the fossil data from Wandakara, weighted by their relative abundance. The combined data suggest that Kasenda has been sensitive to climate over much of this period, and has shown substantial fluctuations in conductivity, while Wandakara has had a more muted response, likely due to the increasing dominance of human activity as a driver of change within the lake and catchment over the length of our record. Evidence from both records, however, supports the idea that lake levels were low from $\sim$ AD 700 to AD 1000, with increasing aridity from AD 1100 to 1600, interrupted by brief wet phases around AD 1000 and 1400. Wetter conditions are recorded in the 1700s, but drought returned by the end of the century and into the early 1800s, becoming wetter again from the mid-1800s. Comparison with other records across East Africa suggest that while some events are widespread (e.g., aridity beginning $\sim$ AD 1100), at other times there is a more complex spatial signature (e.g., in the 1200s–1300s, and from the 1400s–1600s). This study highlights the important role of catchment-specific factors in modulating the sensitivity of proxies, and lake records, as indicators of environmental change, and potential hazards when regional inference is based on a single site or proxy.

With surface areas between 0.04 and 5.1 km², the lakes in the studies discussed so far are sufficiently small and shallow for chironomid paleoecologists to assume more-or-less complete spatial integration of chironomid death assemblages across the lake before permanent burial. Under those conditions, if no differential preservation among taxa occurs, and the relative abundances of chironomid species and ecological groups present in the fossil assemblages will reflect the fractional distribution of various types of bottom habitat at the time of deposition. These conditions are clearly not met in larger African lakes, as demonstrated by Eggermont et al. (2007). This study investigated the within-lake variability of surface-sediment chironomid assemblages in the fairly large (~ 100 – 170 km^2 in 1983) and shallow ($Z_{\text{max}} = 5$ – 8 m) fluctuating tropical lake basin of Lake Naivasha, and compared the patterns observed with those in its two smaller satellite basins, one similarly shallow (Lake Oloidien, 5.1–5.7 km², 5–8 m), the other deep and stratified (Crescent Island Crater, 1.9 km², 14–17 m). This study showed that although sedimentation dynamics in the wind-stressed main basin of Naivasha are dominated by frequent resuspension and random sediment distribution, the near-to-offshore gradient in chironomid habitat remains imprinted on subfossil assemblages. The same is true for the African great lakes, such as Lake Victoria and Lake Tanganyika, where fossil assemblages must be taken to only represent habitat conditions close to their area of deposition at the time of their deposition, contrasting with other (distant) locations within the same lake. Indeed, Eggermont et al. (2008a) analyzed the distribution of chironomid larvae along a depth transect (0–80 m) at Kigoma Bay in Lake Tanganyika, and showed a fairly high correspondence between inferred habitat preferences of fossil and live collections. In these conditions, surface-sediment death assemblages deposited recently at various depths and microhabitats within the lake can be used as modern reference data for comparison with fossil assemblages deposited at a master core location during certain

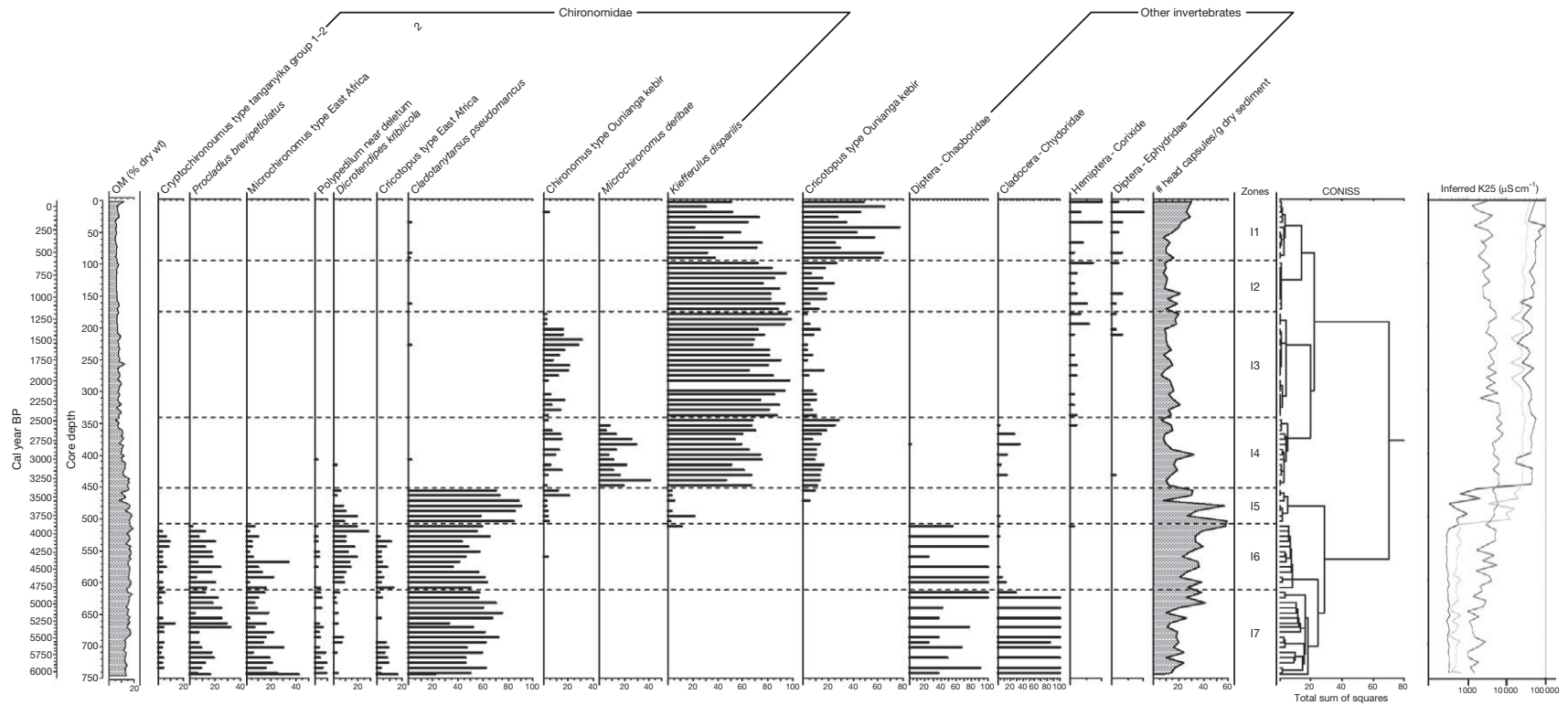


Figure 2 Stratigraphic distribution of aquatic invertebrate remains in Lake Yoa (northeastern Chad). Taxa are arranged from left to right according to their increasing weighted-average conductivity optimum. Horizontal lines indicate invertebrate zone boundaries. Reproduced from Eggermont H, Verschuren D, Fagot M, Rumes B, Van Bocxlaer B, and Kröpelin S (2008b) Aquatic community response in a groundwater-fed desert lake to Holocene desiccation of the Sahara. *Quaternary Science Reviews* 27: 2411–2425.

times in the past. Verschuren et al. (2002) and Eggermont and Verschuren (2003) exploited this principle in their use of fossil Chironomidae as tracers of anthropogenic impact on Lake Victoria and Lake Tanganyika.

Verschuren et al. (2002) addressed the question of whether the eutrophication and water-quality loss of Lake Victoria during recent decades resulted primarily from food-web alterations following the population explosion of Nile perch (*Lates niloticus*) in the early 1980s, or alternatively from increased nutrient inputs due to large-scale agricultural conversion of its drainage basin over the past half century. Fossil chironomid assemblages extracted from a mid-lake sediment core (Figure 3) traced the evolution of deep-water oxygenation through time. To strengthen their inferences, the authors first calibrated the relationship between the fossil abundance ratio of *Procladius brevipetiolatus* to *Chironomus* nr. *alluaudi* (in this paper identified as *Chironomus imicola*) and lake depth, or more specifically the duration of seasonal bottom anoxia, which increases with depth. They did so by analyzing recently deposited chironomid death assemblages in a transect of surface-sediment samples between 48 and 68 m water depth. The core analyses revealed that the hypolimnion of Lake Victoria was adequately oxygenated year-round until about 1960. Seasonal depletion of deep-water oxygen started shortly after that time and expanded strongly in the 1970s, hence, possibly contributing to the 1980s' collapse of indigenous fish stocks by eliminating suitable habitat for deep-water cichlids. Concentration data on diatom silica in the same core profile (Figure 3) further indicated that the primary (algal) productivity of Lake Victoria had already started increasing

from the 1930s onwards, that is, well before the introduction of Nile perch, and in proportion to historical agricultural production in the Lake Victoria region. The combined data thus suggested that the eutrophication of Lake Victoria, characterized by blooms of noxious algae, is due mainly to the excess nutrient input resulting from anthropogenic landscape disturbance.

Eggermont and Verschuren (2003) used fossil chironomid remains as model organisms to investigate the impact of sediment pollution (due to agricultural soil erosion) on the benthic invertebrate fauna of Lake Tanganyika. These authors did not look at stratigraphic changes in chironomid faunal composition through time, but compared the species composition of recently buried death assemblages in bottom habitat adjacent to pristine and disturbed watersheds. The analysis involved 36 surface-sediment samples from between 53 and 189 m water depth, each representing an area of nearshore sublittoral habitat adjacent to different size classes (<10 km², 30–50 km², 50–162 km², >162 km²) of either pristine or disturbed drainages. This data set was explored with uni- and multivariate techniques to evaluate relationships between species distribution and selected environmental variables. In this analysis, water depth was used as a proxy for the oxygen regime at the sampling site, and sediment composition (organic matter and carbonate content, fractions of coarse plant debris and sand) as a proxy for local substrate and food quality. The distance to shore, the surface area of the adjacent drainage, and its land-use regime were used as proxies for the susceptibility of each particular site to excess sedimentation. The authors found that in each of the three smaller drainage size classes (of the largest size class, no pristine example remained

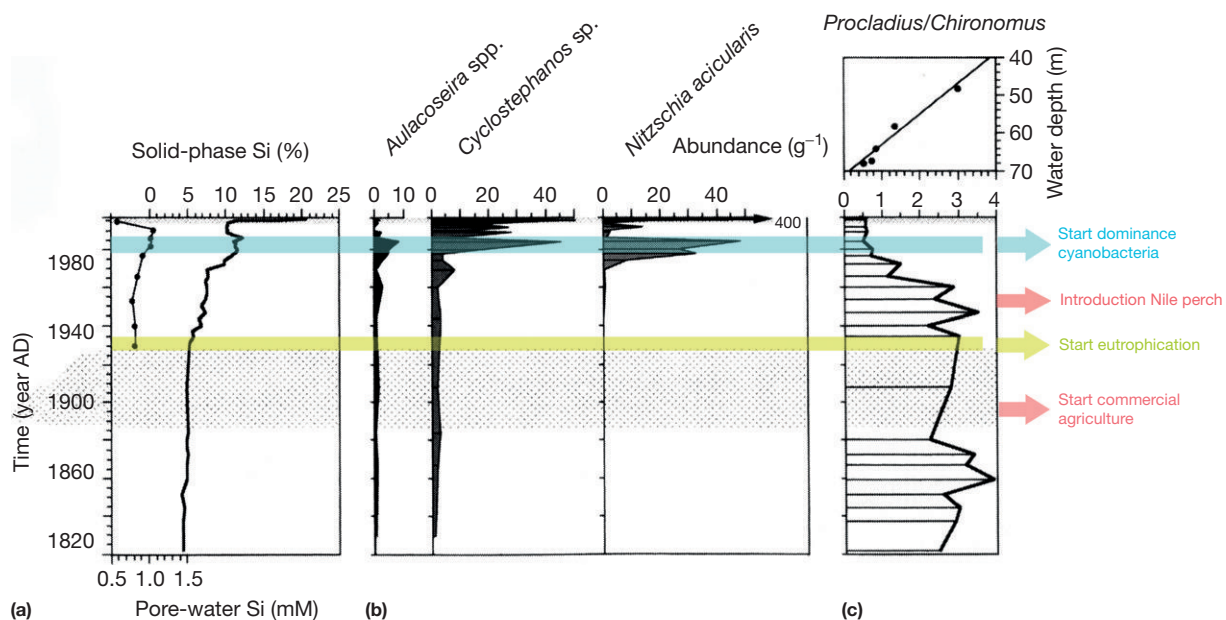


Figure 3 Evidence for lake-wide ecosystem change in Lake Victoria, East Africa. (a) stratigraphy of biogenic silica (mostly fossil diatoms) and dissolved silica in pore waters; the stippled zone reflects the inferred core hiatus corresponding to an unconformity in unsupported ²¹⁰Pb activity at 25 cm depth. (b) Fossil diatom stratigraphy; peak concentrations in flocculent surface muds are diatoms subject to further dissolution before permanent burial. (c) Abundance ratio of the anoxia-intolerant *Procladius brevipetiolatus* versus the anoxia-tolerant *Chironomus* nr. *alluaudi* in mid-lake fossil assemblages compared with its modern-day gradient in relation to water depth as determined by the local duration of seasonal bottom anoxia. Reproduced from Verschuren D, Johnson TC, Kling HJ, et al. (2002) History and timing of human impact on Lake Victoria, East Africa. *Proceedings of the Royal Society of London B* 269: 289–294.

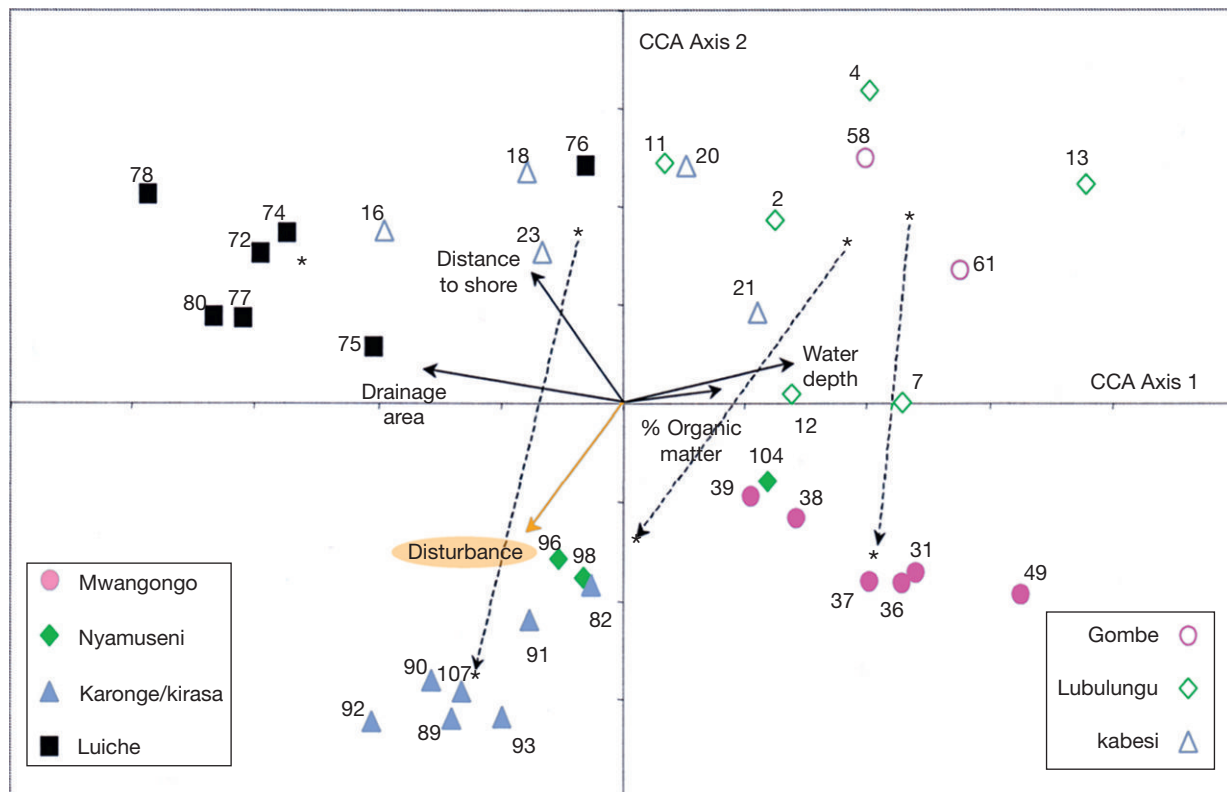


Figure 4 Canonical correspondence analysis (CCA) ordination of surface-sediment chironomid faunal assemblages in 38 sublittoral bottom environments of northeastern Lake Tanganyika, East Africa, adjacent to pristine (open symbols) and disturbed (closed symbols) drainages ranging from very small ($<10 \text{ km}^2$; circles) to very large ($>162 \text{ km}^2$; squares). Excess clastic sedimentation due to human disturbance of adjacent stream drainages resulted in directional shifts in faunal composition (arrows), indicating habitat change equivalent to reductions in water depth and distance to shore. Reproduced from Eggermont H and Verschuren D (2003) Impact of soil erosion in disturbed tributary drainages on the benthic invertebrate fauna of Lake Tanganyika, East Africa. *Biological Conservation* 113: 99–109.

in the study area), catchment disturbance had resulted in similar shifts in chironomid faunal composition (Figure 4), which apparently reflected an expansion of coarse, low-organic substrates into deeper water, at the expense of the soft organic muds which normally characterize the deep-water environment. Because the area of habitable lake bottom in Lake Tanganyika is bounded by the permanent oxycline at $\sim 100 \text{ m}$ depth, these results indicated that sediment pollution is already negatively affecting oxygenated soft-bottom environments in Lake Tanganyika, and that intensifying soil erosion in its tributary drainages may eventually threaten the survival of endemic fish and mollusc species restricted to those same environments. This study established the value of chironomid death assemblages as biological indicators of ecosystem health on the scale at which human impact occurs, which is that of entire drainage basins.

Methodological Advances in African Chironomid Paleoecology

Qualitative Inferences Using the Indicator-Species Approach

Comprehensive lake surveying and collecting efforts in tropical Africa often face considerable logistical problems, sometimes

compounded by political obstacles. Consequently, until recently no large uniform reference data sets existed on the distribution of chironomid species across modern climate and ecological gradients. This obviously limited paleoecologists' opportunities for chironomid-based quantitative reconstruction of past environmental change. Existing autecological information on local species is also mostly anecdotal, and in contrast with other biological proxy indicators such as diatoms, it can be used for paleoecological inferences only when both adult and larval morphology are known. As a result, workers interpreted stratigraphic change in fossil chironomid assemblages on the basis of traditional ecological typology and genus-level autecological data transferred from the Holarctic region. Prominent in this context is *Chironomus*, a genus of which most member species in both tropical and Holarctic regions are tolerant of poor oxygen conditions (as is often associated with anthropogenic nutrient enrichment, but occurs naturally in many stratified tropical lakes). Another example is the subfamily Diamesinae, the members of which are largely restricted to cold-water environments and therefore characteristic indicators of cold climatic episodes during the last deglaciation in north-temperate Europe and North America. Fossil Diamesinae have also been reported from south-temperate New Zealand, but not yet from Africa. However, the occurrence

of at least three *Diamesa* species in alpine lakes and streams on Africa's highest mountains suggests they might become valuable tracers of tropical glacier advance and retreat.

Among chironomid-based paleoclimate research in Africa using the indicator-species approach, studies on past salinity change in lowland lakes (Mees et al., 1991; Verschuren et al., 1999b) took advantage of a fair amount of preexisting autecological information on the local species (Verschuren, 1997); the study dealing with montane peat-bog fauna (Verschuren et al., 1996) relied almost exclusively on ecological data from related Holarctic taxa.

Development of Quantitative Inference Models

Quantitative reconstruction of lake and climate history based on fossil chironomid assemblages requires calibration of the relationship between chironomid species distribution and modern environmental conditions in an appropriate number of physically, chemically, and/or ecologically diverse lakes. Initially lacking a traditional surface-sediment calibration data set where this relationship is determined using recent death assemblages of larval remains extracted from surface sediments, the first exploratory model for quantitative salinity inference (Verschuren et al., 2004) was developed using published distribution data based on regional surveys of the living fauna (adults and/or larvae). Using the computationally simple method of weighted-averaging, the authors produced inference models with a statistical performance comparable to that of weighted-averaging salinity-inference models based on African diatoms (Gasse et al., 1995). For reconstruction of lake-level change in the Kenya Rift Valley, Verschuren et al. (2000a) then performed a hybrid procedure in which this inference model, calibrated with presence-absence distributional records, was applied to abundance-weighted fossil data. This ensured that not only the appearance or disappearance of taxa but also changes in their relative abundance would generate a proxy climate signal. Because of the rather small number of calibration data (20 taxa recovered from 32 sites, or 147 species-conductivity combinations only) the model's statistical power is somewhat limited. Regression of inferred over observed conductivity in the 32 reference lakes also suggested a threshold response of African chironomid communities at the transition between freshwater and salt-lake environments ($\sim 3000 \mu\text{S cm}^{-1}$), as opposed to the more continuous species replacement along the salinity gradient seen in diatoms. To evaluate whether this does indeed reflect ecological reality, or is an artifact of the inference model due to the use of presence-absence calibration data that perhaps do not adequately cover the regional salinity gradient, Eggermont et al. (2006) constructed a true paleoenvironmental calibration data set based on percent-abundance data of 65 chironomid morphotypes recovered from the surface sediments of 67 lakes in equatorial East Africa (Kenya, Uganda, Ethiopia, and Tanzania), and then applied various numerical techniques to this calibration data set to develop a series of salinity-inference models (Figure 5(a)). These new models are evidently more statistically robust than the exploratory model developed previously, but the observed threshold response of African chironomid communities to increasing salinity remained. To improve model sensitivity, the authors experimented with modified

calibration data sets in which the regional salinity gradient was extended further (by adding dilute high-elevation lakes from Mt. Kenya and the Ruwenzori Mountains.) or in which fresh and saline lakes were represented in equal numbers (to shift upward the salinity optimum of salt-tolerant taxa). Neither experiment appreciably improved the inference models' statistical performance.

Validation of the new African chironomid-salinity inference models in the time domain, in part based on direct comparison of chironomid-inferred salinity trends with those based on fossil diatoms, revealed significant variation among models in reconstructed salinity trends through time, due to their different sensitivity to the presence or relative abundance of certain key chironomid taxa. Models based on simple weighted-averaging and the modern analog techniques were inferior to more sophisticated models (such as maximum likelihood models) on the basis of error statistics. However, these simpler models produced the most trustworthy salinity reconstructions. The authors suggested that this occurred exactly because their step change of inferred conductivity near $3000 \mu\text{S cm}^{-1}$ mirrors the rapid climate-driven transitions between fresh and saline lake phases actually happening in these shallow closed-basin lakes. In a broader context, Eggermont et al. (2006) stressed that selection of the most appropriate transfer function for environmental reconstruction using fossil chironomids (or any other biological proxy indicator) should not only compare their statistical performance, but also consider whether models produce ecologically meaningful results with high signal content.

Verschuren and Eggermont (2007) compared the performance of the above-mentioned salinity-inference models, which are all based on $>100 \mu\text{m}$ size fractions of surface-sediment fossil assemblages, with models based on the $>150 \mu\text{m}$ size fractions only. They found that despite $\sim 35\%$ additional fossil loss, restriction to $>150 \mu\text{m}$ data did not appreciably affect individual taxon optima and tolerances, and resulted in only modest reduction of mean taxon richness per lake. Parameters of statistical model performance were as good as, or better than those of models based on the $>100 \mu\text{m}$ data. Also, the application of inference models based on $>150 \mu\text{m}$ or $>100 \mu\text{m}$ data to a 200-year fossil record from Lake Abiyata, Ethiopia, produced very similar trends and amplitudes of inferred past salinity change. Using the larger mesh size reduced sample processing time by up to 50%, by removing a significant proportion of visually obstructive organic debris, and allowing a greater fraction of chironomid fossils to be identified directly in the sorting tray. The fraction of first instar group taxa in surface-sediment samples was cut from 13% to 3%, increasing the mean taxonomic resolution of fossil assemblages, and thus their ecological specificity.

Fossil assemblages of chironomid larvae are well-established paleothermometers in north-temperate and boreal regions, but their potential for temperature reconstruction in tropical regions had not been assessed until recently. Eggermont et al. (2010a) surveyed subfossil chironomid assemblages in the surface sediments of 65 (fresh) lakes and permanent pools in southwestern Uganda (including the Ruwenzori Mountains) and central and southern Kenya (including Mt. Kenya) to document the modern distribution of African chironomid communities along a regional temperature

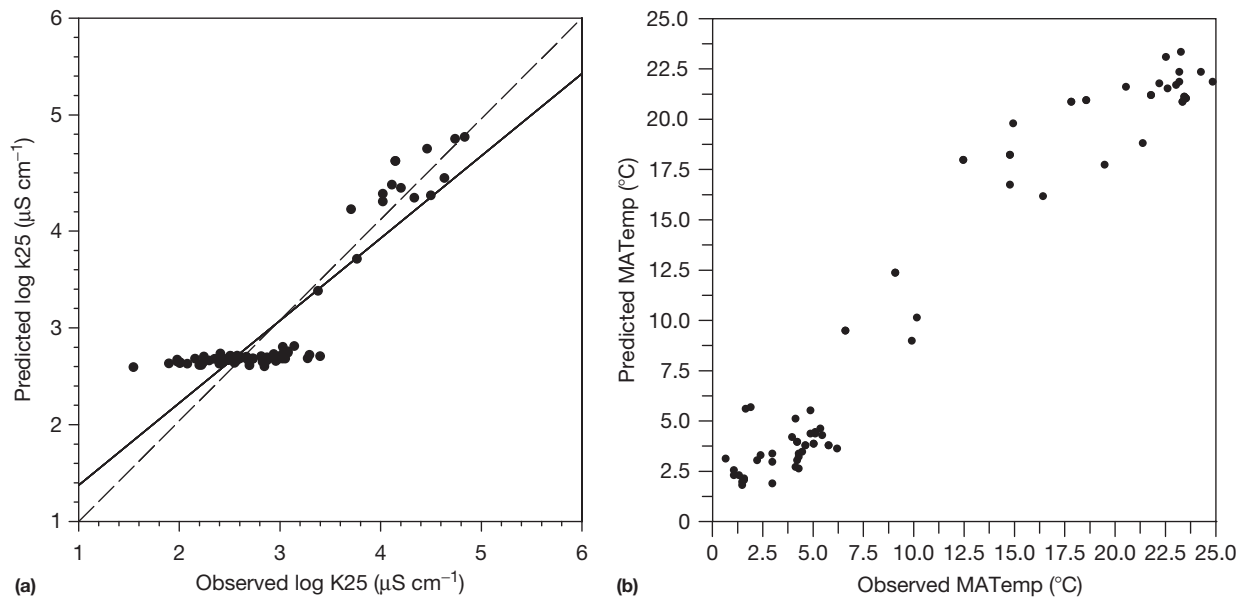


Figure 5 African chironomid-based inference models using weighted-averaging for (a) salinity reconstruction and (b) mean annual air temperature (MATemp) reconstruction. Illustrated are the scatter plots of predicted versus observed values. Data from H. Eggermont.

gradient covered by lakes between 489 and 4575 m a.s.l. (mean annual air temperature range: 1.1–24.9 °C; surface water temperature range: 2.1–28.1 °C). Similar to the salinity trial models, the authors applied various numerical techniques to this calibration data set to develop a series of temperature-inference models, and also ran trials with subsets of lakes with varying gradient lengths of temperature and other environmental variables. All inference models calibrated against temperature produced error statistics ranking among the best when compared to similar work on other continents (Figure 5(b)). Models based on calibration over the full temperature gradient suffer somewhat from the limited number of study sites at intermediate elevation (2000–3000 m), and from the presence of morphologically indistinguishable but ecologically distinct taxa. Calibration confined to high-elevation sites (>3000 m) has poorer statistics, but is less susceptible to biogeographical and taxonomic complexities.

Testing the Response of African Chironomidae to Climate Change

Verschuren et al. (1999a, 2000b) used dated sediment records linked to long time series of historical data to investigate the operative limits of African Chironomidae as biological climate proxies. These studies employed the sediment records of satellite basins of Lake Naivasha in Kenya, which, because of their hydrological connections, have all experienced the same 120-year history of documented lake-level change. Instrumentally recorded lake depth was linked to the sediment record by ^{210}Pb dating, whereas past mixing regime, salinity, and benthic habitat conditions were reconstructed using lithological indicators, fossil diatoms, and plant macrofossils, respectively. In both studies, chironomid species-assemblage changes through time were entirely consistent with available autecological data and known salinity tolerance ranges. Salinity change and the

temporal occurrence of littoral papyrus swamp together explained about half of the historical variation in benthic community composition, and their effects appear to be largely independent from each other. Most importantly, these studies illustrated the complexity of biological response to short-term (decade-scale) water-balance fluctuations. Specifically, the direct physiological problem of osmotic stress caused by salinity change appears to affect African zoobenthos communities only at the extremes of the salinity gradient. A significant portion of the observed correlation between salinity and invertebrate community structure is in fact an indirect effect of diffuse covariance between salinity, lake level, and the distribution of various types of benthic habitat.

Eggermont et al. (2010b) assessed whether high-altitude lakes in the Rwenzori Mountains (Uganda – DR Congo) are sensitive to climate-driven environmental change of the same order of magnitude as that expected to result from current and future anthropogenic global warming. This was done by comparing the species assemblages of larval chironomid remains deposited recently in lake sediments with those deposited at the base of short cores (dated to within of shortly after the Little Ice Age) in 16 lakes. Chironomid-based temperature-reconstruction models were made using the East-African temperature-inference models mentioned above (Eggermont et al., 2010a). Excluding one atypical mid-elevation lake, the authors found a three-to-one ratio in cases of inferred warming versus cases of inferred cooling. The chironomid-inferred temperature changes mostly fell within the error range of the regional temperature-inference models, but a generalized linear mixed model analysis of the combined result from all lakes nevertheless indicated significantly warmer mean annual temperatures (on average $+0.38 \pm 0.11$ °C) at present compared to between ~85 and ~645 years ago. This result was independent of basal core age and of whether lakes were located in glaciated or nonglaciated catchments, suggesting that at least part of the

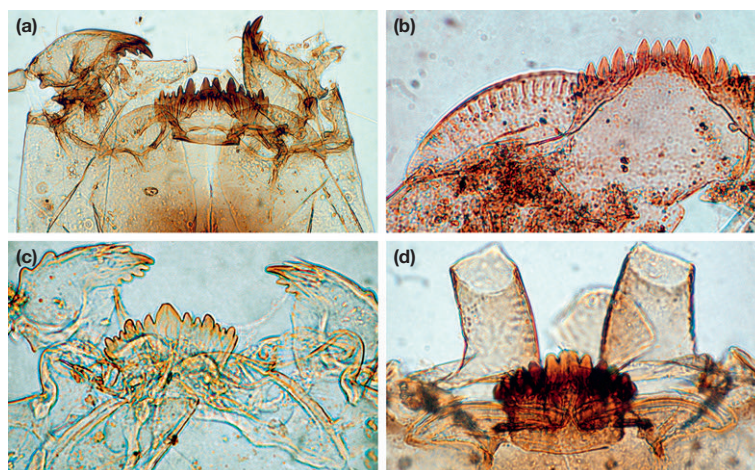


Figure 6 Subfossil remains of larval Chironomidae recovered from African lake sediments. (a) *Polypedilum* near *deletum* Goetghebuer, (b) *Conochironomus* type Gagouba, (c) *Chironomini* indet. type *Tanganyika* sp. 1., (d) *Tanytarsini* indet. type Large Hall Tarn. Photos by H. Eggermont.

signal is due to relatively recent, anthropogenic warming. Analysis of the patterns of faunal change at each lake in relation to established species–environment relationships does suggest that part of the observed shifts in species composition reflect lake-specific evolution in habitat features other than temperature, such as nutrients, pH or oxygen regime, which covary with temperature to greater or lesser extent.

Taxonomy and Biogeography of African Chironomidae

Progress in the discrimination of subfossil morphotypes of African chironomid larvae (Eggermont and Verschuren, 2004a,b, 2007; Eggermont et al., 2005, 2008a; Verschuren, 1997; Figure 6) continues to increase the taxonomic diversity of fossil assemblages. Judging from comparison with morphological diversity within and between genera in the better known Holarctic fauna, most fossil larval morphotypes currently distinguished in African contexts must be functionally equivalent to biological species, or groups of closely related species. Maintaining a strong link to official chironomid taxonomy and systematics based on the morphology of adult midges ensures compatibility between ecological data derived from living and fossil assemblages, and justifies the use of chironomid fossils as biological proxy indicators. Given this level of taxonomic resolution, the African fossil material can also be exploited to address biogeographical issues about the living fauna, such as its patterns of alpha- and beta-diversity (taxonomic richness of lakes and regions, e.g., Eggermont et al., 2005). It can also be exploited to assess the biological uniqueness of particular aquatic ecosystems in the context of biodiversity conservation. Most of all, in regions such as Africa where we have only fragmentary taxonomic and ecological knowledge of individual chironomid species, taxon-discrimination criteria based on fossil material can also improve species discrimination in the living material. If this stimulates new applied ecological research on the benthic invertebrate communities of Africa's modern aquatic ecosystems, the knowledge gained may eventually feed back into improved interpretation of

chironomid stratigraphy for the purpose of paleoenvironmental reconstruction.

See also: [Chironomid Records: Chironomid Overview; Postglacial Europe.](#)

References

- Bonnefille R, Roeland JC, and Guiot J (1990) Temperature and rainfall estimates for the past 40,000 years in Equatorial Africa. *Nature* 346: 347–349.
- Eggermont H, De Deyne P, and Verschuren D (2007) Spatial variability of chironomid death assemblages in surface sediments of a shallow fluctuating tropical lake (Lake Naivasha, Kenya). *Journal of Paleolimnology* 38: 309–328.
- Eggermont H, Heiri O, Russell J, Vuille M, Audenaert L, and Verschuren D (2010) Paleotemperature reconstruction in tropical Africa using fossil Chironomidae (Insecta: Diptera). *Journal of Paleolimnology* 43: 413–435.
- Eggermont H, Heiri O, and Verschuren D (2006) Fossil Chironomidae (Insecta: Diptera) as quantitative indicators of past salinity in African lakes. *Quaternary Science Reviews* 25: 1966–1994.
- Eggermont H, Kennedy D, Hasiotis ST, Verschuren D, and Cohen AS (2008) Distribution of larval Chironomidae (Insecta: Diptera) along a depth transect at Kigoma Bay, Lake Tanganyika: Implications for paleoecology and paleoclimatology. *African Entomology* 16: 162–184.
- Eggermont H and Verschuren D (2003) Impact of soil erosion in disturbed tributary drainages on the benthic invertebrate fauna of Lake Tanganyika, East Africa. *Biological Conservation* 113: 99–109.
- Eggermont H and Verschuren D (2004a) Subfossil Chironomidae from East Africa. 1. Tanytarsini and Orthocladiinae. *Journal of Paleolimnology* 32: 383–412.
- Eggermont H and Verschuren D (2004b) Subfossil Chironomidae from East Africa. 2. Chironomini and Tanytarsini. *Journal of Paleolimnology* 32: 413–455.
- Eggermont H and Verschuren D (2007) Taxonomy and diversity of Afroalpine Chironomidae (Insecta: Diptera) on Mount Kenya and the Rwenzori Mountains, East Africa. *Journal of Biogeography* 34: 69–89.
- Eggermont H, Verschuren D, Audenaert L, et al. (2010) Limnological and ecological sensitivity of Rwenzori mountain lakes to climate warming. *Hydrobiologia* 648: 123–142.
- Eggermont H, Verschuren D, and Dumont H (2005) Taxonomic diversity and biogeography of Chironomidae (Insecta: Diptera) in lakes of tropical West Africa, using sub-fossil remains extracted from surface sediments. *Journal of Biogeography* 32: 1063–1083.
- Eggermont H, Verschuren D, Fagot M, Rumes B, Van Bocxlaer B, and Kröpelin S (2008) Aquatic community response in a groundwater-fed desert lake to Holocene desiccation of the Sahara. *Quaternary Science Reviews* 27: 2411–2425.

- Gasse F, Juggins S, and Khelifa LB (1995) Diatom-based transfer functions for inferring hydrochemical characteristics of African paleolakes. *Palaeogeography, Palaeoclimatology, Palaeoecology* 177: 31–54.
- Hoelzmann P, Gasse F, Dupont LM, et al. (2004) Palaeoenvironmental changes in the arid and subarid (Sahara-Sahel-Arabian Peninsula) from 150 kyr to present. In: Batterbee RW, Gasse F, and Stickley CE (eds.) *Past Climate Variability Through Africa and Europe. Developments in Paleoenvironmental Research* 6: 638.
- Kroepelin S, Verschuren D, Lézine AM, et al. (2008) Climate-driven ecosystem succession in the central Sahara: The last 6000 years. *Science* 320: 765–768.
- Mees F, Verschuren D, Nijs R, and Dumont HJ (1991) Holocene evolution of the crater lake at Malha, Northwest Sudan. *Journal of Paleolimnology* 5: 227–253.
- Merritt RW and Cummins KW (eds.) (1984) *An Introduction to the Aquatic Insects of North America*, p. 722. Dubuque, IA: Kendall/Hunt Publishing Company.
- Rosenberg DM (ed.) (1993) *Freshwater Biomonitoring and Benthic Macroinvertebrates*, p. 488. New York: Chapman & Hall.
- Ryner M, Gasse F, Rumes B, and Verschuren D (2007) Climatic and hydrological instability in semi-arid equatorial East Africa during the late Glacial to Holocene transition: A multiproxy reconstruction of aquatic ecosystem response in northern Tanzania. *Palaeogeography, Palaeoclimatology, Palaeoecology* 248: 440–458.
- Ryves DB, Mills K, Bennike O, et al. (2011) Environmental change over the last millenium recorded in two contrasting crater lakes in western Uganda, East Africa (Lakes Kasenda and Wandakara). *Quaternary Science Reviews* 30: 555–569.
- Verschuren D (1997) Taxonomy and ecology of subfossil Chironomidae (Insecta: Diptera) from Rift Valley lakes in central Kenya. *Archiv für Hydrobiologie Beihefte* 107: 467–512.
- Verschuren D, Cocquyt C, Tibby J, Roberts N, and Leavitt PR (1999) Long-term dynamics of algal and invertebrate communities in a fluctuating tropical soda lake. *Limnology and Oceanography* 44: 1216–1231.
- Verschuren D, Cumming BF, and Laird KR (2004) Quantitative reconstruction of past salinity variations in African lakes using fossil midges (Diptera: Chironomidae): Assessment of inference models in space and time. *Canadian Journal of Fisheries and Aquatic Sciences* 61: 986–998.
- Verschuren D, Dumont HJ, and Armengol-Diaz J (1996) Utilisation de cladocères et chironomides fossiles pour reconstruire l'évolution hydrologique de leur habitat marécageux dans la tourbière de Kashiru (Burundi) depuis 40.000 ans BP. *Paleoecology of Africa and the Surrounding Islands* 24: 109–120.
- Verschuren D and Eggermont H (2006) Quaternary paleoecology of aquatic Diptera in tropical and Southern Hemisphere regions, with special reference to the Chironomidae. *Quaternary Science Reviews* 25: 1926–1947.
- Verschuren D and Eggermont H (2007) Sieve mesh size and quantitative chironomid paleoclimatology. *Journal of Paleolimnology* 38: 329–345.
- Verschuren D, Johnson TC, Kling HJ, et al. (2002) History and timing of human impact on Lake Victoria, East Africa. *Proceedings of the Royal Society of London B* 269: 289–294.
- Verschuren D, Laird KR, and Cumming BF (2000) Rainfall and drought in equatorial East Africa during the past 1100 years. *Nature* 403: 410–414.
- Verschuren D, Tibby J, Leavitt PR, and Roberts CN (1999) The environmental history of a climate-sensitive lake in the former 'White Highlands' of central Kenya. *Ambio* 28: 494–501.
- Verschuren D, Tibby J, Sabbe K, and Roberts CN (2000) Effects of lake level, salinity and substrate on the invertebrate community of a fluctuating tropical lake. *Ecology* 81: 164–182.
- Walker IR (1995) Chironomids as indicators of past environmental change. In: Armitage P, Cranston PS, and Pinder LVC (eds.) *The Chironomidae: The Biology and Ecology of Non-biting midges*, pp. 405–422. London: Chapman & Hall.
- Walker IR (2001) Midges: Chironomidae and related Diptera. In: Smol J, Birks HJB, and Last W (eds.) *Tracking Environmental Change using Lake Sediments: Zoological Indicators. Developments in Paleoenvironmental Research* 4: 43–66.

Late Pleistocene of Europe

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Introduction

Chironomids, nonbiting midges, are among the most versatile indicators of paleoenvironmental change. They have been used to reconstruct, both quantitatively and qualitatively, a variety of environmental variables. They have proved to be particularly useful as temperature proxies. Until the 1990s, paleoecologists used chironomids in qualitative studies of climate change. They assessed changes in chironomid assemblages, from those dominated by thermophilic taxa to those dominated by cold stenothermic taxa, as a reflection of relative changes in climate. However, with the advent of numerical techniques, which take the modern temperature optima of taxa and transfer these to fossil assemblages, it is now possible to use chironomids quantitatively to reconstruct past climatic change at high temporal resolution with an error of about ± 1.0 °C. In this article, the use of chironomids as climate proxies for sequences during the late Pleistocene in Europe is reviewed. First, papers in which chironomids were used as qualitative indicators are reviewed and then an account of studies in which chironomids have been used quantitatively to estimate changes in mean July air temperature is presented. Finally, the results of these studies are compared with results obtained from other climate proxies and the relative performance of chironomids as climate proxies is assessed.

Chironomids as Qualitative Indicators of Climate Change in the late Pleistocene

Chironomids have been used for 80 years to investigate past environmental change. However, like Gams' (1927) pioneering work, the majority of these studies have focussed on Holocene sequences. Nevertheless, among the earliest uses of chironomids in a paleoecological context was the work of Andersen (1938) who investigated a late-glacial sequence from Denmark. However, it was not until the 1970s–1980s that the Late Glacial once more became the focus of paleoenvironmental studies using chironomids, when a series of papers were published by Wolfgang Hofmann.

Hofmann (1978) found a diverse late-glacial chironomid fauna at Grosser Segeberger See in northern Germany. In this study, the Younger Dryas was dominated by cold stenothermic taxa of the *Tanytarsus lugens* community. During the Bølling, *Corynocera ambigua* comprised over 40% of the chironomid fauna. Other taxa included *Microtendipes* and *Procladius*. In several other studies (see Section 'Chironomids as Quantitative Indicators of Climate Change in the late Pleistocene'), *C. ambigua* rapidly dominates late-glacial chironomid assemblages and just as rapidly declines after a relatively short period of time. Hofmann found similar trends in late-glacial chironomid assemblages at Schöhsee (Hofmann, 1971). Insufficient data were available from the Allerød to draw any firm conclusions.

Lobsigensee, on the Swiss Plateau, was the focus of a study into the response of midge assemblages to climate change during the Late Glacial (Hofmann, 1983a). The Oldest Dryas was characterized by an assemblage dominated by *Sergentia coracina* and *Chironomus anthracinus* type. Early in this period *Microtendipes*, *Paratanytarsus*, and *C. ambigua* were also abundant, but toward the end of the Oldest Dryas these taxa declined and were replaced by *Psectrocladius*, which coincided with the establishment of *Betula nana* in the catchment. These changes are consistent with climatic warming from relatively cold conditions. The cold-adapted fauna included cold stenotherms such as *S. coracina* and *C. ambigua*, and *Microtendipes*, which is cool tolerant. This fauna was replaced by a temperate climate fauna in which the thermophilic taxon *Psectrocladius* thrived. A major change in the fauna occurred at the onset of the Bølling when the fauna became dominated by just three thermophilic taxa, namely *Microtendipes*, *Chironomus*, and *Psectrocladius*. In the Allerød, the fauna became even less diverse when 89% of the specimens were either *Chironomus* or *Tanytarsus* sp. C (Hofmann, 1971) but there was also a large increase in *Dasyhelea*, a ceratopogonid biting midge. This suggests that while temperate conditions continued to prevail, water levels in the lake also fell. The Lobsigensee core was taken in the littoral zone of the lake which would make it particularly sensitive to lake level change. Thus, changes in the chironomid fauna apparently indicated both changes from cold to temperate conditions and also a decline in precipitation. The core was sampled at relatively coarse intervals and so it was not possible to differentiate any response to relatively short-duration climatic oscillations.

In order to determine whether late-glacial climate was driving changes in chironomid assemblages at other sites, Hofmann (1983b) analyzed sediments from a littoral core from Poolsee, a shallow lake in northern Germany. Once more he found that the early late-glacial chironomid fauna was dominated by taxa such as *Micropsectra*, *Heterotrissocladius*, *Protanypus*, *S. coracina*, and *Paracladius*, which are typical of cold, oligotrophic conditions. Hofmann argued that because these taxa, which occur in the profundal zone of temperate lakes, were occurring in the littoral zone of the lake, they must be indicative of a cold climate rather than trophic conditions. These taxa were replaced by thermophilic taxa as temperature warmed in post glacial times. Similar patterns emerged from studies at other sites. The early late-glacial chironomid fauna at Meerfelder Maar, southern Germany (Hofmann, 1984), was dominated by cold stenothermic taxa such as *Diamesa*, *Paracladopelma nigrifulva*, and *Micropsectra*, but these were later replaced by taxa associated with more thermophilic conditions such as *Psectrocladius* and *Paratanytarsus*.

In a study of cores from three lakes in the Eifel region of southern Germany and a lake from the Massif Central in central France, Hofmann (1990; 1991) found that Upper Pleniglacial sediments had relatively few chironomid remains but those present in the littoral zone of lakes (*Diamesa*, *Protanypus*,

Paracladopelma, and *Micropsectra*) were all indicative of cold conditions. During the Middle-Lower Pleniglacial and Early Glacial the increase in abundance and changes in dominance between *Micropsectra* and *Paracladius* indicated that climatic conditions may have been slightly warmer. During the Eemian Interglacial increasing species diversity and the presence of thermophilic species, including *Glyptotendipes*, indicated that conditions had become considerably warmer.

At Lago Albano, central Italy, the response of chironomids to climatic change between 28 and 17 ka BP was investigated as part of a multiproxy study (Guilizzoni et al., 2000; Manca et al., 1996). Between 28 and 26 ka BP chironomids were abundant and the fauna was dominated by *Paracladopelma* and *Micropsectra*. These taxa are indicative of cool, oligotrophic conditions. From 26 to 17 ka BP, very few chironomids were recovered but most were *Micropsectra*. Numbers increased once more during the Late Glacial and the fauna became more diverse, including *Micropsectra* but in addition temperate, littoral taxa such as *Dicoretendipes* and *Cladotanytarsus* (Manca et al., 1996). Increases in the overall abundance and diversity of chironomids were probably in response to increases in temperature, while cooler, drier periods resulted in a decrease in lake productivity and thus chironomid abundance.

Brooks (2000) studied the Late Glacial–early Holocene chironomid assemblages of four lakes on an altitudinal transect in the Swiss Alps (Gerzensee, 603 m; Leysin, 1230 m; Regenmoos, 1260 m; Hérémenche, 2300 m). This was a high-resolution study, and samples were analyzed at 0.5 cm intervals in some sections of the cores, in order to detect the chironomid response to low amplitude, short-duration cold oscillations. Because the lakes were each at different altitudes they supported quite different chironomid assemblages. However, if the chironomids were influenced by regional climate change then a synchronous response would be expected in all the lakes and this could be disentangled from local responses to in-lake environmental change, which would not affect the chironomid assemblages of all the lakes at the same time. A detailed chronology was obtained by wiggle-matching high-resolution oxygen isotope records from each lake against the Greenland ice core (GRIP) oxygen isotope record. Only Gerzensee and Leysin will be considered here since late-glacial sediments were not well-represented at the other two sites. A high-resolution record was obtained from Gerzensee where the sequence between 13 440 and 11 154 cal years BP was analyzed at approximately 10–30 years intervals (Figure 1). The late Interstadial was dominated by a temperate taxon, *Parakiefferiella bathophila*, which accounted for 40–60% of the chironomid assemblage. However, during the Gerzensee Oscillation (13 200–12 800 cal years BP), there was a significant increase in the abundance of a second *Parakiefferiella* species, which later dominated the Younger Dryas assemblage. This taxon was referred to as *Parakiefferiella* type A, and is the same as *Parakiefferiella* sp. B of Walker et al. (2003). At the start of the Younger Dryas *P. bathophila* decreased abruptly, and finally disappeared in the second half of this period. It was not until the early Holocene that *P. bathophila* returned and rapidly became a dominant part of the fauna once more. This taxon was apparently close to its thermal threshold at Gerzensee and therefore, was a sensitive indicator of climatic conditions at this site. At Leysin (Figure 2) *P. bathophila* was absent. Presumably temperatures were too low at this altitude to support a

population. *Parakiefferiella* type A became more abundant again during the Gerzensee Oscillation and Younger Dryas, but at this site *Einfeldia* was apparently close to its thermal threshold and was only present in the fauna during the cold oscillations. This study clearly demonstrated that chironomids could be useful paleoclimatic indicators because they were responding to shifts in regional climate. Synchronous responses to cool oscillations, which were independently identified from an oxygen isotope signal, were apparent in each lake from different elements of the chironomid fauna. Unfortunately, despite the development of a quantitative chironomid-temperature inference model based on a modern training set of Swiss lakes (Heiri, 2001; Heiri et al., 2003) it has not been possible to estimate temperatures from either of these two sites because *Parakiefferiella* type A was not represented in the training set. This taxon is present in southern Norway today but has apparently been eliminated from the Swiss fauna since the Late Glacial.

Gandouin and Franquet (2002) described a Late Glacial and Holocene chironomid fauna from Lac Long Inférieur in south-east France. The Late Glacial was not examined in detail but the fauna during this period was dominated by *T. lugens*-type and other cold stenotherms including *Paracladius* and *Diamesa* which indicate that cool, oligotrophic conditions prevailed. A peak in the abundance of *Micropsectra* suggested a cold oscillation in the mid-Interstadial but the chronology of the sequence was not sufficiently resolved to determine which Greenland Interstadial event this may reflect. A gradual decline in *T. lugens*-type and an increase in taxon diversity throughout the Interstadial may indicate an increase in temperature. However, this is contrary to most temperature reconstructions for this period. Another possible explanation for this result may be a decrease in lake level.

A relatively new and promising development has been the analysis of stable oxygen isotopes from the chitin in chironomid head capsules. This provides the attractive potential of obtaining two independent climate records from the same proxy. Work by Wooller et al. (2004) and Verbruggen et al. (2010a) demonstrated that there was a strong relationship between chironomid $\delta^{18}\text{O}$ and precipitation $\delta^{18}\text{O}$. Verbruggen et al. (2010b) then went on to show that chironomid $\delta^{18}\text{O}$ could be used to derive a late-glacial record from Rotsee, Switzerland, which was a good reflection of the $\delta^{18}\text{O}$ record based on bulk carbonate sediments from the lake. This work demonstrates that chironomids can be used to produce a $\delta^{18}\text{O}$ record from lakes on carbonate-poor bedrock and low in organic material, thus significantly expanding the number of lake sediment records that can be studied for variations in past lake water $\delta^{18}\text{O}$. However, Verbruggen et al. (2010b) caution that more information is required on the extent to which chironomid $\delta^{18}\text{O}$ is influenced by temperature before the technique can be used as a reliable temperature proxy.

Chironomids as Quantitative Indicators of Climate Change in the late Pleistocene

The development of quantitative chironomid-mean July air temperature inference models has revolutionized paleoclimate research. Now it is possible to obtain high resolution, low error, summer air temperature estimates from lake sediment

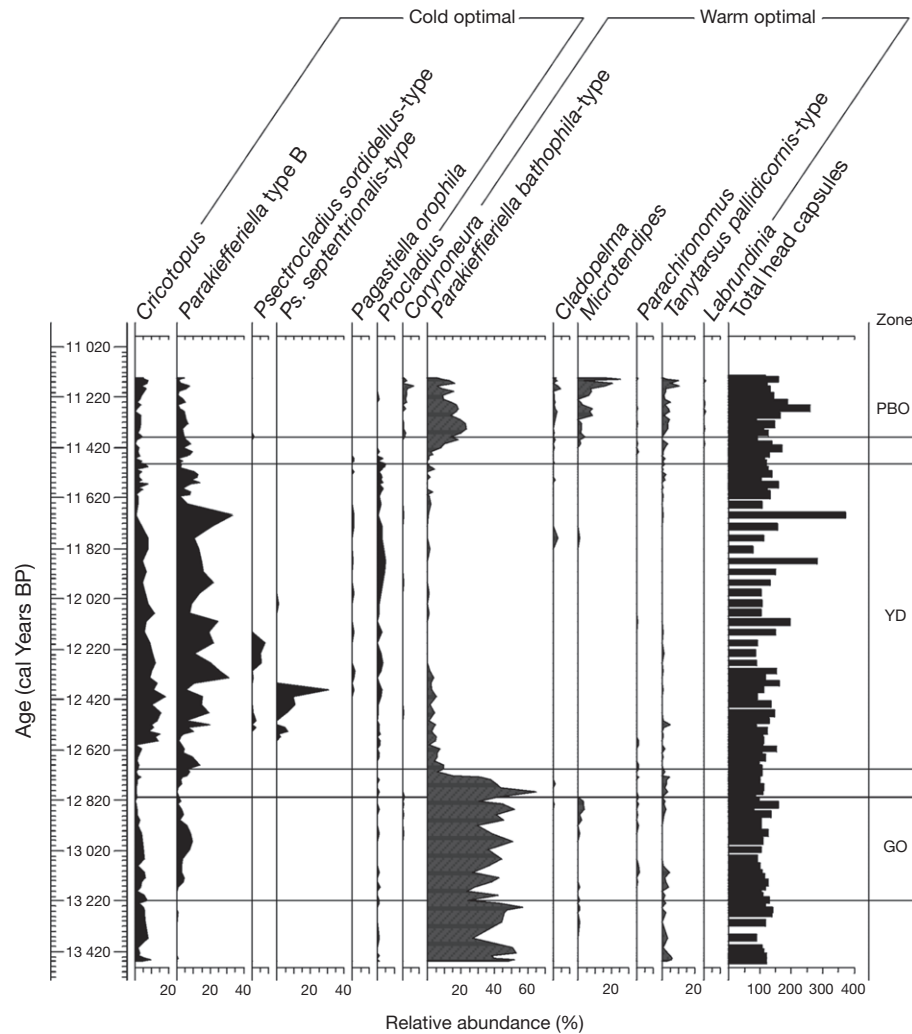


Figure 1 Response of Chironomidae to climate change around the time of the Younger Dryas at Gerzensee, Switzerland. GO, Gerzensee Oscillation; PBO, Preboreal Oscillation; YD, Younger Dryas. Reproduced from Brooks SJ (2000) Late-glacial fossil midge stratigraphies (Insecta: Diptera: Chironomidae) from the Swiss Alps. *Palaeogeography, Palaeoclimatology, Palaeoecology* 159: 261–279.

samples with the knowledge that the biological proxy has completed its development *in situ*. At the sampling resolution of most studies, the response of midge taxa to climate change shows little or no delay by virtue of the long-distance, wind-assisted, rapid dispersion of chironomid adults across the landscape. These chironomid-inferred temperature reconstructions can be compared with high resolution oxygen isotope records and quantitative reconstructions based on other proxies such as pollen and beetle assemblages to confirm their reliability.

The first late-glacial chironomid-inferred summer temperature records from Europe were published as part of a multiproxy study of Kråkenes Lake, western Norway (Brooks and Birks, 2000a). The chironomid-inferred temperatures were derived from a modern Norwegian 44-lake training set. The training set covered a mean July air temperature range of 5.7–14.0 °C but in fact only two lakes in the training set experienced mean July air temperatures of less than 8 °C and so the chironomid-inferred temperatures estimates derived from this model were likely to

overestimate the coldest temperatures of the Younger Dryas. Details of the chironomid stratigraphy are shown in Figure 3. The early Interstadial chironomid fauna was dominated by cold stenothermic taxa such as *Diaesinae*, *Heterotrissocladius grimshawi*, *T. lugens*-type, *Micropsectra radialis*, *S. coracina*, and *Arctopelopia* but also included more temperate taxa such as *Cricotopus*, *Psectrocladius sordidellus*-type, and *Microtendipes*. As the Interstadial climate continued to warm, the proportion of temperate taxa increased at the expense of the cold stenotherms, and *Diaesinae* disappeared from the fauna altogether. The onset of the Younger Dryas caused a sudden and major change in the chironomid fauna. The temperate taxa and some of the cold stenothermic taxa, such as *Sergentia*, *T. lugens*-type, and *Arctopelopia* were eliminated. *M. radialis* and *H. grimshawi* continued to dominate the fauna, *Diaesinae* returned, but several cold stenothermic taxa appeared in the fauna for the first time, including *H. subpilosus* and *Paracladius alpicola*. These faunal changes indicate that the lake went from cool, mesotrophic to very cold and ultraoligotrophic conditions.

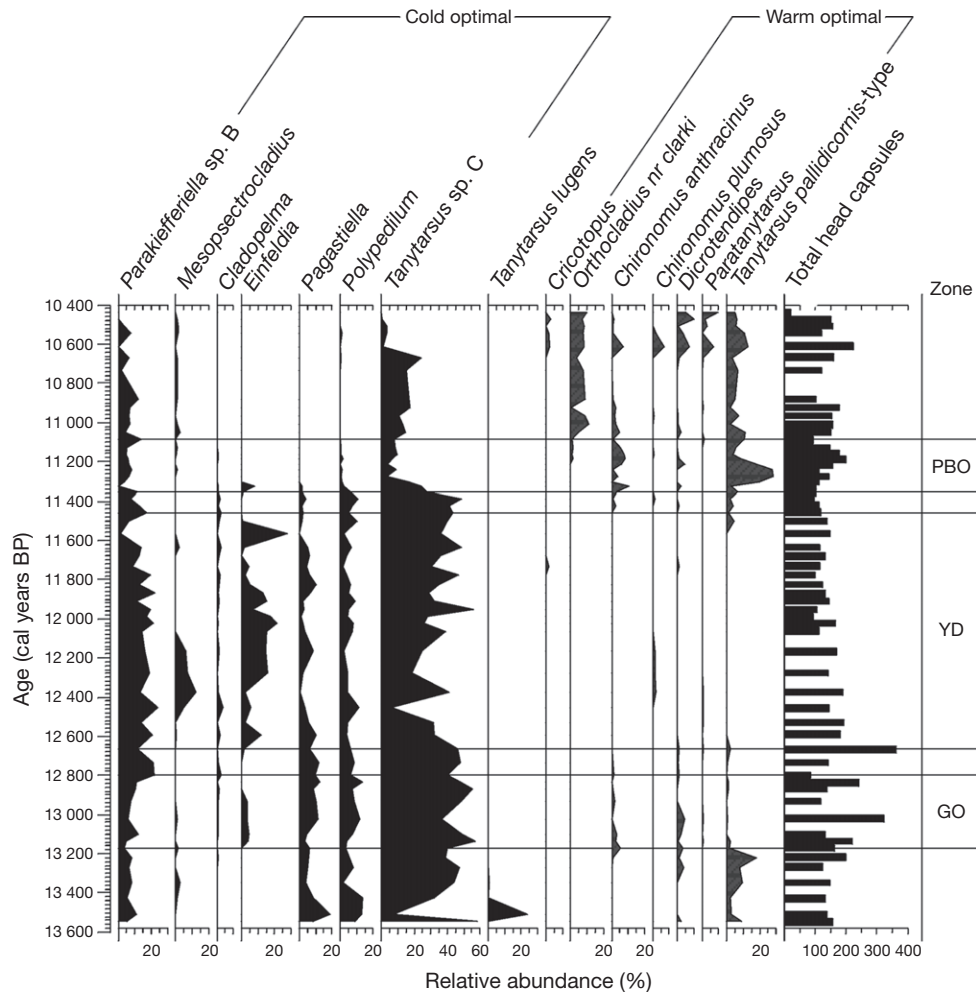


Figure 2 Response of Chironomidae to climate change around the time of the Younger Dryas at Leysin, Switzerland. GO, Gerzensee Oscillation; PBO, Preboreal Oscillation; YD, Younger Dryas. Reproduced from Brooks SJ (2000) Late-glacial fossil midge stratigraphies (Insecta: Diptera: Chironomidae) from the Swiss Alps. *Palaeogeography, Palaeoclimatology, Palaeoecology* 159: 261–279.

Mean July air temperatures of 9–10 °C were inferred from the chironomid assemblage for the Younger Dryas which were about 5 °C higher than those inferred from plant macrofossils. This indicated that the range of temperatures covered by the chironomid training set was not sufficient to reconstruct reliable temperatures for the late-glacial period. Therefore, an additional 65 lakes were added, which extended the range of temperatures in the modern training set to 3.5–15.6 °C, and in addition provided a more even distribution of lakes across the temperature gradient (Brooks and Birks, 2001). Using the expanded training set, new temperature estimates for the Late Glacial and early Holocene were obtained that closely supported the temperatures indicated by the plant macrofossils (Figure 4).

Gradually warming temperatures from about 9.5 to 10.5 °C are indicated during the Interstadial, followed by a sudden drop of about 3.5 °C over a 60-year period at the start of the Younger Dryas. This fall is still not as great as that inferred from vegetational changes which suggest that temperatures fell by as much as 6 to 4–5 °C (Birks, 1994). Chironomid-inferred temperatures began to increase before the end of the Younger

Dryas, shortly after 12 000 cal years BP, and continued to rise rapidly into the early Holocene.

The 109-lake Norwegian training set was also used to reconstruct late-glacial temperatures at Whitrig Bog in southeast Scotland (Brooks and Birks, 2000b; 2001). High resolution analysis of the core meant that a highly detailed picture of climatic fluctuations emerged. Following the deglacial period when the chironomid assemblage was dominated by cold stenotherms, the Interstadial fauna was dominated for the most part by temperate taxa such as *Psectrocladius*, *Chironomus*, *Tanytarsus* spp., and *Pentaneurini* (Brooks et al., 1997). However, there were two relatively short periods when there was a resurgence of cold-water taxa. In the earliest period, eight cold taxa were present and there were declines in most of the temperate taxa, including *Chironomus* (Figure 5). During the second cool period only three cold-water taxa were present and, although several temperate taxa declined, *Chironomus* did not decline in abundance. At the onset of the Younger Dryas there was a sudden replacement of the entire temperate fauna by a cold water fauna, dominated by *M. radialis*-type, which had not previously occurred in the lake, and several

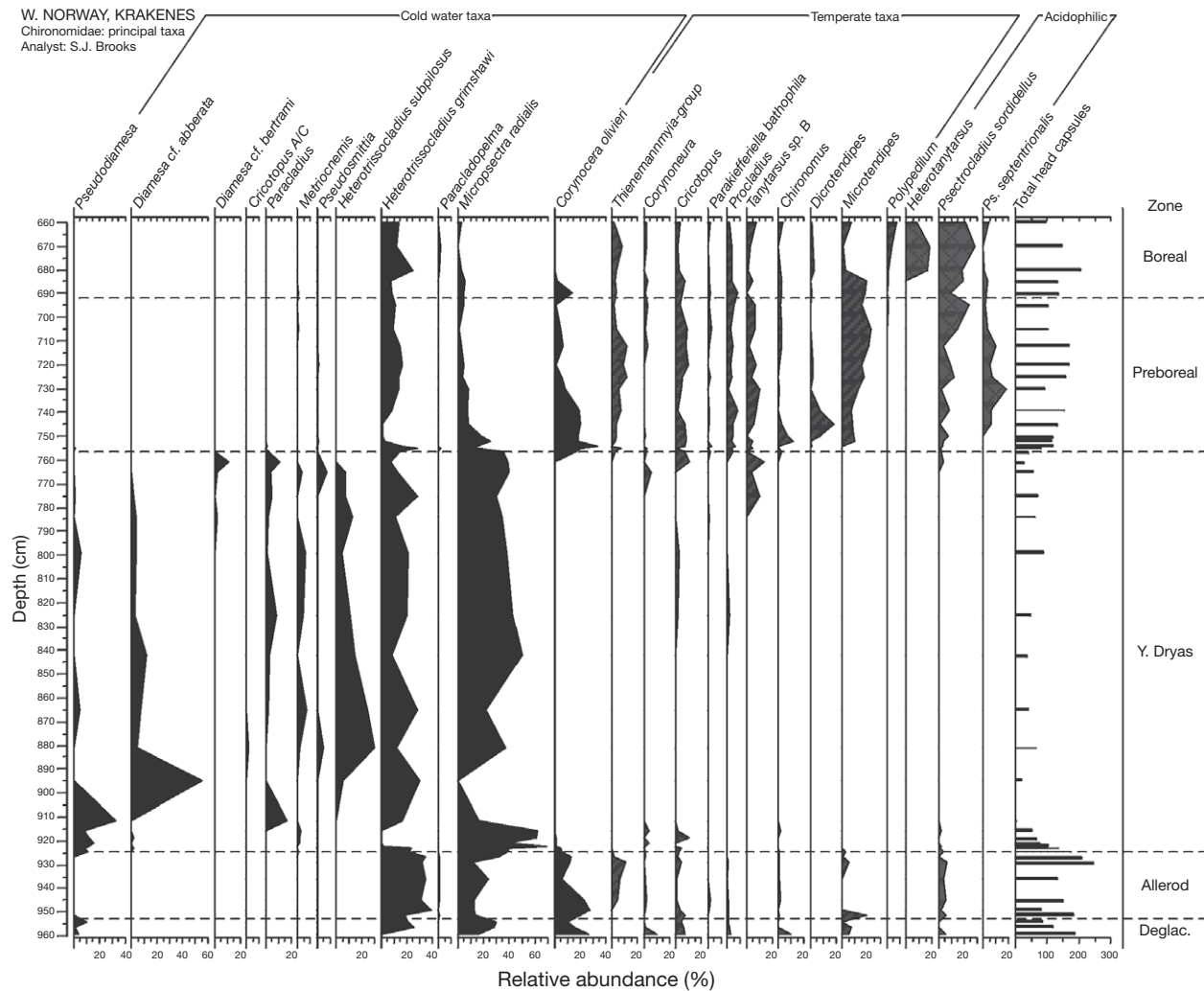


Figure 3 Response of Chironomidae to climate change during the Late Glacial and early Holocene at Kråkenes, Norway. Reproduced from Brooks SJ (1997) The response of Chironomidae (Insecta: Diptera) assemblages to late-glacial climatic change in Kråkenes Lake, western Norway. *Quaternary Proceedings* 5: 49–58.

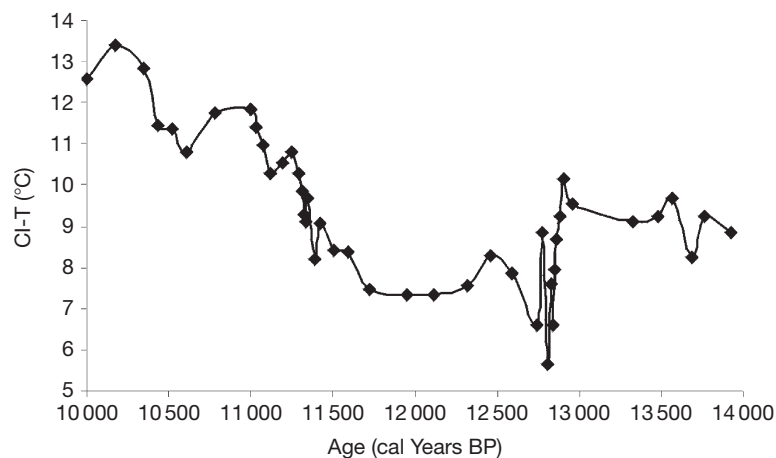


Figure 4 Chironomidae-inferred mean July air temperature (CI-T) reconstruction of the Late Glacial and early Holocene at Kråkenes, Norway. Reproduced from Brooks SJ and Birks HJB (2001) Chironomid-inferred air temperatures from late-glacial and Holocene sites in north-west Europe: Progress and problems. *Quaternary Science Reviews* 20: 1723–1741.

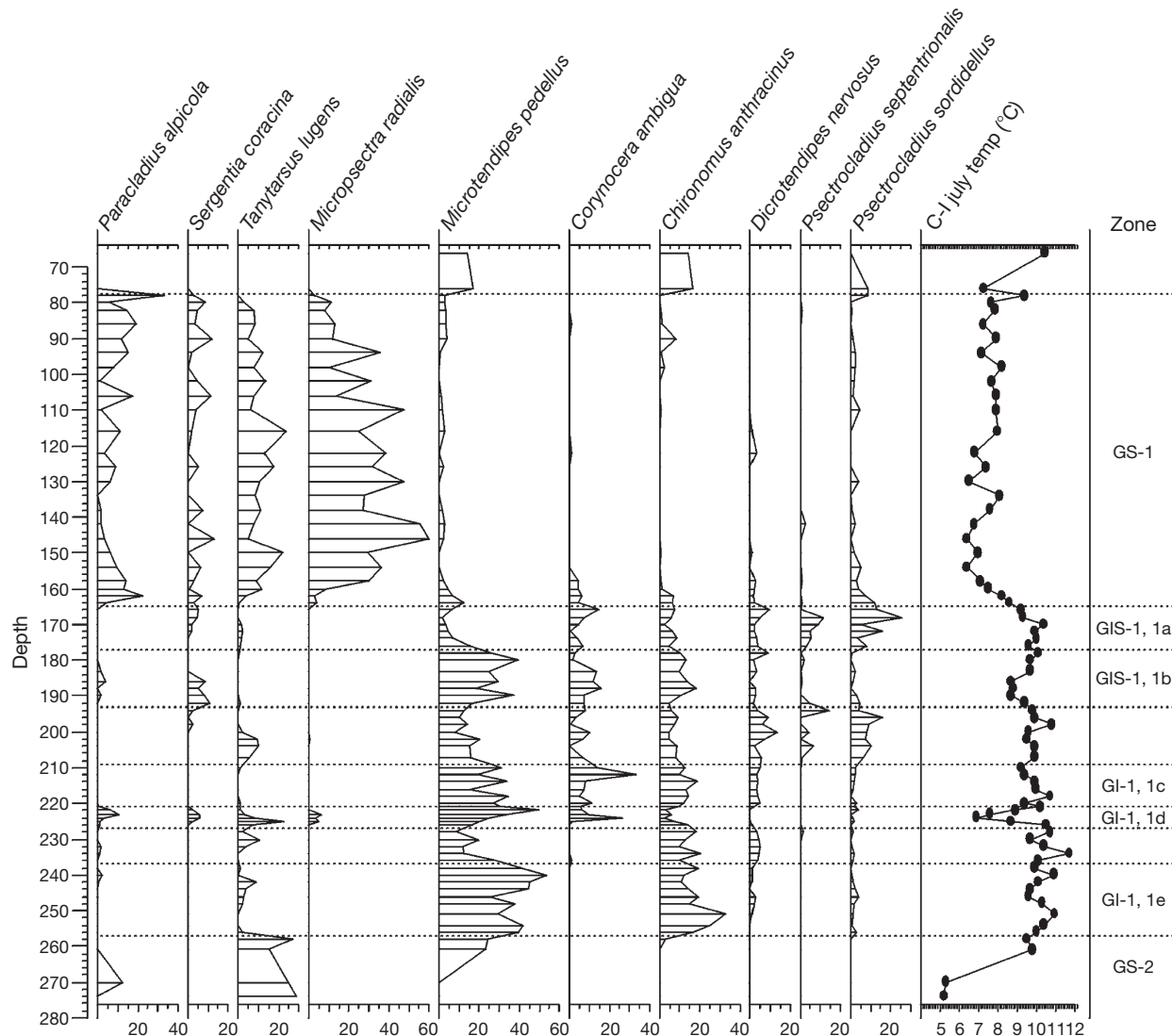


Figure 5 Response of Chironomidae to climate change during the Late Glacial and early Holocene at Whitrig Bog, southeast Scotland. Reproduced from Brooks SJ and Birks HJB (2000) Chironomid-inferred Late-glacial air temperatures at Whitrig Bog, southeast Scotland. *Journal of Quaternary Science* 15: 759–764.

other taxa which had only occurred at low abundance during the previous cold periods. These faunistic changes implied that the Younger Dryas was the coldest period in the Late Glacial, that temperatures during the early cool oscillation were a little less cold, but that they were considerably colder than the second cool oscillation.

These qualitative inferences were confirmed when the 109-lake Norwegian transfer function was applied to the data. Interstadial mean July air temperatures gradually fell from ~ 11.5 to 11.0 °C but were punctuated by two major cold oscillations. During the first oscillation, temperatures fell to ~ 8.0 °C. This oscillation is referable to Greenland Interstadial event 1d (GI-1d) which has been recorded in northern Switzerland as the Aegelsee Oscillation (Lotter et al., 1992) and elsewhere in Europe (Björck et al., 1998). The second oscillation, when temperatures fell to ~ 10.0 °C, is referable to Greenland Interstadial event 1b (GI-1b) and has

also been recorded elsewhere in Europe as the Gerzensee or Amphi-Atlantic Oscillation (Bedford et al., 2004; Levesque et al., 1993; Lotter et al., 1992). Early in the Younger Dryas, temperatures fell to ~ 7.5 °C but gradually increased to ~ 8.5 °C at the end of this period. These trends are closely reflected in the GRIP ice core oxygen isotope record (Figure 6), in particular, the gradual warming during the Younger Dryas, the two major cold oscillations in the Interstadial, together with the suggestion of two additional minor Interstadial oscillations just before and just after GI-1d. However, there are two obvious differences between the records. First, the cooling trend in Interstadial temperatures is more pronounced in the ice core record than the Whitrig Bog record. Second, GI-1d is warmer than GI-1b in the GRIP record, whereas at Whitrig Bog the reverse is indicated. The chironomid-inferred temperature estimates are closely corroborated by temperature estimates derived from plant macrofossils

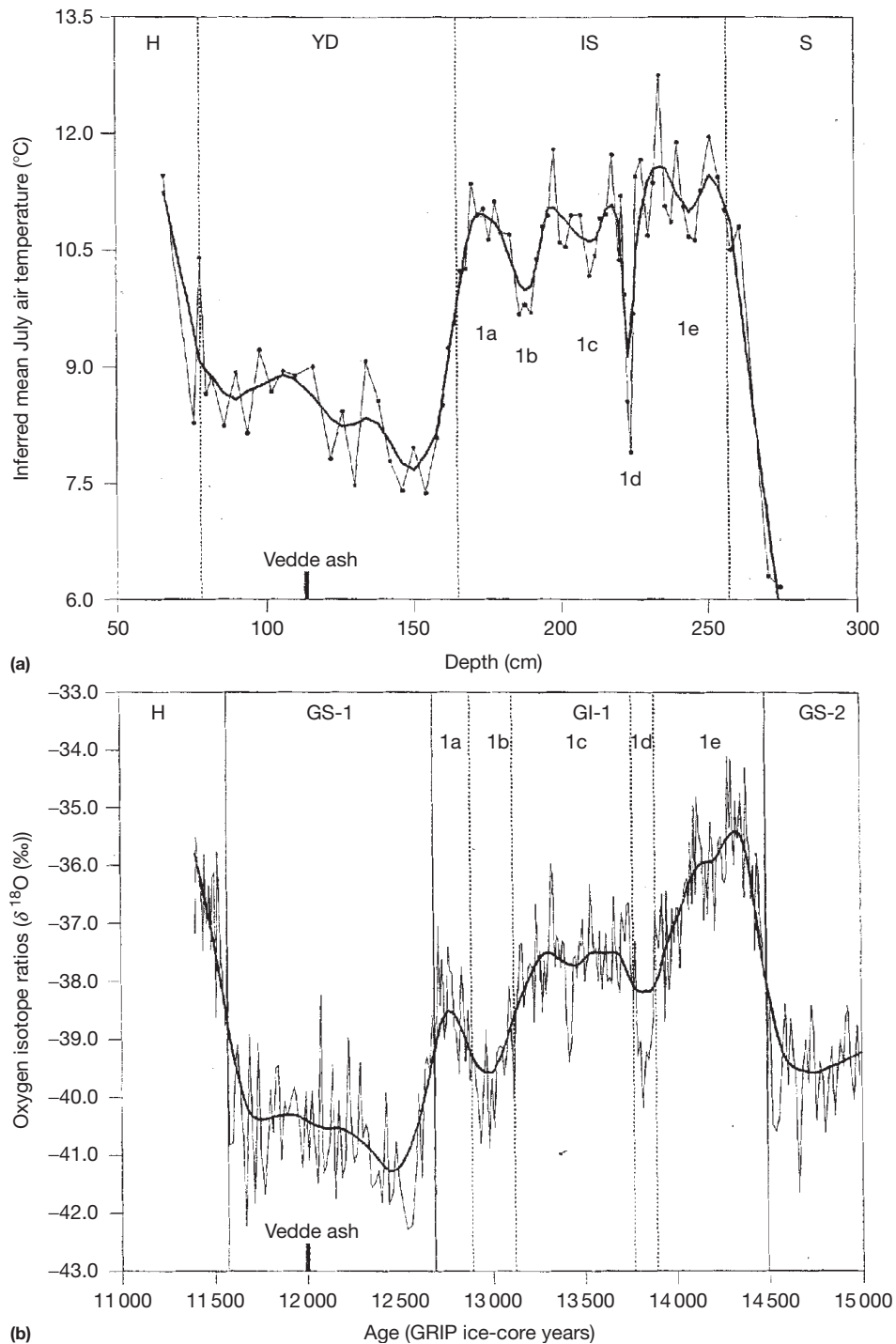


Figure 6 Comparison of (a) chironomid-inferred mean July air temperatures for the Late Glacial at Whitrig Bog, southeast Scotland and (b) the oxygen isotope record from the Greenland (GRIP) ice core (Johnsen et al., 1992). Reproduced from Brooks SJ and Birks HJB (2000) Chironomid-inferred Late-glacial air temperatures at Whitrig Bog, southeast Scotland. *Journal of Quaternary Science* 15: 759–764.

(Birks, unpublished; Conolly, 1961; Sheldrick, 1997), although warmer than those derived from glacial data (Ballantyne, 1989; Sissons, 1979; Sutherland, 1984) for lowland Scotland. However, late-glacial beetle-inferred estimates for southern Scotland are consistently about 2 °C higher than the

chironomid estimates (Atkinson et al., 1987; Coope and Lemdahl, 1995; Coope et al., 1998). Reasons for this discrepancy are not obvious but may be related to differences in the response of aquatic and terrestrial insects to temperature or in the numerical approaches used to derive temperatures.

A second British quantitative reconstruction of late-glacial temperatures based on chironomids is available from Hawes Water in northwest England (Bedford et al., 2004). The chironomid assemblage (Figure 7) is similar to that of Whitrig Bog, even though Hawes Water is a hard water lake whereas Whitrig Bog has soft water. The Interstadial is once more dominated by temperate taxa such as *Microtendipes*, *Chironomus*, *Psectrocladius*, and *Pentaneurini*. However, there are two periods during the Interstadial in which the cold stenotherm, *Sergentia*, increases in abundance. The Younger Dryas is dominated by a cold water fauna very similar in composition to that at Whitrig Bog. The main difference between the two assemblages is that *C. ambigua* is most abundant during the Younger Dryas at Hawes Water, whereas at Whitrig Bog this taxon reaches maximum abundance during the Allerød. This may reflect in-lake differences between the sites since *C. ambigua* is often associated with charophyte beds (Brodersen and Lindegaard, 1999). The Norwegian training set was used to reconstruct chironomid-inferred temperatures. Interstadial mean air July temperatures gradually fell from ~13.0 to 12.0 °C but were punctuated by two abrupt cold oscillations. The first, when temperatures fell by about 2 to ~11.0 °C was more severe than the second when temperatures fell by about 1 to ~11.5 °C. Temperatures fell rapidly at the start of the Younger Dryas to ~7.5 °C but then gradually increased to ~9.5 °C at the end of this period. The chironomid-inferred temperature trend at Hawes Water therefore closely reflects that inferred at Whitrig Bog. The main difference is that temperatures are consistently warmer at Hawes Water than Whitrig Bog, but this probably reflects the more southerly location of the former site. An oxygen isotope record is also available from Hawes Water (Figure 8) and this closely matches the chironomid-inferred trends. Notably, the Hawes Water isotope record, like the chironomid-inferred temperature records, indicates that the Aegelsee Oscillation was colder than the Gerzensee Oscillation and that there was only a gradual long-term cooling trend during the Interstadial. These trends are also similar to the oxygen isotope record from Ammersee in southern Germany (Von Graffenstein et al., 1999) but differ from the GRIP record.

Additional late-glacial studies have recently been published from England (Lang et al., 2010) and Ireland (Watson et al., 2010), and others from Scotland and Norway are currently in progress by Brooks and coworkers which will be published over the next few years. All these records show similar trends as those described above and further support the view that there are important and consistent differences in detail between the Greenland oxygen isotope records and chironomid-inferred temperature records and isotope records from Europe.

Following a detailed account by Millet et al. (2003) of the late-glacial chironomid fauna at Lac Lautrey (Jura, eastern France), the Swiss chironomid-temperature inference model (Heiri, 2001; Heiri et al., 2003) was used to reconstruct Late Glacial mean July air temperatures at this site by Heiri and Millet (2005). This high resolution record (Figure 9) begins at about 16000 cal years BP when the fauna was dominated first by the cold stenotherm *Stictochironomus* and then by *C. ambigua* and mean July air temperatures were inferred at about 11–12 °C (Figure 10). There was an abrupt rise in

temperature to 14.5 °C at the beginning of the Bølling (14600 cal years BP) when a diverse fauna of widely distributed and lowland taxa replaced the alpine and subalpine fauna. During the second half of the Bølling (14200–13700 cal years BP), temperatures continued to rise more gradually to reach 16.5 °C as temperate taxa increased in abundance. These estimates are within the range of pollen-based reconstructions from central Europe (Renssen and Isarin, 2001). There was no apparent response to the Aegelsee cool fluctuation (at ca. 14000 cal years BP), however, two centennial-scale cold oscillations of 1.5–2.0 °C were indicated at 13200 and 12900 cal years BP which appear to be synchronous with the beginning and end of the Gerzensee Oscillation. Interstadial mean July air temperatures peaked at about 18 °C toward the end of the Allerød, which is warmer than present-day values from northern Switzerland and considerably warmer than late Allerød temperatures inferred from Cladocera and pollen assemblages (Lotter et al., 2000).

Mean July air temperatures declined to 14 °C at the start of the Younger Dryas but then gradually increased and were about 1 °C higher by the end of the Younger Dryas, which reflects the trends in other European records (Brooks and Birks, 2000b; Von Graffenstein et al., 1999). When corrected for altitude, the Lac Lautrey Younger Dryas estimates agree closely with chironomid-inferred records from the northern Swiss Alps (Heiri et al., 2003) but are warmer than pollen-inferred estimates (Lotter et al., 2000). A warming of about 1.5 °C was indicated at the Younger Dryas/Holocene transition, which is in close agreement with other proxy records from the region (Heiri et al., 2003; Lotter et al., 2000; Magny et al., 2001). The Lac Lautrey temperature record differs in detail from the GRIP oxygen isotope record but shares similarities with the Ammersee isotope record from southern Germany (Von Graffenstein et al., 1999). Most notably, chironomid-inferred temperatures before the Bølling warming were cooler than those during the Younger Dryas, there is a two-step increase in early Bølling temperatures, the GI-1d event is not clearly reflected in the records and there is not a strong temperature decline during the Interstadial.

Heiri et al. (2007) reported a late-glacial chironomid record from Hijkermeer, The Netherlands. Both the GI-1d and GI-1b cold oscillations are apparent in the Hijkermeer sequence, however, the temperature decline is greater during GI-1b than GI-1d, which is the opposite of the trends in the British records. The Hijkermeer record also indicates that peak Interstadial temperatures were reached after the GI-1d oscillation, at around 13.5 ka BP, which also reflects the results found at Lac Lautrey and in the Ammersee oxygen isotope record. This is later than the British records in which peak temperatures are reached early in the Interstadial, before the GI-1d oscillation. Similar trends to those at Hijkermeer are also apparent in two further records from central Switzerland, one from the Maloja Pass (Ilyashuk et al., 2009) and the other from Egelsee (Larocque-Tobler et al., 2010). However, a record from Lago Piccolo di Avigliana in the southwest Alps (Italy; Larocque and Finsinger, 2008) indicates that the warmest Interstadial temperatures occurred prior to the GI-1d cold oscillation. These records suggest that there were differences in the spatial patterns of late-glacial temperature trends across Europe, as inferred from fossil beetle records from this interval (Coope

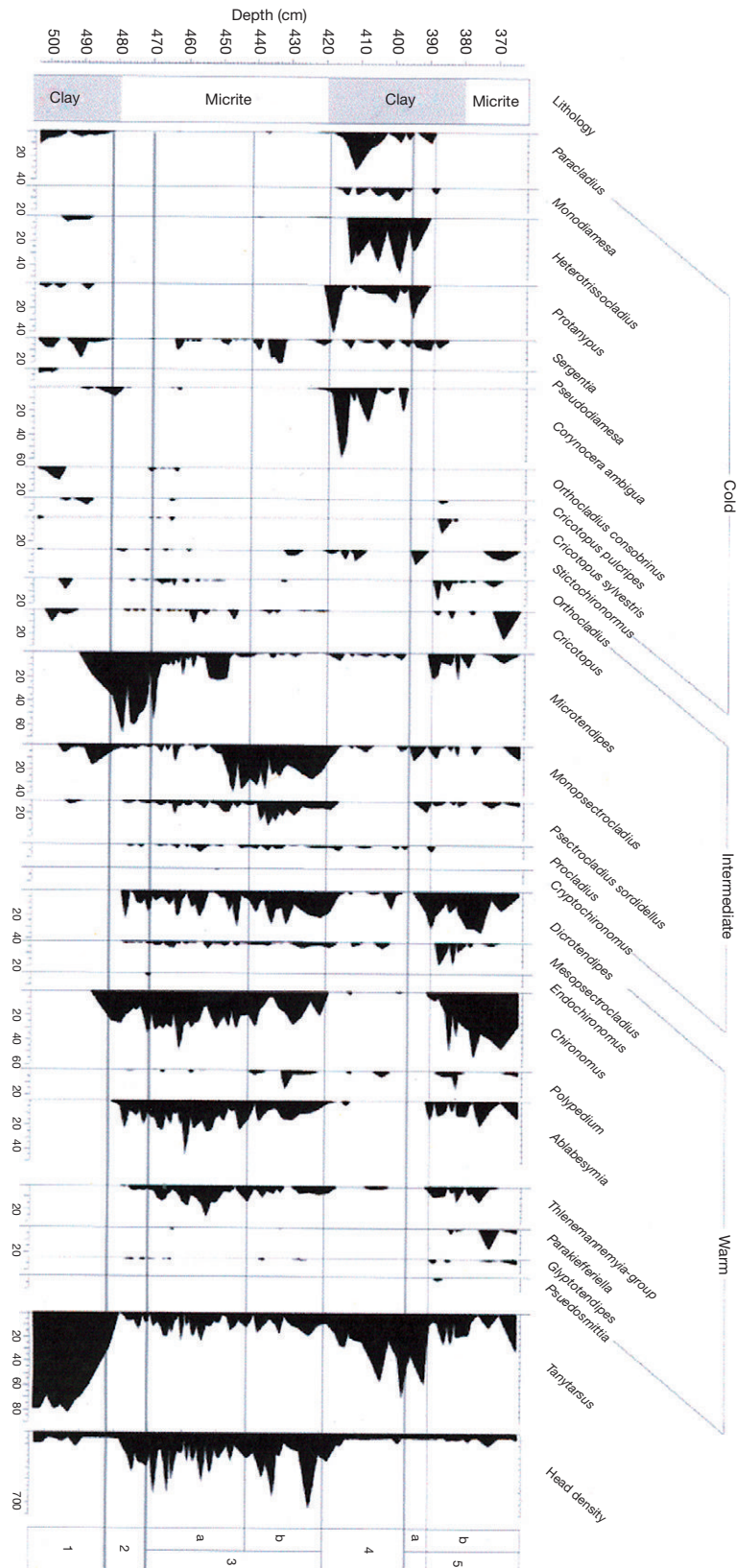


Figure 7 Response of Chironomidae to climate change during the Late Glacial and early Holocene at Hawes Water, northwest England. Reproduced from Bedford A, Jones RT, Lang B, Brooks SJ, and Marshall JD (2004) A late-glacial chironomid record from Hawes Water, N.W. England. *Journal of Quaternary Science* 19: 281–290.

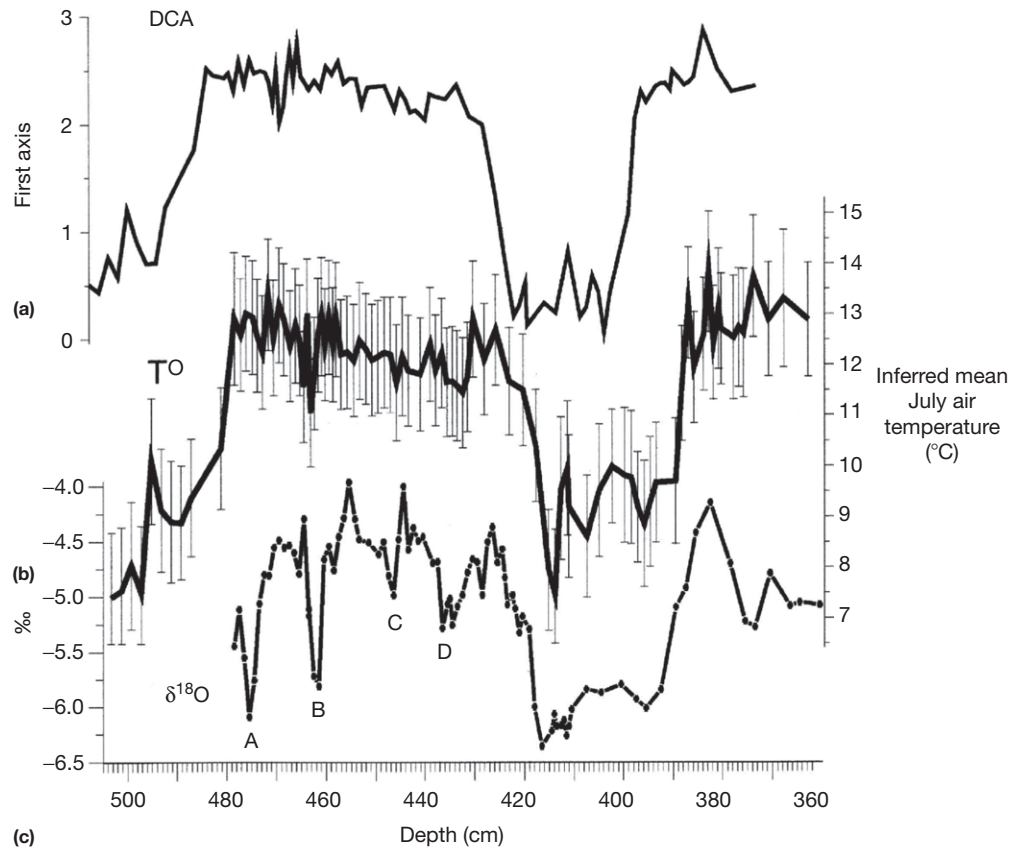


Figure 8 Comparison of chironomid-inferred mean July air temperatures for the Late Glacial at Hawes Water, northwest England, the detrended correspondence analysis (DCA) axis 1 sample scores and the oxygen isotope record from Ammersee, southern Germany (Von Graffenstein et al., 1999). Reproduced from Bedford A, Jones RT, Lang B, Brooks SJ, and Marshall JD (2004) A late-glacial chironomid record from Hawes Water, N.W. England. *Journal of Quaternary Science* 19: 281–290.

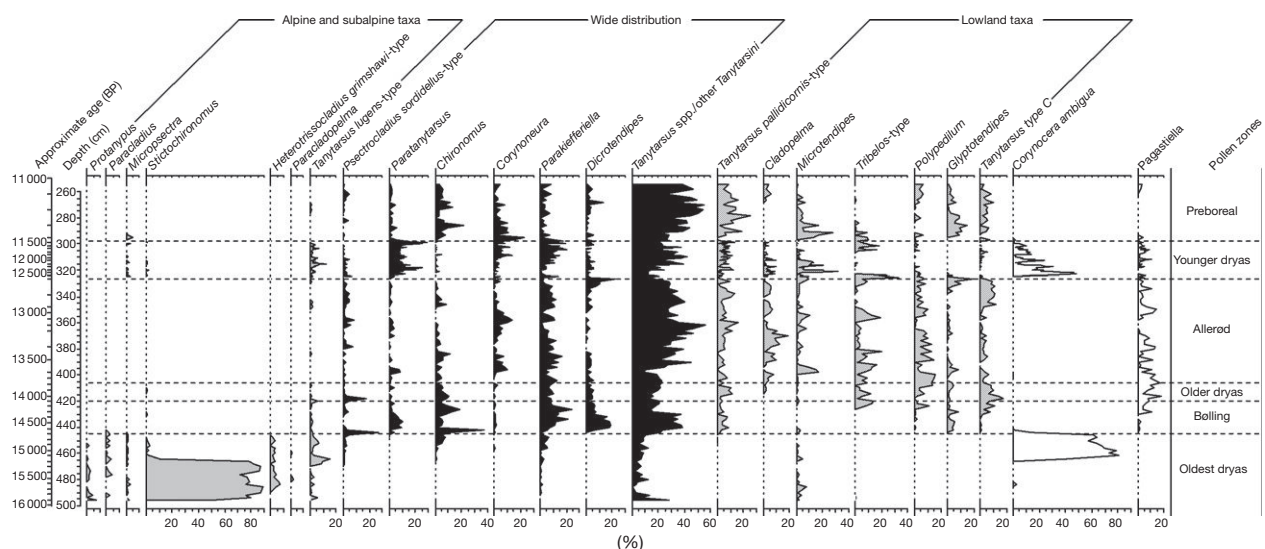


Figure 9 Response of Chironomidae to climate change during the Late Glacial and early Holocene at Lac Lautrey, Jura, eastern France. Reproduced from Heiri O and Millet L (2005) Reconstruction of Late Glacial summer temperatures from chironomid assemblages in Lac Lautrey (Jura, France). *Journal of Quaternary Science* 20: 33–44.

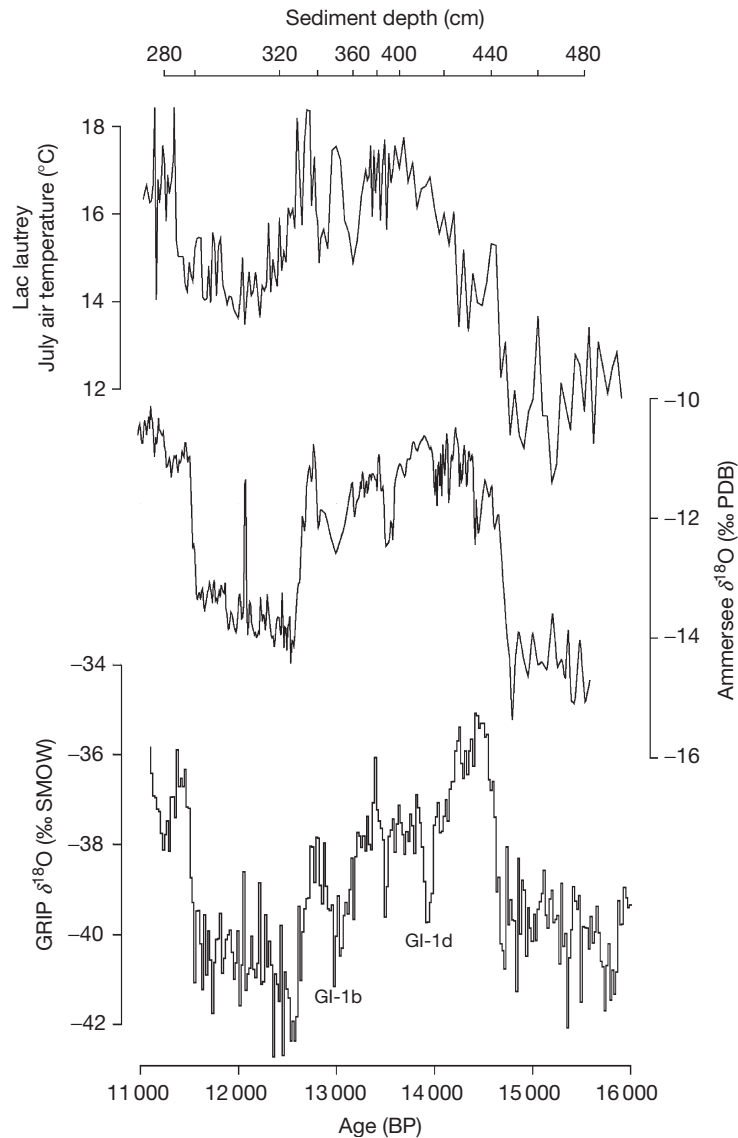


Figure 10 Comparison of chironomid-inferred mean July air temperatures for the Late Glacial at Lac Lautrey, Jura, eastern France, the oxygen isotope record from Ammersee, southern Germany (Von Graffenstein et al., 1999) and the oxygen isotope record from the Greenland (GRIP) ice core (Johnsen et al., 1992). Reproduced from Heiri O and Millet L (2005) Reconstruction of Late Glacial summer temperatures from chironomid assemblages in Lac Lautrey (Jura, France). *Journal of Quaternary Science* 20: 33–44.

et al., 1998), although more data is required to clarify the apparent discrepancies between them and confirm whether or not there are consistent regional patterns.

Conclusion

At present relatively few quantitative chironomid-inferred temperature reconstructions are available from the Late Glacial in Europe. However, it is clear from those that have been published that such sequences can provide valuable high resolution, low error records of climate change during this period. The chironomid-inferred temperature trends largely reflect those derived from oxygen isotope records, which suggests that the chironomid records are reliable, and so provide

quantitative details of short-lived, small amplitude climate fluctuations as well as long term trends. However, differences between the GRIP oxygen isotope record and the isotope records from Hawes Water and Ammersee are also reflected in the chironomid-inferred temperature records from Whitrig Bog, Hawes Water, Hijkermeer, and Lac Lautrey, which more closely resemble the local isotope trends than the GRIP record. This suggests that climate change was not uniform throughout the North Atlantic region and central Europe and that it may be simplistic to rely on the GRIP ice core record to provide an 'event stratigraphy' for the whole region (Björck et al., 1998). Similarly, late-glacial chironomid-inferred temperature estimates are largely corroborated by estimates derived from other proxies. Where conflicts do exist there are currently insufficient records to resolve them. Therefore, it is particularly

important that more chironomid-derived climate records become available, which will help to underline consistencies and inconsistencies between them and other proxy records. This will be necessary to improve our understanding of the limitations of the different proxy records. In particular, discrepancies between chironomid- and beetle-inferred estimates need to be resolved. However, the current lack of same site quantitative reconstructions that include both beetles and chironomids impedes the resolution of this problem.

See also: **Beetle Records:** Postglacial Europe; Late Pleistocene of Europe. **Chironomid Records:** Africa; Chironomid Overview; Postglacial Europe; Postglacial Southern Hemisphere. **Ice Core Records:** Greenland Stable Isotopes.

References

- Andersen FS (1938) Spätglaziale Chironomiden. *Meddelelser Dansk Geologisk Forening* 9: 320–326.
- Atkinson TC, Briffa KR, and Coope GR (1987) Seasonal temperatures in Britain during the past 22,000 years reconstructed using beetle remains. *Nature* 352: 587–592.
- Ballantyne CK (1989) The Loch Lomond readvance on the Isle of Skye, Scotland: Glacier reconstruction and palaeoclimatic implications. *Journal of Quaternary Science* 4: 95–108.
- Bedford A, Jones RT, Lang B, Brooks SJ, and Marshall JD (2004) A late-glacial chironomid record from Hawes Water, N.W. England. *Journal of Quaternary Science* 19: 281–290.
- Birks HH (1994) Late-glacial vegetational ecotones and climatic patterns in western Norway. *Vegetation History and Archaeobotany* 3: 107–119.
- Björck S, Walker MJC, Cwynar LC, et al. (1998) An event stratigraphy for the last termination in the north Atlantic region based on the Greenland ice-core record: A proposal by the INTIMATE group. *Journal of Quaternary Science* 13: 283–292.
- Brodersen KP and Lindegaard C (1999) Mass occurrence and sporadic distribution of *Corynocera ambigua* Zetterstedt (Diptera, Chironomidae) in Danish lakes. Neo- and palaeolimnological records. *Journal of Paleolimnology* 22: 41–52.
- Brooks SJ (2000) Late-glacial fossil midge stratigraphies (Insecta: Diptera: Chironomidae) from the Swiss Alps. *Palaeogeography, Palaeoclimatology, Palaeoecology* 159: 261–279.
- Brooks SJ and Birks HJB (2000a) Chironomid-inferred late-glacial and early-Holocene mean July air temperatures for Kråkenes Lake, western Norway. *Journal of Paleolimnology* 23: 77–89.
- Brooks SJ and Birks HJB (2000b) Chironomid-inferred Late-glacial air temperatures at Whitrig Bog, southeast Scotland. *Journal of Quaternary Science* 15: 759–764.
- Brooks SJ and Birks HJB (2001) Chironomid-inferred air temperatures from late-glacial and Holocene sites in north-west Europe: Progress and problems. *Quaternary Science Reviews* 20: 1723–1741.
- Brooks SJ, Lowe JJ, and Mayle FE (1997) The Late Devensian lateglacial palaeoenvironmental record from Whitrig Bog, S.E. Scotland. 2. Chironomidae (Insecta: Diptera). *Boreas* 26: 297–308.
- Conolly AP (1961) Some climatic and edaphic indications from the late-glacial flora. *Proceedings of the Linnean Society of London* 172: 56–62.
- Coope GR and Lemdahl G (1995) Regional differences in the Lateglacial climate of northern Europe based on coleopteran analysis. *Journal of Quaternary Science* 10: 391–395.
- Coope GR, Lemdahl G, Lowe JJ, and Walkling A (1998) Temperature gradients in northern Europe during the last glacial-interglacial transition (14–9 ¹⁴C kyr BP) interpreted from coleopteran assemblages. *Journal of Quaternary Science* 13: 419–433.
- Gams H (1927) Die Gessichte der Lunzer Seen, Moore und Wälder. *Internationale Revue der Gesamten Hydrobiologie und Hydrographie* 18: 305–387.
- Gandouin E and Franquet E (2002) Late glacial and Holocene chironomid assemblages in 'Lac Long Inférieur' (southern France, 2090 m): Palaeoenvironmental and palaeoclimatic implications. *Journal of Paleolimnology* 28: 317–328.
- Guilizzoni P, Marchetto A, Lami A, et al. (2000) Evidence for short-lived oscillations in the biological records from the sediments of Lago Albano (Central Italy) spanning the period ca. 28 to 17 k yr BP. *Journal of Paleolimnology* 23: 117–127.
- Heiri O (2001) *Holocene paleolimnology of Swiss mountain lakes reconstructed using subfossil chironomid remains: Past climate and prehistoric human impact on lake ecosystems*. PhD Thesis, University of Bern.
- Heiri O, Cremer H, Engels S, Hoek WZ, Peeters W, and Lotter AF (2007) Lateglacial summer temperatures in the Northwest European lowlands: A chironomid record from Hijkermeer, the Netherlands. *Quaternary Science Reviews* 26: 2420–2437.
- Heiri O, Lotter AF, Hausmann S, and Kienst F (2003) A chironomid-based Holocene summer air temperature reconstruction from the Swiss Alps. *The Holocene* 13: 477–484.
- Heiri O and Millet L (2005) Reconstruction of late glacial summer temperatures from chironomid assemblages in Lac Lautrey (Jura, France). *Journal of Quaternary Science* 20: 33–44.
- Hofmann W (1971) Die postglaziale Entwicklung der Chironomiden- und *Chaoborus*-Fauna (Dipt.) des Schönhsees. *Archiv für Hydrobiologie Supplement* 40: 1–74.
- Hofmann W (1978) Analysis of animal microfossils from the Großer Segeberger See (F.R.G.). *Archiv für Hydrobiologie* 82: 316–346.
- Hofmann W (1983a) Stratigraphy of subfossil Chironomidae and Ceratopogonidae (Insecta: Diptera) in late glacial littoral sediments from Lobsigensee (Swiss Plateau). Studies in the Late Quaternary of Lobsigensee 4. *Revue de Paléobiologie* 2: 205–209.
- Hofmann W (1983b) Stratigraphy of Cladocera and Chironomidae in a core from a shallow North German lake. *Hydrobiologia* 103: 235–239.
- Hofmann W (1984) Stratigraphie subfossiler Cladocera (Crustacea) und Chironomidae (Diptera) in zwei Sedimentprofilen des Meerfelder Maars. *Courier Forschungs Institut Senckenberg* 65: 67–80.
- Hofmann W (1990) Weichselian chironomid and cladoceran assemblages from maar lakes. *Hydrobiologia* 214: 207–211.
- Hofmann W (1991) Stratigraphy of Chironomidae (Insecta: Diptera) and Cladocera (Crustacea) in Holocene and Würm sediments from Lac du Bouchet (Haute Loire, France). *Documents du C.E.R.L.A.T.* 2: 363–386.
- Ilyashuk B, Gobet E, Heiri O, et al. (2009) Lateglacial environmental and climatic changes at the Maloja Pass, Central Swiss Alps, as recorded by chironomids and pollen. *Quaternary Science Reviews* 28: 1340–1353.
- Johnsen SJ, Hammer CU, Iversen P, et al. (1992) Irregular glacial interstadials recorded in a new Greenland ice core. *Nature* 359: 311–313.
- Lang B, Brooks SJ, Bedford A, Jones RT, Birks HJB, and Marshall J (2010) Regional consistency in Lateglacial chironomid-inferred temperatures from five sites in north-west England. *Quaternary Science Reviews* 29: 1528–1538.
- Larocque I and Finsinger W (2008) Late-glacial chironomid-based temperature reconstructions for Lago Piccolo di Avigliana in the southwestern Alps (Italy). *Palaeogeography, Palaeoclimatology, Palaeoecology* 257: 207–223.
- Larocque-Tobler I, Heiri O, and Wehrli M (2010) Late glacial and Holocene temperature changes at Egelsee, Switzerland, reconstructed using subfossil chironomids. *Journal of Paleolimnology* 43: 649–666.
- Levesque AJ, Mayle FE, Walker IR, and Cwynar LC (1993) The Amphi-Atlantic Oscillation: A proposed late-glacial climatic event. *Quaternary Science Reviews* 12: 629–643.
- Lotter AF, Birks HJB, Eicher U, Hofmann W, Schwander J, and Wick L (2000) Younger Dryas and Allerød summer temperatures at Gerzensee (Switzerland) inferred from fossil pollen and cladoceran assemblages. *Palaeogeography, Palaeoclimatology, Palaeoecology* 159: 349–361.
- Lotter AF, Eicher U, Birks HJB, and Siegenthaler U (1992) Late-glacial climatic oscillations as recorded in Swiss lake sediments. *Journal of Quaternary Science* 7: 187–204.
- Magny M, Guiot J, and Scoellammer P (2001) Quantitative reconstruction of Younger Dryas to mid-Holocene paleoclimates at Le Locle, Swiss Jura, using pollen and lake-level data. *Quaternary Research* 56: 170–180.
- Manca M, Nocentini AM, Belis CA, Comoli P, and Corbella L (1996) Invertebrate fossil remains as indicators of late Quaternary environmental changes in Latium crater lakes (L. Albano and L. Nemi). *Memorie dell'Istituto Italiano Idrobiologia* 55: 149–176.
- Millet L, Verneaux V, and Magny M (2003) Lateglacial paleoenvironmental reconstruction using subfossil chironomid assemblages from Lake Lautrey (Jura, France). *Archiv für Hydrobiologie* 156: 405–429.
- Renssen H and Isarin RFB (2001) The two major warming phases of the last deglaciation at ~14.7 and ~11.5 ka cal. BP in Europe: Climate reconstruction and AGCM experiments. *Global and Planetary Change* 30: 117–153.

- Sheldrick C (1997) *Plant macrofossil records from UK lake sediments spanning the last glacial–interglacial transition, ca. 14–9 ^{14}C ka BP*. PhD Thesis, University of London.
- Sissons JB (1979) Palaeoclimatic inferences from former glaciers in Scotland and the Lake District. *Nature* 278: 518–521.
- Sutherland DG (1984) Modern glacier characteristics as a basis for inferring former climates with particular reference to the Loch Lomond stadial. *Quaternary Science Reviews* 3: 291–309.
- Verbruggen F, Heiri O, Reichert G-J, de Leeuw JW, Nierop KGJ, and Lotter AF (2010a) Effects of chemical pretreatments on $\delta^{18}\text{O}$ measurements, chemical composition, and morphology of chironomid head capsules. *Journal of Paleolimnology* 43: 857–872.
- Verbruggen F, Heiri O, Reichert GJ, and Lotter AF (2010b) Chironomid delta O-18 as a proxy for past lake water delta O-18: A Lateglacial record from Rotsee (Switzerland). *Quaternary Science Reviews* 29: 2271–2279.
- Von Graffenstein U, Erlenkeuser H, Brauer A, Jouzel J, and Johnsen S (1999) A mid-European decadal isotope climate record from 15,500 to 5000 years BP. *Science* 284: 1654–1657.
- Walker IR, Levesque AJ, Pienitz R, and Smol JP (2003) Freshwater midges of the Yukon and adjacent Northwest Territories: A new tool for reconstructing Beringian paleoenvironments? *Journal of the North American Benthological Society* 22: 323–337.
- Watson J, Brooks SJ, Whitehouse NJ, Reimer PJ, Birks HJB, and Turney C (2010) Chironomid-inferred Late-glacial summer air temperatures from Lough Nadourcan, Co. Donegal, Ireland. *Journal of Quaternary Science* 25: 1200–1210. <http://dx.doi.org/10.1002/jqs.1399>.
- Woolfer MJ, Francis D, Fogel ML, Miller GH, Walker IR, and Woolfe AP (2004) Quantitative paleotemperature estimates from $\delta^{18}\text{O}$ of chironomid head capsules preserved in arctic lake sediments. *Journal of Paleolimnology* 31: 267–274.

Postglacial Europe

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Introduction

The Chironomidae (nonbiting midges) are a family of the Diptera (true flies) whose larvae are abundant in most freshwater habitats (Figure 1). Individual lakes can include more than 100 different chironomid species, many of which have a restricted distribution with respect to water chemistry parameters (e.g., nutrient condition, oxygen, pH) or temperature. The strongly sclerotized larval head capsules of chironomids preserve well in lake sediments and remain identifiable, usually to species-group or genus level. Therefore, past changes in the chironomid fauna of individual lakes can be reconstructed based on the fossil record preserved in sediment cores. The lacustrine environment in which different fossil chironomid ‘morphotypes’ (or taxa) occur can be assessed by analyzing the taxonomic composition of chironomid assemblages in the surface sediments of lakes. Commonly, the surficial 1 cm of sediment will be analyzed in each lake, with the samples usually originating from the lake center. These samples incorporate chironomid remains deposited within the past few years to a few decades at most and provide an integrated picture of the chironomid taxa inhabiting a lake. Surface-sediment samples analyzed from lakes with different limnological conditions indicate that, even though the remains of many chironomids cannot be identified to species level, most taxa identifiable in fossil assemblages provide useful indications of past environmental conditions.

In Europe, fossil chironomid remains from postglacial lake deposits have increasingly been used as paleoecological and paleoenvironmental indicators. They are now used as proxy indicators to assess the natural development of lakes, the effect of pollution on lake ecosystems, and quantitative reconstruction of temperature, dissolved oxygen, chlorophyll *a*, total phosphorous, lake depth, stream flow, and distance to littoral vegetation. In addition, the stable oxygen isotope composition of chironomid head capsules has been used to reconstruct past variations in lake water $\delta^{18}\text{O}$ as a proxy for temperature changes. Here, we focus on four research themes: (1) the natural development of lakes, (2) the effects of pollution on lake ecosystems and the reconstruction of past changes in water quality, (3) the reconstruction of summer temperatures during the past 11.5 ka, and (4) the effect of past environmental conditions on the down-core chironomid assemblages.

Reconstructions of Lake Ontogeny

The use of chironomids as a proxy for past environmental conditions originates from the pioneer work of August Friedrich Thienemann (1882–1960). Thienemann studied the biota of lakes in relation to physical lake characteristics. He found that lakes can be classified according to the chironomid fauna where

certain taxa are likely to occur under a certain set of environmental conditions (Thienemann, 1954). Early applications of fossil chironomid analysis were built on this work. Lacustrine chironomid records were mainly intended to reconstruct changes in lake nutrient- and oxygen conditions related to the natural development of lake ecosystems. Since chironomids were considered a key indicator group for the classification of lakes according to their trophic state, fossil chironomid records were used to explore and test hypotheses related to lake ontogeny (i.e., lake origin and development). Examples include the natural eutrophication of lakes, decreasing hypolimnic oxygen conditions in lakes as a consequence of lake infilling, and the effects of natural acidification on lake ecosystems (see Walker, 1987 for an extensive review of these topics). Although most of these studies originate from the early days of chironomid-based paleoenvironmental reconstruction, chironomids are still used to reconstruct the natural development of lakes. Some studies quantitatively interpret chironomid stratigraphy using transfer functions to infer environmental conditions and also qualitatively to interpret lake ontogeny. For example, Ilyashuk et al. (2005) studied the postglacial chironomid stratigraphy at Lake Berkut, southern Kola Peninsula, Russia. These chironomid assemblages were used to quantify postglacial temperature changes, and in addition, the ecological demands of individual taxa formed the basis for a qualitative interpretation of the lake's ontogeny in terms of changes in lake-level, hypolimnic oxygen and acidification (Figure 2).

Effects of Pollution on Lake Ecosystems and Reconstruction of Water Quality

Modern assessment of pollution and monitoring of aquatic ecosystems are often based on reference conditions, as indicated by the biota. The reference condition is defined as the observed values of biological variables in the absence of significant human impact. A restoration goal can therefore be defined on the basis of knowledge of the present and past conditions before the onset of human influence. Fossil chironomid records have been used to reconstruct and assess the effects of human-induced nutrient pollution on benthic macroinvertebrate assemblages in lakes. One commonly used approach is to analyze chironomids in the framework of a multiproxy study, usually including analyses of a number of other biotic and abiotic indicators. Changes in chironomid assemblages are then examined in view of the ecology of the chironomid taxa found and in the context of the pollution history of the lake as inferred by the other proxy records analyzed in the same sediment core. In Europe, chironomids have been widely used to study the response of eutrophication and acidification (Bítušík et al., 2009; Heiri and Lotter, 2003;



Figure 1 (a) Coring of lacustrine sediments in Norway from the lake ice surface using a Nesje piston corer (picture by G. Velle); (b) imago of *Chironomus anthracinus* male (picture by K.P. Brodersen); (c) larvae of *C. anthracinus* where the arrow indicates the head capsule (picture by K.P. Brodersen); (d) chironomid head capsule (*Cricotopus* spp. B), the heavy sclerotized mentum (me) and the mandibles (ma) are important characters for the identification of subfossil larvae (picture by G. Velle); (e) subfossil head capsule (*Micropsectra* spp.) where many of the mouth-parts have been lost, as is often the case for subfossil material (SEM picture by K.P. Brodersen).

Henrikson et al., 1982; Ilyashuk et al., 2003). For example, Meriläinen and Hamina (1993) analyzed subfossil chironomid assemblages in a sediment core from Lake Päijänne, southern Finland, encompassing the past ~150 years. Based on the identified chironomid taxa, these authors calculated the benthic quality index (BQI), a semiquantitative water-quality index defined for living chironomid assemblages in deep stratified lakes (Wiederholm, 1980), in order to trace the local eutrophication history of the lake (Figure 3). The results from Meriläinen and Hamina (1993) indicate that the studied sub-basin of Lake Päijänne experienced the strongest phase of eutrophication during the 1970s when *Chironomus anthracinus* increased in abundance, and that the trophic status of the lake has slightly improved since.

Often, the pollution history of a lake can be further constrained by using other sources, for example, published water-sampling series or historical records. If the surface sediment

layers can be adequately dated (e.g., using ^{137}Cs or ^{210}Pb dating), the down-core chironomid record can be directly compared with instrumental measurements, or records of past industrial pollution. In the case of the chironomid-based reconstruction from Lake Päijänne, Meriläinen and Hamina (1993) were able to show that during the period of highest pollution, the deterioration of the water quality as indicated by the BQI was in close agreement with the biochemical oxygen demand (BOD) of discharge from pulp mills located close to the lake basin where the sediment core was taken. However, BQI values tended to lag changes in BOD-loading by several years (Meriläinen and Hamina, 1993). In a later study, Verbruggen et al. (2011) used the relationship between the BQI and nutrient concentrations apparent in European lakes to produce a quantitative estimate of past changes in concentrations of total phosphorus (TP) in Lake Päijänne, based on Meriläinen and Hamina's (1993) results. This reconstruction agreed closely with measurements of TP available from historical sources. When using reference conditions as a restoration goal, correct estimation and validation of the reference conditions are critical. Jyväsjärvi et al. (2010) assessed whether reference conditions for Finnish lakes defined from modern studies agreed with the preimpact state as assessed by the BQI from fossil chironomid assemblages. The authors compared the performance of different numerical approaches for defining the reference states in 73 semipristine lakes. They concluded that a regression model taking into account morphometric characteristics of lakes outperformed assigning undisturbed reference conditions based on the environmental characteristics of the lakes.

During recent years, transfer functions have been produced that directly relate water-quality variables such as chlorophyll *a*, TP, or oxygen to subfossil chironomid assemblages from surface sediments of lakes (see Brodersen and Quinlan, 2006 and references therein). Down-core applications of water-quality inference models have provided promising results, but the inference models and results arising from them need further validation (see below).

Chironomid-Based Summer Temperature Reconstruction

Fossil chironomids have received by far the most attention in recent years as quantitative paleotemperature indicators. Along long temperature gradients in temperate and subarctic latitudes in Europe, the distribution of chironomid taxa in lake surface sediments is closely related to summer air- and water temperature. Hence, chironomid-temperature transfer functions can be developed that predict temperature with a prediction error typically in the range of 1–1.5 °C. At present, chironomid-summer air or water temperature transfer functions are available from different regions in Europe, including Finland (e.g., Olander et al., 1999), northern Sweden (e.g., Larocque et al., 2001), Norway (e.g., Brooks and Birks, 2001), the Swiss Alps (e.g., Heiri and Lotter, 2010), and Iceland (Langdon et al., 2008). Chironomids have been used to reconstruct postglacial temperatures at a number of sites in Scandinavia, Spitsbergen, the British Isles, Iceland, southern France, the Kola Peninsula, central Spain and different regions in the Alps. Chironomid-based temperature reconstructions have shown

Age (cal years BP)	Chironomid zone	Lake conditions			Climatic conditions			
		Avg VWHO (mg O ₂ ·l ⁻¹)	pH status	Lake level change	Humidity	Tjul (°C)		
0	CZ-III	<2.0	Acidic		High	Increasing from 11.3 to 12.2		
1000		3.5–2.0						
1500						CZ-II	3.5–4.5	Circumneutral
2400	Low	Decreasing from 12.1 to 11.3						
3500			Increasing from 11.8 to 12.1					
4000		Significant lowering		~11.8				
5000			Gradual rise		High			
6000	CZ-I	<1.5		12.3–12.5				
7000								
7500								
8000								
8400								
9000								
9200								
10 100								

Figure 2 Chironomid-inferred climatic and environmental changes at Lake Berkut, Kola Peninsula. VWHO refers to average ‘end-of-summer’ volume-weighted hypolimnetic oxygen. Reproduced from Ilyashuk EA, Ilyashuk BP, Hammarlund D, and Larocque I (2005) Holocene climatic and environmental changes inferred from midge records (Diptera: Chironomidae, Chaoboridae, Ceratopogonidae) at Lake Berkut, southern Kola Peninsula, Russia. *The Holocene* 15: 897–914.

great potential for inferring past summer temperature during the major climatic fluctuations during the Late Glacial period. For some postglacial sites, chironomids have provided excellent results when evaluated against instrumental temperature records from meteorological stations or against tree-line fluctuations. For example, Larocque-Tobler et al. (2010) used chironomids to reconstruct temperatures from varved sediment with near-annual resolution at Lake Silvaplana in the Swiss Alps. First, the chironomid-inferred temperatures were validated against instrumental data of the last 150 years (Figure 4). Then, the reconstructions from the last 420 years were validated with local and regional reconstructions based on other proxies. And finally, encouraged by the promising results from the first studies, chironomids were used to reconstruct temperatures for the last 1000 years (see Larocque-Tobler et al., 2010 and references therein). The reconstructed temperatures indicate that the last part of the ‘Medieval Climate Anomaly’ (ca. AD 1032–1262), was 1 °C warmer than the climate reference period (1961–90).

Further, that the ‘Little Ice Age’ was separated into three cold phases that were about 1 °C colder than the reference period (Larocque-Tobler et al., 2010; Figure 4).

Several chironomid records have been described that support that, as in the Lake Silvaplana record, chironomid assemblages can respond to summer temperature changes smaller than the prediction error of available chironomid–temperature inference models. It is therefore tempting to interpret even relatively small amplitude changes in chironomid-inferred temperature. At the same time, several examples are available which demonstrate that fossil chironomid assemblages can be strongly influenced by factors unrelated, or only very indirectly related to climate, such as human activity or large scale vegetation changes. At least some chironomid records have been developed which also clearly demonstrate that such changes in catchment or in-lake environmental conditions can lead to biases and unrealistic variations in chironomid-based temperature estimates (Heiri and Lotter, 2003), although

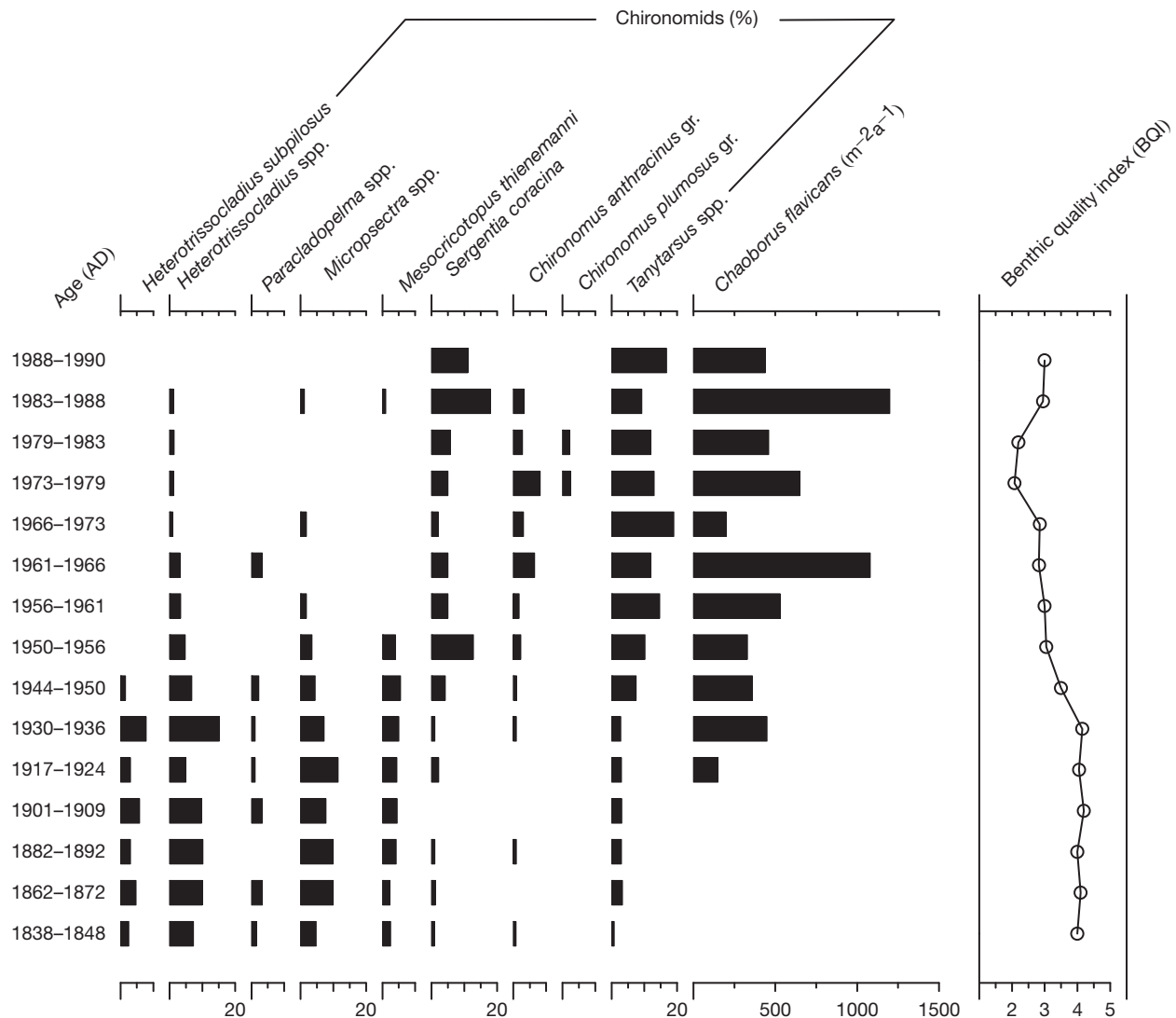


Figure 3 Relative abundance of selected fossil chironomid taxa and *Chaoborus flavicans* mandibles (as accumulation rates) in the uppermost sediments of Lake Päijänne, south Finland. The benthic quality index (BQI) was calculated for individual samples based on the fossil chironomid assemblages. Data and BQI values from Meriläinen JJ and Hamina V (1993) Recent environmental history of a large, originally oligotrophic lake in Finland: A palaeolimnological study of chironomid remains. *Journal of Paleolimnology* 9: 129–140.

these biases do not usually seem to exceed the prediction error of the applied transfer functions. As a consequence, opinions differ concerning the extent to which chironomid-based temperature records should be used to reconstruct postglacial climate change and whether temperature changes within the prediction error of the models should be interpreted without additional independent evidence to support these inferences.

The study of Velle et al. (2005) exemplifies some of the problems that can be encountered when applying chironomid-based temperature reconstruction to postglacial records. These authors used a chironomid–July air temperature transfer function to infer past temperature change from six postglacial chironomid profiles. There were differences in the chironomid-inferred temperature development between the six records. They concluded that these discrepancies are either related to different microclimates at the study sites, to a different

sensitivity of the chironomid fauna of the individual lakes to changes in summer temperature, or to chronological uncertainties (see Figure 5). In another study from Norway, Brooks (2003) presented a chironomid-based summer temperature record from Holebudalen, southern Norway. In this reconstruction, a distinct cooling in inferred July air temperature of $\sim 1.2^{\circ}\text{C}$ is reconstructed at ca. 8.3 cal ka BP (Figure 6). This cooling most likely represents the 8.2 ka cooling event and is clearly expressed in the chironomid-based temperature reconstruction. However, the episode is accompanied by only comparatively minor changes in the chironomid fauna (Figure 6). Brooks (2003) indicates that this decrease in inferred temperatures is largely driven by the absence of *C. anthracinus* type from the fossil chironomid assemblages at ca. 8.3 cal ka BP, a taxon at present restricted to comparatively warm Norwegian lakes.

In Swedish and Finnish Lapland, a number of Holocene temperature reconstructions have similarly been developed

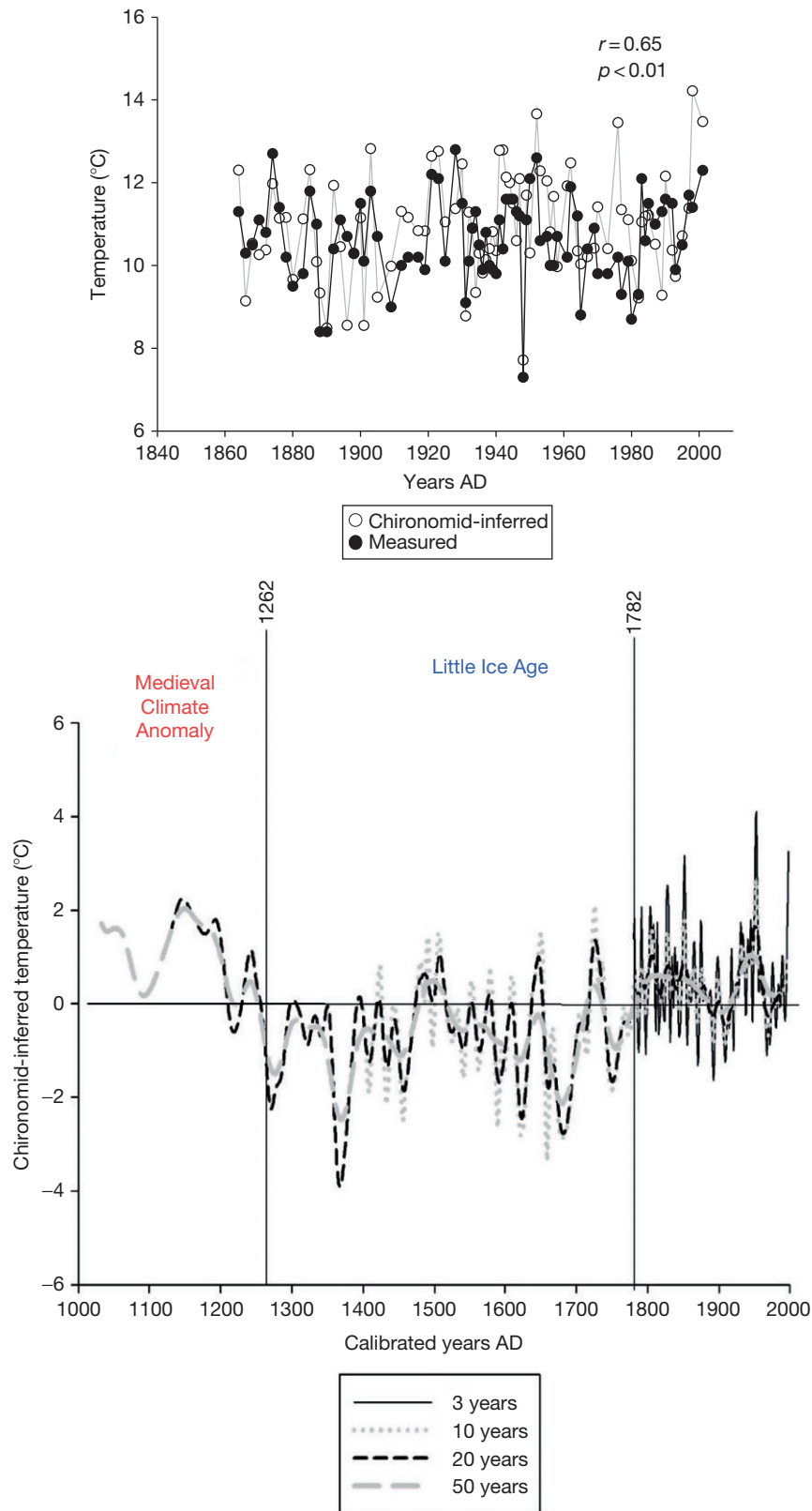


Figure 4 Top: Chironomid-inferred mean July air temperatures in Lake Silvaplana (Switzerland) at near-annual resolution and instrumental temperature measurements. Modified from Larocque I, Grosjean M, Heiri O, Bigler C, and Blass A (2009) Comparison between chironomid-inferred July temperatures and meteorological data AD 1850–2001 from varved Lake Silvaplana, Switzerland. *Journal of Paleolimnology* 41: 329–342. Bottom: Chironomid-inferred mean July air temperatures in Lake Silvaplana (Switzerland) with a 3-, 10-, 20-, and 50-year triangular filters. Modified from Larocque-Tobler I, Grosjean M, Heiri O, Trachsel M, and Kamenik C (2010) Thousand years of climate change reconstructed from chironomid subfossils preserved in varved lake Silvaplana, Engadine, Switzerland. *Quaternary Science Reviews* 29: 1940–1949.

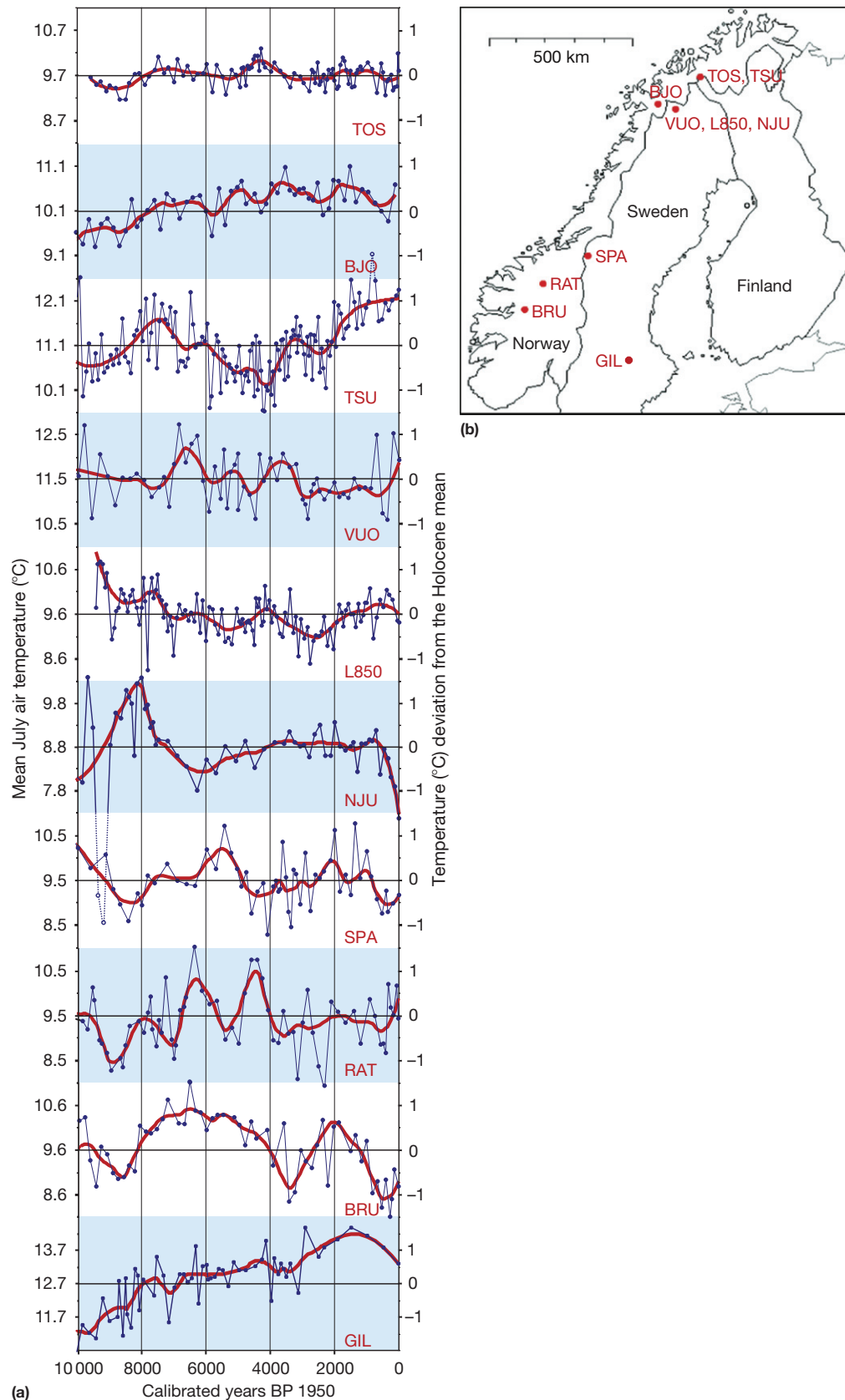


Figure 5 (a) Chironomid-inferred mean July air temperatures (thin line with dots) adjusted for glacio-isostatic rebound and with a Loess smoother standardized to include the nearest ± 1000 years fitted to the inferred values (bold line). Sites include Giltjärnen (GIL), Brurskardet (BRU), Råtåsjøen (RAT), Spåime (SPA), Vuoskkujávri (VUO), Lake 850 (L850), Njulla (NJU), Toskaljavi (TOS), Tsuolbmajavri (TSU), and Bjørnfieltjønn (BJO). (b) Map indicating the position of sites in Fennoscandia. Reproduced from Velle G, Brodersen KP, Birks HJB, and Willassen E (2010) Midges as quantitative temperature indicator species: Lessons for palaeoecology. *The Holocene* 20: 989–1002.

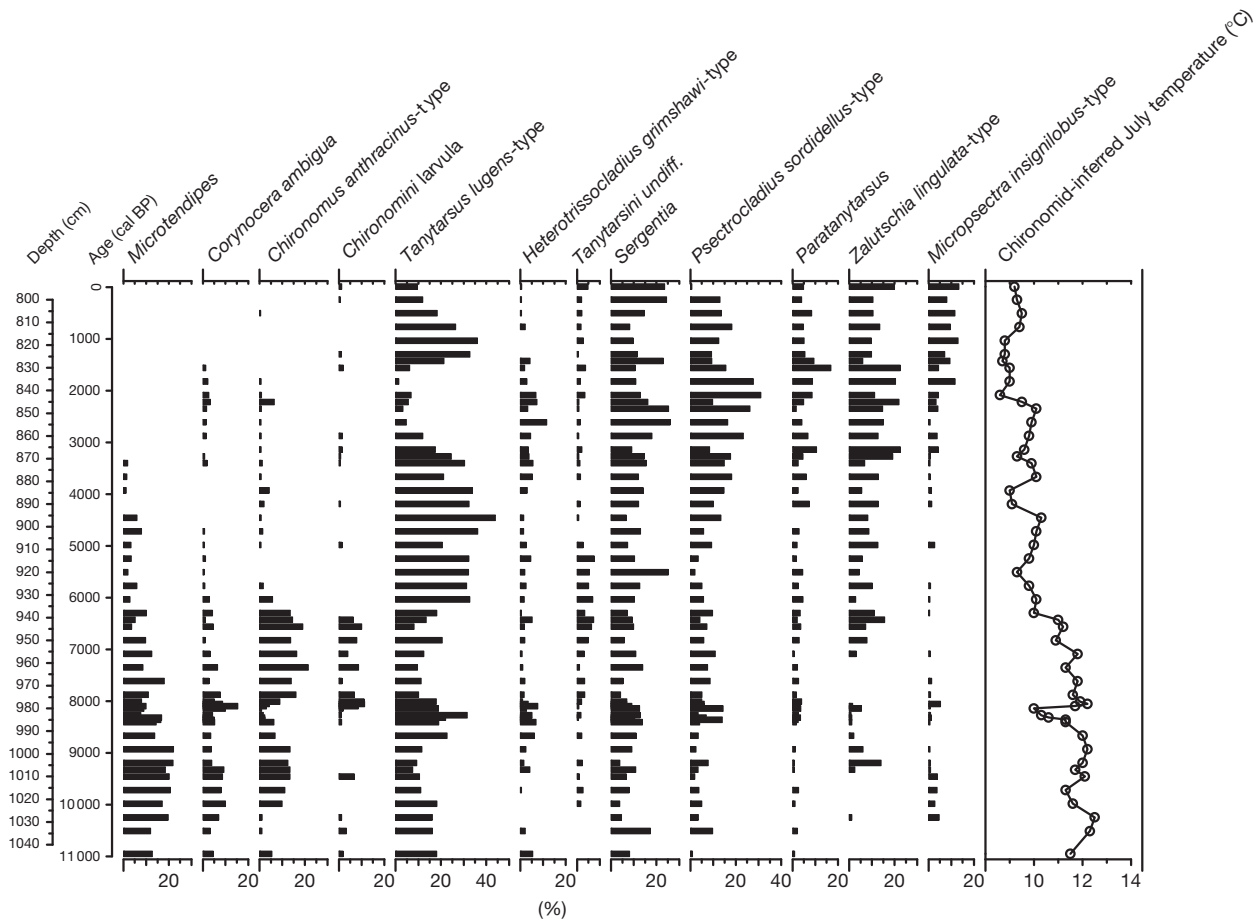


Figure 6 Chironomid assemblages in the Holocene sediments of Hølebudalen, a small Alpine lake in western Norway, and chironomid-inferred July air temperatures (°C) for individual assemblages. Data from Brooks SJ (2003) Chironomid analysis to interpret and quantify Holocene climate change. In: Mackay A, Batterbee RW, Birks HJB, and Oldfield F (eds.) *Global Change in the Holocene*, pp. 328–341. London: Arnold; Velle G, Brooks SJ, Birks HJB, and Willassen E (2005) Chironomids as a tool for inferring Holocene climate: An assessment based on six sites in southern Scandinavia. *Quaternary Science Reviews* 24: 1429–1462.

using two different chironomid–temperature transfer functions (Korhola et al., 2002; Larocque and Hall, 2004; Seppä et al., 2002). All three reconstructions from northern Sweden (Larocque and Hall, 2004) and one record from Finnish Lapland (Seppä et al., 2002) show a more-or-less continuous summer temperature decrease from 9 cal ky BP to the present (Figure 5). This would be expected for high northern latitudes during the mid- and late Holocene due to a decrease in summer insolation in this region related to orbital changes. The second chironomid-based summer temperature record from Finnish Lapland also indicates higher temperatures at ca. 9 ka BP than at present (Korhola et al., 2002). However, it also features periods of distinctly cooler inferred temperatures between ca. 6 and 4 ka BP and, depending on the numerical method used for the transfer function, possibly another cooling episode at ca. 3–2 ka BP. These coolings are not expressed as distinctly in the other chironomid-based temperature reconstructions from northern Fennoscandia, and it is difficult to decide whether these cold episodes in one of the five records are related to regional differences in postglacial climate in northern Scandinavia, whether they reflect biases related to the use of

different numerical methods and transfer functions, or whether they are the consequence of factors other than temperature that may have influenced the chironomid assemblages.

Causative Environmental Cues for the Postglacial Chironomid Communities

Environmental reconstruction should be examined within the context of the prediction error of the inference model and the sample-specific error estimates. If the influences of confounding variables exceed the reconstruction error, then the reconstruction could be compromised. The interest in chironomids as quantitative indicators of past temperature change has therefore also led to increased research on nonclimatic factors affecting fossil chironomid assemblages. Since temperature variations in the postglacial period are smaller than in the earlier glacial periods and often of a similar magnitude as the prediction errors of the applied transfer functions, much research on nonclimatic factors has focused on the postglacial period. Velle et al. (2010b) discussed how changes in lake

catchments and lake water chemistry can influence the postglacial development of chironomid assemblages. For example, natural changes in catchment vegetation and soils can affect nutrient conditions in lakes. As discussed above, chironomid assemblages react to changes in lake trophic state. It can therefore be expected that natural changes in lake primary productivity in the past may have affected fossil chironomid records and also influenced the temperature reconstruction, given that temperature and productivity shift independently (Figure 7). Similarly, vegetation changes, especially on acid bedrock, can affect pH and dissolved organic carbon concentrations of lake water and thereby affect the chironomid faunal composition of a lake. Some open questions remain, however: (1) How strongly are chironomid-based temperature records affected by such water chemistry changes; (2) How widespread are the effects of past water chemistry changes on postglacial chironomid records (e.g., with respect to geological or climatic conditions); and (3) Are chironomid records from different lake types affected to the same extent by past changes in water chemistry, productivity, or other potentially confounding factors?

The examples discussed above indicate that it is important to take into account the ecological preferences of the taxa involved when performing environmental reconstructions. Also, consider the appropriateness of the training set involved, especially when interpreting relatively small temperature changes inferred from chironomid assemblages. For example, in a study from Anterne in the French Alps, Millet et al. (2009) used chironomids to reconstruct temperatures during the last two millennia. A comparison with other temperature records in the region suggested that temperatures were successfully reconstructed, apart from the last hundred years. This period was marked by a peculiar fauna coinciding with diverging temperature patterns compared to other climate records. The fauna was notably characterized by the disappearance of *Procladius*, a genus with a warm temperature optimum compared with other taxa in the record. Arctic char (*Salvelinus alpinus*) was introduced into the lake in 1905 and the authors link the disappearance of *Procladius* with fish predation. In another study, Velle et al. (2005) reported a phase with dominance of *C. anthracinus* type during the early Holocene in Norway coinciding with what seemed to be markedly high-inferred temperatures. The authors interpreted this phase as reflecting (1) warm temperatures, (2) high production, (3) oxygen depletion, (4) lack of competition as *C. anthracinus* type then may thrive under suboptimal temperatures, or (5) a high feeding efficiency giving the taxon a competitive advantage in nutrient poor sediments. In a study from Stora Vidarvatn in northeast Iceland, Axford et al. (2009) compared the results from a temperature reconstruction based on a chironomid-temperature training set and that of the detrended correspondence analysis (DCA) sample scores to a 170 year instrumental temperature record. They found that the DCA scores were useful for inferring past temperatures at this site and that the chironomid-temperature transfer function developed for northwestern Iceland did not perform very well in reconstructing instrumental temperature shifts. The poor performance of the training set was possibly due to the lake's large size and depth relative to the calibration sites or to the limited resolution of the subfossil taxonomy.

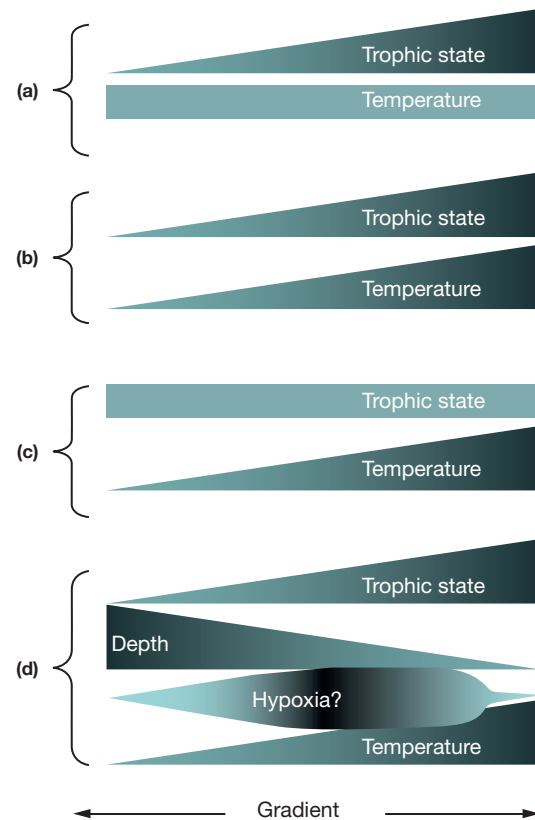


Figure 7 Possible scenarios that indicate how nutrient conditions could co-vary with temperature in lakes and potentially affect temperature reconstructions: (a) Gradient of lake productivity (trophic state) in a geographic area with constant and equal temperature. This scenario is suitable for trophic evaluation. (b) The commonly observed positive relation between temperature and lake trophic conditions. The scenario calls for careful interpretation of the results. (c) The ideal situation for building a temperature transfer function. However, such a scheme is rarely observed within a restricted geographical area. (d) For all previous scenarios, the confounding influence of oxygen availability (hypoxia). Reproduced from Velle G, Brodersen KP, Birks HJB, and Willassen E (2010) Midges as quantitative temperature indicator species: Lessons for palaeoecology. *The Holocene* 20: 989–1002.

Approaches that seem to be especially robust when performing paleoenvironmental reconstructions include multisite chironomid-based reconstructions (Larocque and Hall, 2004; Velle et al., 2005), single-site multiproxy reconstructions (see examples given in Brooks, 2003; Nyman et al., 2008; Velle et al., 2010a; see designated sections in the book for other proxies), multiproxy multisite reconstructions (e.g., Heiri et al., 2004), or reconstructions that are validated against instrumental or known environmental records (e.g., Larocque-Tobler et al., 2010). An option including environmental reconstruction by use of more than one proxy seems particularly promising. Such an approach allows the combination of a chironomid-based approach to temperature reconstruction with proxies that have been shown to be sensitive to past temperature change, but remain difficult to use as quantitative temperature indicators on their own during certain periods (e.g., pollen or oxygen isotopes in lake marl). In many combinations, the different paleotemperature proxies would

be affected by past temperature change via fundamentally different mechanisms. A close agreement between different paleotemperature records based on different proxies and from different localities in the same region would therefore increase the confidence that past temperatures rather than other environmental variables were the driving force for the chironomid assemblages. The role of the chironomid-based reconstruction is then to quantify past summer temperature changes, whereas the additional nonchironomid proxies provide independent support for the chironomid-based inferences or point toward problematic parts of the record. For example, Heiri et al. (2004) present a chironomid-based July air temperature reconstruction for the early and mid-Holocene from the Swiss Alps which shows a close agreement with an independently dated pollen-based reconstruction of past tree-line altitude (Figure 8). At present, tree-line altitude in the Alps is largely constrained by temperatures during the growing-season. The close agreement between the chironomid-based July temperature record and the reconstruction of past tree-line altitude from two sites more than 60 km apart suggests that, in this case, the chironomid-based reconstruction successfully captured past centennial-scale changes in summer temperature. In another example, Velle et al. (2010a) used chironomids and pollen from Lake Brurskardet in Norway to infer Holocene July air temperatures. There were incongruences between the inferred temperatures from the two proxies and in the light of the in-lake and catchment-related processes as assessed by diatoms, plant macrofossils and sediment composition, the authors discussed which reconstruction is most likely correct at any one time. The chironomid-inferred temperatures closely resembled results from an ordination analysis of the chironomids during the earliest Holocene. At this time, the pollen suggested unrealistically high temperatures, most likely owing to a high proportion of long-distance transported pollen. The chironomids suggest a Holocene maximum around 9.8 cal ka BP, 800 years prior to the pollen-inferred Holocene maximum. However, according

to the ordination techniques, the chironomid assemblage changed to a limited extent along the main temperature gradient at the time, and to a larger extent along the conductivity gradient. This may suggest that the chironomids responded to changes in conductivity and sediment load. There was coincident substantial shift in sediment input to the lake from inorganic silt to more organic gyttja (Velle et al., 2010a).

In the postglacial period, human impact on lake ecosystems can further complicate the use of chironomids for paleotemperature reconstruction purposes. During the historical period, human land use and industrial activity had led to a widespread eutrophication of lakes in many parts of Europe. Furthermore, prehistoric agriculture and land-use changes may have influenced lacustrine ecosystems as well. Prehistoric anthropogenic effects on lakes were probably not as widespread as lake eutrophication during the historical period. However, evidence suggests that these effects may have been considerable, locally. For example, Heiri and Lotter (2003) present a fossil chironomid record covering the past 9 ka from the Bernese Alps, Switzerland. During three distinct phases, the earliest starting at ca. 4 cal ka BP, the accumulation rate of the dominant lacustrine chironomid taxa was strongly reduced in the record, whereas most running water and littoral taxa showed no changes in accumulation rate. A comparison with other records produced within the framework of a multiproxy study indicated that these decreases in lacustrine chironomids occurred during phases for which geochemical measurements indicate reduced oxygen conditions in the lake. Furthermore, for all three of these phases, paleobotanical proxies showed clear evidence of increased human activity in the lake catchment, such as forest clear-cutting and pasturing. Combined, these records suggest that prehistoric human activity led to anoxic conditions during three distinct periods in the lake's past and reduced the abundance of, or even eliminated, some of the chironomid taxa living in the lake. When a chironomid-temperature transfer function was applied to this record,

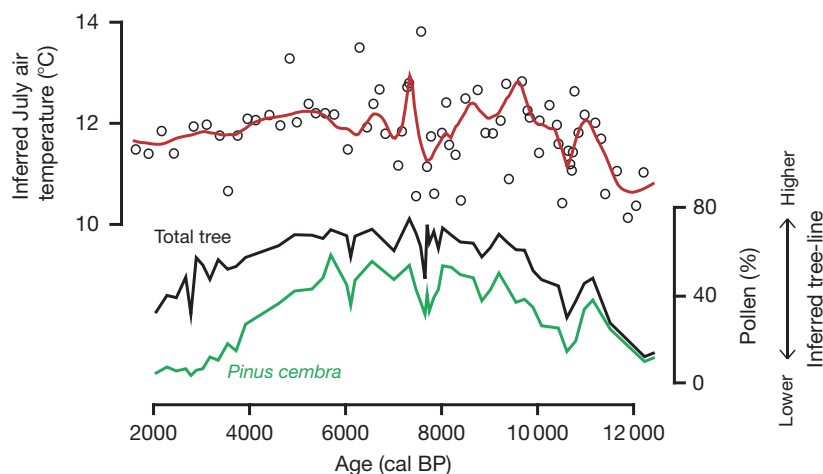


Figure 8 Chironomid-inferred temperatures and reconstruction of relative shifts in tree-line altitude from the Swiss Alps. July air temperatures are reconstructed from chironomid assemblages in the sediments of Hinterburgsee, as small subalpine lake in the Northern Swiss Alps. The record is smoothed using a locally weighted regression (LOWESS) smoother. Relative changes in past tree-line altitude were reconstructed based on total tree and *Pinus cembra* pollen in the sediments of Gouille' Rion, a small lake situated at present-day tree-line in the Central Swiss Alps. Data from Heiri O, Tinner W, and Lotter AF (2004) Evidence for cooler European summers during periods of changing melt-water flux to the North Atlantic. *Proceedings of the National Academy of Sciences of the United States of America* 101: 15285–15288.

these episodes were accompanied by distinct coolings in the chironomid-inferred temperatures, and Heiri and Lotter (2003) concluded that these apparent coolings were most likely an artifact of prehistoric human impact on the lake ecosystem rather than a genuine temperature signal.

Future Developments

In recent years, most research on fossil chironomids has focused on their use as paleotemperature proxies. Temperature transfer functions are most easily developed in regions with large temperature gradients in a geographically restricted area. Therefore, most available transfer functions and many of the available postglacial temperature reconstructions in Europe originate from mountainous regions. Considering the ongoing public interest in climate research, and the need for more information on natural climate variability during the pre-instrumental period, it is expected that postglacial chironomid-based temperature reconstruction will continue to remain a focus of research. By optimizing the gradient length for environmental variables other than temperature, chironomids also respond significantly to those variables in modern training sets. Hence, studies are emerging that explore the use of chironomids as quantitative proxies for effective moisture through lake

depth, for example, in Finland (Luoto, 2009), for hypolimnetic oxygen content of the lake, for example, on the Kola peninsula (Ilyashuk et al., 2005), and for TP and chlorophyll *a*, for example, in the United Kingdom (Langdon et al., 2006). As such studies increase in popularity, their resulting paleoenvironmental reconstructions will hopefully be rigorously tested against additional sediment records and against known fluctuations from instrumental records.

Next to methodological advancements, such as the development of new numerical tools for transfer function development (e.g., Velle et al., 2011) and reconstruction validation (Telford and Birks, 2011), and an increase in the taxonomic resolution of fossil chironomid records (e.g., Heiri and Lotter, 2010), one of the main foci will most likely be the expansion of the spatial coverage of chironomid-based temperature records. This may lead to taxonomic harmonization and cross-evaluation of both existing transfer functions and forthcoming temperature reconstructions. In addition, studies that merge our fundamental knowledge of chironomid ecology based on empirical and experimental data with traditional paleolimnological research may provide additional information, for example, when interpreting chironomid stratigraphies, and aid in identifying confounding factors in environmental reconstructions (see Brodersen and Quinlan, 2006 and references therein).

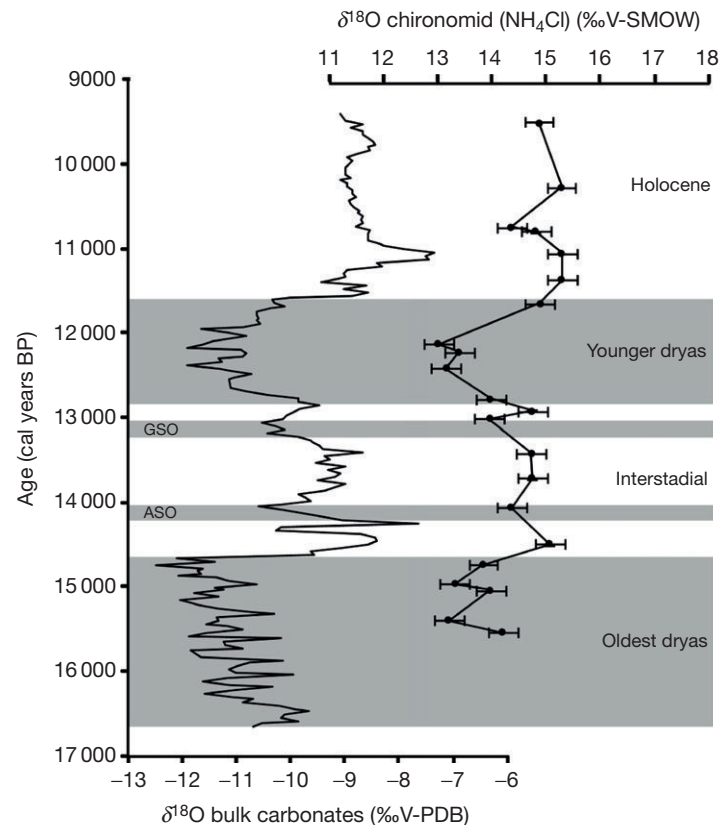


Figure 9 Comparison between stable oxygen isotope records from bulk carbonates (left curve) and from chironomids (right curve). $\delta^{18}\text{O}$ data from the chironomids are per mille relative to the Vienna-Standard Mean Ocean Water (V-SMOW). Gray areas indicate cold periods. GSO, Gerzensee Oscillation; ASO, Aegelsee Oscillation. Reproduced from Verbruggen F, Heiri O, Reichert GJ, and Lotter AF (2010) Chironomid $\delta^{18}\text{O}$ as a proxy for past lake water $\delta^{18}\text{O}$: A Late Glacial record from Rotsee (Switzerland). *Quaternary Science Reviews* 29: 2271–2279.

Chironomid remains are highly resistant to microbial degradation and chemical alteration after their burial in the sediments. Recently, isotopes have been successfully measured on chironomid fossil head capsules to produce records of $\delta^{18}\text{O}$ as an indirect proxy for climatic changes (Verbruggen et al., 2010; Wooller et al., 2004; Figure 9) and $\delta^{13}\text{C}$ to trace dietary links and carbon cycling in the food webs (van Hardenbroek et al., 2010). These methods can potentially provide an alternative to stable isotope measurements on other lacustrine sediment components, such as lake marl and diatoms, and significantly expand the applicability of fossil chironomids as paleoclimatic and paleoecological indicators. There is also great potential to validate paleoenvironmental reconstructions by information gained from isotope studies, for example, to examine how temporal changes in trophic position can influence the reconstructions.

Next to paleoclimate reconstruction, a research field within paleolimnology that has received an increasing amount of attention is the reconstruction of preanthropogenic water quality and ecosystem-baseline conditions for freshwater ecosystems. In Europe, this research has obtained further relevance due to recent developments in European legislation, since the ecological status of lakes and rivers is now defined in relation to the expected state of these ecosystems under minimal anthropogenic influence. Compared with other freshwater indicators such as diatoms, fossil chironomids have received comparatively little attention in reconstructing parameters reflecting past water-quality conditions. However, the examples mentioned above (see Section 'Effects of Pollution on Lake Ecosystems and Reconstruction of Water Quality') indicate that chironomids have potential as quantitative indicators of past nutrient conditions and productivity in lakes. Furthermore, chironomids are one of the few benthic macroinvertebrate groups living in profundal environments that preserve well and remain identifiable in most lacustrine sediments. For restoration purposes, there is therefore considerable potential for using chironomids to reconstruct the natural faunal assemblages in European lakes during the preanthropogenic period and to use them to establish water-quality reference conditions. This potential has been recently demonstrated by the pioneering work of Jyväsjärvi et al. (2010).

See also: **Carbonate Stable Isotopes:** Lake Sediments.
Chironomid Records: Chironomid Overview; Late Pleistocene of Europe.
Plant Macrofossil Methods and Studies: Treeline Studies.
Pollen Methods and Studies: Numerical Analysis Methods.

References

- Axford Y, Geirsdóttir Á, Miller G, and Langdon P (2009) Climate of the Little Ice Age and the past 2000 years in northeast Iceland inferred from chironomids and other lake sediment proxies. *Journal of Paleolimnology* 41: 7–24.
- Bitušik P, Kubovčík V, Štefková E, Appleby P, and Svitok M (2009) Subfossil diatoms and chironomids along an altitudinal gradient in the High Tatra Mountain lakes: A multi-proxy record of past environmental trends. *Hydrobiologia* 631: 65–85.
- Brodersen KP and Quinlan R (2006) Midges as palaeoindicators of lake productivity, eutrophication and hypolimnetic oxygen. *Quaternary Science Reviews* 25: 1995–2012.
- Brooks SJ (2003) Chironomid analysis to interpret and quantify Holocene climate change. In: Battarbee RW, Oldfield F, Birks HJB, and Mackay A (eds.) *Global Change in the Holocene*, pp. 328–341. London: Arnold.
- Brooks SJ and Birks HJB (2001) Chironomid-inferred air temperatures from Late-glacial and Holocene sites in north-west Europe: Progress and problems. *Quaternary Science Reviews* 20: 1723–1741.
- Heiri O and Lotter AF (2003) 9000 years of chironomid assemblage dynamics in an Alpine lake: Long-term trends, sensitivity to disturbance, and resilience of the fauna. *Journal of Paleolimnology* 30: 273–289.
- Heiri O and Lotter A (2010) How does taxonomic resolution affect chironomid-based temperature reconstruction? *Journal of Paleolimnology* 44: 589–601.
- Heiri O, Tinner W, and Lotter AF (2004) Evidence for cooler European summers during periods of changing meltwater flux to the North Atlantic. *PNAS* 101: 15285–15288.
- Henrikson L, Olofsson JB, and Oscarson HG (1982) The impact of acidification on Chironomidae (Diptera) as indicated by subfossil stratification. *Hydrobiologia* 86: 223–229.
- Ilyashuk B, Ilyashuk E, and Dauvalter V (2003) Chironomid responses to long-term metal contamination: A paleolimnological study in two bays of Lake Imandra, Kola Peninsula, northern Russia. *Journal of Paleolimnology* 30: 217–230.
- Ilyashuk EA, Ilyashuk BP, Hammarlund D, and Larocque I (2005) Holocene climatic and environmental changes inferred from midge records (Diptera: Chironomidae, Chaoboridae, Ceratopogonidae) at Lake Berkut, southern Kola Peninsula, Russia. *The Holocene* 15: 897–914.
- Jyväsjärvi J, Nyblom J, and Hämäläinen H (2010) Palaeolimnological validation of estimated reference values for a lake profundal macroinvertebrate metric (benthic quality index). *Journal of Paleolimnology* 44: 253–264.
- Korhola A, Vasko K, Toivonen HTT, and Olander H (2002) Holocene temperature changes in northern Fennoscandia reconstructed from chironomids using Bayesian modelling. *Quaternary Science Reviews* 21: 1841–1860.
- Langdon PG, Holmes N, and Caseldine CJ (2008) Environmental controls on modern chironomid faunas from NW Iceland and implications for reconstructing climate change. *Journal of Paleolimnology* 40: 273–293.
- Langdon PG, Ruiz Z, Brodersen KP, and Foster IDL (2006) Assessing lake eutrophication using chironomids: Understanding the nature of community response in different lake types. *Freshwater Biology* 51: 562–577.
- Larocque I and Hall RI (2004) Holocene temperature estimates and chironomid community composition in the Abisko Valley, northern Sweden. *Quaternary Science Reviews* 23: 2453–2465.
- Larocque I, Hall RI, and Grahn E (2001) Chironomids as indicators of climate change: A 100-lake training set from a subarctic region of northern Sweden (Lapland). *Journal of Paleolimnology* 26: 307–322.
- Larocque-Tobler I, Grosjean M, Heiri O, Trachsel M, and Kamenik C (2010) Thousand years of climate change reconstructed from chironomid subfossils preserved in varved lake Silvaplana, Engadine, Switzerland. *Quaternary Science Reviews* 29: 1940–1949.
- Luoto TP (2009) A Finnish chironomid- and chaoborid-based inference model for reconstructing past lake levels. *Quaternary Science Reviews* 28: 1481–1489.
- Meriläinen JJ and Hamina V (1993) Recent environmental history of a large, originally oligotrophic lake in Finland: A palaeolimnological study of chironomid remains. *Journal of Paleolimnology* 9: 129–140.
- Millot L, Arnaud F, Heiri O, Magny M, Verneaux V, and Desmet M (2009) Late-Holocene summer temperature reconstruction from chironomid assemblages of Lake Anzerne, northern French Alps. *The Holocene* 19: 317–328.
- Nyman M, Weckström J, and Korhola A (2008) Chironomid response to environmental drivers during the Holocene in a shallow treeline lake in northwestern Fennoscandia. *The Holocene* 18: 215–227.
- Olander H, Birks HJB, Korhola A, and Blom T (1999) An expanded calibration model for inferring lakewater and air temperatures from fossil chironomid assemblages in northern Fennoscandia. *The Holocene* 9: 279–294.
- Seppä H, Nyman M, Korhola A, and Weckström J (2002) Changes of treelines and alpine vegetation in relation to post-glacial climate dynamics in northern Fennoscandia based on pollen and chironomid records. *Journal of Quaternary Science* 17: 287–301.
- Telford R and Birks HJB (2011) A novel method for assessing the significance of quantitative reconstructions inferred from biotic assemblages. *Quaternary Science Reviews* 30: 1272–1278.
- Thienemann A (1954) *Chironomus*. Leben, Verbreitung und wirtschaftliche Bedeutung der Chironomiden. *Binnengewässer* 20: 1–834.
- van Hardenbroek M, Heiri O, Grey J, Bodelier P, Verbruggen F, and Lotter A (2010) Fossil chironomid $\delta^{13}\text{C}$ as a proxy for past methanogenic contribution to benthic food webs in lakes? *Journal of Paleolimnology* 43: 235–245.
- Velle G, Bjune AE, Larsen J, and Birks HJB (2010a) Holocene climate and environmental history of Brurskardstjønni, a lake in the catchment of Øvre Heimdalsvatn, south-central Norway. *Hydrobiologia* 642: 13–34.

- Velle G, Brodersen KP, Birks HJB, and Willassen E (2010b) Midges as quantitative temperature indicator species: Lessons for palaeoecology. *The Holocene* 20: 989–1002.
- Velle G, Brooks SJ, Birks HJB, and Willassen E (2005) Chironomids as a tool for inferring Holocene climate: An assessment based on six sites in southern Scandinavia. *Quaternary Science Reviews* 24: 1429–1462.
- Velle G, Kongshavn K, and Birks HJB (2011) Minimizing the edge-effect in environmental reconstructions by trimming the calibration set: Chironomid-inferred temperatures from Spitsbergen. *The Holocene* 21: 417–430.
- Verbruggen F, Heiri O, Meriläinen JJ, and Lotter AF (2011) Subfossil chironomid assemblages in deep, stratified European lakes: Relationships with temperature, trophic state and oxygen. *Freshwater Biology* 56: 407–423.
- Verbruggen F, Heiri O, Reichert GJ, and Lotter AF (2010) Chironomid $\delta^{18}\text{O}$ as a proxy for past lake water $\delta^{18}\text{O}$: A Lateglacial record from Rotsee (Switzerland). *Quaternary Science Reviews* 29: 2271–2279.
- Walker IR (1987) Chironomidae (Diptera) in Paleoecology. *Quaternary Science Reviews* 6: 29–40.
- Wiederholm TE (1980) Use of benthos in lake monitoring. *Journal of Water Pollution Control Federation* 52: 537–547.
- Wooler MJ, Francis D, Fogel ML, Miller GH, Walker IR, and Wolfe AP (2004) Quantitative paleotemperature estimates from delta O-18 of chironomid head capsules preserved in arctic lake sediments. *Journal of Paleolimnology* 31: 267–274.

Postglacial Southern Hemisphere

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Introduction

The use of fossil chironomid remains as indicators of past environmental conditions is now a well established approach (Walker, 2001). From reconstructing trophic status to climate, chironomids have been deemed among the best indicators for their ability to provide quantifiable estimates of change. The use of chironomids in postglacial environmental reconstructions in Australia, New Zealand, and South America has not been extensive, but in recent years has undergone a resurgence of interest with many workers following the success of chironomid-based reconstructions from Europe and North America. A review by Verschuren and Eggermont (2006) provides a detailed account of chironomid paleoecology in the Southern Hemisphere prior to 2006. In this article, we review the published information about postglacial fossil chironomids from Australia, New Zealand, and South America and outline the directions that chironomid research has followed in these areas. We also discuss the importance of chironomid research in the Southern Hemisphere, identifying directions for future research.

Australasian–South American Fossil Chironomid Studies

Australia

Several paleoecological studies that primarily focus on chironomids as paleoindicators have been published from Australia. Paterson and Walker (1974) investigated changes in paleoproductivity. Using the environmental range of tolerance of the chironomid *Tanytarsus barbitarsis*, the authors reconstructed historical salinity changes in Lake Werowrap (38°15'S, 143°30'E) in western Victoria (Figure 1). Based on experiments identifying the range of tolerance of *T. barbitarsis* to total dissolved solids (salinity) and changes in its relative abundance throughout the Lake Werowrap sequence, the authors concluded that freshwater conditions occurred in the early part of the lake's history with a shift to more saline conditions midway through the sequence to the present day. The time span for the Lake Werowrap sequence remains unknown due to the lack of chronology, but it is likely to represent changes since the mid-to late Holocene.

The first study in Australia to provide a quantitative reconstruction of Holocene climate using chironomids was carried out by Dimitriadis and Cranston (2001) from Lake Barrine (17°15'S, 145°38'E), on the Atherton Tablelands, northeast Queensland. This study examined the relationship between the physical and chemical parameters of more than 68 water bodies and their modern chironomid fauna to assess the relative importance of climate and environmental variables on eastern Australian chironomids. They verified a significant

relationship between climate and chironomids, with a correlation (r^2) of 0.94 with mean air temperature of the warmest quarter (TWARM) and 0.59 with mean precipitation of the driest quarter. Through the use of a mutual climatic range method, rather than the more commonly used weighted averaging method, Holocene temperatures were reconstructed for Lake Barrine over the last 9 cal ka. Temperature reconstructions indicate that annual temperatures were warmer by ~2.5 °C from 9.01 to 6.89 cal ka BP but may have been up to 6.6 °C cooler in the summer months between 6.89 and 6.09 cal ka BP. Temperatures remained ~2.5 °C warmer than the current mean from 6.09 cal ka BP to the present day. Precipitation estimates indicate drier climates by ~500 mm of moisture per year from 6.89 to 6.09 and 5.13 to 2.65 cal ka BP, but large ranges associated with these estimates make these reconstructions equivocal. Although estimates of 6.6 °C temperature depressions seem extreme for periods during the Holocene, evidence from regional pollen records provide some support for the chironomid-based estimates and imply similar changes in temperature at these times during the Holocene.

More recently, there have been new developments in Australia using chironomids as quantitative indicators of temperature. Rees et al. (2008) presented a temperature transfer function based on subfossil assemblages of chironomids, Ceratopogonidae (biting midges), and Chaoboridae (phantom midges) from 47 Tasmanian lakes. The chironomid distributions surveyed in this study were strongly influenced by pH, air temperature of the TWARM, annual radiation, Mg^{2+} , precipitation, SiO_2 , and water depth. The observed temperatures were significantly correlated with temperatures predicted from the model, indicating that midges are an appropriate tool for modeling the temperature history of Tasmania. This investigation provides the first such transfer function built for Australia, providing new tools for testing climate hypotheses in the Southern Hemisphere.

Successive studies have focused on chironomid-based environmental reconstructions. The first (Rees and Cwynar, 2010a), shows a temperature reconstruction of the TWARM from two sites Platypus and Eagle Tarn (42°40'S, 146°35'E) in Mount Field National Park, Tasmania. The authors attempt to answer several paleoclimatic questions, including the existence of the Antarctic Cold Reversal (ACR; 14.5–12.5 cal ka BP) in Tasmania, the magnitude of temperature change associated with it, and the temperature differences between the Holocene thermal maximum and modern temperatures. The chironomid records from both sites indicated early warmth with TWARM values at modern levels during the ACR. However, the decrease in the organic content of the loss-on-ignition (LOI) profile from Eagle and Platypus Tarns corresponds with the timing of the ACR in the deuterium content of the EPICA Dome C ice-core. A mid-Holocene thermal maximum, based on the chironomid-inferred temperature estimates, corresponds with

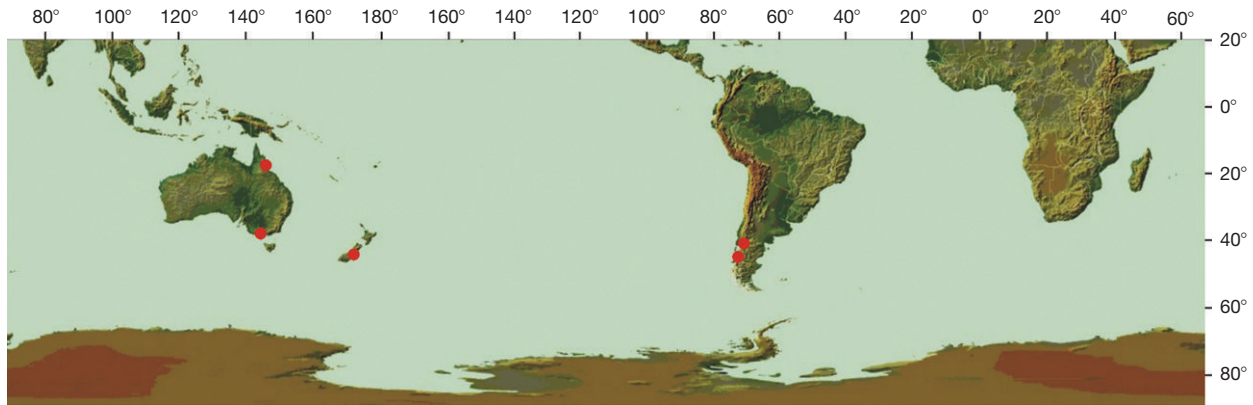


Figure 1 Main regions for postglacial chironomid studies in Australia, New Zealand, and South America.

warming inferred from pollen records in the region. During this period, temperatures were considered to be similar to or 1.5 °C warmer than modern. In a second study, [Rees and Cwynar \(2010b\)](#) investigated the response of chironomids to a pH gradient in Tasmania. The inference model suggests that higher species richness at low pH may result from the large pool of tropical taxa that have adapted to naturally occurring acidic conditions. The authors suggest that during glacial episodes an isthmus would have formed between mainland Australia and Tasmania, allowing for the migration of plants and animals. This could have been a potential route of colonization for the suite of acid-adapted taxa from mainland Australia and Southeast Asia into Tasmania. This model is among the few that have been developed to estimate pH from chironomid remains.

New Zealand

The first paleoecological studies involving chironomids from New Zealand have primarily been qualitative investigations focusing on paleolimnological changes such as lake productivity or catchment and lake system development. Early work was carried out by [Deevey \(1955\)](#) from a swamp deposit sequence in the upper Pyramid Valley (42°58' S, 172°35' E), central South Island. In this study, [Deevey \(1955\)](#) used a range of terrestrial and aquatic macrofossils including cladocera, ostracods, mollusks, plant macrofossils, charophytes, green algae, pollen, diatoms, and chironomids to reconstruct lake history. The dynamics of these proxies were interpreted to indicate trophic changes and lake-level fluctuations during the past ca. 3.9 cal ka BP. Interpretations from this study are generalized, mainly due to a poor taxonomic resolution and ecological understanding of New Zealand freshwater biota at the time.

In two other studies, [Schakau \(1986, 1990\)](#) examined chironomids from Holocene sequences in two lakes: Lakes Taylor (42°46' S, 172°14' E) and Grasmere (43°59' S, 171°45' E), both located in the central South Island of New Zealand. In Lake Taylor ([Schakau, 1986](#)), the chironomid assemblages spanning the last ca. 3.2 cal ka indicated a higher abundance of Tanytarsini in the older sediments and a shift in dominance from Tanytarsini to Chironomini after ca. 2.5 cal ka BP. This shift was not statistically significant and therefore not considered to be a response to climate change or lake trophic status.

At Lake Grasmere ([Schakau, 1990](#)), changes in chironomid assemblages were used to reconstruct lake paleoenvironmental history over the last 6.8 cal ka. The sequence is dominated by *Corynocera* sp. but shows marked shifts in faunal abundance and composition, characterized by increases in Chironomini, Orthocladinae, and Tanytarsini genera. The observed changes were not interpreted to be driven by climate as pollen analysis from the site did not indicate any major vegetation changes that might be driven by climate change during this time. The dominance of *Corynocera* sp. and the relatively stable faunal assemblage suggest a period of general environmental stability over the past 6.8 ka. Distinct changes in chironomid structure throughout this time were thought to reflect short-term changes in lake productivity, hydrological inflow, and rate and type of sedimentation.

More recently, [Woodward and Shulmeister \(2005\)](#) combine pollen, Trichoptera, chironomids, and algal abundance to reconstruct the development of coastal Lake Forsyth (43°48' S, 172°44' E), South Island, from a tidal bay to a barrier-dammed lake. The main fluctuations recorded in chironomids (mainly *Chironomus* sp.) and other aquatic taxa during this time were interpreted as a response to changes in salinity, catchment deforestation, and eutrophication throughout the past 7 cal ka. This study indicates a significant relationship between the abundance of aquatic taxa and regional vegetation, and proposes catchment deforestation resulting from human disturbance as the main cause of biological change in the lake history. Increases in abundance and diversity of chironomids after AD 1840 suggested an increase in sediment and nutrient input into the lake and a decrease in salinity as a result of deforestation. In the late 1800s, a significant peak in freshwater algae and a drop in abundance and diversity of chironomids and other aquatic invertebrates were interpreted as a period of lake eutrophication. This study emphasizes the advantages of using multiproxy analyses to interpret environmental change.

Since these initial studies, there have been major improvements both in the statistical approaches and the subfossil taxonomic identification applied to New Zealand chironomids, culminating in the development of several chironomid-temperature inference models ([Dieffenbacher-Krall et al., 2007; Vandergoes et al., 2008; Woodward and Shulmeister, 2006a,b, 2007](#)). The first chironomid-based transfer function

developed in the Southern Hemisphere was published recently by Woodward and Shulmeister (2006a,b). They analyzed chironomids and environmental variables from 46 New Zealand lakes, identifying temperature (February mean air temperature) and lake productivity (chlorophyll *a*) as the main drivers of chironomid distribution. The temperature model was more robust with an r^2 (jack) of 0.77 and a prediction error (root mean squared error of prediction, RMSEP jack) of 1.31. A less reliable model was produced for chlorophyll *a* with a correlation r^2 (jack) of 0.49 and a prediction error (RMSEP jack) of 0.46 Log¹⁰ mg l. In this paper, the authors identify the problematic distribution of the ubiquitous *Chironomus* morphotype.

More recently, Dieffenbacher-Krall et al. (2007) provided a crucial cornerstone for the development of modern chironomid–temperature calibration sets and the use of chironomids to quantify postglacial climatic change in New Zealand. The study used data from 61 lakes and included a number of lakes from high altitudes in South Island, New Zealand. The multivariate analysis showed that summer air temperature (SmT) explains the greatest amount of variance in chironomid distribution of any of the measured variables. This investigation presents a robust inference model (RMSEP jack of 1.27 °C, r^2 jack 0.80) to reconstruct mean SmTs from subfossil chironomids. Soon after, Vandergoes et al. (2008) applied the Dieffenbacher-Krall et al. (2007) chironomid-inferred temperature model to a down-core chironomid study of Boundary Stream Tarn (44°02'S, 170°07'E) in the Southern Alps, New Zealand (Figure 2(a)). In this study, pollen and chironomids were compared and the chironomid model was used to quantify Late Glacial climate change and explore the existence of cooling associated with the ACR. The results indicate a ca.1000-year disruption to the Late Glacial warming trend and an overall cooling consistent with the ACR. The main interval of chironomid-inferred summer temperature depression (2–3 °C), indicated by increases in cool climate *Chironomus* and *Tanytarsus* sp., lasted about 700 years during the ACR. Following this cooling event, both proxies indicate a warming step to temperatures slightly cooler than present during the Younger Dryas (YD) chronozone (12.9–11.5 cal ka BP; Figure 2(b)). Minor differences between the two proxies were attributed to seasonal climate variations and a possible lag in vegetation response to abrupt climate

change. The chironomid–temperature reconstruction at this site represents the first of its kind for the Late Glacial in New Zealand, and provides a quantitative assessment of climate change for the period associated with the ACR.

South America

Pioneer chironomid work has been carried out in the Neotropics and three studies dealing with fossil midges have been published. Goulden (1966a,b) studied fossil midges from Laguna de Petenxil and another pond in Guatemala. Interpretations from both studies are vague, mainly due to the low abundance and poor preservation of the remains and the lack of an accurate chronology. In Venezuela, Binford (1982) carried out a paleolimnological investigation in Lake Valencia using animal remains and some chemical and physical variables to reconstruct the last 12.5 ka of lake history. Results of this work indicate that lake-level fluctuations and trophic changes occurred during the time span of the core. More recently, fossil chironomid investigations have been conducted in the southern Andes of Argentina and Chile (Ariztegui et al., 1997; Bianchi et al., 1997, 2000; Corley and Massafiero, 1998; Massafiero, 2000; Massafiero and Brooks, 2002; Massafiero and Corley, 1998; Massafiero et al., 2004, 2005, 2008, 2009). These studies were concentrated in two areas: in Northern Patagonia, Argentina, within the limits of the Nahuel Huapi National Park (41°S) and in Southern Chile on the Peninsula de Taitao and the Chonos Archipelago (46°S). Here, we review the main focus of these publications: paleoclimate, and paleo-productivity reconstructions.

Paleoclimatic reconstructions

Studies focusing mainly on Late Glacial climate reconstructions address evidence of Late Glacial Northern Hemisphere climate signals (i.e., YD cooling) in South America. To date, results of investigations at four sites using fossil chironomids to reconstruct past climate during the Late Glacial period in Argentinean and Chilean Patagonia have been published. In Ariztegui et al. (1997), changes in pollen, fossil chironomids, and geochemistry during the last 15 cal ka BP at Lake Mascardi (41°20'S, 71°34'W; Figure 3) were interpreted as a response to a climate cooling coincident with the timing of the YD

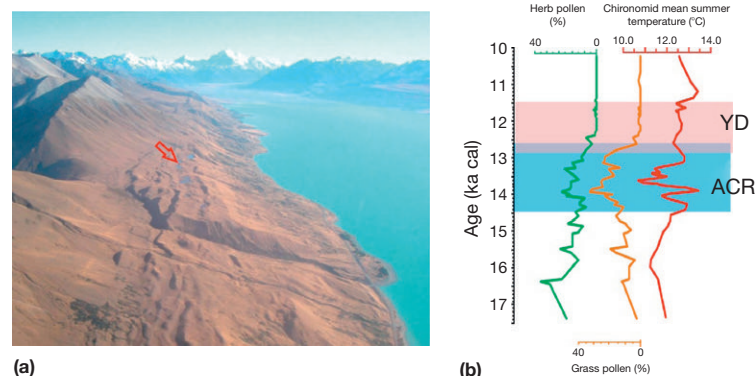


Figure 2 (a) Location of Boundary Stream Tarn, a kettle lake within lateral moraines of the Lake Pukaki Last Glacial Maximum ice advances. Aoraki (Mount Cook) in the background. (b) Late Glacial cooling expressed in pollen and chironomid proxies at Boundary Stream Tarn. Main period of cooling in both proxies coincides with the timing of the ACR and is 2–3 °C below modern (13.4 °C).



Figure 3 Eastern end of Lago Mascardi (800 m a.s.l., 30 km² surface area, 200 m water depth) located ~15 km from the Mt Tronador ice cap that feeds the lake. Glacial ice from Mt Tronador filled this valley during the Last Glacial Maximum. The fossil core site is located outside the photograph.

Northern Hemisphere cold event. Results showed that between 13.47 and 11.96 cal ka BP, there was a decrease in the total pollen influx as a consequence of an increase in inorganic sedimentation followed by a sharp decrease in the hydrogen index. These results were interpreted as a response to cooling accompanied by an increase in subglacial erosion due to an advance of the Tronador ice cap that feeds the proglacial lake. The disappearance of the warm-adapted *Chironomus* and the decline in the total chironomid abundance at this time also suggested a climatic deterioration.

In a multiproxy paleolimnological study carried out in Lake El Trebol (41°04'9"S, 71°30'9"W), Bianchi et al. (1997) analyzed fossil chironomids, pollen, and geochemistry to reconstruct paleoenvironmental history. Changes in these proxies were interpreted to indicate climatic fluctuations during the last 14 cal ka BP. The increase of fossil midges, organic matter, biogenic silica, and nutrients from ca. 12.42 cal ka BP was interpreted as a change in lake productivity due to climatic amelioration. Subsequently, Bianchi et al. (1997) combined pollen, chironomids, rare earth elements, heavy metals, and fossil pigments to show the postglacial evolution of the landscape and biota in the area surrounding Lake El Trebol, providing the basis for future accurate paleoenvironmental reconstructions. Based on changes in the analyzed proxies, three different periods were recognized: the Late Glacial, a transition, and the Holocene period. Chironomids showed a gradual replacement of littoral Orthocladinae fauna, abundant in the transition zone, by profundal Tanytarsini taxa, present in the Holocene, indicating more autochthonous lake production. Changes in pollen and paleopigments were consistent with changes in midge assemblages. These results suggested a climatic amelioration during the Late Glacial/Holocene transition. Climate interpretations were consistent with those observed at Lake Mascardi (Ariztegui et al., 1997). Chironomid assemblages and biogeochemical parameters also indicated changes in lake productivity during Holocene times. Nevertheless, the interpretation of climate changes in both

investigations remained speculative due to the lack of resolution and a precise chronology.

The first high-resolution late Quaternary environmental reconstruction using chironomids in South America was published by Massafiero and Brooks (2002). Changes in the chironomid assemblages at Laguna Stibnite (46°25'S, 74°24'W) in the Taitao Peninsula in Chile suggested that the climate in southern Chile was drier and at its coolest during the YD chronozone. Between ca. 13 and 11.2 cal ka BP, the cold-stenothermic Podonominae were consistently present in every sample, reaching a peak abundance during this period. This suggests that lake waters were cool and oligotrophic during the YD. The low concentration of head capsules also reflects low lake productivity. The dynamics of the late Quaternary chironomid fauna of Lake Stibnite shows a remarkably cyclical pattern, which was interpreted as being linked to changes in precipitation. During the Holocene, the presence of a *Limmophyes/Chironomus/Phaenopsectra/Parachironomus* assemblage between 10.6 and 7.2 cal ka BP and 2.5 and 1.5 cal ka BP suggested dry periods with low lake levels. Within a regional context, results from beetles from Puerto Edén (Ashworth et al., 1991) located 300 km south of Taitao Peninsula also suggested changes in precipitation during the Holocene, although they were almost out-of-phase to those inferred at Laguna Stibnite. If these interpretations are both correct, changes in precipitation patterns may be a consequence of latitudinal shifts of the Westerlies (Massafiero and Brooks, 2002). However, the climatic interpretation that suggested cooling during the YD is not in agreement with other investigations carried out in the lake. Pollen records from Lake Stibnite (McCulloch et al., 2000) and other lakes in this region of Chile show no evidence of cooling during the Late Glacial period (e.g., Haberle and Bennet, 2004). In contrast, other sites in Patagonia show changes in pollen assemblages at the end of the Late Glacial that have been interpreted as a response to climatic cooling (e.g., Heusser et al., 1996; Moreno, 1997). The inconsistencies between these interpretations have contributed to the ongoing controversy concerning the presence of a YD cooling in South America.

Massafiero et al. (2005) carried out a study on pollen and fossil chironomids in Laguna Facil (44°19'S, 74°17'W), located in the Chonos Archipelago in Chile. In this article, the authors argue that chironomids respond to local rather than regional environmental changes. Chironomid stratigraphical changes coincided closely but not synchronously with changes in the pollen and charcoal. The lag in the response of midges compared with the forest trees would suggest that the midge fauna responds to changes brought about by the migration and colonization of these trees in the lake catchment. This could be in response to changes in the structure of the forest canopy and amount of tree shade close to the shores of the lake, or to changes in water chemistry brought about by soil development. However, a change in the chironomid assemblage at ca. 7.2 cal ka BP, preceding the resurgence of *Pilgerodendron*, suggests a midge response to regional climatic change.

Despite the faunal similarities and the close proximity of Lake Facil to Lake Stibnite, no response to a cooling event coinciding with the YD was recorded in the chironomid assemblages from Lake Facil. The explanation for this (Massafiero et al., 2005) pointed to the more northerly location of Lake Facil than Laguna Stibnite, which is less likely to

have been influenced by any resurgence of Andean glaciers during the YD. Geographic variability in glacial activity would indicate that climatic conditions were possibly insufficiently intense and/or of insufficient duration to uniformly affect regional paleobiotic changes; therefore, a response during the YD may be recognizable at some sites but not at others.

Recently, [Massaferro et al. \(2009\)](#) combined pollen and fossil chironomid records from the Huelmo site, Chile ($41^{\circ}31'S$, $73^{\circ}00'W$) to elucidate the timing and structure of climate changes during the final portion of the Last Glacial Maximum and during the last glacial termination (LGT) in NW Patagonia. During the Late Glacial/Holocene transition, chironomid assemblages show cooling pulses at 14 and 13.5 cal ka BP ([Figure 4](#)). The first cooling was interpreted as a low-magnitude precursor event, contemporaneous with the ACR, giving support for the Huelmo Mascardi Cold Reversal (HMCR; [Hajdas et al., 2003](#)). This study provides one of the first detailed chironomid and pollen records in the Southern Hemisphere spanning the culmination of the Last Glacial Maximum through the LGT and into the Holocene. The results reported in this paper clearly demonstrate the sensitivity of fossil chironomids to temperature and moisture fluctuations during the last transition from glacial maximum to interglacial conditions in NW Patagonia.

Paleoproductivity reconstructions

A number of studies from South America focus on the relationships between chironomids and lake productivity. [Massaferro and Corley \(1998\)](#) studied the chironomid paleodiversity from Lake Mascardi. The study demonstrates how major environmental episodes such as volcanic eruptions, climatic cooling/warming, and changes in lake level occurring during the Late Glacial period affected midge diversity patterns. The authors suggest that long-term studies reconstructing the history of lake diversity provide important information for present-day conservation management. In conjunction with this study, [Corley and Massaferro \(1998\)](#) published the first fossil insect community dynamics study in Patagonia. Using a long-term apparent turnover rate, the authors show chironomid community

extinction/immigration patterns along a 10 m sediment core representing 15 ka of lake history. This kind of work is important from an ecological perspective to facilitate the interpretation of current trends in diversity and succession patterns. Subsequently, [Massaferro \(2000\)](#) published the complete chironomid assemblage from Lake Mascardi, covering the last 15 ka BP. In this study, changes in chironomid species abundance are compared to diversity indices and also to changes in total organic carbon. Sampling resolution was not sufficiently high to discern short-term chironomid responses to environmental change; however, changes in midge assemblages were interpreted to be related to temperature and also to changes in lake productivity. [Massaferro et al. \(2004\)](#) studied geochemical and chironomid records from a short core from Lake Morenito ($41^{\circ}39'S$, $71^{\circ}31'W$), located near the city of San Carlos de Bariloche in Patagonia. Due to the expansion of the population in recent years, Lake Morenito and surrounding areas have experienced moderate human impact such as the deforestation, fish introduction, and building construction. This work shows the response of chironomids and geochemistry to natural and anthropogenic (European) environmental disturbances during the last 200 years. The main fluctuations recorded in chironomid assemblages during the time span of the core were interpreted as a response to human disturbances in the catchment area, demonstrating the value of chironomids in reconstructing paleoproductivity. It also demonstrates the utility of combining multiple proxies (chironomids, tephra and sediment analysis, organic matter, and chronology) in the interpretation of such changes.

Hemispheric and Interhemispheric Comparison

Although more high-resolution paleotemperature studies are warranted for the Southern Hemisphere, some tentative comparisons with other regions of the globe can be made from those that have already addressed the Late Glacial–interglacial climate change in the mid-latitudes of the Southern Hemisphere. This topic has been extremely controversial in the

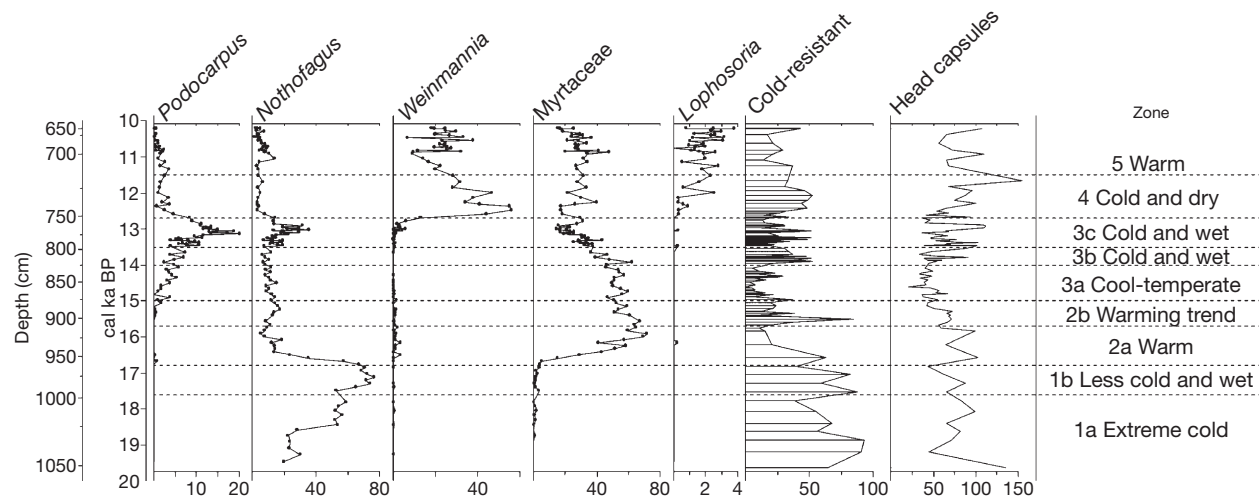


Figure 4 Key pollen indicator taxa and cold-tolerant chironomid taxa from the Huelmo site. The Huelmo Mascardi Cold Reversal encompasses zones 3c and 4.

past, with some paleoclimate records indicating changes that are more coincident with cooling noted in Antarctic ice-core records, that is, the ACR, than with cooling noted from the North Atlantic region (associated with the YD chronozone). Other records have suggested gradual warming with no reversals in the Southern Hemisphere. Studies from lacustrine environments of New Zealand, Tasmania, and South America can help resolve this debate. Chironomid records from Tasmania do not provide clear evidence of an ACR or YD cooling during the Late Glacial while in New Zealand there seems to be clearer evidence of a 2–3 °C cooling during the time of the ACR. Chironomid records suggest that a distinct climatic reversal similar to the YD cool phase of the North Atlantic region occurred in sites from Central Chile and Argentina. This mid-latitude cooling was recently identified as the HMCR (Hajdas et al., 2003). Nevertheless, chironomid assemblages also indicate that a cooling trend started earlier than the HMCR, at ca. 14.4 cal ka BP. This cooling corresponds with the timing of the ACR (Massaferro et al., 2009). Comparison of chironomid results with both EPICA Antarctica and GISP2 Greenland ice-core records indicates similar Late Glacial climatic events, but their full interpretation awaits the development of a robust inference model (South America) and the development of more fossil chironomid records from sites across the Southern Hemisphere.

From this review it is clear that much progress have been made in the Australasian region in the use of chironomids as quantitative indicators of past conditions. Major achievements have occurred in New Zealand (Dieffenbacher-Krall et al., 2007; Vandergoes et al., 2008; Woodward and Shulmeister, 2006a,b, 2007) and Tasmania (Rees et al., 2008, Rees and Cwynar, 2010a,b). Identification keys for the modern chironomid fauna continue to be developed or New Zealand (Boothroyd, personal communication), and for the subfossil remains of New Zealand (Dieffenbacher-Krall et al., 2008 online), Tasmania (Cwynar and Rees, personal communication), and South America (Massaferro, personal communication).

Future Research Directions

The fact that chironomids have the potential to provide detailed paleotemperature reconstructions is extremely important in the Southern Hemisphere where quantitative estimates of temperature change are limited. Combined with other proxies including pollen, diatoms, and isotopic measurements, chironomids can contribute substantially to multiproxy environmental reconstructions to produce more robust estimates of climate change. They can also be used to calibrate regional paleotemperature estimates derived from nonecological indicators such as snowline reconstructions and ice cores. The importance of attaining quantitative temperature estimates from landmasses within the Southern Hemisphere is critical to clarify climate relationships between the Northern and Southern Hemispheres, and therefore identify possible coupling mechanisms. Much progress has been made in the study of fossil midges from the Southern Hemisphere; however, further work is required in two main areas: chironomid taxonomy and ecology and quantitative models.

Modern Taxonomy and Ecology

Early studies by Edwards (1931) and Brundin (1966) were the first to recognize a circum-Antarctic chironomid faunal connection and make significant contributions to the knowledge of Southern Hemisphere chironomid taxonomy and biogeography. Since then, other researchers from New Zealand (e.g., Boothroyd, 2002; Forsyth, 1971; Sublette and Wirth, 1980) and South America (e.g., Andersen and Contreras-Ramos, 1999; Cranston, 2000b; Fitkau and Reiss, 1973; Paggi, 1998) have made notable advances in furthering our knowledge of the taxonomy and distribution of Southern Hemisphere chironomids, while work by Cranston (2000a) has made an exceptional contribution to the identification and ecology of the Australian chironomid larval fauna. Even with these recent advances, knowledge of chironomid taxonomy of Australia, New Zealand, and South America remains limited. The number of species currently documented is considered conservative and many taxa remain undescribed (Boothroyd, personal communication; Cranston, 2000b). A further problem is that the existing taxonomic descriptions are based on characters found in adults or pupae, while the taxonomy of the larval stages has largely been ignored. As a consequence, in New Zealand and South America, many of the existing taxa can only be identified to the generic level (Figure 5) and identification keys for the larval stages of chironomids in these regions are scarce. Many chironomid taxa remain undescribed because of a lack of ecological investigations. The result is that the knowledge of ecological requirements, tolerances, and species distribution of these insects in Australia, New Zealand, and South America is incomplete. These factors hamper our understanding of the causes of change in modern chironomid assemblages and inhibit the use of fossil chironomid assemblages, which relies solely on the identification of subfossil larval head capsules for environmental reconstructions. There is a need for the improvement of taxonomic resolution of modern and subfossil chironomids. In the Southern Hemisphere, there is just one key for the identification of the chironomid fauna from Australia (Cranston, 2000a), but similar keys have yet to be published for New Zealand or South America. For the first time, Massaferro and Brooks (2002) described specimens of the subfossil chironomid fauna from a Chilean lake and identified additional taxonomic groups that could be ecologically related to the European taxa. However, more taxonomic and ecological work is needed in the Southern Hemisphere, especially regarding the distribution and habitat requirements of chironomids. A standardized taxonomic approach, together with an improvement in taxonomic resolution, will allow direct comparisons between chironomid investigations from Australia, New Zealand, South America, and Africa, providing an invaluable tool for biogeographical and paleolimnological research.

Chironomid–Temperature Inference Models

Quantitative estimates of paleotemperature can only be achieved through the development of chironomid–temperature inference models specific to the geographic region of the

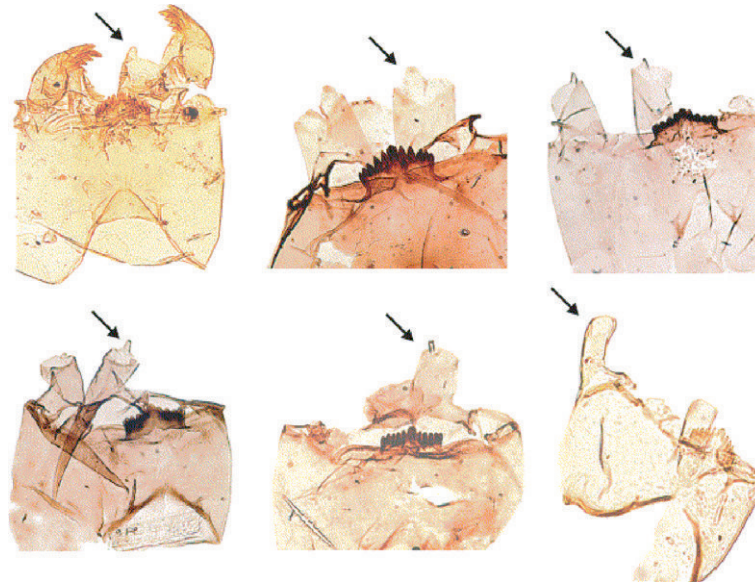


Figure 5 Head capsule remains of South American *Tanytarsus* sp. yet to be classified below generic level. Distinction of subgroups based on the differing characteristics of the spur on the antenna pedestal (arrows). Photographs by J. Massafiero.

paleoclimate analysis. In the Northern Hemisphere, a number of studies have used the training set approach to determine how environmental variables, including summer temperature, are important in determining chironomid community composition. The recent development of inference models from Tasmania (Rees et al., 2008; Rees and Cwynar, 2010a,b), New Zealand (Dieffenbacher-Krall et al., 2007; Woodward and Shulmeister, 2006a,b) represents a significant advance in the use of chironomids as paleoenvironmental indicators. However, improvements to this initial calibration data set, including a more refined taxonomy and the addition of more lakes evenly distributed along a wider temperature gradient, are needed to improve the performance of these models. At present, the lack of chironomid–temperature inference models from South America is a serious impediment in providing robust, quantifiable estimates of temperature from existing Late Glacial and Holocene chironomid records. Progress in this area is underway and Massafiero (2010) presented the results of a training set of 60 chironomid taxa, 6 environmental variables and 102 lakes from Patagonia (Argentina and Chile) located along a transect from 40 to 54° S from southern South America. Multivariate analysis indicates that temperature (December/January mean air temperature) and mean annual precipitation explains the greatest amount of variance in chironomid distribution of any of the measured variables. However, the influence of low head capsule abundance in ultraoligotrophic lakes from Patagonia currently limits the performance of the transfer functions and work is currently underway to refine these models.

See also: **Beetle Records:** Late Pleistocene of Europe. **Chironomid Records:** Chironomid Overview. **Introduction:** History of Quaternary Science. **Vertebrate Records:** Late Pleistocene of South America.

References

- Andersen T and Contreras-Ramos A (1999) First record of *Antillocladius* Sæther from continental South America (Chironomidae, Orthocladinae). *Acta Zoologica Academiae Scientiarum Hungaricae* 45(2): 149–154.
- Ariztegui D, Bianchi MM, Massafiero J, Lafargue E, and Niessen F (1997) Interhemispheric synchrony of Late-glacial climatic instability as recorded in proglacial Lake Mascardi, Argentina. *Journal of Quaternary Science* 12: 333–338.
- Ashworth AC, Markgraf V, and Villagran C (1991) Late Quaternary climatic history of the Chilean Channels based on fossil pollen and beetle analysis, with an analysis of the modern vegetation and pollen rain. *Journal of Quaternary Science* 6: 279–291.
- Bianchi MM, Massafiero J, Roman Ross G, del Valle R, Tatur A, and Amos A (1997) The Pleistocene–Holocene boundary from cores of Lake El Trébol, Patagonia Argentina. *Archiv für Limnologie* 26: 805–808.
- Bianchi MM, Massafiero J, Roman Ross G, and Lami A (2000) Geochemical and biological tracers of climate fluctuations during Late Pleistocene and Holocene in Lake El Trébol, Northern Patagonia, Argentina. *Journal of Paleolimnology* 22: 137–148.
- Binford MW (1982) Ecological history of Lake Valencia, Venezuela: Interpretation of animal microfossils and some chemical, physical and geological features. *Ecological Monographs* 52: 307–333.
- Boothroyd IKG (2002) *Cricotopus* and *Paratricocladius* (Chironomidae: Insecta) in New Zealand, with description of *C. hollyfordensis* n. sp., and redescription of adult and immature stages of *C. zealandicus* and *P. plurisetialis*. *New Zealand Journal of Marine and Freshwater Research* 36: 775–788.
- Brundin L (1966) Transantarctic relationships and their significance, as evidenced by chironomid midges, with a monograph of the subfamilies Podonominae and Aphroteniinae and the austral Heptagyiinae. *Kungl Svenska Vetenskapsakademiens Handlingar* 11: 1–472.
- Corley JC and Massafiero JI (1998) Long term turnover of a fossil community of chironomids (Diptera) from Lake Mascardi (Patagonia, Argentina). *Journal of the Kansas Entomological Society* 71: 407–413.
- Cranston PS (2000a) An electronic guide to the Chironomidae of Australia. <http://entomology.ucdavis.edu/chiropage/index.html>.
- Cranston PS (2000b) *Parapsectrocladius*: A new genus of Orthocladinae Chironomidae (Diptera) from Patagonia, the southern Andes. *Insect Systematics and Evolution* 31(1): 103–120.
- Deevey ES (1955) Paleolimnology of the upper swamp deposit, Pyramid Valley. *Records of the Canterbury Museum* 6: 291–344.
- Dieffenbacher-Krall AC, Vandergoes MJ, and Denton GH (2007) An inference model for mean summer air temperatures in the Southern South Island, New Zealand. *Quaternary Reviews* 26: 2487–2504.

- Dieffenbacher-Krall AC, Vandergoes MJ, Woodward CA, and Boothroyd IKG (2008) *Guide to Identification and Ecology of New Zealand Subfossil Chironomids Found in Lake Sediment*. Orono, ME: University of Maine, Climate Change Institute <http://www.climatechange.umaine.edu/Research/facilities/perl/nzguide.html>.
- Dimitriadis S and Cranston PS (2001) An Australian Holocene climate construction using Chironomidae from a tropical volcanic maar lake. *Palaeogeography, Palaeoclimatology, Palaeoecology* 176: 109–131.
- Edwards FW (1931) Fascicle 5 – Chironomidae. *Diptera of Patagonia and South Chile*, vol. 2, pp. 233–331. London: Natural History Museum.
- Fitkau EJ and Reiss F (1973) Amazonische Tanytarsini (Chironomidae Diptera). I. Die riopreto Gruppe der Gattung Tanytarsus. *Studies in Neotropical Fauna* 8: 1–16.
- Forsyth DJ (1971) Some New Zealand Chironomidae. *Journal of the Royal Society of New Zealand* 1: 113–144.
- Goulden CE (1966a) La Aguada de Santa Ana Vieja: An interpretive study of the cladoceran microfossils. *Archiv für Hydrobiologie* 62: 373–404.
- Goulden CE (1966b) The animal microfossils of Laguna Petenxil, Guatemala. *Memoirs of the Connecticut Academy of Arts and Sciences* 17: 84–120.
- Haberle SG and Bennet KD (2004) Postglacial formation and dynamics of North Patagonian rainforest in the Chonos Archipelago, Southern Chile. *Quaternary Science Reviews* 23: 2433–2452.
- Hajdas I, Bonani G, Moreno P, and Ariztegui D (2003) Precise radiocarbon dating of Late-Glacial cooling in mid-latitude South America. *Quaternary Research* 59: 70–78.
- Heusser CJ, Lowell TV, Heusser LE, Hauser A, Andersen BG, and Denton GH (1996) Full-glacial–late-glacial palaeoclimate of the Southern Andes: Evidence from pollen, beetle, and glacial records. *Journal of Quaternary Science* 11: 173–184.
- Massaferro J (2000) Fossil assemblages in an oligotrophic lake of Patagonia (Lago Mascaradi, Argentina). In: Hoffrichter O (ed.) *Late 20th Century Research on Chironomidae*, pp. 127–131. Aachen, Germany: Shaker Verlag.
- Massaferro J (2010) A chironomid temperature inference model to interpret patterns of climate changes in southern South America. *Proceedings VI Southern Connection Congress Bariloche, Argentina, Febrero 15–19*.
- Massaferro J, Ashworth A, and Brooks S (2008) Quaternary fossil insects from South America. In: Rabassa J (ed.) *The Late Cenozoic of Patagonia and Tierra del Fuego. Developments on Quaternary Sciences*, pp. 393–409. Amsterdam: Elsevier.
- Massaferro J and Brooks SJ (2002) The response of Chironomids to Late Quaternary environmental change in the Taitao Peninsula, Southern Chile. *Journal of Quaternary Sciences* 17(2): 101–111.
- Massaferro J, Brooks SJ, and Haberle SG (2005) The dynamics of chironomid assemblages and vegetation during the Late Quaternary at Laguna Facil, Chonos Archipelago, southern Chile. *Quaternary Science Reviews* 24: 2510–2522.
- Massaferro J and Corley J (1998) Environmental disturbance and chironomid paleodiversity: 15 kyr BP of history at Lake Mascaradi (Patagonia, Argentina). *Aquatic Conservation: Marine and Freshwater Ecosystem* 8: 315–323.
- Massaferro J, Moreno PI, Denton GH, Vandergoes M, and Dieffenbacher-Krall A (2009) Chironomid and pollen evidence for climate fluctuations during the Last Glacial Termination in NW Patagonia. *Quaternary Science Reviews* 28: 517–525.
- Massaferro J, Ribeiro Guevara S, Rizzo A, and Arribere M (2004) Short term environmental changes in Lake Morenito, (41°S, Patagonia, Argentina) from the analysis of subfossil chironomids. *Aquatic Conservation: Marine and Freshwater Ecosystems* 14: 123–134.
- McCulloch RD, Bentley MJ, Purves RS, et al. (2000) Climatic inferences from glacial and palaeoecological evidence at the last glacial termination, southern South America. *Journal of Quaternary Science* 15: 409–417.
- Moreno PI (1997) Vegetation and climate near Lago Llanquihue in the Chilean lake district between 20,200 and 9,500 14C yr BP. *Journal of Quaternary Sciences* 12: 485–500.
- Paggi AC (1998) Chironomidae. In: Morrone JJ and Coscarón S (eds.) *Biodiversidad de Artrópodos argentinos. Una perspectiva biotaxonomica*, ch. 31. pp. 327–337. La Plata, Argentina: Ediciones Sur.
- Paterson CG and Walker KF (1974) Recent history of *Tanytarsus barbitarsus* Freeman (Diptera: Chironomidae) in the sediments of a shallow saline lake. *Australian Journal of Marine and Freshwater Research* 25: 151–165.
- Rees ABH and Cwynar L (2010a) A test of Tyler's Line – Response of chironomids to a pH gradient in Tasmania and their potential as a proxy to infer past changes in pH. *Freshwater Biology* 55(12): 1–20.
- Rees ABH and Cwynar L (2010b) Evidence for early postglacial warming in Mount Field National Park, Tasmania. *Quaternary Science Reviews* 29: 443–454.
- Rees ABH, Cwynar L, and Cranston PS (2008) Midges (Chironomidae, Ceratopogonidae, Chaoboridae) as a temperature proxy: A training set from Tasmania, Australia. *Journal of Paleolimnology* 40: 1159–1178.
- Schakau BL (1986) Preliminary study of the development of the subfossil chironomid fauna (Diptera) of Lake Taylor, South Island, New Zealand, during the younger Holocene. *Hydrobiologia* 43: 287–291.
- Schakau BL (1990) Stratigraphy of the fossil Chironomidae (Diptera) from Lake Grasmere, South Island, New Zealand, during the last 6000 years. *Hydrobiologia* 214: 213–221.
- Sublette JE and Wirth WW (1980) The Chironomidae and Ceratopogonae (Diptera) of New Zealand's subantarctic islands. *New Zealand Journal of Zoology* 7: 299–378.
- Vandergoes MJ, Dieffenbacher-Krall AC, Newnham RM, Denton GH, and Blaauw M (2008) Cooling and changing seasonality in the Southern Alps, New Zealand during the Antarctic Cold Reversal. *Quaternary Science Reviews* 27: 589–601.
- Verschuren D and Eggermont H (2006) Quaternary paleoecology of aquatic Diptera in tropical and Southern Hemisphere regions, with special reference to the Chironomidae. *Quaternary Science Reviews* 25(15–16): 1926–1947.
- Walker IR (2001) Midges: Chironomidae and related diptera. In: Smol JP, Birks HJB, and Last WM (eds.) *Tracking Environmental Change Using Lake Sediments. Zoological Indicators*, vol. 4, pp. 43–67. Dordrecht: Kluwer Academic.
- Woodward CA and Shulmeister J (2005) A Holocene record of human induced and natural environmental change from Lake Forsyth (Te Wairewa), New Zealand. *Journal of Paleolimnology* 34(4): 481–501.
- Woodward CA and Shulmeister J (2006a) New Zealand chironomids as proxies for human-induced and natural change: Transfer functions for temperature and lake production (chlorophyll *a*). *Journal of Paleolimnology* 36(4): 381–501.
- Woodward CA and Shulmeister J (2006b) New Zealand chironomids as proxies for human-induced and natural environmental change: Transfer functions for temperature and lake production (chlorophyll *a*). *Journal of Paleolimnology* 36(4): 407–429.
- Woodward CA and Shulmeister J (2007) Chironomid-based reconstructions of summer air temperature from lake deposits in Lyndon Stream, New Zealand spanning the MIS 3/2 transition. *Quaternary Science Reviews* 26: 142–154.

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Cladocera Studies *see* **Paleolimnology**: Cladocera

Climate Change *see* **Paleoclimate**: Introduction; Timescales of Climate Change; **Paleoclimate Modeling**: Data–Model Comparisons; Quaternary Environments; The Last Interglacial; Last Glacial Maximum GCMs; Paleo-ENSO; **Paleoclimate Reconstruction**: Approaches; Pliocene Environments; Paleodroughts and Society; Sub-Milankovitch (DO/Heinrich) Events; Paleotempestology; Younger Dryas Oscillation, Global Evidence; The Last Millennium; Historical Climatology; Paleoclimate Relevance to Global Warming

Climate Modeling, Quaternary *see* **Paleoclimate Modeling**: Quaternary Environments

Coccolithophore Studies *see* **Paleoceanography, Biological Proxies**: Alkenone Paleothermometry Based on the Haptophyte Algae; Coccolithophores

Coleoptera Fossil Records *see* **Beetle Records**: Overview; Late Tertiary and Early Quaternary Records; Middle Pleistocene of Western Europe; Late Pleistocene of Australia; Late Pleistocene of Europe; Late Pleistocene of Japan; Late Pleistocene of North America; Late Pleistocene of South America; Late Pleistocene of New Zealand; Late Pleistocene of Northern Asia; Postglacial Europe; Postglacial North America

Coral Studies *see* **Paleoceanography, Biological Proxies**: Corals, Sclerosponges and Mollusks; **Sea Level Studies**: Coral Records of Relative Sea-Level Changes

COSMOGENIC NUCLIDE DATING

Contents

Overview

Methods

Cosmic Ray Interactions in Minerals

Exposure Geochronology

Landscape Evolution

Overview

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Introduction

The advent of atmospheric cosmogenic radiocarbon dating by Libby in the 1950s brought high precision ages of important

geological and archaeological events and solutions to long-standing marine and terrestrial Quaternary problems. Over the next three decades, methods were developed that could measure cosmogenic isotopes in minerals at the precision

necessary for surface exposure dating. However, since the mid-1980s, there has been a comparable revolution in Quaternary science due to applications of terrestrial cosmogenic nuclides (TCN) produced *in situ*. Many previously intractable geochronology questions are being addressed, many new questions are being asked, and there has been an explosion of ideas related to rates of surface processes.

Synthesis of TCN applications

Recent reviews of applications of TCN methods are provided by Bierman and Nichols (2004), Cockburn and Summerfield (2004), T. J. Dunai (2010), Gosse and Phillips (2001), Muzikar et al. (2003), and Niedermann (2002). To date, TCN methods comprise six routinely-utilized isotope-mineral systems: ^3He , ^{21}Ne , ^{10}Be , ^{26}Al , ^{36}Cl , and ^{14}C . The former two are stable noble gases and the latter four are listed in order of decreasing half lives – 1.5 Ma (currently debated), 0.7 Ma, 0.3 Ma, and 5.7 ka. Some TCNs are calibrated for specific minerals due to:

- (1) diffusion constraints (e.g., ^3He diffuses too quickly from quartz to make it useful as a geochronology tool, but is retained in minerals such as pyroxenes and olivine);
- (2) production rates (^{36}Cl cannot be produced in pure quartz); and
- (3) other issues, such as high native concentrations of the isotope or its comparative isotope for AMS measurement (for instance, measurements of ^{26}Al in an aluminosilicate are difficult).

So far, one isotope has been calibrated for measurement in whole rock samples (^{36}Cl), so virtually any rock type can be used for TCN research. The stability of the noble gases permits the measurements of the oldest surfaces on Earth (tens of millions of years). An advantage of the radionuclides is that their concentration is dependent on decay rate as well as exposure duration, offering an extra degree of freedom for calculating long exposure durations or avoiding difficulties associated with inheritance. For instance, the rapid decay of ^{14}C offers a useful geochronology tool to date glaciated surfaces (or sediment derived from them) that have not been deeply eroded. Those surfaces would retain some TCN concentration from exposure prior to ice cover. However, if the ice cover was of sufficient duration (say 40 ka), most of the ^{14}C would be decayed and its concentration measured today would reflect the timing of exposure since the surface was uncovered. The youngest and still experimental member of the TCN toolbox is the noble gas ^{38}Ar . However, other new isotope and mineral systems are being developed to provide more applications (see **Methods**).

Every TCN experiment depends on the past century (Gosse and Phillips, 2001) of efforts of theoretical and experimental physics, analytical chemistry, geochemistry, geomorphology, and geophysics. Sample selection must consider geometry effects and the angular dependence of cosmic radiation. Sample preparation requires planning to ensure the proper isotope-mineral system is selected based on lithology and mineral abundance. To achieve the desired precision, the analytical limits of the isotope mass spectrometry, exposure duration, site production rate and sample depth, mass of sample,

number of samples per site, abundance of non-cosmogenic species of the same isotope, and complexity of exposure history, must all be considered. The analyses of these nuclides require high precision atomic-grade noble gas mass spectrometers or accelerator mass spectrometers.

Besides physical improvements to both of these systems, spectrometry methods and target preparation procedures have improved significantly over the past decade (both techniques routinely analyze TCN samples with 2 to 4% 1σ precision). Finally, when a TCN method is applied, numerous calculations and sensitivity assessments are required before interpreting the concentration as an age or an erosion rate. Cosmic ray physics and empirical measurements must be relied upon to scale production rates from sea level high latitude to the sample site. Decisions must be made regarding which production rate to use, and whether to compensate for geomagnetic field effects, snow cover, bulk density variations with time, and other factors. The intensity of sample specific calculations and analyses is a significant departure from most Quaternary dating methods, which often only require a thorough consideration of stratigraphic context before the data can be used.

A major premise of all TCN methods is that the nuclides are produced at known rates. Production of cosmogenic isotopes in minerals exposed to secondary radiation depends on the nuclide and mineral system and other environmental factors. Secondary radiation is produced when primary galactic cosmic rays (primarily fast protons) interact with nuclei in the Earth's atmosphere or within the upper few meters of the lithosphere. TCN production rates vary spatially as a function of geomagnetic field effects (dipolar and non-dipolar), and atmospheric depth. TCN production also varies temporally, in a way similar to atmospheric radiocarbon production variation under stronger or weaker magnetic field intensities. Additionally, the concentration of TCNs in rock depends on erosion, movement, or burial of the surface. Production rates for most locations are probably known to within 15% at the 1σ confidence level. However, the uncertainty is not well defined due to the number of variables involved and the difficulty in establishing a highly accurate production rate for a given time and place. The principal objective of two concurrent major international research programs, CRONUS-Earth in North America and CRONUS-EU in Europe, was to improve knowledge of TCN production rates. Results from these studies are in the process of publication, however, results from the intercomparison of different sites in the CRONUS program indicates that the level of uncertainty in production rates can be reduced to a few percent for ^{10}Be , but substantial challenges remain for other radionuclides.

TCN methods can be used to directly date rock or sediment landforms. By 2006, hundreds of lava flows and thousands of boulders on moraines have been dated directly with TCN measurements. Exposure of surfaces in hyper-arid regions, such as the Antarctic dry valleys and the Atacama Desert (Dunai et al., 2005) record exposure histories since the Pliocene and Miocene. The upper time range of the method is limited by surface stability although for long (hundred million years) exposures, changes in the diffusion of the noble gases and the production of non-cosmogenic species can become the limiting factors.

Secondary cosmic radiation (primarily fast, epithermal, and thermal neutrons and muons) is attenuated with depth below the surface at a rate dependent on the particle attributes (type, charge, energy) and the bulk density of the rock or sediment. Production of TCN isotopes is routinely detected at depths exceeding 5 m. The attenuation of the cosmic ray secondaries with mass has many useful implications. For instance, the differences in the decrease in total production rate of ^3He and ^{38}Ar with atmospheric depth may lead to future uses of their ratio for paleo-altimetry (surface uplift) measurements. Furthermore, the predictable decrease in concentration with depth creates an important advantage for TCN over other geochronology methods. Because the concentration is sensitive to depth, the method can be used to determine directly the rate and style of erosion of landforms and landscapes. A third implication of the penetration of cosmic rays through mass is that the definition of 'surface' can include surface evolution, so that it is possible to yield a minimum exposure age on an eroding surface, although the original surface may have been changed or completely stripped.

TCN exposure methods have already made lasting impacts on society and science outside Quaternary geology. For instance, Partridge et al. (2003) have provided $^{26}\text{Al}/^{10}\text{Be}$ burial ages on sediment at Sterkfontein, South Africa, showing that associated hominid fossils were deposited in the Lower Pliocene. The capacity to date surfaces with millions of years of exposure has provided an important extension of the billion-to million-year timescale of rock exhumation history derived from high- and low-temperature thermochronology methods, such as apatite (U-Th)/He.

CRONUS Calculator

An important development as a result of the many studies conducted over the last decade has been the development of a calculator for cosmogenic nuclide studies, so that standardized computations of exposure age and erosion rates can be estimated for ^{10}Be and ^{26}Al measurements (Balco et al. 2008). This calculator can be found at the web address hess.ess.washington.edu.

Preface

The purpose of the following four articles is to provide a review of the current state of knowledge concerning:

- (1) the theory of cosmic ray interactions with minerals;
- (2) recent developments in production systematics and analytical and geochemical methodologies;
- (3) applications of the technique for geochronology; and

- (4) applications of the technique for landscape evolution studies.

Lal (see [Cosmic Ray Interactions in Minerals](#)) provides a systematic review of the important interactions that produce TCN. The article describes the origin, composition, and energy spectra of cosmic rays, and their fate from origin to Earth's surface. Lal provides a summary of the founding physics theory and experimentation upon which the TCN technique is based, and discusses some of the shortcomings of knowledge that remain as important sources of systematic error in the methods. Schaefer and Lifton (see [Methods](#)) provide a thorough review of the recent progress that has been made in TCN production systematics and the iso- topic procedures. They highlight one of many exciting strategies undertaken by the international CRONUS initiative to improve TCN techniques. Ivy-Ochs and Kober (see [Exposure Geochronology](#)) summarize a wide range of applications of TCN methods for dating surfaces and events, and include caveats and limitations to different approaches. Granger (see [Landscape Evolution](#)) demonstrates how the TCN methods offer more than geochronology, because their concentrations can be interpreted directly as erosion rates and burial histories to provide insights into landscape evolution that could otherwise not be resolved.

See also: [Cosmogenic Nuclide Dating: Cosmic Ray Interactions in Minerals](#); [Exposure Geochronology](#); [Landscape Evolution](#); [Methods](#).

References

- Balco G, Stone JO, Lifton NA, and Dunai TJ (2008) A complete and easily accessible means of calculating surface exposure ages or erosion rates from ^{10}Be and ^{26}Al measurements. *Quaternary Geochronology* 3: 174–195.
- Bierman P and Nichols KK (2004) Rock to sediment – slope to sea with ^{10}Be -rates of landscape change. *Annual Reviews of Earth Planetary Science* 32: 215–255.
- Cockburn HAP and Summerfield MA (2004) Geomorphological applications of cosmogenic isotope analysis. *Progress in Physical Geography* 28: 1–42.
- Dunai TJ (2010) *Cosmogenic Nuclides: Principles, Concepts and Applications in the Earth Surface Sciences*. Cambridge: Cambridge University Press.
- Dunai TJ, Gonzales Lopez GA, and Juez-Larre J (2005) Oligocene-Miocene age of aridity in the Atacama Desert revealed by exposure dating of erosion-sensitive landforms. *Geology* 33: 321–324.
- Gosse JC and Phillips FM (2001) Terrestrial *in situ* cosmogenic nuclides: Theory and application. *Quaternary Science Reviews* 20: 1475–1560.
- Muzikar P, Elmore D, and Granger D (2003) Accelerator Mass Spectrometry in Geologic Research. *Geological Society of America Bulletin* 115: 643–654.
- Niedermann S (2002) Cosmic-ray-produced noble gases in terrestrial rocks: Dating tools for surface processes. In: Porcelli D, Ballentine CJ, and Wieler R (eds.) *Noble Gases in Geochemistry and Cosmochemistry: Reviews in Mineralogy and Geochemistry*, vol. 47, pp. 731–784.
- Partridge TC, Granger DE, Caffee MW, and Clarke RJ (2003) Lower Pliocene hominid remains from Sterkfontein. *Science* 300: 607–612.

Methods

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Introduction

During the last 10–15 years, some 50 years after their initial discovery (Schaeffer and Davis, 1956), terrestrial cosmogenic nuclide (TCN) applications have had a revolutionary impact on various Earth science disciplines including Quaternary geology, geomorphology, sedimentology, volcanology, and paleoclimatology. The rapid progress has been fueled by recent methodological developments in TCN science, creating a quantitative foundation for studying a broad range of sedimentary processes on the Earth's surface. Simultaneously, it has become clear that the production systematics of TCNs on Earth are complex and are influenced by many parameters, known to varying degrees, all of which need to be taken into account for accurate interpretations in TCN applications. TCN research at the beginning of the twenty-first century is driven by the enormous potential of this tool in a rapidly growing number of applications in Earth science and beyond, but at the same time the effects of numerous uncertainties must be considered to avoid overinterpretation of TCN measurements.

A primary application of TCNs is the field of surface exposure dating (SED). The basic principle of SED is simple: secondary cosmic rays (mostly high-energy neutrons) interact with target atoms in rocks and sediments on the Earth's surface, producing cosmogenic nuclides *in situ* (dominantly via spallation reactions). Several of these *in situ* cosmogenic nuclides can be measured in excess of the natural background. If the rate of production per unit time is known, if the exposure of the surface has been continuous, and if the surface to be sampled has not been eroded, the measured cosmogenic nuclide concentrations can be directly translated into an exposure age. In nature, the application of TCNs is more complex because of the uncertainties in the quantitative knowledge of production rates and because of several natural processes modifying the surface with time (e.g., erosion) or interrupting the exposure (e.g., sediment/snow cover).

There are various recent review articles summarizing SED methods. The comprehensive paper by Gosse and Phillips (2001) presented the first complete overview of TCN science including overall uncertainty discussions. More recent reviews were provided by Blard et al. (2006) and Ivy-Ochs and Kober (2010). Balco (2011) reviews the potential of SED on glacial chronology. Niedermann (2002) reviews in detail the cosmogenic noble gas method and presents an overview of modern applications. The reviews by Cerling and Craig (1994), Kurz and Brook (1994), and Lal (1991) are already introductory classics, presenting the basic principles of the method and outlining the immense potential of the SED method. The rapid progress in the cosmogenic dating field is arguably best reflected in the breadth and depth of innovation reported since these classic reviews. Many of these advances have resulted both directly and indirectly from a major international

collaborative effort from 2004 to 2011 to better define the systematics of TCN production – the Cosmic-Ray Produced Nuclide Systematics on Earth (CRONUS-Earth, www.physics.purdue.edu/cronus) project, and its sister project from Europe (CRONUS-EU, www.cronus-eu.net/). This very recent progress is the focus of this short article. The new findings are presented and discussed in two different sections, namely, nuclide-specific news and production-rate novelties.

TCN-Specific Progress

One of the major areas of recent progress is the increase in sensitivity of cosmogenic nuclide methods, including extension of the exposure age timescale into the historical period, and the broadening of the target minerals that can be used for SED. The recent developments for each TCN are briefly discussed below.

High-Sensitivity ^{10}Be Analysis from Quartz

In situ ^{10}Be produced in quartz has become the workhorse of SED in the last several years, evidenced by a dramatic increase in published ^{10}Be studies. A few recent studies introduced progress in ^{10}Be SED by dating historical moraines as young as a few hundred years and resolving up to five individual glacier culminations during the last millennium, making the case for centennial resolution (Licciardi et al., 2009; Schaefer et al., 2009; Stone et al., 2003b). The internal consistency of ^{10}Be ages of boulders from one moraine also shows that the problem of inherited nuclides in SED samples can be close to zero in some cases (Figure 1).

^3He from Olivine, Pyroxene, Fe/Ti-Oxide Minerals, and Garnet

^3He is a TCN that is widely used as a result of the well-calibrated production rate (Goehring et al., 2010; Niedermann, 2002) and the wide applicable time range from 1 ka (e.g., Gayer et al., 2004) to more than 10 My (Schäfer et al., 1999). Most studies assume 100% spallation from high-energy nucleons as the production pathway for cosmogenic noble gases (i.e., assuming muogenic and thermal neutron production to be negligible). Cosmogenic helium is routinely analyzed from pyroxene and olivine (e.g., Kurz, 1986; Licciardi et al., 2001). Some success in measuring cosmogenic helium from quartz was reported early from cold and dry Antarctica (Brook and Kurz, 1993), but most studies since have found that quartz does not quantitatively retain cosmogenic ^3He (Niedermann, 2002 and references therein). One of the challenges is to separate the cosmogenic helium from noncosmogenic helium (e.g., mantle helium) trapped in the mineral, but several studies have shown convincingly that isotopic analysis of the helium released by cold

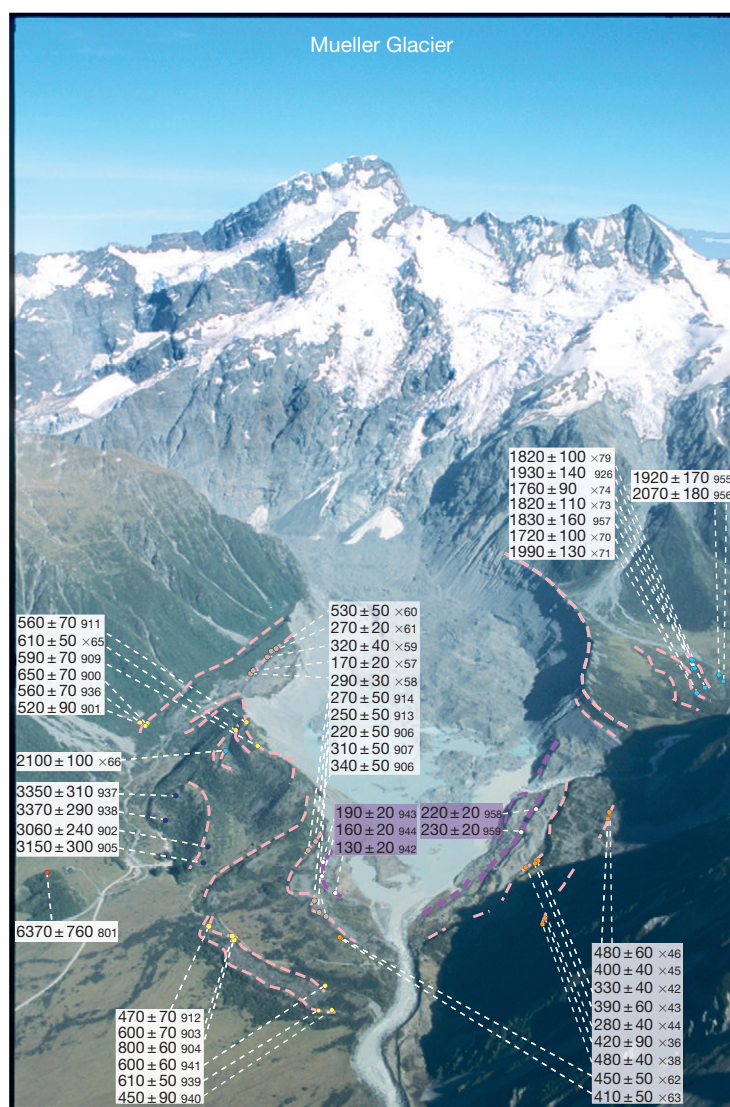


Figure 1 High-frequency advances of Mueller Glacier, New Zealand's Southern Alps, reconstructed by high-sensitivity ^{10}Be dating of moraine boulders, including five distinct advances during the last millennium. After Schaefer et al. (2009).

crushing and subsequent analysis of the remaining helium released by heating allows accurate determination of the cosmogenic ^3He concentration (Kurz and Brook, 1994; Niedermann, 2002 and references therein).

^3He from Fe/Ti-oxide minerals and garnet

In 2002, Bryce and Farley (2002) presented cosmogenic ^3He data from magnetite at the Goldschmidt Conference in Davos. Subsequently, Kober et al. (2005) presented a convincing cosmogenic ^3He dataset from Fe/Ti-oxide minerals. Using the ^{21}Ne exposure ages from quartz from the same surfaces, they calibrated the production rate on five different samples (Table 1), resulting in much higher production rates from Fe/Ti-oxide minerals than previously suggested (Bryce and Farley, 2002; Margerison et al., 2005). Kober et al. (2005) also modeled the ^3He production rates from the main target elements (i.e., O, Mg, Al, Si, Fe, Ni; Table 1) and reported excellent agreement between experimentally

and theoretically derived values. The main findings were the high production-rate values from O, Mg, and Al, indicating that elemental analysis for each sample is a requirement for all cosmogenic ^3He studies. This has not been done routinely since cosmogenic ^3He was assumed to be produced at very similar rates to the main target elements given above. Although the discussion about the production rate of cosmogenic ^3He in Fe/Ti-minerals remains controversial for now, this new tool is very promising.

Gayer et al. (2004) reported cosmogenic ^3He data from Himalayan garnets and compared those with ^{10}Be data from coexisting quartz. Their ^3He production-rate calibration results show a trend towards higher $^3\text{He}/^{10}\text{Be}$ production ratio with increasing altitude, which they interpret as an altitude dependence of this production ratio. While certainly more studies are needed to answer the question of element-dependent scaling factors, this study introduces cosmogenic ^3He from garnet as a new SED tool.

Table 1 Recently published TCN production rates at the Earth's surface at SLHL

<i>Path–target–nuclide</i>	<i>Method</i>	<i>Production rate (g⁻¹ year⁻¹)</i>	<i>Reference</i>	<i>Comment</i>
Helium				
S–(Fe/Ti-oxide)– ³ He	Cross-calibrated from cosmogenic ²¹ Ne in coexisting quartz	120 ± 11	Kober et al. (2005)	Valid for present-day magnetic field
S–(Fe/Ti-oxide)– ³ He	Experimental	69–77	Bryce and Farley (2002)	
S–(Fe/Ti-oxide)– ³ He	Calculated	124 ± 24	Kober et al. (2005)	
S–(O)– ³ He	Calculated	191	Kober et al. (2005)	
S–(Mg)– ³ He	Calculated	182		
S–(Al)– ³ He	Calculated	160		
S–(Si)– ³ He	Calculated	129		
S–(Fe)– ³ He	Calculated	95		
S–(Ni)– ³ He	Calculated	91		
Neon				
S–(sanidine)– ²¹ Ne	Cross-calibrated from cosmogenic ²¹ Ne in coexisting quartz	30.4 ± 5.4	Kober et al. (2005)	Valid for present-day magnetic field
S–(sanidine)– ²¹ Ne	Calculated	28 ± 5.6	Kober et al. (2005)	Valid for present-day magnetic field
S–(Na)– ²¹ Ne	Calculated	208	Kober et al. (2005)	
S–(Mg)– ²¹ Ne	Calculated	189		
S–(Al)– ²¹ Ne	Calculated	60		
S–(Si)– ²¹ Ne	Calculated	44		
S–(Ca)– ²¹ Ne	Calculated	17		
S–(Fe)– ²¹ Ne	Calculated	1		
Argon				
S–(calcium)– ³⁸ Ar		~200	Niedermann et al. (2005)	
Beryllium				
S–(sanidine)– ¹⁰ Be	Cross-calibrated from cosmogenic ¹⁰ Be in coexisting quartz	4.16 ± 0.39	Kober et al. (2005)	Valid for present-day geomagnetic field; the value given here is based on the ¹⁰ Be production rate of 5.1 at g ⁻¹ year ⁻¹
S–(sanidine)– ¹⁰ Be	Calculated	4.55 ± 0.91	Kober et al. (2005)	
S–(O)– ¹⁰ Be	Calculated	6.5	Kober et al. (2005)	
S–(N)– ¹⁰ Be	Calculated	24.5		
S–(Mg)– ¹⁰ Be	Calculated	1.9		
S–(Al)– ¹⁰ Be	Calculated	1.6		
S–(Si)– ¹⁰ Be	Calculated	3.3		
S–(Ca)– ¹⁰ Be	Calculated	1.1		
S–(Fe)– ¹⁰ Be	Calculated	1.9		
S–(Ni)– ¹⁰ Be	Calculated	1.5		
S–(carbonate)– ¹⁰ Be	Cross-calibrated from cosmogenic ³⁶ Cl	37.9	Braucher et al. (2005)	
In situ ¹⁴C				
S–(quartz)– ¹⁴ C	17.4 ± 0.3 ka	15.1 ± 0.2 16.3 ± 0.2	Lifton et al. (2001), Miller et al., 2006), Lifton et al. (ms in prep.)	Assumes 100% spallation production Assumes 83% spallation, 2% fast muon, 15% slow muon production

This table is complementary to Table 4 in Gosse and Phillips (2001). 'S' means 'spallation.'

²¹Ne from Quartz, Olivine, Pyroxene, and Sanidine

Cosmogenic neon is a more complex system but has specific potential compared to cosmogenic helium. The production rate of cosmogenic ²¹Ne (²¹Ne is preferred over other Ne isotopes because it has the lowest isotopic abundance in air) is less well calibrated, primarily relying on the studies by Niedermann and Graf (Balco and Shuster, 2009; Niedermann, 2000; Niedermann et al., 1994). However, the useful age range is the largest of all TCNs, ranging from 10 ka (Niedermann et al., 1994) to more than 30 Ma (Dunai et al., 2005).

Traditionally, cosmogenic ²¹Ne has been analyzed from quartz (Niedermann et al., 1994), olivine (Poreda and Cerling, 1992), and pyroxene (e.g., Bruno et al., 1997).

²¹Ne from sanidine

Kober et al. (2005) reported successful measurement of ²¹Ne from sanidine (for production-rate values see Table 1) showing that cosmogenic ²¹Ne is quantitatively retained in sanidine. A very interesting aspect of this work is the apparently pure two-component mixture of atmospheric and cosmogenic neon

for all samples analyzed. If nucleogenic neon is less abundant or absent in sanidine, implying that the separation of cosmogenic neon from the neon fraction trapped in the mineral lattice (mostly nucleogenic neon) is not required, cosmogenic neon in sanidine allows for dating of young surfaces. For all studies, analysis of the elemental composition of the mineral separates is required because of large differences between elemental production rates (Table 1). Several elemental production rates rely on modeling so far, which is an additional uncertainty. The analytical measurement of cosmogenic neon is challenging because of doubly-charged ions (e.g., $^{40}\text{Ar}^{++}$, CO_2^{++}) interfering with the single-charged neon isotopes in the collector of the mass spectrometer, which might explain the relatively small number of published cosmogenic neon studies.

^{38}Ar from Calcium

Cosmogenic ^{38}Ar is produced in terrestrial surface rocks by spallation of target nuclides, in particular K and Ca. Though the presence of cosmogenic argon in Ca-rich minerals has been demonstrated earlier (Renne et al., 2001), it has proven difficult to establish its production rate because of problems connected to ^{36}Ar production by ^{35}Cl neutron capture and different production rates from K and Ca. A strategy to circumvent this problem is to analyze essentially Cl- and K-free minerals, as done by Niedermann et al. (2007) using pyroxene, which is low in potassium (Table 1). Although this methodology is still in the developmental stage, calibrating the ^{38}Ar SED method is important because it would allow the use of the many existing Ar/Ar mass spectrometers for this SED technique.

^{10}Be from Calcite, and Sanidine

^{10}Be from calcite

Braucher et al. (2005) report a first attempt to measure *in situ* ^{10}Be from calcite. By comparing the *in situ* ^{10}Be –*in situ* ^{36}Cl concentrations from a depth profile in carbonate rocks, they find constant $^{10}\text{Be}/^{36}\text{Cl}$ ratios for all samples, which suggests that all ^{10}Be was produced *in situ*. They determine a sixfold higher ^{10}Be production rate from calcite than from quartz (see Table 1). In their study, they also report that simple rinsing in deionized water over 24 h removes the meteoric ^{10}Be completely, implying that meteoric ^{10}Be does not adsorb on carbonates. Simple *in situ* ^{10}Be measurements from carbonates would represent a significant advance, particularly interesting for ^{10}Be – ^{36}Cl burial dating techniques. However, as the authors state themselves, “it is a first attempt that must be verified in future studies.”

^{10}Be from sanidine

In addition to the noble gas measurements mentioned above, Kober et al. (2005) measured the ^{10}Be from sanidine following a chemical extraction procedure similar to that routinely done for ^{10}Be from quartz. Again, the production is element-dependent, that is, the elemental composition of each mineral separate has to be measured.

^{26}Al from Quartz

^{26}Al is produced in quartz at a rate approximately 6.8 times higher than that of ^{10}Be (Nishiizumi et al., 1989). *In situ* ^{26}Al and ^{10}Be can be extracted in parallel from the same quartz separate without any significant extra effort during the chemical processing, resulting in a large number of combined ^{26}Al and ^{10}Be studies. The ^{26}Al production rate is nearly as well known as the ^{10}Be production rate. Its useful age range is from a few 1000 years to 2.5 My. The analytical uncertainty is higher than for the ^{10}Be , typically some 5–7%. Another methodological problem is the need for extremely pure, essentially Al-free quartz separates (only 0.5% of feldspar in the quartz separate compromises the ^{26}Al analysis). Its most important recent application is arguably the burial dating technique discussed elsewhere in this encyclopedia.

In Situ ^{14}C from Quartz and Olivine

Over the last decade, the AMS group at the University of Arizona has developed a revolutionary tool – *in situ* ^{14}C . Lifton et al. (2001) report successful routine measurements of *in situ* ^{14}C from quartz. The major challenge, namely, the separation of the atmospheric ^{14}C from the *in situ* ^{14}C , is overcome by a combination of acid etching and stepwise combustion with a degassed oxidizing flux. The major advantage of this new tool is its dating range covering the Holocene, where traditional TCN methods become problematic (with the exception of ^{10}Be see above).

Recent studies combine *in situ* ^{14}C with ^{10}Be SED (Goehring et al., 2011; Miller et al., 2006), introducing pro- and subglacial bedrock as climate and glaciologic archives. This new line of research has a lot of potential, but there are still only a handful of laboratories worldwide equipped with an *in situ* ^{14}C extraction system; also, *in situ* ^{14}C sample processing remains complex, slow, and expensive. However, the Arizona group recently automated the major steps in the extraction process, potentially enabling a fourfold increase in throughput when complete. Work to fine-tune the automation equipment and procedures is ongoing.

An additional application of *in situ* ^{14}C forms one of the core components of the CRONUS-Earth initiative (Figure 2). After about 4–5 half-lives, the production and the decay of *in situ* radionuclides reach secular equilibrium. Therefore, surfaces older than approximately 25 ka can be used to estimate the local ^{14}C production rate since the nuclide inventory under those conditions is only a function of the production rate and the decay constant ($t_{1/2}$ for ^{14}C is 5.73 ka). Doing so along transects in altitude and latitude allows one to constrain time-integrated scaling factors.

^{14}C from olivine

Recently, Pigati et al. (2010) have demonstrated a method for extracting *in situ* ^{14}C from olivine separates, extending the *in situ* ^{14}C method to mafic lithologies. They take an approach similar to that of Lifton et al. (2001), in that they use chemical pretreatments followed by stepped combustion with an oxidizing flux. However, recovery of *in situ* ^{14}C with this method is dependent on the Mg:Fe ratio of the olivine, apparently resulting from chemical interactions in the melted flux. A large

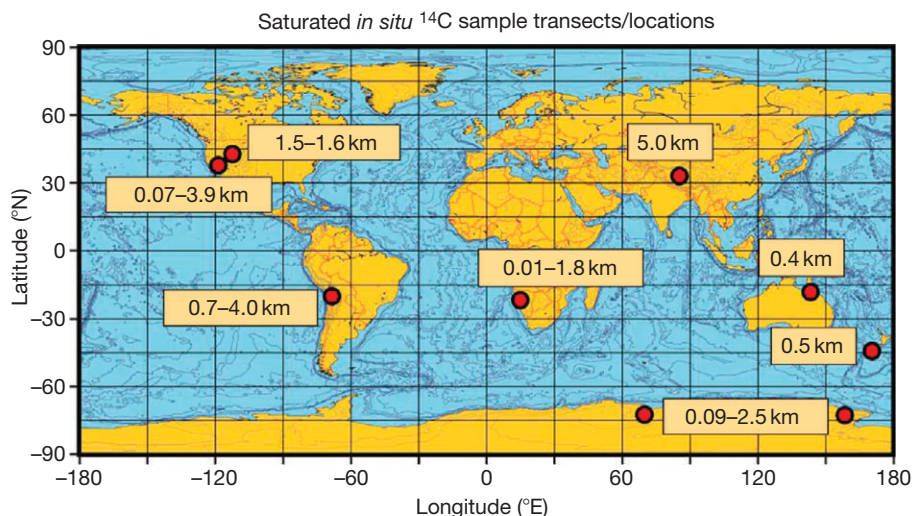


Figure 2 Map of the *in situ* ^{14}C survey within the CRONUS-Earth project.

empirical correction based on this ratio, however, brings the recovered *in situ* ^{14}C concentrations into agreement with *in situ* ^{14}C measurements in quartz samples from similar locations. As with several other methods described here, this is an emerging technique that needs to be verified in future studies.

^{36}Cl from Whole Rock and Feldspar

^{36}Cl production in rocks is complex and dominated by multiple production pathways at distinct energies. A particular problem is the production via natural chlorine, in rocks. This problem is reflected in the large inconsistencies of published production rates, which can differ by up to a factor of 2 (Phillips et al., 1996; Stone et al., 1996; Swanson, 1996; Swanson and Caffee, 2001). One promising way to reduce the natural chlorine background is to use mineral separates (Schimmelpfennig et al., 2009; Stone et al., 1996). Schimmelpfennig et al. (2009) also report that partial dissolution of 10–20% of the mineral separates prior to the total digestion can eliminate the natural chlorine almost completely. The ^{36}Cl tool holds large potential due to the combination of using a relatively simple chemical extraction procedure on whole-rock samples and a wide age range allowing dating of surfaces as young as several hundred years up to about 1 My. However, more methodological work is needed to improve the understanding of production rate systematics and potential problems with water mobilization of ^{36}Cl by weathering processes.

Production of Cosmogenic Nuclides in Terrestrial Rocks

As mentioned above, the rapid increase in applications of TCN methods is paralleled by a better understanding of the complexity of TCN production physics. In this section, we will briefly present these new findings impacting the TCN production systematics and their scaling with altitude and latitude, as well as recently published temporal corrections for TCN

production rates, such as corrections for geomagnetic field variations or long-term solar modulations.

Calculation of an absolute surface exposure age using TCNs at a given location on Earth requires the knowledge of the local TCN production rate through time. Time-integrated TCN production rates are estimated by (i) measuring the cosmogenic nuclide concentrations in samples from surfaces of well-constrained exposure histories or (ii) measuring the cosmogenic radionuclide concentration of an essentially uneroded surface that is old enough for the radionuclide concentration to approach saturation/secular equilibrium (i.e., a steady state between production and decay – typically 4–5 half-lives). There are relatively few such sites, however; hence, estimating the TCN production rate at a given study typically relies on scaling the production rates from the calibration locations with theoretical and empirical models of cosmic ray flux variation in time and space. Production rates are typically referenced to a hypothetical location at sea level and high geomagnetic latitude (SLHL) as a basis for comparison. Such reference production rates depend on the calibration site values as well as the particular scaling model used – i.e., values calculated with a given set of scaling parameters are not generally comparable to those calculated with another set.

TCN research in the late twentieth century relied primarily on the relatively simple production rate scaling model of Lal (1991). However, the years 2000–2005 yielded several alternative models of generally increasing complexity, in attempts to better understand the spatial and temporal variations in TCN production rates around the globe. Unfortunately, the increased complexity also yielded a corresponding increase in confusion among the user community, as it was very difficult to compare the predictions of each model directly in a consistent framework. To remedy this situation, the CRONUS-Earth project developed a set of online calculators for exposure ages and erosion rates that provide results for each published scaling model, based on an internally consistent geomagnetic and atmospheric framework using a common geochemical standardization for each nuclide (Balco et al., 2008; Goehring et al., 2010). These calculators provide users with an easy way

to compare large numbers of measurements made at different times, from different studies, in different locations. For example, they have enabled new paleoclimatic interpretations from multiple large datasets of TCN measurements (e.g., Owen et al., 2008).

Table 1 gives an overview of the recently published production rates for the TCNs ^3He , ^{21}Ne , ^{38}Ar , ^{14}C , and ^{10}Be . This table is meant to be a complementary addition to Table 4 in the review by Gosse and Phillips (2001). The timescale of SED currently ranges from a few 100 years (Stone et al., 2003a) to several tens of My (Dunai et al., 2005; Margerison et al., 2005; Schäfer et al., 1999), and the variety of lithologies that can be cosmogenically dated by various TCNs is steadily growing. Despite the numerous experiments, the production rates of TCNs on the Earth's surface and their variation in time and space represented the biggest remaining uncertainties of the method. The CRONUS-Earth and CRONUS-EU initiatives have made great strides in reducing these uncertainties.

For example, a major advance arising from the CRONUS-Earth project (and other studies) is a more precise and accurate estimate of the global ^{10}Be production rate. The greatly reduced scatter over previously published datasets (Balco et al., 2008) has resulted from careful attention to calibration site selection and independent age control, at sites spanning a range of altitudes and latitudes, as well as improvements in geochemical processing and accelerator mass spectrometry (AMS) measurement techniques. This new generation of calibration sites includes shoreline features associated with Pleistocene Lake Bonneville, Utah, Late Glacial landslide and moraine deposits in northwestern Scotland, recessional moraines in the north-eastern United States (a subset of those sampled in Balco et al. (2009)), Late Glacial deposits in Norway and near the Quelccaya ice cap in Peru, a particularly well-dated debris flow deposit in New Zealand (Putnam et al., 2010), and well-dated deposits in Patagonia (Kaplan et al., 2011) and Greenland (Briner et al., 2012). This, in turn, has allowed for previously unattainable temporal resolution in SED studies using ^{10}Be (e.g., Schaefer et al., 2009). Another key milestone in ^{10}Be applications has been the recent convergence of ^{10}Be half-life estimates (Chmeleff et al., 2010; Korschinek et al., 2010; Nishizumi et al., 2007), resolving a longstanding controversy.

Production of *in situ* cosmogenic nuclides in near-surface rocks is complex, as TCN production is a multiparameter function of the cosmic ray flux at the Earth's surface. Variations in this flux are consequences of (i) variations in the primary galactic cosmic ray flux, (ii) solar modulation of the GCR flux, (iii) changes in atmospheric shielding, and (iv) variations in the intensity and configuration of the Earth's magnetic field. While the low-frequency variation (i) is of minor importance for the application of TCN on timescales of 10^3 – 10^6 years, the other variations have to be considered. This led to a number of studies designed to characterize the various effects. The online CRONUS calculators (Balco et al., 2008; Goehring et al., 2010) each incorporate the latest advances in our understanding of these variations.

Geomagnetic Variation and TCN Production

There are several recent studies investigating the impact of geomagnetic variations in intensity and configuration on the

TCN production on the Earth's surface (Desilets and Zreda, 2003; Dunai, 2000; Gosse and Phillips, 2001; Masarik et al., 2001; Pigati and Lifton, 2004). While each predicts somewhat different impacts, the CRONUS online calculators utilize a common geomagnetic framework based largely on the model of Lifton et al. (2008), which captures the Holocene effects on TCN production of time-dependent spherical harmonic models developed by Korte and Constable (2005) and subsequent studies (Korte et al., 2009).

Solar Modulation of TCN Production

A recent study by Lifton et al. (2005) introduced correction factors for millennial-scale solar modulation of TCN production rates based on model sunspot numbers over the last 11.4 ka. They find effects on spallogenic nucleon scaling factors for SLHL of up to 10%, supporting the view that long-term solar modulation must be included in secondary cosmic ray scaling procedures.

Changes in Atmospheric Shielding

Stone (2000) introduced the effects of regional air pressure regimes different from the standard atmosphere air pressure. The biggest changes are in Antarctica, where air pressure is about 25 hPa lower on average than elsewhere, resulting in up to 30% higher production rates. This paper has also published a map of global correction factors relative to standard atmospheric values, which should be used in all TCN studies and which are included in the online CRONUS calculators. The influence of variations in the local air pressure regime during glacial periods towards lower air pressure was recently suggested by Ackert et al. (2003) as an explanation for their relatively high ^3He production rates constrained from $^{39}\text{Ar}/^{40}\text{Ar}$ -dated lava some 70 and 110 ka in age, and by Licciardi et al. (2006) in a study calibrating ^3He production rates in Iceland.

Outlook

This article focuses on the most recent progress in the method of cosmogenic surface exposure dating over the last several years, highlighting the rapid progress in the field. Cosmogenic nuclide surface exposure dating has fundamentally changed our perspectives on quantifying Earth surface processes. However, prior to 2005, controversies about TCN production systematics threatened to limit the range of potential applications. The multidisciplinary CRONUS-Earth and CRONUS-EU projects were designed to improve significantly the accuracy and reliability of TCN science while making the required production-rate systematics easier to understand and apply for the broader user community. CRONUS-Earth is winding down, but it has achieved much toward its goal of a robust foundation for quantitative surficial processes and Quaternary geologic science with reliably defined uncertainties. A particular success has been the production rate of the TCN workhorse, namely, ^{10}Be , which is now known to be better than 10% globally, and which appears to be approximately 10% lower than previously thought. Tighter constraints on ^{36}Cl production are another

positive outcome. Finally, a better understanding of scaling systematics is within reach as a result of this work. While it remains the prime tool for modern Earth sciences research, future progress in this field can expand TCN applications to disciplines such as archaeology.

See also: Dating Techniques. **Cosmogenic Nuclide Dating:** Cosmic Ray Interactions in Minerals; Exposure Geochronology; Landscape Evolution; Overview. **Paleoclimate:** The Younger Dryas Climate Event. **Paleoclimate Reconstruction:** The Last Millennium; Younger Dryas Oscillation, Global Evidence.

References

- Ackert RP, Singer BS, Guillo H, Kaplan MR, and Kurz MD (2003) Long-term cosmogenic ^3He production rates from $^{40}\text{Ar}/^{39}\text{Ar}$ and K–Ar dated Patagonian lava flows at 47°S. *Earth and Planetary Science Letters* 210: 119–136.
- Balco G (2011) Contributions and unrealized potential contributions of cosmogenic-nuclide exposure dating to glacier chronology 1990–2010. *Quaternary Science Reviews* 30: 3–27.
- Balco G, Briner JP, Rayburn J, Ridge JC, and Schaefer JM (2009) Regional beryllium-10 production rate calibration for northeastern North America. *Quaternary Geochronology* 4: 93–107.
- Balco G and Shuster DL (2009) Production rate of cosmogenic ^{21}Ne in quartz estimated from ^{10}Be , ^{26}Al , and ^{21}Ne concentrations in slowly eroding Antarctic bedrock surfaces. *Earth and Planetary Science Letters* 281: 48–58.
- Balco G, Stone JO, Lifton NA, and Dunai TJ (2008) A complete and easily accessible means of calculating surface exposure ages or erosion rates from Be-10 and Al-26 measurements. *Quaternary Geochronology* 3: 174–195.
- Blard P, Bourles D, Lave J, and Pik R (2006) Applications of ancient cosmic-ray exposures: Theory, techniques and limitations. *Quaternary Geochronology* 1(1): 59–73.
- Braucher R, Benedetti L, Bourles D, Brown ET, and Chardon D (2005) Use on in-situ produced Be-10 in carbonate-rich environments: A first attempt. *Geochimica et Cosmochimica Acta* 69: 1473–1478.
- Briner JP, Young NE, Goehring B, and Schaefer JM (2012) Constraining Holocene ^{10}Be production rates in Greenland. *Journal of Quaternary Science* 27: 2–6.
- Brook EJ and Kurz MD (1993) Surface-exposure chronology using in situ cosmogenic ^3He in Antarctic quartz sandstone boulders. *Quaternary Research* 39: 1–10.
- Bruno LA, Baur H, Graf T, Schlüchter C, Signer P, and Wieler R (1997) Dating of Sirius Group tillites in the Antarctic Dry Valleys with cosmogenic ^3He and ^{21}Ne . *Earth and Planetary Science Letters* 147: 37–54.
- Bryce JG and Farley KA (2002) He-3 exposure dating of magnetite. *Geochimica et Cosmochimica Acta* 66: A108.
- Cerling TE and Craig H (1994) Geomorphology and in-situ cosmogenic isotopes. *Annual Review of Earth and Planetary Sciences* 22: 273–317.
- Chmieleff J, von Blanckenburg F, Kossert K, and Jakob D (2010) Determination of the Be-10 half-life by multicollector ICP-MS and liquid scintillation counting. *Nuclear Instruments and Methods in Physics Research Section B: Beam Interactions with Materials and Atoms* 268: 192–199.
- Desilets D and Zreda M (2003) Spatial and temporal distribution of secondary cosmic-ray nucleon intensities and applications to in situ cosmogenic dating. *Earth and Planetary Science Letters* 206: 21–42.
- Dunai JT (2000) Scaling factors for production rates of in-situ produced cosmogenic nuclides: A critical re-evaluation. *Earth and Planetary Science Letters* 176: 157–169.
- Dunai TJ, Gonzalez-Lopes GA, and Juez-Larre J (2005) Oligocene–Miocene age of aridity in the Atacama Desert revealed by exposure dating of erosion-sensitive landforms. *Geology* 33: 321–324.
- Fenton CR, Hermanns RL, Blikra LH, et al. (2011) Regional (10)Be production rate calibration for the past 12 ka deduced from the radiocarbon-dated Grotlandsura and Russenes rock avalanches at 69 degrees N, Norway. *Quaternary Geochronology* 6: 437–452.
- Gayer E, Pik R, Lave J, France-Lanord C, Bourles D, and Marty B (2004) Cosmogenic He-3 in Himalayan garnets indicating an altitude dependence of the He-3/Be-10 production ratio. *Earth and Planetary Science Letters* 229: 91–104.
- Goehring B, Kurz M, Balco G, Schaefer J, Licciardi J, and Lifton N (2010) A reevaluation of in situ cosmogenic ^3He production rates. *Quaternary Geochronology* 5: 410–418.
- Goehring B, Schaefer JM, Schluechter C, et al. (2011) The Rhone Glacier was smaller than today for most of the Holocene. *Geology* 39: 7.
- Gosse JC and Phillips FM (2001) Terrestrial in situ cosmogenic nuclides: Theory and application. *Quaternary Science Reviews* 20: 1475–1560.
- Ivy-Ochs S and Kober F (2010) Surface exposure dating with cosmogenic nuclides. *Quaternary Science Journal* 57: 157–189.
- Kaplan MR, Strelin J, Schaefer JM, et al. (2011) The ^{10}Be production rate in southern South America and the coherency of late glacial chronologies. *Earth and Planetary Science Letters* 309: 21–32.
- Kober F, Ivy-Ochs S, Leya I, et al. (2005) In situ cosmogenic ^{10}Be and ^{21}Ne in sanidine and in situ cosmogenic ^3He in Fe–Ti-oxide minerals. *Earth and Planetary Science Letters* 236: 404–418.
- Korschinek G, Bergmaier A, Faestermann T, et al. (2010) A new value for the half-life of ^{10}Be by heavy-ion elastic recoil detection and liquid scintillation counting. *Nuclear Instruments and Methods in Physics Research, B* 268(2): 187–191.
- Korte M and Constable CG (2005) The geomagnetic dipole moment over the last 7000 years – New results from a global model. *Earth and Planetary Science Letters* 236: 348–358.
- Korte M, Donadini F, and Constable CG (2009) Geomagnetic field for 0–3 ka: 2. A new series of time-varying global models. *Geochemistry, Geophysics, Geosystems* 10: Q06008.
- Kurz MD (1986) Cosmogenic helium in a terrestrial igneous rock. *Nature* 320: 435–439.
- Kurz MD and Brook EJ (1994) Surface exposure dating with cosmogenic nuclides. In: Beck C (ed.) *Dating in Exposed and Surface Contexts*. Albuquerque: University of New Mexico Press.
- Lal D (1991) Cosmic ray labeling of erosion surfaces: In situ nuclide production rates and erosion models. *Earth and Planetary Science Letters* 104: 424–439.
- Licciardi JM, Clark PU, Brook EJ, et al. (2001) Cosmogenic ^3He and ^{10}Be chronologies of the late Pinedale northern Yellowstone ice cap, Montana, USA. *Geology* 29: 1095–1098.
- Licciardi JM, Kurz MD, and Curtice JM (2006) Cosmogenic He-3 production rates from Holocene lava flows in Iceland. *Earth and Planetary Science Letters* 246: 251–264.
- Licciardi JM, Schaefer JM, Taggart JR, and Lund DC (2009) Holocene glacier fluctuations in the Peruvian Andes indicate northern climate linkages. *Science* 325: 1677–1679.
- Lifton NA, Biebel JW, Clem JM, et al. (2005) Addressing solar modulation and long-term uncertainties in scaling secondary cosmic rays for in situ cosmogenic nuclide applications. *Earth and Planetary Science Letters* 239: 140–161.
- Lifton NA, Jull AJT, and Quade J (2001) A new extraction technique and production rate estimate for in situ cosmogenic ^{14}C in quartz. *Geochimica et Cosmochimica Acta* 65: 1953–1969.
- Lifton N, Smart DF, and Shea MA (2008) Scaling time-integrated in situ cosmogenic nuclide production rates using a continuous geomagnetic model. *Earth and Planetary Science Letters* 268: 190–201.
- Margerison HR, Phillips WM, Stuart FM, and Sugden DE (2005) Cosmogenic ^3He concentrations in ancient flood deposits from the Coombs Hills, northern Dry Valleys, East Antarctica: Interpreting exposure ages and erosion rates. *Earth and Planetary Science Letters* 230: 163–175.
- Masarik J, Frank M, Schäfer JM, and Wieler R (2001) Correction of in situ cosmogenic nuclide production rates for geomagnetic field intensity variations during the past 800,000 years. *Geochimica et Cosmochimica Acta* 65: 2995–3003.
- Miller GH, Briner JP, Lifton NA, and Finkel RC (2006) Limited ice-sheet erosion and complex in situ cosmogenic Be-10, Al-26, and C-14 on Baffin Island, Arctic Canada. *Quaternary Geochronology* 1: 74–85.
- Niedermann S (2000) The ^{21}Ne production rate in quartz revisited. *Earth and Planetary Science Letters* 183: 361–364.
- Niedermann S (2002) Cosmic-ray-produced noble gases in terrestrial rocks dating tools for surface processes. In: Porcelli D, Ballentine CJ, and Wieler R (eds.) *Noble Gases in Geochemistry and Cosmochemistry*. USA: Mineralogical Society of America.
- Niedermann S, Graf T, Kim JS, Kohl CP, Marti K, and Nishiizumi K (1994) Cosmic-ray-produced ^{21}Ne in terrestrial quartz: The neon inventory of Sierra Nevada quartz separates. *Earth and Planetary Science Letters* 125: 341–355.
- Niedermann S, Schaefer JM, Wieler R, and Naumann R (2005) Experimental determination of the cosmogenic Ar production rate from Ca. *Eos, Transactions American Geophysical Union* 86: Fall Meet. Supplement, Abstract U33A-0003.
- Niedermann S, Schaefer JM, Wieler R, and Naumann R (2007) The production rate of cosmogenic ^{38}Ar from calcium in terrestrial pyroxene. *Earth and Planetary Science Letters* 257: 596–608.

- Nishiizumi K, Imamura M, Caffee MW, Southon JR, Finkel RC, and McAninch J (2007) Absolute calibration of ^{10}Be AMS standards. *Nuclear Instruments and Methods in Physics Research B* 258: 403–413.
- Nishiizumi K, Winterer EL, Kohl CP, et al. (1989) Cosmic ray production rates of ^{10}Be and ^{26}Al in quartz from glacially polished rocks. *Journal of Geophysical Research* 94: 17907–17915.
- Owen LA, Caffee MW, Finkel RC, and Seong YB (2008) Quaternary glaciation of the Himalayan–Tibetan orogen. *Journal of Quaternary Science* 23: 513–531.
- Phillips FM, Zreda MG, Elmore D, and Sharma P (1996) A reevaluation of cosmogenic ^{36}Cl production rates in terrestrial rocks. *Geophysical Research Letters* 23: 949–952.
- Pigati JS and Lifton NA (2004) Geomagnetic effects on time-integrated cosmogenic nuclide production with emphasis on in-situ ^{14}C and ^{10}Be . *Earth and Planetary Science Letters* 226: 193–205.
- Pigati JS, Lifton NA, Jull AJT, and Quade J (2010) Extraction of in situ cosmogenic C-14 from Olivine. *Radiocarbon* 52: 1244–1260.
- Poreda RJ and Cerling TE (1992) Cosmogenic neon in recent lavas from the Western United States. *Geophysical Research Letters* 19: 1863–1866.
- Putnam A, Schaefer JM, Barrell D, et al. (2010) A high precision ^{10}Be production rate calibration in New Zealand's Southern Alps. *Quaternary Geochronology* 5: 392–409.
- Renne PR, Farley KA, Becker TA, and Sharp WD (2001) Terrestrial cosmogenic argon. *Earth and Planetary Science Letters* 188: 435–440.
- Schaefer JM, Denton GH, Kaplan M, et al. (2009) High frequency Holocene glacier fluctuations in New Zealand differ from the northern signature. *Science* 324: 622.
- Schaeffer OA and Davis R (1956) Chlorine-36 in nature. *Nuclear Processes in Geologic Settings*. Washington, DC: National Academy of Sciences.
- Schäfer JM, Ivy-Ochs S, Wieler R, et al. (1999) Cosmogenic noble gas studies in the oldest landscape on Earth: Surface exposure ages of the Dry Valleys, Antarctica. *Earth and Planetary Science Letters* 167: 215–226.
- Schimmelpfennig I, Benedetti L, Finkel R, et al. (2009) Sources of in-situ Cl-36 in basaltic rocks. Implications for calibration of production rates. *Quaternary Geochronology* 4: 441–461.
- Stone J (2000) Air pressure and cosmogenic isotope production. *Journal of Geophysical Research* 105: 23753–23759.
- Stone JO, Allan GL, Fifield LK, and Cresswell RG (1996) Cosmogenic chlorine-36 from calcium spallation. *Geochimica et Cosmochimica Acta* 60: 679–692.
- Stone JO, Balco G, Sugden DE, et al. (2003) Holocene deglaciation of Mary Bird Land, West Antarctica. *Science* 299: 99–102.
- Stone JO, Balco GA, Sugden DE, et al. (2003) Holocene deglaciation of Marie Byrd Land, West Antarctica. *Science* 299: 99–102.
- Swanson T (1996) Determination of Cl-36 production rates from the deglaciation history of Whidbey and Fidalgo Islands, Washington. *Radiocarbon* 38: 172.
- Swanson TW and Caffee MW (2001) Determination of ^{36}Cl production rates derived from the well-dated deglaciation surfaces of Whidbey and Fidalgo Islands, Washington. *Quaternary Research* 56: 366–382.

Cosmic Ray Interactions in Minerals

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Introduction

Nuclear interactions occur everywhere in the universe as natural radiations are present everywhere. The types of interactions and their magnitude show a wide variety as would be expected. Nuclear interactions, particularly the inelastic variety, in which energy transfers occur between the incident particle and the target, have been increasingly used in the past few decades to study the history of evolution of planetary matter. Considering the topics covered in this encyclopedia, the emphasis is on nuclear interactions caused by cosmic radiation in Earth's atmosphere and in upper layers of the lithosphere, with a view to use interaction products as tracers for studying geomorphic processes. It must be pointed out here that this is a field that has now become accessible to quantitative modeling as a result of the recent advances that have made it possible to measure the ultralow concentrations of nuclear products arising from nuclear interactions in surficial and deep-seated materials (Gosse and Phillips, 2001; Lal, 1988a, 1991; Tuniz et al., 1998). Prior to this development, only sketchy information based on indirect assessments, such as coatings on surfaces and thermoluminescence, was available.

To understand the principles and uncertainties of terrestrial *in situ* cosmogenic nuclide (TCN) methods, it is necessary to appreciate the theory and experimental basis for our understanding of primary cosmic ray interactions in the atmosphere and the subsequent interactions between cosmic ray secondaries and matter. Although the theoretical framework allows fairly accurate estimations of the temporal variations in the cosmic ray flux at the top of the atmosphere for hypothetical changes in the geomagnetic field and solar activity, and although the basic features of nuclear interactions are fairly well understood, it is indeed a difficult task to accurately estimate the total rates of nuclear disintegrations and particular nuclear interaction rates giving rise to specified nuclides at different locations in the atmosphere.

Cosmic rays attenuate as they interact with mass, at a rate referred to as an interaction mean free path. Earth's atmosphere represents more than ten interaction mean free paths for the geometric nuclear interaction of nucleons (protons or neutrons) with air nuclei, about 90 g cm^{-2} . Within the lithosphere, the character of nuclear interacting particles changes dramatically, from primarily nucleons to slow and fast muons. Within the atmosphere, the total rate of nuclear disintegrations changes by a factor of 10^2 at the equator and 1.5×10^3 at the poles, between the top of the atmosphere and sea level (Lal and Peters, 1967). It is quite manifest that the atmosphere imposes a heavy demand on any nuclear systematics used to treat the nuclear cascade development. Nuclear interactions are complex, with emission of different particles over a wide range of energies, and with a complex angular distribution (with respect to the incident particle). This requires parametrization of

secondary particle emission characteristics as a function of energy for MeV to GeV energies. As an example, even if one makes an error of 10% in the assignment of the value of one of the relevant parameters, this would lead to an error of $>100\%$ in taking the nucleonic cascade to sea level (Lal, 2000). Even a 10% error in the assignment of geometric nuclear interaction cross-section (the probability of interaction) produces the same effect. The propagation of the generating component of nucleons through the atmosphere depends critically on the inelasticity parameter, which fortunately does not show a strong energy dependence. The observations of cosmic ray neutrons show that at low latitudes the secondary nucleonic component at atmospheric depths greater than 200 g cm^{-2} is appreciably harder than at the poles, which is not unexpected because of the much harder energy spectrum of the primary radiation incident at the equator (protons of $E > 14 \text{ GeV}$ and He nuclei above 6.6 GeV/n). The absorption mean free paths lie in the range of $220\text{--}160 \text{ g cm}^{-2}$ between the equator and the poles. A 10% uncertainty in the mean free paths corresponds to a 35% uncertainty in scaling factors between sea level at 60° and 3 km at the equator.

There is a need for accurate estimates of production rates of nuclides in the atmosphere and at underground depths. Even if the total nuclear disintegration rates can be estimated, it is not generally feasible to make accurate estimations of nuclide production rates. One of the difficulties, in particular, is a lack of knowledge of partial cross-sections for the formation of nuclides in different target elements as a function of neutron energy, because in the atmosphere, neutrons are the principal producers of nuclear interactions.

It is very difficult to depend entirely on theory to accomplish the tasks. One has to employ suitable 'process normalization' techniques, treating finite size chunks of matter as black boxes, without going into details of the interaction. In this treatment, as an example, one assumes that the more energetic component of nucleons is absorbed following certain observed features in high-energy interactions, whereby, after an interaction, the incident nucleon continues with an appreciable energy. The absorption of an incident nucleon (neutron or proton) in matter is then governed by simple absorption rules, again guided by experimental data. It further follows that within a few interaction mean free paths in the target (atmosphere or rock), there would exist certain relationships between the propagating nucleon components in different energy bands, with near-equilibrium between high- and low-energy nucleons. In fact, accepting this makes it possible to make very accurate predictions of fluxes of secondary nucleons in a target.

However, it must be mentioned that apparently some numerical codes are able to predict the buildup of nucleonic cascades in the atmosphere, and sometimes they also predict the partial cross-sections for the formation of different nuclides as a function of energy (Masarik and Reedy, 1995). However,

the codes must also rest heavily on normalization techniques as data are simply not available to treat in any detail the processes that occur in nuclear interactions, as mentioned above.

A practical approach, therefore, is to consider the dynamics of nuclide production theoretically and to obtain absolute nuclide production rates wherever possible by exposing suitable targets to cosmic radiation at different altitudes and latitudes (Lal et al., 1960), as was first shown in the case of radionuclides of half-lives <3 months. With the development of the accelerator mass spectrometry technique, it is now possible to extend this method to long half-life radionuclides (cf. Nishiizumi et al., 1996). Currently utilized production rates of TCN have therefore been empirically based on measurements of the isotope for targets as short as a year to targets as long as tens of thousands of years. The longer duration experiments make use of geological surfaces with exposure ages that have been determined independently (e.g., radiocarbon-dated lava).

Two approaches are currently being used in applications of the cosmogenic nuclides as tracers for production: one to determine the scaling factors for neutrons of different energies at different altitudes and latitudes, and the other to determine the absolute production rates of a nuclide in the atmosphere or in targets exposed in the atmosphere at a few sites. The latter is very necessary, partly because of lack of data on partial cross-sections for formation of nuclides from different target elements. Additionally, semiempirical approaches have to be used to estimate nuclide production within rocks of different sizes exposed at different elevations in the atmosphere (Lal and Chen, 2005; Masarik and Wieler, 2003).

Cosmic Radiation: The Omnipresent Energetic Nuclear Radiation

Among the four principal radiations, which are present everywhere in the universe, electromagnetic radiation, cosmic radiation, neutrinos, and gravitational radiation, the former two occupy a large band in energy, >15 orders of magnitude. By this fact alone, they serve as windows to the universe to map the distribution of gases, dust, and different forms of matter including stars. Cosmic radiation, discovered in 1912 by Victor Hess by carrying his charged-particle detector instrument to an altitude of 17 500 ft in a gondola, consists of nuclei of elements in the periodic table with energies of a few electron volts to $>10^{20}$ eV. In cosmic ray physics, it is customary to measure nuclear particle energies in electron volts. Further, because one must often deal with nuclear matter moving with speeds comparable to the speed of light, it becomes necessary to use relativistic equations derived from Lorentz transformations. The magnitude of the electron volt and the relativistic relations between particle's mass, momentum, and energy are briefly outlined in Appendix A.

As discussed below, most of the cosmic ray flux incident at the Earth lies in the region of a few thousand MeV. Since the binding energy of nucleons (neutrons or protons) in matter also lie in the range of few MeV, the cosmic ray particles are capable of fragmenting nuclei with which they interact. Indeed, they produce appreciable nuclear transformations in the Earth's atmosphere and even in the lithosphere down to great depths. This results in production of new forms of nuclei, some not initially present in the targets. In contrast, the

electromagnetic radiation, which interacts very weakly with matter by virtue of its coulomb field, does not produce any appreciable changes in the nuclear composition of matter.

In this article, we discuss the salient features of cosmic radiation, their composition and energy spectra, development of the secondary cosmic ray beam in the atmosphere and the lithosphere, and production of nuclides in this process by the principal nuclear active particles. The incident cosmic ray flux at the top of the atmosphere is latitude dependent since it is strongly modulated by (1) the Earth's geomagnetic field and (2) solar plasma. These modulations show temporal variability because both geomagnetic field and solar activity are variable. We explicitly consider the nature of this variability.

Composition and Energy Spectrum of Cosmic Rays

'Cosmic rays' are composed of ordinary forms of matter accelerated by electromagnetic processes in certain flares in stars, and by shock processes in explosions of supernovas to relativistic energies. (The term cosmic radiation was originally adopted to reflect our lack of knowledge of their acceleration mechanisms and also the stellar sources primarily responsible for their acceleration. It remains today as the term of choice because of the multiplicity of its sources in the universe, from stars to supernovas.)

Although cosmic rays are accelerated in several astrophysical settings, the principal source of energy for galactic cosmic ray (GCR) acceleration seems to be shock waves driven by the explosions of supernovas (see Higdon et al., 1998; Ramaty et al., 1998; and references therein). Most (80–90%) of the galactic supernovas are type II and Ib/C (core collapses of massive O and B stars). Since the average supernova energy release is about 10^{50} erg, and their frequency is ~ 3 per century, the average cosmic ray power, 10^{41} erg s $^{-1}$, needed to maintain the level of 'galactic' cosmic rays is met by supernovas.

As would be expected, the cosmic ray beam is composed of fast nuclei up to relativistic energies, completely stripped of their electrons in the early stages of acceleration. Since hydrogen is the most abundant element in the universe, most of the flux is comprised of protons. Next in abundance are He, C, N, O, and heavier nuclei. Li, Be, B ions are rare in the universe. Their presence in the cosmic ray beam arises primarily due to nuclear interactions of heavier nuclei.

The wide range of cosmic ray energies requires that their energy spectra be described in terms of a power law function in kinetic energy, E_k , as differential or integral forms:

$$\frac{dn}{dE_k} = j(E_k) \quad [1]$$

$$N(> E_k) = \int_{E_k} j(E_k) dE_k \quad [2]$$

E_k is defined as kinetic energy of the particle. In the case of complex nuclei, such as He and heavier nuclei (C, N, O, Si, Fe, etc.), it is often defined as kinetic energy per nucleon. A typical form of $j(E_k)$ over a wide range of energies is: $(a + E_k)^{-\Gamma}$, except at low energies, where it is of the form: $(a + E_k)^{-\Gamma}$, where a is a constant. At kinetic energies >3 GeV, the value of Γ is ~ 2.5 . For further details, see Peters (1958), Simpson (1983, 1992), and Ziegler (1998).

Primary GCR Energy Spectra and its Modulation by Solar Plasma

The differential energy spectra of H, He, C, and Fe nuclei in the energy region $10\text{--}10^4$ MeV per nucleon are shown in **Figure 1**, obtained using instruments flown in a large number of balloon flights at the top of the atmosphere (Simpson, 1992). The scatter in the measured data, taken at different periods, arises from varying modulation of the cosmic ray beam by the solar plasma. The primary GCR flux incident at the top of Earth's atmosphere is anticorrelated with solar activity (with some phase lag). See Gleeson and Axford (1968) for a convection-diffusion theoretical framework for the transport of GCR particles through the heliosphere, and see Castagnoli and Lal (1980) and Papini et al. (1996) for the essential features of the experimental data on solar modulation of primary cosmic ray flux.

In terms of the solar modulation formulism (Gleeson and Axford, 1968), the effective change in the energy of a charged particle moving from a point P on the boundary along a dynamical trajectory to an interior point Q is $Ze\phi$, irrespective

of its path to Q. If the measured kinetic energy of a charged particle at Q is E_k per nucleon, then it follows (Randall and Van Allen, 1986) that its kinetic energy per nucleon at P must have been $(E_k + Ze\phi)$. As shown by Randall and Van Allen (1986), this leads to the following simple equation relating differential fluxes at points P and Q:

$$\frac{j_Q(T)}{T(T + 2mc^2)} = \frac{j_P(T + (Ze\phi/A))}{(T + (Ze\phi/A))(T + Ze\phi/A + 2mc^2)} \quad [3]$$

In eqn [3], A is the mass number, T is defined as the kinetic energy per nucleon, the total kinetic energy being AE_k . If the point P is taken at the heliosphere boundary, and the point Q at 1 AU, then the charged particle is modulated through a potential difference of ϕ MeV per nucleon.

For more than four decades, the solar modulation of cosmic ray flux has been studied using satellites/spacecraft in orbit around Earth, and more recently on deep space probes. The range of observed variations in the modulated near-Earth differential flux of GCR protons until the late 1970s is shown in **Figure 2** for protons, as described by different curves designated by the modulation parameter, ϕ (Castagnoli and Lal, 1980; Randall and Van Allen, 1986). The hypothetical curve, $\phi=0$, corresponds to the predicted shape of the interstellar GCR proton flux, that is, the GCR flux outside the outer heliosphere. Information about near-Earth changes in flux have come largely from the charged-particle monitoring experiment aboard the IMP8 satellite since 1973. For a recent summary of these data, reference is made to Simunac and Armstrong (2004).

Our knowledge of modulation of GCR through the outer heliosphere comes from deep space probes (Fuji and McDonald, 1997; Lockwood and Webber, 1995) which transmitted *in situ* cosmic ray data up to distances of more than 70 AU (Pioneers 10 and 11, Voyagers 1 and 2). In **Figure 3** are plotted the results of measurements of time variations in the differential particle fluxes of H ions of 130–225 MeV and He ions of 10–21 and 180–450 MeV/n kinetic energies (Fuji and McDonald, 1997), at distances of up to ~ 60 AU during 1972–95. In the same figure, the measured changes in the counting rates of the neutron monitor at Climax are also presented. Finally, **Figure 4** shows the data on energy spectra of protons measured by Pioneer 10 and Voyager 2 spacecrafts from 1978 to 1992 for GCR protons of >70 MeV kinetic energy and corresponding near-Earth fluxes based on IMP satellite (Lockwood and Webber, 1995). The radial distances of Pioneer 10 and Voyager 2 were ~ 60 and 40 AU, respectively, in 1992.

Figures 2–4 show a consistent behavior of modulation of cosmic ray flux at radial distances of 1–60 AU space in the heliosphere, corresponding to modulation parameter, ϕ values ranging between ~ 350 and 1300 MeV/n, inversely correlated with solar activity (often defined using the Zurich sunspot numbers). During sustained solar minimum periods, as during the Maunder Minimum, the value of ϕ may have been ~ 100 MeV/n (Castagnoli and Lal, 1980).

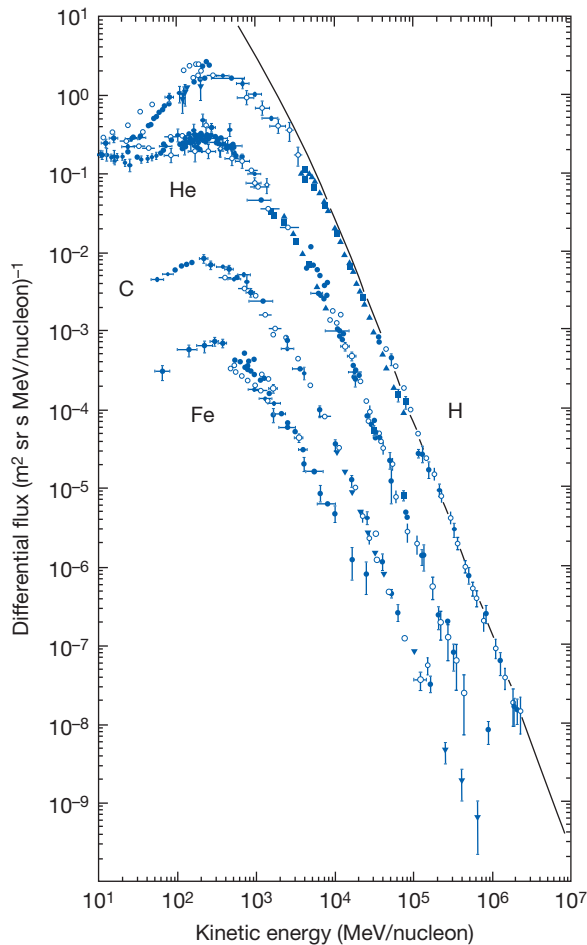


Figure 1 The energy spectra of elements in cosmic rays (from top: H, He, C, and Fe) measured near Earth. The solid curve shows the hydrogen spectrum extrapolated for interstellar space, for no solar modulation. Reproduced from Simpson JA (1992) Cosmic radiation: Particle astrophysics in the heliosphere. *Annals of the New York Academy of Sciences* 655: 95–137.

Solar Cosmic Ray Energy Spectra

The energy spectra of energetic particles accelerated during certain solar flares are often represented as a power law (see Aschwanden, 2002; Lal, 1972, 1988b):

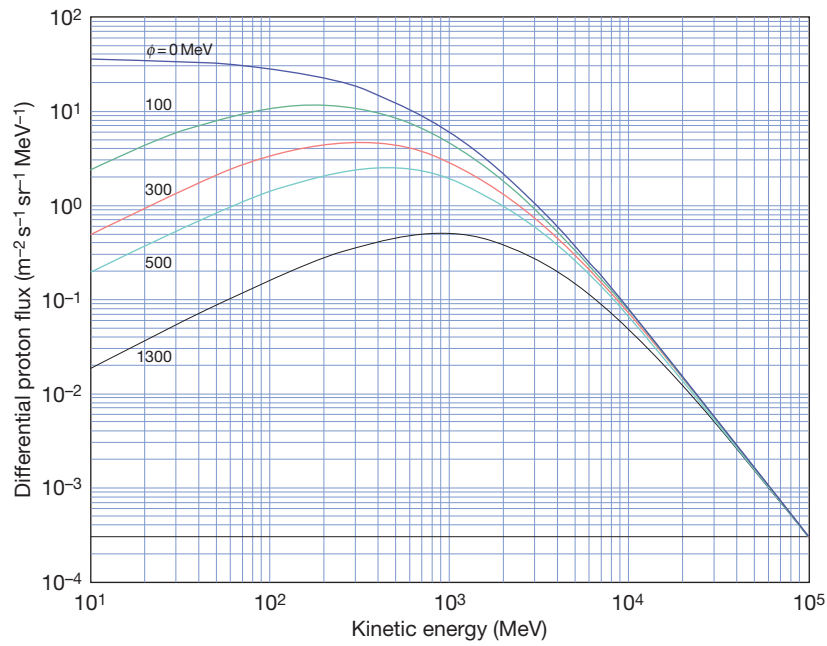


Figure 2 Estimated proton spectra for different solar modulation intensities as defined by ϕ and guided by observed proton spectra during different years (see text). Cosmic ray fluxes for different ϕ values are described well by eqn [3].

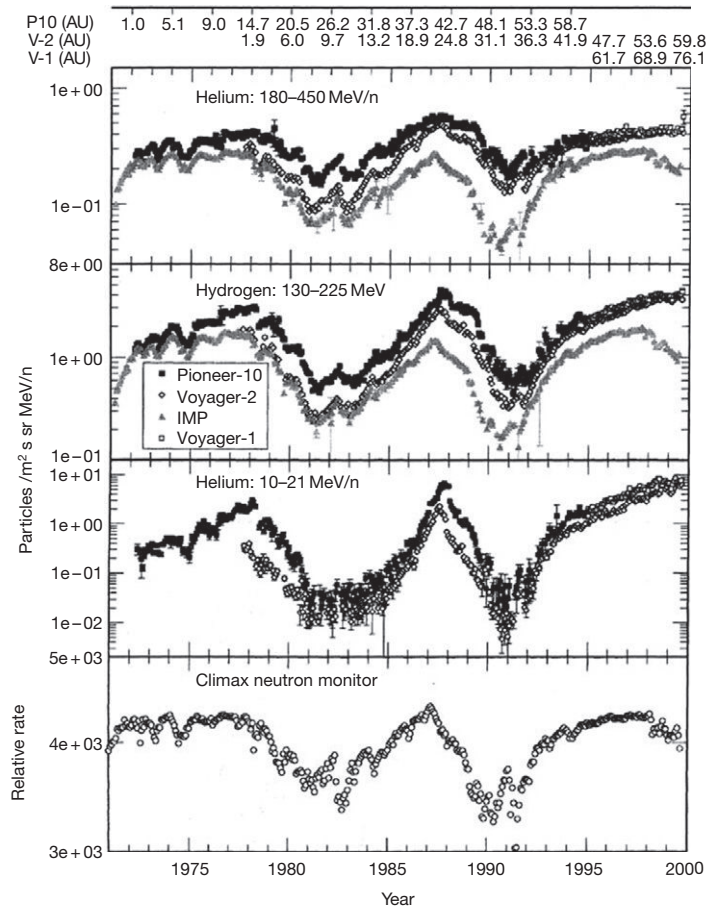


Figure 3 Time histories of the 26-day average galactic cosmic ray intensities of H and He, and of thermal neutrons measured at Climax. Reproduced from Fuji Z and McDonald FB (1997) Radial intensity gradients of galactic cosmic rays (1972–1995) in the heliosphere. *Journal of Geophysical Research* 102(A11): 24201–24208.

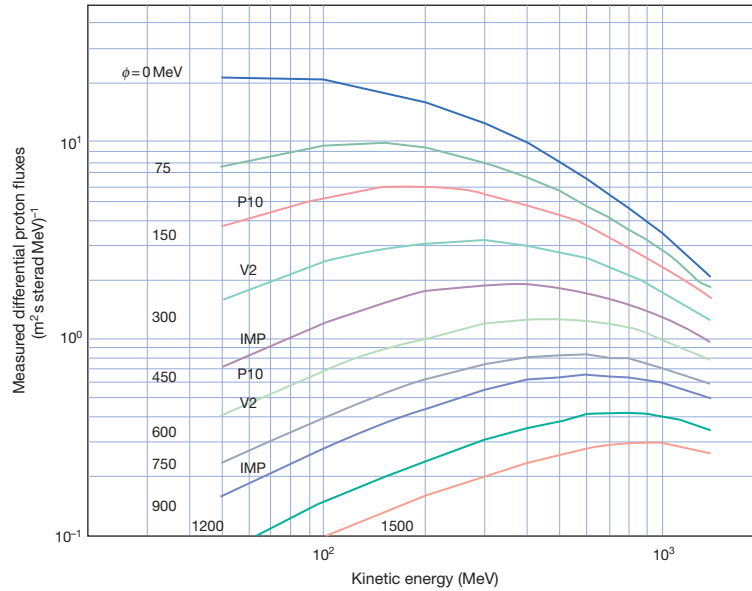


Figure 4 Observed cosmic ray fluxes of protons of kinetic energy >70 MeV by Pioneer 10 and Voyager 2 spacecrafts from 1978 to 1992 for GCR protons of 70 MeV kinetic energy, and near-Earth fluxes based on IMP satellite. The radial distances of Pioneer 10 and Voyager 2 were ~ 60 and 40 AU, respectively, in 1992. Redrawn from Lockwood JA and Webber WR (1995) An estimate of the location of modulation boundary for $E > 70$ MeV galactic cosmic rays using voyager and pioneer spacecraft data. *The Astrophysical Journal* 442: 852–860.

$$\frac{dn}{dE_k} = \text{constant} \times E_k^{-\Gamma} \quad [4]$$

However, the value of Γ changes rapidly with energy in the interval (1:100) MeV/n; it usually lies between 2 and 4. A much better fit is the exponential rigidity spectrum:

$$\frac{dn(R)}{dR} = \text{constant} \cdot e^{(-R/R_0)} \quad [5]$$

where R is the rigidity (Appendix A) and R_0 is a constant for the time-averaged spectrum of the solar particles in a flare event. The value of R_0 ranges between 50 and 200 MV in most flares (Lal, 1988b).

The solar cosmic ray (SCR) energy spectra are very soft compared to the GCR spectra, as can be seen from Figure 5, which presents the range of temporally observed variations in the GCR and SCR fluxes (Lal, 1988b). A quiet-time GCR spectrum in Figure 5 implies that the spectrum was not affected by SCR protons at the time of measurement.

As can be seen from Figure 5, the energy of the bulk of SCR proton flux is >100 MeV. The range of a proton of 100 MeV in air is <10 g cm^{-2} (Ziegler, 2004), only about 10% of an interaction mean free path for its nuclear interaction (Appendix B, Figure 6). Because of the geomagnetic cutoff (Appendix C), SCR protons are only incident at geomagnetic latitudes $>80^\circ$. Further, they are lost by nuclear interactions in the upper atmosphere. Any neutrons produced by them would however be slowed down reaching greater depths in the atmosphere and would mostly give rise to ^{14}C by getting captured in ^{14}N nucleus.

In view of the above, the principal effect of SCR is to produce ^{14}C in the atmosphere. Even here, detailed calculations by Castagnoli and Lal (1980) show that the SCR production of ^{14}C is generally negligible compared to its production

by GCR, but significant amounts could be produced in very large SCR events in some flares.

Development of the Secondary Particles of the Cosmic Radiation in the Earth's Atmosphere

A great deal is known about the basic processes which occur when energetic H, He, and heavier nuclei interact with matter, based on the early observations of nuclear interactions, and disintegrations of unstable particles as observed in photographic nuclear emulsions (Powell, 1959) and subsequent observations of high-energy particles accelerated up to ~ 30 GeV in particle accelerators. In Figures 7 and 8, we present examples of nuclear interactions of energetic protons and in Figure 9, an example of decay of an unstable particle (figures taken from Powell et al. (1959)). Several examples of first observations of interactions of multiply charged particles can be found in Bradt and Peters (1950).

A detailed description of the salient features of nuclear interactions is given by Lal and Peters (1967). We present here some of the important aspects of the propagation of the cosmic rays through the atmosphere.

The secondary cosmic ray phenomena are quite complex depending on the energy of the primary cosmic ray particle. Considering a group of interactions of an ensemble of cosmic ray primaries, the following is a good description of the secondary particles. The first-generation secondary particles are nucleons, pions, and a variety of unstable (elementary) particles. The nucleons and charged pions produce further nuclear interactions. Because of the low density of the atmosphere, most charged pions decay to muons before they interact. Neutral pions decay to photons which give rise to soft cascades (electrons and positrons). For properties of the elementary

particles, see [Review of Particle Physics \(2004\)](#). We note here that the elementary particles are classified under the headings hadrons and leptons. Mesons (π and K) are bosons with spin 0, while nucleons, hyperons, leptons, and neutrinos are fermions

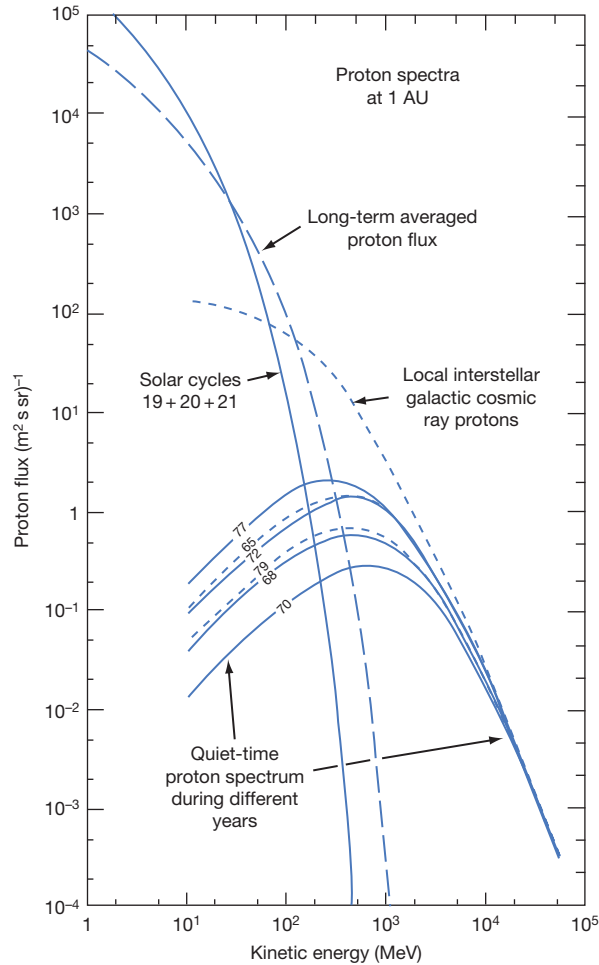


Figure 5 Observed solar modulated GCR fluxes of protons during different years (the last two digits of the year are shown), and the inferred local interstellar GCR proton spectrum are shown alongside solar cosmic ray (SCR) proton spectra. The time-averaged SCR proton spectra are shown for the solar cycles 19, 20, and 21 (1954–86) along with the long-term average spectrum based on observations of lunar rocks. Reproduced from Lal D (1988) Temporal variations of low energy cosmic ray protons on decadal and million year time scales: Implications of their origin. *Astrophysics and Space Science* 144: 337–346.

with spin 1/2. Among leptons are electrons, positrons, muon, and tauon. Note that muon is not called a meson because it is not a boson.

The fluxes of various cosmic ray secondary particles (nucleons, mesons, and electrons) change with depth in the atmosphere at geomagnetic latitude 45° and longitude 80° west (Lal and Peters, 1967; Peters, 1958; Figure 10). In this figure, the γ -ray flux is not shown; the photon:electron ratio is about unity at sea level and increases to a value of about 2 at 3500 m altitude. For approximate rates of nuclear disintegrations by nucleons and mesons, see Lal (1988a). It is seen from Figure 10 that at atmospheric depths of ~ 150 – 200 g cm^{-2} , there occurs a transition in the fluxes of nucleons, mesons, and electrons when these fluxes build up to a maxima due to nuclear cascades. At depths exceeding 200 g cm^{-2} , nucleons in different energy regions reach an equilibrium, the nucleonic cascade being sustained by high-energy nucleons of $E_k \geq 3 \text{ GeV}$.

Very detailed information on the generation of the secondary component in nuclear interactions became available from observations of secondary particles in nuclear emulsions exposed at different altitudes and latitudes in the atmosphere. From measurements of grain densities of tracks in nuclear interactions, Camerini et al. (1949) deduced the energy spectra of secondary particles produced in these interactions. These data are very useful for estimating the development of secondary cosmic ray particles in the atmosphere, and nuclear disintegration rates (Lal, 1988a; Powell et al., 1959; Rossi, 1952). From these data, Rossi (1952) has deduced the relative energy spectra of neutrons and protons at a depth of 700 g cm^{-2} in the atmosphere. This is in good agreement with the direct measurements of proton spectra, and the corresponding neutron fluxes based on the observed rates of nuclear disintegrations in nuclear emulsions by neutrons and protons (Lal, 1958; Lal and Peters, 1967). In Figure 11, we show the differential spectra of protons calculated in this way along with the measured proton spectrum at geomagnetic latitude 50° and atmospheric depth, 680 g cm^{-2} . The nuclear emulsion data provide the most direct means of obtaining neutron spectra in the atmosphere for kinetic energies in the MeV–GeV region; it is very difficult to obtain reliable neutron spectra using detectors, as was attempted by Hess et al. (1959).

It is seen from Figure 11 that the fluxes of protons and neutrons are about the same at $E_k \geq 1000 \text{ MeV}$, but proton fluxes are progressively smaller at lower energies than those of neutrons. They are produced in about the same numbers in nuclear interactions. However, protons produce fewer interactions as low-energy protons are brought to rest by ionization

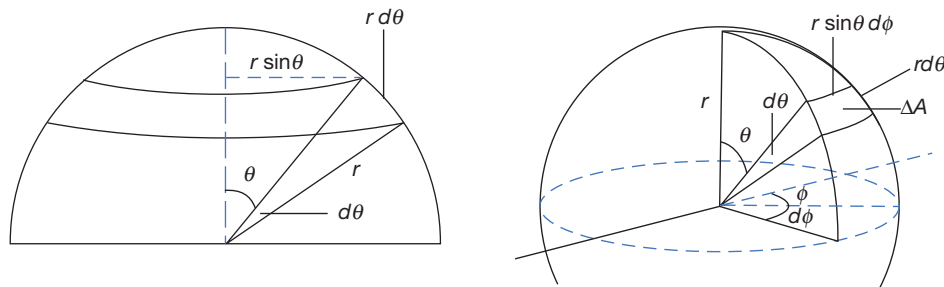


Figure 6 Coordinate system used in defining arrival directions of cosmic ray particles.

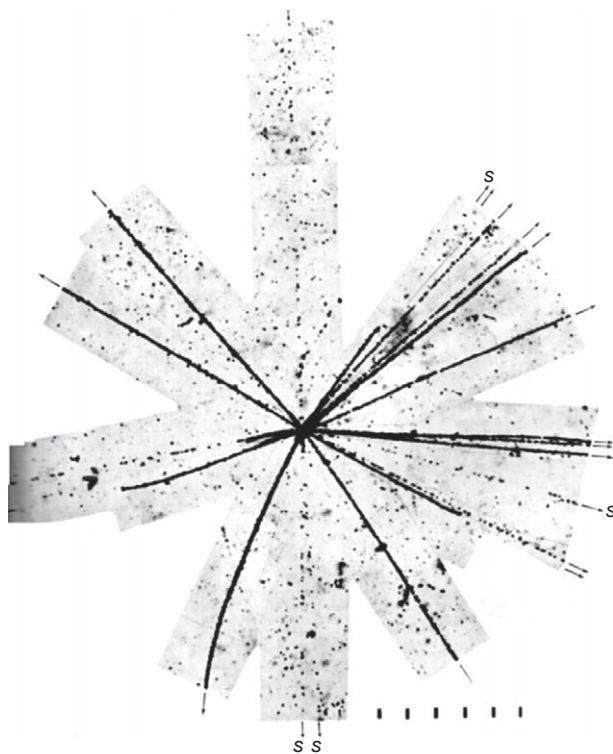


Figure 7 An example of a high-energy cosmic ray proton producing a nuclear interaction with a silver or bromine nucleus in the photographic nuclear emulsion. The incident proton track (shown by an arrow on the top) and other similarly appearing 'gray tracks' (five marked S for secondary) are due to energetic particles (protons and unstable particles) of low ionization rate. The 16 darker tracks, on the other hand, are due to slow protons and heavier ionization rates. Reproduced from Powell CF, Fowler PH, and Perkins DH (1959) *The Study of Elementary Particles by the Photographic Method*, p. 669. New York: Pergamon.

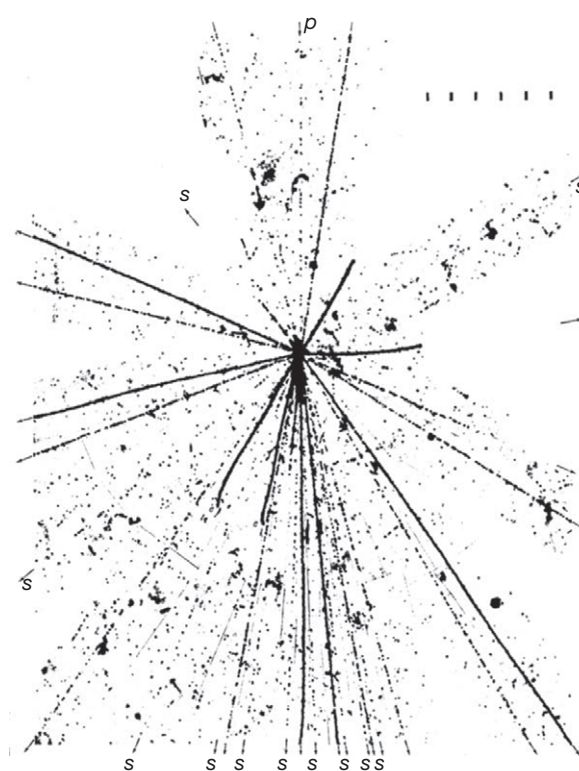


Figure 8 Example of an energetic nuclear interaction produced by a cosmic ray proton (estimated to be of ~ 30 GeV kinetic energy) with a silver or bromine nucleus in the photographic nuclear emulsion. The incident proton track is shown by an arrow labeled P on the top. Nine tracks of energetic mesons (gray tracks) and 22 darker tracks of different ionization rates are seen. Reproduced from Powell CF, Fowler PH, and Perkins DH (1959) *The Study of Elementary Particles by the Photographic Method*, p. 669. New York: Pergamon.

losses before they can interact. The ranges of a proton of 100, 500, and 1000 MeV in air are ~ 10 , 150, and 500 g cm^{-2} , respectively (Ziegler, 2004). The nuclear geometric and absorption mean free paths of nucleons are ~ 90 and 160 g cm^{-2} , respectively (Appendix B).

In the atmosphere, neutrons produce 90% of the nuclear interactions. The contribution of γ -radiation to nuclear interactions is about 5% at atmospheric depth of $\sim 200 \text{ g cm}^{-2}$ where its flux reaches a maximum (Figure 10); however, it can be neglected when considering the atmosphere as a whole. The contribution of muons to nuclear interactions becomes important below the atmosphere. In the following section, we present nuclear disintegration rates in the atmosphere and in the rocks and sediment of the upper tens of meters realm in the lithosphere.

Nuclear Disintegration Rates in the Atmosphere

The total rate of nuclear disintegrations in the atmosphere have been estimated, based on observed rates of nuclear disintegrations in nuclear emulsions and cloud chambers at different altitudes and latitudes, the measured low-energy neutron fluxes using neutron monitors located at several sites, and neutron fluxes measured in the atmosphere at different altitudes and

latitudes using boron trifluoride counters flown by balloons (Lal and Peters, 1967). In the upper atmosphere ($< 200 \text{ g cm}^{-2}$ depth), low-energy neutrons are not in equilibrium with fast nucleons. Furthermore, in this region, slow neutrons escape from the atmosphere. Even in the equilibrium region of the atmosphere ($> 200 \text{ g cm}^{-2}$ depth), only nucleons of $E_k < 400 \text{ MeV}$ are in equilibrium with slow neutrons. Taking into account all these factors, the total rates of nuclear disintegrations produced in the atmosphere with energy release $\geq 40 \text{ MeV}$ have been estimated (Lal, 1958, 1988c; Lal and Peters, 1967), as shown in Figure 12. These rates refer to a period of high solar activity since the neutron data come from observations during 1948–49.

The appreciable differences in the rate of nuclear interactions with latitude arise due to the modulation of cosmic ray flux by the Earth's geomagnetic field. As examples, the minimum kinetic energy which a vertically incident proton should have at geomagnetic latitudes of 0° , 30° , 60° , and 80° are, respectively, 14.0, 7.49, 0.384, and 0.0001 GeV (Appendix C).

The effect of geomagnetic field changes on nuclear disintegration rates can be seen from Figure 13, which shows the global mean nuclear disintegration rate as a function of geomagnetic field strength (relative to present field) and solar modulation parameter, ϕ .

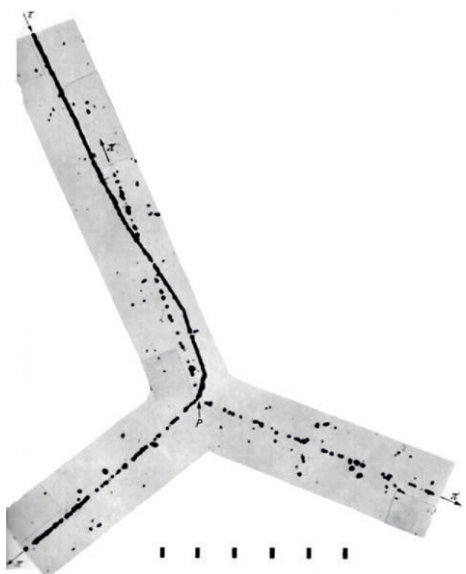


Figure 9 Example of decay of a τ -meson into three π -mesons in photographic nuclear emulsion. The darker wiggly track is due to τ -meson which comes to rest by ionization loss at the point P and decays giving rise to three π -meson tracks. Reproduced from Powell CF, Fowler PH, and Perkins DH (1959) *The Study of Elementary Particles by the Photographic Method*, p. 669. New York: Pergamon.

The minimum kinetic energy of a vertically incident proton at geomagnetic latitudes of 0° , 10° , 20° , 30° , 40° , 50° , 60° , 70° , 80° , and 90° are, respectively, 14.0, 13.1, 10.7, 7.49, 4.28, 1.77, 0.384, 0.02, 0.0001, and 0 GeV (Appendix C). The differential primary fluxes at these energies are given in Figure 2 for different states of solar modulation.

Temporal Variations in Cosmic Ray Nuclear Disintegration Rates

The incident cosmic ray flux at the top of the atmosphere is latitude dependent since it is strongly modulated by (1) the Earth's geomagnetic field and (2) solar plasma. Temporal variations in these parameters are therefore the two principal expected causes of changes in nuclear disintegration rates on the Earth. We do not expect to see any appreciable changes in the GCR intensity, except if a supernova explosion occurs nearby. This is supported by direct measurements of long-term constancy of cosmic radiation based on meteoritic evidence (Arnold et al., 1961), and the long galactic confinement of cosmic rays in the galaxy (~ 20 Ma; Simpson and Connell, 1998), combined with the fact that cosmic rays are replenished continually by supernovas whose frequency is quite high (~ 3 per century).

Temporal variations in nuclear disintegration rates due to changes in solar activity would depend strongly on altitude

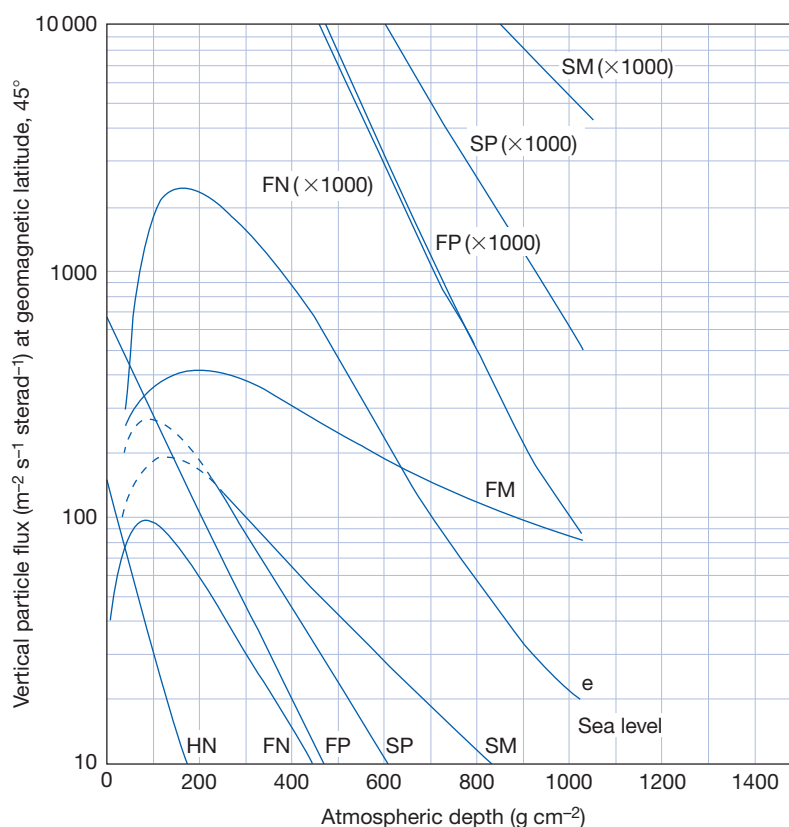


Figure 10 Composition of cosmic ray secondaries as a function of atmospheric depth at geomagnetic latitude 45° and longitude 80° west. The symbols in the figure are: HN (multiply charged (heavy) nuclei); FN (fast neutrons of $E_k \geq 3$ GeV); FP (fast protons of ≥ 3 GeV); SP (slow protons of $0.4 < E_k < 3$ GeV); FM (fast mesons of ≤ 220 MeV kinetic energy); SM (slow protons of $27 < E_k < 220$ MeV), and e ($E_k \geq 10$ MeV). Redrawn from Lal D and Peters B (1967) Cosmic ray produced radioactivity on the Earth. In: *Handbuch der Physik*, vol. 46(2), pp. 551–612. Berlin: Springer; see also Peters (1958).

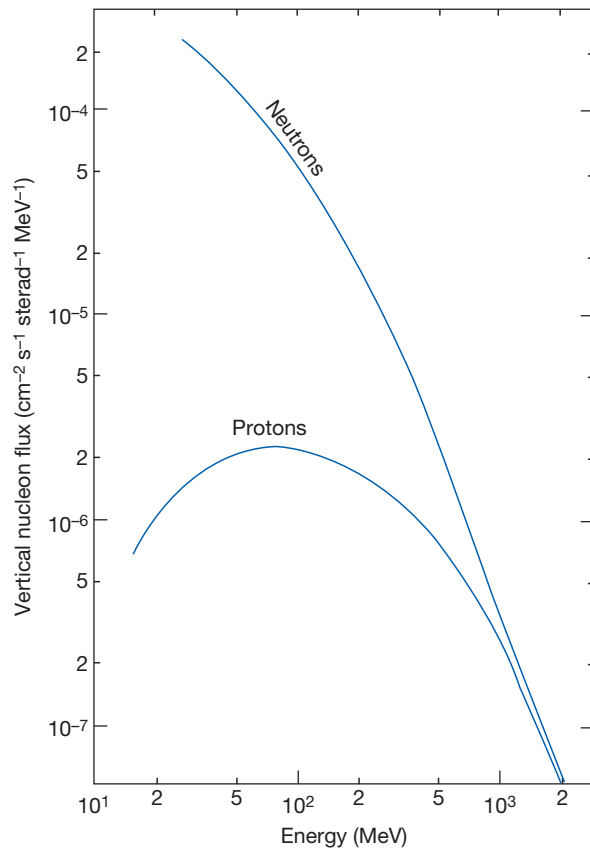


Figure 11 Differential kinetic energy spectra of protons (measured) and neutrons (estimated; see text) in the atmosphere at geomagnetic latitude 50° and depth 680 g cm^{-2} . Figure taken from Lal D and Peters B (1967) Cosmic ray produced radioactivity on the Earth. In: *Handbuch der Physik*, vol. 46(2), pp. 551–612. Berlin: Springer.

and latitude. Since the largest solar modulation effect is on the low energies, the temporal variations would be the greatest in the polar regions at high altitudes. As an example, the magnitude of change in cosmic ray intensity at different depths in the atmosphere between 1954 (a period of low solar activity) and 1958 (a period of unusually high solar activity) is shown in **Figure 15(b)**, page 579 of Lal and Peters (1967).

The effect of geomagnetic field changes on nuclear disintegration rates can be seen from **Figure 13**, which shows the global mean nuclear disintegration rate as a function of geomagnetic field strength (relative to present field) and solar modulation parameter, ϕ .

Nuclear Disintegration Rates in the Upper Lithosphere

As can be seen from **Figure 10**, the flux of fast neutrons decreases appreciably at sea level, decreasing with an absorption mean free path of $\sim 160 \text{ g cm}^{-2}$ (Gosse and Phillips, 2001; Lal and Peters, 1967), but the flux of fast mesons remains appreciable. In the atmosphere, most of the fast π -mesons decay before interacting, giving rise to fast muons. The latter component is absorbed slowly in the atmosphere; muons lose energy primarily by ionization, *bremsstrahlung* (braking radiation), and inelastic nuclear scattering. Loss of fast muons by nuclear interactions is not significant because they interact

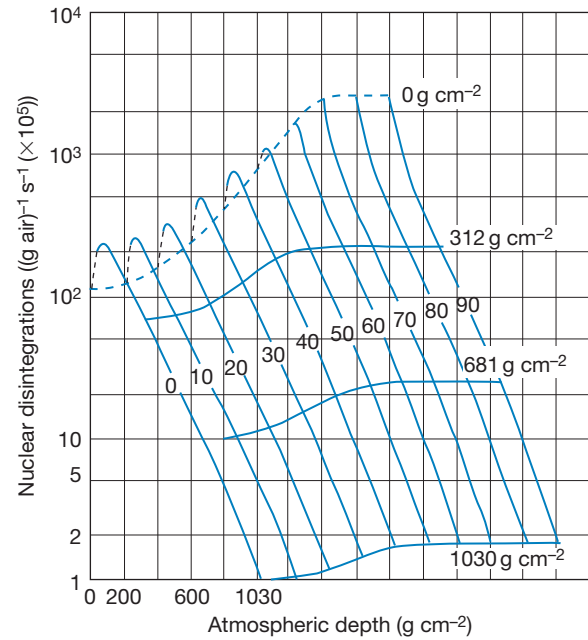


Figure 12 The total rate of nuclear disintegrations in the atmosphere, with energy release 40 MeV, is plotted as a function of altitude and geomagnetic latitude, in steps of 10° . Curves for different latitudes have been successively displaced along the abscissa by 1 km (based on Lal and Peters, 1967).

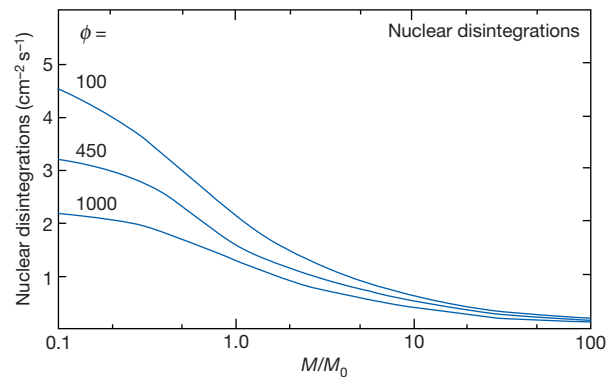


Figure 13 The global mean nuclear disintegration rate is plotted as a function of geomagnetic field strength, M (relative to present field, M_0) and solar modulation parameter $\phi = 100, 450$, and 1000 MeV .

with an effective (weakly energy dependent) Coulombic (photonuclear) cross-section of $\sim 3 \times 10^{-30} \text{ cm}^2$ per nucleon (cf. Heisinger et al., 2002a; Wolfendale et al., 1972). Compare this with geometric nuclear cross-sections, $\sim 4.5 \times 10^{-26} \text{ Å}^{2/3} \text{ cm}^2$ (Appendix B, eqn [B.5]). Negative muon captures by nuclei result in nuclear disintegrations, with energy release of a few tens of MeV at most. Energy release in nuclear interactions by fast muons is about an order of magnitude greater than in the capture of negative muons at rest.

Cosmic ray literature is extensive on cosmic rays underground (cf. Bhat and Ramana Murthy, 1973; Menon and Ramana Murthy, 1967). Cosmic ray muon intensities have been measured to underground depths of $\sim 3.7 \text{ km}$. In **Figure 14**, we present rates of nuclear disintegrations underground in

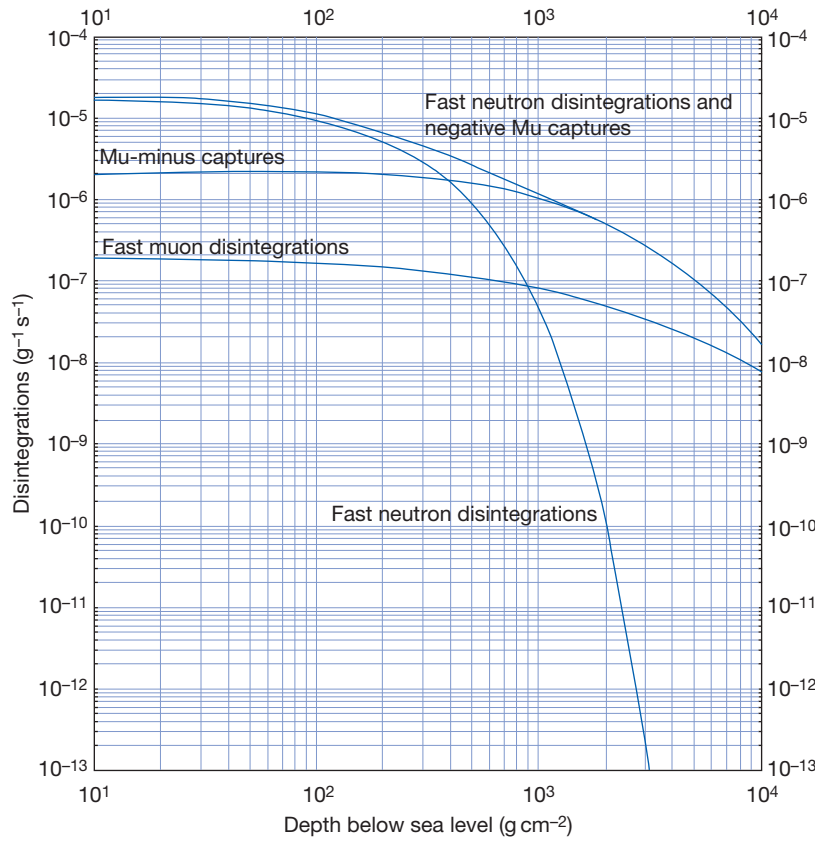


Figure 14 Nuclear disintegration rates due to fast neutrons and muons (fast muons and slow negative muons) in standard rock (density = 3.09 g cm^{-3} ; $Z^2/A = 6.3$). Reference is made to Lal (1987) for sources of data and the procedures used to determine the disintegration rates. Depths given are below sea level (1030 g cm^{-2}).

standard rock (density = 3.09 g cm^{-3} ; $Z^2/A = 6.3$), produced by neutrons and muons. For sources of data and the procedures used to determine the disintegration rates, reference is made to Lal (1987).

Rates of Nuclear Interactions in Targets Exposed at Different Altitudes in the Atmosphere

Thus far, we have presented rates of nuclear disintegrations in the atmosphere and in standard rock at underground depths. If a target is exposed in or at the bottom of the atmosphere, the rates of nuclear disintegrations in any mineral can be estimated. At altitudes within the atmosphere, nuclear interactions of fast muons are not important. Below, we discuss how nuclear disintegration rates for other particles can be estimated in minerals.

Fast Nucleons

The rates of nuclear disintegrations by fast nucleons in the target can be obtained by scaling the geometric cross-section according to eqn [B.5]. For a given omnidirectional cosmic ray flux, the number of interactions per gram target is proportional to the product $n\sigma$ (eqn [B.3]) (where n is the number of target atoms per g), and σ is the geometric nuclear interaction

cross-section. If, therefore, the rate of nuclear interactions due to fast neutrons is known for air, the nuclear interaction rate in the target would be the product of the nuclear reaction rate in air multiplied by a factor, f :

$$f = \left(\frac{A_{\text{target}}}{\bar{A}_{\text{atmosphere}}} \right)^{-1/3} \quad [6]$$

where $\bar{A}_{\text{atmosphere}}$ is the abundance weighted mean mass number of air, since n scales as A^{-1} and σ scales as $A^{2/3}$ (eqn [B.5]).

Slow Negative Muons

In the atmosphere, the rate of stopping negative muons, S_{μ^-} is given by:

$$S_{\mu^-} (\text{g}^{-1} \text{s}^{-1}) = 5.1 \times 10^{-6} e^{(-X/A_a)} \quad [7]$$

where X is the atmospheric depth (g cm^{-2}) and A_a is the absorption mean free path for negative muons (cf. Lal, 1988a,b,c); the value of A_a is 247 g cm^{-2} for $160 < X < 1030 \text{ g cm}^{-2}$. In light nuclei, the rate of capture of negative muons by the nucleus depends on the target mass number. Negative muons are quickly captured in the K-shell of the atom but in light nuclei a fraction of the decay occurs before they are captured by the nucleus. For details, reference is made to Figure 4 in Lal (1988b).

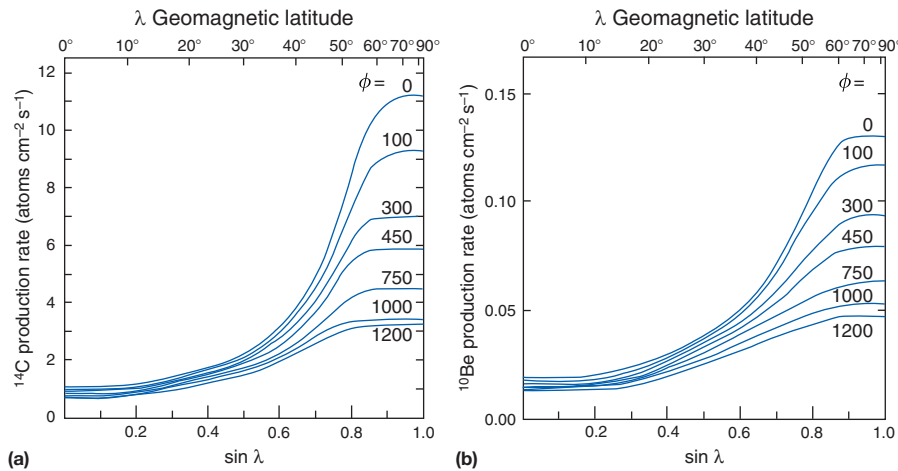


Figure 15 Total column production rates ($\text{cm}^{-2} \text{s}^{-1}$) of (a) ^{14}C and (b) ^{10}Be in the atmosphere are shown as a function of latitude for discrete values of the solar modulation parameter, ϕ .

Thermal Neutrons

In the atmosphere, most of the thermal neutrons are captured by N as it has the greatest abundance and highest cross-section among the atmospheric constituents. Rates of production of ^{14}C in the atmosphere during solar minimum and maximum periods observed during 1950–60 (Lingenfelter, 1963) are presented in Figures 11 and 12 in Lal and Peters (1967). From these data, one can calculate the thermal neutron fluxes within the atmosphere.

The differences in the strength of interaction of these three components of secondary radiation have been utilized to help determine the exposure ages and the erosion rates of a wide range of landforms and landscapes over a wide timescale (see Nishiizumi et al., 1993).

Rates of Production of Nuclides at Different Altitudes in the Atmosphere and at Depths in the Lithosphere

Detailed calculations of production rates of several nuclides are presented as a function of latitude and altitude by Lal and Peters (1967). The column-integrated global average production rates are also given by these authors. These estimates are based on three methods: (1) suitable (energy-dependent) normalization to nuclear disintegration rates in cases where absolute production rates are available at some point(s) in the atmosphere; (2) observed distribution of different-sized nuclear disintegrations in nuclear emulsions, and the partitioning of isotopes in nuclear spallation at different energies; and (3) partial cross-sections and the energy spectra of the nucleonic component. The first method is the most reliable; the second method fares much better than the third method because it is closest to the nuclear reaction dynamics and because the relevant cross-sections for neutrons, which produce most of the nuclides in the atmosphere, are generally few or are not available.

The latitudinal column production rates ($\text{cm}^{-2} \text{s}^{-1}$) of ^{14}C and ^{10}Be in the atmosphere are shown in Figure 15 as a function of latitude for discrete values of the solar modulation parameter, ϕ (Lal, 1988c).

At underground depths, fluxes of slow and fast muons are presented by Heisinger et al. (2002a,b), along with a large

number of cross-sections for production of a large number of nuclides from different targets up to depths of 10^5 g cm^{-2} . This is the most comprehensive summary available to date, based on cosmic ray and cross-section data. These data allow fairly detailed calculations of the production rates of the frequently measured radionuclides, ^{10}Be , ^{14}C , and ^{26}Al , by fast muons, and slow negative muons. The production of nuclides by fast neutrons is important only at depths $< 10^3 \text{ g cm}^{-2}$ (Figure 15; i.e., about 4 m of granite or 5 m of gravel).

The flux of thermal neutrons at sea level or near the atmosphere–lithosphere interface does not show a simple absorption. In fact, the flux is enhanced due to back-scattering neutrons produced in the crust/rock. Also, a depth transition occurs within the denser material in the neutron flux at depths of $< 50 \text{ g cm}^{-2}$ (Phillips et al., 2001), after which the thermal neutron flux decreases, following the absorption of the (parent) high-energy neutrons.

Rates of Production of Nuclides in Landforms Exposed at Different Altitudes in the Atmosphere and Lithosphere

The rate of production of a nuclide in a target is given by eqn [B.4]. As the rates of nuclear disintegration in the atmosphere are known, one can derive the nuclide production rates in any target using any of the three methods discussed in the previous section. In large targets, such as boulders or tors, however, one has to take into account the absorption of the secondary cosmic ray beam.

The relative nuclide production rates at a point, $P(x, y, z)$, within a body can then be obtained by integrating the following equation (Lal and Chen, 2005):

$$P(x, y, z) = \iint F(\theta, \text{surface}) \sin(\theta) \cdot e^{-\rho \vec{r}(x, y, z, \theta, \phi)/\Lambda} d\theta d\phi \quad [8]$$

where $\vec{r}(x, y, z, \theta, \phi)$ is the particular path length vector to reach the point (x, y, z) in the direction of incidence of the cosmic ray particle, θ , ϕ (see Appendix B for definitions of symbols). The parameter ρ signifies the mean density of the matrix and Λ is the mean absorption coefficient for nuclear interacting particles. A compilation of measured values of Λ

can be found in Gosse and Phillips (2001). From studies of neutrons and nuclear interactions in the equilibrium region of the atmosphere ($\chi > 200 \text{ g cm}^{-2}$), the angular dependence of F in the atmosphere is known to be fairly well represented by

$$F(\theta, \text{surface}) - F(0, \text{surface}) \cos^n(\theta) \quad [9]$$

in terms of the vertical flux, $F(0, \text{surface})$ in the atmosphere, that is, the cosmic ray flux on the surface of the object. The value of n is 2.3 ± 0.3 (Conversi and Rothwell, 1954; Lal, 1988a). Equation [3] readily adapts itself to an object whose dimension is changing as a result of erosion (Lal and Chen, 2005).

The distribution of path length vectors within a target can easily be derived. This task has been carried out earlier in similar contexts (Bhattacharya et al., 1973; Lal, 1972; Lal and Chen, 2005). For calculations of nuclide production rates in large boulders, see Masarik et al. (2000), Masarik and Wieler (2003), and Lal and Chen (2005). The latter authors have also considered relative nuclide production rates within rectangular bodies. They have concluded that since cosmic ray interactions are highly dependent on the geometry of exposure of the target, an examination of nuclide concentrations within a large target can be used to constrain the exposure history of objects, as their surfaces evolve due to weathering and erosion. This is a unique feature of the TCN 'exposure age dating,' vis-É-vis other nuclear methods (Lal, 1991).

Conclusions

In the past five decades and beyond, a great deal has been learned about the composition and fluxes of cosmic ray particles on the Earth, as well as about their most significant origins in the galaxy. The primary cosmic ray beam is composed of the ordinary matter we see in the universe, mostly H nuclei, since this is most abundant in the cosmic matter. The incidence of the cosmic radiation on the Earth's atmosphere leads to a very complex cascade of nucleons, mesons, muons, and a variety of elementary particles and soft cascades composed of electrons and positrons. In this article, the discussion has been confined to the primary cosmic rays and their composition; the composition of the secondary particles and their inelastic nuclear interactions leading to changes in the nuclear composition of matter. Knowledge of the primary cosmic radiation and its temporal variation due to solar modulation, and the development of the secondary cosmic ray beam are essential for accurately deducing the production of new forms of matter in cosmic ray interactions in the atmosphere as well as in solids in the lithosphere.

One may wonder how it is possible to deal practically with the above problem in spite of its complexity. The text shows that a fair amount of success has been achieved in this attempt, and there is no doubt that further improvements will be made in the future.

Appendix A Units Used in Cosmic Ray Physics

Since we are dealing with nuclear size particles, and since we generally deal with relativistic particles (velocities comparable to the speed of light, c), we work with a special unit system

appropriate to particle masses and energies. The units and important relations between particle velocities, momentum, rigidity, and energy are defined below.

The unit of energy is electron volt, the kinetic energy gained by an electron when it is raised in potential by 1 V (the energy gained by an He nucleus in being raised by 1 V would be 2 eV). One electron volt equals 1.6×10^{-12} erg of energy (one electron charge = 4.8×10^{-10} e.s.u. (1.6×10^{-20} e.m.u.) and $1 \text{ V} = 1/300$ e.s.u. (10^8 e.m.u.)).

At relativistic energies, when particle velocities are comparable to speed of light, the following relations hold between momentum, p , and rigidity, R (m is the particle mass at rest, and c is the speed of light in vacuo):

$$E = mc^2\gamma \quad [A.1]$$

$$\gamma = \frac{1}{\sqrt{1 - (v/c)^2}} \quad [A.2]$$

$$p = mv\gamma; \quad pc = mc^2\left(\frac{v}{c}\right)\gamma \quad [A.3]$$

$$R = \frac{pc}{Ze} \quad [A.4]$$

E is defined as the total energy (rest mass + kinetic energy, E_k). When vc , $\gamma \approx 1$, the total energy is close to its rest mass: m_0c^2 . At Newtonian velocities, vc , the kinetic energy can be seen to be equal to $(1/2)mv^2$, from binomial expansion of eqn [A.1].

The particle rigidity is so defined (eqn [A.4]), based on the movement of a charged particle in magnetic field, balancing the centripetal force on the particle with the Lorentz force (Rossi and Olbert, 1970). The units of E , p , and R are electron volt (eV), eV/ c , and volt (V), respectively.

Energy and momentum are related to each other by a very useful relation, as can be seen by using eqns [A.1] and [A.3]:

$$E^2 = (E_k + mc^2)^2 = (pc)^2 + (mc^2)^2 \quad [A.5]$$

Appendix B Definitions of Cosmic Ray Flux, and Geometric and Absorption Nuclear Interaction Rates

Crossing and Omnidirectional Fluxes

Unidirectional flux, $j(\theta, \phi)$, of cosmic ray particles arriving from any specified direction is defined as the number of particles per unit time, per unit solid angle, striking 1 cm^2 of a plane surface-oriented perpendicular to the incident direction; see Figure 6 for definitions of θ (zenith angle) and ϕ (azimuthal angle) shown schematically around a sphere of radius, r . It then follows (since differential solid angle, $d\omega = \Delta A/r^2$) that the 'total' number of particles striking a horizontal area, 1 cm^2 per unit time, F is given by:

$$F(\text{cm}^{-2} \text{ s}^{-1}) = \int_{\theta} \int_{\phi} j(\theta, \phi) \cos(\theta) \sin(\theta) d\theta d\phi \quad [B.1]$$

The $\cos(\theta)$ term in eqn [A.1] arises from the definition of 'flux.' It is often convenient to define an 'omnidirectional flux,' F_{omni} , the total number of particles striking an unit sphere of projected area 1 cm^2 from all directions, based on the definition of the unidirectional flux, $j(\theta, \phi)$:

$$F_{\text{omni}}(\text{cm}^{-2} \text{s}^{-1}) = \int_{\theta} \int_{\phi} j(\theta, \phi) \sin(\theta) d\theta d\phi \quad [\text{B.2}]$$

The secondary cosmic ray fast nucleon flux in the atmosphere at depths $>200 \text{ g cm}^{-2}$ shows a steep dependence with θ ($\cos^{2.3}(\theta)$) since particles incident at greater zenith angle traverse greater amount of matter in the atmosphere (cf. [Conversi and Rothwell, 1954](#)).

Nuclear Interaction Rates

The rate of nuclear reactions resulting in the production of a nuclide of interest in 1 g target, q (nuclei/g target/s), by particles of kinetic energy, E , in the specified target atoms are given by the relation

$$q(\text{nuclei/g target/s}) = F_{\text{omni}} n \sigma_i(E_k) \quad [\text{B.3}]$$

$$= n \int_{\theta} \int_{\phi} \int_{E_k} j(\theta, \phi, E_k) \sigma_i(E_k) r \sin(\theta) r d\theta d\phi \quad [\text{B.4}]$$

where n is the number of specified atoms in 1 g of the target, and $\sigma_i(E)$ is the kinetic energy-dependent partial cross-section for the formation of a nuclide. Note that in eqn [B.4], the term $\cos(\theta)$ in the double integral of eqn [B.1] is missing since particles incident at zenith angle, θ , have a longer path length in the unit target volume by a factor of $\sec(\theta)$. Here lies the utility of the concept of omnidirectional flux, F_{omni} .

Geometric Nuclear Interaction and Nuclear Absorption Cross-Sections

For fast nucleons (neutrons or protons), the total geometric nuclear interaction cross-section, σ_g , of a nucleus of mass number A is given by the following relation:

$$\sigma_g = \pi r^2 = \pi r_0^2 A^{2/3} \quad [\text{B.5}]$$

where r is the effective nuclear radius and r_0 is the nucleon radius, taken to be $1.20 \times 10^{-13} \text{ cm}$ ([Lang, 1980](#)). Equation [B.5] assumes constant density of nuclear matter in complex nuclei. The geometric mean free path for nuclear interaction, λ , is defined as (see eqn [A.3])

$$\lambda = \frac{1}{n\sigma_g} \quad [\text{B.6}]$$

The effective absorption cross-section for fast nucleons is smaller than the geometric nuclear interaction cross-section as the incident nucleon retains a large fraction of its initial kinetic energy (see text). If the elasticity, that is, the fraction of energy retained by the nucleon after interaction, is η , then it can be shown that for a power law nucleon differential spectrum of the form $a \times E_k^{-\Gamma}$ (the emerging nucleons retain the same spectral form, and are attenuated with an absorption mean free path, Λ :

$$\Lambda = \frac{\lambda}{1 - \eta^{\Gamma-1}} \quad [\text{B.6a}]$$

The corresponding effective absorption cross-section, σ_a is

$$\sigma_a = \sigma_g(1 - \eta^{\Gamma-1}) \quad [\text{B.7}]$$

The above equations are valid for treatment of propagating flux of secondary high-energy nucleons in the atmosphere produced from primary protons of $E_k \geq 3 \text{ GeV}$.

The calculated value of the geometric mean free path of nucleons in air is 90 g cm^{-2} . The corresponding values of the absorption mean free path of nucleons in air for $\eta = 0.50, 0.55$, and 0.60 are $155, 169$, and 187 g cm^{-2} , respectively. The value of Γ has been taken to be 2.5 in these calculations.

Appendix C Modulation of Cosmic Ray Flux by the Geomagnetic Field; Cutoff Rigidity and Kinetic Energy

Before entering the Earth's atmosphere, primary cosmic ray particles being charged are influenced by the Earth's magnetic field. The present geomagnetic field is well represented by a dipole of strength, $M = 8.1 \times 10^{25} \text{ G cm}^3$, located 342 km from the Earth's geometric center, with its axis intersecting the Earth's surface at 78.50° N , 69.00° W and 78.50° S , 111.00° E ([Sandström, 1965](#)). The geomagnetic field acts as a rigidity analyzer. For any hypothetical direction of arrival, one can define a particle momentum, p_u , such that particles of this momentum arriving from infinity can reach the given point from all directions. Similarly, one can also define a rigidity value, p_c , so that particles of lower momentum cannot arrive the given point in a given direction. For particle rigidities between p_u and p_c , there are 'forbidden' and 'allowed' bands of momenta. Reference is made to [Peters \(1958\)](#), [Sandström \(1965\)](#), and [Pomerantz \(1971\)](#) for further details.

If one neglects the fact that the magnetic dipole is shifted from the Earth's geometric center, and also the fact that the Earth is not a perfect sphere, the minimum cutoff momentum, p_c , which a particle must have is given by

$$p_c = \frac{M}{r_e^2} \frac{\cos^4(\lambda)}{[1 + \sqrt{1 - \sin(\theta) \cos(\phi) \cos^2(\lambda)}]^2} \quad [\text{C.1}]$$

$$= 59.6 \frac{\cos^4(\lambda)}{[1 + \sqrt{1 - \sin(\theta) \cos(\theta) \cos^2(\lambda)}]^2} \text{ GeV/c} \quad [\text{C.2}]$$

where r_e is the Earth's radius, θ and ϕ are respectively the zenith angle and azimuthal angles of the particle, and λ is the geomagnetic latitude. For vertical incidence, the corresponding cutoff rigidity, R_c (rigidity being defined as $R = p_c/Ze$; see [Appendix A](#)), is given by:

$$R_c = 14.9 \cos^4(\lambda) \text{ GV} \quad [\text{C.3}]$$

The corresponding vertical cutoff kinetic energy, E_c (kinetic), of the particle can be estimated using relation [A.5]. The minimum kinetic energy of a vertically incident proton at geomagnetic latitudes of $0^\circ, 10^\circ, 20^\circ, 30^\circ, 40^\circ, 50^\circ, 60^\circ, 70^\circ, 80^\circ$, and 90° are respectively $14.0, 13.1, 10.7, 7.49, 4.28, 1.77, 0.384, 0.02, 0.0001$, and 0 GeV .

The geomagnetic coordinates in the centered dipole approximation can be estimated from the corresponding geographical coordinates (cf. [Hopper, 1964](#); [Sandström, 1965](#)).

See also: [Cosmogenic Nuclide Dating: Exposure Geochronology](#); [Landscape Evolution](#); [Methods](#); [Overview](#).

References

- Arnold JR, Honda M, and Lal D (1961) Record of cosmic ray intensity in the meteorites. *Journal of Geophysical Research* 66: 3519–3531.
- Aschwanden MJ (2002) Particle acceleration and kinematics in solar flares. *Space Science Reviews* 101: 1–227.
- Bhat PN and Ramana Murthy PV (1973) Intensity measurements of low energy cosmic ray muons underground. *Journal of Physics A: Mathematical, Nuclear and General Physics* 6: 1960–1973.
- Bhattacharya SK, Goswami JN, Gupta SK, and Lal D (1973) Cosmic ray effects induced in a rock exposed on the moon or in free space: Contrast in patterns for tracks and isotopes. *The Moon* 8: 253–286.
- Bradt HL and Peters B (1950) The heavy nuclei of the primary cosmic radiation. *Physical Review* 77: 54–75.
- Camerini U, Coor T, Davies JH, et al. (1949) Nuclear transmutations produced by cosmic ray particles of great energy. *Philosophical Magazine* 40: 1073.
- Castagnoli G and Lal D (1980) Solar modulation effects in terrestrial production of carbon 14. Proceedings of the 10th International Radiocarbon Conference. *Radiocarbon* 22: 133–158.
- Connell JJ, DuVernois MA, and Simpson JJ (1998) The cosmic-ray radioactive nuclide ^{36}Cl and its propagation in the galaxy. *The Astrophysical Journal* 509: L97–L100.
- Conversi M and Rothwell P (1954) Angular distributions in cosmic ray stars at 3500 meters. *Nuovo Cimento* 12: 191–211.
- Fuji Z and McDonald FB (1997) Radial intensity gradients of galactic cosmic rays (1972–1995) in the heliosphere. *Journal of Geophysical Research* 102(A11): 24201–24208.
- Gleeson LJ and Axford WI (1968) Solar modulation of galactic cosmic rays. *The Astrophysical Journal* 423: 426–431; 154: 1011–1026.
- Gosse JC and Phillips FM (2001) Terrestrial *in situ* cosmogenic nuclides: Theory and application. *Quaternary Science Reviews* 20: 1475–1560.
- Heisinger B, Lal D, Jull ATJ, et al. (2002a) Production of selected cosmogenic radionuclides by muons. 1: Fast muons. *Earth and Planetary Science Letters* 200: 345–355.
- Heisinger B, Lal D, Jull ATJ, et al. (2002b) Production of selected cosmogenic radionuclides by muons. 2: Capture of negative muons. *Earth and Planetary Science Letters* 200: 357–369.
- Hess WN, Patterson HW, Wallace R, and Chupp EL (1959) Cosmic-ray neutron energy spectrum. *Physical Reviews* 116: 445–457.
- Higdon JC, Lingenfelter RE, and Ramaty R (1998) Cosmic ray acceleration from supernova ejecta in superbubbles. *The Astrophysical Journal* 509: L33–L36.
- Hopper VD (1964) New Jersey: Prentice Hall. p. 234.
- Lal D (1958) Investigations of Nuclear Interactions Produced by Cosmic Rays, p. 90. PhD Thesis, University of Bombay.
- Lal D (1972) Hard rock cosmic ray archaeology. *Space Science Reviews* 14: 3–102.
- Lal D (1987) Production of ^3He in terrestrial rocks. *Chemical Geology* 66: 89–98.
- Lal D (1988a) Temporal variations of low energy cosmic ray protons on decadal and million year time scales: Implications of their origin. *Astrophysics and Space Science* 144: 337–346.
- Lal D (1988b) *In-situ* produced cosmogenic isotopes in terrestrial rocks. *Annual Review of Earth and Planetary Science Letters* 16: 355–388.
- Lal D (1988c) Theoretically expected variations in the terrestrial cosmic ray production rates of isotopes. In: Castagnoli GC (ed.) *Solar Terrestrial Relationships and the Earth Environment in the Last Millennia, Italian Physical Society, Bologna*. Proceedings of the International School of Physics, ‘Enrico Fermi’ Course XCV, Varenna, June, 1985, pp. 216–233. Amsterdam: North Holland.
- Lal D (1991) Cosmic ray tagging of erosion surfaces: *In situ* production rates and erosion models. *Earth and Planetary Science Letters* 104: 424–439.
- Lal D (2000) Cosmogenic nuclide production rate systematics in terrestrial materials: Present knowledge, needs and future action for improvement. *Nuclear Instruments and Methods in Physics Research B* 172: 772–781.
- Lal D, Arnold JR, and Honda M (1960) Cosmic ray production rates of Be^7 in Oxygen and P^{32} , P^{33} , S^{35} in Argon at mountain altitudes. *Physical Reviews* 118: 1626–1632.
- Lal D and Chen J (2005) Cosmic ray labeling of erosion surfaces. II: Special cases of exposure histories of boulders, soils and beach terraces. *Earth and Planetary Science Letters* 236(3–4): 797–813.
- Lal D and Peters B (1967) Cosmic ray produced radioactivity on the Earth. In: Sitte K (ed.) *Handbuch der Physik*, vol. 46(2), pp. 551–612. Berlin: Springer.
- Lang KR (1980) *Astrophysical Formulae*, p. 783. Heidelberg: Springer.
- Lingenfelter RE (1963) Production of carbon-14 by cosmic ray neutrons. *Reviews of Geophysics* 1: 35–55.
- Lockwood JA and Webber WR (1995) An estimate of the location of modulation boundary for $E > 70\text{MeV}$ galactic cosmic rays using voyager and pioneer spacecraft data. *The Astrophysical Journal* 442: 852–860.
- Lukasiak A, Ferrando P, McDonald FB, and Webber WR (1994) The isotopic composition of cosmic aery beryllium and its implication for the cosmic ray age. *The Astrophysical Journal* 423: 426–431.
- Masarik J, Kollar D, and Vanya S (2000) Numerical simulation of *in situ* production of cosmogenic nuclides: Effects of irradiation geometry. *Nuclear Instruments and Methods B* 172: 786–789.
- Masarik J and Reedy RC (1995) Terrestrial cosmogenic-nuclide production systematics calculated from numerical simulations. *Earth and Planetary Science Letters* 176: 381–595.
- Masarik J and Wieler R (2003) Production rates of cosmogenic nuclides in boulders. *Earth and Planetary Science Letters* 216: 201–208.
- Menon MGK and Ramana Murthy PV (1967) Cosmic ray intensities deep underground. In: Wilson JG and Wothers SA (eds.) *Progress in Elementary and Cosmic Ray Physics*, vol. 9, pp. 163–243. Amsterdam: North Holland.
- Muzikar P, Elmore D, and Granger DE (2001) *Accelerator Mass Spectrometry in Geologic Research*, vol. 9, pp. 163–243. Amsterdam: North Holland.
- Nishiizumi K, Finkel RC, Klein J, and Kohl CP (1996) Cosmogenic production of ^7Be and ^{10}Be in water targets. *Journal of Geophysical Research* 101: 22225–22232.
- Nishiizumi K, Kohl CP, Dorn R, et al. (1993) Role of in-situ cosmogenic nuclides ^{10}Be and ^{26}Al in the study of diverse geomorphic processes. *Earth Surface Processes and Landforms* 18: 407–425.
- Nishiizumi K, Winterer EL, Kohl CP, et al. (1989) Cosmic ray production rates of ^{10}Be and ^{26}Al in quartz from glacially polished rocks. *Journal of Geophysical Research* 94(17): 907–916.
- Papini P, Grimani C, and Stephens SA (1996) An estimate of the secondary proton spectrum at small atmospheric depths. *Nuovo Cimento* 19C(3): 367–388.
- Peters B (1958) Cosmic rays. In: Condon EU and Odishaw H (eds.) *Handbook of Physics*, ch. 12, pp. 201–244. New York: McGraw-Hill.
- Phillips FM, Stone WD, and Fabryka-Martin JT (2001) An improved approach to calculating low-energy cosmic-ray neutron fluxes near the land/atmosphere interface. *Chemical Geology* 175: 689–701.
- Pomerantz MA (1971) *Cosmic Rays*, p. 186. New York: Van Nostrand Reinhold.
- Powell CF, Fowler PH, and Perkins DH (1959) *The Study of Elementary Particles by the Photographic Method*, p. 669. New York: Pergamon.
- Ramaty R, Kozlovsky B, and Lingenfelter RE (1998) *Physics Today* 4: 30–35.
- Randall BA and Van Allen JA (1986) Heliocentric radius of the cosmic ray modulation boundary. *Geophysical Research Letters* 13: 628–631.
- Review of Particle Physics (2004) In: Alvarez-Gaume L, et al. (ed.) *Physics Letters B* 592: 1–1109.
- Rossi B (1952) *High-Energy Particles*, p. 561. Englewood Cliffs: Prentice Hall.
- Rossi B and Olbert S (1970) *Introduction to the Physics of Space*, p. 454. New York: McGraw-Hill.
- Sandström AE (1965) *Cosmic Ray Physics*, p. 421. Amsterdam: North Holland.
- Simpson JA (1983) Elemental and isotopic composition of the galactic cosmic rays. *Annual Review of Nuclear and Particle Science* 33: 323–382.
- Simpson JA (1992) Cosmic radiation: Particle astrophysics in the heliosphere. *Annals of the New York Academy of Sciences* 655: 95–137.
- Simpson JA and Connell JJ (1998) Cosmic ray ^{26}Al and its decay in the galaxy. *The Astrophysical Journal* 497: L85–L88.
- Simunac KDC and Armstrong TP (2004) Solar cycle variations in solar and interplanetary ions observed with Interplanetary Monitoring Platform. *Journal of Geophysical Research* 109: A10101.
- Tuniz C, Bird JR, Fink D, and Herzog GF (1998) *Accelerator Mass Spectrometry*, p. 371. Washington, DC: CRC Press.
- Wolfendale AW, Young ECM, and Davis R (1972) Indirect observations of the photonuclear cross section above 20 GeV. *Nature Physical Science* 238: 130–131.
- Ziegler JF (1998) Terrestrial cosmic ray intensities. *IBM Journal of Research and Development* 42(1): 117–139.
- Ziegler JF (2004) SRIM 2003. *Nuclear Instruments and Methods* 219–220: 1027–1036.

Exposure Geochronology

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Introduction

Cosmogenic nuclides are produced *in situ* within minerals exposed at the earth's surface as a result of interactions of cosmic rays with the nuclei of target elements. Because cosmogenic nuclides build up predictably with time, measuring their concentrations allows the determination of how long a rock surface or sediment has been at or near the surface of the earth (Gosse and Phillips, 2001). The radionuclides ^{10}Be , ^{14}C , ^{26}Al , and ^{36}Cl and the noble gases ^3He and ^{21}Ne are the most commonly used cosmogenic nuclides today. With cosmogenic nuclide methods, rock surfaces and sediment can be directly dated (Figure 1). This is a unique and powerful tool available only to geomorphologists for roughly the last 25 years. Bedrock landforms, such as fluvially or glacially polished bedrock surfaces, fault footwall faces, and landslide bedrock detachment surfaces, can be dated directly. Sedimentary units such as moraines, landslide debris, fluvial terraces, debris-flow deposits, or alluvial fans can be directly dated by sampling boulder surfaces or by taking samples made up of numerous clasts or by taking sand samples. Geochronological questions that were previously difficult to solve due to the limitation of finding organic material for radiocarbon dating or by the 50-ka upper limit of radiocarbon dating can now be addressed with cosmogenic nuclides. Similarly, depositional landforms whose ages far exceed the upper limit or whose sediment does not fulfill the requirements of luminescence techniques (complete bleaching of the quartz or feldspar grains) can be studied. But more can be learned from the concentrations of cosmogenic nuclides than just the age. The dates themselves are used as a tool to glean fundamental information about rates and styles of geomorphic processes. In this article, the ways in which cosmogenic nuclides are used to date various landforms are described. In addition, examples where the obtained ages are used to calculate the rates of geomorphic processes are highlighted. A number of review articles provide excellent overviews as well as comprehensive reference lists (Bierman and Nichols, 2004; Bierman et al., 2002; Cockburn and Summerfield, 2004; Gosse and Phillips, 2001; Ivy-Ochs and Schaller, 2010; Morris et al., 2002).

Although the focus here is on the dating of landforms, cosmogenic nuclides have a largely untapped potential in archaeology, for example, the dating of man-made structures and lithic artifacts (Akçar et al., 2008 and references therein) and use of burial dating of sediments to date early hominids (Partridge et al., 2003).

Dating Rocks and Sediment with Cosmogenic Nuclides

The cosmogenic nuclide methods available to date landforms may be broadly divided into two categories: buildup and decay methods. The former includes exposure dating of boulders or

clasts and depth-profile dating with clast or sand samples. Decay methods include burial and isochron dating. The appropriate cosmogenic nuclide method for dating a given landform is largely dependent on the character of the deposit, which in turn depends on the depositional processes and the age and state of preservation of the landform; where possible, several methods are combined.

If large boulders are present on the top surface of the landform, these are often the most suitable sample material. In order to use the exposure age of a boulder surface to date a landform, the rock surface must have undergone single-stage (no preexposure), continuous (not covered) exposure in the same position (not shifted) since deposition. Thus, limitations are imposed on the landforms that can be dated; they must have formed rapidly and be no longer active. Exposure dating is then used to establish the end of the period of activity. In older deposits, surface degradation and boulder spalling and crumbling often lead to smoothing out of the original surface. In these cases, individual or amalgamated clast samples from the surface or along depth profiles are analyzed. For even older deposits (more than several 100 ka), burial or isochron dating methods are called upon.

In dating, the possibility of inherited nuclide concentrations (preexposure) must be considered. Inheritance may have built up in the bedrock setting or in the storage of sediment prior to deposition on the landform of interest. The amount of inheritance acquired in the bedrock setting is related to how deep in the bedrock the boulder originated and how long that bedrock surface was exposed. The effect on the calculated age (how much 'too old') depends on what proportion of the final measured concentration can be attributed to inherited nuclides. The inherited component will be proportionately greater in rocks stemming from long-exposed bedrock surfaces that are finally deposited in very young (Holocene) landforms.

Rock surfaces are constantly undergoing weathering (erosion). As a rock surface weathers, the outermost layers with the highest concentrations of cosmogenic nuclides are removed. This lowers the concentration and, consequently, the calculated apparent exposure age. As an aside, we note that, because of the shape of depth profile of ^{36}Cl production, ^{36}Cl ages on eroding surfaces may sometimes be apparently 'older' (see Gosse and Phillips, 2001). For calculating exposure ages from a single nuclide concentration, either an erosion rate of zero or a reasonable erosion rate for that rock surface is assumed. Crystalline rock erosion rates are generally less than 5 mm ka^{-1} (Cockburn and Summerfield, 2004).

Radionuclide concentrations build up in exposed minerals until secular equilibrium (saturation) is reached. At secular equilibrium, the number of nuclides produced per unit time is equal to the number that decays; the concentration is at steady state. This is shown by the flattening out of the nuclide



Figure 1 Examples of exposure dated landforms and rock surfaces. (a) Boulder dated from the Kromer moraine (Austria; Kerschner et al., 2006); view is down valley; moraine ridge continues to the right. (b) Pahoehoe flow top from the ca. 90 000-year-old (^3He) Bar Ten flow, Arizona (Fenton et al., 2001). (c) Beryllium-10 dating of boulders indicates that this lobe of the Shephard Creek fan in Owens Valley (California) is Holocene in age (Dühnforth et al., 2007); view is up into the catchment. (d) Continuous sampling of the Kaparelli limestone fault surface (Greece) for ^{36}Cl exposure dating (Benedetti et al., 2003).

concentration growth versus time curve after 3–4 half-lives (Figure 2). Both erosion and decay lead to loss of nuclides, and saturation is reached earlier (after a shorter exposure period; lower lines on Figure 2 for several different erosion rates).

In principle, when the concentrations of two nuclides from the same rock surface are measured, exposure time, erosion rate, and the presence of past coverage (complex exposure) may be revealed. An elegant way of portraying two-nuclide data is the erosion-island plot (Figure 3(a) and 3(b); Lal, 1991). Data for two nuclides with relatively long half-lives are well suited to study landforms in settings characterized by very low erosion rates and, thus, old rock surfaces (e.g., as in arid regions; see below). The erosion-island plot is especially useful for quick and clear identification of data points that represent continuous, single-stage exposure (plot in or near the erosion island) versus those that reflect complex exposure by burial. As a result of the difference in the half-lives of two nuclides (e.g., ^{26}Al and ^{10}Be), the length of time that sediments have been buried can be determined. When sediment is completely shielded, the changes in concentrations are dependent only on the half-lives. Aluminum decays faster, so the ratio $^{26}\text{Al}/^{10}\text{Be}$

decreases with time and the data points plot below the erosion island in the Al–Be erosion-island plot (Figure 3(c)) and above in the Be–C and Ne–Be plots (Figure 3(a) and 3(b)). It is important to note that there is no location on the diagram that can be reached by only one pathway. Data points for rock surfaces or sediment that were exposed and then buried can arrive back inside the erosion island after a long enough period of (re-)exposure. For points below the erosion island in the Al–Be plot, the minimum time to reach that point is made up of continuous exposure along the outer line followed by burial with complete shielding (zero production). All other possible scenarios (partial shielding or repeated burial and exposure) require more time (cf. Bierman et al., 1999). As seen in Figure 3(b), the use of ^{21}Ne (stable) along with ^{10}Be provides increased sensitivity for older, more slowly eroding surfaces. The recent addition of ^{14}C (Figure 3(a)) to the nuclide palette (Lifton et al., 2001) allows periods of burial during the last 10–15 thousand years to be identified (Goehring et al., 2011; Miller et al., 2006). Such recent intervals of burial cannot be accessed with the longer lived (or stable) nuclides.

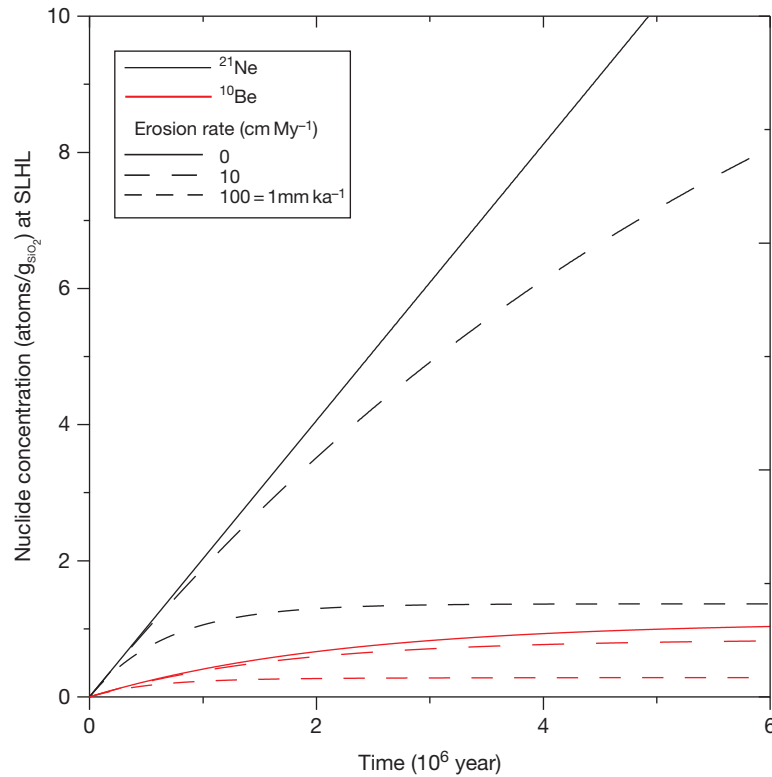


Figure 2 Increase in ^{10}Be and ^{21}Ne concentrations with time. With erosion, saturation is reached earlier and at lower concentrations. ^{10}Be production as in Balco et al. (2008), ^{21}Ne production rate after Kober et al. (2011), both at sea level and high latitude (SLHL). In the case of saturation, a minimum age or a maximum erosion rate can be derived.

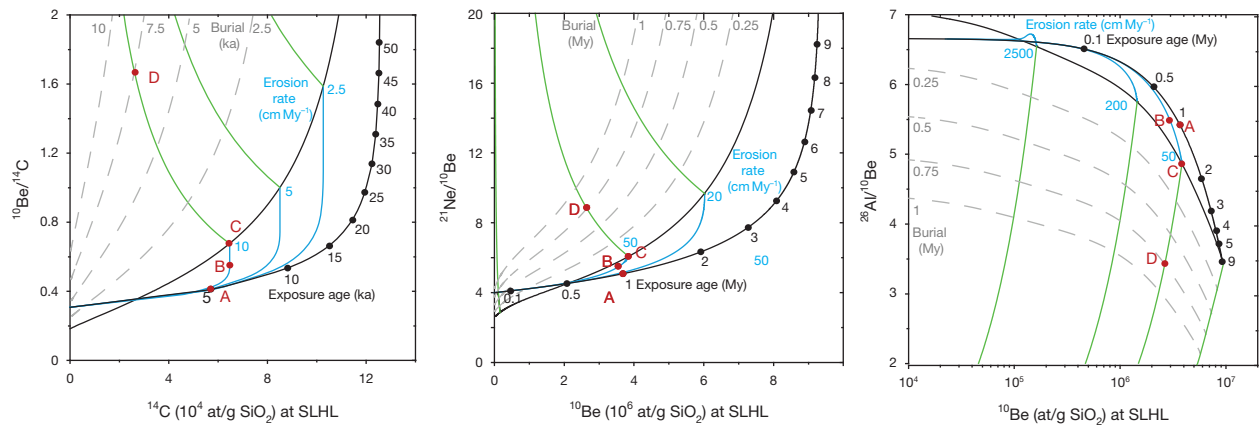


Figure 3 Erosion-island plots: $^{10}\text{Be}/^{14}\text{C}$ versus ^{10}Be (a); $^{21}\text{Ne}/^{10}\text{Be}$ versus ^{10}Be (b); $^{26}\text{Al}/^{10}\text{Be}$ versus ^{10}Be (c); all at sea level and high latitude (SLHL). With time, nuclide concentrations increase and, as the several nuclides decay with different rates, the nuclide ratio decreases (Al/Be) or increases (Ne/Be and Be/C) with time. For a continuously exposed rock surface with no erosion and no coverage, a data point would follow on the 'zero erosion' trajectory (sample point A). Periods of exposure are shown as black points on the outer black line. The blue lines inside the erosion island are trajectories for rocks continuously exposed and undergoing continuous erosion (sample point B). A data point on the erosion saturation line (inner black line) indicates secular equilibrium of both nuclides with respect to a specific erosion rate (sample point C). The field between the zero erosion line and the erosion saturation line is referred to as the erosion island. Data points may plot in the field below (Al/Be) or above (Ne/Be and Be/C) the erosion island if the rock surface has been buried (e.g., by ice or sediment) and/or if thick slabs (tens of centimeters) have spalled off (Bierman et al., 1999; Kober et al., 2007). Green lines show data point trajectories after burial. Note the very different temporal ranges that are accessible when ^{14}C is included. The combination $^{14}\text{C}/^{10}\text{Be}$ has its strongest advantage in complex exposure history studies covering the last 10 000 years.

The Spectrum of Dated Landforms

Glacial and Periglacial Landscapes

Glaciers respond rapidly to changes in temperature and precipitation and therefore are very sensitive indicators of climate change. A striking example is the nearly synchronous behavior of glaciers worldwide during the Little Ice Age. Prior to exposure dating, the timing of glacier size variations was approximated on the basis of correlation with local radiocarbon-dated records or with regional or hemispheric records, such as ice cores or marine sediment cores. But this fitting into an established chronology meant that the information on past climate change contained in the glacial record itself was lost. The direct dating of boulders on moraines with cosmogenic nuclides is one of the most powerful tools for reconstructing the timing of past glacier volume changes (Gosse, 2005).

The time of stabilization of a moraine is determined by measuring the concentration of cosmogenic nuclides in the surface of the largest boulders on the crest of the moraine (Figure 1(a)). Ideally, the signal recorded is the point of abandonment of the moraine by the glacier. Moraines ranging in age from hundreds of years (Schaefer et al., 2009) to hundreds of thousands of years have been dated. Limiting factors are the weathering of the boulder surface and the degradation of the moraine itself. Degradation of moraines as well as the associated exhumation and instability of boulders poses a severe limitation to accurate moraine dating (Putkonen and Swanson, 2003). In contrast, preexposure in moraine boulders is uncommon (Heyman et al., 2011; Putkonen and Swanson, 2003) but has been encountered (Briner et al., 2005). Reflecting the long-term instability of the landform, on older moraines that contain fewer boulders the spread in ages is greater (Kaplan et al., 2005). In dry climate regimes, for example in Patagonia, clasts on outwash deposits have been dated (Hein et al., 2009). The success of this approach highlights the inherent stability of the flat outwash deposit versus the instability of moraines. Similarly, as degradation leads to a change in the shape of a moraine with time, depth-profile dating in the till has proven difficult. Schaller et al. (2009) determined ages tens of thousands of years younger with depth profiling than ages determined originally by exposure dating of boulders on the moraine crest (Gosse, 2005).

Dating of glacially modified bedrock surfaces, such as roches moutonnées, may provide information about the timing of initial deglaciation of the valley bottom, from which retreat rates of glaciers are calculated (Guido et al., 2007). However, results from bedrock surfaces should be viewed with caution. Often, where excellent glacial polish is observed, the rock surface has been covered and protected, making it unsuitable for exposure dating (determined ages are less than the true age of formation). Reflecting the spatial variability of subglacial erosion, bedrock surfaces may contain inherited nuclides where less than about 2 m of rock was removed. Fabel et al. (2004) used this fact to estimate glacial erosion rates of bedrock at the bottom and along the sides of U-shaped valleys in the Sierra Nevada and Wind River Ranges. The presence of inheritance in glacially modified bedrock prompted Dühnforth et al. (2010) to highlight the importance of lithological control on glacial erosion. Two-nuclide data, for example, $^{26}\text{Al}/^{10}\text{Be}$ ratios, plot below the erosion island when

exposure was not continuous and the bedrock was (repeatedly) covered but not eroded by the ice (Bierman et al., 1999; Staiger et al., 2005). With *in situ* ^{14}C , periods of ice cover in the Holocene are determined (Goehring et al., 2011; Miller et al., 2006).

Cosmogenic nuclides are used to date the end of periods of activity and thus to pinpoint the melting out of permafrost from periglacial landforms, such as rock glaciers (Ivy-Ochs et al., 2009a) and block streams (Barrows et al., 2004). This affords the previously unrealized opportunity to decipher the climate signal contained in periglacial landforms. Periods of enhanced talus production are also determined with cosmogenic nuclides (Ballantyne and Stone, 2009).

Volcanic Terrains

Cosmogenic nuclides can be used to provide information about eruptive sequences of lava and ignimbrite flows. In addition, the dating of volcanic landforms allows the determination of fluvial incision rates and, when they are offset along a fault, slip rates. Exposure dating in mafic volcanic terrains has been carried out with the cosmogenic noble gases ^3He and/or ^{21}Ne (Cerling and Craig, 1994), because they are more suitable for analysis in mafic phenocrysts (olivine and pyroxene). Chlorine-36 is suitable for whole rock determinations on volcanic rocks, thus allowing dating of aphyric lavas (Phillips, 2003). To determine the eruptive age of the lava flow and develop a chronology for the flows in a volcanic field, original eruptive surfaces must be clearly identified; primary eruptive features, such as spatter and frothy glassy texture of the cooling rinds, are sought (Blard et al., 2005; Cerling and Craig, 1994; Fenton et al., 2001; Phillips, 2003; Figure 1(b)).

Alluvial Fans

Surface exposure dating is used to date the time of abandonment of alluvial fan lobe surfaces. Lobe abandonment is the result of fans moving laterally along a fault away from the locus of deposition, or of avulsion or fan head incision events. Alluvial fans build up through a variety of processes, including fluvial flow, overland flow, debris flow, rock avalanche, and rockfall. The architecture of alluvial fan sediments often records changes in the catchment driven by either climate or tectonics. The cosmogenic nuclide method appropriate for dating a given fan depends on the depositional processes and the age of the fan. Fan surfaces are dated to determine slip rates along faults (Meriaux et al., 2005; Siame et al., 1997) or in studies aimed at elucidating temporal variation of catchment processes (Dühnforth et al., 2007). Dates from fan abandonment times based on cosmogenic nuclides are compared with luminescence dates, which give the time elapsed since burial of the dated sediment (Guralnik et al., 2011). Much older fans are dated with cosmogenic nuclides (Nishiizumi et al., 2005) rather than with radiocarbon or luminescence dating.

Effective dating of fans requires that they are built up over a relatively short time interval, on the order of thousands of years (Hancock and Anderson, 2002). On younger surfaces, the largest boulders are dated (Figure 1(c); Dühnforth et al., 2007). On older alluvial fans (more than one or two hundred thousand years), degradation of fan surfaces and boulder

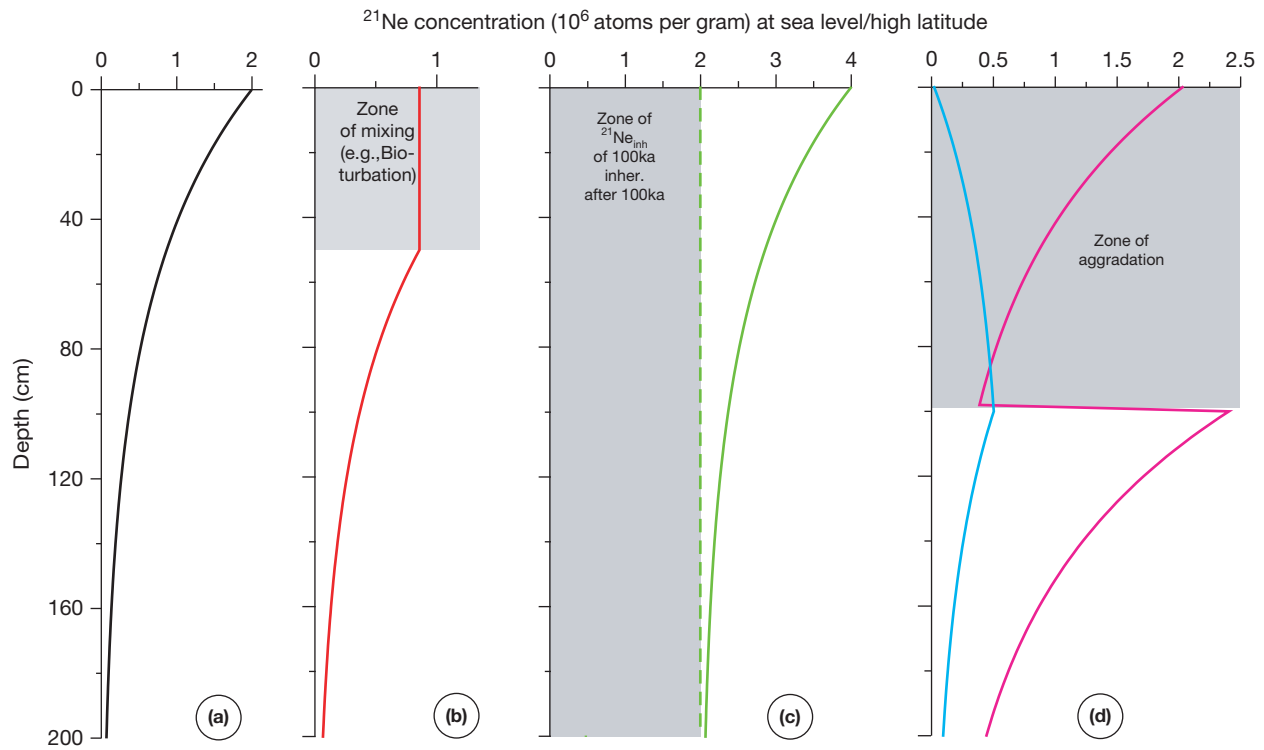


Figure 4 Depth profiles of ^{21}Ne reflecting various geomorphic processes. Depth profiles for alluvial landforms: erosion rate of top surface is zero. (a) Depth profile showing nuclide buildup with time. (b) A flat profile in the upper few tens of centimeters may indicate soil or regolith production, bioturbation, cryoturbation, or creep processes (Phillips et al., 1998). (c) The presence of inherited nuclide concentrations in the sampled sediment is indicated by the offset to higher concentrations at several meters depth (compare with (a)). (d) Two periods of aggradation; the blue line shows continuous exposure of the first unit (the lower section) followed by aggradation on top of it at a rate of centimeter per thousand years. The pink line portrays two separate single aggradation cycles each leaving 1 m of sediment.

weathering lead to smoothing out of the fan surface (Matmon et al., 2005). Such fans are dated with samples comprising individual clasts, tens of clasts, or sand samples from the surface or in depth profiles (Guralnik et al., 2010; Meriaux et al., 2005). For even older deposits (more than several 100 ka), burial or isochron dating methods are called upon. Samples are collected from the highest and flattest part of the surface away from diffusing slopes. Inheritance is a serious concern in the dating of alluvial fans. It could have been acquired during hillslope weathering in the catchment especially where if denudation rates are low. Also, clasts may have been exposed during intermittent storage in terraces in the catchment before final deposition on the fan. Cosmogenic nuclide results also show that many clasts do not come from the catchment during the event from which they are deposited. They may be scavenged from higher elevation fan lobes upstream (Ward et al., 2005).

Two considerations are crucial: (i) the amount of inheritance in the clasts and (ii) the degradation or aggradation rate of the top surface. These issues are addressed by measuring depth profiles that extend to several meters (Anderson et al., 1996; Braucher et al., 2009; Hidy et al., 2010). The shape of the concentration versus depth curve depends on the exposure age, the erosion (or aggradation) rate, and the amount of inheritance (Figure 4(c) and 4(d)). Where field relationships indicate that the top of the depositional surface has undergone little erosion (flat tread morphology as opposed to convex terrace

tops) or little aggradation (minimal silt layer below the top surface pavement), the shape of the depth curve reveals both age and inheritance (Figure 4(a)). Often, mixing of the upper 40–50 cm of sediment leads to the lack of a decrease in nuclide concentrations in the upper several tens of centimeters (Figure 4(b); Phillips et al., 1998). Depth-profile results that indicate separate depositional units (Figure 4(d)) are evaluated in the light of field evidence.

The dating of fluvial landforms is used to determine incision rates. The height and age of abandoned bedrock strath terraces (Burbank et al., 1996) or fluvial cut-fill (Ward et al., 2005) located along a given reach of a river provide information about fluvial incision rates and their variation with time. Sampled bedrock surfaces bear scars of fluvial erosion, provided they have not been covered for any extended period by sediment. Boulders or amalgamated clast samples on debris-flow deposits on ancient straths are dated to determine fluvial incision rates (Fenton et al., 2004). Fluvially sculpted channel walls may provide a more continuous record of incision than strath terraces (Schaller et al., 2005).

Rates of Tectonic Activity

Cosmogenic nuclides are used to determine rates of tectonic activity by means of two approaches: (i) by indirectly measuring the slip rates along faults or growing folds by determining

the ages of deformed or offset abandoned landforms; and (ii) by direct dating of bedrock fault surfaces produced during rupture events. Determined rates can be compared along various segments of a fault as well as through time.

Slip rates on strike-slip faults or rates of uplift on normal faults are determined by taking the amount of offset of the dated feature and its age. Offset alluvial fans (see above), fluvial terraces, river channels, diffused hillslopes, moraines, and lava flows have been dated for slip-rate determinations (Fenton et al., 2001; Meriaux et al., 2005).

With cosmogenic nuclides, the limestone bedrock surfaces created during normal faulting are dated directly with ^{36}Cl . Continuous sampling is done along an exposed bedrock fault face (Figure 4(d)). Nuclide concentrations are plotted versus height above the base of the scarp whereby a pattern of steps in concentration is revealed. The steps are interpreted to represent periods of exposure separated by earthquake-related rupture events (Benedetti et al., 2003; Mitchell et al., 2001; Zreda and Noller, 1998). Recent improvements to the calculations allow statistical evaluation of the concentration jumps (Schlenger et al., 2011).

Landslide Dating

Landsliding is a key process for the adjustment of steep (often glacially oversteepened) slopes, for modification of valley cross profiles, as well as for overall landscape denudation in mountainous terrain. By determining the timing of recurrence intervals, especially for high-magnitude events, we can begin to understand the mechanisms of large-scale downslope movement of rock, as well as the interplay between the various possible causes and triggers. New rock surfaces suitable for exposure dating are created during large deep-seated gravitational movements. Each possible sampling site (head scarps, sliding planes, large boulders in the deposit) has its own advantages and disadvantages with respect to exposure dating (Ivy-Ochs et al., 2009b). In the detachment scarp, post-slide rock fall can be difficult to identify. Shallow sliding planes may yield too young ages as a result of coverage by snow and/or vegetation, although when sliding planes dip steeply, consistent results may be obtained. Data from a number of sites where landslide boulders contain inheritance provide direct confirmation of modeling, which shows that large boulders are carried along on the surface during movement. Preexposure in moraine boulders is rare (Heyman et al., 2011), while in landslide boulders it is common (Ivy-Ochs et al., 2009b), especially in regions that were not glaciated, thus with higher nuclide concentrations in the pre-slide bedrock. Recent results dating historic landslides (Akcar et al., 2012) underline the importance of landslide dating for determining nuclide production rates.

Dating of Paleoshorelines

Ancient shorelines may be marked by wave-cut bedrock platforms or by scattered loose clasts aligned along ancient strandlines. To access the climate signal contained in the record of higher sea level or high lake stands in arid regions, a time frame

is required. When the time of formation is known, the height of uplifted bedrock platforms can be used to estimate coastal rock uplift rates (Alvarez-Marrón et al., 2007). Exposure dating of bedrock platforms can be confounded by the presence of inherited nuclides. This may mean that not enough rock was eroded by wave action to re-zero the bedrock (a minimum of several meters depending on nuclide), or that the platform was reoccupied during successive high sea levels, or both. At Lake Lisan, problems with inheritance, pointing to intermediate storage of clasts, were encountered in dating the paleostrandlines (Matmon et al., 2003).

Arid Regions

Cosmogenic nuclide methods are well suited to the dating of rock surfaces in the arid and hyperarid deserts of southern Africa (Fleming et al., 1999), south-central Australia (Belton et al., 2004), western South America (Kober et al., 2007; Nishiizumi et al., 2005), and Antarctica (Sugden et al., 2005). High concentrations of long-lived radionuclides (^{10}Be , ^{26}Al) and stable nuclides (^3He , ^{21}Ne) measured on bedrock or low-relief alluvial surfaces confirm the antiquity of these landscapes. Multistage exposure (with intervals of burial) and erosion rate histories can be deciphered by measuring different cosmogenic nuclides with different half-lives from individual surfaces (Kober et al., 2007). At several sites, data points plot within the erosion island, pointing to single-stage exposure history, but do not plot on the erosion saturation line (Figure 3) as would be expected for continuous exposure for millions of years. It is important to recognize that both burial and slab breakoff can lead to such nuclide ratios (Bierman et al., 1999; Kober et al., 2007). Maximum erosion rates for the rock surfaces are calculated by assuming that nuclide concentrations are in steady state. At several of these sites, erosion rates of 1 mm ka^{-1} or even less have been determined (Cockburn and Summerfield, 2004).

Cosmogenic nuclides have also been used to study sand dunes (Nishiizumi et al., 1993; Vermeesch et al., 2010). By analyzing ^{10}Be , ^{26}Al , and ^{21}Ne in sand grains, Vermeesch et al. (2010) found that the Namibia sand sea prevailed despite major variations in climate during the Quaternary.

Summary and Outlook

The ability to use cosmogenic nuclides to determine how long rock surfaces have been exposed at the surface of Earth provides an unrivaled tool for determining the ages of landforms and therefore the rates of geomorphic processes. Rocks on landforms that have been exposed for hundreds of years (landslides and moraines), hundreds of thousands of years (alluvial fans and river terraces), or to millions of years (ancient bedrock landforms) can be dated with cosmogenic nuclides. Only younger landforms (less than several hundred thousand years) are dated with exposure dating of boulders. For older landforms, surface amalgamated clast samples, sand clasts in depth profiles, or burial dating is required. This broad spectrum of available dating methods allows the dating of numerous depositional landforms as well as many bedrock

landforms. As knowledge of production rate and scaling formalities improves, there is improvement in the precision of obtained ages. But the accuracy of the ages remains a question of geological uncertainties. Both rock surfaces and landforms degrade with time. This imposes clear limitations on the useful time range as well as the accuracy of dating. Similarly, the natural variability of samples depends on landform morphology and age. Consequently, the obtained exposure ages must be evaluated individually. The ages must conform to field relationships such as terrace or moraine stratigraphy and regional morphostratigraphic relationships, as well as independent age constraints for the same or correlative features.

See also: **Cosmogenic Nuclide Dating: Methods; Overview.**

References

- Akçar N, Ivy-Ochs S, and Schlüchter C (2008) Application of in-situ produced terrestrial cosmogenic nuclides to archaeology: A schematic review. *Eiszeitalter und Gegenwart* 57: 226–238.
- Akcar N, Deline P, Ivy-Ochs S, et al. (2012) The 1717 AD rock avalanche deposits in the upper Ferret Valley (Italy): A dating approach with cosmogenic ^{10}Be . *Journal of Quaternary Science* 27: 383–392.
- Alvarez-Marrón J, Hetzel R, Niedermann S, Menéndez R, and Marquinez J (2007) Origin, structure and exposure history of a wave-cut platform more than 1 Ma in age at the coast of northern Spain: A multiple cosmogenic nuclide approach. *Geomorphology* 93: 316–334.
- Anderson RS, Repka JL, and Dick GS (1996) Explicit treatment of inheritance in dating depositional surfaces using ^{10}Be and ^{26}Al . *Geology* 24: 47–51.
- Balco G, Stone JO, Lifton NA, and Dunai TJ (2008) A complete and easily accessible means of calculating surface exposure ages or erosion rates from Be-10 and Al-26 measurements. *Quaternary Geochronology* 3: 174–195.
- Ballantyne CK and Stone JO (2009) Rock-slope failure at Baosbheinn, Wester Ross, NW Scotland: Age and interpretation. *Scottish Journal of Geology* 45: 177–181.
- Barrows TT, Stone JO, and Fifield LK (2004) Exposure ages for Pleistocene periglacial deposits in Australia. *Quaternary Science Reviews* 23: 697–708.
- Belton DX, Brown RW, Kohn BP, Fink D, and Farley KA (2004) Quantitative resolution of the debate over antiquity of the central Australian landscape: Implications for the tectonic and geomorphic stability of cratonic interiors. *Earth and Planetary Science Letters* 219: 21–34.
- Benedetti L, Finkel R, King G, et al. (2003) Motion on the Kaparelli fault (Greece) prior to the 1981 earthquake sequence determined from Cl-36 cosmogenic dating. *Terra Nova* 15: 118–124.
- Bierman PR, Caffee MW, Davis PT, et al. (2002) Rates and timing of earth surface processes from in-situ produced cosmogenic ^{10}Be . In: Grew E (ed.) *Beryllium: Mineralogy, Petrology, and Geochemistry. Reviews in Mineralogy and Geochemistry* 50: 147–204.
- Bierman PR, Marsella KA, Patterson C, Davis PT, and Caffee M (1999) Mid-Pleistocene cosmogenic minimum-age limits for pre-Wisconsinan glacial surfaces in southwestern Minnesota and southern Baffin Island: A multiple nuclide approach. *Geomorphology* 27: 25–39.
- Bierman P and Nichols KK (2004) Rock to sediment – Slope to sea with ^{10}Be -rates of landscape change. *Annual Reviews of Earth and Planetary Sciences* 32: 215–255.
- Blard P-H, Lave J, Pik R, Quidelleur X, Bourles D, and Kieffer G (2005) Fossil cosmogenic ^3He record from K–Ar dated basaltic flows of Mount Etna volcano (Sicily, 38° N): Evaluation of a new paleoaltimeter. *Earth and Planetary Science Letters* 236: 613–631.
- Braucher R, Del Castillo P, Siame L, Hidy AJ, and Bourlés DL (2009) Determination of both exposure time and denudation rate from an in situ-produced ^{10}Be depth profile: A mathematical proof of uniqueness. Model sensitivity and applications to natural cases. *Quaternary Geochronology* 4: 56–67.
- Briner JP, Kaufmann DS, Manley WF, Finkel RC, and Caffee MW (2005) Cosmogenic exposure dating of late Pleistocene moraine stabilization in Alaska. *Geological Society of America Bulletin* 117: 1108–1120.
- Burbank DW, Leland J, Fielding E, Anderson RS, Brozovic N, Reid MR and Duncan C (1996) Bedrock incision, rock uplift and threshold hillslopes in the northwestern Himalayas. *Nature* 379: 505–510.
- Cerling TE and Craig H (1994) Geomorphology and in-situ cosmogenic isotopes. *Annual Reviews of Earth and Planetary Sciences* 22: 273–317.
- Cockburn HAP and Summerfield MA (2004) Geomorphological applications of cosmogenic isotope analysis. *Progress in Physical Geography* 28: 1–42.
- Dühnforth M, Anderson RS, Ward D, and Stock GM (2010) Bedrock fracture control of glacial erosion processes and rates. *Geology* 38: 423–426.
- Dühnforth M, Densmore AL, Ivy-Ochs S, Allen PA, and Kubik PW (2007) Timing and patterns of debris flow deposition on Shepherd and Symmes creek fans, Owens Valley, California, deduced from cosmogenic Be-10. *Journal of Geophysical Research-Earth Surface* 112: F03S15.
- Fabel D, Harbor J, Dahms D, et al. (2004) Spatial patterns of glacial erosion at a valley scale derived from terrestrial cosmogenic Be-10 and Al-26 concentrations in rock. *Annual of Association of American Geographers* 94: 241–255.
- Fenton CR, Poreda R, Nash BP, Webb RH, and Cerling TE (2004) Geochemical discrimination of five Pleistocene lava-dam outburst-flood deposits, western Grand Canyon, Arizona. *Journal of Geology* 112: 91–110.
- Fenton CR, Webb RH, Pearthree PA, Cerling TE, and Poreda RJ (2001) Displacement rates on the Toroweap and Hurricane faults: Implications for Quaternary downcutting in the Grand Canyon, Arizona. *Geology* 29: 1035–1038.
- Fleming A, Summerfield MA, Stone JO, Fifield LK, and Cresswell RG (1999) Denudation rates for the southern Drakensberg escarpment, SE Africa, derived from in-situ-produced cosmogenic Cl-36 : Initial results. *Journal of the Geological Society* 156: 209–212.
- Goehring BM, Schaefer JM, Schluechter C, et al. (2011) The Rhone Glacier was smaller than today for most of the Holocene. *Geology* 39: 679–682.
- Gosse JC (2005) The contributions of cosmogenic nuclides to unraveling alpine paleo-glacier histories. In: Huber UM, Bugmann HKM, and Reaser MA (eds.) *Global Change and Mountain Regions. An Overview of Current Knowledge. Advances in Global Change Research*, vol. 23, pp. 39–50. Dordrecht: Springer.
- Gosse JC and Phillips FM (2001) Terrestrial in situ cosmogenic nuclides: Theory and application. *Quaternary Science Reviews* 20: 1475–1560.
- Guido ZS, Ward DJ, and Anderson RS (2007) Pacing the post-Last Glacial Maximum demise of the Animas Valley glacier and the San Juan Mountain ice cap, Colorado. *Geology* 35: 739–742.
- Guralnik B, Matmon A, Avni Y, and Fink D (2010) ^{10}Be exposure ages of ancient desert pavements reveal Quaternary evolution of the Dead Sea drainage basin and rift margin tilting. *Earth and Planetary Science Letters* 290: 132–141.
- Guralnik B, Matmon A, Avni Y, Porat N, and Fink D (2011) Constraining the evolution of river terraces with integrated OSL and cosmogenic nuclide data. *Quaternary Geochronology* 6: 22–32.
- Hancock GS and Anderson RS (2002) Numerical modeling of fluvial strath-terrace formation in response to oscillating climate. *Geological Society of America Bulletin* 114: 1131–1142.
- Hein AS, Hulton NRJ, Dunai TJ, et al. (2009) Middle Pleistocene glaciation in Patagonia dated by cosmogenic-nuclide measurements on outwash gravels. *Earth and Planetary Science Letters* 286: 184–197.
- Heyman J, Stroeve AP, Harbor JM, and Caffee MW (2011) Too young or too old: Evaluating cosmogenic exposure dating based on an analysis of compiled boulder exposure ages. *Earth and Planetary Science Letters* 302: 71–80.
- Hidy AJ, Gosse JC, Pederson JL, Mattern JP, and Finkel RC (2010) A geologically constrained Monte Carlo approach to modeling exposure ages from profiles of cosmogenic nuclides: An example from Lees Ferry, Arizona. *Geochemistry, Geophysics, Geosystems* 11: Q0AA10.
- Ivy-Ochs S, Kerschner H, Maisch M, Christl M, Kubik PW, and Schlüchter C (2009a) Latest Pleistocene and Holocene glacier variations in the European Alps. *Quaternary Science Reviews* 28: 2137–2149.
- Ivy-Ochs S and Schaller M (2010) Examining processes and rates of landscape change with cosmogenic radionuclides. In: Froehlich K (ed.) *Environmental Radionuclides – Tracers and Timers of Terrestrial Processes*, pp. 232–294. Amsterdam: Elsevier.
- Ivy-Ochs S, von Poschinger A, Synal H-A, and Maisch M (2009b) Surface exposure dating of the Flims rockslide, Graubünden, Switzerland. *Geomorphology* 103: 104–112.
- Kaplan MR, Douglass DC, Singer BS, and Caffee MW (2005) Cosmogenic nuclide chronology of pre-last glacial maximum moraines at Lago Buenos Aires, 46 degrees S, Argentina. *Quaternary Research* 63: 301–315.
- Kerschner H, Hertl A, Gross G, Ivy-Ochs S, and Kubik PW (2006) Surface exposure dating of moraines in the Kromer valley (Silvretta Mountains, Austria) – Evidence for glacial response to the 8.2 ka event in the Eastern Alps? *The Holocene* 16: 7–15.

- Kober F, Alfimov V, Ivy-Ochs S, Kubik PW, and Wieler R (2011) The cosmogenic ^{21}Ne production rate in quartz evaluated on a large set of existing ^{21}Ne – ^{10}Be data. *Earth and Planetary Science Letters* 302: 163–171.
- Kober F, Ivy-Ochs S, Schlunegger F, Baur H, Kubik PW, and Wieler R (2007) Denudation rates and a topography-driven rainfall threshold in northern Chile: Multiple cosmogenic nuclide data and sediment yield budgets. *Geomorphology* 83: 97–120.
- Lal D (1991) Cosmic ray labeling of erosion surfaces: In-situ nuclide production rates and erosion models. *Earth and Planetary Science Letters* 104: 424–439.
- Lifton NA, Jull AJT, and Quade J (2001) A new extraction technique and production rate estimate for in situ cosmogenic ^{14}C in quartz. *Geochimica et Cosmochimica Acta* 65: 1953–1969.
- Matmon A, Crouvi O, Enzel Y, et al. (2003) Complex exposure histories of chert clasts in the late Pleistocene shorelines of Lake Lisan, southern Israel. *Earth Surface Processes and Landforms* 28: 493–506.
- Matmon A, Schwartz DP, Finkel R, Clemmens S, and Hanks T (2005) Dating offset fans along the Mojave section of the San Andreas fault using cosmogenic Al-26 and Be-10. *Geological Society of America Bulletin* 117: 795–807.
- Meriaux AS, Tapponnier P, Ryerson FJ, et al. (2005) The Aksay segment of the northern Altyn Tagh fault: Tectonic geomorphology, landscape evolution, and Holocene slip rate. *Journal of Geophysical Research-Solid Earth* 110(B4): B04404.
- Miller GH, Briner JP, Lifton NA, and Finkel RC (2006) Limited ice-sheet erosion and complex exposure histories derived from in situ cosmogenic ^{10}Be , ^{26}Al , and ^{14}C on Baffin Island, Arctic Canada. *Quaternary Geochronology* 1: 74–85.
- Mitchell SG, Matmon A, Bierman PR, Enzel Y, Caffee M, and Rizzo D (2001) Displacement history of a limestone normal fault scarp, northern Israel, from cosmogenic Cl-36. *Journal of Geophysical Research-Solid Earth* 106: 4247–4264.
- Morris JD, Gosse JC, Brachfield S, and Tera F (2002) Cosmogenic Be-10 and the solid earth: Studies in Geomagnetism, subduction zone processes, and active tectonics. In: Grew E (ed.) Beryllium: Mineralogy, Petrology, and Geochemistry. *Reviews in Mineralogy and Geochemistry* 50: 207–270.
- Nishiizumi K, Caffee M, Finkel RC, Brimhall G, and Mote T (2005) Remnants of a fossil alluvial fan landscape of Miocene age in the Atacama Desert of northern Chile using cosmogenic nuclide exposure age dating. *Earth and Planetary Science Letters* 237: 499–507.
- Nishiizumi K, Kohl JR, Arnold JR, et al. (1993) Role of in situ cosmogenic nuclides ^{10}Be and ^{26}Al in the study of diverse geomorphic processes. *Earth Surface Processes* 18: 407–425.
- Partridge TC, Granger DE, Caffee MW and Clarke RJ (2003) Lower Pliocene hominid remains from Sterkfontein. *Science* 300: 607–612.
- Phillips FM (2003) Cosmogenic Cl-36 ages of Quaternary basalt flows in the Mojave Desert, California, USA. *Geomorphology* 53: 199–208.
- Phillips WM, McDonald EV, Reneau SL, and Poths J (1998) Dating soils and alluvium with cosmogenic ^{21}Ne depth profiles: Case studies from the Pajarito Plateau, New Mexico, USA. *Earth and Planetary Science Letters* 160: 209–223.
- Putkonen J and Swanson T (2003) Accuracy of cosmogenic ages for moraines. *Quaternary Research* 59: 255–261.
- Schaefer JM, Denton GH, Kaplan M, et al. (2009) High-frequency Holocene glacier fluctuations in New Zealand differ from the northern signature. *Science* 324: 622–625.
- Schaller M, Ehlers TA, Blum JD, and Kallenberg MA (2009) Quantifying glacial moraine age, denudation, and soil mixing with cosmogenic nuclide depth profiles. *Journal of Geophysical Research* 114: F01012.
- Schaller M, Hovius N, Willet SD, Ivy-Ochs S, Synal HA, and Chen H-C (2005) Fluvial bedrock incision in the active mountain belt of Taiwan from in situ-produced cosmogenic nuclides. *Earth Surface Processes and Landforms* 30: 955–971.
- Schlagenhauf A, Manighetti I, Benedetti L, et al. (2011) Earthquake supercycles in Central Italy, inferred from ^{36}Cl exposure dating. *Earth and Planetary Science Letters* 307: 487–500.
- Siame LL, Bourles DL, Sebrier M, et al. (1997) Cosmogenic dating ranging from 20 to 700 ka of a series of alluvial fan surfaces affected by the El tigre fault, Argentina. *Geology* 25: 975–978.
- Staiger JKW, Gosse JC, Johnson JV, et al. (2005) Quaternary relief generation by polythermal glacier ice. *Earth Surface Processes and Landforms* 30: 1145–1159.
- Sugden DE, Balco G, Cowdery SG, Stone JO, and Sass LC (2005) Selective glacial erosion and weathering zones in the coastal mountains of Marie Byrd Land, Antarctica. *Geomorphology* 67: 317–334.
- Vermeesch P, Fenton CR, Kober F, Wiggs GFS, Bristow CS, and Xu S (2010) One million year residence time of Namib dune sand measured with cosmogenic nuclides. *Nature Geoscience* 3: 862–865.
- Ward DJ, Spotila JA, Hancock GS, and Galbraith JM (2005) New constraints on the late Cenozoic incision history of the New River, Virginia. *Geomorphology* 75: 54–72.
- Zreda M and Noller JS (1998) Ages of prehistoric earthquakes revealed by cosmogenic chlorine-36 in a bedrock fault scarp at Hebgen Lake. *Science* 282: 1097–1099.

Landscape Evolution

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Glossary

Cosmogenic nuclide: A type of atom produced by a nuclear reaction with a secondary cosmic ray.

Muon: A type of secondary cosmic ray particle.

Nucleon: A neutron or a proton; a type of secondary cosmic ray particle.

Radioactive mean life: The time taken for a radioactive material to decay to $1/e$ of its abundance.

Saprolite: Weathered bedrock that retains some original structure.

Spallation: A type of nuclear reaction in which nucleons are lost from a nucleus in a collision.

Introduction

Landscapes evolve over time as material is eroded from bedrock, transported down hillslopes and river systems, and deposited in valleys or basins. River incision and aggradation transmit base-level change through a watershed and help determine how landscapes respond to external factors such as tectonic uplift or climate change. To understand how landscapes change over time, it is important to characterize both erosion rates and valley lowering rates. It is also important to establish the timing of major events in the landscape's history to understand its dynamic response to changes in tectonics, climate, or other factors. Cosmogenic nuclides have played a key role in quantifying both erosion rates and the timing of landscape evolution.

Until recently, the task of measuring erosion rates had been very difficult. This was because eroding landscapes leave little sedimentary evidence of their past. Erosion rates, after all, are a measure of how fast material has been removed from a landscape. Most of the eroded material is either transported away or dissolved. It is possible to estimate short-term erosion rates from sediment and solute loads in streams, but sediment loads are susceptible to rapid fluctuations and their measurements are often fraught with error. Moreover, modern sediment loads are strongly influenced by land use. Prior to cosmogenic nuclide methods, the issue of land use was so daunting that it was considered questionable that erosion rates could ever be measured reliably in large parts of the world.

Today, the application of cosmogenic nuclide methods allows us to circumvent the problems of land use and to measure erosion rates over timescales of thousands of years. This is because the cosmogenic nuclides present in rock or soil today were produced by secondary cosmic rays passing through the rock or sediment above them in the past. Cosmogenic nuclide concentrations allow us to infer the presence of an overburden that is no longer there and how quickly it disappeared. Moreover, because cosmogenic nuclides persist in mineral grains for millions of years, they allow us to determine paleo-erosion rates in the distant past and to date important events in landscape evolution.

Cosmogenic Nuclide Buildup and Decay

Cosmogenic nuclides are such valuable monitors of surface processes because their production is rapidly attenuated within about a meter of the ground surface (Lal, 1991). Their buildup

in a mineral grain thus indicates how much time that grain has spent within the production zone. Many different cosmogenic nuclides are produced in different mineral grains, depending on the mineral composition. Cosmogenic nuclides used in landscape evolution include stable nuclides such as ^3He and ^{21}Ne , as well as radioactive nuclides such as ^{10}Be , ^{14}C , ^{26}Al , and ^{36}Cl (Cerling and Craig, 1994; Gosse and Phillips, 2001; Muzikar et al., 2003; Nishiizumi et al., 1993). Each of these nuclides can be used in a slightly different way, taking advantage of their production rates and decay times. The great majority of landscape evolution studies use the nuclides ^{10}Be and ^{26}Al in quartz. This is because production rates of these nuclides have a simple depth dependence, and because quartz is a common mineral grain that is resistant to chemical and physical weathering, allowing it to persist over long periods of erosion and transport (Lal and Arnold, 1985).

Aluminum-26 and ^{10}Be are produced by two different types of secondary cosmic ray particles that result from a nuclear cascade through the atmosphere (Lal and Peters, 1967). The first type of particle is a fast nucleon (these are mainly neutrons but include some protons) that causes a nuclear spallation reaction within mineral grains. For example, in quartz (SiO_2), a nuclear collision can cause ^{28}Si to undergo spallation to the lighter nuclide ^{26}Al through the net loss of a proton and a neutron. Likewise, ^{16}O can be transformed to ^{10}Be through the net loss of three protons and three neutrons. Spallation reactions such as these occur very near the ground surface, because fast nucleons are rapidly attenuated as they pass through matter. The penetration length of a fast nucleon depends on the density of the material that it is passing through, and is approximately 160 g cm^{-2} . For example, in typical rock of density 2.6 g cm^{-3} , the nucleons travel 60 cm; in typical soil of density 1.5 g cm^{-3} , the nucleons travel over a meter.

The second type of particle that produces ^{26}Al and ^{10}Be is a muon. Muons are less reactive than nucleons, so they produce fewer cosmogenic nuclides at the ground surface, only 2–3% of total production (Brown et al., 1995a); however, muons penetrate more deeply and dominate production at depths greater than a few meters. Although muons do not contribute much near the ground surface, they can account for up to 20% of the total cosmogenic nuclide inventory in an eroding terrain (Granger et al., 2001). Muons are very important for dating sedimentary deposits and for determining paleo-erosion rates because they continue to produce ^{10}Be and ^{26}Al even when sediment is buried at depth. A proper interpretation of

cosmogenic nuclide concentrations in a sedimentary deposit must account not only for the concentration inherited from the time of deposition, but also for postdepositional production by muons.

Determining Erosion Rates

To a close approximation, the depth dependence of ^{26}Al and ^{10}Be production can be described by the equation

$$P(z) = P_{\text{spallation}}(0)e^{-\rho z/\Lambda} + P_{\text{muon}}(0)e^{-\rho z/L} \quad [1]$$

where P is the production rate at depth z , the subscripts spallation and muon indicate production by the two separate mechanisms, ρ is the density, and Λ ($=160 \text{ g cm}^{-2}$) and L ($\sim 1300 \text{ g cm}^{-2}$) are the penetration lengths of fast nucleons and muons, respectively (Brown et al., 1995a; Lal, 1991). Although the depth dependence of muons varies from this simple approximation at depths greater than about 5 m (e.g., Heisinger et al., 2002), this equation will suffice for the purposes of discussion.

In an eroding landscape, mineral grains are exposed to an increasingly intense secondary cosmic ray flux as they are gradually exhumed from bedrock. In a slowly eroding landscape, mineral grains spend a lot of time near the surface and build up a large cosmogenic nuclide inventory; in contrast, in a rapidly eroding landscape, the mineral grains approach the surface quickly and are eroded away with only a small cosmogenic nuclide inventory. More quantitatively, the cosmogenic nuclide concentration at a rock surface can be found by integrating eqn [1] over time (Lal, 1991). For the case of long-term steady-state erosion, the concentration of a cosmogenic nuclide is given by eqn [2]

$$N = \frac{P_{\text{spallation}}}{(1/\tau) + (\rho E/\Lambda)} + \frac{P_{\text{muon}}}{(1/\tau) + (\rho E/L)} \quad [2]$$

where N is the concentration, E is the erosion rate, and τ is the radioactive mean life (a mean life is equal to a half-life divided by the natural logarithm of 2). The mean lives of ^{26}Al and ^{10}Be are approximately 1.0 and 2.0 million years, respectively.

It is important to realize that the timescale over which cosmogenic nuclides measure erosion rates is approximately the time taken to erode one fast nucleon penetration length (about 60 cm in rock; Lal, 1991). In most soil-mantled landscapes, erosion rates are averaged over a time that is comparable to that of soil formation. Cosmogenically derived erosion rates are usually averaged over a much longer timescale than that of agricultural land use and so primarily reflect a preagriculture rate (Brown et al., 1995b). The exception to this rule comes in rapidly eroding areas with a long history of land use, where several penetration lengths may have eroded away since the beginning of agriculture thousands of years ago. Cosmogenic nuclides not only circumvent the problem of land use in determining erosion rates, but also provide a useful tool for determining natural erosion rates as a baseline against which to ascertain the impacts of land use.

There are two different approaches to measuring erosion rates with cosmogenic nuclides. The first is to measure the erosion rate of a discrete site by analyzing material from the

surface of eroding bedrock or sediment. This can be either an eroding outcrop or a material from bedrock covered by soil (Figure 1). The latter case has been very important for understanding how erosion rates depend on soil cover.

The second approach is to measure erosion rates by analyzing sediment carried by streams. Streams sample eroded material from the entire watershed, providing a natural averaging mechanism (Figure 2). It can be shown that the average cosmogenic nuclide concentration in stream sediment reflects the average residence time of quartz in the watershed, even in a watershed eroding at different rates in different places (Granger et al., 1996). This is because of a self-compensating tendency in sediment mixing. Areas that erode quickly produce a lot of sediment, but with a low cosmogenic nuclide concentration. Areas that erode slowly do not contribute much sediment, but that which they do produce contains a high cosmogenic nuclide concentration. These two effects balance so that the average cosmogenic nuclide concentration indicates the average erosion rate of the watershed, over the time interval required to erode about one penetration length, or 60 cm. There are many complications that can lead to errors. Care must be taken to ensure that the source of quartz in the watershed is known, and that stream sediment is indeed derived from the whole watershed, not just from a local landslide, for example. Nonetheless, repeated studies have shown that basin-averaged

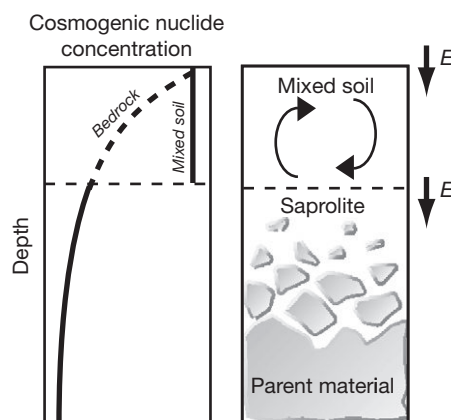


Figure 1 Cosmogenic nuclide concentrations in both bedrock (dashed line) and well-mixed soil (solid line) reach a value at the ground surface determined by the erosion rate (eqn [2]). Erosion rates can be inferred from cosmogenic nuclide concentrations in samples collected at the surface of eroding rock or soil, as well as from saprolite collected beneath a soil cover, as long as soil depth is known.

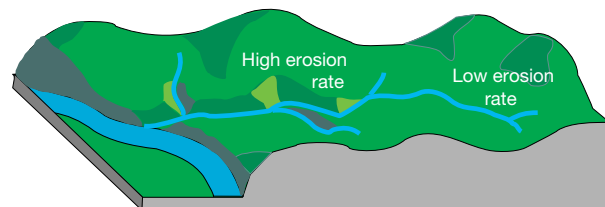


Figure 2 Cosmogenic nuclide concentrations in well-mixed stream sediment are determined by averaging a small amount of sediment from slowly eroding areas with a large amount of sediment from rapidly eroding areas. Neglecting radioactive decay, the average cosmogenic nuclide concentration is approximately given by eqn [2].

erosion rates can be accurately determined by this method (for a review, see [von Blanckenburg, 2005](#)).

Following are examples of each of these methods. The first example shows that bedrock erosion is highly dependent on the depth of soil cover. The second example shows that basin-averaged erosion rates are much more dependent on the long-term tectonic regime than on climate.

Example 1: Erosion Rates Beneath a Soil Cover – The Soil Production Function

It has long been postulated that the rate at which bedrock converted to soil depends on soil depth. Erosional processes such as animal burrowing and tree throw are more vigorous under a thin soil cover, but decline for thick soils. Likewise, freeze–thaw and wet–dry cycles are more effective under a thin soil cover. The functional form by which soil production depends on soil depth, however, has remained a matter of debate. Cosmogenic nuclides provide a way to determine this relationship directly, by measuring erosion rates in a single landscape.

[Heimsath et al. \(1997, 2010 and references therein\)](#) measured ^{10}Be concentrations in weathered bedrock beneath soil to determine the depth dependence of soil production rates. A successful experiment for determining the soil production function requires a landscape where both soil depths and erosion rates are likely to vary considerably. The landscapes they examined had thin soil on hill crests and ridges, and thicker soil in valley bottoms and hollows. All of their measurements were made on divergent hillslopes where it is most likely that soil is neither accumulating nor evacuating over time. The rate of soil production recorded by cosmogenic nuclides in this case is not the rate of chemical weathering, but is instead the rate of mechanical disruption of preweathered bedrock into mobile regolith. For a steady-state soil depth, this is equivalent to the denudation rate.

[Heimsath et al.](#) found that soil production rates decrease exponentially with soil depth, reaching a maximum under very shallow soils. This implies that soil production and erosion is stabilized by a negative feedback, so that soil depths tend to vary smoothly across a landscape and regulate erosion rates. If soils become thin in a particular place, then bedrock is decomposed more quickly to produce a thicker soil. Likewise, thick soils inhibit erosion so that soils eventually thin.

The fact that [Heimsath et al.](#) found thin soils and fast erosion rates on hill crests and ridges indicates that the landscapes where they worked are not in steady state, and that their relief is decreasing through time as the hill crests lower faster than the surrounding hillslopes, at least locally. This is not necessarily the case in all landscapes, however. In areas with more uniform soil depths or with large areas of exposed bedrock, the relationship between erosion rates and soil depth is less apparent ([Wilkinson et al., 2005](#)). Also, it should be realized that the rate of bedrock weathering to produce saprolite may have a different soil depth dependence than the mechanical disruption of saprolite into soil. Indeed, exposed bedrock erodes far more slowly than soil-mantled saprolite in many terrains ([Bierman, 1994; Granger et al., 2001; Heimsath et al., 2000](#)). A transition from soil mantled to bedrock topography is

one way that climate can affect erosion rates in arid environments ([Heimsath et al., 2010](#)).

Example 2: The Influence of Climate and Tectonics on Weathering and Erosion

Landscape-scale weathering and erosion rates depend on a wide variety of factors, including climatic parameters such as rainfall and temperature which have fluctuated dramatically through the Quaternary, as well as tectonic uplift rates, which vary at considerably slower rates. The issue of whether climate is more important than tectonics in determining regional erosion rates has been difficult to resolve using traditional methods. Many models have suggested that erosion rates may vary by orders of magnitude, for example, as rainfall increases from semiarid to humid environments. A suite of cosmogenic nuclide studies, however, has shown that although climate largely controls the degree of chemical alteration of rock as it is eroded, the net export of material from a watershed is closely matched to the long-term exhumation rate dictated by tectonics (see reviews by [Granger and Riebe, 2007; von Blanckenburg, 2005](#)).

The influence of climate

In a wide-ranging set of experiments, [Riebe et al.](#) analyzed ^{10}Be in sediment from nonglaciated granitic watersheds throughout a range of temperature and precipitation regimes to determine erosion rates (summarized by [Granger and Riebe, 2007; Riebe et al., 2003, 2004a](#)). They also inferred the degree of chemical weathering from the depletion of various elements relative to immobile zirconium.

[Riebe et al. \(2004a\)](#) found that although erosion rates were highly variable within a given climate regime, there is remarkably little variation in erosion rates between climate regimes. In other words, erosion rates show little dependence on climate. For example, in granitic sites throughout North America, extending over a mean annual temperature range from 2 to 25 °C and an annual precipitation gradient from 22 to 420 cm year⁻¹, they found no apparent pattern in erosion rates.

Although erosion rates do not depend on climate, the degree of chemical weathering does. For example, measurements at various altitudes in the Santa Rosa Mountains, Nevada, show that as vegetation decreases and snow cover increases, the degree of chemical depletion decreases from 20% to 0%, while the total erosion rate determined from ^{10}Be remains unchanged ([Riebe et al., 2004b](#)).

One implication of these studies is that the export of solutes from a watershed is generally limited by the supply of fresh minerals through physical erosion. A quickly eroding watershed in an arid environment supplies more solutes to the ocean than a slowly eroding watershed in a tropical environment ([Riebe et al., 2004a](#)).

The influence of tectonics

If erosion rates show such little dependence on climate, then what controls long-term erosion rates? A host of studies from throughout the world indicate that the long-term uplift and exhumation rate due to tectonics is the primary

control of short-term erosion rates (Riebe et al., 2001; von Blanckenburg, 2005).

One example where erosion rates have been quantitatively compared to known uplift rates is in Italy (Cyr and Granger, 2008; Cyr et al., 2010). Uplift rates have been quantified from well-dated marine terraces that ring the Italian peninsula and islands, and range from 0.2 mm year^{-1} in the north to 1.6 mm year^{-1} near the Straits of Messina in the south. Cyr et al. (2010) collected quartz sediment from streams draining landscapes with a variety of uplift rates. Although there was considerable scatter in the data reflecting stochastic erosional processes, the average erosion rate and the average uplift rate agree remarkably well in any given area. Deviations from this pattern occur where stream profiles showed strong convexities, indicating a landscape still undergoing a transient response to uplift.

Measuring River Incision and Valley Lowering Rates

River incision is intimately linked to basin-scale erosion. Rivers are the primary means for transporting sediment and solute from an eroding terrain. Rivers set the base level to which a watershed can erode. Because rivers can incise more quickly than hillslopes erode, they transmit base-level lowering signals to the basin's headwaters, thus regulating hillslope erosion in response to tectonic uplift and subsidence. Rivers also aggrade when they are choked by excess sediment from rapid hillslope erosion; thus alluvial sediments often provide a record of fluctuations in erosion rates. Cosmogenic nuclides can be used to study the history of river incision and aggradation by dating alluvial sediments, dating terraces, and dating caves alongside the river.

Over timescales of tens to hundreds of thousands of years, cosmogenic nuclide exposure dating can be used to date bedrock surfaces or sediments on river terraces. Over longer timescales, however, surfaces are seldom preserved well enough for exposure dating. Yet it is important to understand how river incision varies through time over millions of years.

Cosmogenic Nuclide Burial Dating

Cosmogenic nuclide burial dating measures the decay of cosmogenic nuclides after sediment has been buried. If sediment containing cosmogenic nuclides is buried so that it is shielded from secondary cosmic radiation, then the inherited cosmogenic nuclides decay over time according to eqn [3]

$$N(t) = N_{\text{inh}}(0)e^{-t_{\text{burial}}/\tau} \quad [3]$$

where t_{burial} is the time since burial and N_{inh} is the inherited cosmogenic nuclide concentration at the time of burial (Lal and Arnold, 1985). For a landscape eroding at steady state, N_{inh} can be determined using eqn [2]. Burial dating works exceptionally well with the cosmogenic nuclide pair ^{26}Al – ^{10}Be in quartz, because these nuclides are produced at an effectively constant ratio, and they decay at different rates subsequent to burial. By solving eqn [3] simultaneously for both ^{26}Al and ^{10}Be in the same sample of quartz, both the burial time and the paleo-erosion rate can be determined. The burial dating method is typically

limited to the past 5 million years, because ^{26}Al is usually undetectable beyond that time (Granger and Muzikar, 2001).

Burial dating requires that the sediment is shielded from secondary cosmic rays, including muons. The degree of shielding necessary to use this method depends on the age of the sediment being dated. For sediments older than a million years, burial depths of 10–20 m of sediment or rock are usually needed. Two situations that are appropriate for burial dating are thick alluvial terraces, where sediment packages are deposited in excess of 10 m, and caves, where sediment is deeply shielded underground.

Using Caves to Learn About River Incision Rates

Caves provide a nearly perfect scenario for burial dating because sediment washed into the cave entrance is instantly shielded beneath tens of meters of solid rock. Because caves are formed by flowing water, they can also provide information about the water table at the time of sediment deposition. Quartz sediment can be washed into caves either by a surface river that passes near a preexisting cave opening (Figure 3) or by a sinking river that carries sediment underground as the cave is being formed. In either case, the burial date of quartz in the cave indicates the last time that water flowed through the



Figure 3 River sediment deposited in caves such as this one above the New River, Virginia, can be dated with cosmogenic nuclides to infer long-term river incision rates.

passage. As a river cuts through bedrock, the cave is abandoned by flowing water. The fossil cave passage, together with its sediment, remains an indicator of when the river was at that particular elevation. In this way, caves are very similar to terraces; they both indicate past river levels and they both contain alluvial sediment. Caves sometimes offer an advantage over terraces, however, because the sediment within is shielded from weathering and erosion. Cave sediments can thus persist much longer than terrace sediments on the surface. Furthermore, burial in caves is instantaneous, whereas burial under younger terrace sediment may, in some cases, have the disadvantage of only partial burial for an initial period.

Example: The Siebenhengste–Hohgant Cave System and Glacial Valley Lowering in Switzerland

The Siebenhengste–Hohgant cave system in Switzerland has a relief of >1.3 km, stretching from the top of the mountains in the Alpine front to the bottom of the Aare glacial valley. The cave has developed in at least 14 distinct levels, each of which has been identified from the vadose–phreatic transition evident in the cave’s morphology. These cave levels formed primarily during interglacials, when water was undersaturated with respect to calcite, and they drained to the level of the adjacent valley floor. The cave levels therefore record the long-term incision of the Aare glacier. Most of the cave levels contain sediment that was washed into the system from the surface. By dating this sediment, we can learn about the rates of glacial valley incision.

Haeuselmann et al. (2007) dated 20 samples of quartz-bearing sediment from seven levels within the cave. Sediment ages range up to 4 million years, predating major glaciation in the Alps. The ages indicate gradual lowering of the valley floor, at a rate of about 120 m My^{-1} , until the middle Pleistocene. Beginning about 800 000 years ago, valley incision accelerated dramatically, lowering the valley floor by a kilometer. Incision rates increased by an order of magnitude. The reason for such rapid glacial valley incision remains unclear; perhaps rapid incision is a regional pattern due to climate change at the mid-Pleistocene transition (a global shift of glacial periodicity from 41 000 to 100 000 years), or alternatively, the signal could be limited to glacial dynamics in the Aare valley alone. In either case, the data show that the spectacular topography of the Aare glacial trough developed almost entirely over the past million years.

Additional Applications

There are many applications of cosmogenic nuclides in landscape evolution that have not been discussed here. For example, it is possible to use cosmogenic nuclides in buried sediments to infer not only the burial age, but also the paleo-erosion rate. For example, data from the Siebenhengste–Hohgant cave system, mentioned above, show that paleo-erosion rates in the Alpine front were about 100 m My^{-1} , in equilibrium with valley lowering rates in the Pliocene and early Pleistocene (Haeuselmann et al., 2007). Paleo-erosion rates in the Italian Apennines over the past million years also show steady erosion, in equilibrium with regional uplift rates (Cyr and Granger, 2008).

Another important application of cosmogenic nuclides is the discrimination of eroding versus noneroding areas beneath ice sheets and glaciers. It has long been suspected that cold-based glaciers, frozen to their beds, do not erode bedrock, while warm-based glaciers that slide over their beds are very effective at scouring bedrock. The degree of glacial erosion, though, has previously been difficult to measure. Cosmogenic nuclides provide a method for quantifying glacial erosion. This is because areas of bedrock that are eroded by less than a few meters retain cosmogenic nuclides generated prior to glaciation. Subsequent to deglaciation, these areas have an excess cosmogenic nuclide concentration relative to areas that are deeply scoured. This cosmogenic nuclide method has been used to demonstrate that significant areas of nonerosive ice existed in the Fennoscandian Ice Sheet (Fabel et al., 2002) as well as in northern Canada (e.g., Staiger et al., 2005).

Conclusion

The study of landscape evolution has historically been divided into two related but distinct approaches: denudation chronology and process geomorphology. Denudation chronology is concerned with establishing a sequence of events responsible for creating today’s landscape. These events may be associated with, for example, tectonic uplift, climate change, or drainage capture. However, process geomorphology is less concerned with the sequence of events in the past and instead focuses on the physical processes of how landscapes are changing today. Cosmogenic nuclides are contributing to both of these approaches. Measurements of erosion rates, soil production functions, and the influence of climate on erosion all increase our understanding of present-day processes that are shaping the landscape. However, river incision rates and paleo-erosion rates contribute more to understanding landscape history and denudation chronology. A fuller understanding of landscape evolution and the dynamic response of hillslopes to climate, tectonics, and land use change require both these historical and mechanistic perspectives.

See also: **Introduction:** History of Dating Methods; History of Quaternary Science; Understanding Quaternary Climatic Change. **Paleoclimate:** Introduction.

References

- Bierman PR (1994) Using in situ produced cosmogenic isotopes to estimate rates of landscape evolution: A review from the geomorphic perspective. *Journal of Geophysical Research* 99: 13885–13896.
- Brown ET, Bourles DL, Colin F, Raisbeck GM, Yiou F, and Desgarceaux S (1995) Evidence for muon-induced in situ production of ^{10}Be in near-surface rocks from the Congo. *Geophysical Research Letters* 22: 703–706.
- Brown ET, Stallard RF, Larsen MC, Raisbeck GM, and Yiou F (1995) Denudation rates determined from the accumulation of in situ-produced ^{10}Be in the Luquillo Experimental Forest, Puerto Rico. *Earth and Planetary Science Letters* 129: 193–202.
- Cerling TE and Craig H (1994) Geomorphology and in-situ cosmogenic isotopes. *Annual Review of Earth and Planetary Science* 22: 273–317.
- Cyr AJ and Granger DE (2008) Dynamic equilibrium among erosion, river incision, and coastal uplift in the northern Apennines, Italy. *Geology* 36: 103–106.

- Cyr AJ, Granger DE, Olivetti V, and Molin P (2010) Quantifying rock uplift rates using channel steepness and cosmogenic nuclide-determined erosion rates: Examples from northern and southern Italy. *Lithosphere* 2: 188–198.
- Fabel D, Stroeven AP, Harbor J, Kleman J, Elmore D, and Fink D (2002) Landscape preservation under Fennoscandian ice sheets determined from in situ produced ^{10}Be and ^{26}Al . *Earth and Planetary Science Letters* 201: 397–406.
- Gosse JC and Phillips FM (2001) Terrestrial in situ cosmogenic nuclides: Theory and application. *Quaternary Science Reviews* 20: 1475–1560.
- Granger DE, Kirchner JW, and Finkel RC (1996) Spatially averaged long-term erosion rates measured from in situ-produced cosmogenic nuclides in alluvial sediment. *Journal of Geology* 104: 249–257.
- Granger DE and Muzikar PF (2001) Dating sediment burial with cosmogenic nuclides: Theory, techniques, and limitations. *Earth and Planetary Science Letters* 188: 269–281.
- Granger DE and Riebe CS (2007) Cosmogenic nuclides in weathering and erosion. In: Drever JL (ed.) *Surface and Groundwater, Weathering and Soils*. In: Turekian KK and Holland HD (eds.) *Treatise on Geochemistry (online update)*, vol. 5, p. 43. Oxford: Elsevier-Pergamon.
- Granger DE, Riebe CS, Kirchner JW, and Finkel RC (2001) Modulation of erosion on steep granitic slopes by boulder armoring, as revealed by cosmogenic ^{26}Al and ^{10}Be . *Earth and Planetary Science Letters* 186: 269–281.
- Haeuselmann P, Granger DE, Jeannin P-Y, and Lauritzen S-E (2007) Abrupt glacial valley incision at 0.8 Ma dated from cave deposits in Switzerland. *Geology* 35: 143–146.
- Heimsath AM, Chappell J, Dietrich WE, Nishiizumi K, and Finkel RC (2000) Soil production on a retreating escarpment in southeastern Australia. *Geology* 28: 787–790.
- Heimsath AM, Chappell J, and Fifield K (2010) Eroding Australia: Rates and processes from Bega Valley to Arnhem Land. *Geological Society of London, Special Publications* 346: 225–241.
- Heimsath AM, Dietrich WE, Nishiizumi K, and Finkel RC (1997) The soil production function and landscape equilibrium. *Nature* 388: 358–361.
- Heisinger B, Lal D, Jull AJT, et al. (2002) Production of selected cosmogenic radionuclides by muons: 2. Capture of negative muons. *Earth and Planetary Science Letters* 200: 357–369.
- Lal D (1991) Cosmic ray labeling of erosion surfaces: In situ nuclide production rates and erosion models. *Earth and Planetary Science Letters* 104: 424–439.
- Lal D and Arnold JR (1985) Tracing quartz through the environment. *Proceedings of the Indian Academy of Sciences: Earth and Planetary Sciences* 94: 1–5.
- Lal D and Peters B (1967) Cosmic ray produced radioactivity on the earth. In: Lal D and Peters B (eds.) *Handbuch der Physik*, pp. 551–612. Berlin: Springer-Verlag.
- Muzikar P, Elmore D, and Granger D (2003) Accelerator mass spectrometry in geologic research. *Geological Society of America Bulletin* 15: 643–654.
- Nishiizumi K, Kohl CP, Arnold JR, et al. (1993) Role of in situ cosmogenic nuclides ^{10}Be and ^{26}Al in the study of diverse geomorphic processes. *Earth Surface Processes and Landforms* 18: 407–425.
- Riebe CS, Kirchner JW, and Finkel RC (2003) Long-term rates of chemical weathering and physical erosion from cosmogenic nuclides and geochemical mass balance. *Geochimica et Cosmochimica Acta* 67: 4411–4427.
- Riebe CS, Kirchner JW, and Finkel RC (2004a) Erosional and climatic effects on long-term chemical weathering rates in granitic landscapes spanning diverse climate regimes. *Earth and Planetary Science Letters* 224: 547–562.
- Riebe CS, Kirchner JW, and Finkel RC (2004b) Sharp decrease in long-term chemical weathering rates along an altitudinal transect. *Earth and Planetary Science Letters* 218: 421–434.
- Riebe CS, Kirchner JW, Granger DE, and Finkel RC (2001) Strong tectonic and weak climatic control of long-term chemical weathering rates. *Geology* 29: 511–514.
- Staiger JKW, Gosse JC, Johnson JV, et al. (2005) Quaternary relief generation by polythermal glacier ice. *Earth Surface Processes and Landforms* 30: 1145–1159.
- von Blanckenburg F (2005) The control mechanisms of erosion and weathering at basin scale from cosmogenic nuclides in river sediment. *Earth and Planetary Science Letters* 237: 462–479.
- Wilkinson MT, Chappell J, Humphreys GS, Fifield LK, Smith B, and Hesse P (2005) Soil production in heath and forest, Blue Mountains, Australia: Influence of lithology and paleoclimate. *Earth Surface Processes and Landforms* 30: 923–934.

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Dating Techniques

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Introduction

A number of important dating methods are used for studies in the Quaternary. Many of these subjects are discussed in detail in sections of this encyclopedia. The purpose of all of these methods is to estimate the age of the material of interest. Sometimes, the need is for the age determination to be very precise; in other cases, relative age or approximate age may be sufficient. Hence, the purpose of this article is to summarize those methods and discuss briefly their advantages and deficiencies. [Figure 1](#) summarizes the age ranges covered by different methods.

These methods can be subdivided into three main categories:

1. Radioactive nuclide methods
2. Radiative dosimetry methods and
3. Qualitative and comparative methods

Each of these sections is subdivided into the specific methods, and cross references to the more detailed articles that deal with these methods are also given.

Radioactive Nuclides

The most important collection of methods for Quaternary dating involve radioactive decay, or in some cases, the buildup of radioactive species. As radioactive nuclides will increase or decrease with time, we can use the concentration of the specific nuclide to estimate the age of the material ([Table 1](#)).

Radioactive species can be produced by two different kinds of nuclear processes: radioactive decay of a parent nuclide or

production by nuclear reactions as a result of exposure to cosmic or other radiation.

Radioactive Decay

For example, in the simple case of the study of a nuclide present in its maximum concentration at the creation of the material, such as radiocarbon dating, the number of ^{14}C atoms will decay with time, according to the radioactive decay equation, originally shown by Ernest Rutherford ([Dickin, 1995](#)):

$$dn/dt = -\lambda N \quad [1]$$

where the decay rate is proportional to the number of radioactive atoms present (N). The 'decay constant' λ defines the rate of decay. It is usual to refer to the 'half-life' of the radionuclide, which is the time for one-half of the remaining atoms to decay, which is related directly to the decay constant:

$$t_{1/2} = \frac{\ln 2}{\lambda} \quad [2]$$

If eqn [1] is integrated, within the limits $N=N_0$ at $t=0$ and $N=0$ at $t=\infty$, the result is a more common form:

$$\frac{N}{N_0} = e^{-\lambda t} \quad [3]$$

where N_0 is the number of atoms at time $t=0$ and N is the number of atoms at any time t .

In-Growth of a Radionuclide

In-growth of a radioactive species occurs when it is produced either from the decay of another nuclide or directly in the

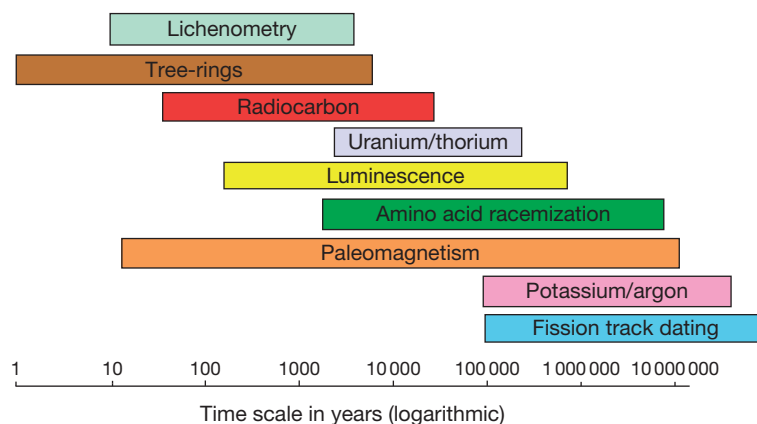


Figure 1 Time scales of different Quaternary geochronology methods. Courtesy Scott Elias (Royal Holloway College, University of London).

Table 1 Radioactive Quaternary dating methods

Isotopic system	Parent nuclide or process	Half-life	Time scale	Material	References
^{14}C	$^{14}\text{C}(\text{n,p})\ ^{14}\text{N}$	5730 year	0–50 000 year	Organic material, carbonates	Taylor (1987)
^{230}Th	^{238}U and ^{234}U	75 200 year	1000–500 000 year	Carbonates	Schwarcz (1989)
^{210}Pb	U-series decay sequence	22.3 year	0–150 year	Fine-grained sediments	Olsson (1986)
U–He	U isotopes α -decay	Depends on parent nuclides	0–300 000 year	Tephra	Kohn et al. (2000)
^{137}Cs	U fission	30	After AD 1950	Fine-grained sediments	Pennington et al. (1973)
^{40}K – ^{40}Ar	^{40}K β -decay	1.25×10^9 year	10^4 – 10^9 year	K-bearing minerals, bones, tephra	McDougall and Harrison (1988)

sample. The latter example occurs often in the case of *in situ* cosmogenic nuclide studies.

An example of a process where the in-growth of a radionuclide is the ^{231}Th – ^{235}U system. In this case, ^{231}Pa increases with time due to the decay of the parent nuclides ^{235}U , so that (Ku, 1968):

$$N_{231} = N_{235}(1 - e^{-\lambda t}) \quad [4]$$

A more complex case occurs for ^{230}Th . In this case, ^{230}Th increases with time due to the decay of the parent nuclides ^{234}U , which is produced from ^{238}U decay (see U–Th section). Because the half-life of ^{234}U is very low compared to either the parent ^{238}U or daughter ^{230}Th , in principle, it should drop out of the equation. This equation would be correct if the intermediate daughter product of ^{238}U , ^{234}U , was in secular equilibrium with the ^{238}U (Ku, 2000). Hence, the amount of ^{230}Th produced is a product of the decay of both the excess ^{234}U and the ^{238}U . However, it is not and the result is a more complex equation, described in detail in the section on U–Th dating.

Radiocarbon Dating

As discussed in the sections on radiocarbon dating, this method is widely used for reconstructing the age of various kinds of carbon-containing materials. This involves a considerable amount of pretreatment chemistry (Figure 2). Here, organic or inorganic materials in equilibrium with the



Figure 2 Sample preparation chemistry at the University of Arizona AMS Laboratory.

production of ^{14}C in the atmosphere and its removal into the oceans, establish a consistent level of ^{14}C . When the animal or plant dies, it is removed from this quasi-equilibrium and so the level of ^{14}C decays according to eqn [1]. There are a vast number of applications of ^{14}C , which can be measured either by counting the decays of the radioactive ^{14}C or by direct atom counting by accelerator mass spectrometry (AMS). Most measurements these days are measured by AMS (see Figure 3). More details of ^{14}C dating are given in the sections by Cook



Figure 3 The 3 MV accelerator mass spectrometer at the University of Arizona.

and van der Plicht, who discuss decay counting of ^{14}C and Jull, who discusses AMS. Other chapters by van de Plicht and Burr discuss the causes of ^{14}C variations over time.

U-Series Methods

Studies of uranium-series nuclides (Wagner, 1995) rely on the extensive decay sequence in the uranium decay series. We can take some of these pairs such as ^{234}U – ^{238}U , ^{230}Th – ^{238}U , ^{231}Pa – ^{235}U , and ^{226}Ra – ^{230}Th as chronometers for measurements of time on various time scales, depending on the radioactive pair chosen (Ku, 2000). Problems can arise, such as where there is some initial ^{230}Th . Hence, calculations of the age must take into account if there is disequilibrium in the ^{238}U – ^{234}U system (Cohen, 2005). U-series nuclides have a wide variety of applications to dating of sediments, soil horizons, peat, bones, corals, and carbonates of all kinds.

Other Dating Methods Based on U-Series Decay

Although sometimes listed as a separate method, in U–He dating, one observes the in-growth of the He daughter product of U. This method is sometimes used for dating bones, mollusks, and corals (Wagner, 1995). Similarly, the U-daughter nuclide ^{210}Pb can be used for dating of sediments in the last ~200 year (Cohen, 2005; Wagner, 1995).

K–Ar Dating

The K–Ar method and its companion, the $^{40}\text{Ar}/^{39}\text{Ar}$ method, are based on the radioactive decay of ^{40}K to ^{40}Ar (Renne, this volume). ^{40}K undergoes a branching decay to both ^{40}Ar and ^{40}Ca , so it is important to revise the simple radioactive decay equation to take into account the decay via different processes. The ^{40}Ar – ^{39}Ar process is based on the same idea as that of the K–Ar method, except that neutron activation is used to estimate the amount of ^{39}K by activation to ^{39}Ar . The in-growth of ^{40}Ar is a variant of the in-growth eqn [4], since there is actually a branching decay to ^{40}Ca and ^{40}Ar from ^{40}K :

$$^{40}\text{Ar} = ^{40}\text{Ar}_{\text{initial}} + 0.105 ^{40}\text{K} (1 - e^{-\lambda t}) \quad [5]$$

where t is the age and 0.105 is the branching ratio to ^{40}Ar . K–Ar dating has many applications, particularly during the Quaternary to the dating of bones and tephra.

Cosmogenic Radionuclides

Similar equations to those for in-growth of a nuclide also apply to the case of the buildup of a cosmogenic radionuclide such as ^{10}Be :

$$N = \frac{P}{\lambda} (1 - e^{-\lambda t}) \quad [6]$$

Here, P is the production rate in atoms per year and λ is the decay constant (in years $^{-1}$). More complex scenarios can arise where the sample is buried and re-exposed. The application of cosmogenic nuclide for dating of Quaternary surfaces and deposits has increased dramatically in the past decade (Dickin, 1995; Fifield and Fink, 2010; Zreda and Phillips, 2000), as presented elsewhere in this volume.

In the case where the cosmogenically produced nuclide is stable, the number of atoms produced is simply $N = Pt$. There have been a number of studies of stable noble gas nuclides in various rock surfaces.

Postnuclear Tracers

The radionuclides produced as the result of anthropogenic nuclear activities can be used as tracers for recent processes. Nuclides such as ^{137}Cs , ^{90}Sr , and postbomb levels of ^{14}C can be used to date event horizons since AD 1950. These markers are often of great use in the study of recently deposited sediments and other materials. Recent sediment studies are often done in conjunction with in-growth of ^{210}Pb from U-series decay (Cohen, 2005).

Radiative Dosimetry Methods

Radiative dosimetry methods are basically studies by which we investigate the buildup of radiation damage. This is usually measured by studying some luminescence phenomenon. Common methods are thermoluminescence, optically stimulated luminescence, and electron spin resonance (ESR) techniques. The techniques rely on the production of radiation damage in the substrate (usually, a silicate mineral). When these solids are exposed to ionizing radiation, electron/hole pairs are formed and some of these become trapped in the solid.

The luminescence age appears simple enough and is constrained by the equation:

$$\text{Age} = \frac{\text{ED}}{\sum D}$$

where ED is the 'equivalent dose' (in units of grays, 1 Gy = 100 rad) and $\sum D$ is the sum of annual doses from α , β , γ and cosmic radiation (Forman, 2000). ED is determined in several different ways, including estimates of the chemical composition of radioactive sources in the sample, regeneration of an additional dose of the luminescence signal in the laboratory, and partial or total bleaching (i.e., removal of the luminescence signal by exposure to light).

Table 2 Nonradiometric Quaternary dating methods

<i>Technique</i>	<i>Process</i>	<i>Time scale</i>	<i>Material</i>	<i>References</i>
Fission-track dating	α -particle tracks in minerals	10^5 – 10^9 year	Tephtras, glass	Westgate and Naeser (1995)
Amino acid racemization	Rate of equilibration of D and L isomers of amino acids	10 – 10^6 year	Mollusks, ostracods, egg shells, some fine-grained sediments	Wehmiller and Miller (2000)
Thermoluminescence	Luminescence resulting from heating to rest electronic traps caused by radiation damage	100 – 10^6 year	Fine-grained quartz or feldspar in sediments	Aiken (1998)
Optically stimulated luminescence	Luminescence resulting from optical stimulation to rest electronic traps caused by radiation damage	100 – 10^6 year	Fine-grained quartz or feldspar in sediments	Stokes (1999)
Electron spin resonance	Intensity of ESR signal proportional to the number of radiation-damage electron traps	100 year–50 ky	Fine-grained quartz	Lee and Schwarcz (2000)
Varve chronologies	Annual varve deposits in sediment	Any age, up to 45 ka correlated with radiocarbon chronology	Fine-grained sediments	O'Sullivan (1983), Wohlfarth et al. (1993), and Hughen et al. (1996)
Paleomagnetism	Changes in magnetic intensity and direction	10^2 – 10^8 year	Fine-grained sediments and volcanics	Thompson and Oldfield (1986) and Verosub (1988)
Tephrochronology	Volcanic ash layers	Varies	Tephtras	Westgate and Naeser (1995)
Biostratigraphy	Correlation of similar macrofossils	Any age		

The main problem with luminescence methods is the number of possible complications which can arise due to the fact that the original sediment was not totally bleached by sunlight before the accumulation of the dose under study, or that later partial bleaching occurs, or there is an unexplained loss of the signal, called 'anomalous fading' (Wintle, 1973).

There are two main methods which use radiative dose: fission-track dating and luminescence methods, which include thermoluminescence and optically stimulated luminescence (Aiken, 1998). Another radiative dose method which can sometimes be used is the study of the buildup of electronic 'traps' by ESR dating (Lee and Schwarcz, 1993). All of these methods have limitations due to the necessity of accurately estimating the levels of the source of the radiation, usually α particles (from the decay of U) and sometimes γ radiation (Wagner, 1995). These topics are all discussed in various articles of this encyclopedia and the references listed in Table 2.

Amino Acid Racemization

The technique of amino acid racemization relies on the kinetics of changes in the symmetry of amino acid molecules (Miller and Brigham-Grette, 1989). As amino acids can have two different optically active isomers, there is one biologically preferred form (usually the L-form). After initial deposition, these amino acids react over time to approach an equal mixture of D and L isomers, which is called a racemic mixture (Figure 4). The decay rates of many of these reactions are known, although they are sensitive to changes in temperature, as are nearly all chemical reactions. Hence, in order to calculate an age from an amino acid composition, it is important

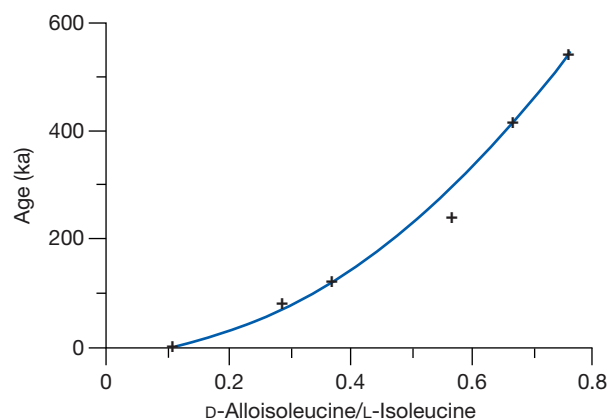


Figure 4 Exchange rates of amino acids at different temperatures. Reproduced from Hearty PS and Kaufman DS (2000) Whole-rock aminostratigraphy and Quaternary sea-level history of the Bahamas. *Quaternary Research* 54: 163–173.

to understand the reaction rates and the dependence of the reaction rates on temperature and other parameters (Wehmiller and Miller, 2000). The amino acid composition is usually expressed as the D/L ratio of the amino acid of interest. If the goal is absolute dating, it is usually necessary to compare the rates of change with some independent dating method such as ^{14}C . The method can also be used in conjunction with independent age estimates to measure temperature (Miller et al., 1997). Otherwise, serious errors can result, such as the very early age assignments of early humans in the New World, which were later shown to be in error by radiocarbon dating (Stafford et al., 1984).

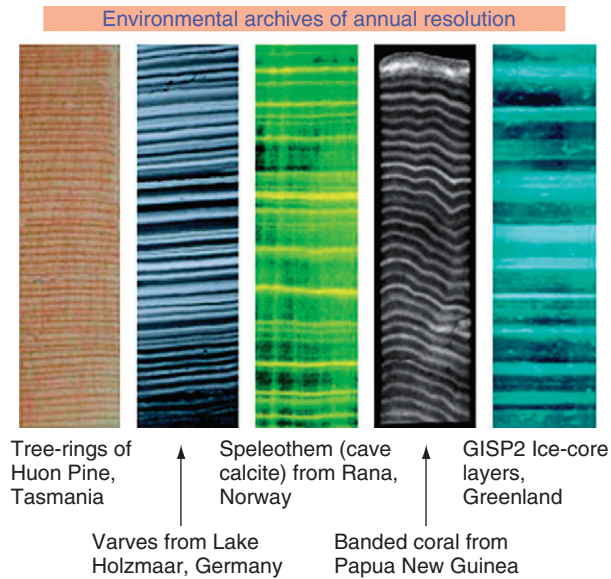


Figure 5 Examples of different Quaternary archives of annual resolution. Courtesy Royal Holloway College, University of London. <http://www.gg.rhul.ac.uk/cqr/MSc/courses/GG5217.html>

Qualitative and Comparative Methods

There are several methods used for dating of geological materials and the methods rely on what are best described as relative or incremental techniques (Figure 5). These methods rely on various observations to estimate the passage of time:

- Deposition of regular additions of sediment on a cyclical fashion, called rhythmites. These are often called 'varves' if annually deposited. Lake varve chronologies were originally proposed as a dating technique by the Swedish geologist De Geer. These chronologies have also been proposed as ways to cross-correlate to other chronologies. In general, there is a degree of scepticism about some of these chronologies since it is difficult to establish a priori that the deposition is annual.
- Long varve chronologies have been established for marine sediments (e.g., Hughen, this volume) and these chronologies have been incorporated into some radiocarbon calibration schemes. These varved sediments in principle have the same problems as lake sediments, but in the case of the Cariaco Basin chronology (Hughen et al., 2004), there is good evidence that the annual deposition assumption can be taken as valid over certain parts of the time scale.
- Lichenometry. The incremental growth of certain lichen species has been used to estimate the age of a rock surface. The technique was pioneered by Bull (2000) who used measurements of the diameter of the largest lichens on surfaces as a way of estimating the age of the surface. This approach appeared to give approximate ages for some surfaces which could be confirmed by other methods.
- Dendrochronology. Trees produce annual growth rings and hence, the variation of the thickness of tree rings can be cross-correlated with established chronologies (Figure 6).



Figure 6 A Bristlecone pine from the White Mountains of California. Courtesy of the University of Arizona Tree Ring Laboratory.

In cases where a piece of wood can be dendrochronologically dated, results can be precise to a few years.

- Paleomagnetic dating. The remanent magnetization in magnetically susceptible materials in sediments and ceramics can sometimes be correlated with known geomagnetic fluctuations (Butler, 1992). Both the intensity of the geomagnetic field and its direction change with time (Cohen, 2005). The sequence of geomagnetic signals in a sediment sequence can then be used to estimate the age of the sequence.
- Correlation of isotopic signals is also an important method for Quaternary dating. The long record of $\delta^{18}\text{O}$ in ice cores and marine sediments is often used to cross-correlate between different marine and ice core records, and therefore establish definite time markers. An excellent example is the ability to cross-correlate dramatic changes in climate from glacial to interglacial, or a return to glacial-like conditions, such as the Younger Dryas event at about 13 000 years ago (Markgraf, 2001), through the oxygen-isotope record (Mayewski et al., 1997).
- Correlation of changes of flora and pollen can sometimes also be correlated between different sites in the same region or over wider areas (Bradbury, 2001).

See also: K/Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ Dating; U-Series Dating. **Cosmogenic Nuclide Dating:** Exposure Geochronology. **Glacial Landforms:** Glaciofluvial Landforms of Erosion. **Introduction:** Understanding Quaternary Climatic Change. **Luminescence Dating:** Electron Spin Resonance Dating; Thermoluminescence. **Paleoclimate:** Introduction. **Radiocarbon Dating:** AMS Radiocarbon Dating; Causes of Temporal ^{14}C Variations; Conventional Method; Variations in Atmospheric ^{14}C .

References

- Aiken MJ (1998) *Thermoluminescence Dating*. New York: Academic Press.
 Bradbury JP (2001) Full and late glacial lake records along the PEP1 transect: Their role in developing interhemispheric paleoclimate interactions.

- In: Markgraf V (ed.) *Interhemispheric Climate Linkages*, pp. 265–291. San Diego: Academic Press.
- Bull WB (2000) Lichenometry: A new way of dating and locating prehistorical earthquakes. In: Noller JS, et al. (ed.) *Quaternary Geochronology: Methods and Applications*, pp. 157–176. Washington, DC: American Geophysical Union.
- Butler R (1992) *Paleomagnetism: Magnetic Domains to Geologic Terrains*. Boston: Blackwell.
- Cohen AS (2005) *Paleolimnology: The History and Evolution of Lake Systems*. Oxford: Oxford University Press.
- Dickin AP (1995) *Radiogenic Isotope Geology*. Cambridge: Cambridge University Press.
- Fifield LK and Fink D (2010) Environmental applications of accelerator mass spectrometry. In: Beauchemin D and Matthews DE (eds.) *Elemental and Isotope-ratio Mass Spectrometry. Encyclopedia of Mass Spectrometry*, vol. 5, pp. 629–640. Elsevier: Amsterdam.
- Forman S (2000) Thermoluminescence dating. In: Noller JS, et al. (ed.) *Quaternary Geochronology: Methods and Applications*, pp. 157–176. Washington, DC: American Geophysical Union.
- Hearty PS and Kaufman DS (2000) Whole-rock aminostratigraphy and Quaternary sea-level history of the Bahamas. *Quaternary Research* 54: 163–173.
- Hughen KA, Baillie MGL, Bard E, et al. (2004) Marine04 marine radiocarbon age calibration, 0–26kyr. *Radiocarbon* 46: 1059–1086.
- Hughen KA, Overpeck JT, Peterson LC, and Anderson RF (1996) The nature of varved sedimentation in the Cariaco basin, Venezuela, and its palaeoclimatic significance. In: Kemp AES (ed.) *Palaeoclimatology and Palaeoceanography from Laminated Sediments, Geological Society Special Publication* 116: 171–183.
- Kohn BP, Farley KA, Pillans B (2000) (U-th)/He and fission track dating of the Pleistocene Rangitawa tephra, North Island, New Zealand: A comparative study. In *9th International Conference on Fission Track Dating and Thermochronology*, Lorne, Australia, pp. 207–208.
- Ku TL (2000) Noller JS, et al. (ed.) *Quaternary Geochronology: Methods and Applications*. Washington, DC: American Geophysical Union.
- Ku TL (1968) Protoactinium-231 method of dating coral from Barbados Island. *Journal of Geophysical Research* 73: 2271–2276.
- Ku TL (2000) Uranium-series methods. In: Noller JS, Sowers JM and Lettis WR (eds.) *Quaternary Geochronology: Methods and Applications, American Geophysical Union Reference Shelf* 4: 101–114. Washington: American Geophysical Union. <http://dx.doi.org/10.1029/RF004p0101>.
- Lee H and Schwarcz H (2000) Electron Spin Resonance dating of fault rocks. In: Noller JS, et al. (ed.) *Quaternary Geochronology: Methods and Applications*, pp. 177–186. Washington, DC: American Geophysical Union.
- Markgraf V (2001) *Interhemispheric Climate Linkages*. San Diego: Academic Press.
- Mayewski PA, Meeker LD, Twickler MS, et al. (1997) Major features and forcing of high-latitude northern hemisphere atmospheric circulation using a 110,000-year-long glaciochemical series. *Journal of Geophysical Research* 102: 26,345–26,366.
- McDougall I (1995) Potassium-argon dating in the Pleistocene. In: Rutter NW and Catto NR (eds.) *Dating Methods for Quaternary Deposits*, pp. 1–14. Ottawa: Geological Survey of Canada.
- McDougall I and Harrison TM (1988) *Geochronology and Thermochronology by the $^{40}\text{Ar}/^{39}\text{Ar}$ Method*. New York: Oxford University Press.
- McSween HY, Richardson SM, and Uhle ME (2003) *Geochemistry: Pathways and Processes*. New York: Columbia University Press.
- Miller GH and Brigham-Grette J (1989) Amino acid geochronology – Resolution and precision in carbonate fossils. *Quaternary International* 1: 111–128.
- Miller GH, Magee JW, and Jull AJT (1997) Subtropical southern hemisphere Pleistocene cooling deduced from racemization in emu eggshells from Australia. *Nature* 385: 241–244.
- O'Sullivan PE (1983) Annually-laminated sediments and the study of Quaternary environmental changes – A review. *Quaternary Science Reviews* 1: 245–313.
- Olsson IU (1986) Radiometric dating. In: Berglund BE (ed.) *Handbook of Holocene Palaeoecology and Palaeohydrology*. London: Wiley.
- Pennington W, Cambray RS, and Fisher EM (1973) Observations on lake sediments using ^{137}Cs . *Nature* 242: 324–326.
- Schwarcz H (1989) Uranium series dating. *Quaternary International* 1: 7–17.
- Schwarcz H, Buhay W, Grün R (1988) Electron spin resonance (ESR) dating of fault gouge. *US Geological Survey Open File Report* 87–673, pp. 50–64.
- Stafford TW Jr., Jull AJT, Zabel TH, et al. (1984) Holocene age of the Yuha burial: Direct radiocarbon determinations by accelerator mass spectrometry. *Nature* 308: 446–447.
- Stokes S (1999) Luminescence dating applications in geomorphological research. *Geomorphology* 29: 153–171.
- Taylor RE (1987) *Radiocarbon Dating: An Archaeological Perspective*. New York: Academic Press.
- Tuniz C, Bird JR, Fink D, and Herzog GF (1998) *Accelerator Mass Spectrometry: Ultrasensitive Analysis for Global Science*. Boca Raton, FL: CRC Press.
- Verosub KL (1988) Geomagnetic and secular variation and the dating of Quaternary sediments. In Easterbrook DJ (ed.) *Dating Quaternary Sediments*. Geological Society of America Special Publication Paper 227: 123–128.
- Wagner GA (1995) *Age Determination of Young Rocks and Artifacts*. Berlin: Springer.
- Wehmiller JF and Miller GH (2000) Aminostratigraphic dating methods in Quaternary geology. In: Noller JS, et al. (ed.) *Quaternary Geochronology, Methods and Applications*, pp. 187–222. Washington, DC: American Geophysical Union.
- Westgate JA and Naeser ND (1995) Tephrochronology and fission-track dating. In: Rutter NW and Catto NR (eds.) *Dating Methods for Quaternary Deposits*, pp. 15–28. Ottawa: Geological Survey of Canada.
- Wintle AG (1973) Anomalous fading of thermoluminescence in mineral samples. *Nature* 245: 143–144.
- Wohlfarth B, Björck S, Possnert G, et al. (1993) AMS dating of Swedish varved clays of the last glacial/interglacial transition and potential difficulties of calibrating Late Weichselian 'absolute' chronologies. *Boreas* 22: 113–128.
- Zreda M and Phillips F (2000) Cosmogenic nuclide buildup in surficial materials. In: Noller JS, et al. (ed.) *Quaternary Geochronology, Methods and Applications*, pp. 61–76. Washington, DC: American Geophysical Union.

Further Reading

- Hearty PS and Kaufman DS (2000) Whole-rock aminostratigraphy and Quaternary sea-level history of the Bahamas. *Quaternary Research* 54: 163–173.
- McDougall I (1995) Potassium-argon dating in the Pleistocene. In: Rutter NW and Catto NR (eds.) *Dating Methods for Quaternary Deposits*, pp. 1–14. Ottawa: Geological Survey of Canada.
- McSween HY, Richardson SM, and Uhle ME (2003) *Geochemistry: Pathways and Processes*. New York: Columbia University Press.
- Schwarcz H, Buhay W, and Grün R (1988) Electron spin resonance (ESR) dating of fault gouge. *US Geological Survey Open File Report* 87–673, pp. 50–64.
- Tuniz C, Bird JR, Fink D, and Herzog GF (1998) *Accelerator Mass Spectrometry: Ultrasensitive Analysis for Global Science*. Boca Raton, FL: CRC Press.

Dendrochronology

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Introduction

Dendrochronology is the science that deals with the dating and study of the annual growth increments, or tree rings, in woody trees and shrubs. When the radial growth of trees is affected by a common limiting factor, tree-ring variations are replicated across forest stands and cross-dating, or pattern matching, of these series can provide an accurate record of annual growth trends. These data can be used to date events or describe variations in environmental conditions. Due to the annual resolution and spatial breadth of tree-ring records, dendrochronological analyses provide both reliable and ubiquitous archives for dating past events and for paleoenvironmental reconstruction.

Origins and History

Early in the fifteenth century, Leonardo da Vinci noted the annual nature of tree rings, recognizing a relationship between the size of individual rings and precipitation (Stallings, 1937). The first recorded application of dendrochronology occurred in 1737 when Du Hamel and De Buffon observed a prominent frost-damaged tree ring in recently felled trees (Studhalter, 1955, 1956). By the mid-1800s, the ecological foundations of dendrochronology were established following Theodore Hartig's study of wood structures and annual tree-ring development (Schweingruber, 1988). By the end of the nineteenth century Hartig's son, Robert, had published papers on the anatomy and ecology of tree rings, using them to date hail events, frost years, and insect damage in trees (Schweingruber, 1988).

In 1904, astronomer Andrew Ellicott Douglass began examining the annual rings of Ponderosa pine trees in the Southwestern United States as a potential source for pre-instrumental climate records. He observed that narrow tree rings formed in very dry years over large geographical areas, and confirmed that cross-dating of tree-ring variability between sites could be used to assign calendar dates to individual tree rings (Douglass, 1909). By 1914 Douglass had developed a 500-year cross-dated tree-ring chronology and established a positive correlation between ring width and precipitation in the winter preceding growth (Douglass, 1914). His chronology was used to date wood samples from prehistoric pueblo sites in the American southwest and established dendrochronology as an effective dating tool (Dean, 1997).

Prior to pioneering the subfields of dendroclimatology and dendrohydrology, Schulman (1945) began working with Douglass to develop better methods of tree-ring analysis. In his quest to improve the length and sensitivity of tree-ring records,

Schulman (1954) discovered the oldest living trees in the world; the more than 4000-years-old bristlecone pines of the California White Mountains. His discovery was integral to the calibration of atmospheric ^{14}C records and set the foundation for radiocarbon dating methods (LaMarche and Harlan, 1973).

Working in Europe after World War II, German botanist Bruno Huber pioneered an ecological approach to dendrochronology. His discovery that the year-to-year variation in tree rings was not as pronounced in temperate species as it was in those from semiarid zones led to a more quantitative approach to tree-ring research. Huber developed a statistical test of similarity, the Gleichlaufigkeit (Eckstein and Pilcher, 1990; Huber, 1943), which remains an integral component of statistical cross-dating in European laboratories.

In the early 1960s, Hal Fritts established the basic physiology of radial tree growth by employing a dendrograph (Haasis, 1933) to measure the daily growth of tree rings. As detailed in his seminal publication *Tree Rings and Climate*, Fritts (1976) was able to describe many of the complex relationships between tree rings and climate and used these insights to provide annually resolved climate reconstructions for eastern and western North America (Fritts, 1991). Polge (1966) applied X-ray densitometry to the study of tree rings, showing that density records provided higher resolution dendroclimatic insights than did records developed from solely from annual ring-width increments (Lenz et al., 1976).

In the last 50 years, dendrochronology has evolved from a tool primarily employed to date monuments and archaeological pueblos, to a discipline with a variety of applications including studies of: the history and effects of pollution (Schweingruber, 1988); geological dating (Winchester and Harrison, 2000); earthquakes (Jacoby, et al., 1997); snow avalanches (Smith et al., 1994); glacier movement (Luckman and Villalba, 2001); volcanoes (Jones et al., 1995); river flooding (St George and Nielson, 2003); and insect life cycles (Swetnam and Lynch, 1993). In addition to the continued application of the techniques pioneered by Douglass, new analytical techniques (i.e., X-radiography, X-ray densitometry, neutron activation analysis) are providing insights into climate-induced morphological and chemical changes within tree rings and wood cells, and are being used to describe trace element and isotope signatures (Hughes, 2011; Kuniholm, 2001).

Biological Foundations

Tree-ring dating is possible due to the physiological behavior of trees growing in seasonal climates. In these environments

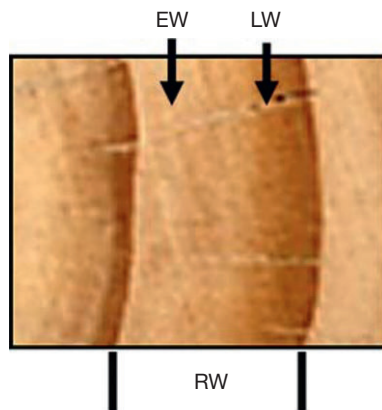


Figure 1 A cross-section from a young subalpine fir (*Abies amabilis*) tree illustrating the annual concentric ring pattern around the stem of tree. Highlighted are the annual ring width (RW), latewood (LW), and earlywood (EW) components.

radial tree growth occurs in annual increments, or 'tree rings,' whose anatomical structures vary in response to fluctuating environmental variables (Figure 1). Although there are differences between the anatomical structures of angiosperms and gymnosperms, both trees are commonly used in dendrochronology.

Radial tree growth occurs as vascular cambium cells divide. During the winter months cambium is dormant and cell production ceases until warmer and/or wetter spring conditions reactivate cambial activity (Fritts, 1976). Xylem cells formed at the beginning of the season, referred to as 'earlywood,' tend to be large and thin walled. With the onset of hotter and/or drier summer conditions cambial activity slows and smaller, thick-walled xylem ('latewood') cells are produced (Fritts, 1976). It is the marked difference between earlywood and latewood cells that make it possible to discern annual growth increments (Figure 2).

Although tree rings usually form every growing season, a tree may fail to develop an annual ring or the ring may be discontinuous around the circumference of the tree stem (i.e., a missing ring; Speer, 2010). Intra-annual growth bands, or false rings, are sometimes produced if seasonal growing conditions temporarily deteriorate (Hoffer and Tardif, 2009). False rings are usually easily detectible, as the transition from one cell type to another is gradual. In contrast, the resumption of earlywood cell production after winter dormancy produces a distinct boundary (Figure 2). Missing rings are best identified by cross-dating comparison with established tree-ring series.

Principles of Dendrochronology

Seven widely accepted principles govern the application of dendrochronology. While some are specific to dendrochronology, others have been borrowed and modified from other disciplines.

Uniformitarian Principle

The principle of uniformitarianism was proposed by James Hutton in 1785. In the context of dendrochronology, it



Figure 2 Cross-sectional view of tree rings from a mountain hemlock (*Tsuga mertensiana*) tree.

suggests that the physical and biological processes that link environmental conditions to variations in radial tree growth in the present are the same ones that existed in the past. This does not mean that past conditions were the same as present ones, but rather that tree growth was influenced in the same manner in the past as it is at present (Fritts, 1976). Hence, by understanding present climate–radial growth relationships, dendrochronologists are able to reconstruct paleoenvironmental records from tree-ring data.

Principle of Limiting Factors

Annual tree-ring growth cannot proceed faster than permitted by the most limiting environmental factor (Fritts, 1976). For instance, in arid and semiarid regions, a lack of precipitation is often limiting and the annual radial increment is a direct function of the quantity of precipitation. Other limiting factors include air temperature, snowpack, and nonclimatic parameters such as soil condition or insect infestation. This principle is central to dendrochronological research in that tree rings may be cross-dated only if one or more environmental factors becomes significantly limiting, persists long enough, and acts over a wide enough geographical area to cause widespread common patterns of ring width variation.

Principle of Aggregate Tree Growth

According to this principle, tree-ring growth is an aggregate of several factors. In a simplified linear model of radial growth Cook (1987) notes that:

$$R_t = A_t + C_t + D1_t + D2_t + E_t$$

where the observed tree-ring width (R) in any one year (subscript t) is a factor of: age-related growth trends (A); the influence of climate (C); the pulse of a localized disturbance factor



Figure 3 Illustration of effects of the principles of ecological amplitude and site selection. Shown is the radial growth response of trees to limiting environmental factors under different site conditions. The top ring-width pattern (a) is from a tree found growing near its limit and exhibiting variable ring width patterns as a result of climatic stress. The sample below, (b), is from a tree near the center of its ecological range and not limited by environmental factors. As a result, the tree is not stressed and is producing complacent tree rings with little variability.

(endogenous disturbance, i.e., blowdown) (D1); stand-wide disturbance factors (exogenous disturbance, i.e., fire) (D2); and unexplained year-to-year variability not accounted for by these variables (E). To maximize the desired growth component (signal), the other factors (noise) should be minimized. For instance, to highlight a climate signal age-related growth trends must be removed (see *Standardization*), and trees and sites selected to minimize the influence of internal and external ecological processes on tree growth (see section ‘*Principle of Site Selection*’).

Principle of Ecological Amplitude

Tree-ring growth is particularly sensitive to environmental factors at the altitudinal and latitudinal limits of a species’ range (Figure 3). For example, growth is frequently restricted by average summer air temperature in high-mountain and high-latitude settings (Luckman and Wilson, 2005).

Principle of Site Selection

Tree-ring-width series that are highly variable and controlled by a limiting environmental factor can be selected for based on site-specific characteristics. For example, trees found growing on bare rocky outcrops are often sensitive to drought conditions. Conversely, trees growing in mesic environments are usually characterized by complacent (unvarying) growth trends (Figure 3).

Principle of Cross-Dating

Cross-dating procedures ensure that individual tree rings are assigned an exact year of formation. Cross-dating provides a means to date floating tree-ring series and to develop tree-ring chronologies that extend back beyond the age of living trees. Undated wood samples can be assigned ages by cross-dating to chronologies of known age (Figure 4). For instance, the date when a tree died can be determined by matching its ring patterns with those within a chronology developed from living trees and spanning, at least in part, the same interval of time (Figure 4).

Principle of Replication

An environmental signal of interest can be maximized, and the amount of ‘noise’ minimized, by sampling more than one stem radius per tree, more than one tree per site, and/or several sites in a region (Cook, 1987; Fritts, 1976). Cross-dating samples from multiple trees minimizes the likelihood that missing or false rings from individual trees will be incorporated into a chronology. Obtaining more than one sample per tree reduces the amount of ‘intratree variability,’ or the nondesirable environmental signal particular to a given tree.

Dendrochronological Methodologies

The primary objective in developing tree-ring chronologies is to establish stationary time series that are directly comparable and appropriate for statistical analysis (Fritts, 1976). Following the collection of tree-ring samples (Speer, 2010) chronology development generally follows a five-step procedure:

(i) *Sample preparation*: Increment cores are normally glued to slotted mounting boards. The surface of the cores are then sanded and polished at increasingly finer grits until the annual ring boundaries are clearly visible. If necessary, stem cross-sections (‘cookies’) are glued to preserve their structural integrity and are then similarly polished.

(ii) *Measurement*: Assessment of annual tree-ring increments was originally undertaken using manual (Stokes and Smiley, 1968), or more recently electronic (Cropper, 1979), skeleton plotting. According to this method the widths of unusually narrow and wide rings are represented by bars of varying length. Skeleton plotting is still commonly used for basic assessments of tree growth (Brookhouse, 2006; Pelfini and Santilli, 2008; Wils et al., 2009).

Numerous instruments are now commercially available to measure annual tree-ring parameters (Maxwell et al., 2011). These include widely employed plot/sliding linear-encoding measurement stages equipped with either high powered ocular microscopes or video cameras connected to a microscope. The latter display high-resolution images on computer monitors that allow for measurement accuracies of 0.001 mm. Semi-automated analysis systems employ high-resolution scanners to capture tree-ring images that are used to demarcate anatomical tree-ring boundaries. An advantage of the latter system is

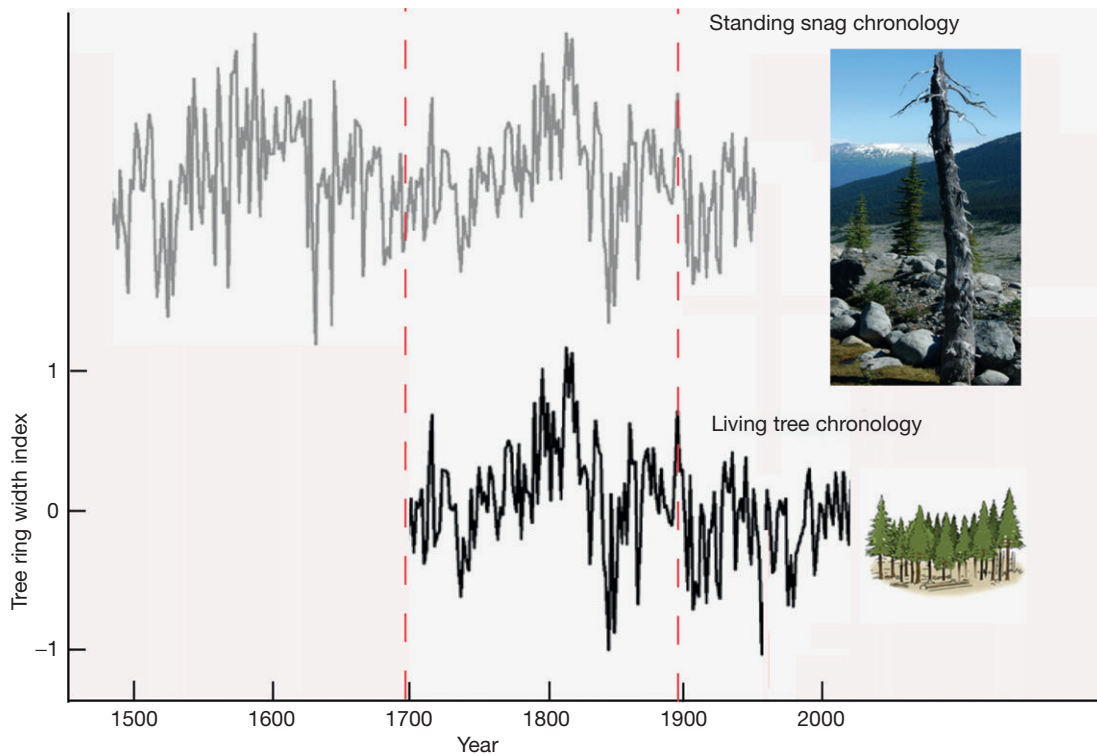


Figure 4 Illustration of how a calendar date for the growth rings of an undated wood sample can be established by cross-dating with a dated living chronology. By anchoring floating chronologies to the living chronology, it is possible to determine when a sample was killed and to extend the duration of a tree-ring chronology.

that measurement pathways can be edited and saved as image files that can be revisited during the cross-dating process.

X-ray densitometry provides the opportunity to delineate and measure intra-annual tree-ring parameters, characteristics that are valuable for describing subannual relationships between variables such as earlywood/latewood density and seasonal environmental conditions (Cleaveland, 1986; D'Arrigo et al., 1992; Robertson and Jozsa, 1988; Schweingruber et al., 1978; Zhang, 2007). More recently the dendrochronological attributes of wood microstructures have been explored (Gureyev and Evans, 1997). High-precision assessments of annual to subannual variations in density (Evans et al., 2000; Robertson and Jozsa, 1988), microfibril angle (Sarén et al., 2004), cell wall mass (Silkin and Kirdyanov, 2003), and cell dimensions (Wood et al., 2009) are providing innovative insights into the role of climate in wood anatomy.

Differentiated image brightness variables also provide a measure of wood fiber features. Reflected-light imaging techniques, such as minimum blue intensity, provide proxy density measurements useful for developing paleoclimate records (Campbell et al., 2007, 2011; Sheppard et al., 1996). Similar approaches have successfully linked wood structure characteristics to specific disturbance incidents such as flooding (Yanosky and Robinove, 1986).

Measurement of annual variations in stable carbon, oxygen, and hydrogen isotopes in tree rings have been correlated to climate and climate-related parameters (i.e., soil moisture, summer irradiance, and air temperature) to provide both annual (McCarroll and Loader, 2004; Schleser et al., 1999) and intra-annual (Helle and Schleser, 2003; Porté and Loustau,

2001) insights useful for paleoclimatic reconstruction. Recent methodological advances in sample processing have substantially reduced the costs of isotopic analyses, making them increasingly attractive for application in dendrochronological research (McCarroll and Loader, 2004).

(iii) *Cross-dating*: A pattern-matching methodology used to assign dates to individual tree-ring chronologies. Absolute dating is possible when the age of at least one cross-dated sample is known (i.e., taken from a living tree). The presence of bark on a sample confirms that the last year of growth is present (Baillie, 1982). Series of unknown age are cross-dated by matching annual tree-ring parameters to those within a dated series (Figure 4). Cross-dating serves as a check for establishing measurement errors, or for recognizing the presence of false or missing rings (Stokes and Smiley, 1968). The application of quality checking software programs, such as COFECHA (Holmes, 1983) and CDendro (Larsson, 2003), provides a statistical measure of cross-dating quality.

(iv) *Standardization*: Annual tree rings and their anatomical characteristics typically vary in response to one or more limiting growth factors, as well as to traits such as tree age, stem height above the ground, site-specific characteristics, and general productivity (Fritts, 1976). To isolate the proportion of annual growth that is controlled by the limiting factor, the impact of these external parameters is statistically removed by standardizing and transforming each variable to an index (Fritts, 1976). The application of standardization procedures requires careful consideration of the morphological properties of individual tree-ring series and the statistical analyses to be performed (Cook and Kairiukstis, 1990; Speer, 2010).

(v) *Master chronology development*: Accomplished by combining transformed time series into dimensionless, site-specific, tree-ring-width indices (Cook and Holmes, 1986; Speer, 2010). A master chronology is considered representative of the tree growth trends at a site, and can subsequently be correlated to the environmental parameters limiting that growth.

Concluding Remarks

The broad geographical distribution of trees, and the annual to subannual resolution of their annual growth rings in temperate latitudes, has permitted dendrochronological methodologies to be employed in a wide array of applications. Towner (2013) illustrates how dendroarchaeological studies continue to allow researchers to understand the impact of paleoenvironments on human activity. In paleoenvironmental studies, Luckman (2013) describes how dendroclimatologists have used tree rings to reconstruct millennia-long records of temperature, precipitation, and related variables such as streamflow at various spatial and temporal scales. Gärtner and Heinrich (2013) provide examples of how tree rings afford an annually resolved methodology for documenting prehistorical geomorphic activity and landform dynamics; long-term records that provide critical insights for modeling future landscape responses to shifting climates. The accelerating retreat of glaciers in many settings is exposing the remains of subfossil trees that dendroglaciologists are examining to date and reconstruct Holocene glacier activity. As Coulthard and Smith (2013) describe, tree rings are also providing detailed multicentury mass balance records useful for understanding the long-term glaciological response of glaciers to changing climates. Future dendrochronological research opportunities include the development of new measurement approaches and instruments, exploring multiproxy dendrochronological assessments of climate and stand disturbance, giving focused attention to trees growing in tropical latitudes (Speer, 2010), and evaluating the physiology of nonstationary radial growth responses to variable climates (Esper and Frank, 2009).

See also: Dendroclimatology. **Glacial Landforms, Tree Rings:** Dendrogeomorphology; Dendroglaciology. **Plant Macrofossil Methods and Studies:** Dendroarchaeology.

References

- Baillie MGL (1982) *Tree-Ring Dating and Archaeology*. London, UK: Croom Helm.
- Brookhouse M (2006) Eucalypt dendrochronology: Past, present and potential. *Australian Journal of Botany* 54: 435–449.
- Campbell R, McCarroll D, Loader NJ, Grudd H, Robertson I, and Jalkanen R (2007) Blue intensity in *Pinus sylvestris* tree-rings: Developing a new palaeoclimate proxy. *The Holocene* 17: 821–828.
- Campbell R, McCarroll D, Robertson I, Loader NJ, Grudd H, and Gunnarson B (2011) Blue intensity in *Pinus sylvestris* tree-rings: A manual for a new palaeoclimate proxy. *Tree-Ring Research* 67: 127–134.
- Cleaveland MK (1986) Climatic response of densitometric properties in semiarid site tree rings. *Tree-Ring Bulletin* 46: 13–29.
- Cook ER (1987) The decomposition of tree-ring series for environmental series. *Tree-Ring Bulletin* 47: 37–59.
- Cook ER and Holmes RL (1986) Users manual for program ARSTAN. In: Holmes RL, Adams RK, and Fritts HC (eds.) *Tree-Ring Chronologies of Western North America: California, Eastern Oregon and Northern Great Basin*, pp. 50–65. Tucson, AZ: Laboratory of Tree-Ring Research, University of Arizona.
- Cook ER and Kairiukstis L (1990) *Methods of Dendrochronology*. Boston, MA: Kluwer Academic Publishers.
- Coulthard BL and Smith DJ (2013) Dendroglaciology. In: Smith D (ed.) *Encyclopedia of Quaternary Science*, 2nd edn. Oxford: Elsevier.
- Cropper JP (1979) Tree ring skeleton plotting by computer. *Tree-Ring Bulletin* 39: 47–59.
- D'Arrigo RD, Jacoby GC, and Free RM (1992) Tree-ring width and maximum latewood density at the North American tree line: Parameters of climatic change. *Canadian Journal of Forest Research* 22: 1290–1296.
- Dean JS (1997) Dendrochronology. In: Aitken RE and Taylor MJ (eds.) *Chronometric Dating in Archaeology*, pp. 31–64. New York, NY: Plenum Press.
- Douglass AE (1909) Weather cycles in the growth of big trees. *Monthly Weather Review* 37: 225–237.
- Douglass AE (1914) A method for estimating rainfall by the growth of trees. In: Huntington E (ed.) *The Climatic Factor as Illustrated in Arid America*, pp. 101–121. Washington, DC: Carnegie Institute of Washington.
- Eckstein D and Pilcher JR (1990) Dendrochronology in Western Europe. In: Cook ER and Kairiukstis LA (eds.) *Methods of Dendrochronology: Applications in the Environmental Sciences*, pp. 11–13. Boston, MA: Kluwer Academic Publishers.
- Esper J and Frank D (2009) Divergence pitfalls in tree-ring research. *Climatic Change* 94: 261–266.
- Evans R, Ilic J, and Matheson C (2000) Rapid estimation of solid wood stiffness using SilvScan. *Proceedings 26th Forests Products Research Conference: Research Developments and Industrial Applications and Wood Waste Forum*, pp. 49–50. Clayton, Victoria, Australia, 19–21 June.
- Fritts HC (1976) *Tree Rings and Climate*. New York, NY: Academic Press.
- Fritts HC (1991) *Reconstructing Large-Scale Climate Patterns from Tree-Ring Data: A Diagnostic Analysis*. Tucson, AZ: University of Arizona Press.
- Gärtner H and Heinrich I (2013) Dendrogeomorphology. In: Smith D (ed.) *Encyclopedia of Quaternary Science*, 2nd edn. Oxford: Elsevier.
- Gureyev TE and Evans R (1997) A method for measuring vessel-free density distribution in hardwoods. *Wood Science and Technology* 33: 31–42.
- Haasis FW (1933) Shrinkage and expansion in woody cylinders of living trees. *American Journal of Botany* 20: 85–91.
- Helle G and Schleser GH (2003) Seasonal variations of stable isotopes from tree-rings of *Quercus petraea*. In: Schleser GH, Winiger M, and Bräuning A, et al. (eds.) *Tree Rings in Archaeology, Climatology, and Ecology*, vol. 1, pp. 66–70. Proceedings of the Dendrosymposium 11–13 April 2003.
- Hoffer M and Tardif JC (2009) False rings in Jack pine and black spruce trees from eastern Manitoba as indicators of dry summers. *Canadian Journal of Forest Research* 39: 1722–1736.
- Holmes RL (1983) Computer-assisted quality control in tree-ring dating and measurement. *Tree-Ring Bulletin* 43: 69–75.
- Huber B (1943) Über die Sicherheit jährlichchronologische Datierung. *Holz als Roh und Werkstoff* 6: 263–268.
- Hughes MK (2011) Dendroclimatology in high-resolution paleoclimatology. In: Hughes MK, Swetnam TW, and Diaz HF (eds.) *Dendroclimatology: Developments in Paleoenvironmental Research*, pp. 17–36. New York, NY: Springer.
- Jacoby GC, Bunker DE, and Benson BE (1997) Tree-ring evidence for an A.D. 1700 Cascadia earthquake in Washington and northern Oregon. *Geology* 25: 999–1002.
- Jones PD, Briffa KR, and Schweingruber FH (1995) Tree-ring evidence of the widespread effects of explosive volcanic eruptions. *Geophysical Research Letters* 22: 1333–1336.
- Kuniholm P (2001) Dendrochronology and other applications of tree-ring studies in archaeology. In: Brothwell DR and Pollard AM (eds.) *The Handbook of Archaeological Sciences*. London: John Wiley & Sons, Ltd.
- LaMarche VC Jr. and Harlan TP (1973) Accuracy of tree-ring dating of bristlecone pine for calibration of the radiocarbon time scale. *Journal of Geophysical Research* 78: 8849–8858.
- Larsson LA (2003) *CDendro: Cybis Dendro dating program*. Available at <http://www.cybis.se>.
- Lenz O, Schär E, and Schweingruber FH (1976) Methodische Probleme bei der radiographisch-densitometrischen Bestimmung der Dichte und der Jahrringbreiten von Holz. *Holzforschung* 30: 114–123.
- Luckman BH (2013) Dendroclimatology. In: Smith D (ed.) *Encyclopedia of Quaternary Science*, 2nd edn. Oxford: Elsevier.

- Luckman BH and Villaiba R (2001) Assessing the synchronicity of glacier fluctuations in the Western Cordillera of the Americas during the last millennium. In: Markgraf V (ed.) *Interhemispheric Climate Linkages*, pp. 119–140. San Diego, CA: Academic Press.
- Luckman BH and Wilson RJ (2005) Summer temperatures in the Canadian Rockies during the last millennium: A revised record. *Climate Dynamics* 24: 131–144.
- Maxwell RS, Wixom JA, and Hessler AE (2011) Technical note: A comparison of two techniques for measuring and crossdating tree rings. *Dendrochronologia* 29: 237–243. <http://dx.doi.org/10.1016/j.dendro.2010.12.002>.
- McCarroll D and Loader NJ (2004) Stable isotopes in tree rings. *Quaternary Science Reviews* 23: 771–801.
- Pelfini M and Santilli M (2008) Frequency of debris flows and their relation with precipitation: A case study in the central Alps, Italy. *Geomorphology* 101: 721–730.
- Polge H (1966) Etablissement des courbes de variation de la densité du bois par exploration densitométrique de radiographies d'échantillons prélevés à la tarière sur des arbres vivants. Applications dans les domaines technologiques et physiologiques. *Annales des Sciences Forestières* 23: 1–206.
- Porté A and Loustau D (2001) Seasonal and interannual variations in carbon isotope discrimination in a maritime pine (*Pinus pinaster*) stand assessed from the isotopic composition of cellulose in annual rings. *Tree Physiology* 21: 861–868.
- Robertson EO and Jozsa LA (1988) Climatic reconstruction from tree rings at Banff. *Canadian Journal of Forest Research* 18: 888–900.
- Sarén MP, Serimaa R, Andersson S, Saranpää P, Keckes J, and Fratzl P (2004) Effect of growth rate on mean microfibril angle and cross-sectional shape of tracheids of Norway spruce. *Trees – Structure and Function* 18: 354–362.
- Schleser GH, Helle G, Lücke A, and Heinz V (1999) Isotope signals as climate proxies: The role of transfer functions in the study of terrestrial archives. *Quaternary Science Reviews* 18: 927–943.
- Schulman E (1945) Tree-ring hydrology of the Colorado Basin. *Laboratory of Tree-Ring Research Bulletin* 16: 1–51.
- Schulman E (1954) Longevity under adversity in conifers. *Science* 119: 396–399.
- Schweingruber FH (1988) *Tree Rings: Basics and Applications of Dendrochronology*. Dordrecht, The Netherlands: D. Reidel.
- Schweingruber FH, Fritts HC, Braker OU, Drew LG, and Schar E (1978) X-ray techniques as applied to dendroclimatology. *Tree-Ring Bulletin* 38: 61–91.
- Sheppard PR, Graumlich LJ, and Conkey LE (1996) Reflected-light image analysis of conifer tree rings for reconstructing climate. *The Holocene* 6: 62–68.
- Silkin PP and Kirdyanov AV (2003) The relationship between variability of cell wall mass of earlywood and latewood tracheids in larch tree-rings, the rate of tree-ring growth and climatic changes. *Holzforschung* 57: 1–7.
- Smith DJ, McCarthy DP, and Luckman BH (1994) Snow-avalanche impact pools in the Canadian Rocky Mountains. *Arctic and Alpine Research* 26: 116–127.
- Speer J (2010) *Fundamentals of Tree-Ring Research*. Tucson, AZ: University of Arizona Press.
- Stallings J (1937) Some early papers on tree-rings. *Tree-Ring Bulletin* 3: 27–28.
- St George S and Nielson E (2003) Palaeoflood records for the Red River, Manitoba, Canada, derived from anatomical tree-ring signatures. *The Holocene* 13: 547–555.
- Stokes MA and Smiley TL (1968) *An Introduction to Tree-Ring Dating*. Chicago, IL: University of Chicago Press.
- Studhalter RA (1955) Tree growth: I. Some historical chapters. *The Botanical Review* 21: 1–72.
- Studhalter RA (1956) Early history of crossdating. *Tree-Ring Bulletin* 21: 1–4.
- Swetnam TW and Lynch AM (1993) Multicentury, regional-scale patterns of western spruce budworm outbreaks. *Ecological Monographs* 63: 399–424.
- Towner R (2013) Dendroarchaeology. In: Smith D (ed.) *Encyclopedia of Quaternary Science*, 2nd edn. Oxford: Elsevier.
- Wils THG, Robertson I, Eshetu Z, Touchan R, Sass-Klaassen U, and Marcin K (2009) Crossdating *Juniperus procera* from North Gondar, Ethiopia. *Trees – Structure and Function* 25: 71–82.
- Winchester V and Harrison S (2000) Dendrochronology and lichenometry: Colonization, growth rates and dating of geomorphological events on the east side of the North Patagonian Icefield, Chile. *Geomorphology* 34: 181–194.
- Wood L, Hartley ID, and Watson P (2009) Determining hybridization in Jack pine and lodgepole pine from British Columbia. *Wood and Fiber Science* 41: 386–395.
- Yanosky TM and Robinove CJ (1986) Digital image measurement of the area and anatomical structure of tree rings. *Canadian Journal of Botany* 64: 2896–2902.
- Zhang SY (2007) Variations and correlations of various ring width and density features in European oak: Implications in dendroclimatology. *Wood Science and Technology* 31: 63–72.

Dendroclimatology

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Introduction

Tree-ring series provide the most widely distributed and easily accessible archive of annually resolved proxy climate data. In regions with well-defined seasonal growth, the annual growth rings of trees provide both chronological control and a continuous time series of proxy environmental variables. The year-to-year variability of the physical (e.g., width and density) and chemical properties of these annual rings provides potential proxies for the environmental factors that influence tree growth. Dendroclimatology can be defined simply as 'the science that uses tree rings to study present climate and reconstruct past climate' (Grissino-Meyer, n.d.) and during the past 30 years has become a major tool in the reconstruction of climates of the past millennium in many areas of the world. A detailed review of recent developments in the field is given in Hughes et al. (2011a,b).

Basic Principles and Methods

The process of matching patterns of tree-ring width between series (cross-dating) is one of the fundamentals of dendrochronology. It is used to verify the dating of tree-ring series and, in some cases, to demonstrate that the series are annual. For cross-dating to be successful, the ring parameters must vary synchronously among series, reflecting a common response to an external variable. Tree growth is influenced by many factors, but often only one is strongly and consistently limiting. Successful dendroclimatological studies target sampling sites where a single climate parameter is, or is expected to be, the limiting factor of growth and therefore the primary control of interannual variability in ring width. The presence and strength of the common signal between tree-ring series within a site is a measure of the degree of climate control of growth: correlation between tree-ring series from different sites across a broad area almost inevitably reflects a common climate control, as no other controlling factors vary systematically at these spatial scales. Therefore successful cross-dating usually establishes the presence of a climate signal in tree-ring series.

Trees growing at their range limits are often the most sensitive to climate variability, whereas trees in mid-range or interior forest sites may show 'complacent' ring-width series with little interannual variability. Classically, latitudinal and altitudinal treeline sites are most sensitive to temperature variations, whereas lower treeline and/or forest border sites are moisture sensitive; however, there are exceptions. Bald cypress trees growing in swamps in the southeastern United States are sensitive to the water level (and therefore precipitation), and growth at many treeline sites in arid mountains is sensitive to both precipitation and temperatures (e.g., Bristlecone pines (*Pinus aristata* and *P. longaeva*)). The key to dendroclimate research is careful site selection, as the limiting factor to growth

can also vary based on microsite considerations. Dendroclimate studies further assume that the growth-limiting factor at a site does not change over time, and therefore contemporary tree-ring climate relationships can be used to reconstruct past climate conditions.

Models of cambial processes in trees have recently been developed and used to simulate tree-ring widths and cell development in response to selected climatic and site variables (see Vaganov et al., 2011 for a review). However, most dendroclimate studies have been carried out using ring-width data. Annual tree rings can be divided into earlywood (larger, thin-walled cells) and latewood (smaller cells with thicker walls). The earlywood-latewood transition is usually gradational but the latewood-earlywood boundary is sharply defined and marks the end of the growth year. Traditionally, ring width is the primary variable used for dendroclimate work, but more recently other variables such as earlywood width or latewood width have been successfully used for seasonal precipitation reconstructions (e.g., Meko and Baisan, 2001; Stahle et al., 2000). Tree-ring densitometry was initially developed by Hubert Polge (1966) in France and applied to climate studies by Parker and Henschel (1971). Traditionally, precise thin sections are cut from the cores, x-rayed, and the negatives scanned with a high-resolution densitometer to produce a continuous density trace orthogonal to the rings. Several ring-width and density variables can be obtained from these scans. Maximum latewood density (MXD) reflects the thickening of cell walls at the end of the growing season and therefore represents a narrower climatic window than more traditional ring-width parameters (Schweingruber et al., 1978). MXD in northern conifers is very strongly related to summer half-year temperatures and has been used in several important temperature reconstructions. Recently non-film-based densitometric systems have been developed for tree-ring studies (see <http://www.geog.uvic.ca/uvtr/uvtr.htm>), and other approaches such as blue intensity measures have been developed as alternates (Campbell et al., 2007; McCarroll et al., 2010).

Chronology Development

The fundamental data in dendroclimatology are tree-ring chronologies – average series of, for example, tree-ring widths from a stand or sample site of a single species growing under similar conditions. Individual chronologies usually consist of paired radii from at least 10–20 erect trees without growth anomalies. The age assigned to each measured ring is verified by the process of cross-dating the series against each other and/or reference sequences (matching ring patterns either visually or by computer software) to ensure secure dating and to check for the presence of 'false' or locally absent rings. Prior to averaging, each tree-ring series is converted to indices by removing low-frequency age-related trends. This 'standardization' process is

usually accomplished by dividing the measured ring parameter for each year by values derived from a theoretical curve (commonly a negative exponential or linear regression) that converts these data to a stationary series with a mean of 1 and similar variance throughout the series. This approach ensures that each sample contributes equally to the average series by removing differences in mean ring widths between series (e.g., wider rings in relatively young trees, narrower rings in older trees). Several approaches to standardization are available (see, e.g., Cook and Briffa, 1990; Fritts, 1976) and are the subject of continuing discussion because standardization can modify the common (climate) signal retained within these chronologies. This is a significant concern in the construction of long chronologies from overlapping cross-dated segments of living and subfossil material where standardization may remove climatically related low-frequency trends from tree-ring series. Recent studies indicate that the frequencies retained may be as little as one-third of the mean segment length in the chronology (the 'segment length curse', Cook et al., 1995). Special techniques have been developed to try and retain this low frequency signal (see Briffa and Melvin, 2011; Briffa et al., 2001; Esper et al., 2003a,b).

In any chronology, sample depth decreases further back in time as there are fewer older trees. Expressed population signal (EPS) is an objective measure of signal strength, based on intercore correlation and sample size, that can be used to identify those parts of a chronology (usually the oldest sectors) where signal strength falls below acceptable levels. Wigley et al. (1984) suggest that an EPS value of 0.85 is a reasonable threshold for adequate signal strength. The effective sample size for chronology development depends on the strength of the common signal between these series; for example, at sites where intercore correlation averages approximately 0.6, four or five trees may provide adequate signal strength but over 20 trees would be needed at sites with mean intercorrelation values of approximately 0.2 (Briffa and Jones, 1990).

Reconstruction of Climate Variables

Identification and capture of the climate signal in tree-ring series is accomplished through the process of sampling and chronology development. Isolation of the climate signal is best described by the 'Conceptual Linear Aggregate Model' developed by Cook (1990) following Graybill (1982). In this model, the variation in ring width can be decomposed into several components, namely, $R_t = A_t + C_t + \delta D1_t + \delta D2_t + E_t$. In this expression, R_t is the ring-width series, A_t is the age-size related trend of ring width, C_t is the climatically related environmental signal, $\delta D1_t$ and $\delta D2_t$ are local ($D1$) or stand-wide ($D2$) nonclimatic disturbances, and E_t is unexplained year-to-year variability (Cook, 1990). Ideally, the E term is random and minor, and can be minimized by averaging sufficient tree-ring series. The disturbance effects may be present or absent and can be avoided by examination of the individual series ($D1$), by careful site selection prior to sampling ($D2$), or by comparison of the record with that from adjacent undisturbed sites. The age-related term is removed by standardization. Averaging of the indexed tree-ring series emphasizes the common climate signal, and conceptually the equation simplifies to $R_t \approx C_t$.

Identification of the climate signal in tree-ring series requires calibration against instrumental climate data that overlap part of the tree-ring series. These climate data should be spatially representative and homogeneous (i.e., high-quality series that are continuous, complete, and corrected for any changes in station location, instrumentation, measurement practices, etc.). They may be single-station data or aggregated at appropriate spatial scales that vary with the parameter being investigated. Although proximal sites are ideal, proximity is more significant for precipitation than temperature data because of the greater spatial coherence of temperature fields (Jones and Thompson, 2003). Exploratory response function or correlation analyses are used to identify the strongest relationships between tree growth and a wide range of monthly, seasonal, or annual temperature or precipitation variables for periods up to 2 years prior to the growth year (see Fritts, 1976). Subsequent analyses develop a transfer function model between the most significantly related climate variable (annual precipitation, summer temperatures, spring precipitation, etc.) and a series of candidate tree-ring-derived variables (ring width, ring width of the previous year, maximum density, etc.). The tree-ring data are then entered into the model to reconstruct or hindcast the climate variable. Dendroclimate reconstructions are unique in that they are routinely verified against data withheld from the initial calibration and error terms can be applied to these reconstructions (Cook and Kairiukstis, 1990). Ideally, if the calibration dataset is long, then split-period verification is used. Calibration (transfer function development) is carried out using one-half of the data and the predicted values compared ('verification') against the data withheld from calibration. Subsequently, the procedure is reversed and, if these results are satisfactory (i.e., the relationship is time-stable) and pass a series of statistical tests, the entire joint dataset is used to derive the transfer function used to reconstruct the climate variable for the entire chronology. Reconstructions are generally modeled linearly using regression-based procedures. For short calibration datasets, the 'leave out one' procedure (Gordon, 1982) may be used. In addition to simple reconstructions for individual sites, extensive single- or multispecies chronology networks allow the reconstruction of temperature, precipitation, or atmospheric fields over large areas. An interesting discussion of correlation and response functions and the derivation of chronology statistics is given in Cook and Pederson (2011).

Historical Development

Although there are scattered examples from Europe of tree-ring counting dating back several hundred years (see Schweingruber, 1988), the 'father' of dendrochronology is generally considered to be the astronomer A.E. Douglass who first developed the principles of cross-dating and built chronologies based on drought-sensitive trees from the American Southwest, ultimately using them to date archaeological sites. In 1937, he established the Laboratory of Tree-Ring Research (LTRR) in Tucson, Arizona, which was the dominant center for dendrochronology in the Americas for the next 50 years. Edmund Schulman made significant early developments in dendroclimate research at Tucson during the 1930–50s,

'discovering' the ancient Bristlecone Pines in the White Mountains that were used for the first calibration of ^{14}C dating. The oldest yet known pine was 4862 years old, and the oldest presently living tree is >4765 years. Schulman also carried out pioneer reconnaissance studies of trees in similar semiarid environments in Canada, Argentina, Chile, and Mexico. In Europe, Bruno Huber took up Douglass' ideas and with his students undertook studies of trees within the temperate mountains of Europe between the late 1930s and 1960s. Modern quantitative dendroclimatology, while drawing on these earlier studies, stems largely from the classic work of Hal Fritts (1976) at LTRR, who documented many of the basic principles of the science and led the application of computing techniques and statistics to tree-ring analysis which revolutionized the discipline. Much of the classic work from Tucson focused on ring-width studies of arid land conifers. However, in the last 35 years, the research field has expanded rapidly and diversified with the development of new laboratories, techniques, and research centers (see listings in Grissino-Meyer, n.d.), engendered in part by the climate change debate and the need for long, annually resolved proxy climate records (Hughes, 2002, 2011). Dendroclimatology is now applied to a broad range of species and environments that cover the globe from forest tundra at 72°N in Siberia to scrubby rainforest at 56°S at Cape Horn, or from sea level in Alaska to higher than 4900 m on the Bolivian Altiplano.

Although dendroclimatology has typically utilized physical measures of tree-ring characteristics, increasing attention has been paid to the study of stable isotopes (mainly $\delta^{13}\text{C}$, $\delta^{18}\text{O}$ and δD) in tree rings over the last 20 years. Gagen et al. (2011) provide a recent review of this work, noting, in particular, that isotopic series are seen to have higher signal strength, no long-term age-related trends, and potentially different climate signals than traditional width and density variables, especially in tropical trees that lack annual rings.

Recognition of the increasing utility of tree-ring records for paleoclimate study and the expanding scales of investigation created the need for the archiving and exchange of data and the development of an easily accessible international database with standard protocols for format. The International Tree-Ring Data Bank (ITRDB) was founded in 1974 and has been housed at the NOAA Paleoclimatology Data Center in Boulder, Colorado, USA, since 1992. Initially based on the LTRR collections, it is now a truly global (if unevenly distributed) data base that presently holds over 2000 chronologies from over 150 species on 6 continents. These data are freely accessible online at www.ncdc.noaa.gov/paleo/paleo.html together with statistical programs for the analysis of tree-ring data from LTRR and other laboratories. The website also provides access to many tree-ring-based temperature and precipitation reconstructions plus links to other collections of tree-ring data.

Applications of Dendroclimatology

Dendroclimatological techniques have been used to reconstruct a wide range of paleoclimate variables that have progressively expanded from local to regional and continental scales. In providing examples of these applications, this article proceeds from relatively simple single-chronology examples to the

results from large-scale networks. For a more comprehensive review, see Hughes et al. (2011a,b).

Single-Site Reconstructions

The earliest recognition of climate sensitivity in North American tree rings came from the American Southwest, where Douglass first exploited the relationships between narrow rings and precipitation to infer major droughts based on tree-ring records from precipitation-sensitive trees. Drought-sensitive conifers show some of the strongest relationships with ring-width values (Figure 1). Many reconstructions have been developed from such relationships to show seasonal or annual precipitation totals. In some cases where there is more than one 'wet season,' seasonal precipitation amounts may be reconstructed separately based on differences between the earlywood, latewood, or total ring chronologies. For example, latewood widths from Douglas fir (*Pseudotsuga menziesii*) in Central Mexico are strongly correlated with the April–June rainfall, which is critical for the rain-fed maize crop. Therefore, the reconstruction of maize yields in Figure 2 is essentially a precipitation reconstruction in which the major droughts are verified by historical records (Therrell et al., 2006). Drought magnitude and frequency can also be determined from reconstructions of climate indices such as the Palmer Drought Severity Index (PDSI).

Reconstructions of temperature from ring-width data are usually for summer season, or annual values. Jacoby and Cook (1981) were the first to demonstrate strong linear relationships between spruce ring widths and June–July temperatures for a site in the central Yukon (65°N). This record shows strong low-frequency variability and, together with 19 other treeline chronologies from North American, Scandinavia, and Russian treelines, provided one of the early reconstructions of a Northern Hemisphere temperature record (D'Arrigo and Jacoby, 1993). Significant summer temperature reconstructions have also been developed for northern conifers based on MXD data or combined ring-width and MXD data (see for example, Luckman and Wilson, 2005).

Routinely recoverable tree-ring records in many parts of the world are less than 2–300 years, as heart rot, fire, or other disturbances limit the age of trees. Long, climate-sensitive chronologies are therefore of particular importance, and attention has been given to sites with long-lived species, well-preserved subfossil wood, or both. Many of these chronologies come from the range limits of species, particularly at Northern Hemisphere treelines with old slow-growing trees and/or good preservation of wood at dry or cold sites (Briffa, 2000). The greatest concentration of these long chronologies is in the SW United States (Hughes and Graumlich, 1996) but millennial chronologies are now developed in Northern Europe (see Briffa and Matthews, 2002), Canada (e.g., Luckman and Wilson, 2005), Mexico (Stahle et al., 2011), the Alps (Büntgen et al., 2005), and Asia (e.g., D'Arrigo et al., 2001; Esper et al., 2003a,b). These long chronologies, combined with other high-resolution proxy climate series, are primary inputs for several Northern Hemisphere annual or summer temperature reconstructions (e.g., Briffa et al., 2004; Esper et al., 2002; Mann et al., 1999) for the last millennium. Although these show a fairly similar pattern for the last 3–400 years (Figure 3), there

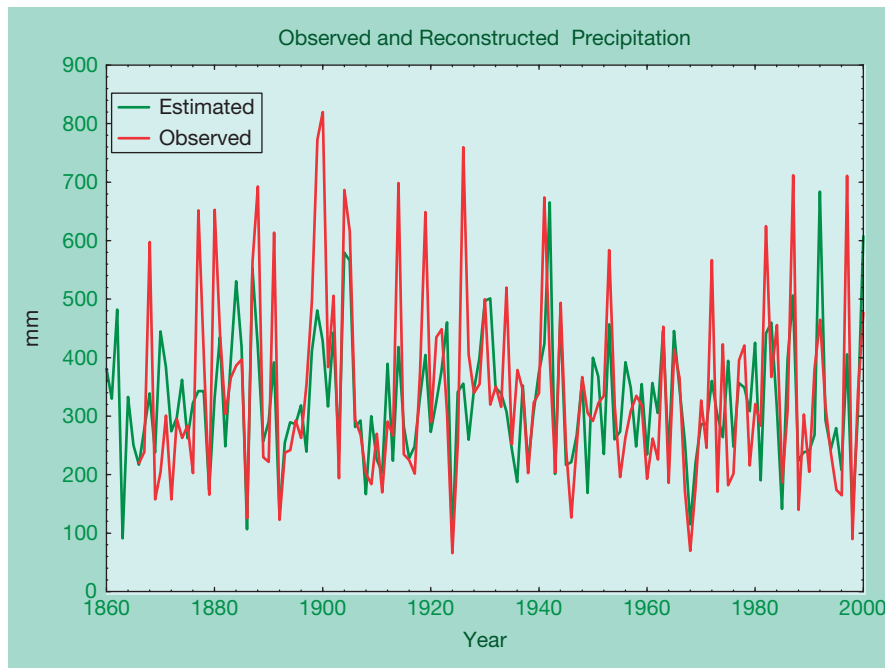


Figure 1 Actual and reconstructed precipitation for Santiago, Chile, 1866–2000, based on ring widths of *Austrocedrus chilensis* (J.A. Boninsegna, personal communication). The adjusted r^2 value is 0.54.

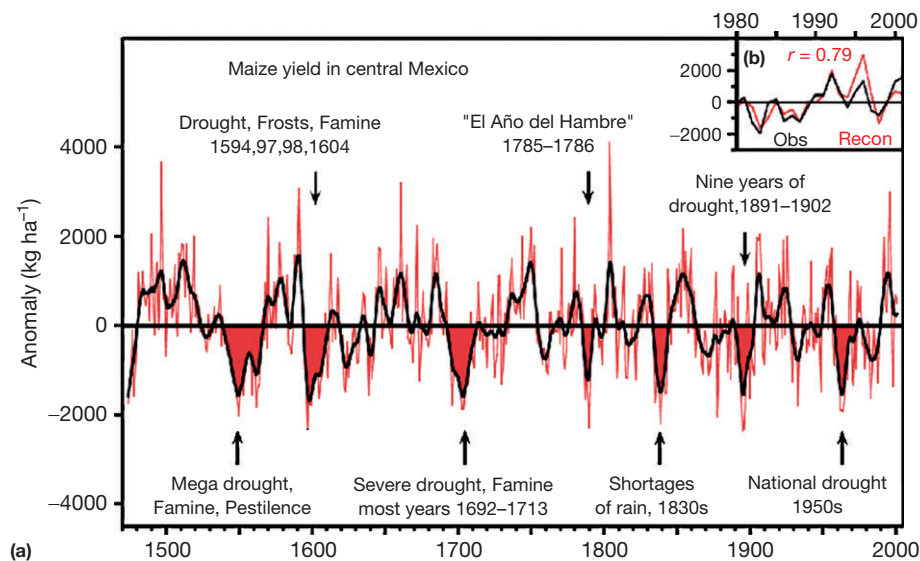


Figure 2 Reconstructed crop yield for central Mexico from 1480 to 2000 based on latewood ring width of Douglas fir growing at Cuauhtemoc la Frangua, Mexico. Maize cultivation is 'rain-fed' and the yield is strongly correlated with April–June rainfall. Dating for the major droughts is based on documentary records. Reproduced from Therrell et al. (2006).

are differences in the earlier records related to the relatively sparse and unevenly distributed sample base and the methods of standardization used (see Rutherford et al., 2005). This situation should improve as more chronologies and a more widespread distribution of long chronologies become available; however, it is clear from reconstructions of the last few hundred years that, although some variations may be globally

synchronous, for example, those forced by solar activity, there is considerably spatial variability at all timescales.

Currently, there are fewer long chronologies (and related temperature reconstructions) from the Southern Hemisphere, in part because of the restricted land area. However, recent work has significantly increased coverage in Tasmania, New Zealand, and Southern South America (see Boninsegna et al.,

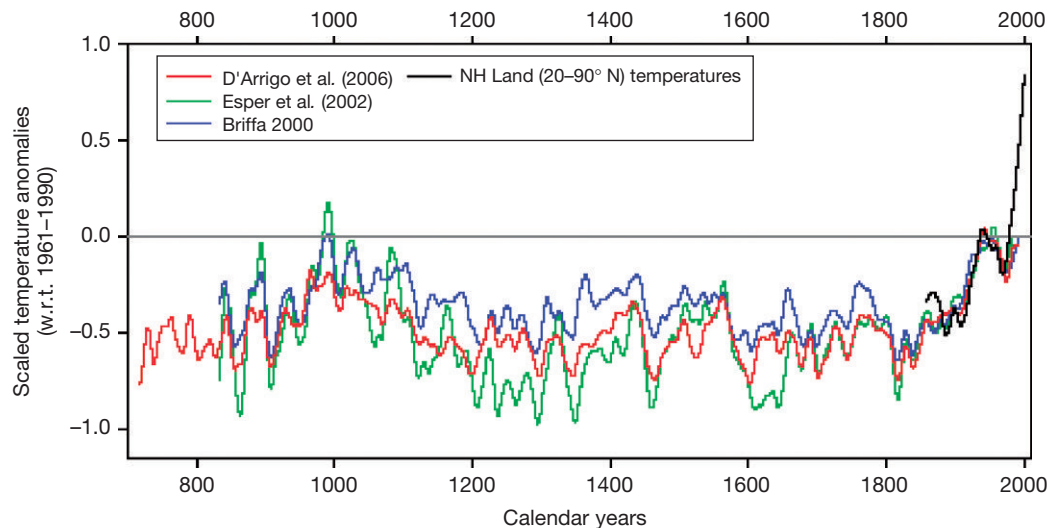


Figure 3 Comparison of three millennial-length Northern Hemisphere (NH) temperature reconstructions based exclusively on tree-ring data and standardized to emphasize low-frequency variance. The reconstructions are Briffa (2000), Esper et al. (2002), and D'Arrigo et al. (2006a,b,c) and are scaled to NH land temperatures (R. Wilson, personal communication).

2009; Roig et al., 2010). Millennial and multimillennial temperature reconstructions have been made from the Huon Pine in Tasmania (Cook et al., 2000), silver pine in New Zealand (Cook et al., 2002a,b) and from *Fitzroya cupressoides* (Alerce) in Chile and Argentina (Lara and Villalba, 1993; Lara et al., 2000). Precipitation reconstructions 6–800 years long are now available using *Austrocedrus chilensis* in Chile (Christie et al., 2010; Le Quesne et al., 2006).

Tree-Ring Networks

The widespread distribution of suitable sites for dendroclimatology has resulted in considerable effort to build single or multichronology networks to reconstruct spatial patterns of climate variability. Reconstructions based on multiple sites also reduce the influence of local site factors and enhance the common regional signal in the reconstructed record. The first explicit attempt was made by Harold Fritts (1976, 1991) and colleagues who used a network of 65 moisture-sensitive ring-width chronologies to reconstruct spatial patterns of temperature, precipitation, and atmospheric pressure across western North America. In a parallel endeavor, Fritz Schweingruber and colleagues built a circumpolar network of over 387 MXD and ring-width chronologies (Schweingruber and Briffa, 1996), which permit the development of hemispheric-scale warm season temperature histories and the study of spatial patterns over the last 300 years. Assigning 303 chronologies from this network to 92, $5^\circ \times 5^\circ$ temperature grid boxes has yielded maps of circumpolar summer temperature patterns for each year since 1600 (Figure 4) (Briffa et al., 2002). The composite Northern Hemisphere (20–70° N) series developed from these data clearly shows the global impact of major volcanic eruptions on summer temperatures over the last 600 years (Figure 5). Cool and/or wet summers, following major volcanic eruptions, have poorly developed latewood and are characterized by 'light' rings in MXD chronologies. Decreased

temperatures following major volcanic activity may also be inferred by the presence of 'frost rings' which show cell disruption due to freezing at certain high-elevation sites (LaMarche and Hirschboeck, 1984). The chronological precision of these tree-ring dates is comparable to documentary sources and superior to that available from ice cores or radiocarbon dating for events prior to the historic record, enabling tentative calendar dates to be ascribed to major volcanic or possibly cometary impact events in the last 2–3000 years (Baillie, 1996; Grudd et al., 2000).

The major recent developments using networks of tree-ring data have targeted drought-sensitive records, usually presented in a gridded data base. Expansion of the Fritts network, incorporating results from many investigators, led to the development of a network of 155 PDSI reconstructions for a 2° latitude by 3° longitude grid that essentially covered the coterminous United States (Cook et al., 1999). This was subsequently expanded to 286 grid points ($2.5^\circ \times 2.5^\circ$) that cover most of North America from the Yukon to Yucatan (Cook et al., 2004). These reconstructions provide a powerful tool for the analysis of hydroclimate variability across North America (Figures 6 and 7) and exploration of the spatial extent and duration of drought, placing twentieth-century records in a long-term context (Fye et al., 2003). In the western United States, this new reconstruction has 68 grid squares with records that extend back to AD 800, showing the extent of significant droughts in the twelfth and thirteenth centuries and even a major eighth-century drought that had important impacts on the cultural history in pre-Hispanic America (Acuna-Soto et al., 2005; Stahle and Dean, 2011). The North American Drought Atlas derived from these data is freely accessible (<http://www.ncdc.noaa.gov/paleo/pdsi.html>) and provides a useful tool for exploratory analyses that demonstrate potential influence of selected atmospheric circulation patterns on precipitation and growth of tree-rings across North America (e.g., St. George et al., 2009). Cook et al. (2010) have published a similar gridded PDSI data base (also available from the NOAA paleoclimate website) covering a large area in SE Asia which is

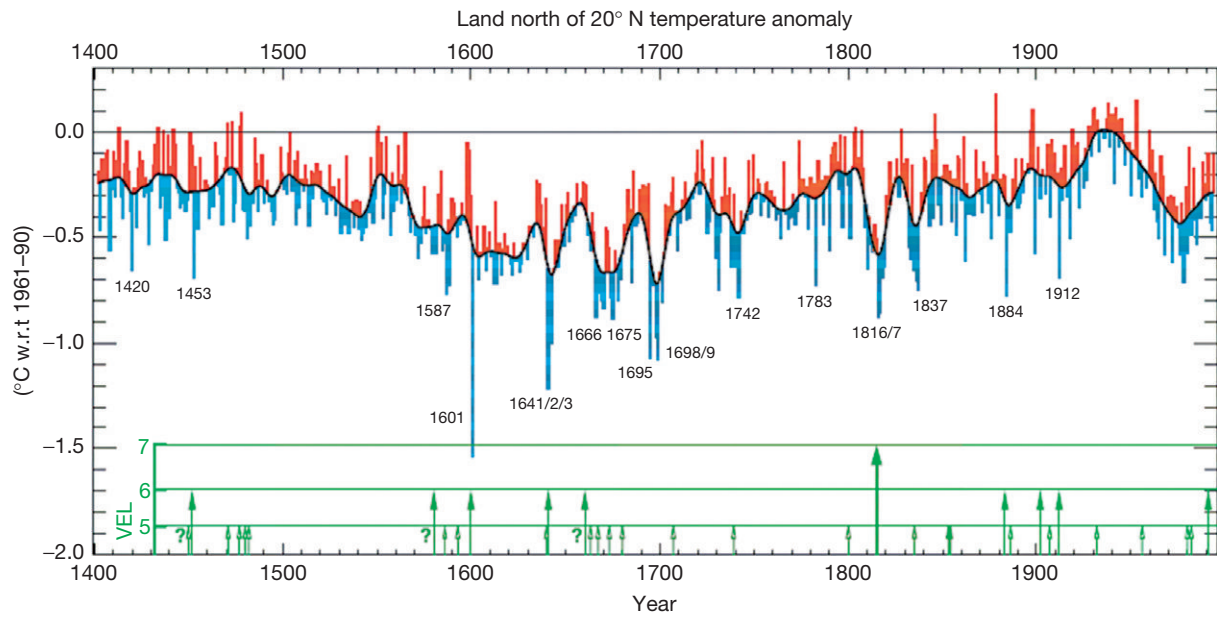


Figure 4 Estimated April–September temperature anomalies for the land area between 20° and 70° N, smoothed with a 25-year filter and based on MXD. The lower diagram shows values of the volcanic explosivity index. Eruptions with uncertain dating are indicated by question marks. Note the strong cooling trend reconstructed for the late twentieth century, reflecting a loss of temperature signal in these data. Reproduced from Briffa KR, Osborn TJ, and Schweingruber FH (2004) Large-scale temperature inferences from tree rings: A review. *Global and Planetary Change* 40: 11–26.

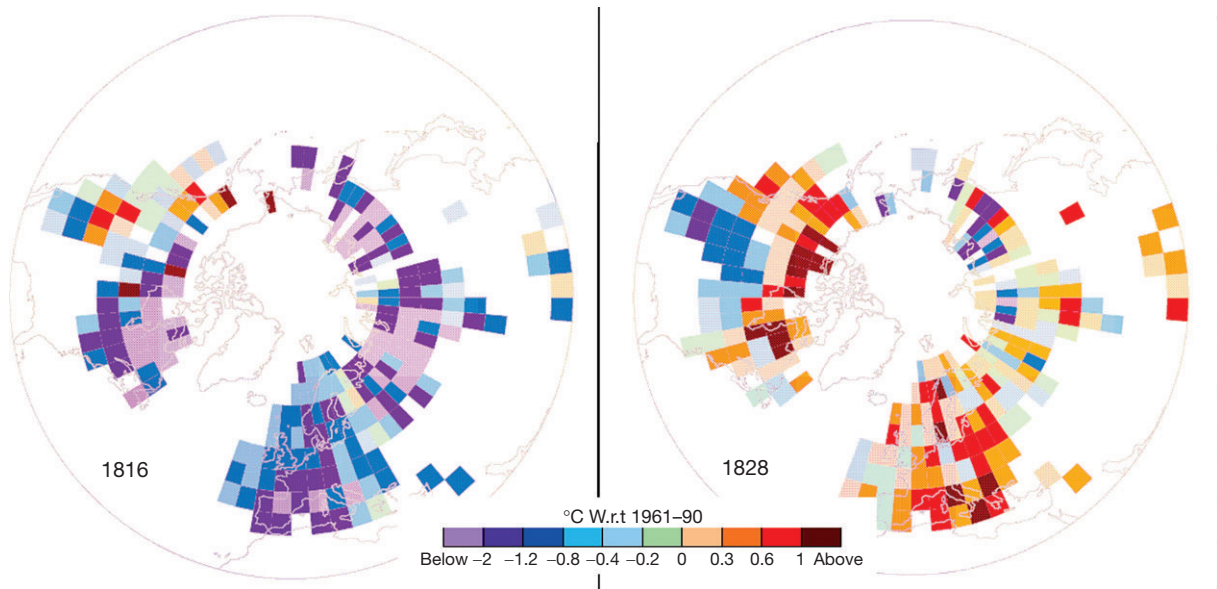


Figure 5 Mean April–September gridded temperature anomalies for 1816 and 1828 estimated from MXD tree-ring data. These are two of the most extreme years in the nineteenth century (1816 follows the Tambora eruption). Note the regional differences in temperatures (K. Briffa, personal communication, see Briffa et al., 2002, 2004).

affected by the Asian Monsoon. This database used 328 chronologies to develop data for 534, $2.5^\circ \times 2.5^\circ$ grid squares over the period from 1300 to 2005 and provides the first long-term picture of monsoonal variability prior to the instrumental climate record. Recently, Neukom et al. (2010a,b) have used new tree-ring collections and other proxy data to reconstruct separate summer and winter surface air temperature and precipitation fields for a $0.5^\circ \times 0.5^\circ$ grid that covers southern

($>20^\circ$ S) South America and provides records between 300 and 900 years long. Collectively, these gridded reconstructions provide annually resolved data on major elements of the global climate system and its long-term variability, particularly for the circum-Pacific region. These studies are critical to documenting whether the decadal and longer variability seen in available instrumental records is characteristic of the long-term behavior of the climate system.

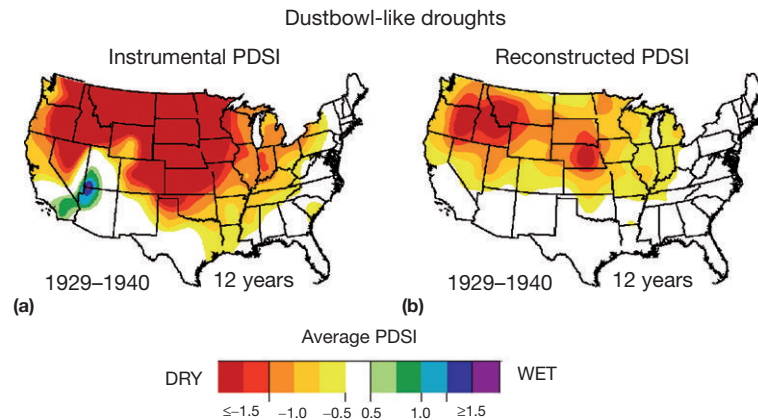


Figure 6 Comparison of instrumental and reconstructed PDSI data for the United States during the 'dustbowl' drought of 1929–1940. Reproduced from Fye FK, Stahle DW, and Cook ER (2003) Paleoclimate Analogs to 20th century moisture regimes across the US. *Bulletin of the American Meteorological Society* 84: 901–909.

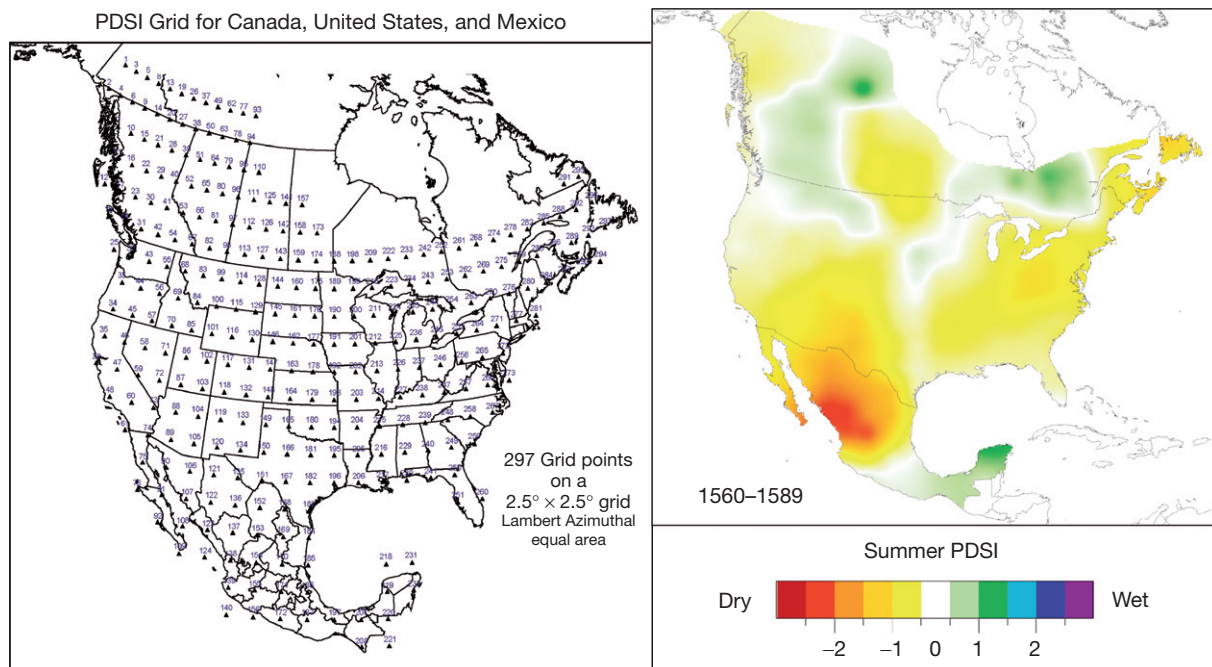


Figure 7 Coverage of the $2.5^\circ \times 2.5^\circ$ gridded PDSI dataset for North America based on tree-ring data. The diagram on the right shows the average reconstructed PDSI for the major drought between 1560 and 1589 (David Stahle, personal communication; see Cook et al., 2004).

Noncontiguous sets of tree-ring data have been assembled to examine spatial patterns and teleconnections associated with large-scale atmospheric phenomena. This is the basis for the field of 'synoptic dendroclimatology,' which examines climate variability through reconstructing changes in the atmospheric circulation patterns that control temperature and precipitation fields (Hirschboeck et al., 1996) and attempts to reconstruct climate fields from limited proxy data (Evans et al., 2001). The strong relationships between winter precipitation, El Niño Southern Oscillation (ENSO), and ring width in the SW United States and adjacent Mexico first led to the recognition of a strong ENSO signal in many of these chronologies. Combination of these data with coral, ice-core, and tree-ring data (including a teak chronology from Indonesia) allowed the reconstruction of

the Southern Oscillation Index between 1707 and 1977 (Stahle et al., 1998). A subsequent tree-ring-based reconstruction of the Niño3 Index, calibrated on sea-surface temperatures, extends back to 1408 (D'Arrigo et al., 2005) and long precipitation-sensitive chronologies from *Austrocedrus* in Chile and Argentina allow the reconstruction of ENSO patterns back over 1500 years in this region (J.A. Boninsegna, 2003, Personal communication). Chronologies from *Polylepis tarapacana* (queñoa) growing at over 4500 m on the Bolivian Altiplano also show a strong correlation with ENSO (Christie et al., 2009; Villalba et al., 2011). Similar studies have reconstructed the Pacific Decadal Oscillation (PDO, Gedalof et al., 2002), North Atlantic Oscillation (NAO; Cook et al., 2002a), and other climate indices from tree-ring data. These large-scale investigations have also

been important in the exploration of teleconnection patterns over time. For example, [St. George et al. \(2010\)](#) have been able to demonstrate the relationships between drought on the Canadian Prairies and several circulatory patterns (ENSO, PDO, AMO, etc.). At a larger scale, [Villalba et al. \(2001, 2011\)](#) demonstrate that the phase relations between tree-ring series from Patagonia and Coastal Alaska show similar patterns at different frequencies over the last 300 years.

Climate-Related Reconstructions

As available databases and investigations diversify, dendroclimatological studies are expanding into several related fields. Temperature-sensitive chronologies have been used to reconstruct sea-surface temperatures (e.g., [D'Arrigo et al., 2005, 2006c](#)). An obvious extension of work on moisture-sensitive sites has been the reconstruction of streamflow. In an early study, [Stockton \(1975\)](#) used 17 chronologies in the Colorado River basin to develop a long streamflow reconstruction which demonstrated that the 1564–1961 mean flow was approximately $2.5 \times 10^9 \text{ m}^3$ less than the 1896–1960 mean and considerably less than the $22 \times 10^9 \text{ m}^3$ base period on which allocation of water rights was based in the 1922 Colorado Compact between the United States and Mexico. In a warming world, more attention is being paid to water supply and streamflow reconstructions and these types of calculations will become increasingly important in watershed planning (e.g., [Meko and Woodhouse, 2005, 2011](#)). [Watson and Luckman \(2004\)](#) used both temperature and precipitation-sensitive tree-ring series to reconstruct the mass balance of Peyto Glacier in Canada over the last 322 years ([Figure 8](#)). Such studies are important in determining the relative roles of precipitation and temperature variability in the past and future behavior of these glaciers in response to climate change. In several mountain areas, summer streamflows are mainly fed by runoff from winter snowpack, and [Pederson et al. \(2011\)](#) have used snowpack-sensitive high-elevation chronologies and lower elevation moisture-sensitive sites to reconstruct snowpack for several large drainage basins in the western Americas.

Tree-ring data have been extensively used to reconstruct 'event' chronologies that show the temporal and spatial

distribution of significant geomorphic or ecological events such as fire, insect outbreaks, major floods, etc. The availability of regional or proximal long temperature and precipitation reconstructions allows examination of the potential climatic controls of such events. Dendrochronology has been widely used to reconstruct forest fire history for individual sites, but studies combining networks of such reconstructions with proxy climate records allow investigation of the climatic controls of these events at interannual, decadal, and centennial timescales ([Figure 9](#)). Synchronous landscape responses to climatic events at continental scales may result in significant ecological changes, which have been characterized as 'ecologically-effective climate change' in the western United States (see [Swetnam and Brown, 2011](#)).

Investigations in the Tropics

One of major gaps in dendroclimatic studies concerns the tropics where few trees have annual rings. In these areas, annual ring development is mainly found in regions with marked seasonality in precipitation ([Worbes, 2002](#)). Recent work has extended chronologies into the subtropics at high elevation sites in the Americas ([Villalba et al., 2011](#)). Precipitation-sensitive chronologies up to 600 years long have been developed from *Polylepis tarapacana* on the Bolivian Altiplano between 18° and 22°S ([Argollo et al., 2004; Christie et al., 2009](#)) and chronologies have been developed from newly investigated tropical and subtropical species east of the Andes in Bolivia and Argentina ([Ferrero and Villalba, 2009; López and Villalba, 2011](#)). Recently, there have been major new collections in SE Asia which are beginning to yield excellent results ([Buckley et al., 2007; D'Arrigo et al., 2006b,c](#)) using teak (*Tectona grandis*) and *Fokienia hodginsii* (locally known as Po Mu) in Vietnam and Cambodia ([Buckley et al., 2010; Sano et al., 2009](#)). Long chronologies have also been developed from the tropics and subtropics in Mexico using both Douglas fir and Montezuma bald cypress (*Taxodium mucronatum*) ([Stahle et al., 2011](#)). There is also promise that isotope dendroclimatology can yield proxy

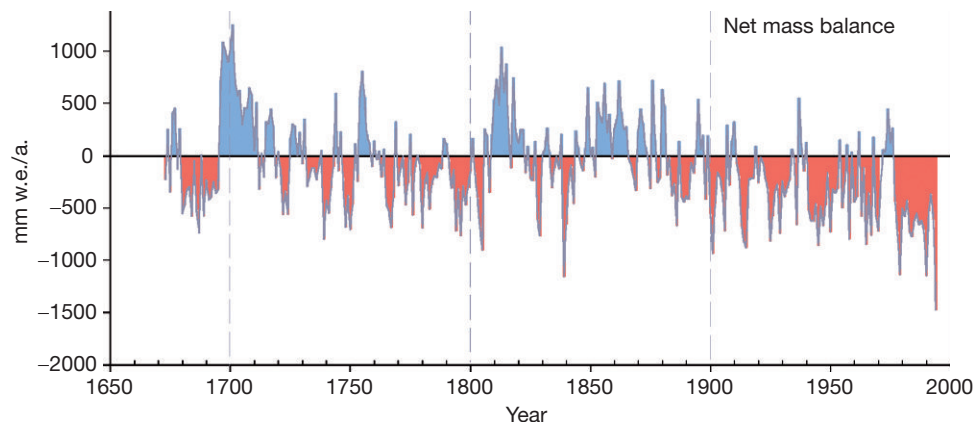


Figure 8 Reconstructed glacier mass balance for Peyto Glacier, Alberta, Canada, based on separate reconstructions of winter and summer balances using precipitation- and temperature-sensitive tree-ring chronologies. Reproduced from Watson E and Luckman BH (2004) Tree-ring estimates of Mass Balance at Peyto Glacier for the last three centuries. *Quaternary Research* 62: 9–18.

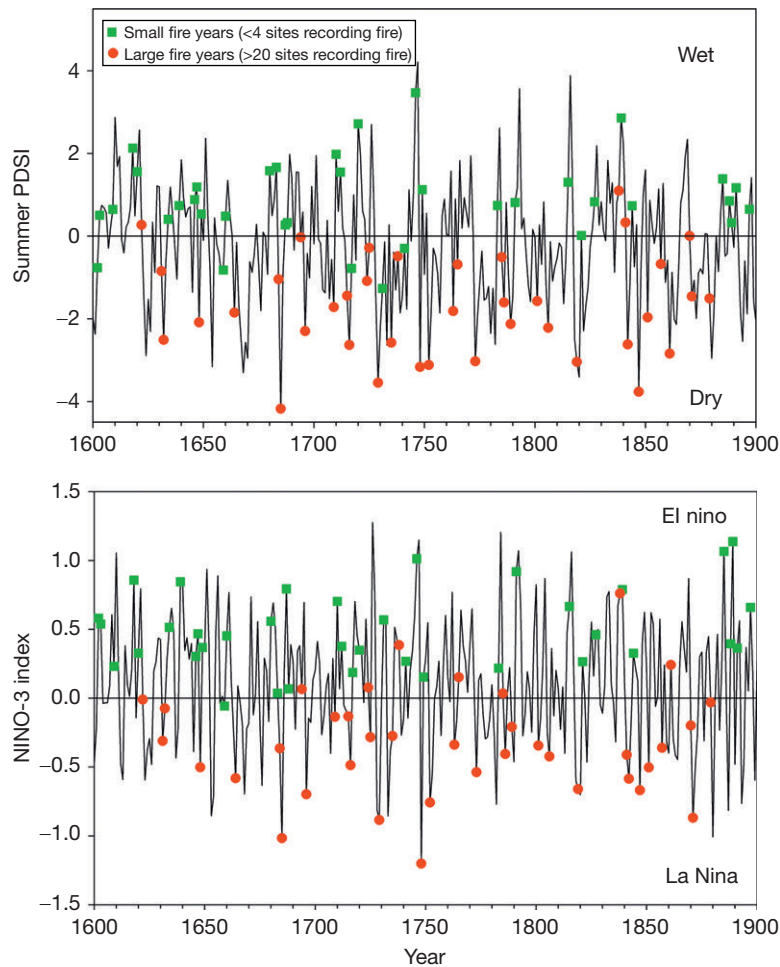


Figure 9 Relationship between regional fires in the southwestern United States and potential climate drivers. *Upper diagram*: tree-ring-reconstructed July–August PDSI (solid line, from Cook et al., 2004) with the largest and smallest regional fire years from a 120-site fire scar network. *Lower diagram*: The same set of fire years plotted with tree-ring-reconstructed Niño-3 index of sea surface temperatures (Cook, 2000) (Tom Swetnam, personal communication, after Swetnam and Brown, 2011, Fig. 3).

climate records from species that lack annual rings (Evans and Schrag, 2004; Gagen et al., 2011).

Changing Signals

Dendroclimatic reconstructions, like many other paleoclimatic techniques, are based on uniformitarian assumptions. However, several authors have noted a decline in the temperature-related signal in both ring width (e.g., Jacoby and D'Arrigo, 1995), and MXD series from high-latitude Northern Hemisphere sites (Briffa et al., 2004) and other studies report changes in signal relating to atmospheric loading effects (e.g., Wilson and Elling, 2004). A variety of causes have been suggested, including increased moisture sensitivity once summer temperatures pass a critical threshold (D'Arrigo et al., 2004), possible ozone-related effects (Briffa et al., 2004), or even chronology development techniques (Wilmking et al., 2005). Although not universal (e.g., Wilson et al., 2007), these effects need careful scrutiny and further study to resolve this critical issue which potentially challenges the validity of

the fundamental uniformitarian assumptions of dendroclimate reconstructions. The potential causes of divergence have been reviewed most recently by D'Arrigo et al. (2008), and Visser et al. (2010) provide a possible technique for identifying these problems. However, it is important to remember that the establishment of 'divergence' requires the demonstration of a change in signal rather than the absence of a temperature signal in the tree-ring record examined.

Concluding Remarks

Dendroclimatological studies are providing new, annually resolved, and precisely dated proxy climate records for large areas of the Earth's land surface and parts of the adjacent oceans. These allow the reconstruction of climate history, trends, and patterns at an expanding range of temporal and spatial scales. The increasing availability of long chronologies from major continental areas can address climate variability at decadal to millennial timescales while retaining annual resolution. The increasing density and distribution of chronologies

provide networks that examine the spatial variability and teleconnection patterns associated with major modes of global climate variability such as ENSO, PDO, NAO, and the Asian Monsoon. As these networks of annually resolved data are minimally at least double or triple the length of instrumental records (in many remote areas they may be the only records), they can be used to establish whether these patterns are time-stable or how they have responded to past changes in forcing. In conjunction with other proxy climate data, they provide the best records to benchmark and understand the natural modes of climate variability at decade-to-century scales over much of the Earth's surface. The growing availability of long individual or gridded temperature and precipitation reconstructions also allows exploration of the relationships of between critical episodes in human history and large-scale climate events. [Stahle and Dean \(2011\)](#) provide many examples of the human impact of drought in North America over the last millennium. More recently, [Büntgen et al. \(2011\)](#) have examined the relationships between significant events in European History and the temperature and precipitation signals derived from European oak and pine chronologies of the last 2500 years. PDSI and drought reconstructions have also shed light on possible causes for the collapse of the Khmer ([Buckley et al., 2010](#)), Mayan, Toltec, and Aztec civilizations ([Stahle et al., 2011](#)). Thus, whilst dendroclimatology plays a central role in documenting climate variability, it can also be used to understand the important role these changes play in geomorphic, ecological, and human histories of the last millennium.

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See also: [Dating Techniques](#). [Paleoclimate: Introduction](#). [Paleoclimate Modeling: Paleo-ENSO](#). [Paleoclimate Reconstruction: Approaches](#). [Radiocarbon Dating: \$^{14}\text{C}\$ of Plant Macrofossils](#).

References

- Acuna-Soto R, Therrell MD, Gomex Chavez S, and Cleaveland MK (2005) Drought, epidemic disease, and the fall of classic period cultures in MesoAmerica (AD 750–950). Hemorrhagic fevers as a cause of massive population loss. *Medical Hypotheses* 65: 405–409.
- Argollo J, Soliz C, and Villalba R (2004) Potencialidad dendrocronológica de *Polylepis tarapacana* en los Andes Centrales de Bolivia. *Ecología en Bolivia* 39: 5–24.
- Baillie MGL (1996) Extreme environmental events and the linking of tree-ring and ice-core records. In: Dean J, Meko DM, and Swetnam TW (eds.) *Tree Rings, Environment and Humanity*, pp. 703–711. Tucson, AZ: Radiocarbon.
- Boninsegna JA, Argollo J, Aravena JC, et al. (2009) South American tree rings as climate proxy records. *Palaeogeography Palaeoclimatology Palaeoecology* 281: 210–228. Special Issue on Regional high-resolution multiproxy climate reconstruction for South America: State of the art and perspectives.
- Briffa KR (2000) Annual variability in the Holocene: Interpreting the message of ancient trees. *Quaternary Science Reviews* 19: 87–105.
- Briffa KR and Jones PD (1990) Estimating the chronology signal strength. In: Cook ER and Kairiukstis LA (eds.) *Methods of Dendrochronology: Applications in the Environmental Sciences*, pp. 137–152. Dordrecht The Netherlands: Kluwer.
- Briffa KR and Matthews JA (2002) Analysis of dendrochronological variability and associated natural climates in Eurasia (Advance-10K Special Issue). *The Holocene* 12: 639–794.
- Briffa KR and Melvin T (2011) A closer look at regional curve standardization of tree-ring records: Justification of the need, a warning of some pitfalls, and suggested improvements in its application. In: Hughes MK, Swetnam TW, and Diaz HF (eds.) *Dendroclimatology, Progress and Prospects*, pp. 113–146. Dordrecht, The Netherlands: Springer.
- Briffa KR, Osborn TJ, and Schweingruber FH (2004) Large-scale temperature inferences from tree rings: A review. *Global and Planetary Change* 40: 11–26.
- Briffa KR, Osborn TJ, Schweingruber FH, et al. (2001) Low-frequency temperature variations from a northern tree-ring network. *Journal of Geophysical Research* 106: 2929–2941.
- Briffa KR, Osborn TJ, Schweingruber FH, et al. (2002) Tree-ring width and density data around the Northern Hemisphere: Part 2, Spatio-temporal variability and associated climate patterns. *The Holocene* 12: 759–789.
- Buckley BM, Anchukaitis KJ, Penny D, et al. (2010) Climate as a contributing factor in the demise of Angkor, Cambodia. *Proceedings of the National Academy of Sciences of the United States of America* 107: 6748–6752.
- Buckley BM, Palakit K, Duangsathaporn K, Sanguantham P, and Prasomsin P (2007) Decadal scale droughts over northwestern Thailand over the past 448 years: Links to the tropical Pacific and Indian Ocean sectors. *Climate Dynamics* 29: 63–71.
- Büntgen U, Esper J, Frank DC, Nicolussi K, and Schmidhalter M (2005) A 1052-year tree-ring proxy for Alpine summer temperatures. *Climate Dynamics* 25: 141–153.
- Büntgen U, Tegel W, Nicolussi K, et al. (2011) 2500 years of European climate variability and human susceptibility. *Science* 331: 578–581. <http://dx.doi.org/10.1126/science.1197175>.
- Campbell R, McCarroll D, Loader NJ, Grudd H, Robertson I, and Jalkanen R (2007) Blue intensity in *Pinus sylvestris* tree-rings: Developing a new palaeoclimate proxy. *The Holocene* 17: 821–828.
- Christie DA, Boninsegna JA, Cleaveland MK, et al. (2010) Aridity changes in the Temperate-Mediterranean transition of the Andes since AD 1346 reconstructed from tree-rings. *Climate Dynamics* 36: 1505–1521. <http://dx.doi.org/10.1007/s00382-009-0723-4>.
- Christie DA, Lara A, Barichivich J, et al. (2009) El Niño–Southern Oscillation signal in the world's highest-elevation tree-ring chronologies from the Altiplano, Central Andes. *Palaeogeography Palaeoclimatology Palaeoecology* 281: 309–319.
- Cook ER (1990) A conceptual linear aggregate model for tree rings. In: Cook ER and Kairiukstis LA (eds.) *Methods of Dendrochronology: Applications in the Environmental Sciences*, pp. 98–103. Dordrecht, The Netherlands: Kluwer.
- Cook ER (2000) NINO3 index reconstruction. International Tree-Ring Data Bank. IGBP PAGES/World Data Center-A for Paleoclimatology data contr.ser #2000-052.
- Cook ER, Anchukaitis KJ, Buckley BM, et al. (2010) Asian monsoon failure and megadrought during the last millennium. *Science* 328: 486–489.
- Cook ER and Briffa KR (1989) A comparison of some tree-ring standardization methods. In: Cook ER and Kairiukstis LA (eds.) *Methods of Dendrochronology: Applications in the Environmental Sciences*, pp. 153–162. Dordrecht, The Netherlands: Kluwer.
- Cook ER, Briffa KR, Meko DM, Graybill DM, and Funkhouser G (1995) The “segment length curse” in long tree-ring chronology development for paleoclimate studies. *The Holocene* 5: 229–237.
- Cook ER, Buckley BM, D'Arrigo RD, and Peterson MJ (2000) Warm-season temperatures since 1600 B.C. reconstructed from Tasmanian tree rings and their relationship to large-scale sea surface temperature anomalies. *Climate Dynamics* 16: 79–91.
- Cook ER, D'Arrigo RD, and Mann ME (2002a) A well-verified, multiproxy reconstruction of the winter north Atlantic oscillation index since A.D. 1400. *Journal of Climate* 15: 1754–1764.
- Cook ER, Palmer JG, and D'Arrigo RD (2002b) Evidence for a Medieval Warm Period in a 1100 year tree-ring reconstruction of past Austral summer temperatures in New Zealand. *Geophysical Research Letters* 29. <http://dx.doi.org/10.1029/2001GL014580>.
- Cook ER and Kairiukstis LA (eds.) (1990) *Methods of Dendrochronology: Applications in the Environmental Sciences*. Dordrecht, The Netherlands: Kluwer.
- Cook ER, Meko DM, Stahle DW, and Cleaveland MK (1999) Drought reconstructions for the continental United States. *Journal of Climate* 12: 1145–1162.
- Cook ER and Pederson N (2011) Uncertainty, emergence, and statistics in dendrochronology. In: Hughes MK, Swetnam TW, and Diaz HF (eds.) *Dendroclimatology, Progress and Prospects*, pp. 77–112. Dordrecht, The Netherlands: Springer.
- Cook ER, Woodhouse CA, Eakin CM, Meko DM, and Stahle DW (2004) Long-term aridity changes in the western United States. *Science* 306: 1015–1018.

- D'Arrigo RD, Cook ER, Wilson RJ, Allan R, and Mann ME (2005) On the variability of ENSO over the past six centuries. *Geophysical Research Letters* 32: L03711. <http://dx.doi.org/10.1029/2004GL022055>.
- D'Arrigo RD and Jacoby GC (1993) Secular trends in high northern latitude temperature reconstructions based on tree rings. *Climatic Change* 25: 163–177.
- D'Arrigo RD, Jacoby GC, Frank D, et al. (2001) 1738 years of Mongolian temperature variability inferred from a tree-ring width chronology of Siberian pine. *Geophysical Research Letters* 28: 543–546.
- D'Arrigo RD, Kaufmann RK, and Davi N (2004) Thresholds for warming-induced growth decline at elevational treeline in Yukon Territory, Canada. *Global Biogeochemical Cycles* 18: GB3021. <http://dx.doi.org/10.1029/2004GB002249>.
- D'Arrigo RD, Wilson R, and Jacoby G (2006a) On the long-term context for late 20th century warming. *Journal of Geophysical Research* 111: D03103/1–D03103/12.
- D'Arrigo RD, Wilson R, Palmer J, et al. (2006b) Monsoon drought over Java, Indonesia during the past two centuries. *Geophysical Research Letters* 33: L04709. <http://dx.doi.org/10.1029/2005GL025465>.
- D'Arrigo RD, Wilson R, Palmer J, et al. (2006c) Reconstructed Indonesian warm pool SSTs from tree rings and corals: Linkages to Asian monsoon drought and ENSO. *Paleoceanography* 21: PA3005. <http://dx.doi.org/10.1029/2005PA001256>.
- D'Arrigo RD, Liepert B, and Cherubini P (2008) On the 'divergence problem' in northern forests: A review of the tree-ring evidence and possible causes. *Global and Planetary Change* 60: 289–305.
- Esper J, Cook ER, Krusic PJ, Peters K, and Schweingruber FH (2003a) Tests of the RCS method for preserving low-frequency variability in long tree-ring chronologies. *Tree Ring Research* 39: 81–98.
- Esper J, Shiyatov SG, Mazepa VS, et al. (2003b) Temperature-sensitive Tien Shan tree-ring chronologies show multi-centennial growth trends. *Climate Dynamics* 21: 699–706.
- Esper J, Cook ER, and Schweingruber FH (2002) Low-frequency signals in long tree-ring chronologies for reconstructing past temperature variability. *Science* 295: 2250–2253.
- Evans MN, Kaplan A, Cane MA, and Villalba R (2001) Globality and optimality in climate field reconstructions from proxy data. In: Markgraf V (ed.) *Interhemispheric Climate Linkages*, pp. 53–72. New York: Academic Press.
- Evans MN and Schrag DP (2004) A stable isotope-based approach to tropical dendroclimatology. *Geochimica et Cosmochimica Acta* 2004(68): 3295–3305.
- Ferrero ME and Villalba R (2009) Potential of *Schinopsis lorentzii* for dendrochronological studies in subtropical dry Chaco forests of South America. *Trees* 2009. <http://dx.doi.org/10.1007/s00468-009-0369-1>.
- Fritts HC (1976) *Tree Rings and Climate*. New York: Academic Press.
- Fritts HC (1991) *Reconstructing Large-Scale Climatic Patterns from Tree-Ring Data*. Tucson, AZ: University of Arizona Press.
- Fye FK, Stahle DW, and Cook ER (2003) Paleoclimate analogs to 20th century moisture regimes across the US. *Bulletin of the American Meteorological Society* 84: 901–909.
- Gagen M, McCarroll D, Loader NJ, and Robertson I (2011) Stable isotopes in dendroclimatology: Moving beyond 'potential'. In: Hughes MK, Swetnam TW, and Diaz HF (eds.) *Dendroclimatology, Progress and Prospects*, pp. 147–172. Dordrecht, The Netherlands: Springer.
- Gedalof Z, Mantua NJ, and Peterson DL (2002) A multi-century perspective of variability in the Pacific Decadal Oscillation: New insights from tree rings and coral. *Geophysical Research Letters* 29: 2204. <http://dx.doi.org/10.1029/2002GL015824>.
- Gordon GA (1982) Verification of dendroclimatic reconstructions. In: Hughes MK, Kelly PM, Pilcher JP, and LaMarche VC Jr. (eds.) *Climate from Tree Rings*, pp. 58–61. Cambridge: Cambridge University Press.
- Graybill DA (1982) Chronology development and analysis. In: Hughes MK, Kelly PM, Pilcher JP, and LaMarche VC Jr. (eds.) *Climate from Tree Rings*, pp. 21–28. Cambridge: Cambridge University Press.
- Grissino-Meyer, H. (n.d.) The Ultimate Tree-Ring Web Site. <http://web.utk.edu>.
- Grudd H, Briffa KR, Gunnarson BE, and Linderholm HW (2000) Swedish tree-rings provide new evidence in support of a major widespread environmental disruption in 1628 B.C. *Geophysical Research Letters* 27: 2957–2960.
- Hirschboeck K, Ni F, Wood ML, and Woodhouse CA (1996) Synoptic dendroclimatology: Overview and outlook. In: Dean J, Meko DM, and Swetnam TW (eds.) *Tree Rings, Environment and Humanity*, pp. 205–223. Tucson, AZ: Radiocarbon.
- Hughes MK (2002) Dendrochronology in climatology. The state of the art. *Dendrochronologia* 20: 95–116.
- Hughes MK (2011) Dendroclimatology in high resolution paleoclimatology. In: Hughes MK, Swetnam TW, and Diaz HF (eds.) *Dendroclimatology, Progress and Prospects*, pp. 17–34. Dordrecht, The Netherlands: Springer.
- Hughes MK, Diaz HF, and Swetnam TW (2011a) Tree rings and climate: Sharpening the focus. In: Hughes MK, Swetnam TW, and Diaz HF (eds.) *Dendroclimatology, Progress and Prospects*, pp. 331–353. Dordrecht, The Netherlands: Springer.
- Hughes MK, Swetnam TW, and Diaz HF (eds.) (2011b) *Dendroclimatology: Progress and Prospects Developments in Paleoenvironmental Research*, vol. 11, p. 365. Dordrecht, The Netherlands: Springer.
- Hughes MK and Graumlich LJ (1996) Multimillennial dendroclimatic studies from the western United States. In: Jones PD, Bradley RS, and Jouzel J (eds.) *Climatic Variations and Forcing Mechanisms of the Last 2000 Years. NATO ASI Series 1; Global Environmental Change*, vol. 41, pp. 109–124. New York: Springer.
- Jacoby GC and Cook ER (1981) Past temperature variations inferred from a 400-year tree-ring chronology from Yukon Territory, Canada. *Arctic and Alpine Research* 13: 409–418.
- Jacoby GC and D'Arrigo RD (1995) Tree ring width and density evidence of climatic and potentially forest change in Alaska. *Global Biogeochemical Cycles* 9: 227–234.
- Jones PD and Thompson R (2003) Instrumental records. In: MacKay A, Battarbee R, Battarbee R, Birks J, and Oldfield F (eds.) *Global Change in the Holocene*, pp. 140–158. London: Hodder Arnold.
- LaMarche VC Jr. and Hirschboeck KK (1984) Frost rings in trees as records of major volcanic eruptions. *Nature* 307: 121–128.
- Lara A and Villalba R (1993) A 3,620 year temperature reconstruction from *Fitzroya cupressoides* in Southern South America. *Science* 260: 1104–1106.
- Lara A, Villalba R, Aravena JC, Wolodarsky A, and Neira E (2000) Desarrollo de una red de cronologías de *Fitzroya cupressoides* (Alerce) para Chile y Argentina. In: Roig F (ed.) *Dendrocronología en América Latina*, pp. 217–244. Universidad Nacional de Cuyo, Mendoza, Argentina: EDIUNC.
- Le Quesne C, Stahle DW, Cleaveland MK, Therrell MD, Aravena JC, and Barichivich J (2006) Ancient cipres tree-ring chronologies used to reconstruct central Chile precipitation variability from A.D. 1200–2000. *Journal of Climate* 19: 5731–5744.
- López L and Villalba R (2011) Climate influences on the radial growth of *Centrolobium microchaete*, a valuable timber species from the tropical dry forests in Bolivia. *Biotropica* 43: 41–49.
- Luckman BH and Wilson RJS (2005) Summer temperature in the Canadian Rockies during the last millennium – A revised record. *Climate Dynamics* 24: 131–144.
- Mann ME, Bradley RS, and Hughes MK (1999) Northern hemisphere temperatures during the past millennium: Inferences, uncertainties and limitations. *Geophysical Research Letters* 26: 759–762.
- McCarroll D, Tuovinen M, Campbell R, et al. (2010) A critical evaluation of multi-proxy dendroclimatology in northern Finland. *Journal of Quaternary Science* 25: 1–8.
- Meko DM and Baisan CH (2001) Pilot study of latewood-width of conifers as an indicator of variability of summer rainfall in the North American Monsoon region. *International Journal of Climatology* 21: 697–708.
- Meko DM and Woodhouse CA (2005) Tree-ring footprint of joint hydrologic drought in Sacramento and Upper Colorado River Basins, western USA. *Journal of Hydrology* 308: 196–213.
- Meko DM and Woodhouse CA (2011) Application of streamflow reconstruction to water resources management. In: Hughes MK, Swetnam TW, and Diaz HF (eds.) *Dendroclimatology, Progress and Prospects*, pp. 231–262. Dordrecht, The Netherlands: Springer.
- Neukom R, Luterbacher J, Villalba R, et al. (2010a) Multiproxy summer and winter surface air temperature field reconstructions for southern South America covering the past centuries. *Climate Dynamics* 37: 35–51. <http://dx.doi.org/10.1007/s00382-010-0793-3>.
- Neukom R, Luterbacher J, Villalba R, et al. (2010b) Multi-centennial summer and winter precipitation variability in southern South America. *Geophysical Research Letters* 37: L14708. <http://dx.doi.org/10.1029/2010GL043680>.
- Parker ML and Henoch WES (1971) The use of Engelmann spruce latewood density for dendrochronological analysis. *Canadian Journal of Forest Research* 1: 90–98.
- Pederson GT, Gray ST, Woodhouse CA, et al. (2011) Evidence from tree rings highlights the unusual nature of recent snowpack declines. *Science* 333: 332–335. <http://dx.doi.org/10.1126/science.1201570>.
- Polge H (1966) Etablissement des Courbes de Variation de la Densité du Bois par Exploration Densitométrique de Radiographies d'Échantillons Prélevés à la Tarière sur des Arbres Vivants. *Annales des Sciences Forestières* 23: 1–206.
- Roig FA, Palmer JG, Cook ER, et al. (2010) Longues séries dendrochronologiques dans l'hémisphère Sud. In: Payette S and Filion L (eds.) *La Dendroécologie: Principes, méthodes et applications*, pp. 647–682. Québec: Presses de l'Université Laval.
- Rutherford S, Mann ME, Osborn TJ, et al. (2005) Proxy-based northern hemisphere surface temperature reconstructions: Sensitivity to method, predictor network, target season, and target domain. *Journal of Climate* 18: 2308–2329.
- Sano M, Buckley BM, and Sweda T (2009) Tree-ring based hydroclimate reconstruction over northern Vietnam from *Fokienia hodginsii*: Eighteenth century mega-drought and tropical Pacific influence. *Climate Dynamics* 33: 331–340.
- Schweingruber FH (1988) *Tree Rings*. Dordrecht, The Netherlands: D. Reidel.

- Schweingruber FH and Briffa KR (1996) Tree-ring density networks for climate reconstruction. In: Jones PD, Bradley RS, and Jouzel J (eds.) *Climate Variations and Forcing Mechanisms of the Last 2000 Years*, pp. 43–66. Berlin: Springer.
- Schweingruber FH, Fritts HC, Braeker OU, Drew LG, and Schaer E (1978) The X-ray technique as applied to dendroclimatology. *Tree-Ring Bulletin* 38: 61–91.
- St. George S, Meko DM, and Cook ER (2010) The seasonality of precipitation signals embedded within the North American Drought Atlas. *The Holocene* 20: 983–988.
- St. George S, Meko DM, Girardin MP, et al. (2009) The tree-ring record of drought on the Canadian prairies. *Journal of Climate* 22: 689–710.
- Stahle DW, D'Arrigo RD, Krusic PJ, et al. (1998) Experimental dendroclimatic reconstruction of the southern oscillation. *Bulletin of the American Meteorological Society* 79: 2137–2152.
- Stahle DW and Dean JS (2011) North American tree-rings, climatic extremes, and social disasters. In: Hughes MK, Swetnam TW, and Diaz HF (eds.) *Dendroclimatology, Progress and Prospects*, pp. 297–327. Dordrecht, The Netherlands: Springer.
- Stahle DW, Villanueva J, and Cleaveland MK (2000) Recent tree-ring research in Mexico. In: Roig F (ed.) *Dendrochronología en América Latina*, pp. 285–306. Universidad Nacional de Cuyo, Mendoza, Argentina: EDIUNC.
- Stahle DW, Villanueva-Diaz J, Burnette DJ, et al. (2011) Major Mesoamerican droughts of the past millennium. *Geophysical Research Letters* 38: L05703. <http://dx.doi.org/10.1029/2010GL046472>.
- Stockton CW (1975) *Long Term Streamflow Records Reconstructed from Tree Rings*. Paper 5, Laboratory of Tree-Ring Research. University of Arizona Press.
- Swetnam TW and Brown PM (2011) Climatic inferences from dendroecological reconstructions. In: Hughes MK, Swetnam TW, and Diaz HF (eds.) *Dendroclimatology, Progress and Prospects*, pp. 263–296. Dordrecht, The Netherlands: Springer.
- Therrell MD, Stahle DW, Cleaveland MK, Villanueva-Diaz J, and Cornejo Oviedo E (2006) Tree-ring reconstructed maize yield in Central Mexico. *Climatic Change* 74: 493–504.
- Vaganov EA, Anchukaitis KJ, and Evans MN (2011) How well understood are the processes that create dendroclimate records? A mechanistic model of climatic control on conifer tree-ring growth dynamics. In: Hughes MK, Swetnam TW, and Diaz HF (eds.) *Dendroclimatology, Progress and Prospects*, pp. 37–76. Dordrecht: Springer.
- Villalba R, D'Arrigo RD, Cook ER, Jacoby GC, and Wiles G (2001) Decadal-scale climatic variability along the extratropical western coast of the Americas: Evidence from tree-ring records. In: Markgraf V (ed.) *Interhemispheric Climate Linkages*, pp. 155–172. New York: Academic Press.
- Villalba R, Luckman BH, Boninsegna JA, et al. (2011) Dendroclimatology from regional to continental scales: Understanding regional processes to reconstruct large-scale climatic variations across the Western Americas. In: Hughes MK, Swetnam TW, and Diaz HF (eds.) *Dendroclimatology, Progress and Prospects*, pp. 175–230. Dordrecht, The Netherlands: Springer.
- Visser H, Buntgen U, D'Arrigo R, and Petersen AC (2010) Detecting instabilities in tree-ring proxy calibration. *Climate of the Past* 6: 367–377.
- Watson E and Luckman BH (2004) Tree-ring estimates of Mass Balance at Peyto Glacier for the last three centuries. *Quaternary Research* 62: 9–18.
- Wigley TML, Briffa KR, and Jones PD (1984) On the average value of correlated time series with applications in dendroclimatology and hydrometeorology. *Journal of Climate and Applied Meteorology* 23: 201–213.
- Wilkinson M, D'Arrigo RD, Jacoby GC, and Juday GP (2005) Increased temperature sensitivity and divergent growth trends in circumpolar boreal forests. *Geophysical Research Letters* 32: L15715. <http://dx.doi.org/10.1029/2005GL023331>.
- Wilson RJS, D'Arrigo RD, Buckley B, et al. (2007) A matter of divergence: Tracking recent warming at hemispheric scales using tree-ring data. *Journal of Geophysical Research* 112: D17103. <http://dx.doi.org/10.1029/2006JD008318>.
- Wilson RJS and Elling W (2004) Temporal instabilities of tree-growth/climate response in the Lower Bavarian Forest Region: Implications for dendroclimatic reconstruction. *Trees: Structure and Function* 18: 19–28.
- Worbes M (2002) One hundred years of tree-ring research in the tropics – A brief history and an outline of future challenges. *Dendrochronologia* 20: 217–231.

Diatom Introduction

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The Biology of Diatoms

The word 'diatom' derives from the Greek, meaning cut in two. Diatoms (Bacillariophyta) are unicellular, eukaryotic (complex cells where the genetic material is found in membrane-bound nuclei) microscopic organisms belonging to the Kingdom Protista. They are pigmented and usually appear golden brown in color; they are photosynthetic although a few species are facultative or obligate heterotrophs (i.e., feeding on other organisms). The characteristic feature of the diatom is its cell wall, which is almost always heavily impregnated with silica ($\text{SiO}_2 \cdot n\text{H}_2\text{O}$).

Globally, diatoms are very important as members of the planktonic food chain for large parts of the planet. They play a key role in the global carbon cycle. Diatom photosynthesis in the ocean yields as much as 40% of the organic carbon produced each year in the sea, and their role in global carbon cycling is predicted to be comparable to that of all the terrestrial rain forests combined, fixing about 20% of the carbon each year. Over geological time, diatoms may have influenced global climate by changing the flux of atmospheric carbon dioxide into the oceans.

Diatoms reflect a fundamentally different evolutionary history from the higher plants that dominate photosynthesis on land. They share biogeochemical features of both plants and animals. The marine centric diatom *Thalassiosira pseudonana* was chosen as the first eukaryotic marine phytoplankton species for whole-genome sequencing. The genome is relatively small at 34 mega base pairs (Armbrust et al., 2004). There is a well-documented fossil record of the diatoms from the middle Cretaceous with some records from the early Jurassic gap. There are a large number of Eocene/Miocene diatom fossil deposits situated around the world, which are mined commercially as diatomite; Flower (this volume) deals with the Quaternary diatomites. Further details of diatom biology are given in Round et al. (1990) and van den Hoek et al. (1995).

The Structure of the Diatom Frustule

The silica cell wall of diatoms is box-like in structure, always made of two halves, called valves. The upper one (the epivalve or epitheca) is slightly larger than the lower valve (hypovalve or hypotheca; Figure 1) and they are linked together with girdle bands. The valves are intricately sculptured and when organic material is removed by cleaning, the frustule splits into two. Most Quaternary applications are based on the observation of these fragments.

Diatoms are the most species-rich group of algae and there may be as many as 200 000 species. Although traditionally most diatom systematic studies have been based on valve features observed under the light microscope, recent developments through electron microscopy, the examination of living material, and molecular techniques have resulted in many nomenclatural changes, and consequently the taxonomy can be confusing.

Round et al. (1990) propose three classes of diatom: the Centrics, (Coscinodiscophyceae), pennates without raphe (Fragilariophyceae), and pennates with a raphe (Bacillariophyceae). These are shown in schematic form in Figure 1.

Features of the Valve

Most diatoms contain areolae (Figure 1) that appear to be simple perforations through the cell wall, although internally they are occluded by a finely perforated velum. These areolae are usually arranged in rows of striae. The characteristics of these are used extensively for identification purposes, for example, the spacing between striae (i.e., the striae density) and the pitch of the striae. The raphe is one of the most important features used to separate suborders or classes. It is an elongated fissure or pair of fissures through the valve wall. The naviculoid raphe (Figure 1(c)) runs along the central axis of the valve and is divided in two, separated by the central nodule.

Diatom shape varies, usually being round or boat shaped, but is sometimes triangular, square, or elliptical. The cell wall is highly patterned with an array of pores, ribs, spines, and processes that can be used for identification purposes. Their spectacular forms have been appreciated since the invention of the microscope and have formed the subjects of many wonderful illustrations (Figure 2).

Cells are solitary (Figure 3) or united into colonies (Figure 4) which may be linked by siliceous structures such as spines or bipolar elevations (e.g., *Eucampia*) or by mucilage (e.g., in tubes in genera such as *Nitzschia*).

Motility is limited to those pennate species that have a raphe, and occurs through secretion of mucopolysaccharide filaments which become attached to the solid substrate enabling the diatom to 'ooze' along a surface. A unique way of moving is displayed by the pennate diatom *Bacillaria paradoxa* (= *Bacillaria paxillifera*); this species forms small colonies in which individual cells slide back and forth over one another so the shape of the colony can change from a square to a linear form and back again.

The uptake and deposition of silica is one of the characteristic features of diatoms. This leads to rigidity of the cell walls and preservation as fossils. The characteristics of the silica cell wall are also used in their taxonomy. Leng and Barker (this volume) give a more detailed account of the uptake of silica and Tyler (this volume) discusses the use of isotopes of silica in the Quaternary.

Lifecycles and Life Form

Diatoms are unusual in that vegetative cell division usually results in the size of the population becoming successively smaller and smaller. This is because the daughter cell is formed within the confines of the rigid parent cell. In some species, this

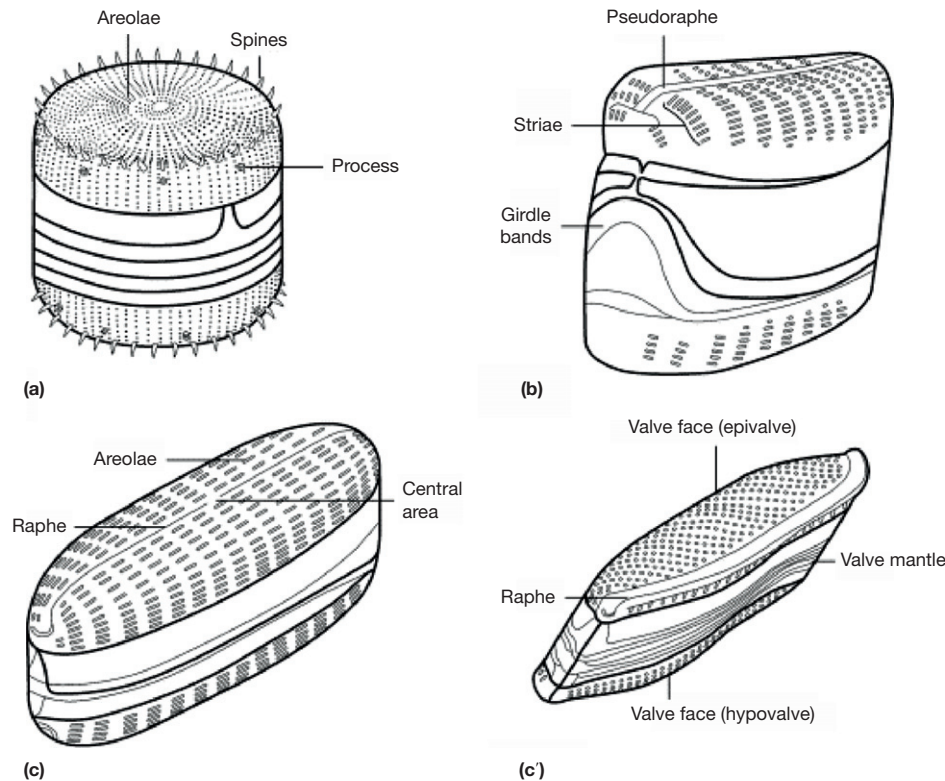


Figure 1 A schematic diagram illustrating the three main classes of diatom: (a) a typical centric form Coccinodiscophyceae, showing radial symmetry (e.g., *Stephanodiscus*) and pattern of areolae radiating from the center of the valve; (b) a typical pennate diatom without a raphe Fragilariophyceae (e.g., *Staurosira*) with areolae are usually arranged in rows or striae, in this case striae are parallel; (c) a typical pennate diatom with a raphe Bacillariophyceae (e.g., *Pinnularia*) with a naviculoid raphe running along the central axis of the valve which is divided into two separated by the central area; (c') a pennate diatom Bacillariophyceae (e.g., *Nitzschia*) with an eccentric raphe. Adapted from Round FE, Crawford RM, and Mann DG (1990) *The Diatoms: Biology and Morphology of the Genera*. Cambridge: Cambridge University Press, with kind permission from Cambridge University Press.

can, over many generations, result in a considerable reduction in size which can lead to problems in their identification and classification as the full-size ranges of most species are unknown. In centric species, the main features of the valve are unaltered, but in pennate species, the width often decreases less than the length (see Figure 5). Although size reduction is often ignored by Quaternary scientists, there are potential applications (e.g., Cortese and Gersonde, 2007; Simola, this volume).

Size is restored after sexual reproduction, and diatoms show a range of different types of sex. The Centrics are oogamous (i.e., they have eggs and flagellate sperm) while the pennates have morphologically similar gametes, which are nonflagellate. In all cases, copulation takes place between the two haploid gametes producing a diploid zygote, the auxospore. The auxospore then swells and when it has enlarged fully, a new vegetative cell is formed within it. The complete lifecycle has rarely been observed in natural populations, but in some species at least, it seems to take a considerable time, for example, at least 6 years for *Nitzschia sigmaidea*, and 7–8 years (based on observations of specimens from varved sediments) for *Tabelaria fenestrata* (Mann, 1988).

Diatom resting spores are dormant cells that are produced as a response to unfavorable conditions, mainly by centric diatoms. The spores contain stored energy in the form of photosynthetic products and have tough, thickened cell walls. They sink to the bottom of the water column and can return

to normal cell functioning if favorable conditions return. Some are morphologically distinct (e.g., *Chaetoceros* spp.) while others are not (e.g., *Aulocoseira subarctica*). In *Chaetoceros*, resting spore formation is associated with nutrient exhaustion in the water column, after mass multiplication of vegetative cells following an upwelling event. *Chaetoceros* resting spores, therefore, provide a record of paleobloom events in the ocean (Pike and Stickley, this volume). In continental saline waters, the very heavily silicified resting spores are often all that remain of a diatom assemblage after dissolution.

Diatom Habitats

Diatoms are distributed in all waters except the hottest and most hypersaline and most samples obtained from aquatic habitats contain some diatom cells.

Marine Habitats

Diatoms occur in all oceans from the poles to the tropics, but are most abundant in polar-to-temperate regions. These are mostly planktonic species. True planktonic species spend all of their lifecycle in open water, depending primarily on water currents for their buoyancy. Large blooms sometimes descend the water column *en masse* as mats (see Pike and Stickley, this

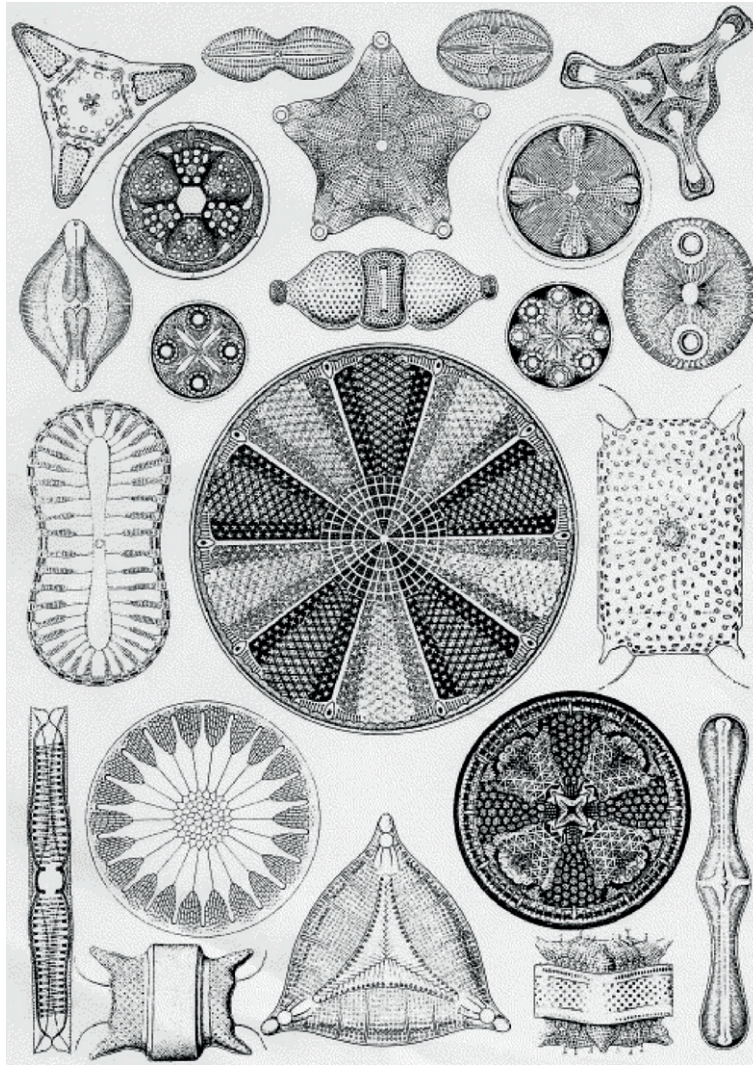


Figure 2 The variety of marine diatoms as illustrated by E Haeckel (1904; *Artforms in Nature*. Dover Press Pictorial Archive Series, Copyright-free).

volume). As well as growing as plankton in open waters, they grow attached to various substrates in the coastal zone, for example, as epiphytes on seaweeds or as epilithon on rocks. Diatoms also form biofilms on the surface of intertidal mudflats; the motile forms show a remarkable tidal rhythm. On mudflats, diatoms influence sediment stability by increasing sediment cohesion through the excretion of extracellular polysaccharides. By reducing erosive stress, diatoms promote the silt content of the sediment, and high densities of diatoms significantly reduce the frequency and magnitude of erosion events in some coastal areas.

In the coastal zone, Quaternary scientists have used diatom salinity preferences and life forms for environmental reconstruction (Cooper et al., 2010; Snoeijs and Weckström, 2010; Trobajo and Sullivan, 2010) and used diatoms to determine paleoceanographic conditions (Koizumi, this volume). Other applications include sea-level reconstruction, studies of basin isolation, paleotidal reconstructions, studies of salt marsh development, inference of tsunami events, and biological indicators of pollution (e.g., eutrophication, radionuclide contamination) which are detailed extensively in

Smol and Stoermer (2010). Zong and Horton (1999) modeled the relationship between diatom assemblages and tidal levels and developed a diatom-based tidal-level transfer function which was applied successfully to estimate paleotidal levels from fossil diatom data recorded in late Holocene coastal sequences.

A unique habitat is formed by the seasonal expansion and contraction of sea ice at the poles. Diatoms can grow attached to the underside of the ice as well as interstitially and attached to ice platelets. These different communities can be distinguished and have been used to trace the development of sea ice in the Quaternary (Stickley et al., this volume; Koç et al., this volume). Sea ice diatoms can leave biomarkers in marine sediments and these have been used as a proxy for past sea ice presence (see Belt et al., this volume).

Freshwater Habitats

Planktonic diatoms

As in the oceans, planktonic diatoms are common in open water. In lakes, however, many forms are not truly planktonic

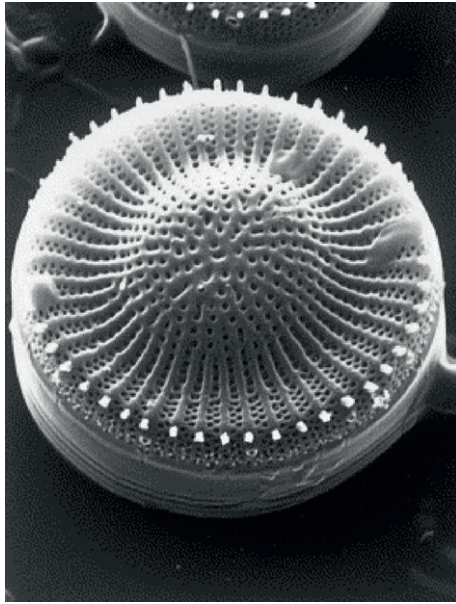


Figure 3 A scanning electron micrograph of *Cyclostephanos tholiformis*, a typical centric diatom. Photograph by C. Sayer.



Figure 4 A light micrograph of a colony of *Meridion circulare*, a common stream diatom, showing the silica cell walls and chloroplasts. Photograph by S. Spaulding.

as they spend some of their lifecycle resting on the sediment (meroplanktonic; e.g., some *Aulocoseira* spp.). However, planktonic species from both lakes and oceans are subject to two special influences: the availability of silicate and the tendency of diatoms to sink as they are as dense as or denser than water. However, the photic zone is often turbulent due to winds, currents, and convection; therefore, many diatoms exhibit features such as a long cylindrical shape or long chains which slow their sinking velocity. In the northern temperate lakes, there is a well-established pattern of seasonal succession in the plankton with a spring bloom, a subsequent decline (as silica is depleted), and a second peak in autumn. The species are often distinct, due mainly to nutrient differences between spring and autumn, and therefore this information has been used to interpret climate and weather patterns. In Elk Lake, Bradbury et al. (2002) interpreted changes in the ratio of *Stephanodiscus*

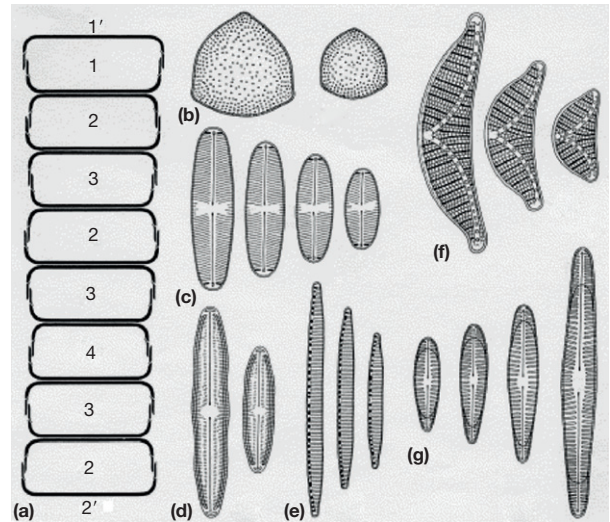


Figure 5 Diatom vegetative reproduction. (a) Change in valve size following normal vegetative (mitotic) cell division 1' and 2' indicate the epivalve and hypovalve of the original cell: (b)–(g) examples of changes in valve shape and size following division (b) *Stictodiscus*, (c) *Sellaphora*, (d) *Brachysira*, (e) *Nitzschia*, (f) *Epithemia*, and (g) *Rhocosphenia*. Reproduced from Round FE, Crawford RM, and Mann DG (1990) *The Diatoms: Biology and Morphology of the Genera*. Cambridge: Cambridge University Press, with kind permission from Cambridge University Press.

minutulus and *Fragilaria crotonensis* as indicative of shifts between long, dry springs and warm, stormy summers. In addition, wavelet analysis suggested strong periodic correlations between individual diatom species and various sunspot cycles such as the 11-year Schwabe cycle (see also Mackay et al., 2003).

In both marine and freshwater habitats, the distinction between plankton and benthos is not absolute, as well as the meroplanktonic forms; species which have their true habitat in the benthos can often be found resuspended in the water column (tychoplanktonic).

Benthic diatoms

Benthic diatoms are often very diverse and can live either free on or in the sediments, or attached to the substratum, for example, rocks (epilithon), sand grains (epipsammon), and plants (epiphyton). Diatoms living on soft sediments (epipelion) where the chances of burial are high are often motile species that are able to migrate up and down in the sediment matrix. Attached communities can be adnate (i.e., closely attached to the substrate; e.g., *Cocconeis*) or attached by mucilaginous stalks or pads (e.g., *Gomphonema*). Often diatoms can group themselves into three-dimensional (3D) communities forming complex microscopic 'forests' (Figure 6). Diatoms can also form part of the diverse 3D microbial communities of filamentous algal mats which form in streams and lakes, in some cases forming perennial layers up to 30-cm thick.

Although diatoms rarely have very discrete habitats (there are often overlaps between species from different communities), habitat information is still useful for Quaternary scientists. For example, Sayer (2001) suggested that a recent increase in *Cocconeis placentula*, a very common epiphyte in Groby Pool,

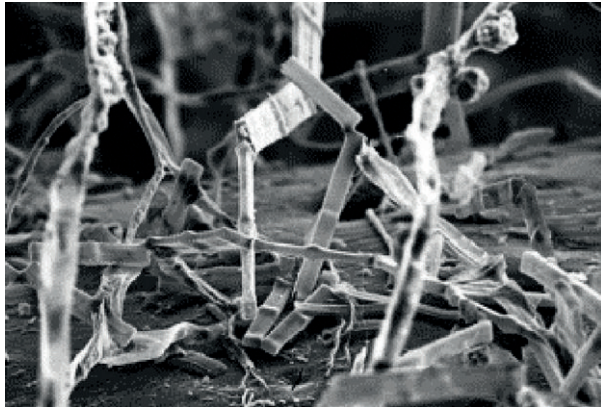


Figure 6 A colony of *Tabellaria quadriseptata* frozen *in situ* and examined by scanning electron microscopy. Photograph by R. Flower.

was due to an increase in macrophytes in the lake over the last 50 years caused by increasing nutrient enrichment. In this study, a shift from benthic *Fragilaria* communities to planktonic communities dominated by *Stephanodiscus parvus*, *Cyclotella tholiformis* (Figure 3) and *Cyclotella dubius* was also seen as a consequence of nutrient enrichment. Although the ratio of benthic to planktonic species can give useful information about the changing trophic state of some lakes, there has been a shift to the use of quantitative statistical methods (see Section ‘Diatoms and Quantitative Inferences of Past Water Quality Using Transfer Functions’) for reconstructing lake eutrophication.

Aerophilic diatoms

Diatoms can also grow as subaerial forms (aerophilic) on damp rock faces, terrestrial soils, or even on the leaves of plants growing in damp environments. These include species such as *Hantzschia amphioxys*, *Diadesmis* spp., some *Pinnularia* spp., and large chain-forming *Orthoseira* spp.; some of these are obligate aerophiles (Johansen, 2010). When these are preserved in sediments, they can provide useful information for environmental scientists, indicating drying out of a water body, for example.

River diatoms

In rivers, diatoms can grow attached to substrates, and in large rivers they can form a planktonic community. Some pennate species such as *Meridion circulare* (Figure 4) and *Hannaea arcus* are characteristic of running waters while a characteristic centric species of running water is *Melosira varians*. Stream velocity is an important determinant of both species composition and the physical structure of communities. Quaternary applications in lotic ecosystems are clearly limited, owing to the lack of continually accumulating sediments. However, undisturbed sediments can be found in reservoirs or deltas and in paleochannels and floodplain lakes in lowland river systems. The floodplains of relatively large, lowland rivers often have old channel systems (paleochannels) and associated oxbow cutoff lakes (or billabongs). Such systems are low energy and thus provide suitable sedimentary environments for paleoecological study using diatoms (Reavie and Edlund, 2010). Thoms

et al. (1999) determined the impact of agriculture and river regulation on water quality in billabongs of southeastern Australia using the fossil diatom record. Reavie and Smol (1997) developed reliable transfer functions for reconstructing past habitat conditions based on diatom habitat preferences (rock, *Cladophora*, macrophytes) in the St. Lawrence River, Canada. The models were applied to a fluvial sediment sequence and showed a marked increase in macrophyte-associated diatom taxa which the authors attributed to accelerated nutrient inputs. Ludlam et al. (1996) used a simple indicator for the relative contribution of running-water communities in an arctic lake. The Lotic Index showed a clear positive relationship to sedimentation rate as recorded in the varves, and therefore seem to be a good indicator of stream flow.

Biogeography and Endemism

There has often been an emphasis on the cosmopolitan nature of diatoms. While some species have a global distribution, for example, *Achnanthes minutissimum*, others may not be as widespread as has been suggested. This is due to a number of factors: for example, the uncritical use of floras from other regions and taxonomic clumping of closely related taxa. Certainly it seems that the numbers of diatom species have been underestimated, and when careful taxonomic work is done, several species that cannot be attributed to published species are almost always identified for the first time. Endemism describes diatoms that are restricted to a particular location. The best-known examples of endemic diatom floras are island floras and long-lived or ancient freshwater lake systems such as those found in Lake Baikal, Lake Ochrid (Macedonia), or the African Rift Lakes (Mackay et al., 2010). The endemic flora of Lake Baikal and its application to paleoclimate reconstruction is discussed by Mackay (this volume).

Collecting and Studying Live Diatoms

Live diatom material can be collected from different habitats using a variety of means; attached diatoms can simply be scraped off rocks or plants using a brush, a known area of substrate can be used if density data are required. A lens tissue can be used to collect motile forms. Planktonic forms can be sampled with a bottle or plankton net. Diatoms can also be frozen *in situ* to preserve their 3D architecture (Figure 6). Normally, though, samples are preserved with a few drops of Lugol's iodine to retain the form and position of the chloroplasts. Even though most taxonomy is based on the cleaned frustule, it is possible to identify living diatoms using cell and colony features (Cox, 1996). Although often ignored by Quaternary scientists the identification of live versus dead cells can be informative as to whether modern diatoms collected from a particular habitat are actually living there or whether they have been transported there after death. Diatomists working with lake sediments need to sample both living and fossil communities to develop transfer functions and to understand the relationship between source communities and fossil assemblages. Further details of field sampling and coring are given in Battarbee et al. (2001).

Cleaning and Sample Preparation

In most Quaternary applications fossil diatoms need to be separated from the sedimentary matrix. The precise steps depend on the nature of the material and care needs to be taken not to lose or damage valves. For example, delicate diatoms can easily be broken by rapid centrifugation. Figure 7 illustrates the

basic steps needed, but often the removal of organic matter by oxidation with either hydrogen peroxide, strong acids, or by incineration is all that is required. Once cleaned material is obtained, permanent slides can be made by mounting in a suitable medium with a high refractive index (>1.7) such as Naphrax (refractive index 1.73). Diatom concentrations can be calculated by adding a known quantity of markers such as

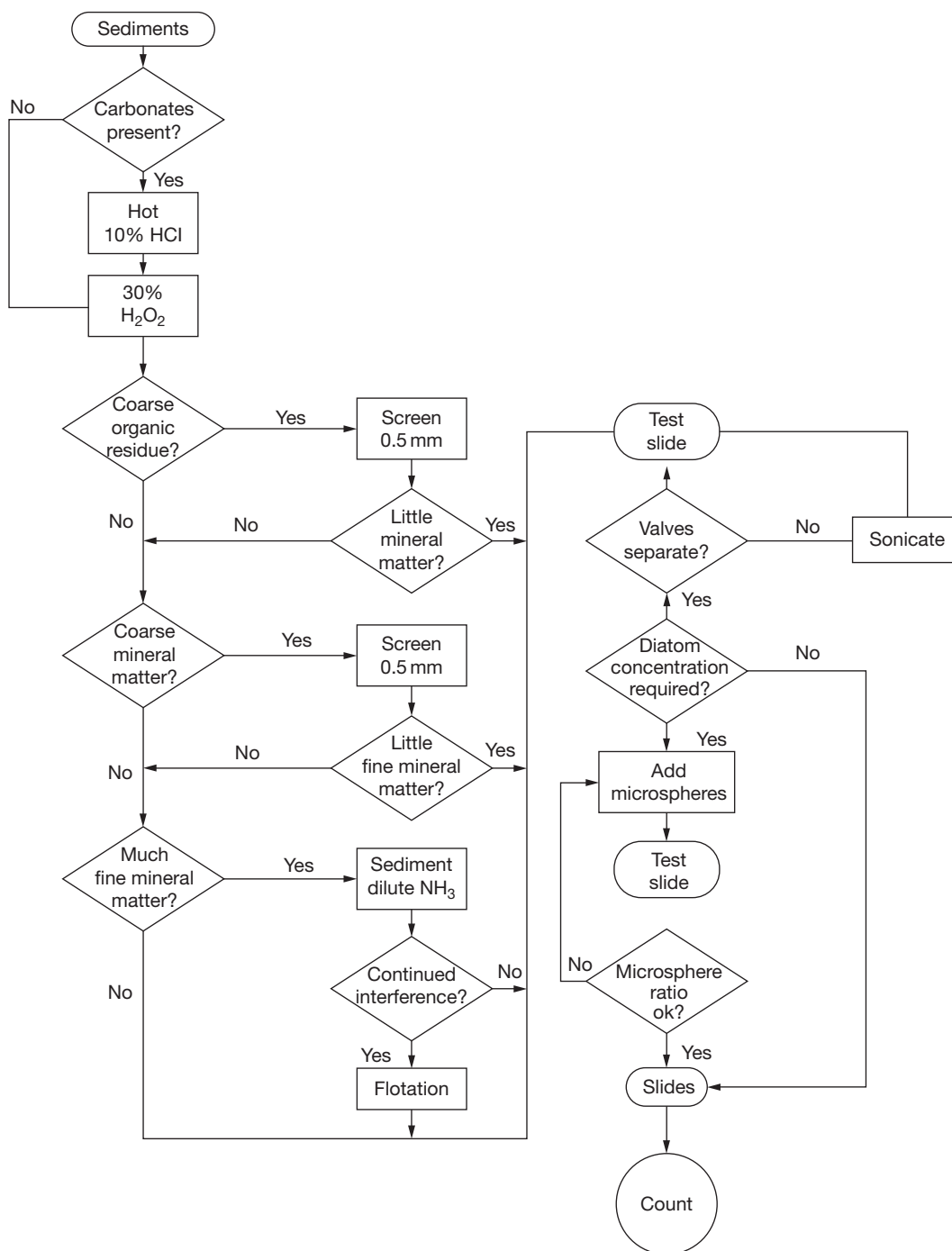


Figure 7 Flow chart for sample and slide preparation. Reproduced from Battarbee RW, Jones VJ, Flower RJ, et al. (2001) Diatoms. In: Smol JP, Birks HJB, and Last WM (eds.) *Tracking Environmental Change Using Lake Sediments. Terrestrial, Algal and Siliceous Indicators*, vol. 3, pp. 155–202. Dordrecht: Kluwer Academic, with kind permission from Springer Science and Business Media.

divinylbenzene microspheres which are commercially available in the 5–10 μm size range. For further details on preparation techniques, see [Battarbee et al. \(2001\)](#).

Identification

Most diatom assemblages are diverse and some expertise is needed for identification. Identification to the generic level is relatively straightforward although it requires some familiarity with terminology. Recent keys to genera include [Barber and Haworth \(1981\)](#), [Kelly \(2000\)](#), and [Tomas \(1996\)](#). Once familiar with the genera, more specialized floras are needed (e.g., [Krammer and Lange-Bertalot, 1986](#)); ideally these should be generated from the same geographic region as the diatom sample. In addition, there are more specialized floras for specific ecosystems or specific genera which are available as books, CD-ROMs, and on the Internet. Databases such as the European Diatom Database (<http://craticula.ncl.ac.uk/Eddi/jsp>) combine information on the distribution of diatoms in surface sediment datasets, including images, with the ability to perform online reconstructions.

Enumerating Diatoms

Most routine counts are based on the light microscope using oil immersion objectives and magnifications of 750–1500 $\times$. Bright field, phase contrast, or differential interference contrast illumination can be used, and an eyepiece graticule is useful for measuring. Digital cameras with image capture software are increasingly being used to manipulate and store images.

Diatoms species counts are commonly expressed as a percentage of a total diatom sum or as a concentration value. Commonly the sum comprises all taxa but occasionally it is useful to exclude some groups, for example, when these can be identified as clear contaminants from reworked material. Appropriate counting strategies need to be developed for individual sequences and to ensure that a representative proportion of the coverslip is examined and that broken fragments of valves are identified and enumerated. Concentration data can be expressed as the number of valves per gram of sediment, or more usefully they can be used to produce a diatom accumulation rate in valves per gram per year. As diatoms vary considerably in size, this statistic may not be particularly meaningful unless it is converted to biovolume.

Preservation and Taphonomy

The preservation of diatoms in sediments ranges from excellent in diatomites (see [Flower, this volume](#)) to poor in some saline lakes ([Fritz, this volume](#)) and in extreme cases no diatoms are preserved in the sediment. Factors affecting preservation include salinity, temperature, alkalinity, and sediment accumulation rate as well as the silica content of the cell and the concentration gradient of dissolved silica between the sediment and overlying water. Differential dissolution can occur leading to bias in the composition of the assemblage to more robust forms ([Ryves et al., 2003](#)). A key question is whether diatoms preserved in the sediment actually reflect the communities from which they are derived. Taphonomic processes

include removal from the outflow of lakes, resuspension and reworking of older sediments, losses through grazing, bioturbation in the sediment, and dissolution. Some studies show a good agreement between the composition of live communities and the sediment record (e.g., [Cameron, 1995](#)) while some studies show considerable differences between diatoms found in the water column and the sediment record. For example, in Lake Baikal the majority of diatoms dissolve before reaching the sediment surface (see [Mackay, this volume](#)).

Diatoms as Paleoenvironmental Indicators

Besides the types of habitats available, there are other controls on the composition of the diatom flora. Physical controls include temperature, turbulence, and light. Chemical controls include factors such as pH, nutrients, and salinity. Biological controls such as grazing and parasitism are also important.

The ubiquitous nature of diatoms is one reason why diatoms are widely used as biological indicators in Quaternary paleoecology ([Smol and Stoermer, 2010](#)) and they have been used to provide information specifically for archaeologists (see [Cameron, this volume](#)) as well as a wide range of past environmental and ecological changes, such as climatic and lake-level changes, water pollution, eutrophication, and acidification.

Fossil diatoms have been used as indicators of environmental change on a range of time scales from decadal to millennial. Many studies have focused on Late Glacial and Holocene sequences, but long diatom records encompassing several glacial–interglacial cycles are available from marine and freshwater sequences (e.g., see [Koc, this volume](#); [Mackay, this volume](#)). In recent decades, progress has been made in our methods of interpreting diatom stratigraphical data through powerful numerical techniques that allow accurate quantitative reconstructions of environmental change to be made ([Birks, 2010](#)).

Diatoms are excellent paleoindicators in Quaternary studies ([Korhola, 2007](#); [Smol and Stoermer, 2010](#)), because:

- They are common in aquatic systems.
- They are diverse, and communities are usually represented by many taxa which increase the ecological information gained and the confidence of environmental inferences.
- Diatoms show a broad range of tolerance along environmental and ecological gradients, with individual species having specific ecological and water quality requirements that are usually quantifiable. [Figure 8](#) shows how different species respond along a pH gradient.
- They have short generation times so can respond rapidly to environmental change.
- Their siliceous frustules are resistant to degradation and they are often well preserved in different sedimentary environments.
- The taxonomy of diatoms is relatively well known and allows identification of fossil diatom assemblages to the species or even subspecies level.

Diatoms and Quantitative Inferences of Past Water Quality Using Transfer Functions

Diatoms have been used to quantitatively infer past conditions with a range of techniques. The early applications were

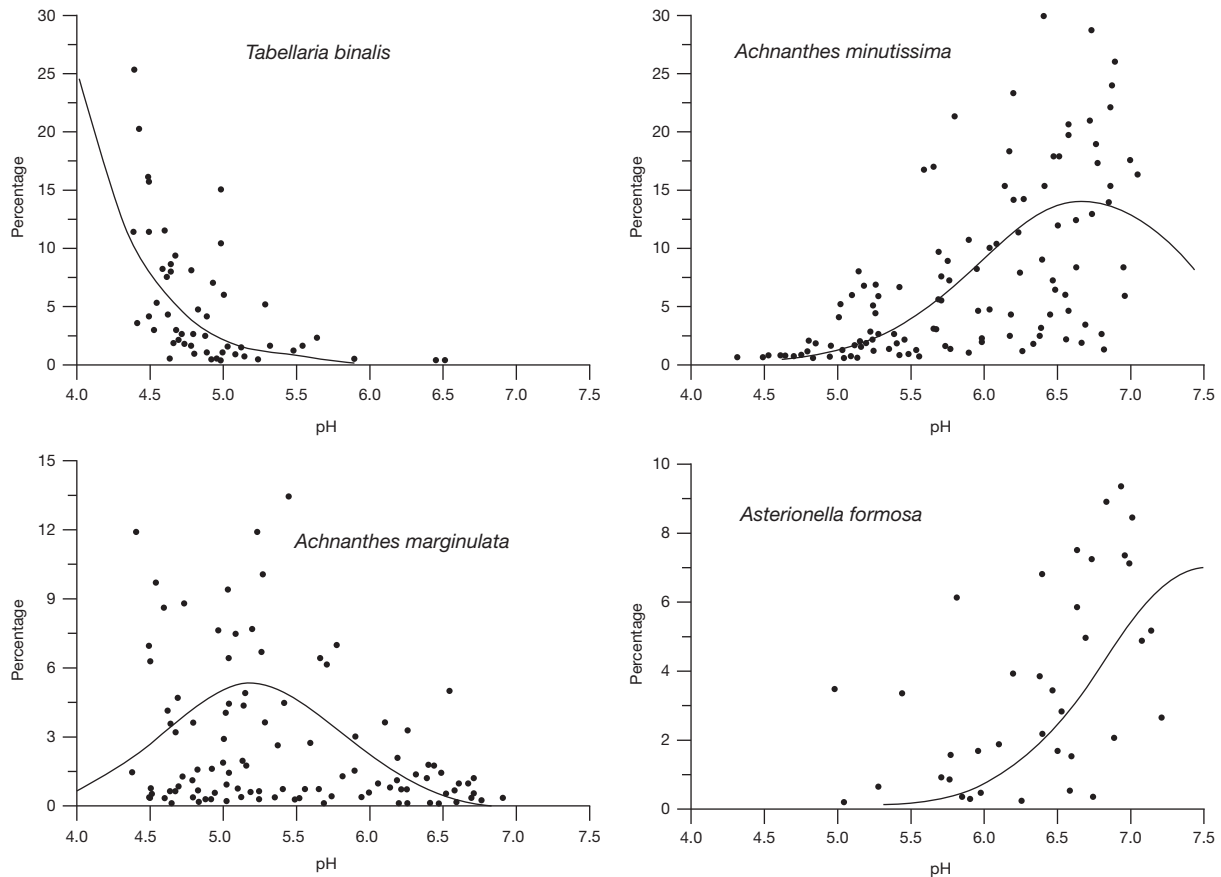


Figure 8 Fitted Gaussian response curves for four diatom species along a pH gradient. Data are from the SWAP dataset of 167 lakes from England, Wales, Scotland, Norway, and Sweden (Stevenson et al., 1991), with kind permission from Cambridge University Press.

developed for pH inference and were based on species categories and indices, later multiple linear regression models were developed to estimate pH according to the proportions of the established pH-preference groups (Battarbee et al., 2010). More recently, paleoecologists have used the ecological optima and tolerances of diatom assemblages to construct transfer functions that allow environmental conditions to be inferred from fossil assemblage data (Birks, 2010). Transfer functions have been used for a range of inferences and Quaternary applications including reconstructing nutrients (Bennion et al., 2010), salinity (Fritz, this volume), conductivity (Mackay, this volume) and sea-surface temperature (Koç, this volume).

Transfer functions are used to describe the relationship between diatom assemblages and their environment and then this relationship is applied to past assemblages to quantitatively infer past conditions. The basic approach is illustrated as a hypothetical example in Figure 9 (this is Figure 6 from Korhola, 2007) and involves two steps: (1) the regression step in which the responses of taxa to the contemporary environment are modeled and (2) a calibration step in which the environmental variable of interest is predicted from the fossil compositional data. Transfer functions are established using modern diatom assemblages and measured environmental variables from a calibration dataset (also called a training set; Figure 9(a)) which are created by sampling a large (between 30

and 200) number of sites (e.g., lakes, wetlands, estuaries, and marine environments) along the environmental gradient of interest (e.g., pH, nutrient, and water temperature) for water chemistry and surface sediments (the uppermost sediment representing recent sediment accumulation; Figure 9(b)). The transfer function approach is based on the implicit assumption that the modern training sets cover the range of environmental conditions likely to be represented by the fossil assemblages. Hence, calibration datasets should ideally span all biologically relevant environmental gradients.

There are now many numerical techniques for deriving transfer functions. Simple two-way weighted average regression and calibration, weighted-averaging partial least squares (WA-PLS) regression and calibration, Gaussian logit regression (GLR), and maximum likelihood (ML) calibration are relatively statistically robust and computationally economical and have been widely adopted (Birks, 2010). Ideally the reliability of transfer functions can be assessed by comparison with measured historical data (e.g., Fritz, 1990) but such datasets are rare. Alternatively, diatom-based reconstructions can be compared with independent paleoenvironmental data. Without either historical validation, or independent data, evaluation needs to be made indirectly using numerical criteria which include analog measures, 'goodness-of-fit' statistics, and sample-specific errors (Birks, 2010).

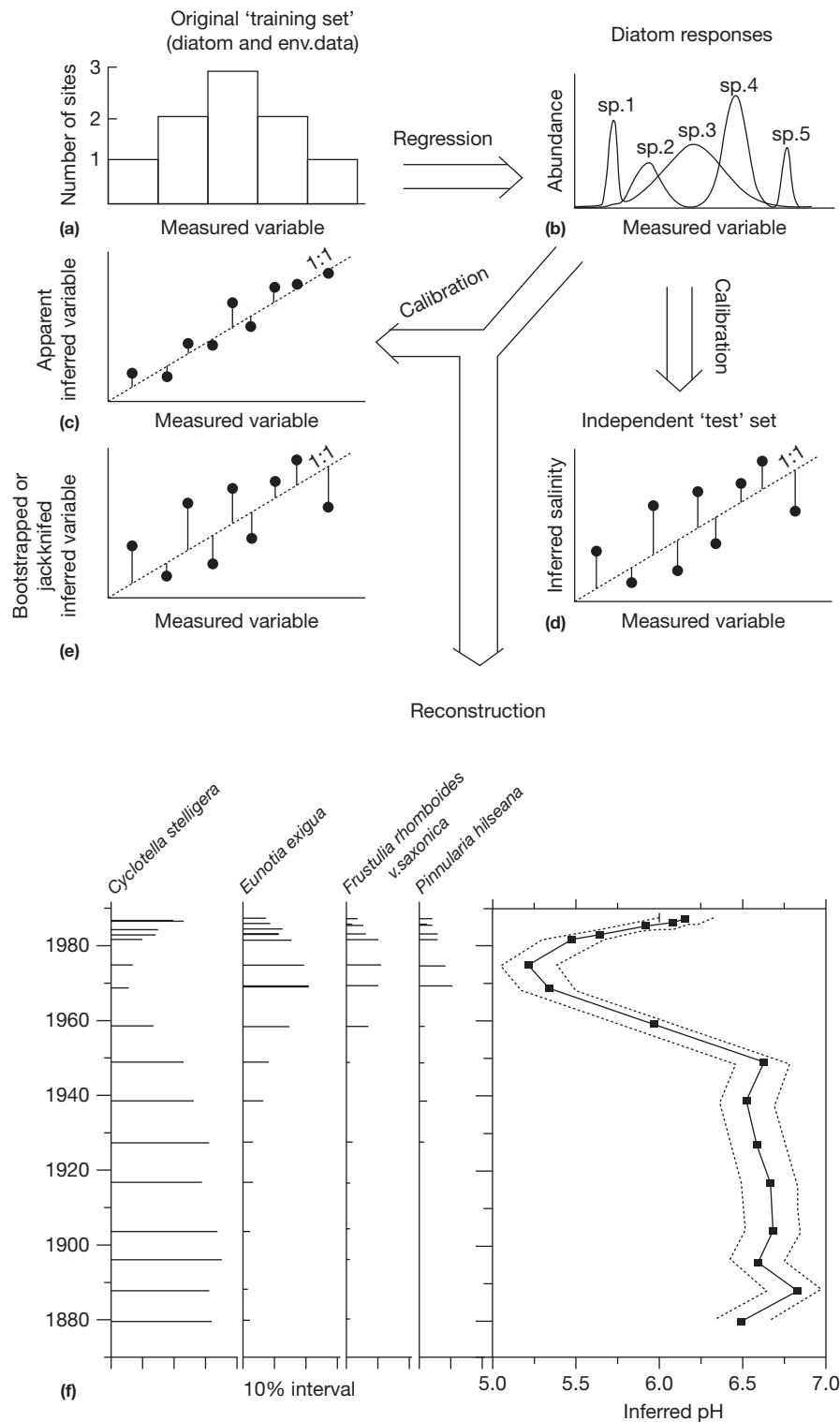


Figure 9 Simplified illustration of the typical steps involved in developing and assessing the predictive ability of a diatom-based calibration model and its application to fossil diatom data including: (a) selection of a modern training dataset; (b) estimation of taxon responses based on their distribution in the training set (regression); (c) generation of a transfer function on the basis of the training set (calibration); (d) selection of an independent test set to assess the transfer function (calibration); (e) use of computer resampling techniques, such as jackknifing or bootstrapping, to assess the transfer function (calibration); and (f) reconstruction of parameter value from fossil diatom assemblages and estimation of sample-specific prediction errors by bootstrapping. Modified from Fritz SC, Cumming BF, Gasse F, and Laird KR (1997) Diatoms as indicators of hydrologic and climatic change in saline lakes. In: Stoermer EF and Smol JP (eds.) *The Diatoms: Applications for the Environmental and Earth Sciences*, 1st edn., pp. 41–72. Cambridge: Cambridge University Press.

See also: **Diatom Methods:** $\delta^{18}\text{O}$ Records; Diatomites: Their Formation, Distribution, and Uses; Salinity and Climate Reconstructions from Continental Lakes; Use in Archaeology. **Diatom Records:** Antarctic Waters; Diatom Fossil Records from Marine Laminated Sediments; Freshwater Laminated Sequences; Large Lakes; North Atlantic and Arctic; Pacific.

References

- Armbrust EV, Berges JA, Bowler C, et al. (2004) The genome of the diatom *Thalassiosira pseudonana*: Ecology, evolution, and metabolism. *Science* 306: 79–86.
- Barber HG and Haworth EY (1981) *A Guide to the Morphology of the Diatom Frustule*. Ambleside: Freshwater Biological Association.
- Battarbee RW, Charles DF, Bigler C, Cumming BF, and Renberg I (2010) Diatoms as indicators of surface water acidity. In: Smol JP and Stoermer EF (eds.) *The Diatoms: Applications for the Environmental and Earth Sciences*, 2nd edn., pp. 98–121. Cambridge: Cambridge University Press.
- Battarbee RW, Jones VJ, Flower RJ, et al. (2001) Diatoms. In: Smol JP, Birks HJB, and Last WM (eds.) *Tracking Environmental Change Using Lake Sediments. Terrestrial, Algal and Siliceous Indicators*, vol. 3, pp. 155–202. Dordrecht: Kluwer Academic.
- Bennion H, Sayer CD, Tibby J, and Carrick HJ (2010) The diatoms: applications for the environmental and earth sciences. In: Smol JP and Stoermer EF (eds.) *Diatoms as Indicators of Environmental Change in Shallow Lakes*, 2nd edn., pp. 152–173. Cambridge: Cambridge University Press.
- Birks HJB (2010) Numerical methods for the analysis of diatom assemblage data. In: Smol JP and Stoermer EF (eds.) *The Diatoms: Applications for the Environmental and Earth Sciences*, 2nd edn., pp. 23–54. Cambridge: Cambridge University Press.
- Bradbury P, Cumming B, and Laird K (2002) A 1500-year record of climatic and environmental change in Elk Lake, Minnesota III: Measures of past primary productivity. *Journal of Paleolimnology* 27: 321–340.
- Cameron NG (1995) The representation of diatom communities by fossil assemblages in a small acid lake. *Journal of Paleolimnology* 14: 185–223.
- Cooper SR, Gaiser E, and Wachnicka A (2010) Estuarine paleoenvironmental reconstructions using diatoms. In: Smol JP and Stoermer EF (eds.) *The Diatoms: Applications for the Environmental and Earth Sciences*, 2nd edn., pp. 324–345. Cambridge: Cambridge University Press.
- Cortese G and Gersonde R (2007) Morphometric variability in the diatom *Fragilariopsis kerguelensis*: Implications for Southern Ocean paleoceanography. *Earth and Planetary Science Letters* 257: 526–544.
- Cox EJ (1996) *Identification of Freshwater Diatoms from Live Material*. London: Chapman and Hall.
- Fritz SC (1990) Twentieth-century salinity and water level fluctuations in Devils Lake, N. Dakota: A test of a diatom-based transfer function. *Limnology and Oceanography* 35: 1844–1856.
- Johansen JR (2010) Diatoms of aerial habitats. In: Smol JP and Stoermer EF (eds.) *The Diatoms: Applications for the Environmental and Earth Sciences*, 2nd edn., pp. 465–472. Cambridge: Cambridge University Press.
- Kelly M (2000) Identification of common benthic diatoms in rivers. *Field Studies* 9: 583–700.
- Korhola A (2007) Data interpretation. In: Elias S (ed.) *Encyclopedia of Quaternary Science*, 1st edn. pp. 494–507. Amsterdam: Elsevier.
- Krammer K and Lange-Bertalot H (1986) *Bacillariophyceae*. Stuttgart: Gustav Fisher Verlag.
- Ludlam SD, Feeney S, and Douglas MSV (1996) Changes in the importance of lotic and littoral diatoms in a high arctic lake over the last 191 years. *Journal of Paleolimnology* 16: 187–204.
- Mackay AW, Edlund MB, and Khursevich G (2010) Diatoms in ancient lakes. In: Smol JP and Stoermer EF (eds.) *The Diatoms: Applications for the Environmental and Earth Sciences*, 2nd edn., pp. 209–228. Cambridge: Cambridge University Press.
- Mackay AW, Jones VJ, and Battarbee RW (2003) Approaches to Holocene climate reconstruction using diatoms. In: Mackay AW, Battarbee RW, Birks HJB, and Oldfield F (eds.) *Global Change in the Holocene*, pp. 294–309. London: Arnold.
- Mann DG (1988) Why didn't Lund see sex in *Asterionella*? A discussion of the diatom life cycle in nature. In: Round FE (ed.) *Algae and the Aquatic Environment*, pp. 383–412. Bristol: Biopress.
- Reavie ED and Edlund MB (2010) Diatoms as indicators of long-term environmental change in rivers, fluvial lakes, and impoundments. In: Smol JP and Stoermer EF (eds.) *The Diatoms: Applications for the Environmental and Earth Sciences*, 2nd edn., pp. 86–97. Cambridge: Cambridge University Press.
- Reavie ED and Smol JP (1997) Diatom-based model to infer past littoral habitat characteristics in the St. Lawrence River. *Journal of Great Lakes Research* 23: 339–348.
- Round FE, Crawford RM, and Mann DG (1990) *The Diatoms: Biology and Morphology of the Genera*. Cambridge: Cambridge University Press.
- Ryves DB, Jewson DH, Sturm M, et al. (2003) Quantitative relationships between planktonic diatom communities and diatom assemblages in sedimenting material and surface sediments in Lake Baikal. *Limnology and Oceanography* 48: 1643–1661.
- Sayer CD (2001) Problems with the application of diatom-total phosphorus transfer functions: Examples from a shallow English lake. *Freshwater Biology* 46: 743–757.
- Smol JP and Stoermer EF (2010) *The Diatoms: Applications for the Environmental and Earth Sciences*, 2nd edn. Cambridge: Cambridge University Press.
- Snoeijs P and Weckström K (2010) Diatoms and environmental change in large brackish water ecosystems. In: Smol JP and Stoermer EF (eds.) *The Diatoms: Applications for the Environmental and Earth Sciences*, 2nd edn., pp. 287–308. Cambridge: Cambridge University Press.
- Stevenson AC, Juggins S, Birks HJB, et al. (1991) The Surface Waters Acidification Project Palaeolimnology Programme: Modern Diatom/Lake-Water Chemistry Data-Set. London: Ensis Ltd. 86 pp.
- Thoms MC, Ogden RW, and Reid MA (1999) Establishing the condition of lowland floodplain rivers: A palaeo-ecological approach. *Freshwater Biology* 41: 407–423.
- Tomas CME (1996) *Identifying Marine Diatoms and Dinoflagellates*. San Diego: Academic Press.
- Trobajo R and Sullivan M (2010) Applied diatom studies in estuaries and shallow coastal environments. In: Smol JP and Stoermer EF (eds.) *The Diatoms: Applications for the Environmental and Earth Sciences*, 2nd edn., pp. 309–323. Cambridge: Cambridge University Press.
- van den Hoek C, Mann DG, and Jahns HM (1995) *Algae: An Introduction to Phycology*. Cambridge: Cambridge University Press.
- Zong Y and Horton BP (1999) Diatom-based tidal-level transfer functions as an aid in reconstructing Quaternary history of sea-level movements in the UK. *Journal of Quaternary Science* 14: 153–167.

DIATOM METHODS

Contents

$\delta^{18}\text{O}$ Records

Data Interpretation

Diatomites: Their Formation, Distribution, and Uses

**Salinity and Climate Reconstructions from Continental Lakes
Use in Archaeology**

$\delta^{18}\text{O}$ Records

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Introduction

Diatom silica is a form of biogenic opal ($\text{SiO}_2 \cdot n\text{H}_2\text{O}$, [Figure 1](#)) that contains oxygen isotopes and can be used in lacustrine and marine paleoenvironmental studies. The oxygen isotope ratio ($^{18}\text{O}/^{16}\text{O}$) of diatom silica is expressed on the delta scale (δ) in terms of per mille (or per mil) (‰):

$$\delta = [(R_{\text{sample}}/R_{\text{reference}}) - 1] \times 1000\text{‰} \quad [1]$$

where R is $^{18}\text{O}/^{16}\text{O}$, and 'reference' means the appropriate universally accepted reference material. The ' δ ' for each element takes its name from the heavy isotope, thus $\delta^{18}\text{O}$. For diatom oxygen, the reference is Vienna Standard Mean Ocean Water (VSMOW), a specially prepared distilled seawater. There is no universally accepted standard material for oxygen isotopes from diatom silica ($\delta^{18}\text{O}_{\text{diatom}}$), although most laboratories analyze it alongside the NBS28 (quartz) and their own standard diatomites. VSMOW and NBS28 are held and distributed by the International Atomic Energy Agency (IAEA) in Vienna.

Diatoms form a shell or frustule, following a seasonal pattern defined by the variability of climate, nutrient supply, mixing regimes, and at high latitudes the period of ice cover. The isotope signature acquired by diatoms will, therefore, be skewed toward their major growing season, which will be specific to the lake or oceanic region under consideration.

Analytical Considerations

Analytical techniques for measuring $\delta^{18}\text{O}_{\text{diatom}}$ will generally liberate oxygen from all the components in the sediment, for example, silt, clay, tephra, carbonates, other siliceous organisms and organic matter. Since the presence of contaminants

will distort the $\delta^{18}\text{O}_{\text{diatom}}$ signal, it is important that the sample is almost entirely composed of diatom frustules. Similarly, only pristine samples should be analyzed due to the risk of isotopic fractionation occurring during diagenesis and dissolution, leading to an increase in $\delta^{18}\text{O}_{\text{diatom}}$ ([Moschen et al., 2005](#)). Published methods of removing contamination involve various stages of chemical attack (including HCl, H_2O_2 , HNO_3), sieving, differential settling ([Figure 2\(a\)](#) and [2\(b\)](#)) and the use of heavy liquids and gravitational split-flow thin (SPLITT) fractionation ([Figure 3](#)).

Disequilibrium effects (commonly known as 'vital' effects in the case of biogenic carbonates) include a variety of rate effects and microenvironment-induced changes that cause the mineral to have an isotope composition that is different from that predicted purely by thermodynamics. A number of studies on both marine and lacustrine taxa have failed to find any evidence for such a process in $\delta^{18}\text{O}_{\text{diatom}}$ outside of analytical error (e.g., [Brandriss et al., 1998](#); [Juillet-Leclerc and Labeyrie, 1987](#); [Moschen et al., 2005](#); [Sancetta et al., 1985](#); [Schiff et al., 2009](#); [Schmidt et al., 2001](#); [Shemesh et al., 1995](#)). An exception is a stratigraphic record from the North Pacific Ocean which displays significant variations between different size fractions, each containing different taxa ([Swann et al., 2007, 2008](#)). Further work is required to fully understand the processes behind these changes, demanding the need for caution when analyzing $\delta^{18}\text{O}_{\text{diatom}}$ in bulk samples containing numerous taxa. Attempts are ongoing to develop a means of routinely extracting individual species for isotope analysis.

Analytical Methods

The chemical structure of a diatom can be described by defining Q_n where Q represents a silicon atom surrounded by n

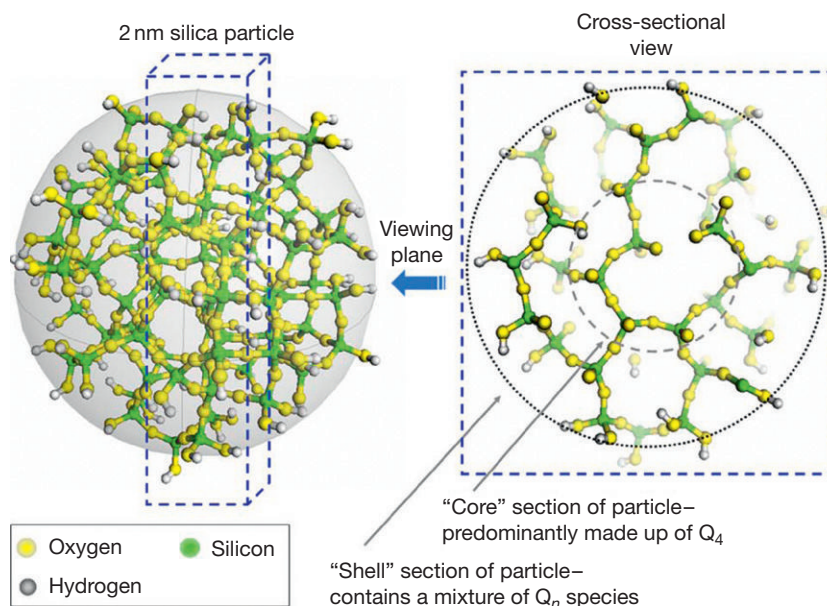


Figure 1 Diatom silica is a form of biogenic opal and has the chemical formula $\text{SiO}_2 \cdot n\text{H}_2\text{O}$. Here the structure of biogenic silica is shown in a 2-nm particle. The core of the particle is the only part predominantly made up of Q_4 , while subsequent layers contain a mixture of Q_n species. Q represents a silicon atom surrounded by four oxygen atoms while suffix n gives the number of surrounding oxygen atoms (out of 4) that are bonded to another silicon atom.

oxygen atoms. Frustules contain a Q_4 , tetrahedrally bonded silica (Si—O—Si) layer, surrounded by more hydrous, Q_3 , material (Si—OH). The exact distribution of oxygen in the frustule, however, is more complex than simple layering and is related to the mode of frustule formation and its differential porosity. Oxygen in the hydrous layer, amounting to ~5–40% of the frustule, is freely exchangeable with any water the diatom silica comes into contact with during sample preparation and must be removed prior to analysis using oxidizing reagents and/or high temperatures.

Breaking the internal Si—O bond requires considerable energy, and four main methods have been developed to obtain the $\delta^{18}\text{O}_{\text{diatom}}$ paleoenvironmental signal stored within the tetrahedrally bonded, Q_4 , layer: controlled isotope exchange (CIE), stepwise fluorination (SWF) techniques, inductive high-temperature carbon reduction (iHTR) and inert gas flow dehydration (iGFD) method. iHTR and iGFD are based on the thermal reduction and dehydration of silica at high temperatures (1100–1550 °C for biogenic silica; Chapligin et al., 2010; Lücke et al., 2005). The other two techniques involve the use of a fluorine-based compound such as F_2 , ClF_3 , or BrF_5 (extremely powerful oxidizing reagents) to liberate the oxygen from the diatom silica. The CIE method uses the fact that the surficial hydrous silica (Si—OH) is reactive under much lower energy conditions than the inner tetrahedrally bound silica, which allows oxygen atoms of known isotope composition to be exchanged with the unstable hydrous material while leaving the skeletal silica intact (Labeyrie and Juillet, 1982). The method exchanges oxygen in Q_3 bond, hydrous silica, with oxygen of a known isotope value, and then through mass balance calculations determines the composition of the non-exchangeable silicate oxygen after fluorination. The SWF method involves reacting the silica with a fluorine compound in a series of steps (Figure 4) and was first described by

Haimson and Knauth (1983). Oxygen from Q_3 bonds are disassociated from the biogenic silica in the first step leaving the structurally bound silica for subsequent reactions (Leng and Sloane, 2008).

Typically, between 1.5 and 6.5 mg of material is required for a single analysis depending on the analytical procedure being followed. Interlaboratory comparisons demonstrate comparable results between each technique (B. Chapligin, personal communication). However, the difficulties of purifying the material and the extraction of formation water oxygen means that the analytical reproducibility of diatom silica oxygen isotope ratios is generally quoted as ~0.2–0.3‰. This is significantly higher than for carbonate $\delta^{18}\text{O}$ which has a precision in the range of $\pm 0.05\text{‰}$ (for pure carbonate minerals) to $\pm 0.1\text{‰}$ (for authigenic carbonates) within sediments.

Diatom Oxygen Systematics

Changes in $\delta^{18}\text{O}_{\text{diatom}}$ can be used to infer changes in either the temperature or the $\delta^{18}\text{O}$ composition of the water in which they grew ($\delta^{18}\text{O}_{\text{water}}$; Figure 5). As outlined in the sections below, the controls on $\delta^{18}\text{O}_{\text{water}}$ will depend on the nature of the site for both marine and lacustrine environments.

Temperature

A number of water column and culture experiments have suggested a $\delta^{18}\text{O}_{\text{diatom}}$ –temperature relationship in lacustrine taxa of $\sim 0.2\text{‰} \text{ } ^\circ\text{C}^{-1}$ (Brandriss et al., 1998; Crespin et al., 2010; Dodd and Sharp, 2009; Moschen et al., 2005). The temperature dependence in marine taxa has not been as rigorously derived with measurements based on bulk rather than species or genera-specific samples. In the first of these studies, Juillet-

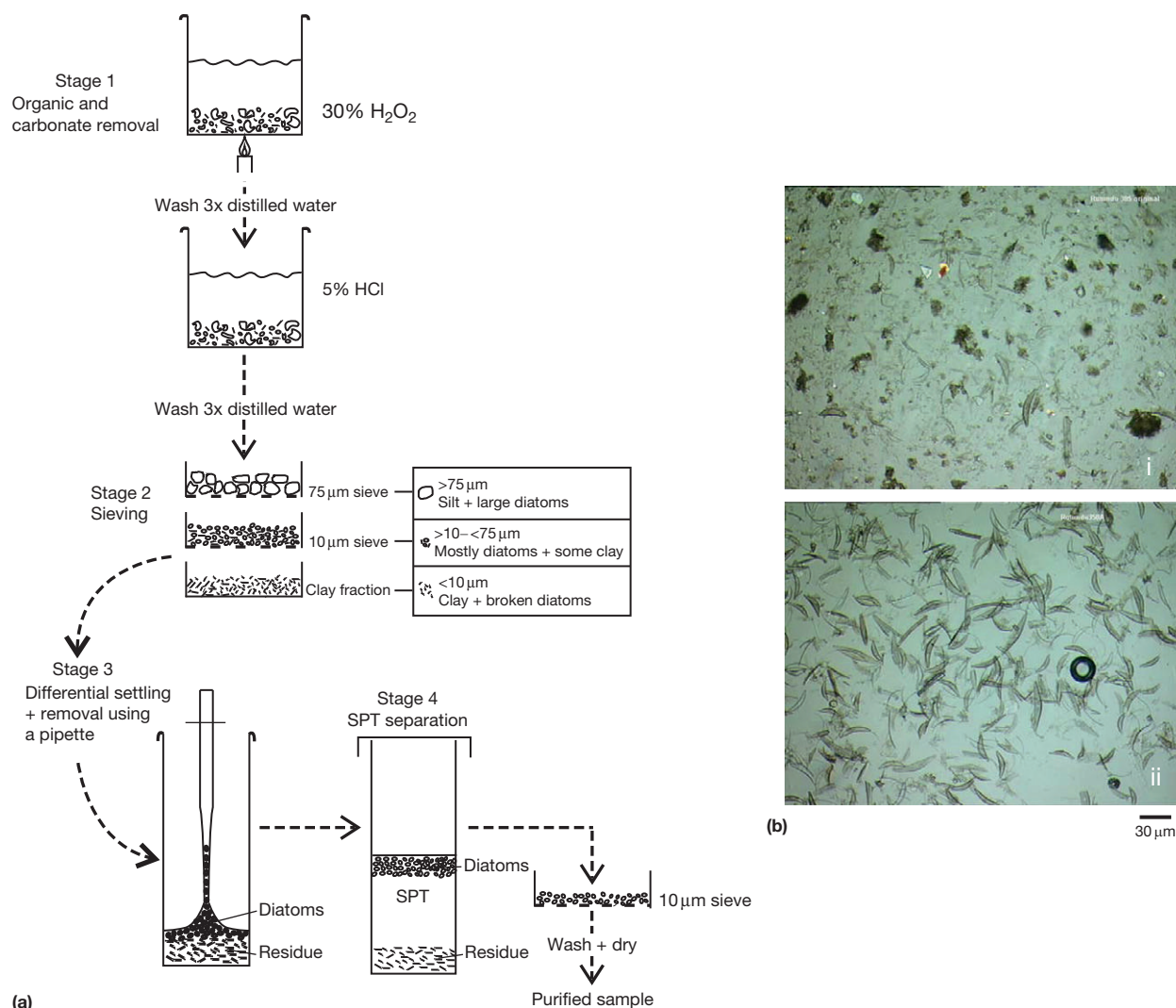


Figure 2 (a) Diagram showing the four-stage cleaning method for concentrating diatom for oxygen isotope analysis from lake sediments. SPT, Sodium polytungstate. Reproduced from Morley DW, Leng MJ, Mackay AW, Sloane HJ, Rioual P, and Battarbee RW (2004) Cleaning of lake sediment samples for diatom oxygen isotope analysis. *Journal of Palaeolimnology* 31: 391–401, with kind permission from Springer Science and Business Media. (b) Lake sediment samples from Lake Rutundu, Mount Kenya before and after cleaning, (i) shows the sample after standard diatom preparation using HCl and H_2O_2 and (ii) shows the cleaned sample after a staged cleaning approach. Species diversity has been reduced by the cleaning procedure, leaving the sample dominated by large *Cymbella* and *Epithemia* spp.

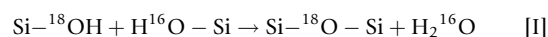
Leclerc and Labeyrie (1987) conducted a global calibration using surface-sediment assemblages to produce a calibration between sea surface temperatures (SST) of 1.5 and 24 °C:

$$T(^{\circ}\text{C}) = 17.2 - 2.4(\delta^{18}\text{O}_{\text{diatom}} - \delta^{18}\text{O}_{\text{water}} - 40) - 0.2(\delta^{18}\text{O}_{\text{diatom}} - \delta^{18}\text{O}_{\text{water}} - 40)^2 \quad [2]$$

Although this work was supported by Matheney and Knauth (1989), Shemesh et al. (1992) reported that the calibration significantly overestimates temperature in high-latitude sites. By selecting only the high-latitude data points from Juillet-Leclerc and Labeyrie (1987), Shemesh et al. (1992) suggested that a calibration of $-0.5\text{‰}^{\circ}\text{C}^{-1}$ was more appropriate for such regions. However, most work points to a coefficient of approximately $-0.2\text{‰}^{\circ}\text{C}^{-1}$, although further work is needed in marine systems to establish an accurate $\delta^{18}\text{O}_{\text{diatom}}$ -temperature calibration.

Silica Maturation

When analyzing $\delta^{18}\text{O}_{\text{diatom}}$ from sediments, consideration must be given to the role of silica maturation, often referred to as secondary isotope exchange, in which isotopically light ^{16}O from the $-\text{Si}-\text{OH}$ layer is released with ^{18}O forming isotopically enriched Q_4 bonds (Figure 1) in the $-\text{Si}-\text{O}-\text{Si}$ layer of the frustule (Brandriss et al., 1998; Dodd and Sharp, 2009; Moschen et al., 2005; Schmidt et al., 1997; 2001):



At present, the implications of these changes remain under investigation. On the one hand, silica maturation should erode any surface water $\delta^{18}\text{O}$ signal to produce a more muted record reflecting a combination of both surface and bottom water conditions. However, as outlined below, numerous studies

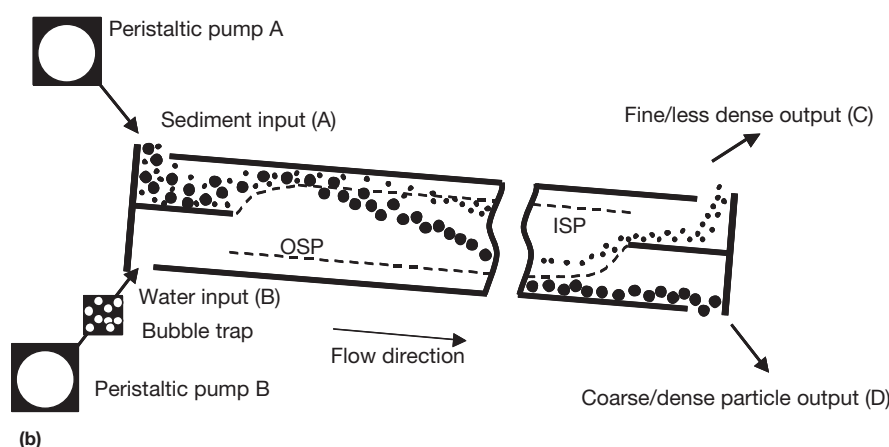
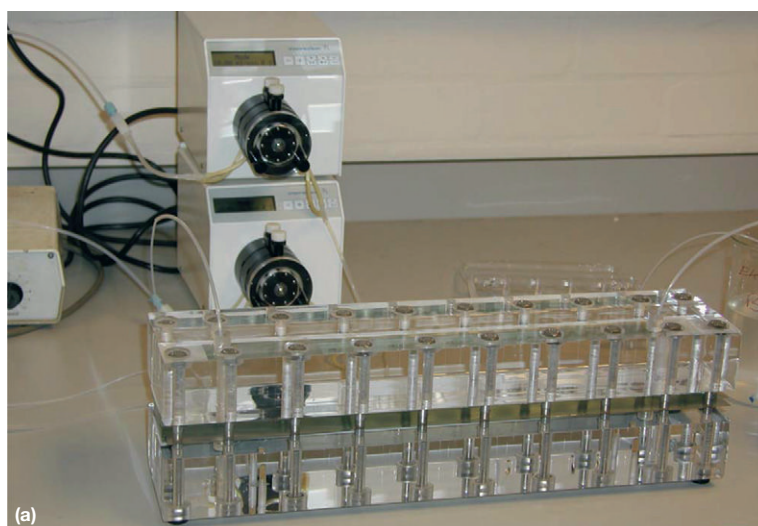


Figure 3 (a) An alternative approach to heavy liquid separation for cleaning diatom samples is gravitational split-flow thin fractionation (SPLITT) developed by J.C. Giddings at the University of Utah and first applied to the separation of diatoms by [Schleser et al. \(2001\)](#) and [Rings et al. \(2004\)](#). (b) SPLITT works by introducing a sample (A) into a thin channel where it meets a carrier fluid (B, usually water for diatom samples) in a laminar flow. The inlet splitter plane (ISP) and outlet splitter plane (OSP) divide the flow. The velocity of the two flows is controlled to separate the sample into two streams: an upper A sample containing finer, less dense, and more hydrodynamic particles passes through one outlet (C), while the remaining particles are collected through a lower outlet (D). The SPLITT cell is inclined to reduce deposition of sediment onto the splitter plate at low flows.

have demonstrated strong correlations between $\delta^{18}\text{O}_{\text{diatom}}$ and other surface water records with no evidence that silica maturation prevents the use of $\delta^{18}\text{O}_{\text{diatom}}$ to reconstruct surface water conditions ([Leng and Barker, 2006](#); [Leng and Swann, 2010](#)).

Lake reconstructions

Temperature

In the lacustrine environment, the interpretation of the measured oxygen isotope data from diatom silica in terms of paleotemperatures requires an understanding of two temperature effects that have opposing effects on the composition of diatom silica. There is the fractionation between diatom silica and water, described above, which has a gradient of $-0.2\text{‰}\text{°C}^{-1}$. There is also the relationship between precipitation and temperature (the so-called Dansgaard relationship $\delta p/dT$; [Dansgaard, 1964](#)) which is approximately $+0.6\text{‰}\text{°C}^{-1}$ at intermediate and high latitudes but has a worldwide range of between $+0.2$ and $+0.7\text{‰}\text{°C}^{-1}$ ([Dansgaard, 1964](#)). Ideally,

it is important to establish the relationship between the oxygen isotope composition of precipitation (δp) and temperature at each site (see [Leng et al. \(2005a\)](#) and the IAEA-WMO Global Network of Isotopes in Precipitation (GNIP) database); it is worth noting here that a Dansgaard relationship is not common in coastal or monsoonal regions or where rainfall comes from two air masses.

For many lake records, the diatom oxygen isotope response to temperature will be dominated by the change in the isotope composition of precipitation and effectively ‘damped’ by the opposing effect of mineral-water fractionation. For diatoms, the measured isotope composition will, therefore, covary with temperature – with an increase of $\sim 0.1\text{--}0.4\text{‰}\text{°C}^{-1}$ (cf. [Leng and Marshall, 2004](#)). This assumes that $\delta p/dT$ always changes according to the Dansgaard relationship. It is usually very difficult to derive quantitative paleotemperature/ δp data with any certainty. It might be more realistic to use $\delta^{18}\text{O}_{\text{diatom}}$ in areas where there are likely to have been significant changes in the isotope composition of the lake water due to changes in

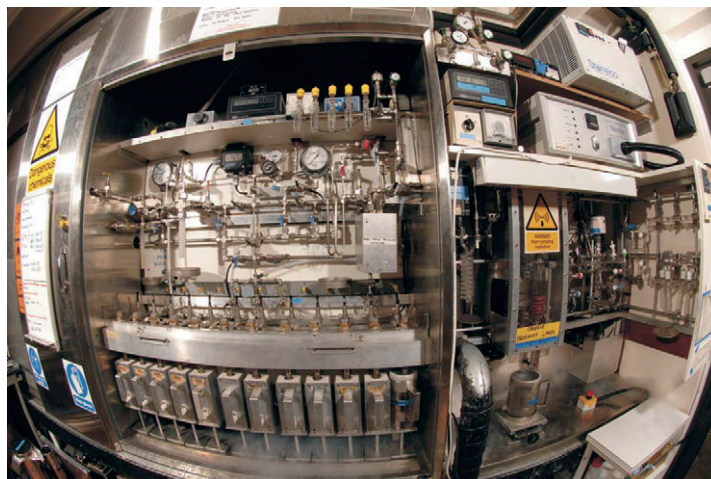
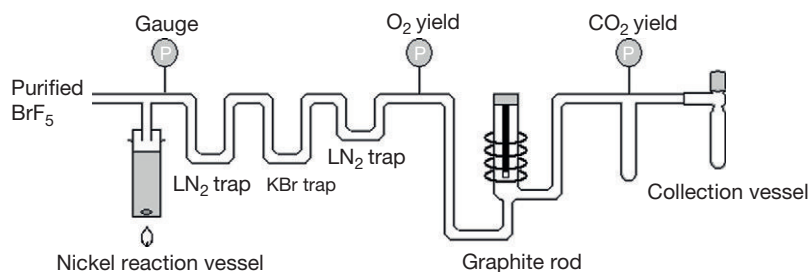


Figure 4 The fluorination line used at the NERC Isotope Geosciences Laboratory for the extraction of oxygen from diatom silica for isotope analysis was built by P.B. Greenwood.

the precipitation/evaporation balance or in the source of precipitation (see below). The change in $\delta^{18}\text{O}_{\text{diatom}}$ due to these factors is normally greater than temperature alone.

$\delta^{18}\text{O}_{\text{water}}$

Changes in the isotope composition of the lake water can be due to either evaporation or changes in the source of the rainfall. In closed (or terminal) lakes, evaporation of $\delta^{18}\text{O}_{\text{water}}$ will usually be far greater than changes due to temperature or δp . $\delta^{18}\text{O}_{\text{water}}$ depends on the balance between the isotope composition of inputs (including the source and amount of precipitation, surface runoff, and groundwater inflow) and outputs (evaporation and groundwater loss). Unless there is significant groundwater seepage, most closed lakes will lose water primarily through evaporation, the rate of which is controlled by wind speed, temperature, and humidity. The phase change of evaporation results in light isotopes of oxygen (^{16}O) being preferentially evaporated from water bodies leaving water that is relatively enriched in the heavier isotope (^{18}O). Any effects of varying precipitation source or temperature on $\delta^{18}\text{O}$ are often small in comparison to evaporative concentration and measured $\delta^{18}\text{O}$ (and δD) values become elevated above those of ambient precipitation. In extreme circumstances, evaporation in the lake catchment area or from the surface of a lake can lead to significantly elevated $\delta^{18}\text{O}_{\text{water}}$ values (Leng et al., 2005a).

The degree to which evaporation will increase $\delta^{18}\text{O}_{\text{water}}$ depends on the residence time of the lake (lake volume/throughput rate). Changes to a lake's residence time, caused

by changes in basin hydrology or varying groundwater fluxes, will also influence the degree of enrichment, as will changes in the nature of catchment vegetation and soils. These factors have been considered important in the interpretation of a $\delta^{18}\text{O}_{\text{diatom}}$ record from Lake Tilo in East Africa which is interpreted as changes in aridity (Lamb et al., 2005). They showed that high temperatures and a constant supply of solutes from hydrothermal springs enabled diatoms to grow throughout the year. The early Holocene lake level was close to the crater rim with a slow but progressive fall through the early Holocene. Major changes are recorded simultaneously by $\delta^{18}\text{O}_{\text{diatom}}$ and $\delta^{18}\text{O}_{\text{calcite}}$, although larger errors are associated with $\delta^{18}\text{O}_{\text{diatom}}$, and especially noteworthy is that the $\delta^{18}\text{O}_{\text{diatom}}$ data do not record the well-documented 7.8 ka BP arid interval found in neighboring Ethiopian lakes.

Certain diatom records have been shown to be sensitive to other aspects of climate, such as amount or source of precipitation. For example, abrupt shifts of up to 18‰ in $\delta^{18}\text{O}_{\text{diatom}}$ have been found in a 14 kyr-long record from two Alpine lakes on Mount Kenya (Barker et al., 2001), which cannot be entirely temperature related given the current knowledge of the diatom-temperature fractionation. Instead, the variations have been interpreted as a moisture balance effect related to changes in Indian Ocean SST through the Holocene. Episodes of heavy convective precipitation in the $\delta^{18}\text{O}_{\text{diatom}}$ record correlate with enhanced alkenone-based SST estimates.

There have been several studies of $\delta^{18}\text{O}_{\text{diatom}}$ from lakes in Northern Europe (see Jonsson et al., 2010; Leng and Barker, 2006). These lakes have similarities, in that many are open,

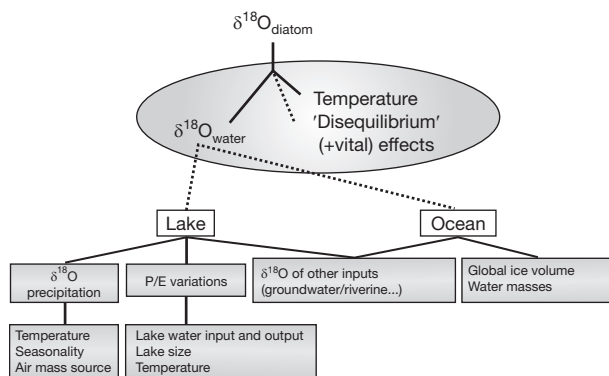


Figure 5 Controls on the oxygen isotope composition of diatom silica ($\delta^{18}\text{O}_{\text{diatom}}$). If precipitated in isotopic equilibrium, $\delta^{18}\text{O}_{\text{diatom}}$ depends entirely on temperature and the isotope composition of the lake water ($\delta^{18}\text{O}_{\text{water}}$). Disequilibrium effects, commonly known as 'vital effects' in biogenic precipitates, caused by local changes in microenvironment or rate of precipitation can induce systematic or nonsystematic offsets in the $\delta^{18}\text{O}_{\text{diatom}}$ signal. Interpretation of $\delta^{18}\text{O}_{\text{diatom}}$ in terms of temperature depends on knowledge of the environment and time of year in which the diatom silica forms. For example, at high-latitude sites, diatoms typically grow in the summer months in the surface waters. Quantitative interpretation similarly demands an understanding of the local controls on the isotope composition of the ambient water ($\delta^{18}\text{O}_{\text{water}}$). In marine localities, changes in $\delta^{18}\text{O}_{\text{water}}$ are primarily driven by changes in global ice volume, ocean currents, and meltwater inputs. In some lakes, there is a simple relationship to $\delta^{18}\text{O}_{\text{precipitation}}$, but in others the water composition is strongly influenced by processes such as evaporation within the catchment and within the lake itself. $\delta^{18}\text{O}_{\text{precipitation}}$ is increasingly being shown to be an important indicator of climate change: it typically changes with mean annual temperature, but nonlinear responses to climate change are increasingly being recognized. See text for more discussion of this figure. Adapted from Leng MJ and Barker PA (2006) A review of the oxygen isotope composition of lacustrine diatom silica for palaeoclimate reconstruction. *Earth Science Reviews* 75: 5–27.

through-flow systems with minimal evaporation. Changes in the stratigraphic record of $\delta^{18}\text{O}_{\text{diatom}}$ are interpreted as changes in the summer isotope composition of the lake water. In Northern Sweden, a 5000-year $\delta^{18}\text{O}_{\text{diatom}}$ record from a proglacial lake showed a general lowering trend of 3.5‰ with spikes of >1‰ and a ~1000-year cyclicity associated with sedimentary proxies for glacial advance. The authors suggest that a persistent change in the atmospheric circulation pattern could potentially have caused the changes in $\delta^{18}\text{O}_{\text{diatom}}$ because different air masses influence this area and these air masses have characteristic $\delta^{18}\text{O}$ signatures of their precipitation. They infer a steady increase but fluctuating dominance of colder and $\delta^{18}\text{O}$ -depleted air masses from the north/northeast. The $\delta^{18}\text{O}_{\text{diatom}}$ depletion and glacier events occur at times of relative ice rafted debris maxima in the North Atlantic, consistent with cold conditions and changes in surface wind directions. The Northern European studies show some similarities in the diatom isotope records, in particular, a late Holocene cooling, although there are regional differences, notably the relatively cold early Holocene on the Kola Peninsula is not seen in the records from Northern Sweden.

Changes in the source of precipitation have also been suggested as one reason for variations in $\delta^{18}\text{O}_{\text{diatom}}$ in a small,

spring fed lake in the Central Mexican highlands (Leng et al., 2005b). The lake has become more shallow, and today is at one of its lowest lake levels, but the modern isotope data suggest that the modern water is not evaporated, indicating that hydrology or recharge rates, rather than evaporation is controlling lake level. Two main possibilities might have controlled variations in $\delta^{18}\text{O}$ in the past. Since Central Mexico gets rainfall that originates from the tropical Atlantic/Gulf of Mexico and also from both the tropical and extratropical Pacific (in summer and winter, respectively), the authors postulate that changes in $\delta^{18}\text{O}_{\text{precipitation}}$ are connected to temperature and salinity variations in the Gulf of Mexico and/or changing contributions from these two different air masses (Gulf of Mexico vs. Pacific-derived hurricanes).

There are only a few studies of oxygen isotope ratios in lacustrine diatom silica that are older than the Holocene; these include interglacial investigations in France (Rioul et al., 2001), investigations into the MIS 11 period in Lake Baikal (Mackay et al., 2008), and the late-glacial–Holocene sequence from Lake El'gygytyn in North East Siberia (Swann et al., 2010).

Marine reconstructions

In the marine environment, records of $\delta^{18}\text{O}_{\text{diatom}}$ can be used in a similar way to that of planktonic Foraminifera; $\delta^{18}\text{O}$ ($\delta^{18}\text{O}_{\text{foram}}$) being sensitive to changes in both temperature and the $\delta^{18}\text{O}_{\text{water}}$ of surrounding water, a function of ocean currents/masses and riverine/continental/meltwater inputs. Accordingly, $\delta^{18}\text{O}_{\text{diatom}}$ can be employed as a direct replacement for $\delta^{18}\text{O}_{\text{foram}}$ in oceanic regions that are depleted in CaCO_3 and so do not preserve calcareous Foraminifera. Planktonic Foraminifera are capable of living at depths of hundreds of meters and, therefore, tend to have an isotopic signature typical of deep water, whereas diatoms are restricted to the surface ocean because of their need to photosynthesize, and, therefore, reflect isotopic conditions from the uppermost layers of the surface ocean.

Existing records of marine $\delta^{18}\text{O}_{\text{diatom}}$ primarily originate from high-latitude regions, such as the Southern Ocean, where Foraminifera are generally not well preserved in the sedimentary record. However, in one early study, records of $\delta^{18}\text{O}_{\text{diatom}}$ were obtained from the Gulf of California over the last 3 ka (Juillet-Leclerc and Schrader, 1987). By using modern-day oceanographic data to account for localized changes in $\delta^{18}\text{O}_{\text{water}}$, $\delta^{18}\text{O}_{\text{diatom}}$ was used to detect SST changes of up to 8 °C. In conjunction with taxonomic studies, cold SST were linked to changes in regional upwelling intensity, in turn tied to atmospheric patterns related to El Niño events. These results indicate a period of enhanced upwelling and intensified trade winds in North Pacific Ocean and reduced El Niño activity between 1.5 and 2.0 ka BP. Subsequent work focusing on the last 100 years has verified these findings by revealing a strong relationship between $\delta^{18}\text{O}_{\text{diatom}}$ and historical records of El Niño events (Juillet-Leclerc et al., 1991).

Combined $\delta^{18}\text{O}_{\text{diatom}}/\delta^{18}\text{O}_{\text{foram}}$ studies

Where both diatoms and Foraminifera co-occur, $\delta^{18}\text{O}_{\text{diatom}}$ and $\delta^{18}\text{O}_{\text{foram}}$ can be combined to provide unique insights

into paleoceanographic conditions. For example, Shemesh et al. (1992) combined $\delta^{18}\text{O}_{\text{diatom}}$ with planktonic $\delta^{18}\text{O}_{\text{foram}}$ to account for variations in $\delta^{18}\text{O}$ that arose from changes in global ice volume. By assuming that the incorporation of oxygen into the Foraminifera calcite shell and diatom frustule occurred in the same season and water depth and using $\delta^{18}\text{O}$ –temperature coefficients for diatoms and Foraminifera, an $\sim 2.0^\circ\text{C}$ SST increase was observed in Southern Ocean surface waters during the shift from glacial to Holocene conditions with a concordant 1.2‰ decrease in $\delta^{18}\text{O}_{\text{water}}$. Records of $\delta^{18}\text{O}_{\text{diatom}}$ and planktonic $\delta^{18}\text{O}_{\text{foram}}$ have also been combined in the subarctic North West Pacific Ocean during the late Pliocene (Haug et al., 2005). On the basis of modern-day data, the Foraminifera were assumed to have lived and precipitated their shells during the spring months, whereas the diatoms were primarily blooming during autumn/early winter (Swann et al., 2006). Divergent trends in the two records at the intensification of Northern Hemisphere Glaciation (~ 2.75 – 2.73 Ma) are, therefore, interpreted as reflecting interseasonal conditions with increases in $\delta^{18}\text{O}_{\text{foram}}$ suggesting a significant, 7.5°C , cooling of the spring surface waters while a 4.6‰ decrease in $\delta^{18}\text{O}_{\text{diatom}}$ equated to a significant freshening and warming of the autumn surface waters related to the development of a halocline (salinity driven, stratification boundary).

Meltwater

Other records of $\delta^{18}\text{O}_{\text{diatom}}$ have documented notable changes in surface water $\delta^{18}\text{O}$ arising from meltwater influx. Results from the Atlantic and Indian sectors of the Southern Ocean have demonstrated periodic decreases of ~ 1 ‰ in the Holocene, changes of up to 2–3‰ during the last glacial and 5‰ variations over the last 430 000 years (Shemesh et al., 1994, 1995, 2002). Relating these changes to temperature contradicts other proxy records for temperature from this region. Instead, the periodic marked decreases in $\delta^{18}\text{O}_{\text{diatom}}$ are thought to be connected to meltwater influxes originating from Antarctica, releasing water containing an inherently lower $\delta^{18}\text{O}$ than seawater (Shemesh et al., 1995). While these studies highlight the potential sensitivity of the Southern Ocean to meltwater fluxes from Antarctica, it is not currently possible to fully understand the timing, mechanism, or implication of these meltwater releases due to the low resolution nature of the $\delta^{18}\text{O}_{\text{diatom}}$ records.

Records from the North Pacific Ocean and Bering Sea have also shown significant changes in meltwater influx. Using mixed species samples, Sancetta et al. (1985) detected 5–10‰ variations in the Bering Sea over the last glacial cycle. By using changes in the sedimentary diatom assemblages to distinguish between changes in SST and salinity, the isotope variations were partitioned to reflect changes in both sea-ice extent as well as meltwater influx with a 1.9‰ equivalent increase in SST between the last glacial and the present day. Further south in the subarctic North West Pacific Ocean, large changes of ~ 4 ‰ are apparent over a series of glacial–interglacial cycles during the late Pliocene and Early Quaternary (Swann, 2010). By accounting for other processes that may alter $\delta^{18}\text{O}_{\text{diatom}}$ such as temperature, changes in global ice volume, and freshwater inputs from other sources, these changes were related to meltwater input from glacials surrounding the North Pacific and Bering Sea basins over similar timescales.

Conclusions

Since fluorination and other analytical techniques will liberate oxygen from all the components in the sediment, analysis of the oxygen isotope composition of diatom silica requires samples that are almost pure diatomite. There is a generally acceptable protocol involving chemistry, sieving and settling techniques, and more recently laminar flow separation for extracting clean diatoms for isotope analysis. However, all samples require a dedicated procedure and every sample must be scrutinized with microscopy. Diatom silica analysis requires the removal of the outer hydrous layer before a reaction to break the tetrahedrally bonded silica skeleton (Si—O—Si). To date, there are four main methods (both thermal dehydration, isotope exchange, and stepped fluorination techniques) to do this.

Diatom $\delta^{18}\text{O}$ in sediments can be an extremely useful indicator of paleoenvironmental change and tends to be most successful in areas where there are likely to have been large changes in the isotope composition of the host water. In lakes, this can be due to changes in the precipitation/evaporation balance, or source of precipitation, and in oceans it can be due to variations in meltwater flux. The change in $\delta^{18}\text{O}_{\text{diatom}}$ due to these factors is normally greater than temperature and methodological errors.

As with all isotope studies, the use of $\delta^{18}\text{O}_{\text{diatom}}$ in paleoenvironmental studies should include a study of the contemporary environment and the modern diatom silica. The presence of a systematic relationship between temperature, water isotope composition, and diatom isotope composition for the host water enables quantitative interpretation of the sediment record. It is particularly useful to know when and where the diatom silica ‘bloom’ occurs in the modern environment, so details of diatom ecology are important. Modern studies will tell us whether the diatom record reflects specific seasons or year-round conditions.

See also: Diatom Introduction. Paleobotany: Silicon Isotopes in Diatoms.

References

- Barker PA, Street-Perrott FA, Leng MJ, et al. (2001) A 14 ka oxygen isotope record from diatom silica in two alpine tarns on Mt Kenya. *Science* 292: 2307–2310.
- Brandriss ME, O’Neil JR, Edlund MB, and Stoermer EF (1998) Oxygen isotope fractionation between diatomaceous silica and water. *Geochimica et Cosmochimica Acta* 62: 1119–1125.
- Chapligin B, Meyer H, Friedrichsen H, Marent A, Sohns E, and Hubberten H-W (2010) A high-performance, safer and semi-automated approach for the $\delta^{18}\text{O}$ analysis of diatom silica and new methods for removing exchangeable oxygen. *Rapid Communications in Mass Spectrometry* 24: 2655–2664.
- Crespin J, Sylvestre F, Alexandre A, Sonzogni C, Paillès C, and Perga ME (2010) Re-examination of the temperature-dependent relationship between $\delta^{18}\text{O}_{\text{diatoms}}$ and $\delta^{18}\text{O}_{\text{lake water}}$ and implications for paleoclimate inferences. *Journal of Paleolimnology* 44: 547–557.
- Dansgaard W (1964) Stable isotopes in precipitation. *Tellus* 16: 436–468.
- Dodd JP and Sharp ZD (2009) A laser fluorination method for oxygen isotope analysis of biogenic silica and a new oxygen isotope calibration of modern diatoms in freshwater environments. *Geochimica et Cosmochimica Acta* 74: 1381–1390.
- Haimson M and Knauth LP (1983) Stepwise fluorination – A useful approach for the isotopic analysis of hydrous minerals. *Geochimica et Cosmochimica Acta* 47: 1589–1595.

- Haug GH, Ganopolski A, Sigman DM, et al. (2005) North Pacific seasonality and the glaciation of North America 2.7 million years ago. *Nature* 433: 821–825.
- Jonsson CE, Andersson S, Rosqvist GC, and Leng MJ (2010) Reconstructing past atmospheric circulation changes using oxygen isotopes in lake sediments from Sweden. *Climate of the Past* 6: 46–62.
- Juillet-Leclerc A and Labeyrie L (1987) Temperature dependence of the oxygen isotopic fractionation between diatom silica and water. *Earth and Planetary Science Letters* 84: 69–74.
- Juillet-Leclerc A, Labeyrie LD, Reyss JL, and Schrader H (1991) Temperature variability in the Gulf of California during the last Century: A record of the recent strong El Niño. *Geophysical Research Letters* 18: 1889–1892.
- Juillet-Leclerc A and Schrader H (1987) Variations of upwelling intensity recorded in varved sediment from the Gulf of California during the past 3,000 years. *Nature* 329: 146–149.
- Labeyrie LD and Juillet A (1982) Oxygen isotopic exchangeability of diatom valve silica: Interpretation and consequences for paleoclimatic studies. *Geochimica et Cosmochimica Acta* 46: 967–975.
- Lamb AL, Leng MJ, Sloane HJ, and Telford RJ (2005) A comparison of the paleoclimate signals from diatom oxygen isotope ratios and carbonate oxygen isotope ratios from a low latitude crater lake. *Palaeogeography, Palaeoclimatology, Palaeoecology* 223: 290–302.
- Leng MJ and Barker PA (2006) A review of the oxygen isotope composition of lacustrine diatom silica for palaeoclimate reconstruction. *Earth-Science Reviews* 75: 5–27.
- Leng MJ, Lamb AL, Heaton THE, et al. (2005a) Isotopes in lake sediments. In: Leng MJ (ed.) *Isotopes in Palaeoenvironmental Research*, pp. 147–184. Dordrecht: Springer.
- Leng MJ, Metcalfe SE, and Davies SJ (2005b) Investigating late Holocene climate variability in Central Mexico using carbon isotope ratios in organic materials and oxygen isotope ratios from diatom silica within lacustrine sediments. *Journal of Paleolimnology* 34: 413–431.
- Leng MJ and Marshall JD (2004) Palaeoclimate interpretation of stable isotope data from lake sediment archives. *Quaternary Science Reviews* 23: 811–831.
- Leng MJ and Sloane HJ (2008) Combined oxygen and silicon isotope analysis of biogenic silica. *Journal of Quaternary Science* 23: 313–319.
- Leng MJ and Swann GEA (2010) Stable isotopes in diatom silica. In: Smol JP and Stoermer EF (eds.) *The Diatoms: Applications for the Environmental and Earth Sciences*, p. 667. Cambridge: Cambridge Press.
- Lücke A, Moschen R, and Schleser GH (2005) High temperature carbon reduction of silica: A novel approach for oxygen isotope analysis of biogenic opal. *Geochimica et Cosmochimica Acta* 69: 1423–1433.
- Mackay AW, Karabanov E, Leng MJ, et al. (2008) Reconstructing hydrological variability in Lake Baikal during MIS 11: An application of oxygen isotope analysis of diatom silica. *Journal of Quaternary Science* 23: 365–374.
- Matheney RK and Knauth LP (1989) Oxygen-isotope fractionation between marine biogenic silica and seawater. *Geochimica et Cosmochimica Acta* 53: 3207–3214.
- Moschen R, Lucke A, and Schleser GH (2005) Sensitivity of biogenic silica oxygen isotopes to changes in surface water temperature and palaeoclimatology. *Geophysical Research Letters* 32: L07708. <http://dx.doi.org/10.1029/2004GL022167>.
- Rings A, Lucke A, and Schleser GH (2004) A new method for the quantitative separation of diatom frustules from lake sediments. *Limnology and Oceanography: Methods* 2: 25–34.
- Rioual P, Andrieu-Ponel V, Battarbee RW, et al. (2001) High-resolution record of climate stability in France during the last interglacial period. *Nature* 413: 293–296.
- Sancetta C, Heusser L, Labeyrie L, Sathy NA, and Robinson SW (1985) Wisconsin–Holocene paleoenvironment of the Bering Sea: Evidence from diatoms, pollen, oxygen isotopes and clay minerals. *Marine Geology* 62: 55–68.
- Schiff C, Kaufman DS, Wolfe AP, Dodd J, and Sharp Z (2009) Late Holocene storm-trajectory changes inferred from the oxygen isotope composition of lake diatoms, south Alaska. *Journal of Paleolimnology* 41: 189–208.
- Schleser GH, Lücke A, Moschen R, et al. (2001) Separation of diatoms from sediment and oxygen isotope extraction from their siliceous valves – A new approach. *Terra Nostra*, 2001/3, *Schriften der Alfred-egener-tiftung* (6th Workshop of the European Lake Drilling Programme, Potsdam) 187–191.
- Schmidt M, Botz R, Rickert D, Bohrmann G, Hall SR, and Mann S (2001) Oxygen isotopes of marine diatoms and relations to opal – A maturation. *Geochimica et Cosmochimica Acta* 65: 201–211.
- Schmidt M, Botz R, Stoffers P, Anders T, and Bohrmann G (1997) Oxygen isotopes in marine diatoms: A comparative study of analytical techniques and new results on the isotope composition of recent marine diatoms. *Geochimica et Cosmochimica Acta* 61: 2275–2280.
- Shemesh A, Burckle LH, and Hays JD (1994) Meltwater input to the Southern Ocean during the Last Glacial Maximum. *Science* 266: 1542–1544.
- Shemesh A, Burckle LH, and Hays JD (1995) Late Pleistocene oxygen isotope records of biogenic silica from the Atlantic sector of the Southern Ocean. *Paleoceanography* 10: 179–196.
- Shemesh A, Charles CD, and Fairbanks RG (1992) Oxygen isotopes in biogenic silica: Global changes in ocean temperature and isotopic composition. *Science* 256: 1434–1436.
- Shemesh A, Hodell D, Crosta C, Kanfoush S, Charles C, and Guilderson T (2002) Sequence of events during the last deglaciation in Southern Ocean sediments and Antarctic ice cores. *Paleoceanography* 17: 1056. <http://dx.doi.org/10.1029/2000PA000599>.
- Swann GEA (2010) Salinity changes in the North West Pacific Ocean during the late Pliocene/early Quaternary from 2.73 Ma to 2.52 Ma. *Earth and Planetary Science Letters* 297: 332–338.
- Swann GEA, Leng MJ, Juschus O, Melles M, Brigham-Grette J, and Sloane HJ (2010) A combined oxygen and silicon diatom isotope record of Late Quaternary change in Lake El'gygytyn, North East Siberia. *Quaternary Science Reviews* 29: 774–789.
- Swann GEA, Leng MJ, Sloane HJ, and Maslin MA (2008) Isotope offsets in marine diatom $\delta^{18}\text{O}$ over the last 200 ka. *Journal of Quaternary Science* 23: 389–400.
- Swann GEA, Leng MJ, Sloane HJ, Maslin MA, and Onodera J (2007) Diatom oxygen isotope vital effects: Evidence from the palaeo record. *Geochemistry, Geophysics, Geosystems* 8: Q06012.
- Swann GEA, Maslin MA, Leng MJ, Sloane HJ, and Haug GH (2006) Diatom $\delta^{18}\text{O}$ evidence for the development of the modern halocline system in the subarctic northwest Pacific at the onset of major Northern Hemisphere glaciation. *Paleoceanography* 21: PA1009. <http://dx.doi.org/10.1029/2005PA001147>.

Data Interpretation

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Introduction

Diatoms are single-celled, microscopic algae (Bacillariophyta) whose siliceous, glass-like cell walls preserve well in various sedimentary environments allowing reconstructions of past environments from subfossil diatom assemblages. They are a very successful group of organisms that can be found in virtually every aquatic habitat. Accordingly, diatom analysis is applicable across a wide range of aquatic ecosystem types, including freshwater, brackish, estuarine, and ocean, and includes lentic and lotic environments, wetlands and their associated damp, marginal, and littoral zones. This is why diatoms are one of the most widely used biological indicators in Quaternary paleoecology (Stoermer and Smol, 1999; see Overview).

Diatoms have played an important role in providing information on a wide spectrum of past environmental and ecological changes, such as climatic and lake-level changes, water pollution, eutrophication, and acidification, while their usefulness is also demonstrated in archaeology, oil and gas exploration, and forensic science (Stoermer and Smol, 1999 and Battarbee et al., 2001; see Use in Archaeology). Most studies using fossil diatoms as indicators of environmental change have focused on the Late Glacial and Holocene sequences, although long diatom records encompassing several glacial-interglacial cycles are available from marine and freshwater sequences (e.g., see North Atlantic and Arctic; Large Lakes) and offer great potential for a very long-term perspective of environmental change. The standard method of inferring environmental conditions from fossil diatoms consists of an analysis of assemblage composition and consideration of the relevant autecological characteristics of the taxa present (indicator species). In recent decades, tremendous progress has been made in our methods of interpreting diatom stratigraphical data through powerful numerical techniques that allow surprisingly accurate quantitative reconstructions of environmental change to be made.

Why Are Diatoms Good Paleoindicators?

There are many reasons why diatoms are such excellent paleoindicators in Quaternary studies (Dixit et al., 1992 and Stoermer and Smol, 1999):

- Diatoms comprise a large proportion of total algal biomass in various aquatic systems and their assemblages are typically taxon rich. The large number of taxa enhances the ecological information gained, and hence increases the confidence of environmental inferences.
- They collectively show a broad range of tolerance along environmental and ecological gradients, with individual

species having specific ecological and water quality requirements that are usually quantifiable.

- They have one of the shortest generation times of all biological indicators (~2 weeks), for which reason they reproduce and respond rapidly to environmental change.
- They are characterized by having an opaline, siliceous frustule that is resistant to degradation and hence is well preserved in different sedimentary environments.
- The taxonomy of diatoms is comprehensively documented, allowing species or even subspecies level identification of fossil diatom assemblages.

Analytic and Interpretative Procedures

Diatom Concentrations and Accumulation Rates

A prerequisite for the correct interpretation of stratigraphic diatom data is the careful collection, processing, and analysis of the initial diatom samples. Sampling and laboratory methods are discussed in detail by Battarbee et al. (2001). Diatom counts from sediments are most often expressed as percentage values, and paleoecologists generally base their environmental reconstructions on the assumption (that may not always be valid) that the percentage composition of diatoms in sediment assemblages accurately reflects the relative abundance of different taxa across the different living communities in the lake. Such proportional data lend themselves well to most numerical data treatment procedures and quantitative approaches. Estimation of diatom concentrations per unit weight (frustules per gram dry weight⁻¹) or unit volume (frustules cm⁻³) can be used as a record of diatom paleoproductivity and possibly to indicate net primary productivity. If diatom concentrations are required, an easy method is to add a known quantity of external markers (e.g., microspheres) in cleaned sediment slurry (Battarbee et al., 2001). Diatom concentration values *per se* have rather limited use in paleoecology, but they can be further converted to diatom accumulation rates (DAR; frustules per cm² per year, also referred to as diatom valve flux; Fig. 1). However, an excellent chronology is needed for this.

One has to bear in mind that different factors such as diatom preservation status (i.e., dissolution and diatom fragmentation during grazing or sample preparation) as well as potential variations in the rate of sediment accumulation, make interpretation of diatom concentration results more-or-less tentative. Also, because diatom cell sizes vary tremendously, simple concentrations may produce a misleading depiction of total diatom productivity. To minimize such biases, species-specific correction factors can be developed that allow the composition of the source populations to be reconstructed. Also, diatom concentrations can be converted into biovolumes per cm per year, using taxon-specific measurements of whole-cell volumes. Surprisingly, the latter approach

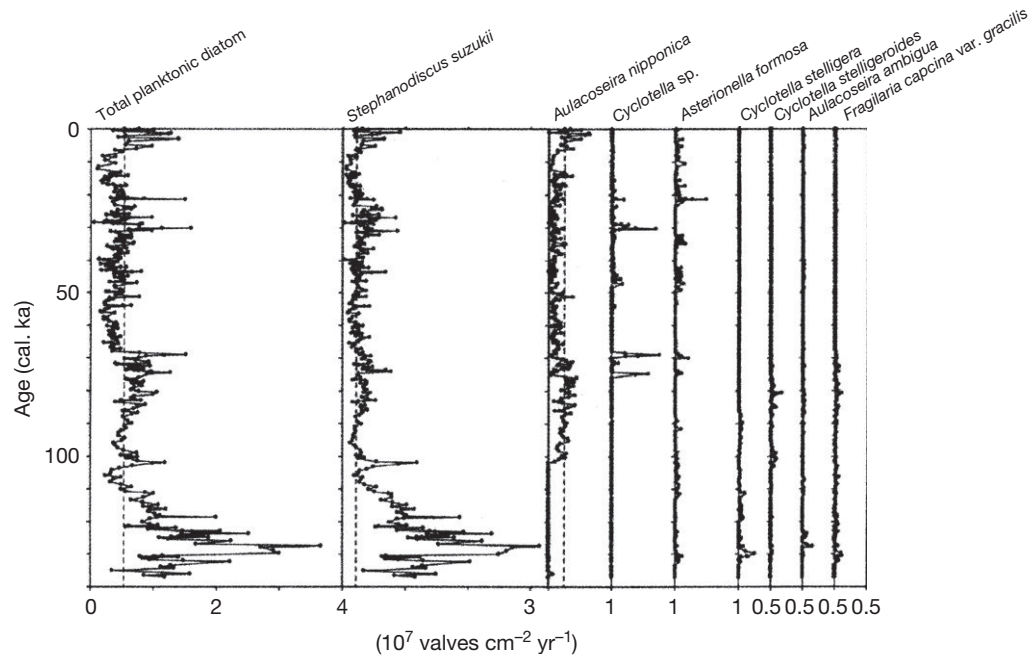


Figure 1 Records of total planktonic diatom flux (valves $\text{cm}^{-2} \text{yr}^{-1}$) and most important individual species fluxes from a 140 m long core taken from Lake Biwa, Shiga Prefecture, Japan, covering ca. 140 ka. The dashed line indicates the present level of diatom flux for the total planktonic taxa and each individual taxa, respectively. Based on characteristics of past and modern diatom responses to possible climate variables, the authors interpreted changes in total planktonic diatom flux to reflect a combination of summer precipitation, winter temperature, and snowfall values. During the 140–101 ka interval, high values of *Stephanodiscus suzukii* productivity and the disappearance of *Aulacoseira nipponica* were interpreted to indicate weaker vertical mixing, possibly caused by increased temperatures and decreased snowfall levels in winter. During the 101–70 ka interval, the record showed levels near or above those observed in present times, indicating that winter water temperatures fell within the optimal range for *A. nipponica* to prosper. Low *Aulacoseira* fluxes during the 70–7 ka interval were interpreted to indicate weak winter vertical mixing and cold winters. From Kuwae M, Yoshikawa S, Tsugeki N, and Inouchi Y (2004) Reconstruction of a climate record for the past 140 ka based on diatom valve flux data from Lake Biwa, Japan. *Journal of Paleolimnology* 32: 19–39.

has rarely been implemented in fossil data, despite the fact that algal biovolumes have long been a standard and routine unit of measure in modern phytoplankton ecology. The concentration estimates can be further reinforced by multiple core and multiple indicator analyses. In conjunction with diatom studies, analysis of fossil pigments, organic carbon and chemical determination of biogenic silica may help reconstruct the biomass of diatom populations in a more reliable way. For example, Abrantes (2000) compared DAR records from several Atlantic upwelling sites with other independent productivity proxies such as silicoflagellates, radiolaria, total opal, and abundances and accumulation rates of organic carbon and benthic Foraminifera, and suggested that maximum productivity of the past 200 ka in the eastern Atlantic occurred at the Last Glacial Maximum (LGM), consistent with the models that attribute glacial atmospheric CO_2 decrease to higher low-latitude productivity.

Ecological and Habitat Classification – Qualitative Approaches

Despite the rapid development of quantitative reconstruction techniques during the past few decades, qualitative interpretations based on the relevant autecological characteristics of the individual taxa and the habitat and ecological classification of

groups of taxa still form the basis for understanding and interpreting the stratigraphic diatom record. Many diatom species have clearly defined, and relatively narrow, preferences for particular habitats and water quality conditions. These species are indicative of several variables, including nutrient availability, acidity, light and thermal conditions, dissolved oxygen concentrations, and lake levels. As a result, a broad range of specific paleoenvironmental interpretations can be made on the basis of the life histories, habitats, and known ecological preferences of indicator species. Interpretation of diatom records is fortunate in this respect, because there is a considerable amount of tested species ecological data available. In particular, categorization of diatom taxa according to their acidity or salinity tolerances has been successfully used to interpret changes in surface water acidity, sea level, and hydrology of saline lakes (see, e.g., Palmer and Abbott (1986) and Stoermer and Smol (1999)).

Diatom species also have affinities for particular physical habitats for example there are open water (planktonic) and benthic forms. See [Diatom Introduction](#) for more details. In stratigraphic studies dealing with dynamic coastal environments, classification of diatoms according to habitat has also proved very useful since there are distinct communities associated with the different salt-marsh habitats ([Fig. 2](#)).

Qualitative interpretations based on expert knowledge have both advantages and disadvantages. Qualitative approaches

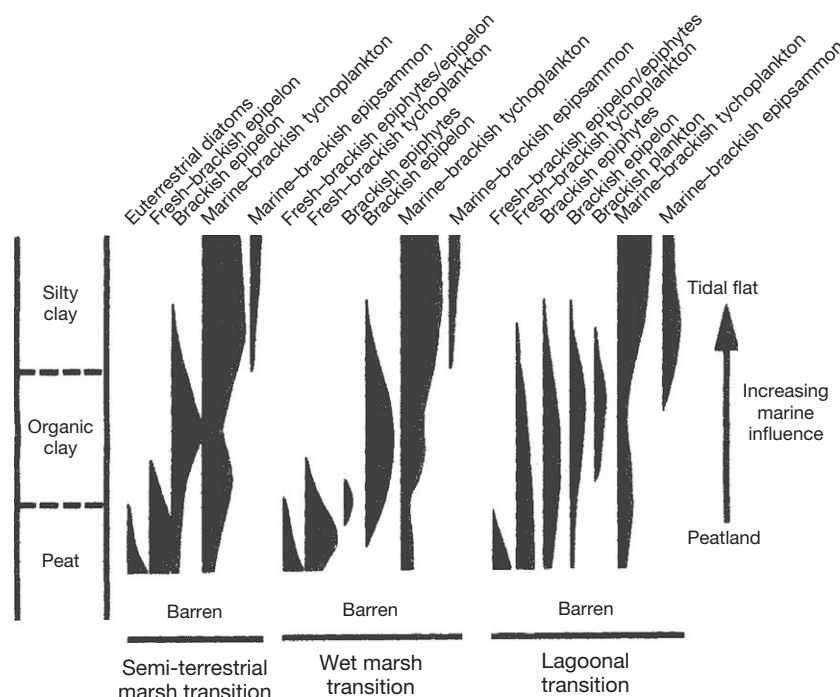


Figure 2 Idealized scheme of some possible diatom assemblage shifts occurring at the transgressive overlap of a peat with tidal deposits under different drainage conditions. Accuracy of paleotidal inference decreases to the right. Note that the relative representation of different groups will vary according to local conditions and input of external (allochthonous) material. From Denys L and De Wolf H (1999) *Diatoms as indicators of coastal paleoenvironmental and relative sea-level change*. In: Stoermer EF and Smol JP (eds.) *The Diatoms: Applications for the Environmental and Earth Sciences*, pp. 277–297. Cambridge: Cambridge University Press.

may pursue a variety of theory-generating aims, including identifying potentially important variables or concepts, recognizing patterns and relationships, and generating coherent theories and hypotheses. However, qualitative studies rely heavily on subjective interpretation and expert knowledge. They do not typically generate answers but rather provide narrative accounts.

Toward More Objective Assessment

A great variety of classification and descriptive multivariate tools have been developed that can be used to more effectively manage the rich mosaic of ecological information inherent in diatom stratigraphic data. Typically, an important first step is the partitioning or zoning of the initial diatom diagram with the aim of identifying uniform segments in the data. A favored method for paleoecological data has been the constrained incremental sum-of-squares clustering (CONISS; Grimm, (1987)). However, it suffers from having no numerical basis for deciding on the number of reliable or statistically significant zones within a sequence. Bennett (1996) has resolved the problem by using the broken-stick and randomization models to assess the reliability of different zonations against a model of random distribution of zones within a sequence. For example, Korhola and Weckström (2004) applied Bennett's broken-stick model to core data from several lakes in Finnish Lapland to identify stratigraphic levels in which diatom floristic changes have been most distinct. Such analysis helped them elucidate

the relative role of local versus regional environmental changes in explaining the changes in fossil diatom assemblages. However, Bennett's method for determining the number of zones is based on the use of a single randomized sample as a comparison. Techniques that use a large set of randomized samples might provide more realistic means of making statistical evaluation of possible zonations.

Ordination techniques can be used to arrange species and/or samples along gradients. Ordination is especially useful for visualizing large and complex data sets in reduced dimensions that are ecologically easier to interpret. For example, stratigraphic sample scores derived from an indirect gradient analysis technique can be plotted chronologically, and when related together with species scores to ordination axes, used to determine the major ecological drivers of the temporal variation. Stager et al. (1997) used correspondence analysis (CA) to show that changes in the fossil diatom assemblages of a core from Lake Victoria occurred in response to climatic fluctuations during the Holocene. First, a CA ordination biplot of major diatom taxa distributed them into qualitative assemblage groups by quadrant (Fig. 3A). Based on available diatom habitat information, axis 1 was interpreted to approximate a gradient of water column stability and axis 2 to approximate a gradient of precipitation:evaporation (P:E) ratios or lake depth. When plotted stratigraphically, the sample scores of the two first CA axes were then able to be used as proxies of windiness, rainfall patterns, and monsoon variability (Fig. 3B). Large Lakes gives a more detailed account of the Lake Victoria diatom record and its use in paleoclimatic reconstruction.

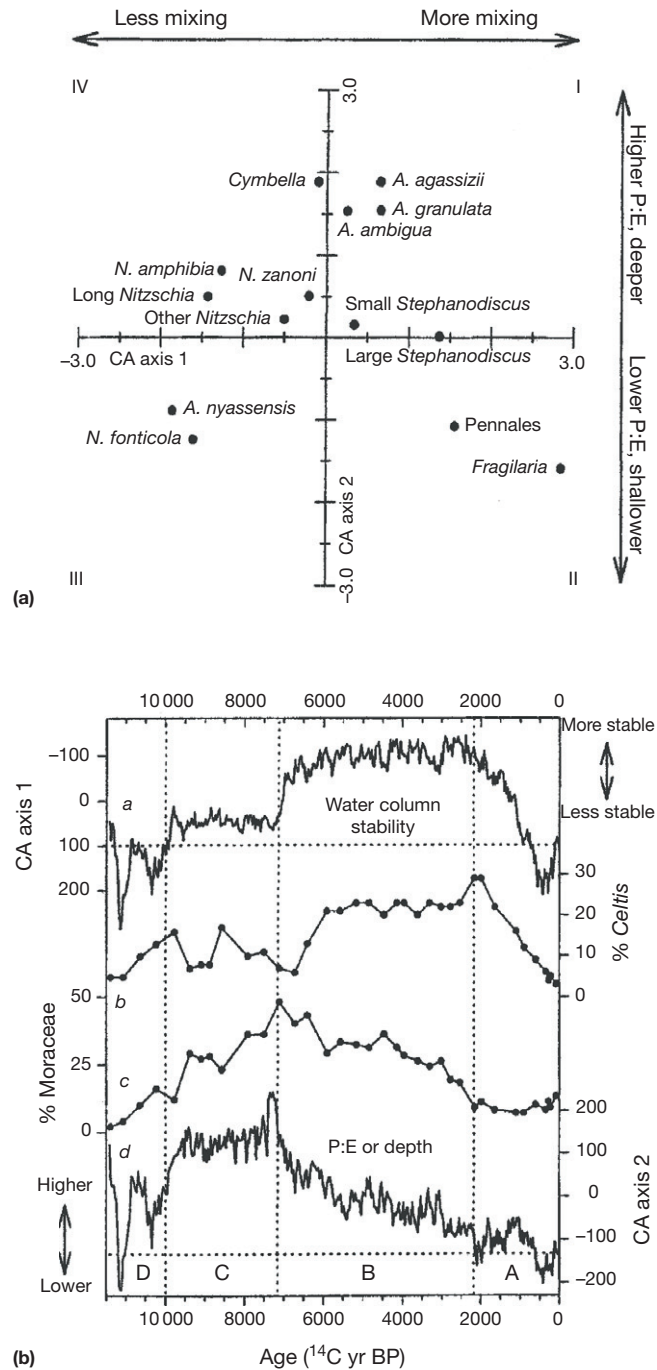


Figure 3 Interpretation of a diatom record of a core from Lake Victoria's Damba Channel. (a) CA biplot showing major diatom taxa of the core. (b) CA axis plots and pollen stratigraphy. Inverted CA axis 1 represents water column stability and resembles the *Celtis* (seasonally dry forest) pollen record. CA axis 2 represents P:E or lake depth, and resembles the Moraceae (moist forest) pollen record. Horizontal dotted lines mark present conditions. Vertical dotted lines delimit diatom zone boundaries. From Stager et al. (1997).

Ordination procedures can also aid in the comparison and synthesis of multiple data sets. For example, Sorvari et al. (2002) summarized the diatom records from several short cores from Arctic Lapland using principal components analysis (PCA), and showed that the changes in diatom assemblages were significantly correlated with long-term variations in regional spring temperatures. Smol et al. (2005) used

detrended canonical correspondence analysis (DCCA) to summarize the amount of species turnover in paleolimnological records (mostly diatoms) from 55 lakes in the circumpolar Arctic. They found widespread, parallel changes in species compositions that were likely to be driven by climate warming.

The multiple proxy indicator approach enables some independent variables representing different environmental

processes to be identified. Variance decomposition (Borcard et al., 1992) can then be effectively used to distinguish the independence and relative strength of different variables in explaining diatom species compositional changes through time. In each variance partitioning experiment, the diatoms are the response variable and other independent variables are used in different combinations as explanatory variables. The effect of time-dependent ecological and environmental processes can be further removed by introducing sample age as a covariable in all experiments. Finally, a variance decomposition experiment is performed with each of the predictor variables assigned as covariables to cancel out their effect on one another and assess redundancy proportions between variables. The explanatory strength of various variables (or their combinations) is assessed by testing the significance of the first ordination axis under different model conditions using a constrained Monte Carlo permutation test. For example, Barker et al. (2000) used partial redundancy analysis (RDA) to test whether the effect of tephra deposition 1.19 ka had a statistically significant impact on the diatom assemblages in Lake Massoko, a volcanic crater lake in southern Tanzania, and discriminated between floral changes in response to tephra and changes in climate (as derived from the pollen data), or those induced by physical catchment disturbance (as revealed by the magnetic properties of the sediment).

Given a modern reference data set (see below), canonical ordination can also be a useful visual aid in environmental reconstruction. Fossil diatom assemblages can be introduced in the ordination analysis of modern diatom data as

supplementary (passive) samples and hence positioned into the ordination plane of modern training set samples on the basis of similarities in their biological composition. Changes in the diatom assemblages within a core can therefore be compared directly to the diatom-environment gradients within the modern data set to follow core trajectories. Engstrom et al. (2000) used this methodology effectively when studying chemical and biological trends during lake evolution in Glacier Bay, Alaska (Fig. 4). They summarized the limnological trends in lakes of varying age by projecting their fossil diatom assemblages onto the DCCA biplot of the modern samples. The core data of most lakes showed a net negative trajectory along DCCA axis 1, suggesting that the main trend of lake development with increasing age was a progressive loss of pH, alkalinity, and base cations, and a corresponding increase in dissolved organic carbon (DOC).

Diatom core data can be subjected to several numerical data manipulations to evaluate diatom community stability and diversity. For the former, rate of change analysis (Jacobson and Grimm, 1986) has proved useful (Figures 5 and 9). This technique simply calculates the dissimilarity between successive samples divided by the estimated period of deposition elapsed between the two, in order to produce measures of any assemblage changes per unit time. Chord distance is the dissimilarity metric most commonly applied to diatom data because of its statistical properties concerning the weighting of species with low percentages. The rate of change method is critically dependent on the time standardization unit (TSU) used to standardize the estimated dissimilarity, so good

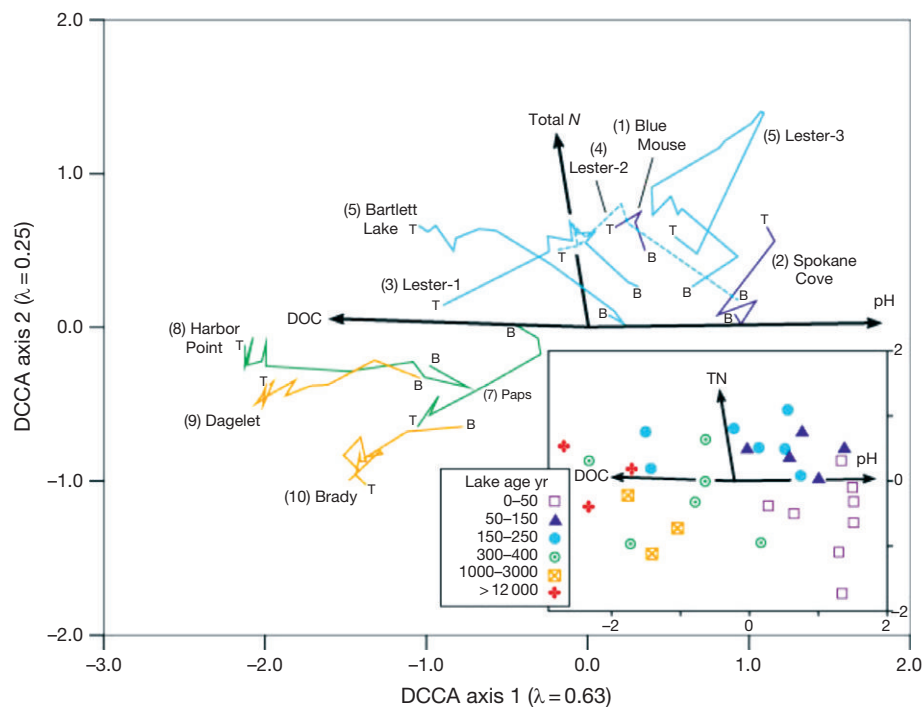


Figure 4 Detrended canonical correspondence analysis of modern and fossil diatom assemblages from the chronosequence of lakes in the Glacier Bay region of southeastern Alaska. In the inset, ordination of modern diatom assemblages (from surface sediments) and associated water-chemistry data are shown. The arrows depict the direction of increase and maximum variation of the measured environmental variables. Lakes plotting close to one another have similar diatom composition. In the main figure, the time-trajectories from 10 sediment cores (B, bottom; T, top) were projected onto the biplot by adding the core assemblages to the analysis as passive samples. From Engstrom et al. (2000).

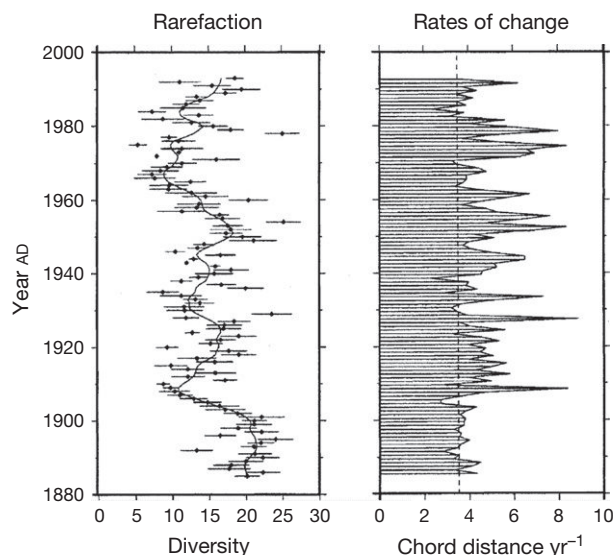


Figure 5 Diatom diversity as assessed by rarefaction analysis and rate of diatom assemblage change in the annually laminated sediment of Baldeggersee, Switzerland, between 1885 and 1993. The diatom diversity closely mirrors the different phases of nutrient enrichment in the lake with decreasing numbers of taxa during the eutrophication phases (cf. Fig. 9). The timing and direction of diatom assemblage changes also compare well with sedimentological and limnological changes (cf. Fig. 9). The error bars in the rarefaction diagram represent the 95% confidence interval, whereas the dashed line in the rates of change diagram indicates the approximate significance level (2.5th percentile). The line through the rarefaction estimates is based on a LOWESS smoother. From Lotter AF (1998) *The recent eutrophication of Badeggersee (Switzerland) as assessed by fossil diatom assemblages. The Holocene* 8: 395–405.

chronological control is essential for this. As sediment accumulation rates are variable, sample intervals are unequally distributed through time. In order to correct this systematic bias, re-sampling by interpolation of new samples at even time intervals can be done. The resulting synthetic data set can then be subjected to smoothing and the chord distance coefficient calculated (Laird et al., 1998). When analyzing large and complex data matrices, such as the multitaxon diatom core data sets, often are, it is sometimes necessary to reduce the dimensionality in the data (e.g., by PCA) prior to estimating the rates of change (Barker et al., 2000).

The assessment of past ecological diversity as deduced from the sedimentary record is a challenging task. Besides variables such as differential preservation of diatom fossils, one of the major issues in assessing past diversity is the choice of a sound measure that takes into account richness and equitability. The diversity of subfossil organisms cannot be investigated unless the count size of the samples is constant or standardized, because the number of identified taxa usually increases as the fossil count size increases. Rarefaction analysis (Heck et al., 1975) has been shown to be very suitable for fossil diatom data (Fig. 5). Rarefaction uses a strategy of random selection without replacement, typically selecting a sample of the same size as the smallest count from the entire population. In this way realistic estimates of richness are produced without any bias associated with the variability of individual sample count

sizes. However, when dealing with non-laminated sediments, differences in sediment accumulation rate need further to be taken into account when interpreting the diversity records.

Estimating Parameter Values Using Diatoms

Numerical modeling, with the aim of quantitative paleoenvironmental reconstruction from diatom assemblages, was first applied to reconstruct pH. The early applications were based on ecological species categories and pH indices, and later multiple linear regression models were developed to estimate pH according to the proportions of the established pH-preference groups (see Battarbee and Charles, 1987 and Davis, 1987 for extensive reviews). More recently, paleoecologists have cataloged the ecological optima and tolerances of diatom assemblages and have constructed transfer functions that allow environmental conditions to be inferred from fossil assemblage data (Birks et al., 1990). Quantitative approaches using diatoms are now being applied to a wide array of environmental problems and questions, with many new methodologies currently being implemented.

In diatom paleoecology, transfer functions simply describe the relationship between diatom assemblages and their environment with the ultimate goal to use the present-day ecology of diatoms to quantitatively infer past conditions. The basic approach is illustrated as a hypothetical example in Figure 6. The transfer function approach involves two steps: (1) the regression step in which the responses of taxa to the contemporary environment are modeled, and (2) a calibration step in which the environmental variable of interest is predicted from the fossil compositional data. Transfer functions are established using modern diatom assemblages and measured environmental variables from a calibration data set (also called a training set; Fig. 6A). These calibration sets involve sampling a range of sites (e.g., lakes, wetlands, estuaries, marine environments, usually >40 sites) for a suite of critically considered environmental variables, which are then related to the diatom assemblages preserved in the surface sediments (top 0.5–1 cm representing the last few years of sediment accumulation) of the systems concerned using various numerical procedures (Fig. 6B). The transfer function approach is based on the implicit assumption that the modern training sets cover the range of environmental conditions likely to be represented by the fossil assemblages. Hence, calibration data sets should ideally span all biologically relevant environmental gradients.

The mathematical methods for establishing a transfer function are commonly based on nonlinear regression techniques. The most popular approaches in diatom studies have been weighted averaging (WA; ter Braak and van Dam (1989)) and the WA partial least squares (WAPLS; ter Braak and Juggins (1993)) regression, because they perform well with noisy, species-rich, compositional data, with many zero values and some rogue observations (outliers). In both methods, the underlying response model is the Gaussian function, that is, each taxon has a bell shape (Gaussian) distribution along an environmental variable. From the Gaussian curve, the species optimum (the peak) and its tolerance (the standard deviation) can be estimated (Fig. 7).

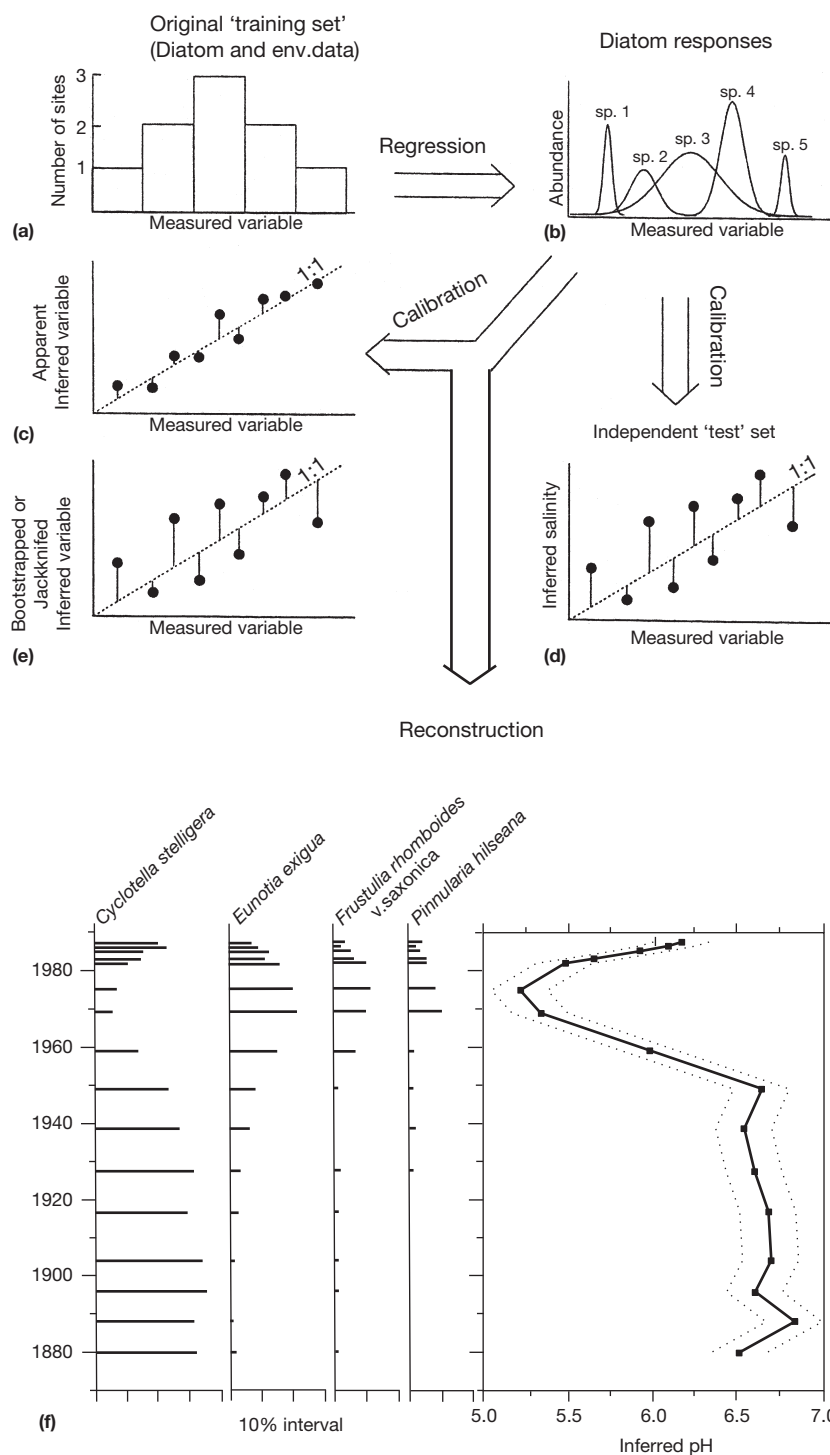


Figure 6 Simplified illustration of the typical steps involved in developing and assessing the predictive ability of a diatom-based calibration model and its application to fossil diatom data including: (a) selection of a modern training data set; (b) estimation of taxon responses based on their distribution in the training set (regression); (c) generation of a transfer function on the basis of the training set (calibration); (d) selection of an independent test set to assess the transfer function (calibration); (e) use of computer re-sampling techniques, such as jackknifing or bootstrapping, to assess the transfer function (calibration); and (f) reconstruction of parameter value from fossil diatom assemblages and estimation of sample-specific prediction errors by bootstrapping. Modified from Fritz et al. (1999) Fritz SC, Cumming BF, Gasse F, and Laird KR (1997) Diatoms as indicators of hydrologic and climatic change in saline lakes. In: Stoermer EF and Smol JP (eds.) *The Diatoms: Applications for the Environmental and Earth Sciences*, pp. 41–72. Cambridge: Cambridge University Press.

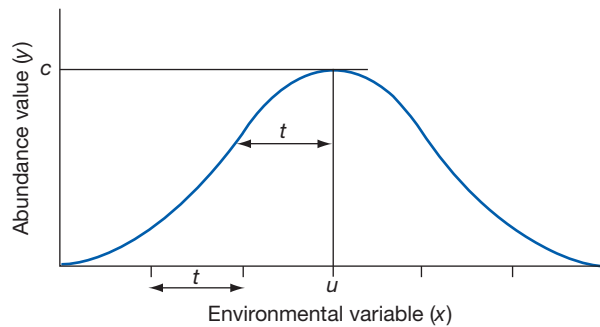


Figure 7 Gaussian response curve for the changing abundance of a species along an environmental gradient. t = tolerance; u = optimum; c = maximum. From ter Braak CJF (1987) *Unimodal Models to Relate Species to Environment*, 152pp. Unpublished PhD Thesis, University of Wageningen.

In the WA methods, the estimate for the species optimum (u_k) is obtained as the average of the k th taxon in the calibration dataset (y_{ik}) weighted by the environmental variable (x_i) of interest:

$$\hat{u}_k = \frac{\sum_{i=1}^n y_{ik} x_i}{\sum_{i=1}^n y_{ik}} \quad [1]$$

WA can also be used to estimate tolerance values (Fig. 7) for diatoms and these values can be used to down-weight the importance of taxa with a wide tolerance relative to those with a narrow tolerance. Figure 8 shows the response models and the estimated WA temperature optima and tolerances of the most frequent diatom genera and species for the Arctic Lapland temperature training set. These estimated optima and tolerances form the transfer function, which can then be used to infer past temperature from fossil diatom assemblages.

In addition to WA, other routinely used calibration techniques include the maximum likelihood (ML), modern analog technique (MAT), and Gaussian logit regression (GLR) (reviewed in Birks, 1995 and Birks, 1998). Recently, artificial neural networks (ANNs) have been introduced to diatom paleoecological research. However, Telford and Birks (2005) found that transfer functions based on a unimodal species–environment response are more robust to spatial autocorrelation of environmental variables in the training set (a serious problem for transfer functions) than MAT or ANN models, which are likely to over-fit the data unless a spatially independent test set (see below) is available to help model selection. Of other novel techniques, Bayesian inference is shown to have many advantages over conventional frequency distribution methods in environmental reconstruction, including reduction of inherent systematic bias, more precise error estimation, the possibility of utilizing existing ecological and environmental information, and improved extrapolation capacity (Toivonen et al., 2001; Vasko et al., 2000; Korhola et al., 2002).

Different modeling tools have their inherent philosophical and mathematical assumptions as well as limitations, so the prediction ability of the transfer models must be carefully evaluated. Ideally, independent data sets that are similar in size and environmental gradients would best suit this purpose

(Fig. 6D), but are seldom available because they are laborious and expensive to collect. Instead, computer-intensive re-sampling techniques, such as jackknifing (leaving one-out substitution) and bootstrapping (creation of pseudoreplicate data sets by re-sampling), are used (Fig. 6E); these should provide more realistic assessments of the prediction ability of a transfer model than the pure or apparent error (Fig. 6C). The output of the reconstruction procedure will be a reconstructed parameter value for each core sample plus sample-specific prediction errors, commonly estimated by bootstrapping (Fig. 6F).

Evaluation of the Reconstruction

There are no means or ‘skilled techniques’ to prove that a reconstruction is correct. The most robust evaluation is to compare diatom-inferred values with measured long-term data series where these exist (Figures 9 and 10). Close agreement between reconstructed and measured values increases confidence in the reconstruction whereas deviation between the two sets of data warrants caution. However, close agreement during the monitoring period does not necessarily imply accurate reconstructions for earlier periods, especially if assemblages have a very different composition. Conversely, divergence for the monitoring period does not necessarily imply that reconstructions for earlier periods are in error.

Another way to evaluate the robustness of a diatom-based reconstruction is to compare the inferred values with the ones reconstructed using other proxy sources (Fig. 9). Close agreement between several proxy indicators increases their reliability. In addition, indirect gradient analysis (PCA, CA, and DCA) can be used to determine whether the inferred environmental variable is related to the main directions of variation in the fossil diatom assemblages (i.e., the sample scores on the first unconstrained axis). The strength of the correlation between the unconstrained axis 1 sample scores and the inferred values indicates the extent to which the principal changes in diatom community structure can be explained by the inferences of the parameter in question.

Implicit in the transfer function approach is that the modern training set samples provide good analogs for the fossil diatom assemblages. This could simply be done by highlighting sediment samples that contain significant numbers of taxa that are absent from the training set. A more robust assessment is to find the closest modern analogs for each fossil sample using MAT (Fig. 9). MAT quantifies the dissimilarity between modern and fossil samples using various distance measures. Another way is to assess the goodness-of-fit between modern and fossil species data using constrained canonical correspondence analysis (CCA) with the variable of interest as the sole constraint. A fossil sample with a residual distance equal to or larger than the residual distance of the extreme 10% of the calibration data set is typically considered to have a poor fit to the desired variable (Birks et al., 1990).

Here are two examples in which multiple data manipulation and validation tools have been efficiently used to extract information from fossil diatoms. Wolfe (2003) examined different methods of interpreting diatom stratigraphic data from the late Holocene sediments of a small Arctic lake on Baffin Island, Nunavut, Canada, with the objective of gaining

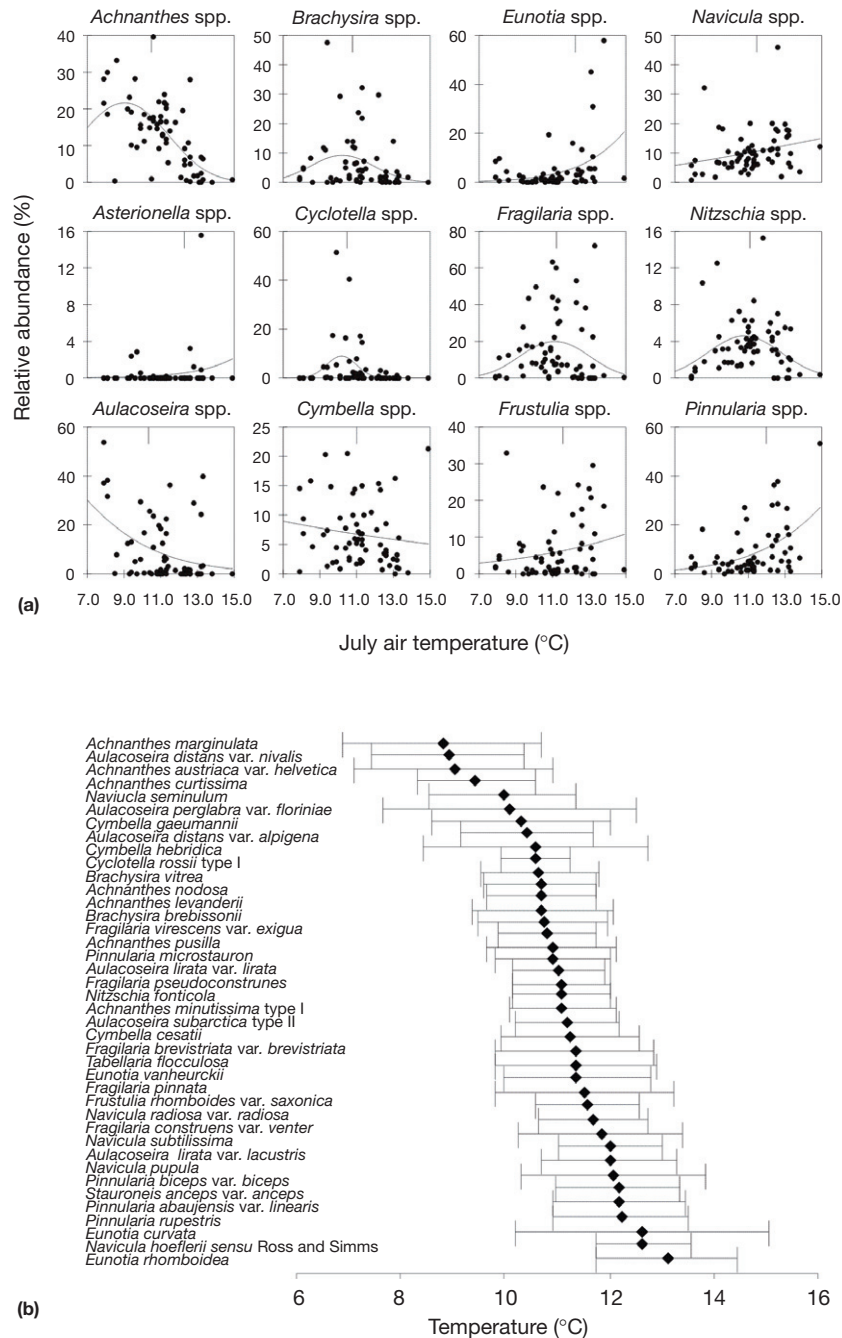


Figure 8 Responses of diatoms to the site-specific air temperature from a training set of 64 lakes in Arctic Lapland. (a) Distribution of the most dominant diatom genera along the temperature gradient. The estimated weighted averaging (WA) temperature optimum for each genus is indicated as a tick. The trend lines are based on a Gaussian logit model. (b) The estimated temperature optima (solid squares) and tolerances (bars) of the 40 most abundant diatom taxa in the data set arranged according to temperature optima. From Weckström J and Korkola A (2001) Patterns in the distribution, composition and diversity of diatom assemblages in relation to ecoclimatic factors in Arctic Lapland. *Journal of Biogeography*, 28, 31–46.

paleoclimatic insights from fossil assemblages. In this study, diatom relative frequencies and absolute abundances were subjected to several data manipulations, including application of a WA transfer function for summer water temperature, conversion of valve concentrations to cell biovolumes, biostratigraphic rate of change, and taxonomic richness estimated by rarefaction analysis. Hence, from the primary diatom data, a

second data set was derived on the basis of the data manipulations, consisting of seven new variables. The major directions of variability within the derived diatom data were then summarized using PCA and the results were considered in relation to independent paleoclimatic proxies to enable an objective assessment of the response of freshwater diatom communities to climatic change.

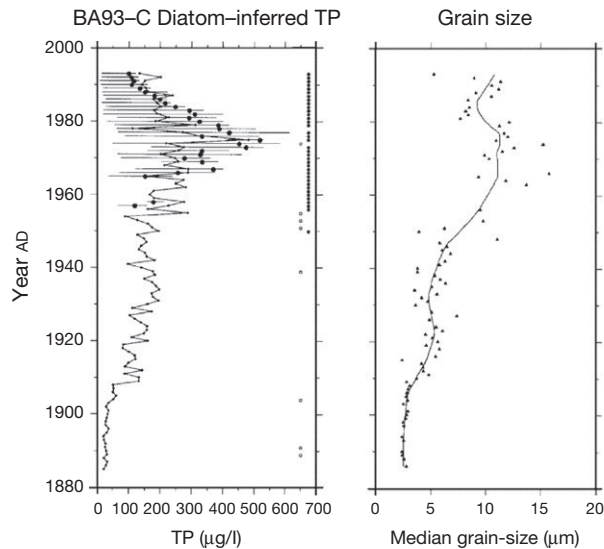


Figure 9 Measured total phosphorus (TP) during spring circulation (filled circles) compared to diatom-inferred TP values and median grain-size distribution in the Baldeggersee annual layers. The horizontal lines represent the annual TP range in the uppermost 15 m of the water column. The dots on the right hand side of the graph represent samples with close (filled dots; 2nd percentile) and good modern analogs (open dots; 5th percentile). Grain-size analyses in annual layers show variations in calcite size (a proxy for PO_4 concentration in the lake water) that parallel the diatom-inferred TP concentrations, hence giving independent support to the reconstruction. From Lotter AF (1998) The recent eutrophication of Badeggersee (Switzerland) as assessed by fossil diatom assemblages. *The Holocene* 8: 395–405.

The second example comes from Laajalahti, a semi-enclosed embayment of the eastern Baltic Sea that has been studied for diatoms to reconstruct its eutrophication history (Weckström et al., 2004). Figure 10 shows the summary diatom diagram, together with diatom-inferred total dissolved nitrogen (TDN) concentrations, diversity estimates, analog measures, and monitoring data of TN for the last 30 yr (Weckström et al., 2004). The diagram shows the transition from a flora dominated by benthic *Fragilaria* taxa in the late nineteenth century to one dominated by a succession of planktonic *Cyclotella* and *Thalassiosira* species today. All fossil samples in this case have squared chord distances less than 10% of the distribution of all distances among modern samples in the calibration set, as assessed by MAT, suggesting that the samples have good floristic analogs in the eastern Baltic training set. However, one sample had a poor fit to TDN as assessed by the constrained CCA. Comparison between the reconstructed values and monitoring data show that the transfer function is able to reproduce trends and absolute values for the last ca. 10 yr but underestimates concentrations during the most eutrophic period during the 1970s. This underestimation is probably due to the lack of high TN sites in the training set. The down core diversity of diatoms closely follows the pattern of eutrophication deduced from the diatom TDN transfer function. As expected, the diversity decreases with increasing concentrations on TDN, following a negative logarithmic function. These patterns give clear evidence that nutrient enrichment leads to a loss in diatom diversity at this site. There is more variability in the species richness than in the overall species compositions (as deduced on the basis of DCA). These patterns

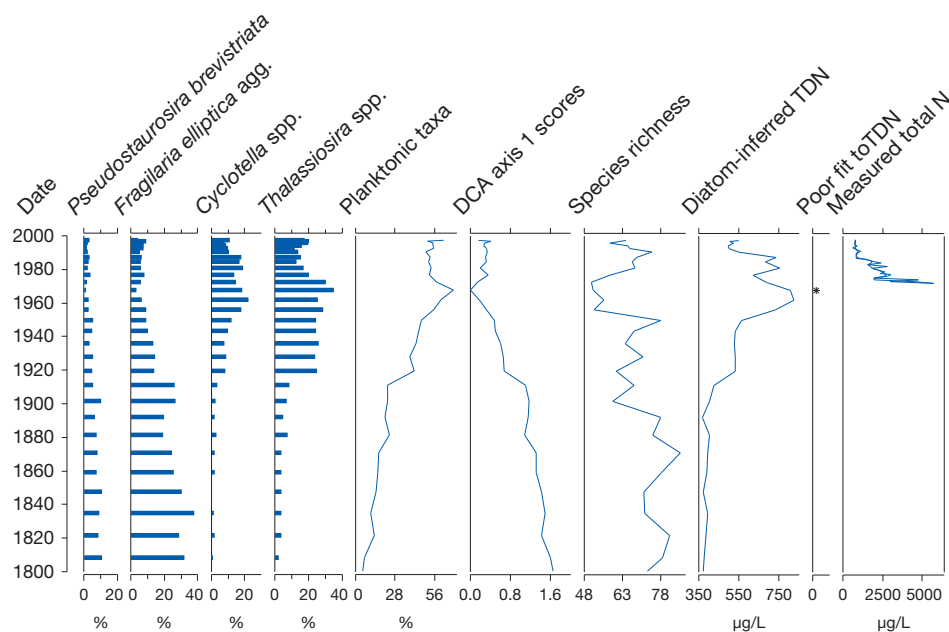


Figure 10 Summary diatom diagram showing major diatom taxa, proportion of planktonic taxa, detrended correspondence analysis (DCA) axis 1 scores, rarefaction species richness, diatom-inferred annual mean TDN and measured TN concentrations for Laajalahti embayment, Helsinki. The inferred TDN values are based on the two-component WAPLS model. The measured TN concentrations are presented as annual, seasonally weighted mean values. The sole sample having a poor CCA fit to TDN is shown by asterisk. Modified from Weckström K, Juggins S, and Korhola A (2004) Quantifying background nutrient concentrations in the coastal zone of the Baltic Sea. *Ambio* 33: 324–327 and Weckström K (2006) Assessing recent eutrophication in coastal waters of the Gulf of Finland (Baltic Sea) using subfossil diatoms. *Journal of Paleolimnology* 35: 571–592.

reinforce the general axiom that aquatic biological diversity tends to decrease with increasing nutrient levels.

Conclusions

Stratigraphic diatom data are complex, multivariate, and temporally ordered, most commonly forming irregularly spaced discontinuous time series. Such proxy data thus have special statistical properties and their numerical analysis requires techniques that take particular account of these unique data properties. Much progress has been made in the development of quantitative approaches to infer environmental variables from diatom assemblages. Robust diatom-based quantitative transfer functions have been created for pH, nutrients, salinity, organic carbon, water and air temperature, ice cover duration, lake level and other environmental parameters. However, quantitative inference techniques have their own shortcomings and limitations, so all transfer function approaches have to be assessed critically (Birks, 1998). We need to be sure that our interpretations are not becoming too mechanistic and straightforward, as the transfer functions will always produce results, whether correct or not. Also, it should be remembered that the transfer function approach is not the sole method for gleaning useful information from diatom stratigraphic data. As shown by the study of Wolfe (2003), highly valuable environmental information can be extracted from diatom stratigraphic data when subjected to many kinds of creative numerical data manipulations. Other indicators that can be measured in sediments will further strengthen conclusions drawn from the diatom data. In addition, the use of multiple indicators also allows questions of ecological response to environmental forcing to be addressed using diatoms. Perhaps in the future, our research should focus more on using fossil diatom compositions as response variables and early warning indicators rather than using them to reconstruct environmental change *per se*. For this, however, a more comprehensive understanding of the role of diatoms in the functioning of aquatic ecosystems is required.

See also: [Diatom Introduction](#). [Quaternary Stratigraphy: Continental Biostratigraphy](#).

References

- Abrantes F (2000) 200,000 yr diatom records from Atlantic upwelling sites reveal maximum productivity during LGM and a shift in phytoplankton community structure at 185 000 yr. *Earth and Planetary Science Letters* 176: 7–16.
- Barker P, Telford R, Merdaci O, et al. (2000) The sensitivity of a Tanzanian crater lake to catastrophic tephra input and four millennia of climate change. *The Holocene* 10: 303–310.
- Battarbee RW and Charles DF (1987) The use of diatom assemblages in lake sediment as a means of assessing the timing, trends, and causes of lake acidification. *Progress in Physical Geography* 11: 552–580.
- Battarbee RW, Jones VJ, Flower RJ, Cameron NG, Bennion H, and Carvalho L (2001) Diatoms. In: Smol JP, Birks HJB, and Last WM (eds.) *Tracking Environmental Change Using Lake Sediments: Terrestrial, Algal and Siliceous Indicators* Volume 3, pp. 155–202. Dordrecht: Kluwer.
- Bennett K (1996) Determination of the number of zones in a biostratigraphical sequence. *New Phytologist* 132: 155–170.
- Birks HJB (1995) Quantitative palaeoenvironmental reconstructions. In: Maddy D and Brew JS (eds.) *Statistical Modelling of Quaternary Science Data. Quaternary Research Association Technical Guide 5*, pp. 161–254. Cambridge: Quaternary Research Association.
- Birks HJB (1998) Numerical tools in palaeolimnology – Progress, potentialities, and problems. *Journal of Paleolimnology* 20: 307–332.
- Birks HJB, Line JM, Juggins S, Stevenson AC, and ter Braak CJF (1990) Diatoms and pH reconstruction. *Philosophical Transactions of the Royal Society of London B* 327: 263–278.
- Borcard D, Legendre P, and Drapeau P (1992) Partialling out the spatial component of ecological variation. *Ecology* 73: 1045–1055.
- Davis RB (1987) Paleolimnological diatom studies of acidification of lakes by acid rain: An application to quaternary science. *Quaternary Science Reviews* 6: 147–163.
- Denys L and De Wolf H (1999) Diatoms as indicators of coastal paleoenvironmental and relative sea-level change. In: Stoermer EF, Smol JP, Stoermer EF, and Smol JP (eds.) *The Diatoms: Applications for the Environmental and Earth Sciences*, pp. 277–297. Cambridge: Cambridge University Press.
- Dixit SS, Smol JP, Kingston JC, and Charles DF (1992) Diatoms: Powerful indicators of environmental change. *Environmental Science and Technology* 26: 23–33.
- Engstrom DR, Fritz SC, Almendinger JE, and Juggins S (2000) Chemical and biological trends during lake evolution in recently deglaciated terrain. *Nature* 408: 161–166.
- Fritz SC, Cumming BF, Gasse F, and Laird KR (1997) Diatoms as indicators of hydrologic and climatic change in saline lakes. In: Stoermer EF and Smol JP (eds.) *The Diatoms: Applications for the Environmental and Earth Sciences*, pp. 41–72. Cambridge: Cambridge University Press.
- Grimm EC (1988) Data analysis and display. In: Huntley B, Webb T, Huntley B, and Webb T (eds.) *Vegetation History*, pp. 43–76. Dordrecht: Kluwer.
- Heck KL, van Belle G, and Simberloff DS (1975) Explicit calculation of the rarefaction diversity measurement and the determination of sufficient sample size. *Ecology* 56: 1459–1461.
- Jacobson GL Jr. and Grimm EC (1986) A numerical analysis of Holocene forest and prairie vegetation in central Minnesota. *Ecology* 67: 958–966.
- Korhola A, Vasko K, Toivonen HTT, and Olander H (2002) Holocene temperature changes in northern Fennoscandia reconstructed from chironomids using Bayesian modelling. *Quaternary Science Reviews* 21: 1841–1860.
- Korhola A and Weckström J (2004) Paleolimnological studies in Arctic Fennoscandia and the Kola Peninsula (Russia). In: Pienitz R, Douglas MSV, and Smol JP (eds.) *Long-Term Environmental Change in Arctic and Antarctic Lakes*, pp. 381–418. Dordrecht: Kluwer.
- Kuwae M, Yoshikawa S, Tsugeki N, and Inouchi Y (2004) Reconstruction of a climate record for the past 140 kyr based on diatom valve flux data from Lake Biwa, Japan. *Journal of Paleolimnology* 32: 19–39.
- Laird KR, Fritz SC, Cumming BF, and Grimm EC (1998) Early-Holocene limnological and climatic variability in the northern Great Plains. *The Holocene* 8: 275–285.
- Lotter AF (1998) The recent eutrophication of Badeggersee (Switzerland) as assessed by fossil diatom assemblages. *The Holocene* 8: 395–405.
- Palmer AJM and Abbott WH (1986) Diatoms as indicators of sea-level change. In: van de Plassche O (ed.) *Sea-Level Research*, pp. 457–487. Amsterdam: Free University.
- Smol JP, Wolfe AP, Birks HJB, et al. (2005) Climate-driven regime shifts in Arctic lake ecosystems. *Proceedings of the National Academy of Science* 102: 4397–4402.
- Sorvari S, Korhola A, and Thompson R (2002) Lake diatom response to recent Arctic warming in Finnish Lapland. *Global Change Biology* 8: 153–163.
- Stager JC, Cumming B, and Meeker L (1997) A high-resolution 11,400-yr diatom record from Lake Victoria, East Africa. *Quaternary Research* 47: 81–89.
- Stoermer EF and Smol JP (eds.) (1999) *The Diatoms: Applications for the Environmental and Earth Sciences*, p. 469. Cambridge: Cambridge University Press.
- Telford RJ and Birks HJB (2005) The secret assumption of transfer functions: Problems with spatial autocorrelation in evaluating model performance. *Quaternary Science Reviews* 24: 2173–2179.
- ter Braak CJF (1987) *Unimodal Models to Relate Species to Environment*. University of Wageningen p. 152. Unpublished PhD Thesis.
- ter Braak CJF and Juggins S (1993) Weighted averaging partial least squares regression (WA-PLS): An improved method for reconstructing environmental variables from species assemblages. *Hydrobiologia* 269/270: 485–502.
- ter Braak CJF and van Dam H (1989) Inferring pH from diatoms: A comparison of old and new calibration methods. *Hydrobiologia* 178: 209–223.
- Toivonen H, Mannila H, Korhola A, and Olander H (2001) Applying Bayesian statistics to organism-based environmental reconstruction. *Ecological Applications* 11: 618–630.
- Vasko K, Toivonen HTT, and Korhola A (2000) A Bayesian multinomial Gaussian response model for organism-based environmental reconstruction. *Journal of Paleolimnology* 24: 243–250.

- Weckström K (2006) Assessing recent eutrophication in coastal waters of the Gulf of Finland (Baltic Sea) using subfossil diatoms. *Journal of Paleolimnology* 35: 571–592.
- Weckström K, Juggins S, and Korhola A (2004) Quantifying background nutrient concentrations in the coastal zone of the Baltic Sea. *Ambio* 33: 324–327.

- Weckström J and Korhola A (2001) Patterns in the distribution, composition and diversity of diatom assemblages in relation to ecoclimatic factors in Arctic Lapland. *Journal of Biogeography* 28: 31–46.
- Wolfe A (2003) Diatom community responses to late-Holocene climatic variability, Baffin Island, Canada: A comparison of numerical approaches. *The Holocene* 13: 29–37.

Diatomites: Their Formation, Distribution, and Uses

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Introduction

Diatoms are ubiquitous microalgae; they grow in both terrestrial and aquatic habitats and occur in most regions of the world. Because these algae possess resistant siliceous cell walls, their remains are often preserved in sedimentary environments and, where they are sufficiently abundant, diatomaceous deposits can accumulate. Such deposits are composed predominantly of diatom valves or frustules and are collectively referred to as diatomite. Diatomites vary considerably in texture and composition and are known by a variety of names, including kieselguhr, diatomaceous earth, fuller's earth (bentonite), and tripolite. Despite differences in precise composition that are attributable in part to the nature of the diatoms preserved (diatom valves can vary from about 5 to 200 μm in diameter), the word diatomite is generally used to describe any sedimentary deposit where the proportion of diatom remains or of diatom silica is high. When extracted, diatomite can contain a considerable proportion of water as well as non-biogenic clastics and organic material. The term diatomite is often used for sediment where silica exceeds 50% of the bulk sediment, and commercially useful diatomite contains more than 80% silica (e.g., Kadey, 1983). Diatomite deposits have a worldwide distribution and can be either freshwater or marine in origin, ancient or modern, and can usually be readily identified in the field as gray sediment (friable when dry) with a low specific density. In this article, the formation, preservation, distribution, and uses of diatomite are considered.

Diatomite Formation

Whether in past or contemporary environments, the process of diatomite formation is essentially the same: prolific diatom growth in either fresh water or seawater and the subsequent accumulation of their siliceous remains into a sedimentary deposit. Key aspects of diatomite formation are high diatom productivity supported by adequate light and dissolved nutrient availability (especially of dissolved silica), good diatom preservation (see Section 'Preservation of Diatoms'), and perhaps most importantly, the lack of any significant terrigenous, or other biogenic or clastic materials diluting the diatom component of the accumulating sediment (Harwood, 1999). Terrestrial diatomites are all fossil deposits and enjoy a worldwide distribution; both freshwater and marine fossil diatom deposits are common.

The formation of diatom-rich deposits is indicated schematically in Figure 1. The figure indicates two extreme processes for diatom deposition. On the left of Figure 1, the basic mode of diatomaceous sediment formation in a deep water environment is indicated. Here, diatom phytoplankton grows in the illuminated upper part of the water column. As the crop develops, diatoms begin to sink and mass sinking events mark the culmination of growth periods as the diatom

phytoplankton bloom collapses. Following a period of sinking (which depends on water depth), the dead and senescent diatoms are deposited onto the ocean floor or lake bed. Once deposited on the surface sediment, the diatoms are subjected to taphonomic processes (see below) as they are incorporated into the permanent sediment record. In shallow water environments, diatomaceous sediments are formed rather differently. On the right of Figure 1, a shallow water environment is indicated and here diatoms grow within a short water column as well as in biofilms on subaqueous sediment and aquatic plant surfaces. Hence, the sinking depth is short or nonexistent, and diatom species are predominantly benthic or tychoplanktonic in habit, rather than planktonic. At the sediment surface, the diatoms are again subjected to taphonomic processes and, in some shallow water environments, these can be intensive, depending on the local environmental conditions. These shallow water deposits are normally formed only in freshwater habitats where the inclusion of aquatic animal and plant macrofossils (such as fish and charophyte remains) often confirm the shallow water depth of the system.

In an oceanic environment, diatom-rich sediments are referred to as diatom oozes and they occupy significant areas of the world's ocean surface sediments, especially in the Southern Hemisphere. These oozes occur most extensively in cold water regions, notably around the productive Antarctic convergence zone but also in the central eastern Pacific where high diatom productivity is supported by nutrient-rich upwelling currents. These pelagic sediments are composed of at least 30% biogenic silica and the diatoms are nearly all planktonic. In deep freshwater lakes, the formation of diatomaceous sediment is somewhat similar to that observed in the oceans; for instance, diatom ooze accumulates in Lake Baikal, Siberia. This is especially so for the north basin of this great lake. In shallow lakes, benthic diatom production requires that the sediment surface lies within the photic zone and that the diatoms are growing in fairly clear water with only small amounts of terrigenous material entering the system. Interestingly, many shallow freshwater diatom deposits are closely associated with geologies that potentially supply abundant dissolved silica for diatom growth. Hence, locations where these diatomites have formed are often associated with characteristic geological formations (basalts, tephra, and volcanic deposits) that are rich in silica. Such associations are found in locations as disparate as Lillcur (Australia), Lomocastro (Costa Rica), Leizhou (China), and Toome Bridge (Northern Ireland).

Irrespective of whether a freshwater or marine environment is considered, diatoms deposited onto or growing on a subaqueous sediment surface are subjected to a variety of sedimentary processes. Those that affect diatom fossilization are referred to as taphonomic processes. These include sediment disturbances of various kinds (sediment resuspension and bioturbation, sediment dewatering) that can break or dissolve at least some of the deposited diatoms. Surface-sediment

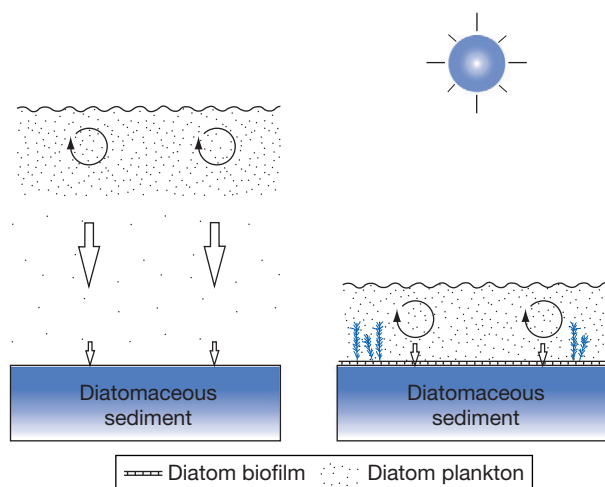


Figure 1 Formation of diatomaceous sediments. Left indicates the processes involved in the formation of diatomaceous sediment in a deep water environment. Right indicates the processes involved in the formation of diatomaceous sediment in a shallow water environment. The essential difference between these environments is light availability; in shallow water, light promotes growth of both benthic (as a biofilm on surfaces) and planktonic (free-floating) diatoms, but in deep water, diatom plankton growth is restricted to the upper illuminated surface water. Closed arrows indicate water circulation and open arrows indicate sinking diatoms.

taphonomic processes can operate until permanent burial; then slower sedimentary processes (compaction and diagenesis) take over and transform the diatomaceous sediment into diatomite. Most importantly for diatomite formation, the overall sedimentary environment must be conducive to the preservation of diatoms. However, after incorporation into permanent sediments, diatom remains can persist for millions of years.

Preservation of Diatoms

Diatoms are prone to damage and dissolution and these processes can combine to diminish the quantity and quality of diatoms accumulating in sediments. Most processes inimical to the preservation of a high portion of sedimentary diatom remains operate within the water column or at the mud water interface on the sea or lake floor. Diatoms will tend to dissolve whenever the surrounding water is undersaturated with regard to dissolved silica (e.g., Shemesh et al., 1989), and preservation will be further reduced by breakage, often encountered in high-energy environments and promoted by grazing (cf. Flower, 1993). Depending on water column depth, the degree of silicification of individual species of diatom, and a variety of water quality characteristics, diatoms will be preferentially preserved, but major initial losses by dissolution can occur, even in areas with sediments rich in diatoms. In Lake Baikal, for example, over 99% of diatoms produced in the euphotic zone dissolve before being incorporated into sediments. Some dissolution occurs in the water column, but most of it occurs as a result of surface-sediment processes (Ryves et al., 2003). Diatom remains in diatomites can show a range of preservational states from good to very poor and, although dissolution and breakage occurs to various extents during sediment formation, the

preservational condition of some terrestrial diatomites decreases from aerial exposure and weathering.

Over geological time, diatom silica changes state; the physical frustular structures disappear as the initial opaline silica of the diatom cell wall loses water and changes state, eventually becoming chrysotbolite. Despite differing formation processes, we can generalize that diatomites form in shallow or deep water environments as a result of moderate to high diatom productivity, adequate preservation and an absence of diluting detrital materials. Because of the instability of opaline silica, diatomite deposits are indicative of the most recent geological period, the Cenozoic.

Diatomite Distributions and Age

Terrestrial diatomite deposits occur on all the major continental land masses and, although Quaternary deposits are most frequent, many date from the Miocene Period. The Pre-Quaternary deposits are often substantial, usually marine in origin, and are particularly frequent around the margins of the Pacific Ocean (Reading, 1978). A well-known example is the extensive Miocene Monterey Formation in California, but other similar examples exist in Japan, South Korea, Eastern Russia, and in South America. Although the Monterey formation is large and variable, the Toro Road exposure (Barron, 1976) consists of almost pure marine diatoms, but preservation is often poor with many valves being broken. Aquitanian-Burdigalian and Messinian diatomites occur in countries bordering the Mediterranean, and Reading (1978) lists Romania, Spain, Sicily, Algeria, and northern Italy as regions with abundant marine Miocene diatomite deposits. The Miocene seems to have been a time of particularly extensive diatomite formation and the world's largest (marine) diatomite deposit of this period is probably at Lompoc in California (Dolley, 2002). Reading (1978) attributes the prevalence of Miocene diatomites to both tectonic and global climatic changes but Kidder and Gierlowski-Kordesh (2005) suggested that the nonmarine Miocene diatomite deposits are associated with the rise of grasslands and changes in the silica cycle. However, increased widespread weathering of silica-rich source rocks such as basalt may be a more likely source of dissolved silica at that time.

Freshwater diatomites are common around the Tertiary/Quaternary boundary and those dating across the Plio-Pleistocene periods occur in sections often tens of meters thick at several locations around the world. Notably, such freshwater deposits occur in Central (Mathers et al., 1991) and North America (Bradbury and Krebs, 1982), China (Qu and Chen, 2005), and Africa (Gasse, 1980). See Trauth et al. (2005) for a summary of diatomite sequences in East Africa. One Quaternary diatomite exposure (Figure 2) in central Kenya is over 30 m deep; it is near ancient Lake Kariandisi and has been dated to 0.97–0.974 Ma (Trauth et al., 2005). Its existence illustrates the development of wet periods in the East African Rift Valley zone at that time (M. Maslin pers. comm.). Also in Africa, extensive diatomite deposits occur around the Lake Chad area and these extend into the Holocene epoch. Diatomite samples collected as part of the 'ACACIA' (Arid Climate, Adaptation, and Cultural Innovation in Africa) Project from the former Lake Megachad and elsewhere in this



Figure 2 Exposed outcrop of freshwater diatomite from ancient Lake Kariandusi in Kenya. The section dates to 0.97–0.974 Ma and illustrates a former wet period in the East African Rift Valley system (see Trauth et al., 2005). Note that several caves have been excavated in the basal diatomite section. Photograph by M. Maslin.

region contain abundant freshwater plankton such as *Aulacoseira granulata*, *Cyclotella* (= *Discostella*) *stelligera*, and *C. ocellata* and benthic diatoms (especially *Cocconeis placentula*, *Denticula*, and *Epithemia*), as well as tephra deposits (Gasse, 2002; D. Verschuren, pers. comm.). These diatom species indicate the former presence of clear dilute alkaline water of varying water depths. The Bodélé depression (northern Chad) contains large areas of exposed diatomite, which are a globally significant source of wind-blown dust (Giles, 2005).

Another region with extensive Pliocene–Pleistocene diatomites is Central America. Deposits of freshwater (mainly planktonic) diatoms in Costa Rica can reach 100 m in thickness (Mathers et al., 1991). These also often contain abundant macrofossils of fish and other aquatic organisms. Other very extensive diatomite deposits occur in the Far East, particularly China, where they are often associated with volcanic formations and tephra deposits. In China, they typically date from the Pliocene to Holocene (Shang, 2003). The commercial diatomite deposit in the Leizhou Peninsula (North China) is reported to be of middle Pleistocene age (Qu and Chen, 2005) as are those in southwestern China, especially in Yunnan Province (Shang, 2003).

Holocene freshwater diatom deposits occur globally, but they tend to be thin, often being only several meters or less thick, and are often localized in distribution. Only a few lakes seem to be accumulating diatomite-quality sediment today. Lake Myvatn in Iceland is one lake where diatomaceous sediment is accumulating and where diatomite was being extracted commercially until recently (Kjaran et al., 2004). Many recent diatomite deposits, however, seem to have formed during the early to mid-Holocene. They are regional in distribution but very widespread; Eardley-Wilmot (1928) gives a thorough account and maps such deposits in Canada (notably in British Columbia and Nova Scotia). In the United Kingdom, Holocene diatomites occur at sites such as Toome Bridge in Northern Ireland (Wilson, 1972), in Loch Cuithir and other basins on the Isle of Skye in Scotland (Robertson, 1945), and



Figure 3 The exposed and wind-eroded surface of a Holocene freshwater diatomite deposit near Lake Qarun (Egypt). Note that the desiccation fissures in the exposed diatomite surface are filled with wind-blown sand. Photograph by R.J. Flower.

in England, notably at Kentmere (e.g., Round, 1957). Today, human activities have accelerated soil erosion and promoted cultural eutrophication of inland water bodies, all processes that tend to diminish the proportion of diatoms in accumulating sediments.

One rather local but environmentally significant Holocene diatomite deposit occurs in the Faiyum Depression (Egypt) to the north and southwest of modern Lake Qarun (Flower et al., 2006). This material was explored by Aleem (1958); it occurs as large expanses of white sediment, several meters thick in places, and is interspersed with blown sands (Figure 3). Fish remains are abundant and according to Wendorf and Schild (1976) these calcareous diatomites were deposited in several stages when the present lake was much larger. However, dates are few and not directly made on the diatomite, but on organic matter within diatomite sections. Radiocarbon dates indicate the deposit is of early Holocene age, with some sections younger than 5550 years BP. Other diatomite deposits occur around the northern fringes of the Sahara and especially in Algeria (Baudrimont, 1973), and some date to the Neolithic period. Collectively, they indicate a freshwater crisis during the middle Holocene as climate change and aridity affected much of Northern Africa.

The Use of Diatomites

Diatomites have two principal uses, one is commercial and entails extraction and processing into various diatomite products and the other relates to their value as an archive of past environments and of environmental change. Regarding the former, diatomites have been used by people for several thousands of years, but it is only during the last 200 years that diatomite deposits have been extracted on a commercial scale.

Diatomite Extraction

Use of diatomite materials first requires extraction of a suitable deposit, followed by fabrication into materials appropriate for distribution. Whether one considers ancient or modern uses,

the activities involved are similar and entail both extraction and processing. The ancient methods seemed to have focused on cutting blocks of diatomite sediment or manufacturing diatomaceous mud bricks. For the latter, diatomites were mixed (presumably by hand) with water and plant materials (usually straw), molded, and dried in the sun (Flower, 2006; Spencer, 1979). More recently, traditional diatomite extraction in the United Kingdom was largely done by hand and the techniques depicted by Wilson (1972) for diatomite products from the Toome Bridge area of Northern Ireland are typical. Diatomite bricks were cut by hand, stacked into piles for air drying and then milled to produce a fine dry powder product for bagging and sale. A commercial company (W. Kenyon Ltd.) exploited these deposits from around the outflow of Lough Neagh until recently and the remaining deposits are now protected under environmental legislation. Small-scale diatomite extraction was also formerly undertaken in Scotland, especially on Skye where eight diatomite sites occur in the Trotternish area. The Loch Cuithir site was worked on until 1914 and again after 1937 (<http://canmore.rachms.gov.uk/en/1131/>). Nevertheless, during the 1950s and 1980s, commercial extraction of diatomite effectively ceased in the United Kingdom, even though the feasibility of small operations is still debated. Elsewhere in the world, diatomite extraction is currently big business, and this is particularly so in the United States, China, and the former Soviet Union countries, where advanced plant machinery and high-temperature drying of the product are deployed. Other countries with commercially active diatomite extraction industries include Australia, Denmark, Spain, and countries in South and Central America (see Dolley, 2002; Eardley-Wilmot, 1928). World production of diatomite averaged about 1.65 million tons per year from 1988 to 1992 (British Geological Survey, 1992).

Ancient Uses of Diatomites

Archaeology has indicated some interesting applications of diatom-rich materials but the extent of their use is unclear. Nevertheless, the value of diatomite as a lightweight, easily handled material seems to have attracted interest. Dolley (2002) noted that lightweight building bricks of diatomite were used by the ancient Greeks and that diatomite blocks were used in the construction of the sixth century Hagia Sophia church, Istanbul. In ancient Egypt, the Ptolemaic inhabitants of the archaeological site of Dimai in the Faiyum also seemed well aware of at least some of the properties of diatomite deposits. Mud bricks at Dimai were fabricated from local diatomite deposits and used in the construction of dwellings, a funerary complex, and official buildings (Flower, 2006). Diatomite can be found frequently exposed near Dimai (Figure 4) but whether this material was actively sought for making mud bricks is unknown. It is thought, however, that the insulating properties of diatomite could well have helped to keep buildings internally cool in the hot climate. Diatoms in building materials are known in archaeological sites elsewhere, but they seem to be present as passive tracers of source materials. Caton-Thompson and Gardner (1934) described a region, also near Dimai, of siliceous (diatomite) deposits that were irrigated and used as a soil base for crop cultivation in Ptolemaic times.



Figure 4 A partly eroded ridge of a Holocene freshwater diatomite near Dimai (Faiyum, Egypt). Note that the exposure is fissured and faulted and is severely wind eroded. Photograph by R.J. Flower.

Diatomites in Modern Commercial Usage

It is the basic structure of diatomite, its high porosity, large surface area, abrasive quality, insulation capability, and refractory nature that makes it so useful commercially. The principal commercial uses of processed diatomite are in the food and building industries but it also has applications in the chemical, agricultural, and related industries (Eardley-Wilmot, 1928).

In the food industry, diatomite-derived products (usually powders) are useful as a filtrant or as an absorbent for liquids produced for consumption. There are a variety of liquid clarification aids produced commercially and these are all designed to remove impurities and to clarify liquids. Hence, wine and beers are often treated and clarified with processed diatomite as are certain oils and fats. In the chemical industry, diatomite products are used as filler and as a valuable additive to many paints, rubber, and plastics. Diatomite particles are an effective absorption medium for some macromolecules and can be used to selectively capture some dissolved organic compounds. Because of the mildly abrasive nature of diatomite, it is also used in the manufacture of polishes and in some toothpastes. In agriculture, processed diatomite has several uses as a soil additive; it improves soil quality (mainly through its ability to take up and retain water) and it can act as a snail and slug repellent or to protect stored grain from insect pest attack (Arnaud et al., 2005). In some agricultural locations, natural surface soil consists largely of diatomite and the resulting light and freely draining soils are favored in Northern Ireland for growing root crops (Cruickshank, 1982).

Some twentieth century uses of diatomite are being replaced by modern synthetic materials, but it is in the building industry that diatomite retains considerable potential. Traditionally, diatomite has been the material of choice in the manufacture of fire bricks, where the fine-grained siliceous material confers high insulation value. A range of brick and tile diatomite products have been developed with varying degrees of success (Eardley-Wilmot, 1928). However, more recent research on the addition of diatomite products to building materials, cements, and building bricks, has shown

that the structural properties of these materials can be improved. Addition of diatomite to fabrication processes of modern building bricks can influence the pozzolanic properties of the product (Wild et al., 1997). The pozzolanic properties of high-reactivity silica in diatomite produce cements with increased strength and compressibility resistance, and their benefits are similarly useful for ceramics. The properties of various diatomites are being tested as an additive in modern light weight building bricks where strength and insulation are needed (Pimraksa and Chindapsirt, 2009). The reaction of the silica surfaces in diatomite with aromatic compounds is the focus of interesting new research and Goa et al. (2005) showed that surface-modified diatomite can be a very effective trap for aqueous phenol. Hence, treated diatomite could have potential applications for combating water pollution.

Diatomites in Environmental Science

Diatom-rich sediment sections in marine diatom deposits have long been used for exploring paleodiversity in diatoms. Well-known sites are in New Zealand (the Oamaru Deposit, Grove and Sturt, 1853) and in California (the Moreno Shales, Hanna, 1927). Later, diatom-rich marine sediment cores were examined to demark biostratigraphic facies, to help define correlations and sedimentary chronozones, and to demonstrate species evolution (e.g., Strelikova, 1990). Similarly, terrestrial deposits of freshwater diatoms have been evaluated since the mid-nineteenth century. These were examined initially from the point of view of diatom diversity, and it was not until the 1980s that these deposits were used to develop a worldwide lacustrine chronobiostratigraphy and to infer climate change. Bradbury and Krebs (1982) studied Miocene to Holocene diatomites in Idaho and Oregon and indicated that at least seven taxa were identified as being common to lacustrine deposits at various locations in the Northern Hemisphere. Working in Ethiopia, Gasse (1980) used the diatom stratigraphy of a 31-m section of diatomite to show that the diatom record was indicative of past environments where water quality and climate fluctuated.

On a more recent time scale, diatomite has also been employed as a tracer in freshwater systems (e.g., Lund, 1974) and as a stratigraphic marker to estimate sediment accumulation in both lakes and salt marshes. Diatomites have also been examined to reveal species changes according to past lake-level variations and, normally being terrestrial deposits, they can be readily sampled at exposures or cored with hand-operated equipment (Figure 5). Holocene diatomites near extant lakes such as Lough Neagh in Northern Ireland (Plunket et al., 2004) and Lake Qarun in Egypt (Wendorf and Schild, 1976) have been used to document lake-level fluctuations in relation to the establishment of early human occupation sites.

Conclusions

Diatomites around the world have provided people with a rich source of material for both commercial exploitation and paleoenvironmental science. Commercial uses of diatomite are changing, and some applications are diminishing due to



Figure 5 Sampling a Holocene freshwater diatomite deposit in the Toome Bridge area of Northern Ireland. Note the exposed cut face of the diatomite deposit in the foreground and the upper 20–30 cm of soil cover. The semi water-logged deposit is about 1.5-m thick and a Hiller corer is being used to collect a stratigraphically ordered sequence of samples for paleoenvironmental analysis. Photograph by R.J. Flower.

the development of cheaper alternatives. Yet diatomite has considerable potential as an additive to building materials and it also has promising applications in the organic chemistry industry. As for paleoenvironmental research, there are rich opportunities for further exploitation of the biostratigraphies of diatomite deposits as recorders of past climates, human–aquatic ecosystem interactions, and of diatom species evolution.

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See also: [Diatom Introduction](#), [Diatom Methods: Use in Archaeology](#), [Diatom Records: Large Lakes; Pacific](#), [Paleolimnology: Overview of Paleolimnology](#).

References

- Aleem AA (1958) A taxonomic and palaeoecological investigation of the diatom flora of the extinct Fayoum Lake (Upper Egypt). *University of Alexandria Bulletin* 2: 17–44.
- Arnaud L, Lan HTT, Brostaux Y, and Haubruge E (2005) Efficacy of diatomaceous earth formulations admixed with grain against populations of *Tribolium castaneum*. *Journal of Stored Products Research* 41: 121–130.
- Barron JA (1976) Marine diatom and silicoflagellate biostratigraphy of the Delmontain Stage and the type Bolivina obliqua Zone, California. *Journal of Research of the U.S. Geological Survey* 4: 339–351.
- Baudrimont R (1973) Recherches sur la diatomées des eaux continentales de l'Algérie: Ecology and paleoecology. *Mémoires de la Société d'Histoire Naturelle de l'Afrique du Nord* 12: 1–249.
- Bradbury JP and Krebs WN (1982) Neogene and Quaternary lacustrine diatoms of the western Snake River basin Idaho-Oregon, USA. *Acta Geologica Academiae Scientiarum Hungaricae* 25: 97–122.
- British Geological Survey (1992) World Mineral Production 1988–1992. *Preliminary statistics*. British Geological Survey.
- Caton-Thompson GW and Gardner E (1934) *The Desert Fayum*. London: Royal Anthropological Institute.
- Cruickshank JG (1982) Soils. In: Cruickshank JG and Wilcock DN (eds.) *Environment and Natural Resources*, pp. 165–184. Belfast: The Queens University of Belfast.
- Dolley TP (2002) Diatomite. *Metals and Minerals*, vol. 1, 25.1–25.6. US Geological year Book.
- Eardley-Wilmot VL (1928) Diatomite, its Occurrence, Preparation and Uses. *Report for the Department of Mines (Mines Branch)*. Ottawa: F. A. Acland.
- Flower RJ (1993) Diatom preservation: Experiments and observations on dissolution and breakage in modern and fossil material. *Hydrobiologia* 269(270): 473–484.
- Flower RJ (2006) Diatoms in ancient building materials, application of diatom analysis to Egyptian mud bricks. *Nova Hedwigia, Beiheft* 130: 245–264.
- Flower RJ, Stickley C, Rose NL, Peglar S, Fathi AA, and Appleby PG (2006) Environmental changes at the desert margin: An assessment of recent paleolimnological records in Lake Qarun, Middle Egypt. *Journal of Paleolimnology* 35: 1–26.
- Gasse F (1980) Les diatomées lacustres Plio-Pleistocene du Gadeb (Ethiopie) Revue Algologique. *Mémoire hors-série* 3: 1–249.
- Gasse F (2002) Diatom inferred salinity and carbonate oxygen isotopes in Holocene waterbodies of the Western Sahara and Sahel (Africa). *Quaternary Science Reviews* 21: 737–767.
- Giles J (2005) The dustiest place on earth. *Nature* 434: 816–819.
- Goa BJ, Jiang PF, An FQ, Zhao SY, and Ge Z (2005) Studies on the surface modification of diatomite with polyethylenimine and trapping effect of the modified diatomite for phenol. *Applied Surface Science* 250: 273–279.
- Grove E and Sturt G (1853) On a fossil marine diatom deposit from Oamaru, Otago New Zealand. *Journal of the Quekett Microscopical Club* 2: 321–330.
- Hanna DG (1927) Cretaceous diatoms from California. *Occasional Papers of the California Academy of Sciences* 13: 5–48.
- Harwood DM (1999) Diatomite. In: Stoermer EF and Smol JP (eds.) *The Diatoms: Applications for the Environmental and Earth Sciences*, pp. 436–446. Cambridge: Cambridge University Press.
- Kadey FL (1983) Diatomite. In: Lefond SJ (ed.) *Industrial Minerals and Rocks*, 5th edn., pp. 677–708. New York: Society of Mining Engineers of the American Institute of Mining, Metallurgical and Petroleum Engineers, Inc.
- Kidder DL and Gierlowski-Kordesh EH (2005) Impact of grassland radiation on nonmarine silica and Miocene diatomite. *Palaos* 20: 198–206.
- Kjarn SP, Holm SL, and Myer EM (2004) Lake circulation and sediment transport in Lake Myvatn. *Aquatic Ecology* 38: 145–162.
- Lund JWG (1974) Phytoplankton as indicators of change in lakes. *Environment and Change* 2: 273–281.
- Mathers SJ, Chavez L, Alvarado L, and Inglethorpe DJ (1991) Detailed investigations of selected Costa Rican diatomites. British Geological Survey, Nottingham. *Unpublished report to Recope S. A. San Jose*, 74 pp.
- Pimraksa K and Chindaprasit P (2009) Lightweight bricks made of diatomaceous earth, lime and gypsum. *Ceramics International* 35: 471–478.
- Plunket GM, Whitehouse NJ, Hall VA, Brown DM, and Baille MGL (2004) A precisely-dated lake-level rise marked by diatomite formation in northeastern Ireland. *Journal of Quaternary Science* 19: 3–7.
- Qu R and Chen S (2005) Geological factors controlling ore and mineralization of diatomaceous earth deposits in Leizhou Peninsula. *West China Exploration and Engineering* 107: 88–89.
- Reading HG (ed.) (1978) *Sedimentary Environments and Facies*. London: Blackwell Scientific Publications.
- Robertson RHS (1945) Diatomite in Scotland. *The Quarry Managers' Journal* (April Issue): 451–458.
- Round FE (1957) The late-glacial and post-glacial diatom succession in the Kentmere Valley deposit. *New Phytologist* 56: 98–126.
- Ryves DB, Jewson DH, Sturm M, et al. (2003) Quantitative relationships between planktonic diatom communities and diatom assemblages in sedimenting material and surface sediments in Lake Baikal. *Limnology and Oceanography* 48: 1643–1661.
- Shang Y (2003) Tengchong diatomite deposit and its genesis. *Yunnan Geology* 22: 418–425.
- Shemesh AL, Burkle LH, and Froelich PN (1989) Dissolution and preservation of Antarctic diatoms and the effects on sediment thanatocoenoses. *Quaternary Research* 31: 288–308.
- Spencer AJ (1979) *Brick Architecture in Ancient Egypt*. Warminster: Aris & Phillips.
- Strelikova NI (1990) Evolution of diatoms during the Cretaceous and Paleogene periods. In: Simola H (ed.) *Proceedings of the 10th International Diatom Symposium*, pp. 195–204.
- Trauth MH, Maslin MA, Deino A, and Strecker MR (2005) Late Cenozoic moisture history of East Africa. *Science* 309: 2051–2053.
- Wendorf F and Schild R (1976) *Prehistory of the Nile Valley*. New York: Academic Press.
- Wild S, Gailius A, Hansen H, Pederson L, and Szwabowski J (1997) Pozzolan properties of a variety of European clay bricks. *Building Research and Information* 25: 170–175.
- Wilson HE (1972) *Regional Geology of Northern Ireland*. Belfast: Her Majesty's Stationary Office.

Salinity and Climate Reconstructions from Continental Lakes

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Introduction

Diatoms are widely used to reconstruct past environments because of their sensitivity to a range of environmental variables (see [Diatom Introduction](#)). Among the variables that exert a strong influence on diatoms are ionic concentration, expressed as salinity or conductivity, and ionic composition. Because the ionic concentration and composition of inland water bodies is affected by changes in effective moisture (precipitation minus evaporation, $P-E$), changes in the abundance of diatom species that vary in their salinity tolerances provide a tool for the reconstruction of past climate.

The use of diatoms to infer changes in salinity driven by climate relies on settings where effective moisture variation is moderate to large and produces relatively large changes in lake salinity. These conditions are most common in semiarid and arid regions, where evaporative forcing is strong, and in topographically closed lakes where changes in $P-E$ have the potential to concentrate or dilute dissolved salts. These conditions are widespread in the dry interiors of continents, in subtropical deserts, and in dry polar regions. Thus, diatom-inferred salinity can be used to reconstruct effective moisture from lake deposits in many regions of the globe.

In practice, the inference of climate from diatom-inferred salinity is complicated by the diverse controls on the hydrologic budgets of lakes and, occasionally, the accurate inference of salinity from diatoms may be obscured by other ecological influences. Nonetheless, diatom records have played a critical role in developing the understanding of the extent to which twentieth-century climate patterns are representative of long-term climate variation and of the patterns and controls on climate variability at decadal to orbital time scales.

Diatoms as Indicators of Salinity and Ion Composition

Salinity Classification of Diatoms

Diatoms have long been recognized as indicators of ionic concentration (salinity). [Kolbe \(1927\)](#), and subsequently [Hustedt \(1953\)](#), developed a system to classify diatoms according to their distribution in waters of varied salinity, the so-called halobion system. This system was primarily based on distributional patterns in sodium chloride (NaCl)-dominated marine and coastal environments, from marshes and estuaries receiving fresh surface runoff to the open ocean where salinity reaches $\sim 35 \text{ g l}^{-1}$. Although this halobion classification was applied to lacustrine diatoms in the mid-1900s, it is not well suited for diatoms in continental lakes because of the much broader range of salinity and ion composition found there.

Studies to characterize the salinity ranges of lacustrine diatoms expanded in the 1970s and 1980s, when the potential to reconstruct past climate change from lake sediments became more broadly recognized ([Ehrlich, 1978](#); [Gasse, 1977](#)). During the last few decades, studies of the distribution of diatoms

relative to gradients of salinity and ion composition have been carried out on all the continents and in lakes from the high latitudes to the tropics and have fueled the use of diatoms as paleoclimatic indicators.

Diatoms occur in the full range of lakes from dilute freshwater to hypersaline brines ([Figure 1](#)), and classification schemes reflect this diversity of environments. One of the most commonly applied classification systems splits lakes and their associated diatom floras into freshwater ($<0.5 \text{ g l}^{-1}$), subsaline ($0.5\text{--}3.0 \text{ g l}^{-1}$), hyposaline ($3.0\text{--}20 \text{ g l}^{-1}$), mesosaline ($20\text{--}50 \text{ g l}^{-1}$), and hypersaline ($>50 \text{ g l}^{-1}$) waters. Many species have distributional limits near the overlap of the subsaline and hyposaline ranges, and the diversity of taxa is low in mesosaline and hypersaline systems ([Fritz et al., 2010b](#)).

Although the distribution patterns of many diatoms are correlated with salinity, the mechanisms for this correlation are not well understood. Salinity imposes an osmotic gradient and hence an energetic cost on cells. At high salinity, diatoms manufacture proline or other osmolytes to deal with the osmotic gradient; many of these compounds are amino acids and rich in nitrogen. Thus, under high salinity, nutrient requirements may be higher ([Saros and Fritz, 2000](#)). Other compounds that are correlated with salinity gradients also may affect nutrient uptake by diatoms. For example, high concentrations of sulfate may interfere with molybdate uptake, a critical component of enzymes involved in nitrogen utilization. Many saline lakes in grasslands also have high dissolved organic carbon concentrations (DOC) and low chlorophyll relative to ambient nutrient concentrations ([Salm et al., 2009](#)). This suggests that DOC may form complexes with a critical nutrient that controls primary production and thereby affects diatom productivity and species composition. Thus, several lines of evidence suggest that one of the major effects of changing salinity may be via its impact on nutrient availability, uptake, and requirements.

Diatoms also can be classified according to their distribution in waters of differing anion composition, and some species are characteristic of chloride, sulfate, or carbonate-bicarbonate-dominated waters ([Figure 2](#)). The reasons for differences in anion association also are unclear but are likely related partly to anion impacts on nutrient availability ([Saros and Fritz, 2000](#)). Shifts in anion composition of lacustrine systems can be indicative of changes in source waters, particularly groundwater, or of progressive changes in brine composition driven by climate.

Quantitative Approaches for Salinity Reconstruction

Since the mid-1980s, transfer functions applied to fossil diatom assemblages have increasingly been used in paleoenvironmental studies to quantitatively reconstruct salinity, conductivity, or some other environmental variable of interest. These transfer functions are based on modern calibration data sets, which sample a group of contemporary lakes that span a salinity

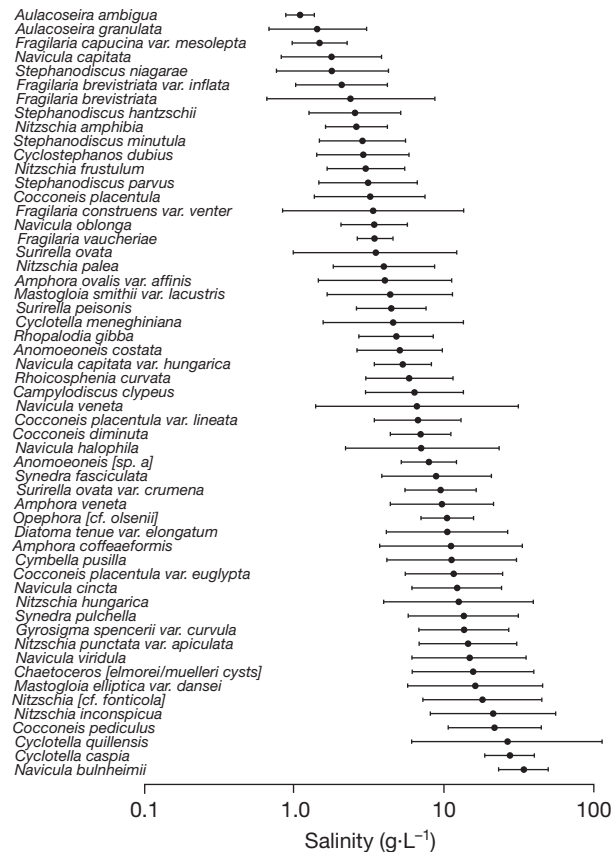


Figure 1 The salinity ranges (horizontal line) and salinity optima (dot) of selected diatom species from the Great Plains region of North America, generated from a survey of water chemistry and diatoms in the surface sediments of 66 regional lakes. Weighted averaging regression was used to calculate ranges and optima (see [Diatom Introduction](#)). Species are arranged from top to bottom in order of increasing salinity optima. Reproduced from Fritz SC, Juggins S, and Battarbee RW (1993) Diatom assemblages and ionic characterization of lakes in the northern Great Plains, N.A.: A tool for reconstructing past salinity and climate fluctuations. *Canadian Journal of Fisheries and Aquatic Sciences* 50: 1844–1856.

gradient and use these lakes to characterize the salinity range of diatom species (see [Diatom Introduction](#)) and the so-called salinity optimum, the salinity at which each diatom species is most abundant ([Figure 1](#)). A variety of statistical techniques can be used to generate transfer functions for the reconstruction of salinity or other environmental variables ([Birks, 2010](#)). The transfer function approach does not assume that salinity, or another reconstructed variable, is the proximate or physiological variable affecting diatom distribution, but simply that diatom distribution is highly correlated with salinity and that this relationship is stationary through time. In most cases, a regional transfer function cannot be independently tested by applying the function to a new group of lakes or by comparing a reconstruction from a sediment core with historic measurements of salinity from the same lake, although occasionally these sorts of data are available (e.g., [Fritz, 1990](#)). Therefore, statistical techniques for error estimation, such as bootstrapping or jack-knifing, are commonly used to evaluate transfer function performance ([Birks, 2010](#); [Fritz et al., 2010b](#)).

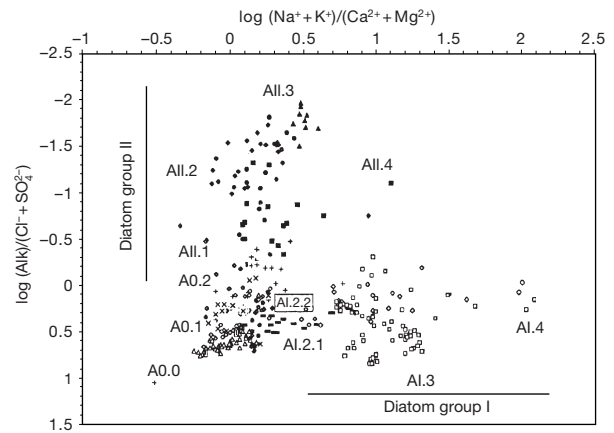


Figure 2 Diatom-inferred cation composition (x-axis) and anion composition (y-axis) inferred from a series of diatom samples from sediment cores from 12 African lakes. Sediment samples from each of the 12 lakes are represented by a different symbol. A weighted average calibration model developed from over 300 modern lakes in Africa was used to reconstruct cation and anion composition from these sediment samples (see [Diatom Introduction](#)). The groups A0, AI, AII represent different combinations of lake water cation and anion dominance, as indicated in the figure. Subgroups 1–4 represent progressively increasing salinity within a given group. Reproduced from Gasse F (2002) Diatom-inferred salinity and carbonate oxygen isotopes in Holocene waterbodies of the western Sahara and Sahel (Africa). *Quaternary Science Reviews* 21: 737–767.

Transfer functions for salinity reconstruction are effective in lakes that have undergone moderate to large fluctuations in salinity that exceed the standard error of prediction. Quantitative techniques also work best in systems where other water-chemistry variables that have a strong impact on diatoms, such as pH or nutrients, are less likely to show large changes over time. In theory, statistical approaches, such as variance partitioning, can account for the impact of multiple environmental influences on species distribution. In practice, however, modern data sets are constructed to optimize the environmental gradient of interest, in this case salinity, and other ecologically important variables may not be adequately accounted for. In these cases, the assumption that the relationship between species and salinity does not change with time may be invalid, such as has been demonstrated for diatom-based transfer functions for temperature ([Bigler et al., 2002](#)). Because of these issues, salinity reconstructions are most robust when they are corroborated by inferences from other proxies, such as ostracodes, chironomids, sedimentology, or sediment geochemistry ([Aebly and Fritz, 2009](#); [Gasse, 2002](#); [Kropelin et al., 2008](#)).

Quantitative reconstructions depend on having a modern calibration data set that accurately defines the environmental ranges and tolerances of diatom species. For most regions, the size of calibration data sets is less than 100 lakes, and although such a sample size can generate a statistically robust transfer function, it may not adequately capture the true optima of all taxa. In some regions, such as the tropical Andes, the density of modern samples is not yet sufficient to allow quantitative reconstructions, and the large number of apparently endemic species makes generalizations from the floras of other regions difficult. Finally, given the large range of environmental

change in the past, some regions may no longer contain aquatic environments that are modern analogs for past environmental conditions. Thus, in Africa and Mexico, some of the dominant diatom taxa in fossil records do not occur in the modern flora (Gasse, 2002; Metcalfe et al., 2002), even given large spatially distributed modern data sets.

Diatom Preservation in Saline Lakes

Although diatoms are a common component of the algal community in lakes from low to high salinity, they are not always well preserved in sedimentary records. Diatom dissolution is highest in sodium carbonate solutions, where high pH (>9) enhances the dissociation of silicic acid, but dissolution also occurs in other salt solutions (Barker et al., 1994). Brines can induce the formation of diagenetic silicates (e.g., zeolites and smectite), leading to alteration or dissolution of diatom assemblages. These processes can alter not only total diatom abundance in sediments, but also species composition because diatom species are differentially sensitive to dissolution and diagenesis (Barker et al., 1994; Ryves et al., 2001). In some cases, dissolution indices can be developed to correct for some of the bias imposed by taphonomic processes (Ryves et al., 2009).

Lakes, Salinity, and Climate

Climate affects the water budget of lakes via its influence on precipitation, surface evaporation, and groundwater flow. Shifts in a lake's water balance, in turn, alter lake salinity through concentration or dilution of ions in solution. The magnitude of the climatic imprint depends on the morphometry of the lake and its drainage basin; both lake surface area and volume, for example, affect the degree to which precipitation or evaporation influence lake level and lake salinity. Overall, the best candidates for climate reconstruction from diatom-inferred salinity techniques are subsaline to hyposaline lakes in semiarid to arid climates because variations in precipitation and evaporation cause salinity changes that are sufficiently large to produce shifts in diatom species composition. The rate of lake response to climate is a product of the lake's water residence time and the relative balance of surface flow relative to groundwater. In lakes where groundwater is a large component of the water budget, lake response is governed by the response time of the groundwater system, and short-term climatic change may have very little impact on the water and budget (Reed et al., 1999). Connection to a groundwater flow system also allows for loss of salts via subsurface outflow, whereas in a terminal basin (a basin without surface or subsurface outflow), salts build up over time.

Globally, subsaline and saline lakes outnumber freshwater lakes (Meybeck, 1995), and systems appropriate for paleoclimatic study are common in most of the regions of the world where water availability is limited. This includes the continental interiors of the Americas, Africa, Australia, and Eurasia, as well as the high-latitude regions of Antarctica and the Arctic. In the next section, examples of a range of lake systems are discussed in which diatom-inferred salinity has been used to infer paleoenvironmental and paleoclimatic history during the Quaternary.

Quaternary Environmental History as Inferred from Diatoms

Quaternary Climate Variation at Orbital and Millennial Time Scales

Shifts between freshwater and saline diatom taxa have been used effectively to infer moisture variation on orbital and millennial time scales in extant and former lake basins. In the Owens Lake basin in western North America (Bradbury, 1997), peaks in freshwater planktic diatoms and inferred deepwater conditions are broadly correlated with cold periods in the SPECMAP oxygen isotope record after ~400 000 years BP, when tectonic processes likely produced an increase in basin depth (Figure 3). Prior to that time, the evidence suggests that the basin was much shallower, and the correlation between intervals of increased planktic diatoms and cold climatic periods is not as strong. In Lake Titicaca in the tropical Andes, the increased abundance of freshwater planktic diatoms (Figure 4)

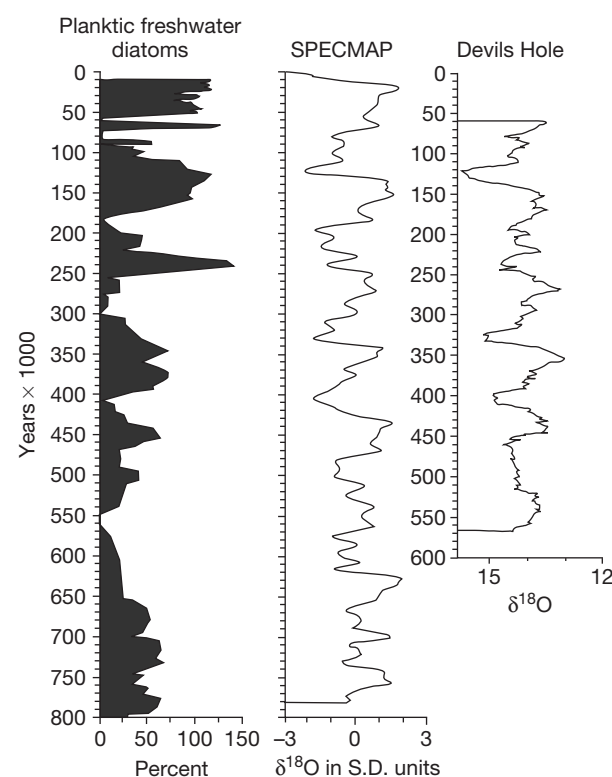


Figure 3 A comparison of the abundance of freshwater planktic diatoms in a drill core from Owens Lake, California compared with the SPECMAP $\delta^{18}\text{O}$ record of global temperature change – high isotopic values represent global cold periods, whereas depleted values are warm periods. The $\delta^{18}\text{O}$ record from a calcite vein at Devils Hole, Wyoming also provides a chronology of temperature change; in this case with more depleted (smaller) values representing colder periods. The relative proportion of diatoms in each sample was calculated using a standard sample count of 300 diatom valves. In a few samples with very high diatom concentrations, more than 300 valves were counted and thus percentages exceed 100%. Reproduced from Bradbury JP (1997) A diatom-based palaeohydrologic record of climate change for the past 800 k.y. from Owens Lake, California. *Geological Society of America, Special Paper* 317: 99–112.

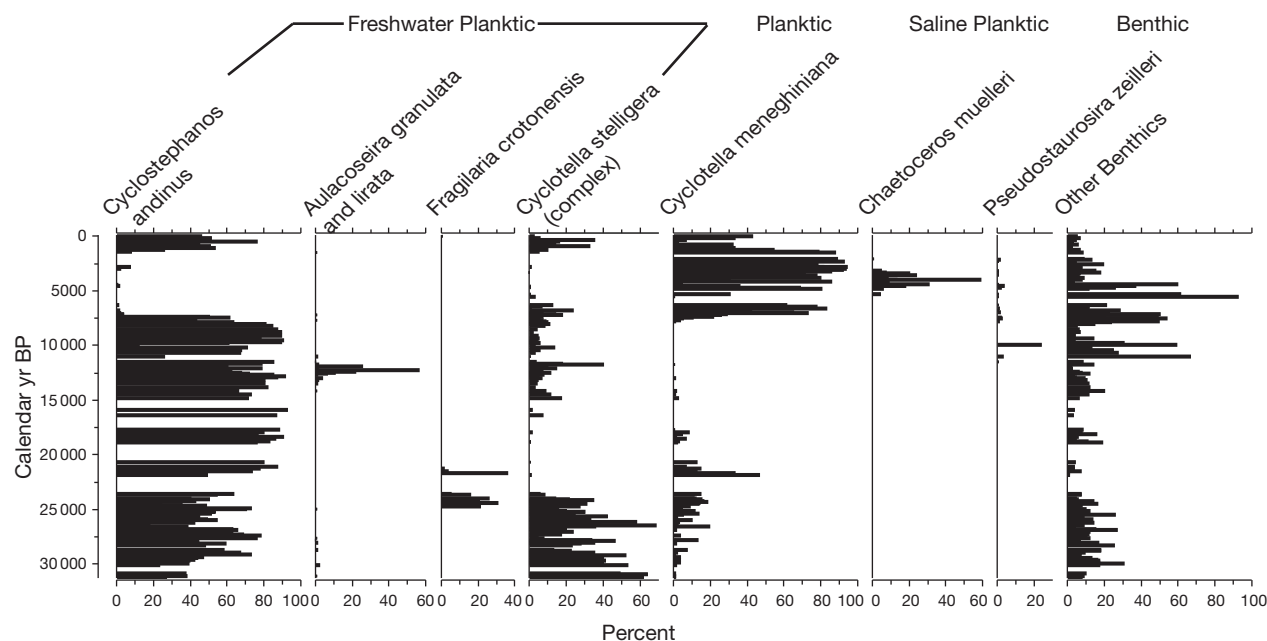


Figure 4 The diatom stratigraphy of the last ~30 000 years from Lake Titicaca, Bolivia-Peru. The dominance of freshwater planktic diatoms is inferred to represent times when the lake was fresh and overflowing, whereas benthic (*Pseudostaurosira zelleri* and other taxa) and saline (*Chaetoceros muelleri*) diatoms dominate when the lake-level fell below the outlet threshold (see [Baker et al., 2001](#) for additional information).

indicates fresh overflowing conditions associated with increased precipitation, whereas benthic diatoms and the saline planktic taxon, *Chaetoceros muelleri*, indicate periods when lake-level fell below the outlet and salinity built up. In this case, the wet intervals are coincident with global cold periods, including the Last Glacial Maximum, when cold temperatures reduced evaporation and high summer insolation and cold Atlantic Ocean sea-surface temperatures (SSTs) enhanced the South American summer monsoon. During the contrasting conditions of global interglacial periods, lake-level declines of up to 240 m (modern water depth is 285 m) were sufficient to more than double lake water salinity ([Baker et al., 2001](#); [Fritz et al., 2007](#)); more discussion of the Lake Titicaca record is given in Mackay (this volume). Millennial-scale variations in lake level also are evident in Lake Titicaca and other lakes in the drainage basin and can be linked to climate variation in the North Atlantic region ([Ekdahl et al., 2008](#); [Fritz et al., 2010a](#)). In Africa, massive reductions in lake level are evident during the late Pleistocene in the diatom record from Lake Malawai, as well as in other proxies analyzed in the same core ([Stone et al., 2010](#)). Similarly, a 50 000-year diatom record from Tanzania shows the influence of global temperature change on regional hydrology ([Barker et al., 2003](#)). Here changes in lake level and lake conductivity show a strong relationship to SST change in the SW Indian Ocean ([Figure 5](#)), which forces changes in the position of the intertropical convergence zone (ITCZ) and hence influences precipitation in the African tropics.

Late-Glacial and Holocene Climate Change

Changes in salinity inferred from diatom records have played a prominent role in documenting hydrologic changes in the late-glacial period through the Holocene, particularly in

the Americas and Africa. In northern Mexico and the Great Basin of the southwestern United States, diatom records suggest that the late-glacial period was wet as a result of the southerly displacement of the westerly storm tracks, followed by drying in the early to mid-Holocene as the storm tracks moved northward ([Bradbury et al., 2001](#); [Metcalf et al., 2002](#)). In central Mexico, the opposite climate pattern is evident, with a dry late-glacial period because of a southward displacement of the ITCZ and a reduction in trade wind sources of moisture. Holocene salinity and moisture patterns across Mexico are spatially complex, as a result of the differential influence of rising sea level, dating uncertainties, and increasing human impact on the landscape in the late Holocene ([Metcalf et al., 2000](#)).

In Africa, diatom analysis has been instrumental in documenting massive precipitation changes during the late Quaternary, including the greening of the Sahara and Sahel in the early to middle Holocene ([Gasse, 2002](#)). At Bougdouma in Niger, for example, shifts between alkiliphilous benthic diatoms and freshwater planktic species suggest fluctuating water levels following the last period of glaciation ([Figure 6](#)). Stable wet conditions developed in the early Holocene (~10 300 years BP), followed by drying after 7300 years BP, a trend that is widespread throughout the Sahara and Sahel. Several wet pulses that produced a rising water table interrupted the drying trend, but conductivity and inferred aridity increased after 3800 years BP, with desiccation of the basin during the last 1000 years. These trends reflect variation in monsoon intensity across northwest Africa, but regional differences in the timing of lake salinity change reflect differences in the rate of response of different linked aquifer-lake systems.

The impacts of groundwater on lake response to climate have been highlighted in several paleolimnological studies. In some lakes of the Ethiopian Rift Valley, diatom-inferred water

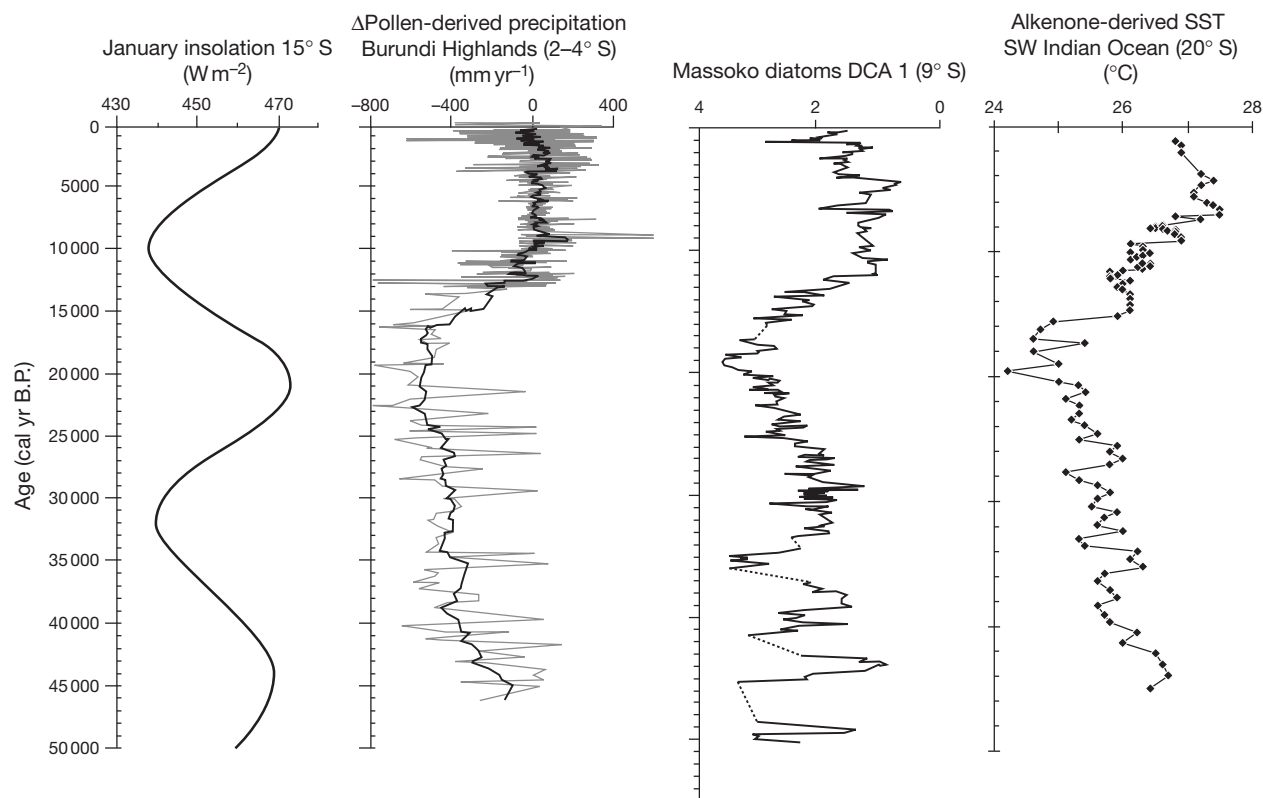


Figure 5 The diatom stratigraphy of Lake Massoko, Tanzania compared with insolation and sea-surface temperature forcing. The detrended correspondence axis (DCA) scores summarize the patterns of change in the diatom species assemblage – samples with high DCA axis 1 scores are samples with high diatom-inferred conductivity and represent times of lowered P – E, an inference that is corroborated by pollen data. Reproduced from Barker P, Williamson D, Gasse F, and Gibert E (2003) Climatic and volcanic forcing revealed in a 50,000-year diatom record from Lake Massoko, Tanzania. *Quaternary Research* 60: 368–376.

chemistry does not show patterns of salinity change that would be predicted based on independent evidence for precipitation variation (Telford et al., 1999). For example, an interval of high diatom-inferred conductivity from Lake Awassa corresponds with strand lines in an adjacent basin that document a contemporaneous interval of high precipitation (Figure 7). Increased inputs of saline groundwater to one lake but not the other are the likely explanation. Similarly, a karst lake in the Konya Basin of Turkey showed little change in diatom-inferred conductivity during the late Pleistocene, probably because of relatively constant inflow of fresh groundwater, whereas a neighboring lake with limited groundwater flow showed evidence of changes in conductivity and lake level indicative of changes in effective moisture (Reed et al., 1999).

Complex salinity responses to climate change also are evident in intermittently meromictic lakes in West Greenland. Here, the amplitude of salinity variation is apparently affected by the stability of stratification, producing a nonlinear response of lake salinity to climate change (Aebly and Fritz, 2009; McGowan et al., 2003). In some intervals, increased diatom-inferred salinity correlates with evidence of increased precipitation from paleoshorelines. Complex groundwater interactions are unlikely in this region because of permafrost. In this setting, the observed response likely resulted from the coalescence of a freshwater lake with a lake of higher salinity

during times of increased precipitation or from the reworking of shoreline salt deposits as lake level rose.

The use of diatom-inferred salinity to infer changes in effective moisture also can be affected by the sensitivity of diatom assemblages to changes in salinity driven by climate. At high salinity, diatom diversity is commonly quite low, and the ecological tolerances of many high-salinity taxa are broad. Thus, species composition may not change as salinity changes. At Moon Lake, North Dakota, USA, for example, diatom-inferred salinity is very high during the mid-Holocene (Figure 8), indicative of the dry climate that was characteristic of the mid-latitudes of North America. But, the low variability in mid-Holocene diatom-inferred salinity probably results because changes in salinity caused by precipitation variation are not sufficiently large to change the composition of the diatom assemblage (Laird et al., 1996).

Diatom records from the North American Great Plains also provide an example of the use of diatom-inferred salinity to evaluate natural drought variation and the degree to which the twentieth century climate patterns are representative of the natural variability of climate. Records from a suite of lakes in the United States and the Canadian prairies clearly show that droughts equivalent in magnitude to the Dust Bowl droughts of the 1930s and 1940s were recurrent during the last millennium or two (Laird et al., 2003). These records also suggest that

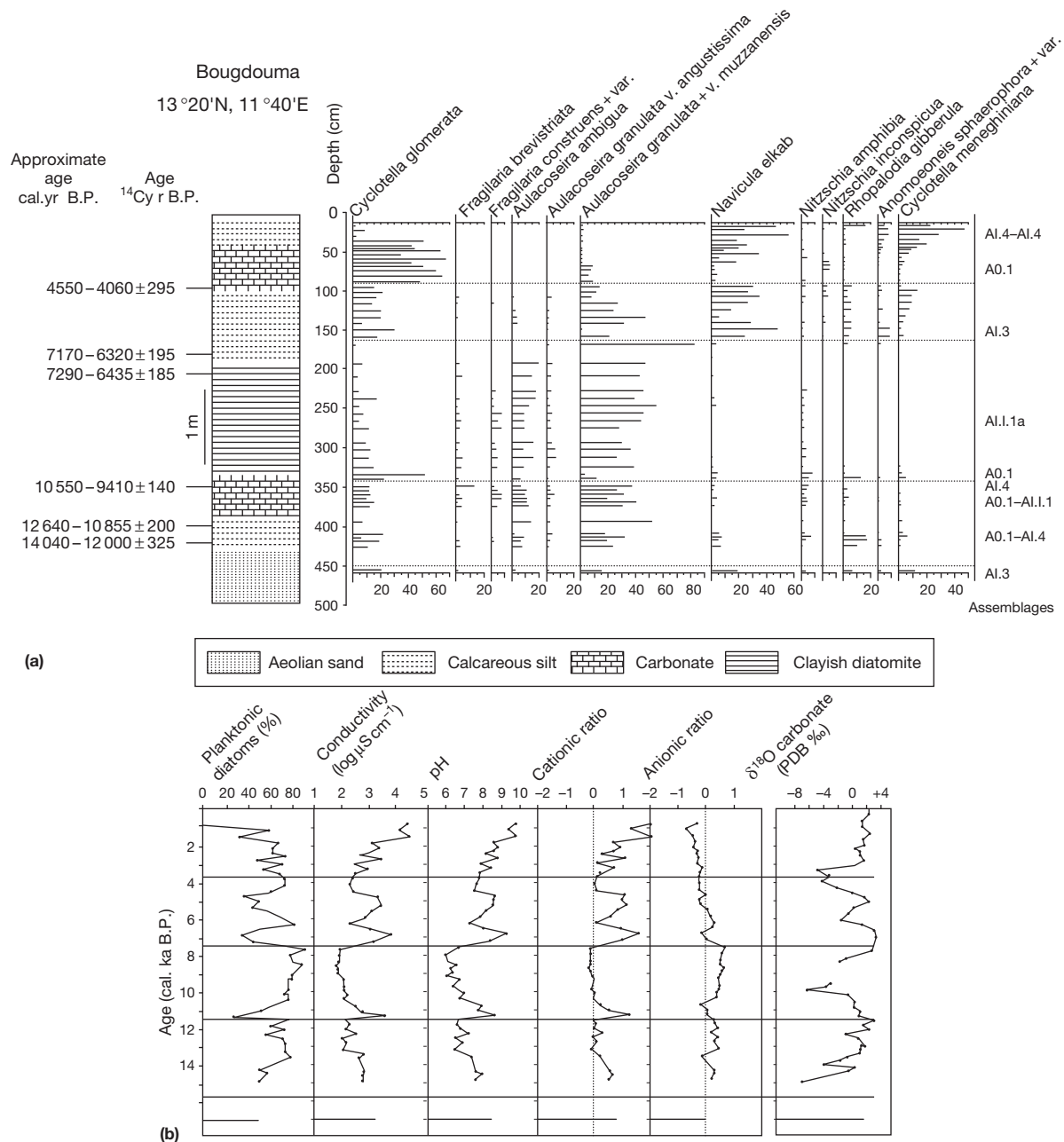


Figure 6 The (a) diatom stratigraphy of Bougdouma in southern Niger and (b) various water-chemistry variables inferred from the diatom stratigraphy using transfer functions (lower figure). The cationic and anionic ratios are defined in Figure 2. The assemblages defined on the right-hand part of the figure (A0, AI, All) represent the inferred cation and anion water types as defined in Figure 2. Reproduced from Gasse F (2002) Diatom-inferred salinity and carbonate oxygen isotopes in Holocene waterbodies of the western Sahara and Sahel (Africa). *Quaternary Science Reviews* 21: 737–767.

drought was more prolonged in the recent past than during the twentieth century; some droughts apparently persisted for a century or more. One of the major difficulties with these high-resolution records, however, is in the correlation of the pattern of change among multiple sites; the errors associated with a radiocarbon chronology prohibit a clear correlation of changes at multidecadal scales. Even directional shifts that are persistent can be difficult to interpret in climatic terms. Thus, a major shift at the onset of Little Ice Age about 800 years BP is manifested as an increase in salinity at one site and a decrease at another nearby lake (Figure 9), and it is unclear whether the

differences reflect a sharp climatic gradient, or whether they are more likely due to differing hydrological responses to the same climatic forcing (Fritz et al., 2000).

Conclusions

The use of diatoms to infer past moisture variation in continental settings has increased greatly during the last two decades, and appropriate basins are widespread in semiarid and arid regions worldwide. In these regions, lacustrine records

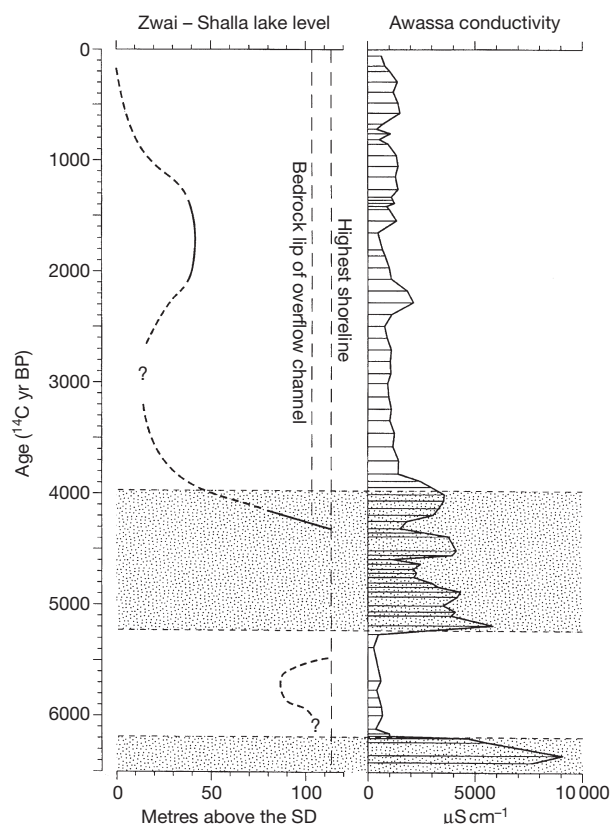


Figure 7 The diatom-inferred conductivity of a lake in Ethiopia compared with shoreline evidence for lake-level change in a nearby basin. Reproduced from Telford RJ, Lamb HF, and Mohammed MU (1999) Diatom-derived palaeoconductivity estimates for Lake Awassa, Ethiopia: Evidence for pulsed inflows of saline groundwater? *Journal of Paleolimnology* 21: 409–421.

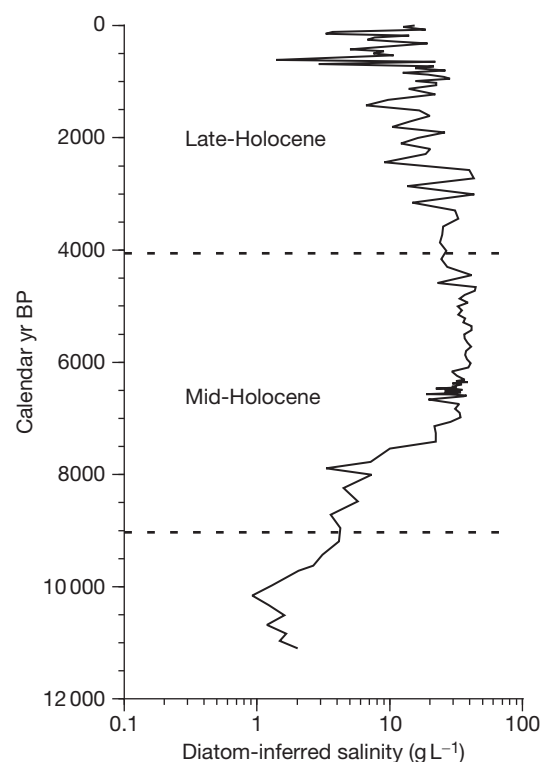


Figure 8 The diatom-inferred salinity of Moon Lake, North Dakota, USA during the last ~12000 years (Laird et al., 1996).

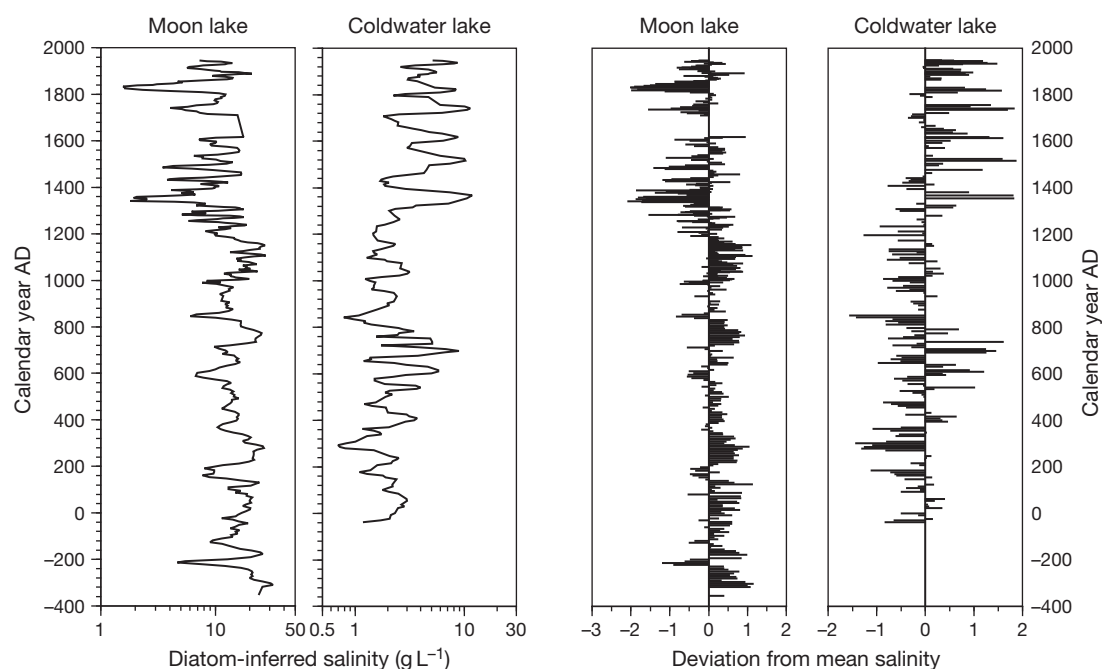


Figure 9 Diatom-inferred salinity spanning the last ~2000 years of two lakes in North Dakota, USA that are within 200 km of one another. The right-hand panel shows the deviation from the long-term mean salinity. Note that Moon Lake became fresher at ~AD 1200, whereas Coldwater Lake becomes saltier at about the same time (Fritz et al., 2000).

often provide the best archives of long-term continental climate variation. Although some lakes in extremely dry climates may have discontinuous sedimentary records, in many basins, sedimentation rates are sufficiently rapid to allow the sedimentary record to be resolved at a subdecadal scale.

Although more studies are needed on how salinity and other ecologically important variables interact to influence diatoms, in many settings knowledge of diatom ecological tolerances is sufficient to enable robust reconstructions of qualitative changes in lake salinity. In some regions, calibration data sets have been developed to allow quantitative salinity reconstructions that are reasonably good at identifying the direction and relative magnitude of salinity change. In general, salinity change can be reconstructed from diatom assemblages with a high degree of confidence.

The inference of climate from salinity change is considerably more difficult, because of the potentially complex series of influences on the hydrologic budget of individual lakes. Although many of the issues involved in disentangling climate from biological and geochemical records are recognized (Fritz, 2008), more frequent application of modeling and empirical studies to quantify their potential impact in individual sites or regions would improve many paleolimnological interpretations. The tandem use of other proxies can help to constrain environmental interpretation from diatom records, although in some cases, reconciliation of apparently divergent inferences is not clear. Such cases also point to the need for more frequent integration of paleolimnology with modern process studies and modeling efforts. Overall the number of lacustrine records is still relatively limited, particularly at high temporal resolution and for periods much older than the Last Glacial Maximum. Thus, although the diatom paleoclimatology field has expanded in recent years, there is still a considerable distance to travel to unravel the patterns of environmental change across space and time, and the hierarchy of factors that produce them.

See also: Diatom Introduction. **Diatom Records:** Large Lakes. **Lake Level Studies:** Africa during the Late Quaternary; Latin America; North America; Overview. **Paleoclimate:** Introduction. **Paleolimnology:** Overview of Paleolimnology.

References

- Aebly F and Fritz SC (2009) The paleohydrology of West Greenland for the past 8000 years. *The Holocene* 19: 91–104.
- Baker PA, Seltzer GO, Fritz SC, et al. (2001) The history of South American tropical precipitation for the past 25,000 years. *Science* 291: 640–643.
- Barker P, Fontes J-C, Gasse F, and Druart J-C (1994) Experimental dissolution of diatom silica in concentrated salt solutions and implications for paleoenvironmental reconstruction. *Limnology and Oceanography* 39: 99–110.
- Barker P, Williamson D, Gasse F, and Gibert E (2003) Climatic and volcanic forcing revealed in a 50,000-year diatom record from Lake Massoko, Tanzania. *Quaternary Research* 60: 368–376.
- Bigler C, Larocque I, Peglar SM, Birks HJB, and Hall RI (2002) Quantitative multiproxy assessment of long-term patterns of Holocene environmental change from a small lake near Abisko, northern Sweden. *The Holocene* 12: 481–496.
- Birks HJB (2010) Numerical methods for the analysis of diatom assemblage data. In: Smol JP and Stoermer EF (eds.) *The Diatoms: Applications for the Environmental and Earth Sciences*, 2nd edn., pp. 23–56. Cambridge: Cambridge University Press.
- Bradbury JP (1997) A diatom-based paleohydrologic record of climate change for the past 800 k.y. from Owens Lake, California. *Geological Society of America*, 317: 99–112. Special Paper.
- Bradbury JP, Grosjean M, Stine S, and Sylvestre F (2001) Full and late glacial lake records along the PEP 1 transect: Their role in developing interhemispheric paleoclimate interactions. In: Markgraf V (ed.) *Interhemispheric Climate Linkages*, ch. 216, pp. 265–292. Boulder, CO: Academic Press.
- Ehrlich A (1978) The diatoms of the hyperhaline Solar Lake (NE Sinai). *Israel Journal of Botany* 27: 1–13.
- Ekdahl E, Fritz SC, Baker PA, Riggsby CA, and Coley K (2008) Holocene multi-decadal to millennial-scale hydrologic variability on the South American Altiplano. *The Holocene* 18: 867–876.
- Fritz SC (1990) Twentieth-century salinity and water-level fluctuations in Devils Lake, N. Dakota: A test of a diatom-based transfer function. *Limnology and Oceanography* 35: 1771–1781.
- Fritz SC (2008) Deciphering climate history from lake sediments. *Journal of Paleolimnology* 39: 5–16.
- Fritz SC, Baker PA, Ekdahl E, Seltzer GO, and Stevens LR (2010a) Millennial-scale climate variability during the last glacial period in the tropical Andes. *Quaternary Science Reviews* 29: 1017–1024.
- Fritz SC, Cumming BF, Gasse F, and Laird KR (2010b) Diatoms as indicators of hydrologic and climatic change in saline lakes. In: Smol JP and Stoermer EF (eds.) *The Diatoms: Applications for the Environmental and Earth Sciences*, 2nd edn., pp. 186–208. Cambridge: Cambridge University Press.
- Fritz SC, Baker PA, Seltzer GO, et al. (2007) Quaternary glaciation and hydrologic variation in the South American tropics as reconstructed from the Lake Titicaca drilling project. *Quaternary Research* 68: 410–420.
- Fritz SC, Ito E, Yu Z, Laird KR, and Engstrom DR (2000) Hydrologic variation in the northern Great Plains during the last two millennia. *Quaternary Research* 53: 175–184.
- Gasse F (1977) Evolution of Lake Abhe (Ethiopia and TFAI) from 70,000 BP. *Nature* 2: 42–45.
- Gasse F (2002) Diatom-inferred salinity and carbonate oxygen isotopes in Holocene waterbodies of the western Sahara and Sahel (Africa). *Quaternary Science Reviews* 21: 737–767.
- Hustedt F (1953) Die Systematik der Diatomeen in ihren Beziehungen zur Geologie und Ökologie nebst einer Revision des Halobien-systems. *Botanisk Tidskrift* 47: 509–519.
- Kolbe RW (1927) Zur Ökologie, Morphologie und Systematik der Brackwasser-Diatomeen. *Pflanzenforschung* 7: 1–146.
- Kropelin S, Verschuren D, Lezine AM, Eggemont H, Cocquyt C, and Francus P (2008) Climate-driven ecosystem succession in the Sahara: The past 6000 years. *Science* 320: 765–768.
- Laird KR, Cumming FF, Wunsam S, et al. (2003) Lake sediments record large-scale shifts in moisture regimes across the northern prairies of North America during the past two millennia. *Proceedings of the National Academy of Sciences* 100: 2483–2488.
- Laird KR, Fritz SC, Grimm EC, and Mueller PG (1996) Century-scale paleoclimatic reconstruction from Moon Lake, a closed-basin lake in the northern Great Plains. *Limnology and Oceanography* 41: 890–902.
- McGowan S, Ryves DB, and Anderson NJ (2003) Holocene records of effective precipitation in West Greenland. *The Holocene* 13: 239–250.
- Metcalfe SE, O'Hara SL, Caballero M, and Davies SJ (2000) Records of Late Pleistocene-Holocene climatic change in Mexico – A review. *Quaternary Science Reviews* 19: 699–721.
- Metcalfe S, Say A, Black S, McCulloch R, and O'Hara S (2002) Wet conditions during the last glaciation in the Chihuahuan Desert, Alta Babicora, Basin, Mexico. *Quaternary Research* 57: 91–101.
- Meybeck M (1995) Global distribution of lakes. In: Lerman A, Imboden D, and Gat J (eds.) *Physics and Chemistry of Lakes*, pp. 1–35. Berlin: Springer Verlag.
- Reed JM, Roberts N, and Leng MJ (1999) An evaluation of the diatom response to Late Quaternary environmental change in two lakes in the Konya Basin, Turkey, by comparison with stable isotope data. *Quaternary Science Reviews* 18: 631–646.
- Ryves DB, Battarbee RW, and Fritz SC (2009) The dilemma of disappearing diatoms: Incorporating diatom dissolution data into palaeoenvironmental modelling and reconstruction. *Quaternary Science Reviews* 28: 120–136.
- Ryves DB, Juggins S, Fritz SC, and Battarbee RW (2001) Experimental diatom dissolution and the quantification of microfossil preservation in sediments. *Palaeogeography, Palaeoclimatology, Palaeoecology* 172: 99–113.
- Salm CR, Saros JE, Fritz SC, Osburn CL, and Reineke DM (2009) Phytoplankton productivity across prairie saline lakes of the Great Plains (USA): A step toward deciphering patterns through lake classification models. *Canadian Journal of Fisheries and Aquatic Sciences* 66: 1435–1448.

- Saros JE and Fritz SC (2000) Changes in the growth rates of saline-lake diatoms in response to variation in salinity, brine type and nitrogen form. *Journal of Plankton Research* 22: 1071–1083.
- Stone JR, Westover KS, and Cohen AS (2010) Late Pleistocene paleohydrography and diatom paleoecology of the central basin of Lake Malawi, Africa. *Palaeogeography, Palaeoclimatology, Palaeoecology* 303: 51–70. <http://dx.doi.org/10.1016/j.palaeo.2010.1001.1012>.
- Telford RJ, Lamb HF, and Mohammed MU (1999) Diatom-derived palaeoconductivity estimates for Lake Awassa, Ethiopia: Evidence for pulsed inflows of saline groundwater? *Journal of Paleolimnology* 21: 409–421.

Use in Archaeology

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Introduction

Diatom analysis has been used in archaeological investigations for almost 60 years (for discussion and references see the review of [Battarbee, 1988](#)). Like other Quaternary paleoecological techniques, the results from off-site diatom paleoecological studies are often relevant to environmental archaeology. However, only diatom research that is integrated with field archaeology is considered relevant here. Specific collaboration between field archaeologists excavating ancient sites and archaeological sediments and diatomists interested in site environments and other aspects of archaeology has led to an increasing use of the technique in environmental archaeology. Unfortunately, the majority of published diatom-archaeological studies are of necessity placed in the appendices of archaeological site reports, making it difficult to gain an impression of the frequency with which the technique is applied in this context. However, diatom analysis is now more commonly known and used in archaeology, and often, particularly in the prehistoric context, it forms part of environment-led archaeological investigations ([Nayling and Caseldine, 1997](#)).

The applications of diatom analysis in archaeology have been reviewed by several authors ([Battarbee, 1988](#); [Juggins and Cameron, 2010](#); [Mannion, 1987](#); [Miller and Florin, 1989](#)). In this article, a number of case studies of on-site diatom-archaeological investigations are presented as examples.

Diatoms as Indicators of the Site Environment

As mentioned earlier, many diatom-paleoecological studies related to Holocene environmental reconstruction make a reference to catchment archaeology (human impact) as a factor in driving environmental change (e.g., [de Wolf and Cleveringa, 1996](#)) and indeed may discuss the archaeological or historical documentary evidence. Diatom environmental studies integrated with archaeological excavations are less common. An example of such a project is that of [Douglas et al. \(2004\)](#) who investigated the impact of prehistoric Inuit whalers upon Arctic freshwater ecosystems. Despite their low population densities and the nomadic nature of Inuit whalers, it was shown through a combination of archaeological and paleolimnological data that human settlement markedly changed pond water ecology from an early date. Winter settlements were constructed from the bones of bowhead whales on Somerset Island in the high Canadian Arctic between about 1200 and 1600. Archaeological excavation, recovery of artifacts, and radiocarbon dating document this occupation and allow for estimation of the population size. Diatom analysis ([Figure 1](#)) indirectly records marked changes in water chemistry and diatom habitat. Although the whalers abandoned the area over 400 years ago, the pond today retains elevated nutrient concentrations and biota atypical of the region.

In diatom-based studies of Danish lakes, [Bradshaw et al. \(2006\)](#) have shown that in the Bronze Age and Iron Age there was moderate eutrophication. Large increases in nutrient levels accompanied changes in agriculture and the retting of hemp and flax in the Medieval period. These studies demonstrate that human activities in lake catchments had a long-term (millennial) impact and that significant eutrophication is not restricted to the modern period (post-1800).

It is more common to find diatom analysis used in archaeological studies from urban contexts in towns in Europe. An example of this kind of analysis is work carried out by [Demiddele and Erynck \(1993\)](#) using diatoms as ecological indicators in an excavation of Roman and Medieval Oudenberg in the west of Flanders. During the Roman period, waterlain sediments in the town showed a high proportion of marine species reaching the town through tidal creeks. However, during the Medieval period, freshwater and low salinity species dominated the assemblages, indicating the disappearance of a marine influence with implications for water transport to the town. In addition, the diatoms from the medieval contexts corroborate evidence from other sources that there was heavy organic pollution of the water courses. Diatom analysis was therefore of relevance here, both in looking at the influence of tidal waters on human activities and on considering the impact of humans on site water quality. Diatoms have also been used for archaeological environmental reconstruction over larger areas in landscape archaeology projects. This is exemplified by work carried out over several seasons of excavations on wetlands bordering the Severn Estuary, UK, where diatom analysis supported evidence for large-scale Roman land reclamation ([Rippon, 2000](#)).

Diatom Analysis of Artifacts and Ancient Building Materials

Diatom analysis has been used in the direct study of artifacts, particularly in the investigation of pottery clay provenance and pottery typology, and also in identifying the sources of materials and the manufacturing techniques used in ancient buildings.

Where the firing temperature is low enough to allow the survival of diatom valves, diatom analysis of pottery can assist in the identification of clay sources or in the understanding of the provenance of finished pottery. Where adequate diatom assemblages are preserved, diatom assemblages can also be used to assist in the establishment of typologies for pottery or in the comparison of typologies with pottery style-based classification schemes ([Alhonen et al., 1980](#)). In addition to diatoms derived directly from clay, the mixture of diatoms derived from tempering materials and from water used in the manufacturing process can form part of the species assemblage. The use of diatoms in the analysis of pottery, which,

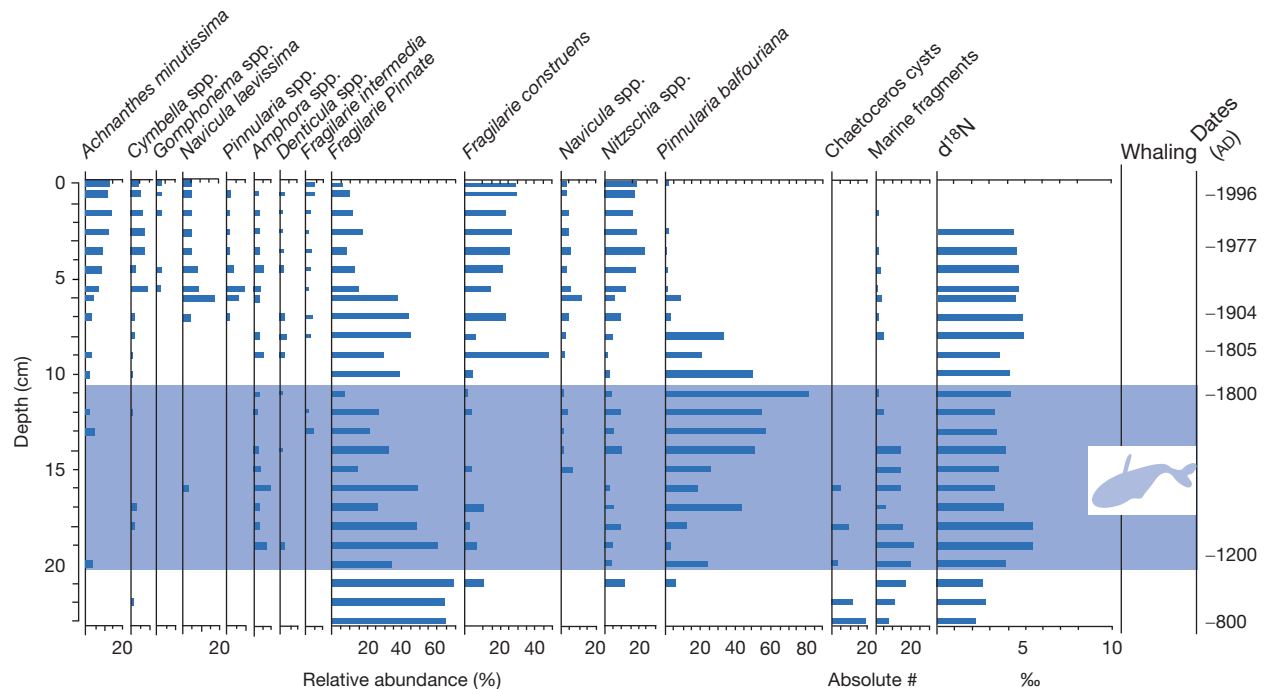


Figure 1 Diatom analysis of lake sediments from the Canadian Arctic showing the impact of Inuit whalers on water chemistry and diatom habitat. Reproduced from Douglas MSV, Smol JP, Saville JM, and Blaise JM (2004) Prehistoric Inuit whalers affected Arctic freshwater ecosystems. *Proceedings of the National Academy of Sciences* 101: 1613–1617. Copyright (2004) National Academy of Sciences, USA, with kind permission.

because of the requirement for a low-temperature firing, is usually of prehistoric age, has been reviewed by Battarbee (1988) and by Juggins and Cameron (2010). Stilborg et al. (1997) published a microfossil analysis of late Roman Iron Age pottery (AD 200–400) from a region of present-day Denmark. Just the presence or absence of siliceous and calcareous microfossils was used to group and separate pottery sherds. Two groups were established and the diatomaceous class, known as ‘D-ware,’ was further subdivided into sherds containing exclusively nonmarine diatoms and a subgroup containing a mixture of marine and nonmarine species. From this classification, it was also possible to infer the clay sources used for the pottery. In some of the classes of pottery, it was established that a mixture of clays was used during manufacture.

In a similar way, diatom analysis has been used to examine ancient building materials (Flower, 2006). Flower examined the diatom content of ancient mud bricks. These were used in ancient Egypt from approximately 5000 BP (Figure 2) and such bricks continue to be made in this way in the region today. The bricks are made from a mixture of mud, sand, water, and plant material. Diatom analysis of the bricks thus offers the possibility of investigating the provenance and uses of these building materials. Analysis of diatoms from a mud brick from a temple at Dimai showed that the diatom ‘mud’ component was almost certainly extracted from somewhere nearby, now terrestrial, diatomite that had accumulated in a former large lake. In contrast, the few, predominantly planktonic diatoms found in brick from a 12th Dynasty Pyramid at Hawara were thought to have been introduced from Nile water during manufacture. Flower notes that his results are preliminary, being based on a small number of samples. However, he



Figure 2 Buildings at the Ptolemaic Temple at Dimai, Egypt, dating and showing mud brick construction. Photograph taken by Dr. Roger Flower.

demonstrates that diatoms in ancient Egyptian mud bricks (and potentially other mud bricks) can provide important clues to the selection of raw materials, the manufacturing processes and the past environments associated with the raw materials. Flower observes that it is unclear whether ancient Egyptian builders exercised any preference for diatom-rich sources for their bricks. However, the well known insulating properties of diatomite would be advantageous in the construction of buildings. He emphasizes that diatom analysis of ancient building materials is also potentially useful not only for mud brick typification but also for making space-time correlations between mud brick facies, within and between monuments.

Salt Extraction Sites

Salterns are sites where salt is produced by the evaporation of salt water in shallow pools, and sometimes with the aid of hearths. In Europe, from the Iron Age through to Medieval times, such sites were found near salt water in places such as estuaries and salt marshes and also near inland sources of mineral salt. As a result of their sensitivity to salinity, diatoms have been used to assist in the environmental analysis of such sites and in the identification of the functions of specific areas of such sites (e.g., Juggins, 1992).

Moats and Medieval Fish Ponds

Man-made water bodies such as moats and fish ponds dating from Roman to Medieval times are associated with sites such as settlements, villas or farmhouses, fortifications, and monastic sites and can provide local and, sometimes, well-dated diatom-based environmental evidence for the effects of human activities. An example of this is provided by diatom analysis carried out in association with the archaeological excavations at the Tower of London Moat (Keevill, 2004). At the Tower of London a number of sources were found to influence water quality over a 600-year period from 1240 to 1843. The input of estuarine water from the tidal River Thames was evaluated using ecological groups of diatom-salinity preferences. Saline water entered this moat through the Traitor's Gate and the input varied through time. Although there was little archaeological evidence for the dumping of refuse directly into the moat, organic pollution was derived from drainage from the Tower, and also from waste disposal into a water course, the City Ditch, which drained from the external City settlement into the Moat. Using a diatom-phosphorus transfer function, total phosphorus (TP) levels indicated nutrient enrichment (Figure 3) in excess of that found at pre-Medieval sites elsewhere in London that have been analyzed for diatoms. Diatom analysis has been used in a similar context to examine sediments from two profiles taken from a ditch draining the moat surrounding fortifications in Prague, initiated in AD 1230 (Beneš et al., 2002). Here the diatom analyses revealed that a relatively clean, fast-flowing stream tended to flush organic pollutants quickly away from the urban environment. Only toward the end of its function as a gutter, when the feature became clogged with sediment, did slowing of the current lead to stagnation. Even then the degree of nutrient enrichment

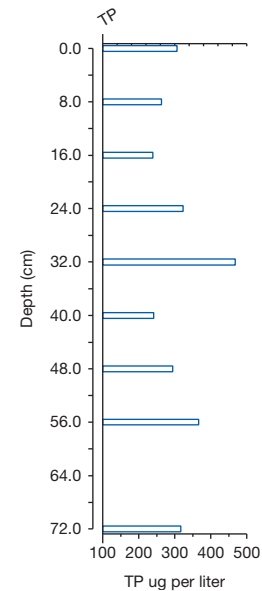


Figure 3 Section through Medieval sediments from the Tower of London moat, showing variation in the diatom-based TP reconstruction. Reproduced from Keevill G (2004) *The Tower of London Moat. Archaeological Excavations 1995–9*, pp. 315. Historic Royal Palaces Monograph No. 1. Oxford: Oxford Archaeology with Historic Royal Palaces. Copyright N. G. Cameron.

remained surprisingly low, with water quality comparable to 'recreational water supply' according to present-day norms. Diatom analysis of sediments from a ditch and moat-like feature from the Castle of Kastelholm, Åland, Finland was also used by Nunez and Paabo (1990) to ascertain whether these features were connected with the sea. The presence of coastal diatom species in the samples seemed to confirm that this was the case; however, as is common with archaeological sediments, the low concentration and fragmentary state of the diatoms from the ditch suggested that in this feature they represented secondary deposition from the ditch bank when the castle island was below sea level. Like many archaeological contexts, the interpretation of the diatoms at this site was therefore based on the state of diatom preservation in addition to the diatom composition. In much of the routine environmental archaeology evaluation work carried out in the United Kingdom, for example, the aims of a diatom assessment are to determine if diatoms are present and to assess the potential for possible percentage analysis of the diatom assemblages and its value. Before more detailed analyses are considered, sample characteristics examined include diatom valve concentrations, the quality of diatom preservation (the degree of breakage and dissolution), and the diversity of taxa, as well as the types of diatom assemblage and the environmental preferences of the diatoms present.

Crannogs

Crannogs are man-made islands constructed in lakes and in sea lochs that support buildings. These structures are found in northern and northwestern Europe and date from the Bronze

Age to late Medieval times. Diatom analysis has been used as part of a multiproxy investigation of a crannog in a lough in central Ireland (O'Brien et al., 2005). Here the crannog was constructed in the seventh century with the most intensive occupation from the beginning of the twelfth century. At this time, diatoms show evidence for eutrophication and aquatic habitat changes with increases in nonplanktonic taxa. The changes in the diatom communities are attributed to the building, rebuilding, and extension of the crannog. In addition, changes in water quality were linked to changes in agricultural practices and more intensive livestock grazing in the catchment of the lough associated with the crannog's inhabitants in the Medieval period. In addition to inputs from the lough's catchment, there would have been a direct input of nutrients to the lough from the crannog itself.

Waterfronts

The collaboration between diatomists and archaeologists concerned with the reconstruction of waterfront environments is exemplified by a series of detailed studies in London where developer-funded excavations over two decades have stimulated archaeological research projects involving diatoms (Juggins, 1992; Milne et al., 1983; Sidell, 2003; Sidell et al., 2000). Prior to these studies, Roman remains lying 4 m below

the present high tide level in the outer Thames estuary led to a common assumption that the estuary lay below London during the Roman period. However, within the city the pre-Roman foreshore is buried to the north of the present riverbank. There is a time-space sequence of waterfront structures covering the last 2000 years and preserved beneath the modern waterfront (Figure 4). The halophilous, estuarine, and allochthonous marine diatoms present in sediments accumulating by well-dated quay structures clearly demonstrated that the river was tidal in the City of London during the Roman period. Juggins (1992) developed models relating diatom assemblages to salinity and taking into account the taphonomy of the diatom assemblages. This allowed the past salinity of the Thames in the city to be better estimated. More recent studies have greatly expanded the number of archaeological sites with well-dated diatom records and added to the picture of a rapidly changing river (e.g., Sidell et al., 2000). Analyses of sediments, including diatom analysis, show that the upstream progression of the tidal head of the Thames through London began in the Late Bronze Age. However, this rapid transgression is followed by a period of regression and a second transgression. During the Roman period, this resulted in continuous rebuilding of the waterfront to maintain The Port of London (Sidell, 2003). In addition, signs of possible nutrient enrichment and the effects of catchment deforestation have been detected in sediments from other waterfront and wetland archaeological sites in

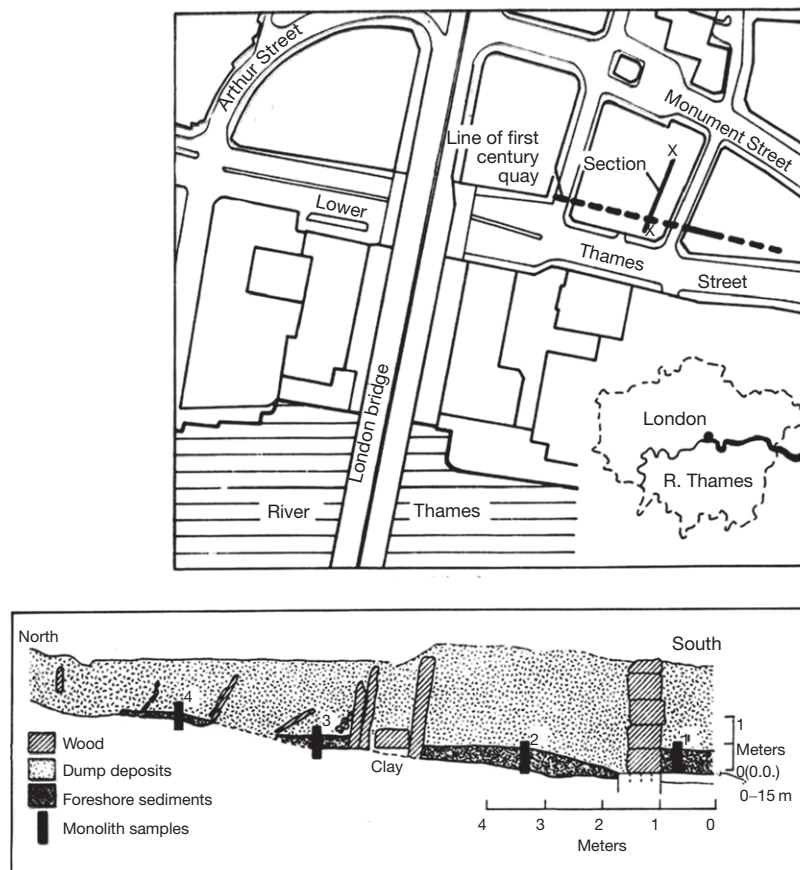


Figure 4 Location and stratigraphy of Roman waterfront sediments from the City of London. Reprinted from Battarbee RW (1988) The use of diatom analysis in archaeology: A review. *Journal of Archaeological Science* 15: 621–644, with permission from Elsevier.

London. The proximity of human activities to accumulating sediments can sometimes be an unwanted complication when interpreting diatom analyses of archaeological site sediments. However, the use of archaeological artifacts and dendrochronological dating of waterfront structures has the advantage of allowing relatively accurate chronological control for many of the reconstructions.

Ancient Boats

The increasing threat of sea-level rise and coastal erosion to ancient monuments has led to an increasing interest in coastal, estuarine, and intertidal archaeology and the recent discoveries of a number of ancient boats ranging in age from the Bronze Age to the Medieval period (Figure 5). Again, dendrochronology has in many cases allowed precise dating of the boat planks (strakes) and other structural wood such as oak frames. Archaeological deductions have led to interpretation of the uses of these vessels. There is convincing evidence that several relatively large early boat finds represent sea-going craft with implications for long distance transport and trade at an early



Figure 5 (a) Shows the aft plan of the Newport Medieval (fifteenth century) Ship found sealed in estuarine alluvium on the bank of the River Usk in South Wales, UK. (b) Shows the Magor Pill wreck (AD 1240) found on the foreshore of the Gwent Levels, South Wales, UK (see [Nayling, 1998](#)). Both photographs are by kind permission of Nigel Nayling, Department of Archaeology and Anthropology, University of Wales, Lampeter.

date in prehistoric Europe. For these finds diatom analysis has been used to reconstruct the sedimentary environments in which these ancient boats came to lie and, in some cases, to look at the provenance of materials used in boat construction ([Clark, 2004](#); [Nayling, 1998](#); [Nayling and Caseldine, 1997](#); [Nayling and McGrail, 2004](#)). Diatom analysis has also been applied to ancient harbors, exemplified by analysis carried out on samples from a core sequence from Caesarea's ancient harbor with the aim of reconstructing the general salinity conditions of the sedimentary environment ([Cameron, 1999](#)).

Other Archaeological Contexts

Many unusual archaeological contexts have been analyzed for diatoms in order to assist in their interpretation. For example, diatoms from Pleistocene postoccupation layers from a cave in the Western Kalahari Desert ([Robbins et al., 1996](#)) were found to indicate flowing water. Structures such as sediments in wells ([Neely et al., 1995](#)), pits, post holes, and around fish traps ([Godbold and Turner, 1993](#)) have been analyzed for diatoms to examine the functions and environments of such features. Similarly [Denham et al. \(2009\)](#) have used diatom analysis as part of a multiproxy geoarchaeological study to interpret the fills of Holocene archaeological features at a site in New Guinea. Diatom analysis has also been used in experimental archaeology, for example, to look at the drinking habits of livestock and excretion of diatom valves to aid in the soil micromorphological interpretation of habitation floor layers. The examination of diatoms has also been applied to the investigation of ancient burials ([Foged, 1985](#)). Like the application of the methods of archaeological excavation in forensic science (forensic archaeology), diatom analysis has also become more frequently employed as a forensic technique ([Peabody and Cameron 2010](#)).

See also: Diatom Introduction. **Diatom Records:** Antarctic Waters; Freshwater Laminated Sequences; Large Lakes; Diatoms.

References

- Alhonen P, Kokkonen J, Mäskäinen H, and Vuorinen A (1980) Applications of AAS and diatom analysis and stylistic studies of Finnish Subneolithic pottery. *Bulletin of the Geological Society of Finland* 52: 193–206.
- Battarbee RW (1988) The use of diatom analysis in archaeology: A review. *Journal of Archaeological Science* 15: 621–644.
- Beneš J, Kaštoký J, Kočárová R, et al. (2002) Archaeobotany of the Old Prague Town defence system, Czech Republic: Archaeology, macro-remains, pollen, and diatoms. *Vegetation History and Archaeobotany* 11: 107–119.
- Bradshaw EG, Nielsen A-B, and Anderson NJ (2006) Using diatoms to assess the impacts of prehistoric, pre-industrial and modern land use on Danish lakes. *Regional Environmental Change* 6: 17–24.
- Cameron NG (1999) Diatoms from the sediments of the Inner Harbour (Area 114) 343–345. In: Holum KG, Raban A, and Patrick J (eds.) *Caesarea Papers 2: Herod's Temple, The Provincial Governor's Praetorium and Granaries, The Later Harbor, A Gold Coin Hoard, and Other Studies*. The Journal of Roman Archaeology, Supplement Series Number 35. Rhode Island: Portsmouth.
- Clark P (ed.) (2004) *The Dover Bronze Age Boat*, 340 p. Swindon: English Heritage Publications.

- De Wolf H and Cleveringa P (1996) The impact of the beer industry on medieval water quality. In: Mayama S and Koizumi I (eds.) *Proceedings of the 14th Diatom Symposium*, pp. 511–522.
- Demiddele H and Erynck A (1993) Diatomeen als ecologische indicatoren in de Vlaamse archeologie: Romeins en middeleeuws Oudenburg (prov. West-Vlaanderen). In *Archeologie in Vlaanderen*, vol. III, pp. 217–231. Instituut voor Het Archeologisch Patrimonium.
- Denham T, Sniderman K, Saunders KM, Winsborough B, and Pierret A (2009) Contiguous multi-proxy analyses (X-radiography, diatom, pollen, and microcharcoal) of Holocene archaeological features at Kuk Swamp, Upper Wahgi Valley, Papua New Guinea. *Geoarchaeology* 24(6): 715–742.
- Douglas MSV, Smol JP, Savelle JM, and Blais JM (2004) Prehistoric Inuit whalers affected Arctic freshwater ecosystems. *Proceedings of the National Academy of Sciences* 101: 1613–1617.
- Flower RJ (2006) Diatoms in ancient building materials, application of diatom analysis to Egyptian mud bricks. *Nova Hedwigia* 130: 245–264.
- Foged N (1985) Diatoms in a tomb from the early Bronze Age. *Nova Hedwigia* 41: 471–482.
- Godbold S and Turner RC (1993) *Second Severn Crossing Archaeological Response: Phase 1: The Intertidal Zone in Wales*. Brentwood: Westbury Press.
- Juggins S (1992) *Diatoms in the Thames Estuary, England: Ecology, Palaeoecology, and Salinity Transfer Function*, pp. 216. Berlin/Stuttgart: Cramer.
- Juggins S and Cameron NG (2010) Diatoms and archaeology. In: Stoermer EF and Smol JP (eds.) *The Diatoms: Applications for the Environmental and Earth Sciences*, 2nd edn., pp. 514–522. Cambridge: Cambridge University Press.
- Keevill G (2004) *The Tower of London Moat. Archaeological Excavations 1995–9*, pp. 315. Historic Royal Palaces Monograph No.1. Oxford: Oxford Archaeology with Historic Royal Palaces.
- Mannion AM (1987) Fossil diatoms and their significance in archaeological research. *Oxford Journal of Archaeology* 6: 131–147.
- Miller U and Florin M-B (1989) Diatom analysis. Introduction to methods and applications. In: Hackens T and Miller U (eds.) *Geology and Palaeoecology for Archaeologists*, PACT 24: 133–157.
- Milne G, Battarbee RW, Straker V, and Yule B (1983) The River Thames in London in the mid 1st century AD. *Transactions of the London and Middlesex Archaeological Society* 34: 19–30.
- Nayling N (1998) *The Magor Pill Medieval Wreck*, 173p. Council for British Archaeology Research Report 115. York: Council for British Archaeology.
- Nayling N and Caseldine A (1997) *Excavations at Caldicot, Gwent: Bronze Age Palaeochannels in the Lower Nedern Valley*, 348p. Council for British Archaeology Research Report 108. York: Council for British Archaeology.
- Nayling N and McGrail S (2004) *The Barland's Farm Romano-Celtic Boat*, 325p. Council for British Archaeology Research Report 115. York: Council for British Archaeology.
- Neely JA, Caran SC, Winsborough BM, Sorensen FR, and Valastro S Jr. (1995) An early Holocene hand-dug water well in the Tehuacan valley of Puebla Mexico. *Current Research in the Pleistocene* 12: 38–40.
- Nunez M and Paabo K (1990) Diatom analysis. *Norwegian Archaeological Review* 23: 128–130.
- O'Brien C, Selby K, Ruiz Z, et al. (2005) A sediment-based multiproxy palaeoecological approach to the environmental archaeology of lake dwellings (crannogs), central Ireland. *The Holocene* 15: 707–719.
- Peabody AJ and Cameron NG (2010) Forensic science and diatoms. In: Stoermer EF and Smol JP (eds.) *The Diatoms: Applications for the Environmental and Earth Sciences*, 2nd edn., pp. 534–539. Cambridge: Cambridge University Press.
- Rippon S (2000) The Romano-British exploitation of coastal wetlands: Survey and excavation on the North Somerset Levels, 1993–7. *Britannia* XXXI: 69–204.
- Robbins LH, Murphy ML, Stevens NJ, et al. (1996) Palaeoenvironment and archaeology of Drotsky's Cave: Western Kalahari Desert, Botswana. *Journal of Archaeological Science* 23: 7–22.
- Sidell EJ (2003) *Holocene Sea Level Change and Archaeology in the Inner Thames Estuary. London, UK*. Unpublished PhD Thesis, University of Durham, 614 pp.
- Sidell J, Wilkinson K, Scaife R, and Cameron N (2000). The Holocene Evolution of the London Thames. Archaeological Excavations (1991–1998) for the London Underground Limited Jubilee Line Extension Project. Museum of London Archaeological Service Monograph 5, London, 144 pp.
- Stilborg O, Olsson S, and Håkansson H (1997) *Shards of Iron Age Communications*, 326 pp. Monographs on Ceramics, Keramiska Forskingslaboratoriet (ser.). Sweden: University of Lund.

Diatoms

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Introduction

Diatoms are single-celled microscopic algae, characterized by their silica cell walls, composed of two valves and separating girdle bands. For further details, see Jones, Diatom Introduction (this volume). Diatoms are found in most aquatic habitats and in damp terrestrial habitats and in many of these habitats the diatoms represent the most abundant and diverse algal class. Diatoms are important components of the phytoplankton, benthos, and attached algal communities of marine and fresh waters. Species composition is highly sensitive to water quality and many species are habitat specific. Further, diatoms grow rapidly and under favorable conditions produce many cells within a short period. They colonize submerged surfaces quickly and many taxa show seasonality of growth. The diatom valve is persistent and allows the majority of both living and fossil taxa to be identified to the species or subspecies level. For these reasons, they have been used in a number of areas in forensic science (Peabody and Cameron, 2010).

The involvement of diatom specialists (diatomists) in forensic investigations extends back to one of the earliest recorded geoforensic studies. In Germany, during the mid-nineteenth century, the geologist and diatom taxonomist Christian Gottfried Ehrenberg (Williams and Huxley, 1998) used variation in types of sand as forensic trace evidence (Ruffell, 2010). However, as Ruffell indicates, whether Ehrenberg used his knowledge of diatom taxonomy in any subsequent forensic investigation has not been established. Although used less often in forensic ecology than techniques such as pollen analysis (Mildenhall et al., 2006), diatom analysis has become an established bioforensic technique.

The most frequent and best-known application of diatom analysis in forensic science is to assist pathologists in the diagnosis of death by drowning.

Diagnosis of Drowning

Drowning is a common cause of death. In some cases, the pathological symptoms of drowning, or the circumstances of a person's death are unclear. If drowning is suspected, or to be eliminated as a contributory cause of death, diatom analysis may be employed to provide evidence to assist the pathologist in reaching a diagnosis of the cause of death.

The diatom test for drowning has become an established forensic technique to assist in the diagnosis of drowning (Peabody and Burgess, 1984). When water enters the lungs, small particles suspended in the water may also be carried into the bloodstream and potentially throughout the body. Such particles are likely to be deposited in the capillaries of the major organs. In most natural water bodies such as streams, rivers, ponds, lakes, and in the sea, this particulate material will

include diatoms. Therefore, where diatom valves are present in sufficiently high concentrations, their presence may assist in the diagnosis of drowning and, further, in identifying the aquatic environment in which drowning occurred (see below). It should be noted that it is not uncommon to find low concentrations of diatoms in the organs of nondrowned subjects. This 'background' flora of diatoms may be taken into the body during the person's lifetime. However, the diatom species most usually found in the nondrowned can be identified and separated from those species of value in the diagnosis of drowning. For this reason, it is important that an experienced diatomist carries out the analysis.

The earliest work on diatoms for the purpose of investigating suspected cases of drowning was carried out in Hungary during the 1940s (Incze, 1942; see Peabody, 1980). The application of diatom analysis to the investigation of drowning and specialized techniques for the preparation of highly organic and potentially biohazardous samples has been reviewed recently (Peabody and Cameron, 2010).

It is important to note that the presence of diatoms in nondrowned subjects has misled many investigators unfamiliar with diatoms, and has prompted them to suggest that diatom analysis is not useful in the diagnosis of drowning (Peabody and Cameron, 2010). For example, fragments of taxa including *Coscinodiscus* Ehrenberg, *Hantzschia amphioxys* (Ehrenberg) Grunow, and *Aulacoseira granulata* (Ehrenberg) Simonsen are ubiquitous. The successful investigator must be able to recognize these and other 'naturally occurring' diatoms for what they are, and to separate the background flora from those diatoms that may be present as a result of drowning (Peabody and Cameron, 2010).

Comparison of test samples (from organs such as liver, kidney, brain, or bone marrow) with control samples (from lungs, water and possibly nonplanktonic diatom communities) representing the diatom assemblage at the suspected site of drowning and elimination of possible background diatom flora is an essential part of using diatoms to assist in the diagnosis of drowning. Figures 1 and 2 show comparisons of diatom images of two species originating from water in which drowning occurred: (1) a control sample (water and lung samples), and (2) photographs of diatom valves of the same species taken from a slide prepared from the liver. In this case, there were other indications that drowning had occurred, but diatom analysis provided supporting evidence for the diagnosis of drowning as a contributory cause of death.

In addition to the diagnosis of drowning, diatoms may be employed specifically in locating the site of drowning in cases where water currents may have moved a body some distance from its original position, or where there are indications that the body has deliberately been moved from the site of death (Ludes et al., 1999). Differences in diatom composition will reflect variation in water quality in all aquatic environments

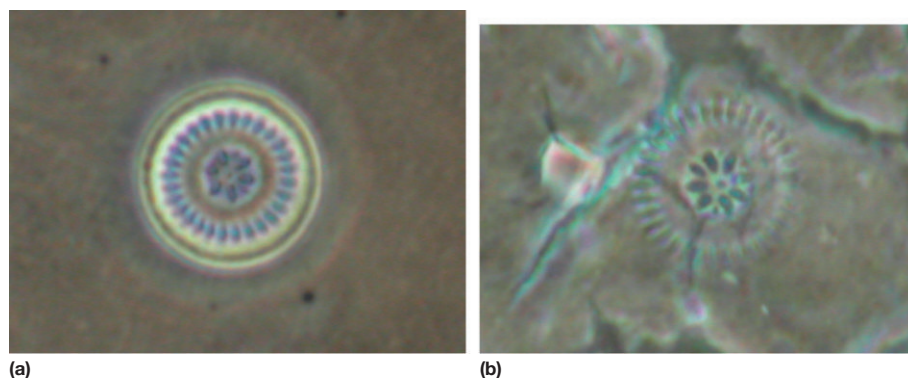


Figure 1 (a) *Cyclotella stelligera* from a test water sample and (b) *C. stelligera* from a slide prepared from liver.

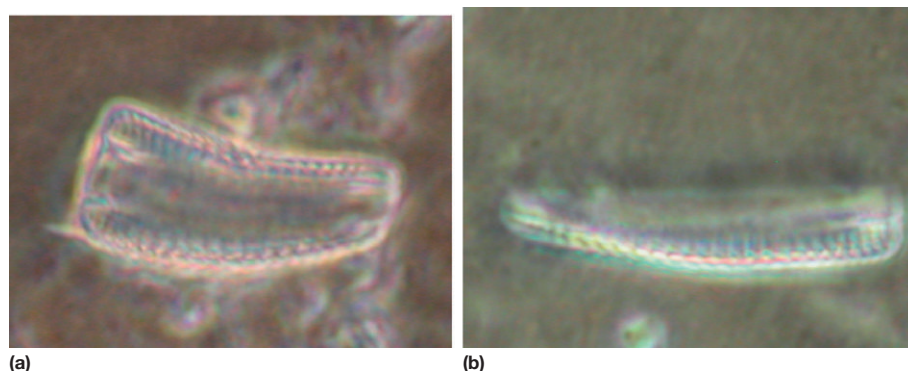


Figure 2 (a) *Rhoicosphaenia curvata* (girdle view of frustule) from a slide prepared from lung and (b) *R. curvata* (girdle view of convex dorsal valve) from a slide prepared from liver.

where there are strong environmental gradients or differences in chemistry such as pH, nutrient concentrations, salinity, or point sources of pollution such as heavy metals.

Provenance of Questioned Samples

Where materials have been submerged or there is contact with littoral or riparian sediment or vegetation, diatom analysis of sediments or other diatomaceous traces present on clothing, footwear, or vehicles can be used to identify the type of habitat involved.

Compared with nonsiliceous algae, diatoms are especially resistant to degradation and are not usually visible to the naked eye. They are, therefore, less likely to be removed from potential exhibits by an alleged offender.

In New England, USA, [Siver et al. \(1994\)](#) were able to link suspects with a lake that was the scene of a serious assault. They compared the diatom content of mud taken from suspects' and victim's shoes and freshwater pond sediment from the site of the crime. Diatom abundance, presence, or absence showed marked similarities between the samples.

Similarly diatom analysis was used to associate a murder suspect's clothing and footwear with the diatom flora of part of the River Avon, United Kingdom in which the body of a murder victim was found ([Cameron, 2004](#)). Ordination was

employed as a means of exploring and presenting the species data graphically to the jury in order to show the best analogs from among a group of test samples. Multivariate analysis of diatom data was also employed by the defence team in this case to challenge the prosecution diatom evidence ([Horton et al., 2006](#)). However, the overall concurrence of prosecution and defence diatom analytical results provided supporting evidence to assist in securing the conviction of the suspect.

Diatom analysis played a controversial role in the official investigation into the loss at sea of a British fishing vessel the FV Gaul. The Gaul and all 36 crew members were lost without trace in January 1974 in the Arctic Ocean north of Norway. In May 1974, a lifebuoy from the Gaul was recovered from the sea and diatom analysis was carried out on material taken from the lifebuoy. The diatom species present on the lifebuoy indicated an absence of deep-water marine diatoms and the presence of fresh water diatoms. At the time this led to controversy about the fate of the FV Gaul with the suggestion that the lifebuoy had been planted to disguise the fact that the loss of the ship, during the Cold War, was not accidental. However, the wreck of the Gaul was located in 1997 in the area where the ship and her crew were presumed to have been lost in heavy seas ([Steel, 2004](#)).

The standard laboratory preparation techniques developed for diatom ecological and paleoecological work ([Battarbee, 1986](#); [Battarbee et al., 2001](#)) can often be adapted for these

kinds of forensic study. However, [Uitdehaag et al. \(2010\)](#) made a systematic comparison of the efficacy of different methods used to extract diatoms from cotton clothing. Rinsing items with ethanol was found to be an effective and reproducible extraction method for both qualitative and quantitative analysis of diatoms deposited on cotton clothing.

Diatomite

The use of diatomite, both in manufacturing processes and as a component of finished products, leads to potential uses in forensic investigations. In addition, other raw materials may contain diatoms, for example, diatomaceous clays used in the manufacture of ceramics. Pottery, tile, or brick, for example, may retain diatoms where firing temperatures are low. However, the most important diatomaceous raw material is diatomite. This is a porous, low-density, sedimentary rock formed by the accumulation and compaction of diatoms which may comprise 90% or more of the dry weight of the rock. The physical properties of diatomite have led to its use in many commercial applications, the most common being for the filtration of liquids or as a filler, for example, in paints. Other uses of diatomite include insulation (fire bricks), fine abrasion (in some toothpaste and polishes) and as a pesticide. Freshwater and marine diatomites are mined in various parts of the world and range in age from the late Cretaceous to the Holocene so there may be a great deal of variation in the diatom assemblages present. The estimated total annual world production of diatomite is approximately 2 million metric tons ([Harwood, 2010](#)).

Diatomite derived from various materials has, therefore, been employed by forensic scientists to provide evidence of association. At present, commercial sources of diatomite are restricted to a few large sites. Consequently, the same raw material, potentially having a similar fossil diatom flora, can be used in the manufacture of a large number of products. However, forensic scientists have, for example, utilized diatoms from paint and polishes to provide evidence of association ([Peabody and Cameron, 2010](#)).

Diatomite was formerly used as a fireproof packing material between the outer casing and inner compartment of safes. If this type of safe is broken open, diatomite may contaminate clothing with a high concentration of diatom valves. The comparison of diatom assemblages in such cases will produce highly distinctive traces because diatomite safe ballast is likely to have been taken from former diatomite workings and it is unlikely that an individual would have this material on clothing by chance ([Peabody and Cameron, 2010](#)). In the past, nitroglycerine was absorbed in diatomite as a means of safe transport and this form of the explosive is still used occasionally ([Harwood, 2010](#)) and, therefore, this has the potential for use as forensic trace evidence.

[Ruffell and McKinley \(2008\)](#) describe a case where material suspected to be diatomite was presented to forensic geologists in Northern Ireland. Seized from a fuel storage depot, the possible diatomite might have been used to absorb red dye used in the United Kingdom to mark diesel fuel that has a lower tax duty and is allowed for use only in approved industrial applications. It was claimed by the suspected 'fuel launderers' that the material was clay – Fuller's Earth – used for the

legitimate purpose of soaking up oil spills. However, in this case the definitive separation of Fuller's Earth from diatomite was complicated by the presence of a proportion of siliceous fossils (including diatoms) in the Fuller's Earth, and by a fraction of clay in the diatomite. It is not indicated whether microscopic examination of the disputed substance was made to determine the diatom species composition.

[Ruffell and McKinley \(2008\)](#) also cite the use of diatom analysis to investigate a 1967 case of kidnapping in Beverly Hills, CA ([Cleveland, 1973](#)). After the release of the victim, a footprint on the carpet of an abandoned vehicle used to collect the ransom, was found to be of diatomite. Two common sources of diatomite in the western United States are fossil freshwater diatom assemblages from dry lakes (e.g., in Nevada) and Miocene fossil assemblages from the Monterey diatomite, a marine formation. A mixture of both these diatomite types was present in the diatomite footprint, as well as a smaller number of contemporary diatoms. A search identified an abandoned quarry and a testing laboratory where fossil marine diatomite had been quarried and freshwater diatomite was also imported for analysis. During storage at the site, mixing of the two diatomite types occurred, as well as the contemporary growth of diatoms in rainwater pools on the ground. Examination of one suspect's footwear, the imprint matching that in the abandoned car, and another vehicle with a similar mix of diatoms on the exterior, associated this suspect with the quarry where the kidnap victim had been held. Diatom analysis, therefore, provided some supporting evidence in the conviction of the kidnapper in this case.

Soils and Aerophilous Diatoms

Diatoms are not restricted to growing in fully aquatic habitats and the diatom communities of soils and other damp habitats are dominated by specialized desiccation-tolerant taxa (for a review see [Johansen, 2010](#)). Therefore, diatoms that are present as a component of forensic soil samples ([Pye, 2007](#); [Rawlins et al., 2006](#)) do not necessarily have an aquatic origin and detailed identifications of the species concerned are required to determine the species provenance. Although careful diatom analysis combined with statistical analysis of this type of soil diatom assemblage has been used successfully to provide evidence of association in forensic cases (van Dam, unpublished data), this area of diatom study has probably been underemployed elsewhere. Further work on the ecology of such aerophilous diatom communities is required. [Sugita et al. \(2004\)](#) examined microalgae in three soil samples taken from roads for the forensic discrimination of soils. Culture media were used as a means of growing and separating the algal species from the soil matrix. However, in this case cyanobacteria and green algae were observed in all the samples while diatoms occurred in only one of them. The three soil samples were successfully discriminated by the comparison of algae.

Period of Immersion

The use of diatoms in estimating the time elapsed since a body (postmortem submersion interval) or object was first

immersed in water has been the subject of some recent studies. These are analogous to methods used in diatom ecology where the growth of diatoms on inert substrata submerged in water is used as a means of environmental monitoring. This is similar in principle to the use of forensic entomology in estimating the time of death when a significant period has elapsed before the discovery of a body. Experimental work on forensic pathology (Casamatta and Verb, 2000; Zimmerman and Wallace, 2008) indicates that species diversity and sequence of diatom species and other algal taxa colonizing submerged animal carcasses has some potential for estimating the period of submersion of a body or other materials.

Conclusion

A number of characteristics of diatoms, including their wide-spread presence in water, sensitivity to water quality, high species diversity, habitat specificity, their use in manufacture, and good potential for preservation, suggest that this group of algae has further potential for use in forensic science. The invisibility of valves to the naked eye and the fact that diatoms are not well known to the layman are also useful attributes. Mildenhall et al. (2006) point out the importance of careful subsampling to avoid compromising evidential and control samples through destruction or contamination, particularly where a number of complementary forensic techniques are to be applied (e.g., diatom and pollen analysis). The use of diatom biogeographic data from large databases (e.g., Battarbee et al., 2000) prepared for environmental research has potential for use in future forensic investigations.

See also: [Use of Quaternary Proxies in Forensic Science: Analytical Techniques in Forensic Palynology](#); [Use of Geology in Forensic Science: Search to Locate Burials](#).

References

- Battarbee RW (1986) Diatom analysis. In: Berglund BE (ed.) *Handbook of Holocene Palaeoecology and Palaeohydrology*, pp. 527–570. Chichester: Wiley.
- Battarbee RW, Jones VJ, Flower RJ, et al. (2001) Diatoms. In: Smol JP, Birks HJB, and Last WM (eds.) *Tracking Environmental Change Using Lake Sediments, Vol. 3: Terrestrial, Algal, and Siliceous Indicators*, pp. 155–202. Dordrecht: Kluwer Academic.
- Battarbee RW, Juggins S, Gasse F, Anderson NJ, Bennion H, and Cameron NG (2000) European diatom database (EDDI). *An Information System for Palaeoenvironmental Reconstruction*, pp. 1–10. European Climate Science Conference. Vienna City Hall, Vienna, Austria, 19–23 October 1998.
- Cameron NG (2004) The use of diatom analysis in forensic geoscience. In: Pye K and Croft DJ (eds.) *Forensic Geoscience: Principles, Techniques and Applications*, pp. 277–280. London: Geological Society of London. Special Publications 232.
- Casamatta DA and Verb RG (2000) Algal colonization of submerged carcasses in a mid-order woodland stream. *Journal of Forensic Sciences* 45: 1280–1285.
- Cleveland GB (1973) http://web.mst.edu/rogersda/forensic_geology/cdmg_helps_find_kidnapper.pdf.
- Harwood DM (2010) Diatomite. In: Stoermer EF and Smol JP (eds.) *The Diatoms: Applications for the Environmental and Earth Sciences*, 2nd edn., pp. 570–574. Cambridge: Cambridge University Press.
- Horton BP, Boreham S, and Hillier C (2006) The development and application of a diatom-based quantitative reconstruction technique in forensic science. *Journal of Forensic Sciences* 51: 643–650.
- Incze G (1942) Fremdkörper in Blutkreislauf Ertrunkener. *Zentralblatt für allgemeine Pathologie und pathologische Anatomie* 79: 176.
- Johansen JR (2010) Diatoms of aerial habitats. In: Stoermer EF and Smol JP (eds.) *The Diatoms: Applications for the Environmental and Earth Sciences*, 2nd edn., pp. 465–472. Cambridge: Cambridge University Press.
- Ludes B, Coste M, North N, Doray S, Tracqui A, and Kinz P (1999) Diatom analysis in victim's tissues as an indicator of the site of drowning. *International Journal of Legal Medicine* 112: 163–166.
- Mildenhall DC, Wiltshire PEJ, and Bryant VM (2006) Forensic palynology: Why do it and how it works. *Forensic Science International* 163: 163–172.
- Peabody AJ (1980) Diatoms and drowning – A review. *Medicine, Science and the Law* 20: 254–261.
- Peabody AJ and Burgess RM (1984) Diatoms in the diagnosis of death by drowning. In: Mann DG (ed.) *Proceedings of the Seventh International Diatom Symposium*, pp. 537–541. Koenigstein: Otto Koeltz.
- Peabody AJ and Cameron NG (2010) Forensic science and diatoms. In: Stoermer EF and Smol JP (eds.) *The Diatoms: Applications for the Environmental and Earth Sciences*, 2nd edn., pp. 534–539. Cambridge: Cambridge University Press.
- Pye K (2007) *Geological and Soil Evidence: Forensic Applications*, p. 335. London: CRC Press.
- Rawlins BG, Kemp SJ, Hodgkinson EH, et al. (2006) Potential and pitfalls in establishing the provenance of earth-related samples in forensic investigations. *Journal of Forensic Sciences* 51: 832–845.
- Ruffell A (2010) Forensic pedology, forensic geology, forensic geoscience, geoforensics and soil forensics. *Forensic Science International* 202: 9–12.
- Ruffell A and McKinley J (2008) *Geoforensics*, p. 332. Chichester: Wiley.
- Siver PA, Lord WD, and McCarthy DJ (1994) Forensic limnology: The use of freshwater algal community ecology to link suspects to an aquatic crime scene in southern New England. *Journal of Forensic Sciences* 39: 847–853.
- Steel D (2004) *Report of the Re-opened Formal Investigation into the Loss of the FV Gaul*, p. 371. London: Her Majesty's Stationery Office.
- Sugita R, Shinichi S, and Marumo Y (2004) Validity of micro algae for forensic soil identification. *Japanese Journal of Forensic Science and Technology* 9: 113–122.
- Uitdehaag S, Dragutinovic A, and Kuiper I (2010) Extraction of diatoms from (cotton) clothing for forensic comparisons. *Forensic Science International* 200: 112–116.
- Williams DM and Huxley R (eds.) (1998) *Christian Gottfried Ehrenberg (1795–1876): The Man and His Legacy*. London: Linnean Society. Special Publications No. 1.
- Zimmerman KA and Wallace JR (2008) The potential to determine a postmortem submersion interval based on algal/diatom diversity on decomposing mammalian carcasses in brackish ponds in Delaware. *Journal of Forensic Sciences* 53: 935–941.

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DIATOM RECORDS

Contents

Antarctic Waters

Freshwater Laminated Sequences

Large Lakes

Diatom Fossil Records from Marine Laminated Sediments

North Atlantic and Arctic

Pacific

Antarctic Waters

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Marine Antarctic Waters: The Southern Ocean Siliceous Ooze Belt

The southern high latitudes play a critical role in the global climate system. Advances in the understanding of the paleoceanographic and cryospheric evolution of the Southern Ocean and Antarctica are helping us comprehend global climate change on a number of timescales. Diatoms are arguably the best fossil group for unraveling Antarctic climate change because they are prolific in the Southern Ocean – the area between the coast of Antarctica ($\sim 70^\circ$ S) and the subtropical front (STF; $\sim 40^\circ$ S; [Figure 1](#)) – contributing up to 75% of Southern Ocean production, despite this being a high-nitrate low-chlorophyll (HNLC) region. Such high diatom production in the surface waters, combined with a high export flux to the seafloor, has characterized the Southern Ocean as the world's main sink for biogenic opal (biosilica) since at least the early Pliocene (~ 4.5 million years ago, my; [Cortese et al., 2004](#)), and therefore a key regional component in the silica cycle. In particular, the area beneath the Antarctic Circumpolar Current (ACC), a major link in the thermohaline circulation, comprises a continuous belt of opal completely encircling Antarctica between 45° and 60° S ([Figure 2](#)). Diatomaceous sedimentation has been occurring within this belt more or less continuously since at least late Miocene times. Today, the nutrient-rich Southern Ocean supports a unique flora comprising at least six planktonic endemic species according to [Zielinski and Gersonde \(1997\)](#): *Eucampia antarctica*, *Fragilariopsis curta*, *F. kerguelensis* ([Figure 3](#)), *F. separanda*, *Shionodiscus gracilis* (cited as *Thalassiosira*), and *Thalassiosira lentiginosa*. Here, the Polar Frontal Zone (PFZ; [Figure 1](#)) acts as a major

geographic barrier to the migration of planktonic diatoms, separating cooler water Antarctic species to the south from warmer water subantarctic species to the north.

The Quaternary fossil diatom record of the Southern Ocean is generally well known, comprising mostly, but not exclusively, extant (living) species in high abundance. For example, [Zielinski and Gersonde \(1997\)](#) report diatom concentrations of $50\text{--}200 \times 10^6$ valves g^{-1} within surface sediment of the Atlantic sector beneath the ACC. Nutrient availability, temperature, water column stability, influence of sea ice, and grazing pressure are the main factors controlling distribution (e.g., [Crosta et al., 2005](#)). A siliceous ooze belt where diatoms are well preserved is bound north and south by sediments containing less well-preserved diatoms ([Figure 2](#)); carbonate-rich diatomaceous sediments to the north; and silty, terrigenous-rich diatomaceous clay to the south including the shelf regions. These boundaries are defined by oceanographic parameters – the modern PFZ ([Figure 1](#)) to the north and the mean position of the winter sea-ice edge (WSI; [Figure 1](#)) to the south. A similar relationship is inferred for the past, for which diatom records indicate a dynamic, climate-controlled history. Oceanographic parameters are sensitive to climate change and have shifted position, notably north or south, during the past, thereby influencing the position of the siliceous ooze belt. An added complexity is that these past frontal shifts and changes to the position of the sea-ice edge were not uniform or synchronous in all sectors of the Antarctic region. Therefore, through time, climate change has also indirectly altered the shape, or pattern, of the siliceous ooze belt to a certain degree. South of the siliceous ooze belt – across the Antarctic continental shelf – biogenic silica production and deposition also occurred during

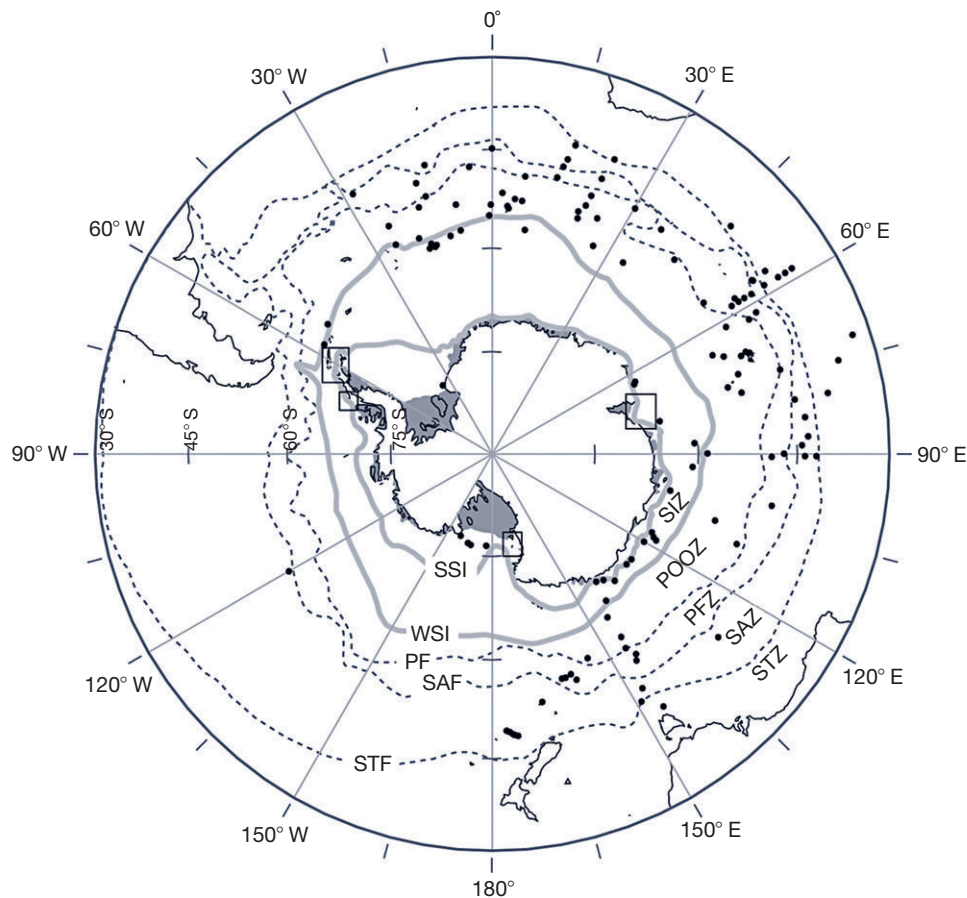


Figure 1 Oceanographic fronts and zones in the modern Southern Ocean referred to in this article. See Armand et al. (2005) for references to the position of the fronts and zones and of the summer and winter sea-ice edges. STF, subtropical front; SAF, subantarctic front; PF, Antarctic polar front; WSI, maximum average winter sea-ice edge; SSI, maximum average summer sea-ice edge; SZI, sea-ice zone; POOZ, permanently open ocean zone; PFZ, Polar Frontal Zone; SAZ, subantarctic zone; STZ, subtropical zone. Black dots – the location of the 228 surface samples represented in DD228 of Armand et al. (2005), Crosta et al. (2005), and Romero et al. (2005). Boxed regions represent regions of denser surface sample coverage, which are detailed in Armand et al. (2005). Reprinted from Armand LK, Crosta X, Romero OE, and Pichon J-J (2005) The biogeography of major diatom taxa in Southern Ocean surface sediments: 1. Sea ice related species. *Palaeogeography, Palaeoclimatology, Palaeoecology* 223: 93–126. Copyright (2005), with permission from Elsevier.

the Quaternary, but records are patchier, diluted by high amounts of terrigenous material, and generally less well preserved than in open ocean regions due to the influence of glacial and shelfal current processes. However, there are unique regions on the shelf where diatom productivity and flux were exceptionally high during the last deglaciation and Holocene resulting in distinct seasonal layering.

Applications of Quaternary Marine Antarctic Diatoms

Despite their preservation issues, marine diatoms are remarkably useful tools for climate reconstruction because they reflect their life environment and are normally the first microplankton group to respond to environmental change. Their use in paleoenvironmental reconstruction is perhaps most valuable for marine sediments recovered from below the calcium carbonate compensation depth (CCD) and in the high-latitude regions such as the Southern Ocean. In these environments, siliceous microfossils (mostly diatoms, and silicoflagellates

and radiolarians as well) prevail over calcareous microfossils (nannofossils and planktonic Foraminifera) because carbonate is not preserved below the CCD (preservational advantage), and reproduction of calcareous-walled microplankton is severely inhibited in polar waters (environmental advantage). Therefore, Antarctic marine diatoms are arguably the most reliable microfossil group for reconstructing past Southern Ocean and Antarctic shelf environments. Of critical interest are past fluctuations in sea-surface temperature (SST; e.g., Crosta et al., 2004; Zielinski et al., 1998), latitudinal oceanic frontal migration (e.g., Burckle, 1984; Gersonde et al., 2005; Kemp et al., 2010), sea-ice fluctuations (e.g., Armand and Leventer, 2009; Crosta et al., 1998; Gersonde et al., 2005), productivity and opal depositional history (e.g., Cortese et al., 2004), the role of diatoms in biogeochemical processes such as the cycling of major nutrients (i.e., silica, carbon, nitrogen, phosphorous), and the carbon pump, involving stratification (e.g., Kemp et al., 2006) and/or upwelling-induced carbon dioxide drawdown. To this end, there has been significant progress in recent days in assessing modern or

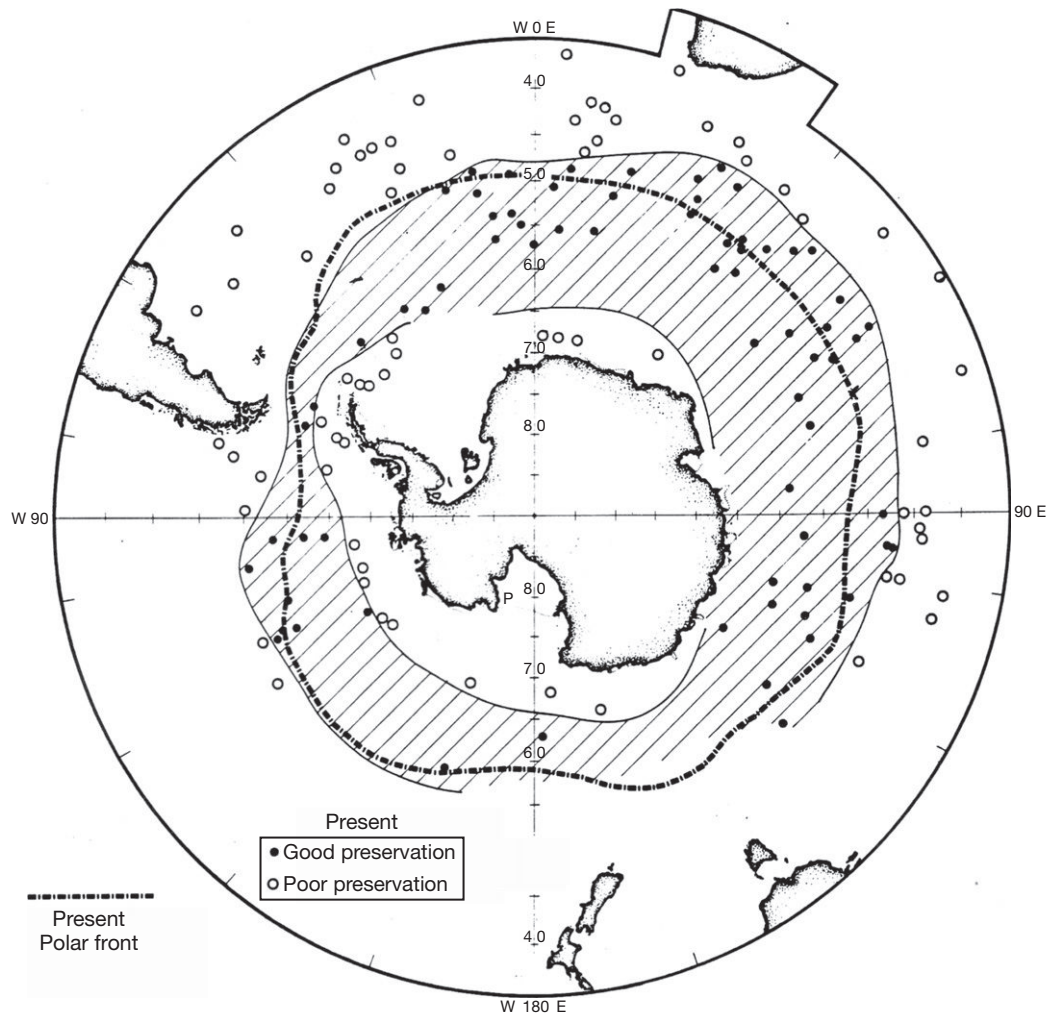


Figure 2 Siliceous ooze belt of well-preserved diatoms in surface sediments around Antarctica (black dots within the shaded area). Poorly preserved diatoms are indicated by white dots. Reprinted from Burckle (1984) Diatom distribution and paleoceanographic reconstruction in the Southern Ocean – Present and Last Glacial Maximum. *Marine Micropaleontology* 9: 241–261. Copyright (1984), with permission from Elsevier.

submodern species distribution patterns (biogeography) and other modern parameters, and in the refinement of Antarctic diatom-based statistical techniques that link the past with the present. These techniques have advanced our understanding of the main environmental drivers controlling species abundance. In this way, Antarctic diatoms are the key to elucidating the climatic and oceanographic history of the Southern Ocean. Further, the sort of information they provide is valuable for feeding back into, and improving, climate models that ‘hind-cast’ past environmental conditions. Importantly, because most Quaternary Antarctic species are extant, ecological-based applications can be applied to the recent past with confidence. Quaternary marine Antarctic diatoms are also widely used in biostratigraphy for both offshore regions and the Antarctic shelf (e.g., Barron, 2003; Cody et al., 2008). Across the shelf, they can also help elucidate sea-level changes, past glacial processes, and ice-sheet history (e.g., Naish et al., 2009). North of the Southern Ocean, an interesting but somewhat underdeveloped application is in their use for tracing Antarctic-source bottom water in the world’s ocean (e.g., Stickley et al., 2001).

Encompassing the world’s richest source of biogenic opal, the Southern Ocean and Antarctic shelf provide enormous opportunity to understand past Antarctic environments and climate change, promoting a wealth and diversity of studies involving Quaternary (and older) marine Antarctic diatoms (see Jordan and Stickley, 2010; Leventer et al., 2010 for comprehensive reviews). This has been made possible by several highly publicized, large-scale multinational drilling expeditions to the Southern Ocean and Antarctic shelf over the past 40 years or so (Figure 4): most recently IODP (and its predecessors ODP and DSDP), ANDRILL (and earlier Ross Sea drilling by CRP, CIROS, MSSTS, and DVDP), SHALDRIL, IMAGES (see bibliography for explanation of acronyms and project websites for references), as well as numerous smaller scale multinational and nationally funded expeditions.

It is not possible to discuss all these studies here; instead, the reports and areas of research that have widespread importance for the understanding of the Quaternary history of Antarctica and the Southern Ocean are summarized. Further reading on these and other topics from these publications and

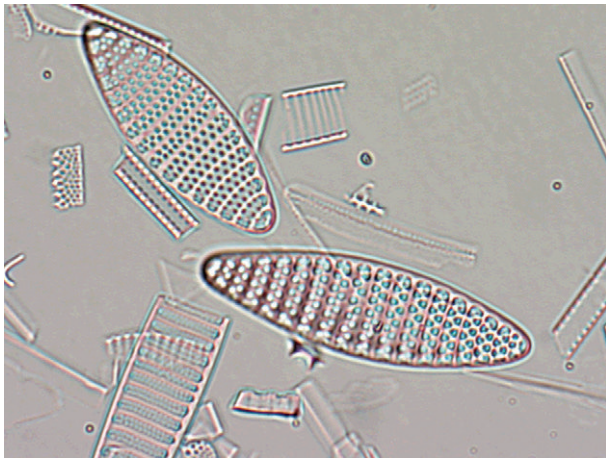


Figure 3 Light microscope image (plane polarized light) of endemic diatom *Fragilariopsis kerguelensis*, one of the main contributors to biogenic opal in Southern Ocean sediments. Image by E.J. Maddison (Cardiff University, UK) taken from late Quaternary laminated sediments of the East Antarctic Margin. Lower specimen is 50 μm in length.

references therein is encouraged. The focus is on (1) reconstructions of past sea-surface conditions including biogeography and primary productivity, sea-ice history, SST, and oceanic frontal migration; (2) Antarctic diatoms as tracers of bottom water; (3) ice-sheet history and glacial processes; and (4) biostratigraphy.

Reconstructions of Past Sea-Surface Conditions

Advances have been made in recent days in mapping the biogeography of modern (or submodern) marine diatoms in the Southern Ocean and Antarctic shelf seafloor sediments with respect to environmental controls. These ‘core-top’ sediments are taken to reflect ‘modern day’ sedimentation that may in reality represent sedimentation averaging the last several hundred years or so, depending on location – this fact is usually acknowledged in the better reports, and is generally acceptable for the resolution and timescales involved in most environmental reconstructions for the Quaternary. The assessment of hydrographic and cryospheric parameters controlling the temporal and spatial patterns in the productivity of marine diatom species living in the overlying surface waters is a valuable and necessary goal for reproducible paleoenvironmental reconstructions. These modern diatom data sets, or training sets, have been analyzed by a variety of statistical techniques and transfer functions have been created. Techniques such as the modern analog technique and the generalized additive model (see Crosta and Koç, 2007 for review and references) have enabled quantitative reconstructions of SSTs and seasonal sea-ice patterns for the Quaternary. These methods follow the assumption that the biogeographic variability of species preserved in the surface sediments reflect the surface-water hydrology and, therefore, may be used as a proxy for paleoenvironmental variability. The merits of assessing past SSTs and sea-ice patterns are clear: the warming and cooling of the ocean surface is a primary gauge for climate change, while sea ice (coverage, duration, and concentration) is especially important

for climate models because of its critical feedback effects and the energy exchanges associated with its growth and decay, as well as its impact on thermohaline circulation. As most diatom flux to seafloor sediments of the Southern Ocean occurs during austral summer (the ‘growing season’), relationships between diatom distributions and environmental parameters are best made with reference to summer conditions. However, reference to winter conditions is equally important as it provides a means to contrast seasonal extremes as well as factoring in the controllers for particular species distributions. Care must be taken, however, in generalizing for all areas of the circum-Antarctic because as Pike et al. (2008) showed in their statistical comparative study of surface water and sea-ice samples, and core-top sediments on the West Antarctic Peninsula shelf, seafloor sediments of this region more closely resemble spring diatom production rather than summer. In another example, Armand et al. (2008) describe the north-eastern Kerguelen Plateau as having an ‘island mass effect’ on diatom species distributions in core-top sediments that deviate from those previously reported for the permanently open ocean zone (POOZ; Figure 1; see below) that includes the Kerguelen Plateau.

Biogeography and primary productivity

Until the last decade or so, understanding of the relationship between the distribution of Antarctic diatoms in core-top sediments and surface-water hydrology was limited to a few early reports from the 1960s and 1970s. Although pioneering, these reports either accounted for a limited area of the Southern Ocean or covered a wider area in less detail. One of the more useful later reports is that by Burckle (1984). He mapped species distributions and preservation from core tops in all sectors of the Southern Ocean, covering its entire latitudinal range from nearshore to just north of the present day position of the STF (Figures 1 and 2). Although the dataset was relatively small (55 core tops; nine significant species) and the data points relatively scattered (Figure 2), Burckle (1984) produced one of the first distributional studies to cover the entire circum-Antarctic, and he attempted to statistically quantify the relationship between species distribution and environmental parameters. Later, Zielinski and Gersonde (1997) provided the first truly comprehensive modern species-specific distributional dataset for the Atlantic Sector of the Southern Ocean. They clearly showed that the biogeography and abundance of 35 significant diatom species in this region are closely related to summer SSTs and the extent of sea-ice cover, and therefore concluded that distributional patterns undoubtedly have paleoenvironmental significance.

A more extensive biogeographical dataset in terms of both coverage and detail is provided in a set of linked research papers by Armand et al. (2005), Crosta et al. (2005), and Romero et al. (2005). Their comprehensive modern database (Diatom Database 228 or DD228) highlights the importance of using fossil diatoms to reconstruct past environments and indicates that maximum abundances (indicating high primary productivity) of individual diatom species reflect narrow ranges of environmental conditions. Appropriately, Armand et al. (2005) broaden the use of sea-ice control by factoring in its concentration and duration, and not merely its presence or absence. This gives a much more sophisticated and

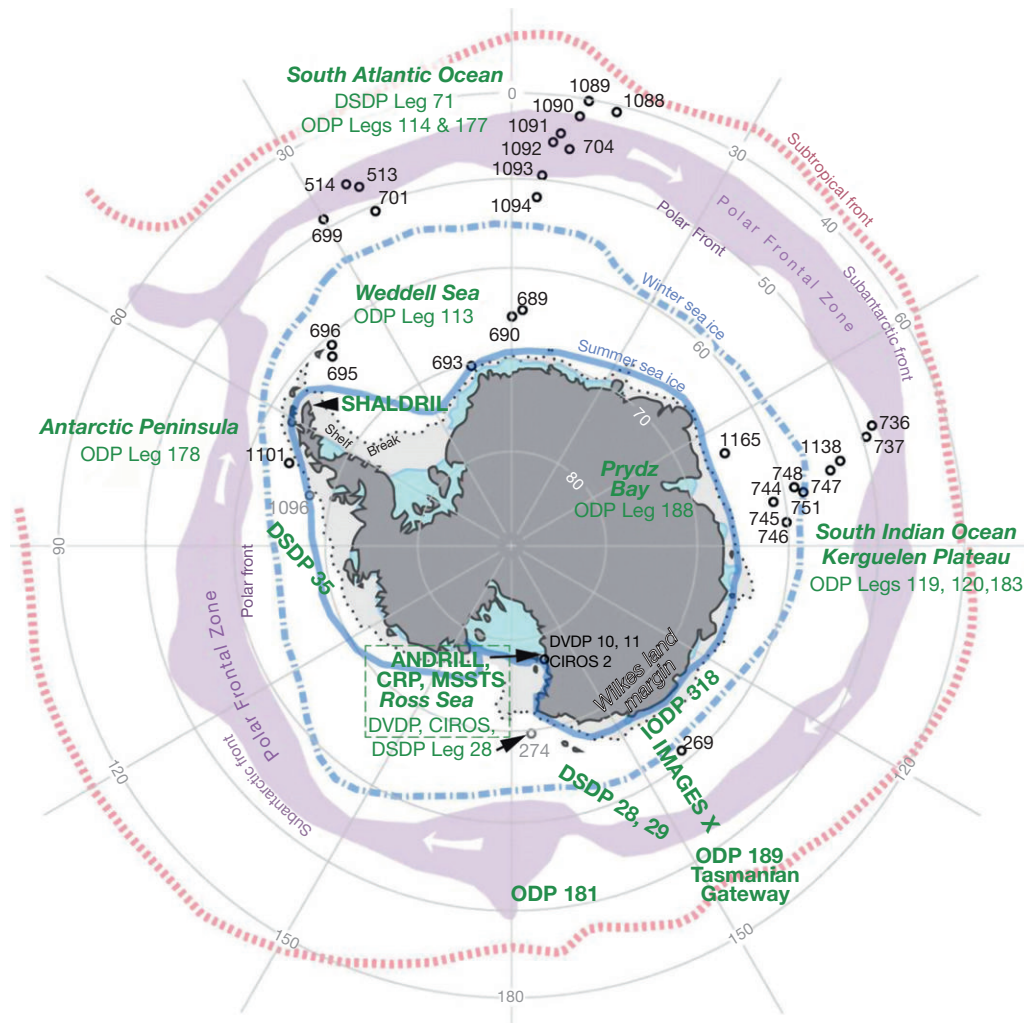


Figure 4 Location of key drill sites and drilling regions of the Southern Ocean and Antarctic shelf mentioned in the text, relative to physiography and modern oceanography. Adapted from [Figure 1](#) by [Cody et al. \(2008\)](#) with some additions. See 'International drilling projects cited in text' for acronyms. Reprinted from Cody RD, Levy RH, Harwood DM, and Sadler PM (2008) Thinking outside the zone: High-resolution quantitative diatom biochronology for the Antarctic Neogene. *Palaeogeography, Palaeoclimatology, Palaeoecology* 260: 92–121. Copyright (2008), with permission from Elsevier.

complete view of the relationships between diatom distribution and environmental parameters. These three interlinked papers also cover a much wider area than previous studies, and the biogeography of 32 major diatom species/taxa in 228 core-top sediment samples from all sectors of the Southern Ocean from the coast to just north of the STF was mapped ([Figure 1](#)). For example, the distribution pattern of the abundance of the endemic diatom *F. kerguelensis* ([Figure 3](#)), one of the main contributors to biogenic opal in Southern Ocean sediments, indicates its highest abundance within a SST range of between +1 and +8 °C ([Figure 5](#)). In DD228, each diatom species is defined as having the highest affinity with one of three zonally distinct regions: (1) the seasonal sea-ice zone (SIZ; [Figure 1](#)) between the maximum average winter and summer sea-ice edges, where sea ice occurs annually; (2) the POOZ between the maximum average WSI edge and the subantarctic front (SAF; [Figure 1](#)); and (3) the tropical/subtropical zone (STZ;

[Figure 1](#)) north of the SAF. Although all 32 diatom taxa occur throughout the study region, each to a greater or lesser degree, the correlation of their geographic patterns with sea-surface parameters illustrates the principal environmental control on the occurrence and abundance of each species (see [Table 1](#), sea-ice diatoms). For the SIZ and POOZ, these controls are summer (February) SSTs and sea-ice coverage in terms of (i) its concentration (% of water covered) during seasonal extremes (summer and winter) and (ii) its annual duration (months per year). In the STZ, the main controller is warmer SSTs and minimal influence by sea ice.

Another widely applied indicator of primary productivity is the abundance and distribution of *Hyalochaete Chaetoceros* resting spores (CRS). One of the first applications in the Antarctic realm was in the study of laminated sediments from the Antarctic Peninsula. High abundances of CRS are normally taken to reflect an influx of nutrients causing a bloom of vegetative *Hyalochaete Chaetoceros* that leads to resting spore

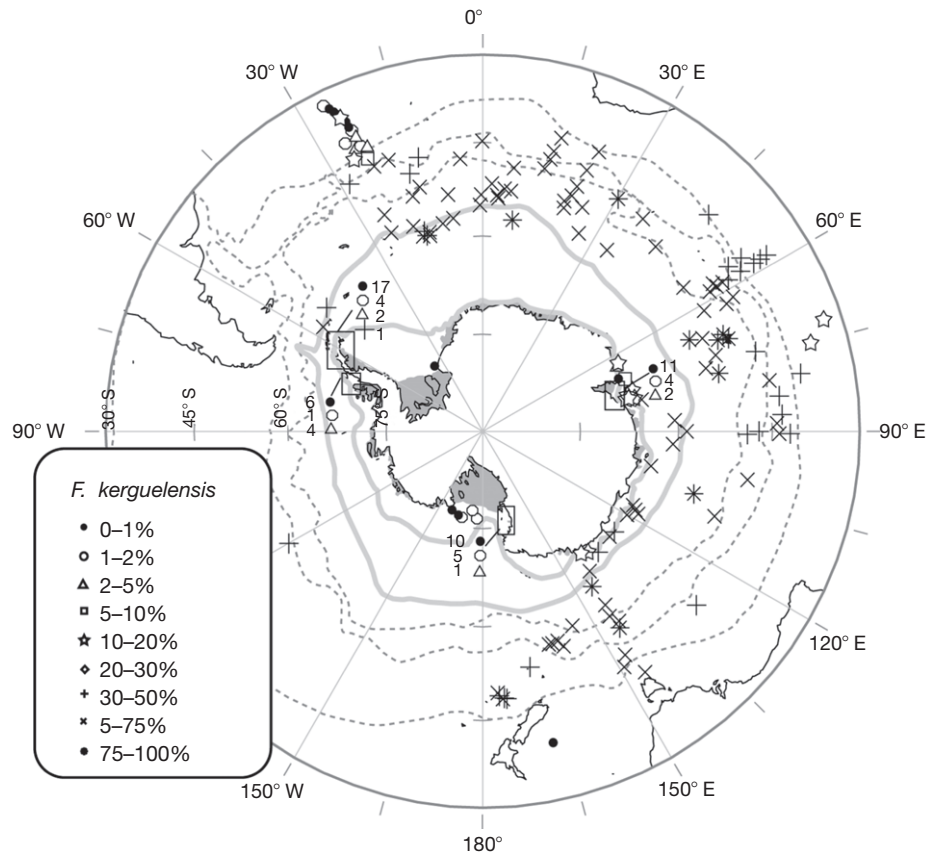


Figure 5 Surface sediment distribution of *Fragilariopsis kerguelensis* relative abundances in DD228. See [Figure 1](#) for oceanographic fronts and zones. Reprinted from Crosta X, Romero OE, Armand LK, and Pichon J-J (2005) The biogeography of major diatom taxa in Southern Ocean surface sediments: 2. Open ocean related species. *Palaeogeography, Palaeoclimatology, Palaeoecology* 223: 66–92. Copyright (2005), with permission from Elsevier.

Table 1 Temperature and sea-ice parameter ranges and maximum relative abundances of the 14 sea-ice diatom species and groups in DD228

Sea-ice diatoms (data from Table 1, Armand et al., 2005)	Maximum relative abundance (MRA, %)	February SST at MRA (°C)	February SST range (°C)	Maximum sea-ice duration at MRA (months per year)	Maximum summer (February) sea-ice concentration at MRA (%)	Maximum winter (September) sea-ice concentration at MRA (%)
<i>Actinocyclus actinochilus</i>	2.9	0–1	–1.3–9	9	2	82
<i>Chaetoceros</i> resting spores	91.8	–0.5–1.5	–1.3–3.5	7	43	62
<i>Fragilariopsis curta</i>	64.6	0.5–1	–1.3–2.5	10	20	83
<i>Fragilariopsis cylindrus</i>	2.9	0.5–1	–1.3–3	8.5	0	87
<i>Fragilariopsis obliquecostata</i>	10.2	–1–0	–1.3–2	8.5	0	90
<i>Fragilariopsis rhombica</i>	3.8	0.5–1	–1.3–11.5	9	1	93
<i>Fragilariopsis ritscheri</i>	2.9	0–3	–1.3–7.5	0 or 9	0 or 28	0 or 72
<i>Fragilariopsis separanda</i>	6.9	–0.5–1.5	–1.3–1.8	8.5	0	90
<i>Fragilariopsis sublinearis</i>	6.1	0–1	–1.3–2.5	7.5	24	61
<i>Porosira glacialis</i>	6.4	0–0.5	–1.3–2	8	0	90
<i>Porosira pseudodenticulata</i>	2.2	0–0.5	–1.3–2	9.5	28	81
<i>Stellarima microtrias</i>	3.2	–0.5–0.5	–1.3–3.5	8.5	0	88
<i>Thalassiosira antarctica</i> group	31.8	0–0.5	–1.3–3.5	8.5	3	92
<i>Thalassiosira tumida</i>	2	–0.5–0.5	–1.3–7	8.5	0	90

Source: Reprinted from Armand LK, Crosta X, Romero OE, and Pichon J-J (2005) The biogeography of major diatom taxa in Southern Ocean surface sediments: 1. Sea ice related species. *Palaeogeography, Palaeoclimatology, Palaeoecology* 223: 93–126. Copyright (2005), with permission from Elsevier.



Figure 6 Multiyear sea ice off the western Antarctic Peninsula in austral summer (December) 2003. Photo by C.E. Stickley taken aboard the *RSS James Clark Ross* of the British Antarctic Survey.

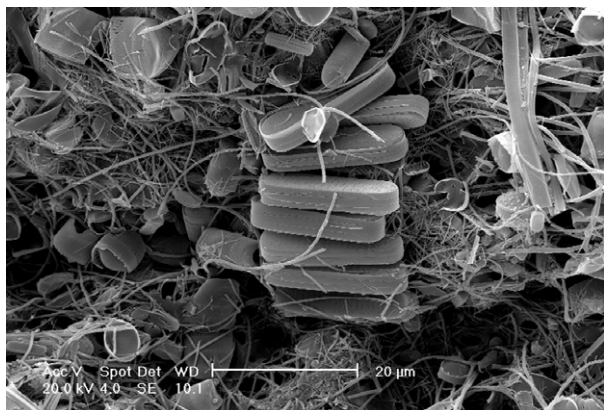


Figure 7 Secondary electron scanning electron microscope image of a colony (eight frustules) of the bipolar diatom *Fragilariopsis cylindrus*, used to trace past winter sea-ice patterns in the circum-Antarctic. Taken from late Quaternary laminated sediments of the East Antarctic Margin. Figure 4(c) of Stickley et al. (2005). Scale bar = 20 μm . Reprinted from Stickley CE, Pike J, Leventer A, et al. (2005) Deglacial ocean and climate seasonality in laminated diatom sediments, Mac.Robertson Shelf, Antarctica. *Palaeogeography, Palaeoclimatology, Palaeoecology* 227: 290–310. Copyright (2005), with permission from Elsevier.

formation and flux. CRS are often associated with upwelling systems and oceanic fronts in the world oceans, but in Antarctica, they are more likely to indicate ocean stability associated with sea ice (Armand and Leventer, 2009).

Sea-ice history, SSTs, and ocean frontal migration

Past sea-ice patterns and SSTs should be considered together as they are interrelated, albeit not necessarily linearly over long timescales. For example, Crosta et al. (2004) showed, in their quantitative diatom study of the Indian sector of the Southern Ocean, that during the transition from interglacials to glacials, sea-ice expansion lagged the SST decrease by approximately one thousand years (ka), while sea-ice retreat was almost synchronous with the SST increase at glacial–interglacial terminations.

Sea ice (Figure 6) provides an important and vast habitat for cryophilic Antarctic diatoms, supporting rich assemblages of these uniquely adapted extremophiles or the so-called sea-ice diatoms throughout the growing season. At the height of their growth, populations of sea-ice diatoms dwelling in the lower layers of sea ice can become so dense that they discolor the underside of the ice a yellow-brown or yellow-green color. Austral summer (February) sea ice (or multiyear ice) covers an area of approximately $3\text{--}4 \times 10^6 \text{ km}^2$, expanding in winter (August) to approximately $19 \times 10^6 \text{ km}^2$ (Figure 1) on average. Although sea ice inhibits normal surface water production and sedimentation, it stimulates huge blooms of diatoms at the edge of the sea ice and in the surrounding waters (the marginal SIZ) during the spring melt. The so-called sea-ice-related (or sea-ice-associated) taxa involved in these spring melt blooms, such as *F. curta* and *F. cylindrus* (Figure 7), are routinely used to trace winter sea-ice patterns for the Quaternary via a number of techniques, ranging from simple species presence/absence to more complex quantitative methods. Similarly, the presence of *F. obliquecostata* and a general fall in biogenic sedimentation may be used to trace summer sea-ice extent (e.g., Crosta and Koç, 2007; Gersonde et al., 2005). All techniques assume that the paleoecology of the extant species was the same as that observed for its living counterpart (e.g., Armand and Leventer, 2009). Leventer et al. (2010) list a number of other important sea-ice-related diatom taxa widely used to reconstruct sea-ice patterns for the Quaternary and Holocene, and go on to describe some of the pioneering and highly cited reconstructions, a detailed and vast subject in itself. The discussion elsewhere in this encyclopedia highlights studies where ultra-high resolution (i.e., seasonal) sea-ice reconstruction is possible from laminated circum-Antarctic sediments. Sea-ice diatoms also produce specific biomarkers that have been used to infer past sea ice. Further, one of the more interesting aspects of sea ice and sea-ice-related diatoms is that they are an important source of dimethylsulfide (DMS), a molecule implicated in the formation of clouds. Therefore, an understanding of sea-ice-related diatom production may have important implications for atmospheric climate, as well as cryospheric, feedbacks.

Zielinski et al. (1998) developed a rigorously tested SST transfer function of summer SST estimates for a core located in the Atlantic and western Indian sector of the Southern Ocean over the last ~ 200 ka. Bianchi and Gersonde (2004) later used this transfer function to reconstruct detailed trends in SST from the South Atlantic over the past 18 ka. Qualitative SST and sea-ice reconstructions are also possible through the documenting of changes in valve morphometrics, ornamentation, areolation, and/or degree of silicification of some species, for example, *F. kerguelensis* and *F. curta* (see Leventer et al., 2010, for review).

Other studies have concentrated on reconstructing conditions during particular time-slices, such as the Last Glacial Maximum (LGM). For example, Crosta et al. (1998) were the first to use the modern analog technique to quantitatively reconstruct circum-Antarctic sea-ice patterns from marine diatoms in terms of annual duration (months per year) at the LGM. In their comprehensive ‘state-of-the-art’ compilation of published and new datasets, Gersonde et al. (2005) provide the most up-to-date estimation of summer SSTs and both summer and winter sea-ice distribution patterns in the

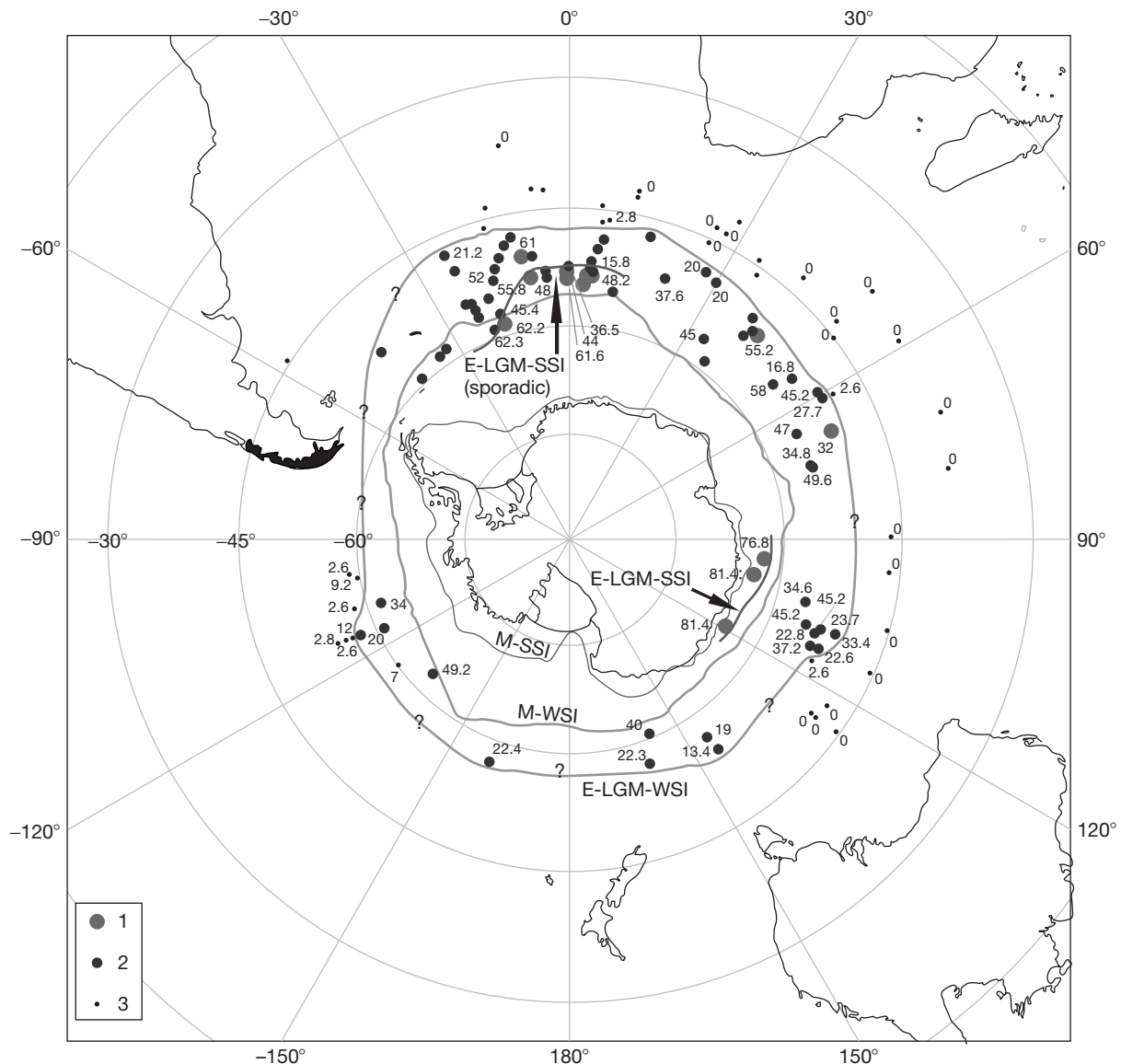


Figure 8 Sea-ice distribution at the EPILOG Last Glacial Maximum (E-LGM) time-slice compared to modern conditions. E-LGM-WSI, maximum extent of winter sea ice at the LGM (September, concentration >15%). M-WSI, maximum extent of modern winter sea ice (September, concentration >15%, see Gersonde et al. (2005) for reference). Values indicate estimated winter (September) sea-ice concentration in percent derived with the modern analog technique and the generalized additive model. Legend: (1) Large dots – concomitant occurrence of cold-water indicator *Fragilariopsis obliquecostata* (>1% of diatom assemblage) and summer sea ice (February, concentration >0%) interpreted to represent sporadic occurrence of E-LGM-SSI (= maximum extent of summer sea ice at the LGM); (2) medium dots – presence of winter sea ice (September, concentration >15%, diatom winter sea-ice indicators >1%); (3) small dots – no winter sea ice (September, concentration <15%, diatom winter sea-ice indicators <1%). Data compilation is presented in Table 3 of Gersonde et al. (2005). M-SSI, maximum extent of modern summer sea ice. Reprinted from Gersonde R, Crosta X, Abelmann A, and Armand L (2005) Sea-surface temperature and sea-ice distribution of the Southern Ocean at the EPILOG Last Glacial Maximum – A circum-Antarctic view based on siliceous microfossil records. *Quaternary Science Reviews* 24: 869–896. Copyright (2005), with permission from Elsevier.

Southern Ocean at the LGM, based on quantitative analysis of diatoms and radiolarians. Based on new data and reference datasets of several previously published works, Gersonde et al. (2005) applied three statistical methods to 122 marine cores located in all sectors of the Southern Ocean to quantitatively reconstruct the glacial Southern Ocean at the EPILOG (environment processes of the Ice Age: land, ocean, glaciers) LGM time-slice (23–19 calendar years before present, cal ka BP).

Their results indicate that at the EPILOG LGM, the maximum winter sea-ice extent (at a concentration of >15%) reached as far north as nearly 47° S in the Atlantic and Indian sectors, but only to 57° S in the Pacific (Figure 8). This equates to a total, asymmetric, EPILOG LGM circum-Antarctic winter sea-ice field of approximately 39 million km², a doubling of the area compared to the present, that agrees with earlier estimations (e.g., climate long-range investigation, mapping,

and prediction, CLIMAP). However, unlike previous studies, they suggest increased seasonality at the LGM based on the large differences between the extent of the winter and postulated summer sea-ice fields and those of today (Figure 8). In addition, their estimation of EPILOG LGM summer SSTs indicates a northward migration of the LGM polar front (PF; Figure 1) by 4° and 5°–10° latitude in the Atlantic and Indian sectors, respectively (Figure 8), with the strongest cooling (4–6 °C decrease) occurring in these sectors in the modern subantarctic zone (SAZ; Figure 1). These are larger estimates than previously reported (e.g., CLIMAP), and Gersonde et al. (2005) conclude a similar northward shift in the LGM SAF in the Atlantic and Indian sectors but only a relatively minimal northward shift of the LGM STF, leading to compression of the SAZ and steepened thermal gradients. However, their data from the Pacific sector implies reduced cooling here and therefore possible nonuniform cooling (and asymmetric frontal movement) of the glacial Southern Ocean at the LGM.

The authors advocate using integrated multiproxy approaches (i.e., diatom proxies used in conjunction with other biological as well as nonbiological proxies) for paleoceanographic reconstruction and suggest consultation of the reviews by Röthlisberger et al. (2010) and Armand and Leventer (2009) on the use of sea-ice proxies, and that by the MARGO Project Members (2009) on SST proxies at the LGM.

Antarctic Diatoms as Tracers of Bottom Water

Marine planktonic diatoms endemic to the Southern Ocean can be transported out of the Antarctic waters into the world's oceans via Antarctic Bottom Water (AABW). This is possible where surface-dwelling diatoms are entrained into downwelling AABW during its formation in the Weddell and Ross Sea gyres and on the shelves around the Antarctic continent. Entrained diatoms subsequently settling out along the flow path become effective tracers of this deep current. The process has clearly operated in recent times, as demonstrated by the presence of Antarctic endemic diatoms in core-top sediments along the AABW flow path north of the STF in the southwest and central Pacific (Stickley et al., 2001) and the southwest Atlantic (Jones and Johnson, 1984). In these locations, displaced endemic diatoms can comprise the majority of the sediment assemblage, while they are absent from core-top sediments located in overlying water masses (e.g., the North Atlantic Deep Water, NADW). Endemic diatoms displaced by AABW have also been noted in surface and core sediments as far as 60° N (L.H. Burckle and C.E. Stickley unpublished data).

Significant abundances of displaced endemic diatoms have been found in late Quaternary (last ~200 ka) core sediments within the AABW flow path in the southwest Pacific (Stickley et al., 2001) and southwest Atlantic oceans (e.g., Jones and Johnson, 1984). Along with the stable isotopic signal from biogenic carbonate, these assemblages help provide information on the production and flow history of AABW to these oceanic basins. This is of interest as AABW plays a key role in the climate system, being one of the main transporters of heat and salts around the world's oceans. A higher abundance of displaced diatoms indicates higher AABW flow. For example, Jones and Johnson (1984) found that for the southwest Atlantic, displaced diatoms are more abundant during

interglacials (higher AABW flow) than during glacial maxima (lower AABW flow). Unlike stable isotopes, displaced endemic Antarctic diatoms provide unequivocal evidence of bottom water provenance in the North Atlantic. Furthermore, in the southwest Pacific, down-core variations in the abundance of displaced open-water diatoms (e.g., *F. kerguelensis*; Figure 3; *T. lentiginosa*) versus sea-ice-related diatoms (e.g., *F. curta*, *F. cylindrus*; Figure 6) may indicate temporal switches in AABW source (open-water-derived versus shelf-derived, respectively) through the late Quaternary (Stickley et al., 2001).

Ice-Sheet History and Glacial Processes

The long history of Antarctic shelf drilling programs, mostly in the Ross Sea region (see bibliography), and also along the East Antarctic Margin, have recovered diatom-bearing deposits, which are helping to elucidate both long-term and shorter term cryospheric evolution and development through the current 'icehouse' world, including variations in both the East and West Antarctic ice sheets. Results have revealed Milankovitch to seasonal pacing in ice-sheet dynamics. Some of the more recent and groundbreaking multiproxy-based research studies on ice-sheet history employ species and total abundance changes in Antarctic diatoms over long and shorter time intervals.

Anticipated results from IODP Expedition 318 to the adjacent Wilkes Land Margin in 2010 (Expedition 318 Scientists, 2010) are expected to provide high quality seasonal to longer timescale changes in the East Antarctic Ice Sheet, whereas on-going results (including numerous diatom studies) from the high profile ANDRILL program are unlocking the evolution of the West Antarctic Ice Sheet on a range of timescales. During its first drilling season (2006–7) alone, ANDRILL drilled over 1280 m of sediment core (the AND-1B core) with overall 98% recovery, representing the longest and most complete geological record from the Antarctic continental margin to date, archiving obliquity-paced (40 ka cycles) Pliocene and Pleistocene West Antarctic ice-sheet oscillations (Naish et al., 2009). Upper Miocene to Pleistocene sediments were recovered that comprise alternating diatomites of ~1–100 m thickness and glacial diamictites, with episodic volcanics. The upper ~600 m of the AND-1B core are diatom-rich (Pliocene–Pleistocene). The diamictites represent glacial advance (glacials), while the diatomites, most of which are nearly pure biogenic silica and often laminated, record high diatom productivity in open-water marine environments during climatic phases that were generally warmer than present (interglacials) with minimal summer sea ice (Naish et al., 2009; Scherer et al., 2007). The diatoms in the AND-1B core are generally low in diversity and often fragmented, consistent with subglacial or sedimentary compaction (Scherer et al., 2007). A careful analysis of the abundance and style of fragmentation may reveal glacial processes (see Jordan and Stickley, 2010 for review and references). Analysis of diatom fragments from diamictites in other localities and experiments indicate that valve characteristics remain identifiable even after degradation by grounded ice and subglacial transportation, that is, sub-ice streams even if shearing is intense. The relative proportion of centric to pennate diatom valves may distinguish glacially sheared sediments from undisturbed hemipelagic sediments, and from sediments fragmented by normal stress compaction on the Antarctic

continental shelf. Unequal changes in absolute and relative diatom abundances occur with compaction and shearing, depending on the shape and size of the diatom; also, the ratio of unbroken centric to pennate diatoms may gauge past sub-ice stream shearing (proportionally more centrics indicate more shearing). Further, by analyzing the relative proportion and type of reworked diatoms to *in situ* diatoms, in combination with whole to fragmented valves, it is possible to classify and distinguish between late Quaternary tills deposited beneath ice streams and those emplaced by slow-moving ice, information that is important in ice-sheet reconstructions (see [Jordan and Stickley, 2010](#), for references).

Southern Ocean Diatom Biostratigraphy

Since the rise of the diatoms in the Mesozoic Era, climate change and tectonism have influenced their evolution, diversity, migration, and biogeography, as niches and environments are created and destroyed (see [Jordan and Stickley, 2010](#) for review and references). Planktonic marine diatoms have undergone several relatively rapid, stepwise, turnover (evolutionary) events during the Cenozoic and as such are excellent tools for biostratigraphy. In turn, excellent biostratigraphy (biochronology) is necessary to elucidate the timing of climatic and oceanographic events. Diatoms are perhaps the most useful and widely used fossil group in age determination for Neogene (last ~23 my) Southern Ocean sediments. Since the early 1970s, several diatom zonation schemes have been, and continue to be, developed through offshore drilling by DSDP and ODP, and more recently, by IODP, in early 2010 ([Expedition 318 Scientists, 2010](#)). The more recent and useful Southern Ocean zonation schemes are integrated with magnetostratigraphy, allowing diatom events to be tied to the geomagnetic polarity timescale (GPTS), giving geochronologic as well as chronostratigraphic information. Such calibrations are key to understanding the timing of climate events through a multiproxy approach, and can highlight diachroneity in biostratigraphic events across regions, leading in turn to a better understanding of evolution and biogeography. For example, due to latitudinal gradients in surface-water parameters, the distribution of late Neogene and Quaternary marker diatoms is not the same north and south of the PFZ ([Figure 1](#)), leading to separate late Pliocene and Pleistocene (Quaternary) zonations for sites north and south of the PFZ (see [Cody et al., 2008](#), for references).

High resolution diatom biochronologies constructed by the methods used by [Barron \(2003\)](#) and [Cody et al. \(2008\)](#) allow a reasonable estimation of the rates of evolutionary turnover of planktonic diatom species for the Neogene (including Quaternary) Southern Ocean, as well as an estimation of their diversity, in approaches that incorporate reduced preservational and recording bias. The best biostratigraphic events (datums) are those marking the range bases (or the first stratigraphic occurrence, FO) of abundant and widespread species. Range tops (or the last stratigraphic occurrence, LO) of species are also useful, but reworking can mask the true stratigraphic position of these bioevents, so more skill is required in the assessment of LO datums. Compared to the Southern Ocean offshore regions, diatom biostratigraphy for marginal marine Antarctic shelf sediments is more problematic and therefore relatively

underdeveloped – a reflection of poorer preservation, lack of oceanic marker species, reworking, and the effects of glacial processes such as ice-grounding events that truncate lithological units. Despite these issues, significant progress in developing Antarctic-shelf diatom zonation schemes has been made in recent years, largely through the unprecedented success of the first two ANDRILL shelf drilling seasons in the Ross Sea region during 2006–7 and 2007–8 (e.g., [Scherer et al., 2007](#); [Winter et al., 2010](#)). These works add to and build on earlier studies from a suite of previous Antarctic shelf drilling programs in the Ross Sea (CRP, CIROS, MSSTS, DVD, DSDP; see [Scherer et al., 2007](#) and [Winter et al., 2010](#), for references) and drilling in the NW Weddell Sea (SHALDRILL).

The comprehensive high-resolution quantitative diatom biochronology for the Antarctic Neogene by [Cody et al. \(2008\)](#) is currently the most robust biostratigraphic scheme for the Southern Ocean for any fossil group. Their scheme allows age calibration to within an average of ± 80 ka – an order of magnitude greater in temporal resolution than earlier approaches. Further, unlike zonation schemes that tend to be geographically limited, the outputs of the quantitative approach by [Cody et al. \(2008\)](#) are applicable over a much wider area of the Southern Ocean, from the shelf to north of the PF. [Cody et al. \(2008\)](#) statistically integrate diatom biostratigraphy, magnetostratigraphy, and tephrostratigraphy from 32 offshore and shelf drill sites around the Southern Ocean, and present the correlation of these parameters as a constrained optimization (CONOP) – similar in approach to that for multidimensional graphic correlation. Their approach establishes a precise, reproducible biochronology that takes into account the quality of the original records (removing unreliable data) such as recovery and preservation, and gives confidence intervals on age assignments. Two complementary models are presented, based on different assumptions about reworking, which give independent estimates of average local ranges and total regional ranges of 116 fossil diatom taxa in the southern high latitudes covering ~20–0.06 my (~60 ka; [Figure 9](#)). Sixty-three diatom datums resolve the Quaternary (taking the base of the Pleistocene to be 2.59 my). As the CONOP model incorporates biostratigraphies from nearshore and offshore sites, it heralds better correlation between shelf and offshore sedimentary sections. Indeed, the ‘on-going fluid model’ presented by [Cody et al. \(2008\)](#) is constantly being improved as new diatom biostratigraphic data become available, for example, those from ANDRILL (e.g., [Scherer et al., 2007](#); [Winter et al., 2010](#)) and those anticipated for IODP Expedition 318 to the Wilkes Land Margin, East Antarctica (e.g., [Expedition 318 Scientists, 2010](#)).

Nonmarine Antarctic Waters: Freshwater Diatoms

Less than 1% of the Antarctic continent is ice free and here diatoms can grow in lakes and streams as well as in terrestrial habitats. The majority of the lakes are situated on the coastal oases (e.g., McMurdo Dry Valleys, Vestfold Hills, Larsemann Hills, and Schirmacher Oasis), the ice-free areas of the Antarctic Peninsula (e.g., James Ross Island, King George Island, and Livingston Island) and the Antarctic Islands (e.g., Signy Island). Modern ecological and distributional data have been collected

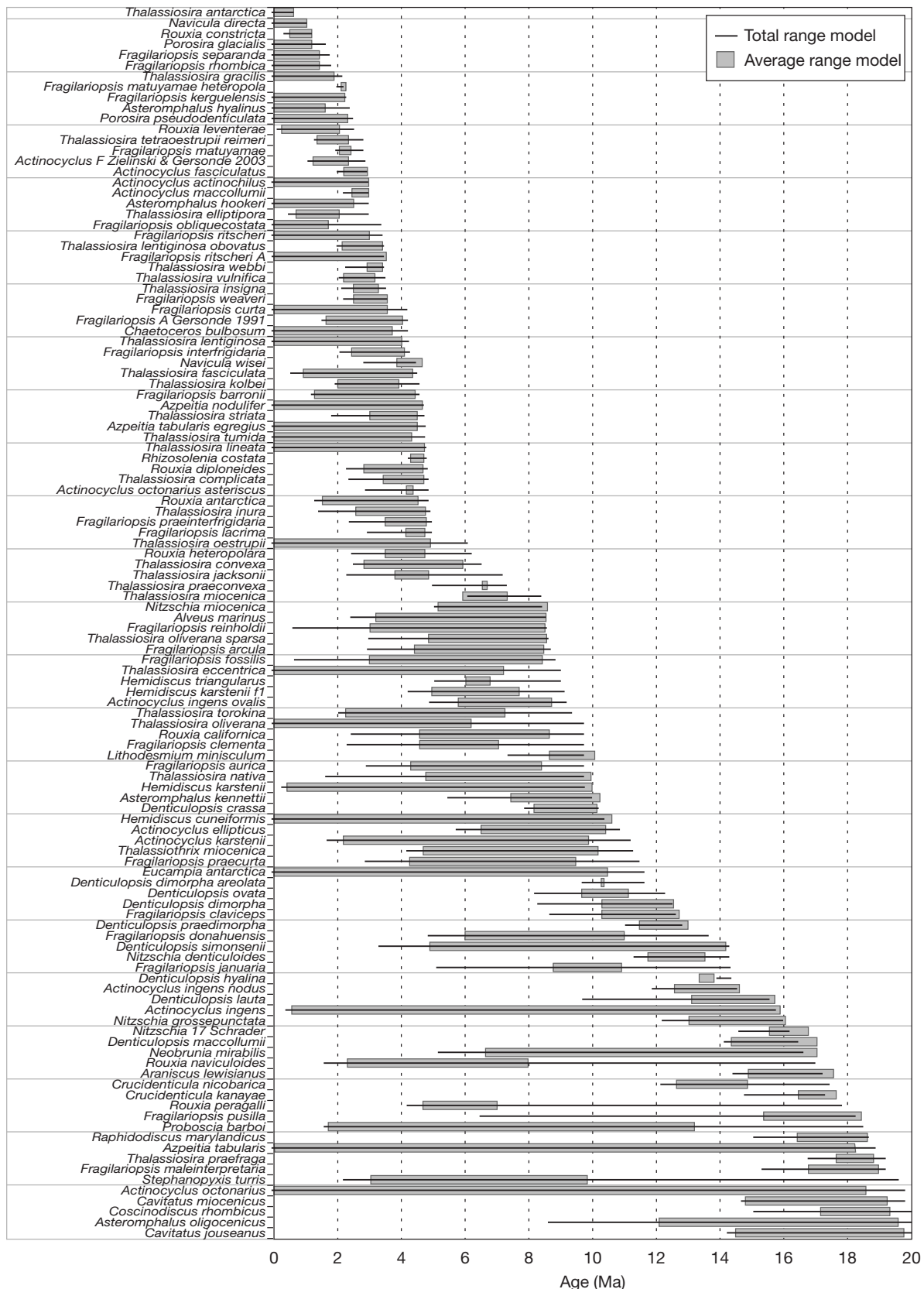


Figure 9 Composite range chart of 116 diatom taxa showing the total ranges (dark gray) and the average ranges (light gray). Each range end is plotted at the midpoint of its best-fit interval (see Cody et al., 2008). Data are presented in Table 3 of Cody et al. (2008). Reprinted from Cody RD, Levy RH, Harwood DM, and Sadler PM (2008) Thinking outside the zone: High-resolution quantitative diatom biochronology for the Antarctic Neogene. *Palaeogeography, Palaeoclimatology, Palaeoecology* 260: 92–121. Copyright (2008), with permission from Elsevier.

from several areas; for reviews see Jones (1996), van de Vijver and Beyens (1999), and Spaulding et al. (2010). The species diversity of nonmarine diatoms in Antarctica is low compared to Arctic regions, with few species in common, probably due to the physical isolation of the continent. The Antarctic Peninsula has a more diverse flora with a general trend of decreasing diversity southward (Jones, 1996). Many Antarctic diatoms are classed as aerophilic but are also found in aquatic habitats subject to desiccation or freezing; others are found in association with terrestrial soils and mosses. Some taxa are widespread whilst others have a regional distribution, and some are endemic (see Jones, 1996). Spaulding et al. (2010) review the ecology and distribution of diatoms in terrestrial and freshwater habitats from Antarctica and also provide more detailed information on Antarctic diatoms that have been used in environmental reconstruction.

Pre-Holocene records from Antarctica are rare but Hodgson et al. (2006) report a diatom record from marine isotope stage (MIS) 5e from East Antarctica. At this site, the diatoms from the last interglacial suggest a productive environment. The species are similar to those found in the subantarctic at present, and include species that are not found at present in east Antarctica. Hodgson et al. (2006) suggest a warm and wet interglacial environment consistent with the warmer temperatures shown by Antarctic ice cores.

Holocene stratigraphic records have accumulated in many lakes, and due to the low diversity of Antarctic freshwaters and the absence of many other proxies conventionally used in temperate regions (e.g., pollen), diatoms have been extensively used for environmental reconstruction. Hodgson et al. (2004) provide a recent review of Antarctic paleolimnology in general and Spaulding et al. (2010) review the uses of diatoms in particular. Diatoms have been used as indicators of climate and ecosystem change. They have also been used as indicators of salinity, which in some lakes can be a proxy for climate while in others it is related to changes in sea level. Diatoms have also been used as indicators of the changing extent of lake ice cover and lake depth.

Transfer functions have been developed for a range of environmental variables; for example, Jones and Juggins (1995) created a transfer function for nutrients (chlorophyll *a*) based on surface sediments obtained from Signy and Livingston Islands. This was applied to Holocene cores from Signy Island and, together with other proxy data, it showed a Holocene climate optimum between 3800 and 1300 ¹⁴C year BP (Jones et al., 2000) when the productivity of the lakes was higher. Recent changes in the diatom flora related to nutrient enrichment from birds and seals were also observed.

Transfer functions for salinity have been developed for diatom floras in the Vestfold and Larsemann Hills. The water chemistry of lakes in these areas responds quickly to changes in the moisture balance, with increasing salinity and decreasing lake water level during dry periods, and the opposite pattern in wet periods. Roberts et al. (2001) used a transfer function approach to obtain a high-resolution record of evaporation for the last 650 years from Ace Lake (Vestfold Hills), which was also highly correlated to the seasonal oxygen isotope signal preserved in an ice core from Law Dome, suggesting that the diatom-based reconstructions provide a robust paleoclimate signal.

The McMurdo Dry Valleys, located on the western coast of the Ross Sea, are amongst the coldest and driest places on Earth. Lakes here have some of the harshest aquatic environments in the world. The lakes mostly have a perennial ice cover (2.8–6 m thick) over liquid water. The volume of these lakes has fluctuated greatly over the last 20 000 years and, in recent times, lake levels have been rising. Lake Hoare is a perennially ice-covered, closed-basin lake where diatoms form part of the benthic microbial mats, with characteristic species occurring in shallow-water, mid-depth and deep-water communities (Spaulding et al., 1997). The lake has a unique sedimentary environment where, due to lack of water circulation, the sediments accumulate *in situ* and so record the history of a particular site rather than integrating events over the whole lake basin. Analysis of the sedimentary diatoms from Lake Hoare indicates that the rapid increase in lake level is a recent event (Spaulding et al., 1997).

Diatoms have also been found in Antarctic ice cores such as Dome C and Taylor Dome. These are a mixture of freshwater and marine taxa, which have been blown in by long-distance transport (Kellogg and Kellogg, 1996a). At Taylor Dome, Kellogg and Kellogg (1996b) found that low abundances of diatoms occur during glacial intervals (MIS 3–5), and higher abundances occur during interstadials (MIS 5b, 5d, and 6). This is due to the conditions in the source areas; in glacials, the West Antarctic Ice Sheet covers most of the Ross Sea continental shelf and thus the nearest open-water source for marine diatoms would be far offshore. The extent of ice-free areas in the dry valleys would also have been limited by lobes of grounded ice pushing into the valleys from the sea and by large ice-covered lakes greatly reducing possible sources for reworked marine diatoms (Kellogg and Kellogg, 1996b).

See also: Diatom Introduction. **Diatom Records:** Diatom Fossil Records from Marine Laminated Sediments. **Glacial Climates:** Thermohaline Circulation. **Glaciations:** Late Quaternary of Antarctica. **Paleoceanography, Biological Proxies:** Coccolithophores; Marine Diatoms; Planktic Foraminifera; Radiolarians and Silicoflagellates. **Paleoceanography, Physical and Chemical Proxies:** Oxygen-Isotope Stratigraphy of the Oceans. **Paleoclimate:** Timescales of Climate Change. **Quaternary Stratigraphy:** Chronostratigraphy; Continental Biostratigraphy; Overview.

References

- Armand LK, Crosta X, Quéguiner B, Mosseri J, and Garcia N (2008) Diatoms preserved in surface sediments of the northeastern Kerguelen Plateau. *Deep-Sea Research II* 55: 677–692. <http://dx.doi.org/10.1016/j.dsr2.2007.12.032>.
- Armand LK, Crosta X, Romero OE, and Pichon J-J (2005) The biogeography of major diatom taxa in Southern Ocean surface sediments: 1. Sea ice related species. *Palaeogeography, Palaeoclimatology, Palaeoecology* 223: 93–126.
- Armand LK and Leventer A (2009) Palaeo sea ice distribution and reconstruction derived from the geological records. In: Thomas DN and Dieckmann GS (eds.) *Sea Ice*, pp. 469–530. Oxford: Wiley-Blackwell.
- Barron JA (2003) Planktonic marine diatom record of the past 18 m.y.: Appearances and extinctions in the Pacific and Southern Oceans. *Diatom Research* 18: 203–224.

- Bianchi C and Gersonde R (2004) Climate evolution at the last deglaciation: The role of the Southern Ocean. *Earth and Planetary Science Letters* 228: 407–424.
- Burckle LH (1984) Diatom distribution and paleoceanographic reconstruction in the Southern Ocean – Present and Last Glacial Maximum. *Marine Micropaleontology* 9: 241–261.
- Cody RD, Levy RH, Harwood DM, and Sadler PM (2008) Thinking outside the zone: High-resolution quantitative diatom biochronology for the Antarctic Neogene. *Palaeogeography, Palaeoclimatology, Palaeoecology* 260: 92–121. <http://dx.doi.org/10.1016/j.palaeo.2007.08.020>.
- Cortese G, Gersonde R, Hillenbrand C, and Kuhn G (2004) Opal sedimentation shifts in the World Ocean over the last 15 Myr. *Earth and Planetary Science Letters* 224: 509–527.
- Crosta X and Koç N (2007) Diatoms: From micropaleontology to isotope geochemistry. In: Hillaire-Marcel C and de Vernal A (eds.) *Proxies in Late Cenozoic Paleooceanography. Developments in Marine Geology Series*, vol. 1, pp. 327–369. Amsterdam, The Netherlands: Elsevier.
- Crosta X, Pichon J-J, and Burckle LH (1998) Application of modern analog technique to marine Antarctic diatoms: Reconstruction of maximum sea-ice extent at the Last Glacial Maximum. *Palaeoceanography* 13: 284–297.
- Crosta X, Romero OE, Armand LK, and Pichon J-J (2005) The biogeography of major diatom taxa in Southern Ocean surface sediments: 2. Open ocean related species. *Palaeogeography, Palaeoclimatology, Palaeoecology* 223: 66–92.
- Crosta X, Sturm A, Armand L, and Pichon JJ (2004) Late Quaternary sea ice history in the Indian sector of the Southern Ocean as recorded by diatom assemblages. *Marine Micropaleontology* 50: 209–223.
- Expedition 318 Scientists (2010) Wilkes Land Glacial History: Cenozoic East Antarctic Ice Sheet evolution from Wilkes Land margin sediments. *IODP Preliminary Report* 318: doi:10.2204/iodp.pr.318.2010.
- Gersonde R, Crosta X, Abellmann A, and Armand L (2005) Sea-surface temperature and sea ice distribution of the Southern Ocean at the EPILOG Last Glacial Maximum – A circum-Antarctic view based on siliceous microfossil records. *Quaternary Science Reviews* 24: 869–896.
- Hodgson DA, Doran P, Roberts D, and McMinn A (2004) Paleolimnological studies from the Antarctic and subantarctic islands. In: Pienitz R, Douglas MSV, and Smol JP (eds.) *Long-Term Environmental Change in Arctic and Antarctic Lakes*, pp. 419–474. Dordrecht: Springer.
- Hodgson DA, Verleyen E, and Squier H (2006) Interglacial environments of coastal east Antarctica: Comparison of MIS1 (Holocene) and MIS5e (last interglacial) lake-sediment records. *Quaternary Science Reviews* 25: 179–197.
- Jones VJ (1996) The diversity, distribution and ecology of diatoms from Antarctic inland waters. *Biodiversity and Conservation* 5: 1433–1449.
- Jones VJ, Hodgson DA, and Chepstow-Lusty AJ (2000) Palaeolimnological evidence for marked Holocene environmental changes on Signy Island, Antarctica. *The Holocene* 10: 43–60.
- Jones GA and Johnson DA (1984) Displaced Antarctic diatoms in Vema Channel sediments: Late Pleistocene/Holocene fluctuations in AABW flow. *Marine Geology* 58: 165–186.
- Jones VJ and Juggins S (1995) The construction of a diatom-based chlorophyll a transfer function and its application at three lakes on Signy Island (maritime Antarctic) subject to differing degrees of nutrient enrichment. *Freshwater Biology* 34: 433–445.
- Jordan RW and Stickley CE (2010) Chapter 22: Diatoms as indicators of paleoceanographic events. In: Smol J and Stoermer E (eds.) *The Diatoms: Applications for the Environmental and Earth Sciences*, 2nd edn., p. 686. Cambridge: Cambridge University Press 9780521509961.
- Kellogg DE and Kellogg TB (1996a) Diatoms in South Pole ice: Implications for eolian contamination of Sirius Group deposits. *Geology* 24: 118.
- Kellogg DE and Kellogg TB (1996b) Glacial/interglacial variations in the flux of atmospherically transported diatoms in Taylor Dome ice core. *Antarctic Journal of the United States Review* 1996(31): 68–70.
- Kemp AES, Grigorov I, Pearce RB, and Naveira Garabato AC (2010) Migration of the Antarctic Polar Front through the mid-Pleistocene transition: Evidence and climatic implications. *Quaternary Science Reviews* 29: 1993–2009.
- Kemp AES, Pearce RB, Grigorov I, et al. (2006) Production of giant marine diatoms and their export at oceanic frontal zones: Implications for Si and C flux from stratified oceans. *Global Biogeochemical Cycles* 20: GB4S04. <http://dx.doi.org/10.1029/2006GB002698>.
- Leventer A, Crosta X, and Pike J (2010) Chapter 21: Holocene marine diatom records of environmental change. In: Smol J and Stoermer E (eds.) *The Diatoms: Applications for the Environmental and Earth Sciences*, 2nd edn., p. 686. Cambridge: Cambridge University Press 9780521509961.
- Margo Project Members (2009) Constraints on the magnitude and patterns of ocean cooling at the Last Glacial Maximum. *Nature Geoscience* 2: 127–132. <http://dx.doi.org/10.1038/NGEO411>.
- Naish T, Powell R, Levy R, et al. (2009) Obliquity-paced Pliocene West Antarctic ice sheet oscillations. *Nature* 458: 322–328. <http://dx.doi.org/10.1038/nature07867>.
- Pike J, Allen CS, Leventer A, Stickley CE, and Pudsey CJ (2008) Comparison of contemporary and fossil diatom assemblages from the western Antarctic Peninsula shelf. *Marine Micropaleontology* 67: 274–287.
- Roberts D, van Ommen TD, McMinn A, Morgan V, and Roberts JL (2001) Late Holocene East Antarctic climate trends from ice core and lake sediment proxies. *The Holocene* 11: 117–120.
- Romero OE, Armand LK, Crosta X, and Pichon J-J (2005) The biogeography of major diatom taxa in Southern Ocean surface sediments: 3. Tropical/subtropical species. *Palaeogeography, Palaeoclimatology, Palaeoecology* 223: 49–65.
- Röthlisberger R, Crosta X, Abram NJ, Armand L, and Wolff EW (2010) Potential and limitations of marine and ice core sea ice proxies: An example from the Indian Ocean sector. *Quaternary Science Reviews* 29: 296–302.
- Scherer R, Hannah M, Maffioli P, et al. (2007) Paleontologic characterisation and analysis of the AND-1B Core, ANDRILL McMurdo Ice Shelf Project, Antarctica. *Terra Antarctica* 14: 223–254.
- Spaulding SA, McKnight DM, Stoermer EF, and Doran PT (1997) Diatoms in sediments of perennially ice-covered Lake Hoare, and implications for interpreting lake history in the McMurdo Dry Valleys of Antarctica. *Journal of Paleolimnology* 17: 403–420.
- Spaulding SA, van der Vijver B, Hodgson DA, McKnight DM, Verleyen E, and Stanish L (2010) Diatoms as indicators of environmental change in Antarctic and subantarctic freshwaters. In: Smol JP and Stoermer E (eds.) *The Diatoms: Applications for the Environmental and Earth Sciences*, pp. 267–283. Cambridge: Cambridge University Press.
- Stickley CE, Carter L, McCave IN, and Weaver PPE (2001) Lower Circumpolar Deep Water flow through the SW Pacific Gateway for the last 190 ky: Evidence from Antarctic diatoms. In: Seidov D, Haupt BJ, and Maslin MA (eds.) *The Oceans and Rapid Climate Change, Past, Present and Future. Geophysical Monograph Series* 126, pp. 101–116. Washington: AGU.
- van der Vijver B and Beyens L (1999) Biogeography and ecology of freshwater diatoms in subantarctica: A review. *Journal of Biogeography* 26: 993–1000.
- Winter D, Sjunnskog C, Scherer R, Maffioli P, Riesselman C, and Harwood D (2010) Pliocene–Pleistocene diatom biostratigraphy of nearshore Antarctica from the AND-1B drillcore, McMurdo Sound. *Global and Planetary Change* <http://dx.doi.org/10.1016/j.gloplacha.2010.04.004>.
- Zielinski U and Gersonde R (1997) Diatom distribution in Southern Ocean surface sediments (Atlantic Sector): Implications for paleoenvironmental reconstructions. *Palaeogeography, Palaeoclimatology, Palaeoecology* 129: 213–250.
- Zielinski U, Gersonde R, Sieger R, and Fütterer D (1998) Quaternary surface water temperature estimations: Calibration of a diatom transfer function for the Southern Ocean. *Palaeoceanography* 13: 365–384.

Relevant Websites

- andril.org/ – ANDRILL: Antarctic Geological Drilling.
- www.arf.fsu.edu/projects/ – CRP: Cape Roberts Project, also see this website for earlier drilling projects in the Ross sea: CIROS: Cenozoic Investigations in the western Ross Sea; MSSTS: McMurdo Sound Sediment and Tectonic Study; DVDP: Dry Valley Drilling Project, and on the Antarctic Peninsula: SHALDRIL: Shallow Drilling on the Antarctic Continental Margin.
- www.images-pages.org/ – IMAGES: International Marine Past Global Changes Study.
- www.iodp.org/ – IODP: Integrated Ocean Drilling Program.
- www-odp.tamu.edu/ – ODP: Ocean Drilling Program; DSDP: Deep Sea Drilling Project.

Freshwater Laminated Sequences

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Introduction

Varves and Diatoms – Ideal Objects for Paleoecological Studies

Varved or annually laminated sediments contain high-precision records of sedimentation with yearly and even seasonal resolution (Lamoureux, 2001). Varves, especially in unconsolidated sediments, may be difficult to discern or may be easily disturbed by coring, so they are best sampled by freeze coring (Glew et al., 2001). Varve formation is a consequence of low-energy sedimentation, and it can therefore also indicate good preservation of various constituents of the sediment. A particular benefit of sediment varves is that they provide a precise chronology for the sediment sequence. The chronology may be floating or absolute (calendar), depending on the possibility to fix the series to the present day or any historical event, and more or less accurate, depending on the regularity and ease of identification of the annual cycles.

Diatoms are often abundant and are usually well preserved in lake sediments. Sedimentary diatom analysis is a versatile tool for tracking environmental and climatic history (e.g., Battarbee et al., 2001; Smol, 2004). Diatoms in varved sediments provide environmental records of very high quality, but the relative rarity of varved sediments sets limits on their routine use as environmental archives. In many cases, however, it is possible to improve the interpretation of nonvarved sediment records by comparisons with varved ones.

Friedrich Nipkow – Splendid Pioneer

Friedrich Nipkow (1886–1963) is worth remembering as an early pioneer of paleolimnology, and especially for his studies of diatom records in varved sediments. By profession a pharmacist, he studied algae as a hobby in his own laboratory in Zurich, Switzerland. He utilized a simple tube sampling device to collect profundal sediment cores from the deep lakes Zürichsee and Baldeggersee, which at that time were becoming strongly eutrophic from the input of untreated sewage effluents.

Nipkow (1927) presented excellent photographs of the varved sediment samples (Figure 1), as well as microphotographs of his diatom slides showing the seasonal assemblages and cell size variation observed within the varves. Varve formation had started in both the lakes a few decades earlier as a consequence of hypolimnetic anoxia. Nipkow convincingly verified the laminae as true varves by their seasonal diatom microstratigraphy and by correlation of species occurrences in the varves with previous plankton records. Furthermore, he could connect several turbidite layers observed in the sediments with historically known landslides on the lake shores.

Nipkow meticulously dissected each varve in his samples into 5–10 seasonal layers and analyzed the diatoms in each in great detail. Besides demonstrating the size reduction–auxosporulation cycle (a phase in the life cycle of diatoms in

which they form specialized reproductive cells) for several species (Figure 2), he also described year-to-year plankton community succession in both the lakes and discussed the potential effects of environmental conditions on the diatom development. There is also a good discussion about the advantages provided by the continuous and quantitative sediment records over the laborious long-term monitoring of plankton in water samples.

Diatoms as a Structural Component in Sediment Varves

The siliceous frustules of diatoms are a quantitatively important constituent in many lake sediments. Rapid and straightforward sedimentation of seasonal blooms of planktonic diatoms may result in distinct diatom-rich layers in the sediment. Often yellowish or greenish in color, these may be pure enough to be classified as diatomite. Such layers are often regular and visually significant structural elements in varves (Figure 3; Simola, 1992).

Often the annual character of rhythmically laminated sediment can be demonstrated by the microstratigraphy of diatoms, reflecting the regular yearly succession and the seasonal blooms of planktonic diatoms (Simola, 1992). In other cases, the absence of such seasonality may indicate a nonannual rhythmite.

Combining diatom microstratigraphy with other interannual or sublaminar sediment characteristics (facies units, their composition, and thickness) may allow detailed reconstruction of the typical seasonal patterns of catchment, and climate events and changes over time (e.g., Kienel et al., 2005; St Jacques et al., 2009). In very deep lakes, the highly variable settling rate of different types of sedimentary particles must be taken into account when looking at the composition of the varve subunits.

Methods of Analyzing Diatom Microstratigraphy

Numerous methods have been developed for studying diatom microstratigraphy, the choice depending on the objectives of the particular study. Basically, there are two approaches: (i) dissection of the sediment core into thin horizontal subsamples, in which the constitution of each separated layer can be studied individually, and (ii) preparing of vertical sections of the core, which enables the study of stratigraphic details in a single sample.

Nipkow (1927) was able to manually dissect varves in his fresh sediment cores into less than 1-mm increments and could estimate even the relative diatom frustule concentrations in these. Cocquyt and Israël (2004) described a microtome device allowing sediment subsampling optimally in 0.1-mm thick horizontal slices. These techniques are best suited for fairly compact sediments, for example, calcareous sediments.

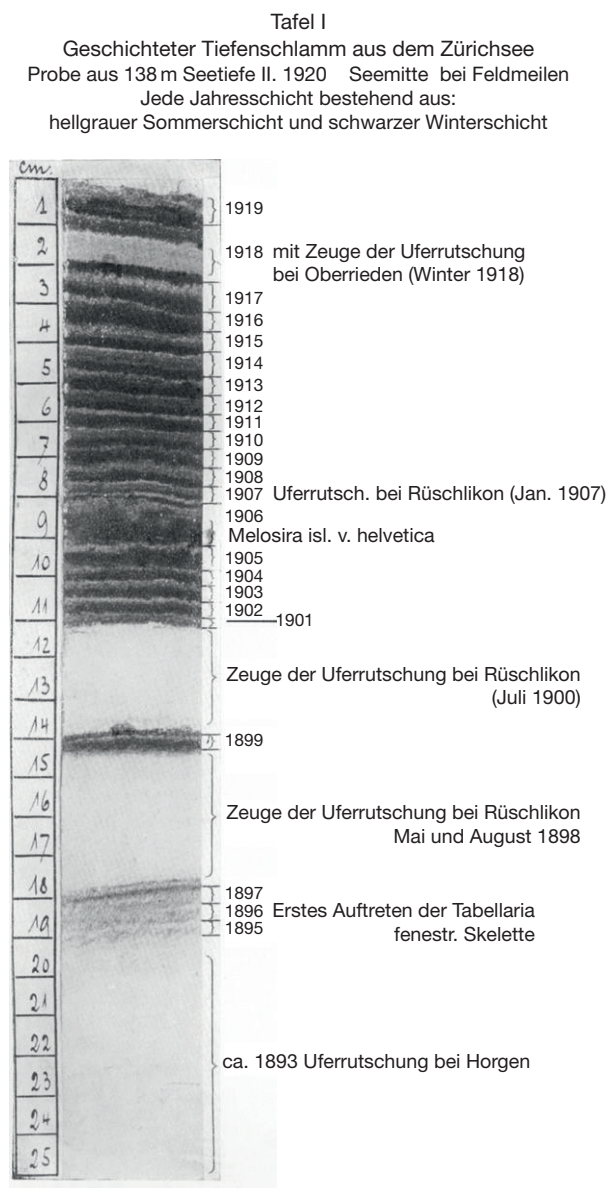


Figure 1 Friedrich Nipkow's photographic documentation and interpretation of the varved sediment of Lake Zurich (Nipkow, 1927).

It may not be possible to reach similar precision with soft organic sediments, which may be better suited to freezing and freeze drying. Schmidt et al. (1995) described a suction sampler for fine-resolution subsampling of freeze-dried varves and their structural components.

An advantage of fine-resolution horizontal slicing is that the material can be properly treated for the preparation of high-quality diatom slides. Furthermore, sampling of a somewhat extended horizontal area (at least some square millimeters) helps overcome the problem of uneven distribution of diatom frustules in each individual layer. A drawback of this approach is, of course, that it is very laborious, especially when dealing with longer varve sequences.

The tape peeling technique (see Simola, 1992) is a simple method for obtaining vertical sediment sections for

microscopy. From the cleaned surface of a frozen and slightly freeze-dried sediment core, a thin layer of sediment particles is adhered on strips of adhesive tape. The tape strips can be directly mounted onto microscope slides with a suitable liquid mounting medium, for example, Canada balsam. Unmounted, such tape strips are also suitable for direct examination by scanning electron microscopy (SEM; Dean et al., 1999). For clean diatom slides of tape peels, Davidson (1988) transferred the sediment particles onto a microscope slide spread with rubber glue, which he then incinerated and mounted with diatom mounting medium.

A standard technique for making permanent thin sections of soft sediment samples involves replacement of the interstitial water via acetone into a low-viscosity epoxy that is polymerized into a solid and transparent matrix (Kemp et al., 2001), after which the samples are ground into thin slides, analogous to petrographic thin sections. As pointed out by Cocquyt and Israël (2004), identification and counting of long, spindle-shaped diatoms may be especially difficult in such thin sections.

Diatom Assemblages Provide High-Resolution Records of Environmental Changes

Human Impact and its Relaxation

Ever since Nipkow (1927), a large number of paleolimnological varve studies have detailed the eutrophication and pollution history of lakes caused by human influence (for a general summary, see Hall and Smol, 2010). In many cases, eutrophication, leading to hypolimnetic anoxia, has actually triggered the onset of varve formation in the sediment.

The early studies were mainly based on empirical knowledge of the significance of individual taxa in diatom communities as indicators of water conditions such as trophic status, saprobity (the physiological and biochemical characteristics of an organism that permit it to live in water that is somewhat polluted), or pH of water bodies. The accumulated knowledge of diatoms as indicators of pH proved invaluable, when acid rain was recognized as a widespread and serious environmental problem in the 1980s (Hall and Smol, 2010). The modern approach to diatom-based environmental reconstruction typically involves a calibration data set of at least some tens of lakes, by which statistical transfer functions between species assemblages and the relevant environmental parameters are worked out. The transfer functions are then applied to make paleoecological inferences.

As a positive sign, increasing numbers of recent studies have been able to demonstrate aquatic ecosystem recovery from the effects of eutrophication or acidification. Alefs and Müller (1999) identified a four-phase eutrophication succession of sedimentary diatom assemblages within the varved sediments of two prealpine lakes in South Germany. The plankton diatom assemblages appear rather similar in both lakes, but varve dating reveals temporal differences in the onset, progress, and relaxation of the eutrophy between the lakes. This study topic will gain even more importance in the near future, when the enforcement of active water protection policies (e.g., The EU Water Framework Directive) will require knowledge of the natural or predisturbance status of impacted water bodies.

Größenwechsel der sedimentierten Planktondiatomeen im geschichteten Tiefenschlamm des Zürichsees aus dem Zeitraum 1896—1924

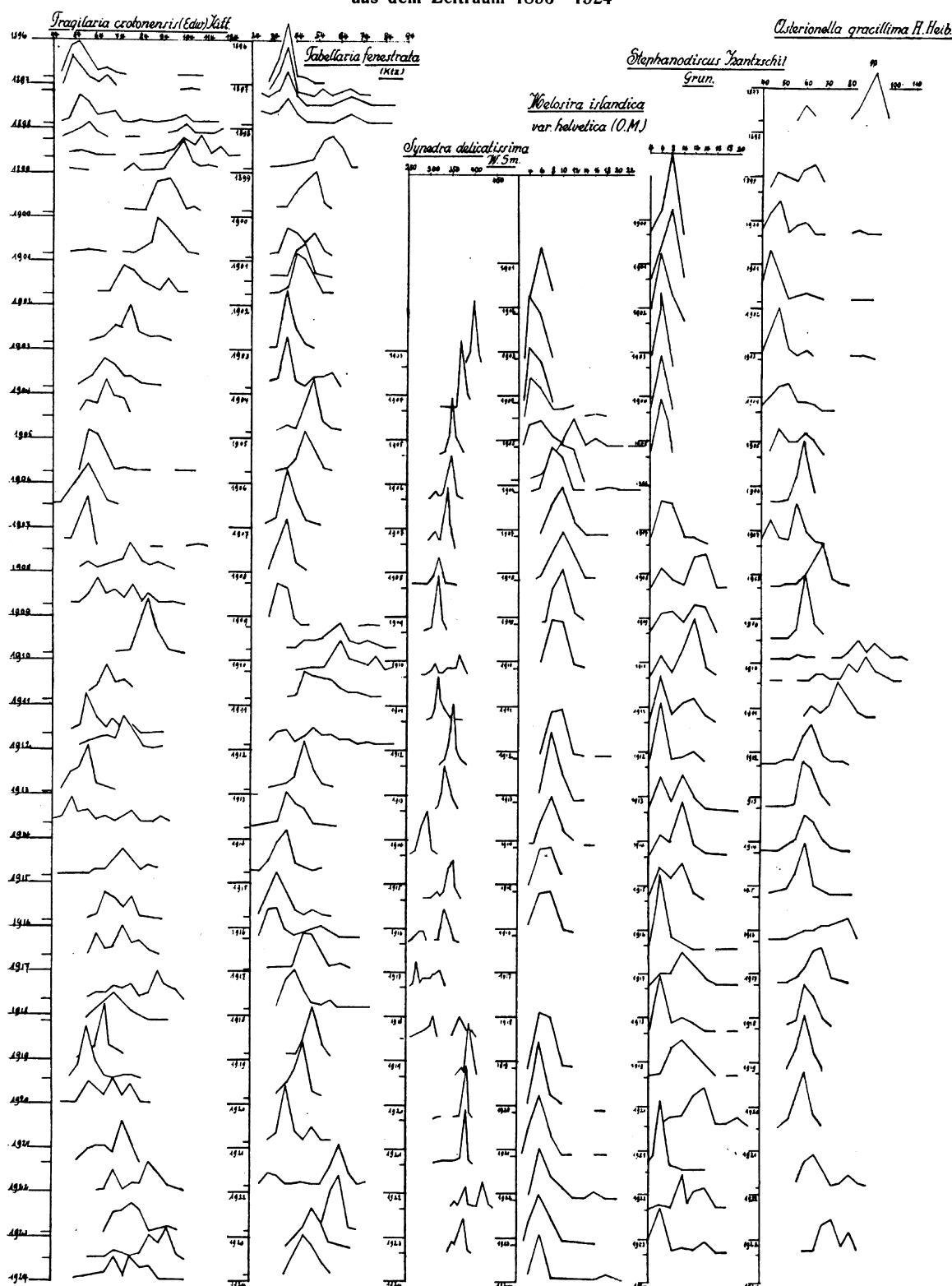


Figure 2 Nipkow's graph of yearly size distributions in the populations of major plankton diatoms in the varved sediment of Lake Zurich (Nipkow, 1927). Note that the time-scale on the vertical axis is inverted, so that the oldest varve (1896) is at the top and the last (1914) at the bottom. For each varve, one or two size distributions have been determined from individually dissected seasonal layers for each of the six planktonic taxa presented. The horizontal scales mark the measured size for each taxon: frustule length for the araphids (*Fragilaria crotonensis*, *Tabellaria fenestrata*, *Synedra delicatissima*, and *Asterionella gracillima*), and diameter for the centrics (*Melosira islandica* and *Stephanodiscus hantzschii*). Cycles of abrupt cell size increase and subsequent gradual decrease are evident in all the populations.



Figure 3 A tape peel of sediment varves of Lake Laukunlampi, East Finland. The varves are delimited by the dark brown humus-rich winter layers; each varve contains one or more light-colored diatom-rich strata. The photographed sequence spans the period 1964–1981; scale in cm.

The baseline conditions need to be known so as to obtain a realistic goal for the management or restoration measures, and advanced paleolimnology is often the sole provider of this information (e.g., Flower et al., 1997; Miettinen et al., 2005).

Geological Processes and Catchment Events

In suitable cases, the study of long varve series will allow precise dating and high-resolution diatom analyses to assess lake ecosystem dynamics on geological time scales. Long varve series are, in fact, fairly commonly found in the early postglacial sediments on northern European lakes (e.g., Ralska-Jasiewiczowa et al., 1998). Unfortunately, however, these varve records seldom extend to the present day. Typically, they provide floating chronologies for the gradual oligotrophication (decline in the level of nutrients) of the water bodies as a consequence of catchment evolution since the end of the last glaciation through processes of landscape, vegetation, and soil development. Gradual filling in of the basin and declining productivity may have led to weakened stratification and lessened anoxic tendencies, and thus eventual cessation of varve deposition in the typical development of such lakes.

Coastal meromictic lakes are a special category of basins with a high tendency for varve formation. Especially in areas of isostatic land uplift, such as the Baltic coasts in Northern Europe, and northeastern North America, there are numerous recently isolated coastal lakes, with residual or recurring salt water-induced meromixis (lack of mixing of water layers

in a lake). The diatom records in the sediments of such lakes may, of course, reveal and date the isolation process itself with great accuracy, but particularly in the arctic areas, they may also serve exceptionally well as archives of catchment development and climatic events, as demonstrated by the multidisciplinary Taconite Inlet Lakes Project (Douglas et al., 1996).

Inferring Past Climatic Changes

Much recent research has been focused on tracking the climatic signal in sedimentary diatom records. There is plenty of evidence that diatoms are indeed sensitive indicators of climatic factors, either directly or indirectly, in different climatic zones (Battarbee et al., 2001). Even here, varved sequences are of great value; besides serving as a dating tool, they even provide a record of seasonal patterns and year-to-year variability of the sedimentation process. Detailed analysis of currently depositing varve series may enable verification of the climate signal through correlations with instrumental records (e.g., Blass et al., 2007; Boes and Fagel, 2008; Lamoureux, 1999; St Jacques et al., 2009).

Overall, the main problem is to distinguish the specific climatic signals from those of catchment processes and especially from human disturbance, which has a long, variable, and often poorly documented history in most lake catchments of the world (e.g., Kienel et al., 2005; Lotter and Birks, 1997).

The climate signal appears fairly straightforward in some arctic varved lakes, in which the annual diatom production and sedimentation is typically associated with the length of the ice-free period, hence, summer temperature, and varve thickness often records snowmelt runoff, or winter snow quantity (Douglas et al., 1996; Lamoureux, 1999). In lower latitude climate zones, such relationships tend to be more complicated, but some similar patterns may be discerned. Thus, in regularly freezing dimictic lakes (lakes that mix from top to bottom twice per year), abundant spring diatoms appear to indicate a warm spring with an early ice break-up and prolonged water circulation, whereas copious winter snow would be likewise reflected in a thick minerogenic spring flood layer (e.g., Baier et al., 2004; Ojala et al., 2008). There is plenty of random variation, however, as shown by Gälman et al. (2006) in a comparison of three varve series deposited in close proximity to each other and in very similar conditions (two separate basins in one lake and one in a nearby lake), representing the late twentieth century. Even though the varve series could be correlated year by year, there was still great variance in the thickness and constitution of individual varves representing the same year.

St Jacques et al. (2009) made use of the seasonal occurrence of diatoms within varves to interpret seasonal climate patterns in a 900-year varved sequence in a lake in Minnesota, USA. They determined the relative abundances of sedimentary diatom taxa and chrysophyte scales in the dark (winter-spring) and light (summer) laminae of short sequences of typical varves, and combined this algal seasonality with pollen-inferred temperatures and other proxy data for the sequence. They found inferences for two separate limnological features, namely for the intensity of spring mixing (characterized by the abundance of *Stephanodiscus minutulus*, indicating mild winters, early ice out), and for the severity of summer stratification

(e.g., *Cyclotella pseudostelligera* and low diatom:chrysophyte scale index; indicating summer epilimnetic nutrient depletion). The varying patterns of spring mixing and summer stratification, reflecting the essentially unrelated winter and summer weather conditions, also explain the seasonal variations in nutrient levels, which may appear confusing when inferred from nonseasonally analyzed diatom assemblages.

Avenues for Further Research

Species-Level Studies of Diatoms in Sedimentary Records

The current paleoecological assessments are mainly based on community-level analysis. Inferences are drawn of observed correlations between environmental parameters and gross diatom assemblages. The ecological inference potential could be further refined by incorporation of species-level autecological information into the models. Basic algological research is still needed, especially concerning planktonic taxa and the factors affecting their population dynamics, that is, species responses to seasonal limnological events. Actual monitoring of lakes with ongoing varve deposition, combined with follow-up analysis of the sediment record, has great potential in providing this kind of improved ecological understanding of single species and of typical assemblages (Köster and Pienitz, 2006).

Detailed studies of single-species dynamics within varved sediments could, in turn, serve basic algological and ecological research. Research questions, partly outlined already by Nipkow (1927), include diatom cell-cycle dynamics (size reduction–auxosporulation cycles and their timing); within-population morphology variation; and possible genetic studies in sediments with well-preserved diatom cells. Such studies could lead to a better understanding of the species concept, especially for common taxa with diffuse and difficult species identification (e.g., *Aulacoseira*, *Stephanodiscus*).

Simola et al. (1990) is an example of microstratigraphic diatom analysis conducted with a primary aim to study the population dynamics and species interactions of planktonic taxa. This study of a 418-year long regularly varved sequence provided data on the seasonality and year-to-year dynamics of the analyzed taxa, but, unavoidably, detailed interpretation of the results was hampered by the lack of corresponding environmental data (as an aside, the 418-year population histories of some plankton diatoms presented in this article have ended up in a biostatistics textbook as typical examples of ‘chaotic’ time series).

Revisiting Well-Documented Varves for a Better Understanding of Postdepositional Processes

Taphonomic processes, or the numerous factors and events that tend to alter the representativity of a fossil assemblage in comparison with the original living communities, are a major issue in paleoecological reconstructions. However, this is rarely accounted for in routine analyses, which usually take the analyzed assemblages at their face value.

Postdepositional dissolution is a very significant factor shaping the fossil assemblages or indeed completely destroying diatom frustules in sediment deposits (see **Large Lakes**

Figure 2). Experimental studies (e.g., Battarbee et al., 2001; Flower, 1993) have demonstrated quite dramatically that the species composition of a fossil assemblage may show very little resemblance to the original one, if only a few robust forms remain while all the common ones have dissolved. In extreme cases, the remaining fossil assemblage may even be dominated by taxa that, owing to their initial rarity, would not have been recorded in the analysis of the original sample. Dissolution is exemplified by the common plankton diatoms *Urosolenia* and *Attheya*, which are hardly ever included in calibration data sets, because their very thin frustules are rarely encountered in lake sediments (their cysts may be observed, but usually not in abundance).

Revisiting well-documented varve deposits would provide very useful information for the typical postdepositional dissolution process, particularly its duration and timing, and the way it affects different species in different types of sediments. This kind of information could considerably improve the interpretation accuracy and potential in paleoecological inference studies.

See also: [Diatom Introduction](#); [Geomagnetic Excursions and Secular Variations](#); [Varved Lake Sediments](#). **Diatom Records:** [Large Lakes](#).

References

- Alefs J and Müller J (1999) Differences in the eutrophication dynamics of Ammersee and Starnberger See (Southern Germany), reflected by the diatom succession in varved sediments. *Journal of Paleolimnology* 21: 395–407.
- Baier J, Lücke A, Negendank J, Schleser G-H, and Zolitschka B (2004) Diatom and geochemical evidence of mid- to late Holocene climatic changes at Lake Holzmaar, West-Eifel (Germany). *Quaternary International* 113: 81–96.
- Battarbee R, Jones V, Flower R, et al. (2001) Diatoms. In: Smol JP, Birks HJB, and Last WM (eds.) *Tracking Environmental Change Using Lake Sediments. Vol. 3: Terrestrial, Algal, and Siliceous Indicators*, pp. 155–202. Dordrecht, The Netherlands: Kluwer.
- Blass A, Bigler C, Grosjean M, and Sturm M (2007) Decadal-scale autumn temperature reconstruction back to AD 1580 inferred from the varved sediments of Lake Silvaplana (southeastern Swiss Alps). *Quaternary Research* 68: 184–195.
- Boes X and Fagel N (2008) Relationships between southern Chilean varved lake sediments, precipitation and ENSO for the last 600 years. *Journal of Paleolimnology* 39: 237–252.
- Cocquyt C and Israël Y (2004) A microtome for sectioning lake sediment cores at a very high resolution. *Journal of Paleolimnology* 32: 301–304.
- Davidson GA (1988) A modified tape-peel technique for preparing permanent qualitative microfossil slides. *Journal of Paleolimnology* 1: 229–234.
- Dean JM, Kemp AES, Bull D, Pike J, Patterson G, and Zolitschka B (1999) Taking varves to bits: Scanning electron microscopy in the study of laminated sediments and varves. *Journal of Paleolimnology* 22: 121–136.
- Douglas M, Ludlam S, and Feeney S (1996) Changes in diatom assemblages in Lake C2 (Ellesmere Island, Arctic Canada): Response to basin isolation from the sea and to other environmental changes. *Journal of Paleolimnology* 16: 217–226.
- Flower R (1993) Diatom preservation: Experiments and observations on dissolution and breakage in modern and fossil material. *Hydrobiologia* 269/270: 473–484.
- Flower RJ, Juggins S, and Battarbee RW (1997) Matching diatom assemblages in lake sediment cores and modern surface sediment samples: The implications for lake conservation and restoration with special reference to acidified systems. *Hydrobiologia* 344: 27–40.
- Gälman V, Petterson G, and Renberg I (2006) A comparison of sediment varves (1950–2003 AD) in two adjacent lakes in northern Sweden. *Journal of Paleolimnology* 35: 837–853.
- Glew J, Smol J, and Last W (2001) Sediment core collection and extrusion. In: Last WM and Smol JP (eds.) *Tracking Environmental Change Using Lake Sediments*.

- Vol. 1: *Basin Analysis, Coring and Chronological Techniques*, pp. 73–105. Dordrecht, The Netherlands: Kluwer.
- Hall RI and Smol JP (2010) Diatoms as indicators of lake eutrophication. In: Smol JP and Stoermer EF (eds.) *The Diatoms: Applications for the Environment and Earth Sciences*, 2nd edn., pp. 122–151. Cambridge, UK: Cambridge University Press.
- Kemp A, Dean J, Pearce R, and Pike J (2001) Recognition and analysis of bedding and sediment fabric features. In: Last WM and Smol JP (eds.) *Tracking Environmental Change Using Lake Sediments. Vol. 2: Physical and Geochemical Methods*, pp. 7–22. Dordrecht, The Netherlands: Kluwer.
- Kienel U, Schwab M, and Schettler G (2005) Distinguishing climatic from direct anthropogenic influence during the past 400 years in varved sediments of Lake Holzmaar (Eifel, Germany). *Journal of Paleolimnology* 33: 327–347.
- Köster D and Pienitz R (2006) Seasonal diatom variability and paleolimnological inferences – A case study. *Journal of Paleolimnology* 35: 395–416.
- Lamoureux S (1999) Spatial and interannual variation in sedimentation patterns recorded in nonglacial varved sediments from the Canadian High Arctic. *Journal of Paleolimnology* 21: 73–84.
- Lamoureux S (2001) Varve chronology techniques. In: Last WM and Smol JP (eds.) *Tracking Environmental Change Using Lake Sediments. Vol. 1: Basin Analysis, Coring and Chronological Techniques*, pp. 247–260. Dordrecht, The Netherlands: Kluwer.
- Lotter A and Birks HJB (1997) The separation of the influence of nutrients and climate on the varve time-series of Baldeggersee, Switzerland. *Aquatic Sciences* 59: 362–375.
- Miettinen JO, Kukkonen M, and Simola H (2005) Hindcasting baseline values for water colour and total phosphorus concentration in lakes using sedimentary diatoms – Implications for lake typology in Finland. *Boreal Environmental Research* 10: 31–43.
- Nipkow F (1927) Über das Verhalten der Skelette planktischer Kieselalgen im geschichteten Tiefenschlamm des Zürich- und Baldeggersees. *Zeitschrift für Hydrologie* 4: 71–120. <http://dx.doi.org/10.3929/ethz-a-000115971>.
- Ojala AEK, Heinsalu A, Kauppi T, Alenius T, and Saarnisto M (2008) Characterizing changes in the sedimentary environment of a varved lake sediment record in southern central Finland around 8000 cal. yr BP. *Journal of Quaternary Science* 23: 765–775.
- Ralska-Jasiewiczowa M, Goslar T, Madeyska T, and Starkel L (eds.) (1998) *Lake Gościąg, Central Poland, a Monographic Study, Part 1*. Warsaw, Poland: W. Szafer Institute of Botany, Polish Academy of Science.
- Schmidt R, Höllerer H, and Wallner G (1995) A vacuum sampler for subsampling freeze-dried laminated sediments with the application to *in situ* frozen varves of Mondsee, Austria. *Journal of Paleolimnology* 14: 93–96.
- Simola H (1992) Structural elements in varved lake sediments. In: Saarnisto M and Kahra A (eds.) *INQUA Workshop on Laminated Sediments, Lammi, Finland, June 1990*, pp. 5–9. Finland: Geological Survey of Finland. Special Paper 14.
- Simola H, Hanski I, and Liukkonen M (1990) Stratigraphy, species richness and seasonal dynamics of plankton diatoms during 418 years in Lake Lovojärvi, South Finland. *Annales Botanici Fennici* 27: 241–259.
- Smol JP (2004) *Pollution of Lakes and Rivers: A Paleoenvironmental Perspective*, p. 280. Oxford, UK: Arnold and Oxford University Press.
- St Jacques J-M, Cumming P, and Smol JP (2009) A 900-yr diatom and chrysophyte record of spring mixing and summer stratification from varved Lake Mina, west-central Minnesota, USA. *The Holocene* 19: 537–547.

Large Lakes

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Introduction

Large lakes are generally defined as those with a surface area of over 500 km² (Tilzer and Serruya, 1990). They are irregularly spread out across the globe and contain over 90% of global surface freshwater resources (Herdendorf, 1982). However, perhaps a more reliable metric of a large lake is the volume of water it contains, as significant changes to surface area can occur if the lake is shallow and hydrological inputs alter. For example, extraction of water for irrigation in central Asia has resulted in the Aral Sea losing 75% of its surface area and 90% of its volume in recent decades, and the lake now exists as several separate water bodies (Micklin, 2007). The 20 largest lakes (by volume of water) are summarized in Table 1 and their geographical position is shown in Figure 1. Half of these lakes are tectonic in origin while the other half were formed from glacial activity. These lakes show some specific geographical distributions, which have important implications for Quaternary reconstructions using diatoms and other proxies. For example, all but one of the large, glacial, ice-scour lakes are found in North America north of 40° N latitude between young sedimentary strata and more northerly Precambrian shield rocks (the remaining glacial scour lake is Lake Ladoga in western Russia). Moreover, only one large lake (Hamoun i Helmand) is found between 20° N and 40° N, mainly because aridity prevents their development between these latitudes.

By definition, large glacial lakes contain only sedimentary diatom deposits formed since deglaciation, and records, therefore, commonly span less than 20 ka BP. Longer records are therefore best extracted from either extant lakes with deep basins in active rift zones, or paleolakes that have since in-filled or dried up, but the presence of which once spanned the Quaternary. The long sedimentary records of tectonic lakes are being increasingly exploited. Quaternary diatom studies are most advanced from Lake Baikal, due to the Baikal Drilling Programme (BDP), which began extracting long records in the early 1990s (Williams et al., 2001). However, under the auspices of the Intercontinental Drilling Program (<http://www.icdp-online.de>), other large lakes in tectonic basins have recently been cored, for example, Lake Titicaca in 2001 and Lake Malawi in 2005, and plans are underway for future coring expeditions, for instance in Lake Issyk-Kul.

Diatoms as Paleoenviromental Proxies

Interpretation of the climate record held by diatom proxies in the sediments of large lakes is enhanced in cases where there is knowledge of population growth and succession in the water column, and an understanding of how diatoms are transported to the sediment surface and ultimately incorporated into the sedimentary record (Mackay et al., 2003). In this respect, the quality of diatom preservation can be assessed and key,

controlling climatic variables evaluated. Diatom-inferred climatic interpretations therefore require knowledge of taphonomic processes, most notably dissolution (Ryves et al., 2001), which can be used to improve paleoenvironmental reconstructions (Ryves et al., 2009).

The types of diatom communities that can be found in large lakes depend on lake morphology (i.e., is the lake deep or shallow?), age of the lake, and prevailing climate. For example, the sensitivity of large lakes to changing water balance can often be reflected through the ratio of benthic to planktonic diatoms. In Lake Titicaca, high abundances of benthic taxa (often approaching 100%) are indicative of shallow closed-basin conditions during periods of reduced regional glaciation (Fritz et al., 2007). During periods of increased glacial extent around Lake Titicaca, water balance is much higher, the lake water is fresh, and planktonic diatoms dominate sedimentary assemblages (often >80%; *ibid.*). Planktonic diatoms in large, deep lakes may also be adapted to other hydrological processes different from those in large, shallow lakes such as increased time spent out of the photic zone due to deep-water mixing (e.g., in Lake Baikal), or the importance of internal cycling over external inflows on nutrient budgets over short timescales (e.g., in Lake Victoria; Tilzer, 1990).

Large tectonic lakes are undoubtedly some of the world's oldest water bodies, and this great age tends to result in planktonic diatom assemblages with low diversity, especially in comparison to large, glacially formed lakes, which can contain species assemblages representing many planktonic genera (Stoermer and Edlund, 1999). Attempts to explain low plankton diversity in lakes such as Lake Baikal include competitive exclusion (e.g., Stoermer and Edlund, 1999) and resource gradient and competition theory (Edlund, 2006).

Case Studies

The range of diatom-based paleolimnological studies is vast; however, these studies cover mostly Late Glacial–Holocene periods. There are rather few diatom records from large lakes covering the whole of the Quaternary, in comparison to marine records. Four case studies are considered here because they represent ongoing challenges in the field of paleoclimatology/paleolimnology over the Quaternary. Case study 1 is from Lake Baikal in Russia, situated within an active tectonic region of the Eurasian plate (51°28′–55°47′N and 103°43′–109°58′E; Figure 1). Quaternary diatom records, associated trends in biogenic silica (BioSi), and $\delta^{18}\text{O}$ analysis of diatom silica (e.g., Mackay et al., 2011) from this lake have greatly increased our understanding of climate variability in continental regions far from oceanic influences, hence the comprehensive treatment of this lake here. Case study 2 is from Lake Titicaca, which lies between Bolivia and Peru in South America (15°45′S 69°30′W; Figure 1). This lake is found to the north

Table 1 Adapted from [Herdendorf \(1982\)](#) but updated from <http://www.worldlakes.org/>

<i>Volume rank</i>	<i>Name</i>	<i>Volume (km³)</i>	<i>Latitude</i>	<i>Longitude</i>	<i>Origin</i>	<i>Country</i>
1	Caspian	78 200	42°00' N	50°00' E	Tectonic	Azerbaijan Iran Kazakhstan Russian Federation Turkmenistan
2	Baikal	23 600	54°00' N	109°00' E	Tectonic	Russia
3	Tanganyika	19 000	6°00' S	29°30' E	Tectonic	Burundi Congo (Democratic Republic) Tanzania Zambia
4	Superior	12 100	47°30' N	88°00' W	Glacial	Canada USA
5	Malawi	7 775	12°00' S	34°30' E	Tectonic	Malawi Mozambique Tanzania
6	Michigan	4 920	44°00' N	87°00' W	Glacial	Canada USA
7	Huron	3 540	45°00' N	81°15' W	Glacial	Canada USA
8	Victoria	2 760	01°00' S	33°00' E	Tectonic	Kenya Tanzania Uganda
9	Great Bear	2 292	66°00' N	121°00' W	Glacial	Canada
10	Great Slave	2 088	62°40' N	114°00' W	Glacial	Canada
11	Issyk-Kul	1 738	42°30' N	77°15' E	Tectonic	Kyrgyzstan
12	Ontario	1 640	43°30' N	77°45' W	Glacial	Canada USA
13	Titicaca	932	15°45' S	69°30' W	Tectonic	Bolivia Peru
14	Ladoga	908	61°00' N	31°30' E	Glacial	Russia
15	Hamoun i Helmand	510	31°00' N	61°15' E	Tectonic	Afghanistan Iran
16	Erie	484	42°15' N	81°15' E	Glacial	Canada USA
17	Hovsgol	380	51°00' N	100°30' E	Tectonic	Mongolia
18	Winnipeg	371	52°30' N	97°45' W	Glacial	Canada
19	Kivu	333	2°13' S	29°10' W	Tectonic	Congo (Democratic Republic) Rwanda
20	Nipigon	320	49°50' N	88°31' W	Glacial	Canada

of the Altiplano of Peru, the world's second highest plateau (after Tibet), and as such Lake Titicaca is the highest large lake in the world. Paleoclimate records for this site are available for much of the late Quaternary period. The third case study uses diatom records from one of the world's largest tropical lakes, Lake Victoria in Africa (01°00' S 33°00' E; [Figure 1](#)), bordered by Kenya, Tanzania, and Uganda. Lake Victoria is relatively shallow, and observations in the 1960s revealed that the lake had dried out completely toward the end of the last glacial ([Kendall, 1969](#)). It was this observation that challenged conventional orthodoxy at the time, which held that tropical regions experienced wetter climates during the Late Glacial. Although there has been much debate since then on the extent and timing of this desiccation event, recent work indicates that it occurred between approximately 15.9 and 14.2 cal ka BP ([Stager et al., 2002](#)). The final case study focuses on sediments recently extracted from Lake Malawi ([Figure 1](#); spanning approximately the last 144 ka) as part of the IDP 2005 programme ([Stone et al., 2010](#)). Lake Malawi (9–14° S) is of

tectonic origin, and is the world's fifth largest lake ([Table 1](#)). It is bordered by Malawi, Mozambique, and Tanzania, and is the southernmost of the East African Great Lakes.

Lake Baikal

Lake Baikal, situated in southeast Siberia ([Figure 1](#)) has long been recognized as one of the world's most unusual freshwater ecosystems, especially in terms of its age (in excess of 20 Myr), depth (~1640 m) and volume (23 600 km³), its uninterrupted sedimentary record and the high degree of endemism that exists within the lake. Another important feature that sets Lake Baikal apart from other large, deep lakes is that its entire water column is saturated throughout with oxygen, due to regular renewal of the deep waters every spring and autumn. This results in the oxidation of even the deepest bottom waters, allowing an extensive and almost wholly endemic deep-water fauna to survive.



Figure 1 Map showing the distribution of the world's 20 most voluminous lakes. The numbers given correspond to the ranking of the lakes, as indicated in Table 1.

Diatoms are important primary producers and are crucial to the ecosystem functioning of Lake Baikal (Popovskaya, 2000). The most common planktonic diatoms in Lake Baikal are a mixture of endemic taxa (e.g., *Aulacoseira baicalensis*, *Cyclotella minuta*, and *Stephanodiscus meyerii*) and cosmopolitan taxa (e.g., *Synedra acus* and its varieties, and *Nitzschia acicularis*). The distribution of Lake Baikal diatoms exhibits significant spatial variability across the lake, dependant on a number of factors including climate (e.g., ice and snow cover) and hydrological factors such as water depth and fluvial input (Mackay et al., 2006). As much as 40% of the lake's diatoms may be endemic although this figure is being revised upward as more comprehensive studies in deeper littoral zones (>20 m) are carried out (Flower, 2005) and centers of hyperendemicity become recognized around the lake. The diatom assemblages in Lake Baikal contain several examples of species flocks, and relict distributions of previously more widespread taxa (ibid.). Genkal and Bonderenko (2006) further proposed that many of the common endemic planktonic diatoms found in Lake Baikal today should be reclassified as relicts because they are morphologically very similar to neighboring mountain lake species.

The paleoclimatic archive of Lake Baikal sediments is of global importance, given its position in one of the most continental regions of the world where continuous, high-resolution records spanning the Quaternary are otherwise sparse. Lake Baikal diatoms and associated BioSi have proved

to be extremely useful paleoclimate proxies, in part because of the intense seasonality of the region, and also because of the size of the lake, which covers 4° latitude, across which a distinct climatic gradient exists. Given the depth and volume of Lake Baikal, it is important to assess factors that may affect diatom integrity before their final incorporation in the sedimentary record.

As in the marine environment, the entire water column of Lake Baikal is undersaturated with respect to silica (Falkner et al., 1991), and therefore diatom valve dissolution has always been an important process through glacial and interglacial cycles. In a study comparing recent diatom fluxes through the water column with fluxes being incorporated into the sediments, Ryves et al. (2003) demonstrated that in the south basin of Lake Baikal, only ~1% of the valves are preserved in the sedimentary record. Once incorporated into the sediments, dissolution of diatom silica continues into the pore waters, until saturation of pore waters by Si is reached, thereby limiting further dissolution. During glacial periods, few diatoms were preserved in the sedimentary record. This is most likely due to more intense relative dissolution processes operating on fewer diatom cells growing during colder climates (Carter and Colman, 1994; Mackay, 2007; Swann and Mackay, 2006).

Although the impact of dissolution on the diatom records from Lake Baikal has been recognized since some of the earliest studies, only recently have the quantitative impacts been

determined (Battarbee et al., 2005; Mackay et al., 2005; Ryves et al., 2003). Dissolution is problematic from a paleoenvironmental perspective because partially dissolved valves are often difficult to identify – see Figure 2. Furthermore, assemblages may be biased toward more robust, heavily silicified species, thus influencing paleoclimatological reconstructions, or more seriously, taxa may be removed from the record altogether.

One of the first studies to relate changes in Lake Baikal diatom composition to changes in orbital forcing was by Grachev et al. (1998) who worked on BDP cores extracted from the Academician Ridge. They highlighted the observation that relatively few planktonic taxa tended to dominate the sedimentary assemblages during interglacials. Moreover, they also noted the successional nature of the changing flora, especially during approximately the last 1 Myr, as diatom abundances suggested progressively longer, more severe glaciations, with different species perhaps dominating the record at different stages for less than 1 ka. Notably, although many of these taxa are now extinct, several of these diatom assemblages include species commonly found in the lake today, such as *C. minuta*, *A. baicalensis*, and *Aulacoseira skvortzowii*.

However, the most detailed study on diatom species responses to orbital forcing indicates that over the last 0.8 Myr (since the onset of the mid-Pleistocene revolution) 31 local diatom assemblage zones can be characterized, corresponding to MIS 1–19 (Khursevich et al., 2001; Figure 3). This study is important because Khursevich et al., make the link between diatom evolution and insolation changes, thereby adding to the debate on evolutionary pressures on relatively short time-scales. More specifically, species responses varied depending

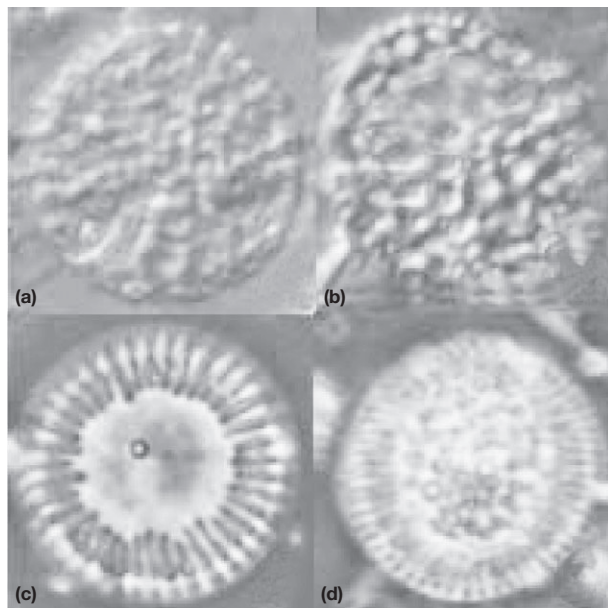


Figure 2 Dissolved and near pristine images of two endemic, planktonic diatoms found in Lake Baikal sediments. Images (a) and (b) are dissolved valves of *C. gracilis* and *C. minuta*, respectively, found in a marine isotope stage (MIS) 3 sequence extracted from the north basin of Lake Baikal. Images (c) and (d) are of pristine valves of the same species, this time found in the MIS 5e sequence of the same core. Photographs courtesy of George Swann and Patrick Rioual.

on whether insolation amplitude was small (e.g., as occurred between MIS 15a and 11, corresponding to the almost continuous presence of *Stephanodiscus flabellatus* var. *distinctus* Khursevich and its varieties) or large (e.g., as occurred between MIS 19 and 15e, during which time each precessional cycle was dominated by a different species, which subsequently became extinct). Furthermore, during the last 0.8 Myr, there were also apparent differential speciation rates between the genera, with 16 new species of *Stephanodiscus* being recorded in comparison to only four for *Cyclotella* (Khursevich et al., 2000, 2001). Importantly, during this time, ice volume increased during each subsequent glaciation, indicative of prolonged glacial periods. Such diatom responses to climate during the Pleistocene have not been witnessed in any other freshwater or marine environment (Khursevich et al., 2001). Detailed records also exist for changes in diatom assemblage composition prior to the last 0.8 Myr, with the onset of Northern Hemisphere Glaciation influencing extinctions in at least two genera in the

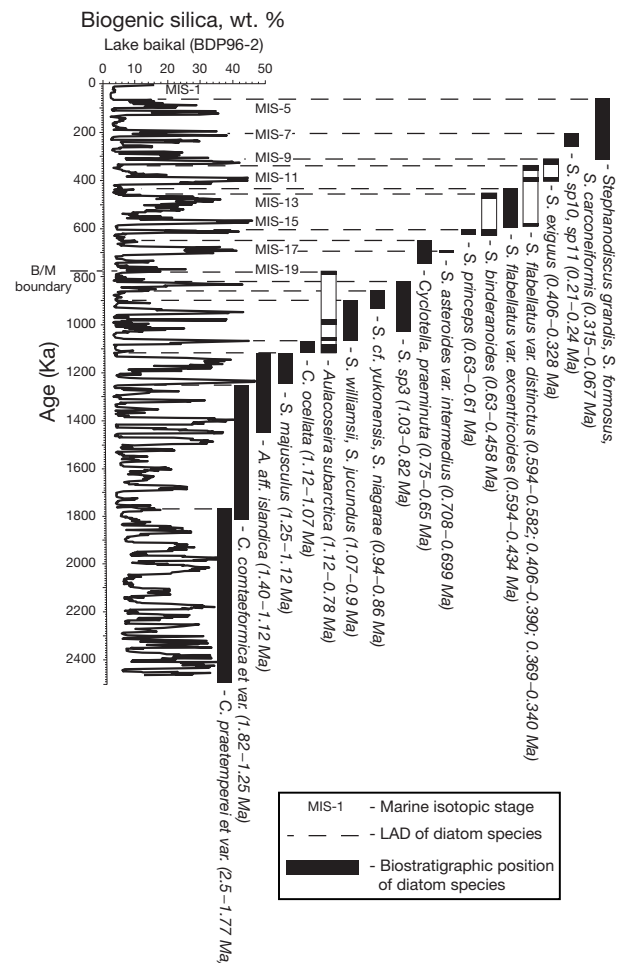


Figure 3 Biostratigraphic position of major diatom species plotted against a biogenic silica record over the entire Quaternary in Lake Baikal. Redrawn from Khursevich GK, Karabanov EB, Williams DF, Kuzmin MI, and Prokopenko AA (2000) Evolution of freshwater centric diatoms within the Baikal rift zone during the late Cenozoic. In: Minoura K (ed.) *Lake Baikal: A Mirror in Time and Space for Understanding Global Change Processes*, pp. 146–154. Amsterdam: Elsevier.

lake toward the beginning of the Quaternary. Following this time, endemic *Cyclotella* species developed and dominated the diatom flora, including *C. praetemperei*, until approximately 1.7 Myr, followed by *C. comtaeformica* up to approximately 1.25 Myr (Khursevich et al., 2000).

Lake Titicaca

Lake Titicaca was formed by tectonic activity approximately 3 Myr ago. This large, monomictic lake is an important site for Quaternary studies because (i) it is the only large, deep, freshwater lake in South America; (ii) the lake itself is sensitive to changes in precipitation (Tapia et al., 2003); and (iii) its sedimentary archive extends back through many glacial-interglacial cycles (Fritz et al., 2007), and hence offers the possibility of high-resolution paleoclimate reconstructions in this understudied region of the globe (Baker et al., 2001).

Detailed diatom analyses of late Quaternary sediments from Lake Titicaca have contributed significantly to the debate on tropical precipitation trends since the last glaciation, especially within the Amazon Basin. In general terms, diatom-based reconstructions have demonstrated that during the Late Glacial period, the climate over the Altiplano was wet, whereas a significant shift to a more prevalent drier climate occurred during the early to middle Holocene (Baker et al., 2001).

Although these scenarios are based on qualitative interpretations of diatom communities present in the lake, they are robust because they are confirmed by other proxies found in the sediment archive (e.g., Cross et al., 2000). Initially, Baker et al. (2001) grouped diatoms according to their preferred habitat of living in the planktonic or littoral regions of the lake. High proportions of sedimentary benthic or littoral taxa were therefore attributed to prevailing arid climates resulting in lower lake levels. Conversely, high proportions of planktonic taxa were indicative of higher lake levels (increased lake volume) due to prevailing wetter climates. High lake levels in Lake Titicaca and wet Amazonian conditions correspond today to

cold, northern equatorial Atlantic sea-surface temperatures. On this basis, changing diatom assemblages highlight that Lake Titicaca has been sensitive to changes in the North Atlantic thermohaline circulation during both the Younger Dryas and even Bond cycles through the Holocene (Baker et al., 2001).

Tapia et al. (2003) considered these diatom responses in more detail. Periods of deep water in the lake, indicative of wetter climate, were characterized by planktonic diatoms, especially the *Cyclotella stelligera* complex between 30 and 24 ka BP and *Cyclotella andina* throughout the Late Glacial period up to approximately 8 ka BP. Two further diatom groups were recognized in the sedimentary profiles: saline indifferent taxa, especially *Cyclotella meneghiniana*, which has a very wide tolerance to varying lake salinities, and saline taxa including *Chaetoceros muelleri*, which is indicative of low lake levels and corresponding paleoclimate aridity.

Using these groups, a paleoclimatic synthesis was developed (Tapia et al., 2003; Figure 4). Between 26 and 12.5 ka BP, freshwater planktonic taxa (mainly *C. andina* and *C. stelligera*) dominated, suggesting that lake levels were high and prevailing climate was therefore wetter, especially during the period coincident with the Younger Dryas in the Northern Hemisphere. High lake levels were interrupted by a period of aridity causing a decline in lake water volume during the transition from the Late Glacial into the early Holocene (as characterized by an increase in benthic taxa). However, this arid period was short lived, as planktonic diatoms again dominated the lake between approximately 10 and 8 ka BP, indicating that high lake-level stands can be compared to the wet Coipasa event in the central Altiplano (Sylvestre et al., 1999). The shift to prevailing drier climate during the early to mid-Holocene (as suggested by Baker et al. (2001) above) is characterized by marked increases in benthic taxa and *C. meneghiniana*, suggesting that a decline in lake level was accompanied by small increases in salinity. This lake-level decline culminated in the appearance of saline taxa for the first time in Lake Titicaca (*C. muelleri*) from 6 to 3.5 ka BP. However, low lake-level stands were temporary, and from approximately

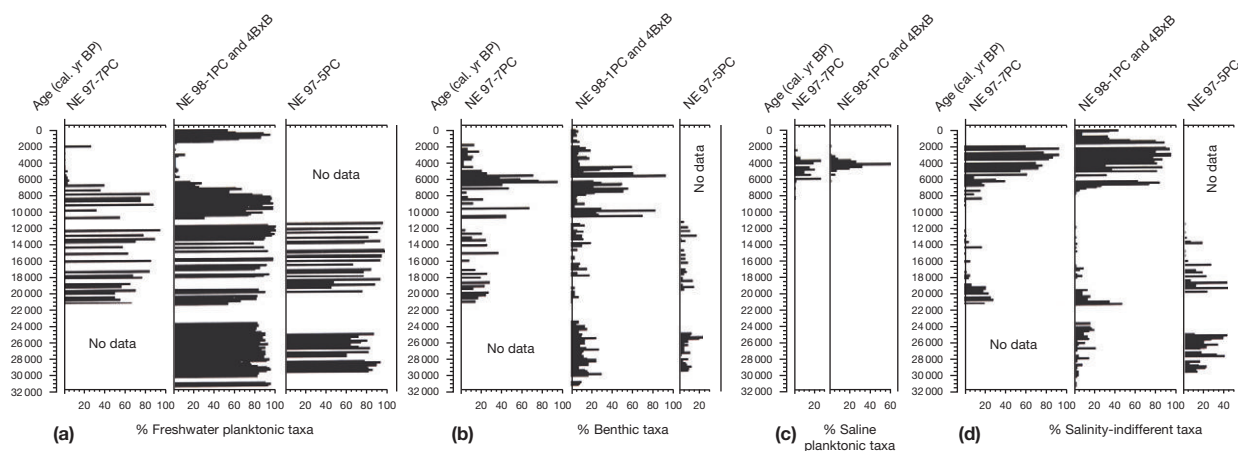


Figure 4 Changes in diatom abundance for the past 31.5 ka BP in Lake Titicaca. Diatom groups represent the four ecological groups highlighted in the text. Gaps in the stratigraphy are zones barren of diatom valves or where diatom concentrations were too low to generate an accurate diatom count. Note that core NE97-7PC does not extend beyond 21 ka BP, while core NE97-5PC does not extend through the Holocene period. Reproduced from Tapia PM, Fritz SC, Baker PA, Seltzer GO, and Dunbar RB (2003) A late Quaternary diatom record of tropical climatic history from Lake Titicaca (Peru and Bolivia). *Palaeogeography, Palaeoclimatology, Palaeoecology* 194: 139–164.

4 cal ka BP, freshwater planktonic taxa began to grow and then dominate the flora again.

In 2001, new overlapping drill cores of ~136 m were extracted from Lake Titicaca as part of the GLAD-800 coring system. These sediments span approximately the last 370 ka (Fritz et al., 2007), and again diatom species were grouped as above (i.e., freshwater plankton, saline plankton, saline indifferent, and benthic species). Periods of positive water balance are characterized by high abundances of freshwater diatoms and low abundances of benthic taxa. These periods coincide with increased glacial extent, linked to higher rainfall and lower temperatures (Fritz et al., 2007). Low lake levels in Titicaca were inferred during MIS 5e, due to high abundances of the saline species *C. muelleri* – it is likely that the lake was a closed system at this time, influenced by the prevailing warm and dry conditions.

Lake Victoria

Lake Victoria was also formed by tectonic activity. Although Lake Victoria has an outlet into the Nile River, this accounts for only ~10% of water loss from the lake each year; the remaining ~90% is lost through evaporative processes. Lake Victoria is relatively shallow, with a mean water depth of ~40 m, and consequently small fluctuations in precipitation have marked effects on the lake's hydrology. Lake Victoria is monomictic and overturn occurs during the dry months of June–September, driven by the trade winds. This overturn period determines nutrient availability in the lake and is therefore crucial in controlling algal productivity, allowing diatom populations (including *Aulacoseira* spp. and *Stephanodiscus* spp.) to thrive in the lake. In recent years, however, primary production is increasingly determined by other algal groups, including diatoms indicative of more stable water conditions (e.g., *Nitzschia* spp.) and cyanobacteria (Hecky, 1993; Verschuren et al., 2002). Recent diatom analyses suggests that eutrophication in Lake Victoria is spatially complex and time transgressive, and may be related to longer term nutrient enrichment and climate instability (Stager et al., 2009).

Throughout the Late Glacial–Holocene period, Lake Victoria has been shown to be a sensitive recorder of climate variability because its hydrological regime is mainly controlled by precipitation/evaporation. Paleoenvironmental interpretations are based on the knowledge of the ecologies of different diatom taxa. For example, *Aulacoseira* species (e.g., *A. granulata*) are generally heavily silicified, and they therefore require considerable turbulence in the water column to keep their valves suspended in the photic zone. *Nitzschia* species (e.g., *N. acicularis*), on the other hand, are often more lightly silicified and have a large surface-area-to-volume ratio, which allows them to remain buoyant in more stable water conditions. Finally, benthic taxa, including *Fragilaria* species, are again indicative of shallow-water conditions, and when their numbers increase in lake sediment archives, this is often indicative of declining lake levels. Using these characteristics, diatom-inferred arid phases in this region of tropical Africa are coincident with weakening of Afro-Asian monsoon systems, linked to changes in North Atlantic thermohaline circulation associated with such phenomena as Heinrich events and meltwater pulses.

One of the best high-resolution records of Late Glacial diatom-inferred climate changes clearly demonstrates that

water levels in Lake Victoria were much lower during the Younger Dryas than at the start of the Holocene (Figure 5; Stager et al., 2002). Relative lake level is inferred from correspondence analysis axis-2 diatom sample scores from a core taken in the Damba Channel in the north of Lake Victoria. Stager et al. (1997) suggest that these benthic assemblages (consisting of taxa such as *Fragilaria* spp.) are indicative of lake depth. Periodicities in the diatom record from Lake Victoria are further interpreted as being indicative of Dansgaard–Oeschger cycles operating on precipitation trends in tropical Africa (Stager et al., 1997).

Further research exploiting Lake Victoria diatoms has reconstructed past change at very high resolution (1–6 years) for the last 1 ka. Diatoms were once again used to infer past changes in lake level, but a diatom-inferred conductivity model was also developed and applied to a 2-m core with a robust chronology (Stager et al., 2005; Figure 6). Increases in diatom-inferred lake-level changes over the last millennium were coincident with declines in solar activity, especially during the period commonly referred to as the Little Ice Age. Although solar fluctuations on these timescales are low in amplitude, the authors suggest that when insolation is reduced, cloud cover is increased, resulting in a decrease in evaporation from Lake Victoria. However, in the last 200 years, higher lake levels are positively associated with increased sunspot numbers, and so interactions are complex

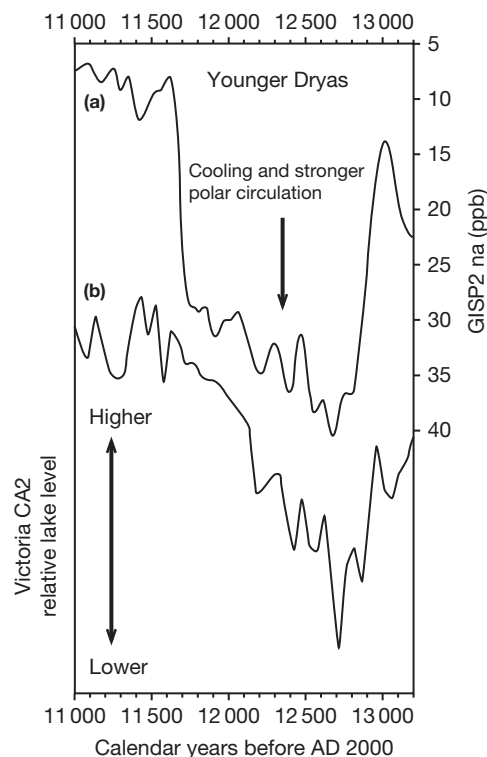


Figure 5 Details of the Younger Dryas cooling episode registered in the (a) GIISP2 Na (ppb) record and (b) Lake Victoria's relative lake level, inferred from correspondence analysis, axis 2 diatom assemblage sample scores. Reproduced from Stager JC, Cumming B, and Meeker L (1997) A high-resolution 11,400-Yr diatom record from Lake Victoria, East Africa. *Quaternary Research* 47: 81–89.

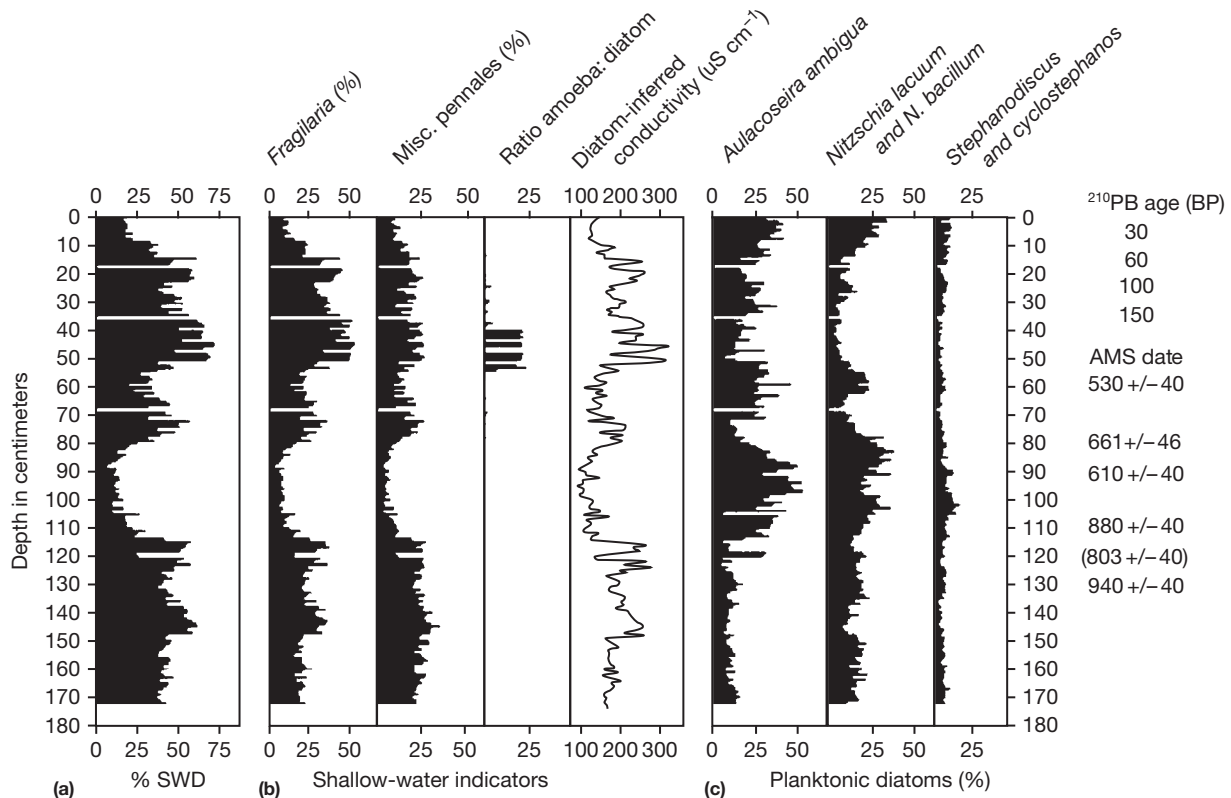


Figure 6 Major diatom taxa in core P2K-1 from Lake Victoria (Stager et al., 2005). (a) Percentages of shallow-water diatoms (SWDs) in fossil assemblages versus depth; (b) shallow-water indicator taxa and diatom-inferred conductivity; and (c) planktonic diatoms indicating greater lake depths and reduced conductivity. To the right are uncorrected AMS dates and selected ^{210}Pb chronology. Reproduced from Stager JC, Ryves D, Cumming BF, Meeker LD, and Beer J (2005) Solar variability and the levels of Lake Victoria, East Africa, during the last millennium. *Journal of Paleolimnology* 33: 243–251.

and undoubtedly confounded by changes in thermohaline circulation in both the Atlantic and Pacific oceans.

Lake Malawi

Lake Malawi has a very seasonal climate: summers (December–April) are warm and wet, and are influenced by the southward migration of the intertropical convergence zone (ITCZ). Winters on the other hand (May–August) are much cooler and drier because the ITCZ migrates north toward the equator. Although Lake Malawi has an outlet (the Shire River in the southern basin), over 80% of water is lost through evaporation (Spigel and Coulter, 1996). Moreover, although Lake Malawi is fed by six major inflows, over 60% of hydrological input is from rain, which falls directly on the surface of the lake (Spigel and Coulter, 1996). Currently, anoxic conditions prevail below a water depth of ~230 m, which results in a sedimentary archive that sensitively records changes in the lake's hydrological balance (Stone et al., 2010). During winter, diatoms can comprise the majority of phytoplankton in Lake Malawi, but for the rest of the year, nutrients in surface waters are low, leading to low levels of diatom production (ibid.).

The longest diatom record obtained for Lake Malawi (as part of the Lake Malawi Scientific Drilling Project) spans the interval from approximately 144 ka to the onset of the

Holocene. However, many species in the early sediment record are not found in the lake today. Interpretation of past assemblages therefore takes account of diatom life strategies in the open water (e.g., euplankton, meroplankton, and tychoplankton) and known ecologies of dominant groups (e.g., if species are stephanodiscoid or cyclotelloid) (Stone et al., 2010). Principal components analysis (PCA) of major diatom groups (along with other biological and mineral proxies) clearly demonstrates a water-depth gradient along the first axis, reflecting lake-level changes of many hundreds of meters. During the early part of the record, prior to 70 ka, proxy evidence suggests that Lake Malawi was often more shallow and better mixed than it is today. For example, between 133 and 125 ka, the diatom assemblage was dominated by diatom species tolerant of saline lake water, for example, *Cyclotella* compare with *quirlensis*, while between 109 and 97 ka, shallow-water *Aulacoseira* species dominate assemblages (*A. ambigua*) together with abundant ostracods, indicative of a very shallow lake with complete ventilation of the water column (Stone et al., 2010). Greater variability in lake level prior to 70 ka occurred at a time when precessional forcing was strong due to high eccentricity. Post 70 ka, Lake Malawi was consistently deep, apart from two abrupt lowstands between approximately 64–62 and 47–30 ka, which were marked by, among others, shallow-water *Aulacoseira* spp in high abundance.

Conclusion

The case studies highlighted here indicate that diatoms make powerful proxies of past environmental change over the Quaternary period. For large lakes, by far the majority of the studies are focused on Late Glacial–Holocene changes. This is in part because half of the world's 20 largest lakes are glacial in origin and all but one of these lie in North America. Paleoenvironmental interpretations based on diatoms can be made using both qualitative aspects of diatom autecologies (e.g., habitat preference, morphology) and quantitative inferences, including the use of transfer function methodologies. In the field of paleoclimatology, increasingly sophisticated questions are being asked with respect to the magnitude of climate variability over different timescales and across different regions of the world. High-resolution diatom analyses from large lakes are now considered one of the most robust and useful tools in the field of Quaternary research.

See also: [Diatom Introduction](#). **Diatom Methods:** [Salinity and Climate Reconstructions from Continental Lakes](#). **Diatom Records:** [Diatom Fossil Records from Marine Laminated Sediments](#); [Freshwater Laminated Sequences](#).

References

- Baker P, Seltzer G, Fritz S, et al. (2001) The history of South American tropical precipitation for the past 25,000 years. *Science* 291: 640–643.
- Battarbee RW, Mackay AW, Jewson D, Ryves DB, and Sturm M (2005) Differential dissolution of Lake Baikal diatoms: Correction factors and implications for palaeoclimatic reconstruction. *Global and Planetary Change* 46: 75–86.
- Carter SJ and Colman SM (1994) Biogenic silica in Lake Baikal sediments: Results from 1990–1992 American cores. *Journal of Great Lakes Research* 20: 751–760.
- Cross SL, Baker PA, Seltzer GO, Fritz SC, and Dunbar RB (2000) A new estimate of the Holocene lowstand level of Lake Titicaca, central Andes, and implications for tropical paleohydrology. *The Holocene* 10: 21–32.
- Edlund MB (2006) Persistent low diatom plankton diversity within the otherwise highly diverse Lake Baikal ecosystem. *Nova Hedwigia Beihefte* 130: 339–356.
- Falkner KK, Measures CI, Herbelin SE, and Edmond JM (1991) The major and minor element geochemistry of Lake Baikal. *Limnology and Oceanography* 36: 413–423.
- Flower RJ (2005) A review of diversification trends in diatom research with special reference to taxonomy and environmental applications using examples from Lake Baikal and elsewhere. *Proceedings of the Californian Academy of Sciences* 56: 107–128.
- Fritz SC, Baker PA, Seltzer GO, et al. (2007) Quaternary glaciation and hydrologic variation in the South American tropics as reconstructed from the Lake Titicaca drilling project. *Quaternary Research* 68: 410–420.
- Genkal SI and Bonderenko NA (2006) Are the Lake Baikal diatoms endemic? *Hydrobiologia* 568: 143–153.
- Grachev MA, Vorobyova SS, Likhoshvay YV, et al. (1998) A high-resolution diatom record of the palaeoclimates of east Siberia for the last 2.5 My from Lake Baikal. *Quaternary Science Reviews* 17: 1101–1106.
- Hecky RE (1993) The eutrophication of Lake Victoria. *Verhandlungen Internationale Vereinigung Limnologie* 25: 39–48.
- Herdendorf CE (1982) Large lakes of the world. *Journal of Great Lakes Research* 8: 379–412.
- Kendall RL (1969) An ecological history of the Lake Victoria basin. *Ecological Monograph* 39: 121–176.
- Khursevich GK, Karabanov EB, Prokopenko AA, et al. (2001) Insolation regime in Siberia as a major factor controlling diatom production in Lake Baikal during the past 800,000 years. *Quaternary International* 80–81: 47–58.
- Khursevich GK, Karabanov EB, Williams DF, Kuzmin MI, and Prokopenko AA (2000) Evolution of freshwater centric diatoms within the Baikal rift zone during the late Cenozoic. In: Minoura K (ed.) *Lake Baikal: A Mirror in Time and Space for Understanding Global Change Processes*, pp. 146–154. Amsterdam: Elsevier.
- Mackay AW (2007) The paleoclimatology of Lake Baikal: A diatom synthesis and prospectus. *Earth-Science Reviews* 82: 181–215.
- Mackay AW, Jones VJ, and Battarbee RW (2003) Approaches to Holocene climate reconstruction using diatoms. In: Mackay AW, Battarbee RW, Birks HJB, and Oldfield F (eds.) *Global Change in the Holocene*, ch. 20, pp. 294–309. London: Arnold.
- Mackay AW, Ryves DB, Battarbee RW, et al. (2005) 1000 years of climate variability in central Asia: Assessing the evidence using Lake Baikal diatom assemblages and the application of a diatom-inferred model of snow thickness. *Global and Planetary Change* 46: 281–297.
- Mackay AW, Ryves DB, Morley DW, Jewson DJ, and Rioual P (2006) Assessing the vulnerability of endemic diatom species in Lake Baikal to predicted future climate change: A multivariate approach. *Global Change Biology* 12: 2297–2315.
- Mackay AW, Swann GEA, Brewer T, et al. (2011) A reassessment of Lateglacial–Holocene diatom oxygen isotope records from Lake Baikal using a mass balance approach. *Journal of Quaternary Science* 26: 627–634.
- Micklin P (2007) The Aral Sea disaster. *Annual Review of Earth and Planetary Sciences* 35: 47–72.
- Popovskaya GI (2000) Ecological monitoring of phytoplankton in Lake Baikal. *Aquatic Ecosystem Health and Management* 3: 215–225.
- Ryves DB, Battarbee RW, and Fritz SC (2009) The dilemma of disappearing diatoms: Incorporating diatom dissolution data into palaeoenvironmental modelling and reconstruction. *Quaternary Science Reviews* 28: 120–136.
- Ryves DB, Jewson DH, Sturm M, et al. (2003) Quantitative and qualitative relationships between planktonic diatom communities and diatom assemblages in sedimenting material and surface sediments in Lake Baikal, Siberia. *Limnology and Oceanography* 48: 1643–1661.
- Ryves DB, Juggins S, Fritz SC, and Battarbee RW (2001) Experimental diatom dissolution and the quantification of microfossil preservation in sediments. *Palaeogeography, Palaeoclimatology, Palaeoecology* 172: 99–113.
- Spigel RH and Coulter GW (1996) Comparison of hydrology and physical limnology of the East African Great Lakes: Tanganyika, Malawi, Kivu and Turkana (with reference to some North American Great Lakes). In: Johnson TC and Odada EO (eds.) *The Limnology, Climatology and Paleoclimatology of the East African Lakes*, pp. 103–139. Amsterdam: Gordon & Breach.
- Stager JC, Cumming B, and Meeker L (1997) A high-resolution 11,400-Yr diatom record from Lake Victoria, East Africa. *Quaternary Research* 47: 81–89.
- Stager JC, Hecky RW, Grzesik D, Cumming B, and Kling H (2009) Diatom evidence for the timing and causes of eutrophication in Lake Victoria, East Africa. *Hydrobiologia* 636: 463–478.
- Stager JC, Mayewski PA, and Meeker LD (2002) Cooling cycles, Heinrich event 1, and the desiccation of Lake Victoria. *Palaeogeography, Palaeoclimatology, Palaeoecology* 183: 169–178.
- Stager JC, Ryves D, Cumming BF, Meeker LD, and Beer J (2005) Solar variability and the levels of Lake Victoria, East Africa, during the last millennium. *Journal of Paleolimnology* 33: 243–251.
- Stoermer EF and Edlund MB (1999) No paradox in the plankton? – Diatom communities in large lakes. In: Mayama S, Idei M, and Koizumi I (eds.) *Proceedings of the 14th International Diatom Symposium, Tokyo*, pp. 49–61. Koenigstein: Koeltz Scientific Books.
- Stone JF, Westover KS, and Cohen AS (2010) Late Pleistocene paleohydrography and diatom paleoecology of the central basin of Lake Malawi, Africa. *Palaeogeography, Palaeoclimatology, Palaeoecology* 303: 51–70. <http://dx.doi.org/10.1016/j.palaeo.2010.01.012>.
- Swann GEA and Mackay AW (2006) Potential limitations of biogenic silica as an indicator or abrupt climate change in Lake Baikal, Russia. *Journal of Paleolimnology* 36: 81–89.
- Sylvestre F, Servant M, Servant-Vildary S, Causse C, Fournier M, and Ybert J-P (1999) Lake-level chronology on the Southern Bolivian Altiplano (18°–23°S) during late-Glacial time and the early Holocene. *Quaternary Research* 51: 54–66.
- Tapia PM, Fritz SC, Baker PA, Seltzer GO, and Dunbar RB (2003) A late Quaternary diatom record of tropical climatic history from Lake Titicaca (Peru and Bolivia). *Palaeogeography, Palaeoclimatology, Palaeoecology* 194: 139–164.
- Tilzer MM (1990) Environment and physiological control of phytoplankton productivity in large lakes. In: Tilzer MM and Serruya C (eds.) *Large Lakes: Ecological Structure and Function*, pp. 339–367. Berlin: Springer-Verlag.
- Tilzer MM and Serruya C (eds.) (1990) *Large Lakes: Ecological Structure and Function*. Berlin: Springer.
- Verschuren D, Johnson TC, Kling HJ, et al. (2002) History and timing of human impact on Lake Victoria, East Africa. *Proceedings of the Royal Society of London B* 269: 289–294.
- Williams DF, Kuzmin MI, Prokopenko AA, Karabanov EB, Khursevich GK, and Bezrukova EV (2001) The Lake Baikal drilling project in the context of a global lake drilling initiative. *Quaternary International* 80(81): 3–18.

Diatom Fossil Records from Marine Laminated Sediments

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Introduction

Historical Perspective

The study of seasonally laminated sediments from the geological record was brought to the fore by Gerard De Geer in the early twentieth century (De Geer, 1912). He introduced the term 'varve' to describe an annually repetitive sediment deposit within Swedish glacial marine clays. These late Quaternary varves comprise a silt-rich lamination, related to higher energy spring glacier melt, and a clay-rich lamination, related to lower energy in-wash and biological productivity in the summer. The description of these varves and the seasonal-scale proxy data that they contained, together with the realization that packages of varves could be correlated over tens of kilometers, demonstrated, for the first time, that an annual-scale chronology was possible and led to the first such chronology for the past 12 ka being erected in Scandinavia (De Geer, 1912). However, after De Geer's work, the study of laminated sediments, varves, and seasonal-scale proxy records was slow to progress. The geological community was skeptical about whether annual resolution was possible from the geological record. In the 1960s, Roger Anderson and his colleagues revived research into laminated sediment records (e.g., Anderson and Kirkland, 1960) and re-established the credibility of varves as proxy climate archives with the highest resolution.

In the marine realm, research was hampered until the 1970s by the technological challenge of recovering long sediment cores from the sea floor. However, early workers first recovered short cores of diatom-rich, marine laminated sediments from the Gulf of California, NW Mexico (Figure 1), and interpreted these as marine varves (Calvert, 1966, and references therein), relating the alternation between lighter and darker olive-green laminae to seasonally varying diatom flux to the sea floor. High sedimentation rates associated with laminated marine sediments (millimeters to centimeters per year), together with immature coring technologies, meant that the recovery of full Quaternary sequences, or even complete Holocene records, was not possible until the inception of the Deep Sea Drilling Project (DSDP), and the introduction of the Hydraulic Piston Corer (HPC). The HPC was first deployed during DSDP Leg 64 in 1978, in the Gulf of California, recovering 152 m (~250 ka) of marine laminated diatom ooze, interbedded with homogeneous diatom ooze, of late Pleistocene to Holocene age (Schrader et al., 1980; see Figure 2(a)). DSDP Site 480 represented the first complete Quaternary marine laminated sediment sequence recovered. The development of the HPC into the Advanced Piston Corer (APC) during the Ocean Drilling Program (ODP) and Integrated ODP (IODP) led to the recovery of long cores of laminated diatom ooze, including an almost 200 m long, ~160 ka sequence from Santa Barbara Basin, California margin (Behl and Kennett, 1996), a 118 m long sequence from Saanich Inlet, British Columbia, with approximately 60 m of

Holocene laminated diatom-rich sediments (Blais-Stevens et al., 2001) and, most recently, a ~200 m long Holocene sequence from the Adélie Basin, East Antarctic margin (Expedition 318 Scientists, 2010; Figure 1).

Advances in analytical techniques during the 1990s have realized the potential of laminated, diatom-rich marine sediments to provide seasonal-scale proxy ocean and climate records. Almost surgical precision in sampling has been introduced, including small, discrete samples for quantitative diatom abundance work (Barron et al., 2005; Leventer et al., 2002; Sancetta, 1995), and scanning electron microscope (SEM) analysis of intact sequences of laminations on the micron scale to investigate diatom species successions through the seasons and relationships between succeeding laminae (Kemp, 1990). The introduction of SEM-scale analysis to the study of laminated, diatom-rich sediments has unlocked Quaternary proxy paleo-ocean and -climate records that can now be compared with modern instrumental data retrieved from sediment traps and satellite remote sensing of the sea surface (Thunell et al., 1993). The seasonal variations in diatom flux and terrigenous input into the laminated sediments can often be extrapolated to the interpretation of longer time-scale variations through a sequence (Anderson, 1986), and hence laminations can provide analysis of variations from seasonal through to millennial timescales (Kemp, 2003).

Sites of Accumulation of Quaternary Laminated Diatom Ooze

There are two requirements for the formation of laminated diatom ooze in marine sediments: (1) a temporally varying flux of either diatoms or terrigenous grains, or both; and (2) a mechanism for preserving the primary sediment fabric. In the Gulf of California (Figure 1), a classic site of laminated diatom-rich sediment deposition, early workers proposed that either diatom flux varied through the year with terrigenous flux remaining constant, or terrigenous flux varied with biogenic flux remaining constant. The introduction of sediment traps to the study of lakes (e.g., Anderson, 1977), and then to the study of oceans (e.g., Honjo, 1982), made it possible to relate compositionally different fluxes with their seasons of production. This technological advance made it possible to address debates over which flux varied and which remained constant through the year and, in the case of the Gulf of California, it was discovered that both biogenic and terrigenous fluxes varied, with diatoms having two peak flux events per year in the spring and autumn/early winter (Thunell et al., 1993) related to the 'spring bloom' and 'fall dump' of diatoms, respectively (Kemp et al., 2000). The application of SEM-scale analytical techniques to Gulf of California sediments further revealed that the fossil diatom assemblage record reflected the seasonal variations in flux through the water column, with up

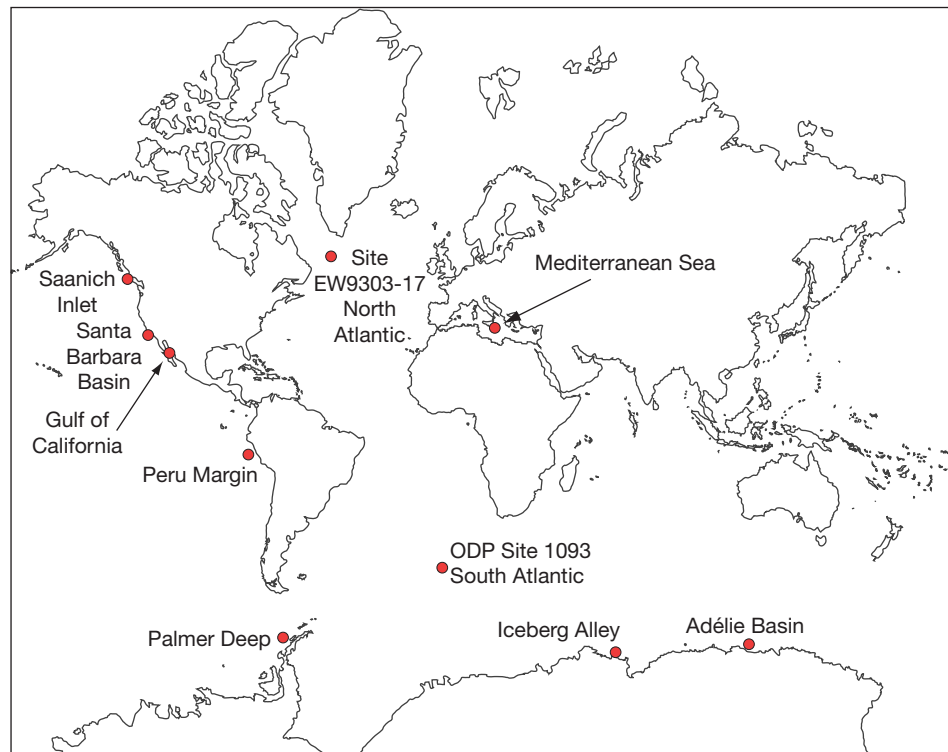


Figure 1 Schematic world map showing the location of sites of accumulation of Quaternary diatom-rich laminated sediments discussed in this article.

to five discrete diatom productivity events identified in sediments per year (Pike and Kemp, 1997).

Typical sites of accumulation of laminated diatom-rich sediments occur beneath areas of coastal upwelling. Seasonal variations in the strength or direction of alongshore winds influences the influx of nutrients into the euphotic zone from deeper water, hence the temporal abundance and species composition of the diatoms living in the surface waters. Such sites range from continental shelf silled basins (Figure 3(a)), such as Santa Barbara Basin (e.g., Schimmelmann and Lange, 1996), to open continental slope sites such as along the California and Peru margins (e.g., Anderson et al., 1987; Brodie and Kemp, 1994), to semienclosed seas such as the Gulf of California (Pike and Kemp, 1997). Mechanisms for the preservation of laminated sediments that accumulate in shelfal environments are related to levels of dissolved oxygen at the sea floor. Restricted water circulation in silled basins, such as Santa Barbara Basin (Figure 1), together with high surface water rain of organic material leads to the rapid consumption of dissolved oxygen below sill depth (concentrations below 0.2 ml l^{-1}), the restriction of an active macrofauna, and the preservation of the seasonally varying flux as primary laminations (e.g., Schimmelmann and Lange, 1996, and references therein). Along the continental slope, laminated sediments are preserved where the slope intersects the midwater depth oxygen minimum zone (OMZ; see Figure 3(b)). The oceanic OMZ is strengthened in regions of coastal upwelling due to bacterial degradation of the high flux of organic debris from the nutrient-rich surface waters. Where this OMZ intersects the continental slope, the macrofauna is restricted and variations in flux are preserved in the sediments (e.g., Anderson et al.,

1987). In semienclosed seas, the intersection of the basin slopes with the OMZ also will lead to the preservation of diatom-rich laminations.

During the Quaternary, laminated diatom-rich sediments have also been preserved in the deep-sea beneath areas of open ocean distant from the coasts. Typically, in the deep sea, levels of benthic dissolved oxygen are high enough to support an active, bioturbating benthic faunal community. When laminated diatom ooze is preserved, a mechanism other than prohibitively low levels of dissolved oxygen is required for the preservation of laminations. In the Quaternary, laminated diatom oozes have been described from the North Atlantic and Southern Ocean (Bodén and Backman, 1996; Grigorov et al., 2002). Both of these regions are associated with convergent oceanic frontal systems, which concentrate tangled mats of pennate diatoms and lead to a mass flux of preformed diatom mat laminations to the sea floor (Figure 3(c)), overwhelming the benthic community and preventing bioturbation (Kemp and Baldauf, 1993). In this case, a physical mechanism of diatom-rich lamina preservation is invoked as opposed to the chemical mechanism involved with the occurrence of low benthic dissolved oxygen concentrations.

Diatom fossil records from Quaternary laminated marine sediments can provide us with a wealth of the highest resolution proxy environmental data. This article describes the nature of these records over (i) seasonal and interannual timescales, relating to variations in species composition and abundance through the year, and (ii) orbital timescales, relating to variations in abundance and the presence or absence of the diatom-rich laminations as controlled by large-scale oceanographic and climatic processes.

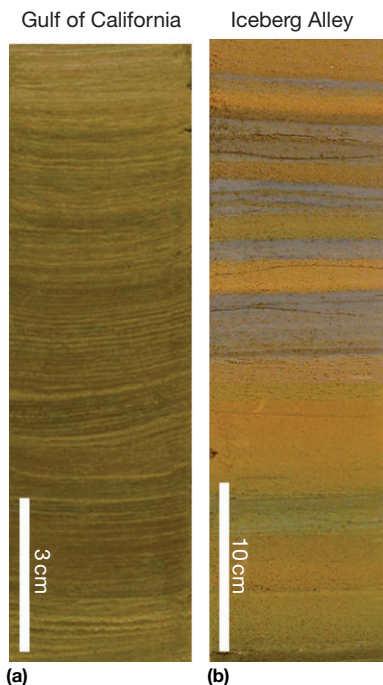


Figure 2 Examples of Quaternary diatom-rich varved marine sediments. (a) Guaymas Basin, Gulf of California (JPC56, 40–70 cm). (b) Iceberg Alley, Mac.Robertson Shelf, East Antarctica (NBP0101 JPC43B, 23.50–23.80 m).

Intra- to Interannual Diatom Productivity Events from Quaternary Laminated Sediments: Marine Varves

Enclosed and Semienclosed Seas

Diatoms that are highly sensitive to changes in environmental conditions, and laminated marine sediments that contain a well-preserved record of seasonal sediment flux, are a very powerful combination for reconstructing past environments. Investigation of Holocene laminated diatom ooze from enclosed and semienclosed seas, such as the Gulf of California and laminated Pleistocene sapropels from the Mediterranean Sea (Figure 1), highlighted the process of the fall dump (Kemp et al., 2000) as a major mechanism of diatom flux, silica and carbon export to the sediments. In the Gulf of California, sediment trap (Thunell et al., 1993) and fossil diatom assemblage analysis (Pike and Kemp, 1997) showed that there were two silica flux maxima per year (Figure 4(a)). In spring, coastal upwelling associated with strong NW winds promotes high primary productivity along the eastern coast of the gulf that results in a layer of *Hyalochaete Chaetoceros* spp. resting spores (CRS) in the sediments of the deep basins, for example, Guaymas Basin, where the basin slopes intersect the OMZ. In the fossil record of coastal oceans, CRS are typically associated with strong spring coastal upwelling. However, sediment traps in Guaymas Basin showed that the silica flux maxima for the year occurred in late autumn/early winter (Thunell et al., 1993), and the diatom fossil record associated with this event, as recorded by SEM-scale sediment fabric analysis, consisted of large centric diatoms such as *Coscinodiscus asteromphalus*, *Coscinodiscus oculus-iridis*, *Coscinodiscus granii*, *Rhizosolenia bergonii*, and *Stephanopyxis palmeriana*. Pike and Kemp (1997)

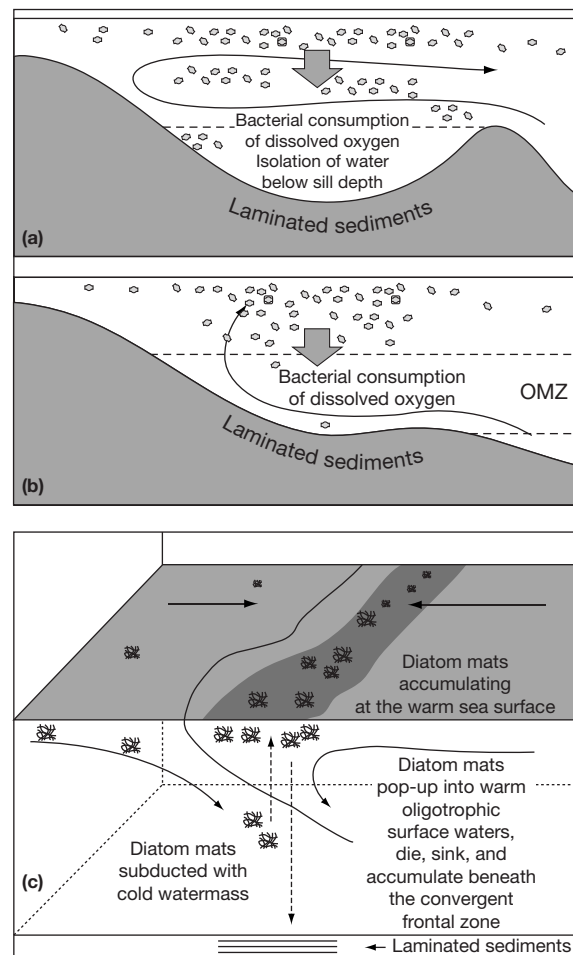


Figure 3 Schematic representation of different diatom-rich laminated sediment accumulation sites. (a) Silled basin, continental shelf. (b) Intersection of the oceanic oxygen minimum zones with the continental shelf. (c) Convergent oceanic frontal zone systems, open ocean settings.

concluded that these diatoms had been growing slowly at depth during the summer when there was a well-stratified water column and low intensity southerly winds. When the autumn/early winter storms mixed the water column as the winds shifted to become more northerly, these diatoms were deposited en masse, creating the opal flux maxima record in the sediment traps. A similar interpretation was reached for well-preserved Pleistocene sapropels (organic carbon-rich deposits) from the Mediterranean (Kemp et al., 1999). Here, diatoms such as *Rhizosolenia* spp. and *Hemiaulus hauckii*, which form long chains that tangle together to form ‘mats,’ grow during the summer season of strong water column stratification. These diatoms take advantage of nutrients trapped at depth in the stratified water column and then, by regulating their buoyancy, move up to the surface to photosynthesize (Kemp et al., 2000). Hence, the study of these low-latitude diatom fossil records from the Gulf of California and the Mediterranean raised awareness that high diatom export production need not be closely coupled to coastal or oceanic upwelling, as has been the classic interpretation, but could

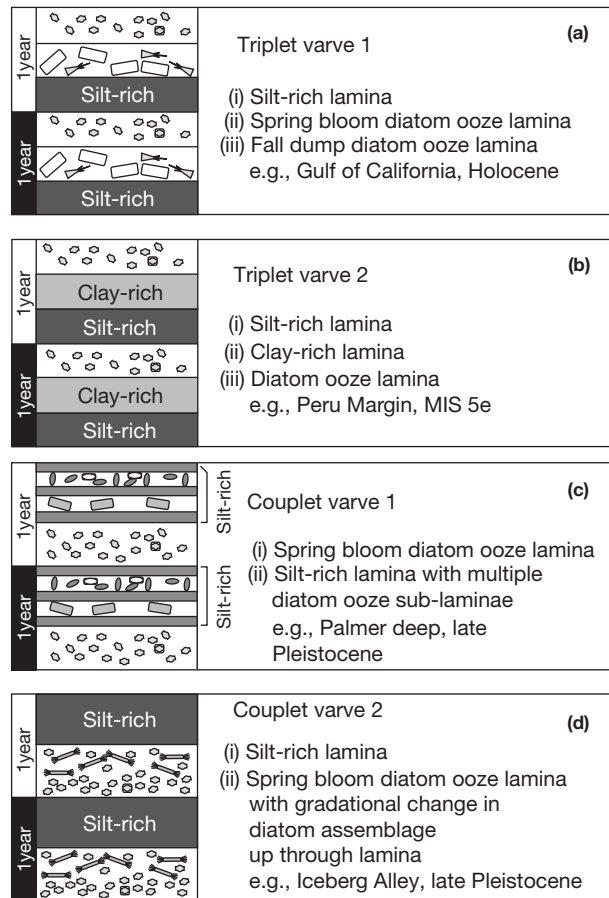


Figure 4 Schematic representation of the different types of marine varves. (a) Triplet varves comprising one terrigenous lamina and two diatom ooze laminae. (b) Triplet varves comprising two terrigenous laminae and one diatom ooze lamina. (c) Couplet varves comprising one diatom ooze lamina and one more terrigenous-rich lamina with the terrigenous lamina containing sublaminae of diatom ooze. (d) Couplet varves comprising one diatom ooze lamina and one terrigenous lamina with the diatom ooze lamina showing a gradational change in diatom assemblage from bottom to top.

also be the result of deposition following the breakdown of summer water column stratification and a fall dump of a diatom population that had been growing slowly over the summer in a deep chlorophyll maximum (DCM; Kemp et al., 2000).

In the Pleistocene Mediterranean Sea (Figure 1), the primary method of promulgating opal and organic carbon flux to the sediments, to form sapropels, was through the mat-forming diatoms *Rhizosolenia* spp. and *H. hauckii*. Although this was not the case for the Holocene Gulf of California laminated diatom-rich sediments, a typical mat-forming diatom was recorded here – *Thalassiothrix longissima*. Analysis of the stratigraphic position of *Thalassiothrix longissima* diatom mats within the Holocene varve led to the interpretation that *Thalassiothrix longissima* was transported into the gulf with subtropical surface waters during summer, but was deposited in the autumn along with the fall dump flora as wind speeds increased and wind direction changed to become more northerly (Pike and Kemp, 1997). Today, subtropical Pacific surface waters are advected far into the Gulf of California during El

Niño years, as the warm El Niño water mass flows northwards along the western edge of continental North America. Frequency analysis of the abundance of *Thalassiothrix longissima* diatom mats within an early Holocene section of laminations revealed three dominant frequencies: ~11 years, ~22–24 years, and ~50 years (Pike and Kemp, 1997). These frequencies are related to cycles in solar activity and, hence, the diatom record would suggest that in the early Holocene in the eastern Pacific Ocean, the frequency of El Niño, hence the import of *Thalassiothrix longissima* diatom mats into the gulf, was in part regulated by solar activity (Pike and Kemp, 1997).

Continental Shelf Basins

There is an extensive Quaternary laminated sediment fossil diatom record from silled continental shelf basins from both low and high latitudes. A well-known site for the accumulation of varved sediments is the Santa Barbara Basin on the California margin (Figure 1), a silled basin situated beneath a region of strong coastal upwelling that supports seasonally high primary production of diatoms. The Holocene record has been well documented (Schimmelmann and Lange, 1996 and references therein); however, the recovery of a long drill core, ODP Site 893, extended the record back into the late Pleistocene, over 160 ka (e.g., Bull et al., 2000). Laminations are preserved during warm interglacials or interstadials when bottom water ventilation of the basin is weak and bottom water dissolved oxygen concentrations are low (Behl and Kennett, 1996). The laminations form varves, which are either couplets of silt-rich and silt-poor laminae, or triplets of silt-rich, silt-poor, and diatom ooze laminae (Bull and Kemp, 1995). Lithogenic material is delivered primarily as terrestrial run-off during the winter rainy season whereas diatom flux occurs mainly during the spring and summer, related to the intensity of coastal upwelling over the basin. The diatom ooze laminae range in thickness from <20 µm to 2 mm, with a mean thickness of 0.25–0.75 mm, and are composed of vegetative remains and resting spores of *Hyalochaete Chaetoceros* spp., predominantly *Chaetoceros radicans*. Some thicker laminations show a succession of decreasing vegetative remains and increasing resting spores upward, with the very thickest laminae containing sublaminae of other genera (Bull and Kemp, 1995). The range in thickness of the diatom ooze laminae reflects increased or decreased diatom productivity, hence, efficiency of coastal upwelling, whereas the thickness of the terrigenous components is linked to the amount of rainfall and subsequent run-off from the continent. This decoupling of formation processes between diatom and terrigenous laminae means that in Santa Barbara Basin, the varves can be used to assess the concurrence of changes in marine productivity and terrestrial run-off in relation to oceanographic/climatic events such as El Niño (Bull et al., 2000). Bull et al. (2000) investigated the cyclicity and phase relationships between different laminae types from an interval of triplet laminations dated at 160.2 ka (marine isotope stage, MIS, 6.2). They showed that 3.1, 8.4, and 8.8 year cycles in terrigenous lamina thickness and total varve thickness could be related to similar frequencies in modern El Niño time series. During modern El Niño, there is increased rainfall over the California coast and, hence, increased terrestrial run-off onto the shelf. Cycles of 3.5, 7.3, and

7.6 years in the diatom laminations could also be related to El Niño and reflect reduced marine productivity associated with warmer waters expanding northward along the North American coast and reducing the effectiveness of upwelling to reach cold, nutrient-rich deeper waters (Bull et al., 2000). Further, the changes in thickness of the terrigenous laminae and diatom ooze laminae are antiphased which is expected because the impact of El Niño is to simultaneously increase run-off from land and decrease marine productivity.

In recent years a number of late Quaternary laminated diatom ooze deposits have been recovered from continental basin and trough sites around Antarctica (Barker et al., 1999; Expedition 318 Scientists, 2010; Leventer et al., 2006). These basins and troughs were sites of exceptionally high sedimentation, particularly during the rapid warming following the last glacial period. For example, Palmer Deep (Figure 1), a funnel-shaped depression on the western Antarctic Peninsula shelf which acts as a large natural sediment trap, had an average radiocarbon-derived linear sedimentation rate of $0.25 \text{ cm year}^{-1}$ during the last deglaciation (Domack et al., 2001; as recalibrated in Maddison et al., 2005). Despite this low sedimentation rate, individual varves can be in excess of 5 cm thick in the early deglaciation (Maddison et al., 2005). Iceberg Alley, a linear-shaped trough on the MacRobertson Shelf, East Antarctic margin (Figure 1), had a radiocarbon-derived linear sedimentation rate of 0.5 cm year^{-1} during a similar postglacial interval (Stickley et al., 2005), but with individual varves being over 8 cm in thickness (Figure 2(b)). Compared with average coastal high sedimentation rates measured in cm per thousand years, sedimentation rates of cm per year represent much higher biological productivity in the surface waters, and hence higher flux of diatoms to the sediments. Analysis of the seasonal sequence of diatom assemblages recorded in these sediments reveals evidence of dynamic changes around the coast of Antarctica during deglaciation. In the Palmer Deep postglacial laminated diatom ooze, thick laminations composed of CRS (see Figure 5(a) and 5(b)) are deposited following the spring bloom of vegetative *Hyalochaete Chaetoceros* spp. In Antarctic coastal waters, as opposed to coastal upwelling at low latitudes, *Chaetoceros* spp. (Hyalochaete) are typical indicators of spring waters at the sea ice edge. As the sea ice melts, a freshwater cap traps nutrients in the surface waters leading to a bloom of *Chaetoceros* spp. (Hyalochaete) diatoms. When the sea ice has completely melted and the nutrients in the water are depleted, CRS are deposited to form a thick lamination in the sediments (Maddison et al., 2005). In the summer, as the water column becomes increasingly oligotrophic, a more typical mixed Antarctic coastal assemblage takes over, comprising diatoms such as *Corethron pennatum*, *Coscinodiscus bouvet*, *Odontella weissflogii*, *Fragilariopsis* spp., and *Thalassiosira antarctica*. During the immediate postglacial times, the summer laminations are often characterized by a succession of sublaminae dominated by different species, related to pulses of upwelled circumpolar deep water replenishing nutrients on the shelf, or to advected water masses (Figure 4(c)). CRS sublaminae occur in earliest summer when the influx of new nutrients restimulates the waning spring *Chaetoceros* (Hyalochaete) population. *Corethron pennatum* sublaminae occur early in summer, perhaps deposited from a water column that is still partially stratified but with no sea ice

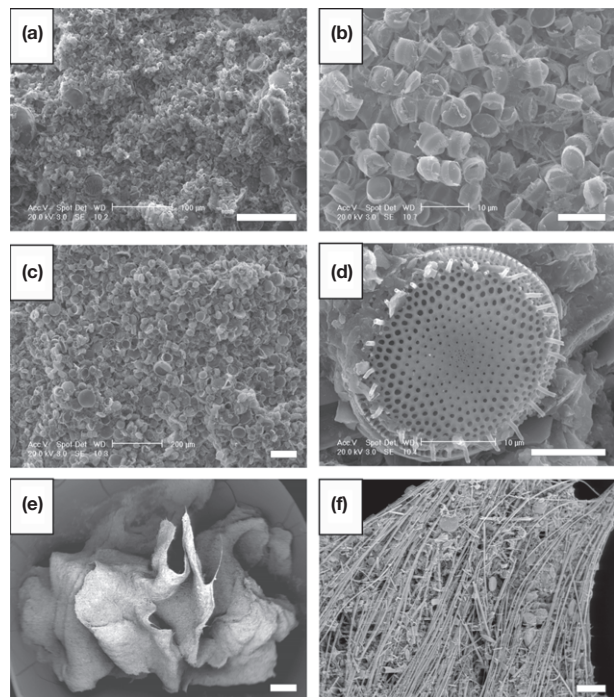


Figure 5 Scanning electron microscope secondary electron images of examples of Quaternary diatom laminations. (a) Low magnification image of an *Chaetoceros* (Hyalochaete) resting spore lamina-parallel sediment surface (ODP Hole 1098A, 44.1341–44.1261 mcd; scale bar = 100 μm). (b) High magnification image of a *Chaetoceros* (Hyalochaete) resting spore lamina (scale bar = 10 μm). (c) Low magnification image of a *Thalassiosira antarctica* resting spore lamina-parallel sediment surface (ODP Hole 1098A, 44.3965–44.3735 mcd; scale bar = 100 μm). (d) *Thalassiosira antarctica* resting spore (scale bar = 10 μm). (e) Low magnification image of a pennate diatom mat demonstrating the inherent strength of the mat that enables it to be sculpted (NBP0101 KC09, 255 cm; scale bar = 1 mm). (f) Higher magnification image of a pennate diatom mat showing the long thin frustules of *Thalassiothrix antarctica* that inhibit bioturbation (NBP0101 KC09, 255 cm; scale bar = 100 μm).

remaining. Sublaminae of *Coscinodiscus bouvet* are most likely deposited from advected water masses from further north along the peninsula, associated with summer storms. Near the top of the summer laminations, *Thalassiosira antarctica* sublaminae often occur as the last bloom of the year before light levels drop and sea ice reforms over the region (Maddison et al., 2005). It seems from this, and other laminated records recovered from around the Antarctic margin, that *Thalassiosira antarctica* is a good indicator in the sediment record for autumn diatom flux (Pike et al., 2009).

In Iceberg Alley, MacRobertson Shelf, the postglacial diatom record is similar to that of Palmer Deep; however, here the seasonal succession of diatoms is expressed within the thick 'spring' laminae, as opposed to the summer laminae in Palmer Deep (Figure 4(d)). The spring laminae are dominated by CRS as in Palmer Deep; however, near the top of the CRS laminae, the concentration of diatoms such as *Corethron pennatum* and *Rhizosolenia* spp. increases (Stickley et al., 2005). The early postglacial environment was characterized by *Corethron pennatum*, whereas the later postglacial was characterized by increased abundance of *Rhizosolenia* spp. reflecting the gradual move to

relative oligotrophy and less persistent sea ice when compared with the early postglacial. This increase in *Corethron pennatum* and *Rhizosolenia* spp. toward the top of the spring lamination suggests that as the *Chaetoceros* spp. (Hyalochaete) exhaust the nutrients in the surface waters the other, more oligotrophic, diatoms gradually take over until summer (i.e., the second lamination of the couplet), when the water column stratification eventually breaks down and the typical Antarctic mixed flora of diatoms takes over (Stickley et al., 2005). During the postglacial in Iceberg Alley, it is suggested that a calving bay reentrant existed (a fjord-like indentation in the ice-sheet edge that forms a open water bay surrounded by cliff of ice and, into which, icebergs calve), surrounding the trough and trapping nutrients and flux within the trough (Leventer et al., 2006), thus the extremely high sedimentation rates and diatom flux can be attributed to both high biological productivity resulting from high nutrients in the water column, and sediment focusing within the trough. The investigation of these laminated sediments from Iceberg Alley, together with others (laminated and nonlaminated) from the East Antarctic margin (Adélie Basin, Mertz Ninnis Trough and Svenner Channel) have led to the description of a new proxy for annual sea ice persistence and winter sea concentration (Pike et al., 2009). When the ratio of abundance of *Porosira glacialis* resting spores:*Thalassiosira antarctica* resting spores is less than 0.1, winters and springs are relatively warm, with annual sea ice duration of ~7.5 months per year and winter sea ice concentration of ~70%. When this ratio is greater than 0.1, winters and springs are relatively cool, with much greater than 7.5 months of annual sea ice cover and winter sea ice concentrations of ~80%. This relative abundance threshold between the two diatoms increases at the transition between the relatively warm Antarctic hypsithermal and the cool Neoglacial (transition at approximately 4000 cal year BP along the East Antarctic margin), indicating the sensitivity of this diatom proxy to recent climate change (Pike et al., 2009).

High-Latitude Fjords

Thickly laminated diatom-rich sediments also sometimes accumulate in fjords, for example, Effingham Inlet and Saanich Inlet (Figure 1), British Columbia, Canada (Blais-Stevens et al., 2001; Chang and Patterson, 2005). Saanich Inlet is a silled fjord with freshwater input at its head which results in weak flushing of the surface waters and isolation of the waters below sill depth. High surface water productivity leads to severely reduced benthic dissolved oxygen levels and the preservation of sedimentary laminations. ODP drilling in Saanich Inlet recovered ~60 m of Holocene diatom-rich laminated sediments. Centimeter-scale varves consist of couplets of diatom ooze or diatom mud deposited in spring and summer, with silty clay deposited in winter (Dean et al., 2001). Saanich Inlet varves may have up to 19 sublaminae composed of various diatom species, or assemblages. Following from the winter terrigenous lamina, up to two early spring sublaminae characterized by *Thalassiosira pacifica* or *Chaetoceros* spp. (Hyalochaete) may be deposited as light levels increase and water column stratification begins to be established. In spring, with a strong pycnocline (a zone of rapid increase in density within a layer of the ocean) and high nutrient levels in the surface waters, *Chaetoceros* spp. (Hyalochaete) bloom and a sublamina

is deposited, followed by a sublamina characterized by *Skeletonema costatum*. The summer portion of the varve is characterized by one to many sublaminae of *Chaetoceros* spp. (Hyalochaete). The number of summer sublaminae, or blooms in the water column, is related to a tidally controlled input of new nutrients and diatom seedstock into the fjord. As the water column stability is broken down in autumn, there may be a final bloom and subsequent sublamina of *Chaetoceros* spp. (Hyalochaete) deposited (Dean et al., 2001). Interannual variability in the varved sequence is revealed by presence or absence of *Thalassiosira pacifica* and/or *S. costatum* sublaminae. *Thalassiosira pacifica* is abundant when there is a greater input of freshwater into the fjord during El Niño events, whereas *S. costatum* tends to be more abundant during the cooler periods between El Niño events and negative phases of the Pacific–North American index (Dean et al., 2001).

Centennial- to Orbital-Scale Diatom Records from Quaternary Laminated Sediments

Quaternary laminated sediment sections not only record information about year-on-year variations in diatom assemblages which can be used to reconstruct the annual cycle of hydrographic conditions (Pike and Kemp, 1997), they also record variations over interannual through to millennial timescales (Leventer et al., 2010). In the Gulf of California, investigations of diatom species variations revealed long-term variations in the hydrography and climate of the gulf over the last 15 ka (Barron et al., 2005, and references therein). For example, *Azpeita nodulifera*, a tropical species, was abundant in the Guaymas Basin, central Gulf of California (Figure 1), during the Bølling/Allerød interstadial (14.6–12.9 ka) indicating that the seasonal incursion of warm, tropical waters from the Pacific to the south was strong. *A. nodulifera* was still common during the Younger Dryas stadial (12.9–11.6 ka), but it is associated with an overall reduction in diatom abundance suggestive of tropical, oligotrophic conditions extending throughout the year (i.e., seasonality was reduced; there were no laminations preserved during the Younger Dryas). Between ~11.0 and 6.2 ka, diatom abundance increased, but *A. nodulifera* almost disappears from the records. This indicates intensification in the strength or duration of northwesterly winter winds driving high productivity associated with coastal upwelling and effectively excluding Pacific-derived tropical forms from entering the gulf. The return of *A. nodulifera* to the record during the time period 6.2–5.4 ka indicates the return of southerly sourced tropical water into the gulf and the onset of modern monsoonal conditions with seasonally reversing winds resulting in coastal upwelling in the spring and incursions of warm, tropical waters in the summer. Hence, the diatoms preserved in the Gulf of California laminated sediments can be used to look at long-term, large-scale changes in hydrography and climate of the region.

Along the Peru margin diatom records from laminated sediments have been used to investigate the occurrence of ocean-climate events such as El Niño during earlier Quaternary climates (Brodie and Kemp, 1994). Laminated diatom-rich sediments have accumulated intermittently at ODP Site 686, West Pisco Basin (Figure 1), which is a continental shelf basin

and part of the rapidly subsiding forearc basin associated with the Peru–Chile Trench. Well-developed triplet laminations (Figure 4(b)) comprising an alternation of silt-rich and silt-poor terrigenous laminae with occasional near-monospecific *S. costatum* or, less often, *Hyalochaete* CRS diatom ooze laminations, accumulated during the last interglacial, MIS 5e (Brodie and Kemp, 1994). The triplet laminations represent marine varves whereby the silt-rich and silt-poor terrigenous laminae represent annual variations in continental run-off, and the intermittent diatom ooze laminations are interpreted as mega-blooms associated with anti-El Niño conditions, such as La Niña, when upwelling intensity increased along the Peru margin. Landward of the West Pisco Basin today, however, there is very little annual rainfall and run-off is intermittent, whereas to the north it is believed that only El Niño is capable of producing any significant surface run-off (Brodie and Kemp, 1994 and references therein). The evidence from the diatom-rich triplet laminations and other data such as from alkenone sea-surface temperature reconstructions (Brodie and Kemp, 1994 and references therein) lead to the conclusion that surface waters along the Peru margin were generally warmer during the last interglacial, which would have promoted a more regular rainfall pattern over the continent with more consistent annual run-off – hence the preservation of silt-rich and silt-poor laminations. As stated above, the intermittent *S. costatum* diatom ooze laminae represent deposition from extremely large productivity events likely to be associated with strong La Niña years (Brodie and Kemp, 1994).

Presence or Absence of Diatom-Rich Laminations: Oceanographic Frontal Zones

Accumulations of Quaternary diatom-rich laminations in the deep sea that cannot simply be ascribed a seasonal origin, are associated with the position of major convergent oceanic frontal systems (Figure 3(c)). The taxa associated with these laminated oozes are mat-forming diatoms from the genus *Thalassiothrix* (e.g., Bodén and Backman, 1996; Grigorov et al., 2002). The physical oceanographic mechanism responsible for the production of accumulations of laminated, diatom mat-rich sediments was first described for the Neogene sediments of the eastern equatorial Pacific (Kemp and Baldauf, 1993); however, Quaternary occurrences of diatom mat deposits have since been described and ascribed a similar mechanism. In the North Atlantic at Site EW9303-17 on the west flank of the Reykjanes Ridge (Figure 1), a 3-m-thick interval of laminated diatom ooze was described as being composed of monospecific *Thalassiothrix longissima* diatom mats (Bodén and Backman, 1996). Preservation of these diatom mats at the sea floor is not due to low sea-floor dissolved oxygen levels as the deep Atlantic is well oxygenated, but rather to the physical suppression of the benthic faunal burrowers by the impenetrable, mesh-like nature of the diatom mats themselves. (Figure 5(e) and 5(f) show examples from the Antarctic that demonstrate the strength of these pennate diatom mats.) This interval accumulated during MIS 5e, that is, the last interglacial. The accumulation of these diatom mats at this site, at this time, is related to the position of the subarctic convergence where the warm, salty North Atlantic and Irminger Currents meet cold, less saline Labrador Sea water. *Thalassiothrix*

longissima mats (colonies) carried in the colder water get subducted at the frontal zone to then ‘pop-up’ to the surface on the warmer more oligotrophic side of the front in order to photosynthesize. Here, they exhaust the already low nutrients, the diatoms die and sink to the sea floor to accumulate in a narrow band beneath the front. Oceanic frontal systems are often dynamic and migrate with changing climates, so the presence or absence of the laminated *Thalassiothrix longissima* diatom mat deposits can be used to track the frontal systems over time; for example, the subarctic convergence was likely positioned over site EW9303-17 during the last interglacial (Bodén and Backman, 1996) but was elsewhere, perhaps further to the south, during the ensuing glacial period.

In the South Atlantic, Pleistocene laminated diatom mat deposits have accumulated beneath the Polar Front at approximately 50° S, and were recovered from ODP Site 1093 (Figure 1), among others, during ODP Leg 177 (Grigorov et al., 2002). These Southern Ocean laminated diatom mat sediments are composed of *Thalassiothrix antarctica*. Similar to the North Atlantic, the laminated sediments appear to have been deposited during either interglacials (e.g., MIS 11 and 29) or the transitions into interglacials (e.g., MIS12/11 transition). In contrast to the diatom record from the North Atlantic, the laminated diatom ooze of the South Atlantic appears to contain a regular succession of diatom species in couplets, for example, *Thalassiothrix antarctica* diatom mats followed by a mixed centric/pennate diatom lamina in the MIS 12/11 interval. In the older interglacial (MIS 29), *Thalassiothrix antarctica* is followed by either by a mixed centric/pennate diatom lamina or *Fragilariopsis kerguelensis* lamina, and then centric diatom lamina. These successions imply that the laminated diatom mat deposits from the Polar Front region of the Southern Ocean could be recording some seasonal variations as well as the physical position of the Polar Front itself (Grigorov et al., 2002).

Summary

Diatom fossil records from Quaternary laminated marine sediments reveal evidence for climate and ocean variability over subannual, interannual, millennial, and orbital timescales. Seasonal diatom species or assemblage successions can reveal information about the nature of the prevailing water column process producing the silica flux, for example, upwelling or breakdown of water column stratification. Interannual variations in the occurrence of particular species, or in the make up of the whole diatom assemblage can provide a valuable indication of the behavior of phenomena such as the El Niño–Southern Oscillation during past climate states. In locations where long, late Quaternary laminated sediments sequences are recovered, the changes in the diatom ooze lamina assemblage over centennial to millennial timescales can reveal changes in local and region-scale ocean circulation and climate. Laminated diatom ooze also accumulates in the open ocean in association with convergent oceanic frontal systems. Here the laminations are composed of tangled diatom mats and are not necessarily seasonally deposited, however, the presence or absence of the laminated diatom mat deposits can be used to track the movement of the oceanic fronts over glacial–interglacial timescales.

See also: [Diatom Introduction](#). [Diatom Records: Antarctic Waters](#). [Paleoceanography](#), [Biological Proxies: Marine Diatoms](#).

References

- Anderson RY (1977) Short term sedimentation response in lakes in western United States as measured by automated sampling. *Limnology and Oceanography* 22: 423–433.
- Anderson RY (1986) The varve microcosm: Propagator of cyclic bedding. *Paleoceanography* 1: 373–382.
- Anderson RY, Hemphill-Haley E, and Gardner JV (1987) Persistent late Holocene seasonal upwelling and varves of the coast of California. *Quaternary Research* 28: 307–313.
- Anderson RY and Kirkland DW (1960) Origin, varves, and cycles of Jurassic Todiito Formation, New Mexico. *Association of American Petroleum Geologists Bulletin* 44: 37–52.
- Barker PF, Camerlenghi A, Acton GD, et al. (1999) *Proceedings of ODP, Initial Reports*, 178 (CD-ROM). Available from Ocean Drilling Program, Texas A&M University, College Station, TX 77845–9547.
- Barron JA, Bukry D, and Dean WE (2005) Paleoceanographic history of the Guaymas Basin, Gulf of California, during the past 15,000 years based on diatoms, silicoflagellates, and biogenic sediments. *Marine Micropaleontology* 56: 81–102.
- Behl RJ and Kennett JP (1996) Brief interstadial events in the Santa Barbara basin, NE Pacific, during the past 60 kyr. *Nature* 379: 243–246.
- Blais-Stevens A, Bornhold BD, Kemp AES, Dean JM, and Vaan AA (2001) Overview of late Quaternary stratigraphy in Saanich Inlet, British Columbia: Results of Ocean Drilling Program Leg 169S. *Marine Geology* 174: 3–26.
- Bodén P and Backman J (1996) A laminated sediment sequence from northern North Atlantic Ocean and its climatic record. *Geology* 24: 507–510.
- Brodie I and Kemp AES (1994) Variation in biogenic and detrital fluxes and formation of laminae in late Quaternary sediments from the Peruvian coastal upwelling zone. *Marine Geology* 116: 385–398.
- Bull D and Kemp AES (1995) Composition and origins of laminae in late Quaternary and Holocene sediments from the Santa Barbara Basin. In: Kennett JP, Baldauf JG, and Lyle M (eds.) *Proceedings of ODP, Scientific Results*, vol. 146 (Part 2), pp. 77–87. College Station, TX: Ocean Drilling Program.
- Bull D, Kemp AES, and Weedon GP (2000) A 160-k.y.-old-record of El Niño southern Oscillation in marine production and coastal run-off from Santa Barbara Basin, California, USA. *Geology* 28: 1007–1010.
- Calvert SE (1966) Origin of diatom-rich varved sediments from the Gulf of California. *Journal of Geology* 74: 546–565.
- Chang AS and Patterson RT (2005) Climate shift at 4400 years BP: Evidence from high-resolution diatom stratigraphy, Effingham Inlet, British Columbia, Canada. *Palaeogeography, Palaeoclimatology, Palaeoecology* 226: 72–92.
- De Geer G (1912) A Geochronology of the last 12,000 years. *Comptes Rendus du XI Congrès Géologique International* (Stockholm 1910), pp. 241–253.
- Dean JM, Kemp AES, and Pearce RB (2001) Palaeo-flux records from electron microscope studies of Holocene laminated sediments, Saanich Inlet, British Columbia. *Marine Geology* 174: 139–158.
- Domack E, et al. (2001) Chronology of the Palmer Deep site, Antarctic Peninsula: A Holocene paleoenvironmental reference for the circum-Antarctic. *The Holocene* 11: 1–9.
- Expedition 318 Scientists (2010) Wilkes Land Glacial History: Cenozoic East Antarctic Ice Sheet Evolution from Wilkes Land Margin Sediments. *IODP Preliminary Report* 318: doi:10.2204/iodp.pr.318.2010.
- Grigorov I, Pearce RB, and Kemp AES (2002) Southern Ocean laminated diatom ooze: Mat deposits and potential for palaeo-flux studies, ODP leg 177, Site 1093. *Deep-Sea Research II* 49: 3391–3407.
- Honjo S (1982) Seasonality and interaction of biogenic and lithogenic particulate flux at the Panama Basin. *Science* 218: 883–884.
- Kemp AES (1990) Sedimentary fabrics and variation in lamination style in Peru continental margin upwelling sediments. In: Suess E and von Heune R (eds.) *Proceedings of ODP, Scientific Results*, vol. 112, pp. 43–58. College Station, TX: Ocean Drilling Program.
- Kemp AES (2003) Evidence for abrupt climate changes in annually laminated marine sediments. *Philosophical Transactions of the Royal Society, London A* 361: 1851–1870.
- Kemp AES and Baldauf JG (1993) Vast Neogene laminated diatom mat deposits from the eastern equatorial Pacific Ocean. *Nature* 362: 141–144.
- Kemp AES, Pearce RB, Koizumi I, Pike J, and Rance SJ (1999) The role of mat-forming diatoms in formation of the Mediterranean sapropels. *Nature* 398: 57–61.
- Kemp AES, Pike J, Pearce RB, and Lange CB (2000) The 'Fall dump' — A new perspective on the role of a 'shade flora' in the annual cycle of diatom production and export flux. *Deep-Sea Research II* 47: 2129–2154.
- Leventer A, Crosta X, and Pike J (2010) Holocene marine diatom records of environmental change. In: Smol JP and Stoermer EF (eds.) *The Diatoms: Applications for the Environmental and Earth Sciences*, pp. 400–422. Cambridge: Cambridge University Press.
- Leventer A, Domack E, Barkoukis A, McAndrews B, and Murray J (2002) Laminations from the Palmer Deep: A diatom-based interpretation. *Paleoceanography* 17: PAL 3 1–15.
- Leventer A, et al. (2006) Marine sediment record from the East Antarctic margin reveals dynamics of ice sheet recession. *GSA Today* 16(12): 4–10. <http://dx.doi.org/10.1130/GSAT01612A.1>.
- Maddison EJ, Pike J, Leventer A, and Domack EW (2005) Deglacial seasonal and sub-seasonal diatom record from Palmer Deep, Antarctica. *Journal of Quaternary Science* 20: 435–446.
- Pike J and Kemp AES (1997) Early Holocene decadal-scale ocean variability recorded in Gulf of California laminated sediments. *Paleoceanography* 12: 227–238.
- Pike J, et al. (2009) Observations on the relationship between the Antarctic coastal diatoms *Thalassiosira antarctica* Comber and *Porosira glacialis* (Grunow) Jørgensen and sea ice concentrations during the late Quaternary. *Marine Micropaleontology* 73: 14–25.
- Sancetta C (1995) Diatoms in the Gulf of California: Seasonal flux patterns and the sediment record for the past 15,000 years. *Paleoceanography* 10: 67–84.
- Schimmelmann A and Lange CB (1996) Tales of 1001 varves: A review of Santa Barbara Basin sediment studies. In: Kemp AES (ed.) *Palaeoclimatology and Palaeoceanography from Laminated Sediments*, pp. 121–141. London: Geological Society of London Special Publication 116.
- Schrader H, et al. (1980) Laminated diatomaceous sediments from the Guaymas Basin slope (central Gulf of California): 250,000-year climate record. *Science* 207: 1207–1209.
- Stickley CE, et al. (2005) Deglacial ocean and climate seasonality in laminated diatom sediments, Mac.Robertson Shelf, Antarctica. *Palaeogeography, Palaeoclimatology, Palaeoecology* 227: 290–310.
- Thunell R, Pride C, Tappa E, and Muller-Karger F (1993) Varve formation in the Gulf of California: Insights from time series sediment trap sampling and remote sensing. *Quaternary Science Reviews* 12: 451–464.

North Atlantic and Arctic

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Introduction

Marine planktonic diatoms are single-celled, photosynthetic, siliceous-walled algae. In the oceans, diatoms are restricted to cold, nutrient-rich regions where silicic acid is not limiting, such as in the polar regions, the coastal and equatorial upwelling systems, and in coastal areas. The most important factors that determine the distribution and abundance of diatoms in surface waters are sea-surface temperatures (SSTs), sea-ice conditions, macro- and micronutrient levels, stability of the surface water layer, light levels, and grazing. Surface-water productivity and dissolution in the water column and at the water-sediment interface are major processes dictating the diatom flux to the seafloor and preservation of diatom valves in sediments.

Preservation of diatoms in sediments of the North Atlantic Ocean through geologic time has been highly variable. The North Atlantic Ocean and the Nordic Seas were regions of high biosiliceous productivity until the intensification of Northern Hemisphere glaciations during the late Miocene and Pliocene. Diatoms are one of the most abundant microfossil groups preserved in sediments of Paleogene and early to mid-Neogene age in this region, organic-walled palynomorphs are also abundant, whereas calcareous microfossils are less abundant. In the central Arctic, diatoms were prolific in the late Cretaceous (Davies et al., 2009) and middle Eocene (Stickley et al., 2008, 2009; Suto et al., 2009). Since diatom species have undergone rapid evolution through the Cenozoic, it has been possible to establish a high-resolution biostratigraphy for the area. Based on the Deep Sea Drilling Project (DSDP) Legs 81 and 94, Baldauf (1984, 1987) established a Cenozoic diatom biostratigraphy for the middle- to high-latitude North Atlantic Ocean. Most of the fossil diatom species of the Nordic Seas and central Arctic are, however, endemic to these areas. Therefore, a separate diatom biostratigraphy has been developed for the Nordic Seas, and is underway for the central Arctic based on the DSDP Leg 38 material. Schrader and Fenner (1976) and Dzinoridze et al. (1978) proposed diatom biostratigraphies for the Nordic Seas. Meanwhile, development of drilling techniques and availability of reliable paleomagnetic stratigraphy enabled the development of a new, high-resolution Neogene diatom biostratigraphy for the Nordic Seas based on the Ocean Drilling Program (ODP) Leg 151 material (Koç and Scherer, 1996). The late Paleogene diatom biostratigraphy of Scherer and Koç (1996) from ODP Leg 151 material vastly improved and built on earlier works but calibration of these diatom datums to the geomagnetic polarity time scale (Gradstein et al., 2004) has yet to be achieved. Currently, Paleogene dinoflagellate cyst stratigraphy is better defined than that for diatoms in the Nordic seas (e.g.,

Eldrett et al., 2004); and future integrated stratigraphies are anticipated.

Until 2004, little information existed on central Arctic pre-Quaternary diatoms. What was known was limited to a few short piston cores from the Alpha Ridge such as the USGS cores FL-437 (late Cretaceous) and FL-422 (middle Eocene; Dell'Agnese and Clark, 1994) and the Canadian CESAR 6 core (late Cretaceous; Barron et al., 1985). Recently, Davies et al. (2009) made significant advances in our understanding of pre-Cenozoic Arctic diatom ecosystems by describing late Cretaceous seasonality from diatomaceous laminations within the CESAR 6 core. However, in 2004, Integrated ODP (IODP) Expedition 302, 'The Arctic Coring Expedition' (ACEX) unearthed arguably the most significant find of Paleogene siliceous microfossils in nearly 2 decades. Hundred meters of early middle Eocene, organic-rich, finely laminated sediments were recovered that contain abundant marine, sea ice, and freshwater siliceous microfossils (Onodera et al., 2008; Stickley et al., 2008, 2009; Suto et al., 2009) indicating high production and allowing intriguing insights into central Arctic paleoenvironments during the start of Cenozoic cooling. Largely endemic diatoms including the oldest known sea-ice diatoms (Stickley et al., 2009) other diatoms, ebridians, and silicoflagellates were preserved, along with very high abundances of chrysophyte cysts, the endogenously formed resting stage of freshwater algae. Continuing work on the siliceous microfossils of the ACEX cores, including examination of the laminations (cf. Stickley et al., 2008), is helping us understand the consequences of Cenozoic cooling and the role the Arctic played in important climate feedbacks.

Pleistocene Diatom Record

Biostratigraphy

Pleistocene sediments of the Nordic Seas are mostly barren of diatoms. In contrast to this, sites south of Iceland such as ODP Site 919 from the Irminger Basin and Site 983 from the Gardar Drift are sites with high Pleistocene accumulation rates and well preserved, abundant diatoms (Figure 1). Based on these sites, the resolution of the high-latitude North Atlantic Pleistocene diatom biostratigraphy established by Baldauf (1987) has been refined (Koç and Flower, 1998; Koç et al., 1999, 2001). Eight Pleistocene diatom datum events are identified and, for the first time, tied directly to the oxygen isotope record and paleomagnetic stratigraphy at these sites (Figure 2). These datum events are:

- the last occurrence (LO) of *Proboscia curvirostris* at 0.3 Ma,
- the LO of *Thalassiosira jouseae* at 0.3 Ma,

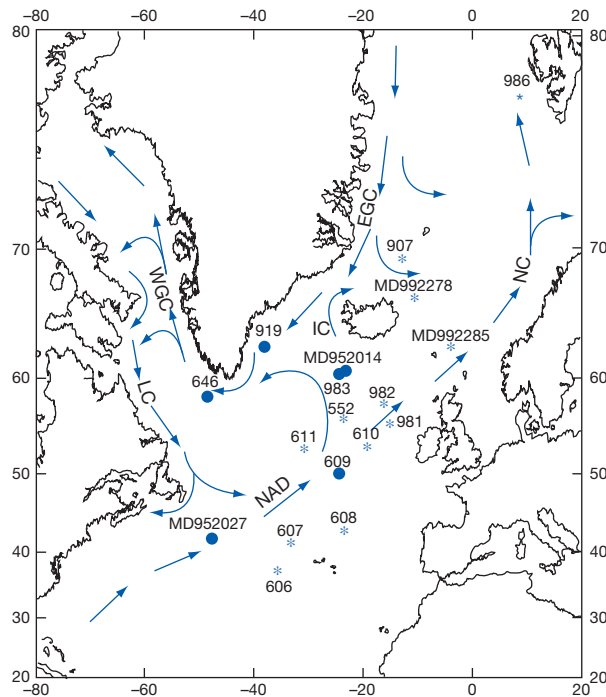


Figure 1 General surface circulation of the northern North Atlantic Ocean and the location of the cores cited for Pleistocene diatom records. In cores marked with dots, the last occurrence of *Proboscia curvirostris* is correlated to the oxygen isotope stratigraphy of the core. Stars depict sites where *P. curvirostris* has been observed. MD depicts cores taken by the French RV Marion Dufresne during the IMAGES cruises. All other cores are Deep Sea Drilling Project (DSDP) or Ocean Drilling Program (ODP) cores. Surface currents are illustrated with arrows as follows: NAD, North Atlantic Drift; IC, Irminger Current; NC, Norwegian Current; EGC, East Greenland Current; WGC, West Greenland Current; LC, Labrador Current (Koç et al., 2001).

- the LO of *Nitzschia reinholdii* at 0.6 Ma,
- the LO of *Nitzschia fossilis* at 0.68 Ma,
- the LO of *Neodenticula seminae* at 0.84 Ma,
- the first occurrence (FO) of *Neodenticula seminae* at 1.25 Ma,
- the FO of *Proboscia curvirostris* at 1.53 Ma, and
- the FO of *Pseudoeunotia doliolus* at 1.89 Ma.

Most of these datums are isochronous (i.e., occur at the same time) across the middle–high latitudes of the North Atlantic and the North Pacific. Based on these datums four high-latitude North Atlantic diatom zones have been proposed for the Pleistocene: the *Nitzschia reinholdii* zone, the *Neodenticula seminae* zone, the *Proboscia curvirostris* zone, and the *Thalassiosira* (*Shionodiscus*) *oestrupii* zone (Koç et al., 1999).

Paleoceanography

Fluctuating Pleistocene climatic conditions had a major impact on diatom productivity in the high-latitude North Atlantic. Diatom abundance and preservation at Site 919 and 983 covary with the oxygen isotope record of the sites in response to changing oceanographic conditions affecting diatom productivity (Koç and Flower, 1998; Koç et al., 1999; Figure 2). Generally, interglacials relate to high diatom productivity, whereas

sediments representing glacial stages are often barren of diatoms. However, during glacial stages 18, 20, and 30, there was significant diatom production and preservation.

During the late Quaternary glacial periods, Northern Hemisphere oceanic polar fronts and sea-ice migrated southward, with some of the most dramatic changes occurring in the mid- to high-latitude North Atlantic (CLIMAP Project Members, 1981). Reduced surface-water productivity during glacial periods indicates that these were the intervals during which the Polar Front migrated southeastward, thereby covering Sites 919 and 983 with sea ice. Considering the modern proximity of the sea-ice margin to the area, it is highly probable that these diatom-barren glacial intervals reflect periods when these sites were covered by year-round sea ice. On the other hand, the presence of significant diatom production during glacial intervals associated with marine isotope stages (MIS) 18, 20, and 30 indicates open marine conditions at sites south of Iceland during these times. These glacial stages fall within the first 100 ka cycles following the mid-Pleistocene transition, and the benthic oxygen isotope records indicate that these glaciations were not as severe as the late Quaternary glacial periods.

Mid-Pleistocene transition

Neodenticula seminae is part of the modern diatom assemblage of the middle- and high-latitude North Pacific. In the North Atlantic, this species is limited to the time interval of 0.84–1.26 Ma (Koç et al., 1999; Figure 3), but has recently been found in plankton samples from the North Atlantic – possibly a consequence of diminishing sea ice in the modern Arctic facilitating open water exchange between the Pacific and Atlantic (Reid et al., 2007). However, Baldauf (1986) interpreted the occurrence of *Neodenticula seminae* in North Atlantic Quaternary sediments as evidence for the presence of cool, low-salinity surface waters in the central North Atlantic during the early Quaternary. The interval of *Neodenticula seminae* in the North Atlantic straddles the transition from the dominance of 41 ka cycles in the climate records to the dominance of 100 ka cycles. It is, therefore, possible to take the FO of *Neodenticula seminae* in the North Atlantic as an indicator for cooling, which started 1.26 Ma, leading to the establishment of the 100 ka cycles and intensified glaciation. The presence of *Neodenticula seminae* in North Atlantic sediments is, therefore, attributable to the unique conditions related to the mid-Pleistocene transition. Lessons from modern studies (e.g., Reid et al., 2007) should, however, also be considered.

LO of *Proboscia curvirostris*

The youngest diatom event observed in North Atlantic sediments is the LO of *Proboscia curvirostris* (Jousé) Jordan and Priddle (Koç et al., 2001). *Proboscia curvirostris* is a robust species that can easily be identified in fossil assemblages, and therefore is a practical biostratigraphic tool. Its distribution stretches from 40°N to 80°N in the North Atlantic (Figure 1).

The extinction of *Proboscia curvirostris* is latitudinally diachronous through MIS 9–8 within the North Atlantic (Figure 4); a phenomenon closely related to the paleoceanographic development of the area. *Proboscia curvirostris* first disappeared within interglacial MIS 9 (324 ka) from the northern

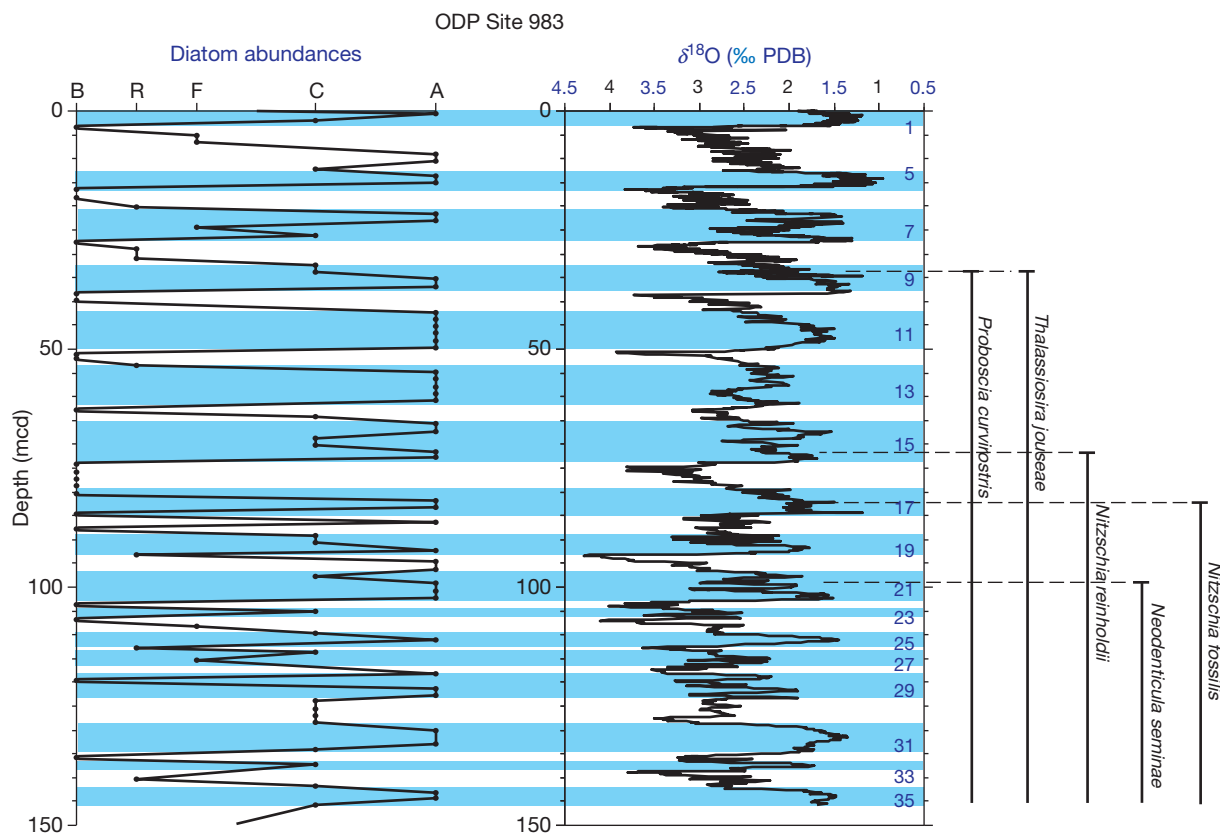


Figure 2 ODP Site 983 relative diatom abundances (B, barren; R, rare; F, few; C, common; A, abundant) plotted together with the planktonic $\delta^{18}\text{O}$ record against composite depth. Interglacial stages are shaded and numbered. Diatom biostratigraphic events and their ranges are indicated, and correlated to the $\delta^{18}\text{O}$ record (Koc et al., 1999).

areas that are most sensitive to climatic forcing, such as the East Greenland Current (EGC) and the sea-ice margin. It survived in the mid-latitudes North Atlantic until the conditions of the MIS 8 glaciation became too severe (260 ka). In the North Pacific at ODP Site 883, the LO of *Proboscia curvirostris* also falls within MIS 8. This apparent correlation between the North Atlantic and the North Pacific strongly suggests that the extinction of *Proboscia curvirostris* is isochronous between these oceans. The LO of *Proboscia curvirostris* is therefore potentially a highly valuable datum for correlation between the high-latitude North Atlantic and the North Pacific Oceans.

Last glacial period and Heinrich layers

Diatoms are absent in the sediments from the last glacial period in the Nordic Seas. They are also found very infrequently in the sediments of the northern North Atlantic, probably due to extensive sea-ice cover inhibiting diatom production. In contrast, diatom records from the Atlantic upwelling sites reveal maximum productivity during the Last Glacial Maximum (LGM; Abrantes, 2000).

The last glacial period is perturbed by millennial-scale climate variability in the form of Heinrich events (HE) and Dansgaard-Oeschger cycles and there has been a debate on the degree to which massive iceberg discharges associated with HE affected diatom productivity. Sancetta (1992) argued for increased diatom production in the North Atlantic linked

to HE, suggesting that the presence of numerous icebergs would have supported high production through physical (turbulent mixing) and/or biogeochemical (nutrient input) mechanisms. However, a recent detailed study on HE1 and HE4 shows the opposite effect, namely the reduction of diatom production in relation to massive iceberg discharges (Nave et al., 2007). Heinrich layers are deposited mainly between 40°N and 55°N , within the so-called, Ruddiman belt (Ruddiman, 1977). North of the Ruddiman belt during HE4, iceberg discharges had a strong impact on productivity, resulting in a drastic decrease, whereas during HE1, productivity showed a progressive increase through the deglaciation (Nave et al., 2007). Within the Ruddiman belt, massive iceberg discharges had a major effect on productivity during both HE4 and HE1. The possibility that diatom concentrations in ocean sediments had been diluted by ice-rafted detritus was tested by normalization to a constant ^{230}Th flux, and this showed that the enhancement of particle rain rate does not explain the decrease of diatom flux (Nave et al., 2007). The hypothesis that increased nutrient supply from southern source waters would cause an increase in productivity south of the area of maximum iceberg influence was also tested, and no indication for a productivity increase during HE1 and HE4 south of the Ruddiman belt was found. In summary, the massive iceberg discharges associated with HE1 and HE4 had strong impacts on surface ocean productivity, causing its drastic decrease.

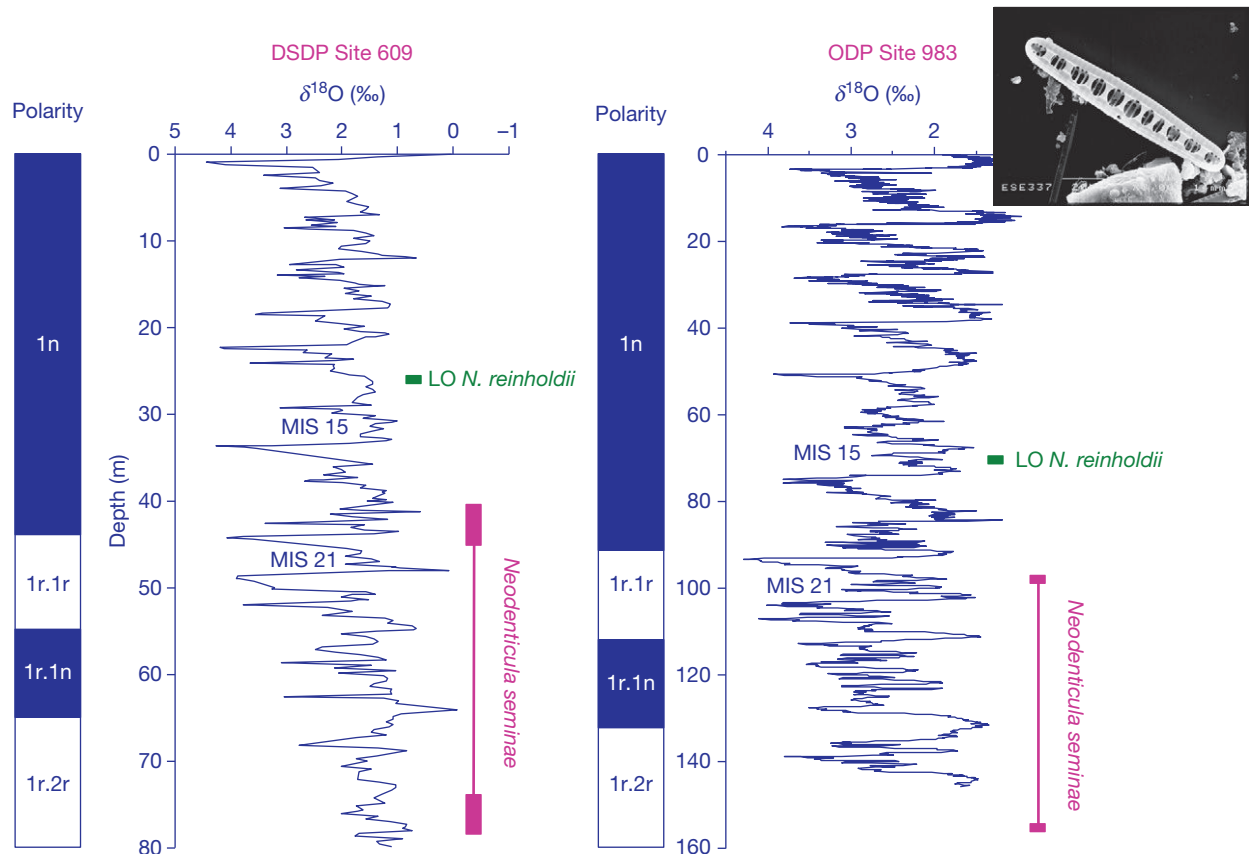


Figure 3 Planktonic $\delta^{18}\text{O}$ records of DSDP Site 609 (L. Labeyrie, unpublished data) and ODP Site 983 (Hodell and Kleiven, unpublished data) shown together with their respective paleomagnetic stratigraphies (Koç et al., 1999). In order to compare the timing of events between the middle- and high-latitude North Atlantic, the LO of *Nitzschia reinholdii*, and the last and first occurrence of *Neodenticula seminiae* are plotted at both sites. MIS, marine isotope stage. Inset, scanning electron microscope (SEM) picture of *N. seminiae* (Simonsen and Kanaya) Akiba and Yanagisawa from sample 162-983A-13H-7, 49–50 cm (from plate 1 in Koç et al., 1999).

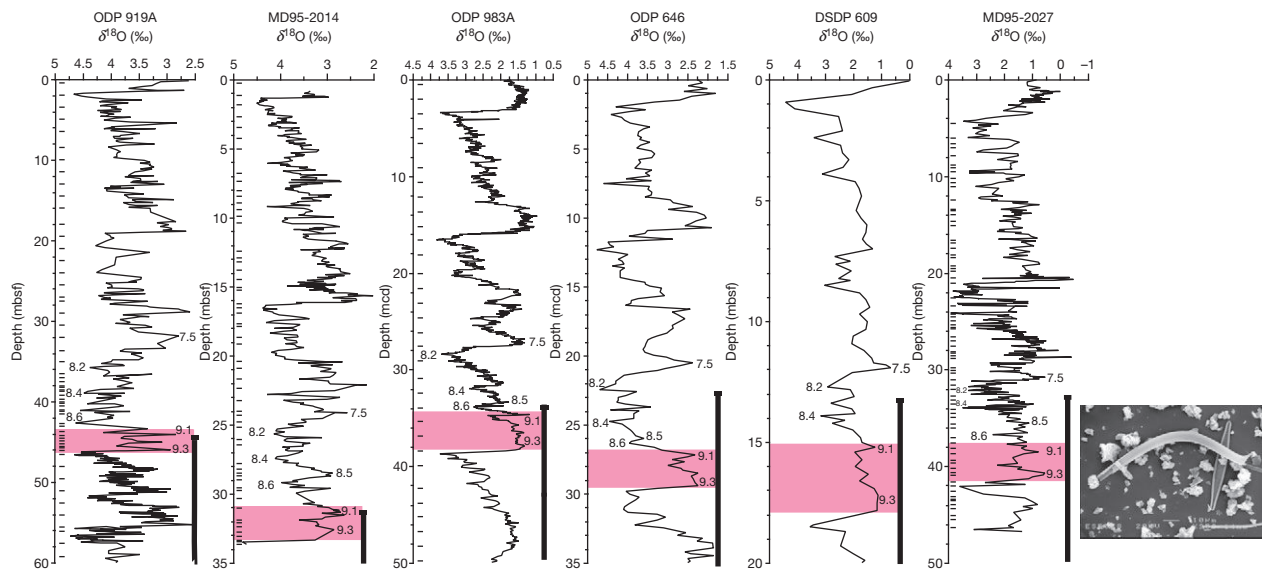


Figure 4 The LO of *Proboscia curvirostris* plotted on the oxygen isotope stratigraphy of each core (Koç et al., 2001). MIS 9 is shaded, and the substages numbered according to Prell et al. (1986). Sampling intervals are plotted as bars on the left side of the records. Scanning electron microscope (SEM) photograph of *P. curvirostris* Jordan and Priddle and *Nitzschia fossilis* (Frenguelli) Kanaya on the right side, Sample 162-983A-9H-3, 49–50 cm (from plate 2 in Koç et al., 2001).

Diatoms in Surface Sediments

To gain knowledge about present-day diatom ecology and biogeography, the distribution of diatom species and/or assemblages in the surface sediments of the North Atlantic Ocean and the Arctic Ocean have been mapped, and the resulting information has been applied to fossil records to reconstruct past ocean conditions. The high diversity of diatoms in polar and arctic environments makes them one of the foremost tools available for paleoclimatic reconstructions of the sea-ice margin, oceanic fronts, distribution of water masses and quantitative estimates of, for example, past SSTs. Several types of transfer functions, based on different statistical techniques, have been applied to quantify the relationship between the assemblages and the physical parameters. The most common ones are the Imbrie and Kipp technique (Imbrie and Kipp, 1971), the modern analog technique (Hutson, 1980), weighted averaging partial least square (WA-PLS; ter Braak and Juggins, 1993), maximum likelihood, and the artificial neural network approach (Malmgren and Nordlund, 1997).

North Atlantic and the Nordic Seas

Diatoms have proved to be an excellent paleoclimatic tool in the Nordic Seas and the northern North Atlantic. Koç and Schrader (1990) mapped the modern distribution of diatoms in this area based on 87 surface sediment samples and demonstrated that diatoms show good analog relations to modern oceanic conditions and are linearly related to SST. Recently, Andersen et al. (2004a) expanded this calibration set to 139 surface samples and refined the SST equation.

Q-mode factor analysis was applied to this set which defined eight different factors (Figure 5(a)–5(h)). The Arctic Greenland Assemblage (factor 1) with *Thalassiosira angustilineata* and *Thalassiosira trifurcata* mirrors Arctic water masses and has its highest factor loadings in the Greenland Sea. The North Atlantic Assemblage (factor 2) with *Shionodiscus* (*Thalassiosira*) *oestrupii* reflects the influence of the warm North Atlantic Drift. The sub-Arctic Assemblage (factor 3) with *Rhizosolenia hebetata* f. *semispina* and *Rhizosolenia borealis* has its highest loadings in mixed areas between Arctic and Atlantic waters, both northeast and southwest of Iceland. The Norwegian Atlantic Current (NwAC) Assemblage (factor 4) with *Thalassionema nitzschioides*, *Proboscia alata*, and *Thalassiosira angulata* can be used as an indicator of the NwAC. The Sea Ice Assemblage (factor 5) with *Fragilariopsis oceanica*, *Fragilariopsis cylindrus*, and *Thalassiosira hyalina* corresponds to the spring sea-ice limit. The Arctic Assemblage (factor 6) with *Thalassiosira gravida* spores and vegetative cells has its highest loadings in waters between the Polar and the Atlantic waters north of Iceland, and in an intermediate area southwest of Iceland. The East and West Greenland Current Assemblage (factor 7) with *Thalassiosira gravida* vegetative cells has its highest factor loadings under the EGC and the West Greenland Current, continuing into the Labrador Sea. A small plume of this assemblage is also seen in close vicinity to the Arctic front, west of the NwAC. The Mixed Water Masses Assemblage (factor 8) with *Rhizosolenia borealis* has the highest loadings in a conjunction area between the Labrador Sea and the North Atlantic.

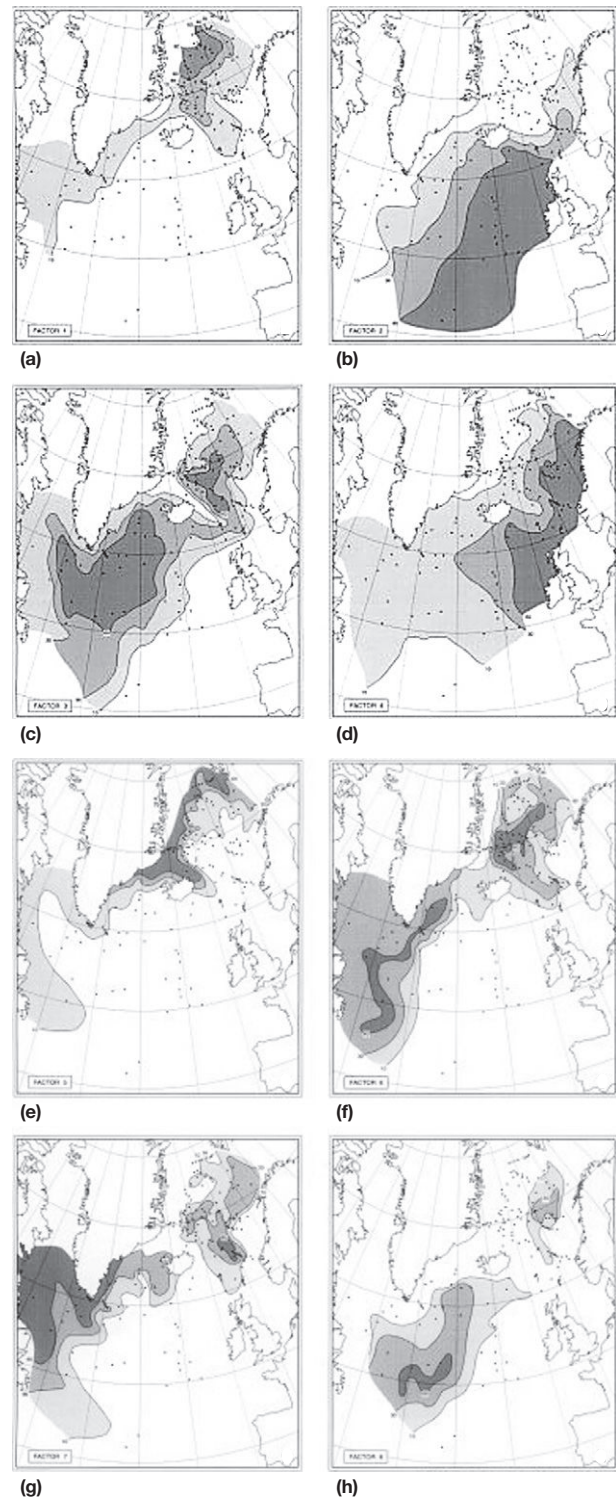


Figure 5 The modern geographic distribution and maximum factor loadings ($\times 100$) (shaded for factor loadings 10, 30, and 60) of: (a) the Arctic Greenland Assemblage (factor 1); (b) the North Atlantic Assemblage (factor 2); (c) the sub-Arctic Assemblage (factor 3); (d) the Norwegian Atlantic Current Assemblage (factor 4); (e) the Sea-Ice Assemblage (factor 5); (f) the Arctic Assemblage (factor 6); (g) the East and West Greenland Current Assemblage (factor 7); and (h) the Mixed Water Masses Assemblage (factor 8) (Andersen et al., 2004b).

Because of their close affinity to modern hydrographic regimes, these factors provide us with a tool to reconstruct details of past surface current variability.

Portuguese Margin

The entire Portuguese west coast and western Algarve shelf are areas of upwelling. Diatom assemblages of the Portuguese continental margin have been mapped from 169 surface sediment samples (Abrantes, 1988). The results of the cluster and principal component analyses indicate that diatom species and assemblages are good indicators of the changing character of the upwelling system in this area. For example, small species of the genus *Thalassiosira* occur in areas of persistent upwelling.

Baffin Bay and Davis Strait

The Baffin Bay and Davis Strait area is dominated by a counter-clockwise circulation driven by the cold Baffin Current and the West Greenland Current. Baffin Bay is covered by winter sea ice, and pack ice can last until August in west central Baffin Bay. Analysis of 74 surface sediment samples resulted in the definition of five diatom assemblages in this area (Williams, 1986). These assemblages are closely related to discrete water masses and ice conditions in the area. Baffin Current Assemblage (factor 1) is almost totally dominated by spores of *Thalassiosira gravida*. The Summer Pack Ice Assemblage (factor 2) is dominated by *Actinocyclus curvatulus* with a secondary contribution of *Thalassiosira trifulta*. The West Greenland Current Assemblage (factor 3) is dominated by *Thalassiosira hyalina*. The Fast Ice Assemblage (factor 4) is characterized by *Porosira* sp. Spores, and the Marine Littoral Assemblage (factor 5) is dominated by *Thalassionema nitzschioides* and *Porosira glacialis*.

Laptev and Kara Seas

The modern distribution pattern of diatoms in the Laptev Sea has been mapped from 89 surface sediment samples (Cremer, 1999), and grouped by Q-mode factor analysis into five factors (diatom assemblages): the sea-ice diatom assemblage of the central region of the Laptev Sea, the *Chaetoceros* assemblage of the eastern and southeastern shelf, the *Thalassiosira antarctica* assemblage of the continental slope and deep sea, the freshwater diatom assemblage in the vicinity of river mouths and deltas, and the *Thalassiosira nordenskiöldii* assemblage which shows a patchy occurrence on the central Laptev Sea shelf. The main factor controlling the distribution of diatoms in the Laptev Sea is the input of riverine freshwater during the summer, which strongly affects the salinity conditions and the sea-ice extent.

The distribution of diatom assemblages from the outer Ob and Yenisei estuaries and the adjacent inner Kara Sea shelf have been mapped from 62 surface sediment samples (Polyakova, 2003). A total number of 260 diatom species and varieties were identified in these samples and grouped according to their known salinity tolerances and their phytogeographical distributions. The overall distribution pattern of these groups mainly reflects summer surface salinity and the pattern of sea-ice distribution in the area.

SSTs in the North Atlantic and the Nordic Seas

Diatom-based quantitative SSTs have been developed using Imbrie and Kipp's (1971) 'factor analysis regression' and ter Braak and Juggins's (1993) WA-PLS. Both transfer functions have low root mean square error of prediction ($\sim 1^\circ\text{C}$), a high coefficient of determination r^2 (~ 0.9), and a maximum bias of approximately 1°C (Andersen et al., 2004a; Birks and Koç, 2002).

Transfer functions in the high-latitude North Atlantic have primarily been applied to the deglaciation (Koç and Jansen, 1992) and the Holocene periods (Andersen et al., 2004a,b; Birks and Koç, 2002; Jiang et al., 2005; Koç et al., 1993). Due to extensive sea-ice cover during the glacial periods, it has been impossible to obtain continuous, long diatom records from the high-latitude North Atlantic.

Deglaciation

Diatoms are absent in the LGM sediments of the northern North Atlantic and the Nordic Seas. Diatoms first occurred in the sediments at the start of the deglaciation as the sea-ice cover retreated. Reconstruction of SSTs show that the initial warming of the surface North Atlantic after the LGM happened synchronously between 50°N and 63°N at 13.4^{14}C ka , but was delayed by 1 ka at 72°N (Koç et al., 1996). While a small ice-free corridor opened along the Norwegian coast during the deglaciation, the rest of the Nordic Seas were still covered by sea ice (Figure 6; Koç et al., 1993). The Bølling–Allerød period was a climatically unstable time interval punctuated by several SST coolings (Koç and Jansen, 1992). The magnitudes of these brief coolings were strongly subdued in the North Atlantic compared to the eastern Norwegian Sea. A strong (9°C) gradient (the Arctic front) between the Norwegian Sea and the North Atlantic was present throughout the Bølling–Allerød period. This gradient became even stronger (15°C) during the Younger Dryas period and nearly disappeared during the early Holocene. During the Younger Dryas period, diatom assemblages off the Norwegian coast were dominated by sea-ice species indicating the proximity of the marginal ice zone to the area.

Holocene

The Younger Dryas–early Holocene transition in the Vøring Plateau (the Norwegian Sea), was characterized by a temperature increase of $\sim 3^\circ\text{C}$ within 100 years (Figure 7(a); Berner et al., 2010). The sea-ice species of the Younger Dryas period were very rapidly replaced by warm Atlantic species at the Younger Dryas–Holocene transition. In the Nordic Seas, the early Holocene is dominated by the North Atlantic Assemblage, with its dominant species *Shionodiscus* (*Thalassiosira*) *oestrupii* indicating the invasion of the central and eastern Nordic Seas by the warm Atlantic waters (Figure 8; Andersen et al., 2004a; Koç et al., 1993). After 6.5 cal ka BP, the influence of the North Atlantic Assemblage gradually decreased and Arctic species dominated the assemblage. The diatom-based SSTs indicate a division of the Holocene in the Nordic Seas into three periods: the Holocene climate optimum (HCO)

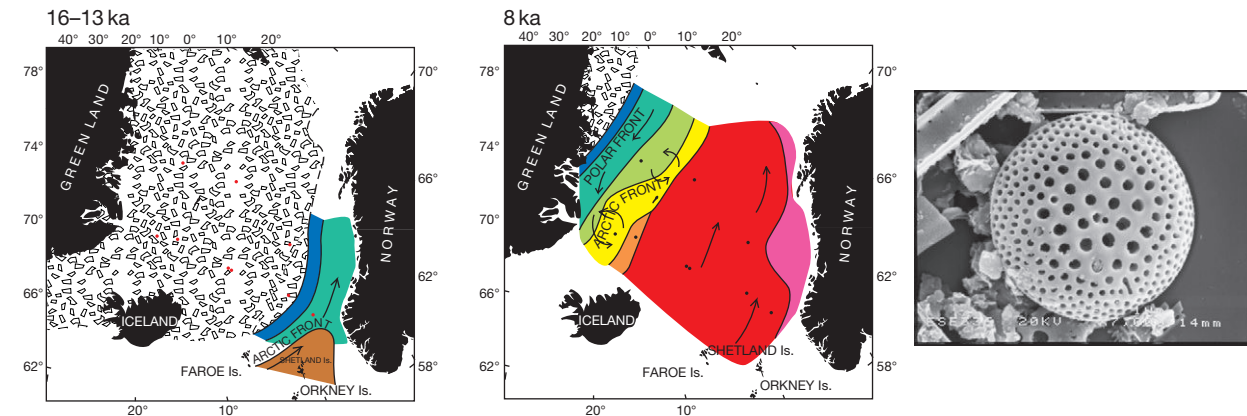


Figure 6 Reconstructions of paleoceanographic conditions of the Nordic Seas during the deglaciation between 16 and 13 ka (calibrated years) and during the Holocene Climate Optimum at 8 ka together with the dominant Atlantic water species *Thalassiosira oestrupii*. Plotted are the sites of the cores used (dots), polar and Arctic fronts with surface ocean circulation (denoted by arrows) and sea-ice cover (broken pattern). Also plotted are the reconstructions of the areal distribution of the various factors described in Koç and Schrader (1990); factor 1, Norwegian Atlantic Current assemblage (pink); factor 2, Arctic Waters assemblage (green); factor 3, sea-ice assemblage (blue); factor 4, Arctic-Norwegian Waters Mixing assemblage (yellow); factor 5, Atlantic assemblage (red); factor 6, Norwegian-Arctic Waters Mixing assemblage (brown). Areas of overlaps are indicated by the mixture of the two overlapping colors (Koç et al., 1993).

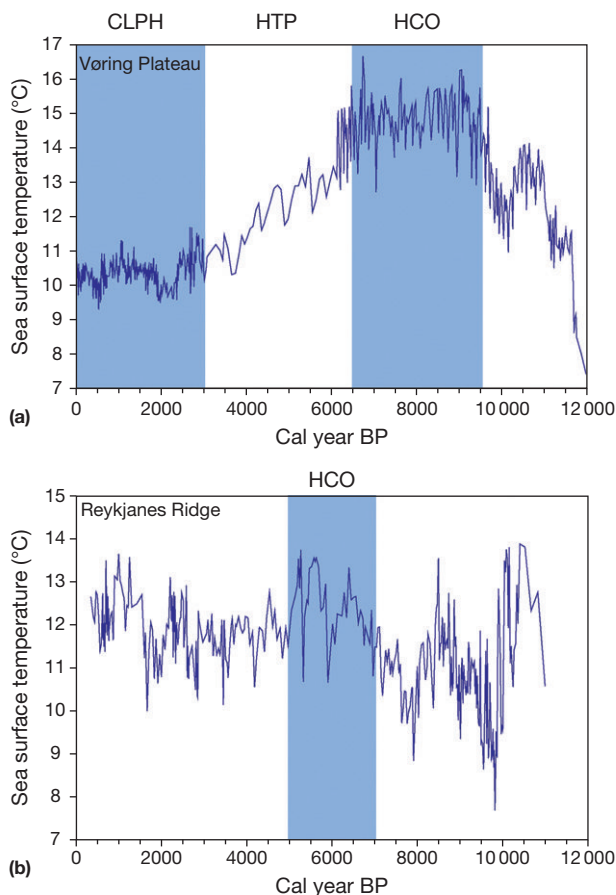


Figure 7 Holocene sea-surface temperature development at (a), the Vøring Plateau, NE Norwegian Sea and (b), the Reykjanes Ridge in the subpolar North Atlantic. The Vøring Plateau record is divided into the HCO, the HTP, and the CLPH (Andersen et al., 2004a; Berner et al., 2008; Berner et al., 2011).

period between 9.5 and 6.5 cal ka BP with SSTs 4 °C warmer than the modern, the Holocene transition period between 6.5 and 3 cal ka BP displaying a gradual SST decrease toward modern values, and the cool late Holocene period between 3 cal ka BP and the present with temperatures around modern SST values (Figure 7(a); Andersen et al., 2004a; Berner et al., 2011). The close similarity of the climate development in the Nordic Seas to high-latitude Northern Hemisphere insolation indicates that it has mainly been driven by insolation forcing (Koç and Jansen, 1994).

The western subarctic gyre is an area of intense mixing of warm and cold waters, which enhances biological productivity. The Holocene sediments of this area are characterized by abundant diatoms. Several levels within the Holocene are dominated by *Rhizosolenia borealis* and *Thalassiothrix longissima*, often as diatom mats (Andersen et al., 2004a,b).

The development of surface ocean climate in the subpolar North Atlantic deviates from the development in the Nordic Seas, both in terms of the timing of the HCO and the stability of SSTs. Over the Reykjanes area, the thermal optimum was not reached until 7–5 cal ka, lagging the insolation maximum by 4 cal ka (Figure 7(b); Andersen et al., 2004b; Berner et al., 2008). SSTs document climatic oscillations with 600–1000-, ~1500-, and 2500-year periodicities, with a time-dependent dominance of different periodicities through the Holocene; a clear change in variability occurred about 5 cal ka BP (Berner et al., 2008). SSTs also provide evidence for Holocene cooling events that, in some cases, correlate to documented southward intrusions of ice into the North Atlantic.

For the last 230 years, August SSTs show a warming trend of ~0.5 °C in the Reykjanes Ridge area (subpolar North Atlantic, Figure 9; Miettinen et al., 2011). The interval from the AD 1770s to the AD 1830s was the coldest period, whereas AD 1860–80 was the warmest period during the last 230 years. The last three decades from the AD 1970s are characterized by a

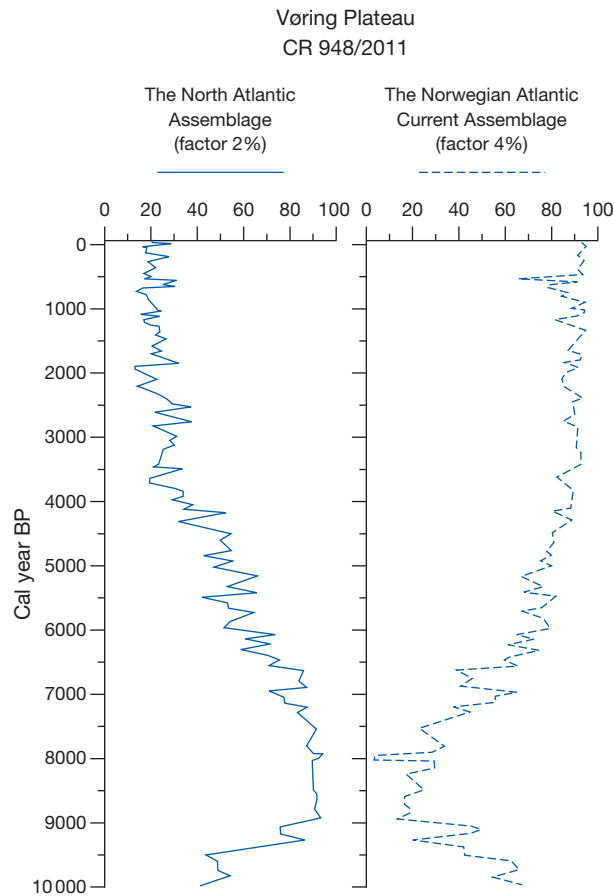


Figure 8 Changes in Holocene dominance of diatom assemblages (factors 2 and 4) from the Vøring Plateau, NE Norwegian Sea (Andersen et al., 2004a).

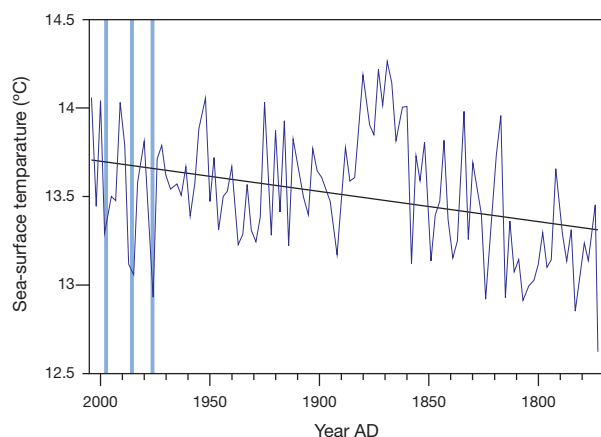


Figure 9 Sea-surface temperature record for the last 230 years from the Reykjanes Ridge, the subpolar North Atlantic (Miettinen et al., 2011). Black line shows a linear warming trend, and the great salinity anomalies in the mid-1970s, the 1980s, and the 1990s are shaded.

warming trend showing strong decadal SST variability with several warm years, but also the coldest years since the AD 1820s. These cold years correspond with the documented great salinity anomalies in the North Atlantic, suggesting increased fluxes of cold, low-salinity waters from the Arctic. SST and the August North Atlantic Oscillation (NAO) index (Luterbacher et al., 2002) show similar but reversed multidecadal-scale variability indicating a close coupling between the oceanic and atmospheric patterns. On the multidecadal scale, cold SSTs prevailed during the positive August NAO trends and vice versa. Therefore, the wind-driven variations in volume fluxes of the North Atlantic surface waters could be the major mechanism behind the observed SST trends during the last centuries (Miettinen et al., 2011).

During the Holocene, climate development in the subpolar North Atlantic in the proximity of the Sub-Arctic Front seems to have been mainly affected by (1) incursions of meltwater from the melting remnants of the Laurentide Ice Sheet, (2) increased outflow of polar water from the Canadian Arctic, (3) variations in the production rate of North Atlantic Deep Water (NADW), (4) variations in solar insolation, (5) variations in the strength of the Atlantic Meridional Overturning Circulation, and (6) atmospheric forcing (the NAO) on the multidecadal scale.

Conclusion

Numerous studies have shown that, due to their high diversity in polar and Arctic environments, diatoms are one of the foremost tools available for biostratigraphic and paleoceanographic studies in the mid- to high-latitude North Atlantic. Their high sensitivity to surface ocean conditions have rendered them invaluable especially for reconstructing the sea-ice margin, oceanic frontal migration, distribution of surface water masses and providing quantitative estimates of, for example, past SSTs. They are especially useful for reconstructing subtle SST changes in the interglacials (e.g., the Holocene). Great potential therefore exists in using diatoms for exploring the natural variability of the mid- to high-latitude North Atlantic using quantitative methods to extract information on the magnitudes and frequency of past ocean changes.

See also: **Diatom Methods:** $\delta^{18}\text{O}$ Records. **Paleoceanography, Biological Proxies:** Marine Diatoms. **Paleoclimate Reconstruction:** Sub-Milankovitch (DO/Heinrich) Events.

References

- Abrantes F (1988) Diatom assemblages as upwelling indicators in surface sediments off Portugal. *Marine Geology* 85: 15–39.
- Abrantes F (2000) 200,000 yr diatom records from Atlantic upwelling sites reveal maximum productivity during LGM and a shift in phytoplankton community structure at 185,000 yr. *Earth and Planetary Science Letters* 176: 7–16.
- Andersen C, Koç N, Jennings A, and Andrews JT (2004a) Nonuniform response of the major surface currents in the Nordic Seas to insolation forcing: Implications for the Holocene climate variability. *Paleoceanography* 19: PA2003. <http://dx.doi.org/10.1029/2002PA000873>.

- Andersen C, Koç N, and Moros M (2004b) A highly unstable Holocene climate in the subpolar North Atlantic: Evidence from diatoms. *Quaternary Science Reviews* 23: 2155–2166.
- Baldauf JG (1984) Cenozoic diatom biostratigraphy and paleoceanography of the Rockall Plateau region, North Atlantic, Deep Sea Drilling Project, Leg 81. *Initial Reports of the Deep Sea Drilling Project* 81: 439–478. Washington, DC: US Government Printing Office.
- Baldauf JG (1986) Diatom biostratigraphic and paleoceanographic interpretations for the middle to high latitude North Atlantic Ocean. In: Summerhays CP and Chackleton NJ (eds.) *North Atlantic Paleoceneanography*, pp. 243–252. London: Geological Society. Special Publication 21.
- Baldauf JG (1987) Diatom biostratigraphy of the middle and high latitude North Atlantic Ocean, Deep Sea Drilling Project, Leg 94. *Initial Reports of the Deep Sea Drilling Project* 94: 729–762. Washington, DC: US Government Printing Office.
- Barron JA (1985) Diatom biostratigraphy of the CESAR 6 core, Alpha Ridge. In: Jackson HR, Mudie PJ, and Blasco SM (eds.) *Initial Geological Report on CESAR – The Canadian Expedition to Study the Alpha Ridge*, pp. 137–148. Ottawa, ON: Geological Survey of Canada. Paper 84–22, Report 10.
- Berner KS, Koç N, Divine D, Godtliebsen F, and Moros M (2008) A decadal-scale Holocene sea surface temperature record from the subpolar North Atlantic constructed using diatoms and statistics and its relation to other climate parameters. *Paleoceanography* 23: PA2210. <http://dx.doi.org/10.1029/2006PA001339>.
- Berner KS, Koç N, and Godtliebsen F (2010) High frequency climate variability of the Norwegian Atlantic Current during the early Holocene period and a possible connection to the Gleissberg cycle. *The Holocene* 20: 245–255.
- Berner KS, Koç N, Godtliebsen F, and Divine D (2011) Holocene climate variability of the Norwegian Atlantic Current during high and low solar insolation forcing. *Paleoceanography* 26: PA2220.
- Birks CJA and Koç N (2002) A high-resolution diatom record of late-Quaternary sea-surface temperatures and oceanographic conditions from the eastern Norwegian Sea. *Boreas* 31(4): 323–344.
- Cremer H (1999) Distribution patterns of diatom surface sediment assemblages in the Laptev Sea (Arctic Ocean). *Marine Micropaleontology* 38: 39–67.
- Davies A, Kemp AES, and Pike J (2009) Late Cretaceous seasonal ocean variability from the Arctic. *Nature* 460: 254–259. <http://dx.doi.org/10.1038/nature08141>.
- Dell'agnese DJ and Clark DL (1994) Siliceous microfossils from the warm Late Cretaceous and Early Cenozoic Arctic Ocean. *Journal of Paleontology* 68: 31–47.
- Dzinoridze RN, Jouse AP, Koroleva-Golikova GS, et al. (1978) Diatom and radiolarian Cenozoic stratigraphy, Norwegian Basin; DSDP Leg 38. In: Talwani M and Udintsev G, et al. (eds.) *Initial Reports of the Deep Sea Drilling Project*, Supplement to vols. 38, 39 and 41, pp. 289–428. Washington, DC: US Government Printing Office.
- Eldrett JS, Harding IC, Firth JV, and Roberts AP (2004) Magnetostratigraphic calibration of Eocene–Oligocene dinoflagellate cyst biostratigraphy from the Norwegian–Greenland Sea. *Marine Geology* 204: 91–127.
- Gradstein FM, Ogg JG, and Smith A (eds.) (2004) *A Geologic Time Scale 2004*, p. 589. Cambridge, UK: Cambridge University Press.
- Hutson WH (1980) The Agulhas current during the Late Pleistocene: Analysis of modern faunal analogs. *Science* 207: 64–66.
- Imbrie J and Kipp NG (1971) A new micropaleontological method for quantitative micropaleontology: Application to a late Pleistocene Caribbean core. In: Turekian K (ed.) *Late Cenozoic Glacial Ages*, pp. 71–181. New Haven, CT: Yale University Press.
- Jiang H, Eiriksson J, Schulz M, Knudsen KL, and Seidenkrantz MS (2005) Evidence for solar forcing of sea-surface temperature on the North Icelandic Shelf during the Late Holocene. *Geology* 33(1): 73–76.
- Koç N and Flower B (1998) High-resolution Pleistocene diatom biostratigraphy and paleoceanography of Site 919 from the Irminger Basin. In: Saunders AD, Larsen HC, and Wise SW Jr. (eds.) *Proc. ODP Sci. Results* 152: 209–219. College Station, TX: Ocean Drilling Program.
- Koç N, Hodell DA, Kleiven H, and Labeyrie L (1999) High-resolution Pleistocene diatom biostratigraphy of Site 983 and correlations to isotope stratigraphy. In: Jansen E, Raymo M, and Blum P (eds.) *Proc. ODP Sci. Results*. 162: 51–62. College Station, TX: Ocean Drilling Program.
- Koç KN and Jansen E (1992) A high resolution diatom record of the last deglaciation from the SE Norwegian Sea: Documentation of rapid climatic changes. *Paleoceanography* 7: 499–520.
- Koç N and Jansen E (1994) Response of the high-latitude Northern Hemisphere to orbital climate forcing: Evidence from the Nordic Seas. *Geology* 22: 523–526.
- Koç N, Jansen E, and Hafliðason H (1993) Paleoceanographic reconstructions of surface ocean conditions in the Greenland, Iceland and Norwegian Seas through the last 14ka based on diatoms. *Quaternary Science Reviews* 12: 115–140.
- Koç N, Jansen E, Hald M, and Labeyrie L (1996) Late glacial-Holocene sea surface temperatures and gradients between the North Atlantic and the Norwegian Seas: Implications for the Nordic heat pump. In: Andrews JT, Austin WEN, Bergsten H, and Jennings AE (eds.) *Late Quaternary Paleoceneanography of the North Atlantic Margins*, pp. 177–185. London: Geological Society. Special Publication No. 111.
- Koç N, Labeyrie L, Manthé S, Flower BP, Hodell DA, and Aksu A (2001) The last occurrence of *Proboscya curvirostris* in the North Atlantic marine isotope stages 9–8. *Marine Micropaleontology* 41: 9–23.
- Koç N and Scherer R (1996) Neogene diatom biostratigraphy of the Iceland Sea Site 907. In: Thiede J and Myhre A (eds.) *Proc. ODP Sci. Results* 151: 61–74. College Station, TX: Ocean Drilling Program.
- Koç KN and Schrader H (1990) Surface sediment diatom distribution and Holocene paleotemperature variations in the Greenland, Iceland and Norwegian Sea. *Paleoceanography* 5(4): 557–580.
- Luterbacher J, Xoplaki E, Dietrich D, et al. (2002) Extending North Atlantic Oscillation Reconstructions Back to 1500. *Atmospheric Science Letters* 2: 114–124.
- Malmgren BA and Nordlund U (1997) Application of artificial neural networks to paleoceanographic data. *Palaeogeography, Palaeoclimatology, Palaeoecology* 136: 359–373.
- Miettinen A, Koç N, Hall IR, Godtliebsen F, and Divine D (2011) North Atlantic sea surface temperatures and their relation to the North Atlantic Oscillation during the last 230 years. *Climate Dynamics* 36: 533–543. <http://dx.doi.org/10.1007/s00382-010-0791-5>.
- Nave S, Labeyrie L, Gherardi J, et al. (2007) Primary productivity response to Heinrich events in the North Atlantic Ocean and Norwegian Sea. *Paleoceanography* 22: PA3216. <http://dx.doi.org/10.1029/2006PA001335>.
- Onodera J, Takahashi K, and Jordan RW (2008) Eocene silicoflagellate and ebridian paleoceanography in the central Arctic ocean. *Paleoceanography* 23: PA1S15. <http://dx.doi.org/10.1029/2007PA001474>.
- Polyakova YI (2003) Diatom assemblages in surface sediments of the Kara Sea (Siberian Arctic) and their relationship to oceanological conditions. In: Stein R, et al. (ed.) *Siberian River Run-off in the Kara Sea: Characterization, Quantification, Variability and Environmental Significance. Proceedings in Marine Science*, pp. 375–399. New York: Springer-Verlag.
- Prell WL, Imbrie J, Martinson DG, et al. (1986) Graphic correlation of oxygen isotope stratigraphy: Application to the Late Quaternary. *Paleoceanography* 1: 137–162.
- CLIMAP Project Members (1981) Seasonal reconstruction of the earth's surface at the last glacial maximum. Geological Society of America, Map and Chart Series MC-36.
- Reid P, Johns D, Edwards M, Starr M, Poulin M, and Snoeijis P (2007) A biological consequence of reducing Arctic ice cover: Arrival of the Pacific diatom *Neodenticula seminae* in the North Atlantic for the first time in 800 000 years. *Global Change Biology* 13: 1910–1921.
- Ruddiman WF (1977) Late Quaternary deposition of ice-rafted sand in the subpolar North Atlantic (lat 40° to 65° N). *Journal of Foraminiferal Research* 19: 253–267.
- Sancetta C (1992) Primary production in the glacial North Atlantic and North Pacific Oceans. *Nature* 360: 249–251.
- Scherer R and Koç N (1996) Late Paleogene diatom biostratigraphy of the northern Norwegian–Greenland Sea, ODP Leg 151. In: Thiede J and Myhre A (eds.) *Proc. ODP Sci. Results* 151: 75–99. College Station, TX: Ocean Drilling Program.
- Schrader H-J and Fenner J (1976) Norwegian Sea Cenozoic diatom biostratigraphy and taxonomy. In: Talwani M and Udintsev G, et al. (eds.) *Init. Repts. DSDP* 38: 921–1099. Washington, DC: US Government Printing Office.
- Stickley CE, Koç N, Brumsack H-J, Jordan RW, and Suto I (2008) A siliceous microfossil view of middle Eocene Arctic paleoenvironments: A window of biosilica production and preservation. *Paleoceanography* 23: 1–19. <http://dx.doi.org/10.1029/2007PA001485> PA1S14.
- Stickley CE, St. John K, Koç N, et al. (2009) Evidence for middle Eocene Arctic sea ice from diatoms and ice-rafted debris. *Nature* 460: 376–379. <http://dx.doi.org/10.1038/nature08163>.
- Suto I, Jordan RW, and Watanabe M (2009) Taxonomy of middle Eocene diatom resting spores and their allied taxa from IODP sites in the central Arctic Ocean (the Lomonosov Ridge). *Micropaleontology* 55: 259–312.
- ter Braak CJF and Juggins S (1993) Weighted averaging partial least squares regression (WA-PLS): An improved method for reconstructing environmental variables from species assemblages. *Hydrobiologia* 269(270): 485–502.
- Williams KM (1986) Recent Arctic marine diatom assemblages from bottom sediments in Baffin Bay and Davis Strait. *Marine Micropaleontology* 10: 327–341.

Introduction

Throughout much of the middle- to high-latitude North Pacific and equatorial Pacific, diatom fossils occur more frequently in sediments than other microfossils, with a greater diversity of species. Since the 1960s, diatom biostratigraphy and paleoceanography has been well established for the various areas of the Pacific. The diatom records in the Pacific Ocean show correlation with the Earth's orbital parameters and also reveal large and abrupt climatic changes within century to millennial time scales.

Many marine diatom species are endemic to a given region, whereas others have a more cosmopolitan distribution. Modern ecological and biogeographic information needs to be constructed from data on diatom fluxes, species composition of diatom assemblages, hydrographic data sets, and seafloor sediments representing present conditions.

Micropaleontological and paleoceanographic studies on the Quaternary siliceous sediments recovered from the middle to high latitudes of the North Pacific Ocean have been mainly based on diatoms. Studies on the diatoms in these regions have been reviewed by Koizumi (2006). The aim here is to provide an up-to-date account of how diatoms are used in marine environments in the Pacific for the inference of past sea-surface temperatures (SSTs) and for studies of biostratigraphy and phylogenetic evolution.

The modern North Pacific Ocean is subdivided into three different marine environments by the North Pacific Current across the Pacific: the Alaska Gyre containing cold nutrient-rich waters, the Subtropical Gyre containing warm nutrient-depleted waters, and a mixed zone of both subarctic and central water masses between 40° and 42° N (Figure 1). The Kuroshio Current encounters the Oyashio cold water in the mid-western area and forms a western boundary current. The California Current is the related eastern boundary current diverging from the Alaskan Current. The northwest Pacific has three marginal seas: the Japan, Okhotsk, and Bering seas. The Tsushima Warm Current (TWC), a branch of the Kuroshio-bounded western side of the anticyclonic Subtropical Gyre, traverses the Japan Sea. Circulation in the subarctic Pacific is a cyclonic gyre with secondary gyres in the Okhotsk and Bering seas. They are sources for Subarctic Water created by vertical and horizontal mixing. Diatom valves are generally well preserved in sediments under the subarctic Alaska Gyre and the Equatorial Current System. However, diatoms are poorly preserved in sediments under the central portion of the Subtropical Gyre in the North Pacific Ocean.

Paleoceanographic Proxies

Paleoceanographic proxies of diatoms have been constructed by using statistical methods comparing the spatial distribution of diatoms in the surface sediments to the primary production,

SST (°C), salinity, and other physical–chemical parameters in the modern upper-ocean conditions. Calibrated relationships have applied to the diatom assemblages in cores to reconstruct past oceanographic conditions.

Sea-Surface Temperature (°C)

Around the Japanese Islands, Northwest Pacific Ocean

Surface water masses around the Japanese Islands show distinct surface-flow patterns, as documented by their physical properties, especially temperature and salinity. Understanding the current patterns of the Tohoku Area off the northeast part of Honshu Island is critical for understanding the complex hydrographic relations between waters in the subtropical and subarctic North Pacific Ocean.

Very complicated hydrographic conditions occur in the 'Perturbed Area' or the 'Mixed Water Region' between the Kuroshio Front to the North of the Kuroshio Extension and the Oyashio Front at latitudes 35–40° N. Numerous meanders and eddies occur along the northern boundary of the Kuroshio Extension (Figure 2). In the Japan Sea, the fan-shaped Subarctic Front at latitude 38–43° N between Cheongjin and Sado Island off the northwestern coast of Honshu Island separates the southern warm-water mass brought in by the TWC from the northern cold-water mass region brought in by the Liman Current.

Regression analysis was performed to estimate the relationship between the ratio of warm- and cold-water diatoms (Td' ratio) in 123 surface sediments around the Japanese Islands and the mean annual SSTs (°C) (Koizumi, 2008). Td' ratios were calculated as follows: $Td' = [(Xw + XW) / (Xw + XW + Xc + XC)]^{-1} \times 100$, where Xw and XW are the frequency of warm-water species, and Xc and XC are that of cold-water species (Koizumi et al, 2004). Figures 3–5 illustrate indicative holoplanktonic species. The annual SSTs (°C) were calculated from the means of the observed SST (°C) data in summer and winter (Figure 6).

There is a strong relationship between Td' and SST for the Tohoku Area and the Japan Sea (Figure 7). There is no systematic bias in the equations. The inferred SSTs (°C) in the Tohoku Area were generally higher than in the Japan Sea despite lower Td' values because the warm-water species *Fragilariopsis doliolus* (Wallich) Medlin and Sims is abundant only in the TWC in the Japan Sea.

Diatom Assemblages Associated with Oceanic Conditions

The Okhotsk Sea

Today, the Okhotsk Sea is a locus of the North Pacific Intermediate Water and also plays a major role as a source area of glacial North Pacific Deep Water formation. The formation of

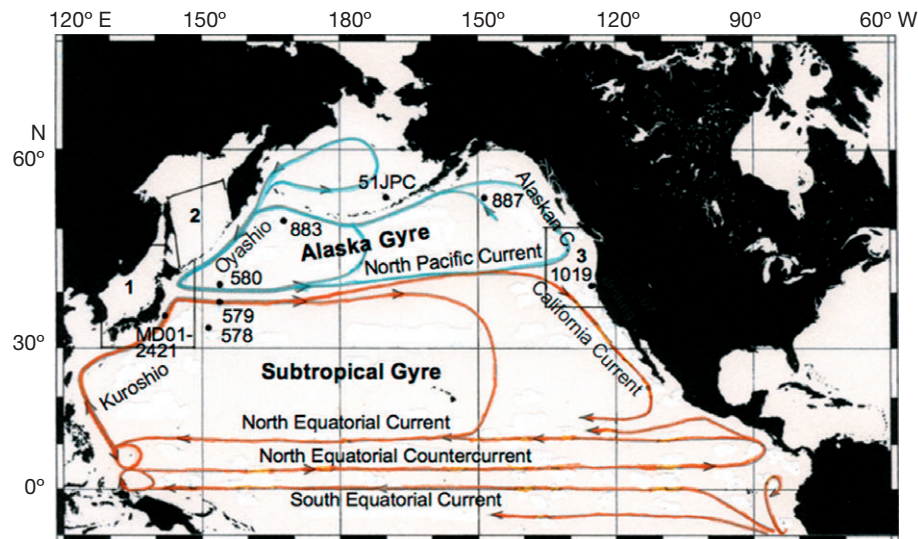


Figure 1 Map showing the location sites of cores mentioned in the text and surface currents in the modern North Pacific Ocean. Orange line, warm-water surface current; blue line, cold-water surface current. The areas of the studied surface sediments are marked by boxes. 1, Koizumi, 2008; 2, Shiga and Koizumi, 2000; 3, Lopes et al., 2006.

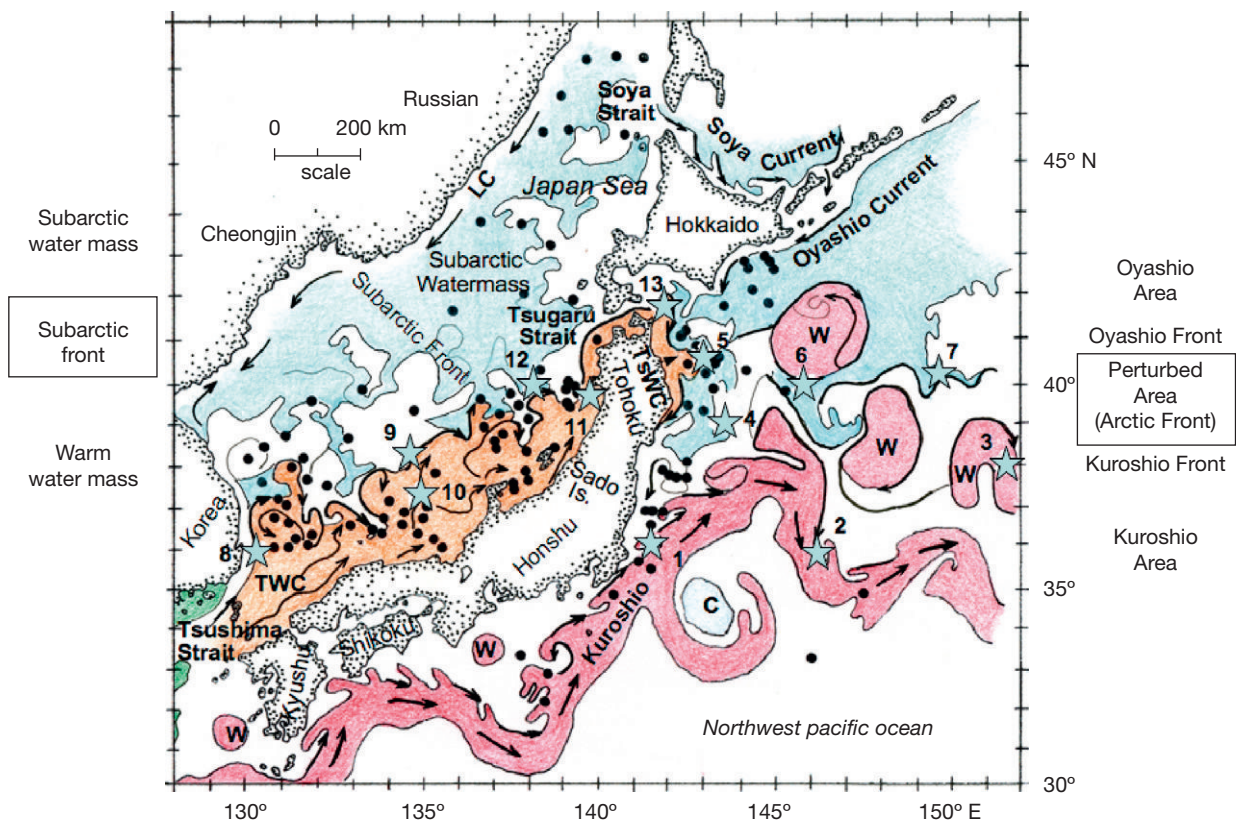


Figure 2 Map showing the location sites of the 123 surface-sediment samples (black circles) and 13 cores (blue stars) around the Japanese Islands. 1, MD01-2421; 2, MR02-03-2; 3, MR99-04-3; 4, ODP Hole 1150A; 5, MR97-04-1; 6, MR00-05-2; 7, MR99-04-2; 8, POI-J3; 9, ODP Site 797; 10, MD01-2407 and DGC-6; 11, CGC-8; 12, KT94-15-5; 13, MD01-2409. Major current systems and the generalized distribution of surface water masses are also indicated. LC, Liman Current. ECSCW, East China Sea Coastal Water: green color; TWC, Tsushima Warm Current and TSWC, Tsugaru Warm Current: orange color; Kuroshio Warm Core and Kuroshio Extension Warm Current: red color; Subarctic water mass: blue color. Arrows indicate flow of the current. W, warm water; C, cold water. The most complicated hydrographic conditions are encountered in the Perturbed Area or Mixed Water Region between the Kuroshio Front and the Oyashio Front at latitudes $\sim 39\text{--}40^\circ\text{N}$ in the Tohoku Area. The fan-shaped Subarctic Front at $39\text{--}43^\circ\text{N}$ separates the southern warm water mass brought in by the TWC from the northern cold-water mass region brought in by the LC.

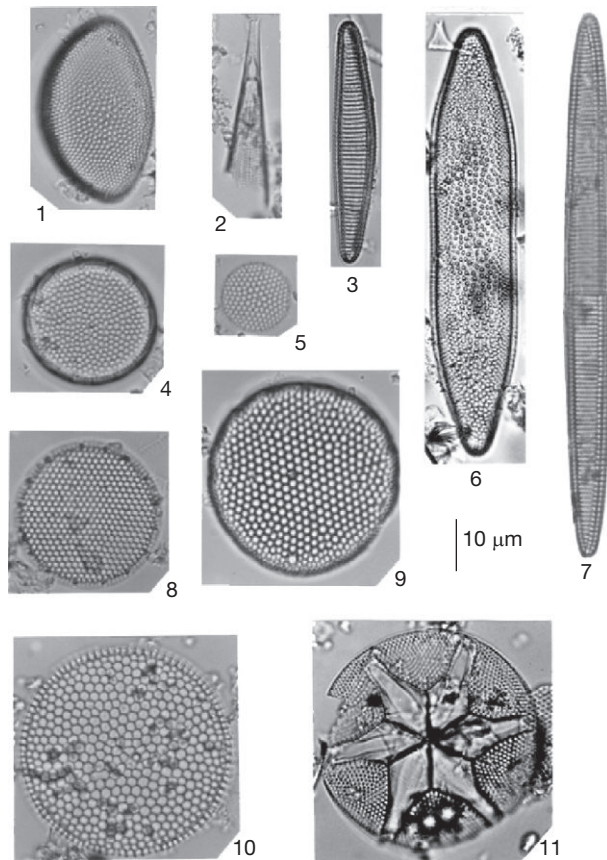


Figure 3 Warm-water species (Xw). (1) *Hemidiscus cuneiformis* Wallich, DSDP 579A-6-4 (25–26 cm). (2) *Rhizosolenia bergonii* Peragallo, DSDP 579-2-1 (123–124 cm). (3) *Fragilariopsis doliolus* (Wallich) Medlin and Sims, DSDP 580-5-4 (40–41 cm). (4) *Azpeitia africana* (Janisch) Fryxell and Watkins, DSDP 579A-3-4 (115–116 cm). (5) *Planktoniella sol* (Wallich) Schtt, DSDP 579-1-3 (97–98 cm). (6) *Actinocyclus elongatus* Grunow, DSDP 579A-12CC. (7) *Alveus marinus* (Grunow) Kaczmarek and Fryxell, DSDP 579-1CC. (8) *Thalassiosira leptopus* (Grunow) Hasle and Fryxell, DSDP 579A-1-5 (14–15 cm). (9) *Roperia tessellata* (Roper) Grunow, DSDP 580-4-1 (121–122 cm). (10) *Azpeitia nodulifera* (Schmidt) Fryxell and Sims, DSDP 579A-1-1 (14–15 cm). (11) *Asterolampra marylandica* Ehrenberg, DSDP 581-6-2 (100–101 cm). Reproduced from Koizumi I (2008) Diatom-derived SSTs (T_d' ratio) indicate warm seas off Japan during the middle Holocene (8.2–3.3 ka BP). *Marine Micropaleontology* 69: 263–281.

the Okhotsk Intermediate Water is linked with seasonal sea-ice cover. Diatoms are by far the most abundant microfossils found in the sediments of the Okhotsk Sea and are promising for interpreting past environments (Koizumi et al., 2003).

Four characteristic diatom assemblages were identified from the spatial distribution of 13 major characteristic diatom species and R-mode principal component analysis in 15 surface sediments (Figure 8; Shiga and Koizumi, 2000). These species reflect environmental factors controlling surface water conditions in the present Okhotsk Sea. The assemblages are (1) A sea-ice assemblage, including *Bacterosira fragilis* Gran, *F. cylindrus* (Grunow) Helmck and Krieger, *F. oceanica* (Cleve) Hasle, and *Thalassiosira gravida* Cleve. This assemblage is present within the floating ice or attached to its under side in the area covered with sea ice, mainly in coastal areas in the western

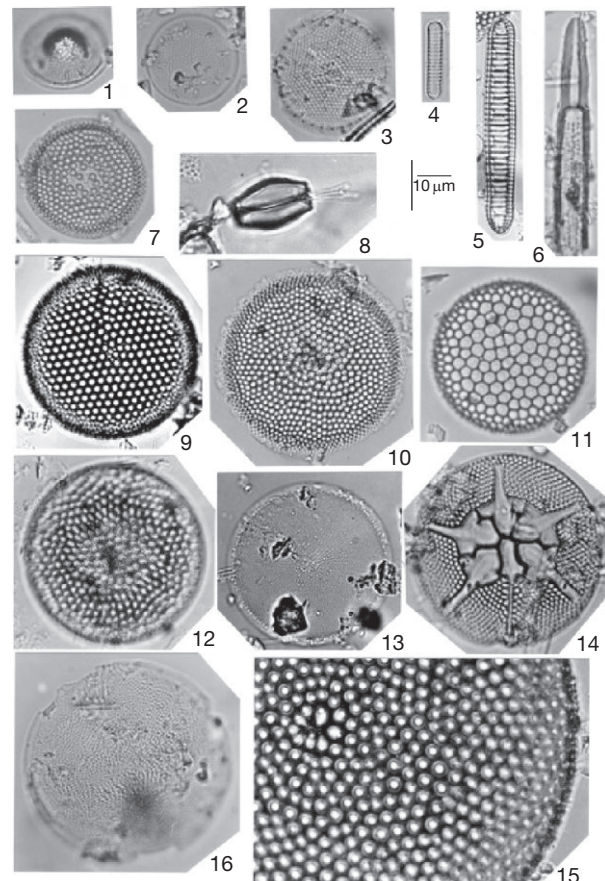


Figure 4 Cold-water species (Xc). (1) *Bacterosira fragilis* Gran, DSDP 580-2-2 (128–129 cm). (2) *Thalassiosira nordenskiöldii* Cleve, DSDP 579-1-2 (97–98 cm). (3) *Thalassiosira kryophila* (Grunow) Joergensen, P333 (52 cm). (4) *Fragilariopsis cylindrus* (Grunow) Krieger, GH76-2-82 (231 cm). (5) *Neodenticula seminae* (Simonsen and Kanaya) Akiba and Yanagisawa, DSDP 580-1-3 (2–3 cm). (6) *Rhizosolenia hebetata* (Bailey) Gran forma hiemalis Gran, DSDP 579-1-5 (97–98 cm). (7) *Actinocyclus ochotensis* Jousé, DSDP 579A-4-3 (12–13 cm). (8) *Chaetoceros furcellatus* Bailey, DSDP 580-1-3 (2–3 cm). (9) *Thalassiosira trifulta* Fryxell, DSDP 580-15-6 (2–3 cm). (10) *Actinocyclus curvatulus* Janisch, DSDP 580-9-2 (23–24 cm). (11) *Coscinodiscus marginatus* Ehrenberg, DSDP 579-1-3 (97–98 cm). (12) *Thalassiosira gravida* Cleve, DSDP 580-2-2 (128–129 cm). (13) *Thalassiosira hyalina* (Grunow) Gran, DSDP 580-7CC. (14) *Asteromphalus robustus* Castracane, DSDP 581-4-2 (98–99 cm). (15) *Coscinodiscus oculus-iridis* Ehrenberg, DSDP 580-9-2 (98–99 cm). (16) *Porosira glacialis* (Grunow) Joergensen, GH76-2-P82 (91 cm). Reproduced from Koizumi I (2008) Diatom-derived SSTs (T_d' ratio) indicate warm seas off Japan during the middle Holocene (8.2–3.3 ka BP). *Marine Micropaleontology* 69: 263–281.

and northern parts of the Okhotsk Sea. *Thalassionema nitzschoides* Grunow is also abundant in the southern part of the Okhotsk Sea under the influence of the warm Kuroshio and/or Soya Current from the Japan Sea (Figure 1). (2) Indicator species for high productivity and oceanic waters, including *Neodenticula seminae* (Simonsen and Kanaya) Akiba and Yanagisawa and *Thalassiothrix longissima* (Cleve) Cleve and Grunow. These are abundant in the central area of the Okhotsk Sea. (3) *Paralia sulcata* (Ehrenberg) Cleve is most common on the continental shelf as a benthic littoral species, but it also occurs in plankton. (4) Oceanic species, *Actinocyclus curvatulus*

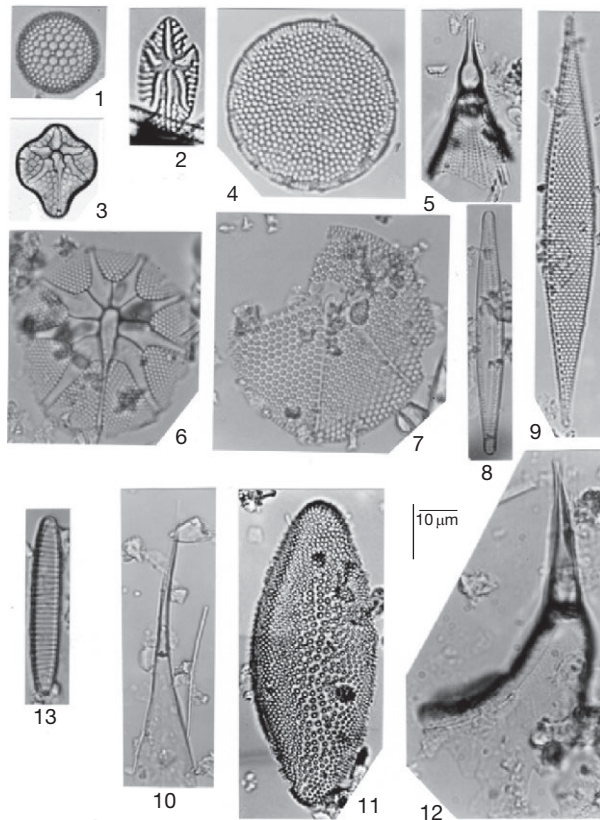


Figure 5 Warm-water species (XW) and cold-water species (XC). (1) *Thalassiosira oestrupii* (Ostenfeld) Hasle, DSDP 579A-1-3 (97–98 cm). (2) *Asteromphalus petterssonii* (Kolbe) Thorning-Smith, DSDP 579A-9-5 (130–131 cm). (3) *Asteromphalus sarcophagus* Wallich, African coast of the Indian Ocean Station 167 (Simonsen, 1974). (4) *Azpeitia tabularis* (Grunow) Fryxell and Sims, GH76-2-P82 (111 cm). (5) *Rhizosolenia acuminata* (Peragallo) Gran, MD01-2421 (0–1 cm). (6) *Asteromphalus flabellatus* (Brebisson) Greville, DSDP 579A-9-3 (30–31 cm). (7) *Asteromphalus arachne* (Brebisson) Ralfs, DSDP 579-1CC. (8) *Nitzschia interruptestriata* (Heiden) Simonsen, DSDP 579A-1-1 (14–15 cm). (9) *Nitzschia kolaczekii* Grunow, DSDP 580-6-6 (17–18 cm). (10) *Rhizosolenia calcar-avis* (Schultze) Sundstrom, DSDP 580-3-6 (25–26 cm). (11) *Actinocyclus ellipticus* Grunow, DSDP 581-3-4 (90–91 cm). (12) *Rhizosolenia imbricata* Brightwell, MD01-2421 (5–6 cm). (13) *Fragilariopsis oceanica* (Cleve) Hasle, P333 (52 cm). Reproduced from Koizumi I (2008) Diatom-derived SSTs (T_d' ratio) indicate warm seas off Japan during the middle Holocene (8.2–3.3 ka BP). *Marine Micropaleontology* 69: 263–281.

Janisch, *Rhizosolenia hebetata* (Bailey) Gran forma *hiemalis* Gran, and *R. styliformis* Brightwell, occur abundantly in the central area of the Okhotsk Sea.

The Bering Sea

The Bering Sea is a gateway between the Pacific and Arctic oceans (Figure 1). The Bering Strait was closed and most of the Bering continental shelf was exposed during intervals of Pleistocene glaciation due to the growth of continental ice sheets, and consequent lowering of sea level (Sancetta, 1983).

In the Bering Sea, the relative percentages of *Fragilariopsis* spp., including *F. cylindrus* (Grunow) Helmck and Krieger and

F. oceanica (Cleve) Hasle, vary from 0% to 76% in 153 core tops from locations that experience between 0 and 11 months of sea ice, (Caissie et al., 2010; Sancetta, 1981). Relative percentages of *Fragilariopsis* spp. have been related to the presence of sea ice (1) less than 3.5% indicating perennially ice-free conditions, (2) between 13% and 18% indicating that sea ice was present 1–4 months per year, and (3) greater than 19% indicating that sea ice was present more than 4 months per year.

Off the Oregon and California Coasts, Northeast Pacific Ocean

Diatom assemblages in regions of high productivity in transitional climate regimes off the coasts of southern British Columbia to northern California in the northeast Pacific Ocean (Figure 1) preserve paleoceanographic information about the coastal and open-ocean upwelling, and major river inputs.

Comparisons between core tops and sediment traps show that diatom remains reaching the sea floor generally reflect the export of resistant robust diatom species from the upper ocean in the northeast Pacific Ocean (Lopes et al., 2006). The relative abundances of *Chaetoceros* spp. appear to respond to both coastal upwelling and to open-ocean upwelling. In open-ocean settings, *Chaetoceros* spp. track annual average conditions, whereas in the coastal region, *Chaetoceros* spp. are associated with summer/fall production.

Q-mode factor analysis from 30 surface-sediment samples with abundant diatoms defines five diatom assemblages that are associated with oceanic conditions (Figure 9): (1) *Chaetoceros* spores are related with coastal upwelling in the area east of 127°W and curl-driven open-ocean upwelling in the area northwest of 46°N. (2) *Thalassionema nitzschiioides* is associated with the warmer subtropical waters. (3) *R. hebetata* is associated with the Subpolar Front in the open ocean. (4) *N. seminae* is related to the intrusions of subarctic water masses into the upwelling system. (5) Freshwater diatoms are related to the influence of freshwater from the continents.

Holocene Deglaciation

Around the Japanese Islands, Northwest Pacific Ocean

Diatom abundance (expressed as diatom valves/g dry weight of sediment) is mainly dependent on the primary productivity of diatoms in the offshore region. Diatom abundances in the Oyashio Area are twice as high as those in the Kuroshio Area because of mixing of the Oyashio intrusion waters and warm-water eddies (Figures 2 and 10). These abundances decreased during the Younger Dryas (YD) and fluctuated during the middle to late Holocene, with an overall increase from 7 to 3 ka and a decrease from 3 ka to the present.

The littoral–neritic cold-water species *Odontella aurita* increased in abundance along the Frontal Zone in the northern margin of the Perturbed Area in the Tohoku Area. In the Japan Sea, the relative abundances of the East China Sea Coastal Water indicator species *P. sulcata* decreased in the northern part more than in the southern part during the YD and Holocene, and also decreased after ~8–9 ka, following the inflow of the TWC carrying oceanic warm-water species. *F. doliolus* was dominant only in the path of the TWC, with its first occurrence

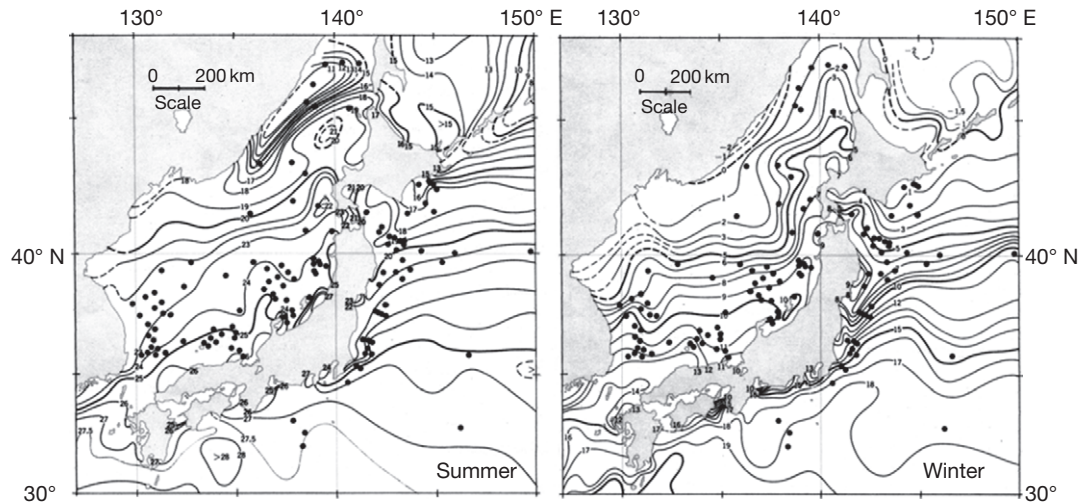


Figure 6 Maps showing the location sites of the 123 surface-sediment samples (black circles) in relation to modern surface water temperatures (°C) in summer and winter around the Japanese Islands. Reproduced from Koizumi I (2008) Diatom-derived SSTs (Td' ratio) indicate warm seas off Japan during the middle Holocene (8.2–3.3 ka BP). *Marine Micropaleontology* 69: 263–281.

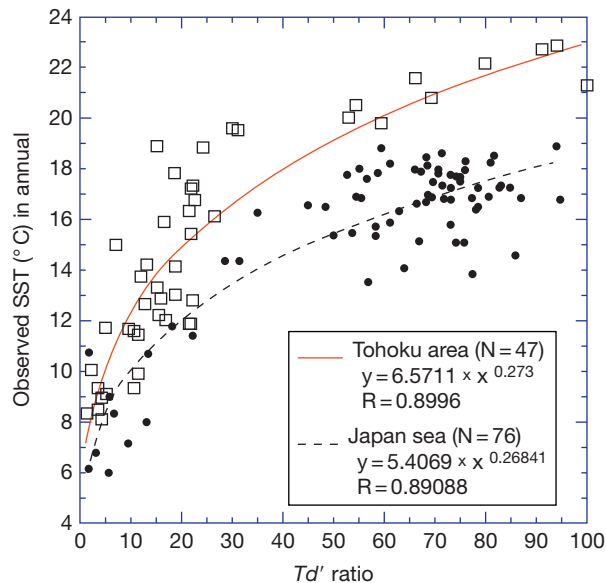


Figure 7 Graph showing the comparison between Td' ratios and calculated annual sea-surface temperatures (SSTs) which is the mean of the observed SSTs (°C) in summer and winter around the Japanese Islands (Koizumi, 2008). The paleo-SSTs (°C) can be estimated using the equations obtained by the regression analysis. Open squares, Tohoku Area; filled circles, Japan Sea. Reproduced from Koizumi I (2008) Diatom-derived SSTs (Td' ratio) indicate warm seas off Japan during the middle Holocene (8.2–3.3 ka BP). *Marine Micropaleontology* 69: 263–281.

(FO) at 8.5–8.0 ka in the southern part, but at 7.5 ka in the northern part (Figure 10).

The Okhotsk Sea

The diatom records from six cores from the Okhotsk Sea showed that during the Last Glacial Maximum (LGM; 21–18 ka), sea ice covered the southwest part of the sea

throughout the whole year, but the eastern and northern parts were characterized by open-sea conditions. Perennial sea ice extended northward, and sea ice was present throughout the whole year over a great part of the sea except for the eastern part during the interval of 18–17 ka. The sea-ice area extended greatly to the south during the interval of 17–11 ka. During the interval of 11–8 ka, there was no perennial ice cover over the whole area, and the ice-free area appeared as the seasonal ice area began to extend to the western part. After 8 ka, the ice-free waters contracted to the present extent (Koizumi et al., 2003; Shiga and Koizumi, 2000).

The Bering Sea

Diatom records in core 51JPC from the Umnak Plateau in the southeastern Bering Sea showed that during the LGM through the early deglacial period (23–17 ka), sea ice was present for more than 6 months each year, with resting spores of *Thalassiosira antarctica* (Fryxell) Doucet and Hubbard (= *Thalassiosira gravis* Cleve) and *Fragilariopsis* spp. dominating. During deglaciation, perennial sea-ice conditions transitioned to ice-free conditions (17–10 ka) coinciding with a high diversity of *Thalassiosira hyalina*, *Thalassiosira nordenskiöldii*, *Thalassionema nitzschoides*, *O. aurita*, *Chaetoceros* resting spores, *Thalassiosira trifurcata*, and *N. seminae* between 15.1 and 13.7 ka. During the Holocene (10–0 ka), a decrease of resting spores of *Chaetoceros* and an increase of *N. seminae* has occurred since the Holocene hypsithermal at 11.5–9 ka (Caissie et al., 2010).

Climate Changes Based on the Diatom-Inferred SSTs (°C)

Around the Japanese Islands, Northwest Pacific Ocean

The Td' -derived paleo-SSTs (°C) in five cores representing the Kuroshio and TWC (Figure 2) decreased during the YD chronozone due to weakening of both currents (Figure 11). On average SSTs (°C) in the YD were 7–9 °C lower than present-day

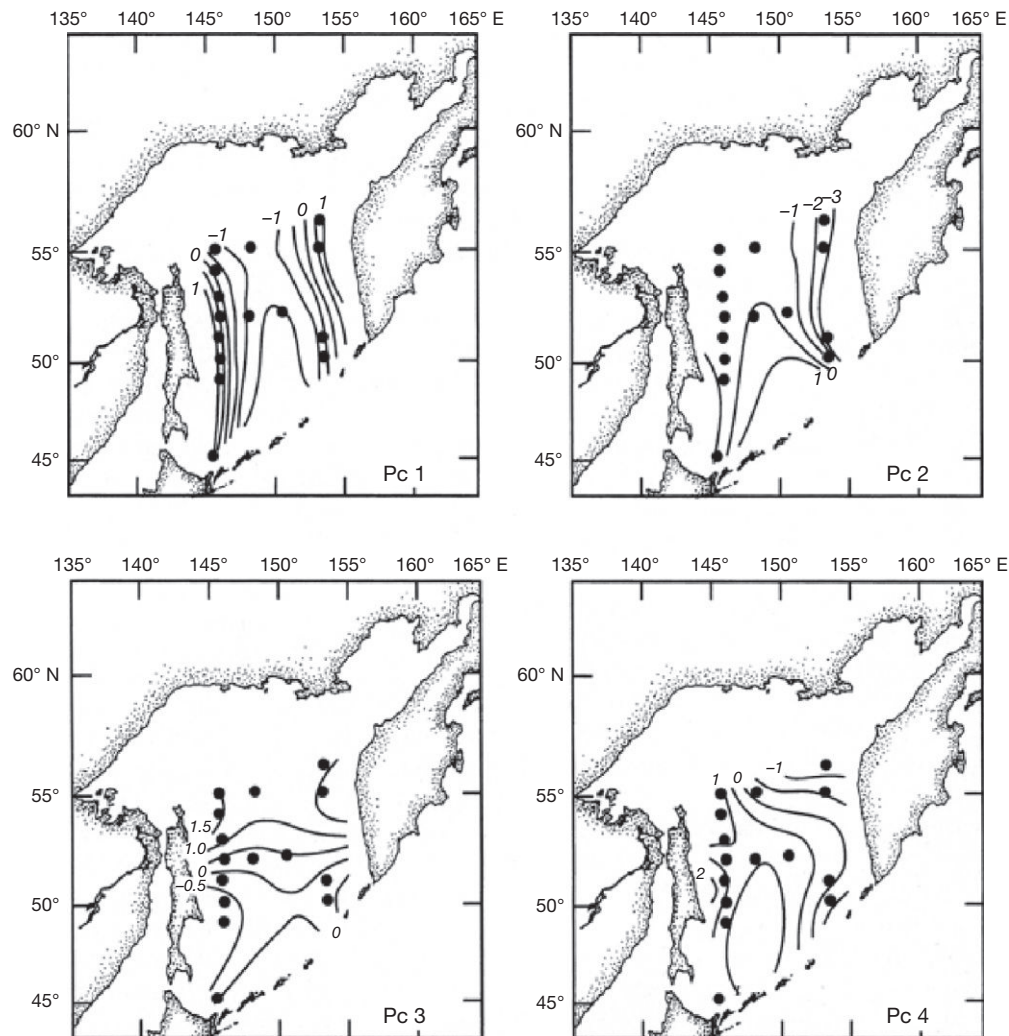


Figure 8 Maps showing the spatial distribution of scores for each component of R-mode principal component analysis of the diatom assemblage in the surface sediments of the Okhotsk Sea. Pc 1, sea-ice assemblage; Pc 2, subarctic oceanic assemblage; Pc 3, neritic assemblage; Pc 4, assemblage dominated by *Actinocyclus curvatus* Janisch. Reproduced from Shiga K and Koizumi I (2000) Latest Quaternary oceanographic changes in the Okhotsk Sea based on diatom records. *Marine Micropaleontology* 38: 91–117.

temperatures in the Tohoku Area, and 6–9 °C lower than those in the Japan Sea. The amplitudes of fluctuation of the Td' -derived SSTs (°C) after the YD were within 7–9 °C in the Tohoku Area and within 6–10 °C in the Japan Sea. The early Holocene (11.6–8.2 ka) was a transitional period characterized by a long-term increasing trend of temperatures punctuated by several cooling events, while the middle Holocene (8.2–3.3 ka) coincided with the Holocene hypsithermal period and was 1–2 °C warmer than the earlier and later parts of Holocene. The late Holocene (3.3–0 ka) neoglacal period was marked by a generally decreasing trend of the paleo-SSTs (°C) (Koizumi, 2008), reflecting the decreased solar insolation in the Northern Hemisphere summer (see Figure 12).

The middle Holocene warm period was also recognized in the alkenone records in the Tohoku Area (Isono et al., 2009) and indicates the presence of millennial scale, antiphased SST relationships with coastal northern California as shown by the alkenone SSTs, and the abundances of the warm-water diatom species *F. doliolus* (Figure 13; Barron and Anderson, 2010;

Barron et al., 2003). Such antiphase SST variations between east-to-west margins of the mid-latitude North Pacific Ocean are similar to the behaviors of the long-term orbital-scale El Niño-Southern Oscillation (ENSO; Clement et al., 1999; Yamamoto et al., 2004). The antiphase SST variations also correspond to the patterns seen in the modern Pacific-Decadal Oscillation (PDO), where cooler SSTs in the central north-west Pacific Ocean contrast with warmer SSTs in the eastern North Pacific Ocean and eastern equatorial Pacific Ocean during periods of positive PDO (Barron and Anderson, 2010; Mantua et al., 1997).

The Td' -SSTs (°C) fluctuate with a possible periodicity of century to millennial timescales, indicating a strong and regular flow of the Kuroshio Current in the Tohoku Area and TWC in the Japan of the northwest Pacific Ocean after ~10 ka. Those fluctuations are synchronous with abrupt climate events reported from the different paleoclimatic proxy records in many regions of the Northern Hemisphere (Koizumi and Yamamoto, 2011). Decreases of Td' -SSTs (°C) correspond to

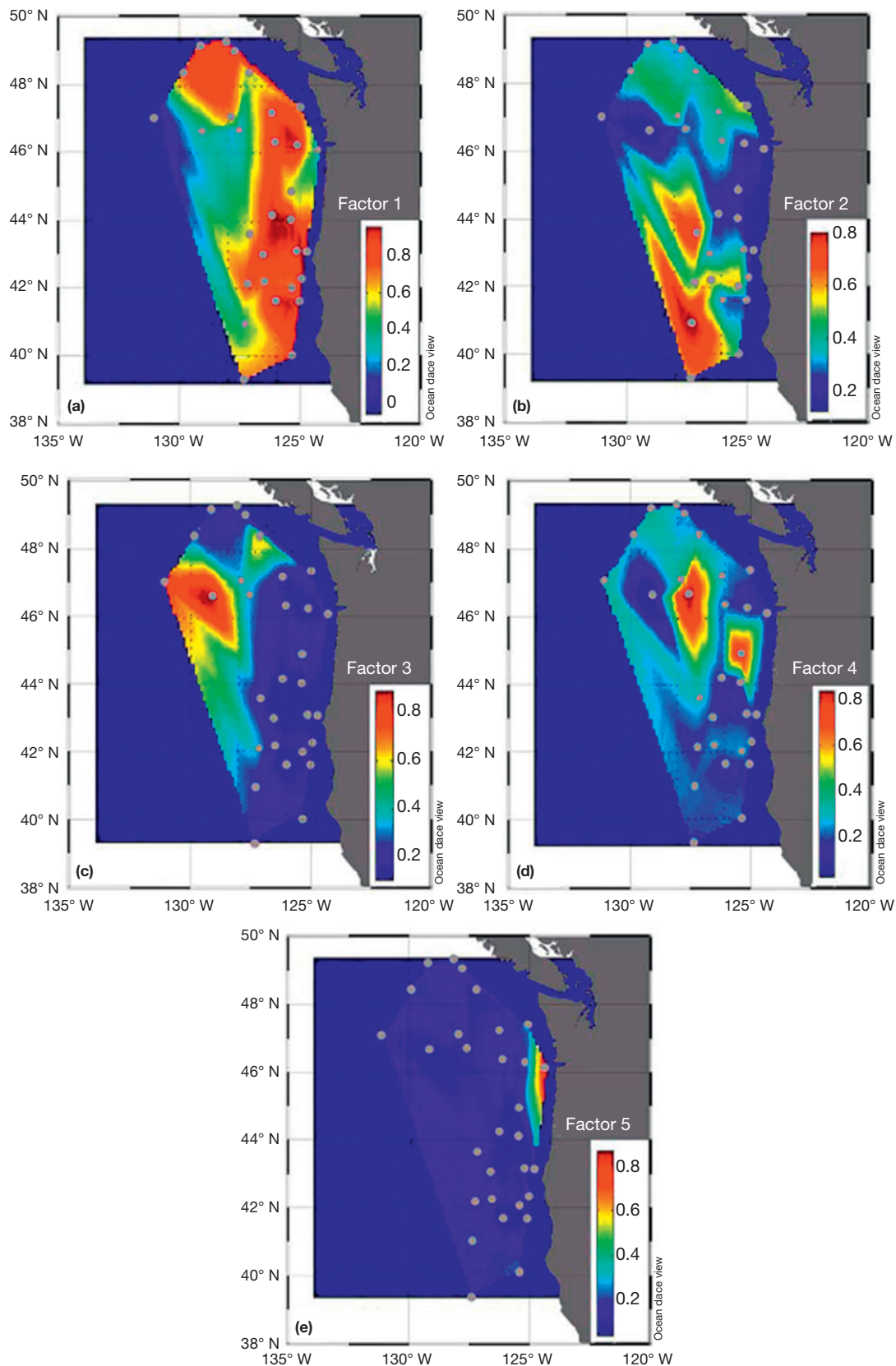


Figure 9 Maps showing the spatial distribution of factor loadings for the five factors of Q-mode factor analysis of the diatom assemblage in the surface sediments of the northeast Pacific Ocean. Factor 1, upwelling (*Chaetoceros* spores); Factor 2, subtropical (*Thalassionema nitzschioides* Grunow); Factor 3, subarctic (*Rhizosolenia hebetata* (Bailey) Gran); Factor 4, transitional (*Neodenticula seminae* (Simonsen and Kanaya) Akiba and Yanagisawa); Factor 5, freshwater (freshwater diatoms). Reproduced from Lopes C, Mix AC, and Abrantes F (2006) Diatoms in northeast Pacific surface sediments as paleoceanographic proxies. *Marine Micropaleontology* 60: 45–65.

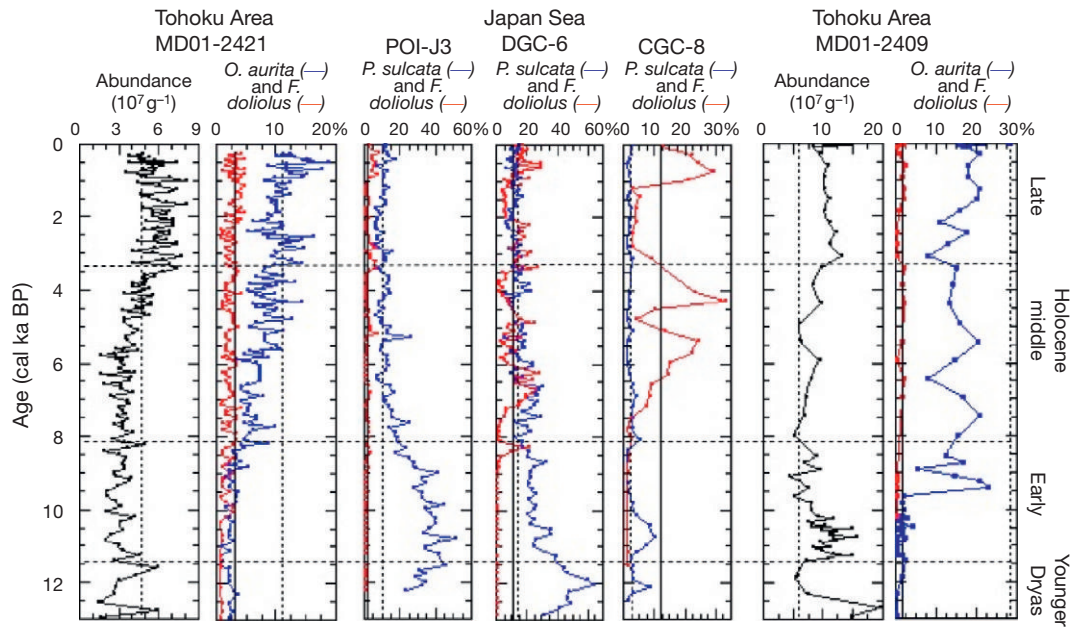


Figure 10 Graphs showing the spatial and temporal variations of the diatom abundances (10^7 valves/dry g of sediment) and relative abundances (%) of characteristic diatom species in the Tohoku Area, and the relative abundances (%) of characteristic diatom species in the Japan Sea (Koizumi, 2008). *Odontella aurita* (Lyngbye) Agardh (blue line) is an indicator of the littoral–neritic cold-water, *P. sulcata* (Ehrenberg) Cleve (blue line) is an indicator of East China Sea Coastal Water, and *Fragilariopsis doliolus* (Wallich) Medlin and Sims (red line) is an oceanic warm-water species, especially dominant in the Tsushima Warm Current. Vertical lines indicate the modern values. Modified from Koizumi I (2008) Diatom-derived SSTs (T_d' ratio) indicate warm seas off Japan during the middle Holocene (8.2–3.3 ka BP). *Marine Micropaleontology* 69: 263–281.

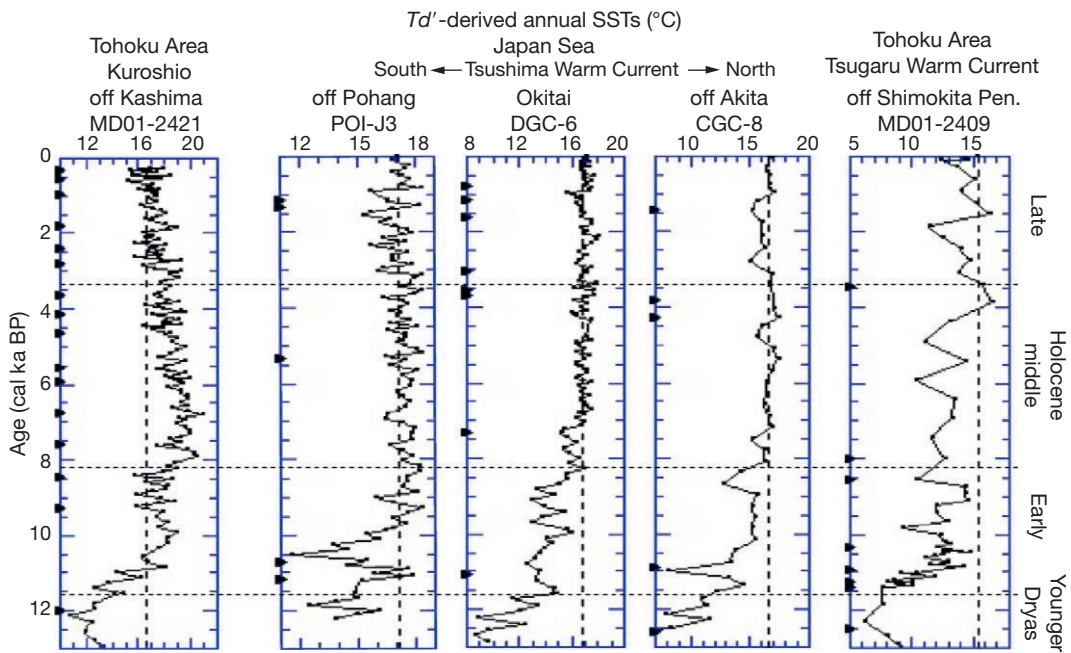


Figure 11 Graphs showing the spatial and temporal variations of the T_d' -derived SSTs ($^{\circ}\text{C}$) in the Tohoku Area and the Japan Sea (Koizumi, 2008). Vertical dotted lines indicate the modern values. Modified from Koizumi I (2008) Diatom-derived SSTs (T_d' ratio) indicate warm seas off Japan during the middle Holocene (8.2–3.3 ka BP). *Marine Micropaleontology* 69: 263–281.

cooling and to the triple events of high ^{14}C values in the atmospheric residual ^{14}C records (Stuiver et al., 1991), as well as the Bond events in the North Atlantic Ocean (Bond et al., 2001; Figure 12).

The T_d' -inferred SSTs ($^{\circ}\text{C}$) agreed with alkenone-derived summer SSTs ($^{\circ}\text{C}$) in the same core MD01-2421 from a site off the eastern coast of central Japan (Koizumi, 2008). However, the comparison between the diatom-based T_d' - and the

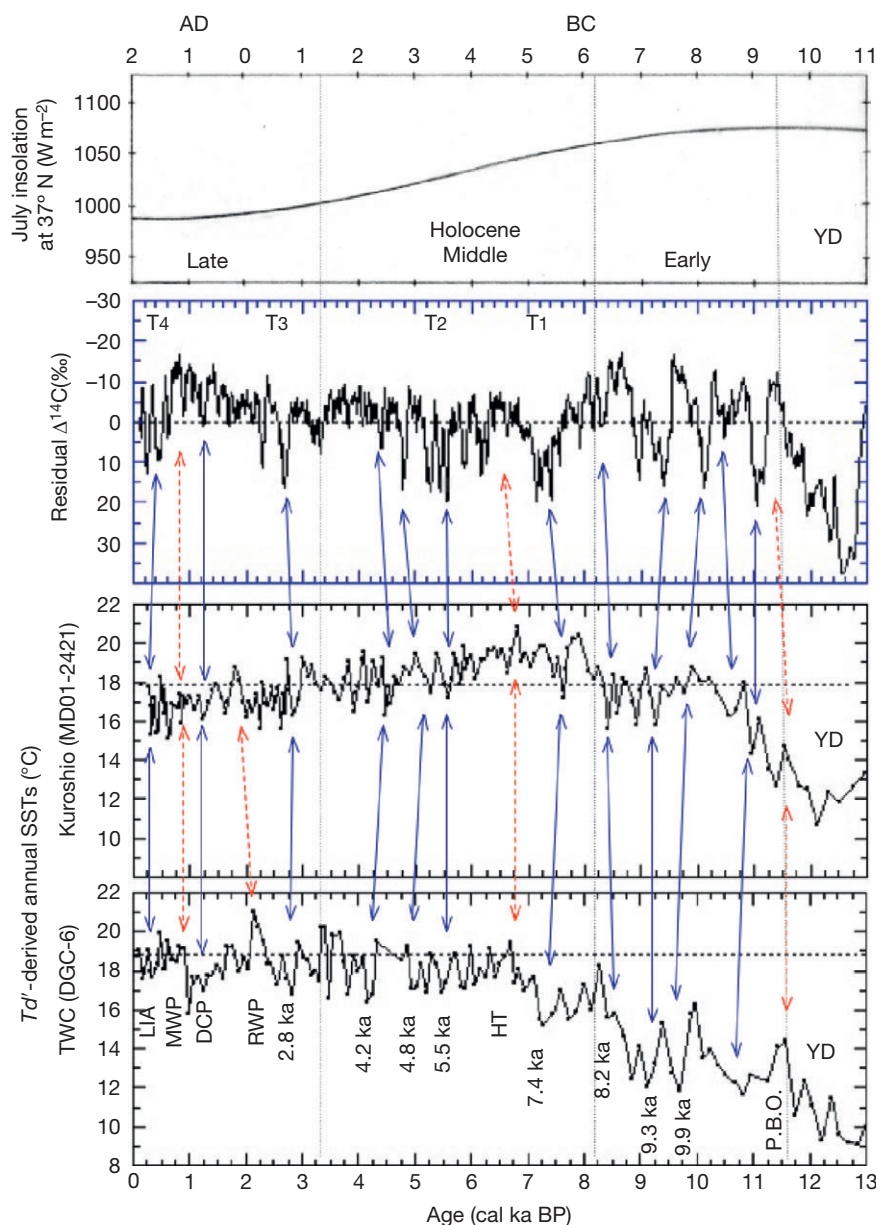


Figure 12 Graphs showing the comparison of variations of the Td' -derived SSTs ($^{\circ}\text{C}$) in cores MD01-2421 and DGC-6, the residual $\Delta^{14}\text{C}$ (‰) (Stuiver et al., 1991), and July insolation at 37°N (W m^{-2}) (Berger, 1978) with the climatic events during the last 13000 years. Blue lines (cold events) and red lines (warm events) indicate the corresponding changes. BC, before Christ. AD, *Anno Domini*; YD, Younger Dryas; PBO, Preboreal Oscillation; HT, Hypsithermal; RWP, Roman Warm Period; DCP, Dark Age Cold Period; MWP, Medieval Warm Period; LIA, Little Ice Age; T1–T4, Triple events comparing Maunder and Spörer perturbations; TWC, Tsushima Warm Current (Stuiver et al., 1991). Reproduced from Koizumi I and Sakamoto T (2010) Synchronous Td' -derived SSTs ($^{\circ}\text{C}$) off Japan with climatic events in the Northern Hemisphere. *Journal of Geography* 119: 489–509 (in Japanese with English abstract).

alkenone-derived (Isono et al., 2009) SSTs ($^{\circ}\text{C}$) in the north-west Pacific Ocean with SSTs ($^{\circ}\text{C}$) estimations in the northeast Pacific Ocean (Barron et al., 2003, 2010) indicate antiphase SST relationships on a millennial scale in the mid-latitude North Pacific Ocean (Isono et al., 2009; Koizumi, 2008).

The cross-spectral analysis of the Td' -derived annual paleo-SSTs ($^{\circ}\text{C}$) in cores MD01-2421 and residual $\Delta^{14}\text{C}$ (‰) indicate a link with fluctuations in atmospheric ^{14}C values of 2400, 1600, 950, and 400–300-year oscillations (Figure 14; Koizumi and Sakamoto, 2010). According to Debret et al. (2007), 1000- and

2500-year cycles are directly forced by solar activity, whereas the 1600-year cycle corresponds to internal oceanic forcing.

The interval from the YD to the early Holocene in the Tohoku Area and the Japan Sea was punctuated by several cooling events: the YD, the Preboreal Oscillation, the 9.9 ka event and the 9.3 ka event. The hypsithermal mid-Holocene in the Tohoku Area and the Japan Sea was punctuated by a series of significant climatic events which fluctuated at ~ 1.5 ka within $1\text{--}2^{\circ}\text{C}$: the 7.4 ka event corresponded with T1 in the triple events of atmospheric high ^{14}C value (Stuiver et al.,

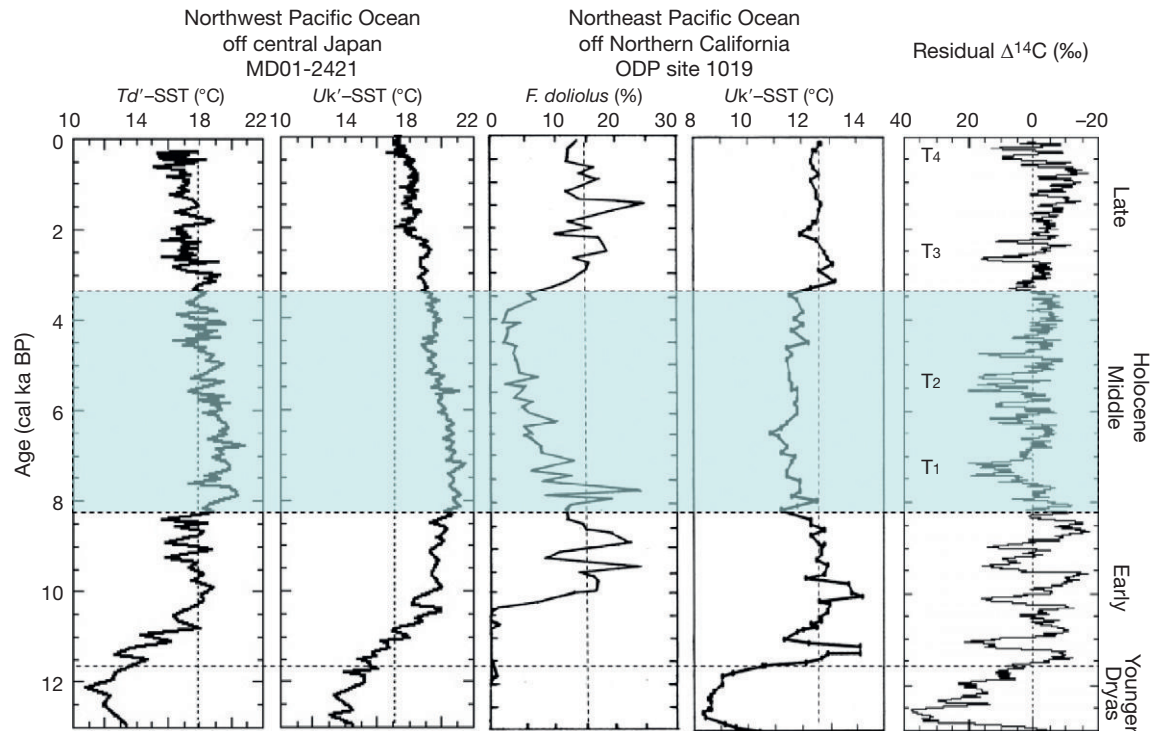


Figure 13 Graphs showing the comparison of the Td' -derived SSTs ($^{\circ}\text{C}$) and $UK'37$ -derived SSTs ($^{\circ}\text{C}$) (Isono et al., 2009) in the Tohoku Area of northwest Pacific Ocean, and the relative abundance (%) of warm-water diatom species *Fragilariopsis doliolus* (Wallich) Medlin and Sims (Barron and Anderson, 2010) and $UK'37$ -derived SSTs ($^{\circ}\text{C}$) (Barron et al., 2003) off northern California of the northeast Pacific Ocean, and the triple events (T1–T4) of similar nature to Maunder and Spörer type perturbations in the residual $\Delta^{14}\text{C}$ (Stuiver et al., 1991).

1991), the 5.5 ka event with T2 (ibid.). The neoglacal late Holocene period in the Tohoku Area and the Japan Sea consists of a 2.8 ka event, the Roman Warm Period (Yayoi Warm Stage), the Dark Age Cold Period (Kofun Cold Stage), the Medieval Warm Period (MWP), and the Little Ice Age (LIA). The antiphase SST variations occurring during the interval between the MWP and LIA in the northwest (Isono et al., 2009; Koizumi, 2008) and northeast (Barron et al., 2010) Pacific Ocean correspond to the combination of behaviors of the ENSO and PDO (D'Arrigo et al., 2001; Gedalof et al., 2002).

Glacial–Interglacial Cycles

Diatom records in the cores from the northwest Pacific Ocean have shown oscillations between warm and cold Pleistocene intervals and have recorded glacial–interglacial cycles twice over the past 150,000 years. Diatom associations and Td' -SSTs ($^{\circ}\text{C}$) imply that the paleohydrography of the northwest Pacific Ocean has corresponded to the environmental changes on an orbital timescale.

The Tohoku Area off Northeast Japan

Diatom assemblages in the southern margin of the Perturbed Area have been controlled by the Kuroshio–Kuroshio Extension, but in the northern margin the composition of the assemblages has changed in response to the expansion of the warm-water eddies detached from the Kuroshio Extension and

the southward Oyashio Intrusion (Figure 2; Koizumi and Yamamoto, 2010).

Reworked and/or displaced diatoms derived from the seafloor are dominant during the glacial periods due to a lower sea level in the near-shore, northern part of the Perturbed Area. The relative abundances of the littoral–neritic association diatom species are higher in the cores in the Kuroshio Extension than in the cores in the northern margin of the Perturbed Area because the flow of the Kuroshio Current northeast along the near-shore of southwest Honshu transports littoral–neritic diatoms into the southern margin of the Perturbed Area.

On the other hand, the oceanic diatom abundances were generally higher during the interglacial phase than during the glacial phase (Figure 15). The abundances of oceanic diatoms in the northern margin of the Perturbed Area were twice as high as those in the cores in the stream axis of the Kuroshio Extension because of the occurrence of regions of localized abundant diatoms associated with mixing of Oyashio Intrusions and warm-water eddies along the Kuroshio Extension that developed in the northern margin area. The periodicities of 41 ka recognized in the variability of the oceanic diatoms correspond to the obliquity (tilt) period, while those at a 23-ka periodicity correspond to the precession band. Vertical and lateral mixing between cold and warm waters associated with glacioeustatic sea level rise occurring since the last glacial period is likely to have the primary production of oceanic diatoms. Fluctuations of oceanic diatom abundances in core MR97-04-1 from the outer margin of the Tsugaru Warm Current show a reversed saw-tooth pattern composed of lower

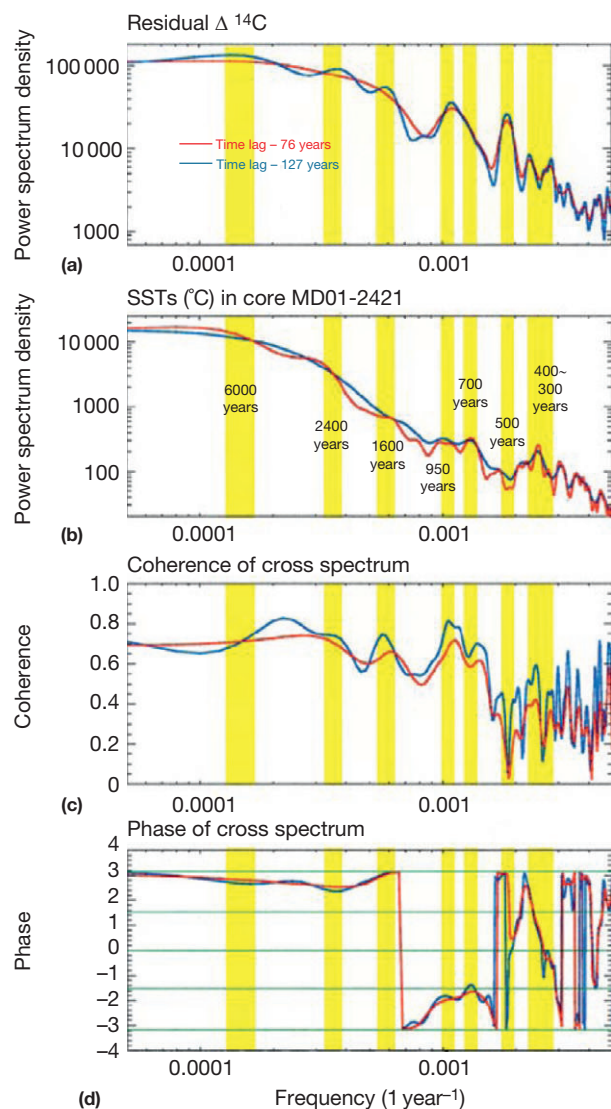


Figure 14 Graphs showing the cross spectrum between the Td' -derived SSTs ($^{\circ}\text{C}$) in core MD01-2421 and the residual $\Delta^{14}\text{C}$ (‰) by the Blackman–Tukey method during the last 13000 years (Koizumi and Sakamoto, 2010). (a) Power spectrum of the residual $\Delta^{14}\text{C}$ (‰) (Stuiver et al., 1991). (b) Power spectrum of the Td' -derived SSTs ($^{\circ}\text{C}$) in core MD01-2421. (c) Coherence of cross spectrum. (d) Phase of cross spectrum. Reproduced from Koizumi I and Sakamoto T (2010) Synchronous Td' -derived SSTs ($^{\circ}\text{C}$) off Japan with climatic events in the Northern Hemisphere. *Journal of Geography* 119: 489–509 (in Japanese with English abstract).

abundance in shorter cycles and higher values in longer cycles at 132, 100, 70, and 30 ka (Figure 16(a)). The wavelet analysis of oceanic diatom abundances reveals low frequencies at 96–112, 44–64, and 24 ka periods (Figure 16(b)). The higher frequencies at ~ 2.0 , 2–5.6, and 3.2–9.6 ka periods have been recognized over the last 150 000 years.

High Td' -derived paleo-SSTs ($^{\circ}\text{C}$) in middle MIS 5e (Marine Isotope Event 5.5) and MIS stage 1 of the interglacial phase have been recognized in cores in the Tohoku Area, off northeast Japan (Figure 17). SSTs ($^{\circ}\text{C}$) in MIS 5e of the

last interglacial period were $\sim 5^{\circ}\text{C}$ higher than modern values from near-shore cores, but in the off-shore cores the differences are only up to $\sim 1.5^{\circ}\text{C}$. A remarkable decline of the Td' -derived paleo-SSTs ($^{\circ}\text{C}$) has also been recognized in seven cores also corresponding to the Marine Isotope Event 6.0 at the MIS 6/5 boundary, Event 5.2 in MIS 5b (Martinson et al., 1987), and to the ages of the Heinrich events 1–6 in the northern North Atlantic Ocean (Heinrich, 1988).

Fluctuations in the Td' -derived SSTs ($^{\circ}\text{C}$) over 150 000 years show a reversed saw-tooth pattern with the duration of ~ 30 ka in each saw-tooth in the southern core MD01-2421, but a shape of a saw-tooth at ~ 40 ka periods in the northern core MR97-04-1 (Figure 18(a)). The Td' -SSTs ($^{\circ}\text{C}$) records fluctuate at 118-, 60-, 30-, and 23-ka periods (Koizumi and Yamamoto, 2008, 2010). The lower frequencies in 96–112 and 7.2–12.8 ka periods correspond to Bond cycles (Figure 18(b); Alley, 1998). A periodicity ~ 32 ka corresponds to the periodicity of the long-term variation of the ENSO (Clement et al., 1999). The shorter period of 1.4–2.1 ka corresponds to Heinrich events (Heinrich, 1988), Dansgaard–Oeschger cycles (Alley, 1998), the fluctuations of residual $\Delta^{14}\text{C}$ record with variance on ~ 2 ka (Stuiver et al., 1991), and the ~ 2 ka cycle in the variability of ENSO activity (Moy et al., 2002).

The Japan Sea

Diatom abundances in the southern core MD01-2407 were about two times greater on average than those in the northern core ODP Site 797 over the last 150 000 years (Koizumi and Yamamoto, 2011). Core KT94-15-5, located in the northern area of the Subarctic Boundary, showed almost negligible abundances except during MIS 5e, 2, and 1.

Td' -derived paleo-SSTs ($^{\circ}\text{C}$) in core MD01-2407 were $\sim 2^{\circ}\text{C}$ higher than the mean value 15.6°C during ~ 10 ka long intervals in the last interstadial phase MIS 5e and modern interglacial phase MIS 1, and decreased during the period from the middle part of MIS 5e to the middle part of MIS 2 (Figure 19). Diatom-based SSTs ($^{\circ}\text{C}$) in cores ODP Site 797 and KT94-15-5 indicate fluctuations similar to those recognized during the last and modern interstadial phases in core MD01-2407, but the SSTs ($^{\circ}\text{C}$) in the stadial phase could not be reconstructed as very few marker diatoms were found.

Td' -derived SSTs ($^{\circ}\text{C}$) in core MD01-2407 fluctuated at the 48-ka periods between 160–90 and 45–0 ka due to the orbital-obliquity (tilt) cycles, and at 23 ka periods, orbital precession cycles, during the 140–100 ka interval. The predominant periodicities of 2- and 4-ka periods occurred at intervals of 20 and 40 ka over the past 160 000 years, respectively (Koizumi and Yamamoto, 2011).

Mega-Floods During the LGM in Northeast Pacific Ocean

The freshwater diatom abundances in marine sediments off northern California have been used to infer massive discharges of freshwater from the Columbia River and low sea-surface

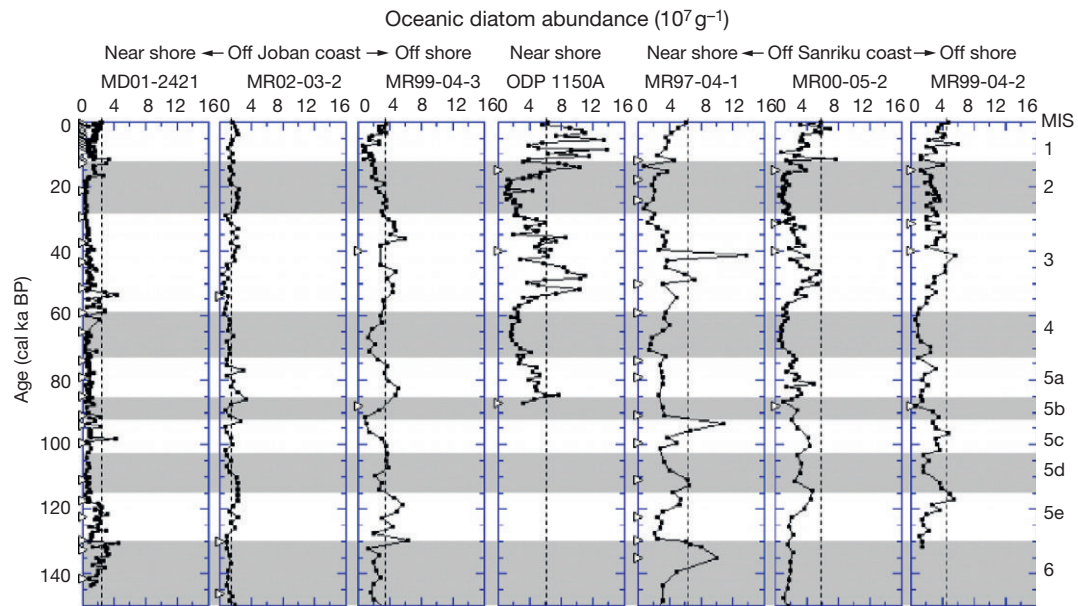


Figure 15 Graphs showing the spatial and temporal variations of the oceanic diatom abundances (10^7 g^{-1} of dried sediment) in seven cores in the Tohoku Area, northwest Pacific (Koizumi and Yamamoto, 2010). Diatom abundances increase in the interglacial phase, and in the cores recovered from near-shore off the Sanriku coast. The fluctuations show changes from lower abundance variations at shorter cycles to higher values in longer cycles at about 30 ka periods (see Figure 16(a)). Reproduced from Koizumi I and Yamamoto H (2010) Paleoclimatological evolution of North Pacific surface water off Japan during the past 150,000 years. *Marine Micropaleontology* 74: 108–118.

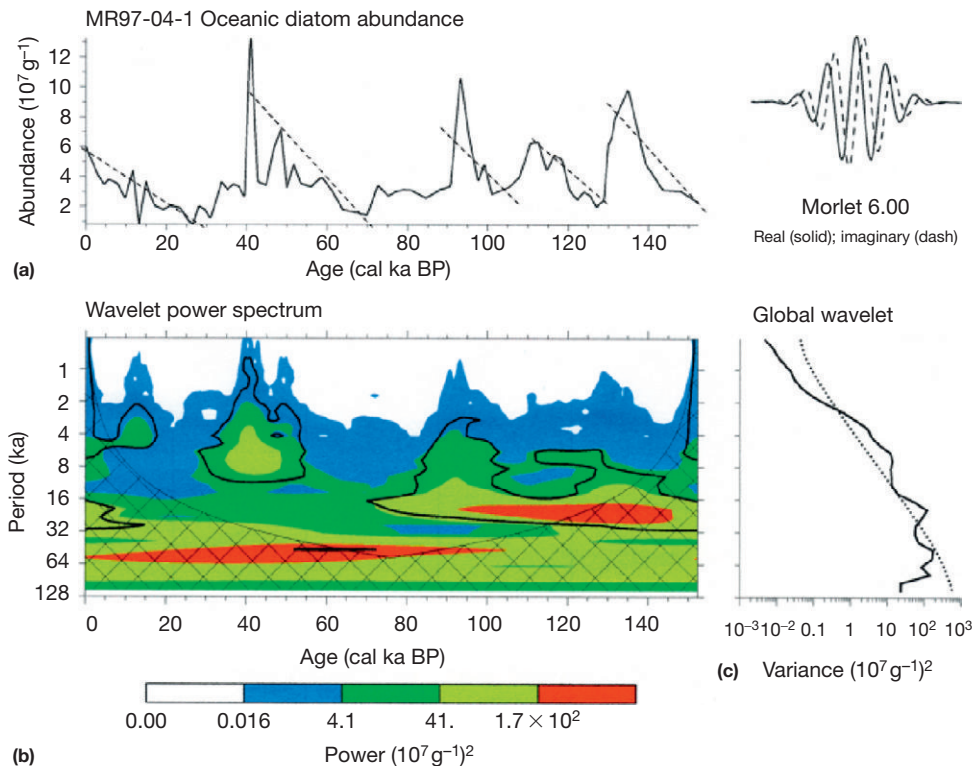


Figure 16 Graphs showing the wavelet analysis for the values of the band ratio oceanic diatom abundance in core MR97-04-1 (Koizumi and Yamamoto, 2010). (a) Time series. (b) Wavelet power spectrum. Solid black line indicates the 10% confidence level (power significance). The contour levels are chosen for 75%, 50%, 25%, and 5% of the wavelet power. More intense colors indicate higher power. The crosshatched region is the cone of influence, where zero padding has reduced the variance. (c) The global wavelet power spectrum (black line). The dashed line indicates the significance for the global wavelet spectrum, assuming the same significance level and background as in (b). The wavelet software provided by Torrence and Compo (1998) is available at URL: <http://paos.colorado.edu/research/wavelet/>. Reproduced from Koizumi I and Yamamoto H (2010) Paleoclimatological evolution of North Pacific surface water off Japan during the past 150 000 years. *Marine Micropaleontology* 74: 108–118.

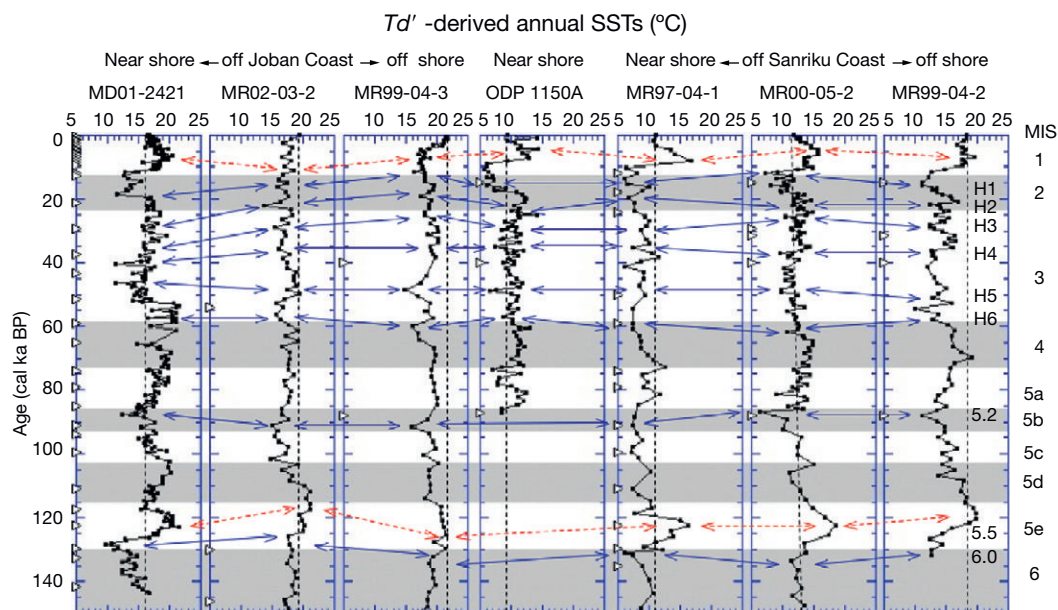


Figure 17 Graphs showing the spatial and temporal variations of the Td' -derived SSTs ($^{\circ}\text{C}$) in seven cores in the Tohoku Area, northwest Pacific Ocean (Koizumi and Yamamoto, 2010). The high Td' -derived SSTs ($^{\circ}\text{C}$) in middle MIS 5e (Marine Isotope Event 5.5) and MIS 1 of the interglacial phase are correlated among cores (red dotted lines). The remarkable decline of SSTs ($^{\circ}\text{C}$) recognized in seven cores correspond to the ages of the Heinrich events 1–6 in the northern North Atlantic (Heinrich, 1988) and for the Marine Isotope Event 5.2 in MIS 5b and 6.0 in MIS 6 (blue lines; Martinson et al., 1987). White triangles on the left side of each column indicate the age-control points for age model. Reproduced from Koizumi I and Yamamoto H (2010) Paleocceanographic evolution of North Pacific surface water off Japan during the past 150,000 years. *Marine Micropaleontology* 74: 108–118.

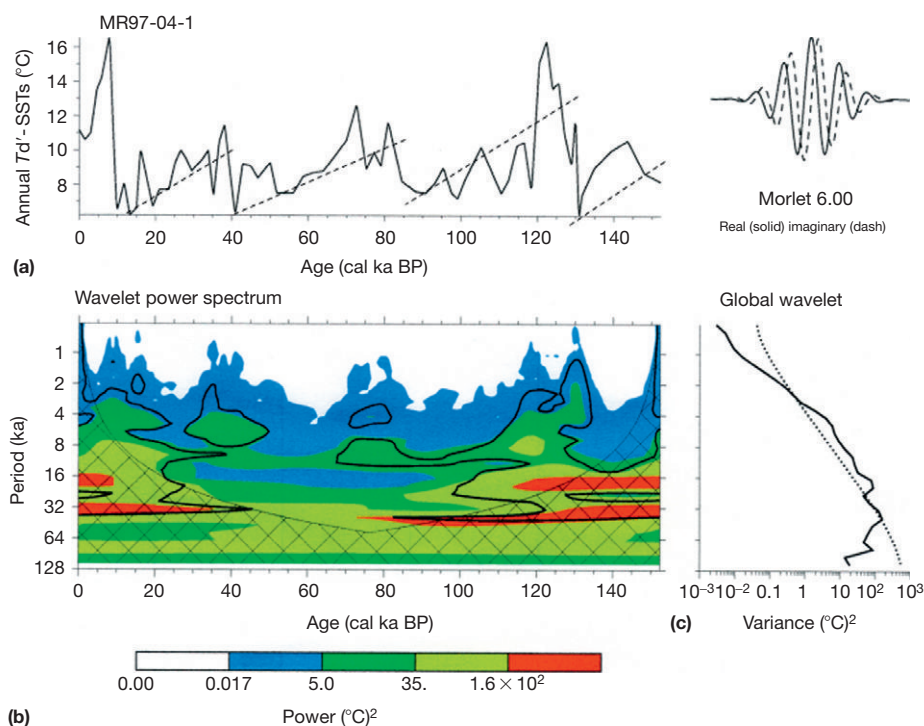


Figure 18 Graphs showing the wavelet analysis for the values of the band ratio Td' -derived SSTs ($^{\circ}\text{C}$) in core MR97-04-1 (Koizumi and Yamamoto, 2010). (a) Time series. The fluctuations indicate the shape of saw-tooth at about 40 ka periods since 150,000 years. (b) Wavelet power spectrum. Solid black line indicates the 10% confidence level (power significance). The contour levels are chosen for 75%, 50%, 25%, and 5% of the wavelet power. More intense colors indicate higher power. The crosshatched region is the cone of influence, where zero padding has reduced the variance. (c) The global wavelet power spectrum (black line). The dashed line indicates the significance for the global wavelet spectrum, assuming the same significance level and background as in (b). Reproduced from Koizumi I and Yamamoto H (2010) Paleocceanographic evolution of North Pacific surface water off Japan during the past 150,000 years. *Marine Micropaleontology* 74: 108–118.

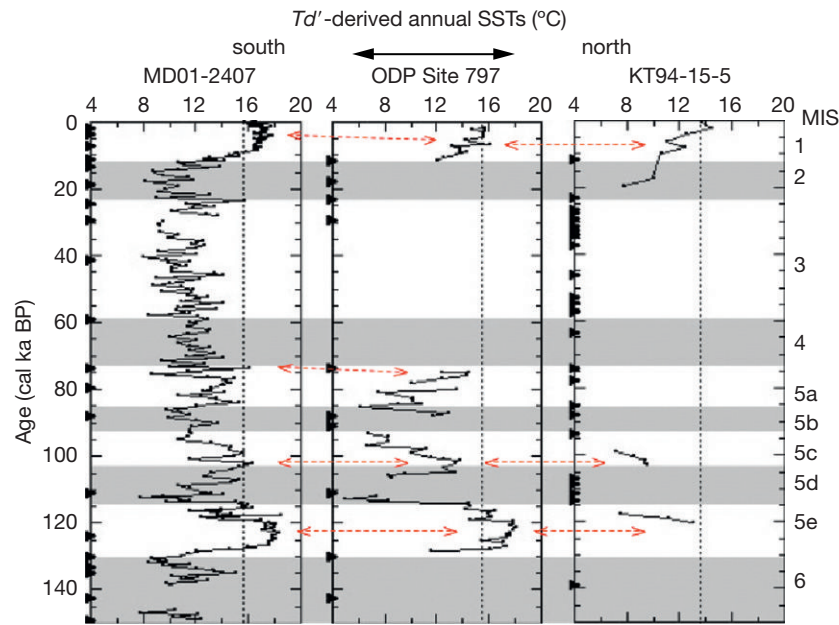


Figure 19 Graphs showing the spatial and temporal variations of the Td' -derived SSTs ($^{\circ}\text{C}$) in three cores over the last 150,000 years in the Japan Sea, northwest Pacific Ocean (Koizumi and Yamamoto, 2011). Td' -derived SSTs ($^{\circ}\text{C}$) in cores ODP Site 797 and KT94-15-5 cannot be reconstructed due to the absence of marker diatom species. The high Td' -derived annual SSTs ($^{\circ}\text{C}$) in middle MIS 5e, base 5c and 1 of the warm-climate phase are correlated between cores (red dotted lines). Broken lines indicate the SSTs ($^{\circ}\text{C}$) in the surface sediment of each core. Black triangles on the left side of each column indicate the age-control points for the age model. Reproduced from Koizumi I and Yamamoto H (2011) Oceanographic variations over the last 150,000 yr in the Japan Sea and synchronous Holocene with the Northern Hemisphere. *Journal of Asian Earth Sciences* 40: 1203–1213.

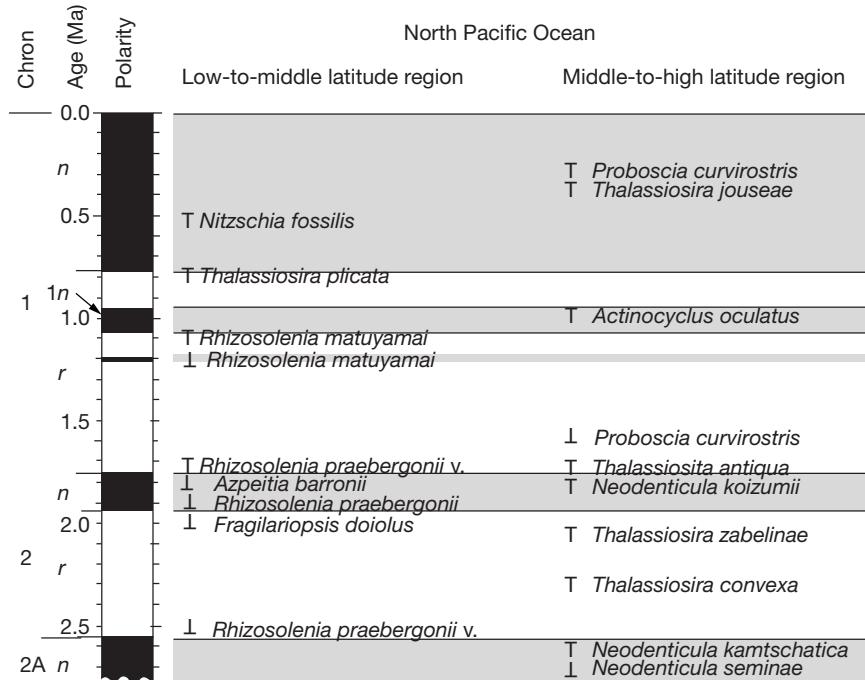


Figure 20 Graph showing the biostratigraphic Quaternary diatom datums in the low-to-middle latitude and middle-to-high latitude regions in the North Pacific Ocean compared to the magnetostratigraphy of Cande and Kent (1995). $\perp$, first occurrence; $\top$, last occurrence.

salinity anomalies more than 400 km away near 46°N between 31 and 16 ka (Lopes and Mix, 2009).

Sea-surface salinity (in psu) was estimated using the formula, $S = 32.323 - 0.1231 \times \text{FD}$ ($r^2 = 0.8$), where FD is

freshwater diatom-relative abundances to total diatoms in 41 core tops. The freshwater diatoms abundantly occurred at 30.5, 27, 23, 20, and 17.5 ka during most of MIS 2 and constituted more than 40% of assemblages in core MD02-2499 off

northern California. The freshwater diatom assemblages were stratigraphically different; in the events older than 25 ka, benthic freshwater diatoms were mainly *Surirella linearis* W. Smith which occurred with relatively high percentages, consistent with a source dominated by either shallow lakes or running water, whereas in the younger events, planktonic freshwater diatoms dominated, including *Aulacoseira granulata* (Ehrenberg) Simonsen, *A. islandica* (O. Muller) Simonsen, *Cyclotella ocellata* Pantocsek and *C. compta* (Ehrenberg) Kützing, which are common in large lakes. This suggests different freshwater sources and/or flooding mechanisms at different times (Lopes and Mix, 2009).

Biostratigraphy

Datum Levels

Deep-sea drilling in the North Pacific Ocean has recovered complete Quaternary marine sequences by drilling multiple holes at a single site, offering excellent material for the study of diatom biostratigraphy and paleoceanography. Significant progress has been made on Quaternary diatom datum levels (biohorizons) that are directly tied to magnetic reversals. Within the low-to-middle latitudes of the North Pacific Ocean, the FO and last occurrence (LO) of *R. praebergonii* Muknina var. *robusta* Burckle and Trainer are placed at 2.5 and 1.7 Ma, the FO of *F. doliolus* at 2.0 Ma, the LO of *R. praebergonii* Muknina at 1.9 Ma, the FO and LO of *R. matuyamai* Burckle at 1.2 and 1.1 Ma, the LO of *Thalassiosira plicata* Schrader at 0.8 Ma, and the LO of *Nitzschia fossilis* (Frenquelli) Kanaya ex Schrader at 0.5 Ma (Figure 20). Within the middle-to-high latitudes, the FO of *N. seminae* is estimated at 2.7 Ma, the LO of *N. kamtschatica* (Zabelina) Akiba and Yanagisawa at 2.6 Ma, the LO of *Thalassiosira convexa* Mukhina at 2.3 Ma, the LO of *Thalassiosira zabelinae* Jousé at 2.1 Ma, the LO of *N. koizumii* Akiba and Yanagisawa at 1.8 Ma, the LO of *Thalassiosira antiqua* (Grunow) Cleve-Euler at 1.7 Ma, the FO and LO of *Proboscia curvirostris* (Jousé) Jordan and Priddle at 1.6 and 0.3 Ma, the LO of *A. oculatus* Jousé at 1.0 Ma, and the LO of *Thalassiosira jouseae* Akiba at 0.3 Ma (Barron, 2003; Koizumi, 2009). Figure 21 illustrates the diagnostic diatom species.

Phylogenetic Evolution

Two events in the phylogenetic evolution of the *Thalassiosira oestrupii* (Osterfeld) Proshkina-Lavrenko subgroup have been related to warming phases during the intervals of 3.4–3.2 and 2.55–2.2 Ma, indicating the southern shifts of the Subarctic Front in the northwest Pacific Ocean (Figure 22; Shiono and Koizumi, 2000). The five members of the *Thalassiosira oestrupii* subgroup have a central fuloportula (a tubular process of some centric diatoms, usually associated with chitin secretion) with three satellite pores and a trifultate structure (Shiono and Koizumi, 2000).

In this subgroup, *Thalassiosira praeoestrupii* Dumont, Baldauf and Barron is distinguished from other members in that it possesses an underdeveloped trifultate fuloportula

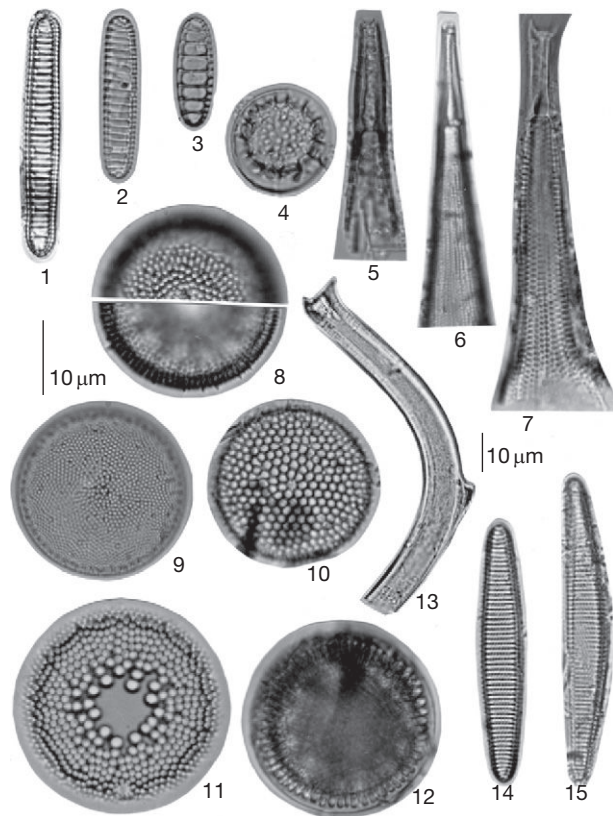


Figure 21 Age diagnostic Quaternary diatom species. (1) *Neodenticula seminae* (Simonsen and Kanaya) Akiba and Yanagisawa, DSDP 580-1-3 (2–3 cm); (2) *Neodenticula koizumii* Akiba and Yanagisawa, DSDP 433A-2-1 (27–29 cm); (3) *Neodenticula kamtschatica* (Zabelina) Akiba and Yanagisawa, DSDP 433A-5-5 (8–10 cm); (4) *Thalassiosira jouseae* Akiba, DSDP 579A-2-2 (13–15 cm); (5) *Rhizosolenia matuyamai* Burckle, DSDP 579A-5-5 (25–27 cm); (6) *Rhizosolenia praebergonii* Muknina, DSDP 579A-12-5 (16–18 cm); (7) *Rhizosolenia praebergonii* Muknina var. *robusta* Burckle and Trainer, DSDP 579A-7-1 (23–25 cm); (8) *Thalassiosira convexa* Mukhina, DSDP 433A-2CC; (9) *Thalassiosira plicata* Schrader, DSDP 433A-3-4 (18–20 cm); (10) *Thalassiosira antiqua* (Grunow) Cleve-Euler, DSDP 579A-8-2 (15–17 cm); (11) *Actinocyclus oculatus* Jousé, DSDP 433A-1-1 (66–68 cm); (12) *Thalassiosira zabelinae* Jousé, DSDP 579A-14-2 (2–4 cm); (13) *Proboscia curvirostris* (Jousé) Jordan and Priddle, DSDP 193-2CC; (14) *Nitzschia fossilis* (Frenquelli) Kanaya ex Schrader, DSDP 580-10-5 (110–112 cm); (15) *Fragilariopsis doliolus* (Wallich) Medlin and Sims, GH76-2, P82 (61 cm).

accompanied with operculate basal structure (Shiono and Koizumi, 2001). *Thalassiosira praeoestrupii* is inferred to be the ancestor species of the subgroup because *Thalassiosira praeoestrupii* occurred at about 4.8 Ma. *Thalassiosira praeoestrupii* forma *juvenis* Shiono and *Thalassiosira trifulta*, which have a complete trifultate fuloportula without the operculate basal structure, evolved from *Thalassiosira praeoestrupii* and differentiated from them during the interval of 3.4–3.2 Ma (Figure 22). *Thalassiosira oestrupii* forma *vetus* Shiono is differentiated from *Thalassiosira trifulta* at 2.55 Ma with a longer distance between the fuloportulae on the margin. *Thalassiosira oestrupii* sensu stricta diverged from *Thalassiosira oestrupii* f. *vetus* at 2.2 Ma.

The Subarctic Front, which is a mixed zone of subarctic and central water masses, occurs at 40–42° N of the North Pacific

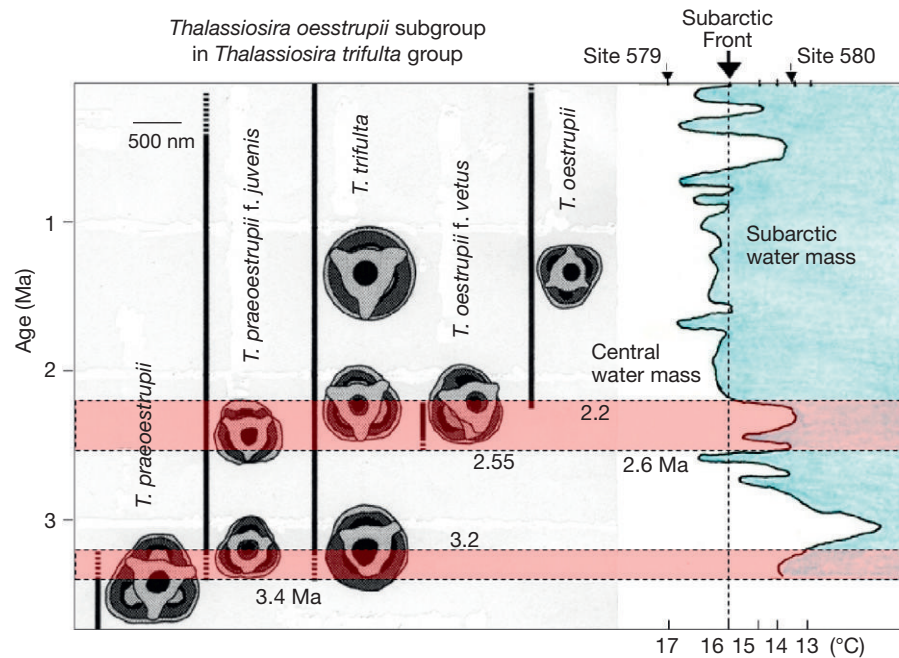


Figure 22 Graph showing the phylogenetic evolution of the *Thalassiosira oestrupii* (Osterfeld) Proshkina-Lavrenko subgroup are compared to the shift of the Subarctic Front and fluctuation of the surface water temperature (°C) in the northwest Pacific Ocean (Koizumi, 1986; Shiono and Koizumi, 2001). The significant glaciations in mid-northern latitudes at 2.6 Ma as the base of the Quaternary Period are inserted between the intervals of 3.4–3.2 and 2.55–2.2 Ma corresponding to the warm-climate phase.

Ocean (Figure 1). Temporal and spatial shifts of the Subarctic Front have been reconstructed based on diatom paleoclimatic data at DSDP Sites 579 and 580 in the middle latitude of the northwest Pacific Ocean (Figure 22; Barron, 1992; Koizumi, 1985, 1986). The interval centered around 3.0 Ma was an extremely warm interval and the Subarctic Front shifted further to the north than its present location prior to the onset of significant Northern Hemisphere glaciation at 2.6 Ma, the newly recognized base of the Quaternary Period (Head et al., 2008). During the interval between 2.8 and 2.4 Ma, diatom assemblages fluctuated, with cold peaks dated at 2.75, 2.6, and 2.45 Ma indicating the drastic north-to-south shift of the Front. Another warm interval is recognized about 2.3 Ma until 2.45 Ma. Diatom accumulation rates declined sharply due to the increase of terrigenous components, resulting from rapid cooling and the global fall of sea level after about 2.7 Ma in the northeast Pacific Ocean (Barron, 1998; Maruyama, 2000). High diatom productivity shifted toward the northwest Pacific Ocean and became higher by enhanced upwelling along the coastal regions. During the interval of 2.7–2.4 Ma, the diatom assemblages dominated by *Coscinodiscus marginatus* Ehrenberg and *N. kamtschatica* rapidly and drastically shifted to one rich in *Neodenticula* complex, *N. koizumi* + *N. seminae*, due to changes in paleoceanographic conditions in the subarctic region (Shimada et al., 2009).

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See also: Diatom Introduction; Structures and Applications of Biomarkers from Arctic Sea Ice Diatoms. **Diatom Records:** North Atlantic and Arctic. **Paleoclimate:** Timescales of Climate Change.

References

- Alley RB (1998) Icing the North Atlantic. *Nature* 392: 335–337.
- Barron JA (1992) Paleoenvironmental and tectonic controls on the Pliocene diatom record of the northeastern Pacific. *Pacific Neogene-Environment, Evolution, and Events*, pp. 25–41. Tokyo: Tokyo University Press.
- Barron JA (1998) Late Neogene changes in diatom sedimentation in the North Pacific. *Journal of Asian Earth Sciences* 16: 85–95.
- Barron JA (2003) Planktonic marine diatom record of the past 18 m.y.: Appearances and extinctions in the Pacific and southern oceans. *Diatom Research* 18: 203–224.
- Barron JA and Anderson L (2010) Enhanced late Holocene ENSO/PDO expression along the margins of the eastern North Pacific. *Quaternary International* 235: 3–12. <http://dx.doi.org/10.1016/j.quaint.2010.02.026>.
- Barron JA, Bukry D, and Field D (2010) Santa Barbara Basin diatom and silicoflagellate response to global climate anomalies during the past 2200 years. *Quaternary International* 215: 34–44.
- Barron JA, Heusser L, Herbert T, and Lyle M (2003) High resolution climatic evolution of coastal northern California during the past 16,000 years. *Paleoceanography* 18: 20–1–20–7, 1020. <http://dx.doi.org/10.1029/2002PA000768>.
- Berger AL (1978) Long-term variations of caloric insolation resulting from the Earth's orbital elements. *Quaternary Research* 9: 139–167.
- Bond G, Kromer B, Beer J, et al. (2001) Persistent solar influence on North Atlantic climate during the Holocene epoch. *Nature* 419: 821–824.
- Caissie BE, Brigham-Grette J, Lawrence KT, and Herbert TD (2010) Last Glacial Maximum to Holocene sea surface conditions at Umnak, Bering Sea, as inferred from diatom, alkenone, and stable isotope records. *Paleoceanography* 25: PA1206. <http://dx.doi.org/10.1029/2008PA001671>.
- Cande SC and Kent DV (1995) Revised calibration of geomagnetic polarity time scale for the Late Cretaceous and Cenozoic. *Journal of Geophysical Research* 100: 6093–6095.
- Clement AC, Seager R, and Cane MA (1999) Orbital controls on the El Niño/Southern oscillation and the tropical climate. *Paleoceanography* 14: 441–456.

- D'Arrigo R, Villalba R, and Wiles G (2001) Tree-ring estimates of Pacific decadal climate variability. *Climate Dynamics* 18: 219–224.
- Debet MD, Bout-Roumazeilles V, Crousset F, et al. (2007) The origin of the 1500-year climate cycles in Holocene North-Atlantic records. *Climate of the Past* 3: 569–575.
- Gedalof Z, Mantua NJ, and Peterson DL (2002) A multi-century perspective of variability in the Pacific Decadal Oscillation: New insight from the tree rings and coral. *Geophysical Research Letters* 29: 2204–2207.
- Head MJ, Gibbard P, and Salvador A (2008) The Quaternary: Its character and definition. *Episodes* 31: 234–238.
- Heinrich H (1988) Origin and consequences of cyclic ice rafting in the Northeast Atlantic Ocean during the past 130,000 years. *Quaternary Research* 29: 142–152.
- Isono D, Yamamoto M, Irino T, et al. (2009) The 1500-year climate oscillation in the midlatitude North Pacific during the Holocene. *Geology* 37: 591–594. <http://dx.doi.org/10.1130/G25667A.1>.
- Koizumi I (1985) Late Neogene paleoceanography in the western North Pacific. *Initial Reports of the Deep Sea Drilling Project* 86: 429–438. Washington, DC: US Government Printing Office.
- Koizumi I (1986) Pliocene and Pleistocene diatom datum levels related with paleoceanography in the northwest Pacific. *Marine Micropaleontology* 10: 309–325.
- Koizumi I (2006) Diatom fossil records of the Pacific. *Encyclopedia of Quaternary Sciences*, vol. 1, pp. 576–598. The Netherlands: Elsevier Science.
- Koizumi I (2008) Diatom-derived SSTs (T_d' ratio) indicate warm seas off Japan during the middle Holocene (8.2–3.3 ka BP). *Marine Micropaleontology* 69: 263–281. <http://dx.doi.org/10.1016/j.marmicro.2008.08.004>.
- Koizumi I (2009) Diatom biostratigraphy and chronology during the Quaternary. *Digital Book: Progress in Quaternary Research in Japan*, pp. 1–15. Tokyo: Japan Association for Quaternary Research (in Japanese with English abstract).
- Koizumi I, Irino T, and Oba T (2004) Paleoceanography during the last 150 kyr off central Japan based on diatom floras. *Marine Micropaleontology* 53: 293–365.
- Koizumi I and Sakamoto T (2010) Synchronous T_d' -derived SSTs ($^{\circ}\text{C}$) off Japan with climatic events in the northern hemisphere. *Journal of Geography* 119: 489–509 (in Japanese with English abstract).
- Koizumi I, Shiga K, Irino T, and Ikehara M (2003) Diatom record of the late Holocene in the Okhotsk Sea. *Marine Micropaleontology* 49: 139–156.
- Koizumi I and Yamamoto H (2008) Paleohydrography of the Kuroshio Extension in the Tohoku Area based on fossils diatoms. *JAMSTEC Reports of Research and Development* 7: 1–10.
- Koizumi I and Yamamoto H (2010) Paleoceanographic evolution of North Pacific surface water off Japan during the past 150,000 years. *Marine Micropaleontology* 74: 108–118. <http://dx.doi.org/10.1016/j.marmicro.2010.01.003>.
- Koizumi I and Yamamoto H (2011) Oceanographic variations over the last 150,000 years in the Japan Sea and synchronous Holocene with the Northern Hemisphere. *Journal of Asian Earth Sciences*. 40: 1203–1213. <http://dx.doi.org/10.1016/j.jseas.2010.06.013>.
- Lopes C and Mix AC (2009) Pleistocene megafloods in the northeast Pacific. *Geology* 37: 79–82.
- Lopes C, Mix AC, and Abrantes F (2006) Diatoms in northeast Pacific surface sediments as paleoceanographic proxies. *Marine Micropaleontology* 60: 45–65. <http://dx.doi.org/10.1016/j.marmicro.2006.02.010>.
- Mantua NJ, Hare SR, Zhang Y, Wallace JM, and Francis RC (1997) A Pacific interdecadal climate oscillation with impacts on salmon production. *Bulletin of the American Meteorological Society* 78: 1069–1079.
- Martinson DG, Pisias NG, Hays JD, Imbrie J, Moore TC, and Shackleton NJ (1987) Age dating and the orbital theory of the Ice Ages: Development of a high-resolution 0 to 300,000-year chronostratigraphy. *Quaternary Research* 27: 1–29.
- Maruyama T (2000) Middle Miocene to Pleistocene diatom stratigraphy of Leg 167. *Proceedings of the Ocean Drilling Program, Scientific Results*, vol. 167, pp. 63–110. College Station, TX: Ocean Drilling Program.
- Moy CM, Seltzer GO, Rodbell DT, and Anderson DM (2002) Variability of El Niño/Southern Oscillation activity at millennial timescales during the Holocene epoch. *Nature* 420: 162–165.
- Sancetta C (1981) Oceanographic and ecologic significance of diatoms in surface sediments of the Bering and Okhotsk seas. *Deep-Sea Research Part A* 28: 789–817.
- Sancetta C (1983) Effect of Pleistocene glaciation upon oceanographic characteristics of the North Pacific Ocean and Bering Sea. *Deep-Sea Research Part A* 30: 851–869.
- Shiga K and Koizumi I (2000) Latest Quaternary oceanographic changes in the Okhotsk Sea based on diatom records. *Marine Micropaleontology* 38: 91–117.
- Shimada C, Sato T, Yamasaki M, Hasegawa S, and Tanaka Y (2009) Drastic change in the late Pliocene subarctic Pacific diatom community associated with the onset of the Northern Hemisphere Glaciation. *Palaeogeography, Palaeoclimatology, Palaeoecology* 279: 207–215. <http://dx.doi.org/10.1016/j.palaeo.2009.05.015>.
- Shiono M and Koizumi I (2000) Taxonomy of the *Thalassiosira trifulta* group in late Neogene sediments from the northwest Pacific Ocean. *Diatom Research* 15: 355–382.
- Shiono M and Koizumi I (2001) Phylogenetic evolution of the *Thalassiosira trifulta* group (Bacillariophyceae) in the northwestern Pacific Ocean. *Journal of the Geological Society of Japan* 107: 496–514.
- Stuiver M, Braziunas TF, Becker B, and Kromer B (1991) Climatic, solar, oceanic, and geomagnetic influences on late-glacial and Holocene atmospheric $^{14}\text{C}/^{12}\text{C}$ change. *Quaternary Research* 35: 1–24.
- Torrence C and Compo GP (1998) A practical guide to wavelet analysis. *Bulletin of the American Meteorological Society* 79: 61–78.
- Yamamoto M, Oba T, Shimamura J, and Ueshima T (2004) Orbital-scale anti-phase variation of sea surface temperature in mid-latitude North Pacific margins during the last 145,000 years. *Geological Research Letters* 31: L16311. <http://dx.doi.org/10.1029/2004GL020138>.

Structures and Applications of Biomarkers from Arctic Sea Ice Diatoms

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Introduction

Polar seas are characterized by the presence of multiyear and annually formed sea ice which contributes considerably to the reduction of heat and gas exchange between the atmosphere and the upper layers of the ocean. The enhanced albedo of sea ice relative to the open ocean results in a dramatic reduction of solar radiation at the sea surface, an effect that is further enhanced when the sea ice is snow covered. In either case, the resulting insulation strongly influences changes in climate, both locally and globally. Further, during melting, the outflow of low-salinity surface water affects the climate by influencing deep oceanic circulation. The rapid and significant changes to sea ice extent and thickness in the Arctic and Antarctic in recent decades has stimulated a significant amount of new research into the likely impacts this will have on the future climate of the Earth and the complex ecosystems of the polar regions. At the base of the food chain, sea ice provides a physical support for the development and growth of a productive community of ice algae and a wide diversity of heterotrophic eukaryotes, all of which interact with the underlying surface waters (Caron and Gast, 2010; Róžańska et al., 2009).

Sea Ice Biota

Within sea ice, three distinct habitats can be distinguished mainly according to the position of autotrophic and heterotrophic eukaryotes within the ice matrix: the surface community, the interior community, and the bottom community (Horner et al., 1992) (Note: Prokaryotes are also present in sea ice but these are not discussed here). Surface communities occur at the snow–ice interface resulting from the flooding of the ice surface with seawater, while interior communities correspond to a remnant assemblage of the previous bloom season. The bottom community develops during the spring in the lowest ice layer or at the interface with the ocean. The biota within this bottom ice is further divided into interstitial and sub-ice communities; the latter consisting of algae, such as the Arctic centric diatom *Melosira arctica* Dickie, which floats beneath, or is attached to, the underside of the ice-forming mats and decimeter-long strands (Ambrose et al., 2005). Interstitial communities occur in the bottom 10 cm of annually formed sea ice and are usually associated with small ice crystals, brine pockets, and a well-developed network of channels and capillaries (Arrigo et al., 2010; Horner et al., 1992). This colonization of diatoms in the lower section of annually formed sea ice can be clearly observed in the intense coloration of these layers in cored samples, especially during the spring bloom (Figure 1).

The growth, production, and biomass of the bottom ice algal communities are, in part, controlled by the thickness of sea ice and snow cover, which deeply affect the amount of transmitted light through the ice matrix (Arrigo et al., 2010), and by the availability of macronutrients present in the underlying surface waters (Róžańska et al., 2009). Gosselin et al. (1997) estimated the contribution of sea ice algae to the total primary production to be in the region of 57% in the central Arctic Ocean, while Legendre et al. (1992) reported values between 3% and 25% for the Arctic shelves. Ice algae constitute the main food source for ice-associated and pelagic herbivorous protists (Michel et al., 2002) and metazoans (Nozais et al., 2001), and contribute significantly to carbon cycling throughout the Arctic regions (Michel et al., 2006). Ice algal blooms usually occur before the phytoplankton bloom (Arrigo et al., 2010).

Several names have been used to describe organisms living in close association with the sea ice, but Horner et al. (1992) proposed the term *sympagic community* to describe the group of autotrophic and heterotrophic unicellular eukaryotes that inhabit the ice. In terms of abundance, the main groups of bottom ice eukaryotes are diatoms, flagellates, and dinoflagellates. While diatoms are mainly photosynthetic, a large fraction of the flagellated community, including some dinoflagellates, can be heterotrophic (i.e., feeding on other organisms). Arctic bottom ice algal communities are mainly represented by colonial and solitary pennate diatoms during the bloom period (Róžańska et al., 2009). Among the colonial pennate diatoms, the most abundant and frequently reported species belong to the genera *Entomoneis* Ehrenberg, *Fragilariopsis* Hustedt, *Navicula* Bory, and *Nitzschia* Hassall, while *Nitzschia frigida* Grunow can be considered a key species of bottom ice across all the Arctic regions (Poulin et al., 2011; Róžańska et al., 2009). Solitary pennate diatoms are mainly represented by the genus *Navicula*, followed by a few species of *Nitzschia* and *Pinnularia* Ehrenberg (Poulin, 1990; Róžańska et al., 2009). *Haslea* Simonsen also occurs in low abundance (e.g. *Haslea crucigeroides* (Hustedt) Simonsen, *H. kjellmanii* (Cleve) Simonsen, *H. spicula* (Hickie) Lange-Bertalot and *H. vitrea* (Cleve) Simonsen).

There are approximately 25 recognized species of *Haslea*, mostly reported from tropical and temperate regions, with a few in polar environments. The genus is mainly benthic and mostly present on hard substrates, except for the pelagic *H. wawriakae* (Hustedt) Simonsen. Six species have been reported, though in low abundance, from annually formed sea ice in both the Arctic (*H. crucigeroides*, *H. kjellmanii*, *H. spicula*, *H. vitrea*) (Figure 2) and the Antarctic (*H. trompii* (Cleve) Simonsen, *H. major* (Heiden) Simonsen).

Apart from species identification, investigations into bottom ice eukaryotes have been carried out for a number of

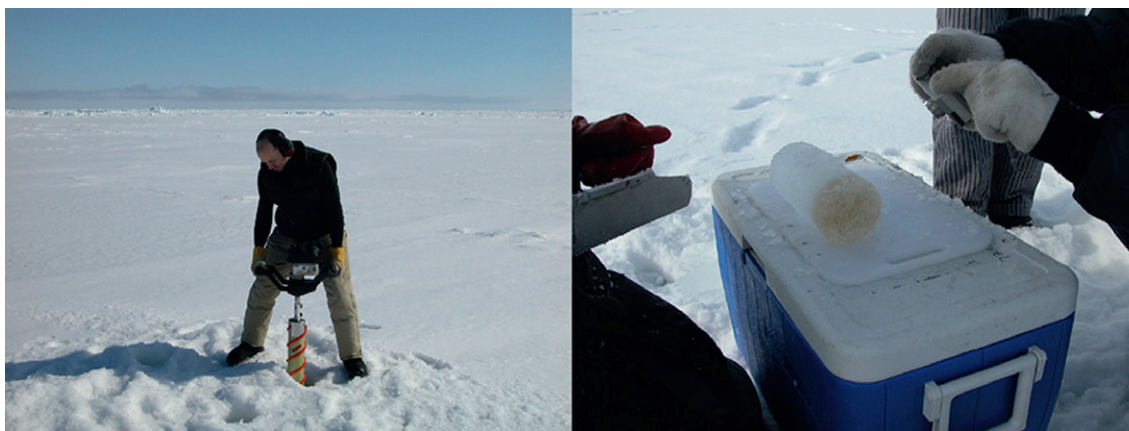


Figure 1 Examples of sea ice coring (left) and bottom ice sections (right) showing the intense brown coloration resulting from the spring diatom bloom, Hudson Bay, Canada, May 2004 (images: M. Poulin).

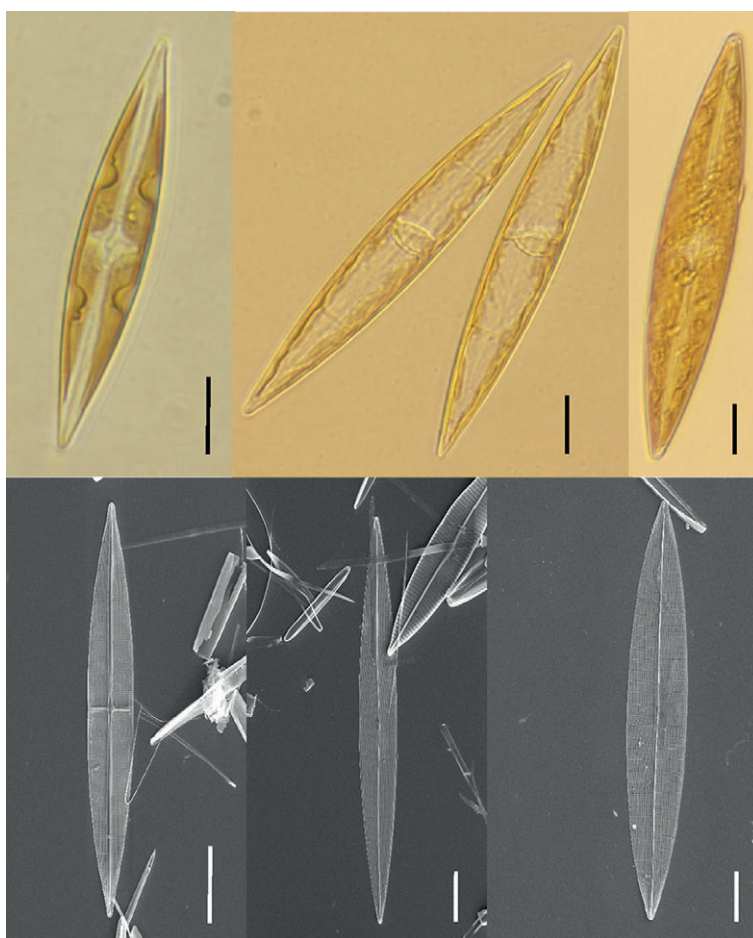


Figure 2 Light (top) and scanning electron micrographs (bottom) of selected Arctic sea ice diatoms. Left: *Haslea crucigeroides*; Middle: *Haslea vitrea*; Right: *Haslea kjellmanii*. Scale bars = 20 μm .

reasons, including the determination of species distributions across the Arctic (Poulin et al., 2011) and Antarctic, an assessment of the role of sea ice algae within polar food webs and the influence of environmental factors on species

diversity and abundance. In the latter case, the potential for rapid climate change to substantially influence sea ice algae production has provided a strong additional motivation for study.

Analysis of Sea Ice Diatoms

Measurements of algal biomass can be carried out through chemical analysis, while the determination of species abundances and distributions clearly relies on identification of specific taxa (Arrigo et al., 2010). Chemical analysis is most routinely performed by spectrophotometric or fluorometric analysis of chlorophyll *a*, largely since pigment analysis is relatively straightforward and, in addition, can provide a reasonable approximation to organic carbon production (Arrigo et al., 2010). This type of proxy analysis of sea ice autotrophic eukaryotes can also be achieved through the identification and quantification of individual biomarkers. Biomarkers are chemicals that are produced by all biota over a wide range of taxonomic levels. They can represent general classes of chemicals (e.g., lipids and pigments) or, alternatively, correspond to individual molecules. Further, they may be primary or secondary metabolites and are generally identified following isolation from their sources and analysis using a combination of purification and spectroscopic methods such as chromatography and mass spectrometry or nuclear magnetic resonance (NMR) spectroscopy, respectively. Lipids such as fatty acids and sterols are amongst the most common biomarkers to investigate, since they are relatively abundant, common to most organisms and straightforward to analyze and identify.

Diatom Lipids

Fatty acids

Saturated and unsaturated fatty acids are important storage lipids in diatoms and they also help maintain cell fluidity. Cellular concentrations of fatty acids are generally high compared to other lipid classes, whilst in marine diatoms, the saturated compounds are generally more abundant than their polyunsaturated fatty acid counterparts (Volkman et al., 1998). The distinctive chromatographic and mass spectral characteristics of fatty acids, combined with their relatively high abundances, makes them attractive biomarkers to analyze; although the ubiquity of these lipids can be limiting, particularly if source identification is required. Despite this, some broad classifications of individual fatty acids have been used in the identification of source organisms, with studies carried out on single species and mixed assemblages of phytoplankton and ice algae showing that C₁₄, C_{16:1ω7}, C₁₆, C₁₈, and C_{20:5ω3} fatty acids are representative of marine species (e.g., Volkman et al., 1998). Discriminating between the origin of a sea ice and planktonic diatom for these compounds is not really possible, although changes in fatty acid distributions have been used to infer events such as algal blooms (e.g., Reuss and Poulsen, 2002).

Sterols

A second group of biomarkers that are commonly used to monitor primary productivity in the ocean and sea ice are sterols (Volkman et al., 1998). Phytosterols are abundant and important structural constituents of the cell membranes of terrestrial and marine plants (Barenholz, 2002), including microalgae such as diatoms, which mainly contain sterols with between 27 and 29 carbon atoms. In an analysis of 14

species of marine diatoms, a range of sterol distributions was observed, although none were considered to be specific to diatoms (Barrett et al., 1995). Although the nonspecific nature of sterols limits detailed determination of source organisms, the analysis of sterols, like that of fatty acids, can still be useful with respect to the measurement of primary productivity.

Highly branched isoprenoid alkenes

A third class of biomarker chemical produced by some diatoms are simple nonfunctionalized lipids known as highly branched isoprenoid (HBI) alkenes which, unlike fatty acids and sterols, are restricted in terms of their sources, making them highly suitable chemicals to be analyzed as biomarkers. HBIs were first reported in the 1970s following isolation from sediments worldwide, although the precise structures of these chemicals were unknown then. As part of a wider investigation into diatom lipids, Volkman et al. (1994) were the first to show that HBIs are made by diatoms, with C₂₅ and C₃₀ HBIs produced by *H. ostrearia* (Gaillon) Simonsen and *Rhizosolenia setigera* Brightwell, respectively. Since this initial identification, a number of further HBI-producing diatoms have been identified, including other *Haslea* spp., *Navicula* spp., and *Pleurosigma* spp. (e.g., Belt et al., 2000; Massé, 2003). The structures of numerous HBI alkenes have been determined through large-scale culturing of individual diatom species, bulk lipid extraction, purification of individual compounds and analysis by ¹H and ¹³C NMR spectroscopy (e.g., Belt et al., 1996, 2000; Massé, 2003). As a result, more than 20 individual structures of C₂₅ and C₃₀ HBI alkenes have been reported, together with some related monocyclic compounds (Massé, 2003). C₂₅ and C₃₀ HBIs occur with varying degrees of unsaturation and in different isomeric forms (see Figure 3 for typical HBI structures and numbering scheme used throughout the remainder of the text).

The majority of C₂₅ HBIs isolated from diatoms range in unsaturation from 2 to 5, although more polyunsaturated compounds have also been reported. The positions and stereochemistries of the double bonds are controlled by the biosynthetic mechanisms responsible for their formation and each of these structural features exhibits some variability. For example, while a double bond at the terminal (vinyl) position is present in virtually all C₂₅ and C₃₀ HBI alkenes, the variability in the positions of the other double bonds leads to regioisomerism, with greater unsaturation resulting in the largest number of regioisomers. Thus, while the structures of only two diene isomers (2,3) have been characterized from diatoms, numerous tetraenes (e.g., 7,8) have been reported, following isolation from large-scale cultures. The presence of both unsaturation and chiral centers in C₂₅ and C₃₀ HBIs can potentially result in stereoisomerism in these compounds and this is indeed observed. In most cases, the stereochemistries of the trisubstituted double bonds in HBIs have the *E* configuration as is common for simple biogenic lipids of this type. However, there are exceptions to this general observation, with both C₂₅ and C₃₀ compounds exhibiting *E* and *Z* isomerism from *Pleurosigma intermedium* W. Smith (5,6) and *R. setigera* (10,11), respectively (Belt et al., 2000, 2001). Some variability in the stereochemistries of the chiral or asymmetric centers in HBIs has also been reported although, to date, this kind of investigation has not been carried out for most HBIs. The reasons for the observed

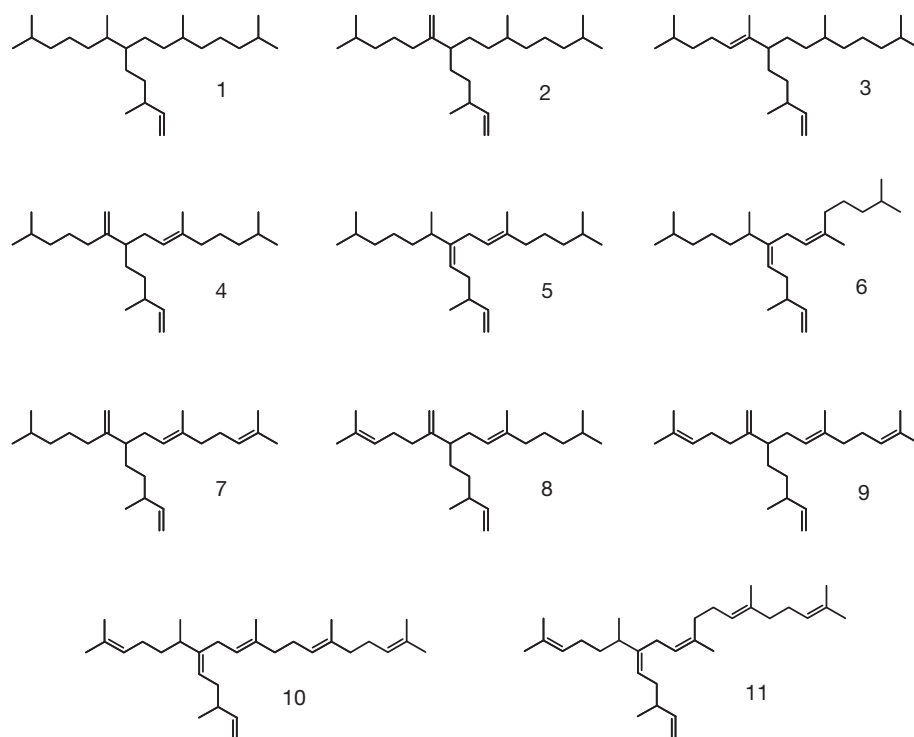


Figure 3 Representative structures and numbering scheme of C_{25} and C_{30} highly branched isoprenoid alkene biomarkers cited in the text.

variability in both regio- and stereoisomerism is, as yet, unknown, but probably relates to the functions of these lipids which are not well understood. What is clear, however, is that both the mevalonate and methylerythritol phosphate isoprenoid biosynthetic mechanisms are responsible for the formation of HBIs, which probably reflects the different locations and roles of these biomarkers in diatom cells (Massé, 2003).

The existence of C_{25} and C_{30} HBIs in different structural and isomeric forms provides a potential association between structure and source, especially as the formation of C_{30} HBIs (and their monocyclic analogs) appears to be restricted to biosynthesis by *R. setigera* (Belt et al., 2001; Volkman et al., 1994). For the C_{25} HBIs, isomers containing double bonds in either of the C5–C6 or C6–C17 positions (e.g., 2–4, 7–9) are common for *Haslea* spp. and *Navicula* spp. (Belt et al., 1996, 2000; Massé, 2003), while *Pleurosigma* spp. are responsible for biosynthesis of isomers with double bonds in the C7–C20 position (e.g. 5,6; Belt et al., 2000). However, these source–structure relationships are not exclusive since, for example, *R. setigera* also produces C_{25} HBI alkenes with C7–C20 double bonds, albeit only at specific intervals within its life cycle (Massé, 2003). It is also clear that variability in distributions of HBIs also exists across different populations of the same species. Several studies have attempted to investigate this, largely by culturing single species under a range of phenotypic conditions including light, salinity, and temperature (Massé, 2003; Rowland et al., 2001). In the case of *H. ostrearia*, salinity was found not to significantly influence either production or distribution of C_{25} HBIs although light intensity was shown to have some control over lipid production, as expected (Massé, 2003).

To date, however, most studies have resulted in few definitive outcomes, but it seems likely that controls on HBI

distributions are not simply limited to growth conditions, but depend on the source organism concerned, as well as other factors. For example, in the case of *R. setigera*, distributions of C_{25} and C_{30} HBIs have been shown to be strongly dependent on the physiological state of the diatoms within their life cycle, with C_{25} HBIs produced mainly during the auxospore phase (Massé, 2003).

The most striking influence on C_{25} HBI distribution in *H. ostrearia* has been shown to be growth temperature (Rowland et al., 2001). In most cultures, *H. ostrearia* produces C_{25} HBIs with unsaturation ranging from 2 to 5. Significantly, when *H. ostrearia* was cultured over a range of temperatures, it was found that the more unsaturated isomers had the greatest abundances at the higher growth temperatures, while the relatively saturated compounds had the greatest concentrations at lower temperatures (Rowland et al., 2001). Thus, at 25 °C, the most abundant isomers were tetraenes and pentaenes, while at 5 °C, only dienes were present. At intermediate temperatures (e.g., 15 °C), trienes were the most abundant HBIs, although all isomers were present to some extent. *H. ostrearia* failed to grow under more extreme conditions (e.g., below 5 °C or above 30 °C) preventing further examination of the temperature influence on isomer distribution.

IP₂₅: A Proxy for Arctic Sea Ice

Structure and Properties of IP₂₅

The clear temperature influence on C_{25} HBI distributions from 5 to 25 °C observed for *H. ostrearia* (Rowland et al., 2001) prompted Belt et al. (2007) to hypothesize that some polar sea ice diatoms may be capable of producing relatively

saturated C₂₅ HBI isomers, especially as some *Haslea* spp. had previously been reported in Arctic and Antarctic diatom communities, even if such species were only present as relatively minor constituents (Poulin, 1990). In addition, it was further suggested that these relatively saturated sea ice-derived isomers could have the potential to serve as proxies for past sea ice occurrence when detected in dated marine sediments in the polar regions, especially as they would probably be more stable in sediments than their polyunsaturated counterparts (Belt et al., 2007).

In order to further test these hypotheses, Belt et al. (2007) sampled and analyzed sea ice from three regions of the Canadian Arctic during the spring algal bloom period. In each case, *Haslea* spp. were detected in the bottom sections of the sea ice cores, together with several C₂₅ HBIs previously reported (Belt et al., 2007). In addition to the presence of a C₂₅ diene (2) as expected, a previously unreported monounsaturated isomer (1) was also detected. The structure of this novel biomarker was established by synthesis from the related diene and characterization by ¹H and ¹³C NMR spectroscopy. Following the structural elucidation of this monounsaturated compound, chromatographic and mass spectral analysis showed it to be different from that of other C₂₅ HBI monoenes previously reported. As a result, the term IP₂₅ (ice proxy with 25 carbon atoms) was proposed for this monounsaturated C₂₅ HBI biomarker (1).

As part of this initial investigation into the potential for IP₂₅ to be used as a proxy for past Arctic sea ice occurrence, Belt et al. (2007) also sampled surface sediment material from the Canadian Arctic Archipelago (CAA) and sub-Arctic regions and found IP₂₅ to be present in all cases, thus providing evidence for efficient transfer of this biomarker from the sea ice to underlying sediments. IP₂₅ was also detected from older sediments radiocarbon dated at ca. 9000 years old and its stable isotopic composition (δ¹³C) was later shown to be consistent with a sea ice origin (Belt et al., 2008).

Paleo Arctic Sea Ice Reconstruction Using IP₂₅

Although a number of proxies have been used to infer past environmental or climatic conditions on Earth, it is clearly important to understand the limitations of individual proxies, in order for their measurements to be fully evaluated and correctly interpreted. In general, proxies rely to a large extent on various forms of calibration, with the added assumption that recent observations are representative of conditions in the past. In the case of the sea ice biomarker proxy IP₂₅, strict calibrations between sedimentary abundances and associated sea ice conditions have not yet been carried out, partly as a result of a paucity of well-defined or accurate recent sea ice conditions, both temporally and spatially. Despite this, some meaningful interpretations of the abundances of IP₂₅ have been made through comparisons of sedimentary concentrations and fluxes with historical records of sea ice in different Arctic regions.

Initially, Massé et al. (2008) determined the concentrations of IP₂₅ in a marine sediment core (MD99–2275) obtained from the North Icelandic Shelf and compared these findings with known sea ice occurrence and other climate conditions documented in historical records covering the last 1000 years.

This study demonstrated a number of strong correlations between the sedimentary concentrations of IP₂₅ and documented climate conditions, particularly during cold intervals, including those in which extensive sea ice cover had been reported. Good agreement between biotic proxies was observed between the sea ice biomarker data (IP₂₅) and an independent sea surface temperature reconstruction using a diatom-based transfer function (Figure 4). Thus, on a centennial scale, sedimentary IP₂₅ concentrations were at their highest during the seventeenth and nineteenth centuries, previously reported as the coldest centuries of the last millennium in the Northern Hemisphere (Jones et al., 2001). At a higher temporal resolution, IP₂₅ abundances were greatest during decades considered to represent either extremely cold intervals (e.g., AD 1690–1700) or intervals when extensive sea ice cover had been well documented. This study by Massé et al. (2008) also provided proxy evidence for sea ice conditions for intervals when historical records were absent. The absence of such records might be explained by few people visiting in periods of severe weather. Further, observations of, consecutively, lower and higher IP₂₅ concentrations during the first and second half of this 1000-year study period were consistent with the occurrence of the well-known Medieval Warm Period (MWP) and Little Ice Age (LIA) epochs.

A subsequent investigation of a second sediment core from the North Icelandic Shelf (MD99–2263), obtained from a more westerly location, revealed very similar outcomes, as would be expected given the position of the Polar Front around northern Iceland during the last millennium. In contrast, and importantly from a calibration point of view, analysis of a sediment core from southwest Iceland showed that IP₂₅ was absent from all sediment horizons, consistent with this location being south of the Polar Front and free of sea ice during the study interval (Andrews et al., 2009).

A second study aimed at calibrating the IP₂₅ proxy was carried out by Vare et al. (2010) who analyzed sediment samples from three locations in the Barents Sea for which historical sea ice records existed for the last few hundred years, principally as a result of old whaling records and more recent observational accounts made through aerial reconnaissance. In this case, study locations were chosen to represent periods of continuous, variable or sea ice-free conditions, especially during the summer. Consistent with the observational records, IP₂₅ abundances and fluxes in cores obtained from the north and southeast regions of the Barents Sea demonstrated periods of frequent sea ice occurrence during the last two centuries, with greater variability for the more southeasterly location. In contrast, for the study location in the southwest Barents Sea, the absence of IP₂₅ from all sedimentary horizons covering the last ca. 300 years was consistent with the failure of sea ice to encroach into this region in recent centuries.

These two sets of proxy-based investigations into sea ice occurrence around Iceland and the Barents Sea during the last millennium have, therefore, provided reasonable evidence that sedimentary abundances of the IP₂₅ biomarker can be used to provide at least qualitative (e.g., presence/absence) paleo-sea ice reconstructions for the Arctic. As a result, new paleo-sea ice investigations are being conducted over timescales that precede historical records and for regions of the Arctic where previous records are incomplete, absent, or for which existing proxy data are contradictory.

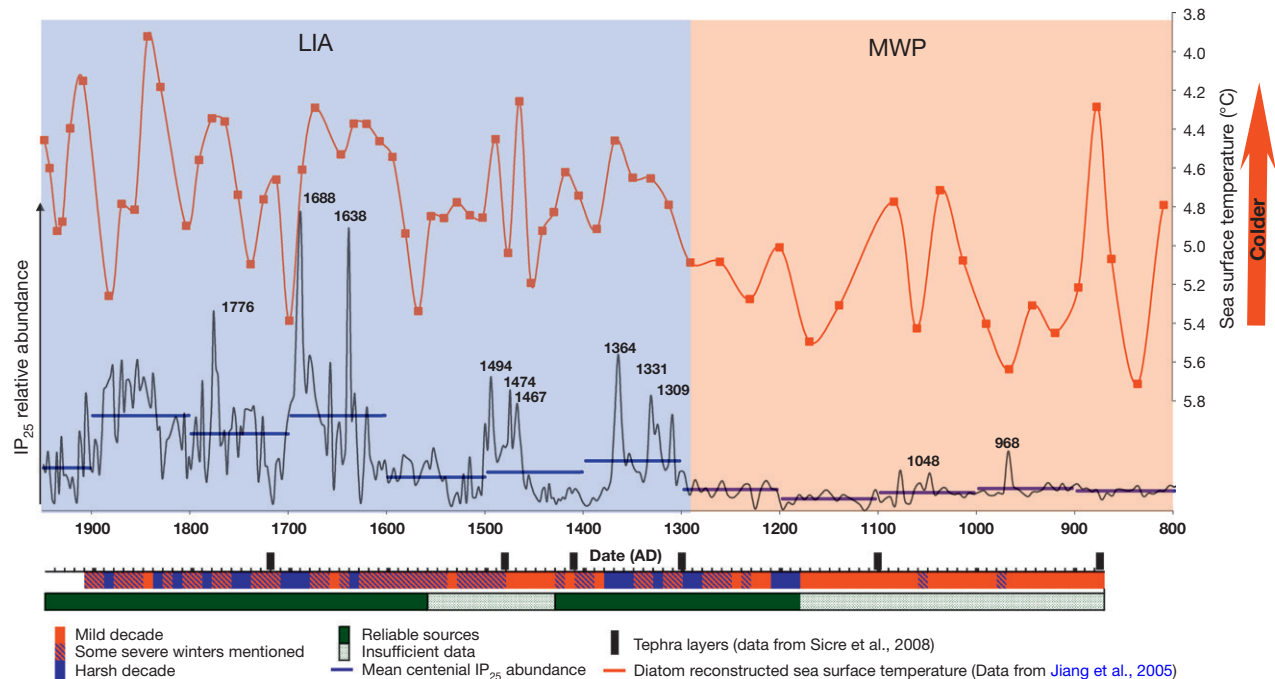


Figure 4 Relative abundances of IP_{25} found in the core MD99-2275 for the period AD 800–1950 plotted against historical records of Icelandic sea ice interpreted from Ogilvie (1992) and Ogilvie and Jónsson (2001) (bottom scales) and diatom-based reconstructed sea surface temperature (Jiang et al., 2005). Reproduced from Massé G, Rowland SJ, Sicre M-A, Jacob J, Jansen E, and Belt ST (2008) Abrupt climate changes for Iceland during the last millennium: Evidence from high resolution sea ice reconstructions. *Earth and Planetary Science Letters* 269: 565–569. Copyright Elsevier.

The CAA contains a vast array of islands and channels that connect the Arctic Ocean with the Pacific and Atlantic oceans, yet, until recently, relatively little was known about the sea ice record of this key oceanographic region other than what has been inferred from neighboring terrestrial records or other components of the marine record (e.g., water masses). In modern times, the CAA is characterized by frozen channels during the winter months following the onset of sea ice formation during the late autumn. Sea ice melt is variable throughout the channels, making navigation possible but often hazardous. In the longer term past, sea ice occurrence has almost certainly been more severe, at least in recent centuries, blocking the Northwest Passage. Following collection of several marine sediment cores during the 2005 ArcticNet oceanographic campaign through the CAA, paleo-sea ice records covering the Holocene have been constructed, based mainly on the variable occurrence of IP_{25} (Belt et al., 2010; Vare et al., 2009; Figure 5).

Initially, Vare et al. (2009) interpreted the variable abundance of IP_{25} in a core obtained from Barrow Strait as representing changes in sea ice conditions throughout the Holocene for this region. This study revealed a number of key outcomes, previously unknown for this important region. Firstly, radiocarbon dating of the core demonstrated that it covered the entire Holocene epoch or the last 10 000 years. This chronological analysis also revealed some variations in the sedimentation rate within the core, which are likely to be a result of changes in local climate. Importantly, the IP_{25} biomarker was present in each of the 650 horizons analyzed, providing evidence for the presence of at least partial sea ice conditions on a decadal to semicentennial timescale throughout the

Holocene. Given the variability in sedimentation rates, Vare et al. (2009) converted all sedimentary IP_{25} concentrations to annual fluxes in order to facilitate more meaningful and consistent temporal comparisons (Figure 6). IP_{25} fluxes were consistent and relatively low during the early and mid-Holocene, before increasing around 3000–4000 years before present (BP), after which they remained high but with much greater variation compared to the early Holocene. These observations were interpreted as indicating reduced sea ice cover in the early to mid-Holocene, with more extensive sea ice cover after ca. 3000 years BP. The co-occurrence of a related HBI diene with an extremely similar flux profile provided complementary evidence for such changes in climate, especially as the proportion of the more unsaturated HBI diene decreased after ca. 3000 years BP, in line with the previously reported temperature dependence of HBI distributions on growth temperature (Rowland et al., 2001). Further, the occurrence of significant changes to sea ice conditions throughout the Holocene, especially around 3000 years BP, was consistent with trends in the extent of melt from the Agassiz ice cap (north of Barrow Strait) and with other proxy evidence for the termination of the so-called Holocene Thermal Maximum during the second half of the Holocene.

In a subsequent study, Belt et al. (2010) constructed paleo-sea ice records for Victoria Strait and Dease Strait, also within the CAA, based on the occurrence of IP_{25} (Figure 6). This study aimed to identify any consistencies or differences in the spatial and temporal distribution of spring sea ice within the CAA throughout the Holocene. The two examined cores did not cover as much of the Holocene as the cores from the Barrow

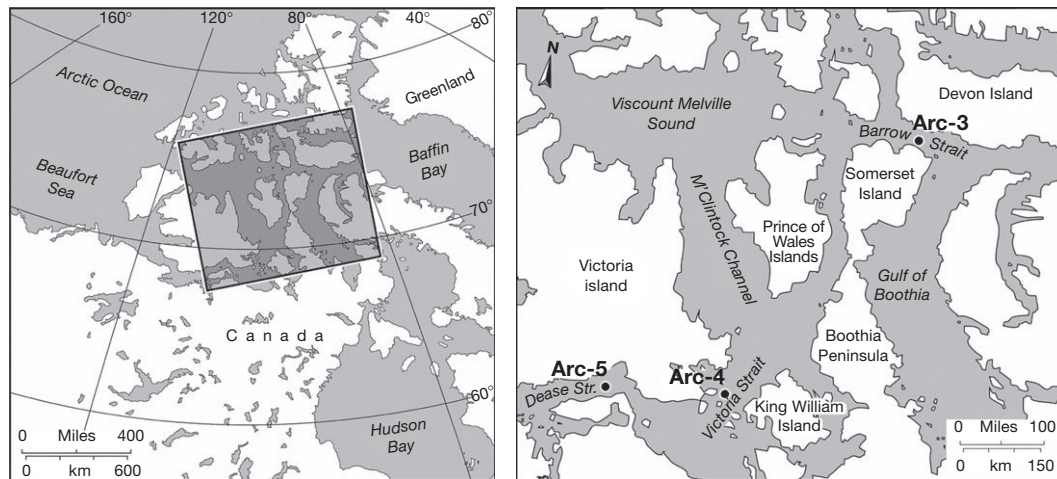


Figure 5 Map of the Canadian Arctic Archipelago and the sampling locations (ARC-3,4,5) described in Vare et al. (2009) and Belt et al. (2010). Reproduced from Belt ST, Vare LL, Massé G, et al. (2010) Striking similarities in temporal changes to spring sea ice occurrence across the central Canadian Arctic Archipelago over the last 7000 years. *Quaternary Science Reviews* 29: 3489–3504. Copyright Elsevier.

Strait study, especially in the early part of the epoch, but some interesting comparisons between all three study locations were nonetheless made. Most significantly, the sea ice biomarker IP₂₅ was again present in all of the horizons examined from the Victoria Strait and Dease Strait cores, consistent with the observations made previously for Barrow Strait (Vare et al., 2009). The IP₂₅ fluxes were also quite variable, but lowest in the mid-Holocene, before increasing at ca. 3000 years BP, suggesting some similarity in sea ice conditions between the three locations. Indeed, statistical analysis of the IP₂₅ fluxes demonstrated significant correlations between sites, especially in the early and mid-Holocene (Belt et al., 2010). However, in contrast to the data reported for Barrow Strait, the IP₂₅ fluxes decreased from their maximum values at ca. 3000 years BP during a relatively brief (500–1000 years) interval around 1000 years BP, indicative of lower sea ice occurrence. Belt et al. (2010) noted that this coincided with a period of warmer climate in the more southerly locations as predicted from terrestrial proxy studies and from the documented migration of ancestral Inuit peoples.

The measurement of IP₂₅ in marine sediments to determine past Arctic sea ice conditions has recently been extended to other regions of the Arctic and for longer timescales. Fram Strait represents a major Arctic gateway that connects the Arctic and Atlantic oceans and channels temperate water from the North Atlantic along the western continental margin of Spitsbergen. In contrast, cold water and sea ice are transported south through Fram Strait via the East Greenland current. By analyzing for IP₂₅ and another biomarker, brassicasterol, derived from open water phytoplankton, Müller et al. (2009) established a 30 000-year sea ice record for Fram Strait. Perhaps of greatest significance, Müller et al. (2009) used the combination of the two biomarkers to highlight the occurrence of perennial ice, and ice-free conditions as representing extremes in climatic conditions, together with evidence for spring sea ice occurrence, mainly through the IP₂₅ biomarker. This study also illustrated the potential for this biomarker-based approach to distinguish between more subtle changes in sea ice conditions,

such as those associated with a regular or variable spring ice edge.

Refining the IP₂₅ Sea Ice Proxy and Future Directions

Collectively, this relatively small number of case study determinations of paleo-sea ice conditions in the Arctic has nonetheless shown the potential to use IP₂₅ to obtain meaningful datasets that can be interpreted alongside other proxy records to establish paleoclimate records. In order to improve on the scope and accuracy of these reconstructions, it will be necessary to further investigate the nuances of the IP₂₅ biomarker proxy, particularly with respect to interpretation of quantitative data and a move toward an increased understanding of how the presence of IP₂₅ in sediments may be influenced by other environmental or climatic factors. For example, it will be important to establish whether sedimentary abundances or fluxes of IP₂₅ can be calibrated so as to yield more quantitative assessments of sea ice occurrence, ice type or thickness. In addition, since sedimentation rates in the Arctic are typically on the order of tens of years per cm (at best), determining interannual variations is probably beyond the scope of the IP₂₅ proxy. However, having a greater insight into the production of IP₂₅ by sea ice diatoms in the first place may help interpret sedimentary distributions, even at these lower resolutions. With this in mind, Brown et al. (2011) investigated the time interval over which IP₂₅ was present in Arctic sea ice by sampling sea ice cores collected from the southeast Beaufort Sea from late winter through to the ice melt in late spring/early summer. This study showed that the IP₂₅ content of the sea ice was absent or below the limit of detection during the winter period as expected, before increasing during the early spring and reaching a maximum concentration during May, coincident with the spring algal bloom as evidenced from parallel cell abundance and pigment determinations. This investigation demonstrated the strong seasonal influence on IP₂₅ production, with the majority (~90%) of this biomarker being

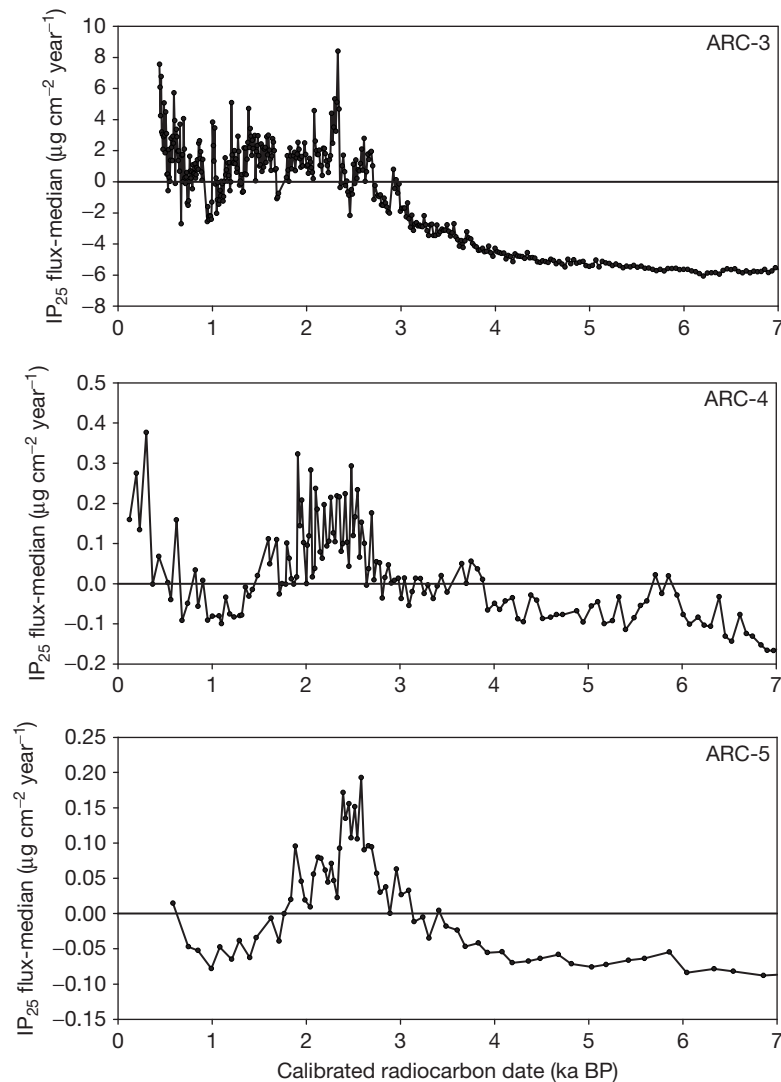


Figure 6 IP_{25} flux profiles (value – median) for ARC-3 (Barrow Strait), ARC-4 (Victoria Strait) and ARC-5 (Dease Strait). Reproduced from Belt ST, Vare LL, Massé G, et al. (2010) Striking similarities in temporal changes to spring sea ice occurrence across the central Canadian Arctic Archipelago over the last 7000 years. *Quaternary Science Reviews* 29: 3489–3504. Copyright Elsevier.

produced in a relatively short (4–6 weeks) time interval during the spring (Brown et al., 2011).

Previously, a strict sea ice diatom origin for IP_{25} had been proposed on the basis of its presence in bulk Arctic sea ice samples, its absence from phytoplankton collected under open water conditions from nearby locations, and its distinctive stable isotopic composition in sea ice, in sediment trap material and in sediments (Belt et al., 2007, 2008). In their more recent and detailed study looking at sectioned sea ice cores, Brown et al. (2011) also showed that the IP_{25} biomarker was largely restricted to intervals of the sea ice cores that had brine volume fractions capable of supporting colonization and growth of diatoms (>5%; Golden et al., 2007) and that the highest IP_{25} concentrations were found in sections of sea ice distal to the ice–seawater interface; both of these observations are consistent with the production of IP_{25} by sea ice-associated diatoms.

In summary, Arctic sea ice diatoms produce a suite of biomarker chemicals that are straightforward to identify and quantify by routine analytical methods. The analysis of fatty acids and sterols is somewhat limited to determinations of primary production since they occur too commonly in other organisms to permit more targeted measures, such as source identification. In contrast, HBI alkenes are biosynthesized by a limited number of diatoms and can, therefore, be used to identify sources when detected in the marine environment, including sediments. One such monounsaturated HBI alkene, IP_{25} , is biosynthesized exclusively by Arctic sea ice diatoms and the presence of IP_{25} in well-dated marine sediments is proving extremely valuable in determining paleo-sea ice records across the Arctic. To date, IP_{25} has not been detected in Antarctic diatoms or sediments, although it is feasible that other HBI compounds may have the potential to be used as paleo-sea ice and climate proxies.

See also: Diatom Introduction; Diatoms. **Diatom Records:** Diatom Fossil Records from Marine Laminated Sediments. **Diatom Methods:** $\delta^{18}\text{O}$ Records; Salinity and Climate Reconstructions from Continental Lakes; Use in Archaeology. **Diatom Records:** Antarctic Waters; Freshwater Laminated Sequences; Large Lakes; North Atlantic and Arctic; Pacific. **Paleobotany:** Silicon Isotopes in Diatoms. **Paleoceanography, Biological Proxies:** Biomarker Indicators of Past Climate; Marine Diatoms.

References

- Ambrose WG Jr., von Quillfeldt C, Clough LM, Tilney PVR, and Tucker T (2005) The sub-ice algal community in the Chukchi sea: Large- and small-scale patterns of abundance based on images from a remotely operated vehicle. *Polar Biology* 28: 784–795.
- Andrews JT, Belt ST, Olafsdottir S, Massé G, and Vare LL (2009) Sea ice and climate variability for NW Iceland/Denmark Strait over the last 2000 cal. yr BP. *The Holocene* 19: 775–784.
- Arrigo KR, Mock T, and Lizotte MP (2010) Primary producers and sea ice. In: Thomas DN and Dieckmann GS (eds.) *Sea Ice*, pp. 283–325. Chichester: Wiley-Blackwell.
- Barenholz Y (2002) Cholesterol and other membrane active sterols: From membrane evolution to 'rafts'. *Progress in Lipid Research* 4: 1–5.
- Barrett SM, Volkman JK, and Dunstan GA (1995) Sterols of 14 species of marine diatoms (Bacillariophyta). *Journal of Phycology* 31: 360–369.
- Belt ST, Allard WG, Massé G, Robert J-M, and Rowland SJ (2000) Highly branched isoprenoids (HBLs): Identification of the most common and abundant sedimentary isomers. *Geochimica et Cosmochimica Acta* 64: 3839–3851.
- Belt ST, Allard WG, Massé G, Robert J-M, and Rowland SJ (2001) Structural characterisation of C_{30} highly branched isoprenoid alkenes (rhizenes) in the marine diatom *Rhizosolenia setigera*. *Tetrahedron Letters* 42: 5583–5585.
- Belt ST, Cooke DA, Robert J-M, and Rowland SJ (1996) Structural characterisation of widespread polyunsaturated isoprenoid biomarkers: A C_{25} triene, tetraene and pentaene from the diatom *Haslea ostrearia* Simonsen. *Tetrahedron Letters* 37: 4755–4758.
- Belt ST, Massé G, Rowland SJ, Poulin M, Michel C, and LeBlanc B (2007) A novel chemical fossil of palaeo sea ice: IP_{25} . *Organic Geochemistry* 38: 16–27.
- Belt ST, Massé G, Vare LL, et al. (2008) Distinctive ^{13}C isotopic signature distinguishes a novel sea ice biomarker in Arctic sediments and sediment traps. *Marine Chemistry* 112: 158–167.
- Belt ST, Vare LL, Massé G, et al. (2010) Striking similarities in temporal changes to seasonal sea ice conditions across the central Canadian Arctic Archipelago during the last 7,000 years. *Quaternary Science Reviews* 29: 3489–3504.
- Brown TA, Belt ST, Philippe B, et al. (2011) Temporal and vertical variations of lipid biomarkers during a bottom ice diatom bloom in the Canadian Beaufort Sea: Further evidence for the use of the IP_{25} biomarker as a proxy for spring Arctic sea ice. *Polar Biology* 34(12): 1857–1868.
- Caron DA and Gast RJ (2010) Heterotrophic protists associated with sea ice. In: Thomas DN and Dieckmann GS (eds.) *Sea Ice*, pp. 327–356. Chichester: Wiley-Blackwell.
- Golden KM, Eicken H, Heaton AL, Miner J, Pringle DJ, and Zhu J (2007) Thermal evolution of permeability and microstructure in sea ice. *Geophysical Research Letters* 34: L16501.
- Gosselin M, Levasseur M, Wheeler PA, Horner RA, and Booth BC (1997) New measurements of phytoplankton and ice algal production in the Arctic Ocean. *Deep-Sea Research Part II* 44: 1623–1644.
- Horner R, Ackley SF, Dieckmann GS, et al. (1992) Ecology of ice biota 1. Habitat, terminology, and methodology. *Polar Biology* 12: 412–427.
- Jiang H, Eiriksson J, Schultz M, Knudsen KL, and Seidenkrantz MS (2005) Evidence for solar forcing of sea surface temperature on the North Icelandic Shelf during the late Holocene. *Geology* 33: 73–76.
- Jones PD, Osborn TJ, and Briffa KR (2001) The evolution of the climate over the last millennium. *Science* 292: 662–667.
- Legendre L, Martineau M-J, Theriault J-C, and Demers S (1992) Chlorophyll *a* biomass and growth of sea-ice microalgae along a salinity gradient (southeastern Hudson Bay, Canadian Arctic). *Polar Biology* 12: 445–453.
- Massé G (2003) *Highly Branched Isoprenoid Alkenes from Diatoms: A Biosynthetic and Life Cycle Investigation*. PhD thesis, University of Plymouth.
- Massé G, Rowland SJ, Sicre M-A, Jacob J, Jansen E, and Belt ST (2008) Abrupt climate changes for Iceland during the last millennium: Evidence from high resolution sea ice reconstructions. *Earth and Planetary Science Letters* 269: 565–569.
- Michel C, Ingram RG, and Harris LR (2006) Variability in oceanographic and ecological processes in the Canadian Arctic Archipelago. *Progress in Oceanography* 71: 379–401.
- Michel C, Nielsen TG, Nozais C, and Gosselin M (2002) Significance of sedimentation and grazing in ice micro- and meiofauna for carbon cycling in annual sea ice (northern Baffin Bay). *Aquatic Microbial Ecology* 30: 57–68.
- Müller J, Massé G, Stein R, and Belt ST (2009) Variability of sea-ice conditions in the Fram Strait over the past 30,000 years. *Nature Geoscience* 2: 772–776.
- Nozais C, Gosselin M, Michel C, and Tita G (2001) Abundance, biomass, composition and grazing impact of the sea-ice meiofauna in the North Water, northern Baffin Bay. *Marine Ecology Progress Series* 217: 235–250.
- Ogilvie AEJ (1992) Documentary evidence for changes in the climate of Iceland AD 1500 to 1800. In: Bradley RS and Jones PD (eds.) *Climate Since AD 1500*, pp. 92–117. London and New York: Routledge.
- Ogilvie AEJ and Jónsson T (2001) "Little Ice Age" research: A perspective from Iceland. *Climatic Change* 48: 9–52.
- Poulin M (1990) Ice diatoms: The Arctic. In: Medlin LK and Priddle J (eds.) *Polar Marine Diatoms*, pp. 15–18. Cambridge: British Antarctic Survey.
- Poulin M, Daugbjerg N, Gradinger R, Ilyash L, Ratkova T, and von Quillfeldt C (2011) The pan-Arctic biodiversity of marine pelagic and sea-ice unicellular eukaryotes: A first-attempt assessment. *Marine Biodiversity* 43: 13–28.
- Reuss N and Poulsen LK (2002) Evaluation of fatty acids as biomarkers for a natural plankton community. A field study of a spring bloom and a post-bloom period off West Greenland. *Marine Biology* 141: 423–434.
- Rowland SJ, Belt ST, Wraige EJ, Massé G, Roussakis C, and Robert J-M (2001) Effects of temperature on polyunsaturation in cytotstatic lipids of *Haslea ostrearia*. *Phytochemistry* 56: 597–602.
- Rózańska M, Gosselin M, Poulin M, Wiktor JM, and Michel C (2009) Influence of environmental factors on the development of bottom ice protest communities during the winter-spring transition. *Marine Ecology Progress Series* 386: 43–59.
- Sicre M-A, Jacob J, Ezat U, et al. (2008) Decadal variability of sea surface temperatures off North Iceland over the last 200 yrs. *Earth and Planetary Science Letters* 268: 137–142.
- Vare LL, Massé G, and Belt ST (2010) A biomarker-based reconstruction of sea ice conditions for the Barents Sea in recent centuries. *The Holocene* 40: 637–643.
- Vare LL, Massé G, Gregory TR, Smart CW, and Belt ST (2009) Sea ice variations in the central Canadian Arctic Archipelago during the Holocene. *Quaternary Science Reviews* 28: 1354–1366.
- Volkman JK, Barrett SM, Blackburn SI, Mansour MP, Sikes EI, and Gelin F (1998) Microalgal biomarkers: A review of recent research developments. *Organic Geochemistry* 29: 1163–1179.
- Volkman JK, Barrett SM, and Dunstan GA (1994) C_{25} and C_{30} highly branched isoprenoid alkenes in laboratory cultures of two marine diatoms. *Organic Geochemistry* 21: 407–414.

Diatoms, Marine *see* **Paleoceanography, Biological Proxies:** Marine Diatoms

Dinoflagellates *see* **Paleoceanography, Biological Proxies:** Dinoflagellates

DNA, Fossil Animal *see* **Vertebrate Studies:** Ancient DNA

DNA, Fossil Plants *see* **Paleobotany:** Ancient Plant DNA

Drought History Reconstruction *see* **Paleoclimate Reconstruction:** Paleodroughts and Society

DUNE FIELDS

Contents

High Latitudes

Mid-Latitudes

Low Latitudes

High Latitudes

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Glossary

Cold-climate environment: Includes periglacial environments, where the mean annual air temperature is less than +3 °C, as well as regions with snowy climates, where the coldest mean monthly temperature is below −3 °C.

High-latitude dune fields: Sand dunes found in more northerly and southerly latitudes and typically associated with seasonal cold-climate conditions.

Parabolic dunes: A dune in the shape of a parabola or horseshoe, in plan view, with a concave windward slope and a convex leeward slope. The trailing arms of parabolic dunes are commonly vegetated.

Sand dunes: Distinct constructional eolian landforms that generally exceed several meters in height and exhibit slipfaces. They typically comprise stacked cosets of large-scale, high-angle cross strata, representing net deposition from climbing bedforms.

Sand sheets: A landform composed of eolian sand of varying thickness but without pronounced relief typically with subhorizontal to low-angle cross-stratification formed by the migration of low-amplitude wind ripples or by sedimentation in the presence of vegetation. The cover sands of the European sand belt are an extensive area of sand sheet deposits.

Introduction

Quaternary dune fields are widespread throughout high-latitude regions of the Northern Hemisphere including Canada, Alaska, Europe, and Russia. Much of North America north of approximately 45° N latitude, and Europe north of 50° N latitude, was glaciated during the late Quaternary (Wisconsin and Weichselian glacial period, respectively). Most of these areas also presently experience seasonal cold-climate conditions, though there are few large dune fields active today. Cold-climate eolian conditions presently occur in Antarctica, though the few dune fields there are small.

Dune activity is largely controlled by a combination of the three interacting factors of transport capacity (i.e., windiness), sediment supply, and sediment availability. Sediment supply represents the total amount of sandy source material, whereas sediment availability represents the proportion of unprotected, sandy material, that is, that available for transport, and may be reduced by vegetation cover, moisture, snow, ice, surface lag deposits, and other factors. In high latitudes, sediment supply appears to be the limiting factor on the size of dune fields. Individual dune fields in high latitudes are rarely larger than a few 100 km²,

owing to the relatively small amount of sediment. In comparison, sediment availability, as influenced by the amount of vegetation cover, appears to be a significant control on the degree of dune activity in high latitudes. Transport capacity is also a factor, as high-latitude continental areas may have been windier during Late Glacial and early Holocene times than they are today.

Primary Dune Fields

North America

Dune fields occur throughout Canada (about 26 000 km²) and the northern contiguous USA (Figure 1). The majority of dune fields (about 90%) are in central Canada owing to the availability of suitable sandy sediments and strong postglacial winds. The Athabasca Sand Dunes of northern Alberta and Saskatchewan is the largest dune field in Canada (4800 km²), and it contains the largest area of active dunes (385 km²) in the high latitudes (Figure 2). Other large dune fields include the Wood Buffalo Sand Hills, Brandon Sand Hills, Great Sand Hills, and Minot Sand Hills (Figure 1), though they contain only limited areas of active dunes.

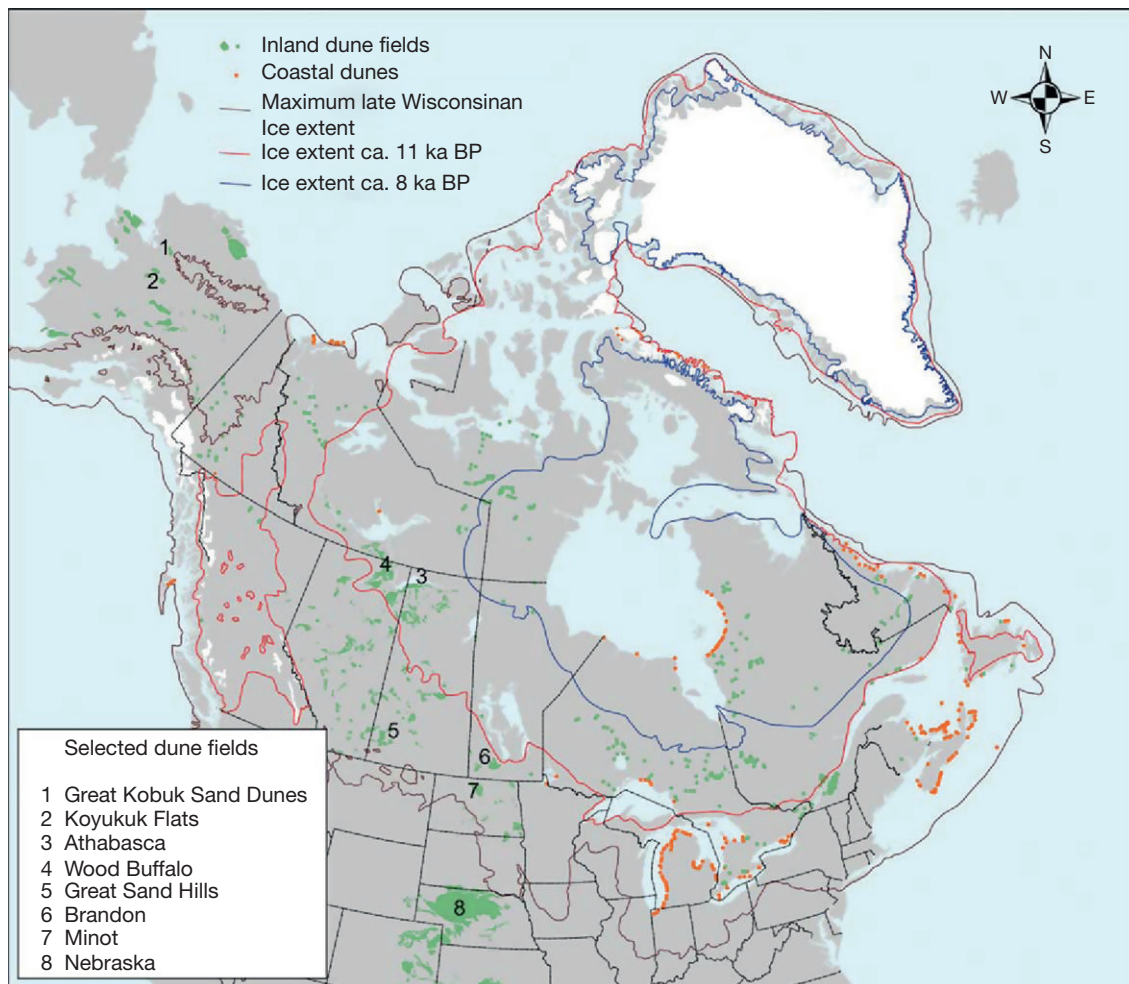


Figure 1 High-latitude dune fields of northern North America. Distribution of dune fields shown in relation to the maximum late Wisconsin glacial ice extent between 18.0 and 15.5 ka BP, and to later ice extents at 11.0 and 8.0 ka BP (uncalibrated radiocarbon years).

Eolian dune fields and sheet sands are also widespread in Alaska (Figure 1), occurring within all major lowlands and alluvial plains beyond the limit of the last glaciation. The most notable dune field is the Great Kobuk Sand Dunes of north-western Alaska (650 km²), which contains areas of active dunes (Figure 3).

Europe and Russia

Eolian sand deposits form a continuous belt that extends through north-west, central, and east Europe, known as the European sand belt (Figure 4). Stabilized dune fields are found inland or along river systems in the UK, Belgium, The Netherlands, Germany, Poland, Belarus, Ukraine, Lithuania,



Figure 2 Active transverse dunes in the Athabasca sand dunes, northern Saskatchewan, Canada (photograph by D.G. Smith).



Figure 3 Large active dune area, approximately 17 km long, of The Great Kobuk Sand Dunes, Alaska (NASA Alaska High Altitude Aerial Photography, courtesy of the Geophysical Institute, University of Alaska Fairbanks).

Latvia, Estonia, and the Russian Plain of European Russia. In addition, relatively small but widespread dune fields are scattered throughout Norway, Sweden, and Finland (Fennoscandia; Figure 5), north of the Arctic Circle.

Other Dunes and Dune Fields

Small active dune fields occur in Antarctica, most notably in the ice-free Victoria Valley of southern Victoria Land. This dune field is about 3.5 km long and <1 km wide and consists of sand sheets, and single and compound barchanoid dunes forming transverse ridges between 70 and 600 m long (Figure 6). Other dunes described as ‘whalebacks’ occur without slipfaces and mantle underlying ground moraine. The dunes contain interbedded snow and sand, and are underlain by permafrost. In addition to inland dunes, active coastal dunes occur in high-latitude, cold-climate, and temperate regions along sandy shorelines in areas of moderate-to-high wind regimes. Active high-latitude marine coastal dunes are found in Europe, Alaska, and the Beaufort Sea, Atlantic and Pacific coasts of Canada. Stabilized coastal dunes also occur inland of present-day marine and lake shorelines due to a decline in relative sea level as a result of isostatic uplift or drainage of postglacial lakes during the late Quaternary and Holocene. These raised, stabilized coastal dunes may be used to document the timing of former higher lake and sea levels.

Types

The most common dune type in high latitudes is the parabolic dune, both simple and compound forms. Parabolic dunes are indicative of moist sand and vegetation, which modify sediment availability and transport. Most high-latitude parabolic dunes typically have partially vegetated heads with lee side slipfaces that are convex downwind in plan view, and trailing arms that are partially or completely vegetated (Figure 7).

Transverse and longitudinal dunes may also occur in high-latitude areas, though some ‘longitudinal’ dunes actually represent the remnant arms of parabolic dunes. Barchanoid dunes are found in some cold arid regions due to the absence of vegetation and moisture. Blowouts are common where vegetation cover on dune fields is high, and often modify stabilized dunes. Many temperate and cold-climate dune fields of high latitudes are presently stabilized by vegetation, suggesting that they are formed under climatic regimes different from those of today.

Although most high-latitude dunes are inactive, they are presently active in some areas of Europe and North America. Small portions of dune fields on the northern Great Plains of North America are active due to moderate-to-strong wind regimes, and periodic droughts that reduce vegetation cover. More northerly dune fields such as the Kobuk and Athabasca dune fields may be active today due to a combination of factors including moderate present-day wind regimes, moisture stress, and limited growing seasons, and periodic sediment input from rivers, deltas, or lakeshores. Along eastern Hudson Bay, Canada, and in the Lapland region of Finland, small dune areas in subarctic settings are active due to disturbance by fire, and overgrazing and trampling by reindeer, respectively.



Figure 4 The European sand belt and selected ice extents. Reproduced from Zeeberg J (1998) The European sand belt in eastern Europe – And comparison of Late Glacial dune orientation with GCM simulation results. *Boreas* 27: 127–139.

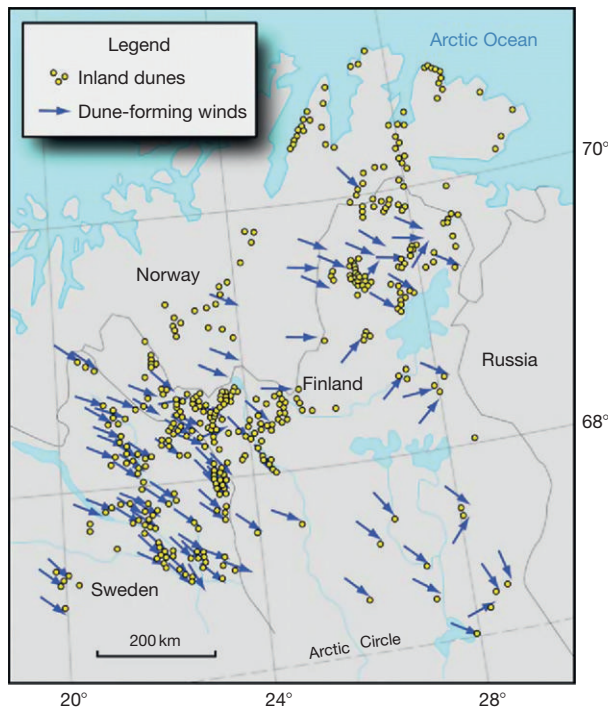


Figure 5 High-latitude dune fields and inferred dune-forming winds in Fennoscandia. Reproduced from Seppälä M (2004) *Wind as a Geomorphic Agent in Cold Climates*. Cambridge: Cambridge University Press.

North America

Most dune fields in Canada and the northern contiguous USA consist of parabolic dunes and blowouts, with a few dune fields also containing transverse and longitudinal dunes. Various forms of parabolic dunes occur. In northern Canada, very elongate and narrow stabilized parabolic dunes occur on otherwise bare rock of the Canadian Shield (Figure 8). These dunes are typically <10 m high, but with trailing arms up to

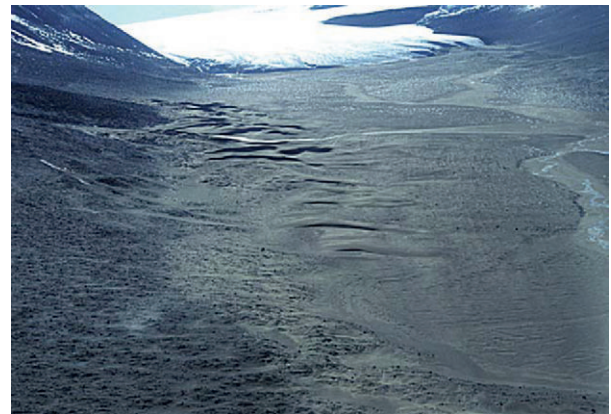


Figure 6 Active transverse dunes of the Victoria Valley, Antarctica (photograph by N. Lancaster).



Figure 7 Active parabolic dune in southwestern Saskatchewan, Canada. Dominant transporting winds are from left to right. The head of the dune is partially vegetated whereas the trailing upwind arms are stabilized by vegetation (photograph by S.A. Wolfe).

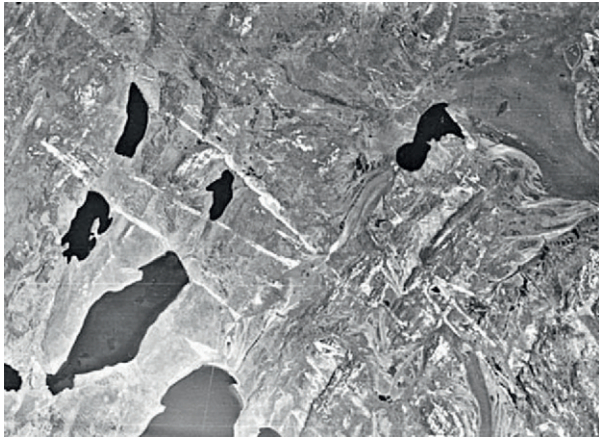


Figure 8 Stabilized elongate parabolic dunes of northern Saskatchewan, Canada. Dune at center left is approximately 3 km long and is oriented towards the northwest. The dunes have migrated over bedrock surfaces with low-sediment supplies. Transporting winds were from the southeast, indicating that anticyclonic winds associated with the Laurentide Ice Sheet were responsible for their formation. See text and **Figure 12** for details (National Air Photo Library of Canada photograph A13385-97).



Figure 9 A field of densely concentrated, stabilized circular parabolic dunes, Great Sand Hills, southwestern Saskatchewan, Canada. These dunes were active in the 1800s, possibly as a result of drought in the late 1700s (Wolfe et al., 2001; photograph by S.A. Wolfe).

8 km long. In southwestern Saskatchewan, parabolic dunes tend to be more densely concentrated and less elongate (**Figure 9**). The Athabasca Sand Dunes contain a wide variety of active dunes, including parabolic, transverse (**Figure 2**), and 'hybrid' transverse-longitudinal dunes. Well-formed stabilized transverse dunes up to 25 m high also occur in the Holmes Crossing sand hills of central Alberta, Canada (**Figure 10**). Many dune fields in the northern Great Plains of Canada and the USA are modified by blowouts that have been active at varying times due to drought and disturbance.

Dune fields in Alaska include stabilized parabolic, longitudinal, and compound dunes, ranging from 1.5 to 60 m high. Active dunes in the Kobuk Valley and Koyukuk Flats include

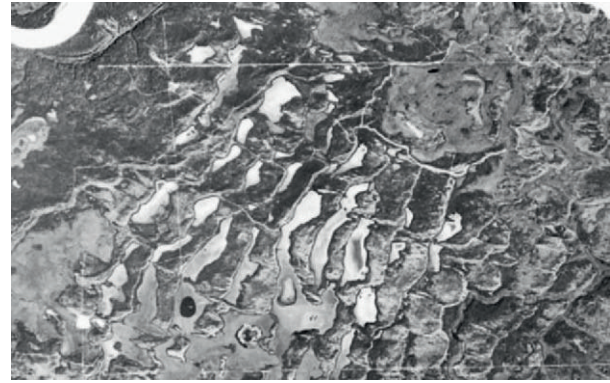


Figure 10 Stabilized transverse dunes, Holmes Crossing, central Alberta, Canada. Optical dating indicates these dunes were last active approximately 15 000 years ago. Transporting winds were from the northwest (upper left), and image view is 7 km wide (National Air Photo Library of Canada photograph A25853-51).

transverse and barchanoid dunes (**Figure 3**). Other areas include partially stabilized parabolic and blowout dunes, and rounded hillocks and hollows or broad undulating ridges and swales with no distinct dune type.

Europe and Russia

Stabilized dune fields occur on the northern portions of the European sand belt. In the United Kingdom, the dunes are typically low hummocks; in The Netherlands, they are low ridges of parabolic form; in Germany, parabolic dunes are typically only a few meters high, and up to 12 m; and in Poland, parabolic, longitudinal, transverse, and, rarely, barchanoid dunes occur. Present-day eolian activity in Europe is mostly restricted to areas of human or animal disturbances.

Dune fields on the Russian Plain include individual and compound parabolic dunes, transverse dunes, and blowouts, and may also be associated with sand sheets. Dune heights typically range from 1.5 to 10.0 m, and may reach 20.0 m. Some parabolic dunes are several kilometers long.

Stabilized parabolic dunes are common in Fennoscandia, and may occur in clusters that are also parabolic in form (**Figure 11**). Individual dunes are typically <1300 m long, up to 150 m wide, and up to 12 m high, but more commonly 4–6 m high. Active parabolic dunes are found in only a few locations, although active blowouts on stabilized dunes are relatively common beyond the treeline.

Origins

The origins of high-latitude dune fields are tied to the interrelated factors of sediment supply, sediment availability, and wind regime. Many dune fields of high latitudes are derived from localized sandy sediments of fluvial, lacustrine, or glacial origin that were deposited by glacial meltwater. Typical source areas include flood plains, glacial outwash plains, braided river valleys, deltas, lake shorelines, eskers, kames, and some till plains and moraines. The size of most high-latitude dune fields is ultimately controlled by the amount of sandy sediment in

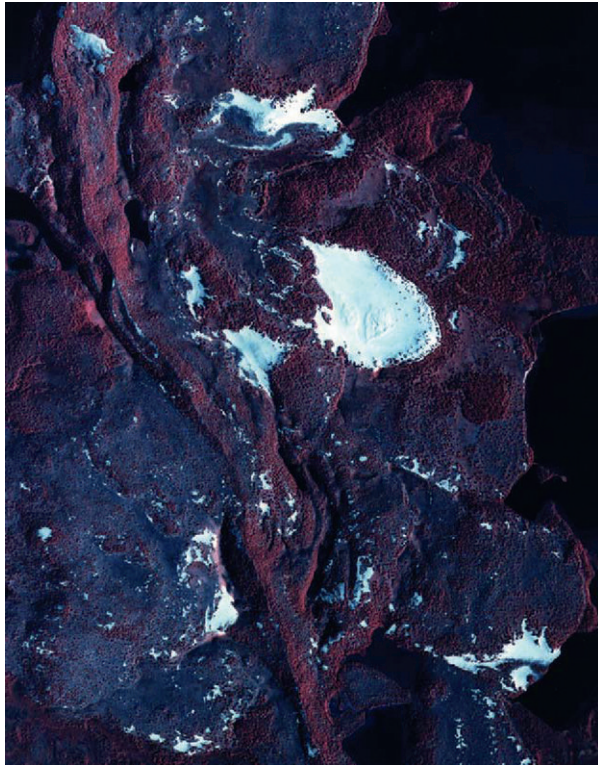


Figure 11 Active and stabilized parabolic dunes and blowouts at Melajärvi (Paddle Lake), in northern Finland. The source material for the dunes is an esker ridge and surrounding glaciolacustrine/deltaic sediments. The largest dune is about 550 m across and the effective wind direction is from the west-northwest. The area is north of the coniferous treeline, and the surface is vegetated by mountain birch (*Betula pubescens*) dwarf shrubs and lichens (Topographic Division of the Finnish Defense Forces, photograph V9120/111).

these deposits. However, the timing of dune activity is more related to periods when these sandy sediments were available for transport by wind. Within or proximal to formerly glaciated areas, initial dune field construction may have occurred shortly after glacial retreat and deposition of the source sediments, when the sandy terrain was sparsely vegetated. It is also hypothesized that winds during these initial periods of dune construction were stronger than today in parts of northern North America and Europe, due to the proximity of continental ice sheets and their effects on pressure gradients and atmospheric circulation patterns.

Subsequent episodes of dune reactivation during the Holocene may be related to climatic variability, affecting available moisture and vegetation cover, or disturbance by fire, animals, or humans.

North America

Within the glaciated regions of Canada and northern USA, most dune fields are derived from discrete sandy glaciofluvial and glaciolacustrine deposits associated with the retreat of Wisconsin glacial ice. Here, the contrast between relatively small dune fields of formerly glaciated North America and much larger dune fields of unglaciated regions to the south

(e.g., the Nebraska Sand Hills of the central Great Plains, USA) is evident (Figure 1). The largest of these northern dune fields, the Athabasca Sand Dunes, was originally supplied by sandy meltwater-derived sediments deposited into glacial Lake Athabasca during the retreat of the Laurentide Ice Sheet about 9500 calendar years (cal) before present (BP). The dunes have since been supplied by river, delta, and the lakeshore sediments as lake levels declined due, in part, to isostatic rebound.

In the unglaciated lowlands of central and southwestern Alaska, dune fields are derived largely from outwash sediments of major rivers. However, most exposures within these dunes are underlain by subhorizontally stratified sand and subordinate silt, interpreted as eolian sheet sands formed during full-glacial (Wisconsin) conditions. Therefore, the dunes may not have formed from direct reworking of older outwash sediments, but as a result of reworking of eolian sand sheet sediments since the Last Glacial Maximum.

Europe and Russia

Similar to high-latitude dune fields of North America, the European sand belt is closely tied to the extent of glaciation, though the relation is somewhat less distinct. The northerly edges of the sand belt containing most of the dune fields and sand sheets coincides roughly with the maximum position of the late Weichselian ice sheet, whereas finer-grained cover sands and loess approximately coincide with the maximum limit of Pleistocene glaciation (Figure 4). Extensive ice-marginal deposits, such as outwash plains, glaciofluvial, and glaciolacustrine deposits, and abandoned river floodplains and spillways channels supplied sediment for eolian redeposition. Similarly, dune fields on the Russian Plain are located predominantly in areas of outwash plains and meltwater channels, and are derived from glaciofluvial or glaciolacustrine sands. In Fennoscandia, dune fields are evidently derived from glaciofluvial sources, including eskers (Figure 5), kames, deltas, valley trains, ice-dammed lake deposits, and outwash plains.

Ages

High-latitude dune fields throughout much of northern North America and Europe occur within the most recent glacial limits and, therefore, their maximum ages postdate the time of deglaciation. Beyond these limits, dune fields may be older, though it appears that most either originated, or were significantly reworked during the Last Glacial Maximum (late Wisconsin and late Weichselian, respectively). A common problem with determining the age of dune fields is that of partial or complete reworking under conditions of changed climate or disturbance. Thus, reworking may result in superimposed dunes that are potentially misleading indicators of the age and source of the underlying deposits.

North America

The maximum extent of the late Wisconsin glacial ice is the primary factor limiting the age of dune fields in Canada and the northern contiguous USA, as dune fields within this region

must postdate deglaciation. **Figure 1** illustrates the maximum extent of the ice sheet between 18.0 and 15.5 ka BP, along with the ice extent at later times during deglaciation, at about 11.0 and 8.0 ka BP. This shows that the maximum limiting ages of dune fields within the glacial limit are potentially time transgressive.

In accordance with glacial retreat, the oldest dune fields are expected to be located near the maximum extent of the Wisconsin ice sheet. However, most ages of eolian sands and paleosols in northern Great Plains dune fields along the southerly extents of the Laurentide Ice Sheet relate to late Holocene dune activity that may be associated with episodes of increased aridity in relation to drought. Such a difference in expected ages of eolian sediments highlights the issue of reworking of dune fields in high latitudes.

In contrast, stabilized parabolic dunes have been found in northern Saskatchewan (David, 1981), and southern Quebec, Canada (Filion, 1987), indicating postglacial dune activity initiated at about 10 ka BP. Dune activity is interpreted as having occurred under cool, dry conditions, and with dune-building winds coming from anticyclonic air circulation induced by the retreating Laurentide Ice Sheet (**Figure 12**). Dune activity ceased when circulation patterns similar to present were established following continued retreat of the

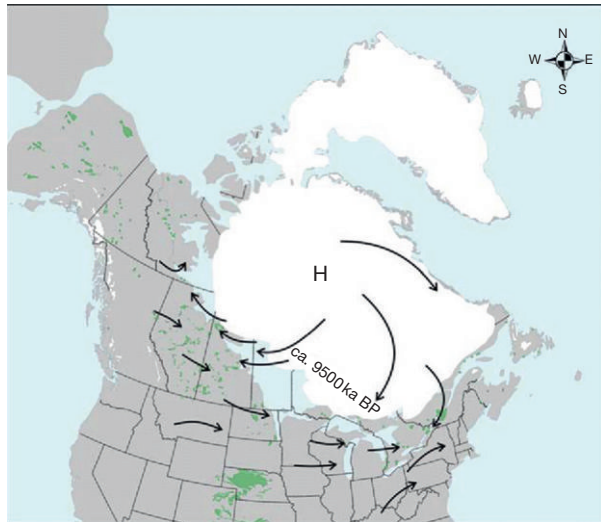


Figure 12 Map of the ice frontal position in North America, ca. 9.5 ka BP, at the time of initial dune formation in northern Saskatchewan and southern Quebec, Canada. Near the end of the Wisconsin glaciation, the retreating continental glacier controlled the distribution of regional air masses by effectively blocking the southward flow of polar air. Instead, it produced its own gravity airflow that descended radially towards the margins of the ice sheet and was gradually deflected clockwise by the Coriolis force. Upon reaching the ice margin, the dry, cold winds were blowing southwestward in the St. Lawrence Lowlands of Quebec in the east and northwestward in northern Saskatchewan and Alberta in the west, producing correspondingly oriented dunes in these areas. Large arrows show principal airflow directions associated with anticyclonic circulation in relation to the high-pressure center (H) over the Laurentide Ice Sheet. Reproduced from David PP (1981) Stabilized dunes ridges of northern Saskatchewan. *Canadian Journal of Earth Sciences* 18: 286–310.

ice sheet. At both localities, however, activity continued for a short time under modified winds and climate prior to stabilization.

Despite the preceding observations, the oldest recognized late Quaternary dune deposit in Canada is the Kittigazuit Formation of the Tuktoyaktuk Coastlands along the southern Beaufort Sea, Northwest Territories. Radiocarbon dates indicate that these eolian sands were deposited during the mid-Wisconsin, between 37 and 33 ka BP. These sediments were subsequently overridden by glacial ice that advanced over the area in the late Wisconsin (Dallimore et al., 1997).

In Alaska, most dune fields occur in areas that were not glaciated during the late Wisconsin and, therefore, the maximum ages of these dune fields are not constrained by the extent of glacial ice. It is hypothesized that many of the subhorizontally stratified deposits represent full-glacial eolian sand sheets; radiocarbon ages indicate that sand sheet aggradation in some lowlands occurred during the Last Glacial Maximum (ca. 25–14 ka BP), whereas many surface dunes may reflect Late Glacial and postglacial reworking (ca. 14–8.5 ka BP), and later. The surface morphology and orientations of most dune fields may, therefore, reflect reworking since the Last Glacial Maximum, rather than full-glacial paleowind directions or the age of the underlying deposits. By example, paleosols dating to ca. 24 ka BP within eolian sands in the Kobuk Valley, Alaska, may indicate regional stabilization of the dunes during the mid-Wisconsin, but with dune activity both before and after this period. There have also been repeated episodes of eolian stability and activity of at least some of these dunes in the past 8000 years.

Europe and Russia

Many detailed chronostratigraphic divisions of European dune sand, cover sand, and loess sequences have been published in the last few decades although a firm chronology of phases of eolian activity for this region is still lacking. Deposition of eolian sediments associated with the European sand belt has probably occurred over at least six Quaternary glacial periods. Eolian cover-sand sediments in The Netherlands have been dated by optically stimulated luminescence (OSL) to the last interglacial and the last two glacial stages. Three major phases of eolian deposition have been described within central and northwestern Europe. The first phase occurred during the late Weichselian glacial maximum, ca. 25–15 ka BP, with predominant fluvioeolian deposition occurring during periglacial conditions with widespread permafrost. The second, during the waning stages of glaciation, ca. 15–12 ka BP, represents the major period of sand sheet and low-dune deposition, with dominant sand-transporting winds from the northwest and a climate that was not much cooler than today. The last phase occurred between the Younger Dryas and early Holocene, ca. 11–9 ka BP. Dunes formed under west-southwesterly winds, in areas proximal to interfluvies and in river valleys, in relation to changing river channel morphology. Dune formation in the earlier phases was probably induced by a combination of factors including an abundance of unconsolidated sediment, absence of topographic obstruction, sparse vegetation cover, and a high wind-energy regime.

On the Russian Plain, radiocarbon ages of paleosols separating dune deposits from underlying alluvial sediments indicate that dune building occurred after about 15.0–12.5 ka BP, associated with the cold Late Glacial stages of the Older and Middle Dryas.

In the UK, Holocene dune activity has taken place episodically during the last 7000 years in relation to reductions in available moisture and vegetation, and disturbance by humans and livestock.

In Fennoscandia, deglaciation of the ice sheet took place between about 12.0 and 8.5 ka BP. Initial dune building probably began soon after local deglaciation. Stabilization may have occurred in several phases, beginning prior to 7 ka BP with colonization by pine trees, and continuing until about 4 ka BP with increased effective moisture and a regional rise in groundwater. Short-lived late Holocene eolian episodes have occurred due to forest fires. Increased dune activity in the last 800 years may be related to changes in climate–vegetation parameters in association with the ‘Little Ice Age’ and to the effects of trampling by reindeer.

Paleoenvironmental Significance and Future Directions

High-latitude dune fields represent informative sedimentary environments with respect to understanding Late Glacial and Holocene paleoclimates. Wind directions can be inferred from dune morphology and orientation, the dip of forest beds, and the relation of dunes to known sediment sources. These provide evidence of past localized wind regimes and regional/continental circulation patterns. Detailed chronostratigraphic sequences of eolian deposits provide information regarding the relative importance of vegetation, niveo–eolian processes, and other cold-climate (i.e., permafrost and periglacial) processes in high latitudes.

In Europe, where the greatest number of chronostratigraphic studies has been conducted, a firm chronology of the phases of dune activity is still lacking. For many other high-latitude dune fields detailed chronostratigraphic studies are still needed. For example, many stabilized dune fields in northern Canada appear to retain morphologies indicating sediment transport directions related to Late Glacial (anticyclonic) and early Holocene (westerly) wind conditions. However, no detailed chronological studies have been conducted to determine the timing of these events.

Detailed stratigraphic studies provide the means for determining the relative importance of cold-climate eolian processes in dune development in high latitudes. From previous work, it has been suggested that cold (semi-) arid conditions, rather than snow-rich climates favor niveo–eolian transport. However, despite the significance of cold-climate processes in the development of high-latitude dune fields, few analog studies are available for understanding relict eolian features and deposits. Studies are needed that document modern cold-climate eolian processes and sedimentary structures, as well as stratigraphic studies of late Wisconsin eolian deposits in permafrost environments.

In high-latitude areas, it is important to understand how the factors of transport capacity, sediment supply, and sediment availability have affected dune fields through time, and

how this compares to mid- to low-latitude dune fields. However, the inaccessibility of many high-latitude dune fields has caused our understanding of them to lag behind that of other areas. Process studies on active transverse and parabolic dunes in high-latitude areas are required in order to understand relative transport rates and sedimentary processes in these environments, and to compare to those in coastal and low-latitude areas.

In the past, dating of eolian deposits has posed a problem due to the reliance on organic materials for radiocarbon dating, which are often scarce in eolian sediments, and which typically provided ages (up to a maximum of about 27 ka BP) that only bracket the time of eolian deposition. The advent of luminescence dating, especially OSL, provides for potentially more detailed chronologies of Mid- to Late Quaternary-aged dune fields based on the eolian sediments themselves.

Coupled with more precise chronology, improved paleoenvironmental reconstructions from high-latitude dune fields are needed by integrating eolian chronostratigraphy with adjacent lake records of biostratigraphy, oxygen-isotopes, and mineralogy.

Finally, studies are needed to determine the relation between eolian sands and various source-sediment deposits. In Alaska, for example, uncertainty still exists about the relation between surficial dune deposits and underlying subhorizontally stratified deposits, variously interpreted as eolian by some workers, and alluvial or lacustrine by others. Further studies are needed here to determine the origin and age of these eolian deposits. In addition, detailed stratigraphic studies of deposits underlying and adjacent to dune deposits, and the relative amount of reworking of source materials are needed here and in other high-latitude dune fields. In this respect, mineralogical and geochemical studies that have provided insight into the origin and evolution of dune fields in mid-to-low latitudes should be applied to high-latitude dune fields.

See also: [Dune Fields: Low Latitudes; Mid-Latitudes.](#)
[Luminescence Dating: Optical Dating.](#)

References

- Bateman MD and Godby SP (2004) Late-Holocene inland dune activity in the UK: A case study from Breckland, East Anglia. *The Holocene* 14: 2004.
- Black RF (1951) Eolian deposits in Alaska. *Arctic* 4: 89–111.
- Calkin PE and Rutford RH (1974) The sand dunes of Victoria Valley, Antarctica. *Geographical Review* 64: 189–216.
- Carson MA and McLean PA (1986) Development of hybrid aeolian dunes: The William River dune field, northwest Saskatchewan, Canada. *Canadian Journal of Earth Sciences* 23: 1974–1990.
- Dallimore SR, Wolfe SA, Matthews JV Jr., and Vincent JS (1997) Mid-Wisconsinan eolian deposits of the Kittigazuit Formation, Tuktoyaktuk Coastlands, Northwest Territories, Canada. *Canadian Journal of Earth Sciences* 34: 1421–1441.
- David PP (1981) Stabilized dunes ridges of northern Saskatchewan. *Canadian Journal of Earth Sciences* 18: 286–310.
- David PP (2000) Aeolian landforms. In: Marsh JH (ed.) *The Canadian Encyclopedia*, p. 2573. Toronto, ON: McLelland and Stewart.
- Dijkman JWA (1990) Niveo–aeolian sedimentation and resulting sedimentary structures; Søndre Strømfjord area, West Greenland and their palaeoenvironmental implications. *Permafrost and Periglacial Processes* 1: 83–96.

- Drenova AN, Timireva SN, and Chikolini NI (1997) Late glacial dune-building in the Russian Plain. *Quaternary International* 41/42: 59–66.
- Fillion L (1987) Holocene development of parabolic dunes in the Central St. Lawrence Lowland, Canada. *Quaternary Research* 28: 196–209.
- Fillion L and Morisset P (1983) The late Holocene record of aeolian and fire activity in northern Québec, Canada. *The Holocene* 1: 201–208.
- Hamilton TD, Galloway JP, and Koster EA (1987) Late Wisconsinan eolian activity and related alluvium, central Kobuk valley. *United States Geological Survey Circular* 1016: 39–43.
- Hansen EC, Arbogast AF, Packman SC, and Hansen B (2002) Port-Nipissing origin of a backdune complex along the southeastern shore of Lake Michigan. *Physical Geography* 23: 233–244.
- Kasse C (2002) Sandy aeolian deposits and environments and their relation to climate during the Last Glacial Maximum and Lateglacial in northwest and central Europe. *Progress in Physical Geography* 26: 507–532.
- Käyhkö JA, Worsley P, Pye K, and Clarke ML (1999) A revised chronology for aeolian activity in subarctic Fennoscandia during the Holocene. *The Holocene* 9: 195–205.
- Koster EA (1988) Ancient and modern cold-climate aeolian sand deposition: A review. *Journal of Quaternary Science* 3: 69–83.
- Koster EA (2005) Recent advances in luminescence dating of Late Pleistocene (cold-climate) aeolian sand and loess deposits in western Europe. *Permafrost and Periglacial Processes* 16: 131–143.
- Lea PD and Waythomas CF (1990) Late-Pleistocene eolian sand sheets in Alaska. *Quaternary Research* 34: 269–281.
- Mann DH, Heiser PA, and Finney BP (2002) Holocene history of the Great Kobuk Sand Dunes, northwestern Alaska. *Quaternary Science Reviews* 21: 709–731.
- Muhs DR (2004) Mineralogical maturity in dunefields of North America, Africa, and Australia. *Geomorphology* 50: 247–269.
- Muhs DR and Wolfe SA (1999) Sand dunes of the northern Great Plains of Canada and the United States, p. 183–197. In: Lemmen DS and Vance RE (eds.) *Holocene Climate and Environmental Change in the Palliser Triangle; A Geoscientific Context for Evaluating the Impacts of Climate Change on the Southern Canadian Prairies*, Geological Survey of Canada Bulletin, vol. 534.
- Murton JB, French HM, and Lamothe M (1997) Late Wisconsinan erosion and eolian deposition, Summer Island area, Pleistocene Mackenzie Delta, Northwest Territories: Optical dating and implications for glacial chronology. *Canadian Journal of Earth Sciences* 34: 190–199.
- Raup HM and Argus GW (1982) *The Lake Athabasca Sand Dunes of Northern Saskatchewan and Alberta, Canada. I: The Land and Vegetation*. Ottawa: National Museums of Canada Publications in Botany, No. 12, p. 96.
- Ruz MH and Allard M (1994) Foredune development along a subarctic emerging coastline, eastern Hudson Bay. *Marine Geology* 117: 57–74.
- Schokker J, Cleveringa P, and Murray AS (2004) Palaeoenvironmental reconstruction and OSL dating of terrestrial Eemian deposits in the southeastern Netherlands. *Journal of Quaternary Science* 19: 193–202.
- Selby MJ, Rains RB, and Palmer WP (1974) Eolian deposits of the ice-free Victoria Valley, southern Victoria Land, Antarctica. *New Zealand Journal of Geology and Geophysics* 17: 543–562.
- Seppälä M (2004) *Wind as a Geomorphic Agent in Cold Climates*. Cambridge: Cambridge University Press.
- Wolfe SA, Huntley DJ, David PP, et al. (2001) Late 18th century drought-induced sand dune activity, Great Sand Hills, Saskatchewan. *Canadian Journal of Earth Sciences* 38: 105–117.
- Wolfe SA and Nickling WG (1997) Sensitivity of eolian processes to climate change in Canada. *Geological Survey of Canada Bulletin* 421: 30.
- Wolfe SA, Ollerhead J, and Lian OB (2002) Holocene eolian activity in south-central Saskatchewan and the southern Canadian Prairies. *Géographie physique et Quaternaire* 56: 215–227.
- Zeeberg J (1998) The European sand belt in eastern Europe – And comparison of Late Glacial dune orientation with GCM simulation results. *Boreas* 27: 127–139.
- Zieliński P (2003) The depositional conditions of longitudinal dunes based on investigations in the western part of the Lubin Upland, SE Poland. *Landform Analysis* 4: 65–73.

Mid-Latitudes

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Introduction

Large dune fields, or sand seas, are landscapes often thought to be found only in deserts beneath the great, subtropical high-pressure zones, where subsiding air suppresses rainfall. Dune fields are also quite common in mid-latitude regions, to the north and south of subtropical deserts (Figure 1). Two major characteristics distinguish many mid-latitude dune fields from the sand seas of lower latitudes. One difference is that many of those in mid-latitudes are in semiarid, rather than arid climates, and therefore are not presently active. In this article, 'active' refers to eolian sand bodies that are not covered with vegetation, and where particles are currently being transported by the wind. Evidence of active sand transport often takes the form of well-expressed sand ripples on a dune's surface. Where sand is vegetated and particles are not being transported by the wind, a sand body is 'stable.'

A second characteristic of mid-latitude dune fields is that many are situated near mountain ranges or at least areas of relatively high relief. Glaciation in high mountains reduces local bedrock to sand-sized particles, and additional reduction of rock to sand sizes takes place through colluvial and fluvial processes quickly in terrains of steep relief. Through these processes, new sediment becomes available to feed growing dune fields much more rapidly than in areas of low relief, which characterize lower latitude regions.

Because of their occurrence in semiarid climates, mid-latitude dune fields are particularly sensitive indicators of shifts in the regional moisture balance. Sand dunes move only where there is wind, sand-sized sediment, and a lack of vegetation cover. A shift to drier conditions may result in loss of vegetation cover and activation of sand by wind. A shift back to wetter conditions may stabilize dunes because of enhanced vegetation cover. The geologic record of this alternation of dry and moist climates tends to be preserved in many mid-latitude dune fields as eolian sand–paleosol sequences. In areas downslope of high sediment production, such as mountains, fresh sediment may accumulate in a dune field during periods of eolian activity, bury the preexisting landscape, and preserve former surface soils as buried soils (or paleosols). In arid, low-latitude environments of low relief, previously stabilized eolian sand may be continually reworked without addition of much new sand, and formerly stable surfaces, which would be marked by paleosols in semiarid climates, may not be preserved.

Throughout this article, we correlate eolian dune-building events, where they can be dated or where the age can be inferred, to the marine oxygen-isotope stratigraphic framework of Martinson et al. (1987). Where the abbreviation 'MIS' occurs, this refers to 'marine isotope stage,' using the numbering system and approximate ages in Martinson et al. (1987).

Dune Fields in China

Introduction

Dune fields in China form an east–west trending belt in the mid-latitude interior (latitude 35–50° N, longitude 75–125° E; Figure 2). The total desert area is 1 308 000 km², or 13.6% of China's land area (Zhu et al., 1980). These dune fields can be divided into two parts by the north–south trending Helan Mountains (Figure 2). Dune fields, west of the Helan Mountains, include the Taklimakan, Gurbantunggut, Kumtag, Qaidam, Badain Jaran, Tengger, and Ulan Buh deserts, which are situated in inland basins or adjacent to high mountains. The formation of these dune fields is a function of their inland geographic location, far from moisture sources. Climate in these deserts is controlled by the westerlies. Dune fields east of the Helan Mountains, situated in the East Asian monsoonal zone, are mainly stabilized and semistabilized at present.

Dune Types in China

The Taklimakan Desert, which is surrounded by the Tianshan and Kunlunshan mountains, is in the Tarim Basin and dominated (85%) by active sand dunes (Figure 2). The climate is extremely arid, with a mean annual precipitation of only 15–40 mm. Local residents call the Taklimakan Desert the 'dead sand sea.' Dune types are quite diverse in this desert (Figure 3), consisting mainly of complex dune chains and transverse dunes. Dunes with heights exceeding 50 m occupy 80% of the area in its interior.

The Gurbantunggut Desert is the second largest dune field in China (48 800 km²), situated in the Junggar Basin (Figure 2). The mean annual precipitation in the Gurbantunggut Desert is 70–150 mm, which is the highest among the deserts in northwestern China. Stabilized and semistabilized sand dunes occupy 97% of this desert, and linear (or longitudinal) dunes are the dominant type (Figure 4). There are also erosional landforms in the northwestern part of the desert, called 'yardangs' (Figure 5).

The Badain Jaran Desert is the third largest dune field in China (44 300 km²). The climate is extremely arid, with a mean annual precipitation of less than 50 mm. There are two remarkable properties of the landscape in this desert. First, the desert consists mainly of complex sand hills, dune chains, and pyramidal landforms called 'star' dunes, with an average height of 200–300 m. Second, as many as 144 small lakes are found in the interdune lowlands of the desert (Zhu et al., 1980). The source of these lakes is melting water of the Qilian Mountains, ~500 km to the southwest, which reaches the desert through a subsurface fault system (Chen et al., 2004). The Kumtag Desert also has complex dune chains and star-shaped pyramidal dunes. In this sand sea, heights of the dunes vary between 100 and 200 m (Figure 6).

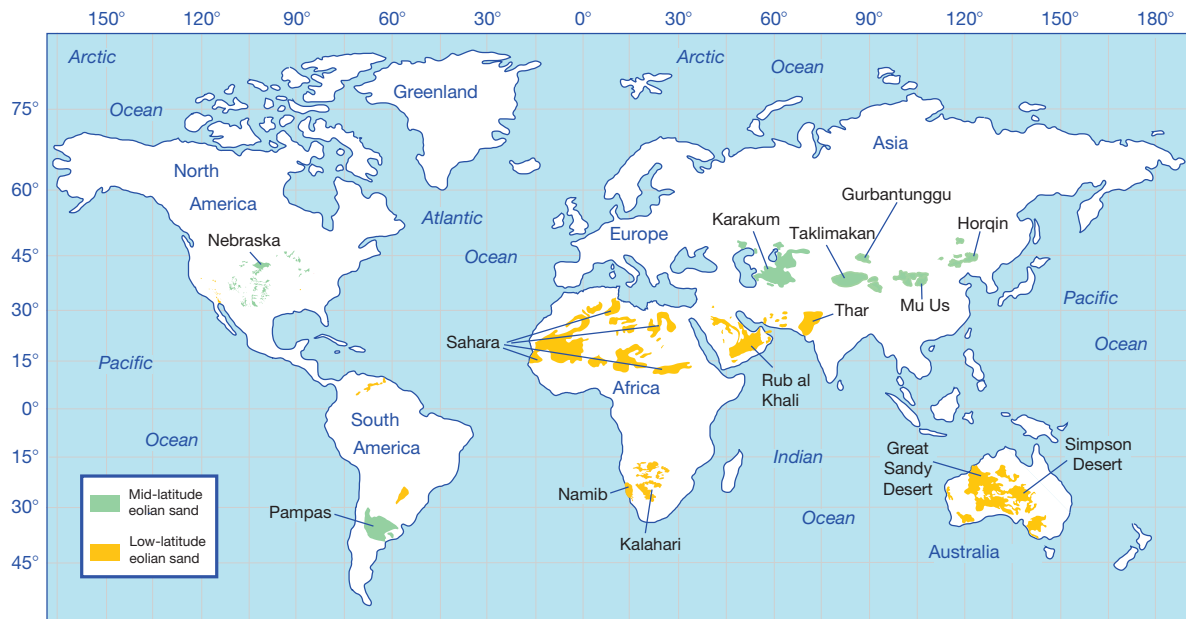
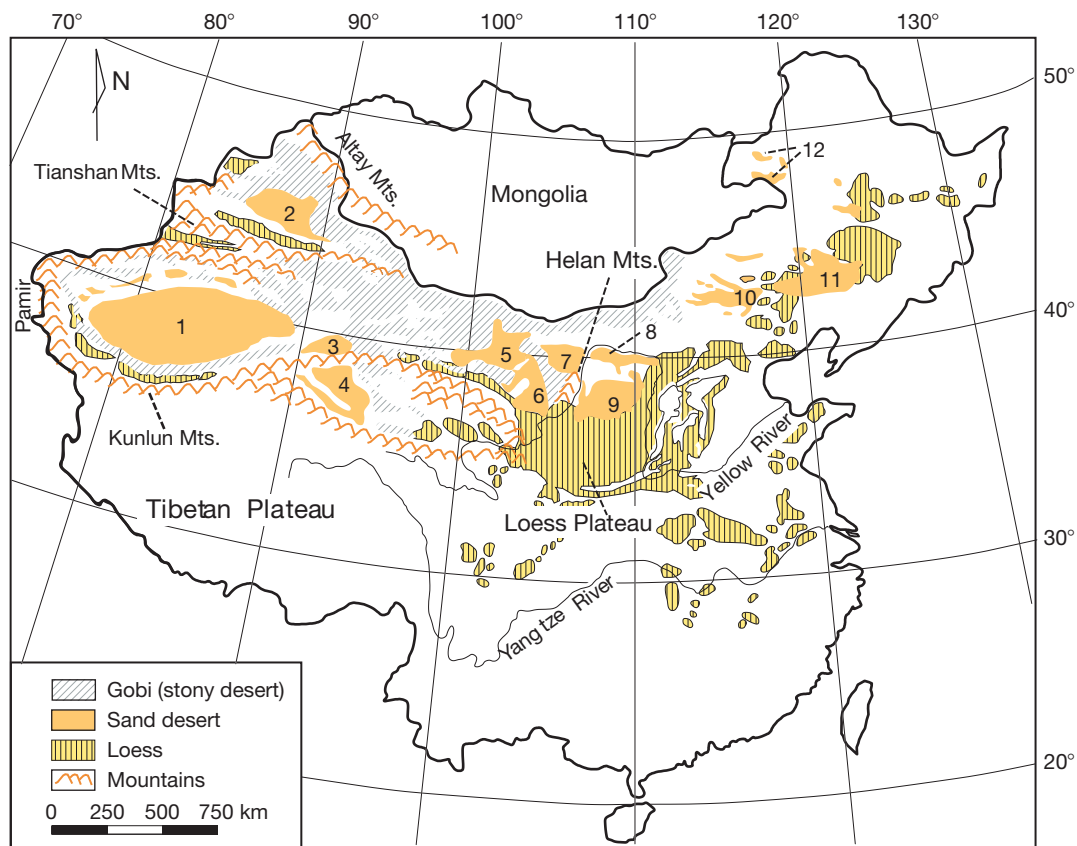


Figure 1 Map showing the global distribution of eolian sand, whether active or stabilized. Sources: North America: [Muhs and Zárate \(2001\)](#), [Muhs and Holliday \(1995\)](#), and [Thorp and Smith \(1952\)](#); South America: [Muhs and Zárate \(2001\)](#) and sources therein; Africa: [Wilson \(1973\)](#), [Lancaster \(1988\)](#), and [Thomas and Shaw \(1991\)](#); Central Asia and India: [Wilson \(1973\)](#); China: Sun and Muhs (this article) and sources therein; Australia: [Bowler et al. \(2001\)](#). Mid-latitude dune fields reviewed in this article are shown in green; low-latitude dune fields reviewed by N. Lancaster (this volume) are shown in yellow.



- 1 Taklimakan 2 Gurbantungut 3 Kumtag 4 Qaidam 5 Badain Jaran 6 Tengger
7 Ulan Buh 8 Hobq 9 Mu Us 10 Hunshandake 11 Horqin 12 Hulun Buir

Figure 2 Map showing gobi (stony desert), dune fields, and loess distribution in China.



Figure 3 Sand sea of the Taklimakan Desert. About 80% of this dune field is active at present and precipitation-to-potential-evapotranspiration (P/PE) values are generally less than 0.2 (Wang et al., 2005).



Figure 4 Semistabilized linear dunes in the Gurbantunggut Desert.



Figure 5 Deflation landforms (yardangs) in the northwestern Gurbantunggut Desert.

The Tengger Desert is the fourth largest desert in China (42700 km²). Sand dunes occupy 71% of the desert, and 93% of the dunes are active. The dominant dune type is transverse dune, with heights usually between 10 and 20 m.

The Qaidam Desert is located in the Qaidam Basin (Figure 2). Yardangs in its interior make up 67% of the desert's



Figure 6 Star-shaped pyramidal sand dunes in the Kumtag Desert.



Figure 7 Deflation landforms (yardangs) in the interior of the Qaidam Desert.



Figure 8 Stabilized or semistabilized sand dunes are the dominant dune types in the deserts, distributed to the east of the Helan Mountains.

area (Figure 7). Sand dunes are mainly distributed in the marginal areas, dominated by transverse dunes.

Transverse dunes dominate other dune fields in China, especially those east of the Helan Mountains. Most dunes are stabilized or semistabilized (Figure 8), and vary in height between 5 and 15 m.

Origin of Chinese Dune Fields

Origins of the dune fields in China can be divided into two categories. Dune fields to the west of the Helan Mountains are situated in the inland basins near the Tianshan, Kunlunshan, and Qilianshan mountains. The mean annual precipitation varies from 20 to 150 mm. Geomorphic processes in the adjacent mountains (especially glacial grinding) have played an important role in producing the vast amounts of sand-sized materials for the formation of these dune fields. Elevations of the Tianshan, Kunlunshan, and Qilianshan mountains range mostly from ~3000 m to ~5500 m above sea level. The high elevation gives rise to extremely cold climatic conditions, thus favoring glacial grinding and frost weathering. This leads to the physical weathering of rocks in the surrounding regions, and thus the production of clastic materials (Sun, 2002a,b). The high relief and steep gradients of these mountains give rise to high potential energy of melting water. Thus, the clastic materials can be transported to the piedmont areas, forming large alluvial fans. Examination of multispectral Landsat thematic mapper (TM) imagery indicates that a series of huge alluvial fans have formed in the piedmont of the Qilian Mountains (Figure 9). The zonal distributions of the gobi (stony desert), dune fields, and loess bodies are the result of wind sorting of the piedmont fluvial materials (Sun, 2002a).

Deserts to the east of the Helan Mountains are situated in the high plains of northeastern China. Because this region lies in the East Asian monsoonal zone, the mean annual precipitation ranges from 200 to 450 mm. Most of the dunes are stabilized and semistabilized. However, due to poor land-use practices in historic time, active sand dunes also occur in these deserts, largely due to reworking of last glacial age (MIS 2) sand dunes (Sun, 2000). The provenance of the sediments in these dunes is dominantly from fluvial sediments of several large rivers.

Age of Desert Dunes

One approach for studying long-term dune field history is to examine the chronology of loess deposition, as there are few



Figure 9 Landsat thematic mapper (TM) image of alluvial fans in the piedmont of the Qilian Mountains, China. Sediments in these large alluvial fans are important sources of the dune sand and loess in China.

age estimates of the oldest dunes in China. In the northern piedmont of the Kunlun Mountains, dust derived from the Taklimakan Desert accumulates as loess sediment. Because the Taklimakan Desert lies within the rainshadow of the Tibetan Plateau, which began rising in the late Cenozoic (Gansser, 1964; Harrison et al., 1992; Molnar and Tapponnier, 1975), it is possible that dunes have existed in this basin since the latest Cenozoic.

Recent studies of eolian silt on the Chinese Loess Plateau indicate that the source areas are the gobi (stony) desert in southern Mongolia and the adjoining gobi and sandy deserts (including the Badain Jaran Desert, Tengger Desert, Ulan Buh Desert, Hobq Desert, and Mu Us Desert) in China (Figure 10). These gobi and sand deserts only serve as 'holding areas' of dust and silt, rather than the original source of loess particles. Mountain processes (especially glacial grinding) in the high elevations of Asia have played an important role in producing the vast amounts of loess-sized material for forming the Loess Plateau (Sun, 2002a). Because the basal age of the loess in the Loess Plateau is 2.58 Ma (Liu, 1985; Liu and Ding, 1993), and the loess materials are transported by the near-surface winds from the deserts listed above, these basins have been arid environments for at least 2.58 Ma.

Relations of Desert Dunes to Loess Deposits

Because Chinese loess is derived from the deserts, desert expansions and contractions, in response to Quaternary glacial-interglacial cycles, may be recorded by eolian deposits at the margins of deserts. Among the 12 deserts in China, only the Mu Us Desert is geographically proximal to the Chinese Loess Plateau. Eolian deposits on the southern margin of the Mu Us Desert are ideal for studying the response of the desert to climatic changes of the Quaternary.

The eolian deposits in the southern marginal Mu Us Desert are characterized by alternations of eolian sand, loess, and paleosols, making it a potential region for recording the ecosystem changes during the Quaternary. Eolian sand accumulated when the Mu Us Desert expanded southeasterly to the Loess Plateau during glacial maxima (Sun et al., 1999).

Paleosols developed under warm and humid conditions of interglacial periods, as a response to the enhanced summer East Asian monsoon winds. The existence of the paleosols in the desert marginal regions indicates that the dunes in the Mu Us Desert were stabilized by vegetation during interglacial maxima.

The loess beds imply intermediate climate conditions, between full glacial and full interglacial climates. The nearly unweathered character of these loess layers suggests that they were deposited under the dominant winter monsoon winds but with greatly reduced wind energy than what is necessary for sand transportation.

The most representative section of these complex eolian deposits is located at Shimao, in the southern marginal Mu Us Desert (Sun et al. (1999); see Figure 10 for location). The eolian sequences of this section consist of 40 eolian stratigraphic units, of which 13 are sands, 13 are loess, and 14 are paleosols. Based on paleomagnetic polarity studies and a comparison with the composite marine oxygen-isotope records (Shackleton and Pisias, 1985; Shackleton et al., 1990),

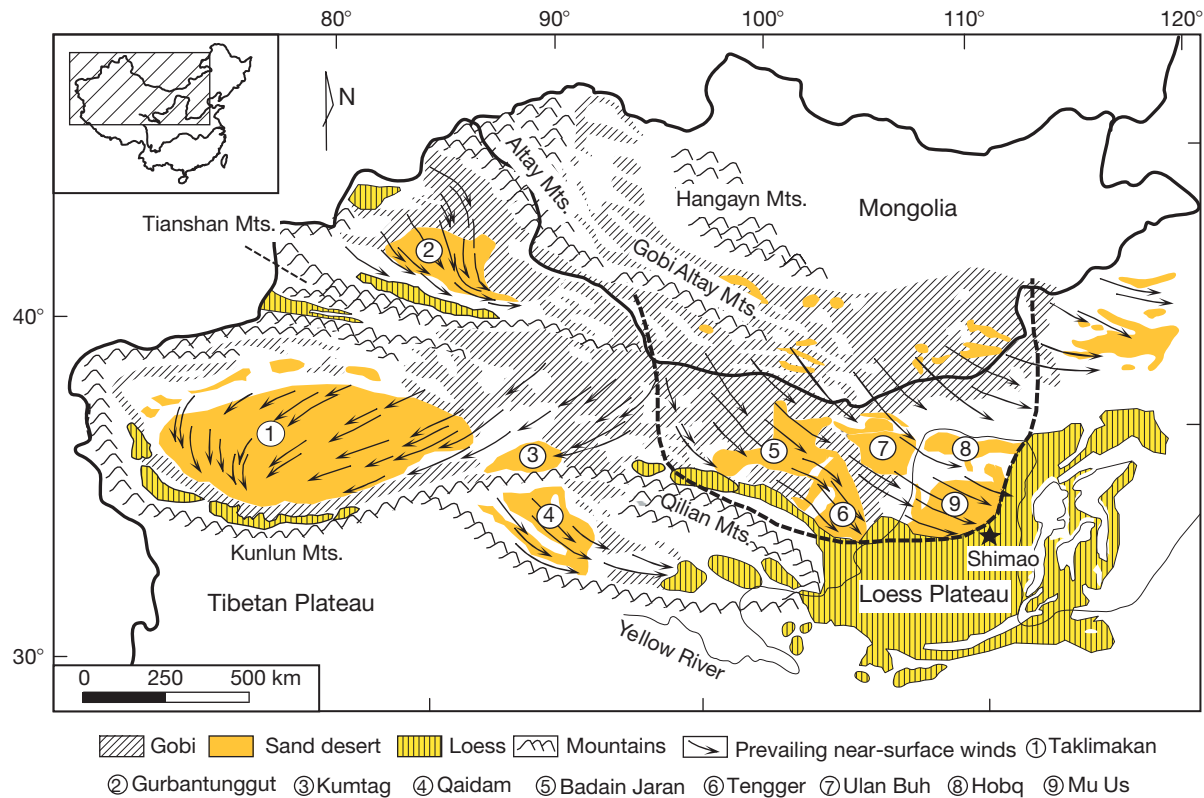


Figure 10 Map showing the near-surface wind directions (arrows) based on dune orientation in the studied regions. Area circled by dashed line indicates the source region of the Loess Plateau (Sun, 2002a).

the basal age of this section is about 0.63 Ma (Figure 11). The eolian deposits at Shimao and their correlation with the marine records yield two important implications.

First, during the past 0.63 Ma, there have been at least 13 times of desert expansion in the Mu Us Desert, and the desert expansion occurred not only during glacial maxima, but also during the cold substages of interglacial periods, such as those of MIS 7, 13, and 15. Intercalated eolian sand in the interglacial paleosols S2 and S5 support this conclusion (Figure 11).

A second major implication is that the stratigraphy of the section at Shimao shows that except for the Holocene paleosol (S0), all the other paleosols (S1, S2, S3, S4, and S5) are pedocomplexes, with eolian sand or loess beds intercalated within the soil complexes. The sand or loess intercalations indicate that the warm climate during each interglacial time was interrupted by cold-dry episodes.

Desert Distributions During the Holocene Optimum and the Last Glacial Maximum in China

Recent studies indicate that dust from Asia dominates the entire Pacific Ocean north of the intertropical convergence zone (Duce et al., 1991; Merrill et al., 1989). China is considered as one of the two dominant source regions for atmospheric mineral aerosols originating from Asia (Rea, 1994). The waxing and waning of the summer monsoon during the Quaternary must have had a great effect on the expansion and contraction of the deserts in northeastern China, which in turn modulated the flux of dust from Asia to the Pacific Ocean.

Former Extent of Deserts During the Last Glacial Maximum

Variation in global ice volume may have played a key role in modulating and pacing the strength of the East Asian monsoon (Ding et al., 1994). During the Last Glacial Maximum (LGM), the enlarged high-latitude ice sheets of the Northern Hemisphere greatly strengthened the Siberian High, giving rise to an enhanced cold-dry northwest winter monsoon. Under winter monsoonal winds, strong eolian erosion, transportation, and dune building characterized the arid and semiarid regions of northern China.

Since the 1980s, 44 profiles with LGM dune sands and overlying Holocene sandy loam soils have been found in northeastern China, in the monsoonal climatic zone. Thus, during the LGM, active sand dunes occurred not only in the arid inland areas of western China, but also in the semiarid to subhumid areas, east of the Helan Mountains. The existence of the buried LGM sands implies that the southern limit of the LGM desert extended south to nearly latitude 36° N, and east to longitude 125° E (Sun et al. (1998); Figure 12(a)).

Former Extent of Deserts During the Holocene Optimum

During the mid-Holocene climatic optimum, from ~9 to ~3 ka, the East Asian summer monsoon was strong. Sandy loam soils on stabilized dunes are widely distributed in desert regions, especially east of the Helan Mountains. Radiocarbon ages and optically stimulated luminescence (OSL) dating indicate that these loamy soils developed mainly between 3.0 and

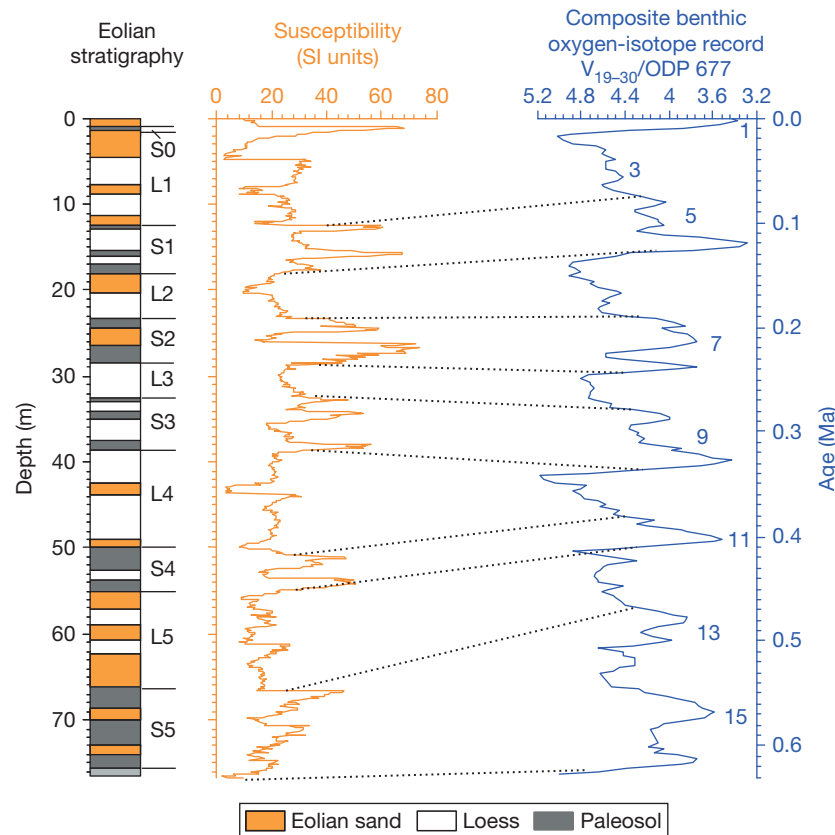


Figure 11 Pedostratigraphy and the magnetic susceptibility curve of the Shima section and its correlation with the $\delta^{18}\text{O}$ records from the Pacific Ocean. The isotope curve is a composite record of V_{19-30} (0–0.34 My, [Shackleton and Pisias, 1985](#)) and ODP 677 (0.34–0.63 My, [Shackleton et al., 1990](#)). Modified from Sun JM, Ding ZL, Rokosh D, and Rutter, N (1999) 580 000 year environmental reconstruction from eolian deposits at the Mu Us Desert margin, China. *Quaternary Science Reviews* 18: 1351–1364.

9.0 ka BP, coincident with the Holocene optimum ([Li et al., 2002](#); [Zhou et al., 2001](#)). Pollen analysis of these soils indicates that the vegetation was dominated by steppe taxa, implying that sand dunes in northeastern China were stabilized during the Holocene optimum ([Sun et al., 1998](#)).

In contrast to the deserts in northeastern China, the distinctive Holocene sandy loam soil has not been found in the Taklimakan, Qaidam, and Badain Jaran deserts. Its absence implies that during the Holocene optimum, the much-enhanced summer monsoon did not penetrate into the interior of the western desert regions, and thus active dune fields existed then (and still exist today). Compared with the desert of the LGM, the extent of desert in the Holocene was greatly reduced, with the eastern limit of desert retreating from longitude 125°E (LGM time, MIS 2) to about 105°E , a distance of well over 1600 km ([Figure 12\(b\)](#)).

Dune Fields in Central Asia and South America

Central Asia

To the west of China, dune fields occupy large areas of Central Asia, in the desert basins of Kazakhstan, Uzbekistan, and Turkmenistan. The largest dune fields are situated mostly southeast of the Caspian Sea, Aral Sea, and Lake Balkhash

([Figure 13](#)). [Wilson \(1973\)](#) estimates these dune fields to encompass $\sim 380\,000\text{ km}^2$ (Karakum sand sea, southeast of the Caspian Sea), $\sim 276\,000\text{ km}^2$ (Kyzylkum sand sea, southeast of the Aral Sea), and $\sim 38\,000\text{ km}^2$ (Myunkum sand sea, southeast of Lake Balkhash). The region is semiarid to arid, and precipitation ranges from ~ 100 to 400 mm year^{-1} ([Goudie, 2002](#); [Lioubimtseva, 2002](#)).

The main dune forms in the region are longitudinal (linear) dunes, which are easily visible on Landsat imagery ([Figure 13](#)). [Goudie \(2002\)](#) reports that the linear dunes are as much as 30 m high and are separated by interdune areas floored with clay-rich sediments. [Wilson \(1973\)](#) indicates that most or all of these dunes are active at present. At the broadest scale, orientations of the dunes ([Figure 13](#)) agree fairly closely with modern, dominant wind directions ([Letollé and Mainguet, 1993](#)), but in detail, there are some differences, as pointed out by [Lioubimtseva \(2002\)](#). Southeast of the areas of linear dunes, barchans are the dominant landforms, and many of these are also active at present ([Goudie, 2002](#); [Lioubimtseva, 2002](#)).

High mountain ranges in Afghanistan, Tajikistan, and China are drained by rivers that flow westward to the basins occupied by the Caspian and Aral Seas. The Volga River, the largest river in Europe, carried sediment from the Fennoscandian ice sheet that covered northern Europe during the LGM (MIS 2). Thus, abundant sediment is delivered to the desert

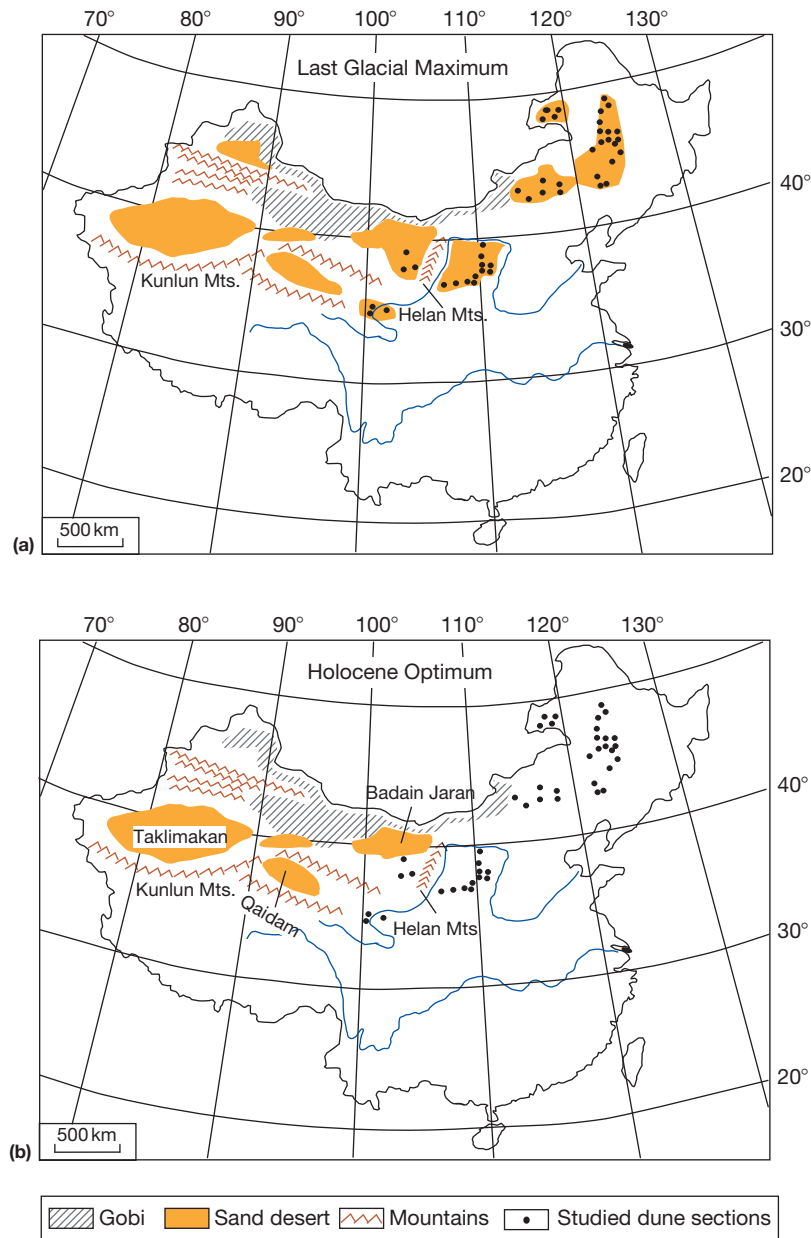


Figure 12 Desert distributions in China during the Last Glacial Maximum (a) and the Holocene Optimum (b) the filled black dots indicating the sites where Holocene loam soils and the underlying sand dunes of the LGM are found (Sun et al., 1998).

basins of Central Asia at present and it is likely that even more sediment was transported there during glacial periods of the Pleistocene. Because sources of specific dune fields in the region are little studied, it is not known which rivers are most important as dune sand sources.

The paleoclimatic significance of dunes in Central Asia is difficult to assess, as there has been little study of dune stratigraphy, and numerical ages are lacking. However, as with China, some comparisons can be made with chronologies developed for loess deposits that are downwind of the dune fields. Loess deposits are extensive in the southern margins of the desert regions in Central Asia (Figure 13). Dodonov (1991) suggests that this loess is derived from a combination of glaciogenic sources in the neighboring mountains and dust sources in the

desert basins. Long-distance transport from desert basin sources is supported by the relatively fine grain size of loess particles in sections exposed in Tadjikistan (Frechen and Dodonov, 1998). Luminescence dating indicates that there was considerable loess deposition during both the early (MIS 4) and later (MIS 2) parts of the last glacial period (Frechen and Dodonov, 1998). These data suggest that winds were strong and desert basin source areas were relatively dry and minimally vegetated during the last glacial period. Nevertheless, during the early part of the last glacial period (MIS 4), the Caspian Sea was connected to the Aral Sea, whereas during the last interglacial period (MIS 5), it was of lesser extent than during the Holocene (Chepalyga, 1984). If the level of the Caspian Sea is at least a partial indicator of the overall moisture

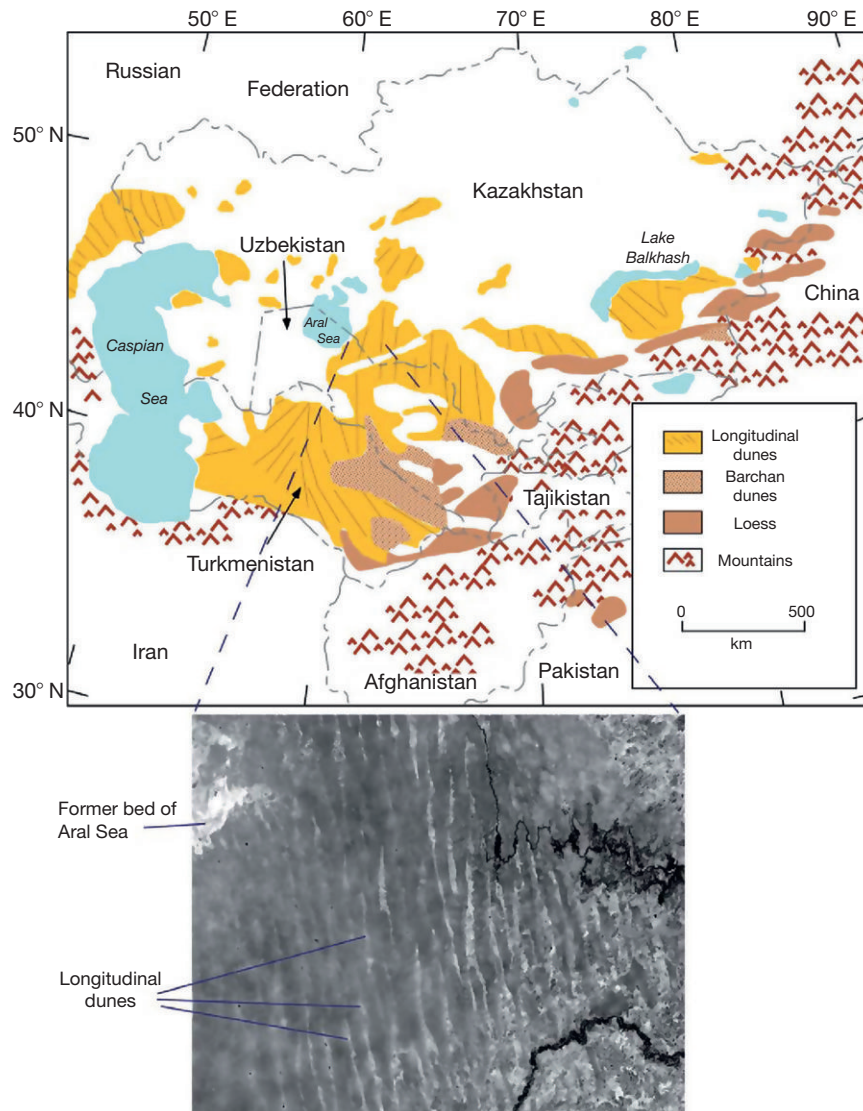


Figure 13 Upper: map showing the distribution of eolian sand and loess in Central Asia (redrawn from data in Lioubimtseva (2002) and sources therein for sand; from Dodonov (1991) for loess). Lower: Landsat TM image (path 159, row 029; band 1, visible blue; image taken 5 August 1987) of longitudinal dunes southeast of the Aral Sea, Kazakhstan and Uzbekistan.

balance of the arid basins of the region, these observations suggest that the dune fields may be active primarily during interglacial, rather than glacial periods.

South America

There is a large area of stabilized dunes in the mid-latitudes of South America, primarily in the Pampas region of Argentina. The precipitation gradient in this region is steep, with present mean annual precipitation ranging from less than 500 mm (southwest) to more than 1000 mm (northeast). In parts of Argentina (particularly in La Pampa and Buenos Aires provinces), there are longitudinal (linear) dunes over 100 km long and 2–5 km wide (Figure 14). Iriondo (1999) refers to this large dune field as the Pampean Sand Sea. The largest dunes in this sand sea are arranged as extensive southwest-to-northeast arch-like forms and are presently stabilized by vegetation. Although

the longitudinal dune forms are easily recognizable on Landsat TM imagery (Figure 14), the low relief of these features makes them difficult to identify in the field (Martinez et al., 2001). A large field of parabolic dunes, most of them also stabilized and oriented in a southwest-to-northeast direction, is situated south of the longitudinal dunes (Figure 14). The parabolic dunes may have developed from reworking of preexisting sands.

Zárate and Blasi (1993) present a model for eolian deposition in Argentina that explains the origin of both dunes and loess. They report sedimentological and mineralogical data that lead them to infer that the main dune sand sources, as well as loess sources, are the Colorado and Negro River floodplains, which are situated to the southwest of the Pampean Sand Sea. The Colorado and Negro rivers drain the Andes, and thus volcanoclastic sediments from these mountains are the ultimate source of much or all of the sediment in the sand sea and loess bodies. Zárate and Blasi (1993) point out that the

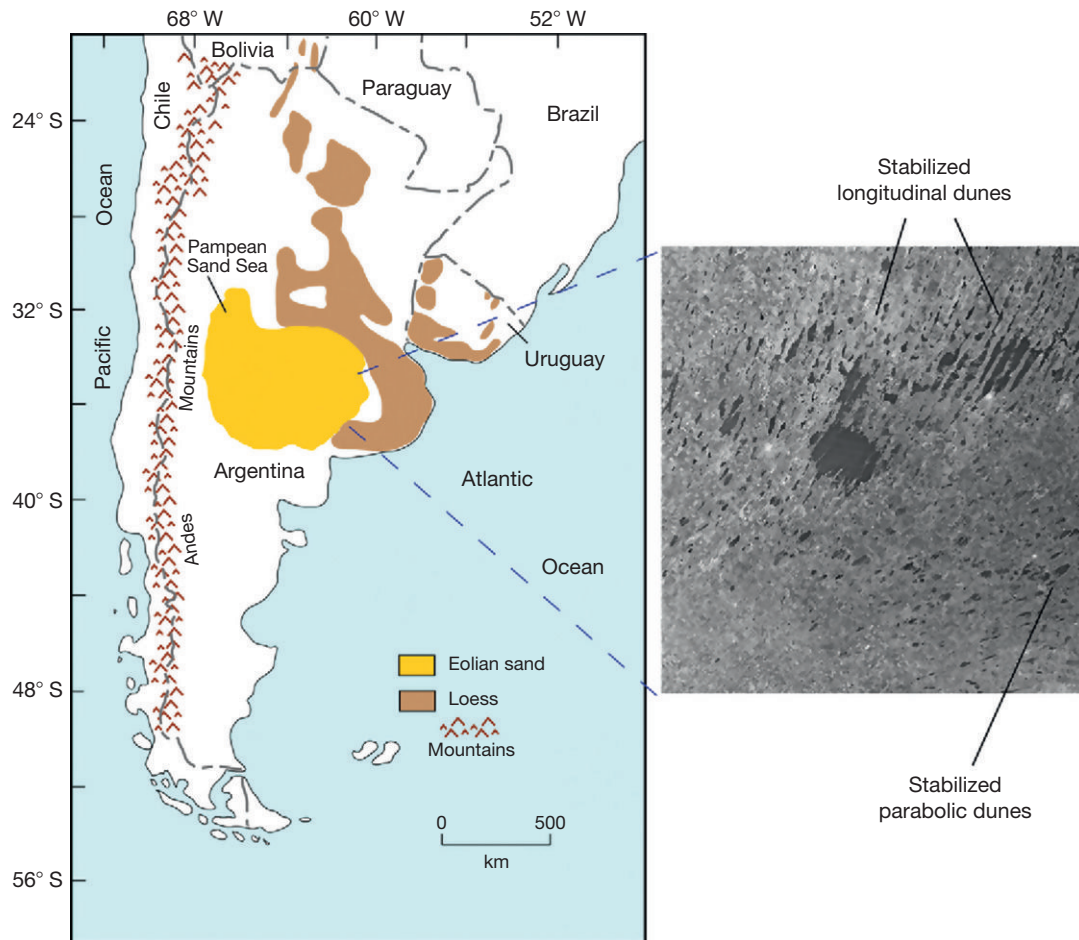


Figure 14 Left: map showing the distribution of eolian sand and loess in southern South America (taken from Zárate (2003), and sources therein). Right: Landsat TM image (path 227, row 085; band 1, visible blue; image taken 17 February 1988) of linear and parabolic dunes in Buenos Aires province, Argentina.

fluvial regime supplying the particles was dependant, to a large degree, on the extent of glacial cover in the Andes. The Colorado and Negro rivers are presently underfit and have a meandering pattern. In the last glacial period, they apparently had a broad, braided pattern, and during seasonal melting periods, both rivers carried large volumes of meltwater and sediment. During periods of low flow, sediment availability was large, and provided the source of both sand to build dunes and silt for loess.

There are few data on the ages of eolian sand in Argentina. Iriondo (1999) considers the longitudinal dunes that occur in northwestern Buenos Aires province and surrounding areas to be of late Pleistocene (early last glacial, or MIS 4) age. Thermoluminescence (TL) ages reported by Kröhling (1999) indicate that the earliest part of the Pampean Sand Sea may have formed during the early part of the last glacial period (MIS 4), with reworking of this sand occurring between ~52 and ~36 cal ka BP. Kröhling (1999) also recognizes an eolian sand facies in a loess-dominated unit called the Tezanos Pinto Formation. Bracketing TL ages of ~32 and ~9 cal ka BP are reported for the loess facies of the Tezanos Pinto Formation. If these ages are correct, they imply a major period of eolian sand movement during the latter part of the last glacial period (MIS 2).

The inference of eolian sand movement occurring during major glacial periods (MIS 4 and 2) is consistent with the sediment origin model of Zárate and Blasi (1993). However, eolian sand movement may also have occurred during the Holocene. For example, Iriondo (1999) considers the parabolic dunes that are found to the southeast of the longitudinal dunes (Figure 14) to be of late Holocene age. Several workers (Iriondo, 1999; Kröhling, 1999; Zárate and Blasi, 1993) use loess and dune formation to infer a relatively dry period for the Pampas region in the late Holocene, although not as dry as during the last glacial period.

Dune Fields in North America

Introduction

Dunes occupy a considerable portion of the mid-latitude regions of North America, particularly in the Great Plains region of the central United States and the Colorado Plateau region of the southwestern United States (Figure 15). The dune fields of the Great Plains have been intensively studied since the 1980s and much is now known about their geographic extent, geomorphic expression, source sediments, chronology, and

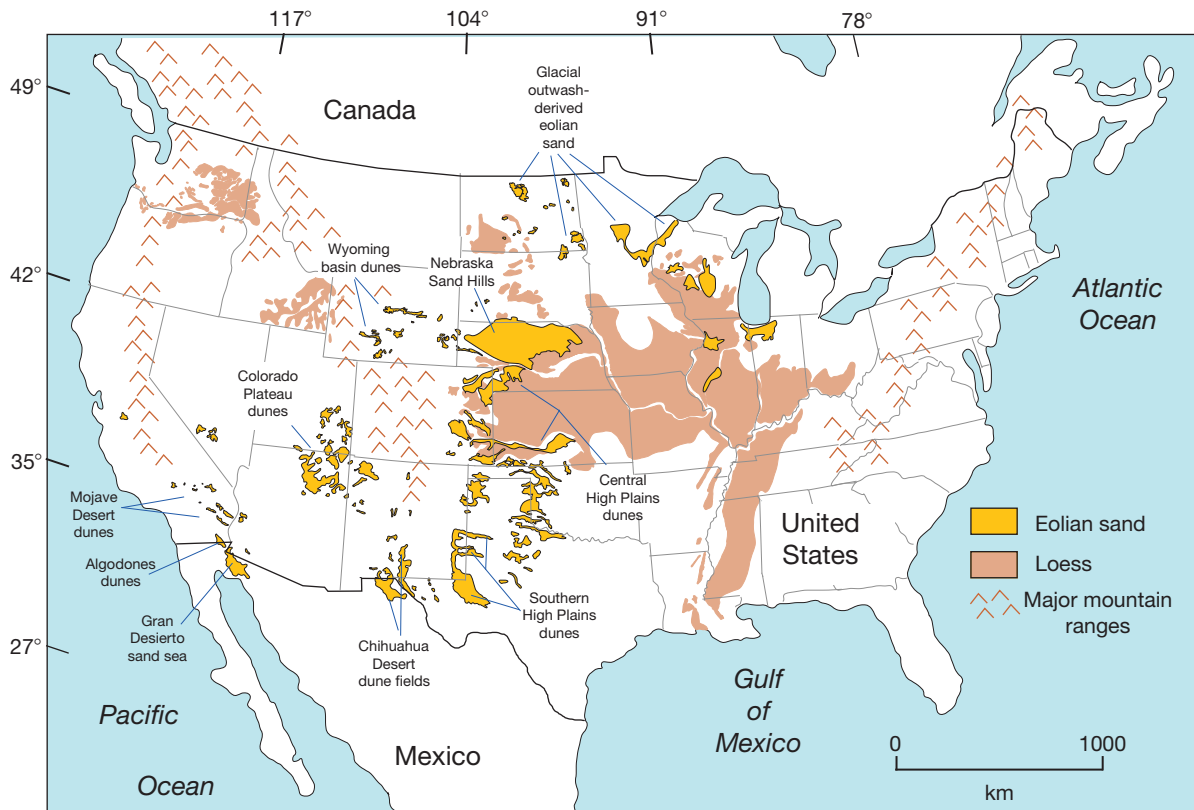


Figure 15 Map showing the distribution of eolian sand and loess in the United States and adjacent parts of Mexico. Redrawn from Thorp and Smith (1952), Muhs and Holliday (1995), Muhs and Zárate (2001), Bettis et al. (2003) and sources therein.

paleoclimatic significance. The dunes of the Colorado Plateau have been studied less extensively, but work in progress (e.g., Redsteer and Block (2004)) promises important new information about these smaller sand seas.

Dune Forms in Mid-latitude Regions of North America

There is a wide range of dune forms in the Great Plains region of North America. The greatest variety of dune forms is found in the Nebraska Sand Hills. In this large ($\sim 50\,000\text{ km}^2$) dune field, barchanoid ridges, many 10–20 km long, are present over extensive areas (Figure 16). Most of these barchanoid ridges have smaller, superimposed parabolic dunes, indicating multiple episodes of eolian deposition. In other areas of the Nebraska Sand Hills, barchans are the main dune form (Figure 16), but again superimposed parabolic dunes indicate that the barchan forms were generated during earlier phases of dune formation. Linear (longitudinal) dunes are present over extensive areas of the Nebraska Sand Hills, many of which appear to be superimposed on older dune forms (Figure 16). The Nebraska Sand Hills subregion is semiarid and mean annual precipitation ranges from about 400 mm in the western part to about 600 mm in the eastern part. With the exception of small blowouts on dune crests, all of these dunes are presently stabilized by vegetation (Figure 17).

Dune fields farther south, in Colorado, Kansas, New Mexico, and Texas, are numerous, but are much smaller than the Nebraska Sand Hills. In eastern Colorado, for example, the

total area occupied by dunes is $\sim 25\,000\text{ km}^2$ (Madole et al., 2005). As with Nebraska, the region is semiarid and mean annual precipitation ranges from about 300 mm (west) to about 770 mm (east). The most common landforms in these dune fields are stabilized parabolic dunes, often separated by stabilized eolian sand sheets (Figure 18). In many places, compound dunes are common, with small parabolic dunes superimposed on larger, older parabolic dunes.

Still farther south, however, in the Chihuahuan Desert of southern New Mexico, western Texas, and northern Mexico, active barchanoid ridges, barchans, and parabolic dunes are found (Figure 18). Mean annual precipitation in this region is lower than on the Great Plains, and ranges from about 220 to 320 mm. Farther west, on the Colorado Plateau, many different dune forms exist in close proximity to one another. For example, in the Moenkopi Plateau of northeastern Arizona, barchan, linear, and parabolic dunes can all be found within a relatively small area (Figure 19). Mean annual precipitation on this high plateau region is lower still, and ranges from about 150 to 300 mm. As a result, while some dunes are stable, many are active. In places, for example, stabilized eolian sand sheets separate active linear dunes (Figure 19).

Dune Origins in North America

In the past, it was commonly assumed that dunes in most parts of central North America, particularly the Great Plains, were derived from local sandstone bedrock, notably the Tertiary

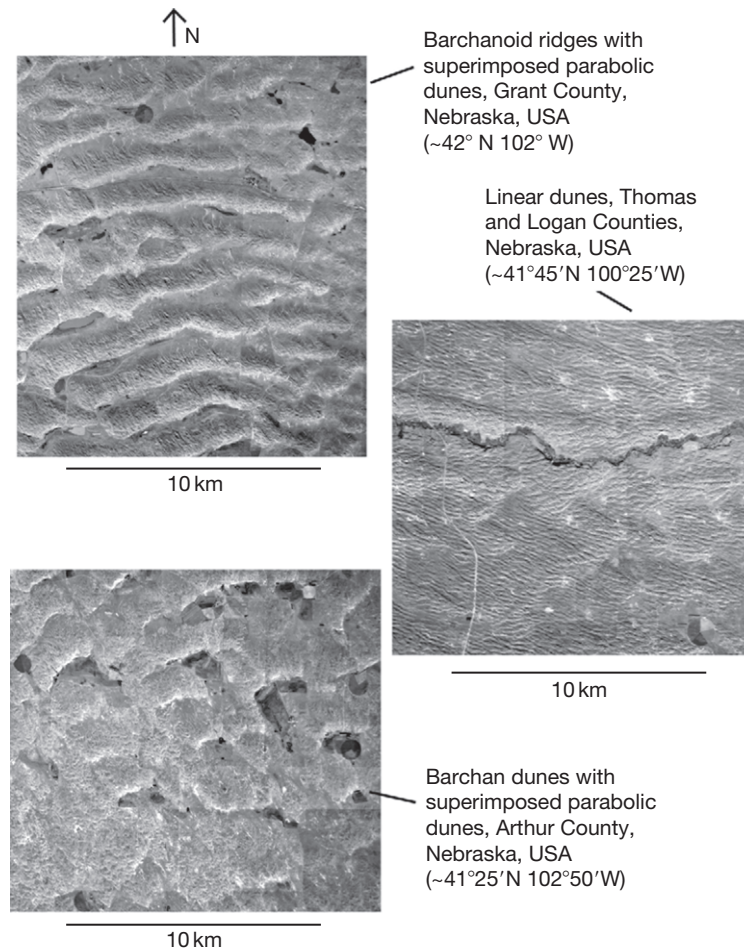


Figure 16 Dune landforms in the Nebraska Sand Hills, USA. All aerial photographs are from the US Geological Survey National High Altitude Aerial Photography (NHAP) program. Frames: Grant County (HAP 182-77, 25 October 1984); Thomas and Logan Counties (HAP 152-52, 31 October 1981); Arthur County (HAP 242-184, 26 September 1985).

Ogallala Formation. This formation occurs over much of the region, crops out close to the land surface, and is rich in clean sand. It is likely that many early investigators were struck by the close geographic match between the distribution of dune fields and the Ogallala Formation and inferred a genetic link. Nevertheless, much or all of the Ogallala Formation is anchored by a thick, well-cemented calcrete caprock and is not easily eroded.

Later investigators questioned the importance of bedrock as a source sediment for Great Plains dunes. Mineralogical and geochemical studies suggest that in much of the Central High Plains of Colorado and Kansas, dune sediments are derived from major rivers that head in the Rocky Mountains and flow eastward onto the Great Plains. In the Central High Plains, these fluvial sources include the South Platte River, which supplies sediment to dune fields in northeastern Colorado and southwestern Nebraska (not the Nebraska Sand Hills, however), and the Arkansas River, which supplies sediment to dunes in southeastern Colorado and western Kansas (Arbogast, 1996; Muhs et al., 1996). In the Southern High Plains of Texas and New Mexico, dune fields appear to be derived mainly from an older Pleistocene eolian sheet sand called the Blackwater Draw Formation (Muhs and Holliday, 2001), which is, in turn, probably derived from the Pecos River.

One of the greatest unanswered questions about dune fields in North America is the origin of the Nebraska Sand Hills, the largest dune field on the continent. No major rivers are situated upwind (to the northwest) of this dune field. The best available evidence, based on isotopic studies, is that it has multiple sources, most of which are sand-rich Tertiary rocks that underlie, or occur upwind of the dune field (Aleinikoff et al., 1994). Whatever their ultimate sources, dunes of the Nebraska Sand Hills are more mineralogically mature (i.e., quartz rich and depleted in feldspars) than any of their possible source sediments (Muhs, 2004; Muhs et al., 1997). One explanation for this maturity is that most feldspars have been reduced to silt-sized sediments through ballistic impacts, an idea that has support in both theory and laboratory evidence (Dutta et al., 1993). If this explanation is correct, it suggests that the Nebraska Sand Hills dune field has a long history of repeated dune activity, with little input from fresh sand sources.

Dune Ages in North America

For decades, many investigators assumed, with no numerical age control, that dunes in the Great Plains of the United States were last active during full glacial time (MIS 2). More recent studies

Western Nebraska Sand Hills, USA
 $P \sim 490 \text{ mm year}^{-1}$; $P/PE = 0.76$



Andrews dune field, western Texas, USA
 $P \sim 350 \text{ mm year}^{-1}$; $P/PE = 0.36$



Figure 17 Contrasting degrees of dune activity in the Great Plains of the United States as a function of moisture balance (P/PE) in the western Nebraska Sand Hills (upper) and Andrews dune field, Texas (lower). Photographs by D.R. Muhs.

confirm that the last glacial period was indeed a time of eolian sand activity over much of the region (Goble et al., 2004; Loope et al., 1995; Madole et al., 2005; Muhs et al., 1996; Swinehart and Diffendal, 1990). However, many studies also show that much eolian sand in the Great Plains region is of late Holocene age. In the Great Plains, the areal extent of late Holocene eolian sand is much greater than that of last glacial sand. For example, dunes of the Nebraska Sand Hills cover $\sim 50,000 \text{ km}^2$, yet all have only simple A/AC/C soil profiles (Entisols) that develop in a few hundred or few thousand years. This observation, along with late Holocene radiocarbon and OSL ages at numerous localities, suggests that most or all of the Nebraska Sand Hills region has been active at some time in the late Holocene. However, all parts of the dune field were not necessarily active simultaneously.

late Holocene eolian sand deposition has now been well documented in Nebraska, Colorado, Kansas, Texas, and New Mexico (Arbogast, 1996; Forman et al., 2005; Goble et al., 2004; Holliday, 2001; Madole et al., 2005; Muhs et al., 1997; Swinehart and Diffendal, 1990). Most of these studies have stratigraphic data indicating multiple periods of eolian activity in the late Holocene (Figure 20). The small number of radiocarbon and OSL ages and their large analytical uncertainties preclude testing hypotheses of regionally synchronous activity. However, these observations show that eolian sand in this region has been active under modern climatic regimes. Explorers in the nineteenth century observed active dune sand in many parts of the Great Plains where it is now stable (Muhs and Holliday, 1995).

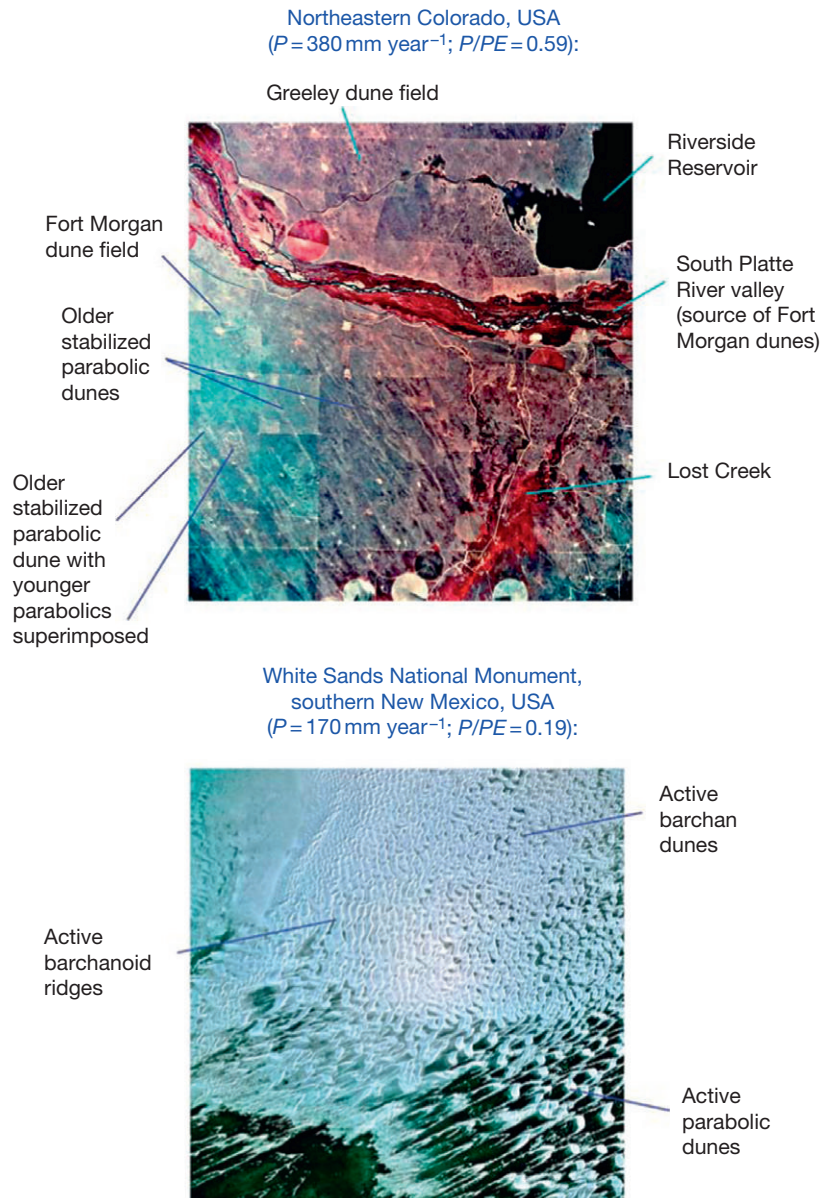


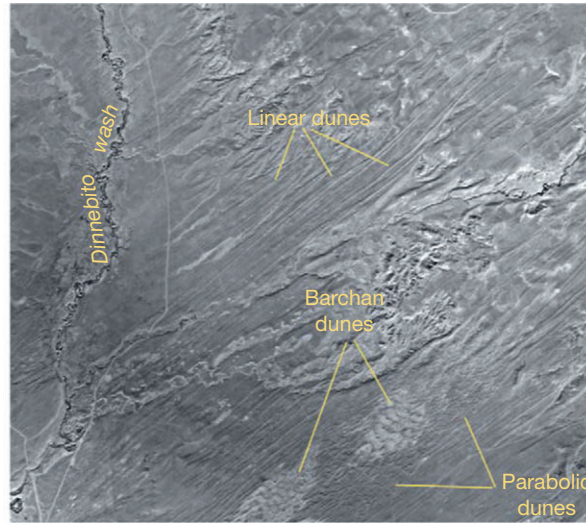
Figure 18 Dune forms in the Central High Plains of Colorado (upper) and northernmost Chihuahuan Desert of New Mexico (lower). Upper photograph is false-colored IR aerial photograph (USGS NHAP 61-107, 22 June 1985); lower photograph is natural-color aerial photograph (USGS NAPP 9524-114, 6 October 1996).

Paleoclimatic Significance of Dunes in North America

One of the most important paleoclimatic inferences made from stabilized, mid-latitude dune fields is that periods of dune activity indicate drier-than-present climate. Many factors influence the degree of dune activity, including wind strength, sediment availability, and degree of surface disturbance by animals or human activity. Nevertheless, several studies show that in many regions, including southern Africa (Lancaster, 1988; Thomas et al., 2005), China (Wang et al., 2005), and North America (Muhs and Holliday, 1995; Wolfe, 1997), the degree of dune activity is influenced to a great extent by the amount of plant cover, which in turn is influenced significantly by moisture balance. Moisture balance can conveniently be

calculated, at least to a first approximation, by the ratio of precipitation (P) to potential evapotranspiration (PE). Values less than 1.0 indicate a net moisture deficit and are typical of regions with semiarid, arid, or hyperarid climates. In areas where P/PE is greater than about 0.35–0.40, dunes are stabilized by vegetation. When P/PE is less than about 0.35–0.40, dunes are at least partially active, and as this ratio diminishes, dune activity increases (Figure 17). In desert regions of the southwestern United States and northwestern Mexico, P/PE is less than about 0.25. Thus, dunes in the Chihuahuan Desert, Algodones, and Gran Desierto dune fields (Figure 15) are quite active (Figure 18). In the Taklimakan Desert of China, eolian sand is active over most of the dune field (Figure 3) and P/PE

Dunes on the Moenkopi plateau, northeastern Arizona, USA (~35°42' N; 110°55' W)



Near Tolani Lakes, Arizona, USA,
~35°30' N; 110°55' W:

Moenkopi Plateau, ~35°47' N; 111°04' W:

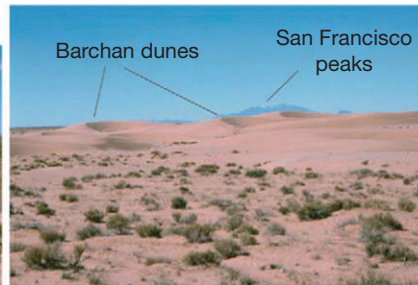
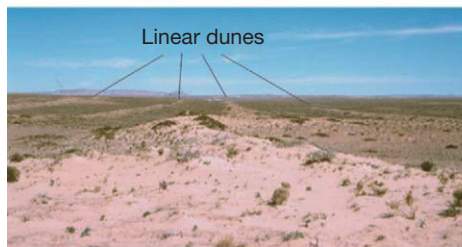


Figure 19 Dune forms on the Colorado Plateau, United States. Upper: aerial photograph showing dune types (USGS NAPP 5240-178, 22 September 1992). Lower: active linear dunes separated by stable eolian sand sheets (left) and active barchan dunes (right). Lower photographs by D.R. Muhs.

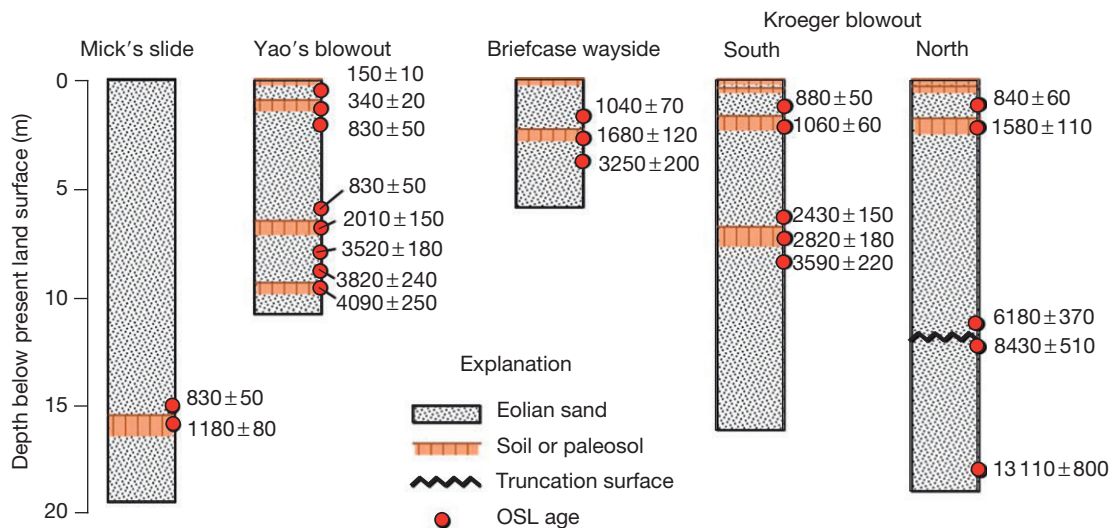


Figure 20 Stratigraphic sections showing optically stimulated luminescence (OSL) ages, in calendar years, of eolian sand in the Nebraska Sand Hills. Redrawn from Goble RJ, Mason JA, Loope DB, and Swinehart JB (2004) Optical and radiocarbon ages of stacked paleosols and dune sands in the Nebraska Sand Hills, USA. *Quaternary Science Reviews* 23:1173–1182.

values for stations in and around this sand sea are less than 0.2 (Wang et al., 2005). With these observations in mind, it can be inferred that in areas now characterized by stable dunes, past periods, represented by eolian sand, were more arid than present. In contrast, periods represented by paleosols were times that may have had a moisture balance similar to the present.

One of the remarkable observations of many North American Great Plains dune field records is the presence of multiple paleosols, all dating to the Holocene, intercalated within eolian sand (Figure 20). The evidence for multiple periods of dune activity and stability in the past few thousand years indicates that Great Plains eolian sand is quite sensitive to small changes in the overall moisture balance and degree of vegetation cover. As in the Mu Us Desert in China, the Great Plains region of North America is highly sensitive to fluctuations in the strength of a monsoonal moisture source. In central North America, the main monsoonal flow of moist air is from the Gulf of Mexico in summer. When this flow is diminished,

summer precipitation is lower and vegetation cover is reduced. Dry periods in the Great Plains are characterized not only by decreased summer monsoonal flow from the Gulf of Mexico, but also by an increase in zonal, or westerly flow of Pacific air (typically a winter pattern). Pacific air is usually dry by the time it reaches the Great Plains. Moisture in Pacific air masses is lost on north–south trending mountain ranges between the Pacific Ocean and the Great Plains (Figure 15). Reactivation of stabilized dune sand or mobilization of new sand by wind is enhanced under conditions of reduced vegetation cover during periods of decreased monsoonal flow and increased zonal flow.

Another important application of dunes to paleoclimate studies in mid-latitude North America is wind direction inferred from dune orientation. Most of the late Holocene dunes in the midcontinent of North America are parabolic forms, which are excellent paleowind indicators. Parabolic dunes have arms that point upwind, with noses that face downwind, as seen when they are active in areas such as

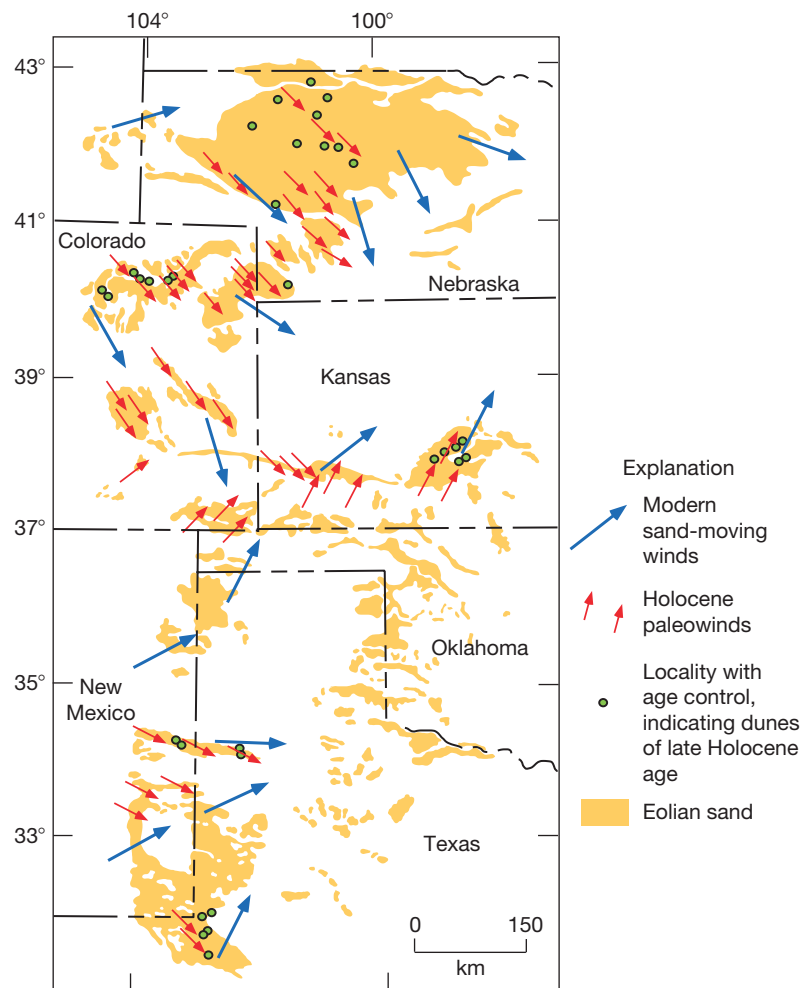


Figure 21 Map showing the distribution of eolian sand in the Great Plains, United States, direction of modern, sand-moving winds, and late Holocene paleowinds based on dune orientations. Modified from Muhs DR and Zárate M (2001) Late Quaternary eolian records of the Americas and their paleoclimatic significance. In: Markgraf V (ed.) *Interhemispheric Climate Linkages*, pp. 183–216. San Diego: Academic Press. Also shown (green circles) are localities where there are late Holocene radiocarbon or OSL ages for eolian sand. From Swinehart and Diffendal (1990), Arbogast (1996), Muhs et al. (1997), Holliday (2001), Goble et al. (2004), Forman et al. (2005), Madole et al. (2005). Modern sand-moving winds constructed using methods of Fryberger and Dean (1979).

White Sands, New Mexico (Figure 18). On the Great Plains of North America, orientations of many late Holocene dunes indicate northwesterly winds in Nebraska, and northern and central Colorado (Figure 18), similar to modern, sand-moving wind regimes of fall, winter, and spring (Figure 21). Such orientations are consistent with the scenario of increased zonal or westerly flow discussed above. Late Holocene dunes in southeastern Colorado and Kansas have orientations that indicate southwesterly paleowinds, again similar to present. However, in eastern New Mexico and western Texas, dunes have orientations that indicate northwesterly winds, whereas modern dune-forming winds are dominantly from the southwest. One explanation is that during the periods of late Holocene dune formation, there was less of an influence from southwestern (Gulf of California), monsoonal airflow, and a greater influence from zonal (westerly) circulation.

Conclusion

Large dune fields, or sand seas, are widespread in the interiors of continental mid-latitudes. Most of them are found in China, Central Asia, Argentina, and the United States. Mid-latitude dune fields in the United States are generally smaller than those of low latitudes, but those in China, Central Asia, and Argentina are quite large. Where climates are arid, such as in western China and Central Asia, active dunes dominate these sand seas. Other dune fields, in the semiarid climates of eastern China, Argentina, and the United States, have mostly stabilized dunes. Mid-latitude dune fields are often near mountains, which supply much of the sediment to the sand seas on a regular basis, particularly during glacial periods. Many different kinds of eolian landforms are found in mid-latitude sand seas, including transverse dunes, longitudinal (linear) dunes, star dunes, parabolic dunes, and eolian sand sheets. Where dune fields of mid-latitudes occur in semiarid climates, they are particularly sensitive indicators of shifts in climate. In China, records of dune field expansion and contraction extend back hundreds of thousands of years.

See also: Eolian Records, Deep-Sea Sediments; Loess Deposits: Origins and Properties. **Dune Fields:** High Latitudes; Low Latitudes. **Loess Records:** Central Asia; China; North America; South America. **Luminescence Dating:** Optical Dating. **Paleosols and Wind-Blown Sediments:** Nature of Paleosols.

References

- Aleinikoff JN, Muhs DR, and Walter M (1994) U-Pb evidence for provenances of Holocene sand dunes, northeastern Colorado and the Nebraska Sand Hills. *US Geological Survey Circular* 1107: 2.
- Arbogast AF (1996) Stratigraphic evidence for Late Holocene eolian sand mobilization and soil formation in south-central Kansas, USA. *Journal of Arid Environments* 34: 403–414.
- Bettis EA III, Muhs DR, Roberts HM, and Wintle AG (2003) Last glacial loess in the conterminous USA. *Quaternary Science Reviews* 22: 1907–1946.
- Bowler JM, Wyrwoll K-H, and Lu Y (2001) Variations of the northwest Australian summer monsoon over the last 300,000 years: The paleohydrological record of the Gregory (Mulan) Lakes System. *Quaternary International* 83–85: 63–80.
- Chen JS, Li L, Wang JY, et al. (2004) Water resources: Groundwater maintains dune landscape. *Nature* 432: 459.
- Chepalysga AL (1984) Inland sea basins. In: Velichko AA, Wright HE Jr., and Barnosky CW (eds.) *Late Quaternary Environments of the Soviet Union*, pp. 229–247. Minneapolis: University of Minnesota Press.
- Ding ZL, Yu ZW, Rutter NW, and Liu TS (1994) Towards an orbital time scale for Chinese loess deposits. *Quaternary Science Reviews* 13: 39–70.
- Dodonov AE (1991) Loess of Central Asia. *GeoJournal* 24(2): 185–194.
- Duce RA, Liss PS, Merrill JT, et al. (1991) The atmospheric input of trace species to the world ocean. *Global Biogeochemical Cycles* 5: 193–259.
- Dutta PK, Zhou Z, and dos Santos PR (1993) A theoretical study of mineralogical maturation of eolian sand. *Geological Society of America, Special Paper* 284: 203–209.
- Forman SL, Marín L, Pierson J, Gómez J, Miller GH, and Webb RS (2005) Aeolian sand depositional records from western Nebraska: Landscape response to droughts in the past 1500 years. *The Holocene* 15: 973–981.
- Frechen M and Dodonov AE (1998) Loess chronology of the Middle and Upper Pleistocene in Tadjikistan. *Geologische Rundschau* 87: 2–20.
- Fryberger SG and Dean G (1979) Dune forms and wind regime. In: McKee ED (ed.) *A Study of Global Sand Seas*, vol. 1052, pp. 137–169. Washington, DC: US Government Printing Office. US Geological Survey Professional Paper 1052.
- Gansser A (1964) *Geology of the Himalayas*. London: Wiley.
- Goble RJ, Mason JA, Loope DB, and Swinehart JB (2004) Optical and radiocarbon ages of stacked paleosols and dune sands in the Nebraska Sand Hills, USA. *Quaternary Science Reviews* 23: 1173–1182.
- Goudie AS (2002) *Great Warm Deserts of the World*. Oxford: Oxford University Press.
- Harrison TM, Copeland P, Kidd WSF, and Yin A (1992) Raising Tibet. *Science* 255: 1663–1670.
- Holliday VT (2001) Stratigraphy and geochronology of upper Quaternary eolian sand on the Southern High Plains of Texas and New Mexico, United States. *Geological Society of America Bulletin* 113: 88–108.
- Iriondo M (1999) Climatic changes in the South American plains: Records of a continental-scale oscillation. *Quaternary International* 57/58: 93–112.
- Kröhling DM (1999) Upper Quaternary geology of the lower Carcarañá Basin, North Pampa, Argentina. *Quaternary International* 57/58: 135–148.
- Lancaster N (1988) Development of linear dunes in the southwestern Kalahari, southern Africa. *Journal of Arid Environments* 14: 233–244.
- Letollé R and Mainguet M (1993) *Aral*. Paris: Springer.
- Li SH, Sun JM, and Zhao H (2002) Optical dating of dune sands in the northeastern deserts of China. *Palaeogeography, Palaeoclimatology, Palaeoecology* 181: 419–429.
- Lioubimtseva E (2002) Arid environments. In: Shahgedanova M (ed.) *The Physical Geography of Northern Eurasia*, pp. 267–283. Oxford: Oxford University Press.
- Liu TS (1985) *Loess and the Environment*. Beijing: China Ocean Press.
- Liu TS and Ding ZL (1993) Stepwise coupling of monsoon circulation to global ice volume variations during the late Cenozoic. *Global and Planetary Change* 7: 119–130.
- Loope DB, Swinehart JB, and Mason JP (1995) Dune-dammed paleovalleys of the Nebraska Sand Hills: Intrinsic versus climatic controls on the accumulation of lake and marsh sediments. *Geological Society of America Bulletin* 107: 396–406.
- Madole RF, VanSistine DP, and Michael JA (2005) Distribution of late Quaternary wind-deposited sand in eastern Colorado. US Geological Survey Scientific Investigations Map: 2875.
- Martínez GA, Martínez-Arca J, Gwyn QHJ, and Bernasconi MV (2001) Combined use of RADARSAT-1 and Landsat TM data for geomorphological applications in lowlands of Buenos Aires Province, Argentina. *Canadian Journal of Remote Sensing* 27: 638–642.
- Martinson DG, Pisias NG, Hays JD, Imbrie J, Moore TC Jr., and Shackleton NJ (1987) Age dating and the orbital theory of the ice ages: Development of a high-resolution 0 to 300,000-year chronostratigraphy. *Quaternary Research* 27: 1–29.
- Merrill JT, Bleck R, and Uematsu M (1989) Meteorological analysis of long-range transport of mineral aerosols over the North Pacific. *Journal of Geophysical Research* 94: 8584–8598.
- Molnar P and Tapponnier P (1975) Cenozoic tectonics of Asia: Effects of a continental collision. *Science* 189: 419–426.
- Muhs DR (2004) Mineralogical maturity in dune fields of North America, Africa and Australia. *Geomorphology* 59: 247–269.
- Muhs DR and Holliday VT (1995) Evidence for active dune sand on the Great Plains in the 19th century from accounts of early explorers. *Quaternary Research* 43: 198–208.
- Muhs DR and Holliday VT (2001) Origin of late Quaternary dune fields on the Southern High Plains of Texas and New Mexico. *Geological Society of America Bulletin* 113: 75–87.

- Muhs DR, Reynolds R, Been J, and Skipp G (2003) Eolian sand transport pathways in the southwestern United States: Importance of the Colorado River and local sources. *Quaternary International* 104: 3–18.
- Muhs DR, Stafford TW, Cowherd SD, et al. (1996) Origin of the late Quaternary dune fields of northeastern Colorado. *Geomorphology* 17: 129–149.
- Muhs DR, Stafford TW Jr., Swinehart JB, et al. (1997) Late Holocene eolian activity in the mineralogically mature Nebraska Sand Hills. *Quaternary Research* 48: 162–176.
- Muhs DR and Zárte M (2001) Late Quaternary eolian records of the Americas and their paleoclimatic significance. In: Markgraf V (ed.) *Interhemispheric Climate Linkages*, pp. 183–216. San Diego: Academic Press.
- Rea DK (1994) The paleoclimatic record provided by eolian deposition in the deep sea: The geological history of wind. *Reviews of Geophysics* 32: 159–195.
- Redsteer M and Block D (2004) Drought conditions accelerate destabilization of sand dunes on the Navajo Nation, southern Colorado Plateau. *Geological Society of America Abstracts with Programs* 36(5): 171.
- Shackleton NJ, Berger A, and Peltier WR (1990) An alternative astronomical calibration of the Lower Pleistocene timescale based on ODP Site 677. *Transactions of the Royal Society of Edinburgh: Earth Science* 81: 251–261.
- Shackleton NJ and Pisias NG (1985) Atmospheric carbon dioxide, orbital forcing, and climate. *Geophysical Monograph Series* 32: 412–417.
- Sun JM (2000) Origin of eolian sand mobilization during the past 2300 years in the Mu Us Desert, China. *Quaternary Research* 53: 73–88.
- Sun JM (2002a) Provenance of loess material and formation of loess deposits on the Chinese Loess Plateau. *Earth and Planetary Science Letters* 203: 845–859.
- Sun JM (2002b) Source regions and formation of the loess sediments on the high mountain regions of northwestern China. *Quaternary Research* 58: 341–351.
- Sun JM, Ding ZL, and Liu TS (1998) Desert distribution during the glacial maximum and climatic optimum: Example of China. *Episodes* 21: 28–31.
- Sun JM, Ding ZL, Rokosh D, and Rutter N (1999) 580,000 year environmental reconstruction from eolian deposits at the Mu Us Desert margin, China. *Quaternary Science Reviews* 18: 1351–1364.
- Swinehart JB and Diffendal RFV Jr. (1990) Geology of the pre-dune strata. In: Bleed A and Flowerday C (eds.) *Resource Atlas No. 5a: An Atlas of the Sand Hills*, pp. 29–42. Lincoln: University of Nebraska.
- Thomas DSG, Knight M, and Wiggs GFS (2005) Remobilization of southern African desert dune systems by twenty-first century global warming. *Nature* 435: 1218–1221.
- Thomas DSG and Shaw PA (1991) *The Kalahari Environment*. Cambridge: Cambridge University Press.
- Thorp J and Smith HTU (1952) *Pleistocene eolian deposits of the United States, Alaska, and parts of Canada*. National Research Council Committee for the Study of Eolian Deposits, Geological Society of America. scale 1: 2,500,000.
- Wang X, Dong Z, Yan P, Zhang J, and Qian G (2005) Wind energy environments and dunefield activity in the Chinese deserts. *Geomorphology* 65: 33–48.
- Wilson IG (1973) Ergs. *Sedimentary Geology* 10: 77–106.
- Wolfe SA (1997) Impact of increased aridity on sand dune activity in the Canadian Prairies. *Journal of Arid Environments* 36: 421–432.
- Zárte MA (2003) Loess of southern South America. *Quaternary Science Reviews* 22: 1987–2006.
- Zárte MA and Blasi A (1993) Late Pleistocene-Holocene eolian deposits of the southern Buenos Aires Province, Argentina: A preliminary model. *Quaternary International* 17: 15–20.
- Zhou WJ, Head MJ, and Deng L (2001) Climatic changes in northern China since the late Pleistocene and its response to global change. *Quaternary International* 83–85: 285–292.
- Zhu ZD, Wu Z, Liu S, and Di XM (1980) *An Outline on Chinese Deserts*. Beijing: Science Press.

Low Latitudes

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Low-latitude sand seas (also called ergs) are extensive areas of sand dunes located in the tropical and subtropical deserts of the world ([Figure 1](#)), where they occupy as much as one-third of the area classified as arid. Major low-latitude sand seas occur in the Sahara and Arabian Deserts, the Thar Desert of India, the interior of Australia, and in southern Africa (Namib and Kalahari Deserts). Large sand seas are absent in the Americas, where areas of dunes cover less than 1% of the area classified as arid. Important studies and surveys of low-latitude sand seas include those by [Wilson \(1973\)](#) and [McKee \(1979\)](#). [Lancaster \(1999\)](#) provided a review of modern paradigms of sand sea geomorphology. Important summaries of chronologic data include [Munyikwa \(2005\)](#) for Southern Hemisphere sand seas, [Glennie and Singhvi \(2002\)](#) for Arabia, and [Swezey \(2001\)](#) for the Sahara.

Sand seas contain large volumes of sand (tens to hundreds of cubic kilometers) and have accumulated episodically during the Quaternary. Their construction has therefore been determined by climatic, tectonic, and sea-level changes that have affected sand supply, availability, and mobility, as well as the preservation of deposits and landforms from prior episodes of eolian construction. As significant sedimentary deposits in low-latitude desert regions, sand seas also provide an archive of the effects of climate change on these areas.

Sand seas and dune fields typically form part of well-defined regional- and local-scale sediment transport systems in which sand is moved by wind from source areas (e.g., distal fluvial deposits and sandy beaches) via transport pathways to depositional sinks. Sand seas accumulate downwind of source zones wherever sand transport rates decrease as a result of regional and/or local changes in wind speed and directional variability so that the influx of sand exceeds outflux. A conceptual model for the response of low-latitude (and other) sand seas to changes in external forcing factors was provided by [Kocurek and Lancaster \(1999\)](#). In this model, episodes of sand sea construction are viewed as the product of changes in the supply, availability, and mobility of sand-sized sediment. Sediment supply is the deposition of sediment that is a source for the eolian system either contemporaneously or at some later time. Availability of sediment for transport by the wind (the probability of entrainment of sand for transport) is strongly influenced by variations in vegetation cover and soil moisture. In addition, changes in sea level and regional hydrology may make the sediment stored in shallow marine or lacustrine environments available for transport. The mobility of sediment is determined by the transport capacity of the wind (the potential rate of sediment transport).

Evidence for Paleoenvironments in Low-Latitude Sand Seas

Sand seas provide paleoenvironmental information from three main sources: (a) dune systems, especially those stabilized by

vegetation; (b) interdune deposits; and (c) penetration of extra-dune fluvial deposits into the margins of the sand sea. The quality of paleoclimatic data provided by sand seas is frequently much better than that obtained from adjacent rocky desert regions, because sand dune areas tend to accentuate the effects of both dry and wet phases ([Rognon, 1982](#)). In periods of increased rainfall, the high infiltration capacity and porosity of dune sands favor the growth and persistence of vegetation, which may lead to the partial or complete stabilization of the dunes and formation of soils on dune sands ([Figure 2](#)). If the periods of increased rainfall are of sufficient magnitude and duration, then water tables may rise, leading to the formation of pond and marsh deposits in interdune areas. In contrast, periods of aridity will give rise to a very unfavorable biotic environment in which vegetation cover is reduced as a result of low precipitation and active sand movement. In addition, dune systems provide some of the best indicators of past wind regimes and atmospheric circulation patterns (e.g., [Lancaster et al., 2002](#)).

Many sand seas contain a variety of dune types or dunes on different trends, often forming complex patterns. Satellite image analysis, combined with studies of dune geomorphology and sedimentology, can be used to analyze the dune patterns, identify genetically distinct generations of dunes, and interpret the conditions in which they formed. The understanding of the history of sand seas and dunes has been revolutionized in recent years by the application of luminescence techniques to directly date periods of dune formation and/or reactivation ([Stokes, 1997](#)). In addition, large dunes, especially those of linear form, appear to be composite features, reflecting many generations of dune construction and stability. Recent application of ground-penetrating radar techniques has enabled identification of major sedimentary units and bounding surfaces prior to luminescence dating of these units ([Bristow et al., 2005](#)).

Sand seas may also contain evidence of periods of climates wetter than present, in which dunes were stabilized by vegetation and soil formation and marshes and shallow lakes formed in interdune areas. Many of these deposits contain fossil biota and/or archaeological materials that can be used to reconstruct past hydrology and vegetation communities and provide ages for periods of wetter climates ([Gasse, 2002](#); [Teller et al., 1990](#)). Stratigraphic relations between such deposits and those of dunes have traditionally been used as a major source of chronologic information for sand seas in the Sahara ([Swezey, 2001](#)) and elsewhere.

In many areas, the evidence for environmental change from the sand seas is corroborated by records of dust flux (indicating arid, windy conditions) from marine sediment cores in adjacent ocean and shelf areas, which frequently provide a more continuous and better dated record than that of the sand seas ([deMenocal et al., 2000](#); [Stuut and Lamy, 2004](#)).

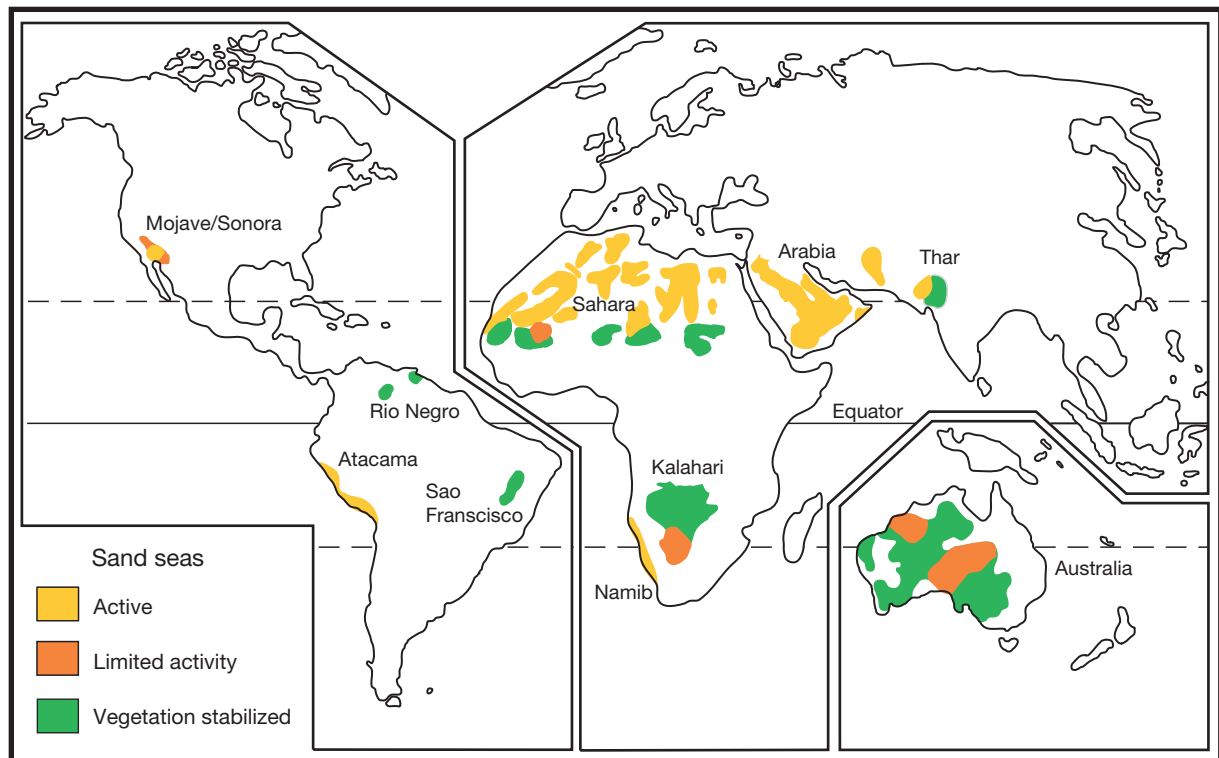


Figure 1 Distribution of major low-latitude sand seas, showing areas of active, partly active, and vegetation-stabilized dunes. For names and details of sand seas, see regional maps.

Regional Summaries

Sahara Desert

The Sahara Desert contains a significant proportion of the global inventory of wind-blown sand as well as some of the world's largest sand seas, which cover approximately 28% of this desert. There is a large amount of paleoenvironmental information on Saharan sand seas, much of it based on analyses of dated interdune lacustrine deposits and their stratigraphic relations to dune deposits (Gasse and Van Campo, 1994; Swezey, 2001). However, there are few examples of episodes of eolian deposition that have been directly dated by luminescence techniques.

The major sand accumulations are situated on each side of the central Saharan uplands, where there is very little sand (Figure 3). In the Sahara, sand is moved long distances between apparent source areas and depositional sinks as well as from one sand sea to another. In many instances, the source of sand is not clear or not known. The high degree of mineralogical maturity of sand from Libyan sand seas (Muhs, 2004) suggests long-distance transport, repeated periods of transport and reworking over short distances, or a source in the mineralogically mature Nubian Sandstone that crops out extensively in the eastern Sahara.

The sand transport pathways, clearly visible on satellite images (Breed et al., 1979; Mainguet, 1984), reflect the dominance of the trade wind circulation that tends to move sand from the sand-poor eastern and central parts of the desert toward thick sand accumulations in the Sahel and from the piedmont of the Atlas Mountains and the Mediterranean coast

to the northern and western sand seas (Mainguet and Chemin, 1983; Wilson, 1971). The location of the major Saharan sand seas is controlled both by patterns of winds and by topographic obstacles. The Grand Erg Occidental and the Ubari and Murzuq sand seas of Libya have accumulated against the northern slopes of the central Saharan uplands (Fryberger and Ahlbrandt, 1979). Sand seas in the Akchar of Mauritania and the Fachi-Bilma region of Tchad lie where winds converge in the lee of escarpments or mountain masses. The thick sand accumulations of the southern Sahara and Sahel have developed because wind energy is reduced on the margins of the trade wind circulation. Many northern Saharan sand seas (e.g., the Grand Erg Occidental and Oriental) occur in areas of complex or multidirectional wind regimes influenced by trade wind and mid-latitude cyclonic circulations as well as local topographically induced winds (Wilson, 1971).

Dune patterns in Saharan sand seas are summarized in Table 1. The northern Saharan sand seas are characterized by complex crescentic and star dunes developed in a multidirectional wind regime that results from the effects of both trade wind and mid-latitude circulations. By contrast, sand seas in the western and southern Sahara are mainly composed of linear dunes, with small areas of crescentic dunes. The dominance of linear dunes in these areas reflects the persistence of trade wind circulations. Eolian accumulations in the eastern Sahara are dominated by extensive sand sheets and areas of gently undulating dunes with a chevron pattern. The formation of extensive sand sheets is attributed to the abundance of coarse sand derived from the Mesozoic Nubian Sandstone that underlies much of the area (Stokes et al., 1998).

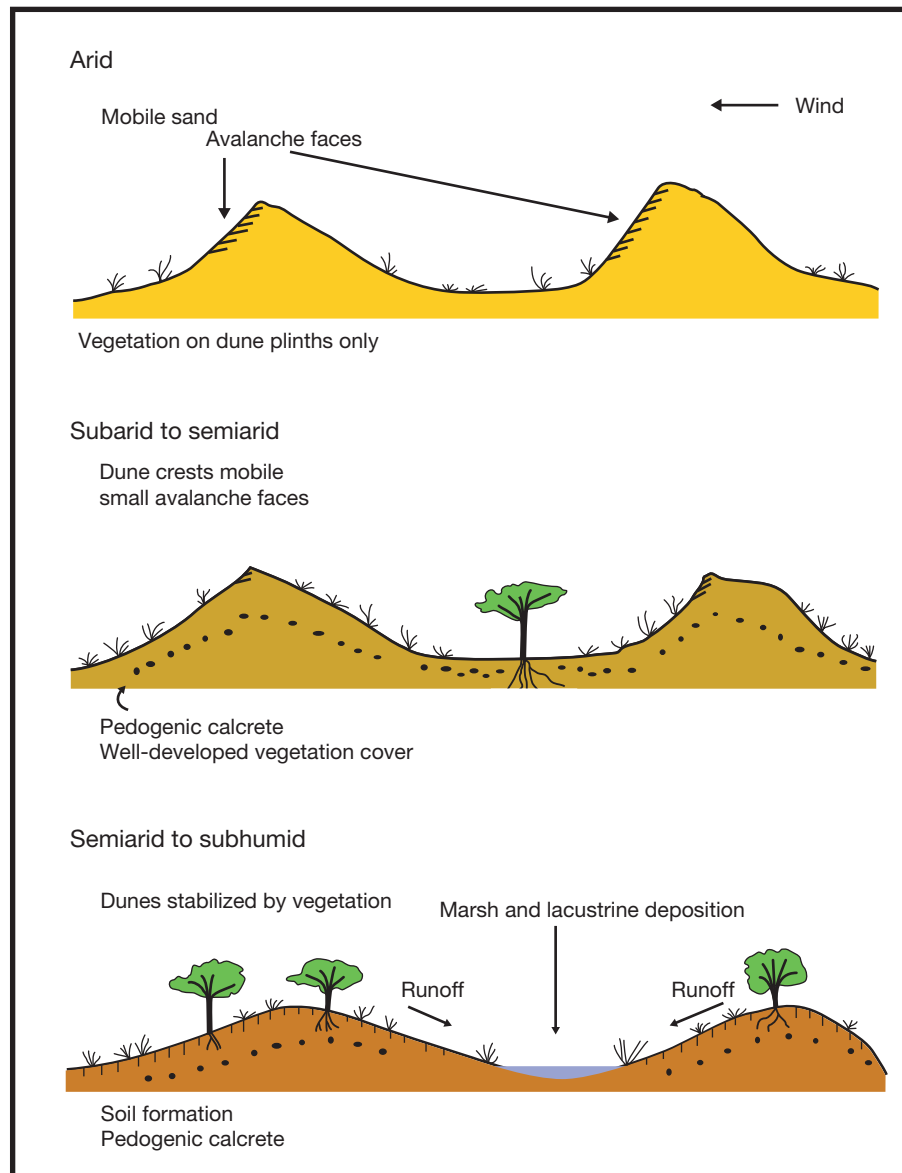


Figure 2 A conceptual model for the response of dune areas to climate change. Note the contrasts between arid and more humid conditions. Reproduced from Rognon P (1982) Pluvial and arid phases in the Sahara: The role of non climatic factors. *Palaeoecology of Africa* 12: 45–62.

In addition to the largely active sand seas of the Sahara, satellite images show the existence of large areas of vegetation-stabilized dunes on the southern margins of the desert, extending south into Sahel and beyond (Figure 4). Extensive systems of mainly linear dunes occur in areas that today receive as much as 1000 mm of rainfall per year. These dunes, which are today extensively degraded and covered by savanna vegetation, were first recognized on aerial photographs of northern Nigeria and southern Mauritania and subsequently mapped throughout the Sahel (Grove and Warren, 1968). A regionally extensive paleosol is present within these dunes (Mauz and Felix-Henningsen, 2005). This paleosol indicates a past period of stability, similar to the present one. In areas to the north, this paleosol underlies late Holocene to modern dunes, many of them of crescentic form.

The age of the stabilized dunes in the Sahel has been hypothesized to be last glacial, spanning the period 20–13 ka (Ogolian period in the western Sahara), mostly by stratigraphic relations to overlying lacustrine deposits dated by radiocarbon methods (Swezey, 2001). Direct age estimates for these dunes using optically stimulated luminescence (OSL) are limited to those in northeastern Nigeria (Stokes and Horrocks, 1998), where maximum rates of construction occurred 18–13 ka, and in the Sahel of Mali, with an age of 12.7 ka.

In western Mauritania, where dunes are today largely active, combined geomorphic and OSL dating studies (Lancaster et al., 2002) show that three main generations of linear dunes were formed during the periods 25–15 ka (centered around the Last Glacial Maximum – LGM), 13–10 ka (spanning the Younger Dryas event) and after 5 ka. The wind

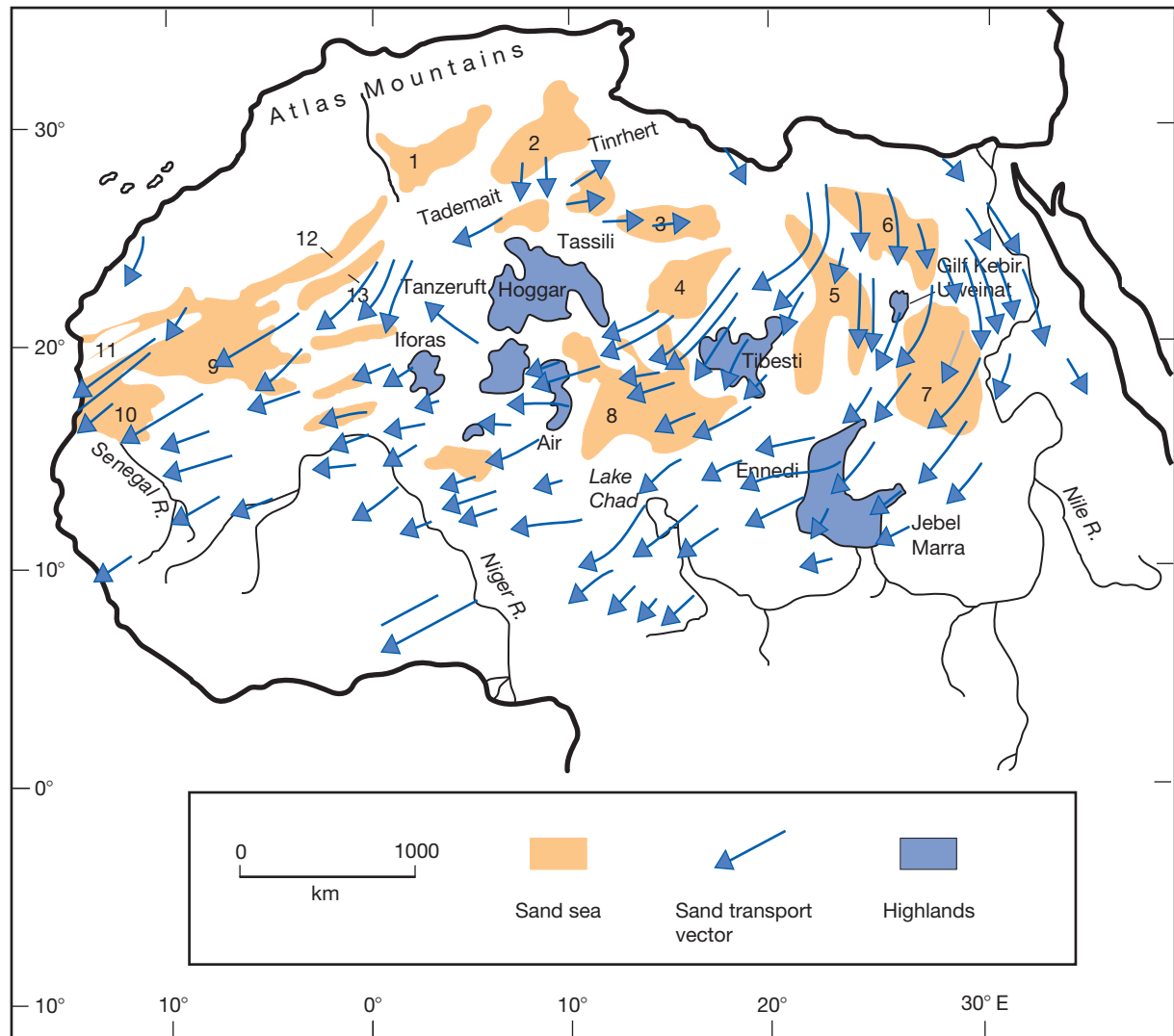


Figure 3 Location of Saharan sand seas and vectors of sand transport. Sand seas are numbered as follows: 1. Grand Erg Occidental, 2. Grand Erg Oriental, 3. Ubari, 4. Murzuq, 5. Calanscio, 6. Great Sand Sea, 7. Selima, 8. Fachi-Bilma and Tenéré, 9. Majabat al Koubra, 10. Aouker, 11. Akchar, 12. Iguidi, and 13. Chech. Reproduced from Mainguet M and Chemin MC (1983) Sand seas of the Sahara and Sahel: An explanation of their thickness and sand dune type by the sand budget principle. In: Brookfield ME and Ahlbrandt TS (eds.) *Eolian Sediments and Processes. Developments in Sedimentology*, pp. 353–364. Amsterdam: Elsevier; Mainguet M (1984) Space observations of Saharan aeolian dynamics. In: El Baz F (ed.) *Deserts and Arid Lands*, pp. 59–77. The Hague: Nyhoff.

regimes that occurred during each of these periods were significantly different, leading to the formation of dunes on three distinct superimposed trends – northeast, north-northeast, and north (Figure 5). The period from approximately 9 to 5 ka was characterized by formation of shallow lakes in interdune areas and formation of soils on the dunes (Kocurek et al., 1991).

Sedimentary and OSL dating studies of the Selima sand sheet in the eastern Sahara (Stokes et al., 1998) indicate two major episodes of accumulation at 20–15 and 4–3 ka, separated by an early to mid-Holocene pluvial period during which a regionally extensive paleosol formed and local wadi and lacustrine deposits were laid down.

Directly dated information on the response of sand seas on the northern margins of the Sahara to Quaternary climate change is sparse, but a detailed luminescence-dated record

from the Chott Rharsa area of southern Tunisia suggests formation of gypsum-rich dunes and low lake levels 12–10, approximately 7.5, and 6.2–5.6 ka (Swezey et al., 1999). In summary, OSL ages, together with stratigraphic relations between eolian sands and ^{14}C -dated lacustrine deposits, indicate widespread eolian activity and dune formation in the Sahara from 25 to 11 ka, with extensive stabilization of dunes in the southern and western Sahara in the period 11–5 ka (the African Humid Period, deMenocal et al., 2000). The African Humid Period was apparently not uniformly humid throughout this region. Gasse and Van Campo (1994) summarized evidence for a widespread period of aridity 8.8–7.8 ka, with some areas experiencing dry conditions at 6.8 ka. The period after 5 ka appears to have been arid, with renewed formation of dunes and/or reworking of existing deposits.

Table 1 Distribution of dunes in selected sand seas

	Percentage of dune area covered								
	Sahara				Kalahari	Namib	Arabia	Thar	Australia
	South	North	Northeast	West					
Linear dunes	24.08	22.84	17.00	35.50	85.55	32.55	49.81	13.96	95.00
Crescentic dunes	28.37	33.34	14.50	19.20		11.80	14.91	25.61	
Star dunes		7.92	23.90			9.92	5.34		
Sand sheets and streaks	47.54	35.92	39.20	45.30	13.56	45.44	23.24	31.75	5.00
Parabolic dunes					0.60			28.68	
Undifferentiated			4.50				6.71		

Data from Fryberger and Goudie (1981).

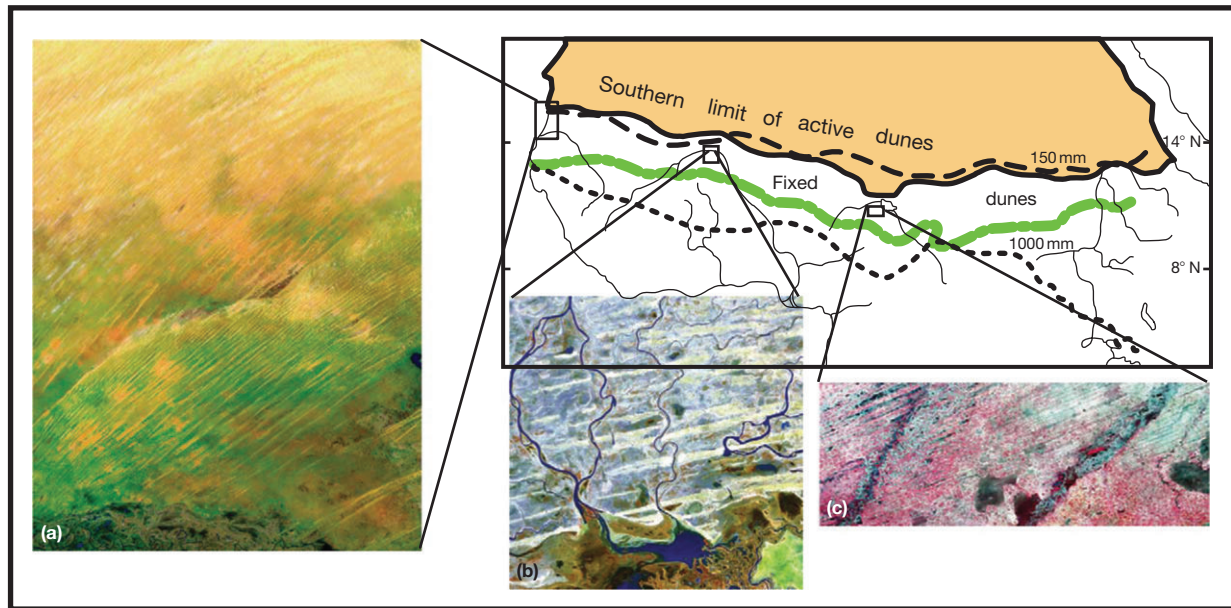


Figure 4 The Sahel and southern Sahara – limits of active and relict vegetation-stabilized dunes. (a) Landsat image of Sahara–Sahel transition in southern Mauritania. Note the change in dune trend and increasing vegetation cover to the south. (b) Aster image of relict dunes and river systems in Mali. (c) Vegetation-stabilized linear dunes west of Lake Chad.

Kalahari Desert

Extensive systems of dunes, mostly of linear form and stabilized by vegetation, occur in the northern and southwestern Kalahari (Figure 6). The dunes can be divided into three groups (Thomas and Shaw, 1991). The eastern and northern groups of dunes are variably degraded and covered by savanna vegetation. They are also truncated by some of the major river systems in the area as well as by the Okavango Delta. The southern group of dunes occurs in the southwestern Kalahari in a 100–200-km-wide belt that extends from the highlands of Namibia to the Orange River. Dunes consist of 2–15-m-high simple and compound linear ridges. Vegetation cover on the dunes consists of grasses with scattered trees and shrubs, but the dune crests are commonly unvegetated and active sand movement takes place seasonally.

Until the recent application of luminescence dating techniques, the age(s) of the Kalahari dune systems was unknown or inferred only by relations to dated lacustrine and fluvial

deposits. Luminescence dating and stratigraphic investigations show that the last major period of linear dune development in the southern dune system occurred between 16 and 10 ka and probably involved reworking of sediment deposited in a previous period of dune building from 30 to 23 ka that may represent the initial deposition of sand in this region (Stokes et al., 1997b). Holocene dune activity in the region was localized in extent and occurred at 6 and –2 to 1 ka (Thomas et al., 1997), although the final stabilization of linear dunes may have occurred as recently as 8–9 ka in drier western areas of the region.

Parallel studies of dune chronology in the northern and eastern dunes indicate that multiple periods of dune building, each spanning several thousand years, occurred during the late Pleistocene and Holocene (Stokes et al., 1997c). OSL ages for these periods are 115–95, 46–41, and 26–20 ka. Short-lived periods of aridity were separated by much longer (40–20 ka) periods of humid conditions. Late Glacial and Holocene eolian

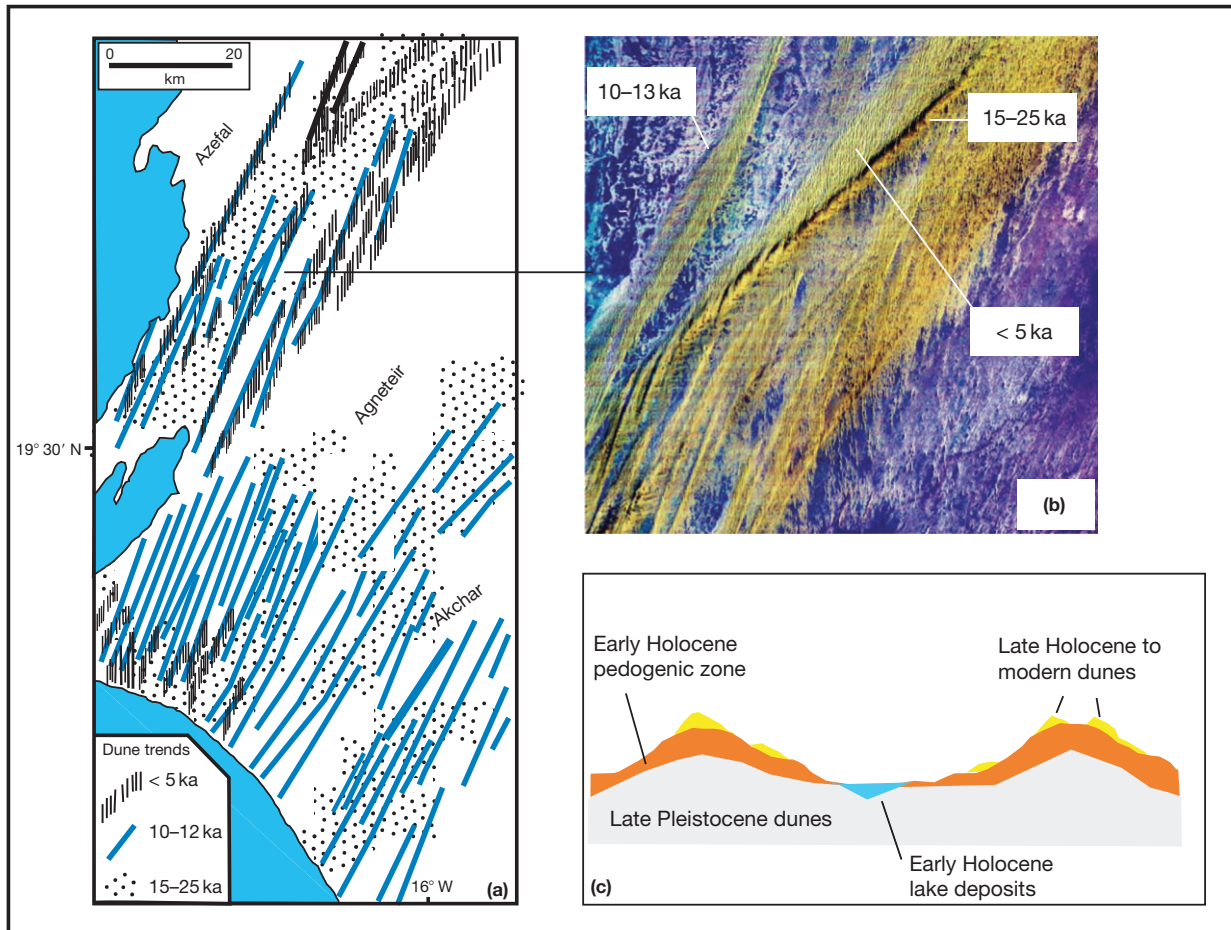


Figure 5 Effects of climate change on Mauritania dune systems. (a) Major dune trends. Reproduced from Lancaster N, et al. (2002) late Pleistocene and Holocene dune activity and wind regimes in the western Sahara of Mauritania. *Geology* 30(11): 991–994. (b) Landsat image of the Azefal, showing three dune generations. (c) Schematic cross-section of dunes in this area, showing superimposition of dune genetic units. Reproduced from Kocurek G, Havholm KG, Deynoux M, and Blakey RC (1991) Amalgamated accumulations resulting from climatic and eustatic changes, Akchar Erg, Mauritania. *Sedimentology* 38(4): 751–772.

activity was restricted to reworking of the crestal areas of the linear dunes in the area of Hwange National Park in the period –16 to 10 ka and to western Zambia, where linear dune construction from local sediment sources occurred 32–27, 16–13, and 10–8 ka (O'Connor and Thomas, 1999). Dune construction is indicated for the periods 48–41, 36–29, and 23–21 ka in the linear dunes of northern Namibia (Thomas and Shaw, 2002).

Namib Desert

The main Namib sand sea occupies an area of approximately 34 000 km² between Luderitz and the Kuiseb River and the coast and the base of the Great Escarpment of Namibia (Figure 6). The sand sea is characterized by crescentic dunes near the coast, with north–south trending linear dunes inland and clusters of star dunes and reticulate or network linear dunes inland (Lancaster, 1989). Despite the existence of multiple generations of dunes in this sand sea, there is little information on their age. OSL dating studies on the northern margin of the sand sea give ages for the base of one linear

dune of between 0.34 ± 0.02 and 1.57 ± 0.07 ka, indicating a dune migration rate of approximately 0.12 My^{-1} during the past 1600 years (Bristow et al., 2005). Studies of interdune lacustrine deposits (Teller et al., 1990) indicate widespread conditions favorable for shallow lakes and ponds in the periods 12–8, 24–20, and 32–26 ka, although the state of the adjoining dunes is not clear.

Additional small dune fields dominated by crescentic dunes occur in the northern Namib in the area known as the Skeleton Coast. A Holocene age for the Skeleton Coast dune field is suggested by the small sand thickness represented by the dunes and their simple patterns. Aerial photographs and satellite images indicate that the Engo–Kunene and Baia dos Tigres–Curoca sand seas are characterized by a similar assemblage of dunes, but increased reddening inland suggests a greater chronologic complexity for this area.

Australian Deserts

Dunes cover as much as 40% of the Australian landscape, with the majority in the Simpson–Strzelecki and Tirarai Deserts

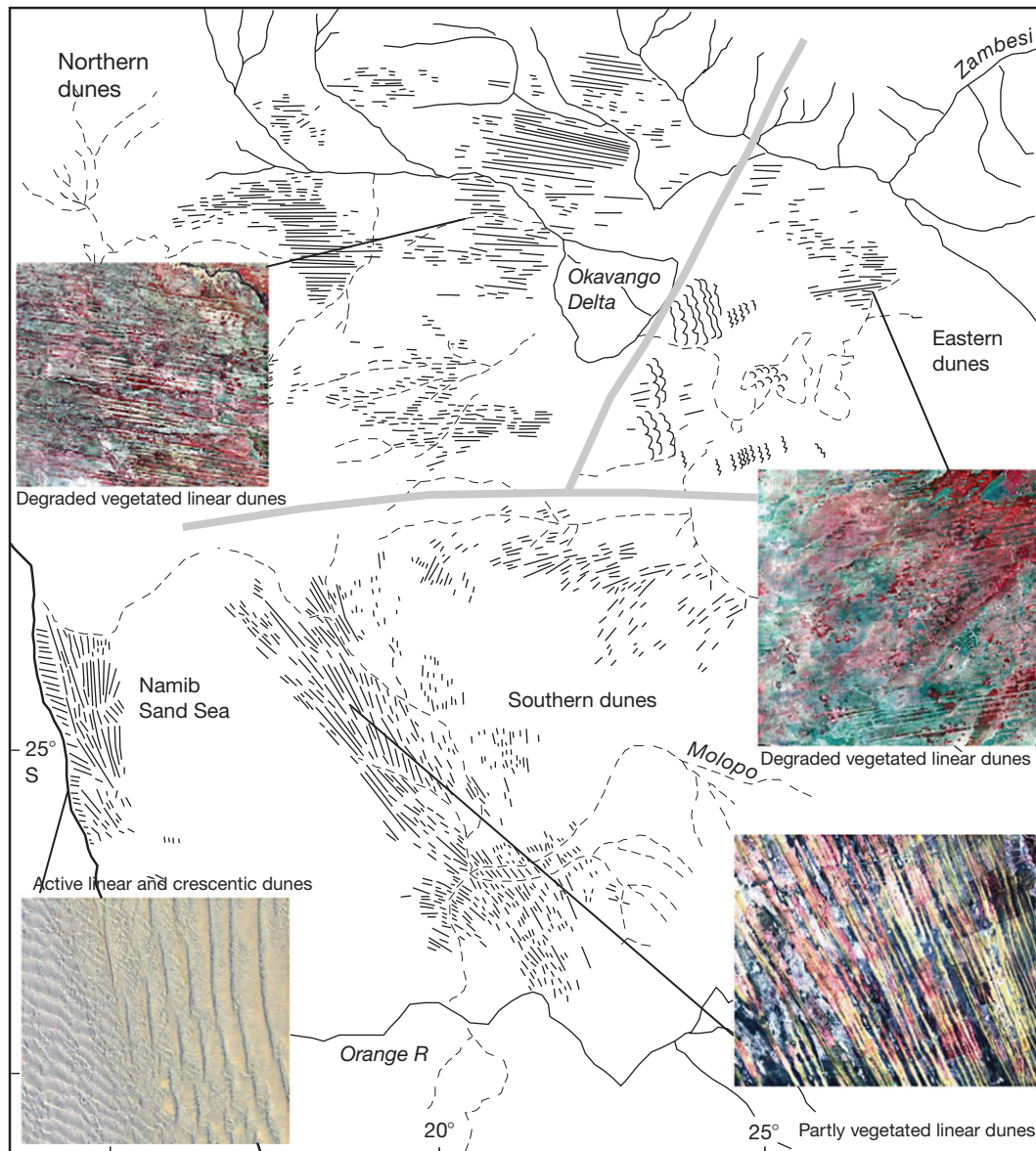


Figure 6 Dune systems in southern Africa, with Landsat images of examples of linear dunes in different areas. Dunes after Thomas DSG and Shaw PA (1991) *The Kalahari Environment*. Cambridge: Cambridge University Press.

(Figure 7). Continent-wide patterns of dunes in Australian sand seas and dune fields have been documented (Wasson et al., 1988). Although linear dunes dominate, short, narrow-crested varieties appear to occur where sand transport is limited by vegetation cover; broad-crested linear dunes are found in the northern parts of the dune system, and narrow-crested varieties are found in the south. Network dunes occur in small basins in the central parts of the continent where the directional variability of the wind regime is high. Mapping of patterns of sediment thickness, together with sedimentological and mineralogical evidence, shows that most sand in the Australian deserts has only moved short distances from its source, although the dune pattern suggests an organized continent-wide sediment transport system (Pell et al., 2000; Wasson, 1988).

A variety of luminescence dating studies indicate discontinuous dune construction in the Simpson–Strzelecki dune field

between 35 and 9 ka, with older periods of activity between 65 and 45 ka and at approximately 78 ka (Munyikwa, 2005). Localized periods of eolian activity also occurred during the Holocene. In all instances, periods of eolian construction were coeval with periods of low lake levels in the region (Wasson, 1983).

Arabian Deserts

The Arabian Peninsula is dominated by extensive dune systems, including the world's largest sand sea, the Rub al Khali, the Nafud, and Ad Dahna, all influenced by the northwesterly Shamal wind system (Figure 8). The Wahiba sand sea of Oman comprises a separate system on the coast, influenced by the southwest monsoon circulation.

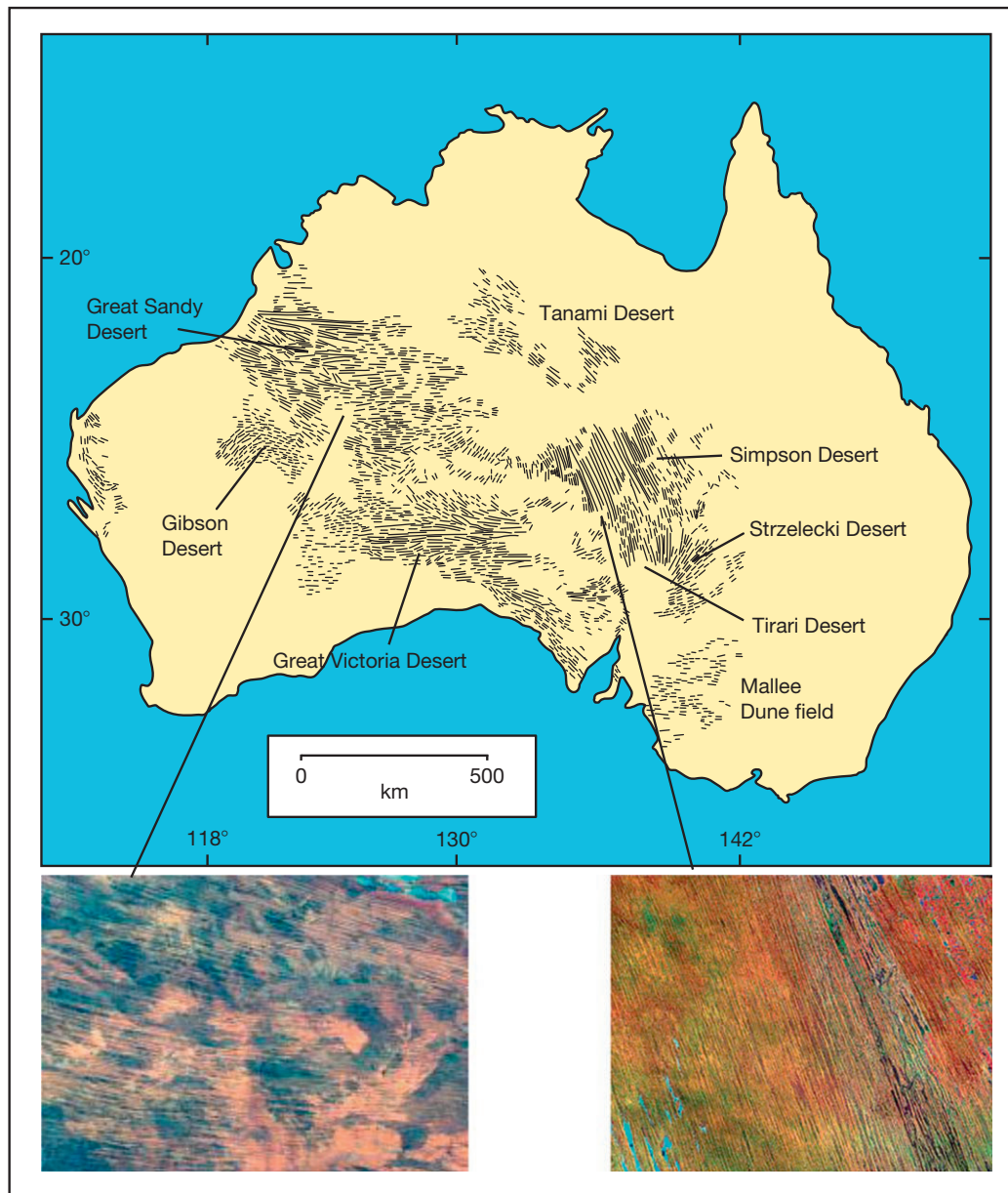


Figure 7 Dune systems in Australia. Landsat images of examples of linear dunes in the Great Sandy and Simpson–Strzelecki deserts. Dune trends and extent after Jennings (1968).

A characteristic feature of eolian sedimentation in the Wahiba sand sea and the United Arab Emirates (UAE) is the existence of thick sequences of carbonate-cemented sand (eolianite) that underlie the more quartz-rich dunes. Limited stratigraphic evidence and clustering of OSL ages on the eolianites in the UAE indicate that they formed during the periods 164–129, 113–99, 64–34, and 22–18 ka from bioclastic sediments deflated from the shelf areas of the Arabian Sea during periods of low sea levels (Glennie and Singhvi, 2002). Wind directions during these periods of dune construction were from the northwest/north-northwest. In the Wahiba, periods of accumulation of eolianites occurred 160–140, 34–24, 16–13, and approximately 9–8 ka (Preusser et al., 2005).

The southern parts of the Rub al Khali are dominated by large complex linear dunes, whereas its northern parts exhibit a complex pattern of dune construction, reactivation, and stabilization in response to changes in sea level and climate. In the UAE and adjacent areas, the largest and most widespread dunes are west-to-east-oriented complex linear dunes, with superimposed crescentic dunes in the north and east and star dune peaks in the southeast of the UAE. These dunes are composed mainly of quartz sand and are being reworked in northern areas by northwest winds into crescentic dunes and obliquely oriented small linear dunes that are superimposed on the linear dunes. The linear dunes appear to have been formed (or reformed) during the LGM, as indicated by OSL ages of 42–31, 18–15, and 12–9 ka in the UAE.

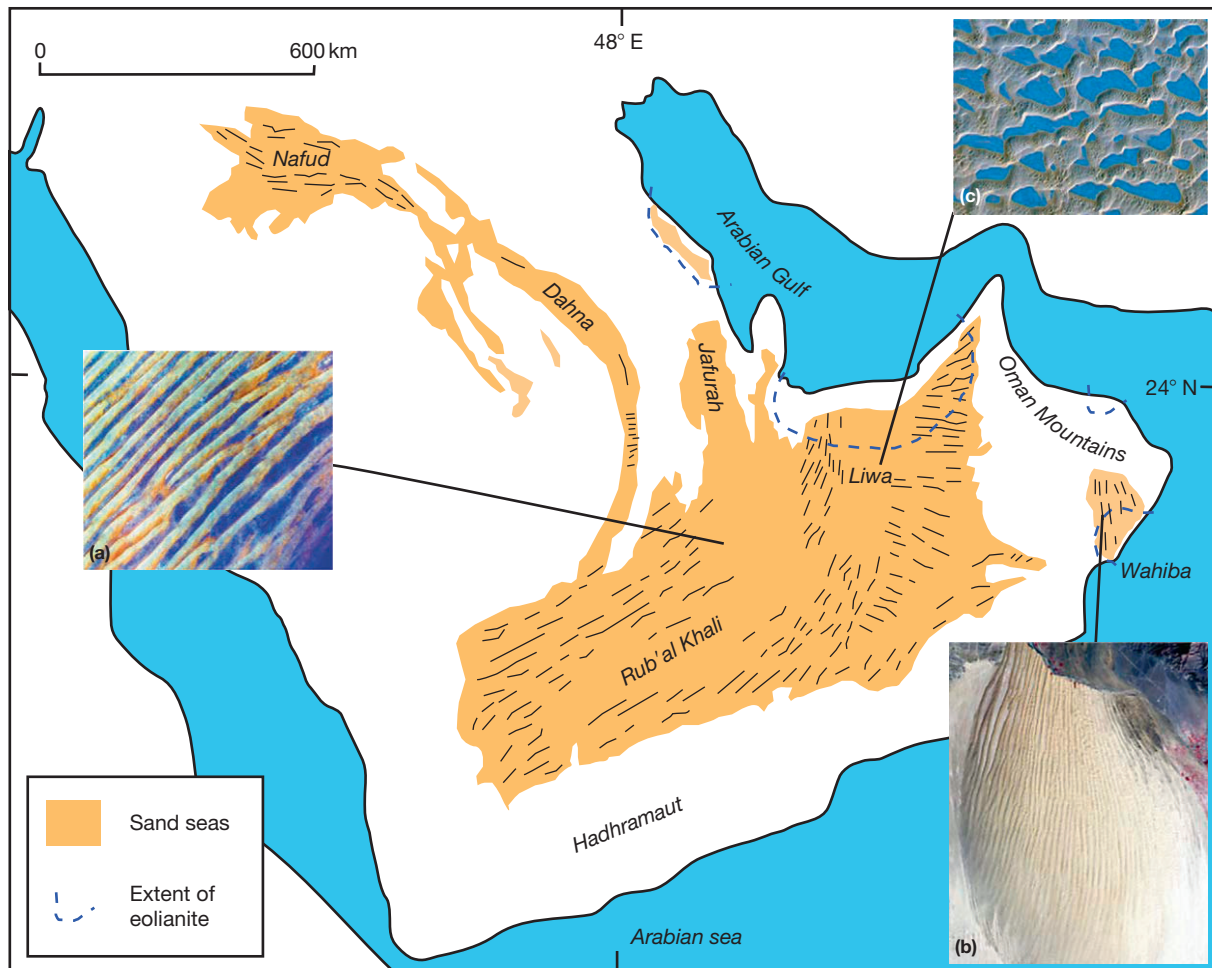


Figure 8 Sand seas and dune fields of the Arabian Peninsula. Insets: (a) aster image of linear dunes in the Rub' al Khali, (b) compound crescentic dunes in the Liwa area, and (c) linear dunes in the Wahiba Sand Sea. Dune extent and schematic dune trends after Glennie KW and Singhvi AK (2002) Event stratigraphy, paleoenvironment and chronology of SE Arabian deserts. *Quaternary Science Reviews* 21: 853–869.

In the Liwa area, large, quartz-rich compound crescentic dunes overlie inland sabkhas. Stratigraphic and geomorphic evidence, together with OSL dating of eolian sands, indicates deposition of eolian sand in the period 130–75 ka (Stokes and Bray, 2005) and after 7.8 ka, peaking at 2.8 ka, but no record from the LGM. The water table in this area was several meters higher than present during the period 41–29 ka (Wood and Imes, 1995).

In the Wahiba sand sea of Oman, a large inland part of the sand sea is characterized by 50–100-m-high south–north trending complex linear dunes. In places, the crestal areas of these dunes are overlain by crescentic and network dunes. To the south of the large linear ridges are smaller (10–30 m high) vegetated linear dunes with 300–500-m spacing. There are three groups of simple and compound crescentic and network dunes near the coast. These groups of dunes overlie cemented carbonate sands (eolianites), which crop out on the southern and western edges of the sand sea. An extensive program of drilling and luminescence dating of dune sands and eolianites in the central parts of the sand sea indicates that periods of eolian accumulation occurred 24–11, 78–65, 120–100, and 160–140 ka (Preusser et al., 2002; Radies et al., 2004).

The main south–north-oriented linear dunes were formed at approximately 20 ka. The coastal part of the Wahiba exhibits a different timing of eolian accumulation, with dune building at 34–24, 16–13, and approximately 9–8 ka (Preusser et al., 2005). In contrast to the main dune area, there was little dune activity at the LGM and during the Younger Dryas. The area experienced wetter conditions with high groundwater levels during the early Holocene.

Thar Desert

The Thar Desert in northwestern India (Figure 9) is dominated by compound parabolic dunes, with smaller areas of linear, crescentic, reticulate, or network dunes and sand sheets (Kar, 1993). The dunes are underlain in part by carbonate-cemented eolianites, which range in age from 21 to 210 ka. Dune trends indicate their formation by winds associated with the southwest monsoon. Many of the dunes are partially or completely stabilized by vegetation, and the limit of stabilized dunes extends into areas where the mean annual rainfall exceeds 500 mm.

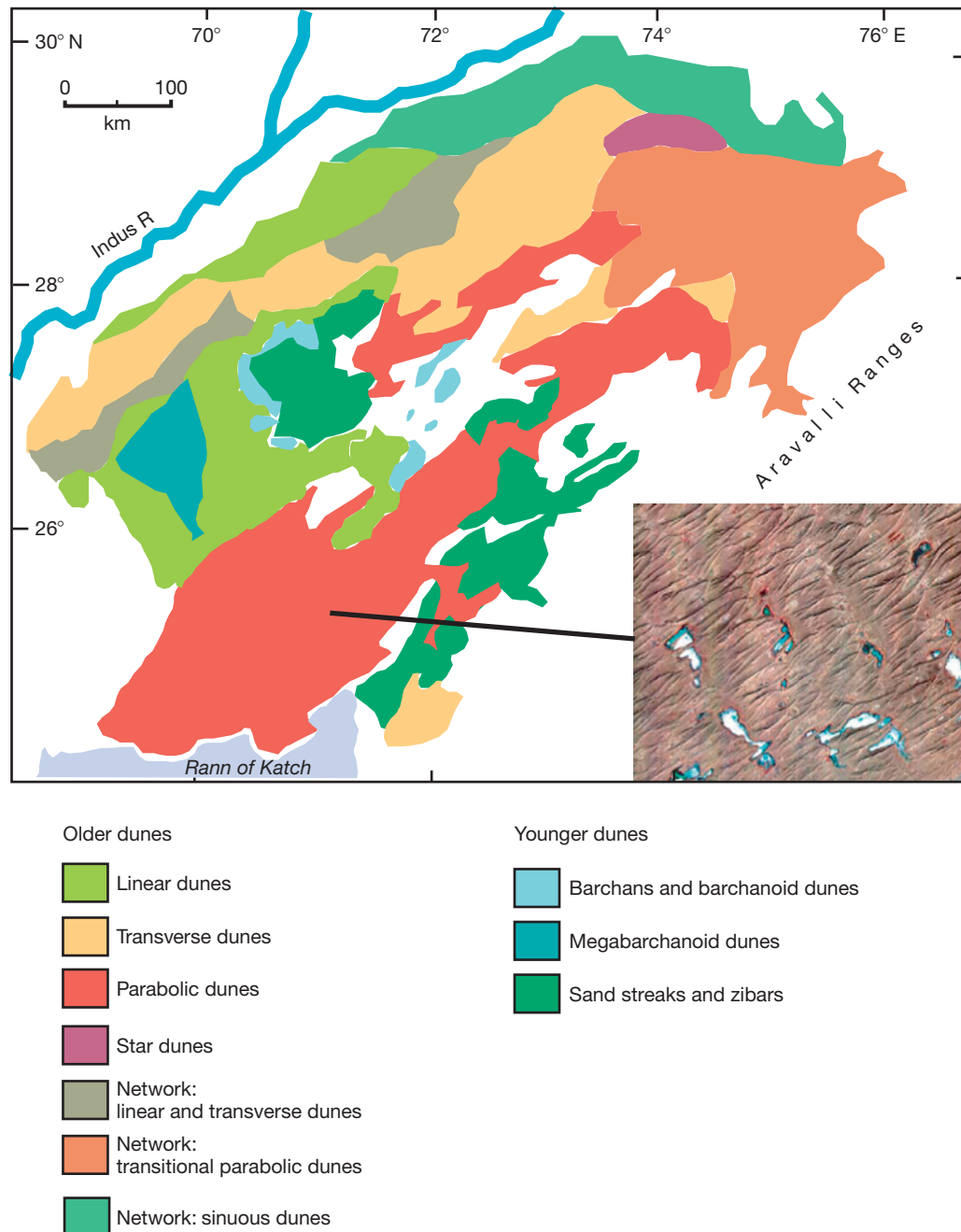


Figure 9 Dunes in the Thar Desert. Inset: aster image of parabolic dunes. Reproduced from Singhvi AK and Kar A (2004) The aeolian sedimentation record of the Thar Desert. *Proceedings of the Indian Academy of Sciences (Earth Sciences)* 113(3): 371–401.

The Thar Desert was the site of some of the first luminescence dating studies on eolian deposits (Singhvi et al., 1982), and subsequent studies have resulted in the development of a detailed chronology of dune-building episodes in the region (Singhvi and Kar, 2004). Eolian deposition in the area dates to 150 ka or older; major periods of construction occurred 115–100, ~75, ~55, and between 30 and 25 ka.

The available data suggest that despite an arid climate and an abundant sediment supply from the exposed shelf of the

Arabian Sea, late Pleistocene dune building in the Thar Desert did not occur coeval with the LGM but instead took place in a narrow window of time (14–10 ka) as the southwest monsoon was reestablished at the Pleistocene–Holocene transition. Major eolian activity occurred in the Holocene, but with important spatial variability. Dune building ceased in the southern Thar Desert in the early Holocene (10 ka) but continued until 5 ka in the north. In the core arid region of western Rajasthan, dune activity continued after 2 ka to 0.7–0.8 ka.

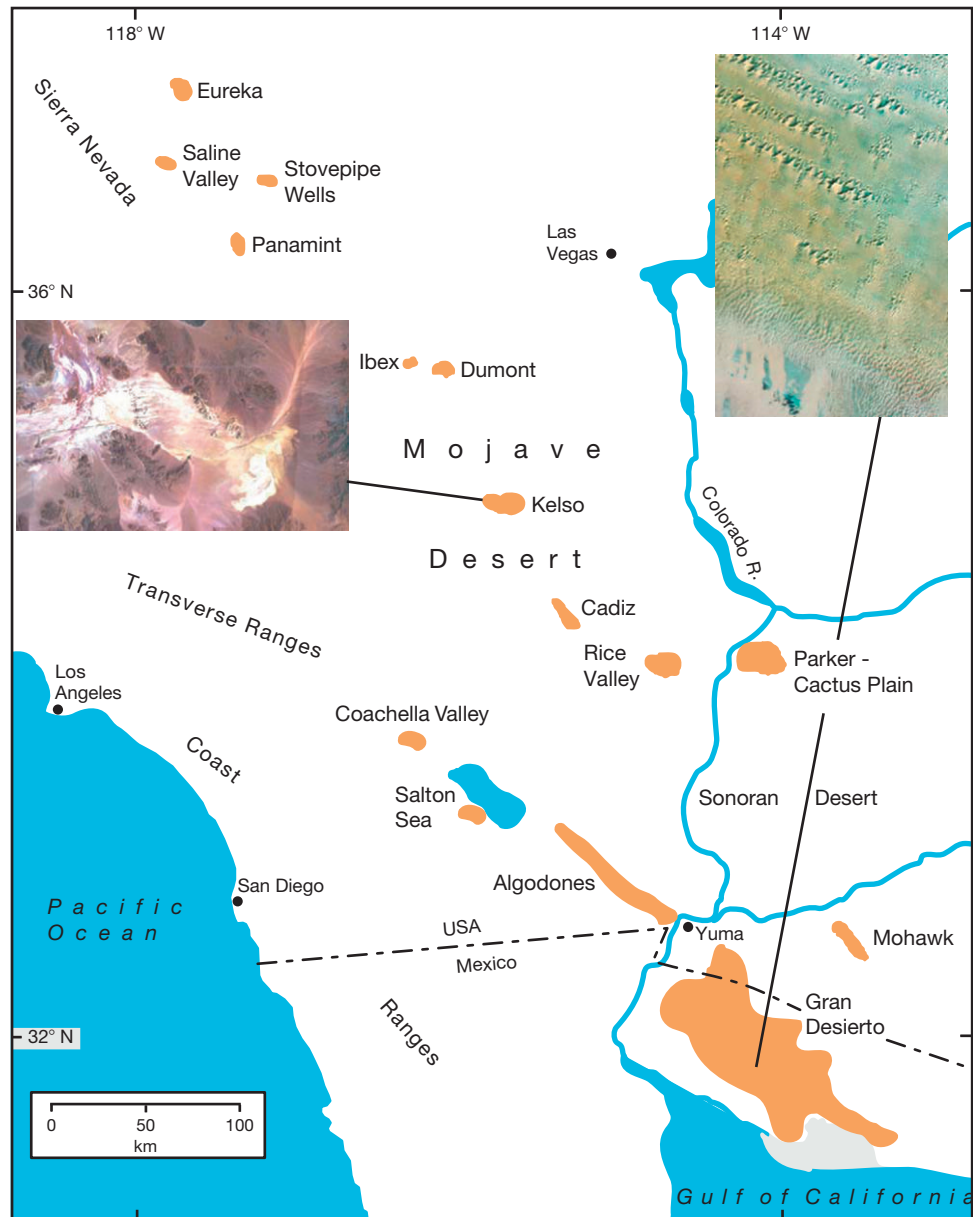


Figure 10 Distribution of dune fields in the Mojave and Sonoran deserts, with Landsat images of selected dune fields. Gran Desierto image shows star dunes in linear chains (26–12 ka) and crescentic dunes formed ~12 ka.

North American Deserts

Dune fields are limited in extent in the deserts of the southwestern United States and adjacent areas of Mexico (Figure 10). Sand deposits are composed of sand sheets, small dune fields, and sand ramps (amalgamated accumulations of eolian sand, and alluvial and colluvial deposits on the slopes of desert mountains).

Mojave Desert

Eolian sediments and landforms in the Mojave Desert of southern California and adjacent areas of Nevada and Arizona are concentrated along well-defined eolian sediment transport corridors that are characterized by areas of active and inactive

dunes and sand sheets, together with sand ramps. Geochemical and mineralogical studies indicate several major (e.g., the Mojave River and Colorado River) and many local sources of sediment, but there is a general west-to-east transport of sand.

The history of Late Quaternary eolian activity along the major sand transport corridors has been documented using thermoluminescence and infrared-stimulated luminescence (IRSL) dating of dunes and eolian deposits in sand ramps (Clarke and Rendell, 1998; Lancaster and Tchakerian, 2003). These indicate periods of eolian activity 30–20 and 15 or 20–7 ka. Mid- and late Holocene eolian deposits occur in the Kelso Dunes. Interstratified paleosols in many sand ramps (e.g., near Kelso Dunes and Rice Valley dune field) suggest one major and a number of minor periods of regional-scale

eolian stability 20–15, 14, and 4 ka. Comparison of the record of periods of eolian construction and stability in the Kelso Dunes system with the lacustrine record from Lake Mojave suggests that periods of eolian stability generally coincide with the highest lake levels, whereas periods of eolian construction are linked to periods of fluctuating or desiccating lakes but also overlap with periods of high lake levels. These observations indicate a strong influence of sediment supply on eolian processes in this region.

Multiple, short-duration episodes of late Holocene eolian accumulation are recorded close to modern sand source areas. IRSL ages for sediments in the Kelso Dunes (including the Cronese basins) cluster around 0.20–0.14, 0.5–0.4, 0.85–0.7, 1.7–1.4, 3–2, 4.2–3.5, and 6.7–5.4 ka. Episodes of eolian deposition correlate well with the record of major floods in the Mojave River system and likely reflect the effect of enhanced sediment supply from the Mojave River (Muhs et al., 2003).

Colorado and Sonoran Deserts

There are three main areas of eolian deposits in the Colorado and lower Sonoran Deserts: the Coachella Valley, the Algodones Dunes, and the Gran Desierto sand sea of northern Mexico (Lancaster, 1995).

The largest dune field in the Sonoran Desert lies in the Gran Desierto of northern Mexico, which has a spatially complex pattern of multiple generations of linear, crescentic, and star dunes, as well as extensive sand sheets (Figure 10), sourced mainly from Colorado River sediments. Recent OSL dating and geomorphic studies show that elements of the complex pattern represent short-lived eolian constructional events since ~25 ka. The simple dune patterns consist of late Pleistocene relict linear dunes formed between 25 and 12 ka, degraded crescentic dunes formed at ~12 ka, eastern crescentic dunes emplaced at ~7 ka, early Holocene western crescentic dunes, and star dunes formed during the past 3 ka (Beveridge, 2004). In the Algodones, Stokes

et al. (1997a) identified periods of eolian activity >30 ka, ~3.1 and <0.4 ka.

South America

Low-latitude dune fields occur in two settings in South America: (a) the largely active dunes of the Atacama Desert on the west coast and (b) the vegetation-stabilized relict dune fields of the Rio Negro and Sao Francisco areas of Brazil and the northern Pampa in Argentina. Chronologic data for the relict dune fields are summarized by Munyikwa (2005) and suggest eolian activity prior to 69 ka and at 63, 33, 28–27, 23–22, and 17.2–8 ka, with limited Holocene activity.

Conclusion

The available chronologic information shows that most sand seas have experienced episodic accumulation during the past 120 ka, with long periods of stability or nondeposition separated by relatively brief periods in which dune construction occurred (Figure 11). Luminescence dating indicates widespread dune activity in low-latitude deserts during the periods 120–100, 35–10, and since 5 ka. However, subsequent periods of eolian activity have reworked older deposits so that the preserved record in many areas is quite short.

The episodic nature of sand sea construction in low-latitude deserts is a direct consequence of the interplay between sand supply, availability, and mobility (Figure 12). Thus, sediment supplied to source areas during periods of increased fluvial activity is mobilized when it becomes available for transport in subsequent drier episodes, as occurred in the Simpson–Strzelecki Deserts in the late Pleistocene, the Mojave Desert in the early Holocene, and throughout the Sahara in the late Holocene. In other areas (e.g., UAE and Wahiba), sediment

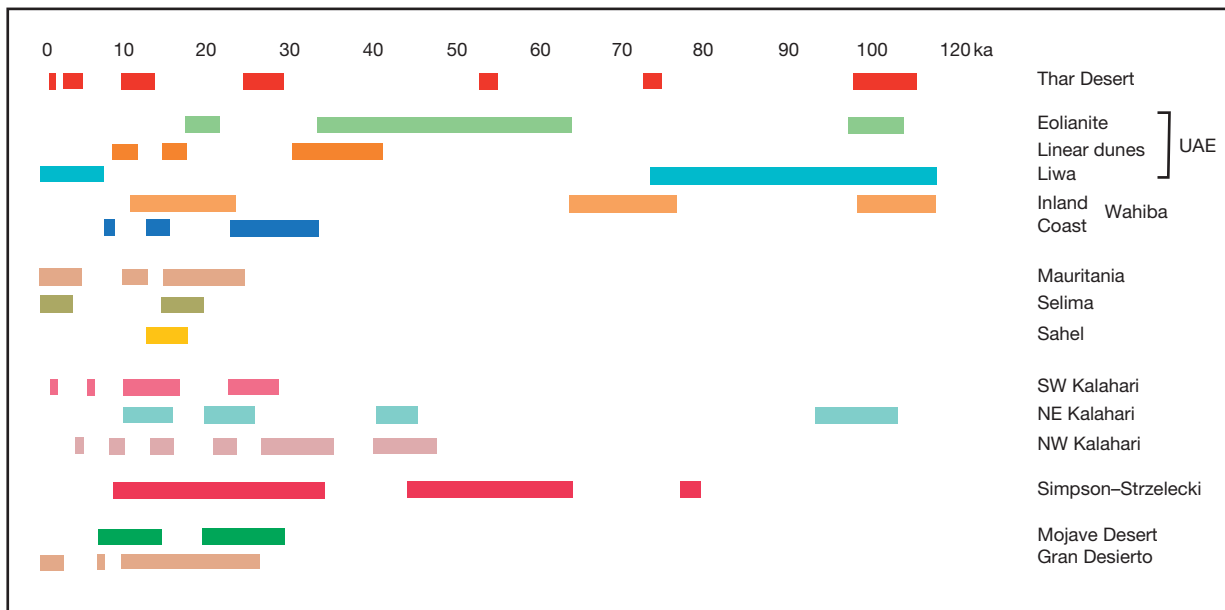


Figure 11 Chronological summary of periods of dune construction and/or activity during the past 120 ka, directly dated using luminescence methods. Compiled from a variety of sources referenced in text.

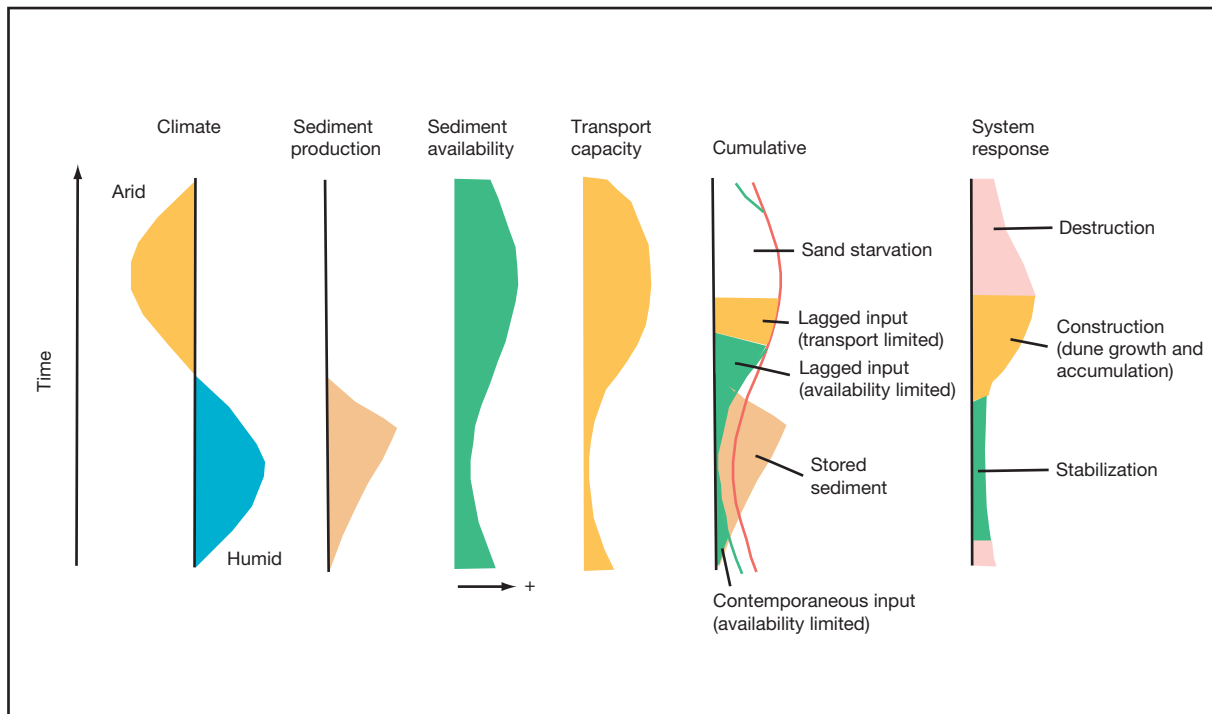


Figure 12 Schematic of the response of sand seas to climate change based on the principles of sediment configuration and sediment state (Kocurek and Lancaster, 1999). Diagram after Kocurek G (1998) Aeolian system response to external forcing factors – A sequence stratigraphic view of the Saharan region. In: Alsharan AS, Glennie KW, Whittle GL, and Kendall CGSC (eds.), *Quaternary Deserts and Climatic Change*, pp. 327–338. Rotterdam: Balkema.

supplied to shelf zones by biogenic activity became available for transport to dunes as sea level fell. In the case of the Wahiba sand sea and the Thar Desert, this sediment was only mobilized as the strength of the southwest monsoon increased following the LGM.

Sarnthein (1978) summarized the then available evidence for Late Quaternary eolian activity in low-latitude arid regions. He concluded that dune fields were much more extensive at the LGM (~20 ka). The recent infusion of chronologic information from luminescence dating studies broadly confirms Sarnthein's hypothesis but suggests that the temporal and spatial patterns of dune construction are more complex than those proposed by Sarnthein (Figure 11). For example, there was no dune construction in the northern Kalahari and Thar Deserts at this time.

Many areas of dunes, especially those composed of linear dunes, were formed or reformed during the LGM. Arid conditions and eolian activity increased in extent and/or intensity during this period, as corroborated by the record of dust flux in adjacent ocean cores. This was apparently the result of increased intensity and persistence of the trade wind circulation in the Sahara and elsewhere giving rise to increased sediment mobility, coupled with a decrease in the strength of monsoon circulations (and summer rainfall) in Africa and Asia, thereby increasing sediment availability as vegetation cover and soil moisture decreased. In many areas, increased sediment availability as a result of decreased rainfall and enhanced mobility (as a result of increased wind strength) coincided, resulting in widespread dune formation. However, sand seas in which

the dominant winds are related to the southwest monsoon (especially the Wahiba and Thar sand seas) experienced reduced activity at this time.

The response of low-latitude sand seas to Quaternary climate change is becoming much clearer, in large part as a result of the application of luminescence dating techniques, coupled with new paradigms (especially concepts of dune generations and eolian sediment state) for the geomorphic and sedimentary development of these extensive depositional systems. Future work will be increasingly focused on understanding spatial patterns in the response of sand seas to climate change; the relations between sediment supply, availability, and mobility; and the role of sub-Milankovitch-scale climate change in sand sea evolution.

See also: **Dune Fields:** High Latitudes; Mid-Latitudes.
Luminescence Dating: Optical Dating; Thermoluminescence.

References

- Beveridge CA (2004) *The Origin and Evolution of the Gran Desierto Sand Sea, Sonora, Mexico*. Master's Thesis, University of Texas at Austin, Austin.
- Breed CS, et al. (1979) Regional studies of sand seas using LANDSAT (ERTS) imagery. In: McKee ED (ed.) *A Study of Global Sand Seas*, pp. 305–398. Washington, DC: U.S. Geological Survey.
- Bristow CS, Duller GA, and Lancaster N (2005) Combining ground penetrating radar (GPR) surveys and optical dating to determine dune migration in Namibia. *Journal of the Geological Society of London* 162(2): 315–322.

- Clarke ME and Rendell HM (1998) Climatic change impacts on sand supply and the formation of desert sand dunes in the south-west U.S.A.. *Journal of Arid Environments* 39(3): 517–532.
- deMenocal P, et al. (2000) Abrupt onset and termination of the African Humid Period: Rapid climate responses to gradual insolation forcing. *Quaternary Science Reviews* 19(1–5): 347–361.
- Fryberger SG and Ahlbrandt TS (1979) Mechanisms for the formation of aeolian sand seas. *Zeitschrift für Geomorphologie* 23: 440–460.
- Gasse F (2002) Diatom-inferred salinity and carbonate oxygen isotopes in Holocene waterbodies of the western Sahara and Sahel (Africa). *Quaternary Science Reviews* 21: 737–767.
- Gasse F and Van Campo E (1994) Abrupt post-glacial events in West Africa and North Africa monsoon domains. *Earth and Planetary Science Letters* 126: 435–456.
- Glennie KW and Singhvi AK (2002) Event stratigraphy, palaeoenvironment and chronology of SE Arabian deserts. *Quaternary Science Reviews* 21: 853–869.
- Grove AT and Warren A (1968) Quaternary landforms and climate on the south side of the Sahara. *The Geographical Journal* 134(Part 2): 189–208.
- Jennings JN (1968) A revised map of the desert dunes of Australia. *Australian Geographer* 10: 408–409.
- Kar A (1993) Aeolian processes and bedforms in the Thar Desert. *Journal of Arid Environments* 25: 83–96.
- Kocurek G (1998) Aeolian system response to external forcing factors – A sequence stratigraphic view of the Saharan region. In: Alsharan AS, Glennie KW, Whittle GL, and Kendall CGSC (eds.) *Quaternary Deserts and Climatic Change*, pp. 327–338. Rotterdam: Balkema.
- Kocurek G, Havholm KG, Deynoux M, and Blakey RC (1991) Amalgamated accumulations resulting from climatic and eustatic changes, Akchar Erg, Mauritania. *Sedimentology* 38(4): 751–772.
- Kocurek G and Lancaster N (1999) Aeolian sediment states: Theory and Mojave Desert Kelso Dunefield example. *Sedimentology* 46(3): 505–516.
- Lancaster N (1989) *The Namib Sand Sea: Dune Forms, Processes, and Sediments*. Rotterdam: Balkema.
- Lancaster N (1995) Origin of the Gran Desierto Sand Sea: Sonora, Mexico: Evidence from dune morphology and sediments. In: Tchakerian VP (ed.) *Desert Aeolian Processes*, pp. 11–36. New York: Chapman & Hall.
- Lancaster N (1999) Sand seas. In: Goudie AS, Livingstone I, and Stokes S (eds.) *Aeolian Environments, Sediments, and Landforms*, pp. 49–70. New York: Wiley.
- Lancaster N and Tchakerian VP (2003) Late Quaternary eolian dynamics, Mojave Desert, California. In: Enzel Y, Wells SG, and Lancaster N (eds.) *Paleoenvironments and Paleohydrology of the Mojave and Southern Great Basin Deserts*, pp. 231–249. Boulder, CO: Geological Society of America.
- Lancaster N, et al. (2002) Late Pleistocene and Holocene dune activity and wind regimes in the western Sahara of Mauritania. *Geology* 30(11): 991–994.
- Mainguet M (1984) Space observations of Saharan aeolian dynamics. In: El Baz F (ed.) *Deserts and Arid Lands*, pp. 59–77. The Hague: Nyhoff.
- Mainguet M and Chemin MC (1983) Sand seas of the Sahara and Sahel: An explanation of their thickness and sand dune type by the sand budget principle. In: Brookfield ME and Ahlbrandt TS (eds.) *Eolian Sediments and Processes. Developments in Sedimentology*, pp. 353–364. Amsterdam: Elsevier.
- Mauz B and Felix-Henningsen P (2005) Paleosols in Saharan and Sahelian dunes of Chad: Archives of Holocene North African climate changes. *The Holocene* 15(3): 453–458.
- McKee ED (ed.) (1979) *A Study of Global Sand Seas*. Washington, DC: U.S. Geological Survey Professional Paper No. 1052.
- Muhs DR, Reynolds RR, Been J, and Skipp G (2003) Eolian sand transport pathways in the southwestern United States: Importance of the Colorado River and local sources. *Quaternary International* 104: 3–18.
- Munyikwa K (2005) Synchrony of Southern Hemisphere late Pleistocene arid episodes: A review of luminescence chronologies from arid aeolian landscapes south of the equator. *Quaternary Science Reviews* 24: 2555–2583.
- O'Connor PW and Thomas DSG (1999) The timing and environmental significance of late Quaternary linear dune development in western Zambia. *Quaternary Research* 52: 44–55.
- Pell SD, Chivas AR, and Williams IS (2000) The Simpson, Strzelecki and Tirari Deserts: Development and sand provenance. *Sedimentary Geology* 130: 107–130.
- Preusser F, Radies D, Driehorst F, and Matter A (2005) Late Quaternary history of the coastal Wahiba Sands, Sultanate of Oman. *Journal of Quaternary Science* 20(4): 395–405.
- Preusser F, Radies D, and Matter A (2002) A 160,000-year record of dune development and atmospheric circulation in southern Arabia. *Science* 296: 2018–2020.
- Radies D, Preusser F, Matter A, and Mange M (2004) Eustatic and climatic controls on the development of the Wahiba Sand Sea, Sultanate of Oman. *Sedimentology* 51(6): 1359–1386.
- Rognon P (1982) Pluvial and arid phases in the Sahara: The role of non climatic factors. *Palaeoecology of Africa* 12: 45–62.
- Sarnthein M (1978) Sand deserts during glacial maximum and climatic optima. *Nature* 272: 43–46.
- Singhvi AK and Kar A (2004) The aeolian sedimentation record of the Thar Desert. *Proceedings of the Indian Academy of Sciences (Earth Sciences)* 113(3): 371–401.
- Singhvi AR, Sharma YP, and Agrawal DP (1982) Thermoluminescence dating of sand dunes in Rajasthan, India. *Nature* 295: 313–315.
- Stokes S (1997) Dating of desert sequences. In: Thomas DSG (ed.) *Arid Zone Geomorphology*, pp. 607–637. New York: Wiley.
- Stokes S and Bray HE (2005) Late Pleistocene eolian history of the Liwa region, Arabian Peninsula. *Geological Society of America Bulletin* 117(11–12): 1466–1480.
- Stokes S and Horrocks J (1998) A reconnaissance survey of the linear dunes and loess plains of northwestern Nigeria: Granulometry and geochronology. In: Alsharan AS, Glennie KW, Whittle GL, and Kendall CGSC (eds.) *Quaternary Deserts and Climatic Change*, pp. 165–174. Rotterdam: Balkema.
- Stokes S, Kocurek G, Pye K, and Winspear NR (1997a) New evidence for the timing of aeolian sand supply to the Algodones dunefield and East Mesa area, southeastern California, USA. *Palaeogeography, Palaeoclimatology, Palaeoecology* 128: 63–75.
- Stokes S, Maxwell TA, Haynes CV, and Horrocks J (1998) Latest Pleistocene and Holocene sand sheet construction in the Selima Sand Sheet, Eastern Sahara. In: Alsharan AS, Glennie KW, Whittle GL, and Kendall CGSC (eds.) *Quaternary Deserts and Climatic Change*, pp. 175–184. Rotterdam: Balkema.
- Stokes S, Thomas DSG, and Shaw PA (1997b) New chronological evidence for the nature and timing of linear dune development in the southwest Kalahari Desert. *Geomorphology* 20(1–2): 81–94.
- Stokes S, Thomas DSG, and Washington R (1997c) Multiple episodes of aridity in southern Africa since the last interglacial period. *Nature* 388: 154–158.
- Stuut J-B and Lamy F (2004) Climate variability at the southern boundaries of the Namib (southwestern Africa) and Atacama coastal deserts during the last 120,000 yr. *Quaternary Research* 62(3): 301–309.
- Swezey C (2001) Eolian sediment responses to late Quaternary climate changes: Temporal and spatial patterns in the Sahara. *Palaeogeography, Palaeoclimatology, Palaeoecology* 167: 119–155.
- Swezey C, et al. (1999) Response of aeolian systems to Holocene climatic and hydrologic changes on the northern margin of the Sahara: A high resolution record from the Chott Rharsa basin, Tunisia. *The Holocene* 9(2): 141–148.
- Teller JT, Rutter NW, and Lancaster N (1990) Sedimentology and palaeohydrology of late Quaternary lake deposits in the northern Namib Sand Sea, Namibia. *Quaternary Science Reviews* 9: 343–364.
- Thomas DSG and Shaw PA (1991) *The Kalahari Environment*. Cambridge: Cambridge University Press.
- Thomas DSG and Shaw PA (2002) Late Quaternary environmental change in central southern Africa: New data, synthesis, issues and prospects. *Quaternary Science Reviews* 21: 783–797.
- Thomas DSG, Stokes S, and Shaw PA (1997) Holocene aeolian activity in the southwestern Kalahari Desert, southern Africa: Significance and relationships to late-Pleistocene dune-building events. *The Holocene* 7(3): 273–281.
- Wasson RJ (1983) The Cainozoic history of the Strzelecki and Simpson dunefields (Australia), and the origin of the desert dunes. *Zeitschrift für Geomorphologie Supplement* 45: 85–115.
- Wasson RJ, Fitchett K, Mackey B, and Hyde R (1988) Large-scale patterns of dune type, spacing, and orientation in the Australian continental dunefield. *Australian Geographer* 19: 89–104.
- Wilson IG (1971) Desert sandflow basins and a model for the development of ergs. *The Geographical Journal* 137(2): 180–199.
- Wilson IG (1973) Ergs. *Sedimentary Geology* 10: 77–106.
- Wood WW and Imes JL (1995) How wet is wet? Precipitation constraints on late Quaternary climate in the Arabian Peninsula. *Journal of Hydrology* 164: 263–268.

El Niño Southern Oscillation *see* **Paleoceanography, Records:** Postglacial South Pacific; **Paleoclimate Modeling:** Paleo-ENSO

Electron Spin Resonance Dating *see* **Luminescence Dating:** Electron Spin Resonance Dating

Eolian Sediments *see* **Paleosols and Wind-Blown Sediments:** Overview; Nature of Paleosols; Mineral Magnetic Analysis; Soil Micromorphology; Weathering Profiles; Soil Morphology in Quaternary Studies; Biogeochemical Role of Dust in Quaternary Climate Cycles

Eolian Records, Deep-Sea Sediments

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Glossary

Biogenic ooze: Sediment composed of the remains of surface-dwelling plankton. These sediments are some combination of calcium carbonate, CaCO_3 and biogenic opal, $\text{SiO}_2 \cdot n\text{H}_2\text{O}$.

Hemipelagic mud: Silty sediment delivered to the ocean by rivers and transported off the continental shelf in the water column. The mud grains settle out across the outer continental shelf and the continental slope. This is the sediment that eventually forms shale and mudstone.

Milankovitch variability: Milankovitch hypothesized that variations in the orbit of Earth associated with precession

(at periods of 19 and 23 ka), tilt of the spin axis (41 ka), and eccentricity of the orbit (100 and 400 ka) had a pronounced influence on climate.

Red or pelagic clay: A common seafloor deposit that consists of mineral dust grains and authigenic oxides and hydroxides. These deposits accumulate very slowly and are devoid of all biogenic material except fossil fish teeth.

Spectral analysis: A mathematical technique that identifies nonrandom variability in a sample set and discovers the amount and frequency of such fluctuations.

Introduction

The transport of eolian grains and their deposition in the deep sea has been of interest to scientists for over 150 years, beginning with Darwin's observations made aboard the *Beagle* (Darwin, 1846). Radczewski (1939) examined the reddish quartz grains in the Cape Verde Basin, downwind from the Sahara, and noted the importance of dust transport in bringing those grains to the central Atlantic. Rex and Goldberg (1958) explained the quartz in abyssal sediments of the North Pacific by dust transport from Asia. Since the 1960s, studies of the transport of mineral aerosols have abounded. The reader would find overview papers by Prospero et al. (1983, 2002),

Rea et al. (1985), Duce et al. (1991), Rea (1994), and Grousset and Biscaye (2005) and books by Pèwè (1981), Pye (1987), and Leinen and Sarnthein (1989) to provide useful depictions of the transport and deposition process. More recently, interest in dust transport and deposition has spread to the biologic oceanographic community, with the formulation of the 'iron hypothesis,' which holds that dust supplies the limiting nutrient iron to otherwise unproductive ocean surface waters (Martin, 1990; Martin and Fitzwater, 1988). Research on various aspects of this topic remains vigorous today (Jickells et al., 2005). Dust is also important in understanding the effects of radiative forcing on the Earth and has received considerable attention from atmospheric scientists (Sokolik et al., 2001).

The demonstration that availability of dust varies with the climate of the source region (Prospero and Nees, 1977, 1986; Prospero et al., 1970) caused marine geologists to begin to look down core into the sediments for this useful record of past continental climates. The first such studies were conducted in the North Atlantic, downwind from the Sahara (Parkin, 1974; Parkin and Padgham, 1975; Parkin and Shackleton, 1973). Samthein et al. (1981, 1982) have studied the same system and provided good overviews of the geologic history of Saharan dust deposition (Ruddiman et al., 1989). Clemens and Prell (1990, 1991), Sirocko and Samthein (1989), and Sirocko et al. (1991) have studied the monsoon-related dust transport from the deserts of Arabia to the northwestern Indian Ocean. The work of the authors, some of which is described below, has been largely in the Pacific Ocean and stems from the early studies of Heath and Leinen (Heath, 1969; Leinen, 1989; Leinen and Heath, 1981).

Nature of the Data

Arid and semiarid continental surfaces are the source of dust in the atmosphere and the ocean. Studies of modern climatology show that most of the wind-borne dust transport each year occurs in a few large storms, commonly in the spring or summer. Satellite data are now available that allow us to track a single dust storm from the Sahara to the Caribbean, or the Gobi to California. The dust entrained by these large storms shows a mixture of particle sizes, with the large particles falling out quickly along the transport path, and the fine grains remaining to be transported transoceanic distances. These fine grains end up in the ocean sediments and, in the absence of other sedimentary components, form the red or pelagic clay deposits that characterize half of the area of the deep-sea floor.

Much of the work on eolian sediment has focused on the fundamental sedimentology of the dust: the transport and depositional processes and the history of those processes provided by the deep-sea record. More recently, the geochemistry of dust has played an increasing role in the overall research effort. The overarching goal of paleoclimate studies based on the record of dust deposition is to understand the geologic history of continental climates and the nature and intensity of the zonal wind systems. There are other ways to approach questions of continental climate, but there are no other direct records of the history and patterns of the wind systems (Rea, 1994).

Students of eolian deposition are faced with several problems from the outset. First, dust makes up only 1 or 2% of the $15\text{--}20 \times 10^{15}$ g of terrigenous sediment brought to the oceans every year. This mainly river-borne material becomes the hemipelagic mud (clayey silts that move offshore in suspension) that drapes the continental margins and extends out to sea for unknown distances. Since 1 or 2% of the river loading can overwhelm the eolian signal in a core, one must be certain that sediments used for eolian information are from cores raised far from land or are otherwise shielded from hemipelagic processes. The most useful sediment for paleoceanographic analyses is biogenic ooze, the fossil remains of sea-surface plankton. In such calcite- and/or opal-rich sediments, the dust makes up only a small percent of the total content, and so must be isolated from the bulk sediment by a somewhat lengthy laboratory chemical procedure (Clemens and Prell, 1990; Hovan, 1995;

Rea and Janecek, 1981). Upon completion of the extraction process, the mineral grains are available for determinations of flux (mass accumulation rate – MAR, commonly given in $\text{mg cm}^{-2} \text{ ka}^{-1}$), particle-size analysis, mineralogy, and geochemistry. It is particularly important to ‘clean’ the grains prior to geochemical analysis because the oxide coatings on the particles carry their own geochemical signature. As the MAR values are a product of the weight percent of the dust component, the ambient linear sedimentation rate, and the dry bulk density of the sediment, resulting values depend on the quality of all three determinations. In new or well-cured cores, the sedimentation rate is the least well known of these values; the percent and bulk density values are generally good to $\pm 5\%$ or so.

Interpretation of the Data

The MAR of eolian dust is a good measure of source area aridity. Humid, vegetated regions put very little dust into the air, while semiarid and arid regions supply considerable amounts. There are hyperarid areas, where there is not enough water to break down the continental feldspars into clays, that generate little dust today; such regions have been deflated. We have been able to map the MAR of dust in the deep sea as being based on the sedimentary record (Rea, 1994). These maps are similar to those provided by Duce et al. (1991) based on modern mineral aerosol transport and show dust flux maxima usually in excess of $1000 \text{ mg cm}^{-2} \text{ ka}^{-1}$ downwind of the main source regions in central and eastern Asia, the Sahara and Sahael, and Arabia and the Horn of Africa. Since most of the land is in the Northern Hemisphere, most of the dust is found there too. The South Pacific, South Indian, and Southern Ocean are sites of minimal dust input, on the order of $1 \text{ mg cm}^{-2} \text{ ka}^{-1}$ or less.

Eolian dust has a grain-size distribution that peaks at about $2 \mu\text{m}$, with essentially no material coarser than $10 \mu\text{m}$ (Rea and Hovan, 1995). All the work on eolian grain size has assumed that the grain size is related to the intensity of the transporting wind, but clear evidence of this was not produced until recently when Clemens reported on sediment trap studies in the Indian Ocean, downwind from the Arabian deserts (Clemens, 1998). The size of the dust grains changes very little downwind, after the coarser materials have settled out, such that after perhaps 2000 km from the original dust storm site, the size is uniform. Maps of eolian grain size show this remarkable uniformity of size across the entire North Pacific Ocean (Rea and Hovan, 1995). Another working hypothesis is that these grains are in equilibrium with the transporting winds, and that the stronger winds will transport a coarser equilibrium grain size of material.

There is one more point to make about these data. At the distal locations sampled in our deep-sea cores, the MAR and grain size of dust vary independently. The closest analogy is with the competence and capacity of rivers. A large river like the Mississippi might carry a lot of mud, but it will all be fine grained. Conversely, a mountain stream might move sand and gravel, but not in Mississippi-like quantities. Cross-plots of data from those cores, which have good flux and size information, show no correlation between the two kinds of data (Rea, 1994). This means that the suggestion that more dust and stronger winds are directly related is a misconception.

Spectral analysis of eolian grain size and flux records have demonstrated that both respond to orbital forcing, commonly at the periodicities of ~ 100 ka (reflecting changing eccentricity of the orbit), 41 ka (tilt of the spin axis with respect to the orbital plane), and 20 ka (precession around the orbit), originally calculated by Milankovitch (1941). Milankovitch variability in dust grain-size data was shown originally by Janecek and Rea (1984, 1985). The first detailed study was based on a core from the equatorial Pacific and demonstrated that the variability in the eolian grain-size data was coherent and in phase with a radiolarian-based proxy for upwelling, quantitatively linking wind intensity and equatorial upwelling in this Quaternary core. The main spectral peak in the grain size and upwelling proxy data occurs at about 30 ka, a cross-product of eccentricity and tilt periods rather than one of the primary orbital periods (Pisias and Rea, 1988). Orbital forcing at approximately 30 ka is a common finding in a variety of spectral-analysis studies of atmospheric and surface ocean proxy records (Rea, 1994).

Studies of a North Pacific core directly downwind from the China Loess Plateau documented orbital variability in both the flux data, which matches loess deposition (see below), and the dust grain-size data (Hovan et al., 1991). The dust

accumulation rate record showed spectral power at the primary orbital periods of about 100, 41, and 19 ka, but eolian grain-size data from this same core exhibit spectral peaks at 100, 50, and 25–35 ka (Hovan et al., 1991). Clemens and Prell (1990, 1991) analyzed sediments downwind from the Arabian Desert to understand the monsoonal influence in the Gulf of Arabia. They show pronounced orbital power at the frequencies associated with tilt and precession, along with noticeable power at about 35 and 54–59 ka in the grain-size record. The MAR of eolian dust varies on longer, 100 and 19–23 ka periods. This last observation is important because it means that wind intensity is not linked only to the 100-ka glacial–interglacial cycles of the later Quaternary but varied on shorter periodicities, an observation consistent with the earlier work of Pisias and Rea (1988). The existence of important amounts of variability in the atmospheric transport process at periods other than the primary periodicities determined by Milankovitch (1941) remains a puzzle and is a continuing topic of research.

Much work has concentrated on determining the geochemistry of the dust, as a way to determine both provenance and, if there is a long record, changes in the nature of the eolian source region. Both elemental, largely rare-earth element information

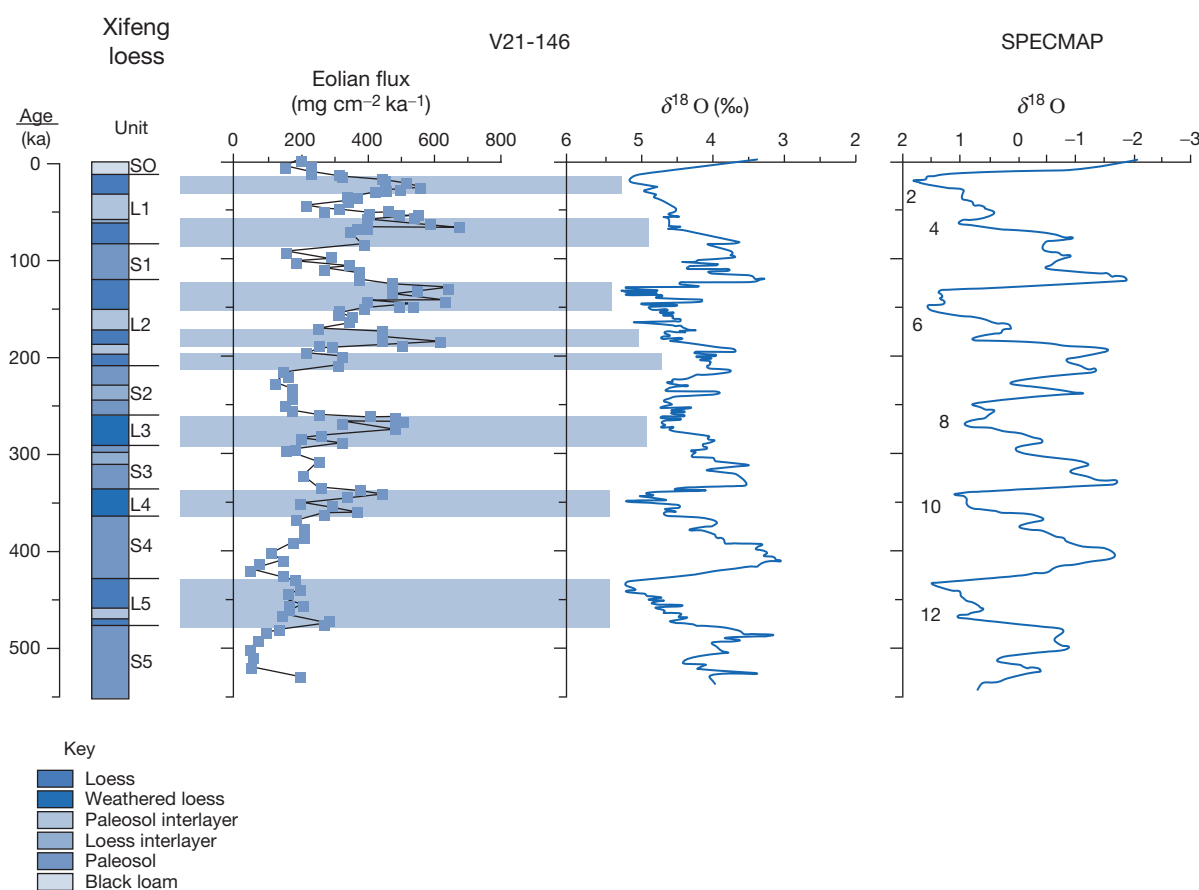


Figure 1 Dust accumulation record spanning the past 500 ka from core V21-146 in the northwestern Pacific Ocean. Loess (L) and soil (S) horizons which are shown at left are from the section at Xifeng. Dust flux maxima faithfully match the loess horizons. $\delta^{18}\text{O}$ records from the core, generated originally by Shackleton, show that dust flux maxima occur at glacial maxima as indicated by heavy values on the isotope curve. The global average isotope curve, the SPECMAP stack, is shown for comparison. Glacial ages are identified with even numbers on the SPECMAP record. Reproduced from Hovan SA, Rea DK, and Pisias NG (1991) Late Pleistocene continental climate and oceanic variability recorded in northwest Pacific sediments. *Paleoceanography* 6: 349–370.

(Olivarez et al., 1991; Weber et al., 1996), and isotopic, based on the isotopes of Nd, Sr, and Pb (Jones et al., 1994, 2000; Pettke et al., 2000, 2002), data have been presented. The first major finding of this work is that maps of the isotopic values of the mineral component of surficial sediments demonstrate that dust from Asia dominates North Pacific seafloor sediments, from China almost all the way east to North America and south to the Intertropical Convergence Zone (ITCZ) (Jones et al., 1994, 2000). North America supplies surprisingly little dust to the North Pacific (Jones et al., 1994, 2000). South of the ITCZ, dust has the geochemical signature of young andesite, implying a Central or South American source. This transition at the ITCZ is of interest because the first good downcore geochemistry on a red clay core, by Kyte et al. (1993), showed a change from a younger Asian source to an older American dust source some time during the late Oligocene. Since the core studied now lies at about 30° N and can be backtracked to about 20° N, this implies that the ITCZ once lay far north of its present location. One of the best examples of using the geochemistry of dust to 'fingerprint' the source regions has been the identification of the source of dust now found in the Antarctic (Argentine Patagonia; Grousset et al., 1992) and Greenland (central Asia; Biscaye et al., 1997) ice cores.

Quaternary Records from the North Pacific

A long-standing and continuing goal of paleoclimatologists is to link together records from the land surface, and the ocean, the main engine of climate change. The primary paleoceanographic record of climate change, linked to every paleoceanographic proxy, is the oxygen isotope record. The classic land record of late Cenozoic climate change is the sequence of loess and soil horizons such as those that cover the Loess Plateau in China. The loess horizons are formed in cold, dusty glacial times, and the soils are the product of warmer and more humid, thus vegetated, interglacials. Loess deposits form a remarkably detailed record of climate in Asia and elsewhere. The only way in which the $\delta^{18}\text{O}$ and loess records can be joined is in a marine core that can be tied to a loess sequence and has a good $\delta^{18}\text{O}$ record. Hovan et al. (1989, 1991) were presented with this circumstance in the form of piston core V21-146 from Shatsky Rise (37°41' N, 163°02' E) in the northwestern Pacific Ocean. The dust MAR record in V21-146 matches well with the loess-soil records from China and enables us to confirm by direct observation that loess maxima occur during the glacial maxima and that, within the accuracy of the data sets, the $\delta^{18}\text{O}$ indication of maximum glaciation matches the loess deposition peak (Figure 1; Hovan et al., 1989, 1991).

Cenozoic Records of Atmospheric Circulation and Continental Climate

All the good, longer records of dust accumulation are from sediment cores that now underlie the North Pacific-prevailing westerlies. The first of these records to have been generated is still the most revealing, that from giant piston core LL44-GPC3 (30°19' N, 157°49' W). Janecek and Rea (1983) determined the 70-million-year-long record of continental aridity and wind intensity from this red clay core. The dust flux data (Figure 2) showed little input in Oligocene and older times, a doubling of the dust flux at about 24 Ma, and a large increase associated

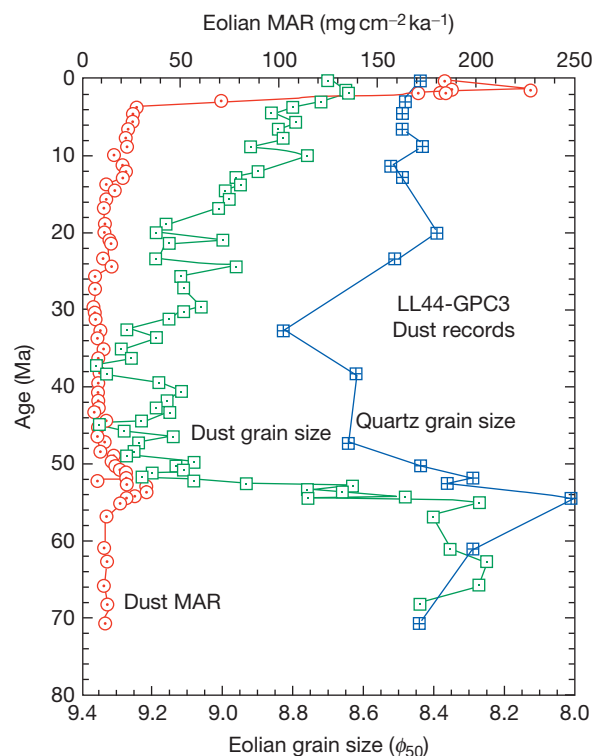


Figure 2 Dust accumulation rate (open circles) and grain size of dust (open squares) and quartz grains (crossed squares) extracted from piston core LL44-GPC3, raised from the central North Pacific Ocean. Size is calculated as median grain size and shown in phi units (ϕ_{50}).

with late Cenozoic cooling and drying of the Northern Hemisphere. It was the grain-size record from this core that provided the important new information in this study. The middle and late Cenozoic grain size shows a coarsening beginning sometime in the late Eocene and continuing, perhaps in steps, until the present. The surprise in this record is the large change in grain size that occurs lower in the core. This large change in dust size occurs in other North Pacific cores and has been dated as occurring exactly at the Paleocene–Eocene boundary (see Rea et al., 1990). It records a remarkable reduction in wind intensity from latest Cretaceous and Paleocene winds similar to present to a regime of very sluggish atmospheric circulation in the Eocene. These data are one of the fundamental pieces of information that underlie the grand quandary of Eocene climate: how, when both atmospheric and oceanic circulation appear sluggish, is it possible to transport enough heat to keep the polar regions warm enough to support crocodiles and palm trees? The question remains unanswered.

The details of late Cenozoic Asian aridity and wind intensity are best found in sediments recovered at North Pacific ODP Site 885/886 (44°41' N, 168°15' W). The section recovered at Site 885/886 is in biogenic ooze with a good magnetic reversal stratigraphy and so has far better age control than do the red clay cores farther south. Information from this site (Figure 3; Rea et al., 1998) shows a number of important aspects of late Cenozoic climate change. First, the dust flux data show an order of magnitude increase at 3.6 Ma, 1 My before the onset of major Northern Hemisphere glaciation at about 2.6 Ma.

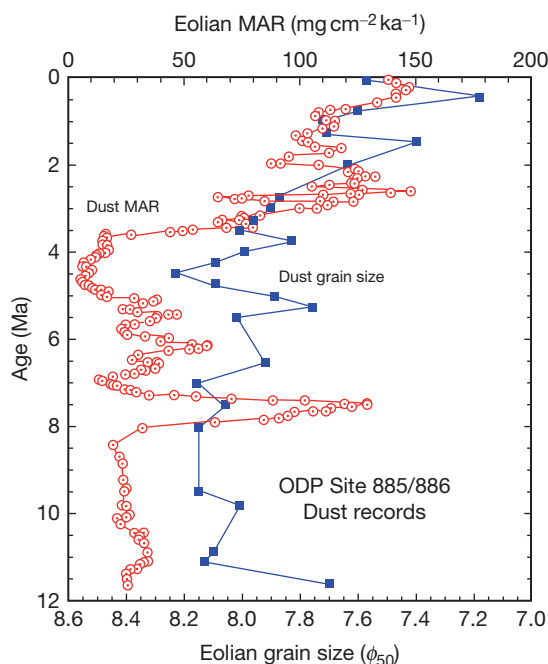


Figure 3 Dust accumulation rate (open circles) and grain size (solid squares) for the past 12 My from the North Pacific. Size is calculated as median grain size and shown in phi units (ϕ_{50}).

This increase reflects the rapid drying of the central Asian basins north of the Tibetan uplift, as these regions were cut off by the rising plateau from their Indian Ocean moisture supply. Other work has shown that conglomeratic sediments sweep north from the north edge of the Tibetan Plateau at this time, a clear indication of rapid uplift then (Fang et al., 2005; Zheng et al., 2004).

The eolian grain-size information from Site 885/886 tells a different story. Size and flux data vary independently from each other, as noted above. The record from this site, along with that from GPC3, is a good example of this at longer timescales. The grain-size data show that the increase in late Cenozoic wind intensity, presumably a response to high-latitude cooling, begins to increase at about 4.5 Ma, 1 My before the dust flux increased. Together, these records show that Northern Hemisphere glaciation did not occur suddenly and without prior indications of change. Rather a series of events, perhaps 2 or 3 My in duration, led up to the rapid buildup of continental ice sheets at about 2.6 Ma.

Summary

Dust forms only a small percentage of the sediment of the ocean floor, but in broad regions away from continents, the terrigenous component of deep-sea sediment is dominated by dust. These materials provide several kinds of paleoenvironmental information: a history of the aridity of the eolian source region in the MAR data, a history of the strength of the transporting winds in the grain-size data, a record of provenance in the dust geochemistry and knowing provenance, the nature of past zonal winds. There is no other kind of direct proxy for the

nature of past atmospheric circulation (Rea, 1994). As the atmosphere is responsible for half of the global heat transport, this kind of information is particularly valuable in paleoclimatic reconstructions.

See also: Loess Deposits: Origins and Properties. Glacial Climates: Effects of Atmospheric Dust. Loess Records: China.

References

- Biscaye PE, Grousset FE, Revel M, et al. (1997) Asian provenance of last glacial maximum dust in the GISP-2 ice core, Summit, Greenland. *Journal of Geophysical Research* 102: 26765–26781.
- Clemens SC (1998) Dust response to seasonal atmospheric forcing: Proxy evaluation and calibration. *Paleoceanography* 13: 471–490.
- Clemens SC and Prell WL (1990) Late Pleistocene variability of Arabian Sea summer monsoon winds and continental aridity: Eolian records from the lithogenic component of deep-sea sediments. *Paleoceanography* 5: 109–145.
- Clemens SC and Prell WL (1991) Late Quaternary forcing of Indian Ocean summer monsoon winds: A comparison of Fourier model and general circulation model results. *Journal of Geophysical Research* 96: 22683–22700.
- Darwin C (1846) An account of the fine dust which often falls on vessels in the Atlantic Ocean. *Quarterly Journal of the Geological Society of London* 2: 26–30.
- Duce RA, Liss PS, Merrill JT, et al. (1991) The atmospheric input of trace species to the world ocean. *Global Biogeochemical Cycles* 5: 193–259.
- Fang X, Yan M, Van der Voo R, et al. (2005) Late Cenozoic deformation and uplift of the NE Tibetan Plateau: Evidence from high-resolution magnetostratigraphy of the Guide Basin, Qinghai Province, China. *Geological Society of America Bulletin* 117: 1208–1225.
- Grousset FE and Biscaye PE (2005) Tracing dust sources and transport patterns using Sr, Nd and Pb isotopes. *Chemical Geology* 222: 149–167.
- Grousset FE, Biscaye PE, Revel M, et al. (1992) Antarctic ice core dusts at 18 ky BP, Isotopic constraints on origins and atmospheric circulation. *Earth and Planetary Science Letters* 111: 175–182.
- Heath GR (1969) Mineralogy of Cenozoic deep-sea sediments from the equatorial Pacific Ocean. *Geological Society of America Bulletin* 80: 1997–2018.
- Hovan SA (1995) Late Cenozoic atmospheric circulation intensity and climatic history recorded by eolian deposition in the Eastern Equatorial Pacific, Leg 138. In: Pisias NG, Mayer LA, Janecek TR, Palmer Juslon A, and van Andel TH (eds.) *Proceedings of the Ocean Drilling Program, Scientific Results*, vol. 138, pp. 615–625. College Station, TX: Ocean Drilling Program.
- Hovan SA, Rea DK, and Pisias NG (1991) Late Pleistocene continental climate and oceanic variability recorded in northwest Pacific sediments. *Paleoceanography* 6: 349–370.
- Hovan SA, Rea DK, Pisias NG, and Shackleton NJ (1989) A direct link between the China loess and marine $\delta^{18}\text{O}$ records: Aeolian flux to the north Pacific. *Nature* 340: 296–298.
- Janecek TR and Rea DK (1983) Eolian deposition in the northeast Pacific Ocean: Cenozoic history of atmospheric circulation. *Geological Society of America Bulletin* 94: 730–738.
- Janecek TR and Rea DK (1984) Pleistocene fluctuations in Northern Hemisphere tradewinds and westerlies, Part I. In: Berger A, Imbrie J, Hays J, Kukla G, and Saltzman B (eds.) *Milankovitch and Climate*, pp. 331–347. Dordrecht: D. Reidel Pub. Co.
- Janecek TR and Rea DK (1985) Quaternary fluctuations in the northern hemisphere tradewinds and westerlies. *Quaternary Research* 24: 150–163.
- Jickells TD, An ZS, Andersen KK, et al. (2005) Global iron connections between desert dust, ocean biogeochemistry, and climate. *Science* 308: 67–71.
- Jones CE, Halliday AN, Rea DK, and Owen RM (1994) Neodymium isotopic variations in North Pacific modern silicate sediment and the insignificance of detrital REE contributions to seawater. *Earth and Planetary Science Letters* 127: 55–66.
- Jones CE, Halliday AN, Rea DK, and Owen RM (2000) Eolian inputs of lead to the North Pacific. *Geochimica et Cosmochimica Acta* 64: 1405–1416.
- Kyte FT, Leinen M, Heath GR, and Zhou L (1993) Cenozoic sedimentation history of the central North Pacific: Inferences from the elemental geochemistry of core LL44-GPC3. *Geochimica et Cosmochimica Acta* 57: 1719–1740.
- Leinen M (1989) The pelagic clay province of the North Pacific Ocean. In: Winterer EL, Hussong DM, and Drecker RW (eds.) *The Eastern Pacific Ocean and Hawaii. The Geology of North America*, pp. 323–335. Boulder, CO: Geological Society of America.

- Leinen M and Heath GR (1981) Sedimentary indicators of atmospheric activity in the northern hemisphere during the Cenozoic. *Palaeogeography, Palaeoclimatology, Palaeoecology* 36: 1–21.
- Leinen M and Sarnthein M (eds.) (1989) *Paleoclimatology and Paleometeorology: Modern and Past Patterns of Global Atmospheric Transport*, p. 909. Dordrecht: Kluwer Academic.
- Martin JH (1990) Glacial–interglacial CO₂ change: The iron hypothesis. *Paleoceanography* 5: 1–13.
- Martin JH and Fitzwater SE (1988) Iron deficiency limits phytoplankton growth in the north-east Pacific subarctic. *Nature* 331: 341–343.
- Milankovitch M (1941) *Kanon der Erdbestrahlung und seine Anwendung auf das Eiszeitenproblem*, pp. 1–633. Beograd: Koniglich Serbische Academie. (English translation by Israel Program for Scientific Translations, Jerusalem, 1969).
- Olivarez AM, Owen RM, and Rea DK (1991) Geochemistry of eolian dust in Pacific pelagic sediments: Implications for paleoclimatic interpretations. *Geochimica et Cosmochimica Acta* 55: 2147–2158.
- Parkin DW (1974) Trade-winds during the glacial cycles. *Proceedings of the Royal Society of London A* 337: 73–100.
- Parkin DW and Padgham RC (1975) Further studies on trade winds during glacial cycles. *Proceedings of the Royal Society of London A* 346: 245–260.
- Parkin DW and Shackleton NJ (1973) Trade wind and temperature correlations down a deep-sea core off the Saharan coast. *Nature* 245: 455–457.
- Pettke T, Halliday AN, Hall CM, and Rea DK (2000) Dust production and deposition in Asia and the North Pacific Ocean over the past 12 Myr. *Earth and Planetary Science Letters* 178: 1–17.
- Pettke T, Halliday AN, and Rea DK (2002) Cenozoic evolution of Asian climate and sources of Pacific seawater Pb and Nd derived from eolian dust of sediment core LL44-GPC3. *Paleoceanography* 17: 3.1–3.13.
- Pêwè TL (ed.) (1981) *Desert Dust: Origin, Characteristics and Effect on Man*, p. 303. Boulder, CO: Geological Society of America. Special Paper 186.
- Pisias NG and Rea DK (1988) Late Pleistocene paleoclimatology of the central Equatorial Pacific: Sea-surface response to the southeast tradewinds. *Paleoceanography* 3: 21–37.
- Prospero JM, Bonatti E, Schubert C, and Carlson TN (1970) Dust in the Caribbean atmosphere traced to an African dust storm. *Earth and Planetary Science Letters* 9: 287–293.
- Prospero JM, Charlson RJ, Mohnen V, et al. (1983) The atmospheric aerosol system: An overview. *Reviews of Geophysics and Space Physics* 21: 1607–1629.
- Prospero JM, Ginoux P, Torres O, Nicholson SE, and Gill TE (2002) Environmental characterization of global sources of atmospheric soil dust identified with the Numbus 7 total ozone mapping spectrometer (TOMS) absorbing aerosol product. *Reviews of Geophysics* 40(1): 1002. <http://dx.doi.org/10.1029/2000RG000095>.
- Prospero JM and Nees R (1977) Dust concentration in the atmosphere of the equatorial North Atlantic: Possible relationship to Sahelian drought. *Science* 196: 1196–1198.
- Prospero JM and Nees RT (1986) Impact of the North African drought and El Niño on mineral dust in the Barbados trade winds. *Nature* 320: 735–738.
- Pye K (1987) *Aeolian Dust and Dust Deposits*, p. 334. London: Academic Press.
- Radczewski OE (1939) Eolian deposits in marine sediments. In: Trask PD (ed.) *Recent Marine Sediments*, pp. 496–502. Tulsa, OK: American Association of Petroleum Geologists.
- Rea DK (1994) The paleoclimatic record provided by eolian deposition in the deep sea: The geologic history of wind. *Reviews of Geophysics* 32: 159–195.
- Rea DK and Hovan SA (1995) Grain size distribution and depositional processes of the mineral component of abyssal sediments: Lessons from the North Pacific. *Paleoceanography* 10: 251–258.
- Rea DK and Janecek TR (1981) Mass accumulation rates of the non-authigenic inorganic crystalline (eolian) component of deep-sea sediments from the western Mid-Pacific Mountains, Deep Sea Drilling Project Site 463. In: Thiede J and Vallier TL, et al. (eds.) *Initial Reports of the Deep Sea Drilling Project*, vol. 62, pp. 653–659. Washington, DC: U.S. Government Printing Office.
- Rea DK, Leinen M, and Janecek TR (1985) Geologic approach to the long-term history of atmospheric circulation. *Science* 227: 721–725.
- Rea DK, Snoeckx H, and Joseph LH (1998) Late Cenozoic eolian deposition in the North Pacific: Asian drying, Tibetan uplift, and cooling of the Northern Hemisphere. *Paleoceanography* 13: 215–224.
- Rea DK, Zachos JC, Owen RM, and Gingerich PD (1990) Global change at the Paleocene–Eocene boundary: Climatic and evolutionary consequences of tectonic events. *Palaeogeography, Palaeoclimatology, Palaeoecology* 79: 117–128.
- Rex RW and Goldberg ED (1958) Quartz contents of pelagic sediments of the Pacific Ocean. *Tellus* 10: 153–159.
- Ruddiman WF, Sarnthein M, Backman J, et al. (1989) Late Miocene to Pleistocene evolution of climate in Africa and the low-latitude Atlantic: Overview of Leg 108 results. In: Ruddiman W and Sarnthein M, et al. (eds.) *Proceedings of the Ocean Drilling Program, Scientific Results*, vol. 108, pp. 463–484. College Station, TX: Ocean Drilling Program.
- Sarnthein M, Tetzlaff G, Koopmann B, Wolter K, and Pflaumann U (1981) Glacial and interglacial wind regimes over the eastern Subtropical Atlantic and North-West Africa. *Nature* 293: 193–196.
- Sarnthein M, Thiede J, Pflaumann U, et al. (1982) Atmospheric and oceanic circulation patterns off Northwest Africa during the past 25 million years. In: von Rad U, Hinz K, Sarnthein M, and Seibold E (eds.) *Geology of the Northwest African Continental Margin*, pp. 545–604. Berlin: Springer.
- Sirocko F and Sarnthein M (1989) Wind-borne deposits in the northwestern Indian Ocean: Record of Holocene sediments versus modern satellite data. In: Leinen M and Sarnthein M (eds.) *Paleoclimatology and Paleometeorology: Modern and Past Patterns of Global Atmospheric Transport*, pp. 401–433. Dordrecht: Kluwer Academic.
- Sirocko F, Sarnthein M, Lange H, and Erlenkeuser H (1991) Atmospheric summer circulation and coastal upwelling in the Arabian Sea during the Holocene and the last glaciation. *Quaternary Research* 36: 72–93.
- Sokolik IN, Winker DM, and Bergametti G, et al. (eds.) (2001) Quantifying the radiative impacts of mineral dust (DUST). *Journal of Geophysical Research D: Atmospheres* 106: 18015–18493.
- Weber ET II, Owen RM, Dickens GR, Halliday AN, Jones CE, and Rea DK (1996) Quantitative resolution of eolian continental crustal material and volcanic detritus in North Pacific surface sediment. *Paleoceanography* 11: 115–127.
- Zheng H, Powell C, Rea DK, Wang J, and Wang P (2004) Late Miocene onset and mid-Pliocene enhancement of the East Asian Monsoon as viewed from land and sea. *Global and Planetary Change* 41: 147–155.

Equilibrium Line Altitude (ELA) Reconstruction *see* **Glacial Landforms, Ice Sheets**: Paleo-ELAs

Erratics, Glacial *see* **Glacial Landforms, Sediments**: Glacial Erratics and Till Dispersal Indicators

Extinctions, Quaternary Vertebrates *see* **Vertebrate Records**: Late Pleistocene Megafaunal Extinctions

F

Felsenmeer (Blockfields) *see* **Permafrost and Periglacial Features**: Block/Rock Streams

Fission-Track Dating

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Introduction

A fission track is a linear region of intense radiation damage formed by the passage of an energetic, charged nuclear particle through an insulating solid. According to a popular hypothesis, known as the 'ion explosion spike' model ([Fleischer et al., 1975](#)), the charged particle collides with atoms in its path, displaces them and removes their electrons, producing a narrow zone rich in positive ions, which repulse each other into interstitial positions and form vacancies. An elastic adjustment of the crystal or glass then occurs, straining the undamaged matrix ([Fleischer, 2004](#)). Another idea, called the 'thermal spike' model, sees the fast-moving fission fragment producing intense heat along its trajectory. After passage of the charged particle, the track core is quenched by heat loss to the adjacent lattice, leaving it in a disordered state. According to [Chadderton \(2003\)](#), it is likely that both mechanisms are involved in track formation. Maximum density damage exists at the site of fission and decreases to the track termini. This damaged zone persists in the solid after the fission fragment has come to a rest and is called a latent track. Its length is equal to the sum of the range of the two charged particles formed by nuclear fission, each moving in opposite directions. ^{238}U fission events form latent tracks that are 6–10 nm wide and 10–20 μm long and

account for most tracks in natural, terrestrial materials. Latent tracks can only be observed with a transmission electron microscope but can be made visible by chemical etching ([Young, 1958](#)) and readily seen under an optical microscope. Etching takes place by rapid dissolution of the disordered region of the track core, which exists in a state of higher free energy than the undamaged material. As can be seen in [Figure 1](#), the shape of the etched fission tracks varies in different materials, due mainly to differences in symmetry and magnitude of the difference between the track and bulk etch rates. Fission tracks in glass are relatively wide, conical pits due to its isotropic properties and a track etch rate that is not much greater than the bulk etch rate. Micas have diamond-shaped tracks, whereas tracks in zircon and apatite are needle-like because of the large difference between track and bulk etch rates, the former being much greater.

Fission tracks have varied uses ranging from applications in the fields of engineering, medicine, nuclear physics, and space sciences to the earth sciences ([Fleischer, 1998](#)). The most important applications in the earth sciences are thermochronology ([Reiners and Ehlers, 2005](#)) and geochronology. Thermochronology – the more active field at present – focuses on understanding the thermal history of rocks and duration and rates of geological processes, such as crustal uplift.

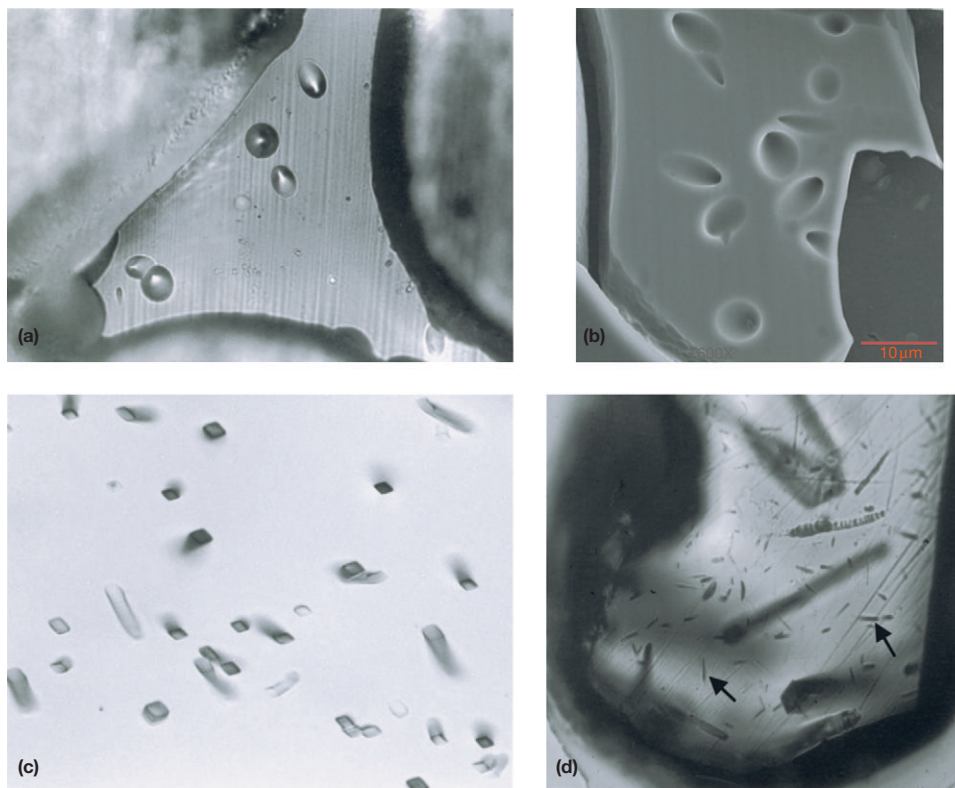


Figure 1 Fission tracks in glass shards, mica, and zircon. (a) Fission tracks (6–8 µm wide) in glass shard as seen using an optical microscope; (b) SEM image of fission tracks in glass shard (darker areas are the deepest part of the conical pits); (c) fission tracks in mica (~5 µm wide); (d) needle-like fission tracks in zircon from the Bishop Tuff, California (crystal is ~0.2 mm wide).

Geochronology, on the other hand, is primarily concerned with formation age, the topic of interest to us here.

Zircon and glass are the most suitable phases for dating Quaternary deposits by fission-track (FT) methods, although apatite and sphene may also be useful. The most desirable situation for dating Quaternary sediments is where they are associated with silicic tephra beds because not only can they be reliably dated using their zircon and glass shards, but their widespread distribution and common distinctiveness make tephra beds excellent markers, linking the stratigraphies of isolated sites, sometimes over distances exceeding 1000 km (see [Major Scale Forms](#)).

We outline in this article the basic principles of FT dating, detail the different methods used for glass and zircon, describe the correction procedures for partial track fading (PTF) in glasses, comment on the accuracy and precision of FT dates, and compare the results obtained to those of other dating methods, especially the $^{40}\text{Ar}/^{39}\text{Ar}$ technique. Finally, we illustrate the application of FT dating to the Quaternary Sciences by reference to a few case studies.

Age Equation

In principle, FT dating is similar to other radioisotopic dating methods except that the decay product of interest is the spontaneous fission track, not the daughter nuclide. If the rate of

decay of ^{238}U by spontaneous fission is known (the decay constant, λ_F), then we need only to know the U concentration and abundance of spontaneous fission tracks, assuming all are preserved, to calculate an age ([Price and Walker, 1963](#)). The basic age equation for an FT age, t , given by [Wagner and Van den haute \(1992\)](#) is:

$$t = 1/\lambda_D \ln \{1 + (\lambda_D/\lambda_F)(\rho_s/\rho_i)QGI\sigma_F\varphi\} \quad [1]$$

where λ_D is the total decay constant for ^{238}U ($1.551 \times 10^{-10} \text{ year}^{-1}$), λ_F is not well determined (values cluster around $7 \times 10^{-17} \text{ year}^{-1}$ and $8.5 \times 10^{-17} \text{ year}^{-1}$; [Bigazzi, 1981](#)), ρ_s is the areal track density of etched spontaneous tracks, ρ_i is the areal track density of etched induced fission tracks, Q is a procedural factor involving track registration and observation efficiency ($Q=1$ if track etching and counting procedures are identical), G refers to the initial geometry ratio of the etched surfaces for counting spontaneous and induced tracks ($G=1$ or 0.5 depending on the dating procedure used), I is the isotopic abundance of $^{235}\text{U}/^{238}\text{U}$ (7.2527×10^{-3}), σ_F is the cross section for induced nuclear fission of ^{235}U by thermal neutrons ($580.2 \times 10^{-24} \text{ cm}^2$), and φ is the thermal neutron fluence, measured by the induced FT density on a U-doped glass (e.g., NIST 612), ρ_D , irradiated together with the sample, and is equal to $B\rho_D$, where B is a calibration constant that is empirically determined. The irradiation procedure is used to enable the U concentration to be determined through the known relative abundance of ^{235}U – ^{238}U because only ^{235}U fissions when

irradiated with thermal neutrons. Hence, given these constants, an FT age is obtained by measuring three areal track densities, ρ_s , ρ_i , and ρ_D . Due to the uncertainties in the values of λ_F and σ_F , and the dependency of φ on the nuclear reactor characteristics, especially the degree of thermalization, use of the zeta calibration based on analysis of age standards was recommended by Hurford and Green (1983) and is now widely used. The zeta age calibration factor, ζ , is defined by:

$$\zeta = B(I\sigma_F/\lambda_F) \quad [2]$$

Substituting eqn [2] in eqn [1], and considering $G=1$ and $Q=1$, gives:

$$t = 1/\lambda_D \ln \{1 + \lambda_D(\rho_s/\rho_i)\zeta\rho_D\} \quad [3]$$

Rearranging eqn [3] for a standard of known age (t_{std}) gives:

$$\zeta = \{ \exp(\lambda_D t_{std}) - 1 \} / \{ \lambda_D(\rho_s/\rho_i)_{std} \rho_D \} \quad [4]$$

The value of ζ can then be used in eqn [3] to give the age of a sample.

If the grains within a sample have a common age, the uncertainty on the age is given by (Green, 1981):

$$\sigma(t)/t = \{ 1/N_s + 1/N_i + 1/N_D + [\sigma(\zeta)/\zeta]^2 \}^{1/2} \quad [5]$$

where $\sigma(t)$ is the $\pm 1\sigma$ uncertainty on the age, $\sigma(\zeta)$ is the $\pm 1\sigma$ uncertainty on ζ ; N_s , N_i , and N_D are the total number of tracks counted for ρ_s , ρ_i , and ρ_D , respectively, and the area is determined using an eyepiece graticule. If the area is estimated by use of the point-counting technique, commonly the case with glass shards, an additional experimental error is involved. In this case, the relative standard error (se) on the age (t) is given by (Bigazzi and Galbraith, 1999):

$$se(t)/t \approx \{ 1/Y_s + [1 - (X_s/n_s)]/X_s + 1/Y_i + [1 - (X_i/n_i)]/X_i + 1/Y_D \}^{1/2} \quad [6]$$

where Y_s , Y_i , and Y_D are the total number of spontaneous and induced tracks counted in the sample, and the number of tracks counted in the fluence monitor, respectively. X_s and X_i are the total number of points on glass in determining ρ_s and ρ_i , and n_s and n_i are the corresponding totals for the fields of view.

Prerequisite Conditions for FT Geochronology

The mineral or glass to be dated by the FT method must meet several criteria (Dumitru, 2000; Faure, 1977). The U concentration must be sufficient to give a statistically significant age. Tracks, once formed, must be stable at ambient temperatures and not fully anneal soon after formation. Partial fading of tracks is common in some minerals and glasses, but correction procedures have been developed to account for this condition, particularly in glass. Ideally, the material should be free of microlites and other inclusions, defects, and lattice dislocations, which, upon etching, can resemble fission tracks and make track identification and counting difficult (Storzer and Wagner, 1982). The crystal or glass surface area must be large enough to permit an accurate measurement of the areal track densities. For example, the pumiceous grains in Figure 2(a) cannot be dated, but the poorly vesicular, blocky glass shards in Figure 2(b) can be readily dated by fission tracks. The lower limit in grain size is $\sim 80 \mu\text{m}$. Ideally, U must be distributed uniformly throughout

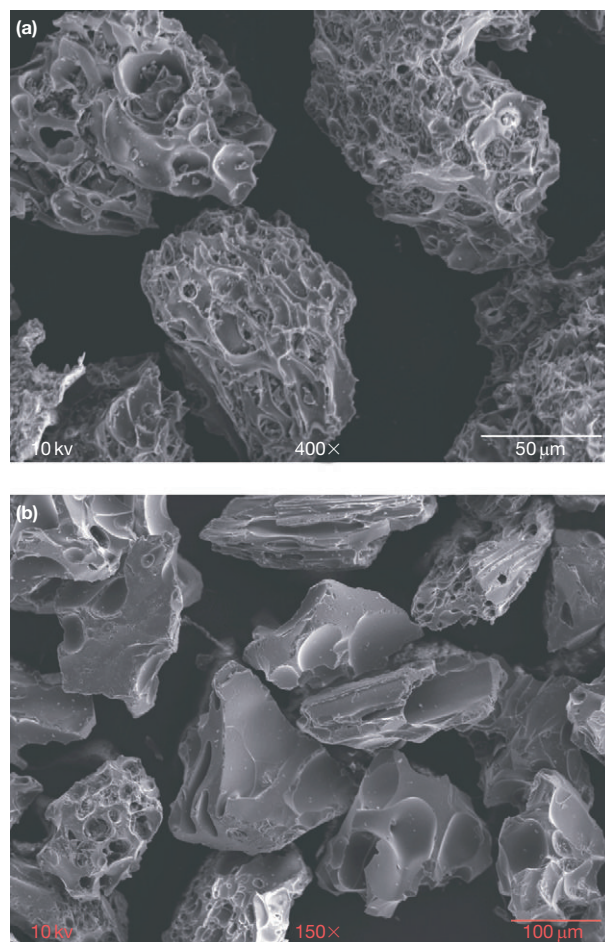


Figure 2 (a) Pumiceous grains from the late Pleistocene Sheep Creek-K tephra, Yukon, Canada. (b) Poorly vesicular, blocky glass shards of the middle Pleistocene Gold Run tephra, Yukon, Canada. Sheep Creek-K tephra cannot be dated by glass FT methods because of insufficient glass surface area. Gold Run tephra, on the other hand, is readily dated by fission tracks (Table 1).

the material. In some materials, this requirement can now be assessed directly by laser ablation inductively coupled plasma-mass spectrometry (LA-ICP-MS) (Pearce et al., 2004). The EDM of FT dating, described below, can compensate for the inhomogeneous U distribution typical of minerals such as zircon. The sample must not have suffered any post-emplacement (depositional) heating that would cause resetting of its fission tracks. The nature of the sample's geologic environment would offer insights on the likelihood of such an event, as does the size distribution of its spontaneous tracks. The mineral or glass must be unaltered and form a primary constituent of the rock or sediment being dated. For example, tephra deposits, commonly dated by the FT method, typically contain foreign grains of xenocrystic, lithic, and detrital origin, which must be avoided during the track-counting operation. This requires use of grain-specific methods for FT counting. As with all geochronologic studies, the stratigraphic, and, in some cases, sedimentologic context of the deposit must be well established. Materials that satisfy these criteria are glass (obsidian, tektites, volcanic glass shards, and glasses made by humans), zircon, apatite, and sphene, although the

latter two are seldom used to date Quaternary events because of low U and rarity.

Glass-FT Methods

General Considerations

One of the earliest successful examples of FT dating involved the use of glass shards to determine the age of Bed 1 at Olduvai Gorge (Fleischer et al., 1965a,b). The FT age of 2.0 ± 0.3 Ma agreed with the K–Ar age of 1.75 ± 0.05 Ma, obtained earlier by Leakey et al. (1961). Silicic, hydrated glass shards are an ideal medium for FT dating. They are the most abundant phase of tephra beds; they are mostly homogeneous and seldom contain microlites or inclusions; they are isotropic with uniform etching characteristics; U contents typically range from 2 to 8 ppm for rhyolitic to dacitic glasses, enough to produce a sufficient number of spontaneous tracks for a statistically useful Quaternary age; rapid cooling during the eruption and subsequent storage at near-surface temperatures mean that acceptable formation ages can be obtained; proven correction methods for PTF exist; and glass shards as small as ~ 80 μm can be dated, opening up the geochronology of regions distant from volcanic centers (especially those of calc–alkaline composition where K-rich minerals are rare) because their parent silicic tephra beds are commonly very widespread. It must be admitted, however, that FT dating late Quaternary events by rhyolitic glass shards is a challenge, although possible, if long counting times are accepted (e.g., Kohn, 1979; 20.3 ± 7.1 ka), or the glass is unusually rich in U (Alloway et al., 2004b; 50 ± 3 ka). Similarly, glass shards of andesitic and basaltic composition commonly have U contents < 1 ppm U and cannot be used to date Quaternary events. Obsidians are much more suitable in this respect because of the large surface areas for counting that are available. Bigazzi and Bonadonna (1973) dated Lipari obsidians in the range of 1.4–11.4 ka, and U-rich archaeological glasses of late historic age have been dated by fission tracks (Wagner and Van den haute, 1992).

Whereas the problem of contamination in obsidian and tektite glasses is negligible, it is an issue when FT dating glass shards and zircon from tephra beds, which are very susceptible to reworking (Figure 3). Grain-specific methods for compositional characterization and FT dating must be used to neutralize this problem. Tight compositional clustering of glass shards, as seen in Old Crow tephra (Figure 4), show little if any contamination, but compositional outliers exist in the Dominion Creek tephra and must be addressed before an accurate FT age can be determined (Figure 4). In the latter case, the contaminating glass shards possessed a notably higher track density, allowing them to be recognized and ignored. Samples that show multiple compositional clustering due to magma-mixing or post-depositional reworking cannot be reliably dated by glass-FT methods. Samples with minor contamination are tractable, however, because glass-FT dating methods are grain-specific in that the track data are accumulated by examination of individual grains so that any foreign glass shards that are encountered during the counting process can be ignored. In such cases, the exotic glass would have to have an age and composition that is very different to that of the

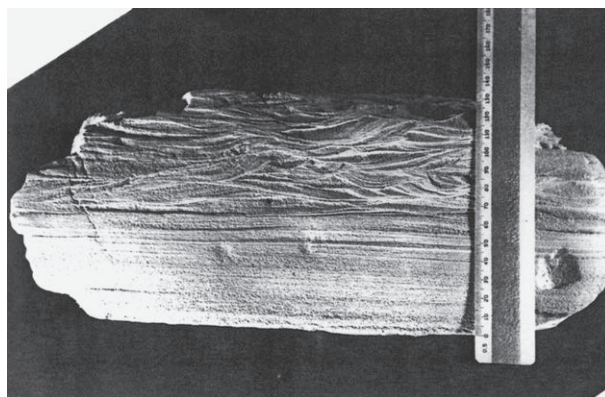


Figure 3 Photograph of Huckleberry Ridge tephra. The lower part of the bed is primary air-fall tephra with horizontal bedding. The upper, cross-bedded part is reworked material washed into the site, and is, therefore, more likely to be contaminated by detrital grains. Scale is mm ruler. Photograph by Glen A. Izett.

host tephra, resulting in a distinctive areal density and size of their fission tracks.

A unimodal major-element composition of glass shards in a sample (e.g., Old Crow tephra, Figure 4) argues for U homogeneity. However, a direct measurement of U homogeneity on a shard-to-shard basis can now be made by LA-ICP-MS (Pearce et al., 2004), and two examples are shown in Figure 5. Uranium contents range from 4 to 12 ppm for both the Lava Creek B and Huckleberry Ridge tephra beds with the average U values obtained by solution ICP-MS and LA-ICP-MS being very similar. In FT dating, U is conventionally measured by irradiating an aliquot of the sample with a specified dose of thermal neutrons that induce fission of a portion of the ^{235}U , which can then be used to determine the amount of ^{238}U in the sample because the ratio of ^{238}U – ^{235}U is a fixed number (I in eqn [1]). Thousands of glass shards are scanned to give ρ_i so that a very good average value of the ^{238}U content is obtained, despite the large range in U values seen in Figure 5. Further, ρ_s is determined under identical conditions so that ρ_s/ρ_i in age eqn [1] can be regarded as accurately determined, and proven to be so by the excellent agreement between corrected glass-FT and $^{40}\text{Ar}/^{39}\text{Ar}$ dating methods for Huckleberry Ridge tephra (1.97 ± 0.22 and 2.003 ± 0.014 Ma, respectively). In a recent development, FT ages of apatite grains are now being determined using U concentrations determined by LA-ICP-MS (Hasebe et al., 2004). Future investigations should explore the viability of this approach for glass shards, obsidians, and tektites, using both average U values from a large number of LA-ICP-MS analyses as well as solution ICP-MS.

There is only one formally accepted glass age standard, Moldavite (Balestrieri et al., 1998). Because the I.U.G.S. Subcommittee on Geochronology specified the use of splits from a single Moldavite (Hurford, 1990), and no supplier is known who can provide a sample large enough for distribution to the FT community, in essence, no formal glass age standard presently exists! Other necessary attributes of a formal age standard are that it contains sufficient tracks to give a statistically meaningful age, and that it has not suffered from PTF. Whether such a glass exists or not is debatable. The candidate that comes closest to these requirements is the Jankov Moldavite. Other potential age

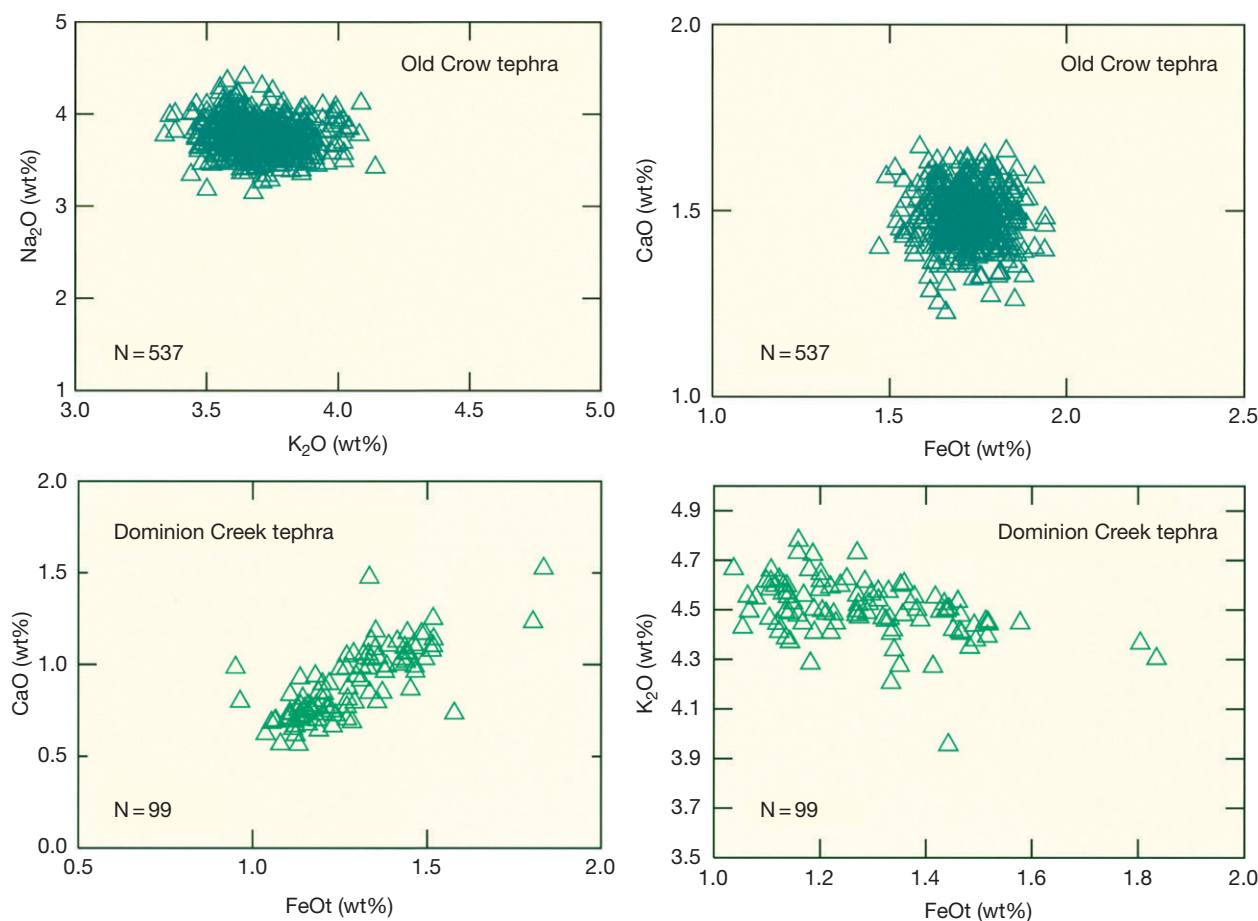


Figure 4 Oxide-variation diagrams showing major-element composition of glass shards from late Pleistocene Old Crow and Dominion Creek tephra beds, Yukon Territory, Canada. Tight clustering in Old Crow tephra demonstrates that contamination is very minor, if present. Several compositional outliers occur in Dominion Creek tephra indicating presence of contamination that must be addressed before an accurate FT age can be obtained. N is the number of individual glass-shard microprobe analyses.

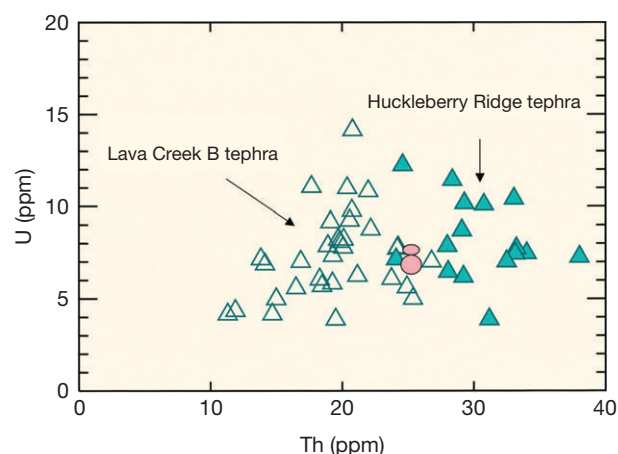


Figure 5 U and Th contents in glass shards from Lava Creek B and Huckleberry Ridge tephra beds, USA, as determined by LA-ICP-MS. Average values for these elements given by solution ICP-MS are shown by pink circle (HRT) and pink oval (LCBT). Average U concentrations by each method are very similar: HRT – 8.2 ppm by LA-ICP-MS, 6.7 ppm by solution ICP-MS; LCBT – corresponding values are 7.4 and 7.5 ppm. Agreement in the average Th values by these two methods is obviously not as good.

standards recommended by Balestrieri et al. (1998) are Roccastrada R2V, Mancusanite, and the Japanese glass JAS-G1. The University of Toronto Laboratory currently uses the Huckleberry Ridge tephra from Meade County, Kansas, USA, to monitor accuracy of its FT age determinations (Table 1). This tephra bed has large glass shards with no inclusions, high U content (Figure 5), an $^{40}\text{Ar}/^{39}\text{Ar}$ age of 2.003 ± 0.014 (2σ) Ma (Ganseccki et al., 1998), and abundant spontaneous fission tracks, making it a useful informal reference glass. PTF has occurred in its glass shards ($D_s/D_i = 0.81 \pm 0.03$ based on five samples; no laboratory heat treatment), but well-established correction procedures are available (see below).

Population FT Method

The glass-population method can be used for homogeneous samples. In the case of glass shards, the coarsest, least vesicular fraction is obtained using sieves combined with magnetic and heavy-liquid techniques. Obsidian and tektite samples are gently crushed and sieved to concentrate the 0.25–0.125 mm fraction. The sample is split into two aliquots, one of which is heated to remove the spontaneous tracks and irradiated. Both aliquots are then mounted on glass slides with epoxy, polished

Table 1 Glass-fission-track age of Gold Run tephra, Klondike goldfields, Yukon Territory, Canada

Sample number	Date irradiated	Spontaneous track density (10^2 t cm^{-2})	Corrected spontaneous track density (10^2 t cm^{-2})	Induced track density (10^5 t cm^{-2})	Track density on muscovite detector over dosimeter glass (10^5 t cm^{-2})	Etching conditions (HF:temp:time; % $^{40}\text{C:s}$)	D_s (μm)	D_i (μm)	D_s/D_i or D_i/D_s	Age (Ma)
<i>Gold Run tephra</i>										
UT1791	21/01/2002	4.09 $\pm$ 0.28 (211)		1.10 $\pm$ 0.01 (12976)	5.15 $\pm$ 0.03 (31171)	24:22:110	5.39 $\pm$ 0.10 [155]	6.24 $\pm$ 0.06 [848]	0.86 $\pm$ 0.02	0.61 $\pm$ 0.06
UT1791 ^a	21/01/2002		4.75 $\pm$ 0.33 (211)	1.10 $\pm$ 0.01 (12976)	5.15 $\pm$ 0.03 (31171)	24:22:110	5.39 $\pm$ 0.10 [155]	6.24 $\pm$ 0.06 [848]	1.16 $\pm$ 0.02 ^b	0.71 $\pm$ 0.07
UT1907 ^c	25/06/2004	3.01 $\pm$ 0.28 (119)		0.60 $\pm$ 0.01 (6780)	4.90 $\pm$ 0.04 (15640)	24:21:200	6.46 $\pm$ 0.18 [94] Weighted mean age	6.81 $\pm$ 0.09 [398] 0.74 $\pm$ 0.06	0.95 $\pm$ 0.03	0.78 $\pm$ 0.09
<i>Internal Standard: Huckleberry Ridge tephra</i>										
UT1366	21/01/2002	27.35 $\pm$ 1.27 (520)		2.21 $\pm$ 0.02 (14722)	5.01 $\pm$ 0.03 (31171)	24:22:160	5.89 $\pm$ 0.10 [212]	6.26 $\pm$ 0.07 [597]	0.94 $\pm$ 0.01	1.97 $\pm$ 0.22

Ages calculated using the zeta approach and $\lambda_D = 1.551 \times 10^{-10} \text{ year}^{-1}$. Zeta value is 318 ± 3 based on six irradiations at the McMaster Nuclear Reactor, Hamilton, Ontario, using the NIST SRM 612 glass dosimeter and the Moldavite tektite glass (Lhenice locality) with an $^{40}\text{Ar}/^{39}\text{Ar}$ plateau age of $15.21 \pm 0.15 \text{ Ma}$ (Staudacher et al., 1982). Standard error ($\pm 1\sigma$) on age estimate is calculated according to Bigazzi and Galbraith (1999). Weighted mean age based on UT1791^a and UT1907^c. Area estimated using the point-counting method (Naeser et al. 1982). D_s = mean spontaneous track diameter, D_i = mean induced track diameter. Number of tracks counted is given in brackets. Number of tracks measured given in square brackets. The single-crystal (sanidine) laser-fusion $^{40}\text{Ar}/^{39}\text{Ar}$ age of Huckleberry Ridge tephra, the internal standard, is $2.003 \pm 0.014 \text{ Ma}$ (2σ error) (Gansecki et al. 1998). The low value of D_s/D_i for UT1366 and its associated small error suggest that the plateau condition was not reached, although the calculated age is very close to its $^{40}\text{Ar}/^{39}\text{Ar}$ age. The population-subtraction method was used. These glass-fit ages have not been corrected according to our new calibration scheme. See notes in Table 4.

^aSample corrected for partial track fading by the track-size (DCFT) method (Sandhu and Westgate, 1995); uncorrected age noted simply by sample number.

^b D_i/D_s value used to correct spontaneous track density.

^cSamples corrected for partial track fading by thermal pretreatment, the ITPFT method (Westgate, 1989); samples heated for 30 days at 150°C .

to expose internal surfaces within the grains, and etched in HF to reveal the tracks. The spontaneous areal track density, ρ_s , is obtained from the unheated shards, and ρ_i from the heated, irradiated shards. Acceptable ages were determined on the Macusanite obsidian using this method (Osorio et al., 2003), the samples being heated for 10 h at 560 °C in order to erase the spontaneous tracks. However, for a number of reasons, this method is not recommended for hydrated glass shards. The differential heat treatment means that track-etching efficiencies for ρ_s and ρ_i are different. Heating above 200 °C drives off water and produces brittle glass shards, which fracture easily when polished, thereby reducing the glass surface area available for counting because of the larger surface area exposed to chemical attack during etching. Crazed surfaces are also produced and hinder track recognition (Naeser et al., 1980).

Population-Subtraction Method

The important distinction of this method is that the two aliquots are not subjected to any heat treatment prior to irradiation. The induced areal track density, ρ_i , is calculated by determining the areal track density of the irradiated sample and then subtracting ρ_s . An example of a FT age determination using this method is shown in Table 1, which also demonstrates the information that should be provided in the documentation of a glass-FT age determination.

Correction Methods for PTF in Glasses

Introduction

When a glass or mineral with latent tracks is held at high temperatures for a sufficient period of time, the disordered structure becomes restored, presumably by a diffusion mechanism. A number of factors influence the stability of latent nuclear tracks in solids, but temperature is by far the most important (Fleischer et al., 1965a,b). As latent tracks are progressively heated, they gradually shorten. For internally distributed tracks, the observed areal track density at a surface depends on the etchable length of the tracks and so this length reduction is accompanied by an areal track density reduction. Thus, if the range of spontaneous fission tracks has been lowered by heating, the calculated age will be younger than the true age. Glasses suffer from PTF, which can occur even at ambient surface temperatures (Naeser et al., 1980), so that it is always necessary to apply a correction for this fading effect before a meaningful FT age can be obtained.

Track-Size Correction Methods

The track-size correction method involves measuring the diameters of spontaneous tracks in the natural sample and comparing them to the diameters of induced tracks in the irradiated aliquot. If the mean track diameter of spontaneous tracks, D_s , is less than the mean diameter of induced tracks, D_i , PTF must have affected the sample (Figure 6). A correction for the lowered areal track density is made by reference to a calibration curve, which shows the correlation between reduction in areal track density, ρ/ρ_o , to reduction in mean track diameter, D/D_o , where ρ_o and D_o are the original areal track density and mean track diameter, respectively. The points defining this curve are derived from a series of

annealing experiments on the irradiated aliquots (Arias et al., 1981; Storzer and Wagner, 1969). Sandhu and Westgate (1995) developed a modified version of the size correction method in which apparent ages of glass shards from tephra beds are corrected for PTF by using a 1:1 relationship between ρ/ρ_o and D/D_o – an approach supported by experimental data. If the sample is etched to give D_i in the range of 6–8 μm , the corrected spontaneous areal track density (ρ_{sc}) is given by:

$$\rho_{sc} = \rho_s(D_i/D_s) \quad [7]$$

The near-linear relationship observed by Sandhu and Westgate (1995) contrasts sharply with the distinctly nonlinear correlation found by Storzer and Wagner (1969; Figure 7). Van den haute's study (1985) explains this difference as due to the different etching conditions used. He theoretically calculated a set of curves showing how ρ/ρ_o varies with D/D_o with progressively longer etch times (i.e., increase in h , the thickness of the glass layer removed by the etchant). At low values of h , the curves are strongly nonlinear, but become near-linear with larger values of h (Figure 7). According to these data, the average track diameter in Storzer and Wagner's study (1969) is $\sim 5 \mu\text{m}$, whereas Sandhu and Westgate (1995) etched their glasses to give a mean track diameter in the range of 6–8 μm .

Plateau Correction Methods

The step-heating plateau method (Storzer and Poupeau, 1973) takes advantage of the higher thermal stability of partially faded spontaneous tracks. The FT age of a sample is proportional to ρ_s/ρ_i . If PTF has occurred, this ratio will be too low. However, when pairs of the natural sample and its irradiated aliquot are heated together for a given time at progressively higher temperatures, ρ_s/ρ_i increases until a plateau level is reached. The value of ρ_s/ρ_i at this plateau level is used to give an FT age corrected for PTF. An example of this approach is given in Arias et al. (1981). An alternative method is to keep a constant temperature and change the heating time (Burchart et al., 1975). An excellent example is the work of Bonetti et al. (1998) on an Australian obsidian (Figure 8). The steady increase in apparent age with continued heating at 140 °C is clearly shown, as is the plateau condition, which is reached after 50 h and indicates an age of 85 Ma. Attainment of the plateau condition is also shown by $D_s/D_i = 1$. A modified version of the plateau technique has been developed by Westgate (1989), who showed that a single heat treatment of 150 °C for 30 days is sufficient to correct fully for PTF. Again, the plateau position is recognized by $D_s = D_i$. In practice, an error is associated with D_s/D_i . For example, taking eight recently determined glass-ITPFT ages, the range in D_s/D_i values is 1.03–0.94 with a mean value of 0.99 ± 0.03 . This procedure has been successfully applied to hydrated, silicic glass shards of many tephra beds, as noted below. Following up on this work, Kitada and Wadatsumi (1995a) showed that by increasing the temperature to 165 °C only a 5-day heat treatment was needed to give a plateau age (Figure 9). Other workers have used different time-temperature combinations to reach the plateau (Table 2). Plateau-annealing conditions should be such as to ensure minimal structural damage to the glass with no deleterious effects on its etching characteristics. Consequently, a

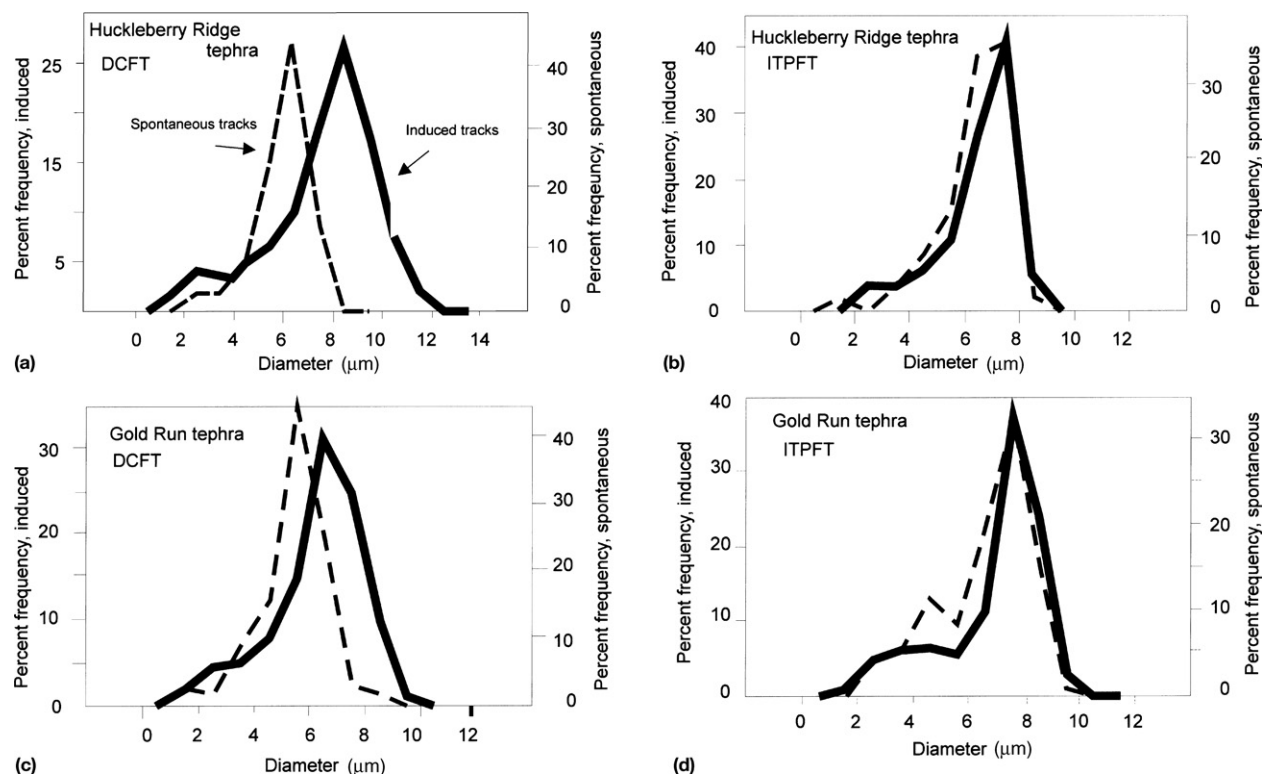


Figure 6 Size frequency curves of FT diameters in glass shards from the Huckleberry Ridge and Gold tephra beds showing partial track fading. DCFT samples unheated; ITPFT samples heated 150 °C for 30 days. (a) Mean track diameter of spontaneous tracks is $5.98 \pm 0.06 \mu\text{m}$ ($n=329$); mean diameter of induced tracks is $7.63 \pm 0.09 \mu\text{m}$ ($n=597$), $D_s/D_i=0.78 \pm 0.01$. Corresponding values for other samples are: (b) $6.41 \pm 0.19 \mu\text{m}$ ($n=50$), $6.41 \pm 0.11 \mu\text{m}$ ($n=199$), 1.00 ± 0.03 ; (c) $5.39 \pm 0.10 \mu\text{m}$ ($n=155$), $6.24 \pm 0.06 \mu\text{m}$ ($n=848$), 0.86 ± 0.02 ; (d) $6.46 \pm 0.18 \mu\text{m}$ ($n=94$), $6.81 \pm 0.09 \mu\text{m}$ ($n=398$), 0.95 ± 0.03 .

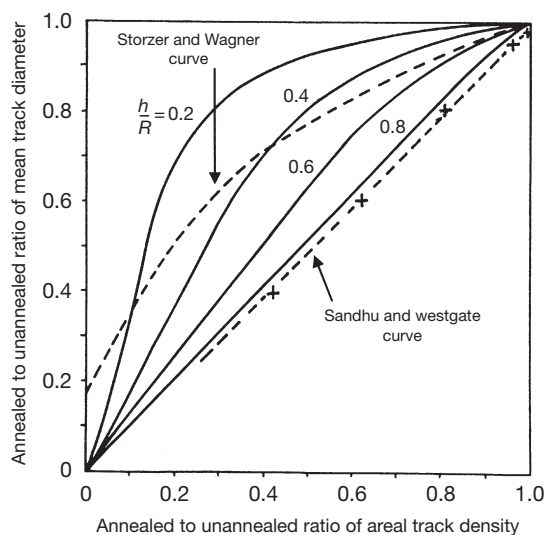


Figure 7 Calculated curves (solid lines) representing the annealed to unannealed ratio of mean track diameter versus areal track density for different etching conditions in an internal glass surface. The thickness of the glass layer removed by the etching is h , and R is the etchable range of track of a single fission fragment. Experimental curves of Storzer and Wagner (1969) and Sandhu and Westgate (1995) are indicated by the dashed lines. Modified from Van den haute P (1985) The density and the diameter of fission tracks in glass with respect to age interpretation. *Nuclear Tracks* 10: 335–348.

low temperature–long time combination is favored, although, in the light of the data shown in Table 2, a heating period of over 15 days at 150 °C is probably excessive.

Comparison of Glass-FT Ages Corrected by the Diameter and Plateau Methods

Glass-FT ages have been determined for many tephra beds at the University of Toronto FT Laboratory using the diameter-correction (glass-DCFT) method of Sandhu and Westgate (1995) and isothermal plateau (glass-ITPFT) method of Westgate (1989; Table 3, Figure 10). Excellent agreement exists, adding confidence to the validity of both methods. Optimal etch times are longer for samples that have been heated, as in the ITPFT method, and limit application of this method to glass shards $\geq 120 \mu\text{m}$. On the other hand, glass shards as small as $\sim 80 \mu\text{m}$ can be dated by the DCFT technique, which requires no heat treatment.

Accuracy and Precision of Glass-FT Ages

The glass-FT ages shown in Table 4 range from 50 to ~ 15 Ma. All are within $\sim 5\%$ of the age determined by other methods, mainly $^{40}\text{Ar}/^{39}\text{Ar}$, and also K–Ar, thermoluminescence, ^{14}C , zircon FT, and orbitally tuned ages (Table 5). In particular, both the DCFT and ITPFT ages on reference glasses mentioned by Balestrieri et al. (1998) agree very well with the $^{40}\text{Ar}/^{39}\text{Ar}$

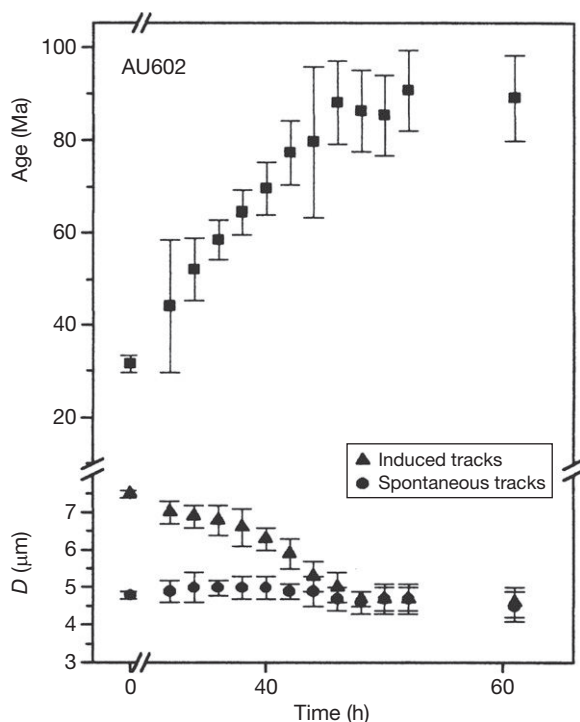


Figure 8 Plateau age for Australian obsidian AU602 obtained by using isothermal treatment at $T = 140^\circ\text{C}$ (upper). Induced and spontaneous FT diameters at each step of thermal treatment (lower). Plateau age of 85 Ma is reached at 50 h when $D_s/D_i = 1$. Reproduced from Bonetti R, Guglielmetti A, Malerba F, Migliorini E, Oddone M, and Bird R (1998) Age determination of obsidian source samples from North Queensland and New South Wales, Australia. In: Van den haute P and De Corte F (eds.) *Advances in Fission-Track Geochronology*, pp. 305–312. Dordrecht: Kluwer. Copyright 1998, with kind permission of Springer Science and Business Media.

results. The glass surface area in all these samples was estimated by the point-counting technique, so that the se on each glass-FT age is calculated from eqn [6] (Bigazzi and Galbraith, 1999). For the most part, the $\pm 1\sigma$ uncertainty is $\leq 10\%$. An improved precision can be achieved if the sample is dated by several persons, a weighted mean and associated error being determined by the standard inverse variance method. It can be seen, therefore, that glass-FT methods can be successfully applied to volcanic glass shards, obsidians, and tektites, giving accurate and precise ages. Of special note, these grain-specific methods are ideally suited to the dating of distal tephra beds, which are typically thin, discontinuous, fine grained, and contaminated by detrital material. They will greatly facilitate the development of detailed chronologies of tephra-bearing sedimentary sequences located far from volcanic centers.

Zircon Fission-Track Method

As with glass, zircon fission-track (ZFT) dating is a grain-specific method in which individual grains are scanned and counted, making it easier to recognize and eliminate contaminating grains (Figure 11). For zircon, an age is obtained on each

grain that is counted. If there is a statistically significant difference in age between the primary and contaminating grains, the latter can be identified by their anomalous ages and eliminated from the analysis (e.g., Galbraith, 1990; Galbraith and Green, 1990; Kowallis, 1988; Seward and Kohn, 1997). Attached glass is a good indicator of a primary zircon (Figure 11), but euhedral zircons are not necessarily primary because they can retain their euhedral form through several erosional cycles.

Zircon has several advantages over glass for FT dating, mainly its significantly higher annealing temperatures. Fission tracks in glass can fade at ambient surface temperatures (Naeser et al., 1980; Seward, 1979; Seward and Kohn, 1997), resulting in an anomalously young age, but zircon does not begin to lose tracks until temperatures of $\sim 160^\circ\text{C}$ or higher, depending on the duration of heating and the α -radiation damage in the zircon (Brandon et al., 1998). Zircon has the added advantage that it can still be dated after glass has altered to clay. On the other hand, it is not present in all tephra beds—silicic tephra beds are more likely to yield usable zircons than mafic tephra deposits. Furthermore, zircons may be too small ($\sim 80\ \mu\text{m}$) to be dated by fission tracks, particularly in tephra beds sampled far from their source vents, and, because zircon is so rare in tephra deposits, much larger samples need to be collected compared to glass. If tephra beds are being sampled from drill cores, glass is probably the only phase that can be dated with fission tracks because of the limited amount of material available.

Unlike glass, zircons from a single source tend to have a very inhomogeneous U distribution, both within and between individual crystals, and therefore U content must be determined from exactly the same area of a crystal in which spontaneous tracks are counted. This is accomplished using the EDM (Naeser and McKee, 1970; Figure 12). In the EDM, spontaneous tracks are etched, normally in a eutectic KOH–NaOH melt (Gleadow et al., 1976), and counted in the image of the crystal recorded in a plastic or low-U ($<10\ \text{ppb}$) muscovite detector that was in direct contact with the crystal during neutron irradiation (Figure 13). An age is obtained for each grain that is analyzed, but experience indicates that a minimum of six grains should be dated to determine a sample age. Seward and Kohn (1997) recommend that for young zircons with low-track densities, at least 20 crystals should be dated. An alternative method of determining U content by LA-ICP-MS is being developed in several laboratories (Cox et al., 2000; Donelick et al., 2005; Hasebe et al., 2004). LA-ICP-MS eliminates the need for neutron irradiation.

Precautions are necessary when dating zircon, particularly from young tephra beds. Care must be taken to ensure that tracks are fully etched because even slight underetching can result in ages that are too young (Gleadow, 1980; Kohn et al., 1992; Naeser et al., 1987; Seward and Kohn, 1997). This is a common pitfall in dating young, low-track-density zircons in which the lack of tracks makes it difficult to judge correct etching time. A solution suggested by Gleadow (1980) is irradiating a separate test mount of zircons from the sample with ^{252}Cf tracks and then etching the two mounts together, using the ^{252}Cf tracks to determine correct etching time. Zircon populations in young samples typically include many grains with no spontaneous tracks. It is critical that induced-track

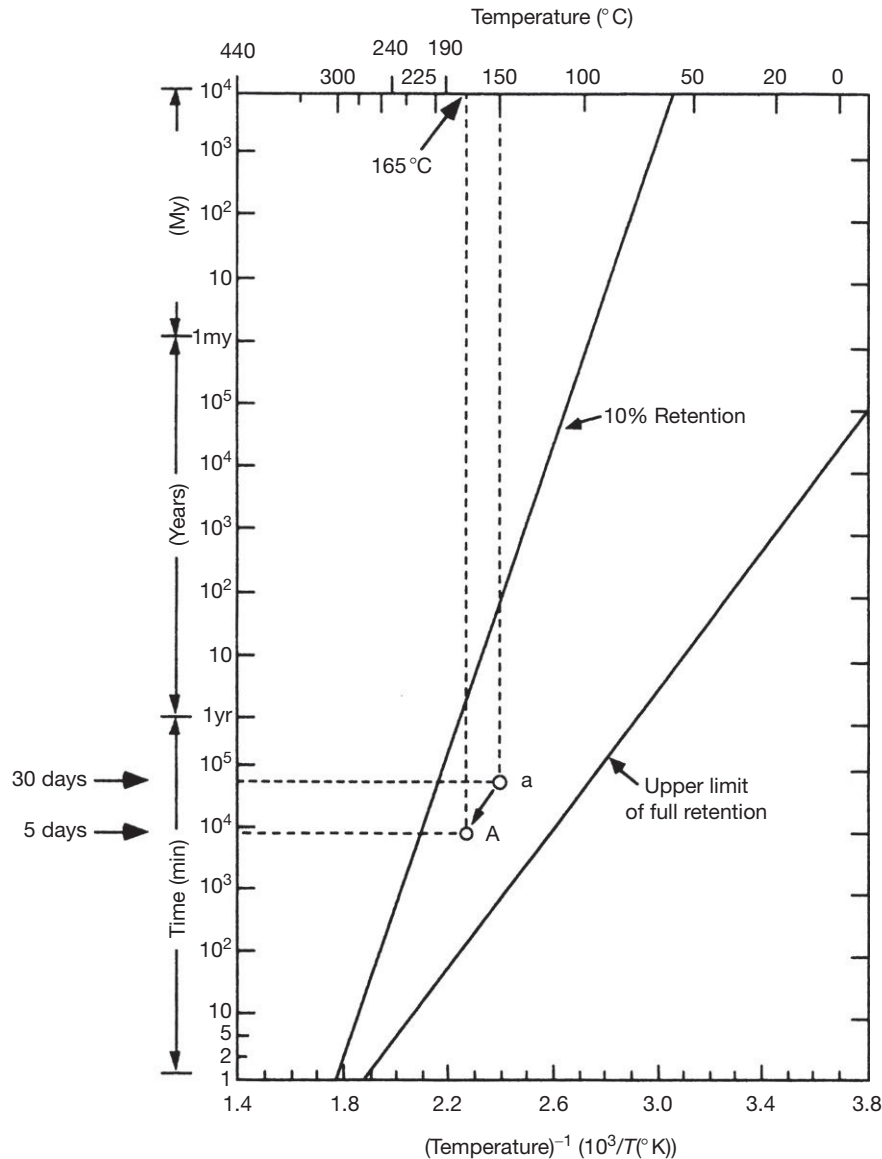


Figure 9 Arrhenius plot of FT annealing experiments for Clearlake glass from Fleischer et al. (1968) showing positions of (a) annealing conditions for glass shards used by Westgate (1989), and (A) annealing conditions used by Kitada and Wadatsumi (1995a). Reproduced from Kitada N and Wadatsumi K (1995a) Annealing properties of natural glass fission tracks at the controlled temperature and its implication on the isothermal plateau dating technique. *Journal of Geosciences, Osaka City University* 38: 4.

counts from these grains be included in the age calculation. Zero-track grains are a real part of low-track-density zircon populations. Failure to include them can result in a calculated age that is significantly older than the true age. A classic illustration is provided by the early attempt to date the ~1.9 Ma KBS Tuff (Gleadow, 1980; Hurford et al., 1976).

The precision of zircon and glass-FT ages is similar, and larger than the typical precision on K–Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ ages, which can be a few percent or less, depending on the material dated and the age of the sample (McDougall, 1995; Table 6 and 7). However, one of the advantages of fission tracks is that many tephra beds that do not contain suitable material for K–Ar or $^{40}\text{Ar}/^{39}\text{Ar}$ dating can be dated by FT using glass and/or zircon.

Applications

New Zealand

Quaternary sequences in New Zealand and in surrounding oceans up to a distance of 1000 km contain abundant silicic tephra beds erupted from the Taupo Volcanic Zone (TVZ) in central North Island (e.g., Shane, 2000). The TVZ is considered to be the most active rhyolitic center on Earth and has been the locus of numerous large ($>100\text{ km}^3$) caldera-forming ignimbrites and plinian eruptions. Dispersal of volcanoclastic material around the TVZ is strongly controlled by prevailing westerly winds, position of fluvial drainage systems, and Mesozoic-aged axial ranges forming topographic obstructions (Figure 14). In New

Table 2 Plateau annealing conditions for glasses

Material	Temperature (°C)	Time	References
Obsidian	250	24 h	Osorio et al. (2003)
Obsidian	220	4 h	Osorio et al. (2003)
Obsidian	250	5 h	Miller and Wagner (1981)
Obsidian	245	24 h	Miller and Wagner (1981)
Obsidian	242	5 h	Miller and Wagner (1981)
Glass shards	165	5–10 days	Kitada and Wadatsumi (1995a)
Obsidian	165	5–10 days	Kitada and Wadatsumi (1995a)
Obsidian	140	50–70 h	Bonetti et al. (1998)
Obsidian	205	1–2 h	Dorighel et al. (1998)
Glass shards	150	30 days	Westgate (1989)
Glass shards	150	15 days	Sandhu et al. (1993)
Glass shards	200	4 h	Bigazzi et al. (2000)
Glass shards	210	8 h	Bigazzi et al. (2000)
Glass shards, obsidian, tektite	150–350	1 h	Naeser et al. (1980)
Glass shards, chilled rim of basalt flows	200	1 h	Macdougall (1976)
Glass shards	150–220	2–3 h	Arias et al. (1981)

In most cases, the plateau condition is recognized by equivalence of the mean spontaneous track diameter (D_s) and mean induced track diameter (D_i).

Table 3 A comparison of glass-DCFT and glass-ITPFT ages

Sample name	Glass-DCFT age	Glass-ITPFT age	References
Gold Run tephra	0.71 ± 0.07	0.78 ± 0.09	This article
Lost Chicken tephra	2.81 ± 0.18	2.90 ± 0.20	Matthews et al. (2003)
Quartz Creek tephra	3.00 ± 0.33	2.93 ± 0.36	Westgate et al. (2003)
Maninjau Ignimbrite	49.5 ± 3.5 ka	51.0 ± 6.0 ka	Alloway et al. (2004a)
Toba tephra	84 ± 17 ka	68 ± 7 ka	Westgate et al. (1998)
Fort Selkirk tephra	1.47 ± 0.14	1.48 ± 0.19	Westgate et al. (2001)
Huckleberry Ridge tephra	2.00 ± 0.21	2.02 ± 0.12	This article
Jankov moldavite	15.10 ± 0.47	15.38 ± 0.52	This article
Roccastrada R2V glass	2.58 ± 0.09	2.50 ± 0.11	This article
Old Crow tephra	0.14 ± 0.03	0.14 ± 0.01	Sandhu and Westgate (1995)
Mamaku Ignimbrite	0.23 ± 0.02	0.23 ± 0.01	Shane et al. (1994)
Ester Ash	0.77 ± 0.12	0.81 ± 0.07	Westgate et al. (1990)
Lake Tapps tephra	0.86 ± 0.11	1.06 ± 0.11	Westgate et al. (1987)
Pakihikura Pumice	1.53 ± 0.22	1.63 ± 0.15	Alloway et al. (1993)
BKT-3 tephra	2.28 ± 0.20	2.05 ± 0.20	Hart et al. (1992)
SHT tephra	3.42 ± 0.29	3.53 ± 0.37	Walter and Aronson (1993)
Cindery Tuff	3.32 ± 0.22	3.70 ± 0.14	Westgate (1989)

All ages in Ma unless otherwise noted. Error given is $\pm 1\sigma$. All ages determined at the University of Toronto FT Laboratory.

Zealand, the predominant tropospheric westerly winds promote dispersal of tephra toward the east coast region of North Island and beyond into the adjacent Pacific Ocean. In sedimentary basins (e.g., Wanganui Basin) located adjacent to the TVZ and outside the main atmospheric dispersal axis, the tephra record is dominated by water-supported mass flow, hyperconcentrated flow, and fluvially transported deposits, which can be considered near-contemporaneous with the proximal eruptive deposits.

In New Zealand, FT dating of tephra and pumice-rich horizons has traditionally played a significant role in advancing the stratigraphy and chronology of Plio–Pleistocene sediments in the southern half of the North Island (e.g., Boellstorff and Te Punga, 1977; Milne, 1973; Seward, 1974, 1976). Zircon and glass are the important phases in this respect, although the use of zircon in some cases is thwarted by low abundance and fine-grain size. On the other hand, hydrated glass shards are typically the most abundant component of tephra beds and commonly the only datable phase present.

The glass-ITPFT technique was first applied in New Zealand to three widespread and stratigraphically important rhyolitic horizons from Wanganui Basin (Alloway et al., 1993). In that study, the glass-ITPFT ages, in conjunction with a revised magnetostratigraphy, demonstrated that the then existing Plio–Pleistocene marine chronology required significant age

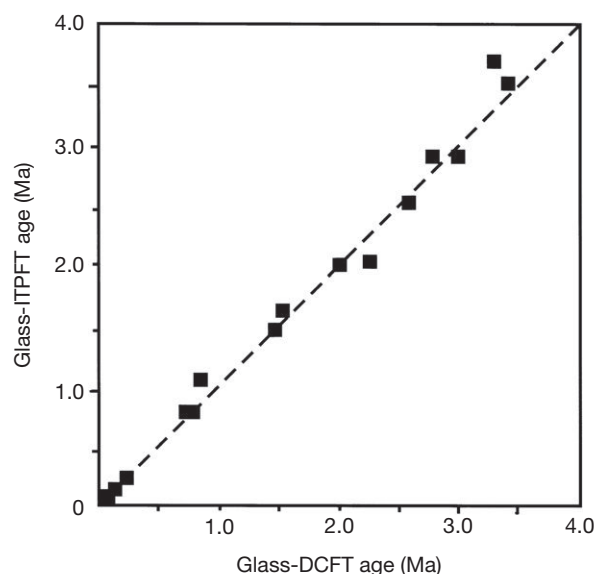


Figure 10 Close correspondence between glass-DCFT and glass-ITPFT ages determined in University of Toronto FT Laboratory. Dashed line is 1:1 line. Regression line through data points has $r^2 = 0.989$.

Table 4 Accuracy and precision of glass-fission-track ages: a comparison with other numerical dating methods

<i>Sample</i>	<i>Material</i>	<i>FT method</i>	<i>Glass-FT age (Ma) ($\pm 1\sigma$)</i>	<i>Age using other methods (Ma)</i>	<i>References</i>
Jankov Moldavite (Czech R.)	Tektite	DCFT	15.10 $\pm$ 0.47	Ar–Ar	Staudacher et al. (1982)
		ITPFT	15.38 $\pm$ 0.52	15.21 $\pm$ 0.15	
Roccastrada R2V glass (Italy)	Glass shards	DCFT	2.58 $\pm$ 0.09	Ar–Ar	Bigazzi et al. (1993)
		ITPFT	2.50 $\pm$ 0.11	2.44 $\pm$ 0.02	
JAS-G1 glass (Japan)	Obsidian	ITPFT (4)	1.01 $\pm$ 0.03	Ar–Ar	Kitada and Wadatsumi (1995b)
				0.947 $\pm$ 0.005	
Huckleberry Ridge tephra (USA)	Glass shards	DCFT	2.00 $\pm$ 0.21	Ar–Ar	Ganseccki et al. (1998)
		ITPFT (3)	2.02 $\pm$ 0.12	2.003 $\pm$ 0.014	
Potaka tephra (New Zealand)	Glass shards	ITPFT (6)	1.00 $\pm$ 0.03	Ar–Ar	Alloway et al. (2004b)
				1.04 $\pm$ 0.03	
Fort Selkirk tephra (Canada)	Glass shards	DCFT	1.47 $\pm$ 0.14	K–Ar	Westgate et al. (2001)
		ITPFT (2)	1.48 $\pm$ 0.19	>1.47 $\pm$ 0.11	
				<1.60 $\pm$ 0.08	
Quartz Creek tephra (Canada)	Glass shards	DCFT	3.00 $\pm$ 0.33	Ar–Ar	Westgate et al. (2003)
		ITPFT	2.93 $\pm$ 0.36	>2.71	
				$\leq$ 3.01	
Ongatiti Ignimbrite (New Zealand)	Glass shards	ITPFT	1.14 $\pm$ 0.06	2.64 $\pm$ 0.24	Alloway et al. (2004c)
				Ar–Ar	
				1.21 $\pm$ 0.04	
Toba tephra (India)	Glass shards	ITPFT (3)	79 $\pm$ 8 ka	Ar–Ar	Westgate et al. (1998)
				74 $\pm$ 2 ka	Oppenheimer (2002)
Maninjau Ignimbrite (Sumatra, Indonesia)	Glass shards	ITPFT	51 $\pm$ 6 ka	^{14}C	Alloway et al. (2004b)
		DCFT (2)		49.5 $\pm$ 3.5 ka	
53 $\pm$ 2 ka					
Old Crow tephra (Canada)	Glass shards	ITPFT (7)	140 $\pm$ 10 ka	TL	Westgate et al. (1990)
				>128 $\pm$ 22 ka	Berger (2003)
				<147 $\pm$ 16 ka	
Lake Tapps tephra (USA)	Glass shards	ITPFT	1.06 $\pm$ 0.11	Ar–Ar	Westgate et al. (1987)
				1.15 $\pm$ 0.01	Hildreth (1996)
Bishop tuff (USA)	Glass shards	ITPFT	0.74 $\pm$ 0.04	Ar–Ar	Sarna-Wojcicki et al. (2000)
				0.759 $\pm$ 0.002	
Cindery Tuff (Ethiopia)	Glass shards	ITPFT	3.70 $\pm$ 0.14	Ar–Ar	Hall et al. (1984)
				3.94 $\pm$ 0.05	
Rangitawa Tephra (New Zealand)	Glass shards	ITPFT (6)	0.35 $\pm$ 0.02	Orbitally tuned	Pillans et al. (1996)
				0.35 $\pm$ 0.01	
Mamaku Ignimbrite (New Zealand)	Glass shards	DCFT	0.23 $\pm$ 0.02	Ar–Ar	Houghton et al. (1995)
		ITPFT	0.23 $\pm$ 0.02	0.22 $\pm$ 0.01	
SHT tephra (Ethiopia)	Glass shards	DCFT	3.42 $\pm$ 0.29	Ar–Ar	Walter and Aronson (1993)
		ITPFT	3.53 $\pm$ 0.37	3.40 $\pm$ 0.03	
Moiti tephra (Ethiopia)	Glass shards	ITPFT	3.89 $\pm$ 0.25	K–Ar	Hart et al. (1992)
				4.1	
BKT-3 tephra (Ethiopia)	Glass shards	ITPFT	2.05 $\pm$ 0.20	ZFT	Hart et al. (1992)
				2.3	

FT ages determined at the University of Toronto FT Laboratory. Number in brackets indicates number of samples dated. All the glass-FT ages in this table are based on a calibration using $\lambda_D = 1.551 \times 10^{-10} \text{ year}^{-1}$, a zeta value of 318 ± 3 based on six irradiations at McMaster Nuclear Reactor, Hamilton, Ontario, the NIST SRM 612 glass dosimeter, and the Moldavite glass with an $^{40}\text{Ar}/^{39}\text{Ar}$ age of $15.21 \pm 0.15 \text{ Ma}$ (Staudacher et al., 1982). We now use a new calibration scheme that is based on a more precise $^{40}\text{Ar}/^{39}\text{Ar}$ age of $14.34 \pm 0.08 \text{ Ma}$ for the Moldavite glass (Laurenzi et al., 2003, 2007) giving a zeta value of 301 ± 3 . Mean age estimates using this new calibration are $\sim 5\%$ younger than the ages noted in this table – that is, within the error due to counting statistics for most of the Pleistocene tephra beds.

revision and confirmed that the previously published uncorrected glass ages were up to 50% too young (Figure 15). As discussed in Seward and Kohn (1997), the glass ages in Seward (1974, 1979) were determined before the effects of annealing on glass were fully recognized, and they demonstrate the extent to which annealing can lower uncorrected glass-FT ages. This revision had important implications for the evolution of

several other sedimentary basins in southern North Island that contained the same tephra beds.

Since that study, the technique has also been used to confirm the age of stratigraphically important rhyolitic horizons (i.e., Potaka Tephra, Shane, 1994; Rangitawa Tephra, Pillans et al., 1996; Waipuru and Vinegar Hill tephra, Naish et al., 1995; Ototoka Tephra, Alloway et al., 2004a), and has

Table 5 Comparison of glass-ITPFT ages for tephra beds in Wanganui Basin, New Zealand, with paleomagnetic and astronomically calibrated age estimates

Tephra	Glass-ITPFT age	Paleomagnetic age	Astronomical age
Kupe	0.63 ± 0.08	<0.78	0.65
Kaukatea	0.86 ± 0.08	0.78–0.99	0.9
Potaka	1.00 ± 0.03	0.99–1.07	0.99
Rewa	1.20 ± 0.14	ca. 1.19	1.19
Mangapipi	1.51 ± 0.16	1.19–1.77	1.54
Pakihikura	1.58 ± 0.08	1.19–1.77	1.60
Table Flat	1.71 ± 0.12	1.19–1.77	1.65
Vinegar Hill	1.75 ± 0.20	1.19–1.77	1.75
Waipur	1.79 ± 0.15	1.77–1.95	1.83

Data from Pillans et al. (2005). All ages in Ma with $\pm 1\sigma$.

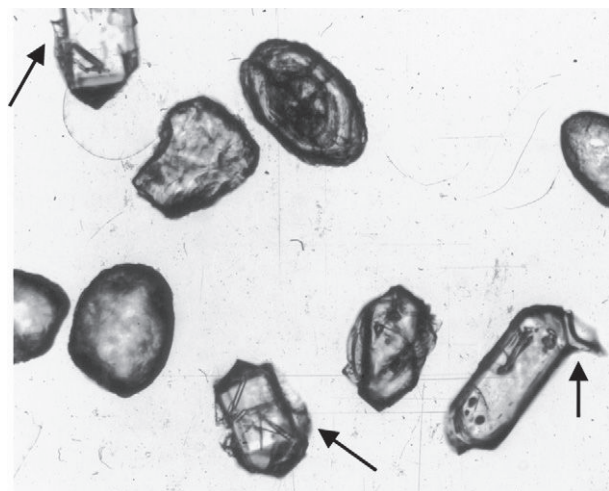


Figure 11 Photomicrograph of zircons separated from a contaminated tephra bed. Primary zircons (arrowed) can be identified by glass adhering to them. Detrital zircons in this tephra are generally darker, more rounded, and lack glass mantles. Zircons are $\sim 150\text{-}\mu\text{m}$ long. Modified from Naeser CW, Briggs ND, Obradovich JD, and Izett GA (1981) *Geochronology of Quaternary tephra deposits*. In: Self S and Sparks RSJ (eds.) *NATO Advanced Studies Institute Series C. Tephra Studies*, pp. 13–47. Dordrecht: Reidel.

been pivotal in establishing a robust Plio–Pleistocene tephrostratigraphic framework throughout North Island, New Zealand (Alloway et al., 2004c; Naish et al., 1998; Pillans et al., 1994, 2005; Shane et al., 1996; Figure 14). More recently, ITPFT ages have been determined on glass shards from eight stratigraphically important macroscopic tephra layers from ODP-core 1124C (Alloway et al., 2005; Carter et al., 2004).

In a careful interlaboratory comparison, Seward and Kohn (1997) reported zircon FT ages for five of the tephra beds in Wanganui Basin, four of which (Pakihikura, Potaka, Vinegar Hill, and Rewa tephtras) are in good statistical agreement with previously determined glass-ITPFT ages. Two of the tephtras (Rewa and Mangapipi) contain two age populations of zircons,

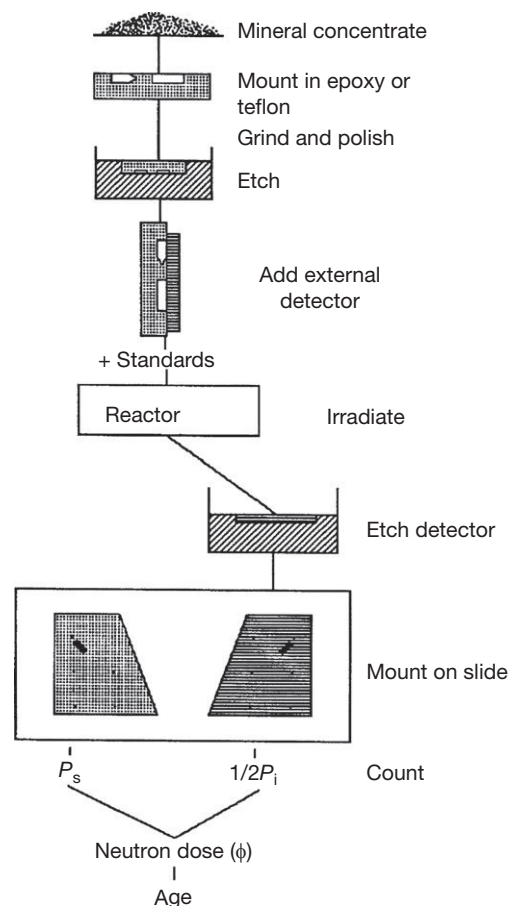


Figure 12 Flow diagram of steps in the external detector method of dating minerals. Details of laboratory procedures are in Naeser (1976). Modified from Naeser CW, Briggs ND, Obradovich JD, and Izett GA (1981) *Geochronology of Quaternary tephra deposits*. In: Self S and Sparks RSJ (eds.) *NATO Advanced Studies Institute Series C. Tephra Studies*, pp. 13–47. Dordrecht: Reidel.

the older interpreted by Seward and Kohn (1997) as detrital contamination, and the younger interpreted as the eruption age. The disagreement in glass and zircon FT ages for the Mangapipi tephra remains to be resolved.

The glass-ITPFT technique is well suited to the dating of distal silicic tephra and has permitted a direct and detailed comparison of stratigraphic and chronologic data in a variety of sedimentary environments in New Zealand. Furthermore, the technique has significantly improved the cross-correlation between tephrochronology, paleomagnetic chronology, and $\delta^{18}\text{O}$ stratigraphy in marine cores.

Indonesia

Introduction

Indonesia is situated adjacent to the Sunda subduction system, which has produced a 1600-km-long volcanic arc with as many as 129 historically active volcanoes, extending from Sumatra in the northwest to Sulawesi in the southeast of the Indonesian

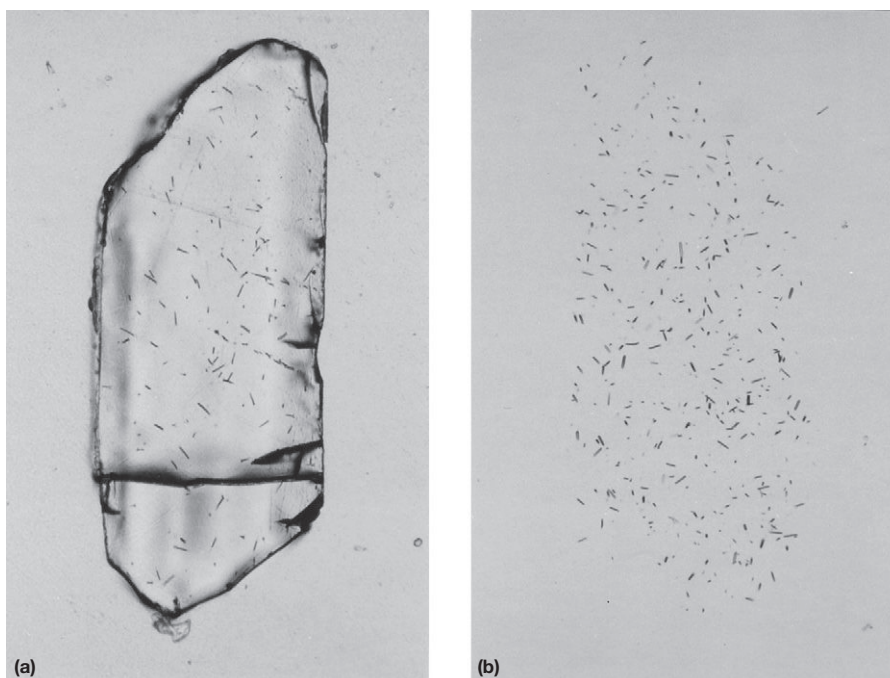


Figure 13 (a) Photomicrograph of crystal (left) and (b) image of the crystal in a muscovite detector (right). In the external detector method, spontaneous fission tracks are counted in the crystal (in this example, an apatite grain), and induced tracks are counted in the detector. Reproduced from Naeser CW and Naeser ND (1989) Laboratory procedures and techniques. In: Crowley KD, Naeser CW, and Naeser ND (eds.) *Geological Society of America Short Course Notes: Fission-Track Analysis: Theory and Applications*, pp. 3-1–3-17. Boulder, CO: Geological Society of America.

archipelago. Sumatra, in particular, contains a great variety and number of volcanoes. Many of these volcanoes are dormant but a few have erupted during historic time (i.e., Marapi and Singgalang–Tandikat). To date, much of the volcanological research conducted in Sumatra has focused on the eruptive history of Toba caldera (e.g., Chesner et al., 1991), and the best record of Indonesian arc volcanism is from tephra layers preserved in ODP-sedimentary cores retrieved from the north-eastern Indian Ocean adjacent to Sumatra (e.g., Dehn et al., 1991; Lee et al., 2004). Many of these tephra layers have yet to be correlated to their onshore eruptive source areas, confirming that there is little known about the eruptive history of many volcanic centers located in Sumatra. In Indonesia, the glass-FT dating technique has been applied to tephra erupted from Toba and Maninjau calderas, situated in northern and west-central Sumatra, respectively (Figure 16).

Toba

The Toba caldera complex consists of four overlapping calderas, the youngest of which is the world's largest caldera (100×30 km) and envelopes the three older calderas. An estimated $2500\text{--}3000$ km³ of dense-rock-equivalent pyroclastic material, termed the Youngest Toba Tephra (YTT), was erupted from the youngest caldera during one of the largest single volcanic eruptions in geological history (Rose and Chesner, 1987). Glass-FT ages for YTT have been determined from sites in Malaysia (68 ± 7 ka; Chesner et al., 1991) and distally in northwest India (84 ± 16 ka; Westgate et al., 1998) and are indistinguishable from an $^{40}\text{Ar}/^{39}\text{Ar}$ age of 73 ± 4 ka (Chesner et al., 1991) and K–Ar ages of 75 ± 12 ka (on biotite) and 74 ± 3 ka (on sanidine) (Ninkovitch et al., 1978). Certainly,

the very widespread distribution of YTT through southern southeast Asia and across the Indian Ocean makes it a very useful marker for dating archaeological sites on the Indian subcontinent (Westgate et al., 1998).

Maninjau

Maninjau is a volcanic edifice situated in the Padang Highlands ~ 300 km to the south of Toba in west-central Sumatra (Figure 16). The central part of this edifice is occupied by a collapsed caldera 20 km long and 8 km wide, and is now occupied by a lake with an estimated volume of 100 km³. The caldera lake parallels the Sumatran volcanic front and is located within the northwest–southeast trending Sumatran fault zone, which runs the entire length of Sumatra. Glass-FT and radiocarbon techniques were used by Alloway et al. (2004b) in an attempt to determine an accurate age for the Maninjau eruptive succession. Glass-ITPFT and -DCFT ages give a weighted mean of 50 ± 3 ka ($n=3$), and, together with concordant ^{14}C ages, give a timing for the latest silicic eruptive activity at 52 ± 3 ka, an age supported by the occurrence of an underlying silicic tephra bed that is geochemically indistinguishable from the ca. 75 ka YTT. This new age for the Maninjau eruptives is significantly different from previous and disparate FT (ca. 70–80 ka, Nishimura et al., 1978) and K–Ar (ca. 0.28 ka, Leo et al., 1980) determinations.

Concluding Statement

We have demonstrated that FT-dating methods applied to volcanic glass and zircon can produce accurate and precise

Table 6 Comparison of representative glass and zircon fission-track and sanidine K-Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ ages of the Pearlette family ash beds and correlative Yellowstone Group tuffs, western North America

Tephra bed/tuff	Pearlette family ash beds		Yellowstone Group tuffs		
	Fission-track glass age ^a (Ma)	Fission-track zircon age (Ma)	Fission-track zircon age (Ma)	K-Ar sanidine age (Ma)	$^{40}\text{Ar}/^{39}\text{Ar}$ sanidine age (Ma)
Lava Creek B	—	0.7 ± 0.1^b	0.6 ± 0.2^b	—	—
	0.60 ± 0.02^c	0.68 ± 0.15^c	—	0.616 ± 0.008^d	—
	0.61 ± 0.12^e	—	—	0.67 ± 0.01^f	0.60 ± 0.01^f
	0.60 ± 0.08^g	—	—	0.63 ± 0.01^f	0.61 ± 0.01^f
	—	—	—	—	0.602 ± 0.002^h
Mesa Falls	—	—	—	—	0.639 ± 0.002^i
	—	—	—	1.27 ± 0.10^d	—
	—	—	—	1.26 ± 0.01^f	1.29 ± 0.01^f
	—	—	—	1.55 ± 0.04^f	1.30 ± 0.01^f
	—	—	—	—	1.293 ± 0.006^h
Huckleberry Ridge	—	1.9 ± 0.1^b	1.9 ± 0.2^b	—	1.285 ± 0.004^i
	1.39 ± 0.08^j	1.93 ± 0.16^j	—	2.02 ± 0.08^d	—
	—	1.91 ± 0.25^e	—	2.01 ± 0.16^f	2.09 ± 0.01^f
	1.3 ± 0.2^k	—	—	2.02 ± 0.03^f	2.04 ± 0.01^f
	1.21 ± 0.07^l	—	—	2.05 ± 0.02^f	2.05 ± 0.01^f
	2.04 ± 0.16^m	—	—	—	2.003 ± 0.006^h
	2.06 ± 0.15^n	—	—	—	2.059 ± 0.004^i
	1.97 ± 0.22^o	—	—	—	—
	—	—	—	—	—

Notes: Analytical uncertainty is ± 1 sigma unless otherwise noted.

Table 1.

^aAges in italics are corrected ages, determined by either ITPFT or DCFT method (see below, and text).

^bNaeser et al. (1973), Wilcox and Naeser (1992): reported zircon age of Lava Creek B tephra ('Pearlette type O') is an average of four age determinations; age of Huckleberry Ridge tephra ('Pearlette type B') is an average of seven determinations; uncertainty is ± 1 standard error of the mean.

^cWestgate et al. (1977): ages determined on correlative Wascana Creek Ash. Glass age is average of five age determinations; uncertainty is ± 1 standard error of the mean.

^dJ.D. Obradovich, in Naeser and Naeser (1988).

^eNaeser et al. (1982): glass age of Lava Creek B tephra determined on correlative Wascana Creek Ash; zircon age is on Huckleberry Ridge tephra ('Pearlette type B').

^fObradovich and Izett (1991).

^gAlloway et al. (1993): 'Lava Creek B ash bed'; ITPFT age.

^hGansecki et al. 1998.

ⁱLanphere et al. (2002): age is weighted mean of total-fusion, plateau, and isochron ages (Lava Creek and Huckleberry Ridge Tuffs), or weighted mean of plateau and isochron ages (Mesa Falls Tuff).

^jSeward (1979): 'Borchers Ash'; reported glass age is an average of four age determinations; uncertainty is ± 1 standard error of the mean.

^kNaeser et al. (1980): 'Pearlette type B'.

^lNaeser and Naeser (1988): 'Huckleberry Ridge ash bed'.

^mWestgate (1989): 'Borchers Ash'; ITPFT age.

ⁿSandhu and Westgate (1995): 'Borchers Ash', DCFT age.

^oTable 1.

Table 7 Comparison of representative glass and zircon fission-track and sanidine K-Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ ages of the Bishop Tuff, California

Fission-track glass age ^a (Ma)	Fission-track zircon age (Ma)	K-Ar sanidine age (Ma)	$^{40}\text{Ar}/^{39}\text{Ar}$ sanidine age (Ma)
0.56 ± 0.05^b	0.74 ± 0.03^c	0.727 ± 0.015^d	—
0.72 ± 0.05^e	—	0.741 ± 0.007^f	0.734 ± 0.012^f
0.73 ± 0.03^g	—	0.722 ± 0.007^h	0.757 ± 0.009^i
—	—	0.738 ± 0.003^h	0.764 ± 0.005^j

Analytical uncertainty is ± 1 sigma.

^aAges in italics are corrected by the ITPFT method (see text).

^bNaeser and Naeser (1988).

^cIzett and Naeser (1976).

^dDalrymple et al. (1965), Mankinen and Dalrymple (1979).

^eWestgate (1989).

^fHurford and Hammerschmidt (1985).

^gBishop ash bed, south-east Idaho: weighted mean of two ITPFT ages, 0.72 ± 0.05 Ma (determined by B.V. Alloway) and 0.74 ± 0.04 (J.A. Westgate).

^hIzett (1982), Izett et al. (1988): Izett et al. (1988) also report a K-Ar isochron age of 0.74 Ma for sanidine, plagioclase, biotite, and glass from the Bishop Tuff.

ⁱIzett and Obradovich (1994).

ages of Quaternary events in regions affected by Quaternary silicic volcanism. Although precision levels are inferior to those achieved by the $^{40}\text{Ar}/^{39}\text{Ar}$ method, FT methods are particularly useful for dating K-poor, calc-alkaline volcanics, and highly vitric tephra beds. Other favorable attributes include the small sample requirement for glass-FT ages and the grain-discrete methodologies used, which neutralize the contamination problem. Importantly, setting up a FT dating laboratory is a comparatively inexpensive operation because of its low-technology and low-maintenance requirements. Accordingly, it is very applicable to volcanic regions of the world where monetary resources are limited.

Acknowledgments

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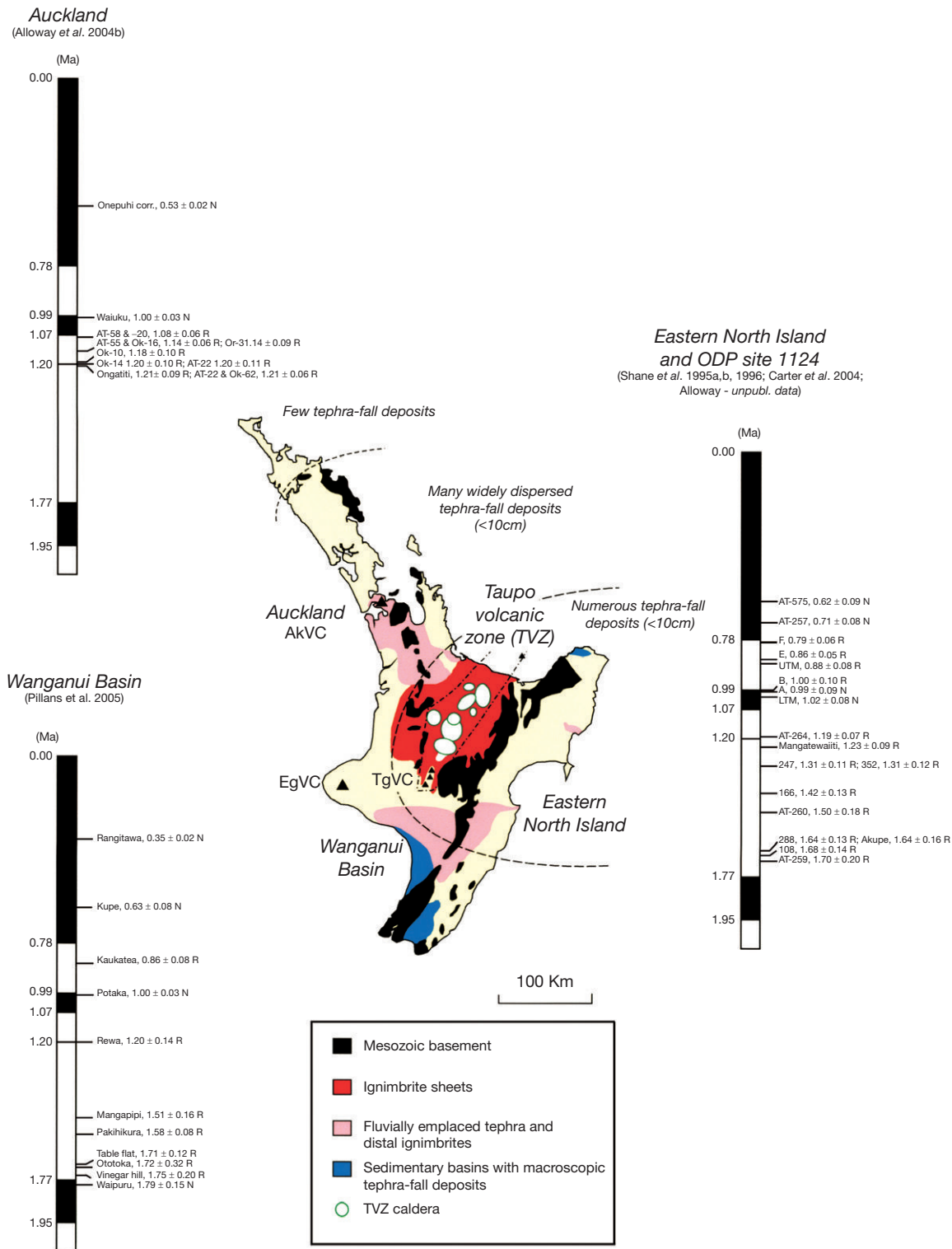


Figure 14 Distribution of Quaternary rhyolitic pyroclastic and associated reworked deposits in North Island, New Zealand and summaries of key Quaternary tephra beds outside of the TVZ that have published glass-ITPFT age control. Paleomagnetic polarity shown as normal (N) or reversed (R) where known. AkVC, Auckland Volcanic Centre; EgVC, Egmont Volcanic Centre; TgVC, Tongariro Volcanic Centre. Modified from Shane PA (2000) Tephrochronology: A New Zealand case study. *Earth-Science Reviews* 49: 223–259.

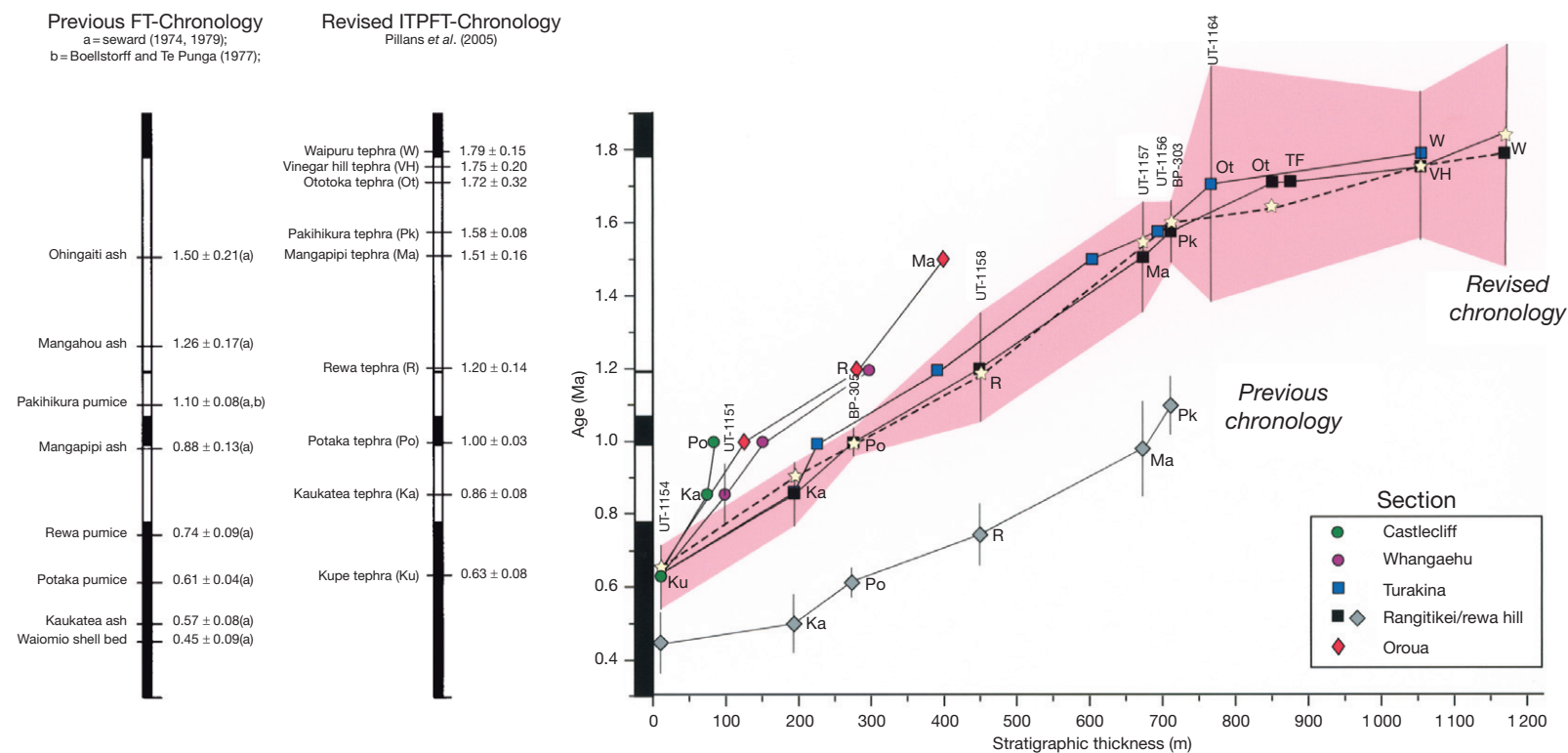


Figure 15 Comparison of previous and revised glass-FT chronologies for middle and lower Pleistocene tephra beds, Wanganui Basin. Shaded area represents $\pm 1\sigma$ on glass-ITPFT dated tephra beds from Rangitikei/Rewa Hill. Stars connected by dashed lines represent astronomically tuned ages determined for the same tephra beds. Reproduced from Pillans BJ, Alloway BV, Naish T, Westgate JA, Abbot S, and Palmer AS (2005) Silicic tephra in Pleistocene shallow marine sediments of Wanganui Basin, New Zealand. *New Zealand Journal of the Royal Society of New Zealand* 35: 43–90.

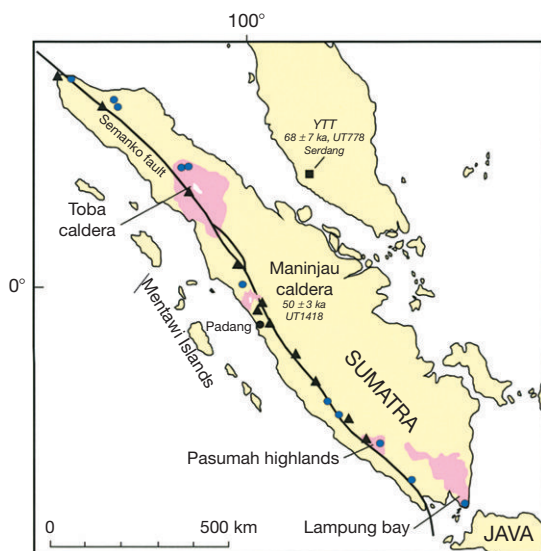


Figure 16 Map of Sumatra showing the position of the Semanko Fault and location of Toba and Maninjau calderas and other volcanic centres. Triangles, volcanoes which have erupted in historic time; circles, volcanoes in fumarolic state. The distribution of silicic volcanoclastic deposits in Sumatra is shown in pink.

National Cooperative Geologic Mapping Program. The manuscript was improved following the critical comments we received from C.W. Naeser (United States Geological Survey).

See also: Dating Techniques; K/Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ Dating. **Luminescence Dating:** Thermoluminescence. **Quaternary Stratigraphy:** Tephrochronology.

References

- Alloway BV, Pillans BJ, Carter L, Naish TR, and Westgate JA (2005) Onshore-offshore correlation of Pleistocene rhyolitic eruptions from New Zealand: Implications for TVZ eruptive history and paleoenvironmental reconstruction. *Quaternary Science Reviews* 24: 1601–1622.
- Alloway BV, Pillans BJ, Naish TR, and Westgate JA (2004a) Age and correlation of Ototoke tephra. Appendix in 'Molluscan biostratigraphy of oxygen isotope stages of the last 2 million years in New Zealand.' Part 1. Revised generic positions and recognition of warm- and cool-water migrants. Beu, A.G. *Journal of the Royal Society of New Zealand* 34: 261–265.
- Alloway BV, Pillans BJ, Sandhu AS, and Westgate JA (1993) Revision of the marine chronology in the Wanganui Basin, New Zealand, based on the isothermal plateau fission-track dating of tephra horizons. *Sedimentary Geology* 82: 299–310.
- Alloway BV, Pribadi A, Westgate JA, et al. (2004b) Correspondence between glass-FT and ^{14}C ages of silicic pyroclastic flow deposits sourced from Maninjau caldera, west-central Sumatra. *Earth and Planetary Science Letters* 227: 121–133.
- Alloway BV, Westgate JA, Pillans BJ, et al. (2004c) Stratigraphy, age and correlation of middle Pleistocene silicic tephra in the Auckland region, New Zealand: A prolific distal record of Taupo Volcanic Zone volcanism. *New Zealand Journal of Geology and Geophysics* 47: 447–479.
- Arias C, Bigazzi G, and Bonadonna FP (1981) Size corrections and plateau age in glass shards. *Nuclear Tracks* 5: 129–136.
- Balestrieri ML, Bigazzi G, Bouska V, et al. (1998) Potential glass age standards for fission-track dating: An overview. In: Van den haute P and De Corte F (eds.) *Advances in Fission-Track Geochronology*, pp. 287–304. Dordrecht: Kluwer.
- Berger GW (2003) Luminescence chronology of late Pleistocene loess-paleosol and tephra sequences near Fairbanks, Alaska. *Quaternary Research* 60: 70–83.
- Bigazzi G (1981) The problem of the decay constant λ_F of ^{238}U . *Nuclear Tracks* 5: 35–44.
- Bigazzi G and Bonadonna F (1973) Fission track dating of the obsidian of Lipari Island (Italy). *Nature* 242: 322–323.
- Bigazzi G, Bonadonna FP, Centamore E, et al. (2000) New radiometric dating of volcanic ash layers in Periadriatic foredeep basin system, Italy. *Palaeogeography, Palaeoclimatology, Palaeoecology* 155: 327–340.
- Bigazzi G, Bonadonna FP, Laurenzi MA, and Tonarini S (1993) A test sample for FT dating of glass shards. *Nuclear Tracks and Radiation Measurements* 21: 489–497.
- Bigazzi G and Galbraith RF (1999) Point-counting technique for fission-track dating of tephra glass shards, and its relative standard error. *Quaternary Research* 51: 67–73.
- Boellstorff JD and Te Punga MT (1977) Fission-track ages and correlation of middle and lower Pleistocene sequences from Nebraska and New Zealand. *New Zealand Journal of Geology and Geophysics* 20: 47–58.
- Bonetti R, Guglielmi A, Malerba F, Migliorini E, Oddone M, and Bird R (1998) Age determination of obsidian source samples from North Queensland and New South Wales, Australia. In: Van den haute P and De Corte F (eds.) *Advances in Fission-Track Geochronology*, pp. 305–312. Dordrecht: Kluwer.
- Brandon MT, Roden-Tice MK, and Garver JI (1998) Late Cenozoic exhumation of the Cascadia accretionary wedge in the Olympic Mountains, northwest Washington State. *Geology* 110: 985–1009.
- Burchart J, Dakowski M, and Galazaka J (1975) A technique to determine extremely high fission track densities. *Bulletin de l'Academie Polonaise des Sciences, Serie des Sciences de la Terre* 23: 1–7.
- Carter L, Alloway B, Shane P, and Westgate JA (2004) Late Cenozoic major rhyolitic eruptions and dispersal – Deep ocean records from off New Zealand. *New Zealand Journal of Geology and Geophysics* 47: 481–500.
- Chadderton LT (2003) Nuclear tracks in solids: Registration physics and the compound spike. *Radiation Measurements* 36: 13–34.
- Chesner CA, Rose WI, Deino A, Drake R, and Westgate JA (1991) Eruptive history of Earth's largest Quaternary caldera (Toba, Indonesia) clarified. *Geology* 19: 200–203.
- Cox R, Kosler J, Sylvester P, and Hodych J (2000) Apatite fission-track (FT) dating by LAM-ICO-MS analysis. *Goldschmidt Journal of Conference Abstracts* 5: 322.
- Dalrymple GB, Cox A, and Doell RR (1965) Potassium-argon age and paleomagnetism of the Bishop Tuff, California. *Geological Society of America Bulletin* 76: 665–674.
- Dehn J, Farrell JW, and Schmincke H-U (1991) Neogene tephrochronology from Site 758 on northern Ninety-east Ridge: Indonesia Arc volcanism of the past 5 Ma. *Proceedings of the Ocean Drilling Programme Scientific Results* 121: 273–295.
- Donelick RA, O'Sullivan PB, and Ketcham RA (2005) Apatite fission-track analysis. In: Reiners PW and Ehlers TA (eds.) *Mineralogical Society of America Reviews in Mineralogy and Geochemistry: Low-Temperature Thermochronology: Techniques, Interpretations, and Applications*, pp. 49–94. Chantilly, VA: The Mineralogical Society of America.
- Doriguel O, Poupeau G, Bellot-Gurlet L, and Labrin E (1998) Fission track dating and provenience of archaeological obsidian artefacts in Colombia and Ecuador. In: Van den haute P and De Corte F (eds.) *Advances in Fission-Track Geochronology*, pp. 313–324. Dordrecht: Kluwer.
- Dumitru TA (2000) Fission-track geochronology. In: Noller JS, Sowers JM, and Lettis WR (eds.) *American Geophysical Union Reference Shelf. Quaternary Geochronology: Methods and Applications*, pp. 131–155. Washington, DC: American Geophysical Union.
- Faure G (1977) *Principles of Isotope Geology*. New York: Wiley.
- Fleischer RL (1998) *Tracks to Innovation: Nuclear Tracks in Science and Technology*. New York: Springer.
- Fleischer RL (2004) Fission tracks in solids – Production mechanisms and natural origins. *Journal of Materials Science* 39: 3901–3911.
- Fleischer RL, Price PB, and Walker RM (1965a) Effects of temperature, pressure and ionization on the formation and stability of fission tracks in minerals and glasses. *Journal of Geophysical Research* 70: 1497–1502.
- Fleischer RL, Price PB, and Walker RM (1975) *Nuclear Tracks in Solids: Principles and Applications*. Berkeley, CA: University of California Press.
- Fleischer RL, Price PB, Walker RM, and Leakey LSB (1965b) Fission track dating of Bed I, Olduvai Gorge. *Science* 149: 383–393.
- Fleischer RL, Viertl JRM, Price PB, and Aumento F (1968) Potassium-argon ages and spreading rates on the Mid-Atlantic ridge at 45° north. *Science* 161: 1338–1342.
- Galbraith RF (1990) The radial plot: Graphical assessment of spread in ages. *Nuclear Tracks and Radiation Measurements* 17: 207–214.
- Galbraith RF and Green PF (1990) Estimating the component ages in a finite mixture. *Nuclear Tracks and Radiation Measurements* 17: 197–206.
- Ganseccki CA, Mahood GA, and McWilliams M (1998) New ages for the climactic eruptions at Yellowstone: Single-crystal $^{40}\text{Ar}/^{39}\text{Ar}$ dating identifies contamination. *Geology* 26: 343–346.

- Gleadow AJW (1980) Fission track age of the KBS Tuff and associated hominid remains in northern Kenya. *Nature* 284: 225–230.
- Gleadow AJW, Hurford AJ, and Quaife RD (1976) Fission track dating of zircon: Improved etching technique. *Earth and Planetary Science Letters* 33: 273–276.
- Green PF (1981) A new look at statistics in fission-track dating. *Nuclear Tracks* 5: 77–86.
- Hall CM, Walter RC, Westgate JA, and York D (1984) Geochronology, stratigraphy and geochemistry of Cindery Tuff in Pliocene hominid-bearing sediments of the Middle Awash, Ethiopia. *Nature* 308: 26–31.
- Hart WK, Walter RC, and WoldeGabriel G (1992) Tephra sources and correlation in Ethiopia: Application of elemental and neodymium isotope data. *Quaternary International* 13(14): 77–86.
- Hasebe N, Barbarand J, Jarvis K, Carter A, and Hurford AJ (2004) Apatite fission-track chronometry using laser ablation ICP-MS. *Chemical Geology* 207: 135–145.
- Hildreth W (1996) Kulshan caldera: A Quaternary subglacial caldera in the North Cascades, Washington. *Geological Society of America Bulletin* 108: 786–793.
- Houghton BF, Wilson CJN, McWilliams MO, et al. (1995) Chronology and dynamics of a large silicic magmatic system: Central Taupo Volcanic Zone, New Zealand. *Geology* 23: 13–16.
- Hurford AJ (1990) Standardization of fission track dating calibration: Recommendation by the fission track working group of the IUGS subcommission on geochronology. *Chemical Geology (Isotope Geoscience Section)* 80: 171–178.
- Hurford AJ, Gleadow AJW, and Naeser CW (1976) Fission-track dating of pumice from the KBS Tuff, East Rudolf, Kenya. *Nature* 263: 738–744.
- Hurford AJ and Green PF (1983) The zeta calibration of fission-track dating. *Isotope Geoscience* 1: 285–317.
- Hurford AJ and Hammerschmidt K (1985) $^{40}\text{Ar}/^{39}\text{Ar}$ and K/Ar dating of the Bishop and Fish Canyon Tuffs: Calibration ages for fission-track dating standards. *Chemical Geology (Isotope Geoscience Section)* 58: 23–32.
- Izett GA (1982) The Bishop ash bed and some older compositionally similar ash beds in California, Nevada, and Utah. *United States Geological Survey Open-File Report* 82-582, 56 pp.
- Izett GA and Naeser CW (1976) Age of the Bishop Tuff of eastern California as determined by the fission-track method. *Geology* 4: 587–590.
- Izett GA and Obradovich JD (1994) $^{40}\text{Ar}/^{39}\text{Ar}$ age constraints for the Jaramillo normal subchron and the Matuyama–Brunhes geomagnetic boundary. *Journal of Geophysical Research* 99(B2): 2925–2934.
- Izett GA, Obradovich JD, and Mehnert HH (1988) The Bishop ash bed (middle Pleistocene) and some older (Pliocene and Pleistocene) chemically and mineralogically similar ash beds in California, Nevada, and Utah. *United States Geological Survey Bulletin* 1675: 37.
- Kitada N and Wadatsumi K (1995a) Annealing properties of natural glass fission tracks at the controlled temperature and its implication on the isothermal plateau dating technique. *Journal of Geosciences, Osaka City University* 38: 1–12.
- Kitada N and Wadatsumi K (1995b) Proposal for a glass age standard 'JAS-G1'. *Bulletin of Liaison and Informations, IUGS Subcommission on Geochronology* 13: 23–27.
- Kohn BP (1979) Identification and significance of a late Pleistocene tephra in Canterbury District, South Island, New Zealand. *Quaternary Research* 11: 78–92.
- Kohn BP, Pillans B, and McGlone MS (1992) Zircon fission track age for middle Pleistocene Rangitawa Tephra, New Zealand: Stratigraphic and paleoclimatic significance. *Palaeogeography, Palaeoclimatology, Palaeoecology* 95: 73–94.
- Kowallis BJ (1988) Fission track dating of contaminated ash beds using age spectra. *Bulletin de Liaison et Informations IGCP Project* 196(7): 53–57.
- Lanphere MA, Champion DE, Christiansen RL, Izett GA, and Obradovich JD (2002) Revised ages for tuffs of the Yellowstone Plateau volcanic field: Assignment of the Huckleberry Ridge tuff to a new geomagnetic polarity event. *Geological Society of America Bulletin* 114: 559–568.
- Laurenzi MA, Balestrieri MI, Bigazzi G, et al. (2007) New constraints on ages of glasses proposed as reference materials for fission-track dating. *Geostandards and Geoanalytical Research* 31: 105–124.
- Laurenzi MA, Bigazzi G, Balestrieri MI, and Bouska V (2003) $^{40}\text{Ar}/^{39}\text{Ar}$ laser probe dating of the central European tektite-producing impact event. *Meteoritics and Planetary Science* 38: 887–893.
- Leakey LSB, Evernden JF, and Curtis GH (1961) Age of Bed I, Olduvai Gorge, Tanganyika. *Nature* 191: 478–479.
- Lee M-Y, Chen C-H, Wei K-Y, Iizuka Y, and Carey S (2004) First Toba supereruption revival. *Geology* 32: 61–64.
- Leo GW, Hedge CE, and Marvin RF (1980) Geochemistry, strontium isotope data and potassium–argon ages of the andesite–rhyolite association in the Pandang area, west Sumatra. *Journal of Volcanology and Geothermal Research* 7: 139–156.
- MacDougall JD (1976) Fission track annealing and correction procedures for oceanic basalt glasses. *Earth and Planetary Science Letters* 30: 19–26.
- Mankinen EA and Dalrymple GB (1979) Revised geomagnetic polarity time scale for the interval 0–5 m.y. B.P. *Journal of Geophysical Research* 84(B2): 615–626.
- Matthews JV Jr., Westgate JA, Ovenden L, Carter DL, and Fouch T (2003) Stratigraphy, fossils, and age of sediments at the upper pit of the Lost Chicken gold mine: New information on the late Pliocene environment of east central Alaska. *Quaternary Research* 60: 9–18.
- McDougall I (1995) Potassium–argon dating in the Pleistocene. In: Rutter NW and Catto NR (eds.) *Dating Methods for Quaternary Deposits*, pp. 1–14. St. John's, NL: Geological Association of Canada.
- Miller DS and Wagner GA (1981) Fission-track ages applied to obsidian artifacts from South America using the plateau-annealing and the track-size age-correction techniques. *Nuclear Tracks* 5: 147–155.
- Milne JDG (1973) Mt. Curl Tephra, a 230,000-year-old marker bed in New Zealand, and its implications for Quaternary chronology. *New Zealand Journal of Geology and Geophysics* 16: 519–532.
- Naeser CW (1976) Fission track dating. *US Geological Survey Open-File Report* 76-190, 65 pp.
- Naeser CW, Briggs ND, Obradovich JD, and Izett GA (1981) Geochronology of Quaternary tephra deposits. In: Self S and Sparks RSJ (eds.) *NATO Advanced Studies Institute Series C: Tephra Studies*, pp. 13–47. Dordrecht: Reidel.
- Naeser CW, Izett GA, and Obradovich JD (1980) Fission-track and K–Ar ages of natural glasses. *United States Geological Survey Bulletin* 1489: 31.
- Naeser CW, Izett GA, and Wilcox RE (1973) Zircon fission-track ages of Pearlette family ash beds in Meade County, Kansas. *Geology* 1: 187–189.
- Naeser CW and McKee EH (1970) Fission-track and K–Ar ages of Tertiary ash-flow tuffs, north-central Nevada. *Geological Society of America Bulletin* 81: 3375–3384.
- Naeser CW and Naeser ND (1988) Fission-track dating of Quaternary events. In: Easterbrook DJ (ed.) *Dating Quaternary Sediments*, pp. 1–11. Boulder, CO: Geological Society of America Special Paper No. 227.
- Naeser CW and Naeser ND (1989) Laboratory procedures and techniques. In: Crowley KD, Naeser CW, and Naeser ND (eds.) *Geological Society of America Short Course Notes: Fission-Track Analysis: Theory and Applications*, pp. 3–13–17. Boulder, CO: Geological Society of America.
- Naeser ND, Westgate JA, Hughes O, and Péwé TL (1982) Fission-track ages of late Cenozoic distal tephra beds in the Yukon Territory and Alaska. *Canadian Journal of Earth Sciences* 19: 2167–2178.
- Naeser ND, Zeitler PK, Naeser CW, and Cervený PF (1987) Provenance studies by fission-track dating of zircon – Etching and counting procedures. *Nuclear Tracks and Radiation Measurements* 13: 121–126.
- Naish TR, Abbott SS, Alloway BV, et al. (1998) Astronomical calibration of a southern Hemisphere Plio–Pleistocene reference section, Wanganui Basin, New Zealand. *Quaternary Science Reviews* 17: 695–710.
- Naish TR, Kamp PJJ, Alloway BV, Pillans BJ, Wilson GS, and Westgate JA (1995) Tephrostratigraphy and integrated chronology for Pliocene–Pleistocene marine cyclothermic strata, Wanganui Basin: Implications for the Pliocene–Pleistocene boundary in New Zealand. *Quaternary International* 34–36: 29–49.
- Ninkovitch D, Shackleton NJ, Abdel-Monem AA, Obradovich JD, and Izett G (1978) K–Ar age of the late Pleistocene eruption of Toba, north Sumatra. *Nature* 276: 574–577.
- Nishimura S, Sasajima S, Hirooka K, Thio KH, and Hehuwat F (1978) *Radiometric Ages of Volcanic Products in Sunda Arc*. Parapat: CCOP/SEATARC Workshop.
- Obradovich JD and Izett GA (1991) $^{40}\text{Ar}/^{39}\text{Ar}$ ages of upper Cenozoic Yellowstone Group tuffs. *Geological Society of America Abstracts with Programs* 23(2): 84.
- Oppenheimer C (2002) Limited global change due to the largest known Quaternary eruption, Toba 74 kyr BP? *Quaternary Science Reviews* 21: 1593–1609.
- Osorio AM, Iunes PJ, Bigazzi G, et al. (2003) Fission-track dating of Macusanite glasses with plateau and size correction methods. *Radiation Measurements* 36: 407–412.
- Pearce NJG, Westgate JA, Perkins WT, and Preece SJ (2004) The application of ICP-MS methods to tephrochronological problems. *Applied Geochemistry* 19: 289–322.
- Pillans BJ, Alloway BV, Naish T, Westgate JA, Abbot S, and Palmer AS (2005) Silicic tephra in Pleistocene shallow marine sediments of Wanganui Basin, New Zealand. *Journal of the Royal Society of New Zealand* 35: 43–90.
- Pillans B, Kohn BP, Berger G, et al. (1996) Multi-method dating comparison for mid-Pleistocene Rangitawa tephra, New Zealand. *Quaternary Science Reviews* 15: 641–653.
- Pillans B, Roberts AP, Wilson GS, and Alloway BV (1994) Magnetostratigraphic, lithostratigraphic and tephrostratigraphic constraints on lower-middle Pleistocene sea level changes. *Earth and Planetary Science Letters* 121: 81–98.
- Price PB and Walker RM (1963) Fossil tracks of charged particles in mica and the age of minerals. *Journal of Geophysical Research* 68: 4847–4862.

- Reiners PW and Ehlers TA (2005) Low-temperature thermochronology: Techniques, interpretations, and applications. *Mineralogical Society of America, Reviews in Mineralogy and Geochemistry* 58: 622.
- Rose WI and Chesner CA (1987) Dispersal of ash in the great Toba eruption, 75 ka. *Geology* 15: 913–917.
- Sandhu AS and Westgate JA (1995) The correlation between reduction in fission-track diameter and areal track density in volcanic glass shards and its application in dating tephra beds. *Earth and Planetary Science Letters* 131: 289–299.
- Sandhu AS, Westgate JA, and Alloway BV (1993) Optimizing the isothermal plateau fission-track dating method for volcanic glass shards. *Nuclear Tracks and Radiation Measurements* 21: 479–488.
- SarnaWojcicki AM, Pringle MS, and Wijbrans J (2000) New Ar-40/Ar-39 age of the Bishop Tuff from multiple sites and sedimentation rate calibration of the Matuyama–Brunhes boundary. *Journal of Geophysical Research – Solid Earth* 105: 21431–21443.
- Seward D (1974) Ages of New Zealand Pleistocene substages by fission-track dating of glass shards from tephra horizons. *Earth and Planetary Science Letters* 24: 242–248.
- Seward D (1976) Tephrostratigraphy of the marine sediments in the Wanganui Basin, New Zealand. *New Zealand Journal of Geology and Geophysics* 19: 9–20.
- Seward D (1979) Comparison of zircon and glass fission-track ages from tephra horizons. *Geology* 7: 479–482.
- Seward D and Kohn BP (1997) New zircon fission-track ages from New Zealand Quaternary tephra: An interlaboratory experiment and recommendations for the determination of young ages. *Chemical Geology* 141: 127–140.
- Shane PA (1994) A widespread, early Pleistocene tephra (Potaka tephra, 1 Ma) in New Zealand: Character, distribution and implications. *New Zealand Journal of Geology and Geophysics* 37: 25–35.
- Shane PA (2000) Tephrochronology: A New Zealand case study. *Earth-Science Reviews* 49: 223–259.
- Shane PA, Alloway BV, Black TS, and Westgate JA (1995a) Isothermal plateau fission-track ages of tephra beds in early–middle Pleistocene marine and terrestrial sequence, Cape Kidnappers, New Zealand. *Quaternary International* 34–36: 49–55.
- Shane PA, Black TM, Alloway BV, and Westgate JA (1996) Early to middle Pleistocene tephrochronology of North Island, New Zealand: Implications for volcanism, tectonism and paleoenvironments. *Geological Society of America Bulletin* 108: 915–925.
- Shane PA, Black T, and Westgate JA (1994) Isothermal plateau fission-track age for a paleomagnetic excursion in the Mamaku Ignimbrite, New Zealand, and implications for late Quaternary stratigraphy. *Geophysical Research Letters* 21: 1695–1698.
- Shane PA, Froggatt P, Black TS, and Westgate JA (1995b) Chronology of Pliocene and Quaternary bioevents and climatic events from fission-track ages on tephra beds, Wairarapa, New Zealand. *Earth and Planetary Science Letters* 130: 141–154.
- Staudacher TH, Jessberger EK, Dominik B, Kirsten T, and Schaeffer OA (1982) $^{40}\text{Ar}/^{39}\text{Ar}$ ages of rocks and glasses from the Nordlinger Ries Crater and the temperature history of impact breccias. *Journal of Geophysics* 51: 1–11.
- Storzer D and Poupeau G (1973) Ages-plateaux de minéraux et verres par la méthode des traces de fission. *Comptes Rendus de l'Académie des Sciences Paris, Série D* 276: 137–139.
- Storzer D and Wagner GA (1969) Correction of thermally lowered fission-track ages of tektites. *Earth and Planetary Science Letters* 5: 463–468.
- Storzer D and Wagner GA (1982) The application of fission track dating in stratigraphy: A critical review. In: Odin GS (ed.) *Numerical Dating in Stratigraphy*, pp. 199–221. New Jersey: Wiley.
- Van den haute P (1985) The density and the diameter of fission tracks in glass with respect to age interpretation. *Nuclear Tracks* 10: 335–348.
- Wagner GA and Van den haute P (1992) *Fission-Track Dating*. Stuttgart: Enke.
- Walter RC and Aronson JL (1993) Age and source of the Sidi Hakoma Tuff, Hadar Formation, Ethiopia. *Journal of Human Evolution* 25: 229–240.
- Westgate JA (1989) Isothermal plateau fission-track ages of hydrated glass shards from silicic tephra beds. *Earth and Planetary Science Letters* 95: 226–234.
- Westgate JA, Christiansen EA, and Boellstorff JD (1977) Wascana Creek Ash (middle Pleistocene) in southern Saskatchewan: Characterization, source, fission-track age, palaeomagnetism and stratigraphic significance. *Canadian Journal of Earth Sciences* 14: 357–374.
- Westgate JA, Easterbrook DJ, Naeser ND, and Carson RJ (1987) Lake Tapps tephra: An early Pleistocene stratigraphic marker in the Puget Lowland, Washington. *Quaternary Research* 28: 340–355.
- Westgate JA, Preece SJ, Froese DG, Walter RC, Sandhu AS, and Schweger CE (2001) Dating Early and Middle (Reid) Pleistocene glaciations in central Yukon by tephrochronology. *Quaternary Research* 56: 335–348.
- Westgate JA, Sandhu AS, Preece SJ, and Froese DG (2003) Age of the gold-bearing White Channel gravel, Klondike district, Yukon. In: Emond DS and Lewis LL (eds.) *Yukon Exploration and Geology 2002*, pp. 241–250. Whitehorse: Yukon Geological Survey.
- Westgate JA, Shane PA, Pearce NJG, et al. (1998) All Toba tephra occurrences across peninsula India belong to the 75 ka eruption. *Quaternary Research* 50: 107–112.
- Westgate JA, Stemper BA, and Péwé TL (1990) A 3 m.y. record of Pliocene–Pleistocene loess in interior Alaska. *Geology* 18: 858–861.
- Wilcox RE and Naeser CW (1992) The Pearlette family ash beds in the Great Plains: Finding their identities and their roots in the Yellowstone Country. In: Westgate JA, Walter RC, and Naeser ND (eds.) *Quaternary International: Tephrochronology: Stratigraphic Applications of Tephra Beds* 13–14: 9–13.
- Young DA (1958) Etching of radiation damage in lithium fluoride. *Nature* 182: 375–377.

FLUVIAL ENVIRONMENTS

Contents

Sediments

Responses to Rapid Environmental Change

Terrace Sequences

Deltaic Environments

Sediments

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Introduction

Rivers are a dominant feature of most landscapes; therefore fluvial sediments are widespread. Fluvial deposits are represented by a continuum of sediment types that range from clay- to gravel-size particles, and include both terrigenous and organic deposits. Fluvial sediments provide an important link between weathering and slope processes in source areas as well as deltaic and coastal processes within depositional basins. Investigations of fluvial sediments are legion; recent syntheses include those by Miall (1996), Blum and Törnqvist (2000), and Bridge (2003). Two additional aspects of fluvial sediments deserve mention. First, these deposits represent economically important reservoirs of oil, natural gas, and water and, include significant coal reserves. Second, fluvial sediments provide critical records of past and present geologic processes and terrestrial environments. This article summarizes the major characteristics of fluvial sediments and discusses key differences between river systems of continental interiors and those associated with subsiding continental margins.

Fluvial Sediments

Fluvial sediments consist predominantly of channel and overbank deposits (Allen, 1965). Channel deposits are dominated by sand and gravel and represent the bed load of rivers. Overbank deposits generally consist primarily of clay, silt, and lesser amounts of very fine to fine sand and represent suspended-load sediments that accumulate on floodplains. Organic sediments such as peat, and chemical sediments such as carbonate and evaporite minerals, are also important constituents in some overbank settings.

Channel Deposits

Channel deposits consist of channel-bar and channel-fill sediments (Bridge, 2003). Channel deposits commonly occur as

isolated lenticular sand bodies (i.e., ribbon sands) or as thick sheet sands associated with channel belts (Figures 1–3). Channel belts are floodplain areas containing active and abandoned channels and channel bars that record the activity of individual channel systems.

Channel-bar deposits

Channel bars typically form in areas where decreasing flow velocities lead to bed-load sedimentation, such as along the inner bend of river meanders (Figure 4(a)). Channel-bar sediments are stratified sands and gravels, which are arranged in sets of large-scale inclined strata (Bridge, 2003; Figure 3). Large-scale inclined strata consist of gently dipping cross strata that reflect episodic migration of major bar forms (i.e., point bars, side bars) in response to channel translation and/or expansion (Figure 4(b)). The base of large-scale inclined strata sets is an erosional bounding surface, which often is overlain by rip-up clasts derived from fine-grained bed and bank materials. Within large-scale inclined strata, smaller-scale sets of cross-stratified sand and/or gravel are present (Figures 3 and 4(c)). These smaller-scale packages of cross-stratified deposits represent bed forms such as dunes and ripples that migrate along the surface of major bars and the bed of channels. Common sedimentary structures associated with these smaller bed forms include small- to medium-scale trough and planar cross-bedding as well as small-scale ripple stratification (Bridge, 2003). Finer-grained sediments such as mud can be present in the upper portion of bar deposits, or occur as mud drapes that separate sets of cross strata.

Channel-fill deposits

Channel-fill deposits are highly variable; sediments range from coarse-grained sand and gravel to fine-grained mud and organic sediments. Specific characteristics of channel fills depend on flow conditions at the time of channel abandonment. In vertical profiles through channel-fill deposits, grain size

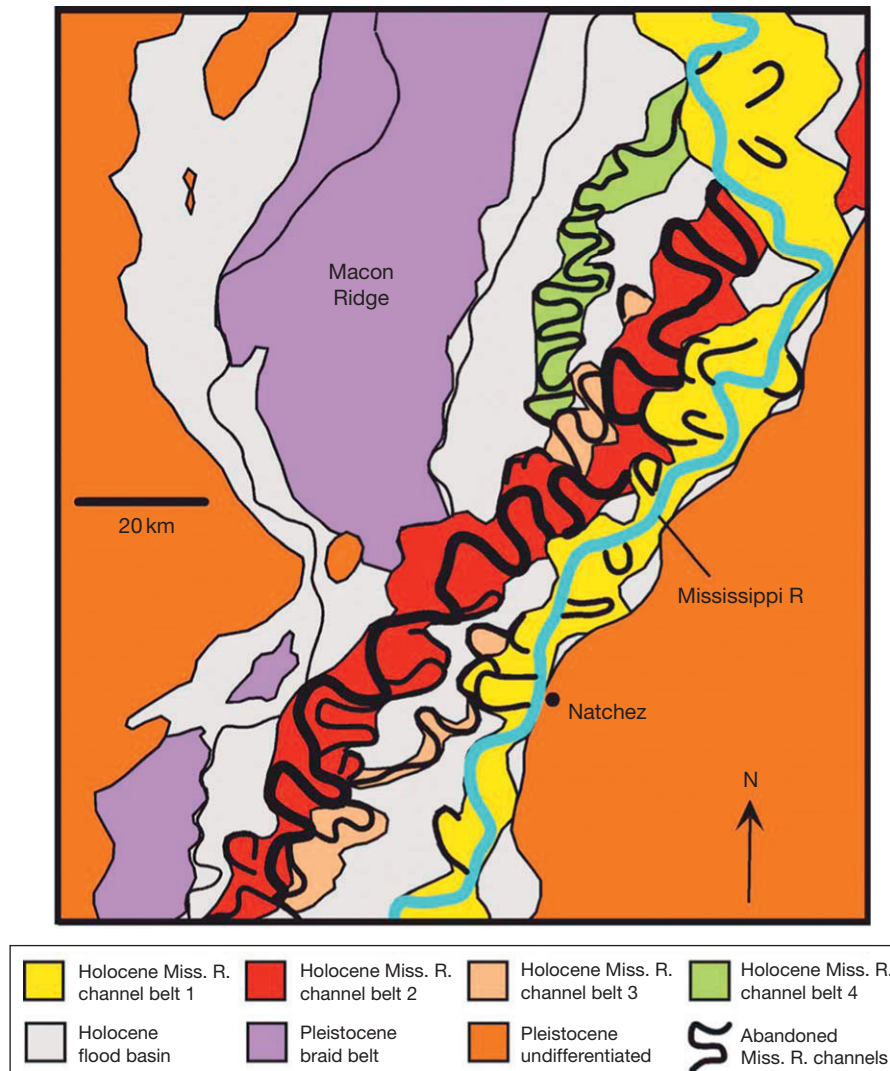


Figure 1 Geologic map of the Mississippi River floodplain near Natchez, Mississippi. Four Holocene Mississippi River channel belts are shown. Modified from Aslan A and Autin WJ (1999) Evolution of the Holocene Mississippi River floodplain, Ferriday, Louisiana: Insights on the origin of fine-grained floodplains. *Journal of Sedimentary Research* 69: 800–815.

generally decreases upwards. Grain size also tends to decrease in a downstream direction. The dominance of coarser-grained bed-load deposits or finer-grained suspended-load sediments depends on whether or not the channel-fill succession represents an active or passive channel fill. Active fills are typically associated with upstream segments of abandoned channels where flow reduction was gradual. In contrast, passive fills are often associated with the central or lower portions of abandoned channels where flow reduction was severe. Active fills are characterized by bed-load deposits that represent the migration of bar forms and smaller-scale dunes and ripples within the upstream portion of the abandoned channel. As flow wanes, active channel-fill deposits transition upward to smaller grain sizes and thinner bed sets representing progressively smaller bars and bed forms. Passive fills are represented by clay, silt, and very fine sand deposited from suspension as well as organic sediment such as peat. These deposits are concentrated in the middle and downstream portions of an

abandoned channel and are associated with rapid decreases in flow such as occurs during neck cutoff and formation of an oxbow lake.

Overbank Deposits

Overbank sediments accumulate during floods in natural-levee, crevasse-splay, and flood-basin environments (Figure 5). Sediment grain size and accumulation rate decreases away from the active channel (Figure 6(a)). Avulsion, the abandonment of all or part of a channel belt in favor of a new course, also plays an important role in the accumulation of overbank deposits (Allen, 1965; Smith et al., 1989). In general, overbank sediments have received considerably less attention than channel deposits (Farrell, 1987). Modern examples of overbank deposits are numerous (Aslan and Autin, 1999; Farrell, 1987; Smith et al., 1989). Classifications of these deposits are presented in Platt and Keller (1992) and Miall (1996). Although

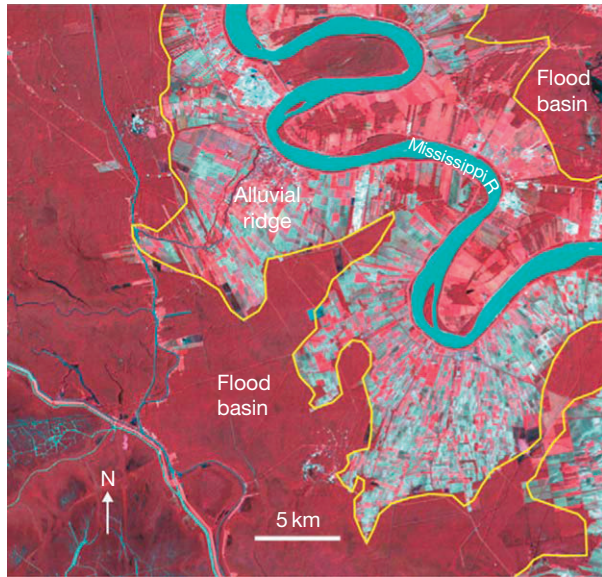


Figure 2 Satellite image of the Mississippi River floodplain south of Baton Rouge, Louisiana. Cultivated areas located adjacent to the river are on natural levees. Flow direction is to the southeast.

overbank deposits are sometimes viewed as monotonous successions of fine-grained alluvium, the recognition of paleosols in ancient overbank sediments has proved useful for subdividing and interpreting these deposits (Kraus, 1999). Additionally, there is increasing recognition that fine-grained sediments in both modern and ancient settings that have traditionally been interpreted as products of repeated overbank flooding, may represent episodes of avulsion (Kraus, 1999; Morozova and Smith, 1999; Slingerland and Smith, 2004; Smith et al., 1989; Stouthamer and Berendsen, 2000).

Natural-levee sediments

Natural levees are asymmetric ridges that slope away from active and abandoned channels toward adjacent flood basins. These features are typically well developed along low-gradient sinuous rivers with large suspended-sediment loads that are subject to deep overbank floods such as the Mississippi (Figure 2). Natural levees can rise up to 5 m above adjacent flood basins and are up to several kilometers wide on floodplains such as the Mississippi (Aslan and Autin, 1999; Törnqvist and Bridge, 2002).

Natural-levee sediments are wedges of sand and silt with minor amounts of clay that are typically thin and fine, away

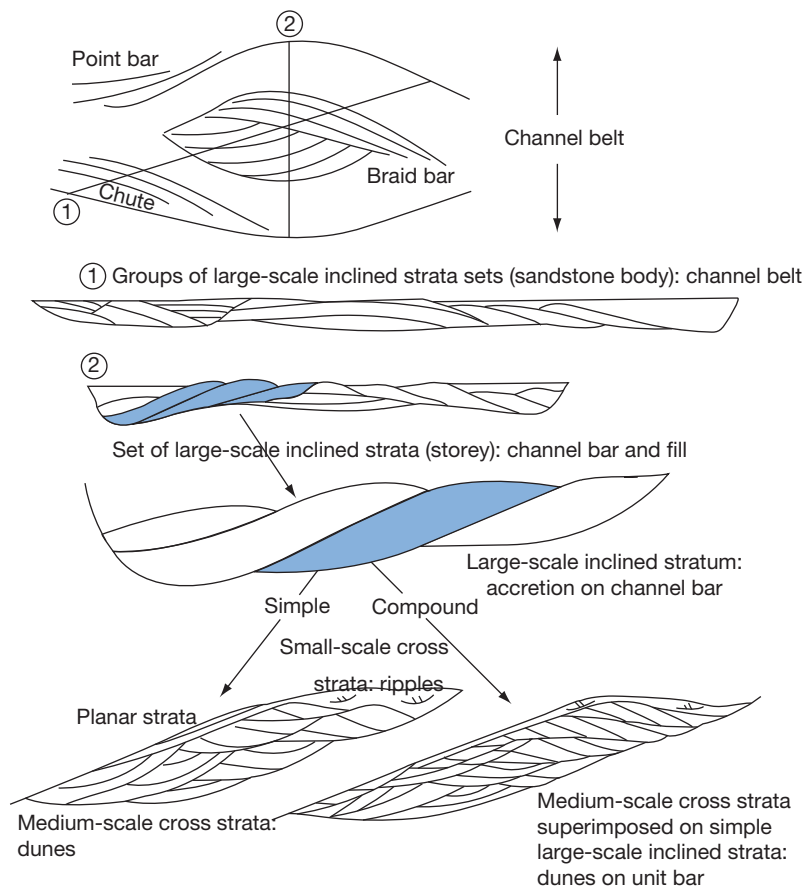


Figure 3 Diagrams showing different scales of fluvial deposits. The upper diagrams show cross-sections through a channel belt composed of multiple depositional units. Large-scale inclined strata (individual examples shown by shaded portions of cross-sections) are the dominant depositional unit of the channel-belt deposits. At the smallest scale, cross strata representing migrating dunes are present within the large-scale inclined strata. Reprinted from Bridge JS (2003) *Rivers and Floodplains, Forms, Processes, and Sedimentary Record*. Oxford: Blackwell Science, with permission from Blackwell Publishing.



Figure 4 (a) Mississippi River and major point bar and chute-channel complex located several kilometers northwest of Vicksburg, Mississippi. The river is approximately 1 km wide. View is upstream. (b) Cross-sectional view through an eroding point bar of the Trinity River in Texas showing a set of large-scale inclined strata. The bank is approximately 6 m tall. Flow direction is to the left. (c) Stacked sets of planar cross-bedded sand representing migrating dunes on a point bar of the Trinity River in Texas. Flow is to the right.



Figure 5 Flooding of the Missouri River in August 1993. Floodwaters have completely inundated the floodplain near Jefferson City, Missouri. Trees on artificial levees outline the sinuous bankfull channel. Areas that are not flooded represent uplands along the margins of the river valley. View is downstream.

from channel margins as a result of decelerating overbank flow (Figure 6(a)). Typical facies are similar to those of upper channel-bar deposits (Bridge, 2003), and consist of horizontal planar sand, small-scale cross-stratified sand, and bioturbated mud (Farrell, 1987). Proximal natural-levee deposits are relatively thick, sandy, and thick bedded. Distal natural-levee sediments are thinner, contain a greater percentage of silt and clay, and show more evidence of bioturbation and pedogenesis.

Natural-levee facies often occur as vertically stacked meter-thick, fining- or coarsening-upward sequences, which reflect periods of episodic overbank flooding and levee abandonment or progradation, respectively.

Crevasse-channel and crevasse-splay sediments

Crevasse channels are breaches in natural levees and transfer water and sediment from mainstem channels to flood basins. Crevasse channels often bifurcate into simple or complex distributary systems that discharge into flood-basin lakes or wetlands (Pérez-Arlucea and Smith, 1999; Smith and Pérez-Arlucea, 1994; Smith et al., 1989). The linear to lobate accumulation of sediment surrounding crevasse channels is referred to as a crevasse splay, and includes the distributary channel network (Figure 7(a)). Where crevasse channels flow into lakes, they form lacustrine deltas. Crevasse channels have natural levees as well as distributary- or terminal-mouth bars (Figure 7(b)). Crevasse-splay channels can extend up to 10 km into flood basins (Smith, 1986; Smith et al., 1989).

Crevasse-channel deposits are typically lenticular bodies of sandy channel-bar and sandy to muddy channel-fill sediments. Grain sizes associated with crevasse-channel deposits are similar to those of the bed-load fraction of the mainstem channel, and are somewhat coarser than other types of overbank sediments. However, muddy fine-grained crevasse-channel fills do occur (Pérez-Arlucea and Smith, 1999; Smith and Pérez-Arlucea, 1994). Stratification types include small- to medium-scale cross bedding and ripple lamination. Evidence of frequent changes in flow stage such as desiccation cracks and bioturbation features may be present (Bridge, 2003). Crevasse-channel deposit thicknesses will generally be smaller than the thickness of the main channel sand bodies.

Crevasse-splay deposits range from tabular to wedge-shaped packages of interbedded sand and silt with lesser amounts of mud. Individual crevasse-splay sand bodies are generally 1–2 m thick. Strata are commonly bioturbated and ripple laminated (Farrell, 1987). Individual splay packages often show upward-coarsening or upward-fining sequences similar to natural-levee deposits (Pérez-Arlucea and Smith, 1999). Upward-coarsening sequences reflect crevasse-splay progradation whereas deposition of overbank silt and clay during waning flood stages produces upward-fining deposits. Crevasse-splay deposits thin or thicken away from channel margins depending on local floodplain topography (Willis and Behrensmeier, 1994). Crevasse-splay sedimentation is

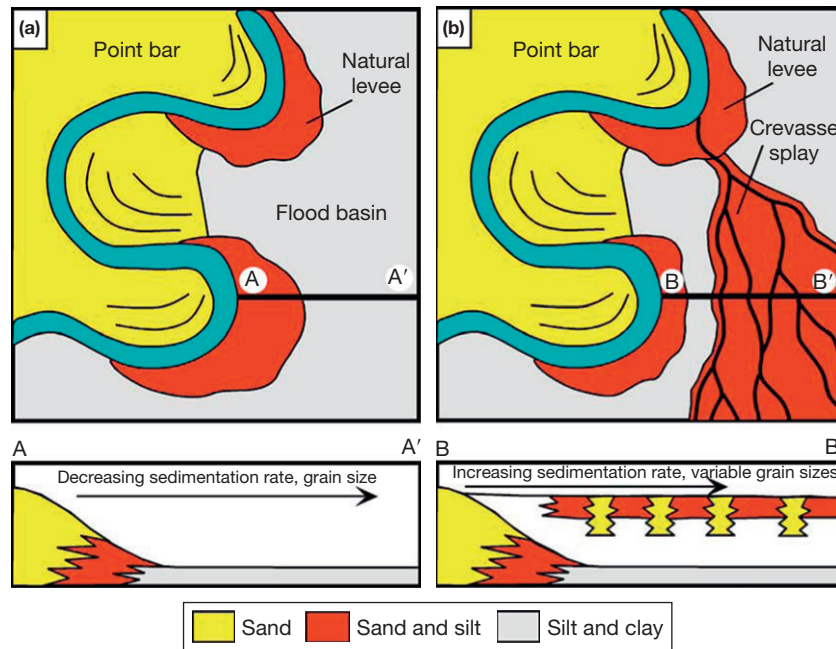


Figure 6 Floodplain maps and cross-sections showing (a) floodplain deposits formed by overbank flooding and sedimentation, and (b) floodplain deposits related to crevassing and avulsion. Note that avulsion deposits thicken away from the active channel belt. Modified from Aslan A and Autin WJ (1999) Evolution of the Holocene Mississippi River floodplain, Ferriday, Louisiana: Insights on the origin of fine-grained floodplains. *Journal of Sedimentary Research* 69: 800–815.

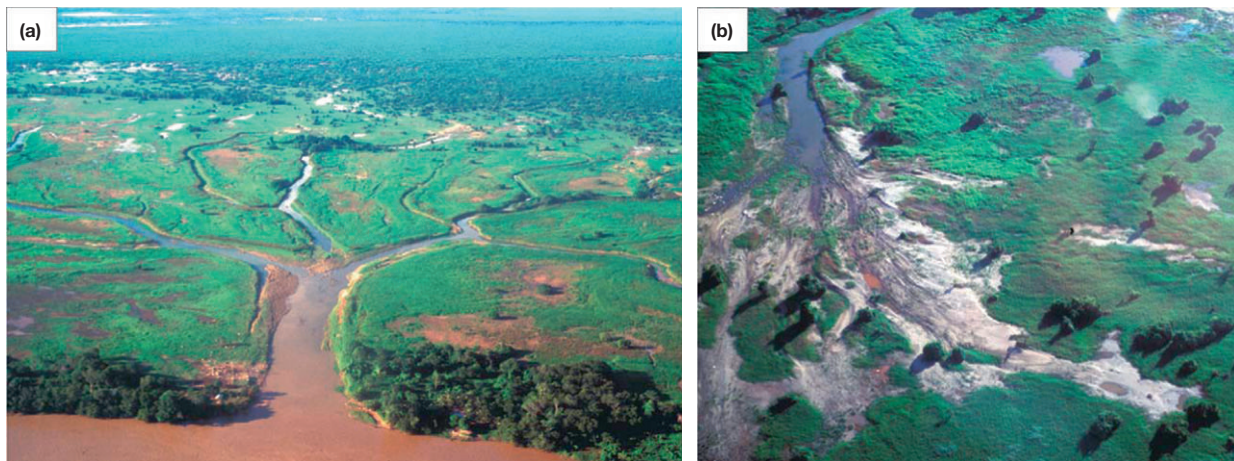


Figure 7 (a) Crevasse splay along Cano Macareo, a major distributary of the Orinoco Delta in Venezuela. Photograph was taken toward the end of the rainy season as water levels began to fall. (b) Sandy distributary mouth-bar deposits at the terminus of one of the crevasse channels shown in (a). The crevasse-splay system discharges into a seasonal wetland.

particularly important within avulsion belts (Smith et al., 1989) and is responsible for rapid fine-grained floodplain aggradation, a key attribute of avulsion.

Flood-basin sediments

Flood basins are broad seasonally flooded depressions that separate channel belts (Figure 8). These areas are characterized by low relief, shallow water tables, and complex drainage channel networks. In humid settings, flood basins are vegetated and contain lakes and perennial or seasonal wetlands. In more arid environments, flood basins are sparsely vegetated

and can have ephemeral lakes. Eolian processes may also be important in arid settings.

Flood-basin deposits are generally tabular accumulations of clay and silt that bury channel deposits or interfinger with sand bodies representing minor floodplain channels and crevasse-splay deposits (Aslan and Autin, 1999; Figures 9 and 10). Flood-basin sediments are often cross-cut by channel-belt sand bodies (Stouthamer and Berendsen, 2000). In Quaternary settings, these deposits are typically meters to tens of meters thick, but can be hundreds of meters thick in pre-Quaternary alluvial successions. Slow sediment accumulation



Figure 8 (a) Seasonal flooding of the Orinoco Delta, Venezuela. Vast tracts of interdistributary flood basins are inundated for several months of each year. During this time, only the crests of natural levees of distributary channels are subaerially exposed. (b) Mud cracks developed in flood-basin silt and clay during the dry season on the Orinoco Delta. This area is completely inundated during the rainy season.



Figure 9 Holocene organic flood-basin mud overlying channel-belt sand on the Mississippi River floodplain near Natchez, Mississippi. Flood-basin mud is approximately 2.5 m thick.

rates in flood basins lead to soil development. Bioturbation features such as root mottles and burrows along with soil nodules and mud cracks are common in these deposits.

Flood-basin sediment characteristics largely depend on climatic conditions. In humid areas with a shallow fluctuating water table (i.e., seasonal wetlands), gray sediment colors and yellow-brown to red color mottles reflect episodes of seasonal wetting and drying. Authigenic minerals such as iron and manganese oxides, calcite, and gypsum can be common in these settings (Aslan and Autin, 1999). If smectitic clays are present, seasonal wetting and drying produces pedogenic slickensides and desiccation cracks. In more perennially saturated flood basins, elevated water levels and low clastic input leads to widespread peat accumulation (Flores, 1981). If clastic input is significant, bioturbated flood-basin sediments will often

interfinger with laminated lacustrine clay and silt. Lacustrine strata are concentrated in topographic depressions where water levels are perennially elevated and reducing conditions prevail. These deposits often contain reducing minerals such as pyrite, vivianite, and jarosite. In more arid settings, flood-basin sediments and playa-lake deposits contain a greater percentage of mud cracks, bioturbation features, and authigenic minerals such as calcite, gypsum, and other evaporite minerals.

Avulsion deposits

Avulsion is common in many low-gradient alluvial river systems and is best known from studies of the Rhine–Meuse delta, the Saskatchewan River, and the Mississippi River and Delta (Aslan et al., 2005; Smith et al., 1989; Stouthamer and Berendsen, 2000). A significant aspect of avulsion is that

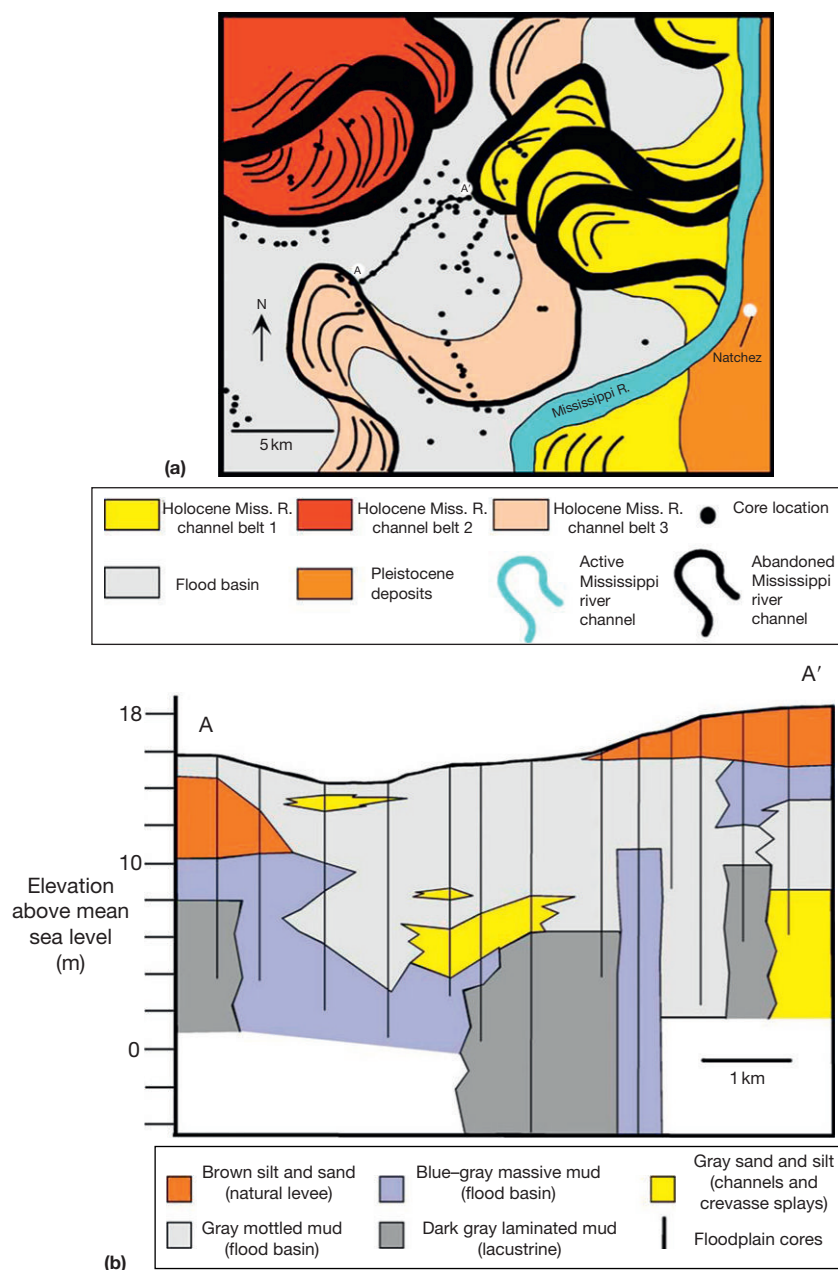


Figure 10 (a) Map of Holocene floodplain deposits of the Mississippi River near Natchez, Mississippi. (b) Cross-section showing typical overbank sediments in the region. Brown silt and sand wedges represent natural-levee deposits. Gray sand and silt represent crevasse-splay sediments (thin sheets to lenticular bodies) whereas the thicker sand accumulations represent floodplain channels. Gray mottled mud and blue-gray massive mud represent flood-basin environments associated with predominantly oxidizing and reducing conditions, respectively. Dark gray laminated mud represents lacustrine strata. Reproduced from Aslan A and Autin WJ (1999) Evolution of the Holocene Mississippi River floodplain, Ferriday, Louisiana: Insights on the origin of fine-grained floodplains. *Journal of Sedimentary Research* 69: 800–815.

this process accounts for large volumes of fine-grained overbank sedimentation on floodplains (Aslan and Autin, 1999; Pérez-Arlucea and Smith, 1999; Smith et al., 1989). During an avulsion, water and sediment escape from the main channel and are transferred to the floodplain through a series of prograding crevasse-splay complexes and/or anastomosing channels (Smith et al., 1989). These channels deposit silt and clay and lesser amounts of sand in local floodplain depressions such as wetlands and lakes. As these topographic depressions fill

with sediment, the locus of floodplain deposition shifts down-valley. Because avulsive deposition occurs at all stages of flow and is not restricted to relatively rare large-magnitude floods, this process results in rapid floodplain aggradation. Texturally, avulsive deposits strongly resemble fine-grained sediments that are typically associated with overbank flooding. However, the stratigraphic architecture of avulsive deposits differs from that of typical overbank deposits (Figure 6(b)). Overbank flooding causes grain size and sediment accumulation rate to

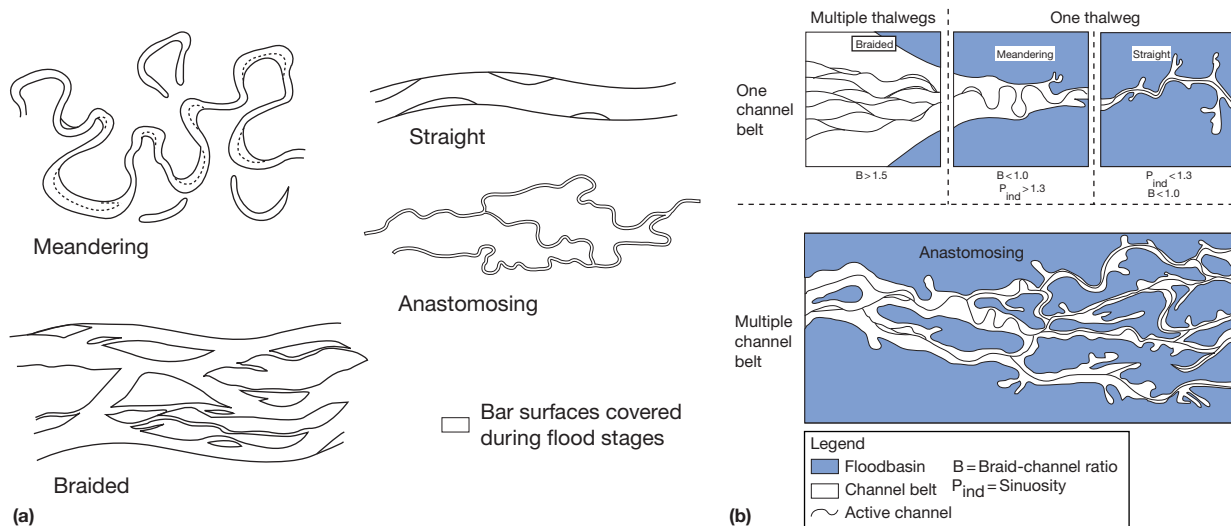


Figure 11 (a) Traditional classification scheme for alluvial channels. Reprinted from Miall AD (1977) A review of the braided river depositional environment. *Earth Science Reviews* 13: 1–62, with permission from Elsevier. (b) An alternative classification of alluvial channels based on the number of channel belts present, braiding characteristics, and channel sinuosity. Reprinted from Makaske B (2001) Anastomosing rivers: A review of their classification, origin, and sedimentary products. *Earth Science Reviews* 53: 149–196, with permission from Elsevier.

decrease away from the active channel. In contrast, avulsion deposit grain sizes do not vary systematically as a function of distance from the active channel. In fact, sedimentation rates can actually increase away from the active channel.

Fluvial Channel Patterns

Fluvial channels have been traditionally subdivided into straight, meandering, and braided (Leopold and Wolman, 1957). More recent classifications recognize anastomosing channels (Miall, 1977, 1996; Rust, 1978; Figure 11(a)). These classifications emphasize two factors: channel sinuosity and bifurcation patterns. For example, braided rivers are characterized by channel bifurcation around bars or islands and low channel sinuosities, whereas anastomosing channels are separated by larger-scale flood basins (Bridge, 2003; Makaske, 2001; Schumm, 1977, 1985). Problems with existing channel classification schemes include measurements of the sinuosity of multichannel rivers, calculations of braiding indices, and delineation of individual channels versus channel belts. Additionally, channel patterns are not mutually exclusive. For example, anastomosing rivers can have braided channel segments (Bridge, 2003). Discrimination of channel patterns using channel characteristics and hydrologic data can be problematic and remains an area of continued research (Lewin and Brewer, 2001; Van den Berg and Bledsoe, 2003). An alternative alluvial river classification scheme proposed by Makaske (2001) subdivides channel systems based on numbers of channel belts, braiding characteristics, and channel sinuosity (Figure 11(b)). Although floodplain classification has received far less attention than channel classification, Nanson and Croke (1992) provide a recent classification of floodplain systems.

Changes in channel patterns are commonly associated with one or more of the following variables (1) discharge, (2) sediment load and type, (3) regional gradient, and (4) bed or

bank materials (Schumm, 1977; 1985). In general, increases in discharge, bed load, and gradient favor the development of low-sinuosity or braided streams. In contrast, decreases in discharge and gradient and increases in suspended-sediment load favor the development of meandering rivers or anastomosing systems. Anastomosing rivers are especially common where regional rates of aggradation are high and bank materials are cohesive (Makaske, 2001; Smith, 1986).

Sedimentary Records of Quaternary Fluvial Systems

Sedimentary records of Quaternary fluvial systems vary dramatically in terms of their abundance, alluvial architecture, and degree of stratigraphic completeness. These differences are demonstrated by comparing fluvial sediments from eroding continental interiors with those associated with subsiding continental margins.

Fluvial Records from Eroding Continental Interiors

Fluvial sediments of continental interiors typically reflect alternating episodes of aggradation, incision, and soil formation, and are located in river valleys. Fluvial sediments in these settings undergo extensive erosion and sediment reworking. Thus, they have a low preservation potential and provide incomplete or discontinuous records of fluvial activity. The erosional nature of fluvial records in continental interiors stems largely from long-term rock and surface uplift driven by erosional isostasy or tectonic processes, which leads to fluvial incision (Pederson et al., 2002). Shorter-term episodes of aggradation and incision associated with glacial-interglacial cycles are superimposed on this long-term uplift and incision. Collectively, these processes lead to the development of flights of down-stepping terraces along many major river valleys of continental interiors (Figure 12).

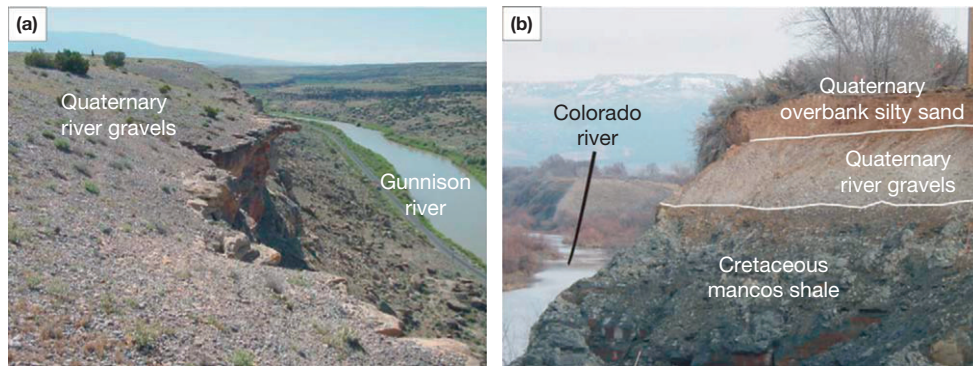


Figure 12 (a) River gravels associated with a Quaternary Gunnison River terrace near Grand Junction, Colorado. The terrace strath is cut into Cretaceous sandstone and is approximately 100 m above the modern Gunnison River. This terrace is part of a flight of down-stepping Quaternary terraces in the region, all of which are associated with abundant gravel. View is upstream. (b) Quaternary river gravels overlain by reddish overbank silty sand deposits of a Quaternary Colorado River terrace near Grand Junction, Colorado. The terrace strath is cut into Cretaceous shale and is approximately 30 m above the modern Colorado River. View is upstream.

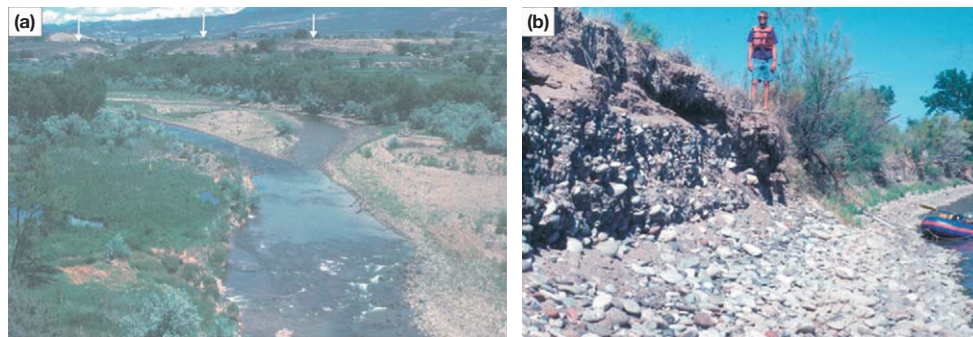


Figure 13 (a) North Fork of the Gunnison River valley near Hotchkiss, Colorado. This river drains volcanic and sedimentary rocks of the central Rocky Mountains, and the channel and floodplain are represented by abundant gravel and sand. Flat benchlands in the background represent river terraces (arrows). View is upstream. (b) Cutbank along the Colorado River near Grand Junction, Colorado showing channel-bar gravel overlain by overbank silt and sand.

Sand and gravel are common constituents of fluvial sediments in continental interiors, and are associated with channel bars and bed-load deposition (**Figure 13(a)**). Overbank sediments are often represented by thin veneers (**Figure 13(b)**). The predominance of coarse-grained sediment reflects the relatively steep slopes of rivers in these settings and low floodplain relief, which combine to inhibit the deposition and preservation of silt and clay. The relative paucity of overbank silt and clay in continental interiors has led some workers to suggest that overbank flooding plays a minor role in the development of floodplains in these settings (**Wolman and Leopold, 1957**).

Fluvial sediments of continental interiors (e.g., Holocene arroyo fills) are often associated with complex alluvial stratigraphies that reflect alternating episodes of aggradation, entrenchment, and soil formation (**Figure 14**). Alluvial successions in the southern Great Plains of the United States provide excellent examples of complex fill histories, and contain regionally correlative stratigraphic units. For example, numerous Holocene valley-fill deposits in this region record aggradation from 5 to 2 ka, soil formation from 2 to 1 ka, and

entrenchment since 1 ka followed by episodic floodplain sedimentation (**Figure 14(b)**). The correlative nature of the alluvial stratigraphic units in these valleys, and similarities in the timing of the major aggradation and incision events suggest that climatic influences and changes in river hydrology have largely controlled the alluvial stratigraphy of the southern Great Plains (**Hall, 1990**) as well as elsewhere in the southwestern United States (**Walters and Haynes, 2001**). Narrow valley widths relative to channel widths cause rivers to rapidly rework older alluvial deposits in these settings. Thus, the length of time represented by the fluvial sediments is limited.

Fluvial Records from Subsiding Continental Margins

Fluvial sediments of continental margins typically accumulate in subsiding sedimentary basins. For example, the Gulf Coast region of the United States represents a subsiding passive margin that consists of stacked successions of fluvial-deltaic Quaternary units (**Figure 15**). Because these sediments are accumulating in a subsiding basin, they have a high preservation potential and the sedimentary records of fluvial activity

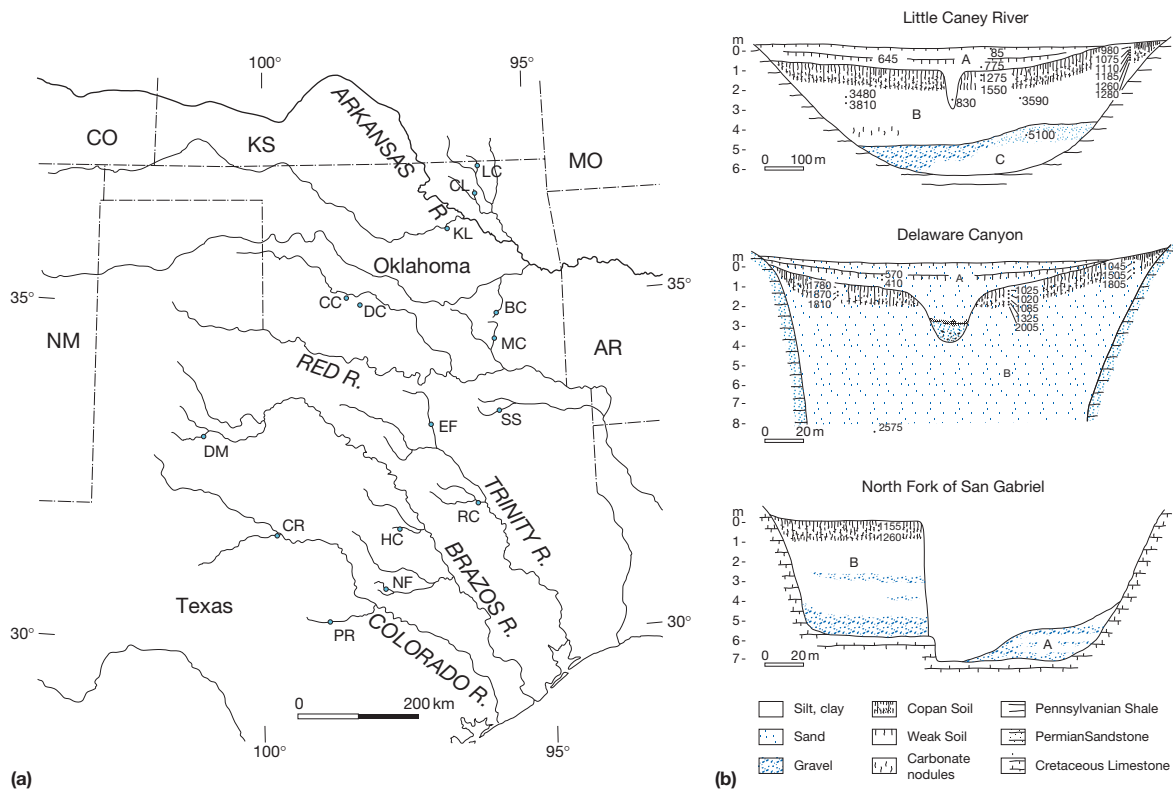


Figure 14 (a) Map showing locations of alluvial rivers that have undergone Holocene filling and entrenchment. BC, Brushy Creek; CC, Carnegie Canyon; CL, Candy Lake; CR, Colorado River; DC, Delaware Canyon; DM, Double Mountain Fork; HC, Hog Creek; EF, Elm Fork of the Trinity River; KL, Keystone Lake; LC, Little Caney River; MC, McGee Creek; NF, North Fork of San Gabriel River; PR, Pedernales River; RC, Richland Creek; SS, South Sulfur Spring. (b) Cross sections showing the Holocene alluvial stratigraphy at three locations in (a). Sand and gravel deposition and Holocene aggradation commenced ~5 ka in all three locations, which produced depositional unit 'B' in the cross sections. From 2 to 1 ka, sedimentation slowed and a cumelic, organic-rich soil (Copan Soil) developed. At 1 ka, channel trenching occurred throughout the southern Great Plains. Channel trenching was followed by episodic floodplain sedimentation, which produced depositional unit 'A' in the cross sections. Ages are in radiocarbon years before present. Premiddle Holocene sediments (depositional unit 'C') are rare in the valley fills. These types of valley-fill fluvial deposits have low long-term preservation potential because of the narrow valley widths and the frequency of channel entrenchment and filling. Reproduced from Hall SA (1990) Channel trenching and climatic change in the southern U.S. Great Plains. *Geology* 18: 342–345.

are more complete than those of continental interiors. Rivers in this region occupy broad alluvial valleys separated by low upland divides. Large floodplain widths relative to channel widths, gentle slopes, and depositional relief created by sedimentation along channels (i.e., elevated alluvial ridges) lead to the accumulation and preservation of significant volumes of fine-grained overbank deposits, in addition to coarse-grained channel sediments (Figure 16).

Late Quaternary fluvial sediments of continental margins predominantly occur as valley-fill complexes associated with glacial–interglacial cycles (spans ~100 ka; Figure 17). Packages of fluvial sediments accumulate in response to changes in sea level and river hydrology that reflect glacial versus interglacial conditions (Blum and Törnqvist, 2000). In the example of the Colorado River of the Texas Gulf Coast region, the river incised and filled its current valley over the past 70 ka (Blum and Price, 1998). The valley fill consists of a basal late Pleistocene sand-dominated unit that is overlain by a mixed sand- and mud-rich succession of Holocene age. The basal sand-dominated unit comprises a series of down-stepping terraces and well-expressed buried soils (Figure 17). During the Holocene, the valley filled with fine-grained

muddy overbank deposits that encase channel and minor overbank sand bodies. Similar muddy overbank sediments are thin or absent from the basal late Pleistocene portion of the valley fill. This major difference in the abundance of overbank sediments in Holocene and late Pleistocene portion of the valley fill could reflect less frequent episodes of deep overbank flooding during the late Pleistocene (Blum and Törnqvist, 2000).

Over longer time scales involving multiple glacial–interglacial cycles, avulsion during late stages of valley filling can lead to abandonment of an alluvial valley (Blum and Törnqvist, 2000). Avulsion causes a shift in the location of the valley axis and future valley incision and filling, which causes the ages of valley-fill complexes to vary both laterally and vertically (Figure 18). In areas of active subsidence, alluvial valley-fill complexes locally onlap and cross cut one another. The basal valley-fill unconformities form a composite unconformity that is overlain predominantly by sandy channel-belt deposits with well expressed paleosols. In contrast, the upper portions of the valley fills are a combination of fine-grained overbank and sandy channel deposits.

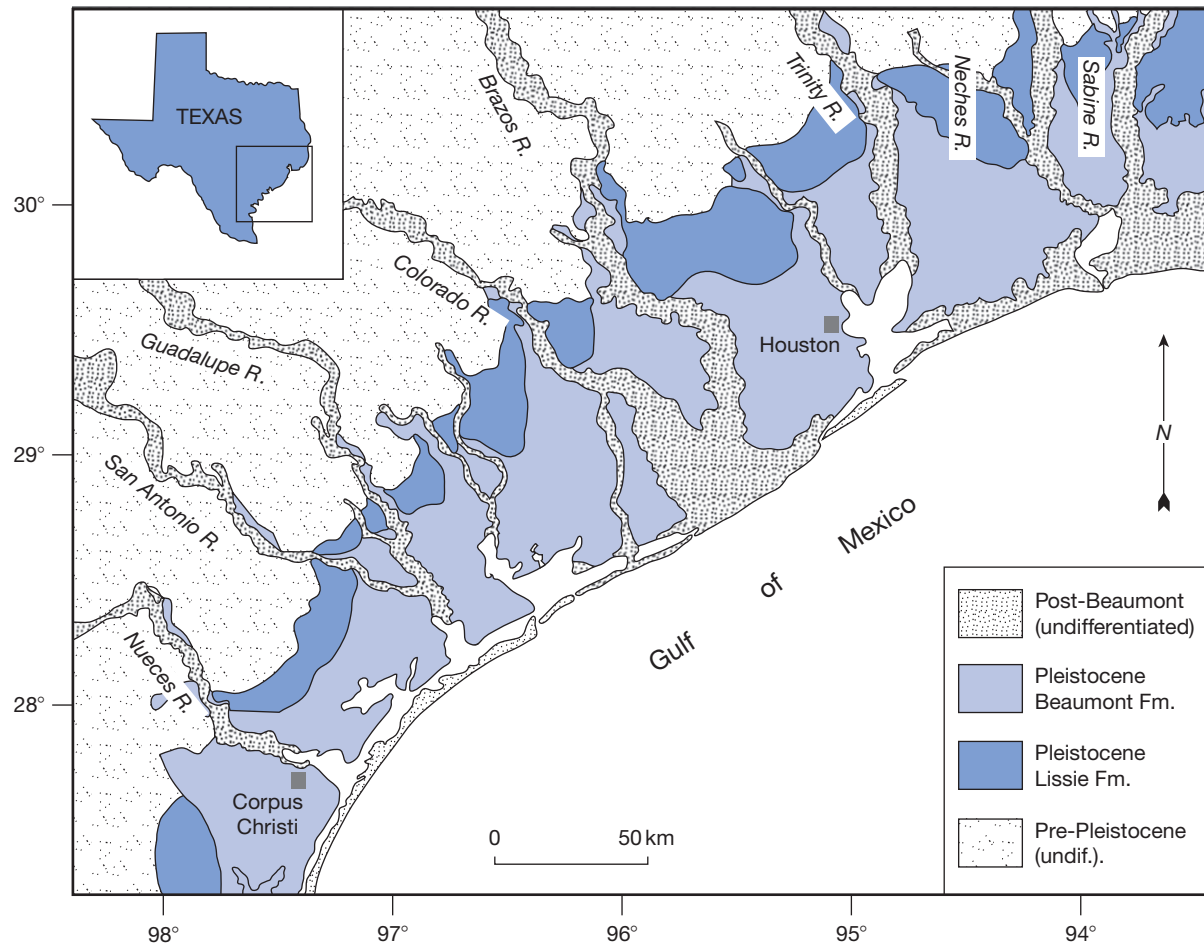


Figure 15 Geologic map of the Texas Gulf Coastal Plain showing the distribution of late Quaternary deposits and major river valleys. Modified from Blum MD and Price DM (1998) Quaternary alluvial plain construction in response to glaci-eustatic and climatic controls, Texas Gulf Coastal Plain. In: Shanley KJ and McCabe PJ (eds.) *Relative Role of Eustasy, Climate and Tectonism on Continental Rocks*, pp. 31–48. SEPM, Special Publication 59.

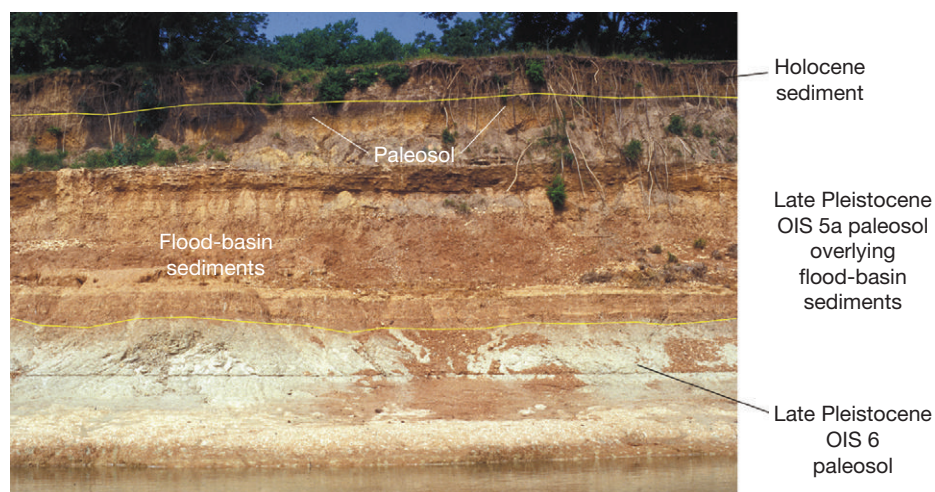


Figure 16 Holocene and late Pleistocene fluvial deposits of the Colorado River of south Texas. Holocene overbank deposits veneer a late Pleistocene paleosol that represents the upper boundary of fluvial sediments deposited during the oxygen isotope stage (OIS) 5a highstand approximately 70 ka. Most of the sediments below this paleosol accumulated during a period of rising sea level. The light colored zone at the base of the section shows a paleosol that formed during OIS 6, which represented a period of alternating episodes of floodplain construction and valley incision.

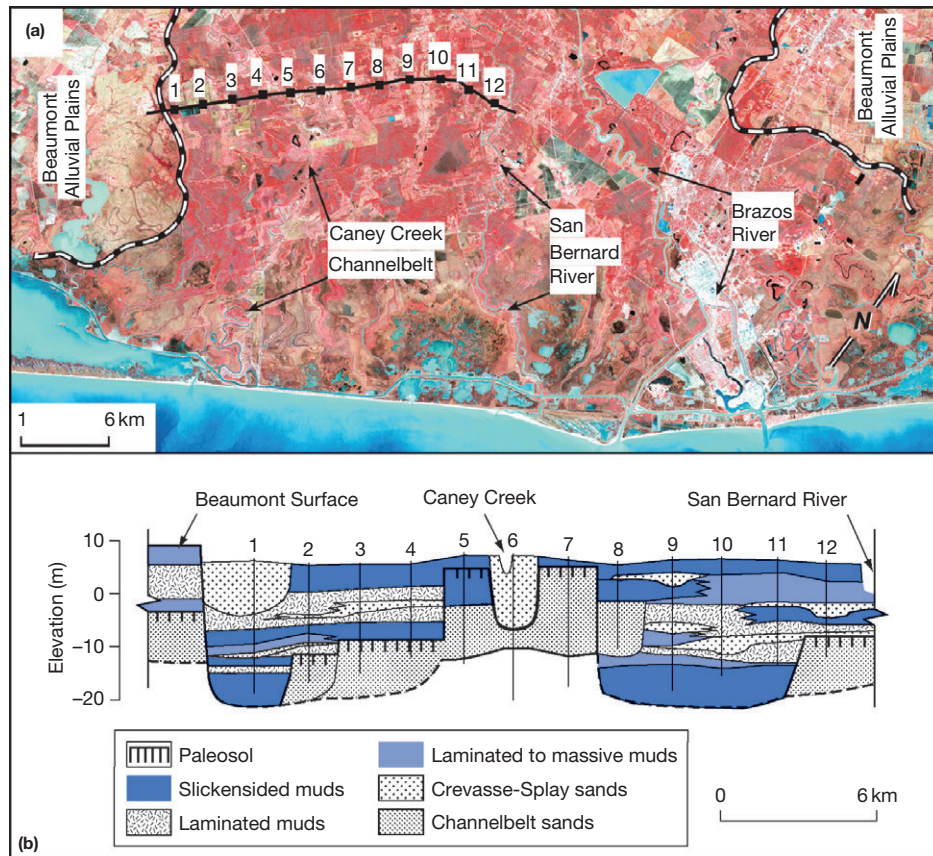


Figure 17 (a) Satellite image of the Lower Colorado River alluvial plain and the transect line and core-sampling sites for the cross section shown in (b). (b) Stratigraphic cross section showing the facies architecture of the late Pleistocene to Holocene Colorado River valley-fill complex. The valley fill consists of a series of down-stepping terraces underlain by channel sands and separated from one another by paleosols and erosional unconformities. The sands range in age from approximately 60–20 ka. These coarse-grained basal valley-fill deposits are overlain by a mixture of channel and overbank sediments that were deposited during late Pleistocene to Holocene sea-level rise. Reprinted from Blum MD and Törnqvist TE (2000) Fluvial responses to climate and sea-level change: A review and look forward. *Sedimentology* 47: 2–48, with permission from Blackwell Publishing.

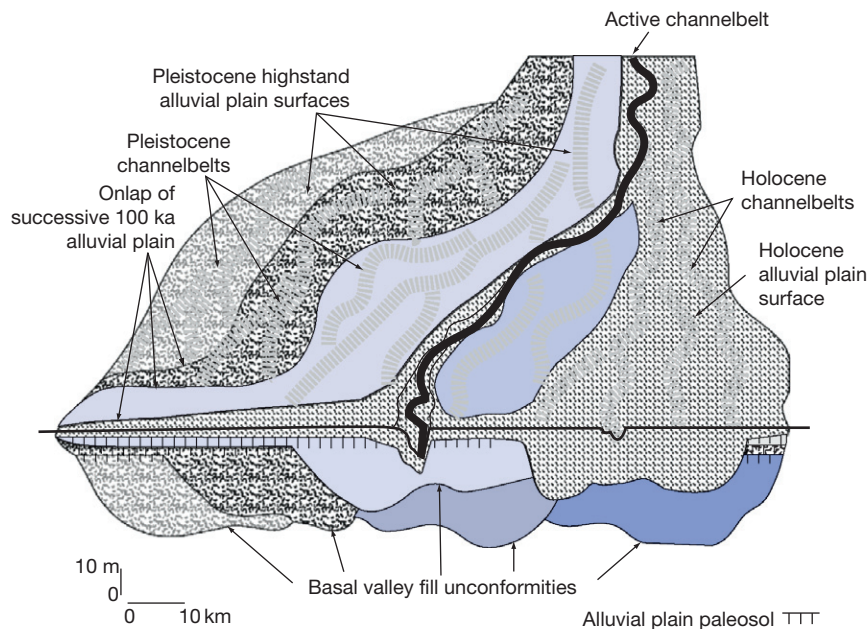


Figure 18 Model showing alluvial-plain construction and the distribution of valley-fill complexes along a subsiding passive margin. Reprinted from Blum MD and Törnqvist TE (2000) Fluvial responses to climate and sea-level change: A review and look forward. *Sedimentology* 47: 2–48.

Summary

Fluvial sediments provide important records of terrestrial environments that are critical for interpreting geologic histories of continental settings. These sediments are also important for evaluating petroleum, coal, and water resources. Fluvial sediments are traditionally subdivided into coarse-grained channel deposits and fine-grained overbank sediments. Channel deposits consist of channel-bar and channel-fill sediments, and accumulate in response to downstream progradation of large-scale bar forms and periodic channel abandonment. Overbank deposits accumulate in natural-levee, crevasse-splay, and flood-basin environments through a combination of overbank flooding and avulsion. Fluvial sediments vary in terms of abundance, architecture, and stratigraphic completeness between continental interiors and margins. In eroding continental interiors, fluvial sediments are represented by flights of terraces underlain by coarse-grained deposits. Fine-grained overbank sediments are relatively rare. These fluvial sediments reflect long-term incision and common sediment reworking, and thus, they preserve relatively incomplete records of fluvial activity. In subsiding continental margins, thick accumulations of both sandy channel and muddy overbank deposits are preserved. These thick fluvial successions provide relatively complete long-term records of fluvial activity.

See also: **Fluvial Environments:** Deltaic Environments; Responses to Rapid Environmental Change; Terrace Sequences. **Glacial Landforms, Sediments:** Glaciofluvial Landforms of Deposition. **Paleoceanography, Physical and Chemical Proxies:** Terrigenous Sediments. **Paleosols and Wind-Blown Sediments:** Nature of Paleosols. **Quaternary Stratigraphy:** Morphostratigraphy/Allostratigraphy; Sequence Stratigraphy.

References

- Allen JRL (1965) A review of the origin and characteristics of recent alluvial sediments. *Sedimentology* 5: 91–191.
- Aslan A and Autin WJ (1999) Evolution of the Holocene Mississippi River floodplain, Ferriday, Louisiana: Insights on the origin of fine-grained floodplains. *Journal of Sedimentary Research* 69: 800–815.
- Aslan A, Autin WJ, and Blum MD (2005) Causes of river avulsion: Insights from the late Holocene avulsion history of the Mississippi River, USA. *Journal of Sedimentological Research* 75: 648–662.
- Blum MD and Price DM (1998) Quaternary alluvial plain construction in response to glaci-eustatic and climatic controls, Texas Gulf Coastal Plain. In: Shanley KJ and McCabe PJ (eds.) *Relative Role of Eustasy, Climate and Tectonism on Continental Rocks*, pp. 31–48. Tulsa, OK: SEPM. Special Publication 59.
- Blum MD and Törnqvist TE (2000) Fluvial responses to climate and sea-level change: A review and look forward. *Sedimentology* 47: 2–48.
- Bridge JS (2003) *Rivers and Floodplains, Forms, Processes, and Sedimentary Record*. Oxford: Blackwell Science.
- Farrell KM (1987) Sedimentology and facies architecture of overbank deposits of the Mississippi River, False River region, Louisiana. In: Ethridge FG and Flores RM (eds.) *Recent Developments in Fluvial Sedimentology*, pp. 111–120. Tulsa, OK: SEPM. Special Publication 39.
- Flores RM (1981) Coal deposition in fluvial paleoenvironments of the Paleocene Tongue River Member of the Fort Union Formation, Powder River area, Powder River Basin, Wyoming and Montana. In: Ethridge FG and Flores RM (eds.) *Recent and Ancient Nonmarine Depositional Environments: Models for Exploration*, pp. 161–190. Tulsa, OK: SEPM. Special Publication 31.
- Hall SA (1990) Channel trenching and climatic change in the southern U.S. Great Plains. *Geology* 18: 342–345.
- Kraus MJ (1999) Paleosols in clastic sedimentary rocks: Their geologic applications. *Earth-Science Reviews* 47: 41–70.
- Leopold LB and Wolman MG (1957) River channel patterns: Braided, meandering and straight. *US Geological Survey Professional Paper* 282-B: 39–85.
- Lewin J and Brewer PA (2001) Predicting channel patterns. *Geomorphology* 40: 329–339.
- Makaske B (2001) Anastomosing rivers: A review of their classification, origin, and sedimentary products. *Earth-Science Reviews* 53: 149–196.
- Miall AD (1977) A review of the braided river depositional environment. *Earth-Science Reviews* 13: 1–62.
- Miall AD (1996) *The Geology of Fluvial Deposits*. Berlin: Springer.
- Morozova GS and Smith ND (1999) Holocene avulsion history of the Lower Saskatchewan fluvial system, Cumberland marshes, Saskatchewan-Manitoba, Canada. In: Smith ND and Rogers J (eds.) *Fluvial Sedimentology VI, International Association of Sedimentologists*, pp. 231–249. International Association of Sedimentologists. Special Publication 28. Oxford: Blackwell Publishing.
- Nanson GC and Croke JC (1992) A genetic classification of floodplains. *Geomorphology* 4: 459–486.
- Pederson JL, Mackley RD, and Eddleman JL (2002) Colorado Plateau uplift and erosion evaluated using GIS. *GSA Today* 12: 4–10.
- Pérez-Arlucea M and Smith ND (1999) Depositional patterns following the 1870s avulsion of the Saskatchewan River (Cumberland Marshes, Saskatchewan Canada). *Journal of Sedimentary Research* 69: 62–73.
- Platt NH and Keller B (1992) Distal alluvial deposits in a foreland basin setting – The lower Freshwater Molasse (Lower Miocene), Switzerland: Sedimentology, architecture and palaeosols. *Sedimentology* 39: 545–565.
- Rust BR (1978) A classification of alluvial rivers. In: Miall AD (ed.) *Fluvial Sedimentology, Memoir Canadian Society Petroleum Geologists*, vol. 5, pp. 187–198. Stacs Data Service Ltd.
- Schumm SA (1977) *The Fluvial System*. New York: John Wiley and Sons.
- Schumm SA (1985) Patterns of alluvial rivers. *Annual Review of Earth and Planetary Sciences* 13: 5–27.
- Slingerland R and Smith ND (2004) River avulsions and their deposits. *Annual Review of Earth and Planetary Sciences* 32: 257–285.
- Smith DG (1986) Anastomosing river deposits, sedimentation rates and basin subsidence, Magdalena River, northwestern Colombia, South America. *Sedimentary Geology* 46: 177–196.
- Smith ND, Cross TA, Diffiey JP, and Clough SR (1989) Anatomy of an avulsion. *Sedimentary* 36: 1–23.
- Smith ND and Pérez-Arlucea M (1994) Fine-grained splay deposition in the avulsion belt of the lower Saskatchewan River. *Canadian Journal of Sedimentary Research* B64: 159–168.
- Stouthamer E and Berendsen HJA (2000) Factors controlling Holocene avulsion history of the Rhine-Meuse delta (The Netherlands). *Journal of Sedimentary Research* A70: 1051–1064.
- Törnqvist TE and Bridge JS (2002) Testing a 3-D process-based alluvial architecture model with new field data: The control of spatial variation in floodplain deposition rate on avulsion frequency. *Sedimentology* 49: 891–905.
- Van den Berg JH and Bledsoe BP (2003) Predicting channel patterns; discussion and reply. *Geomorphology* 53: 333–342.
- Walters MR and Haynes CV (2001) Late Quaternary arroyo formation and climate change in the American Southwest. *Geology* 29: 399–402.
- Willis BJ and Behrensmeier AK (1994) Architecture of Miocene overbank deposits in northern Pakistan. *Journal of Sedimentological Research* B64: 60–67.
- Wolman MG and Leopold LB (1957) River flood plains: Some observations on their formation. *US Geological Survey Professional Paper* 282-C: 87–109.

Responses to Rapid Environmental Change

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Introduction

The recognition of centennial to millennial scale environmental change, notably abrupt climate change, has sparked a great interest in how such rapidly changing boundary conditions influence landscape processes. Rivers are widely considered geomorphic entities that are highly sensitive to environmental change, but it is increasingly being recognized that the fluvial response is extremely complex, with considerable spatiotemporal variability (Knox, 1983). This article provides an overview of the variety of fluvial responses to rapid environmental change and is intimately linked to the preceding article that focuses on longer-term (glacial–interglacial scale) fluvial responses. Primarily, the article will discuss the effects of rapid climate change on fluvial systems in the alluvial reach (i.e., that portion of the fluvial system that is most likely to preserve – at least in a temporary fashion – a stratigraphic record); bedrock rivers are discussed in more detail elsewhere. The first part consists of a discussion of general concepts and criteria for the recognition of fluvial responses, followed by a selection of case studies that provide examples of the different manifestations of late Quaternary fluvial responses to rapid environmental change.

General Concepts

It is usual to distinguish between climate change, base-level (commonly sea-level) change, and tectonism as the three categories of external (or allogenic) forcing of fluvial systems. In addition, a distinction has been made between upstream and downstream controls. While climate change is typically an upstream control (i.e., the forcing propagates downstream through the drainage basin), sea-level change provides a distinct downstream control and can only affect the lowermost reaches of the fluvial system. Considerable debate has centered on the question of how far upstream sea-level control can propagate and it is likely that this varies greatly, depending on variables such as slope and sediment supply (Blum and Törnqvist, 2000). In principle, tectonism can affect any segment of the fluvial system. In addition, the various allogenic controls commonly operate simultaneously.

Modes of Response to Allogenic Forcing

There are different mechanisms of fluvial response to allogenic forcing:

1. changes in the discharge regime;
2. changes in channel pattern (fluvial style); and
3. changes in the longitudinal profile by means of aggradation or degradation (incision).

Perhaps the most obvious and direct response to climate forcing is the change of discharge regime and associated changes in hydraulic geometry parameters, such as channel width, depth, and sinuosity. Such adjustments may in some cases lead to full transformations of fluvial style.

Changes in fluvial style have been the subject of research for a long time, commonly focusing on the time period around the Pleistocene–Holocene transition (Kozarski, 1983; and numerous others). Many of these investigations concern the transformation from braided to meandering channel patterns. While there is a lively ongoing debate on the boundary conditions that cause rivers to braid or meander (Lewin and Brewer, 2001; Van den Berg, 1995), it appears likely that braiding is favored by a high bed load and a high stream power (i.e., discharge and valley slope). In addition, several other factors have been mentioned in the literature (e.g., limited bank stability). Rapid climate change can modify boundary conditions that control fluvial style, but changes in fluvial style due to climate forcing can be spatially nonuniform.

Perhaps the most prolific response of the fluvial system to allogenic forcing is the adjustment of the longitudinal profile, either by aggradation or by degradation (incision). The concept of grade (Mackin, 1948) postulates that alluvial rivers strive to attain an equilibrium profile characterized by a more or less balanced sediment budget, typical of stable boundary conditions in terms of base level, discharge, and sediment supply. Disruptions of this delicate balance commonly occur by means of changes in the proportion between water discharge (Q_w) and sediment supply (Q_s). It is important to note that sediment supply in this context refers strictly to bed load, given the fact that rivers are virtually never saturated with respect to the transport of suspended load, while the supply of bed load at times exceeds sediment-transport capacity. Figure 1 shows how the balance between Q_w and Q_s dictates whether a fluvial reach will incise, aggrade, or maintain equilibrium. In this context, it is important to reiterate that disruptions in the Q_w/Q_s ratio may be accommodated by other adjustments (such as sinuosity), rather than by lowering or raising the longitudinal profile.

Recognizing Fluvial Incision

Fluvial aggradation and incision not only reflect changes in the sediment budget but also leave the most tangible imprint in the stratigraphic record. In fact, many studies of fluvial archives aim to identify fluvial unconformities resulting from phases of incision with regional significance that may be ascribed to allogenic forcing. However, the proper recognition of regional unconformities representing widespread incision has proven challenging for a number of reasons. This is due to the simple fact that river channels are erosional features by definition. Following Salter (1993) and Best and Ashworth (1997), it is

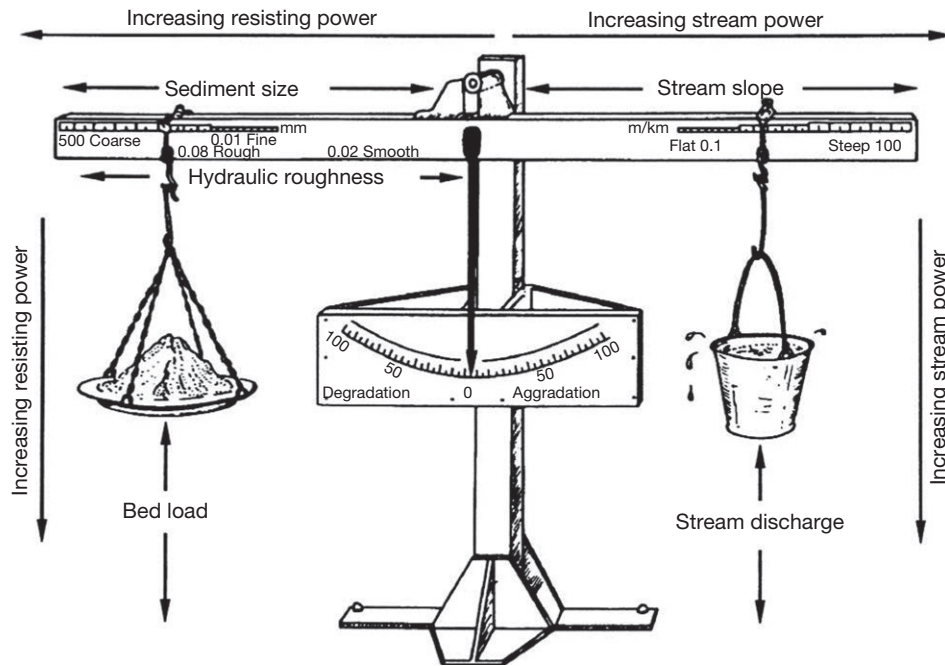


Figure 1 Balance model for aggradation and degradation (incision) in alluvial rivers, as dictated by the relative proportion of water discharge and sediment supply, referred to by Bull (1991) as the balance between stream power and resisting power. Channels aggrade when sediment supply (bed load: amount and grain size) exceeds sediment-transport capacity as determined primarily by discharge and slope, and incise when the reverse is the case. The role of base-level change can also be inferred from this conceptualization. Base-level rise tends to reduce the slope and therefore promote aggradation, whereas base-level fall may increase the slope leading to incision (also depending on the morphology of the surface that is exposed). Originated from an unpublished diagram by W. Borland. Reprinted from Bull WB (1991) *Geomorphic Responses to Climatic Change*. New York: Oxford University Press, with permission from Oxford University Press.

therefore critical that distinction be made between fluvial scour, the autogenic (intrinsic) process that characterizes any alluvial channel (Figure 2) versus allogenic incision. Best and Ashworth (1997), working in the vertically aggrading Ganges–Brahmaputra Delta, showed that channel scour can lead to reworking of the channel belt to depths as much as five times the mean channel depth. They go on to show how the resulting stratigraphic phenomena can easily be mistaken for fluvial incision with regional significance. In fact, numerous examples can be gleaned from the literature where channel scour in overall aggradational settings may have been misinterpreted as fluvial incision.

So how can fluvial incision and aggradation be recognized in the stratigraphic record? Given the discussion above, this is a critical, yet by no means a trivial question. Distinction should be made as to whether incision/aggradation is measured relative to the channel bed or the floodplain surface. Incision of the channel bed can be the result of base-level change, vertical crustal movements, or changes in Q_w relative to Q_s (Figure 3(a)). However, changes in the discharge regime alone (i.e., the ‘flashiness’ of discharge) can cause floodplain aggradation or degradation, without vertical changes in the channel bed. Floodplains that are frequently inundated by deep overbank floods have the potential to aggrade to levels well above the channel bed (Figure 3(b)). If lateral channel migration goes along with a reduction of the peak discharge (e.g., as a result of a decrease in the frequency of major storms) a new floodplain will, over time, be established at a lower level, leaving the previous floodplain behind as a terrace.

These two mechanisms of fluvial terrace formation are shown schematically in Figure 4(a) and 4(b), respectively, and are further illustrated by means of animations available from the supplementary materials. Given these different mechanisms of terrace formation, a variety of combinations of processes are capable of producing – at least in theory – a plethora of stratigraphic results that may prove challenging to unravel (Blum and Törnqvist, 2000). For example, it cannot be ruled out that different allogenic forcings operate simultaneously and cause opposite stratigraphic results (Figure 4(c)). In this hypothetical example, floodplain aggradation due to an increase in peak discharge would occur simultaneously with channel-bed incision, for instance due to a slight base-level drop, and vice versa. Finally, it should be stressed that incision as defined here is broadly defined so as to encompass floodplain degradation in the absence of channel-bed incision. Obviously, long-term incision over vertical intervals that substantially exceed mean channel depth cannot occur without channel-bed incision.

The overarching conclusion is that inferring fluvial incision and aggradation from the stratigraphic record can be most successfully done when data are available on both floodplain surfaces and channel bases. Unfortunately, this presents the problem that in the naturally incomplete stratigraphic record channel bases tend to have a much higher preservation potential than floodplain surfaces. Therefore, it is absolutely critical that the criteria for recognition of fluvial incision and aggradation be explicitly discussed. Unfortunately, often this is not the case, and the literature is riddled with examples of studies

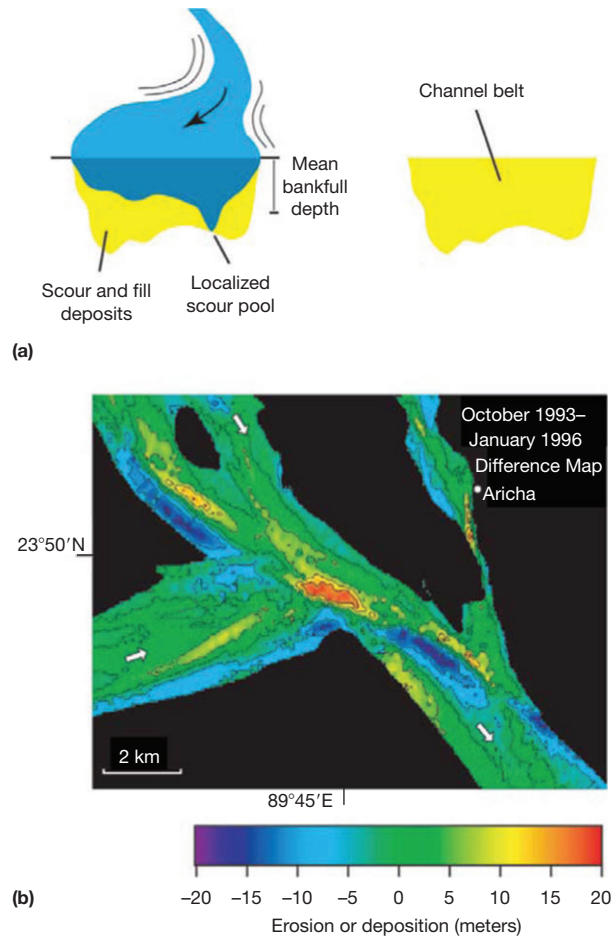


Figure 2 (a) Channel scour and fill in alluvial rivers leads to continuous reworking and deepening of the channel belt, exceeding mean bankfull channel depth. Modified from Salter T (1993) Fluvial scour and incision: Models for their influence on the development of realistic reservoir geometries. In: North CP and Prosser DJ (eds.) *Characterization of Fluvial and Aeolian Reservoirs: Geological Society Special Publication*, vol. 73, pp. 33–51. (b) Repeated bathymetric surveys of the Ganges–Brahmaputra confluence show the dynamic nature of alluvial channels and their capability to form scour holes about five times the average channel depth that are likely to rework channel belts over annual to decadal timescales. Reproduced from Best JL and Ashworth PJ (1997) Scour in large braided rivers and the recognition of sequence stratigraphic boundaries. *Nature* 387: 275–277. Caution must be taken not to confuse such phenomena with fluvial incision and associated floodplain degradation. Reprinted by permission from Macmillan Publishers Ltd.

that infer longitudinal-profile adjustments while potentially misinterpreting autogenic scour phenomena as reflecting regional incision (e.g., due to a focus exclusively on channel-belt bases rather than floodplain surfaces).

Complex Response and Response Time

A potentially large obstacle in the interpretation of fluvial stratigraphic records is the possibility of the occurrence of nonlinear responses. An experimental study by Schumm and Parker (1973) showed how a simple, rapid drop in base level can lead to alternating fluvial incision and aggradation.



Figure 3 (a) The Sabarmati River, Gujarat, India, about 100 km upstream from Ahmadabad, bordered by a ~40-m-high terrace. The uppermost terrace deposits date to ~12 ka, whereas the bars are ~4.5 ka in age (Srivastava et al., 2001), indicating spectacular incision that has been attributed to the early Holocene increase in monsoon activity that affected large sections of South Asia during this time (Tandon et al., 2006). Photograph courtesy Martin Gibling. Reprinted by permission from SEPM (Society for Sedimentary Geology). (b) The Amite River in southeast Louisiana after a peak discharge event. Note the high-water mark recognizable by the mud-covered vegetation. The lower arrow shows the approximate thickness of fine-grained overbank deposits overlying sandy point-bar strata. The upper arrow indicates the minimal amount of continued floodplain aggradation that can be expected if the current discharge regime remains unaltered.

The concept of ‘complex response’ was subsequently discussed at length by Schumm (1977), who advocated that such non-linearity can be caused by a variety of perturbations (sediment supply, tectonism), both upstream and downstream. The cyclic nature of the Schumm and Parker (1973) experiment was later substantiated by mathematical modeling (Slingerland and Snow, 1988).

Regardless of the possibility of complex response, it has long been recognized that fluvial responses to environmental forcing exhibit time lags of variable duration. For example, Bull (1991) defines the geomorphic response time as the sum of the reaction time (time between the perturbation and the initial adjustment) and the relaxation time (time between the initial adjustment and renewed equilibrium). The spirit of these concepts is not unlike the theoretical work by Paola et al. (1992), who introduced the concept of equilibrium time (T_{eq}), defined as

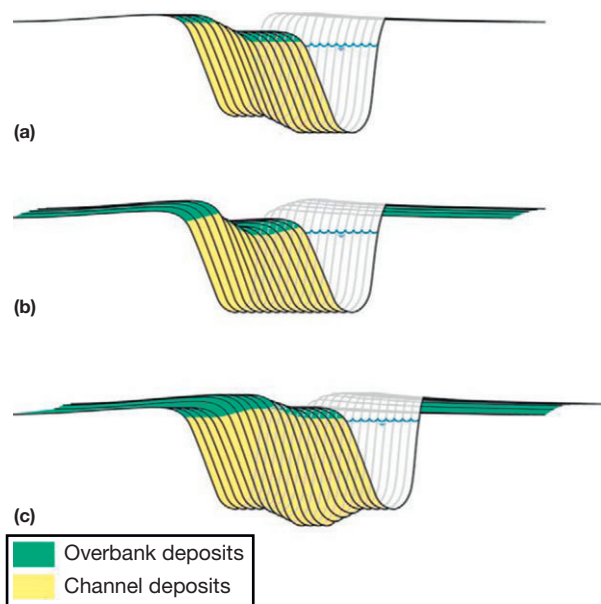


Figure 4 Hypothetical cross-sections illustrating contrasting mechanisms of fluvial incision, floodplain abandonment, and terrace formation (Blum and Törnqvist, 2000). (a) Equilibrium floodplain with lateral migration (from left to right) and minimal vertical aggradation, followed by a short phase of incision characterized by degradation of both the channel bed and the floodplain and a subsequent return to equilibrium conditions. This could be the result of either a reduction in bed load relative to discharge, a drop in base level, or tectonic uplift (or combinations thereof). (b) Fluvial system characterized by lateral migration and vertical aggradation of the floodplain due to frequent overbank flooding. A phase of floodplain degradation due to a reduction in peak discharge is followed by continued conditions of lateral migration and floodplain aggradation at a lower level. Note that the channel bed remains unchanged and that terrace formation is entirely the result of a drastic decrease in peak discharge and resulting floodplain degradation. (c) A combination of the processes operating in the previous two scenarios, where channel-bed degradation occurs simultaneously with floodplain aggradation due to increased overbank flooding. Subsequently, channel-bed aggradation coincides with reduced overbank flooding, resulting in the counterintuitive situation of simultaneous terrace formation and channel-bed aggradation.

$$T_{eq} = L^2/v$$

where L is the basin (river) length (a proxy for the size of the drainage basin) and v is the diffusivity (efficiency of sediment transport). It follows that small drainage basins that permit rapid sediment transfer are more likely to record abrupt, short-term climate changes than large fluvial systems. Recently, [Castelltort and Van Den Driessche \(2003\)](#) have applied this theory to 93 large rivers worldwide, suggesting that tens to hundreds of thousands of years are needed to fully accommodate changes in sediment supply and restore equilibrium conditions. In addition, these authors suggest that upstream forcing is in many cases accommodated in the fluvial transfer zone, implying that perturbations would not propagate to the river mouth. Nevertheless, as will be exemplified below, several cases have been documented of fairly rapid fluvial responses to

environmental forcing, including examples from large drainage basins.

Case Studies

This article focuses on climate controls, because sea-level fluctuations that affect the downstream portions of the longitudinal profile commonly act over longer time intervals. With the exception of episodic seismic events, tectonism also concerns longer timescales. The last glacial and its transition into the Holocene have witnessed numerous abrupt climate changes that have affected considerable parts of the globe. These changes manifest themselves as drastic fluctuations in temperature and precipitation, particularly well documented in the Greenland ice-core records. The Dansgaard–Oeschger events serve as some of the most prolific examples that were of specific importance in the North Atlantic region, but lower latitudes have experienced variations in monsoonal activity as well as higher frequency changes such as the El Niño–Southern Oscillation. In addition, solar forcing has been proposed as another mechanism for short-term climate change.

In many respects, the Pleistocene–Holocene transition can be viewed as one of the most abrupt climate changes of the last glacial–interglacial cycle, in essence corresponding to Termination I. This is particularly the case in temperate regions of the Northern Hemisphere that were strongly influenced, either directly or indirectly, by glaciation. These are also the environments that have the longest tradition of studies of late Quaternary fluvial evolution, so the discussion below will open with a brief overview of some of those findings. The subsequent sections present selected examples of fluvial responses to climate forcing over different timescales, with a premier emphasis on longitudinal-profile adjustments. It should be noted that the subdivision of examples reflects interpretations of the stratigraphic record by the respective authors, based on a variety of observational data. The examples range from monsoonal fluctuations that operate over 10^3 – 10^4 year timescales, to Dansgaard–Oeschger style abrupt (10^2 – 10^3 years) climate changes that have been particularly prominent in the North Atlantic region, as well as higher frequency and lower amplitude (10^0 – 10^3 years) climate fluctuations with variable causes that characterize the Holocene. The examples provided here not only concern different timescales but also contrasting environmental settings, ranging from the tropics to temperate regions (with the latter predominantly subarctic during the last glacial).

Pleistocene–Holocene Transition

A long tradition exists of studies of fluvial channel-pattern changes across the last glacial to interglacial transition, particularly in temperate regions outside of the former centers of glaciation. Such investigations have been numerous in the northwest European lowlands that experienced an overall transformation from periglacial conditions with sparse vegetation cover to temperate, deciduous forests with climate conditions roughly comparable to those of the present. For example, detailed mapping by [Berendsen and Stouthamer \(2001\)](#) of the floodplains and adjacent terraces of the lower Rhine and

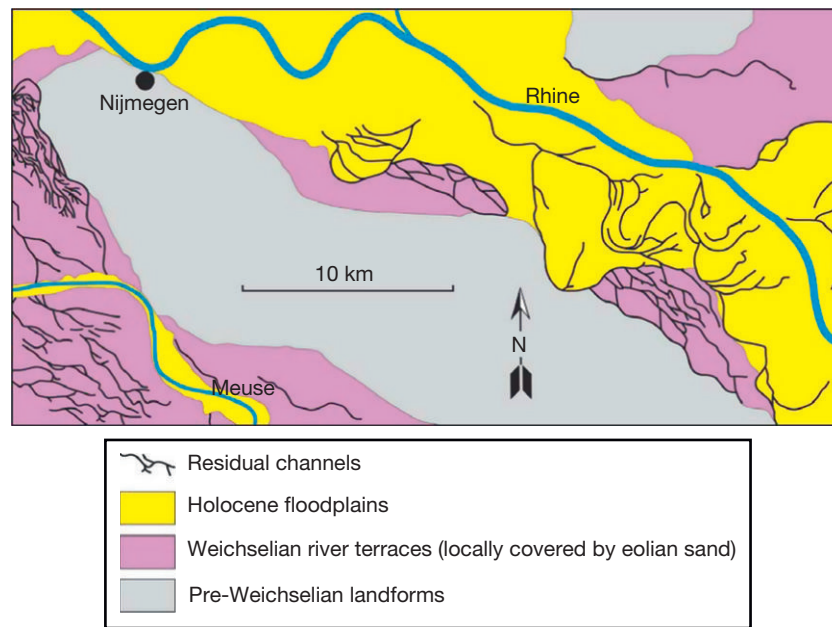


Figure 5 Channel patterns along the Rhine and Meuse Rivers during the late Pleistocene and Holocene near the Dutch–German border. Note the braided channel patterns on the Weichselian (late Pleistocene) terraces, as opposed to the meandering Holocene fluvial style. Reproduced from Berendsen HJA and Stouthamer E (2001) *Palaeogeographic Development of the Rhine–Meuse Delta, The Netherlands*. Assen: Koninklijke Van Gorcum.

Meuse Rivers near the Dutch–German border (**Figure 5**) shows that braided channel patterns dominated during the last glacial, rapidly giving way to meandering planforms in the Holocene. Numerous investigators have argued that braiding was promoted by a high supply of bed load due to a lack of vegetation cover and active periglacial hillslope processes. In contrast, increased stabilization of soils and riverbanks by vegetation during the Holocene favored the transition to meandering patterns. These findings have been mostly substantiated by recent modeling (Bogaart et al., 2003a).

It is interesting to note that strikingly similar changes in fluvial style have been mapped along the Rhine and Meuse (**Figure 5**), despite the different character of these two rivers around the Pleistocene–Holocene transition, with the Rhine – unlike the Meuse – draining a substantially glaciated hinterland. Similar observations have been made in North America, with braided channel patterns dominating both glaciated drainage basins such as the Mississippi River (Fisk, 1944; Saucier, 1994) and ice-free settings along the Atlantic Coastal Plain (Leigh et al., 2004) during the last glacial. As will be discussed in more detail below, higher frequency changes in fluvial style and longitudinal-profile adjustments occurred during the Pleistocene–Holocene transition in northwest Europe.

Monsoonal Forcing

Precession-dominated fluctuations in monsoonal activity in low-latitude regions are well documented. A number of recent studies highlight the associated dramatic responses of fluvial systems over millennial or slightly longer timescales, including extensive work in the Himalayan Foreland Basin. Tandon et al. (2006) present a large body of data from the drainage basin of the Ganges River, suggesting that widespread fluvial incision (commonly on the order of tens of meters), floodplain

abandonment, and terrace formation was especially prominent during time intervals with a strengthened monsoon, particularly the period from about 15 to 5 ka. Similar processes operated farther west in India (**Figure 3(a)**). The preceding period around the Last Glacial Maximum, when the southwestern monsoon was suppressed and discharge was substantially lower, was characterized predominantly by fluvial aggradation. Goodbred (2003) pointed out similar relationships and demonstrated the intimate connection between accelerated incision in the drainage basin of the Ganges River during the hypsithermal (notably the early Holocene) and accelerated sedimentation rates in the Ganges–Brahmaputra Delta. The significance of this finding is that this giant sediment-dispersal system was capable of transmitting a climate signal over a very short (millennial) timescale, from source to sink. Studies like the ones discussed here have been possible because of rapid advances in optically stimulated luminescence dating and its application to strata that previously lacked robust age control.

Dansgaard–Oeschger Style Forcing

The North Atlantic region could be viewed as the type area for abrupt climate change, particularly during the last glacial. The Dansgaard–Oeschger events, characterized by sudden warming that lasted for no more than a few thousand years before reversing back to glacial conditions, are likely to have caused major landscape changes. By far the best-documented Dansgaard–Oeschger event is the most recent one (Bølling–Allerød–Younger Dryas), constituting the final episode of the last glacial. Numerous studies of northwest European rivers have revealed at least some commonality in changes of fluvial styles and terracing during this time interval. The Meuse River in The Netherlands has been particularly extensively studied (Huisink, 1997; Tebbens et al., 1999). Although these two

investigations use conflicting age models they share a number of findings, including a rapid (Tebbens et al., 1999) or gradual (Huisink, 1997) transition from full-glacial braided channel patterns to a meandering fluvial style during the Allerød, reversing back to braiding during the Younger Dryas. In addition, several phases of incision and aggradation have been inferred, although some of these have been subsequently questioned by Bogaart et al. (2003b) who drew the attention to the same 'scour versus incision' problem as discussed above. Nevertheless, Bogaart et al. (2003b), using a novel modeling approach aimed explicitly at understanding centennial to millennial scale climatic forcing of fluvial systems, show the plausibility of short-lived incision associated with abrupt climate change, comparable to the pioneering work by Knox (1972) that is discussed below. This finding is consistent with floodplain degradation at the Allerød–Younger Dryas transition along the Meuse, as inferred by Huisink (1997).

One intriguing challenge in studies of fluvial evolution is the possible impact of the numerous earlier Dansgaard–Oeschger events of the last glacial. Studies of fluvial successions from this time interval in The Netherlands (Van Huissteden and Kasse, 2001) have suggested that prior to the last Dansgaard–Oeschger event discussed above, fluvial systems exhibited mostly autogenic behavior with only few episodes of more widespread erosion. They point out that the nature of fluvial responses is dependent on overall environmental boundary conditions, with a particularly prominent role for vegetation and permafrost. Only during the final episode of the last glacial an alternation between open shrub tundra and woodland allowed more dramatic fluvial responses due to the vegetational control on the hydrologic cycle. Nevertheless, there is ample room for more studies of fluvial records of the last glacial, especially in light of the current feasibility of better chronological frameworks.

Holocene Climate Forcing

The potentially nonlinear nature of fluvial response to abrupt environmental change was discussed by Knox (1972), who argued that high-order tributaries in southwest Wisconsin located near a major ecotone responded in a highly sensitive way to Holocene shifts in atmospheric circulation. An abrupt transformation from prairie to hardwood forest ~6000 ¹⁴C years BP was associated with rapid alluviation along these streams. As shown in Figure 6, changes in fluvial activity are particularly concentrated in the relatively short interval following an abrupt shift in climate, largely due to the time lag in vegetational response. More specifically, a transition from arid to humid conditions provides a small time window with increased discharge and sparse vegetation cover, allowing accelerated hillslope erosion due to a limited infiltration capacity. This results in increased sediment yield and 'geomorphic work.' Knox (1972) raises the point that the exact nature of the fluvial response, in terms of incision versus aggradation, is highly dependent on location within the drainage network.

Somewhat different conditions prevailed in environments that are characterized by more flashy discharge regimes, with peak discharges as much as two or three orders of magnitude higher than mean annual discharge. This is illustrated by a number of rivers in central Texas, including the Colorado

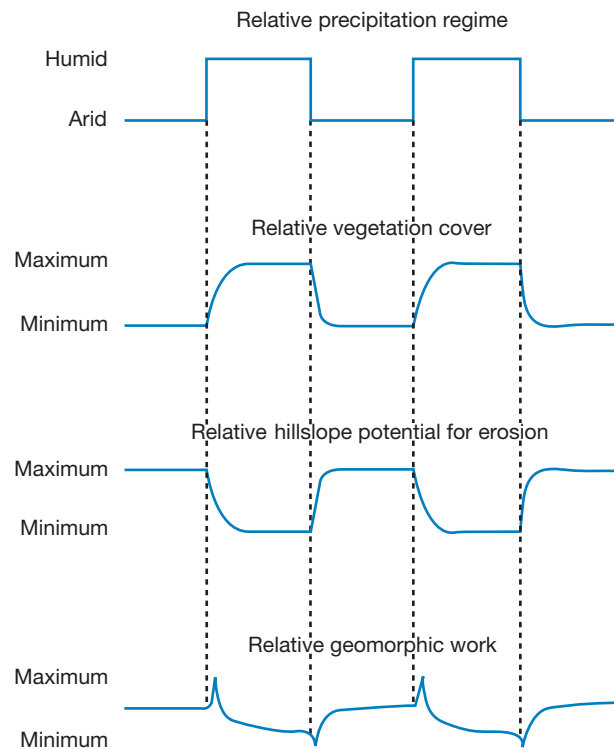


Figure 6 Conceptual model of nonlinear fluvial response to climate forcing. Note that this model is intended to apply specifically to mid-latitude regions with an annual precipitation of 250–1500 mm. Also note that 'geomorphic work' could refer to both incision and aggradation, depending on location within the drainage basin. Reproduced from Knox JC (1972) Valley alluviation in southwestern Wisconsin. *Annals of the Association of American Geographers* 62: 401–410.

River (Blum et al., 1994). Figure 7 shows a cross-section of this fluvial system featuring three Holocene allostratigraphic units. While the base of each unit is found at a roughly constant level, their surfaces vary widely, with the early Holocene unit entirely draped by younger strata. Blum et al. (1994) explain this by variations in the persistence of deep flooding, which was especially prolific during the relatively humid time period 2500–1000 ¹⁴C years BP. In contrast, the past millennium was drier and led to the establishment of an inset floodplain at a lower level, along with pedogenesis and terrace formation.

Daniels and Knox (2005) provide an excellent example of a study of century-scale fluvial response, in their case with respect to the drought that characterized the interior of North America during the Medieval Warm Period. They show that tributaries in a small drainage basin in southwest Nebraska incised rapidly around 800–1100 ¹⁴C years BP, roughly synchronous to rivers elsewhere in Nebraska and Kansas. Drought in this semiarid region is associated with dramatic reductions in vegetation cover and a more variable discharge regime with less frequent, yet more powerful, floods. This would enable high erosion rates and the establishment of new floodplain levels several meters lower, within a time period of a few centuries or less. It appears that fluvial incision in this environment is correlated with drought, whereas in examples given

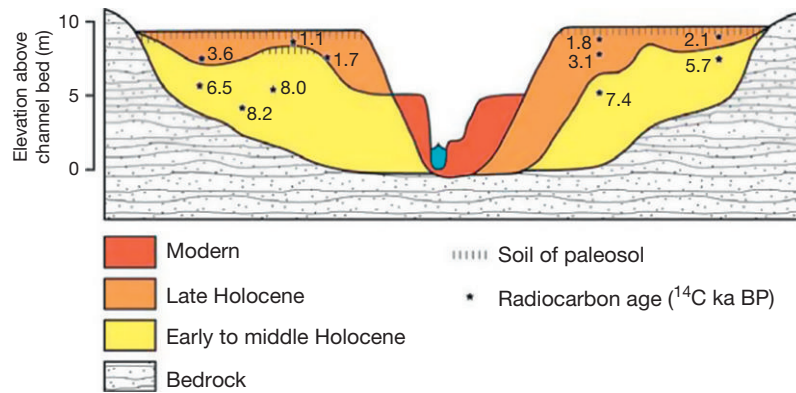


Figure 7 Three Holocene allostratigraphic units along the upper Colorado River, central Texas. For explanation see text. Reproduced from Blum MD, Toomey III RS, and Valastro S Jr. (1994) Fluvial response to Late Quaternary climatic and environmental change, Edwards Plateau, Texas. *Palaeogeography, Palaeoclimatology, Palaeoecology* 108: 1–21.

above, such as the Meuse River, early Holocene incision was associated with an increase in discharge (Bogaart et al., 2003b). This testifies again to the complexity of the fluvial response and its strong dependence on regionally specific boundary conditions.

Relatively rare are stratigraphic studies that focus on the response of the fluvial sediment transfer system to annual to decadal scale climate forcing. Naturally, it is most practical to study such relationships in the most recent sedimentary records, given the need for exceptionally high temporal resolution. Only a few environments remain that have been essentially unaffected by human intervention over the past decades or century. Recent work in the pristine foreland basin of the Amazon drainage (Aalto et al., 2003) suggests that floodplain aggradation is not a gradual process, but rather consists of rapid pulses of sedimentation separated by multiyear intervals of quiescence. These authors documented a strong correlation with the El Niño–Southern Oscillation, with rapid aggradation in the southern part of the Amazon Basin occurring during rainy La Niña intervals.

Summary

The interest in fluvial responses to rapidly changing climatic boundary conditions is here to stay, and will likely increase given the concerns about the impact of future global change. While there exists a rich literature of investigations of fluvial records for the past ~20 ka, studies over shorter (decadal or centennial) timescales are less common (Blum and Törnqvist, 2000). One likely reason for this is the formidable task of developing high-resolution geochronological frameworks in fluvial successions, environments that commonly pose serious challenges for numerical dating.

The diverse collection of examples presented herein shows that there is a wide variety of fluvial responses to climate forcing. For example, the impact of increased aridity or humidity can be opposite in different geographic regions, to a large extent depending on initial boundary conditions. As a result, the scientific literature on the topic covered by this article is somewhat overwhelming and confusing, with numerous

conflicting data and interpretations. To a considerable extent these conflicts are probably real, reflecting spatial variability, differences in sensitivity, and contrasting response times. However, it is likely that some of these conflicts simply stem from erroneous stratigraphic interpretations.

There are also discrepancies between field evidence and theory, given that large rivers may respond swiftly to environmental forcing, as so eloquently illustrated by the Ganges system. Thus, major challenges lie ahead. The emergence of better and more rigorous concepts that are more strongly rooted in theory, as well as better analytical tools (notably dating methods), will likely offer new opportunities to tackle these important issues.

See also: Glacial–Interglacial Scale Fluvial Responses; Paleohydrology. **Fluvial Environments:** Deltaic Environments; Sediments; Terrace Sequences. **Ice Core Records:** Greenland Stable Isotopes. **Luminescence Dating:** Optical Dating. **Paleoclimate Reconstruction:** Sub-Milankovitch (DO/Heinrich) Events; The Last Millennium. **Quaternary Stratigraphy:** Morphostratigraphy/Allostratigraphy; Tephrochronology.

References

- Aalto R, Maurice-Bourgoin L, Dunne T, Montgomery DR, Nittrouer CA, and Guyot J-L (2003) Episodic sediment accumulation on Amazonian flood plains influenced by El Niño/Southern Oscillation. *Nature* 425: 493–497.
- Berendsen HJA and Stouthamer E (2001) *Palaeogeographic Development of the Rhine-Meuse Delta, the Netherlands*. Assen: Koninklijke Van Gorcum.
- Best JL and Ashworth PJ (1997) Scour in large braided rivers and the recognition of sequence stratigraphic boundaries. *Nature* 387: 275–277.
- Blum MD, Toomey RS III, and Valastro S Jr. (1994) Fluvial response to Late Quaternary climatic and environmental change, Edwards Plateau, Texas. *Palaeogeography, Palaeoclimatology, Palaeoecology* 108: 1–21.
- Blum MD and Törnqvist TE (2000) Fluvial responses to climate and sea-level change: A review and look forward. *Sedimentology* 47(supplement 1): 2–48.
- Bogaart PW, Van Balen RT, Kasse C, and Vandenberghe J (2003a) Process-based modelling of fluvial system response to rapid climate change – I: Model formulation and generic applications. *Quaternary Science Reviews* 22: 2077–2095.
- Bogaart PW, Van Balen RT, Kasse C, and Vandenberghe J (2003b) Process-based modelling of fluvial system response to rapid climate change II. Application to the

- River Maas (The Netherlands) during the Last Glacial–Interglacial Transition. *Quaternary Science Reviews* 22: 2097–2110.
- Bull WB (1991) *Geomorphic Responses to Climatic Change*. New York: Oxford University Press.
- Castelltort S and Van Den Driessche J (2003) How plausible are high-frequency sediment supply-driven cycles in the stratigraphic record? *Sedimentary Geology* 157: 3–13.
- Daniels JM and Knox JC (2005) Alluvial stratigraphic evidence for channel incision during the Mediaeval Warm Period on the central Great Plains, USA. *The Holocene* 15: 736–747.
- Fisk HN (1944) *Geological Investigation of the Alluvial Valley of the Lower Mississippi River*. Vicksburg: Mississippi River Commission.
- Goodbred SL Jr. (2003) Response of the Ganges dispersal system to climate change: A source-to-sink view since the last interstade. *Sedimentary Geology* 162: 83–104.
- Huisink M (1997) Late-glacial sedimentological and morphological changes in a lowland river in response to climatic change: The Maas, southern Netherlands. *Journal of Quaternary Science* 12: 209–223.
- Knox JC (1972) Valley alluviation in southwestern Wisconsin. *Annals of the Association of American Geographers* 62: 401–410.
- Knox JC (1983) Responses of river systems to Holocene climates. In: Wright HE Jr. (ed.) *Late-Quaternary Environments of the United States. The Holocene*, vol. 2, p. 26. Minneapolis: University of Minnesota Press.
- Kozarski S (1983) River channel adjustment to climatic change in west central Poland. In: Gregory KJ (ed.) *Background to Palaeohydrology. A Perspective*, p. 355. Chichester: John Wiley.
- Leigh DS, Srivastava P, and Brook GA (2004) Late Pleistocene braided rivers of the Atlantic Coastal Plain, USA. *Quaternary Science Reviews* 23: 65–84.
- Lewin J and Brewer PA (2001) Predicting channel patterns. *Geomorphology* 40: 329–339.
- Mackin JH (1948) Concept of the graded river. *Geological Society of America Bulletin* 59: 463–511.
- Paola C, Heller PL, and Angevine CL (1992) The large-scale dynamics of grain-size variation in alluvial basins, 1: Theory. *Basin Research* 4: 73–90.
- Salter T (1993) Fluvial scour and incision: Models for their influence on the development of realistic reservoir geometries. In: North CP and Prosser DJ (eds.) *Characterization of Fluvial and Aeolian Reservoirs*, vol. 73, pp. 33–51. London: Geological Society. Special Publication.
- Saucier RT (1994) *Geomorphology and Quaternary Geologic History of the Lower Mississippi Valley*. Vicksburg: Mississippi River Commission.
- Schumm SA (1977) *The Fluvial System*. New York: John Wiley.
- Schumm SA and Parker RS (1973) Implications of complex response of drainage systems for Quaternary alluvial stratigraphy. *Nature Physical Science* 243: 99–100.
- Slingerland RL and Snow RS (1988) Stability analysis of a rejuvenated fluvial system. *Zeitschrift für Geomorphologie, Neue Folge, Supplementband* 67: 93–102.
- Srivastava P, Juyal N, Singhvi AK, Wasson RJ, and Bateman MD (2001) Luminescence chronology of river adjustment and incision of Quaternary sediments in the alluvial plain of the Sabarmati River, north Gujarat, India. *Geomorphology* 36: 217–229.
- Tandon SK, Gibling MR, Sinha R, et al. (2006) Alluvial valleys of the Ganga Plains, India: Timing and causes of incision. In: Dalfymlle RW, Leckie D, and Tillman RW (eds.) *Incised-Valley Systems in Time and Space*. Tulsa, OK: SEPM, Special Publication.
- Tebbens LA, Veldkamp A, Westerhoff W, and Kroonenberg SB (1999) Fluvial incision and channel downcutting as a response to Late-glacial and early Holocene climate change: The lower reach of the River Meuse (Maas), the Netherlands. *Journal of Quaternary Science* 14: 59–75.
- Van den Berg JH (1995) Prediction of alluvial channel pattern of perennial rivers. *Geomorphology* 12: 259–279.
- Van Huissteden J and Kasse C (2001) Detection of rapid climate change in Last Glacial fluvial successions in the Netherlands. *Global and Planetary Change* 28: 319–339.

Terrace Sequences

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Introduction

Fluvial terraces are important chronicles of (1) climatically driven changes in sediment supply, discharge, or base level that result in channel incision or aggradation, and (2) tectonically driven long-term uplift that results in channel incision and warping or faulting of terrace surfaces. A fluvial terrace, a relatively flat bench occurring along and above a river, is an abandoned floodplain surface that is separated from the modern floodplain (or a lower river terrace) by a steep slope, known as the terrace riser. River valleys often contain flights of multiple terraces that reflect long periods of terrace formation, a history that can be unraveled through mapping, surveying, and dating of fluvial terraces. As such, terraces are useful in studies of Quaternary climate change and the effects of climate change on landscape processes, such as increased upland erosion and the consequent increased sediment supply. In tectonic studies, terraces are useful for assessing both the rate of rock uplift at a given reach and differential rates of deformation along river valleys.

The causes and processes of terrace formation are briefly summarized and then case studies of terraces in Asia, New Zealand, and North America are reviewed. As demonstrated by these examples, both climatic and tectonic processes have influenced the formation of fluvial terraces over time periods of tens to hundreds of thousands of years during the Quaternary period. The studies cited here used various methods of absolute and relative dating, including cosmogenic nuclide exposure dating, radiocarbon dating, correlation to dated tephra and loess stratigraphies, optically stimulated luminescence (OSL) dating, and relative age dating from degree of soil development and thickness of weathering rinds on gravel clasts.

Types of Fluvial Terraces and Terrace Terminology

The two basic types of fluvial terraces are aggradational and degradational ([Figure 1](#)). Aggradation, or net deposition of sediment along a river channel and its floodplain over a prolonged period of time, leads to the formation of an alluvial fill terrace. At the crossing of a geomorphic threshold from either deposition or stability (i.e., no net deposition) to net degradation, alluvium is incised and a fill terrace is both created and abandoned ([Bull, 1991](#)). Radiocarbon dating of organic matter in alluvium from an aggradational terrace along Cajon Creek, a stream that crosses the San Andreas fault in southern California, USA, reveals that it can take as long as 10 ka for aggradation to form a fill terrace over a distance of ~10 km ([Weldon, 1986](#)). Furthermore, the time since the onset of filling, abandonment, and incision decreases upstream along Cajon Creek, indicating a time-transgressive terrace surface.

If incision of an aggradational terrace is punctuated by pauses that allow the stream to cut laterally rather than vertically, then cut terraces are carved into alluvial fill. Cut terraces (sometimes called cut-fill) have a distinctly different origin than the higher fill terrace into which they are carved. It is possible that a subsequent cycle of aggradation and degradation might begin again after some period of vertical downcutting, resulting in an inset fill terrace that is difficult to distinguish from a cut terrace. Only careful stratigraphic work and dating can help to make the distinction between the two possible origins of the terraces.

Lateral cutting of bedrock followed by vertical degradation, or incision, results in the formation of a strath terrace. When abandoned, the floodplain associated with the bedrock stream is left behind as well, leaving a veneer of coarse sediment on the strath surface that is proportional to the scour depth and bedload thickness associated with bankfull discharge (e.g., [Lavé and Avouac, 2001](#); [Pazzaglia and Brandon, 2001](#)). Whereas the coarse sediment on a strath is generally thought to have been associated with bedload transport, the finer sand, silt, and clay often capping a strath and its gravel veneer probably are deposited by a combination of overbank sedimentation during high-magnitude floods ([Merritts et al., 1994](#)) and colluvial deposition from side slopes ([Lavé and Avouac, 2000](#)).

Through extensive radiocarbon dating of strath terrace sediments, [Merritts et al. \(1994\)](#) have shown that overbank deposition of fine sediment can occur during rare, high-magnitude floods several thousand years after a strath is formed by incision and abandonment of a river's previous channel level. The duration of overbank deposition after strath formation depends on the rate of vertical incision and the stochastic distribution of floods along the river, with low rates of incision and large-magnitude floods favoring the possibility of episodic deposition on the older surface.

Causes and Processes of Fluvial Terrace Formation

Fluvial terraces are formed by one or more changes in three dominant controls on a river's longitudinal profile: base level, water discharge, and sediment supply. When one or more of these factors changes, the river adjusts its channel to the new conditions and sometimes forms a new channel bed and floodplain at a lower or higher level than the original ones. Changes in the controlling parameters of discharge, sediment supply and caliber, or base level essentially are perturbations that lead to waves of erosion or deposition that can propagate along a stream.

Changes in base level (and local gradient) occur primarily through tectonic processes, as with rock uplift, differential tilting, faulting, and folding. Changes in water discharge and sediment supply occur primarily through changes in prevailing

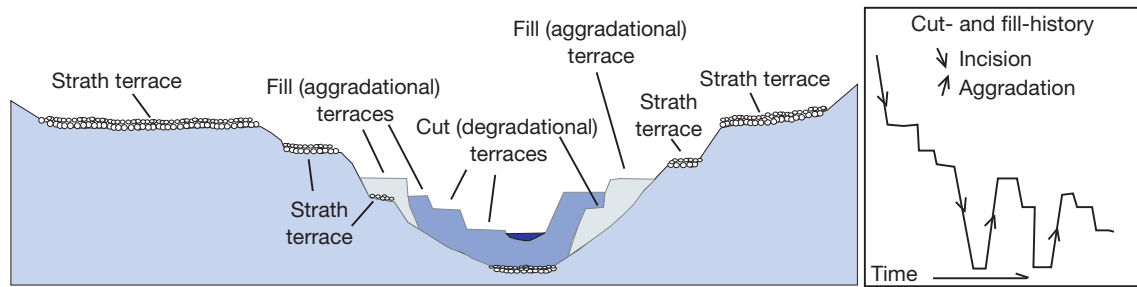


Figure 1 Valley cross-section illustrating a complex sequence of aggradational and degradational (cut and strath) terraces. The history of aggradation and degradation that led to the terrace sequence is outlined in the hand box. Reproduced from Burbank DW and Anderson RS (2001) *Tectonic Geomorphology*, p. 274, Figure 2.11C. Oxford: Blackwell Science.

climatic conditions, which affect vegetation type and cover, and rates of soil erosion and runoff of water from upland slopes. For example, when falling sea level or an uplifting land mass lowers the relative base level of a coastal river and increases the river's gradient near the coast, the river incises and abandons its former floodplain level, leaving a terrace remnant that records the earlier position and gradient of the channel and floodplain. If a cooling climate results in a decrease in vegetation cover on slopes and an increase in sediment supply that exceeds stream-carrying capacity (or perhaps a decrease in effective discharge), then aggradation might occur along the river channel and floodplain. Aggradation continues until the stream gradient is sufficiently steep to transport the prevailing load of incoming sediment. When aggradation ceases, perhaps during subsequent warming and an increase in vegetation cover, the fill surface is abandoned and incised, producing a fill terrace at a higher elevation than the original channel and floodplain.

Previous workers have recognized that a stream can adjust to carry the flux of water and sediment by altering its width, depth, roughness, and/or gradient (Leopold and Bull, 1979). We are most interested in the last of these, adjustments in channel gradient, as these lead to terrace formation. The adjustment of channel long profiles by aggradation or degradation is thought to occur because streams maintain an equilibrium long profile, or graded profile, over sufficiently long time periods (graded timescale of Schumm and Lichty (1965)). A graded stream is one that has sufficient stream power to transport sediment and water supplied from upstream, without aggrading or degrading along its bed (Bull, 1991; Knox, 1975; Leopold and Bull, 1979; Mackin, 1948). In general, long-term degradation of the channel bed and floodplain occurs if stream power is decreased (e.g., decreased discharge or gradient) or if the ability to transport sediment is reduced (e.g., increased bed roughness or grain size).

The graded stream is thought to evolve toward a steady-state profile in which rates of rock uplift and river incision are closely balanced along the length of the long profile of the stream (Bull, 1991; Burbank et al., 1996; Pazzaglia and Brandon, 2001; Pazzaglia et al., 1998). This close connection between uplift and incision implies that the difference in elevation between long profiles of streams and terraces can be used to assess rock uplift, much as marine terraces are used to assess coastal uplift.

Terraces often form during vertical incision by knickpoint, the process by which a steepened reach of channel – the knickpoint – propagates upstream (Gardner, 1983; Zaprowski

et al., 2001). As the knickpoint propagates upstream, it leaves an abandoned strath or cut surface in its wake downstream, such that the degradational terrace is graded to the crest of a knickpoint and is time transgressive in age (Merritts et al., 1994; Seidl and Dietrich, 1992). Experimental work has shown that the rate of knickpoint migration is proportional to discharge, with migration rates diminishing toward head-water reaches. Studies that utilize cosmogenic nuclide exposure dating indicate that knickpoints formed in basalt in Hawaii have migrated upstream at rates of $>2 \text{ mm year}^{-1}$ (Seidl et al., 1994), whereas those cutting metamorphic and igneous rocks in the Himalayan Mountains might be as high as 1 m year^{-1} (Burbank et al., 1996).

Strath terraces in particular have been used as geomorphic markers that record progressive displacement where rivers cross active faults and folds, and in some cases have provided fault slip and fold growth rates (Lavé and Avouac, 2001; Molnar et al., 1994; Rockwell et al., 1984). In the case of faulting that ruptures the surface, each higher terrace records greater displacement than lower terraces (Figure 2).

Case Studies of Terrace Formation

Although both climatic and tectonic processes can drive terrace formation simultaneously, the two can often be separated as they operate over different timescales of change. Long-term rock uplift along a tectonically active margin, for example, might operate at roughly uniform rates over a period of hundreds of thousands to millions of years whereas alternating glacial–interglacial cycles of the Quaternary period have cycles of $\sim 120 \text{ ka}$. While long-term rock uplift drives long-term river incision, shorter-term variations in climate and associated changes in sea level (i.e., base level) might cause episodes of aggradation or accelerated degradation that produce terraces superimposed on the terrace sequence created by the long-term uplift response (see Hancock and Anderson (2002), for a modeling approach to fluvial terrace formation in response to changing climatic conditions).

Tectonically Driven Terrace Formation

We begin the case studies of tectonically driven terrace formation with four examples of long-term incision and strath terrace formation in response to moderate to rapid rock uplift at two tectonically active plate margins: the Himalayan continent–continent collision zone between the Indian and

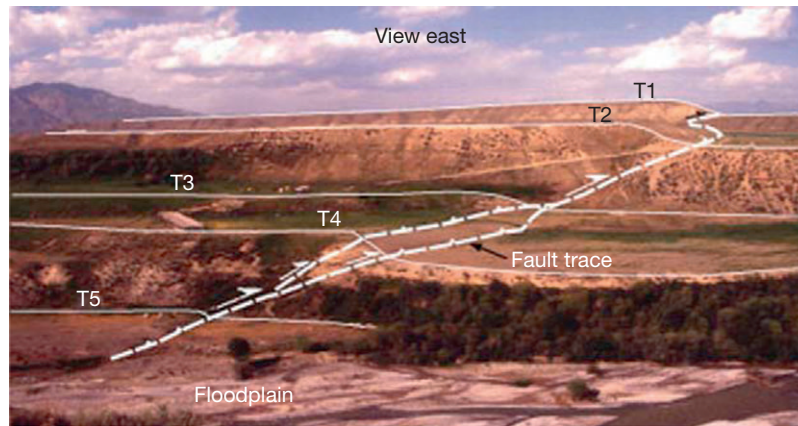


Figure 2 Offset terrace treads (T1 to T5) reveal the trace of a thrust fault (up on the left) and subsidiary splay in the Tien Shan, Kyrgyz Republic. Each higher, older terrace has displaced successively greater amounts. Deformation of such surfaces can be used to estimate rates of faulting and crustal shortening. Reproduced from Burbank DW and Anderson RS (2001) *Tectonic Geomorphology*, p. 274, Figure 9.6. Oxford: Blackwell Science.

Eurasian plates, and the ocean–continent Cascadia subduction zone boundary between the Juan de Fuca and North American plates. The fifth case study in this section is an example of strath formation in eastern North America, a low-uplift rate passive plate margin where late Quaternary climate change might be the cause of accelerated incision and strath formation superposed on long-term incision that correlates with post-Miocene falling sea level.

Long-Term Incision and Strath Formation, Tectonically Active Plate Margins

Burbank et al. (1996) surveyed and dated strath surfaces along the Indus River near Nanga Parbat in northern Pakistan, where the river crosses the suture zone between the colliding Indian and Eurasian plates in the northwestern Himalayas. Cosmogenic radionuclide exposure ages of seven abandoned strath levels, ranging in height from 65 to 410 m above the Indus River bed (low discharge), span a range in ages from 4 to 68 ka (Figure 3). The terrace ages and altitudes yield corresponding bedrock incision rates that range from 2 to 12 mm year⁻¹, some of the highest known rates of bedrock incision in the world. Apatite fission-track ages along the same river reaches generally yield similar denudation rates, indicating that incision and bedrock uplift appear to be balanced.

Early work in Nepal on rivers flowing southward across the Himalayan Mountains documented that terrace profiles on the hanging wall block above the Main Central Thrust, a south-vergent fault, are tilted down to the north and diverge strongly to the south, consistent with terrace formation above a rising block with differential uplift parallel to the river's long profile (Iwata, 1987). Lavé and Avouac (2000) mapped and radiometrically dated four warped and folded Holocene strath terraces along two rivers draining the Siwalik Hills in central Nepal along the Main Frontal Thrust fault in the foothills of the Himalayas (Figure 4). Combining structural mapping and analysis of active fault bend folding with warped profiles of the dated terraces, they derived rates of bedrock uplift from river incision (as high as 15 mm year⁻¹) and estimated that much of the shortening across the Himalayas is accommodated by the Main Frontal Thrust fault.

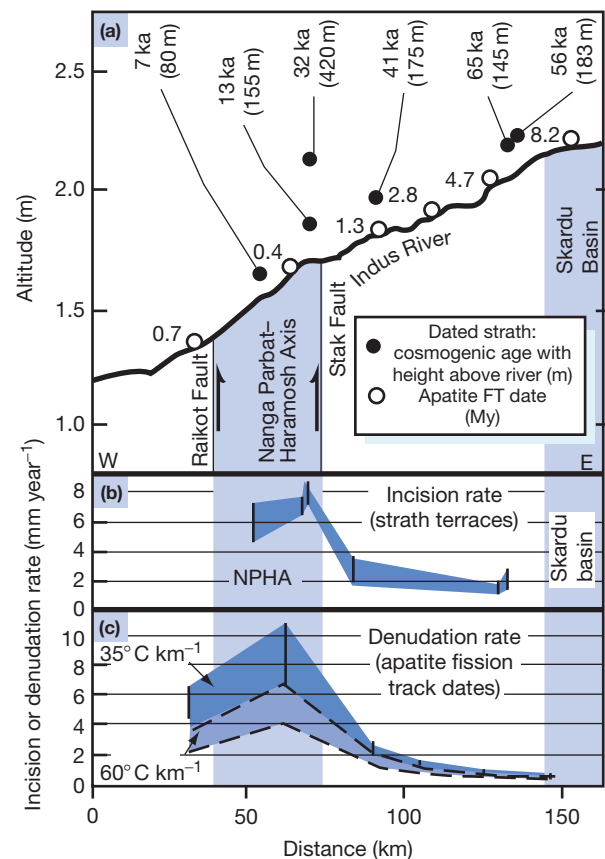


Figure 3 (a) Strath terrace remnants, with their cosmogenic nuclide exposure ages and heights above the modern Indus River level in the Himalayas of Pakistan, are shown with apatite fission-track ages from samples taken along the base of the Indus gorge. (b) Incision rates plotted versus distance along the Indus River, with incision rates calculated from cosmogenic nuclide exposure ages. Highest rates correspond with the steepest river gradient. (c) Rates of denudation based on apatite fission-track data and two different geothermal gradients (35 and 60 °C km⁻¹). The zone of maximum long-term denudation coincides with the zone of most rapid river incision. Reproduced from Burbank DW, Leland J, Fielding E, et al. (1996) Bedrock incision, rock uplift and threshold hillslopes in the northwestern Himalayas. *Nature* 379: 505–510.

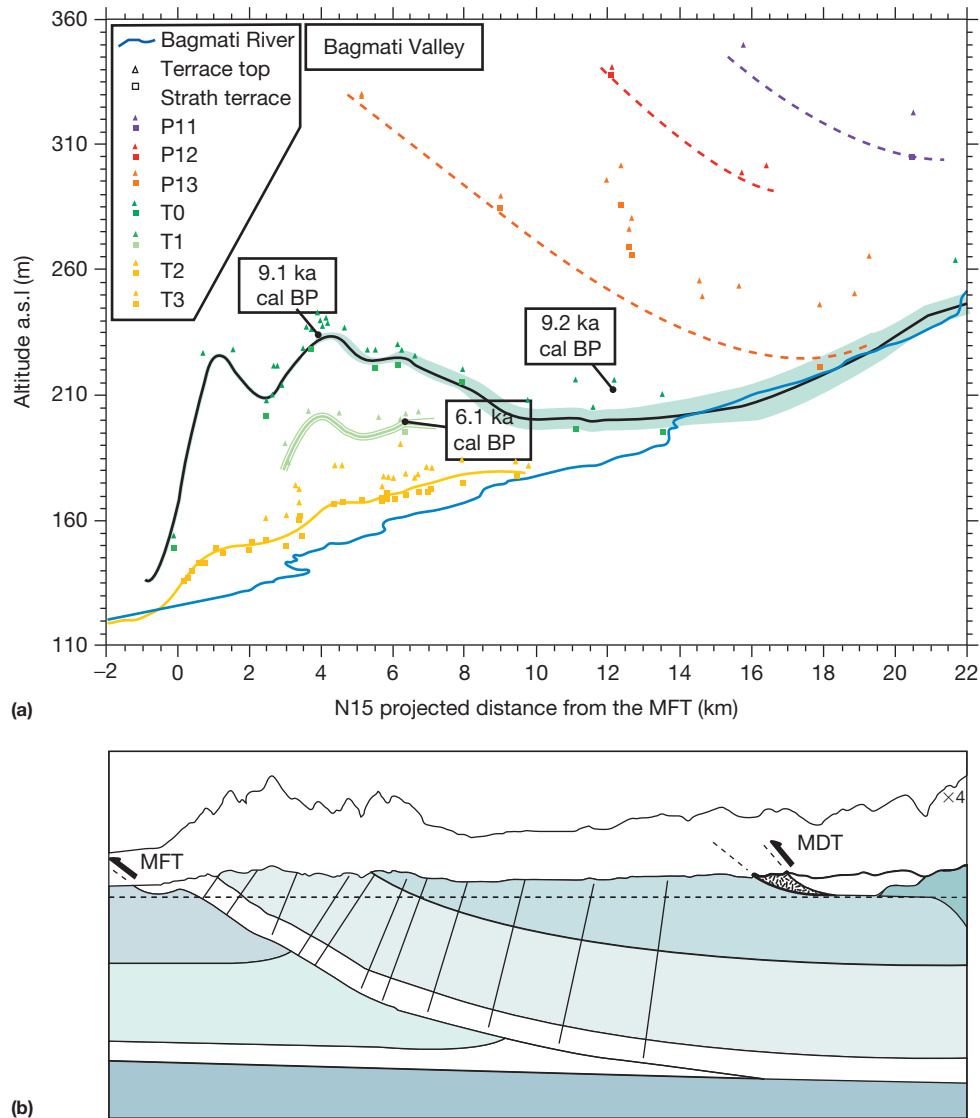


Figure 4 (a) Strath terraces projected along the Bagmati River in the central Himalayas, Nepal. Strath elevations computed from digital elevation model and field measurements (solid thin lines and dashed lines) are shown with individual field measurements. Solid thick lines represent Holocene terraces, and dashed lines represent hypothetical levels of late Pleistocene terraces that have similar soil color and depth of weathering. Note that terraces are folded above a growing thrust fault at 0–14 km. (b) Topographic and structural cross-sections, with the same horizontal scale, are shown for comparison with the terrace profiles. The terraces reveal sustained Holocene and late Pleistocene activity of the fold growing above the Main Frontal Thrust fault (MFT). In contrast, the terraces indicate that little offset has occurred across the Main Dun Thrust fault (MDT) during the late Quaternary. Reproduced from Lavé J and Avouac JP (2000) Active folding of fluvial terraces across the Siwaliks Hills, Himalayas of central Nepal. *Journal of Geophysical Research B* 105(3): 5735–5770, Plate 7A.

Similar to the work of Personius (1995) in Oregon, USA, and Merritts et al. (1994) (discussed below) in northern California, USA, Lavé and Avouac (2000) obtained much younger radiocarbon dates from silty sediments at the tops of terrace deposits than from underlying gravel veneers resting directly on bedrock straths. Corresponding with the above mentioned US research, the authors concluded that radiocarbon dates from the base of the silt or from within the gravel (bedload) itself are better estimates of the timing of strath abandonment, and attribute the overlying silty sediments to postabandonment depositional processes.

Pazzaglia and Brandon (2001) studied six late Quaternary strath terraces along the Clearwater River in northwestern Washington State, USA, in order to compare long-term uplift and incision rates across the Cascadia forearc high formed inboard of the Cascadia subduction zone. Three approaches were used to constrain terrace ages: weathering rind thickness on graywacke gravel clasts; radiocarbon dating; and correlation to dated coastal glaciofluvial deposits and the global eustatic sea-level curve. Strath terraces formed in the dominantly unglaciated Clearwater drainage while alpine glaciers were advancing in adjacent drainages. During the transition to interglacial

times, an increased sediment supply due to local deglaciation, combined with rising base level at the coast, resulted in aggradation along the Clearwater River. Nevertheless, the river returned to the same valley profile during each cycle of strath-cutting. Bedrock incision rates increase inland from $<0.1 \text{ m ka}^{-1}$ at the coast to $\sim 0.9 \text{ m ka}^{-1}$ near the headwaters in the Olympic Mountains, and are similar to long-term denudation rates estimated from fission-track cooling ages. As incision keeps pace with rock uplift, terraces diverge upstream. Pazzaglia and Brandon (2001) concluded that the Clearwater River valley has maintained a steady-state profile and a long-term balance between incision and rock uplift over a period of tens of thousands of years.

In a region of relatively high uplift rates at the southernmost tip of the Cascadia subduction zone, where the North American, Pacific, and Juan de Fuca plates join at the Mendocino triple junction, Merritts et al. (1994) surveyed and radiometrically dated late Pleistocene and Holocene fill and strath terraces along three rivers in northern California. This work determined that lower reaches of the rivers are dominated by the effects of oscillating sea level, primarily aggradation and formation of fill terraces during sea-level high stands, and deep incision during sea-level low stands (Figure 5). A eustasy-driven depositional

wedge that tapers to zero thickness extends tens of kilometers upstream on all three rivers. In contrast, middle to upper reaches of each river are dominated by the effects of long-term uplift, primarily lateral and vertical erosion and formation of steep, unpaired strath terraces exposed only upstream of the depositional wedge. Vertical incision at a rate similar to that of uplift ($1\text{--}2 \text{ m ka}^{-1}$) has occurred even during the present sea-level high stand along the two rivers with highest uplift rates. Strath terraces have steeper gradients than the modern channel bed and do not merge with marine terraces at the river mouths.

Long- and Short-Term Incision and Strath Formation, North American Passive Plate Margin

Three ongoing processes have produced a relative, long-term base-level fall at the mouth of the Susquehanna River along the eastern North American passive plate margin: (1) flexural up-warping of the Appalachian Piedmont in response to denudational unloading in the Susquehanna watershed, (2) downward flexural warping offshore in response to sediment deposition on the continental shelf, and (3) protracted eustatic sea-level fall beginning in the middle Miocene

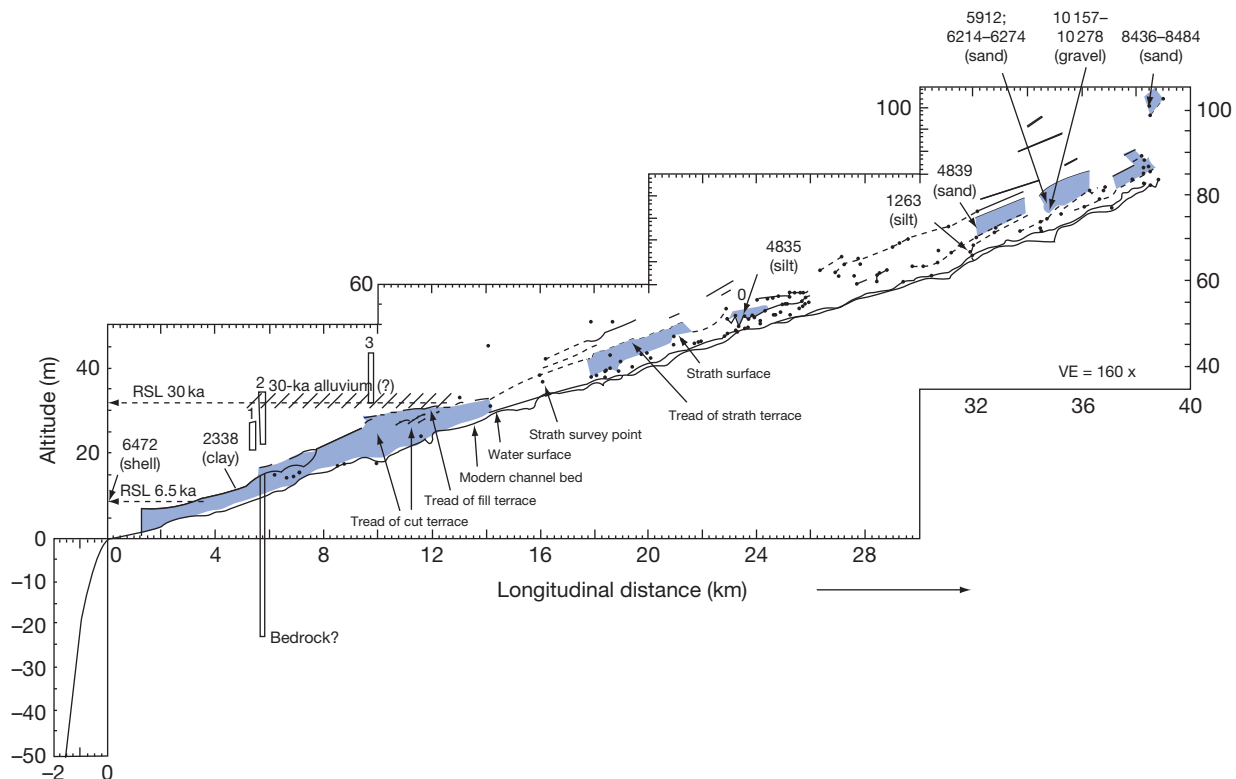


Figure 5 Longitudinal profiles of channel bed, water surface, strath surfaces, and treads of fill terraces from detailed total geodetic station along the Mattole River, northern California. Treads of cut-and-fill terraces are indicated by fine lines (hundreds of survey points not shown), dashed where correlations are tentative, while strath surfaces are indicated by both heavy lines, dashed where correlations are tentative, and survey points (crosses). Stippled pattern downstream of 18 km indicates alluvium where depth can be estimated from well data (drill holes shown as 1, 2, and 3). Multiple unpaired strath terraces occur upstream of 18 km, shown by a bedrock surface (heavy line) and overlying alluvial tread (fine line). Deposits are shown as stippled pattern at several sites to indicate corresponding straths and treads, but not all are included because straths and treads overlap one another. Radiocarbon dates are centroids of calibrated age ranges in years before present. Relative sea level (RSL) is determined from both sea-level change and landmass uplift. Reproduced from Merritts DJ, Vincent KR, and Wohl EE (1994) Long river profiles, tectonism and eustasy: A guide to interpreting fluvial terraces. *Journal of Geophysical Research* 99: 14031–14050.

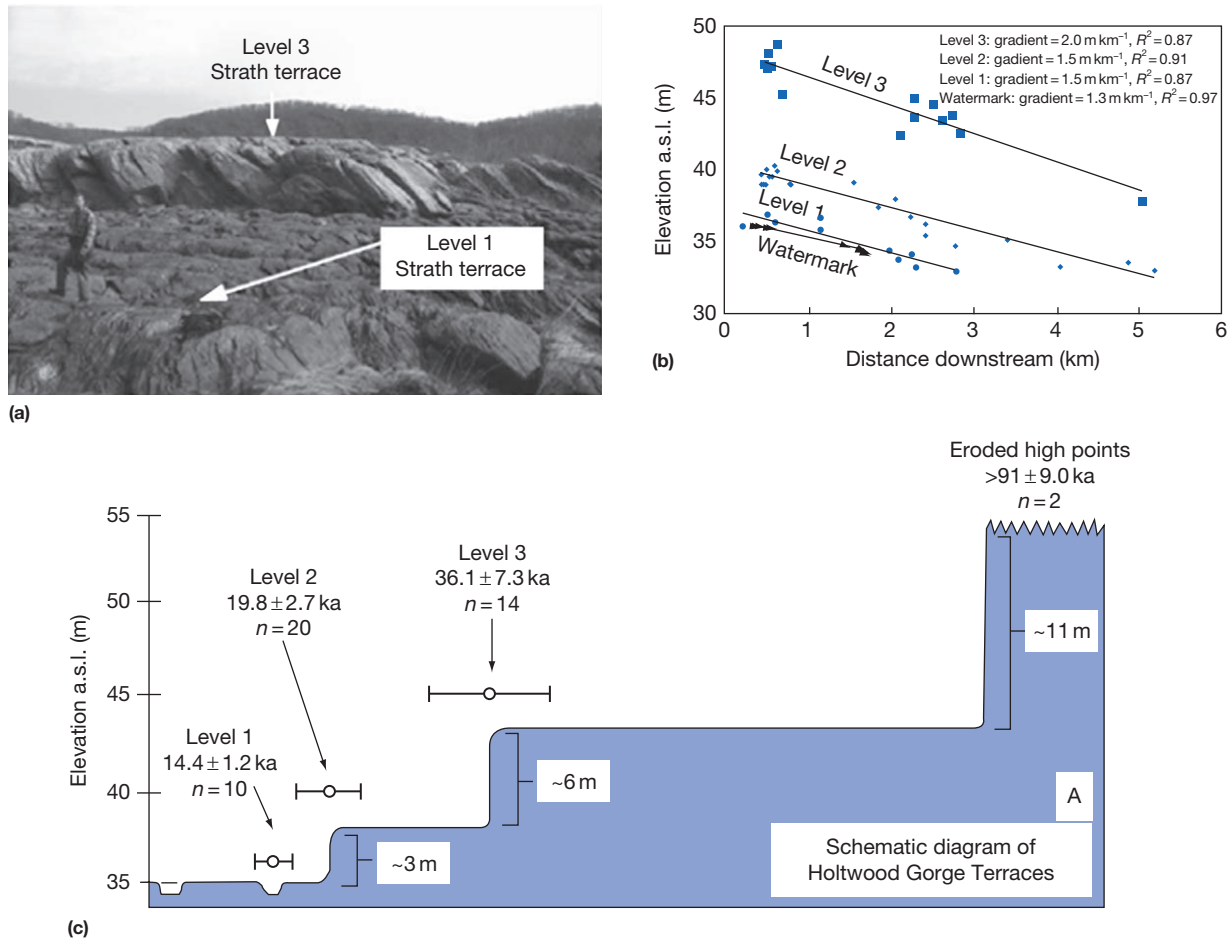


Figure 6 (a) Strath terrace levels 1 and 3 in the Holtwood Gorge along the Susquehanna River at low flow conditions. Level 2 is not preserved at this location. At higher discharges, strath level 1 is submerged. (b) Paleogradients for strath terrace levels 1, 2, and 3 and a prominent watermark in July 2002, were acquired with differential global positioning system (GPS). The watermark is representative of the modern river gradient. (c) The timing of incision and terrace abandonment within Holtwood Gorge is shown diagrammatically with respect to time (x-axis) and in relation to documented changes in climate and sea level. Reprinted from Reusser LJ, Bierman P, Pavich M, Larsen J, and Finkel R (2006) late Pleistocene bedrock channel incision along the U.S. Atlantic passive margin, Holtwood Gorge, Susquehanna River, Pennsylvania. *American Journal of Science*, Figures 4A, 5, and 17A, with permission from the American Journal of Science.

(Pazzaglia and Gardner, 1993; Pazzaglia et al., 1998). The unglaciated, lower Susquehanna River has a convex upward long profile with steep gradients as a result of these processes. Recent cosmogenic exposure dating of three low strath terraces (referred to as levels 1–3) within the Holtwood Gorge along the lower Susquehanna River indicates that rates of vertical incision dramatically increased between ~ 35 and $\sim 14 \text{ ka}$ (Figure 6). Prior to $\sim 35 \text{ ka}$, maximum vertical incision was $\sim 0.2 \text{ m ka}^{-1}$, but from ~ 35 to $\sim 20 \text{ ka}$ the rate of incision accelerated to 0.45 m ka^{-1} . The rate of incision increased again to $\sim 0.52 \text{ m ka}^{-1}$ between ~ 20 and $\sim 14 \text{ ka}$, when significant incision ceased. An upstream-younging age gradient of $\sim 1.4 \text{ ka km}^{-1}$ on the level 2 terrace is consistent with knick-point retreat, but no significant time-transgressive relationship exists on either the highest (level 3) or lowest (level 1) terraces, suggesting that these surfaces were abandoned rapidly along the length of the Holtwood Gorge.

Reusser et al. (2004) interpret the high rates of late Pleistocene incision to be the result of a prolonged period of climate instability from ~ 45 to 14 ka , and the cessation of rapid

incision to correspond with the transition from Pleistocene full-glacial conditions to the warmer, more climatically stable Holocene Epoch. Using climate data for the North Atlantic region, they note that a marked increase in storminess beginning at $\sim 50 \text{ ka}$ and lasting until $\sim 10 \text{ ka}$ correlates well with the initiation and cessation of incision within Holtwood Gorge. Furthermore, they suggest that increases in the frequency, magnitude, and duration of flood events that are capable of exceeding critical geomorphic thresholds for detachment of rock might have increased rates of bedrock channel incision during the last glacial cycle. Because of the timing of increased incision, they conclude that climate is likely to be a more important external factor than the low sea level and base-level fall associated with full-glacial episodes.

Climatically Driven Terrace Formation

The previous example indicates that incision driven by long-term, slow uplift and relative sea-level fall can be accelerated

during cold conditions, as during full-glacial episodes, but many studies in other areas have documented a switch from incision to aggradation during previous glacial periods. [Blum and Tornqvist \(2000\)](#) have reviewed this mismatch between the role of climate and sea level in aggradation and degradation of coastal fluvial systems. Paired aggradational terraces throughout the world have been associated with Quaternary glacial-interglacial cycles, with aggradation commonly thought to have occurred during glacial periods and incision during subsequent warmer periods (cf. [Bridgland, 2000](#); [Penck and Brückner, 1999](#)). In some cases, fill terraces were correlated with glacial moraines and interpreted to be of similar age regionally, so they were correlated with one another among basins, even on unglaciated rivers. Recent work on nonglaciated streams in New Zealand supports the premise of these earlier studies,

and is described here as an example of regional aggradational terrace formation driven by climatic change. Other examples of aggradational river terraces documented to have been formed by Quaternary climate change in nonglaciated regions are described in [Bull \(1991\)](#).

Terrace Formation during Quaternary Glacial–Interglacial Cycles

Along the actively uplifting Hikurangi Margin along the eastern side of the North Island in New Zealand, [Litchfield and Berryman \(2005\)](#) and [Litchfield and Rieser \(2005\)](#) extended the results of previous studies to correlate four aggradational terraces among eight nonglaciated drainage basins ([Figure 7](#)). The work of [Litchfield and Rieser \(2005\)](#) represents one of the

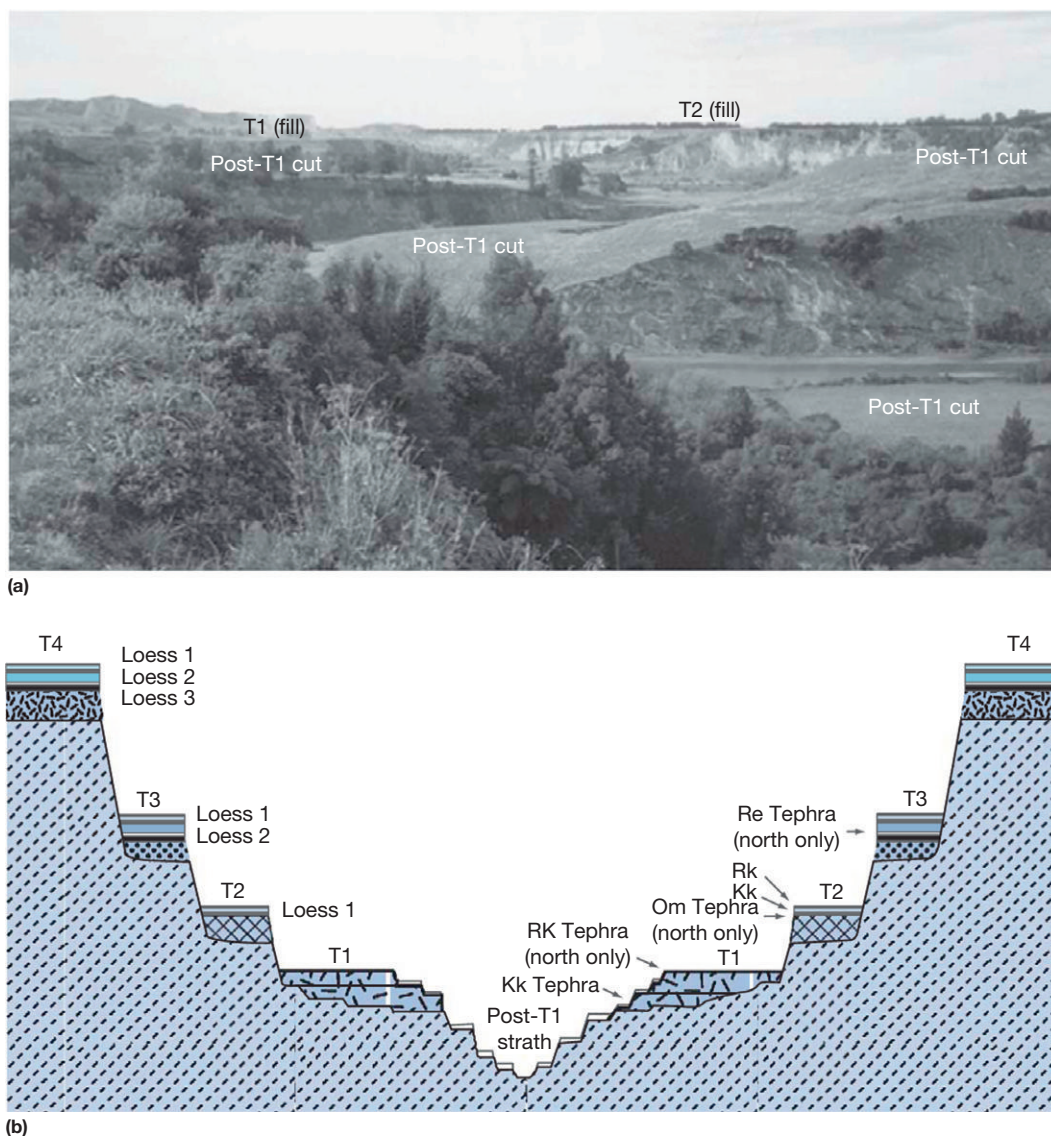


Figure 7 (a) A suite of fill (T1 and T2) and cut terraces along a river on the North Island of New Zealand. The gravel of T2 is visible above a strath cut into early Pleistocene siltstone bedrock (white). (b) Schematic cross-section illustrating the relations among the terraces and overlying loess and tephra coveredbeds. Reproduced from Litchfield NJ and Berryman KR (2005) Correlation of fluvial terraces within the Hikurangi Margin, New Zealand: Implications for climate and base-level controls. *Geomorphology* 68: 291–313, Figure 3. Copyright 2005 with permission from Elsevier.

most extensive analyses of the use of OSL dating for river terraces. OSL dating measures the time that has elapsed since sediment was last exposed to light, presumably at the time of deposition. [Litchfield and Rieser \(2005\)](#) used it for both silts and sands near the tops of aggradational terraces, and for loess coverbeds overlying the terraces.

Based on consistent results from tephra chronology, loess stratigraphy, radiocarbon dating, and OSL dating, terrace ages are linked to the following late Pleistocene dry, cool stadial, and/or glacial periods: 15–30, 31–40, 55–70, and 115 ka. These results confirm that aggradation and loess deposition were widespread on the North Island of New Zealand during both full glacial and interstadial periods. As each aggradational terrace formed, the riverbed was the source of loess blown onto higher terraces, so that each higher, older terrace has one more loess cover deposit than the next lower aggradational terrace. [Litchfield and Berryman \(2005\)](#) conclude that limited grass and shrub vegetation cover, combined with periglacial processes, produces an increased sediment supply that led to the widespread aggradation associated with cool periods, whereas reestablishment of forested slopes during postglacial periods led to incision of fill deposits, producing abandoned fill terraces.

Since the formation of the lowest aggradational terrace, T1 at 15–30 ka, when global eustatic sea level was at its lowest (~ 120 m), North Island streams have been degrading and carving strath terraces into bedrock, despite a rising Holocene sea level. The intriguing conclusion in this region is that climatic controls on sediment and water discharge are more significant to terrace formation than rising or falling base levels.

Cycles of Aggradation and Degradation in Arroyos of the Semiarid American Southwest

Currently incised channel systems in the semiarid, nonglaciated southwestern United States, known as arroyos (Spanish for gulch), have repeatedly aggraded and incised during the Quaternary period, leaving a complex record of cut-and-fill terraces (cf. [Bull, 1991](#); [Cooke and Reeves, 1976](#)). Modern arroyos currently are in a phase of cutting that dates to the nineteenth century, possibly due to grazing and other recent land-use changes, and carry large volumes of sediment during ephemeral to intermittent flows ([Cooke and Reeves, 1976](#)). During repeated episodes throughout the Quaternary period, however, arroyos have been unable to transport their sediment load and have aggraded, forming fill terraces.

Decades of investigations of arroyos have led many workers to attribute the cyclicity of arroyo filling and incision to alternations between relatively wet, pluvial climates during glacial periods and drier climates during interglacial and interstadial periods. Wetter conditions are thought to be associated with production of greater sediment yields from upland slopes (e.g., [Bull, 1991](#)). However, full understanding of the 'arroyo cycle' is complicated by the fact that repeated cycles of cutting and filling have occurred during the Holocene interglacial warm period ([Cooke and Reeves, 1976](#)). Some regional forcing mechanism is thought to be responsible for generally synchronous episodes of cutting and filling that extend over broad areas. Some workers have suggested that complex interactions

within arroyo systems, as well as the initial state of the arroyo system at the onset of climatic change, are critical to the response of arroyos to climatic forcing (cf. [Schumm et al., 1984](#)).

Conclusions

Recent studies of long profiles of rivers and fluvial terraces have illuminated the relation between crustal and surficial processes of uplift and incision, respectively. As described in four examples herein, researchers used altitudinal differences between dated straths and the modern channel to estimate rates of vertical incision at a given stream reach, and then compared these estimates with rates of rock uplift or denudation that were derived independently ([Burbank et al., 1996](#); [Lavé and Avouac, 2001](#); [Merritts et al., 1994](#); [Pazzaglia and Brandon, 2001](#)). In all four cases, the rates of uplift and incision are roughly balanced, supporting earlier ideas of steady-state long profiles for graded streams that are adjusted to long-term uplift. While these four examples were in areas of relatively high rates of uplift along tectonically active plate margins, a recent study along the eastern North American passive margin, on the lower Susquehanna River, provides clues to the effect of climate change on river incision. Cosmogenic exposure dating of strath surfaces indicates that relatively short-term climatic change which leads to colder and possibly stormier conditions can result in pulses of accelerated incision, even where long-term uplift rates are low.

Although the Susquehanna River example links glacial climatic conditions to accelerated bedrock incision and strath abandonment, numerous studies in Europe, North America, and New Zealand have documented a switch from incision to aggradation during glacial conditions. [Merritts et al. \(1994\)](#) and [Pazzaglia and Brandon \(2001\)](#), in contrast, describe aggradation during Late Glacial and interglacial times along lower reaches of the coastal rivers that they studied. [Blum and Tornqvist \(2000\)](#) have reviewed this mismatch between the role of climate and sea level (i.e., base-level change) in aggradation and degradation within coastal fluvial systems. One of the best constrained examples, with extensive dating using multiple techniques that include radiocarbon dating, tephra chronology, loess chronology, and OSL dating, is described here for North Island, New Zealand ([Litchfield and Berryman, 2005](#); [Litchfield and Rieser, 2005](#)). This work demonstrates conclusively that regional aggradation occurred along coastal rivers during full-glacial times, regardless of a markedly lower sea level, and that the Holocene is marked by incision and the formation of cut and strath terraces.

As with the case of the cycles of arroyo cutting and filling described here, it might be that each river system responds differently to climate change, depending on its initial conditions at the onset of external change and on the magnitude of change in sediment supply from upland slopes. Where relief is high and slopes are steep, as in New Zealand, climate cooling and periglacial conditions seem to overwhelm streams with sediment, whereas the low-uplift Susquehanna River basin, with generally low relief and gentle upland slopes, does not generate sufficient excess sediment to cause aggradation along major trunk streams during cold periods. New techniques of dating terraces and terrace deposits, and multiple new sites

studied around the world, are providing a growing body of evidence that fluvial responses to climate change might be similar regionally, but have significant differences among regions.

See also: Glacial–Interglacial Scale Fluvial Responses. **Fluvial Environments:** Responses to Rapid Environmental Change; Sediments. **Luminescence Dating:** Optical Dating.

References

- Blum MD and Tornqvist TE (2000) Fluvial responses to climate and sea-level change: A review and look forward. *Sedimentology* 47(supplement 1): 2–48.
- Bridgland DR (2000) River terrace systems in north-west Europe: An archive of environmental change, uplift and early human occupation. *Quaternary Science Reviews* 19: 1293–1303.
- Bull WB (1991) *Geomorphic Responses to Climate Change*. New York: Oxford University Press.
- Burbank DW, Leland J, Fielding E, et al. (1996) Bedrock incision: Rock uplift and threshold hillslopes in the northwestern Himalayas. *Nature* 379: 505–510.
- Cooke RU and Reeves RW (1976) *Arroyos and Environmental Change in the American South-West*, p. 213. Oxford: Clarendon Press.
- Gardner TW (1983) Experimental study of knickpoint and longitudinal profile evolution in cohesive, homogenous material. *Geological Society of America Bulletin* 94: 664–672.
- Hancock GS and Anderson RS (2002) Numerical modeling of fluvial strath-terrace formation in response to oscillating climate. *Geological Society of America Bulletin* 114: 1131–1142.
- Iwata S (1987) Mode and rate of uplift of the central Nepal Himalaya. *Zeitschrift für Geomorphologie v. Supp. Bd.* 63: 972–984.
- Knox JC (1975) Concept of the graded stream. In: Melhorn WN and Flemal RC (eds.) *Theories of Landform Development. Proceedings of the 6th Geomorphology Symposium*, pp. 169–198. Binghamton: State University of New York at Binghamton. Publications in Geomorphology.
- Lavé J and Avouac JP (2000) Active folding of fluvial terraces across the Siwaliks Hills, Himalayas of central Nepal. *Journal of Geophysical Research B* 105(3): 5735–5770.
- Lavé J and Avouac JP (2001) Fluvial incision and tectonic uplift across the Himalayas of central Nepal. *Journal of Geophysical Research* 106: 26,561–26,591.
- Leopold LB and Bull WB (1979) Base level aggradation and grade. *Proceedings of the American Philosophical Society* 123: 168–202.
- Litchfield NJ and Berryman KR (2005) Correlation of fluvial terraces within the Hikurangi Margin, New Zealand: Implications for climate and base-level controls. *Geomorphology* 68: 291–313.
- Litchfield NJ and Rieser U (2005) Optically stimulated luminescence age constraints for fluvial aggradation terraces and loess in the eastern North Island, New Zealand. *New Zealand Journal of Geology and Geophysics* 48: 581–589.
- Mackin JH (1948) Concept of the graded river. *Geological Society of America Bulletin* 59: 463–512.
- Merritts DJ, Vincent KR, and Wohl EE (1994) Long river profiles, tectonism and eustasy: A guide to interpreting fluvial terraces. *Journal of Geophysical Research* 99: 14031–14050.
- Molnar P, Brown ET, Burchfiel BC, et al. (1994) Quaternary climate change and the formation of river terraces across growing anticlines on the north flank of the Tien Shan, China. *Journal of Geology* 102: 583–602.
- Pazzaglia F and Brandon M (2001) A fluvial record of long-term steady-state uplift and erosion across the Cascadia forearc high, western Washington State. *American Journal of Science* 301: 385–431.
- Pazzaglia F and Gardner T (1993) Fluvial terraces of the lower Susquehanna River. *Geomorphology* 8: 83–113.
- Pazzaglia FJ, Gardner TW, and Merritts DJ (1998) Bedrock fluvial incision and longitudinal profile development over geologic time scales determined by fluvial terraces. In: Tinkler KJ and Wohl EE (eds.) *Rivers over Rock: Fluvial Processes in Bedrock Channels, American Geophysical Union Monograph*, vol. 107, pp. 207–235. Washington, DC: American Geophysical Union.
- Penck W and Brückner E (1909) *Die Alpen im Eiszeitalter*, p. 1199. Leipzig: Tauchnitz.
- Personius SF (1995) Late Quaternary stream incision and uplift in the forearc of the Cascadia subduction zone, western Oregon. *Journal of Geophysical Research B* 100: 20193–20210.
- Reusser LJ, Bierman P, Pavich M, Zen E-A, Larsen J, and Finkel R (2004) Rapid late Pleistocene incision of Atlantic passive margin river gorges. *Science* 305: 499–502.
- Rockwell T, Keller EA, Clark MN, and Johnson DL (1984) Chronology and rates of faulting of Ventura River terraces, California. *Geological Society of America Bulletin* 95: 1466–1474.
- Schumm SA, Harvey MD, and Watson CC (1984) *Incised Channels: Morphology, Dynamics and Control*, p. 200. Littleton: Water Resources Publications.
- Schumm SA and Lichty RW (1965) Time, space, and causality in geomorphology. *American Journal of Science* 263: 110–119.
- Seidl MA and Dietrich WE (1992) The problem of channel erosion into bedrock. *Catena Supplement* 23: 101–124.
- Seidl MA, Dietrich WE, and Kirchner JW (1994) Longitudinal profile development into bedrock: An analysis of Hawaiian channels. *Journal of Geology* 102: 457–474.
- Weldon RJ (1986) *Late Cenozoic Geology of Cajon Pass: Implications for Tectonics and Sedimentation Along the San Andreas Fault*. Pasadena: California Institute of Technology PhD Dissertation.
- Zaprowski BJ, Evenson EB, Pazzaglia FJ, and Epstein JB (2001) Knickzone propagation in the Black Hills and northern High Plains: A different perspective on the late Cenozoic exhumation of the Laramide Rocky Mountains. *Geology* 29: 547–550.

Deltaic Environments

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Introduction

Definition

Deltas are constructional coastal landforms with both subaerial and subaqueous components that are genetically associated with rivers discharging into a standing body of water such as a lake, estuary, lagoon, sea, or the open-ocean shelf (**Figure 1**). A delta is usually built by a single river, but exceptions exist (e.g., Ganges–Brahmaputra; Tigris–Euphrates). Reworking of sediments accumulated at the river mouth by basinal processes (e.g., waves, tides, and currents) should be slow enough to allow delta building to proceed. The river is the main source for sediment delivered to the delta, although in some wave-dominated settings (e.g., Danube, Sao Francisco, and Rio Doce), a significant portion may be transported by wave-driven currents from remote sources (**Figure 2**).

Importance and Significance of Deltas

Deltas are among the most environmentally and economically important coastal sedimentary environments. Significant oil, gas, and groundwater resources have been exploited from deltaic formations. Historically, modern deltas have been prized by human civilizations for their high natural and agricultural productivity, rich biodiversity, and for the abundance of waterways, which provide easy means of transportation. Deltas are ubiquitous: 21 of the world's 25 largest rivers, which deliver 31% of the total fluvial sediment reaching the ocean, have formed well-expressed deltas at the coast (**Meade, 1996**). As a result, ~25% of the world's population lives within deltaic and wetland coastal systems (**Syvitski et al., 2005**), some in large urban centers such as Shanghai, Bangkok, Dhaka, Saigon, Rangoon, Lagos, Alexandria, Cairo, and New Orleans. River deltas act as filters, repositories, and reactors for a suite of continental materials including sediments, organic carbon, nutrients, and pollutants, significantly affecting both the regional environment at the continent–ocean boundary as well as global biogeochemical cycles (e.g., **McKee et al., 2004**). However, deltas are fragile geomorphic features that can change dramatically with modest modifications in their boundary conditions. Currently, the largest threat to the world's deltaic systems is the trapping of sediment behind dams erected on delta-building rivers and their tributaries. Deltas also face the threat of accelerated relative sea-level rise due to climate change and fluid extraction.

History of Delta Research

The term 'delta' was introduced by Herodotus (fifth century BC), by analogy to the Greek letter Δ, to describe the shape of the lowlands between the forking distributaries of the river Nile in Egypt (**Figure 1(g)**). In 1832, in his *Principles of Geology*, Charles Lyell first defined the modern geological meaning of

the term 'delta' as 'an alluvial land, formed by a river at its mouth, without reference to its precise shape.' The first comprehensive review of modern deltas was written by Credner in 1878, wherein the author correctly identified many of the processes affecting the form, structure, and evolution of deltas. Early landmark papers on deltas also include those of **Gilbert (1885)**, **Fisk et al. (1954)**, and **Wright and Coleman (1973)**, among others.

In the 1980s, research on deltas moved from the development of depositional models based on modern highstand deltas, which are ultimately controlled by the delta-building rivers, to sequence stratigraphic interpretations of deltaic evolution through multiple sea-level cycles (e.g., **Van Wagoner et al., 1988**). More recent research on modern deltas has been spurred by concerns related to human influences on the larger riverine systems, either directly through damming and water consumption, or indirectly via climate changes. Historical stages in deltaic research can be followed in successive comprehensive volumes edited by **Morgan (1970)**, **Broussard (1975)**, **Oti and Postma (1995)**, and **Giosan and Bhattacharya (2005)**, as well as periodic review papers such as **Bhattacharya and Walker (1992)**.

River Deltas

Controls and Processes

The stratigraphic architecture of deltas is the end result of complex interactions among upstream catchment processes that regulate the location and magnitude of the fluvial sediment discharge, and downstream basinal controls that include the shape and quantity of accommodation space for sediment accumulation and the type and energy of coastal processes that redistribute these sediments. Together, all of these controls determine the mode and degree of partitioning of sediment between the delta and the wider receiving basin (e.g., **Wright and Coleman, 1973**).

The sediment discharge of a river is determined by (1) climate that regulates the precipitation–evaporation characteristics of the watershed, and thus water flow; (2) the nature and abundance of vegetation that may inhibit erosion; and (3) the tectonic regime and lithology that determines elevation, slope, and degree of erodability of rocks in the watershed. Geology also controls the water gained or lost by a river as groundwater. Accommodation space in the receiving basin is dependent on (1) the relief characteristics of the basin, such as the extent of the shallow sector of the basin (or shelf) and (2) the degree of connection to the open ocean (e.g., a closed lake, indented estuary, or a straight coast). Such basin characteristics may also change drastically as variations in base level expose more or less of the margin. Base-level changes during the Quaternary were predominantly driven by the growth and

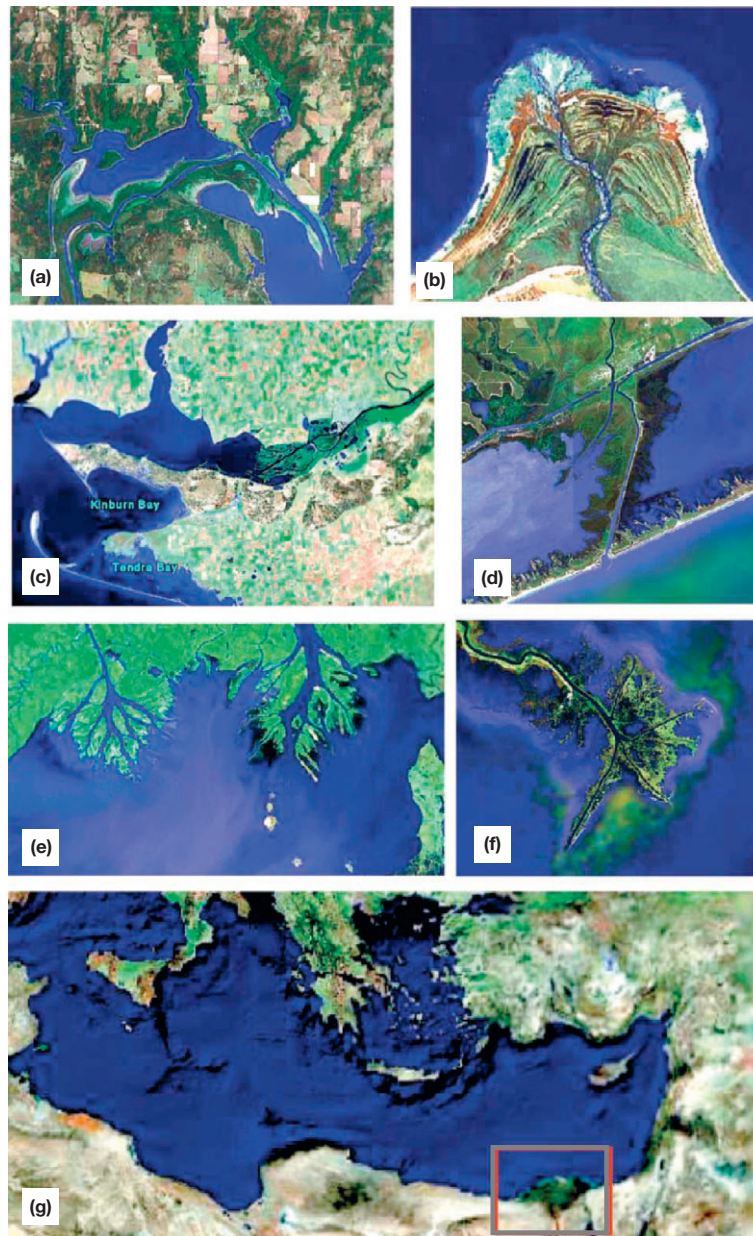


Figure 1 Google Earth satellite images of deltas (<http://earth.google.com/>). Lacustrine deltas: (a) Red River delta in Lake Texoma (Texas, USA; image © DigitalGlobe); (b) William River delta in Lake Athabasca (Canada; image © DigitalGlobe and MDAEarthSat). Deltas in estuaries, lagoons, and bays: (c) bayhead delta in the Southern Bug estuary in the Black Sea (Ukraine; image © MDAEarthSat and DigitalGlobe); (d) Colorado River delta in the Matagorda lagoon (Texas, USA; image © DigitalGlobe); (e) Wax Lake and Atchafalaya deltas in the Atchafalaya Bay (Louisiana, USA; image © MDAEarthSat and DigitalGlobe). The open coast, fluvial dominated (f) Balize lobe of the Mississippi delta (image © MDAEarthSat and DigitalGlobe); and (g) the extensive, wave-dominated Nile delta in the eastern Mediterranean Sea (image © MDAEarthSat).

decay of continental ice sheets, but in the case of enclosed basins such as the Black Sea and Caspian Sea, major base-level changes were forced by climatically driven fluctuations in their water budget. Additional components of relative base-level change are caused by tectonics, compaction, and isostasy.

Sediment in suspension is discharged as a plume at the river mouth, which is the single most important location for partitioning of sediment between the delta and the basin. Some rivers also transport a significant fraction of coarse bedload sands to the coast, where these less mobile sediments are

preferentially stored within the delta edifice. As the river plume enters the receiving basin, flow expansion and deceleration exert a strong control on river-mouth sedimentation, which is dependent on the outflow inertia, turbulent bed friction, and plume buoyancy. The latter may be positively, neutrally, or negatively buoyant, depending on the density contrast between the plume and basin waters. Waves and tides are also important in controlling deposition at a river mouth, especially in high-energy coastal settings. Tidal and wind-driven currents, as well as Coriolis rotation, may strongly

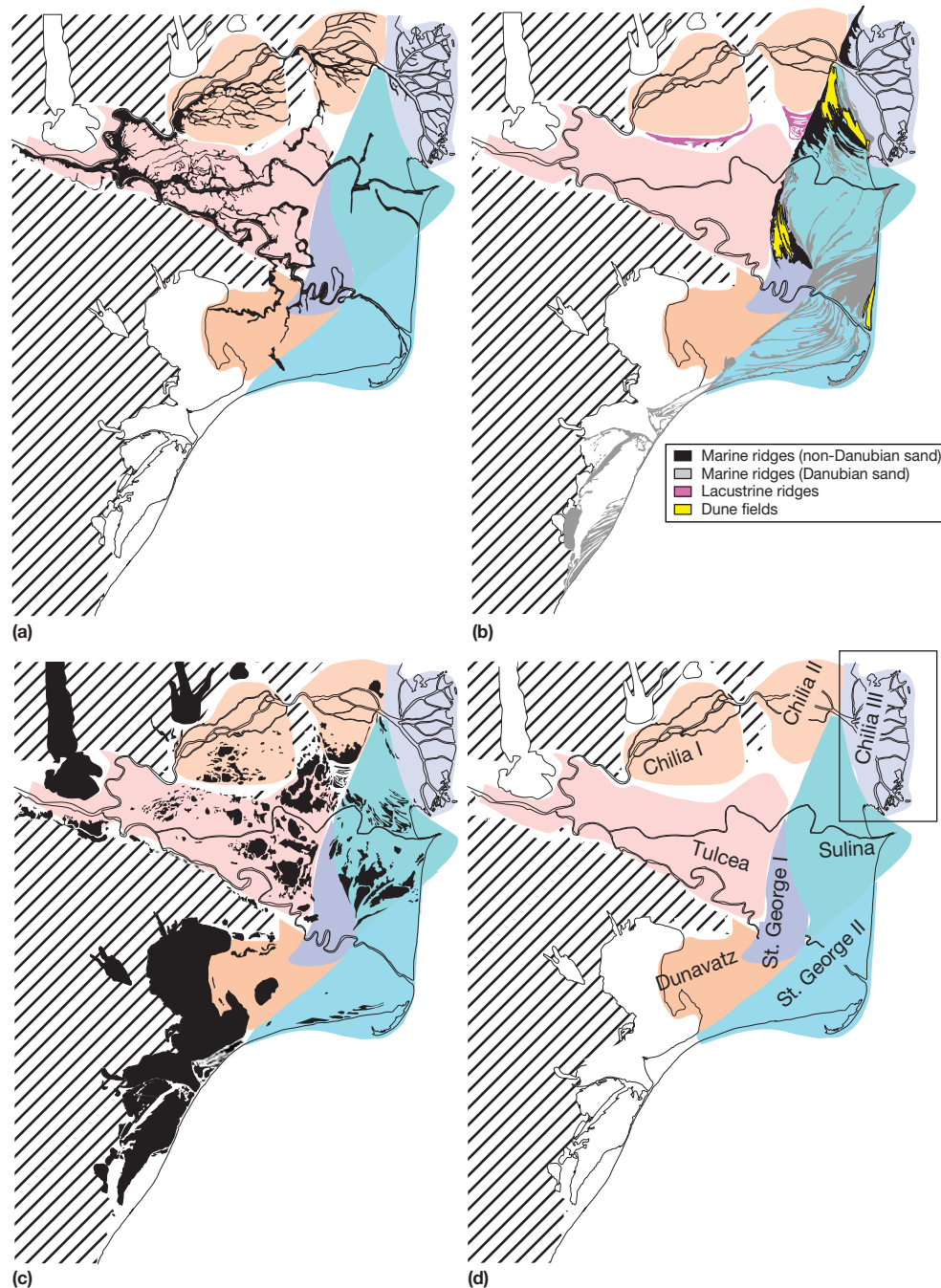


Figure 2 Morphology and depositional environments of the Danube delta in the Black Sea: (a) channels and natural levees; (b) sandy beach ridges; (c) floodbasins, lakes, and lagoons; and (d) delta lobes (the box indicates the location of Chilia III lobe described later in [Figure 5](#)). The delta started as a bayhead delta (Tulcea lobe), followed by the development of other fluvial-dominated lobes in shallow lakes, and lagoons left unfilled by the Tulcea lobe (Chilia I and II; Dunavatz). Once reaching the open coast, several wave-dominated (St. George I and II and Sulina) or fluvial-dominated (Chilia III) lobes formed. Note that the sand comprising the beach-ridge plains built updrift of open coast lobes was delivered via the longshore drift system. Reproduced from Giosan L, Donnelly JP, Vespremeanu E, and Buonaiuto FS (2005) River delta morphodynamics: Examples from the Danube delta. In: Giosan L and Bhattacharya J (eds.) *River Deltas: Concepts, Models, Examples*, pp. 393–411. Tulsa, OK: SEPM. Special Publication 83.

influence the behavior and path of the plume after exiting the river mouth region. The relative importance of these processes, as well as their history (i.e., frequency of occurrence and intensity), are also important in determining the patterns of sediment deposition across the delta and, ultimately, its stratigraphic architecture.

Delta Sediments, Morphology, and Morphodynamics

Deltas are complex sedimentary systems comprising a multitude of depositional environments. At the apex of the delta plain, which is the subaerial part of a delta, the river usually bifurcates into distributaries that, in many cases, can build

their own delta lobes (Figure 2). The number of lobes tends to decrease in wave-dominated settings where the number of distributaries is minimal. Lobes can rarely be distinguished in settings where tides are strong, because tidal currents favor channel stability and suppress avulsions. The entire delta can sometimes change location through large-scale avulsions to build separate subdelta complexes. This is the case for the Huanghe (Yellow) River, which completely switched its course at least twice to build deltas in both the Bohai and South Yellow seas (Saito et al., 2001). The Mississippi River has also built six separate subdeltas comprising 16 individual lobes (see review in Roberts, 1997).

Based on general morphology of the subaerial delta plain, deltas have typically been classified as fluvial, wave, or tide dominated according to the dominant processes affecting sediment delivery, deposition, and dispersal (Figure 3; Galloway, 1975; Orton and Reading, 1993). A typical fluvial-dominated delta has an elongate to lobate plan-view morphology with highly crenulated shorelines; mouth bar and channel fill sands are interspersed within fine-grained deposits of the interdistributary region. In contrast, classic wave-dominated deltas assume a cusped morphology and are mainly composed of beach-ridge sands. Tide-dominated deltas, which appear mostly irregular in plain view, are primarily composed of estuarine and tidal shoals, or islands, flanked by mudflats.

However, a continuum of settings with overlapping processes rarely makes this classification operational at a scale larger than an individual delta lobe (Bhattacharya and Giosan, 2003). Even at the lobe scale, an in-depth look at wave influence on deltas shows that morphology and facies distribution is dependent more on the influence of the wave-driven long-shore sediment transport relative to fluvial discharge, rather than on wave energy itself (Figure 4). In multilobe deltas, one lobe may assume a vastly different morphology than an adjacent lobe. Of the two youngest lobes of the Danube delta, the Chilia, which is built with ~60% of the river-sediment discharge, is fluvial dominated, whereas the St. George is clearly wave dominated (Figure 2). Temporal variability in sediment discharge or basin energy can also result in drastic morphological changes in a delta/lobe (Giosan et al., 2005). The Po delta lobes had been wave dominated before a drastic increase of sediment discharge during the Little Ice Age (ca. AD 1450–1850) resulted in a fluvial-dominated outbuilding phase (Correggiari et al., 2005). Even without changes in total sediment discharge, the developing Chilia lobe of the Danube delta started to show morphological modifications typical of wave-influenced settings (e.g., straightening of the shoreline, buildup of barrier spits and islands) due to a decrease in sediment delivered by the river per unit shoreline (Figure 5).

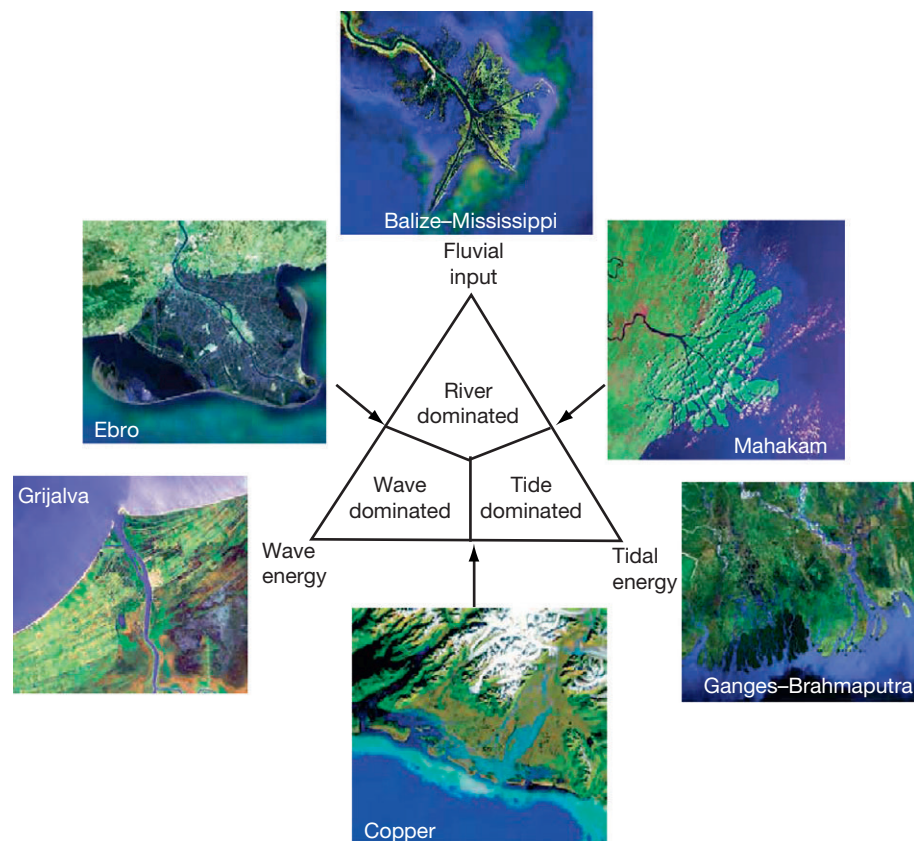


Figure 3 Process-based ternary classification of delta morphology as a function of fluvial discharge and wave and tidal energy. Satellite images of deltas are from Google Earth (<http://earth.google.com/>; Balize, Ebro, Mahakam, Grijalva, and Copper images © MDAEarthSat and DigitalGlobe; Ganges–Brahmaputra image © MDAEarthSat). Reproduced from Galloway WE (1975) Process framework for describing the morphologic and stratigraphic evolution of deltaic depositional systems. In: Broussard ML (ed.) *Deltas: Models for Exploration*. Houston, TX: Houston Geological Society.

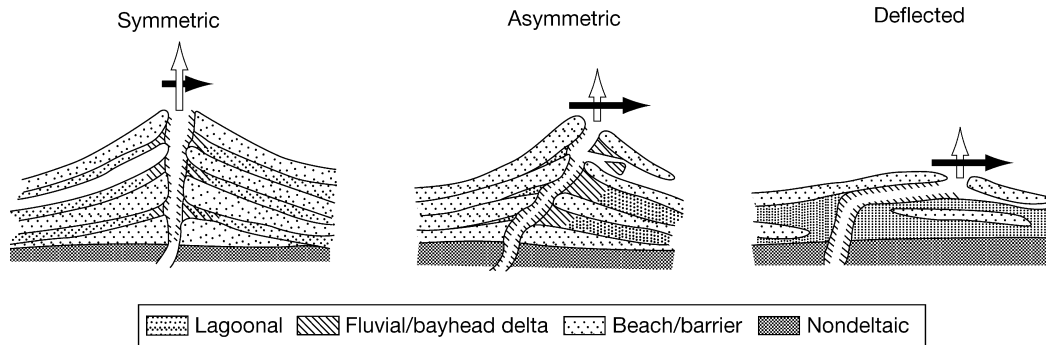


Figure 4 Morphology and depositional environments for mud-rich wave-dominated deltas as a function of the ratio between longshore drift and fluvial discharge (relative magnitudes indicated by black and white-filled arrows, respectively). Unlike the classical symmetric wave-dominated deltas, where sediment supplied by the feeding river is reworked on either side by waves, asymmetric deltas develop where the longshore drift becomes important relative to the river discharge and updrift beach-ridge plains develop. The downdrift side is a succession of elongate sandy ridges separated by mud-filled troughs. The mouth of the river gets deflected in extreme cases of asymmetry resulting in a series of quasiparallel sand spits and channel fills. Reproduced from Bhattacharya JP and Giosan L (2003) Wave-influenced deltas: Geomorphologic implications for facies reconstruction. *Sedimentology* 50: 187–210.

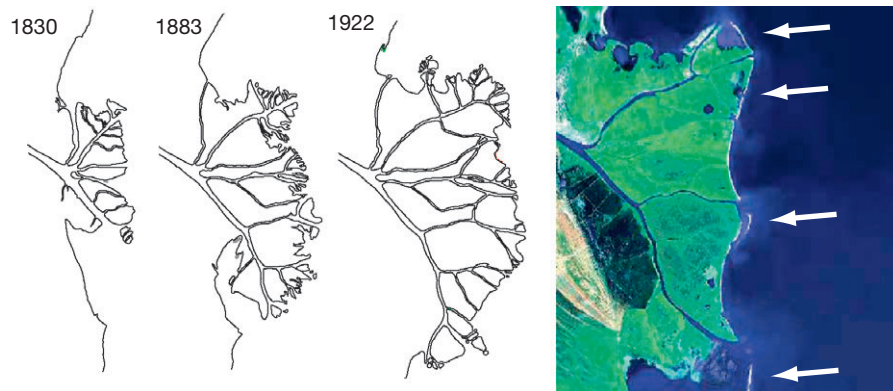


Figure 5 Delta-plain evolution of the Chilia III lobe of the Danube delta. The morphology of the lobe had been fluvial dominated until recently, when the coast began to straighten and barrier spits and islands started to develop along the coast (indicated by arrows). (Satellite image from Google Earth: <http://earth.google.com/>; © MDA EarthSat and DigitalGlobe.) Reproduced from Giosan L, Donnelly JP, Vesprieanu E, and Buonaiuto FS (2005) River delta morphodynamics: Examples from the Danube delta. In: Giosan L and Bhattacharya J (eds.) *River Deltas: Concepts, Models, Examples*, pp. 393–411. Tulsa, OK: SEPM. Special Publication 83.

Delta morphology is also greatly controlled by the geometry of the receiving basin (Wright and Coleman, 1973). Because tides are absent in lakes, lacustrine deltas (Figure 1(a)) are either fluvially dominated, when they build in shallow, low-energy lakes, or may become wave dominated in larger lakes (Figure 1(b)), where a longer fetch allows larger waves to form. Bayhead and lagoonal settings are typically low-energy environments that favor fluvial and tidal influences over the effect of waves (Figure 1(c)–1(e)). Morphology of deltas building along straight, open coasts is strongly dependent on the shelf configuration (i.e., width, steepness), which changes not only with the type of continental margin (i.e., active vs. passive) but also with sea-level changes that regulate the vertical accommodation space and the degree of coastal indentation. For example, as deltas prograde toward the middle and outer shelf, their morphology may be affected by increasing wave influence associated with deeper nearshore waters. In contrast, a broad, low-gradient inner shelf can attenuate incident waves and amplify tides, thereby favoring more tidally influenced delta systems.

In cross section, the idealized delta morphology and stratigraphy has been described as a prograding clinoform composed of an upward coarsening facies succession that starts with a basal muddy prodelta, progresses into a sandy delta front and is topped by texturally variable delta plain deposits (Figure 6). The subaqueous portion of a delta comprises both the distal prodelta and the steeper, prograding delta front, whereas the delta plain is predominantly subaerial. The muddy prodelta itself often acquires a ‘clinoform’ geometry, leading to a compound clinoform architecture for the delta that consists of a nearshore sandy clinoform associated with the shoreface and an offshore muddy clinoform (Figure 7) associated with the prodelta (Kuehl et al., 2005). Development of prodelta clinoforms is not universal and appears to be a function of fluvial input and basin hydrodynamics (Swenson et al., 2005). In such settings, gravity-driven flows of suspended mud commonly feed the growth of prodelta clinoforms in the presence of episodic or periodic high-energy processes, such as spring-neap tidal cycles or during storm events.

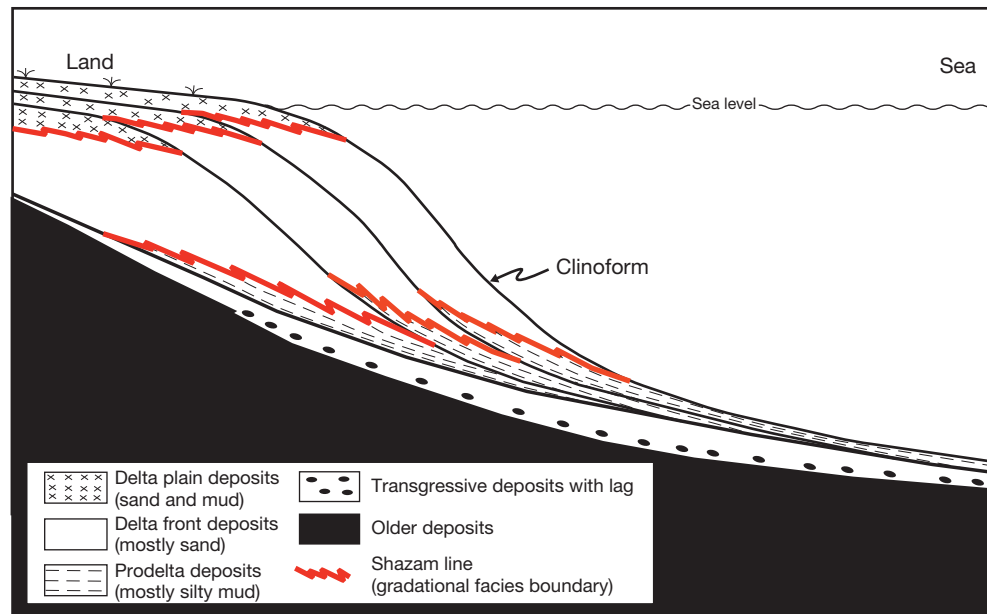


Figure 6 Idealized cross-section showing internal bedding geometry and facies architecture of a prograding deltaic clinoform. Reproduced from Gani MR and Bhattacharya JP (2005) Lithostratigraphy vs. chronostratigraphy in facies correlations of Quaternary deltas: Application of bedding correlations. In: Giosan L and Bhattacharya J (eds.) *River Deltas: Concepts, Models, Examples*, pp. 31–48. Tulsa, OK: SEPM. Special Publication 83.

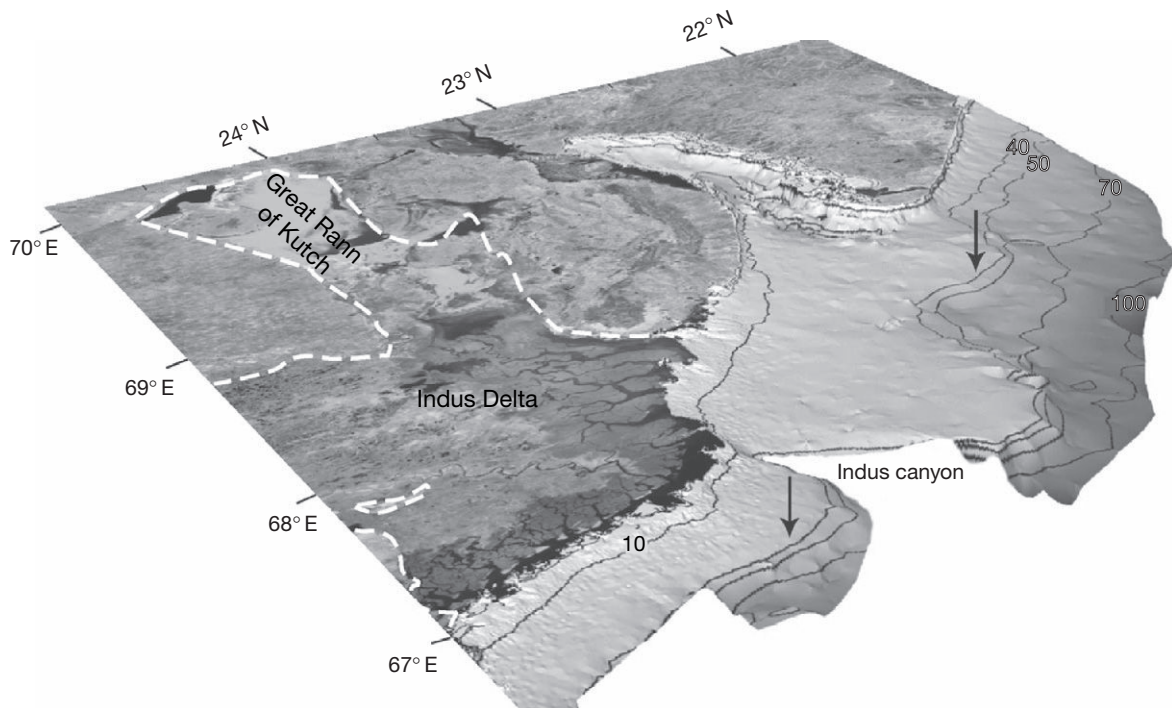


Figure 7 Digital elevation model for the Indus shelf based on the nineteenth century surveys showing the compound clinoform morphology for the subaqueous Indus delta. The nearshore clinoform extends to the ~10 m isobath. Offshore clinoform lobes are indicated by arrows. Inland geography is visualized on a satellite image from 2000 (the limits of the Indus delta and the adjacent mud flats of the Great Rann of Kutch are indicated by the white dashed line).

Morphodynamic response in deltas involves feedbacks between the evolving morphology of a delta and its fluvial and basal hydrodynamics. Early models identified rotation of the shoreline as responsible for changes in longshore drift along

deltaic coasts and affecting the rate of progradation. Asymmetry of the wave climate increases the shoreline instability on the downdrift delta wing, promoting development of spits detached from the main delta coast (Bhattacharya and Giosan, 2003).

Hydrodynamic–morphodynamic feedback loops have also been identified in relation to bedload sedimentation at river mouths. The hydraulic groin effect of river plumes is accompanied by the groin effect of the subaqueous delta or of the delta plain, blocking longshore drift at the river mouth (Giosan et al., 2005). On the downdrift wing of asymmetric wave-dominated deltas, periodic emergent barrier islands provide steep, continuous shorefaces for the longshore drift, leading to rapid expansions of the subaqueous delta in the alongshore direction.

Deltas in the Quaternary

Whereas waves and tides are key controls of deltaic systems over decades to centuries, delta evolution at longer timescales is most significantly controlled by (1) sea level, (2) climate, and (3) tectonics. Changes in these first two controls, sea level and climate, are hallmarks of the Quaternary period and its succession of glacial and interglacial phases. As a consequence, deltaic systems have effectively experienced continuous cycles of formation, change, and destruction throughout the Quaternary. Because the major controls of sea level, climate, and tectonics are interrelated, their relative impacts on deltaic systems remain unclear (Goodbred, 2003). Nevertheless, traditional models posit that the development of major deltaic systems is restricted to periods of sea-level highstand, with flooded continental margins and slow to moderate rates of sea-level rise. However, in the past few decades Quaternary deltaic systems have been discovered at a variety of shelf positions and been linked to numerous states of sea-level change. In the past two decades, the advent of multibeam sonar imaging and improved seismic reflection processing has allowed such deltaic shelf deposits to be discovered. A growing knowledge of climate history has in many cases provided explanations (e.g., high sediment supply due to increased precipitation, geomorphic landscape disequilibrium) for delta formation even under rapid rates of sea-level rise.

Influence of Climate on Deltas

Deltas can largely be considered a function of (1) riverine water and sediment discharge to the continental margin and (2) the interaction of these fluxes with coastal and marine processes. In this way, climate plays a major role in both the catchment-based forcing of sediment production and fluvial transport and the modulation of sea level and coastal weather conditions in the receiving basin. As such, there is strong climatic overprinting on deltaic systems, but precise signals are difficult to quantify because of complex feedbacks and secondary influences.

In the terrestrial realm of a fluviodeltaic dispersal system (i.e., the fluvial catchment) perhaps the most prevalent role that climate plays lies in controlling precipitation. With regard to sediment production, precipitation is a primary control on glacial and fluvial incision, weathering, and slope failure. Although highly variable, sediment is readily produced across a range of climates, from arid to humid and polar to tropical. However, one of the main factors that differs among these settings in terms of climate is runoff (=precipitation – evapotranspiration) and thus, the ability to transport sediment

to the delta. In other words, sediment can be produced in many settings, but is most efficiently transported given sufficient water discharge, which is primarily controlled by climate.

Other important characteristics of precipitation include its phase (liquid vs. frozen), seasonality, and episodicity. First, the phase of precipitation controls the relative role of glacial and fluvial erosion in sediment production, as well as the timing and nature of water discharge (i.e., the hydrograph). Each affects sediment-load grain size and weathering characteristics and, hence, the downstream development of delta stratigraphy. Second, the seasonality of river discharge affects the relative interaction of fluvial and marine processes, whereby strongly seasonal systems are subject to specific periods of river and marine-dominated conditions during the year. Similarly, the episodic discharge typified by storm-dominated or semiarid river systems leaves a strong imprint on deltas, because sediments are delivered in large but infrequent pulses. Finally, precipitation exhibits a strong secondary control on delta systems through the type and density of catchment vegetation, which affects runoff via overland flow and evapotranspiration rates as well as soil erosion rates.

From the marine side of a delta, climate change plays the primary role in modulating eustatic sea level through the orbitally controlled advance and retreat of continental ice sheets. This waxing and waning of ice sheets, a defining characteristic of the Quaternary period, has forced repeated, generally rapid, large-magnitude cycles of sea-level change. For delta systems, being principally coastal features, such glacioeustatic changes in sea level have been a first-order control on their development. In this way, traditional conceptual models suggest that prominent delta systems form only under relatively stable sea-level conditions and more specifically during highstands when fluvial sediment is likely to be trapped on the inner shelf. In contrast, during lowstands of sea level it was considered that most fluvial sediment was discharged beyond the shelf edge, either by off-shelf transport of gravity flows or through bypassing via canyon systems. Intervening periods of sea-level rise were assumed to preclude the development of deltas because of rapid coastal transgression or regression.

Site Cases from the Quaternary

Until recently, few studies have specifically investigated climate signals and impacts in delta systems. The current knowledge base on this subject comes from several modeling groups, some field studies, and also by revisiting existing data with a better understanding of climate change provided by secular records (e.g., loess deposits, speleothems, and lacustrine sediments). Although the site cases discussed below are not exhaustive, two relevant trends emerge. First, the number of modeling studies equals or exceeds those based on field investigations, and second, the majority of delta systems in which climate has been investigated are located in the subtropics (the Rhine/Meuse a notable exception). Subtropical systems tend to be highly seasonal, receiving nearly all of their precipitation during the wet summer monsoon when the intertropical convergence zone (ITCZ) is most proximal. At orbital timescales, both the position and strength of the ITCZ vary in response to global boundary conditions (i.e., extent of continental ice sheets) and maximum summer insolation,

respectively. Thus, such systems have exhibited among the most prominent responses to Quaternary climate variability.

Yellow river delta

Extensive databases from a few areas of the world have led to the discovery of significant deltaic deposits on the middle and outer portions of the shelf (see e.g., [Figure 7](#)). In the North Yellow Sea, a compilation of seismic data by [Liu et al. \(2004\)](#) detailed a massive deltaic clinoform comprising $\sim 400 \text{ km}^3$ of sediment from the Yellow (Huanghe) River. The Shandong mud wedge, as the feature is known, is 20–40 m thick and situated on the mid-shelf at water depths of 30 to $>50 \text{ m}$. Prominent within the deposit is a marine flooding surface that indicates the delta formed in two separate stages. Dating of the two phases of transgression and deposition suggests that these responses are associated with rapid changes in the rate of sea-level rise during deglaciation, perhaps related to meltwater pulses. In such case, brief but rapid transgressions were followed by progradation of the Yellow River delta, infilling accommodation space generated by the transgression.

This early Holocene pattern of Yellow River delta formation challenges traditional models in several ways. First, the growth occurs during a period of unstable sea level punctuated by sudden changes, yet the location of primary deposition remains largely unchanged during this time. [Liu et al. \(2004\)](#) ascribe this in large part to the stability of oceanographic circulation patterns in the North Yellow Sea. Second, size and thickness of the delta far exceed that expected for conditions of rapid Postglacial sea-level rise, when models might suggest formation of a thin transgressive sand sheet. In this case, Postglacial transgression in the early Holocene coincided with intensification of the Asian monsoon, and an apparent increase in fluvial sediment discharge. The notion that climatically induced increases in sediment discharge may be sufficient to force delta progradation during rapid sea level opens many new directions for research on how, where, and when such climate–sea level–delta interactions occur.

Ganges–Brahmaputra river delta

The Ganges and Brahmaputra rivers drain the foreslope and backslope of the Himalayan front range, respectively. The two rivers coalesce near the Bengal margin to form the world's largest subaerial delta and deep-sea fan system. The rivers are also strongly driven by the summer SW monsoon, when over 80% of water and 95% of sediments are discharged in the 4–5-month wet season. Given the global significance of the Himalayan/Tibetan uplift and its associated drainage systems, many portions of the Ganges fluvial delta system have been independently investigated in the past 15 years, revealing important variations in glacial activity, alluvial fan and flood-plain development, delta evolution, and regional oceanography. Taken together, the observed patterns show the immense Ganges dispersal system to respond to millennial-scale ($<10^4$ years) climate change in a system wide and largely contemporaneous manner, and major sedimentary signals to be transferred rapidly from source to sink with little apparent attenuation ([Goodbred, 2003](#)).

Some of the major signals of climate change recorded in the delta stratigraphy include (1) a thick (20–30 m) coastal mangrove facies that was deposited during rapid sea-level rise in the

early Holocene and (2) a massive volume of sediment trapped during Postglacial transgression, reflecting a sediment discharge at least 200% higher than the world's largest modern load of 1 billion tons ([Goodbred and Kuehl, 2000](#)). This high-discharge period corresponds to the early Holocene hypsithermal, when peak regional insolation supported a stronger than present monsoon and associated precipitation. In contrast, signals recorded in the delta during the Last Glacial Maximum (LGM) 18–20 ka reflect aridification and very low river discharge. Carbonate ooids, typical of shallow, evaporative tropical waters, formed at this time on the Bengal shelf directly adjacent to the river mouth. Similarly, reconstructed planktonic Foraminifera assemblages from the northern Bay of Bengal (the Ganges receiving basin) indicate peak surface-water salinities and apparently arid conditions at the LGM. These records from the delta and margin correlate well with geomorphic and stratigraphic results from the fluvial catchment, and together reflect the dominance of climate change in controlling this large fluviodeltaic system during the Quaternary ([Goodbred, 2003](#)).

Colorado, Brazos, and Trinity deltas (Gulf of Mexico)

These three rivers comprise a series of small catchments draining the semiarid to moderately humid hills of central Texas, USA and discharging to the northern Gulf of Mexico ([Anderson and Fillon, 2004](#)). Presently, sediment discharge is relatively small and delta progradation is driven by episodic floods associated with humid phases under El Niño conditions. However, during midstands and lowstands of sea level in the late Quaternary, these drainage basins coalesced to form expansive fluviodeltaic sequences on the middle and outer shelf ([Figure 8](#)). The size of these shelf deltas and their connection to underfit streams occupying large fluvial valleys reflect formation under higher-than-present discharge, implying significant changes in past climate.

On the Texas shelf, the position and architecture of relict deltas also indicate their formation during a variety of sea-level stages and transitions. Seismic-based reconstructions of the shelf deposits reveal large, forced-regressive sequences formed during the falling stages of sea level at marine isotope stages 5 and 3 ([Figure 8](#); [Anderson and Fillon, 2004](#)). During these stages, the relative highstand and midstand of sea level, respectively, placed the deltaic lobes near the modern inner to middle shelf. Overlying these regression-phase deltas are a group of smaller backstepping deltas formed during rapid Postglacial transgression after the LGM. Though smaller than the earlier delta complexes, formation of these systems under rapid rates of sea-level rise was in large part a consequence of a more humid climate and increased sediment discharge in the early Holocene. Together, delta sequences of the Texas shelf have formed under a variety of climatic and sea-level conditions, indicating the critical role that both of these forcings play.

Modeling studies

Quaternary climate change has been recognized as an important control on fluviodeltaic systems. However, it has been difficult to confidently link observed stratigraphic and sedimentological signatures to climatic forcings, given the strong overprinting of sea-level and marine processes on delta morphology. Such limits of field investigations are reflected in

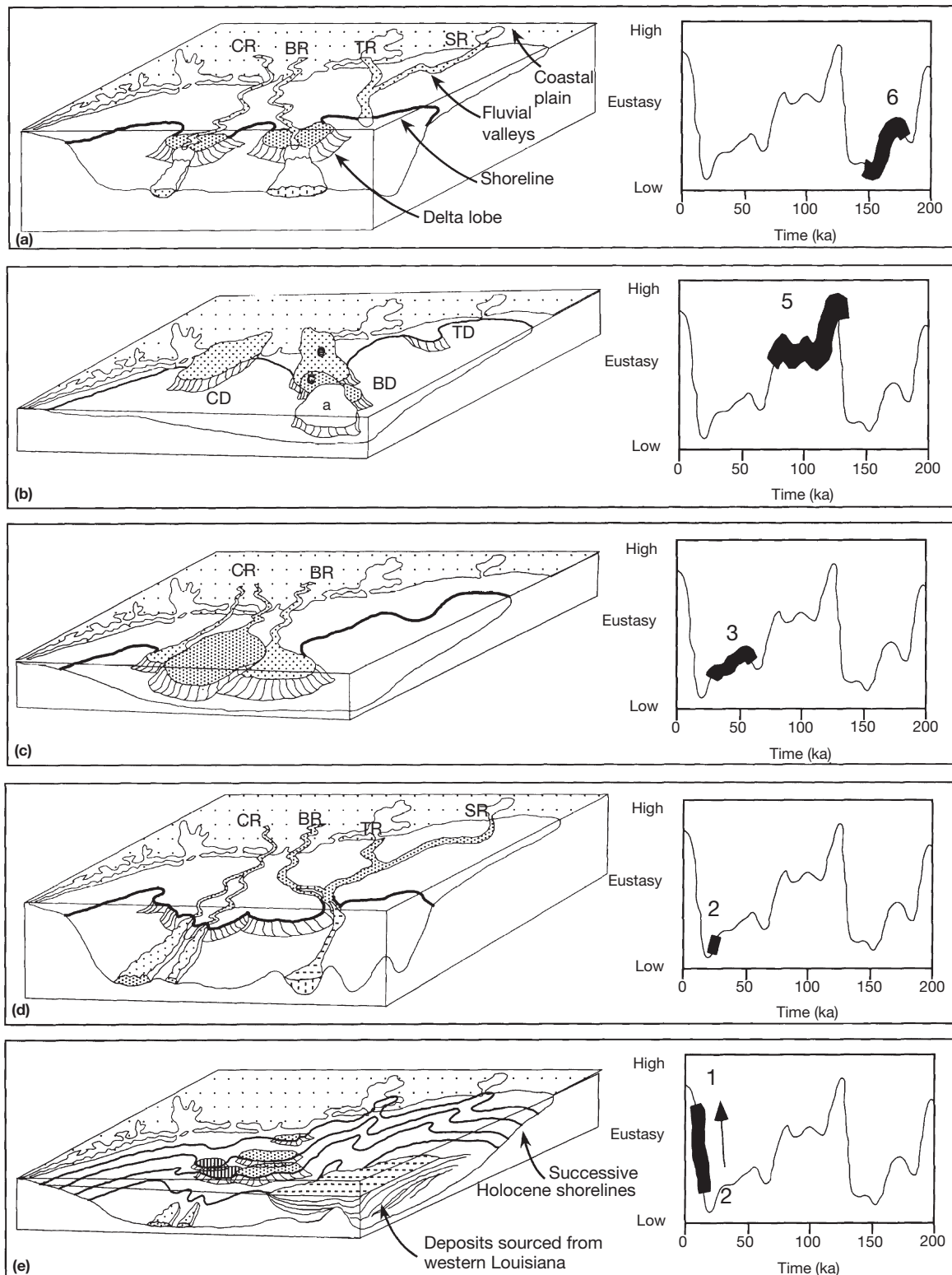


Figure 8 Reconstructed late Quaternary history for the rivers and deltas of the Colorado (CR, CD), Brazos (BR, BD), Trinity (TR, TD), and Sabine (SR) fluviodeltaic systems on the northern Gulf of Mexico shelf. Phases of delta progradation, incision, and backstepping are shown with corresponding phases of sea level, and are also associated with periods of climate change and varying sediment discharge. Note larger-than-present deltaic lobes building on the mid-shelf during MIS-3 interstade, under apparently wetter climatic conditions. Climate and sea level are dominant controls on delta formation and have varied somewhat in-phase during the late Quaternary (see Figure 9), giving rise to complex fluviodeltaic responses. Reproduced from Anderson JB and Fillon RH (eds.) (2004) *Late Quaternary Stratigraphic Evolution of the Northern Gulf of Mexico Margin*, p. 314 Tulsa, OK: SEPM. Special Publication 79.

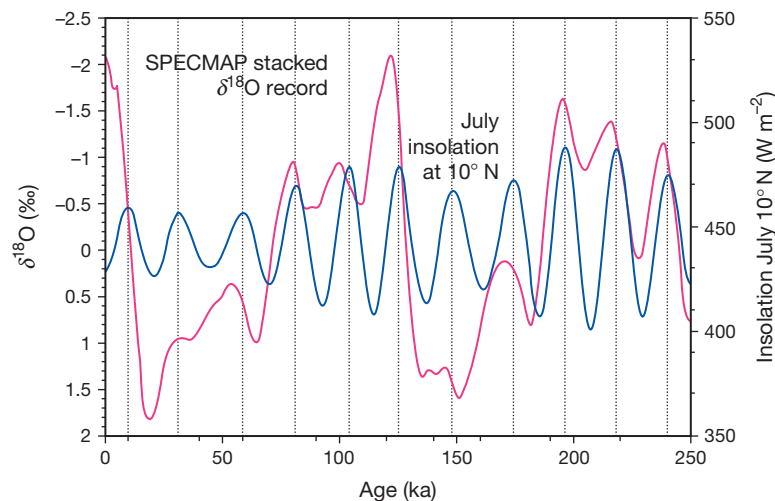


Figure 9 SPECMAP $\delta^{18}\text{O}$ and Milankovitch-tuned insolation records for the late Quaternary compiled by Imbrie, and Berger and Loutre, respectively. SPECMAP data represent a proxy for sea-level change and insolation serves as a coarse proxy for wetness in the subtropical (monsoon-influenced) Northern Hemisphere. Intended to show the general cophasing of peak insolation (generally wetter) with sea-level transgression, with the implied consequence of high sediment discharge corresponding with rapid sea-level rise. This pattern has been observed in the Yellow, Yangtze, and Ganges–Brahmaputra delta systems, west African rivers, and most likely in the Texas Gulf of Mexico rivers.

the relatively limited number of studies, many of which are discussed above. However, the exponential increase in computing power has allowed major advances to be made in the numerical modeling of fluviodeltaic systems. Several current models offer the opportunity to compare the consequences of forcing of fluvial delta systems by any combination of climate, eustasy, and tectonics, thereby helping elucidate the specific character of climate impacts on stratigraphic sequences (Overeem et al., 2005). This is especially relevant given that climate change and glacioeustasy are phased through most of the Quaternary (Figure 9), making it extremely difficult to derive unique interpretations of these forcings from field data alone.

One system that has been a focus of Quaternary response modeling is the Meuse river, which drains central Europe and discharges to the southern North Sea coast. Much of the research on this system during the past decade has had a focus on understanding the signatures of climate and eustasy contained in the late Quaternary stratigraphy and geomorphology (e.g., Veldkamp and Tebbens, 2001). Situated in the periglacial environments of northwest Europe during the Quaternary, climatically driven variations in the magnitude, timing, and nature of discharge appear to have played a role in shaping the fluviodeltaic system upstream of the hinge line. Terrace development in this river-dominated reach of the system is recognized as a primarily climate-controlled process through both modeling and field studies. However, the downstream impacts of such hydrological changes in the Meuse system appear to be insufficient to overcome the dominant influence of glacioeustasy during the Quaternary. Although there are likely to be impacts of climate in the delta, they remain indistinguishable at the resolution of current data and modeling capabilities.

Another system, the Niger delta, has been investigated using an industry-based stratigraphic model (van der Zwan, 2002). This investigation was aimed at testing the impact of presumed

Milankovitch-scale variations in sediment discharge over the Neogene. The orbitally linked forcing of regional climate and sediment occurred via global climate, insolation control of monsoon strength, and vegetation. Results suggest that sediment supply does change due to climatic variability at both large-scale (Ma) and Milankovitch-scale timeframes, and that there was a modest impact on stratigraphy. However, under icehouse conditions, fluctuations in climate-forced sediment supply were greatly overshadowed by rapid glacioeustatic sea-level change. These findings echo those of other modeling studies, in that climate is often found to be an important, but secondary, control on fluviodeltaic sequences. Furthermore, most modeling studies suggest that climatic signals preserved at the continental margin greatly lag the timing of the actual forcing in the fluvial catchment (e.g., Castellort and Van Den Driessche, 2003). This notion conflicts with recent field evidence from several of the subtropical rivers discussed above. Much of these discrepancies likely arise from our limited knowledge of appropriate input values for pre-Modern climate, riverine, and coastal conditions.

See also: [Fluvial Environments: Responses to Rapid Environmental Change; Sediments.](#)

References

- Anderson JB and Fillon RH (eds.) (2004) *Late Quaternary Stratigraphic Evolution of the Northern Gulf of Mexico Margin*, p. 314. Tulsa, OK: SEPM. Special Publication 79.
- Bhattacharya JP and Giosan L (2003) Wave-influenced deltas: Geomorphologic implications for facies reconstruction. *Sedimentology* 50: 187–210.
- Bhattacharya JP and Walker RG (1992) Deltas. In: Walker RG and James NP (eds.) *Facies Models – Response to Sea Level Change*, pp. 157–177. St Johns, NL: Geological Association of Canada.
- Broussard ML (1975) *Deltas, Models for Exploration*. Houston, TX: Houston Geological Society.

- Castelltort S and Van Den Driessche J (2003) How plausible are high-frequency sediment supply-driven cycles in the stratigraphic record? *Sedimentary Geology* 157: 3–13.
- Correggiari A, Cattaneo A, and Trincardi F (2005) Depositional patterns in the late-Holocene Po delta system. In: Giosan L and Bhattacharya J (eds.) *River Deltas: Concepts, Models, Examples*, pp. 364–392. Tulsa, OK: SEPM. Special Publication 83.
- Fisk HN, McFarlan E, Kolb CR, and Wilbert LJ (1954) Sedimentary framework of the modern Mississippi delta. *Journal of Sedimentary Petrology* 24: 76–99.
- Galloway WE (1975) Process framework for describing the morphologic and stratigraphic evolution of deltaic depositional systems. In: Broussard ML (ed.) *Deltas: Models for Exploration*. Houston, TX: Houston Geological Society.
- Gilbert GK (1885) The topographic features of lake shores. *Annual Report US Geological Survey* 5: 75–123.
- Giosan L and Bhattacharya J (eds.) (2005) *River Deltas: Concepts, Models, Examples*, p. 502. Tulsa, OK: SEPM. Special Publication 83.
- Giosan L, Donnelly JP, Vespremeanu E, and Buonaiuto FS (2005) River delta morphodynamics: Examples from the Danube delta. In: Giosan L and Bhattacharya J (eds.) *River Deltas: Concepts, Models, Examples*, pp. 393–411. Tulsa, OK: SEPM. Special Publication 83.
- Goodbred SL (2003) Response of the Ganges dispersal system to climate change: A source-to-sink view since the last interstade. *Sedimentary Geology* 162: 83–104.
- Goodbred SL and Kuehl SA (2000) The significance of large sediment supply, active tectonism, and eustasy on margin sequence development: Late Quaternary stratigraphy and evolution of the Ganges–Brahmaputra delta. *Sedimentary Geology* 133: 227–248.
- Kuehl SA, Allison MA, Goodbred SL, and Kudrass H (2005) The Ganges–Brahmaputra delta. In: Giosan L and Bhattacharya J (eds.) *River Deltas: Concepts, Models, Examples*, pp. 413–434. Tulsa, OK: SEPM. Special Publication 83.
- Liu JP, Milliman JD, Gao S, and Cheng P (2004) Holocene development of the Yellow River's subaqueous delta, North Yellow Sea. *Marine Geology* 209: 45–67.
- McKee BA, Aller RC, Allison MA, Bianchi TS, and Kineke GC (2004) Transport and transformation of dissolved and particulate materials on continental margins influenced by major rivers: Benthic boundary layer and seabed processes. *Continental Shelf Research* 24: 899–926.
- Meade RH (1996) River-sediment inputs to major deltas. In: Milliman JD and Haq BU (eds.) *Sea Level Rise and Coastal Subsidence*, pp. 63–85. Dordrecht: Kluwer.
- Morgan JP (1970) *Deltaic Sedimentation: Modern and Ancient*, p. 312. Tulsa, OK: SEPM. Special Publication 15.
- Orton GJ and Reading HG (1993) Variability of deltaic processes in terms of sediment supply, with particular emphasis on grain size. *Sedimentology* 40: 475–512.
- Oti MN and Postma G (1995) *Geology of Deltas*, p. 315. Rotterdam: A.A. Balkema.
- Overeem I, Syvitski JPM, and Hutton EWH (2005) Three-dimensional numerical modeling of deltas. In: Giosan L and Bhattacharya J (eds.) *River Deltas: Concepts, Models, Examples*, pp. 13–30. Tulsa, OK: SEPM. Special Publication 83.
- Roberts HH (1997) Dynamic changes of the Holocene Mississippi River delta plain: The delta cycle. *Journal of Coastal Research* 13: 605–627.
- Saito Y, Yang ZS, and Hori K (2001) The Huanghe (Yellow River) and Changjiang (Yangtze River) deltas: A review on their characteristics, evolution and sediment discharge during the Holocene. *Geomorphology* 41: 219–231.
- Swenson JB, Paola C, Pratson L, Voller VR, and Murray AB (2005) Fluvial and marine controls on combined subaerial and subaqueous delta progradation: Morphodynamic modeling of compound-clinoform development. *Journal of Geophysical Research* 110. F02013. <http://dx.doi.org/10.1029/2004JF000265>.
- Syvitski JPM, Harvey N, Wolanski E, Burnett WC, Perillo GME, and Gornitz V (2005) Dynamics of the Coastal Zone. In: Crossland C (ed.) *LOICZ: Coastal Change and the Anthropocene*. Berlin, Germany: Springer.
- Van der Zwan CJ (2002) The impact of Milankovitch-scale climatic forcing on sediment supply. *Sedimentary Geology* 147: 271–294.
- Van Wagoner JC, Posamentier HW, Mitchum RM, et al. (1988) An overview of the fundamentals of sequence stratigraphy and key definitions. In: Wilgus CK, Hastings BS, Kendall CGC, Posamentier HW, Ross CA, and Van Wagoner JC (eds.) *Sea-Level Changes: An Integrated Approach*, pp. 39–45. Tulsa, OK: SEPM. Special Publication 42.
- Veldkamp A and Tebbens LA (2001) Registration of abrupt climate changes within fluvial systems: Insights from numerical modelling experiments. *Global and Planetary Change* 28: 129–144.
- Wright LD and Coleman JM (1973) Variations in morphology of major river deltas as functions of ocean wave and river discharge regimen. *Bulletin of the American Association of Petroleum Geologists* 57: 370–398.

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Foraminifera Studies *see* **Paleoceanography**: Paleoceanography An Overview; **Paleoceanography, Biological Proxies**: Benthic Foraminifera; Planktic Foraminifera; **Paleoceanography, Physical and Chemical Proxies**: Oxygen-Isotope Stratigraphy of the Oceans; **Paleoceanography, Biological Proxies**: Temperature Proxies, Census Counts

Fossil Mites *see* Oribatid Mites

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Geomagnetic Excursions and Secular Variations

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Introduction

Earth is the only terrestrial planet to have a strong, largely dipolar magnetic field (i.e., a field similar to that of a bar magnet, with poles near both geographic poles). Since the time of William Gilbert (AD 1600), it has been widely acknowledged that the geomagnetic field has its origins within the Earth. Discovery of field variations through time dispelled the notion that the field could be due to a permanent magnetization. In the twentieth century, the idea of a dynamo operating in Earth's liquid iron-rich outer core became accepted. However, it has only been in the past two decades that computer simulations have demonstrated that such a dynamo mechanism can reproduce the observed features of the field (e.g., [Glatzmaier and Roberts, 1995](#)). The core fluid is in constant motion, driven by both thermal and compositional convection combined with Earth's rotation. Movement of the conducting fluid in a magnetic field causes electromagnetic induction of electric currents in the fluid. The currents themselves are a source of magnetic field, and under the right circumstances they can reinforce the original field, which results in a self-sustaining geodynamo ([Figure 1](#)).

Paleomagnetic evidence demonstrates that the geomagnetic field has existed from the Earth's early history (e.g., [Tarduno et al., 2006, 2010](#)) and that it is dynamic, with a broad spectrum of time variations. Calculations based on the attenuation of electromagnetic waves traveling through the mantle suggest that, for variations originating in the core to be observed at the surface of the Earth, they must have periods longer than about 1 year. Higher frequency geomagnetic variations are observed, but are external in origin (e.g., daily variations due to Earth's rotation, variations due to changes in solar activity, and interactions of the solar wind with the magnetic field). The sunspot cycle also contributes signals with periods of 11 and 22 years.

The range of field variability of relevance to this article originates in the geodynamo and occurs on timescales of years to millennia, and is known as 'geomagnetic secular variation.' Secular variation refers to the gradual changes in direction and intensity observed at all locations on Earth. The amplitude of the changes in field direction may be up to about 30°, and the field intensity can change by a factor of about 2. Occasionally, and apparently randomly, much larger changes occur, where the field intensity decreases drastically and the field direction undergoes large swings, after which the field settles back to its former state. This style of field behavior is referred to as a 'geomagnetic excursion.' Even more rarely the field destabilizes, in much the same manner as during an excursion, but instead of reverting to its former polarity state, the field recovers with the opposite polarity. In such cases, a polarity reversal has occurred. Geomagnetic field variability can be described by magnetohydrodynamic equations (e.g., [Glatzmaier and Roberts, 1995](#)); these equations do not depend on polarity. The paleomagnetic record indicates that, since the formation of Earth, the field has spent approximately equal amounts of time in the normal (i.e., like today) and reversed polarity states ([Merrill et al., 1998](#)). The longest geomagnetic polarity intervals are known as 'superchrons,' which represent periods of tens of millions of years when the geomagnetic field did not reverse polarity. Only two superchrons are well documented (the Permian or Kiaman reversed superchron and the Cretaceous normal superchron), while a third superchron has been postulated in the Ordovician ([Pavlov and Gallet, 2005](#)). Although the frequency of reversals has been highly variable throughout Earth's history, on average two or three reversals have occurred per million years throughout the Cenozoic. In this article, we describe how paleosecular variation (PSV), including changes in geomagnetic field intensity, geomagnetic excursions, and polarity reversals, can be used in Quaternary stratigraphy and geochronology.

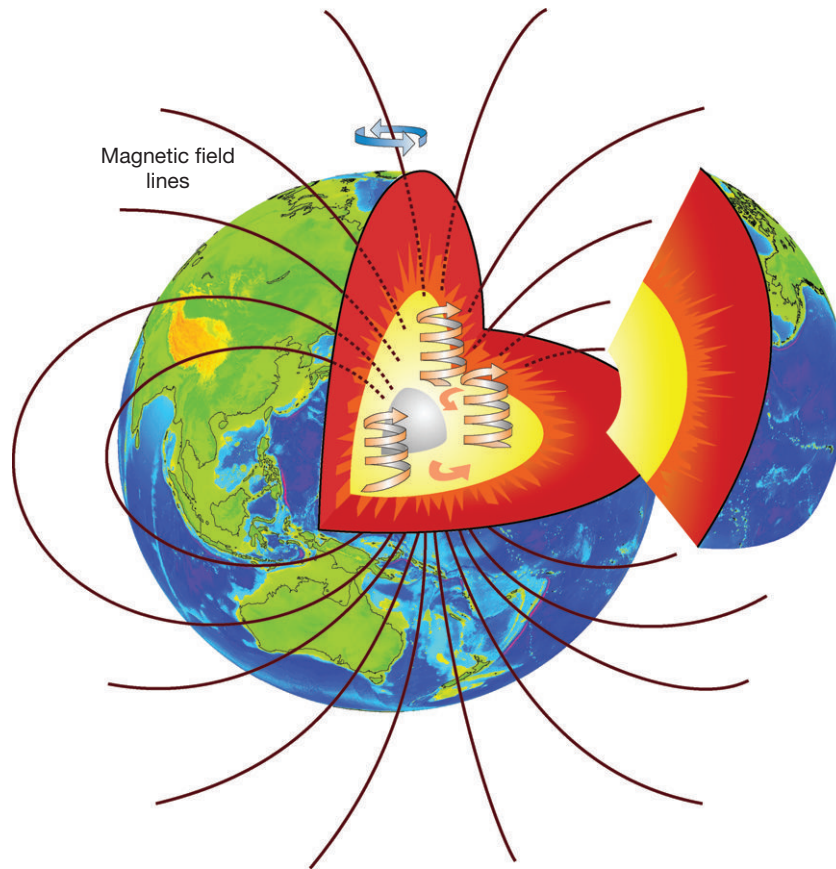


Figure 1 Cut-away cartoon of Earth's interior with a view of the solid inner core and fluid outer core from which geomagnetic flux lines emerge. The helical lines represent flux emerging from columnar convection cells that occur within the larger scale flow of the outer core that drives the geodynamo. These cells transfer heat from the inner core–outer core boundary to the core–mantle boundary. The overall field is symmetrical about Earth's rotation axis. Over time, the field averages to that of a geocentric axial dipole.

Geomagnetism

The Present-Day Field

At most locations on Earth's surface, the compass needle does not point exactly to the north. The angle between true north and magnetic north (the direction of the compass needle) is known as the declination, while the angle of the magnetic field vector below the horizontal (or of a dip needle) is known as the inclination (**Figure 2(a)**). If the geomagnetic field were due simply to a geocentric axial dipole (GAD), then the declination would be 0 at all locations on the globe, and the inclination would depend only on latitude, with $\text{In}_{\text{GAD}} = \tan^{-1}(2 \tan \lambda)$. The field intensity is given by $\text{Int}_{\text{GAD}} = \frac{\mu_0 m}{4\pi r^3} (1 + 3 \sin^2 \lambda)^{1/2}$, where λ is the latitude, m is the dipole moment, r is Earth's radius and μ_0 is the permeability of free space (constant). While the geomagnetic field undergoes constant change, a snapshot of the field in AD 2010 is shown in **Figure 3**, where the declination, inclination, and intensity of the International Geomagnetic Reference Field (IGRF, International Association of Geomagnetism and Aeronomy, Working Group V-MOD, 2010) are contoured on a map of the world. The dominance of the GAD is seen by the fact that the intensity increases from between about 25 000 and 40 000 nT (nano-Tesla) around the equator to about 60 000 or 65 000 nT at high latitudes, and

that the inclination generally steepens from the equator toward the poles. However, in detail the field is clearly not axisymmetric. The best fitting dipole to the present-day field is tilted 10° with respect to Earth's rotation axis. The poles of this best fitting dipole, which are known as the 'geomagnetic poles,' are located at 80.0° N , 72.2° W , and at 80.0° S , 107.8° E . This dipole field accounts for about 90% of the total field. It is the residual 10%, which represents the non-dipole field, that accounts for the irregularities and anomalies seen in global plots of declination, inclination, and intensity (**Figure 3**). For example, the magnetic poles, the points at which the field is found to be vertical (inclination $= \pm 90^\circ$), do not coincide with the geomagnetic poles, but are displaced from them by amounts that depend on the local non-dipole field. The discrepancies are different in the two hemispheres, so that the magnetic poles are not antipodal. In 2010, the north magnetic pole was in northern Canada (85.0° N , 132.6° W), while the south magnetic pole lay to the south of Australia near the coast of Antarctica (64.4° S , 137.3° E). These are also the points from which declination contours radiate out (**Figure 3(a)**). Anomalous foci in inclination and intensity charts, such as the intensity minimum over South America and the southwest Atlantic (**Figure 3(c)**), and the steep inclination off South Africa (**Figure 3(b)**) are further evidence of the non-dipole field components, as also are the complexities of the declination chart (**Figure 3(a)**).

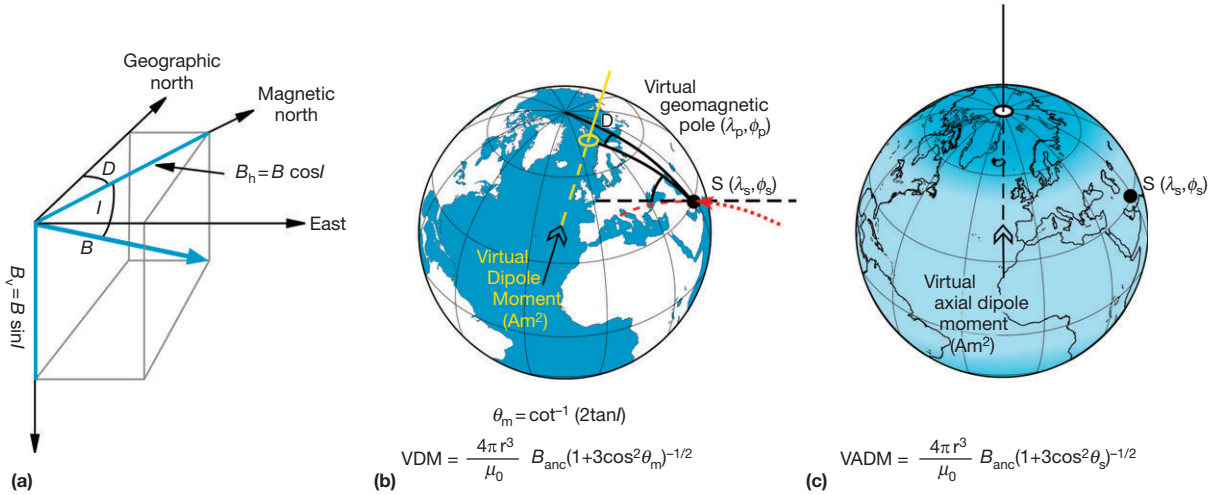


Figure 2 Definition of terms used to describe the geomagnetic field. (a) Three-axis coordinate system (north, east, and down), with definition of the declination and inclination of the geomagnetic field vector. Declination (D) is the angle of the horizontal component of the vector from true north, measured clockwise; inclination (I) is the angle of the vector from the horizontal (positive down); and intensity (B) is the length of the vector. Field measurements at any point can be described in terms of these components. (b) If the field is assumed to be due to a geocentric dipole, a magnetic measurement at any site on Earth's surface (S), with latitude, λ_s , and longitude, ϕ_s , will give rise to a virtual geomagnetic pole (VGP) that describes the geomagnetic field at that instant in time. The intensity of that field is given by the virtual dipole moment (VDM). (c) Definition of the virtual axial dipole moment (VADM), which is determined using the site co-latitude (θ_m) and the measured intensity (or paleointensity) at S . Equations are given for the VDM and VADM below (b) and (c), respectively. (b) and (c) are adapted from Tauxe L (2010) *Essentials of Paleomagnetism*. Berkeley, CA: University of California Press.

Further insight as to the cause of such features is gained by examining the field at the core–mantle boundary (CMB), which is calculated by the mathematical process of downward continuation (Figure 4). At the CMB, the highest field intensities are not at the poles, but occur at latitudes of between 70° and 80° . Furthermore, there are two high-intensity foci in each hemisphere. These four high-latitude ‘flux lobes’ are responsible for the predominantly dipolar field observed at the surface.

Virtual Geomagnetic Poles, Virtual Dipole Moments, and Virtual Axial Dipole Moments

The geomagnetic poles are the locations at which the axis of the best fitting dipole intersects Earth's surface. It is possible to take any single observation or paleomagnetic direction, and from it and the latitude and longitude, λ_s and ϕ_s , of the site from which it was obtained, calculate the poles of the geocentric tilted dipole, which would yield the observation. Such a pole is called a ‘virtual geomagnetic pole’ (VGP). The north VGP lies on the great circle defined by the observed declination D at a distance from the site corresponding to the ‘virtual’ or magnetic colatitude θ_m , which is calculated from the observed inclination I , as shown in Figure 2(b). Spherical trigonometric equations for calculating VGPs from λ_s , ϕ_s , D , and I are given in standard texts on paleomagnetism (e.g., Tauxe, 2010). If the paleointensity and inclination are known, it is possible to calculate the magnetic moment of the virtual dipole: the virtual dipole moment (VDM). If paleointensity is known, but a sample has not been oriented, the direction of the paleomagnetic field remains unknown. In such cases, the colatitude of the sampling site, θ_s , must be used instead of θ_m , and the result is known as a virtual axial dipole moment (VADM) as shown in Figure 2(c).

Conventionally, models of the field at the surface of the Earth are based on spherical harmonic analysis of the

geomagnetic potential. The potential is represented by a summation of a series of terms that generally decrease in magnitude and importance:

$$U = a \sum_{l=1}^{\infty} \sum_{m=0}^l \left(\frac{a}{r}\right)^{l+1} P_l^m(\cos \theta) (g_l^m \cos m\phi + h_l^m \sin m\phi)$$

where a is the radius of the Earth, r , θ , and ϕ are the radial distance from the pole, the colatitude and longitude of the site, respectively, P_l^m are associated Legendre polynomials, and g_l^m and h_l^m are known as the Gauss coefficients of the various surface harmonic components. The most widely used model of the geomagnetic field, the IGRF, is calculated every 5 years by finding the values of the Gauss coefficients that produce the best fit of such a model to observatory, satellite, and other data. In the most recent models, the coefficients are calculated to degree, l , and order, $m=13$. The first and largest term in the summation for the potential, $\frac{a^3}{r^2} g_1^0 \cos \theta$, represents a GAD. The next two terms, $\frac{a^3}{r^2} g_1^1 \cos \phi \sin \theta$ and $\frac{a^3}{r^2} h_1^1 \sin \phi \sin \theta$, represent dipoles in the equatorial plane along longitudes of zero and 90° , respectively. Together, these three terms represent the best fitting tilted dipole, and the orientation of this dipole (i.e., the geomagnetic pole positions) can be found from the coefficients g_1^0 , g_1^1 , and h_1^1 . Degree 2 terms correspond to quadrupole fields, degree 3 to octupole fields, etc., with each successive degree adding greater complexity to the model.

Secular Variation

Gauss pioneered a global network of geomagnetic observatories in the early nineteenth century. INTERMAGNET now

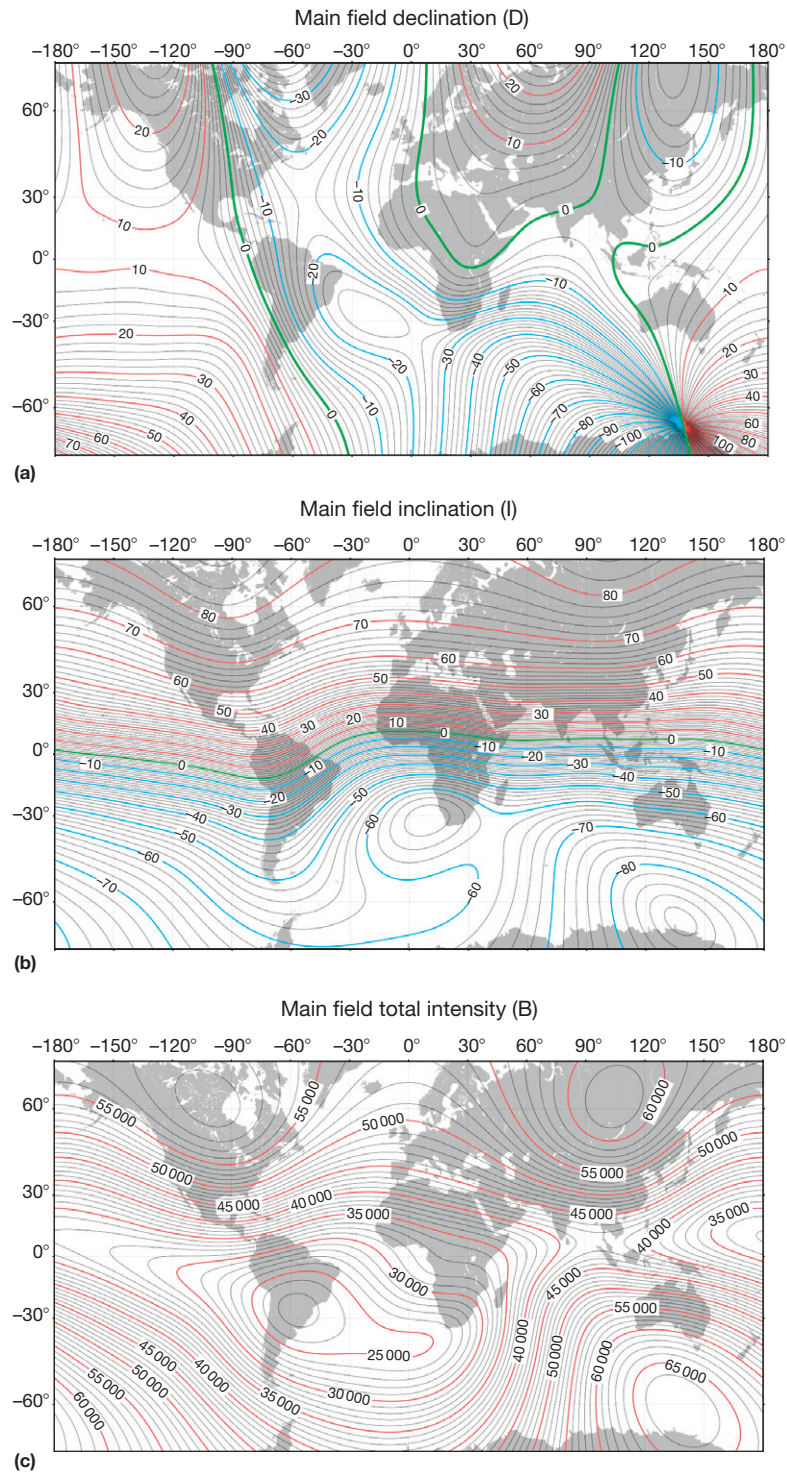


Figure 3 Global plots of the geomagnetic field at the surface of the Earth, calculated from the International Geomagnetic Reference Field for 2010. (a) Declination (in degrees; the green contour is the agonic (zero declination), with red and blue contours indicating eastward and westward declinations, respectively); (b) inclination (in degrees; the green contour is the magnetic equator (zero inclination), with red and blue contours indicating downward and upward inclinations, respectively). The contours are shown at 2° intervals in each case. (c) Intensity (in 1000 nT contour intervals). The intensity generally rises with increasing latitude. The maps are plotted in Mercator projections. From www.ngdc.noaa.org.

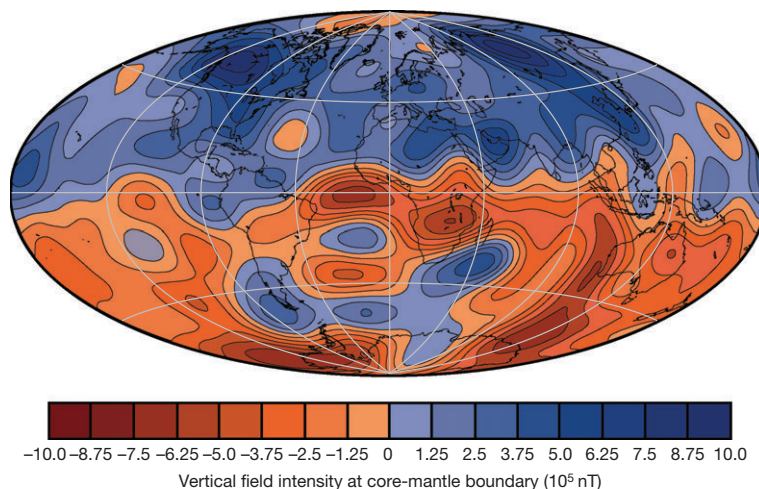


Figure 4 Vertical (or radial) magnetic field at the core–mantle boundary (CMB) in 1990, from the gufm1 model of Jackson et al. (2000). Blue represents flux entering the core and orange-red represents flux exiting from the core, while deeper colors indicate stronger field intensities. Note the two high-latitude flux concentrations in each hemisphere, and also the reverse polarity (blue) flux patch in the south Atlantic (after Jackson and Finlay (2007)). Note the much stronger field intensities at the CMB compared to Earth’s surface (cf. Figure 3(c)).

records and collates not only secular variation data, but also the shorter period variations of external origin. Amongst the longest records are the directional records from London, Paris, and Rome, which date back over 400 years (Figure 5(a)): variations at all three locations are clearly related, with the magnetic field vector describing what looks like three-quarters of a clockwise loop. However, the average inclination at Rome is some degrees lower than that at Paris, and at Paris the average inclination is slightly shallower than that at London. This again is evidence of the dominance of the main dipole field for which inclination is latitude-dependent. A much shorter record from a Southern Hemisphere site, Melbourne (Figure 5(b)), has no clear correlation with European records for the same period. This indicates that the changes are largely related to features of the non-dipole field, which are coherent over regions of continental extent only. Early examinations of such historical records, particularly of those from Europe and North America, led to the suggestion of westward-drifting eddy-like features in Earth’s outer core. Bullard et al. (1950) estimated an average drift rate of 0.18° per year for the Atlantic region between 1907 and 1945. This westward drift is seen clearly in animated global charts of declination changes over the past 400 years (Jackson et al., 2000; <http://jupiter.ethz.ch/~cfinlay/gufm1.html>). However, such animations also demonstrate that westward drift is not ubiquitous. Over other areas of the globe, features grow and decay without drifting significantly, or they even drift eastward, while there is little secular variation over the Pacific hemisphere.

From the late eighteenth century, scientists and mariners began to obtain relative measures of field intensity by measuring the period of oscillation of a dip needle. It was not until 1832, however, that Gauss devised a method of calibrating measurements to give absolute values of field intensity. At most, but not all, locations on Earth, the field intensity has decreased significantly over the past two centuries, although the decrease has not been steady or uniform.

Paleomagnetism

How Rocks Become Magnetized

Detailed historical records of geomagnetic secular variation are available only for the last 400 years (Jackson et al., 2000). In order to retrieve information about variations of the magnetic field beyond this period of historical observations, it is necessary to use paleomagnetic techniques. Use of paleomagnetism requires an understanding of how geological materials record geomagnetic information. The stability of the remanent magnetization in question is largely dependent on the grain size of the magnetic particles. Ideal magnetic recording occurs in particles that have uniform magnetization. Such particles have a magnetic single-domain (SD) state. In magnetite, which is the most common terrestrial ferrimagnetic rock-forming mineral, the particle size range for stable SD behavior is 30–80 nm (Tauxe, 2010). Magnetizations do not remain stable forever: they decay over time. However, for stable SD particles, the so-called relaxation time exceeds the age of the Earth if the rock containing the SD particles remains at ambient temperatures. It is this high magnetic recording fidelity of SD grains that provides the basis for the usefulness of paleomagnetism in Earth science. Different materials acquire a long-lived, or remanent, magnetization through different mechanisms. In Quaternary geology, many materials can be analyzed paleomagnetically; therefore it is important to consider the main mechanisms by which remanent magnetizations are acquired.

A thermoremanent magnetization (TRM) is acquired as a material cools from high temperatures. A TRM is typically responsible for the magnetization of igneous rocks, rocks that have been in contact with igneous bodies (baked margins) or that have been heated by burial, or archaeological artifacts that have been subjected to firing (e.g., pottery). When a ferro or ferrimagnetic material is heated above its Curie temperature, increased thermal energy causes the magnetic moments of

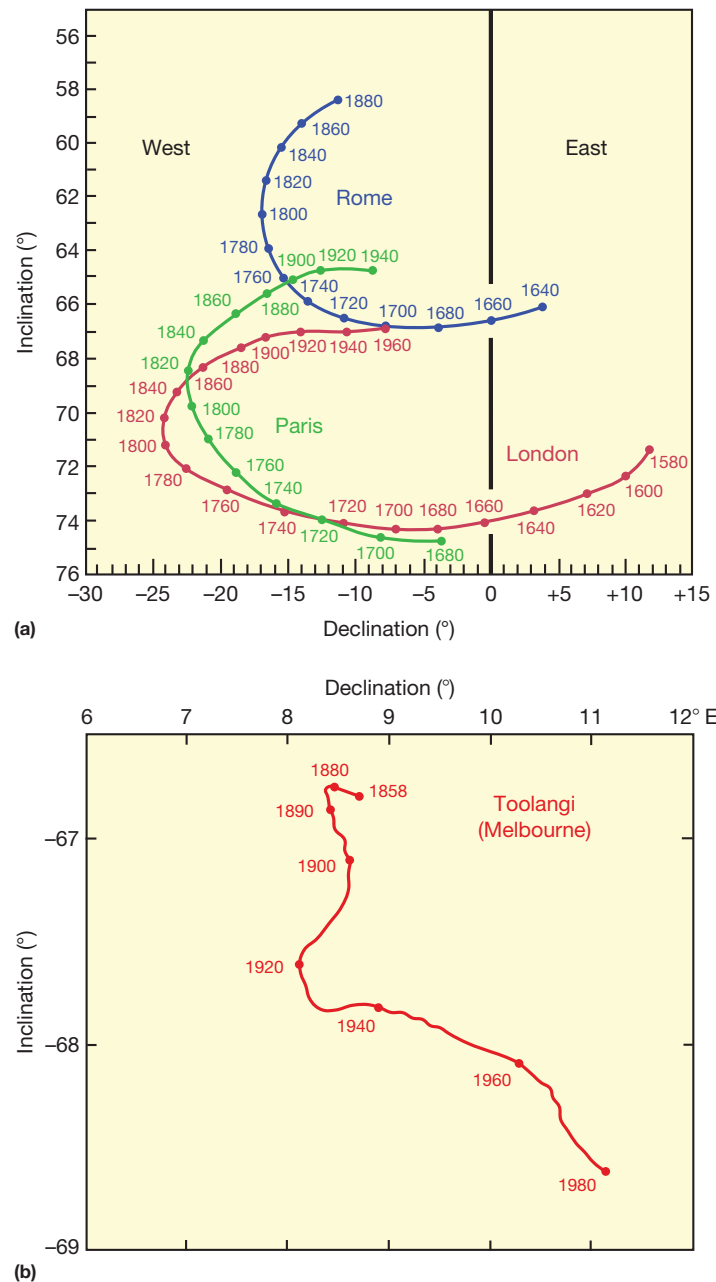


Figure 5 Historical records of the direction of the geomagnetic field at (a) London, Paris, and Rome for the past 400 years, and at (b) Melbourne since 1858. Inclination is plotted against declination, so each curve traces the movement of the tip of the magnetic vector when viewed along the vector direction. Similar clockwise loops for London, Paris, and Rome are indicative of regionally coherent secular variation. The secular variation pattern in Melbourne is completely different from that in Europe. During the twentieth century, increased electromagnetic interference from cultural sources has meant that observatories in all cities shown have been moved to more magnetically quiet locations. Each record is therefore a composite from different observatories, adjusted to a common location.

individual atoms to randomize (Figure 6(a)). As the material is cooled through the Curie temperature, the interatomic distance decreases and the resulting magnetic exchange interaction causes spontaneous alignment of atomic magnetic moments. This enables TRM acquisition, with the rock recording an overall magnetization that is parallel to the ambient geomagnetic field direction (Figure 6(b)). A TRM is generally

highly stable, and fine-grained volcanic rocks often provide ideal paleomagnetic records.

When igneous rocks are eroded, magnetic minerals are transported along with other mineral detritus until they are eventually deposited as sediment. During deposition in a quiet environment, magnetic particles will have a torque exerted on them by the geomagnetic field, which causes

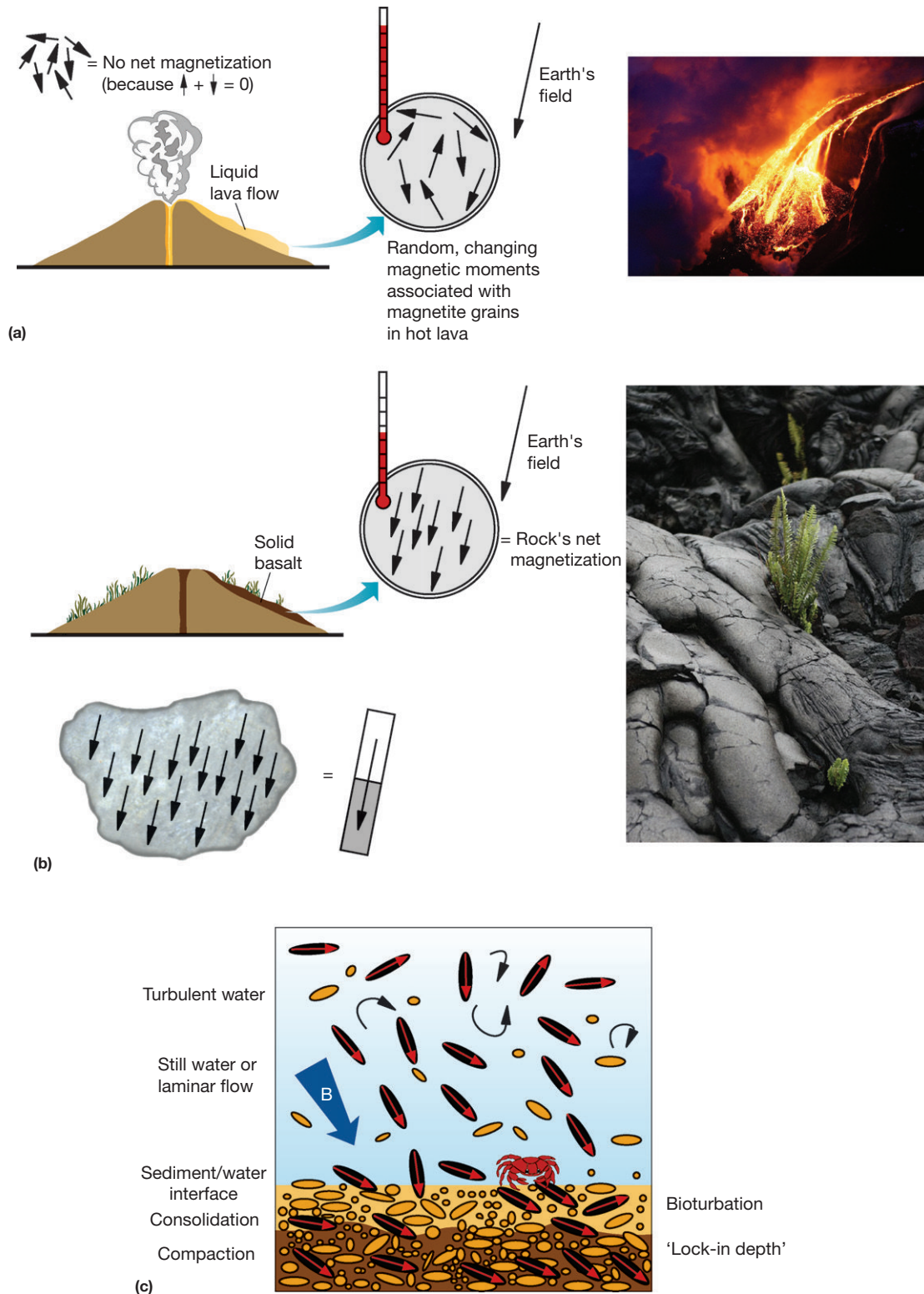


Figure 6 Illustration of how rocks acquire a remanent magnetization. (a) When lava flows are hot (i.e., above the Curie or blocking temperatures of the magnetic minerals in the lava), thermal energy exceeds any atomic magnetic exchange energy and the magnetic moments are randomized. (b) When the lava cools through the Curie and blocking temperatures, a thermoremanent magnetization (TRM) is acquired, which is aligned parallel to the ambient geomagnetic field. (c) When magnetic particles are eroded from a source rock and are deposited as a sediment, the particles may align with the geomagnetic field to give rise to a depositional remanent magnetization (DRM). If the sediment is remobilized (e.g., by bioturbation), high water contents may allow the magnetic particles to realign with the geomagnetic field via the torque exerted on such particles by the field, which gives rise to a postdepositional remanent magnetization (PDRM). Figures (a) and (b) are after Marshak (2008); Figure (c) is modified after Tauxe and Yamazaki (2007).

their magnetic moments to rotate toward the ambient field direction (**Figure 6(c)**). Such a magnetization is referred to as a 'depositional remanent magnetization' (DRM). If deposition occurs in the presence of even a minor hydro/aerodynamic disturbance, the perturbation will exceed the geomagnetic torque and magnetic particles will not be ideally aligned with the geomagnetic field (e.g., [Heslop, 2007](#)). Particles can also rotate when they settle onto the sedimentary substrate, or be flattened into the bedding plane by compaction, or can be disrupted and moved by bioturbation. The low likelihood that a pure DRM will survive physical disturbance during deposition, bioturbation, or Brownian motion can be overcome by invoking a postdepositional remanent magnetization (PDRM). PDRM acquisition is argued to occur within the unconsolidated uppermost sediments in which water contents are high. In this environment, the geomagnetic field can exert a torque on magnetic particles after their last disturbance. This realigns the particles with their magnetic moments parallel to the geomagnetic field and imparts a PDRM, which becomes locked in at some depth below the sediment–water interface (**Figure 6(c)**). This depth can vary and is an elusive parameter that is rarely well constrained. It is usually assumed to be less than 20 cm (e.g., [Roberts and Winklhofer, 2004](#)).

In some environments, chemical changes give rise to the growth of magnetic minerals that record a chemical remanent magnetization (CRM). Under oxic diagenetic conditions, hematite can form after deposition and give rise to a CRM when grains grow through the critical SD blocking volume. The origin and timing of magnetization acquisition in red sediments can be uncertain and has been debated intensely (e.g., [Larson and Walker, 1975](#); [Tauxe and Kent, 1984](#)). Under anoxic diagenetic conditions, postdepositional growth of the magnetic iron sulfide, greigite, can lead to either faithful or complex paleomagnetic records depending on the timing of CRM acquisition (e.g., [Roberts et al., 2011](#)). If monoclinic pyrrhotite forms within sediments, it is likely to cause a later diagenetic remagnetization (e.g., [Hornig and Roberts, 2006](#)). CRM acquisition therefore causes complications, which require considerable care when interpreting paleomagnetic records.

Assessing the Reliability of Paleomagnetic Data

When using paleomagnetism to date Quaternary rocks and sediments, the most desirable magnetizations that are likely to give rise to high-fidelity recording of the geomagnetic field are a TRM or a DRM/PDRM. After sampling, it is necessary to subject samples to rigorous laboratory demagnetization procedures to assess the stability of the recorded magnetizations before any interpretation can be made in terms of field behavior or chronology. The most common laboratory procedures involve a progressive, stepwise alternating field or thermal demagnetization. The efficacy of one technique compared to the other depends largely on the nature of the sample, the magnetic minerals present, and the range of grain sizes in which the magnetic minerals occur. In many cases, it is important to undertake a pilot study to determine the most effective technique to use. After completion of stepwise demagnetization, the paleomagnetic data can be inspected using vector demagnetization diagrams, which enable identification of secondary remanence components and the most stable

'characteristic' remanent magnetization (ChRM) component. Use of the word 'characteristic' avoids premature linking of the documented magnetization to an interpretation of whether it is primary or not. Where possible, it is highly desirable to use a range of paleomagnetic field tests to assess whether the recorded ChRM is an ancient or later magnetization. These tests include the fold test, the conglomerate test, the baked contact test, and the reversals test. For details of demagnetization procedures and field tests, the reader is referred to a recent comprehensive online text ([Tauxe, 2010](#)). In many cases, however, investigated rocks or sediments may not be folded, and they may not contain intraformational conglomerates, baked contacts, or geomagnetic reversals. In such cases, one is more dependent on the coherence of the recorded paleomagnetic signal and how well it compares with other well-documented geomagnetic information, such as continuous records of PSV or relative geomagnetic paleointensity. It is important to emphasize that successful interpretation of paleomagnetic data in terms of geochronology depends on rigorous field and laboratory work, and that unidentified problems associated with CRM acquisition, failure to identify a robust ChRM through detailed stepwise demagnetization, physical disturbance of the paleomagnetic record, or other issues may lead to difficulties that preclude geochronological interpretation.

While TRM and DRM/PDRM might be the most useful remanence acquisition mechanisms for geochronological applications, it is also useful to consider their limitations. For example, while a TRM is usually considered to be the ideal type of magnetic remanence, even volcanoes with the most frequent eruptions will not provide the continuous paleomagnetic records needed to document a wide spectrum of field behavior. Significant time gaps between eruptions can also complicate chronological interpretation. In contrast, sedimentary sequences can provide virtually continuous records of field variability. However, a DRM is not an efficient recorder of paleomagnetic information, and inclination errors, which represent significant divergences from the ambient geomagnetic field direction, are commonly associated with DRM acquisition (e.g., [King, 1955](#); [Tauxe, 2005](#)). This can be especially important in hematite-bearing sediments (e.g., [Tauxe and Kent, 1984](#)) and in eolian sediments (e.g., [Zhao and Roberts, 2010](#)). A PDRM might be considered an improvement on a pure DRM because it provides an opportunity for magnetic particles to become reoriented with the ambient geomagnetic field shortly after deposition. However, it is likely that a PDRM is locked in over a depth range rather than at one specific and narrow horizon, which will cause smoothing of the paleomagnetic signal (e.g., [Verosub, 1977](#)). This smoothing, at minimum, serves as a low-pass filter, so that the highest frequency field variations are removed from the paleomagnetic record (e.g., [Roberts and Winklhofer, 2004](#)). Despite these complexities and limitations, paleomagnetism provides the principal means of reconstructing geomagnetic field variations beyond the last 400 years, and can be a robust technique for Quaternary geochronology.

Determining Geomagnetic Field Intensity

The geomagnetic field is a vector, with both a direction and intensity (**Figure 2(a)**). Many paleomagnetic studies simply use the field direction to obtain, for example, a magnetic

polarity stratigraphy or for tectonic reconstructions. But the geomagnetic field intensity also contains valuable information concerning field variability that is relevant to Quaternary stratigraphy.

Volcanic rocks and fired archaeological artifacts provide the most reliable records of geomagnetic field intensity. When a TRM is acquired during cooling, the strength of the TRM is proportional to the ancient field intensity (B_{anc}). It is necessary to undertake a series of laboratory experiments to determine the proportionality factor between the TRM intensity and the absolute geomagnetic intensity (Thellier and Thellier, 1959). A second TRM is imparted in the laboratory by heating and cooling the sample in a known applied laboratory field (B_{lab}). The paleofield intensity is then determined from $B_{anc} = (TRM_{anc}/TRM_{lab})B_{lab}$. Absolute paleointensities have been widely used to date archaeological materials via comparison with regional archaeomagnetic master curves (see below). Time series of intensity variations from volcanic rocks are also useful for understanding geomagnetic field variability over longer timescales. However, as stated previously, volcanic records are not continuous and it is desirable to obtain continuous records of geomagnetic paleointensity variations for Quaternary dating. Sedimentary records provide the greatest potential for obtaining continuous paleointensity records, but they do not allow determination of the absolute strength of the ancient geomagnetic field.

The relationship between geomagnetic field intensity and DRM or PDRM intensity is much more complex than for rocks that carry a TRM. The DRM intensity is affected not only by the strength of the ambient magnetic field and the concentration of magnetic minerals but also by the nature of the magnetic mineral that records the paleomagnetic signal, its grain size, and the mechanism by which the magnetization was acquired. The effects of most of these variables can be offset by normalizing the natural remanent magnetization (NRM) of sedimentary samples by an artificial laboratory-induced magnetization and by imposing strict rock magnetic selection criteria. These criteria require that magnetite must be the only magnetic mineral present and that it occurs within a narrow grain size range with a small range between its maximum and minimum concentrations (cf. Tauxe, 1993). By applying these strict selection criteria, it has been demonstrated that the geomagnetic field varies in a globally coherent manner on millennial timescales (and on smaller timescales where such variations are recorded in high-resolution datasets, e.g., St-Onge et al., 2003) and that it is possible to use these variations to date sedimentary sequences (e.g., Channell et al., 2008; Hornig et al., 2003; Laj et al., 2000; Stoner et al., 2002; Tauxe and Yamazaki, 2007; Valet et al., 2005; Yamazaki and Oda, 2005; Ziegler et al., 2011). The potential for paleointensity-assisted chronology in Quaternary sediments is explored in more detail below (section 'Relative Geomagnetic Paleointensity Variations').

Paleosecular Variation Records

Secular Variation Recorded in Archaeological Artifacts and Lava Flows

As a result of the meticulous work of many research groups over the past several decades, detailed records of changes in field direction and intensity have been compiled from

archaeomagnetic studies. These records cover only parts of the world and time periods during which pottery was made; few records extend back before 1000 BC. Some of the most detailed, best resolved records come from Europe – a selection of which are shown in Figure 7. Retrieval of the direction of the paleomagnetic field requires knowledge of the orientation of the artifact at the time of its firing and subsequent cooling. For incomplete pottery fragments, the orientation is often not available. In such cases, only the paleointensity can be determined. Even with complete objects, the orientation may be only partially known; for example, it may be known that a pot was fired upright or upside down, but its azimuthal orientation may be uncertain. Hence, inclination, but not declination, might be retrievable. Each archaeomagnetic determination involves a discrete object, therefore each artifact must be dated individually, and each age estimate carries its own uncertainty. Various dating methods are commonly used, ranging from primary methods such as radiocarbon and thermoluminescence dating of contextual material to secondary methods that include historical records of site occupation, pottery design, etc. Each dating method has its own limitations and uncertainties. Regional master curves are compiled by combining all available data that meet carefully chosen reliability criteria from within a certain region. For example, the Western Europe curves of Gallet et al. (2002) include data from France, Great Britain, and Southern Italy, while the curve of Schnepf and Lanos (2006) contains data from within a 500-km radius of the town of Radstadt, Austria (Figure 7). Individual declinations and inclinations are 'reduced' to a common location, Paris and Radstadt in the two examples of Figure 7, to take account of geographic spread. This is usually done by calculating a VGP position (see separate text box), and then calculating the expected direction (D and I) at the common location. Also shown in Figure 7 are lake sediment data (see below) from Great Britain (Turner and Thompson, 1981, 1982), which have also been reduced to the latitude and longitude of Paris. The data agree well (Figure 7), which indicates that these independent datasets record a common underlying geomagnetic signal. Algorithms are now available that enable statistical comparison of the paleomagnetic direction and intensity of an artifact of unknown age with these regional archaeomagnetic master curves of declination, inclination, and intensity (Lanos et al., 2005; Schnepf and Lanos, 2006).

An alternative recently developed approach involves production of regional field models from archaeomagnetic data. The data are not transformed to a common location, but are fitted to a series of mathematical functions in a manner similar to the global models based on spherical harmonic functions (see Section 'The Present-Day Field'). At least two different types of functions are currently being investigated – spherical cap harmonic functions (Haines, 1985; Pavón-Carrasco et al., 2009, 2010; Thébaud et al., 2006) and Slepian functions (Dahlen and Simons, 2008). In principle, once such a model has been calculated for a region, and regularized over time, it is then possible to calculate a secular variation curve for any chosen location within the region. In Figure 7, declination and inclination curves for the latitude and longitude of Paris, which are calculated from the SCHA.DIF.3K spherical cap harmonic model of Pavón-Carrasco et al. (2009) for Western Europe, are also shown. Finally, much smoother curves are obtained from the global CALS3k model (Korte et al., 2009),

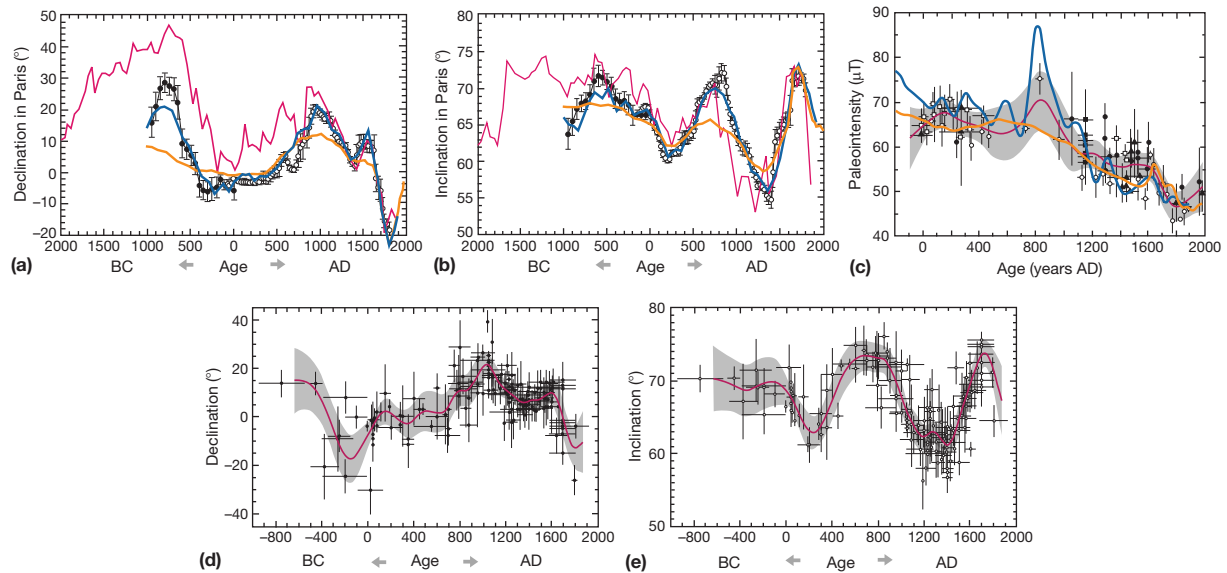


Figure 7 Archaeomagnetic secular variation records from Europe. (a, b) Declination and inclination records from France. Each point represents an average over a 160-year window, calculated at 50-year intervals (error bars represent 95% confidence intervals associated with the archaeomagnetic directions). All directions were referred to Paris using an intermediate VGP transformation (open symbols are from Bucur (1994) and closed symbols are from Gallet et al. (2002)). The solid red lines denote the British lake sediment record of Turner and Thompson (1981, 1982), which is also referred to the latitude and longitude of Paris (after Gallet et al. (2002)). Blue and yellow lines represent curves from the regional and global models of Pavón-Carrasco et al. (2009) and Korte et al. (2009), respectively. (c) Archaeomagnetic paleointensity measurements from France, Spain, Norway, and Denmark, all referred to the latitude of Paris, compiled by Gómez-Paccard et al. (2008). The solid red line and shading represent the mean and 95% confidence interval. The blue and yellow lines are as in (a) and (b). (d, e) Archaeomagnetic declination and inclination master curves for Austria with mean and 95% confidence curves obtained by Bayesian modeling. Compiled from data within a 500-km radius of the town of Radstadt, Austria (Schnepp and Lanos, 2006).

which is based on spherical harmonic analysis of worldwide data. Such models are useful for predicting secular variation curves from regions where data are sparse but are not yet well enough resolved to provide a high-resolution dating tool.

Secular Variation Recorded in Lake and Marine Sediments

Development of purpose-designed piston corers has enabled collection of long lake and marine sediment cores without disturbing, compressing, or twisting the recovered sediments. Since the 1970s and 1980s, detailed PSV records have been obtained as a result of continuous subsampling of such cores, followed by detailed paleomagnetic analysis. Several cores are usually taken from a lake, and can be correlated using magnetic susceptibility signatures and lithological markers. The respective PSV records can then be stacked or averaged to provide a local or regional PSV master curve. Age control is usually provided by radiocarbon dating of organic material, palynology, and/or identification of known tephra. Examples of lacustrine regional PSV records include the British record of Turner and Thompson (1981, 1982), the Southeast Australian record of Barton and McElhinny (1981, 1982), the Minnesota record of Lund and Banerjee (1985), the eastern United States stack of King and Peck (2001), and the Fennoscandian stack of Snowball et al. (2007), all of which cover most of the Holocene. In periods where the records overlap, the best records compare extremely well with archaeomagnetic records from the same region. Several lacustrine and marine PSV records that extend from the west coast of North America to

Europe are shown in Figure 8. Correlation, particularly of declination records over the past 3000–4000 years, is striking. There is, however, no obvious east–west time lag, as would be expected if westward drift was a significant feature of the secular variation. Especially prominent in nearly all the records is the large eastward swing in declination at about 2700 BP, labeled by Turner and Thompson (1981) as ‘f.’ This coincides with relatively high inclinations, which makes the overall direction particularly distinctive for dating purposes. Swing ‘f’ is not well represented in either the global CALS7K model (Korte and Constable, 2005), which was compiled from archaeomagnetic and lake sediment records, or in the European regional SCHA8K model (Pavón-Carrasco et al., 2010), possibly because both models are restricted to low degree and order functions. The current generation of global models is also compromised by data scarcity from the Southern Hemisphere, particularly the South and Southwest Pacific. Even though much work is needed to retrieve high-resolution PSV records in such regions to constrain and improve global models, comparison with regional PSV master curves provides an effective means of dating Holocene sediments at decadal to millennial-scale resolution.

Long-Timescale Relative Paleointensity Variations

Since the early 1990s, much effort has gone into obtaining records of relative geomagnetic paleointensity spanning tens or hundreds of thousands of years. Globally coherent

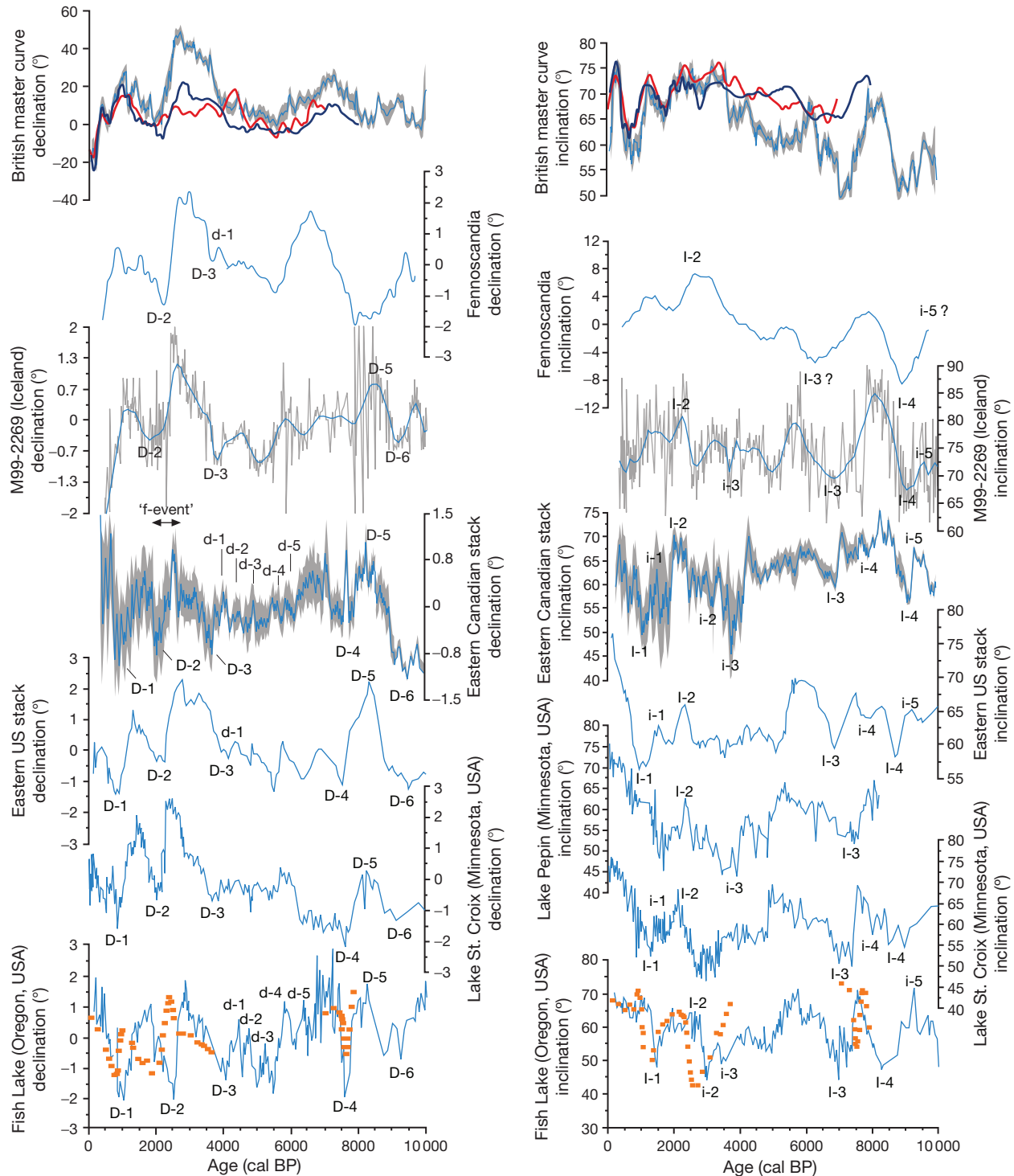


Figure 8 Comparison of Northern Hemisphere Holocene PSV records. Declinations are plotted on the left and inclination records are plotted on the right (see Barletta et al. (2010) for sources of each record). The records demonstrate the use of PSV features for wide-scale regional correlation. The British lake sediment PSV record of Turner and Thompson (1981) is shown above the other records, with the global CALS7K (red) and regional SCHA8K (blue) model curves superimposed (calculated for the latitude and longitude of Loch Lomond). Except for the British record, the declinations have been standardized by subtracting the arithmetic mean declination for each record from each measured declination, and then dividing by the standard deviation. Adapted from Figure 7 of Barletta F, St-Onge G, Stoner JS, Lajeunesse P, Locat J (2010) A high-resolution Holocene paleomagnetic secular variation and relative paleointensity stack from eastern Canada. *Earth and Planetary Science Letters* 298: 162–174.

paleointensity variations are well represented in a series of stacked regional (Laj et al., 2000; Stoner et al., 2002; Yamazaki and Oda, 2005) and global (Tauxe and Yamazaki, 2007; Valet et al., 2005; Ziegler et al., 2011) records (Figure 9). The robustness of these paleointensity stacks is improving with time, and we now have paleointensity time series that collectively span the Quaternary (Channell et al., 2008; Horng et al., 2003; Tauxe and Yamazaki, 2007; Valet et al., 2005; Yamazaki and Oda, 2005; Ziegler et al., 2011). While it is always desirable to have an independent chronology, relative paleointensity records are now recognized to be useful for developing age models for sedimentary sequences. Many paleoceanographic proxies that are used to develop age models for marine sediments are dependent on seawater chemistry, and millennial-scale synchronization of such records can be complex (e.g., Skinner and Shackleton, 2005). In contrast, long-period geomagnetic variations provide a globally synchronous geophysical parameter that is independent of seawater chemistry, which can potentially be used to develop independent millennial-scale chronologies. Continuous relative paleointensity variations also provide a context within which geomagnetic excursions can be more robustly correlated from site to site. There is therefore considerable potential for using paleointensity variations in chronostratigraphy. Nevertheless, as stated previously, magnetically uniform sediments are needed to satisfy strict selection criteria for relative paleointensity studies and there are many environments and sediment types from which it is not possible to obtain high-quality relative paleointensity data.

Geomagnetic Excursions

A geomagnetic excursion is usually defined as a deviation of VGPs by more than 40–45° from the geographic pole (Merrill and McFadden, 1994) or as a deviation of VGPs away from the normal range of geomagnetic secular variation (Vandamme, 1994). Using geomagnetic excursions for the purposes of stratigraphic correlation assumes that certain key questions are answered. For example, are excursions local or global in extent? Do they represent dominantly dipolar or non-dipolar field variations? Are they rare or frequent? Do they exhibit characteristic styles of field behavior that enable unique identification of one particular excursion compared to another? Are they related to geomagnetic secular variation or are they aborted polarity reversals? Progress in answering these questions has recently been summarized by Roberts (2008), but they remain subjects of intensive research to which complete answers are still not available. A repeated problem in using Quaternary excursions for stratigraphic dating is that anomalous data have often been emphasized at the expense of less interesting but more robust data, which, combined with poor chronological constraints, means that dubious ‘excursion’ records have often been used to estimate the age of a sedimentary sequence. This has led to a ‘reinforcement syndrome’ (Thompson and Berglund, 1976) that complicates geomagnetic excursion research and that has wreaked widespread chronostratigraphic havoc in Quaternary research. In providing reasons for why excursions should not be used for stratigraphic correlation, Merrill and McFadden (2005) argued that excursions are not

global phenomena, and therefore that correlation of excursions should be limited to locations separated by angular distances of less than 30° on Earth’s surface. However, detailed high-resolution records obtained for the best studied excursions indicate that at least some excursions are recorded on a global scale (Laj et al., 2006). Lack of recording of excursions in many high-resolution sedimentary paleomagnetic records could result from smoothing associated with PDRM acquisition (Roberts and Winklhofer, 2004). Issues associated with potential PDRM filtering need to be borne in mind among other pitfalls of using excursions for stratigraphic correlation.

Only two geomagnetic excursions, the Laschamp (~40 ka) and Iceland Basin excursions (~190 ka), have been studied in sufficient resolution to provide any idea of whether excursions exhibit characteristic transitional field behavior (Laj et al., 2006). Similarities in the excursive VGP pathways for these two excursions mean that excursions are unlikely to provide a unique basis for correlation based on field behavior. Therefore, avoiding spurious reinforcement of anomalous paleomagnetic data requires that excursion records should be used with extreme caution in Quaternary geochronology without additional high-quality chronostratigraphic data.

With the above warnings in mind, it is worth considering the evidence for different excursions when they can be placed within a precise chronostratigraphic framework. As outlined by Roberts (2008), high-resolution paleomagnetic (including relative paleointensity) and $\delta^{18}\text{O}$ analysis of marine sediments and radioisotopic dating of volcanic rocks over the last decade have verified the precise age of some excursions (e.g., Mono Lake, Laschamp). New excursions have been identified (e.g., Iceland Basin, Bjorn, Gardar), and poorly dated excursions have been precisely dated (e.g., Pringle Falls, West Eifel; Singer et al. (2008a,b)). Unverified excursions have been discredited (e.g., Lake Mungo, Gothenburg, Biwa, Jamaica), and others remain to be fully validated and accepted (e.g., Norwegian-Greenland Sea, Calabrian Ridge 0, 1, 2, and 3). Following the critical reassessment of the evidence for documented excursions by Laj and Channell (2007), seven Brunhes Chron excursions are shown as validated in Figure 9, with six others indicated as ‘possible.’ This revised excursion catalog provides an update to the compilations of Langereis et al. (1997) and Lund et al. (2001). All ‘validated’ and ‘possible’ excursions are associated with paleointensity minima (Figure 9), as are all full polarity reversals. The geomagnetic field is evidently dynamic with constantly changing paleointensity and abundant collapses in field intensity (Figure 9). Excursions occur during such periods of weak paleointensity, and it is now clear that excursions are an intrinsic and frequent component of field behavior. A high-quality excursion catalog is likely to grow in future as detailed studies are performed for progressively older time intervals. We emphasize that excursions are best used in Quaternary stratigraphy when they are well constrained by independent chronological data, including continuous relative paleointensity records.

Geomagnetic Reversals

Of the range of documented geomagnetic field variations, polarity reversals are the most spectacular. With recent revision of

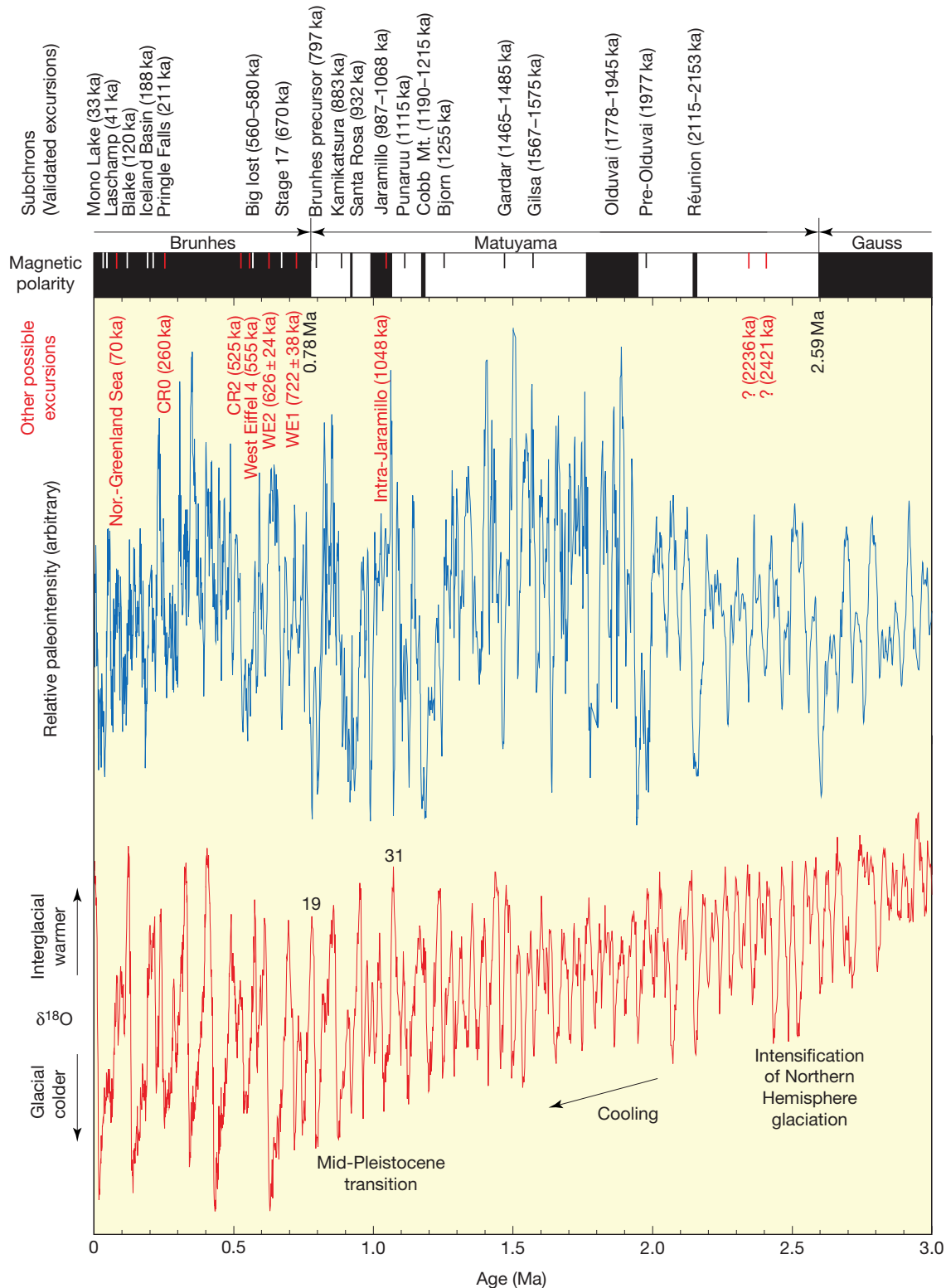


Figure 9 Geomagnetic polarity and relative paleointensity variations and excursions through the Quaternary. Polarity is indicated at the top of the figure, with black=normal and white=reversed polarity. Validated excursions are indicated along with their respective ages above the polarity log in italics, with ‘possible’ excursions that have yet to be fully validated indicated in red below the polarity log (from [Laj and Channell, 2007](#)). Each polarity reversal and excursion coincides with a paleointensity minimum. The relative paleointensity record of [Horng et al. \(2003\)](#) is used for the last 2 My because it has a high-quality, astronomically tuned chronology, and the record of Yamazaki and Oda (2005) is used from 3 to 2 Ma because it is one of few records that span this time interval. The climatic context of the geomagnetic variations is illustrated using the global stacked benthic $\delta^{18}\text{O}$ record of [Lisiecki and Raymo \(2005\)](#); oxygen isotope stages mentioned in the text are numbered (19 and 31).

the definition of the Quaternary Period, the boundary between the Neogene and Quaternary is closely associated with the Gauss–Matuyama reversal (Head et al., 2008; Ogg and Pillans, 2008), which makes paleomagnetic dating fundamentally important in defining the Quaternary (Figure 9). Compared to the other types of geomagnetic field variability discussed above, the number of full polarity reversals that have occurred during the Quaternary is small. Paleointensity-assisted chronology is therefore a welcome development for extending the usefulness of paleomagnetism in high-resolution Quaternary dating. Nevertheless, geomagnetic reversal stratigraphy is still extremely useful in Quaternary geochronology. For example, identification of the last reversal, namely, the Matuyama–Brunhes (M/B) boundary, provides an important marker in many studies including those concerning the presence of ancient humans (e.g., Muttoni et al., 2011; Parfitt et al., 2010). The M/B boundary occurred during a warm interglacial period, namely, marine isotope stage (MIS) 19 (e.g., Tauxe et al., 1996), which is increasingly being studied as a potential analog for future climate change (e.g., Tzedakis et al., 2009). The period between the Jaramillo subchron and the early Brunhes Chron is important for understanding the mid-Pleistocene transition (Figure 9) in which climate variability changed from low-amplitude orbital-obliquity-dominated (41 ka) to high-amplitude eccentricity-dominated (100-ka) cyclicity (e.g., Mudelsee and Schulz, 1997). Likewise, identification of the lower Jaramillo reversal provides an indication of the presence of MIS 31 (Figure 9), which was one of the warmest Quaternary interglacials and is another potential analog for future climate change (e.g., Scherer et al., 2008). The Gauss–Matuyama boundary coincided with intensification of Northern Hemisphere glaciation and is now fundamental to defining the Quaternary. Development of the astronomically calibrated geomagnetic polarity timescale means that identification of a polarity reversal provides a time marker that is precise to within 3 ka (Lourens et al., 2004). Thus, while reversal frequency is not high in the Quaternary, the globally synchronous nature of polarity reversals and the high chronological precision afforded by identification of reversals make magnetic polarity stratigraphy extremely powerful for addressing many problems in Quaternary science.

Paleomagnetic Dating in Quaternary Stratigraphy

High-fidelity paleomagnetic recording is most likely to occur in rocks and sediments that are magnetically dominated by SD or slightly coarser magnetic particles. Such particles are most often found in fine-grained volcanic rocks and in fine-grained organic-carbon-poor sediments. As summarized previously, a range of geomagnetic phenomena can be used to date Quaternary rocks and sediments. These phenomena are applicable to different age intervals and have variable requirements in terms of the types of materials from which suitable paleomagnetic records can be obtained.

Paleosecular Variation

Short-period secular variations can provide decadal to centennial age resolution. Achievement of this level of resolution is limited

to the Holocene, which is the only time interval for which we have high-quality regional PSV master curves. PSV studies are also best undertaken on continuously and rapidly deposited fine-grained lithogenic sediments. Extending such detailed records further back in time is often difficult because of changes in sedimentation in many depositional environments at the Last Glacial Maximum. Nevertheless, development of longer term PSV master curves represents an important target for future work.

Relative Geomagnetic Paleointensity

Millennial-scale geomagnetic variability provides a powerful global-scale phenomenon that can be used to develop paleointensity-assisted chronologies. Significant recent developments now enable the use of relative geomagnetic paleointensity throughout the Quaternary (e.g., Figure 9). Such records can only be obtained from fine-grained sediments that pass strict criteria for magnetic uniformity (e.g., Tauxe, 1993), which limits the applicability of this technique. Nevertheless, while it is ideal to constrain relative paleointensity chronologies with other high-quality chronostratigraphic information, paleointensity has been used as a principal dating technique in high-latitude marine sediments that are corrosive to calcium carbonate and which preclude determination of $\delta^{18}\text{O}$ -based chronologies (e.g., Sagnotti et al., 2001). Relative geomagnetic paleointensity analysis therefore represents an important advance in the paleomagnetic armory for dating Quaternary rocks and sediments.

Geomagnetic Excursions

Large-amplitude, short-period geomagnetic features such as excursions are among the most problematic geomagnetic phenomena for use in Quaternary stratigraphy. Without precise independent chronological constraints, such as a well-constrained relative paleointensity context, use of excursions to date sediments is an extremely precarious activity. For example, glacial deformation of sediments can produce anomalous paleomagnetic directions, which could be interpreted as a geomagnetic excursion without rigorous field and independent chronological constraints (e.g., Verosub, 1975). Excursions are among the most abused geomagnetic phenomena in Quaternary stratigraphy. We therefore urge extreme caution when interpreting anomalous paleomagnetic directions in terms of geomagnetic excursions. Lack of unique known geomagnetic field behavior that can be related to specific excursions leads us to recommend that they should not be used as a basis for dating sediments.

Geomagnetic Reversals

Unambiguous identification of a geomagnetic reversal provides a globally isochronous marker horizon. Magnetic reversals are therefore extremely powerful for dating. Many types of rocks and sediments, even those that are not ideal for studies of PSV or paleointensity, provide information about paleomagnetic polarity, which means that reversals have been used successfully in a wide range of Quaternary studies. However, the relatively low reversal frequency in the Quaternary (12 reversals in the last 2.59 Ma; Figure 9) means that

paleomagnetic reversal stratigraphy is a low-resolution dating tool. For many purposes, however, magnetic reversal stratigraphy can help to address important questions.

Conclusions

Diverse geological materials can be subjected to paleomagnetic investigation in Quaternary studies. Not all materials are suitable for paleomagnetic investigation, but careful study of appropriate materials can enable paleomagnetic documentation of decadal to millennial-scale geomagnetic variations, which have considerable potential for providing precise age constraints. These geomagnetic phenomena range from PSV (which is most useful in the Holocene), relative paleointensity variations (which are useful throughout the Quaternary), geomagnetic excursions (which are generally not useful for stratigraphic correlation unless accompanied by precise additional chronological constraints), and polarity reversals (which are useful between 2.59 and 0.78 Ma). A powerful aspect of using these geomagnetic variations is that they can generally provide chronological precision to within a few ka (or better in some instances). Paleomagnetic analysis therefore provides a key dating technique in Quaternary science.

See also: **Introduction:** History of Dating Methods; **Radiocarbon Dating:** Causes of Temporal ^{14}C Variations; Variations in Atmospheric ^{14}C .

References

- Bartletta F, St-Onge G, Stoner JS, Lajeunesse P, and Locat J (2010) A high-resolution Holocene paleomagnetic secular variation and relative palaeointensity stack from eastern Canada. *Earth and Planetary Science Letters* 298: 162–174.
- Barton CE and McElhinny MW (1981) A 10000 year geomagnetic secular variation record from three Australian maars. *Geophysical Journal of the Royal Astronomical Society* 67: 465–486.
- Barton CE and McElhinny MW (1982) Time-series analysis of the 10000 yr geomagnetic secular variation record from SE Australia. *Geophysical Journal of the Royal Astronomical Society* 68: 709–724.
- Bucur I (1994) The direction of the terrestrial magnetic field in France during the last 21 centuries. Recent progress. *Physics of the Earth and Planetary Interiors* 87: 95–109.
- Bullard EC, Freedman C, Gellman H, and Nixon J (1950) The westward drift of the Earth's magnetic field. *Philosophical Transactions of the Royal Society of London A* 243: 67–92.
- Channell JET, Hodell DA, Xuan C, Mazaud A, and Stoner JS (2008) Age calibrated relative paleointensity for the last 1.5 Myr at IODP Site U1308 (North Atlantic). *Earth and Planetary Science Letters* 274: 59–71.
- Dahlen FA and Simons FJ (2008) Spectral estimation on a sphere in geophysics and cosmology. *Geophysical Journal International* 174: 774–807.
- Gallet Y, Genevey A, and Le Goff M (2002) Three millennia of directional variation of the Earth's magnetic field in western Europe as revealed by archaeological artefacts. *Physics of the Earth and Planetary Interiors* 131: 81–89.
- Glatzmaier GA and Roberts PH (1995) A three-dimensional self-consistent computer simulation of a geomagnetic field reversal. *Nature* 377: 203–209.
- Gómez-Paccard M, Chauvin A, Lanos P, and Thiriot J (2008) New archeointensity data from Spain and the geomagnetic dipole moment in western Europe over the past 2000 years. *Journal of Geophysical Research* 113: B09103. <http://dx.doi.org/10.1029/2008JB005582>.
- Haines GV (1985) Spherical cap harmonic analysis. *Journal of Geophysical Research* 90: 2583–2591.
- Head MJ, Gibbard P, and Salvador A (2008) The Quaternary: Its character and definition. *Episodes* 31: 234–238.
- Heslop D (2007) Are hydrodynamic shape effects important when modelling the formation of depositional remanent magnetization? *Geophysical Journal International* 171: 1029–1035.
- Hornig CS and Roberts AP (2006) Authigenic or detrital origin of pyrrhotite in sediments?: Resolving a paleomagnetic conundrum. *Earth and Planetary Science Letters* 241: 750–762.
- Hornig CS, Roberts AP, and Liang WT (2003) A 2.14-million-year astronomically-tuned record of relative geomagnetic paleointensity from the western Philippine Sea. *Journal of Geophysical Research* 108: 2059. <http://dx.doi.org/10.1029/2001JB001698>.
- International Association of Geomagnetism and Aeronomy, Working Group V-MOD, Finlay CC, Maus S, et al. (2010) International Geomagnetic Reference Field: The eleventh generation. *Geophysical Journal International* 183: 1216–1230.
- Jackson A and Finlay C (2007) Geomagnetic secular variation and its applications to the core. In: Schubert G (ed.) *Treatise on Geophysics: Geomagnetism*, vol. 5, pp. 147–193. Oxford: Elsevier.
- Jackson A, Jonkers ART, and Walker MR (2000) Four centuries of geomagnetic secular variation from historical records. *Philosophical Transactions of the Royal Society of London A* 358: 957–990.
- King RF (1955) The remanent magnetism of artificially deposited sediments. *Monthly Notes of the Royal Astronomical Society Geophysical Supplement* 7: 115–134.
- King J and Peck J (2001) Use of paleomagnetism in studies of lake sediments. In: Last WM and Smol JP (eds.) *Tracking environmental change using lake sediments. Basin analysis, coring and chronological techniques*, vol. 1, pp. 371–390. Dordrecht, The Netherlands: Kluwer Academic Publishers.
- Korte M and Constable CG (2005) Continuous geomagnetic field models for the past 7 millennia. *Geochemistry, Geophysics, Geosystems* 6: Q02H16. <http://dx.doi.org/10.1029/2004GC000801>.
- Korte M, Donadini F, and Constable CG (2009) Geomagnetic field for 0–3 ka: 2. A new series of time-varying models. *Geochemistry, Geophysics, Geosystems* 10: Q06008. <http://dx.doi.org/10.1029/2008GC002297>.
- Laj C and Channell JET (2007) Geomagnetic excursions. In: Schubert G (ed.) *Treatise on Geophysics: Geomagnetism*, vol. 5, pp. 373–416. Oxford: Elsevier.
- Laj C, Kissel C, Mazaud A, Channell JET, and Beer J (2000) North Atlantic palaeointensity stack since 75 ka (NAPIS-75) and the duration of the Laschamp event. *Philosophical Transactions of the Royal Society of London A* 358: 1009–1025.
- Laj C, Kissel C, and Roberts AP (2006) Geomagnetic field behavior during the Iceland Basin and Laschamp geomagnetic excursions: A simple transitional field geometry? *Geochemistry, Geophysics, Geosystems* 7: Q03004. <http://dx.doi.org/10.1029/2005GC001122>.
- Langereis CG, Dekkers MJ, de Lange GJ, Paterne M, and van Santvoort PJM (1997) Magnetostratigraphy and astronomical calibration of the last 1.1 Myr from an eastern Mediterranean piston core and dating of short events in the Brunhes. *Geophysical Journal International* 129: 75–94.
- Lanos P, Le Goff M, Kovacheva M, and Schnepf E (2005) Hierarchical modelling of archaeomagnetic data and curve estimation by moving average technique. *Geophysical Journal International* 160: 440–476.
- Larson EE and Walker TR (1975) Development of chemical remanent magnetization during early stages of red-bed formation in Late Cenozoic sediments, Baja California. *Geological Society of America Bulletin* 86: 639–650.
- Lisiecki LE and Raymo ME (2005) A Pliocene-Pleistocene stack of 57 globally distributed benthic $\delta^{18}\text{O}$ records. *Paleoceanography* 20: PA1003. <http://dx.doi.org/10.1029/2004PA001071>.
- Lourens L, Hilgen F, Shackleton NJ, Laskar J, and Wilson D (2004) The Neogene period. In: Gradstein F, Ogg J, and Smith AG (eds.) *A Geologic Time Scale*, pp. 409–440. Cambridge: Cambridge University Press.
- Lund SP and Banerjee SK (1985) Late Quaternary paleomagnetic field secular variation from two Minnesota lakes. *Journal of Geophysical Research* 90: 803–825.
- Lund SP, Williams T, Acton GD, Clement B, and Okada M (2001) Brunhes Chron magnetic field excursions recovered from Leg 172 sediments. In: Keigwin LD, Rio D, Acton GD, and Arnold E (eds.) *Proceedings of the Ocean Drilling Program Scientific Results* 172: 1–18.
- Marshak S (2008) *Earth: Portrait of a Planet*, 3rd edn., pp. 832. New York: W.W. Norton.
- Merrill RT, McElhinny MW, and McFadden PL (1998) *The Magnetic Field of the Earth: Paleomagnetism, the Core, and the Deep Mantle*. San Diego: Academic Press.
- Merrill RT and McFadden PL (1994) Geomagnetic field stability: Reversal events and excursions. *Earth and Planetary Science Letters* 121: 57–69.
- Merrill RT and McFadden PL (2005) The use of magnetic field excursions in stratigraphy. *Quaternary Research* 63: 232–237.
- Mudelsee M and Schulz M (1997) The Mid-Pleistocene climate transition: Onset of 100 ka cycle lags ice volume build-up by 280 ka. *Earth and Planetary Science Letters* 151: 117–123.

- Muttoni G, Scardia G, Kent DV, et al. (2011) First dated human occupation of Italy at 0.85 Ma during the late Early Pleistocene climate transition. *Earth and Planetary Science Letters* 307: 241–252.
- Ogg JG and Pillans B (2008) Establishing Quaternary as a formal Period/System. *Episodes* 31: 230–233.
- Parfitt SA, et al. (2010) Early Pleistocene human occupation at the edge of the boreal zone in northwest Europe. *Nature* 466: 229–233.
- Pavlov V and Gallet Y (2005) A third superchron during the Early Paleozoic. *Episodes* 28: 78–84.
- Pavón-Carrasco FJ, Osete ML, and Torta JM (2010) Regional modeling of the geomagnetic field in Europe from 6000 BC to 1000 BC. *Geochemistry, Geophysics, Geosystems* 11: Q11008. <http://dx.doi.org/10.1029/2010GC003197>.
- Pavón-Carrasco FJ, Osete ML, Torta JM, and Gaya-Piqué LM (2009) A regional archaeomagnetic model for Europe for the last 3000 years, SCHA.DIF.3K: Applications to archaeomagnetic dating. *Geochemistry, Geophysics, Geosystems* 10: Q03013. <http://dx.doi.org/10.1029/2008GC002244>.
- Roberts AP (2008) Geomagnetic excursions: Knowns and unknowns. *Geophysical Research Letters* 35: L17307. <http://dx.doi.org/10.1029/2008GL034719>.
- Roberts AP, Chang L, Rowan CJ, Horng CS, and Florindo F (2011) Magnetic properties of sedimentary greigite (Fe₃S₄): An update. *Reviews of Geophysics* 49: RG1002. <http://dx.doi.org/10.1029/2010RG000336>.
- Roberts AP and Winkhofer M (2004) Why are geomagnetic excursions not always recorded in sediments? Constraints from post-depositional remanent magnetization lock-in modelling. *Earth and Planetary Science Letters* 227: 345–359.
- Sagnotti L, Macrí P, Camerlenghi A, and Rebecco M (2001) Environmental magnetism of Antarctic late Pleistocene sediments and interhemispheric correlation of climatic events. *Earth and Planetary Science Letters* 192: 65–80.
- Scherer RP, Bohaty SM, Dunbar RB, et al. (2008) Precession-driven warming and Antarctic ice sheet retreat during Marine Isotope Stage 31 (1.07 million years ago). *Geophysical Research Letters* 35: L03505. <http://dx.doi.org/10.1029/2007GL032254>.
- Schnepf E and Lanos P (2006) A preliminary secular variation reference curve for archaeomagnetic dating in Austria. *Geophysical Journal International* 166: 91–96.
- Singer BS, Hoffman KA, Schnepf E, and Guillou H (2008a) Multiple Brunhes Chron excursions recorded in the West Eifel (Germany) volcanics: Support for long-held mantle control over the non-axial dipole field. *Physics of the Earth and Planetary Interiors* 169: 28–40.
- Singer BS, Jicha BR, Kirby BT, Geissman JW, and Herrero-Bervera E (2008b) ⁴⁰Ar/³⁹Ar dating links Albuquerque Volcanoes to the Pringle Falls excursion and the Geomagnetic Instability Time Scale. *Earth and Planetary Science Letters* 267: 584–595.
- Skinner LC and Shackleton NJ (2005) An Atlantic lead over Pacific deep-water change across Termination I: Implications for the application of the marine isotope stage stratigraphy. *Paleoceanography* 24: 571–580.
- Snowball I, Zillén L, Ojala A, Saarinen T, and Sandgren P (2007) FENNOSTAK and FENNOPRIS: Varve dated Holocene palaeomagnetic secular variation and relative palaeointensity stacks for Fennoscandia. *Earth and Planetary Science Letters* 255: 106–116.
- Stoner JS, Laj C, Channell JET, and Kissel C (2002) South Atlantic and North Atlantic geomagnetic paleointensity stacks (0–80 ka): Implications for inter-hemispheric correlation. *Quaternary Science Reviews* 21: 1141–1151.
- St-Onge G, Stoner JS, and Hillaire-Marcel C (2003) Holocene paleomagnetic records from the St. Lawrence Estuary, eastern Canada: Centennial- to millennial-scale geomagnetic modulation of cosmogenic isotopes. *Earth and Planetary Science Letters* 209: 113–130.
- Tarduno JA, Cottrell RD, and Smirnov AV (2006) The paleomagnetism of single silicate crystals: Recording geomagnetic field strength during mixed polarity intervals, superchrons, and inner core growth. *Reviews of Geophysics* 44: RG1002. <http://dx.doi.org/10.1029/2005RG000189>.
- Tarduno JA, Cottrell RD, Watkeys MK, et al. (2010) Geodynamo, solar wind, and magnetopause 3.4 to 3.45 billion years ago. *Science* 327: 1238–1240.
- Tauxe L (1993) Sedimentary records of relative paleointensity of the geomagnetic field: Theory and practice. *Reviews of Geophysics* 31: 319–354.
- Tauxe L (2005) Inclination flattening and the geocentric axial dipole hypothesis. *Earth and Planetary Science Letters* 233: 247–261.
- Tauxe L (2010) *Essentials of Paleomagnetism*. Berkeley: University of California Press.
- Tauxe L, Herbert T, Shackleton NJ, and Kok YS (1996) Astronomical calibration of the Matuyama–Brunhes boundary: Consequences for magnetic remanence in marine carbonates and the Asian loess sequences. *Earth and Planetary Science Letters* 140: 133–146.
- Tauxe L and Kent DV (1984) Properties of a detrital remanence carried by hematite from study of modern river deposits and laboratory redeposition experiments. *Geophysical Journal of the Royal Astronomical Society* 77: 543–561.
- Tauxe L and Yamazaki T (2007) Paleointensities. In: Schubert G (ed.) *Treatise on Geophysics. Geomagnetism*, vol. 5, pp. 509–564. Oxford: Elsevier.
- Thébault E, Schott JJ, and Manda M (2006) Revised spherical cap harmonic analysis (R-SCHA): Validation and properties. *Journal of Geophysical Research* 111: B01102. <http://dx.doi.org/10.1029/2005JB003836>.
- Thellier E and Thellier O (1959) Sur l'intensité du champ magnétique terrestre dans la passé historique et géologique. *Annales de Geophysique* 15: 285–376.
- Thompson R and Berglund B (1976) Late Weichselian geomagnetic 'reversal' as a possible example of the reinforcement syndrome. *Nature* 259: 490–491.
- Turner GM and Thompson R (1981) Lake sediment record of the geomagnetic secular variation in Britain during Holocene times. *Geophysical Journal of the Royal Astronomical Society* 65: 703–725.
- Turner GM and Thompson R (1982) Detransformation of the British geomagnetic secular variation record for Holocene times. *Geophysical Journal of the Royal Astronomical Society* 70: 789–792.
- Tzedakis PC, Raynaud D, McManus JF, Berger A, Brovkin V, and Kiefer T (2009) Interglacial diversity. *Nature Geoscience* 2: 751–755.
- Valet JP, Meynadier L, and Guyodo Y (2005) Geomagnetic dipole strength and reversal rate over the past two million years. *Nature* 435: 802–805.
- Vandamme D (1994) A new method to determine paleosecular variation. *Physics of the Earth and Planetary Interiors* 85: 131–142.
- Verosub KL (1975) Paleomagnetic excursions as magnetostratigraphic horizons: A cautionary note. *Science* 190: 48–50.
- Verosub KL (1977) Depositional and post-depositional processes in the magnetization of sediments. *Reviews of Geophysics and Space Physics* 15: 129–143.
- Yamazaki T and Oda H (2005) A geomagnetic paleointensity stack between 0.8 and 3.0 Ma from equatorial Pacific sediment cores. *Geochemistry, Geophysics, Geosystems* 6: Q11H20. <http://dx.doi.org/10.1029/2005GC001001>.
- Zhao X and Roberts AP (2010) How does the Chinese loess become magnetized? *Earth and Planetary Science Letters* 292: 112–122.
- Ziegler LB, Constable CG, Johnson CL, and Tauxe L (2011) PADM2M: A penalized maximum likelihood model of the 0–2 Ma palaeomagnetic axial dipole moment. *Geophysical Journal International* 184: 1069–1089.

Relevant Websites

http://www.gfz-potsdam.de/portal/gfz/Struktur/Departments/Department+2/sec23/topics/models/CALS3k_ARCH3k_SED3k-CALS3K

<http://www.gfz-potsdam.de/portal/gfz/Struktur/Departments/Department+2/sec23/topics/models/CALS7K-CALS7K>

<http://jupiter.ethz.ch/cfinlay/gufm1.html> – GUFM1, including animations of global secular variation for the last 400 years.

<http://www.ngdc.noaa.gov/AGA/vmod/igrf.html> – IGRF.

<http://pc213fis.fis.ucm.es/scha.dif.3k/index.html> – SCHA3K.

<http://pc213fis.fis.ucm.es/scha.dif.8k/index.html> – SCHA8K.

<http://www.intermagnet.org> – INTERMAGNET.

GLACIAL CLIMATES

Contents

Biosphere Feedbacks

Effects of Atmospheric Dust

Thermohaline Circulation

Volcanic and Solar Forcing

Biosphere Feedbacks

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“Life is not an external or accidental phenomenon of the Earth’s crust. It is closely bound to the structure of the crust, forms part of its mechanism, and fulfils functions of prime importance to the existence of this mechanism. Without life, the crustal mechanism would not exist.”

(Vladimir Vernadsky, *The Biosphere*, 1926)

Introduction

A current paradigm in plant ecology is that the geographical distribution of dominant land plant forms is mainly controlled by climate (Woodward, 1987). If climate alters, vegetation cover closely follows the change in the climatic patterns. In turn, changes in spatial distribution and composition of terrestrial vegetation (as well as marine biology) alter climate through modifications of heat and water fluxes, atmospheric gases, and aerosol composition. Biogeochemical and biophysical mechanisms form numerous feedbacks between biosphere and climate that were active during Earth’s geological past and continue to operate at present.

Quaternary changes in terrestrial and marine biology are recorded in sediments around the planet. A picture emerging from pollen and macrofossil reconstructions is that forest cover was greatly reduced at the maxima of Pleistocene glaciations, while during the warmest phases of interglacials, expanded forests covered almost a half of the ice-free land surface. These dramatic changes in forest cover had several important implications for climate dynamics. Coldness and aridity of glacial climates were amplified by land-cover changes through increase in land-surface albedo (the fraction of solar radiation reflected back into the atmosphere), decrease in transpiration, and enhanced dust formation on exposed bare ground.

Fewer forests and lower productivity during glacial periods should have led to reduced terrestrial biomass and soil carbon

storages, and the consequent release of carbon dioxide (CO₂) to the atmosphere that slightly offset the radiative cooling. This stabilizing effect is invisible in ice-core CO₂ records due to counteracting amplifying feedback between temperature and greenhouse gases that led to the lowering of atmospheric CO₂ levels during glacial periods. One of the hypothesized biological mechanisms that receive some support from proxy data is an iron fertilization effect on glacial marine productivity due to increased levels of dust depositions. The role of slow biogeochemical processes such as coral reef growth and continental weathering in atmospheric CO₂ dynamics is not yet well constrained.

Main Mechanisms of Climate–Biosphere Interaction

Photosynthesis

The process of photosynthesis that has existed on Earth for more than 3 billion years is a focal point of climate–biosphere interaction. Terrestrial plants capture CO₂ from the atmosphere while marine phytoplanktons take up carbon from seawater, which exchanges CO₂ with the atmosphere (House et al., 2005). Respiration of living organisms returns carbon to the atmosphere. In the ocean, a part of dead organic matter (so-called export production) escapes the ocean surface layer and sinks toward the ocean floor. Because of long turnover time of deep ocean waters, the amount of carbon stored in the ocean due to this process (called the biological pump) is comparable to the organic carbon storage on land. Changes in terrestrial biomass and soil organic matter, as well as in volume and composition of marine export production, affect atmospheric CO₂ concentration directly on annual-to-millennial timescales. Altered CO₂ concentration modifies atmospheric radiative properties and, via climatic change and direct CO₂ effect, feeds back on ecosystem production.

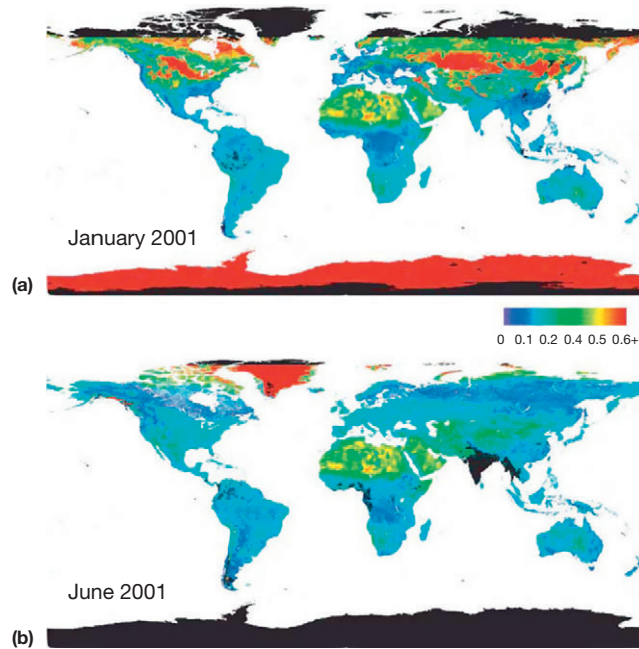


Figure 1 Map of land-surface albedo captured by the MODIS Satellite Instrument (Lucht et al., 2000; Schaaf et al., 2002). Regions where there were no data available, for example, due to clouds, are indicated in black. (a) January 2001: In the northern areas during the winter season, snow albedo is very high (up to 0.8, red). The boreal forest belt can be clearly seen in blue and green as trees mask snow, reduce albedo, and warm the surface air during the snow season. (b) June 2001: In comparison with January 2001, the northern land areas have a much lower albedo due to the absence of snow. In this map, the area with the highest albedo (up to 0.5, green and yellow) is the Sahara desert. High albedo in this region suppresses rainfall during the summer rainy season. Reproduced from House JI and Brovkin V (2005) Climate and air quality. In: Hassan R, Scholes R, and Ash N (eds.) *Ecosystems and Human Well-Being: Current State and Trends*, vol. 1, pp. 355–390. Washington, DC: Island Press, with permission.

Land-Surface Biophysics

In the planetary boundary layer up to 2.5 km above Earth's surface, vertical profiles of temperature and humidity strongly depend on the partitioning of energy between sensible heat and latent heat. This partitioning is, to a large extent, controlled by terrestrial ecosystems through biophysical mechanisms (Pitman et al., 2004). Over dry, bare ground, the energy is transported via sensible heat, resulting in relatively high surface-air temperatures. Vegetation canopies transpire water extracted from the root zone, increase latent heat flux upwards, and cool the surface air during the day. Transpiration is tightly coupled to photosynthesis. Both transpiration and photosynthesis are regulated by the opening of stomatal pores on the leaf's surface. Low atmospheric CO_2 levels during glacial periods led to increased stomatal conductance, lower water use efficiency (productivity per unit volume of water transpired), and reduced levels of photosynthesis.

Increased air humidity due to daily transpiration reduces heat loss in the near-surface atmosphere (greenhouse effect), helping to smooth diurnal fluctuations in temperature. Vegetation canopies also intercept part of the rainfall and effectively reduce runoff, especially during heavy rain events. Moisture that is transported to continental interiors has been transpired and precipitated several times on its way from the ocean; therefore, reduction in vegetation cover during glacial periods enhances intracontinental aridity.

Vegetation affects surface albedo, and in most cases, it reduces albedo compared to bare ground and snow. Forests are more effective than grasslands in trapping solar radiation.

This effect is particularly strong in snow-covered regions where trees extend above the snow pack, while herbaceous vegetation is covered (see Figure 1). The snow-masking effect of tree cover was likely more important in glacial climates when snow cover lasted longer in spring.

Long-Term Carbon Cycle Processes

Carbonate sedimentation

Marine-calcifying organisms such as corals and coccolithophorids have skeletons or shells of calcium carbonate (CaCO_3). After an organism's death, CaCO_3 does not dissolve but forms reefs or drops to the sea floor forming sediments. In the deep seas, carbonates are well preserved above the lysocline (about 3–4 km depth), but they dissolve below this depth. On timescales of 5000–10 000 years, carbonate sedimentation in the ocean is constrained by weathering. Since carbonate deposition in continental shelf regions (e.g., coral reef production) changes with sea level during glacial–interglacial cycles, this affects deep-sea sedimentation and leads to changes in the lysocline depth. A temporal imbalance between weathering and sedimentation causes changes in the total inventory of ocean alkalinity. Acidic water dissolves less CO_2 than basic water, and in this way, changes in CaCO_3 sedimentation affect atmospheric CO_2 concentration. This process is complicated by the fact that export flux contains not only carbonate but also organic matter that becomes mineralized on the seafloor. The latter process releases CO_2 , making sediment pore water more acidic and leading to

carbonate dissolution. Changes in organic-to-carbonate ratio (rain ratio) in export flux affect CaCO_3 sedimentation considerably, and, through lysocline dynamics, the atmospheric CO_2 level.

Anaerobic decomposition

Some dead organic matter is trapped in wetlands or marine sediments in anaerobic conditions. Decomposition of this organic matter by methane (CH_4)-producing bacteria may lead to a release of CH_4 . Most of this CH_4 oxidizes in the water column, but a part escapes to the atmosphere. Atmospheric CH_4 concentration is small but important for climate since CH_4 is 20–30 times more potent a greenhouse gas than CO_2 . CH_4 sink in the atmosphere is regulated by atmospheric chemistry that in turn is affected by plant emissions of volatile organic compounds (VOCs) such as isoprene.

Numerical Modeling as a Methodology of Feedback Analysis

The climatic effect of planting a single tree is negligibly small. Even for large-scale land-cover changes, their climatic consequences are very difficult to detect in observations because several climatic forcings operate simultaneously in the presence of substantial climate variability. In addition, regional climate can be influenced not only by local changes but also by changes in remote regions. Carrying out numerical experiments with climate and ecosystem models is one of the principle methods of testing feedback between climate and biosphere. In the usual design of sensitivity experiments, the simulations are performed twice, once when feedback is neglected and a second time when feedback is included. A difference between these two simulations indicates the importance of feedback for climate. Biophysical feedbacks studied with global climate models are hereafter referred to as biogeophysical feedbacks to stress their significance for global-scale processes.

Hot Spots of Climate–Biosphere Interaction in Quaternary

The Holocene

'Green' Sahara

Multiple proxy data reveal that during the so-called Holocene optimum, ca. 9000–6000 years BP, the vegetation cover of the Sahel was greatly extended to the north. During that period, the Sahara's climate was much wetter; many lakes and rivers, including the greatly extended Lake Chad, were present in the region. A conventional explanation of the 'green' Sahara phenomenon in the mid-Holocene, supported by many modeling experiments, is based on the long-term changes in Earth's orbital parameters. In the early-to-mid Holocene, the northern hemisphere received considerably more solar radiation during the summer. In the northern subtropical regions, this led to stronger warming over the continent than over the ocean, an increased temperature gradient between land and ocean, and, consequently, intensified monsoon-type circulation in summer which led to increased rainfall over the Sahel/Sahara region.

The North African desert differs from many other subtropical deserts. The radiative balance in the region is negative, that is, there is a net radiative heat loss over the desert. This loss is particularly due to the high albedo of the desert, as up to 50%

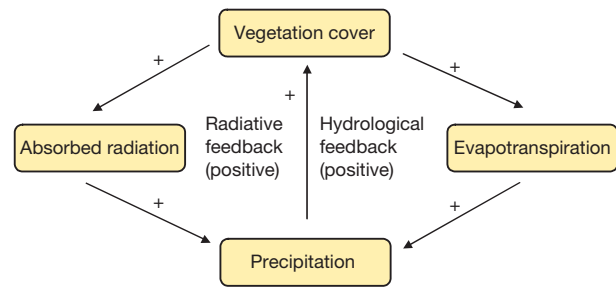


Figure 2 Idealized sketch of atmosphere–vegetation interactions in the Sahel/Sahara region. The signs attached to the arrows indicate the effect of the outgoing box on the incoming box; for example, increased vegetation cover leads to more radiation absorbed at the land surface. Both radiative and hydrological feedbacks are positive, that is, they amplify external changes and therefore may cause instabilities. Reproduced from Claussen M, Cox PM, Zeng X, et al. (2004) The global climate. In: Kabat P, Claussen M, Dirmeyer PA, et al. (eds.) *Vegetation, Water, Humans and the Climate: A New Perspective on an Interactive System*. Global Change: The IGBP Series, pp. 33–57. Berlin, Germany: Springer, with permission.

of incoming radiation is reflected back into space (Figure 1). The radiative mechanism for the Sahara was first explored by Jule Charney in the 1970s. He pointed out that the heat loss due to high albedo leads to a temperature gradient and induces an atmospheric circulation which maintains the sinking motion of dry atmospheric masses and suppresses rainfall over the region. Low precipitation results in little vegetation cover, and the surface albedo is determined by bare ground with a high albedo. This positive feedback supports a desert that is self-sustaining. Alternatively, if there is more precipitation, there is more vegetation. Vegetation canopy is darker than sand, so the albedo is lower, the surface temperature is higher, and the gradient in temperature between land and ocean increases, amplifying monsoon circulation and upward motion over the desert. As a result, the summer rainfall in the region increases. This radiative feedback loop gets additional support through hydrological feedback, as vegetation recycles water to the atmosphere, reduces runoff, and enhances rainfall in comparison with bare ground (Figure 2).

Both radiative and hydrological feedbacks are positive, that is, they amplify changes induced by external forcings or internal variability. Could these positive feedbacks have affected the stability of the regional climate in the past? In the early 1990s, Martin Claussen applied coupled atmosphere–vegetation model ECHAM2–BIOME to investigate the effect of different initial land-surface conditions on regional climate. In these simulations, vegetation cover responded to climate simulated by the model and vice versa. In the first simulation with present-day boundary conditions (Figure 3(a)), which started from the present-day potential vegetation, the desert in North Africa remained stable. However, when the Sahara was initially covered by forest in simulation B (Figure 3(b)), the system converged to another solution, called the 'green' Sahara. The Western Sahara is covered by a mixture of shrubs and grass; in addition, vegetation cover in the Sahel region is enhanced. Both desert and green solutions are stable for present-day conditions. At the same time, the coupled model revealed the only steady solution, 'green' Sahara, for the mid-Holocene orbital forcing (Figure 3(c)). In the dynamical system theory,

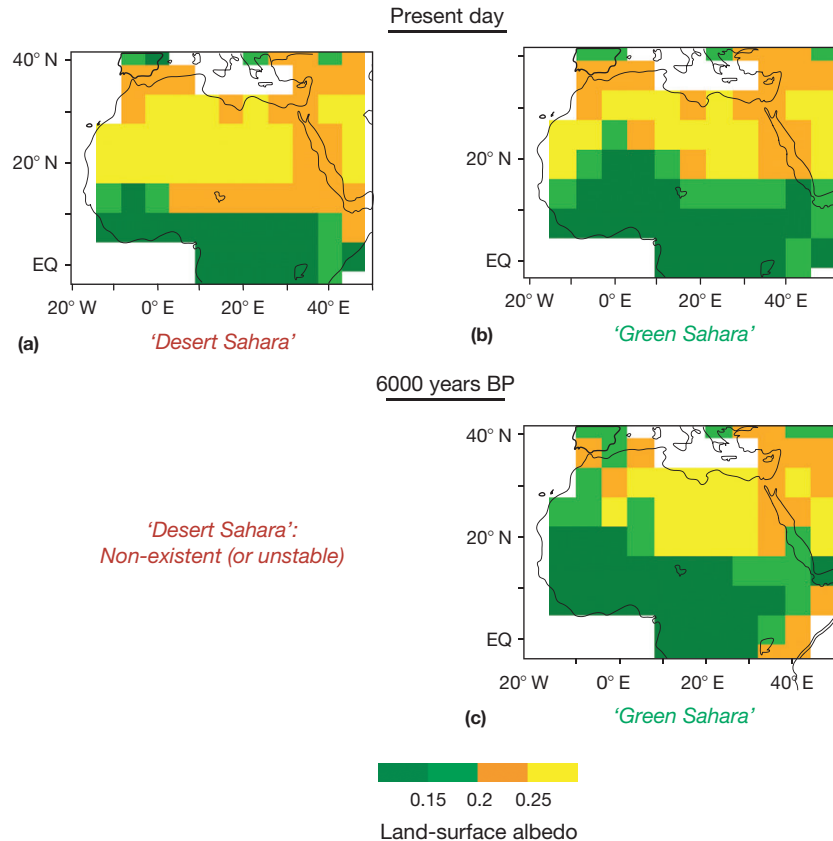


Figure 3 Summary of results obtained with an interactive atmosphere–vegetation ECHAM/BIOME model. For present-day climate, two states are stable in the model: desert Sahara as observed now (a) and ‘green Sahara’ state with enhanced vegetation cover in the Western Sahara (b) in accordance with Claussen (1994). (c) For 6000 years ago, ‘green Sahara’ is the only stable state in the model (Claussen and Gayler, 1997). Lower albedo corresponds to a higher fraction of vegetation cover. EQ, equator. Reproduced from Brovkin V, Claussen M, Petoukhov V, and Ganopolski A (1998) On the stability of the atmosphere–vegetation system in the Sahara/Sahel region. *Journal of Geophysical Research-Atmospheres* 103: 31613–31624, with permission from American Geophysical Union.

the fact that the system has one steady state for one parameter value (mid-Holocene insolation) and two states for another value (present-day orbital parameters) means that in between these points the system has undergone a bifurcation that could be associated with abrupt climate changes. Indeed, proxy data suggest abrupt changes of vegetation cover in the Western Sahara between 5000 and 6000 years BP (deMenocal et al., 2000). Later simulations with another coupled model that included interactive ocean, ECBILT-CLIO-VECODE, supported the fact that the system underwent bifurcation around 4000–6000 years BP as the green state lost its stability (Renssen et al., 2003). Some other coupled models do not show multiple states, but they all reveal a high sensitivity of rainfall in the Sahel/Sahara region to vegetation cover. In summary, strong positive vegetation–atmosphere feedback in North Africa is an important amplifier of externally induced changes in the region during the Quaternary. In particular, this feedback had contributed to abrupt desertification of the region about 5000–6000 years ago.

Northern treeline dynamics

During the mid-Holocene, the northern tundra–forest boundary was located north of its present position in many regions.

In particular, in central Siberia, the treeline (the forest limit) was shifted to the north by up to 200 km (Bigelow et al., 2003), presumably because of warmer summers, as treeline position corresponds reasonably well with July temperature isotherms of about 10–12 °C. In the northern high latitudes, changes in forest cover affect regional climate in several ways. Forests, even deciduous ones, significantly reduce the albedo of snow-covered surfaces. This is the basis for the radiative feedback between forest and surface-air temperature. Increased forest cover within large regions leads to decreased surface albedo during the snow season. This results in elevated air temperature and earlier snow melt, and a longer and warmer growing season, which favors further increase in forest area. This feedback (sometimes called taiga–tundra feedback) amplifies the climate system response to the original external forcing (Figure 4). The rate at which forest-cover changes depends on many local factors such as disturbances (fire), permafrost, soil parent material, and pH. The radiative feedback is dampened during the growing season, when trees have a denser, more productive canopy. Trees also have deeper roots than herbaceous plants and moss and transpire more water. This cools the surface air in forests as compared to tundra during summer. The hydrological feedback, although negative, is secondary

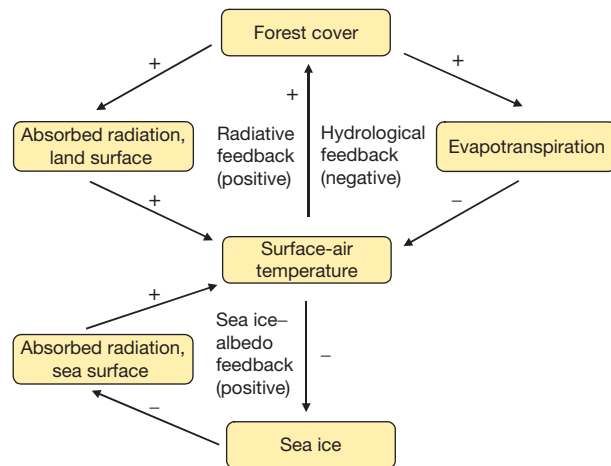


Figure 4 Simplified scheme of climate–vegetation interactions in the northern high latitudes. The signs attached to the arrows indicate the effect of the outgoing box on the incoming box; for example, increased forest (tree) cover leads to more radiation absorbed at the land surface during snow-covered periods. Changes in surface-air temperature due to forest-cover changes are amplified through sea ice–albedo feedback in the Arctic.

to the radiative feedback on the annual average. Besides, enhanced transpiration from deeper soil horizons improves soil drainage and aeration of soil mineral layers, which increases viability of tree establishment. In the Arctic environment, the radiative feedback of forests is closely linked to sea ice–albedo feedback. Elevated surface-air temperature due to increased forest cover leads to higher sea-surface temperature (SST) and decreased sea-ice area, decreased sea-surface albedo, and, finally, elevated surface-air temperature (Figure 4). Model simulation suggests that taiga–tundra feedback and its amplification by sea ice–albedo feedback led to additional warming of northern high latitudes in response to the changes in orbital forcing during the mid-Holocene (Figure 5).

Historical land-cover changes

At present, about one-third of the Earth's land cover has been modified by agricultural and forestry activities. Human-induced land-cover changes probably began in the early Holocene with development of slash-and-burn agriculture. The magnitude of these changes and their temporal dynamics on large regional scales in the preindustrial period are mostly unknown. Most models have been used to evaluate the climatic effect of reconstructed potential vegetation versus present-day land cover, and some models have been applied for testing the climatic effect of the last millennium land-cover changes using simplified assumptions about such changes prior to AD 1700. Most model simulations agree that historical deforestation led to some increase in atmospheric CO_2 and consequent warming. At the same time, an increase in albedo due to deforestation of regions covered with snow in winter, such as northern Europe, led to a radiative cooling (Figure 6). On a global scale, biogeophysical cooling was likely compensated by biogeochemical warming, but consequences for regional climate could be substantial. Biogeophysical effects due to the increasing deforestation rate could have contributed to an observed stabilization of

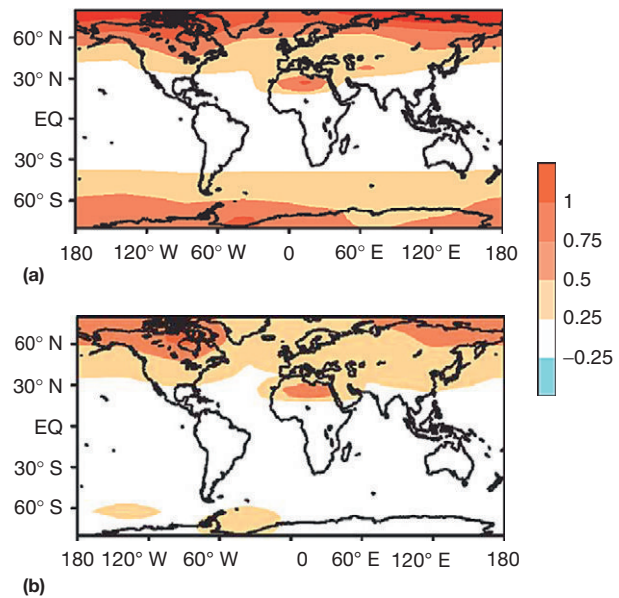


Figure 5 Role of vegetation feedback in the mid-Holocene climate in terms of annually averaged surface-air temperature ($^{\circ}\text{C}$) in CLIMBER-2 model simulations of the mid-Holocene climate (6000 years BP). (a) Difference between the mid-Holocene and the present-day simulations. (b) Contribution of the biogeophysical feedback (difference between mid-Holocene simulations with interactive vegetation and vegetation prescribed from present-day simulation). EQ, equator. Adapted from Ganopolski A, Kubatzki C, Claussen M, Brovkin V, and Petoukhov V (1998) The influence of vegetation–atmosphere–ocean interaction on climate during the mid-Holocene. *Science* 280: 1916–1919, with permission.

the temperature during the second half of the nineteenth century and damped the CO_2 warming in the twentieth century.

Last Glacial Maximum

The Last Glacial Maximum (LGM) was characterized by cold and dry conditions that led to a great reduction in forest cover in the Northern Hemisphere. Boreal evergreen forests and temperate deciduous forests were fragmented, while European and East Asian steppes were greatly extended (Prentice et al., 2000). Tropical forests in the Americas and Africa were reduced as well. These changes toward herbaceous and sparser vegetation cover had considerable implications for climate.

Biogeophysical feedbacks

Model simulations using glacial boundary conditions suggest that reduction in forest cover amplified LGM coldness and aridity. Southward displacement and fragmentation of forest biomes in North America and Eurasia led to increased albedo during the snow period and a regional decrease in temperature by several degrees (Figure 6). Global cooling effect of reduced vegetation cover during the LGM is estimated in 0.6°C (Crucifix and Hewitt, 2005; Jahn et al., 2005). Reduction in forest cover led to less transpiration and increased aridity of the continental interiors. Presumably, physiological effects of lowered CO_2 levels during the LGM resulted in increased canopy conductance of C_3 plants and a consequent drop in

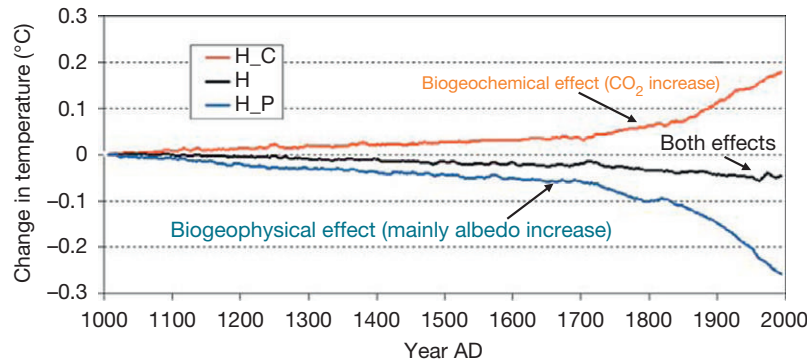


Figure 6 Effects of historical land-cover changes on global mean annual temperature ($^{\circ}\text{C}$) simulated by CLIMBER-2 model. H_P , biogeophysical effect only (mainly due to albedo changes); H_C , biogeochemical effect only in response to atmospheric CO_2 increase; H , both biogeophysical and biogeochemical effects. Shown here are the differences between simulations with and without land-cover changes. Reproduced from Brovkin V, Sitch S, von Bloh W, et al. (2004) Role of land cover changes for atmospheric CO_2 increase and climate change during the last 150 years. *Global Change Biology* 10: 1–14, with permission of Blackwell Publishing.

water-use efficiency (water used per unit of photosynthetic production; **Figure 7**; Harrison and Prentice, 2003).

Terrestrial CO_2 feedback

Cold and dry climates as well as low CO_2 levels led to decreased plant productivity and biomass storage during the LGM. Land areas in the Arctic were more extensive due to lowered sea levels, but regional ecosystems were not very productive. To some degree, reduction in boreal forests was compensated by the establishment of new forests along the exposed shelf regions in tropical regions such as southeast Asia, but on a global scale, terrestrial carbon storage was drastically reduced. Vegetation models applied to LGM climate-change scenarios reveal a range of decrease in terrestrial storages, from several hundreds to thousands of petagrams of carbon (PgC ; 10^{15} g C). A central estimate of about 600–700 PgC is higher than changes suggested by a decline in deep ocean $\delta^{13}\text{C}$ (about 500 PgC). The later proxy should be carefully interpreted since oceanic $\delta^{13}\text{C}$ is also affected by changes in oceanic circulation and marine productivity. Decrease in northern peat storage due to expansion of ice sheets and increased aridity also contributed to the release of land carbon to the atmosphere, but the magnitude and timing of peat changes are uncertain. A reduction in the terrestrial carbon pool during glacial periods provides a negative (stabilizing) feedback to climate by increasing atmospheric CO_2 concentration. This feedback is rather weak since most of the released terrestrial carbon ends up in the ocean. After carbonate compensation on timescales of 5000–10000 years, 500 PgC of land carbon source corresponds to about a 20 parts per million by volume (ppmv) increase in atmospheric CO_2 concentration (Archer et al., 2000).

Dust feedback

In semiarid regions, reduced vegetation cover means more bare ground exposed to wind erosion. During glacial periods, these sparsely vegetated regions were a source of mineral aerosols for the atmosphere. Higher eolian dust transport during glacial periods is supported by sediment data (Kohfeld et al., 2005). Mineral aerosols diffuse light and reduce the amount of solar

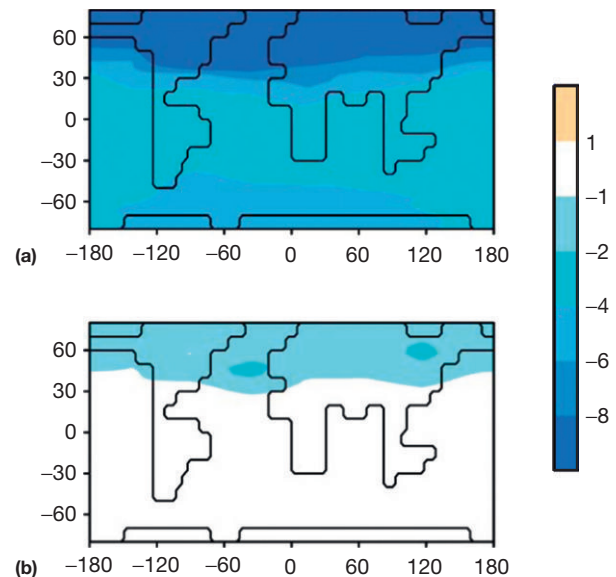


Figure 7 Role of biogeophysical feedback in the Last Glacial Maximum (LGM) climate as simulated by CLIMBER-2 model. Shown are annually averaged surface-air temperature differences ($^{\circ}\text{C}$). (a) A difference between the LGM and the present-day simulations. (b) Change in temperature due to the biogeophysical feedback. Adapted from Jahn A, Claussen M, Ganopolski A, and Brovkin V (2005) Quantifying the effect of vegetation dynamics on the climate of the last glacial maximum. *Climate of the Past* 1: 1–7, with the authors permission.

radiation that reaches the surface. Radiative transfer models estimate that increased atmospheric dust concentrations during the LGM led to about $2\text{--}3 \text{ W m}^{-2}$ reduction in radiative forcing in the tropics, which roughly corresponds to the radiative effect of lower glacial CO_2 levels in the region (Claquin et al., 2003). As a result, sparser vegetation cover amplified LGM cooling through the dust feedback.

Marine biota feedback

Several hypotheses aimed at explaining glacial–interglacial changes in atmospheric CO_2 invoke changes in marine

ecosystem structure and export production. Sub-Antarctic proxy data suggest that export production in the sub-Antarctic Atlantic was enhanced during the LGM, presumably because of iron fertilization in response to increased eolian dust deposition. Some other regions, such as the Pacific, show few changes, while productivity in Antarctic Ocean was reduced. Iron fertilization could be responsible for some drawdown of atmospheric CO_2 levels at the LGM, but it cannot explain the whole range of 80–100 ppmv change (Kohfeld et al., 2005). Increased organic to CaCO_3 ratio in export production (e.g., due to silica leakage) could have substantially lowered the atmospheric CO_2 through the sediment response. However, recent data show that the ratio of organic to CaCO_3 flux that reached the seafloor is almost constant. Coral reef production was reduced during the LGM in response to lowered sea levels and surface cooling (Kleypas, 1997). Reduced shallow-water sedimentation could have led to an increase in oceanic total alkalinity, but this effect might also have been offset by decreased continental weathering during the LGM. In summary, the marine biosphere was likely to have provided positive feedback to the atmosphere by amplifying LGM cooling through lowering of atmospheric CO_2 , but the magnitude of this feedback and the precise mechanisms are still not well known.

Weathering feedback

On a timescale longer than several thousand years, atmospheric CO_2 is strongly affected by changes in carbonate and silicate weathering. Increased weathering lowers atmospheric CO_2 concentration through oceanic carbonate compensation. Vascular plants contribute to the weathering as roots' respiration and litter mineralization lead to elevated soil CO_2 concentrations and the formation of carbonic acid, which enhances weathering of parent soil material. Model estimates of changes in weathering rates during the LGM are equivocal as several factors influencing global weathering flux, such as increased ice-sheet areas, shelf exposure, and decreased runoff, almost nullify each other (Munhoven, 2002).

Methane and atmospheric chemistry feedback

During the LGM, atmospheric CH_4 concentration (about 400 ppbv) was much lower than in the Holocene (600–750 ppbv). Decreased concentration of CH_4 led to small but not negligible additional cooling. The plausible explanation of the lowered CH_4 concentration is that terrestrial CH_4 sources (e.g., wetlands and biomass burning) were lower during the LGM. In particular, the extent of wetlands was reduced in the cooler and drier LGM climate, especially in the northern high latitudes, while this reduction might have been offset by the establishment of wetlands on exposed tropical shelves. Another important factor could be a change in atmospheric chemistry. Model simulations suggest that emissions of VOC were lower during the LGM due to reduction in forest cover, resulting in increased atmospheric concentration of the OH radical (an atmospheric 'detergent'), and this may have led to an increased sink for CH_4 in the atmosphere (Valdes et al., 2005). In summary, biospheric feedback amplified LGM cooling through lowering of CH_4 concentration,

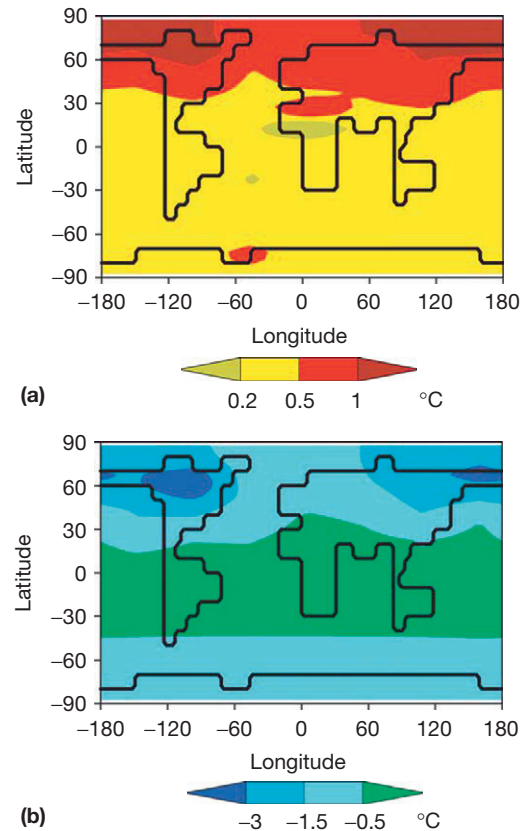


Figure 8 Role of biogeophysical feedback in the climate of last interglacial and glacial inception as simulated by CLIMBER-2 model. Shown are annual temperature differences ($^{\circ}\text{C}$) between transient experiments with interactive vegetation and vegetation cover prescribed from present-day simulation. Ice sheets are interactive. (a) Warming due to interactive vegetation at 125 000 years BP. (b) Cooling induced by vegetation changes at 115 000 years BP. Reproduced from Calov R, Ganopolski A, Petoukhov V, et al. (2005) Transient simulation of the last glacial inception. Part II: Sensitivity and feedback analysis. *Climate Dynamics* 24: 563–576, with permission of Springer Science and Business Media.

while CH_4 -regulating mechanisms involving VOC emissions are not yet fully explored.

The Last Interglacial and Glacial Inception

During the last interglacial, the increase in summer insolation in the northern hemisphere was stronger than during the Holocene. Pollen evidence and model simulations suggest that summer climate in northern regions was substantially warmer than at present and that forest cover was more extensive in the temperate and high northern latitudes. The vegetation feedback amplified the warming in model simulations (Figure 8(a)). This temperature increase anomaly is mainly due to the northward advance of boreal forest during the warmer climate of the last interglacial. Conversely, the later cooling trend triggered by the decreasing boreal summer insolation was facilitated by the vegetation response (Figure 8(b)). Most climate model simulations suggest that a southward shift

of northern treeline is crucial for the initiation of glacial inception: if vegetation cover is prescribed to the values simulated for the last interglacial, then cooling in the north due to declining insolation is not sufficient to initiate ice sheet growth. Treeline retreat amplifies cooling through increased albedo, fostering permanent snow cover on land during the glacial inception.

Conclusions

Analysis of climate–biosphere feedbacks and their role in climate system dynamics during the Quaternary is a new, rapidly developing scientific field. Hypotheses about roles of biological mechanisms in past climate dynamics usually emerge from the analysis of geological data, but quantification of feedback is impossible without modeling. Today, the main methodology of feedback analysis is numerical modeling using climate system models of different complexities, from high-resolution general circulation models (GCMs) to simplified conceptual models. Model results are tested against proxy data for biogeochemistry and vegetation cover.

A picture emerging from model simulations and proxy data analysis is that biospheric feedback played a substantial role in the evolution of Quaternary climate, but many details are uncertain. In some cases, such as glacial inception, model simulations agree that the dynamics of northern treeline is an important mechanism amplifying cooling due to summer insolation decrease in the northern high latitudes. Analysis of climate–vegetation interactions in North Africa suggests that biospheric feedback might be strong enough to cause abrupt, irreversible changes in the region. At the same time, the role of biological mechanisms in glacial–interglacial CO₂ changes remains far from clear.

The understanding of the role of biospheric feedback in the past has important implications for future climate projections. The challenge in Earth system science is to constrain feedback mechanisms within the models based on available – and newly developing – proxies for biological activities in the past.

See also: Carbonate Stable Isotopes: Overview. Ice Core Methods: Methane Studies. Paleocceanography, Physical and Chemical Proxies: Dissolution of Deep-Sea Carbonates. Paleoclimate: The Younger Dryas Climate Event. Paleoclimate Modeling: The Last Interglacial. Pollen Records, Late Pleistocene: Africa.

References

- Archer DE, Winguth A, Lea D, and Mahowald N (2000) What caused the glacial/interglacial atmospheric pCO₂ cycles? *Reviews of Geophysics* 38: 159–189.
- Bigelow NH, Brubaker LB, Edwards ME, et al. (2003) Climate change and arctic ecosystems: I. Vegetation changes north of 55°N between the last glacial maximum, mid-Holocene, and present. *Journal of Geophysical Research-Atmospheres* 108: 8170.
- Bonan G (2002) *Ecological Climatology*, p. 690. Cambridge: Cambridge University Press.
- Brasseur GP, Artaxo P, Barrie LA, et al. (2003) An integrated view of the causes and impacts of atmospheric changes. In: Brasseur GP, Prinn RG, and Pszenny AAP (eds.) *Atmospheric Chemistry in a Changing World*, pp. 207–230. Heidelberg: Springer-Verlag.
- Brovkin V, Claussen M, Petoukhov V, and Ganopolski A (1998) On the stability of the atmosphere–vegetation system in the Sahara/Sahel region. *Journal of Geophysical Research-Atmospheres* 103: 31613–31624.
- Brovkin V, Sitch S, von Bloh W, et al. (2004) Role of land cover changes for atmospheric CO₂ increase and climate change during the last 150 years. *Global Change Biology* 10: 1–14.
- Calov R, Ganopolski A, Petoukhov V, et al. (2005) Transient simulation of the last glacial inception. Part II: Sensitivity and feedback analysis. *Climate Dynamics* 24: 563–576.
- Claquin T, Roelandt C, Kohfeld KE, et al. (2003) Radiative forcing of climate by ice-age atmospheric dust. *Climate Dynamics* 20: 193–202.
- Claussen M (1994) On coupling global biome models with climate models. *Climate Research* 4: 203–221.
- Claussen M, Cox PM, Zeng X, et al. (2004) The global climate. In: Kabat P, Claussen M, and Dirmeyer PA, et al. (eds.) *Vegetation, Water, Humans and the Climate: A New Perspective on an Interactive System Global Change: The IGBP Series*, pp. 33–57. Berlin: Springer.
- Claussen M and Gayler V (1997) The greening of Sahara during the mid-Holocene: Results of an interactive atmosphere–biome model. *Global Ecology and Biogeography Letters* 6: 369–377.
- Crucifix M and Hewitt CD (2005) Impact of vegetation changes on the dynamics of the atmosphere at the Last Glacial Maximum. *Climate Dynamics* 25: 447–459.
- deMenocal P, Ortiz J, Guilderson T, et al. (2000) Abrupt onset and termination of the African Humid Period: Rapid climate responses to gradual insolation forcing. *Quaternary Science Reviews* 19: 347–361.
- Dickinson RE (1992) Changes in land-use. In: Trenberth KE (ed.) *Climate System Modeling*, pp. 689–700. Cambridge: Cambridge University Press.
- Ganopolski A, Kubatzki C, Claussen M, Brovkin V, and Petoukhov V (1998) The influence of vegetation–atmosphere–ocean interaction on climate during the mid-Holocene. *Science* 280: 1916–1919.
- Harrison S and Prentice IC (2003) Climate and CO₂ controls on global vegetation distribution at the last glacial maximum: Analysis based on paleovegetation data, biome modelling and paleoclimate simulations. *Global Change Biology* 9: 983–1004.
- House JI and Brovkin V (2005) Climate and air quality. In: Hassan R, Scholes R, and Ash N (eds.) *Ecosystems and Human Well-Being: Current State and Trends*, vol. 1, pp. 355–390. Washington, DC: Island Press.
- Jahn A, Claussen M, Ganopolski A, and Brovkin V (2005) Quantifying the effect of vegetation dynamics on the climate of the last glacial maximum. *Climate of the Past* 1: 1–7.
- Kleypas JA (1997) Modeled estimates of global reef habitat and carbonate production since the last glacial maximum. *Paleoceanography* 12: 533–545.
- Kohfeld KE, Le Quéré K, Harrison SP, and Anderson RF (2005) Role of marine biology in glacial–interglacial CO₂ cycles. *Science* 308: 74–78.
- Lucht W, Schaaf CB, and Strahler AH (2000) An algorithm for the retrieval of albedo from space using semiempirical BRDF models. *IEEE Transactions on Geoscience and Remote Sensing* 38: 977–998.
- Munhoven G (2002) Glacial–interglacial changes of continental weathering: Estimates of the related CO₂ and HCO₃[−] flux variations and their uncertainties. *Global and Planetary Change* 33: 155–176.
- Pitman AJ, Dolman H, and Kruijt B (2004) The climate near the ground. In: Kabat P, Claussen M, and Dirmeyer PA, et al. (eds.) *Vegetation, Water, Humans and the Climate: A New Perspective on an Interactive System Global Change: The IGBP Series*, pp. 9–19. Berlin: Springer.
- Prentice IC and Jolly D (2000) BIOME 6000 Participants: Mid-Holocene and glacial-maximum vegetation geography of the northern continents and Africa. *Journal of Biogeography* 27: 507–519.
- Prentice IC and Raynaud D (2001) Paleobiogeochemistry. In: Schulze ED, Heimann M, and Harrison SP, et al. (eds.) *Global Biogeochemical Cycles in the Climate System*, pp. 87–93. San Diego, CA: Academic Press.
- Renssen H, Brovkin V, Fichefet T, and Goosse H (2003) Holocene climate instability during the termination of the African Humid Period. *Geophysical Research Letters* 30: 1184.
- Schaaf CB, Gao F, Strahler AH, et al. (2002) First operational BRDF, albedo nadir reflectance products from MODIS. *Remote Sensing of Environment* 83: 135–148.
- Valdes PJ, Beerling DJ, and Johnson CE (2005) The ice-age methane budget. *Geophysical Research Letters* 32: L02704.
- Vernadsky VI (1926) *Biosfera*. Leningrad, Russia: Nauka p. 146. Citation in accordance with English translation: Vernadsky VI (1998) *The Biosphere*, p. 192. New York: Copernicus Books.
- Woodward FI (1987) *Climate and Plant Distribution*, p. 174. New York: Cambridge University Press.

Effects of Atmospheric Dust

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Micrometer-sized dust particles that are lifted from soil surfaces in arid and semiarid regions by strong surface winds contribute a large fraction to the global atmospheric aerosol mass load and aerosol optical thickness (Intergovernmental Panel on Climate Change (IPCC), 2001). They influence the Earth's radiation balance, biogeochemical processes, and atmospheric chemistry in several ways, which are only partly understood and remain largely unquantified (Figure 1). Atmospheric dust reflects and to some extent absorbs incoming solar radiation, and it also absorbs and emits outgoing thermal radiation due to its relatively large particle sizes. Thus, it directly influences top-of-the-atmosphere and surface radiation fluxes as well as atmospheric heating rates and stability. Such changes in heating rates also may influence cloud formation. In addition, dust provides cloud condensation nuclei (CCN) that are particularly effective for ice nucleation and thus potentially influence cloud optical properties, precipitation formation, and cloud lifetimes (indirect aerosol effects). Another potentially important effect of dust is its influence on oceanic export production of marine microorganisms by supplying nutrients into the ocean surface, overcoming limitations of essential nutrients such as iron, or enhancing nitrogen assimilation. Dust fertilizes some terrestrial ecosystems by depositing nutrients into soils, and it influences atmospheric chemistry due to heterogeneous reactions on the particle surfaces. Emission, atmospheric concentration, and deposition of dust are strongly dependent on atmospheric parameters and surface properties, and they are therefore very variable on a wide range of timescales; for example, dust production and dispersal was considerably enhanced in glacial times compared to modern conditions. The strongest emphasis in investigations of the effects of dust aerosol has been on quantification of the direct radiative effect of dust under present-day conditions, whereas only a few studies have examined dust distributions under different climates such as the Last Glacial Maximum (LGM) or with future climate scenarios. Because quantitative data of the atmospheric dust content cannot be obtained from observations alone, such studies must rely on results of large-scale models of dust emission, transport, and deposition.

Atmospheric Dust

Soil dust aerosol particles are produced in arid and semiarid continental regions (Figure 2), including the Sahara and the Sahelian region, the Arabian Peninsula, the Gobi and Taklamakan deserts in Asia, and the Australian and South American deserts (Prospero et al., 2002). Most of the dust contained within the global atmosphere is emitted from the North African (50–70%) and Asian deserts (10–25%). Wind tunnel experiments and theoretical considerations show a dependence of dust emissions to the third or fourth power of the surface

wind friction velocity. Dust production is initiated when the surface wind friction velocity in the source region exceeds a threshold value that depends on surface properties such as the size and distribution of roughness elements, the soil texture, and soil moisture (Marticorena and Bergametti, 1995).

Fine soil particles with diameters smaller than 10 μm , which can be transported by winds over large distances, are released from the soil matrix when larger wind-blown sand grains strike on the surface, releasing the smaller particles through the process of saltation. Dust emissions are particularly strong in areas that contain fine and loose sediments. A good agreement between maximal occurrence of dust over land surfaces observed by satellites and the location of enclosed topographic depressions has been observed (Prospero et al., 2002). In such areas, fine sediment may have accumulated during the past. Either lake sediments have formed in such depressions, in case they were filled with water during wetter climate periods, or these areas may have been end points for riverine transport of fine particles. Dust emission strength in a potential source region depends on the aerodynamic surface roughness length because structural elements such as rocks or vegetation increase the transport of wind energy to the surface but also partly absorb this energy, increasing the threshold wind friction velocity required for dust emission (Marticorena and Bergametti, 1995).

The presence of crusts can reduce dust emission from a soil surface. Also, the loss of fine soil material by wind deflation can lead to a decrease of dust production from a specific area over time. On the other hand, surface disturbances as a consequence of cultivation of soils in dry regions may lead to enhanced dust emissions (e.g., as observed during the 'dust bowl' events in the United States in the 1930s and 1950s). The contribution of this enhanced dust emission to the global dust load may be considerable in semiarid regions, but the magnitude is uncertain on the global scale. Although IPCC (2001) estimated a contribution of this enhanced dust of up to 50% to the global dust load, Tegen et al. (2004) concluded that this contribution is much smaller, probably <10% globally.

Dust injected into high atmospheric levels of up to several kilometers can be transported over horizontal distances of thousands of kilometers by strong wind systems. Remote sensing instruments can clearly distinguish such dust plumes over dark ocean surfaces. Most prominently, dust emitted in the globally strongest dust source, the Sahara desert, can be carried across the North Atlantic to North and Central America (Prospero and Lamb, 2003) and is even found as far downwind as the Amazon basin. Also, large amounts of Asian dust are carried over the North Pacific toward the mid-Pacific islands and North America.

Dust particles are removed from the atmosphere either by wet deposition (i.e., scavenging through precipitation in the water or ice phase) or by dry deposition, which is the gravitational settling in addition to turbulent mixing toward the surface. Close to desert dust source regions, the gravitational

sedimentation is responsible for the major part of dust deposition, whereas wet deposition dominates the removal of far-traveled dust over remote ocean areas (Jickells et al., 2005). The efficiency of the dust scavenging from the atmosphere may increase with distance from the source regions, when soluble components coating the dust surfaces make the particles increasingly hygroscopic. Whether dust is removed from the atmosphere by dry or wet processes affects the solubility of any trace elements transported with the dust. The atmospheric lifetime of dust depends on the deposition velocities of the different particle sizes. Atmospheric lifetimes of dust depend on the particle size and range from a few hours for particles larger than $10\ \mu\text{m}$, which are quickly removed by gravitational settling, to 10–15 days for submicron particles that are mostly removed by wet deposition (Seinfeld and Pandis, 1997).

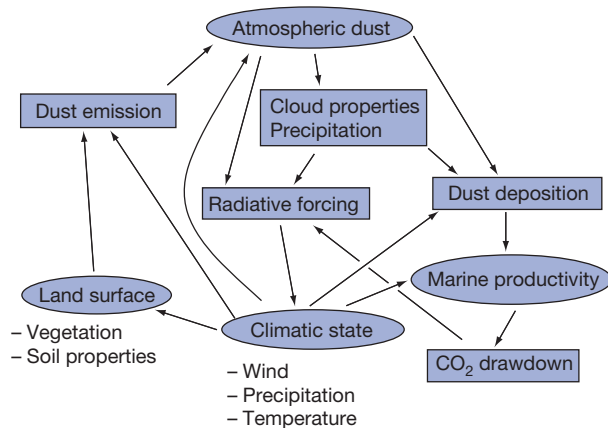


Figure 1 Overview of linkages between atmospheric dust and climate indicating direct and indirect aerosol effects as well as the potentially important role of oceanic dust deposition for marine microorganisms.

Dust Measurements

Although *in situ* observations of airborne and deposited dust particles help us to understand atmospheric dust properties, they cannot describe dust distributions globally because dust is highly variable in space and time and the observations can only cover limited regions. *In situ* observations of dust include multiyear concentration measurements from approximately 30 stations of the Miami oceanic network (Prospero and Lamb, 2003), optical thickness measurements by sun photometers (e.g., the AERONET network), and aerosol lidar observations (Ansmann et al., 2005). In contrast, remote sensing using satellite instruments covers large areas of the globe, but it is nearly impossible to accurately quantify dust properties above land surfaces from such retrievals. Over dark ocean surfaces, retrievals of dust optical thickness and angstrom parameters are possible with some confidence (Kaufman et al., 2001), but such measurements only document dust that is far removed from source areas. Multiyear retrievals of the daily distribution of absorbing aerosols with the Total Ozone Mapping Spectrometer instrument (Herman et al., 1997) have provided useful qualitative information of dust plume occurrences over both ocean and land surfaces (Figure 3).

Measurements of deposited dust provide the only method to obtain information on the atmospheric dust content during past climate periods. Such deposition records include terrestrial loess deposits, marine sediments, and dust concentration in ice cores (Kohfeld and Harrison, 2001). Notable examples of loess deposits that were formed during glacial periods are located in North America, the Eurasian Continent, and the South American Pampa region. Deposition fluxes in loess deposits have been as high as $20000\ \text{g m}^{-2}\ \text{year}^{-1}$ during glacial periods. Dust particles deposited in the loess regions consists of coarser particles ($10\text{--}63\ \mu\text{m}$) that have not traveled far from the source

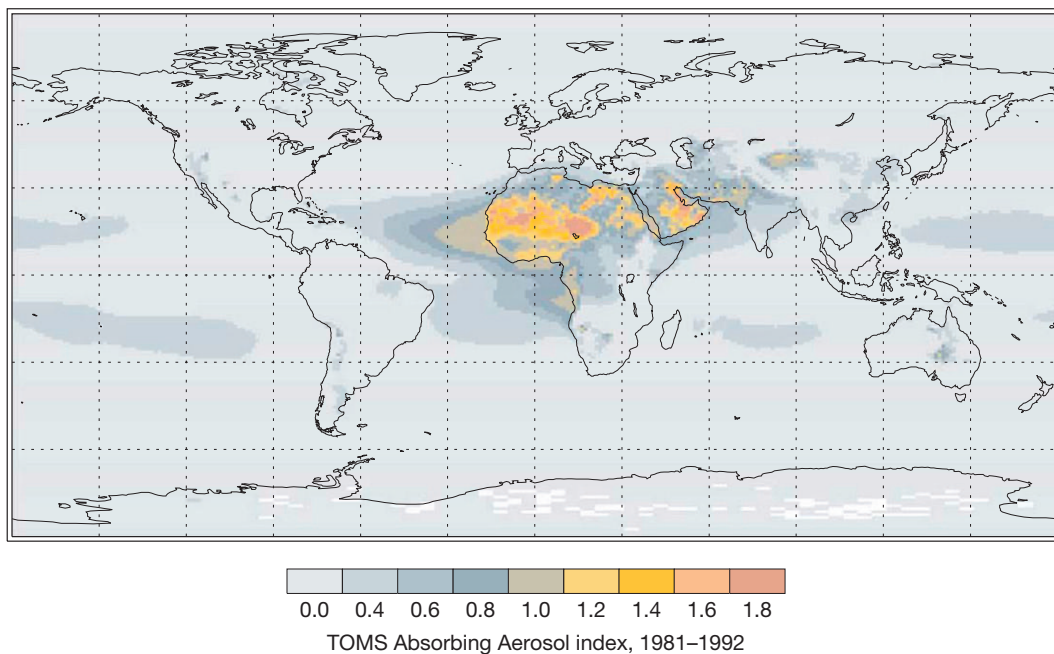


Figure 2 Eleven-year annual average of the Total Ozone Mapping Spectrometer (TOMS) Absorbing Aerosol Index.

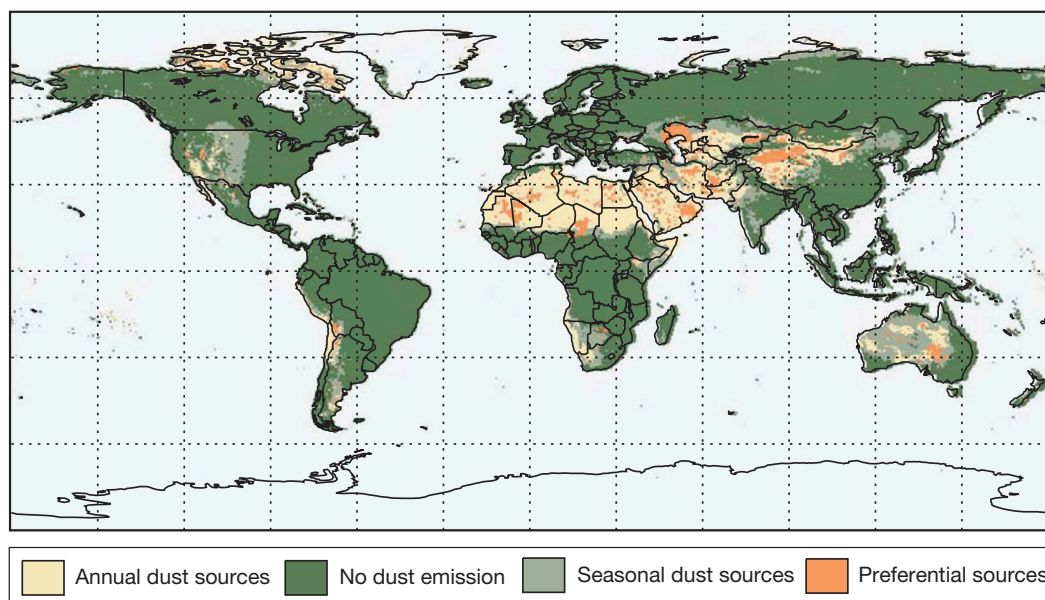


Figure 3 Distribution of global dust source areas, indicating unvegetated desert areas, densely vegetated areas that cannot act as dust sources (dark green), and areas where remote sensing data indicate bare soils for part of the year (light green) and can thus potentially act as seasonal source areas for soil dust emission. Also shown are modeled preferential ‘hot spot’ dust emission areas (brown). This distribution of source areas was used, for example, in the dust emission model described in [Werner et al. \(2003\)](#).

regions. In contrast, marine sediment records are good indicators for intermediate and long-distance transport. Mean grain sizes in marine sediments are typically $<10\ \mu\text{m}$, although larger grains are found in sediments close to continental source areas. Accumulation rates in marine sediments are on the order of $1\text{--}10\ \text{g m}^{-2}\ \text{year}^{-1}$, with maxima as high as $100\ \text{g m}^{-2}\ \text{year}^{-1}$. Marine sediments suggest an approximately twofold increase in dust deposition fluxes in the glacial compared to interglacial periods in the tropics and mid-latitudes ([Kohfeld and Harrison, 2001](#)). Ice core concentration data are records of long-distance transport of dust with particle sizes smaller than $5\ \mu\text{m}$. Records of increases in the LGM indicate that dust concentrations are highest in the Antarctic ice cores that cover several glacial cycles, with an approximately tenfold increase in the Vostok and a 40-fold increase in the European Project for Ice Coring in Antarctica (EPICA) core compared to modern concentrations ([Delmonte et al., 2002](#)). Although the changes in dust fluxes at these locations are very high, the actual fluxes are only on the order of $0.0001\text{--}0.01\ \text{g m}^{-2}\ \text{year}^{-1}$. The isotopic and mineralogical properties of ice core dust allow the determination of source areas and transport. Current studies indicate that glacial dust arriving at Antarctica originates mostly in southern South America, with minor sources in northern New Zealand, or the Antarctic dry valleys; dust arriving in Greenland originates in the Asian deserts ([Basile et al., 1997](#); [Biscaye et al., 1997](#)). Ice core data also show the timing of increase and decrease of dust fluxes compared to changes in other environmental parameters, such as temperature, CO_2 , and CH_4 ([Petit et al., 1990](#)). Dust concentrations increased after the onset of glacial cooling and lowering of atmospheric CO_2 . This timing may lead to explanations of the cause of increased dustiness as well as feedback of dust on climate parameters such as temperature, precipitation, and the radiation balance.

Dust Variability

Soil dust aerosol not only shows distinct spatial patterns but also is temporally very variable at different timescales. Soil dust aerosol distributions respond to changes in global climate conditions, such as the overall intensity of winds that entrain dust into the air, and changes of the hydrological cycle that can affect the extent of source areas as well as the length of time that dust remains in the atmosphere. Within individual source areas, dust emissions mostly follow a seasonal cycle with distinct spring and summer maxima, when strong surface winds occur and the soils are unvegetated ([Figure 4](#)). Dust production can also vary from year to year as a consequence of changes in climate conditions. For example, in the Sahara, dust emissions are found to vary interannually in connection with the Sahelian rainfall deficit ([Prospero and Lamb, 2003](#)). On decadal timescales, an increase in Saharan dust emissions has been observed since the late 1960s through the 1980s, and the emissions have remained at a high level since then. Long-term records of dust deposition measured in ice cores, marine sediments, and loess deposits show a clear change in deposition fluxes between interglacial and glacial climate periods, when dust was approximately 2–10 times increased globally compared to interglacial periods. This increased dustiness is likely due to a combination of different factors, such as increased aridity, reduced vegetation, increased availability of fine sediment, and stronger surface winds in glacial climates. Higher wind speeds could produce stronger dust emission fluxes from soil surfaces compared to modern conditions; increased aridity expands those areas with drier soils and reduced vegetation where soil deflation by wind is possible. Atmospheric dust loads could also have been enhanced in

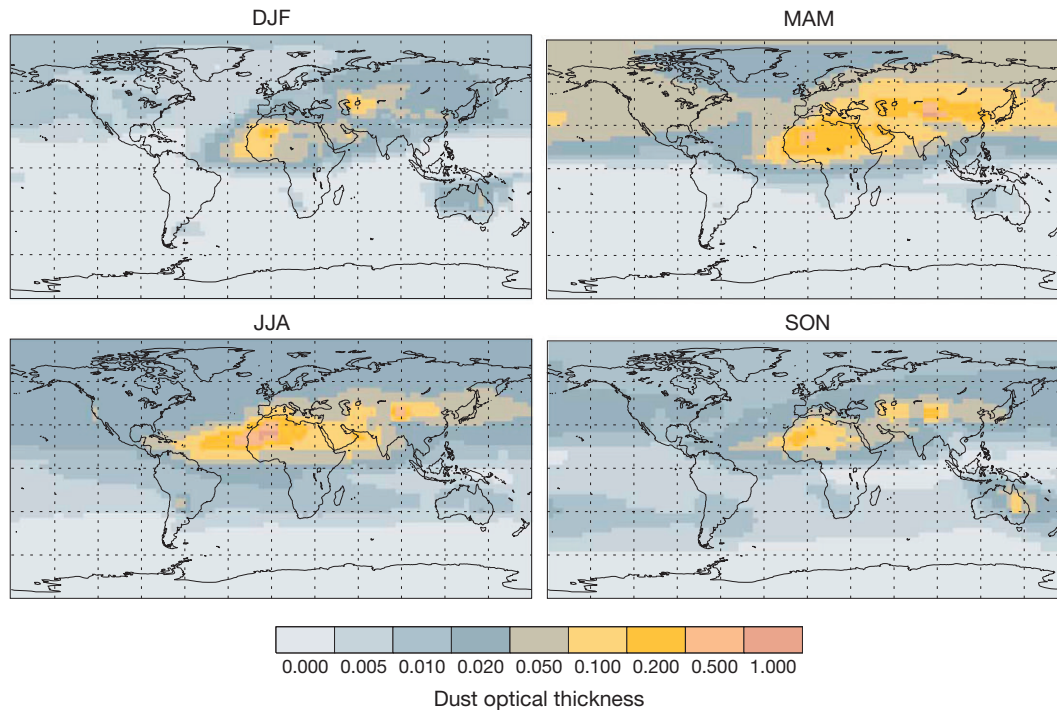


Figure 4 Seasonal distribution of modeled dust aerosol optical thickness based on the results of Tegen et al. (2004).

climate conditions as a consequence of the weakened hydrological cycle with reduced washout of atmospheric dust particles. The best documented period of increased dustiness in the Quaternary is the last glacial period (approximately 17 000–23 000 years ago; Kohfeld and Harrison, 2001).

Models of the Dust Cycle

Increasingly complex regional and global-scale models of the dust cycle have been developed in order to quantify the atmospheric dust distribution with the aim of supporting investigations of dust effects. The atmospheric dust content computed by these models depends on dust emissions, which are determined by surface wind speeds, and surface properties such as roughness length, vegetation cover, soil moisture, and soil texture. Recently developed models explicitly prescribe seasonal vegetation cover (Werner et al., 2003), masking dust sources, as well as the occurrence of topographic depressions in dry, unvegetated areas as indicators of preferential sources containing fine and loose sediment particles that can be easily deflated under strong wind conditions (see Figure 2).

Large-scale data sets of soil properties are difficult to obtain on global scales because the most active dust sources lie in remote areas that cannot easily be accessed. Another major problem in computing dust emissions from wind deflation is insufficient model resolution because peak gusts that are responsible for the strongest dust events remain largely unresolved in large-scale global models. Parameterizations of subgrid scale variability surface winds (Cakmur et al., 2004) provide only moderate improvements in computations of dust emissions.

Table 1 Range of model-based estimates of emission fluxes, atmospheric burden, and atmospheric lifetimes of soil dust aerosols^a

	Dust emission (Mt year ⁻¹)	Atmospheric burden (Mt)	Atmospheric lifetime (days)
Modern	1100–3000	8–36	2.8–7.1
LGM	2400–9000	23–94	3.5–5.5
Ratio LGM/modern	1.4–3.0	1.4–2.8	0.9–1.3

^aGiven are the ranges of minimum and maximum values from model calculations by various authors; summarized in Zender et al. (2004) and Harrison et al. (2001), including also the results from Werner et al. (2003).

Apart from models of present-day dust distributions, the LGM has been a major focus for dust modeling (Table 1). During glacial periods, eolian deposition rates were approximately 2–20 times higher than today (Kohfeld and Harrison, 2001) (Figure 5), and changes in boundary conditions such as ice sheet extent and atmospheric trace gas concentrations are reasonably well known. Models that only consider the influences of increased surface wind speeds and the reduced intensity of the hydrological cycle on dust emission and deposition fluxes at the LGM are unable to reproduce the observed strong increase in dust deposition at high latitudes. The increase in dust source areas caused by exposure of continental shelves is likewise unable to explain the increased high-latitude dust signal. Incorporating increases in dust sources from changes in vegetation cover increases the modeled atmospheric dust loads at high latitudes, consistent with the high dust fluxes observed in polar ice cores (Mahowald et al., 1999; Werner et al., 2003). The potentially large effect of dust aerosol supplied by glacial outwash and the increase in the areal extent of

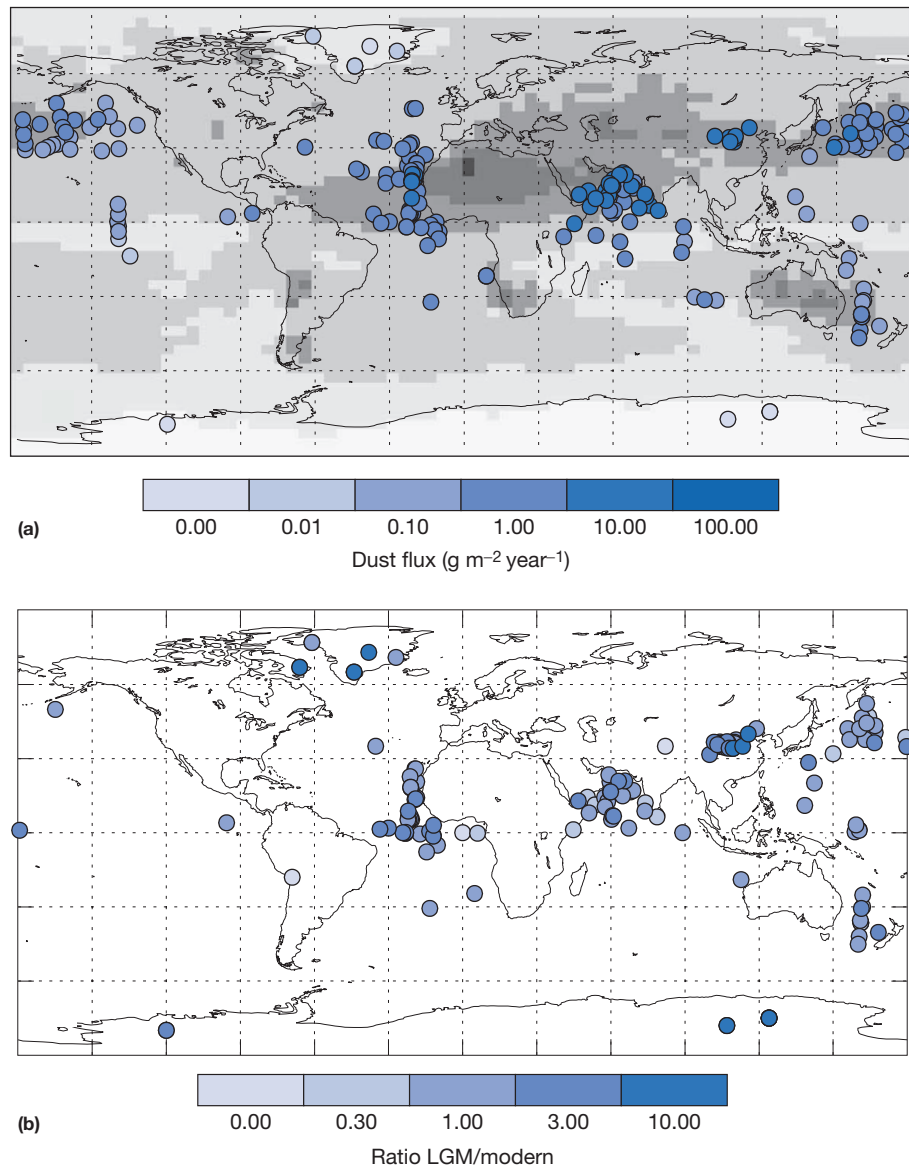


Figure 5 (a) Patterns of modern global dust deposition fluxes comparing a model result (Tegen et al., 2004) with local dust fluxes derived from measurements of marine sediments, sediment traps, terrestrial loess deposits, and ice cores for the late Holocene compiled by Kohfeld and Harrison (2001). (b) To indicate the enhancement of dust deposition fluxes during glacial climate, the ratio of the observed dust fluxes between the Last Glacial Maximum and the late Holocene is shown.

dry lake beds during the LGM have not yet been factored into global dust models.

Global-scale models are not always useful for investigation of regional-scale dust processes, especially in areas close to source regions, since the horizontal resolution of typically $3\text{--}5^\circ$ in global models and the necessarily simplified emission and deposition parameterizations are too crude to resolve small-scale dust processes and enable comparisons to observations. Regional models of dust emission, transport, and deposition exist for key regions, including the Sahara desert and Asia (Sokolik et al., 2001). Such regional-scale models are well suited for simulation of individual dust storm events as well as for comparisons with *in situ* observations made during field experiments. Although the topography, soil conditions, and

small-scale extreme wind events are simulated more realistically in regional compared to global models, uncertainties in the parameterization of key processes and input data on meteorological fields and soil properties remain.

Particle Properties

Dust aerosol particle sizes determine their optical and microphysical properties, as well as atmospheric lifetimes. Close to source regions, the median dust particle diameters are observed to be approximately $60\text{--}100\text{ }\mu\text{m}$; particles in this size range can most effectively be lifted from the ground by surface winds. Because large dust particles settle back to the ground quickly,

dust that has been transported thousands of kilometers away from the source has a median size of approximately $2\ \mu\text{m}$ (Seinfeld and Pandis, 1997).

The mineral composition of dust particles reflects the mineralogy of the rocks at the Earth's surface; some main constituents are quartz, feldspars, carbonates, and clays. Dust particles from different source regions can have specific mineralogical composition and, as a consequence, different chemical composition. Claquin et al. (1999) summarized available information to compile global maps of mineral composition of soils in dry areas by relating soil mineralogy to soil types. They found that quartz, with approximately 70% of the mass, is the most abundant mineral in the silt size fraction of the soil (particles in the size range of $2\text{--}50\ \mu\text{m}$ diameter), and illite is the most abundant clay mineral (contributing up to 50% regionally to the clay fraction). The radiative and chemical properties of dust can change during transport when dust surfaces coat with soluble sulfates and nitrate species and are processed by clouds.

Dust Effects

Direct Effect

Incoming solar radiation is partly reflected and partly absorbed by dust particles in the atmosphere. With mass median diameters of $2\ \mu\text{m}$ or larger, these particles also absorb and emit outgoing terrestrial radiation. The radiative effect of dust particles depends on their optical properties (i.e., their size distribution and complex refractive index), which are either determined by laboratory and field measurements or retrieved by remote sensing (Sokolik et al., 2001). Although direct measurements have a considerable uncertainty range, remote sensing retrievals of dust properties may be influenced by the presence of other aerosol types besides dust or by thin cirrus clouds, or by underestimation of the nonsphericity of dust particles. Dust aerosol radiative properties are strongly influenced by their hematite content, which mostly controls the absorption of incoming solar radiation by the particles. From size and complex refractive indices of the particles, aerosol extinction efficiency (which determines the light attenuation per size fraction), the single scattering albedo (which is the ratio of scattering and total extinction of solar radiation), and the asymmetry parameter (which quantifies the ratio of forward scattered light) can be computed using Mie theory, which is a theory of the scattering of electromagnetic radiation by spherical particles. These parameters are then used to compute the fractions of absorbed and transmitted radiation passing through an aerosol layer. Radiative forcing by dust, defined here as the attenuation of radiation caused by the presence of atmospheric dust, rather than the radiative impact of dust enhancement from human activities, is generally negative over dark ocean surfaces. However, over bright desert surfaces, the net radiative effect of dust depends on their optical properties, and it can be positive for strongly absorbing particles (Figure 6). At the top of the atmosphere, the negative and radiative forcing by dust can partly cancel in the global mean (IPCC, 2001). Measurements of dust optical properties from remote sensing (Kaufman et al., 2001) have led to the conclusion that the net radiative forcing due to dust is negative in the global mean, indicating that enhanced dust aerosol loads would lead to regional cooling.

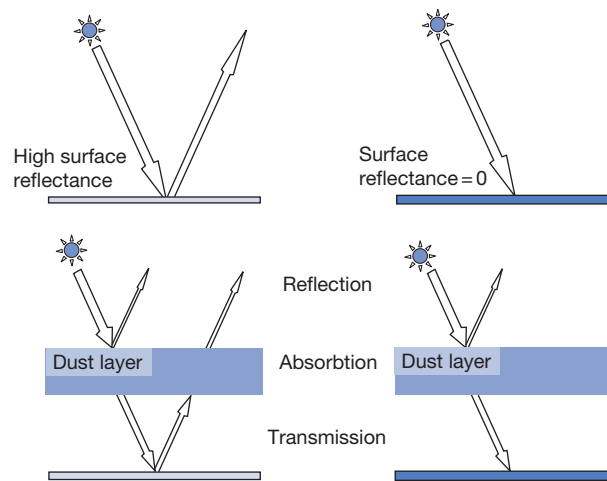


Figure 6 Schematic representation of reflection and absorption of incoming solar radiation in the absence and presence of a layer of dust aerosol, above a bright, highly reflective surface, and a dark absorbing surface. The dust layer partly reflects and absorbs incoming solar radiation; from space the dust appears bright over a dark surface (e.g., the ocean), the dust absorption may be sufficient to appear dark above a bright surface (bright desert or ice cover). Thus, depending on dust optical properties, it may lead to an enhancement or a reduction of energy input into the globe.

The radiative effect of higher dust loads during the LGM has been investigated by Claquin et al. (2003). The authors computed the difference in the top-of-the-atmosphere radiation between the LGM and present-day dust loads from a model simulating past dust distribution, which were obtained from results of a simulation taking into account changes in climate parameters and vegetation density for glacial conditions. The estimated change in top-of-the-atmosphere radiative forcing by the enhanced dust load (Figure 7) ranged from -0.3 to $-0.9\ \text{W m}^{-2}$ at northern high latitudes. This effect is small compared to the radiative effects due to the expansion of ice coverage at the LGM, which results in a radiative forcing of approximately $-19\ \text{W m}^{-2}$ at latitudes between 45° and 90° N. At low latitudes between 15° S and 15° N, Claquin et al. (2003) estimate a decrease in incoming solar radiation fluxes of -1 to $-1.4\ \text{W m}^{-2}$ caused by the approximately doubled dust load at the LGM, which is of comparable magnitude to the decrease in greenhouse forcing by $-1.8\ \text{W m}^{-2}$ due to the reduced CO_2 concentration at the LGM. These results indicate that the enhanced atmospheric dust may have contributed to glacial cooling at low latitudes, whereas it played only a small role in the radiation balance at high latitudes.

Semidirect and Indirect Effects

Radiative forcing by dust not only affects the radiation balance of the earth directly but also influences the energy distribution within the atmosphere. At the surface below a dust cloud, incoming solar radiation is reduced, which mainly causes a decrease in latent heating and therefore a weakening of the hydrological cycle (Miller et al., 2004). Together with atmospheric heating in the presence of a dust cloud, this may lead to a reduction in cloud cover, which is called the semidirect

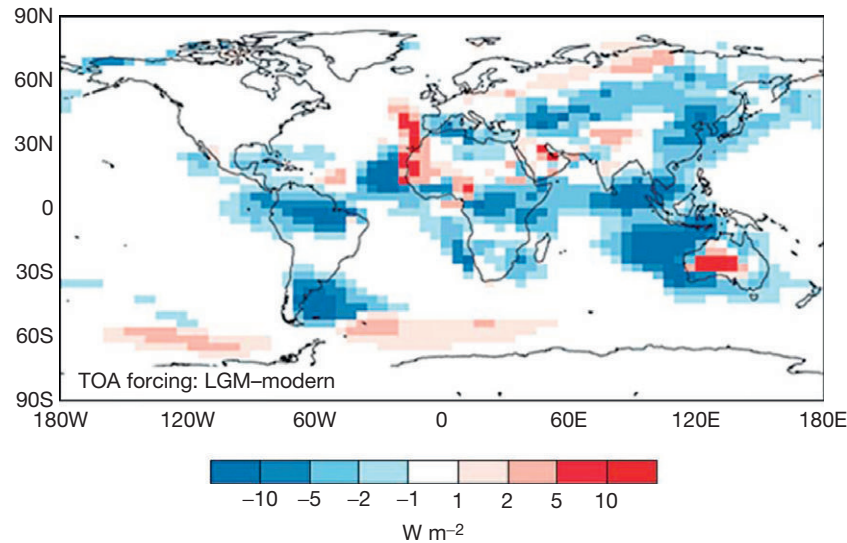


Figure 7 Net changes in solar radiation at the top of the atmosphere by enhanced atmospheric dust loads during the Last Glacial Maximum (LGM). Shown is the difference in radiation fluxes between LGM and modern conditions, taking into account enhanced dust loads at the LGM. Dust optical properties were based on the assumption of an internal mixture of quartz minerals with 2% hematite. Modified from Claquin T, Roelandt C, Kohfeld KE, et al. (2003) Radiative forcing of climate by ice-age atmospheric dust. *Climate Dynamics* 20: 193–202, Figure 3b. Copyright Springer-Verlag.

aerosol effect. Hansen et al. (1997) showed with a simplified global atmospheric model that slightly absorbing aerosols, which have a negative direct radiative effect, may actually cause warming as a consequence of the semidirect effect.

Aerosol particles influence the formation and lifetimes of clouds via the so-called indirect aerosol effect, although the role of soil dust particles in these processes remains unclear. Wurzler et al. (2000) found that the coating of dust by soluble particles enhances their ability to form CCN. In addition, dust particles are effective nuclei for ice clouds; Ansmann et al. (2005) found ice formation in an altocumulus cloud in the presence of dust at 10–20 K higher temperatures than in the absence of insoluble particles, indicating the potentially important role of dust for ice nucleation. This process would not only affect the radiative properties of clouds but also enhance the formation of precipitation. However, the magnitudes of these effects are still unknown.

Effect of Dust on Ocean Biogeochemistry

Deposited soil dust particles are hypothesized to add essential nutrients to terrestrial and marine ecosystems. In particular, the increase in productivity of oceanic microorganisms by addition of iron has received much attention in recent years (Jickells et al., 2005). Microorganisms use iron in a variety of enzyme systems, including those for photosynthesis and nitrogen fixation. Soils contain an average of 3.5% iron. The iron content of dust particles is variable, but this uncertainty is small compared with other uncertainties in the iron cycle. Iron solubility in soil dust particles at source areas is on the order of 1%. Higher solubilities of aerosol iron of locally up to 40% have been reported in atmospheric particles due to photochemical processes (photoreduction of FeIII to FeII) and aerosol cloud processing. In nutrient-rich regions of the ocean, the addition of iron can enhance primary production,

species composition, and CO₂ uptake, whereas in oligotrophic regions such as the subtropics, the nutrient supply from dust aerosol can enhance nitrogen fixation. Several field studies show that plankton productivity and CO₂ drawdown are enhanced in nutrient-rich regions when iron is added, but it is unknown if dust is important for oceanic CO₂ uptake at the global scale. Different biogeochemical models have derived different results; one consequence of dust seems to be a change in planktonic species composition (Bopp et al., 2003).

Glacial–interglacial changes in ocean export production are suspected to have contributed to CO₂ drawdown, assuming that stronger dust deposition fluxes (see Figure 5) enhanced primary productivity. The effects of enhanced dust deposition during glacial periods have been investigated by Kohfeld et al. (2005), who analyzed sediment records of marine productivity at the peak and the middle of the last glacial cycle and found that iron fertilization and associated mechanisms can be responsible for up to half the observed CO₂ drawdown.

Other Effects

Apart from the direct and indirect effects of soil dust aerosol on the climate, dust also affects atmospheric chemistry by providing reactive surface areas for heterogeneous reactions (Dentener et al., 1996), influencing tropospheric ozone concentrations and the atmospheric sulfate and nitrate cycles. Sulfate reactions occur mainly with calcite in the dust aerosol, which is converted to gypsum. Bonasoni et al. (2004) observed that in the Mediterranean region the lowest surface ozone concentration occurred in the presence of coarse dust aerosol, indicating an important role of dust for ozone chemistry.

In addition to these processes, dust storms can adversely affect human quality of life. Dust storms strongly reduce visibility in the atmosphere, causing traffic problems in regions with frequent dust events. Strong exposure to dust has also been related to health problems such as silicosis.

Conclusion

Although the cycle of dust emission, atmospheric transport, and deposition has been well researched and a wide array of data exist on the properties of dust aerosols, their direct and indirect effects on the climate and atmospheric chemistry remain poorly quantified. The strong spatial and temporal variability of the atmospheric content and properties of the dust particles continue to cause challenging problems in quantifying atmospheric dust properties at global scales.

See also: [Paleoclimate Modeling: Last Glacial Maximum GCMs; Quaternary Environments.](#)

References

- Ansmann A, Mattis I, Müller D, et al. (2005) Ice formation in Saharan dust over central Europe observed with temperature/humidity/aerosol Raman lidar. *Journal of Geophysical Research* 110: No. D18S12.
- Basile I, Grousset FE, Revel M, et al. (1997) Patagonian origin of glacial dust deposited in East Antarctica (Vostok and Dome C) during glacial stages 2, 4 and 6. *Earth and Planetary Science Letters* 146: 573–589.
- Biscaye PE, Grousset FE, Revel M, et al. (1997) Asian provenance of glacial dust (stage 2) in the Greenland Ice Sheet Project 2 Ice Core, Summit, Greenland. *Journal of Geophysical Research* 102: 26765–26781.
- Bonasoni P, Cristofanelli P, Calzolari F, et al. (2004) Aerosol–ozone correlations during dust transport episodes. *Atmospheric Chemistry and Physics* 4: 1201–1215.
- Bopp L, Kohfeld KE, LeQuéré C, and Aumont O (2003) Dust impact on marine biota and atmospheric CO₂ during glacial periods. *Paleoceanography* 18: Art No. 1046.
- Cakmur RV, Miller RL, and Torres O (2004) Incorporating the effect of small-scale circulations upon dust emission in an atmospheric general circulation model. *Journal of Geophysical Research* 109: No. D07201.
- Claquin T, Roelandt C, Kohfeld KE, et al. (2003) Radiative forcing of climate by ice-age atmospheric dust. *Climate Dynamics* 20: 193–202.
- Claquin T, Schulz M, and Balkanski Y (1999) Modelling the mineralogy of atmospheric dust sources. *Journal of Geophysical Research* 104: 22243–22256.
- Delmonte B, Petit JR, and Maggi V (2002) Glacial to Holocene implications of the new 27,000-year dust record from the EPICA Dome C (East Antarctica) ice core. *Climate Dynamics* 18: 647–660.
- Dentener FJ, Carmichael GR, Zhang Y, Lelieveld J, and Crutzen PJ (1996) Role of mineral aerosol as reactive surface in the global troposphere. *Journal of Geophysical Research* 101: 22869–22889.
- Hansen JE, Sato M, and Ruedy R (1997) Radiative forcing and climate response. *Journal of Geophysical Research* 102: 6831–6864.
- Harrison SP, Kohfeld KE, Roelandt C, and Claquin T (2001) The role of dust in climate changes today, at the last glacial maximum and in the future. *Earth-Science Reviews* 54: 43–80.
- Herman JR, Bhartia PK, Torres O, et al. (1997) Global distribution of UV-absorbing aerosols from Nimbus 7/TOMS data. *Journal of Geophysical Research* 102: 16911–16922.
- Intergovernmental Panel on Climate Change (2001) Houghton JT, Ding Y, and Griggs DJ, et al. (eds.) *Climate Change 2001: The Scientific Basis*. New York: Cambridge University Press.
- Jickells TD, An ZS, Andersen KK, et al. (2005) Global iron connections between desert dust, ocean biogeochemistry, and climate. *Science* 308: 67–71.
- Kaufman YJ, Tanre D, et al. (2001) Absorption of sunlight by dust as inferred from satellite and ground-based remote sensing. *Geophysical Research Letters* 28: 1479–1483.
- Kohfeld KE and Harrison SP (2001) DIRTMAP: The geological record of dust. *Earth-Science Reviews* 54: 81–114.
- Kohfeld KE, LeQuere C, Harrison SP, and Anderson RF (2005) Role of marine biology in glacial–interglacial CO₂ cycles. *Science* 308: 74–78.
- Mahowald NM, Kohfeld KE, Hansson M, et al. (1999) Dust sources and deposition during the Last Glacial Maximum and current climate: A comparison of model results with paleodata from ice cores and marine sediments. *Journal of Geophysical Research* 104: 15895–16436.
- Marticorena B and Bergametti G (1995) Modeling the atmospheric dust cycle: 1. Design of a soil-derived dust emission scheme. *Journal of Geophysical Research* 100: 16415–16430.
- Miller RL, Tegen I, and Perlwitz J (2004) Surface radiative forcing by soil dust aerosols and the hydrologic cycle. *Journal of Geophysical Research* 109: No. D04203.
- Petit JR, Mounier L, Jouzel J, et al. (1990) Palaeoclimatological and chronological implications of the Vostok core dust record. *Nature* 343: 56–58.
- Prospero JM, Ginoux P, Torres O, Nicholson SE, and Gill TE (2002) Environmental characterization of global sources of atmospheric soil dust identified with the NIMBUS 7 TOMS absorbing aerosol product. *Reviews of Geophysics* 40: No. 1002.
- Prospero JM and Lamb PJ (2003) African droughts and dust transport to the Caribbean: Climate change implications. *Science* 302: 1024–1027.
- Seinfeld JH and Pandis SN (1997) *Atmospheric Chemistry and Physics: From Air Pollution to Climate Change*. New York: Wiley.
- Sokolik IN, Winker DM, Bergametti G, et al. (2001) Introduction to special section: Outstanding problems in quantifying the radiative impacts of mineral dust. *Journal of Geophysical Research* 106: 18015–18027.
- Tegen I, Werner M, Harrison SP, and Kohfeld KE (2004) Relative importance of climate and land use in determining present and future global soil dust emission. *Geophysical Research Letters* 31: No. L05105.
- Werner M, Tegen I, Harrison SP, et al. (2003) Seasonal and interannual variability of the mineral dust cycle under present and glacial climate conditions. *Journal of Geophysical Research* 108: Art. No. 4744.
- Wurzler S, Reisin TG, and Levin Z (2000) Modification of mineral dust particles by cloud processing and subsequent effects on drop size distributions. *Journal of Geophysical Research* 105: 4501–4512.
- Zender C, Miller RL, and Tegen I (2004) Quantifying mineral dust mass budgets: Systematic terminology, constraints, & current estimates. *Eos* 85: 509–512.

Thermohaline Circulation

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"It appears to be extremely difficult, if not quite impossible, to account for this degree of cold at the bottom of the sea in the torrid zone, on any other supposition than that of cold currents from the poles; and the utility of these currents in tempering the excessive heats of these climates is too evident to require any illustration."

Sir Benjamin Thompson (1797)

What is the Thermohaline Circulation?

In 1751, the captain of an English slave-trading ship made the first recorded measurement of deep ocean temperatures – he discovered that the water a mile below his ship was very cold, despite the subtropical location. In 1797, another Englishman, Benjamin Thompson, correctly explained this discovery by cold currents from the poles, as part of what later became known as the thermohaline circulation (THC).

As opposed to wind-driven currents and tides (the latter are due to the gravity of the Moon and the Sun), the THC is that part of the ocean circulation which is driven by fluxes of heat and freshwater across the sea surface and subsequent interior mixing of heat and salt; hence the name thermohaline. (Geothermal heat sources at the ocean bottom play a minor role.) The term THC thus refers to a particular driving mechanism; it is a physical, not an observational, concept.

The distinction of thermohaline versus wind-driven circulation originates in a nineteenth-century dispute on whether ocean currents are primarily due to the wind pushing along the water or whether they are 'convection currents' due to heating and cooling, or evaporation and precipitation. In 1908, Johan Sandström performed a series of classic tank experiments at Bornö oceanographic station in Sweden to investigate both possibilities (Sandström, 1908), describing the properties of 'wind-driven circulation' and 'thermal circulation.' To include salinity, the latter was later extended to 'thermohaline circulation,' a term that by the 1920s appeared in the classic oceanography textbook by Defant (1929).

The ocean's density distribution, which determines pressure gradients and thus circulation, is affected by currents and mixing. Thermohaline and wind-driven currents therefore interact in nonlinear ways and cannot be separated by oceanographic measurements. There are thus two distinct physical forcing mechanisms but not two uniquely separable circulations. Changing the wind stress will alter the THC; altering thermohaline forcing will also change the wind-driven currents.

A related, complementary concept is that of meridional overturning circulation (MOC). This refers to the north-south flow as a function of latitude and depth, often integrated in east-west direction across an ocean basin or the globe and graphically depicted as a stream function. The streamlines typically show a large-scale slow overturning motion of the ocean. The MOC can be easily diagnosed from a model, and in principle it can be measured in the ocean.

Although the terms THC and MOC are often inaccurately used as if synonymous, there strictly is no one-to-one relation between the two. The MOC includes clearly wind-driven parts, namely the Ekman cells consisting of the transport in the near-surface Ekman layer and a return flow below it. A direct contribution of wind-driven currents even to the large scale, deeper overturning is being increasingly discussed. On the other hand, the THC is of course not confined to the meridional direction; rather, it is also associated with zonal overturning cells. Hence, care should be taken with the terminology: The term THC should be reserved for a particular forcing mechanism – for example, when discussing the influence of cooling or freshwater forcing on the ocean circulation. The term MOC should be used when describing a meridional flow field, for example, from a model, which most often will show a mix of both wind-driven and thermohaline flow.

The following are key features of the THC:

- Deep water formation: The sinking of water masses, closely associated with convection, which is a vertical mixing process. Deep water formation takes place in a few localized areas (Figure 1): the Greenland-Norwegian Sea, the Labrador Sea, the Mediterranean Sea, the Weddell Sea, and the Ross Sea.
- Spreading of deep waters (e.g., North Atlantic Deep Water (NADW) and Antarctic Bottom Water (AABW)), mainly as deep western boundary currents.
- Upwelling of deep waters: This is not as localized as convection and difficult to observe. It is thought to take place mainly in the Antarctic Circumpolar Current region, possibly aided by the wind (Ekman divergence; Figure 2).
- Near-surface currents: These are required to close the flow. In the Atlantic, the surface currents compensating the outflow of NADW range from the Benguela Current off South Africa via the Gulf Stream and the North Atlantic Current into the Nordic Seas off Scandinavia (Figure 3). It is worth noting that the Gulf Stream is primarily a wind-driven current, forming part of the subtropical gyre circulation. The THC – approximated here by the amount of water needed to compensate for the southward flow of NADW – contributes only approximately 20% to the Gulf Stream flow.

Some Observational Data

As explained previously, the THC is not a measurable quantity but, rather, a conceptual idea. However, even many aspects of the MOC are difficult to measure. The vertical motions of sinking and upwelling are too slow to directly capture with current meters. Surface currents are to a large extent part of wind-driven horizontal circulations, so measurements do not easily yield values for the surface component of the MOC. Deep western boundary currents, tracer data, and inverse calculations combining various data sources and physical constraints give the best information on the MOC (Figure 4).

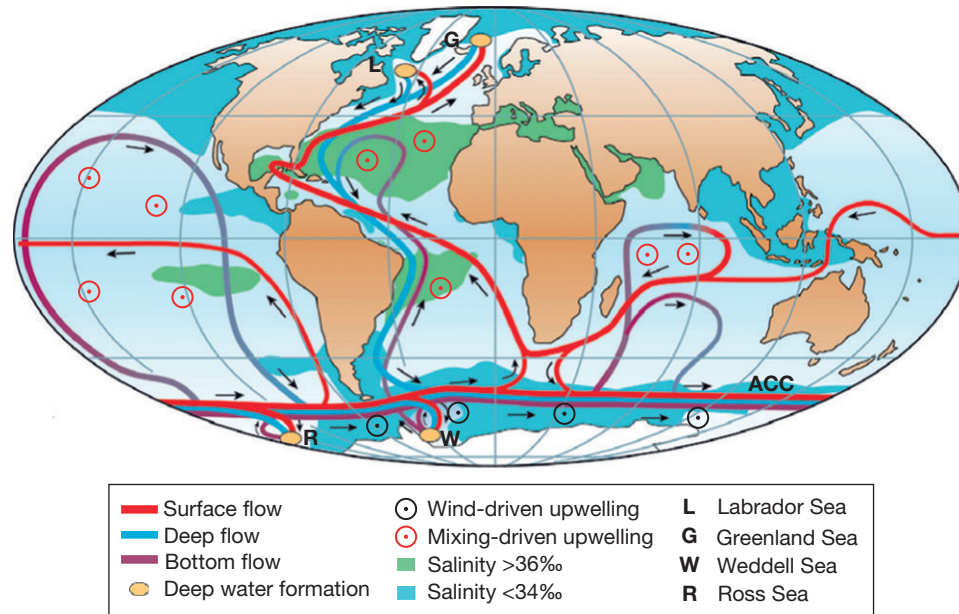


Figure 1 Schematic representation of the global thermohaline circulation (THC). Surface currents are shown in red, deep waters in light blue, and bottom waters in dark blue. The main deep water formation sites are shown in orange. Modified from Broecker WS (1991) The great ocean conveyor. *Oceanography* 4: 79–89; Kuhlbrodt T, et al. (2007) On the driving processes of the oceanic meridional overturning circulation. *Reviews of Geophysics* 45: RG2001.

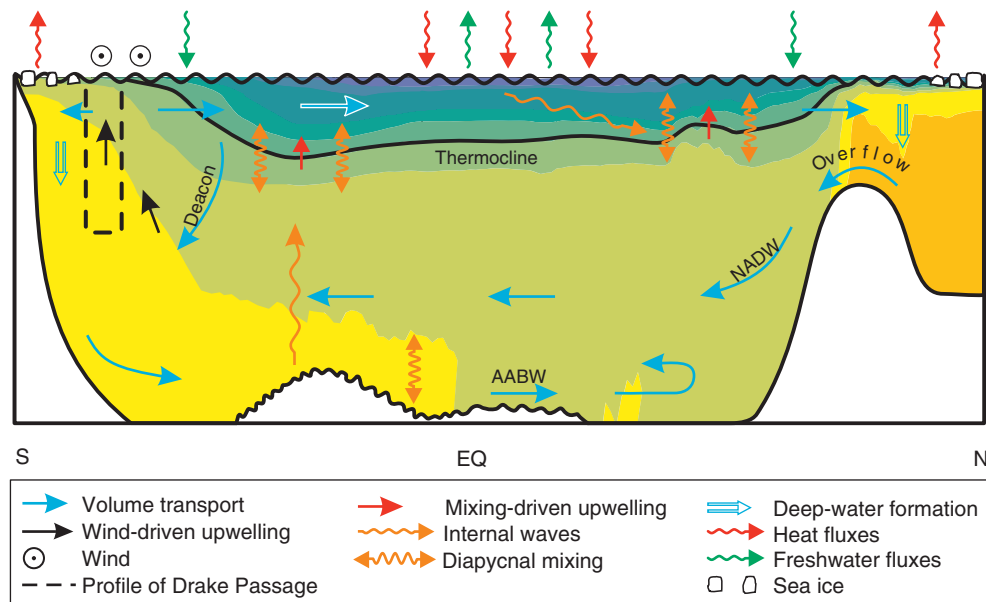


Figure 2 Side view of the circulation in the Atlantic showing various flow components and mechanisms discussed in the text. Color shading shows the observed density stratification, with lightest waters in blue and densest in orange. Reproduced from Kuhlbrodt T, et al. (2007) On the driving processes of the oceanic meridional overturning circulation. *Reviews of Geophysics* 45: RG2001.

The volume transport of the overturning circulation at 24°N in the Atlantic has been estimated from hydrographic section data (Roemmich and Wunsch, 1985) as 17 Sv ($1\text{ Sv} = 10^6\text{ m}^3\text{ s}^{-1}$) and its heat transport as 1.2 PW ($1\text{ PW} = 10^{15}\text{ W}$). Talley et al. (2003) estimated 18 ± 5 Sv of NADW formation, and an inverse model (Ganachaud and Wunsch, 2000) yielded 15 ± 2 Sv NADW overturning in the high latitudes.

Narrow channels provide a good opportunity to measure deep water flows. The overflows from the Nordic Seas have been measured as transporting ~ 3 Sv each between Greenland and Iceland and between Iceland and Scotland (Figure 5), whereas the deep water formation south of these sills, in the Labrador Sea, is estimated as 2–4 Sv. Combined, these numbers do not add up to the total NADW flow estimates given

previously because the volume transport of dense water increases through mixing along the way in a process called entrainment (at the expense of being diluted; i.e., the core density decreases).

Less information is available about the second major deep water formation region of the global oceans, namely the

Southern Ocean around Antarctica (Figures 6 and 7). Tracer data suggest another ~ 15 Sv of deep water forming there, bringing the global total to slightly more than 30 Sv. More recent chlorofluorocarbon data suggest only 5 Sv sinking from the surface around Antarctica, which may be reconciled either by a change over time or, more likely, again by entrainment.

What Drives the THC?

The simplest answer to this question would be high-latitude cooling. In cold regions the highest surface water densities are reached; this causes sinking of water, which in turn drives the circulation. This is a thermally dominated circulation in that it is the coldest waters that sink to fill the deep oceans, and for the moment we will ignore the effects of salinity, returning to them in section 'Nonlinear Behavior of the THC.'

Reality is more complex, however. Cooling a previously warmer ocean at the surface will lead to sinking of the coldest, densest water, which will spread out along the bottom causing upwelling elsewhere, according to what is termed 'filling box dynamics.' However, eventually (after ~ 1000 years at current deep water formation rates) the box would be full – the ocean basins would be filled with dense water and the 'filling box' circulation would come to a halt.

Maintaining a steady-state circulation thus requires more than cold high latitudes. It requires either the source water to get continually denser to keep replacing the water in the box (which contravenes the idea of a steady state) or a mechanism that continually works to make the water in the box less dense so that the source water remains denser than the water already in the box. This allows a steady state, with newly formed dense waters spreading in the ocean tending to increase the density while the other mechanism tends to reduce the

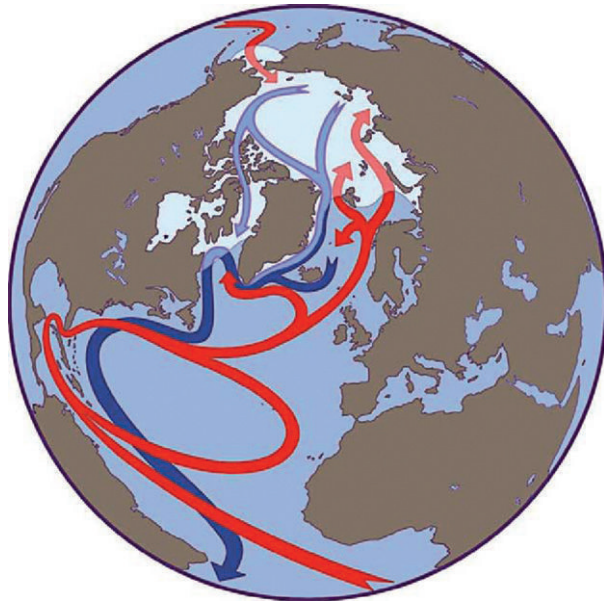


Figure 3 Circulation of the northern Atlantic and Arctic oceans. This simplified figure shows surface currents in red and North Atlantic Deep Water (NADW) in blue. The winter sea ice cover (white) is held back in the Atlantic sector by the warm North Atlantic Current. Figure modified for the Arctic by G. Holloway.

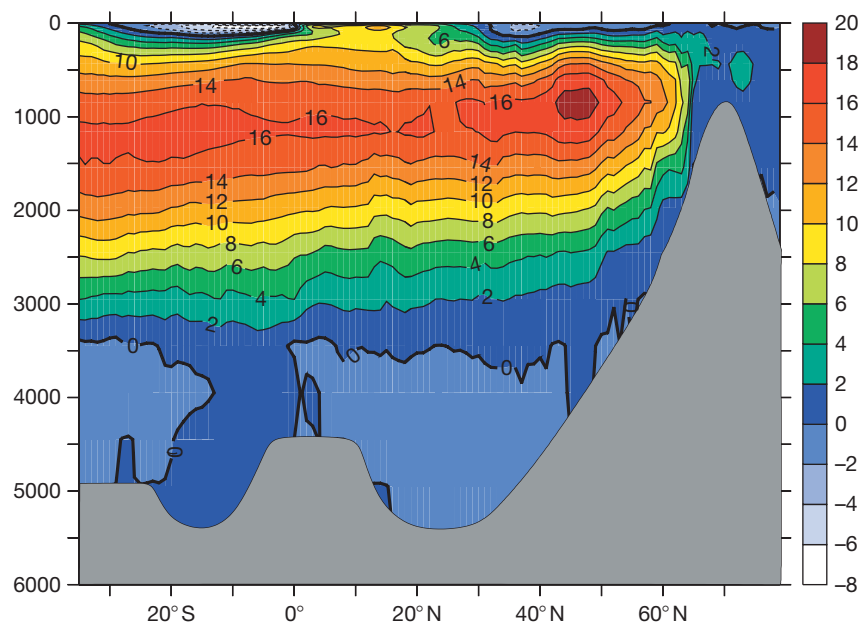


Figure 4 Stream function of meridional overturning in the Atlantic based on a model constrained by observational data. Reproduced from Köhl A, et al. (2007) Interannual to decadal changes in the ECCO global WOCE synthesis. *Journal of Physical Oceanography* 37: 313–337. Visible are the Ekman cells in the upper 500 m, the NADW cell down to 3500 m, and the AABW cell near the bottom.

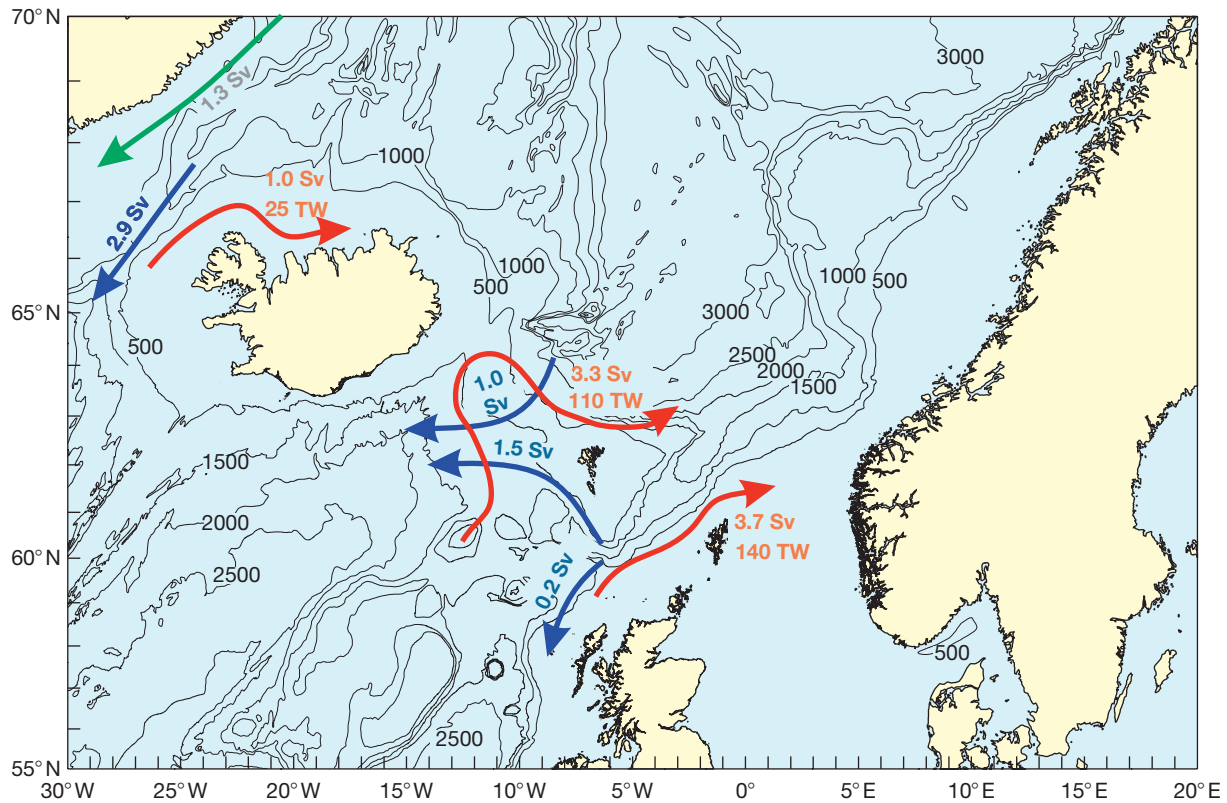


Figure 5 Volume and heat transports across the ridge separating the Nordic Seas from the northern Atlantic. Red marks the surface inflow into the Nordic Seas, blue the deep outflow across the sill, and green the East Greenland Current. Reproduced from Blindheim J (2004) Oceanography and climate. In: Skjoldal HR (ed.) *The Norwegian Sea Ecosystem*. Trondheim: Tapir Academic Press.

density. The flow field adjusts so that both processes are in equilibrium and the density field remains steady.

This basic constraint was understood by Sandström: He found that heating and cooling his tank at the surface could not drive a steady-state, deep overturning flow. Rather, heating at greater depth than cooling was required – a conclusion commonly known as Sandström’s theorem. This heating at depth is a mechanism that makes the deep waters continually less dense, allowing an ongoing replacement by colder water from above.

What is this mechanism in the real ocean, where heating occurs at (or very near) the surface due to the Sun? As Sandström speculated, it is the downward penetration of heat by mixing. The result is the classic advection–diffusion balance: In steady state, at any given point in the depths of the oceans, the slow diffusion of heat by turbulent mixing is balanced by the equally slow transport of cold waters from the poles.

Downward mixing of heat requires energy to overcome frictional dissipation and to raise the potential energy stored in the water column, as heat penetrating downward expands the deep water and lifts up the waters above. This is the energy supply of the THC: In an energetic sense, it is driven by turbulent mixing in the ocean’s interior. This is why mixing appears in the definition of THC given at the outset of this article (quoted from Rahmstorf, 2002).

The power supply of the turbulence that causes this mixing is both by tidal motions and by the winds. The energy supply

needed for generating the observed global overturning motion of ~ 30 Sv can be estimated from the advection–diffusion balance and the observed density field as ~ 0.4 TW (Munk and Wunsch, 1998). This can be compared with the power input by winds and tides, estimated as approximately 1 TW each, although these estimates are rather uncertain. With a mixing efficiency of 20%, 0.4 TW would be available for driving the THC (the remaining 80% is dissipated).

Although the energy supply of the THC thus ultimately comes from winds and tides, it is useful to keep the established term ‘thermohaline circulation’ (as we keep calling ‘steam engine’ a machine that is powered by coal or wood) in order to distinguish this driving mechanism from a completely different kind: the direct generation of large-scale currents by the frictional action of wind on the water surface. Apart from obviously driving surface currents, this mechanism could also be involved in driving the deep-reaching MOC, in a manner proposed by Toggweiler and Samuels (1995).

This mechanism works as follows: Westerly winds over the Southern Ocean, in conjunction with the Coriolis force, lead to a divergence of surface currents in this region and hence to upwelling. In contrast to the majority of the ocean, where wind-driven upwelling is confined to the upper ocean, the upwelling here comes from deep waters. The reason is a peculiar dynamic constraint on the rotating Earth, termed the ‘Drake Passage Effect’: Due to the lack of topographic barriers at the latitude band of Drake Passage, no east–west pressure

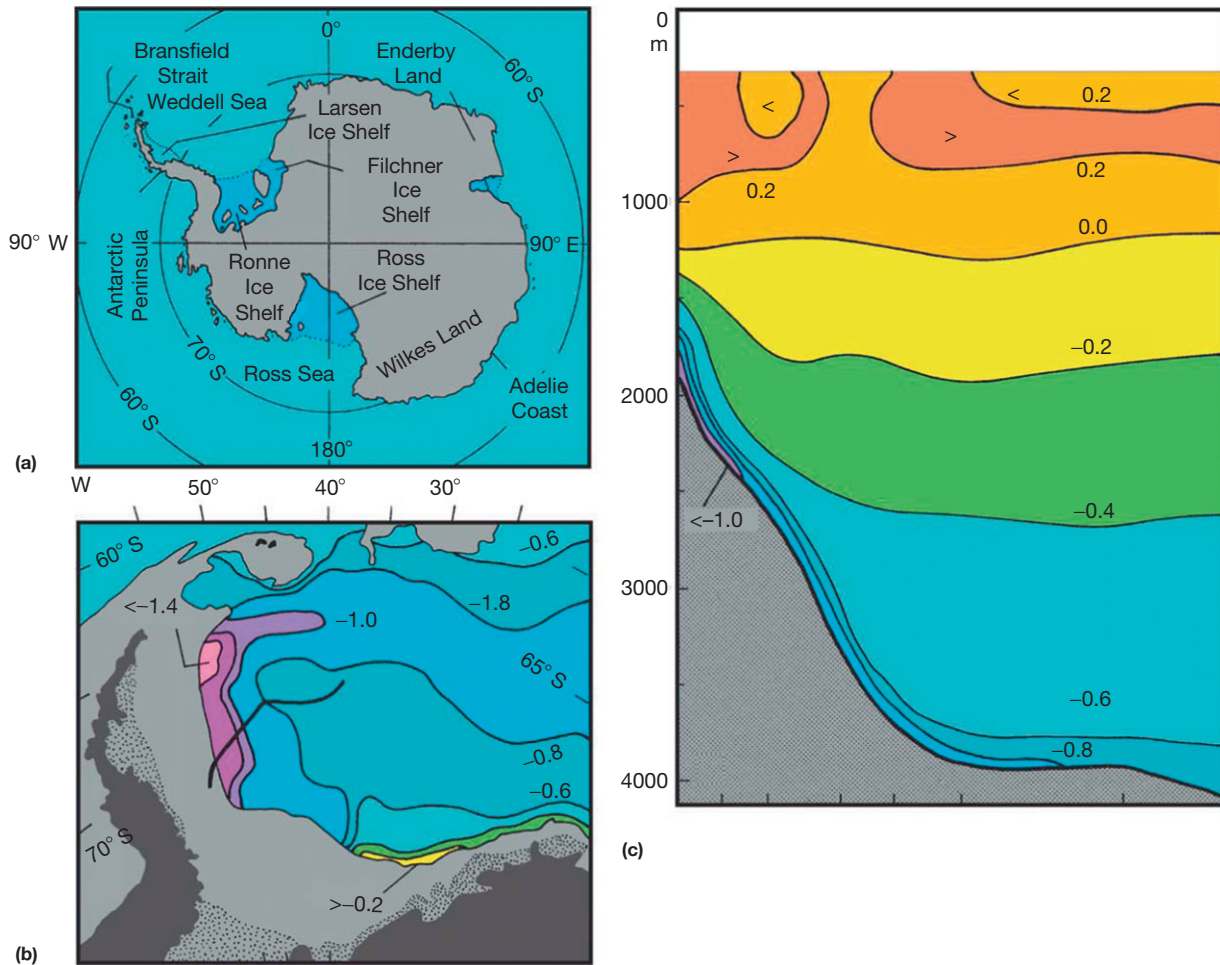


Figure 6 Bottom water formation near Antarctica. (a) Locality map. (b) Bottom potential temperature ($^{\circ}\text{C}$) in the Weddell Sea; the stippled area indicates ice shelf, and the edge of the shaded region is the approximate 3000 m contour. (c) A vertical section of potential temperature, along the heavy line in (b). Reproduced from Tomczak M and Godfrey JS (2003) *Regional Oceanography: An Introduction*, 2nd edn. Available at www.es.flinders.edu.au/~mattom/regoc; Warren BA (1981) Deep circulation of the world ocean. In: Warren BA and Wunsch C (eds.) *Evolution of Physical Oceanography*, pp. 6–40. Cambridge: MIT Press.

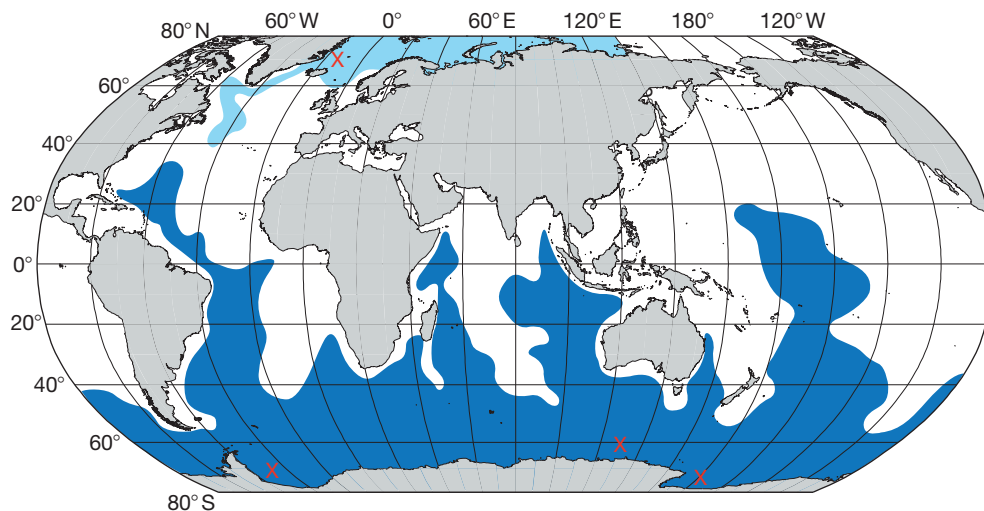


Figure 7 A representation of the spread of Antarctic Bottom Water (AABW, dark blue; based on the $\sigma_4 = 45.92 \text{ kg m}^{-3}$ density surface). Reproduced from Talley LD (1999) Some aspects of ocean heat transport by the shallow, intermediate and deep overturning circulations. In: Clark PU, et al. (eds.) *Mechanisms of Global Climate Change at Millennial Time Scales*, pp. 1–22. Washington, DC: American Geophysical Union.

gradient and hence no meridional flow outside frictional boundary layers can be maintained there. This means that the surface water pushed north by the wind (northward Ekman transport resulting from westerly winds) can only be returned to the south at great depth, below 2500 m, where topographic features exist to support a southward flow. Toggweiler and Samuels (1995) argued that the downward branch required to connect the northward surface flow with this deep southward flow can only occur in the high-latitude north Atlantic because this is the region with the highest surface densities and stable stratification precludes deep water formation elsewhere. In this mechanism, the energy input occurs in the Southern Ocean through the wind creating an Ekman divergence at the surface and convergence at depth, directly 'lifting up' water there, with the MOC looping through the Atlantic to fulfill a continuity requirement.

Which of these two alternative mechanisms actually drives the observed MOC in the world oceans? This discussion is strangely reminiscent of the nineteenth-century debate on wind-driven versus convection currents. Recent literature suggests that both mechanisms play a role (a detailed discussion of the issues raised here and the observational evidence is found in the review article of Kuhlbrodt et al. (2007)). The MOC would thus be composed of both a thermohaline and a wind-driven component (Figure 8).

This view is supported by models: If wind forcing at the ocean surface is turned off in a standard climate model, eliminating the second mechanism, the MOC weakens somewhat but qualitatively remains the same. If interior mixing is reduced, weakening the first mechanism, then the MOC also weakens somewhat but then remains unchanged as one goes to extremely small mixing values. Thus, either mechanism is apparently capable of driving a MOC not unlike that observed. Determining which of them dominates in the real ocean requires further study.

It should be noted that these considerations apply to the mechanism for driving a long-term steady flow, not to the transient response to rapid thermohaline forcing such as a big meltwater inflow (discussed later). The mechanism for interrupting the flow by a density perturbation is different from the driving mechanism; the latter will only come into play on longer timescales.

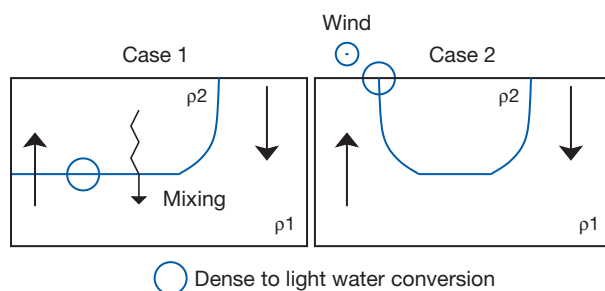


Figure 8 Schematic view of the two driving mechanisms of the meridional overturning circulation discussed in the text. Left: Thermohaline circulation. Right: Wind-driven circulation through the 'Drake Passage' effect. Reproduced from Kuhlbrodt T, et al. (2007) On the driving processes of the oceanic meridional overturning circulation. *Reviews of Geophysics* 45: RG2001.

Nonlinear Behavior of the THC

The highest surface densities in the oceans are reached where water is very cold, whereas lower densities are found in the saltier but warmer tropical and subtropical areas. The THC is thus thermally driven. Nevertheless, the influence of salinity is important and gives rise to an interesting nonlinearity of the system.

Defant's (1929) textbook discussed the fact that cooling and freshwater input have an opposing effect on density in the northern Atlantic and, hence, on the THC. This results in a bistability of the system, first described in a classic paper by Stommel (1961) with the help of a simple box model. Salinity is involved in a positive feedback: Higher salinity in the deep water formation area enhances the circulation, and the circulation in turn transports higher salinity waters into the deep water formation regions (because these are regions where precipitation exceeds evaporation, freshwater would accumulate and surface salinity would drop if the circulation stopped). In Stommel's model the high-latitude salinity increases linearly with the flow, and the flow increases linearly with high-latitude salinity, which combined results in a quadratic equation for the flow as function of freshwater input. This leads to two possible equilibrium states: The system is bistable in a certain parameter range.

The situation can be described with a simple stability diagram showing strength of the MOC as a function of the freshwater input into the North Atlantic. Figure 9 shows the bistable regime and a bifurcation point S, where the circulation breaks down. If freshwater input to the northern Atlantic is systematically increased and decreased, a hysteresis loop between transitions a and d results. This hysteresis behavior appears to be a

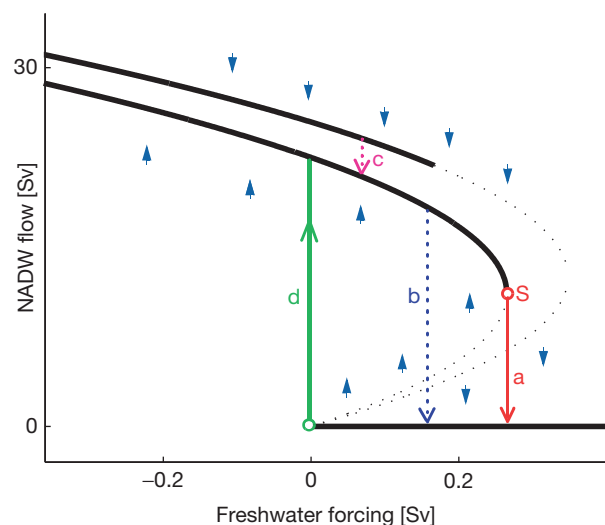


Figure 9 Schematic stability diagram for the Atlantic THC, with solid black lines indicating stable equilibrium states and dotted black lines unstable states. Transitions are indicated by arrows: (a) advective spindown, (b) convective shutdown, (c) transition between different convection patterns, and (d) restart of convection. 'S' marks the Stommel bifurcation beyond which no NADW formation can be sustained. Adapted from Rahmstorf S (2000) The thermohaline ocean circulation – A system with dangerous thresholds? *Climatic Change* 46: 247–256.

robust feature that has been found in many different climate models (Figure 10; Rahmstorf et al., 2005). Although all models tested in this study show qualitatively similar hysteresis behavior, only some have their present-day climate residing in the bistable parameter regime. Whether the real climate system is bistable is unknown.

The salt transport feedback is not the only feedback rendering the system nonlinear. The convective mixing process is itself a highly nonlinear, self-sustaining process. In models this leads to multiple stable convection patterns. On the one hand, this can cause artifacts related to the coarse model grid; on the other hand, this may be part of a real mechanism for shifts in convection location, as have apparently occurred during glacial times.

The Circulation's Effect on Climate

The climatic effect of the MOC is due to its large heat transport of up to 1 PW in the North Atlantic. Back-of-the-envelope calculations suggest that this amount of heat transported into the northern North Atlantic (north of 24° N) should warm this

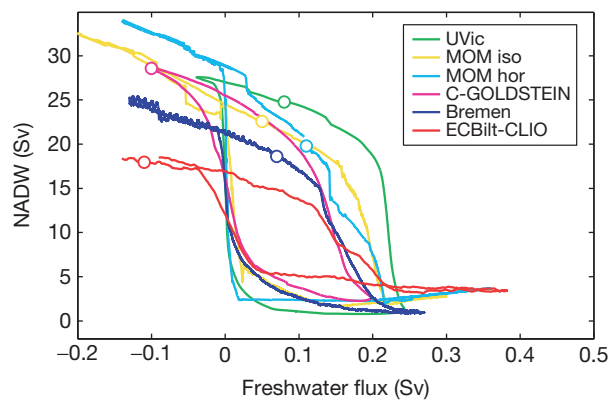


Figure 10 Hysteresis curves from a range of different intermediate-complexity climate models including three-dimensional ocean models. The curves are derived by slowly increasing the freshwater influx to the north Atlantic, then decreasing it again, from a present-day initial state of each model marked by an open circle. Reproduced from Rahmstorf S, et al. (2005) Thermohaline circulation hysteresis: A model intercomparison. *Geophysical Research Letters* 32: L23605.

region by ~ 5 K. This is indeed approximately the difference between sea surface temperature in the North Atlantic compared to the North Pacific at similar latitudes. An examination of sea ice margins suggests that they are pushed back by the warm surface currents in the Atlantic sector compared to the North Pacific (Figure 3), which in turn leads to reduced reflection of sunlight and thus warming (albedo feedback). Examination of observed surface air temperatures is suggestive (Figure 11): Over the three main deep water formation regions of the global oceans, air temperatures are warmer by up to $\sim 10^\circ\text{C}$ compared to the latitudinal mean. These observations, however, are no quantitative proof of the climatic effect of the THC, and other explanations can be partly invoked, such as planetary waves in the atmosphere that are locked in place by the geography.

One way to estimate the effect of the THC is to switch off NADW formation in coupled climate models (by adding a large freshwater anomaly to the northern Atlantic). Roughly, this leads to a cooling with a maximum of typically ~ 10 K over the Nordic Seas (Figure 12). The maximum tends to occur near the sea ice margin due to the positive ice albedo feedback, which causes air temperatures to cool much more than sea surface temperatures. The amplitude and spatial extent of the cooling are somewhat model dependent. Most models tend to cool temperatures over land in northwestern Europe (Scandinavia and Britain) by a few degrees, and some show strong cooling further west affecting Canada.

In models the Northern Hemisphere cools and the Southern Hemisphere warms if the THC is brought to a halt because the cross-equatorial heat transport in the ocean is reduced. This change in heat partitioning between the two hemispheres shifts the thermal equator to the south and thus the intertropical convergence zone (ITCZ) and the associated tropical rainfall belts. This is a robust phenomenon seen in models and consistent with paleoclimatic data.

Finally, a change in circulation also leads to a change in sea level. Currently, sea level in the northern Atlantic is almost 1 m lower than in comparable parts of the North Pacific because deep water forms only in the former. This is due to the geostrophic balance of surface currents and sea surface slopes. Any change in surface currents is associated with a rapid dynamic adjustment of the sea surface topography (with no effect on the global average sea level); in case of a shutdown of NADW formation, the sea level around the northern Atlantic would

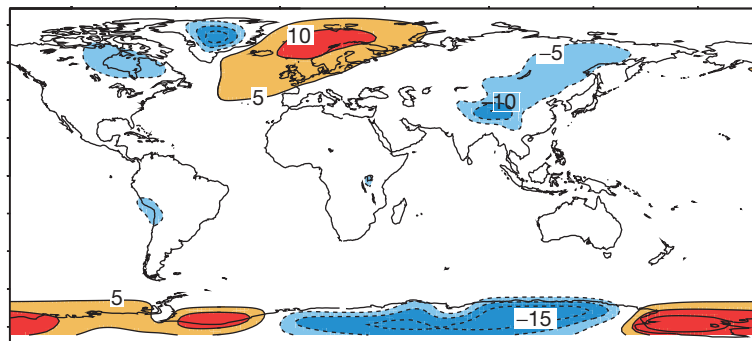


Figure 11 Deviation of surface air temperature from zonal mean ($^\circ\text{C}$). Reproduced from Rahmstorf S (2000) The thermohaline ocean circulation – A system with dangerous thresholds? *Climatic Change* 46: 247–256.

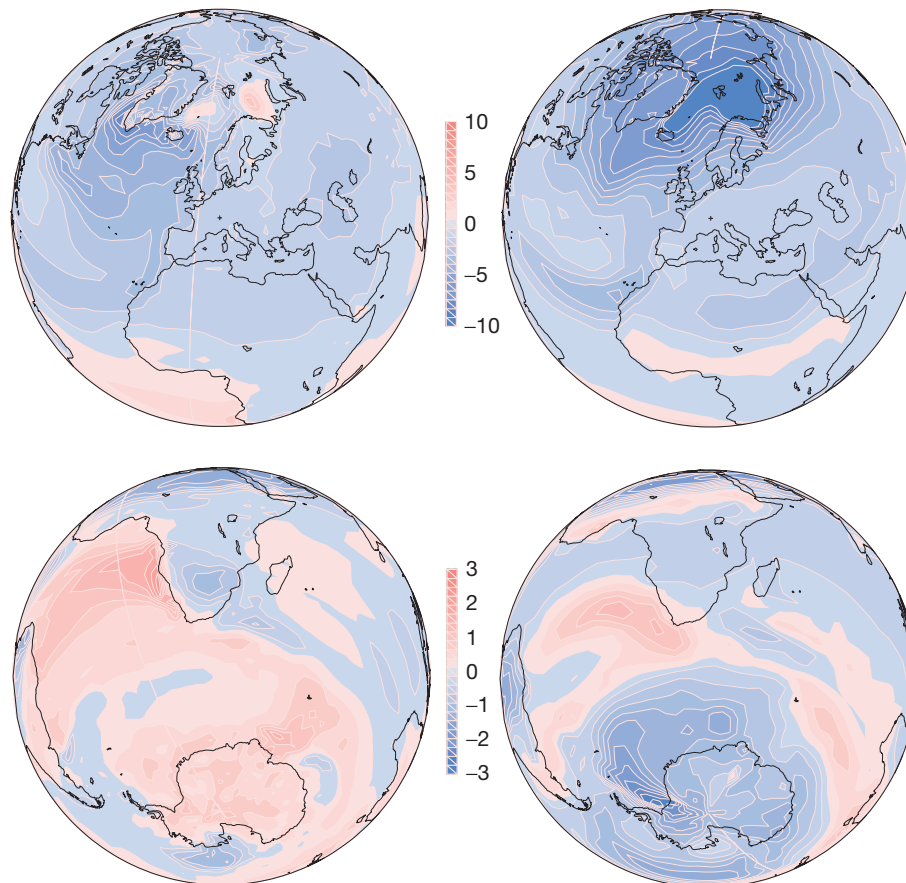


Figure 12 Change in surface air temperature after an induced shutdown of NADW formation in two different climate models. Reproduced from Stocker TF (2002) North-south connections. *Science* 297: 1814–1815.

quickly rise by up to 1 m (Figure 13). A further, slower and global sea-level response would result as the deep ocean temperature changes.

The Role of the THC in Quaternary Climate Changes

Sediment data document that the THC has undergone major changes in the Quaternary (Clark et al., 2002; Rahmstorf, 2002). Three major circulation modes were identified: a warm mode similar to the present-day Atlantic, a cold mode with NADW forming south of Iceland in the Irminger Sea, and a switched-off mode without significant formation of NADW (Figure 14). The latter appears to have occurred after major input of freshwater, either from surging glacial ice sheets (Heinrich events) or in the form of meltwater floods (e.g., Younger Dryas event). The most dramatic climate events recorded in Greenland, the Dansgaard-Oeschger (DO) events, could be explained by north-south shifts in convection location (i.e., transitions between warm and cold modes of the Atlantic THC). The coincidence of changes in ocean circulation and in surface climate, as well as the climate effects and stability properties of the THC discussed previously, suggest that ocean circulation changes can play a key dynamical role in abrupt glacial climate change. Proxy data show that the South Atlantic cooled when the north warmed and vice versa, a

see-saw of Northern and Southern Hemisphere temperatures that is indicative of an ocean heat transport change.

Heinrich events are thought to result from ice sheet instability: An episodic iceberg discharge would have provided a large freshwater addition of the order of 0.1 Sv to the Atlantic for several centuries (Roche et al., 2004). Models suggest this could have been enough to stop all NADW formation due to the reduced density of ocean surface waters; this in turn would explain the cooling found in proxy data, especially around the midlatitude Atlantic. In Greenland, Heinrich events appear to have had little effect, presumably because they occurred during cold phases (cold mode, Figure 14) when the THC already did not reach far enough north to warm Greenland climate.

DO events are dramatic warm events with a large amplitude in Greenland (8–16 °C within approximately a decade; Severinghaus et al., 2003) and areas surrounding the northern Atlantic. A large coinciding increase in salinity in the Irminger Sea suggests a northward push of warm, salty Atlantic waters into the Nordic Seas (Krevelde et al., 2000), which could explain the large warming. What triggered these ocean circulation changes is unknown. Data analysis and modeling have suggested the intriguing possibility that stochastic resonance may have played a role (Alley et al., 2001). This is a physical mechanism that can lead to threshold transitions in systems forced by a weak regular forcing (e.g., a solar cycle) and stochastic variability.

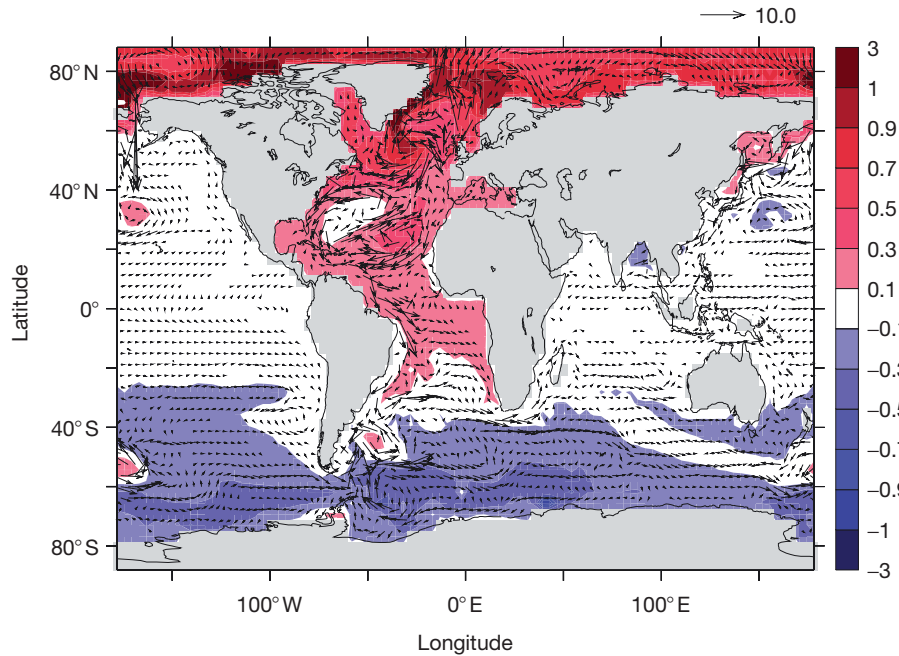


Figure 13 Dynamic sea-level change in a model experiment in which NADW formation was shut down. The global mean of this sea-level change is zero; any global mean sea-level changes resulting from the changed climate occur in addition to the shown redistribution of mass. Reproduced from [Levermann A, et al. \(2005\)](#) Dynamic sea level changes following changes in the thermohaline circulation. *Climate Dynamics* 24: 347–354.

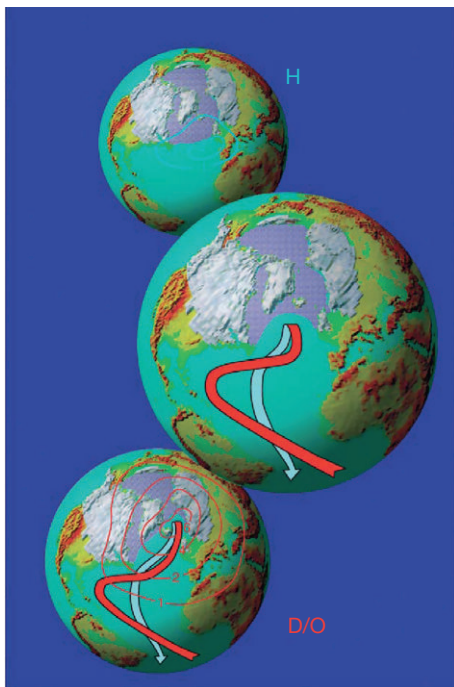


Figure 14 Schematic of three circulation modes of the glacial Atlantic: the prevailing cold mode (middle), the warm mode associated with Dansgaard–Oeschger events (bottom), and the ‘off’ mode occurring after Heinrich events (top).

The role of ocean circulation changes on longer timescales (i.e., for the glacial–interglacial cycles) is still under discussion. Reconstructions of the rate of deep water formation and MOC during glacial times are still tentative but are consistent with prevailing roughly similar, perhaps somewhat weaker overturning rates compared to those in modern times (except during events). Models run for glacial conditions also suggest similar or somewhat weaker overturning rates ([Figure 15](#)) (although a few indicate moderate strengthening). More important for climate than the rate of overturning is the latitude at which deep water forms. Data and some models indicate a southward shift of the main deep water formation sites from the Nordic Seas to a location south of Iceland, which implies a reduced ocean heat transport into high latitudes and an expansion of sea ice. Models suggest that these factors may have amplified glacial cooling, although ocean circulation changes are probably not a crucial factor for glacial inception.

During deglaciation, melting ice sheets surrounding the North Atlantic have apparently led to episodic influx of meltwater into the ocean. The effect of this meltwater on the THC may explain the oscillations and abrupt cold reversals seen, for example, in Greenland ice cores during this time, including the Younger Dryas and the 8 K cold events. This contrasts with the much smoother history of deglaciation found in Antarctic ice cores.

The Future of the THC

Global warming can affect the THC in two ways: surface warming and surface freshening, both reducing the density of high-latitude surface waters and thus inhibiting deep water formation. Most models predict a significant weakening of

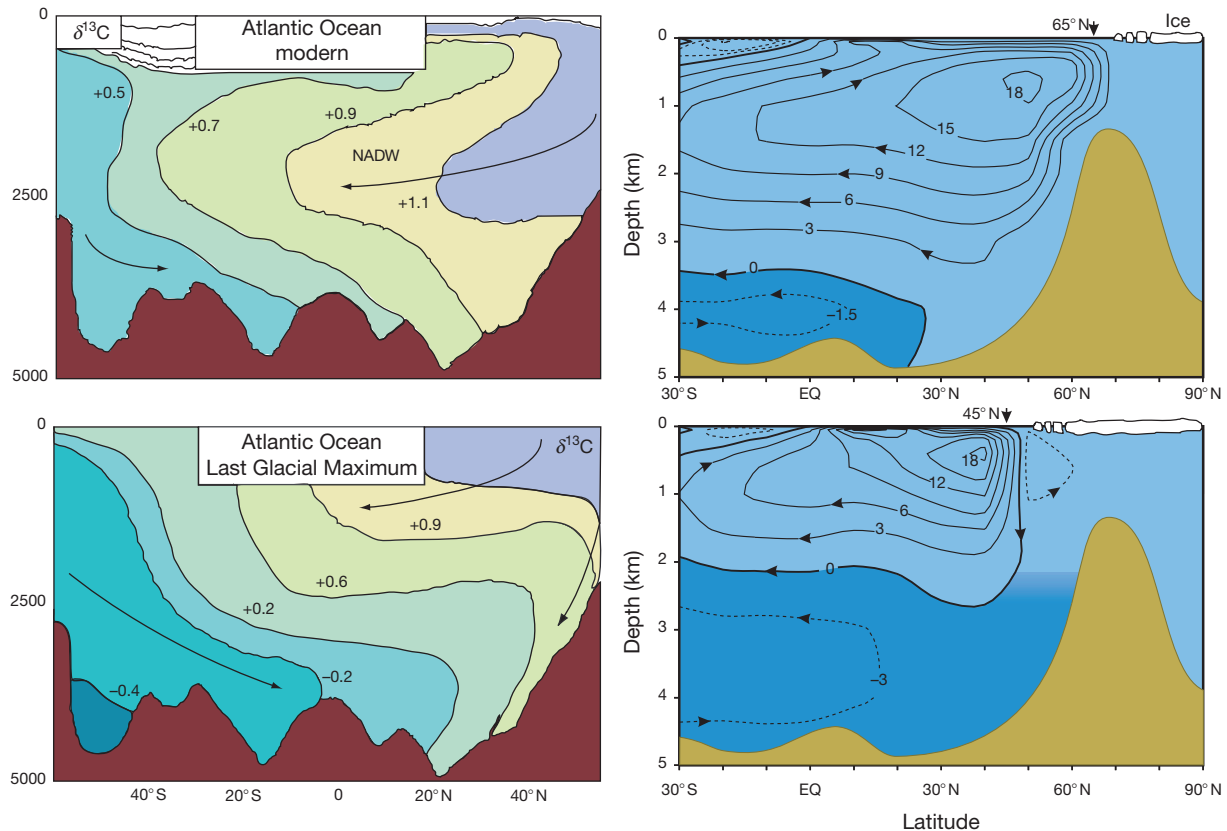


Figure 15 Circulation of modern and glacial Atlantic, as reconstructed by ^{13}C tracer data (left; reproduced from Duplessy J-C, et al. (1988) Deep water source variations during the last climatic cycle and their impact on the global deep water circulation. *Paleoceanography* 3: 343–360) and from a model simulation (right; reproduced from Ganopolski A, et al. (1998) Simulation of modern and glacial climates with a coupled global model of intermediate complexity. *Nature* 391: 351–356). Note the shallower flow of NADW and the upward and northward expansion of AABW (darker blue) in glacial times, as well as the southward shift of deep water formation in this model.

NADW formation (by 20–50%) in response to anthropogenic global warming during the twenty-first century (Intergovernmental Panel on Climate Change, 2001). Some also find a reduction in AABW formation.

These models did not include meltwater runoff from the Greenland ice sheet, which is difficult to predict. A melting of the ice sheet during 1000 years would imply an average meltwater influx of ~ 0.1 Sv, similar to that estimated for Heinrich events. Hence, the future evolution of the Atlantic THC is probably closely tied to the fate of the Greenland ice sheet.

Model simulations – even those that lead to a complete shutdown in the future – find that the influence of anthropogenic warming on the THC until today should be smaller than the natural variability. Therefore, any variations observed to this date are likely related to natural oscillations. However, observations indicate a widespread freshening trend in the northern Atlantic, which, should it continue, could contribute to a future weakening of the THC in the Atlantic (Curry and Mauritzen, 2005).

A major weakening or shutdown of NADW formation could have serious impacts on marine ecosystems, sea level, and surface climate, including a shift in ITCZ and tropical rainfall belts. A THC collapse is widely discussed as one of a number of ‘low probability–high impact’ risks associated with global warming.

See also: **Glacial Landforms, Ice Sheets: Growth and Decay.**
Introduction: Understanding Quaternary Climatic Change.
Paleoclimate Reconstruction: Sub-Milankovitch (DO/Heinrich) Events.

References

- Alley RB, et al. (2001) Stochastic resonance in the North Atlantic. *Paleoceanography* 16: 190–198.
- Blindheim J (2004) Oceanography and climate. In: Skjoldal HR (ed.) *The Norwegian Sea Ecosystem*. Trondheim: Tapir Academic Press.
- Broecker WS (1991) The great ocean conveyor. *Oceanography* 4: 79–89.
- Clark PU, et al. (2002) The role of the thermohaline circulation in abrupt climate change. *Nature* 415: 863–869.
- Curry R and Mauritzen C (2005) Dilution of the northern North Atlantic Ocean in recent decades. *Science* 308: 1772–1774.
- Defant A (1929) *Dynamische Ozeanographie*. Berlin: Springer.
- Duplessy J-C, et al. (1988) Deep water source variations during the last climatic cycle and their impact on the global deep water circulation. *Paleoceanography* 3: 343–360.
- Ganachaud A and Wunsch C (2000) Improved estimates of global ocean circulation, heat transport and mixing from hydrographic data. *Nature* 408: 453–457.
- Ganopolski A, et al. (1998) Simulation of modern and glacial climates with a coupled global model of intermediate complexity. *Nature* 391: 351–356.
- Intergovernmental Panel on Climate Change (2001) *Climate Change 2001*. Cambridge: Cambridge University Press.

- Köhl A, et al. (2007) Interannual to decadal changes in the ECCO global WOCE synthesis. *Journal of Physical Oceanography* 37: 313–337.
- Kreveld Sv, et al. (2000) Potential links between surging ice sheets, circulation changes, and the Dansgaard–Oeschger cycles in the Irminger Sea, 60–18 kyr. *Paleoceanography* 15: 425–442.
- Kuhlbrodt T, et al. (2007) On the driving processes of the oceanic meridional overturning circulation. *Reviews of Geophysics* 45: RG2001.
- Levermann A, et al. (2005) Dynamic sea level changes following changes in the thermohaline circulation. *Climate Dynamics* 24: 347–354.
- Munk W and Wunsch C (1998) Abyssal recipes II: Energetics of wind and tidal mixing. *Deep Sea Research* 45: 1977–2010.
- Rahmstorf S (2000) The thermohaline ocean circulation – A system with dangerous thresholds? *Climatic Change* 46: 247–256.
- Rahmstorf S (2002) Ocean circulation and climate during the past 120,000 years. *Nature* 419: 207–214.
- Rahmstorf S, et al. (2005) Thermohaline circulation hysteresis: A model intercomparison. *Geophysical Research Letters* 32: L23605.
- Roche D, et al. (2004) Constraints on the duration and freshwater release of Heinrich event 4 through isotope modelling. *Nature* 432: 379–382.
- Roemmich DH and Wunsch C (1985) Two transatlantic sections: Meridional circulation and heat flux in the subtropical North Atlantic Ocean. *Deep Sea Research* 32: 619–664.
- Sandström JW (1908) Dynamische Versuche mit Meerwasser. *Annalen der Hydrographie und Maritimen Meteorologie* 36: 6–23.
- Severinghaus JP, et al. (2003) A method for precise measurement of argon 40/36 and krypton/argon ratios in trapped air in polar ice with applications to past firn thickness and abrupt climate change in Greenland and at Siple Dome, Antarctica. *Geochimica et Cosmochimica Acta* 67: 325–343.
- Stocker TF (2002) North–south connections. *Science* 297: 1814–1815.
- Stommel H (1961) Thermohaline convection with two stable regimes of flow. *Tellus* 13: 224–230.
- Talley LD (1999) Some aspects of ocean heat transport by the shallow, intermediate and deep overturning circulations. In: Clark PU, et al. (ed.) *Mechanisms of Global Climate Change at Millennial Time Scales*, pp. 1–22. Washington, DC: American Geophysical Union.
- Talley LD, et al. (2003) Data-based meridional overturning stream functions for the global ocean. *Journal of Climatology* 16: 3213–3226.
- Toggweiler JR and Samuels B (1995) Effect of Drake Passage on the global thermohaline circulation. *Deep Sea Research* 42: 477–500.
- Tomczak M and Godfrey JS (2003) *Regional Oceanography: An Introduction*, 2nd edn. Delhi: Daya Publishing House. Available at: www.es.flinders.edu.au/~mattom/regoc.
- Warren BA (1981) Deep circulation of the world ocean. In: Warren BA and Wunsch C (eds.) *Evolution of Physical Oceanography*, pp. 6–40. Cambridge: MIT Press.

Volcanic and Solar Forcing

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Solar Variability

Prior to large-scale industrialization, climate forcing was primarily due to natural variations in the Sun's output and in volcanic eruptions for timescales up to several thousand years. The Sun also plays a role in orbital forcing, which is driven by changes in the relative position and orientation of the Earth with respect to the Sun (rather than changes in the output of the Sun). The systematic variations in the Earth's precession, obliquity, and eccentricity create three distinct cycles (Milankovitch cycles) that combine to determine the orbital forcing. Summer insolation reaching northern middle latitudes, which plays a major role in ice-sheet growth and retreat, is highly correlated at long timescales ($>20,000$ years) with long-term climate including glacial–interglacial transitions. This orbital forcing is distinct from solar forcing, which refers instead to variations in the irradiance output from the Sun. Over long timescales for which orbital forcing is substantial, solar and volcanic forcing are thought to be comparatively much smaller and are therefore usually neglected.

Solar forcing takes place because the Sun's output varies on many timescales. Some of the most distinct are the 27-day rotation period of the Sun and the approximately 11-year solar cycle (or Schwabe cycle), also called the sunspot cycle and corresponding to half the 22-year cycle of the Sun's magnetic field. A longer cycle of approximately 80 years, called the Gleissberg cycle, modulates both the amplitude and the length of the sunspot cycle. It is difficult to measure the Sun's output precisely from the Earth's surface due to interference in the atmosphere from clouds and particulate, for example. Precise measurements have been obtained from satellites since 1979, and these clearly show variations up through the ~ 11 -year cycle, over which output varies by approximately 0.1% (Fröhlich and Lean, 2004; Willson and Mordvinov, 2003). However, temporal gaps between instrumental records make estimates of longer term changes difficult.

It is clear from the satellite observations that irradiance variations are not uniform across the spectrum, with variations over the ~ 11 -year solar cycle increasing from slightly $<0.1\%$ in the visible to 1–10% in the ultraviolet (UV) and shorter wavelengths. Although the Sun's output at these wavelengths is much less than in the visible, the larger percentage changes still imply that a substantial fraction of the integrated energy variations (~ 15 –20%) takes place at these wavelengths. This radiation is largely absorbed by oxygen and ozone before reaching the Earth's surface, or even the troposphere. However, it has the potential to alter the climate and composition of the stratosphere and above. In particular, UV radiation can produce ozone when it is absorbed by an oxygen molecule in the stratosphere so that increased UV irradiance leads to an increase in stratospheric ozone. This phenomenon has been recorded in both satellite and ground-based ozone data. The

altered ozone as well as the altered irradiance both affect stratospheric temperatures, primarily in sunlit regions, leading to changes in circulation that could in principle affect surface climate. Thus, it is important to know not only how total solar irradiance has varied but also how that variability has depended on wavelength. Unfortunately, even less information is available about past variations in the solar spectrum than the limited 1979–present total irradiance data set, and proxies offer little promise of reconstructing the spectral variation of past irradiance changes.

Some longer term data that appear to be related to total solar output are available. Sunspots records extend back to the seventeenth century, and the modern satellite data confirm that sunspots are highly correlated with irradiance over ~ 11 -year solar cycles. The correlation is an inverse one because the dark sunspots are accompanied by bright regions called faculae during periods of high solar activity, and the faculae exert a stronger influence on total solar output. However, exactly how the sunspot number, a value related to both the number of individual spots and the number of sunspot groups, is related to solar output is unclear. Even more confounding is that there is no good evidence as to whether additional changes in solar output that are not reflected in sunspots took place in periods such as the Maunder Minimum from approximately 1650 to 1710, during which very few sunspots were observed. If the Sun was simply in an extended minimum comparable to that seen in the ~ 11 -year solar cycle, then the reduction in irradiance relative to the mean value in typical cycling conditions would be only one half of one solar cycle's variations, or approximately 0.05%. Many estimates during the past several decades have put the value between several times to more than an order of magnitude greater than this under the assumption that other long-term 'background' changes also took place (Lean, 2000). These were often based on astronomical comparisons with other 'Sun-like' stars, which are now regarded as fairly poor analogs to our own Sun, and recent estimates tend to find substantially reduced long-term solar variations (e.g., Foukal et al., 2004).

Other proxy data have also been used to infer past solar output. The cosmogenic proxies ^{10}Be from ice cores and ^{14}C from tree rings have both been used to reconstruct historical irradiance (Stuiver and Braziunas, 1989). Both isotopes are produced by galactic cosmic rays striking the Earth's atmosphere. The flux of these rays reaching the Earth is modulated by the strength of the solar magnetic field, causing cosmogenic isotope production to vary inversely with solar output. Records of both isotopes show periodicities at many scales, and ^{10}Be , the higher resolution of the two, captures the ~ 11 -year solar cycle reasonably well. Both proxies show periods of extensive lows, of which the Wolf (~ 1280 – 1350), Spörer (~ 1450 – 1540), and Maunder Minima (~ 1645 – 1715) are the most prominent during the past millennium. However,

an absolute calibration of these proxies remains elusive because neither the precise relationship between solar output and the solar magnetic field (and hence isotope production) nor the confounding influence of other factors, such as climate change altering the transport of proxies and their deposition onto ice sheets, are well characterized. Nevertheless, the cosmogenic proxies appear to be reasonably good indicators of when solar irradiance varied despite their limited use in determining the magnitude of such variations. It has been suggested that these proxies show that the Sun is now at its brightest level in many millennia, but further research has demonstrated that this inference appears to result from a flawed calibration technique used in the initial analysis. Thus, the various proxy data sets do not provide a reliable estimate of solar variability during the Quaternary. From the solar physics side, although there has been much progress in helioseismometry (the study of the internal oscillations of the Sun) in recent decades, solar models are still in the early stages of development. Although they have made some notable progress (Wang et al., 2005), they are not able to reproduce some of the known features of the solar dynamo or to spontaneously enter a noncycling state such as the Maunder Minimum. Thus, it is not possible to reliably estimate the precise magnitude of past solar irradiance changes, even for the past two decades, although the time dependence of these changes is known fairly accurately.

Volcanism

The other main natural forcing of climate at timescales of thousands of years or fewer is volcanism. As with solar variability, modern measurements of the optical properties of volcanic aerosols injected into the atmosphere have been available only for the past several decades, with the 1991 eruptions of Mount Pinatubo in the Philippines being by far the best observed. However, reliable optical data on volcanic aerosols (particulate) injected into the upper atmosphere extend back considerably farther than for solar irradiance (Sato et al., 1993) because the observations are technically much simpler due to a much larger signal-to-noise ratio. Prior to the direct optical measurements, a range of historical, geological, and proxy data are used to estimate explosive volcanism. Documentary evidence recording the location of large eruptions has been available for the past several centuries. Geologic evidence allows volcanologists to estimate the explosivity of eruptions, from which the volume of material ejected into the atmosphere can be estimated. One of the most useful techniques is using particulates found in polar ice cores to reconstruct past volcanism. The technique relies on the presence of ash layers in ice cores from the same years in both Greenland and Antarctica, which then implies an eruption with a worldwide impact (Crowley, 2000). In practice, one is more confident if the ash layer can be related to a large, known eruption so that the coincidental eruption of two high-latitude volcanoes near the ice core locations can be ruled out. Since Greenland receives ash from nearby Iceland quite frequently, such a coincidence is not as unlikely as it might seem. Chemical analysis of the ash could also definitively link the ash layers from the two poles, but it is expensive and difficult. Most important, the thickness of the ash layer must then be used to estimate the global impact

of the volcanic aerosols on the Earth's radiation budget (the forcings), an inherently challenging task with substantial uncertainties. Thus, as with solar forcing, the timing of past volcanic forcing is much better known than its magnitude. The records derived from ice cores can be calibrated against modern optical measurements during the period of overlap, however, giving us reasonable confidence in the magnitude of volcanic forcing for at least the past several centuries.

To affect climate, eruptions typically have to inject material into the tropical stratosphere, from which it can be distributed worldwide within a year or two at most. Eruptions that deposit material only in the troposphere usually do not noticeably perturb the global mean climate because the material affects only a portion of the globe and is rapidly removed by the hydrologic cycle. Volcanic aerosols, from eruptions that inject material into the stratosphere but occur poleward of the two subtropical jet streams located at approximately 30° latitude, typically spread out closer toward the pole but not into the tropics or the other hemisphere, and thus seem to have much less global impact. Volcanic aerosols are removed from the stratosphere with an exponential decay time of approximately 1.5–2 years so that a large tropical eruption typically affects climate for 1–3 years. Catalogs of known eruptions now document thousands of events (Simkin and Siebert, 1994). The largest eruption in the past 500 years was that of Tambora in Indonesia during 1815, which caused the famous 'year without a summer' the following year. The largest eruption during the past millennium took place in 1259, but its location has not been determined. In the more distant past, supereruptions such as the Toba eruption ~75 000 years ago (in Indonesia) may have had a major impact on climate. Proxy evidence for the period following Toba's eruption, for example, suggests that it coincided with a cooling epoch lasting ~1000 years, although the connection with volcanism is not firmly established. Similarly, extended periods of enhanced volcanic emissions such as those that created the Deccan or Siberian Traps could have induced climate change by increasing tropospheric aerosols for millennia.

Climate Modeling

The climate impact of solar and volcanic forcing can be investigated by a combination of data analysis and modeling studies. Given an estimate of either solar or volcanic forcing – that is, how the radiative balance of the Earth was changed – a climate model can be used to examine the magnitude, pattern, and timing of the climate response and to explore the underlying physical mechanisms that drive it. These models, of varying complexity, include mathematical representation of features of the climate system, such as the land, oceans, atmosphere, and cryosphere, and the physical, chemical, and biological processes in those regions that affect climate. Energy balance models (EBMs) compute the surface temperature response using a one-dimensional column calculation of changes in incoming and outgoing radiation in response to external forcing. Although simple, these models can tell us much about the global mean annual average response (Crowley, 2000). It is clear from these studies, for example, that volcanic eruptions are the primary cause of externally

forced interannual variations in global mean annual average surface temperature. To fully understand the response of the climate system, however, requires the use of a three-dimensional general circulation model (GCM). These models generate their own inherent variability due to nonlinear chaotic interactions in the highly coupled climate system, which are referred to as internal or unforced variability. This unforced variability provides a baseline against which the response to forcings from outside the climate system, such as solar variations, volcanic eruptions, or greenhouse gases, can be compared. Investigations with these models provide information on the spatial pattern and timing of the climate response to external forcing as well as its magnitude. Studies with state-of-the-art GCMs are limited by available computer power but have advanced dramatically during recent decades. Simulations with GCMs can be compared with observations or proxy data for a wide range of spatial and temporal scales, and the model response can be analyzed to determine the most important processes at work in a particular case. Such comparisons are discussed next.

Proxy Data

The role of solar and volcanic forcing in climate change during the Holocene has been studied more than any other period during the Quaternary. One reason for this is that climate data covering much of the globe are more widely available, especially for the past 1000 years. This allows the proxy data to be compared with estimates of past solar and volcanic forcing, and correlations between the proxy climate records and the forcings can then be compared with the results from climate models used to examine the response to these forcings. Proxies with high temporal resolution, typically from decadal to annual or even seasonal resolution, are generally best for comparison with model simulations. Comparisons for the past millennium can be made with reconstructions of surface temperatures based on networks of proxy data. Of these, tree rings are the most abundant proxy. The width and density of tree rings are related to summer temperatures, and these records can be accurately dated to individual years. Ice core oxygen-18 isotopic ratios also serve as a temperature proxy. These extend back to nearly 1 Ma, making them one of the oldest high-resolution proxies. In sites with high snow accumulation rates, annual temperatures can be retrieved. The limitations of this proxy are primarily that ice cores are restricted largely to the polar regions, and assigning temperatures to calendar years has significant uncertainty. Coral growth rates help fill in the spatial gaps of the ice core records by providing information from the tropics. Documentary evidence is another useful source for the past millennium. Like tree rings, it can be matched exactly with calendar years; however, it is restricted primarily to Europe and East Asia. Annual layers of lake sediments (varves) and annual accumulation in stalagmites (speleotherms) are additional proxies. Finally, deep ocean sediment cores extend farther back in time than any other source, although their interpretation is less straightforward and dating is more uncertain than for the other sources. Additionally, because the ocean integrates transport, temperature, and salinity changes over large spatial and temporal scales, these data typically cannot provide information about the response to

short-term perturbations such as volcanism. During the past millennium, data are plentiful enough that multiple data sets can be combined, allowing reconstructions to be developed for this period with confidence in both the global or hemispheric average temperature changes and the spatial patterns (Jones and Mann, 2004). This is not the case further back in time. However, additional proxy records such as pollen can give a qualitative sense of ancient warm or cool periods over middle latitude continental areas that complements the high-latitude ice core and the oceanic sediment core records.

Model/Data Comparisons

The proxy temperatures clearly show the effects of large eruptions such as Huaynaputina in 1600 or Parker in 1641 (Figure 1), or the later large eruptions of Tambora in 1815 or Krakatua in 1883. In general, these are consistent with the direct observations of the response to recent eruptions such as Pinatubo in 1991. GCM simulations have reproduced the magnitude, spatial pattern, and timing of surface temperature changes in response to the Pinatubo eruption (Robock, 2000; Shindell et al., 2004). Analysis of a large number of eruptions during the past millennium, which improves the statistical reliability of the results, shows a distinct pattern of Northern Hemisphere wintertime warming over the northern continents and cooling over the Labrador Sea region and in the Middle East (Robock, 2000; Shindell et al., 2004). This pattern represents a shift toward the positive phase of the dominant naturally occurring mode of Northern Hemisphere extratropical variability, the large-scale pattern known as either the Northern Annular Mode (NAM) or the Arctic Oscillation (AO). This pattern is closely related to the more regional North Atlantic Oscillation so that in the Atlantic sector they are extremely similar. The hemispheric-scale NAM/AO pattern is visible from the surface up into the stratosphere, however, making it a logical choice for describing the response to solar and volcanic forcing because these have their greatest direct impact on the stratosphere. The relatively good agreement between models and observations in the modulation of the NAM/AO and the resulting Northern Hemisphere continental winter warming following large tropical eruptions is shown in Figure 2. The model result shown in the figure is typical of current GCMs that incorporate the stratosphere (where the volcanic aerosols are injected, but a region often left out of climate models used mainly for other types of studies). It shows the best match between the model and the historical data over Eurasia and the Labrador Sea region, both of which are locations where variations in the NAM/AO have a strong influence on surface temperatures. The response over North America is more difficult to simulate or to determine from data because several of the largest recent eruptions coincided with a strong positive phase in the El Niño/Southern Oscillation (ENSO), which has a strong impact in this region. It is a subject of debate as to whether there is a causal link between volcanic eruptions and ENSO events. The winter continental warming response persists with a similar spatial structure into the second winter following large tropical eruptions but has largely faded away by the third.

Proxy data generally show cooler hemispheric-scale temperatures during solar minima, although the magnitude varies

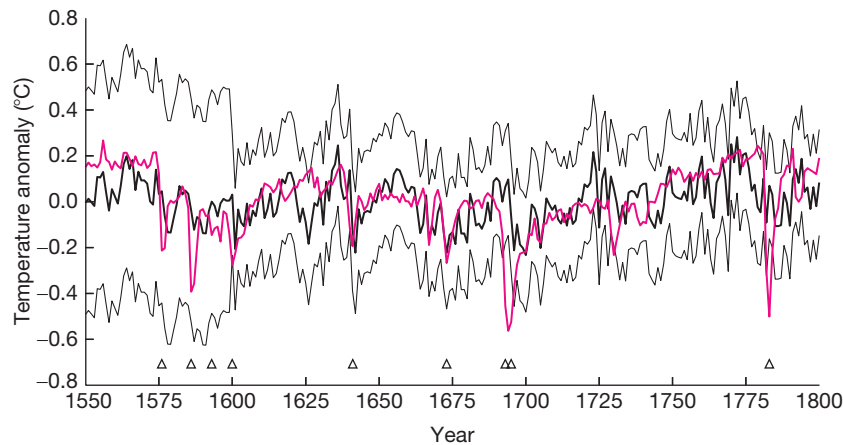


Figure 1 Comparisons between Northern Hemisphere surface temperature anomalies from a proxy reconstruction and a general circulation model (GCM). The proxy reconstruction (Mann et al., 1999) (thick black line) includes an estimate of uncertainty (thin lines bracketing the thick black line), whereas the model results show the mean of six model runs (purple). Large volcanic eruptions included in forcing of the model are marked with triangles, and the response to some of these events is clearly seen in both the model and the proxy data. Adapted from Schmidt GA, Shindell DT, Miller RL, Mann ME, and Rind D (2004) General circulation modelling of Holocene climate variability. *Quaternary Science Review* 23: 2167–2181.

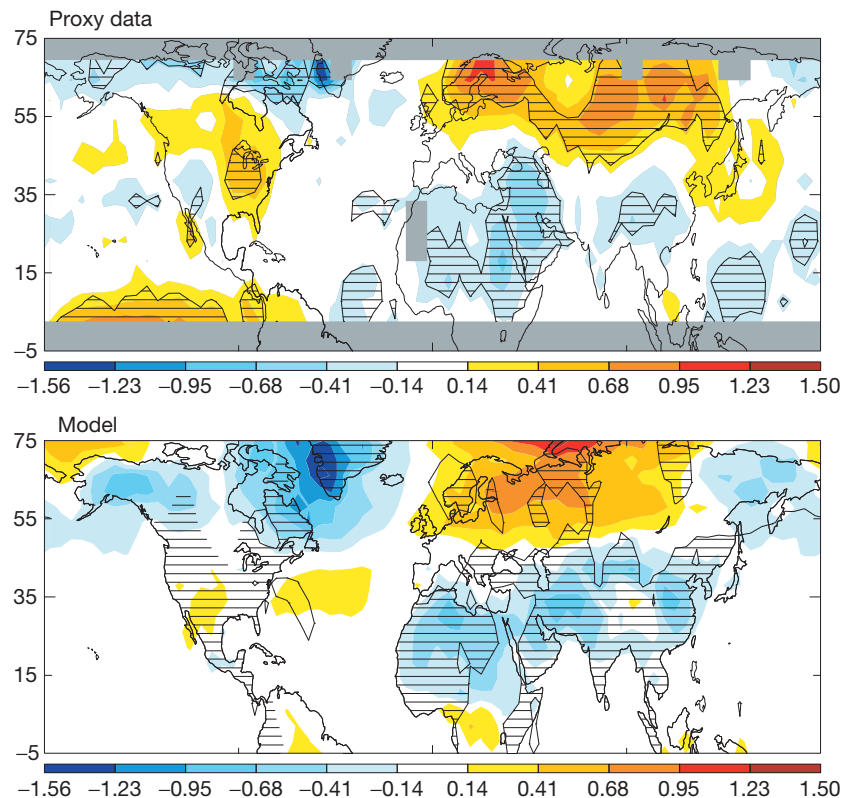


Figure 2 Surface temperature anomalies (°C) during the cold season (October–March) following large tropical volcanic eruptions from proxy reconstructions (top) and in a GCM (bottom). The proxy data are averages of the response to 11 large, well-identified historical eruptions. The model results are the mean cold-season response to 15 simulated eruptions with similar magnitudes to those in the historical analysis. Hatching indicates areas where the response is significant at the 95% confidence level. Modified from Shindell DT, Schmidt GA, Mann ME, and Faluvegi G (2004) Dynamic winter climate response to large tropical volcanic eruptions since 1600. *Journal of Geophysical Research* 109: D05104.

between different reconstructions. EBMs can reproduce the general changes, but given that the magnitude of the changes and the solar forcing are both uncertain, this is not a powerful constraint. GCMs face the same uncertainties in matching global or hemispheric changes. There are two primary ways

around this problem, namely examining the time dependence of climate change in comparison with forcings or examining the spatial pattern of the climate response. In both cases, the importance of uncertainties in the exact magnitude of forcing and response is minimized.

Climate modeling at the NASA Goddard Institute for Space Studies (GISS) examined the global and regional response to volcanic eruptions and solar variability using the GISS climate model. The results were then compared with reconstructions of surface temperatures from historical times, focusing on the change between the late seventeenth and eighteenth centuries. The late seventeenth century, coinciding with the Maunder Minimum, was one of the coldest parts of the period known as the 'Little Ice Age,' when hemispheric average temperatures dropped by a few tenths of a degree. Both reduced solar output and numerous large volcanic eruptions could have contributed to the hemisphere-wide cooling. Regional changes were much larger, however. Europe and eastern North America, for example, were unusually cold, whereas some other areas appear to have been relatively warm.

In the climate model, both the reduced output from the Sun and the increased amount of the volcanic aerosol particles that reflect sunlight away from the Earth contributed a sizeable amount to the Northern Hemisphere cooling seen in temperature reconstructions (Schmidt et al., 2004; Shindell et al., 2003). Their relative importance depends on uncertain estimates of exactly how much each changed relative to a century later. The global or hemispheric average GCM values are quite similar to those obtained in EBM studies (Crowley, 2000).

Regionally, however, solar forcing appears to have played the dominant role in driving annual average temperature changes over periods longer than a year or two (Figure 3). Multiple factors account for this difference. Volcanoes have a major effect on circulation of the atmosphere for a year or two, as discussed previously, but the aerosols fall fairly rapidly out of the atmosphere, resulting in little long-term effect. Furthermore, the effect during the first and second years following eruptions is to increase westerly airflow around the Northern Hemisphere during winter, when circulation responds most strongly to the latitudinal temperature gradients induced by the volcanic aerosols in the lower stratosphere. This enhanced flow brings relatively warm oceanic air over the continents, leading to the continental winter warming shown in Figure 2. In contrast, during the summer, volcanic aerosols lead to cooling via their reflection of sunlight, and this is strongest over the continents, and during summer the volcanic aerosols induce little change in circulation. The regional effects of eruptions during the cold and warm seasons thus tend to cancel one another, leading to a fairly uniform annual average response (Figure 3).

In contrast, the effects of reduced solar output are sustained, lasting for decades, and they also have similar impacts in all seasons. Decreases in solar output lead to reduced heating in the sunlit portion of the stratosphere due to the reduced UV radiation and reduced stratospheric ozone, and ensuing decrease in the meridional gradient of temperature causes decreased wintertime westerly wind flow. This results in less advection of relatively warm maritime air and hence colder Northern Hemisphere continental temperatures (the opposite of the effect following a large tropical eruption) (Shindell et al., 2001). This cooling is reinforced by cooler temperatures everywhere in the summer due to the reduced solar energy, but the land, having a lower heat storage capacity, responds more than the oceans. Thus, during both seasons the land cools in response to reduced solar irradiance, leading to a large regional annual average response.

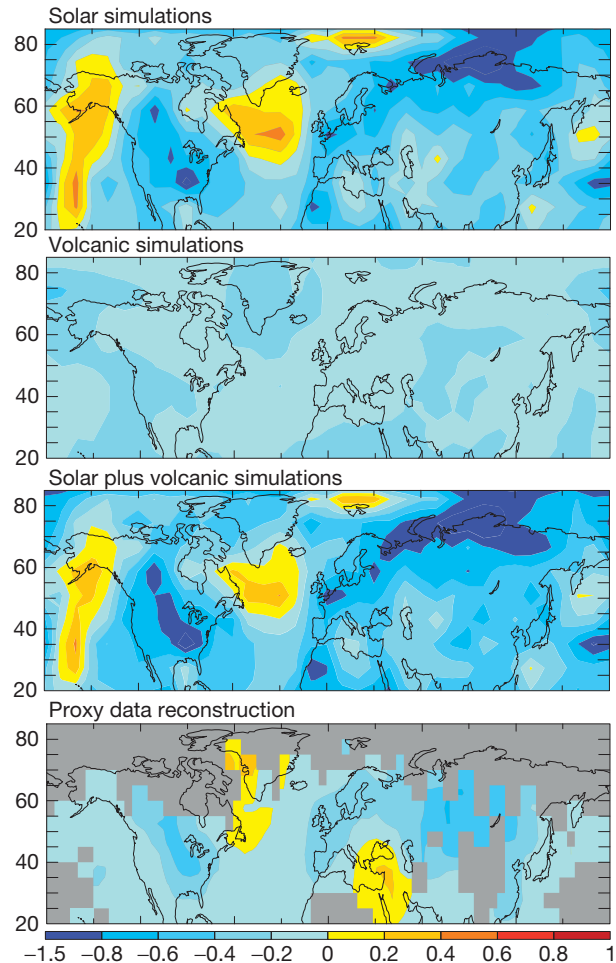


Figure 3 Annual average surface temperature response ($^{\circ}\text{C}$) to decreased solar irradiance (top), substantial volcanic activity (second row), and both forcings (third row) in a GCM. The solar output reduction used in these simulations with the GISS GCM was approximately -0.3 W m^{-2} , whereas the mean volcanic forcing was approximately -0.4 W m^{-2} . Details of the simulations can be found in Shindell et al. (2003). For comparison, the difference between proxy temperature reconstructions averaged over 1680–1710 relative to 1780–1800 is also shown (bottom row). The 1680–1710 period, during the late Maunder Minimum, was a time with both reduced solar irradiance and enhanced volcanic activity relative to a century later.

In the GCM simulations, the annual average response to solar forcing resembles the observations much more closely than the response to volcanic forcing (Figure 3). The response is overall too strong, however, which may indicate an overestimate of how much the Sun dimmed during the late 1600s, an oversensitive model, or both. Nevertheless, it is clear that solar variability seems much more likely to have driven the large regional climate changes seen in historical data. This provides both a useful test of the ability of climate models to simulate patterns of regional changes and important evidence for the importance of solar variability in climate change.

Other modeling studies have demonstrated that the variability in global or hemispheric temperatures induced by volcanic eruptions is fairly well correlated with interannual

variations in the proxy reconstructions. However, at timescales greater than a century, volcanic eruptions are only very weakly correlated with the proxy data, whereas solar irradiance variations are highly correlated. This supports the conclusion that solar changes drive much of the multidecadal and longer variations in Quaternary climate, whereas volcanism is most important at shorter timescales. Further evidence for the role of the Sun in pacing millennial timescale variations in climate comes from records of ocean sediments (Bond et al., 2001) and lake sediments (Hu et al., 2003). These show significant correlations between climate proxy indicators and the solar-related cosmogenic proxies. However, the precise relationship between some of these proxies and climate is unclear, and an understanding of the physical mechanisms underlying such millennial scale interactions between the Sun and climate is lacking.

See also: Paleoclimate Relevance to Global Warming. **Glaciation, Causes:** Astronomical Theory of Paleoclimates. **Paleoclimate:** Introduction. **Paleoclimate Modeling:** Data–Model Comparisons. Last Glacial Maximum GCMs. **Paleoclimate Reconstruction:** The Last Millennium.

References

- Bond G, et al. (2001) Persistent solar influence on North Atlantic climate during the Holocene. *Science* 294: 2130–2136.
- Crowley TJ (2000) Causes of climate change over the past 1000 years. *Science* 289: 270–277.
- Foukal P, North G, and Wigley T (2004) A stellar view on solar variations and climate. *Science* 306: 68–69.
- Fröhlich C and Lean J (2004) Solar radiative output and its variability: Evidence and mechanisms. *Astronomy and Astrophysics Review* 12: 273–320.
- Hu FS, et al. (2003) Cyclic variation and solar forcing of Holocene climate in the Alaskan subarctic. *Science* 301: 1890–1893.
- Jones PD and Mann ME (2004) Climate over past millennia. *Reviews of Geophysics* 42: RG2002.
- Lean J (2000) Evolution of the sun's spectral irradiance since the Maunder Minimum. *Geophysical Research Letters* 27: 2425–2428.
- Mann ME, Bradley RS, and Hughes MK (1999) Northern Hemisphere temperatures during the past millennium: Inferences, uncertainties, and limitations. *Geophysical Research Letters* 26: 759–762.
- Robock A (2000) Volcanic eruptions and climate. *Reviews of Geophysics* 38: 191–219.
- Sato M, Hansen JE, McCormick MP, and Pollack JB (1993) Stratospheric aerosol optical depths, 1850–1990. *Journal of Geophysical Research* 98: 22987–22994.
- Schmidt GA, Shindell DT, Miller RL, Mann ME, and Rind D (2004) General circulation modelling of Holocene climate variability. *Quaternary Science Reviews* 23: 2167–2181.
- Shindell DT, Schmidt GA, Mann ME, and Faluvegi G (2004) Dynamic winter climate response to large tropical volcanic eruptions since 1600. *Journal of Geophysical Research* 109: D05104.
- Shindell DT, Schmidt GA, Mann ME, Rind D, and Waple A (2001) Solar forcing of regional climate change during the Maunder Minimum. *Science* 294: 2149–2152.
- Shindell DT, Schmidt GA, Miller RL, and Mann ME (2003) Volcanic and solar forcing of climate change during the Preindustrial Era. *Journal of Climatology* 16: 4094–4107.
- Simkin T and Siebert L (1994) *Volcanoes of the World: A Regional Directory, Gazetteer, and Chronology of Volcanism during the Last 10,000 Years*, 2nd edn. Tuscon, AZ: Geoscience Press.
- Stuiver M and Braziunas TF (1989) Atmospheric C-14 and century-scale solar oscillations. *Nature* 338: 405–408.
- Wang YM, Lean J, and Sheeley NR (2005) Modeling the sun's magnetic field and irradiance since 1713. *The Astrophysical Journal* 625: 522–538.
- Willson RC and Mordvinov AV (2003) Secular total solar irradiance trend during solar cycles 21–23. *Geophysical Research Letters* 30: 3–6.

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GLACIAL LANDFORMS

Contents

Introduction

Moraine Forms and Genesis

Quaternary Volcanism: Subglacial Landforms

Evidence of Glacier Recession

Glacial Landsystems

Glaciofluvial Landforms of Erosion

Glacitectonic Structures and Landforms

Introduction

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Introduction

The holistic assessment of the character and distribution of glacial sediments and landforms over whole drainage basins or even whole continents facilitates the reconstruction of glacier and ice-sheet dynamics over complete glacial cycles. The accuracy of such paleoglaciological reconstructions relies on analogs with modern examples. Increasing numbers and varieties of modern glacial landsystems are emerging in the research literature based on intensive studies of glacier processes and forms, and these can be utilized by Quaternary scientists interested in paleoenvironmental reconstruction but not specializing in glacial geology. Although paleoglaciological reconstructions are often difficult and sometimes controversial, significant developments in interpreting ancient glaciated landscapes have highlighted complex but coherent patterns of glacier activity over much of the last glacial cycle, and this is now increasingly being linked with regional and global climatic and oceanographic changes. Information on specific glacial processes and their products are reviewed elsewhere in this encyclopedia. This article focuses on basin-wide glacial depositional environments, incorporating interrelated process–form domains in sediment–landform associations that are characteristic of particular topographic and environmental settings or landsystems.

The Subglacial Environment

Ice-Sheet and Glacier Beds

Ice-sheet and glacier beds, rather than being areas devoid of long-term geomorphic records, are actually rich in evidence of ice buildup as well as recession. Large parts of the areas occupied by the former Pleistocene ice sheets are, in effect, fossil

glacier beds, providing a window into subglacial processes and environments that are only poorly known from modern examples, although recent discoveries of landform generation at the base of the Antarctic Ice Sheet are improving the understanding of process–form relationships at the ice–bed interface (e.g., [King et al., 2007, 2009](#)). Regional studies of former glacier beds have shown that the great ice sheets were highly dynamic entities with complex histories of advance and retreat, ice-divide migration, and internal dynamics, and that they shifted their configuration to adapt to climatic change, the position and depth of proglacial water bodies, and basal conditions (e.g., [Boulton and Clark, 1990](#); [Clark, 1993, 1999](#); [Dyke and Prest, 1987](#)). Even former ice streams have been identified in the subglacial bedform record ([Clark and Stokes, 2003](#); [Stokes et al., 2005](#)). With respect to the sedimentary record on former glacier beds, large-scale patterns of ice flow and subglacial conditions may be preserved in the form of distinct till units, glacitectonic deformation, and surfaces of erosion or non-deposition. Assemblages of subglacial deposits and landforms created during the advance and retreat of an ice sheet are typically very complex and often defy analysis by the traditional methods of sedimentary stratigraphy. The products of several glacial advances may be superimposed, and glacitectonic dislocation and subglacial deformation often rework all or part of any preexisting glacial and nonglacial strata. Owing to partial erosion or patchy deposition, tills relating to any one glaciation can be discontinuous so that individual events are only recognizable through the type, style, and orientation of glacitectonic structures.

Theoretical patterns of subglacial deposition and erosion created during an ice-sheet advance and retreat cycle are presented by [Boulton \(1996a\)](#). His model assumes that debris transport below a simple ice sheet occurs entirely by subsole

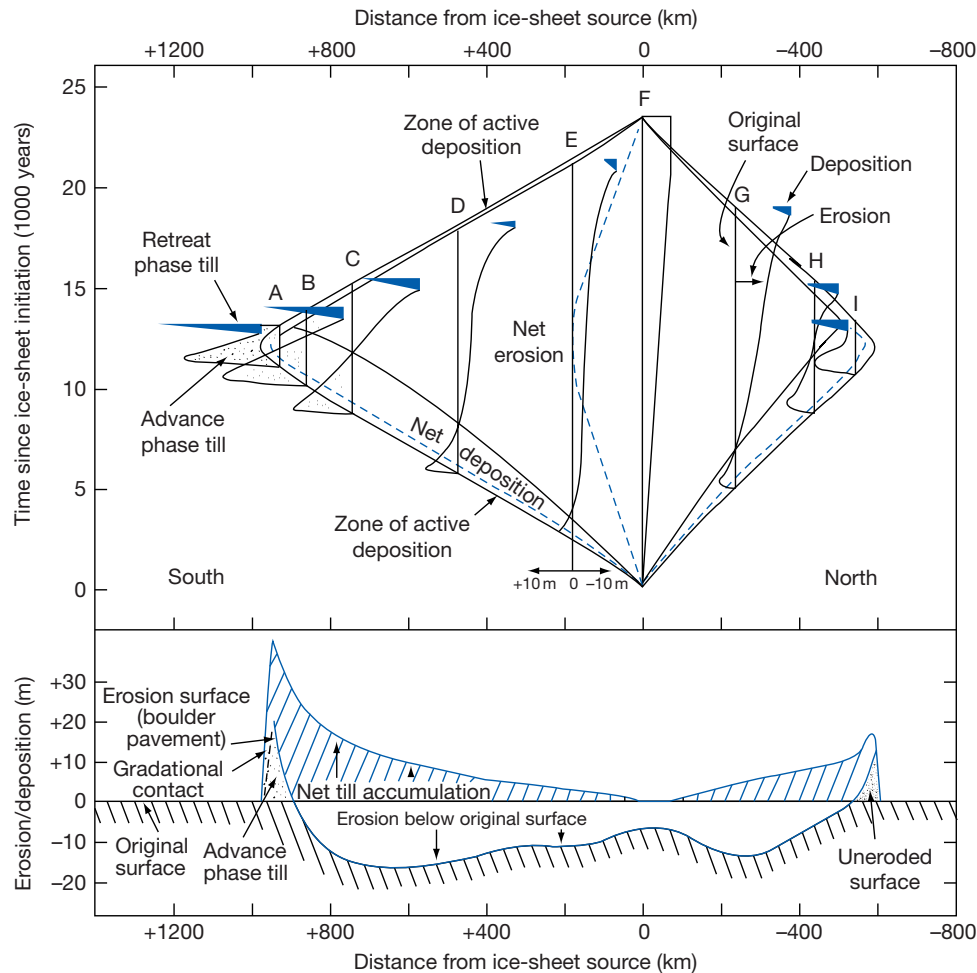


Figure 1 A typical model run from Boulton (1996a), showing subglacial debris transport in a time–space diagram. The ice sheet is initiated at a nucleation point at time $t=0$, and expands to the north and south, reaching its maximum 12 ka after inception. It then contracts into the area where it first developed, finally vanishing at $t=24$ ka. The model predicts that there will be an outer zone of deposition near the ice-sheet margin and an inner zone of erosion, although there will be no erosion below the ice divide where the balance velocity is zero. Reprinted from the *Journal of Glaciology* with permission of the International Glaciological Society.

deformation at rates scaled to the balance velocity of the ice sheet. Although this highly simplified model does not deal explicitly with other important subglacial processes such as the erosion of hard substrata, the entrainment of debris into basal ice layers, and deposition by lodgement and melt-out, it does provide interesting insights into large-scale time-transgressive patterns of erosion and deposition (Figure 1). The model predicts an outer zone of deposition near the ice-sheet margin and an inner zone of erosion, which gives way to a zone of no erosion below the ice divide. The history of deposition and erosion at any given site reflects the migration of the depositional and erosional zones during ice-sheet advance and retreat. At each site, deposition will occur during ice-sheet advance. The resulting till will be removed from most sites by the subsequent erosion phase, although it will be preserved near the ice-sheet limit. As the ice-sheet retreats, the zone of deposition passes over each site in turn, depositing a retreat-phase till. After final deglaciation, the overall pattern of deposition and erosion comprises four major zones: Zone 1 – the ice-divide zone, with slight erosion due to ice-divide migration

and a thin till deposited during the retreat phase (sites E and F); Zone 2 – a zone of strong erosion, in which the advance-phase till and part of the preglacial substratum are removed (the erosion surface is capped by the retreat-phase till (sites B, C, D, G, and H)); Zone 3 – a zone where both advance- and retreat-phase tills are preserved, but with an intervening erosion surface (some of the advance-phase till has been removed, but erosion was insufficient to bite down into the preglacial surface (sites A and I)); Zone 4 – a zone of continuous till deposition and no erosion (this zone only occurs very close to the ice-sheet limit). More complex stratigraphies will result from oscillations of the ice margin and the migration of different basal thermal regimes (Boulton, 1996a,b).

Subglacial Landform–Sediment Associations

At the scale of continental ice sheets, we can identify ‘hard’ and ‘soft’ beds. Former hard beds existed in shield terrain, or ancient continental crust, where subglacial deposits tend to be thin and patchy and commonly streamlined at a wide

range of scales, forming flutings, megaflutings, megascale lineations, rock-cored drumlins, and leeside cavity fills. Bedrock surfaces covered in glacially eroded forms, ranging from microscale features (e.g., striae) to macro- and megascale forms (e.g., roches moutonnees and troughs) are common in such areas. Subglacially streamlined forms such as megaflutings and drumlins commonly record the last (deglacial) ice-flow directions, although earlier ice-flow directions may be recorded in some areas where older lineations were not obliterated by deglacial flow. Ribbed (Rogen) moraine (Dunlop and Clark, 2006) occurs almost exclusively in shield terrains and has been associated with one of three mechanisms: (1) local shearing and thickening of debris-rich basal ice (Bouchard, 1989), (2) partial modification of older streamlined forms following a shift in the ice-divide position (Boulton, 1987; Boulton and Clark, 1990), and (3) a subglacial sediment instability mechanism (Dunlop et al., 2008). Long esker networks are also common in areas of hard substrata, where they record late stages of drainage below ablating ice sheets (Aylsworth and Shilts, 1989; Shilts et al., 1987). Clark and Walder (1994) have argued that eskers are characteristic of hard-rock terrains because channelized, conduit flow is possible at the ice-bed interface. At a continental scale, eskers are more widespread on hard substrata, but they do occur on soft substrata, in some cases superimposed on drumlins and other bedforms. It is possible that on both hard and soft beds, subglacial drainage switched between channelized and distributed systems as ice thickness and gradient, and meltwater supply changed over the course of a glacial cycle.

Soft glacier substrata occur in areas underlain by weak sedimentary rocks or where accumulations of Quaternary sediments mask hard lithologies. Expansive sedimentary bedrock lowlands exist at the margins of the North American and Scandinavian shields (e.g., Kleman et al., 2008), which provided an outer zone of potentially deformable sediment below the great Pleistocene ice sheets. In such areas, the subglacial surface typically consists of fluted and drumlinized terrain underlain by varying thicknesses of subglacial sediments. Late-stage drainage events may be recorded by eskers and/or meltwater channels (Eyles, 1983b; Eyles and Eyles, 1992; Eyles and Menzies, 1983). In many areas occupied by Pleistocene ice sheets, subglacial sedimentary sequences may comprise tens of meters of vertically stacked tills and intervening stratified sediments (Figure 2). Till units may be laterally continuous for tens of kilometers, with remarkably uniform sedimentological characteristics and lithological composition (e.g., Boyce and Eyles, 2000). In many localities, stacked till units record flow interactions between ice lobes (e.g., Eyles et al., 1982; Hicock, 1992). Near the junction between two adjacent ice-flow units, changes in relative ice discharges can cause gradual or abrupt shifts in the ice-flow direction, as one flow unit becomes dominant over the other. As a result, debris from different source areas can be transported to the site, producing marked compositional stratification in which the lithological composition of till units varies systematically up-section. Such compositional changes are commonly paralleled by shifts in the preferred orientation of particle fabric modes, allowing sequences of ice-flow changes to be reconstructed in some detail (Berthelsen, 1978; Boston et al., 2010; Evans et al., 2008; Eyles et al., 1982). Vertical compositional changes can also occur in



Figure 2 A 30-m thick sequence of multiple tills on the Canadian prairies, Milk River, southern Alberta. Tills can be differentiated by color and bounding surfaces marked by clast pavements.

the absence of changes in flow direction. Boulton (1996b) has shown that till composition reflects the duration of glaciation, particularly the length of time available to transport debris from distant locations. Theory suggests that till composition should be dominated by local lithologies near the base and shift to predominantly far-traveled lithologies at the top, reflecting the passage of successive waves of debris from further and further upglacier.

Regional Distributions of Subglacial Landforms

Subglacial sediments and landforms exposed by glacier and ice-sheet recession commonly exhibit systematic spatial patterns, in which different types of landforms occur in distinct zones aligned parallel or transverse to former ice flow. Several factors influence the formation of subglacial sediments and landforms, including strength and hydraulic conductivity of the bed, basal thermal regime and the availability of meltwater, ice velocity, shear stress, and effective overburden pressure. Thus, landform zonation can be viewed as a mosaic produced in a variety of subenvironments beneath the parent ice sheet, either at a single moment in time or as a time-transgressive palimpsest resulting from ice advance or retreat (Kleman et al., 2006; Kleman and Glasser, 2007). Elongate, flow-parallel zones may mark the former position of flow units such as ice streams and marginal lobes, or meltwater drainage pathways, whereas transverse zones may record down-flow variations in basal conditions and their migration through time.

Ice-sheet advance over an area can be recorded by distinctive landform-sediment zones, particularly in regions underlain by weak rocks or unconsolidated sediments. For example, in the northern Great Plains of North America, arcuate belts of hummocky ice-thrust terrain and excavational basins commonly lie at the distal edges of drumlin fields. Some of the thrust belts have a subdued, streamlined appearance, and contain cupola hills and drumlins with low length/width ratios. The excavational basins and thrust ridges document glaciectonic disturbance near the ice margin, probably encouraged by cold-based conditions below the snout or surging (cf. Bluemle and Clayton, 1984; Clayton and Moran, 1974; Evans et al., 1999, 2008), and the streamlined terrain and drumlin belts record the subsequent advance of wet-based ice over the former marginal zone.

In regions where subglacial deformation was intense and prolonged, glactectonized thrust masses may be buried, their upper surfaces intensely sheared and incorporated within the subglacial deforming layer. The amount of subglacial modification commonly decreases down-ice, and landform assemblages exhibit a continuum from drumlins with high length/width ratios, through drumlins with low length/width ratios and cupola hills, to largely unmodified thrust masses. Such landform associations apparently represent progressively less effective or short-lived subglacial molding toward the ice limit. [Boyce and Eyles \(1991\)](#) attribute down-ice changes in drumlin morphometry and sedimentology in the Peterborough drumlin field south of the Dummer Moraine, Ontario, to variations in substratum characteristics and duration of glaciation. Specifically, they consider the down-ice variations in drumlin morphometry to be a function of decreasing duration of subglacial deformation toward the ice margin, with the northern part of the area having been stripped of its preexisting sediment because it was located beneath the ice-sheet interior (*cf.* [Boulton, 1996a](#)).

Cross-Cutting Lineations and Ice-Sheet Dynamics

Regional mapping of flow-parallel and transverse landforms can reveal large-scale patterns of ice-sheet flow and their

changes through time, yielding valuable information on ice-sheet evolution and dynamics. Reconstructed flow patterns for ice-sheet maxima demonstrate that ice sheets were not simple domes from which ice spread radially, but rather comprised multiple accumulation and dispersal centers ([Boulton et al., 1985](#); [Jansson et al., 2003](#); [Punkari, 1995](#); [Stokes et al., 2005](#)). Furthermore, it is possible to reconstruct the migration of ice divides and the switching of ice streams during a glacial cycle by mapping cross-cutting and superimposed subglacial bedforms (see [Kleman et al., 2006](#) for a review), assessing till fabrics and tracing indicator erratics. Analysis of Landsat satellite images and aerial photographs of glaciated terrains has enabled glacial geomorphologists to reconstruct changing ice-sheet flow directions, and numerous ice-flow landform assemblages have been identified, composed of drumlins, megafaultings, and megascale lineations, which are commonly superimposed on or cross-cut one another ([Figure 3](#)). [Clark \(1993\)](#) identified two relative age indicators that enable the sequence of ice flows to be reconstructed: (1) 'simple superimposition,' where one set of streamlined lineations is superimposed over another set with a different orientation, and (2) 'preexisting lineation deformation,' where deformation during a more recent ice-flow phase alters the form or continuity of preexisting lineations. According to Clark, the degree of

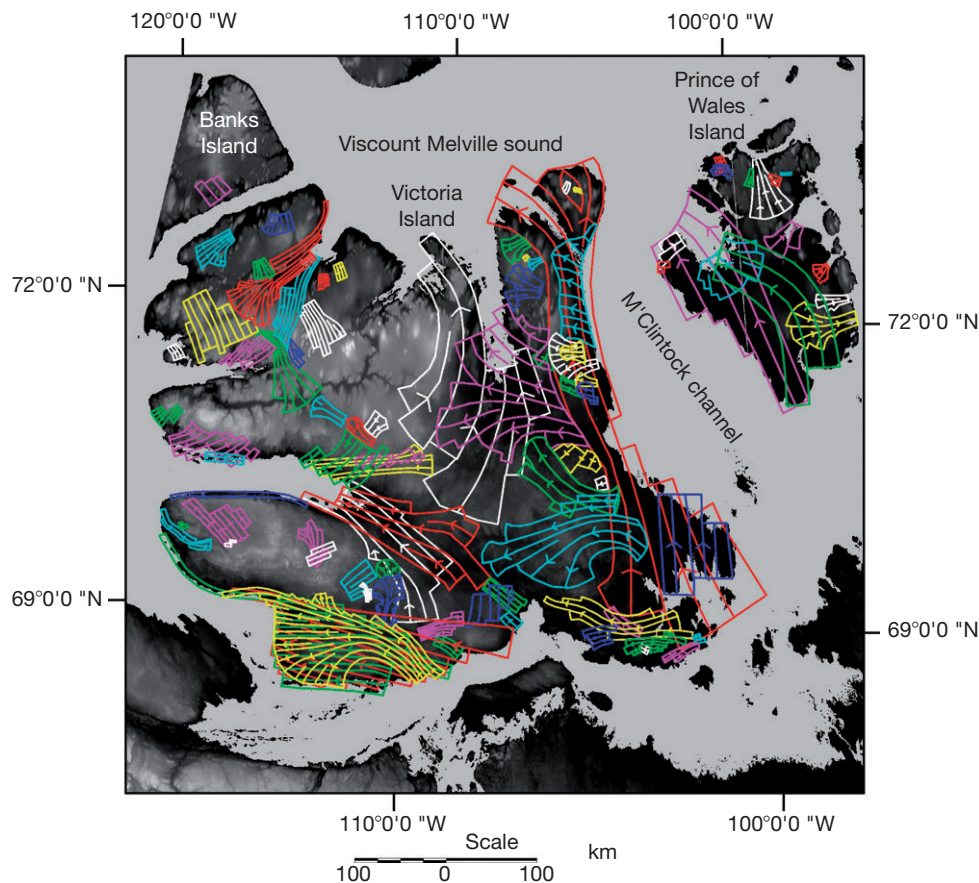


Figure 3 Flow sets comprising coherent assemblages of subglacially streamlined landforms (megalineations) produced during subsequent flow phases at the base of the former Laurentide Ice Sheet over Victoria Island, Arctic Canada. The flow sets record complex cross-cutting by ice streams that changed flow direction due to migrating ice divides. Reproduced from [Stokes CR, Clark CD, Darby DA, and Hodgson DA \(2005\)](#) Late Pleistocene ice export events into the Arctic Ocean from the M'Clure Strait ice stream, Canadian Arctic Archipelago. *Global and Planetary Change* 49: 139–162.

modification of earlier lineations forms a continuum, ranging from (1) a situation where there is no modification of the preexisting lineation to (2) a superimposition of smaller forms on the surface of older lineations to (3) a substantial breaching or deformation of the preexisting lineation and to (4) a total reorganization of sediment into a new orientation.

Ice-Sheet Marginal and Supraglacial Landform–Sediment Associations

At the margins of ice sheets and ice caps, supraglacially entrained debris is rare or absent and therefore debris for the construction of ice-marginal landforms must be derived from the glacier foreland or the bed, and elevated to higher levels by either proglacial or englacial glaciectonic processes, or pressurized meltwater flowing in englacial conduits. The processes and patterns of debris entrainment, elevation, and deposition at glacier margins are strongly influenced by glacier thermal regime and ice-margin activity, allowing some of the climatic and dynamic characteristics of former ice masses to be reconstructed from geological evidence.

Ice-sheet marginal and supraglacial landform–sediment associations can be differentiated on the basis of the nature of the glacier body responsible for their emplacement, particularly the dominant basal thermal regime. [Evans \(2009\)](#) and [Benn and Evans \(2010\)](#) use the concept of a thermal regime process–form continuum to help explain the spatial and temporal variability in landform–sediment associations produced at ice-sheet and glacier margins. At one end of the continuum are warm-based glacier termini, found in mid-latitude settings such as Iceland and southern Norway, where recessional dump and push moraines record margin oscillations on annual or longer timescales. At the other end of the spectrum are the termini of predominantly cold polythermal glaciers, where deformation and thickening of basal ice take place in the marginal cold-based zone, allowing accretion of thick sequences of debris-rich ice and the development of controlled moraine. In the middle of the process–form continuum lie moraines constructed in areas of discontinuous or sporadic permafrost, by glacier margins that can respond to decadal or seasonal climate forcings. Consequently, the typical range of landform–sediment associations can be reviewed under the three categories of active temperate, polythermal, and polar-continental margins.

Active Temperate Glacier Margins

Temperate glacier margins can be defined as those that are mainly wet-based, for at least part of the year, and generally occur in areas of discontinuous permafrost or in permafrost-free areas. A narrow frozen zone develops below the margin of many temperate glaciers during the winter, due to the penetration of a cold wave from the atmosphere through the thin ice at the edge of the glacier ([Kruger, 1994](#)). Additionally, small and discontinuous areas of net freezing can exist below some temperate margins in areas of discontinuous permafrost (e.g., [Lawson, 1979](#)), in some instances amounting to thick basal debris-rich ice sequences where supercooling takes place (e.g., [Alley et al., 1998](#); [Larson et al., 2010](#); [Lawson et al., 1998](#)).

Deposition at temperate glacier margins typically produces small dump moraines or push and squeeze moraines derived from sediment exposed on the glacier foreland. Modern examples have been described from Iceland, where thrusting mainly affects ice-contact fans built up against the ice front but can also relocate slabs of subglacial till ([Boulton, 1986](#); [Evans and Hiemstra, 2005](#); [Humlum, 1985](#); [Kruger, 1985, 1993, 1994](#)). Glaciers without extensive debris cover respond rapidly to seasonal temperature fluctuations, so glacier margins tend to oscillate on an annual basis even during recession, hence the term ‘active temperate.’ As a result, suites of recessional moraines are commonly formed during deglaciation except where glacier velocities are low and/or ablation continues throughout the year ([Boulton, 1986](#); [Kruger, 1994](#); [Sharp, 1984](#)). The areas between recessional moraines represent periods of rapid glacier retreat and commonly exhibit well-preserved subglacial landforms ([Evans and Twigg, 2002](#); [Evans et al., 2009](#); [Figure 4](#)).

Meltwater can have a profound influence on the landforms and sediments formed at temperate glacier snouts, and subglacial, ice-marginal, and supraglacial depositional assemblages are commonly substantially modified by proglacial glaciofluvial deposition and erosion during and following deglaciation. Landforms such as drumlins, eskers, and moraines may be heavily reworked by meltwater and/or buried by outwash sediments so that they are left as remnants or inliers within sandur plains ([Figure 4](#)). Even large ice-marginal moraines can be dissected or partially destroyed by meltstreams directly after emplacement ([Evans and Twigg, 2002](#); [Kruger, 1994](#); [Marren, 2002](#); [Price, 1969](#); [Thompson, 1988](#)). In some cases, virtually the entire depositional legacy of a temperate ice lobe takes the form of glaciofluvial sediments. Where subglacial and englacial drainage systems are well developed, sands and gravels can accumulate within conduits, and meltstreams emerging at the glacier surface can deposit considerable quantities of glaciofluvial sediment in supraglacial positions. Indeed, on some glaciers, glaciofluvial sediment constitutes the major part of the englacial and supraglacial sediment load close to the margins ([Gustavson and Boothroyd, 1987](#)). During deglaciation, uneven ablation of sediment-covered ice can lead to the development of ‘glacier karst’ ([Clayton, 1964](#)). Sediment deposited within conduits and in supraglacial outwash fans and lakes produces a complex assemblage of landforms including esker systems, kame ridges and plateaus, and pitted outwash. Distinctive spatial associations of glaciofluvial landforms deposited during ice wastage have been termed morphosequences ([Koteff and Pessl, 1981](#)). In such sequences, proglacial outwash passes up valley into kame and kettle topography, which is inset within ice-marginal kame terraces and moraines. Kame and kettle topography is locally refashioned into suites of river terraces by proglacial and postglacial streams.

Polythermal Glacier Margins

Polythermal margins are diverse in character as dictated by the extent of frozen and temperate ice at the bed, which in turn is controlled by the environmental conditions and glacier thickness. In essence, they are characterized by an outer cold-based zone and an upglacier wet-based zone. At present, such glacier margins are found mainly in areas of continuous permafrost. The thermal regime and geomorphic impact of glaciers in



Figure 4 Extract from a map of glacial landforms on the foreland of Breidamerkurjokull, Iceland. Tills are orange and their surfaces are fluted (green lines). The flutings are arranged in linear zones separated by push moraines (black line symbol). Overridden push moraines are colored red and eskers are denoted by arrow head symbols. Note that glaciofluvial outwash occupies corridors through the subglacial landforms and sediments, where meltwater excavated channels during deglaciation of the foreland. Reproduced from [Evans DJA and Twigg DR \(2002\)](#) The active temperate glacial landsystem: A model based on Breidamerkurjokull and Fjallsjokull, Iceland. *Quaternary Science Reviews* 21: 2143–2177.

permafrost regions is determined by ice thickness and activity ([Dyke, 1993](#)). Ice caps growing on thick permafrost must initially be entirely cold-based, but wet-based zones can develop in response to subglacial heating from geothermal heat and the strain heating produced by glacier flow. Thawing of the bed is most likely where glacier velocities are highest (e.g., in the vicinity of the glacier equilibrium line and in areas of flow convergence), and therefore, wet-based ice will be most extensive below glaciers with relatively high mass throughput and/or strong converging flow. In the upper parts of glacier accumulation areas, cold-based zones are likely to persist due to

low ice velocities and downward advection of cold firn. Cold-based ice occurs beneath the glacier margins where ice is thin, velocities are low, and proglacial permafrost extends below the glacier, and therefore, small low-activity glaciers in permafrost regions are likely to be cold-based throughout ([Etzelmueller and Hagen, 2005](#)).

The presence of wet-based ice up glacier from a frozen margin in polythermal glaciers has a profound impact on processes of debris entrainment, transport, and deposition. First, erosion below wet-based, sliding ice provides a source of debris which can be transported toward the margin in

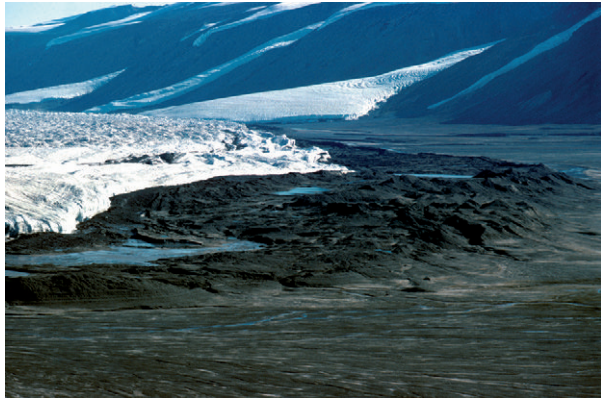


Figure 5 Thrust-block moraine developed in raised marine sediments at the margin of the Eugenie Glacier, Dobbin Bay, Ellesmere Island, Arctic Canada. Note the marine shorelines notched into the front of the moraine attesting to higher sea levels during moraine construction.

regelation ice or a subglacial deforming layer. As this debris is transported into the frozen zone at the glacier margin, sequences of debris-rich basal ice develop as a result of net adfreezing (Hubbard, 1991; Weertman, 1961). These debris sequences are commonly thickened and elevated toward the glacier surface by strong compressive deformation resulting from ice deceleration at the frozen margin (Boulton, 1972; Etzelmuller et al., 1996; Goldthwait, 1951; Hambrey and Huddart, 1995; Knight, 1989; Tison et al., 1993), where bands of controlled moraine can be produced and later develop into ice-cored moraine arcs (Evans, 2009). Thick basal debris sequences also form where glaciers override buried ice dating from earlier glacial events (Evans, 1989b; Hooke, 1973). The long survival of buried ice in permafrost environments (e.g., Dyke and Evans, 2003) means that debris-rich ice may be recycled many times over several glacial cycles as glaciers advance and retreat. Common landforms at polythermal glacier snouts are composite ridges and thrust-block moraines, particularly in wide valley bottoms close to sea level where thick sequences of stratified sediment occur (Etzelmuller et al., 1996; Evans, 1989a,b, 1990; Evans and England, 1991; Hambrey and Huddart, 1995; Kalin, 1971; Figure 5). Where proglacial thrusting does not occur, glaciers commonly terminate in steep ice cliffs along which elevated basal debris is exposed, and frontal aprons, composed of dry calved ice blocks and fallen or in-washed debris, accumulate (Evans, 1989b). Following ice-margin retreat, belts of hummocky and controlled moraine are deposited upvalley from frontal aprons as a result of meltout and reworking of englacial debris. If the debris cover is thicker than the depth of summer melting (the active layer thickness), ice cores may survive indefinitely within frontal aprons and associated hummocky and controlled moraine (Evans, 2009). In cases where bedrock is close to the surface or glacier activity is low, moraines formed at the margins of polythermal glaciers may amount to no more than a line of boulders, attesting to a very low turnover of debris. Recessional moraines are uncommon and basal tills are rare or thin and patchy (Evans, 1990).

The amount of available debris reaches a minimum where polythermal glaciers are predominantly or entirely cold-based. Former glacier margins in such areas are most often marked by lateral meltwater channels cut into bedrock (Dyke, 1993;

Evans, 1990). These lateral channels form because surface meltwater is prevented from penetrating the glacier by low ice temperatures, so that it must drain over the surface and around the margin of the glacier. Where cold glacier tongues descend to the sea or ice-dammed lakes to form glacier ice shelves, they can deposit ice-shelf moraines (Dyke and Prest, 1987; England et al., 1978; Hodgson, 1994; Hodgson and Vincent, 1984; Sugden and Clapperton, 1981).

Polar-Continental Glacier Margins

Distinctive glacial landsystems are formed around the margins of glaciers in the polar deserts (Fitzsimons, 2003), for example, fringing parts of the Antarctic Ice Sheet, where the environment is cold, windy, and has a moisture deficit because sublimation exceeds precipitation. Local glaciers are entirely cold-based and are nearly free of debris, although some of the larger terrestrial outlet tongues of the Antarctic Ice Sheet are partly wet-based with cold-based margins and have stacked basal debris sequences similar to those of subpolar polythermal glaciers (Shaw, 1977a,b). Where thick basal ice sequences crop out, marginal aprons accumulate by gravitational processes, but with very limited reworking by meltwater. Within the marginal aprons, belts of hummocky and controlled moraines can develop from stacked and thrust englacial debris sequences. The topography of controlled moraines is inherited from englacial structures (Figure 6), such as debris bands and folds in debris-rich basal ice, and commonly forms ridges transverse to glacier flow (Shaw, 1977a,b). Entirely cold-based glaciers without stacked basal debris sequences leave very little imprint on the landscape, and end moraines commonly consist of little more than low ridges and, unlike subpolar glacier margins, marginal meltwater channels are rare. Thrust-block moraines occurring in front of some polar-continental glaciers (Fitzsimons, 1997) appear to require special conditions, because most of the proglacial sediment is deeply frozen and ice-cemented and has a high shear strength; it is likely that these moraines form by the proglacial thrusting of unfrozen lake-bottom sediments, because lake floors can remain unfrozen below a lake ice cover until glaciers advance into them.

Subaquatic Glacial Depositional Systems

Subaquatic depositional systems form a very important component of the ancient glacial geological record because (1) large parts of the Pleistocene ice sheets terminated in the sea or large proglacial lakes, depositing huge volumes of sediment in standing water; (2) subaquatic sediments tend to have high preservation potential and commonly preserve a more complete record of glaciation than terrestrial depositional systems; and (3) the study of subaquatic depositional systems yields important insights into the dynamics of former ice sheets, including the patterns of deglaciation, the location of the grounding lines of former ice streams, and episodes of rapid calving retreat and ice rafting as recorded in Heinrich Events.

Stratigraphic Architecture

The large-scale geometry or 'stratigraphic architecture' of subaquatic depositional systems is crucial to understanding environmental change in glacially influenced basins (e.g., Boulton,

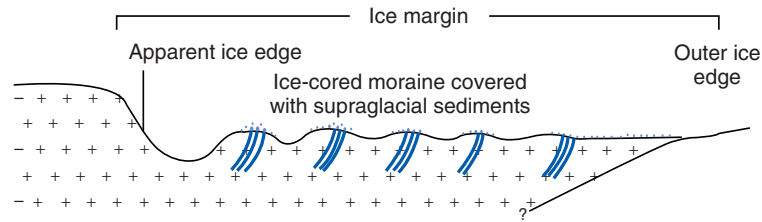


Figure 6 Schematic diagram showing the origin of ice-cored (controlled) moraine ridges at the margin of a polar glacier. Reproduced from Fitzsimons SJ (2003) Ice-marginal terrestrial landsystems: Polar continental glacier margins. In: Evans DJA (ed.) *Glacial Landsystems*, pp. 89–110. London: Arnold, with permission.

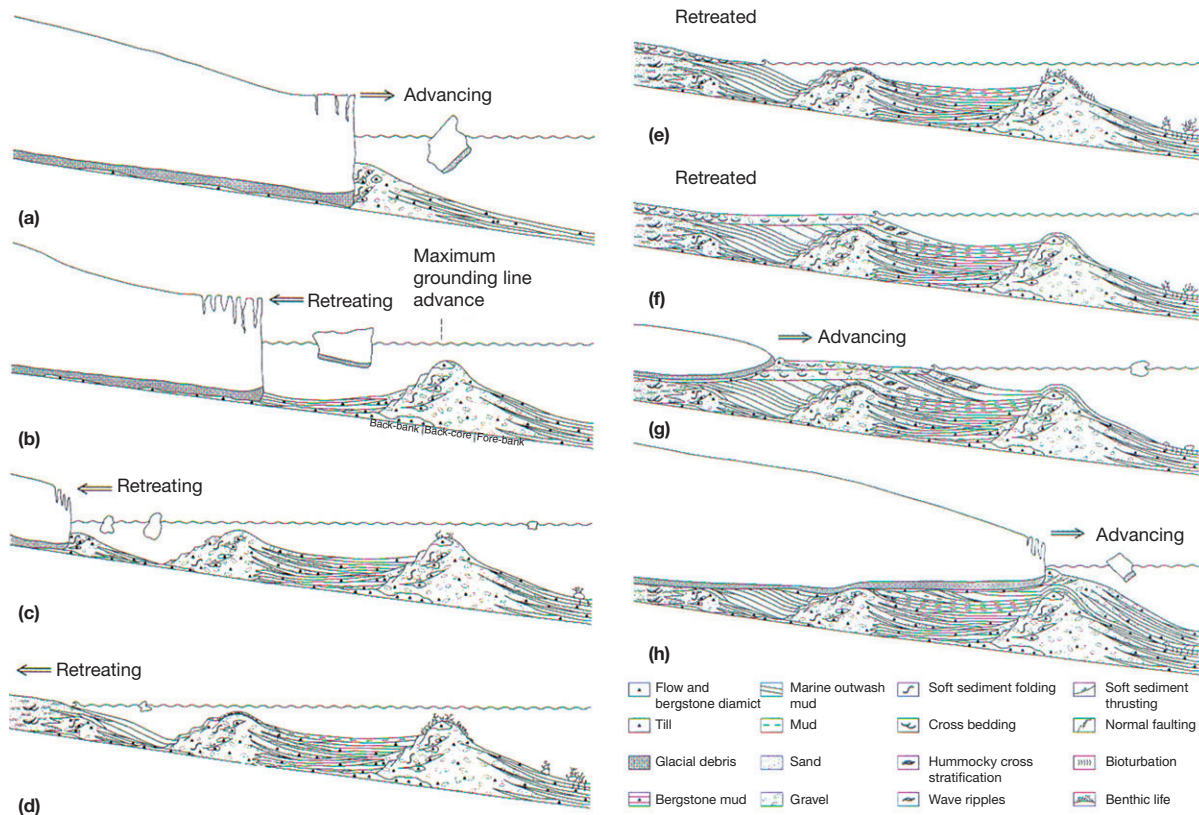


Figure 7 Model of typical facies associations and architecture in glacialmarine environments. The model depicts a glacial advance–retreat–readvance cycle. Reproduced from Powell RD (2003) Subaqueous landsystems: Fjords. In: Evans DJA (ed.) *Glacial Landsystems*, pp. 313–347. London: Arnold, with permission.

1990; Eyles and Eyles, 1992; Hambrey, 1994). A large body of information is available on subaqueous depositional systems for hundreds of formerly glaciated sites, and many depositional models have been proposed (e.g., Ashley, 1995; Eyles et al., 1985; Fyfe, 1990; Gravenor et al., 1984; Molnia, 1983; Powell and Domack, 1995; Figure 7). It has become clear that, although the lateral and vertical relationships between sediment–landform associations such as morainal banks, grounding-line wedges, and distal drapes are very varied, they reflect the interaction of relatively few groups of variables. The most important are (1) the topography and tectonic setting of the basin; (2) the extent, configuration, and dynamics of glacier and floating ice and the position of associated depocenters; and (3) changes in water depth due to eustatic and isostatic

cycles and the life cycle of proglacial lakes (Boulton, 1990; Eyles and Eyles, 1992; Martini and Brookfield, 1995; Powell, 1984, 1990; Powell and Domack, 1995). Subaqueous depositional systems can be classified according to their position of formation relative to the ice front. Powell (1984) defined four zones as subglacial, ice proximal, ice shelf, and ice distal. The subglacial zone lies behind the glacier-grounding line and is reviewed above. Facies deposited in this zone form downglacier-thickening sheets of basal till and associated lenses of water-sorted material. The ice-proximal zone encompasses subaqueous environments adjacent to the glacier-grounding line. The thickness of sediment packages deposited in the ice-proximal zone is related to the glacier snout retreat rate; slow retreat or quasi-stable conditions will encourage the

accumulation of large volumes of sediment, whereas during rapid calving retreat, a large proportion of debris will be rafted away, restricting the amount of ice-proximal deposition.

Sediment associations include push and thrust moraines, grounding-line fans, morainal banks, deltas, DeGeer moraines, and grounding-line fans. Sediment gravity flows are common in the ice-proximal zone due to high sedimentation rates, ice push, iceberg calving, storm waves, tidally induced pressures below ice shelves, and morainal bank collapse. Subaqueous outwash and sediment flows both commonly form prograding beds dipping away from the glacier margin, known as clinoforms. The upper parts of such clinoforms may be erosionally truncated during glacier advance, forming a subhorizontal erosion surface overlain by basal till. The ice-shelf zone occurs in polar climates where ice below the pressure-melting point floats and produces glacier and sea-ice ice shelves. Deposition in this zone is dominated by undermelt and rainout, forming drapes of dropstone diamicton and mud. The ice-distal zone is dominated by ice-rafted debris and muds formed by the settling of suspended mineral grains and microfaunal remains. Rainout, current reworking, and gravitational resedimentation produce sequences of massive and stratified sediment tens to hundreds of meters thick and spanning several glacial cycles. The long-term preservation potential of such sequences is high, and they are well represented in the ancient geological record (Eyles, 1993).

Fjords

Glacial stratigraphic architecture in fjords is strongly influenced by topography. Overdeepened basins act as sediment sinks, whereas bedrock sills provide pinning points where ice-proximal depositional systems can accumulate. Additionally, the presence of steep flanking slopes encourages gravitational reworking of sediment and debris input from ice-free terrain. The sedimentary infill of fjord basins can be usefully divided into three broad units deposited during glacier advance, maximum, and retreat. The advance stage of a glacial cycle is typically represented by a basal till unit or an erosion surface. Till thickness generally increases down fjord toward the glacier limit, reflecting erosion, and downglacier transport of pre-existing sediment and bedrock fragments (Boulton, 1990). Beyond the glacier limit, extensive blankets of mud and diamicton may record distal glacimarine deposition during the advance phase, but they are typically buried by younger glacimarine sediments. Sediments relating to glacial maximum conditions are commonly located at pinning points where reduced calving rates encourage ice-margin stability (Greene, 1992; Warren and Hulton, 1990).

Typical sediment–landform associations include push or thrust moraines, grounding-line fans, morainal banks, and deltas (Powell, 2003). In low-relief fjord basins, lateral moraines may be preserved on the valley sides, but in high-relief settings, such moraines are generally rapidly destroyed by paraglacial reworking after glacier retreat. Lateral moraines with exceptionally low gradients indicate the former extent of floating ice shelves. Beyond the ice limit, blanket-like drapes of fine-grained sediments and diamictic muds record suspension sedimentation from turbid overflows and dumping from icebergs (Powell and Molnia, 1989). During glacier

retreat, depositional zones migrate upfjord, and progressively more distal facies are laid down on top of older units. Pauses in glacier retreat may be marked by push or thrust moraines, grounding-line fans, or morainal banks (Boulton, 1986). Substantial ice-marginal accumulations such as large morainal banks and deltas tend to be associated with topographic pinning points. Glacimarine sediments and landforms in fjords may be raised above sea level by isostatic uplift following deglaciation. The emergence of ice-proximal associations above sea level provides a readily accessible source of information on former ice margins in regions where terrestrial ice-marginal accumulations are commonly not well preserved.

Continental Shelves

The glaciation of shelf areas may arise through the advance of terrestrial margins in a seaward direction, the thickening and grounding of sea ice encouraged by lower sea level, or a combination of these processes. The impact of glaciation on a continental shelf may be direct, where the shelf has hosted a grounded glacier mass, or indirect, where it merely received distal glacimarine sediments during a glacial cycle. Patterns of sedimentation are dependent on a number of interrelated factors, including the extent and configuration of glacier ice, basin history and tectonic setting, bathymetry, oceanography, glacial thermal regime, and sediment supply (e.g., Eyles, 1993; Powell, 1984). Powell (1984) considered the influence of ice-margin configuration and thermal regime on patterns of sedimentation on continental shelves and identified various glacimarine regimes on the basis of the nature of the ice source (valley glacier or ice sheet), glacier thermal regime, and glacier front characteristics (tidewater cliff or ice shelf). Tectonic setting also exerts a strong influence on the style and preservation of glaciogenic sediments on continental shelves (Eyles, 1993; Hambrey, 1994) and includes passive margins, convergent margins, and rift basins (e.g., Eyles et al., 1991; Powell and Domack, 1995; Vorren, 2003).

Large Proglacial Lakes and Epicontinental Seas

During deglaciation of the Pleistocene ice sheets, ice margins were commonly in contact with large water bodies, either proglacial lakes ponded between the ice and topographic barriers, or arms of the sea that flooded isostatically depressed and glacially overdeepened parts of continental interiors (Teller, 2003). The extent and evolution of many of these water bodies has been reconstructed from the sedimentary record, in combination with shoreline evidence, pollen and microfaunal records, and dating programs (e.g., Eronen, 1983; Teller, 1995). These large proglacial water bodies encouraged rapid ice-margin retreat by calving and may have triggered some episodes of ice advance or surges by encouraging low effective pressures below the glacier margin (Fyfe, 1990; Stokes and Clark, 2003). Depositional systems in such settings are similar to those of continental shelves (Figure 7), although there are some important differences. First, fresh lake water and brackish shallow marine conditions reduce the importance of suspended-sediment transport in overflows and interflows, so that underflow deposits are better represented in ice-proximal and distal systems. Second, the seasonal alternation between

deposition by underflows in summer and settling of clay particles in winter produces annual varves in lakes. Third, proglacial water bodies are prone to sudden changes in water level, due to the opening and closing of outlets by glacier margin fluctuations and periodic jokulhlaups and recharge events. The resulting cyclic water-level changes can be recorded by discrete sediment packages bounded above and below by erosion surfaces, and can add considerable complexity to stratigraphic sequences (Brookfield and Martini, 1999; Martini and Brookfield, 1995). Fourth, proglacial lakes and epicontinental seas can receive substantial amounts of sediment from recently deglaciated land surfaces surrounding the basin as well as from the glacier margin (Martini and Brookfield, 1995).

Sediment–Landform Associations of Mountain Glaciation and Glaciated Valleys

The glacial landforms and sediments of mountain environments have been synthesized for a variety of physiographic and latitudinal zones ever since the concept of a glaciated valley landsystem was introduced by Boulton and Eyles (1979) and Eyles (1983c). A variety of depositional settings has been elucidated by Benn et al. (2003) in a model that incorporates the controls of debris supply on landform development. A plateau icefield landsystem has been proposed by Rea et al. (1998) and Rea and Evans (2003) based upon observations on Norwegian and Icelandic ice caps. Glaciated valleys are unique in that they are influenced by valley-side debris inputs and the topographic confinement of depositional basins. The influence of these factors varies from valley to valley, and it is useful to subdivide glaciated valley landsystems into low-relief and high-relief types. Low-relief settings include valleys in ancient mountain belts where the vertical distance between valley floor and summit ridges is generally <1000 m (e.g., the glaciated valleys of Scotland, Norway, and Labrador). High-relief settings are characterized by extensive steep valley sides rising thousands of meters above the valley floor. Such environments are commonly associated with young or tectonically active fold mountains such as the European and New Zealand Alps, the High Andes, and the Himalayas, where relative relief may be 3000 m or more. Features common to both low- and high-relief mountainous terrains are ice-dammed lakes and paraglacially reworked materials.

Low-Relief Mountain Environments

Valley glaciers in low-relief mountain terrain carry relatively small amounts of supraglacial debris, so subglacial landforms commonly escape burial during deglaciation and may be very well exposed at the surface (Figure 8). The upper limits of glacier occupancy in glaciated valleys can be preserved as erosional ‘tidemarks,’ or trimlines, on valley sides. In recently deglaciated terrain, trimlines are often marked by differences in vegetation cover on either side of the former glacier limit, whereas in older glacial landscapes, periglacial trimlines may be preserved where frost-shattered terrain on upper slopes gives way abruptly to ice-scoured bedrock on lower ground (Ballantyne and McCarroll, 1995; Thorp, 1981, 1986). The margins of valley glaciers in low-relief mountain areas are

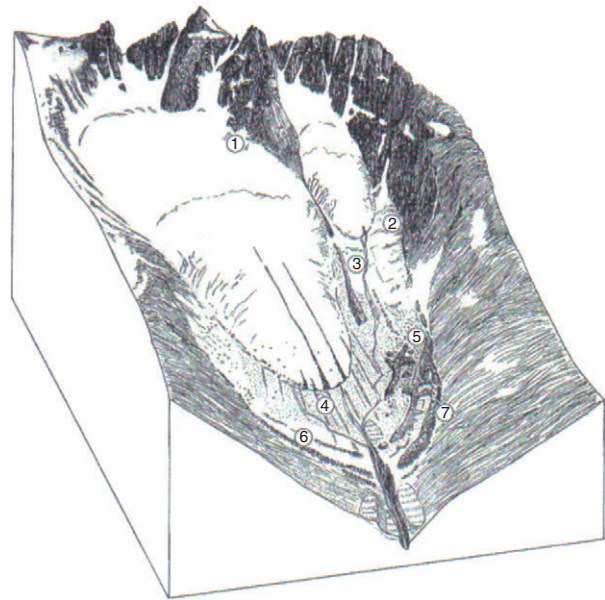


Figure 8 A landsystem model for deposition by valley glaciers with relatively low supraglacial debris concentrations. 1, Supraglacially entrained debris; 2, periglacial trimline; 3, medial moraine; 4, fluted till; 5, paraglacial reworking; 6 and 7, lateral moraine displaying within-valley asymmetry. Reproduced with permission from Benn DI and Evans DJA (1998) *Glaciers and Glaciation*. London: Arnold.

commonly delimited by lateral–frontal dump and push moraines (Benn and Lukas, 2006). Moraine characteristics are strongly influenced by catchment lithology, and in valleys underlain by resistant, crystalline rocks, lateral moraines may be little more than ridges of boulders. In contrast, thrust moraines may occur where glaciers come into contact with thick, unconsolidated sediments such as glacial marine clays and silts. Meltwater deposition at some glacier margins is recorded by ice-marginal ramps and fans (e.g., Lukas, 2005). Lateral moraines may be much larger on one valley side than the other, a phenomenon known as ‘within-valley asymmetry’ due to one or a combination of four factors: (1) the differential extent of extensive, debris-supplying rock walls, (2) glacier pushing of preexisting materials that are unevenly spread over the foreland, (3) cross-valley differences in lithology or structure, and (4) differences in glacier dynamics on either side of a valley (Benn, 1989; Bennett and Boulton, 1993; Evans, 1999; Matthews and Petch, 1982).

The combined use of lateral moraines and other ice-marginal landforms and trimlines allows the dimensions of former valley glaciers and their equilibrium-line altitudes (paleo-ELAs) to be reconstructed with considerable accuracy, yielding important input for glaciological models or paleoclimatic studies (Ballantyne, 1989; Nesje and Dahl, 2000; Thorp, 1991). Glacier retreat in low-relief mountain environments may be recorded by recessional moraines and/or linear hummocky moraine superimposed on subglacial sediments and landforms. The size and spacing of such moraines reflect the availability of debris and the frequency and duration of still stands or readvances of the margin (Benn, 1992; Benn and Lukas, 2006; Bennett and Boulton, 1993; Eyles, 1979). Assemblages of chaotic hummocky moraine may occur in pockets



Figure 9 Lateral moraines and ice-contact fans deposited at the margin of a debris-charged valley glacier in the Lahul Himalaya, India.

within fields of recessional moraines, possibly due to the isolation and stagnation of sediment-covered ice remnants (Benn, 1992). Large areas of hummocky moraine can be produced by incremental marginal stagnation, whereby chaotic moraines are deposited in a 10–500 m wide marginal zone of debris-mantled ice which retreats upvalley during deglaciation (Eyles, 1983a,c). Assemblages indicative of more general stagnation, such as chaotic hummocky moraine inset within ‘staircases’ of subhorizontal kame terraces, occur on some flat-lying valley floors (Benn, 1992).

High-Relief Mountain Environments

On account of their high sediment yields, high-relief mountain environments contain valley floors that are commonly filled by very large volumes of sediment, particularly in tectonically active ranges such as the Southern Alps of New Zealand, the Himalayas, and the Andes (Clapperton, 1993; Kirkbride, 1995; Owen and Derbyshire, 1989, 1993). Valley glaciers in such settings typically have extensive covers of supraglacial debris, and sedimentation around glacier margins forms large lateral-frontal dump moraines and ice-contact fans and ramps (Figure 9; Benn et al., 2003; Boulton and Eyles, 1979; Evans et al., 2010; Hambrey et al., 2008; Humlum, 1978; Owen and Derbyshire, 1989, 1993; Shulmeister et al., 2009; Small, 1983).

Debris-mantled glaciers (Nakawo et al., 2000) have relatively muted responses to climatic change, because large lateral-frontal moraines limit the amount of glacier advance, and extensive supraglacial debris cover inhibits ablation of the underlying ice. Therefore, unlike their low-relief counterparts, valley glaciers in high-relief terrain tend not to deposit numerous dump and push moraines but instead remain at the limits imposed by lateral-frontal moraines until advance or retreat is triggered by significant climatic change; so, the landform record of a retreating debris-mantled glacier consists of major moraine complexes, representing periods of stability, separated by extensive tracts of hummocky moraine deposited during episodes of ice-margin wastage and stagnation. The vast latero-frontal moraine complexes produced during ice wastage commonly form dams for proglacial and supraglacial lakes, giving rise to periodic glacial lake outburst floods when dams are breached by lake water (Cenderelli and Wohl, 2003; Clague and Evans, 1994). The thick supraglacial and ice-marginal sediments of valley glaciers in high-relief terrain largely mask subglacial landforms and sediment so that the former beds of such glaciers are rarely preserved in the geomorphological record.

See also: Varved Lake Sediments. **Glacial Landforms:** Evidence of Glacier Recession; Glacitectonic Structures and Landforms; Glacial Landsystems; Glaciofluvial Landforms of Erosion; Moraine Forms and Genesis; Quaternary Vulcanism: Subglacial Landforms. **Glacial Landforms, Erosional Features:** Major Scale Forms; Micro- to Macroscale Forms. **Glacial Landforms, Ice Sheets:** Evidence of Glacier and Ice Sheet Extent; Growth and Decay; Paleo-ELAs; Trimlines and Paleonunataks. **Glacial Landforms, Sediments:** Clast Form Analysis; Glaciofluvial Landforms of Deposition; Glaciogenic Lithofacies; Glaciolacustrine; Glaciomarine Sediments and Ice-Rafted Debris; Micromorphology of Glacial Sediments; Till Fabric Analysis; Tills. **Permafrost and Periglacial Features:** Paraglacial Geomorphology; Permafrost.

References

- Alley RB, Lawson DE, Evenson EB, Strasser JC, and Larson GJ (1998) Glaciohydraulic supercooling: A freeze-on mechanism to create stratified, debris-rich basal ice. II. Theory. *Journal of Glaciology* 44: 563–569.
- Ashley GM (1995) Glaciolacustrine environments. In: Menzies J (ed.) *Glacial Environments, Vol. 1: Modern Glacial Environments: Processes, Dynamics and Sediments*, pp. 417–444. Oxford: Butterworth-Heinemann.
- Aylsworth JM and Shilts WW (1989) Bedforms of the Keewatin Ice Sheet, Canada. *Sedimentary Geology* 62: 407–428.
- Ballantyne CK (1989) The Loch Lomond Readvance on the Isle of Skye, Scotland: Glacier reconstruction and palaeoclimatic implications. *Journal of Quaternary Science* 4: 95–108.
- Ballantyne CK and McCarroll D (1995) The vertical dimensions of Late Devensian glaciation on the mountains of Harris and southeast Lewis, Outer Hebrides, Scotland. *Journal of Quaternary Science* 10: 211–223.
- Benn DI (1989) Debris transport by Loch Lomond Readvance glaciers in northern Scotland, basin form and the within-valley asymmetry of lateral moraines. *Journal of Quaternary Science* 4: 243–254.
- Benn DI (1992) The genesis and significance of ‘hummocky moraine’: Evidence from the Isle of Skye, Scotland. *Quaternary Science Reviews* 11: 781–799.
- Benn DI and Evans DJA (2010) *Glaciers and Glaciation*. London: Hodder Education.
- Benn DI, Kirkbride MP, Owen LA, and Brazier V (2003) Glaciated valley landsystems. In: Evans DJA (ed.) *Glacial Landsystems*, pp. 372–406. London: Arnold.

- Benn DI and Lukas S (2006) Younger Dryas glacial landsystems in North West Scotland: An assessment of modern analogues and palaeoclimatic implications. *Quaternary Science Reviews* 25: 2390–2408.
- Bennett MR and Boulton GS (1993) A reinterpretation of Scottish 'hummocky moraine' and its significance for the deglaciation of the Scottish highlands during the Younger Dryas or Loch Lomond Stadial. *Geological Magazine* 130: 301–318.
- Berthelsen A (1978) The methodology of kineto-stratigraphy as applied to glacial geology. *Bulletin of the Geological Society of Denmark* 27: 25–38.
- Bluemle JP and Clayton L (1984) Large-scale glacial thrusting and related processes in North Dakota. *Boreas* 13: 279–299.
- Boston CM, Evans DJA, and Ó Cofaigh C (2010) Styles of till deposition at the margin of the Last Glacial Maximum North Sea lobe of the British-Irish Ice Sheet: An assessment based on geochemical properties of glacial deposits in eastern England. *Quaternary Science Reviews* 29: 3184–3211.
- Bouchard MA (1989) Subglacial landforms and deposits in central and northern Quebec, Canada, with emphasis on Rogen moraines. *Sedimentary Geology* 62: 293–308.
- Boulton GS (1972) The role of the thermal regime in glacial sedimentation. In: Price RJ and Sugden DE (eds.) *Polar Geomorphology*, vol. 4, pp. 1–19. London: Institute of British Geographers Special Publication.
- Boulton GS (1986) Push moraines and glacier contact fans in marine and terrestrial environments. *Sedimentology* 33: 677–698.
- Boulton GS (1987) A theory of drumlin formation by subglacial sediment deformation. In: Menzies J and Rose J (eds.) *Drumlin Symposium*, pp. 25–80. Rotterdam: Balkema.
- Boulton GS (1990) Sedimentary and sea level changes during glacial cycles and their control on glacial facies architecture. In: Dowdeswell JA and Scourse JD (eds.) *Glacimarine Environments: Processes and Sediments*, vol. 53, pp. 15–52. London: Geological Society Special Publication.
- Boulton GS (1996a) Theory of glacial erosion, transport and deposition as a consequence of subglacial sediment deformation. *Journal of Glaciology* 42: 43–62.
- Boulton GS (1996b) The origin of till sequences by subglacial sediment deformation beneath mid-latitude ice sheets. *Annals of Glaciology* 22: 75–84.
- Boulton GS and Clark CD (1990) The Laurentide Ice sheet through the last glacial cycle: Drift lineations as a key to the dynamic behaviour of former ice sheets. *Transactions of the Royal Society of Edinburgh – Earth Sciences* 81: 327–347.
- Boulton GS and Eyles N (1979) Sedimentation by valley glaciers: A model and genetic classification. In: Schluchter C (ed.) *Moraines and Varves*, pp. 11–23. Rotterdam: Balkema.
- Boulton GS, Smith GD, Jones AS, and Newsome J (1985) Glacial geology and glaciology of the last mid-latitude ice sheets. *Journal of the Geological Society of London* 142: 447–474.
- Boyce JI and Eyles N (1991) Drumlins carved by deforming till streams below the Laurentide ice sheet. *Geology* 19: 787–790.
- Boyce JI and Eyles N (2000) Architectural element analysis applied to glacial deposits: Internal geometry of a Late Pleistocene till sheet, Ontario, Canada. *Geological Society of America Bulletin* 112: 98–118.
- Brookfield ME and Martini IP (1999) Facies architecture and sequence stratigraphy in glacially influenced basins: Basic problems and water level/glacier input-point controls (with an example from the Quaternary of Ontario, Canada). *Sedimentary Geology* 123: 183–197.
- Cenderelli DA and Wohl EE (2003) Flow hydraulics and geomorphic effects of glacial-lake outburst floods in the Mount Everest region, Nepal. *Earth Surface Processes and Landforms* 28: 385–407.
- Clague JJ and Evans SG (1994) Formation and failure of natural dams in the Canadian Cordillera. *Geological Survey of Canada Bulletin* 464: 35.
- Clapperton CM (1993) *Quaternary Geology and Geomorphology of South America*. Amsterdam: Elsevier.
- Clark CD (1993) Mega-scale lineations and cross-cutting ice-flow landforms. *Earth Surface Processes and Landforms* 18: 1–29.
- Clark CD (1999) Glaciodynamic context of subglacial bedform generation and preservation. *Annals of Glaciology* 28: 23–32.
- Clark CD and Stokes CR (2003) Palaeo-ice stream landsystem. In: Evans DJA (ed.) *Glacial Landsystems*, pp. 204–227. London: Arnold.
- Clark PU and Walder JS (1994) Subglacial drainage, eskers, and deforming beds beneath the Laurentide and Eurasian ice sheets. *Bulletin of the Geological Society of America* 106: 304–314.
- Clayton L (1964) Karst topography on stagnant glaciers. *Journal of Glaciology* 5: 107–112.
- Clayton L and Moran SR (1974) A glacial process-form model. In: Coates DR (ed.) *Glacial Geomorphology*, pp. 89–119. Binghamton: State University of New York.
- Dunlop P and Clark CD (2006) The morphological characteristics of ribbed moraine. *Quaternary Science Reviews* 25: 1668–1691.
- Dunlop P, Clark CD, and Hindmarsh RCA (2008) The Bed Ribbing Instability Explanation (BRIE): Testing a numerical model of ribbed moraine formation arising from coupled flow of ice and subglacial sediment. *Journal of Geophysical Research* 113: F03005. <http://dx.doi.org/10.1029/2007JF000954>.
- Dyke AS (1993) Landscapes of cold-centred Late Wisconsinan ice caps, Arctic Canada. *Progress in Physical Geography* 17: 223–247.
- Dyke AS and Evans DJA (2003) Ice-marginal terrestrial landsystems: Northern Laurentide and Inuitian ice sheet margins. In: Evans DJA (ed.) *Glacial Landsystems*, pp. 143–165. London: Arnold.
- Dyke AS and Prest VK (1987) Late Wisconsinan and Holocene history of the Laurentide Ice Sheet. *Geographie Physique et Quaternaire* XLI: 237–263.
- England J, Bradley RS, and Miller GH (1978) Former ice shelves in the Canadian high Arctic. *Journal of Glaciology* 20: 393–404.
- Eronen M (1983) Late Weichselian and Holocene shore displacement in Finland. In: Smith DE and Dawson AG (eds.) *Shorelines and Isostasy*, pp. 183–208. London: Academic Press.
- Etzelmüller B and Hagen JO (2005) Glacier-permafrost interaction in arctic and alpine mountain environments, with examples from southern Norway and Svalbard. In: Harris C and Murtin JB (eds.) *Cryospheric Systems: Glaciers and Permafrost*, vol. 242, pp. 11–27. London: Geological Society Special Publication.
- Etzelmüller B, Hagen JO, Vatne G, Ødegard RS, and Sollid JL (1996) Glacier debris accumulation and sediment deformation influenced by permafrost: Examples from Svalbard. *Annals of Glaciology* 22: 53–62.
- Evans DJA (1989a) The nature of glaciectonic structures and sediments at subpolar glacier margins, northwest Ellesmere Island, Canada. *Geografiska Annaler* 71A: 113–123.
- Evans DJA (1989b) Apron entrainment at the margins of subpolar glaciers, northwest Ellesmere Island, Canadian high arctic. *Journal of Glaciology* 35: 317–324.
- Evans DJA (1990) The effect of glacier morphology on surficial geology and glacial stratigraphy in a high arctic mountainous terrain. *Zeitschrift für Geomorphologie* 34: 481–503.
- Evans DJA (1999) Glacial debris transport and moraine deposition: A case study of the Jardalen cirque complex, Sogn-og-Fjordane, western Norway. *Zeitschrift für Geomorphologie* 43: 203–234.
- Evans DJA (2009) Controlled moraines: Origins, characteristics and palaeoglaciological implications. *Quaternary Science Reviews* 28: 183–208.
- Evans DJA, Clark CD, and Rea BR (2008) Landform and sediment imprints of fast glacier flow in the southwest Laurentide Ice Sheet. *Journal of Quaternary Science* 23: 249–272.
- Evans DJA and England J (1991) Canadian landform examples. 19. High arctic thrust block moraines. *The Canadian Geographer* 35: 93–97.
- Evans DJA and Hiemstra JF (2005) Till deposition by glacier sub-marginal, incremental thickening. *Earth Surface Processes and Landforms* 30: 1633–1662.
- Evans DJA, Lemmen DS, and Rea BR (1999) Glacial landsystems of the southwest Laurentide Ice Sheet: Modern Icelandic analogues. *Journal of Quaternary Science* 14: 673–691.
- Evans DJA, Shand M, and Petrie G (2009) Maps of the snout and proglacial landforms of Fjallsjökull, Iceland (1945, 1965, 1998). *Scottish Geographical Journal* 125: 304–320.
- Evans DJA, Shulmeister J, and Hyatt O (2010) Sedimentology of latero-frontal moraines and fans on the west coast of South Island, New Zealand. *Quaternary Science Reviews* 29: 3790–3811.
- Evans DJA and Twigg DR (2002) The active temperate glacial landsystem: A model based on Breidamerkurjökull and Fjallsjökull, Iceland. *Quaternary Science Reviews* 21: 2143–2177.
- Eyles N (1979) Facies of supraglacial sedimentation on Icelandic and alpine temperate glaciers. *Canadian Journal of Earth Sciences* 16: 1341–1361.
- Eyles N (1983a) Modern Icelandic glaciers as depositional models for 'hummocky moraine' in the Scottish highlands. In: Evenson EB, Schluchter C, and Rabassa J (eds.) *Tills and Related Deposits*, pp. 47–59. Rotterdam: Balkema.
- Eyles N (1983b) Glacial geology: A landsystems approach. In: Eyles N (ed.) *Glacial Geology*, pp. 1–18. Oxford: Pergamon.
- Eyles N (1983c) The glaciated valley landsystem. In: Eyles N (ed.) *Glacial Geology*, pp. 91–110. Oxford: Pergamon.
- Eyles N (1993) Earth's glacial record and its tectonic setting. *Earth-Science Reviews* 35: 1–248.
- Eyles N and Eyles CH (1992) Glacial depositional systems. In: Walker RG and James NP (eds.) *Facies Models: Response to Sea-level Change*, pp. 73–100. Toronto: Geological Association of Canada.
- Eyles CH, Eyles N, and Lago MB (1991) The Yakataga Formation: A late Miocene to Pleistocene record of temperate glacial marine sedimentation in the Gulf of Alaska. In: Anderson JB and Ashley GM (eds.) *Glacial Marine Sedimentation*:

- Palaeoclimatic Significance*, vol. 261, pp. 159–180. Boulder, CO: Geological Society of America Special Paper.
- Eyles CH, Eyles N, and Miall AD (1985) Models of glaciomarine sedimentation and their application to the interpretation of ancient glacial sequences. *Palaeogeography, Palaeoclimatology, Palaeoecology* 51: 15–84.
- Eyles N and Menzies J (1983) The subglacial landsystem. In: Eyles N (ed.) *Glacial Geology*, pp. 19–70. Oxford: Pergamon.
- Eyles N, Sladen JA, and Gilroy S (1982) A depositional model for stratigraphic complexes and facies superimposition in lodgement tills. *Boreas* 11: 317–333.
- Fitzsimons SJ (1997) Depositional models for moraines in East Antarctic coastal oases. *Journal of Glaciology* 43: 256–264.
- Fitzsimons SJ (2003) Ice-marginal terrestrial landsystems: Polar continental glacier margins. In: Evans DJA (ed.) *Glacial Landsystems*, pp. 89–110. London: Arnold.
- Fyfe GJ (1990) The effect of water depth on ice-proximal glaciolacustrine sedimentation: Salpausselka. I. Southern Finland. *Boreas* 19: 147–164.
- Goldthwait RP (1951) Development of end moraines in east-central Baffin Island. *Journal of Geology* 59: 567–577.
- Gravenor CP, von Brunn V, and Dreimanis A (1984) Nature and classification of waterlain glaciogenic sediments, exemplified by Pleistocene, late Palaeozoic and late Precambrian deposits. *Earth-Science Reviews* 20: 105–166.
- Greene D (1992) Topography and former Scottish tidewater glaciers. *Scottish Geographical Magazine* 108: 164–171.
- Gustavson TC and Boothroyd JC (1987) A depositional model for outwash, sediment sources, and hydrologic characteristics, Malaspina Glacier, Alaska: A modern analog of the southeastern margin of the Laurentide Ice Sheet. *Geological Society of America Bulletin* 99: 187–200.
- Hambrey MJ (1994) *Glacial Environments*, p. 296. London: UCL Press.
- Hambrey MJ and Huddart D (1995) Englacial and proglacial glaciotectonic processes at the snout of a thermally complex glacier in Svalbard. *Journal of Quaternary Science* 10: 313–326.
- Hambrey MJ, Quincey DJ, Glasser NF, Reynolds JM, Richardson SJ, and Clemmens S (2008) Sedimentological, geomorphological and dynamic context of debris-mantled glaciers, Mount Everest (Sagarmatha) region, Nepal. *Quaternary Science Reviews* 27: 2361–2389.
- Hicock SR (1992) Lobal interactions and rheologic superposition in subglacial till near Bradville, Ontario, Canada. *Boreas* 21: 73–88.
- Hodgson DA (1994) Episodic ice streams and ice shelves during retreat of the northwesternmost sector of the late Wisconsinan Laurentide Ice Sheet over the central Canadian Arctic Archipelago. *Boreas* 23: 14–28.
- Hodgson DA and Vincent J-S (1984) A 10,000 yr BP extensive ice shelf over Viscount Melville Sound, Arctic Canada. *Quaternary Research* 22: 18–30.
- Hooke RLeB (1973) Flow near the margin of the Barnes Ice Cap, and the development of ice-cored moraines. *Geological Society of America Bulletin* 84: 3929–3948.
- Hubbard B (1991) Freezing-rate effects on the physical characteristics of basal ice formed by net adfreezing. *Journal of Glaciology* 37: 339–347.
- Humlum O (1978) Genesis of layered lateral moraines: Implications for palaeoclimatology and lichenometry. *Geografisk Tidsskrift* 77: 65–72.
- Humlum O (1985) Genesis of an imbricate push moraine, Hofdabrekkujökull, Iceland. *Journal of Geology* 93: 185–195.
- Jansson KN, Stroeve AP, and Kleman J (2003) Configuration and timing of Ungava Bay ice streams, Labrador–Ungava, Canada. *Boreas* 32: 256–262.
- Kalin M (1971) The active push moraine of the Thompson Glacier, Axel Heiberg Island, Canadian Arctic Archipelago, McGill University, Axel Heiberg Island, Research Report, Glaciology 4, McGill University, Montreal, p. 68.
- King EC, Hindmarsh RCA, and Stokes CR (2009) Formation of megascale glacial lineations observed beneath a West Antarctic ice stream. *Nature Geoscience* 2: 585–596.
- King EC, Woodward J, and Smith AM (2007) Seismic and radar observations of subglacial bedforms beneath the onset zone of Rutford Ice Stream, Antarctica. *Journal of Glaciology* 53: 665–672.
- Kirkbride MP (1995) Processes of transportation. In: Menzies J (ed.) *Glacial Environments, Vol. 1: Modern Glacial Environments: Processes, Dynamics and Sediments*, pp. 261–292. Oxford: Butterworth-Heinemann.
- Kleman J and Glasser NF (2007) The subglacial thermal organization (STO) of ice sheets. *Quaternary Science Reviews* 26: 585–597.
- Kleman J, Hättestrand C, Stroeve A, Jansson KJ, De Angelis H, and Borgström I (2006) Reconstruction of palaeo-ice sheets: Inversion of their glacial geomorphological record. In: Knight P (ed.) *Glaciology and Earth's Changing Environment*, pp. 192–198. Oxford: Blackwell.
- Kleman J, Stroeve AP, and Lunqvist J (2008) Patterns of Quaternary ice sheet erosion and deposition in Fennoscandia and a theoretical framework for explanation. *Geomorphology* 97: 73–90.
- Knight PG (1989) Stacking, of basal debris layers without bulk freezing on: Isotopic evidence from West Greenland. *Journal of Glaciology* 35: 214–216.
- Koteff C and Pessl F (1981) Systematic ice retreat in New England. *United States Geographical Survey Professional Paper* 1179: 20.
- Kruger J (1985) Formation of a push moraine at the margin of Hofdabrekkujökull, south Iceland. *Geografiska Annaler* 67A: 199–212.
- Kruger J (1993) Moraine-ridge formation along a stationary ice front in Iceland. *Boreas* 22: 101–109.
- Kruger J (1994) Glacial processes, sediments, landforms, and stratigraphy in the terminus region of Myrdalsjökull, Iceland. *Folia Geographica Danica* 21: 1–233.
- Larson GJ, Lawson DE, Evenson EB, Knudsen O, Alley RB, and Phanikumar MS (2010) Origin of stratified basal ice in outlet glaciers of Vatnajökull and Öræfajökull, Iceland. *Boreas* 39: 457–470.
- Lawson DE (1979) Sedimentological analysis of the western terminus region of the Matanuska Glacier, Alaska. Hanover: Cold Regions Research and Engineering Laboratory Report 79-9.
- Lawson DE, Strasser JC, Evenson EB, Alley RB, Larson GJ, and Arcone SA (1998) Glaciohydraulic supercooling: A freeze-on mechanism to create stratified, debris-rich basal ice. I. Field evidence. *Journal of Glaciology* 44: 547–562.
- Lukas S (2005) A test of the englacial thrusting hypothesis of 'hummocky' moraine formation: Case studies from the north-west Highlands, Scotland. *Boreas* 34: 287–307.
- Marren PM (2002) Glacier margin fluctuations, Skattafellsjökull, Iceland: Implications for sandur evolution. *Boreas* 31: 75–81.
- Martini IP and Brookfield ME (1995) Sequence analysis of upper Pleistocene (Wisconsinan) glaciolacustrine deposits of the north-shore bluffs of Lake Ontario, Canada. *Journal of Sedimentary Research* B65: 388–400.
- Matthews JA and Petch JR (1982) Within-valley asymmetry and related problems of Neoglacial lateral moraine development at certain Jotunheimen glaciers, southern Norway. *Boreas* 11: 225–247.
- Molnia BF (1983) Subarctic glacial-marine sedimentation: A model. In: Molnia BF (ed.) *Glacial-Marine Sedimentation*, pp. 95–144. New York: Plenum.
- Nakawo M, Raymond CF, and Fountain A (eds.) (2000) *Debris-Covered Glaciers*. Wallingford: International Association of Hydrological Sciences.
- Nesje A and Dahl S-O (2000) *Glaciers and Environmental Change*. London: Arnold.
- Owen LA and Derbyshire E (1989) The Karakoram glacial depositional system. *Zeitschrift für Geomorphologie Supplementband* 76: 33–73.
- Owen LA and Derbyshire E (1993) Quaternary and Holocene intermontane basin sedimentation in the Karakoram Mountains. In: Shroder JF (ed.) *Himalaya to the Sea*, pp. 108–131. London: Routledge.
- Powell RD (1984) Glaciomarine processes and inductive lithofacies modelling of ice shelf and tidewater glacier sediments based on Quaternary examples. *Marine Geology* 57: 1–52.
- Powell RD (1990) Glaciomarine processes at grounding-line fans and their growth to ice-contact deltas. In: Dowdeswell JA and Scourse JD (eds.) *Glaciomarine Environments: Processes and Sediments*, vol. 53, pp. 53–73. London: Geological Society Special Publication.
- Powell RD (2003) Subaquatic landsystems: Fjords. In: Evans DJA (ed.) *Glacial Landsystems*, pp. 313–347. London: Arnold.
- Powell RD and Domack E (1995) Modern glaciomarine environments. In: Menzies J (ed.) *Glacial Environments, Vol. 1: Modern Glacial Environments: Processes, Dynamics and Sediments*, pp. 445–486. Oxford: Butterworth-Heinemann.
- Powell RD and Molnia BF (1989) Glaciomarine sedimentary processes, facies and morphology of the south-southeast Alaska shelf and fjords. *Marine Geology* 85: 359–390.
- Price RJ (1969) Moraines, sandar, kames and eskers near Breidamerkurjökull, Iceland. *Transactions of the Institute of British Geographers* 46: 17–43.
- Punkari M (1995) Glacial flow systems in the zone of confluence between the Scandinavian and Novaya Zemlya ice sheets. *Quaternary Science Reviews* 14: 589–603.
- Rea BR and Evans DJA (2003) Plateau icefield landsystems. In: Evans DJA (ed.) *Glacial Landsystems*, pp. 407–431. London: Arnold.
- Rea BR, Whalley WB, Evans DJA, Gordon JE, and McDougall DA (1998) Plateau icefields: Geomorphology and dynamics. *Quaternary Proceedings* 6: 35–54.
- Sharp MJ (1984) Annual moraine ridges at Skalfellsjökull, south-east Iceland. *Journal of Glaciology* 30: 82–93.
- Shaw J (1977a) Till body morphology and structure related to glacier flow. *Boreas* 6: 189–201.
- Shaw J (1977b) Tills deposited in arid polar environments. *Canadian Journal of Earth Sciences* 14: 1239–1245.

- Shilts WW, Aylsworth JM, Kaszycki CA, and Klassen RA (1987) Canadian Shield. In: Graf WL (ed.) *Geomorphic Systems of North America*, vol. 2, pp. 119–161. Boulder, CO: Geological Society of America Centennial Special.
- Shulmeister J, Davies TRH, Evans DJA, Hyatt OM, and Tovar DS (2009) Catastrophic landslides, glacier behaviour and moraine formation – A view from an active plate margin. *Quaternary Science Reviews* 28: 1085–1096.
- Small RJ (1983) Lateral moraines of Glacier De Tsidjore Nouve: Form, development and implications. *Journal of Glaciology* 29: 250–259.
- Stokes CR and Clark CD (2003) The Dubawnt Lake palaeo-ice stream: Evidence for dynamic ice sheet behavior on the Canadian Shield and insights regarding the controls on ice stream location and vigour. *Boreas* 32: 263–279.
- Stokes CR, Clark CD, Darby DA, and Hodgson DA (2005) Late Pleistocene ice export events into the Arctic Ocean from the M'Clure Strait ice stream, Canadian Arctic Archipelago. *Global and Planetary Change* 49: 139–162.
- Sugden DE and Clapperton CM (1981) An ice-shelf moraine, George VI sound, Antarctica. *Annals of Glaciology* 2: 135–141.
- Teller JT (1995) History and drainage of large ice-dammed lakes along the Laurentide Ice Sheet. *Quaternary International* 28: 83–92.
- Teller JT (2003) Subaquatic landsystems: Large proglacial lakes. In: Evans DJA (ed.) *Glacial Landsystems*, pp. 348–371. London: Arnold.
- Thompson A (1988) Historical development of the proglacial landforms of Svinafellsjökull and Skafafellsjökull, southeast Iceland. *Jökull* 38: 17–31.
- Thorp PW (1981) A trimline method for defining the upper limit of Loch Lomond Readvance glaciers: Examples from the Loch Leven and Glencoe areas. *Scottish Journal of Geology* 17: 49–64.
- Thorp PW (1986) A mountain icefield of Loch Lomond Stadial age, western Grampians, Scotland. *Boreas* 15: 83–97.
- Thorp PW (1991) Surface profiles and basal shear stresses of outlet glaciers from a Lateglacial mountain icefield in western Scotland. *Journal of Glaciology* 37: 77–89.
- Tison J-L, Petit J-R, Barnola J-M, and Mahaney WC (1993) Debris entrainment at the ice–bedrock interface in sub-freezing temperatures, Terre Adélie, Antarctica. *Journal of Glaciology* 39: 303–315.
- Vorren TO (2003) Subaquatic landsystems: Continental margins. In: Evans DJA (ed.) *Glacial Landsystems*, pp. 289–312. London: Arnold.
- Warren CR and Hulton NRJ (1990) Topographic and glaciological controls on Holocene ice margin dynamics, central west Greenland. *Annals of Glaciology* 14: 307–310.
- Weertman J (1961) Mechanism for the formation of inner moraines found near the edge of cold ice caps and ice sheets. *Journal of Glaciology* 3: 965–978.

Moraine Forms and Genesis

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Introduction

Moraines can be generally described as accumulations of sediment/debris on the surface or at the margins of an ice sheet or glacier. They can be constructed at either terrestrial or subaqueous ice margins where internal structures and sedimentary characteristics reflect the processes involved in their emplacement. A range of large-scale morainic features constructed by glaciotectonic processes are excluded from this article because they are classified as glaciotectonic landforms and therefore described elsewhere. This separation of morainic forms is based upon the definitions proposed by Benn and Evans (2010), specifying that moraines are those in which glaciotectonized sediments or bedrock constitute >25% of the unit area of the moraine. Terrestrial moraines include push and squeeze moraines, dump moraines and ice-marginal aprons, and latero-frontal moraines and fans and ramps. Supraglacial debris accumulations form medial moraines and hummocky/controlled moraine. Subaqueous depocenters include morainal banks, coalescent subaqueous fans, and De Geer moraines, and the interface between subaqueous and terrestrial environments can be marked by ice-shelf moraines.

Push and Squeeze Moraines

Push moraines are small moraine ridges, usually <10 m in height, produced by minor (often annual) glacier advances in either subaqueous or terrestrial settings (e.g., Evans and Twigg 2002; Sharp, 1984; Figure 1). Other terms include washboard moraines, cross-valley moraines, and corrugated ground moraines. Although the term 'push moraine' has been employed in the past to refer to large-scale glaciotectonized moraine ridges, more specific terms are now available for such features and so they are not included here. Push moraines vary widely in composition and can consist of subglacial till, mass-movement deposits, water-sorted sediments, or large boulders, depending on the nature of the sediment on the glacier foreland. Where glacier margins are debris covered, push moraines can consist mainly of supraglacial material dumped onto the foreland during the summer and pushed up during winter. Push moraines commonly have asymmetric cross-profiles, with gentle proximal slopes, often covered with small flutings and steep distal slopes. Push moraines are broadly arcuate in planform, but in detail are often irregular and winding, reflecting the morphology of the glacier snout. For example, where the glacier snout is indented by radial crevasses, push moraines have sawtooth planforms, with downvalley-pointing teeth and upvalley-pointing notches (Evans and Twigg, 2002; Evans et al., 2009; Matthews et al., 1979; Price, 1970; Sharp, 1984; Figure 2). It is common to find younger moraines overlying, or completely consuming, parts of older moraines wherever summer retreat is small and/or some winter readvances are more substantial than usual

(Boulton, 1986; Evans and Twigg, 2002; Sharp, 1984). Where glacier snouts are quasistable or have reached a position where they are in equilibrium with their environment, large terminal moraines may accumulate by annual accretion. Similarly, the absence of moraines relating to individual years within a field of annual moraines can be explained by unusually large winter advances or longer-term readvances. These can result in ice advance that exceeds several years of annual retreat and so winter push moraines can be destroyed or incorporated into a larger moraine ridge (Boulton, 1986; Krüger, 1993, 1994). The construction of larger push moraine ridges has been explained in some settings as the result of stacking of till slabs due to submarginal freeze-on of subglacial sediment and its release by melt-out over a seasonal cycle (Evans and Hiemstra, 2005; Krüger, 1996; Matthews et al., 1995; Figure 3). The formation of minor moraine ridges by squeezing of saturated sediment from beneath ice margins has been observed in front of several glaciers whose snouts are predominantly underlain by poorly drained, fine-grained till (Price, 1970; Figure 4). Additionally, subaqueous squeeze moraines have been described by Andrews and Smithson (1966). Newly formed squeeze moraines have steep or vertical sides and are rarely more than 1 m high. Squeeze moraines tend to degrade rapidly and are commonly subject to reworking by ice push; therefore, unmodified squeeze moraines are probably rare in the geomorphological record.

Dump Moraines and Ice-Marginal Aprons

Material accumulating at the surface of a glacier through the melt-out of debris-rich folia is ultimately subject to remobilization by mass flowage, fall, or fluvial transport. Where such material exists at the glacier margin, its remobilization may result in it being dumped onto the adjacent terrain during ice recession. Depending on the rate of debris accumulation and glacier activity, a variety of sediment-landform associations will be created. For retreating glaciers, dumping of supraglacial material onto the former subglacial surface slowly emerging from beneath the ice produces a thin veneer of coarse-grained diamicton or perhaps only sporadic collections of boulders (Eyles, 1979). Dump moraines form where the ice margin remains stationary during debris accumulation (Boulton and Eyles, 1979), although they will be bulldozed into push moraines if the glacier undergoes a subsequent readvance. Dump moraine size is related to supraglacial debris volume and the length of the stillstand period. Small dump moraines will form where glaciers remain stationary during the winter but retreat during the summer, each moraine marking one winter's increment of debris accumulation. The largest and most spectacular dump moraines, however, form at the margins of debris-mantled valley glaciers which occupy similar positions for considerable periods (Figure 5). The dumping of large quantities of supraglacial debris around glacier margins builds up



Figure 1 Contemporary push moraines being constructed at the margin of Flajokull, Iceland.

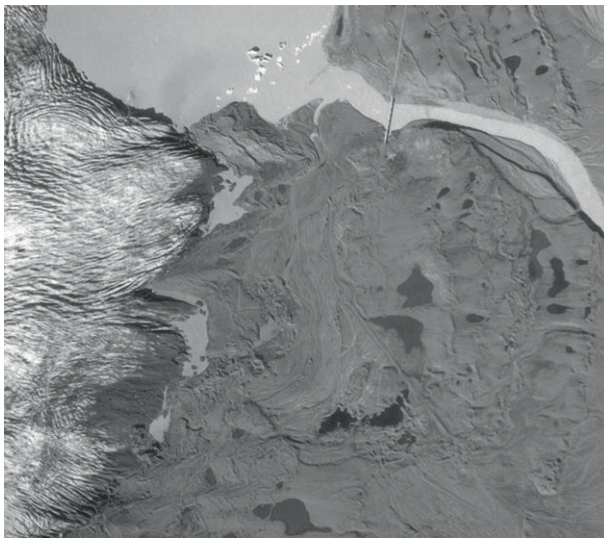


Figure 2 Aerial photograph of the margin of Fjallsjokull, Iceland in 1965, showing a sequence of annual push moraines with 'saw-tooth' planforms (© University of Glasgow and Landmaelingar Islands).

latero-frontal dump moraines (Boulton and Eyles, 1979; Small, 1983, 1987) through the processes of sliding, rolling, flowage, and fall from the ice surface to produce ice-contact scree. In mountain environments, the debris may be predominantly passively transported rockfall material, consisting of coarse, angular clasts with little matrix. However, actively transported debris can also crop out near valley glacier margins due to the elevation of basal debris septa (linear debris concentrations in glacier ice), thereby introducing matrix-rich debris to some dump moraines. On some latero-frontal moraines, there is a tendency for clast roundness to increase in a down-moraine/down-valley direction due to the increasing proportion of actively transported debris in the moraines closer to the former glacier centerline (Benn and Ballantyne, 1994; Evans, 1999; Matthews and Petch, 1982).

Where dump moraines are deposited on steep valley sides, debris derived directly from the glacier is interdigitated with sediment from the valley walls, producing an internal stratigraphy composed of (i) stratified debris flows and fall sorted

debris avalanche material with beds dipping away from glacier margin, (ii) subhorizontally bedded alluvium deposited as kame terraces along the ice margin, and (iii) debris flows and fall sorted screes from valley-side paraglacial fans. Removal of lateral support by glacier recession results in mass failure of the inner slopes of the moraines. In many cases, such failures clearly extend up into the screes which were deposited against the lateral moraines. During glacier thinning, dump moraines are abandoned and their inner faces are subject to collapse and reworking (Figure 6). A new inset dump moraine may be formed within the older one if the glacier restabilizes at a new position, but such moraines will be unstable if they are deposited on top of dead ice masses. Alternatively, renewed thickening of the glacier may dump material on top of the old dump moraine, which may become completely buried. In this case, periods of nondeposition may be recorded by paleosols or even buried trees, which are valuable sources of paleoclimatic data in Alpine environments (Röthlisberger et al., 1980). Localized breaching of latero-frontal moraine loops by advancing glacier snouts can give rise to the development of breach-lobe moraines (Deline, 1999).

Distinctive dump moraines known as frontal- or ice-marginal aprons form around the margins of some Arctic and polar glaciers (Evans, 1989; Fitzsimons, 1997; Shaw, 1977). These moraines form where debris and ice blocks accumulate at the base of steep, terminal ice cliffs by a combination of dry calving and limited melting (Figure 7). When forming, the aprons have steep ice-contact slopes and gentler ice-distal slopes determined by the properties of the debris. However, they will be subject to considerable modification following deglaciation due to the ablation of incorporated ice blocks.

Latero-Frontal Fans and Ramps

Deposition by debris flows and glaciofluvial processes around stationary glacier margins can result in the construction of latero-frontal fans and ramps, consisting of coalescent debris fans, which descend from the glacier snout (Benn et al., 2003; Owen and Derbyshire, 1989, 1993; Figure 8). After glacier recession, such landforms are identifiable by their pronounced asymmetric cross-profile, comprising a distal slope that is much shallower in gradient than on latero-frontal dump moraines and a steep proximal or ice-contact face. Fans deposited predominantly by glaciofluvial processes grade distally into outwash tracts or sandar, producing 'outwash head' landforms (Benn et al., 2003; Kirkbride, 2000; Krzyszkowski, 2002; Krzyszkowski and Zielinski, 2002). In some situations where moraines are absent, fans and ramps provide the only evidence of former ice-marginal positions. Ice-contact fan morphology develops over a considerable period, sometimes involving several generations of deposition with intervening periods of abandonment. During fan accumulation, active depocenters shift position, reflecting changes in the morphology of the ice margin and fan surface. As the fan progrades, surface angles tend to become progressively reduced. Meltwater exiting from englacial (within the ice) positions or draining the glacier surface locally excavates the ice-contact fan/ramp faces to produce inset meltwater fans. In active mountain belts, ice-contact fans can reach heights of hundreds of meters, attesting to very

high rates of debris transfer (Evans et al., 2010). Because debris flow fans and ramps are deposited partly over glacier ice, they may be characterized by pitted or degraded surfaces after ice melt-out. When ice-contact fans are abandoned by glacier retreat, the inner faces are susceptible to collapse and may

provide material for lateral moraine development during ice recession.

The sedimentology of latero-frontal fans/ramps is dominated by massive, often bouldery debris flow diamictos with beds up to tens of meters thick, interbedded with water-sorted

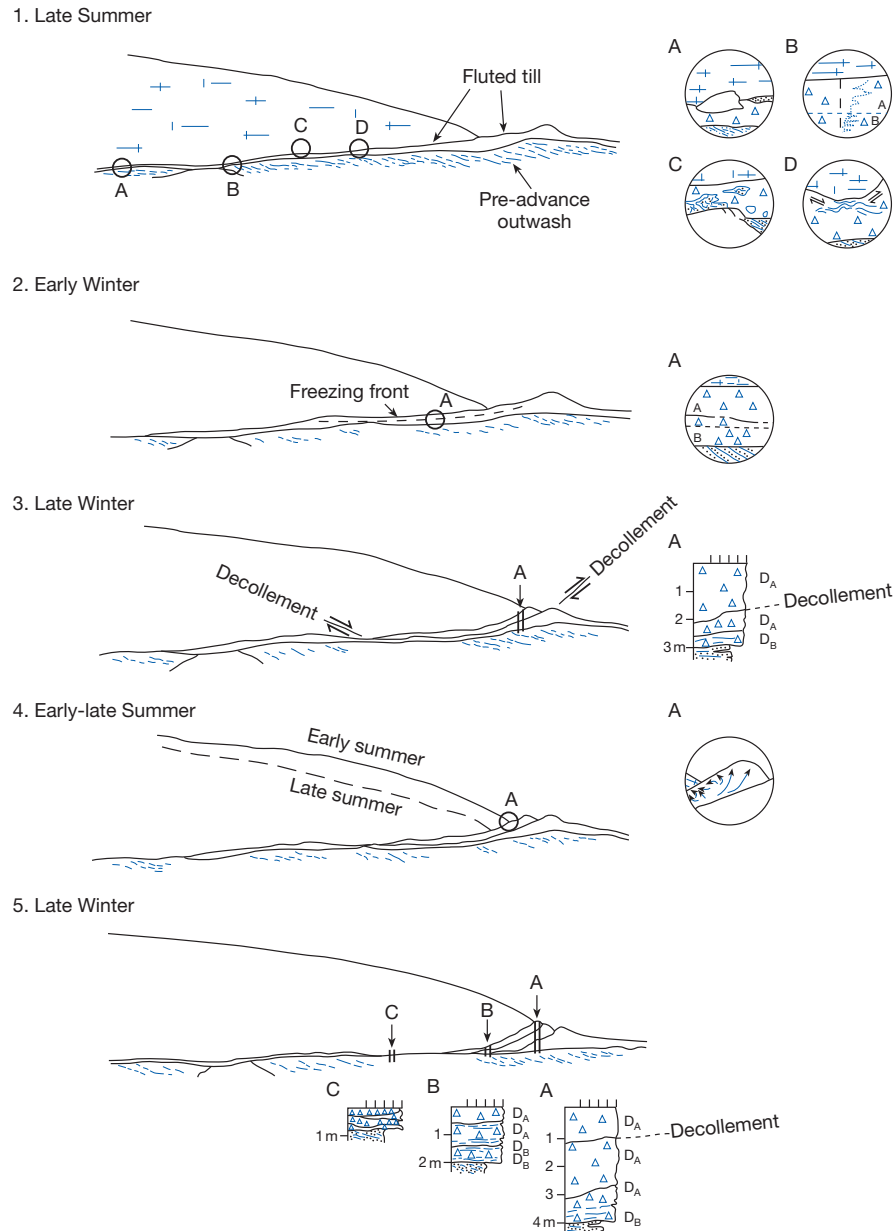


Figure 3 A model of push moraine formation by the incremental stacking of till slabs during an annual cycle of subglacial deformation, freeze-on, and melt-out. Reproduced from Evans DJA and Hiemstra JF (2005) Till deposition by glacier sub-marginal, incremental thickening. *Earth Surface Processes and Landforms* 30: 1633–1662. (1) Situation in late summer at typical Icelandic glacier snout where subglacial processes include lodgement and sliding (A), bedrock and sediment plucking (B), subglacial deformation (C), and ice-keel ploughing (D) in a temporally and spatially evolving process mosaic. (2) During early winter, the thin part of the glacier snout freezes onto part of the subglacial till. The till slab that freezes onto the ice sole is likely to be from the more porous A horizon (A). (3) The later winter readvance initiates failure along a decollement plane within the A horizon or at the junction with the more compact B horizon, resulting in the carriage of A horizon till onto the proximal side of the previous year's push moraine. (4) In the early summer, the melt-out of the till slab (A) initiates porewater migration, water escape, and sediment flow (small arrows) and sediment extrusion due to glaciostatic and glaciodynamic stresses. (5) The late summer situation is again followed by winter freeze-on and marginal stacking of subglacial till produced by the reworking of existing subglacial sediments and fresh materials advected to submarginal locations from up-ice. Repeated reworking of the thin end of submarginal till wedges produces overprinted strain signatures and clast pavements.

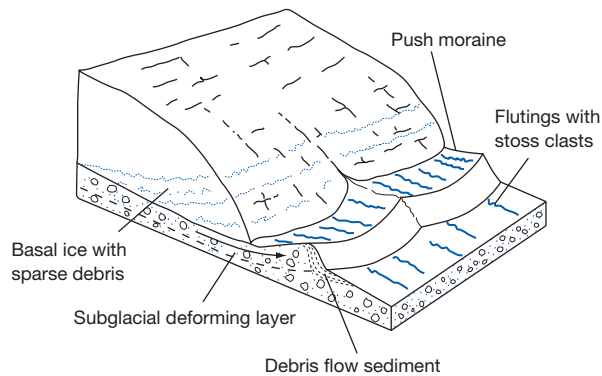


Figure 4 The formation of a push moraine by the combined action of pushing and squeezing from beneath an ice margin. Reproduced from Price RJ (1970) Moraines at Fjallsjokull, Iceland. *Arctic and Alpine Research* 2: 27–42.



Figure 5 Latero-frontal moraines at the margin of the Hooker Glacier, New Zealand, formed by the accumulation of debris dumped at the highly debris-charged ice margin when it stood at its Little Ice Age limit. Some glacier pushing will also have contributed to moraine construction.

sediments and dipping away from the glacier snout (Evans et al., 2010; Krzyszkowski and Zielinski, 2002; Lukas, 2005). These sequences comprise coalescent ice-marginal fans. Pockets of glaciolacustrine sediment, which have in some instances been glaciotectionized, attest to the occurrence of short-lived proglacial lakes ponded between the ice margin and the fans during ice retreat phases.

Medial Moraines

Medial moraines are prominent features on the surfaces of many glaciers and are the surface expression of a medial debris septum, which may be a shallow feature or extend to the base of the glacier (Figure 9). Ablation results in the concentration of the debris septum on the glacier surface. Medial moraines on existing glaciers have been classified by Eyles and Rogerson (1978) based on the relationship between debris supply and morphological development. Three broad types were recognized, including (i) ablation-dominant moraines, which emerge at the surface as the result of the melt-out of englacial debris; (ii) ice stream interaction moraines, which find

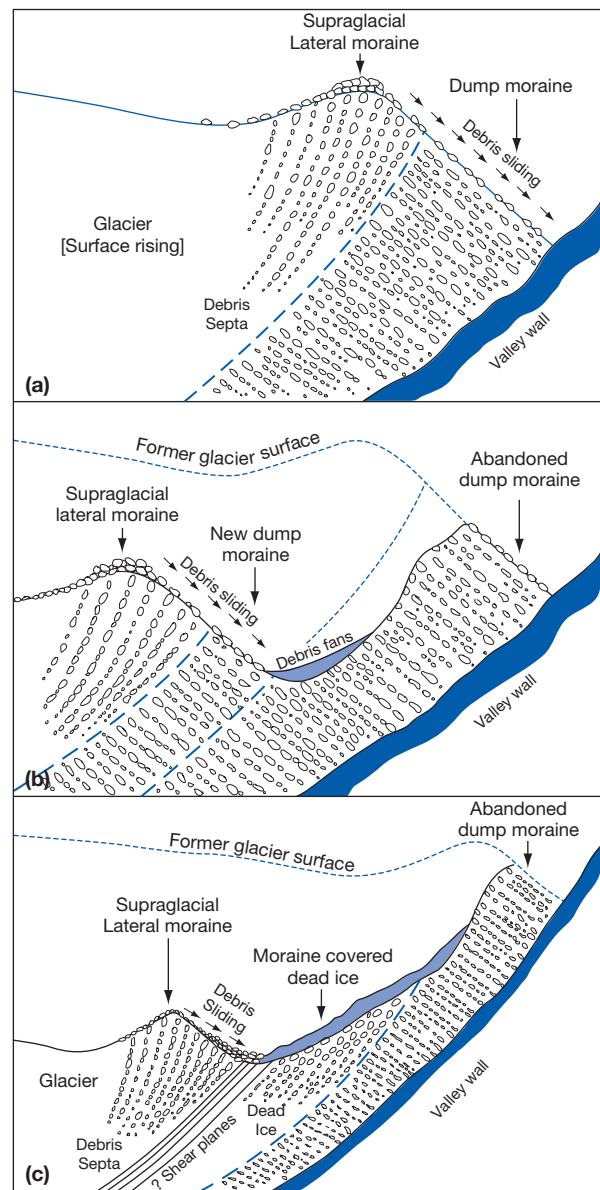


Figure 6 A model of lateral moraine formation by ice-marginal dumping of debris. Reproduced from Small RJ (1983) Lateral moraines of Glacier De Tsidjore Nouve: Form, development and implications. *Journal of Glaciology* 29: 250–259.

immediate surface expression downstream from glacier confluences, often by the merging of two supraglacial lateral moraines; and (iii) avalanche-type moraines, which are transient features formed by exceptional rockfall events onto a glacier surface. Medial moraines do not have good preservation potential because they contain relatively little debris and are subject to intense reworking during glacier melt. They sometimes can be traced from modern glacier surfaces onto proglacial surfaces, where they may be recognizable as diffuse spreads of debris with distinctive lithological composition or particle morphologies (Evans and Twigg, 2002). However, such moraines commonly have no marked surface expression and may



Figure 7 Ice-marginal apron produced by dumping of debris and dry calving of ice blocks from a subpolar glacier snout, Ellesmere Island, Arctic Canada.



Figure 8 Debris flow-dominated ice-marginal fans marking the former margin of the Batal Glacier in the Lahul Himalaya.

be very difficult to recognize once the glacier foreland becomes vegetated. Recognition of ancient medial moraines is easier where they consist of large boulders, which may be traced back to their source outcrops (e.g., Dawson, 1979).

In certain situations, ice stream interaction-type medial moraines have a high preservation potential and may

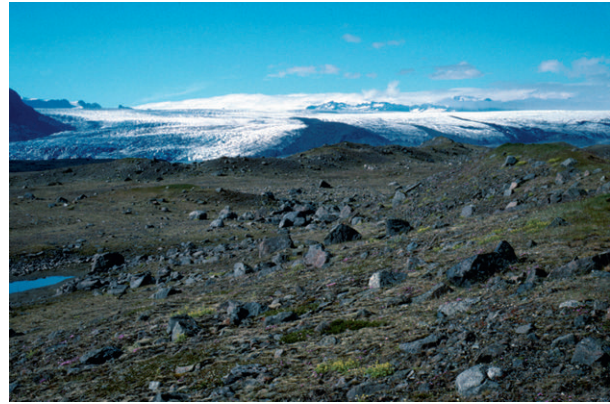


Figure 9 Medial moraines at the margin of Breidamerkurjökull, Iceland. The recession of the glacier snout has resulted in the lowering of the medial moraines on to subglacial landforms to produce a thin rubbly veneer visible in the foreground.

constitute interlobate moraines. For example, on the west coast of New Zealand, very large ice stream interaction medial moraines, which are greater than 100 m high in some areas, have been produced by the coalescence of ancient piedmont glaciers with well-developed lateral moraines. Because these features accumulated essentially as lateral moraines, they contain sediments characteristic of ice-marginal deposition, but the moraine cores commonly display glaciotectionic structures, and in places, carapaces of subglacial till record glacier coalescence and overriding. Benn and Evans (2010) have suggested that wherever such interlobate features become overridden by glacier ice, they may evolve into megaflutings.

Some medial/interlobate moraines may comprise large quantities of glaciofluvial sediment, for example, where lobate margins of an ice sheet coalesce. Two large examples occur at the southern margins of the former Laurentide Ice Sheet. The Harricana Interlobate Moraine, formerly separating the Hudson Bay and Labrador sectors of the receding ice sheet, stretches for 1000 km southward from James Bay to Lake Simcoe, Ontario, Canada (Veillette, 1986, 1988) and comprises a series of ridges up to 10 km wide and 100 m high. The Burntwood-Knife Interlobate Moraine, formerly separating the Hudson Bay and Keewatin sectors of the receding Laurentide Ice Sheet, covers a distance in excess of 500 km on the west side of Hudson Bay and is composed of broad ridges of glaciofluvial sediment and till. Individual segments are up to 12 km long, 4 km wide, and 60 m high (Dredge et al., 1986; Dyke and Dredge, 1989; Klassen, 1986). The large quantities of sands and gravels comprising these moraines were deposited in proglacial lacustrine environments during the final stages of the Laurentide Ice Sheet, attesting to the large-scale reworking of medial moraine debris during glacier recession. Glaciofluvial reworking of medial moraine debris during ice wastage can produce features indicative of former medial moraine locations even though they are strictly different landforms. For example, continuous esker systems may demarcate the former positions of medial moraines in glaciers or ice sheets where debris septum ensures constant sediment supply to englacial drainage networks (e.g., eskers and medial moraines in Breidamerkurjökull, Iceland; Evans and Twigg, 2002; Price,

1969, 1973). Similarly, a low-relief kame and kettle topography centered on the bottom of the Athabasca Valley near Jasper in Alberta, Canada is interpreted as a former supraglacial medial moraine by Levson and Rutter (1989; Figure 10).

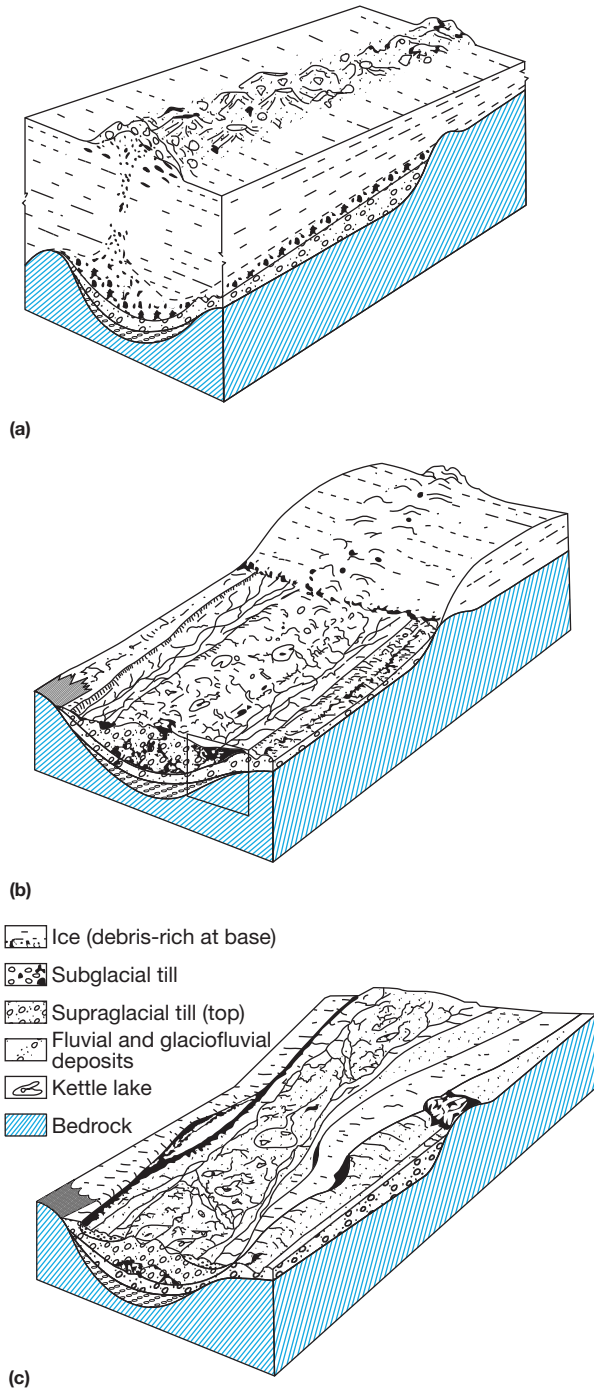


Figure 10 A model of medial moraine formation based on Canadian Rocky Mountain case studies (Levson and Rutter, 1989). Gradual lowering of the medial moraine results in the formation of a linear ice-cored moraine. This develops into a chain of kettle lakes and hummocky terrain comprising rubby supraglacial debris.

Hummocky Moraine and Controlled Moraine

The term 'hummocky moraine' has been employed in a wide range of senses. As a purely descriptive term, it has been applied to moundy, irregular morainic topography exhibiting varying degrees of order, ranging from entirely chaotic assemblages of mounds to suites of nested transverse ridges (Attig and Clayton, 1993; Benn, 1992; Bennett and Boulton, 1993; Sissons, 1967; Figure 11). Genetically, it has been used to encompass landforms with a variety of origins, including deposition in association with active and stagnant ice, and deformation (e.g., Aario, 1977; Bennett and Boulton, 1993; Eyles et al., 1999; Gravenor and Kupsch, 1959; Hoppe, 1952). The term 'hummocky moraine' is now used in a more restrictive sense (Benn and Evans, 2010) to refer to moraines deposited during the melt-out of debris-mantled glaciers (e.g., Hoppe, 1952, 1959; Sharp, 1949, 1985). This genetic definition should only be applied to suites of moraines whose origin has been established by detailed mapping and sedimentological analyses, because generalized descriptions of surface morphology alone are insufficient to determine moraine genesis. For example, many tracts of so-called hummocky moraine in the Highlands of Scotland, originally believed to be chaotic ice-stagnation topography (Sissons, 1967), have subsequently been recognized as nested push and dump moraines, or drumlins draped by supraglacial and ice-marginal deposits (Benn, 1992; Bennett and Boulton, 1993). Hummocky moraine deposited from debris-mantled ice can appear chaotic or linear in pattern, depending on the distribution of debris in the parent glacier and the patterns of debris redistribution and reworking during ice wastage. Occurrences with pronounced transverse linear elements are termed controlled moraines (Evans, 2009). Hummocky moraine is the end product of topographic inversion cycles during the ablation of debris-mantled ice (Figure 12). During ice ablation, debris is transferred away from topographic highs on the glacier surface by mass movements and meltwater, exposing ice cores to renewed melting and creating new depressions in former high points. Further debris reworking and topographic development is achieved by meltwater streams meandering between dirt cones, the collapse of englacial tunnels, and the enlargement of supraglacial lake basins. Several cycles of



Figure 11 Hummocky moraine produced by the melting of a debris-charged surge snout, Tungnaarjökull, Iceland.

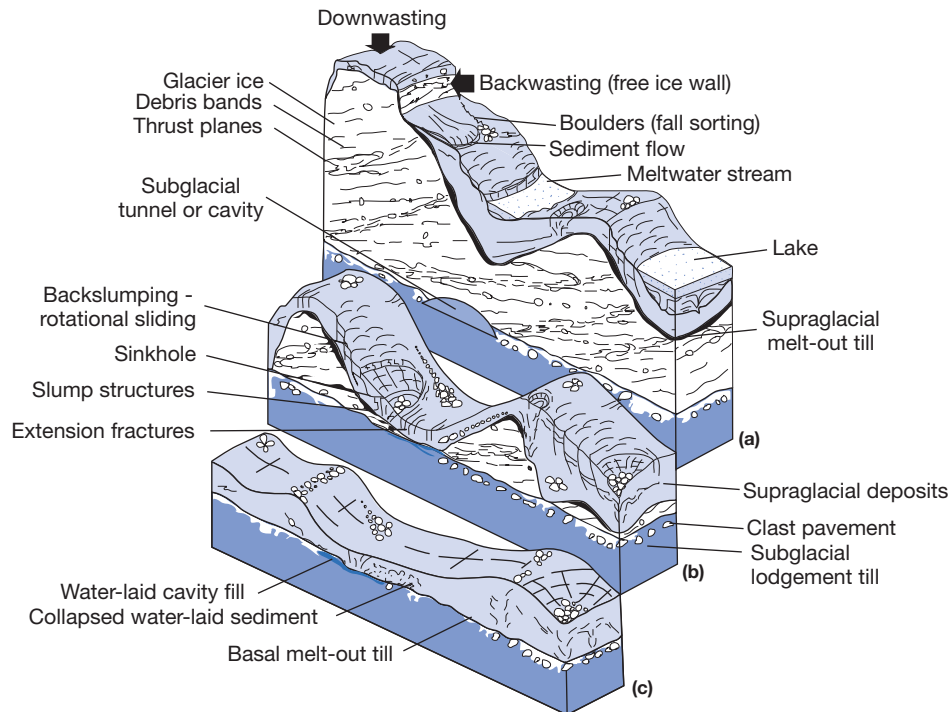


Figure 12 A model of hummocky moraine development as a result of the melting of a debris-charged glacier snout. Reproduced from Kjær KH and Krüger J (2001) The final phase of dead-ice moraine development: Processes and sediment architecture, Kötlujökull, Iceland. *Sedimentology* 48: 935–952.

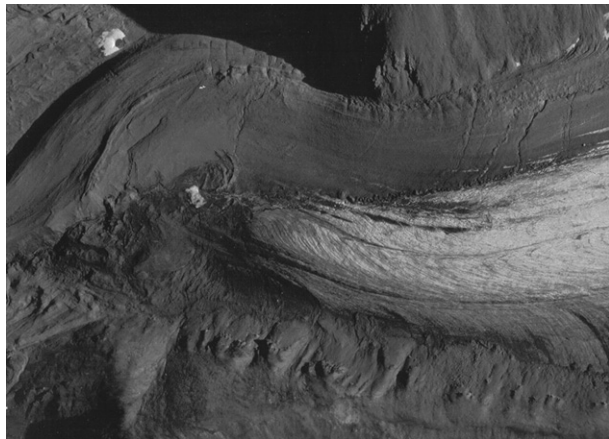


Figure 13 Controlled moraine on the surface of a glacier snout in St. Jonsfjorden, Svalbard. Aerial photograph from the Norsk Polar Institute.

debris reworking and topographic inversion may occur before the debris is finally deposited in a series of irregular mounds and ridges.

The distribution of controlled moraines is inherited from debris concentrations or septa within the parent glacier (Evans, 2009). Transverse debris concentrations close to glacier margins may result from the thrusting of debris along shear planes (Bishop, 1957; Goldthwait, 1951; Souchez, 1967), the emergence of basal debris frozen onto the glacier sole (Boulton, 1970; Swift et al., 2006), or reworked

ice-marginal aprons (Evans, 1989; Hooke, 1970; Hudleston, 1976; Shaw, 1977). Such transverse septa control the pattern of differential ablation on the glacier surface, producing linear ice-cored ridges which eventually become separated from the snout to form ice-cored moraines (Dyke and Evans, 2003; Evans, 2009, 2010, 2011; Østrem, 1964). The morphological expression of englacial structures is always most striking while the moraines retain ice cores (Figure 13). However, uneven sediment redistribution during ice-core wastage means that the final deposits tend to consist of discontinuous transverse ridges with intervening hummocks, preserving only a weak impression of the former englacial structure. Reworking by supraglacial meltwater systems also acts to destroy large tracts of controlled moraine before they can be deposited on the ground surface. Patterns of debris reworking and deposition can also be controlled by crevasses or other lines of weakness within the ice, which can guide the final stages of glacier disintegration (Gravenor and Kupsch, 1959).

The sedimentology of hummocky moraine is complex, reflecting multiple cycles of redeposition during its formation (Kjær and Krüger, 2001). Typical facies associations consist of interbedded debris flows and other mass-movement deposits, laminated lacustrine sediments, and glaciofluvial sands and gravels, in varying degrees of disturbance (Eyles, 1979; Paul, 1983). The constituent debris may be actively or passively transported, depending on the position of the moraine relative to debris transport paths in the parent glacier. In some cases, debris from different sources and transport paths may remain segregated within the moraine, giving rise to compositional stratification (Benn, 1992).

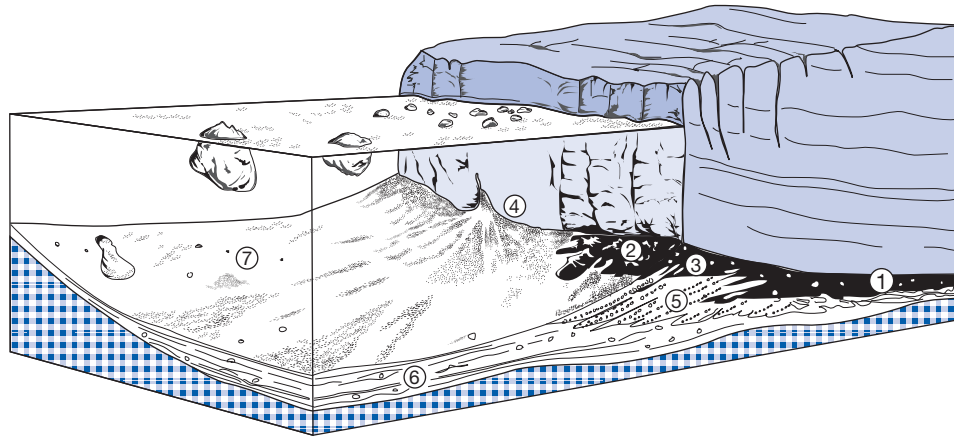


Figure 14 A reconstruction of the depositional environment in which morainal banks accumulate at the glacier grounding line, based on the example at Achnasheen, Scotland (from [Benn, 1996](#)). (1) Subglacial till, (2) mass flows, (3) mass flow diamictons, (4) subglacial meltwater portal, (5) subaqueous outwash, (6) laminated sediments, and (7) iceberg dump mound.

Morainal Banks and Coalescent Subaqueous Fans

Wherever glacier grounding lines remain quasistable for long periods of time and/or sediment fluxes to the glacier margin are high, subaqueous moraines will form by the accumulation of sediment along the glacier margin. Debris may be delivered from the ice surface, the melt-out of debris-rich ice and icebergs, the release of unfrozen till from a subglacial deforming layer, or reworking of older deposits ([Figure 14](#)). This debris can be reworked in conveyor belt fashion by erosion off the subglacial face and deposition on the subaqueous slope ([Powell, 1984](#)).

Subaqueous moraines that extend up into shallow water can also be exposed to wave energy, leading to winnowed and sorted gravels and sands in their upper parts ([Powell, 1984](#)). Some subaqueous moraines mainly consist of coalescent grounding-line fans, deposited at the same time at the efflux points (exit locations) of adjacent meltwater conduits. For example, the Salpausselka Moraines of Finland consist of transverse chains of ice-contact fans and deltas deposited during stillstands of the southern margins of the Scandinavian ice sheet where it contacted deep water ([Fyfe, 1990](#)). Similarly, some large-scale moraines, comprising partially glaciotectionized grounding-line fans, occur in eastern Maine, USA ([Ashley et al., 1991](#)) where they are arranged laterally along former grounding-line positions and are associated with tunnel mouth and esker deposits. A similar depositional scenario is envisaged by [Sharpe et al. \(2007\)](#) for the Oak Ridges Moraine in Ontario, Canada.

Some subaqueous moraines are formed mainly by dumping and other gravitational processes at the foot of terminal ice cliffs. This is especially the case where subpolar glaciers with thick basal debris sequences contact the sea, releasing large amounts of debris in environments where there is only limited subglacial meltwater ([Powell, 1984](#)). At many glacier margins, however, gravitational deposits are interbedded with subaqueous outwash, particularly at the margins of temperate glaciers where meltwater is abundant. For example, [McCabe et al. \(1984, 1987\)](#), [McCabe \(1985, 1986\)](#), and [Eyles and](#)

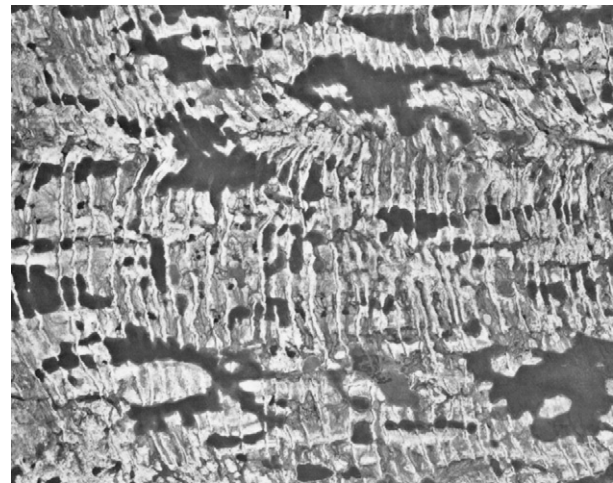


Figure 15 Aerial photograph of De Geer moraines in Quebec, Canada (National Aerial Photograph Library A14882-91).

[McCabe \(1989\)](#) present Pleistocene examples of interbedded gravitational and outwash deposits from the east central coast of Ireland in which channelized subaqueous outwash and debris flow facies interdigitate with ice-rafted dropstone muds. These sediments show evidence of glaciotectionic deformation and underlie arcuate belts of hummocky topography.

De Geer Moraines

Named after Swedish geologist Gerard De Geer, De Geer moraines commonly occur in fields of closely spaced ridges in association with subaqueous sediments and mark the intermittent retreat of water-terminating glaciers ([Figure 15](#)). Many models have been proposed to account for De Geer moraine formation, but the consensus is that they mark former ice-margin positions, where sediment is deposited or pushed up during brief stillstands or minor readvances. In common with terrestrial push moraines, De Geer moraines are characterized

by steeper distal than proximal slopes, lobate forms with acute interlobe angles pointing up glacier, and truncated sections due to partial overriding by later moraines (Boulton, 1986). The formation of subaqueous moraines with steep ice-distal slopes and gentle ice-proximal slopes has also been linked to the calving of floating glacier snouts, perhaps on a yearly basis (Aartolahti, 1972; Holdsworth, 1973). In contrast, De Geer moraines may result from the squeezing of saturated sediment out from beneath the ice margin (Andrews and Smithson, 1966).

The notion that several De Geer moraines may be produced simultaneously has been proposed by Hoppe (1957), Lundqvist (1989), and Zilliacus (1989). This implies formation in basal crevasses a short distance behind glacier margins in a similar fashion to terrestrial crevasse fills (cf. Solheim, 1986), and indeed De Geer moraines are often juxtaposed with probable crevasse-squeeze ridges (cf. Ottesen and Dowdeswell, 2006; Ottesen et al., 2008; Todd et al., 2007). Specifically, transverse crevasses are formed in areas of extending flow and then widened during partial flotation of the ice margin. Moraines are thought to form when the water level drops and the glacier is lowered onto the substratum, which is squeezed up into the basal crevasses. Tabular icebergs would later break off along the transverse crevasses, thus ensuring that the moraines were not later overridden by the glacier. Such an explanation actually implies a crevasse-squeeze origin for the ridges and suggests that the glacier snouts responsible for their construction underwent draw-down and fracturing (Lundqvist, 1989) possibly in response to surging (Zilliacus, 1989). Such an origin would make the ridges directly comparable with terrestrial crevasse-fill ridges on surging glacier forelands (Evans and Rea, 1999, 2003; Sharp, 1985).

Similarly, features referred to as 'lift-off moraines' have been identified in seismostratigraphic investigations of the

Scotian and mid-Norwegian shelves by King and Fader (1986) and King et al. (1991). The production of lift-off moraines is thought to be related to fracture patterns in the base of a retreating glacier snout, involving a genesis similar to that proposed by Lundqvist (1989). It is uncertain whether lift-off moraines are produced individually by single events or simultaneously as groups (King et al., 1991).

Ice-Shelf Moraines

Ice-shelf moraines are formed where debris is deposited or relocated around the margins of ice shelves or floating glacier tongues. The crestlines of such moraines tend to be horizontal or have very low gradients, reflecting the low gradient of ice shelves that have zero or very low basal shear stresses. The gradients of ancient ice-shelf moraines, however, may be steeper due to postglacial isostatic tilting, so their gradients should be corrected with reference to the sea-level record. The recent ice-shelf moraine at the margin of the George VI Sound ice shelf, Antarctica, formed where complex ice pressure ridges were induced by flow against Alexander Island. A horizontal moraine ridge consequently formed over a distance of 120 km along the promontories of Alexander Island where basal debris was brought to the glacier surface by ablation and thrusting (Figure 16). Sediments within the moraine include subglacial debris entrained on Alexander Island and the Antarctic Peninsula, glaciofluvial and lacustrine sediments from ice-marginal streams and ponds, and marine sediments and fauna derived from freezing on at the grounding line on Alexander Island (Sugden and Clapperton, 1981). Similarly, Glasser et al. (2006) report inset moraines of single- or multiple-crested ridges containing a variety of material, including ice cores, constructed on Bratina Island by the McMurdo Ice Shelf.

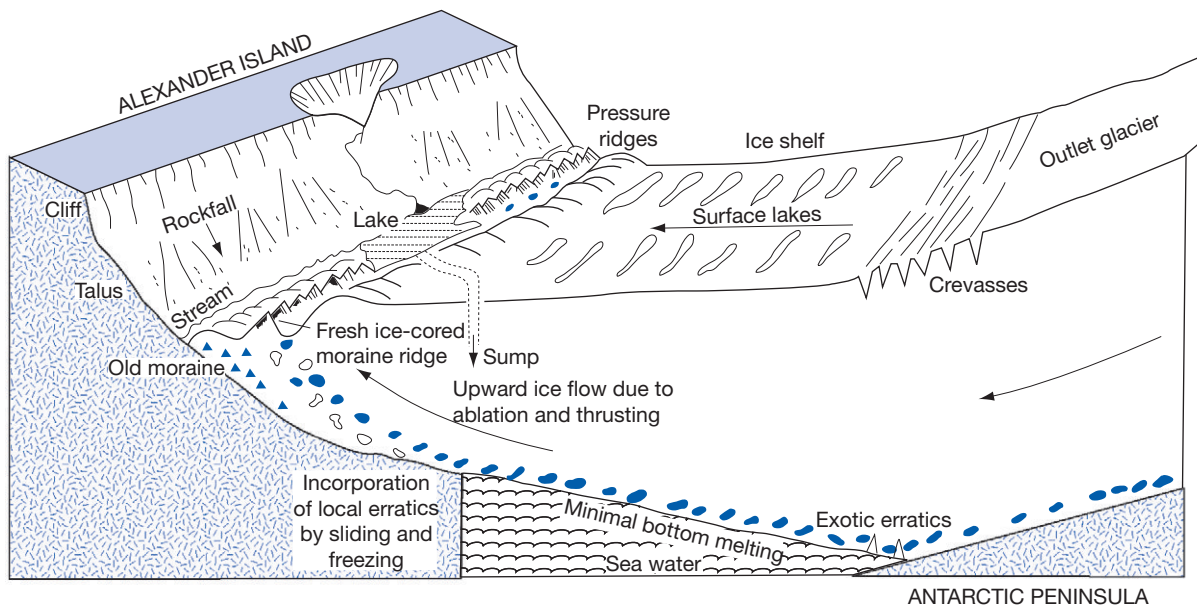


Figure 16 A model for the development of the George VI Sound ice-shelf moraine, Antarctica. Reproduced from Sugden DE and Clapperton CM (1981) An ice-shelf moraine, George VI Sound, Antarctica. *Annals of Glaciology* 2: 135–141.

They propose that ice-shelf moraine production involves the entrainment of debris by anchor ice formation over offshore shoals, which is then mixed with medial moraine debris on the ice-shelf surface as it pushes onshore. Ancient ice-shelf moraines are rare (e.g., England et al., 1978; Evans et al., 2002), but a spectacular example occurs at the former north-western margin of the Laurentide Ice Sheet in Viscount Melville Sound, where extensive linear belts of till (Winter Harbour Till), between the altitudes of 60 and 150 m above present sea level, blanket the shores of Byam Martin, Melville, Banks, and Victoria islands and extend eastward to north Baffin Island (Dyke et al., 1992; England et al., 2009; Hodgson and Vincent, 1984; Klassen, 1993).

See also: **Glacial Landforms:** Glacitectonic Structures and Landforms. **Glacial Landforms, Sediments:** Glaciofluvial Landforms of Deposition; Glaciomarine Sediments and Ice-Rafted Debris.

References

- Aario R (1977) Classification and terminology of moraine landforms in Finland. *Boreas* 6: 87–100.
- Aartolahti T (1972) On deglaciation in southern and western Finland. *Fennia* 114: 1–84.
- Andrews JT and Smithson BB (1966) Till fabrics of the cross valley moraines of north-central Baffin Island, NWT, Canada. *Bulletin of the Geological Society of America* 77: 271–290.
- Ashley GM, Boothroyd JC, and Borns HW (1991) Sedimentology of Late Pleistocene (Laurentide) deglacial phase deposits, eastern Maine; an example of a temperate marine grounded ice-sheet margin. In: Anderson JB and Ashley GM (eds.) *Glacial Marine Sedimentation; Paleoclimatic Significance*, vol. 261, pp. 107–125. Boulder, CO: Geological Society of America, Special Paper.
- Attig JW and Clayton L (1993) Stratigraphy and origin of an area of hummocky glacial topography, northern Wisconsin, USA. *Quaternary International* 18: 61–67.
- Benn DI (1992) The genesis and significance of 'hummocky moraine': Evidence from the Isle of Skye, Scotland. *Quaternary Science Reviews* 11: 781–799.
- Benn DI (1996) Subglacial and subaqueous processes near a glacier grounding line: Sedimentological evidence from a former ice-dammed lake, Achnasheen, Scotland. *Boreas* 25: 23–36.
- Benn DI and Ballantyne CK (1994) Reconstructing the transport history of glacialigenic sediments: A new approach based on the co-variance of clast form indices. *Sedimentary Geology* 91: 215–227.
- Benn DI and Evans DJA (2010) *Glaciers and Glaciation*, 2nd edn. London: Hodder Education.
- Benn DI, Kirkbride MP, Owen LA, and Brazier V (2003) Glaciated valley landforms. In: Evans DJA (ed.) *Glacial Landforms*, pp. 372–406. London: Arnold.
- Bennett MR and Boulton GS (1993) A reinterpretation of Scottish 'hummocky moraine' and its significance for the deglaciation of the Scottish highlands during the Younger Dryas or Loch Lomond Stadial. *Geological Magazine* 130: 301–318.
- Bishop BC (1957) Shear moraines in the Thule area, northwest Greenland. US Snow, Ice and Permafrost Research Establishment, Research Report 17.
- Boulton GS (1970) On the origin and transport of englacial debris in Svalbard glaciers. *Journal of Glaciology* 9: 213–229.
- Boulton GS (1986) Push moraines and glacier contact fans in marine and terrestrial environments. *Sedimentology* 33: 677–698.
- Boulton GS and Eyles N (1979) Sedimentation by valley glaciers: A model and genetic classification. In: Schluchter C (ed.) *Moraines and Varves*, pp. 11–23. Rotterdam: Balkema.
- Dawson AG (1979) A Devensian medial moraine in Jura. *Scottish Journal of Geology* 15: 43–48.
- Deline P (1999) La mise en place de l'amphithéâtre morainique du Miage (Val Veny, Val d'Aoste). *Geomorphologie: Relief, Processus, Environnement* 1: 59–72.
- Dredge LA, Nixon FM, and Richardson RJ (1986) Quaternary geology and geomorphology of northwestern Manitoba. *Geological Survey of Canada, Memoir* 418: 38.
- Dyke AS and Dredge LA (1989) Quaternary geology of the northwestern Canadian shield. In: Fulton RJ (ed.) *Quaternary Geology of Canada and Greenland*, pp. 178–214. Ottawa: Geological Survey of Canada: Geology of Canada, Publication No. 1.
- Dyke AS and Evans DJA (2003) Ice-marginal terrestrial landforms: Northern Laurentide and Innuitian ice sheet margins. In: Evans DJA (ed.) *Glacial Landforms*, pp. 143–165. London: Arnold.
- Dyke AS, Morris TF, Green DEC, and England J (1992) Quaternary geology of Prince of Wales Island, Arctic Canada. *Geological Survey of Canada, Memoir* 433: 142.
- England JH, Bradley RS, and Miller GH (1978) Former ice shelves in the Canadian high Arctic. *Journal of Glaciology* 20: 393–404.
- England JH, Furze MFA, and Doupe JP (2009) Revision of the NW Laurentide Ice Sheet: Implications for paleoclimate, the northeast extremity of Beringia, and Arctic Ocean sedimentation. *Quaternary Science Reviews* 28: 1573–1596.
- Evans DJA (1989) The nature of glaciectonic structures and sediments at subpolar glacier margins, northwest Ellesmere Island, Canada. *Geografiska Annaler* 71A: 113–123.
- Evans DJA (1999) Glacial debris transport and moraine deposition: A case study of the Jurdalen cirque complex, Sogn-og-Fjordane, western Norway. *Zeitschrift für Geomorphologie* 43: 203–234.
- Evans DJA (2009) Controlled moraines: Origins, characteristics and palaeoglaciological implications. *Quaternary Science Reviews* 28: 183–208.
- Evans DJA (2010) Controlled moraine development and debris transport pathways in polythermal plateau icefields: Examples from Tungnafellsjökull, Iceland. *Earth Surface Processes and Landforms* 35: 1430–1444.
- Evans DJA (2011) Glacial landforms of Satujökull, Iceland: A modern analogue for glacial landform overprinting by mountain icecaps. *Geomorphology* 129: 225–237.
- Evans DJA and Hienstra JF (2005) Till deposition by glacier sub-marginal, incremental thickening. *Earth Surface Processes and Landforms* 30: 1633–1662.
- Evans DJA and Rea BR (1999) Geomorphology and sedimentology of surging glaciers: A landform approach. *Annals of Glaciology* 28: 75–82.
- Evans DJA and Rea BR (2003) Surging glacier landform. In: Evans DJA (ed.) *Glacial Landforms*, pp. 259–288. London: Arnold.
- Evans DJA, Rea BR, Hansom JD, and Whalley WB (2002) The geomorphology and style of plateau icefield glaciations in a fjord terrain, Troms-Finnmark, north Norway. *Journal of Quaternary Science* 17: 221–239.
- Evans DJA, Shand M, and Petrie G (2009) Maps of the snout and proglacial landforms of Fjallsjökull, Iceland (1945, 1965, 1998). *Scottish Geographical Journal* 125: 304–320.
- Evans DJA, Shulmeister J, and Hyatt O (2010) Sedimentology of latero-frontal moraines and fans on the west coast of South Island, New Zealand. *Quaternary Science Reviews* 29: 3790–3811.
- Evans DJA and Twigg DR (2002) The active temperate glacial landform: A model based on Breidamerkurjökull and Fjallsjökull, Iceland. *Quaternary Science Reviews* 21: 2143–2177.
- Eyles N (1979) Facies of supraglacial sedimentation on Icelandic and alpine temperate glaciers. *Canadian Journal of Earth Sciences* 16: 1341–1361.
- Eyles N, Boyce JL, and Barendregt RW (1999) Hummocky moraine: Sedimentary record of stagnant Laurentide Ice Sheet lobes resting on soft beds. *Sedimentary Geology* 123: 163–174.
- Eyles N and McCabe AM (1989) Glaciomarine facies within subglacial tunnel valleys: The sedimentary record of glacio-isostatic downwarping in the Irish Sea Basin. *Sedimentology* 36: 431–448.
- Eyles N and Rogerson RJ (1978) A framework for the investigation of medial moraine formation: Austerdalsbreen, Norway, and Berendson Glacier, British Columbia, Canada. *Journal of Glaciology* 20: 99–113.
- Fitzsimons SJ (1997) Depositional models for moraines in East Antarctic coastal oases. *Journal of Glaciology* 43: 256–264.
- Fyfe GJ (1990) The effect of water depth on ice-proximal glaciolacustrine sedimentation: Salpausselkä I, southern Finland. *Boreas* 19: 147–164.
- Glasser NF, Goodsell B, Copland L, and Lawson W (2006) Debris characteristics and ice shelf dynamics in the ablation region of the McMurdo Ice Shelf, Antarctica. *Journal of Glaciology* 52: 223–234.
- Goldthwait RP (1951) Development of end moraines in east central Baffin Island. *Journal of Geology* 59: 567–577.
- Gravenor CP and Kupsch WO (1959) Ice disintegration features in western Canada. *Journal of Geology* 67: 48–64.
- Hodgson DA and Vincent JS (1984) A 10,000 yr BP extensive ice shelf over Viscount Melville Sound, Arctic Canada. *Quaternary Research* 22: 18–30.
- Holdsworth G (1973) Ice deformation and moraine formation at the margin of an ice cap adjacent to a proglacial lake. In: Fahey BD and Thompson RD (eds.) *Research in Polar and Alpine Geomorphology*, pp. 187–199. Norwich: Geo Abstracts.
- Hooke RL (1970) Morphology of the ice-sheet margin near Thule, Greenland. *Journal of Glaciology* 9: 303–324.

- Hoppe G (1952) Hummocky moraine regions, with special reference to the interior of Norrbotten. *Geografiska Annaler* 34: 1–72.
- Hoppe G (1957) Problems of glacial morphology and the ice age. *Geografiska Annaler* 39: 1–18.
- Hoppe G (1959) Glacial morphology and inland ice recession in northern Sweden. *Geografiska Annaler* 41: 193–212.
- Hudleston PJ (1976) Recumbent folding in the base of the Barnes Ice Cap, Baffin Island, Northwest Territories, Canada. *Geological Society of America Bulletin* 87: 1684–1692.
- King LH and Fader G (1986) Wisconsinan glaciation of the continental shelf – Southeast Atlantic Canada. *Geological Survey of Canada, Bulletin* 363: 72.
- King LH, Rokoengen K, Fader GBJ, and Gunleiksrud T (1991) Till tongue stratigraphy. *Bulletin of the Geological Society of America* 103: 637–659.
- Kirkbride MP (2000) Ice marginal geomorphology and Holocene expansion of debris-covered Tasman Glacier, New Zealand. In: Nakawo M, Raymond C, and Fountain A (eds.) *Debris-Covered Glaciers*, vol. 264. pp. 211–217. Wallingford, UK: IAHS Publication.
- Kjær KH and Krüger J (2001) The final phase of dead-ice moraine development: Processes and sediment architecture, Kötlujökull, Iceland. *Sedimentology* 48: 935–952.
- Klassen RW (1986) Surficial geology of north-central Manitoba. *Geological Survey of Canada, Memoir* 419: 57.
- Klassen RA (1993) Quaternary geology and glacial history of Bylot Island, Northwest Territories. *Geological Survey of Canada, Memoir* 429: 93.
- Krüger J (1993) Moraine-ridge formation along a stationary ice front in Iceland. *Boreas* 22: 101–109.
- Krüger J (1994) Glacial processes, sediments, landforms, and stratigraphy in the terminus region of Myrdalsjökull, Iceland. *Folia Geographica Danica* 21: 1–233.
- Krüger J (1996) Moraine ridges formed from subglacial frozen on sediment slabs and their differentiation from push moraines. *Boreas* 25: 57–63.
- Krzyszowski D (2002) Sedimentary successions in ice-marginal fans of the late Saalian glaciation, southwestern Poland. *Sedimentary Geology* 149: 93–109.
- Krzyszowski D and Zielinski T (2002) The Pleistocene end moraine fans: Controls on their sedimentation and location. *Sedimentary Geology* 149: 73–92.
- Levson VM and Rutter NW (1989) Late Quaternary stratigraphy, sedimentology, and history of the Jasper townsite area, Alberta, Canada. *Canadian Journal of Earth Sciences* 26: 1325–1342.
- Lukas S (2005) A test of the englacial thrusting hypothesis of 'hummocky' moraine formation: Case studies from the north-west Highlands, Scotland. *Boreas* 34: 287–307.
- Lundqvist J (1989) Rogen (ribbed) moraine – Identification and possible origin. *Sedimentary Geology* 62: 281–292.
- Matthews JA, Cornish R, and Shakesby RA (1979) Sawtooth moraines in front of Bodalsbreen, southern Norway. *Journal of Glaciology* 22: 535–546.
- Matthews JA, McCarroll D, and Shakesby RA (1995) Contemporary terminal-moraine ridge formation at a temperate glacier: Styggedalsbreen, Jotunheimen, southern Norway. *Boreas* 24: 129–139.
- Matthews JA and Petch JR (1982) Within-valley asymmetry and related problems of Neoglacial lateral moraine development at certain Jotunheimen glaciers, southern Norway. *Boreas* 11: 225–247.
- McCabe AM (1985) Geomorphology. In: Edwards KJ and Warren WP (eds.) *The Quaternary History of Ireland*, pp. 67–93. London: Academic Press.
- McCabe AM (1986) Glaciomarine facies deposited by retreating tidewater glaciers – An example from the Late Pleistocene of Northern Ireland. *Journal of Sedimentary Petrology* 56: 880–894.
- McCabe AM, Dardis GF, and Hanvey PM (1984) Sedimentology of a Late Pleistocene submarine-moraine complex, County Down, Northern Ireland. *Journal of Sedimentary Petrology* 54: 716–730.
- McCabe AM, Dardis GF, and Hanvey PM (1987) Sedimentation at the margins of a Late Pleistocene ice lobe terminating in shallow marine environments, Dundalk Bay, eastern Ireland. *Sedimentology* 34: 473–493.
- Østrem G (1964) Ice-cored moraines in Scandinavia. *Geografiska Annaler* 46: 282–337.
- Ottesen D and Dowdeswell JA (2006) Assemblages of submarine landforms produced by tidewater glaciers in Svalbard. *Journal of Geophysical Research* 111: F01016. <http://dx.doi.org/10.1029/2005JF000330>.
- Ottesen D, Dowdeswell JA, Benn DI, et al. (2008) Submarine landforms characteristic of glacier surges in two Spitsbergen fjords. *Quaternary Science Reviews* 27: 1583–1599.
- Owen LA and Derbyshire E (1989) The Karakoram glacial depositional system. *Zeitschrift für Geomorphologie Supplement Band* 76: 33–73.
- Owen LA and Derbyshire E (1993) Quaternary and Holocene intermontane basin sedimentation in the Karakoram Mountains. In: Shroder JF (ed.) *Himalaya to the Sea*, pp. 108–131. London: Routledge.
- Paul MA (1983) The supraglacial landsystem. In: Eyles N (ed.) *Glacial Geology*, pp. 71–90. Oxford: Pergamon.
- Powell RD (1984) Glaciomarine processes and inductive lithofacies modelling of ice shelf and tidewater glacier sediments based on Quaternary examples. *Marine Geology* 57: 1–52.
- Price RJ (1969) Moraines, sandar, kames and eskers near Breidamerkurjökull, Iceland. *Transactions of the Institute of British Geographers* 46: 17–43.
- Price RJ (1970) Moraines at Fjallsjökull, Iceland. *Arctic and Alpine Research* 2: 27–42.
- Price RJ (1973) *Glacial and Fluvio-glacial Landforms*, p. 242. Edinburgh: Oliver and Boyd.
- Röthlisberger F, Haas P, Holzhauser H, Keller W, Bircher W, and Renner F (1980) Holocene climatic fluctuations – Radiocarbon dating of fossil soils (fAh) and woods from moraines and glaciers in the Alps. *Geographica Helvetica* 35: 21–52.
- Sharp RP (1949) Studies of superglacial debris on valley glaciers. *American Journal of Science* 247: 289–315.
- Sharp MJ (1984) Annual moraine ridges at Skálafellsjökull, south-east Iceland. *Journal of Glaciology* 30: 82–93.
- Sharp MJ (1985) Sedimentation and stratigraphy at Eyjabbakkajökull – An Icelandic surging glacier. *Quaternary Research* 24: 268–284.
- Sharpe DR, Russell HAJ, and Logan C (2007) A 3-dimensional geological model of the Oak Ridges Moraine area, Ontario, Canada. *Journal of Maps* 2007: 239–253.
- Shaw J (1977) Till body morphology and structure related to glacier flow. *Boreas* 6: 189–201.
- Sissons JB (1967) *The evolution of Scotland's scenery*. Edinburgh: Oliver and Boyd.
- Small RJ (1983) Lateral moraines of Glacier De Tsidiore Nouve: Form, development and implications. *Journal of Glaciology* 29: 250–259.
- Small RJ (1987) Moraine sediment budgets. In: Gurnell AM and Clark MJ (eds.) *Glacio-Fluvial Sediment Transfer: An Alpine Perspective*, pp. 165–197. Chichester: Wiley.
- Solheim A (1986) Submarine evidence of glacier surges. *Polar Research* 4: 91–95.
- Souchez RA (1967) The formation of shear moraines: An example from south Victoria Land, Antarctica. *Journal of Glaciology* 6: 837–843.
- Sugden DE and Clapperton CM (1981) An ice-shelf moraine, George VI Sound, Antarctica. *Annals of Glaciology* 2: 135–141.
- Swift DA, Evans DJA, and Fallick AE (2006) Transverse englacial debris-rich ice bands at Kviárjökull, southeast Iceland. *Quaternary Science Reviews* 25: 1708–1718.
- Todd BJ, Valentine PC, Longva O, and Shaw J (2007) Glacial landforms on German Bank, Scotian Shelf: Evidence for Late Wisconsinan ice sheet dynamics and implications for the formation of De Geer moraines. *Boreas* 36: 148–169.
- Veillette JJ (1986) Former southwesterly ice flows in the Abitibi-Timiskaming region: Implications for the configuration of the late Wisconsinan ice sheet. *Canadian Journal of Earth Sciences* 23: 1724–1741.
- Veillette JJ (1988) Deglaciation et evolution des lacs proglaciaires Post-Algonquin et Barlow au Temiscamingue, Quebec et Ontario. *Geographie Physique et Quaternaire* 42: 7–31.
- Zilliacus H (1989) Genesis of De Geer moraines in Finland. *Sedimentary Geology* 62: 309–317.

Quaternary Vulcanism: Subglacial Landforms

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Glossary

Caprock: Subhorizontal layer of resistant rock (e.g., lava) that overlies (caps) less-resistant rocks (usually fragmental material).

Cryolithosphere: Permanently frozen subsurface broadly corresponding to the permafrost zone, where any water present occurs as ice; mainly used in studies of Mars.

Effusive: Said of a volcanic eruption characterized by the relatively nonexplosive ejection of flowing molten lava.

Felsic: Relating to lava (magma) compositions that are often pale colored and are relatively rich in silica and poor in iron and magnesium, such as andesite, dacite, trachyte, phonolite, and rhyolite.

Glaciovolcanic: All types of volcano interactions with ice in all its forms (including snow and firn) and, by implication, any meltwater derived from that ice by volcanic heating.

Lava-fed delta: A distinctive flat-topped volcanic landform that is the volcanic equivalent of a sedimentary delta and formed where lava enters a lake or the sea. It comprises subhorizontal subaerial lava overlying steep-dipping glassy breccias. The presence of lava-fed deltas is what gives mafic tuyas their distinctive morphology.

Mafic: Relating to lava (magma) compositions that are dark colored and are relatively rich in iron and magnesium and poor in silica, such as basalt and basaltic andesite.

Monogenetic: Formed during a single, usually relatively short-lived event (e.g., an eruption).

Morphometry: The measurement and mathematical analysis of landforms and other objects.

Polygenetic: Formed during more than one event (e.g., multiple eruptions).

Stratovolcano: A stratified volcano composed of alternating lava and fragmental rocks (e.g., tephra), commonly long-lived, large and constructed during multiple eruptions (i.e., polygenetic).

Tephra: Volcanic ash and other solidified or molten rock explosively ejected during an eruption.

Tindar: Conical or ridge-like pile dominantly composed of volcanic ash generated by explosive subaqueous eruptions in a glacial environment (meltwater-filled vault or lake) as a result of subglacial eruptions. Some authors restrict the name to ridge-like outcrops only.

Tuff cone: Conical pile with a summit crater, formed of explosively erupted volcanic ash (tephra).

Tuya: Flat-topped steep-sided volcanic landform, the product of volcanic eruptions through a glacier or ice sheet, in the sea or a nonglacial lake. The word is most commonly used for an edifice constructed as a result of subglacial eruptions, particularly those involving mafic magmas with lava-fed deltas, but is sometimes applied to flat-topped volcanic constructs erupted in other environments and involving magmas with more evolved (felsic) compositions.

Volcano–ice interactions are common on Earth (e.g., [Edwards et al., 2009b](#); [Smellie and Chapman, 2002](#), and papers therein). A few eruptions have been observed and there are many extinct edifices of Quaternary age and older. Where ice sheets have shrunk or long since vanished, the subglacial volcanic edifices are frequently highly accessible and permit close-up, detailed field analysis. They are overwhelmingly the products of basaltic eruptions since basaltic volcanism is the dominant mode of volcanic activity on Earth. Edifices formed by more evolved magmas also occur but are much less common and most examples occur in Iceland ([McGarvie, 2009](#)). The internal structure, hydraulic evolution, and eruptive processes have been relatively well investigated (e.g., [Edwards et al., 2009a](#); [Gudmundsson et al., 2004](#); [Schopka et al., 2006](#); [Skilling, 2009](#); [Smellie, 2000, 2001, 2006, 2008](#); [Smellie and Hole, 1997](#); [Smellie and Skilling, 1994](#); [Smellie et al., 1993, 2011a](#); [Stevenson et al., 2009, 2011](#); [Tuffen et al., 2002a,b, 2009](#)). Most of those studies concentrated on documenting the lithofacies present, which were extremely poorly known prior to the 1990s, and using the sequence architecture to reconstruct the mode(s) of edifice construction and associated eruption dynamics. By contrast, the study of subglacial volcanic landforms has not advanced

far and there are very few publications ([Hickson, 2000](#); [Komatsu et al., 2007](#); [Smellie, 2007, 2009](#)).

Subglacial eruptions are now called glaciovolcanic. The term ‘glaciovolcanic’ refers to all types of volcano interactions with ice in all its forms (including snow and firn) and, by implication, any meltwater derived from that ice by volcanic heating ([Smellie, 2006](#)). It includes not only eruptions and their products that took place underneath a glacier or ice sheet (i.e., subglacial), but also subaerial volcanic products (e.g., lavas, pyroclastic flows, etc.) that simply ponded and chilled against an ice barrier or else interacted with snow by overriding or entrainment ([Harder and Russell, 2007](#); [Lescinsky and Sisson, 1998](#); [Mee et al., 2006](#); [Thouret, 1990](#)). The many different types of glaciovolcanic sequences identified so far can be used to obtain a uniquely wide range of critical parameters of past terrestrial ice sheets, for which other (i.e., nonvolcanic) records are typically patchily and poorly preserved and rely on geomorphologic criteria that are highly susceptible to reworking (e.g., moraines, trimlines). They are thus a valuable, but still under-used, proxy for documenting paleoenvironmental changes over time (e.g., [Edwards and Russell, 2002](#); [McGarvie et al., 2006, 2007](#); [Mee et al., 2006](#);

Smellie, 2000; Smellie and Skilling, 1994; Smellie et al., 2006b, 2008, 2009, 2011b; Stevenson et al., 2006). Reliably identifying and documenting the occurrence of subglacially erupted volcanoes and their distribution over geological time is also central to understanding the inventory and dynamic behavior of past and present water on Earth.

Terrestrial landforms also provide our best empirical basis for interpreting remotely observed features on Earth's nearest planetary neighbors (i.e., the Moon, Mars, and Venus) and understanding their geological evolution. Mars, like Earth, also has a substantial inventory of volatile elements, including an unusually thick cryolithosphere (i.e., subsurface ice) and surface ice in the form of small polar ice caps that may have been much more extensive in times past. It is also dominated by basaltic volcanism. At present, the description, quantification, and interpretation of Mars volcanic landforms are wholly reliant on satellite observations and comparisons with possible terrestrial analogs. The Mars landforms include candidate examples of mafic tuyas (volcanic table mountains; see later) and tephra cones and ridges, which have been erupted subglacially (e.g., Chapman, 2003; Chapman and Tanaka, 2001; Head and Pratt, 2001; Ghatan and Head, 2002; Chapman and Smellie, 2007; Martínez-Alonso et al., 2011). Their identification was used to suggest the former existence of a much-expanded south polar ice cap in Hesperian times (~1.8–3.5 Gy) and formerly extensive mid- to low-latitude ice during past high-obliquity periods. However, definitive evidence for glaciovolcanism on Mars remains elusive.

This article focuses on volcanic landforms that were erupted and constructed subglacially. There are no known distinctive landforms that were constructed following volcanic interactions with snow or subsurface ice; subaerial volcanic constructs that simply chilled against surface ice are also excluded (e.g., Harder and Russell, 2007; Lescinsky and Fink, 2000; Lescinsky and Sisson, 1998; Mee et al., 2006; Stevenson et al., 2006). The ability to make robust identifications of volcanic landforms, and understanding the intrinsic and extrinsic environmental controls on their formation, are important research frontiers that are set to become increasingly important for deciphering ancient planetary glacial records (Smellie et al., 2008, 2009, 2011a,b).

Subglacial Volcanic Landforms

Subglacial eruptions on Earth have formed several different types of volcanic landforms (Hickson, 2000; Smellie, 2009). However, our knowledge of these landforms is currently so poor that their classification and description are almost certainly oversimplified. The descriptions that follow also take little account of post eruptive erosional effects, which can be significant especially for glaciovolcanic edifices containing weak piles of waterlogged volcanoclastics that are easily glacially modified and can also undergo multiple postglacial sector collapses once the buttressing effect of the surrounding ice is removed (e.g., Komatsu et al., 2007; Tuffen et al., 2002b). However, they are described here essentially as pristine primary morphologies. The different volcanic landforms reflect differences in (1) lithofacies and sequence architecture; (2) intrinsic properties of the enclosing ice sheet or glacier (i.e., thickness,

thermal regime, and meltwater hydraulics); and (3) magmatic composition, volatile content, rheology, and discharge (Hickson, 2000; Smellie, 2000, 2001, 2009; Tuffen et al., 2002a). All but two of the landforms identified so far are volcano edifices. The two exceptions (called subglacial sheet-like sequences) are also common subglacial volcanic products and they form an additional, although poorly defined, volcanic landform resulting from subglacial eruptions. The several volcanic landform types are illustrated in Figure 1 and summarized in Table 1. The glaciovolcanic data used in this study are presented in Table 2.

Practically all of the published examples of subglacial volcanic landforms were constructed during relatively short-lived monogenetic eruptions beneath temperate or wet-based ice. There are no descriptions of monogenetic edifices formed under cold-based ice, although some of the products of those eruptions have been described (Smellie et al., 2011a). There are also very few descriptions of long-lived (i.e., millions of years) subglacial polygenetic stratovolcanoes or shield volcanoes and there are indications that the compound morphometry of those larger landforms may be very different from those formed by monogenetic eruptions (Edwards and Russell, 2002; Smellie et al., 2008, 2011a). Overall, our fundamental knowledge of subglacial volcanic landforms is at an early stage, and the following descriptions are heavily biased toward the monogenetic examples and should be viewed in that context.

This article is based mainly on reviews by Smellie (2007, 2009). The landforms are divided empirically into mafic examples, typically basalt to basaltic andesite or andesite in composition, and felsic types, principally dacite and more evolved magmas. There is a clear hierarchy of landform types (Figure 2). However, the distinction is unlikely to be so clear-cut in nature and some felsic subglacial volcanoes, particularly those with intermediate compositions, will probably be found that show features more commonly found in mafic systems, and vice versa. A preliminary assessment of the landform morphometry is included in each section, using volume and aspect ratio as empirical parameters. Aspect ratio is defined as the ratio of maximum height (above volcano base) to edifice width. For sheet-like sequences, deposit thickness is used rather than height since the coeval edifice dimensions are either unknown or undiagnostic. Aspect ratio is an admittedly crude empirical parameter and it does not closely reproduce many true volcano profiles, but it is still a useful and informative parameter that can enable distinctions to be made. It cannot resolve the distinctive combination of relatively flat tops and steep perimeter cliffs that characterize mafic tuyas, and it also fails to recognize summit craters. However, this is not always a problem. For example, craters are among the first casualties of post eruptive erosion, particularly when constructed of poorly consolidated tephra.

Mafic Tuyas

Mafic tuyas are the most distinctive and frequently described subglacial volcanic edifices. They are also known nongenetically as table mountains or mesas, although not all table mountains and mesas are tuyas. Tuyas are also the commonest

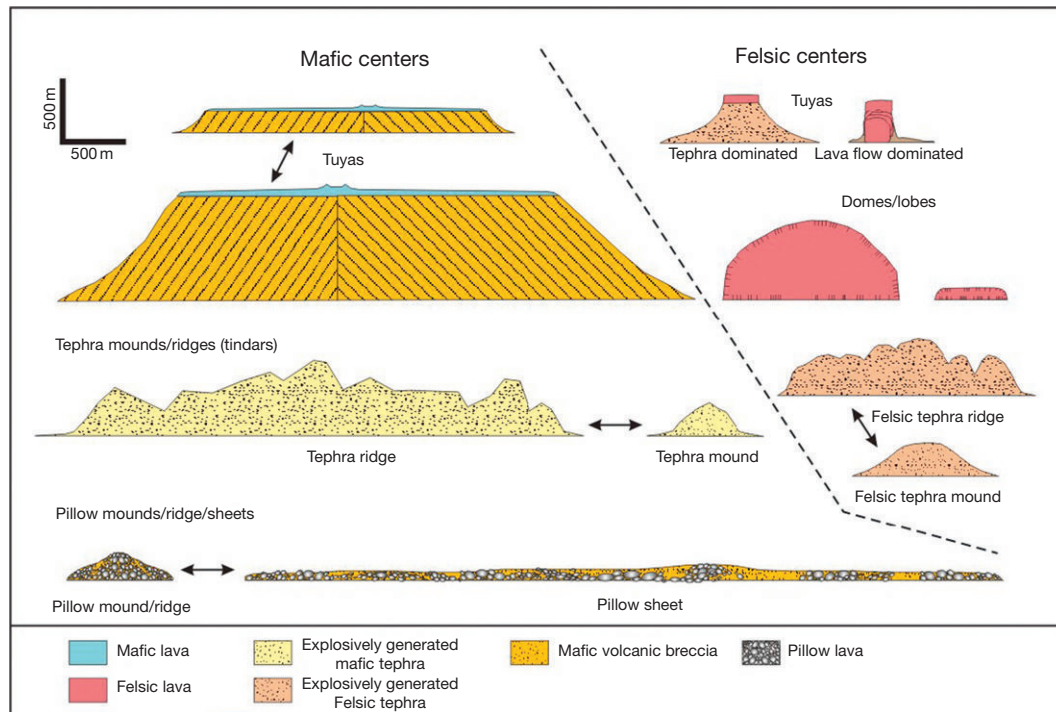


Figure 1 Approximately true-scale vertical profile sketches showing the principal characteristics of the major subglacial volcanic landforms described in this article. Sheet-like sequences are not shown but would resemble pillow sheets in dimensions.

subglacial volcanic landform to have been identified on Mars. The edifices are typically small and steep-sided, with relatively flat or, more typically, gently domed summit regions with or without a crater, and oval to straight-sided plan shapes (van Bemmelen and Rutten, 1955; Jones, 1969; Figure 3). Although commonly regarded as products of monogenetic eruptions, even some volumetrically quite small examples can be polygenetic (Jones, 1969; Skilling, 2009; Smellie, 2001; Werner and Schmincke, 1999; Wörner and Viereck, 1987; Figure 3(c)). Published examples of much larger polygenetic mafic tuya-type volcanoes are scarce, possibly because the edifices would need to be erupted consistently beneath an ice cover that persisted or repeatedly reformed during the lifetime of the volcano (millions of years). The Mount Haddington volcano, on James Ross Island, Antarctica, is a large polygenetic terrestrial example formed from repeated large-volume (typically tens of cubic kilometers) pahoehoe lava-fed deltas (Smellie et al., 2008). It rises 1.6 km, has a basal diameter of 60 km, encircling cliffs up to 600 m high and was constructed in multiple extrusive phases since the latest Miocene (Figure 4(a)). It has a low surface profile similar to a shield volcano but has the steep margins characteristic of confinement by ice. However, those margins are not primary but are a result of modification by numerous arcuate collapses accentuated by subsequent glacierization. Other examples are common in northern Victoria Land, also in Antarctica. They are mainly late Miocene in age (~10–5 Ma) and are shield volcanoes largely constructed from multiple relatively thin aa lava-fed deltas (Smellie et al., 2011a; Figure 4(b)). Aa lava-fed deltas are different in their lithofacies and emplacement characteristics from pahoehoe deltas. Of potential importance here is the

observation that the subaqueously deposited foreset section in each aa delta (formed of massive or poorly bedded lobe-hyaloclastite) typically dips at a lower angle than equivalent hyaloclastite breccia foresets in pahoehoe lava-fed deltas, so the conspicuous steep margins characteristic of pahoehoe deltas are much less pronounced (Smellie et al., 2011a). The volcanoes individually rise to ~2 km and are typically ca. 20–30 km in diameter. Like Mt Haddington, morphologically the Victoria Land edifices seem to simply resemble nonglacial mafic shields and only the lithofacies provide unambiguous evidence for a glacial regime. Because they do not form a distinctive tuya-like landform, neither they nor Mt Haddington are tuyas, despite their construction from multiple superimposed lava-fed deltas.

Monogenetic mafic tuyas are defined by the presence of lava-fed deltas, which give tuyas their distinctive ‘flat’ top. Thus, they require a meltwater lake to form in and, if a lava-fed delta is absent, the landform is unlikely to be a mafic tuya (Figure 5). The deltas represent a late subaerial progradational phase of edifice-building by relatively rapid lateral extension (Skilling, 2002; Smellie, 2000, 2001, 2006) and they have a distinctive sequence architecture comprising a relatively thin caprock (topsets) formed of flat-lying subaerial lava overlying steep-dipping, homoclinal, glassy volcanic breccias (foresets) formed by the mechanical disintegration of the caprock lava during quenching where it reached and entered water, in this case a glacial lake (Figure 5). Two types are known: pahoehoe and aa deltas (cf. Skilling, 2002; Smellie, 2006; Smellie et al., 2011a). They have different lithofacies due to the different emplacement dynamics of the feeding lavas. Described aa lava-fed deltas have less steep delta fronts (typically $\leq 10^\circ$), than pahoehoe deltas (~25–35°). However, only

Table 1 Classification of terrestrial subglacial volcanic landforms and sequences

<i>Landform</i>	<i>Mafic tuya</i>	<i>Felsic tuya (1) Tephra- dominated</i>	<i>Felsic tuya (2) Lava flow dominated</i>	<i>Felsic domes & lobes</i>	<i>Tephra mound/ ridge (tindar)</i>	<i>Pillow mound/ sheet</i>	<i>Sheet-like sequences</i>
Composition	Basalt	Andesite, dacite, rhyolite, trachyte, phonolite	Andesite, dacite, rhyolite, trachyte, phonolite	Andesite, dacite, rhyolite, trachyte, phonolite	Basalt, andesite, dacite, rhyolite	Basalt	Basalt; rare trachyte, phonolite, rhyolite
Morphology	Table mountain, mesa	Table mountain, mesa	Flat-topped column, bladed ridge	Tall steep-sided domes, lobes or irregular	Steep cone or ridge	Low oblate smooth mounds, subdued ridges, extensive sheets	Thin sheets, sinuous ribbons & lobes
Characteristic lithologies	Subaerial lava, glassy breccia, tephra	Sheet lava, tephra, pumice, polymict sediments	Sheet lava, tephra, polymict sediments	Lava, glassy breccia	Tephra	Pillow and (probably) sheet lava	Volcanic sand & gravel, sheet and pillow lava, glassy breccia
Architecture	Lava-fed delta on precursor tephra cone (tindar) and/or pillow mound	Lava-capped table mountain; ash core	Lava column	Lava mound (mainly endogenous?)	Cone/ridge	Mound/sheet	Esker-like sediments, sheet lava & tephra
Deposit type	Vent edifice	Vent edifice	Vent edifice	Vent edifice	Vent edifice	Vent edifice	Outflow facies
Eruptive environment	Ice sheet; glacial lake	Ice sheet	Ice sheet	Alpine glacier, mountain ice cap, ice sheet	Ice sheet; glacial vault/lake	Ice sheet; not well known	Alpine glacier, mountain ice cap; ice sheet
Known distribution	Iceland, Canada, Antarctica	Iceland	Canada	Iceland, Canada	Iceland, Antarctica	Iceland, Canada, Antarctica?	Iceland, Canada, Antarctica

Modified from Smellie JL (2009) Terrestrial sub-ice volcanism: Landform morphology, sequence characteristics and environmental influences, and implications for candidate Mars examples. In Chapman MG and Keszthely L (eds.) *Preservation of Random Mega-Scale Events on Mars and Earth: Influence on Geologic History*, vol. 453, pp. 55–76. Geological Society of America Special Papers.

thin-ice-emplaced aa deltas have been described (Smellie et al., 2011a) and aa deltas emplaced in much thicker ice might have steep flanks comparable with pahoehoe deltas. No monogenetic tuyas constructed from aa lava eruptions have yet been described, whereas descriptions of pahoehoe-fed tuyas are common (e.g., Jones, 1969; Skilling, 2009; Smellie, 2006). Although tuyas commence construction as a subaqueous tuff cone and/or pillow mound (see later), those earlier-formed constructs do not significantly affect the final shape of the edifice. The flat or gently convex top is a subaerial volcanic shield constructed after the ice sheet was melted through completely during the eruption and it is formed of lavas with a summit crater or small pyroclastic cone of scoria or agglutinate. The principal limitation on the lateral extent of a mafic tuya is the volume of magma discharged over time, since deltas will advance as long as space is available and magma is supplied. By contrast, tuyas formed by eruptions within wet-based glaciers might be limited to heights less than 750–1000 m, since meltwater formed during eruptions in much thicker ice sheets should always theoretically float the surrounding glacier, and a stable meltwater lake (necessary for a lava-fed delta to form) will not occur (Smellie, 2009). A height limitation does

not apply to tuyas erupted in cold-based ice since that ice is frozen to its bed and cannot be lifted hydraulically. Although these suggestions are speculative, model based and not yet supported by observed eruptions, it might be significant that the largest wet-ice-emplaced monogenetic mafic tuyas in the present dataset (much expanded from Smellie, 2009) are 900–1000 m high, that is, the theoretical limit although it might simply reflect the rarity of mafic eruptions sufficiently sustained or voluminous to penetrate ice 1 km thick (see also Smellie, 2009, for further explanation and discussion).

Area is plotted against volume in Figure 6. Mafic tuyas have the largest volumes of any subglacial volcanic landform and their areas are also among the largest. They define a moderately well-defined trend with the highest volume:area ratio (mean 0.43) for any of the subglacial volcanic landforms (Figure 6(b)). The trend is curved (concave-up) and fits a power law or polynomial distribution very well (Figure 6(a)). Figure 7(a) shows aspect ratios using, as the horizontal dimension, the basal length of edifices defined as the *maximum* horizontal extent. The length is strongly affected by the mode of eruption: whether from fissures or point sources. Fissure length is a tectonically controlled variable and is not an intrinsic

Table 2 Morphometric parameters of terrestrial subglacial volcanic landforms

<i>Feature</i>	<i>Location</i>	<i>Height (m)</i>	<i>Length (km)</i>	<i>Width (km)</i>	<i>Aspect ratio (1) (ht/length)</i>	<i>Aspect ratio (2) (ht/width)</i>	<i>Area (km²)</i>	<i>Volume (km³)</i>	<i>Sources</i>	<i>Comments</i>
<i>Tuyas, mafic</i>										
Basalt tuya	Sellandafjall, NE Iceland	480	4.5	3	0.107	0.160	10.60	4.453	Iceland 1:300 000 topographical & geological maps	Also Schopka et al. (2006)
Basalt tuya	Blafjall, Iceland	620	12	4	0.052	0.155	37.70	20.452	Iceland 1:300 000 topographical & geological maps	Also Schopka et al. (2006)
Basalt tuya	Hrafnabjorg, SW Iceland	350	2.8	2	0.125	0.175	4.40	1.347	Iceland 1:300 000 topographical & geological maps	Also Werner and Schmincke (1999)
Basalt tuya	Hrutfell, SW central Iceland	700	6	5.5	0.117	0.127	25.92	15.875	Iceland 1:300 000 topographical & geological maps	
Basalt tuya	Laugarvatnsfjall, SW Iceland	400	10	6	0.040	0.067	47.12	16.493	Iceland 1:300 000 topographical & geological maps	Also Werner and Schmincke (1999)
Basalt tuya	Gaesafjoll, NE Iceland	380	5	4.5	0.076	0.084	17.67	5.876	Iceland 1:300 000 topographical & geological maps, Licciardi et al. (2007)	Also Schopka et al. (2006)
Basalt tuya	Herdubreid, Iceland	960	5.3	4	0.181	0.240	16.65	13.986	Iceland 1:300 000 topographical & geological maps, Licciardi et al. (2007)	Widths taken from Werner et al. (1996)
Basalt tuya	Hlodufell, SW Iceland	585	4	2.8	0.146	0.209	8.80	4.503	Iceland 1:300 000 topographical & geological maps, Licciardi et al. (2007)	Also Werner and Schmincke (1999)
Basalt tuya	Blafell, SW central Iceland	650	10.5	9	0.062	0.072	74.22	42.213	Iceland 1:300 000 topographical & geological maps, Licciardi et al. (2007)	
Basalt tuya	Skrida, SW Iceland	430	4.5	2.5	0.096	0.172	8.84	3.324	Iceland 1:300 000 topographical & geological maps, Licciardi et al. (2007)	Also Werner and Schmincke (1999)
Basalt tuya	Hognhofdi, SW Iceland	550	4.2	2.5	0.131	0.220	8.25	3.969	Iceland 1:300 000 topographical & geological maps, Licciardi et al. (2007)	Also Werner and Schmincke (1999)
Basalt tuya	Raudafell, SW Iceland	425	3.5	2.5	0.121	0.170	6.87	2.556	Iceland 1:300 000 topographical & geological maps, Licciardi et al. (2007)	Also Werner and Schmincke (1999)

Basalt tuya	Burfell, Iceland	480	3	2.5	0.160	0.192	5.89	2.474	Iceland 1:300 000 topographical & geological maps, Licciardi et al. (2007)	Also van Bemmelen and Rutten (1955)
Basalt tuya	Hafrafell, Iceland	300	2.2	1.2	0.136	0.250	2.07	0.544	Iceland 1:300 000 topographical & geological maps, Licciardi et al. (2007)	
Basalt tuya	Sandfell, Iceland	300	2.2	2	0.136	0.150	3.46	0.907	Iceland 1:300 000 topographical & geological maps, Licciardi et al. (2007)	
Basalt tuya	Snartarstearnapur, Iceland	280	2.8	2.2	0.100	0.127	4.84	1.185	Iceland 1:300 000 topographical & geological maps, Licciardi et al. (2007)	
Basalt tuya	Hvalfell, Iceland	420	3.2	3	0.131	0.140	7.54	2.771	Iceland 1:300 000 topographical & geological maps, Licciardi et al. (2007)	
Basalt tuya	Geitafell, Iceland	255	2.8	2.2	0.091	0.116	4.84	1.079	Iceland 1:300 000 topographical & geological maps, Licciardi et al. (2007)	
Basalt tuya	Eiríksjökull (W Langjökull), Iceland	1000	12	10	0.083	0.100	94.25	82.467	Iceland 1:300 000 topographical & geological maps	
Basalt tuya	Thorisjökull (W Langjökull), Iceland	600	11	8	0.055	0.075	69.12	36.285	Iceland 1:300 000 topographical & geological maps	
Basalt tuya	Ingólfsjökull (near Hengil), Iceland	400	7	5	0.057	0.080	27.49	9.621	Iceland 1:300 000 topographical & geological maps	
Basalt tuya	Kjalfjall (NE Langjökull), Iceland	400	4	3	0.100	0.133	9.42	3.299	Iceland 1:300 000 topographical & geological maps	
Basalt tuya	Cape Purvis, Dundee Island, Antarctica	500	6.5	5.5	0.077	0.091	28.08	12.284	Smellie et al. (2006c)	
Basalt tuya	Brown Bluff, Tabarin Peninsula, Antarctica	600	6.5	6.5	0.092	0.092	33.18	17.421	Skilling (1994)	Also Skilling (1994); prob near-symmetrical
Basalt tuya	Dobson Dome, James Ross Island, Antarctica	900	13	5	0.069	0.180	51.05	40.203	J. Smellie, unpublished information	Summit elevation inflated by ice cover; feature may include older edifices at one end
Basalt tuya	Shield Nunatak, Victoria Land, Antarctica	250	1.6	1.1	0.156	0.227	1.38	0.302	Wörner and Viereck (1987)	

(Continued)

Table 2 (Continued)

<i>Feature</i>	<i>Location</i>	<i>Height (m)</i>	<i>Length (km)</i>	<i>Width (km)</i>	<i>Aspect ratio (1) (ht/length)</i>	<i>Aspect ratio (2) (ht/width)</i>	<i>Area (km²)</i>	<i>Volume (km³)</i>	<i>Sources</i>	<i>Comments</i>
Basalt tuya	Tuber Hill, BC, Canada	450	2	2	0.225	0.225	3.14	1.237	Kelman et al. (2002)	
Basalt tuya	South Tuya, BC, Canada	500	5	3	0.100	0.167	11.78	5.154	Moore et al. (1995)	
Basalt tuya	Tuya Butte, BC, Canada	400	2.5	1	0.160	0.400	1.96	0.687	Moore et al. (1995)	
Basalt tuya	Hyalo Ridge, BC, Canada	500	4.2	2.4	0.119	0.208	7.92	3.464	Hickson et al. (1995)	
Basalt tuya?	Jack's Jump, BC, Canada	600	5	5	0.120	0.120	19.63	10.308	Hickson et al. (1995)	
Basalt tuya	Mosquito Mound, BC, Canada	300	6.3	3.6	0.048	0.083	17.81	4.676	Hickson et al. (1995)	Height said to be 245–345 (grouped with Gauge Hill, McLeod Hill & Mosquito Mound)
Basalt tuya	Gauge Hill, BC, Canada	300	3.1	1.8	0.097	0.167	4.38	1.150	Hickson et al. (1995)	Height said to be 245–345 (grouped with Gauge Hill, McLeod Hill & Mosquito Mound)
'Basalt' tuya	Sorug-Chushku-Uzu, Tuva Republic, Russia	400	10	2	0.040	0.200	15.71	5.498	Komatsu et al. (2007)	Several Russian tuyas are degraded and not primary landforms
'Basalt' tuya	Prozernyi, Tuva Republic, Russia	400	2.7	1.2	0.148	0.333	2.54	0.891	Komatsu et al. (2007)	Several Russian tuyas are degraded and not primary landforms
'Basalt' tuya	Shivit-Taiga, Tuva Republic, Russia	500	11.8	3.4	0.042	0.147	31.51	13.786	Komatsu et al. (2007)	Several Russian tuyas are degraded and not primary landforms
'Basalt' tuya	Kok-Hemskii, Tuva Republic, Russia	600	4.7	1.8	0.128	0.333	6.64	3.488	Komatsu et al. (2007)	Several Russian tuyas are degraded and not primary landforms
'Basalt' tuya	Ploskii, Tuva Republic, Russia	500	7	2.7	0.071	0.185	14.84	6.494	Komatsu et al. (2007)	Several Russian tuyas are degraded and not primary landforms
'Basalt' tuya	Derbi-Taiga, Tuva Republic, Russia	400	6.5	0.3	0.062	1.333	1.53	0.536	Komatsu et al. (2007)	Several Russian tuyas are degraded and not primary landforms
<i>Tuyas, felsic</i>										
1. Lava-flow-dominated										
Evolved tuya	Slag Hill tuya, BC, Canada	90	0.54	0.4	0.167	0.225	0.17	0.010	Kelman (2005), Unpublished thesis, University of Vancouver	
Evolved tuya	Little Ring Mountain, BC, Canada	140	0.9	0.7	0.156	0.200	0.49	0.046	Kelman (2005), Unpublished thesis, University of Vancouver	
Evolved tuya	Ring Mountain, BC, Canada	370	2.3	1.67	0.161	0.222	3.02	0.744	Kelman (2005), Unpublished thesis, University of Vancouver	

Evolved tuya	Upper Cauldron Dome, BC, Canada	600	2.8	2	0.214	0.300	4.40	1.759	Kelman (2005), Unpublished thesis, University of Vancouver	Kelman (2005) says 'internal stratigraphy is unlike a tuya'
Evolved tuya	Table, BC, Canada	250	0.3	0.18	0.833	1.389	0.04	0.012	Kelman et al. (2002)	
2. Tephra-dominated Evolved tuya	Raudafossafjoll, Torfajokull, central Iceland	400	2.5	1.4	0.160	0.286	2.75	0.733	Tuffen et al. (2002a)	Tephra-dominated
Evolved tuya	Ogmundur, Kerlingarfjoll, central Iceland	350	1.3	1.1	0.269	0.318	1.12	0.262	Stevenson et al. (2011)	Rhyolite; tephra-dominated
Evolved tuya	Hottur, Kerlingarfjoll, central Iceland	360	1.6	1.6	0.225	0.225	2.01	0.483	Stevenson et al. (2011)	Rhyolite; tephra-dominated
Evolved tuya	Maenir, Kerlingarfjoll, central Iceland	350	2.2	1.6	0.159	0.219	2.76	0.645	Stevenson et al. (2011)	Rhyolite; tephra-dominated?
Evolved tuya	Lodmundur, Kerlingarfjoll, central Iceland	300	1	1	0.300	0.300	0.79	0.157	Iceland 1:300 000 topographical & geological maps, McGarvie (2009)	Rhyolite; tephra-dominated? Measurements v approx
Evolved tuya	Jorundur, NE Iceland	250	1.9	1	0.132	0.250	1.49	0.249	Iceland 1:300 000 topographical & geological maps, McGarvie (2009)	Rhyolite; tephra-dominated?
Evolved tuya	Hlidarfjall, NE Iceland	250	1.9	1	0.132	0.250	1.49	0.249	Iceland 1:300 000 topographical & geological maps, McGarvie (2009)	Rhyolite; tephra-dominated?
Evolved tuya	Laufafell, Torfajokull, central Iceland	1164	3.6	2.8	0.32	0.416	7.92	6.143	Iceland 1:300 000 topographical & geological maps, McGarvie (2009)	Probably tephra-dominated
Evolved tuya?	Prestahnukur, western Iceland	570	2.5	2	0.23	0.285	3.93	1.492	McGarvie (2009)	Described as a tuya but wholly subglacial (cf. subglacial domes)
<i>Felsic Domes</i> Subglacial dome	Eenostuck Mass, BC, Canada	90	0.8	0.18	0.11	0.500	0.11	0.007	Kelman et al. (2002)	
Subglacial dome	Watts Point, BC, Canada	100	0.7	0.6	0.14	0.167	0.33	0.022	Kelman et al. (2002)	
Subglacial dome	Ember Ridge Northwest, BC, Canada	100	0.69	0.6	0.14	0.167	0.33	0.022	Kelman (2005), Unpublished thesis, University of Vancouver	Values are approximate
Subglacial dome	Ember Ridge North, BC, Canada	100	0.85	0.44	0.12	0.227	0.29	0.020	Kelman (2005), Unpublished thesis, University of Vancouver	Values are approximate
Subglacial dome	Ember Ridge Southwest, BC, Canada	80	1.08	0.72	0.07	0.111	0.61	0.033	Kelman (2005), Unpublished thesis, University of Vancouver	Values are approximate
Subglacial dome	Ember Ridge West, BC, Canada	60	1.08	0.54	0.06	0.111	0.46	0.018	Kelman (2005), Unpublished thesis, University of Vancouver	Values are approximate

(Continued)

Table 2 (Continued)

<i>Feature</i>	<i>Location</i>	<i>Height (m)</i>	<i>Length (km)</i>	<i>Width (km)</i>	<i>Aspect ratio (1) (ht/length)</i>	<i>Aspect ratio (2) (ht/width)</i>	<i>Area (km²)</i>	<i>Volume (km³)</i>	<i>Sources</i>	<i>Comments</i>
Subglacial dome	Ember Ridge Southeast, BC, Canada	60	0.58	0.5	0.10	0.120	0.23	0.009	Kelman (2005), Unpublished thesis, University of Vancouver	Values are approximate
Subglacial dome	Ember Ridge Northeast, BC, Canada	40	0.25	0.19	0.16	0.211	0.04	0.001	Kelman (2005), Unpublished thesis, University of Vancouver	Values are approximate
<i>Sheet-Like Sequences – all mafic</i>										
Subglacial flow (basalt)	Lithofacies assocns A & B, Eyjafjoll, Iceland	10	8	3	0.001	0.003	24.00	0.126	Sue Loughlin, Personal communication, 2003	Probably not topo confined; width said to be 2–3 km, length 7–8 km
Subglacial flow (basalt)	Unidentified lithof assocn(s), Eyjafjoll, Iceland	10	15	2	0.001	0.005	30.00	0.157	Sue Loughlin, Personal communication, 2003	Said to be clearly valley confined; width 1–2 km, length 10–15 km
Subglacial flow (basalt)	Cheakamus Valley, BC, Canada	20	1.6	0.05	0.01	0.400	0.08	0.001	Kelman et al. (2002)	
Subglacial flow (basalt)	Lithofacies assoc J, Eyjafjoll, Iceland	75	12	8	0.01	0.009	96.00	3.770	Sue Loughlin, Personal communication, 2003	Thickness est. 50–75 m; prob fissure source and thicker near vent; prob valley confined
Subglacial flow (basalt)	Lithofacies assoc C, Eyjafjoll, Iceland	40	9	2.5	0.004	0.016	22.50	0.471	Sue Loughlin, Personal communication, 2003	Probably central vent
Subglacial flow (basalt)	Lithofacies assoc C, Eyjafjoll, Iceland	35	12	2.5	0.003	0.014	30.00	0.550	Sue Loughlin, Personal communication, 2003	Thickness said to be ‘<40 m’; two units
Subglacial flow (basalt)	Dalsheidi, SE Iceland	250	35	1.5	0.01	0.167	52.50	6.872	Walker and Blake (1966)	
Subglacial flow (basalt)	Skaftartunga, E Myrdalsjokull, S Iceland	250	25	5	0.01	0.050	125.00	16.362	K Saemundsson, Personal communication, 2003	
Subglacial flow (basalt)	E of Katla (Kriki hyaloclastite flow), Iceland	100	7	4	0.01	0.025	28.00	1.466	Larsen (2002) and Gudrun Larsen, Personal communication, 2003	
Subglacial flow (basalt)	Unit 3; Sida District, southern Iceland	100	17	14	0.01	0.007	238.00	12.462	Bergh (1985)	Dimensions assume correlated units are correctly mapped (not verified); precise source location unknown assumed fissure just to N; thickness is mean value

Subglacial flow (basalt)	Unit 3a; Sida District, southern Iceland	150	15	12	0.01	0.013	180.00	14.137	Bergh (1985)	Dimensions assume correlated units are correctly mapped (not verified); precise source location unknown assumed fissure just to <i>N</i> ; thickness is mean value
Subglacial flow (basalt)	Unit 5; Sida & Fjotshverfi districts, southern Iceland	150	30	10	0.01	0.015	300.00	23.562	Bergh (1985)	Dimensions assume correlated units are correctly mapped (not verified); precise source location unknown assumed fissure just to <i>N</i> ; thickness is mean value
Subglacial flow (basalt)	Unit 7; Sida District, southern Iceland	150	18	12	0.01	0.013	216.00	16.965	Bergh (1985)	Dimensions assume correlated units are correctly mapped (not verified); precise source location unknown assumed fissure just to <i>N</i> ; thickness is mean value
Subglacial flow (basalt)	Unit 3c; Fjotshverfi District, southern Iceland	50	14	10	0.004	0.005	140.00	3.665	Bergh (1985)	Dimensions assume correlated units are correctly mapped (not verified); precise source location unknown assumed fissure just to <i>N</i> ; thickness is mean value
<i>Tephra Mounds/Ridges – mafic</i>										
Tindar (basaltic)	Kalfstindar(E), SW Iceland	250	8.5	1.2	0.03	0.208	8.01	1.252	Iceland 1:300 000 topographical & geological maps	Also Werner and Schmincke (1999) , but wrongly described as pillow lava/breccia
Tindar (basaltic)	Kalfstindar(W), SW Iceland	200	6	1.5	0.03	0.133	7.07	0.832	Iceland 1:300 000 topographical & geological maps	Also Werner and Schmincke (1999) , but wrongly described as pillow lava/breccia
Tindar (basaltic)	Tindaskagi, SW Iceland	300	7.5	1.5	0.04	0.200	8.84	1.586	Iceland 1:300 000 topographical & geological maps	Also Werner and Schmincke (1999) ,
<i>Tindar (basaltic)</i>	<i>Herdubreidartogl, Iceland</i>	<i>500</i>	<i>5</i>	<i>4.5</i>	<i>0.10</i>	<i>0.111</i>	<i>17.67</i>	<i>1.741</i>	Van Bemmelen and Rutten (1955)	May be lacustrine not glacial (Werner et al., 1996)
Tindar (basaltic)	Vetari-Skogarmannafjoll, Iceland	200	9	1.2	0.02	0.167	8.48	1.062	Van Bemmelen and Rutten (1955)	
Tindar (basaltic)	Austarii-Skogarmannafjoll, Iceland	150	4	0.8	0.04	0.188	2.51	0.255	Van Bemmelen and Rutten (1955)	
Tindar (basaltic)	Storihnukur-Hvammfell, Iceland	300	4.5	1.2	0.07	0.250	4.24	0.782	Van Bemmelen and Rutten (1955)	
<i>Tindar (basaltic)</i>	<i>Ash Mountain, BC, Canada</i>	<i>700</i>	<i>5</i>	<i>4</i>	<i>0.14</i>	<i>0.175</i>	<i>15.71</i>	<i>2.866</i>		<i>Includes pillow lava basal pediment (broadens edifice significantly)</i>
Tindar (basaltic)	Ash Mountain, BC, Canada	400	2	2	0.20	0.200	3.14	0.419	Moore et al. (1995)	Excludes pillow lava basal pediment (tuff cone only)

(Continued)

Table 2 (Continued)

<i>Feature</i>	<i>Location</i>	<i>Height (m)</i>	<i>Length (km)</i>	<i>Width (km)</i>	<i>Aspect ratio (1) (ht/length)</i>	<i>Aspect ratio (2) (ht/width)</i>	<i>Area (km²)</i>	<i>Volume (km³)</i>	<i>Sources</i>	<i>Comments</i>
Tindar? (basaltic)	Spanish Mump, BC, Canada	150	3.3	2	0.05	0.075	5.18	0.352	Hickson et al. (1995)	
Tindar (basaltic)	Pyramid Mountain, BC, Canada	240	1.8	1.3	0.13	0.185	1.84	0.241	Hickson et al. (1995)	
Tindar (basaltic)	Helgafell, SW Iceland	220	1.78	0.93	0.12	0.237	1.30	0.194	Schopka et al. (2006)	
Tindar (basaltic)	Valahnjukar, SW Iceland	80	1.43	0.57	0.06	0.140	0.64	0.043	Schopka et al. (2006)	
'Basalt' tindar	Albine-Boldok, Tuva Republic, Russia	200	1	0.7	0.20	0.286	0.55	0.094	Komatsu et al. (2007)	
'Basalt' tindar	Charash-Dag, Tuva Republic, Russia	400	2	1.5	0.20	0.267	2.36	0.464	Komatsu et al. (2007)	
'Basalt' tindar	Sagan, Tuva Republic, Russia	400	6.5	0.3	0.06	1.333	1.53	0.435	Komatsu et al. (2007)	
<i>Tephra Mounds/Ridges – felsic</i>										
Tindar (evolved)	Rodull, Kerlingarfjoll, central Iceland	180	2	0.7	0.09	0.257	1.10	0.148	J. Smellie, Unpublished information	Rhyolitic
Tindar ('evolved')	Gjalp, Vatnajokull, Iceland; centre edifice	450	7	1	0.06	0.450	5.50	1.586	Gudmundsson et al. (2002)	Basaltic icelandite with v low MgO (3%), hence c. "felsic"; 7 km is length of whole fissure
Tindar ('evolved')	Gjalp, Vatnajokull, Iceland; southern edifice	250	n/a	0.5		0.500			Gudmundsson et al. (2002)	Basaltic icelandite with very low MgO (3%), hence c. "felsic"
<i>Pillow Volcanoes – mafic</i>										
Pillow volcano?	Gjalp, Vatnajokull, Iceland; northern edifice	200		2		0.100			Gudmundsson et al. (2002)	May not be pillow volcano
Pillow sheet	Kaldakvisl Fmn, central Iceland	10	10	2	0.001	0.005	15.71	0.209	S. Snorrason and E. Vilmundardotir, Personal communication, 2003	
Pillow sheet	Miklaglijufur Fmn, central Iceland	50	5	1	0.01	0.050	3.93	0.262	S. Snorrason and E. Vilmundardotir, Personal communication, 2003	
Pillow sheet	Launoldur Fmn, central Iceland	100	15	7	0.01	0.014	82.47	10.996	S. Snorrason and E. Vilmundardotir, Personal communication, 2003	Largest pillow sheet known; thickness unclear, variably 30–150–200 m?

Data are for monogenetic volcanoes only.

Heights are edifice heights, i.e. above surrounding pre-volcanic basement. They are likely maxima.

Tuya surfaces, in particular, may be several tens of meters lower.

Less reliable data (mainly from topographical maps) shown in italics.

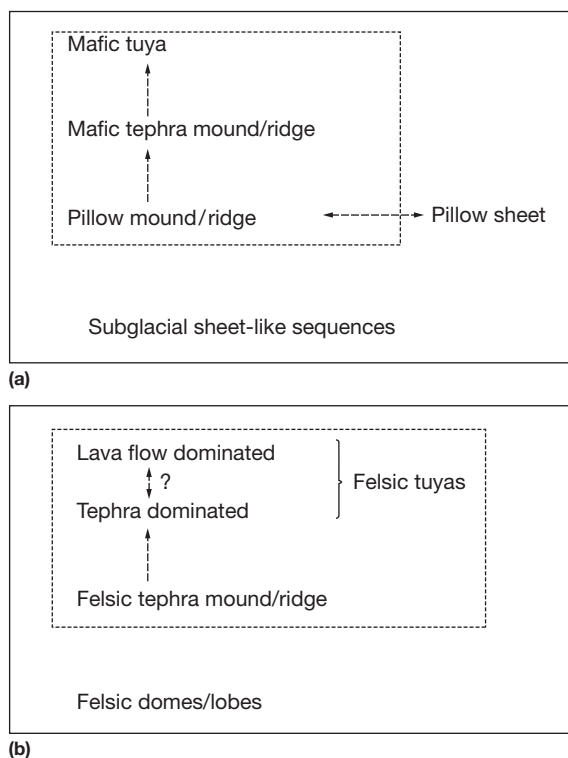


Figure 2 Hierarchy of subglacial volcanic landforms. The dashed arrows indicate possible evolutionary trends between landforms that are commonly related in a single edifice (within dashed-line boxes). With the possible exception of mafic tuyas, each landform can exist without a different precursor landform and some probably always exist in isolation (e.g., felsic domes/lobes).

volcano-determined dimension. The division into discrete fields for each of the different types of volcano morphologies described in this paper, identified by Smellie (2007) using a smaller dataset, are generally not evident in the larger dataset now available. Tuya landforms in subfield 'a' (Figure 7(a)) reflect central-vent (i.e., point source) eruptions and are relatively symmetrical outcrops, while those in subfield 'b' are elongate fissure-erupted tuyas. Intuitively, central-vent eruptions on a sloping bedrock surface might also lead to asymmetry in the resulting tuya as the deltas will flow preferentially downhill and distort the edifice (cf. Smellie et al., 2011b, Figure 4).

In Figure 7(b), aspect ratio is plotted against width, defined as the *minimum* horizontal extent. Because the selected widths are perpendicular to fissure length, they reflect volcano construction processes rather than tectonics. The diagram illustrates that there might be a difference in aspect ratio between mafic and felsic subglacial landforms, with most mafic centers, of any morphological type, characterized by lower ratios ($< \sim 0.23$). There is significant overlap in values, however, although some overlap is probably due to post-eruptive erosional effects modifying the pristine landforms and changing their aspect ratios (especially likely for Russian mafic tuyas and tuff cones; see Komatsu et al., 2007). A subordinate subfield of much wider mafic tuyas is also present in Figure 7(b) (subfield 'b'). The simplest explanation for subfield 'b' in Figure 7(b) is that those edifices were constructed during unusually voluminous

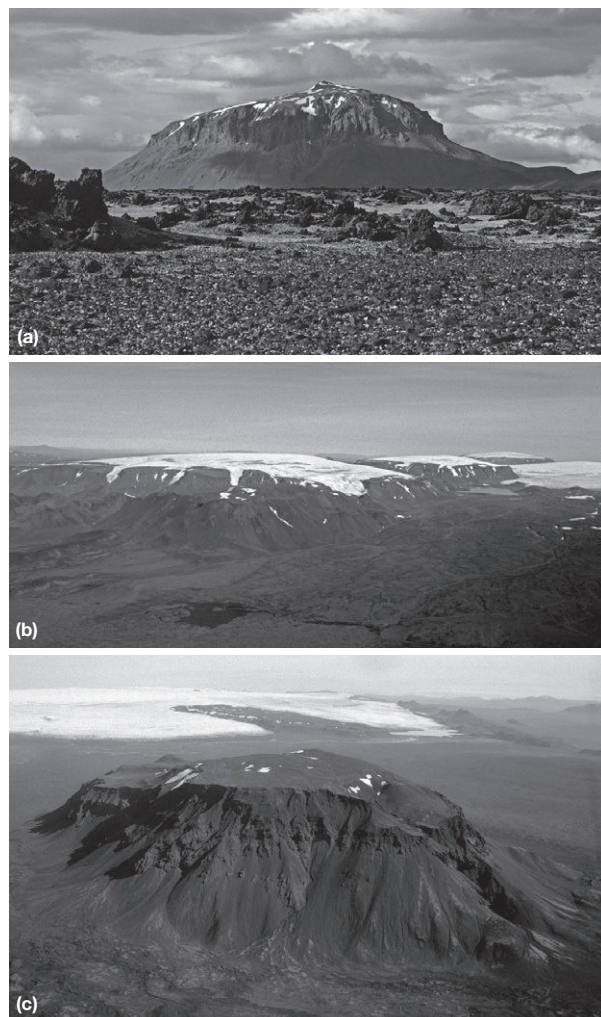


Figure 3 Examples of mafic tuyas. (a) Herdubreid (Iceland) viewed looking north. It rises about 800 m and is probably the best-known mafic tuya. (b) Three large mafic tuyas in Iceland (Thorisjökull, Geitlansjökull, and Eiriksjökull, from left to right) exhumed by the postglacial retreat of the Langjökull ice cap. Thorisjökull tuya is about 600 m high and 10 km wide. The approximately coincident summit elevations suggest that they erupted through the same ice sheet and thus might be of similar age. (c) Hlödufell, a mafic tuya 1200 m high in Iceland. This tuya shows two flat-lying summit surfaces indicating that the edifice was constructed in stages and might be polygenetic (see Skilling, 2009). Note the strong morphological contrast with the narrow ridge (Jahrhettur; a tindar) with multiple peaks, formed of subglacially erupted mafic tephra, seen emerging from the Langjökull icefield in the right background.

eruptions, when the associated lava-fed deltas prograded laterally to a greater extent than tuyas in subfield 'a'. If true, both subfields in Figure 7(b) should in reality be continuous. The separation may be an artifact of the currently available data set, which is still relatively small, and the two fields will probably merge as more data become available.

As stated before, the largest monogenetic mafic tuyas measured so far are 900–1000 m in height, recording eruptions through contemporaneous ice sheets of comparable thickness and those heights are within the range of maximum heights (750–1000 m) predicted for tuyas erupted in wet-based ice.

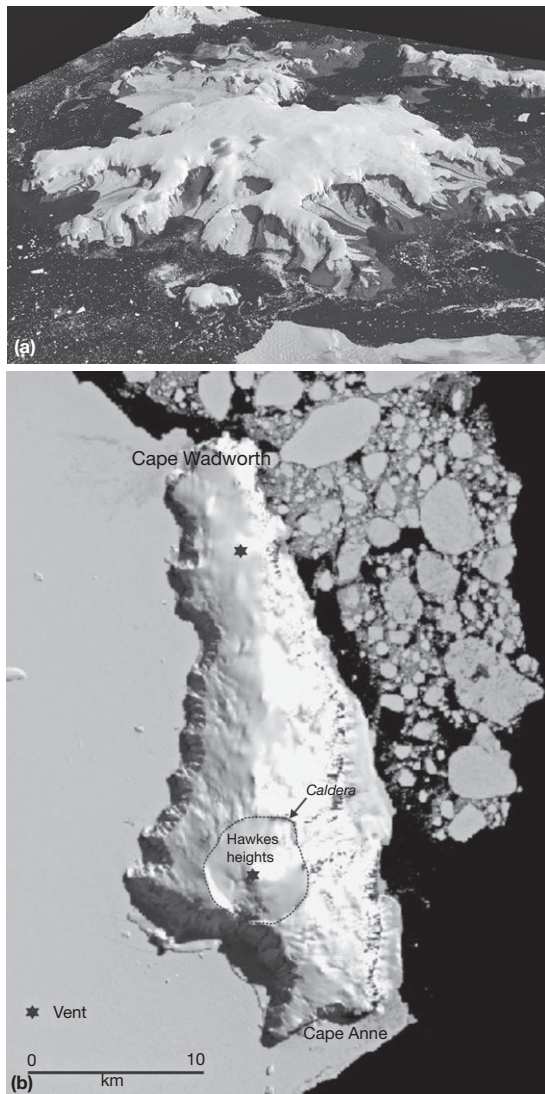


Figure 4 Examples of large subglacially erupted polygenetic volcanoes. (a) Synthetic view of James Ross Island (Antarctica) viewed looking northwest. This is the world's largest-known subglacially erupted basalt shield volcano, about 60 km in diameter and 1.6 km high. It is constructed from multiple superimposed lava-fed deltas that are exposed in tall encircling cliffs. The cliffs formed by major sector collapses and subsequent glacial scouring. ASTER image draped on a DEM. Approximately three times vertical exaggeration. (b) Satellite view of Coulman Island, northern Victoria Land, Antarctica. The volcano margins are now cliffs several hundred meters high formed probably mainly by extensive marine erosion. Although a glaciovolcanic edifice, like Mt Haddington, it resembles other, nonglacial shield volcanoes and a glaciovolcanic origin for both volcanoes can be deduced only from a study of the lithofacies (Smellie et al., 2008, 2011a).

By contrast, the James Ross Island and northern Victoria Land subglacial polygenetic shield volcanoes are considerably taller (1.5–2 km), indicating that the theoretical height limit on mafic tuyas does not apply to polygenetic centers. This is because of the ability of an ice cover to reform on a stratovolcano soon after the cessation of each eruptive phase. Thus each effusive eruption on a shield (or strato-) volcano records the

thickness of the coeval ice draping the volcano, and it may always be much less than 750–1000 m even though the volcano grows to a greater height. This was the case on James Ross Island (1.6 km summit), where the ice sheet never exceeded 850 m (and was typically much less) and it was ≤ 300 m thick on the late Miocene northern Victoria Land polygenetic shields, each of which were constructed to summit elevations of 2000 m (Smellie et al., 2008, 2009, 2011b).

Felsic Tuyas

Some subglacial flat-topped, steep-sided volcanic landforms with felsic compositions have been described as tuyas (Kelman et al., 2002; McGarvie, 2009; Tuffen et al., 2002b). Compared to mafic tuyas, very few felsic examples have been described. Their compositions include andesite, dacite, and rhyolite. Apart from a trachyte example observed by the author on Ross Island, Antarctica, and rare examples with tephriphonolite and mugearite compositions in northern Victoria Land (also Antarctica; Smellie et al., 2011a), it appears that felsic magmas generally do not form lava-fed deltas (necessary for a mafic tuya, above). The internal structure of felsic tuyas is also very different from that of mafic tuyas although they share common characteristics of a flat-lying caprock of subaerial lava overlying subaqueous lithofacies (water-erupted/deposited tephra or water (or ice)-chilled lava; e.g., Tuffen et al., 2002b, Figure 3; Smellie, 2009; Stevenson et al., 2011).

At least two types of felsic tuyas are identified based on their internal structure: (1) tephra-dominated and (2) lava flow dominated. In tephra-dominated felsic tuyas, the flat cap is formed of horizontal or gently dipping, relatively thin (typically a few tens of meters) subaerially emplaced felsic lavas, which overlie a much thicker pile (few hundred meters) of fine-grained tephra (Figure 8). The tephra core is usually obscured by lava scree and is therefore poorly exposed (Stevenson et al., 2011; Tuffen et al., 2002b). By contrast, the lava flow dominated felsic tuyas are distinctive flat-topped columns or short blade-like ridges, also flat-topped. They are formed almost entirely of subhorizontal ice-impounded subaerial felsic lavas (Figure 9), although some of the lavas may also be plastered onto the subvertical tuya flanks (Mathews, 1951). Andesite-dacite lava-dominated systems have mainly been described so far, and fragmental deposits are typically absent or scarce (Kelman et al., 2002; Mathews, 1951). Like mafic tuyas, the lava caprock in both types of felsic tuyas formed after the ice sheet had melted through completely, apart from a unique rhyolitic example described by McGarvie et al. (2007) at Prestahnukur, Iceland, in which the edifice may have developed without the coeval ice carapace being melted through. Moreover, it has a less well-developed flat caprock than occurs in undoubtedly subaerial felsic tuyas. These characteristics are more similar to exogenous subglacial felsic domes (see below). However, other features such as the presence of a 300-m-thick basal pedestal of volcanoclastics (perlitized lobe-bearing breccias rather than tephra) and a much higher volume:area ratio than felsic domes, are possibly more akin to felsic tuyas. The Prestahnukur example illustrates that some glaciovolcanic outcrops can be hard to classify within the range of morphometric types currently recognized.

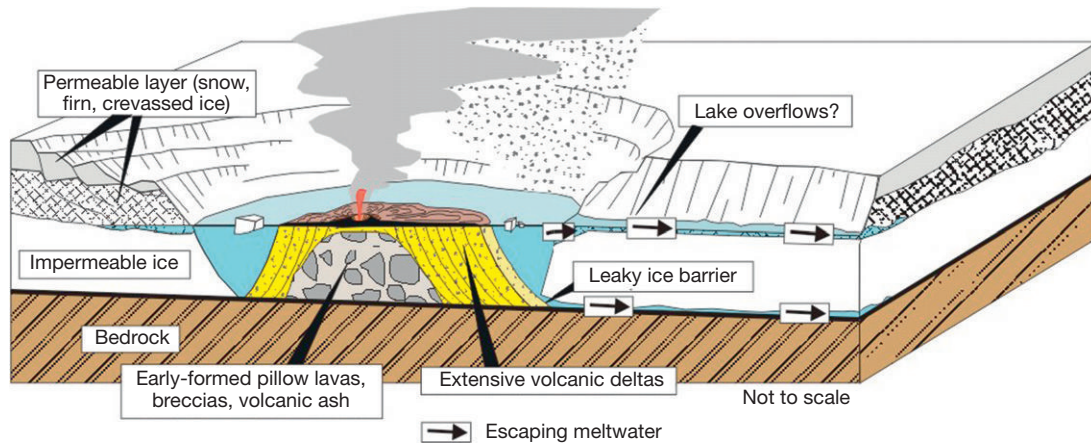


Figure 5 Cartoon of a monogenetic mafic tuya forming in a wet-based ice sheet, illustrating the internal organization of the principal lithofacies that construct the tuya and possible main meltwater discharge routes. The thickness of the overlying permeable layer is exaggerated, for clarity.

There are few data for felsic tuyas so it is difficult to be didactic about their morphometry. The data hint that all felsic tuyas may share similar (high) volume:area ratios to the mafic tuyas (Figure 6(b)). In Figure 7, it is clear that felsic tuyas have slightly to markedly higher aspect ratios than mafic tuyas, and are comparable with those for subglacial felsic domes. Together, they have among the highest aspect ratios of most subglacial volcanic landforms measured so far.

Pillow Mounds, Ridges, and Sheets

Subglacial pillow mounds, ridges, and sheets are formed mainly of pillow lava and less common hyaloclastite breccia. All described examples are basaltic in composition, but there is no intrinsic reason why more andesitic and even dacitic magmas should not form pillows (e.g., Stevenson et al., 2009) although the different rheologies of the more evolved magmas may affect the final edifice dimensions and, therefore, morphometry. Many of these landforms occur in the cores of basaltic tuyas and are therefore largely obscured from view, where pillow lava effusion is usually regarded as a ubiquitous early stage of tuya construction (e.g., Jones, 1969; Smellie and Hole, 1997; Werner and Schmincke, 1999). The landforms include (i) relatively low oblate mounds with moderately dipping sides and smooth profiles in examples erupted from central vents; (ii) subdued ridges with gentle slopes that were erupted from fissures; and (iii) extensive thin sheets (Figures 10 and 11). Examples include the subglacial pillow mounds in British Columbia examined by Dixon et al. (2002), which are formed solely of pillow lava extruded under ice sheets that may have been >2 km thick, and pillow-lava centers draped unconformably by agglutinate at Seal Nunataks, Antarctica, which formed within an ice sheet >600 m thick (Smellie and Hole, 1997). By contrast, pillow sheets are only a few tens of meters thick, yet may extend laterally 10 km or more (Snorrason and Vilmundardóttir, 2000, and unpublished). Pillow sheets are currently only known from Iceland and have not yet been described in detail. Their formation, their internal

structure, and even whether they are dominantly formed of pillow lava, are unclear at present.

Because they have not been studied in detail, there are currently very few morphometric data available for subglacial pillow mounds, ridges, and sheets (Figures 6 and 7). They generally have among the lowest volume:area ratios in the data set, although the data are biased toward the sheet-like examples, which will always have relatively high outcrop areas. Aspect ratios are also very low, comparable only to subglacial sheet-like sequences (Figure 7), and subject to the same caveat.

Tephra Mounds and Ridges

Tephra mounds and ridges have been described for subglacial volcanoes of mafic and felsic compositions. They vary from conical edifices or mounds to linear ridges. Mafic examples are by far the commonest. They are comparatively steep-sided and, when erupted from fissures, form prominent ridges with multiple summit peaks (Figures 10 and 12). The ridge-like occurrences were called *tindars* by Jones (1969; also adopted by Hickson, 2000), an etymologically incorrect use of the Icelandic plural word (*tindur*) meaning pinnacles or summits (singular *tindur*). Its usage was extended informally by Smellie (2000) to encompass indiscriminately any subglacially erupted tephra piles, whether in mounds or ridges. A 'tindar phase' of subaqueous explosive eruptive activity would thus relate to the tephra-dominated cores of edifices that subsequently developed into tuyas. It may be preferable to refer to the landforms nongenetically as either tephra mounds or ridges according to their outcrop morphology (specifying subglacial to indicate eruptive environment). However, *tindar* should be restricted to occurrences (either ridges only, or else both ridges and mounds if a wider definition is preferred) for which a subglacial eruptive origin can be demonstrated. Since they form due to the explosive interaction of magma with water (meltwater in the case of glaciovolcanic eruptions), those erupted from

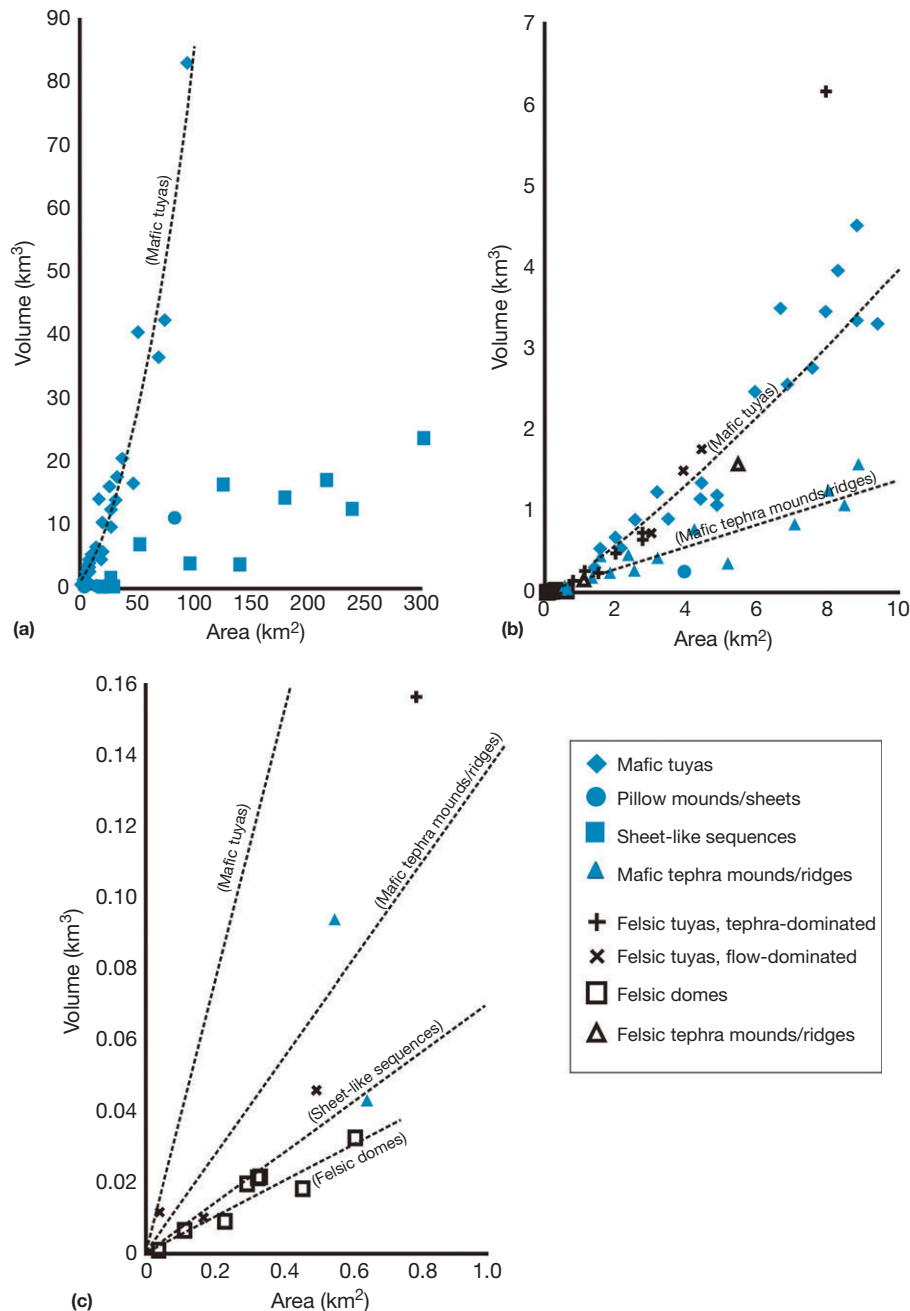


Figure 6 Morphometric data for subglacial volcanic landforms: volume versus area.

central vents may also be described nongenetically as simple tuff cones rather than tephra mounds.

The structure and lithofacies of felsic tephra mounds and ridges are not well described so far, except where they occur below a subaerial lava cap in tephra-dominated felsic (rhyolite) tuyas (Stevenson et al., 2011; Tuffen et al., 2002b). However, they also frequently occur without a lava caprock (e.g., in the Kerlingarfjöll rhyolite complex, central Iceland; Figure 13). They form relatively steep-sided pale-colored successions, commonly with numerous, sometimes voluminous craggy syneruptive felsic intrusions (Stevenson et al., 2011; Tuffen et al., 2007, 2009).

The tephra mounds/ridges have much lower volume:area ratios than the mafic and felsic tuyas, with ratios only lower in the subglacial sheet-like sequences and felsic domes (Figure 6(b) and 6(c)). They also extend to lower aspect ratios than the mafic tuyas on the length versus height diagram, which emphasizes the fissure-related aspects of the tephra landforms (Figure 7(a)). Most of the felsic tephra mounds also have higher aspect ratios than mafic examples, although there are too few data to be definitive (Figure 7(a) and 7(b)). In the width versus height diagram (Figure 7(b)), the tephra mounds/ridges show similar aspect ratios to compositionally analogous tuyas: that is, mafic tephra mounds/ridges with

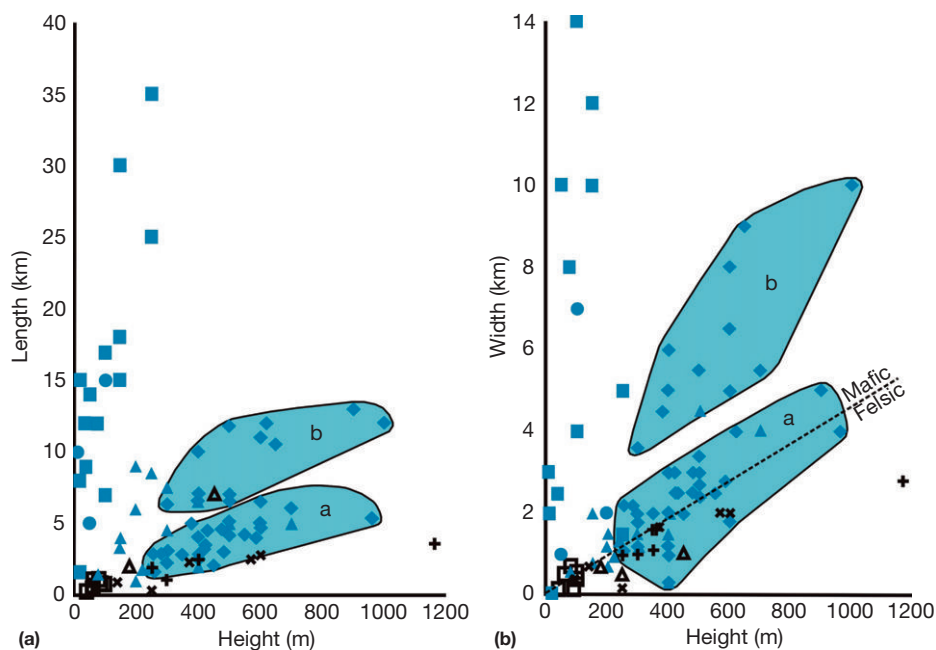


Figure 7 Morphometric data for subglacial volcanic landforms: (a) height versus length; and (b) height versus width (aspect ratio). Symbols as in Figure 6.

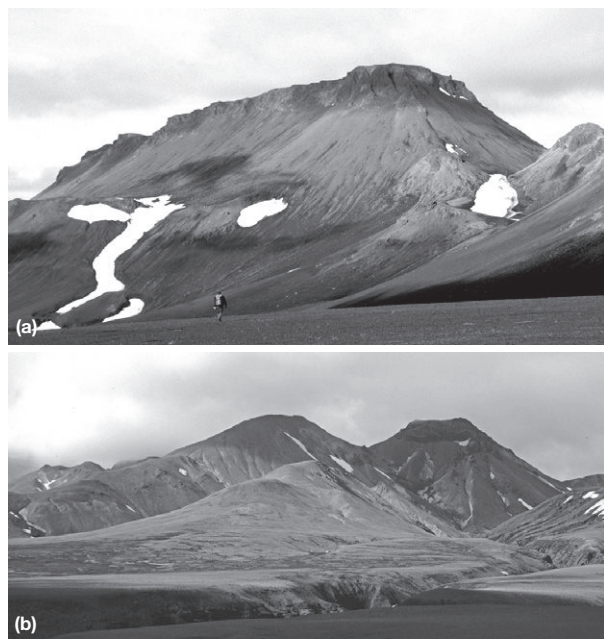


Figure 8 Examples of tephra-dominated felsic tuyas in Iceland. (a) Raudufossafjöll. The summit rises approximately 400 m; (b) Hottur (at right) and unnamed peak, Kerlingarfjöll. Both peaks rise approximately 350 m.

mafic tuyas, and (less well defined) felsic mounds/ridges with felsic tuyas. However, the two landform categories (mounds vs. tuyas) are morphologically very different, with the flat tops of tuyas being the most prominent difference.

The distribution of subglacial mafic tephra mound data overlaps greatly with data for subaqueous mafic tuff cones erupted in unconfined, nonglacial settings, although displaced somewhat to the higher aspect ratio side (Figure 14(a)); subglacial felsic examples have higher aspect ratios than their nonglacial counterparts, but both felsic data sets are still very small. Felsic tephra mounds generally have greater volume:area ratios than their mafic counterparts, irrespective of a glacial or nonglacial setting (Figure 14(b)). The degree of overlapping data for mafic glacial/nonglacial mounds/cones is greater than anticipated since the subglacial examples are predicted intuitively to be generally taller as a consequence of a combination of effects (Smellie, 2009), including:

1. Buttressing of tephra by the confining ice walls (illustrated by Smellie, 2000, Figure 5).
2. Rapid early diagenesis and lithification of the warm wet tephra pile, which may be buffered at several tens of degrees celsius for a substantial period (Gudmundsson et al., 2004), particularly by coeval high-level intrusions.
3. Glacial erosion and oversteepening of the indurated volcano flanks once activity has ceased (Figure 12).

Examination of subglacial examples in Iceland seems to support this prediction. Those tephra landforms shown in Figures 10 and 12 are characterized by steep flanks that frequently cut across stratification in the indurated tephra, which contrasts with associated (underlying) coeval pillow mounds/ridges. By contrast, subaqueous tuff cones formed in unconfined marine or (nonglacial) lake settings generally have lower-gradient asymptotic flanks with surfaces parallel to bedding rather than cross-cutting, because of the long run-out distances possible in unconfined settings (Figure 15). A possible explanation may be that the measured widths of the subglacial

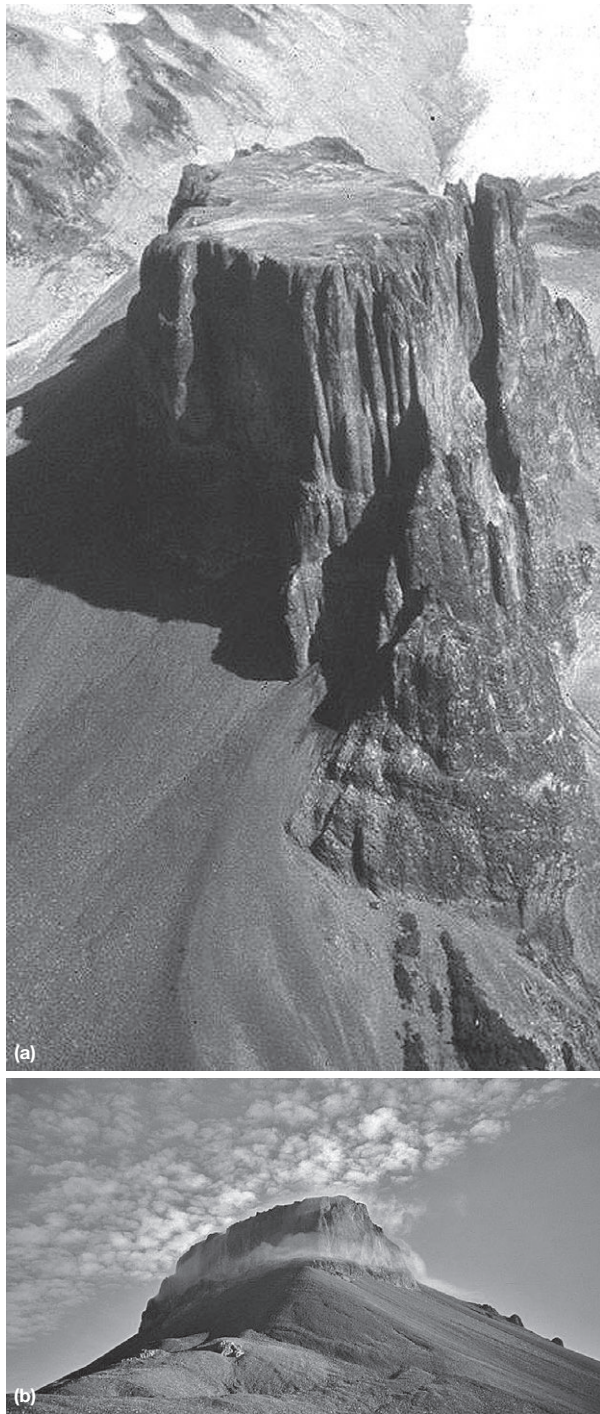


Figure 9 Examples of lava flow-dominated felsic tuyas in British Columbia, Canada. (a) The table (approximately 300 m); (b) Little Ring Mountain (approximately 280 m). Photograph (a) by Cathie Hickson and photograph (b) by Melanie Kelman.

examples are incorrect and inadvertently combine other landforms. Since most of the subglacial examples used in the dataset were culled from the literature, this may come about if the tephra mounds/ridges were constructed on pillow lava mounds (with low aspect ratios) if (1) the pillow lava piles

(commonly obscured by scree) were incorrectly mapped as tephra, or (2) tephra incompletely drapes the underlying, much lower-profile, pillow mounds and incorporates their form, thus creating a misleading composite landform, not one constructed solely by tephra.

Felsic Lobes and Domes

Several steep-sided subglacially erupted andesite, dacite, rhyolite, and trachyte–phonolite lavas with characteristic fine-scale prismatic jointing have been described from British Columbia and Iceland (Edwards and Russell, 2002; Kelman et al., 2002; Tuffen et al., 2002a,b). The radiating joint patterns and associated glassy chilled surfaces are interpreted as ice-contact features, within a subglacial vault. Together they define the original landform shapes, which may be conical, lobate, dome-like, or irregular (Figure 16). Associated fragmental volcanic lithofacies are usually uncommon.

Meltout tills and reworked volcanic sands and gravels are present in an Icelandic example together with ashy-matrix breccias.

The felsic lobes and domes have low volumes relative to their areas (Figure 6(c)) and they appear to have somewhat higher aspect ratios than many of the mafic subglacial volcanic landforms (Figure 7). Conversely, their aspect ratios are effectively indistinguishable from those of most other felsic landforms (i.e., felsic tuyas and tephra mounds).

Subglacial Sheet-Like Sequences

These were also called subglacial ‘sheet flow’ sequences by Smellie (2001) but were renamed subglacial sheet-like sequences by Smellie (2008) to incorporate nongenetic terminology. They are relatively thin deposits, which, as their name suggests, either form narrow, sinuous, anastomosing low outcrops, which are sometimes esker-like (Mathews, 1958; Smellie and Skilling, 1994), or else lobe- or sheet-like features (Smellie, 2008). However, the profile, overall shape, and dimensions of an outcrop will depend on the topography of the underlying bedrock (e.g., incised valley), erupted volume of magma and discharge rate (Loughlin, 2002). Most of the published examples are basaltic, but Edwards and Russell (2002) have described a few trachyte and phonolite examples in British Columbia and there are scarce rhyolite occurrences in Antarctica (Smellie et al., 2011a).

At least two different types of sheet-like sequence can be distinguished, referred to as Mt Pinafore and Dalsheidi types (Smellie, 2008). Their distinction is based on comparatively subtle differences in lithofacies and sequence architecture that have a large impact on their inferred modes of emplacement. It is unclear yet if those differences impact significantly on the resultant landforms and they are provisionally considered as a single landform group here. The differences have important implications for interpreting original ice sheet thicknesses (Smellie, 2008). Unlike all the other subglacial volcanic landforms described in this article, the sheet-like sequences are strata formed away from the erupting vent(s) (i.e., they are ‘outflow facies’) rather than

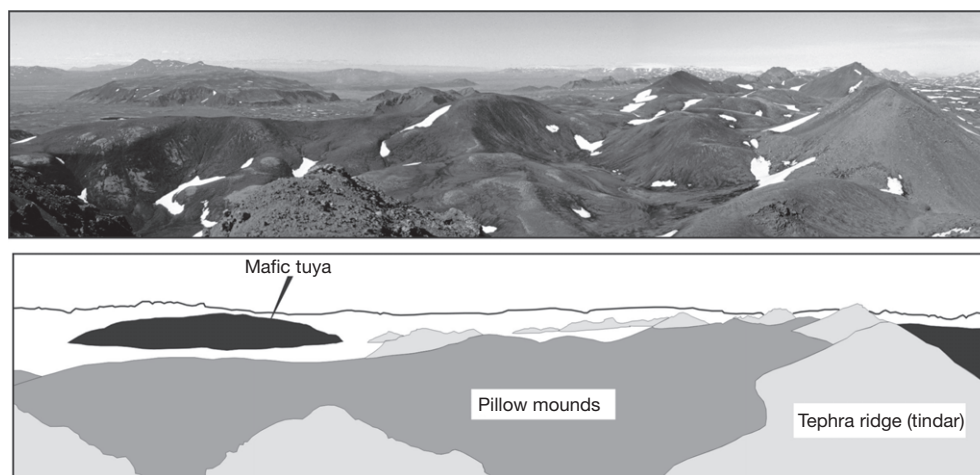


Figure 10 View and explanatory sketch, looking north from Kalfstindar (Iceland) showing contrasting subglacial mafic volcanic landforms, including steep tephra ridges, a tuya, and low-lying smooth-surfaced pillow mounds.

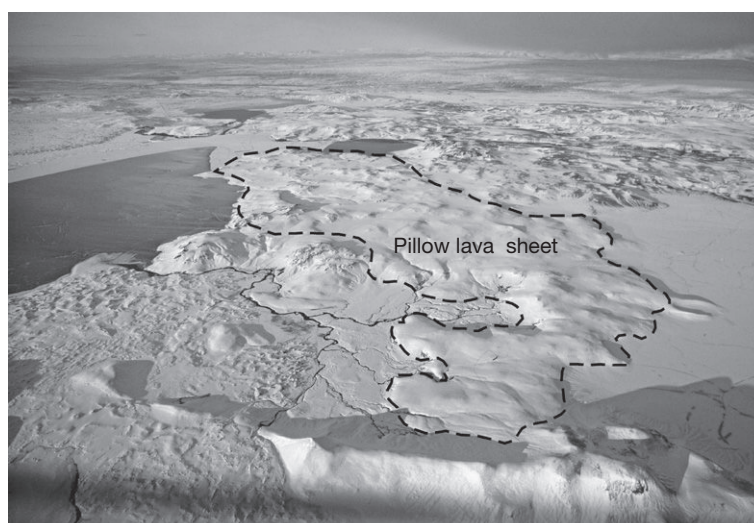


Figure 11 Extensive pillow sheet (dashed-line outcrop) at Thorisvatn, Iceland. The prominent ridge in the foreground is a pillow ridge. Photograph by Snorri Snorrason.



Figure 12 View looking south along the pinnacled ridge crest of Kalfstindar, a mafic tephra ridge in Iceland. Note the very steep sides of the ridge, which cut across tephra bedding in the deposit. The ridge rises approximately 250 m above the surrounding plain. The low subdued mounds in the foreground at right are pillow mounds.

volcanic edifices. The few source vents identified so far are simple pyroclastic cones, undiagnostic, in themselves, of a subglacial setting. The full range of source edifices has probably not been identified.

Subglacial sheet-like sequences range from only a few tens of meters thick to a few hundred meters, with widths up to several kilometers. A wide range of examples has been described in Iceland, Antarctica, and British Columbia (e.g., Kelman et al., 2002; Loughlin, 2002; Smellie, 2008; Smellie et al., 1993, 2011a). There are no published whole-outcrop-scale photographs of the sheet-like sequences since the majority described so far form multiple stacked units that together construct large stratovolcanoes. The morphology is unlikely to be sufficiently distinctive to be identified as glaciovolcanic without detailed lithofacies studies.

Not surprisingly, the sheet-like sequences have the lowest aspect ratios of any subglacial volcanic landform (other than subglacial pillow sheets), and they have a distinctive combination of very high area combined with low volume which,

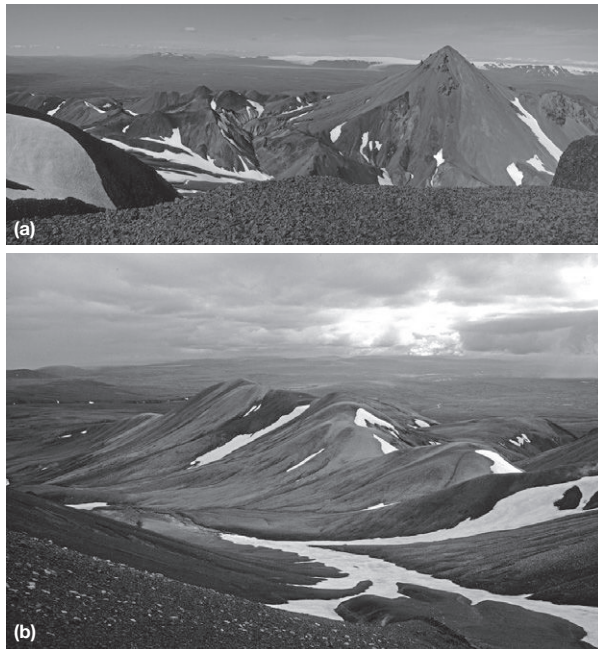


Figure 13 Examples of felsic tephra mounds/ridges at Kerlingarfjoll, Iceland. (a) Looking west to Ogmundur from Hottur; the low mounded ridges at center-left are tephra mounds located along fissures. Ogmundur, at right, is a lava flow-capped felsic tuya that has been eroded to form the present prominent conical peak; (b) subdued, fissure-aligned, multiple overlapping felsic tephra mounds; Rodull.

however, is only slightly different from area:volume ratios for subglacial lobes and domes (Figures 6(c) and 7).

Influence of Glacier Thermal Regime on Subglacial Volcanic Landforms

All of the foregoing is based on an analysis of subglacial, mainly monogenetic, volcanic landforms that were created by eruptions beneath glaciers and/or ice sheets with wet-based (i.e., temperate or subpolar) thermal regimes, apart from aa lava-fed deltas that constructed several large polygenetic shield volcanoes under cold-based (i.e., polar) ice in northern Victoria Land but which probably had only a small influence on their morphometry (Smellie et al., 2011a). Thermal regime is important. It determines the hydraulics of an ice system, which, in turn, strongly influences the lithofacies, sequence architecture and, thus, the landforms ultimately constructed (e.g., Björnsson, 1988; Smellie, 2000, 2001, 2009). Wet-based glaciers are dominant on Earth today, except in parts of the Greenland and Antarctic ice sheets, so the results of the landform analysis in this article are likely to have wide application. There are different consequences for volcanic landforms constructed within a cold-based ice sheet or glacier (i.e., one that is frozen to its bed; Smellie, 2009). In those, water cannot escape subglacially, either in conduits or thin films. Any meltwater created by a subglacial eruption in cold-based ice will therefore be confined in a vault overlying the vent. The system will

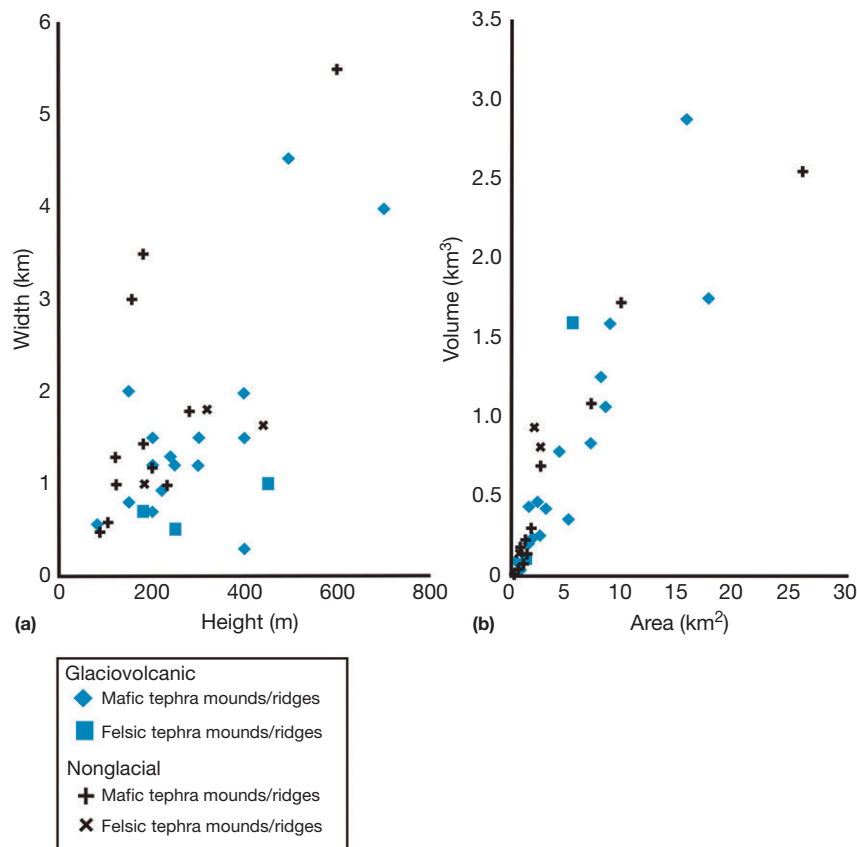


Figure 14 Morphometric data for subglacial and nonglacial tephra mounds/ridges: (a) width versus height; and (b) volume versus area.

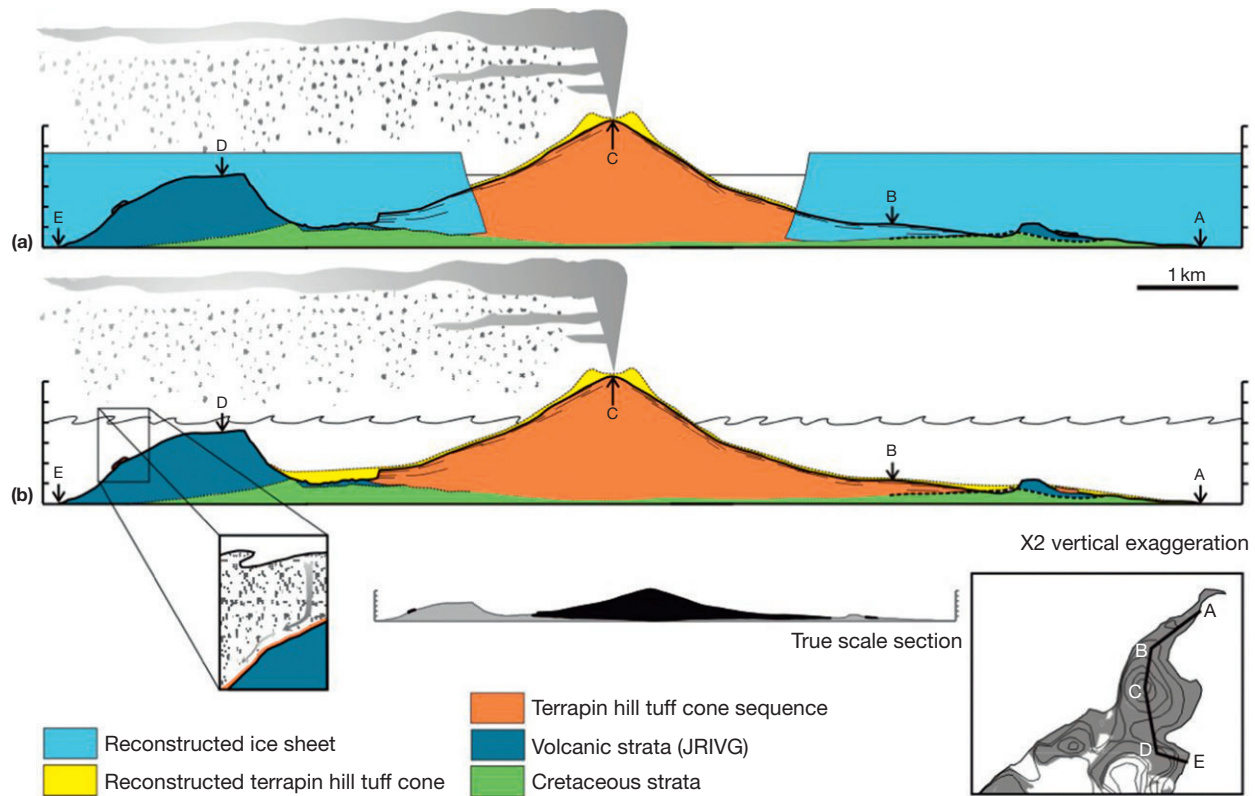


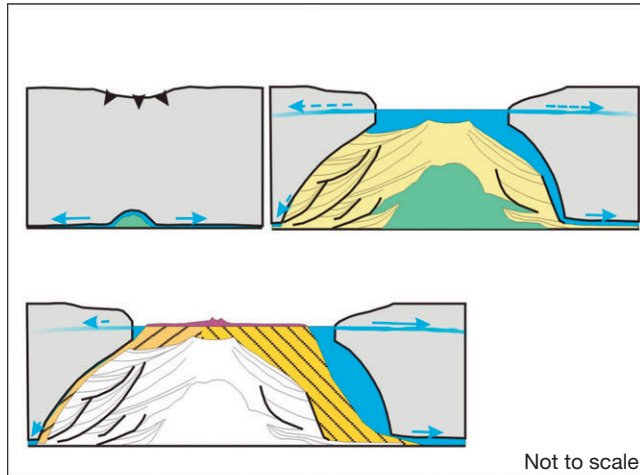
Figure 15 Sketch sections showing the impact of eruptive environment on the morphology of a large tuff cone. (a) An ice sheet is reconstructed. The eruption might have melted a space 3 km in diameter, as sketched. The thermal energy would have been focused in the central parts of the edifice (high heat flow in the vent(s)) and the high thermal inertia of the tephra pile would likely have prevented the melting of a significantly larger space during the eruption. The proximity of an ice sheet would have impinged on the growth of the edifice by buttressing effects of the ice walls (not illustrated), thus leading to an edifice with much steeper flanks, which is not the case. (b) An unconfined marine setting is shown, which is more consistent with the outcrop and can easily explain the presence of isolated outcrops of tephra behind topographical barriers (sketched at left) that cannot have formed had an ice sheet been present at the time. Sections based on a 600 m-high, mafic tuff cone on James Ross Island, Antarctica, which was erupted in a marine (i.e., nonglacial) setting approximately 660 000 years ago. Reproduced from Smellie JL, McArthur JM, McIntosh WC, and Esser R (2006) Late Neogene interglacial events in the James Ross Island region, northern Antarctic Peninsula, dated by Ar/Ar and Sr-isotope stratigraphy. *Palaeogeography, Palaeoclimatology, Palaeoecology* 242: 169–187.



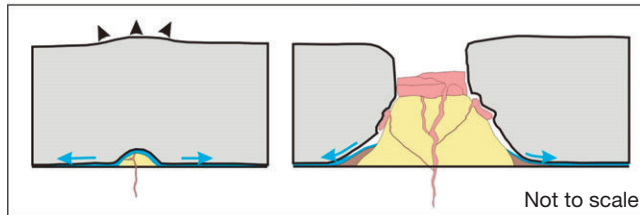
Figure 16 Example of a felsic dome at Ember Ridge northwest, British Columbia, Canada. Despite the prominent post-eruptive erosion, the original dome-like morphology can be reconstructed (broadly following the curved skyline) using the orientations of the prominent columnar cooling joints in the outcrop, which grew perpendicular to the original confining ice surface. Photograph by Melanie Kelman.

ultimately overflow as volcanic mass (magma) is added and the water is displaced. The lower temperatures of cold-based glaciers also strongly influence the ice rheology, which is stronger and deforms much more slowly. The energy requirements for melting are also different. Since more energy is needed than in wet-based systems, simply to bring the ice temperature up to the melting point, the ice walls of a vault or lake will recede more slowly than in a wet-based ice sheet. As a consequence of these combined characteristics, the vault or lake will not enlarge so rapidly as in wet-based ice. The volcanic products will rapidly fill it and spill out to override the surrounding ice sheet surface. Moreover, because of the high thermal inertia of magma and insulating properties of any underlying autobreccia rubble, any supraglacial lava flows may be capable of traveling considerable distances without melting deeply into the underlying ice (Wilson et al., 2012). The edifices will also be buttressed by the slowly receding ice walls, presumably leading to even taller constructs than are possible in wet-based ice, although these will be liable to postglacial gravitational collapse and consequent modification of the edifice shape (e.g., Komatsu et al., 2007).

Mafic tuya



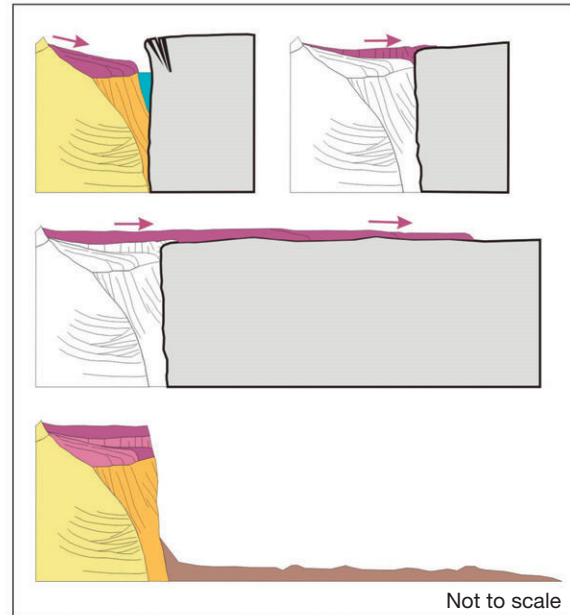
Felsic tuya, tephra dominated



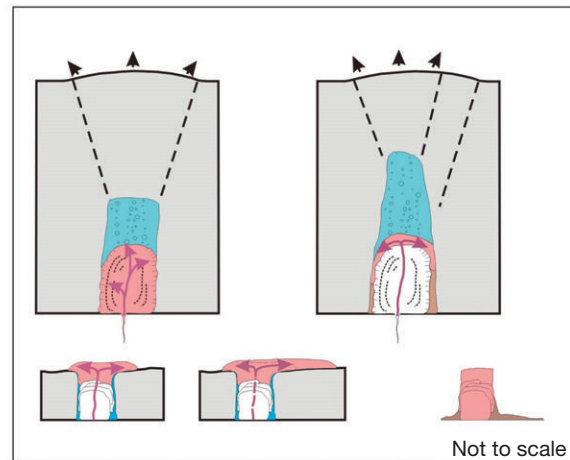
- | | |
|---------------------------|-------------------------|
| Ice sheet | Mafic or silicic tephra |
| Meltwater | Mafic lava |
| Silicic lava | Mafic hyaloclastite |
| Collapse breccia | Mafic pillow lava |
| Flow direction: meltwater | Ice surface deformation |
| Flow direction: lava | Potential ice fractures |

(a)

Mafic tuya



Felsic tuya, lava-flow dominated



(b)

Figure 17 Drawings illustrating the influence of glacier thermal regime on the construction, lithofacies and morphology of mafic and felsic tuyas. (a) Wet-based (temperate or subpolar) glacier regime. (b) Cold-based (polar) glacier regime. The principal differences are shorter lava-fed deltas and taller edifices in cold-based ice compared with those formed in wet-based ice. Reproduced from [Smellie JL \(2009\)](#) Terrestrial sub-ice volcanism: landform morphology, sequence characteristics and environmental influences, and implications for candidate Mars examples. In Chapman MG and Keszthely L (eds.) *Preservation of Random Mega-Scale Events on Mars and Earth: Influence on Geologic History*, vol. 453, pp. 55–76. Geological Society of America Special Papers.

The overall effects of eruptions under ice sheets with differing thermal regimes can, therefore, be broadly predicted and the main results for mafic and felsic tuyas are shown in [Figure 17](#). However, in the absence of any data for monogenetic volcanic landforms constructed beneath cold-based ice sheets, it is currently impossible to provide any real morphometric dimensions with which to compare and contrast with analogous landforms associated with wet-based ice sheets.

See also: [Glacial Climates: Biosphere Feedbacks](#). [Glacial Landforms: Introduction](#).

References

- Bergh S (1985) Structure, depositional environment and mode of emplacement of basaltic hyaloclastites and related lavas and sedimentary rocks: Plio-Pleistocene of

- the eastern volcanic rift zone, southern Iceland. Nordic Volcanological Institute Report No. 8502: 91 pp.
- Björnsson K (1988) Hydrology of ice caps in volcanic regions. *Visindafélag Íslendinga. Societas Scientiarum Islandica* 45: 1–139.
- Chapman MG (2003) Sub-ice volcanoes and ancient oceans/lakes: A Martian challenge. *Global and Planetary Change* 35: 185–198.
- Chapman MG and Smellie JL (2007) Mars interior layered deposits and terrestrial sub-ice volcanoes compared: Observations and interpretations of similar geomorphic characteristics. In: Chapman MG (ed.) *The Geology of Mars. Evidence from Earth-Based Analogs*, pp. 178–210. Cambridge: Cambridge University Press.
- Chapman MG and Tanaka KL (2001) The interior trough deposits on Mars: Sub-ice volcanoes? *Journal of Geophysical Research* 106: 10087–10100.
- Dixon JE, Filiberto JR, Moore JG, and Hickson CJ (2002) Volatiles in basaltic glasses from a subglacial volcano in northern British Columbia (Canada): Implications for ice sheet thickness and mantle volatiles. In: Smellie JL and Chapman MG (eds.) *Volcano–Ice Interaction on Earth and Mars*, pp. 255–271. London: Geological Society. Special Publications No. 202.
- Edwards BR and Russell JK (2002) Glacial influences on morphology and eruptive products of Hoodoo Mountain volcano, Canada. In: Smellie JL and Chapman MG (eds.) *Volcano–Ice Interaction on Earth and Mars*, pp. 179–194. London: Geological Society. Special Publications No. 202.
- Edwards BR, Skilling IP, Cameron B, Haynes C, Lloyd A, and Hungerford JHD (2009a) Evolution of an englacial volcanic ridge: Pillow Ridge tinar, Mount Edziza volcanic complex, NCVP, British Columbia, Canada. *Journal of Volcanology and Geothermal Research* 185: 251–275.
- Edwards BR, Tuffen H, Skilling IP, and Wilson L (2009b) Volcano–ice interactions on Earth and Mars: The state of the science. *Journal of Volcanology and Geothermal Research* 185(4): 247–250.
- Ghatan GJ and Head JW (2002) Candidate subglacial volcanoes in the south polar region of Mars: Morphology, morphometry, and eruption condition. *Journal of Geophysical Research* 107: E7.
- Gudmundsson MT, Palsson F, Björnsson H, and Högnadóttir T (2002) The hyaloclastite ridge formed in the subglacial 1996 eruption in Gjalp, Vatnajökull, Iceland: Present day shape and future preservation. In: Smellie JL and Chapman MG (eds.) *Volcano–Ice Interaction on Earth and Mars*, pp. 319–335. London: Geological Society. Special Publications No. 202.
- Gudmundsson MT, Sigmundsson F, Björnsson H, and Högnadóttir T (2004) The 1996 eruption at Gjalp, Vatnajökull ice cap, Iceland: Efficiency of heat transfer, ice deformation and subglacial water pressure. *Bulletin of Volcanology* 66: 46–65.
- Harder M and Russell JK (2007) Basaltic glaciovolcanism at Llangorse Mountain, northern British Columbia. *Bulletin of Volcanology* 69: 329–340.
- Head JW and Pratt S (2001) Extensive Hesperian-aged south polar ice sheet on Mars: Evidence for massive melting and retreat, and lateral flow and ponding of meltwater. *Journal of Geophysical Research* 106: 12275–12299.
- Hickson CJ (2000) Physical controls and resulting morphological forms of Quaternary ice-contact volcanoes in western Canada. *Geomorphology* 32: 239–261.
- Hickson CJ, Moore JG, Calk L, and Metcalfe P (1995) Intraglacial volcanism in the Wells Gray–Clearwater volcanic field, east-central British Columbia, Canada. *Canadian Journal of Earth Science* 32: 838–851.
- Jones JG (1969) Intraglacial volcanoes of the Laugarvatn region, south-west Iceland—I. *Quarterly Journal of the Geological Society, London* 124: 197–211.
- Kelman MC, Russell JK, and Hickson CJ (2002) Effusive intermediate glaciovolcanism in the Garibaldi Volcanic Belt, southwestern British Columbia, Canada. In: Smellie JL and Chapman MG (eds.) *Volcano–Ice Interaction on Earth and Mars*, pp. 195–211. London: Geological Society. Special Publications No. 202.
- Komatsu G, Arzhannikov SG, Arzhannikova AV, and Ershov K (2007) Geomorphology of subglacial volcanoes in the Azas Plateau, the Tuva Republic, Russia. *Geomorphology* 88: 312–328.
- Larsen G (2002) Holocene eruptions within the Katla volcanic system, south Iceland: Characteristics and environmental impact. *Jökull* 49: 1–28.
- Lescinsky DT and Fink JH (2000) Lava and ice interactions at stratovolcanoes: Use of characteristic features to determine past glacial extents and future volcano hazards. *Journal of Geophysical Research* 105: 23711–23726.
- Lescinsky DT and Sisson TW (1998) Ridge-forming, ice-bounded lava flows at Mount Rainier, Washington. *Geology* 26: 351–354.
- Licciardi JM, Kurz MD, and Curtice JM (2007) Glacial and volcanic history of Icelandic table mountains from cosmogenic ³He exposure ages. *Quaternary Science Reviews* 26: 1529–1546.
- Loughlin SC (2002) Facies analysis of proximal subglacial and proglacial volcanoclastic successions at the Eyjafjallajökull central volcano, southern Iceland. In: Smellie JL and Chapman MG (eds.) *Volcano–Ice Interaction on Earth and Mars*, pp. 149–178. London: Geological Society. Special Publications No. 202.
- Martínez-Alonso S, Mellon MT, Banks ME, Keszthelyi LP, and McEwen AS, the HiRISE Team (2011) Evidence of volcanic and glacial activity in Chryse and Acidalia Planitiae, Mars. *Icarus* 212: 597–621.
- Mathews WH (1951) The Table, a flat-topped volcano in southern British Columbia. *American Journal of Science* 249: 830–841.
- Mathews WH (1958) Geology of the Mount Garibaldi area, southwestern British Columbia. Part II. Geomorphology and Quaternary volcanic rocks. *Geological Society of America Bulletin* 69: 161–178.
- McGarvie D (2009) Rhyolitic volcano–ice interactions in Iceland. *Journal of Volcanology and Geothermal Research* 185: 367–389.
- McGarvie DW, Burgess R, Tindle AG, Tuffen H, and Stevenson JA (2006) Pleistocene rhyolitic volcanism at the Torfajökull central volcano, Iceland: Eruption ages, glaciovolcanism, and geochemical evolution. *Jökull* 56: 57–75.
- McGarvie DW, Stevenson JA, Burgess R, Tuffen H, and Tindle AG (2007) Volcano–ice interactions at Prestahnúkur, Iceland: Rhyolite eruption during the last interglacial–glacial transition. *Annals of Glaciology* 45: 38–47.
- Mee K, Tuffen H, and Gilbert JS (2006) Snow-contact volcanic facies and their use in determining past eruptive environments at Nevados de Chillán volcano, Chile. *Bulletin of Volcanology* 68: 363–376.
- Moore JG, Hickson CJ, and Calk LC (1995) Tholeiitic–alkalic transition at subglacial volcanoes, Tuya region, British Columbia, Canada. *Journal of Geophysical Research* 100: 24577–24592.
- Schopka HH, Gudmundsson MT, and Tuffen H (2006) The formation of Helgafell, SW-Iceland, a monogenetic subglacial hyaloclastite ridge: Sedimentology, hydrology and ice-volcano interaction. *Journal of Volcanology and Geothermal Research* 152: 259–377.
- Skilling IP (1994) Evolution of an englacial volcano: Brown Bluff, Antarctica. *Bulletin of Volcanology* 56: 573–591.
- Skilling IP (2002) Basaltic pahoehoe lava-fed deltas: Large-scale characteristics, clast generation, emplacement processes and environmental discrimination. In: Smellie JL and Chapman MG (eds.) *Volcano–Ice Interaction on Earth and Mars*, pp. 91–113. London: Geological Society. Special Publications No. 202.
- Skilling IPS (2009) Subglacial to emergent basaltic volcanism at Hlōdufell, south-west Iceland: A history of ice-confinement. *Journal of Volcanology and Geothermal Research* 185: 276–289.
- Smellie JL (2000) Subglacial eruptions. In: Sigurdsson H (ed.) *Encyclopedia of Volcanoes*, pp. 403–418. San Diego, CA: Academic Press.
- Smellie JL (2001) Lithofacies architecture and construction of volcanoes in englacial lakes: Icefall Nunatak, Mount Murphy, eastern Marie Byrd Land, Antarctica. In: White JDL and Riggs NR (eds.) *Volcaniclastic Sedimentation in Lacustrine Settings*, pp. 73–98. International Association of Sedimentologists. Special Publication No. 30.
- Smellie JL (2006) The relative importance of supraglacial versus subglacial meltwater escape in basaltic subglacial tuya eruptions: An important unresolved conundrum. *Earth-Science Reviews* 74: 241–268.
- Smellie JL (2007) Quaternary volcanism: Subglacial landforms. In: Elias SA (ed.) *Encyclopedia of Quaternary Sciences*, pp. 784–798. Amsterdam: Elsevier.
- Smellie JL (2008) Basaltic subglacial sheet-like sequences: Evidence for two types with different implications for the inferred thickness of associated ice. *Earth-Science Reviews* 88: 60–88.
- Smellie JL (2009) Terrestrial sub-ice volcanism: Landform morphology, sequence characteristics and environmental influences, and implications for candidate Mars examples. In: Chapman MG and Keszthelyi L (eds.) *Preservation of Random Mega-Scale Events on Mars and Earth: Influence on Geologic History*, vol. 453, pp. 55–76. Geological Society of America. Special Papers.
- Smellie JL and Chapman MG (eds.) (2002) *Volcano–Ice Interaction on Earth and Mars*. London: Geological Society. Special Publications No. 202.
- Smellie JL, Haywood AM, Hillenbrand C-D, Lunt DJ, and Valdes PJ (2009) Nature of the Antarctic Peninsula Ice Sheet during the Pliocene: Geological evidence and modeling results compared. *Earth-Science Reviews* 94: 79–94.
- Smellie JL and Hole MJ (1997) Products and processes in Pliocene-Recent, subaqueous to emergent volcanism in the Antarctic Peninsula: Examples of englacial Surtseyan volcano construction. *Bulletin of Volcanology* 58: 628–646.
- Smellie JL, Hole MJ, and Nell PAR (1993) Late Miocene valley-confined subglacial volcanism in northern Alexander Island, Antarctic Peninsula. *Bulletin of Volcanology* 55: 273–288.
- Smellie JL, Johnson JS, McIntosh WC, et al. (2008) Six million years of glacial history recorded in volcanic lithofacies of the James Ross Island Volcanic Group, Antarctic Peninsula. *Palaeogeography, Palaeoclimatology, Palaeoecology* 260: 122–148.
- Smellie JL, McArthur JM, McIntosh WC, and Esser R (2006a) Late Neogene interglacial events in the James Ross Island region, northern Antarctic peninsula, dated by Ar/Ar

- and Sr-isotope stratigraphy. *Palaeogeography, Palaeoclimatology, Palaeoecology* 242: 169–187.
- Smellie JL, McIntosh WC, and Esser R (2006b) Eruptive environment of volcanism on Brabant Island: Evidence for thin wet-based ice in northern Antarctic Peninsula during the Late Quaternary. *Palaeogeography, Palaeoclimatology, Palaeoecology* 231: 233–252.
- Smellie JL, McIntosh WC, Esser R, and Fretwell P (2006c) The Cape Purvis volcano, Dundee Island (northern Antarctic Peninsula): Late Pleistocene age, eruptive processes and implications for a glacial palaeoenvironment. *Antarctic Science* 18: 399–408.
- Smellie JL, Rocchi S, and Armienti P (2011a) Late Miocene volcanic sequences in northern Victoria Land, Antarctica: Products of glaciovolcanic eruptions under different thermal regimes. *Bulletin of Volcanology* 73: 1–25.
- Smellie JL, Rocchi S, Gemelli M, Di Vincenzo G, and Armienti P (2011b) A thin predominantly cold-based Late Miocene East Antarctic ice sheet inferred from glaciovolcanic sequences in northern Victoria Land, Antarctica. *Palaeogeography, Palaeoclimatology, Palaeoecology* 307: 129–149.
- Smellie JL and Skilling IP (1994) Products of subglacial eruptions under different ice thicknesses: Two examples from Antarctica. *Sedimentary Geology* 91: 115–129.
- Snorrason SP and Vilmundardóttir EG (2000) Pillow lava sheets: Origins and flow patterns. In: Gulick VC and Gudmundsson MT (eds.) *Volcano–Ice Interaction on Earth and Mars*. <http://astrogeology.usgs.gov/Projects/VolcanolceWorkshop/> Conference Abstracts, Reykjavik, Iceland, August 13–18, 2000, 62p.
- Stevenson JA, Gilbert JS, McGarvie DW, and Smellie JS (2011) Explosive rhyolite tuya formation: Classic examples from Kerlingarfjöll, Iceland. *Quaternary Science Reviews* 30: 192–209.
- Stevenson JA, McGarvie DW, Smellie JL, and Gilbert JS (2006) Subglacial and ice-contact volcanism at the Óraefajökull stratovolcano, Iceland. *Bulletin of Volcanology* 68: 737–752.
- Stevenson JA, Smellie JL, McGarvie DW, Gilbert JS, and Cameron BI (2009) Subglacial intermediate volcanism at Kerlingarfjöll, Iceland: Magma–water interactions beneath thick ice. *Journal of Volcanology and Geothermal Research* 185: 337–351.
- Thouret J-C (1990) Effects of the November 13, 1985 eruption on the snow pack and ice cap of Nevado del Ruiz volcano, Colombia. *Journal of Volcanology and Geothermal Research* 41: 177–201.
- Tuffen H, McGarvie DW, and Gilbert JS (2007) Will subglacial rhyolite eruptions be explosive or intrusive? Some insights from analytical models. *Annals of Glaciology* 45: 87–94.
- Tuffen H, McGarvie DW, Gilbert JS, and Pinkerton H (2002a) Physical volcanology of a subglacial-to-emergent rhyolite tuya at Rauufossafjöll, Torfajökull, Iceland. In: Smellie JL and Chapman MG (eds.) *Volcano–Ice Interaction on Earth and Mars*, pp. 213–236. London: Geological Society. Special Publications No. 202.
- Tuffen H, McGarvie DW, Pinkerton H, Gilbert JS, and Brooker R (2009) An explosive-intrusive subglacial rhyolite eruption at Dalakvisl, Torfajökull, Iceland. *Bulletin of Volcanology* 70: 841–860.
- Tuffen H, Pinkerton H, McGarvie DW, and Gilbert JS (2002b) Melting of the glacier base during a small-volume subglacial rhyolite eruption: Evidence from Bláhnúkur, Iceland. *Sedimentary Geology* 149: 183–198.
- Van Bemmelen RW and Rutten MG (1955) *Tablemountains of Northern Iceland*. Leiden, The Netherlands: Brill.
- Walker GPL and Blake DH (1966) The formation of a palagonite breccia mass beneath a valley glacier in Iceland. *Journal of the Geological Society, London* 122: 45–61.
- Werner R, Schmincke H-U, and Sigvaldason G (1996) A new model for the evolution of table mountains: Volcanological and petrological evidence from Herdubreid and Herdubreidartogi volcanoes (Iceland). *Geologische Rundschau* 85: 390–397.
- Werner R and Schmincke H-U (1999) Englacial vs lacustrine origin of volcanic table mountains: Evidence from Iceland. *Bulletin of Volcanology* 60: 335–354.
- Wilson L, Smellie JL, and Head JW (2012) Volcano–ice interactions. In: Fagents SA, Gregg TKP, and Lopes RMC (eds.) *Modeling Volcanic Processes: The Physics and Mathematics of Volcanism*. Cambridge University Press ch. 12.
- Wörner G and Viereck L (1987) Subglacial to emergent volcanism at Shield Nunatak, Mt. Melbourne volcanic field, Antarctica. *Polarforschung* 57: 27–41.

Evidence of Glacier Recession

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Introduction

Today, we live in an age of global glacier recession. The retreat of glaciers and ice sheets worldwide provides unequivocal evidence of climate change and has serious implications for sea-level rise and landscape change in mountainous regions and high latitudes. Many mountain glaciers could disappear in the twenty-first century.

Glacier recession is the stepwise retreat of the terminus of a glacier or ice sheet toward its source area. This occurs when more ice is melted near the terminus in the ablation zone than can be replaced by ice flowing down the glacier from the accumulation zone. Glacier recession usually follows rising air temperatures because an increase in the elevation of the equilibrium line forces the glacier to attempt to reach a new mass balance by decreasing the area of its ablation zone. Glacier recession may also occur because of a decrease in snowfall in the accumulation zone or because of an increase in solar radiation. This also raises the equilibrium line and forces glacier recession. Ablation rates are commonly an order of magnitude higher than accumulation rates; therefore, changes in air temperature usually cause more dramatic changes in glacier areas than do changes in snowfall. Glaciers retreat in a regressive fashion because glacier ice is always actively flowing toward the terminus to replace what is melting. Because glaciers, especially small mountain glaciers, are sensitive to air temperature, snowfall, and solar radiation, both past and present glacier recessions are excellent indicators of climate change. Glaciers may also retreat by widespread downwasting if the terminus is cut off from the accumulation zone as a result of topographic effects or stagnation of ice following a glacier surge. (See Paterson (1994) for a detailed explanation of glacier mass balance and glacier surges.)

Direct observations, field surveys, oblique and aerial photography, and remotely sensed data are the primary evidence for historic glacier recession during the late Holocene. A careful analysis of data gathered in a series of years can provide evidence of both the magnitude and rate of glacier recession. Numerous studies have quantified the rate of glacier recession over the last 50–400 years in nearly every region of the world (e.g., Dyurgerov and Meier, 2000; Grove, 1988; Meier et al., 2003). In the past decade, entirely new methods based on synthetic aperture radar (SAR), laser altimetry or LIDAR (light detection and ranging), and satellite-based gravity measurements have revolutionized the way mass balance is measured in the largest glaciers, including the Antarctic and Greenland ice sheets.

The mapping of landforms and sediments in combination with absolute dating methods are used to determine the magnitude and sometimes the rate of late Pleistocene and Holocene glacier recession. Landforms that can be used to reconstruct the former glacier terminus include moraines, subglacial landforms, eskers, and outwash plains. There are many different types of moraines, which are ridges of glacial sediment formed along the

terminus (end moraines or recessional moraines) or sides of a glacier (lateral moraines). These landforms contain information about both the extent and style of past glaciers. Subglacial landforms such as drumlins, flutes, and striae are commonly exposed during glacier recession and record ice flow direction. Stream sediment deposited in front of, or on top of, the ice produces ice-contact features, outwash plains, fans, and channels. Many of these landforms can be dated with radiocarbon methods or newer methods such as cosmogenic radionuclide exposure and optically stimulated luminescence dating. What follows is a summary of some of the historic and geomorphic evidence for glacier recession and the implications for climate and glacier dynamics.

Evidence for Late Holocene Glacier Recession

Glaciers provide important spiritual, aesthetic, and utilitarian resources for human populations. In many parts of the world, glaciers provide economic benefits through water resources, hydroelectric generation, tourism, and wildlife habitat. The current recession of glaciers has important implications for the future of these resources. Glacier recession also contributes significantly to accelerating global sea-level rise (see World Glacier Monitoring Service report *Global Glacier Changes: Facts and Figures*, <http://www.grid.unep.ch/glaciers/>).

Mountain glaciers all over the world are in full retreat threatening natural resources and providing unequivocal evidence of climate change (Figure 1). About 100 glaciers worldwide have been monitored continuously for their net mass balance for the last 30 years by direct field measurements of accumulation and ablation. Mass balance for a glacier is determined by estimating the total accumulation and ablation in cubic meters of water equivalent. Net mass balance is the change in volume of water equivalent divided by the total area of the glacier. It is commonly reported in millimeters of water equivalent. According to the World Glacier Monitoring System, in the last year reported (2009), 83 of 99 glaciers in this sample decreased in mass, and at least some glaciers have completely disappeared in the last few years. Most years for which we have data show the same trend. The mean net cumulative mass balance for 30 representative glaciers in nine regions shows a decrease in mass for every year except three (1983, 1987, and 1989) in the period 1980–2009, and the three positive years were barely positive (see <http://www.geo.unizh.ch/wgms/>). Unfortunately, most of these are small glaciers and the distribution is weighted toward the Northern Hemisphere and Europe. Even so, it is clear we live in a period dominated by a global glacier recession.

Recession of mountain glaciers adds a significant volume of water to oceans and is an important factor in current sea-level rise (Meier et al., 2003). Dyurgerov and Meier (2005) estimated that there are over 160 000 mountain glaciers and ice caps with a total area of 685 000 km². This does not include the

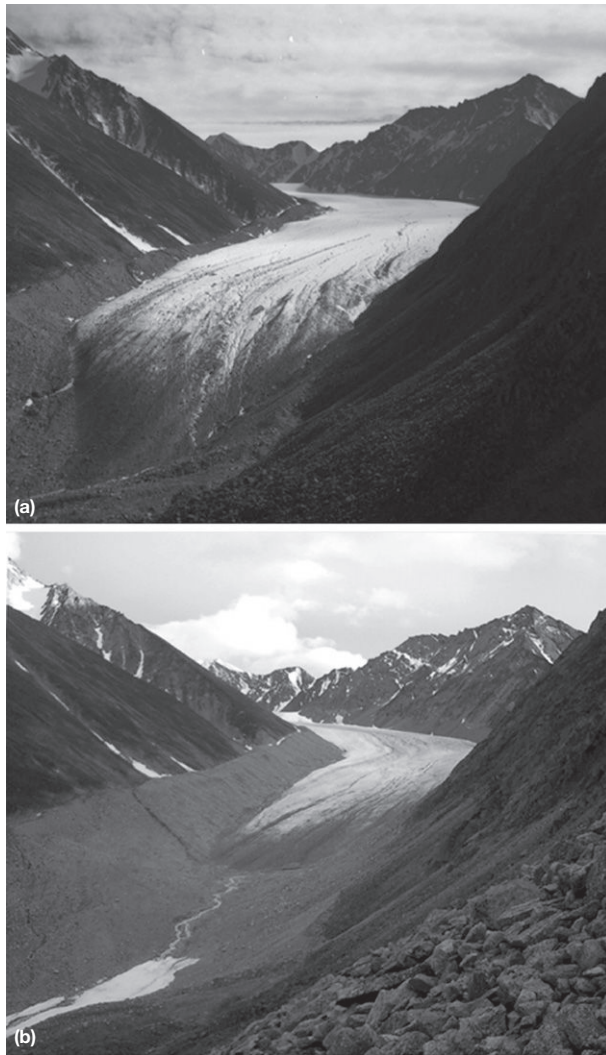


Figure 1 Examples of glacier recession as shown in historical photographs. (a and b) McCall Glacier in 1958 and 2003. This is a small valley glacier located in the Brooks Range of northern Alaska. Photos by Austin Post (1958) and Matt Nolan (2003).

ice sheets of Antarctica and Greenland. Melting of these mountain glaciers alone will likely contribute a minimum of between 18 and 37 cm in sea-level rise by the year 2100 (Bahr et al., 2009). This is in addition to what the ice sheets may contribute in the future.

Glacier monitoring via remotely sensed data is a commonly used method to examine glacier change (Rignot et al., 2003). The Global Land Ice Measurements from Space (GLIMS) sponsored by the United States Geological Survey, the National Snow and Ice Data Center, and academic researchers is currently developing a database of glacier changes using ASTER satellite images compared to other data sources. Monitoring of glacier recession has and will continue to play an important role in understanding global climate change. The current GLIMS system contains outlines for about 60 000 individual glaciers, about one-third of all the known glaciers.

The Greenland and Antarctic ice sheets contain at least 60 m of equivalent sea-level rise. Until very recently, little was known

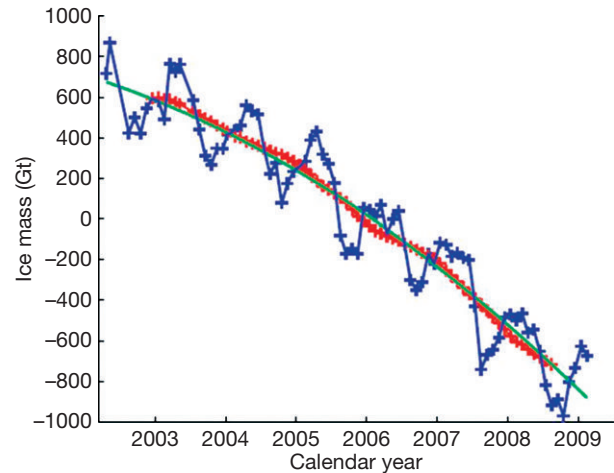


Figure 2 Mass loss from Greenland Ice Sheet between 2002 and 2008 from the GRACE Mission Reproduced from Cazenave A and Chen JL (2010) Time-variable gravity from space and present-day mass redistribution in the Earth system. *Earth and Planetary Science Letters* 298(3–4): 263–274. After Velicogna I (2009) Increasing rates of ice mass loss from the Greenland and Antarctic ice sheets revealed by GRACE. *Geophysical Research Letters* 36: L19503. <http://dx.doi.org/10.1029/2009GL040222>.

about the mass balance of these ice sheets. Recent satellite and airborne missions all indicate that both the Greenland and Antarctic ice sheets are losing mass. The GRACE mission (gravity recovery and climate experiment) beginning in 2003 has provided estimates of mass changes in both ice sheets. Greenland appears to be shrinking rapidly and the mass loss in the most recent period (2006–2009) appears to have accelerated from earlier in the decade (2002–2005). The most recent mass loss appears to be between 267 and 286 Gt year⁻¹ compared to 104–144 Gt year⁻¹ in the earlier period (Figure 2). As per GRACE data, estimates for the rest of Antarctica are complicated by glacial rebound, but they also indicate mass losses. (See Cazenave and Chen (2010) for a complete description of GRACE mission applications.) The most recent SAR data combined with GRACE data indicate that the West Antarctic ice sheet lost ice between 2002 and 2008 (Sasgen et al., 2010).

Besides direct measurements of mass balance changes, many more glaciers have been monitored for terminus position by means of direct observations, photographs, and remotely sensed data (Figure 3). Glacier recession has been reported in nearly every region of Earth (e.g., Cook et al., 2005; Dyurgerov and Meier, 2000; Meier et al., 2003; Rignot et al., 2003). Oerlemans (2005) used 169 glacier retreat records from around the world to reconstruct mean global temperature change over the last 400 years. This climate record extracted from glacier recession records approximated global temperature reconstructions based on other independent paleoclimate proxy records. Other studies show that glacier recession has accelerated in the last two decades of the twentieth century (Dyurgerov and Meier, 2000; Meier et al., 2003). This most recent global glacier recession began at the end of the Little Ice Age. The Little Ice Age was a period of cooler global temperatures that began in the late Middle Ages and ended in the last half of the nineteenth century (Grove, 1988). Since then, we have seen an acceleration of warming during the late twentieth century and this has led to a quickening in the pace of modern glacier recession.

Global sea-level rise since 1900 is estimated to be between 13 and 17 cm or $\sim 1.3\text{--}1.7\text{ mm year}^{-1}$ assuming a linear trend fitted to tide gauge data and satellite measurements. (See Cazenave et al., 2008 for a recent synthesis of global sea-level rise.) This rise is thought to be about half the result of mountain glacier melting and half due to the rise of sea surface temperatures leading to expansion of sea water. Recent satellite data suggest that the rate during the period 1993–2008 is almost twice the long-term rate or between 3.1 and 3.4 mm year^{-1} . Sea-level rise over the next 100 years could be even higher (up to 75 cm) depending on the rate of climate warming, the response of mountain glaciers and ice sheets, and the amount of warming of sea surface waters. These estimates assume that glaciers and ice sheets melt in a progressive fashion without a partial or total collapse of the marine-based West Antarctic ice sheet or parts of the Greenland ice sheet. If the West Antarctic ice sheet were to collapse, it could add another 3–5 m of sea-level rise in perhaps as little as a few hundred years. Rapid sea-level rise has dire implications for societies as a large proportion of the world's population lives close to the sea. Monitoring small mountain glaciers and ice sheets is necessary to help quantify the volume of the water being added to the oceans as well as to monitor regional climate. Current and future satellite remote sensing programs such as GRACE and the future IceSat II missions are critical to the monitoring of glacier and ice sheet recession.

Modern Glacier Recession and Geomorphology

As glaciers retreat, they expose landscapes that were only recently covered by ice. This provides an excellent opportunity to explore and understand how glacial landforms and sediments



Figure 3 This false-color image shows the Gangotri Glacier, located in the Uttarkashi District of Garhwal Himalayas, India. Gangotri Glacier is one of the largest in the Himalayas and is a source for the Ganges River. The glacier is currently 30.2 km long and between 0.5 and 2.5 km wide. Gangotri Glacier has been receding since 1780, although studies show that its retreat accelerated after 1971. (Please note that the contour lines drawn here to show the recession of the glacier's terminus over time are approximate.) Over the past 25 years, this glacier has retreated more than 850 m, with a recession of 76 m from 1996 to 1999 alone. Source: NASA (http://earthobservatory.nasa.gov/Newsroom/NewImages/images.php3?img_id=16584).

are formed during glacier recession. An excellent example of recent glacier recession is the well-studied Breiðamerkurjökull of southeastern Iceland (Evans and Twigg, 2002). This is a large temperate outlet glacier that drains the ice cap Vatnajökull (Figure 4). Breiðamerkurjökull has retreated between 4 and 5 km since 1894 AD (Figure 5). This works out to an average recession rate of approximately $40\text{--}50\text{ m year}^{-1}$ for the last 100 years. While recession has been the overall trend, the glacier did not retreat every year and in fact advanced or remained steady in the periods of 1950–55, 1968–73, and 1975–87. A large end moraine was formed during and after Breiðamerkurjökull reached its Little Ice Age maximum some time between 1886 and 1890. After about 1894, the glacier began to retreat from this Little Ice Age moraine. Breiðamerkurjökull continues to retreat like most other glaciers in Iceland.

The deglaciated foreland in front of Breiðamerkurjökull displays an interesting and varied assemblage of glacial landforms and sediments (Figure 6) that have been mapped in detail by Evans and Twigg (2002) and earlier workers (see references in Evans and Twigg, 2002). The most



Figure 4 This image shows the Vatnajökull ice cap as it appears in late summer. Vatnajökull covers an area of about 8100 km^2 and is the largest ice cap in Europe. One can see how the glacier's accumulation area is still covered in snow. Along the ice cap margin, one can see outlet glaciers that drain the main ice body. Most of the outlet glaciers show up as dark. This is because in the ablation zone, all of the previous winter's snow has melted by the end of summer, and bare ice shows up as dark. One can also see medial and lateral moraines on some of the outlet glaciers. Along the left margin of the photo, one can see some of the other ice caps in Iceland; Hofsjökull in the upper left and Myrdalsjökull in the lower left. The red areas in this false-color image are those areas covered with vegetation. Dark areas are lava fields and water. The large dark area between the ice cap and the coast is the sandur plain of Skeiðarárjökull, one of the largest outlet glaciers that drain Vatnajökull. Some areas along the coast have a bluish tint. This is meltwater carrying silt and discharging into the North Atlantic. (Landsat MSS image – USGS.)

conspicuous of these landforms are numerous small push moraines (Figures 7 and 8). Push moraines are small ridges that are made up of subglacially deposited till plus some outwash deposited in late summer. Ridges are commonly less than 5 m high and 10 m wide (Boulton, 1986). They can be traced in excess of 3 km and commonly have a saw-tooth pattern that reflects their formation along the crevassed terminus (Figures 7 and 8). They tend to be asymmetric in cross-sectional profile with a shallower proximal and steeper distal flank. In late winter (mid-April–mid-May), ablation is very low near the terminus of Breiðamerkurjökull. Because of this factor, the terminus of the glacier advances up to a few meters. This winter readvance pushes up material in front of the ice, forming a push ridge. In summer, the terminus retreats because of the high ablation rates in this part of south Iceland ($\sim 10 \text{ m year}^{-1}$). If the overall retreat in summer is greater than the next year's winter readvance, a small push moraine is preserved. Push moraines may coalesce when a particular year's readvance is more extensive than the previous winter's. Sediment may also be squeezed from underneath the ice margin to form small push ridges. At Breiðamerkurjökull, small push moraines have formed nearly every year since 1965 on the west lobe of the glacier (Figures 7 and 8). Because these ridges can be demonstrated to have formed annually or semiannually, they have

sometimes been called 'annual moraines.' Similar ridges have been exposed by recent glacier recession in Norway, Svalbard, Baffin Island, the Canadian Rockies, Canadian Arctic, southeastern Alaska, and New Zealand.

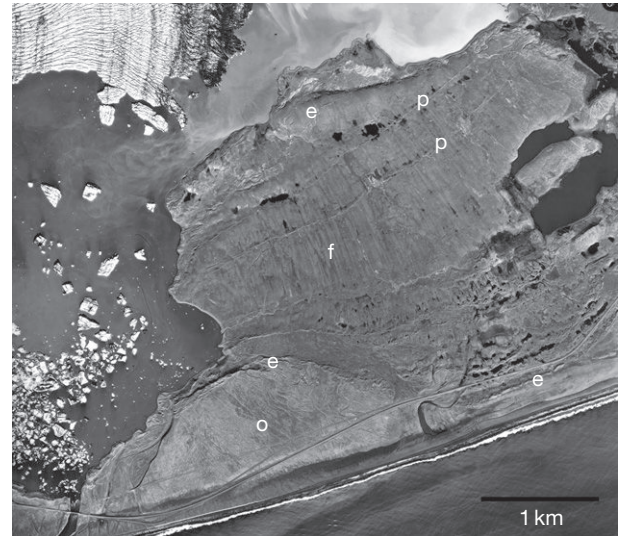


Figure 6 Vertical aerial photograph of the recently deglaciated area in front of the eastern lobe of Breiðamerkurjökull, south Iceland. Features visible include fluted till surface (f), small push moraines (p), larger end moraines (e), and outwash plain (o) (Landmaelingar Islands and University of Glasgow, 1998).

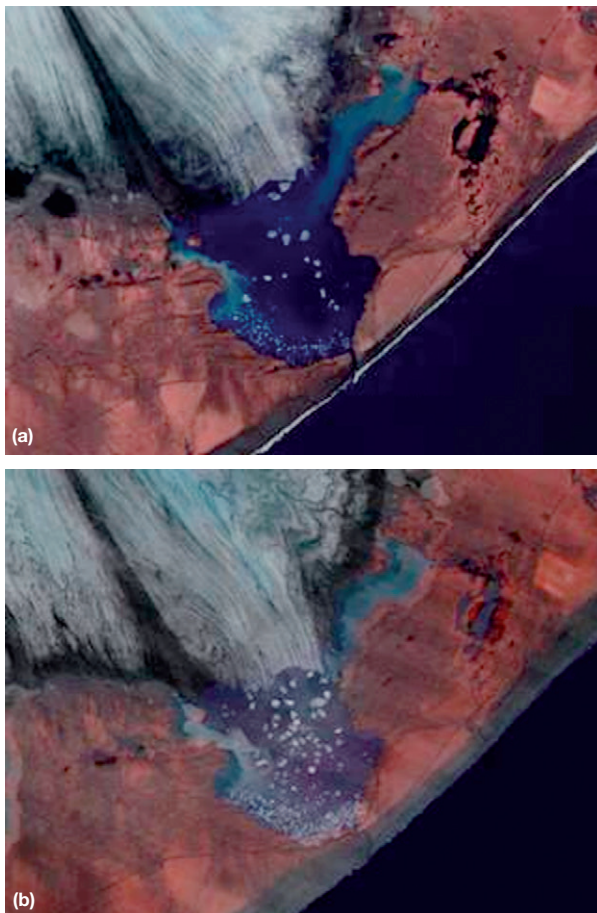


Figure 5 Quicktime movies of the recession of Breiðamerkurjökull from 1973 to 2000. Source: NASA (<http://svs.gsfc.nasa.gov/search/Keyword/Breidamerkurjokull.html>).

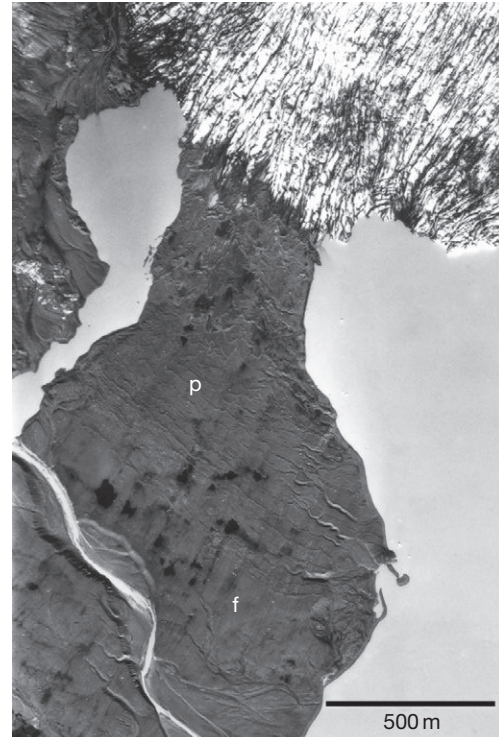


Figure 7 Vertical aerial photograph of the deglaciated surface in front of the western lobe of Breiðamerkurjökull, south Iceland. Features visible are fluted till surface (f) and small push moraines (p) (Landmaelingar Islands and University of Glasgow, 1998).

Larger ridges of glacial sediment up to tens of meters high and hundreds of meters wide formed during more extensive and long-lived readvances or still-stands in the terminus, which commonly results in the stacking and bulldozing of subglacial and proglacial sediments. The larger recessional moraines in front of Breiðamerkurjökull formed during still-stands of the terminus and others are older features that have been overridden by the glacier as it reached its Little Ice Age maximum position. Some of these larger recessional moraines show evidence for the presence of buried ice in the form of hummocky topography. Hummocky topography forms as buried ice within the moraine melts and causes collapse of the overlying sediment.

Flutes and drumlins make up a continuum of streamlined landforms that form under temperate glaciers that are sliding over a temperate bed (Boulton, 1987). Commonly, obstructions such as boulders or rock outcrops will form cavities under the ice (Figure 6). Subglacial sediment may be deformed into these cavities forming a fluted terrain under the ice margin. Larger obstructions and heterogeneity in the subglacial sediment will lead to the formation of streamlined hills called drumlins. At Breiðamerkurjökull, a fluted till surface underlies the small push moraines and larger recessional moraines (Figures 6 and 7). Small drumlins are also present in the landscape. An important observation is that some of the flutes end in individual push moraines (Evans and Twigg, 2002). This indicates that fluted terrain was being formed during recession and that small push ridges and flutes are genetically related. However, drumlins appear to be related to the position of ice at the Little Ice Age maximum and not to recessional positions. Recently, newly formed drumlins have been observed on a recently deglaciated surface in front of Múlajökull, a surge-type outlet glacier of Hofsjökull (Johnson et al., 2010). As modern glaciers retreat, we may learn more about how subglacial features such as drumlins are formed.

Eskers are another common feature in landscapes exposed by glacier recession. At Breiðamerkurjökull, several small eskers are being exposed as the terminus retreats (Figure 9). Numerous subglacial and englacial ice tunnels drain meltwater in summer from the surface of Breiðamerkurjökull. These tunnels can become filled with sorted stream sediments. Many of these tunnels occur englacially and the sediments filling them are poorly preserved because of reworking during the melting of the

underlying ice. Some subglacial tunnels are preserved as esker ridges. Sandar (or outwash plains) have formed at several levels in the proglacial area in front of Breiðamerkurjökull since the glacier began to retreat in 1900. On aerial photographs, the abandoned braided stream pattern can be seen on these surfaces (Figures 6 and 9).

Late Pleistocene Glacier Recession

At the end of the Pleistocene Epoch approximately 10 000 years ago, the great ice sheets that covered North America and northern Europe were rapidly melting. This glacier recession began about 21 000 ^{14}C years BP and accelerated after about 14 500 ^{14}C years BP. By 5000 ^{14}C years BP, the great ice sheets were gone (see Dyke and Prest, 1987 for maps of North America). Most of the evidence we have for the timing of this glacier recession comes from radiocarbon dating of wood, peat, shells, and other organic matter that accumulated on top of glacial sediments deposited during recession (see Dyke et al., 2002, for a recent summary). The timing of major readvances during the deglaciation comes from radiocarbon dated wood buried under till of the readvance phases. Perhaps the most well-known example is the Two Creeks forest bed in the Great Lakes region of the United States. At this site, numerous radiocarbon dates on wood show that ice advance (and the rising lake waters) killed the forest and buried it under a readvancing glacier and its till at about 11 850 ^{14}C years BP (Broecker and Farrand, 1963).

The Laurentide ice sheet, the largest of all the late Pleistocene ice sheets, produced a truly remarkable record of glacier recession along its southern margin in the northern United States. Many of the glacial features we see forming in southeastern Iceland today along the margin of Breiðamerkurjökull were being formed along the southern margin of the Laurentide ice sheet during the late Pleistocene deglaciation (Colgan et al., 2003; Evans, 2003a,b). By comparing landforms and sediments in Iceland with those that formed during the late Pleistocene, we can better understand how the late Pleistocene deglaciation progressed and how rapidly the terminus retreated across the landscape.

Central and southern Wisconsin, an area of roughly 40 000 km², exposes a unique record of glacier recession including landforms such as drumlins and flutes, small push moraines, recessional moraines, eskers, and outwash plains (Colgan, 1999; Colgan and Mickelson, 1997). The distribution of these landforms and their relationships to one another are nearly identical to those forming in southeastern Iceland today. One big difference is that landforms in Wisconsin cover an area hundreds of times larger than that covered in Iceland. Another important difference is the occurrence of features such as hummocky moraines and tunnel channels, features not common in Iceland. These are found along the maximum ice margin and indicate that supraglacial sediment cover and drainage of subglacial meltwater were important processes when ice was at its maximum late Pleistocene position (see Clayton et al., 1985). In the deglaciated foreland of Breiðamerkurjökull, hummocky topography is only a minor landform type and tunnel channels do not appear at all. This difference could be because unlike Breiðamerkurjökull, which



Figure 8 Small push moraines in front of Breiðamerkurjökull, south Iceland. Photograph by P.M. Colgan (September 2000).

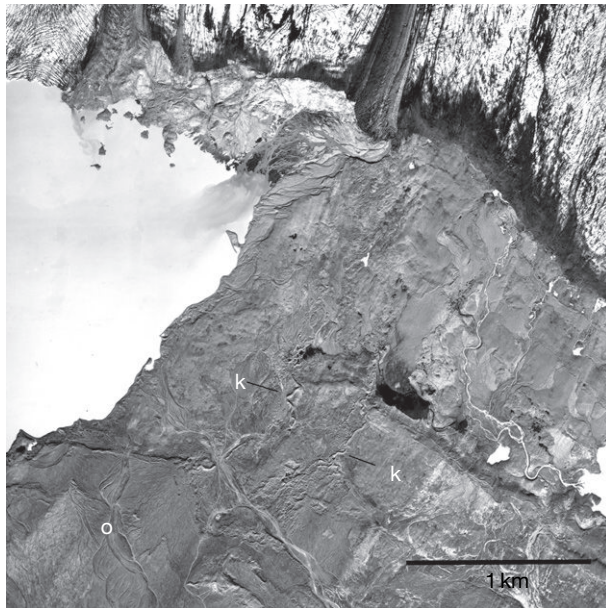


Figure 9 Eskers (k) on the deglaciated foreland in front of Breiðamerkurjökull, south Iceland; outwash plain (o) (Landmaelingar Islands and University of Glasgow, 1998).

is a temperate glacier, these glaciers were polythermal glaciers. This would lead to a frozen margin interacting with permafrost and producing hummocky moraines (see [Evans, 2009](#), for a recent discussion of these landforms).

Small push moraines are common features in southern Wisconsin in the area once covered by the Green Bay lobe of the southern Laurentide ice sheet ([Figure 10](#)). These ridges are usually about 10–50 m wide and less than 2 m high. They are traceable over distances of up to 3 km and commonly overlap or cross-cut. Where ridges are well preserved, they appear to be spaced between 25 and 100 m apart. The ridges are superimposed on fluted and drumlinized terrain just as they are in Iceland. If we accept the hypothesis that these ridges formed annually or semiannually as the small push moraines in Iceland do, then they would indicate a retreat rate of about 50 m year^{-1} for at least hundreds of years between 15 500 ^{14}C years BP and 13 000 ^{14}C years BP. Some especially well-preserved ridges ([Figure 10](#)) are also described in northern Wisconsin, produced by the Wisconsin Valley sublobe of the Laurentide ice sheet ([Ham and Attig, 2001](#)). It is estimated that retreat rates were between 37 and 82 m year^{-1} when these ridges formed and ablation rates near the margin of the ice sheet must have been between 1.7 and 14.5 m year^{-1} . The common seasonal climate patterns both in Iceland and along the southern margin of the Laurentide ice sheet may have been the high summer ablation rates followed by a cold winter when ice could advance and push up a small ridge.

Larger recessional moraines are found along the southern margin of the Laurentide ice sheet and can be traced all the way from New York to the Dakotas. These moraines have a complex history reflecting sedimentation during both glacier advance and glacier recession. Many of the end moraines in Illinois, Indiana, southern Michigan, and Ohio are composed of stacked sequences of basal till, outwash, and lake sediment.

In Illinois, the very massive moraines are thought to have formed mainly by subglacial transport and deposition of deformable subglacial sediment that was being moved toward the terminus by rapidly flowing outlet glaciers perhaps fed by ice streams ([Johnson and Hansel, 1999](#)) with a thin cover of hummocky topography in the end moraine zone ([Curry et al., 2010](#)). In North Dakota, Minnesota, Wisconsin, and Northern Michigan, end moraines are very hummocky and indicate that ice near the margin became stagnant and was buried by thick supraglacial debris ([Ham and Attig, 1996](#); [Johnson and Clayton, 2003](#); [Johnson et al., 1995](#); [Parizek, 1969](#)). Using terminology introduced by [Gravenor and Kupsch \(1959\)](#), [Evans \(2009\)](#) interprets these moraines as ‘controlled moraines.’ He describes them as “supraglacial debris concentrations that become hummocky moraine upon deicing and possess clear linearity due to the inheritance of the former pattern of debris-rich folia in the parent ice.” This supraglacial material must have been elevated from the subglacial bed near the terminus because there are no nunataks that could have dropped debris onto this ice surface in the lowland terrain. Processes such as thrusting and shearing of basal debris as has been observed in subpolar glaciers may have been at work. Upon melting, which could have taken several thousand years, the complex topography of kettles, kames, and ice-walled lake plains formed.

Most of the deglaciated surface in Wisconsin is covered by drumlins and flutes. Other regions covered by extensive fields of drumlins include northcentral Minnesota, Michigan, New York, and New England. The subglacial landforms in these areas are covered by landforms formed during glacier recession such as push moraines, and larger recessional moraines ([Colgan et al., 2003](#)).

Ice-marginal Stagnation Along the Southern Margin of the Laurentide Ice Sheet

In areas along the lowland margins of the great ice sheets in North America and Europe, extensive areas of hummocky moraine suggest that at least in the ice-marginal zones of glaciers, ice wasted from the top down and became stagnant at the end of the last glaciations. Some of these zones were up to 20–40 km wide parallel to the flow direction of the glacier. Blocks of stagnant ice detached from active ice during glacier recession ([Evans, 2009](#); [Johnson and Clayton, 2003](#)). This probably indicates that polythermal glaciers interacted with permafrost. In the western Great Lakes region and the plains states of the United States, large hummocky controlled moraines suggest that ice near the margin was covered by supraglacial sediment and slowly melted out during deglaciation (see [Colgan et al., 2003](#)). Even in areas once thought to be dominated by temperate ice, recent evidence suggests that ice-marginal stagnation and down wasting occurred, forming a karstified landscape of shallow lakes and melting ice ([Curry et al., 2010](#)). The most common landforms produced by these phenomena are hummocks, ring forms, dump ridges, disintegration ridges, and ice-walled lake plains ([Colgan et al., 2003](#); [Evans, 2009](#); [Gravenor and Kupsch, 1959](#); [Johnson and Clayton, 2003](#); [Parizek, 1969](#)). In the Canadian Prairie region and plains region of the United States, there is evidence that large

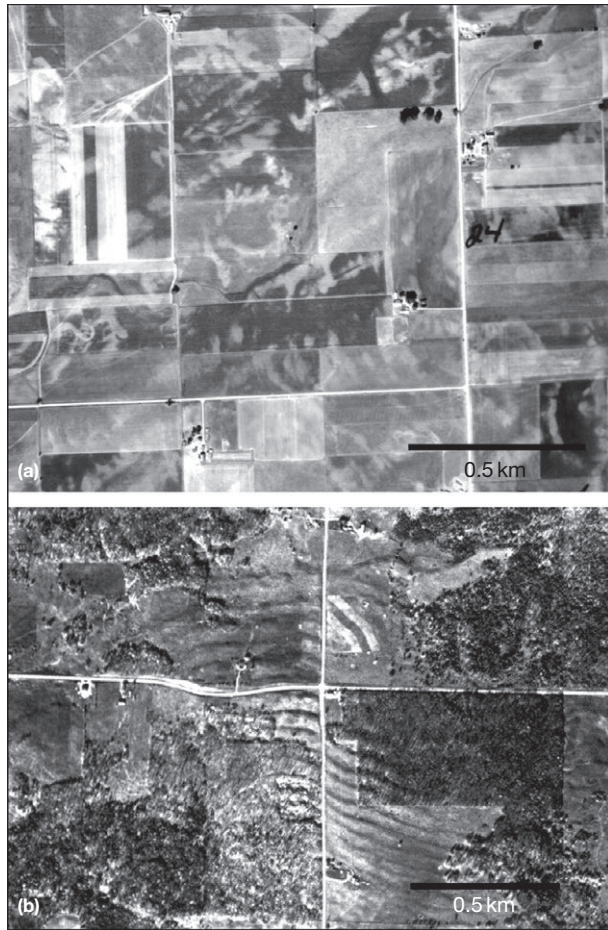


Figure 10 (a) Small push moraines that may have formed annually during the recession of the Green Bay lobe between 15 500 and 13 000 ^{14}C years BP. (b) Small push moraines created by the Wisconsin Valley lobe in north-central Wisconsin during deglaciation. Reproduced from Ham NR and Attig JW (2001) Minor end moraines of the Wisconsin Valley Lobe, north-central Wisconsin, USA. *Boreas* 30: 31–41.

areas of debris-rich stagnant ice covered the landscape during deglaciation in the form of aligned hummocks, ring forms, and disintegration ridges (Gravenor and Kupsch, 1959; Parizek, 1969, and papers in Evans 2003a,b).

Hummocks are rounded knobs, hillocks, or mounds occurring in widespread tracts (Figure 11). Hummocks are usually described as round to oval in map view, 15–400 m in diameter, and spaced from 150 to 500 m apart. The local relief of hummocks is usually described as between 2 and 70 m with most less than 20 m. Hummocks have been classified into categories based on relief, but a continuum of size and form may also exist. A large area of hummocks is commonly called hummocky topography.

Ring forms are circular or doughnut-shaped forms that commonly form a continuum with hummocks. They can be thought of as small hummocks with a central depression. These forms are thought to have formed when stagnant debris-covered ice melts and deposited supraglacial sediment. Ice-walled lake plains form when ice becomes stagnant and meltwater ponds in sinkholes on top of the glacier. The water in the supraglacial lakes melts its way to the bed in areas of

supraglacial debris near the margin. Sediment moved by streams and gravity deposit sediment in these lakes. Lacustrine, fluvial, and deltaic sediment may completely fill the sinkhole formed in the ice. Upon the melting of surrounding ice, the sediment may completely collapse to form a hummocky topography or it may form a plateau of lacustrine sediment surrounded by an ice-contact ridge. Ice-walled lake plains have been described in Canada, North Dakota (Figure 12), Minnesota, Wisconsin, Iowa, and as high kames in southern New England. Ice-walled lake plains are also found in northern Europe in Denmark, Sweden, Germany, Poland, and Russia.

Disintegration ridges (or crevasse-squeeze ridges) are linear ridges from 1 to 10 m high and 10 to 100 m wide that probably form when sediment is dumped into or squeezed into crevasses in stagnant ice. They commonly show the control of a network of crevasses and may display a box work pattern where the original ice blocks outlined by longitudinal and transverse crevasses may be easily visualized. Disintegration ridges are found both in Canada and the northern United States, as well as in northern Europe. A dump-ridge is a pile of debris that forms along the margins of stagnant ice. These ridges are found both in front of stagnant ice bodies and behind large masses of stagnant ice. They have been described in Wisconsin and Minnesota and are probably common wherever stagnant ice was covered by supraglacial debris.

The key environmental factor in developing the features described above is stagnation of supraglacial debris-covered ice near the margin of a glacier during recession. Some have suggested that ice in a zone 2–40 km wide near the ice margin becomes covered by debris and is detached from the actively retreating ice (Ham and Attig, 1996; Johnson and Clayton, 2003; Johnson et al., 1995). This debris-covered ice may then take several thousand years to melt completely. Pleistocene ice-cored moraines in Siberia and in the Mackenzie Delta region of Canada are still melting out and producing topography similar to the forms described above thousands of years after glacier recession.

Estimated Rates of Glacier Recession During the Late Pleistocene

Many workers have attempted to calculate the rate of late Pleistocene ice sheet recession from radiocarbon dating and end moraine mapping. Many of the calculated rates of recession are similar to those seen in modern glaciers today. Kaplan (1999) estimated that the southeastern margin of the Laurentide ice sheet retreated at a rate of about 20 m year^{-1} for the period from 14 000 to 13 000 ^{14}C years BP based on radiocarbon dates and numerous small ridges that formed along a retreating tide-water glacier margin. Gustavson and Boothroyd (1987) compiled a table showing estimates of the glacier recession rates of the southeastern Laurentide ice sheet in southern New England. Most of the rates calculated for the early part of the deglaciation before 13 000 ^{14}C years BP were between 50 and 100 m year^{-1} . Rates of recession calculated from small push moraines in Wisconsin are also in the range of $50\text{--}100 \text{ m year}^{-1}$.

Much higher rates of glacier recession have been calculated for areas in which ice may have stagnated during deglaciation. Clayton et al. (1985) calculated recession rates as high as

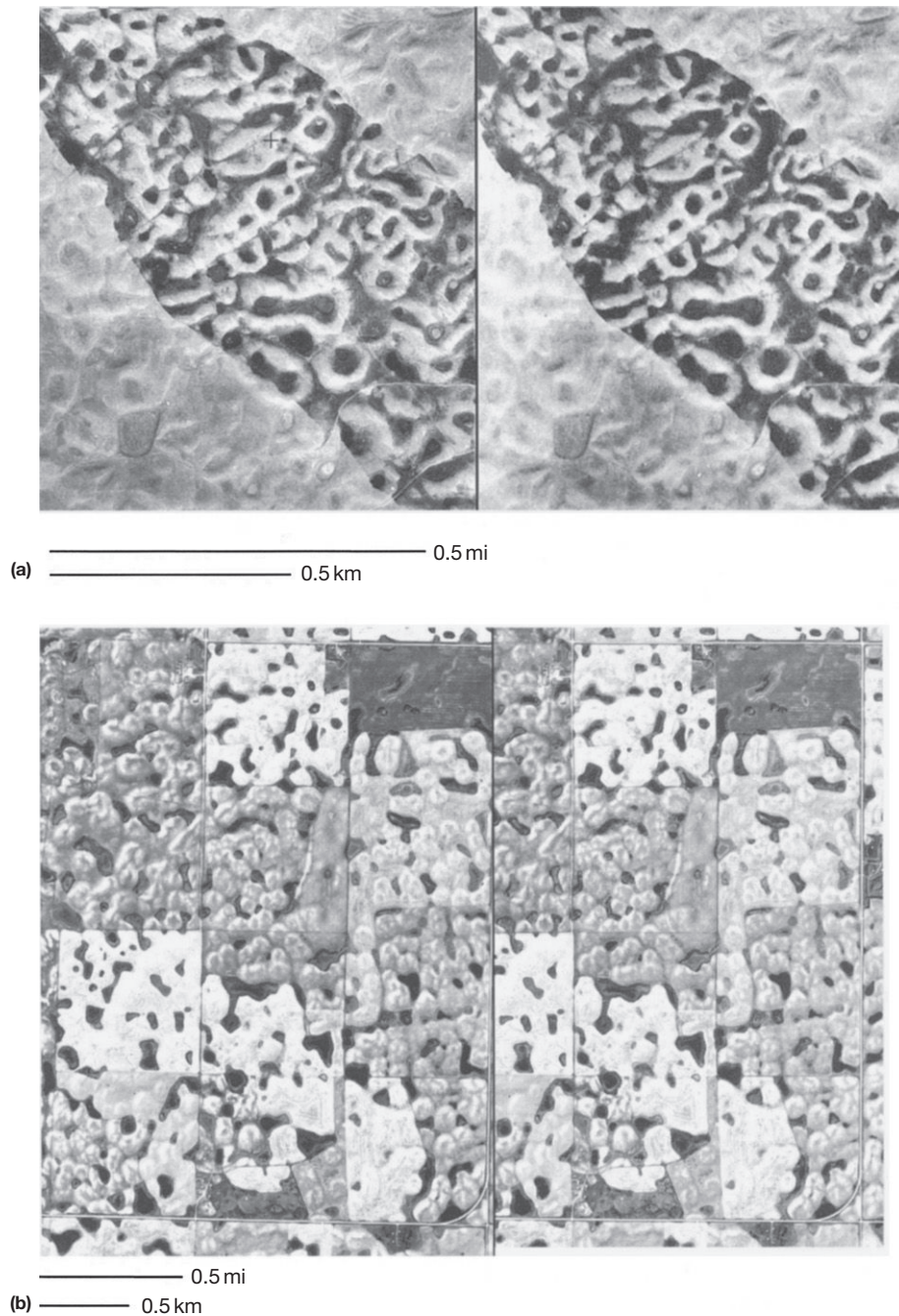


Figure 11 A stereo aerial photo pair of an ice-walled lake plain in Montrail County, North Dakota (BAL-4V-9 and 10, USDA). Ring forms and hummocks surround the ice-walled lake plain.

2000 m year⁻¹ on some of the southwestern lobes of the Laurentide ice sheet. They explained these high rates by regional stagnation of ice lobes after a surge.

Implications of Evidence of Glacier Recession

Landforms and sediments produced during glacier recession provide scientists with a unique perspective on the reconstruction of past glacier recession. As demonstrated earlier,

one way to understand these events is by comparing modern glacier environments and processes to landscapes formed in the past. By these studies and comparisons of past and present events, we can interpret the magnitude and sometimes estimate the rate of glacier recession. The biggest limitation in estimating accurate glacier recession rates is the uncertainty in dating methods (500–1500 years for radiocarbon ages before 10 000 years BP) and the lack of organic material available for radiocarbon dating. A more fundamental problem is that average recession rates do not account for small advances

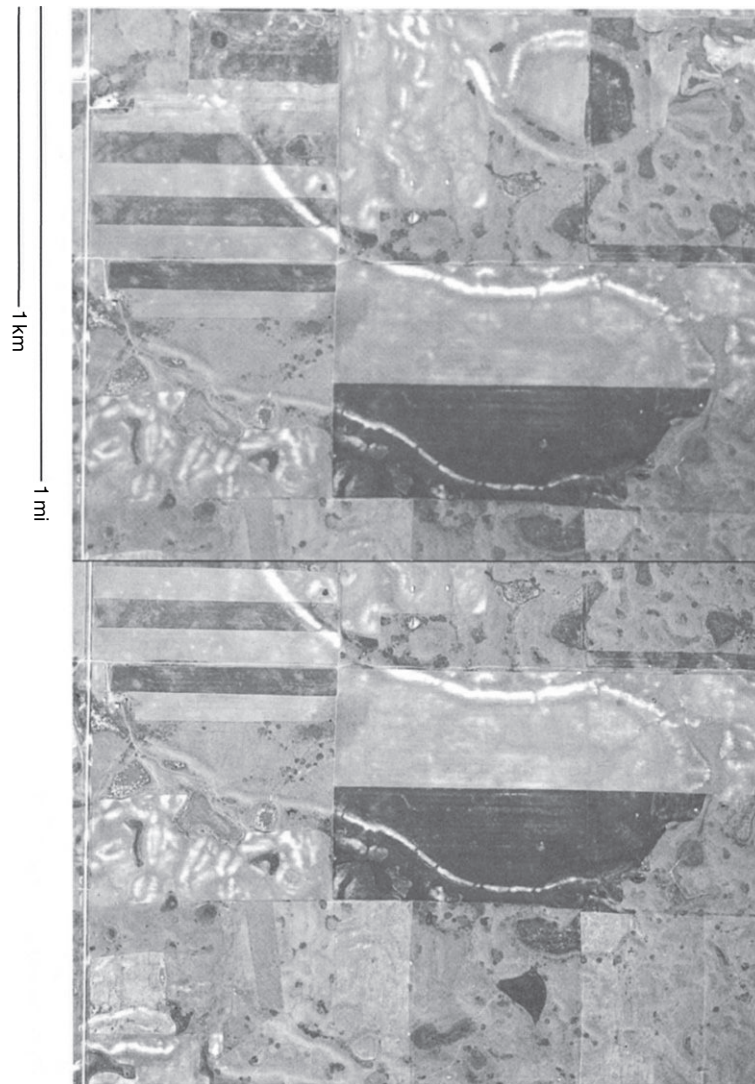


Figure 12 Stereo aerial photographs of an ice-walled lake plain in North Dakota.

during recession or still-stands. This can be clearly seen in Iceland where an average recession rate of $40\text{--}50\text{ m year}^{-1}$ for the period 1894–1998 clearly does not include documented still-stands and minor advances. Clearly, evidence for glacier recession can be used to both monitor modern glacier recession and reconstruct the style and rates of recession of past ice sheets, but it remains to be seen if we can directly compare modern rates with those of late Pleistocene glaciers.

See also: K/Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ Dating; **Glacial Landforms, Sediments:** Micromorphology of Glacial Sediments. **Glacial Landforms, Ice Sheets:** Trimlines and Paleounataks. **Glacial Landforms, Sediments:** Clast Form Analysis; Till Fabric Analysis. **Glaciations:** Late Pleistocene in South America; Overview. **Ice Cores:** Dynamics of the East Antarctic Ice Sheet. **Luminescence Dating:** Electron Spin Resonance Dating. **Permafrost and Periglacial Features:** Cryoturbation Structures. **Radiocarbon Dating:** Charcoal. **Sea Level Studies:** Sedimentary Indicators of Relative Sea-Level Changes – Low Energy.

References

- Bahr DB, Dyurgerov M, and Meier MF (2009) Sea level from glaciers and ice caps: A lower bound. *Geophysical Research Letters* 36: L0350. <http://dx.doi.org/10.1029/2008GL036309>.
- Boulton GS (1986) Push moraines and glacier contact fans in marine and terrestrial environments. *Sedimentology* 33: 677–698.
- Boulton GS (1987) A theory of drumlin formation by subglacial sediment deformation. In: Menzies J and Rose J (eds.) *Drumlin Symposium*, pp. 25–80. Rotterdam: Balkema.
- Broecker WS and Farrand WF (1963) Radiocarbon age of the Two Creeks Forest Bed, Wisconsin. *Geological Society of America Bulletin* 74: 795–802.
- Cazenave A and Chen JL (2010) Time-variable gravity from space and present-day mass redistribution in the Earth system. *Earth and Planetary Science Letters* 298(3–4): 263–274.
- Cazenave A, Lombard A, and Llovel W (2008) Present-day sea level rise: A synthesis. *Comptes Rendus Geoscience* 340(11): 761–770.
- Clayton L, Teller JT, and Attig JW (1985) Surging of the southwestern part of the Laurentide ice sheet. *Boreas* 14: 235–241.
- Colgan PM (1999) Reconstruction of the Green Bay lobe, Wisconsin, United States, from 26,000 to 13,000 radiocarbon years B.P. In: Mickelson DM and Attig JW (eds.) *Glacial Processes Past and Present*, pp. 137–150. Geological Society of America Special Paper 337.
- Colgan PM and Mickelson DM (1997) Genesis of streamlined landforms and flow history of the Green Bay lobe, Wisconsin, USA. *Sedimentary Geology* 111: 7–25.

- Colgan PM, Mickelson DM, and Cutler PM (2003) Ice marginal terrestrial landsystems: Southern Laurentide ice sheet margin. In: Evans DJA (ed.) *Glacial Landsystems*, pp. 111–142. London: Arnold.
- Cook AJ, Fox AJ, Vaughan DG, and Ferrigno JG (2005) Retreating Glacier Fronts on the Antarctic Peninsula over the past half-century. *Science* 308: 541–544.
- Curry BB, Konen ME, Larson TH, et al. (2010) The DeKalb mounds of northeastern Illinois as archives of deglacial history and postglacial environments. *Quaternary Research* 74: 82–90.
- Dyke AS, Andrews JT, Clark PU, et al. (2002) The Laurentide and Innuitian ice sheets during the last glacial maximum. *Quaternary Science Reviews* 21: 9–31.
- Dyke AS and Prest VK (1987) *Paleogeography of northern North America, 18,000–5,000 years ago*. Geological Survey of Canada. Map 1703A, scale 1:12,500,000.
- Dyrugorov MB and Meier MF (2000) Twentieth century climate change: Evidence from small glaciers. *Proceedings of the National Academy of Sciences of the United States of America* 97: 1406–1411.
- Dyrugorov M and Meier MF (2005) *Glaciers and the Changing Earth System: A 2004 Snapshot*. Occasional Paper 58, p. 118. Boulder, CO: Institute of Arctic and Alpine Research, University of Colorado.
- Evans DJA (2003a) *Glacial landsystems*. London: Arnold.
- Evans DJA (2003b) Ice-marginal terrestrial landsystems: Active temperate glacier margins. In: Evans DJA (ed.) *Glacial Landsystems*, pp. 12–43. London: Arnold.
- Evans DJA (2009) Controlled moraines: Origins, characteristics and palaeoglaciological implications. *Quaternary Science Reviews* 28: 183–208.
- Evans DJA and Twigg DR (2002) The active temperate glacial landsystem: A model based on Breiðamerkurjökull and Fjallsjökull, Iceland. *Quaternary Science Reviews* 21: 2143–2177.
- Gravenor CP and Kupsch WO (1959) Ice disintegration features in western Canada. *Journal of Geology* 67: 48–64.
- Grove JM (1988) *The Little Ice Age*. London: Routledge.
- Gustavson TC and Boothroyd JC (1987) A depositional model for outwash, sediment sources, and hydrologic characteristics, Malaspina Glacier, Alaska: A modern analogue for the southeastern margin of the Laurentide ice sheet. *Geological Society of America Bulletin* 99: 187–200.
- Ham NR and Attig JW (1996) Ice wastage and landscape evolution along the southern margin of the Laurentide ice sheet, north-central Wisconsin. *Boreas* 25: 171–186.
- Ham NR and Attig JW (2001) Minor end moraines of the Wisconsin Valley Lobe, north-central Wisconsin, USA. *Boreas* 30: 31–41.
- Johnson MD and Clayton L (2003) Supraglacial landsystems in lowland terrain. In: Evans DJA (ed.) *Glacial Landsystems*, pp. 228–258. London: Arnold.
- Johnson WH and Hansel AK (1999) Wisconsin Episode glacial landscape of central Illinois: A product of subglacial deforming processes. In: Mickelson DM and Attig JW (eds.) *Glacial Processes Past and Present*, pp. 122–135. Geological Society of America Special Paper 337.
- Johnson MD, Mickelson DM, Clayton L, and Attig JW (1995) Composition and genesis of glacial hummocks western Wisconsin, USA. *Boreas* 24: 97–116.
- Johnson MD, Schomacker A, Benediktsson IO, Geiger AJ, Ferguson A, and Ingólfsson O (2010) Active drumlin field revealed at the margin of Múlajökull, Iceland: A surge-type glacier. *Geology* 38: 943–946.
- Kaplan MR (1999) Retreat of a tidewater margin of the Laurentide ice sheet in eastern coastal Maine between ca. 14,000 and 13,000 ¹⁴C yr B.P. *Bulletin of the Geological Society of America* 111: 620–632.
- Meier MF, Dyrugorov MB, and McCabe GJ (2003) The health of glaciers: Recent changes in glacier regime. *Climate Change* 59: 123–125.
- Oerlemans J (2005) Extracting a climate signal from 169 glacier records. *Science* 308: 675–677.
- Parizek RR (1969) Glacial ice-contact rings and ridges. *Geological Society of America Special Paper* 123: 49–102.
- Paterson WSB (1994) *The Physics of Glaciers*. Oxford: Pergamon.
- Rignot E, Rivera A, and Casassa G (2003) Contribution of the Patagonian ice fields of South America to sea level rise. *Science* 302: 437–438.
- Sasgen I, Martinec Z, and Bamber J (2010) Combined GRACE and InSAR estimate of West Antarctic mass loss. *Journal of Geophysical Research* 115: F04010. <http://dx.doi.org/10.1029/2009JF001525>.

Relevant Websites

- <http://www.glims.org/> – Global land Ice Monitoring from Space (GLIMS).
- <http://twitter.com/IceBridge> – IceBridge (NASA).
- <http://icesat.gsfc.nasa.gov/> – IceSat (NASA).
- <http://nsidc.org/> – National Snow and Ice Data Center.
- <http://www.geo.uzh.ch/microsite/wgms/> – World Glacier Monitoring System.

Glacial Landsystems

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Introduction

The earliest glacial landsystem-type investigations include those of [Speight \(1963\)](#), who mapped and grouped landforms according to common origin and age, and [Clayton and Moran \(1974\)](#), who linked process and form in assessing the spatial distribution of glacial features. Glacial landsystems per se were introduced by [Fookes et al. \(1978\)](#) to provide engineers with genetic classifications of glaciogenic landform–sediment assemblages and included the till plain, glaciated valley, and fluvio-glacial and ice-contact deposit landsystems. The component parts of landsystems are land units (e.g., drumlin or fluting field, Rogen moraine belt, and suite of recessional push moraines) and land elements (e.g., individual drumlins, flutings, eskers, kame terraces, and moraines), with the landsystem thus being a recurrent pattern of genetically linked land units. This was refined by [Eyles \(1983a,b\)](#), [Eyles and Menzies \(1983\)](#), and [Paul \(1983\)](#), who introduced the subglacial, supraglacial, and glaciated valley landsystems, and thereby emphasized the location of glaciogenic deposition as the principal factor in land-system classification. Recent developments in the glacial landsystem concept ([Evans, 2003a](#)) have addressed the complexity of glacial depositional environments and relate the variability in landform–sediment assemblages to both the location of deposition and the style of glaciation. Additionally, greater use is now being made of contemporary glacial environments, where landscape evolution can be monitored, to provide modern landsystem analogs for the interpretation of ancient glaciated terrain. Meticulous mapping of glacial landsystems also often informs us of the fact that the characteristics of glacier snouts have changed over the period that they have retreated from a landscape. This is manifest in either land-system overprinting, often associated with glacier readvance (e.g., [Krüger, 1994](#); [Wilson and Evans, 2001](#)), or spatial and temporal change, the latter referring to sequential changes in signature with distance along a single foreland or drainage basin (e.g., [Colgan et al., 2003](#); [Evans, 2011](#); [Evans et al., 2008, 2010](#)).

Landsystem Modern Analogs

Landsystem classifications of modern glacial environments, where landscape evolution can be monitored, provide analogs for the interpretation of ancient glaciated terrain ([Dyke and Savelle, 2000](#); [Evans and Twigg, 2002](#); [Gustavson and Boothroyd, 1987](#); [Kjær and Krüger, 2001](#); [Krüger, 1994](#); [Price, 1969](#)). This type of research on modern glaciers has led to the identification of landform–sediment suites indicative of specific styles of glaciation, including plateau ice fields ([Rea and Evans, 2003](#); [Rea et al., 1998](#)), arid polar glacier margins ([Fitzsimons, 2003](#)), and ice dynamics such as surging glaciers ([Evans and Rea, 1999, 2003](#)) and ice streams

([Clark et al., 2003](#); [Stokes and Clark, 1999, 2001](#)). Once a landform–sediment suite pertaining to a single period of glacier occupancy or activity can be identified, it often becomes possible to differentiate overprinted signatures and landscape palimpsests (e.g., [Clark, 1993, 1999](#); [Dyke and Morris, 1988](#); [Krüger, 1994](#); [Wilson and Evans, 2001](#)). Examples of some modern analogs are reviewed here, followed by a discussion of their application to ancient glaciated landscapes. The terrestrial examples used here are clearly not exhaustive, and the reader is directed to [Evans \(2003a\)](#) and [Benn and Evans \(2010\)](#) for a more comprehensive coverage of the wide range of glacial landsystems that have thus far been identified in both terrestrial and offshore settings.

Active Temperate Glacial Landsystem

Active temperate glacier margins are capable of forward momentum even during overall recession, and this is manifested in the small winter readvances that characterize receding outlet glaciers in places such as Iceland ([Boulton, 1986](#); [Krüger, 1995](#); [Sharp, 1984](#)). Details of the glaciology and associated sediment and landform production at the southern Icelandic glacier snouts are presented in numerous publications ([Benn, 1995](#); [Boulton, 1986](#); [Boulton and Eyles, 1979](#); [Boulton and Hindmarsh, 1987](#); [Evans and Hiemstra, 2005](#); [Krüger, 1985, 1993, 1994](#); [Sharp, 1984](#)), and it is from this expansive literature that a landsystem model for active temperate glaciers has been compiled ([Evans and Twigg, 2002](#); [Evans, 2003b, 2005](#)). This model recognizes three dominant depositional domains with characteristic landform–sediment assemblages ([Figure 1](#)). First, the marginal morainic domain consists of areas of extensive low-amplitude marginal dump, push, and squeeze moraines ([Evans and Twigg, 2000](#); [Evans et al., 1999](#); [Price, 1970](#)) often recording annual ice recession, although short periods of readvance or stability can produce larger push moraines due to stacking of submarginally frozen sediment slabs ([Evans and Hiemstra, 2005](#); [Krüger, 1985, 1993, 1994, 1996](#)), or dump, squeeze, and push mechanisms operating at the same location over several years. Second, the subglacial domain includes assemblages of flutings, drumlins, and overridden push moraines that occur between ice-marginal depocenters and are linked to subglacial deformation and lodgement processes ([Benn, 1995](#); [Benn and Evans, 1996](#); [Boulton, 1987](#); [Boulton and Hindmarsh, 1987](#); [Evans and Hiemstra, 2005](#); [Hart, 1995](#)). Third, the glaciofluvial and glaciolacustrine domain includes (a) the deposits of proglacial outwash streams that drain either away from glacier margins to produce sandur (outwash) fans or parallel to the margin to produce ice margin-parallel outwash tracts and kame terraces, (b) the glaciolacustrine depocenters that have developed in large lake-filled depressions uncovered by glacier recession, and (c) simple and complex anabranching eskers that often grade into recessional ice-contact fans and areas of pitted outwash. In addition to

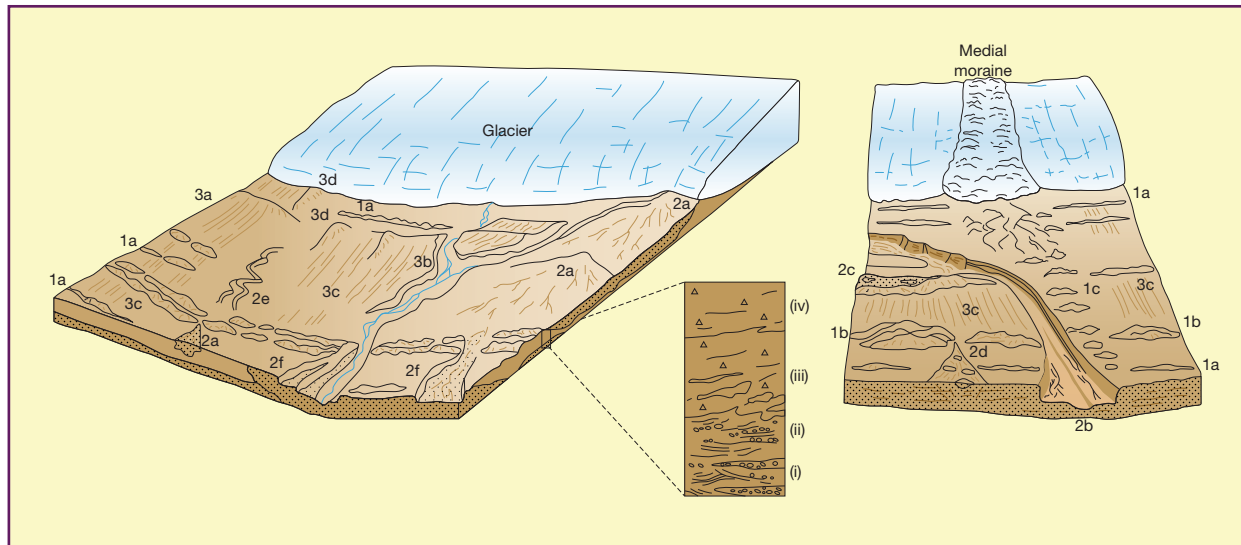


Figure 1 The active temperate glacial landsystem. Landforms are numbered according to their domain: 1, morainic domain; 2, glaciofluvial domain; and 3, subglacial domain): 1a, small, often annual, push moraines; 1b, superimposed push moraines; 1c, hummocky moraine; 2a, ice-contact sandur fans; 2b, spillway-fed sandur fan; 2c, ice margin-parallel outwash tract/kame terrace; 2d, pitted sandur; 2e, eskers; 2f, entrenched ice-contact outwash fans; 3a, overridden (fluted) push moraines; 3b, overridden, readvance ice-contact outwash fan; 3c, flutes; and 3d, drumlins. The idealized stratigraphic section log shows a typical depositional sequence recording glacier advance over glaciofluvial sediments comprising (i) undeformed outwash; (ii) glaciectonized outwash/glacitectonite; (iii) massive, sheared till with basal inclusions of preadvance peat and glaciofluvial sediment; and (iv) massive sheared till with basal erosional contact. Modified from Krüger J (1994) Glacial processes, sediments, landforms and stratigraphy in the terminus region of Myrdalsjökull, Iceland. *Folia Geographica Danica* 21: 1–233; Evans DJA and Twigg DR (2002) The active temperate glacial landsystem: A model based on Breidamerkurjökull and Fjallsjökull, Iceland. *Quaternary Science Reviews* 21: 2143–2177; Evans DJA (ed.) (2003a) *Glacial Landsystems*. London: Arnold.

the three major depositional domains, anomalies are to be expected due to the signatures of erratic/noncyclic or site-specific processes, including jökulhlaup features (Fay, 2002; Maizels, 1997; Russell and Knudsen, 1999).

Glaciated Valley Landsystem

The glaciated valley landsystem was first compiled by Boulton and Eyles (1979) and Eyles (1983b) in order to encompass the landform–sediment associations produced by valley glaciers. Owen and Derbyshire (1989) highlighted the large volumes of supraglacial debris on Karakoram glaciers and illustrated the fact that valley glaciers do not always deposit clear, inset sequences of proglacially constructed latero-frontal moraines as depicted in the Boulton and Eyles (1979) model. Benn et al. (2003) subsequently demonstrated the huge variability in landform and sediment assemblages in highland settings due to their occurrence at every latitude on the Earth's surface, and they proposed a model that comprises a family of landsystems (Figure 2). Recurrent features identified by Benn et al. include (a) the strong influence of topography on glacier morphology, sediment transport paths, and depositional basins; (b) the important role of supraglacial debris sources; and (c) the rate of sediment transport from the glacier to the proglacial environment via the glaciofluvial system. Their family of landsystems lie on a continuum between the two extremes of 'clean' glaciers, with limited supraglacial debris, and debris-covered glaciers/rock glaciers. Additionally, the relief of upland landscapes often results in temporary lake production during glaciation, either by ice damming of valleys or by moraine

damming. The latter are potentially dangerous features in mountain terrain due to their tendency to catastrophically burst through moraine dams and produce glacial lake outburst floods (Cenderelli and Wohl, 2003; Clague and Evans, 2000; Kershaw et al., 2005). The different members of the glaciated valley landsystem family identified by Benn et al. (Figure 2) are defined by the efficiency of glaciofluvial transport between the glacier and its proglacial area. Efficient transport defines a 'coupled' landsystem association in which moraine development is limited and an outwash head/aggrading sandur (Kirkbride, 2000) is common, as typically occurs in the humid mountain ranges of Alaska and New Zealand. Inefficient transport defines a 'decoupled' landsystem in which repeated superposition of large moraines occurs around the margin of a debris-covered glacier, as is common in high-altitude, semiarid to arid mountain ranges such as the central and northern Andes and High Asia.

Plateau Ice Field Landsystem

Plateaus are the loci of glacier initiation and expansion during cooling events (Ives et al., 1975) and tend to be a haven for glaciers during interglacials and interstadials, depending on the latitude and altitude of the plateau and the global climate characteristics at the time (Manley, 1955; Rea et al., 1998). Precipitous drops at the margins of plateaus lead to the development of 'reconstituted glaciers' (Benn and Lehmkuhl, 2000) through dry calving at the plateau cliff edge (Gellatly et al., 1986). During the more advanced stages of local-scale glaciation, glaciers develop in the surrounding valley bottoms, and it

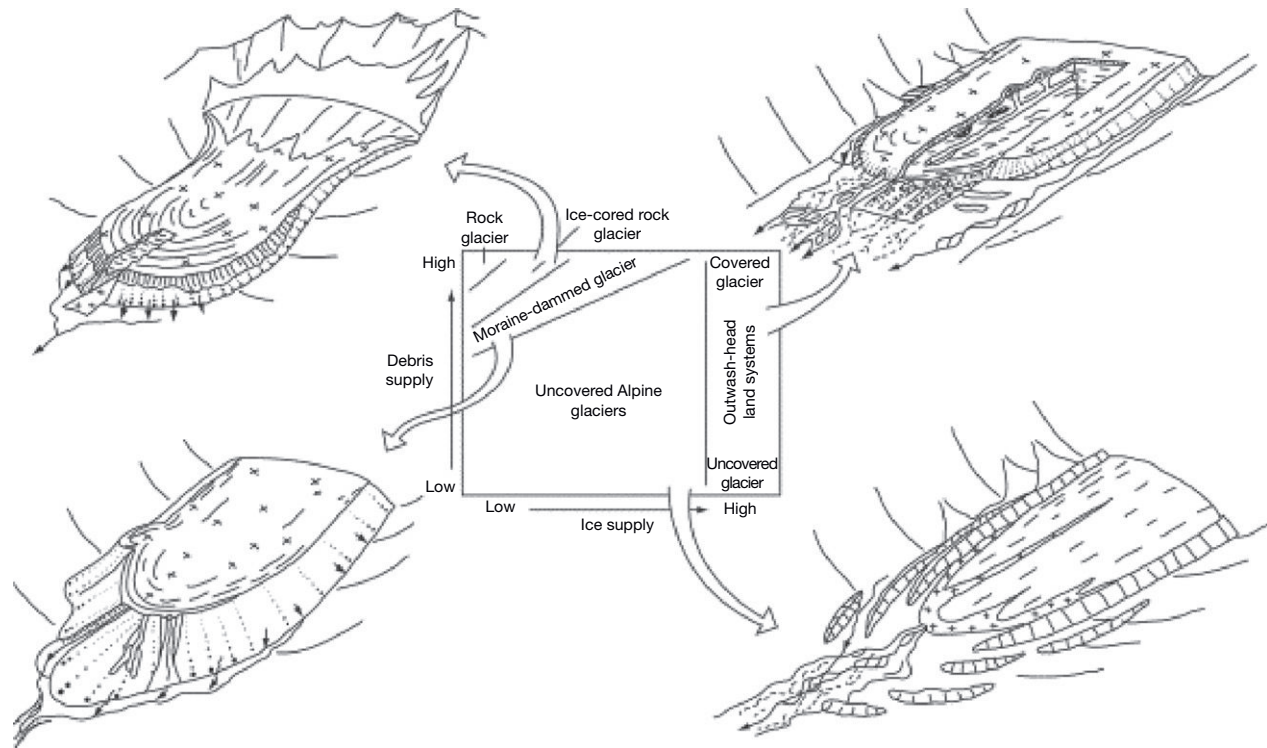


Figure 2 Constraints on landsystem development around valley glaciers, showing development pathways to four landsystem associations. Reproduced from Benn DI, Kirkbride MP, Owen LA, and Brazier V (2003) Glaciated valley landsystems. In: Evans DJA (ed.) *Glacial Landsystems*, pp. 372–406. London: Arnold, with permission.

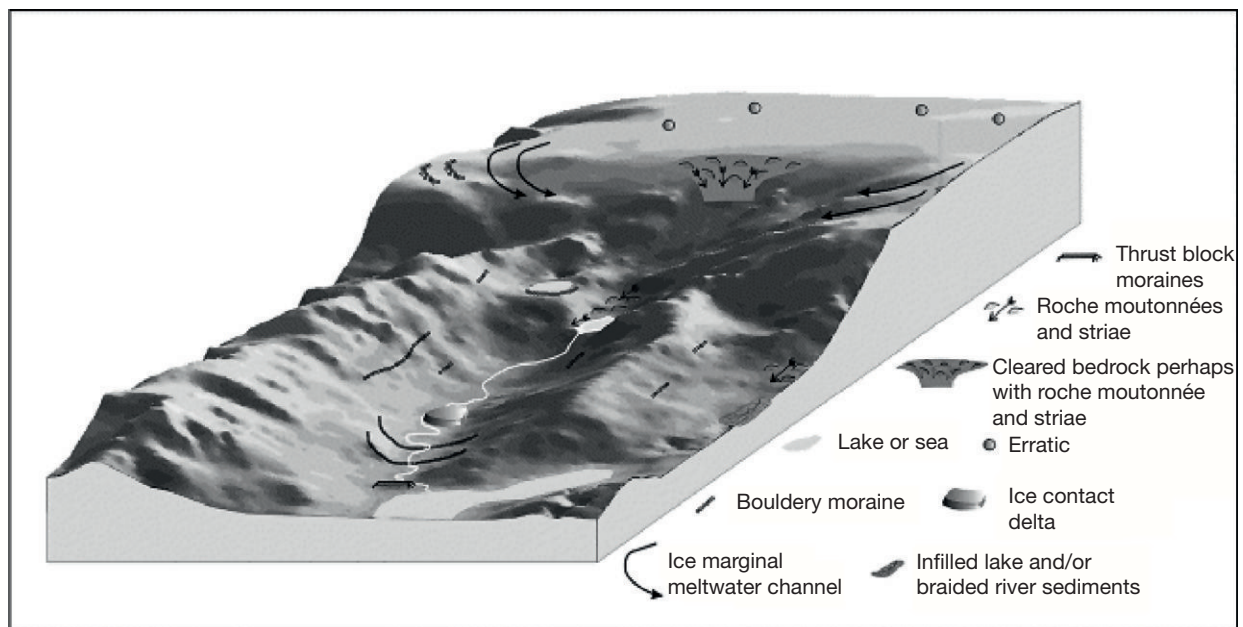


Figure 3 The plateau ice field landsystem. Reproduced from Rea BR and Evans DJA (2003) Plateau icefield landsystems. In: Evans DJA (ed.) *Glacial Landsystems*, pp. 407–431. London: Arnold, with permission.

is during this stage that the most significant landforms are likely to be produced; in most instances, the surrounding valleys contain the bulk of the landform–sediment assemblages used in reconstructing the former glacier cover (Rea

and Evans, 2003). On plateau summits, the thermal and dynamic properties of the ice determine the geomorphic signature, which in many instances, especially where ice was cold based, may be very subtle or nonexistent (Dyke, 1993).

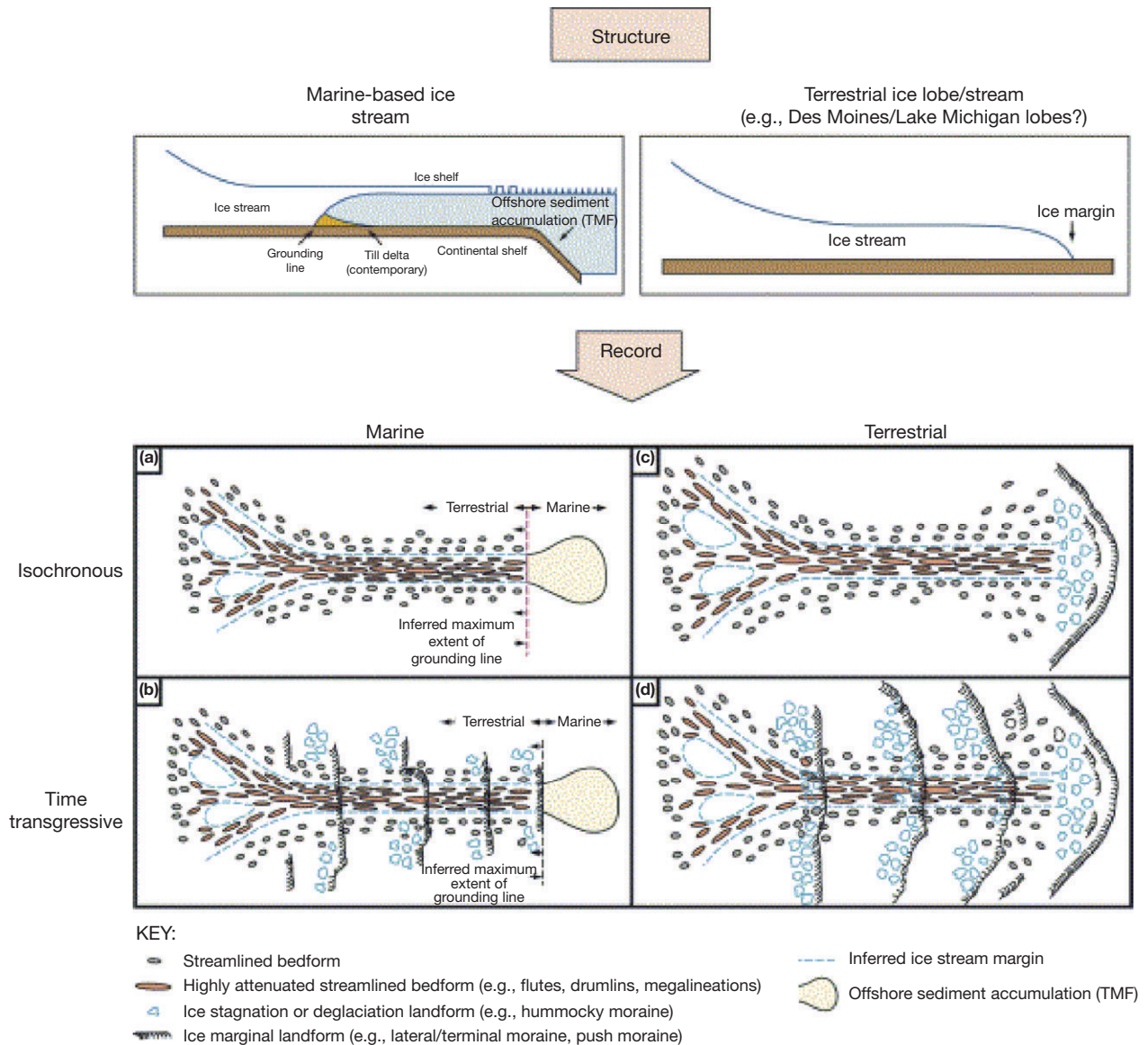


Figure 4 A landsystem model for ice streams showing the four end members of marine-based and terrestrial systems, which are either isochronous/rubber stamped or time transgressive/smudged. Reproduced from Clark CD and Stokes CR (2003) Palaeo-ice stream landsystem. In: Evans DJA (ed.) *Glacial Landsystems*, pp. 204–227. London: Arnold, with permission.

The assemblage of landforms produced by plateau ice fields in the Canadian High Arctic, northern Norway, and Iceland has been used to compile a plateau ice field landsystem (Rea and Evans, 2003; Rea et al., 1998; Figure 3). Plateau summits may contain nothing more than deglacial meltwater channels cutting through preexisting regolith or blockfield if the former ice cover was cold based. Lateral meltwater channels produced by the retreat of outlet glaciers onto plateaus are characteristic of recession by cold-based ice (Dyke, 1993; Maag, 1969). Evidence for localized erosion by warm-based ice (e.g., striated bedrock and roches moutonnées, and the removal of blockfield) occurs in the topographic depressions at plateau edges where ice flow velocities increase due to drawdown into surrounding valleys. Moraines may also be found on plateau edges or leading onto the plateau from the surrounding valley

head outlets if warm-based conditions exist locally, but significant till covers do not form due to the commonly short transport distances to plateau outlets and the dynamics of the plateau ice cover (Evans et al., 2006a). If the ice is cold based throughout, lateral and frontal moraines will be found only in the valleys. Valley moraines tend to be dominated by large, cobble- to boulder-sized, angular material even where there is evidence of basal sliding, indicating that the dominant debris source is rockfall. Many latero-frontal moraines in the valley heads that surround the plateaus are ice cored and display within-valley asymmetry (Benn, 1989). This reflects the distribution of bedrock-free faces and the concomitant variability in rockfall/avalanche and debris flow activity. Active free faces may provide sufficient debris to produce supraglacial lateral moraines. In some circumstances, these may develop into

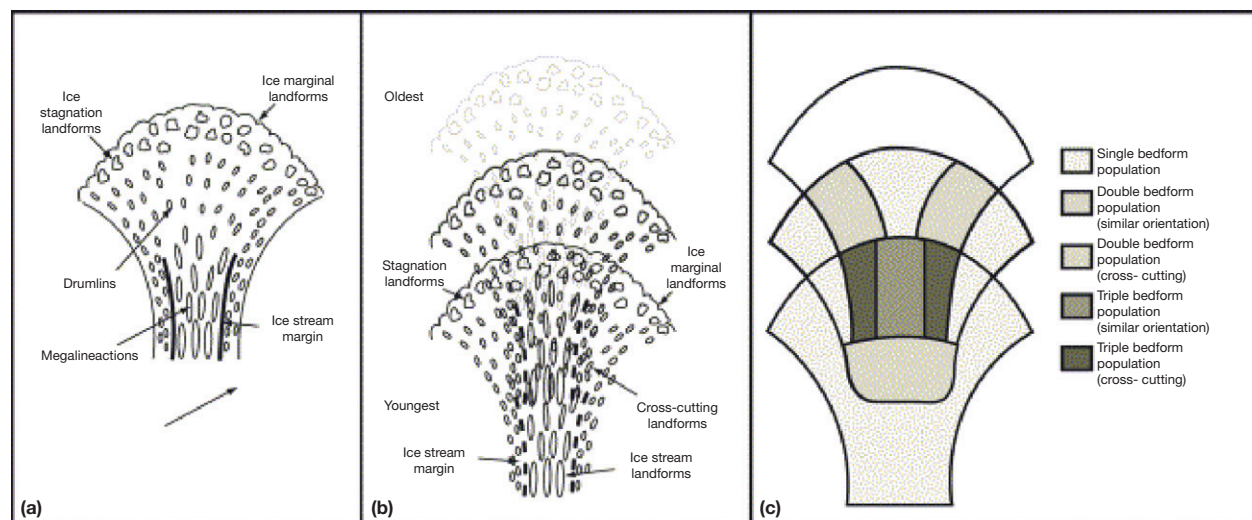


Figure 5 Paleo-ice stream imprints. (a) The imprint of ice stream shut down or the rubber-stamped imprint. (b) The overprinted sequence of a continuously operating ice stream during deglaciation or the smudged imprint. (c) Schematic diagram showing the distribution of different cross-cutting bedform imprints associated with a smudged landsystem. Reproduced from Clark CD and Stokes CR (2003) Palaeo-ice stream landsystem. In: Evans DJA (ed.) *Glacial Landsystems*, pp. 204–227. London: Arnold, with permission.

rock-glacierized ice-cored moraines that may persist after deglaciation of the valley heads (Evans, 2005). The valley floors are characterized by thin, generally patchy tills, which locally thicken in depressions.

Ice Stream Landsystem

Remotely sensed observations on the processes operating at the beds of ice streams in the Antarctic Ice Sheet (Kamb, 2001) combined with evidence of landform–sediment assemblages on the Antarctic continental shelf, across which ice streams formerly flowed, have informed recent compilations of paleo-ice stream landsystems (Canals et al., 2000). Ice streams are spatially restricted regions (arteries) of fast-flowing ice in an ice sheet and are readily identifiable by their surface structures and velocities. A landsystem model for ice streams has been developed by Stokes and Clark (1999) and Clark and Stokes (2003) based predominantly on the subglacial bedform evidence for fast-flowing ice and validated by glaciological data on modern ice streams (Figure 4). The bedform evidence comprises mega-scale glacial lineations (MSGs), which are elongate stream-lined ridges of sediment similar to flutes and drumlins and thought to be indicative of high strain rates and fast ice velocity. The pattern of subglacial bedforms should reflect the dominant flow patterns within an ice stream in that they should converge in an onset zone and lead into a much narrower and well-defined trunk zone where highly attenuated streamlined bedforms (MSGs) record the rapid velocity of the ice stream trunk. The higher elongation ratios of MSGs occur in the trunk rather than in the convergent onset zone, and the highest elongation ratios are in the central axis of the trunk. Additionally, elongation ratios in marine-terminating ice streams should steadily increase downstream toward the grounding line, but in terrestrially terminating ice streams, they decrease toward the outer margins of the lobate terminus. Ice streams also display abrupt lateral margins where fast ice abuts slower

moving ice. This is reflected in the subglacial bedform record, in which a high density of bedforms within the ice stream ceases abruptly. Sediment may concentrate at this boundary, producing an ice stream shear margin moraine (Dyke and Morris, 1988; Dyke et al., 1992; Stokes and Clark, 2002). At the terminus of ice streams, it should be expected that large depocenters will accumulate, such as submarine ‘till deltas’ (Alley et al., 1989; Vorren and Laberg, 1997) and grounding line wedges (Ó Cofaigh et al., 2003; Powell and Domack, 1995). Because ice streams switch on and off and dominant ice flow trajectories change during the life cycle of an ice sheet, ice stream imprints may crosscut each other (Dyke and Morris, 1988). Two end member landsystems are envisaged by Clark and Stokes (2003); Figure 5 to account for possible landform records of paleo-ice streams: (a) the ‘rubber-stamped’ imprint, produced when an ice stream imprint is produced at the same time and represents a snapshot of ice stream activity at a point in time when the ice stream switched off, and (b) the ‘smudged’ or overprinted, time transgressive imprint, produced when an ice stream operates throughout several cycles of advance and retreat or is continuously operating during margin retreat and earlier imprints are modified and overprinted by the younger ice flow patterns.

Applications of Modern Analogs to Quaternary Glacial Landscapes

Spatial and Temporal Change in Active Temperate and Polythermal Glacial Landsystems

Evidence of the recession of a former active temperate glacier lobe at the margin of the southwest Laurentide Ice Sheet has been reported by Evans et al. (1999, 2006b) and Benn and Evans (2006), manifest in the deglacial landforms of the McGregor Lake area of southern Alberta, Canada. Active temperate recession is recorded by numerous closely spaced,

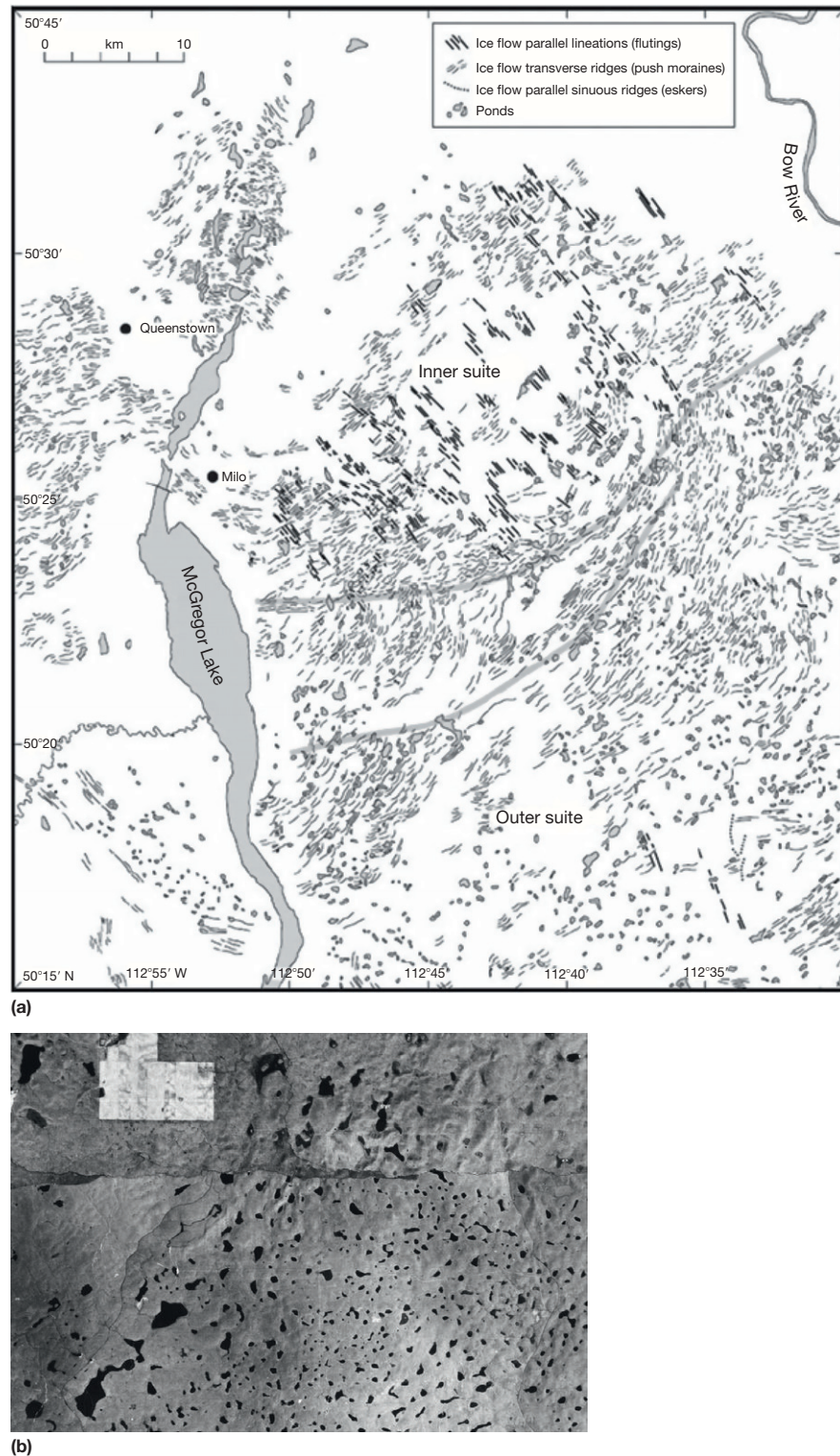


Figure 6 The McGregor Lake moraine belt in Alberta, Canada: (a) glacial geomorphological map of the area showing three partially overprinted suites of moraines and associated landforms (reproduced from [Evans DJA, Rea BR, Hiemstra JF, and Ó Cofaigh C \(2006a\)](#) A critical assessment of subglacial mega-floods: A case study of glacial sediments and landforms in south-central Alberta, Canada. *Quaternary Science Reviews* 25: 1638–1667; [Evans DJA \(2009\)](#) Controlled moraines: origins, characteristics and palaeoglaciological implications. *Quaternary Science Reviews* 28: 183–208); (b) aerial photograph extract of the outermost moraine suite in (a), showing less continuous moraine ridges, numerous kettle holes, and no flutings in the lower right half of the image. The upper left half of the image shows part of the intermediate moraine suite and its larger and more continuous moraine ridges and lack of kettle holes.

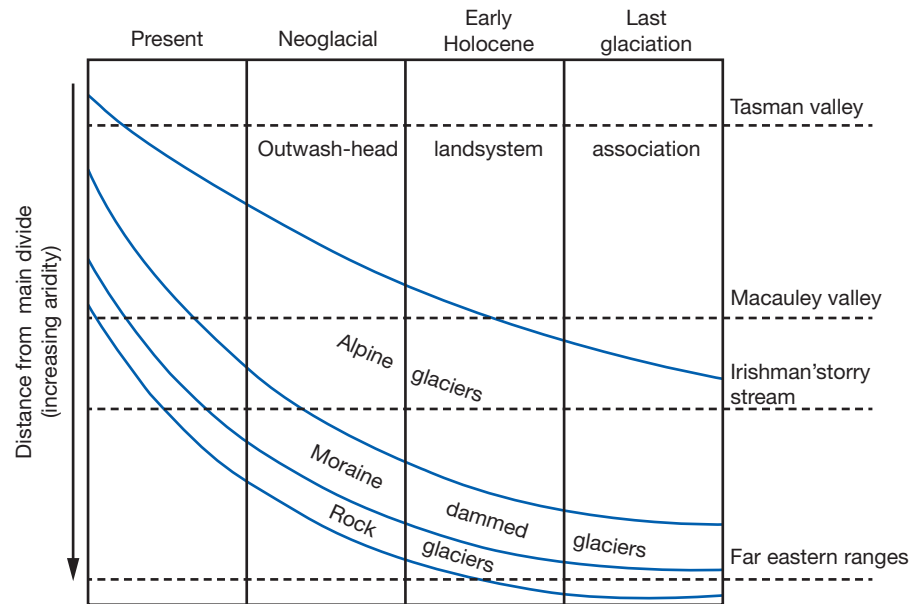


Figure 7 The spatial and temporal occurrence of various members of the glaciated valley landsystem in the Ben Ohau Range, New Zealand, showing the gradual trend from coupled to uncoupled systems. Reproduced from [Benn DI, Kirkbride MP, Owen LA, and Brazier V \(2003\)](#) Glaciated valley landsystems. In: Evans DJA (ed.) *Glacial Landsystems*, pp. 372–406. London: Arnold, with permission.

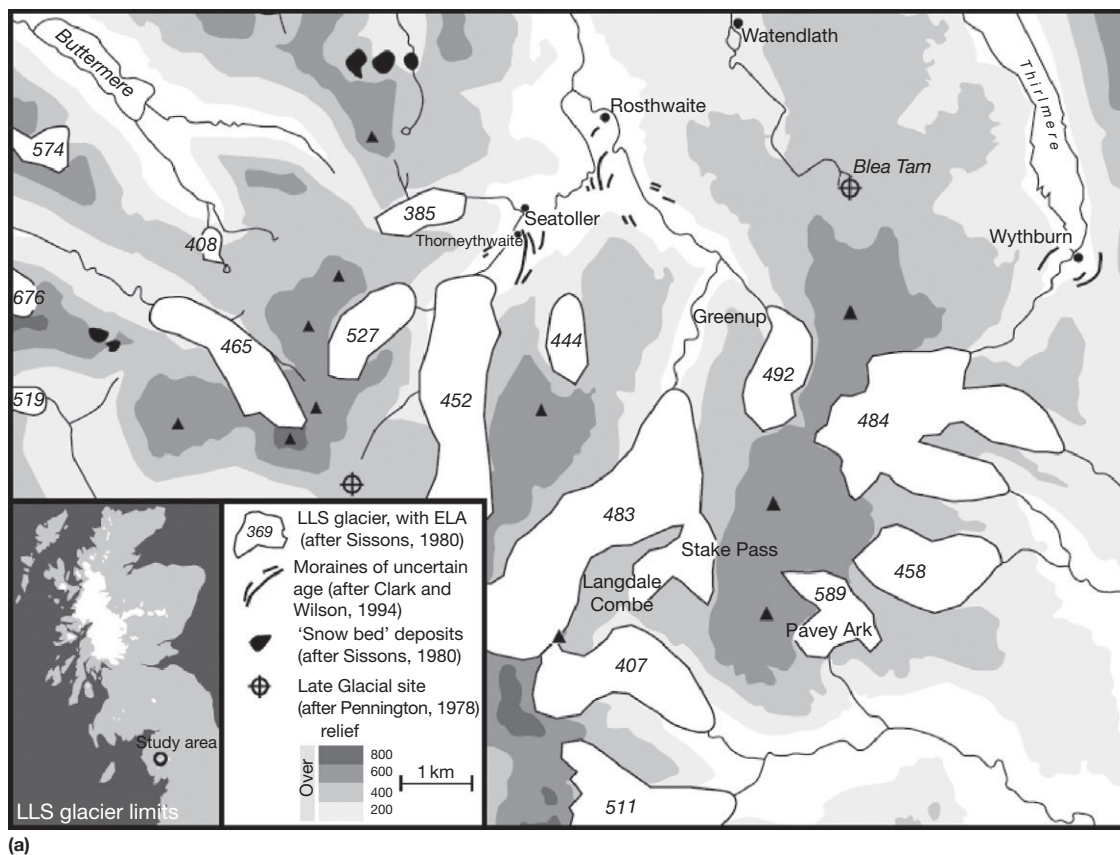


Figure 8 Contrasting paleoglaciological reconstructions of Younger Dryas glacier cover in the English Lake District: (a) alpine style of valley head and cirque glaciers proposed by [Sissons \(1980\)](#); (b) plateau ice field style of glaciation reconstructed by [Rea et al. \(1998\)](#) and [McDougall \(2001\)](#) informed by modern plateau ice field analogs. Reproduced from [McDougall DA \(2001\)](#) The geomorphological impact of Loch Lomond (Younger Dryas) Stadial plateau icefields in the central Lake District, northwest England. *Journal of Quaternary Science* 16: 531–543, with permission.

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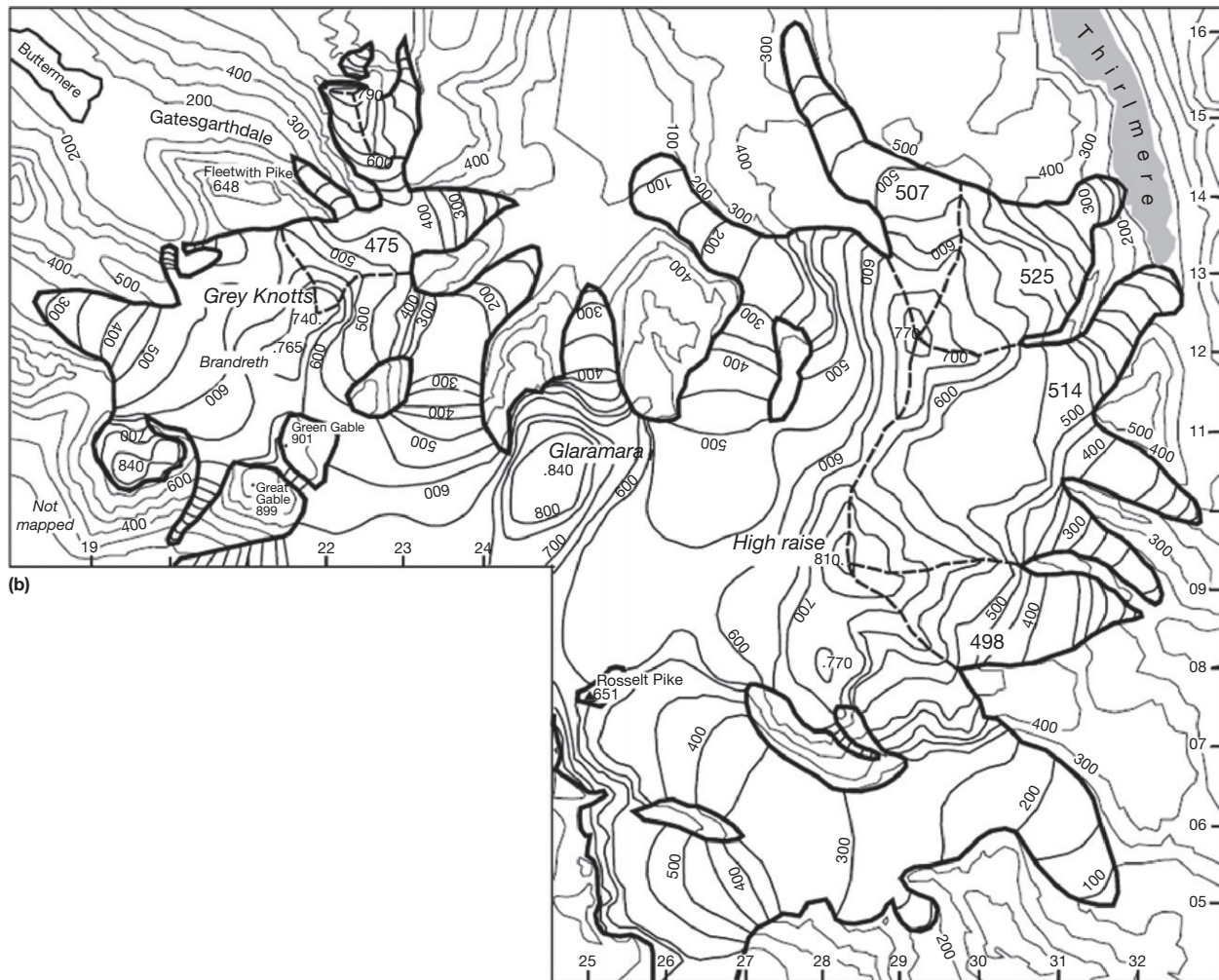


Figure 8 (Continued)

low-amplitude moraine ridges that are inset by zones of flutings trending obliquely to individual moraine crests. This assemblage comprises the innermost of three overprinted 'landform suites' in a larger and more complex moraine belt (Evans, 2009; Figure 6(a)). The landform suites constitute a form of landsystem zonation and appear to record changing thermal regimes in the receding and occasionally readvancing ice sheet margin. In contrast to the innermost, active temperate suite, the outermost suite is characterized by less continuous moraine ridges, numerous kettle holes, and no flutings (Figure 6(b)). The terrain lying between the innermost and outermost suites contains some larger, more continuous moraine ridges, where densely spaced kettle holes are less common. Based upon these characteristics, the outermost suite is interpreted as the remnants of controlled moraine, produced at a time when the ice sheet margin was cold based, whereas the innermost suite appears to be consistent with the interpretation proposed by Evans et al. (2006b) that entails construction of recessional push moraines by a predominantly warm-based ice

margin. Evans (2009) has hypothesized that the intermediate suites of landforms relate to a period when ice-marginal conditions were changing from polythermal to temperate, and patchy permafrost led to the construction of larger push moraines in some areas.

Glaciated Valley Landsystem

The concept of a family of glaciated valley landsystems has been applied by Benn et al. (2003) to the Ben Ohau Range in New Zealand (Brazier et al., 1998). Holocene climate change in an area with a steep precipitation gradient has resulted in rising equilibrium altitudes and a concomitant response in glacial landsystem development due to altering debris supply and transport. Present-day landsystem distribution demonstrates the crucial role that aridity plays in local factors for landform development. The outwash-dominated landsystems of the major Mount Cook valley glaciers occupy large catchments to the north of the range. Whereas relatively clean Alpine

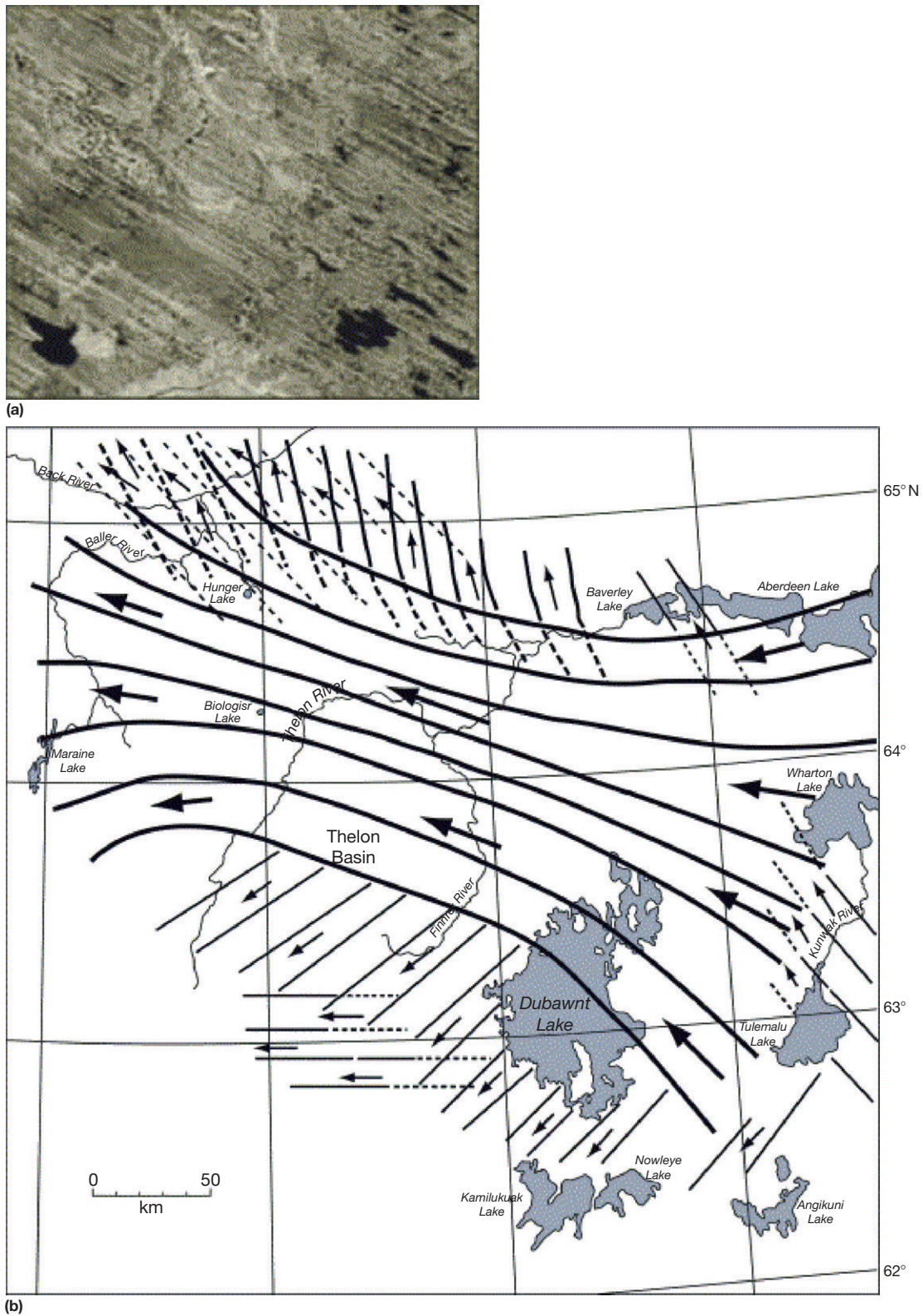


Figure 9 The Dubawnt Lake paleo-ice stream. (a) Satellite image of megascale glacial lineations indicating fast ice flow. (b) Map of the paleo-ice stream showing the extent of former onset zone, trunk, and terminal divergence overprinting earlier flow patterns. Reproduced from Stokes CR and Clark CD (2003) The Dubawnt Lake palaeo-ice stream: Evidence for dynamic ice sheet behaviour on the Canadian Shield and insights regarding controls on ice stream location and vigour. *Boreas* 32: 263–279, with permission.

glaciers occupy the uplands of the humid north, debris-covered cirque glaciers with moraine dams occur in central catchments, and active talus rock glaciers dominate toward the arid south. **Figure 7** shows how process–form domains have shifted in space and time in response to a rise in ELAs since the Last Glacial Maximum, highlighting a shift along the spectrum from coupled to uncoupled landsystems due to reductions in ice-covered area and the importance of glacial transport and runoff.

Plateau Ice Field Landsystem

The plateau ice field landsystem concept has been applied by [Rea et al. \(1998\)](#) and [McDougall \(2001\)](#) to the reconstruction of Younger Dryas glacier cover in the English Lake District. Initial reconstruction by [Sissons \(1980\)](#) featured extended valley head glaciers that lay below large upland massifs, thereby necessitating the proposal for three separate snowfall intensity zones, in combination with the effects of snowblow and aspect, to account for considerable variations in glacier extent and ELAs (the latter varying by 500 m). Recent mapping of the glacial landforms has revealed that, similar to modern glacierized plateau terrains, the English Lake District valley heads contain substantial volumes of glacial sediment in latero-frontal and linear hummocky moraines that extend in many cases onto plateau margins. Evidence of plateau ice is patchy and often restricted to scoured bedrock at plateau edges. Sissons' 'valley' glaciers were therefore actually outlet lobes of a plateau ice field (**Figure 8**). [McDougall's \(2001\)](#) paleoglacier reconstructions for the English Lake District, incorporating plateau ice fields, suggest that ELAs during the Younger Dryas were up to 90 m higher than those calculated by Sissons based on an Alpine style of glaciations with no plateau ice.

Paleo-Ice Stream Landsystem

Evidence for a terrestrially terminating, rubber-stamped, paleo-ice stream has been identified by [Stokes and Clark \(2003\)](#) in the Dubawnt Lake area of the Northwest Territories of Canada (**Figure 9**). The ice stream formerly drained the northwestern portion of the Laurentide Ice Sheet and has been recognized by subglacial bedforms that conform to the patterns expected in zones of converging trunk and terminal diverging flow in a terrestrial ice stream. Additionally, the landsystem includes abrupt lateral margins and MSGs from which high ice flow velocity is inferred. The clarity of the ice stream bedform signature, with no discordances or crosscuts, qualified it as a rubber-stamped record of ice stream shutdown.

Conclusion

Paleoglaciological reconstructions, at scales ranging from local to continental, play a significant part in the development of models of global paleoclimate. However, the great spatial and temporal complexity of glacial processes and forms have necessitated the compilation of process–form models or landsystems that relate to various glaciation styles and systems, which may then be applied to specific physiographic and climatic settings. A number of contemporary glacierized landscapes have been employed as modern analogs for the

landform–sediment assemblages produced by different glaciation styles and ice dynamics in differing climatic, geologic, and topographic settings. The identification of modern landsystems (e.g., those pertaining to active temperate glacier snouts, glaciated valleys, plateau ice fields, and ice streams) provides glacial geomorphologists with the criteria to decipher paleoglacier dynamics and former glacier–climate relationships in ancient glaciated terrain.

See also: [Glacial Landforms: Evidence of Glacier Recession.](#)
Glacial Landforms, Erosional Features: Micro- to Macroscale Forms. **Glacial Landforms, Ice Sheets:** Evidence of Glacier and Ice Sheet Extent. **Glacial Landforms, Sediments:** Glaciofluvial Landforms of Deposition; Glaciomarine Sediments and Ice-Rafted Debris; Till.

References

- Alley RB, Blankenship DD, Rooney ST, and Bentley CR (1989) Sedimentation beneath ice shelves – The view from ice stream B. *Marine Geology* 85: 101–120.
- Benn DI (1989) Debris transport by Loch Lomond Readvance glaciers in northern Scotland, basin form and the within-valley asymmetry of lateral moraines. *Journal of Quaternary Science* 4: 243–254.
- Benn DI (1995) Fabric signature of subglacial till deformation, Breiðamerkurjökull, Iceland. *Sedimentology* 42: 735–747.
- Benn DI and Evans DJA (1996) The interpretation and classification of subglacially-deformed materials. *Quaternary Science Reviews* 15: 23–52.
- Benn DI and Evans DJA (2006) Subglacial floods and the limitations of outrageous hypotheses. In: Knight PG (ed.) *Glacier Science and Environmental Change*. London: Blackwell.
- Benn DI and Evans DJA (2010) *Glaciers and Glaciation*, 2nd edn. London: Arnold.
- Benn DI, Kirkbride MP, Owen LA, and Brazier V (2003) Glaciated valley landsystems. In: Evans DJA (ed.) *Glacial Landsystems*, pp. 372–406. London: Arnold.
- Benn DI and Lehmkuhl F (2000) Mass balance and equilibrium-line altitudes of glaciers in high mountain environments. *Quaternary International* 65–66: 15–29.
- Boulton GS (1986) Push-moraines and glacier-contact fans in marine and terrestrial environments. *Sedimentology* 33: 677–698.
- Boulton GS (1987) A theory of drumlin formation by subglacial deformation. In: Menzies J and Rose J (eds.) *Drumlin Symposium*, pp. 25–80. Rotterdam: Balkema.
- Boulton GS and Eyles N (1979) Sedimentation by valley glaciers: A model and genetic classification. In: Schluchter C (ed.) *Moraines and Varves*, pp. 11–23. Rotterdam: Balkema.
- Boulton GS and Hindmarsh RCA (1987) Sediment deformation beneath glaciers: Rheology and sedimentological consequences. *Journal of Geophysical Research* 92: 9059–9082.
- Brazier V, Kirkbride MP, and Owens IF (1998) The relationship between climate and rock glacier distribution in the Ben Ohau Range, New Zealand. *Geografiska Annaler* 80A: 193–207.
- Canals M, Urgeles R, and Calafat AM (2000) Deep sea-floor evidence of past ice streams off the Antarctic Peninsula. *Geology* 28: 31–34.
- Cenderelli DA and Wohl EE (2003) Flow hydraulics and geomorphic effects of glacial lake outburst floods in the Mount Everest region, Nepal. *Earth Surface Processes and Landforms* 28: 385–407.
- Clague JJ and Evans SG (2000) A review of catastrophic drainage of moraine-dammed lakes in British Columbia. *Quaternary Science Reviews* 19: 1763–1783.
- Clark CD (1993) Mega-scale glacial lineations and cross-cutting ice-flow landforms. *Earth Surface Processes and Landforms* 18: 1–29.
- Clark CD (1999) Glaciodynamic context of subglacial bedform generation and preservation. *Annals of Glaciology* 28: 23–32.
- Clark CD, Evans DJA, and Piotrowski JA (2003) Palaeo-ice streams (Special issue). *Boreas* 32: 1–280.
- Clark CD and Stokes CR (2003) Palaeo-ice stream landsystem. In: Evans DJA (ed.) *Glacial Landsystems*, pp. 204–227. London: Arnold.

- Clayton L and Moran SR (1974) A glacial process-form model. In: Coates DR (ed.) *Glacial Geomorphology*, pp. 89–119. Binghamton, NY: State University of New York.
- Colgan PM, Mickelson DM, and Cutler PM (2003) Ice-marginal terrestrial landystems: Southern Laurentide Ice Sheet margin. In: Evans DJA (ed.) *Glacial Landsystems*, pp. 111–142. London: Arnold.
- Dyke AS (1993) Landscapes of cold-centred Late Wisconsinan ice caps, Arctic Canada. *Progress in Physical Geography* 17: 223–247.
- Dyke AS and Morris TF (1988) Drumlin fields, dispersal trains and ice streams in Arctic Canada. *The Canadian Geographer* 32: 86–90.
- Dyke AS, Morris TF, Green DEC, and England J (1992) Quaternary geology of Prince of Wales Island, Arctic Canada. *Geological Survey of Canada Memoir* 433.
- Dyke AS and Savelle JM (2000) Major end moraines of Younger Dryas age on Wollaston Peninsula, Victoria Island, Canadian Arctic: Implications for paleoclimate and for formation of hummocky moraine. *Canadian Journal of Earth Sciences* 37: 601–619.
- Evans DJA (ed.) (2003a) *Glacial Landsystems*. London: Arnold.
- Evans DJA (2003b) Ice-marginal terrestrial landystems: Active temperate glacier margins. *Glacial Landsystems* 12–43.
- Evans DJA (2005) The glacier-marginal landystems of Iceland. In: Caseldine C, Russell AJ, Hardardottir J, and Knudsen O (eds.) *Iceland - Modern Processes and Past Environments*, pp. 259–288. Amsterdam: Elsevier.
- Evans DJA (2009) Controlled moraines: Origins, characteristics and palaeoglaciological implications. *Quaternary Science Reviews* 28: 183–208.
- Evans DJA (2010) Controlled moraine development and debris transport pathways in polythermal plateau icefields: Examples from Tungnafellsjökull, Iceland. *Earth Surface Processes and Landforms* 35: 1430–1444.
- Evans DJA (2011) Glacial landystems of Satujökull, Iceland: A modern analogue for glacial landystem overprinting by mountain icecaps. *Geomorphology* 129: 225–237.
- Evans DJA, Clark CD, and Rea BR (2008) Landform and sediment imprints of fast glacier flow in the southwest Laurentide Ice Sheet. *Journal of Quaternary Science* 23: 249–272.
- Evans DJA and Hiemstra JF (2005) Till deposition by glacier submarginal, incremental thickening. *Earth Surface Processes and Landforms* 30: 1633–1662.
- Evans DJA, Lemmen DS, and Rea BR (1999) Glacial landystems of the southwest Laurentide Ice Sheet: Modern Icelandic analogues. *Journal of Quaternary Science* 14: 673–690.
- Evans DJA and Rea BR (1999) Geomorphology and sedimentology of surging glaciers: A landystems approach. *Annals of Glaciology* 28: 75–82.
- Evans DJA and Rea BR (2003) Surging glacier landystem. In: Evans DJA (ed.) *Glacial Landystems*, pp. 259–288. London: Arnold.
- Evans DJA, Rea BR, Hiemstra JF, and Ó Cofaigh C (2006a) A critical assessment of subglacial mega-floods: A case study of glacial sediments and landforms in south-central Alberta, Canada. *Quaternary Science Reviews* 25: 1638–1667.
- Evans DJA and Twigg DR (2000) *Breiðamerkurjökull 1998 (1:30,000 scale map)*. Glasgow/Leicestershire: University of Glasgow/Loughborough University.
- Evans DJA and Twigg DR (2002) The active temperate glacial landystem: A model based on Breiðamerkurjökull and Fjallsjökull, Iceland. *Quaternary Science Reviews* 21: 2143–2177.
- Evans DJA, Twigg DR, and Orton C (2010) Satujökull glacial landystem, Iceland. *Journal of Maps* 2010: 639–650.
- Evans DJA, Twigg DR, and Shand M (2006b) Surficial geology and geomorphology of the Þorissjökull plateau icefield, west-central Iceland. *Journal of Maps* 2006: 17–29.
- Eyles N (1983a) Glacial geology: A landystems approach. In: Eyles N (ed.) *Glacial Geology*, pp. 1–18. Oxford: Pergamon.
- Eyles N (1983b) The glaciated valley landystem. In: Eyles N (ed.) *Glacial Geology*, pp. 91–110. Oxford: Pergamon.
- Eyles N and Menzies J (1983) The subglacial landystem. In: Eyles N (ed.) *Glacial Geology*, pp. 19–70. Oxford: Pergamon.
- Fay H (2002) Formation of ice block obstacle marks during the November 1996 glacier-outburst flood (jökulhlaup), Skeiðararsandur, southern Iceland. In: Martini IP, Baker VR, and Garzon G (eds.) *Flood and Megaflood Deposits: Recent and Ancient*, pp. 85–97. International Association of Sedimentologists. Special Publication No. 32.
- Fitzsimons SJ (2003) Ice-marginal terrestrial landystems: Polar continental glacier margins. In: Evans DJA (ed.) *Glacial Landystems*, pp. 89–110. London: Arnold.
- Fookes PG, Gordon DL, and Higginbottom IE (1978) Glacial landforms, their deposits and engineering characteristics. In: *The Engineering Behaviour of Glacial Materials*, Proceedings of Symposium, Geoabstracts, pp. 18–51. Birmingham: University of Birmingham.
- Gellatly AF, Whalley WB, and Gordon JE (1986) Topographic control over recent glacier changes in Southern Lyngen Peninsula, North Norway. *Norsk Geografisk Tidsskrift* 41: 211–218.
- Gustavson TC and Boothroyd TC (1987) A depositional model for outwash, sediment sources, and hydrologic characteristics, Malaspina Glacier, Alaska: A modern analog of the southeastern margin of the Laurentide ice sheet. *Geological Society of America Bulletin* 99: 187–200.
- Hart JK (1995) Recent drumlins, flutes and lineations at Vestari-Hagafellsjökull, Iceland. *Journal of Glaciology* 41: 596–606.
- Ives JD, Andrews JT, and Barry RG (1975) Growth and decay of the Laurentide Ice Sheet and comparisons with Fenno-Scandinavia. *Naturwissenschaften* 62: 118–125.
- Kamb B (2001) Basal zone of the West Antarctic ice streams and its role in lubrication of their rapid motion. In: Alley RB and Bindshadler RA (eds.) *The West Antarctic Ice Sheet: Behavior and Environment Antarctic Research Series* 77, pp. 157–199. Washington, DC: American Geophysical Union.
- Kershaw JA, Clague JJ, and Evans SG (2005) Geomorphic and sedimentological signature of a two-phase outburst flood from moraine-dammed Queen Bess Lake, British Columbia, Canada. *Earth Surface Processes and Landforms* 30: 1–25.
- Kirkbride MP (2000) Ice marginal geomorphology and Holocene expansion of debris-covered Tasman Glacier, New Zealand. In: Nakawo M, Raymond C, and Fountain A (eds.) *Debris-Covered Glaciers*, pp. 211–217. Wallingford, CT: International Association of Hydrological Sciences.
- Kjær KH and Krüger J (2001) The final phase of dead-ice moraine development: Processes and sediment architecture, Kötlujökull, Iceland. *Sedimentology* 48: 935–952.
- Krüger J (1985) Formation of a push moraine at the margin of Hofdabrekkujökull, south Iceland. *Geografiska Annaler* 67A: 199–212.
- Krüger J (1993) Moraine ridge formation along a stationary ice front in Iceland. *Boreas* 22: 101–109.
- Krüger J (1994) Glacial processes, sediments, landforms and stratigraphy in the terminus region of Myrdalsjökull, Iceland. *Folia Geographica Danica* 21: 1–233.
- Krüger J (1995) Origin, chronology and climatological significance of annual-moraine ridges at Myrdalsjökull, Iceland. *The Holocene* 5: 420–427.
- Krüger J (1996) Moraine ridges formed from subglacial frozen on sediment slabs and their differentiation from push moraines. *Boreas* 25: 57–63.
- Maag H (1969) *Ice-Dammed Lakes and Marginal Glacial Drainage on Axel Heiberg Island, Axel Heiberg Island Research Report*. Montreal: McGill University.
- Maizels J (1997) Jökulhlaup deposits in proglacial areas. *Quaternary Science Reviews* 16: 793–819.
- Manley G (1955) On the occurrence of ice-domes and permanently snow-covered summits. *Journal of Glaciology* 2: 453–456.
- McDougall DA (2001) The geomorphological impact of Loch Lomond (Younger Dryas) Stadial plateau icefields in the central Lake District, northwest England. *Journal of Quaternary Science* 16: 531–543.
- Ó Cofaigh C, Taylor J, Dowdeswell JA, and Pudsey CJ (2003) Palaeo-ice streams, trough mouth fans and high-latitude continental slope sedimentation. *Boreas* 32: 37–55.
- Owen LA and Derbyshire E (1989) The Karakoram glacial depositional system. *Zeitschrift für Geomorphologie* 76: 33–73.
- Paul MA (1983) The supraglacial landystem. In: Eyles N (ed.) *Glacial Geology*, pp. 71–90. Oxford: Pergamon.
- Powell RD and Domack EW (1995) Glaciomarine environments. In: Menzies J (ed.) *Glacial Environments – Processes, Sediments and Landforms*, pp. 445–486. Boston, MA: Butterworth-Heinemann.
- Price RJ (1969) Moraines, sandar, kames and eskers near Breiðamerkurjökull, Iceland. *Transactions of the Institute of British Geographers* 46: 17–43.
- Price RJ (1970) Moraines at Fjallsjökull, Iceland. *Arctic and Alpine Research* 2: 27–42.
- Rea BR and Evans DJA (2003) Plateau icefield landystems. In: Evans DJA (ed.) *Glacial Landystems*, pp. 407–431. London: Arnold.
- Rea BR, Whalley WB, Evans DJA, Gordon JE, and McDougall DA (1998) Plateau icefields: Geomorphology and dynamics. *Quaternary Proceedings* 6: 35–54.
- Russell AJ and Knudsen O (1999) Controls on the sedimentology of the November 1996 jökulhlaup deposits, Skeiðararsandur, Iceland. In: Smith ND, Rogers J, and Plint AG (eds.) *Advances in Fluvial Sedimentology*, pp. 315–329. International Association of Sedimentologists. Special Publication No. 28.
- Sharp MJ (1984) Annual moraine ridges at Skálafellsjökull, south-east Iceland. *Journal of Glaciology* 30: 82–93.
- Sissons JB (1980) The Loch Lomond Advance in the Lake District, northern England. *Transactions of the Royal Society of Edinburgh: Earth Sciences* 71: 13–27.

- Speight JG (1963) Late Pleistocene historical geomorphology of the Lake Pukaki area, New Zealand. *New Zealand Journal of Geology and Geophysics* 6: 160–188.
- Stokes CR and Clark CD (1999) Geomorphological criteria for identifying Pleistocene ice streams. *Annals of Glaciology* 28: 67–74.
- Stokes CR and Clark CD (2001) Palaeo-ice streams. *Quaternary Science Reviews* 20: 1437–1457.
- Stokes CR and Clark CD (2002) Ice stream shear margin moraines. *Earth Surface Processes and Landforms* 27: 547–558.
- Stokes CR and Clark CD (2003) The Dubawnt Lake palaeo-ice stream: Evidence for dynamic ice sheet behaviour on the Canadian Shield and insights regarding controls on ice stream location and vigour. *Boreas* 32: 263–279.
- Vorren T and Laberg JS (1997) Trough mouth fans – Palaeoclimate and ice sheet monitors. *Quaternary Science Reviews* 16: 865–881.
- Wilson SB and Evans DJA (2001) Scottish landform example – 24: Coire a' Cheudchnoic, the 'hummocky moraine' of Glen Torridon. *Scottish Geographical Journal* 116: 149–158.

Glaciofluvial Landforms of Erosion

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Introduction

The vast amount of meltwater released from glaciers has enormous erosional capacity. Temporary impoundments of meltwater occur on top of, within, at the base of, and in front of the ice. The magnitude and intensity of erosion is greatest when these impoundments release sediment-deficient meltwater suddenly or rapidly in the form of outbursts. Erosion is facilitated by high hydraulic potential caused by the cryostatic pressure of the ice or the hydraulic head developed behind a barrier in a proglacial lake. The type and resistance of bed material underlying a region also controls the form and size of erosional landforms produced by these meltwater discharges.

Most glaciofluvial landforms of erosion were formed when the great Pleistocene ice sheets of North America and Eurasia produced and ponded vast quantities of meltwater. The erosional landforms include vast channel systems, both subglacial and proglacial, and a variety of forms attributed to subglacial sheetfloods. In some cases, subglacial flows switched from an erosional to a depositional regime upon reaching the glacier terminus, typified by aggrading braided streams, sandar, or subaqueous fans of various types.

The discharges that formed large-scale glaciofluvial erosional forms had significant global impacts because they were of such high magnitude. Many large-scale drainage systems in areas affected by glaciation were formed or largely modified by discrete event, high-discharge meltwater flows. In places, these channels constitute important aquifer systems, some of which are buried beneath multiple sheets of glacial drift from earlier glaciations. In addition, numerous studies have argued that the sudden release of vast quantities of cold meltwater correlates with periods of climatic change, possibly by altering oceanic circulation patterns.

Four categories of processes and landforms are considered in this article because they account for the majority of large-scale glaciofluvial erosional forms: (1) proglacial lake channel systems formed by glacial lake outbursts, (2) tunnel valleys formed by channelized subglacial flows, (3) ice-marginal (lateral) meltwater channels associated primarily with cold-based glaciers, and (4) a variety of erosional forms hypothesized to have been formed by subglacial sheet flows. This review of erosional landforms and related processes includes ideas that are universally accepted, such as a catastrophic flood origin of the Channeled Scabland in the United States, and a discussion of subglacial landforms attributed to fluvial erosion, which are actively debated and highly controversial.

Glacial Lake Outbursts

Overview

Glacial lake outbursts, or jökulhlaups, are very high-magnitude floods caused by the sudden drainage of water from a glacial

lake. Glacial lakes form on, in, under, or in front of glaciers and are inherently unstable. Outbursts occur in many present-day glaciated regions, such as Iceland, Alaska, the Himalayas, and the Andes ([Figure 1](#)). Flood discharges typically exceed background rates by orders of magnitude and, during the Pleistocene, carved out spectacular landscapes by flows of unparalleled magnitude. In fact, glacial lake outbursts account for the top seven floods ever known to have occurred in the world ([O'Connor and Costa, 2004](#)). Floodwaters from Pleistocene glacial lake outbursts were highly erosive and created a characteristic suite of landforms, now widely accepted to have formed by catastrophic floods. The geomorphic record of outbursts is most significant from Pleistocene glaciations because of the high magnitude of flows, in some cases exceeding $10 \times 10^6 \text{ m}^3 \text{ s}^{-1}$ ([Baker et al., 1993](#); [Burr et al., 2009](#)). The emphasis here, therefore, is on Pleistocene glacial lake outburst erosional landforms, although valuable insights into outburst processes have been gained by observations of modern jökulhlaups ([Cenderelli and Wohl, 2003](#)).

Recognition of Pleistocene Glacial Lake Outbursts

For more than a century, the incongruity in size between underfit present-day streams and overfit valleys has been apparent. Furthermore, numerous studies suggested that the overfit valleys were paleochannels – not valleys, as evidenced by uniformity in size over long distances and regular meander wavelengths – that conveyed much higher discharges than present during the Pleistocene. Many of these same channels are now widely recognized to have formed by glacial lake outbursts. This acceptance was not the case when [Bretz \(1923\)](#) first proposed a catastrophic flood origin for the Channeled Scabland: an anastomosing network of channels, covering an area about 40000 km², in the Columbia Plateau of northwestern United States. Bretz, based on years of fieldwork, was compelled to argue for a catastrophic flood, the Spokane Flood, to explain the eroded basalt and loess landscape containing a network of proglacial channels, bedrock cataracts, streamlined islands, dry valleys, and boulder deposits ([Figure 2](#)). His hypothesis was widely rejected for some scientific reasons, including the lack of a known water source for the flood, but more importantly because the idea of a catastrophic origin to a landscape was viewed as an assault on the principle of uniformitarianism – the foundation of modern geology. In the 1960s, Bretz's outrageous hypothesis became widely accepted, now well supported by, among other things, a water source (Glacial Lake Missoula), giant current ripples, and high-altitude imagery ([Figure 3](#)). This fascinating story in the evolution of scientific ideas is well described (e.g., [Baker, 2009](#); [Baker et al., 1987](#)).

Today, dozens of investigations have shown that glacial lake outbursts were not unique to the Channeled Scabland; they were a common, if not inevitable, process during

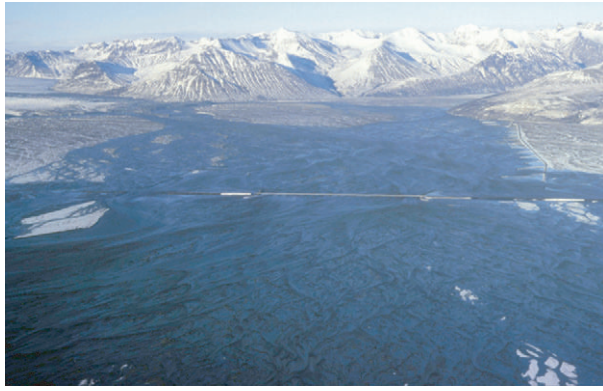


Figure 1 Upstream view of jökulhlaup on Skeiðará river, Iceland. This flood occurred in 1996 and was triggered by a subglacial volcanic eruption. Photograph by Magnus T. Gudmundsson.



Figure 2 Waterfalls (about 60 m high) in cataract and gorge eroded by the Spokane Floods of the Channeled Scabland. Eroded basalt and loess landscape is in the background. Photograph by Mark Lord.

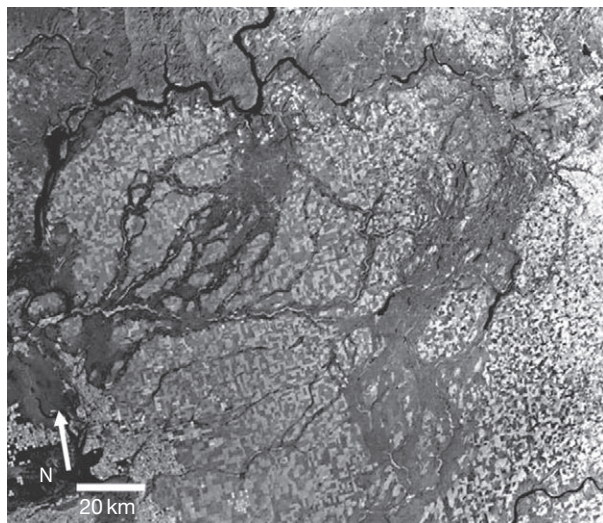


Figure 3 Landsat image of Channel Scabland, Washington State, USA. Flood flows were from the northeast toward the southwest.

Pleistocene glaciations as the land, ice sheets, water sources, and drainage pathways were in a continuous state of flux. Along the margins of the Laurentide Ice Sheet, for example, three dozen outbursts have been documented (Figure 4; Kehew et al., 2009). Channel systems, mostly glacial lake spillways, have been attributed to glacial lake outbursts in Siberia, Iceland, and Scandinavia. Recent investigations of the Antarctic Ice Sheet show that vast amounts of subglacial meltwater exist. Some of this meltwater is highly dynamic and likely formed scabland-like topography beneath the ice sheet in the past (Denton and Sugden, 2005). Characteristic megaflood features also occur on the Martian landscape (Burr et al., 2009).

Glacial Lake Outburst Erosional Landforms

The Channeled Scabland is clearly the most studied area formed by outburst floods. Many landforms now used as indicators of catastrophic floods were first described in this region (Baker, 1973). Although the outburst flood origin of landforms of the Scablands is widely accepted, the number and timing of the outbursts, the magnitude and frequency relationships of the floods, the erosional capability of the outbursts, and the source of flood waters are actively studied (Baker, 2009; Benn and Evans, 2010). Spillway systems that formed by outbursts along the southern margins of the Laurentide Ice sheet, though not as dramatic as the Scabland landscapes, share many of the erosional landforms and have been well documented (Kehew and Lord, 1987). The most ubiquitous erosional landform created by outburst floods is a trench-shape channel with steep sides and a relatively uniform width and depth (Figure 5). The size of the channel varies with flow magnitude, geologic setting, and flow history among other things. Spillways that formed by outbursts in the Great Plains of North America are typically 1–3 km wide and 25–100 m deep (Kehew and Lord, 1987). Common erosional landform types produced by outbursts are summarized in Table 1.

General models for the sequence of development of flood-eroded landscapes have been developed on the basis of studies of the Channeled Scabland, outburst spillways in the Great Plains, and experimental flume studies (respectively, Baker et al., 1987; Carling et al., 2009; Kehew and Lord, 1987; Kehew et al., 2009; Shepard and Schumm, 1974; Wohl and Ikeda, 1997). Initial stages of flooding overwhelm the landscape and cause broad scouring, shallow anastomosing channels, and, in some cases, transverse erosional bedforms. With increased duration, the flow becomes more organized as the bed deforms, which is reflected in longitudinal grooves and streamlined hills. Continued flow leads to selective deepening to form inner channels, which progressively enlarge by capture of more flow laterally and by upstream knickpoint migration. A spillway formed by a long duration or repeated outbursts floods can convey all flow in one channel and may contain only a few very well streamlined hills (Figure 6).

Hydrologic Characteristics of Pleistocene Glacial Lake Outbursts

The cataclysmic failure of ice dams impounding large glacial lakes released huge volumes of water, over 2000 km³ in the Channeled Scabland, in a matter of days to weeks (O'Connor

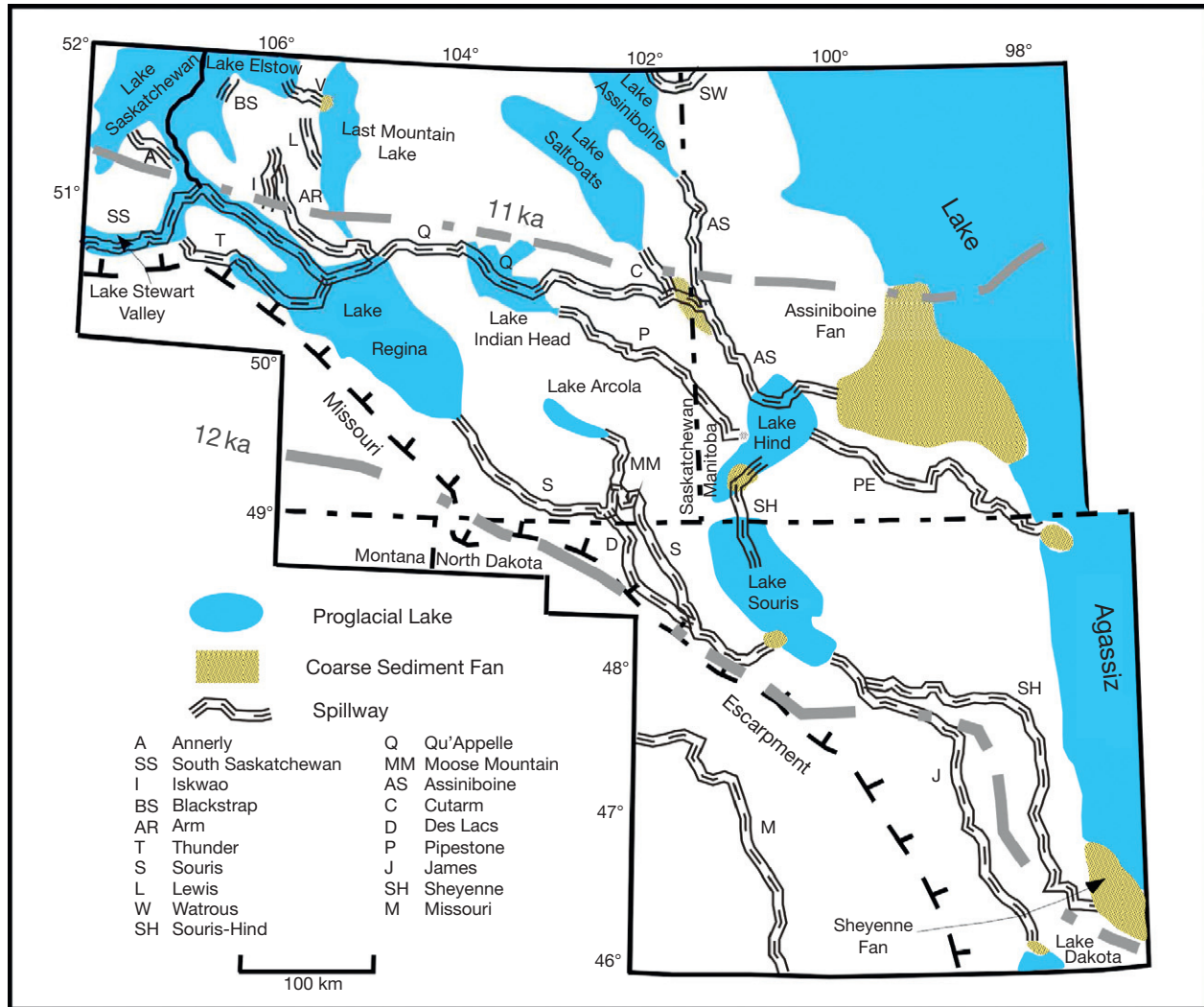


Figure 4 Proglacial lakes and spillways along the southwestern margins of the Laurentide Ice Sheet. Reproduced from Kehew AE and Teller JT (1994) History of Late Glacial runoff along the southwestern margin of the Laurentide Ice Sheet. *Quaternary Science Reviews* 13: 861.



Figure 5 Oblique aerial photograph of inner channel and shallow, anastomosing channels on scoured uplands, Souris Spillway, southeastern Saskatchewan. Flow was to the east (top to bottom). Photography by Mark Lord.

and Baker, 1992). Outburst flows were highly erosive and, at least in much of the Great Plains spillway systems, became hyperconcentrated as they ripped through generally weak, fine-grained materials that underlie the region (Lord and Kehew, 1987). The sedimentary record of outburst floods is scarce in spillways, is present in some glacial lakes that received floodwaters (e.g., Lord, 1991), but is mostly present in the marine offshore record (e.g., Normark and Reid, 2003).

The largest well-documented outburst flood occurred in south-central Siberia in the Altay Mountains, which carved out gorges and produced megaripples in huge gravel bars analogous to those in the Channeled Scabland (Baker et al., 1993). Peak discharge is estimated to have been $18 \times 10^6 \text{ m}^3 \text{ s}^{-1}$ with flows up to 500 m deep and velocities up to 45 m s^{-1} ; the unit stream power generated by this flood is estimated to have been 10^6 W m^{-2} . Paleohydrologic studies of lower magnitude outbursts typically estimate discharges to be between 10^4 and $10^6 \text{ m}^3 \text{ s}^{-1}$ (see Benn and Evans, 2010, for summary). The

Table 1 Erosional landforms produced by glacial lake outbursts

Landforms	Characteristics
Trench-shape channels (Figure 5)	Commonly parallel to ice margins and connect or bisect glacial lake beds
Anastomosing channel pattern	Network of channels may be from 1 to 100 km wide
Streamlined erosional residuals (islands) (Figure 6)	Erosional residuals, given long enough flow, will assume a lemniscate form that minimizes fluid drag
Dry valleys (channels)	Variable sizes, commonly on upland areas above deeper inner channels
Perched (hanging) channels	Variable sizes
Longitudinal grooves (Figure 5)	Flow parallel grooves interpreted to be formed by secondary circulation cells; up to 3 m high in Scablands and up to 2 m high in Great Plains
Cataracts (Figure 2)	Dry Falls in the Channeled Scabland is 5.5 km wide and 120 m high, probably formed by the coalescence of potholes eroded by vertical vortices (Baker et al., 1987)
Potholes and obstacle scour depressions	Occur at scales of meters to hundreds of meters
Transverse erosional ripples	Scarce landform present in well-preserved parts of Great Plains outburst flow paths (up to 1.2 m high with about 40 m wavelengths)
Erosional scarps	Commonly cut into terraces or valley sides

Compiled from **Baker (1973)**, **Baker et al. (1987)**, and **Kehew and Lord (1987)**.



Figure 6 Oblique aerial photograph of streamlined hill in the Souris-Hind spillway, southern Manitoba; flow was to the north (upper right to lower left). Photograph by Mark Lord.

specific characteristics of the erosional landscapes carved by these tremendous flows varied not just with discharge, but also with flow mechanics (e.g., flow regime, cavitation, secondary circulation), flow boundary materials, sediment availability, flow duration, flood hydrograph, and flow history (**Baker, 1973**; **Cenderelli and Wohl, 2003**; **Kehew and Lord, 1987**; **O'Connor and Baker, 1992**). For example, unit stream power has been used successfully to relate hydraulic conditions to different types of erosional processes and landforms.

Tunnel Valleys

Overview

Tunnel valleys, or tunnel channels (also known as tunneldale or rinnentaler), are large meltwater valleys eroded into sediment or rock at the base of an ice sheet. Valley dimensions reach >30 m deep, 4 km wide and more than 100 km long. Tunnel valleys have a wide variety of morphologies and associated landforms, and occur in drainage systems covering extensive areas.

The widespread distribution of tunnel valleys in the area of the Laurentide and Scandinavian Ice Sheets has important implications for ice-sheet dynamics and glacial meltwater regimes (**Brennand and Shaw, 1994**; **Clayton et al., 1999**; **Jørgensen and Sandersen, 2006**; **Mooers, 1989**; **Wright, 1973**). Although literature on tunnel valleys is abundant, controversy exists surrounding the mechanism of tunnel-valley genesis and the implications of meltwater volumes associated with the sizes of some of the valleys. As yet no unified theory on their formation can adequately explain the observed diverse morphologies, landform associations, and substrate lithologies (**O'Cofaigh, 1996**).

The terms 'tunnel valley' and 'tunnel channel' have often been used interchangeably to denote subglacially eroded drainage courses. However, these terms may also have genetic implications (**Clayton et al., 1999**; **Mooers, 1989**). For example, **Clayton et al. (1999)** use the term 'tunnel channel' to refer to erosion during a single or small number of events in which the flow depths and widths are comparable to the size of the valley. The magnitude of discharge inherent in this type of event is usually implied to be short-lived and catastrophic in nature. In this discussion, the term tunnel valley is used to describe landforms eroded by subglacial meltwater without a specific genetic origin.

Morphology

Despite a general orientation subparallel to longitudinal flow lines of the glacier, tunnel-valley drainage patterns are highly variable, including isolated straight valleys, braided and dendritic networks, radial, subparallel valleys (**Figure 7**) and anastomosing and anabranching systems (**Figure 8**) (**Brennand and Shaw, 1994**; **Clayton et al., 1999**; **Mooers, 1989**; **Sjogren et al., 2002**; **Wright, 1973**).

Valley morphologies are equally diverse and include straight-walled channels of uniform width (**Figure 9**), crenulated valleys that expand and contract, and subtle valleys only recognizable as linear chains of lakes or depressions. Buried tunnel valleys with no surface expression occur in the North Sea, Denmark, and the southern margins of the Scandinavian Ice Sheet (**Huuse and Lykke-Andersen, 2000**; **Jørgensen and Sandersen, 2006**). Many tunnel valleys display steep walls similar to outburst spillways, but others display irregular widths and depths and variable degrees of posterosional infill. Longitudinal profiles of flow paths may be undulating and have convex-up or reverse gradients that flow uphill, cross drainage divides, and trend oblique to the regional slope (**Figure 10**) (**Clayton et al., 1999**; **Kozłowski et al., 2005**; **Wright, 1973**). In other areas, tunnel valleys occur as isolated,

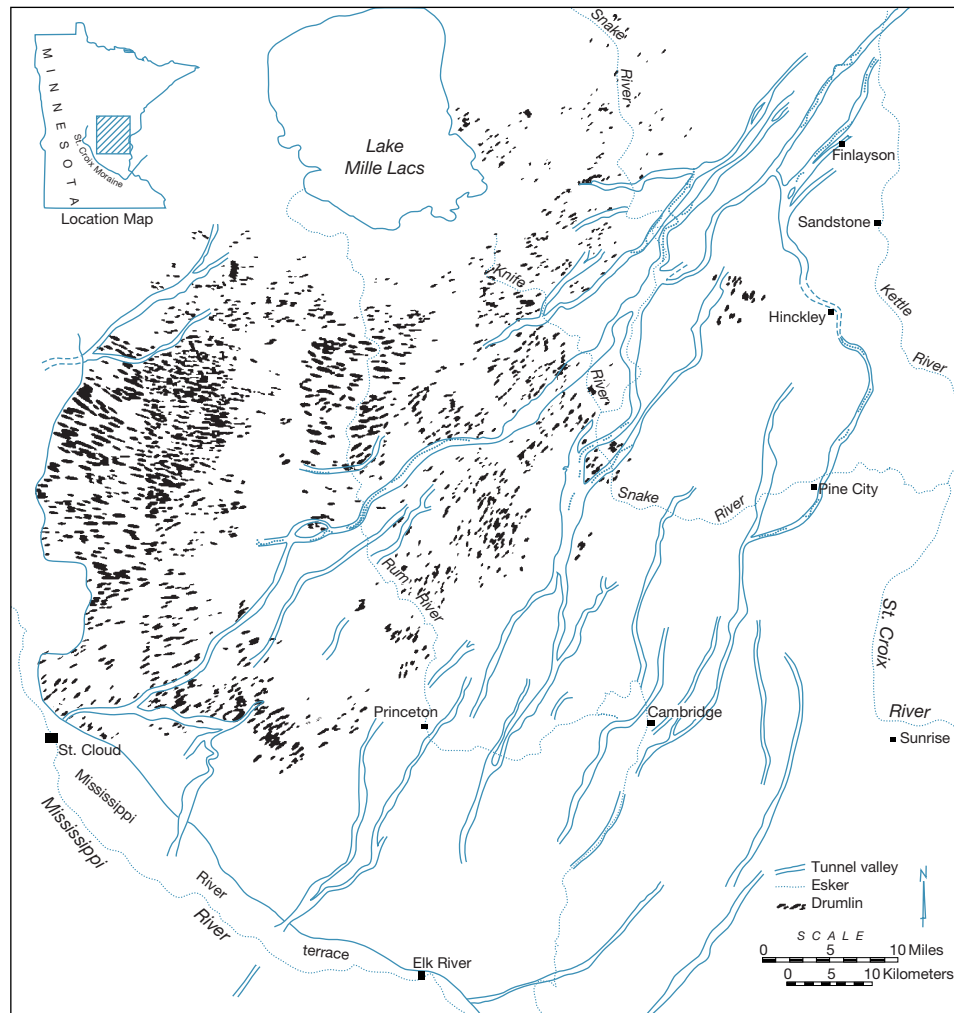


Figure 7 Subparallel tunnel channels and associated landforms of the Superior Lobe, Minnesota, USA. Reproduced from [Wright HE Jr. \(1973\)](#) Tunnel valleys, glacial surges, and subglacial hydrology of the Superior Lobe, Minnesota. In: Black R, Goldthwait R, and Willman H (eds.) *The Wisconsin Stage, Geological Society of America Memoir*, vol. 136, pp. 251–276. Boulder, CO: Geological Society of America.

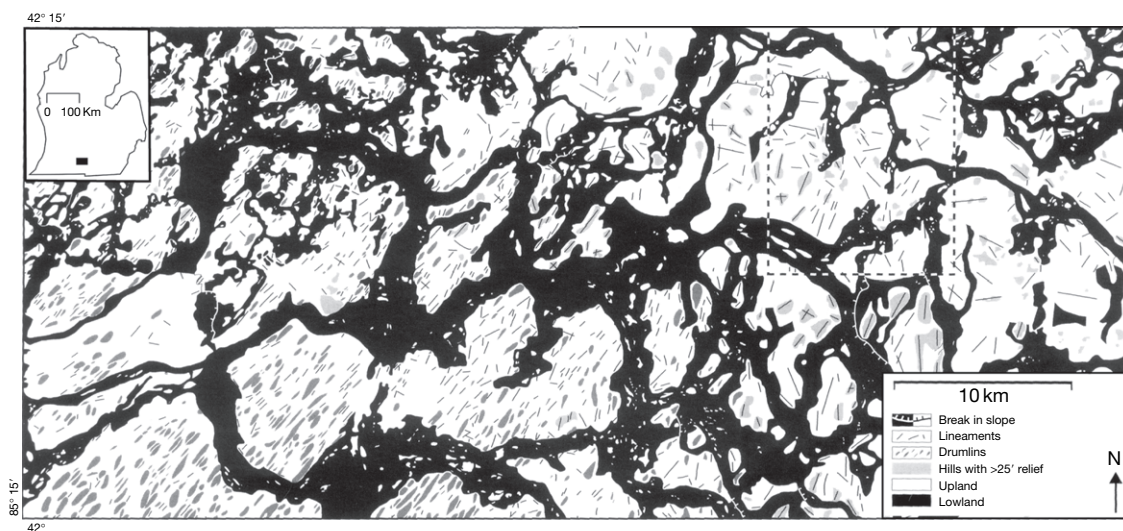


Figure 8 Anabranching system of tunnel channels and associated landforms of the Saginaw Lobe, south-central Michigan, USA. Modified from Sjogren DB, Fisher TG, Taylor LD, Jol HM, and Munro-Stasiuk MJ (2002) Incipient tunnel channels. *Quaternary International* 90: 41–56.

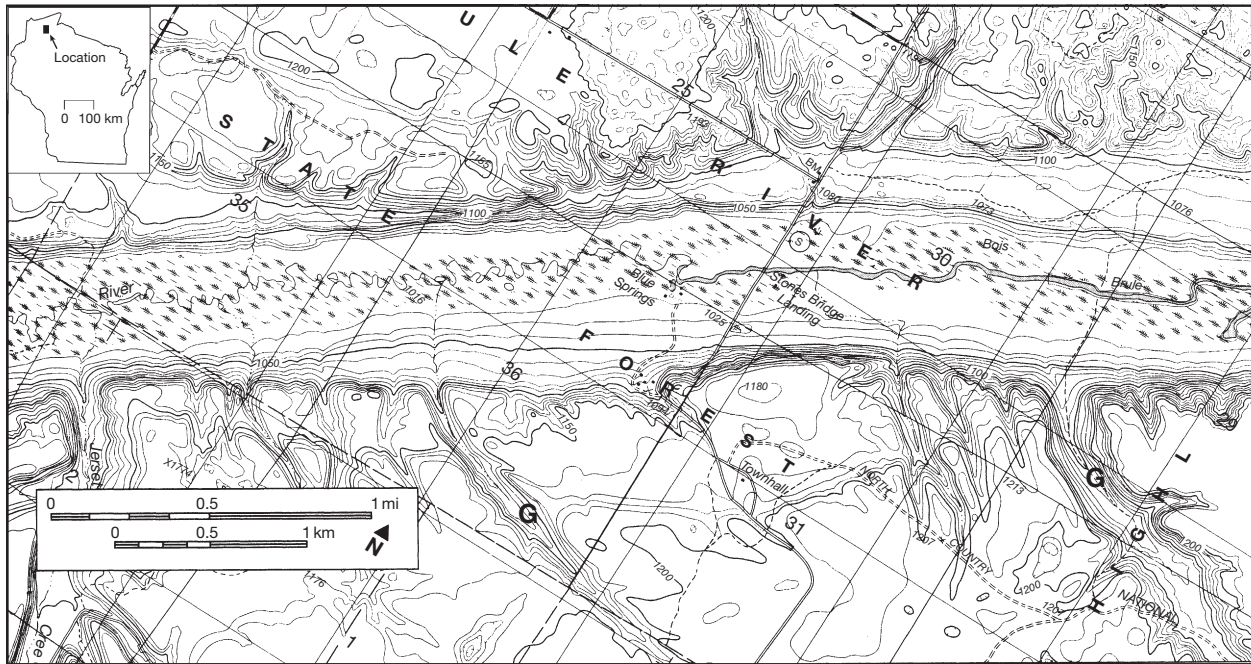


Figure 9 Straight-walled tunnel channel of uniform width in northwestern Wisconsin with postformational gullies (G). Channel is located in the Lake Minnesuing 7.5 min. quadrangle, contour interval 10 ft (approx. 3 m). Modified from Clayton L, Attig JW, and Mickelson DM (1999) Tunnel channels formed in Wisconsin during the last glaciation. In: Mickelson DM and Attig JW (eds.) *Glacial Processes Past and Present*, pp. 69–82, Special Paper 337. Boulder, CO: Geological Society of America.

hanging, channel sections adorning uplands without integration to other channels or modern drainage.

Valley dimensions range from 0.15 to 6 km in width and 2–150 km in length, but the majority of valleys are 0.5–2.5 km wide and 5–40 km long. Valley depths of 10–50 m are common, with most ranging between 20 and 30 m, although subsurface studies of deeply buried tunnel channels document depths in excess of 100 m (Piotrowski, 1994). Other subsurface investigations have indicated that the subtle surface expression of some valleys is attributable to burial or partial burial by thick sequences of younger sediment that obscures the true depth of the valley bottoms (Clayton et al., 1999). Despite a general lack of adequate subsurface information and a scarcity of exposures, available studies suggest that a wide range of sedimentary sequences fills or partially fills tunnel valleys (Benn and Evans, 2010; Brennand and Shaw, 1994; Piotrowski, 1994).

Landform Associations

Tunnel valleys are often closely associated with drumlins, eskers, moraines, glaciofluvial meltwater fans, and hummocky topography. Several studies have noted that tunnel valleys cross-cut drumlins (Brennand and Shaw, 1994), contain eskers (Kehew and Kozłowski, 2007), and end at past ice-marginal positions (Clayton et al., 1999; Cutler et al., 2002), where they cut through the terminal moraine at the apex of large, coarse-grained glaciofluvial fans (Figures 11 and 12). Where active ice margins were in close proximity but not necessarily in synchronicity, some tunnel channels display cross-cutting relationships with younger landforms and have been interpreted as palimpsest tunnel channels (Kehew et al., 1999). These

cross-cutting relationships may be used to establish or constrain relative timing of glacial events in complex glaciated regions where the chronology of events is poorly constrained (Kehew et al., 1999; Kozłowski et al., 2005).

Although widely varying in morphology, drainage pattern, and associated landforms, there is a general consensus that tunnel valleys

- are subparallel to glacial flow lines, drumlins, and eskers;
- are usually straight and lack tributaries except for postglacial gullies;
- are most distinct near ice margins;
- have undulatory longitudinal profiles with uphill reaches;
- commonly contain eskers; and
- often breach moraines and terminate at large, coarse-grained glaciofluvial fans.

Origin

Most of the controversy surrounding tunnel channels centers on the models proposed to explain their formation (O’Cofaigh, 1996) and relationships with various landform associations (Benn and Evans, 2010). There is a fundamental difference in interpretation between those who use the term ‘tunnel channel’ to represent bankfull meltwater flow during one or more discharge events (Clayton et al., 1999) and those who use the term ‘tunnel valley,’ which implies gradual erosion by flows much smaller than the dimensions of the valley (Boulton and Hindmarsh, 1987; Mooers, 1989).

Although multiple models of formation have been proposed, most fall into four categories: (1) catastrophic meltwater floods, (2) time-transgressive erosion, (3) subglacial

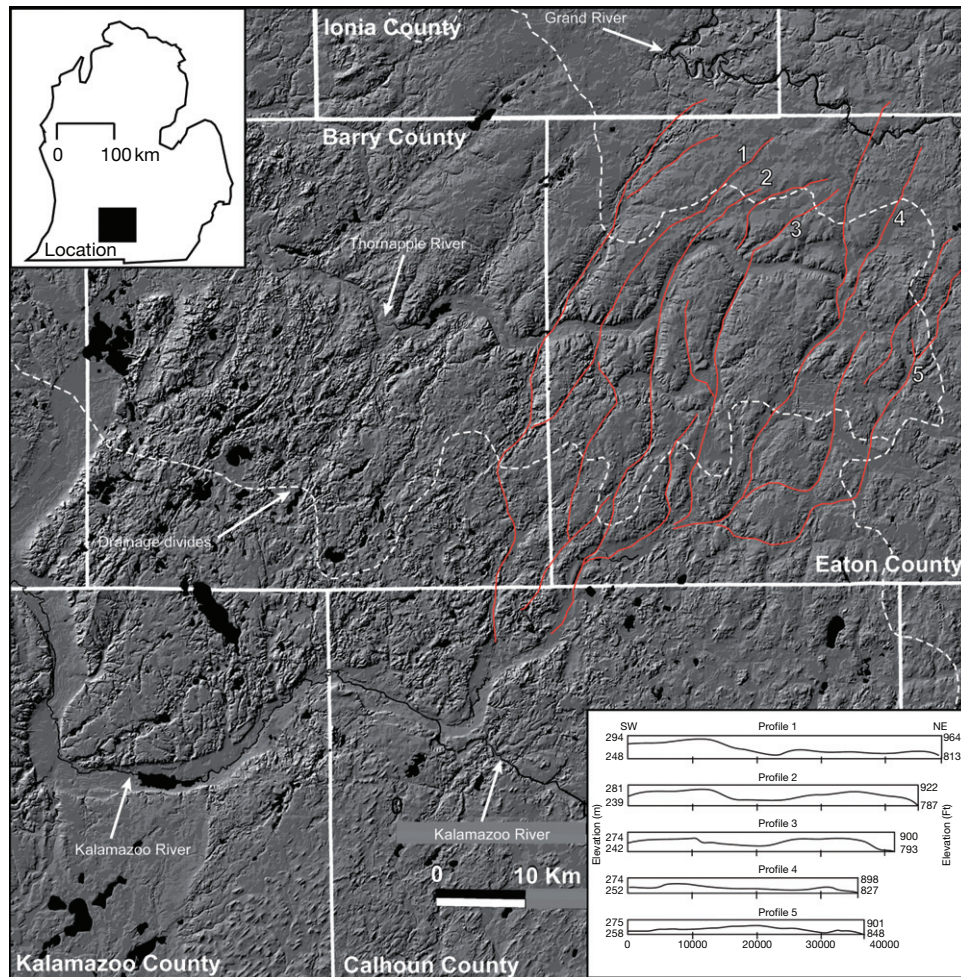


Figure 10 Digital elevation model of south-central Michigan displaying a series of anastomosing Saginaw Lobe tunnel channels (red lines) with segments that flowed uphill and crossed drainage divides (dashed white lines on DEM and inset profiles) to feed a subaerial outburst flood in the Central Kalamazoo River Valley. Modified from Kozłowski AL, Kehew AE, and Bird BC (2005) Outburst flood origin of the Central Kalamazoo River Valley, Michigan, USA. *Quaternary Science Reviews* 24: 2354–2374.

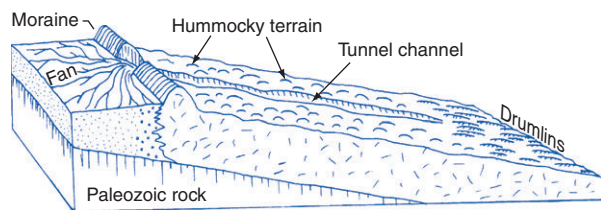


Figure 11 Idealized block model of tunnel channels and spatial relationships of associated landforms observed in Wisconsin, USA. Note up-hill flow path and termination of channel at an outwash fan through a breach in a moraine. Modified from Clayton L, Attig JW, and Mickelson DM (1999) Tunnel channels formed in Wisconsin during the last glaciation. In: Mickelson DM and Attig JW (eds.) *Glacial Processes Past and Present*, pp. 69–82, Special Paper 337. Boulder, CO: Geological Society of America.

sediment deformation and piping, and (4) polygenetic origin (Figure 13). Catastrophic meltwater release is the most commonly suggested mechanism for tunnel channel formation and, although differences in opinion exist over the exact nature

of the flow regime, most authors argue for large-scale transient events. The model shown in Figure 14 illustrates that after erosion by a large discharge, ice will flow under pressure into the valley. Smaller discharges confined to a restricted area of the former channel then form eskers over longer time periods. Several researchers have suggested that formation of tunnel channels is largely controlled by glacier-bed conditions and the presence of a permafrost seal at the glacier margin (Clayton et al., 1999; Wright, 1973). Hooke and Jennings (2006) have developed a model for channel erosion, which begins by discharge of pressurized groundwater at or near the margin. High fluid-pressure gradients lead to erosion of soil particles, a process called piping, and headward extension of the channel. When the channel reaches a reservoir of subglacial meltwater behind the permafrost wedge, it catastrophically drains to the margin (Figure 15). The source of meltwater remains a contentious issue and significant subglacial reservoirs of meltwater are required for rapid erosion of channels of the scale of those observed.

Alternatively, Mooers (1989) suggested that formation of large erosional troughs in Minnesota resulted from

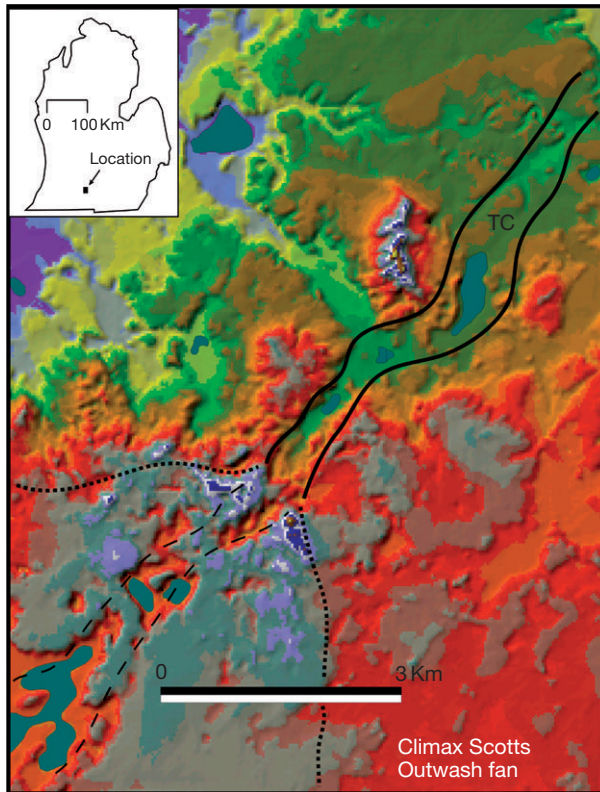


Figure 12 Digital elevation model of a Saginaw Lobe tunnel channel fan in south-central Michigan, USA. Note channel ascends in elevation to the apex of the fan, which is the highest feature on the landscape and delineates a former ice margin. Also, the collapsed appearance of the channel extending through the medial and distal portions of the fan suggests the channel extended beyond this margin and had been occupied prior to burial.

time-transgressive discharge of seasonal supraglacial meltwater that accessed the subglacial system via crevasses and moulins. Valley formation, as opposed to channel formation, resulted incrementally from vertical and lateral planation of subglacial meltwater streams (**Figure 13(b)**). This model accounts for the association of eskers and meltwater fans sometimes observed with tunnel channels, does not require catastrophic flows, and does not require permafrost to act as a hydraulic seal. In this model, small subglacial discharges gradually eroded the observed valleys as the ice margin retreated.

Subglacial deformation in conjunction with groundwater sapping, is another formational model that has been suggested for the development of tunnel valleys (Boulton and Hindmarsh, 1987). In this model, basal meltwater is unable to infiltrate downward through deforming subglacial sediment because of the low transmissivity of the deforming beds. Lateral groundwater flow toward low-pressure tunnels is initiated in the substrate to sustain meltwater discharge and maintain glacier equilibrium. Because of the high fluid-pressure gradients, sediment deforms into the meltwater conduit by creep or piping and is entrained and transported down glacier, initiating or enlarging the tunnel (**Figure 13(c)**). In time, this process leads to a gradual formation of a valley much larger than the hydraulic conduit.

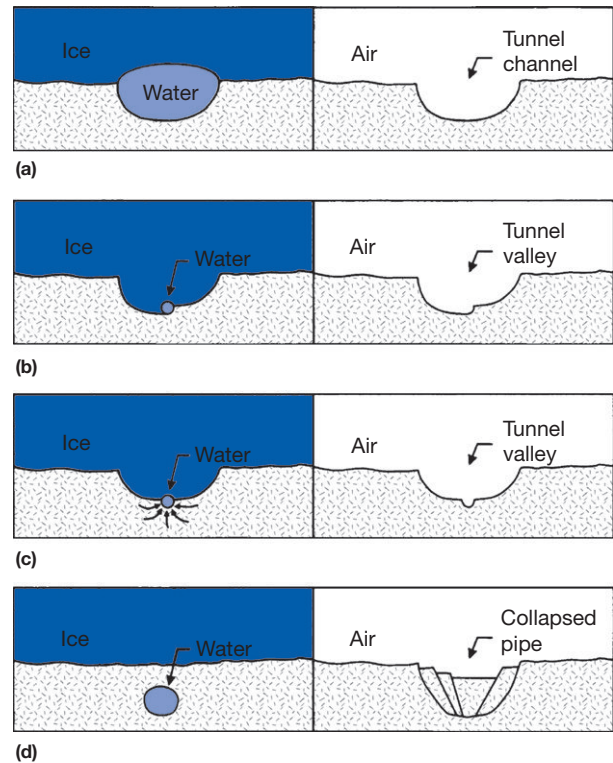


Figure 13 Schematic diagram illustrating proposed models of tunnel channel formation. (a) Catastrophic meltwater floods, (b) time-transgressive erosion, (c) subglacial deformation, and (d) hydraulic sapping or piping and collapse. Modified from Clayton L, Attig JW, and Mickelson DM (1999) Tunnel channels formed in Wisconsin during the last glaciation. In: Mickelson DM and Attig JW (eds.) *Glacial Processes Past and Present*, pp. 69–82, Special Paper 337. Boulder, CO: Geological Society of America.

The diversity in tunnel-valley morphology, drainage patterns, and substrate suggests a polygenetic origin for some valleys, which may involve preglacial, subglacial, glacial, and subaerial erosion (Benn and Evans, 2010; Brennand and Shaw, 1994; Clayton et al., 1999; Jørgensen, and Sandersen, 2006; O’Cofaigh, 1996; Sjogren et al., 2002). Once formed, tunnel valleys may be reoccupied during multiple glacial advances or glaciations (Benn and Evans, 2010). Buried and open tunnel valleys in northwestern Europe date from the Elsterian, Saalian, and Weichselian glaciations (**Figure 16**). The larger valleys are more than 100 m deep and completely buried. Huuse and Lykke-Andersen (2000) suggest that the origin of these large valleys involves a combination of repeated cycles of erosion and downcutting by catastrophic events, separated by steady-state subglacial meltwater flow and glacial erosion when closure by ice creep occurs. Deposition of a variety of sediment types occurs during interglacial periods.

Ice-Marginal (Lateral) Meltwater Channels

Overview

Channels eroded along or adjacent to former ice margins are referred to as lateral meltwater channels. These channels often occur in parallel, inset sequences, delimiting the gradual retreat

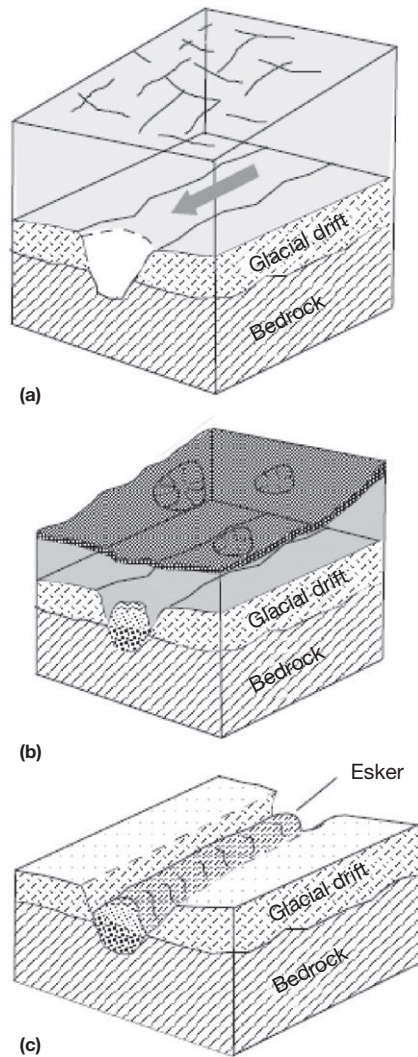


Figure 14 Conceptual model of tunnel-valley formation by high-discharge subglacial meltwater flows (a), partial closure of the valley by ice creep and drainage of lower-discharge flows through a smaller conduit during ice wastage (b), and final melting of glacier leaving an esker in a tunnel valley (c).

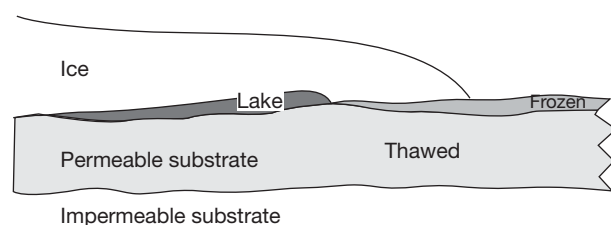


Figure 15 Formation of a subglacial lake behind a permafrost wedge. Groundwater under pressure from the ice overburden would drain toward the margin and break through the permafrost, initiating a conduit by piping. The conduit would extend headward to the lake and drain it catastrophically to form a tunnel channel. Reproduced from Hooke R LeB and Jennings CE (2006) On the formation of the tunnel channels of the southern Laurentide ice sheet. *Quaternary Science Reviews* 25: 1364–1372.

of a former glacier. Although not completely restricted to cold-based glaciers, lateral meltwater channels are best developed where glaciers are frozen to their beds.

Occurrence and Origin

Cold-based glaciers are most common in polar regions such as the Canadian Arctic Islands. Dyke (1993) mapped the landforms on several islands that supported ice caps during the last glaciation. These ice caps are now thought to be part of the Innuitian Ice Sheet (Dyke et al., 2002), which was confluent with the Greenland and Laurentide Ice Sheets. The thermal regime of these glaciers at the time of their maximum extents consisted of cold-based interiors and warm-based marginal areas. During deglaciation, they became entirely cold-based. Because cold-based glaciers have minimal subglacial meltwater, melting and meltwater production is confined to the ice surface. During the ablation season, meltwater from the ice and adjacent areas (if the topography slopes toward the glacier) drains toward the ice margin and then along the margin in the direction of the land slope. Channels are eroded by this meltwater flow. As the ice margin retreats, new channels replace those from the last period of meltwater flow, perhaps even forming on an annual basis (Dyke, 1993). In this way, a series of meltwater channels develops as the ice retreats (Figure 17).

As the result of the very minor amount of subglacial erosion and deposition performed by cold-based glaciers, lateral meltwater channels represent the most significant landforms produced by these ice masses. The closely spaced, parallel orientation of the channels suggests frequent abandonment in favor of new ice-marginal channels during orderly ice margin retreat, and the channels are therefore useful in mapping and defining the pattern of glacial retreat.

Although most common in cold-based glaciers, lateral meltwater channels may also form adjacent to temperate glaciers (Figure 18). Dyke (1993) attributes these occurrences to deflection of surface water to the margin by high subglacial meltwater fluid pressures or to intermittently or permanently frozen basal conditions during ice retreat.

Subglacial Sheetfloods – The Meltwater Hypothesis

Overview

Erosion of glacial substrates by subglacial sheetfloods, also known as the meltwater hypothesis, provides an alternative explanation for the formation of many glacial landforms that are traditionally interpreted as the result of direct action of glaciers upon their beds. A historical summary of the meltwater hypothesis (Shaw, 2002) reviews the major points of the hypothesis and includes a comprehensive citation of the literature.

Bedrock Erosional Forms

Glacier beds composed of hard bedrock exhibit a wide variety of erosional marks and forms. Some, such as striations, clearly formed by glacial abrasion. Other exposures, in which the bedrock surface is shaped into an assemblage of smoothed, streamlined forms, are more difficult to interpret. The name

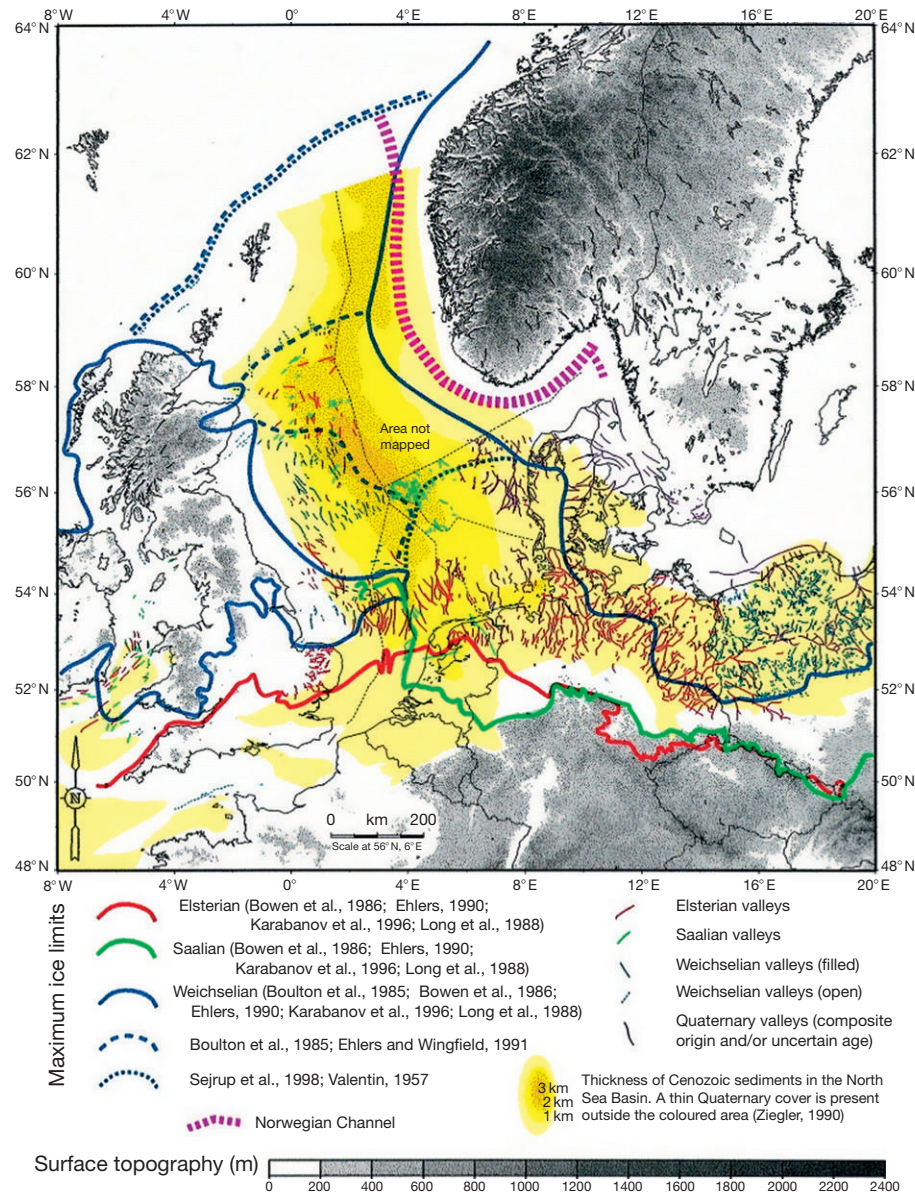


Figure 16 Location and relative age of on-shore and off-shore tunnel valleys in northwestern Europe. Valleys are linked to the Elsterian, Saalian, and Weichselian glaciations. Reproduced from Huuse M and Lykke-Andersen M (2000) Overdeepened Quaternary valleys in the Danish North Sea: Morphology and origin. *Quaternary Science Reviews* 19: 1233–1253.

'P form' (plastically molded) has been applied to these features, with the interpretation that they were produced primarily by the action of plastically deforming ice. Kor et al. (1991) concluded that these forms were instead the product of a turbulent, viscous fluid – sediment-rich subglacial meltwater – and suggested the generic term 'S form' (sculpted) for them. In their study area on the northeastern side of Georgian Bay, Lake Huron, the forms range in size from centimeters to tens of meters. Their classification of the forms is shown in Figure 19. Formation by meltwater rather than plastically deforming ice, or till, is based on form analogy to marks produced by experimental erosion of cohesive or hard substrates by turbulent fluid flows. The process envisioned for the formation of S

forms by meltwater involves the action of vortices and other flow structures impinging on an irregular bed (Kor et al., 1991). One of these flow structures, the horseshoe vortex (Shaw, 1994), is illustrated in Figure 20, and an exposure of sculpted bedrock on Kelley's Island, Lake Erie, is shown in Figure 20. The Georgian Bay flood was estimated by Kor et al. (1991) to have been 70 km wide and 10 m deep, with a velocity of $5\text{--}10\text{ m s}^{-1}$, producing a discharge of $0.4\text{--}0.7 \times 10^7\text{ m}^3\text{ s}^{-1}$.

Pair (1997), in a study of bedrock erosional forms in northern New York, reached the conclusion that, although many of the forms were indicative of meltwater erosion, they were limited to bedrock valleys in which channelized subglacial drainage would have been concentrated. Pair also points out

that plastic deformation of basal ice over an irregular bedrock surface would prevent the kind of broad decoupling of the ice from its bed required by the meltwater hypothesis (Figure 21).

Drumlins, Rogens, Hummocky Terrain, Megaflutes, and Tunnel Valleys

Although the immense discharges required by the meltwater hypothesis attract continued criticism, there is general agreement that at least some of the S forms observed on bedrock surfaces were eroded by turbulent subglacial meltwater. The meltwater hypothesis encounters an even greater degree of controversy, however, in the explanation of drumlins, rogens, and hummocky terrain, landforms occurring in mostly unconsolidated materials, as the result of subglacial



Figure 17 A series of subparallel, nested lateral meltwater channels, Melville Peninsula, Nunavut, Arctic Canada. Channels developed adjacent to a cold-based ice margin. Photograph by Lynda Dredge. Reproduced with the permission of the Minister of Public Works and Government Services Canada, 2006 and Courtesy of Natural Resources Canada, Geological Survey of Canada.

megaflows rather than the traditional explanation of direct glacial action. This extension of the hypothesis was initially applied to the Livingstone Lake drumlin field in northern Saskatchewan. Form analogy of the drumlins to the inverted sole of a turbidite bed led to the inference that the drumlins formed by infilling of erosional marks scoured upward into the base of the ice sheet, perhaps during the waning phases of the flood (Shaw and Kvill, 1984). Shaw et al. (1989) estimated the total volume of water released as $84 \times 10^3 \text{ km}^3$.

The recognition of drumlins with truncated internal structure at land surface led to the broadening of the meltwater hypothesis to include drumlins formed by scour of the subglacial bed by the sheetflood rather than scour of the base of the ice (Shaw and Sharpe, 1987). Form analogy is also used here as a basis for the process, in this case to the large-scale, streamlined erosional hills of the Channeled Scabland and elsewhere. Interpretation of landforms other than drumlins by similar mechanisms was soon to follow. Rogen moraine was explained by infill of scours in basal ice oriented transverse to the flow direction of the sheet flow (Fisher and Shaw, 1992). Munro and Shaw (1997) added hummocky terrain to the hypothesis. This interpretation was based on landscapes in Alberta where hummocks, covered with boulder lags, have internal structures that are truncated at the land surface. The hummocky terrain occurs adjacent to incised 'flood tracts'; the hummocky terrain itself is attributed to less severe erosion on the flanks of the main channels or tracts. Tunnel valleys are usually explained by the collapse of the sheet flows that produce drumlins into channelized flows.

Spirited debate continues on the meltwater hypothesis (Evans et al., 2006). Form analogy as the most fundamental evidence for landforms produced by subglacial sheetfloods has been challenged on several grounds. In the case of cavity infill landforms, confirmation of process by form analogy requires extrapolation over orders of magnitude; in addition, Benn and Evans (2010) have pointed out that dissimilar processes can produce similar landforms. The formation of drumlins, rogen

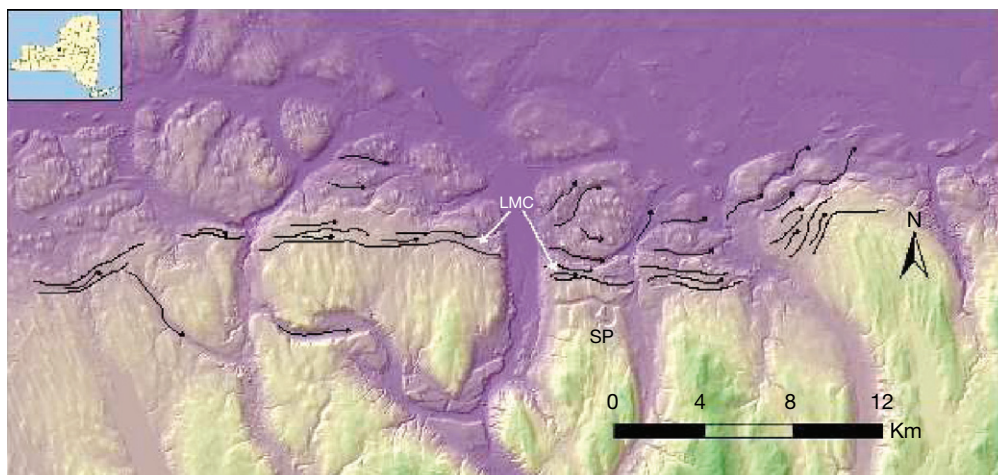


Figure 18 Lateral meltwater channels (LMC) and spillways near Syracuse, NY, USA. Lateral channels developed as meltwater from the Ontario Lobe of the Laurentide Ice Sheet grounded against the Onondaga Escarpment. Meltwater discharges trapped between the ice margin and uplands scoured channels (outlined and arrows) down to bedrock. Some of the deep north-south valleys carved into the escarpment contained proglacial lakes. Spillways (SP) were incised as ice retreated northward and lower outlets opened.

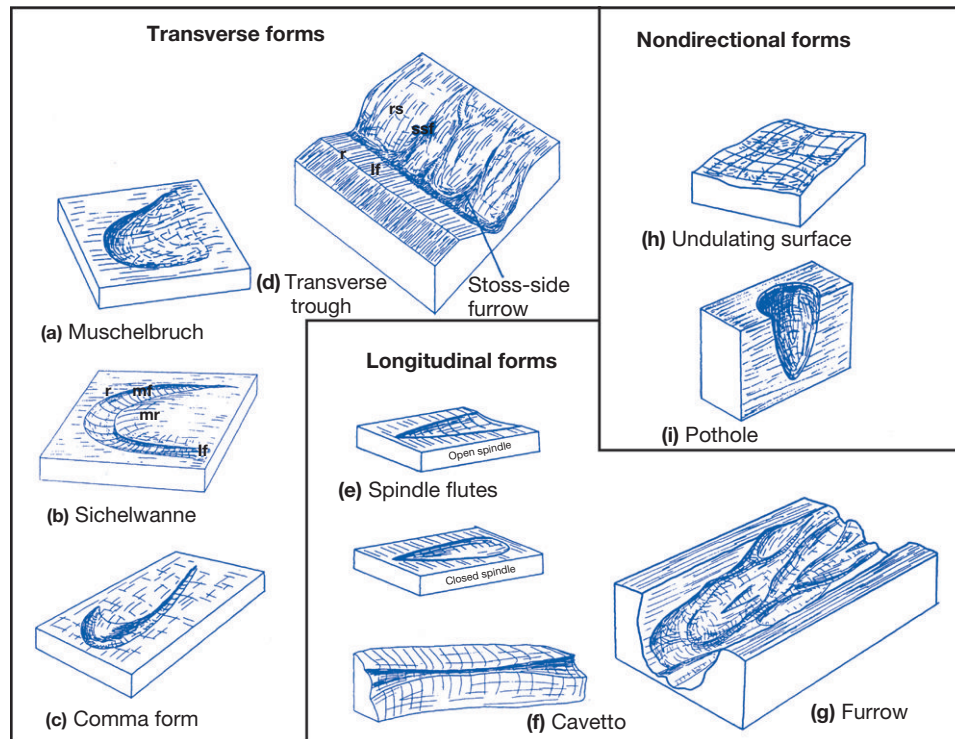


Figure 19 Classification of S forms, interpreted to be eroded on rock surfaces by subglacial meltwater. Reproduced from Kor PSG, Shaw J, and Sharpe DR (1991) Erosion of bedrock by subglacial meltwater, Georgian Bay, Ontario: A regional view. *Canadian Journal of Earth Sciences* 28: 623–642.

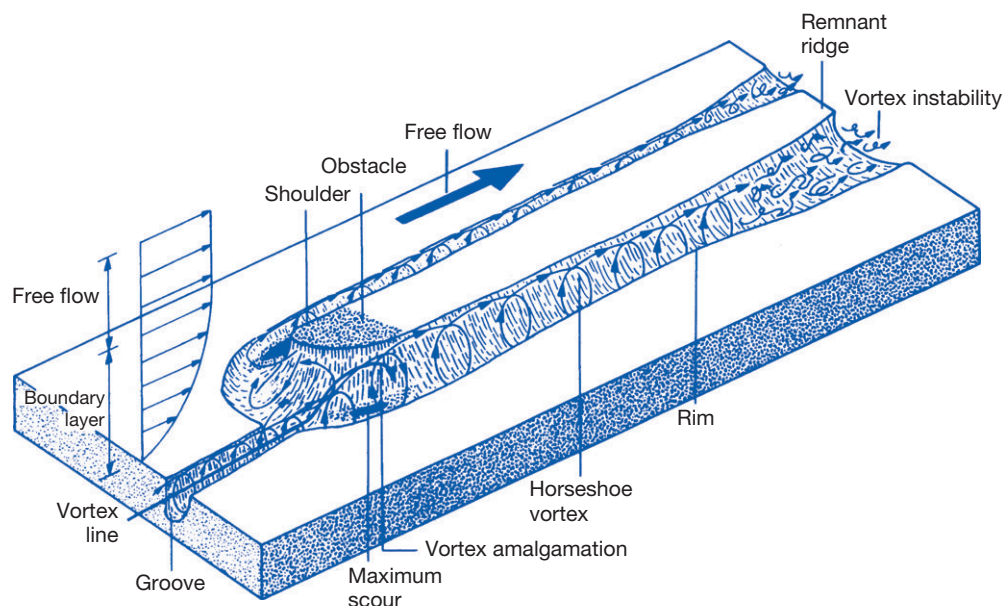


Figure 20 Erosion of bedrock by a horseshoe vortex in subglacial meltwater. Reproduced from Shaw J (1994) Hairpin erosional marks, horseshoe vortices and subglacial erosion. *Sedimentary Geology* 91: 269–283.

moraine, and megaflutes by bed deformation by glacial ice is well established in the literature, although details continue to be controversial. Other problematical aspects of the meltwater hypothesis include the lack of erosional landforms beyond the ice margins from which megafloods emerged and the lack of

evidence for the huge meltwater reservoirs necessary for the generation of megafloods. Clarke et al. (2005) contend that both the production of sufficient meltwater and its storage on top of or beneath ice sheets are unlikely without special circumstances.



Figure 21 Exposure of a sculpted bedrock surface on Kelley's Island, Lake Erie. Photograph by Alan Kehew.

See also: **Glacial Landforms, Sediments:** Glaciomarine Sediments and Ice-Rafted Debris; Till Fabric Analysis. **Glaciations:** Early Quaternary (Pleistocene) and Precursors; Late Quaternary in Highland Asia. **Permafrost and Periglacial Features:** Cryoturbation Structures.

References

- Baker VR (1973) *Paleohydrology and Sedimentology of the Lake Missoula Flooding in Eastern Washington*. Boulder, CO: Geological Society of America Special Paper 144.
- Baker VR (2009) The Channeled Scabland: A retrospective. *Annual Review of Earth and Planetary Sciences* 37: 393–411.
- Baker VR, Benito G, and Rudoy AN (1993) Paleohydrology of late Pleistocene superflooding, Altay Mountains, Siberia. *Science* 259: 348–350.
- Baker VR, Greeley R, Komar PD, Swanson DA, and Waitt RB Jr. (1987) Columbia and Snake River Plains. In: Graf WL (ed.) *Geomorphic Systems of North America*, Vol. 2, pp. 403–468. Boulder, CO: Geological Society of America Centennial Special.
- Benn DI and Evans DJA (2010) *Glaciers and Glaciation*, 2nd edn. Oxon, UK: Hodder Education.
- Boulton GS and Hindmarsh RCA (1987) Sediment deformation beneath glaciers: Rheology and geological consequences. *Journal of Geophysical Research* 92: 9059–9082.
- Brennand TA and Shaw J (1994) Tunnel channels and associated landforms, south-central Ontario: Their implications for ice-sheet hydrology. *Canadian Journal of Earth Sciences* 31: 505–522.
- Bretz JH (1923) The Channeled Scabland of the Columbia Plateau. *Journal of Geology* 31: 617–649.
- Burr DM, Carling PA, and Baker VR (eds.) (2009) *Megaflooding on Earth and Mars*, p. 330. Cambridge, UK: Cambridge University Press.
- Carling PA, Herget J, Lanz JK, Richardson K, and Pacifici A (2009) Channel-scale erosional bedforms in loose granular material: Character, processes and implications. In: Burr DM, Carling PA, and Baker VR (eds.) *Megaflooding on Earth and Mars*, pp. 13–32. Cambridge, UK: Cambridge University Press.
- Cenderelli DA and Wohl EA (2003) Flow hydraulics and geomorphic effects of glacial-lake outburst floods in the Mount Everest region, Nepal. *Earth Surface Processes and Landforms* 28: 385–407.
- Clarke GKC, Leverington DW, Teller JT, Dyke AS, and Marshall SJ (2005) Fresh arguments to comments by David Sharpe on "Paleohydraulics of the last outburst flood from glacial Lake Agassiz and 8200 BP cold event" *Quaternary Science Reviews* 24: 1533–1541.
- Clayton L, Attig JW, and Mickelson DM (1999) Tunnel channels formed in Wisconsin during the last glaciation. In: Mickelson DM and Attig JW (eds.) *Glacial Processes Past and Present*, pp. 69–82. Boulder, CO: Geological Society of America Special Paper 337.
- Cutler PM, Colgan PM, and Mickelson DM (2002) Sedimentologic evidence for outburst floods from the Laurentide Ice Sheet margin in Wisconsin, USA: Implications for tunnel channel formation. *Quaternary International* 90: 23–40.
- Denton GH and Sugden DE (2005) Meltwater features that suggest Miocene ice-sheet overriding of the Transantarctic Mountains in Victoria Land, Antarctica. *Geografiska Annaler* 87A(1): 67–85.
- Dyke AS (1993) Landscapes of cold-centered Late Wisconsinan ice caps, Arctic Canada. *Progress in Physical Geography* 17: 223–247.
- Dyke AS, Andrews JT, Clark PU, et al. (2002) The Laurentide and Innuitian ice sheets during the Last Glacial Maximum. *Quaternary Science Reviews* 21: 9–31.
- Evans DJA, Rea BR, Hiemstra JF, and Cofaigh CO (2006) A critical assessment of subglacial mega-floods: A case study of glacial sediments and landforms in south-central Alberta, Canada. *Quaternary Science Reviews* 25: 1638–1667.
- Fisher TG and Shaw J (1992) A depositional model for rogen moraine, with examples from the Avalon Peninsula, Newfoundland. *Canadian Journal of Earth Sciences* 29: 669–686.
- Hooke R LeB and Jennings CE (2006) On the formation of the tunnel channels of the southern Laurentide ice sheet. *Quaternary Science Reviews* 25: 1364–1372.
- Huuse M and Lykke-Andersen M (2000) Overdeepened Quaternary valleys in the Danish North Sea: Morphology and origin. *Quaternary Science Reviews* 19: 1233–1253.
- Jørgensen F and Sandersen PBE (2006) Buried and open tunnel valleys in Denmark – Erosion beneath multiple ice sheets. *Quaternary Science Reviews* 25: 1339–1363.
- Kehew AE and Kozlowski AL (2007) Tunnel channels of the Saginaw Lobe, Michigan, USA. In: Johannsson P and Sarala P (eds.) *Applied Quaternary Research in the Central Part of Glaciated Terrain*, pp. 69–77. Finland: Geological Survey of Finland Special Paper 46.
- Kehew AE and Lord ML (1987) Glacial-lake outburst along the mid-continent margins of the Laurentide Ice Sheet. In: Mayer L and Nash D (eds.) *Catastrophic flooding, Proceedings of the 18th Annual Binghamton Symposium in Geomorphology*, 95–120.
- Kehew AK, Lord ML, Kozlowski AL, and Fisher TG (2009) Proglacial megaflooding along the margins of the Laurentide Ice Sheet. In: Burr DM, Carling PA, and Baker VR (eds.) *Megaflooding on Earth and Mars*, pp. 104–127. Cambridge, UK: Cambridge University Press.
- Kehew AE, Nicks LP, and Straw TW (1999) Palimpsest tunnel valleys: Evidence for relative timing of advances in an interlobate area of the Laurentide ice sheet. *Annals of Glaciology* 28: 47–52.
- Kor PSG, Shaw J, and Sharpe DR (1991) Erosion of bedrock by subglacial meltwater, Georgian Bay, Ontario: A regional view. *Canadian Journal of Earth Sciences* 28: 623–642.
- Kozlowski AL, Kehew AE, and Bird BC (2005) Outburst flood origin of the Central Kalamazoo River Valley, Michigan, U.S.A. *Quaternary Science Reviews* 24: 2354–2374.
- Lord ML (1991) Depositional record of a glacial-lake outburst: Glacial Lake Souris, North Dakota. *Geological Society of America Bulletin* 103: 663–673.
- Lord ML and Kehew AK (1987) Sedimentology and paleohydrology of glacial-lake outbursts in southeastern Saskatchewan and northwestern North Dakota. *Geological Society of America Bulletin* 99: 663–673.
- Moers HD (1989) On formation of the tunnel valleys of the Superior lobe, central Minnesota. *Quaternary Research* 32: 24–35.
- Munro M and Shaw J (1997) Erosional origin of hummocky terrain in south-central Alberta. *Geology* 25: 1027–1030.
- Normark WR and Reid JA (2003) Extensive deposits on the Pacific Plate from late Pleistocene North American glacial lake bursts. *Journal of Geology* 111: 617–637.
- O'Cofaigh C (1996) Tunnel valley genesis. *Progress in Physical Geography* 20: 1–19.
- O'Connor JE and Baker VR (1992) Magnitudes and implications of peak discharges from Glacial Lake Missoula. *Geological Society of America Bulletin* 104: 267–279.
- O'Connor JE and Costa JE (2004) The world's largest floods, past and present: Their causes and magnitude. *U.S. Geological Survey Circular* 1254: 13 p.
- Pair DL (1997) Thin film, channelized drainage, or sheetfloods beneath a portion of the Laurentide Ice Sheet: An examination of glacial erosion forms, northern New York State. *Sedimentary Geology* 111: 199–216.
- Piotrowski JA (1994) Tunnel-valley formation in northwest Germany – Geology, mechanisms of formation and subglacial bed conditions for the Bornhöved tunnel valley. *Sedimentary Geology* 89: 107–141.
- Shaw J (1994) Hairpin erosional marks, horseshoe vortices and subglacial erosion. *Sedimentary Geology* 91: 269–283.
- Shaw J (2002) The meltwater hypothesis for subglacial bedforms. *Quaternary International* 90: 5–22.
- Shaw J and Kvill D (1984) A glaciofluvial origin for drumlins of the Livingstone Lake Area, Saskatchewan. *Canadian Journal of Earth Sciences* 21: 1442–1459.
- Shaw J, Kvill D, and Rains RB (1989) Drumlins and catastrophic subglacial floods. *Sedimentary Geology* 62: 177–202.

- Shaw J and Sharpe DR (1987) Drumlin formation by subglacial meltwater erosion. *Canadian Journal of Earth Sciences* 24: 2316–2322.
- Shepard RG and Schumm SA (1974) Experimental study of river incision. *Geological Society of America Bulletin* 85: 257–268.
- Sjogren DB, Fisher TG, Taylor LD, Jol HM, and Munro-Stasiuk MJ (2002) Incipient tunnel channels. *Quaternary International* 90: 41–56.
- Wohl EE and Ikeda H (1997) Experimental simulation of channel incision into a cohesive substrate at varying gradients. *Geology* 25: 295–298.
- Wright HE Jr. (1973) Tunnel valleys, glacial surges, and subglacial hydrology of the Superior Lobe, Minnesota. In: Black R, Goldthwait R, and Willman H (eds.) *The Wisconsin Stage. Geological Society of America Memoir*, vol. 136, pp. 251–276. Boulder, CO: Geological Society of America.

Glacitectonic Structures and Landforms

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Glacitectonic Processes

Proglacial glacitectonics involves the large-scale displacement of proglacial materials due to stresses imposed by glacier ice and involves ductile or brittle deformation or a combination of the two. Ductile deformation involves the production of large, open folds in sediments or rocks in front of an advancing glacier, which may develop into overfolds or begin to undergo internal thrusting due to continued ice advance. In contrast, brittle deformation involves the thrusting of semicoherent blocks along discrete planes of failure. In both cases, the lower limit of deformation is usually marked by a basal failure plane or plane of decollement, which often coincides with a sedimentary discontinuity or bedding plane. Although all materials are capable of undergoing both ductile and brittle failure according to variations in the applied stress, temperature, strain rate, and porewater pressure, glacitectonic disruption of frozen sediments appears to be dominated by brittle failure (e.g., Evans and England, 1991; Kalin, 1971; Klassen, 1982).

The compression of proglacial materials by glacitectonic deformation is very similar to the processes that occur during continental collision and mountain building (Aber, 1988; Croot, 1987). Indeed, proglacial glacitectonic landforms can be regarded as scale models of mountain chains and can be studied using well-founded principles of structural geology (see Twiss and Moores, 1992; Figure 1). The formation of large thrust–moraine complexes standing many tens of meters above the surrounding terrain depends on a number of factors, the most important of which are low-strength proglacial sediments and high, glacially imposed stresses. Sediment strength is dependent upon grain size and sorting, the existence of potential planes of failure, and porewater pressures. High porewater pressures result from subglacial drainage through inefficient distributed systems, such as porewater flow, and can be further encouraged by the existence of proglacial permafrost, which acts to confine pressurized porewater in underlying unfrozen aquifers (Mathews and Mackay, 1960).

The presence of weakened sediments, however, is not a sufficient condition for proglacial glacitectonic deformation, which also requires stresses large enough to elevate large masses of sediment above the glacier margin. It is clear that in most situations, the shear stresses beneath glacier margins are too low to produce the observed deformation. The solution to this problem was developed by Rotnicki (1976), van der Wateren (1985), and Aber et al. (1989), and is known as the ‘gravity spreading model.’ According to this model, proglacial sediment failure results from the translation of the weight of the glacier into lateral stresses, which push sediment wedges away from the load and then upward from their original position. The stress field producing a thrust block is therefore the product of the total weight of the ice mass and older thrust blocks. Williams et al. (2001) have demonstrated this concept on the St Bees Moraine in northwest England by applying a

critical wedge model developed in structural geology. In addition to gravity spreading, two other mechanisms may contribute to the development of thrust systems (Aber et al., 1989; Price and Cosgrove, 1990). First, gravity gliding or gravity sliding involves the deformation of sediment blocks as they move downslope under their own weight. This requires the uplifting of the ice proximal zone by gravity spreading to produce a slope upon which blocks can slide. Second, compression or push from behind may occur by the direct shoving of sediment wedges by the forward movement of the glacier. In reality, gravity spreading, gravity gliding, and compression can act in unison in proglacial settings.

The likelihood of glacitectonic disturbance is dictated by various factors (Aber et al., 1989), the most important of which are (i) slope of the proglacial area, whereby ice advance against a topographic obstacle increases the likelihood of proglacial glacitectonics; (ii) the presence of weak layers in the substratum, whereby failure is encouraged by weak layers or potential decollement planes; (iii) the nature of ice/sediment contact in the proglacial area, where partial burial of glacier snouts prior to their advance may lead to the lifting of sedimentary strata; and (iv) subglacial and proglacial drainage, whereby failure is encouraged by high porewater pressures in proglacial and submarginal sediments and rocks.

The shortening of proglacial sediments and rocks during glacitectonic deformation results in complex geological structures analogous in many ways to thrust-fold belts at the collision points of drifting continents, and can be explained using the principles of structural geology (Figure 2). Ductile deformation during glacitectonic compression and shortening results in a wide range of fold types, depending on the magnitude and orientation of the applied stresses and the strain response of the sediments. Folds range from long wavelength or open folds to isoclinal or recumbent (nappe) folds. Other fold types include chevron, sheath, kink band, box, and polyclinal folds. Complexities may arise where different sedimentary layers respond differently to stress and produce disharmonic folds, in which the fold wavelengths of some layers are smaller than those of others. Additionally, the response of water in Quaternary sediments when put under pressure results in a range of small-scale soft sediment deformation structures, hydrofracture fills, and burstout structures (Phillips et al., 2007; Rijdsdijk et al., 1999). Complex deformation histories can be evident in the superimposition of folds on preexisting structures (Berthelsen, 1978), allowing the reconstruction of glacier marginal oscillations (e.g., Phillips et al., 2002). Syntectonic sedimentation can also contribute to significant complexity in Quaternary glacial stratigraphies, an excellent example being that of the tills, glacitectonites, and associated sediments of the Norfolk coast in Britain, where proglacial tectonics led to the construction of thrust ridges and intervening depressions that acted as ‘foreland’ basins for localized sedimentation (Phillips et al., 2008; Figure 3).

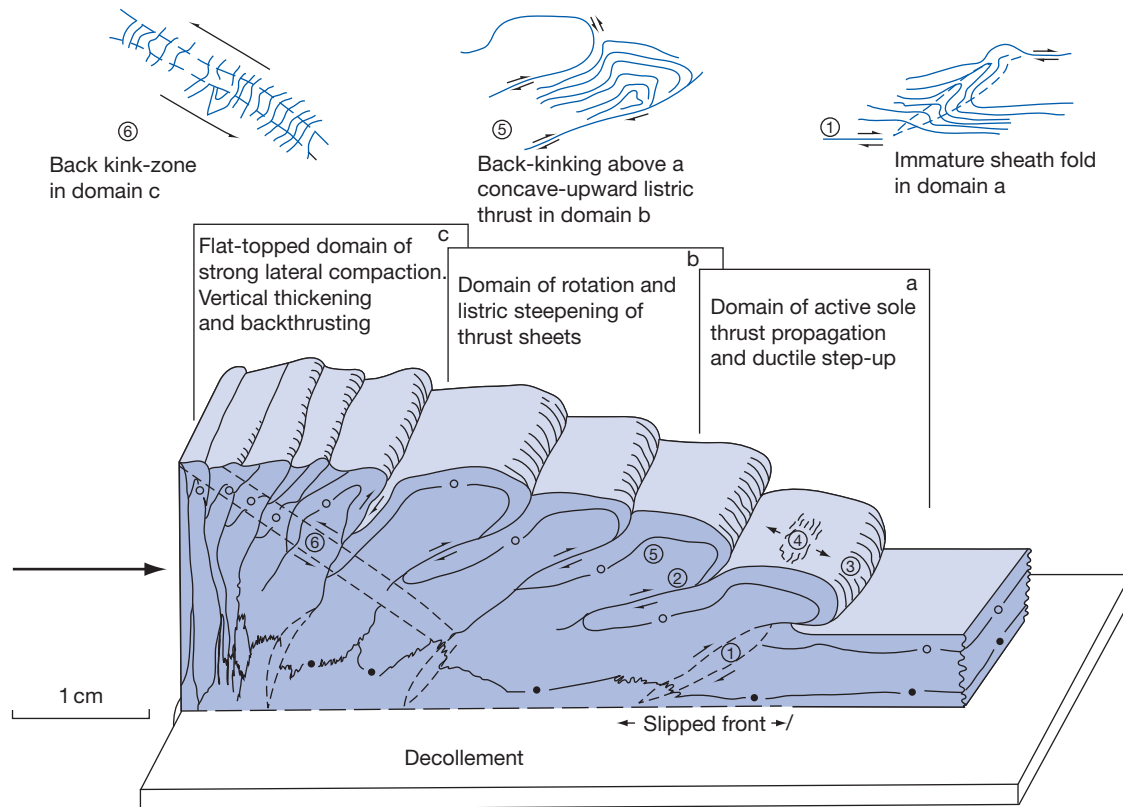


Figure 1 Example of the tectonic structures produced by the thrusting of stratified sand in a squeeze box. These features are typical of moraines constructed by proglacial glacitectonics: (1) initiation of thrust, (2) thrust fault, (3) slump zone, (4) extension fractures, (5) back kink fold, and (6) backthrust zone. Reproduced from Mulugeta G and Koyi H (1987) Three-dimensional geometry and kinematics of experimental piggyback thrusting. *Geology* 15: 1052–1056.

Composite Ridges and Thrust-Block Moraines

The most common and distinctive type of glacitectonic landforms are composite ridges and thrust-block moraines (Figure 4(a)). Composite ridges are composed of multiple slices of upthrust and contorted bedrock and/or unconsolidated sediments, which are commonly interlayered with and overlain by glaciogenic and glaciofluvial sediment (Aber et al., 1989). They are subdivided by Aber et al. (1989) into large (>100 m of relief) and small (<100 m of relief) composite ridges. This size differentiation is essentially arbitrary and the processes involved in the production of both large and small composite ridges are the same. However, Aber et al. (1989) suggested that large ridges usually contain a considerable volume of pre-Quaternary bedrock, whereas small ridges are composed mostly of unconsolidated Quaternary strata. The term thrust-block moraine was introduced by Evans and England (1991) to describe composite ridges comprising proglacially thrust masses of permafrozen blocks of Quaternary sediments at the margins of Arctic subpolar glaciers. Composite ridges are associated with proglacial or submarginal glacitectonics and they mark the positions of glacier stillstands or readvances.

In planform, composite ridges comprise arcuate suites of subparallel ridges and intervening depressions arranged in a concave up-glacier pattern and conforming to the general shape of the glacier margin that produced them. Individual

ridge crests correspond to the crests of internal folds, the upturned ends or edges of thrust blocks, or resistant upturned beds. The depth of disturbance associated with composite ridges is generally a few tens of meters and may be up to 200 m for the very largest features (Aber et al., 1989; Kupsch, 1962). The internal structures of large composite ridges are very similar to the products of thin-skinned tectonics in thrust and folded mountain belts like the Canadian Rockies, even though composite ridges are at least one order of magnitude smaller than true mountain ranges (Aber et al., 1989). The material in composite ridges commonly forms stacked sheets, with individual thrust slices showing varying degrees of internal deformation. Overturned and rootless folds of bedrock and Quaternary sediments in the composite ridges at Flade Klint, Denmark, have been thrust up to 100 m from their probable source depression (Gry, 1940). Probably, the most spectacular exposures through a large composite ridge are located at Møns Klint in Denmark. Several ice advances during the last glaciation have piled up numerous chalk scales with intervening Quaternary sediment to produce a composite ridge characterized by more than 150 m of structural relief (Aber, 1985). The structures within the cliffs at Møns Klint include multiple thrust anticlines and individually folded and stacked chalk floes, which can be traced as a series of ridges on the rugged highland of Høje Møn.

There are numerous examples of small composite ridges, both in front of modern glacier snouts and in the ancient

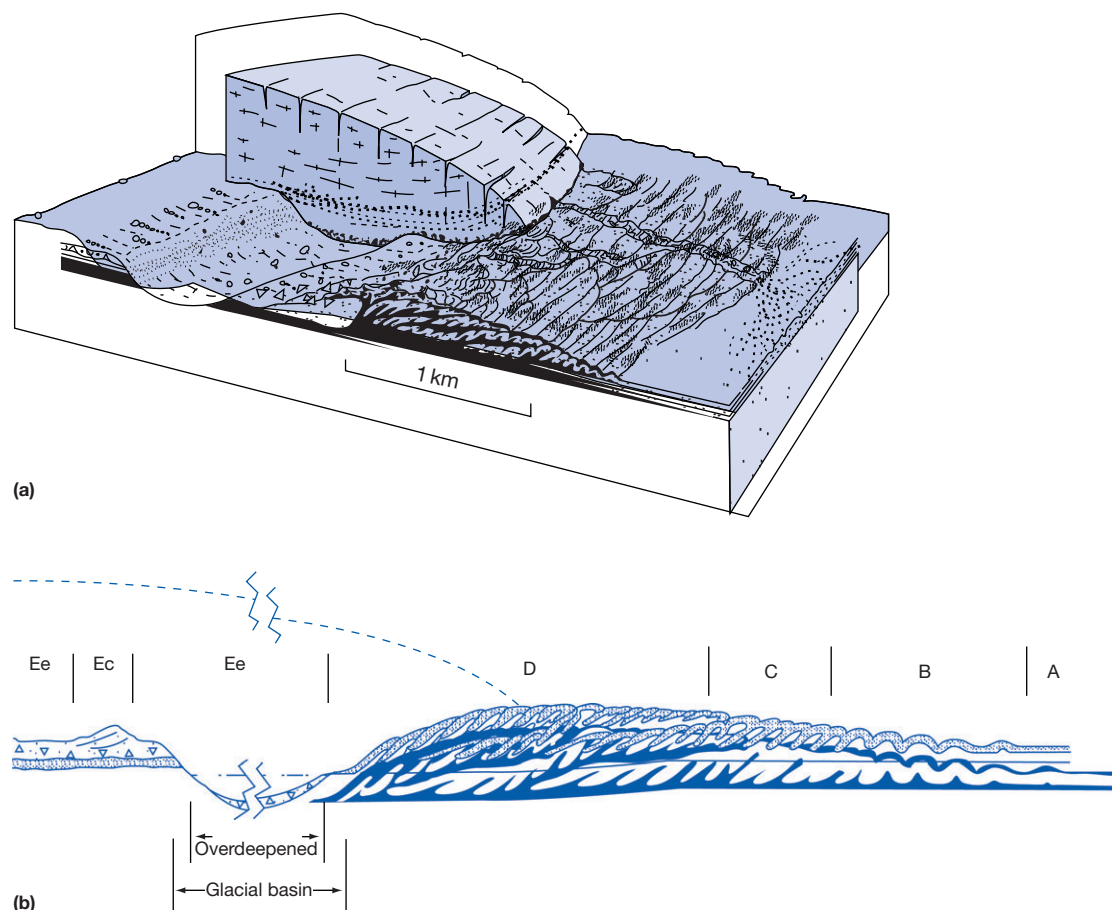


Figure 2 Glacitectonic structures and landforms associated with proglacial thrusting by the snout of Holmstrombreen, Svalbard: (a) block diagram of the thrust and folded sediments that constitute the moraine at the glacier snout; (b) styles of glacitectonic disturbance that can be observed at Holmstrombreen; A = undeformed foreland; B = 'Jura-style' high-angle folding; C = low-angle structures including thrusts and overturned and recumbent folds; D = nappes; E = deformed subglacial sediment/tills where Ec is compressive and Ee is extensional. Reproduced from [van der Wateren DFM \(1995\)](#) Processes of glaciotectionism. In: Menzies J (ed.) *Modern Glacial Environments: Processes, Dynamics and Sediments, Glacial Environments*, vol. 1, pp. 309–335. Oxford: Butterworth-Heinemann.

landform record. Spectacular composite ridges composed of outwash and glacial marine sediments form continuous moraines around the snouts of many Spitsbergen glaciers ([Croot, 1988](#); [Gripp, 1929](#); [Gripp and Todtmann, 1925](#)). In Spitsbergen and Iceland, composite ridges are strongly associated with surging glaciers, suggesting that conditions during surges are particularly conducive to proglacial tectonics ([Croot, 1987, 1988](#); [Evans and Rea, 2003](#); [Sharp, 1985](#)). The most important factors underlying this association are (i) rapid advance of the snout, (ii) extreme compressive deformation at the margin, and (iii) elevated pore-water pressures associated with the release of meltwater. The role of porewater pressures is highlighted by the general absence of composite ridges where proglacial areas slope steeply away from the glacier and water is able to drain away freely. The internal structures of recent composite ridges at Eyjabakkajökull, an Icelandic surging glacier, include low-angle subglacial thrust faults dipping down glacier, over which thrust slices have been transported during overall extension, and proglacial compression structures, such as stacked thrust sheets composed of outwash, tephra, turf, organic soil, and peat beds, which form asymmetric ridges. In the more distal parts of the moraine,

smaller asymmetric ridges are the surface expression of anticlines or overturned anticlines ([Croot, 1988](#)).

Thrust-block moraines produced by proglacial thrusting of blocks of frozen outwash and raised marine sediments have been documented in the Canadian arctic ([Evans, 1989](#); [Evans and England, 1991](#); [Kalin, 1971](#); [Klassen, 1982](#); [Figure 5](#)). In most examples, the bedding within individual blocks and block surfaces dip gently toward the glacier, suggesting that the moraines are composed of either stacked scales or partially rotated deep-seated blocks ([Kalin, 1971](#)). In some cases, however, bedding may dip away from the glacier snout suggesting deep-seated wedging of the advancing snout beneath proglacial sediments ([Evans and England, 1991](#)). The Canadian Arctic thrust-block moraines occur in areas where permafrost is often in excess of 700 m thick, but they are located in valley bottoms either below the marine limit or below former proglacial lake shorelines relating to the last glaciation. This suggests that permafrost may have been partially degraded below deep-water bodies prior to the Holocene advances responsible for thrusting. Therefore, block failure could have been initiated along an unfrozen talik between newly aggraded postglacial

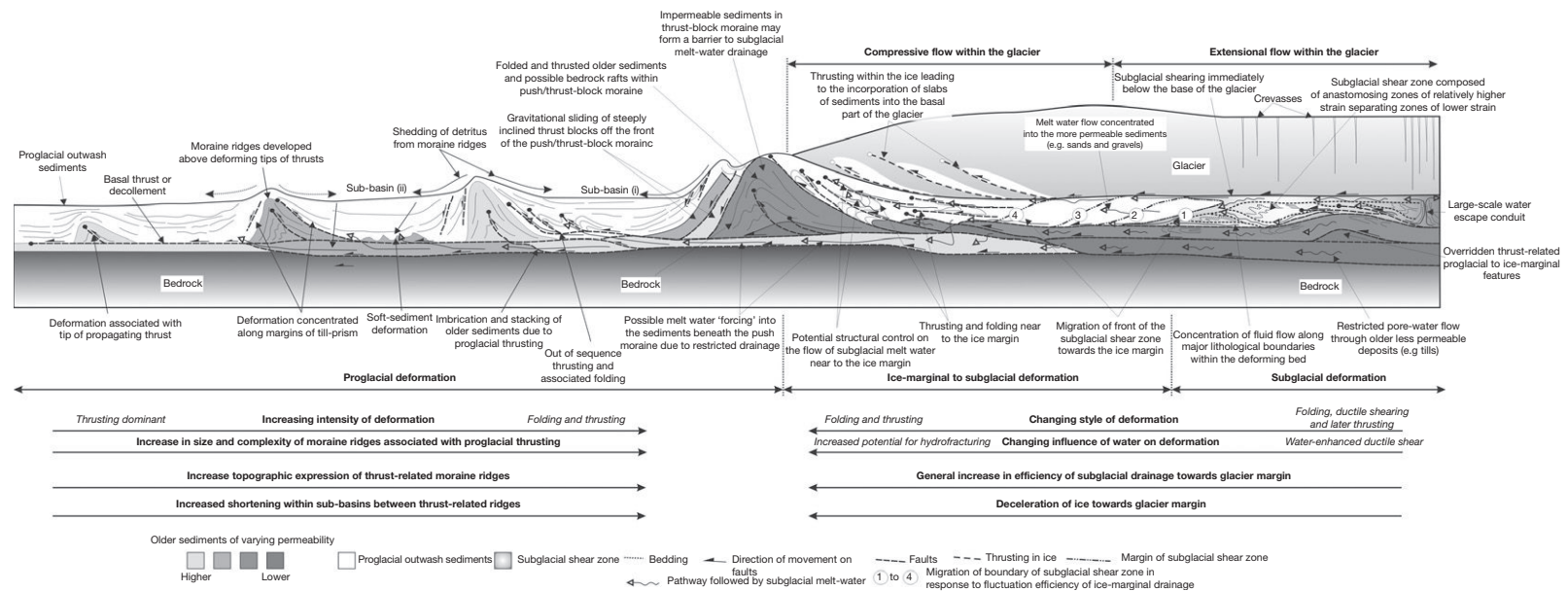


Figure 3 The evolution of complexly deformed ice sheet marginal sediments due to deformation of proglacial outwash, based on the Quaternary deposits of north Norfolk, England. This idealized reconstruction of the processes and products of progressive proglacial to subglacial deformation is based upon the features observed at the cliffs between West Runton and Sheringham. Reproduced from Phillips ER, Lee JR, and Burke H (2008) Progressive proglacial to subglacial deformation and syntectonic sedimentation at the margins of the mid-Pleistocene British Ice Sheet: Evidence from north Norfolk, UK. *Quaternary Science Reviews* 27: 1848–1871.

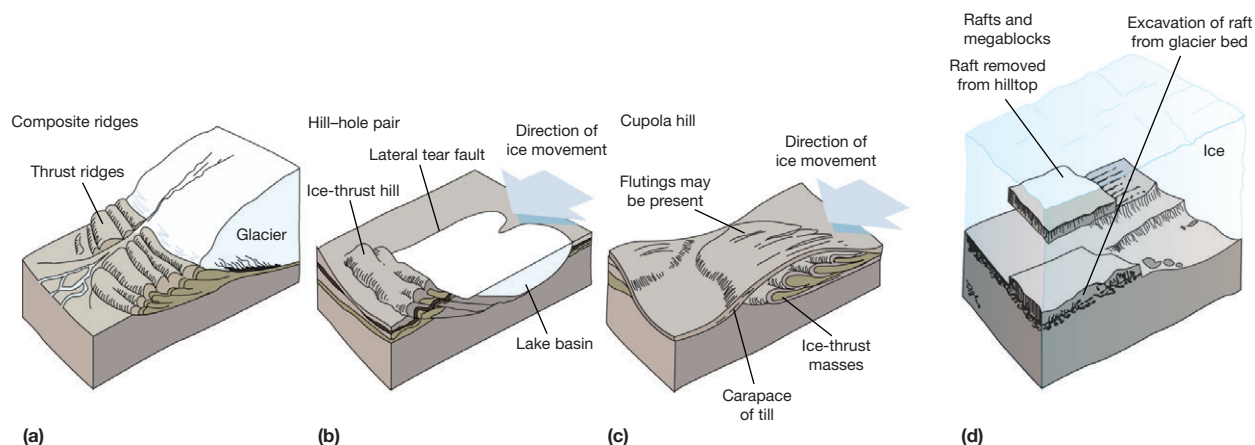


Figure 4 The four main types of glacitectonic landform/structure as defined by Aber et al. (1989). Reproduced from Evans DJA and Benn DI (2001) *Earth's giant bulldozers. Geography Review* 14: 29–33.

surface permafrost and the degraded top of the full glacial permafrost. Thrust-block moraines at the margins of cold-based glaciers in the Dry Valleys area, Antarctica, are interpreted by Fitzsimons (1996) as the product of thrusting by cold-based glaciers overriding weak, unfrozen lake floor sediments. Unfrozen areas occur below either saline lakes or wet-based lakes which have a permanent ice cover but do not undergo full-depth freezing. As the glacier enters the lake, sediment blocks become frozen onto the base of the glacier, but can be thrust forward because their deeper layers remain unfrozen. In summary, therefore, the formation of thrust-block moraines in permafrost regions is thought to occur only within areas where unfrozen or only partially frozen sediments occur.

Hill-Hole Pairs

A hill-hole pair is defined as “a discrete hill of ice-thrust material, often slightly crumpled, situated a short distance down-glacier from a depression of similar size and shape” (Bluemler and Clayton, 1984; Figure 4(b)). Ice-thrust hills may be found at distances of up to 5 km from source depressions, but are usually closer (Bluemler and Clayton, 1984; Figure 6). However, source depressions can become infilled with younger sediment and are not always visible as surface features. Conversely, glacitectonic depressions are sometimes found without associated hills (Andriashek and Fenton, 1989), perhaps because positive relief features have been destroyed by subglacial erosion.

The typical morphology of a hill-hole pair is as follows (Aber et al., 1989): (i) the hill has an arcuate or crescentic planform, which is concave up-glacier; (ii) the surface of the hill is traversed by a series of transverse, subparallel ridges and depressions; (iii) the hill has an asymmetric cross-profile, with the highest point and steeper slopes on the convex or down-glacier side; and (iv) the topographic depression is approximately the same shape and area as the hill, located on the concave or up-glacier side. These morphologic features may be less obvious where prolonged subglacial modification has taken place after formation. The largest hill-hole pair so

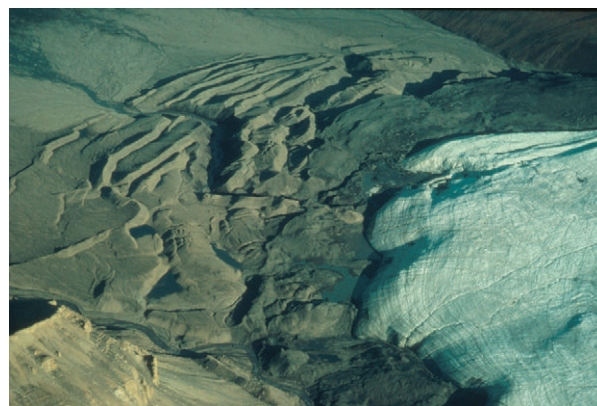


Figure 5 Thrust-block moraine at the margin of a subpolar glacier in the Canadian high Arctic.

far recognized on the Earth's surface is Herschel Island in Mackenzie Bay, Canada, with an area of more than 100 km² (Mackay, 1959; Rampton, 1982) and composed of preglacial Pleistocene sediments thrust up from a now submerged source depression. Its internal structures include low-angle thrust faults, synclines, and anticlines, tilted beds which occasionally approach subvertical angles, overturned folds, repeated strata, slickensides, and segregated ice lenses which originally formed horizontally but now parallel the deformed enclosing strata (Mackay, 1959; Mackay et al., 1972).

Cupola Hills

Cupola hill is the English translation of the Danish term *kupelbakke*, which was introduced by Smed (1962) to describe irregular hills with the general characteristics of glacitectonic landforms but lacking hill-hole pair relationship and/or the transverse ridged morphology of composite ridges (Figure 4). The characteristics of cupola hills can be summarized as (i) they have a dome-like morphology varying from near circular to oval shapes with lengths of 1–15 km and heights of



Figure 6 Bittern Lake, AB, Canada, an example of a hill-hole pair. The glaciotectionally disturbed Cretaceous bedrock that comprises the ice-thrust hill is visible in the road cut.

20–100 m; (ii) internally, they are composed of detached and deformed floes of Quaternary sediments, older strata, or bedrock; and (iii) they have a carapace of till which truncates the underlying structures. These characteristics indicate that cupola hills are subglacially overridden hill-hole pairs or composite ridges and represent early stages in the development of drumlins or megafloodings. In cases where the subglacial modification has been slight, transverse ridge morphology may be partially preserved and the cupola form will tend to lie transverse to former ice flow. With increasing subglacial modification, ridge forms tend to be replaced by more smoothed, elongated forms. Viewed from the air, cupola hills may appear aligned in arcuate or transverse belts which conform to the former margins of glaciers. Where such belts occur, they represent the positions of former large composite ridges that have been modified by subglacial molding.

Examples of cupola hills composed entirely of Quaternary sediments include the island of Ven in Sweden and Ristinge Klint in Denmark (Aber et al., 1989). The island of Ven is composed of gentle synclines disturbed by numerous thrust faults and overturned folds. The glaciotectional disturbance responsible for these structures has produced two major transverse ridges, the surface of which is muted by a cover of discordant till. In contrast, the topography of Ristinge Klint is subdued and drumlinized similar to cupola hills on the nearby island of Aero (Smed, 1962). Cliff sections reveal that the area of Ristinge is underlain by more than 30 stacked scales of Quaternary sediments, each scale being up to 20 m thick and containing the same stratigraphic sequence of Saalian, Eemian, and Weichselian strata (Sjorring, 1983). The internal stratification of each scale is little disturbed except for prominent drag folds associated with the thrust faults between scales. Thrusting of the scales is thought to have occurred at the ice margin during glacier advance and deposition of the upper, discordant tills; surface streamlining took place during subsequent ice overriding. Other excellent examples of cupola hills occur in the late Wisconsinan end moraine systems that make up the offshore islands of New England, USA. The islands consist of glaciotectionally disturbed floes or thrust scales of Cretaceous, Tertiary, and Quaternary strata (Oldale and O'Hara, 1984).

Sections at Gay Head Cliff, Martha's Vineyard expose a series of layered scales, each 20–30 m thick and dipping up-glacier, interrupted by overturned folds, low-angle thrust faults, and occasional down-dropped blocks or grabens. In places, the strata have been uplifted by over 160 m.

Megablocks and Rafts

Megablocks and rafts are dislocated slabs of rock and unconsolidated strata which have been transported from their original position by glacier action (Ruszczynska-Szenajch, 1987; Sauer, 1978; Stalker, 1973, 1976; Figure 4(d)). Traditionally, such rafts were assumed to have been frozen on to the base of cold-based ice sheets, although it is now recognized that failure may occur along a basal decollement in a subglacial deforming layer. Megablocks may originate from uprooted cupola hills, although it is often difficult if not impossible to demonstrate detachment from the parent substrate unless there is exceptionally good exposure. Where they are subhorizontal, megablocks and rafts may form flat-topped buttes or plateaus and can be mistaken for bedrock outliers. Similarly, buried megablocks can be mistaken for bedrock if borehole logs or exposures are not available. Although megablocks and rafts contain coherent masses of bedrock or sediment, they are also commonly traversed by shear zones, faults, and brecciated zones, and may be folded.

Rafts of Jurassic clay more than 20 m thick at Lukow, Poland, are derived from Lithuania, more than 300 km to the northeast, and are thought to have been transported in a frozen state (Ruszczynska-Szenajch, 1987). A similar transport distance has been calculated for the Cooking Lake megablock near Edmonton, AB, Canada, which is 10 m thick and covers an area of approximately 10 km² (Stalker, 1976). A good example of the subsurface expression of a megablock occurs in the vicinity of the Qu'Appelle Valley, Saskatchewan. This block of Cretaceous shale is approximately 1000 km² in area with a maximum thickness of 100 m, and was discovered by borehole logging and bedrock contour mapping (e.g., Sauer, 1978). A well drilled at the western end of the megablock penetrated 80 m of brecciated and deformed bedrock of the Riding Mountain Formation and then 2 m of till. Steeply dipping rafts are more likely to have some surface expression, although they may be buried and difficult to identify without extensive borehole data. A series of detached chalk rafts occur in till overlying limestone in quarry exposures at Kvarnby in Sweden, suggesting that they were removed from the Baltic Sea bed and transported more than 25 km before being stacked at angles of up to 80° (Ringberg, 1983). These subsurface rafts lie beneath a structurally controlled, low-amplitude hummocky topography which would be difficult to differentiate from a cupola hill without exposures. Similarly, chalk rafts in the Anglian till of

Hertfordshire, England, have been identified in borehole records and as pale-toned patches in darker, freshly seeded fields on aerial photographs (Hopson, 1995). Unlike the Swedish examples, the Hertfordshire chalk rafts have been displaced only a short distance (mostly <1 km) from their origin by the southerly flowing Anglian ice sheet.

See also: **Glacial Landforms:** Moraine Forms and Genesis.
Glacial Landforms, Sediments: Tills.

References

- Aber JS (1985) The character of glaciectonism. *Geologie en Mijnbouw* 64: 389–395.
- Aber JS (1988) *Structural Geology Exercises with Glaciectonic Examples*, p. 140. Winston-Salem, NC: Hunter Textbooks Inc.
- Aber JS, Croot DG, and Fenton MM (1989) *Glaciectonic Landforms and Structures*, p. 200. Dordrecht: Kluwer.
- Andriashek LD and Fenton MM (1989) *Quaternary Stratigraphy and Surficial Geology, Sand River Map Sheet 73L*. Alberta: Terrain Sciences Department, Alberta Research Council. Bulletin 57.
- Benn DI and Evans DJA (1998) *Glaciers and Glaciation*. London: Arnold.
- Benn DI and Evans DJA (2010) *Glaciers and Glaciation*, 2nd edn. London: Hodder Education.
- Berthelsen A (1978) The methodology of kineto-stratigraphy as applied to glacial geology. *Bulletin of the Geological Society of Denmark* 27: 25–38.
- Bluemle JP and Clayton L (1984) Large-scale glacial thrusting and related processes in North Dakota. *Boreas* 13: 279–299.
- Croot DG (1987) Glacio-tectonic structures: A mesoscale model of thin-skinned thrust sheets? *Journal of Structural Geology* 9: 797–808.
- Croot DG (1988) Morphological, structural and mechanical analysis of neoglaciated ice-pushed ridges in Iceland. In: Croot DG (ed.) *Glaciectonics: Forms and Processes*, pp. 33–47. Rotterdam: Balkema.
- Evans DJA (1989) The nature of glaciectonic structures and sediments at subpolar glacier margins, northwest Ellesmere Island, Canada. *Geografiska Annaler* 71A: 113–123.
- Evans DJA and England J (1991) Canadian landform examples 19: High arctic thrust block moraines. *The Canadian Geographer* 35: 93–97.
- Evans DJA and Rea BR (2003) Surging glacier landsystem. In: Evans DJA (ed.) *Glacial Landsystems*. London: Arnold.
- Fitzsimons SJ (1996) Formation of thrust-block moraines at the margins of dry-based glaciers, south Victoria Land, Antarctica. *Annals of Glaciology* 22: 68–74.
- Gripp K (1929) Glaciologische und geologische Ergebnisse der Hamburgischen Spitzbergen-Expedition 1927. *Naturwissenschaftlichen Verein in Hamburg Abhandlungen aus dem Gebiete der Naturwissenschaften* 22: 146–249.
- Gripp K and Todtmann EM (1925) Die Endmorane des Green-Bay Gletschers auf Spitzbergen, eine studie zum Verstandnis norddeutscher Diluvial-Gebilde. *Mitteilungen aus Geographischen Gesellschaft in Hamburg* 37: 45–75.
- Gry H (1940) De istektoniske forhold i moleromraadet. *Meddelelser Dansk Geologisk Forening* 9: 586–627.
- Hopson PM (1995) Chalk rafts in Anglian till in north Hertfordshire. *Proceedings of the Geologists' Association* 106: 151–158.
- Kalin M (1971) The active push moraine of the Thompson Glacier, Axel Heiberg Island, Canadian Arctic Archipelago, p. 68. Montreal: McGill University McGill University, Axel Heiberg Island, Research Report, Glaciology 4.
- Klassen RA (1982) Glaciectonic thrust plates, Bylot Island, District of Franklin. *Geological Survey of Canada, Paper* 82-1A: 369–373.
- Kupsch WO (1962) Ice-thrust ridges in western Canada. *Journal of Geology* 70: 582–594.
- Mackay JR (1959) Glacier ice-thrust features of the Yukon coast. *Geographical Bulletin* 13: 5–21.
- Mackay JR, Rampton VN, and Fyles JG (1972) Relic Pleistocene permafrost, western Arctic, Canada. *Science* 176: 1321–1323.
- Mathews WW and Mackay JR (1960) Deformation of soils by glacier ice and the influence of pore pressures and permafrost. *Transactions of the Royal Society of Canada* 54: 27–36, Section 3.
- Oldale RN and O'Hara CJ (1984) Glaciectonic origin of the Massachusetts coastal end moraines and a fluctuating late Wisconsinan ice margin. *Bulletin of the Geological Society of America* 95: 61–74.
- Phillips ER, Evans DJA, and Auton CA (2002) Polyphase deformation at an oscillating ice margin following the Loch Lomond Readvance, central Scotland, UK. *Sedimentary Geology* 149: 157–182.
- Phillips ER, Lee JR, and Burke H (2008) Progressive proglacial to subglacial deformation and syntectonic sedimentation at the margins of the mid-Pleistocene British Ice Sheet: Evidence from north Norfolk, UK. *Quaternary Science Reviews* 27: 1848–1871.
- Phillips ER, Merritt J, Auton CA, and Golledge N (2007) Microstructures in subglacial and proglacial sediments: Understanding faults, folds and fabrics, and the influence of water on the style of deformation. *Quaternary Science Reviews* 26: 1499–1528.
- Price NJ and Cosgrove JW (1990) *Analysis of Geological Structures*, p. 502. Cambridge: Cambridge University Press.
- Rampton VN (1982) Quaternary geology of the Yukon Coastal Plain. *Geological Survey of Canada, Bulletin* 317: 49.
- Rijsdijk KF, Owen G, Warren WP, McCarroll D, and van der Meer JJM (1999) Clastic dykes in over-consolidated tills: Evidence for subglacial hydrofracturing at Killiney Bay, eastern Ireland. *Sedimentary Geology* 129: 111–126.
- Ringberg B (1983) Till stratigraphy and glacial rafts of chalk at Kvarnby, southern Sweden. In: Ehlers J (ed.) *Glacial Deposits in North-West Europe*, pp. 151–154. Rotterdam: Balkema.
- Rotnicki K (1976) The theoretical basis for and a model of glaciectonic deformations. *Quaestiones Geographicae* 3: 103–139.
- Ruszczynska-Szenajch H (1987) The origin of glacial rafts: Detachment, transport, deposition. *Boreas* 16: 101–112.
- Sauer EK (1978) The engineering significance of glacier ice thrusting. *Canadian Geotechnical Journal* 15: 457–472.
- Sharp MJ (1985) Sedimentation and stratigraphy at Eyjabakkajökull – An Icelandic surging glacier. *Quaternary Research* 24: 268–284.
- Sjorring S (1983) Ristinge Klint. In: Ehlers J (ed.) *Glacial Deposits in North-West Europe*, pp. 219–226. Rotterdam: Balkema.
- Smed P (1962) Studier over den fynske ogrundens glacielle landskabsformer. *Meddelelser Dansk Geologisk Forening* 15: 1–74.
- Stalker A MacS (1973) The large interdrift bedrock blocks of the Canadian Prairies. *Geological Survey of Canada, Paper* 75-1A: 421–422.
- Stalker A MacS (1976) Megablocks, or the enormous erratics of the Albertan Prairies. *Geological Survey of Canada, Paper* 76-1C: 185–188.
- Twiss RJ and Moores EM (1992) *Structural Geology*. New York: Freeman.
- van der Wateren DFM (1985) A model of glacial tectonics, applied to the ice-pushed ridges in the central Netherlands. *Bulletin of the Geological Society of Denmark* 34: 55–74.
- van der Wateren DFM (1995) Processes of glaciectonism. In: Menzies J (ed.) *Glacial Environments, Vol. 1 – Modern Glacial Environments: Processes, Dynamics and Sediments*, pp. 309–335. Oxford: Butterworth-Heinemann.
- Williams GD, Brabham PJ, Eaton GP, and Harris C (2001) Late Devensian glaciectonic deformation at St Bees, Cumbria: A critical wedge model. *Journal of the Geological Society of London* 158: 125–135.

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GLACIAL LANDFORMS, EROSIONAL FEATURES

Contents

Major Scale Forms

Micro- to Macroscale Forms

Major Scale Forms

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Introduction

Forms of glacial erosion are both the most obvious of landforms, recognized after a few classes in school, and yet the most difficult to fully explain and to date (Glasser and Bennett, 2004). Medium-sized glacial landforms, such as roches moutonnées, whalebacks, and rock drumlins are discussed in another article in this encyclopedia. Most of these are decametric and hectometric in scale but some, such as giant roches moutonnées, are kilometric. Here, we deal with the largest forms: cirques, troughs, rock basins, fjords, and landscapes of glacial erosion. These are on kilometric and larger scales; the smallest glacial cirques are on the order of 200 m in length and breadth.

Cirques show the former presence of cirque glaciers or the source areas of valley glaciers. Where ice from several cirques converged, forming a valley glacier, a trough was generally eroded. Further glacier growth in areas of high relief produced systems of tributary troughs hanging above major troughs, often with the development of rock basins. Ice cap or ice sheet development led to more complex systems of branching troughs, some formed by ice diffuent across a ridge from a valley with more ice to one with less (Linton, 1963). Ice streams further enlarged major troughs oriented in a direction permitting ice flow to lower or marine areas.

Further references can be found in Benn and Evans (2010), Evans (1997) and Glasser and Bennett (2004). Embleton (1972) and Evans (2008) discuss the older literature.

Glacial Cirques

Processes

Glacial cirques are concave landforms formed at the sources of mountain glaciers. They are doubly concave hollows, open downstream but bounded upstream by the convex crest of a steep headwall (Evans and Cox, 1974). The headwall is arcuate in plan around a more gently sloping floor that should show the effects of glacial erosion, if not buried under moraine and postglacial accumulations. The word is related to 'circus,' but

cirques must be open at one side. Their presence shows that a mountain range rose above the snowline long enough for small glaciers to form and to erode their beds. Cirques vary greatly in their degree of development, and it is useful to recognize five grades (Figure 1).

As rates of glacial erosion are believed to increase with basal shear stress or velocity, and thus with ice thickness and surface slope (Hallet et al., 1996), it may come as a surprise that such striking forms are produced at glacier sources, where discharge and basal velocity are low. The 'problem of cirque erosion' is to explain the concentration of erosion at the foot of the headwall. This is necessary to convert a nonglacial valley head into a cirque, and to produce a cliffed headwall. Periglacial and ice-marginal processes such as freeze-thaw cannot achieve this; they would cause retreat of the upper part, reducing headwall gradient.

Factors contributing to this explanation include duration of glacier cover, positive feedback on glacier balance, rotational flow and water pressure variations. As the snowline falls over a mountain range, the first glaciers will form in high cirques. And as glaciers waste away, those in sheltered, shady cirques are likely to survive the longest (Figure 2). The more deeply a cirque is eroded, the greater the shelter and (if pole facing) the shade provided, increasing the trapping of wind-blown snow and reducing the rate of melting. The increased mass balance gives a larger, thicker glacier, which may erode more effectively enlarging the cirque further and providing positive feedback. Extra input of snow at the top of a glacier, by wind drifting or avalanching, exaggerates the wedge of net accumulation and is effective in increasing velocity and tendency to rotate.

Cirque glaciers with gradients of roughly 10°–30° tend to flow in a rotational manner, as a wedge of net accumulation in the upper half flows to replace a wedge of net ablation in the lower. In warm-based ice, this maximizes basal sliding and concentrates erosion at the foot of the headwall, deepening the cirque floor and helping develop a closed basin (Lewis, 1960). This depends on ice thickness and surface gradient, and is not found in the Canadian glacier studied by Sanders et al. (2010). Many cirque glaciers are polythermal, and the

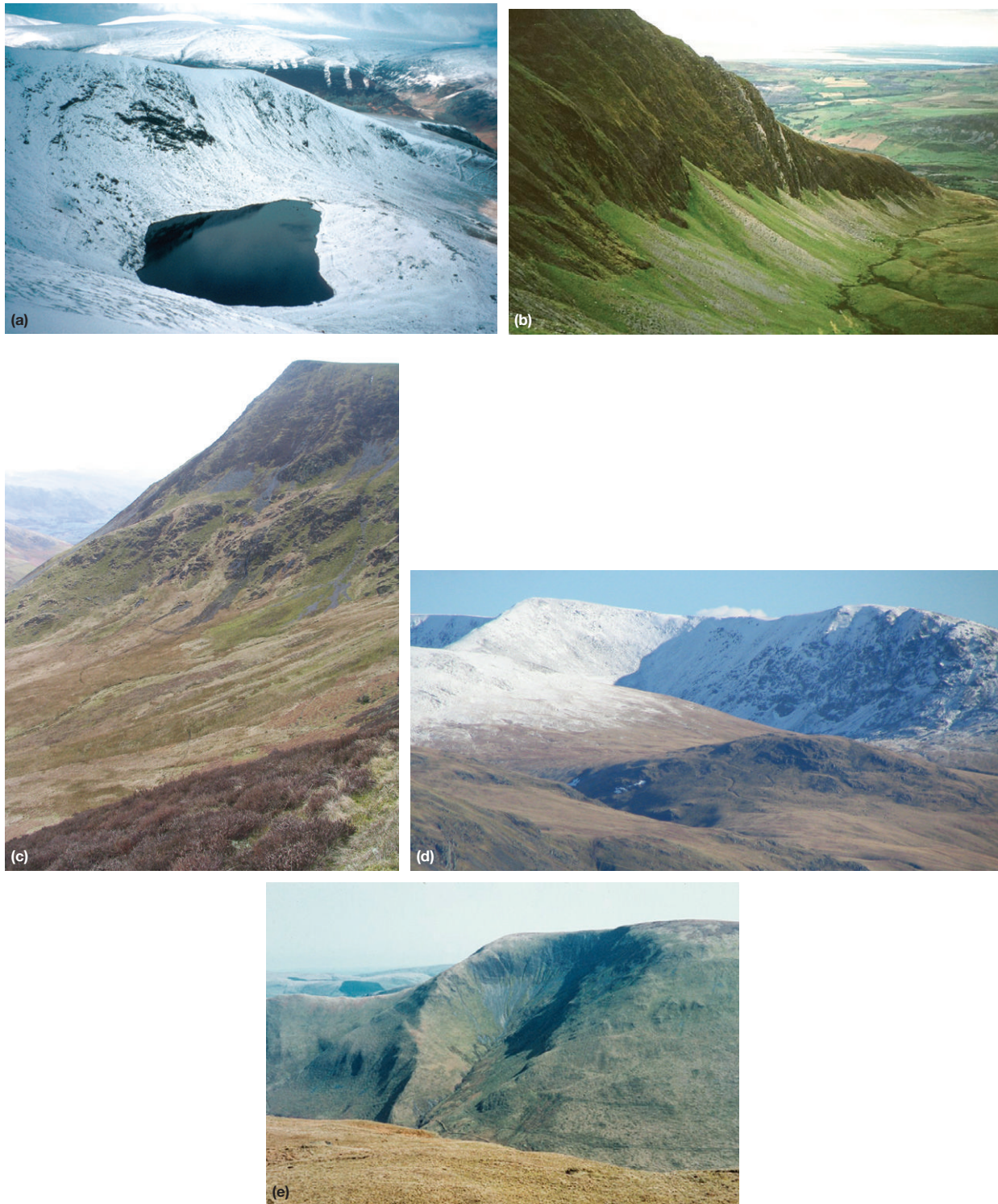


Figure 1 Cirques of different grades. (a) Bowscale Tarn, a classic (grade 1) cirque in the northern Lake District, England. The headwall consists of slates with varying degrees of metamorphism: the lake is up to 16.5 m deep. The unusually large moraine (bottom right) was produced by a Loch Lomond Stadial (Younger Dryas) glacier with an equilibrium line altitude (ELA) of about 536 m and a surface slope of 17°; it was up to 95 m thick. (b) Craig Trum y Ddysgl, the west side of a well-defined (grade 2) cirque in tuffaceous Ordovician sandstones above the Nantlle, North Wales. (c) Lonscale Cove, a definite (grade 3) cirque in slates east of Skiddaw in the northern Lake District, England. (d) Brown Cove Crags, a poor (grade 4) cirque in Ordovician volcanic rocks on the northwest slopes of Helvellyn, eastern Lake District, England. (e) Craig Rhiw-erch, a cirque in tuffaceous Ordovician siltstones, on Maesglase in the upper Dyfi basin, Mid Wales. It is marginal (grade 5) because of its steep (although drift-filled) floor (photographs by I.S. Evans).



Figure 2 Coalescent north-facing cirques with remnant Little Ice Age (LIA) glaciers at the head of Truax Creek, Bendor Range, British Columbia Coast Mountains. The three summits are at 2466, 2517 and 2589 m altitude: rocks are thermally metamorphosed mudrocks (photograph by I.S. Evans).

boundary between ice frozen to the bed and that which is sliding gives possible instabilities that facilitate quarrying (Richardson and Holmlund, 1996). This is most effective in sedimentary or well-jointed rock, in which thrust sheets may develop (Bennett et al., 1999). Continued erosion where ice is thickest, while peripheral ice is frozen to the headwall, may be a further factor undercutting the headwall. It is, however, limited to colder climates (continental and polar), as cold ice is not found near equilibrium line altitude (ELA) in snowy, maritime climates (see below).

Finally, the bergschrund and other crevasses formed as ice moves away from the headwall permit the periodic entry of meltwater. This causes variation in water pressure in subglacial pockets, facilitating the extension and opening of joints and the quarrying (plucking) of blocks of rock. This water pressure mechanism (Hooke, 1991; Iverson, 1991) has replaced Lewis' meltwater and freeze-thaw hypotheses, as temperatures beneath warm ice vary very little. Although abrasion is sensitive to basal velocity, quarrying is not so directly affected and is important on lee slopes under shallow ice flowing over bumps.

Rapid glacial erosion near the sources of mountain glaciers is thus explicable, and has the further effect of steepening the headwalls. This causes a threshold of stability to be exceeded, and the headwall retreats by episodic collapse in rock avalanches (Evans, 1997; McSaveney, 2002). These may occur when the glacier is active, but by removing support ('debudding' the slope) glacier recession makes them more likely. Again we note that frost action is not necessary; this is simply a slope stability problem, with slope gradient and height dependent on rock mass strength (RMS) (Augustinus, 1992). Figure 3 shows a cirque floor covered by two rock avalanche deposits, the results of headwall collapse in ultrabasic rocks. Headwall collapse is a normal part of cirque erosion, as basal glacial erosion steepens them beyond their maximum stable angle and height.

As many mountain areas have undergone tens of major glacial episodes in the Quaternary (and late Pliocene), considerable time is available for cirques to develop. Larsen and Mangerud (1981) studied deposits from a small cirque glacier on resistant mica-rich gneiss in western Norway. It eroded



Figure 3 Rock avalanche covering a cirque floor in the Shulaps Range, British Columbia Coast Mountains. The outer deposit (left) is covered by the 2500 BP. Bridge River Ash; the inner deposit is younger. The rock is serpentinised peridotite (harzburgite) (photograph by I.S. Evans).

118 000 m³ during 700 a of the Younger Dryas, equivalent to vertical lowering of the 0.21 km² cirque by 0.8 mm a⁻¹. At this modest rate, 200 m would be eroded in 250 ka of glaciation, which is sufficient to account for development of most cirques. Glaciation must, however, be of the appropriate style; ice should be wet-based, at least in part, and with steep surface gradients in cirques. The rapidity of climatic changes suggests that a prolonged phase of snowpatch activity (nivation) is unlikely before glaciers form. The nivation process-complex erodes more slowly than do warm-based glaciers; its importance has often been exaggerated, as has freeze-thaw action.

Distribution and Development

Spatially, glacial ELAs currently vary by 6000 m globally, by 600–2000 m between wet and dry sides of major mountain ranges, and by several hundred meters locally (Evans, 2006a). Glacial cirques likewise occur at lower altitudes on the snowy, windward sides of mid-latitude mountains. Globally, cirque floors are found down to an altitude below, and roughly paralleling, present ELA. As the zero (or other) isotherms follow a different pattern, both these observations relate cirques to glacial erosion rather than to frost action or any other factor directly related to atmospheric temperature. On a local scale, cirque glaciers and thus cirques favor shady and leeward slopes. In most areas they face poleward (Figure 4), and in the west-wind belt, their mean aspect is deflected toward the east, for example, in Europe, Tasmania, the Falklands, and the American Rockies. Thus, the distribution of cirques provides valuable evidence of former climates, including winds and sunniness as well as ELA (Mindrescu et al., 2010).

All of these spatial relationships lead to the rejection of a proposal by Turnbull and Davies (2006) that large cirques in general are formed essentially by deep-seated, often coseismic rock slope failures. The scars of such failures do indeed provide excellent locations for glacier initiation, but the distribution of such rock avalanche scars is very different from that of cirques, and sometimes complementary. The feasibility of mapping a 'paleo-glaciation level' separating mountains above, with cirques, from mountains below, without cirques, and the way its

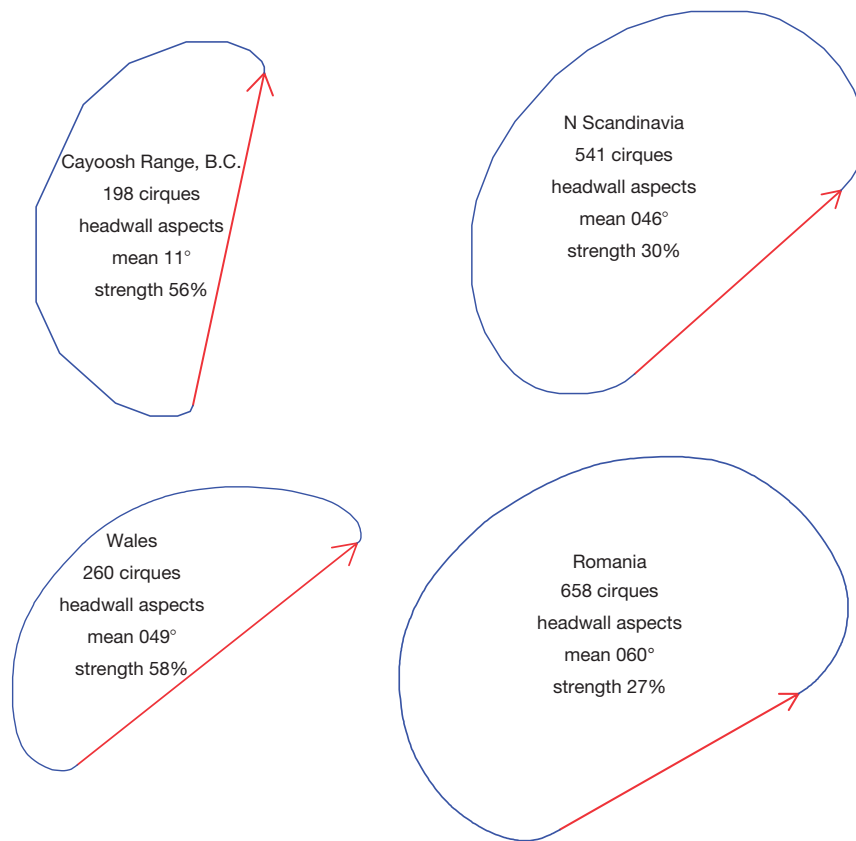


Figure 4 Vector plots of cirque aspect. The resultant vector is in red, the individual aspects in blue, plotted in cumulated vector form. Scales vary. The Cayoosh Range, with a resultant a little east of north, is typical of fairly dry mountains of high relief, as in Central Asia and parts of British Columbia. The other three plots are all from the west-wind belt in Europe and show eastward tendencies as strong as northward, and stronger in Romania. Data for northern Scandinavia kindly supplied by S. Hassinen; for Romania, by M. Mindrescu (graphs by I.S. Evans).

altitude varies with precipitation, supports a strong climatic control. The relation of degree of cirque development to length and intensity of exposure to glaciations is also important.

Techniques for defining and measuring cirques are presented by [Evans and Cox \(1995\)](#) and, using Geographic Information Systems (GIS), by [Federici and Spagnolo \(2004\)](#). Where cirques are eroded into a plateau, their outlines are broadly rounded and the headwall crest approaches a semi-circular outline. Classically simple forms are found in horizontally bedded rocks such as basalts ([Figure 5](#)) or well-bedded sandstones. In sharp-crested mountains coalescence of cirques gives more angular outlines ([Figure 2](#)). Cirques are scale-specific forms, commonly between 200 m and 4 km wide and long, with the upper limit of width dictated by valley spacing. They are rarely over 2 km in Britain; larger cirques found in high-mountain areas tend to be complex, with smaller cirques nested within. Cirques around Mount Everest reach widths of 4 km, and some alpine valley heads in the Antarctic ([Haynes, 1998](#)) are over 8 km wide and long. These reflect much longer periods of alpine glaciation during the last 36 Ma. After deep glacial erosion, high-alpine cirque forms develop around trough heads ([Figure 6](#); see also [Figure 2](#) in [Evans, 1997](#)).

Cirque development involves flattening of the floor (eventually developing a reversed slope), steepening of the headwall, increased closure in plan as the headwall recedes more deeply



Figure 5 A trough-side cirque eroded into the basalt plateau above Isafjordur, northwest Iceland (photograph by D.J.A. Evans).

into the mountain mass, and increased length, width, and depth. As cirques develop, as expressed by their overall size, it appears that they increase in length faster than in width, and in depth least of all ([Evans, 2006b](#); [Federici and Spagnolo, 2004](#); [Gordon, 1977](#); [Figure 7](#)). Cirque development is thus allometric rather than isometric. For part of the Tien Shan, [Oskin and Burbank \(2005\)](#) have invoked cirque headwall

retreat two to three times faster than downward erosion. On the other hand, in the Ben Ohau Range, New Zealand, cirques seem to lengthen and deepen faster than they widen (Brook et al., 2006).

Haynes (1998) invoked long-term branching and coalescence keeping width and length in balance, so that alpine valley heads attain an equilibrium shape in plan: continual addition of new small cirques maintains positively skewed frequency distributions as older cirques widen and lengthen. Attainment of a steady state, however, requires downward erosion to match headwall retreat not only in cirques but in

their surroundings, including any glacial troughs below. This is possible only after cirque coalescence has eliminated gently-sloping higher surfaces, producing a landscape of arêtes and horns. Arêtes are sharp ridges produced initially between cirques side-by-side on the glacially favored side of a ridge. Those between cirques back-to-back are more likely to produce deep cols, but this requires glaciation that is more symmetrical between slopes of different aspects. Locally symmetrical glaciation (Evans, 2006a) is found either where ELA is well below ridge altitudes or where the climate is neither sunny nor very windy, and in theory it may permit steady state erosion once all headwalls are coalescent. In general, however, this situation has not been reached and the form of cirques changes (flattens) as they enlarge: they are not equilibrium forms. The factors discussed above permit quite rapid cirque erosion both downward and headward, and this is an important part of the 'glacial buzzsaw' (discussed below).



Figure 6 High-alpine cirques around a trough-head south of Tzoonie Mountain, 66 km NNW of Vancouver, British Columbia (photograph by I.S. Evans).

Glacial Troughs

Calibration

Popularly termed 'U-shaped valleys,' glacial troughs are the channels of present or former glaciers or ice streams. Like rivers, glaciers obey the necessary geometric relationship:

$$Q = Av \cong wdv$$

Discharge equals mean velocity (v) times cross-sectional area (A); the latter is approximately width (w) times depth (d).

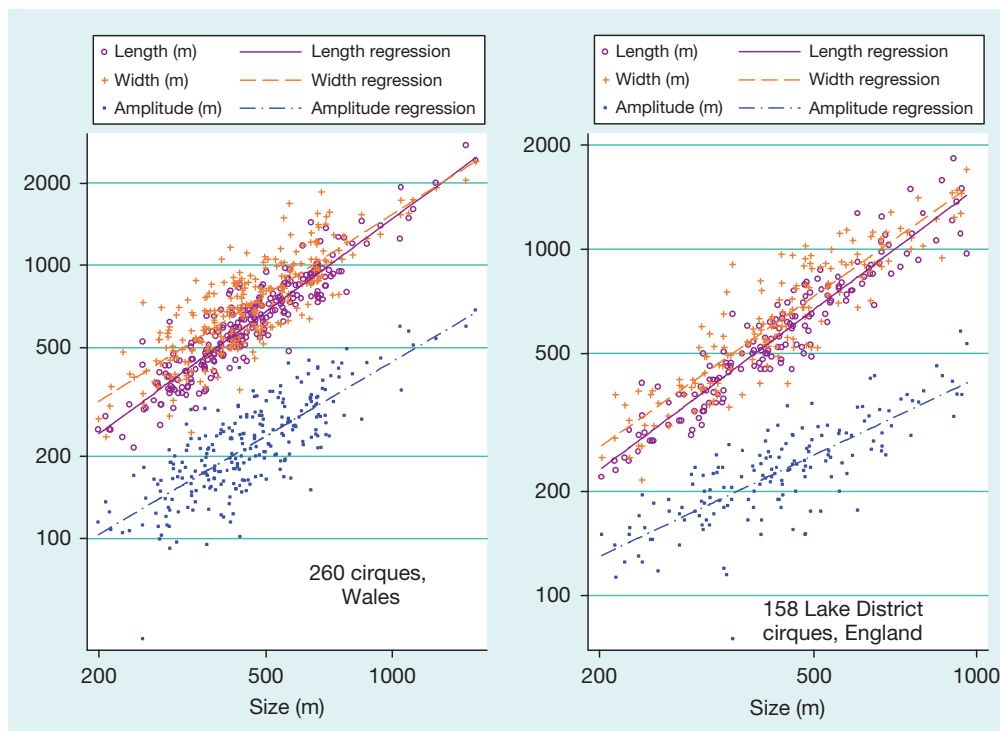


Figure 7 Allometric plots of length, width and height (amplitude) against overall size, for cirques in Wales and the English Lake District (scales differ and are logarithmic). In both regions the gradient for amplitude is significantly less than those for length and width (graphs and data by I.S. Evans).

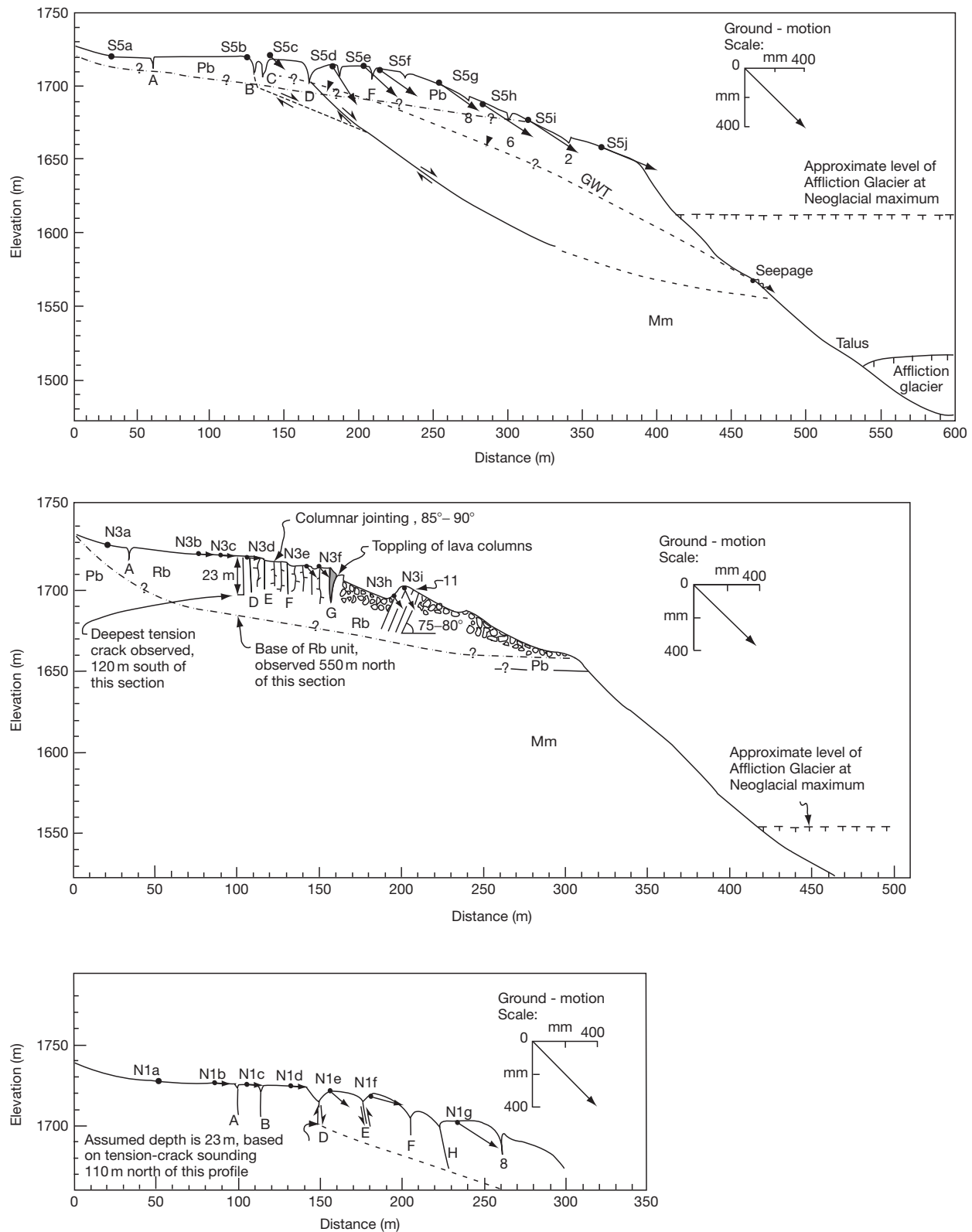


Figure 9 Debuttressing of a trough side in the Mount Meager massif by post-LIA glacier recession. Arrows are ground motion vectors between 1982 and 1986. S5, N3, and N1 are three survey lines. GWT, Ground water table; Mm, Miocene monzonite; Pb, Pleistocene basalt; Rb, recent basalt. Reproduced from Bovis MJ (1990) Rock-slope deformation at Affliction Creek, southern Coast Mountains, British Columbia. *Canadian Journal of Earth Sciences* 27: 243–254.



Figure 10 (a, b) Two views of the trough of Keary Creek, Bendor Range, British Columbia Coast Mountains. The foreground (of both photos) is on granodiorite; beyond the lake, the slopes are on thermally metamorphosed mudrocks. The upper slopes are gullied; talus covers those below a mid-slope trimline. These are typical trough sides: Keary Lake is a ribbon lake in a rock basin (photographs by I.S. Evans).

influenced by geology (e.g., joint spacing) and by the initial, preglacial profile. Initial variations in gradient are exaggerated by a positive feedback, basins are excavated in flatter sections, and the height of steps is increased. Steps are exaggerated because fluctuations in water pressure on lee slopes aid quarrying. Many steps (riegels, verrous) relate to zones of more resistant (usually less jointed) rock, but others relate to position within the system of glacial channels; as troughs tend

to be calibrated to ice discharge, steps are also expected at the confluences of major tributaries. Where the troughs are glacially calibrated, the degree of 'hang' relates to the difference in ice discharge: large tributaries hang less than small (Figure 13). This development of 'hanging valleys' has been simulated by MacGregor et al. (2000) (Figure 14). When simulating long profiles, erosion is proportional either to basal velocity or to its rate of change, depending whether erosion is 'production-limited' or 'transport-limited.'

MacGregor et al. (2009) simulated long profile erosion over 400 ka by a glacier some 20 km long, separating abrasion and quarrying in some runs. Abrasion was varied with clast concentration and angularity and with the square of basal sliding velocity. Quarrying was varied with bed gradient and sliding velocity. Long-term erosion lowered the glacier, reducing its mass balance and thus its length. The more realistic versions (with temperature varying over glacial cycles) did not develop closed basins in the upper reaches. Headwalls were not realistically steep, perhaps because frost action above the glacier was overemphasized compared with subglacial erosion by quarrying. The latter is essential to headwall development, but may be aided by rapid uplift (MacGregor et al., 2009).

Rock Basins and Fjords

The glacier bed can be irregular, with reversed slopes enclosing rock basins. The glacier surface must slope continuously down-valley to provide the necessary potential energy for flow (via shear stress). Of course, basins are found also in river beds, but on a much smaller scale. Glacier flow is approximately laminar whereas most rivers are turbulent; velocities are lower for glaciers than for rivers but pressures and stresses are higher. Glaciers are able to transport large blocks of rock up reversed bed slopes and are better placed than rivers to excavate fractured material from their beds: there is no limit to their competence.

There are three exceptions to the widely accepted view that glacial erosion is greatest around the EL. First, in multi-branched valley glacier systems, maximum discharge and erosion may be below confluences, some way below EL. Second, emergent flow in the lower ablation zone (to replace surface ice lost by melting, Figure 15) carries material upward and tends to excavate a rock basin near the terminus, so the lower ablation zone may be the locus of maximum erosion. This is most evident for tidewater glaciers (Figure 15), where flow (and basal velocity) accelerates toward the calving terminus and exceeds that at the EL. Third, under a polythermal ice sheet or ice cap, erosion is concentrated in warm-based areas with faster sliding: these are in the lower reaches of valleys, well below EL (Figure 6 in Jamieson et al., 2008).

Rock basins thus relate to position, and not only to lithology or structure. Rock basins are larger and deeper in the west of Scotland than in east, because of greater ice discharge due to heavier snowfall. Around the Alps it is notable how many deep lake basins lie just within the terminal moraines of major Pleistocene glaciers.

Modern seismic work on sediments in large lakes (mainly valley rock basins) has shown great thicknesses of glacial, glaciofluvial and postglacial sediment. The rock basins are

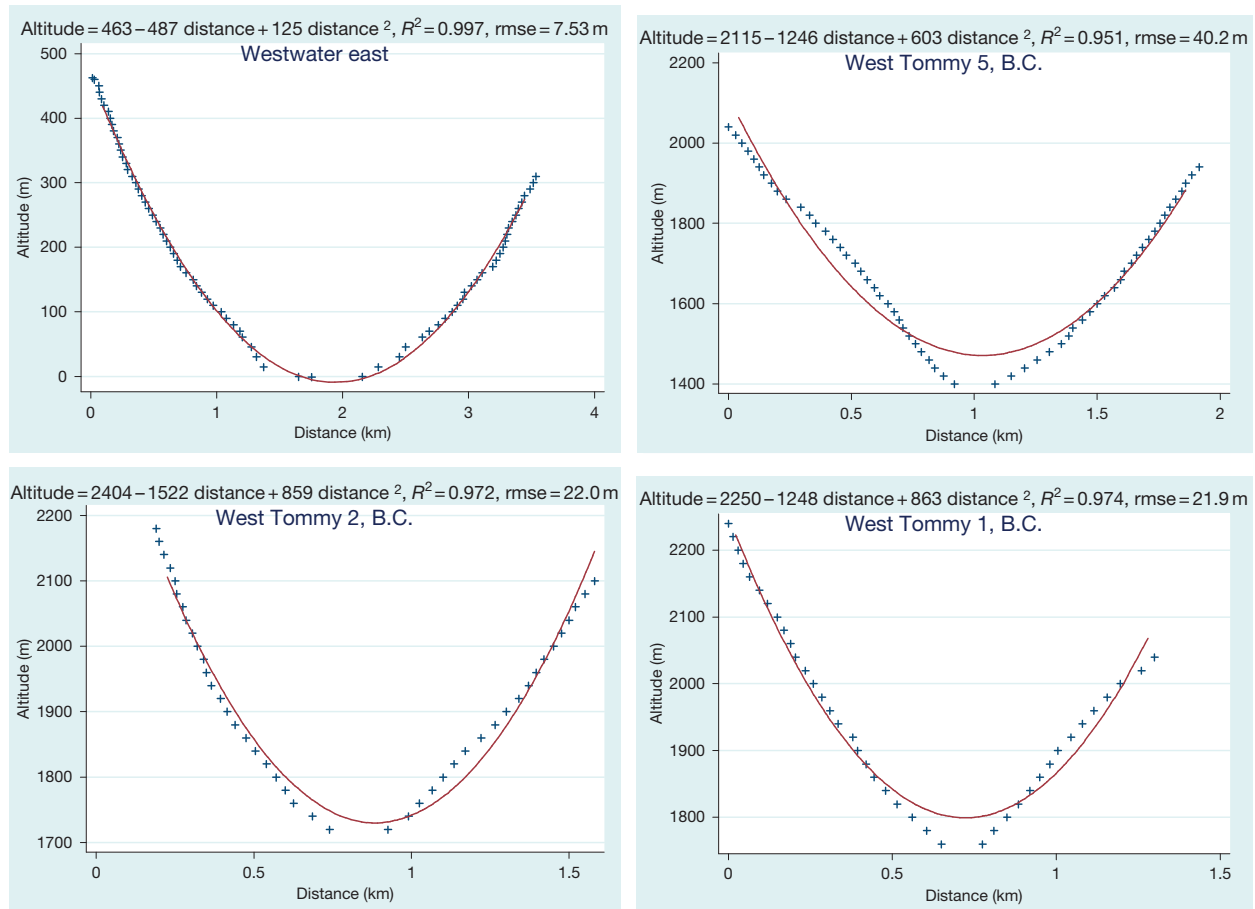


Figure 11 Fitting quadratic curves to a glacial trough cross-section. Each point plots the intersection of a contour with the profile. Note that an R^2 of 0.95 is obtained for a poor fit (top right), and even an R^2 of 0.97 (bottom two) does not guarantee a good fit, though 0.997 (top left) does. Quadratic curves are more robust than power curves (for which a complex iterative procedure is recommended) (graphs by I.S. Evans).

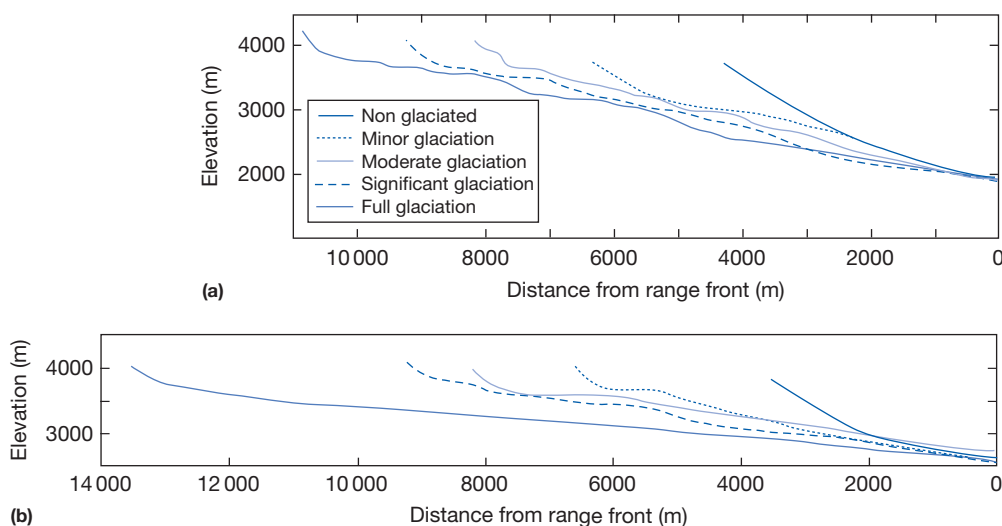


Figure 12 Long profiles of streams on the east slope of (a) the Sierra Nevada, California, and (b) the Sangre de Cristo Mountains, Colorado: no vertical exaggeration. These show the effects of glacial erosion in deepening and, especially, extending valleys, in areas of marginal glaciation (valley glaciers). Reproduced from Brocklehurst SH and Whipple KX (2006) Assessing the relative efficiency of fluvial and glacial erosion through simulation of fluvial landscapes. *Geomorphology* 75: 283–299.

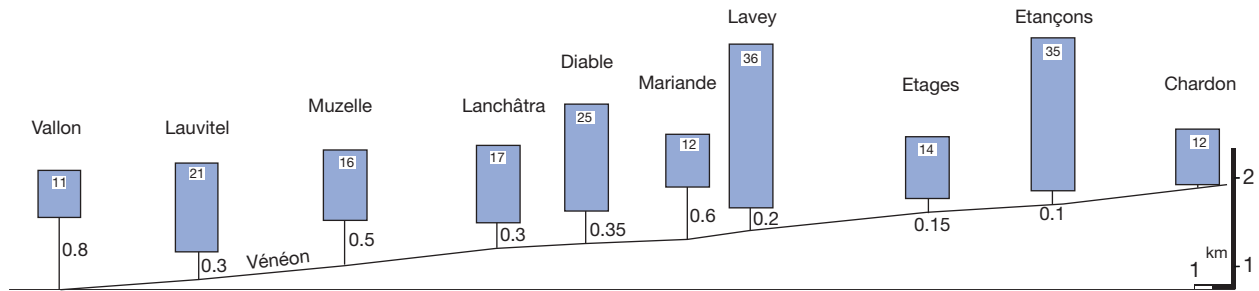


Figure 13 On uniform granitic rocks, large tributaries of the Vénéon (Oisans, French Alps) hang less than small. Shaded bars are proportional to tributary catchment areas in km²; numbers below give step height in km. Step height also increases downstream as the trunk catchment area increases. Reproduced from Monjuvent G (1974) *Considérations sur le relief glaciaire a propos des Alpes de Dauphiné. Revue de Géographie Physique et de Géologie Dynamique, Série 2*, 16: 466–502.

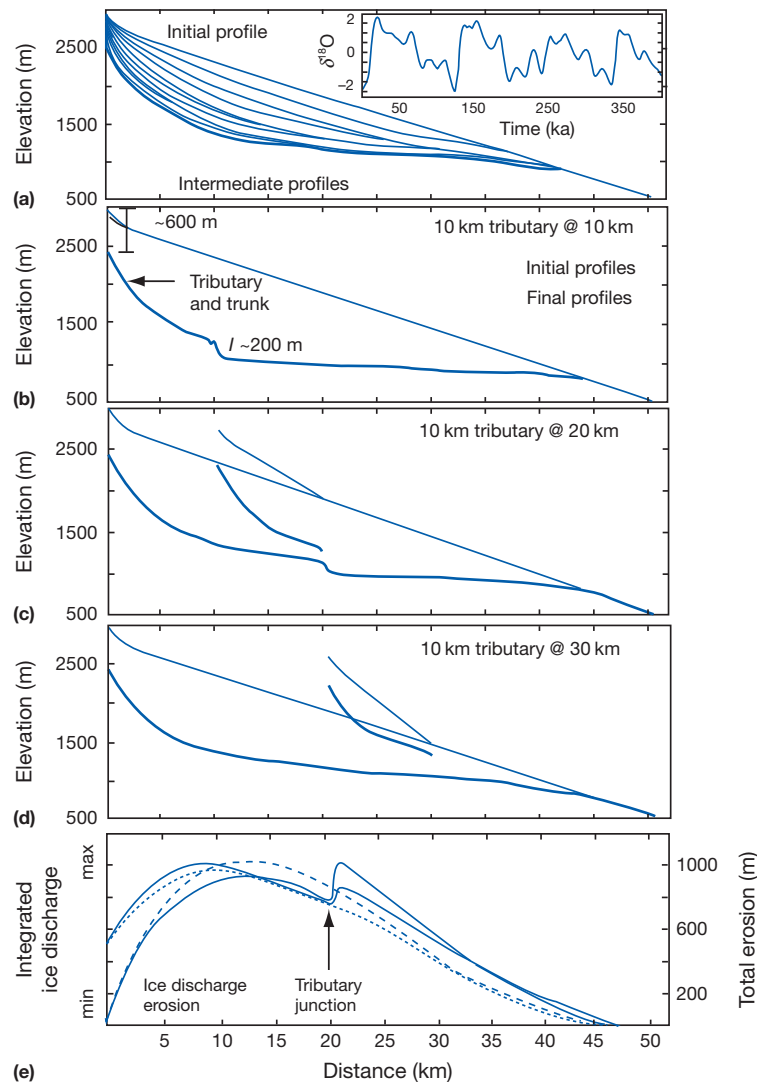


Figure 14 Simulation of long profiles, with glaciers varying over the last 400 ka of environmental conditions and erosion proportional to basal sliding velocity. (a) Single glacier with no tributary; profiles at 40 ka intervals. INSET marine $\delta^{18}\text{O}$ curve used to simulate temperature. (b) Initial and final profiles with tributary joining at 10 km from sources, that is, equal to trunk; both 'hang' by 200 m. (c) Tributary at 20 km; a 100-m step is generated, with the tributary hanging by a further 100 m, and a shallow rock basin below. (d) Tributary at 30 km hangs by 200 m and trunk profile is little affected. (e) Long-term ice discharge (top line) and total erosion (bottom line) in trunk valley, for tributary at 20 km (solid lines) and for no tributary (dashed lines). Reproduced from MacGregor KR, Anderson RS, Anderson SP, and Waddington ED (2000) Numerical simulations of glacial-valley longitudinal profile evolution. *Geology* 28: 1031–1034.

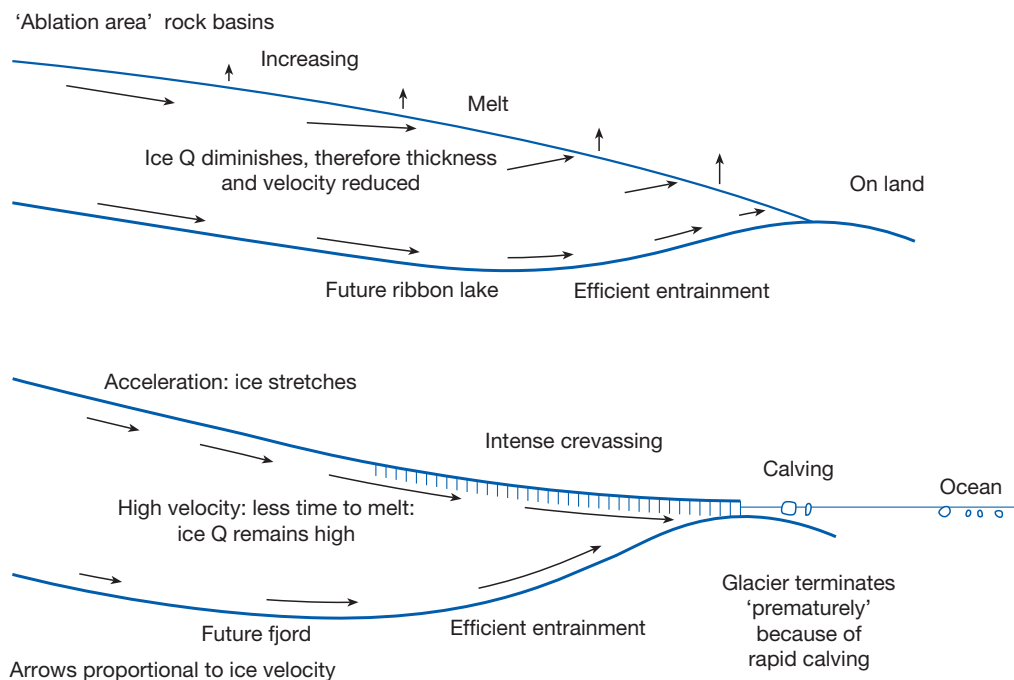


Figure 15 Erosion of 'ablation area' rock basins; similarities and differences between land- and water-terminating glaciers, producing ribbon lakes and fjords, respectively. The upward flow tendency of a normal ablation area aids entrainment and excavation of a rock basin as the glacier thins toward the snout. This is enhanced in a tidewater glacier as it accelerates toward its calving front (sketch by I.S. Evans).

even deeper than formerly thought (often twice as deep as the water: [Figure 16](#)), and they are often dammed by massive rock thresholds, strengthening the case for glacial excavation. However, some of the deeper sediment is pre-last glaciation, showing that deepest excavation was by larger, earlier glaciers.

Fjords ([Figures 5 and 17](#)) are spectacular troughs, often with deep rock basins ([Holtedahl, 1967](#)), and shallow water near their mouths ([Figure 18](#)). Their thresholds are often where ice flow can diverge, at the edge of the mountains. Flotation and calving of ice reduce discharge more abruptly than does surface melting, and provide an extra explanation for the reversed slope. Never the less, in Scandinavia, British Columbia, Patagonia and New Zealand, the west coast fjords are almost matched by long ribbon lakes on the landward slope, often extending beyond the mountains: these must have formed largely under ablation zones, initially without calving.

Fjords are not tectonic landforms due to earth movements, but they may be structural landforms due to erosion exploiting fracture systems ([Figure 19](#); [Holtedahl, 1967](#)). Thus, fjords sometimes follow geological structures, but they often cut across contrasting rock types. On crystalline rocks around 72° N in East Greenland, ice-sheet erosion over the last 7 Ma deepened troughs and produced overdeepened basins and some interconnections, but the sinuosity of preglacial fluvial valleys was retained: low-gradient high areas eroded very little. On sedimentary rocks toward the coast, troughs were widened by ice streams and have smoother long profiles ([Swift et al., 2008](#)).

Classic fjords are transverse to coasts, but remarkable longitudinal fjords are found in southeast Alaska, British Columbia, and southern Chile. These reflect diffluent flow of ice, probably eroding lines of structural weakness. Longitudinal fjords may be characteristic of orogenic coasts (active

margins): they are largely absent from the passive margins around the North Atlantic.

Ages and Rates

Dating Forms

Some troughs appear to have developed solely by glacial erosion, but most have developed by the deepening and widening of preglacial valleys or graben. It is difficult to date large erosional forms, as most are modified over a series of glaciations. Allowing for duration, periods of greatest basal ice velocity will be most influential unless bedrock is protected by a cover of low-strength material. Discharge has varied greatly throughout the Quaternary, so the timing of formative discharge is an important question. Some troughs and basins, for example, around the periphery of the Alps, could develop only around times of glacial maxima. Others relate to intermediate, 'average Quaternary' conditions of much-repeated valley glaciation in mountains.

[Kleman et al. \(2008, Figure 9\)](#) mapped the durations of varying extents of ice, expected in Fennoscandia over the last 2.6 Ma. Interglacial conditions with glaciers in the mountains lasted 1050 ka, and mountain ice sheets (MIS, reaching no farther than the Gulf of Bothnia) lasted 1120 ka, much longer than the 440 ka of more extensive Fennoscandian Ice Sheets (FIS). The rock-basin ribbon lakes on the eastern side of the mountains could not have developed under FIS conditions, when they were near the main divide, with sluggish ice flow. They must have been eroded by MIS, which were especially important before 1 Ma ago. Similar conditions may apply in British Columbia (on the east side of the Coast Mountains, and

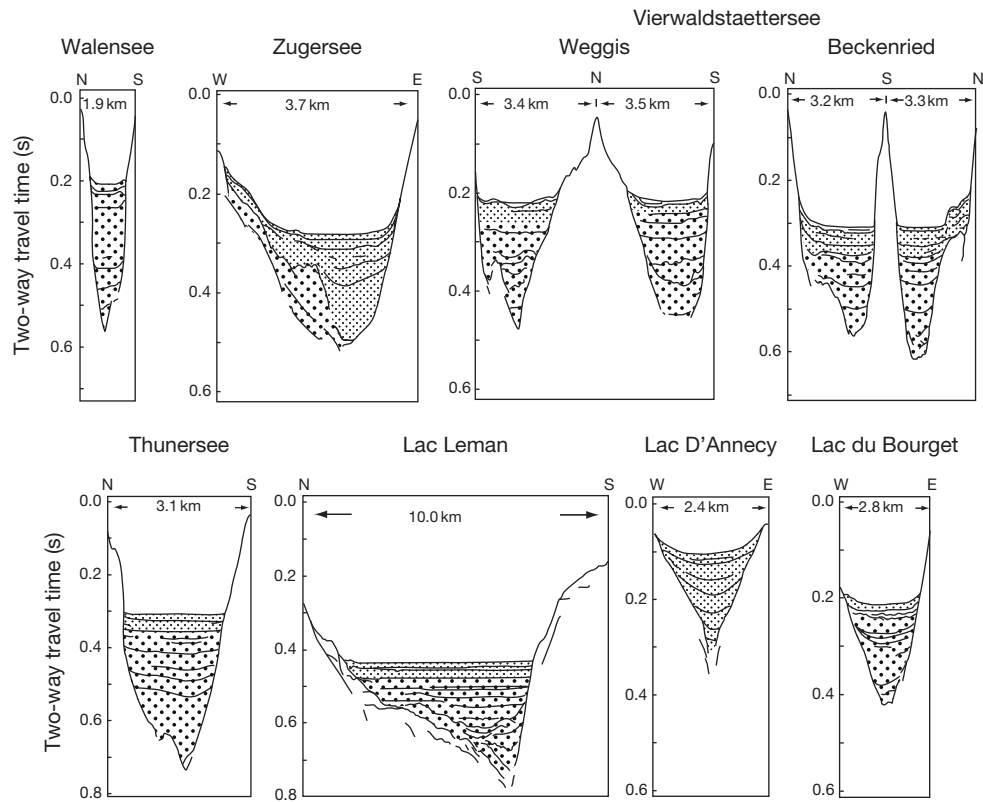


Figure 16 Cross-profiles of lake basins and sediments in the northwestern Alps (Switzerland and France). Basin depth is often twice water depth, as rapid sedimentation during deglaciation tends to fill the sediment trap formed by a lake. Coarse dots: glacially overridden deposits. Fine dots: Late- and postglacial deposits. Reproduced from [Finckh P, Kelts K, and Lambert A \(1984\)](#) Seismic stratigraphy and bedrock forms in perialpine lakes. *Bulletin of the Geological Society of America* 95: 1118–1128.

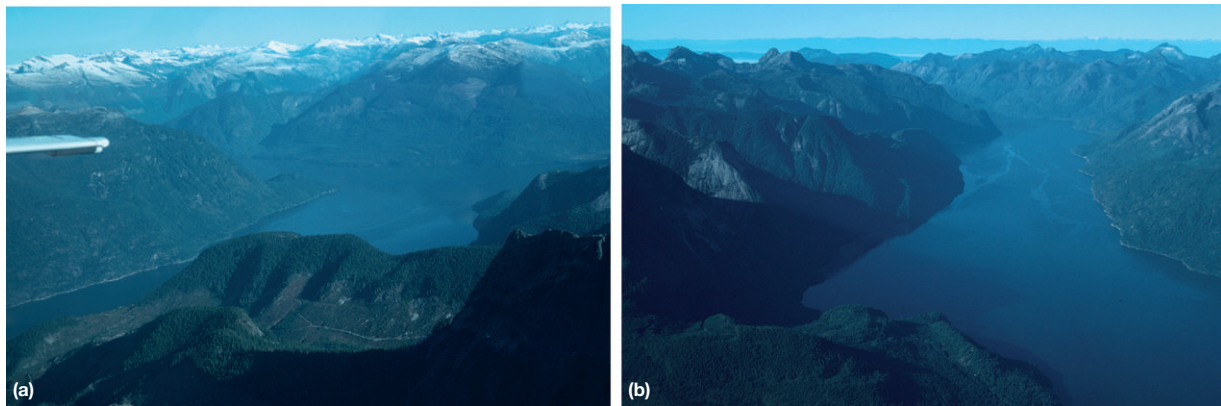


Figure 17 A British Columbian fjord: Jervis Inlet. (a) Queen's Reach, looking north: altitudes rise to 1982 m (left) and 1753 m (right); (b) Princess Royal Reach, looking southwest: altitudes rise to 2282 m on the right and around 2600 m in the distance (photographs by I.S. Evans).

the west side of the Columbia Mountains), but not in Patagonia or New Zealand where recent glacial maxima did not extend far beyond the ribbon lakes. Norwegian fjords could continue developing under both MIS and FIS conditions ([Kleman et al., 2008](#)), helping to explain their basins being deeper and their sides steeper than those of the ribbon lakes.

Full trough development requires erosion of 50–1000 m, which can be achieved at $0.1\text{--}2\text{ mm a}^{-1}$ for several hundred thousand years. These are the rates observed even for small

European glaciers in recent decades, as discussed below. Erosion is expected to be greater early in glacial epochs, as glaciers adjust the landscape to forms better adapted to ice discharge. But if ice becomes warm-based at a later stage, or more meltwater reaches the bed, erosion may accelerate late in glaciations, and abundant meltwater may evacuate stored sediment ([Koppes and Montgomery, 2009](#)).

If troughs are calibrated, are they equilibrium forms (unlike most cirques)? Once a well-calibrated trough has developed,

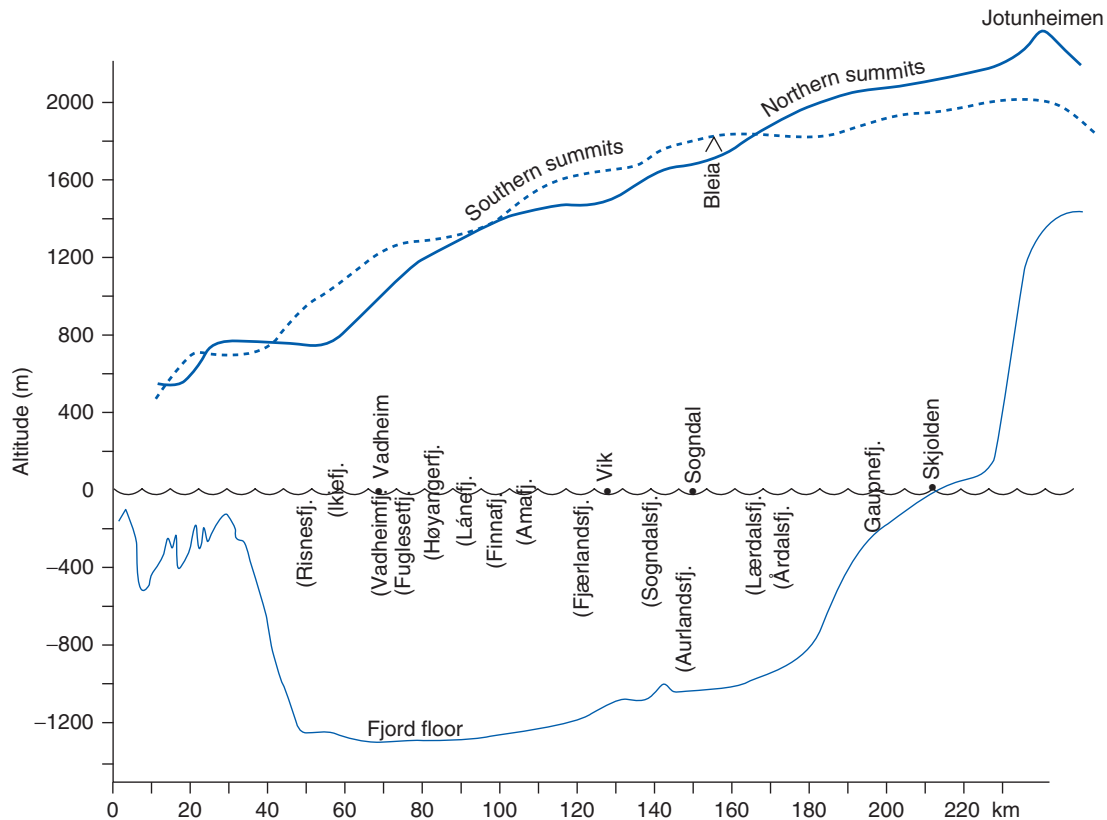


Figure 18 Long profile of Sognefjord (with its Lustrafjord source), western Norway. The level of hanging tributary fjord floors is shown by (Name. Profiles of summit levels on both sides are shown. The fjord floor is covered by an average thickness of 200 m of sediment. Reproduced from Nesje A and Whillans IM (1994) Erosion of Sognefjord, Norway. *Geomorphology* 9: 33–45.

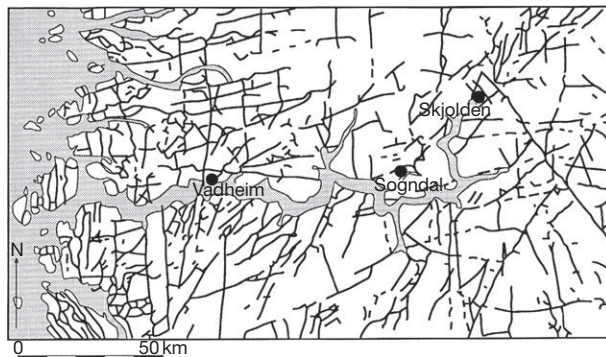


Figure 19 Map showing the relation of fjords to the fracture system around Sognefjord, western Norway. Reproduced from Nesje A and Whillans IM (1994) Erosion of Sognefjord, Norway. *Geomorphology* 9: 33–45.

erosion may slow down and sediments may protect bedrock until greater ice discharges or velocities are achieved. Diminishing land-based glaciers require smaller A and v and are expected to be much less effective unless there is a change of thermal state, gradient or basal hydrology permitting more rapid sliding. Smaller, slower glaciers are likely to erode less bedrock, but abrading and quarrying (as well as slope collapse) do continue during deglaciation. Trough-widening is a one-way process, unlikely to be matched by downward erosion in a

steady state; the positive feedbacks involved make attainment of equilibrium unlikely.

Recent and Holocene Rates of Erosion

A consensus has emerged that wet-based mountain glaciers erode rapidly (Benn and Evans, 2010; Evans, 1997), especially in tectonically active areas, whereas completely cold-based glaciers are much less effective. Short-term (years or decades) measures of glacial erosion are based on suspended sediment concentrations in proglacial streams. Hallet et al. (1996, Table 1: see also Evans, 1997, Table 7.1) collated rates of $0.06\text{--}1.0\text{ mm a}^{-1}$ for small glaciers on crystalline rocks in Norway, $0.4\text{--}1.7\text{ mm a}^{-1}$ for small glaciers in Switzerland, $0.2\text{--}5.0\text{ mm a}^{-1}$ in central Asia (increasing with glacier size: Evans, 1997), and $5\text{--}60\text{ mm a}^{-1}$ for large tidewater glaciers in southeast Alaska. The latter estimates, however, are based on post-LIA (Little Ice Age) sediment volumes in fjords, accumulated during a period of rapid glacial recession associated with accelerated basal sliding. Koppes and Hallet (2006) conclude that longer-term (1000 years) rates are five times slower for the Muir Glacier and three times slower for the Tyndall. Making their proposed corrections, the Alaskan rates of $1\text{--}15\text{ mm a}^{-1}$ remain very high, yet are reasonable for an active orogenic area with high relief and high precipitation. They at least match rates of uplift and exhumation (Spotila et al., 2004).

Koppes et al. (2009) found a similar situation for a tidewater glacier in the Cordillera Darwin, Chilean Tierra del Fuego, which in 45 years has receded 13 km up a fjord and is now 21 km long. Glaciér Marinelli has eroded an average of 39 mm a^{-1} of bedrock over this time, but for stable conditions a rate of 10 mm a^{-1} is extrapolated, and it should be noted that the recession phase of fjord glaciers is much shorter than the advancing phase. As corrected, Hallet et al.'s (1996) data show very high erosion rates in glacier basins larger than 200 km^2 , and rather lower rates for those below 60 km^2 . On a broader scale, Sheaf et al. (2003) calculated the volume of Holocene sediment offshore from the $47\,500 \text{ km}^2$ Pacific drainage of the St. Elias Range, inferring accelerated denudation of at least 5.1 mm a^{-1} over 10 000 a.

The Norwegian glaciers have been more stable during the period of sediment measurement: Nigardsbreen has eroded at an average rate of 0.15 mm a^{-1} for 25 years, a period long enough to smooth out interannual variations in sediment evacuation by subglacial streams. Bedrock erosion is expected to be proportional to the rate of basal sliding, but Riihimäki et al. (2005) show that even a low rate of sliding ($1\text{--}2 \text{ m a}^{-1}$) can erode $1\text{--}2 \text{ mm a}^{-1}$ beneath a small valley glacier.

Areas with more sluggish throughput of snow and ice, such as the dry end of the Pyrenees, have lower erosion rates. Delmas et al. (2009) demonstrated temporal and spatial variations in glacial and paraglacial erosion, with more erosion during deglaciation due to supplies of supraglacial debris, and with cirque headwall recession faster than overall averaged denudation.

Longer-Term Late Cenozoic Exhumation

Over a longer term, rates of surface denudation generally increased in the Late Cenozoic, at times of inception or extension of glaciation. Berger et al. (2008) found that denudation and exhumation of the windward (ocean) side of the St. Elias orogen (southeast Alaska) accelerated around 1 Ma ago, from 0.3 to 3 or 4 mm a^{-1} . They attributed this to the more extensive glaciation (greater ice discharge and near-total cover to the edge of the continental shelf, during maxima) following the mid-Pleistocene transition from dominantly 40 to 100 ka glacial cycles. Glaciation has affected orographic kinematics, producing a narrowing of the orogenic wedge; although the St. Elias may be extreme in both tectonic activity and intensity of glaciation, similar changes are expected in other orogenic belts (Berger et al., 2008). Southeast Alaska is one of several regions where broad glacial troughs ('sea valleys') cross the continental shelf, as recognized off Labrador long ago by F. P. Shepard (Evans, 2008).

Similarly in British Columbia, Shuster et al. (2005) found a sixfold increase in erosion around 1.8 Ma, deepening the Klinaklini trough by 2 km. Although the Klinaklini discharged ice from the interior of British Columbia, this glacial acceleration of erosion was widespread in the southern Coast Mountains (Ehlers et al., 2006). The Late Cenozoic increase in denudation rates, especially 2–4 Ma ago, was global. It is attributed partly to expanded glaciations and to lowered sea level exposing continental shelves, but also to the increased amplitude and frequency of environmental change, preventing the establishment of equilibrium states (Peizhen et al., 2001).

Long-Term Rates: Landform Evidence

In terms of landform development, the overall rate of denudation is less important than the spatial pattern. The more selective, concentrated impacts of glacial erosion, compared with the even surface lowering in mature fluvial landscapes (Jamieson et al., 2008) explains the dominance of glacial forms in glaciated mountains. This provides strong evidence of the effectiveness of glacial erosion: the distinctive landforms of glaciated mountains have survived over 11 000 years of fluvial and slope processes. Small glaciers are often concentrated on one side of a ridge, either the shadier or the sheltered side (Evans, 2006a). This leads to cirque headwall retreat: Evans (2006b) cited examples where headward erosion around north-facing glacier sources has pushed divides southward, for example, by 1 or 2 km in parts of the British Columbia Coast Mountains. Brocklehurst and Whipple (2006) suggested that cirque headwall retreat is important on the east slope of the Californian Sierra Nevada, causing divide recession of up to 2.5 km beyond that expected from fluvial erosion (Figure 12).

Oskin and Burbank (2005) proposed that as mountains are uplifted above the snowline, glacial erosion increases local relief by deepening and widening fluvial valleys. In the Kyrgyz Range of the Tien Shan this starts from the north slope, where the snowline is some 200 m lower, and the divide is pushed 0.9–4.4 km southward. Erosion is localized at the bases of cirque headwalls, and cirque headwall retreat is two to three times the rate of vertical erosion. "Cirque retreat can effectively bevel across an elevated alpine plateau..." (Oskin and Burbank, 2005, p. 936).

In 11 east-trending valleys in the Bitterroot Mountains of Idaho, north-facing glaciers have deepened tributary valleys by 90 m more than fluvial erosion on south-facing slopes, and headwall retreat has pushed divides on average 400 m southward (Naylor and Gabet, 2007). Amerson et al. (2008) showed that in both central Idaho and the Olympic Mountains, for a given drainage area and comparable lithology, glaciated valleys are around 32% deeper than fluvial valleys, and around 37% wider. On the granitic rocks of central Idaho, overall valley-side slope gradients exceed 35° only in glaciated valleys. Although Koppes and Montgomery (2009) find that river incision in active orogens can be as fast as glacial denudation, the above results are evidence of glacial erosion exceeding fluvial during the Late Quaternary.

In Norway and British Columbia, great contrasts in precipitation between coastal and inland areas have affected the degree of glacial activity. Rudberg (1997) made estimates of the volume of rock removed to produce various landforms in central Norway, comparing the $15\,100 \text{ km}^2$ coastal province of Møre og Romsdal with an adjacent comparable area inland:

Volume in km^3 :	Coastal	Interior
V-shaped fluvial valleys	15	22
Cirques	60	4
Trough valleys	2150	330
Fjords/lakes	1290	4
Sounds and submarine valleys	265	
Strandflat	270	

Although very rough, these estimates clearly show the much more intense glacial transformation of the land surface in the

coastal area, of the order of tenfold. This is attributed to high snowfall producing faster flow and greater erosion in the windward, coastal area.

Landscapes of Glacial Erosion

Buzzsaw Hypothesis

Based on this demonstration that glacial denudation is more effective than fluvial (so long as glaciers are wet-based), and a second assumption that basal sliding and thus erosion are greatest around the ELA, recent papers have likened glacial erosion to a buzzsaw (chain saw) operating at ELA, and near-truncating mountain ranges so that only a small proportion of mountain area rises above ELA. While noting that recent sediment yields in southeast Alaska have accelerated, [Spotila et al. \(2004\)](#) found that spatial patterns of denudation over the long term (million years) are strongly correlated with the distribution of glaciers, and the southward decrease in ELA correlates with a decrease in mean elevation. [Berger and Spotila \(2008\)](#) found that exhumation over the last half million years peaks at some 4 mm a^{-1} , between the intersections of the present and the Last Glacial Maximum (LGM) ELAs with the topography of the Chugach–St. Elias Range. Exhumation does not vary laterally with glacier size, implying that erosion by both larger and smaller valley glaciers has kept pace with Late Quaternary rock uplift. [Enkelmann et al. \(2009\)](#) related the very young rock cooling ages around the Seward–Malaspina glacier system to intense localized exhumation at an active orogenic ‘corner’: rapid glacial erosion is coupled with active tectonic uplift.

In part of the Washington Cascades, [Mitchell and Montgomery \(2006\)](#) noted a parallelism between three trends: nonvolcano summit altitudes, cirque floors, and modern glacier median altitudes. Each rises eastward at $9\text{--}15 \text{ m km}^{-1}$. Only about 10% of each section rises above the highest cirque floors, and few peaks rise more than 600 m above. Rock exhumation rates are greatest 30–40 km west of the highest peaks, and suggest vertical erosion of 2–5 km in the last 15 Ma. This is not the pattern expected from fluvial or slope erosion, so Mitchell and Montgomery proposed a glacial buzzsaw of greatest glacial erosion at the average Quaternary ELA (represented by the cirque floors), where ice discharge and velocity were greatest. This increased the slope gradients above cirque floors to over 30° , causing slope failure. Thus, both vertical and headward erosion in cirques is considered to dominate landscape development at high altitudes.

Recent work has used hypsometric (area-altitude) analysis to support the Buzzsaw Hypothesis. The most ambitious was by [Egholm et al. \(2009\)](#) and [Pedersen et al. \(2010\)](#), using SRTM data for most of the world (up to 60°N and 55°S). For 1029 one-degree data tiles with mountains believed to rise above present ELA, they establish a pronounced maximum of area some 660 m below present ELA. This applies across a broad range of ELAs, from 1000 to 6000 m. It is supported by a series of minima in average slope gradient, between present ELA and 1000 m lower. In that range, (tangent) gradients average 0.27–0.29, whereas they soon rise to 0.32–0.42 on either side (Figure 4A in [Pedersen et al., 2010](#)). Where good ELA data were available from the World Glacier Inventory, clear area maxima were 600 m below present ELA in Tibet

(with the Himalayas and Tien Shan), 700 m below in southern Norway, 850 m in Mongolia and southwest Siberia, and 1050 m in coastal Alaska and B.C. Area maxima for individual data tiles are mainly between present and former (Late Quaternary) ELAs. Only in a few regions do peaks rise more than 1500 m above present ELA (Figure 2 in [Egholm et al., 2009](#)). [Egholm et al.](#) were also able to model the development of the hypsometric maxima.

Van der Beek and Bourbon (2008) analyzed the hypsometry and clinometry of part of the Dauphiné Alps, France, and confirmed a strong glacial imprint. For seven major drainage basins, area peaks at altitudes between present (2900 m) and glacial maximum (1800 m) ELA: 61% of the area of the nine basins studied lies within this band, which is 41% of the total altitude range. In the most heavily glaciated basins (the Upper Romanche, Séveraisse, and Vénéon), this is accompanied by a shallow minimum in mean slope gradient.

The positional and lithological reasons for some peaks rising hundreds of meters above the rest were discussed by [Foster et al. \(2010\)](#) for the Teton Range of Wyoming, where headward erosion is important and the Buzzsaw Hypothesis is otherwise applicable. They did, however, suggest a discharge threshold: glaciers with basins over 20 km^2 incise much deeper troughs than smaller glaciers. Small glaciers ($<10 \text{ km}$ long at maximum), in several Idaho ranges uplifting at 0.3 mm a^{-1} or less, have eroded effectively and limited maximum altitudes, which trend parallel to cirque floor altitudes and to estimates of LGM ELA ([Foster et al., 2008](#)). There are some hypsometric maxima near LGM ELA, and steeper gradients above, also supporting the Buzzsaw Hypothesis. Small glaciers in the Teton Range were apparently less effective in eroding the highest summits, perhaps because the Range is rising at a fast rate of 2 mm a^{-1} ([Foster et al., 2008](#)).

Simulations by [Anderson et al. \(2006\)](#) confirmed that valley glacier erosion should be greatest near the long-term mean ELA, so long as this is below the summits and the ice is warm-based. They proposed a more robust rule, that erosion peaks at 30–40% of the distance from source to maximum glacial limit (if within a valley). Note also the three situations discussed above where valley glacier erosion may reach a maximum well below ELA, and that on a low-gradient glacier the ELA erosion maximum is broad, due to discharge initially rising and falling slowly either side of EL. [MacGregor et al.’s \(2009\)](#) simulation of cirque development produced cirque floors hundreds of meters below ELA.

The Buzzsaw Hypothesis can explain why so many mountain ranges have small glaciers, while very few have extensive icecaps or icefields; outside polar regions, the latter are confined to Patagonia, the Saint Elias and the Karakoram. Each of these appears to be undergoing rapid uplift, sufficient perhaps to take substantial areas quickly through the zone of varying ELA to altitudes where ice is cold-based and less erosive. The glacial buzzsaw limits peak altitudes in many glacially dissected regions (with intersecting cirque headwalls). That cannot apply where smooth summit surfaces are only partly consumed by cirque development, as in the Colorado Rockies (Front Range) but, even there, hypsometric maxima may relate to cirque floors. Together with the asymmetric divides discussed in the previous section, such evidence confirms the effectiveness of glacial erosion in many mountain ranges.

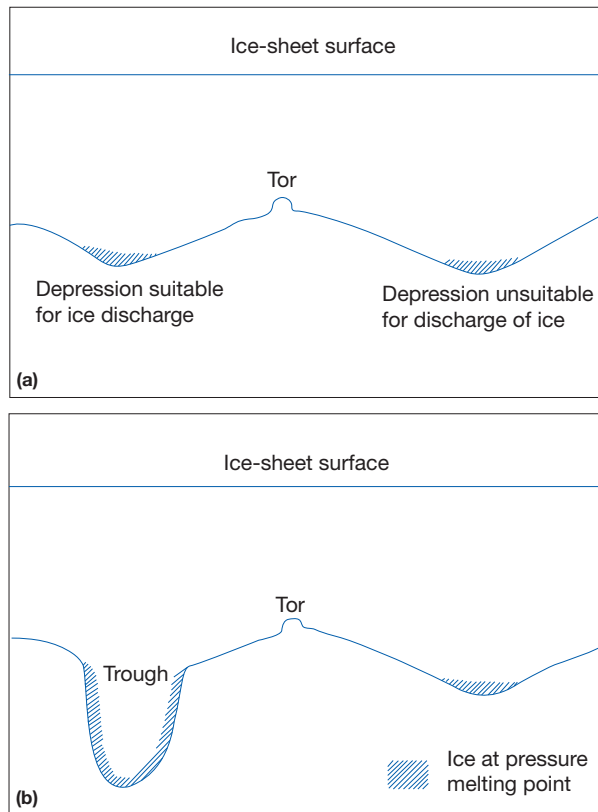


Figure 20 Selective linear erosion of troughs requires basal ice at pressure melting point, in a depression that can channel ice flow. Reproduced from Sugden DE (1974) Landscapes of glacial erosion in Greenland and their relationship to ice, topographic and bedrock conditions. In: Brown EH and Waters RS (eds.) *Progress in Geomorphology*. Institute of British Geographers Special Publication 7: 177–195.

Selectivity

Sugden (1974) noted the importance of relief and basal temperature in developing different landscapes of glacial erosion. Where ice had remained frozen to its bed, there was hardly any erosion unless meltwater from upstream froze on to the glacier base. Where deeper valleys had permitted basal melting and thus sliding, they were deepened further while intervening uplands were unaffected (Figure 20). Valleys focusing ice flow and facilitating evacuation to areas of net ablation permitted faster flow and generation of more heat and more basal melting; positive feedback enhanced the landscape contrasts. Where a whole landscape was covered by warm-based ice, it was scoured into rocky hills and lake basins. These three landscape types were labeled, respectively, landscapes of no erosion; of selective linear erosion; and of areal scouring. They contrast also with a fourth, alpine landscape, of cirques and troughs.

Kleman and Stroeven (1997) demonstrated the selectivity of glacial erosion in the Swedish mountains, and Kleman et al. (2008) applied the concept throughout the FIS area. Recent studies using terrestrial cosmogenic nuclide (TCN) concentrations confirm the selectivity of glacial erosion. For example, from striated surfaces near the downstream ends of two glacial troughs, some 2.5 m has been eroded during the last major

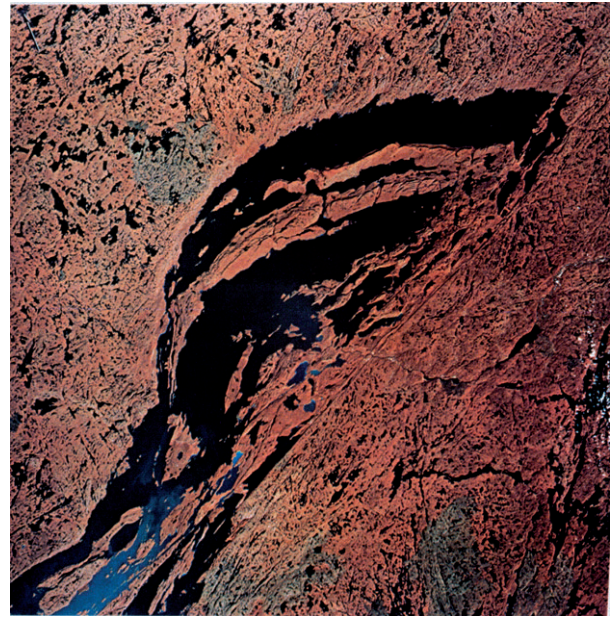


Figure 21 NASA Landsat image of Canadian Shield around the eastern arm of Great Slave Lake, Canada. Cuestas in the Lake show differential glacial erosion among bedded volcanic rocks. Elsewhere the pattern of small lakes shows erosion of thousands of rock basins in Precambrian (Archaean) igneous and metamorphic rocks by the Laurentide Ice Sheet.

glaciation near the trough center, declining to lesser depths near the margins (Fabel et al., 2004), as predicted by Harbor's model. This study was based on abraded areas, not quarried. Further upstream, the depth of scour was greater, with complete removal of inherited nuclides throughout the cross-section. TCN work in Labrador also confirms selectivity of erosion (Staiger et al., 2005); summit plateaus where ice has always been cold have been lowered <3 m over 20 glacial cycles, whereas valleys with wet-based ice have been lowered at least 50 m by glacial erosion. Delmas et al. (2009) find limited and selective glacial erosion at the drier, eastern end of the Pyrenees.

Thus, many recent studies, including Swift et al. (2008), have confirmed the applicability of Sugden's model to landscape development on tectonic passive margins around the North Atlantic. Erosion on these margins is also much slower than in high-precipitation orogenic areas such as southeast Alaska, Patagonia and much of the Himalaya.

Areal Scouring

Landscapes of areal scouring might be expected to show streamlining and, thus, lineation. This is common on weak rocks, for example, around the Forth in central Scotland (Linton, 1963). On resistant metamorphic and igneous rocks of the Canadian and Fennoscandian Shields, there is strong structural control of glacial erosion with lake basins along fault and major joint lines, and on weaker rocks (Figure 21). This leads some to infer that glaciation has simply removed a mantle of weathered rocks, which is clearly the case in some marginal areas of decelerating flow. Elsewhere, in the absence of deeply weathered material, erosion of bedrock may have been considerable but is difficult to measure. Kleman et al. (2008) suggested that

ice-scoured areas developed quickly, under rapid ice flow toward calving margins: some could develop only under more extensive, FIS conditions.

See also: **Glacial Landforms:** Glaciofluvial Landforms of Erosion. **Glacial Landforms, Erosional Features:** Micro- to Macroscale Forms.

References

- Amerson BE, Montgomery DR, and Meyer G (2008) Relative size of fluvial and glaciated valleys in central Idaho. *Geomorphology* 93: 537–547.
- Amundson JM and Iverson NR (2006) Testing a glacial erosion rule using hang heights of hanging valleys, Jasper National Park, Alberta, Canada. *Journal of Geophysical Research* 111: F01020. <http://dx.doi.org/10.1029/2005JF000359>.
- Anderson RS, Molnar P, and Kessler MA (2006) Features of glacial valley profiles simply explained. *Journal of Geophysical Research* 111: F01004. <http://dx.doi.org/10.1029/2005JF000344>.
- Augustinus PC (1992) The influence of rock mass strength on glacial valley cross-profile morphometry: A case study from the Southern Alps, New Zealand. *Earth Surface Processes and Landforms* 17: 39–51.
- Benn DI and Evans DJA (2010) *Glaciers and Glaciation*, 2nd edn. London: Hodder Education.
- Bennett MR, Huddart D, and Glasser NF (1999) Large-scale bedrock displacement by cirque glaciers. *Arctic, Antarctic, and Alpine Research* 31: 99–107.
- Berger AL and Spotila JA (2008) Denudation and deformation in a glaciated orogenic wedge: The St. Elias orogen, Alaska. *Geology* 36: 523–526.
- Berger AL, et al. (2008) Quaternary tectonic response to intensified erosion in an orogenic wedge. *Nature Geoscience* 1: 793–799.
- Bovis MJ (1990) Rock-slope deformation at Affliction Creek, southern Coast Mountains, British Columbia. *Canadian Journal of Earth Sciences* 27: 243–254. (postscript, 2000; 1998 collapse; *Canadian Journal of Earth Sciences* 37: 1321–1334).
- Brocklehurst SH and Whipple KX (2006) Assessing the relative efficiency of fluvial and glacial erosion through simulation of fluvial landscapes. *Geomorphology* 75: 283–299.
- Brook MS, Kirkbride MP, and Brock BW (2006) Cirque development in a steadily uplifting range: Rates of erosion and long-term morphometric change in alpine cirques in the Ben Ohau Range, New Zealand. *Earth Surface Processes and Landforms* 31: 1167–1175.
- Delmas M, Calvet M, and Gunnell Y (2009) Variability of erosion rates in the eastern Pyrenees during the last glacial cycle – A global perspective on the impact of glacial erosion on mountain landscapes. *Quaternary Science Reviews* 28: 484–498.
- Egholm DL, Nielsen SB, Pedersen VK, and Lesemann J-E (2009) Glacial effects limiting mountain height. *Nature* 460(7257): 884–887. <http://dx.doi.org/10.1038/nature08263>.
- Embleton CE (ed.) (1972) *Glaciers and Glacial Erosion (Readings and Translations)*. London: Macmillan.
- Enkelmann E, Zeitler PK, Pavlis TL, Garver JL, and Ridgway KD (2009) Intense localized rock uplift and erosion in the St Elias orogen of Alaska. *Nature Geoscience* 2: 360–363. <http://dx.doi.org/10.1038/NGEO502>.
- Evans IS (1997) Process and form in the erosion of glaciated mountains. In: Stoddart DR (ed.) *Process and Form in Geomorphology*, pp. 145–174. London: Routledge.
- Evans IS (2006a) Local aspect asymmetry of mountain glaciation: A global survey of consistency of favoured directions for glacier numbers and altitudes. *Geomorphology* 73: 166–184.
- Evans IS (2006b) Allometric development of glacial cirque form: Geological, relief and regional effects on the cirques of Wales. *Geomorphology* 80: 245–266.
- Evans IS (2008) Glacial erosional processes and forms: Mountain glaciation and glacier geography, ch. 11. In: Burt TP, Chorley RJ, Brunsden D, Cox NJ, and Goudie AS (eds.) *The History of the Study of Landforms or the Development of Geomorphology, Vol. 4: Quaternary and Recent Processes and Forms (1890–1960s) and the Mid-Century Revolutions*, pp. 413–494. London: The Geological Society. ISBN 978-1-86239-249-6.
- Evans IS and Cox NJ (1974) Geomorphometry and the operational definition of cirques. *Area* 6: 150–153 (IBG, London).
- Evans IS and Cox NJ (1995) The form of glacial cirques in the English Lake District, Cumbria. *Zeitschrift für Geomorphologie, N.F.* 39: 175–202.
- Fabel D, Harbor J, Dahms D, et al. (2004) Spatial patterns of glacial erosion at a valley scale derived from terrestrial cosmogenic ^{10}Be and ^{26}Al concentrations in rock. *Annals of the Association of American Geographers* 94: 241–255.
- Federici PR and Spagnolo M (2004) Morphometric analysis on the size, shape and area distribution of glacial cirques in the Maritime Alps (Western French-Italian Alps). *Geografiska Annaler* 86A: 235–248.
- Finckh P, Kelts K, and Lambert A (1984) Seismic stratigraphy and bedrock forms in perialpine lakes. *Bulletin of the Geological Society of America* 95: 1118–1128.
- Foster D, Brocklehurst SH, and Gawthorpe RL (2008) Small valley glaciers and the effectiveness of the glacial buzzsaw in the northern Basin and Range, USA. *Geomorphology* 102: 624–639.
- Foster D, Brocklehurst SH, and Gawthorpe RL (2010) Glacial-topographic interactions in the Teton Range, Wyoming. *Journal of Geophysical Research* 115: F01007. <http://dx.doi.org/10.1029/2008JF001135>.
- Glasser NF and Bennett MR (2004) Glacial erosional landforms: Origins and significance for palaeoglaciology. *Progress in Physical Geography* 28: 43–75.
- Gordon J (1977) Morphometry of cirques in the Kintail–Affric–Cannich area of NW Scotland. *Geografiska Annaler* 59A: 177–194.
- Hallet B, Hunter L, and Bogen J (1996) Rates of glacial erosion and sediment evacuation by glaciers: A review of field data and their implications. *Global and Planetary Change* 12: 213–235.
- Harbor JM (1992) Numerical modeling of the development of U-shaped valleys by glacial erosion. *Bulletin of the Geological Society of America* 104: 1364–1375.
- Haynes VM (1998) The morphological development of alpine valley heads in the Antarctic Peninsula. *Earth Surface Processes and Landforms* 23: 53–68.
- Holtedahl H (1967) Notes on the formation of fjords and fjord-valleys. *Geografiska Annaler* 49A: 188–203.
- Le B Hooke R (1991) Positive feedbacks associated with erosion of glacial cirques and overdeepenings. *Bulletin of the Geological Society of America* 103: 1104–1108.
- Iverson NR (1991) Potential effects of subglacial water-pressure fluctuations on quarrying. *Journal of Glaciology* 37(125): 27–36.
- Jamieson SSR, Hulton NRJ, and Hagdorn M (2008) Modelling landscape evolution under ice sheets. *Geomorphology* 97: 91–108.
- Jarman D (2006) Large rock slope failures in the Highlands of Scotland: Characterisation, causes and spatial distribution. *Engineering Geology* 83: 161–182.
- Klemm J and Stroeven AP (1997) Preglacial surface remnants and Quaternary glacial regimes in northwestern Sweden. *Geomorphology* 19: 35–54.
- Klemm J, Stroeven AP, and Lundqvist J (2008) Patterns of Quaternary ice sheet erosion and deposition in Fennoscandia and a theoretical framework for explanation. *Geomorphology* 97: 73–90.
- Koppes M and Hallet B (2006) Erosion rates during rapid deglaciation in Icy Bay, Alaska. *Journal of Geophysical Research* 111: F02023. <http://dx.doi.org/10.1029/2005JF000349>.
- Koppes M, Hallet B, and Anderson J (2009) Synchronous acceleration of ice loss and glacier erosion, Glaciar Marinelli, Chilean Tierra del Fuego. *Journal of Glaciology* 55(190): 207–220.
- Koppes M and Montgomery DR (2009) The relative efficacy of fluvial and glacial erosion over modern to orogenic timescales. *Nature Geoscience* 2: 644–647. <http://dx.doi.org/10.1038/NGEO616>.
- Larsen E and Mangerud J (1981) Erosion rate of a Younger Dryas cirque glacier at Kråkenes, western Norway. *Annals of Glaciology* 2: 153–158.
- Lewis WV (ed.) (1960) *Norwegian Cirque Glaciers. R.G.S. Research Series 4*, 104 pp. London: Royal Geographical Society.
- Li YK, Liu GN, and Cui ZJ (2001) Glacial valley cross profile morphology, Tian Shan Mountains, China. *Geomorphology* 38: 153–166.
- Linton DL (1963) The forms of glacial erosion. *Transactions of the Institute of British Geographers* 33: 1–28.
- MacGregor KR, Anderson RS, Anderson SP, and Waddington ED (2000) Numerical simulations of glacial-valley longitudinal profile evolution. *Geology* 28: 1031–1034.
- MacGregor KR, Anderson RS, and Waddington ED (2009) Numerical modeling of glacial erosion and headwall processes in alpine valleys. *Geomorphology* 103: 189–204.
- McSaveney MJ (2002) Recent rockfalls and rock avalanches in Mount Cook National Park, New Zealand. In: Evans SG and DeGraff JV (eds.) *Catastrophic Landslides: Effects, Occurrence, and Mechanisms, Reviews in Engineering Geology*, vol. XV, pp. 35–70. Boulder, CO: Geological Society of America.
- Mindrescu M, Evans IS, and Cox NJ (2010) Climatic implications of cirque distribution in the Romanian Carpathians: Palaeowind directions during glacial periods. *Journal of Quaternary Science* 25: 875–888.
- Mitchell SG and Montgomery DR (2006) Influence of a glacial buzzsaw on the height and morphology of the Cascade Range in central Washington State, USA. *Quaternary Research* 65: 96–107.

- Monjuvent G (1974) Considérations sur le relief glaciaire a propos des Alpes de Dauphiné. *Revue de Géographie Physique et de Géologie Dynamique, Série 2* 16: 466–502.
- Naylor S and Gabet EJ (2007) Valley asymmetry and glacial versus non-glacial erosion in the Bitterroot Range, Montana, USA. *Geology* 35: 375–378.
- Nesje A and Whillans IM (1994) Erosion of Sognefjord, Norway. *Geomorphology* 9: 33–45.
- Oskin M and Burbank DW (2005) Alpine landscape evolution dominated by cirque retreat. *Geology* 33: 933–936.
- Pedersen VK, Egholm DL, and Nielsen SB (2010) Alpine glacial topography and the rate of rock column uplift: A global perspective. *Geomorphology* 122: 129–139.
- Peizhen Z, Molnar P, and Downs WR (2001) Increased sedimentation rates and grain sizes 2–4 Myr ago due to the influence of climate change on erosion rates. *Nature* 410: 891–897.
- Richardson C and Holmlund P (1996) Glacial cirque formation in northern Scandinavia. *Annals of Glaciology* 22: 102–106.
- Riihimäki CA, MacGregor KR, Anderson RS, Anderson SP, and Loso MG (2005) Sediment evacuation and glacial erosion rates at a small alpine glacier. *Journal of Geophysical Research* 110: F03003. <http://dx.doi.org/10.1029/2004JF000189>.
- Rudberg S (1997) Glacial and interglacial erosion in Scandinavian mountains in a west–east comparison including an approach to a quantitative calculation. *Zeitschrift für Geomorphologie* 41: 183–204.
- Sanders JW, Cuffey KM, MacGregor KR, Kavanaugh JL, and Dow CF (2010) Dynamics of an alpine cirque glacier. *American Journal of Science* 310: 753–773.
- Sheaf MA, Serpa L, and Pavlis TL (2003) Exhumation rates in the St. Elias Mountains, Alaska. *Tectonophysics* 367: 1–11.
- Shuster DL, Ehlers TA, Rusmore ME, and Farley KA (2005) Rapid glacial erosion at 1.8 Ma revealed by $^4\text{He}/^3\text{He}$ thermochronometry. *Science* 310(1668): 1668–1670.
- Spotila JA, Buscher JT, Meigs AJ, and Reiners PW (2004) Long-term glacial erosion of active mountain belts: Example of the Chugach–St. Elias Range, Alaska. *Geology* 32: 501–504.
- Staiger JKW, Gosse JC, Johnson JV, et al. (2005) Quaternary relief generation by polythermal glacier ice. *Earth Surface Processes and Landforms* 30: 1145–1159.
- Sugden DE (1974) Landscapes of glacial erosion in Greenland and their relationship to ice, topographic and bedrock conditions. In: Brown EH and Waters RS (eds.) *Progress in Geomorphology, Institute of British Geographers Special Publication 7*: 177–195.
- Swift DA, Persano C, Stuart FM, Gallagher K, and Whitham A (2008) A reassessment of the role of ice sheet glaciation in the long-term evolution of the East Greenland fjord region. *Geomorphology* 97: 109–125.
- Turnbull JM and Davies TRH (2006) A mass movement origin for cirques. *Earth Surface Processes and Landforms* 31: 1129–1148.

Micro- to Macroscale Forms

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Micro- to macroscale forms of glacial erosion represent some of the most widely recognized indicators of glaciation and pay testament to the efficacy of glaciers as erosional agents (Figure 1). The use of micro- to macroscale forms of glacial erosion was vital in the evolution of Quaternary science with regard to the development of the 'glacial theory' in the mid-1800s and the wide acceptance that significant areas of the mid-latitudes supported extensive glaciers and ice sheets in the past (Agassiz, 1831; Forbes, 1843).

The suite of features produced by glacial erosion, from micro- through macroscale dimensions, forms a continuum shown in Table 1. In relation to Quaternary science, these erosional forms play a number of roles. Their presence in the landscape provides unequivocal evidence of former glaciation and that the ice was, at least at some time, warm-based (Benn and Evans, 1998). All ice-streamlined erosional forms can be used to identify ice-flow directions, but must be interpreted in conjunction with other information (e.g., topography, erratic trains, and paleogeography) as many are bidirectional indicators of ice flow (Table 1). The ultimate goal is to link erosional forms with ice dynamics in order to facilitate detailed reconstructions of former ice masses. The first requirement, however, is to understand the processes of subglacial erosion.

Fundamentals of Glacial Erosion

The first and most fundamental consideration related to all the landforms detailed in Table 1 is that they are only produced by ice that is warm-based/wet-based/at the pressure melting point and sliding across its bed (Benn and Evans, 1998). Although there are reports of basal sliding at subpressure melting point temperatures, the amount of erosion that may be accomplished is likely to be minimal. All the forms discussed here are eroded in bedrock and are only produced where there is 'direct' (cavities included) contact between the ice and the bed or there is a thin (probably discontinuous) sediment layer between the bedrock and the glacier sole. Two processes of glacial erosion are responsible for the formation of all the forms listed in Table 1, abrasion and quarrying (Boulton, 1974). The erosion of bedrock by flowing subglacial water (mechanical and/or chemical) is not considered explicitly here, although it may play some part in the formation of a number of the landforms described. Subglacial bedrock erosion rates by abrasion and quarrying are generally assumed to be dominant over subglacial fluvial processes.

Abrasion

Abrasion is the process of (frictional) wear, produced by surfaces rubbing against each other, and is achieved in the subglacial environment by sliding of debris-charged ice across the bed. The key is the presence of debris (clasts) in the basal ice of

the glacier. Ice-on-rock will not produce mechanical wear of the bed, although it may produce chemical erosion through dissolution. Rather, it is at the contact between clasts entrained in the basal ice and the bedrock where abrasion/wear occurs. The classic illustration of subglacial abrasive wear are striae (Figures 1 and 2), which are ubiquitous on bedrock that has been overridden by debris-charged warm-based ice. Chamberlin (1888) undertook the first and most comprehensive field investigation, providing very detailed descriptions of striae and friction cracks.

Careful observations of striae provide key indicators of the formation process. The sides of striae are often jagged and rough, with partial ring fractures present in the bottom that are most often concave in the direction of flow (Figure 3). This demonstrates that the plowing of the asperity (a point protuberance from a clast that contacts the bed) into the bed proceeds by brittle fracture. The stress fields created beneath static and sliding contacts are well known from fracture and rock mechanics, and they provide a complete explanation of striae formation (Frank and Lawn, 1967; Lawn, 1967; Swain and Lawn, 1976). A brief description is provided here, and a more detailed explanation can be found in Iverson (1995). Figure 4 shows the bed stresses and likely crack paths developed beneath static spherical, static point, and sliding spherical asperities, associated with a loading-unloading cycle. The important points to note are the ring, central, and median crack paths and, when sliding, how the stress trajectories are modified such that partial ring fracture formation is most likely at the trailing edge contact. Close observation of striae in the field often reveals the presence of partial ring fractures (Figure 3). Figure 5 presents an experimental striation indicating how the trailing edge tensile stresses and crack paths are controlled by contact geometry. Here the contact was oriented oblique to movement with cracks becoming curved (concave down-ice) laterally beyond the trailing edge contact (Figure 4(c)). Figure 6 shows the relationship between contact geometry and stress in the bed for plowing asperities (Iverson, 1991a). The expected crack paths are at the trailing edge of the contact for low plowing angles (similar to the sliding spherical indenter in Figure 4(c)) but extend into and ahead of a high plowing angle contact. Thus, for high plowing angles the asperity is more likely to plow down into the bed (Figure 6(c)). Parameters that potentially control striae formation (abrasion) are the force pressing the asperity against the bed (F_n), the hardness of the clast relative to the bed, the geometry of the asperity-bed contact, and the relative velocity of clast to ice.

Over the length scales of striae, $\leq 1\text{--}2\text{ m}$ (Table 1), none of these parameters are likely to change significantly except the contact geometry (Iverson, 1991a, 1995), and this is the primary control on striae morphology. F_n is a critical parameter and there is divergence of opinion regarding the contributing factors. Boulton (1974) suggested that it is a function of the effective pressure acting on the bed. Problematic with this

theory is that at high effective stress, clasts are prone to lodgement (increased bed friction), and therefore there is difficulty explaining the formation of valleys and fjords that contained thick ice masses. Hallet (1979, 1981) argued that the Boulton model could not promote the deviatoric stress required to induce bed fracture because the same stress is transmitted to the bed via the ice-bed or basal water-bed contact. Hallet proposed that F_n is the sum of the buoyant weight of the clast and the viscous drag imparted on the clast by the component of ice flow that is directed toward the bed, and it is essentially independent of the ice thickness. In his model, thin, warm, fast sliding ice is the most erosive. Hindmarsh (1996) argued that the presence of discontinuities in meltwater film pressures developed around clast edges can isolate cavities around clasts by effectively pinching out/pressure sealing them from the wider basal hydraulic system. The cavity size is controlled by the relative velocity of the clast and the ice and the effective pressure in the cavity. As the cavity grows, the ratio of horizontal to vertical forces acting on the clast increases and motion may be permitted. F_n is then related to the relative velocity of clast to the ice through the water pressure gradient on the stoss side of the clast. Cohen et al. (2005) demonstrated, from subglacial experiments in the Engabreen subglacial laboratory,

that the simple friction law implied by the Boulton model could not account for the high shear tractions (100–500 kPa) they measured, but their results did demonstrate a dependence of shear traction on effective normal stress. Their data failed to demonstrate a clear dependence of shear traction on melt rate,



Figure 2 Heavily striated bedrock, Øksfjordjøkelen, Norway. Ice flow was from right to left. Photograph by B. R. Rea.



Figure 1 A striated bedrock exposure from Dwyka, South Africa, showing familiar results of subglacial abrasion (ruler is 15 cm). This striated surface dates from the Permo-Carboniferous glaciation of Gondwana approximately 300 Ma. Photograph by W. B. Whalley.



Figure 3 A large (type 2) striation showing partial ring fractures (three are marked for clarity), concave in the downflow direction. Note how the fractures extend beyond the contact zone in some places. Photograph by B. R. Rea.

Table 1 Size classification of glacial erosional forms

Relief type	Relief shape	Micro	Scale				Macro	
		m^{-2} (1 cm)	m^{-1}	m^0 (1 m)	m^1	m^2 (100 m)	m^3 m^4 (10 km)	m^5
Eminence	Streamlined			← Whaleback →	← Rock drumlin →	← Spur →		
			← Mini crag and tail/rat tails →		← Crag and tail →			
Depression	Part-streamlined			← Roche mounnée- Flyggborg →				
	Streamlined		← Striae →	← Friction cracks →		← Groove →		
	Part-streamlined			← p-form →				
						← Rock basin →		

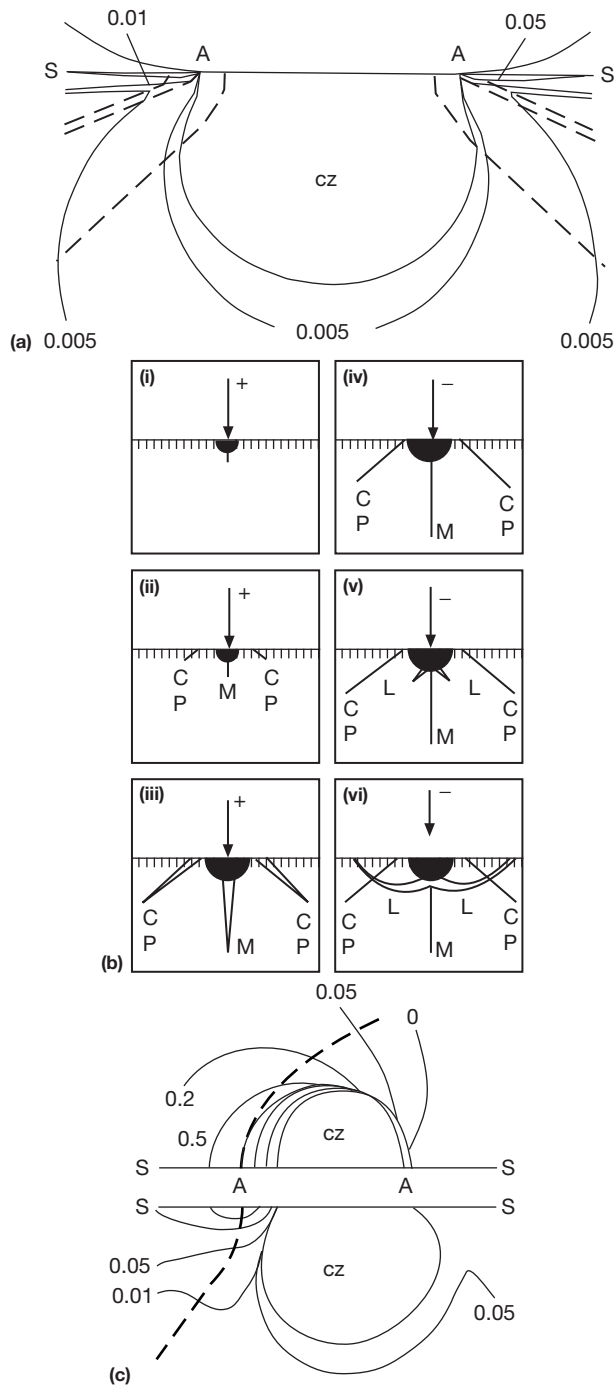


Figure 4 (a) Stress field developed in the bed beneath a static spherical indenter. Outside of the compressive zone (cz) the contour values indicate the tensile stresses relative to the mean pressure between the indenter and the bed. The maximum tensile stress is developed at the edge of the circle of contact (A–A). Dashed lines show σ_3 stress trajectories which control the downward penetration of a cone crack. Redrawn from Frank FC and Lawn BR (1967) On the theory of Hertzian fracture. *Proceedings of the Royal Society, Series A* 299: 291–306, Figure 1; with permission of the Royal Society. (b) Cracks developed beneath a static point indenter during a complete loading (i–iii)/unloading (iv–vi) cycle. M, median crack; CP, cone or prism crack; L, lateral crack. Redrawn from Swain MV and Lawn BR (1976) Indentation fracture in brittle rocks and glasses. *International Journal of Rock Mechanics and*

indicating that the Hallet model did not correctly account for F_n either. Using their empirical data they developed a model where F_n is a function of both the effective normal stress (Boulton) and the viscous drag from ice flow toward the bed, but differed in its parameterization of this from the Hallet model. Modeling results suggested that a limited particle size range, 0.007–0.1 m, contributed approximately 90% of the bed friction and that sufficiently large and small (<200 μm) particles tended to become lodged. The lodgement of large clasts may well have implications for the formation of friction cracks and p-forms (see below).

Abrasive wear, through indentation fracture and plowing, is responsible for all bedrock streamlining by mechanical subglacial erosion, and thus contributes to the formation of all the streamlined forms described below.

Quarrying

The effects of abrasion can best be viewed by looking in the direction of ice flow to see the characteristic smoothed bed (Figure 2), whereas the effects of quarrying are best seen looking in the opposite direction, upflow (Figure 7). Quarrying is generally accepted as a two-component system comprising fracture (breakdown of intact rock) and plucking/entrainment (removal) (Boulton, 1974; Iverson, 1995).

Fracture

Fractured lee sides of bedrock protuberances are, in many instances, controlled by preexisting joint systems (Matthes, 1930;

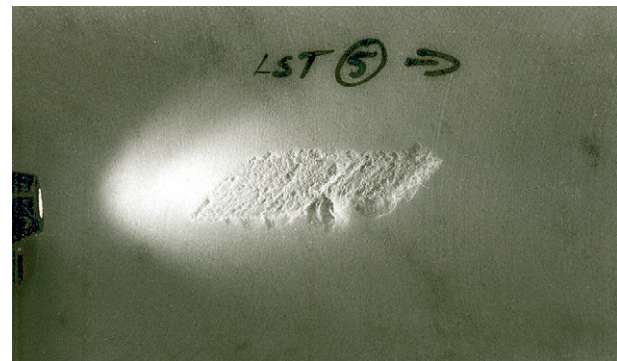


Figure 5 Experimental striation showing how the geometry of the clast–bed contact controls the trailing edge stress pattern and thus the crack geometry. The oblique-to-flow clast–bed contact produced similar oriented cracks, which become concave downflow (like a partial ring fracture) beyond the contact. Photograph by B. R. Rea.

Mining Sciences and Geomechanics Abstracts 13: 311–319, Figure 1; with permission from Elsevier. (c) Stress field developed in the bed beneath a sliding spherical indenter (low plowing angle), viewed from the surface (top) and in cross-section (bottom). Contour values are as in (a) with σ_2 (top) and σ_3 (bottom) stress trajectories which control the likely crack path shown as dashed line. Redrawn from Lawn BR (1967) Partial cone crack formation in a brittle material loaded with a sliding spherical indenter. *Proceedings of the Royal Society, Series A* 299: 307–316, Figure 1; with permission of the Royal Society.

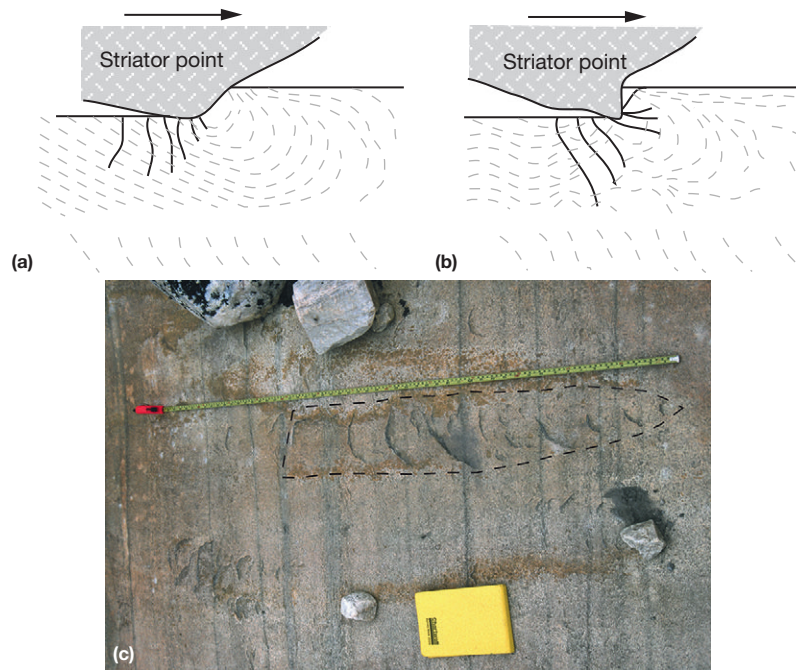


Figure 6 Stresses and likely crack paths developed beneath clasts, making different plowing angles with the bed. (a) 45° , a shallow plowing angle concentrates tensile stresses at the trailing edge contact. (b) 90° , a steep plowing angle concentrates tensile stresses ahead of the contact, forward and downward into the bed, facilitating plowing into the bed with displacement. Reproduced from Iverson NR (1991) Morphology of glacial striae: Implication for abrasion of glacier beds and fault surfaces. *Geological Society of America Bulletin* 103: 1308–1316, with permission of the Geological Society of America. (c) A large type 1 striation containing multiple crescentic gouges (contained within the striation margins – dashed line) along its length. Ice flow was from left to right, the striation is ~ 0.9 m long, 0.15 m wide (max.) and 0.009 m deep (max.). Photograph by B. R. Rea.



Figure 7 The effects of the abrasion and quarrying are more clearly evident when looking in down- and upflow directions, respectively. This view is looking upflow showing clearly the 'rough' appearance in this direction. The stoss-lee forms, resulting from structural control and subglacial erosion, are clearly evident. Ice-flow direction is indicated by the white arrow. The smoothing effect of abrasion is clearly seen in the downflow direction as shown in Figure 2 (the look direction is oblique to ice flow across the rock step with the white arrow). Photograph by B. R. Rea.

Rastas and Seppälä, 1981). Subaerial weathering, either chemical or mechanical, undoubtedly presents a rockhead of open joints and fractures that are easily plucked from the bed. Once these have been removed, further bed erosion proceeds by quarrying of more resistant bedrock. It occurs in massive bedrock (Gordon, 1981; Figure 8) and lodged boulders (Boulton,



Figure 8 Massive bedrock showing the effects of quarrying. The fracture bears no relation to any obvious preexisting or inherent bedrock structure. Reproduced from Gordon JE (1981) Ice-scoured topography and its relationships to bedrock structure and ice movement in parts of northern Scotland and West Greenland. *Geografiska Annaler* 63(A): 55–65, with permission.

1974), indicating that the process may be effective even in the absence of joints and bedding planes. Controlling factors in the process are the bed roughness, the presence of subglacial cavities and water pressure fluctuations within them and inherent flaws and microcracks in the bedrock or boulder. A brief discussion is provided here, and the reader is directed to Iverson (1995) and Hallet (1996) for further details. Iverson

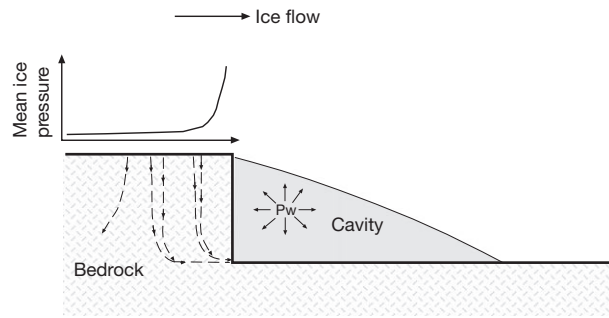


Figure 9 The importance of the stoss-lee form and related water-filled cavity in controlling bed stresses. Cracks are likely to develop parallel to the contours of greatest principal stress (dashed lines), maintaining the stoss-lee form. Rapid drainage of the cavity increases pressure on the bed and may enhance rates of crack growth. Modified from Iverson NR (1991) Potential effects of subglacial water-pressure fluctuations on quarrying. *Journal of Glaciology* 37: 27–36.

(1991b) demonstrated the role of cavities in enhancing stress in the bed and highlighted the potential role of water pressure fluctuations in promoting bedrock fracture. Figure 9 illustrates the projected principal stress directions (and likely crack trajectories) created by a water-filled lee side cavity. Although the stresses calculated by Iverson (1991b) were unlikely to induce critical (instantaneous propagation) crack growth, subcritical (slow propagation) crack growth was likely. Rapid water pressure reductions result in significant increases in the mean bed pressure, thereby increasing the tendency for bed failure. Fracture potential is further enhanced for water-filled cracks as water pressure within them is less responsive than in the cavity (Iverson, 1991b, 1995). Subcritical crack growth is fundamental to the process because it allows bed fracture (quarrying) to proceed under stresses that do not exceed the tensile strength (fracture toughness) of the bedrock (Atkinson, 1982). Loading must be sufficient so that stresses are raised above a minimum threshold, and given this condition and sufficient loading time, subcritical crack growth should result in bed failure. Hallet (1996) investigated the stress state in a series of rock steps superimposed on a horizontal bed (Figure 10) under varying levels of ice–bed separation (i.e., cavity length). As ice–bed separation increases, so does the normal stress supported across the contact areas. More important for bed fracture, tensile stresses are developed in the bed just upglacier from the point of cavity closure, and these drive crack growth and fracture. Note that the stoss-lee form is self-reinforcing in that fractures are likely to grow subvertically down into the bed, maintaining the stepped appearance as the lee faces migrate upflow.

Plucking/Entrainment

For completion of the quarrying process, loose material has to be removed from the bed. This requires displacement of the material from *in situ*, which is easiest for very loose material that can be removed by bulldozing and overriding during the initial ice advance. Robin's (1976) heat pump effect provides a mechanism to 'freeze on' the glacier sole to the bed, but the low shear strength of this bond makes it unlikely that such a

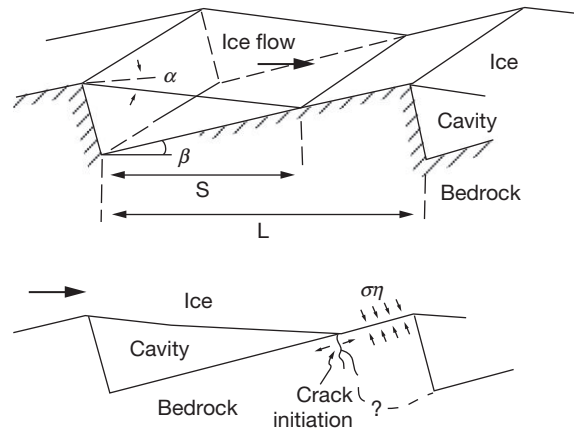


Figure 10 A horizontal bed comprising a series of rock steps with cavities developed in the lee-side. Bed stresses are elevated at the areas of ice–bed contact, and tensile stresses are developed in the bed, just upflow from the closure of each cavity. Reproduced from Hallet B (1996) Glacial quarrying: A simple theoretical model. *Annals of Glaciology* 22: 1–8, with permission of the International Glaciological Society.

mechanism will produce significant amounts of plucking alone. The hydraulic jacking effect of Röthlisberger and Iken (1981) integrates the heat pump concept with water pressure fluctuations in subglacial cavities. Upglacier from the cavity, as water pressure falls, so ice pressure increases. This increased pressure may lead to the expulsion of intercrystalline and regelation meltwater from basal ice and the ice–bed interface. Subsequently, with a rise in cavity water pressure, the mean ice pressure on the bed is reduced (i.e., the water supports more of the ice overburden), raising the pressure melting point, resulting in refreezing of available meltwater. If the expelled water cannot be replaced, for example, from an adjoining cavity, heat conduction from the bedrock leads to freezing on. With the hydraulic jack, ice–bed adhesion occurs under reduced effective stress which should be more effective than the heat pump effect alone (i.e., reduced effective stress reduces friction along bedrock joints and fractures; Iverson, 1995). The only other method for removing material is through the viscous drag induced by ice flow over the bed, which increases with the upglacier dip of the stoss surfaces and ice velocity. However, the joint shear strength (the force resisting removal) also increases with these factors. Rea and Whalley (1996) showed that drag-induced removal of joint-bounded blocks was possible from steeply upglacier dipping bedrock steps in air-filled cavities under thin ice (Figure 11(a) and 11(b)). The squeezing of ice along the lateral and rear bounding joints of the blocks (Figure 11(c)) reduced joint friction and assisted block removal. Sediment injection into bedrock joints and fractures has also been proposed, which would facilitate, in an analogous manner, the plucking of bedrock beneath thin and patchy till cover.

It is generally accepted that quarrying is volumetrically of greater significance than abrasion in terms of total glacial erosion (Iverson, 1995), with abrasion never exceeding 40% of the total erosion (Hallet, 1996). Quarrying is the only source of new abrasive beneath glaciers and ice sheets once preexisting regolith has been removed.

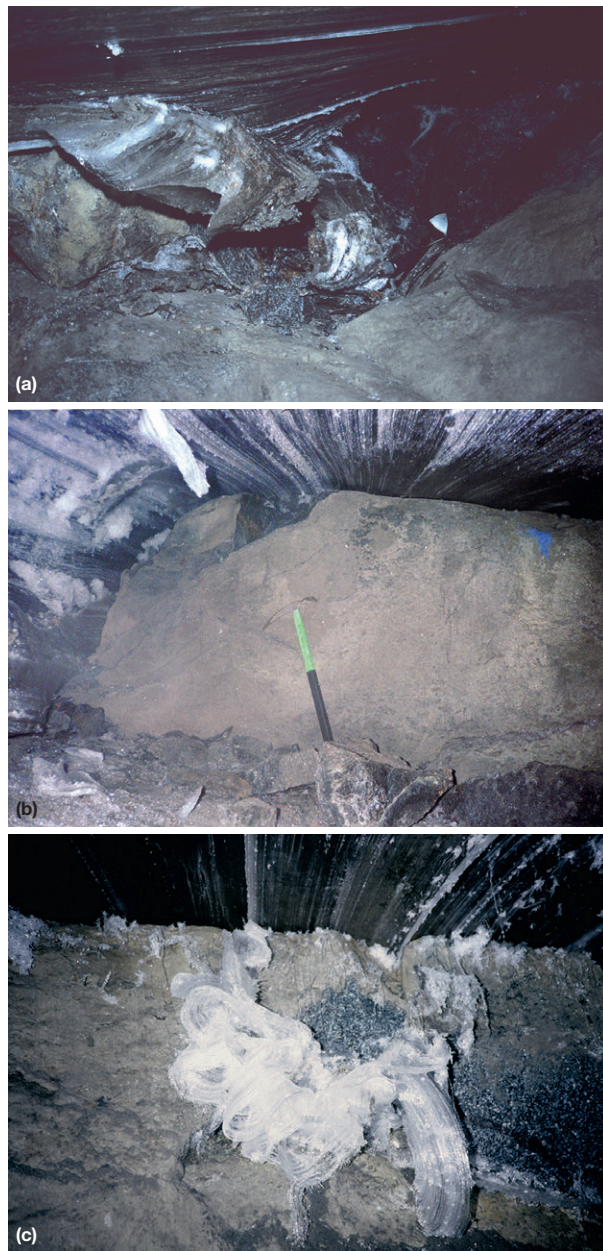


Figure 11 (a) Active quarrying of a block (left side of image) from the lee-side of a roche moutonnée. (b) A lee-side view of a roche moutonnée looking upflow. Note the space at the ice–bed separation point where a joint-controlled block has recently been removed (above 75-cm-long ice axe). (c) Squeezing of ice around a joint-controlled block at the crest of a roche moutonnée. This reduces friction along the joints, facilitating plucking. All images are from cavities beneath Øksfjordjøkelen, North Norway. Photographs by B. R. Rea.

Glacial Erosional Forms

Striae

In accordance with the first descriptions of these forms by Chamberlin (1888), types 1–3 are used throughout the remainder of this article as descriptors of striae morphology (Figure 12(a)). The abrupt end to type 1 striae results from the asperity either breaking off or rotating completely out of

the striation due to increased coupling (shear stress supported) across the clast–bed contact (Figures 6(c) and 12(b)). If, instead of asperity fracture or complete outrotation, the plowing angle decreases due to gradual clast rotation and/or wear of the asperity, the striation will gradually shallow with displacement (following indentation and plowing – type 1), forming a type 2 striation (Figure 3). Type 3 striae are formed by static indentation followed by plowing with a reducing plowing angle (i.e., the second part of a type 2 striation; Figure 12(c)).

Friction Cracks

These are a family of forms that tend to be found on horizontal to upglacier dipping, smoothed abraded bedrock surfaces and include crescentic fractures, lunate fractures, crescentic gouges, and conchoidal fractures (Figure 13(a)). Crescentic fractures (Figure 13(a)(i) and 13(b)) tend to form en echelon, sometimes with decreasing or increasing diameter in the ice-flow direction. These are most easily explained as partial ring fractures formed at the trailing edge contact of a rounded (low plowing angle) sliding asperity (Figure 4(c)). Lunate fractures (Figure 13(a)(ii) and 13(c)) and crescentic gouges (Figure 13(a)(iii) and 13(c)) are assumed to be formed by a similar process, with the differing orientation of the horns resulting from the clast–bed contact geometry. Common to both forms is a shallow down ice-dipping principal fracture (Figure 13(a)(ii,iii)) and intersecting secondary fracture defining the wedge of rock removed. Iverson (1991a) suggested that the primary fracture could form ahead of a plowing asperity (Figure 6(b)), in a manner similar to a Riedel (R)-type shear produced along shearing fault surfaces. The only shallow outward dipping fractures that form beneath asperities are either the lateral crack formed during unloading beneath a static contact (Figure 4) or fractures formed ahead of a steep plowing contact (Figure 6(b)). The geometry of lunate fractures and crescentic gouges allow some conclusions to be drawn regarding their formation. The primary fracture of the lunate fracture has a shallow dip in the direction of ice flow from a concave downflow contact, which is most easily interpreted as a rounded trailing edge contact. The primary crack must then have formed below the asperity–bed contact, analogous to an R-type shear, with the secondary crack forming subsequently as a partial ring fracture under a static or sliding contact beneath the same asperity. The primary fracture of the crescentic gouge has a shallow dip in the direction of ice flow (Figure 13(a)(iii)), from a convex downflow leading edge contact, indicating that it formed ahead of the asperity–bed contact. This could represent either an R-type shear produced ahead of a steep plowing contact (Iverson, 1991a) or a lateral crack produced beyond the contact area of a blunt static contact. The secondary fracture results from either the termination of the asperity plowing (i.e., the crescentic gouge represents a very short type 1 striation) or a partial ring fracture developed under static loading, beneath the same blunt asperity. Experimental evidence of brittle fracture in glass suggests that the orientation may have some relation to the dynamics of the contact (i.e., sliding may produce concave and convex downflow forms, whereas rotating contacts produce only concave downflow forms; Smith, 1984). Conchoidal fractures are

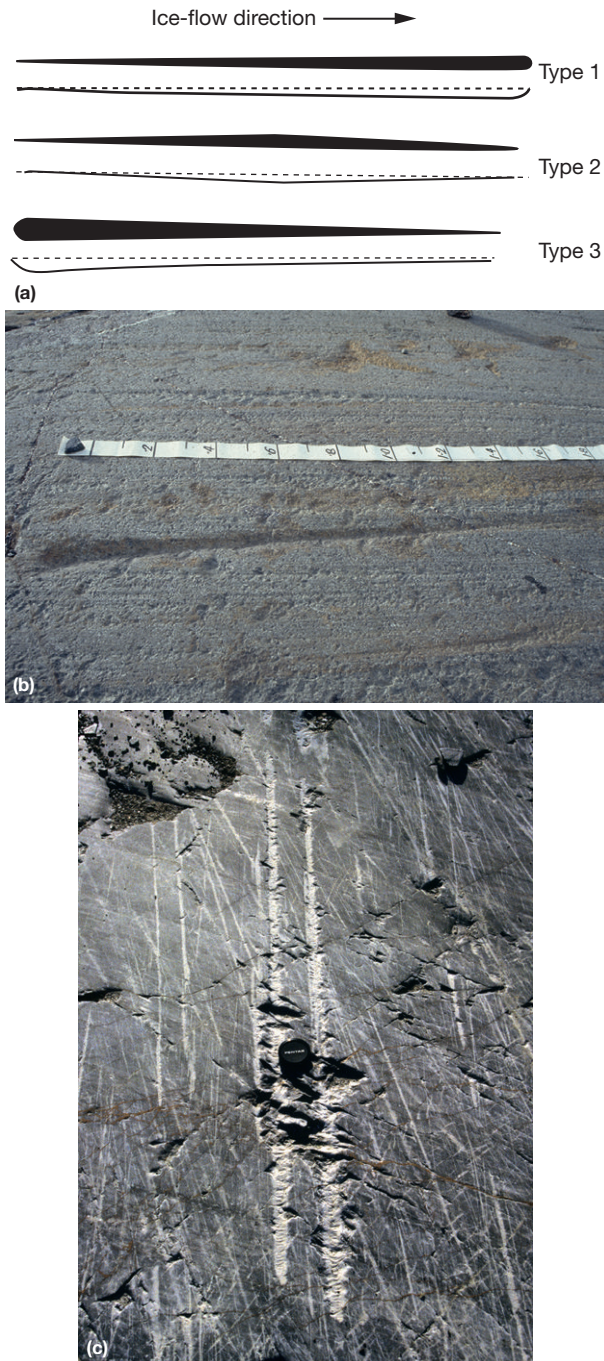


Figure 12 Examples of typical glacial striae. (a) The striae type classification based on the descriptions of Chamberlin (1888). (b) Type 1 striation, widening and deepening until the abrupt termination is reached. Reproduced from Rea BR and Whalley WB (1994) Subglacial observations from Øksfjordjøkelen, North Norway. *Earth Surface Processes and Landforms* 19: 659–673. Copyright John Wiley and Sons Ltd., with permission. (c) Two type 3 striae starting at an abrupt initiation point and shallow and narrow to the end point. Note the concave downflow partial ring fractures in the base of the striae. Photographs by B. R. Rea.

reported less often in the literature but are likely to originate close to the slope break, which they cross-cut, in a manner similar to a lunate fracture.

p-Forms

Dahl (1965) introduced the term plastically molded forms to describe a range of smoothed bedrock erosional features, including transverse, longitudinal, and nondirectional forms. A number of processes have in the past been suggested for the origin of p-forms, but only two are currently advocated. One theory favors a subglacial fluvial origin, with the striae that cover their surface representing a minor 'ornamentation' component in their formation (Kor et al., 1991). The alternative theory views some p-forms as undoubtedly of a subglacial fluvial origin (e.g., potholes), with others resulting from quarrying and abrasion, perhaps with additional fluvial erosion (Boulton, 1974; Goldthwait, 1979; Rea et al., 2000). The quarrying and abrasion origin only is of relevance to this article. Boulton (1974) suggested that streaming of ice around bed obstacles focuses erosion along and downstream of the flanks of the obstacle (Figure 14(a)), leading to the formation of sichelwannen, which could be especially effective in the lee side of smaller bed obstacles (Iverson, 1995). Alternatively, following the modeling of Cohen et al. (2005), large boulders will tend to lodge on the bed and streaming around the flanks of the contact, in a manner similar to that envisaged by Boulton (1974), could, produce a sichelwannen or muschelbruch form. Goldthwait (1979) cited convergent striae into deep limestone grooves and the lack of percussion marks (fluvial) on Kelleys Island, Lake Erie, as evidence for formation by ice abrasion. Rea et al. (2000) presented detailed observations of contemporaneous subglacial multidirectional ice flow and linked these to proglacial bedrock striation patterns, indicating how glacial erosion can exploit preexisting bedrock joint patterns to produce transverse troughs (Figure 14(b)), furrows, and comma forms (Figure 14(c)). It seems that there are some p-forms which are undoubtedly of subglacial fluvial origin, but many others are of a subglacial mechanical origin or polygenetic with a minor to significant input from both processes.

Polish

Table 1 indicates that striae are formed at the centimeter to meter scale, although this is not strictly true because they are found down to the micrometer scale. This small-scale abrasive wear produces a 'polished' appearance to freshly exposed bedrock. The polish may be developed on the bedrock by the gradual smoothing by shearing of clasts and fines (sand and silt) across the bed (without significant indentation/plowing) or within the finer-grained material (rock flour) undergoing shear between clast-bed (Figure 15) and ice-bed contacts (Lee and Rutter, 2004; Rea, 1996). The polish will be short lived if it is developed in the rock floor or longer lived if it is developed on the bedrock. Exposure to subaerial weathering will gradually roughen the surface.

Roches Moutonnées

Roches moutonnées are the classic example of subglacial bedrock erosion, with abrasion dominating on the upstream (stoss) side and quarrying producing the (sub)vertical lee side (Figure 11(b)). The component of ice flow toward the bed is greatest approximately half way up the stoss side (Iverson, 1995), enhancing clast-bed contact stresses and favoring the

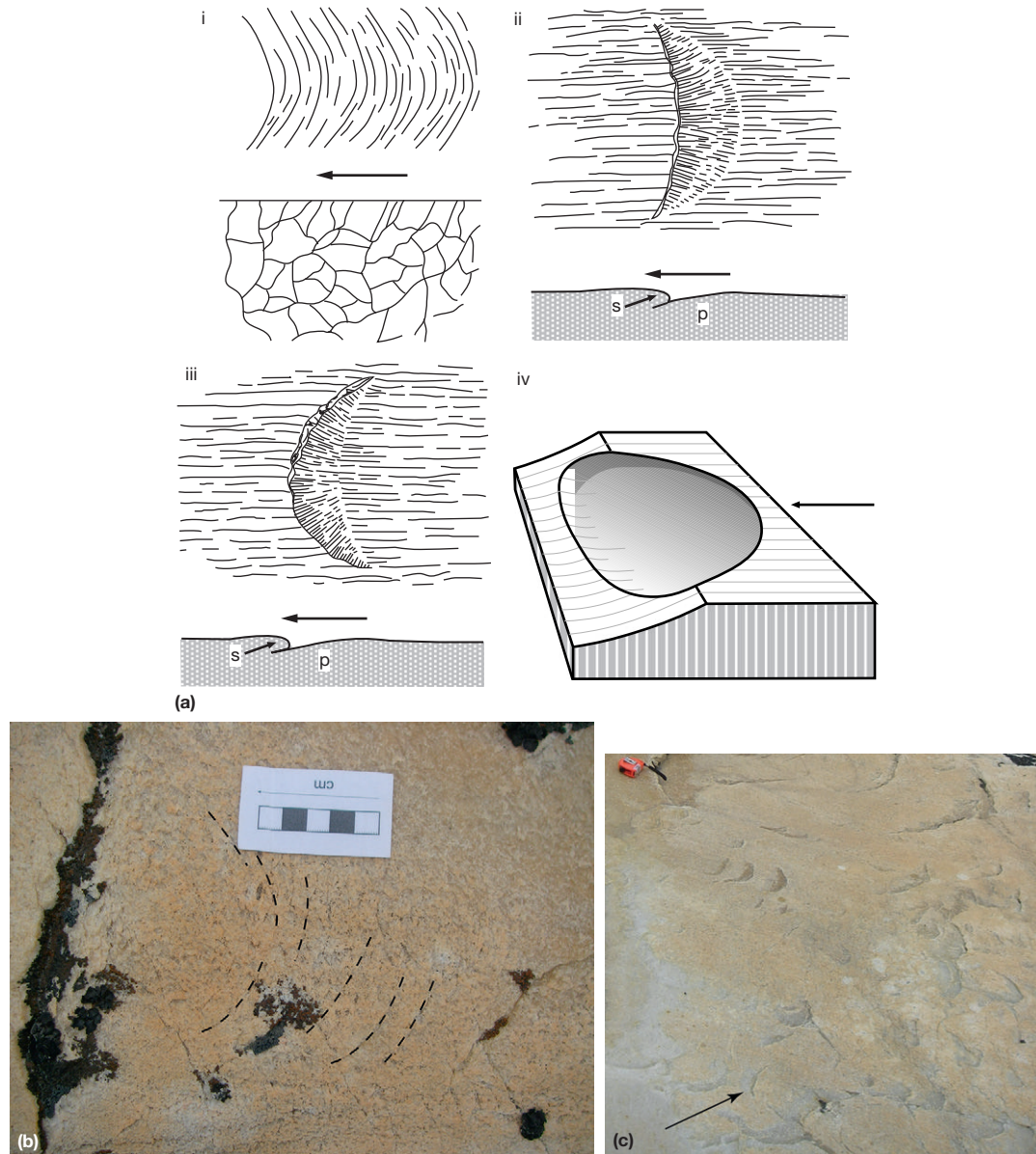


Figure 13 (a) Morphology of friction cracks, comprising (i) crescentic fractures, (ii) lunate fractures, (iii) crescentic gouges, and (iv) conchoidal fractures (p, primary crack and s, secondary crack). Redrawn from Embleton C and King CAM (1975) *Glacial Geomorphology*. London: Edward Arnold. (b) A series of crescentic fractures on quartzite bedrock (some are highlighted by dashed lines) where ice flow was from right to left. (c) Exposure showing multiple crescentic gouges and one lunate fracture (arrowed) where ice flow was approximately top left to mid right. Photographs by B. R. Rea.

production of deep striae (Rastas and Seppälä, 1981). Bed-normal stresses are lower along the flanks and so striae should be shallower (Rastas and Seppälä, 1981). Bedrock jointing often controls the orientation of the lee sides (Matthes, 1930; Rastas and Seppälä, 1981; Rea et al., 2000). Above however, it has been shown that the presence of the (sub)vertical lee sides focus bed stresses, facilitating quarrying even in the absence of preexisting joints (Figures 9 and 10). Friction cracks are more frequently found toward the top of roches moutonnées (Rastas and Seppälä, 1981). This may seem counterintuitive because F_n is highest halfway up the stoss sides; however, the largest clasts are likely to make contact only with the tops of roches

moutonnées (Iverson, 1995) due to the shadowing effect of successive roche moutonnée (rock steps; Figure 10). As indicated in Table 1, roches moutonnées (including flyggbergs, which are large-scale asymmetrical hills) cover a large range of scales, such that smaller roches moutonnées may be superimposed on the stoss sides of larger ones.

Whalebacks and Rock Drumlins

If bedrock quarrying is suppressed and abrasion still occurs, elongate and smoothed bedrock protuberances may be produced. Where the form is asymmetric with a steeper stoss side

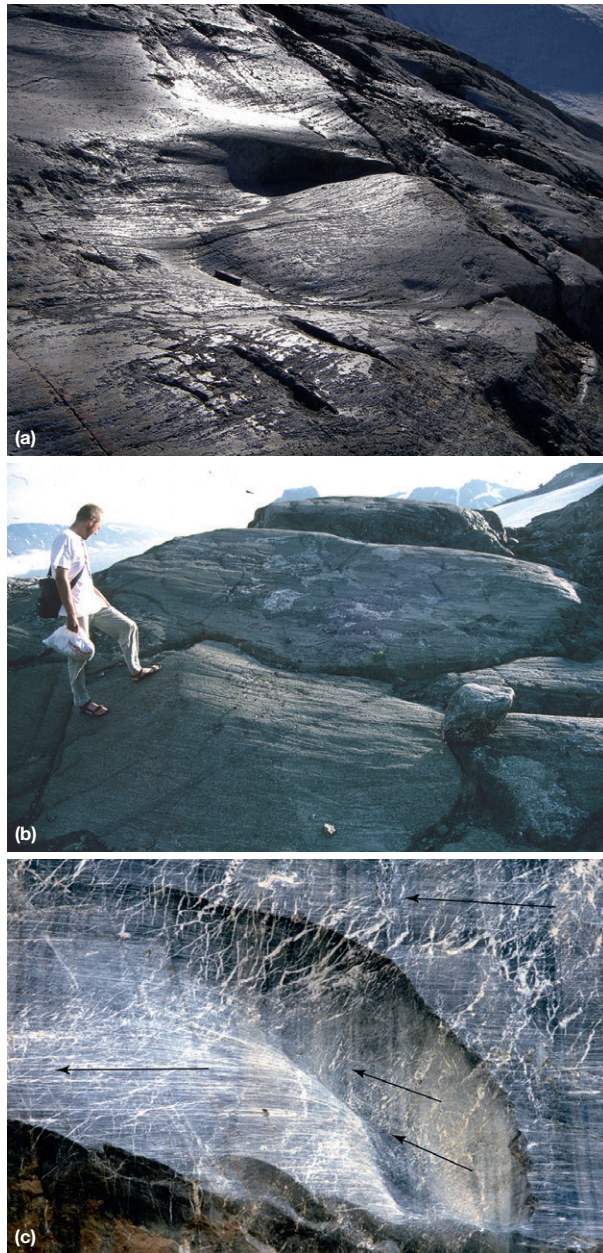


Figure 14 (a) Bedrock protuberance with evidence for enhanced erosion along the flanks as proposed by Boulton (1974; note penknife for scale). Photographs by B. R. Rea. (b) Joint-controlled channel (containing conforming striae) on a stoss surface. Photograph by B.R. Rea. (c) A comma p-form with conformable striae. (b) and (c) Reproduced from Rea BR, Evans DJA, Dixon TS, and Whalley WB (2000) Contemporaneous, localized, basal ice-flow variations: Implications for bedrock erosion and the origin of p-forms. *Journal of Glaciology* 46: 470–476, with permission of the International Glaciological Society.

(like a drumlin), it is known as a rock drumlin (Figure 16(a)), and the more symmetrical forms are known as whalebacks (Figure 16(b)). The suppression of quarrying is likely linked to the absence of lee side cavities, intransient basal water pressures, and mechanically strong, massive bedrock (Evans, 1996; Gordon, 1981). High effective stresses (thick ice and low basal

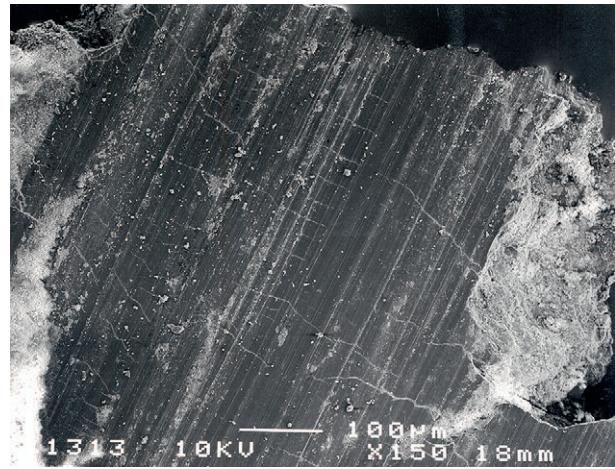


Figure 15 Polish developed between a shearing clast and the bed in the bottom of an experimental striation. Note the striated surface of the polish and the granular composition below (right side of micrograph). Micrograph by B. R. Rea.

water pressures) and low sliding velocities suppress ice–bed separation; thus, quarrying by either of the mechanisms discussed previously (Figures 9 and 10) is unlikely. It has been suggested that whalebacks are best developed beneath thick fast-flowing topographically controlled ice streams so that ice–bed separation is suppressed (Evans, 1996; Roberts and Long, 2005).

Crag and Tail

As indicated in Table 1, crag and tails cover a significant scale range and are represented in microforms as mini crag and tails. At all scales, they comprise a resistant core forming the steep stoss side, with an elongate tail formed in the lee (Figure 17(a)). The classic example is in Edinburgh, UK, with Edinburgh Castle built on the crag and the Royal Mile being the tail (Benn and Evans, 1998). The crag is a volcanic plug and the tail comprises less resistant sedimentary rocks so that the tail is protected in the lee of the resistant plug in a direction paralleling the direction of former ice flow. In many instances, the tail is composed of unlithified sediments, deposited in the lee of the crag, so strictly speaking, these are not true erosional forms.

At a smaller scale, changes in bedrock structure or composition may produce a protective crag. At the miniscale, resistant inclusions such as chert nodules in carbonate rocks (Figure 17(b)) form the crag and there are often troughs cut into the bed around the crag, making a sickle-shaped depression. In less well-developed situations, this is likely to resemble a sichelwannen.

Applications in Quaternary Science

All of the forms described previously provide unequivocal evidence for the presence of former warm-based glaciers and ice sheets. All of the forms associated with abrasive wear indicate that the ice contained debris in traction across the bed. Roches moutonnées and crag and tail provide clear unidirectional ice-flow evidence, whereas other streamlined



Figure 16 (a) Rock drumlin that has undergone subsequent lee-side quarrying. The original surface is shown as a dashed white line. Ice flow was from right to left. Photograph by B. R. Rea. (b) Whaleback from the southern edge of Jakobshavns Isfjord, west Greenland. Ice flow is from right to left. Photograph by D. H. Roberts.

forms provide equivocal or bidirectional ice-flow directions (i.e., may be streamlined by ice flow from one of two directions) and require additional information (e.g., erratic trains) to define absolutely the former ice-flow direction. Crossing striae (or striae directions) are regularly observed in the field and require careful interpretation (Figure 18). In some instances, they represent contemporaneous ice flow (Rea et al., 2000), and in others they represent changing dynamics and flow directions within a single or multiple glaciations. If cross-cutting relationships can be defined at the outcrop scale it may be possible to determine a relative chronology (Figure 18). It has been suggested, for example, if fine striae are found cross-cutting deeper striae, then the deeper striae are older (Sharp et al., 1989), but this only holds if the fine striae can be traced across the base of larger striae. However, if only multiple direction relationships are recorded on a bedrock surface, care must be taken. For example, if the direction of fine striae on the lee sides of a roche moutonnée are at odds with the direction of stoss side striae, they are not necessarily of a different age. The fine striae may be older having been preserved/protected in the lee sides of the roche moutonnée, or they are synchronous and represent contemporaneous

multidirectional ice flow (Rea et al., 2000). The primary fracture in lunate fractures and crescentic gouges (Figure 13(a)(ii, iii) and 13(c-e)) invariably dips in the direction of ice flow and so may be used to reconstruct former ice-flow directions. Wintges (1985) performed a photogrammetric survey of friction cracks, differentiating populations formed under multiple phases of glaciation. A most interesting result was an apparent nonlinear dependence of crack size on ice thickness. Presumably, this is at least partially due to its contribution to the effective pressure, but components of the ice dynamics, such as basal melt rate will also contribute to F_n (Cohen et al., 2005), which is the major control on their formation. The origin of p-forms is still disputed, and linkages with ice dynamics, beyond the ice being warm based, have yet to be made.

The presence/absence or juxtaposition of roches moutonnées and whalebacks provide a potentially fruitful line of inquiry for determining, at least at a broad scale, the dynamics of former ice masses. Evans (1996) and Roberts and Long (2005) suggest that high effective pressures developed beneath thick, topographically confined, 'ice streams' led to cavity suppression, prohibiting quarrying and allowing abrasion to dominate. Cavity suppression will undoubtedly limit quarrying, however, for

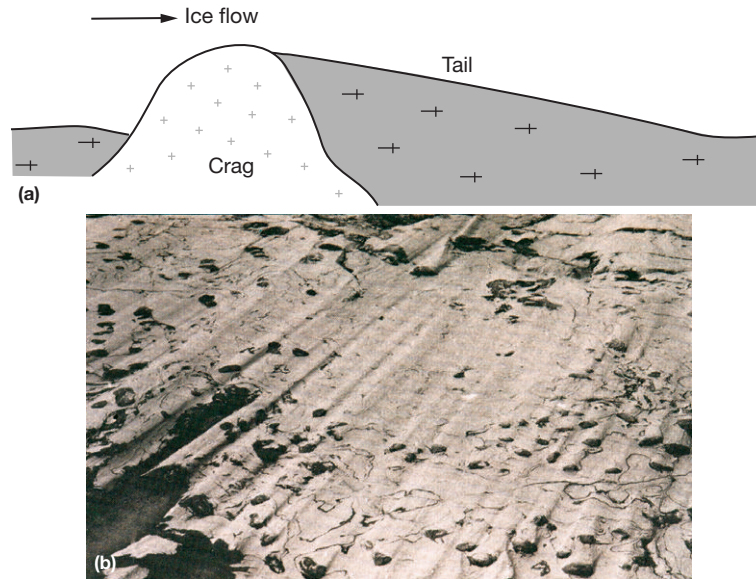


Figure 17 (a) Generalized configuration of a crag and tail. The simple requirement is a harder crag forming the stoss end and less resistant surrounding rock forming the tail. (b) Mini crag and tails (rat tails) formed by the protection of weaker dolomite in the lee of chert nodules. Reproduced from Prest VK (1983) Canada's Heritage of Glacial Features. Miscellaneous Report No. 28. Vancouver: Geological Survey of Canada, with permission of the Geological Survey of Canada.



Figure 18 (a) Bedrock exposure showing three generations of striae. Detailed inspection of the striae relationship at the outcrop scale failed to yield conclusive evidence for their age relationship. However, based on an interpretation of increasing topographic control during deglaciation the chronology is 1 (oldest) through 3 (youngest). (b) The large deep striation (parallel to the knife) appears to cross-cut the top right-bottom left trending striae, but again proving a cross-cutting relationship from the outcrop alone is nontrivial. Photographs by B. R. Rea.

abrasion to occur there still needs to be sufficient force pressing clasts against the bed in the lee side of the whaleback. This is somewhat problematic for the 'symmetry' of whalebacks as ice flow is directed away from the bed in the lee side which will reduce all components contributing to F_n so the abrasion rate should be lower in the lee than on the stoss. Therefore, it is theoretically difficult to have high rates of lee side abrasion, but in order for a symmetrical whaleback to form, stoss and lee side abrasion rates must be comparable, or the form is an inherited landform, which seems unlikely. As ice thins during deglaciation, effective pressures become transient and quarrying may proceed, producing superimposed roches moutonnées (Roberts and Long 2005).

It is through the use of combined data sets that the best reconstructions of former ice dynamics are made. An excellent example is that of Sharp et al. (1989), who integrated detailed bedrock erosional information, including striae directions and depths, frequency and orientation of quarried faces, ice thickness estimates, rock fracture relationships, and glaciological theory, to interpret the dynamics and erosional styles beneath ice filling Llyn Llydaw, North Wales, during the late Devensian glaciation (ice sheet) and the Loch Lomond Stadial (Younger Dryas; valley glacier) glaciation.

Further advancement in our understanding of subglacial erosional process will undoubtedly lead to the use of micro- to macroscale forms of glacial erosion to allow more rigorous and

quantitative reconstructions of the dynamics of former glaciation. A better understanding of the formation/evolution of whalebacks in particular, probably holds important information regarding the dynamics of former ice streams and the dichotomy of apparent fast ice flow with an absence of lee side ice-bed separation and equity of stoss and lee side abrasion rates.

See also: **Glacial Landforms:** Glacial Landsystems; Glaciofluvial Landforms of Erosion. **Glacial Landforms, Erosional Features:** Major Scale Forms. **Glacial Landforms, Sediments:** Glacial Erratics and Till Dispersal Indicators.

References

- Agassiz L (1831) On the polished and striated surfaces of the rocks which form the beds of glaciers in the Alps. *Proceedings of the Geological Society of London* 3: 321–322.
- Atkinson BK (1982) Subcritical crack propagation in rocks: Theory, experimental results and applications. *Journal of Structural Geology* 4: 41–56.
- Benn DI and Evans DJA (1998) *Glaciers and Glaciation*. London: Edward Arnold.
- Boulton GS (1974) Processes and patterns of glacial erosion. In: Coates DR (ed.) *Glacial Geomorphology*, pp. 41–87. New York: State University of New York.
- Chamberlin TC (1888) The rock scorings of the great ice invasions. In: *Geological Survey of America Annual Report*, No. 7, pp. 155–248.
- Cohen D, Iverson NR, Hooyer TS, Fischer UH, Jackson M, and Moore PL (2005) Debris-friction of hard-bedded glaciers. *Journal of Geophysical Research* 110: F02007. <http://dx.doi.org/10.1029/2004JF000228>.
- Dahl R (1965) Plastically sculpted detail forms on rock surfaces in northern Nordland, Norway. *Geografiska Annaler* 47(A): 83–140.
- Evans IS (1996) Abraded rock landforms (whalebacks) developed under ice streams in mountain areas. *Annals of Glaciology* 22: 9–16.
- Forbes JD (1843) *Travels Through the Alps of Savoy*. Edinburgh: Adam and Charles Black.
- Frank CF and Lawn BR (1967) On the theory of Hertzian fracture. *Proceedings of the Royal Society, Series A* 299: 291–306.
- Goldthwait RP (1979) Giant grooves made by concentrated basal ice streams. *Journal of Glaciology* 23: 297–307.
- Gordon JE (1981) Ice-scoured topography and its relationships to bedrock structure and ice movement in parts of northern Scotland and West Greenland. *Geografiska Annaler* 63(A): 55–65.
- Hallet B (1979) A theoretical model of glacial abrasion. *Journal of Glaciology* 23: 39–50.
- Hallet B (1981) Glacial abrasion and sliding: Their dependence on the debris concentration in basal ice. *Annals of Glaciology* 2: 23–28.
- Hallet B (1996) Glacial quarrying: A simple theoretical model. *Annals of Glaciology* 22: 1–8.
- Hindmarsh RCA (1996) Cavities and the effective pressure between abrading clasts and the bedrock. *Annals of Glaciology* 22: 32–40.
- Iverson NR (1991a) Morphology of glacial striae: Implication for abrasion of glacier beds and fault surfaces. *Geological Society of America Bulletin* 103: 1308–1316.
- Iverson NR (1991b) Potential effects of subglacial water-pressure fluctuations on quarrying. *Journal of Glaciology* 37: 27–36.
- Iverson NR (1995) Processes of erosion. In: Menzies J (ed.) *Glacial Environments: Vol. 1. Modern Glacial Environments. Processes, Dynamics and Sediments*, pp. 241–259. London: Butterworth Heinemann.
- Kor PSG, Shaw J, and Sharpe DR (1991) Erosion of bedrock by subglacial meltwater, Georgian Bay, Ontario: A regional view. *Canadian Journal of Earth Sciences* 28: 623–642.
- Lawn BR (1967) Partial cone crack formation in a brittle material loaded with a sliding spherical indenter. *Proceedings of the Royal Society, Series A* 299: 307–316.
- Lee AAG and Rutter EH (2004) Experimental rock-on-rock frictional wear: Application to subglacial abrasion. *Journal of Geophysical Research* 109: B0902. <http://dx.doi.org/10.1029/2004JB003359>.
- Matthes FE (1930) *Geological History of Yosemite Valley, Professional Paper No. 160*. Sierra Nevada, CA: Geological Survey of America.
- Prest VK (1983) *Canada's Heritage of Glacial Features. Miscellaneous Report No. 28*. Vancouver: Geological Survey of Canada.
- Rastas J and Seppälä M (1981) Rock jointing and abrasion forms in roche moutonnées, SW Finland. *Annals of Glaciology* 2: 159–163.
- Rea BR (1996) A short note on the experimental production of a glacially polished surface. *Glacial Geology and Geomorphology*, SC01 (electronic journal).
- Rea BR, Evans DJA, Dixon TS, and Whalley WB (2000) Contemporaneous, localized, basal ice-flow variations: Implications for bedrock erosion and the origin of p-forms. *Journal of Glaciology* 46: 470–476.
- Rea BR and Whalley WB (1994) Subglacial observations from Øksfjordjøkelen, North Norway. *Earth Surface Processes and Landforms* 19: 659–673.
- Rea BR and Whalley WB (1996) An investigation of the role of bedrock topography, ice dynamics, and pre-glacial weathering in controlling subglacial erosion beneath a high latitude, maritime icefield. *Annals of Glaciology* 22: 121–125.
- Roberts DH and Long AJ (2005) Streamlined bedrock terrain and fast ice flow, Jakobshavn Isbrae, West Greenland: Implications for ice stream and ice sheet dynamics. *Boreas* 34: 25–42.
- Robin GDQ (1976) Is the basal ice of a temperate glacier at the pressure melting point? *Journal of Glaciology* 16: 183–195.
- Röthlisberger H and Iken A (1981) Plucking as an effect of water-pressure variations at the glacier bed. *Annals of Glaciology* 2: 57–62.
- Sharp MJ, Dowdeswell JA, and Gemmell JC (1989) Reconstructing past glacier dynamics and erosion from glacial geomorphic evidence: Snowdon, North Wales. *Journal of Quaternary Science* 4: 115–130.
- Smith JM (1984) Experiments relating to the fracture of bedrock at the ice–rock interface. *Journal of Glaciology* 30: 123–125.
- Swain MV and Lawn BR (1976) Indentation fracture in brittle rocks and glasses. *International Journal of Rock Mechanics and Mining Sciences and Geomechanics Abstracts* 13: 311–319.
- Wintges T (1985) Studies on crescentic fractures and crescentic gouges with the help of close-range photogrammetry. *Journal of Glaciology* 31: 340–349.

GLACIAL LANDFORMS, ICE SHEETS

Contents

Growth and Decay

Evidence of Glacier and Ice Sheet Extent

Evidence of Glacier Flow Directions

Paleo-ELAs

Trimlines and Paleonunataks

Growth and Decay

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Paleoenvironmental records reveal the climate of the past 2 million years to be characterized by large-scale oscillations in global temperature. Short periods of relative warmth (interglacials, which last for several thousand years) are separated by lengthy periods of cooling (glacial periods, which last for several tens of thousands of years) followed by rapid warming and then return to an interglacial state. The records also show that global ice volume, sea level, and atmospheric composition are intrinsically linked with these temperature changes (Petit et al., 1999). The climate oscillations have three distinct frequencies, which relate strongly to cyclical variations in Earth's orbit around the Sun (as predicted by Croll in the nineteenth century and later refined by Milankovitch in the twentieth century). This association has led to a consensus among glaciologists that ice ages (i.e., the growth and decay of ice sheets) are forced by orbital variations. The consequence of this view is that significant modification to Earth's climate results from external influences, and that such influences are inevitable in the future. However, because the percentage changes in solar inputs resulting from orbital variations are low, it is unreasonable to expect that orbital variations alone can explain the growth and decay of huge Quaternary ice sheets. In other words, solar forcing of ice sheet behavior needs to be amplified. Such amplification is thought to result from a number of 'feedback' systems operating among a variety of Earth surface (or near surface) processes (Figure 1). Comprehension of glacial feedback mechanisms requires a holistic appreciation of these processes and their interaction.

The last glacial cycle is characterized by a general increase in global ice volume from the Eemian interglacial ~125 ka, when sea level was ~3 m higher than at present and there was less ice in the world than today, to the Last Glacial Maximum (LGM) ~21 ka years ago (Figure 2), when sea level was 120 m lower. In between, there were several global ice volume peaks, corresponding to solar insolation minima. By the early

Holocene, ~8 ka, the distribution of ice about Earth was fairly similar to today, with the notable exception of the West Antarctic Ice Sheet, which took until the mid-Holocene to retreat to its current position.

Glacier Mass Balance, Equilibrium Line Altitudes, and the Growth and Decay of Terrestrial Ice Sheets

Ice sheets build mass through several accumulation processes (Paterson, 1994; Figure 3). In maritime climates, direct snowfall contributes the bulk of accumulation. In polar desert environments, on the other hand, accumulation can be restricted to the freezing of condensation precipitated from air on the ice surface (Rignot and Thomas, 2002). After a number of years (several tens of years at the center of large ice sheets), the surface material is compressed and turned to glacial ice. Ice accumulates across the highest, coldest regions of an ice mass, known as the accumulation zone, and is lost at the lower, warmer end of the ice mass within the ablation zone. If an ice mass terminates in water, ice can also be lost from marine melting (at the ice margin if grounded, or under the ice if floating) and iceberg calving. The line between the ablation and accumulation zones is called the equilibrium line, and its elevation is referred to as the equilibrium line altitude (ELA). The position of the ELA dictates the size of the accumulation and ablation zones. The process of ice flow acts to transfer mass from the accumulation zone into the ablation zone. An ice mass is said to be 'in balance' when the volume of ice accumulated equals the volume ablated and the volume transferred beneath the ELA. In this condition, in which the ice mass remains stable, neither growing nor shrinking, the ELA is at a position that allows this 'steady state' to exist. If accumulation exceeds ablation, then the mass balance will be positive and the glacier will grow. On the other hand, if the ablation is larger

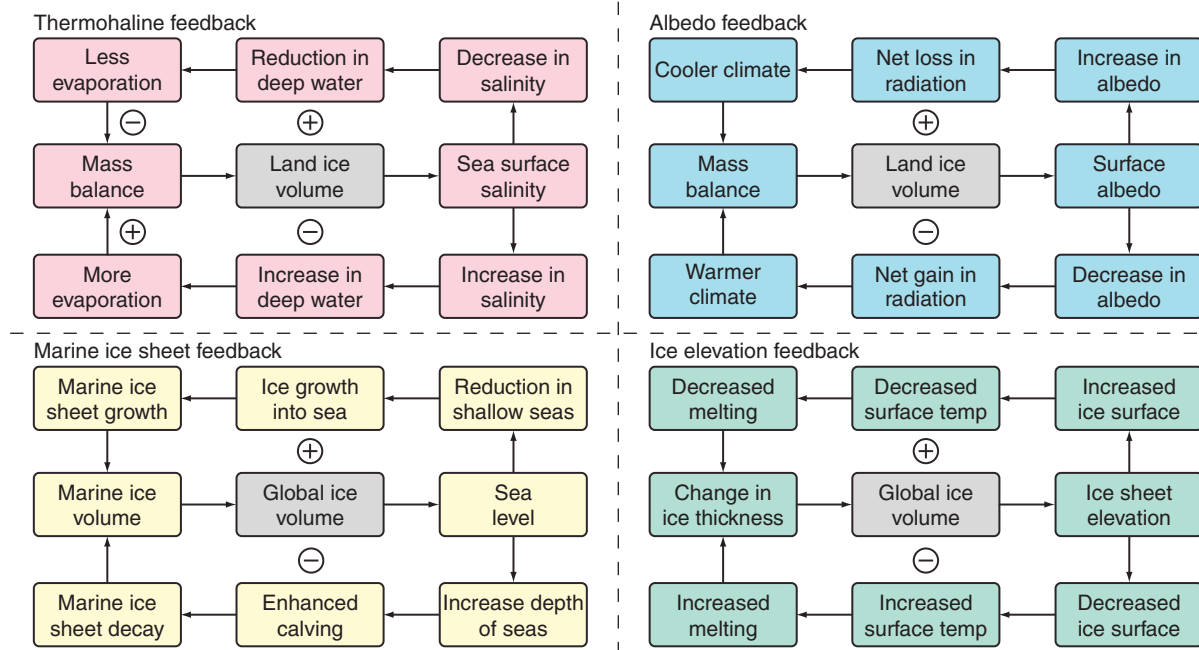


Figure 1 Feedback processes involved in the growth and decay of ice sheets: thermohaline circulation feedbacks, ice sheet surface albedo feedbacks, marine ice sheet and sea-level feedbacks, and ice sheet elevation feedbacks.

than accumulation, the mass balance becomes negative and ice recession occurs. Generally, ice accumulates in the winter and ablates in the summer. In most cases, when glacier mass balance is positive, the summer melt season is cooler, and the ELA is lower. Conversely, warmer summer conditions increase the ELA, resulting in more melting, and if this continues for several years, the ice mass terminus will retreat to high elevation. This simple relation between summer air temperature and mass balance holds as much for large terrestrial ice sheets such as in Greenland as it does for small valley glaciers.

For nonglaciated terrain to become covered with ice, the ELA needs to be lower than the highest topography. This can best be achieved by reducing summer temperatures, thus allowing cirque and small mountain glaciers to form in the first instance. As the ELA is lowered further, the size of the ice masses may increase, and glaciers may merge as single large ice bodies. In this way, large terrestrial ice sheets may be established. Higher summer air temperatures will act to raise the ELA, resulting in increased ablation and ice retreat. Thus, the growth and decay of ice sheets on land can be related to the position of the ELA and, consequently, to summer temperatures.

Because orbital forcing of solar radiation changes is too small to account solely for the growth of continental ice sheets at the LGM, other mechanisms that amplify the solar effects on air temperature are required.

Sea Level and the Growth and Decay of Marine-Based Ice Sheets

Reductions in ELAs may allow ice from land to flow to the sea. Once at the coast, ice flow into the sea will be curtailed by iceberg calving and melting at the ice–water interface. The formation of marine-based ice masses (e.g., the West Antarctic

Ice Sheet) is thought to be assisted by sea-level fall in two ways (Siebert, 2001). First, as a result of sea-level lowering, the shallowest regions of continental seas may become subaerially exposed, after which terrestrial ice sheet formation may take place. Second, as modern glaciers terminating and grounded in water experience loss of ice from iceberg calving at a rate linked to the water depth, sea-level fall may result in a reduction in the rate of iceberg calving of such glaciers and, hence, their growth into formerly deeper water. Sea-level reduction during an Ice Age is caused by an increase in the global ice volume (i.e., the transfer of water from oceans to the continent). Lowering sea level causes marine ice sheets to grow, which results in further reduction in sea level (although marine ice sheets displace part of their weight in water, and so their effect on sea level is less than that of ice sheets on land). A simple positive feedback process is thus defined by which ice growth may result in further growth. Since it is not likely that marine ice sheets can initiate sea-level fall, their growth must be a consequence of the expansion of terrestrial ice masses.

During deglaciation, however, marine ice sheets may experience decay at least in part because of sea-level rise for two reasons. First, as sea levels increase, so too may the rates of calving at grounded ice margins. Second, increasing sea levels result in decreasing friction at the base of marine ice sheets and, hence, enhanced flow of ice to the margin. Another feedback is thus established in which sea-level rise leads to marine ice sheet decay and, in turn, further sea-level rise.

A classic example of how the growth and decay of marine ice sheets are influenced by sea-level changes is the late Weichselian Eurasian ice sheet, which at its maximum occupied the whole of the Barents Sea. During ice sheet growth between 30 ka and the LGM, ice flowed from land masses such as Svalbard and Scandinavia into the shallowing continental seas, forming a large continuous ice mass across the Eurasian Arctic.

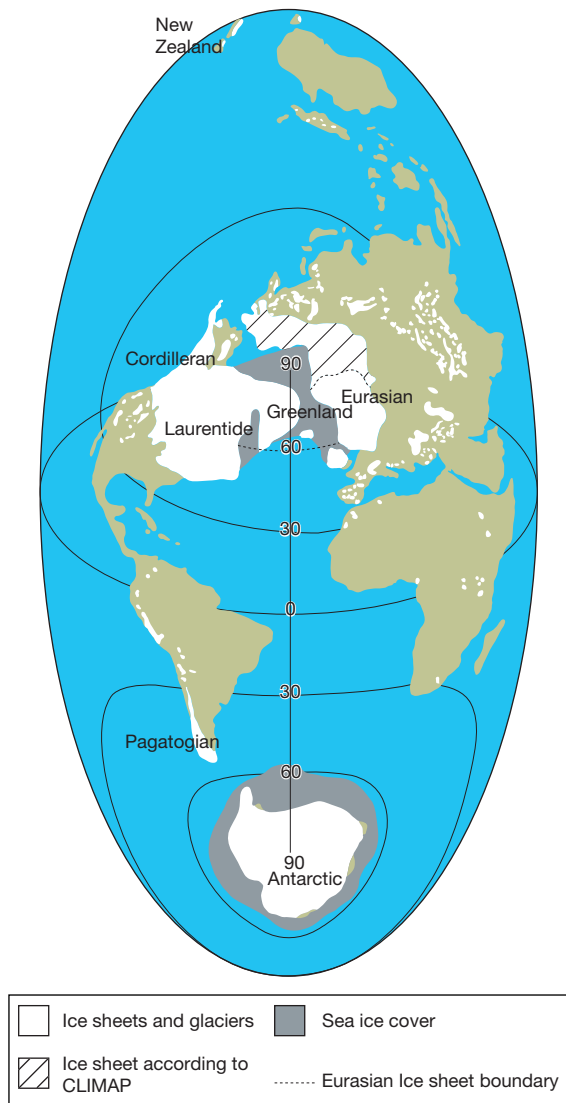


Figure 2 Ice sheets of the Last Glacial Maximum. Note that the size of the Eurasian Ice Sheet across northern Europe and Arctic Russia is shown in both maximum (after the CLIMAP project of the 1970s) and minimum (after the QUEEN project of the 1990s) configurations. The locations of the Antarctic, Greenland, Laurentide, Cordilleran, Eurasian, Patagonian, and New Zealand ice sheets are provided. Adapted from Broecker WS and Denton GH (1990) The role of ocean–atmosphere reorganisations in glacial cycles. *Quaternary Science Reviews* 9: 305–341.

Deglaciation occurred first within the deepest marine sections of the ice sheet soon after the LGM, and the ice front quickly migrated back to the coastal regions, thus rapidly deglaciating the Barents Sea. Numerical modeling results of the growth and decay of the Eurasian ice sheet are provided in [Figure 4](#).

Surface Albedo, Solar Radiation Budget, and Ice Sheet Fluctuations

The albedo of a surface is a measure of its reflectivity to solar radiation ([Paterson, 1994](#)). If the albedo of a surface is high,

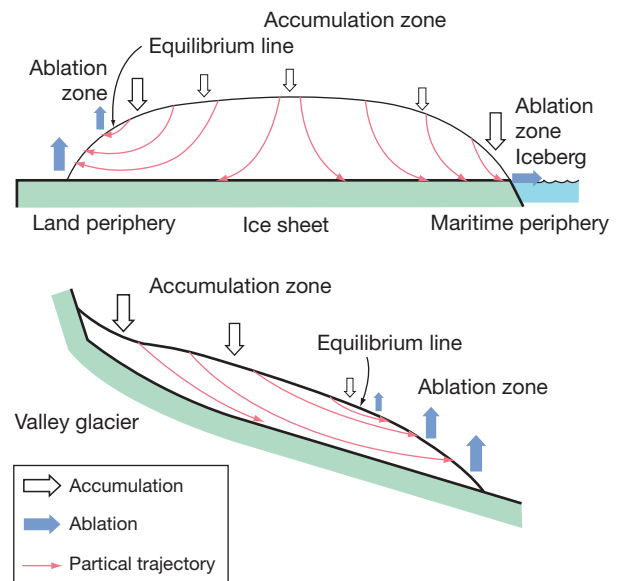


Figure 3 (Top) Ice sheet and (bottom) glacier, mass balance, and flow.

more radiation will be reflected than if the albedo is low. Hence, the average albedo of the planet dictates the levels of solar radiation held and reflected at its surface. Fresh snow has a particularly high albedo. This reflective property results in amplification of the solar forcing of air temperature. As snow fields and ice sheets expand as a result of a reduction in the solar insolation-induced air temperature, the surface albedo will be increased, and the net amount of solar radiation held by the Earth will be reduced. Air temperature will be reduced further, resulting in glacial expansion. At the LGM, when large sections of the Northern Hemisphere were permanently covered with snow and ice, the surface albedo of the planet, and therefore the level of radiation retained, was significantly different from that of today (~ 0.22 compared to ~ 0.14 ; D. Lunt, personal communication). Therefore, through adjustment to the Earth's albedo, ice sheets contributed to global cooling and, hence, their own existence.

Conversely, during deglaciation, this feedback system works to accelerate ice sheet decay. When air temperatures increase and snow and ice melt, the reflectivity of the Earth's surface decreases and, hence, solar radiation is retained. The increase in net radiation leads to increased temperatures and to more ice sheet decay, further lowering the albedo.

Atmospheric Carbon Dioxide Levels and Ice Sheet Changes

Atmospheric carbon dioxide (CO_2) has an impact on climate because it absorbs/reflects longwave radiation emitted from the Earth's surface (i.e., the greenhouse effect). As air bubbles trapped in the ancient ice of deep ice cores demonstrate, when the air gets colder, atmospheric CO_2 is depleted. As the atmosphere cools in summers at the beginning of a glacial cycle (due to solar insolation changes), the CO_2 level drops and, as a result, so too does the absorption of longwave radiation. Hence, heat is lost from the Earth, yielding further cooling. During deglaciation, warmer temperatures are associated with increased levels of atmospheric CO_2 and an enhanced 'greenhouse' effect.

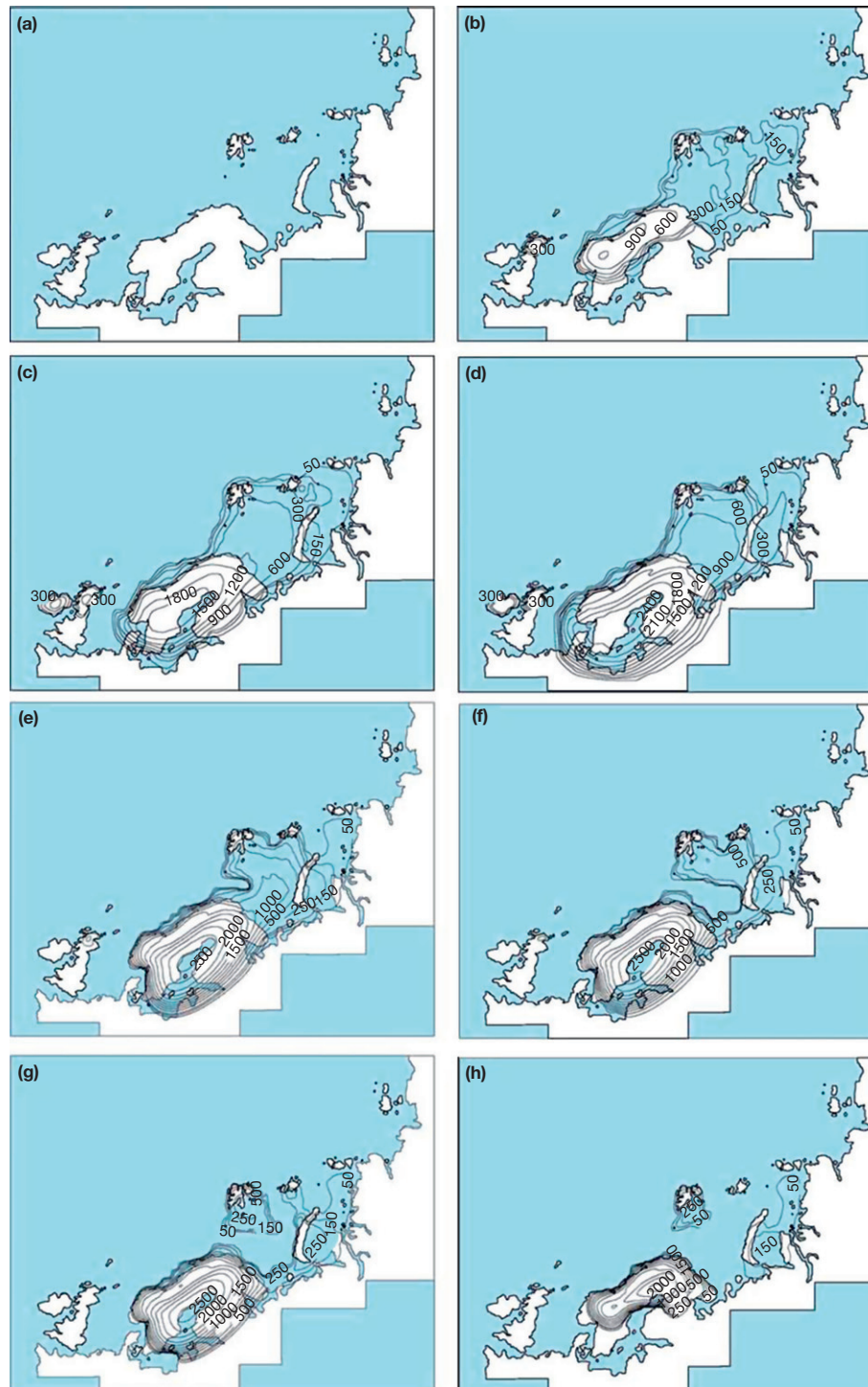


Figure 4 Growth and decay of ice in the Eurasian Arctic during the late Weichselian, as calculated by a numerical ice sheet model (Siegert and Dowdeswell, 2002; Siegert et al., 1999). Maps show the thickness of ice at the following times: (a) 30 ka, (b) 25 ka, (c) 20 ka, (d) 15 ka, (e) 14 ka, (f) 13 ka, (g) 12 ka, and (h) 11 ka. Contours are given in meters.

The reasons why CO_2 varies over glacial cycles are not well understood (Shackleton, 2000). Clearly, biospheric variations will lead to CO_2 alteration. One theory implicates the growth of Antarctic sea ice extent during glacial times, which covered a larger portion of the Southern Ocean than at present, as the reason for the inhibition of exchange of CO_2 between the

ocean and atmosphere. Stephens and Keeling (2000) suggested that this process may be responsible for approximately 80% of the total CO_2 drop recorded in atmospheric proxy records during the last glacial.

Whether atmospheric CO_2 leads or lags climate change is crucial to determining its role in the growth and decay of ice

sheets. Unfortunately, ice core analysis is equivocal on this issue. The problem is that deep ice cores in Antarctica lack the accurate dating required to match CO₂ variations to known climate change events recorded elsewhere. Therefore, although the effects of atmospheric CO₂ levels on climate are established, and CO₂ changes lead ice sheet responses to climate change (Shackleton, 2000), it remains unclear whether CO₂ actually forces climate change over glacial cycles or simply enhances it.

Ice Sheet Elevation and Mass Balance

As ice sheets build, the surface elevation at which snow accumulates increases. Ice growth results in an increase in the area of the ice mass above the ELA. A feedback is then set up, whereby the increase in the accumulation zone results in further glacier growth. Steady state is achieved by enhanced ablation (e.g., the southern margin of the Laurentide Ice Sheet) and/or by a reduction in accumulation rate as a consequence of elevation increase (as in east Antarctica today). Such is the climatic effect of ice sheet orography (surface relief of the Earth's subaerial continents) on mass balance that, it has been argued, if the ice sheet on Greenland were taken away, it would probably not regrow to its current size without climate cooling (Huybrechts, 2002). Thus, many contemporary ice masses owe their current form to the former climate conditions that permitted their initial growth.

In the past, some ice sheets were so large that the hypsometry (measurement of the elevation of land above sea level) of the continents on which they lay were modified substantially (e.g., the Laurentide Ice Sheet and North America at the LGM). Atmospheric circulation across such large glaciated regions would have been affected by ice sheet orography, and this may have had climatic and mass balance consequences for regions downstream (Kageyama and Valdes, 2000).

Ocean Circulation and Ice Sheet Growth and Decay

It is highly likely that the ocean circulation system affected glacier mass balance in the past. Today, the North Atlantic provides an important source of heat to Western Europe because warm water flows from the mid-Atlantic northwards (Figure 5). If this flow of water were suppressed, the North Atlantic would cool and the heat supplied to Western Europe would be taken away, resulting in colder conditions over Europe and, hence, a reduction in the ELA of regional glaciers. Conversely, if the northward drift of mid-Antarctic water were enhanced, it might lead to warmer conditions and an increase in the ELA in Europe. Warmer and colder maritime conditions may also be associated with wetter and drier climates, respectively. This may explain why Scandinavian glaciers display a positive mass balance over the past few decades, in contrast to the majority of other ice masses, which have decreased in size due to global warming (Braithwaite, 2002).

Process Complexity and Ice Sheet Fluctuations

There are likely other feedback processes than those listed previously, and many of these will be interactive. Therefore, although glacial–interglacial cycles may be forced by orbital variations, the climate processes that allow ice sheets to grow and decay over these timescales are interrelated and, consequently, complex. Such complexity may help explain the rapid, short-term variations in ice sheet volume that occur during glacial cycles, particularly deglaciation.

Ice sheets add complexity to the processes governing their growth and decay. One reason for this is that ice sheet flow cannot be assumed to be stable. Ice may flow by three processes: ice deformation, basal sliding, and deformation of water-saturated basal sediments. Ice deformation occurs in all ice masses. The other two processes require the ice sheet base to

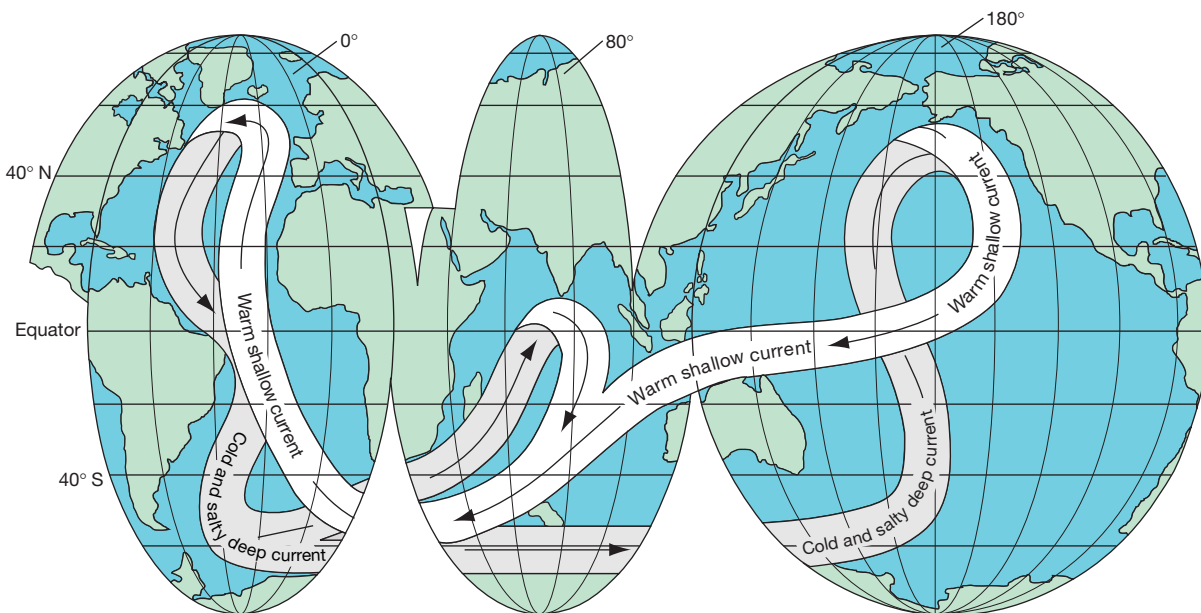


Figure 5 Schematic diagram of the global ocean circulation system.

be at the pressure melting temperature and can result in fast ice flow. The ice base may be warmed by thickening the ice sheet and/or by increasing the ice flow (as the friction associated with ice flow leads to an additional source of heat to the geothermal flux). Thus, as ice sheets grow while frozen to their beds, the basal temperatures increase until a threshold is reached at which subglacial water is generated. This activates subglacial motion and enhanced ice flow. Next, ablation rates likely increase because more ice will be transferred below the ELA. Subsequently, the ice sheet will thin, resulting in reduced basal temperatures and, eventually, the deactivation of enhanced ice flow. This 'binge-purge' model of ice sheet behavior has been used to explain the short-term rapid flow of the Laurentide Ice Sheet, which oscillated at its southern ablation margin to form a series of 'lobes' and across its eastern margin to periodically issue large numbers of icebergs into the North Atlantic, depositing layers of ice rafted debris, known as Heinrich layers (MacAyeal, 1993; Figure 6).

Periods of enhanced ice flow may be associated with the increased production of both meltwater and icebergs. Each of these leads to the transfer of freshwater from the continent to the ocean. In the case of the Laurentide Ice Sheet, the ocean to which such material is delivered is the North Atlantic, the site of deepwater formation and the engine of the thermohaline circulation. Thus, the short-term periodic enhanced flow of the Laurentide Ice Sheet, associated with the deposition of Heinrich layers, may also be linked with perturbations in ocean flow and, consequently, rapid climate change across the landmasses bordering the North Atlantic. This ice-ocean-atmospheric linkage may be particularly important during deglaciation, and it may have led to large-scale reversals in climate such as the Younger Dryas (Tarasov and Peltier, 2005).

A feature of several ice masses that underwent post-LGM retreat is the accumulation of meltwater across their proglacial regions (Mangerud et al., 2001). During ice sheet decay, many such proglacial lakes would have been extensive. In the case of the Laurentide Ice Sheet, the majority of its southern margin was in contact with huge expanses of proglacial meltwater. One massive lake, Lake Agassiz, grew larger as ice decay continued. Being dammed by the ice, a threshold was reached at ~8.2 ka, at which the ice sheet could no longer hold the lake water, resulting in a catastrophic outburst. So much water flowed to the North Atlantic that the sea-level rose by ~20 cm in a matter of days, according to some researchers, and a brief period of cooling resulted, which is recorded in the Greenland ice core (Barber et al., 1999).

The southern margin of the Eurasian Ice Sheet ~90 ka was also associated with a large proglacial lake named Lake Komi (Mangerud et al., 2001). According to numerical modeling of the ice sheet-climate-lake system, Lake Komi caused a reduction in the level of summer melting across the southern margin of the ice sheet, which supported ice sheet growth and delayed ice sheet decay. In contrast, other proglacial lakes have been found to accelerate ice sheet decay by allowing the process of iceberg calving to contribute to the loss of ice in otherwise terrestrial regions of large ice sheets.

Therefore, it seems likely that proglacial lakes may influence growth and decay of large ice sheets and, if they outburst into the oceans, may lead to rapid short-term oscillations in climate and glacier mass balance.

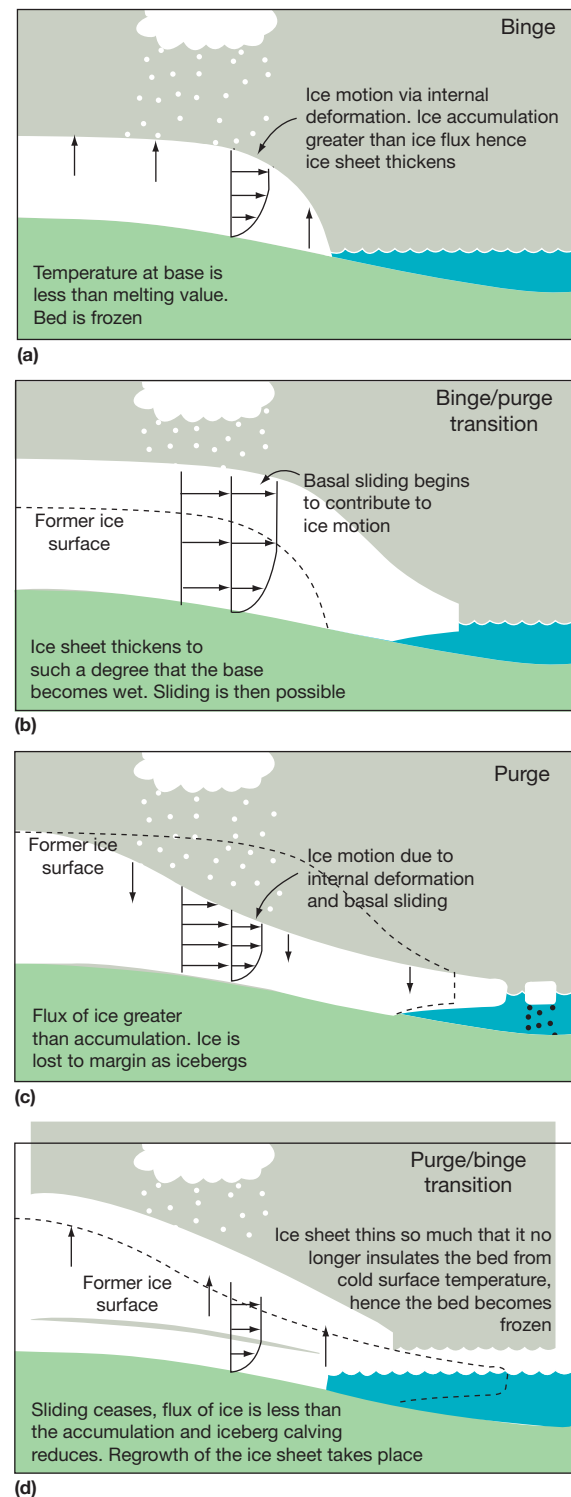


Figure 6 Binge-purge theory of ice sheet growth and decay. (a) Ice sheet growth occurs on a frozen bed. The only mechanism of ice flow is through ice deformation. (b) Further ice sheet buildup results in basal temperatures reaching the pressure melting point. Basal sliding contributes to ice flow and the ice sheet margin expands. (c) Enhanced velocities lead to more ice being lost to iceberg calving than is replenished through accumulation. The ice sheet issues icebergs and meltwater to the ocean and thins as a consequence. (d) Ice sheet thinning and retreat leads again to frozen basal conditions, slow ice flow, and ice sheet regrowth.

There are several large remnants of the last ice age. Of these, possibly the most important to modern environmental processes is the Greenland Ice Sheet, which is much larger now than in the Eemian interglacial (~125 ka; Cuffey and Marshall, 2000). Processes that have been linked with the decay of large ice sheets, such as the perturbation to the ocean circulation system and the associated rapid environmental change, remain relevant today because this ice sheet in particular reacts to climate warming and issues large volumes of freshwater (via runoff and icebergs) directly into the North Atlantic.

See also: **Glacial Climates:** Thermohaline Circulation.
Glaciation, Causes: Astronomical Theory of Paleoclimates.
Glaciations: Overview. **Ice Core Records:** Correlations Between Greenland and Antarctica. **Introduction:** History of Quaternary Science. **Paleoceanography:** Paleoceanography An Overview.
Paleoclimate Modeling: Last Glacial Maximum GCMs.
Paleoclimate Reconstruction: Sub-Milankovitch (DO/Heinrich) Events. **Sea Level Studies:** Overview.

References

- Barber DC, Dyke A, Hillaire-Marcel C, et al. (1999) Forcing of the cold event of 8200 years ago by catastrophic drainage of Laurentide lakes. *Nature* 400: 344–348.
- Braithwaite RJ (2002) Glacier mass balance: The first 50 years of international monitoring. *Progress in Physical Geography* 26: 76–95.
- Cuffey KN and Marshall SJ (2000) Substantial contribution to sea-level rise during the last interglacial from the Greenland ice sheet. *Nature* 404: 591–594.
- Huybrechts P (2002) Sea-level changes at the LGM from ice-dynamic reconstructions of the Greenland and Antarctic ice sheets during the glacial cycles. *Quaternary Science Reviews* 21(1–3): 203–231.
- Kageyama M and Valdes PJ (2000) Impact of the North American ice-sheet orography on the Last Glacial Maximum eddies and snowfall. *Geophysical Research Letters* 27: 1515–1518.
- MacAyeal DR (1993) Binge–purge oscillations of the Laurentide Ice Sheet as a cause of the North Atlantic's Heinrich events. *Palaeoceanography* 8: 775–784.
- Mangerud J, Astakov V, Jakobsson M, and Svendsen JI (2001) Huge ice-age lakes in Russia. *Journal of Quaternary Science* 16: 773–777.
- Paterson WSB (1994) *The Physics of Glaciers*, 3rd edn. New York: Pergamon.
- Petit JR, Jouzel J, Raynaud D, et al. (1999) Climate and atmospheric history of the past 420,000 years from the Vostok ice core, Antarctica. *Nature* 399: 429–436.
- Rignot E and Thomas RH (2002) Mass balance of polar ice sheets. *Science* 297: 1502–1506.
- Shackleton NJ (2000) The 100,000-year ice-age cycle identified and found to lag temperature, carbon dioxide and orbital eccentricity. *Science* 289: 1897–1902.
- Siegert MJ (2001) *Ice Sheets and Late Quaternary Environmental Change*. Chichester, UK: Wiley.
- Siegert MJ and Dowdeswell JA (2002) Late Weichselian iceberg, meltwater and sediment production from the Eurasian High Arctic ice sheet: Results from numerical ice-sheet modelling. *Marine Geology* 188: 109–127.
- Siegert MJ, Dowdeswell JA, and Melles M (1999) Late Weichselian glaciation of the Eurasian High Arctic. *Quaternary Research* 52: 273–285.
- Stephens BB and Keeling RF (2000) The influence of Antarctic sea ice on glacial–interglacial CO₂ variations. *Nature* 404: 171–174.
- Tarasov L and Peltier WR (2005) Arctic freshwater forcing of the Younger Dryas cold reversal. *Nature* 435: 662–665.

Evidence of Glacier and Ice Sheet Extent

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Importance of the Extent of Past Glaciers and Ice Sheets

Reconstruction of past areal and vertical ice extent is important because the volume and the height of ice sheets strongly influence global sea level and patterns of oceanic and atmospheric circulation. Ice sheets are large topographic features that create anomalies in albedo (reflectivity) and radiation balance, and represent significant freshwater reservoirs. They can influence climate by displacing the jet stream, by cooling the air, by reorganizing storm tracks, by releasing freshwater into the ocean, or by changing sea level. Therefore, past vertical and horizontal ice extents are important boundary conditions for paleoclimate modeling. If general circulation models are able to simulate climatic extremes, such as the time of major global glaciation, model-based future predictions will gain greater credibility.

Investigations of past ice extent also help in the understanding of glacial processes such as ice stream behavior (fast-flowing parts of ice sheets, enabled by the presence of large amounts of meltwater at the ice–bed interface). Such knowledge is important for a better prediction of the stability of current ice sheets, especially the West Antarctic Ice Sheet, which ablates dominantly through ice streams. Vertical ice extent in mountainous areas determines the distribution and size of nunataks (mountain tops that rise above the ice surface), and these have been proposed as ice-age refugia for plants and animals.

Areal Ice Extent

Ice Sheets During the Last Glaciation

The vast Laurentide Ice Sheet in North America had several ice domes and divides and was at times partly confluent with the smaller Cordilleran Ice Sheet and with the Innuitian and Greenland Ice Sheets at the Last Glacial Maximum (LGM; between 19 and 23 cal ka BP). In Europe, the Scandinavian (or Fennoscandian) Ice Sheet was the largest ice mass and probably coalesced with the British and Irish Ice Sheets during the time of their maximum extent. From the northern Baltic Sea area, the ice sheet radiated south- and eastward into major and minor ice lobes. Ice complexes in the southern hemisphere included the Patagonian Ice Sheet in South America and the Antarctic Ice Sheet, as well as glaciated areas in southeastern Australia and in Tasmania. Minor ice caps occurred in the tropical/subtropical zone, including East Africa, the Himalayas and the Tibetan Plateau, Japan, Taiwan, New Guinea, New Zealand, the Hawaiian Islands, the tropical Andes (Ecuador, Peru, Bolivia), and the circum-Caribbean region (Mexico, Guatemala, Costa Rica, Colombia, Venezuela).

Detailed Maps of Glacial Extent

The areal extent of Quaternary glaciations has been studied extensively and summarized in numerous publications. While

the ice extent for the LGM is relatively well-known, earlier glaciation limits are more difficult to determine. For the LGM, the nature of Earth's surface has been examined and mapped, and the climate system has been studied within two comprehensive projects: CLIMAP (Climate Long-range Investigation, Mapping, and Prediction; [CLIMAP Project Members, 1981](#)) and EPILOG (Environmental Processes of the Ice Age: Land, Ocean, Glaciers; see special issue of *Quaternary Science Reviews* 21: 2002). A recent overview by [Ehlers and Gibbard \(2004\)](#) comprises three volumes that cover different parts of the world and contain CDs with supporting material (GIS files).

Determination of the Areal Extent of Former Glaciers

The most conspicuous, and earliest recognized, remnants of past ice cover are erratic boulders, large rocks that have been transported by ice far away from the bedrock exposures from which they originated. In the less than 200 years since the recognition that glaciers were much more extensive than they are today, it has become evident that the glacial record is far more complex than this. The glaciers have left behind sediments and landforms, and although they together provide an integrated picture of the nature and extent of former ice sheets, they are discussed separately below.

Sediment record

Glaciers transport sediment on their surface, as debris-rich ice near the bed, and by causing deformation of soft sediment beneath the ice. Typically, glacier-transported sediment is poorly sorted (diamicton) and is called till if deposited directly by the glacier. Although till can be confused with landslide debris or other diamictons, the presence of a variety of lithologies, preferred orientation of clasts, and association with subglacial landforms can generally be used to map the former extent and flow direction of glacial ice. Glacial sediment that accumulates on the ice surface often slides and flows from high to low spots, causing differential melting of the underlying ice. The hummocky topography produced by this process contains a mix of till, debris flow sediment, and stream and pond sediment. Meltwater carries huge amounts of sand and gravel away from the ice margin. This sediment may be deposited far from the ice margin, so it is not a good indicator of ice extent, unless it is clear that the stream sediment was deposited on or under glacial ice (i.e., ice-contact glaciofluvial sediment). Often, signs of collapse of the sediment bedding indicate that ice once lay below, thus providing an indication of former ice extent. Glaciomarine sediments rarely show the former extent of ice but can be an important indicator of glaciation in the geologic record, because marine deposits are more often preserved than terrestrial deposits.

Geomorphological features

As glaciers move, they erode the underlying rock or sediment, and transport it in the direction of ice flow. Most of this sediment is deposited in landforms that vary considerably

depending on, for example, landscape relief, ice thickness, temperature, and presence or lack of water at the glacier bed. Often the nature and distribution of landforms provide a key to glaciological conditions when the landscape formed. One way of classifying these landscapes is by using a landsystems approach. A landsystem is characterized by its specific combination of landforms, subsurface materials, and vegetation types and thus allows the understanding of individual landforms in a broader landscape context. Many different glacial landsystems have been recognized (Evans, 2003); common landform assemblages are described below.

In some areas behind the ice margin, ice sheets erode preglacial soil and weathered and solid rock. Rock surfaces left behind are scoured and sometimes polished and marks indicate the former presence of ice and its flow direction (Figure 1). Pieces of rock or mineral fragments that are incorporated at the glacier base and dragged along leave scratches called glacial striations or striae on the rock surface, probably the most common features left by glacier ice. Deeper marks that are elongate in the direction of ice flow are called glacial grooves.

At places where debris at the ice base chattered over the rock surface instead of moving along smoothly, chatter marks or crescentic fractures develop. Reverse crescentic fractures are concave in an up-ice direction, due to a force applied by a rolling stone on the ice base at the up-ice side of bedrock outcrops. Crag and tail features, left behind by knobs of hard rock protruding from softer rock and thus protecting a tail of the softer rock down-ice from erosion, are valuable indicators of the direction of ice flow.

Larger landforms produced by glacial erosion, especially in alpine areas, provide evidence of the extent of past glaciers. Valleys produced by river downcutting typically have a V-shaped cross-section, whereas U-shaped valleys indicate the former presence of ice (Figure 2). Many of these U-shaped valleys were fed by glaciers in cirques, bowl-shaped basins produced by glacial abrasion and plucking. Often cirque floor elevations are used as one indicator of the position of the former equilibrium line. Rock is commonly gouged more deeply in some places than others, producing depressions that may become lakes. Since stream valleys do not have

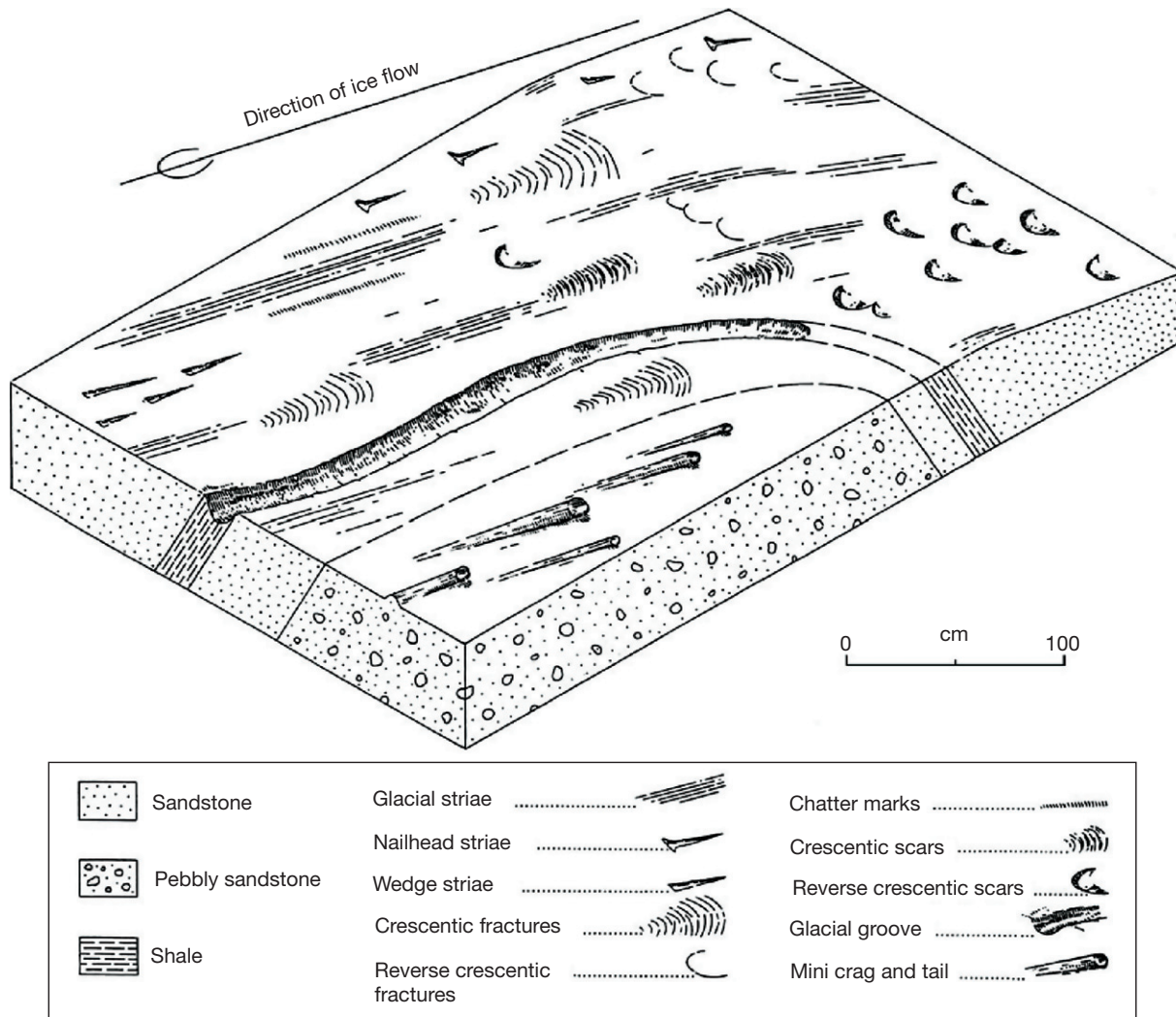


Figure 1 Glacial erosional forms on rock surfaces, showing the direction of ice flow. Reproduced from Prest VK (1983) Canada's heritage of glacial features. *Geological Survey of Canada, Miscellaneous Report 28*, with permission.

natural lakes, the presence of shallow rock basins along the valley bottom proves the extent of past glaciers. In places where several cirque glaciers have eroded headward toward a peak, a horn is produced (Figure 3). U-shaped valleys that have their floors below sea level are called fjords (Figure 4), and glaciers that eroded fjords terminated in water, thus producing landform and sediment assemblages different from those described below (Powell, 2003; Vorren, 2003). Sediment is delivered to alpine glaciers by subglacial erosion, and by avalanche and other mass wasting processes on to the ice surface. Many glaciers have accumulations of sediment along their edges called lateral moraines. Their upper extent marks the upper limit of former glacier ice on the valley wall. Where these moraines meet at the confluence of alpine valleys, sediment is carried on to the ice surface and then down-valley. These medial moraines (Figure 2) are rarely preserved after ice has disappeared. End moraines are ridges of sediment that form in the valley away from valley walls. The position of moraines

indicates the extent of glacier ice when the moraine was being formed, and often successive advances produce sequences of moraines that are progressively older down-valley.

Ice sheets that cover generally low-relief and low-elevation areas also leave behind end moraines that mark the extent of ice at the time the moraine was formed (Figure 5). They are characterized by their hummocky topography, and may contain sand and gravel as well as till and supraglacial diamicton. These are the most common landforms used to map the former extent of glacier ice in low, rolling landscapes. The outermost extent of an ice advance is called a terminal moraine, and moraines formed during ice retreat are recessional moraines. These ridges are generally interpreted as representing positions of stability of the ice margin. Moraines can be formed by several processes, including deposition of basal till, collapse of supraglacial sediment, bulldozing and ice shove, and stacking of debris-rich ice layers that may melt out centuries later. These differences in process result in different sizes and forms

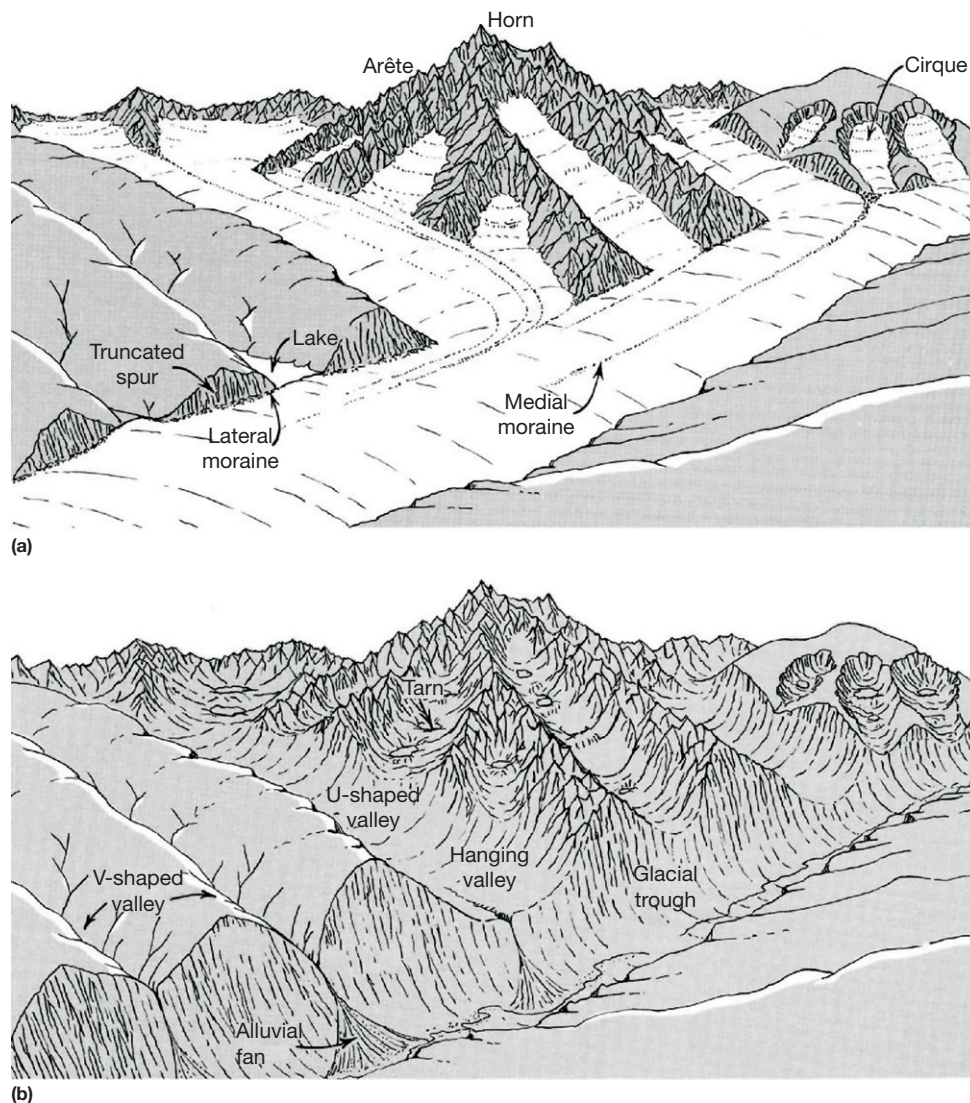


Figure 2 Large-scale features produced by glaciation in alpine areas: (a) during glaciation, (b) after deglaciation. Adapted from Strahler AN and Strahler AH (1974) *Introduction to Environmental Science*. Santa Barbara: Hamilton Publishing, with permission.



Figure 3 View of the Matterhorn, a horn, from above Zermatt, Switzerland.



Figure 4 View of Fjaerlandsfjord, Norway, showing U-shaped valley cross-section.

of moraines. They range from less than a meter high to several tens of meters and from about a meter to several kilometers in width. Although generally composed of diamicton, in some parts of the world some ridges of meltwater-deposited sand and gravel are called moraines. These are often in the shape of coalescing alluvial fans and deltas deposited near the ice margin, and like true moraines, indicate the presence of a stationary ice margin position.

In some environments, the land surface over which the ice rides is streamlined by a combination of erosion and deposition. The resulting elongate hills are called drumlins ([Figure 6](#)), and these can be used to determine the flow direction of former ice masses. Drumlins may indicate not only the extent of glacier ice, but also the conditions at the glacier bed when the streamlining took place. They are generally thought to have been shaped and streamlined when there was water at the bed of the ice and the glacier was sliding. Smaller elongate forms, called flutes, are present on many recently deglaciated surfaces,

but may be difficult to recognize on Pleistocene-age vegetated landscapes. In some subglacial environments, particularly where the bed of the glacier is frozen but wet, soft sediments are present beneath, stresses are propagated into the rock or sediment, and thrust features, called hill-hole pairs or anamoores ([Figure 7](#)), are formed as slabs of frozen sediment or rock are thrust by the glacier.

Associated subglacial meltwater features are tunnel channels and eskers, which are deposited by water flowing beneath the ice. Tunnel channels are primarily stream-cut channels that were partly or wholly filled with meltwater sediment and supraglacial diamicton as ice thinned. They formed in several different subglacial environments, but generally formed in the marginal zones of former ice sheets, perhaps where the base of the glacier was partly frozen to its bed, or where breaking ice dams caused catastrophic flows of subglacial meltwater. Eskers ([Figures 8 and 9](#)) are generally sinuous ridges of gravel that may extend for up to hundreds of kilometers across the

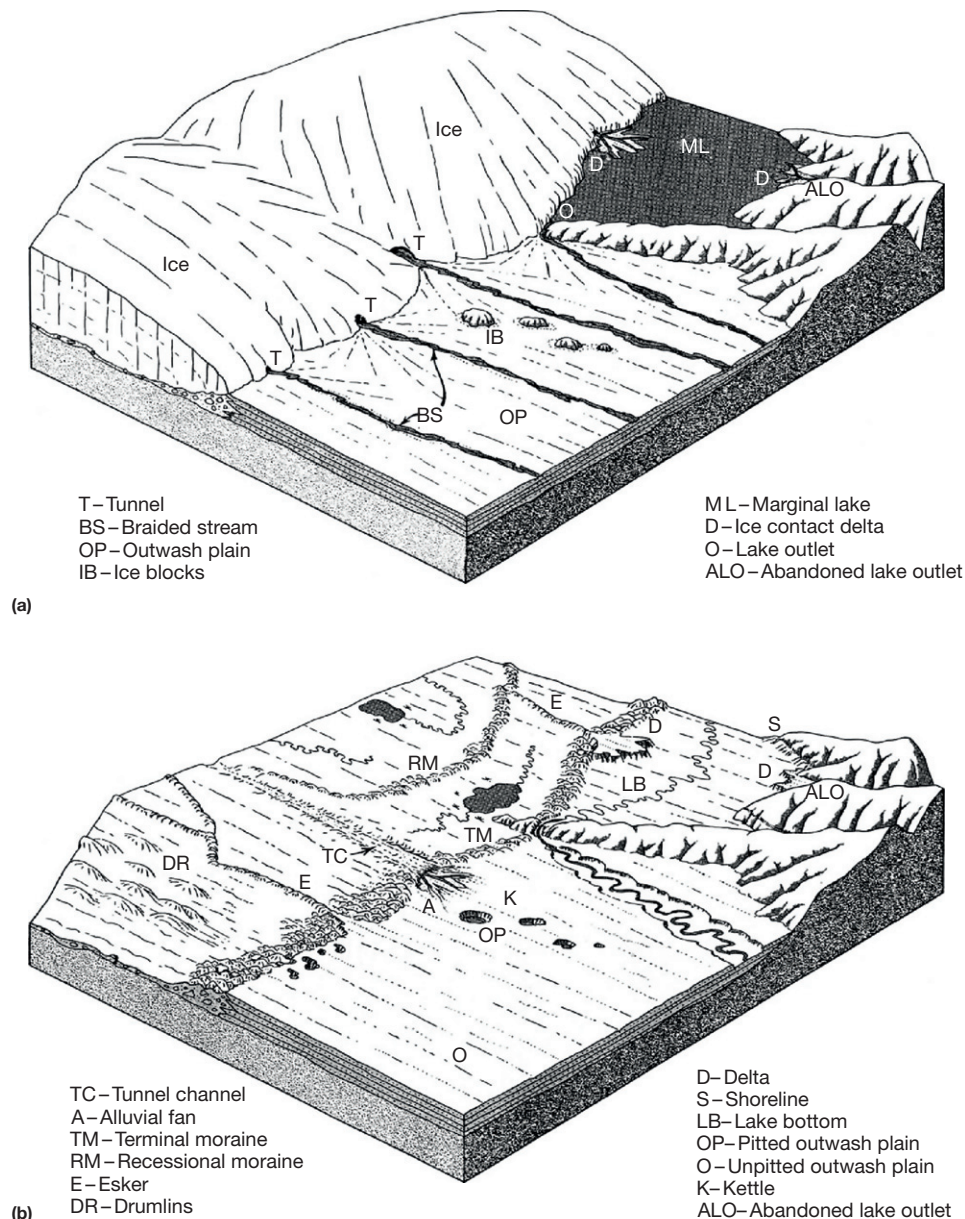


Figure 5 Lowland landform assemblage: (a) during glaciation, (b) after deglaciation. Adapted from Strahler AN and Strahler AH (1974) *Introduction to Environmental Science*. Santa Barbara: Hamilton Publishing, with permission.

landscape. In some areas that were under the thickest parts of ice sheets, and where most sediment has been scoured from bedrock surfaces, eskers provide the clearest record of former glaciation. Stream-deposited sand and gravel accumulates on the glacier surface in places where streams rework sediment from melting debris-rich ice. Collapse of this gravel produces a hummocky surface. Depressions called kettles are produced as the last buried ice melts between hummocks that are sometimes called kames. Water and sediment flowing down a moulin, a vertical shaft in the ice, and accumulating in an inverted cone-shaped hill, produce moulin kames (Figure 10). Where meltwater exited the ice, ice-contact fans or deltas typically formed. In some areas, these are the only indicators of a former ice margin position if no moraine was built. Sand and gravel

plains and terraces may have kettles that demonstrate that ice formerly underlay the surface. In the absence of till, this pitted outwash may be the only indicator of former ice cover.

In areas where the ice margin was against hills at its maximum extent or during deglaciation, lateral meltwater channels commonly mark former ice margin positions. The slopes of the floors of ice-marginal channels indicate the slope of the intersection of ice with the hillside and allow ice extent and ice surface slope to be determined.

In places, ice-marginal lakes can be used to reconstruct previous ice extent even where glacial deposits are lacking. Abandoned beaches and lake outlets can be used to determine the former elevation of ice-dammed lakes. Ice advances typically block outlets, causing lake levels to rise or new lakes to form.



Figure 6 Drumlin in Wisconsin, USA, highlighted by contour plowing. Ice flow direction is from upper right to lower left. Photograph by Donna Harris.



Figure 7 Air photo of anamoose or hill-hole pair, North Dakota, USA. Ice flow from upper left. Distance between prominent north-south (top-bottom) roads is 1 mile or 1.6 km.



Figure 8 Oblique air photo of Dahlen Esker, North Dakota, USA. Water flow direction is from lower right to upper left. Photograph by Lee Clayton.



Figure 9 Cross section of Pine River Esker in New Hampshire, USA.



Figure 10 Moulin kame forming at base of moulin beneath marginal zone of the Burroughs Glacier, Alaska, USA. Vegetation and soil has been removed ahead of mining.

Ice retreat often exposes lower outlets, allowing water level to fall or lakes to entirely disappear. Because these lakes require a glacier dam in order to exist, in the past ice must have extended far enough across a landscape to dam the water.

Most of the above features are studied from air photos and field examination. However, some features can only be well displayed with satellite images (Boulton et al., 2001) or digital elevation models of large areas. These provide overviews of glacial bedforms and their succession, and can be used to map maximum ice extent and deglaciation patterns (Clark, 1997). Figure 11 shows a digital elevation model of the southwestern part of the Laurentide Ice Sheet. The extent of glacier ice can be clearly seen. Streamlined paths due to ice flow are interpreted to be indicators of fast-flowing (streaming) ice (Clark and Stokes, 2003).

The above features allow determination of the extent of glacier ice in many locations, but there are places where former ice left little trace of its presence, leading to debates on whether

these regions were glaciated at all. Even in areas where glacier coverage is obvious, there may be debate about whether it was during the last glacial cycle. Determining the age of glaciation is routinely undertaken using radiocarbon dating, but this is limited to the last glaciation and to deposits containing organic matter. When organic material is lacking in glacial deposits, optically stimulated luminescence and cosmogenic radionuclide surface exposure dating techniques are now commonly used to determine the age of glaciation. Weathering rates and degree of soil development can also be used to indicate the relative age of geomorphic surfaces.

Areas of Uncertain Areal Extent

The extent of glaciation is difficult to detect in places where glaciers did little geomorphic work and, therefore, had little effect on the landscape. The absence of till, meltwater stream deposits, or erratics can be interpreted either as a lack of

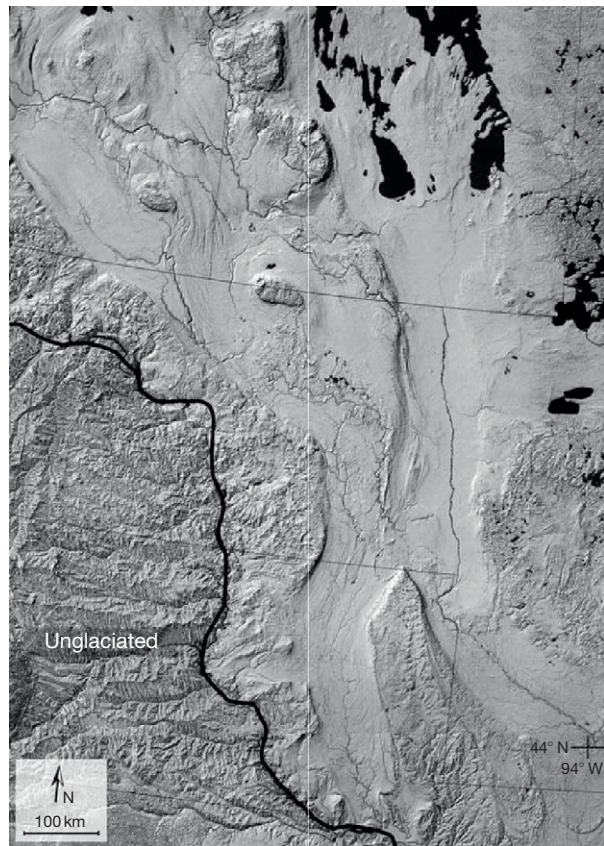


Figure 11 Shaded relief map of the southwestern part of the Laurentide Ice Sheet in the United States and Canada, showing evidence of ice streaming along the western side of the Laurentide Ice Sheet. The central parts of ice lobes flowed parallel to the axis of the lobes, suggesting stagnant or slow moving ice along the edges of the lobes. Note southwestern extent of the Laurentide Ice Sheet is the line separating unglaciated terrain in the West from glaciated terrain to the East and North. Image by Paul Morin, data from NASA Shuttle Radar Topography Mission. See [Jennings \(2006\)](#) for further information.

glaciation or that the ice that covered the area left no deposit behind – it was clean ice. The presence of paleosols, deeply weathered rock, surfaces covered with weathered boulders, and lack of striations or other erosional features, may also be explained in different ways. Conditions at the bed of glaciers differ considerably. Some have abundant water, and slide or ride on a deforming soft-sediment layer. These readily shape the landscape beneath. In some places, there was little erosion, and thick till was deposited. Other glaciers or parts of glaciers are frozen to their beds. Although some basal movement has been measured even in frozen-bed glaciers, occasionally they appear to have relatively little impact on the underlying ground surface; they do not alter the landscape significantly and can thus preserve relicts from former glaciations and interglacials (as has been inferred for the central Fennoscandian Ice Sheet – [Kleman and Hättestrand, 1999](#); [Stroeven et al., 2002](#)). Examples of regions of debatable areal extent are described in the following text and in more detail by [Rutter \(1995\)](#).

Northern Laurentide Ice Sheet

The limit of the Laurentide Ice Sheet during the last glaciation (the Late Wisconsin) is well-established at its southern border, but remains contentious in many other areas ([Dyke et al., 2002](#)). Ice extent on eastern Baffin Island is still debated. Timing of maximum ice extent is not well-known for the northwestern Laurentide margin. Another debate concerns the coalescence of the Laurentide Ice Sheet with the Cordilleran Ice Sheet. It is now well-established that the two ice sheets coalesced at their southern margin, but at the northern part of their respective borders, an ice-free corridor may have existed. The two ice masses seem to have developed out of phase.

Innuitian Ice Sheet

After much debate as to whether an Innuitian Ice Sheet existed over the northern half of the Canadian Arctic Archipelago during the LGM, consensus has emerged that such an ice sheet covered most of this region, based mainly on evidence from flow and from meltwater features ([Dyke et al., 2002](#)). However, the northwestern limit of this ice mass, presumably located on the polar continental shelf, has not been mapped to date.

Barents Sea

An ice sheet that was centered over the Barents and Kara Seas probably expanded on to mainland Russia several times during the Quaternary, but there has been a long-standing debate about whether it existed during the last glaciation. Northern Eurasia has been studied extensively in the QUEEN program (Quaternary Environments of the Eurasian North; see special volume of *Boreas*, 1999, 28(1)). Glaciologic modeling supports a larger ice sheet that covered parts of the Russian mainland, while most field observations and geochronometric data suggest that the area was not overridden by ice during the last glaciation.

Eastern Siberia

Another debate of ice extent centers on northeastern Siberia. Some evidence for a marine-based ice sheet comes from features identified through sea-floor mapping on the Chukchi shelf, but their age has not been determined. Other studies strongly question the existence of an LGM western Arctic ice sheet ([Sher, 1995](#)).

Antarctic Ice Sheet

Evidence is growing that the West and East Antarctic Ice Sheets did not advance and retreat synchronously. Probably the West Antarctic Ice Sheet extended to the shelf edge in most areas, as has been inferred from geophysical surveys of the Antarctic continental shelf, while the East Antarctic Ice Sheet advanced to a mid-shelf position only.

Tropical and subtropical regions

LGM conditions are still controversial in many parts of the tropics and subtropics. Studies have recently been compiled in a special volume of *Quaternary International*, 2005, 138/139).

Vertical Extent

Determination of Vertical Extent

The vertical extent of former glaciers and ice sheets is much more difficult to determine than the areal extent and involves more uncertainties, because there is often no direct geological proxy available. We outline methods that are commonly applied in order to infer the vertical extent of ice masses, especially for the last glaciation.

Isostatic rebound

The ice-water mass exchange during a glacial cycle is sufficient to generate a deformation of the Earth, called glacial isostatic adjustment. Present rates of uplift in formerly glaciated areas can be used to infer former ice thickness distribution using rheologic modeling. Uplift rates observed relative to mean sea level have to be corrected for eustatic changes. Theoretical tilting of shorelines computed in the models is then compared to observed shoreline tilting. The relative sea-level data might, however, primarily reflect the deglaciation period subsequent to the LGM and hence not give a definitive estimate of LGM ice thickness. Also, these data are confined to coastal regions and, therefore, provide little constraint on ice thickness in interior regions. Furthermore, isostatic modeling often relies on the assumption that the postglacial uplift is caused entirely by unloading due to ice sheet melting. Often, possible tectonic components in the uplift are not considered or are difficult to estimate.

Global sea-level change

The magnitude of global sea-level change due to LGM glaciation has been debated for a long time. The Laurentide and the Fennoscandian Ice Sheets contained the majority of ice at the LGM, but smaller ice sheets and glaciers also contributed to global sea-level lowering. CLIMAP presented two possible ice sheet reconstructions: a minimum model, accounting for 127 m of ice-equivalent sea-level lowering, and a maximum model (with extensive marine-based ice sheets), accounting for 163 m of sea-level lowering (Denton and Hughes, 1981). Subsequently, substantially thinner ice sheets have been inferred, for example, through isostatic modeling, suggesting a global sea-level lowering of about 120 m (Peltier, 2002) to 130–135 m (Lambeck et al., 2002; Milne et al., 2002). Coral reef data also yield sea-level records that suggest a sea-level lowering of about 120 m due to glaciation. Geochemical records of oxygen isotopes ($\delta^{18}\text{O}$) from deep-sea cores provide supporting evidence for the LGM ice volume.

Ice-sheet modeling

Observational control on ice sheet thickness changes remains limited in many areas, for example, the Greenland and the Antarctic Ice Sheets, but also for most parts of the Laurentide Ice Sheet. For these locations, glaciological modeling is especially useful (Marshall et al., 2002; Siebert et al., 2001). Most models applied are time-dependent and thermomechanically coupled and account for isostasy. They are based on mathematical relations of mass balance (resulting from snow accumulation and ice ablation), ice flow, and interaction of the ice with the underlying bedrock. Models use paleoclimatic

information, often as output from paleoclimate models or from proxy data, in order to simulate ice extent and ice thickness evolution over time. Often sensitivity experiments are carried out in which parameters are varied within plausible ranges of uncertainty. Model results are constrained by comparison with the geological data available. However, models are generalizations of reality; not all possible processes are represented in models, and parameterizations are simplifications.

Consolidation measurements

The pressure exerted by an ice mass leads to the vertical compression of the underlying soil. Water and air are expelled from pore space in the sediment, soil particles are packed more closely, and deformation of the soil particles takes place. All these processes yield a volume change that is called consolidation. The stress that has been experienced by a soil sample when it was overridden by ice in the past can be inferred by consolidation testing, during which an increasing load is applied to the sample and the resulting deformation is observed. From changes in the response of the sample to the load, the preconsolidation pressure can be inferred. Since former pore water pressure partially supported the ice load, only minimum ice thickness values can be estimated from the consolidation testing results.

Sources of error for this method include disturbances or changes in the moisture content of the sample during collection or storage, fluctuations in the water table during the geologic history of the sampling area, changes in soil structure by freezing and thawing, or the presence of permafrost or meltwater during maximum ice load.

Landmarks

Geomorphic observations can help to identify upper ice sheet/glacier limits. Indicators used include trimlines, separating an upper zone of *in situ* weathered bedrock and a lower zone of glacial erosion and hence representing the upper limit of glaciation in mountainous areas (Figure 12). However, these landmarks are often controversial for several reasons: They could represent a glacial readvance rather than the maximum vertical ice limit, or periglacial weathering may have taken place after deglaciation, or they may indicate the boundary between frozen and temperate conditions at the ice base. Mineral magnetic soil measurements or soil clay-fraction mineralogy may help to address this problem if reliable dating for bedrock above and below the trimline is not available.

Cosmogenic dating

Cosmogenic nuclides (^{26}Al , ^{36}Cl , ^{10}Be , ^3He , ^{21}Ne) are produced in surface rocks (affecting mostly the upper 1 or 2 m) by cosmic radiation. Samples for cosmogenic dating are often taken from quartz-rich clasts or quartz veins from glacial erratic boulders or from glacially smoothed bedrock. Areas that show signs of weathering or surface degradation have to be avoided. Production rates are used for calculating surface exposure ages after corrections for altitude and latitude. However, some uncertainties arise from possible erosion and/or burial of the surface. Erosion rates are sometimes difficult to estimate.



Figure 12 Photo of trimline between clearly glaciated, till-covered surface below and weathered rock-strewn surface above, Nordfjord, Norway. Photograph by M. Hildreth.

Examples for Controversial Areas of Vertical Ice Extent

Evidence for vertical ice sheet extent is often controversial. For example, ice thickness at the LGM of the Scandinavian Ice Sheet in the mountainous areas in southwestern Norway has long been debated. While ice-free regions in western and northern Norway have been postulated on the basis of biological findings (survival of endemic plant and animal species), geomorphic evidence (presence of steep-sided peaks, cirque topography, summit block fields), and cosmogenic dating, thick ice that covered all mountain peaks has been inferred from numerical modeling of glaciological processes, preconsolidation testing, isostatic response, till fabrics, and striation patterns.

A similar controversy has evolved around glacial extent in eastern Baffin Island, Arctic Canada. While some researchers inferred ice-free refugia and large outlets of Laurentide ice that were restricted to fjords and valleys, others suggested a complete ice cover of Baffin Island during the last glaciation (see overview in [Miller et al., 2002](#)).

As methods of dating improve and we gain a better understanding of how cold-based glaciers function and how to interpret the meager traces they leave behind, these controversies will likely be resolved.

See also: **Cosmogenic Nuclide Dating:** Overview. **Glacial Landforms:** Glacial Landsystems; Glaciofluvial Landforms of Erosion; Moraine Forms and Genesis. **Glacial Landforms, Erosional Features:** Major Scale Forms; Micro- to Macroscale Forms. **Glacial Landforms, Ice Sheets:** Trimlines and Paleonunataks. **Glacial Landforms, Sediments:** Glaciofluvial Landforms of Deposition; Glaciomarine Sediments and Ice-Rafted Debris; Till Fabric Analysis; Tills. **Glaciations:** Overview. **Quaternary Stratigraphy:** Overview; **Sea Level Studies:** Eustatic Sea-Level Changes – Glacial–Interglacial Cycles; Eustatic Sea-Level Changes Since the Last Glacial Maximum.

References

- Boulton GS, Dongelmans P, Punkari M, and Broadgate M (2001) Paleoglaciology of an ice sheet through a glacial cycle: The European ice sheet through the Weichselian. *Quaternary Science Reviews* 20: 591–625.
- Clark CD (1997) Reconstructing the evolutionary dynamics of former ice sheets using multi-temporal evidence, remote sensing and GIS. *Quaternary Science Reviews* 16: 1067–1092.
- Clark CD and Stokes CR (2003) Paleo-ice stream landsystem. In: Evans DJA (ed.) *Glacial Landsystems*. New York: Oxford University Press.
- CLIMAP Project Members (1981) *Seasonal reconstruction of the Earth's surface at the Last Glacial Maximum*. Geological Society of America; Map and Chart Series C36.
- Denton GH and Hughes TJ (1981) *The Last Great Ice Sheets*. New York: Wiley.
- Dyke AS, Andrews JT, Clark PU, et al. (2002) The Laurentide and Innuitian ice sheets during the last glacial maximum. *Quaternary Science Reviews* 21: 9–31.
- Ehlers J and Gibbard PL (eds.) (2004) *Quaternary Glaciations – Extent and Chronology, Parts I–III*. Amsterdam: Elsevier.
- Evans DJA (2003) Introduction to glacial landsystems. In: Evans DJA (ed.) *Glacial Landsystems*, pp. 1–11. New York: Oxford University Press.
- Jennings CE (2006) Terrestrial ice streams – A view from the lobe. *Geomorphology* 75: 100–124.
- Kleman J and Hättestrand C (1999) Frozen-bed Fennoscandian and Laurentide ice sheets during the Last Glacial Maximum. *Nature* 402: 63–66.
- Lambeck K, Yokoyama Y, and Purcell T (2002) Into and out of the Last Glacial Maximum: Sea-level change during Oxygen Isotope Stages 3 and 2. *Quaternary Science Reviews* 21: 343–360.
- Marshall SJ, James TS, and Clarke GKC (2002) North American Ice Sheet reconstructions at the Last Glacial Maximum. *Quaternary Science Reviews* 21: 175–192.
- Miller GH, Wolfe AP, Steig EJ, Sauer PE, Kaplan MR, and Briner JP (2002) The Goldilocks dilemma: Big ice, little ice, or 'just-right' ice in the Eastern Canadian Arctic. *Quaternary Science Reviews* 21: 33–48.
- Milne GA, Mitrovica JX, and Schrag DP (2002) Estimating past continental ice volume from sea-level data. *Quaternary Science Reviews* 21: 361–376.
- Peltier WR (2002) On eustatic sea level history: Last Glacial Maximum to Holocene. *Quaternary Science Reviews* 21: 377–396.
- Powell RD (2003) Subaquatic landsystems: Fjords. In: Evans DJA (ed.) *Glacial Landsystems*, pp. 313–347. New York: Oxford University Press.
- Rutter N (1995) Problematic ice sheets. *Quaternary International* 28: 19–37.
- Sher A (1995) Is there any real evidence for a huge shelf ice sheet in East Siberia? *Quaternary International* 28: 39–40.

Siegert MJ, Dowdeswell JA, Hald M, and Svendsen J-I (2001) Modelling the Eurasian Ice Sheet through a full (Weichselian) glacial cycle. *Global and Planet Change* 31: 367–385.

Stroeven A, Fabel D, Hättestrand C, and Harbor J (2002) A relict landscape in the centre of Fennoscandian glaciation: Cosmogenic radionuclide

evidence of tors preserved through multiple glacial cycles. *Geomorphology* 44: 145–154.

Vorren TO (2003) Subaquatic landsystems: Continental margin. In: Evans DJA (ed.) *Glacial Landsystems*, pp. 289–312. New York: Oxford University Press.

Evidence of Glacier Flow Directions

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Introduction and Significance

Knowledge of the flow direction of Quaternary ice sheets is crucial to both paleoglaciology and glaciology and has important socioeconomic implications in terms of mineral prospecting and mining. In paleoglaciology, evidence of former ice flow directions is a key ingredient for accurate reconstructions of the behavior of former ice sheets. Such reconstructions typically span thousands of years and provide valuable knowledge of how ice sheets evolve and both respond to and perturb the ocean-climate system (e.g., Clark et al., 2010; Kleman et al., 1997). These reconstructions, based on glacial geomorphology, are also required to test numerical models of paleo-ice sheets (e.g., Holmlund and Fastook, 1995; Tarasov and Peltier, 2004) and, significantly, provide a key context for assessing the present stability of modern-day ice sheets and the likely magnitude of sea-level change that might be expected under different climate scenarios (Alley et al., 2005).

Evidence relating to former ice flow directions is also intimately linked to subglacial processes that occur at the base of ice sheets and facilitate their flow. Because of the inaccessibility of modern-day ice sheet beds, these processes are poorly constrained, leaving us with a rather rudimentary understanding of how ice interacts with the underlying substrate and how such processes ought to be parameterized in numerical models of ice flow. It is known, however, that former ice flow directions are recorded by suites of subglacial bedforms (e.g., drumlins) created at the ice-bed interface and these have recently been imaged beneath a West Antarctic ice stream (King et al., 2009). These ice flow indicators are a direct manifestation of subglacial processes, and by studying both their morphometry (e.g., Clark et al., 2009) and internal composition (Stokes et al., 2011), it is possible to gain important new insights to complement those being obtained from direct investigation of modern-day ice sheets.

Former ice sheet flow directions are also crucial to regional drift prospecting, and it has long been recognized that various grain sizes in till can be analyzed to determine their mineralogical and chemical composition to delimit dispersal trains and define source areas of potential resources for mining (McMartin and McClenaghan, 2001). Knowledge of former flow directions is therefore crucial to constrain sampling strategies and procedures. Moreover, once identified, defined paths of glacial transport of distinctive bedrock lithologies have resulted in major new insights regarding the dynamic behavior of paleo-ice sheets (e.g., Dyke and Morris, 1988; Shilts, 1980; Shilts et al., 1979).

This article outlines the principles of glacial inversion techniques for reconstructing former ice sheet flow directions, emphasizing the importance of cross-cutting ice flow indicators, ice sheet basal thermal regime, and paleo-ice streaming.

Glacial Inversion of Paleo-Ice Sheet Beds

Information regarding the presence and extent of paleo-ice sheets has been compiled since the nineteenth century, but it was the advent of aerial photography, satellite remote sensing, and various dating techniques in the latter half of the twentieth century that resulted in robust reconstructions of their maximum extent and subsequent timing of margin retreat (e.g., Denton and Hughes, 1981; Dyke and Prest, 1987a,b; Ehlers and Gibbard, 2004). In contrast, ice sheet reconstructions that incorporate the internal properties of the ice sheet (e.g., the changing flow dynamics, velocity, basal thermal regime, and location of ice divides) have been generated only relatively recently.

Numerical ice sheet models are capable of producing fully three-dimensional simulations of ice sheet dynamics (e.g., Holmlund and Fastook, 1995; Tarasov and Peltier, 2004) but, as noted earlier, such models require validation and evaluation against the glacial geomorphological record. As a result, the last few decades have seen much research effort focused on using the glacial geomorphological record to decipher ice sheet evolution through time, with particular attention to the organization of the subglacial thermal regime, subglacial hydrology, evidence of fast ice flow (paleo-ice streams), and an appreciation of the time-integrated nature of subglacial landscapes (Kleman et al., 2006). This process is known as *glacial inversion* (Kleman and Borgström, 1996) and can be simply defined as “the extraction of ice-sheet properties from glacial geomorphology and geology” (Kleman et al., 2006, p. 192).

Ice sheets leave behind a wealth of evidence for glacial inversion techniques (Kleman and Borgström, 1996). Arguably, the most widespread and commonly utilized in terms of ice flow direction are erosional and depositional landforms. Erosional forms sculpted in bedrock range from small-scale rat tails, striae, and chattermarks to larger forms such as roches moutonnées, whalebacks, and crag-and-tails. Some of these forms may provide only an orientation and not direction (e.g., striae), but where they exhibit a clear longitudinal asymmetry, it is presumed to reflect differential erosion processes at the up-ice (stoss) and down-ice (lee) sides of the form, and ice flow direction is more obvious. For example, steeper stoss slopes on crag-and-tails or irregular plucked surfaces on the lee side of roches moutonnées. It should be noted, however, that while the morphometry of bedrock features can be used to reconstruct ice flow direction and even draw inferences regarding ice velocities (e.g., Roberts and Long, 2005), the characteristics of the underlying geology (e.g., spacing of joints) can introduce additional complexity (cf. Roberts et al., 2010).

Erosional or depositional forms may also be created from unconsolidated sediments, and features aligned parallel to ice flow direction are loosely referred to as glacial lineations. These range from small-scale flutings to larger and more ovoid drumlins and even larger and more elongate mega-scale glacial

lineations. Drumlins are, arguably, the most ubiquitous and, perhaps, the most studied of all glacial landforms, and it follows that they are most commonly employed in reconstructing former ice flow directions.

Traditionally, the 'textbook' description of a drumlin describes a blunter, steeper stoss side and a more gently tapering lee side, and this aerodynamic profile and teardrop planform is widely cited in numerous papers and textbooks. However, it has been recognized that drumlin morphometry varies from this classic description and can range from almost perfectly round to highly elongate spindle forms (e.g., Aario, 1977), with some workers suggesting that the diversity of forms can, in fact, be viewed as a glacier bedform continuum, which might also include transverse features such as ribbed moraine (e.g., Menzies, 1987; Rose, 1987). Recent support for this idea comes from measurements of drumlin morphometry, which reveal a unimodal distribution of various parameters, for example, length, width, and elongation (e.g., Clark et al., 2009). Indeed, the recent availability of high-resolution satellite imagery and digital elevation models has allowed, for the first time, tens of thousands of drumlins to be systematically mapped and measured from a range of settings.

Interestingly, these large data sets have been used to question some long-held assumptions regarding drumlin shape and their utility as evidence of ice sheet flow direction. Spagnolo et al. (2010, 2011), for example, argue that the planform and longitudinal profile of drumlins actually tend toward symmetry, with the major implication being that they are not necessarily a reliable indicator of ice flow direction. With this point in mind, it is important to stress that most studies aimed at reconstructing ice flow direction will utilize a range of available evidence and rarely rely on the characteristics of individual landforms. In this respect, it is the overall pattern and spatial arrangement of a 'swarm' or 'flow-set' of drumlins and their relationship with other landforms that prove most informative (Clark, 1999; Kleman and Borgström, 1996). These might include major terminal moraine systems and/or eskers, which are generally aligned parallel to ice surface slopes and directed toward former margin positions. Thus, it is the appreciation of the entire sediment-landform assemblage (cf. land system: Evans, 2003) that provides the most robust evidence of ice flow direction and indeed ice dynamics, rather than individual landforms. This, along with an evaluation of the age of the landscape, is the essence of the glacial inversion technique, and the following section details recent methodological advances in this important area of paleoglaciology.

Cross-Cutting Ice Flow Indicators, Flow-Sets, Fans, and Ice Streams

Multitemporal Evidence of Ice Flow Direction and Cross-Cutting Landforms

The burgeoning availability of aerial photography and satellite remote sensing products in the latter half of the twentieth century permitted the systematic mapping of ice sheet beds and the production of some pioneering reconstructions of the last midlatitude ice sheets that remain benchmark publications (e.g., Dyke and Prest, 1987a,b). These reconstructions, however, often assumed that the mapped ice flow indicators

(e.g., drumlins) were formed in a radial pattern close to the ice margin and mostly during the decay stages of the ice sheet (e.g., Boulton et al., 1985). The resultant reconstruction of ice flow direction, therefore, often depicted a radial pattern of ice flow from one or more relatively stationary ice domes (Figure 1(a)).

In 1990, Boulton and Clark (1990a,b) questioned the assumption of glacial lineations forming close to the ice margin during deglaciation and instead recognized the ice sheet bed as a 'palimpsest' landscape of flow patterns of different ages that were not all formed in an ice marginal environment. Working on the former North American (Laurentide) Ice Sheet, they grouped coherent patterns of glacial lineations into discrete mapped units which they termed 'flow-sets' (Figure 1(b)). They noted how flow-sets typically cross-cut each other (see Figure 2) and were therefore able to assess their relative age of formation based on principles of superimposition, even in the absence of any absolute dating techniques. The resultant ice sheet reconstruction was more sophisticated than earlier attempts in that it clearly revealed a highly mobile ice sheet with major shifts in the location of ice divides and dispersal centers by 1000–2000 km during the glacial period, many of which have since been confirmed by a more detailed mapping of ice directional indicators (e.g., McMartin and Henderson, 2004).

The discovery of cross-cutting ice flow patterns demonstrated major changes in ice flow direction through time and overturned traditional assumptions regarding the long-term behavior and configuration of ice sheets (Clark, 1993). Later work by Clark (1997) suggested that there were potentially four possible situations that might lead to the creation of cross-cutting flow patterns (see Figure 3): (1) migrating ice divides, (2) the switching on and off of ice streams, (3) ice margin retreat, and (4) flow patterns from more than one glaciation.

Classification of Landform Fans and Flow-Sets

Clearly, the recognition of multitemporal evidence permits more detailed reconstructions of ice dynamics, but the geomorphological record contains additional evidence from which it is possible to glean information about ice dynamics at the time of flow-set formation. In this regard, Kleman and Borgström (1996) highlighted the importance of the basal thermal regime and subglacial hydrology (meltwater features) when deciphering the glaciodynamic context of flow-set formation. In a landmark paper that represented the first attempt at formalizing a methodology for glacial inversion techniques, Kleman and Borgström (1996) classified glacial landscapes or 'fans' (equivalent to flow-sets in Boulton and Clark, 1990a,b) according to (1) their age gradient (were the flow traces/glacial lineations formed synchronously or do they get younger or older down-ice?), (2) the presence or absence of aligned meltwater features, and (3) the basal thermal conditions (frozen or thawed?). They proposed that such considerations result in six fan types (flow-sets) that are known or postulated to exist (Figure 4):

1. *Wet-based deglaciation fans* that consist of flow traces (e.g., glacial lineations) overlain by aligned eskers and preserved as the ice retreats
2. *Dry-bed deglaciation fans* that consist solely of a see-through pattern of meltwater features overprinted on a relict surface



Figure 1 Contrasting approaches to the mapping of evidence of ice flow patterns on the bed of former North Laurentide Ice Sheet. In (a), [Boulton et al. \(1985\)](#) assume that all ice flow indicators (e.g., glacial lineations such as drumlins) are formed close to the ice margin during deglaciation, which results in a radial flow pattern from several large dispersal centers (ice domes). In (b), [Boulton and Clark \(1990a,b\)](#) recognize discrete patches of glacial lineations as separate entities, which they term 'flow-sets,' and which they suggest formed at different times during ice sheet buildup and decay. Relative age dating of these flow-sets using their cross-cutting relationships generates a highly mobile reconstruction of ice flow with major shifts in the location of ice dispersal centers.

- that may contain misaligned flow traces from an older flow stage
3. *Synchronous fans* that consist of abundant flow traces but lack aligned eskers that may represent the sites of former ice streams or slow ice sheet flow well inside the margin
 4. *Surge fans* with a distinctive bottleneck pattern thought to form near-synchronously and with aligned flow traces in the distal part but not in the up-ice part
 5. *Contraction fans* created by inward migration of a preservation borderline other than the ice front (e.g., by refreezing of the bed)

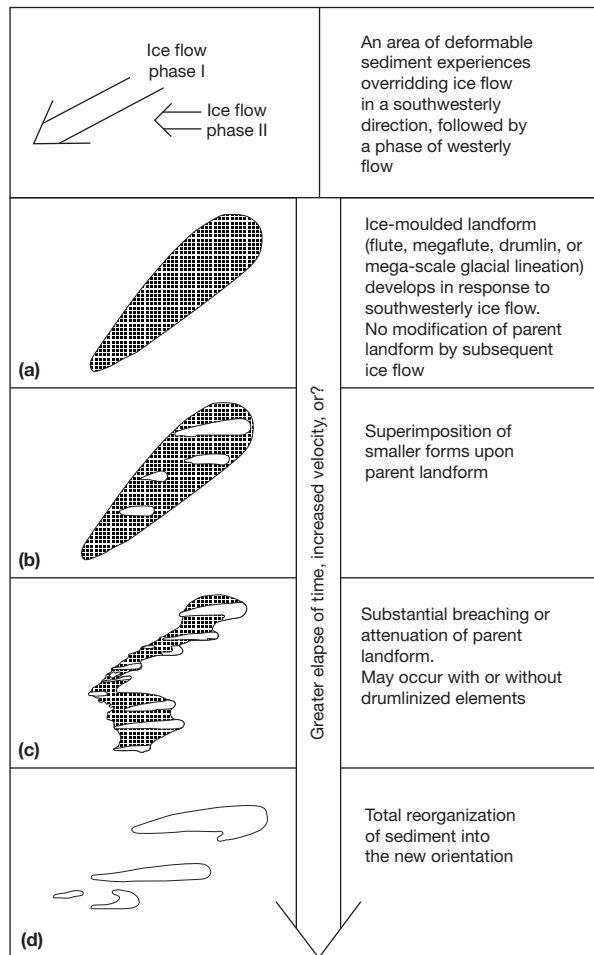
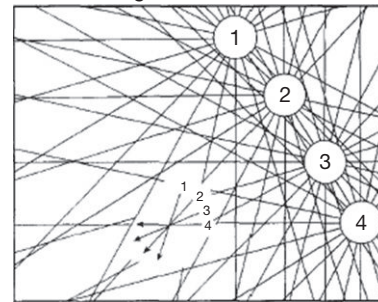


Figure 2 Illustration of the formation of cross-cutting bedform patterns as a result of a switch in ice flow direction. Reproduced from [Clark CD \(1994\)](#) Large-scale ice moulding: A discussion of genesis and glaciological significance. *Sedimentary Geology* 91: 253–268.

6. *Expansion fans* created by outward migration of a preservation borderline other than the ice front (e.g., by refreezing of the bed).

[Kleman and Borgström \(1996\)](#) provided ‘type landscapes’ (cf. type-sites) of fan types 1–4 and outlined the deciphering procedures from defining fan types through to ‘unfolding’ a stack of fans to create a time-slice sequence of ice sheet evolution. An excellent application of these procedures is presented by [Kleman et al. \(1997\)](#), who used the glacial inversion model to reconstruct the evolution of flow pattern in the Fennoscandian Ice Sheet and present time slices from 115 to 9 ka (see [Figure 5](#)). Like [Boulton and Clark \(1990a,b\)](#), their reconstruction identifies several ice sheet configurations with shifts in the location of the main ice divides during buildup and decay. They also note discrepancies between their geomorphologically derived reconstruction and those produced by numerical modeling (e.g., [Holmlund and Fastook, 1995](#)) and highlight the importance of using a

Ice divide migration:



Ice stream activation:



Lobate margin retreat:

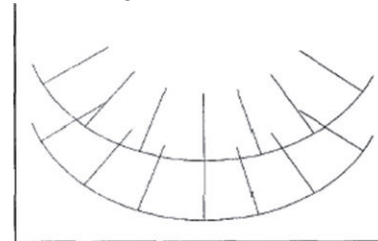


Figure 3 Three conceptual models for the cause of cross-cutting ice flow patterns during a single glacial cycle. A further explanation is that some cross-cutting flow patterns might be produced by more than one glaciation. Reproduced from [Clark CD \(1997\)](#) Reconstructing the evolutionary dynamics of former ice sheets using multitemporal evidence, remote sensing and GIS. *Quaternary Science Reviews* 16: 1067–1092.

geomorphologically based reconstruction as a constraint for a numerical model whose output has the added advantage of providing three-dimensional ice sheet evolution and, critically, ice thicknesses.

Preservation of Geomorphology by Cold-Based Ice and Pre-LGM Ice Flow Direction Indicators

A significant aspect of the [Kleman et al. \(1997\)](#) reconstruction was the recognition of fans that record pre-Last Glacial Maximum (LGM) ice flow directions. Indeed, there are several descriptions in the literature of ‘nonglacial’ terrain that appears to have survived ice sheet occupation with minimal alteration and this has been attributed to the sustained presence of cold-based ice (e.g., [Dyke, 1993](#); [Kleman, 1992, 1994](#); [Kleman and Borgström, 1990, 1994](#); [Lagerbäck, 1988](#)), especially where the nunatak hypothesis can be safely rejected through exposure-age dating ([Hättestrand and Stroeven, 2002](#); [Phillips et al., 2006](#)). [Kleman and Glasser \(2007\)](#) provide an overview of the subglacial thermal organization of ice

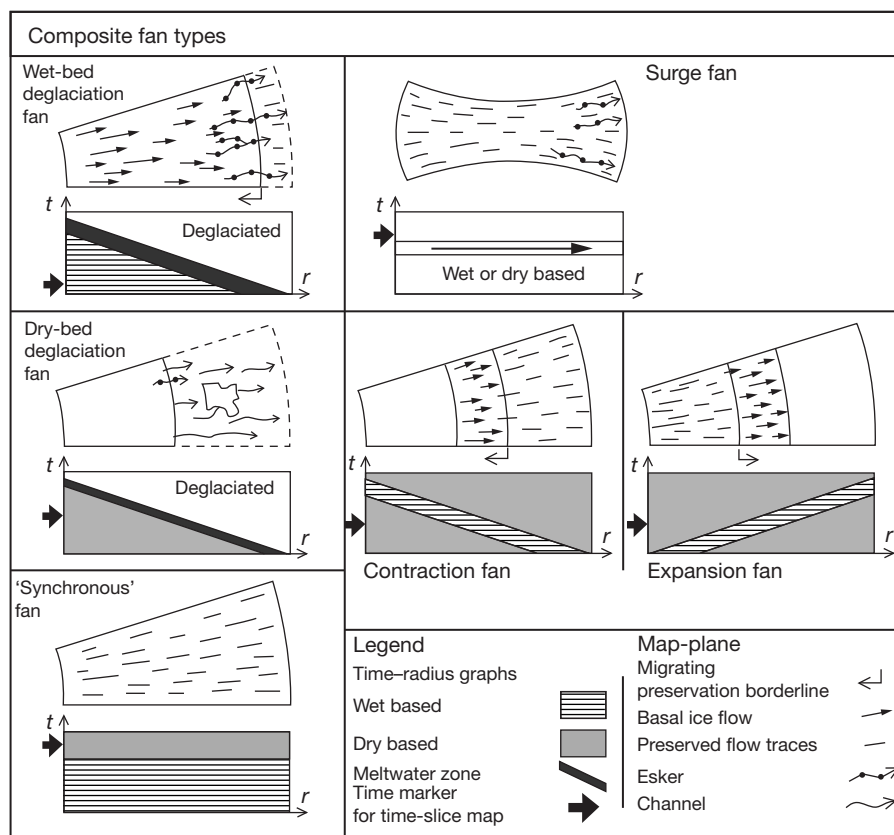


Figure 4 Six fan types (flow-sets) that are considered relevant for use in glacial inversion models for reconstructing paleo-ice sheets. Reproduced from [Kleman J and Borgström I \(1996\)](#) Reconstruction of paleo-ice sheets: The use of geomorphological data. *Earth Surface Processes and Landforms* 21: 893–909.

sheets and describe how frozen-bed patches span a wide range of spatial scales ([Figure 6](#)). In places, they occur as a mosaic of isolated patches or ‘islands’ in upland landscapes, but it is also clear that, at the ice sheet scale, large frozen-bed areas are controlled by dispersal center locations (ice divides) where there is minimal strain heating.

Significantly, some frozen-bed areas preserve ice directional indicators from earlier stages in the glacial cycle and they can be pieced together to reconstruct the inception and buildup of ice sheets prior to the LGM (e.g., [Kleman et al., 1997, 2010](#)). Clearly, compared to the record of deglaciation post-LGM, ice flow directional indicators and associated flow-sets are much more fragmentary, which inevitably results in a spatially incomplete reconstruction. However, the fragmentary record can be integrated with other available evidence (e.g., till stratigraphy, chronological data) to provide a more speculative interpretation of ice sheet outlines ([Kleman et al., 2010](#)). [Kleman et al. \(2010\)](#) provide such a reconstruction for the Laurentide and Cordilleran Ice Sheets on the basis of glacial inversion principles and utilizing flow-sets that predate the LGM. Although somewhat speculative, it is a testable hypothesis and represents one of the first attempts to systematically synthesize the geomorphological and chronological evidence of ice sheet inception and buildup spanning 115–21 ka. Interestingly, [Kleman et al. \(2010\)](#) compare their reconstruction to

those reproduced in numerical modeling experiments and note key areas of discrepancy. For example, while most models generate a rapid growth toward a mono-domed ice sheet early in the last glacial cycle, the geomorphologically derived reconstruction depicts persistent, dynamically independent dispersal centers in Keewatin and Quebec. Furthermore, [Kleman et al. \(2010\)](#) suggest that the location and size of remnant ice masses at the end of interstadials are likely to be an important control on growth centers during subsequent stadials. They also note variations in the southern extent of the ice sheet margin during early/middle Wisconsin stadials, which are not, generally, reproduced by numerical modeling. Again, this highlights the potential for geomorphologically derived reconstructions to inform and test numerical ice sheet modeling.

Paleo-Ice Streaming

Given the importance of rapidly flowing ice streams to ice sheet mass balance and stability (cf. [Bennett, 2003](#)), any ice sheet reconstruction that omits their activity is likely to be deficient ([Stokes and Clark, 2001](#)). Indeed, the highly dynamic behavior of paleo-ice sheets and the associated changes in ice flow direction that have been described earlier were almost certainly driven by ice stream activity. [Kleman et al. \(2006\)](#),

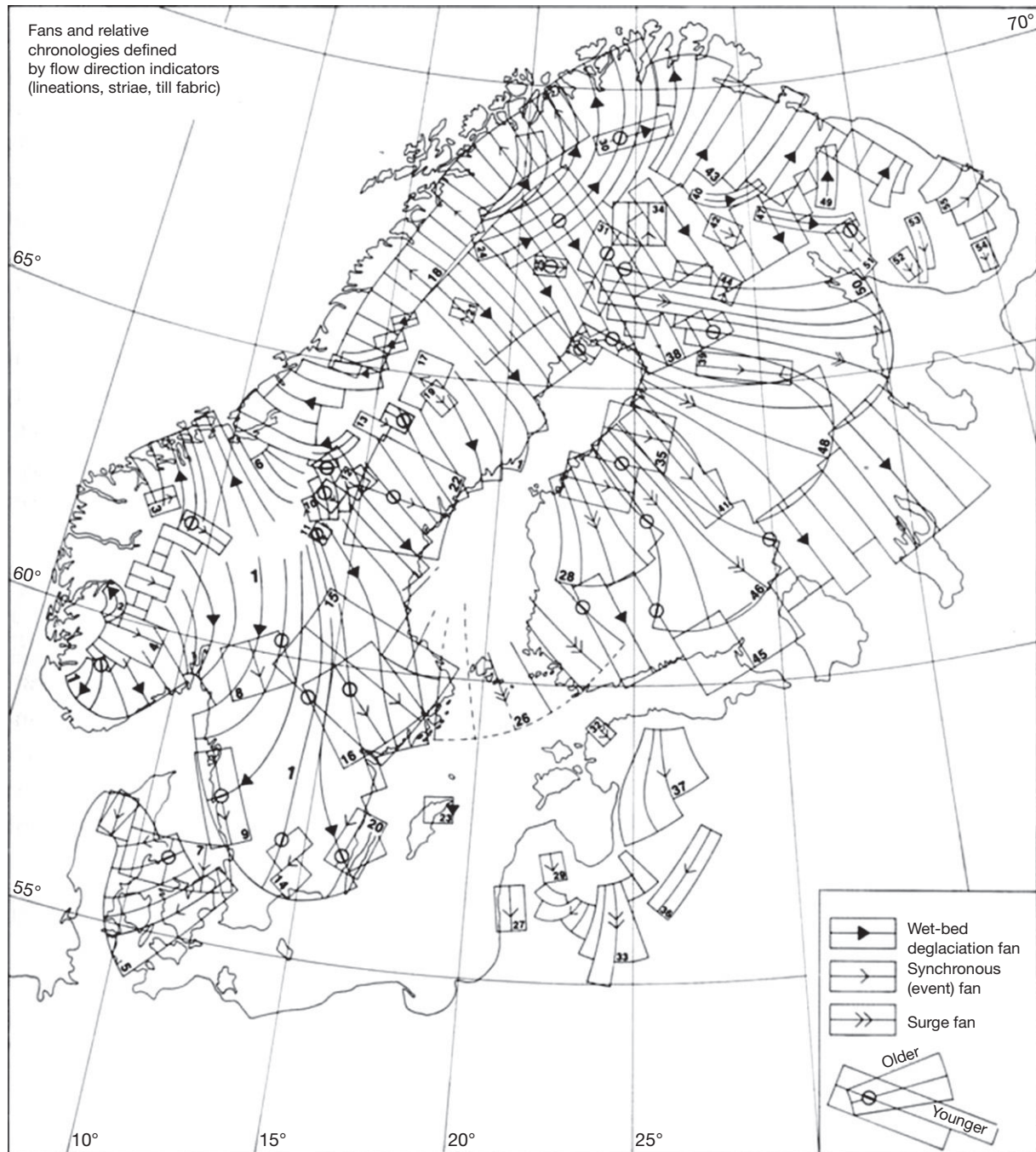


Figure 5 Fans and relative chronologies defined by ice flow direction indicators relating to the Fennoscandian Ice Sheet. These fans are 'unfolded' to produce a time-slice sequence of ice sheet evolution that spans from 115 to 9 ka. Reproduced from [Kleman J, Hättestrand C, Borgström I, and Stroeven, A \(1997\)](#) Fennoscandian palaeoglaciology reconstructed using a glacial inversion model. *Journal of Glaciology* 43: 283–299.

for example, note that the westward migration of the main ice divide in the Fennoscandian Ice Sheet was probably linked to the onset of four major ice streams in Finland during deglaciation (cf. [Boulton et al., 2001](#); [Kleman et al., 1997](#)). Likewise, it has long been recognized that ice streams are required to produce reconstructions of the Laurentide Ice Sheet ([Figure 7](#))

that comply with geological evidence of low-gradient ice surface profiles (cf. [Fisher et al., 1985](#)). It is, perhaps, surprising therefore that explicit recognition of ice streaming has only been incorporated into glacial inversion methods relatively recently.

Early attempts at reconstructing ice stream location (e.g., [Denton and Hughes, 1981](#); [Dyke and Prest, 1987a,b](#)) were

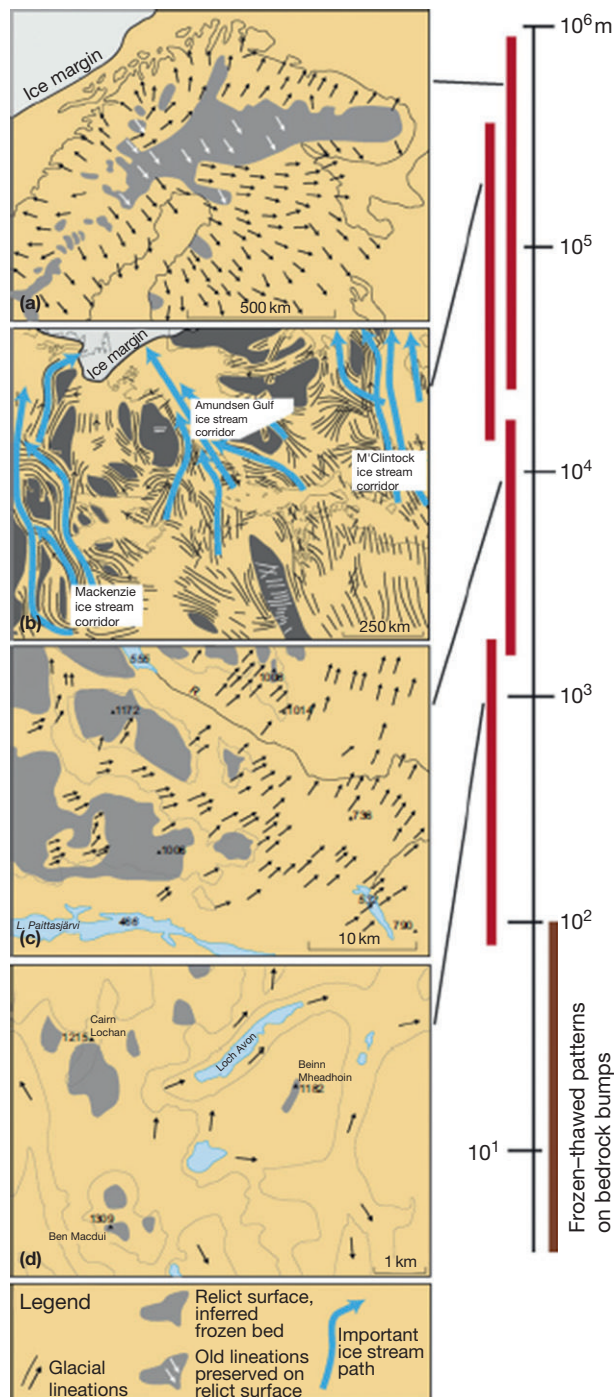


Figure 6 Frozen-bed patches (gray shading) exist over a range of spatial scales from large areas under former ice divides (a) to isolated 'islands' in upland terrain (d). Note how some frozen-bed patches preserve 'old' glacial lineations (white arrows in (a) and (b) that may be related to pre-LGM ice sheet configurations. (a) northern Fennoscandia; (b) northwestern Canada; (c) low mountains west of Kiruna, northern Sweden; (d) the Cairngorm Mountains, Scotland. Reproduced from Kleman J and Glasser NF (2007) The subglacial thermal organisation (STO) of ice sheets. *Quaternary Science Reviews* 26: 585–597.

augmented with pioneering studies that identified specific geological evidence of their activity, such as distinctive erratic dispersal trains with abrupt lateral margins (Dyke and Morris, 1988; Hicock, 1988) (Figure 8). However, Mathews (1991) noted that there were no diagnostic criteria for distinguishing paleo-ice stream flow-sets and, with this in mind, Stokes and Clark (1999) proposed several specific features that would be expected to characterize a paleo-ice stream 'footprint.' These features were based on a convergence of knowledge gleaned from hypothesized paleo-ice stream locations (e.g., Dyke and Morris, 1988; Hicock, 1988) and the characteristics of contemporary ice streams. These 'geomorphological criteria' include a characteristically large and convergent flow-set, highly attenuated mega-scale glacial lineations with abrupt lateral margins, lateral shear margin moraines, distinctive erratic dispersal trains, and glaciotectionic and/or geotechnical evidence of pervasively deformed till that may accumulate as a till delta or trough mouth fan in marine environments. Thus, although Kleman and Borgström (1996) had speculated that their 'synchronous' and 'surge fans' may be produced by ice streams, the criteria proposed by Stokes and Clark (1999) were explicitly describing a paleo-ice stream land system (Figure 9).

The geomorphological criteria act as an observational template for glacial inversion techniques and numerous paleo-ice streams have been identified in the last decade, such that there is now a fairly comprehensive knowledge of their location from most paleo-ice sheets (Figure 10). Moreover, recent work has revealed abrupt changes in the velocity and trajectory of paleo-ice streams, known as 'flow switching' (e.g., Dowdeswell et al., 2006; Ó Cofaigh et al., 2010; Stokes et al., 2009; Winsborrow et al., 2012), which has also been reported in contemporary ice sheets (Conway et al., 2002) and reproduced in numerical models (Payne and Dongelmans, 1997). These reconstructions provide new insights into the long-term behavior of ice streams (and their potential forcing) and often help to reconcile complex cross-cutting flow patterns. Indeed, ice stream flow-sets are often seen to encroach, superimpose, and even 'behead' other types of flow-sets (cf. Dyke and Morris, 1988; Stokes et al., 2009).

Once identified, the 'known' location of ice streams provides a useful observational record to test numerical models of ice streaming. Given the current attention to improving ice sheet models, there is an urgent need to understand the properties (i.e., boundary conditions) that initiate, sustain, or inhibit ice streaming. This can be achieved through a combined approach that evaluates model predictions of ice streaming against independent geological evidence of their activity. Very few studies have attempted a direct data-model comparison for an entire ice sheet, but Stokes and Tarasov (2010) found generally good agreement for the Laurentide Ice Sheet, although there were some issues in reproducing ice streams that terminated on land that will guide future model development. Evans et al. (2009) also found good agreement between their geomorphologically derived reconstruction of the central sector of the British and Irish Ice Sheet and a simple numerical ice sheet model. Both the geomorphology and the model appear to depict continuous migration of ice dispersal centers, and the model has the additional

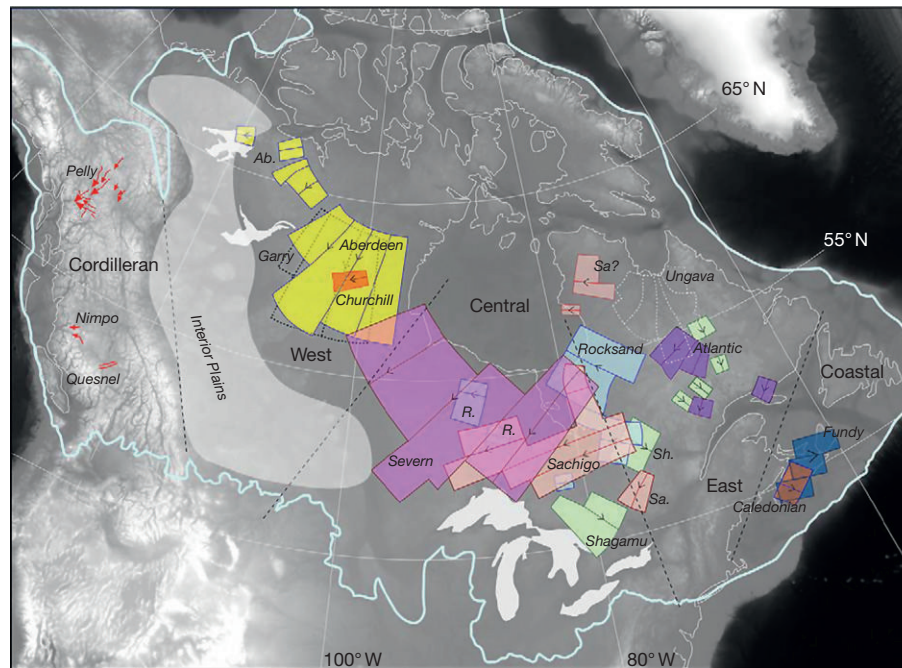


Figure 7 Landform swarms of the former Laurentide Ice Sheet thought to be older than the LGM. Despite their fragmentary nature compared to post-LGM records, these flow-sets can be used to reconstruct the evolutionary sequence of the inception and buildup of the ice sheet (see [Figure 8](#)). Reproduced from [Kleman J, Jansson K, De Angelis H, et al. \(2010\)](#) North American Ice Sheet build-up during the last glacial cycle, 115–21 ka. *Quaternary Science Reviews* 29(17–18): 2036–2051.

advantage of providing a guide to the duration of various flow phases, with some modeled flow reversals taking place in <300 years.

[Stokes and Tarasov \(2010\)](#) identified that a major limitation of reconstructions of paleo-ice streams from geomorphological evidence is the lack of dating control on ice stream flow-sets. In this sense, numerical models can provide useful insights into the likelihood of streaming duration at given locations (cf. [Evans et al., 2009](#)), but there is a clear need for additional chronological constraints on ice stream timing to further refine ice sheet reconstructions. This applies equally to non-streaming flow-sets but, if enough radiocarbon dates are available to provide minimum ages on deglaciation, it is possible to date flow-sets and provide an ice sheet reconstruction incorporating ice stream activity up to 250–500-year time steps (see [Figure 11](#); cf. [Stokes et al., 2009](#)).

Ice Flow Indicators Under Contemporary Ice Sheets

One of the most obvious deficiencies of the glacial inversion methods described earlier is that our process understanding of how various landforms are produced is rather limited. This is what [Kleman et al. \(2006\)](#) refer to as the ‘genetic problem,’ that is, deciphering the processes by which, and the conditions under which, particular landforms are created. As [Kleman et al. \(2006\)](#) note, observations of subglacial processes are often

restricted to marginal or thin ice (<100 m) environments or obtained from borehole or geophysical techniques which are limited by their ‘spot’ location and/or poor resolution. Unfortunately, therefore, the landforms that are most important in glacial inversion methods (e.g., glacial lineations and ribbed moraines) are least understood in terms of their genesis. Nevertheless, accumulated knowledge from glacial geomorphological mapping and inversion techniques has contributed some insights into the conditions under which certain landforms are created. For example, the discovery of mega-scale glacial lineations (cf. [Clark, 1993](#)) and their association with paleo-ice stream tracks (cf. [Stokes and Clark, 1999, 2001](#)) has led some to propose that bedform elongation ratio (length:width) is related to ice velocity ([Hart, 1999](#); [Stokes and Clark, 2002](#)), although precisely how a bedform is attenuated is debatable. In contrast, ribbed moraines are rarely associated with fast-flowing ice and are more likely to represent areas of the ice sheet bed associated with oscillations in the basal thermal regime ([Dyke and Morris, 1988](#); [Hättestrand and Kleman, 1999](#); [Kleman and Hättestrand, 1999](#)) and/or possible sticky spots ([Stokes et al., 2008](#)).

Crucially, recent advances in geophysical observations have allowed, for the first time, bedforms to be imaged beneath existing ice sheets at depths of around 2 km below the ice surface ([King et al., 2007, 2009](#); [Smith and Murray, 2009](#); [Smith et al., 2007](#)). Specifically, drumlins appear to be recorded in the onset zone of the Rutford Ice Stream, West Antarctica, where ice velocities accelerate from 72 to >200 m a⁻¹

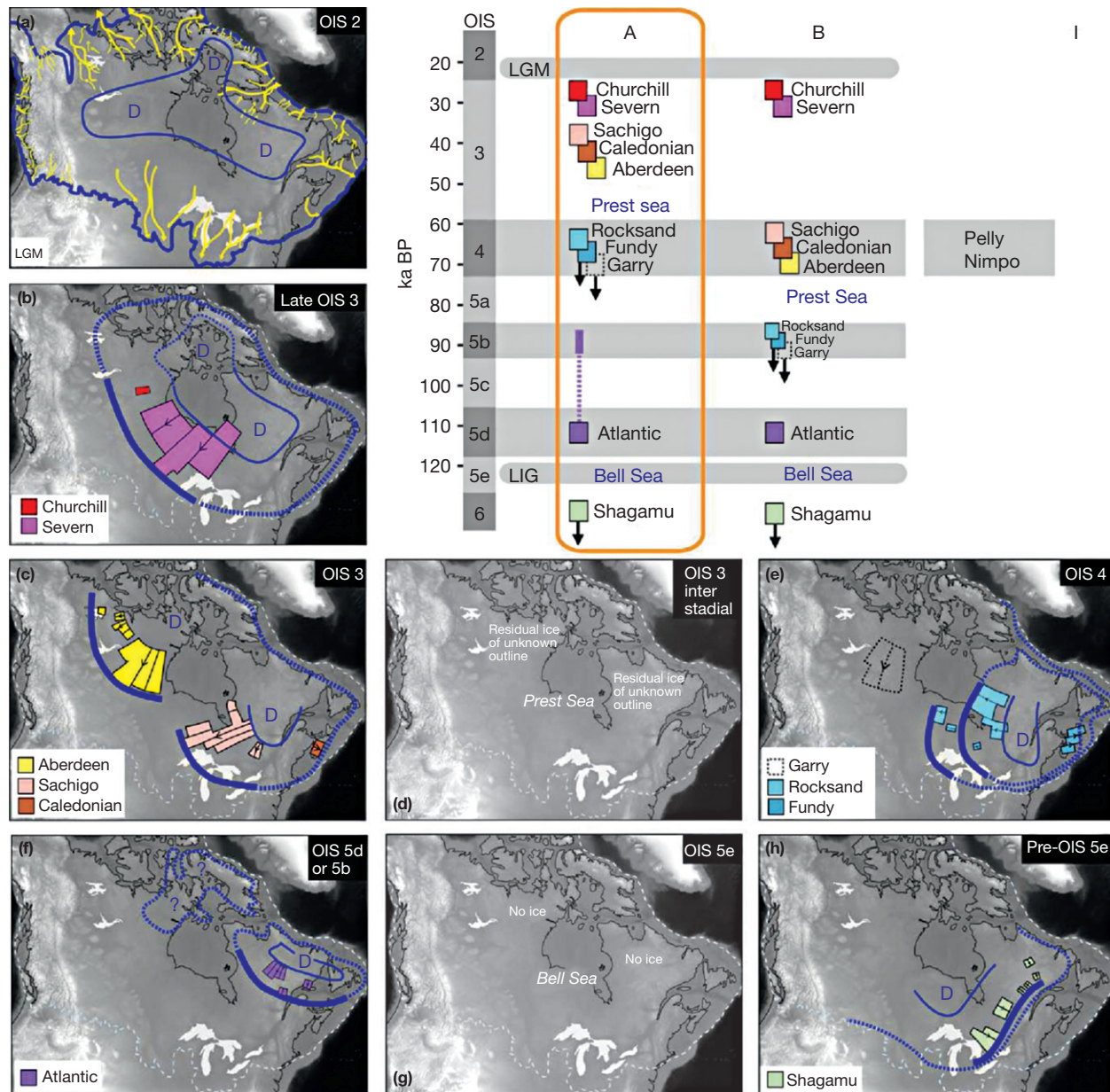


Figure 8 Reconstructed evolutionary history of the Laurentide Ice Sheet (LIS) based on the flow-sets shown in [Figure 7](#) and with extrapolations based on topographic conditions, glaciological plausibility, and stratigraphical data. Column A gives the preferred age interpretation and column B assigns the Rocksand till to oxygen isotope stage (OIS) 5. LGM, Last Glacial Maximum; LIG, Last Inter-Glaciation. Reproduced from [Kleman J, Jansson K, De Angelis H, et al. \(2010\) North American Ice Sheet build-up during the last glacial cycle, 115–21 ka. *Quaternary Science Reviews* 29\(17–18\): 2036–2051.](#)

([King et al., 2007](#)). Further down the ice stream, where velocities increase to around 375 m a^{-1} , [King et al. \(2009\)](#) report the presence of mega-scale glacial lineations that are indistinguishable from those reported on paleo-ice stream beds ([Figure 12](#)). These observations of the bed confirm the inference from paleoglaciology that bedform elongation is related to ice velocity, but, more importantly, repeat surveys have shown that these bedforms are evolving over very short time-scales, with material being eroded at rates of $\sim 1 \text{ m a}^{-1}$ and

remobilized to form bedforms further downstream, with associated changes in subglacial hydrology ([King et al., 2009; Smith et al., 2007](#)).

Arguably, the most important aspect of these observations is that the formation of bedforms can be linked to specific glaciological conditions (e.g., ice thickness, velocity, and subglacial hydrology), which specifically address the genetic problem. In this sense, observations of the bed of modern-day ice sheets provide the ‘missing link’ between glaciology and

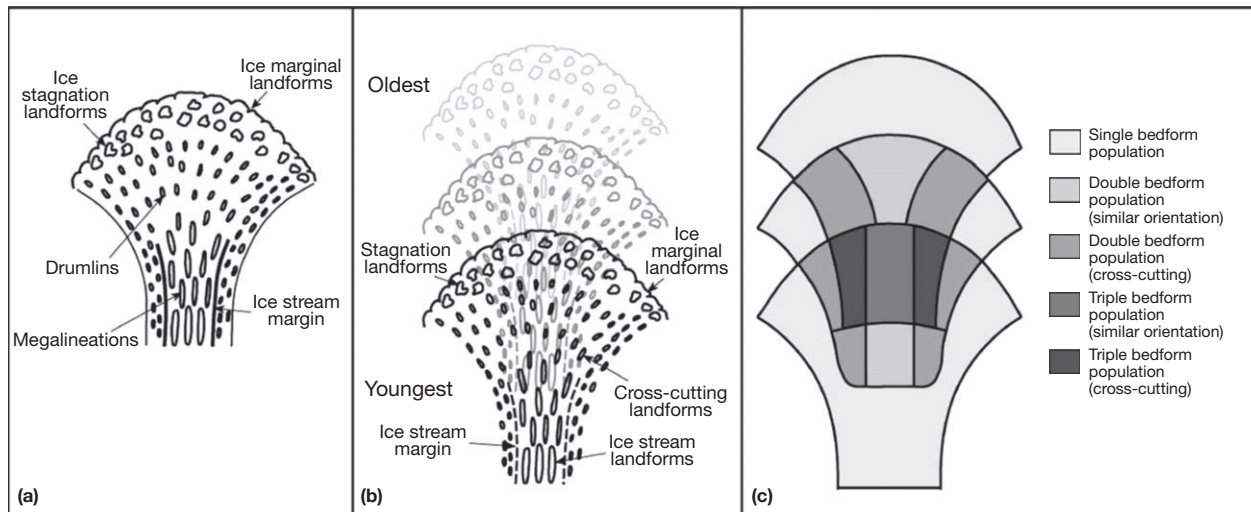


Figure 9 The paleo-ice stream land system of a terrestrially terminating ice stream and the difference between an unaltered imprint (a) and that produced by a retreating ice stream (b) that continually remolds its bed and creates several populations of bedforms (c). Reproduced from Stokes CR and Clark CD (2001) Palaeo-ice streams. *Quaternary Science Reviews* 20: 1437-1457.

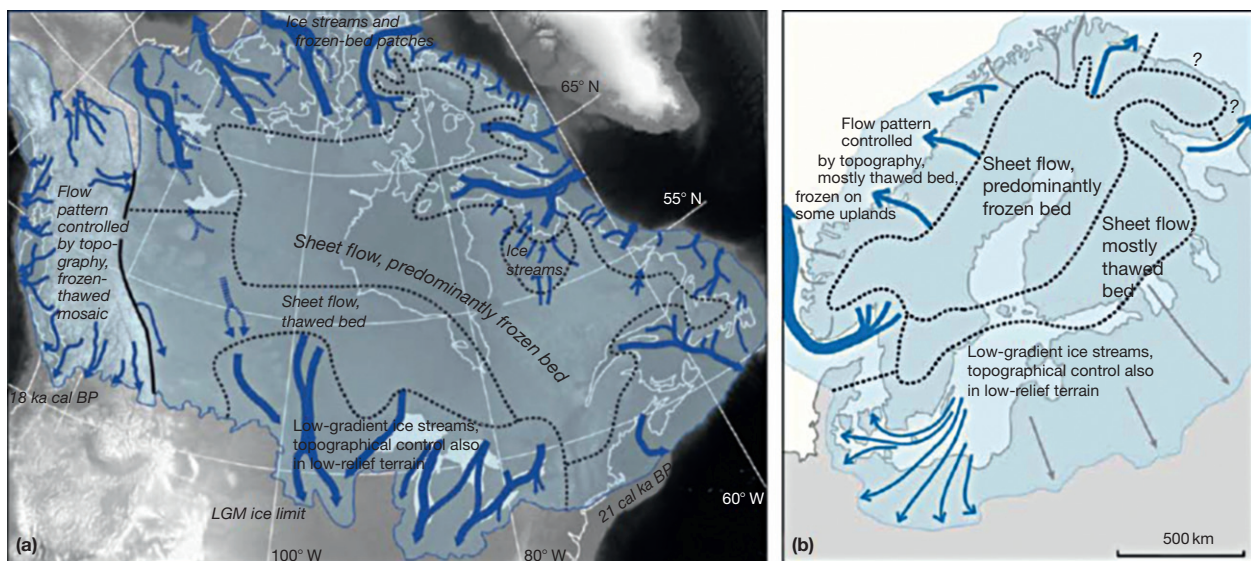
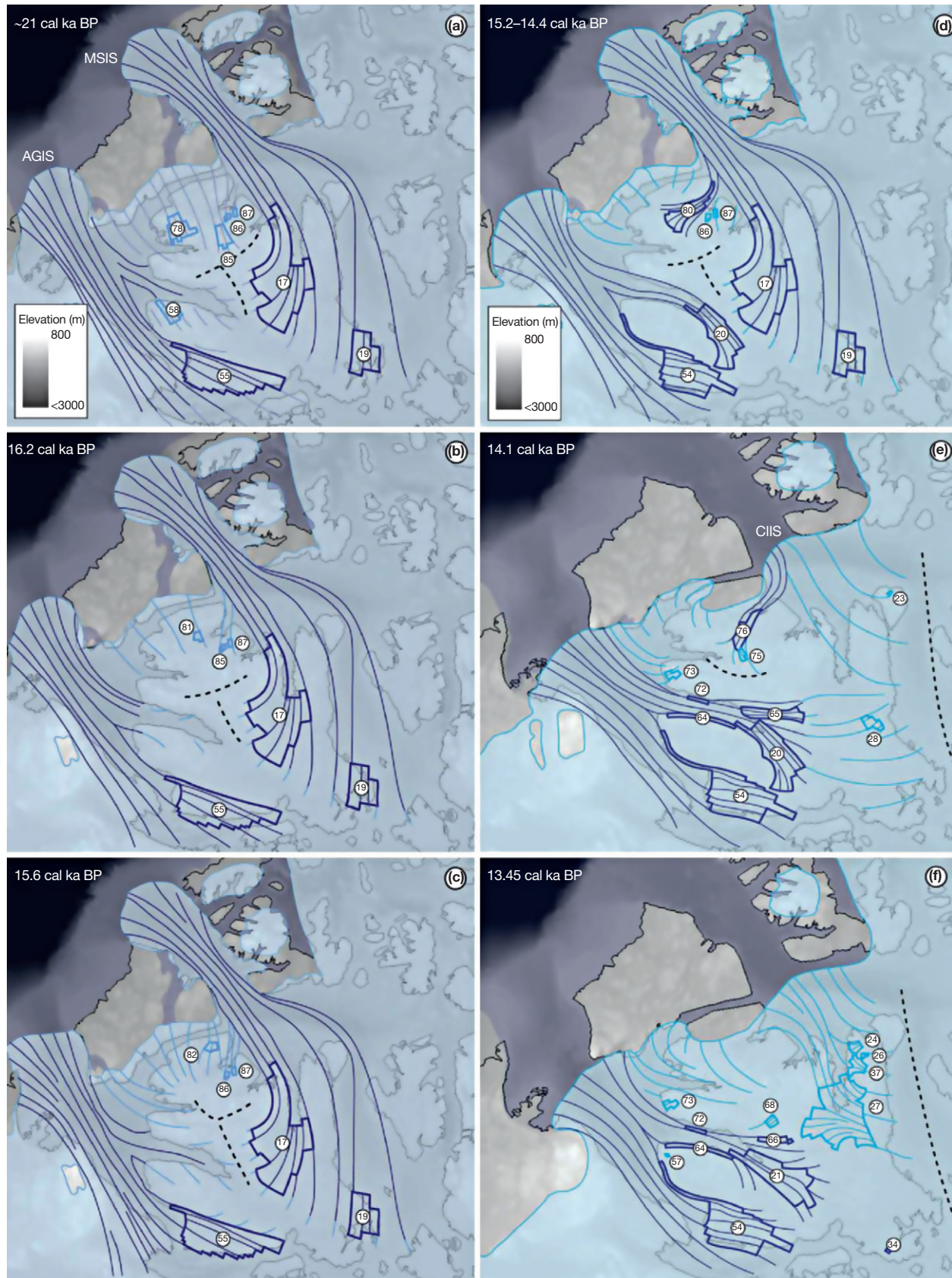


Figure 10 Ice stream locations in the North American and Fennoscandian Ice Sheets at the LGM or near-LGM stages and the subglacial thermal organization of the ice sheet. Reproduced from Kleman J and Glasser NF (2007) The subglacial thermal organisation (STO) of ice sheets. *Quaternary Science Reviews* 26: 585-597.

paleoglaciology and will considerably refine glacial inversion methods. At the same time, one can assume that future technological advances will permit whole landscapes to be imaged beneath the ice in Greenland and Antarctica, and glacial inversion techniques that have been applied to paleo-ice sheets provide the only observational template to interpret what is likely to be an incredibly complex assemblage of cross-cutting landforms, which will reveal important aspects of the ice sheet's evolutionary history.

See also: **Glacial Landforms:** Glaciofluvial Landforms of Erosion; Moraine Forms and Genesis. **Glacial Landforms, Erosional Features:** Major Scale Forms; Micro- to Macroscale Forms. **Glacial Landforms, Ice Sheets:** Evidence of Glacier and Ice Sheet Extent; Growth and Decay. **Glacial Landforms, Sediments:** Glacial Erratics and Till Dispersal Indicators; Glaciofluvial Landforms of Deposition; Tills.

**Figure 11** (Continued)

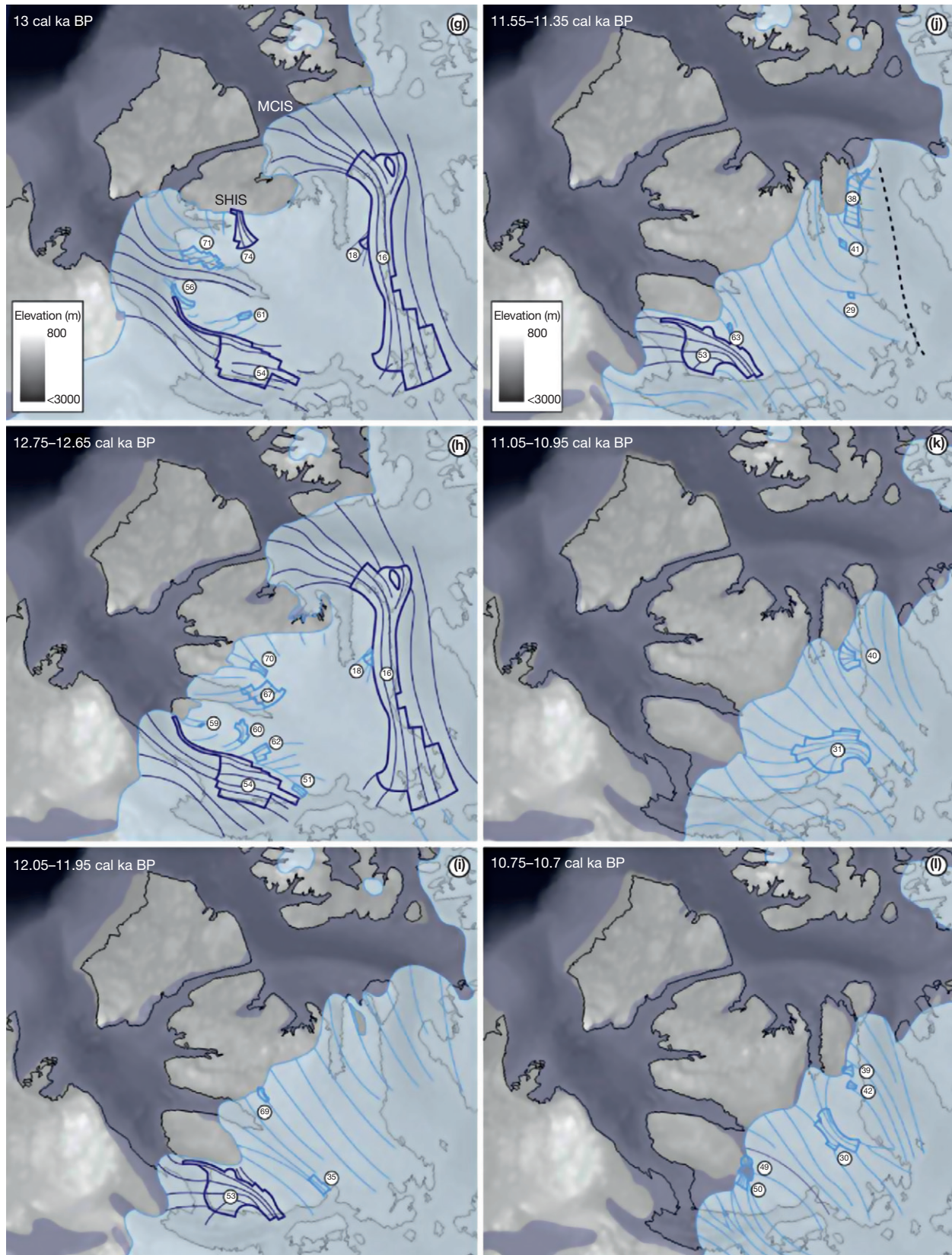


Figure 11 Ice sheet reconstruction at the northwestern extent of the Laurentide Ice Sheet showing how flow-sets can be dated using both cross-cutting relationships and radiocarbon dates and embedded with an ice margin chronology. Dark blue lines represent ice stream flow lines and light blue lines represent sheet flow. Note the contrasting behavior of the Amundsen Gulf Ice Stream (AGIS) and the M'Clure Strait Ice Stream (MSIS) and the rapid on-off switching of the M'Clintock Channel Ice Stream (MCIS). CIIS, Collinson Inlet Ice Stream; SHIS, Saneraun Hills Ice Stream. Reproduced from [Stokes CR, Clark CD, and Storrar R \(2009\)](#) Major changes in ice stream dynamics during deglaciation of the north-western margin of the Laurentide Ice Sheet. *Quaternary Science Reviews* 28: 721–738.

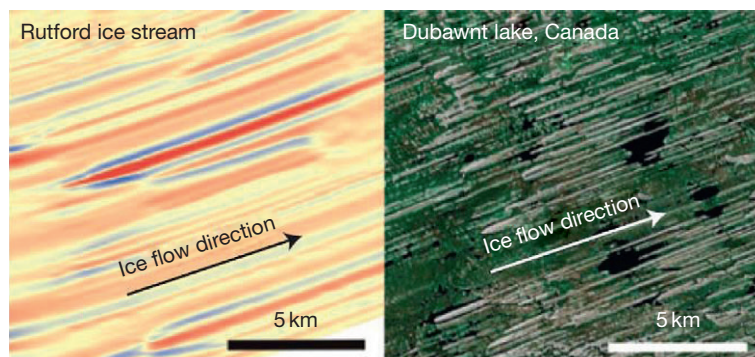


Figure 12 Comparison of mega-scale glacial lineations from beneath Rutford Ice Stream, West Antarctica (King et al., 2009) with those identified on the Dubawnt Lake paleo-ice stream bed in central Canada (Stokes and Clark, 2003).

References

- Aario R (1977) Classification and terminology of morainic landforms in Finland. *Boreas* 6: 87–100.
- Alley RB, Clark PU, Huybrechts P, and Joughin I (2005) Ice-sheet and sea-level changes. *Science* 310: 456–460.
- Bennett MR (2003) Ice streams as the arteries of an ice sheet: Their mechanics, stability and significance. *Earth-Science Reviews* 61(3–4): 309–339.
- Boulton GS and Clark CD (1990a) A highly mobile Laurentide Ice Sheet revealed by satellite images of glacial lineations. *Nature* 346: 813–817.
- Boulton GS and Clark CD (1990b) The Laurentide Ice Sheet through the last glacial cycle: The topology of drift lineations as a key to the dynamic behaviour of former ice sheets. *Transactions of the Royal Society of Edinburgh: Earth Sciences* 81: 327–347.
- Boulton GS, Dongelmans P, Punkari M, and Broadgate M (2001) Palaeoglaciology of an ice sheet through a glacial cycle: The European Ice Sheet through the Weichselian. *Quaternary Science Reviews* 20: 591–625.
- Boulton GS, Smith GD, Jones AS, and Newsome J (1985) Glacial geology and glaciology of the last mid-latitude ice sheets. *Journal of the Geological Society of London* 142: 447–474.
- Clark CD (1993) Mega-scale glacial lineations and cross-cutting ice-flow landforms. *Earth Surface Processes and Landforms* 18: 1–29.
- Clark CD (1994) Large-scale ice moulding: A discussion of genesis and glaciological significance. *Sedimentary Geology* 91: 253–268.
- Clark CD (1997) Reconstructing the evolutionary dynamics of former ice sheets using multi-temporal evidence, remote sensing and GIS. *Quaternary Science Reviews* 16: 1067–1092.
- Clark CD (1999) Glaciodynamic context of subglacial bedform generation and preservation. *Annals of Glaciology* 28: 23–32.
- Clark CD, Hughes ALC, Greenwood SL, Jordan C, and Sejrup HP (2010) Pattern and timing of retreat of the last British–Irish Ice Sheet. *Quaternary Science Reviews* 44: 112–146.
- Clark CD, Hughes ALC, Greenwood SL, Spagnolo M, and Ng FSL (2009) Size and shape characteristics of drumlins, derived from a large sample, and associated scaling laws. *Quaternary Science Reviews* 28(7–8): 677–692.
- Conway H, Catania G, Raymond CF, Gades AM, Scambos TA, and Engelhardt H (2002) Switch of flow in an Antarctic ice stream. *Nature* 419: 465–467.
- Denton GH and Hughes TJ (eds.) (1981) *The Last Great Ice Sheets*. New York: Wiley.
- Dowdeswell JA, Ottesen D, and Rise L (2006) Flow switching and large-scale deposition by ice streams draining former ice sheets. *Geology* 34: 313–316.
- Dyke AS (1993) Landscapes of cold-centred Late Wisconsinan ice caps, Arctic Canada. *Progress in Physical Geography* 17(2): 223–247.
- Dyke AS and Morris TF (1988) Drumlin fields, dispersal trains, and ice streams in Arctic Canada. *The Canadian Geographer* 32: 86–90.
- Dyke AS and Prest VK (1987a) Late Wisconsinan and Holocene history of the Laurentide Ice Sheet. *Géographie Physique et Quaternaire* 41: 237–262.
- Dyke AS and Prest VK (1987b) Palaeogeography of Northern North America, 18000–5000 Years Ago. Ottawa: Geological Survey of Canada, Queen's Printer. Map 1703A.
- Ehlers J and Gibbard P (eds.) (2004) *Quaternary Glaciations – Extent and Chronology, Vol. 1, Europe. Developments in Quaternary Science*. Amsterdam: Elsevier.
- Evans DJA (2003) *Glacial Landscapes*, p. 532. London: Arnold.
- Evans DJA, Livingstone SJ, Vieli A, and Ó Cofaigh C (2009) The palaeoglaciology of the central sector of the British and Irish Ice Sheet: Reconciling glacial geomorphology and preliminary ice sheet modelling. *Quaternary Science Reviews* 28(7–8): 739–757.
- Fisher DA, Reeh N, and Langley K (1985) Objective reconstructions of the Late Wisconsinan Laurentide Ice Sheet and the significance of deformable beds. *Géographie Physique et Quaternaire* 39: 229–238.
- Hart JK (1999) Identifying fast ice flow from the landform assemblages in the geological record: A discussion. *Annals of Glaciology* 28: 59–66.
- Hättestrand C and Kleman J (1999) Ribbed moraine formation. *Quaternary Science Reviews* 18: 43–61.
- Hättestrand C and Stroeven AP (2002) A relict landscape in the centre of the Fennoscandian glaciations: Geomorphological evidence of minimal Quaternary glacial erosion. *Geomorphology* 44: 27–143.
- Hicock SR (1988) Calcareous till facies north of Lake Superior, Ontario: Implications for Laurentide ice streaming. *Géographie Physique et Quaternaire* 42: 120–135.
- Holmlund P and Fastook J (1995) A time-dependent glaciological model of the Weichselian Ice Sheet. *Quaternary International* 27: 53–58.
- King EC, Hindmarsh RCA, and Stokes CR (2009) Formation of mega-scale glacial lineations observed beneath a West Antarctic ice stream. *Nature Geoscience* 2(8): 529–596.
- King EC, Woodward J, and Smith AM (2007) Seismic and radar observations of subglacial bed forms beneath the onset zone of Rutford Ice Stream, Antarctica. *Journal of Glaciology* 53: 665–672.
- Kleman J (1992) The palimpsest glacial landscape in northwestern Sweden: Late Weichselian deglaciation landforms and traces of older west-centred ice sheets. *Geografiska Annaler* 74A: 305–325.
- Kleman J (1994) Preservation of landforms under ice sheets and ice caps. *Geomorphology* 9: 19–32.
- Kleman J and Borgström I (1990) The boulder fields of Mt. Fuluñjället, west-central Sweden – Late Weichselian boulder blankets and interstadial periglacial phenomena. *Geografiska Annaler* 72A(1): 63–78.
- Kleman J and Borgström I (1994) Glacial landforms indicative of a partly frozen bed. *Journal of Glaciology* 135: 255–264.
- Kleman J and Borgström I (1996) Reconstruction of paleo-ice sheets: The use of geomorphological data. *Earth Surface Processes and Landforms* 21: 893–909.
- Kleman J and Glasser NF (2007) The subglacial thermal organisation (STO) of ice sheets. *Quaternary Science Reviews* 26: 585–597.
- Kleman J, Hättestrand C, Stroeven AP, Jansson KN, De Angelis H, and Borgström I (2006) Reconstruction of palaeo-ice sheets – Inversion of their glacial geomorphological record. In: Knight PG (ed.) *Glacier Science and Environmental Change*, p. 527. Oxford: Blackwell.
- Kleman J and Hättestrand C (1999) Frozen-bed Fennoscandian and Laurentide ice sheets during the Last Glacial Maximum. *Nature* 402: 63–66.
- Kleman J, Hättestrand C, Borgström I, and Stroeven A (1997) Fennoscandian palaeoglaciology reconstructed using a glacial inversion model. *Journal of Glaciology* 43: 283–299.
- Kleman J, Jansson K, De Angelis H, et al. (2010) North American Ice Sheet build-up during the last glacial cycle, 115–21 kyr. *Quaternary Science Reviews* 29(17–18): 2036–2051.

- Lagerbäck R (1988) Periglacial phenomena in the wooded areas of Northern Sweden – Relicts from the Tändö interstadial. *Boreas* 17: 487–499.
- Mathews WH (1991) Ice sheets and ice streams: Thoughts on the Cordilleran Ice Sheet symposium. *Géographie Physique et Quaternaire* 45(3): 263–267.
- McMartin I and Henderson PJ (2004) Evidence from Keewatin (Central Nunavut) for paleo-ice divide migration. *Géographie Physique et Quaternaire* 58(2–3): 163–186.
- McMartin I and McClenaghan MB (2001) Till geochemistry and sampling techniques in glaciated shield terrain: A review. In: McClenaghan MB, Bobrowsky PT, Hall GEM, and Cook SJ (eds.) *Drift Exploration in Glaciated Terrain* 185: 19–43. London: Geological Society of London. Geological Society Special Publication.
- Menzies J (1987) Towards a general hypothesis on the formation of drumlins. In: Menzies J and Rose J (eds.) *Drumlin Symposium*, pp. 9–24. Rotterdam: A. A. Balkema.
- Ó Cofaigh C, Evans DJA, and Smith IR (2010) Large-scale reorganization and sedimentation of terrestrial ice streams during late Wisconsinan Laurentide Ice Sheet deglaciation. *Geological Society of America Bulletin* 122: 743–756.
- Payne AJ and Dongelmans PW (1997) Self-organisation in the thermomechanical flow of ice sheets. *Journal of Geophysical Research* 102: 12219–12234.
- Phillips WM, Hall AM, Mottram R, Fifield LK, and Sugden DE (2006) Cosmogenic ^{10}Be and ^{26}Al exposure ages of tors and erratic, Cairngorm Mountains, Scotland: Time scales for the development of a classic landscape of selective linear glacial erosions. *Geomorphology* 73: 222–245.
- Roberts DH and Long AJ (2005) Streamlined bedrock terrain and fast ice flow, Jakobshavns Isbrae, West Greenland: Implications for ice stream and ice sheet dynamics. *Boreas* 34(1): 25–42.
- Roberts DH, Long AJ, Davies BJ, Smipson MJR, and Schnabel C (2010) Ice stream influence on West Greenland Ice Sheet dynamics during the Last Glacial Maximum. *Journal of Quaternary Science* 25(6): 850–864.
- Rose J (1987) Drumlins as part of a glacier bedform continuum. In: Menzies J and Rose J (eds.) *Drumlin Symposium*, pp. 103–116. Rotterdam: A. A. Balkema.
- Shilts WW (1980) Flow patterns in the central North American Ice Sheet. *Nature* 286: 213–218.
- Shilts WW, Cunningham CM, and Kaszycki CA (1979) Keewatin Ice Sheet – Re-evaluation of the traditional concept of the Laurentide Ice Sheet. *Geology* 7: 537–541.
- Smith AM and Murray T (2009) Bedform topography and basal conditions beneath a fast-flowing West Antarctic ice stream. *Quaternary Science Reviews* 28: 584–596.
- Smith AM, Murray T, Nicholls KW, et al. (2007) Rapid erosion, drumlin formation, and changing hydrology beneath an Antarctic ice stream. *Geology* 35: 127–130.
- Spagnolo M, Clark CD, Hughes ALC, and Dunlop P (2011) The topography of drumlins; assessing their long profile shape. *Earth Surface Processes and Landforms* 36: 790–804.
- Spagnolo M, Clark CD, Hughes ALC, Dunlop P, and Stokes CR (2010) The planar shape of drumlins. *Sedimentary Geology* 232: 119–129.
- Stokes CR and Clark CD (1999) Geomorphological criteria for identifying Pleistocene ice streams. *Annals of Glaciology* 28: 67–75.
- Stokes CR and Clark CD (2001) Palaeo-ice streams. *Quaternary Science Reviews* 20: 1437–1457.
- Stokes CR and Clark CD (2002) Are long bedforms indicative of fast ice flow? *Boreas* 31: 239–249.
- Stokes CR and Clark CD (2003) The Dubawnt Lake palaeo-ice stream: Evidence for dynamic ice sheet behaviour on the Canadian Shield and insights regarding the controls on ice stream location and vigour. *Boreas* 32: 263–279.
- Stokes CR, Clark CD, and Storrar R (2009) Major changes in ice stream dynamics during deglaciation of the north-western margin of the Laurentide Ice Sheet. *Quaternary Science Reviews* 28: 721–738.
- Stokes CR, Lian OB, Tulaczyk S, and Clark CD (2008) Superimposition of ribbed moraines on a palaeo-ice stream bed: Implications for ice stream dynamics and shut-down. *Earth Surface Processes and Landforms* 33: 593–609.
- Stokes CR, Spagnolo M, and Clark CD (2011) The composition and internal structure of drumlins: Complexity versus commonality. *Earth-Science Reviews* 107: 398–422.
- Stokes CR and Tarasov L (2010) Ice streaming in the Laurentide Ice Sheet: A first comparison between data-calibrated numerical model output and geological evidence. *Geophysical Research Letters* 37: L01501. <http://dx.doi.org/10.1029/2009GL040990>, 2010.
- Tarasov L and Peltier WR (2004) A geophysically constrained large ensemble analysis of the deglacial history of the North American ice-sheet complex. *Quaternary Science Reviews* 23: 359–388.
- Winsborrow MCM, Stokes CR, and Andreassen K (2012) Ice stream flow-switching during deglaciation of the south-western Barents Sea. *Geological Society of America Bulletin* 124(3/4): 275–290.

Paleo-ELAs

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Glossary

Ablation: All processes that reduce the glacier mass, including calving.

Ablation zone: The part of the glacier where summer melting exceeds winter accumulation. This includes not only the total melting of the snow cover of the last winter but also a layer of glacier ice. A deficit of mass appears in that area. The zone lies at lower altitudes of the glacier surface. The ablation zone meets the accumulation zone at the equilibrium line.

Accumulation: All processes that increase the glacier mass. Winter snowfalls are the most important source of mass gain. Redepositions of snow by wind and avalanche are important factors on cirque glaciers and on glaciers surrounded by large mountain plateaus and steep valley sides.

Accumulation area ratio (AAR): The ratio of the accumulation zone to the entire glacier with respect to any particular year. The AAR is an indicator of the glacier balance state in the observation year. The ratio is expressed as a proportion of the total area of the glacier.

Accumulation zone: The part of the glacier where snow that has accumulated during the winter does not melt completely in the subsequent summer. An increase of mass is observed in this area. The zone lies normally in the upper part of the glacier. The accumulation zone meets the ablation zone at the equilibrium line.

Annual ablation: The mass loss during 1 measurement year in the fixed date system.

Annual accumulation: The mass gain to the glacier during one measurement year in the fixed date system.

Annual balance: The sum of the annual accumulation (positive) and the annual ablation (negative) at the end of the measurement year (balance year). This term is used in the fixed date system of measuring and reporting of mass balance. Total values are averaged over the entire glacier surface and presented in terms of equivalent water layer (in meters).

Balance year: The time between dates of formation of two consecutive summer surfaces, commonly understood as the time between the beginning of the winter accumulation and the end of the ablation in the subsequent summer (the date of the minimum summer balance). The balance year is rarely exactly equal to one calendar year.

Calving: The process of mass loss with respect to tidewater glaciers (glaciers terminating in the sea or in a lake) and ice shelves (detachment of icebergs).

Climatic equilibrium line: The mean annual equilibrium line over a 30-year period.

Cumulative mass balance: The mass balance summed from particular years of an observation period, which indicates a general trend of the glacier mass to have either grown or shrunk.

Equilibrium line: A line joining points on a glacier surface where winter balance equals summer balance. Normally, this is a line or narrow zone where the summer melting entirely removes the winter snow cover but not any older ice or firn below this. The line separates the accumulation zone from the ablation zone.

Equilibrium-line altitude (ELA): The altitude at which the equilibrium line is noted at the end of any particular balance year. Normally, it is an averaged value with respect to the whole glacier. ELA is used as an indicator of the glacier mass balance state; when it is higher, the net balance is lower and vice versa.

Firn: Old, coarse-grained snow that has survived at least one summer melt season.

Glaciation threshold: The critical level where a glacier can form and is normally calculated by means of the 'summit method' (between the lowest mountain carrying a glacier and the highest mountain without a glacier).

Internal accumulation: The water melted out at times of ablation usually drains from the glacier and its mass is thereby reduced. However, in areas with snow or firn temperatures below zero, meltwater percolating through the summer surface can refreeze and thereby add mass to the lower layers of snow or firn.

Mass balance: The change in mass at any point on a glacier surface at any time (positive or negative mass balance). Normally, it means a change in the mass of the entire glacier in a standard unit of time (the balance year or measurement year).

Mass balance amplitude: The winter balance minus the summer balance (i.e., negative) divided by two.

Mass balance gradients: The rates at which annual ablation and accumulation change with altitude are termed ablation gradient and accumulation gradient, respectively. Together, they define the mass balance gradient.

Mass balance ratio: The ratio between the ablation gradient and the accumulation gradient is termed the balance ratio.

Measurement year: The unit of time used in the fixed date system of mass balance study, which is usually taken at the end of the summer or the beginning of winter and lasts 365 days.

Net balance (b_n): The sum of the winter balance (positive) and the summer balance (negative) through the balance year ($b_n = b_w + b_s$). The term is used in the stratigraphic system of measurement and reporting of mass balance (see Figure 1). Total values are averaged over the entire glacier area and presented in terms of an equivalent of water layer (thickness in meters).

Steady-state equilibrium line: The ELA when the net balance (b_n) is zero.

Summer balance (b_s): The change in mass (commonly negative) during the summer season. It is usually measured

at the end of the summer season and time of formation of the summer surface (minimum balance). The term is often used synonymously with summer ablation.

Summer surface: The glacier surface is formed as a result of the summer balance. This represents the surface of the minimum glacier volume during the balance year.

Temporary equilibrium-line altitude: The equilibrium line at an arbitrarily chosen time of the year. During early spring

the temporary ELA is in the lower glacier area, whereas it is higher up the glacier later in the ablation season.

Winter balance (b_w): The maximum balance value (positive) during the balance year in the stratigraphic system (considered synonymous with winter accumulation).

The time when the maximum balance is measured divides the balance year into the winter and summer seasons.

The rate of growth and retreat of a glacier is mainly controlled by the mass balance, which is the difference between the accumulation and ablation rates. This is termed the net mass balance and commonly measured over a 1-year period. Mass balance is therefore the net gain or loss of ice and snow. Mass balance is highly dependent on the climate conditions. The accumulation includes all materials that add mass to the glacier, such as snow, refrozen slush, hail, rain, and avalanched snow and ice. The ablation includes direct ice melt, iceberg calving, wind erosion/deflation, and sublimation. On most glaciers, accumulation is dominant in the winter months, whereas ablation occurs mainly during the summer. Where an ice sheet or glacier terminates in water, however, iceberg calving may occur throughout the year. Commonly, accumulation exceeds ablation on the upper part of a glacier, whereas ablation is larger than the accumulation in the lower parts. For a 'regular' glacier that is fed by snowfall in its accumulation

area and that loses mass by melting of 'clean' ice in its ablation area, altitudinal mass balance gradients (Figure 1) are approximately linear, with ablation gradients tending to be steeper than accumulation gradients (Furbish and Andrews, 1984; Schytt, 1967). The linear mass balance gradient approach, however, apparently works well in many cases (Kaser, 1995; Kuhn, 1984; Mayo, 1984). In practice, mass balance gradients are influenced by local factors, such as irregularities on the glacier surface (heights and depressions), shading by surrounding topography, and areas of supraglacial debris.

The equilibrium-line altitude (ELA) on modern glaciers is the average elevation where accumulation equals ablation over a 1-year period (Paterson, 1994; Figure 2). Rarely, however, glaciers experience zero net mass balance at the same elevation across the entire width of the glacier due to local topographic and climatic variations in accumulation and ablation. Thus, the ELA is the average altitude of the equilibrium line. The

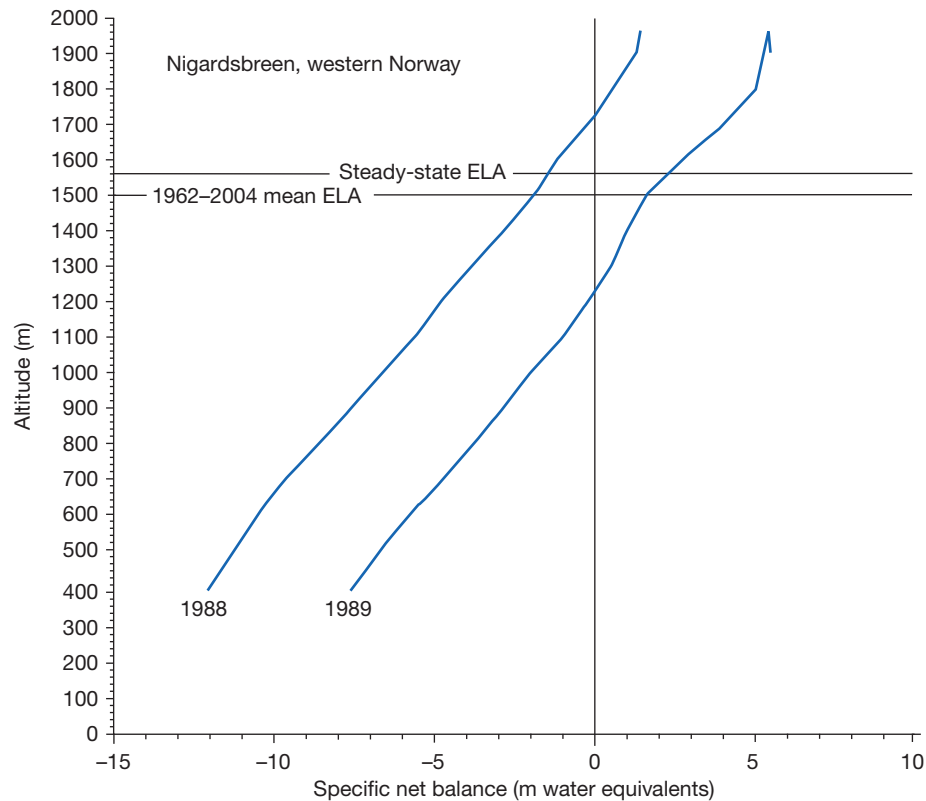


Figure 1 Net balance gradients of Nigardsbreen, an outlet glacier from Jostedalbreen in western Norway, in two balance years with contrasting net mass balance situations (1987–88 and 1988–89). The steady-state ELA and the 1962–2004 ELA are indicated. Data provided by NVE (Oslo, Norway).

'steady-state ELA' is defined as the ELA when the annual net balance is zero, and it can be calculated by linear regression analysis of annual net balance data and corresponding ELAs over some years (Figure 3). When the annual ELA coincides with the steady-state ELA, which rarely occurs, the mass of ice is in equilibrium with the climate. The steady-state ELA is therefore merely a theoretical term. The concept, however, is very useful because it provides a measure of the climatic means related to certain glacier positions and geometries. The 'climatic ELA' is the average ELA over a 30-year period (corresponding to a climate 'normal' – 30-year mean – period).

Climate processes influencing the ELA on glaciers involve ablation (mainly controlled by summer temperature, T) and

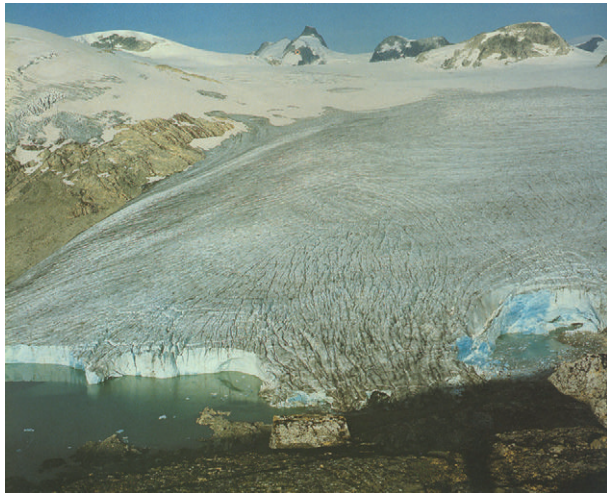


Figure 2 The ELA on the glacier Austdalsbreen, an eastern outlet glacier from Jostedalbreen in western Norway. Photograph by A. Nesje.

direct snowfall (precipitation, P), giving the TP-ELA. In addition, wind (W) transport of dry snow is an important factor for the glacier mass balance, particularly during the accumulation season. On plateau glaciers, snow deflation and drifting dominate on the windward side, whereas snow accumulates on the leeward side. By calculating the mean ELA in all glacier quadrants on a plateau glacier, the influence of wind on plateau glaciers may be negligible. The resulting ELA is therefore defined as the temperature–precipitation ELA (TP-ELA). The TP-ELA reflects the combined influence of the accumulation season precipitation and ablation season temperature (Dahl and Nesje, 1992). On wind-exposed mountain plateaus, the snow may deflate and accumulate in lower cirques and valleys, either by direct accumulation on the cirque/valley glaciers or by avalanching from the mountain slopes. The accumulation on the cirque/valley glaciers may therefore be significantly higher than the local precipitation indicates (Dahl and Nesje, 1992). The mean ELA on a plateau glacier therefore defines the TP-ELA, whereas the ELA on a cirque glacier, commonly influenced by wind-transported snow, gives the TPW-ELA. Therefore, the TPW-ELA is commonly lower than the TP-ELA (Figure 4). Leeward accumulation of windblown snow, including avalanching from the cirque walls, is an important factor influencing the mass balance of cirque glaciers. Due to the potential upslope transport of dry snow by wind, the additional accumulation of snow in a cirque is commonly difficult to estimate. Dahl et al. (1997) therefore introduced a ratio between the catchment area above the TPW-ELA (D) and the reconstructed glacier accumulation area (A) as a rough measure of the potential for additional snow by wind transport and avalanching in cirques. A low D/A ratio indicates a TPW-ELA close to the TP-ELA with only minor additional supply of windblown snow that is typical for cirque glaciers surrounded by steep alpine peaks. A high D/A ratio, on the other hand,

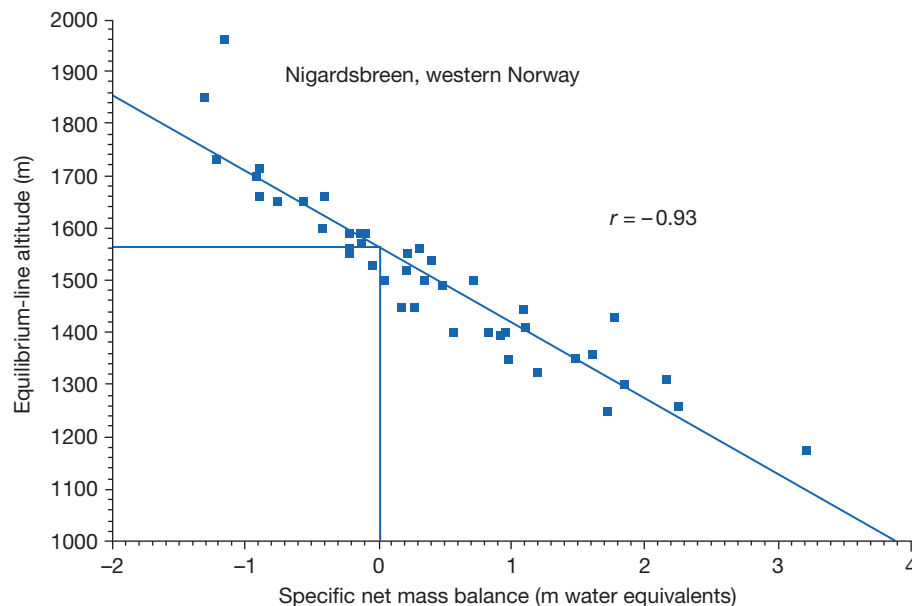


Figure 3 The steady-state ELA (=1560 m; ELA where the net balance is zero) at Nigardsbreen, an eastern outlet glacier from Jostedalbreen, western Norway. Based on data (observation period 1962–2004) provided by the Norwegian Water Resources and Electricity Directorate.

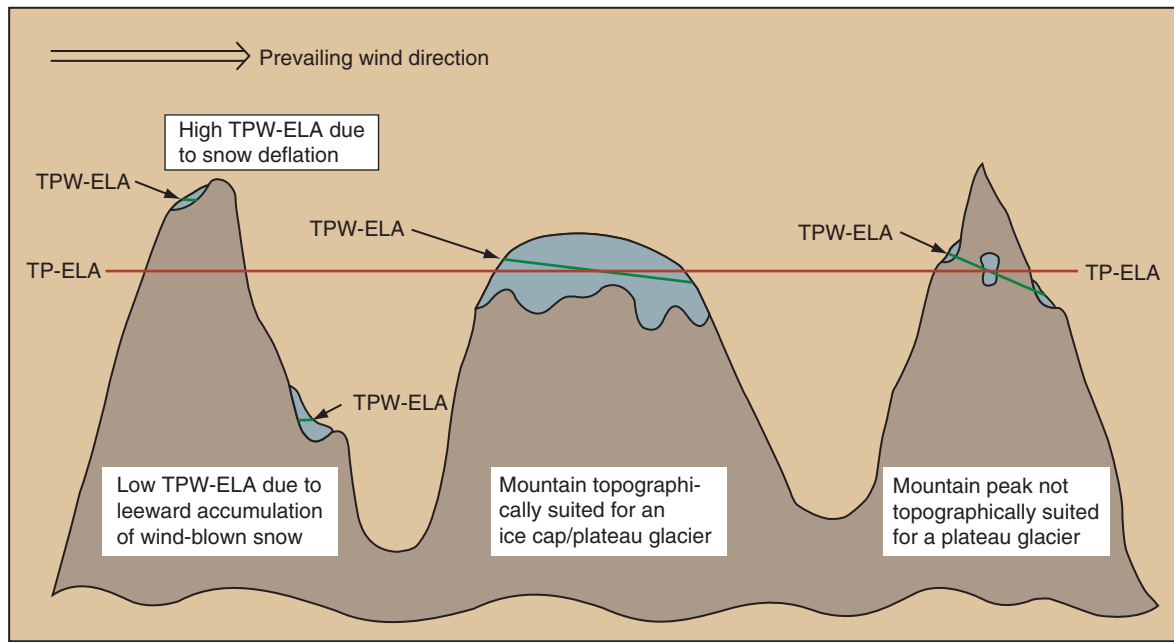


Figure 4 Schematic examples showing the difference between the temperature–precipitation equilibrium-line altitude (TP-ELA) at plateau glaciers and the temperature–precipitation–wind equilibrium-line altitude (TPW-ELA) at cirque glaciers. Modified from Dahl SO and Nesje A (1992) Palaeoclimatic implications based on equilibrium-line altitude depressions of reconstructed Younger Dryas and Holocene cirque glaciers in inner Nordfjord, western Norway. *Palaeogeography, Palaeoclimatology, Palaeoecology* 94: 87–97, with permission.

shows the potential for large amounts of windblown snow onto a cirque glacier and a TPW-ELA well below the TP-ELA. This is typical for a cirque glacier surrounded by a large mountain plateau.

Three equations derived from a close exponential glacier–climate relationship at the ELA of Norwegian glaciers were implemented in a geographical information system (Lie et al., 2003). The equations enable calculation of (1) the minimum altitude of areas climatically suited for present glacier formation, (2) quantification of the glacial build-up sensitivity (GBS) in an area, and (3) calculation of the theoretical climatic TP-ELA (based on temperature and precipitation data from meteorological stations in southern Norway) in presently nonglaciated areas by combining GBS with terrain altitude. The approach is primarily intended for reconstruction of paleo-ELAs in areas with no present glaciers.

There is a very close connection between the ELA and local climate, particularly winter accumulation (snowfall), summer temperature, and wind strength causing snow transport during the accumulation season. The ELA is therefore sensitive to perturbations of these variables; high winter snowfall combined with low summer temperatures the following summer gives a low ELA, whereas low winter accumulation followed by a warm summer yields a high ELA. Fluctuations in the ELA over time (Figure 5) are therefore an important indicator of climate change and commonly allow reconstructions of past climates. Thus, paleoclimatic reconstructions based on former glacier extent commonly make use of reconstructed ELAs. Data on past variations of ELAs therefore provide an important source of proxy paleoclimatic data in glaciated regions throughout the world.

Relationships Between Precipitation–Temperature and the ELA

Relationships between temperature and precipitation at ELAs on glaciers have been established by statistical/analytical approaches (Liestøl in Dahl et al., 1997; Nesje and Dahl, 2000; Ohmura et al., 1992; Shi et al., 1992; Sissons, 1979; see Figure 5). The positive correlation between these two variables for a wide range of glaciers reflects the fact that higher values of mass turnover at the ELA require higher ablation (higher summer temperatures) to balance the annual specific mass budget. The curves in Figure 4 yield fairly similar values and a curve shape in the temperature range of -2 to 4°C and accumulation–precipitation up to 3500 mm. The approaches of Ohmura et al. (1992) and Nesje and Dahl (2000), however, deviate significantly for temperature and precipitation $>4^{\circ}\text{C}$ and >3500 mm, respectively. Some of the discrepancy may be explained by the fact that Ohmura et al. use winter accumulation plus summer precipitation at the ELAs, whereas Nesje and Dahl use only winter (1 October–30 April) precipitation. (For details on the different approaches, see Figure 5.)

The temperature–precipitation relationship also explains the regional rise in glacier ELAs with increasing distance from moisture sources. In areas of lower precipitation, the temperature required to melt the annual accumulation at the ELA does not need to be as high as in areas of high precipitation. Consequently, ELAs tend to be at higher elevations in areas with low ablation season temperatures than in areas with higher ablation season temperatures. If paleo-ELAs have been reconstructed and if paleotemperatures are reconstructed from independent proxies (pollen, faunal remains, etc.) nearby, former

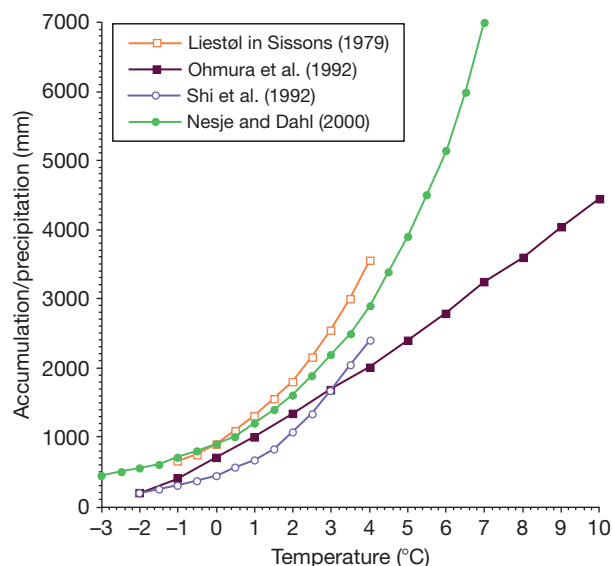


Figure 5 Temperature–precipitation relationships at glacier ELAs. Liestøl in Sissons (1979): mean summer (1 May–30 September) temperatures and winter (1 October–30 April) accumulation at the ELA of ten Norwegian glaciers (Ålfotbreen, Engabreen, Folgefonna, Nigardsbreen, Tunsbergdalsbreen, Hardangerjøkulen, Storbreen, Austre Memurubreen, Hellstugubreen, and Gråsubreen). Ohmura et al. (1992): mean summer temperature in free atmosphere at ELA and annual total (winter accumulation plus summer precipitation) at ELA for 70 mid- and high-latitude glaciers. Shi et al. (1992): mean summer temperature and annual precipitation at the ELA of 17 high Asian glaciers. Nesje and Dahl (2000): winter (1 October–30 April) precipitation and summer (1 May–30 September) temperature at the ELA in corresponding balance years calculated from adjacent meteorological stations. The ELAs at the four glaciers used were Ålfotbreen, Hardangerjøkulen, Hellstugubreen (all three in southern Norway), and Brøggerbreen (Svalbard), together with summer temperature data from adjacent meteorological stations Sandane, Finse, Øvre Tessa, and Isfjord Radio, respectively.

winter or total precipitation at the ELA can be calculated from temperature–precipitation relationships (Dahl and Nesje, 1996). Regional variations of glacier ELAs can be used to infer former precipitation gradients that may allow prevailing moisture sources and circulation patterns to be reconstructed (Lehmkuhl, 1998; Lehmkuhl et al., 1998; Miller et al., 1975). In addition, reconstructed ELAs of former glaciers have been used to correlate moraines in mountain regions (Maisch, 1982). However, the term ELA has been applied to several altitudinal indices calculated for past glaciers. The relationship between some of these indices and climate, however, has not been firmly established. Paleoclimatic reconstructions based on these indices may therefore be inaccurate (Benn and Lehmkuhl, 2000). In some high mountain regions, the relationship between climate and glacier ELAs may be complicated by avalanches and debris cover on the glacier surface.

Approaches to Calculating Paleo-ELAs

The determination of steady-state ELAs on modern glaciers should be based on measurements of glacier mass balance

over several consecutive years (Figure 6). If we want to calculate the ELAs when a glacier was larger than at present or if the glacier has melted completely, other methods have to be used. Several approaches for calculating/estimating the ELAs of former glaciers from geomorphological evidence have been developed (Gross et al., 1976; Meierding, 1982; Torsnes et al., 1993; see review by Benn and Lehmkuhl, 2000). The most common approaches for reconstructing paleo-ELAs are the use of the following:

- The maximum elevation of lateral moraines (MELM)
- The median elevation of glaciers (MEG)
- The toe-to-headwall altitude ratio (THAR)
- The toe-to-summit altitude method (TSAM)
- The accumulation area ratio (AAR) to the total glacier area
- The balance ratio (BR) method
- Cirque floor altitudes
- Proglacial sites

The concepts of the different methods vary significantly, as do their reliability and ease of use. Some approaches are based on detailed evaluation of mass balance and glacier hypsometry (distribution of glacier area over its altitudinal range). Others are based on the large-scale morphology of the glacier catchment, the altitude of marginal moraines, and the cirque floor. A literature survey indicates that ELA depressions from modern values during the Last Glacial Maximum, Younger Dryas, and the ‘Little Ice Age’ were typically in the range of 1000 ± 300 , 500 ± 200 , and 100 ± 50 m, respectively.

Maximum Elevation of Lateral Moraines

Due to the nature of glacier flow toward the center above the ELA and toward the margin of the glacier below the ELA, lateral moraines are theoretically deposited only in the ablation zone below the ELA. As a result, the MELM should ideally reflect the corresponding ELA (Figure 7). This method is considered best suited for long valley glaciers with large, continuous lateral moraines. However, there are several problems with this approach. First, englacial material may not reach the glacier surface immediately below the equilibrium line. Thus, lateral moraines may not form at the actual ELA, but below it. Second, it is difficult to assess whether or not a lateral moraine is preserved entirely in the upper part due to degradation on steep slopes. Consequently, ELA estimates derived from eroded/degraded lateral moraines may be too low. Third, the valley sides may be too steep for deposition of lateral moraines, giving estimates of the former ELA that are too low. Finally, if initial glacier retreat is slow, additional moraine material could be deposited in the prolongation of the former steady-state lateral moraine. A continuous supply of debris from the valley or cirque walls during glacier retreat may lead to estimates of the former ELA that are too high. Despite its problems, the MELM approach is considered the most reliable method in cases in which former glaciers were covered by debris and the mass balance gradient is poorly or not known. Special consideration must be paid to the topographic setting and the preservation of moraines. The MELM approach does not need detailed topographic maps for the entire glacier area. The altitude of the moraines can easily be determined photogrammetrically or by an altimeter in the field.

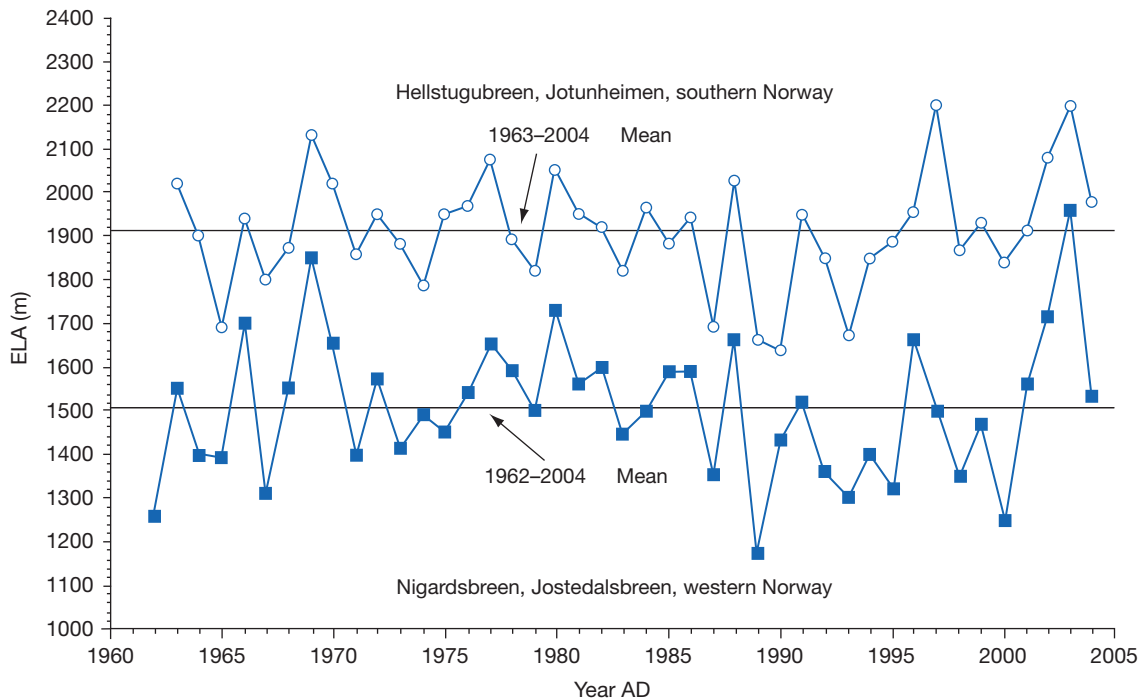


Figure 6 Observed ELAs on Hellstugubreen (Jotunheimen in central southern Norway) and Nigardsbreen (Jostedalbreen in western Norway) from the early 1960s to 2004. Data from NVE (Oslo, Norway).

Median Elevation of Glaciers

This approach places the ELA halfway between the glacier front and the top of the glacier. Empirical evidence from modern glaciers, however, suggests that the MEG overestimates the ELA. In addition, this method fails to take into account variations in valley morphology, which strongly affects the area/elevation distribution of a glacier. However, it works reasonably well for small glaciers with an even area/altitude distribution. The main problem is commonly to define the headward limit of a former glacier.

Toe-to-Headwall Altitude Ratio

This method assumes that the ELA lies at a fixed proportion of the vertical distance between the highest and lowest part of a former glacier. A ratio between the maximum and minimum altitude of a glacier (THAR) has been used as a quick estimate to calculate the ELA. Ratios of 0.35–0.4 were used by Meierding (1982) for glaciers in Colorado. Values of 0.4–0.5 were used by Porter et al. (1983) for ‘clean’ ice glaciers. Debris-covered glaciers in California, however, have rather high THARs of 0.6–0.8 (Clark et al., 1994). A THAR of 0.60 ± 0.12 has been obtained from 15 Norwegian glaciers with modern mass balance measurements and ELA observations (A. Nesje, unpublished data). The THAR approach is considered a rough way of estimating former ELAs because it does not take into account glacier hypsometry and mass balance. In addition, a major problem is to define the headward limit of a former glacier. The THAR approach allows rapid estimates of former ELAs in remote areas with poor or inaccurate maps.

Toe-to-Summit Altitude Method

A way to overcome the problem of defining the upper limit of a former glacier is to use the maximum altitude of the glacier catchment. Louis (1955) suggested that the ELA may be calculated from the arithmetical mean of the altitude of the highest mountain and the terminal moraine. For the European Alps, Gross et al. (1976) demonstrated that this approach yields ELAs that are approximately 100 m too high. For modern glaciers in the northernmost mountain range of the Mongolian Altai, the inferred values are in agreement with snowlines on glaciers observed in the field and on aerial photographs (Lehmkuhl, 1998). The TSAM approach may cause inaccurate ELA estimates where the highest mountain peak is not representative of the catchment as a whole and does not contribute significantly to glacier accumulation. Despite its crudeness, the TSAM method gives rapid estimates of former ELAs in remote areas. ELA estimates based on the TSAM method should be determined for each region separately because there may be significant variations due to variations in glacier type.

AAR to the Total Glacier Area

The ratio of the accumulation area to the total area (AAR) is based on the assumption that the steady-state AAR of former glaciers is typically 0.55–0.65 (Porter, 1975). An AAR of 0.6 ± 0.05 is generally considered to characterize steady-state conditions of valley/cirque glaciers. Ice caps and piedmont glaciers, however, may differ significantly from this ratio. Gross et al. (1976) inferred a mean AAR value of 0.67 for glaciers in the European Alps. The AAR of a glacier varies mainly as a function of its mass balance; ratios less than 0.5 indicate

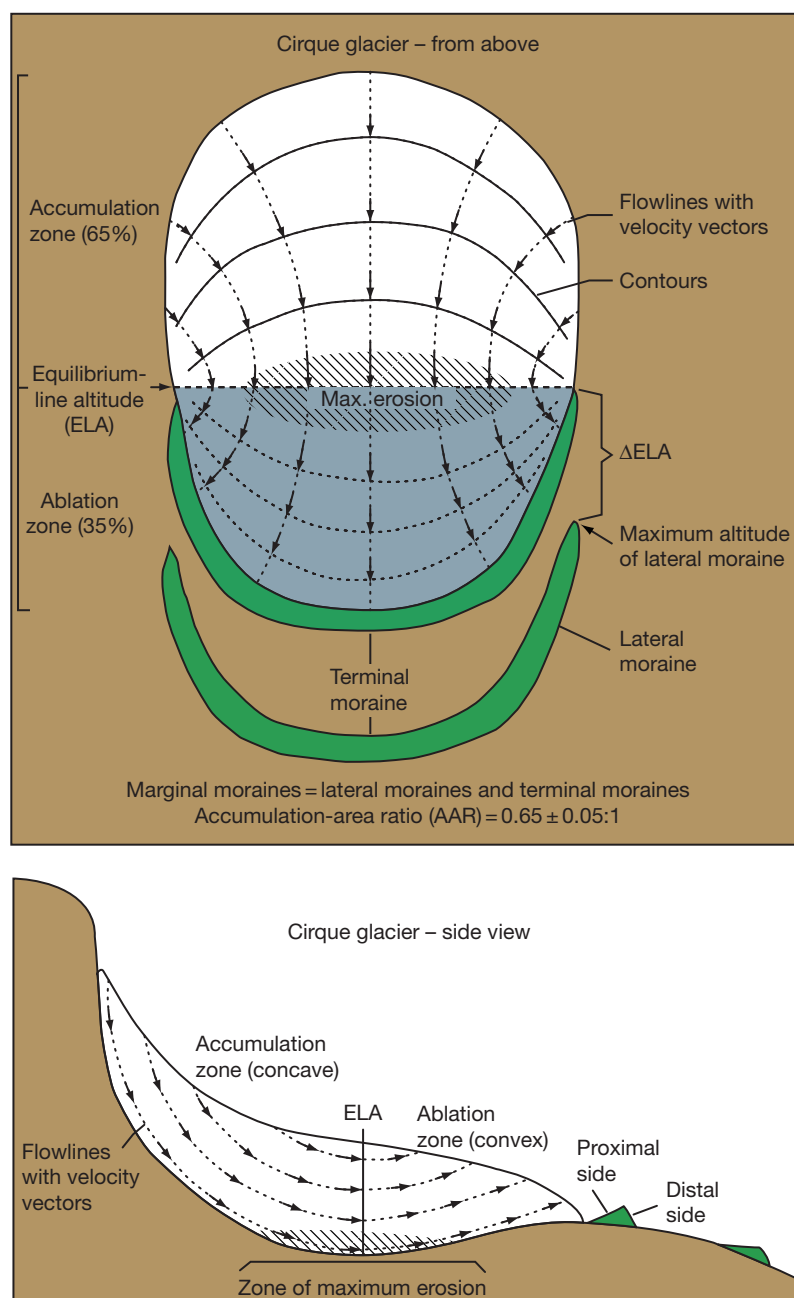


Figure 7 The principle of calculating the depression of the ELA on a cirque glacier based on the maximum elevation of lateral moraines. Dashed lines on the top panel indicate surface contours, and arrows on both panels indicate ice flow direction.

negative mass balance, ratios of 0.5–0.8 correspond to steady-state conditions, and values higher than 0.8 reflect positive mass balance regimes. For modern mid- and high-latitude glaciers, steady-state AARs typically lie in the range 0.5–0.8.

The AAR approach requires contour maps of former glacier surfaces, and therefore, this method can be used only where detailed topographic data/maps are available. Information about mass balance gradients is not required for the AAR method. AARs are commonly determined empirically and should be based on observations of modern glaciers in the region of interest (Gross et al., 1976). Debris-covered glacier tongues have lower AARs than clean glaciers. Thick debris

(>2 cm) lowers the ablation and increases the size of the ablation area required to balance accumulation (Clark et al., 1994). In high mountain regions, AARs commonly vary according to the extent and distribution of debris cover in the ablation area. There is a need to determine the range of AARs associated with different glacier geometries and mass balance characteristics (Benn and Lehmkuhl, 2000).

The largest source of inaccuracy related to the AAR method of determining the ELA on former glaciers is the reconstruction of the surface contours, especially if the glacier margins intersect valley-side topographic contours at small angles or coincide with them for some distance. However, this source of error

is considered to be randomly distributed and is not considered to introduce major deviations from representative conditions. In addition, this method only requires glacier reconstruction as high as the former ELA. A theoretical evaluation of the AAR approach, using changing slope angles and valley morphology on idealized glaciers, shows that glaciers advancing into flat areas underestimate the ELA depression, whereas glaciers moving into areas of increasing slope angle overestimate the climatic ELA difference. Consequently, topographical and morphological effects on calculated ELA depressions on glaciers must be carefully evaluated.

BR Method

The BR method is a refinement of the AAR approach. As discussed previously, one shortcoming of the AAR method and also the MEG approach is that they do not fully account for variations in the hypsometry. To overcome this problem, a BR method was developed by [Furbish and Andrews \(1984\)](#). This approach takes account of both glacier hypsometry and the shape of the mass balance curve and is based on the fact that for glaciers in equilibrium, the total annual accumulation above the ELA must balance the total annual ablation below the ELA. This can be expressed as the areas above and below the ELA multiplied by the average accumulation and ablation, respectively. The BR approach assumes that the accumulation and ablation gradients are approximately linear, that the ratio between the two is known, and that the hypsometry of the glacier is known. The BR method can thus only be used where detailed topographic maps are available and the surface contours of former glaciers can be reconstructed in detail ([Benn and Gemmel, 1997](#)). Because accumulation and ablation gradients are controlled by different climatic variables, the accumulation and ablation gradients generally have different values, with the ablation gradient somewhat steeper than the accumulation gradient. In a study of 22 glaciers in Alaska, [Furbish and Andrews \(1984\)](#) found that BRs clustered around 1.8, meaning that the ablation gradient is 1.8 times greater than the accumulation gradient. BRs are different under different climate conditions; they are lowest in polar areas and highest in tropical regions.

The BR approach is best suited to snow-fed, clean glaciers in areas with detailed maps. However, the method is not appropriate where a significant part of the accumulation is by avalanching from surrounding mountains and/or where debris cover strongly influences the ablation gradient. [Benn and Gemmel \(1997\)](#) published a spreadsheet program that rapidly calculates glacier ELAs using the BR method, which allows hypsometric variations to be accounted for in ELA reconstructions.

Cirque Floor Altitudes

Cirque floor levels or altitudes have been used as indicators of the elevation of former glacier snowlines for Pleistocene glaciers. However, the elevation of the cirque floor/basin does not have a close relationship with the ELA of the glacier that formed it ([Benn and Lehmkuhl, 2000](#)). Cirques develop cumulatively over many glacial cycles and therefore cannot be attributed to any particular glacial event. However, they may give some insight into regional trends in glaciation levels and average climatic conditions over longer timescales.

Proglacial Sites

Several approaches have been applied to record past variations in ELAs based on lacustrine and terrestrial proglacial sites. [Dahl et al. \(2003\)](#) presented a conceptual model of glacier meltwater-induced sedimentation in which a variety of proxies are related to the occurrence and size variations (as a response to variations in the ELA) of a glacier in a catchment.

Validity of Different Approaches for Paleo-ELA Reconstruction

The mass balance characteristics of glaciers in high mountain environments complicate the relationship between glacier ELAs and climatic variables such as precipitation and air temperature. Therefore, methods of ELA reconstruction employed in low-relief environments are commonly not applicable in high mountains or require some modification ([Benn and Lehmkuhl, 2000](#)). [Leonard and Fountain \(2003\)](#) examined the validity of two methods for estimating glacier ELAs from topographic maps. The ELA determined by the inflection of the contour lines on the glacier (kinematic ELA) and the mean elevation of the glacier correlates well with the ELA determined from mass balance data (measured). The data were normalized and a fairly high correlation ($r^2 = 0.59$) was obtained between observed and kinematic ELAs. The mean of the normalized kinematic ELAs was consistently located down-glacier from the observed ELA, in agreement with theory. The normalized mean elevation of the glacier yielded no correlation and suggests that the THAR approach is not a good approximation for the ELA. Kinematic waves had apparently no effect on the position of the kinematic ELA. Topographic maps of glacier surfaces can therefore be used to infer the position of the ELA and provide a method for estimating paleo-ELAs from historic topographic maps and reconstructed paleoglaciers.

Small glaciers have short response times to climate change and they are therefore well suited for climate monitoring. Glacier responses to climate, or their mass change, may be suggested by a change in the ELA. However, several regional climatic reconstructions have neglected the importance of local variations in ELAs in preference for regional trends ([Carrivick and Brewer, 2004](#)). Ignoring the importance of local topographic components in mass balance estimates for small glaciers close to the glaciation threshold/level may lead to erroneous climatic reconstructions. Of 510 small valley and cirque glaciers digitized across northern Scandinavia, 284 were considered suitable for inferring an ELA ([Carrivick and Brewer, 2004](#)). The inferred ELA was derived from the median elevation and several local topographic variables using regression analysis. The glacier elevation, area, slope, and aspect parameters were found to be the best predictors of the local ELA. ELA estimations improved from 77% up to 94% accuracy when topographic parameters for every grid cell within rasters representing glacier surfaces were computed rather than using subjective measurements from topographic maps. Regional ELA trend surfaces, interpolated between the local ELA values, differed in representing the local variability, depending on the distribution and accuracy of the local ELA values. A second-order polynomial trend surface represented ELA variations most accurately across the study area, within the local

measurement accuracy of ± 100 m. Carrivick and Brewer (2004) concluded that current subjective topographic map-based analyses are unlikely to be sufficiently accurate for predicting the regional ELA of small, sensitive, and marginal glaciers, unless GIS-based spatial analyses are made at a reasonable resolution.

Conclusion

The ELA, which marks the area or zone on a glacier where accumulation is balanced by ablation over a 1-year period, is sensitive to variations in winter precipitation, summer temperature, and wind transport of dry snow. When the annual net mass balance is negative, the ELA rises, and when the annual mass balance is positive, the ELA declines.

Fluctuations in the ELA are important indicators of climate change. This may allow reconstructions of accumulation season precipitation, ablation season temperature, and prevailing snow-bearing wind directions. Paleoclimatic reconstructions based on former glacier extent commonly include estimates of ELAs and depression of ELAs from present values. Several methods have been developed to estimate steady-state ELAs of former glaciers in high mountainous regions. The BR method is the only approach that is directly linked to glacier mass balance. The AAR method is related to generalized mass balance relationships. The MELM builds on the relationship between glacier mass balance and the distribution of glaciogenic deposition, whereas the THAR, TSAM, and cirque floor altitudes are related to the large-scale morphology of the glaciated catchments and therefore not directly linked to glacier mass balance. The BR approach is most useful on snowfall-fed, clean glaciers where mass balance gradients were approximately linear. The AAR method can be applied to clean, snowfall-fed glaciers when appropriate values of the steady-state AAR are taken into account. The MELM approach is a quick method of estimating former ELAs in remote, high mountain regions. This method does not require detailed topographic maps. The THAR and cirque floor altitudes are considered as crude estimates of paleo-ELAs. The TSAM method may provide a useful and rapid method in areas where mass balance data are not available. Researchers who study former glaciers should provide detailed descriptions of their information on the method (s) (multiple methods should be used) used to calculate former ELAs. A literature survey indicates that ELA depressions (from modern elevations) during the Late Glacial maximum, Younger Dryas, and the Little Ice Age were typically in the range of 1000 ± 300 , 500 ± 200 , and 100 ± 50 m, respectively.

See also: [Paleoclimate Reconstruction: Younger Dryas Oscillation, Global Evidence.](#)

References

- Benn DI and Gemmel AMD (1997) Calculating equilibrium-line altitudes of former glaciers by the balance ratio method: A new computer spreadsheet. *Glacial Geology and Geomorphology*.
- Benn DI and Lehmkuhl F (2000) Mass balance and equilibrium-line altitudes of glaciers in high-mountain environments. *Quaternary International* 65/66: 15–29.
- Carrivick JL and Brewer TR (2004) Improving local estimation and regional trends of glacier equilibrium line altitudes. *Geografiska Annaler* 86A: 67–79.
- Clark DH, Clark MM, and Gillespie AR (1994) Debris-covered glaciers in the Sierra Nevada, California, and their implication for snowline reconstructions. *Quaternary Research* 41: 139–153.
- Dahl SO, Bakke J, Lie Ø, and Nesje A (2003) Reconstruction of former glacier equilibrium-line altitudes based on proglacial sites: An evaluation of approaches and selection of sites. *Quaternary Science Reviews* 22: 275–287.
- Dahl SO and Nesje A (1992) Palaeoclimatic implications based on equilibrium-line altitude depressions of reconstructed Younger Dryas and Holocene cirque glaciers in inner Nordfjord, western Norway. *Palaeogeography, Palaeoclimatology, Palaeoecology* 94: 87–97.
- Dahl SO and Nesje A (1996) A new approach to calculating Holocene winter precipitation by combining glacier equilibrium line altitudes and pine-tree limits: A case study from Hardangerjøkulen, central southern Norway. *The Holocene* 6: 381–398.
- Dahl SO, Nesje A, and Øvstedal J (1997) Cirque glaciers as morphological evidence for a thin Younger Dryas ice sheet in east-central southern Norway. *Boreas* 26: 161–180.
- Furbish DJ and Andrews JT (1984) The use of hypsometry to indicate long-term stability and response to valley glaciers to changes in long-term stability and response of valley glaciers to changes in mass transfer. *Journal of Glaciology* 30: 199–211.
- Gross G, Kerschner H, and Patzelt G (1976) Metodische Untersuchungen über die Schneegrenze in alpinen Gletschergebieten. *Zeitschrift für Gletscherkunde und Glazialgeologie* 12: 223–251.
- Kaser G (1995) How do tropical glaciers behave? Some comparisons between tropical and mid-latitude glaciers. In: Ribstein P and Francou B (eds.) *Agua Glaciares y Cambios en los Andes Tropicales. La Paz: Conferencias y Posters*, June 13–16, pp. 207–218. Seminario Internacional.
- Kuhn M (1984) Mass budget imbalances as criterion for a climatic classification of glaciers. *Geografiska Annaler* 66A: 229–238.
- Lehmkuhl F (1998) Quaternary glaciations in central and western Mongolia. *Quaternary Proceedings* 6: 153–167.
- Lehmkuhl F, Owen LA, and Derbyshire E (1998) Late Quaternary glacial history of northeast Tibet. *Quaternary Proceedings* 6: 121–142.
- Leonard KC and Fountain AG (2003) Map-based methods for estimating glacier equilibrium-line altitudes. *Journal of Glaciology* 49: 329–336.
- Lie Ø, Dahl SO, and Nesje A (2003) Theoretical equilibrium-line altitudes and glacier buildup sensitivity in southern Norway based on meteorological data in a geographical information system. *The Holocene* 13: 373–380.
- Louis H (1955) Schneegrenze und Schneegrenzebestimmung. *Geographisches Taschenbuch* 1954/1955: 414–418.
- Maisch M (1982) Zur Gletscher – und Klimageschichte des alpinen Spätglazials. *Geographica Helvetica* 37: 93–104.
- Mayo LR (1984) Glacier mass balance and runoff research in the USA. *Geografiska Annaler* 66A: 215–227.
- Meierding TC (1982) Late Pleistocene glacial equilibrium-line in the Colorado front range. A comparison of methods. *Quaternary Research* 18: 289–310.
- Miller GH, Bradley RS, and Andrews JT (1975) The glaciation level and lowest equilibrium line altitude in the high Canadian Arctic: Maps and climatic interpretation. *Arctic and Alpine Research* 7: 155–168.
- Nesje A and Dahl SO (2000) *Glaciers and Environmental Change*. London: Arnold.
- Ohmura A, Kasser P, and Funk M (1992) Climate at the equilibrium line of glaciers. *Journal of Glaciology* 38: 397–411.
- Paterson WSB (1994) *The Physics of Glaciers*, 3rd edn. Oxford: Pergamon.
- Porter SC (1975) Equilibrium line altitudes of late Quaternary glaciers in the Southern Alps, New Zealand. *Quaternary Research* 5: 27–47.
- Porter SC, Pierce KL, and Hamilton TD (1983) Late Pleistocene glaciation in the western United States. In: Porter SC (ed.) *Late Quaternary Environments in the United States*, pp. 71–111. Minneapolis: University of Minnesota Press.
- Schytt V (1967) A study of 'ablation gradient'. *Geografiska Annaler* 49A: 327–332.
- Shi Y, Zheng B, and Li S (1992) Last glaciation and maximum glaciation in the Qinghai-Xizang (Tibet) Plateau: A controversy to M. Kuhle's ice sheet hypothesis. *Zeitschrift für Geomorphologie, B.F.* 84(supplement): 19–35.
- Sissons JB (1979) Palaeoclimatic inferences from former glaciers in Scotland and the Lake District. *Nature* 278: 518–521.
- Torsnes I, Rye N, and Nesje A (1993) Modern and Little Ice Age equilibrium-line altitudes on outlet valley glaciers from Jostedalsgreen, western Norway: An evaluation of different approaches to their calculation. *Arctic and Alpine Research* 25: 106–116.

Trimlines and Paleonunataks

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Introduction

Nunataks are mountain summits surrounded on all sides by glacier ice (Figure 1), and summits formerly surrounded by ice during Quaternary glacial maxima are referred to as paleonunataks. A periglacial trimline (sometimes termed a glacial trimline) represents the upper level to which glacial erosion has removed or 'trimmed' a pre-existing cover of regolith (weathered rock debris) on the flanks of nunataks or other mountain slopes. In many areas, periglacial trimlines have been preserved since late Pleistocene deglaciation. In some circumstances, such trimlines can be employed to delimit the configuration and maximum altitude reached locally by glacier ice during the Last Glacial Maximum (LGM) of ~26–~21 ka BP (Figures 2 and 3), or later readvances of more restricted extent. Not all periglacial trimlines, however, represent the former upper limit of glacier ice. In many areas, the contrast between glacially eroded terrain on low ground and apparently intact tors and regolith cover on plateaus reflects preservation of the latter under cold-based ice within a former ice sheet. Where this is the case, trimlines have important implications for the subglacial thermal regime of former ice cover, but provide only a minimum estimate of former ice-surface altitude. Periglacial trimlines are unrelated to vegetation trimlines. The latter are phytological boundaries marking the extent of recent (Little Ice Age) glacier advances or readvances, and their use in glacier reconstruction is limited to establishing the extent of ice advances within the past millennium.

Periglacial trimlines and their potential for reconstructing the dimensions of former ice sheets were recognized by James Geikie as early as 1873, and for over a century, geomorphological evidence for trimlines has been employed in mountain areas to estimate the maximum vertical dimensions and/or surface configuration of late Pleistocene ice sheets or Late Glacial readvances, particularly in Norway, Scotland, Labrador, and Arctic Canada (Ballantyne et al., 1998; Ives, 1978; Nesje et al., 1988). Over the past three decades, however, an accumulating body of evidence has demonstrated that many trimlines originally interpreted as being of ice-marginal origin actually reflect altitudinal zonation of erosion within former ice sheets that completely buried the summits of presumed paleonunataks, necessitating radical reinterpretation of ice thickness and ice-surface configuration at the LGM (e.g., André, 2004; Ballantyne, 2010a; Ballantyne and Hall, 2008; Briner et al., 2003; Fabel et al., 2002; Fjellanger et al., 2006; Kleman and Glasser, 2007; Marquette et al., 2004; Staiger et al., 2005). Appreciation that trimlines are interpretable in more than one way has led to the development of increasingly sophisticated approaches to date and validate trimlines, and to aid correct interpretation of trimline evidence.

Hypotheses of Trimline Formation

A periglacial trimline represents a weathering limit that separates an upper zone of more advanced rock breakdown from a lower zone characterized by glacially smoothed rock surfaces

that exhibit limited post-deglaciation rock weathering. In glaciated mountains, an upslope change (sharp or transitional) from glacially smoothed bedrock to terrain characterized by angular, often shattered bedrock outcrops and extensive regolith cover is common (Figures 2 and 3). The boundary between the two may potentially be interpreted in five ways (Figure 4).

1. The ice-marginal (nunatak) hypothesis: the weathering limit represents the upper limit of a warm-based (erosive) ice sheet at the LGM, with ice-free nunataks remaining above the ice (Figure 4(a)).
2. The ice-cap hypothesis: the weathering limit represents a thermal boundary between a warm-based (erosive) ice sheet at the LGM and thin, nonerosive cold-based ice caps occupying some or all of the higher ground (Figure 4(b)). This hypothesis represents a transitional case between (Figures 4(a) and 4(c)).
3. The englacial trimline hypothesis: the weathering limit represents a former englacial transition from erosive, warm-based ice occupying surrounding valleys to passive, cold-based ice on summits and plateaus (Figure 4(c)). This hypothesis implies that at the LGM, the highest summits were overridden by the regional ice sheet.
4. The readvance hypothesis: the weathering limit represents the upper limit of warm-based (erosive) ice during a post-LGM readvance (Figure 4(d)).
5. The climatic hypothesis: the weathering limit represents the lower boundary of widespread breakup of rock under periglacial conditions since deglaciation, and is thus unrelated to former ice-sheet or valley-glacier dimensions (Figure 4(e)).

Under hypotheses 1–4, the weathering limit represents the upper limit of glacial erosion and thus may be defined as a periglacial trimline. Cases 1 and 2 imply that the trimline marks the upper limit of regional ice cover at the LGM and thus may be employed to constrain the dimensions of LGM ice cover. Case 4 implies that the trimline represents the upper surface of a later readvance. Case 3 implies that the last and possibly earlier ice sheets were sufficiently thick to cover all high ground, but failed to erode older regolith covers on summits and plateaus. In case 5, the weathering limit is unrelated to glacial erosion and cannot be described as a trimline.

The age of blockfields and other regolith covers on high ground is critical for differentiating hypotheses 1–3 from 4 and 5. Most regolith covers on midlatitude or subarctic mountains are essentially relict features that predate the LGM. In some instances, these may even exhibit features possibly inherited from Neogene saprolite (e.g., André, 2004; Marquette et al., 2004; Paasche et al. 2006), though the characteristics of many blockfields appear consistent with development under Pleistocene periglacial conditions through exploitation of stress-release joints by frost wedging (Ballantyne, 1998, 2010b; Goodfellow et al., 2009). On some well-jointed or fissile lithologies, however, incipient blockfields and other forms of mountain-top regolith



Figure 1 Mount Rea, Marie Byrd Land, Antarctica, a present-day nunatak surrounded by outlet glaciers flowing toward the coast. From Sugden DE, Balco G, Cowdery SG, Stone JO, and Sass LC (2005) Selective glacial erosion and weathering zones in the coastal mountains of Marie Byrd Land, Antarctica. *Geomorphology* 67: 317–334. © Elsevier B.V. reproduced with permission.



Figure 2 Trimline representing the extent of glacier ice at the Last Glacial Maximum (LGM) (Otira Glaciation), Lake Wakatipu, South Island, New Zealand. Below the trimline, the bedrock is strongly ice-molded; above the trimline are outcrops of shattered bedrock and an extensive regolith cover. The trimline descends in altitude from right to left, away from the ice-shed location in the Southern Alps.

cover have developed locally since downwastage of the last ice sheet. In such cases, the lower limit of *in situ* weathered regolith bears no relation to former ice-sheet maxima, though it may represent a trimline cut by a post-LGM readvance. In the western Highlands of Scotland, for example, it is possible at some locations to identify two trimlines: an upper and older trimline representing the approximate maximum altitude of effective glacial erosion at the LGM and a lower and younger trimline that represents the upper limit of glaciation during the Younger Dryas Chron of ~12.9–11.7 ka BP. These two trimlines separate an uppermost (periglacial) weathering zone of advanced rock breakdown from an intermediate (Late Glacial) weathering zone of more limited bedrock disruption and a lowermost weathering zone of ice-molded bedrock exhibiting only superficial weathering (Ballantyne, 1998; Figure 5). The lower trimline has been employed by several authors to delimit the maximum ice-surface altitude and configuration during the Younger Dryas chronozone (e.g., Thorp, 1981, 1986). The clarity of the younger trimline, however, is strongly conditioned by lithology: well-jointed rocks (notably quartzite) have broken down to form incipient blockfields in the interval between ice-sheet downwastage and the Younger Dryas glacier readvance, whereas more massive lithologies such as gneiss have remained essentially intact since ice-sheet downwastage. Present-day blockfield development appears to be limited to permafrost environments that experience severe annual freezing (Ballantyne 2010b).



Figure 3 Granite paleonunatak in Yosemite National Park, California. A clear trimline separates a summit blockfield from ice-molded rock on lower slopes and marks the maximum altitude of glacier ice during the LGM (Tioga Glaciation).

Trimline Geomorphology

The identification of periglacial trimlines is based on the contrast between lower slopes exhibiting evidence of glacial erosion and deposition, and higher areas characterized by regolith cover and unmodified assemblages of periglacial features (Figures 2, 3, 6, and 7). Among the most useful forms of glacial

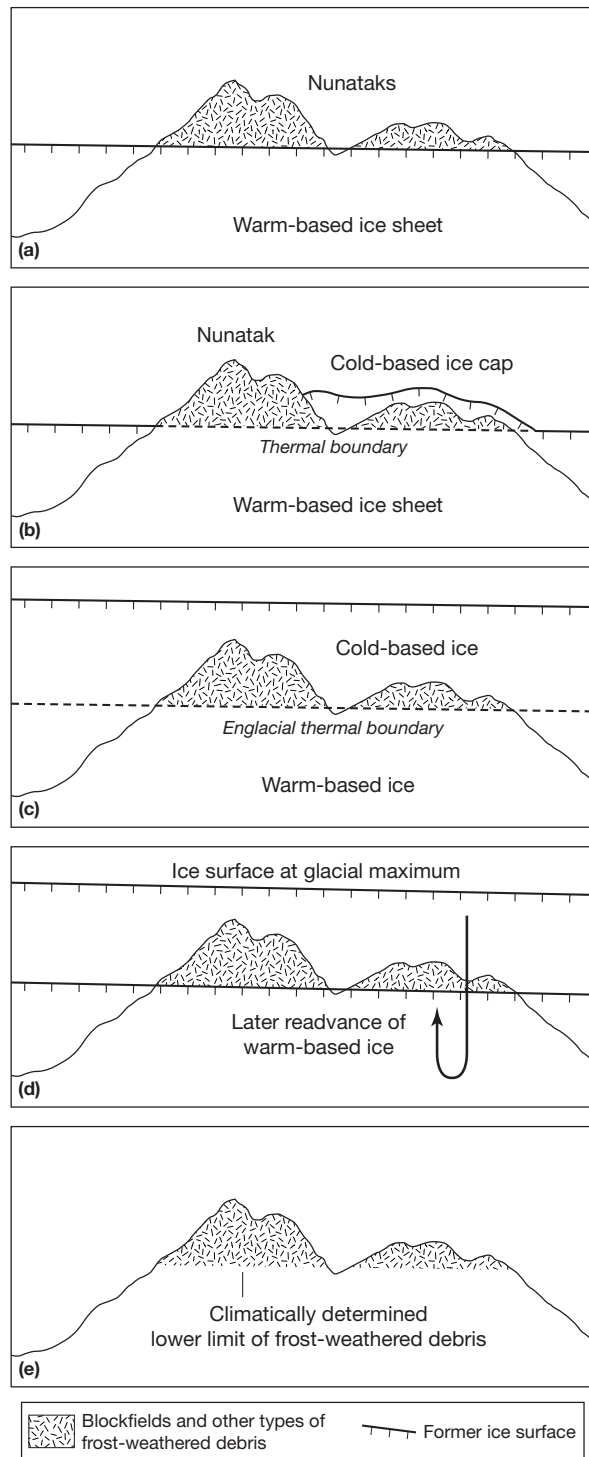


Figure 4 Hypothesis of trimline formation: (a) the ice-marginal (nunatak) hypothesis, (b) the ice-cap hypothesis, (c) the englacial trimline hypothesis, (d) the readvance hypothesis, and (e) the climatic hypothesis. For explanation, see text.

evidence are ice-molded or ice-abraded bedrock forms (roches moutonnées, whalebacks, glacially smoothed bedrock slabs, gouges, striae, and chattermarks) and erratics and other glacially transported perched boulders that rest on bedrock

slabs or on frost-weathered debris (Figure 8). Unmodified periglacial terrain on mountains is characterized by the absence of glacial evidence and by the presence of frost-shattered tors and bedrock outcrops, a thick cover of blockfield (felsenmeer), scree, or other weathered regolith, and a range of relict frost-action landforms that have developed on the regolith cover, such as solifluction sheets, solifluction lobes, and relict patterned ground. Because identification of trimlines is based on contrasts in terrain characteristics, they tend to be clearest in areas where the flow of glaciers or ice streams was topographically constrained and thus accelerated. Some of the clearest trimlines are therefore those cut across bedrock on truncated spurs. Conversely, trimlines are much less evident near former ice sheds (the highest points of former ice sheets) where ice sliding velocities were low. For similar reasons, clear trimlines tend to be better developed in maritime areas formerly occupied by fast-moving, erosive ice than in more arid continental areas where the mass budget of former ice masses was lower and ice movement slower.

Four problems affect the identification and mapping of periglacial trimlines. First, there has often been downslope movement of regolith since deglaciation, so that trimlines are obscured on moderate or steep slopes. This problem may be overcome by mapping the geomorphological evidence on low-gradient areas unaffected by mass movement (cols, shoulders, and summit plateaus) to constrain the former upper limit of glacial erosion and lower extent of unmodified periglacial terrain. Second, as noted previously, some lithologies have experienced disruption by frost action since glacial downwastage, and on such lithologies, incipient blockfields or other types of frost-weathered regolith may have developed below the upper limit of glacial erosion at the LGM (Ballantyne, 1998). In such cases, establishment of the former upper limit of glacial activity must rely on the distribution of erratics, perched boulders, or glacially detached blocks rather than glacial erosional landforms (Sellier, 1995). Third, some mountain landscapes take the form of gently rolling plateaus or plateau remnants that terminate abruptly at the edges of deep glacial troughs, notably in Norway, Sweden, Svalbard, Greenland, Scotland, Québec-Labrador, Arctic Canada, parts of Antarctica, and the Laramide Ranges of the western United States. In such areas, trimlines are usually absent because the troughs are flanked by steep cliffs, though the lowermost limit of blockfields may provide a constraint on the former altitudinal limit of glacier ice provided that there is no evidence for former ice cover on the plateau. In many areas, however, investigation of plateau surfaces has revealed evidence of former glacier ice cover in the form of occasional erratics (Figure 8), glacially transported perched boulders, glacially detached blocks, ablation moraines, or meltwater channels (e.g., Fjellanger et al., 2006; Landvik et al., 2003; Marquette et al., 2004; Sellier, 1995), even though evidence for glacial erosion of plateau surfaces may be absent, very localized, or limited to modification of tors (e.g., André, 2004; Hall and Glasser, 2003; Hall and Phillips, 2006). Such features provide compelling evidence for blockfield survival (though not necessarily at the LGM) under a cover of cold-based glacier ice (hypotheses 2 and 3). Finally, even where a clear weathering limit is identified, its

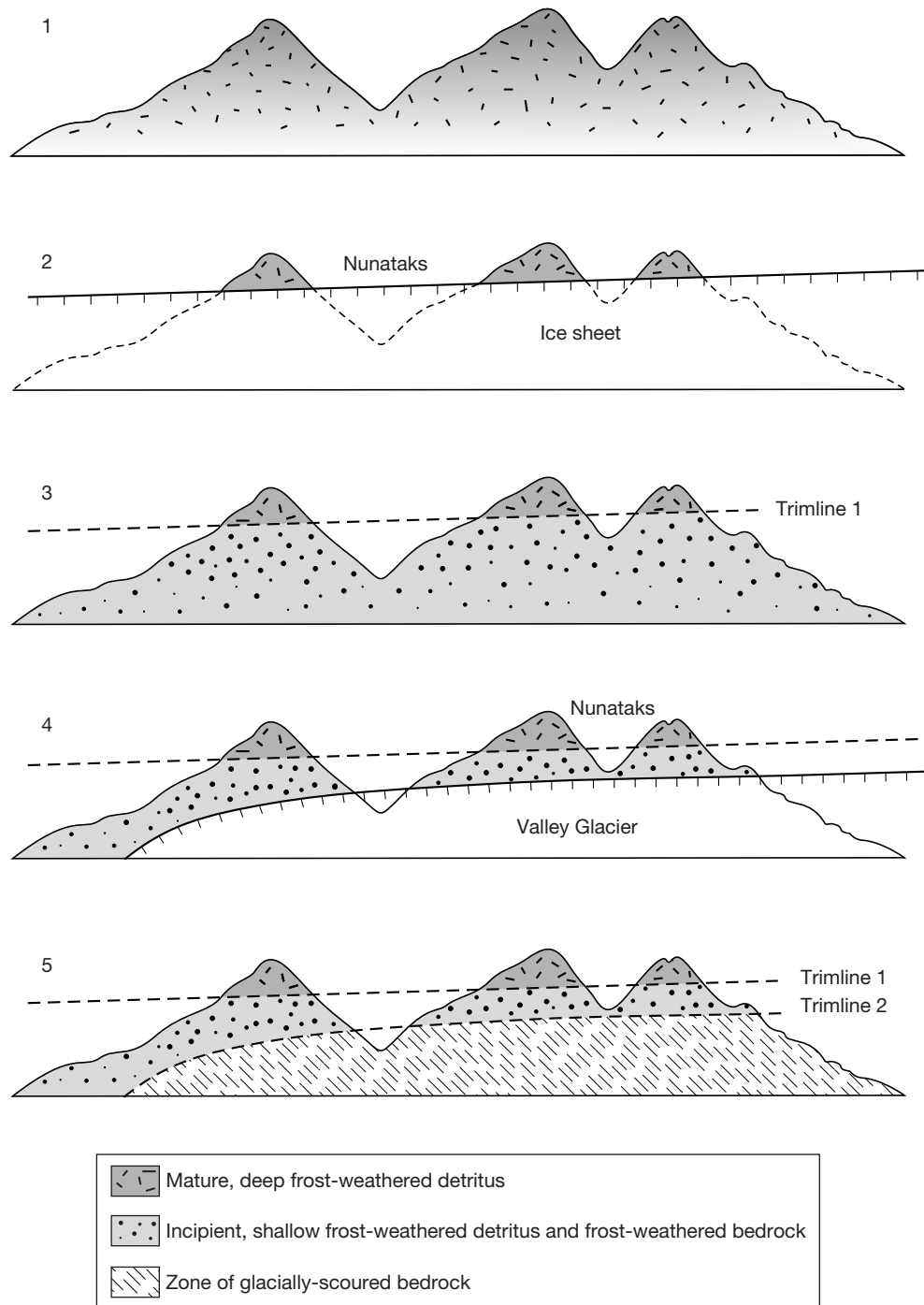


Figure 5 Formation of two trimlines separating three weathering zones. (1) Prolonged weathering results in an extensive regolith mantle. (2) This debris is trimmed by an ice sheet at the LGM. (3) During and after ice-sheet downwastage, weathering attacks bedrock below the trimline. (4) A later ice advance or readvance trims the lower weathering zone. (5) Glacier retreat leaves a zone of relatively unweathered ice-scoured bedrock and glacial drift. Note that in many cases, the upper trimline may have developed within an ice sheet that overtopped the highest summits (Figure 4(c)). From Ballantyne CK and Harris C (1994) *The Periglaciation of Great Britain*, Cambridge, UK. © Cambridge University Press, 1994. Reproduced with permission.

interpretation may be ambiguous (Figure 4), and further analyses are required. Such analyses fall into four categories: relative-age dating techniques, surface exposure dating using cosmogenic isotopes, relationship with other geomorphological evidence, and paleoglaciological considerations.

Relative-Age Dating of Trimlines

Relative-Age Dating of Bedrock Outcrops

The validity of trimlines mapped on the basis of geomorphological evidence has been assessed using three techniques that compare the weathering characteristics of bedrock outcrops



Figure 6 High-level ice-molded landscape at 900 m in the Western Grampian Highlands of Scotland. Rounded bedrock outcrops provide evidence of erosion by warm-based, sliding ice at the LGM, though some detachment of blocks by frost wedging after deglaciation is evident at the foot of the conspicuously rounded outcrop.



Figure 7 Unmodified periglacial landscape of shattered rock outcrops and regolith cover above the maximum level of Otira (LGM) glaciation, Torlesse Range, New Zealand.



Figure 8 Glacially transported perched boulder resting on granite blockfield debris at 796 m on Grytøya, northwest Norway.

above and below proposed trimlines: measurements of rock-surface roughness, Schmidt hammer measurements, and measurements of joint depths. All are based on the assumption that rock surfaces above trimlines have experienced much more prolonged weathering under periglacial conditions than rock surfaces affected by glacial abrasion at lower altitudes.

A microerosion meter was employed by [McCarroll and Nesje \(1993\)](#) to assess differences in rock-surface roughness on augen gneiss at sites above, between, and below two trimlines on Scåla (1843 m), a mountain in western Norway. Their findings indicated a significant increase in surface roughness above the lower trimline, but not across the higher. They concluded that the lower trimline probably represents the upper limit of Late Weichselian glacial abrasion, whereas the upper is of earlier origin. This technique is likely to be successful only for resistant lithologies that develop marked surface irregularities after prolonged exposure.

The Schmidt hammer is a device that records percentage rebound from rock surfaces to blows of constant force; for any lithology, weathered bedrock surfaces generally yield lower rebound values than fresh bedrock surfaces. In NW Scotland, [Ballantyne et al. \(1998\)](#) demonstrated significantly lower aggregate rebound values for quartzite and sandstone bedrock surfaces above mapped trimlines in comparison with those below trimlines ([Figure 9\(a\)](#)). In the same area, they found that the distribution of depths of horizontal stress-release joints exposed on rock faces above and below mapped trimlines differ ([Figure 9\(b\)](#)), with significantly deeper open joints above the trimlines. Both techniques have also been successfully employed to validate a trimline representing the upper limit of LGM glacial erosion in southwest Ireland ([Rae et al., 2004](#)). However, they require large numbers of measurements, and on many lithologies, intact outcrops are rare at above-trimline sites. Moreover, although the three techniques described earlier can be used to validate the geomorphological evidence for trimlines, they offer no conclusive evidence regarding trimline age or genesis.

Soil Weathering Characteristics

An alternative approach to relative-age dating of surfaces above and below proposed trimlines involves comparison of the weathering characteristics of soils. This approach assumes that glacial erosion removes pre-existing soils up to the level of LGM trimlines, so that soils above such trimlines exhibit evidence for more prolonged and advanced weathering than those below.

A method that has been widely employed in confirming mapped trimlines is X-ray diffraction (XRD) analyses of clay-fraction mineralogy of soils above and below mapped trimlines. [Ballantyne et al. \(1998\)](#) summarized the results of 128 analyses carried out on above-trimline and below-trimline soil samples obtained from mountains in northwest Scotland. Their results showed that for most common secondary clay mineral species, there is no difference in representation (presence or absence) above and below mapped LGM trimlines. However, gibbsite, a form of aluminum hydroxide that represents the end product of silicate alteration in a wide range of free-draining environments, is preferentially represented in above-trimline soils, at 78% of sites compared with 10%

below. The presence of gibbsite in samples from below the weathering limit may represent glacial reworking of gibbsitic soils. As gibbsite in Scottish montane soils is believed to be of pre-Devensian (pre-Weichselian) origin, its preferential representation above high-level trimlines indicates that such trimlines represent the upper limit of glacial erosion at the Devensian glacial maximum. Soil samples obtained by [Rae et al. \(2004\)](#) in southwest Ireland also exhibited significant differences in the composition of secondary clay minerals; they detected gibbsite in some above-trimline samples but none of the below-trimline samples.

[Marquette et al. \(2004\)](#) employed a wide range of soil weathering criteria to differentiate soils on blockfield-covered plateaus from those sampled from till and glacially eroded terrain in mountain areas of NE Quebec and Labrador. They demonstrated that the blockfield soils exhibit higher concentrations of crystalline iron and amorphous aluminum and silicium extracted with oxalate, indicating significantly greater weathering of soils above the blockfield limit. They also found that that secondary clay minerals (gibbsite and kaolinite) are also much more abundantly represented above the blockfield limit; gibbsite was detected only in the blockfield soils. They inferred from this evidence and associated cosmogenic isotope analyses that the blockfields probably survived several glacial events under a cover of cold-based ice. [Marquette et al. \(2004\)](#) inferred that the chemical and secondary clay mineral characteristics of blockfield soils on the mountains of northeast Québec and Labrador are “inherited from soil formed under a climate warmer than that prevailing today” (p. 31), and numerous other authors have concluded that the presence of secondary clay minerals in plateau-top regolith indicates inheritance from Neogene saprolite covers that have survived repeated Pleistocene glaciations under a protective cover of cold-based ice (e.g., [André, 2004](#); [Paasche et al., 2006](#)). [Goodfellow et al. \(2009\)](#), however, noted the potential for secondary clay mineralization and gibbsite formation in the early stages of silicate weathering and found only very limited evidence for chemical weathering in blockfield soils at a site in Sweden. It is possible that the secondary clay minerals detected in many blockfields are no older than the last interglacial and reflect regolith weathering under cool temperate or periglacial conditions ([Ballantyne, 2010b](#)).

Comparative analyses of soil weathering characteristics above and below trimlines are useful in identifying areas that escaped significant glacial erosion at the LGM, but do not allow differentiation of ice-marginal trimlines representing the maximum altitude of former ice cover and englacial trimlines representing the former boundary between cold-based and warm-based ice.

Environmental Magnetism

[Walden and Ballantyne \(2002\)](#) have explored the possibility of validating LGM trimlines on the premise that above-trimline soils with prolonged weathering histories may yield significantly different mineral magnetic signatures from those of post-deglaciation soils below trimlines. They found that concentration-dependent mineral magnetic parameters exhibit significant differences for soil samples above and below LGM trimlines, but that the nature of the difference is dependent on

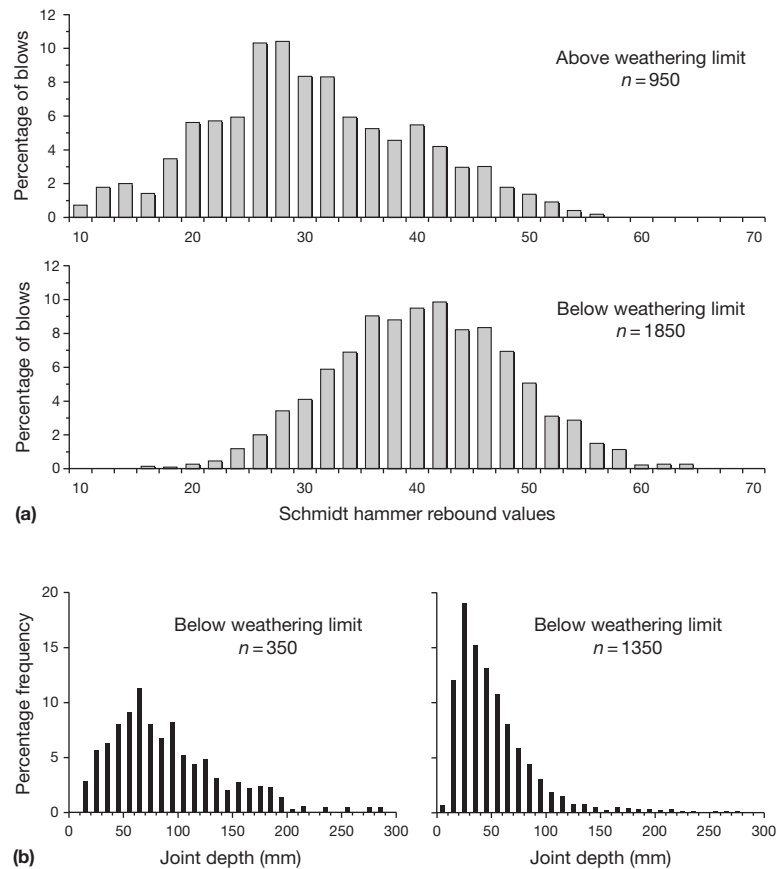


Figure 9 (a) Schmidt hammer rebound response above and below a high-level trimline on sandstone terrain, northwest Scotland. (b) Depths of open (stress-release) joints above and below the same trimline. From Ballantyne CK, McCarroll D, Nesje A, Dahl SO, and Stone JO (1998) The last ice sheet in North-West Scotland: Reconstruction and implications. *Quaternary Science Reviews* 17: 1149–1184. © 1998 Elsevier Science Ltd. Reproduced with permission.

lithology. Basalt soil samples from above the LGM trimline on the Faroe Islands produced significantly higher concentration-dependent magnetic parameters than those below, probably reflecting more pronounced transformation of iron from non-magnetic to magnetic forms. Conversely, for sandstone soils in northwest Scotland, most concentration-dependent magnetic parameters proved significantly lower for samples above the trimline, probably reflecting loss of ferromagnetic minerals through dissolution and oxidation into antiferromagnetic forms. Significant differences in some mineral magnetic parameters were also detected by Rae et al. (2004) in their analyses of soil samples above and below an inferred LGM trimline in southwest Ireland. This approach has considerable promise for validating LGM trimlines, but does not by itself allow differentiation of hypotheses 1–3 above regarding trimline genesis.

Surface Exposure Dating Using Cosmogenic Isotopes

Dating the surface exposure age of bedrock or large boulders above and below mapped trimlines using cosmogenic isotope concentrations (commonly ^{10}Be , ^{26}Al , and ^{36}Cl) represents a powerful methodology not only for validating trimlines, but

also for determining their age and for discriminating between different hypotheses (1–4 above) of trimline origin. Exposure ages obtained from glacially eroded bedrock surfaces and erratic boulders below LGM trimlines should record the approximate timing of deglaciation, whereas those obtained from bedrock or boulders above trimlines should record ages that exceed the age ($\sim 26\text{--}21\text{ ka}$) of the LGM. Similarly, for younger trimlines cut by glaciers of Younger Dryas age, below-trimline surfaces should record ages $< \sim 11.7\text{ ka}$, whereas above-trimline surfaces should record the age ($< \sim 22\text{--}13\text{ ka}$) of earlier ice-sheet downwastage (Colledge et al., 2007).

Two potential drawbacks affect this approach. First, weathering of exposed rock surfaces results in erosional losses, so that measured exposure ages may slightly underestimate the true age of deglaciation. This problem is more acute for above-trimline surfaces where weathering under periglacial conditions has been prolonged, and for this reason, above-trimline exposure ages are generally regarded as minimal. Second, if below-trimline surfaces have not experienced sufficient (2–3 m) removal of rock by glacial erosion, they may retain residual concentrations of cosmogenic nuclides inherited from previous periods of exposure. Erratics and other glacially transported boulders may be similarly affected by complex exposure histories. This problem may be overcome by a careful selection

of bedrock sampling sites (e.g., on the plucked side of roches moutonnées), analyses of multiple samples to identify anomalies, and the use of two unstable isotopes with different half-lives (usually ^{10}Be and ^{26}Al) to establish whether their ratios indicate continuous exposure or complex exposure histories (e.g., Briner et al., 2003; Fabel et al., 2002; Marquette et al., 2004; Staiger et al., 2005; Sugden et al., 2005).

Despite these potential drawbacks, surface exposure dating using cosmogenic isotopes has proved robust not only in establishing the age of trimlines, but also in distinguishing ice-marginal trimlines cut around paleonunataks at the edge of former ice sheets from those representing a former englacial boundary between protective cold-based ice on high ground and erosive ice occupying adjacent valleys. Surface exposure ages obtained for paired above-trimline and below-trimline sites on mountains in Scotland and Ireland provide impressive vindication of the trimline concept: all the 21 samples collected from above-trimline bedrock outcrops or boulders yielded pre-LGM (>26 ka) ages, whereas all the 21 samples collected from ice-molded bedrock or large glacially deposited boulders below trimlines produced post-LGM (<21 ka) exposure ages (Figure 10). This contrast by itself, however, does not resolve the question of whether the trimlines in these areas reflect survival of unglacierized paleonunataks above the surface of the last British-Irish Ice Sheet, or preservation of glacially unmodified bedrock and boulders on high ground under a cover of cold-based ice that was frozen to the underlying substrate. Earlier work favored the first interpretation (e.g., Ballantyne et al., 1998), but the inferred ice-surface altitudes appear inconsistent with recent numerical ice-sheet models driven by proxy climate (Hubbard et al., 2009), evidence that the British-Irish Ice Sheet was much more extensive than previously believed and the presence on some mountains of above-trimline erratics. It now seems probable that most or all trimlines in the British Isles are of englacial origin (Ballantyne, 2010a) and that the areas of periglacial terrain above trimlines represent frozen-bed zones under the former ice sheet (Hall and Glasser, 2003; Kleman and Glasser, 2007), though trimlines delimiting the former maximum altitude of LGM ice may exist in southwest Ireland (Rae et al., 2004; Figure 11).

Elsewhere, however, surface exposure dating has been employed explicitly to test the competing validity of the nunatak hypothesis (Figure 4(a)) and englacial trimline hypothesis (Figure 4(c)). Fabel et al. (2002) employed paired analyses of ^{10}Be and ^{26}Al concentrations from erratics and glacially eroded bedrock on high-level relict surfaces in northern Sweden to demonstrate that these areas were ice-covered at the LGM and thus represent frozen-bed areas within the last Fennoscandian Ice Sheet. Similarly, Marquette et al. (2004) and Staiger et al. (2005) have resolved a long-standing debate on the significance of trimlines on the mountains of Labrador and northeast Quebec by showing that $^{26}\text{Al}/^{10}\text{Be}$ ratios for tors and blockfield boulders imply that these features survived the last and probably earlier glacial maxima under a protective cover of cold-based ice, a conclusion supported by the post-LGM ages of erratics located amid blockfield debris. A similar approach has been used by Sugden et al. (2005) to infer LGM and earlier ice cover on present-day nunataks in Marie Byrd Land, Antarctica (Figure 1). Briner et al. (2003) showed that erratics on weathered tors in eastern Baffin Island yield

exposure ages of 17–11 ka, whereas the tors themselves produced exposure ages >60 ka, demonstrating that the tors survived under a cover of cold-based ice at the LGM. Surface exposure ages obtained for tors on the Cairngorm Mountains of Scotland similarly indicate LGM ice cover despite extensive regolith cover on adjacent plateaus (Phillips et al., 2006). Conversely, Landvik et al. (2003) have employed surface exposure dating to argue that high ground in northwest Svalbard remained ice-free at the LGM, as glacial erratics and other boulders on blockfield-covered plateaus produced pre-LGM exposure ages, in contrast to the post-LGM exposure ages obtained for bedrock and erratic samples on low ground (Figure 12).

Geomorphological Evidence

Geomorphological evidence may, in some circumstances, allow conclusions to be drawn regarding the age and genesis of trimlines. In some cases, valley-side trimlines (Figure 2) may be continued downvalley by lateral moraines or drift limits, demonstrating that they mark the upper limit of coeval glaciation (e.g., Rae et al., 2004; Thorp, 1981, 1986). Consistent decline in trimline altitudes along former flowlines determined by striae, roches moutonnées, and streamlined bedforms also supports their interpretation as delimiting the upper surface of former ice masses but does not exclude an englacial interpretation. Several accounts also suggest that sharp trimlines defining an abrupt altitudinal contrast between ice-scoured and frost-weathered terrain are indicative of an ice-marginal rather than an englacial origin, as in the latter case, the gradual distribution of shear between sliding and cold-bed ice would be expected to produce a transition between weathering zones rather than a sharp boundary, though Marquette et al. (2004, Figure 3) illustrated that even englacial trimlines sometimes form distinct boundaries.

Paleoglaciological Considerations

A final test of the validity of ice-marginal interpretations of trimlines is reconstruction of the basal shear stress (τ_b) of reconstructed ice-surface profiles using

$$\tau_b = F(\rho gh \sin \alpha)$$

where ρ is ice density ($\sim 900 \text{ kg m}^{-3}$), g gravitational acceleration (9.81 m s^{-2}), h reconstructed ice thickness (m), α reconstructed ice-surface slope, and F is a form factor (Paterson, 1994). As glaciers moving over rigid beds yield $\tau_b = 40\text{--}140 \text{ kPa}$, the reconstructed shear stress should fall between these limits. Unfortunately, this constraint is probably too widely defined to prove conclusive in most cases. It is satisfied, for example, by most trimline altitudes and gradients in northwest Scotland (Ballantyne et al., 1998), but these probably represent englacial trimlines defining the altitudinal limit of erosion by warm-based sliding ice within an ice sheet that overtopped all mountains in the area (Ballantyne 2010a; Ballantyne and Hall, 2008). An additional constraint is that the altitudes of ice-marginal trimlines must be consistent with

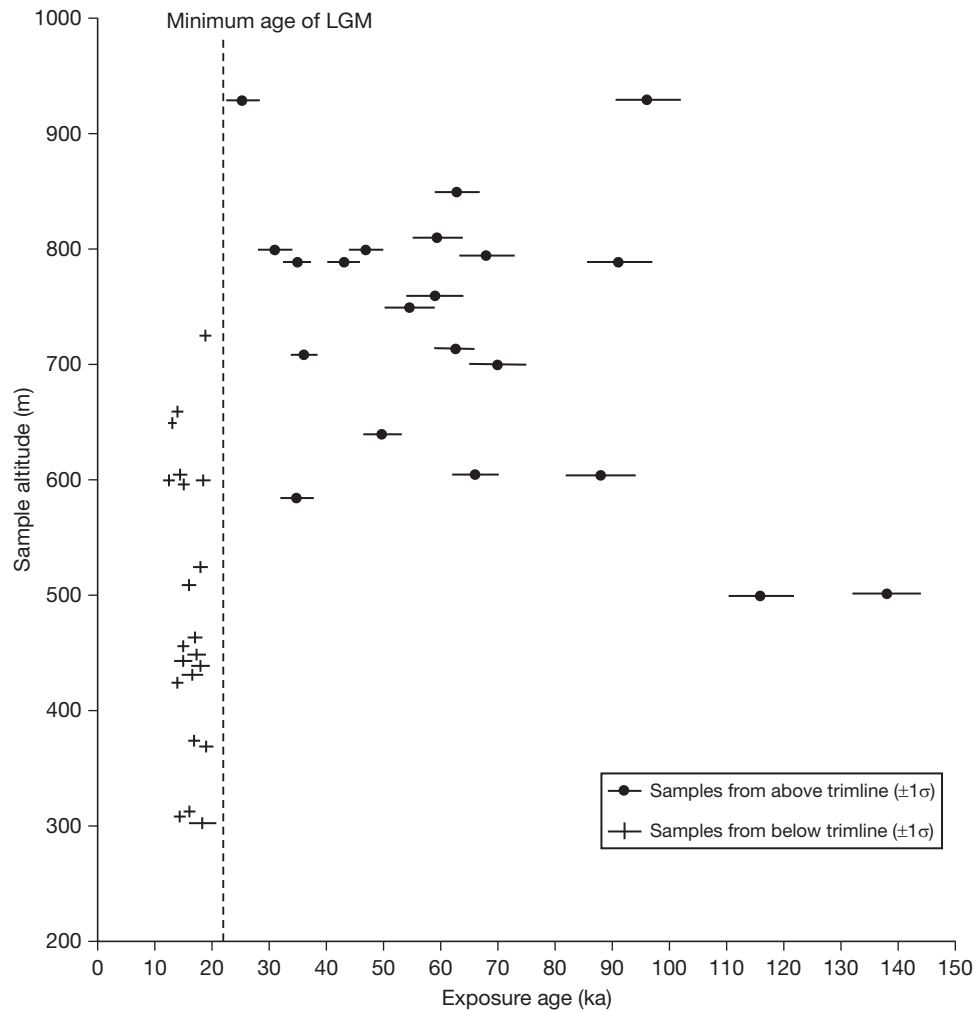


Figure 10 Age–altitude plot of ^{10}Be and ^{36}Cl exposure ages obtained from above and below trimlines that mark the upper limit of glacially modified terrain and the lower boundary of blockfields and shattered bedrock outcrops on mountains in Scotland and Ireland. Horizontal lines represent $\pm 1\sigma$ external uncertainties. All above-trimline ages must be regarded as minima. From Ballantyne CK (2010) Extent and deglaciation chronology of the last British-Irish Ice Sheet: Implications of exposure dating using cosmogenic isotopes. *Journal of Quaternary Science* 25: 515–534. © 2009 John Wiley and Sons Ltd. Reproduced with permission.

the inferred lateral dimensions of former ice masses, particularly with climatic-proxy driven models of ice-sheet dynamics (e.g., Hubbard et al., 2009).

Conclusion

Periglacial (or glacial) trimlines have long been interpreted as representing the approximate upper limit of former glaciers or ice sheets. In many areas, however, it has now been demonstrated that tors, blockfields, and other weathering covers have been preserved under cold-based ice above the upper limit of erosion within a former ice sheet. Interpretation of trimlines as the upper limit of ice masses remains valid, but only if it can be demonstrated that they meet stringent criteria. The most important tests are (1) lack of erratics, perched glacially transported boulders, glacially detached blocks, and glacially modified tors above the trimline; (2) use of $^{26}\text{Al}/^{10}\text{Be}$ ratios to test for uninterrupted exposure of above-trimline boulder

and bedrock surfaces; (3) exposure dating of erratics and/or perched boulders to establish whether these were emplaced at the LGM or by an earlier, thicker ice sheet; and (4) consistency with other geomorphological evidence and the inferred lateral extent of former ice masses. Trimlines that satisfy these conditions permit a detailed reconstruction of former ice dimensions at the LGM in areas where nunataks remained above the ice surface at its maximum thickness, such as the Southern Alps of New Zealand (Figure 2) and Yosemite National Park in California (Figure 3). Trimlines cut across the flanks of paleonunataks and other valley-side slopes may also aid reconstruction of the former dimensions of glaciers that developed during later readvances of more limited extent, such as the outlet glaciers that developed in the Western Highlands of Scotland during the Younger Dryas Chron (e.g., Thorp, 1981, 1986). Conversely, where it can be demonstrated that tors, blockfields, and other weathering covers survived LGM ice cover above an englacial trimline, such evidence not only provides a minimum altitude for former ice cover but also



Figure 11 Trimline on Broaghnbinnia (745 m), southwest Ireland. The summit plateau supports a vegetated blockfield, but lower slopes comprise glacially scoured sandstone bedrock. The trimline separating these zones has been interpreted as representing the maximum altitude of the Cork–Kerry Ice Cap at the LGM, but may represent the uppermost limit of glacial erosion within a thicker ice mass that overtopped the mountain.

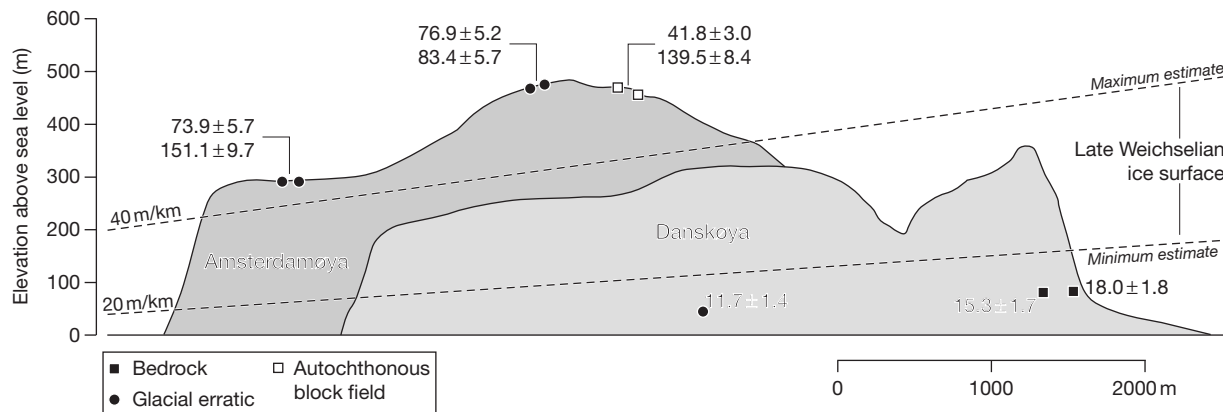


Figure 12 Topographic profiles of Amsterdamøya and Danskoøya in northwest Svalbard, showing locations and ^{10}Be exposure ages for samples from bedrock, erratic boulders, and blockfield boulders, and inferred maximum and minimum Late Weichselian ice-surface profiles. From Landvik JY, Brook EJ, Gualtieri L, Raisbeck G, Salvigsen O, and Yiou F (2003) Northwest Svalbard during the last glaciation: Ice-free areas existed. *Geology* 31: 905–908. © 2003 Geological Society of America. Reproduced with permission.

indicates the location of zones where cold-based ice remained frozen to the underlying substrate. Such evidence may throw light on the former subglacial thermal regime within ice sheets and the location and dimensions of former ice streams within ice sheets, thus permitting a much more detailed empirically based reconstruction of former ice-sheet history and dynamics than has hitherto proved possible (e.g., Hubbard et al., 2009; Sugden et al., 2005).

See also: **Cosmogenic Nuclide Dating: Exposure Geochronology.** **Glacial Landforms, Ice Sheets: Evidence of Glacier and Ice Sheet Extent.** **Glacial Landforms, Sediments: Glacial Erratics and Till Dispersal Indicators.** **Permafrost and Periglacial Features: Blockfields (Felsenmeer); Rock Weathering.**

References

- André M-F (2004) The geomorphic impact of glaciers as indicated by tors in North Sweden (Aurivaara, 68°N). *Geomorphology* 57: 403–421.
- Ballantyne CK (1998) Age and significance of mountain-top detritus. *Permafrost and Periglacial Processes* 9: 327–345.
- Ballantyne CK (2010a) Extent and deglacial chronology of the last British-Irish Ice Sheet: Implications of exposure dating using cosmogenic isotopes. *Journal of Quaternary Science* 25: 515–534.
- Ballantyne CK (2010b) A general theory of autochthonous blockfield evolution. *Permafrost and Periglacial Processes* 21: 289–300.
- Ballantyne CK and Hall AM (2008) The altitude of the last ice sheet in Caithness and east Sutherland. *Scottish Journal of Geology* 44: 169–181.
- Ballantyne CK, McCarroll D, Nesje A, Dahl SO, and Stone JO (1998) The last ice sheet in North-West Scotland: Reconstruction and implications. *Quaternary Science Reviews* 17: 1149–1184.
- Briner JP, Miller GH, Davis PT, Bierman PR, and Caffee M (2003) Last glacial maximum ice sheet dynamics in Arctic Canada inferred from young erratics perched on ancient tors. *Quaternary Science Reviews* 22: 437–444.

- Fabel D, Stroeve AP, Harbor J, Kleman J, Elmore D, and Fink D (2002) Landscape preservation under Fennoscandian ice sheets determined from in situ produced ^{10}Be and ^{26}Al . *Earth and Planetary Science Letters* 201: 397–406.
- Fjellanger J, Sørbel L, Linge H, Brook EJ, Raisbeck GM, and Yiou F (2006) Glacial survival of blockfields on the Varanger Peninsula, northern Norway. *Geomorphology* 82: 255–272.
- Geikie J (1873) On the glacial phenomena of the Long Isle or Outer Hebrides. *Journal of the Geological Society of London* 29: 532–545.
- Golledge NR, Fabel D, Everest JD, Freeman S, and Binnie S (2007) First cosmogenic ^{10}Be age constraint on the timing of Younger Dryas Glaciation and ice cap thickness, western Scottish Highlands. *Journal of Quaternary Science* 22: 785–791.
- Goodfellow BW, Fredin IO, Derron M-H, and Stroeve AP (2009) Weathering processes and Quaternary origin of an alpine blockfield in Arctic Sweden. *Boreas* 38: 379–398.
- Hall AM and Glasser NF (2003) Reconstructing former glacial basal thermal regimes in a landscape of selective linear erosion: Glen Avon, Cairngorm Mountains, Scotland. *Boreas* 32: 191–207.
- Hall AM and Phillips WM (2006) Glacial modification of granite tors in the Cairngorms, Scotland. *Journal of Quaternary Science* 21: 811–830.
- Hubbard A, Bradwell T, Golledge N, et al. (2009) Dynamic cycles, ice streams and their impact on the extent, chronology and deglaciation of the British-Irish ice sheet. *Quaternary Science Reviews* 28: 758–776.
- Ives JD (1978) The maximum extent of the Laurentide ice sheet along the east coast of North America during the last glaciation. *Arctic* 31: 24–53.
- Kleman J and Glasser NF (2007) The subglacial thermal organization (STO) of ice sheets. *Quaternary Science Reviews* 25: 585–597.
- Landvik JY, Brook EJ, Gualtieri L, Raisbeck G, Salvigsen O, and Yiou F (2003) Northwest Svalbard during the last glaciation: Ice-free areas existed. *Geology* 31: 905–908.
- Marquette GC, Gray JT, Gosse JC, et al. (2004) Felsenmeer persistence under non-erosive ice in the Torngat and Kaumajet mountains, Quebec and Labrador, as determined by soil weathering and cosmogenic nuclide exposure dating. *Canadian Journal of Earth Sciences* 41: 19–38.
- McCarroll D and Nesje A (1993) The vertical extent of ice sheets in Nordfjord, western Norway: Measuring degree of rock weathering using Schmidt hammer and rock surface roughness. *Boreas* 22: 255–265.
- Nesje A, Dahl SO, Anda E, and Rye N (1988) Blockfields in southern Norway: Significance for the Late Weichselian ice sheet. *Norsk Geologisk Tidsskrift* 68: 149–169.
- Paasche Ø, Strømsøe JR, Dahl SO, and Linge H (2006) Weathering characteristics of arctic islands in northern Norway. *Geomorphology* 82: 430–452.
- Paterson WSB (1994) *The Physics of Glaciers*, 3rd edn., p. 480. Oxford: Pergamon Press.
- Phillips WM, Hall AM, Mottram R, Fifield LK, and Sugden DE (2006) Cosmogenic ^{10}Be and ^{26}Al exposure ages of tors and erratics, Cairngorm Mountains, Scotland: Timescales for the development of a classic landscape of selective linear erosion. *Geomorphology* 73: 222–245.
- Rae AC, Harrison S, Mighall T, and Dawson AG (2004) Periglacial trimlines and nunataks of the last glacial maximum: The Gap of Dunloe, southwest Ireland. *Journal of Quaternary Science* 19: 87–97.
- Sellier D (1995) Le felsenmeer du Mont Gausta (Telemark, Norvège): environnement, caractères morphologiques et significations paléogéographiques. *Géographie Physique et Quaternaire* 49: 185–205.
- Staiger JKW, Gosse JC, Johnson JV, et al. (2005) Quaternary relief generation by polythermal glacier ice. *Earth Surface Processes and Landforms* 30: 1145–1159.
- Sugden DE, Balco G, Cowderly SG, Stone JO, and Sass LC (2005) Selective glacial erosion and weathering zones in the coastal mountains of Marie Byrd Land, Antarctica. *Geomorphology* 67: 317–334.
- Thorp PW (1981) A trimline method for defining the upper limit of Loch Lomond Advance glaciers: Examples from the Loch Leven and Glencoe areas. *Scottish Journal of Geology* 17: 49–64.
- Thorp PW (1986) A mountain icefield of Loch Lomond Stadial age, western Grampians. *Boreas* 15: 83–97.
- Walden J and Ballantyne CK (2002) Use of environmental magnetic measurements to validate the vertical extent of ice masses at the last glacial maximum. *Journal of Quaternary Science* 17: 193–200.





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2ND EDITION

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DEDICATIONS

The second edition of the *Encyclopedia of Quaternary Science* is dedicated to the memory of three outstanding scientists who have recently passed away: Russell Coope, Bill Watts, and Aleksis Dreimanis.

Russell Coope

G. Russell Coope (1930–2011) was essentially the founding father of Quaternary Entomology, having initiated the modern era of Quaternary insect fossil research in the late 1950s. Prior to that time, and going back to the nineteenth century, there had been a number of false starts and stops in this field. Previous researchers had assumed that any ice-age insect fossils they found would necessarily represent extinct taxa. This, as it turns out, was a completely false assumption. Beginning with the Late Pleistocene Upton Warren and Chelford sites, Russell was able to determine that fossil insects, especially beetles (Coleoptera) from Pleistocene deposits, represent extant species, albeit species that may live in dramatically different regions today than those where they lived in the past. For many years at the start of this endeavor, Russell was ‘a voice in the wilderness,’ having to convince colleagues in both the Quaternary sciences and modern Entomology that it is possible to identify ‘beetle bits’ (individual sclerites of beetle exoskeleton) to the species level, and that fossil beetle assemblages can yield reliable reconstructions of past environments. One of his greatest battles was to convince skeptical palynologists that the interstadial interval in what is now known as MIS 3, and another one toward the end of the last glaciation, were almost as warm as today, in spite of the fact that the vegetation cover represented in pollen records failed to register this. Likewise, Russell championed the idea that climate oscillations, particularly in the late glacial interval and early Holocene, were extremely rapid and dramatic: switching from full glacial to full interglacial conditions in a matter of decades, at most. Russell’s paleoclimate reconstructions were finally given full credence by the scientific community at large when the oxygen isotope records from Greenland ice cores began to be published, showing nearly the exact same amplitude and timing of climate change he had been publishing since the 1960s.

Russell’s research has had important implications for the field of zoogeography. On the basis of his work, we now know that insects respond to climate change chiefly by shifting their distributional ranges, so that they



Figure 1 Russell Coope at the Chelford quarry fossil site, August 1977. Photograph by Alan V. Morgan, used with permission.

track these changes through space. Through his tenacity and hard work, Russell was able to track down the modern ranges of beetle species he found in Pleistocene deposits to such remote locations as northeastern Siberia and the Tibetan Plateau (cold-adapted species), as well as North Africa and the Mediterranean islands (warm-adapted species).

Russell brought a level of curiosity and enthusiasm for natural history and all things scientific that was infectious. He was a brilliant lecturer and storyteller, holding his audiences practically spellbound. He trained many of the paleoentomologists who took the field forward in recent decades. His seemingly boundless energy kept him working and publishing papers, right up to the end of his life. He published more than 200 papers in his long career. He collaborated with dozens of other Quaternary scientists and was one of the first to appreciate and promote multidisciplinary studies of Pleistocene fossil sites. There is no doubt that his research and publishing legacies will continue to inspire future generations of paleoentomologists.

Bill Watts

William (Bill) A. Watts (1930–2010) was a towering figure in Quaternary paleoecology. He was a skilled botanist and ecologist and this was a secret to his success in unraveling the interglacial, late-glacial, and Holocene vegetation history not only in his native Ireland but also in the midwestern, the southeastern, and eastern USA, and in Italy. A major contribution was his reintroduction and development of quantitative plant macrofossil analysis. Macrofossils can often be identified to lower taxonomic levels than pollen, and Bill took full advantage of the complementarity of both methods. Thus, he was able to determine important features of vegetation history by identifying trees to species level and to illuminate the history of aquatic plants and local vegetational developments.



Figure 2 Photograph of Bill Watts, taken at Laguna de la Roya, Spain. Photograph by Fraser Mitchell, used with permission.

Bill always enjoyed fieldwork and microscope work, carrying out the latter in spare time left over from his job as Provost of Trinity College Dublin and his participation on committees dealing with hospitals and conservation in Ireland. Bill's quiet but intense enthusiasm and his great sense of humor added much to discussions, macrofossil identification sessions, and excursions with him. He and his work are inspirations to those who follow him.



Figure 3 Aleksis Dreimanis, photograph by Stephen Hicock.

Aleksis Dreimanis

Aleksis Dreimanis (1914–2011) was widely considered the world's foremost Quaternary glacial geologist in the 1970s and 80s. Aleksis was born in Latvia. While still an undergraduate at the University of Latvia, he published his first paper on the structural geology of deformed glacial deposits along the banks of the River Daugava – a study that he would return to and revise 60 years later. In 1944, Aleksis was conscripted as a military geologist by the Germans who occupied Latvia. For the next two years, he was separated from his family and forced to do strategic geological mapping in eastern Germany and northern Italy until the end of WWII, followed by internment in prisoner of war and displaced persons camps. In 1948, he escaped to London, England, where he met with the President of the University of Western Ontario who gave him a job in London, Canada (his family was able to join him the following year).

Over the next six decades at the University of Western Ontario, Aleksis ascended to the top of his discipline while gaining international recognition through landmark publications and inspired leadership on committees. His 1970s synthesis and definition of the Wisconsin Stage with Paul Karrow, Jan Terasmae, Dick Goldthwait, and Jane Forsyth became a world standard to which other records were compared. In 1989, he expanded this synthesis and introduced the conceptual till tetrahedron that showed the till continuum with end members occupying the corners of an inverted pyramid. He was among the first geologists to use radiocarbon dating, and he provided samples for inter-lab calibration. He was among the first to quantitatively study the relationship between till lithology and till formation, and to investigate how glacial transport and dynamics determine till texture, deformation, and rheologic states. Aleksis produced over 200 publications spanning eight decades and over a wide range of topics. Largely because of his influence, the study of till genesis and its relation to till properties and past glacial behavior has become a major focus of glacial geologic research in recent decades. The contribution and legacy that forms the body of work of Dreimanis will continue as a major benchmark within the diverse fields of scientific enquiry that are sometimes loosely referred to as 'glacial geology.'

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FOREWORD

As with the publication of the first edition in 2007, the publication of the second edition of the *Encyclopedia of Quaternary Science* represents a landmark in the history of publishing in the field of Quaternary Science. Quaternary Science is a multidisciplinary endeavor which seeks to establish as detailed a picture as possible of the manifold environmental changes that have occurred during the most recent geological period, the Quaternary – an interval of time that spans the past 2.59 million years or the past 0.056% of geological time. It is a period of significant climate and environmental change and witnessed the widespread dispersal of our species, *Homo sapiens*, across the planet.

Since Louis Agassiz and Reverend William Buckland traipsed over parts of the Scottish landscape in 1840 in search of evidence of glaciation, a huge literature has emerged on the science of long-term climate and environmental change. Rapid technological advances in the late twentieth century and the proliferation of scientific journals, particularly in an era of electronic publishing, have resulted in an exponential growth and documentation of knowledge on the climate and environmental changes that have occurred during the Quaternary period. More recently, there has been an increasing public interest in applied Quaternary research as a framework for understanding the basis for recent climate changes and for understanding the nature and frequency of geological hazards and vexing issues such as soil erosion and land degradation, and the adverse effects of ocean water temperature increases and acidification on coral reef environments. In a similar manner, the likely magnitude of future sea-level rise and the associated impacts on coastal landscapes in the twenty-first century have attracted wide public interest. In this sense, Quaternary Science is very much on the political agenda and is a critically important subject to address issues of public concern.

The *Encyclopedia of Quaternary Science* edited by Scott Elias of Royal Holloway, University of London, UK, accordingly represents a particularly welcome addition to the literature. The encyclopedia presents an up-to-date and authoritative overview of Quaternary Science. The encyclopedia should enjoy a wide readership as the entries are presented in a very clear and easily readable style. The text of the articles is written at a level that allows undergraduate students to understand the material, while providing active researchers with a ready reference resource for information in the field. Each entry of up to 4000 words covers the salient points of each topic with very clear illustrations. A central theme that pervades the work is the importance of Quaternary Science in providing an historical context for assessing present environmental changes and as basis for modeling potential future changes.

The encyclopedia consists of four volumes in print form and is available electronically. All the entries have been updated and the text totals around 3,500,000 words. As a major reference work, the encyclopedia has a very wide coverage of topics within the Quaternary sciences reflecting the complex and interdisciplinary nature of the science. Each of the major sections begins with a general overview of the topic prepared by a leading expert in the field. The major sections, for example, examine the analytical methods commonly used in paleoenvironmental reconstructions to unravel in a forensic-like manner the nature of former environments and the tempo of environmental change. Accordingly, a great range of topics such as the former extent of ice cover and nature of Quaternary glaciations, the biological responses and resultant fossil records to fluctuating climate, the expansion and contraction of desert environments, and global and local changes in relative sea levels are examined. Other topics covered include dating techniques, Quaternary stratigraphy, fluvial environments, lake level studies, paleosols, paleobotany, ancient DNA, paleolimnology, vertebrate studies, insect fossil studies, paleoceanography, stable isotope studies, ice core records, and human evolution in the Quaternary. One of the new sections in the encyclopedia examines the application of Quaternary proxy evidence in forensic science. All sections provide a clear summary of the latest advances in the fields of research.

This is an outstanding work and the editors and the publisher are to be congratulated for producing an encyclopedia that cogently summarizes the current state of the science.

A handwritten signature in dark ink, appearing to read 'C. Murray-Wallace', with a stylized, flowing script.

Colin V. Murray-Wallace
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INTRODUCTION

It has been my privilege to work with more than 420 leading Quaternary scientists in developing the second edition of the *Encyclopedia of Quaternary Science*. This team of writers and editors represents 28 countries in Europe, Asia, Africa, the Americas, Australia, and New Zealand. Starting with the first edition in 2006, I have had my finger on the pulse of Quaternary science, and this branch of science is truly pulsating! Information now comes from an incredible variety of disciplines: geochemistry, numerical modeling, history, vulcanology, paleobiology, nuclear physics, stratigraphy, sedimentology, climatology, anthropology, archaeology, glacial geology, soil science, ice-stream modeling – the list is staggering. This highly disparate group of people are bound together by one common thread: the desire to know the history of the planet during the last 2.6 million years – the time of the ice ages. For Quaternary scientists, this is a pressing need, not an idle curiosity. Any doubts about this statement can be easily dispelled by a consideration of the lengths to which many of them go to gather the necessary data. Some of them have worked for months in sub-zero temperatures on top of very high mountains, or near the center of polar ice sheets, collecting ice cores. Others have spent many weeks on some of the roughest seas in the world, drilling deep-sea sediment cores. Often the work is more mundane. An oxygen isotope curve for a lengthy marine sediment core represents thousands of hours of patiently picking tiny fossils from layer after layer of sediments, in order to obtain sufficient numbers of calcium carbonate shells to yield samples for isotopic analysis. A map showing proposed ice limits from the last glaciation represents thousands of hours of field mapping of glacial features by dozens of people. Why do all of these people devote their lives to this pursuit of knowledge? Does it really matter so much? The answer becomes clear when you step back and examine the topic of Quaternary science in its proper context.

The world we inhabit has largely been shaped by the events of the Quaternary. All the biological communities that exist today are the end product of a long series of species associations that came together in the past, largely driven by climatic change during the Pleistocene. We cannot properly understand the functioning of modern ecosystems without a solid knowledge of their history, any more than we can understand the plot of a long novel by reading just the final page. We are also living in a time of alarming climate changes. Even though the pace and intensity of some of these changes have not been seen in historical times, there were many rapid, large-scale climatic shifts in the Pleistocene. The best way to predict the effects of global warming on the planet's climate and ecosystems is to look at the history of similarly intense, rapid changes in the prehistoric past. The interval that is most relevant today is the most recent geologic period: the Quaternary. As human populations rise exponentially, increasing numbers of people are exposed to geologic hazards, such as earthquakes (and attendant tsunamis), and volcanic eruptions. These are short-lived events that take place only rarely in any one region. The interval between major events, such as volcanic eruptions, may be centuries or millennia. How do we come to grips with predicting the future likelihood of such erratic phenomena? Again, the answers come from piecing together the ancient history of such events, over many thousands of years.

The Quaternary has been the time when our own species came of age. The beginning of the Quaternary, roughly 2.6 million years ago, was about the time when the earliest member of our genus (*Homo*) first appeared in Africa. Pleistocene environments shaped the course of human evolution, culminating in anatomically modern *Homo sapiens* spreading from Africa throughout most of the world during the last glaciation. Even though human beings largely shape their own environments today, for the vast majority of our species' history, it has been the environment that has shaped us. Our direct ancestors' adaptations to environmental change are deeply ingrained in our genes. Thus, an understanding of the environmental conditions that shaped our species is critical to our understanding of humanity.

Quaternary science is a rapidly changing field, and the articles that appear in this encyclopedia reflect this. New dating techniques, such as cosmogenic nuclide dating, are revolutionizing our understanding of many earth surface processes. The ability to analyze increasingly smaller samples for radiocarbon and stable isotopes of oxygen and hydrogen means that we are gaining a level of precision in the reconstruction of past events that was unheard of just a few years ago. Stable isotope studies of air bubbles trapped in ice cores from Greenland and Antarctica have given Quaternary scientists an entirely new perspective on the rapidity and intensity of climatic change during the last glacial cycle and beyond. Likewise, the discovery of long sequences of annually laminated sediments in both marine and freshwater environments has provided a great leap forward in our ability to resolve the timing of environmental changes in nonpolar regions. The ability to extract and analyze ancient DNA sequences from Pleistocene fossils (both plants and animals) is revolutionizing the field of paleobiology. We are beginning to be able to trace the genetic lineages of a number of different organisms, from beetles to bison. In short, these are very exciting times to be a Quaternary scientist! While it is virtually impossible for any Quaternary researcher or student to keep abreast of all the new discoveries in this multifaceted science, this encyclopedia can be of great help. The articles contained here represent the state of the art in a huge variety of topics, and they offer the opportunity to dig deeper into their respective subjects by providing full citations of the most pertinent literature available. I invite you to come and explore the Quaternary Period in the pages that follow. It is a fascinating story.

Scott A. Elias

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HOW TO USE THE ENCYCLOPEDIA

The material in this encyclopedia is organized into two broad sections. At the start of Volume 1 are four *Introductory Articles* that serve as a broad introduction to the field. It is intended that these articles should serve as the first port of call for readers, to enable them to embark on any of the main entries without the need to consult other material beforehand.

Following the Introductory Articles, the main body of The Encyclopedia is arranged as a series of entries in alphabetical order. These entries fill the remainder of the volumes.

To help realize the full potential of the material in the Encyclopedia we have provided four features to help you find the topic of your choice: a contents list by subject; an alphabetical contents list; cross-references to other articles; and a full subject index.

1. Contents list by subject

Your first point of reference will probably be the contents list by subject. This list appears at the front of each volume, and groups the entries under subject headings describing the broad themes of Quaternary science. This will enable the reader to make quick connections between entries and to locate the entry of interest. Under each main section heading, you will find several subject areas and under each subject area is a list of those entries that covers aspects of that subject, together with the volume and page numbers on which these entries may be found.

2. Alphabetical contents list

The alphabetical contents list, which also appears at the front of each volume, lists the entries in the order in which they appear in the Encyclopedia. This list provides both the volume number and the page number of each entry.

You will find ‘dummy’ entries where obvious synonyms exist for entries or where we have grouped together related topics. Dummy entries appear in both the contents list and the body of the text.

Example

If you were attempting to locate material on *peat studies* via the alphabetical contents list:

PEAT STUDIES *see* PLANT MACROFOSSIL METHODS AND STUDIES: Mire and Peat Macros

The dummy entry directs you to the entry in which *peat* is covered.

If you were trying to locate the material by browsing through the text and you had looked up *Peat Studies*, then the following information would be provided in the dummy entry:

Peat Studies *see* PLANT MACROFOSSIL METHODS AND STUDIES: Mire and Peat Macros

3. Cross-references

All of the entries in the Encyclopedia have been extensively cross-referenced. The cross-references, which appear at the end of the entry, serve three different functions:

- i. To indicate if a topic is discussed in greater detail elsewhere.
- ii. To draw the reader’s attention to parallel discussions in other entries.
- iii. To indicate material that broadens the discussion.

Example

The following list of cross-references appear at the end of the entry GLACIAL LANDFORMS, ICE SHEETS | Growth and Decay

See also: **Introduction:** History of Quaternary Science; **Paleoclimate Reconstruction:** Sub-Milankovitch (DO/Heinrich) Events; **Glacial Climates:** Thermohaline Circulation; **Glaciation, Causes:** Astronomical Theory of Paleoclimates; **Paleoclimate Modeling:** Last Glacial Maximum GCMs; **Glaciations:** Overview; **Sea Level Studies:** Overview; Paleocceanography; **Ice Core Records:** Correlations Between Greenland and Antarctica.

Here you will find examples of all three functions of the cross-reference list: a topic discussed in greater detail elsewhere (e.g., **Paleoclimate Modeling:** Last Glacial Maximum GCMs), parallel discussion in other entries (e.g., **Glacial Climates:** Thermohaline Circulation), and reference to entries that broaden the discussion (e.g., **Glaciations:** Overview).

4. Index

The index provides you with the page number where the material is located. The index entries differentiate between material that is a whole entry, is part of an entry, or is data presented in a figure or a table. Detailed notes are provided on the opening page of the index.

5. Contributors

A full list of contributors is listed at the beginning of each volume.

CONTENTS LIST BY SUBJECT

Location references refer to the volume number and page number (separated by a colon)

Introduction and History of Science

INTRODUCTION | History of Quaternary Science, 1:1
INTRODUCTION | History of Dating Methods, 1:9
INTRODUCTION | Societal Relevance of Quaternary Research, 1:17
INTRODUCTION | Understanding Quaternary Climatic Change, 1:26

Paleoclimatology

Overview and Approaches

PALEOCLIMATE | Introduction, 3:87
PALEOCLIMATE RECONSTRUCTION | Approaches, 3:179
PALEOCLIMATE MODELING | Data-Model Comparisons, 3:135
PALEOCLIMATE | Timescales of Climate Change, 3:93
PALEOCLIMATE | Modern Analog Approaches in Paleoclimatology, 3:102
PALEOCLIMATE | Paleoclimate History of the Arctic, 3:113
PALEOCLIMATE | The Younger Dryas Climate Event, 3:126

Paleoclimate Reconstructions

PALEOCLIMATE RECONSTRUCTION | Pliocene Environments, 3:185
GLACIATION, CAUSES | Tectonic Uplift and Continental Configurations, 2:127
GLACIATION, CAUSES | Astronomical Theory of Paleoclimates, 2:136
PALEOCLIMATE RECONSTRUCTION | Paleodroughts and Society, 3:194
PALEOCLIMATE RECONSTRUCTION | Paleotempestology, 3:209

PALEOCLIMATE RECONSTRUCTION | Sub-Milankovitch (DO/Heinrich) Events, 3:200
PALEOCLIMATE RECONSTRUCTION | Younger Dryas Oscillation, Global Evidence, 3:222
PALEOCLIMATE RECONSTRUCTION | Historical Climatology, 3:237
PALEOCLIMATE RECONSTRUCTION | The Last Millennium, 3:229
PALEOCLIMATE RELEVANCE TO GLOBAL WARMING, 3:244

Climate Forcing Mechanisms

GLACIAL CLIMATES | Thermohaline Circulation, 1:737
GLACIAL CLIMATES | Biosphere Feedbacks, 1:721
GLACIAL CLIMATES | Volcanic and Solar Forcing, 1:748
GLACIAL CLIMATES | Effects of Atmospheric Dust, 1:729

Climate Modelling

PALEOCLIMATE MODELING | Quaternary Environments, 3:147
PALEOCLIMATE MODELING | The Last Interglacial, 3:155
PALEOCLIMATE MODELING | Last Glacial Maximum GCMs, 3:165
PALEOCLIMATE MODELING | Paleo-ENSO, 3:171

Dating Techniques

Radiometric Methods

DATING TECHNIQUES, 1:447
RADIOCARBON DATING | Conventional Method, 4:305
RADIOCARBON DATING | AMS Radiocarbon Dating, 4:316

RADIOCARBON DATING | Sources of Error, 4:324
 RADIOCARBON DATING | Variations in Atmospheric ^{14}C , 4:329
 RADIOCARBON DATING | Causes of Temporal ^{14}C Variations, 4:336
 RADIOCARBON DATING | Calibration of the ^{14}C Record, 4:345
 RADIOCARBON DATING | Charcoal, 4:353
 RADIOCARBON DATING | ^{14}C of Plant Macrofossils, 4:361
 K/Ar AND $^{40}\text{Ar}/^{39}\text{Ar}$ DATING, 2:477
 U-SERIES DATING, 4:567
 FISSION-TRACK DATING, 1:643
 LUMINESCENCE DATING | Thermoluminescence, 2:643
 LUMINESCENCE DATING | Optical Dating, 2:653
 LUMINESCENCE DATING | Electron Spin Resonance Dating, 2:667
 COSMOGENIC NUCLIDE DATING | Overview, 1:407
 COSMOGENIC NUCLIDE DATING | Cosmic Ray Interactions in Minerals, 1:418
 COSMOGENIC NUCLIDE DATING | Methods, 1:410
 COSMOGENIC NUCLIDE DATING | Exposure Geochronology, 1:432
 COSMOGENIC NUCLIDE DATING | Landscape Evolution, 1:440

Geomagnetic Excursions and Secular Variations in the Magnetic Field

Geomagnetic Excursions and Secular Variations, 1:705

Biological Methods

LICHENOMETRY, 2:565

Annual Layer Methods

VARVED MARINE SEDIMENTS, 4:582
 VARVED LAKE SEDIMENTS, 4:573

Chemical Methods

AMINO-ACID DATING, 1:37

Quaternary Stratigraphy

QUATERNARY STRATIGRAPHY | Tephrochronology, 4:277
 QUATERNARY STRATIGRAPHY | Overview, 4:189
 QUATERNARY STRATIGRAPHY | Lithostratigraphy, 4:227

QUATERNARY STRATIGRAPHY | Continental Biostratigraphy, 4:206
 QUATERNARY STRATIGRAPHY | Pedostratigraphy, 4:250
 QUATERNARY STRATIGRAPHY | Morphostratigraphy/Allostratigraphy, 4:243
 QUATERNARY STRATIGRAPHY | Climatostratigraphy, 4:222
 QUATERNARY STRATIGRAPHY | Sequence Stratigraphy, 4:260
 QUATERNARY STRATIGRAPHY | Chronostratigraphy, 4:215

Quaternary Glaciation

Glacial Landforms

GLACIAL LANDFORMS | Introduction, 1:755
 GLACIAL LANDFORMS, ICE SHEETS | Growth and Decay, 1:877
 GLACIAL LANDFORMS, ICE SHEETS | Evidence of Glacier and Ice Sheet Extent, 1:884
 GLACIAL LANDFORMS | Evidence of Glacier Recession, 1:803
 GLACIAL LANDFORMS, ICE SHEETS | PaleoeLAS, 1:909
 GLACIAL LANDFORMS | Glacitectonic Structures and Landforms, 1:839
 GLACIAL LANDFORMS, SEDIMENTS | Glaciogenic Lithofacies, 2:18
 GLACIAL LANDFORMS | Moraine Forms and Genesis, 1:769
 GLACIAL LANDFORMS, SEDIMENTS | Glaciofluvial Landforms of Deposition, 2:6
 GLACIAL LANDFORMS | Glaciofluvial Landforms of Erosion, 1:825
 GLACIAL LANDFORMS, SEDIMENTS | Glacial Erratics and Till Dispersal Indicators, 2:81
 GLACIAL LANDFORMS, SEDIMENTS | Glaciomarine Sediments and Ice-Rafted Debris, 2:30
 GLACIAL LANDFORMS, SEDIMENTS | Tills, 2:62
 GLACIAL LANDFORMS, SEDIMENTS | Till Fabric Analysis, 2:76
 GLACIAL LANDFORMS | Glacial Land Systems, 1:813
 GLACIAL LANDFORMS, EROSIONAL FEATURES | Micro- to Macroscale Forms, 1:865
 GLACIAL LANDFORMS, EROSIONAL FEATURES | Major Scale Forms, 1:847
 GLACIAL LANDFORMS, ICE SHEETS | Trimlines and Palaeonunataks, 1:918
 GLACIAL LANDFORMS | Quaternary Vulcanism: Subglacial Landforms, 1:780
 GLACIAL LANDFORMS, SEDIMENTS | Micromorphology of Glacial Sediments, 2:52

GLACIAL LANDFORMS, SEDIMENTS | Clast Form Analysis, 2:1
 GLACIAL LANDFORMS, ICE SHEETS | Evidence of Glacier Flow Directions, 1:895
 GLACIAL LANDFORMS, SEDIMENTS | Glaciolacustrine, 2:43
 GLACIAL LANDFORMS, SEDIMENTS | Glacial Sequence Stratigraphy, 2:85

Permafrost and Periglacial Features

PERMAFROST AND PERIGLACIAL FEATURES | Permafrost, 3:464
 PERMAFROST AND PERIGLACIAL FEATURES | Active Layer Processes, 3:421
 PERMAFROST AND PERIGLACIAL FEATURES | Paraglacial Geomorphology, 3:553
 PERMAFROST AND PERIGLACIAL FEATURES | Slope Deposits and Forms, 3:481
 PERMAFROST AND PERIGLACIAL FEATURES | Frost Mounds: Active and Relict Forms, 3:472
 PERMAFROST AND PERIGLACIAL FEATURES | Patterned Ground, 3:452
 PERMAFROST AND PERIGLACIAL FEATURES | Talus Slopes, 3:566
 PERMAFROST AND PERIGLACIAL FEATURES | Thermokarst Topography, 3:574
 PERMAFROST AND PERIGLACIAL FEATURES | Ice Wedges and Ice-Wedge Casts, 3:436
 PERMAFROST AND PERIGLACIAL FEATURES | Cryoturbation Structures, 3:430
 PERMAFROST AND PERIGLACIAL FEATURES | Blockfields (Felsenmeer), 3:523
 PERMAFROST AND PERIGLACIAL FEATURES | Block/Rock Streams, 3:514
 PERMAFROST AND PERIGLACIAL FEATURES | Rock Weathering, 3:500
 PERMAFROST AND PERIGLACIAL FEATURES | Rock Glaciers and Protalus Forms, 3:535
 PERMAFROST AND PERIGLACIAL FEATURES | Periglacial Fluvial Sediments and Forms, 3:490
 PERMAFROST AND PERIGLACIAL FEATURES | Permafrost and Glacier Interactions, 3:507
 PERMAFROST AND PERIGLACIAL FEATURES | Yedoma: Late Pleistocene Ice-Rich Syngenetic Permafrost of Beringia, 3:542

Tree-ring section

DENDROCHRONOLOGY, 1:453
 DENDROCLIMATOLOGY, 1:459
 GLACIAL LANDFORMS, TREE RINGS | Dendrogeomorphology, 2:91
 GLACIAL LANDFORMS, TREE RINGS | Dendroglaciology, 2:104

Fluvial Environments

FLUVIAL ENVIRONMENTS | Sediments, 1:663
 FLUVIAL ENVIRONMENTS | Terrace Sequences, 1:684
 Glacial–Interglacial Scale Fluvial Responses, 2:112
 FLUVIAL ENVIRONMENTS | Responses to Rapid Environmental Change, 1:676
 FLUVIAL ENVIRONMENTS | Deltaic Environments, 1:693
 PALEOHYDROLOGY, 3:253

History of Quaternary Glaciations

GLACIATIONS | Overview, 2:143
 GLACIATIONS | Transition from Late Neogene to Early Quaternary Environments, 2:151
 GLACIATIONS | Early Quaternary (Pleistocene) and Precursors, 2:167
 GLACIATIONS | Middle Pleistocene in Eurasia, 2:172
 GLACIATIONS | Mid-Quaternary in North America, 2:180
 GLACIATIONS | Middle Pleistocene Glaciations in the Southern Hemisphere, 2:187
 GLACIATIONS | Late Quaternary of Antarctica, 2:216
 GLACIATIONS | Late Quaternary in North America, 2:245
 GLACIATIONS | Late Quaternary in Highland Asia, 2:236
 GLACIATIONS | Late Quaternary of the Southwest Pacific Region, 2:202
 GLACIATIONS | Late Pleistocene in Eurasia, 2:224
 GLACIATIONS | Late Pleistocene in South America, 2:250
 GLACIATIONS | Late Pleistocene Glacial Events in Beringia, 2:191
 GLACIATIONS | Neoglaciation in Europe, 2:257
 GLACIATIONS | Neoglaciation in the American Cordilleras, 2:269

Sea Level Studies

SEA LEVEL STUDIES | Overview, 4:369
 SEA LEVEL STUDIES | Geomorphological Indicators, 4:377
 SEA LEVEL STUDIES | Coral Records of Relative Sea-Level Changes, 4:409
 SEA LEVEL STUDIES | Use of Cave Data in Sea-Level Reconstructions, 4:460
 SEA LEVEL STUDIES | Eustatic Sea-Level Changes – Glacial–Interglacial Cycles, 4:429
 SEA LEVEL STUDIES | Eustatic Sea-Level Changes Since the Last Glacial Maximum, 4:439

SEA LEVEL STUDIES | Isostasy: Glaciation-Induced Sea-Level Change, 4:452
SEA LEVEL STUDIES | Sedimentary Indicators of Relative Sea-Level Changes – High Energy, 4:385
SEA LEVEL STUDIES | Sedimentary Indicators of Relative Sea-Level Changes – Low Energy, 4:396
SEA LEVEL STUDIES | Microfossil-Based Reconstructions of Holocene Relative Sea-Level Change, 4:419
SEA-LEVELS, LATE QUATERNARY | Late Quaternary Relative Sea-Level Changes in High Latitudes, 4:467
SEA-LEVELS, LATE QUATERNARY | Late Quaternary Sea-Level Changes in Greenland, 4:481
SEA-LEVELS, LATE QUATERNARY | Late Quaternary Relative Sea-Level Changes at Mid-Latitudes, 4:489
SEA-LEVELS, LATE QUATERNARY | Late Quaternary Relative Sea-Level Changes in the Tropics, 4:495
SEA-LEVELS, LATE QUATERNARY | Tectonics and Relative Sea-Level Change, 4:503

Paleosols and Wind-Blown Sediments

Paleosol Research

PALEOSOLS AND WIND-BLOWN SEDIMENTS | Overview, 3:357
PALEOSOLS AND WIND-BLOWN SEDIMENTS | Nature of Paleosols, 3:367
LOESS DEPOSITS: ORIGINS AND PROPERTIES, 2:573
PALEOSOLS AND WIND-BLOWN SEDIMENTS | Weathering Profiles, 3:392
PALEOSOLS AND WIND-BLOWN SEDIMENTS | Soil Micromorphology, 3:381
PALEOSOLS AND WIND-BLOWN SEDIMENTS | Soil Morphology in Quaternary Studies, 3:402
PALEOSOLS AND WIND-BLOWN SEDIMENTS | Mineral Magnetic Analysis, 3:375
PALEOSOLS AND WIND-BLOWN SEDIMENTS | Biogeochemical Role of Dust in Quaternary Climate Cycles, 3:412

Dune Studies

DUNE FIELDS | High Latitudes, 1:597
DUNE FIELDS | Mid-Latitudes, 1:606
DUNE FIELDS | Low Latitudes, 1:623

Loess Studies

LOESS RECORDS | Central Asia, 2:585
LOESS RECORDS | China, 2:595
LOESS RECORDS | Europe, 2:606
LOESS RECORDS | North America, 2:620
LOESS RECORDS | South America, 2:629
EOLIAN RECORDS, DEEP-SEA SEDIMENTS, 1:637

Lake Level Studies

LAKE LEVEL STUDIES | Overview, 2:483
LAKE LEVEL STUDIES | Modeling, 2:558
LAKE LEVEL STUDIES | Africa during the Late Quaternary, 2:499
LAKE LEVEL STUDIES | Asia, 2:506
LAKE LEVEL STUDIES | Australia, 2:524
LAKE LEVEL STUDIES | Latin America, 2:531
LAKE LEVEL STUDIES | North America, 2:537
LAKE LEVEL STUDIES | West-Central-Europe, 2:549

Paleobotany

Overview

PALEOBOTANY | Overview of Terrestrial Pollen Data, 2:699

Pollen Studies

POLLEN ANALYSIS, PRINCIPLES, 3:794
POLLEN METHODS AND STUDIES | Use of Pollen as Climate Proxies, 3:805
POLLEN METHODS AND STUDIES | Surface Samples and Trapping, 3:839
POLLEN METHODS AND STUDIES | The Biome Approach to Reconstructing Past Vegetation, 3:861
POLLEN METHODS AND STUDIES | POLLSCAPE Model, 3:871
POLLEN METHODS AND STUDIES | Reconstructing Past Biodiversity Development, 3:816
POLLEN METHODS AND STUDIES | Numerical Analysis Methods, 3:821
POLLEN METHODS AND STUDIES | Changing Plant Distributions and Abundances, 3:854
POLLEN METHODS AND STUDIES | Stand-Scale Palynology, 3:846
PALEOBOTANY | Paleophytogeography, 2:730
POLLEN METHODS AND STUDIES | Databases and Their Application, 3:831

Charred Particle Analysis

PALEOBOTANY | Charred Particle Analyses, 2:716

Quaternary Pollen Records

POLLEN RECORDS, LAST INTERGLACIAL OF EUROPE, 4:1
POLLEN RECORDS, LATE PLEISTOCENE | Africa, 4:9
POLLEN RECORDS, LATE PLEISTOCENE | Australasia, 4:18
POLLEN RECORDS, LATE PLEISTOCENE | Northern Asia, 4:27

POLLEN RECORDS, LATE PLEISTOCENE | Northern North America, 4:39
 POLLEN RECORDS, LATE PLEISTOCENE | Western North America, 4:72
 POLLEN RECORDS, LATE PLEISTOCENE | South America, 4:52
 POLLEN RECORDS, LATE PLEISTOCENE | Middle and Late Pleistocene in Southern Europe, 4:63
 POLLEN RECORDS, POSTGLACIAL | Africa, 4:85
 POLLEN RECORDS, POSTGLACIAL | Australia and New Zealand, 4:104
 POLLEN RECORDS, POSTGLACIAL | Northeastern North America, 4:115
 POLLEN RECORDS, POSTGLACIAL | Northwestern North America, 4:124
 POLLEN RECORDS, POSTGLACIAL | Southeastern North America, 4:133
 POLLEN RECORDS, POSTGLACIAL | Southwestern North America, 4:142
 POLLEN RECORDS, POSTGLACIAL | South America, 4:156
 POLLEN RECORDS, POSTGLACIAL | Northern Asia, 4:164
 POLLEN RECORDS, POSTGLACIAL | Northern Europe, 4:173
 POLLEN RECORDS, POSTGLACIAL | Southern Europe, 4:179
 POLLEN METHODS AND STUDIES | Archaeological Applications, 3:880

Phytoliths

PHYTOLITHS, 3:582

Ancient Plant DNA

PALEOBOTANY | Ancient Plant DNA, 2:705

Plant Macrofossil Studies

PLANT MACROFOSSIL INTRODUCTION, 3:593
 PLANT MACROFOSSIL METHODS AND STUDIES | Surface Samples, Taphonomy, Representation, 3:684
 PLANT MACROFOSSIL METHODS AND STUDIES | Treeline Studies, 3:690
 PLANT MACROFOSSIL METHODS AND STUDIES | Megafossils, 3:621
 PLANT MACROFOSSIL METHODS AND STUDIES | Paleolimnological Applications, 3:657
 PLANT MACROFOSSIL METHODS AND STUDIES | CO₂ Reconstruction from Fossil Leaves, 3:613
 PLANT MACROFOSSIL METHODS AND STUDIES | Rodent Middens, 3:674

PLANT MACROFOSSIL METHODS AND STUDIES | Mire and Peat Macros, 3:637
 PLANT MACROFOSSIL METHODS AND STUDIES | Use in Environmental Archaeology, 3:699
 PLANT MACROFOSSIL METHODS AND STUDIES | Dendroarchaeology, 3:630
 PLANT MACROFOSSIL METHODS AND STUDIES | Validation of Pollen Studies, 3:725

Plant Macrofossil Records

PLANT MACROFOSSIL RECORDS | Arctic Eurasia, 3:733
 PLANT MACROFOSSIL RECORDS | Greenland, 3:760
 PLANT MACROFOSSIL RECORDS | Arctic North America, 3:746
 PLANT MACROFOSSIL RECORDS | Holocene North America, 3:768
 PLANT MACROFOSSIL RECORDS | Late Glacial Multidisciplinary Studies, 3:785

Diatom Studies

DIATOM INTRODUCTION, 1:471
 DIATOM METHODS | Data Interpretation, 1:489
 DIATOM RECORDS | North Atlantic and Arctic, 1:562
 DIATOM RECORDS | Pacific, 1:571
 DIATOM RECORDS | Antarctic Waters, 1:527
 DIATOM RECORDS | Diatom Fossil Records from Marine Laminated Sediments, 1:554
 DIATOM RECORDS | Freshwater Laminated Sequences, 1:540
 DIATOM RECORDS | Large Lakes, 1:546
 DIATOM METHODS | Salinity and Climate Reconstructions from Continental Lakes, 1:507
 DIATOM METHODS | $\delta^{18}\text{O}$ Records, 1:481
 PALEOBOTANY | Silicon Isotopes in Diatoms, 2:734
 DIATOM METHODS | Use in Archaeology, 1:516
 DIATOM METHODS | Diatomites: Their Formation, Distribution, and Uses, 1:501
 Structures and Applications of Biomarkers from Arctic Sea Ice Diatoms, 1:588

Paleolimnology

PALEOLIMNOLOGY | Overview of Paleolimnology, 3:259
 PALEOLIMNOLOGY | Lake Chemistry, 3:292
 PALEOLIMNOLOGY | Physical Properties of Lake Sediments, 3:300
 PALEOLIMNOLOGY | Cladocera, 3:271
 PALEOLIMNOLOGY | Freshwater Mollusks, 3:281

PALEOLIMNOLOGY | Contributions of
Paleolimnological Research to
Biogeography, 3:313
PALEOLIMNOLOGY | Pigment Studies, 3:326
PALEOLIMNOLOGY | Visible and Infrared
Spectroscopical Applications, 3:349
PALEOLIMNOLOGY | Multiproxy Approaches, 3:339

Vertebrate Studies

VERTEBRATE OVERVIEW, 4:590
VERTEBRATE RECORDS | Early Pleistocene, 4:599
VERTEBRATE RECORDS | Early and Middle
Pleistocene of Northern Eurasia, 4:605
VERTEBRATE RECORDS | Mid-Pleistocene of
Southern Asia, 4:651
VERTEBRATE RECORDS | Mid-Pleistocene of
Africa, 4:615
VERTEBRATE RECORDS | Mid-Pleistocene of
Australia, 4:621
VERTEBRATE RECORDS | Mid-Pleistocene of
Europe, 4:639
VERTEBRATE RECORDS | Mid-Pleistocene of
North America, 4:646
VERTEBRATE RECORDS | Late Pleistocene of
Africa, 4:664
VERTEBRATE RECORDS | Late Pleistocene of North
America, 4:673
VERTEBRATE RECORDS | Late Pleistocene of South
America, 4:680
VERTEBRATE RECORDS | Late Pleistocene of
Southeast Asia, 4:693
VERTEBRATE RECORDS | Late Pleistocene
Megafaunal Extinctions, 4:700
VERTEBRATE RECORDS | Late Pleistocene
Mummified Mammals, 4:713
VERTEBRATE STUDIES | Ancient DNA, 4:719
VERTEBRATE STUDIES | Speciation and
Evolutionary Trends in Quaternary
Vertebrates, 4:723
VERTEBRATE STUDIES | Dwarfing and Gigantism in
Quaternary Vertebrates, 4:733
VERTEBRATE STUDIES | Interactions with
Hominids, 4:748

Insect Fossil Studies

Beetle Studies

BEETLE RECORDS | Overview, 1:161
BEETLE RECORDS | Late Tertiary and Early
Quaternary Records, 1:173
BEETLE RECORDS | Middle Pleistocene of Western
Europe, 1:184

BEETLE RECORDS | Late Pleistocene of
Europe, 1:200
BEETLE RECORDS | Late Pleistocene of North
America, 1:221
BEETLE RECORDS | Late Pleistocene of South
America, 1:235
BEETLE RECORDS | Late Pleistocene of
Australia, 1:191
BEETLE RECORDS | Late Pleistocene of Northern
Asia, 1:255
BEETLE RECORDS | Late Pleistocene of Japan, 1:207
BEETLE RECORDS | Late Pleistocene of New
Zealand, 1:244
BEETLE RECORDS | Postglacial Europe, 1:274
BEETLE RECORDS | Postglacial North America, 1:282

Chironomid Studies

CHIRONOMID RECORDS | Chironomid
Overview, 1:355
CHIRONOMID RECORDS | Late Pleistocene of
Europe, 1:373
CHIRONOMID RECORDS | Africa, 1:361
CHIRONOMID RECORDS | Late Pleistocene of
Europe, 1:373
CHIRONOMID RECORDS | Postglacial
Europe, 1:386
CHIRONOMID RECORDS | Postglacial Southern
Hemisphere, 1:398

Oribatid Mite Studies

ORIBATID MITES, 2:680

Paleoceanography

Paleoceanographic Studies

PALEOCEANOGRAPHY | Paleoceanography An
Overview, 2:745
PALEOCEANOGRAPHY, BIOLOGICAL PROXIES |
Corals, Sclerosponges and Mollusks, 2:795
PALEOCEANOGRAPHY, BIOLOGICAL PROXIES |
Benthic Foraminifera, 2:765
PALEOCEANOGRAPHY, BIOLOGICAL PROXIES |
Planktic Foraminifera, 2:825
PALEOCEANOGRAPHY, BIOLOGICAL PROXIES |
Radiolarians and Silicoflagellates, 2:830
PALEOCEANOGRAPHY, BIOLOGICAL PROXIES |
Coccolithophores, 2:783
PALEOCEANOGRAPHY, BIOLOGICAL PROXIES |
Alkenone Paleothermometry Based on the
Haptophyte Algae, 2:755
PALEOCEANOGRAPHY, BIOLOGICAL PROXIES |
Marine Diatoms, 2:816

PALEOCEANOGRAPHY, BIOLOGICAL PROXIES |
Dinoflagellates, 2:800

PALEOCEANOGRAPHY, PHYSICAL AND
CHEMICAL PROXIES | Terrigenous
Sediments, 2:941

PALEOCEANOGRAPHY, PHYSICAL AND
CHEMICAL PROXIES | Oxygen Isotopic
Composition of Seawater, 2:915

PALEOCEANOGRAPHY, PHYSICAL AND
CHEMICAL PROXIES | Oxygen-Isotope
Stratigraphy of the Oceans, 2:907

PALEOCEANOGRAPHY, PHYSICAL AND
CHEMICAL PROXIES | Dissolution of Deep-Sea
Carbonates, 2:859

PALEOCEANOGRAPHY, PHYSICAL AND
CHEMICAL PROXIES | Mg/Ca and Sr/Ca
Paleothermometry from Calcareous Marine
Fossils, 2:871

PALEOCEANOGRAPHY, BIOLOGICAL PROXIES |
Temperature Proxies Census Counts, 2:841

PALEOCEANOGRAPHY, PHYSICAL AND
CHEMICAL PROXIES | Salinity Proxies
 $\delta^{18}\text{O}$, 2:932

PALEOCEANOGRAPHY, PHYSICAL AND
CHEMICAL PROXIES | Nutrient Proxies, 2:899

PALEOCEANOGRAPHY, PHYSICAL AND
CHEMICAL PROXIES | Carbon Cycle
Proxies, 2:849

PALEOCEANOGRAPHY, PHYSICAL AND
CHEMICAL PROXIES | Radioisotope
Proxies, 2:923

PALEOCEANOGRAPHY, BIOLOGICAL PROXIES |
Biomarkers Indicators of Past Climate, 2:775

PALEOCEANOGRAPHY, PHYSICAL AND
CHEMICAL PROXIES | Magnetic Proxies and
Susceptibility, 2:884

Paleoceanographic Records

PALEOCEANOGRAPHY, RECORDS | Early
Pleistocene, 3:1

PALEOCEANOGRAPHY, RECORDS | Late
Pleistocene North Atlantic, 3:9

PALEOCEANOGRAPHY, RECORDS | Late
Pleistocene South Atlantic, 3:18

PALEOCEANOGRAPHY, RECORDS | Late
Pleistocene North Pacific, 3:33

PALEOCEANOGRAPHY, RECORDS | Postglacial
North Atlantic, 3:55

PALEOCEANOGRAPHY, RECORDS | Postglacial
North Pacific, 3:62

PALEOCEANOGRAPHY, RECORDS | Postglacial
South Pacific, 3:73

PALEOCEANOGRAPHY, RECORDS | Postglacial
Indian Ocean, 3:46

Ice Core Studies

Ice Core Studies

ICE CORE METHODS | Overview, 2:278

ICE CORES | History of Research, Greenland and
Antarctica, 2:435

ICE CORE METHODS | Chronologies, 2:303

ICE CORE METHODS | Stable Isotopes, 2:347

ICE CORE METHODS | Conductivity Studies, 2:319

ICE CORE METHODS | Glaciochemistry, 2:326

ICE CORE METHODS | Biological Material, 2:288

ICE CORE METHODS | Microparticle and Trace
Element Studies, 2:342

ICE CORE METHODS | CO₂ Studies, 2:311

ICE CORE METHODS | Methane Studies, 2:334

ICE CORE METHODS | ¹⁰Be and Cosmogenic
Radionuclides in Ice Cores, 2:353

ICE CORE METHODS | Studies of Firn Air, 2:361

Ice Core Records

ICE CORES | History of Carbon Monoxide and
Ultra-Trace Gases from Ice Cores, 2:463

ICE CORES | History of Nitrous Oxide from Ice
Cores, 2:471

ICE CORES | Dynamics of the Greenland Ice
Sheet, 2:439

ICE CORES | Dynamics of the East Antarctic Ice
Sheet, 2:456

ICE CORES | Dynamics of the West Antarctic Ice
Sheet, 2:448

ICE CORE METHODS | Borehole Temperature
Records, 2:298

ICE CORE RECORDS | Greenland Stable
Isotopes, 2:403

ICE CORE RECORDS | Antarctic Stable Isotopes, 2:395

ICE CORE RECORDS | Ice Margin Sites, 2:416

ICE CORE RECORDS | Correlations Between
Greenland and Antarctica, 2:410

ICE CORE RECORDS | Africa, 2:373

ICE CORE RECORDS | Chinese, Tibetan
Mountains, 2:379

ICE CORE RECORDS | South America, 2:387

ICE CORE RECORDS | Thermal Diffusion
Paleotemperature Records, 2:431

Stable Isotope Research – Carbonates

CARBONATE STABLE ISOTOPES | Overview, 1:291

CARBONATE STABLE ISOTOPES |
Speleothems, 1:294

CARBONATE STABLE ISOTOPES | Terrestrial Teeth
and Bones, 1:304

CARBONATE STABLE ISOTOPES | Terrestrial
Organic Materials, 1:315
CARBONATE STABLE ISOTOPES | Non-Lacustrine
Terrestrial Studies, 1:322
CARBONATE STABLE ISOTOPES | Lake
Sediments, 1:333
CARBONATE STABLE ISOTOPES | Nonmarine
Biogenic Carbonates, 1:341

Humans in the Quaternary

ARCHAEOLOGICAL RECORDS | Overview, 1:49
ARCHAEOLOGICAL RECORDS | 2.7 Myr–300 000
Years Ago in Africa, 1:59
ARCHAEOLOGICAL RECORDS | 2.7 Myr–300 000
Years Ago in Asia, 1:67
ARCHAEOLOGICAL RECORDS | 1.9 Myr–300 000
Years Ago in Europe, 1:83
ARCHAEOLOGICAL RECORDS | Global Expansion
300 000–8000 Years Ago, Africa, 1:91
ARCHAEOLOGICAL RECORDS | Global Expansion
300 000–8000 Years Ago, Asia, 1:98
ARCHAEOLOGICAL RECORDS | Global
Expansion 300 000–8000 Years Ago,
Australia, 1:108

ARCHAEOLOGICAL RECORDS | Global Expansion
300 000–8000 Years Ago, Americas, 1:119
ARCHAEOLOGICAL RECORDS | Neanderthal
Demise, 1:146
ARCHAEOLOGICAL RECORDS | Human Evolution
in the Quaternary, 1:135
ARCHAEOLOGICAL RECORDS | Postglacial
Adaptations, 1:154

Use of Quaternary proxies in forensic science

USE OF QUATERNARY PROXIES IN FORENSIC
SCIENCE | Use of Geology in Forensic Science:
Search to Locate Burials, 4:522
USE OF QUATERNARY PROXIES IN FORENSIC
SCIENCE | Analysis Techniques in Forensic
Palynology, 4:556
USE OF QUATERNARY PROXIES IN FORENSIC
SCIENCE | Soils and Sediment, 4:535
USE OF QUATERNARY PROXIES IN FORENSIC
SCIENCE | The Use of Macroscopic Plant Remains
in Forensic Science, 4:542
USE OF QUATERNARY PROXIES IN FORENSIC
SCIENCE | Insects, 4:548
Diatoms, 1:522

CONTENTS

<i>Dedications</i>	<i>v</i>
<i>Foreword</i>	<i>ix</i>
<i>Introduction</i>	<i>xi</i>
<i>Editorial Advisory Board</i>	<i>xiii</i>
<i>Contributors</i>	<i>xv</i>
<i>How to Use the Encyclopedia</i>	<i>xxvii</i>
<i>Contents List by Subject</i>	<i>xxix</i>

VOLUME 1

INTRODUCTORY ARTICLES 1

History of Quaternary Science <i>S A Elias</i>	1
History of Dating Methods <i>A G Wintle</i>	9
Societal Relevance of Quaternary Research <i>S A Elias</i>	17
Understanding Quaternary Climatic Change <i>J J Lowe, M J C Walker, and S C Porter</i>	26

A

ALKENONE STUDIES *see* PALEOCEANOGRAPHY, BIOLOGICAL PROXIES: Alkenone Paleothermometry Based on the Haptophyte Algae

ALLOSTRATIGRAPHY *see* QUATERNARY STRATIGRAPHY: Morphostratigraphy/Allostratigraphy

Amino Acid Dating <i>G H Miller, D S Kaufman, and S J Clarke</i>	37
---	----

ANATOMICALLY MODERN HUMANS *see* ARCHAEOLOGICAL RECORDS: Global Expansion 300 000–8000 Years Ago, Africa; Global Expansion 300 000–8000 Years Ago, Asia; Global Expansion 300 000–8000 Years Ago, Australia; Global Expansion 300 000–8000 Years Ago, Americas

ARCHAEOLOGICAL RECORDS 49

Overview <i>C Gamble</i>	49
-----------------------------	----

2.7 Myr–300 000 Years Ago in Africa <i>J W K Harris, D R Braun, and M Pante</i>	59
2.7 Myr–300 000 Years Ago in Asia <i>R Dennell</i>	67
1.9 Myr–300 000 Years Ago in Europe <i>J McNabb</i>	83
Global Expansion 300 000–8000 Years Ago, Africa <i>A E Close and T Minichillo</i>	91
Global Expansion 300 000–8000 Years Ago, Asia <i>M D Petraglia and R Dennell</i>	98
Global Expansion 300 000–8000 Years Ago, Australia <i>R Cosgrove</i>	108
Global Expansion 300 000–8000 Years Ago, Americas <i>T Goebel</i>	119
Human Evolution in the Quaternary <i>L Cashmore</i>	135
Neanderthal Demise <i>W Davies</i>	146
Postglacial Adaptations <i>G Bailey</i>	154

B

BEETLE RECORDS	161
Overview <i>S A Elias</i>	161
Late Tertiary and Early Quaternary Records <i>S A Elias and S Kuzmina</i>	173
Middle Pleistocene of Western Europe <i>G R Coope</i>	184
Late Pleistocene of Australia <i>N Porch</i>	191
Late Pleistocene of Europe <i>G Lemdahl and G R Coope</i>	200
Late Pleistocene of Japan <i>M Hayashi</i>	207
Late Pleistocene of North America <i>S A Elias</i>	221
Late Pleistocene of South America <i>A C Ashworth</i>	235
Late Pleistocene of New Zealand <i>M Marra</i>	244
Late Pleistocene of Northern Asia <i>A Sher and S Kuzmina</i>	255

Postglacial Europe <i>P Ponel</i>	274
Postglacial North America <i>S A Elias</i>	282

BERINGIA *see* ARCHAEOLOGICAL RECORDS: Global Expansion 300 000–8000 Years Ago, Americas;
 BEETLE RECORDS: Late Pleistocene of North America; DUNE FIELDS: High Latitudes; GLACIATIONS:
 Late Pleistocene Glacial Events in Beringia; PALEOCEANOGRAPHY, RECORDS: Postglacial North
 Pacific; PLANT MACROFOSSIL RECORDS: Arctic North America; POLLEN RECORDS, LATE
 PLEISTOCENE: Northern North America; VERTEBRATE RECORDS: Late Pleistocene of North America

BIOGENIC CARBONATE STUDIES *see* CARBONATE STABLE ISOTOPES: Nonmarine Biogenic Carbonates

BOND CYCLES *see* PALEOCLIMATE RECONSTRUCTION: Sub-Milankovitch (DO/Heinrich) Events

C

CARBONATE STABLE ISOTOPES 291

Overview <i>H Schwarcz</i>	291
-------------------------------	-----

Speleothems <i>H Schwarcz</i>	294
----------------------------------	-----

Terrestrial Teeth and Bones <i>H Bocherens and D G Drucker</i>	304
---	-----

Terrestrial Organic Materials <i>D McCarroll and N Loader</i>	315
--	-----

Non-Lacustrine Terrestrial Studies <i>J Quade and T Cerling</i>	322
--	-----

Lake Sediments <i>S M Bernasconi and J A McKenzie</i>	333
--	-----

Nonmarine Biogenic Carbonates <i>S J Carpenter</i>	341
---	-----

CARBON DIOXIDE, ATMOSPHERIC CONCENTRATIONS *see* CARBONATE STABLE ISOTOPES:
 Overview; ICE CORE METHODS: CO₂ Studies; PALEOCEANOGRAPHY, PHYSICAL AND CHEMICAL
 PROXIES: Carbon Cycle Proxies ($\delta^{11}\text{B}$, $\delta^{13}\text{C}_{\text{calcite}}$, $\delta^{13}\text{C}_{\text{organic}}$, Shell Weights, B/Ca, U/Ca, Zn/Ca, Ba/Ca);
 PLANT MACROFOSSIL METHODS AND STUDIES: CO₂ Reconstruction from Fossil Leaves

CAVE ART *see* VERTEBRATE STUDIES: Interactions with Hominids

CHARCOAL STUDIES *see* PALEOBOTANY: Charred Particle Analyses

CHIRONOMID RECORDS 355

Chironomid Overview <i>I R Walker</i>	355
--	-----

Africa <i>H Eggermont and D Verschuren</i>	361
---	-----

Late Pleistocene of Europe <i>S J Brooks</i>	373
---	-----

Postglacial Europe <i>G Velle and O Heiri</i>	386
--	-----

Postglacial Southern Hemisphere <i>J Massferro and M Vandergoes</i>	398
--	-----

CLADOCERA STUDIES *see* PALEOLIMNOLOGY: Cladocera

CLIMATE CHANGE *see* PALEOCLIMATE: Introduction; Timescales of Climate Change; PALEOCLIMATE MODELING: Data–Model Comparisons; Quaternary Environments; The Last Interglacial; Last Glacial Maximum GCMs; Paleo-ENSO; PALEOCLIMATE RECONSTRUCTION: Approaches; Pliocene Environments; Paleodroughts and Society; Sub-Milankovitch (DO/Heinrich) Events; Paleotempestology; Younger Dryas Oscillation, Global Evidence; The Last Millennium; Historical Climatology; Paleoclimate Relevance to Global Warming

CLIMATE MODELING, QUATERNARY *see* PALEOCLIMATE MODELING: Quaternary Environments

COCCOLITHOPHORE STUDIES *see* PALEOCEANOGRAPHY, BIOLOGICAL PROXIES: Alkenone Paleothermometry Based on the Haptophyte Algae; Coccolithophores

COLEOPTERA FOSSIL RECORDS *see* BEETLE RECORDS: Overview; Late Tertiary and Early Quaternary Records; Middle Pleistocene of Western Europe; Late Pleistocene of Australia; Late Pleistocene of Europe; Late Pleistocene of Japan; Late Pleistocene of North America; Late Pleistocene of South America; Late Pleistocene of New Zealand; Late Pleistocene of Northern Asia; Postglacial Europe; Postglacial North America

CORAL STUDIES *see* PALEOCEANOGRAPHY, BIOLOGICAL PROXIES: Corals, Sclerosponges and Mollusks; SEA LEVEL STUDIES: Coral Records of Relative Sea-Level Changes

COSMOGENIC NUCLIDE DATING 407

Overview <i>J C Gosse</i>	407
------------------------------	-----

Methods <i>J M Schaefer and N Lifton</i>	410
---	-----

Cosmic Ray Interactions in Minerals <i>D Lal</i>	418
---	-----

Exposure Geochronology <i>S Ivy-Ochs and F Kober</i>	432
---	-----

Landscape Evolution <i>D E Granger</i>	440
---	-----

CUT-MARKED BONE *see* VERTEBRATE STUDIES: Interactions with Hominids

D

DANSGAARD-OESCHGER EVENTS *see* PALEOCLIMATE RECONSTRUCTION: Paleodroughts and Society

Dating Techniques <i>A J T Jull</i>	447
--	-----

DENDROARCHAEOLOGY *see* PLANT MACROFOSSIL METHODS AND STUDIES: Dendroarchaeology

Dendrochronology <i>B L Coulthard and D J Smith</i>	453
--	-----

Dendroclimatology <i>B H Luckman</i>	459
---	-----

Diatom Introduction <i>V J Jones</i>	471
DIATOM METHODS	481
$\delta^{18}\text{O}$ Records <i>M J Leng, P A Barker, G E A Swann, and A M Snelling</i>	481
Data Interpretation <i>A Korhola</i>	489
Diatomites: Their Formation, Distribution, and Uses <i>R J Flower</i>	501
Salinity and Climate Reconstructions from Continental Lakes <i>S C Fritz</i>	507
Use in Archaeology <i>N G Cameron</i>	516
Diatoms <i>N G Cameron</i>	522
DIATOM RECORDS	527
Antarctic Waters <i>C E Stickley, J Pike, and V J Jones</i>	527
Freshwater Laminated Sequences <i>H Simola</i>	540
Large Lakes <i>A W Mackay</i>	546
Diatom Fossil Records from Marine Laminated Sediments <i>J Pike and C E Stickley</i>	554
North Atlantic and Arctic <i>N Koç, A Miettinen, and C E Stickley</i>	562
Pacific <i>I Koizumi</i>	571
Structures and Applications of Biomarkers from Arctic Sea Ice Diatoms <i>S T Belt, G Massé, and M Poulin</i>	588
DIATOMS, MARINE <i>see</i> PALEOCEANOGRAPHY, BIOLOGICAL PROXIES: Marine Diatoms	
DINOFLAGELLATES <i>see</i> PALEOCEANOGRAPHY, BIOLOGICAL PROXIES: Dinoflagellates	
DNA, FOSSIL ANIMAL <i>see</i> VERTEBRATE STUDIES: Ancient DNA	
DNA, FOSSIL PLANTS <i>see</i> PALEOBOTANY: Ancient Plant DNA	
DROUGHT HISTORY RECONSTRUCTION <i>see</i> PALEOCLIMATE RECONSTRUCTION: Paleodroughts and Society	
DUNE FIELDS	597
High Latitudes <i>S A Wolfe</i>	597
Mid-Latitudes <i>J Sun and D R Muhs</i>	606

Low Latitudes <i>N Lancaster</i>	623
-------------------------------------	-----

E

EL NIÑO SOUTHERN OSCILLATION <i>see</i> PALEOCEANOGRAPHY, RECORDS: Postglacial South Pacific; PALEOCLIMATE MODELING: Paleo-ENSO	
ELECTRON SPIN RESONANCE DATING <i>see</i> LUMINESCENCE DATING: Electron Spin Resonance Dating	
EOLIAN SEDIMENTS <i>see</i> PALEOSOLS AND WIND-BLOWN SEDIMENTS: Overview; Nature of Paleosols; Mineral Magnetic Analysis; Soil Micromorphology; Weathering Profiles; Soil Morphology in Quaternary Studies; Biogeochemical Role of Dust in Quaternary Climate Cycles	
Eolian Records, Deep-Sea Sediments <i>D K Rea</i>	637
EQUILIBRIUM LINE ALTITUDE (ELA) RECONSTRUCTION <i>see</i> GLACIAL LANDFORMS, ICE SHEETS: Paleo-ELAs	
ERRATICS, GLACIAL <i>see</i> GLACIAL LANDFORMS, SEDIMENTS: Glacial Erratics and Till Dispersal Indicators	
EXTINCTIONS, QUATERNARY VERTEBRATES <i>see</i> VERTEBRATE RECORDS: Late Pleistocene Megafaunal Extinctions	

F

FELSENMEER (BLOCKFIELDS) <i>see</i> PERMAFROST AND PERIGLACIAL FEATURES: Block/Rock Streams	
Fission-Track Dating <i>J A Westgate, N D Naeser, and B V Alloway</i>	643

FLUVIAL ENVIRONMENTS 663

Sediments <i>A Aslan</i>	663
Responses to Rapid Environmental Change <i>T E Törnqvist</i>	676
Terrace Sequences <i>D J Merritts</i>	684
Deltaic Environments <i>L Giosan and S L Goodbred</i>	693

FORAMINIFERA STUDIES *see* PALEOCEANOGRAPHY: Paleocyanography An Overview; PALEOCEANOGRAPHY, BIOLOGICAL PROXIES: Temperature Proxies, Census Counts; Benthic Foraminifera; Planktic Foraminifera; PALEOCEANOGRAPHY, PHYSICAL AND CHEMICAL PROXIES: Oxygen-Isotope Stratigraphy of the Oceans

FOSSIL MITES *see* Oribatid Mites

G

Geomagnetic Excursions and Secular Variations <i>A P Roberts and G M Turner</i>	705
--	-----

GLACIAL CLIMATES	721
Biosphere Feedbacks <i>V Brovkin</i>	721
Effects of Atmospheric Dust <i>I Tegen</i>	729
Thermohaline Circulation <i>S Rahmstorf</i>	737
Volcanic and Solar Forcing <i>D T Shindell</i>	748
GLACIAL LANDFORMS	755
Introduction <i>D J A Evans and D I Benn</i>	755
Moraine Forms and Genesis <i>D J A Evans</i>	769
Quaternary Vulcanism: Subglacial Landforms <i>J L Smellie</i>	780
Evidence of Glacier Recession <i>P M Colgan</i>	803
Glacial Landsystems <i>D J A Evans</i>	813
Glaciofluvial Landforms of Erosion <i>A E Kehew, M L Lord, and A L Kozlowski</i>	825
Glacitectonic Structures and Landforms <i>D J A Evans</i>	839
GLACIAL LANDFORMS, EROSIONAL FEATURES	847
Major Scale Forms <i>I S Evans</i>	847
Micro- to Macroscale Forms <i>B R Rea</i>	865
GLACIAL LANDFORMS, ICE SHEETS	877
Growth and Decay <i>M J Siegert</i>	877
Evidence of Glacier and Ice Sheet Extent <i>D M Mickelson and C Winguth</i>	884
Evidence of Glacier Flow Directions <i>C R Stokes</i>	895
Paleo-ELAs <i>A Nesje</i>	909
Trimlines and Paleonunataks <i>C K Ballantyne</i>	918

VOLUME 2

GLACIAL LANDFORMS, SEDIMENTS 1

Clast Form Analysis <i>D I Benn</i>	1
Glaciofluvial Landforms of Deposition <i>J L Carrivick and A J Russell</i>	6
Glaciogenic Lithofacies <i>N Eyles and M Lazorek</i>	18
Glaciomarine Sediments and Ice-Rafted Debris <i>C Ó Cofaigh</i>	30
Glaciolacustrine <i>D J A Evans</i>	43
Micromorphology of Glacial Sediments <i>J F Hiemstra</i>	52
Tills <i>D J A Evans</i>	62
Till Fabric Analysis <i>D I Benn</i>	76
Glacial Erratics and Till Dispersal Indicators <i>D J A Evans</i>	81
Glacial Sequence Stratigraphy <i>D J A Evans</i>	85

GLACIAL LANDFORMS, TREE RINGS 91

Dendrogeomorphology <i>H Gärtner and I Heinrich</i>	91
Dendroglaciology <i>B L Coulthard and D J Smith</i>	104
Glacial–Interglacial Scale Fluvial Responses <i>M D Blum</i>	112

GLACIATION, CAUSES 127

Tectonic Uplift and Continental Configurations <i>L A Owen</i>	127
Astronomical Theory of Paleoclimates <i>A Berger, M-F Loutre, and Q Z Yin</i>	136

GLACIATIONS 143

Overview <i>J Ehlers and P L Gibbard</i>	143
Transition from Late Neogene to Early Quaternary Environments <i>M Sarnthein</i>	151
Early Quaternary (Pleistocene) and Precursors <i>J Ehlers, V Astakhov, P L Gibbard, Ó Ingólfsson, J Mangerud, and J I Svendsson</i>	167

Middle Pleistocene in Eurasia <i>J Ehlers, V Astakhov, P L Gibbard, J Mangerud, and J I Svendsen</i>	172
Mid-Quaternary in North America <i>C E Jennings, J S Aber, G Balco, R Barendregt, P R Bierman, C W Rovey II, M Roy, L H Thorleifson, and J A Mason</i>	180
Middle Pleistocene Glaciations in the Southern Hemisphere <i>A Coronato and J Rabassa</i>	187
Late Pleistocene Glacial Events in Beringia <i>S A Elias and J Brigham-Grette</i>	191
Late Quaternary of the Southwest Pacific Region <i>D J A Barrell</i>	202
Late Quaternary of Antarctica <i>Ó Ingólfsson</i>	216
Late Pleistocene in Eurasia <i>J Ehlers, V Astakhov, P L Gibbard, J Mangerud, and J I Svendsen</i>	224
Late Quaternary in Highland Asia <i>L A Owen</i>	236
Late Quaternary in North America <i>J T Andrews and A S Dyke</i>	245
Late Pleistocene in South America <i>A Coronato and J Rabassa</i>	250
Neoglaciation in Europe <i>J A Matthews</i>	257
Neoglaciation in the American Cordilleras <i>S C Porter</i>	269

GLOBAL WARMING *see* Paleoclimate Relevance to Global Warming

H

HEINRICH EVENTS *see* PALEOCLIMATE RECONSTRUCTION: Paleodroughts and Society

HISTORICAL CLIMATE RECORDS *see* PALEOCLIMATE RECONSTRUCTION: Historical Climatology; The Last Millennium

HOLOCENE ENVIRONMENTS *see* Dendroclimatology; ARCHAEOLOGICAL RECORDS: Postglacial Adaptations; BEETLE RECORDS: Postglacial Europe; Postglacial North America; CHIRONOMID RECORDS: Postglacial Europe; Postglacial Southern Hemisphere; DIATOM METHODS: Use in Archaeology; GLACIATIONS: Neoglaciation in Europe; Neoglaciation in the American Cordilleras; ICE CORES: Dynamics of the Greenland Ice Sheet; Dynamics of the West Antarctic Ice Sheet; Dynamics of the East Antarctic Ice Sheet; PALEOCEANOGRAPHY, RECORDS: Postglacial Indian Ocean; Postglacial North Atlantic; Postglacial North Pacific; Postglacial South Pacific; PALEOCLIMATE: Timescales of Climate Change; PALEOCLIMATE MODELING: Paleo-ENSO; PALEOCLIMATE RECONSTRUCTION: Historical Climatology; The Last Millennium; PLANT MACROFOSSIL METHODS AND STUDIES: Treeline Studies; PLANT MACROFOSSIL RECORDS: Holocene North America; POLLEN RECORDS, POSTGLACIAL: Africa; Australia and New Zealand; Northeastern North America; Northwestern North America; Southeastern North America; Southwestern North America; South America; Northern Asia; Northern Europe; Southern Europe

HUMAN EVOLUTION *see* ARCHAEOLOGICAL RECORDS: Overview; 2.7 Myr–300 000 Years Ago in Africa; 2.7 Myr–300 000 Years Ago in Asia; 1.9 Myr–300 000 Years Ago in Europe; Human Evolution in the Quaternary

I

ICE CORE METHODS 277

Overview 278
E J Brook

Biological Material 288
J C Priscu, B C Christner, C M Foreman, and G Royston-Bishop

Borehole Temperature Records 298
K M Cuffey

Chronologies 303
J Schwander

CO₂ Studies 311
T Blunier and T M Jenk

Conductivity Studies 319
R Mulvaney

Glaciochemistry 326
K J Kreutz and B G Koffman

Methane Studies 334
J Chappellaz

Microparticle and Trace Element Studies 342
J R McConnell

Stable Isotopes 347
E J Brook

¹⁰Be and Cosmogenic Radionuclides in Ice Cores 353
R Muscheler

Studies of Firn Air 361
C Buizert

ICE CORE RECORDS 373

Africa 373
L G Thompson and M E Davis

Chinese, Tibetan Mountains 379
C P Wake

South America 387
L G Thompson and M E Davis

Antarctic Stable Isotopes 395
E J Brook

Greenland Stable Isotopes 403
B M Vinther and S J Johnsen

Correlations Between Greenland and Antarctica 410
E J Brook

Ice Margin Sites <i>V V Petrenko</i>	416
Thermal Diffusion Paleotemperature Records <i>A M Grachev</i>	431
ICE CORES	435
History of Research, Greenland and Antarctica <i>M Aydin</i>	435
Dynamics of the Greenland Ice Sheet <i>C S Hvidberg, A Svensson, and S L Buchardt</i>	439
Dynamics of the West Antarctic Ice Sheet <i>R Bindshadler</i>	448
Dynamics of the East Antarctic Ice Sheet <i>E D Waddington and C S Lingle</i>	456
History of Carbon Monoxide and Ultra-Trace Gases from Ice Cores <i>M Aydin</i>	463
History of Nitrous Oxide from Ice Cores <i>A Schilt</i>	471
ICE SHEETS, PLEISTOCENE <i>see</i> GLACIAL LANDFORMS, ICE SHEETS: Growth and Decay; Evidence of Glacier and Ice Sheet Extent; Trimlines and Paleonunataks; GLACIATIONS: Early Quaternary (Pleistocene) and Precursors; Middle Pleistocene in Eurasia; Mid-Quaternary in North America; Middle Pleistocene Glaciations in the Southern Hemisphere; Late Pleistocene Glacial Events in Beringia; Late Quaternary of the Southwest Pacific Region; Late Quaternary of Antarctica; Late Pleistocene in Eurasia; Late Quaternary in Highland Asia; Late Quaternary in North America; Late Pleistocene in South America; ICE CORES: Dynamics of the Greenland Ice Sheet; Dynamics of the West Antarctic Ice Sheet; Dynamics of the East Antarctic Ice Sheet	
ICE WEDGES, ICE WEDGE CASTS <i>see</i> PERMAFROST AND PERIGLACIAL FEATURES: Ice Wedges and Ice-Wedge Casts	
K	
K/Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ Dating <i>J R Wijbrans and K F Kuiper</i>	477
L	
LAKE CHEMISTRY RECONSTRUCTION <i>see</i> PALEOLIMNOLOGY: Lake Chemistry	
LAKE LEVEL STUDIES	483
Overview <i>R T Jones and J T Jordan</i>	483
Africa during the Late Quaternary <i>M E Edwards</i>	499
Asia <i>G Yu, B Xue, and Y Li</i>	506
Australia <i>J Magee</i>	524

Latin America <i>S E Metcalfe</i>	531
North America <i>J R Stone and S C Fritz</i>	537
West-Central Europe <i>M Magny</i>	549
Modeling <i>J Vassiljev</i>	558
Lichenometry <i>D P McCarthy</i>	565

LITHOSTRATIGRAPHY *see* QUATERNARY STRATIGRAPHY: Lithostratigraphy

Loess Deposits: Origins and Properties <i>D R Muhs</i>	573
---	-----

LOESS RECORDS 585

Central Asia <i>A E Dodonov</i>	585
China <i>S C Porter</i>	595
Europe <i>D-D Rousseau, E Derbyshire, P Antoine, and C Hatté</i>	606
North America <i>H M Roberts, D R Muhs, and E A Bettis III</i>	620
South America <i>M A Zárate</i>	629

LUMINESCENCE DATING 643

Thermoluminescence <i>O B Lian</i>	643
Optical Dating <i>O B Lian</i>	653
Electron Spin Resonance Dating <i>A J T Jull</i>	667

M

MAGNETIC POLARITY STUDIES *see* Geomagnetic Excursions and Secular Variations

MAMMALIAN EVOLUTION *see* Vertebrate Overview; VERTEBRATE RECORDS: Early Pleistocene; Mid-Pleistocene of Africa; Mid-Pleistocene of Southern Asia; VERTEBRATE STUDIES: Ancient DNA; Speciation and Evolutionary Trends in Quaternary Vertebrates

MARINE ISOTOPE STAGES *see* PALEOCEANOGRAPHY, PHYSICAL AND CHEMICAL PROXIES: Oxygen-Isotope Stratigraphy of the Oceans

MEGAFAUNA, PLEISTOCENE *see* Vertebrate Overview; VERTEBRATE RECORDS: Mid-Pleistocene of Australia; Mid-Pleistocene of North America; Early Pleistocene; Late Pleistocene of Africa;

Late Pleistocene of North America; Late Pleistocene of South America; Late Pleistocene of Southeast Asia; Late Pleistocene Megafaunal Extinctions; Mid-Pleistocene of Africa; Mid-Pleistocene of Europe; Mid-Pleistocene of Southern Asia; VERTEBRATE STUDIES: Interactions with Hominids

MEGAFAUNAL EXTINCTION *see* VERTEBRATE RECORDS: Late Pleistocene Megafaunal Extinctions; VERTEBRATE STUDIES: Interactions with Hominids

METHANE STUDIES, ICE CORES *see* ICE CORE METHODS: Methane Studies

Mg/Ca AND Sr/Ca STUDIES *see* PALEOCEANOGRAPHY, PHYSICAL AND CHEMICAL PROXIES: Mg/Ca and Sr/Ca Paleothermometry from Calcareous Marine Fossils

MICROMORPHOLOGY OF SEDIMENTS *see* GLACIAL LANDFORMS, SEDIMENTS: Micromorphology of Glacial Sediments; PALEOSOLS AND WIND-BLOWN SEDIMENTS: Soil Micromorphology

MIDGES *see* CHIRONOMID RECORDS: Africa; Chironomid Overview; Late Pleistocene of Europe; Postglacial Europe; Postglacial Southern Hemisphere

MILANKOVITCH THEORY *see* GLACIATION, CAUSES: Astronomical Theory of Paleoclimates

MOLLUSKS, MARINE *see* PALEOCEANOGRAPHY, BIOLOGICAL PROXIES: Corals, Sclerosponges and Mollusks

MORAINES, GLACIAL *see* GLACIAL LANDFORMS: Moraine Forms and Genesis; Evidence of Glacier Recession; Glacial Landsystems; Glacitectonic Structures and Landforms; GLACIAL LANDFORMS, SEDIMENTS: Glacial Erratics and Till Dispersal Indicators

MORPHOSTRATIGRAPHY *see* QUATERNARY STRATIGRAPHY: Morphostratigraphy/Allostratigraphy

N

NEANDERTHAL DEMISE *see* ARCHAEOLOGICAL RECORDS: Neanderthal Demise

NEOGLACIATION *see* GLACIATIONS: Neoglaciation in Europe; Neoglaciation in the American Cordilleras

O

OPTICALLY-STIMULATED LUMINESCENCE DATING *see* LUMINESCENCE DATING: Optical Dating

Oribatid Mites

J M Erickson and R B Platt Jr.

680

OXYGEN ISOTOPE STAGES *see* PALEOCEANOGRAPHY, PHYSICAL AND CHEMICAL PROXIES: Oxygen-Isotope Stratigraphy of the Oceans

P

PACKRAT MIDDENS *see* PLANT MACROFOSSIL METHODS AND STUDIES: Rodent Middens

PALEOANTHROPOLOGY *see* ARCHAEOLOGICAL RECORDS: Overview; 2.7 Myr–300 000 Years Ago in Africa; 2.7 Myr–300 000 Years Ago in Asia; 1.9 Myr–300 000 Years Ago in Europe; Global Expansion 300 000–8000 Years Ago, Africa; Global Expansion 300 000–8000 Years Ago, Asia; Global Expansion 300 000–8000 Years Ago, Australia; Global Expansion 300 000–8000 Years Ago, Americas; Human Evolution in the Quaternary; Neanderthal Demise; Postglacial Adaptations

PALEOBOTANY

699

Overview of Terrestrial Pollen Data

699

R H W Bradshaw

Ancient Plant DNA

705

N Wales, R Allaby, E Willerslev, and M T P Gilbert

Charred Particle Analyses <i>K J Brown and M J Power</i>	716
Paleophytogeography <i>A E Bjune</i>	730
Silicon Isotopes in Diatoms <i>J J Tyler</i>	734
PALEOCEANOGRAPHY	745
Paleoceanography An Overview <i>D M Anderson and K E Lee</i>	745
PALEOCEANOGRAPHY, BIOLOGICAL PROXIES	755
Alkenone Paleothermometry Based on the Haptophyte Algae <i>S L Ho, B D A Naafs, and F Lamy</i>	755
Benthic Foraminifera <i>R Saraswat and R Nigam</i>	765
Biomarker Indicators of Past Climate <i>J P Sachs, K Pahnke, R Smittenberg, and Z Zhang</i>	775
Coccolithophores <i>J-A Flores and F J Sierro</i>	783
Corals, Sclerosponges and Mollusks <i>T M Quinn and B R Schöne</i>	795
Dinoflagellates <i>A de Vernal, A Rochon, and T Radi</i>	800
Marine Diatoms <i>F Abrantes and I M Gil</i>	816
Planktic Foraminifera <i>H J Dowsett and M M Robinson</i>	825
Radiolarians and Silicoflagellates <i>D Lazarus</i>	830
Temperature Proxies, Census Counts <i>J D Ortiz</i>	841
PALEOCEANOGRAPHY, PHYSICAL AND CHEMICAL PROXIES	849
Carbon Cycle Proxies ($\delta^{11}\text{B}$, $\delta^{13}\text{C}_{\text{calcite}}$, $\delta^{13}\text{C}_{\text{organic}}$, Shell Weights, B/Ca, U/Ca, Zn/Ca, Ba/Ca) <i>B Hönisch and K A Allen</i>	849
Dissolution of Deep-Sea Carbonates <i>S Barker</i>	859
Mg/Ca and Sr/Ca Paleothermometry from Calcareous Marine Fossils <i>Y Rosenthal and B Linsley</i>	871
Magnetic Proxies and Susceptibility <i>R G Hatfield and J S Stoner</i>	884
Nutrient Proxies <i>T M Marchitto</i>	899

Oxygen-Isotope Stratigraphy of the Oceans <i>F C Bassinot</i>	907
Oxygen Isotope Composition of Seawater <i>E J Rohling</i>	915
Radioisotope Proxies <i>R Francois</i>	923
Salinity Proxies $\delta^{18}\text{O}$ <i>P D Naidu</i>	932
Terrigenous Sediments <i>A M Franzese and S R Hemming</i>	941

VOLUME 3

PALEOCEANOGRAPHY, RECORDS	1
Early Pleistocene <i>T de Garidel-Thoron</i>	1
Late Pleistocene North Atlantic <i>D M Anderson</i>	9
Late Pleistocene South Atlantic <i>S Mulitza, A Paul, and G Wefer</i>	18
Late Pleistocene North Pacific <i>T Kiefer</i>	33
Postglacial Indian Ocean <i>P D Naidu</i>	46
Postglacial North Atlantic <i>M W Kerwin and K A Huguen</i>	55
Postglacial North Pacific <i>J I Martínez</i>	62
Postglacial South Pacific <i>F Lamy and R de Pol-Holz</i>	73
PALEOCLIMATE	87
Introduction <i>C J Mock</i>	87
Timescales of Climate Change <i>P J Bartlein</i>	93
Modern Analog Approaches in Paleoclimatology <i>C J Mock and J J Shinker</i>	102
Paleoclimate History of the Arctic <i>G H Miller, J Brigham-Grette, R B Alley, L Anderson, H A Bauch, M S V Douglas, M E Edwards, S A Elias, B P Finney, J J Fitzpatrick, S V Funder, A Geirsdóttir, T D Herbert, L D Hinzman, D S Kaufman, G M MacDonald, L Polyak, A Robock, M C Serreze, J P Smol, R Spielhagen, J W C White, A P Wolfe, and E W Wolff</i>	113
The Younger Dryas Climate Event <i>A E Carlson</i>	126

PALEOCLIMATE MODELING	135
Data–Model Comparisons <i>S P Harrison</i>	135
Quaternary Environments <i>C S Jackson</i>	147
The Last Interglacial <i>M Montoya</i>	155
Last Glacial Maximum GCMs <i>A J Broccoli</i>	165
Paleo-ENSO <i>C J Mock</i>	171
 PALEOCLIMATE RECONSTRUCTION	 179
Approaches <i>B N Shuman</i>	179
Pliocene Environments <i>R Z Poore</i>	185
Paleodroughts and Society <i>C J Mock</i>	194
Sub-Milankovitch (DO/Heinrich) Events <i>L Labeyrie, L Skinner, and E Cortijo</i>	200
Paleotempestology <i>K-b Liu</i>	209
Younger Dryas Oscillation, Global Evidence <i>S Björck</i>	222
The Last Millennium <i>M E Mann</i>	229
Historical Climatology <i>M Chenoweth</i>	237
 Paleoclimate Relevance to Global Warming <i>S H Schneider and M D Mastrandrea</i>	 244
Paleohydrology <i>V R Thorndycraft</i>	253
 PALEOLIMNOLOGY	 259
Overview of Paleolimnology <i>M S V Douglas</i>	259
Cladocera <i>M Rautio and L Nevalainen</i>	271
Freshwater Mollusks <i>C G De Francesco</i>	281
Lake Chemistry <i>D Antoniadis</i>	292

Physical Properties of Lake Sediments <i>K R Hodder and R Gilbert</i>	300
Contributions of Paleolimnological Research to Biogeography <i>K A Moser</i>	313
Pigment Studies <i>S McGowan</i>	326
Multiproxy Approaches <i>N Michelutti and J P Smol</i>	339
Visible and Infrared Spectroscopical Applications <i>P Rosén and H Vogel</i>	349
PALEOLITHIC <i>see</i> ARCHAEOLOGICAL RECORDS: Overview; 2.7 Myr–300 000 Years Ago in Africa; 2.7 Myr–300 000 Years Ago in Asia; 1.9 Myr–300 000 Years Ago in Europe; Global Expansion 300 000–8000 Years Ago, Africa; Global Expansion 300 000–8000 Years Ago, Asia; Global Expansion 300 000–8000 Years Ago, Australia; Global Expansion 300 000–8000 Years Ago, Americas; Neanderthal Demise	
PALEOSOLS AND WIND-BLOWN SEDIMENTS	357
Overview <i>D R Muhs</i>	357
Nature of Paleosols <i>J A Mason and P M Jacobs</i>	367
Mineral Magnetic Analysis <i>M J Singer and K L Verosub</i>	375
Soil Micromorphology <i>R A Kemp</i>	381
Weathering Profiles <i>E A Bettis III</i>	392
Soil Morphology in Quaternary Studies <i>L McFadden</i>	402
Biogeochemical Role of Dust in Quaternary Climate Cycles <i>K E Kohfeld</i>	412
PATTERNED GROUND <i>see</i> PERMAFROST AND PERIGLACIAL FEATURES: Patterned Ground	
PEAT STUDIES <i>see</i> PLANT MACROFOSSIL METHODS AND STUDIES: Mire and Peat Macros	
PEDOSTRATIGRAPHY <i>see</i> QUATERNARY STRATIGRAPHY: Pedostratigraphy	
PERMAFROST AND PERIGLACIAL FEATURES	421
Active Layer Processes <i>N I Shiklomanov and F E Nelson</i>	421
Cryoturbation Structures <i>J Vandenberghe</i>	430
Ice Wedges and Ice-Wedge Casts <i>J Murton</i>	436
Patterned Ground <i>C K Ballantyne</i>	452

Permafrost <i>C R Burn</i>	464
Frost Mounds: Active and Relict Forms <i>N Ross</i>	472
Slope Deposits and Forms <i>C Harris</i>	481
Periglacial Fluvial Sediments and Forms <i>J van Huissteden, J Vandenberghe, P L Gibbard, and J Lewin</i>	490
Rock Weathering <i>J Murton</i>	500
Permafrost and Glacier Interactions <i>R I Waller</i>	507
Block/Rock Streams <i>P Wilson</i>	514
Blockfields (Felsenmeer) <i>B R Rea</i>	523
Rock Glaciers and Protalus Forms <i>A Kääb</i>	535
Yedoma: Late Pleistocene Ice-Rich Syngenetic Permafrost of Beringia <i>L Schirrmeister, D Froese, V Tumskoy, G Grosse, and S Wetterich</i>	542
Paraglacial Geomorphology <i>C K Ballantyne</i>	553
Talus Slopes <i>B H Luckman</i>	566
Thermokarst Topography <i>C R Burn</i>	574
PERMAFROST HISTORY <i>see</i> PERMAFROST AND PERIGLACIAL FEATURES: Active Layer Processes; Cryoturbation Structures; Patterned Ground; Permafrost	
Phytoliths <i>M S Blinnikov</i>	582
PIGMENTS, FOSSIL <i>see</i> PALEOLIMNOLOGY: Pigment Studies	
PINGOS <i>see</i> PERMAFROST AND PERIGLACIAL FEATURES: Frost Mounds: Active and Relict Forms	
Plant Macrofossil Introduction <i>H H Birks</i>	593
PLANT MACROFOSSIL METHODS AND STUDIES	613
CO ₂ Reconstruction from Fossil Leaves <i>M Rundgren</i>	613
Megafossils <i>G M MacDonald</i>	621
Dendroarchaeology <i>R H Towner</i>	630

Mire and Peat Macros <i>D Mauquoy and B van Geel</i>	637
Paleolimnological Applications <i>M-J Gaillard and H H Birks</i>	657
Rodent Middens <i>S A Elias</i>	674
Surface Samples, Taphonomy, Representation <i>A C Dieffenbacher-Krall</i>	684
Treeline Studies <i>W Tinner</i>	690
Use in Environmental Archaeology <i>S Jacomet</i>	699
Validation of Pollen Studies <i>S T Jackson and R K Booth</i>	725
PLANT MACROFOSSIL RECORDS	733
Arctic Eurasia <i>F Kienast</i>	733
Arctic North America <i>N H Bigelow, G D Zazula, and D E Atkinson</i>	746
Greenland <i>O Bennike</i>	760
Holocene North America <i>R G Baker</i>	768
Late Glacial Multidisciplinary Studies <i>B Ammann, H H Birks, A Walanus, and K Wasylkowa</i>	785
PLATE TECTONICS <i>see</i> GLACIAL LANDFORMS: Glacitectonic Structures and Landforms; GLACIATION, CAUSES: Tectonic Uplift and Continental Configurations; GLACIATIONS: Late Quaternary in North America	
PLIOCENE ENVIRONMENTS <i>see</i> PALEOCLIMATE RECONSTRUCTION: Pliocene Environments	
Pollen Analysis, Principles <i>H Seppä</i>	794
POLLEN METHODS AND STUDIES	805
Use of Pollen as Climate Proxies <i>S Brewer, J Guiot, and D Barboni</i>	805
Reconstructing Past Biodiversity Development <i>B V Odgaard</i>	816
Numerical Analysis Methods <i>H J B Birks</i>	821
Databases and Their Application <i>E C Grimm, R H W Bradshaw, S Brewer, S Flantua, T Giesecke, A-M Lézine, H Takahara, and J W Williams</i>	831
Surface Samples and Trapping <i>A Poska</i>	839

Stand-Scale Palynology <i>R H W Bradshaw</i>	846
Changing Plant Distributions and Abundances <i>T Giesecke</i>	854
The Biome Approach to Reconstructing Past Vegetation <i>M E Edwards</i>	861
POLLSCAPE Model: Simulation Approach for Pollen Representation of Vegetation and Land Cover <i>S Sugita</i>	871
Archaeological Applications <i>M-J Gaillard</i>	880

VOLUME 4

Pollen Records, Last Interglacial of Europe <i>C Tzedakis</i>	1
--	---

POLLEN RECORDS, LATE PLEISTOCENE 9

Africa <i>M E Meadows and B M Chase</i>	9
Australasia <i>P Kershaw and S van der Kaars</i>	18
Northern Asia <i>A V Lozhkin and P M Anderson</i>	27
Northern North America <i>N H Bigelow</i>	39
South America <i>H Hooghiemstra and J C Berrio</i>	52
Middle and Late Pleistocene in Southern Europe <i>J-L de Beaulieu, P C Tzedakis, V Andrieu-Ponel, and F Guiter</i>	63
Western North America <i>R S Thompson</i>	72

POLLEN RECORDS, POSTGLACIAL 85

Africa <i>A-M Lézine</i>	85
Australia and New Zealand <i>J R Dodson</i>	104
Northeastern North America <i>J W Williams and B N Shuman</i>	115
Northwestern North America <i>D G Gavin and F S Hu</i>	124
Southeastern North America <i>D A Willard</i>	133
Southwestern North America <i>P E Wigand</i>	142

South America	156
<i>H Behling</i>	
Northern Asia	164
<i>A A Andreev and P E Tarasov</i>	
Northern Europe	173
<i>M J Bunting</i>	
Southern Europe	179
<i>L Sadori</i>	

POTASSIUM-ARGON DATING *see* K/Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ Dating

Q

QUATERNARY STRATIGRAPHY 189

Overview	189
<i>B Pillans</i>	
Continental Biostratigraphy	206
<i>T van Kolfschoten</i>	
Chronostratigraphy	215
<i>B Pillans</i>	
Climatostratigraphy	222
<i>P L Gibbard</i>	
Lithostratigraphy	227
<i>W E Westerhoff and H J T Weerts</i>	
Morphostratigraphy/Allostratigraphy	243
<i>P D Hughes</i>	
Pedostratigraphy	250
<i>A Palmer</i>	
Sequence Stratigraphy	260
<i>T R Naish, S T Abbott, and R M Carter</i>	
Tephrochronology	277
<i>B V Alloway, D J Lowe, G Larsen, P A R Shane, and J A Westgate</i>	

R

RADIOLARIAN STUDIES *see* PALEOCEANOGRAPHY, BIOLOGICAL PROXIES: Radiolarians and Silicoflagellates

RADIOCARBON DATING 305

Conventional Method	305
<i>G T Cook and J van der Plicht</i>	
AMS Radiocarbon Dating	316
<i>A J T Jull</i>	
Sources of Error	324
<i>E M Scott</i>	

Variations in Atmospheric ^{14}C <i>J van der Plicht</i>	329
Causes of Temporal ^{14}C Variations <i>G S Burr</i>	336
Calibration of the ^{14}C Record <i>P J Reimer, R W Reimer, and M Blaauw</i>	345
Charcoal <i>M I Bird</i>	353
^{14}C of Plant Macrofossils <i>C Hatté and A J T Jull</i>	361

ROCK GLACIERS *see* PERMAFROST AND PERIGLACIAL FEATURES: Rock Glaciers and Protalus Forms

RODENT MIDDENS *see* PLANT MACROFOSSIL METHODS AND STUDIES: Rodent Middens

S

SEA LEVEL STUDIES 369

Overview <i>I Shennan</i>	369
Geomorphological Indicators <i>P A Pirazzoli</i>	377
Sedimentary Indicators of Relative Sea-Level Changes – High Energy <i>J A G Cooper</i>	385
Sedimentary Indicators of Relative Sea-Level Changes – Low Energy <i>R J Edwards</i>	396
Coral Records of Relative Sea-Level Changes <i>C D Woodroffe</i>	409
Microfossil-Based Reconstructions of Holocene Relative Sea-Level Change <i>W R Gehrels</i>	419
Eustatic Sea-Level Changes – Glacial–Interglacial Cycles <i>C V Murray-Wallace</i>	429
Eustatic Sea-Level Changes Since the Last Glacial Maximum <i>P L Whitehouse and S L Bradley</i>	439
Isostasy: Glaciation-Induced Sea-Level Change <i>G Milne and I Shennan</i>	452
Use of Cave Data in Sea-Level Reconstructions <i>A Dutton</i>	460

SEA-LEVELS, LATE QUATERNARY 467

Late Quaternary Relative Sea-Level Changes in High Latitudes <i>C Ó Cofaigh and M J Bentley</i>	467
Late Quaternary Sea-Level Changes in Greenland <i>S A Woodroffe and A J Long</i>	481
Late Quaternary Relative Sea-Level Changes at Mid-Latitudes <i>A C Kemp, B P Horton, and S E Engelhart</i>	489

Late Quaternary Relative Sea-Level Changes in the Tropics <i>Y Zong</i>	495
Tectonics and Relative Sea-Level Change <i>A R Nelson</i>	503

SEA SURFACE TEMPERATURE RECONSTRUCTION *see* PALEOCEANOGRAPHY: Paleoceanography An Overview; PALEOCEANOGRAPHY, BIOLOGICAL PROXIES: Alkenone Paleothermometry Based on the Haptophyte Algae; Biomarker Indicators of Past Climate; Coccolithophores; Dinoflagellates; Marine Diatoms; Planktic Foraminifera; Temperature Proxies, Census Counts; PALEOCEANOGRAPHY, PHYSICAL AND CHEMICAL PROXIES: Carbon Cycle Proxies ($\delta^{11}\text{B}$, $\delta^{13}\text{C}_{\text{calcite}}$, $\delta^{13}\text{C}_{\text{organic}}$, Shell Weights, B/Ca, U/Ca, Zn/Ca, Ba/Ca); Mg/Ca and Sr/Ca Paleothermometry from Calcareous Marine Fossils; Oxygen-Isotope Stratigraphy of the Oceans; PALEOCEANOGRAPHY, RECORDS: Late Pleistocene North Atlantic; Late Pleistocene North Pacific; Late Pleistocene South Atlantic; Postglacial Indian Ocean; Postglacial North Atlantic; Postglacial North Pacific; Postglacial South Pacific

SILICOFLAGELLATES *see* PALEOCEANOGRAPHY, BIOLOGICAL PROXIES: Radiolarians and Silicoflagellates

SPELEOTHEMS *see* CARBONATE STABLE ISOTOPES: Speleothems

STABLE ISOTOPE STUDIES, CARBONATES *see* CARBONATE STABLE ISOTOPES: Overview; Speleothems; Terrestrial Teeth and Bones; Nonmarine Biogenic Carbonates; Terrestrial Organic Materials; Non-Lacustrine Terrestrial Studies; Lake Sediments

STABLE ISOTOPE STUDIES, DEEP SEA RECORDS *see* PALEOCEANOGRAPHY, PHYSICAL AND CHEMICAL PROXIES: Oxygen-Isotope Stratigraphy of the Oceans; Oxygen Isotope Composition of Seawater; Salinity Proxies $\delta^{18}\text{O}$; PALEOCEANOGRAPHY, RECORDS: Early Pleistocene; Late Pleistocene North Atlantic; Late Pleistocene South Atlantic; Postglacial Indian Ocean; Postglacial North Atlantic; Postglacial North Pacific; Postglacial South Pacific; ICE CORE METHODS: Stable Isotopes; ICE CORE RECORDS: Antarctic Stable Isotopes; Greenland Stable Isotopes; Correlations Between Greenland and Antarctica

STABLE ISOTOPE STUDIES, ICE CORES *see* ICE CORE METHODS: Stable Isotopes; ICE CORE RECORDS: Antarctic Stable Isotopes; Greenland Stable Isotopes; Correlations Between Greenland and Antarctica

SUB-MILANKOVITCH EVENTS *see* PALEOCLIMATE RECONSTRUCTION: Sub-Milankovitch (DO/Heinrich) Events

T

TALUS SLOPES *see* PERMAFROST AND PERIGLACIAL FEATURES: Talus Slopes

TEETH AND BONES, STABLE ISOTOPE STUDIES *see* CARBONATE STABLE ISOTOPES: Terrestrial Teeth and Bones

TEPHROCHRONOLOGY *see* QUATERNARY STRATIGRAPHY: Tephrochronology

THERMOHALINE CIRCULATION OF THE OCEANS *see* GLACIAL CLIMATES: Thermohaline Circulation

THERMOLUMINESCENCE DATING *see* LUMINESCENCE DATING: Thermoluminescence

TILL, GLACIAL *see* GLACIAL LANDFORMS, SEDIMENTS: Till Fabric Analysis; Tills; Glacial Erratics and Till Dispersal Indicators

TREE RINGS *see* Dendrochronology; Dendroclimatology; GLACIAL LANDFORMS, TREE RINGS: Dendrogeomorphology; Dendroglaciology; PLANT MACROFOSSIL METHODS AND STUDIES: Dendroarchaeology

TREELINE RECONSTRUCTION *see* POLLEN METHODS AND STUDIES: Archaeological Applications

U**USE OF QUATERNARY PROXIES IN FORENSIC SCIENCE 522**

Use of Geology in Forensic Science: Search to Locate Burials 522
L Donnelly

Soils and Sediment 535
R C Murray

The Use of Macroscopic Plant Remains in Forensic Science 542
J H Bock

Insects 548
D E Gennard

Analytical Techniques in Forensic Palynology 556
V M Bryant

U-Series Dating 567
W G Thompson

V

Varved Lake Sediments 573
B Zolitschka

Varved Marine Sediments 582
K A Hughen and B Zolitschka

Vertebrate Overview 590
D C Schreve

VERTEBRATE RECORDS 599

Early Pleistocene 599
L Rook, M Delfino, M P Ferretti, and L Abbazzi

Early and Middle Pleistocene of Northern Eurasia 605
I Vislobokova and A Tesakov

Mid-Pleistocene of Africa 615
L C Bishop and A Turner

Mid-Pleistocene of Australia 621
G J Prideaux

Mid-Pleistocene of Europe 639
R Sardella

Mid-Pleistocene of North America 646
C J Bell

Mid-Pleistocene of Southern Asia 651
J de Vos

Late Pleistocene of Africa 664
T E Steele

Late Pleistocene of North America 673
J I Mead

Late Pleistocene of South America <i>M Ubilla</i>	680
Late Pleistocene of Southeast Asia <i>Y Chaimanee</i>	693
Late Pleistocene Megafaunal Extinctions <i>S A Elias and D C Schreve</i>	700
Late Pleistocene Mummified Mammals <i>C R Harington</i>	713
VERTEBRATE STUDIES	719
Ancient DNA <i>I Barnes, R Barnett, and B Shapiro</i>	719
Speciation and Evolutionary Trends in Quaternary Vertebrates <i>A M Lister</i>	723
Dwarfing and Gigantism in Quaternary Vertebrates <i>M R Palombo and R Rozzi</i>	733
Interactions with Hominids <i>K V Boyle</i>	748
VOLCANIC ASH <i>see</i> QUATERNARY STRATIGRAPHY: Tephrochronology	
VULCANISM <i>see</i> GLACIAL CLIMATES: Volcanic and Solar Forcing; GLACIAL LANDFORMS: Quaternary Vulcanism: Subglacial Landforms; QUATERNARY STRATIGRAPHY: Tephrochronology	
W	
WIND-BLOWN SEDIMENTS (LOESS, SAND DUNES) <i>see</i> Eolian Records, Deep-Sea Sediments; Loess Deposits: Origins and Properties; DUNE FIELDS: High Latitudes; Mid-Latitudes; Low Latitudes; LOESS RECORDS: Central Asia; China; Europe; North America; South America	
Y	
YOUNGER DRYAS OSCILLATION <i>see</i> PALEOCLIMATE RECONSTRUCTION: Younger Dryas Oscillation, Global Evidence	
<i>Index</i>	757

GLACIAL LANDFORMS, SEDIMENTS

Contents

Clast Form Analysis

Glaciofluvial Landforms of Deposition

Glaciogenic Lithofacies

Glaciomarine Sediments and Ice-Rafted Debris

Glaciolacustrine

Micromorphology of Glacial Sediments

Tills

Till Fabric Analysis

Glacial Erratics and Till Dispersal Indicators

Glacial Sequence Stratigraphy

Clast Form Analysis

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Introduction

Different processes of erosion, weathering, and transport leave distinctive morphological signatures on sedimentary particles. By associating particular morphologies with particular processes, clast form analysis can be used to reconstruct aspects of debris history, and to shed light on the significance and context of sediment facies. Clast form analysis has been part of the standard toolkit of sedimentologists for many decades, and a broad range of techniques and approaches has been developed. As a result, the literature is large, diverse, and often confusing. An excellent review of the early literature on clast form analysis was provided by [Barrett \(1980\)](#), and a detailed description of field techniques and analytical methods has been presented by [Benn \(2004\)](#). Clast form analysis refers to the study of pebble-sized or larger particles. Analysis of the form of sand-sized or smaller particles is known as micromorphological analysis.

Defining Particle Form

Like every snowflake, every sedimentary particle has a unique form and a unique story to tell. To extract the maximum amount of useful information from this boundless variety, it is useful to think of clast form in terms of a simple hierarchy of measurable characteristics ([Barrett, 1980](#)). Shape, the first-order property, refers to the overall proportions of the particle; roundness, the second-order property, describes the degree of rounding of the particle outline; and texture, the third-order property, concerns its surface characteristics. Conceptually,

these are quite separate properties, although in practice there tend to be consistent associations of shape, roundness, and textural characteristics in clasts with similar histories.

Shape

Shape is defined in terms of the relative dimensions of the long-, intermediate-, and short-particle axes (a , b , c , or L , I , S , respectively). This definition of shape reduces all the craggy detail of individual clasts to simple cuboids, which can be visualized as the smallest box that the particle can be fitted into. All possible shapes form a continuum between three end members: (i) cubic, or equant particles with all three axes of equal length ($a=b=c$); (ii) slabby, or platy particles with the a - and b -axes equal and a very small c -axis ($a=b \gg c$); and (iii) rod-shaped particles with the a -axis much longer than the other two ($a \gg b=c$). This continuum is illustrated in [Figure 1](#), in which all possible cuboidal shapes are represented on an isometric triangular diagram known as the general shape triangle ([Benn and Ballantyne, 1993](#); [Sneed and Folk, 1958](#)). The ratios between the short and long axes ($c:a$) and the intermediate and long axes ($b:a$) form the scales on two sides of the triangle. These two ratios are enough to uniquely define the shape of any cuboid, although other indices of particle shape can be used as additional scales if desired (e.g., [Benn and Ballantyne, 1993](#)). Examples of other potentially useful indices include the disk-rod index (DRI: $(a-b)/(a-c)$), which distinguishes disk- and rod-shaped clasts; and maximum projection sphericity (MPS: $3\sqrt{(c^2/ab)}$), which reflects the balance of drag and gravitational forces acting on a particle immersed

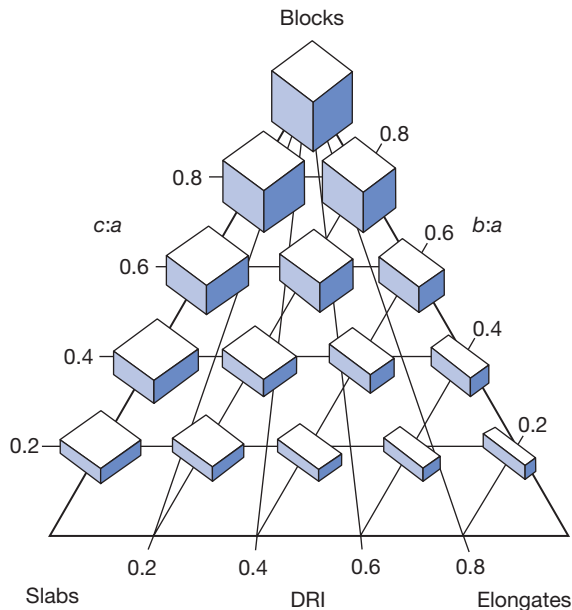


Figure 1 General shape triangle, showing the three-way continuum of clast shapes. Adapted from [Sneed ED and Folk RL \(1958\)](#) Pebbles in the lower Colorado River, Texas, a study in clast morphogenesis. *Journal of Geology* 66: 114–150.

in a fluid. MPS is a good index of the behavior of clasts transported by or settling in, water but is of limited usefulness outside this hydraulic context. A useful spreadsheet program for plotting particle shape data on triangular diagrams has been created by [Graham and Midgley \(2001\)](#).

Roundness

Roundness can be assessed using a variety of approaches. Several quantitative methods have been proposed based on the radius of curvature of particle edges. For example, the Wentworth–Cailleux index ($2000r/a$), compares the radius of curvature of the sharpest edge of the clast (r) with its longest dimension (a). Values of r can be determined from templates, although in practice radii are very difficult to determine for clasts with sharp, unworn edges. Another problem with this approach is that the sharpest edge may not be representative of the roundness of the particle as a whole (as is likely to be the case for rounded clasts that have been broken). For this reason, most researchers use semiquantitative methods, which assign particles to roundness categories using comparison charts or descriptive criteria. [Figure 2](#) shows a widely used comparison chart, in which roundness categories are represented by sets of images. An alternative approach is to employ descriptive criteria as the basis for assigning clasts to a roundness category ([Table 1](#)). Such criteria provide an objective basis for classification, which can be quickly and easily applied in the field. However, as with all roundness classification schemes, it attempts to divide continuous variation into mutually exclusive categories, and different operators may put the class boundaries in different places. Care must be taken to ensure consistency and minimize operator bias, perhaps by keeping a collection of ‘reference clasts’ for comparison.

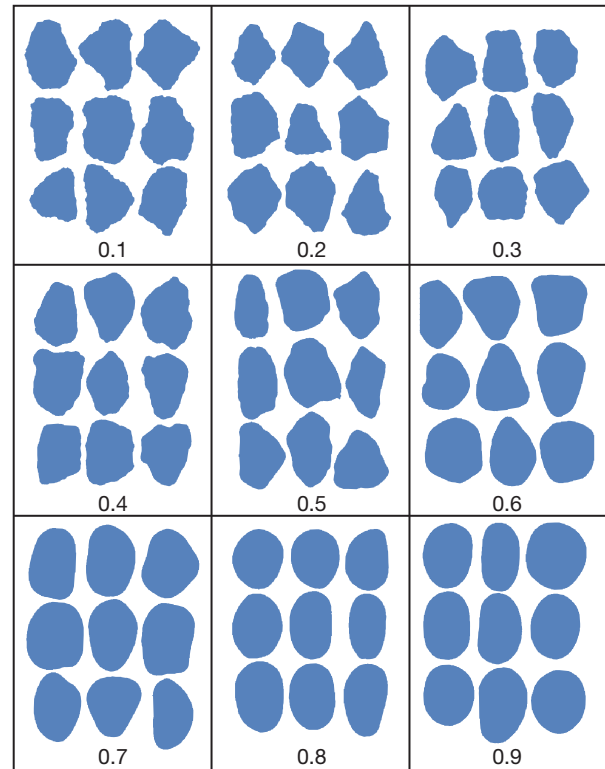


Figure 2 Comparison chart showing nine numerical categories of particle roundness. Adapted from [Krumbein WC \(1941\)](#) Measurement and geological significance of shape and roundness of sedimentary particles. *Journal of Sedimentary Petrology* 11: 64–72.

Texture

Surface texture is an underused source of information, and few systematic studies have been conducted to date. However, certain textures (such as striated surfaces) provide unambiguous records of particular transport and weathering processes, and thus amply repay the investment of time spent recording them. A simple but effective approach is to assign clasts to nominal categories and count frequencies of occurrence. Such categories could include: fresh, unworn fracture surfaces; rough, weathered surfaces; polish; striations; crescentic or conchoidal fractures on polished pebbles, which may indicate high impact forces in shoreface environments; and ventifacts, recording polishing by wind-blown sand.

This list of categories is not exhaustive, and variations can be devised depending on the purpose of the investigation.

Field Measurements

Clast form measurements are usually made in the field, using samples of clasts taken from individual landforms or sediment facies. Normally, each sample comprises 50 clasts, although some investigators take larger or smaller samples. Objective interpretation of sample characteristics can only be accomplished with reference to deposits of known origin, and sampling schemes should therefore include control samples

Table 1 Descriptive clast roundness categories

<i>Class</i>	<i>Description</i>
Very angular (VA)	Edges and faces unworn; sharp, delicate protuberances
Angular (A)	Edges and faces unworn
Subangular (SA)	Faces unworn, edges worn
Subrounded (SR)	Edges and faces worn but clearly distinguishable
Rounded (R)	Edges and faces worn and barely distinguishable
Well rounded (WR)	No edges or faces distinguishable

representing a range of process domains. Well-chosen control samples provide unambiguous baselines against which other samples can be compared, so research design should be informed by a thorough understanding of the system under investigation. For example, where the transport history of glaciogenic sediments is of interest, control samples might include supraglacial debris, basal debris-rich ice, sediments in glaciofluvial transport, and so on, encompassing the range of possible transport paths taken by debris in the catchment. Where the relevant processes are no longer active, such as in deglaciated regions, sediment facies of known origin can be used as proxies. For example, glaciofluvial bedload may be represented by river gravels, supraglacial debris by scree, basal ice by till, and so on. Any errors in the interpretation of such proxies, however, will be propagated throughout the investigation, so great care must be taken with this fundamental aspect of research design. Pilot studies may help to optimize sampling strategies. A range of summary statistics can be used to characterize data distributions and compare groups of samples. These include (i) sample means and standard deviations; (ii) medians and interquartile ranges; and (iii) percentages of data values above or below some reference point. Where particle shape and/or roundness data are not normally distributed, means and standard deviations are inappropriate measures of central tendency and dispersion, so nonparametric statistics ((ii) and (iii)) provide the most flexible means of summarizing and comparing samples. The covariance of two clast form indices provides a more powerful means of characterizing the aggregate particle form characteristics of samples than single indices used in isolation (Benn and Ballantyne, 1994).

Influences on Particle Form

Lithology

Clast form characteristics interpreted with reference to control samples can be used to reconstruct weathering histories, transport paths, and certain depositional and postdepositional processes. Clast form, however, is also strongly influenced by parent lithology. The spacing and distribution of bedding planes, joints, or other discontinuities predetermine lines of fracture and, therefore, influence particle shape. Thin-bedded sandstones and shales, for example, tend to form platy clasts, whereas massive crystalline rocks such as granites, gneiss, and gabbro tend to produce more equant particles. Additionally, rock hardness affects the susceptibility of particles to rounding processes. It is necessary, therefore, to control for lithology

**Figure 3** Striated and faceted clasts from subglacial till.

when sampling, either by limiting measurements to single lithologies if transport processes are of interest, or limiting measurements to single process domains if the influence of lithology is being investigated.

Weathering

Weathering processes can affect particle form in several ways. Macrogelivation (frost shattering) tends to produce angular and very angular clasts, whereas microgelivation (granular disintegration) tends to increase particle roundness by preferentially weathering particle edges. Macrogelivation also affects surface texture by exposing fresh, unweathered surfaces and affects shape by changing the relative dimensions of particle axes, whereas microgelivation produces rough, weathered surfaces but tends to leave overall shape unchanged. Radius of curvature of the edges of granite boulders can be used as a proxy for time elapsed since deposition, on the assumption that radius increases as microgelivation proceeds.

Transport

Transport processes can leave a variety of imprints on sedimentary particles. Some high-energy processes, such as rockfall and avalanching, cause clast fracturing, altering shape characteristics and increasing angularity. Clasts transported by snow avalanches tend to have particularly high angularity due to very large interparticle impact forces. Conversely, abrasion in fluvial and coastal environments increases particle roundness, ultimately producing well-rounded, egg-shaped clasts. Transport in subglacial shear zones at or close to the ice–bed interface produces very distinctive clast morphologies by a combination of fracture and abrasion. Asymmetric application of stresses in subglacial shear zones means that abrasion is focused on the top and upglacier-facing (stoss) surfaces, whereas fracturing preferentially occurs on downglacier-facing (lee) surfaces. This produces characteristic asymmetric clast forms known as ‘flat-irons,’ ‘stoss–lee forms,’ or ‘bullet-shaped clasts’ (Sharp, 1982; Figure 3). Krüger (1984) has described ‘double stoss–lee forms,’ which he attributed to a two-stage process of ploughing followed by lodgement. The asymmetry of clasts shaped during subglacial transport and deposition contrasts with the symmetrical or isotropic wear patterns typical of clasts modified by fluvial and shoreface processes.

Some low-energy processes, such as hillslope creep, debris flow, and englacial or supraglacial transport, may have little or no effect on particle shape, and clasts transported by such

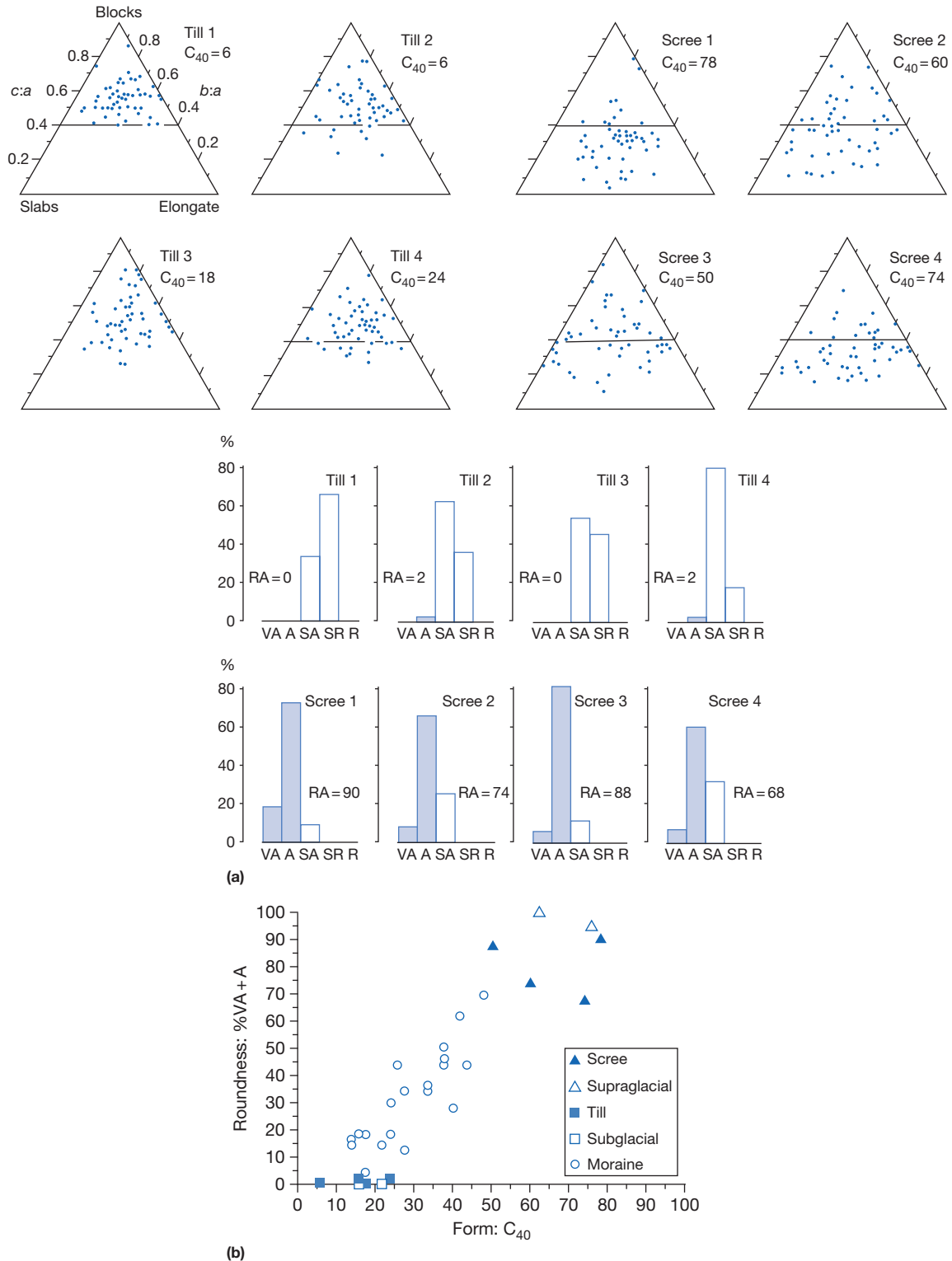


Figure 4 (a) Shape (triangular diagrams) and roundness data (histograms) for till and scree deposits in Jotunheimen, Norway. The clast shape index (C_{40} ; % of clasts with $c:a=0.4$) is high for scree (numerous slabby and elongate shapes) and low for till clasts (numerous blocky shapes); the roundness index (RA; % of angular and very angular clasts) is high for scree and low for till samples. (b) Covariance of C_{40} and RA for control and moraine samples. The moraine samples form a band between the supraglacial and till controls, suggesting that they comprise mixtures of debris from active and passive transport paths. Adapted from [Benn DI and Ballantyne CK \(1994\)](#) Reconstructing the transport history of glacial sediments: A new approach based on the co-variance of clast form indices. *Sedimentary Geology* 91: 215–227.

processes tend to retain characteristics developed during earlier events. Evidence for a lack of modification can, therefore, provide as much information as evidence for particular processes. For example, Boulton (1978) distinguished two contrasting modes of glacial transport: active transport at or close to the bed, and passive transport at higher levels in or on the ice. As already noted, active transport results in clast fracture and abrasion, whereas passively transported debris tends to have shape, roundness, and textural properties inherited from earlier phases of weathering and transport, such as macrogelivation and avalanching. Benn and Ballantyne (1994) found that actively and passively transported clasts can be differentiated using the covariance of their shape and roundness characteristics, providing an efficient means of identifying their relative abundances in glacial deposits and thus the relative importance of different transport modes in former glacier systems (Figure 4). Recent studies have shown that glacial transport systems are more complex than envisaged in Boulton's original classification scheme. First, the associations between supraglacial and passive transport, on the one hand, and subglacial and active transport on the other, does not apply in all cases. On debris-covered glaciers, for example, boulders can grind against each other as they adjust to melting of the underlying ice, and smaller clasts can undergo significant wear during repeated gravitational reworking (Benn and Evans, 1998). Thus, edge-rounded debris does not necessarily imply subglacial transport. Second, the active-passive dichotomy does not encompass all possible modes of transport. For example, Spedding (2000) and Spedding and Evans (2002) used clast form analysis to show that glaciofluvial processes can transport large amounts of sediment through glaciers. The presence of glaciofluvially transported clasts may go undetected if data are analyzed using methods designed to distinguish actively and passively transported debris (see Benn, 2004).

Conclusion

These complications show that uncritical application of clast form analysis can lead to misleading conclusions. With care,

however, subtle differences in clast morphology can be used to reconstruct many aspects of former glacier systems. The key is a flexible approach to research design guided by a thorough knowledge of glacial and extraglacial sediment transport systems. Once attuned to the language of shape, roundness, and texture, you will truly believe a stone can speak.

See also: [Glacial Landforms, Sediments: Glaciogenic Lithofacies; Glaciomarine Sediments and Ice-Rafted Debris; Micromorphology of Glacial Sediments; Tills.](#)

References

- Barrett PJ (1980) The shape of rock particles, a critical review. *Sedimentology* 27: 291–303.
- Benn DI (2004) Clast morphology. In: Evans DJA and Benn DI (eds.) *A Practical Guide to the Study of Glacial Sediments*, pp. 77–92. London: Edward Arnold.
- Benn DI and Ballantyne CK (1993) The description and representation of clast shape. *Earth Surface Processes and Landforms* 18: 665–672.
- Benn DI and Ballantyne CK (1994) Reconstructing the transport history of glaciogenic sediments: A new approach based on the co-variance of clast form indices. *Sedimentary Geology* 91: 215–227.
- Benn DI and Evans DJA (1998) *Glaciers and Glaciation*. London: Edward Arnold.
- Boulton GS (1978) Boulder shapes and grain-size distributions of debris as indicators of transport paths through a glacier and till genesis. *Sedimentology* 25: 773–799.
- Graham DJ and Midgley DG (2001) Graphical representation of clast shape using triangular diagrams: An Excel spreadsheet method. *Earth Surface Processes and Landforms* 25: 1473–1477.
- Krüger J (1984) Clasts with stoss-lee form in lodgement tills: A discussion. *Journal of Glaciology* 30: 241–243.
- Sharp MJ (1982) Modification of clasts in lodgement tills by glacial erosion. *Journal of Glaciology* 28: 475–481.
- Sneed ED and Folk RL (1958) Pebbles in the lower Colorado River, Texas, a study in clast morphogenesis. *Journal of Geology* 66: 114–150.
- Spedding NF (2000) Hydrological controls on sediment transport pathways: Implications for debris-covered glaciers. In: Nakawo M, Fountain A, and Raymond C (eds.) *Debris-Covered Glaciers*, pp. 133–142. Gentbrugge, Belgium: International Association of Hydrological Sciences. Publication No. 264.
- Spedding NF and Evans DJA (2002) Sediments and landforms at Kviarjokull, southeast Iceland: A reappraisal of the glaciated valley landsystem. *Sedimentary Geology* 149: 21–42.

Glaciofluvial Landforms of Deposition

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Introduction: Glaciofluvial Systems

Glaciofluvial landforms are created by glacial meltwater processes, either upon, within, or beneath glaciers and beyond glacial margins. Glaciofluvial landforms can therefore be recognized in both present and past landscapes of alpine and continental glaciation.

Glaciofluvial systems exhibit a distinct magnitude and frequency regime, which contrasts with nonglacially influenced fluvial systems. Flow magnitudes within glaciofluvial systems fluctuate diurnally with snow and ice melt, seasonally with glacial ice ablation and snowmelt, episodically with glacial advances and recessions, and also by the release of stored meltwater. Glacier regime and stability are therefore crucial controlling factors in the glaciofluvial system. Distinction of glaciofluvial systems by climatic zones is perhaps inappropriate, however, because although glaciofluvial landforms are produced at different rates in polar, arctic, and alpine environments, the resultant landform morphology is often similar (Fitzsimons, 2003).

Glacial meltwater can route supraglacially, englacially, and subglacially before exiting from an ice margin into a proglacial zone (Figure 1). Routes depend upon water source, volume and release rate, and any preexisting supraglacial channels, crevasses, englacial passages, or subglacial cavities. Proglacial meltwater tends to follow preexisting topography although frequent channel avulsions occur due to glacier margin fluctuations and internal system adjustments of braided river systems. In the case of an exceptionally high-magnitude glaciofluvial flood, or 'jökulhlaup,' preexisting glacier margin topography and any proglacial river courses can be of little consequence to the route of meltwater (e.g., Russell and Knudsen, 1999). Glaciofluvial landforms can be produced within a range of unusual depositional environments such as glacier ice-walled tunnels or conduits, crevasses or fractures, and other ice-contact positions (Figure 1).

Besides meltwater, sediment flux is very important for determining glaciofluvial landform type. Sediment availability tends to be very high in glacial regions due to high erosion rates, and particularly during episodes of glacial retreat. Glaciofluvial sediment can be derived from valley walls, bedrock, and substrate erosion. Sediment flux affects flow rheology or flow behavior and can modify a flood hydrograph (e.g., Rushmer, 2006). For example, during a high-magnitude flood, sediment supply is likely to become exhausted, resulting in the development of a positive hysteresis loop (e.g., Old and Lawler, 2005) as well as pulses or 'slugs' of sediment-charged meltwater (e.g., Carrivick et al., 2004a,b). High-magnitude glaciofluvial flows can approach a hyperconcentrated state, with a greater tendency for upper flow regime bedform formation.

Owing to the range of scales at which glaciofluvial landforms exist, it is difficult to distinguish glaciofluvial landforms

of deposition from nonglaciofluvial landforms on the basis of morphology alone. It is therefore usually necessary to examine the sedimentary character of glaciofluvial landforms of deposition to identify diagnostic features. For example, ice-proximal glaciofluvial sediment is typically coarse grained, exhibits relatively poor sorting, and features abrupt changes in lithofacies. However, glaciofluvial landforms formed in an ice-contact situation will, with subsequent loss of ice support, feature sedimentary structures that are indicative of postdepositional modification, that is, shearing, faulting, slumping, and subsidence. Ice-contact glaciofluvial landforms may also be compressed and overridden by advancing glacier ice, producing characteristic reverse faulting and folding. Finally, it must also be considered that many glaciofluvial landforms can become reworked or modified by several events and processes. For example, depositional bars can become dissected by subsequent erosion, and bedrock or alluvial channels can become infilled.

Glaciofluvial Landforms of Deposition

Glaciofluvial processes produce a wide range of landforms that are here classified by location of formation, that is, proglacial or ice contact, and geometric scale (Table 1). The geometric scale of glaciofluvial landforms is determined by the size of the channel in which they are situated. This dependence on scale as well as position contrasts with standard fluvial bedform classification schemes (e.g., Leeder, 1999) that imply the action of depositional processes over relatively long periods of time. Proglacial landforms of deposition include outwash plains or 'sandur,' and also a series of bedforms that are found within channels on these sandur. Similar bedforms can also form in proglacial bedrock channels. Ice-contact glaciofluvial landforms of deposition include kames and eskers.

Macroscale glaciofluvial landforms are a cumulative result of numerous depositional events. They span kilometers in planform and include sandur and valley-fill deposits. Meso-scale landforms include bar and bedforms of several hundred meters in plan-form geometry, such as bars and dunes, and tend to be dependent upon reach-scale equilibrium sediment transport. Microscale landforms include ripples; they are generally products of transient hydraulics, and of a planform of the order of one to a few tens of meters.

Proglacial Glaciofluvial Landforms of Deposition

Sandar

The Icelandic term 'sandur' (*singular*) is synonymous with a braided proglacial outwash plain. Southern Iceland is the type-site for extensive outwash plains or 'sandur' (*plural*) such as Skeiðarársandur (Figure 2). Sandur character, in terms of geomorphology and sedimentology, is indicative of meltwater

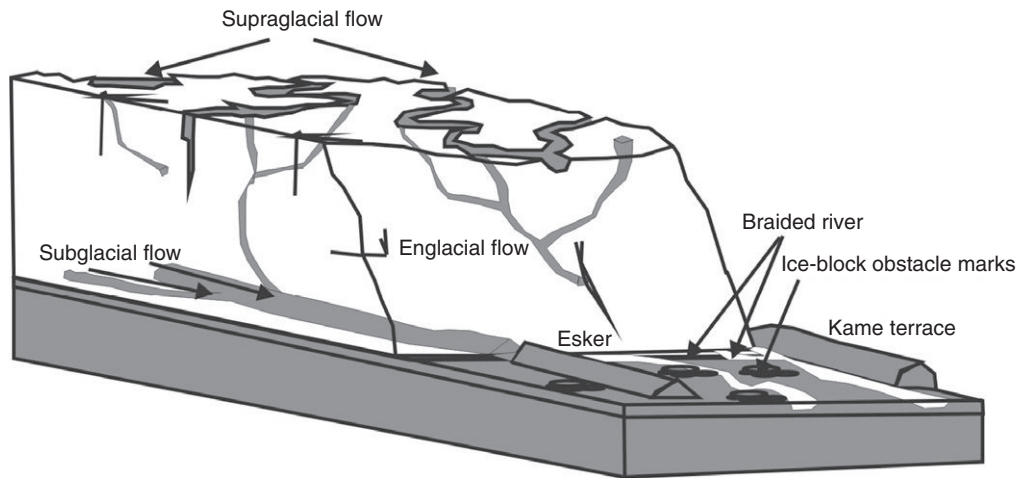


Figure 1 Conceptual diagram of the glaciofluvial system. Note: Not all glaciofluvial processes and landforms are depicted.

Table 1 Glaciofluvial landforms of deposition, in relation to situation and spatial scale

Spatial scale	Glaciofluvial landforms of deposition			
	Proglacial		Ice contact	
	Sandar		Subglacial	Ice marginal
Macroscale	<i>Unconfined</i> Outwash plain	<i>Confined</i> Valley fill		
Mesoscale	<i>Unconfined</i> Braided river bars Kettle holes Ice-block obstacle marks	<i>Confined</i> Gravel dunes Gravel antidunes Expansion bars Eddy bars Pendant bars	<i>Linear</i> Eskers <i>Nonlinear</i> Kames	<i>Linear</i> Kame terraces <i>Nonlinear</i> Subaerial outwash fans
Microscale	Smaller bedforms, such as ripples			

Categories of landforms are italicized.

regimes dominated by either normal ablation-controlled flows, rainfall floods, or glacial outburst floods, ‘jökulhlaups’ (e.g., Carrivick and Rushmer, 2009; Marren, 2005; Russell et al., 2006, 2009; Sambrook-Smith, 2000). Jökulhlaups can be generated either by the release of stored meltwater or by the direct melting of ice and snowfields by subglacial geothermal or volcanic activity. Most sandar are a product of a highly variable discharge regime within a relatively sediment-rich environment dominated by fluvial bedload transport. Consequently, sandar support braided river systems with highly erodible bed and bank materials.

Terraces are a characteristic of many proglacial fluvial systems, including sandar. Terraces constitute abandoned fluvial surfaces that have been formed by incision from more recent flows. Stacks or flights of terraces therefore preserve a record of successive depositional and erosional episodes, which can be due to a change in external forcing such as climate, base level, discharge regime, or glacier snout advance/recession, for example. A single erosional event within a fluvial system may generate a ‘complex response’ within the downstream channel consisting of multiple cycles of deposition and erosion resulting in the formation of multiple terraces (Schumm, 1973).

Sandar are composed of a number of glaciofluvial bar and bed forms. Although unconfined sandar are dominated by aggradation, their detailed surface topography and subsurface sedimentary architecture consist of both erosional and depositional forms. Erosional forms are the product of ‘intrinsic’ processes such as channel confluence scour, bedform migration, and waning flow stage incision or ‘extrinsic’ external factors such as glacier margin fluctuations or base-level change. By contrast, topographically confined sandar are characterized by a greater interaction of flow with the surrounding bedrock topography and a relatively higher and more rapidly varying river level (e.g., Glasser et al., 2009; Russell et al., 2009). It is possible that a large range of both erosional and depositional landforms can occur upon confined sandar (Carrivick, 2007a,b).

Unconfined sandar

Unconfined sandar occur in coastal or piedmont locations, where glaciofluvial sediment aggrades freely in a distributary manner in the relative absence of lateral topographic controls. Frequent overbank deposition, continuous scour and fill, and persistent bar modification produce a continuously

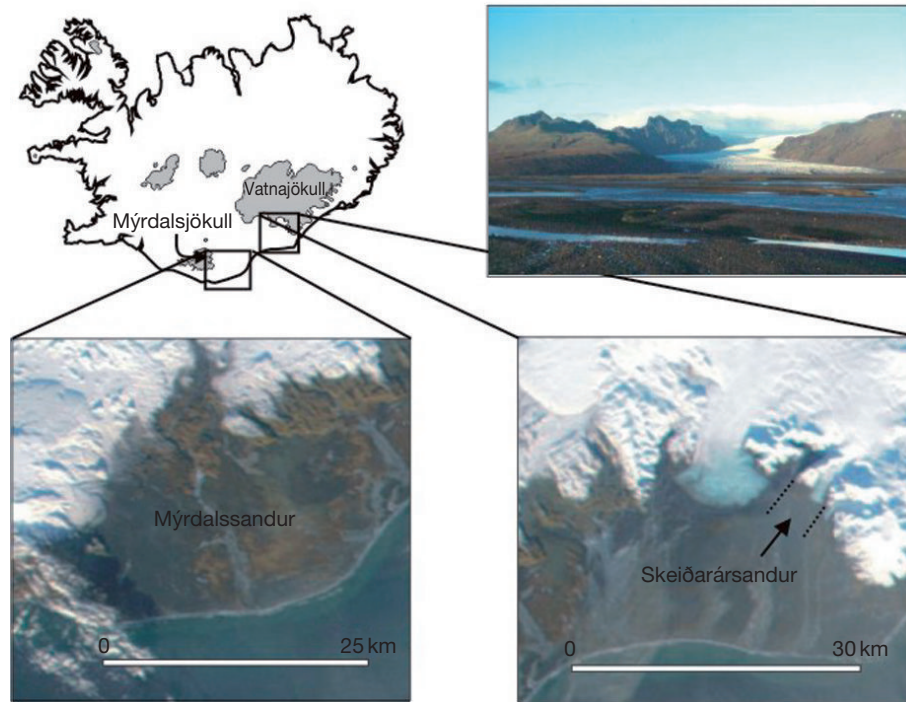


Figure 2 Two examples of large sandar in Iceland. Orientation and limits of photograph of Skaftafellsjökull is indicated by arrow and dashed lines. Photo credit: Prof. Fiona Tweed.

evolving surface in the case of sandar produced by ‘normal’ ablation-dominated meltwater. Sandar dominated by episodic high-magnitude meltwater discharges, such as jökulhlaups, will contain numerous terraces and paleo-bar surfaces, as well as sediments characterized by lenticular and tabular bedsets, all associated with deposition from turbulent, transcritical-supercritical flow (Duller et al., 2008). Bedforms and bars on unconfined sandar vary in size, from meters to hundreds of meters.

Sandar are typically concave in long profile. Marked downstream reduction in sandur gradient combined with progressive increase in channel system width results in a rapid downstream decrease in flow energy. As a result, major proximal–distal changes in grain size and depositional style can be detected on most unconfined sandar (Boothroyd and Nummedal, 1978). Proximal sandar surface morphology commonly comprises high-relief amplitude boulder bars incised by large channels. Ice-contact bar morphology is often controlled by meltwater vent or conduit mouth location and geometry, with bars radiating from an outlet channel or artesian vent. Ice-block obstacle marks and kettle holes also tend to be the largest and most prevalent near a glacier margin. Proximal grain size and lithofacies variations are highly variable. However, with increasing distance downsandur, large-scale bar and channel systems give way to a more complex braid pattern. Generally, clast sizes, clast size variation, and lithofacies variation also decrease downsandur, and clast roundness increases. In Alpine environments, nonglacial sources of water via tributaries from the nonglaciated parts of the catchment dampen discharge fluctuations with increasing distance downsandur.

Confined sandar

Meltwater flows constrained by areas of high relief are constrained by topography and thus are characterized by flows with greater depths and stream powers (Baker and Costa, 1987). This is partly because at high discharges, valley walls form resistant channel boundaries, and downstream constrictions in valley cross-profile may result in temporary backwater zones (Russell and Knudsen, 1999; Russell et al., 2009). Within a jökulhlaup-dominated system, the sedimentary infill of a given reach may be entirely depositional, or may contain both erosional and depositional forms, depending upon the ratio of flood magnitude to local channel gradient and geometry (e.g., Carrivick et al., 2004a,b). Such thick accumulations of sediment can be termed valley-confined sandar, valley-fill sediments (e.g., Church, 1978), or valley train deposits (Figure 3). Giant fluvial bedforms, particularly giant gravel bars, are common in confined sandar.

Many valley-confined sandar are dominated by a proximal flow expansion from either conduits or narrower bedrock-controlled feeder channels (Baker and Kochel, 1988). In medial channel zones, all other depositions tend to be focused at valley margins as longitudinal bars, point bars, or as pendant bars (Baker and Kochel, 1988). Vigorous lateral flow separation occurs within alcoves or reentrants in bedrock valley sides, generating eddy bars. In distal zones, deposits may comprise slackwater sediments emplaced within tributary valley mouths (e.g., Glasser et al., 2009). Main channel distal deposits can result from bedrock topographic constrictions that induce localized backwater ponding. While down-slope reductions in clast size, clast size variation, and lithofacies



Figure 3 Valley-fill sediments, 2 km wide, at Kverkfjöll, Iceland. Note isolated fluvially transported boulders. Thick accumulations of glaciofluvial sediment can exist in predominantly bedrock channels where local channel gradient rapidly decreases or where a valley cross-section widens.

variation are common to all sandar, greater spatial variation of depositional style can be expected within valley-confined sandar (Figure 4).

Giant fluvial bedforms

Glaciofluvial systems provide the best examples of giant fluvial bedforms such as dunes, antidunes, ice-block obstacle marks, and chute and pool complexes. As the morphology of giant bedforms may be confused with glaciofluvial bars, glaciolacustrine deltas, and kame terraces deposited by more frequent low-magnitude flows, great care must be taken when describing and interpreting both surface and subsurface sediments. Therefore, a holistic assessment including location, situation, scale, and sedimentology is emphasized below.

Gravel dunes

Giant gravel dunes are perhaps the most distinctive but least common bedforms attributed to glacial outburst floods. Giant gravel dunes are also referred to in the literature as giant dunes, gravel ripples, giant current ripples, and megaripples. Gravel dunes are sparsely reported in the literature due to comparative rarity of occurrence and generic confusion with antidunes and classes of gravel bars (Carling, 1999). Furthermore, since flow must be hydraulically smooth for the generation of ripples and hydraulically smooth flow is not associated with mobile gravel, it is most likely that observations of gravel ripples are actually incipient or transient dunes (Carling, 1999; Carling et al., 2002). Giant gravel dunes are controlled by flow depth and the size of particles available for transport and are thought to develop within subcritical flows, at Froude numbers of up to 0.75, although due to lag effects, transitional bedforms have been recorded for Froude numbers near the supercritical flow value of 1 (Carling, 1996, 1999).

Giant gravel dunes were first identified in association with the drainage of Lake Missoula (Baker and Bunker, 1985), and particularly within zones of flow expansion. Over 100 'trains' were identified with individual dunes of heights up to 15 m and wavelengths up to 150 m. Since then, there have been reports of giant gravel dunes in the Altai Mountains of Siberia



Figure 4 Confined outwash plain at (a) the Franz Josef glacier and (b) the Tasman valley, both New Zealand. The contrast between the boulder bar and the fine gravel bar in the foreground of (a) demonstrates considerable spatial flow variability. Both outwash plains are confined by steep valley walls and are of a gradual concave long profile. Photo credit: Dr. E. Lucy Rushmer.

with crests rising up to 10 m above their troughs with a wavelength of 50–150 m (Baker et al., 1993).

Besides an obvious flow-transverse undulating morphology, the sedimentology of giant gravel dunes is remarkably consistent between sites. They comprise coarse gravel-sized material that forms large foresets, with cross-strata frequently occurring at angles of up to 35° (Carling, 1996, 1999). Gravel foresets separated by reactivation surfaces are indicative of repeated high discharges (Russell and Marren, 1999). Large-scale gravel foresets record progressive downstream bedform migration by lee-side 'avalanche' processes (Carling et al., 2002). Giant gravel dunes are thought to be initiated by the migration of gravel sheets at high flow stage.

Gravel antidunes

Antidune bedforms develop in transitional-supercritical flows (Froude number ~ 1) in phase with stationary or upstream-migrating surface water waves. Antidunes are products of upstream bedform migration with crests typically orientated transverse to flow. Upstream dipping or 'backset' strata record progressive accretion of the bedform stoss side simultaneously with lee-side scour. Antidunes may also be completely erosional in nature developed within erodible bed materials. Antidune backsets typically dip upstream at a shallow angle ($\sim 2\text{--}3^\circ$). Development of antidunes within gravel (e.g., Duller et al., 2008) requires relatively shallow, supercritical flows with sufficient power and coarse-grained sediment availability. Preservation of gravel antidunes requires either sufficiently rapid flow stage reduction so as to prevent bed reworking or high-enough bed aggradation rates to allow bedform preservation at depth.

Ice-block obstacle marks

Ice blocks can become detached from a glacier snout due to differential ablation or mechanical fracturing. Hollows or pits within a sandar surface formed by the meltout of these ice blocks while they are held within glaciofluvial sediments are called kettle holes (e.g., Kjaer et al., 2004). Ice blocks can also be transported within high-magnitude floods from glaciers; jökulhlaups (Burke et al., 2010; Fay, 2002; Russell, 1993). Subsequent fluvial scour and deposition around an ice block can produce ice-block obstacle marks (e.g., Russell, 1993). Ice-block obstacle marks can be distinguished from kettle holes by a number of geomorphological and sedimentological features (Fay, 2002; Figure 5). For example, ice-block obstacle marks

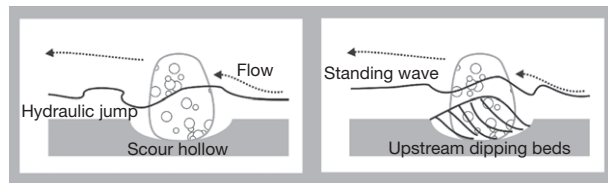


Figure 5 Development of antidune stoss-side strata within an ice-block obstacle mark. Redrawn from Fay H (2002) Formation of ice-block obstacle marks during the November 1996 glacier-outburst flood (jökulhlaup), Skeiðarársandur, southern Iceland. *Special Publication of the International Association of Sedimentologists* 32: 85–97.

tend to occur in association with depositional bars and may exhibit a boulder-like downstream fining in clast size (Fay, 2002). In contrast, buried stagnant glacial ice generally leads to bigger and spatially unsorted kettle holes. In terms of sedimentology, ice-block obstacle marks can feature distinctive antidune stoss-side strata, indicating localized supercritical flow around an ice block (Burke et al., 2010; Maizels, 1995, 1997). They commonly have aggradational obstacle shadows consisting of crudely bedded, matrix-supported, boulder gravel and stoss-side stratified open-work cobbles (Fay, 2002). The presence of well-sorted, stratified, and clast-supported gravel further suggests turbulent fluidal flow (Maizels, 1995, 1997).

Proglacial subaerial mesoforms

Channel macroforms or channel scale bars within subaerial environments can be classified into braided river bars, expansion bars, eddy bars, and pendant bars (Church and Gilbert, 1975). These forms are classified according to distinct processes and mechanisms operating at a channel scale within rivers. Braided river bars can be produced from a continuum of flow magnitudes and regimes. Expansion, eddy, and pendant bars tend to be associated with high-magnitude discharges, and specifically with a hydraulic jump, from subcritical to supercritical flow and the subsequent development of a fluvial shear zone.

Bars

Braided rivers have a multithreaded planform comprising many confluences and diffluences (Miall, 1977). Braid bars and channels can be classified as a hierarchy depending upon their relative size, and three orders of braiding can be seen in proglacial rivers with constant discharge (Williams and Rust, 1969; Figure 6). First-order braid bars are those largest bars that are separated by major channels. First-order bars are made up of a series of smaller, second-order braid bars, which are separated by smaller channels. Third-order braid bars are contained within second-order bars. The complexity of braid pattern is however, strongly stage-dependent, with rivers losing their braided character during peak flood conditions. Mid-channel or central bars, which include crescentic, longitudinal, and transverse bars, bank-attached, point, or unit bars, and larger composite forms known as compound bars or braid bar complexes, have been described in braided river systems on a range of scales ranging from meters to kilometers. All bars are subject to constant dissection via splays and chutes

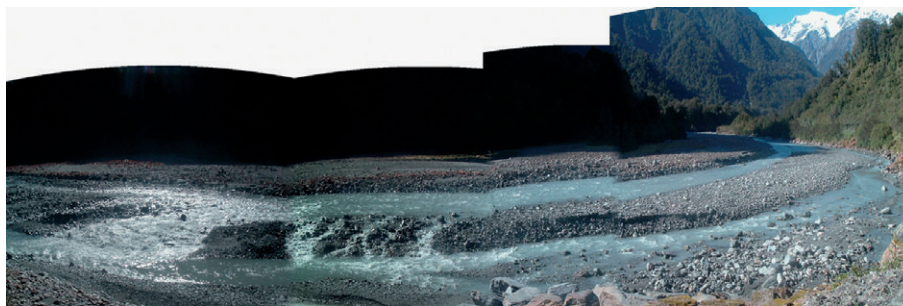


Figure 6 Development of chute channels on a bar upon a small confined outwash plain in New Zealand. Chute channels and bar dissection are products of fluctuating meltwater discharge and sediment flux. Photo credit: Dr. E. Lucy Rushmer.

(Figures 6 and 7), remobilization and migration (Miall, 1977, 1983). Only the largest bar forms produced by the most powerful flow events such as jökulhlaups may become stable over periods of years.

Bar formation can occur by deposition of material or by erosional dissection or preexisting topographic highs (Maizels, 1995). Bar growth occurs when initially small irregularities in a channel are amplified by deposition, creating conditions favorable for continued accumulation of sediment. Erosion and deposition can occur simultaneously, with sediment being gained and lost from different parts of the same bar.

Distinguishing between bars and islands in braided rivers is obviously problematic (Figure 8), as is a clear understanding of the influence of flow stage on braided river classification. For example, it is typically difficult to decide which channel segments should be used to define channel sinuosity and which bars should be used to define channel splitting (Figure 8). Besides river discharge regime, the volume and character of glaciofluvial sediment is a key factor in controlling braiding (Church and Gilbert, 1975).

Expansion bars

Expansion bars are associated with deposition downstream of zones of rapid flow deceleration, such as hydraulic jumps generated at the junction between steep, confined reaches and wide- and low-gradient reaches (Baker, 1984; Figure 9). Expansion bars initiated at flood peak become progressively modified as stage recedes (Church and Jones, 1982) and also during subsequent flow events. Waning stage flows may incise expansion bars, commonly forcing a terraced or channeled surface and a fan-like plan, although incision may be inhibited in fluvial systems dominated by boulder-sized sediment. Identification of expansion bars may be aided by bar-surface cobble and boulder lags, which represent winnowing of fine-grained

clasts during receding flows (e.g., Carrivick et al., 2004a,b). Sediment clast sizes typically show a marked decrease in a downstream direction upon expansion bars and also units that become progressively finer upward.

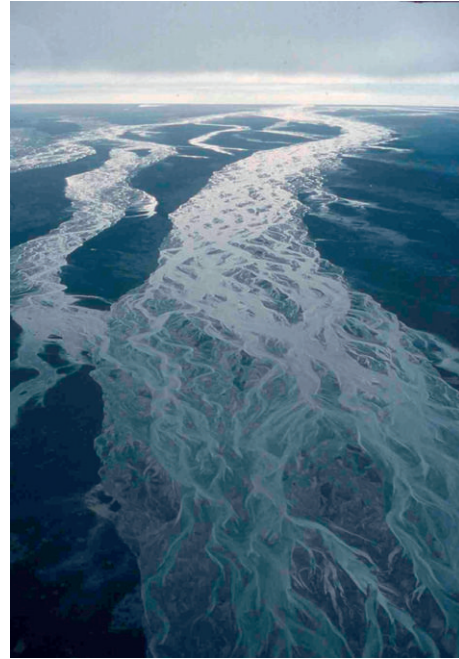


Figure 8 An example of a glaciofluvial braided river comprised of numerous stream threads and constantly altering sand and gravel bars. The 2-km-wide active channel develops a ceaseless pattern of deposition and remobilization.

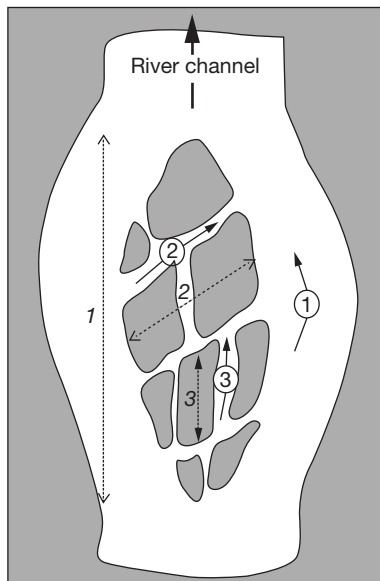


Figure 7 Hierarchy of channels and bars within a braided river system. Redrawn from Williams PF and Rust BR (1969) The sedimentology of a braided river. *Journal of Sedimentary Petrology* 39(2): 649–679. Channels denoted by solid arrows, and bars by dashed lines.

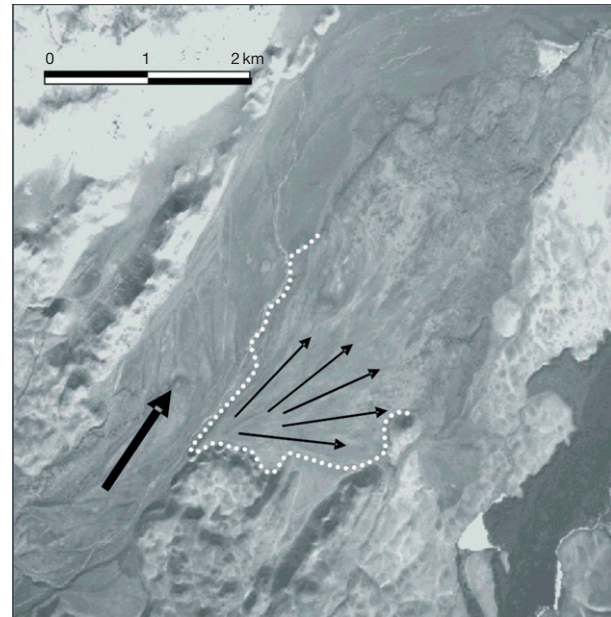


Figure 9 Expansion fan at the mouth of Hraundalur, north-central Iceland. Thick arrow denotes main channel slope; thinner arrows denote direction of expansion fan paleoflow. Dashed white line marks limit of deposition.

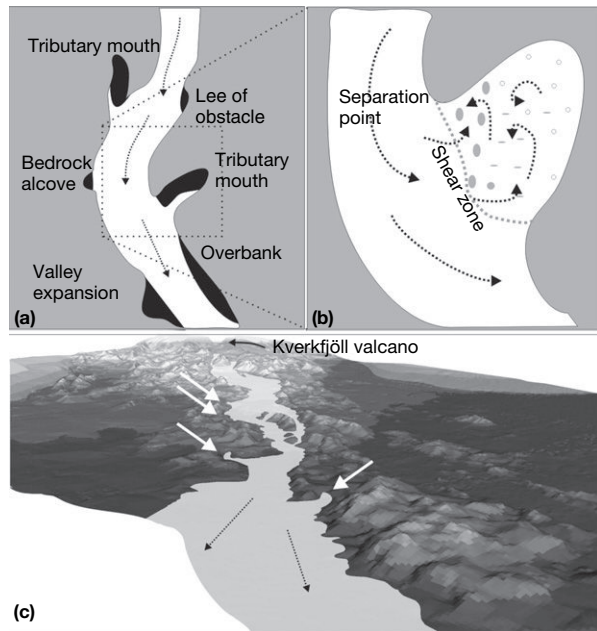


Figure 10 Schematic representation of the most common locations for eddy bar and slackwater deposits in confined valley and bedrock settings (a) (redrawn from Baker VR and Kochel RC (1988) Flood sedimentation in bedrock fluvial systems: In: Baker VR, Kochel RC, and Patton PC (eds.) *Flood Geomorphology*, pp. 123–137. New York: John Wiley and Sons), and typical flow conditions associated with eddy bar formation (b) (adapted from O'Connor JE (1993) Hydrology, hydraulics and geomorphology of the Bonneville Flood. *Geological Society of America Special Paper* 274: 83 p.). Image of a 2D hydrodynamic model of a jökulhlaup from Kverkfjöll volcano, Iceland, routing 25 km downstream (c) (reproduced from Carrivick JL (2006) 2D modelling of high magnitude outburst floods: an example from Kverkfjöll volcano, Iceland. *Journal of Hydrology* 321: 187–199). Dashed arrows indicate flow direction, and major slackwater zones are indicated by white arrows.

Eddy bars

Eddy bar sediments are derived from two different sources; namely from tributary flow and from marginal channel flow (Figure 10(a)). Eddies result between zones of higher and lower velocity flow and promote deposition due to abruptly decreased sediment transport capacity (O'Connor, 1993).

Eddy bars can contain a wide variety of sediment sizes and structures, including poorly sorted, commonly cross-stratified boulder gravel and cross-stratified graded gravel (Church and Jones, 1982; Figure 10(b)). Typically, foresets dip away from the main flow direction, containing well-sorted subrounded boulders and cobbles with an open framework structure (Miall, 1983). Thus, varying directions of transport are distinguished, whereas in a pendant or expansion bar, flow is more unidirectional. Smaller foresets comprise granules and pebbles dipping locally upstream, thus reflecting swirling eddies. Eddy bars at tributary mouths tend to grade upstream into slackwater deposits (Figure 10(b)). Deposition during multiple events may result in interfingering of graded sand-silt units with laminated silt units (Figure 10(b)).

Pendant bars

Pendant bars are narrow, sharp-crested, streamlined landforms of sand and gravel-sized clasts, which are deposited

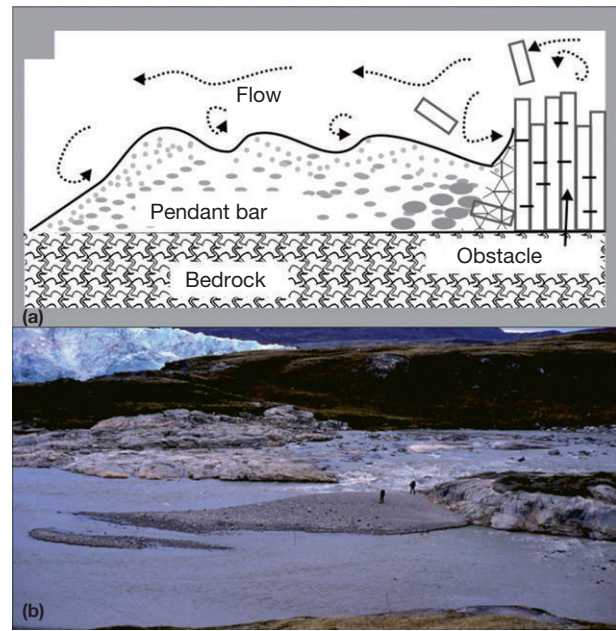


Figure 11 Schematic development and characteristics of pendant bars (a). Redrawn from Baker VR (1978) Large-scale erosional and depositional features of the channelled scabland. In: Baker VR and Nummedal D (eds.) *The Channelled Scabland: Comparative Planetary geology Field Conference Guidebook, Columbia Basin*, pp. 17–35. Washington, DC: National Aeronautical Space Administration. Example of a pendant bar in west Greenland (b); note persons for scale.

immediately downstream of obstacles within a main flow path (Figure 11(a) and 11(b); Miall, 1983; Russell and Marren, 1998). Pendant bars up to 2 km long and 30 m high have been identified in steep channels of uniform gradient within the Missoula flood landscape (Baker, 1978, 1984; Baker and Bunker, 1985; Baker and Kochel, 1988). Deposits in the lee of large bedrock obstacles can be regarded also as pendant bars (Figure 11(a) and 11(b)). Pendant bars are therefore commonly associated with proximal and lateral hollows formed by intense and localized scour. Relative flow depth around or over the obstacle will determine the exact interaction between hydraulics and sediment. If an obstacle is completely submerged at high flow depth, pendant bars may develop as products of a localized hydraulic jump, as flow accelerates over the obstacle (Figure 11(a)). The internal structure of pendant bars can consist of simple large-scale gravel foresets or massive matrix-supported gravel. In the Altai, coarse-grained pendant bar sediments typically consist of high-amplitude gravel foresets interpreted as the product of bar deposition by deep flows (Baker et al., 1993). The predominance of large foresets in the bars suggests growth by downstream progradation, rather than by vertical accretion.

Ice-Contact Glaciofluvial Landforms of Deposition

Glaciofluvial deposition in supraglacial, englacial, subglacial, and proglacial environments can produce ice-contact landforms. Ice-contact depositional landforms are usually classified as linear or nonlinear features depending on the nature and degree of glacier control on meltwater flow pathways. Ice-contact landforms experience postdepositional modification

to varying degrees due to loss of support upon ice melt, or compression by glacier ice advance, or overriding.

Linear features

Kame terraces and glaciofluvial eskers are deposited directly in contact with glacier margins and ice-conduit walls, respectively. Kame terrace and esker morphology is a direct reflection of the linear nature of their respective ice contacts.

Kame terraces

Kame terraces are generally discontinuous and often pitted benches deposited by rivers flowing between a valley side and an ice margin (Figure 1). They are usually located on the distal slopes of recessional marginal 'moraine' ridges and developed during deglaciation, probably due to the interaction of active ice lobes (surges) and masses of dead ice that persisted beyond the ice margin (Bitinas, 2011). Kame terraces, therefore, frequently occur in stacks or flights on either side of a valley. They may not be altitudinally matched (paired), nor have similar gradients, but do tend to have gently sloping surfaces which reflect the gradient of the former ice margin (Figure 12(a)). Parallel kame terraces on a single valley side typically represent successive stages in the lowering of a glacier surface



Figure 12 Parallel kame terraces with gently sloping surfaces representing successive downwasting of the glacier surface (a), and are composed of glaciofluvial sands and gravels (b).

(Figure 12(a)). Kame terraces are predominantly composed of sands and gravels (Figure 12(b)), but can also contain glaciolacustrine deposits from meltwater ponds.

Eskers

Glaciofluvial eskers are named after the Irish 'eiscir,' which means ridge, and are known as 'osar' in Scandinavia (Benn and Evans, 2010). Eskers have many possible origins including sub- and englacial conduit deposits (Figure 1), fracture fills, ice-walled open-roofed channels, and subaqueous outwash fans (Brennand, 2000; Warren and Ashley, 1994). They can be formed by deposition of sediments from 'normal' ablation-controlled meltwater, or during episodic high-magnitude discharges, such as jökulhlaups. Esker ridges often emerge from chaotic downwasting surfaces of ice-cored outwash (e.g., Boulton et al., 2007; Burke et al., 2008; Russell et al., 2006) and can show evidence of continuous deposition over hundreds of kilometers. Such distances might indicate regional ice-sheet stagnation (Brennand, 2000) or a strong coupling of subglacial pathways with groundwater catchments, most likely during deglaciation (Boulton et al., 2007). Segments of an esker may be deposited time-transgressively during glacier retreat (Warren and Ashley, 1994). Hooke et al. (2007) reported that segments of an esker increase in size down-ice as a product of melt rates exceeding tunnel closure rates by an increasing amount toward the ice margin, leading to a reduction in water flow velocity and increased deposition of sediments. Conduit-fill eskers can be found in isolation (Figure 13) or within interconnected anastomosing networks (Figure 14) whose routeways are determined by ice-sheet surface morphology and subglacial topography interactions. Glaciofluvial esker segments can terminate in subaerial or subaqueous outwash fans to form esker beads. A 'till esker' has been identified by Evans et al. (2010) at Sandfellsjökull in southern Iceland and attributed to squeezing of a dilatant till into an elongated cavity or R-channel after meltwater evacuation. This deposit is indirectly linked to glacial meltwater, which must have generated appropriate conditions for nonfluvial emplacement of till. Eskers tend to display considerable downstream variation in morphology,



Figure 13 Example of a ~3 km esker in northern Sweden.

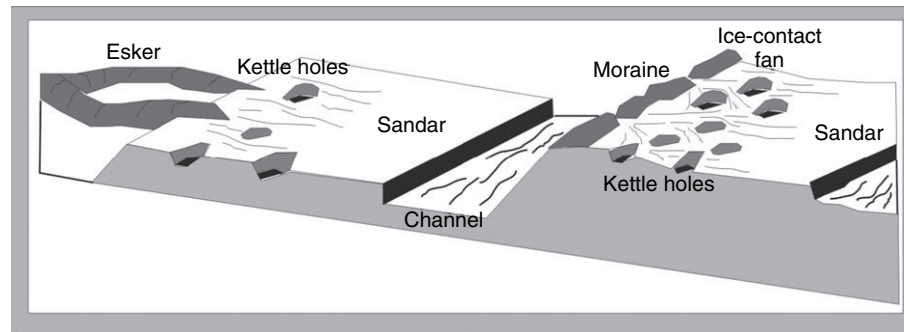


Figure 14 Schematic diagram of the common spatial relationship between eskers, sandar, moraine, and ice-contact fans. Redrawn from Price RJ (1969) Moraines, sandar, kames and eskers near Breiðamerkjökull, Iceland. *Transactions of the Institute of British Geographers* 46: 17–43.

with localized gaps and enlargements, which reflect deposition within expansive sub- and englacial cavities and zones of non-deposition, respectively. Eskers can be composed of a wide range of facies, including silts, sands, gravels, boulders, and matrix-supported diamictos. Englacial esker sediments experience great postdepositional modification during meltout, resulting in less well-defined ridge morphology. Removal of ice support from sub- and englacial eskers produces slumping characterized by normal faults.

Fluvial processes within conduits differ from those in sub-aerial environments in a number of ways. Meltwater within conduits flows down a hydraulic gradient controlled by glacier surface slope, allowing pressurized water to flow up-hill (Röthlisberger and Lang, 1987; Shreve, 1972). Meltwater pressure and velocity within conduits are highly variable due to temporal and spatial variations in meltwater discharge and conduit size. Conduit diameter is a function of the balance between meltwater-driven melting and closure due to plastic deformation of glacier ice. Maximum flow velocities within eskers are thought to be higher than within subaerial fluvial systems due to low conduit roughness.

Conduit eskers are characterized by their distinctive large-scale sedimentary architecture. In cross-section, the characteristic arched or pseudo-anticlinal structure of eskers is believed to develop when deposition is dominated by large-scale vortices, which sweep sediment from the flanks into the middle of the esker ridge (Brennand, 2000). Esker sediments also may fine laterally, away from ridge crests, usually with a transition from large-scale cross-bedded gravel and boulders in the esker core to horizontally bedded sand and laminated silt at the margins. These sedimentary sequences are thought to represent conditions where the conduit is expanding via ice wall melting and/or mechanical erosion. Where conduit geometry is irregular, channel macroforms are deposited at flow expansions (Brennand, 2000; Burke et al., 2008). Macroforms, hundreds of meters in length and tens of meters in thickness, display large up-stream dipping stoss sides and distal foreset beds. Macroforms can be deposited within single events, such as jökulhlaups (Burke et al., 2008, 2010), or can be deposited by repeated events. Large-scale gravel dunes and gravel sheets are commonly superimposed upon esker macroforms. Although esker facies and grain sizes include those commonly found within all glaciofluvial settings, esker core sediments do contain a greater occurrence of matrix-supported gravel and boulders,

which reflects rapid deposition from highly sediment-charged flows (Brennand, 1994, 2000).

Nonlinear features

Nonlinear ice-contact glaciofluvial features include subaerial outwash fans, kames, and hummocky terrain. These landforms are products of deposition at ice margins or within subglacial cavities.

Subaerial outwash fans

Ice-contact outwash fans are fed by either sub-, en-, or supra-glacial meltwater systems, and may therefore be associated with esker ridges (Figure 14). They are suggested to be an end member of a continuum of landforms (e.g., Pisanska-Jamroz, 2006). Outwash fans range from decimeter to kilometer in scale and are usually composite cone-shaped landforms created by aggradation of material deposited from a range of flow events of varying magnitude and frequency. Additionally, outwash fans are created during glacier surges or during jökulhlaups (Russell and Knudsen, 2002; Russell et al., 2001). Outwash fans are a cone of aggradation (Figure 14), radiating from an apex at an ice-marginal meltwater portal. Outwash fans in close proximity to each other coalesce (Figures 14 and 15) to form large glaciofluvial ‘end’ moraine complexes such as the Salpausselkä ridges in southern Finland. Outwash fans have steeper slopes than sandar, which they grade into (Figure 15). Steep fan gradients encourage the



Figure 15 An ice-contact subaerial outwash fan at Skeiðarárjökull, Iceland.



Figure 16 Example of kames, or hummocky meltout topography.

development of supercritical flows generating planar bed conditions and localized antidunes. Outwash fan sediments typically comprise fan surface parallel bedding and localized low-angle backsets (Russell and Knudsen, 1999, 2002).

Kames

The Scottish words ‘cam’ or ‘kaim,’ which mean ‘sinuous steep-sided mound,’ are the origins of the term kame (Benn and Evans, 2010). The term kame refers to a suite of steep-sided mounds that are usually deposited supraglacially or at an ice margin. An example is moulin deposits. Kame and kettle topography (Figure 16) is a zone of ridges and mounds, or ‘kames,’ between kettle holes. While kettle holes are products of the melting of buried glacial ice, kame ridges and mounds are glaciofluvial deposits. Kame deposits can be disarranged or with some alignment, reflecting former glaciofluvial routes and characteristics (Roberts et al., 2001). Although kames comprise the complete spectrum of glaciofluvial deposits, they commonly display signs of postdepositional modification due to the removal of ice support comprising normal faulting and sagging/collapse. Thus, kames, like eskers, can exhibit complexity and interconnectivity on both temporal and spatial scales; that is, in both formation and preservation. Livingstone et al. (2010) have recently described these characteristics for the Brampton kame belt.

Links to Other Systems

Glaciofluvial systems are driven by fluctuations in glacier mass balance and to a lesser extent by rainfall–runoff events. Glacial erosion is a major supplier of glaciofluvial material, and finer-grained sediment is often sourced from aeolian action, which is enhanced by strong thermal gradients in glacial terrain. Landforms of glaciofluvial deposition are influenced by local topography, geology, and situation. Glaciofluvial systems can be directly coupled to glaciolacustrine and glacimarine environments, by conduits within tidewater glaciers, for example. Indirect coupling of glaciofluvial and glaciolacustrine or glaci-marine systems can occur via delta-top environments, such as fjord-head sandar and large coastal sandur–delta complexes, such as Skeiðarársandur in southern Iceland.

Conclusion

Glaciofluvial processes are effective agents of landscape change, eroding, transporting, and depositing large amounts of sediment at a wide range of spatial scales. Macroscale landforms are composite landforms, formed cumulatively by a range of meltwater discharges and sediment flux regimes. Macroforms can, in exceptional cases, be produced by episodic high meltwater discharge, such as from glacier surges, or from glacial outburst floods or ‘jökulhlaups.’ Mesoscale landforms include bedforms of several hundred meters in plan-form geometry, such as dunes and bars, and tend to be dependent upon reach-scale equilibrium sediment transport. Microscale landforms include ripples, are generally products of transient hydraulics, and are of a plan-form of the order of one to a few tens of meters. Glaciofluvial processes are controlled by glacier mass balance and resultant meltwater discharge, and are therefore intimately linked to glacier fluctuations. Glaciofluvial systems transfer vast amounts of sediment, to other fluvial, lacustrine, and marine environments. Sediment is most abundant proximal to active ice margins. Glaciofluvial landforms of deposition can be categorized as either proglacial or ice-contact landforms. On the basis of morphology alone, both types of landforms are difficult to unequivocally identify, without recourse to sedimentary character, depositional setting, and geometric scale. Additionally, ice-contact landforms will inevitably feature secondary deformation as ice support is lost due to ice mass melting. However, recognition of glaciofluvial landforms can permit reconstruction of glacial ‘plumbing’ systems and can imply glacier behavior from meltwater discharge magnitude–frequency regimes. Reconstructions of past glaciofluvial events in space and time, and hence of glacial activity, are important for understanding glacial chronology and glacial responses to, for example, climate change.

See also: Glacial–Interglacial Scale Fluvial Responses. **Fluvial Environments:** Responses to Rapid Environmental Change; **Sediments:** Glacial Landforms: Glacial Landforms; Glaciofluvial Landforms of Erosion; Glacitectonic Structures and Landforms. **Glacial Landforms, Sediments:** Glaciogenic Lithofacies.

References

- Baker VR (1978) Large-scale erosional and depositional features of the channelled scabland. In: Baker VR and Nummedal D (eds.) *The Channelled Scabland: Comparative Planetary geology Field Conference Guidebook, Columbia Basin*, pp. 17–35. Washington, DC: National Aeronautical Space Administration.
- Baker VR (1984) Flood sedimentation in bedrock fluvial systems. In: Koster EH and Steel HJ (eds.) *Sedimentology of Gravels and Conglomerates*, Canadian Society of Petroleum Geologists, Memoir 10: 87–98.
- Baker VR, Benito G, and Rudoy AN (1993) Palaeohydrology of late Pleistocene superflooding, Altay Mountains, Siberia. *Science* 259: 348–350.
- Baker VR and Bunker RC (1985) Cataclysmic late Pleistocene flooding from glacial Lake Missoula: A review. *Quaternary Science Reviews* 4: 1–41.
- Baker VR and Costa JE (1987) Flood Power. In: Mayer L and Nash D (eds.) *Catastrophic Flooding*, pp. 1–25. The Bingham Symposium in Geomorphology.
- Baker VR and Kochel RC (1988) Flood sedimentation in bedrock fluvial systems. In: Baker VR, Kochel RC, and Patton PC (eds.) *Flood Geomorphology*, pp. 123–137. New York: Wiley.
- Benn DI and Evans DJA (2010) *Glaciers and Glaciation*, p. 802. London: Arnold.

- Bitinas A (2011) New insights into the last deglaciation of the south-eastern flank of the Scandinavian Ice Sheet. *Quaternary Science Reviews* 44: 69–80.
- Boothroyd JC and Nummedal D (1978) Proglacial braided outwash: A model for humid alluvial-fan deposits. *Canadian Society of Petroleum Geologist, Memoir* 5: 641–668.
- Boulton GS, Lunn R, Vidstrand P, and Zatsepin S (2007) Subglacial drainage by groundwater-channel coupling, and the origin of esker systems. Part 1 – Glaciological observations. *Quaternary Science Reviews* 26: 1067–1090.
- Brennand TA (1994) Macroforms, large bedforms and rhythmic sedimentary sequences in subglacial eskers, south-central Ontario: Implications for esker genesis and meltwater regime. *Sedimentary Geology* 91: 9–55.
- Brennand TA (2000) Deglacial meltwater drainage and glaciodynamics: Inferences from Laurentide eskers, Canada. *Geomorphology* 32: 263–293.
- Burke MJ, Woodward J, Russell AJ, Fleisher PJ, and Bailey PK (2008) Controls on the sedimentary architecture of a single event englacial esker: Skeiðarárjökull, Iceland. *Quaternary Science Reviews* 27: 1829–1847.
- Burke MJ, Woodward J, and Russell AJ (2010) Sedimentary architecture of large-scale, jökulhlaup-generated, ice-block obstacle marks: Examples from Skeiðarársandur, SE Iceland. *Sedimentary Geology* 227: 1–10.
- Carling PA (1996) Morphology, sedimentology and paleohydraulic significance of large gravel dunes, Altai Mountains, Siberia. *Sedimentology* 43: 647–664.
- Carling PA (1999) Subaqueous gravel dunes. *Journal of Sedimentary Research* 69(3): 534–545.
- Carling PA, Cao Z, and Ervine DA (2002) Flood plain contribution to open channel flow structure. In: Dyer FJ, Thoms MC, and Olley JM (eds.) *The Structure, Function and Management Implications of Fluvial Sedimentary Systems*, pp. 227–237. Alice Springs, Australia: IAHS Press.
- Carrivick JL (2007a) Hydrodynamics and geomorphic work of jökulhlaups (glacial outburst floods) from Kverkfjöll volcano, Iceland. *Hydrological Processes* 21: 725–740.
- Carrivick JL (2007b) Modelling coupled hydraulics and sediment transport of a high-magnitude flood and associated landscape change. *Annals of Glaciology* 45: 143–154.
- Carrivick JL and Rushmer EL (2009) Inter- and Intra-catchment variability in proglacial geomorphology: An example from Franz Josef Glacier and Fox Glacier, South Westland, New Zealand. *Arctic, Antarctic, and Alpine Research* 41(1): 18–36.
- Carrivick JL, Russell AJ, and Tweed FS (2004a) Geomorphological evidence for jökulhlaups from Kverkfjöll volcano, Iceland. *Geomorphology* 63: 81–102.
- Carrivick JL, Russell AJ, Tweed FS, and Twigg D (2004b) Palaeohydrology and sedimentology of jökulhlaups from Kverkfjöll volcano, Iceland. *Sedimentary Geology* 172: 19–40.
- Church M (1978) Palaeohydrological reconstructions from a Holocene valley fill. In: Miall AD (ed.) *Fluvial Sedimentology*, pp. 743–772. Canadian Society of Petroleum Engineers Memoir 5.
- Church M and Gilbert R (1975) Proglacial fluvial and lacustrine sediments. In: Jopling AV and McDonald BC (eds.) *Glaciofluvial and Glaciolacustrine Sedimentation* pp. 22–100. Tulsa, OK: SEPM. Special Publication 23.
- Church M and Jones D (1982) Channel bars in gravel-bed rivers. In: Hey RD, Bathurst JC, and Thorne CR (eds.) *Gravel-Bed Rivers*, pp. 291–338. London: Wiley.
- Duller RA, Mountney NP, Russell AJ, and Cassidy NC (2008) Architectural analysis of a volcanoclastic jökulhlaup deposit, southern Iceland: Sedimentary evidence for supercritical flow. *Sedimentology* 55: 939–964.
- Evans DJA, Nelson CD, and Webb C (2010) An assessment of fluting and 'till esker' formation on the foreland of Sandfellsjökull, Iceland. *Geomorphology* 114: 453–465.
- Fay H (2002) Formation of ice-block obstacle marks during the November 1996 glacier-outburst flood (jökulhlaup), Skeiðarársandur, southern Iceland. *Special Publication of the International Association of Sedimentologists* 32: 85–97.
- Fitzsimons SJ (2003) Ice-marginal terrestrial landsystems: Polar continental glacier margins. In: Evans DJA (ed.) *Glacial Landsystems*, pp. 89–110. London: Arnold.
- Glasser NF, Harrison S, and Jansson KN (2009) Topographic controls on glacier sediment-landform associations around the temperate North Patagonian Icefield. *Quaternary Science Reviews* 28: 2817–2832.
- Hooke R, Le B, and Fastook J (2007) Thermal conditions at the bed of the Laurentide Ice Sheet in Maine during deglaciation: Implications for esker formation. *Journal of Glaciology* 53: 646–658.
- Kjaer KH, Sultan L, Kruger J, and Shomacker A (2004) Architecture and sedimentation of outwash fans in front of the Myrdalsjökull ice cap, Iceland. *Sedimentary Geology* 172: 139–163.
- Leeder MJ (1999) *Sedimentology and Sedimentary Basins: From Turbulence to Tectonics*, p. 592. Oxford: Blackwell.
- Livingstone SJ, Cofaigh C, Evans DJA, and Hopkins J (2010) The Brampton kame belt and Pennine escarpment meltwater channel system (Cumbria, UK): Morphology, sedimentology and formation. *Proceedings of the Geologists Association* 121: 423–443.
- Maizels JK (1995) Sediments and landforms of modern proglacial terrestrial environments. In: Menzies JE (ed.) *Modern Glacial Environments*, pp. 365–416. Oxford: Butterworth-Heinemann.
- Maizels JK (1997) Jökulhlaup deposits in proglacial areas. *Quaternary Science Reviews* 16: 793–819.
- Marren PM (2005) Magnitude and frequency in proglacial rivers: A geomorphological and sedimentological perspective. *Earth-Science Reviews* 70: 203–251.
- Miall AD (1977) A review of the braided river environment. *Earth-Science Reviews* 13: 1–62.
- Miall AD (1983) Glaciofluvial transport and deposition. In: Eyles N (ed.) *Glacial Geology*, pp. 168–183. New York: Pergamon.
- O'Connor JE (1993) Hydrology, hydraulics and geomorphology of the Bonneville Flood. *Geological Society of America Special Paper* 274: 83 pp.
- Old GH and Lawler DM (2005) Flow and sediment dynamics of a glacial outburst flood from the Skafá system in southern Iceland. *Earth Surface Processes and Landforms* 30: 1441–1460.
- Pisarska-Jamroz M (2006) Transitional deposits between the end moraine and outwash plain in the Pomeranian glaciomarginal zone of NW Poland: A missing component of ice-contact sedimentary models. *Boreas* 35: 126–141.
- Roberts MJ, Russell AJ, Tweed FS, and Knudsen Ó (2001) Controls on englacial sediment deposition during the November 1996 Jökulhlaup, Skeiðarárjökull, Iceland. *Earth Surface Processes and Landforms* 26: 935–952.
- Röthlisberger H and Lang H (1987) Glacial hydrology. In: Gurnell AM and Clark MJ (eds.) *Glaciofluvial Sediment Transfer: An Alpine Perspective*, pp. 207–284. Chichester: Wiley.
- Rushmer EL (2006) Sedimentological and geomorphological impacts of the jökulhlaup (glacial outburst flood) in January 2002 at Kverkfjöll, northern Iceland. *Geografiska Annaler* 88A(1): 1–11.
- Russell AJ (1993) Obstacle marks produced by flow around stranded ice blocks during a glacier outburst flood (jökulhlaup) in west Greenland. *Sedimentology* 40: 1091–1111.
- Russell AJ, Carrivick JL, Ingeman-Nielsen T, Yde JC, and Williams M (2009) A new cycle of jökulhlaups at Russell Glacier, Kangerlussuaq, West Greenland. *Journal of Glaciology* 202: 238–246.
- Russell AJ, Knight PG, and Van Dijk TAGP (2001) Glacier surging as a control on the development of proglacial, fluvial landforms and deposits, Skeiðarársandur, Iceland. *Global and Planetary Change* 28: 163–174.
- Russell AJ and Knudsen Ó (1999) Controls on the sedimentology of the November 1996 jökulhlaup deposits, Skeiðarársandur, Iceland. In: Smith ND and Rogers J (eds.) *Fluvial Sedimentology VI*, pp. 315–324. International Association of Sedimentologists Special Publication, 28. Oxford: Blackwell Science.
- Russell AJ and Knudsen Ó (2002) The effects of glacier-outburst flood flow dynamics on ice-contact deposits: November 1996 jökulhlaup, Skeiðarársandur, Iceland. In: Martini P, Baker VR, and Garzon G (eds.) *Flood and Megaflood Processes and Deposits: Recent and Ancient Examples*, pp. 67–84. IAS Special Publication 32. Oxford: Blackwell Science.
- Russell AJ and Marren PM (1998) A Younger Dryas (Loch Lomond Stadial) jökulhlaup deposit, Fort Augustus, Scotland. *Boreas* 27: 231–242.
- Russell AJ and Marren PM (1999) Proglacial fluvial sedimentary sequences in Greenland and Iceland: A case study from active proglacial environments subject to jökulhlaups. In: Jones AP, Tucker ME, and Hart JK (eds.) *The Description and Analysis of Quaternary Stratigraphic Field Sections*, pp. 171–208. London: Quaternary Research Association. Technical Guide 7.
- Russell AJ, Roberts MJ, Fay H, et al. (2006) Icelandic jökulhlaup impacts: Implications for ice-sheet hydrology, sediment transfer and geomorphology. *Geomorphology* 75: 33–64.
- Sambrook-Smith GH (2000) Small-scale cyclicity in alpine proglacial fluvial sedimentation. *Sedimentary Geology* 132: 217–231.
- Schumm SA (1973) Geomorphic thresholds and complex response of drainage systems. In: *Proceedings of the 4th Annual Geomorphology Symposium*, pp. 299–310. Binghamton.
- Shreve RL (1972) Movement of water in glaciers. *Journal of Glaciology* 11: 205–214.
- Warren WP and Ashley GM (1994) Origins of the ice-contact stratified ridges (eskers) of Ireland. *Journal of Sedimentary Research* A64(3): 433–449.
- Williams PF and Rust BR (1969) The sedimentology of a braided river. *Journal of Sedimentary Petrology* 39(2): 649–679.

Further Reading

- Bennett MR and Glasser NF (1996) *Glacial Geology: Ice Sheets and Landforms*. Chichester: Wiley.

- Carrivick JL (2006) 2D modelling of high magnitude outburst floods: An example from Kverkfjöll volcano, Iceland. *Journal of Hydrology* 321: 187–199.
- Evans DJA and Twigg DR (2002) The active temperate glacial landsystem: A model based on Breiðamerkurjökull and Fjallsjökull, Iceland. *Quaternary Science Reviews* 21: 2143–2177.
- Fisher TG, Clague JJ, and Teller JT (2002) The role of outburst floods and glacial meltwater in subglacial and proglacial landform genesis. *Quaternary International* 90: 1–115.
- Hambrey MJ and Alean J (2004) *Glaciers*, 2nd edn. Cambridge: Cambridge University Press.
- Lowe JJ and Walker MJC (1997) *Reconstructing Quaternary Environments*. London: Prentice-Hall.
- Menzies J (ed.) (1995) *Modern Glacial Environments: Processes, Dynamics and Sediments*. Oxford: Butterworth-Heinemann.
- Menzies J (ed.) (1996) *Past Glacial Environments: Sediments, Forms and Techniques*. Oxford: Butterworth-Heinemann.
- Price RJ (1969) Moraines, sandar, kames and eskers near Breiðamerkurjökull, Iceland. *Transactions of the Institute of British Geographers* 46: 17–43.
- Zielinski T and van Loon AJ (2003) Pleistocene sandur deposits represent braidplains, not alluvial fans. *Boreas* 32: 590–611.

Glaciogenic Lithofacies

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The Meaning of Facies

The term glaciogenic refers to any sediment or landform that results from glacial action. In the main, glaciogenic sediment is produced directly by erosion and comminution at the base of glaciers and ice sheets and redistributed by other processes. The word facies means 'appearance of' (hence 'face'). Lithofacies are simply different types of clastic or chemical sediments (and their lithified rock equivalents) produced by gravity, water, ice, or wind in sedimentary environments. Sediments are readily classified informally into facies based simply on color, texture, lithology of clastic particles, internal structure, thickness and geometry, presence or absence of body and trace fossils, and any postdepositional structures (e.g., reddish, fossiliferous, pebbly quartz-rich sandstone with burrows). There are several formal classifications of sedimentary lithofacies (Eyles et al., 1983; Ghilardo, 1992; Walker, 1992). A widely used practice is to study lithofacies accumulating in the present day and use these as keys to interpret ancient deposits and their formative processes. Unfortunately, few sedimentary environments are typified by a single diagnostic lithofacies because most environments are spatially complex, involving the interaction of many different processes. An exception is the environment where large storm waves create distinct hummocky bedforms in shallow water (producing a distinctive hummocky and swaley cross-stratified sand facies). The latter is referred to as a so-called universal facies type because its origin is unique to a particular setting (i.e., wave-dominated shallow water). Another example is that of a graded turbidite bed deposited by turbidity currents flowing downslope under water. So-called common facies types, on the other hand, are produced in many environments (eolian, deep marine, fluvial, etc.) and are not uniquely diagnostic of any one depositional setting (e.g., rippled cross-laminated sands).

The Complex Glacial Environment

Consider the wide range of environments found around glaciers (Figures 1(a), 1(b), and 2). Meltwater emerges from under the ice as large rivers that are often ponded in front of (or even under) glaciers to form lakes in which large fans and deltas are built. Seasonal and long-term variations in the position of the ice front give rise to complex diachronous (time-transgressive) patterns of sedimentation and partial destruction, and reworking of glacial facies either by melt streams or bulldozing by the ice front. In the absence of vegetation and the presence of strong downglacier (katabatic) winds, eolian transport and deposition of fine sediment are important. Steep slopes and melting ice result in mass movement processes. In marine environments where glacial sediment was deposited under water, crustal rebound may lift sedimentary successions either above water or into energetic shallow waters where they will be reworked and redeposited.

Typical glaciogenic sediment is poorly sorted and uniquely fingerprinted by the presence of striated and glacially shaped pebbles or boulders. The greatest yields of such sediment (mostly in the form of silt, so-called rock flour) are from temperate wet-based valley glaciers that are able to slide rapidly across their relatively steep beds (e.g., Hooke and Elverhoi, 1996). These glaciers are capable of eroding their beds to form deep basins often enclosed by rock bars. This creates a typical up-and-down and locally overdeepened long profile (thalweg) in glaciated valleys; glaciated basins often exhibit parts that have been eroded well below sea level (e.g., Great Lake basins in North America). The lower seaward end of glaciated valleys forms narrow, steep-walled, and very deep troughs or fjords.

Glaciers in areas of permafrost (dry-based polar glaciers) are frozen to their beds and move slowly by internal creep and so do not generate a thick depositional record. Ice streams form where parts of polar ice sheets can slip on wet sediment (Bennett, 2003). Subpolar glaciers share characteristics of both wet- and dry-based glaciers. While useful in terms of understanding modern glacial processes and deposits, a thermal regime cannot easily be inferred from ancient glacial lithofacies. A polar thermal regime is transitory; at the end of any one glaciation, dry-based glaciers eventually warm and melt. Polar conditions predominate today in Antarctica but ancient deposits preserved along the continental margin record sediment delivered by wet-based glaciers (e.g., Eyles et al., 2001).

Diamict Versus Till

Diamict (diamictite when a rock) is a nongenetic umbrella term for any poorly sorted deposit regardless of depositional environment (Figures 1(c), 1(d), and 3(a)). Man-made concrete is a diamictite. Sliding glaciers produce a wide range of diamicts that contain admixtures of most textural classes ranging from clay to large boulders. These diamict facies are specifically identified as till because they were deposited directly by ice. A till (tillite: rock) is a specific genetic term for a diamict deposited directly from ice either terrestrially on land or in the glaciomarine realm.

Diamicts form in a wide range of glacial and nonglacial environments such as on the slopes of volcanoes, under water, or on land where debris moves downslope as debris flows or in addition as a consequence of meteorite impact (Figure 3(a) and 3(b)). They are said to be a common facies (not universal facies; see above) because they have no unique environmental interpretation. They look very similar regardless of their origin. It is by the study of other facies with which they are conformably interbedded that valuable information regarding depositional setting is gained. Walther's law states that facies found in a continuous vertical sequence must have been spatially related at the time. If, on the other hand, facies are separated

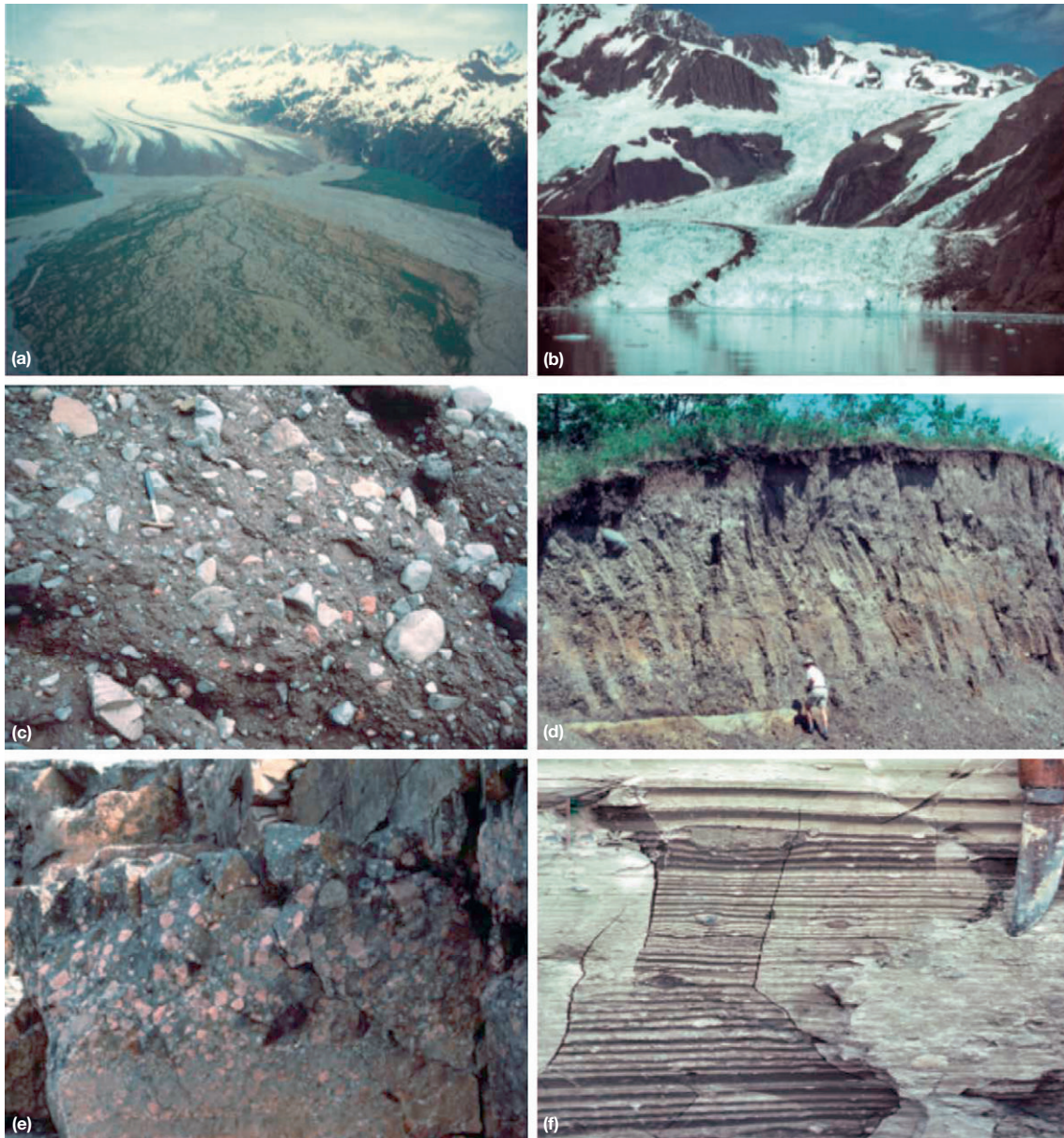


Figure 1 (a) Many processes in areas of wet-based glaciation are dominated not by 'glacial' processes per se but by the activity of meltwater. Here is an extensive braided outwash plain in southern Alaska where meltwaters leaving the ice front have reworked almost all primary glacial sediment and landforms. (b) A calving tidewater ice margin (see [Figure 9](#)). (c) Diamict deposited by debris flow processes. (d) Subglacially deposited deformation till ([Figures 4](#) and [6](#)). (e) Diamict deposited underwater by downslope flow; note systematic coarsening of pebble sizes upward (known as inverse grading developed where debris flows become transitional to turbidites; [Figure 3\(b\)](#)). (f) Classic varves composed of couplets of summer deposited silt (white laminae) and dark winter mud. Numerous white-colored granules of silt (silt clasts) were brought in during summer underflows ([Figure 7](#)).

by disconformities or unconformity surfaces, then the facies above or below a diamict can provide little information. The application of Walther's law yields some indication of the overall depositional environment (whether it was deposited in deep or shallow water; or on a slope; or near a volcano or an impact crater; or onland, in a terrestrial setting, under a glacier perhaps?). A diamict conformably interbedded within undisturbed turbidites cannot be a till but is more reasonably interpreted as a debris flow since these are genetically related to turbidity currents. This sort of contextual analysis greatly constrains the possible origin of the diamict(ite).

Conceptually, till is deposited in three ways. Till accumulates by the passive meltout of englacial debris under stagnant dead ice (meltout till), by meltout of debris from flowing ice (akin to smearing peanut butter on toast: lodgement till) or by subglacial mixing of preexisting sediment (akin to smearing peanut butter on soft bread: deformation till; [Figure 4](#)). Most tills found in mid-latitude areas affected by Pleistocene ice sheets are probably deformation tills; subglacial shearing and mixing of preexisting sediment is the most efficient method of making, transporting, and depositing large quantities of till. Deformation till is commonly massive, rests on deformed

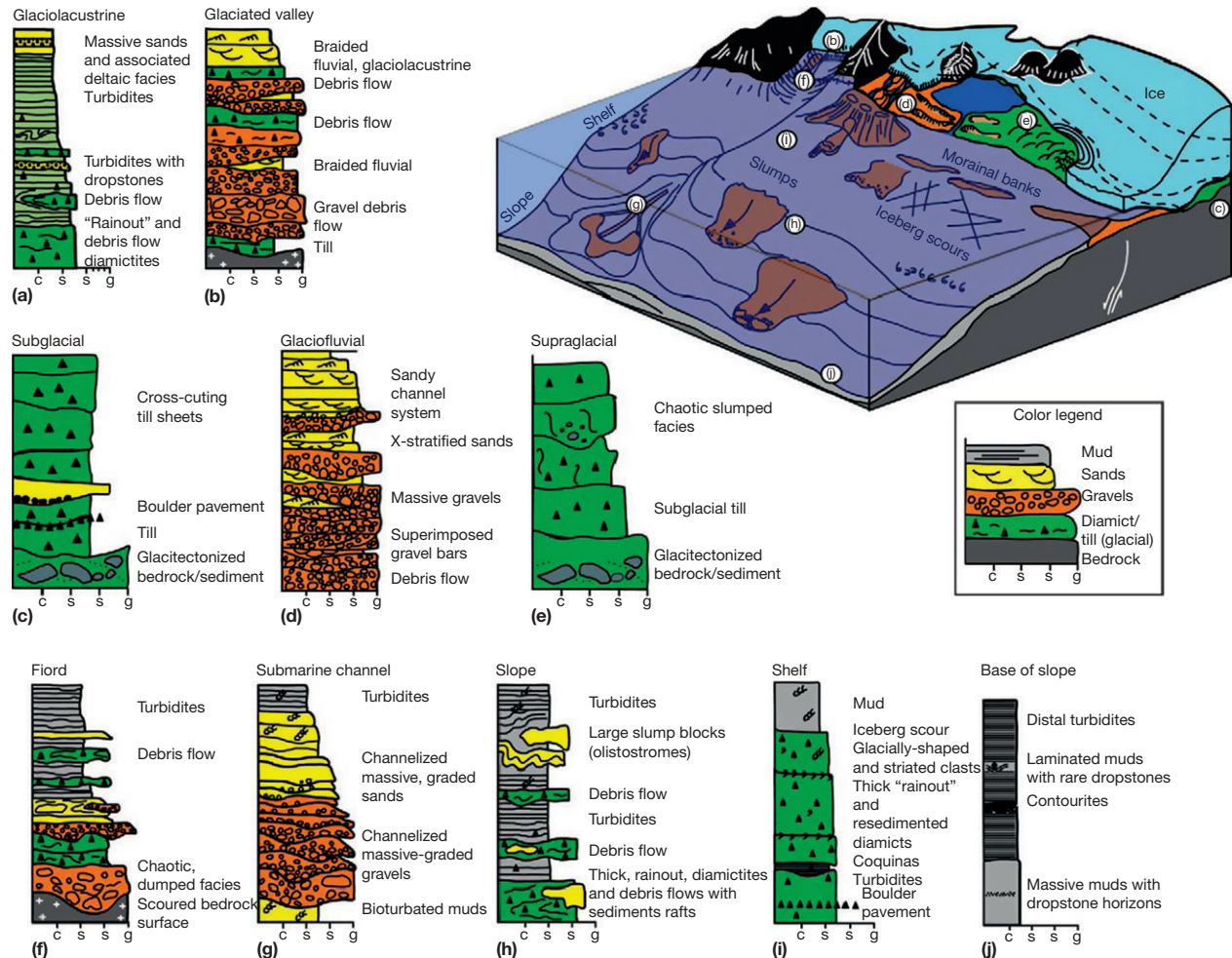


Figure 2 Principal glacial depositional systems associated with large ice sheets that reach sea level. Quaternary glaciations have left a very prominent glacial record on land but most sediment (perhaps as much as 90%) is deposited offshore on continental slopes. Such deposits dominate the record of older glaciations in Earth history because the terrestrial record is easily eroded.

substrate sediments, and usually includes rafts of undigested substrate sediment in its basal parts recording incomplete mixing. Some are held to exhibit lamination and banding similar to gneissic textures in high-grade metamorphic rocks subject to strong shearing. The same facies are produced by debris flows under water, however requiring care in interpretation and consideration of the wider depositional context. At modern glaciers flowing over stiff rock, lodgement till is more important. Meltout will be of local significance where evidence indicates that ice has stagnated and melted *in situ*.

There have been several techniques confidently proposed in the past as a Holy Grail for identifying subvarieties of till. Such methods center on samples examined in the laboratory and include geochemical and grain size characteristics, small-scale structures seen in thin sections, microsurface textures of quartz sand grains, or the orientation of clasts (clast fabric analysis). All of them ultimately proved simplistic because of overlapping properties. In reality, till is deposited by interactions of all the three processes identified above (indeed, processes may change through time at any one site) but one process will be dominant. Volumetrically, most till is

produced by subglacial deformation of other sediments as ice sheets grow and expand outward and cannibalize their own outwash deposits.

Glacial Depositional Systems: Facies Models and Facies Associations

Geologists recognize a wide range of different environments around and under the margins of glaciers, each one creating a distinct depositional system. These systems do not have sharp boundaries but are intimately linked as a consequence of sediment moving from one system to another (Walker, 1992). The system forms a characteristic deposit with a distinct morphology (e.g., a deltaic depositional system) and several recurring systems occur in glacial environments (Figure 2). Terms such as landsystem and landform/sediment association have also been used. One advantage of this approach is that in areas of modern-day and Pleistocene glaciation, it is possible to predict the types of facies found in the subsurface by simply identifying landform types typical of different systems.

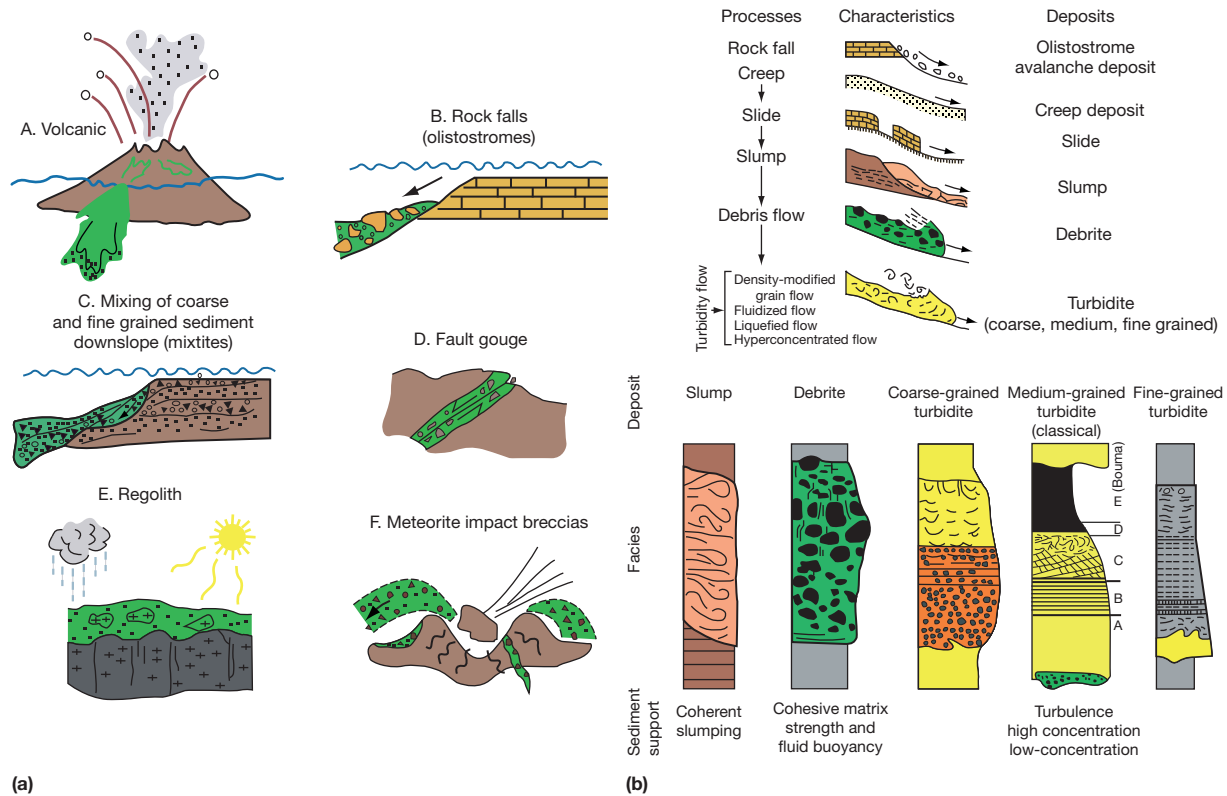


Figure 3 (a) Diamicts form in a wide range of environments. Identifying diamicts deposited directly by glacier ice (till; e.g., [Figure 6](#)) requires analysis of associated facies. (b) Gravity-driven processes and their deposits.

In pre-Pleistocene rocks, however, this cannot be done, and interpretation is reliant on consideration of facies associations.

Facies associations are defined as distinct vertical successions of genetically related lithofacies ([Miall, 1997](#); [Walker, 1992](#)). As emphasized above, it is by recognizing these packages (rather than any one unique facies type) that ancient glacial settings can be recognized. A three-dimensional (3D) block diagram called a facies model is a useful tool for representing depositional systems and their constituent facies associations. These models provide a mental image that can embrace different data derived from studies of outcrops, subsurface drill core, or geophysical data (such as 3D seismic or ground-penetrating radar), and the study of processes and deposits in a controlled laboratory experiment. The model will vary from one particular case to another (such as tectonic setting), but it is a useful tool for distilling large amounts of data, for predicting subsurface conditions and for comparison with stratigraphic successions of unknown origin. Drawing a facies model forces consideration of the 3D distribution of facies.

For each depositional system (and thus facies model), a representative vertical profile can be made showing typical facies and the sedimentary evolution of the system through time ([Figure 5](#)). These profiles suffer from the obvious limitation of depicting 4D conditions in two dimensions, but the detailed assessment of facies in measured outcrops or drill core (called facies logging) is the essential starting point in any analysis. Simple lithofacies codes are available to aid labeling and identification and serve as a reminder of what to look for

in the field. Two-dimensional panels (say of cliff exposures) are very useful in showing lateral variability and 2D geometry of facies (such as those found in channels) and, increasingly, there is a move to consider the 3D architecture of deposits. Much progress has been made in this respect in fluvial, eolian, and deep marine environments (prompted by the search for hydrocarbons and reservoir analysis by petroleum geologists, and aided by geophysical techniques such as 3D seismic). The need to model the movement of groundwater through glacial sediments is a major stimulus ([Boyce and Eyles, 2000](#)).

Terrestrial Glacial Depositional Systems

Subglacial Depositional System

The terrestrial subglacial depositional system is composed of sediments deposited under the ice, and is typically dominated by sheet-like deposits of deformation till ([Figure 6](#)). These tills rest on eroded and deformed remnants of older preexisting sediment. These were glaciotectionized when ice advanced over the sediments and they were remobilized as till. A low-relief landform called a till plain (ground moraine of the older literature) is created. The most distinctive feature of a till plain are bedforms called flutes (long and thin) and drumlins (long and wider) elongated in the direction of ice flow. The origin of these is still contentious but they are increasingly being recognized as a product of either erosion of preexisting sediment (e.g., where they show cores of nontill sediment) or deformation and remobilization of deformation till ([Benn and](#)

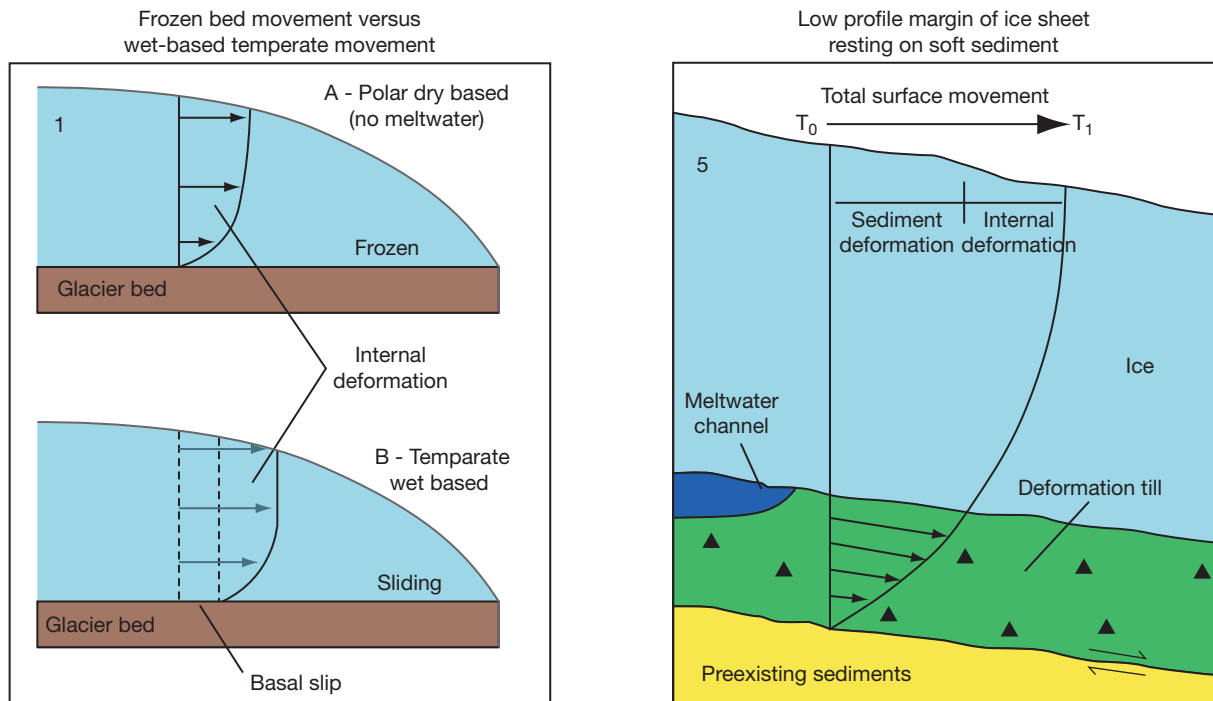
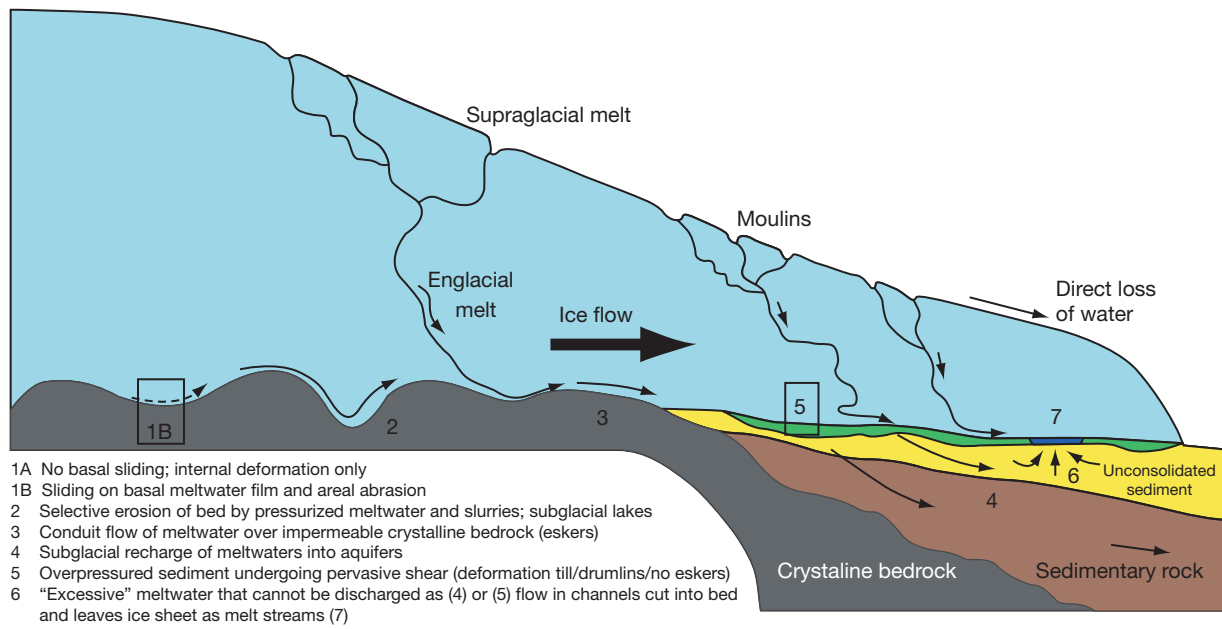


Figure 4 Summary diagram of processes operating below ice of contrasting thermal regime. The generation and movement of till as a deforming layer (5) occurs below the outer margins of a wet-based ice sheet resting on sediment. Most modern glaciers are sliding on rock and are unrepresentative of Quaternary ice sheets.

Evans, 1998; Boulton, 1996; Hart, 1999). Fundamentally, a streamlined till plain is the 'soft bed' of the former glacier, in effect a giant slickenside surface such as along a fault plane veneered by ground-up rock (gouge). End moraines are ridges that form perpendicular to ice-flow direction and mark pauses in ice retreat, creating a ribbed appearance to the till plain. Moraines are built either by the extrusion of soft deforming till from under the ice margin or by bulldozing of sediment by a

readvancing ice front. The gravelly deposits of subglacial rivers are left as sinuous esker ridges (Brennand, 1994).

Till plains may be overlain by extensive spreads of gravelly outwash facies (e.g., Kjaer et al., 2004) and silty glaciolacustrine deposits left from ephemeral lakes ponded against the retreating ice margin. Glacial outwash deposits are dominated by poorly sorted, crudely bedded, sheet-like gravel facies deposited in shallow braided rivers. There is an absence of

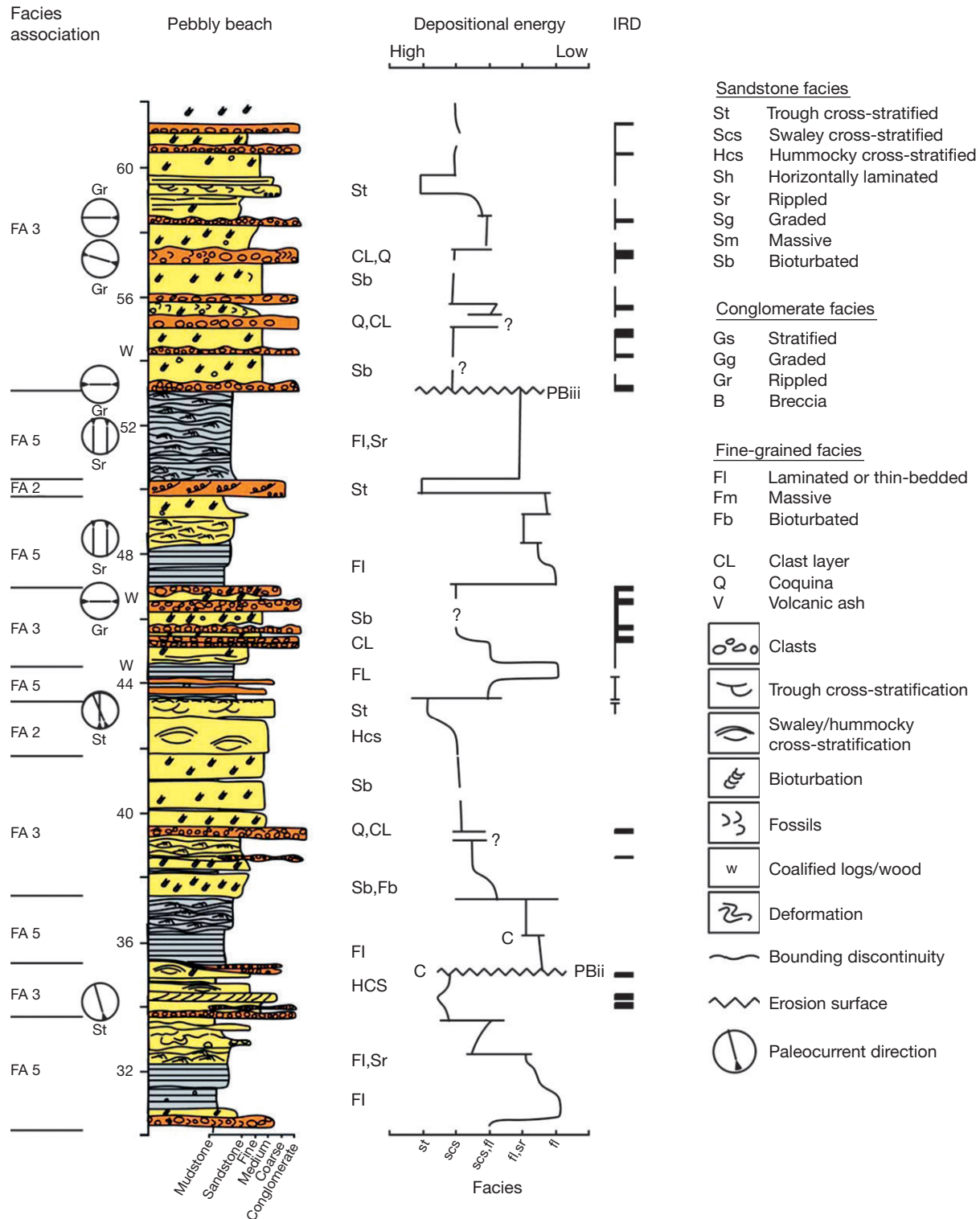


Figure 5 Typical vertical profile showing facies types, paleocurrents (small circles with arrows), and amount of ice-rafted debris (IRD) in a glacially influenced shelf succession (Eyles et al., 1998). Vertical scale is in meters.

large-scale cross-stratified facies because of a lack of deep channels and an abundance of coarse debris (Miall, 1992). Glacial lake facies include distinctive rhythmically laminated silts and clays that record seasonal patterns of sedimentation (varves).

Varved facies dominate glacier-fed lakes (Figure 7). Varve thickness typically decreases upward in measured profiles recording ice retreat. So-called proximal varves may be many meters thick and composed of ripple cross-laminated sands.

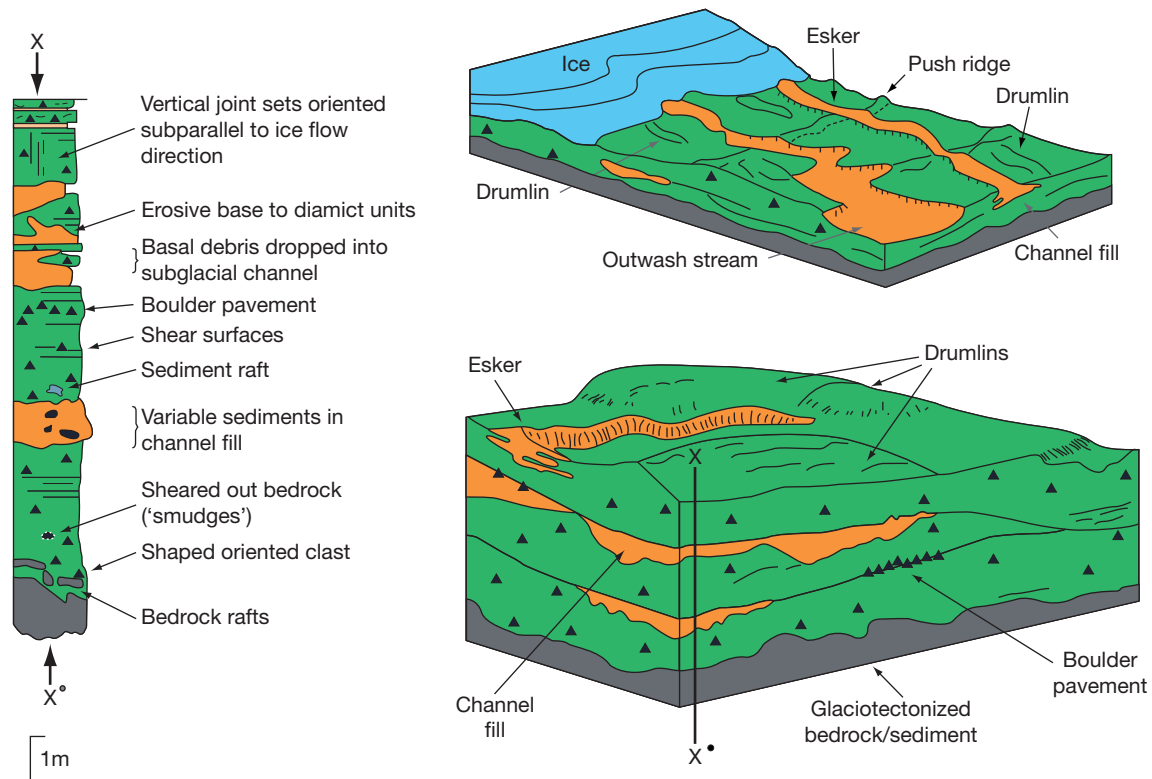


Figure 6 Facies model for subglacial depositional system (Figure 2) involving deposition of successive sheet-like units of deformation till.

Till-like sediments may form in ice-contact glacial lakes where icebergs drop coarse sediment into bottom muds (rainout diamict). These are discriminated from tills proper by consideration of their geometry (typically blanket-like, resting conformably on underlying deposits) and the associated glaciolacustrine facies with which they are conformably interbedded. Local deformation and incision result from the scouring of the lake floor by drifting ice masses such as icebergs (Eyles et al., 2005).

Supraglacial Depositional System

The subglacial depositional system is often partially covered by sediments that were transported on the ice surface (Figure 8). These supraglacial sediments were lowered onto the underlying till plain as the underlying glacier melted. A distinctive hummocky topography marked by craters (kettle holes) forms when buried ice melts and overlying sediment collapses. This topography is common at the margins of valley glaciers that transport substantial quantities of rubble rock fall debris derived from adjacent rock walls. Boulderly supraglacial debris can blanket the immediate ice margin, insulating it from melt and thereby creating a belt of hummocky ice-cored sediments underlain by subglacial sediments and landforms. This is the hallmark of the glaciated valley depositional system. In addition, large volumes of sediment accumulate between the glacier and the valley sidewalls. With ice retreat, lateral moraine ridges left perched on the valley sides are quickly destroyed by mass wasting. Mass flow and meltwater processes rapidly convert the glacial fill of a confined valley into thick accumulations of outwash gravel and sand. The term proglacial has been used in reference to a

short-lived phase of intense reworking and sediment transport by rivers immediately after deglaciation.

There is still much uncertainty regarding the origin of the hummocky terrain left behind in flat mid-continent regions by Pleistocene ice sheets. This distinctive topography extends across huge tracts of the glaciated portion of the mid-continent North American plains underlain by soft Mesozoic rocks. This terrain was initially regarded as a supraglacial deposit akin to that left by valley glaciers, but the process of how thick sediment could accumulate on top of the ice sheet was never clear. The composition of hummocky moraine offered clues; it is composed of the same clay-rich till found in nearby drumlins into which it passes. Indeed, transitional landforms can be recognized where the formerly flat till plain was pressed into hummocky streamlined shapes (humdrums). It is now recognized as the product of the pressing of a soft till substrate, under portions of the ice margin that stagnated (Boone and Eyles, 2001). Here, the extensive development of a soft bed in mid-continent reflects the presence of weak, fine-grained rock. Such topography may identify relatively faster flowing ice streams within the ice sheet (Hart, 1999) moving across a soft bed of overpressured shales. Large rafts of glaciotectionized soft rock were moved below the ice.

Marine Glacial Depositional Systems

The term glaciomarine identifies processes or deposits that occur within a narrow zone in direct contact with the margin of a glacier (or ice sheet) that reaches sea level. The term

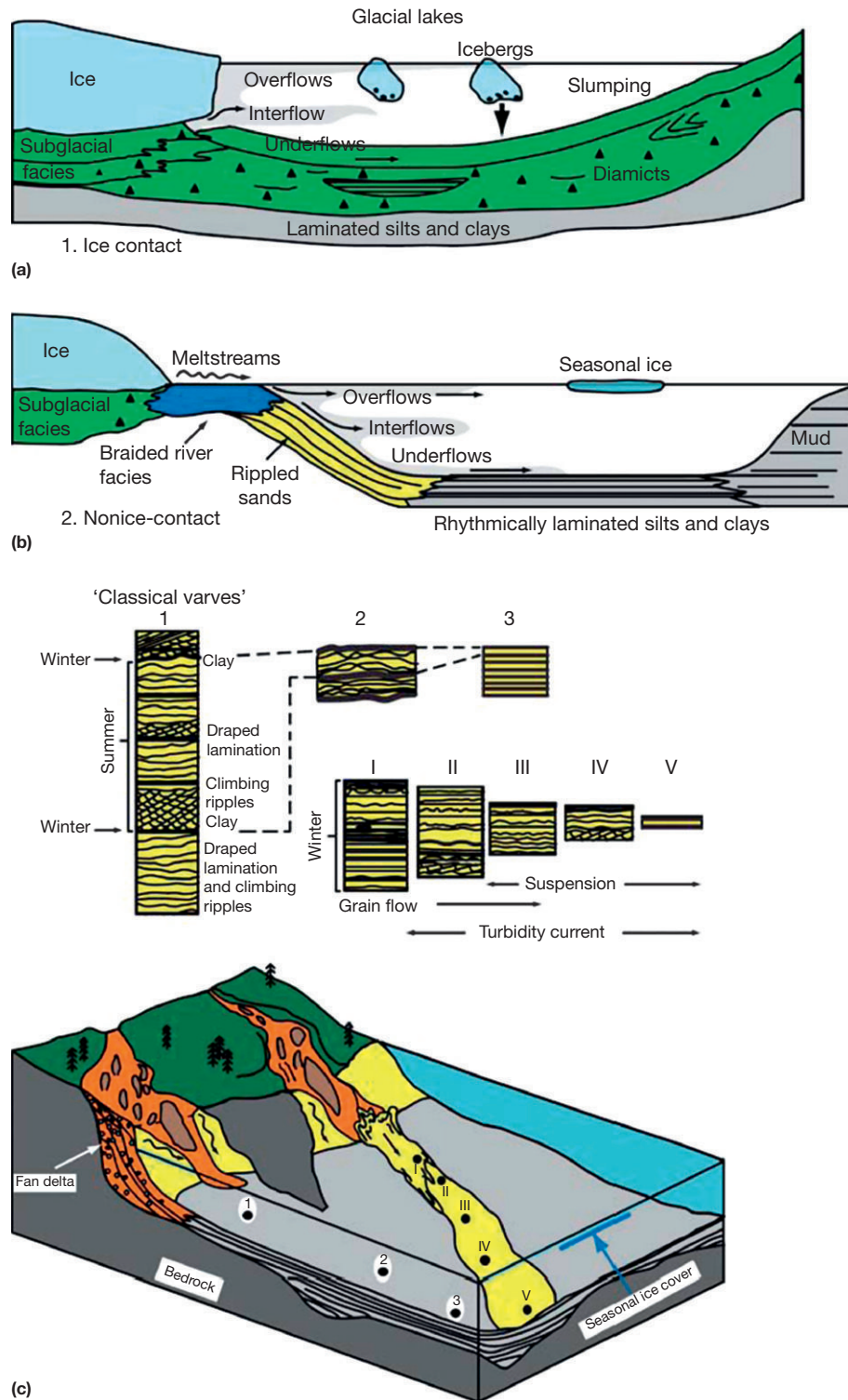


Figure 7 Facies models for glaciolacustrine depositional system (Figure 2) with (a) ice-contact and (b) non-ice-contact types. Typically, the deposits of non-ice-contact lakes are strongly influenced by seasonal variations in the volume of meltwater entering the basin, reflected in varved sedimentation. (c) Slumping on oversteepened delta fronts (I–V) in winter may complicate this.

ice-contact marine or proglacial marine are synonyms for marine settings where glacial influences predominate. Outside this zone and away from direct contact with the ice margin is the glacially influenced marine environment where marine

processes, not glacial, dominate. In areas affected by Quaternary glaciations, glacioisostatic depression and subsequent postglacial uplift of coastal margins have left widespread marine glacial deposits exposed above sea level. Knowledge of

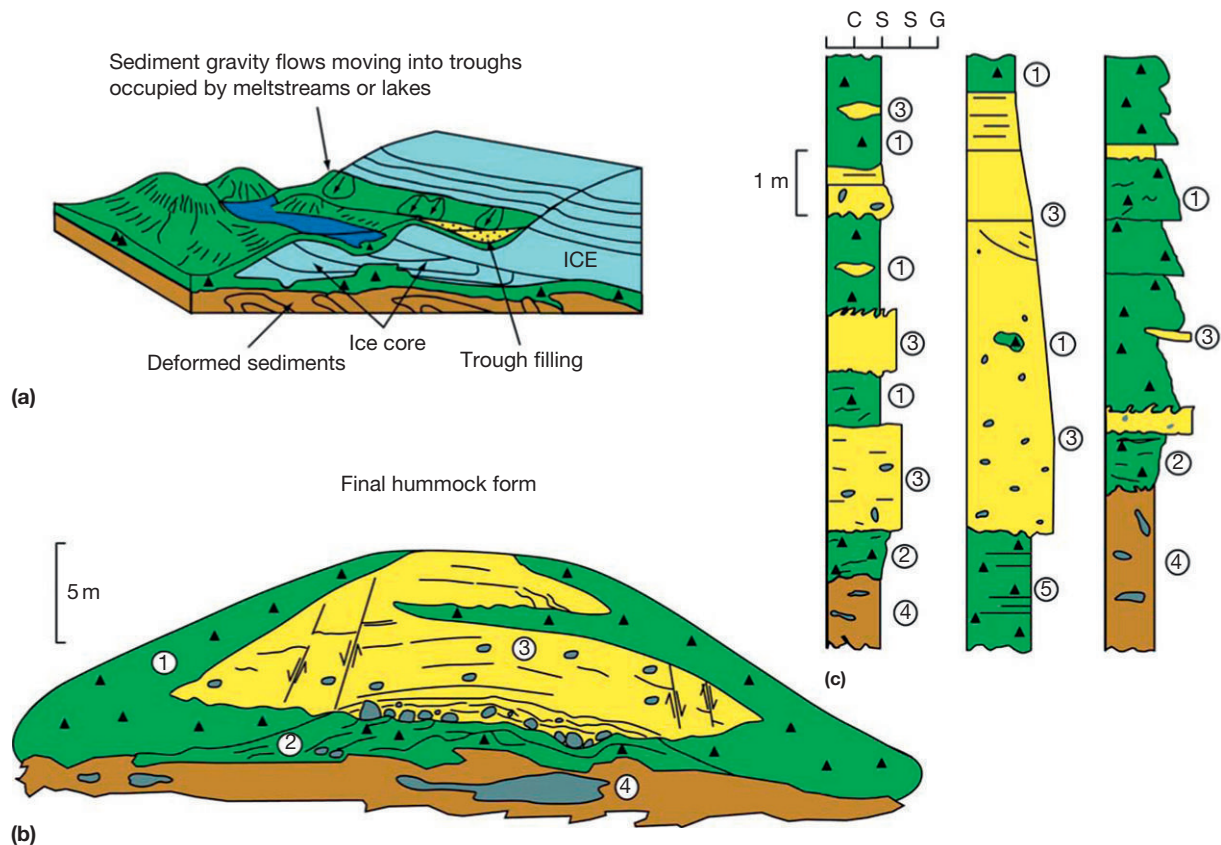


Figure 8 (a) Facies model for supraglacial deposition where ice surface is covered by debris. Melt of buried ice results in widespread slumping and flow of sediment as it is lowered onto subglacial sediments below. Final ice melt produces a resulting in a chaotic hummocky terrain. (b and c) Typical hummock stratigraphy showing: (1) debris flow diamict with rafts of other sediments, (2) diamict melted out from dead ice (meltout till), (3) outwash gravels that accumulated in troughs, (4) glaciotectionically deformed subglacial sediment or rock. Faults and slump structures are common throughout.

marine glacial depositional systems is also key to interpreting the pre-Pleistocene record of glaciation given selective preservation of offshore basinal facies (see below).

Glaciomarine Depositional System

The glaciomarine depositional system occurs where an ice margin terminates in water (Figure 9). When advancing, ice will commonly bulldoze an underwater ridge (morainal bank) in front. This deposit is subsequently overridden to produce thrust sediment and deformation tills (Boulton et al., 1996). A retreating glacier will leave morainal banks on basement highs where ice could ground and stabilize temporarily (e.g., Powell and Domack, 1995). This setting is highly dynamic because the ice margin is partially floating and thus unstable, losing mass by calving icebergs. This environment is dominated by strong jets of meltwaters issuing below water at the ice front, delivering coarse gravel and sand facies deposited on subaqueous meltwater fans (e.g., Hunter et al., 1996). Deposition by turbidity currents results in a wide variety of massive and graded gravel and sand facies found in multistorey cross-cutting channels. Plumes of suspended sediment are also released. These produce blankets of massive mud where mud rains out in quiet water areas; pebbly muds form where

ice-rafted debris is dropped to the seafloor. Tides interact with plumes to produce rhythmically laminated mud facies; seasonal varved deposits may also occur. Ice proximal submarine outwash facies rest unconformably on lowermost subglacially deposited and glaciotectionically deformed facies such as till.

In general, glaciomarine deposits deposited along ice margins in water have a low preservation potential. These accumulations are usually destroyed by shallow marine erosion as a result of glacioisostatic uplift, by iceberg scour, and by downslope mass wasting. Typically, shallow water facies (beach gravels) truncate underlying glaciomarine facies.

Glacially Influenced Marine Depositional System

'Glacially influenced' is an umbrella term for deposits accumulating in the more remote parts of a glaciated basin beyond the direct reach of glaciers (i.e., distal to the glaciomarine realm; Figure 2). The depositional system is fed by glacial sediment that has been extensively reworked by currents or gravity. Areas affected by glacioisostatic movements of the crust created by the weight of the ice sheet, or glacioeustatic sea-level changes (or floating icebergs), are also said to be glacially influenced. The setting necessarily encompasses a much

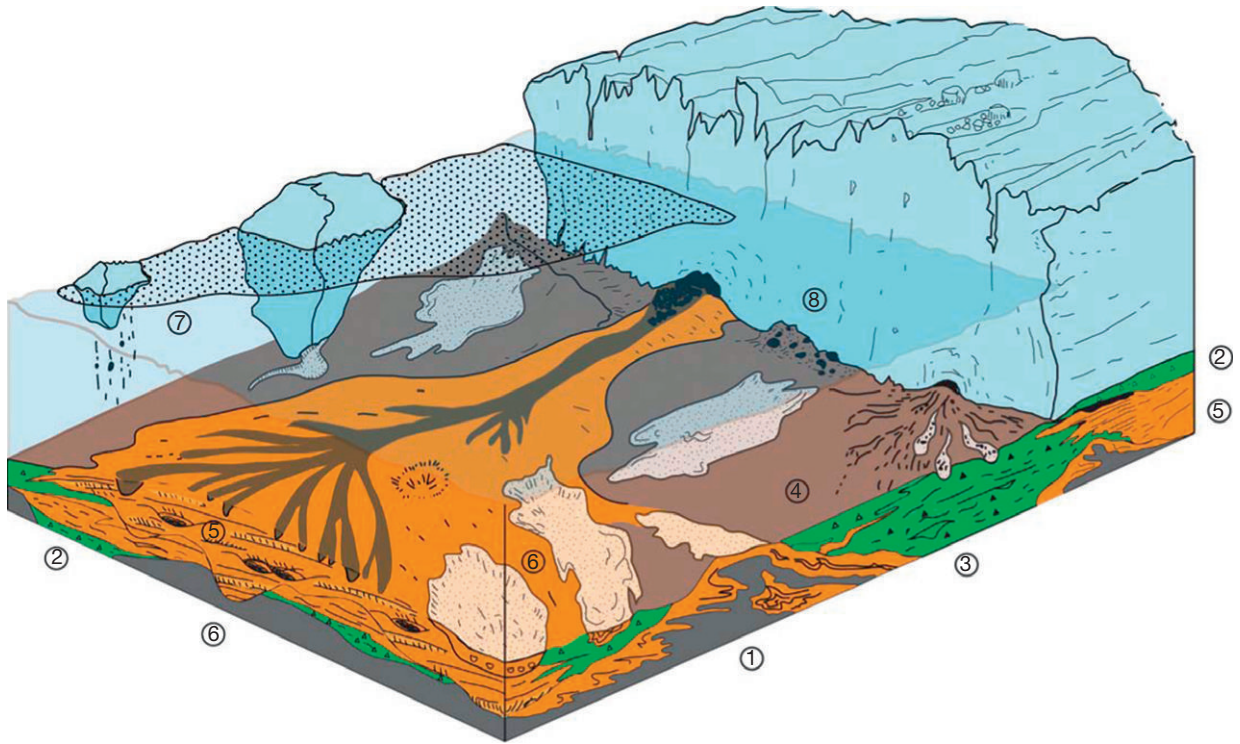


Figure 9 Facies model for glaciomarine depositional system near a tidewater glacier margin (see [Figure 1\(b\)](#)). (1) Glaciotectionized marine sediments or bedrock, (2) till, (3) stratified accumulation of diamict deposited by slumping of till and other debris, (4) mud with ice-rafted debris, (5) channelled gravel and sand of submarine fan, (6) slumps and sediment gravity flow facies ([Figure 3\(b\)](#)), (7) iceberg scour and plume of suspended sediment, (8) push ridge. The entire sediment package commonly takes the form of large ridges (morainal banks) reflecting short-lived pauses in the ice margin as it retreats.

broader range of environments across continental shelves and slopes (e.g., [Anderson, 2002](#); [Eyles et al., 1998](#)). Here, a glacial influence is expressed in the form of rapid changes in water depths brought about by glacioeustasy (the removal of water from the oceans to build ice sheets) and glacioisostasy (the unloading and loading of the lithosphere by removing and returning water from the oceans during glaciations, or direct loading by ice sheets). These two processes act together to create complex spatial and temporal variations in water depths and depositional conditions in different areas of the same sedimentary basin. Detailed examination of glacial facies and disconformities in vertical successions will provide data as to whether water depths were increasing or decreasing ([Miall, 1997](#)). It is dangerous however to interpret these as necessarily reflecting changes in sea level per se; sea level may have remained unchanged with the basin floor itself responding to different loads. For this reason, changes in water depths evident from the study of stratigraphic successions are best described as relative sea-level changes, not in absolute (eustatic) sea level.

Floating ice masses and the introduction of coarse ice-rafted debris, and 'rainout' of mud from plumes of suspended sediment are other significant distal glacial influences on otherwise normal marine sedimentation processes. Several times during the last 100 ka, huge armadas of icebergs were released from rapidly calving margins of Northern Hemisphere ice sheets and are recorded as distinct ice-rafted horizons in mud ([Andrews, 1998](#)); coarser layers of ice-rafted facies are known from the ancient record ([Eyles et al., 1998](#)). Blankets of stony mud

subject to repeated winnowing and scour by iceberg keels (ice keel turbation) cover the mid and outer parts of glacially influenced shelves below storm wave base ([Woodworth-Lynas and Dowdeswell, 1994](#)). These substrates host a wide range of infaunal and epifaunal organisms reflected in diverse ichnofacies types ([Eyles and Vossler, 1992](#)). Deep-water muds pass shoreward into inner shelf areas, and interfinger as fairweather deposits within thick storm-influenced and shore-face sands.

Most glacially influenced continental shelves are not major repositories of glacially influenced marine facies. Water depths are too shallow, so their floors are current-swept and most shelves are undergoing uplift or only slow subsidence. This results in the inability to create space for (accommodate) substantial thicknesses of sediment. Large areas were exposed at times of glacioeustatic sea-level lows and scoured by ice sheets or large melt rivers. Typically, shelves are characterized by condensed successions where deposits of only the last glaciation are found resting on major erosion surfaces. Some glacially influenced continental margins (e.g., Antarctica) are slowly subsiding. These show a well-defined subsurface structure (evident on seismic records) of horizontal 'topsets' composed of relict shelf sediments that have aggraded vertically through time and which may include subglacial sediments ([Anderson, 2002](#); [Dowdeswell and O'Cofaigh, 2002](#)).

Along glaciated continental margins, the bulk of glacial sediment accumulates in deep-water basins on the continental slope or at its base. These experience enhanced sediment

supply at times of lowered sea level when rivers or glaciers discharge sediment directly to the shelf edge (e.g., Hooke and Elverhoi, 1996). Sediments pushed across the shelf under glaciers are transported downslope at the shelf edge by slumping and sediment gravity flows and promote progradation of the slope (Anderson, 1999). The outward growth of the slope during successive glaciations produces large foresets of downslope-thickening wedges of sediment clearly evident on deep seismic records.

Using Glacial Lithofacies to Unravel Pre-Quaternary Glacial Climates

Study of modern Quaternary glacial facies has been instrumental in understanding ancient cold climates in Earth history. The oldest sedimentary rock so far known is of 3.8 Ga and provides the first record of running water. In the sedimentary record of ancient cold climates, there are seven episodes of extended glaciation (glacioeras). There are two Precambrian glacioeras (the late Archean–early Proterozoic ca. 2.8–2.3 Ga and the Neoproterozoic ca. 750–600 Ma). These are followed by the three Phanerozoic glacioeras of the end Ordovician (ca. 440 Ma), the late Paleozoic (the longest: ca. 350–250 Ma), and the late Cenozoic glacioera, which is the best documented (but where major uncertainties still remain as to causal mechanisms). This last glacioera was the culmination of global cooling after 55 Ma and commences with the growth of the Antarctic Ice Sheet at about 40 Ma. Large continental ice sheets only appear much later in the Northern Hemisphere (mostly after 3 Ma). Their extent was controlled by Milankovitch astronomical variables creating Pleistocene glaciations and interglacials. Mountain glaciers probably existed for much of Earth's history but left no record, and the length of apparent nonglacial intervals without significant ice covers (such as the mid-Proterozoic) awaits explanation. The removal of water from the oceans during Ordovician glaciations is considered by some to be the main driver behind extinction events in marine biota (e.g., Brenchley et al., 2003).

The pre-Pleistocene glacial record is fertile ground for research because much of it remains to be investigated using detailed facies analysis. In general, glacial terrestrial depositional models are of limited use in examining this record because glacially influenced marine facies dominate Earth's ancient (pre-Pleistocene) glacial record (e.g., Allen et al., 2004; Eyles, 1993). This simply reflects the selective preservation of offshore basinal facies. The origins of ancient glaciations result from the interplay of tectonics creating high ground (tectonotopography), principally by rifting of large plates, the changing paleogeography and oceanography of the planet arising from rifting and plate migrations, variations in the solar flux, astronomical Milankovitch variables, and reductions in atmospheric CO₂ created by weathering of sediment. One school of thought argues that Precambrian and Phanerozoic glacioeras were radically different, with older glaciations being of global extent (e.g., Hoffman and Schrag, 2002). This model is the subject of much debate and opposition from those in favor of a uniformitarian approach applying knowledge of modern glacial depositional systems and facies to the reconstruction of ancient conditions (Eyles and Januszczak, 2004).

See also: Varved Lake Sediments. **Glacial Landforms, Sediments:** Glaciomarine Sediments and Ice-Rafted Debris; **Tills. Quaternary Stratigraphy:** Morphostratigraphy/Allostratigraphy.

References

- Allen PA, Leather J, and Brasier MD (2004) The Neoproterozoic Fiq glaciation and its aftermath: Huqf Supergroup of Oman. *Basin Research* 16: 507–535.
- Anderson JB (2002) *Antarctic Marine Geology*, p. 330. Cambridge: Cambridge University Press.
- Andrews JT (1998) Abrupt changes (Heinrich events) in late Quaternary North Atlantic marine environments: A history and review of data and concepts. *Journal of Quaternary Science* 13: 3016.
- Benn DI and Evans DJA (1998) *Glaciers and Glaciation*, p. 734. London: Arnold.
- Bennett MR (2003) Ice streams as arteries of an ice sheet: Their mechanics, stability and significance. *Earth-Science Reviews* 61: 309–359.
- Boone S and Eyles N (2001) A geotechnical model for the formation of hummocky moraine by till deformation below stagnant ice. *Geomorphology* 38: 109–124.
- Boulton GS (1996) Theory of glacial erosion, transport and deposition as a consequence of subglacial sediment deformation. *Journal of Glaciology* 42: 43–62.
- Boulton GS, Van der Meer JJ, Hart J, et al. (1996) Moraine emplacement in a deforming bed surge – An example from a marine environment. *Quaternary Science Reviews* 15: 961–987.
- Boyce J and Eyles N (2000) Architectural element analysis applied to glacial deposits: anatomy of a till sheet near Toronto, Canada. *Geological Society of America Bulletin* 112: 98–118.
- Brenchley PJ, Carden GA, Hints L, et al. (2003) High-resolution isotope stratigraphy of Late Ordovician sequences: constraints on the timing of bio-events and environmental changes associated with mass extinction and glaciation. *Geological Society of America Bulletin* 115: 89–104.
- Brennard TA (1994) Macroforms, large landforms and rhythmic sedimentary sequences in subglacial eskers. *Sedimentary Geology* 91: 9–55.
- Dowdeswell JA and O'Cofaigh C (eds.) (2002) *Glacier-Influenced Sedimentation on High-Latitude Continental Margins*, p. 310. London: Geological Society of London. Special Volume.
- Eyles N (1993) Earth's glacial record and its tectonic setting. *Earth-Science Reviews* 35: 1–248.
- Eyles N, Daniels J, Osterman L, and Januszczak N (2001) Ocean Drilling Program Leg 178 (Antarctic Peninsula): Insights in to the sedimentology of continental shelf topsets and foresets. *Marine Geology* 178: 135–156.
- Eyles CH, Eyles N, and Gostin V (1998) Facies and allostratigraphy of high latitude glacially influenced marine strata of the Early Permian southern Sydney Basin, Australia. *Sedimentology* 45: 121–163.
- Eyles N, Eyles CH, and Miall AD (1983) Lithofacies types and vertical profile models; an alternative approach to the description and environmental interpretation of glacial diamict sequences. *Sedimentology* 30: 393–410.
- Eyles N, Eyles CH, Woodsworth-Lynas C, and Randall T (2005) The sedimentary record of drifting ice (early Wisconsin Sunnybrook deposit) in an ancestral ice-dammed Lake Ontario. *Quaternary Research* 63: 171–181.
- Eyles N and Januszczak N (2004) Zipper Rift: Neoproterozoic glaciations and the diachronous break up of Rodinia between 740 and 620 Ma. *Earth-Science Reviews* 65: 1–73.
- Eyles N and Vossler S (1992) Ichnology of a glacially-influenced continental shelf and slope: The Late Cenozoic Gulf of Alaska (Yakataga Formation). *Palaeogeography, Palaeoclimatology, Palaeoecology* 94: 193–221.
- Ghibaudo A (1992) Subaqueous sediment gravity flow deposits: Practical criteria for their field description and classification. *Sedimentology* 39: 423–454.
- Hart JK (1999) Identifying fast ice flow from landform assemblages in the geologic record: A discussion. *Annals of Glaciology* 28: 59–66.
- Hoffman PF and Schrag DP (2002) The Snowball Earth hypothesis: Testing the limits of global change. *Terra Nova* 14: 129–155.
- Hooke RL and Elverhoi A (1996) Sediment flux from a fjord during glacial periods, Isfjorden, Spitsbergen. *Global and Planetary Change* 12: 237–249.
- Hunter LE, Powell RD, and Smith GW (1996) Facies architecture and grounding-line fan processes of morainal banks during the deglaciation of coastal Maine. *Geological Society of America Bulletin* 108: 1022–1038.
- Kjaer KH, Sultan L, Kruger J, and Schomacker A (2004) Architecture and sedimentation of outwash fans in front of Myrdalsjokull ice cap, Iceland. *Sedimentary Geology* 172: 139–163.

- Miall AD (1992) Alluvial models. In: Walker RJ and James NP (eds.) *Facies Models: Response to Sea Level Change*, pp. 119–142. Canada: Geological Association of Canada.
- Miall AD (1997) *The Geology of Stratigraphic Sequences*, p. 437. Berlin: Springer.
- Powell RD and Domack EW (1995) Glaciomarine environments. In: Menzies J (ed.) *Glacial Environments – Processes, Sediments and Landforms*, pp. 445–486. Boston: Butterworth.
- Walker RG (1992) Facies, facies models and modern stratigraphic concepts. In: Walker RG and James NP (eds.) *Facies Models: Response to Sea level Change*, pp. 1–14. Canada: Geological Association of Canada.
- Woodworth-Lynas CMT and Dowdeswell JA (1994) Soft sediment striated surfaces and massive diamicton facies produced by floating ice. In: Deynoux M, Miller JMG, Domack EW, Eyles N, Fairchild IJ, and Young GM (eds.) *Earth's Glacial Record*, pp. 241–259. Cambridge: Cambridge University Press.

Glaciomarine Sediments and Ice-Rafted Debris

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Introduction

The glaciomarine environment refers to that part of the world's seas affected by glaciers and ice sheets, which influence the ocean through the delivery of meltwater, sediment, and icebergs (Figure 1). Today, such environments are only found in high-latitude regions such as Antarctica, Greenland, and Svalbard (Figure 2), but during Quaternary full-glacial periods they were more extensive and many midlatitude oceans were affected by glaciers (Dowdeswell and Scourse, 1990). In general, glaciomarine environments can be characterized by those affected by grounded tidewater glaciers (nonfloating) and those affected by ice shelves (floating) (Anderson and Ashley, 1991). Modern, large ice shelves are only found around Antarctica, and non-floating glacier margins are more common. This pattern appears to have been similar during the Quaternary.

Glaciomarine sediments comprise debris deposited in the marine environment by release from either grounded or floating ice. Such sediments have a greater preservation potential than terrestrial glacial sediments, as the depositional setting of deep-ocean basins situated beyond the ice-sheet margin means that they are less likely to be eroded and reworked during successive glacial advance–retreat cycles. They preserve a sedimentary archive of past environmental change in terms of the waxing and waning of continental ice sheets and associated changes in oceanography and climate. Glaciomarine sediments can be deposited by a variety of processes including the rain-out of ice-rafted debris (IRD), suspension settling of fine-grained sediment from plumes, and a variety of sediment gravity flow processes (Figures 1 and 2). In addition, the effects of postdepositional processes which act to rework glaciomarine sediments, such as ploughing by grounded icebergs (Figure 3) and bioturbation, can also play an important role. The resulting sediments therefore tend to be both texturally and sedimentologically complex.

Glaciomarine Sediments in the Quaternary Record

Depositional Architecture and Sedimentology of Ice-Proximal Settings

Quaternary glaciomarine sediments may be preserved on land above present sea level as a result of glacioisostatic rebound, and they may also be found below present sea level on continental margins where they were deposited during former ice-sheet advances. A variety of glaciomarine landform–sediment assemblages have now been identified in the Quaternary record using sedimentological, geomorphological, and marine geophysical techniques, coupled with reference to modern glaciomarine analogs.

Grounding-line fans

Grounding-line fans (also referred to as 'subaqueous outwash fans') are sediment accumulations deposited where

subglacial meltwater, debouching from an englacial and subglacial tunnel, enters the sea, discharge velocity is rapidly reduced, and sediment is deposited (Powell and Molnia, 1989). The ice-proximal location of grounding-line fans means that they are characterized by texturally and sedimentologically heterogeneous lithofacies. These fans typically have a complex geometry with fine-grained, laminated, and stratified sediments incised by a variety of coarser-grained channelized facies.

Cheel and Rust (1982) have developed a facies model for grounding-line fans based on investigations of late Quaternary glaciomarine sediments (Figure 4). At the fan apex close to the subglacial tunnel mouth, clast-supported, imbricate boulder gravels record deposition from a subaqueous jet emanating from the tunnel (Figure 5). These boulder gravels give way down fan to a series of steep-sided channels containing horizontally stratified sands with discontinuous beds of imbricate gravel, massive sands, and subhorizontally laminated sands. Facies between the channels comprise, proximally, cross-stratified and massive-to-normally graded sands and fine gravels, and, more distally, climbing ripples with occasional silt drapes.

Morainal banks

Morainal banks are subaqueous moraines deposited along grounding lines during intervals of glacier-terminus stability (Figure 6(a)). They are commonly composed of coalescent grounding-line fans (Powell and Molnia, 1989). The elongate morphology of morainal banks reflects their origin by deposition from a series of point sources along the ice-front, as well as by ice-marginal fluctuations that act to bulldoze and squeeze sediment along the grounding line. Like grounding-line fans, lithofacies within late Quaternary morainal banks tend to be sedimentologically and texturally variable, reflecting the dynamic nature of the depositional setting where high ice-proximal sedimentation rates combine with glacier-margin fluctuations to both deposit and rework sediment. This typically results in the formation of thick, stacked sequences of heterogeneous debris ranging from boulder gravels to mud (Figure 6(b)). Sediment gravity flow processes are particularly common within morainal banks. In a detailed sedimentological study of a late Quaternary glaciomarine morainal bank in northeast Ireland, McCabe et al. (1984) showed how the moraine-complex records fan sedimentation against a retreating tidewater glacier margin. Sediments within the morainal bank are dominated by sediment gravity flow facies including massive and variably stratified diamicts deposited as subaqueous debris flows, parallel laminated, and massive sands deposited from density underflows, as well as channelized normal- and inversely graded gravels which record deposition from high-density sediment gravity flows.

Morainal bank size is controlled by the duration of grounding-line stability, debris supply, and ice-proximal

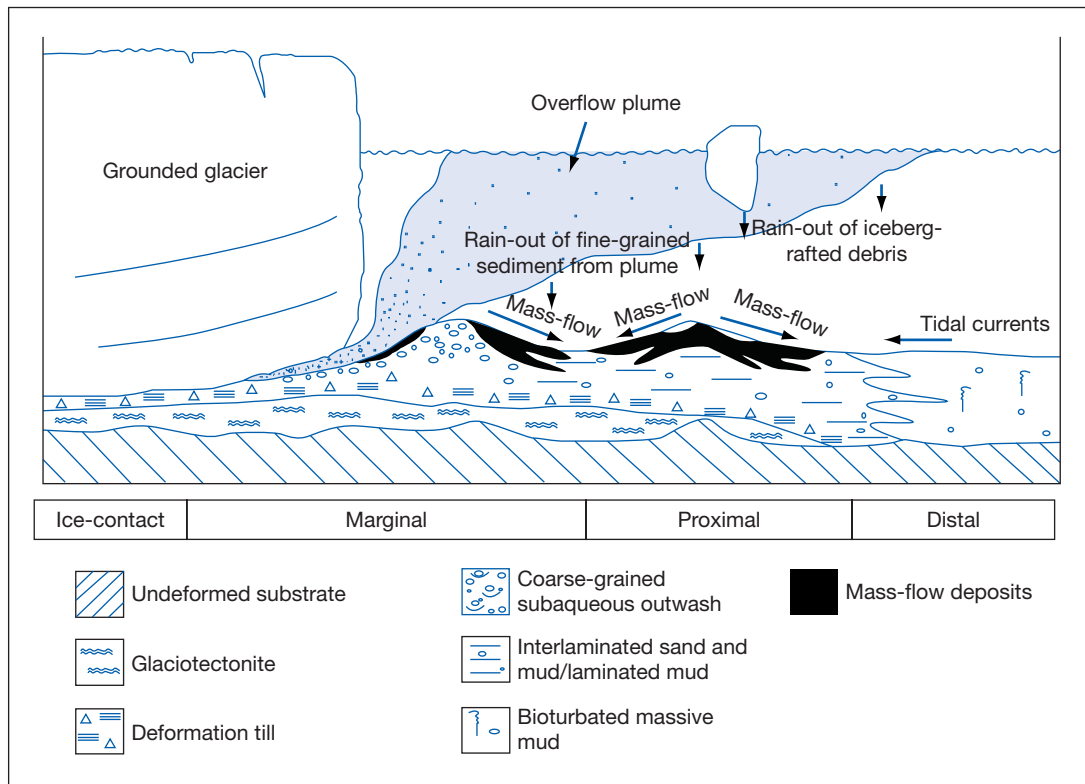


Figure 1 Schematic representation of sedimentation in an ice-contact glaciomarine environment fed by a tidewater glacier (not to scale).

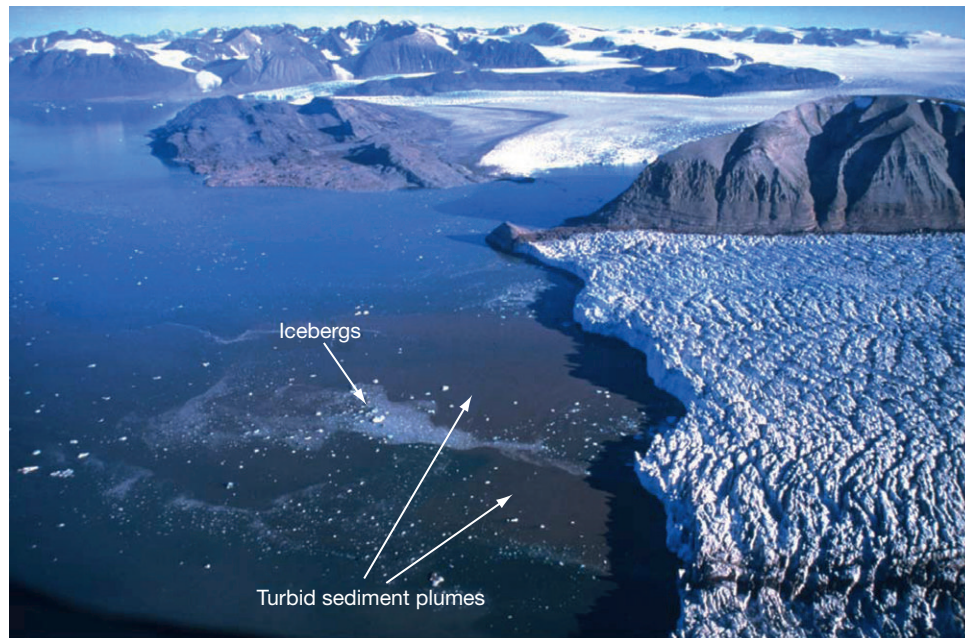


Figure 2 Photograph of suspended sediment plumes emanating from tidewater glacier front, Spitsbergen. Photograph by J.L. Wadham.

sedimentation rates. In the Canadian Arctic, emergent morainal banks from the last glaciation range from diamict veneers that are less than 1 m thick, to over 10 m thick accumulations of mud, sand, and diamict that are the result of the coalescence of grounding-line fans. Sediments include

rhythmically laminated muds deposited by suspension settling from turbid plumes (Figures 1 and 2; Ó Cofaigh and Dowdeswell, 2001), diamict facies ranging from massive tills to weakly graded subaqueous debris-flow deposits, and iceberg-rafted deposits (Figure 6(c) and 6(d)).

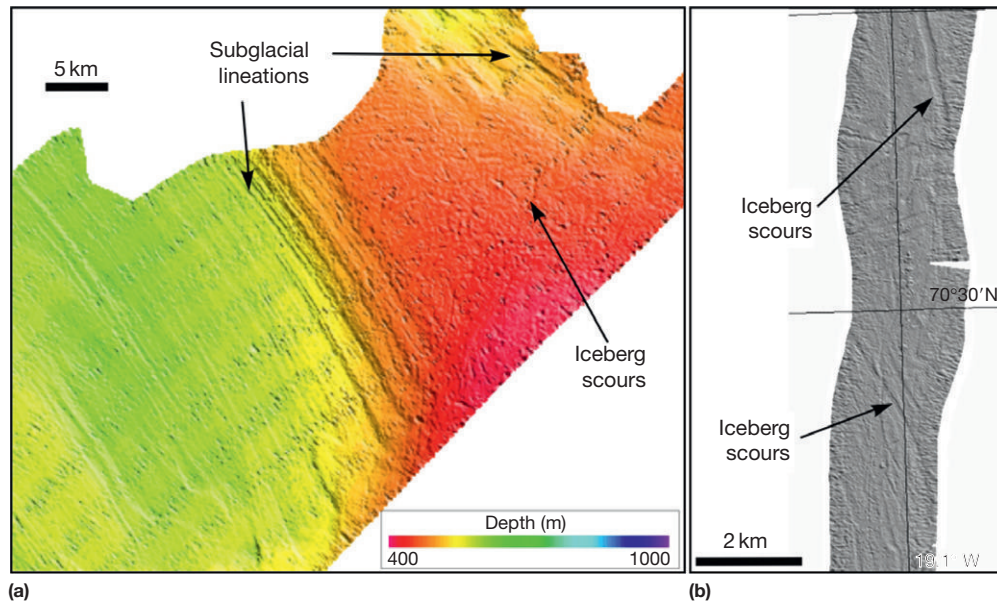


Figure 3 Multibeam swath bathymetric images of iceberg scours on high-latitude continental shelves. (a) Antarctic Peninsula. Modified from Ó Cofaigh C, Dowdeswell JA, Allen CSA, et al. (2005) Flow dynamics and till genesis associated with a marine-based Antarctic paleo-ice stream. *Quaternary Science Reviews* 24: 709–740. Note the irregular pattern of the scours and contrast in form with adjacent streamlined mega-scale glacial lineations. (b) East Greenland. Modified from Ó Cofaigh C, Taylor J, Dowdeswell JA, et al. (2002) Sediment reworking on high-latitude continental margins and its implications for paleoceanographic studies: Insights from the Norwegian-Greenland Sea. In: Dowdeswell JA and Ó Cofaigh C (eds.) *Glacier-Influenced Sedimentation on High-Latitude Continental Margins*, pp. 325–348. London: Geological Society. Special Publication 203.

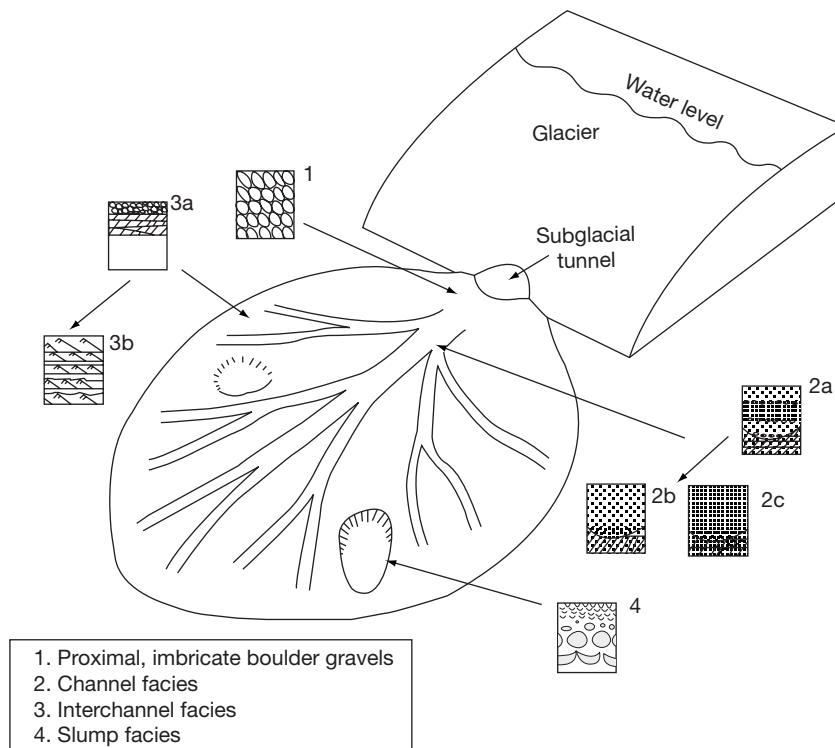


Figure 4 Model of grounding-line fan sedimentation based on Quaternary glaciomarine sediments of the Champlain Sea, Canada. Modified from Cheel RJ and Rust BR (1982) Coarse-grained facies of glaciomarine deposits near Ottawa, Canada. In: Davidson-Arnott R, Nickling W, and Fahey BD (eds.) *Research in Glaciofluvial and Glaciolacustrine Systems*, pp. 279–295. Norwich: Geo Books.



Figure 5 Late Quaternary grounding-line fan sediments, Eastern Ireland. Massive, *a*-axis imbricate, boulder gravels deposited from subaqueous jet, in a proximal grounding-line fan depositional setting.

Deltas

Where sedimentation rates are high, grounding-line fans and morainal banks can aggrade to sea level and form ice-contact deltas (Figure 7(a); Powell, 1990). Such deltas often form isolated flat-topped mounds, located on topographic highs that acted as pinning points during glacier retreat (Figure 7(b)). They are characterized by steep ice-contact slopes and pitted surfaces due to the melt-out of buried ice. Such deltas commonly are Gilbert type with a tripartite internal structure of topsets, foresets, and bottomsets (Figure 7(c)). Topsets are essentially braided river deposits, and consist of subhorizontally bedded, massive gravel. Foreset beds are typically dominated by gravelly and sandy sediment gravity flow facies (Figure 7(d)), with occasional diamict beds deposited by debris flows. They grade downslope into bottomset beds of massive and graded sand and silt (Figure 7(e)). Prodelta distal bottomset beds are composed of more massive finer-grained silts and clays that often contain marine macrofauna (Figure 7(f)). Sediments on ice-proximal slopes typically contain a range of deformation structures related to glaciotectonic disturbance, and the melt-out of buried ice and/or slope collapse associated with glacier retreat.

Grounding-zone wedges

The term grounding-zone wedge (also termed 'till delta' and 'grounding-line wedge') refers to sediment depocenters comprising low-gradient, dipping beds of diamict that are unconformably overlain by horizontal beds of deformation till. Such features are believed to be associated with glaciers that are underlain by a deforming till layer. In the Quaternary record they are being increasingly described in marine geophysical

investigations of high-latitude continental shelves (e.g., Davies et al., 1997). In plan view they appear on geophysical records as sedimentary ridges, orientated transverse to former ice flow. They mark stillstands of the ice-sheet margin during deglaciation and their formation requires that the ice margin was stable for some period of time. Although they are not necessarily associated with floating ice, they are generally believed to mark the location of the transition from grounded to floating ice. Excellent examples of grounding-zone wedges have been imaged on marine geophysical records from Antarctica (e.g., Anderson, 1999; Shipp et al., 2002). There, geophysical profiles show that they are characterized by a back-stepping pattern reflecting progressive ice-sheet retreat (Figure 8). Sediment cores and seismic records suggest that they are composed of poorly sorted diamictic sediment.

Glaciomarine Sedimentation on High-Latitude Continental Margins

High-latitude continental shelves are typically characterized by deep cross-shelf bathymetric troughs separated by intervening shallow banks (Figure 9). These troughs extend across the shelf from fjords, which dissect the mountainous hinterland of adjacent landmasses. Examples of this pattern have been well documented from the Norwegian–Greenland Sea and Antarctica (e.g., Anderson, 1999; Dowdeswell and Ó Cofaigh, 2002; Elverhøi et al., 1998). During the Quaternary, ice sheets expanded across continental shelves through these troughs, in many cases reaching the edge of the continental shelf, and depositing a characteristic suite of sediments and landforms.

Within the troughs, streamlined subglacial bedforms in the form of drumlins, mega-scale glacial lineations and flutings (Figure 10) record the former presence of fast-flowing, ice-sheet outlets (paleo-ice streams) and can be used to infer both the former flow direction and style (streaming in the troughs vs. nonstreaming on adjacent banks; e.g., Canals et al., 2000; Ó Cofaigh et al., 2005). Sediments recording ice-sheet advance on the shelf normally comprise massive overconsolidated till. In the bathymetric troughs, however, these high shear strength tills are characteristically overlain by highly porous, massive tills characterized by low shear strengths, which are the product of subglacial sediment deformation during streaming flow (Dowdeswell et al., 2004). The streamlined subglacial bedforms that record the presence of paleo-ice streams within the troughs are typically composed of these weak tills. These subglacial sediments are overlain by a variety of glaciomarine facies deposited during ice-sheet retreat including grounding-zone deposits in the form of massive diamict and poorly sorted gravel, as well as massive and laminated mud. The precise nature of such sediments will depend on a variety of factors including the rate of retreat and availability of meltwater. Sequences may be capped by organic-rich, open-marine sediments.

Where these ice streams reach the continental shelf edge during successive glacial maxima, large volumes of subglacial sediment are released at the grounding line and delivered to the upper continental slope. This sediment is remobilized downslope under the influence of gravity by debris-flow processes resulting in the formation of low-gradient (in some cases $<1^\circ$) fan-shaped sediment accumulations in front of the trough mouths (Figures 9(b) and 11). These trough-mouth fans are

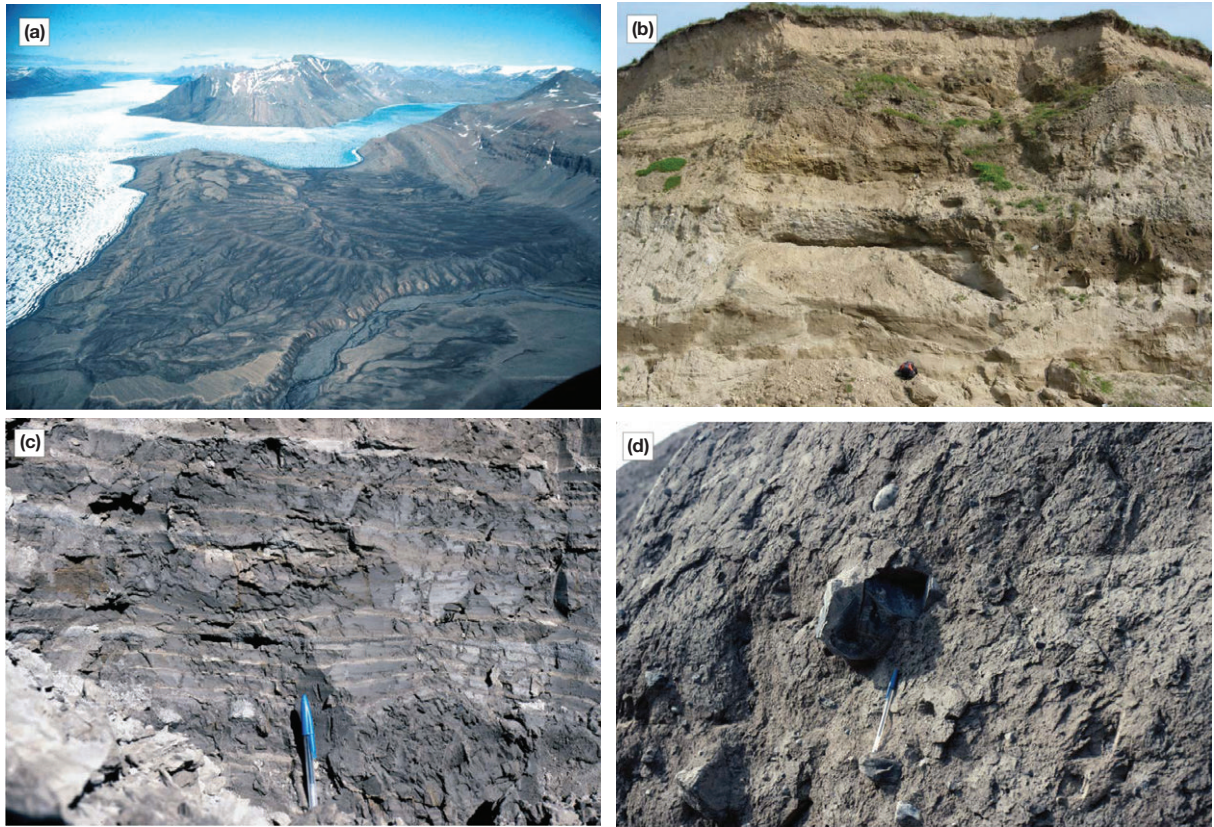


Figure 6 Late Quaternary morainal banks. (a) Arcuate morainal bank inset between fjord side and ice-molded bedrock highs, Ellesmere Island. The morainal bank was deposited by a fjord glacier during retreat from the Last Glacial Maximum. (b) Section in morainal bank, Dundalk Bay, Eastern Ireland. Note the thick sequence of stacked, texturally heterogeneous sediments including diamict, gravel, sand, and silt that is characteristic of the ice proximal, subaqueous depositional setting of morainal banks. (c) Horizontally laminated sand-silt couplets with occasional small dropstones, interpreted as suspension deposits from turbid overflow plumes with background iceberg-rafted debris, Ellesmere Island, Canadian Arctic. (d) Stacked beds of massive, matrix-supported diamict with outsized clasts, deposited by subaqueous cohesive debris flows, Ellesmere Island, Canadian Arctic.

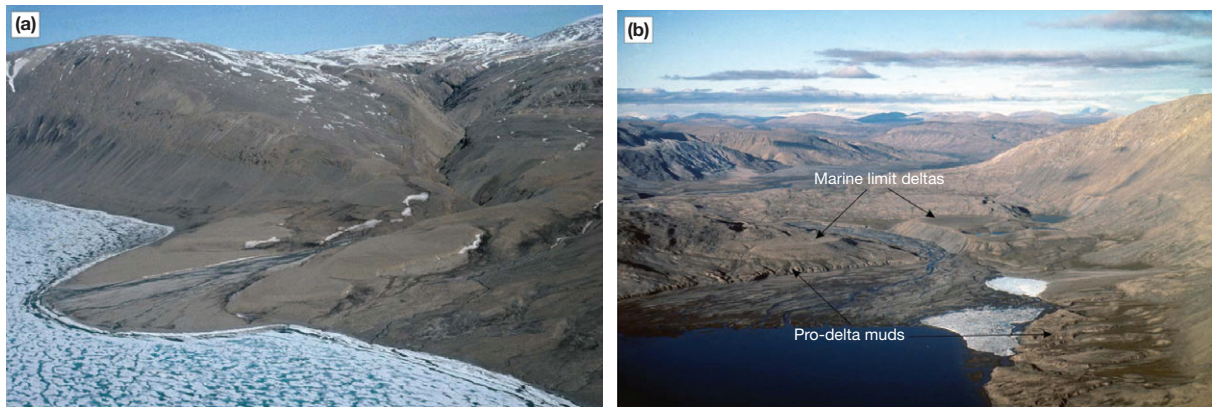


Figure 7 Late Quaternary raised marine, ice-contact deltas and sediments, Canadian Arctic. (a) Terraced, ice-contact fjord delta. (b) Marine limit ice-contact deltas at fjord head, with distal accumulations of pro-delta silty muds. The deltas are banked against ice-molded bedrock highs which acted as pinning points during deglaciation facilitating a stillstand of the retreating glacier margins and delta formation. (c) Internal stratigraphy of ice-contact Gilbert-type raised marine gravel delta. Note the horizontally bedded topset gravels which unconformably overlie foreset beds comprising planar crossbedded gravel. (d) Delta foreset bed composed of normally graded open-work gravels deposited by high-concentration turbidity current. (e) Delta bottomset beds consisting of normally graded sands with silty mud cap. (f) Detail of the sedimentology of distal prodelta muds shown in (b). Massive silty mud with *in situ* paired marine bivalves (arrowed).

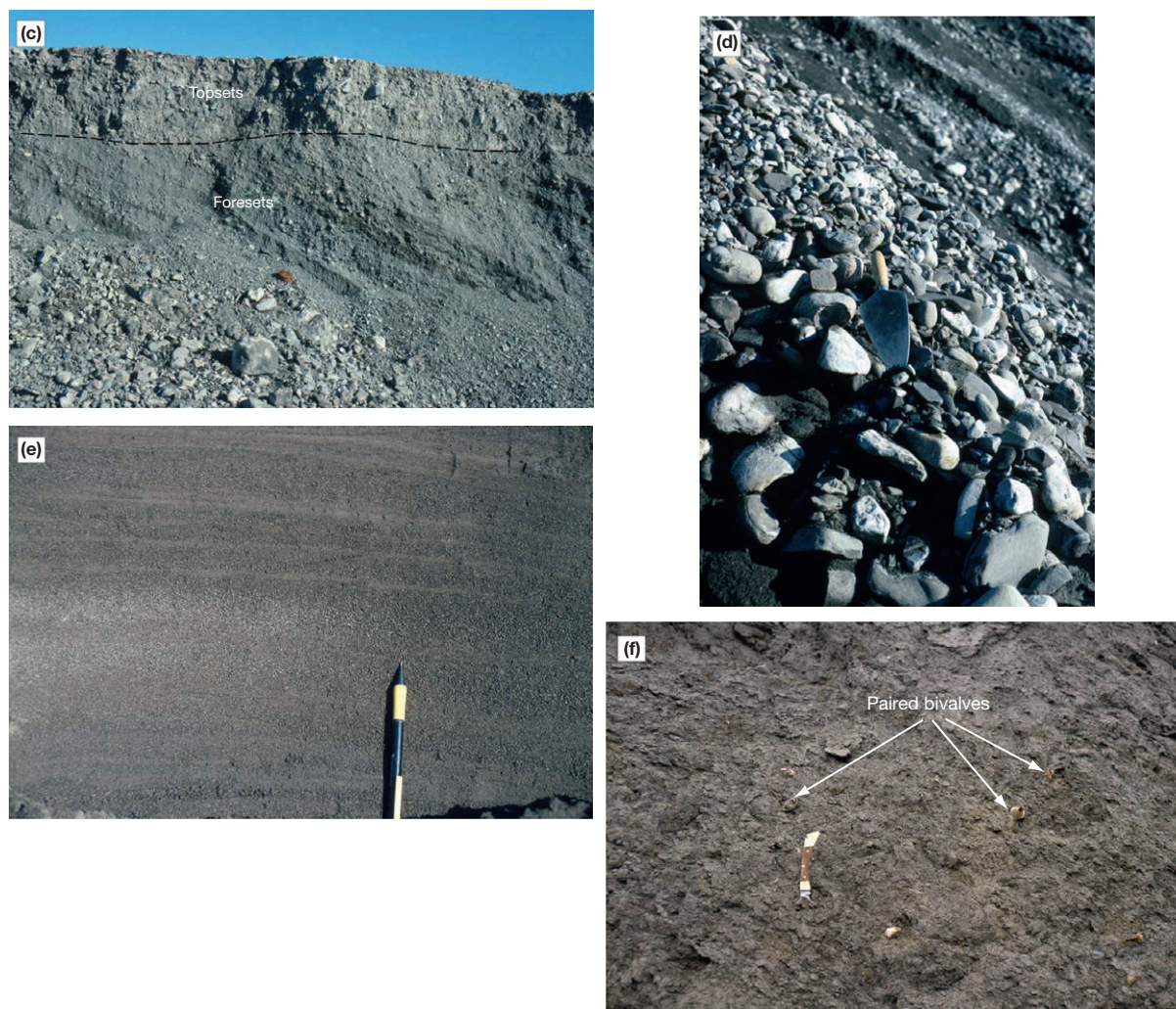


Figure 7 (Continued)

major sedimentary depocenters and are a prominent feature of the Quaternary record of high-latitude continental margins (Dowdeswell et al., 1998; Vorren et al., 1998). Major periods of fan progradation are associated with relatively short intervals when ice streams terminated at the shelf edge during glacial maxima, and their sedimentary record is dominated by stacked debris flows (Figure 11). The debris flows form packages which may be separated by thinner layers of hemipelagic sediments deposited during interglacials or interstadials. Trough-mouth fans therefore contain an archive of past ice-sheet expansion and recession. Adjacent areas of the continental slope can be characterized by sediment slides, although such sediment failures have also been documented on the fans themselves. Slide deposits are often characterized by a blocky chaotic appearance on acoustic records, and associated sediments include turbidites related to emplacement of the slide.

Ice-Rafted Debris

Grain Size of IRD: Definitions and Problems

Transport by floating ice, whether it be icebergs or sea ice, is the dominant mechanism by which glacial sediment is delivered

to ice-distal glaciomarine environments and the deep-ocean basins. The term IRD refers to the material carried on, and within, this floating ice (Figure 12). The more specific terms 'iceberg-rafted debris' and 'sea-ice-rafted debris' differentiate between sediment transported by icebergs and sea ice, respectively. The emphasis here is on debris transported and deposited by icebergs. Initial research using IRD in Quaternary paleoceanographic investigations was mainly focused at Milankovitch timescales. It was not until the identification of the 'Heinrich layers' and recognition that these sediments related to enormous, millennial-scale discharges of icebergs ('Heinrich events') from the last North American ice sheet, that the use of IRD as a proxy by the paleoceanographic community became widespread.

Paleoceanographic studies use one of several grain size envelopes to define IRD. Most investigators use counts of the $>150\ \mu\text{m}$ fraction. Others have focused on the number of visible particles greater than 2 mm in diameter on core X-rays for a specified depth interval (usually 1 or 2 cm). This avoids the potential problem with fine grain sizes, whereby, close to a continental source, delivery of particles $<0.2\ \text{cm}$ in diameter may be by processes other than iceberg rafting. A problem with this, however, is that such coarse-grained material may be rare in

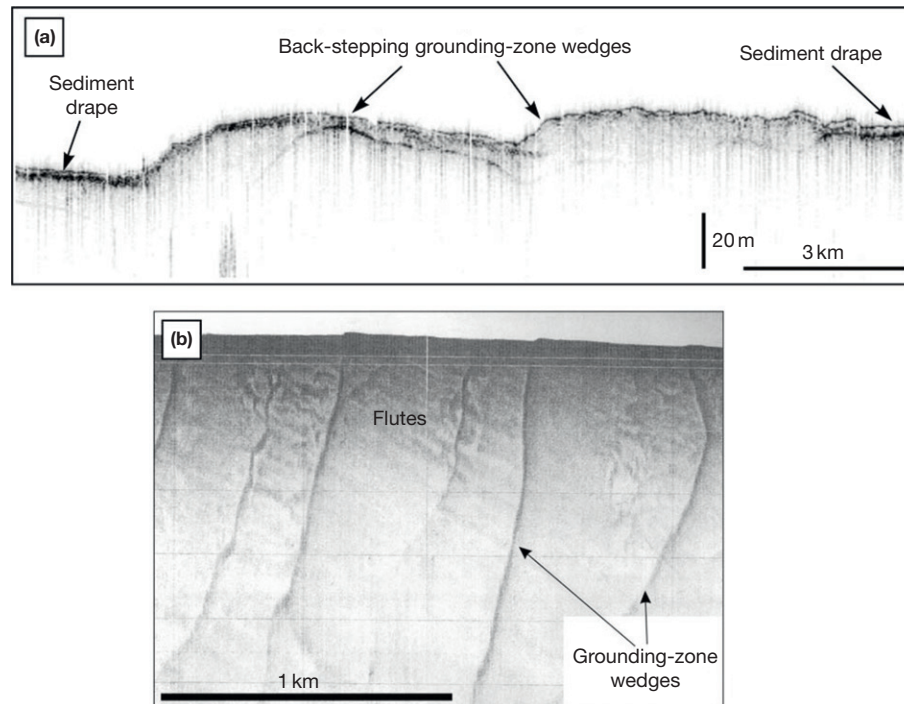


Figure 8 Marine geophysical records of grounding-zone wedges on the continental shelf around Antarctica. (a) TOPAS record of back-stepping grounding-zone wedges deposited during ice-sheet retreat from the Last Glacial Maximum, Marguerite Bay, Antarctic Peninsula. Modified from Ó Cofaigh C, Dowdeswell JA, Allen CSA, et al. (2005) Flow dynamics and till genesis associated with a marine-based Antarctic paleo-ice stream. *Quaternary Science Reviews* 24: 709–740. (b) Side-scan sonar record of back-stepping grounding-zone wedges on the floor of the Ross Sea that were deposited during ice-sheet retreat. Note the presence of flutes between the transverse ridges of the grounding-zone wedges. Modified from Shipp SS, Wellner JS, and Anderson JB (2002) Retreat signature of a polar ice stream: Sub-glacial geomorphic features and sediments from the Ross Sea, Antarctica. In: Dowdeswell JA and Ó Cofaigh C (eds.) *Glacier-Influenced Sedimentation on High-Latitude Continental Margins*, pp. 277–304. London: Geological Society. Special Publication 203.

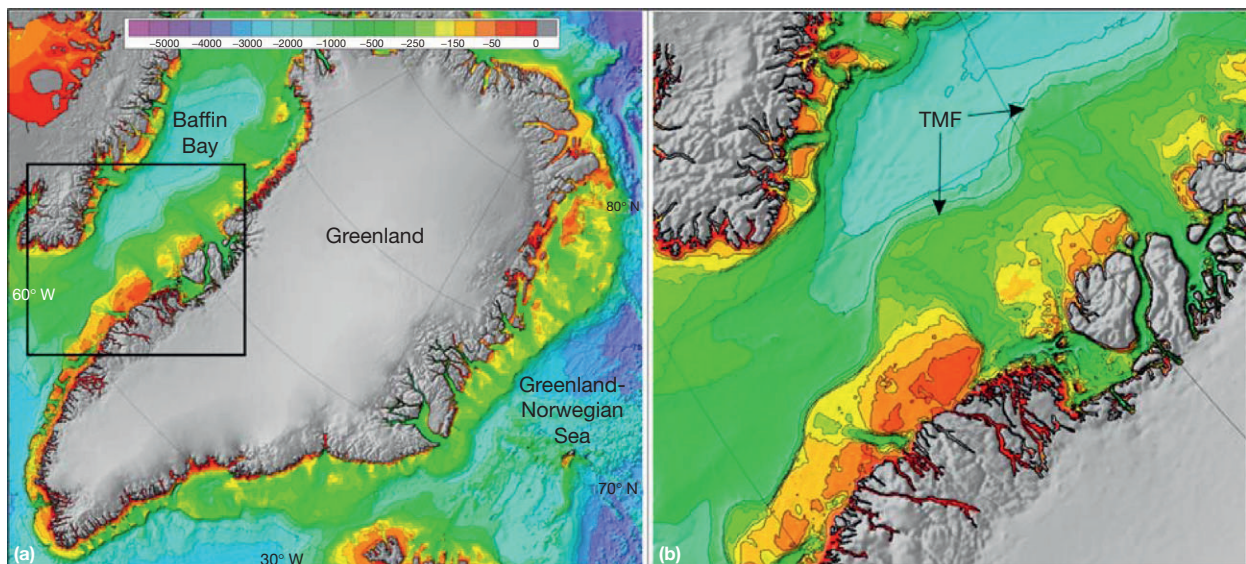


Figure 9 (a) Bathymetry of the seas around Greenland. Note the presence of cross-shelf bathymetric troughs and outbuilding of the submarine contours at the trough mouths recording trough-mouth fans. (b) Blow up to show detail of trough-mouth fans on the west Greenland continental margin (bathymetric data derived from the IBCAO Arctic bathymetry database, Jakobsson et al., 2000).

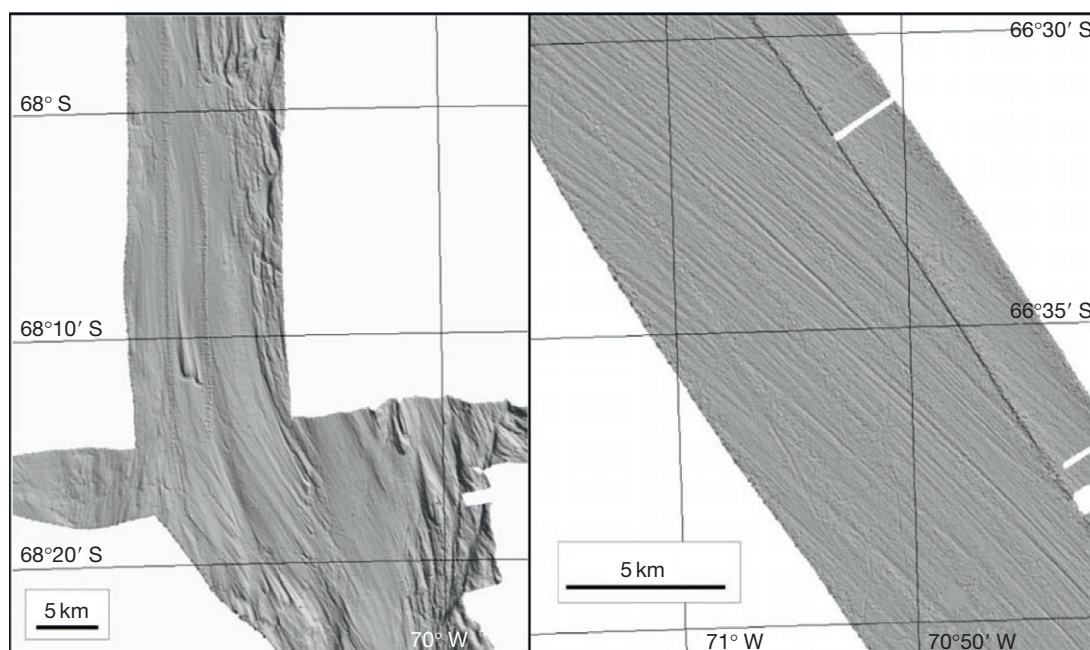


Figure 10 Multibeam swath bathymetry shaded relief images of streamlined subglacial bedforms in a bathymetric trough on the continental shelf of the Antarctic Peninsula. The bedforms record the former flow of a grounded paleo-ice stream at, or immediately following the Last Glacial Maximum. (a) Drumlins and lineations formed in sediment and bedrock. (b) Mega-scale glacial lineations. The lineations have elongation ratios of up to 90:1, a maximum height of 15 m, and widths range from 130 to 300 m. Modified from Ó Cofaigh C, Dowdeswell JA, Allen CSA, et al. (2005) Flow dynamics and till genesis associated with a marine-based Antarctic paleo-ice stream. *Quaternary Science Reviews* 24: 709–740.

the deep sea. Andrews (2000) points out that it is often implicitly assumed in paleoceanographic studies that the relative abundance of IRD is similar for all size fractions from fine-sand through to grains >2 mm across (i.e., a peak in the >2 mm fraction will be matched by a peak in the >250 μm fraction which will be matched by a peak in the >125 μm fraction, and so on). However, as he notes, in reality, this is unlikely, since the 63–2000 μm fraction is least well represented in glacial sediments, whereas the <63 μm fraction is far more abundant.

A further problem is that of distinguishing between sea-ice-rafted debris and iceberg-rafted debris. Limited data suggest that sea-ice-rafted sediment is fine-grained (silt and clay), whereas icebergs are more likely to contain debris of heterogeneous grain size. This reflects the processes of debris entrainment where sea ice predominantly incorporates sediment by eolian processes and freeze-on of turbid seawater. It should be noted however, that anchor ice can incorporate coarse, glacier derived, material in coastal and shallow-marine settings.

Case Study of IRD in the Deep Sea: The Heinrich Layers

Heinrich (1988) first documented the occurrence of six sand-rich IRD layers in deep-sea cores from west of Spain and Portugal. Subsequent studies found similar correlative layers with high carbonate contents in a band approximately coinciding with the IRD belt in the North Atlantic (Andrews and Tedesco, 1992; Bond et al., 1992; Figures 13 and 14). These 'Heinrich layers' record the episodic collapse of the Laurentide Ice Sheet in the Hudson Strait region of eastern Canada on at

least six separate occasions in the last 69000 years, during which enormous numbers of icebergs were delivered into the North Atlantic ('Heinrich events'). The most recent estimate for the duration of Heinrich layer formation is 500 ± 250 years (Hemming, 2004). A Hudson Strait source for the icebergs is supported by the large fraction of detrital carbonate contained within the layers (see Internal characteristics), and the strong distance decay eastward in Heinrich layer thickness from the mouth of Hudson Strait, which is floored by Paleozoic limestone, into the North Atlantic (Dowdeswell et al., 1995; Figures 14 and 15).

Internal characteristics

In the North Atlantic, Heinrich layers are characterized by unusually high concentrations of IRD, low Foraminifera abundance but with the planktonic species *Neogloboquadrina pachyderma* (left-coiling) dominating, sharp lower contacts demonstrating rapid onset of IRD deposition, and detrital carbonate contents of 20–25% (although H6 does not appear to be as carbonate-rich as the other five; Bond et al., 1992; Hemming, 2004; Figure 14). Decreases in planktonic $\delta^{18}\text{O}$ also occur in the layers implying marked reductions in sea surface salinities during their deposition. These characteristics apply to ice-distal Heinrich layers, that is, those deposited in the North Atlantic where the Heinrich layer thickness can be <5 cm. This contrasts with the more ice-proximal regions of the Labrador Sea, close to Hudson Strait. Here, Heinrich layer thickness can be more than 3 m, and detrital carbonate contents are 40–50%. Furthermore, ice-proximal Heinrich layers exhibit considerable sedimentological variability and record a

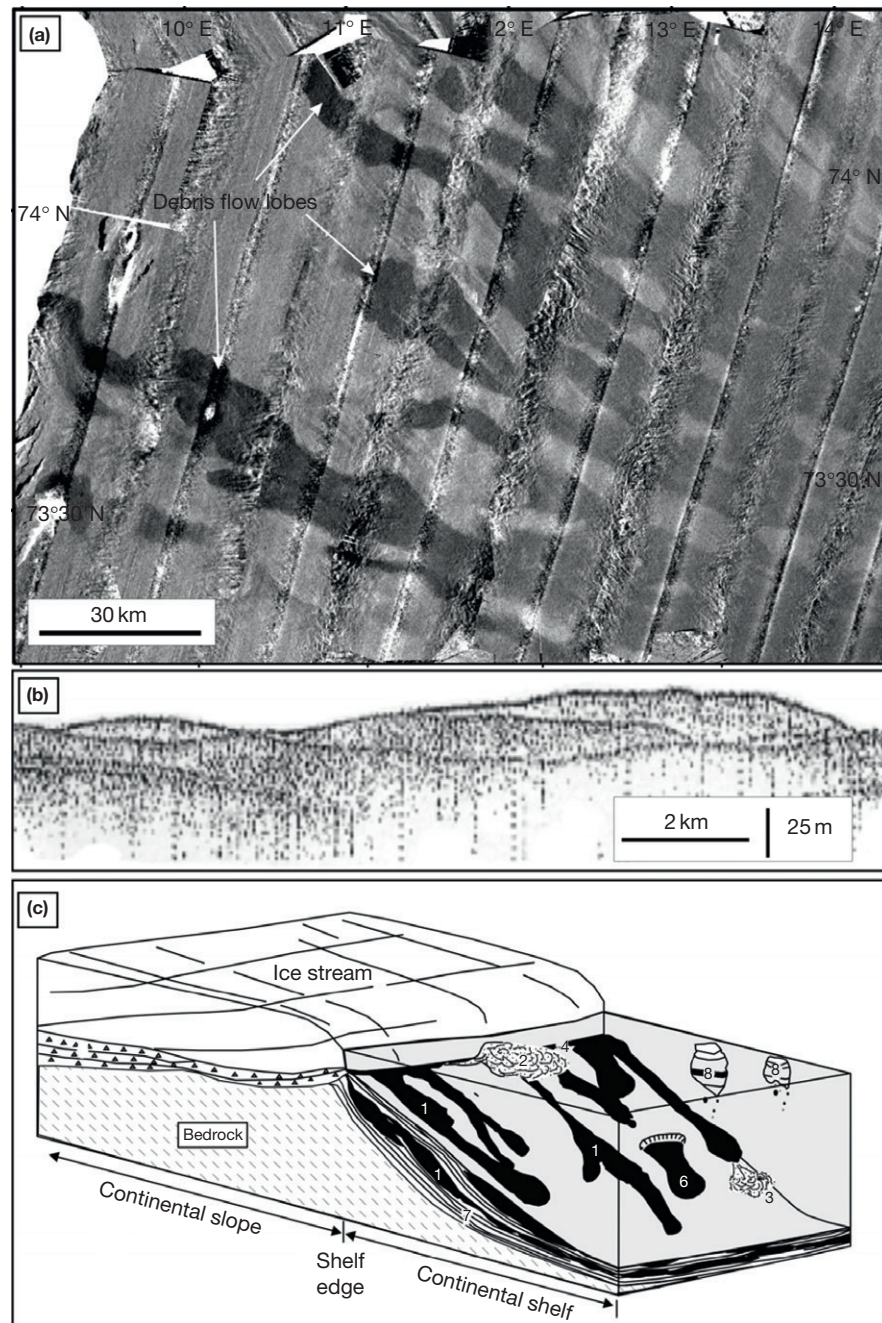


Figure 11 Trough-mouth fan sedimentation on high-latitude continental margins. (a) GLORIA 6.5 kHz side-scan sonar image of debris-flow lobes on the Bear Island Fan. The dark lines running approximately north–south across the image are ship tracks; (b) 3.5 kHz sub-bottom profile through debris flows on the Bear Island Fan, showing the stacked nature of the flows and their convex upper surface. Both (a) and (b) are modified from Dowdeswell JA, Elverhøi A, and Spielhagen R (1998) Glaciomarine sedimentary processes and facies on the Polar North Atlantic margins. *Quaternary Science Reviews* 17: 243–272. (c) Conceptual model for trough-mouth fan deposition in front of an ice stream reaching the continental shelf edge. Debris-flow activity and suspension settling from turbid meltwater plumes dominate sedimentation. (1) Debris flows sourced from glaciogenic sediment at the shelf edge; (2) buoyant turbid meltwater plume; (3) turbidity current formed by downslope evolution from a debris flow; (4) subglacial till extruded along the ice-stream front/grounding line; (5) subglacial (deformation) till; (6) slump generated debris flow; (7) stratified sediments deposited by suspension settling from meltwater plumes, contour-current and minor turbidity-current activity; (8) iceberg rafting.



Figure 12 (a) Debris of heterogeneous grain size on the surface of an iceberg offshore of the Antarctic Peninsula. The largest clasts are about 0.5 m in diameter. Note the subangular to subrounded shape of the clasts which indicates subglacial transport. (b) Core X-radiograph of an IRD layer of gravelly sand contained within weakly laminated, clast-poor mud. Note the sharp upper and lower contacts of the layer.

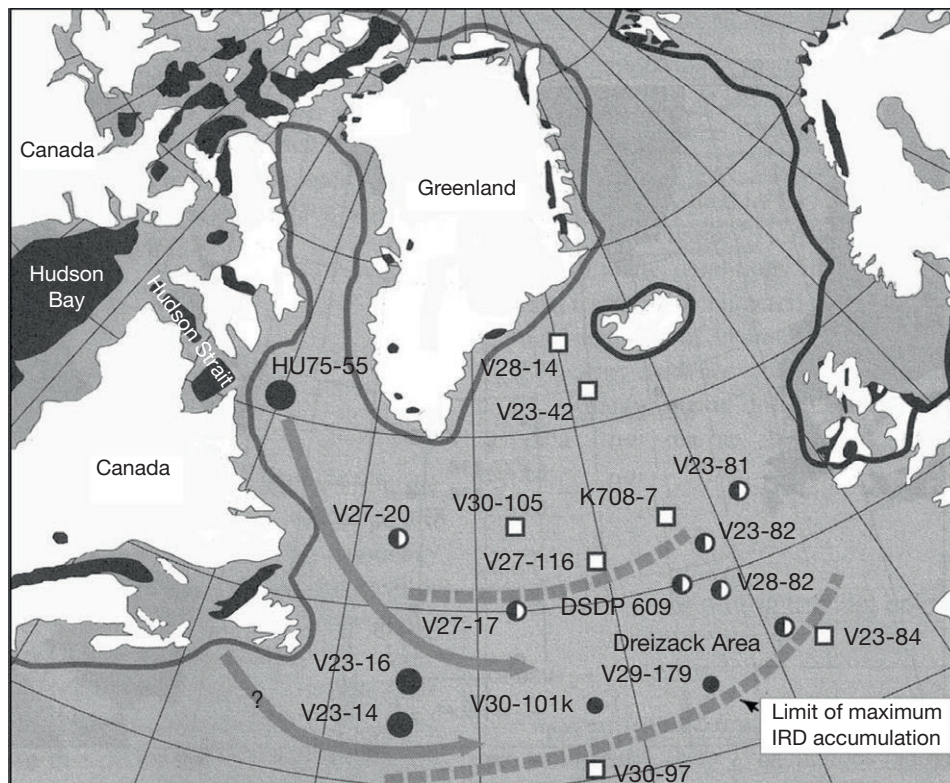


Figure 13 Locations of cores that contain Heinrich layers. Fully black filled circles indicate that carbonate-rich IRD is present in all Heinrich layers identified; half-filled circles indicate that carbonate-rich IRD is absent in some of the Heinrich layers identified; open squares show no carbonate-rich IRD in Heinrich layers. The solid black pattern is the distribution of limestone and dolomite bedrock and the thick solid line is approximate maximum limit of ice sheets during the last glaciation. The westward increase in thicknesses of the Heinrich layers is indicated by the increase in the size of the circles. Modified from Bond G, Heinrich H, Broecker W, et al. (1992) Evidence for massive discharges of icebergs into the North Atlantic during the last glacial period. *Nature* 360: 245–249.

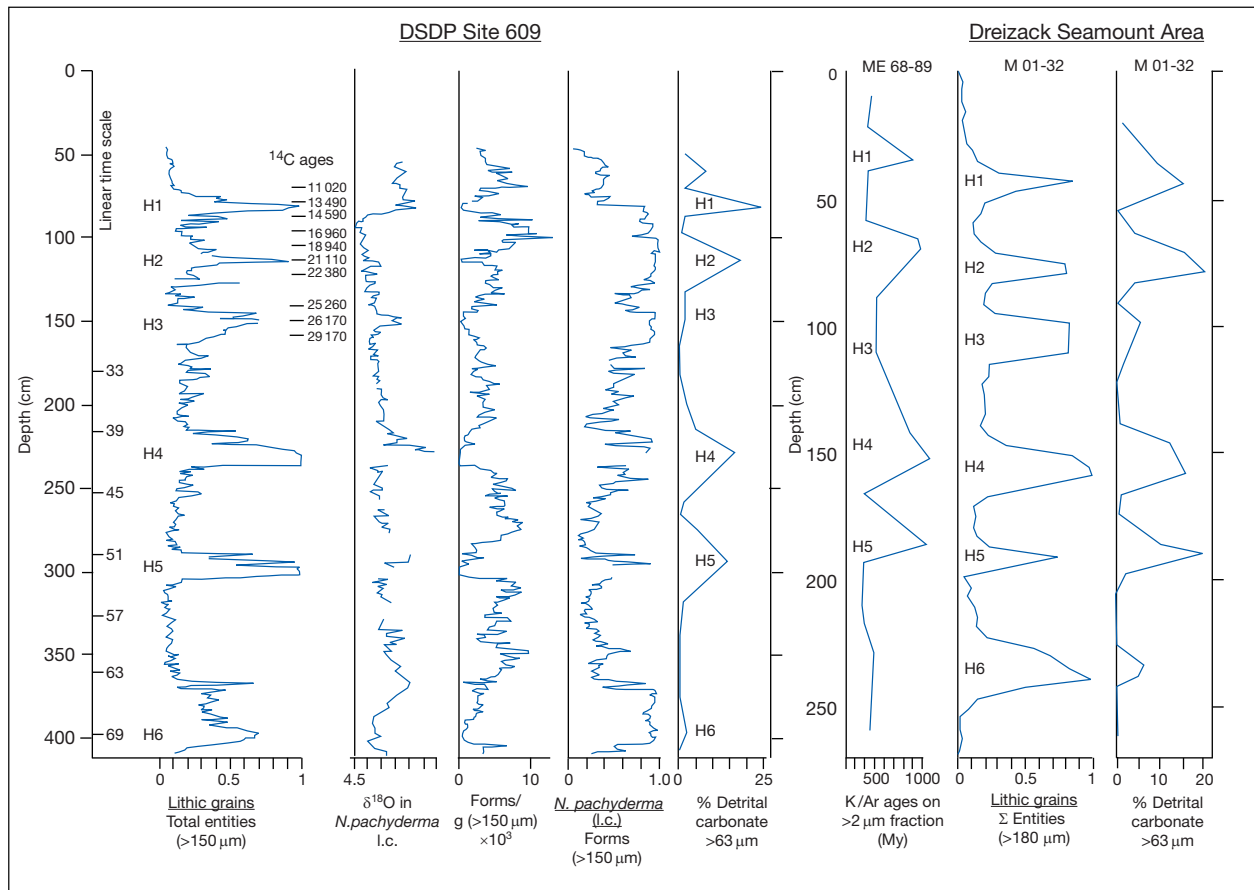


Figure 14 Correlation between cores containing Heinrich layers from the Dreizack Seamount area with DSDP site 609 from the North Atlantic, including summary of radiometric, planktonic, $\delta^{18}\text{O}$, foraminiferal, and lithic measurements. Heinrich layers H1–H6 are marked. Modified from Bond G, Heinrich H, Broecker W, et al. (1992) Evidence for massive discharges of icebergs into the North Atlantic during the last glacial period. *Nature* 360: 245–249.

variety of processes other than iceberg rafting, including turbidity current sedimentation and deposition from the nepheloid layer (Hesse and Khodabakhsh, 1998).

Mechanisms and controls

Three main explanations have been proposed to explain the Heinrich layers:

1. The binge–purge model (MacAyeal, 1993) is a glaciological explanation. A large ice sheet will build up gradually through time (the ‘binge’ stage) until a critical threshold (determined by geothermal heat flux, advection of heat from the ice-sheet surface and strain generated heat) is exceeded and the ice-sheet bed changes from a frozen to a thawed state. This results in rapid ice flow and a major pulse of icebergs – the ‘purge stage’ – giving a Heinrich event. Successive surges of the Laurentide Ice Sheet through Hudson Strait were believed to have been facilitated by soft unconsolidated sediment on the floor of the strait. MacAyeal (1993) calculated a periodicity of about 7000 years for the cycle which agrees with the observed periodicity of the Heinrich events.
2. Johnson and Lauritzen (1995) proposed that repeated outburst floods from a subglacial lake in Hudson Bay produced major pulses of freshwater into the North Atlantic when the ice dam at the mouth of Hudson Strait failed. Such large meltwater discharges would help to explain the abundance of meltwater sediments in ice-proximal Heinrich layers. However, although such a lake could have formed during the early phases of ice-sheet build-up, its presence during the Last Glacial Maximum (LGM) appears to have been unlikely (Andrews, 1998).
3. Hulbe et al. (2004) explain the Heinrich events through the catastrophic collapse of ice shelves fringing the Laurentide Ice Sheet. They argue that ice shelves formed around the ice-sheet margin during extreme cold periods. Subsequent climate warming resulted in their catastrophic collapse and the production of major pulses of icebergs (cf. collapse of the modern Larsen B Ice Shelf in Antarctica). A problem with this model is that available paleoclimatic records show that the Heinrich events began during cold intervals and thus preceded climate warming. Indeed, correlation of the North Atlantic IRD record with the Greenland ice cores shows that each Heinrich event was actually the culmination of a long period of climate cooling and was then followed by an abrupt shift to a warmer climate (Bond et al., 1999).

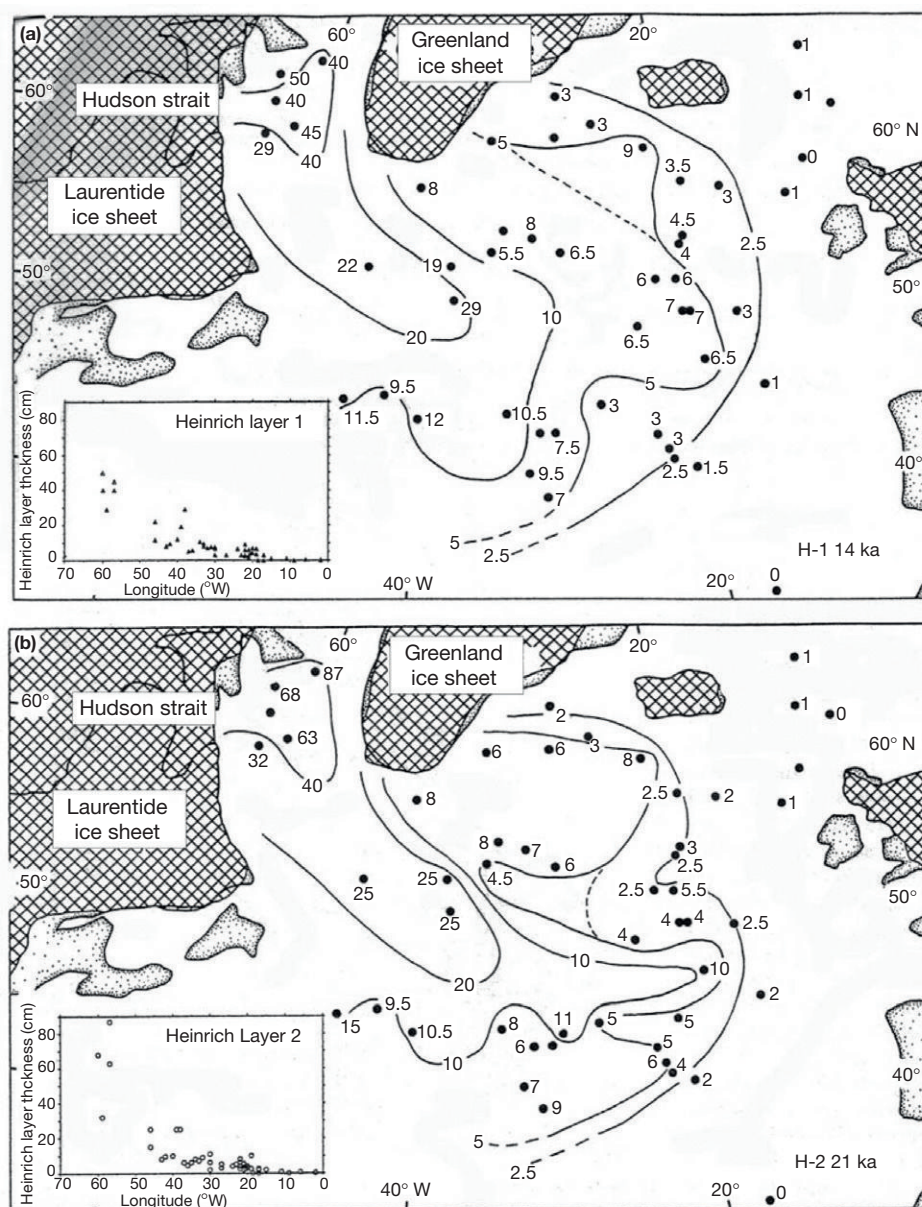


Figure 15 Map showing thickness in centimeters of Heinrich layers H1 (a) and H2 (b) in the North Atlantic based on whole core magnetic susceptibility data. Stipple indicates adjacent landmasses and cross-hatching refers to the generalized locations of ice sheets at the time of Heinrich layer deposition. Modified from Dowdeswell JA, Maslin MA, Andrews JT, and McCave IN (1995) Iceberg production, debris rafting, and the extent and thickness of Heinrich layers (H-1, H-2) in North Atlantic sediments. *Geology* 23: 301–304.

Resolving the question of whether the Heinrich events were triggered by an internal instability within the Laurentide Ice Sheet or alternatively are an ice-sheet response to climate forcing, is critical to determining their underlying cause. Fundamental to this is the timing of the Heinrich events and how they correlate with climate records from the North Atlantic region. A near synchronous response of several ice masses around the North Atlantic might imply an external driver such as climate but this relies upon accurate fingerprinting of IRD to different ice-sheet sources, as well as robust chronological control (cf. the debate over

the ‘precursors’ to Heinrich events from European sources; Hemming, 2004). This topic is likely to remain central to research on IRD and Heinrich layers for some time to come.

See also: **Glacial Landforms:** Moraine Forms and Genesis.
Glacial Landforms, Sediments: Glaciogenic Lithofacies.
Glaciations: Late Quaternary of Antarctica. **Paleoclimate Reconstruction:** Sub-Milankovitch (DO/Heinrich) Events.
Quaternary Stratigraphy: Lithostratigraphy.

References

- Anderson JB (1999) *Antarctic Marine*, p. 289. Cambridge: Cambridge University Press.
- Anderson JB and Ashley GM (eds.) (1991) *Glacial Marine Sedimentation: Paleoclimatic Significance*, p. 232. Boulder, CO: Geological Society of America. Special Paper 261.
- Andrews JT (1998) Abrupt changes (Heinrich events) in late Quaternary North Atlantic marine environments: A history and review of data and concepts. *Journal of Quaternary Science* 13: 3–16.
- Andrews JT (2000) Icebergs and iceberg rafted detritus (IRD) in the North Atlantic: Facts and assumptions. *Oceanography* 13: 100–108.
- Andrews JT and Tedesco K (1992) Detrital carbonate-rich sediments, northwestern Labrador Sea: Implications for ice-sheet dynamics and iceberg rafting (Heinrich) events in the North Atlantic. *Geology* 20: 1087–1090.
- Bond G, Heinrich H, Broecker W, et al. (1992) Evidence for massive discharges of icebergs into the North Atlantic during the last glacial period. *Nature* 360: 245–249.
- Bond G, Showers W, Elliot M, et al. (1999) The North Atlantic's 1–2 ka climate rhythm: Relation to Heinrich events, Dansgaard/Oeschger cycles and the little ice age. In: Clark PU, Webb RS, and Keigwin LD (eds.) *Mechanisms of Global Climate Change at Millennial Time Scales*, pp. 35–68. Washington, DC: American Geophysical Union.
- Canals M, Urgeles R, and Calafat AM (2000) Deep sea-floor evidence of past ice streams off the Antarctic Peninsula. *Geology* 28: 31–34.
- Cheel RJ and Rust BR (1982) Coarse-grained facies of glaciomarine deposits near Ottawa, Canada. In: Davidson-Arnott R, Nickling W, and Fahey BD (eds.) *Research in Glacioluvial and Glaciolacustrine Systems*, pp. 279–295. Norwich: Geo Books.
- Davies TA, et al. (ed.) (1997) *Glaciated Continental Margins: An Atlas of Acoustic Images*, p. 315. London: Chapman and Hall.
- Dowdeswell JA, Elverhøi A, and Spielhagen R (1998) Glaciomarine sedimentary processes and facies on the Polar North Atlantic margins. *Quaternary Science Reviews* 17: 243–272.
- Dowdeswell JA, Maslin MA, Andrews JT, and McCave IN (1995) Iceberg production, debris rafting, and the extent and thickness of Heinrich layers (H-1, H-2) in North Atlantic sediments. *Geology* 23: 301–304.
- Dowdeswell JA and Ó Cofaigh C (eds.) (2002) *Glacier-Influenced Sedimentation on High-Latitude Continental Margins*, p. 378. London: Geological Society. Special Publication 203.
- Dowdeswell JA, Ó Cofaigh C, and Pudsey CJ (2004) Thickness and extent of the subglacial till layer beneath an Antarctic paleo-ice stream. *Geology* 32: 13–16.
- Dowdeswell JA and Scourse JD (eds.) (1990) *Glaciomarine Environments: Processes and Sediments*. London: Geological Society Special Publication 53.
- Elverhøi A, Dowdeswell JA, Funder S, Mangerud J, and Stein R (eds.) (1998) Glacial and Oceanic of the Polar North Atlantic Margins. *Quaternary Science Reviews* 17: 302.
- Heinrich H (1988) Origin and consequences of cyclic ice rafting in the Northeast Atlantic Ocean during the past 130,000 years. *Quaternary Research* 29: 143–152.
- Hemming S (2004) Heinrich events: Massive late Pleistocene detritus layers of the North Atlantic and their global climate imprint. *Reviews of Geophysics* 42: RG1005. <http://dx.doi.org/10.1029/2003RG00128>.
- Hesse R and Khodabakhsh S (1998) Depositional facies of late Pleistocene Heinrich events in the Labrador Sea. *Geology* 26: 103–106.
- Hulbe CL, MacAyeal DR, Denton GH, Kleman J, and Lowell TV (2004) Catastrophic ice shelf breakup as the source of Heinrich event icebergs. *Paleoceanography* 19: PA1004. <http://dx.doi.org/10.1029/2003PA000890>.
- Jakobsson M, Cherkis N, Woodward J, MacNab R, and Coakley B (2000) New grid of Arctic bathymetry aids scientists and mapmakers. *Eos, Transactions of the American Geophysical Union* 81: 89, 93, 96.
- Johnson RG and Lauritzen SE (1995) Hudson Bay–Hudson Strait jökulhlaups and Heinrich events: A hypothesis. *Palaeogeography, Palaeoclimatology, Palaeoecology* 117: 123–137.
- MacAyeal DR (1993) Binge/purge oscillations of the Laurentide Ice Sheet as a cause of the North Atlantic Heinrich events. *Paleoceanography* 8: 775–784.
- McCabe AM, Dardis GF, and Hanvey PM (1984) Sedimentology of a late Pleistocene submarine–moraine complex, County Down, Northern Ireland. *Journal of Sedimentary Petrology* 54: 716–730.
- Ó Cofaigh C and Dowdeswell JA (2001) Laminated sediments in glaciomarine environments: Diagnostic criteria for their interpretation. *Quaternary Science Reviews* 20: 1411–1436.
- Ó Cofaigh C, Dowdeswell JA, Allen CSA, et al. (2005) Flow dynamics and till genesis associated with a marine-based Antarctic paleo-ice stream. *Quaternary Science Reviews* 24: 709–740.
- Ó Cofaigh C, Taylor J, Dowdeswell JA, et al. (2002) Sediment reworking on high-latitude continental margins and its implications for paleoceanographic studies: Insights from the Norwegian–Greenland Sea. In: Dowdeswell JA and Ó Cofaigh C (eds.) *Glacier-Influenced Sedimentation on High-Latitude Continental Margins*, pp. 325–348. London: Geological Society. Special Publication 203.
- Powell RD (1990) Glaciomarine processes at grounding-line fans and their growth to ice-contact deltas. In: Dowdeswell JA and Scourse JD (eds.) *Glaciomarine Environments: Processes and Sediments*, pp. 53–73. London: Geological Society. Special Publication 53.
- Powell RD (1991) Grounding-line systems as second-order controls on fluctuations of tidewater termini of temperate glaciers. In: Anderson JB and Ashley GM (eds.) *Glacial marine sedimentation: Paleoclimate significance*; Special Paper 261. Denver: Geological Society of America. pp. 74–94.
- Powell RD and Molnia BF (1989) Glaciomarine sedimentary processes, facies and morphology of the south-south-east Alaska Shelf and fjords. *Marine Geology* 85: 359–390.
- Shipp SS, Wellner JS, and Anderson JB (2002) Retreat signature of a polar ice stream: Sub-glacial geomorphic features and sediments from the Ross Sea, Antarctica. In: Dowdeswell JA and Ó Cofaigh C (eds.) *Glacier-Influenced Sedimentation on High-Latitude Continental Margins*, pp. 277–304. London: Geological Society. Special Publication 203.
- Vorren TO, Laberg JS, Blaume F, et al. (1998) The Norwegian–Greenland Sea continental margins: Morphology and Late Quaternary sedimentary processes and environment. *Quaternary Science Reviews* 17: 273–302.

Glaciolacustrine

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Introduction

The glaciolacustrine environment includes all ice-contact deep water bodies created by the damming of terrestrial drainage basins by glacier ice. Many small ice-dammed lakes exist in contemporary glacierized landscapes, especially in upland settings where the damming is often related to the abandonment of large moraine loops by receding glacier snouts. Such moraine-dammed lakes are unstable and have a tendency to become potentially hazardous, often creating highly damaging glacial lake outburst floods (GLOFs). Many of the sedimentation processes that operate in the glaciolacustrine environment are similar to those of glaciomarine settings, and so lake processes and forms are emphasized here. Very large, continental scale proglacial lakes existed around the margins of the receding Late Quaternary ice sheets, for example, the Lake Agassiz/McConnell system at the south and west margins of the Laurentide Ice Sheet (Teller, 1995, 2003) and the Baltic Ice Lake and western Russian drainage basins at the southern edge of the Eurasian/Fennoscandinavian ice sheet (Mangerud et al., 2004; Jakobsson et al., 2007). This has given rise to extensive and often thick blankets of glaciolacustrine deposits over formerly glaciated terrains.

Glaciolacustrine Processes

Significant in glaciolacustrine systems is the temperature stratification or thermal structure of lake water and its influence on the patterns of sediment transport and water flow (see Gustavson, 1975; Gilbert and Shaw, 1981; Smith et al., 1982; Smith and Ashley, 1985). The combination of surface heating and vertical mixing in a lake produces a three-layer water column comprising: (a) the epilimnion or upper layer of warm, low-density water; (b) the hypolimnion or lower layer of cold, dense water; and (c) the metalimnion or thermocline, which is a zone of rapid temperature change that separates the epilimnion and hypolimnion. Water inflow into glacier-fed lakes comes from a variety of sources, including direct input from glacier ice via englacial and subglacial portals, glacially fed rivers, and nonglacial rivers. Regions with marked seasonal temperature variations are characterized by maximum discharges of both water and sediment during the summer melt season. Upon entering a lake, the sediment-laden meltwater inflow undergoes an exponential decrease in velocity and thereby produces an expanding and decelerating plume. When viewed from the air, the outer edge of a turbid plume has a very sharp contact with the lake water into which it is discharging, and this marks the position of the 'plunge line' or the point where the plume sinks to reach lake water of similar density. If the density of the diluted inflowing water is less than the density of the epilimnion, it will rise to the surface as an

overflow, but if it is denser than the hypolimnion, it will flow along the bottom of the lake as an underflow. Interflows occur where the density of the inflowing water is intermediate between the dense, cold lake bottom water and the low-density, warm lake surface water, and will move along the thermocline (Figure 1). Other factors that govern the pattern of glacial lake sedimentation include the Coriolis effect, which causes plumes to be deflected to the right in the northern hemisphere and to the left in the southern hemisphere, and bathymetry, which can have a considerable impact on water circulation and associated sedimentation patterns.

Depositional processes operating in glaciolacustrine environments (Figure 2) include deposition from jets emerging from efflux points or subglacial/englacial tunnels (Powell and Domack, 1995), suspension settling, and melt-out/rafting from floating ice and icebergs (Thomas and Connell, 1985). In more proximal settings, the influx of sediment-laden water and the collapsing margins of depocenters give rise to subaqueous debris flows, which are either matrix-rich and cohesive or water-saturated and cohesionless (Lowe, 1976a,b; Nemec, 1990; Postma, 1986; Figure 3). Secondary or syndepositional disturbance and resedimentation can be affected by iceberg scouring (Woodworth-Lynas and Guigné, 1990; Teller, 2003) or the melt-out of buried glacier ice in supraglacial lake settings. The operation of these processes in space gives rise to a rapid proximal to distal fining of sediment assemblages, with the highest sediment concentrations being emplaced close to the efflux point in large depocenters called subaqueous fans, fed by hyperconcentrated flows or bedload traction carpets. Suspension sedimentation is the settling out of fine material from overflows, interflows, and underflows. Unlike marine environments where the salt content of the water aids the settling process, in lake water, suspended particles fall through the water column with a velocity that reflects the buoyant weight of the particle and the drag forces resisting motion, as explained by Stokes's Law.

Glaciolacustrine Sediments

The suspended sediment that gradually settles out from overflows and interflows forms blanket-like drapes, which typically display proximal to distal fining and thinning in lake basins (Smith et al., 1982) with additional patterns of thickening toward the right or left shore in the northern and southern hemispheres, respectively, due to Coriolis deflection (Smith and Ashley, 1985). Below shallow water, overflow and interflow deposits will tend to form massive muds and silts due to sediment mixing by bioturbation, wave disturbance, and wind-generated currents (Smith and Ashley, 1985), but below deeper water, the deposits are typically laminated, fining-up sequences of silt and clay rhythmities that record fluctuations in the

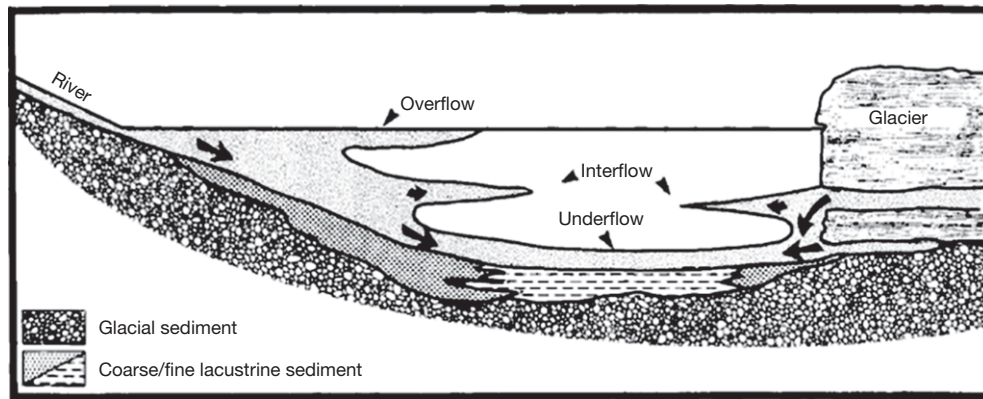


Figure 1 Simplified overview of the main glaciolacustrine processes, showing suspended sediment in light gray being delivered by interflow and underflow from the glacier margin and from a glacier-fed river by underflow, interflow, and overflow. Reproduced from [Teller JT \(1995\)](#) History and drainage of large ice-dammed lakes along the Laurentide Ice Sheet. *Quaternary International* 28: 83–92.

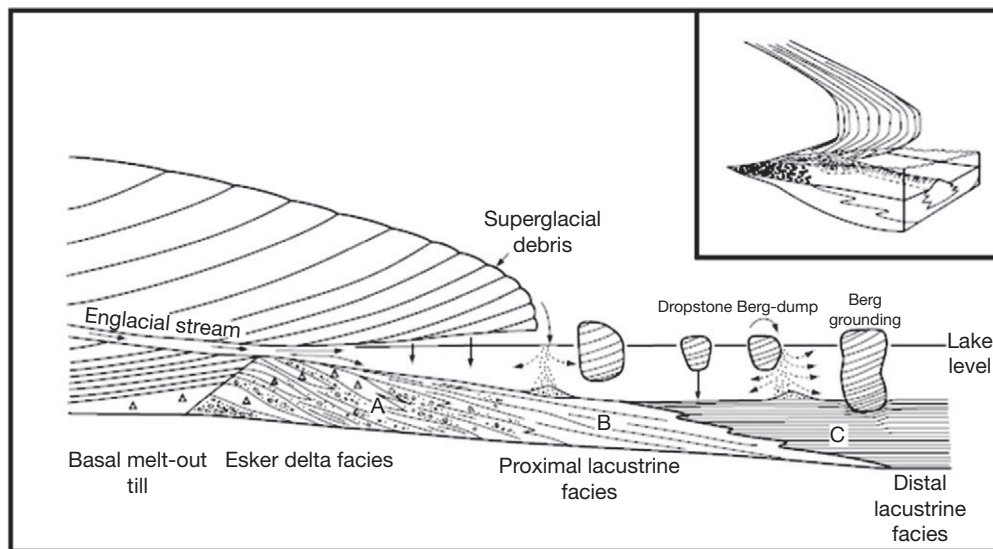


Figure 2 Diagrammatic representation of the range of depositional processes operating in a glaciolacustrine setting. An esker/subaqueous fan complex at 'A' feeds a prograding fan, which emerges from an englacial portal at 'B' and then grades distally into finer-grained rhythmites at 'C'. Iceberg dropstone, dump, and grounding structures are also depicted. Reproduced from [Thomas GSP \(1984\)](#) Sedimentation of a sub-aqueous esker-delta at Strathathie, Aberdeenshire Scottish. *Journal of Geology* 20: 9–20.

grain size and quantity of incoming sediment, due to daily, meteorological, or annual water and sediment discharge cycles ([Church and Gilbert, 1975](#); [Figure 4](#)). Short-term surge currents produce thin, normally graded laminae with sharp basal contacts, whereas annual rhythmic cycles produce distinct silt-clay couplets or varves, which reflect coarse sediment input from overflow and interflow plumes during the ablation season and fine-grained sediment input due to gradual suspension settling during winter, respectively ([Andren et al., 1999](#); [Ashley, 1975](#); [Eyles and Eyles, 1992](#); [Ohlendorf et al., 1997](#); [Palmer et al., 2008](#); [Figure 5](#)). In addition to providing chronological control for deglacial events in some regions, varve sequences have been linked to climatic and meteorological events in Late Quaternary sediment records (e.g., [Hambley and Lamoureux, 2006](#)).

In more proximal locations, the predominance of underflows tends to produce turbidites ([Mulder and Alexander, 2001](#)), which can be interbedded with rhythmites derived from interflows and overflows. Additionally, the operation of subaqueous debris flows gives rise to the accumulation of sequences of interdigitated, discontinuous beds of stratified diamictos. Turbidity currents produce a characteristic graded vertical sequence, known as the Bouma sequence ([Shanmugam, 1997](#); [Kneller and Buckee, 2000](#)), which from the bottom-up includes: (a) massive or normally graded sand or gravel, (b) planar-laminated sand, (c) ripple cross-laminated sand and silt, (d) interlaminated silt and/or clay, and (e) massive clay or silt ([Figure 6](#)). This reflects the gradual decrease in energy of a turbidity current on a lake bottom. Deposits indicative of cohesive debris flows tend to be sheet-like or lobate beds of massive,

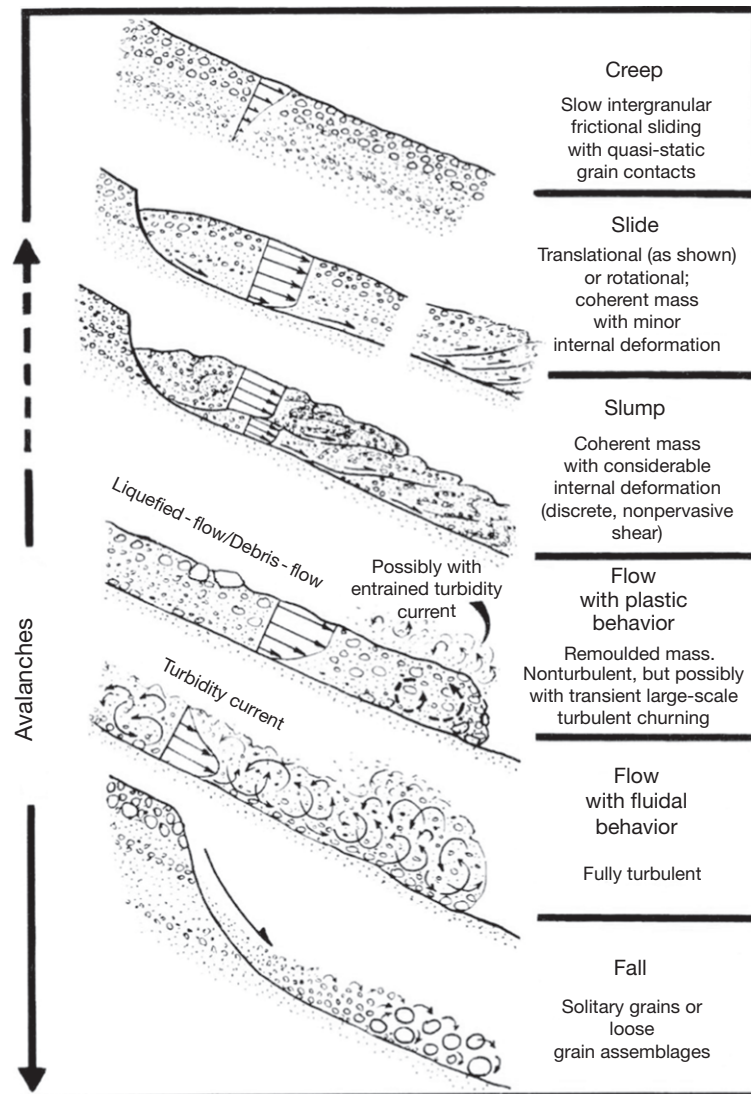


Figure 3 Diagrammatic sketch of the range of gravity-driven sediment transport processes on subaqueous slopes with associated, schematic velocity profiles. Reproduced from Nemec W (1990) Aspects of sediment movement on steep delta slopes. In: Colella A and Prior DB (eds.) *Coarse-Grained Deltas*, pp. 29–73. International Association of Sedimentologists, Special Publication 10.

matrix-supported diamicton or muddy sands and gravels, often containing soft-sediment rafts ripped up from the substrate, and display planar or slightly scoured bases and basal shear zones (Ghibaudo, 1992). Cohesionless debris flow deposits are also typically dipping sheet-like or lobate masses whose bases are commonly erosive and form deep, channelized scours or chutes aligned downslope (Prior and Bornhold, 1990; Figure 7). Internally, they comprise gravel, sand, or pebbly sand, and may be well or poorly sorted, and may have incorporated rafts of underlying material to produce gradational 'welded' basal contacts. Individual flows may appear internally massive or inversely graded and rarely normally graded. Inverse grading at flow bases results from either dispersive pressure generated by colliding grains or kinetic sieving (Legros, 2002; Le Roux, 2003).

Ice-rafted debris (IRD) in glaciolacustrine environments is similar in appearance to that produced in marine settings.

It ranges from isolated clasts or dropstones, dropped onto the lake bed from floating ice, to iceberg dump mounds greater than a meter in height (Figure 8). Smith (2000) reports dropstone A-axis fabrics with high dip angles due to the penetration of stiff bottom sediments by clasts that fall vertically through the water column and hit the bottom nose-first. Thomas and Connell (1985) describe iceberg dump mounds based on Quaternary glaciolacustrine examples as planar-based symmetrical cones of coarse materials, occurring either in isolation or overlapping each other, or even stacked vertically and interstratified with finer-grained sediments. Asymmetrical cones can be produced where multiple dumping events result from repeated overturning of, or basal melt-out from, the same iceberg. A variation on this style of deposition are lake ice conveyor deposits which are linear ridges that mark the former locations of moats in lake ice covers (Hall et al., 2006).

Iceberg or lake ice scouring can produce specific sediment types and structures. Woodworth-Lynas and Guigné (1990) describe massive diamictos or ice-keel turbate produced by ice keel scouring, which can be very difficult to differentiate from nonstratified dropstone diamictos and even some subglacial tills. Some distinctive fault structures can be created by single iceberg grounding or ploughing events (Woodworth-Lynas and Guigné, 1990; Winsemann et al., 2007). These include laterally displaced sediment slices lying on either side of an iceberg plough trace (Woodworth-Lynas and Guigné, 1990) and soft sediment striations or grooves produced by floating lake ice (Eyles et al., 2005). Extensive and often complex disturbance can be introduced to glaciolacustrine sediment sequences where they have been deposited in supraglacial lakes (e.g., Eyles et al., 1987; Mager and Fitzsimons, 2007).



Figure 4 Typical rhythmically bedded sand, silt, and clay laminae, climbing ripples, and draped lamination in a glaciolacustrine depositional sequence on the foreland of Heinabergsjökull, Iceland. Note also the intraformational soft sediment deformation structures.

This includes intraformational soft-sediment deformation, normal faulting, water escape structures, and fluidization tracks, in addition to widespread subaqueous debris flow sedimentation.

Glaciolacustrine Depocenters

In common with marine settings, the landform–sediment associations of the depocenters typically observed in glaciolacustrine depositional environments include grounding-line or subaqueous outwash fans, subaqueous (De Geer) moraines, and deltas. Because of their ice-contact nature, many of the sedimentary sequences of these depocenters may display significant glaciotectionic disturbance (e.g., Phillips et al., 2002; Benn et al., 2004).

Subaqueous outwash fans accumulate at subglacial meltwater efflux points at the glacier–lake interface due to the sudden drop in stream velocity and the progradation of predominantly coarse-grained sediments at the base of the water column (Rust and Romanelli, 1975; Cheel and Rust, 1982; Figure 2). If sedimentation rates are high enough and/or the glacier snout remains relatively stationary, a subaqueous fan can build up to the water surface to form delta (Fyfe, 1990; Figure 9). Their association with subglacial tunnels often results in a close association with eskers where they can demarcate former glacier margins as esker beads (Banerjee and McDonald, 1975; Thomas, 1984; Figure 2). Coarse bedload deposited at the tunnel mouth due to efflux jet deceleration builds up steep, unstable slopes at the glacier margin. Slope failures and renewed sediment discharge from the tunnel mouth then feed cohesionless mass flows that transport sediment radially away from the depocenter and often evolve into turbidity currents. Glaciolacustrine subaqueous fans differ from their glaciomarine counterparts in that they are often finer grained due to the buoyancy and greater transport distances of marine plumes, and tend to contain underflow

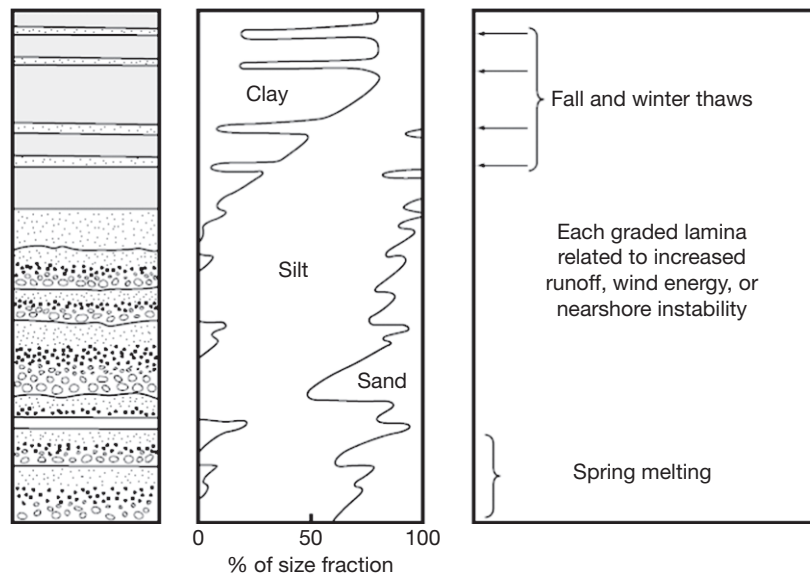


Figure 5 Schematic diagram to show the influence of changes in runoff, wind, or gravitational transfer on the grain size variations in a single varve. The typical yearly signal of overall upward-fining is still visible and records the summer–winter change in sedimentation pattern. Reproduced from Teller JT (2003) Subaquatic landsystems: Large proglacial lakes. In: Evans DJA (ed.) *Glacial Landsystems*, pp. 348–371. London: Arnold.

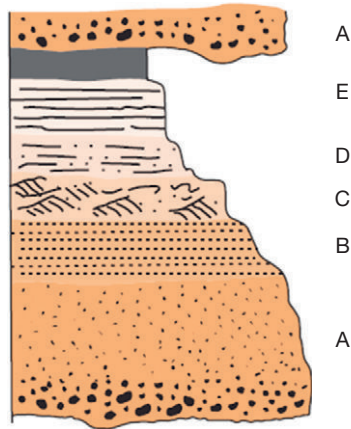


Figure 6 The classic Bouma turbidite sequence with letter-coded units as described in the text.



Figure 7 Stacked sequence of subaqueous debris flow diamictons and intervening stratified lenses in glaciolacustrine sediments at Loch Quoich, Scotland.

Figure 8 Typical iceberg-rafted sediments and structures in glaciolacustrine settings: (a) dropstone-rich rhythmites, SE Ireland; (b) iceberg melt-out structure defined by graben-like fault complex in glaciolacustrine deposits of Lake Pukaki, South Island, New Zealand; (c) iceberg dump and grounding structures observed in Scottish glaciolacustrine deposits: (A) single symmetric structure; (B) stacked asymmetric, laterally overlapping structure; (C) detailed marginal contact; (D) vertically stacked, symmetric structure; and (E) compound, stacked asymmetric structure. Reproduced from [Thomas GSP, Connell RJ \(1985\)](#) Iceberg drop, dump and grounding structures from Pleistocene glacio-lacustrine sediments, Scotland. *Journal of Sedimentary Petrology* 55: 243–249.

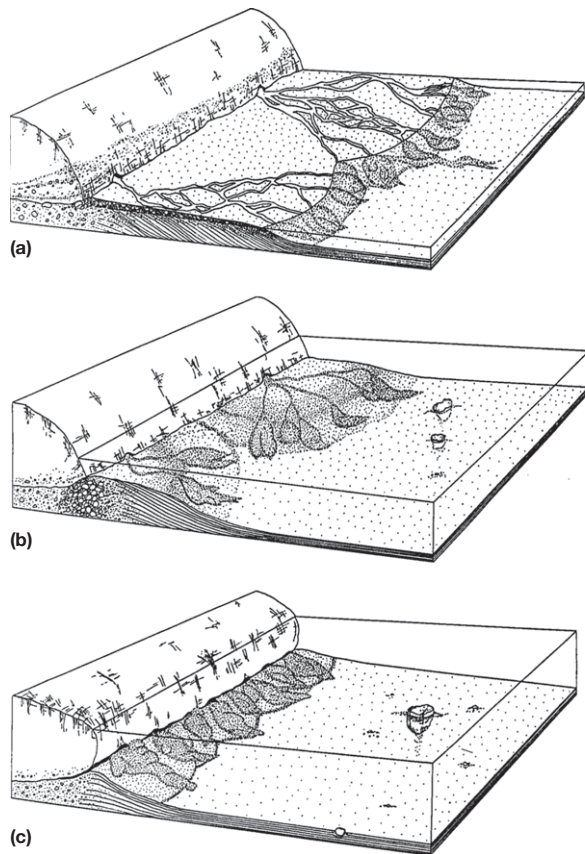


Figure 9 Three styles of glaciolacustrine depocenters based on the Salpausselka moraine assemblage in Finland (from Fyfe, 1990): (a) Gilbert-type delta, (b) subaqueous fan in shallow water, which may aggrade to form a Gilbert delta, and (c) subaqueous fan in deep water.

deposits or turbidites because the sediment-laden meltwater is generally denser than the surrounding lake water. Subaqueous fan sedimentary sequences therefore tend to display cross-bedded clinoforms of interdigitated stratified mass flow diamictites and coarse-grained sands and gravels (Bennett et al., 2002; Cheel and Rust, 1982; Figure 10). The latter comprise crude to well-developed foreset bedding dipping down from the fan apex at angles of 10–30°. The coarser-grained, poorly sorted, massive to normally graded gravels at the fan apex, deposited from high-density, cohesionless debris flows and gravelly traction carpets, grade distally into better sorted, finer grained facies, which document episodic turbidity currents and cohesionless debris flows, triggered by influxes of sediment-laden meltwater or slope failures on the fan surface (Postma et al., 1983).



Figure 10 Stacked sequence of subaqueous outwash fans and associated drapes of rhythmites, documenting the oscillation of a glacier snout in the Rakaia Valley, South Island, New Zealand. Exposure is approximately 15 m high.

Feeder system	Type A	Type B	Type C	Type D
Shallow water deltas	Hjulström-type			
	 1+2 Shallow-water profile		 7+8 Shallow-water profile	
Deep water deltas	Gilbert-type		Mouth bar-type	
	 3+4 Gilbert-type		 9+10 Gilbert-type	
Deep water deltas	Gilbert-type		Mouth bar-type	
	 5 Debris cones		 11 Delta-fed submarine ramp system	
Deep water deltas	Gilbert-type		Mouth bar-type	
	 6 Gravitationally modified Gilbert-type		 12 Delta-fed thalweg and lobe system	

Figure 11 General classification scheme for deltas, showing delta morphology as a function of water depth and the gradient of the feeder river system, classified as (a) very steep, (b) steep, (c) moderate, and (d) low gradient. Reproduced from Postma G (1990) Depositional architecture and facies of river and fan deltas: a synthesis. In: Collella A and Prior DB (eds.) *Coarse-Grained Deltas*, pp. 13–27. Oxford: Blackwell. International Association of Sedimentologists, Special Publication 10.

Subaqueous moraines are deposited at or close to the grounding lines of water-terminating glaciers, and in glaciolacustrine settings have been variably termed sublacustrine moraines, cross-valley moraines, and De Geer moraines. [Benn and Evans \(2010\)](#) classify such moraines as either 'coalescent subaqueous fans', where they comprise large subaqueous moraine ridges formed at glacier grounding lines, or 'De Geer moraines' where they constitute smaller, narrow, and sharp-crested sub-parallel ridges commonly occurring in closely spaced clusters. Spectacular examples of coalescent subaqueous fans are the Salpausselka Moraines of Finland ([Figure 9](#)), which were deposited during stillstands of the southern margins of the Scandinavian Ice Sheet ([Fyfe, 1990](#)). Sediment exposures through such landforms typically exhibit a range of deposits,

from ice-proximal subglacial tills and glacioteconites derived from preexisting glaciolacustrine deposits to ice-distal subaqueous outwash, interbedded with cohesive to cohesionless debris flows (e.g., [Benn, 1996](#)). De Geer moraines commonly occur in fields of closely spaced ridges in association with subaqueous sediments and mark the intermittent retreat of water-terminating glaciers (e.g., [Aartolahti et al., 1995](#); [Andrews and Smithson, 1966](#); [Holdsworth, 1973](#)). They possess many of the morphological characteristics of terrestrial push moraines, such as steeper distal than proximal slopes, lobate forms with acute interlobe angles pointing up glacier, and truncated sections due to partial overriding by later moraines. Like subaqueous fans, De Geer moraines display a wide range of sediment types, from tills to stratified sediments, related to subglacial and glaciotectonic

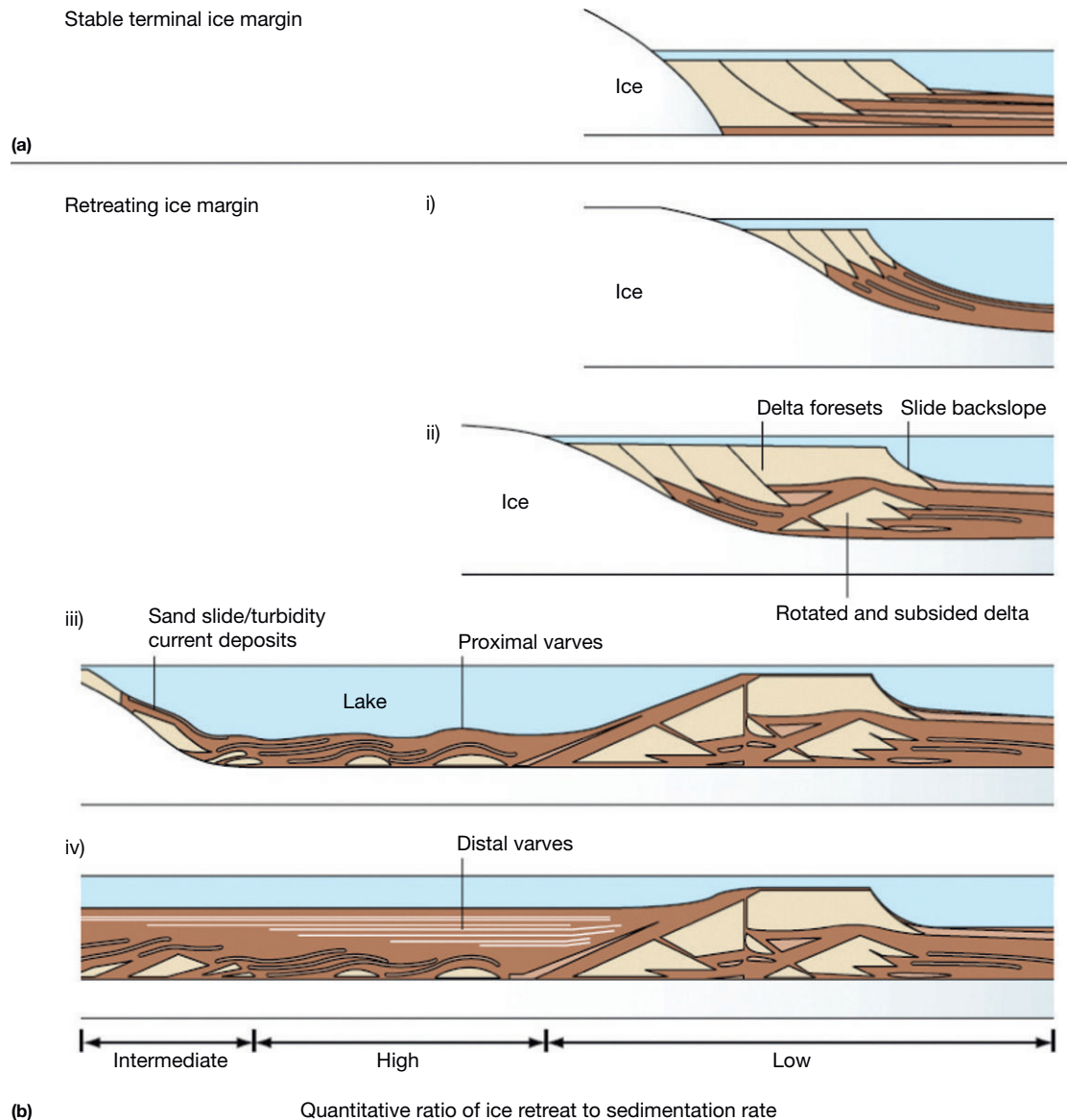


Figure 12 Conceptual model to convey the relationship between ice-contact delta development and glacier retreat rate based on alpine glacial lakes during (a) glacier occupancy and ice-contact delta production, and (b) glacier retreat and ice-contact/supraglacial deposition of deltaic and distal sediments. Note the occurrence of delta collapse and the qualitative ratio of ice retreat to sedimentation rate. Reproduced from [Shaw J \(1977\)](#) Sedimentation in an alpine lake during deglaciation: Oakanagan Valley, British Columbia, Canada. *Geografiska Annaler* 59A: 221–240.

processes on ice-proximal slopes and the interdigitation of sediment gravity flow deposits and glaciolacustrine sediments on ice-distal slopes (e.g., Blake, 2000; Lindén and Möller, 2005).

Deltas form by a combination of fluvial aggradation above water level and progradation of bedload on the delta front due to deceleration of incoming streamflow. Fine-grained suspended sediment is carried beyond the delta by underflows, interflows, and overflows to be deposited in the deeper parts of the lake basin. Delta-front profiles usually display steep upper slopes, giving way to easier angled slopes toward the base of the delta, reflecting proximal-to-distal decreases in sedimentation rates and grain size. In a glacial context, deltas may form either as glacier-fed deltas, which receive sediment from proglacial meltwater streams following transport across an intervening land surface, or as ice-contact deltas, which are built out directly from glacier margins after evolving from subaqueous fans or moraines. Glacier-fed deltas are commonly the shallow-water Hjulström-type with gently sloping fronts or the deeper water Gilbert-type with steep fronts, in addition to the intermediate Salisbury-type (Church and Gilbert, 1975; Figure 11). The most common Gilbert-type deltas are recognizable from their characteristic tripartite sedimentary sequences of fluvial topsets, gravitational foresets, and underflow-fed bottomsets (Gustavson et al., 1975; Martini, 1990; Mastalerz, 1990; Postma, 1995). Ice-contact deltas commonly display similar sedimentary characteristics to those of glacier-fed deltas but, because they are deposited in ice-marginal or supraglacial settings, some important differences are readily apparent. Following glacier recession, an ice-contact delta will form an isolated plateau with a prominent, steep ice-contact slope at its ice-proximal end. They often also lie perched on topographic highs, where the only plausible source of sediment and water is a former glacier margin (Clayton et al., 2008), and will contain extensive surface pitting or kettle holes created by the melt-out of buried glacier ice. Ice-proximal facies will also often display evidence of subglacial deformation (Phillips et al., 2002; Benn et al., 2004) and/or slumping due to the melt-out of buried glacier ice or the removal of support from the ice-contact face. The relationship between ice-contact delta development and glacier retreat rate is depicted by the conceptual model of Shaw (1977) (Figure 12). At relatively stable glacier termini, a normal Gilbert-type delta will develop because of the low ice retreat to sedimentation ratio. If the terminus then retreats a short distance to another stable position, the ice-contact face of the delta will subside and be buried by a subsequent deltaic sequence. In contrast, if the glacier margin retreats rapidly, the increased ice retreat to sedimentation ratio will result in the replacement of delta formation with subaqueous fans and distal rhythmites.

Conclusion

Extensive and often thick blankets of glaciolacustrine deposits characterize the Quaternary sediment-landform assemblages of deglaciated lowland terrains. Ice-proximal and ice-contact depocenters provide important information on the former locations of glacier and ice sheet margins where they oscillated in proglacial lakes, and comprise a spatial and temporal continuum of subaqueous outwash fans, ice-contact deltas, and subaqueous (De Geer) moraines. These ice-proximal

and ice-contact features are distinct from the topset/foreset/bottomset-bedded Gilbert-, Salisbury-, and Hjulstrom-type glacier-fed deltas in that (a) they occur in physiographic settings where the only plausible source of sediment and water is a former glacier margin and (b) they contain a wide range of interdigitated deposits, including subglacial tills and glacio-tectonites, subaqueous outwash gravels and sands, and cohesive to cohesionless debris flow diamictos. Distal sedimentation is evident in iceberg-rafted debris and rhythmite sequences, often clearly recording seasonal variability in the influxes of fine-grained sediments in the form of varves. Evidence of secondary or syndepositional disturbance by glacier ice pushing and melt-out, iceberg and lake ice scouring, and dumping is widespread in the form of a whole range of deformation structures.

See also: Varved Lake Sediments. **Fluvial Environments:** Deltaic Environments. **Glacial Landforms:** Evidence of Glacier Recession; Glacial Landsystems; Glacitectonic Structures and Landforms; Moraine Forms and Genesis. **Glacial Landforms, Sediments:** Glaciofluvial Landforms of Deposition; Glaciogenic Lithofacies; Glaciomarine Sediments and Ice-Rafted Debris; Tills.

References

- Aartolahti T, Koivisto M, and Nenonen K (1995) De Geer Moraines in Finland. Geological Survey of Finland. Special Paper 20, 67–74.
- Andren T, Björck S, and Johnsen S (1999) Correlation of Swedish glacial varves with the Greenland (GRIP) oxygen isotope record. *Journal of Quaternary Science* 14: 361–371.
- Andrews JT and Smithson BB (1966) Till fabrics of the cross valley moraines of north-central Baffin Island, NWT, Canada. *Bulletin of the Geological Society of America* 77: 271–290.
- Ashley GM (1975) Rhythmic sedimentation in glacial lake Hitchcock, Massachusetts-Connecticut. In: Jopling AV and McDonald BC (eds.) *Glaciofluvial and Glaciolacustrine Sedimentation*, pp. 302–320. Society of Economic Paleontologists and Mineralogists Special Publication, No. 23.
- Banerjee I and McDonald BC (1975) Nature of esker sedimentation. In: Jopling AV and McDonald BC (eds.) *Glaciofluvial and Glaciolacustrine Sedimentation*, pp. 132–154. Society of Economic Paleontologists and Mineralogists Special Publication, No. 23.
- Benn DI (1996) Subglacial and subaqueous processes near a glacier grounding line: Sedimentological evidence from a former ice-dammed lake, Achnasheen, Scotland. *Boreas* 25: 23–36.
- Benn DI and Evans DJA (2010) *Glaciers and Glaciation*. London: Hodder Arnold.
- Benn DI, Evans DJA, Phillips ER, Hiemstra JF, Walden J, and Hoey TB (2004) The research project – a case study of Quaternary glacial sediments. In: Evans DJA and Benn DI (eds.) *A Practical Guide to the Study of Glacial Sediments*, pp. 209–234. London: Arnold.
- Bennett MR, Huddart D, and Thomas GSP (2002) Facies architecture within a regional glaciolacustrine basin: Copper River, Alaska. *Quaternary Science Reviews* 21: 2237–2279.
- Blake KP (2000) Common origin for De Geer moraines of variable composition in Raudvassdalen, northern Norway. *Journal of Quaternary Science* 15: 633–644.
- Cheel RJ and Rust BR (1982) Coarse-grained facies of glacio-marine deposits near Ottawa, Canada. In: Davidson-Arnott R, Nickling W, and Fahey BD (eds.) *Research in Glacial, Glaciofluvial, and Glaciolacustrine Systems*, pp. 279–295. Ontario, Canada: University of Guelph.
- Church M and Gilbert R (1975) Proglacial fluvial and lacustrine sediments. In: Jopling AV and McDonald BC (eds.) *Glaciofluvial and Glaciolacustrine Sedimentation*, pp. 22–100. Society of Economic Paleontologists and Mineralogists Special Publication, No. 23.

- Clayton L, Attig JW, Ham NR, Johnson MD, Jennings CE, and Syverson KM (2008) Ice-walled-lake plains: Implications for the origin of hummocky glacial topography in middle North America. *Geomorphology* 97: 237–248.
- Eyles N and Eyles CH (1992) Glacial depositional systems. In: Walker RG and James NP (eds.) *Facies Models: Response to Sea Level Change*, pp. 73–100. Toronto: Geological Association of Canada.
- Eyles N, Clark BM, and Clague JJ (1987) Coarse-grained sediment gravity flow facies in a large supraglacial lake. *Sedimentology* 34: 193–216.
- Eyles N, Eyles CH, Woodworth-Lynas CMT, and Randall TA (2005) The sedimentary record of drifting ice (early Wisconsin Sunnyside deposit) in an ancestral ice-dammed Lake Ontario, Canada. *Quaternary Research* 63: 171–181.
- Fyfe GJ (1990) The effect of water depth on ice-proximal glaciolacustrine sedimentation: Salpausselkä I, southern Finland. *Boreas* 19: 147–164.
- Ghibaudo G (1992) Subaqueous sediment gravity flow deposits: Practical criteria for their field description and classification. *Sedimentology* 39: 423–454.
- Gilbert R and Shaw J (1981) Sedimentation in proglacial Sunwapta Lake, Alberta. *Canadian Journal of Earth Sciences* 18: 81–93.
- Gustavson TC (1975) Sedimentation and physical limnology in proglacial Malaspina Lake, southeastern Alaska. In: Jopling AV and McDonald BC (eds.) *Glaciofluvial and Glaciolacustrine Sedimentation*, pp. 249–263. Society of Economic Paleontologists and Mineralogists, Special Publication, No. 23.
- Gustavson TC, Ashley GM, and Boothroyd JC (1975) Depositional sequences in glaciolacustrine deltas. In: Jopling AV and McDonald BC (eds.) *Glaciofluvial and Glaciolacustrine Sedimentation*, pp. 264–280. Society of Economic Paleontologists and Mineralogists Special Publication, No. 23.
- Hall BL, Hendy CH, and Denton GH (2006) Lake-ice conveyor deposits: Geomorphology, sedimentology and importance in reconstructing the glacial history of the Dry Valleys. *Geomorphology* 75: 143–156.
- Hambley GW and Lamoureux SF (2006) Recent summer climate recorded in complex varved sediments, Nicolay Lake, Cornwall Island, Nunavut. *Journal of Paleolimnology* 35: 629–640.
- Holdsworth G (1973) Ice deformation and moraine formation at the margin of an ice cap adjacent to a proglacial lake. In: Fahey BD and Thompson RD (eds.) *Research in Polar and Alpine Geomorphology*, pp. 187–199. Norwich: Geo Abstracts.
- Jakobsson M, Björck S, Alm G, Andren T, Lindeberg G, and Svensson NO (2007) Reconstructing the Younger Dryas ice-dammed lake in the Baltic Basin: Bathymetry, area and volume. *Global and Planetary Change* 57: 355–370.
- Kneller B and Buckee C (2000) The structure and fluid mechanics of turbidity currents: A review of some recent studies and their geological implications. *Sedimentology* 47(supplement 1): 62–94.
- Le Roux JP (2003) Can dispersive pressure cause inverse grading in grain flows? Discussion. *Journal of Sedimentary Research* 73: 333–334.
- Legros F (2002) Can dispersive pressure cause inverse grading in grain flows? *Journal of Sedimentary Research* 72: 166–170.
- Lindén M and Möller P (2005) Marginal formation of De Geer moraines and their implications to the dynamics of grounding line recession. *Journal of Quaternary Science* 20: 113–133.
- Lowe DR (1976a) Grain flow and grain flow deposits. *Journal of Sedimentary Petrology* 46: 188–199.
- Lowe DR (1976b) Subaqueous liquefied and fluidized sediment flows and their deposits. *Sedimentology* 23: 285–308.
- Mager S and Fitzsimons S (2007) Formation of glaciolacustrine Late Pleistocene end moraines in the Tasman Valley, New Zealand. *Quaternary Science Reviews* 26: 743–758.
- Mangerud J, Jakobsson M, Alexanderson H, et al. (2004) Ice-dammed lakes and rerouting of the drainage of northern Eurasia during the Last Glaciation. *Quaternary Science Reviews* 23: 1313–1332.
- Martini IP (1990) Pleistocene glacial fan deltas in southern Ontario, Canada. In: Colella A and Prior D (eds.) *Coarse-Grained Deltas*, pp. 281–295. International Association of Sedimentologists, Special Publication 10.
- Mastalerz K (1990) Diurnally and seasonally controlled sedimentation on a glaciolacustrine foreset slope: An example from the Pleistocene of eastern Poland. In: Colella A and Prior D (eds.) *Coarse-Grained Deltas*, pp. 297–309. International Association of Sedimentologists, Special Publication 10.
- Mulder T and Alexander J (2001) The physical character of subaqueous sedimentary density flows and their deposits. *Sedimentology* 48: 269–299.
- Nemec W (1990) Aspects of sediment movement on steep delta slopes. In: Colella A and Prior DB (eds.) *Coarse-Grained Deltas*, pp. 29–73. International Association of Sedimentologists, Special Publication 10.
- Ohlendorf C, Niessen F, and Weissert H (1997) Glacial varve thickness and 127 years of instrumental climate data – a comparison. *Climate Change* 36: 391–411.
- Palmer AP, Rose J, Lowe JJ, and Walker MJC (2008) Annually laminated Late Pleistocene sediments from Llangorse Lake, South Wales: A chronology for the pattern of ice wastage. *Proceedings of the Geologists' Association* 119: 245–258.
- Phillips ER, Evans DJA, and Auton CA (2002) Polyphase deformation at an oscillating ice margin following the Loch Lomond Readvance, central Scotland, UK. *Sedimentary Geology* 149: 157–182.
- Postma G (1986) Classification for sediment gravity flow deposits based on flow conditions during sedimentation. *Geology* 14: 291–294.
- Postma G (1995) Causes of architectural variation in deltas. In: Oti MN and Postma G (eds.) *Geology of Deltas*, pp. 3–16. Rotterdam: Balkema.
- Postma G, Roep TB, and Ruegg GJH (1983) Sandy-gravelly mass-flow deposits in an ice-marginal lake (Saalian, Leuvenumsche Beek Valley, Veluwe, The Netherlands), with emphasis on plug-flow deposits. *Sedimentary Geology* 134: 59–82.
- Powell RD and Domack E (1995) Modern glaciomarine environments. In: Menzies J (ed.) *Modern Glacial Environments: Processes, Dynamics and Sediments. Glacial Environments*, vol. 1, pp. 445–486. Oxford: Butterworth-Heinemann.
- Prior DB and Bornhold BD (1990) The underwater development of Holocene fan deltas. In: Colella A and Prior DB (eds.) *Coarse-Grained Deltas*, pp. 75–90. Oxford: Blackwell. International Association of Sedimentologists, Special Publication No. 10.
- Rust BR and Romanelli R (1975) Late Quaternary subaqueous deposits near Ottawa, Canada. In: Jopling AV and McDonald BC (eds.) *Glaciofluvial and Glaciolacustrine Sedimentation*, pp. 177–192. Society of Economic Paleontologists and Mineralogists, Special Publication, No. 23.
- Shanmugam G (1997) The Bouma Sequence and the turbidite mind set. *Earth Science Reviews* 42: 201–229.
- Shaw J (1977) Sedimentation in an alpine lake during deglaciation: Okanagan Valley, British Columbia, Canada. *Geografiska Annaler* 59A: 221–240.
- Smith IR (2000) Diamictic sediments within high arctic lake sediment cores: Evidence for lake ice rafting along the lateral glacial margin. *Sedimentology* 47: 1157–1179.
- Smith ND and Ashley G (1985) Proglacial lacustrine environment. In: Ashley GM, Shaw J, and Smith ND (eds.) *Glacial Sedimentary Environments*, pp. 135–207. Society of Economic Paleontologists and Mineralogists Short Course 16.
- Smith ND, Vendl MA, and Kennedy SK (1982) Comparison of sedimentation regimes in four glacier-fed lakes of western Alberta. In: Davidson-Arnott R, Nickling W, and Fahey BD (eds.) *Research in Glacial, Glaciofluvial, and Glaciolacustrine Systems*, pp. 203–238. Ontario, Canada: University of Guelph.
- Teller JT (1995) History and drainage of large ice-dammed lakes along the Laurentide Ice Sheet. *Quaternary International* 28: 83–92.
- Teller JT (2003) Subaquatic landsystems: Large proglacial lakes. In: Evans DJA (ed.) *Glacial Landsystems*, pp. 348–371. London: Arnold.
- Thomas GSP (1984) Sedimentation of a sub-aqueous esker-delta at Strathathie, Aberdeenshire Scottish. *Journal of Geology* 20: 9–20.
- Thomas GSP and Connell RJ (1985) Iceberg drop, dump and grounding structures from Pleistocene glacio-lacustrine sediments, Scotland. *Journal of Sedimentary Petrology* 55: 243–249.
- Winsemann J, Asprion U, Meyer T, and Schramm C (2007) Facies characteristics of Middle Pleistocene (Saalian) ice-marginal subaqueous fan and delta deposits, glacial Lake Leine, NW Germany. *Sedimentary Geology* 193: 105–129.
- Woodworth-Lynas CMT and Guigné JY (1990) Iceberg scours in the geological record: Examples from Glacial Lake Agassiz. In: Dowdeswell JA and Scourse JD (eds.) *Glaciomarine Environments: Processes and Sediments*, pp. 217–223. Geological Society of London, Special Publication 53.

Micromorphology of Glacial Sediments

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Introduction

The dynamics and spatial variability of processes operating in glacial depositional environments may lead to sedimentary sequences that are highly diverse and complex, in terms of both composition and architecture. This makes the interpretation of many glacial sedimentary (paleo-)records challenging. Specific problems are associated with the interpretation of a range of glacial diamictos, which, although they are known to form in a multitude of ways, may appear identical to the naked eye (see [Glaciogenic Lithofacies](#); [Glaciomarine Sediments and Ice-Rafted Debris](#); [Tills](#)). Particularly where visible exposures of these sediments are limited, such as in sediment cores, it can be notoriously difficult or almost impossible to arrive at reliable interpretations when using conventional field sedimentological techniques alone.

Sometimes there is a need for more detailed information on sediments. For example, not being able to distinguish between subglacially (de)formed diamictos (basal tills) and ice-proximal glaciomarine diamictos in sedimentary records of formerly glaciated continental shelf areas will seriously affect the accuracy of paleo-grounding line reconstructions, which has detrimental effects on reconstructions of ice-sheet volumetric changes over time (see [Glaciomarine Sediments and Ice-Rafted Debris](#)).

In cases such as this, glacial sedimentologists have employed micromorphology. This technique, which has been the standard procedure for decades in rock mineralogical studies and paleopedology and which is starting to get established in other fields as well ([Menzies et al., 2010](#)), entails the analysis of up to 90 cm² thin sections under a transmitted-light petrographic microscope. Using magnifications commonly of 2× to 50×, texture, structure, and other microscale features of undisturbed, impregnated sediment samples can be analyzed in both plane- and cross (or circularly)- polarized light. Unlike other techniques, micromorphology allows for a detailed study of interrelations between sediment constituents, which has proved to be invaluable in genetic interpretations.

Sampling and Preparation of Thin Sections

If glacial sediments are unlithified, which is commonly the case for those of Quaternary age, they need to be sampled with great care. To ensure that disturbance is kept to a minimum and sedimentary and other primary structures are preserved, there is essentially a choice between two sampling methods, both of which can be quite elaborate. The first method is to carefully carve a block of material from an exposure or from a sediment core. The second method makes use of a special double-hinged, double-lidded tin (8 × 6 × 4 cm Kubišna-size, or 15 × 8 × 5 cm, so-called mammoth-size) and is

designed, and therefore particularly suitable, for brittle and non-cohesive materials. Sampling in this case involves the meticulous transfer of sediment into the tin by gently cutting and pushing the tin into the exposure. The tin is needed to support the loose material and to prevent the sample from disintegrating.

Unlike most rock specimens, which can be thin sectioned without preparation, unlithified sediment samples require impregnation before they can be cut, ground, and polished (e.g., [Murphy, 1986](#)). To prepare samples for impregnation, they need to be dried in order to remove all interstitial water. Normally, this involves 1 or 2 weeks of air- and oven-drying (at ~40 °C), but in exceptional cases, it may be necessary to expel excessive amounts of water by means of acetone replacement.

Impregnation is undertaken in a desiccator or sealed tank, in which a vacuum pressure of ~0.6 kPa is generated. The impregnation fluid mixture has crystic or epoxy resin as its most important component but also contains thinners and catalysts to enhance and speed up the process. Because at this stage the samples are still susceptible to disturbance, it is crucial that the fluid is applied in steadily increasing doses and that vacuum pressures are increased gradually in order to allow slow and optimal penetration of the resin into the sediment.

Curing of the impregnated samples normally takes between 4 and 8 weeks at room temperature and atmospheric pressure. Once hard, samples can be treated in much the same way as rock specimens and are ready for machine sawing, lapping, and grinding. In the final stage of producing the thin sections, slabs of impregnated samples are mounted on glass plates, ground, and polished to a thickness of ~30 μm, and covered with thin glass slips.

Thin Section Description

To help with communication between individual users of the technique in its application to glacial sediments, much effort has been put into standardizing the description of thin sections. Although still subject to modifications and refinements, a scheme has been in place since the 1980s, which is now recognized by most glacial micromorphologists. The nomenclature in this scheme, which was introduced by [van der Meer \(1987\)](#), is largely derived from the terminology used in soil micromorphology (cf. [Brewer, 1976](#); see [Soil Micromorphology](#)). Van der Meer recognized that many of the pedological terms, particularly those used to describe orientations and patterns in the silt and clay fraction (see below), would also be appropriate for the characterization of microscale features occurring in glacial sediments ([Figure 1](#)).

Micromorphological analyses of glacial sediments commonly involve observation, description, and interpretation of texture, structure, microfabric, and plasmic fabric (each discussed in detail in the following sections). Ideally, analyses should lead to the identification of microscale features that

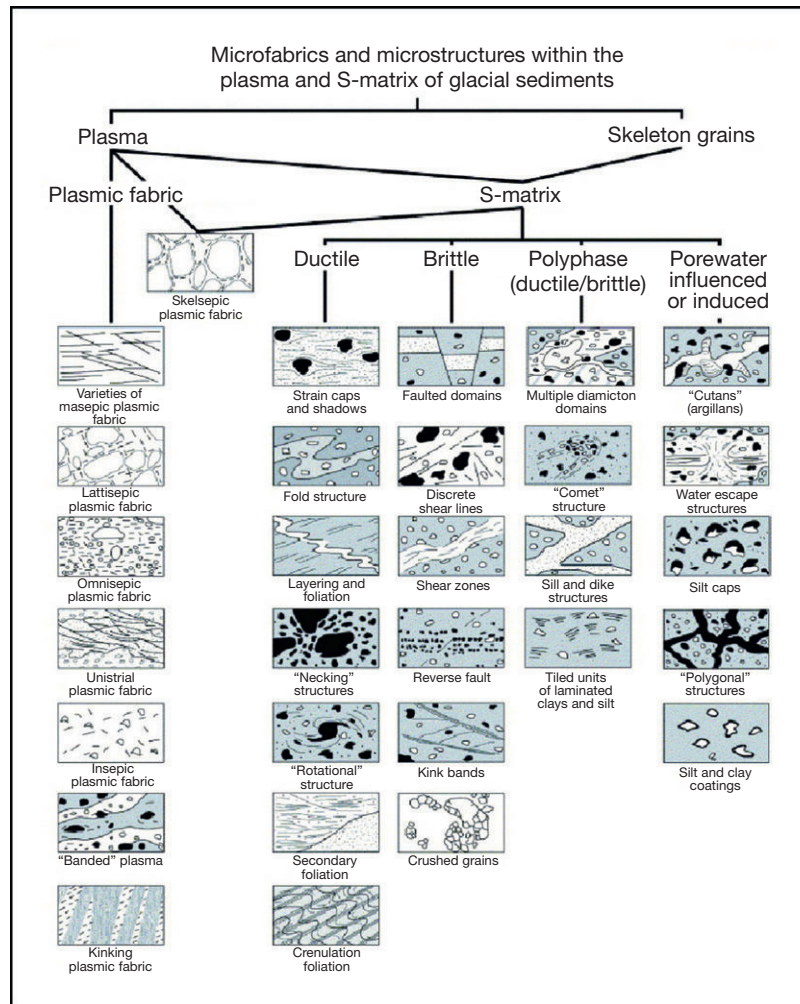


Figure 1 Microstructures and plasmic fabric commonly found within glacial sediments. Reprinted from [Menzies J \(2000\)](#) Micromorphological analyses of microfabrics and microstructures indicative of deformation processes in glacial sediments. In: Maltman AJ, Hubbard B, and Hambrey MJ (eds.) *Deformation of Glacial Materials*, pp. 245–257. London: Geological Society. Special Publication 176. © 2000, with permission from The Geological Society.

indicate a certain depositional process or environment. The reliability of the genetic interpretation of glacial sediments is largely dependent upon how much evidence can be collected in support of this interpretation. It is important to realize in this respect that individual features are hardly ever diagnostic of a certain process or environment.

Textural Analysis

Textural analysis of thin sections involves the assessment (and measurement) of size, shape, distribution, and composition of particles larger (coarser) than $\sim 30\ \mu\text{m}$. In micromorphology, a distinction is made between particles $>\sim 30\ \mu\text{m}$ (defined as 'skeleton') and particles $<\sim 30\ \mu\text{m}$ (referred to as 'plasma'; [Figure 2\(a\)](#)). This threshold is not arbitrarily chosen but equals the thickness of an average thin section. This thickness determines the visual resolution of the technique: Because several layers of plasma may be contained within the thickness of a thin-sectioned sample, particles $<30\ \mu\text{m}$ cannot be examined individually.

While knowledge of the mineralogical composition may be important in establishing the provenance of a sediment, the size and shape of skeleton particles may hold rudimentary information about the processes of erosion, transport, and deposition. For example, observations of a large silt fraction and relatively rounded pebbles and sand grains in a glacial diamicton ([Figure 2\(b\)](#)) may be used to infer that the sediment under consideration has experienced a significant amount of subglacial modification during transport. This would suggest that processes such as abrasion, comminution, and attrition accounted for the production of silt-sized material and the associated edge rounding of the coarse fraction (see [Clast Form Analysis](#)).

Although advanced techniques such as spectrometry and diffraction analysis of fine-grained sediments can provide accurate information on mineralogical composition and grain-size distribution, respectively, neither technique can give a detailed insight into the spatial arrangement of individual constituents, or indeed resolve whether grains of certain size and/or composition are uniformly distributed, arranged in banded patterns or perhaps concentrated in clusters. Herein

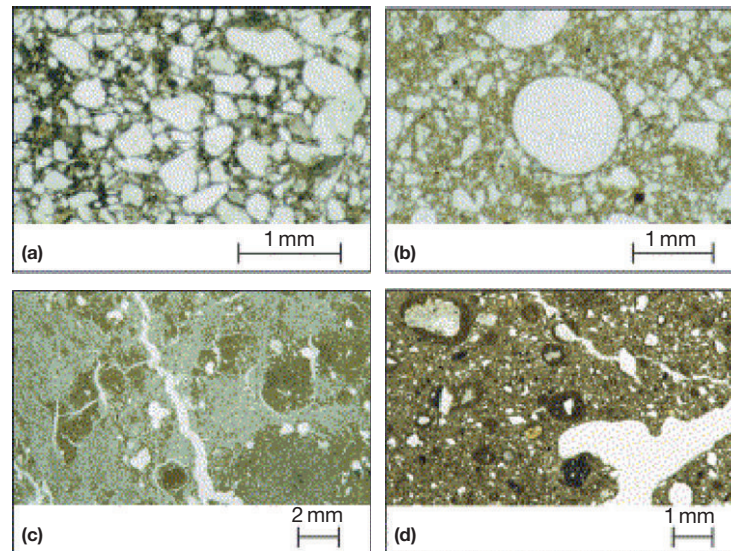


Figure 2 (a) Distinction between skeleton (light particles $>30\ \mu\text{m}$) and plasma (darker matrix, particles $<30\ \mu\text{m}$) in a diamict (basal till) from the Netherlands. Plane-polarized light. (b) Diamict (basal till) from Germany showing plasma that is rich in silt (as reflected by its grainy appearance) and overall subrounded to some very well-rounded skeleton grains. Plane-polarized light. (c) Example of a cored diamict from the Greenland continental shelf showing intraclasts in different stages of development. Two well-rounded intraclasts occur next to features (both to the left and right) that seem to represent the early stages of their development. Plane-polarized light. (d) Intraclasts in a cored diamict recovered from the Antarctic continental shelf. In contrast to the situation in (c), the features here consist of a nucleus (a skeleton grain) and a casing of plasma material. This possibly indicates a different formation process, in which, as a result of rotation of individual skeleton grains, fine material is plastered onto their surfaces. Plane-polarized light.

lies the strength of microscopic texture analysis. Because thin section samples capture the undisturbed sediment, micromorphology allows the investigation of *in situ* particles and thus of the true small-scale sediment architecture. The importance of this is manifest in the study of sediments containing soft sediment intraclasts (or rip-up clasts) (Figure 2(c)–2(d)). Intraclasts, which occur quite commonly in glacial sediments, often indicate reworking of preexisting sediments, caused either by direct glacial action (see van der Meer, 1993) or by resedimentation, for example, as a debris flow (subaerial: see Menzies, 2000; subaqueous: Hiemstra, 2001; Kilfeather et al., 2010). The size, shape, composition, and position of intraclasts in the glacial sediment logically provide information on how and where preexisting sediments were incorporated and, more generally, on which mechanisms played a role in the formation of the glacial sediment as a whole. Such information is not disclosed through methods that make use of bulk samples.

Structural Analysis

In microscopic structural analysis, the arrangement and distribution of primary compound constituents, that is, the interrelations between aggregate particles, are investigated. Structures identified in thin sections of glacial sediment comprise features that result from depositional or deformational processes, or combinations thereof (see Menzies, 2000; van der Meer, 1993). Most glacial microstructures have a macroscopic equivalent. In other words, most can be regarded as smaller versions of known outcrop-scale features (see Figure 1; Glacitectonic Structures and Landforms; Glaciogenic Lithofacies).

Bedding and stratification are common depositional structures, which are often readily visible in sediments in the field. In

such cases, there is obviously no immediate need for micromorphological study. However, certain water-lain sediments are so finely or subtly laminated that investigation under the microscope is imperative, simply because such structures might otherwise go unnoticed, or would be incorrectly interpreted. For example, micromorphology allowed Palmer et al. (2010) to distinguish between sub-mm scale, annually deposited layers (varves) and other rhythmites in a Younger Dryas glaciolacustrine sequence, which enabled them to develop a high-resolution chronology for the area under investigation (see Glaciolacustrine).

Another example is given in Figure 3(a), which shows details of a very finely laminated fjord deposit. It comprises a series of mm-scale couplets consisting of a thin lamina of silt or fine sand (sometimes just one grain thick) that fines upward into a thicker lamina composed of silt and clay (cyclopels: see Hiemstra et al., 2004). This type of glaciomarine deposit is thought to result from suspension settling where turbid, fine particle-laden meltwater plumes interact with tidal marine currents (see Glaciomarine Sediments and Ice-Rafted Debris). Again, due to the subtlety of the structure, microscopic analysis was vital to fully appreciate the nature and history of the sediment.

When carrying out structural analysis of glacial sediments, it is important to realize that not all stratification (or lamination) is necessarily depositional in origin. Where a sequence of sediments is overridden by a glacier, it is quite common to find that deformation and glaciectonic processes produce so-called pseudostratification (pseudolamination) or transposed foliation (e.g., van der Wateren et al., 2000). In situations of low strain, it is possible that preexisting structures are retained or only modified to some degree (Figure 3(b)), so that occurring stratification is a primary feature, but where

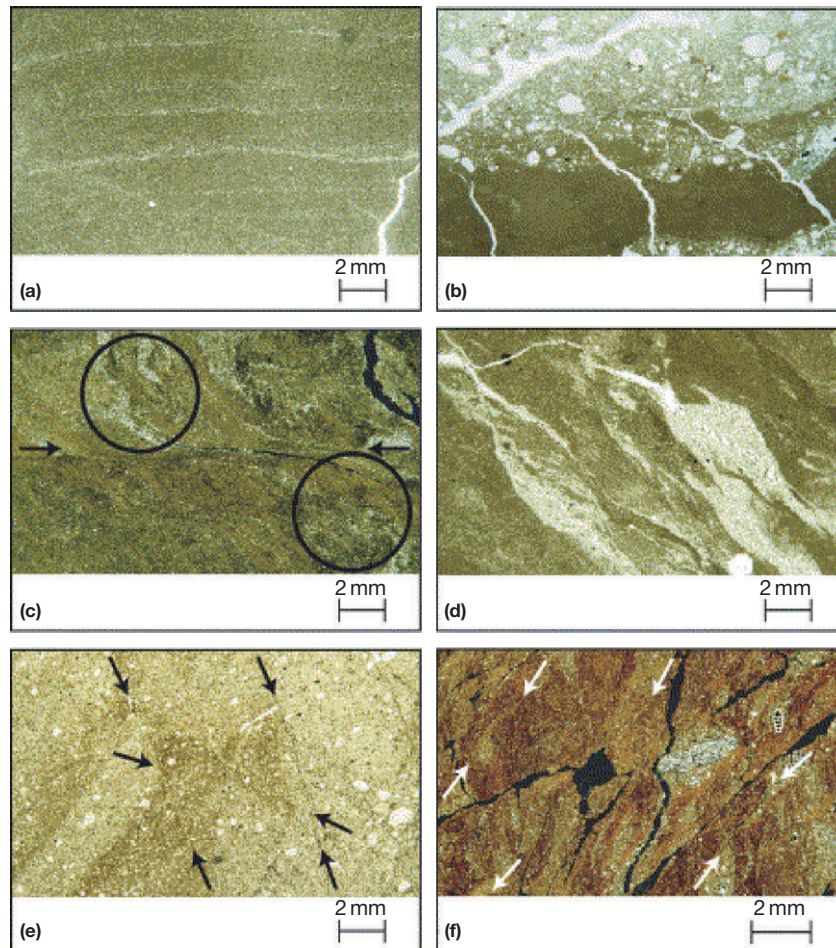


Figure 3 (a) Glacimarine cyclopel sequence recovered from an Alaskan fjord (see text for explanation). Plane-polarized light. (b) Example of a glacimarine deposit from the Antarctic continental shelf in which preexisting stratification is largely preserved (with the exception of a flamelike structure on the right hand side of the image). This reflects a relatively minor disturbance in a low-strain situation. (c) Higher-strain situation in which stratification in the areas indicated was once connected (now separated as a result of discrete shearing: arrowed plane). Also note birefringence in the plasma, both within the shear plane and in the stratified parts (see below). Example from glacimarine fjord deposit, Alaska. Cross-polarized light. (d) Example of pseudo-lamination caused by simple shear. Note streamlined, boudinaged silty laminae. Example from Alaska: glacimarine fjord deposit in plane-polarized light. (e) Disrupted stratification in part of the CIROS-1 sediment core (Antarctica). Brittle deformation: single joint and two faults. Plane-polarized light. (f) Example from a glacitectedonized glacimarine deposit from Spitsbergen showing complex ductile and brittle deformation comprising superimposed joints, faults and conjugate shears (as reflected by planar fractures), and discrete zones of plasmic fabric (arrows). Note that the original, primary stratification can still be discerned. Cross-polarized light.

higher strain occurs, any stratification present is more likely to have developed through attenuation of inhomogeneities in the deforming (subglacial traction) layer of sediment or glaciectonite (Figure 3(c); cf. Benn and Evans, 1996; Evans et al., 2006; see Tills). Such inhomogeneities are commonly introduced into the deforming layer from the immobile sediment substrate below. Once incorporated, inhomogeneities may be boudinaged, that is, develop into sausagelike forms, and eventually be stretched into streamlined lenses and tectonic laminae (Figure 3(d)). Alternatively, incorporations may be folded and rotated, which could lead to the formation of rounded intraclasts (see above).

Clearly, the type of (micro)structures that will develop is largely dependent on the type and nature of the deformation. In glacier submarginal or proglacial settings, sediments are generally compressed, which leads to folding, whereas in subglacial

settings, extensional regimes are prevalent, which means that attenuation is more common. It should be noted, however, that because glacial sediment may undergo several cycles of extension and compression, resultant structural patterns can be extremely complex and may comprise various superimposed features. Microscopic analysis of such deformed sediments greatly facilitates the unraveling of their history and thus aid the interpretation of glacial sequences (Phillips and Auton, 2000; Phillips et al., 2007; Figure 4). Superimposed structures not only reflect consecutive stages of extensional and compressive deformation but may also reflect changes in deformation style and intensity. The latter changes are controlled, for example, by clay and water content (see Phillips et al., 2007; van der Meer et al., 2003), which determine whether a sediment responds in essentially a brittle or a ductile fashion, in other words, whether structures such as fissility, joints, and faults or

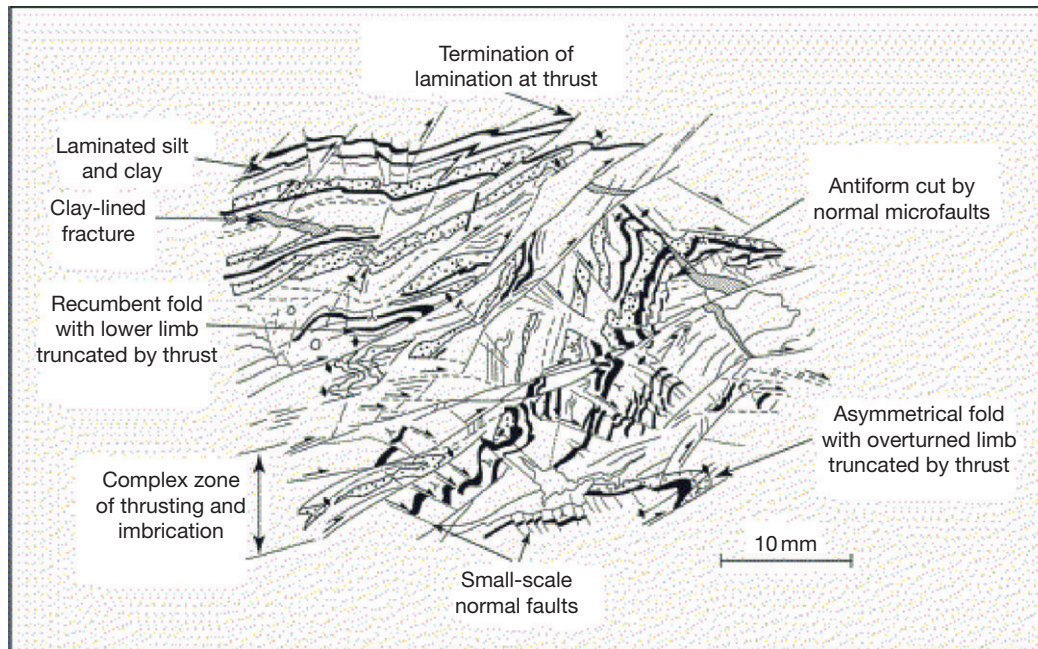


Figure 4 Detailed cartoon of a thin section from Strathspey, Scotland, showing a range of microstructures developed in a laminated fine-grained deposit. Note the complex structural geometry, indicating polyphase deformation. Reworking of primary lamination involved folding, thrusting, and faulting, both normal and reverse. Reprinted from Phillips ER and Auton CA (2000) Micromorphological evidence for polyphase deformation of glaciolacustrine sediments from Strathspey, Scotland. In: Maltman AJ, Hubbard B, and Hambrey MJ (eds.) *Deformation of Glacial Materials*, pp. 279–292. London: Geological Society. Special Publication 176. © 2000, with permission from The Geological Society.

crenulations, folds, and pressure shadows (areas of low strain, associated with flow around rigid objects) dominate (Figures 1 and 3(e)–3(f); see Menzies, 2000).

Where sediments are strained to an extent that all primary sedimentary features are erased and all inhomogeneities are stretched or reworked beyond macroscopic recognition, micromorphology can be indispensable. In such cases, very subtle arrangements of skeleton grains often present the only clues for unraveling a sediment's history. Seemingly insignificant microstructures such as lineaments and turbates, that is, linear, respectively circular arrangements of elongated skeleton grains (Figure 5(a) and 5(b)), have been shown to reflect particle reorientations in response to simple shear (Hiemstra and Rijdsdijk, 2003) and are now quite confidently used to distinguish between deformed and undeformed glacial sediments.

Microfabric Analysis

The occurrence of turbates and lineaments in deformed glacial sediments clearly illustrates that individual grains tend to readjust their positions when subjected to stresses (cf. Larsen et al., 2007). This is the essence of microfabric analysis, which involves the examination of the arrangement and orientation of particles at a microscopic scale. The objective is to reconstruct stress vectors at the time of formation and thus to learn about the depositional environment. Although size limits have never been defined exactly, it is common practice in microfabric analysis of thin sections to examine particles ~100 to ~6 mm in size. The lower limit is chiefly determined by resolution, whereas the upper limit is conveniently set to complement

macrofabric analysis as carried out in the field (see Till Fabric Analysis).

Traditionally, microfabric data have been used to infer former ice flow directions from subglacial sediments (e.g., Evenson, 1971). The idea is that fine particles become aligned with their long axes predominantly parallel to the ice flow while displaying dips in upglacier directions (cf. Thomason and Iverson, 2006). However, it has also been suggested that certain particles may attain positions transverse, that is, perpendicular, to ice flow, lending support to the idea that apart from depositional and deformational processes, the shape and size of particles may also have some control over how they become oriented (Carr and Rose, 2003; Kjør and Krüger, 1998; Figure 6).

In contrast to macrofabric analyses (see Till Fabric Analysis), where clasts can be removed from a sediment exposure to check the degree of elongation and their dimensions and to establish accurately their orientation and dip (or compass bearing and plunge), microfabric measurements have to be taken either directly from a two-dimensional (2D) thin section or from a projected image, which means that orientations of perceived elongated particles are always apparent rather than true. The most common solution to this problem is to produce two or three mutually orthogonal thin sections, perform measurements in all planes, and work out the true microfabric characteristics mathematically (Evenson, 1971).

This method is quite laborious, however, and prone to subjective bias in that the operator has to select the elongated particles. In fact, such limitations have probably been the reason why microfabric analysis is not yet a standard procedure in glacial micromorphology. However, the introduction of

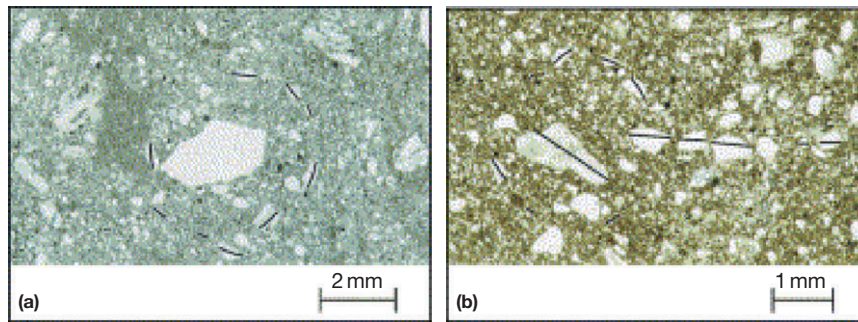


Figure 5 (a) Example of a turbate structure in which the long axes of fine grains are positioned parallel to the surface of the large grain in the center of the image. This structure suggests that the large grain, under influence of shear stresses, has performed a rotation, dragging the fines in its vicinity with it and forcing them to realign. Sample from the Antarctic continental shelf. Plane-polarized light. (b) Example of a lineament structure, which consists of at least five elongated grains which have their long axes in the same (shear) plane. Note that the larger grain on the left has a turbate structure around it, which illustrates the close association between the two structures in terms of genesis. Sample from the Antarctic continental shelf. Plane-polarized light.

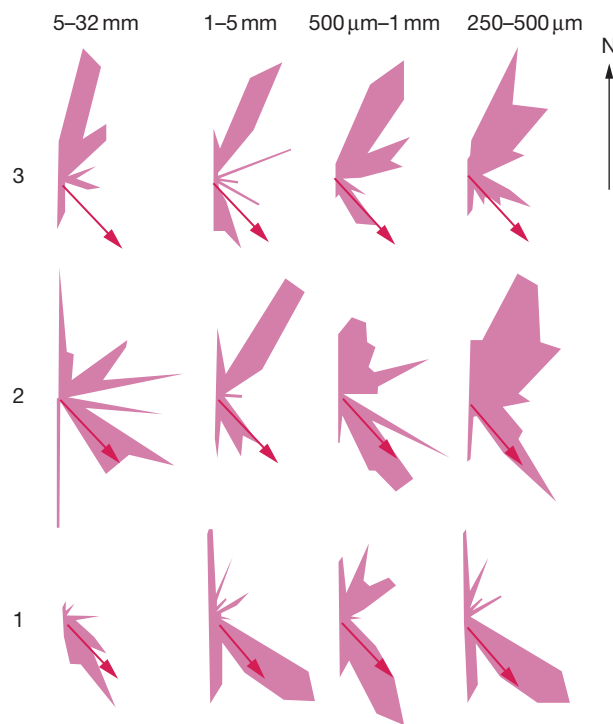


Figure 6 Fabric rose diagrams of three glacial till samples from Balglass, Scotland. Plotted are the 2D orientations of particles in four textural classes. The inferred local ice flow direction is indicated by the red arrow. Carr and Rose (2003) suggested that both stress history and size class may influence particle orientation patterns. Note both transverse and parallel signals. Reprinted from Carr SJ and Rose J (2003) Till fabric patterns and significance: Particle response to subglacial stress. *Quaternary Science Reviews* 22(14): 1415–1426. © 2003, with permission from Elsevier.

automated (i.e., computer-based) analyses has shown potential in facilitating the study of microfabric anisotropy. Derbyshire et al. (1992), for example, introduced the use of Fourier series, and more recently, Stroeven et al. (2005) employed stereological principles demonstrating that the time efficiency and objectivity of the method can be improved dramatically. Stereology, by definition, is a sampling theory for

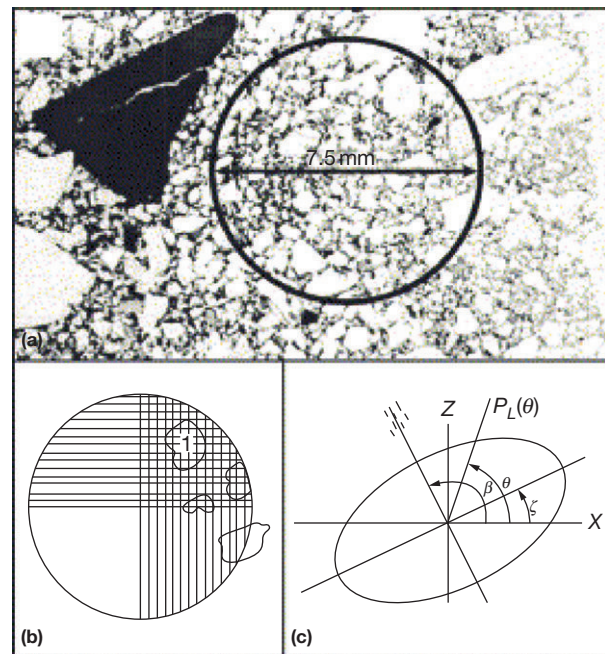


Figure 7 Directed secants analysis. (a) Circled area in a thin section image is subjected to systematic line scanning, which entails the counting of the number of intersections between the secants and the particle traces for each 10° of rotation. (b) The situations at $\theta = 0^\circ$ and $\theta = 90^\circ$ are illustrated. (c) A schematic rose of intersection densities, which is used to identify the preferred fabric orientation of the sample under consideration. Reprinted from Stroeven AP, Stroeven P, and van der Meer JJM (2005) Microfabric analysis by manual and automated stereological procedures: A methodological approach to Antarctic tillite. *Sedimentology* 52: 219–233. © 2005, with permission from Wiley-Blackwell and the International Association of Sedimentologists.

populations having a geometric structure and, in its application to microfabric, allows for the derivation of preferred orientations from measurements on particles in two projection planes (i.e., two orthogonal thin sections). In practice, the orthogonal set of thin sections is subjected to directed secant analysis (Figure 7), in which the number of intersections

between particle traces and a rotating line system is determined. With this information, 3D preferred orientations of the particles can be easily calculated using a set of straightforward trigonometric relations (Stroeve et al., 2005).

Plasmic Fabric Analysis

Plasmic fabric refers to the arrangement of particles $< \sim 30 \mu\text{m}$. Although resolution limitations preclude the study of individual particles (see above), certain zones or domains consisting

of aligned particles may show interference colors under cross- or circularly polarized light, which allows preferred orientations to be studied even at plasma level. As with larger skeleton grains, particles in the plasma (notably clay platelets) tend to accommodate stresses applied during deposition or deformation by (re)adjusting their positions so that plasmic fabric can be used in much the same way as microfabric.

The interference color of the plasma depends on the thickness of the thin section and an optical property called birefringence, which is defined as the difference between the two

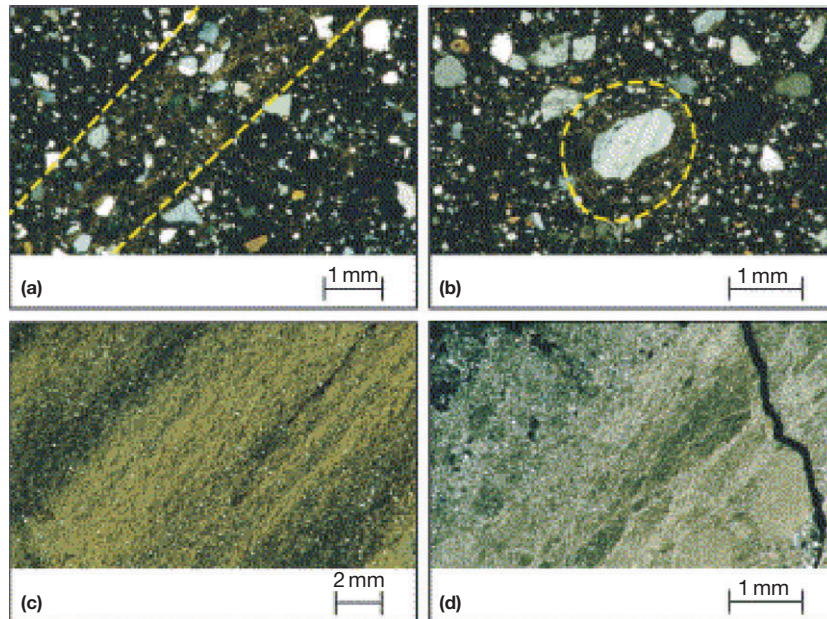


Figure 8 (a) Image of a till recovered from the Antarctic continental shelf showing a band of masepic plasmic fabric. Note that within the indicated zone, birefringence is high, whereas outside, the plasma shows no preferred orientation. Cross-polarized light. (b) Light grain in the center of the image shows a well-defined, thick halo of skelsepic plasmic fabric. Note that the dark grain to the left shows a zone of skelsepic plasmic fabric which is very much thinner. This suggests that this type of birefringence may result from different processes (see text). Cross-polarized light. (c) Pervasive type of masepic plasmic fabric conforming to the lamination in a fine-grained diamicton recovered from the Antarctic continental shelf. Cross-polarized light. (d) Another example from the Antarctic continental shelf showing a shear zone (from lower left to upper right of the image) with thin, continuous unistrial plasmic fabric representing discrete, anastomosing shear planes. Cross-polarized light. Compare shear zone geometry with patterns in [Figure 10](#).

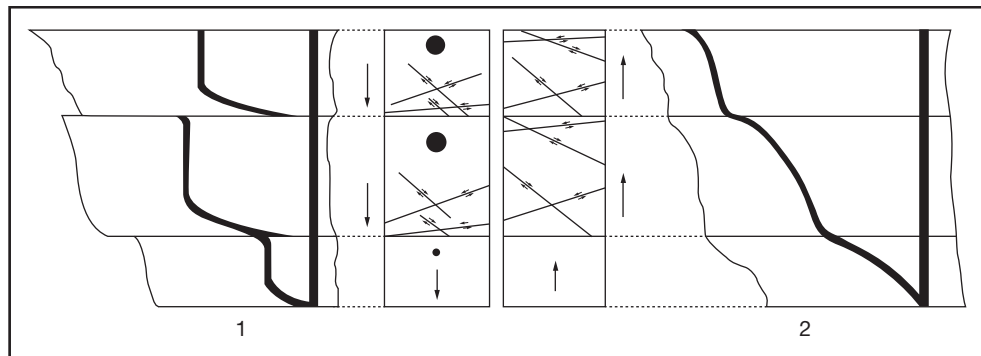


Figure 9 Conceptual model illustrating the strain signature in sediments affected by (1) subaqueous gravity-driven sliding and (2) subglacial deformation. [Hiemstra et al. \(2004\)](#) observed reversed trends in deformation profiles and used these to distinguish between different types of deformation. Note theoretical shear zone configuration (conjugate shear planes in central columns) and black strain markers. Adapted from [Hiemstra JF, Zaniewski K, Powell RD, and Cowan EA \(2004\)](#) Strain signatures of fjord sediment sliding: Micro-scale examples from Yakutat Bay and Glacier Bay, Alaska, USA. *Journal of Sedimentary Research* 74: 760–769. © 2004, with permission from SEPM (Society for Sedimentary Geology).

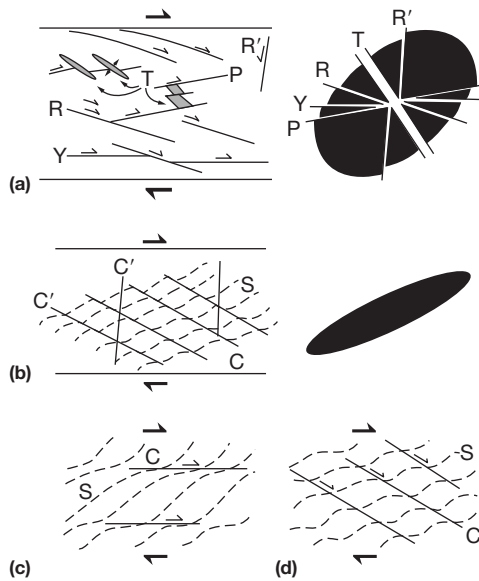


Figure 10 (a) Geometry in a dextral brittle shear zone, showing directions in Riedel shears (Y, P, R, and R') and tension fractures (T). Also shown in (a) is the brittle strain ellipse. (b–d) Possible geometries in a dextral ductile shear zone: S–C–C' fabric, S–C fabric, and S–C' fabric (S stretching direction, C and C' extensional crenulation cleavages). Also shown in (b) is the ductile strain ellipse. Compare with [Figure 8\(d\)](#). Reprinted and adapted from [van der Wateren FM, Kluiving SJ, and Bartek LR \(2000\)](#) Kinematic indicators of subglacial shearing. In: Maltman AJ, Hubbard B, and Hambrey MJ (eds.) *Deformation of Glacial Materials*, pp. 259–278. London: Geological Society. Special Publication 176. © 2000, with permission from The Geological Society.

refractive indices of the anisotropic domain ([Cribble and Hall, 1992](#); [Figure 8\(a\)](#)). Not only is the intensity or brightness of the colors is dependent on the degree of particle orientation, but it is also controlled by the orientation of the thin section relative to these particle orientations. Maximum birefringence is displayed only if both the maximum and minimum refractive indices of a particular anisotropic domain lie within the plane of the thin section.

Following [Brewer's \(1976\)](#) terminology for microscopic phenomena in soils (see above), many glacial micromorphologists employ terms such as skelsepic, unistrial, and masepic plasmic fabric to describe birefringence patterns (see also [Figure 1](#)). Skelsepic plasmic fabric is defined as plasma separations occurring subcutanically around skeleton grains, that is, where the birefringence occurs in haloes parallel to the surface of the grains ([Figure 8\(b\)](#)). There are at least three explanations for this type of plasmic fabric. The first is that clay platelets become realigned due to torque-generated rotational movements of individual skeleton grains within the plasma (e.g., [van der Meer, 1993](#)), an idea that finds support in the frequent observations of their concurrence with turbate structures (e.g., [Hiemstra and Rijdsdijk, 2003](#)). The second explanation is that skelsepic plasmic fabric results from translocation of clays through pore-water movements in simple shear-induced dilating sediment ([Menzies, 2000](#)), while a third proposes swelling of clays during repeated drying–wetting cycles as a possible explanation (e.g., [Jim, 1990](#)). Assuming that all three explanations are equally plausible, it would mean that both anisotropic and isotropic shear regimes could be responsible for the alignments and that, although quite common in, for example, subglacially deformed tills, skelsepic plasmic fabrics are by no means diagnostic of this type of sediment.

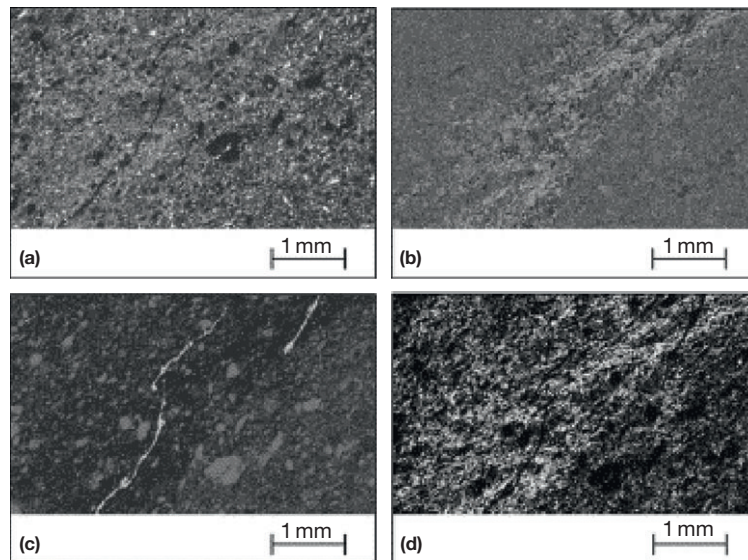


Figure 11 Application of a multispectral classification algorithm to identify the main components of a cross-polarized image of a glacial sediment (unprocessed image (a)). (b) The same image after the first stage of classification. Different shades of gray represent objects of similar appearance. (c) The image after the next stage clearly highlights fractures (white), plasma (dark gray), skeleton grains (lighter shades of gray), and unknown matter (black). (d) The image through a birefringent plasma mask (based on differences in red and green values for individual pixels), allowing identification of plasmic fabric patterns. Reprinted from [Zaniewski K, van der Meer JJM \(2005\)](#) Quantification of plasmic fabric through image analysis. *Catena* 63(1): 109–127, © 2005, with permission from Elsevier.

In fact, this applies to all plasmic fabric types. To date, not one birefringence pattern has been identified that could unequivocally be related to a singular depositional or deformational process. Nevertheless, plasmic fabric analysis may be regarded as one of the more powerful tools in micromorphology. Particularly in deformed glacial sediments, the occurrence of so-called striated birefringence usually provides crucial information about how strain evolved in response to progressive simple shear. Types such as masepic plasmic fabric and its more continuous variant, unistrial plasmic fabric, which show streaks or lines of uniformly oriented plasma (**Figures 1 and 8(c) and 8(d)**), can be analyzed in terms of geometry and lead to very specific information on the type and direction of shear, even if the signature can be studied only in one single thin section plane (**Hiemstra et al., 2004**; see **Figure 9**). Noteworthy in this respect is the application of kinematic principles to striated birefringence analysis (**van der Wateren et al., 2000**). This approach makes use of analogies with deformation structures found in low-grade metamorphic rocks (see **Passchier and Trouw, 1996**). **Van der Wateren et al. (2000)** explained how deformed glacial sediments may reveal plasmic fabrics that are very similar to Riedel shear patterns and ductile shear band cleavages in, for example, mylonites, a similarity which thus allows for the reconstruction of finite strain ellipsoids and for drawing inferences with regard to shear type, direction, and geometry (**Figure 10**; cf. **Larsen et al., 2006; 2007; Thomason and Iverson, 2006**).

It should be noted that, as with microfabric analysis, there is obviously an element of bias involved in plasmic fabric analysis. It is not unlikely that the relative strength of plasmic fabric – potentially useful in the quantification of strain – is rated differently by different observers, and secondly, it is quite possible that subtle birefringence patterns go undetected due to human eyesight limitations. Again, the use of image analysis techniques seems to be the way forward. Through the use of multispectral image classification, for example (see **Figure 11**), **Zaniewski and van der Meer (2005)** made an attempt to control the subjectivity in plasmic fabric analysis. Although the authors themselves state that the quantification of birefringence is still very much in its infancy, there seems to be merit in automated approaches.

Recent and Future Developments in Microscopic Studies

In recent years, improvements to the technique of micromorphology have involved developments in the areas of objectivity and efficiency. These were mainly achieved through automated, that is, computer-assisted analyses of thin section samples (as described earlier: stereology – microfabric, image analysis – plasmic fabric). However, there have also been successful attempts to make micromorphology less ‘descriptive,’ less subjective, and more systematic through ‘manual’ quantification (counting and calculation of microstructure density: **Larsen et al., 2006, 2007; Reinardy et al., 2011**) and the use of multiple observers to reduce or overcome observer bias (**Larsen et al., 2007**).

With regard to observer bias, it is worth noting that there is an increasing awareness that psychology may be more significant in thin section analysis than previously thought. **Leighton et al. (in press)** are looking into the possibility that human perception and identification of patterns (such as turbate microstructures) are at least partially based upon (and controlled by) previous experience and conditioning of the brain.

Another interesting development is, and has been, that of the introduction of microtomography. Although so far few publications have reported on its application to glacial sediments (**Kilfeather and van der Meer, 2008; Tarplee et al., 2011**), the technique may be expected to complement micromorphology in the future, thus allowing for visualization and reconstruction of 3D phenomena on the basis of 2D information.

Microtomography is a technique that is similar to CAT scan imaging as used in medicine but achieves a much higher resolution of $\sim 2\ \mu\text{m}$. It involves exposing natural, untreated sediment samples to high-power x-radiation and obtaining projected images (radiographs) of these samples in planes that are perpendicular to the x-ray beam direction. By rotating individual samples through small angular increments, series of several hundred of such radiograms are acquired, which via mathematical algorithms can be converted into 3D images (**Figure 12**).

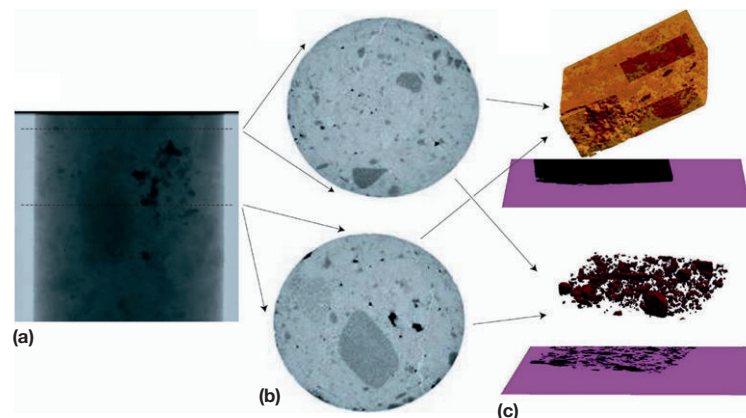


Figure 12 x-Ray computed microtomography. (a) Example of a radiograph of a cylindrical till sample; a total of 400 radiographs cover the entire sample (rotating through 360° at steps of 0.9°). (b) Two examples of reconstructed slices through the cylinder; approximately 1000 of these are reconstructed along the vertical sample axis. (c) Examples of reconstructed 3D blocks of part of the cylinder with pores spaces highlighted. Image courtesy of J.J.M. van der Meer.

Tomographic images are composed of millions of volumetric elements (voxels), each carrying a value representing the amount of x-radiation absorbed. The 3D images can be sliced at any given angle to show cross sections, and every single image – both 2D and 3D – can obviously be subjected to a range of image analysis techniques allowing for detailed investigation of, for example, internal structures, reflected in, for example, fabrics or pore spaces (Figure 12).

See also: **Glacial Landforms:** Glacitectonic Structures and Landforms. **Glacial Landforms, Sediments:** Clast Form Analysis; Glaciogenic Lithofacies; Glaciomarine Sediments and Ice-Rafted Debris; Till Fabric Analysis; Tills. **Paleosols and Wind-Blown Sediments:** Soil Micromorphology.

References

- Benn DI and Evans DJA (1996) The interpretation and classification of subglacially deformed materials. *Quaternary Science Reviews* 15: 23–52.
- Brewer R (1976) *Fabric and Mineral Analysis of Soils*. Huntington, NY: Krieger.
- Carr SJ and Rose J (2003) Till fabric patterns and significance: Particle response to subglacial stress. *Quaternary Science Reviews* 22: 1415–1426.
- Derbyshire E, Unwin DJ, Fang XM, and Langford M (1992) The Fourier-frequency-domain representation of sediment fabric anisotropy. *Computers and Geosciences* 18: 63–73.
- Evans DJA, Phillips ER, Hiemstra JF, and Auton CA (2006) Subglacial till: Formation, sedimentary characteristics and classification. *Earth-Science Reviews* 78(1–2): 115–176.
- Evenson EB (1971) The relationship of macro- and micro-fabric of till and the genesis of subglacial landforms in Jefferson County, Wisconsin. In: Goldthwait RP (ed.) *Till: A Symposium*, pp. 345–364. Columbus, OH: Ohio State University Press.
- Gribble CD and Hall AJ (1992) *Optical Mineralogy: Principles and Practice*. London: UCL Press.
- Hiemstra JF (2001) Microscopic analyses of Quaternary glaciogenic sediments of Marguerite Bay, Antarctic Peninsula. *Arctic, Antarctic, and Alpine Research* 33(3): 258–265.
- Hiemstra JF and Rijdsdijk KF (2003) Observing artificially induced strain: Implications for subglacial deformation. *Journal of Quaternary Science* 18: 373–383.
- Hiemstra JF, Zaniewski K, Powell RD, and Cowan EA (2004) Strain signatures of fjord sediment sliding: Micro-scale examples from Yakutat Bay and Glacier Bay, Alaska, USA. *Journal of Sedimentary Research* 74: 760–769.
- Jim CY (1990) Stress, shear deformation and micromorphological clay orientation: A synthesis of various concepts. *Catena* 17: 431–447.
- Kilfeather AA, Ó cofaigh C, Dowdeswell JA, van der Meer JJM, and Evans DJA (2010) Micromorphological characteristics of glaciomarine sediments: Implications for distinguishing genetic processes of massive diamicts. *Geo-Marine Letters* 30: 77–97.
- Kilfeather AA and van der Meer JJM (2008) Pore size, shape and connectivity in tills and their relationship to deformation processes. *Quaternary Science Reviews* 27: 250–266.
- Kjær KH and Krüger J (1998) Does clast size influence fabric strength? *Journal of Sedimentary Research* 68: 746–749.
- Larsen NK, Piotrowski JA, and Christiansen F (2006) Microstructures and microshears as proxy for strain in subglacial diamicts: Implications for basal till formation. *Geology* 34: 889–892.
- Larsen NK, Piotrowski JA, and Menzies J (2007) Microstructural evidence of low-strain, time-transgressive subglacial deformation. *Journal of Quaternary Science* 22: 593–608.
- Leighton ID, Hiemstra JF, and Weidemann C (in press) Recognition of micro-scale deformation structures in glacial sediments – Pattern perception, observer bias and the influence of experience. *Boreas*. <http://dx.doi.org/10.1111/j.1502-3885.2011.00246.x>.
- Menzies J (2000) Micromorphological analyses of microfabrics and microstructures indicative of deformation processes in glacial sediments. *Deformation of Glacial Materials*, pp. 245–257. London: Geological Society. Special Publication 176.
- Menzies J, van der Meer JJM, Domack E, and Wellner JS (2010) Micromorphology: As a tool in detection, analyses and interpretation of (glacial) sediments and man-made materials. *Proceedings of the Geologists' Association* 121: 281–292.
- Murphy CP (1986) *Thin Section Preparation of Soils and Sediments*. Berkhamsted: AB Academic.
- Palmer AP, Rose J, Lowe JJ, and MacLeod A (2010) Annually resolved events of Younger Dryas glaciation in Lochaber (Glen Roy and Glen Spean), western Scottish Highlands. *Journal of Quaternary Science* 25: 581–596.
- Passchier CW and Trouw RAJ (1996) *Microtectonics*. Berlin: Springer.
- Phillips ER and Auton CA (2000) Micromorphological evidence for polyphase deformation of glaciolacustrine sediments from Strathspey, Scotland. *Deformation of Glacial Materials*, pp. 279–292. London: Geological Society. Special Publication 176.
- Phillips ER, Merritt J, Auton CA, and Golledge N (2007) Microstructures in subglacial and proglacial sediments: Understanding faults, folds and fabrics, and the influence of water on the style of deformation. *Quaternary Science Reviews* 26: 1499–1528.
- Reinardy BT, Hiemstra JF, Murray T, Hillenbrand CD, and Larter RD (2011) Till genesis at the bed of an Antarctic Peninsula palaeo-ice stream as indicated by micromorphological analysis. *Boreas* 40: 498–517.
- Stroeven AP, Stroeven P, and van der Meer JJM (2005) Microfabric analysis by manual and automated stereological procedures: A methodological approach to Antarctic tillite. *Sedimentology* 52: 219–234.
- Tarplee MFV, van der Meer JJM, and Davis GR (2011) The 3D microscopic 'signature' of strain within glacial sediments revealed using X-ray computed microtomography. *Quaternary Science Reviews* 30: 3501–3532.
- Thomason JF and Iverson NR (2006) Microfabric and microshear evolution in deformed till. *Quaternary Science Reviews* 25: 1027–1038.
- van der Meer JJM (1987) Micromorphology of glacial sediments as a tool in distinguishing genetic varieties of till. *Geological Survey of Finland, Special Paper* 3: 77–89.
- van der Meer JJM (1993) Microscopic evidence of subglacial deformation. *Quaternary Science Reviews* 12: 553–587.
- van der Meer JJM, Menzies J, and Rose J (2003) Subglacial till: The deforming glacier bed. *Quaternary Science Reviews* 22: 1659–1685.
- van der Wateren FM, Kluiving SJ, and Bartek LR (2000) Kinematic indicators of subglacial shearing. *Deformation of Glacial Materials*, pp. 259–278. London: Geological Society. Special Publication 176.
- Zaniewski K and van der Meer JJM (2005) Quantification of plasmic fabric through image analysis. *Catena* 63: 109–127.

Tills

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Subglacial Processes

The observations and data available in the existing literature on till sedimentology allow us to identify four main processes of till production in the subglacial environment, including sliding/lodgement, melt-out, deformation, and cavity filling. Additionally, the process of ploughing and the concept of a lodgement–deformation continuum can potentially produce a recognizable till end member.

Subglacial Sliding and Lodgement

Subglacial sliding over a rigid bed is well documented, and soft-bed sliding has been proposed to explain the characteristics of certain till sequences (e.g., Brown et al., 1987; Clark and Hansel, 1989; Piotrowski and Tulaczyk, 1999). This involves glacier bed decoupling due to increased basal water pressures, which prevent the transmission of stress to the soft substrate (Fischer and Clarke, 1997; Fuller and Murray, 2000). Spatial variability in the intensity of sliding may explain the characteristics of certain tills, as proposed by Piotrowski and Kraus (1997), who further suggest that the glacier bed is a mosaic of sliding bed conditions and deforming patches or spots (Figure 1). This is supported by Thorsteinsson and Raymond (2000), who suggest that sliding must be the dominant mechanism of basal motion beneath an ice stream, as opposed to till shearing, unless there is either adhesion of till to the ice sole, short-scale roughness at the ice–till interface, or internal slip boundaries in the till.

Lodgement is defined as the plastering of glacial debris from the base of a sliding glacier on to a rigid or semirigid bed by pressure melting and/or other mechanical processes (Dreimanis, 1989). This includes the lodgement of debris-rich basal ice masses as well as individual clasts (Figure 2). The lodging of preexisting till masses gives rise to the term ‘deformed lodgement till’ (Dowdeswell and Sharp, 1986). The lodgement process preferentially deposits larger clasts while finer grain sizes remain in motion due to deformation (see section ‘Lodgement–Deformation Continuum’). This poses problems when explaining thick fine-grained till sequences due to lodgement. Consequently, a range of contemporaneous subglacial processes have been proposed for temperate glaciers in order to explain the accumulation of thick tills, including simultaneous lodging, ploughing, and shear deformation at the glacier bed, whereby large particles lodge first and smaller particles remain in traction until strain rates fall (Benn, 1994; Boulton and Hindmarsh, 1987; Boulton et al., 1974; Evans and Twigg, 2002; Kruger, 1994).

Subglacial Melt-Out

Subglacial melt-out is the slow and passive release of sediment from debris-rich stagnant basal glacier ice. The process is

defined by theoretical postulates based upon interpretations of ancient tills (e.g., Haldorsen and Shaw, 1982; Munro-Stasiuk, 2000; Piotrowski et al., 2001; Shaw, 1979, 1982, 1983; Figure 3), although support has been provided by observations at modern glacier snouts where basal tills have apparently been released from overlying debris-rich ice (Boulton, 1971; Lawson, 1979a,b; Shaw, 1982). High debris concentrations in melting ice result in low thaw consolidation ratios, and, if drainage is efficient, delicate englacial structures may be preserved (e.g., Carlson, 2004). However, the ideal conditions for melt-out till preservation are probably rare, because poor drainage and high ice contents will produce high porewater pressures in the accumulating melt-out sequence thereby initiating failure, remobilization, and dewatering (Paul and Eyles, 1990). Additionally, the paleoglaciological requirements for melt-out till production often appear unrealistic. For example, Boulton (1996) points out that in order to produce a melt-out till sequence in excess of 10 m (a commonly reported thickness) would require more than 100 m of debris-rich ice of average debris concentration, far in excess of any present-day basal ice sequences.

Subglacial Deformation

The subglacial deformation of sediments and soft rocks accounts for some of the forward motion of many glaciers (Alley, 1989a,b; Alley et al., 1986, 1987a,b; Boulton and Hindmarsh, 1987; Clarke, 1987; Humphrey et al., 1993). The Icelandic subglacial experiment reported by Boulton (1979), Boulton and Hindmarsh (1987), and Boulton and Dobbie (1998) and the high-resolution seismic surveys beneath Ice Stream B, Antarctica (Blankenship et al., 1986, 1987) are benchmark studies providing the most compelling case for substrate deformation. The Icelandic experiment revealed a convex-upward displacement curve in the till, marked by the displacement of strain markers, which was interpreted as viscous response to applied stress (Figure 4). Further studies at the base of modern glaciers have confirmed the role of deforming beds and highlighted the role of fluctuating porewater pressures in controlling sliding versus deformation temporally (e.g., Blake et al., 1992; Fischer et al., 1999, 2001; Iverson, 1999; Iverson et al., 1994, 1995, 1999). Laboratory experiments (e.g., Iverson and Iverson, 2001; Watts and Carr, 2002) have reproduced convex-upward displacement curves through plastic failure of till, with failure taking place along discrete planes, implying that such curves do not necessarily indicate a viscous rheology. The consensus on the rheological behavior of subglacial tills is that they behave plastically but undergo spatially distributed deformation, thereby producing a pattern of displacement that resembles a viscous material (Fowler, 2002, 2003).

Some stratified sediment bodies found in association with basal tills are related to subglacial drainage during till deposition, based upon theoretical reconstructions of subglacial

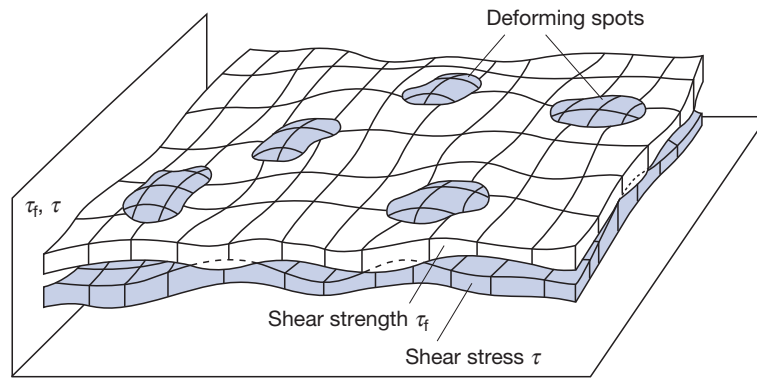


Figure 1 Conceptual model of the subglacial deforming/sliding bed mosaic proposed by Piotrowski and Kraus (1997). The model compares subglacial shear stress (τ) with sediment shear strength (τ_f). The shear stress is lower than the sediment shear strength wherever high basal water pressures cause ice–bed decoupling. Elsewhere the bed develops deforming spots.

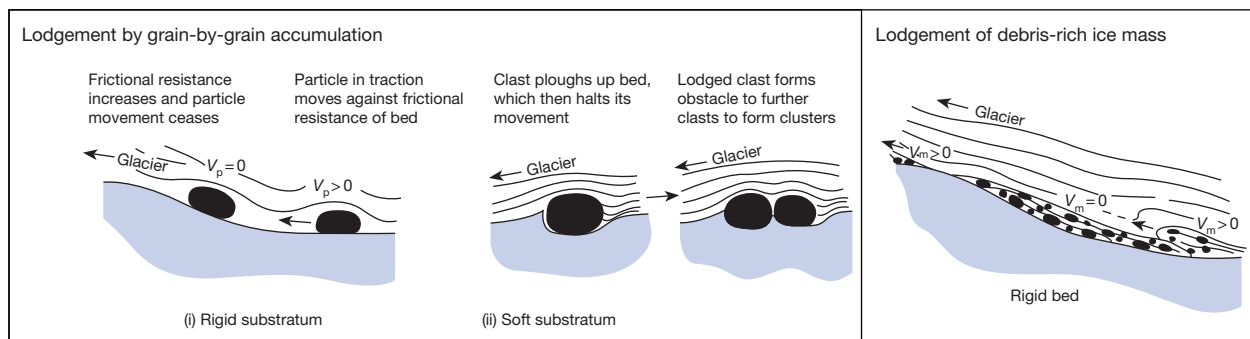


Figure 2 Schematic depictions of the lodgement process. Note that deformation is inextricably linked to lodgement on a soft substratum. Reproduced from Boulton GS (1982) Subglacial processes and the development of glacial bedforms. In: Davidson-Arnott R, Nickling W, and Fahey BD (eds.) *Research in Glacial, Glacio-fluvial and Glacio-lacustrine Systems*, pp. 1–31. Norwich: Geo Books.

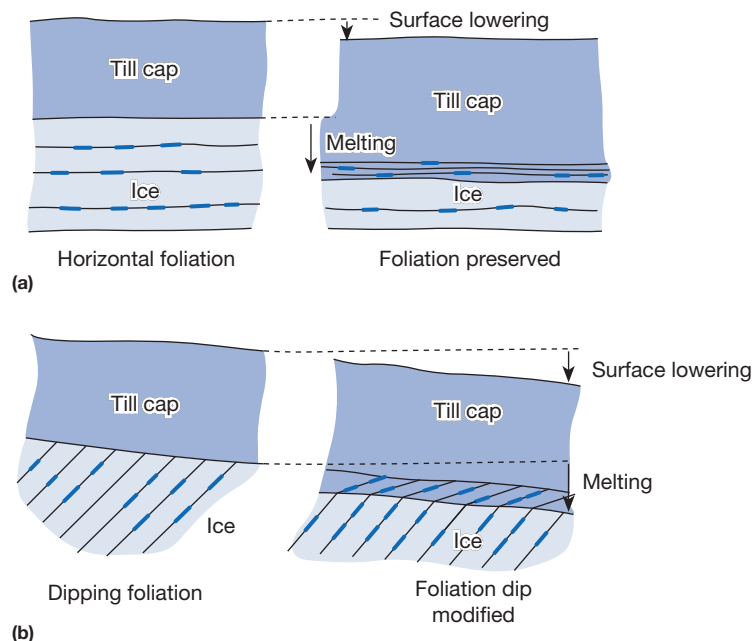


Figure 3 The theoretical concepts of preservation and modification of melt-out till in horizontally bedded (a) and dipping (b) debris-rich ice folia. Reproduced from Boulton GS (1971) Till genesis and fabric in Svalbard, Spitsbergen. In: Goldthwait RP (ed.) *Till: A Symposium*, pp. 41–72. Columbus: Ohio State University Press; Sugden DE and John BS (1976) *Glaciers and Landscape: A Geomorphological Approach*. New York, NY: John Wiley & Sons.

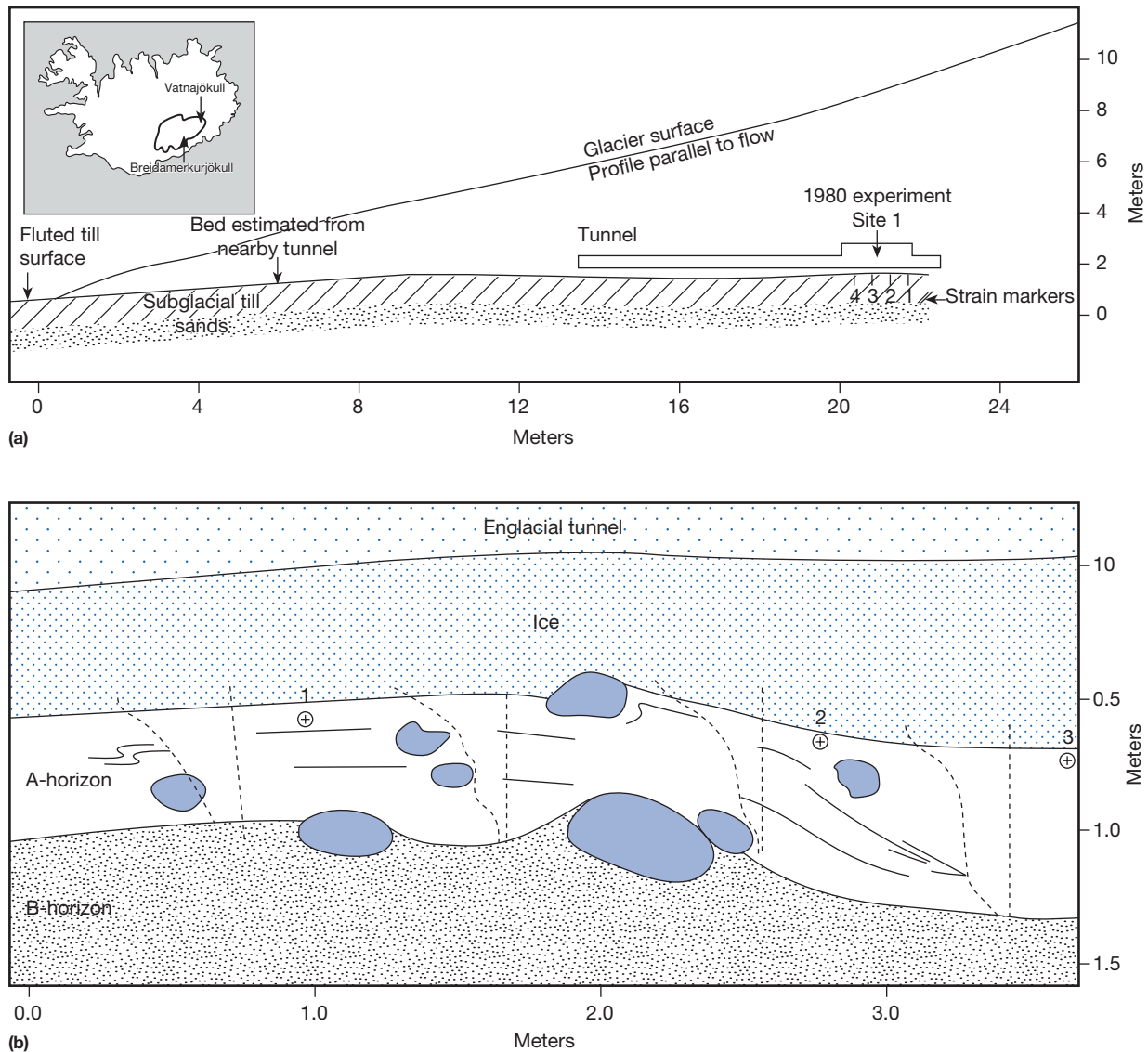


Figure 4 Location map and schematic diagram (from Murray, 1997; after Boulton and Hindmarsh, 1987) of the experimental tunnel below Breydamerkurjökull, Iceland (a) and summary diagram depicting the results of the deformation experiment conducted beneath the tunnel through the insertion of strain markers in the till (b). The dashed lines represent the positions of the strain markers at the beginning of the experiment and after 136 h. Pore pressure transducers are numbered 1–3 and sand wisps in the till A horizon are marked by thin lines. Large clasts are shaded by dark speckles.

hydrological conditions (Figure 5). Walder and Fowler (1994) suggest that drainage beneath a soft-bedded glacier occurs in a series of braided canals (Ng, 2000) at the ice–till interface when discharges are too large to be evacuated through the bed. This leads to ice–bed separation, thereby temporarily inducing sliding and shutting down till deformation/lodgement. Drainage of meltwater through a till, particularly one that is deforming, can occur either by Darcian porewater flow or by pipe flow. Many subglacial tills also contain vertical dykes or injection structures composed of nontill sediment. Such features are interpreted as the products of meltwater movement through till due to subglacial water escape (van der Meer et al., 1999) or hydrofracture fills relating to the burst out of pressurized groundwater (Boulton and Caban, 1995; Mandl and Harkness, 1987; Rijdsdijk et al., 1999).

Lee-Side Cavity Filling

The infilling of lee-side cavities, as observed at the irregular beds of modern glaciers with rigid substrates, involves a range of processes that interact with all of the above sediment emplacement mechanisms (Boulton, 1982; Boulton and Paul, 1976; Vivian, 1975). This includes slurring, expulsion and fall, decollement, and debris-rich ice ‘curl’ detachment and fluvial inwashing (Figure 6). The cavity-fill deposits are often heavily overprinted by the sedimentary and structural signatures of other till-forming processes due to the recoupling of glacier ice with its bed. Hence, Boulton (1971) regards cavity filling as essential to the smoothing of the irregularities in the bed of a glacier depositing till by other subglacial traction/deposition processes.

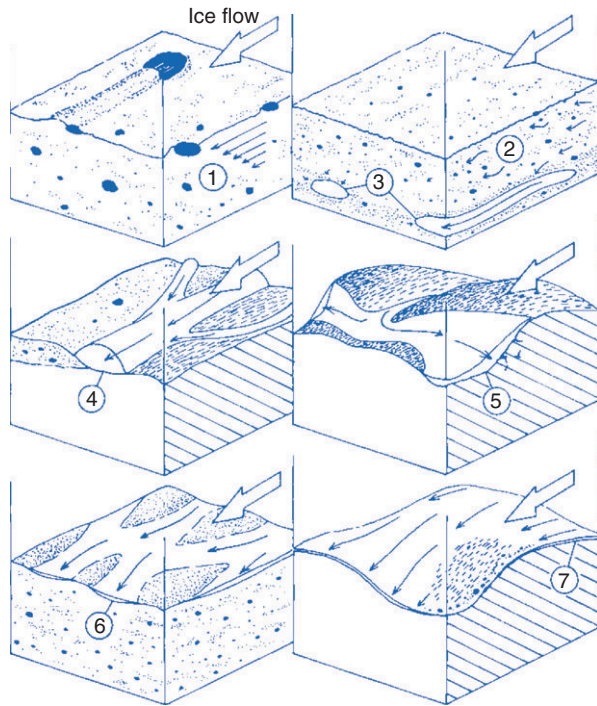


Figure 5 Various subglacial drainage pathways for rigid and deformable glacier beds: (1) bulk movement with deforming bed; (2) Darcian porewater flow; (3) pipe flow; (4) dendritic channel network; (5) linked cavity system; (6) braided canal network; and (7) water film at ice-bed interface. Reproduced from Benn DI and Evans DJA (1996) *The interpretation and classification of subglacially deformed materials. Quaternary Science Reviews* 15: 23–52.

Subglacial Ploughing

Similar to the ploughing of boulders through subglacial sediment (e.g., Brown et al., 1987; Clark and Hansel, 1989; Weertman, 1964), it has been hypothesized that the grooved base of a glacier can plough through soft substrates (Clark et al., 2003; Tulaczyk et al., 2001). The ice ploughing model has been proposed because previous models of subglacial deformation were increasingly unable to explain: (i) the results of laboratory experiments on subglacial tills, in which a Coulomb-plastic rheology has been demonstrated; and (ii) field measurements in which basal sliding and/or shallow deformation is apparent. Ploughing involves disturbance and transportation of subglacial till up to a few meters thick by the keels in a sliding, bumpy glacier sole. Tulaczyk et al. (2001) propose an analogy with the fault gouge model of Eyles and Boyce (1998) in which the sliding ice base acts as the rigid upper fault plate. The asperities in the fault plate, whether clasts or bumps/keels in the ice sole, plough through the soft substrate, generating the deformable till (fault gouge layer) above the underlying, rigid strata (lower fault plate).

Lodgement–Deformation Continuum

The spatial and temporal variability in ice–bed coupling due to changes in pressure distribution of the basal water system leads to variability in basal motion mechanics (Fischer and Clarke, 2001; Fischer et al., 1999). This gives rise to a patchwork of sticky spots and areas of stick–slip sliding, the extent of which vary in response to changes in the basal meltwater pressure. Such processes have been measured by Boulton

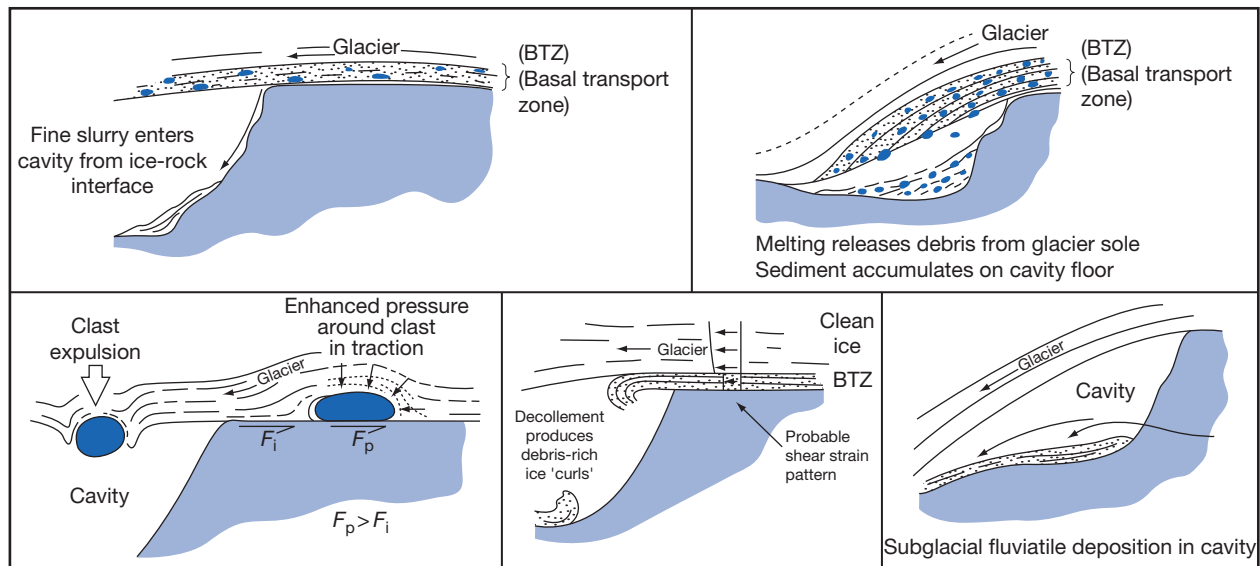


Figure 6 Schematic diagrams representing observations of sediment accumulation in subglacial cavities. Reproduced from Boulton GS (1982) *Subglacial processes and the development of glacial bedforms*. In: Davidson-Arnott R, Nickling W, and Fahey BD (eds.) *Research in Glacial, Glacio-fluvial and Glacio-lacustrine Systems*, pp. 1–31. Norwich: Geo Books.

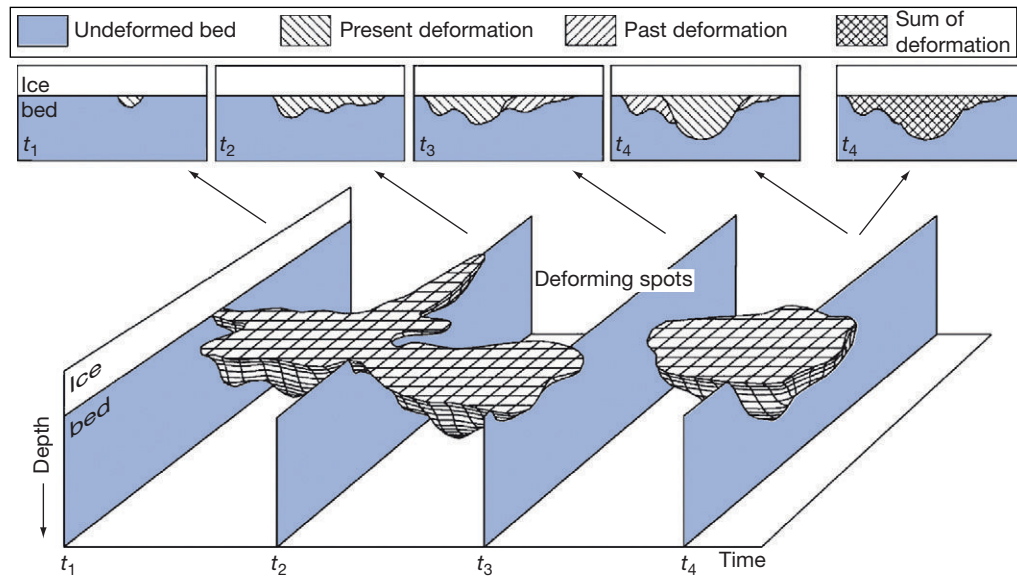


Figure 7 Schematic reconstruction of the concept of the stable/deforming subglacial bed mosaic concept of Piotrowski JA, Larsen NK, Frank FW, and Junge W (2004) Reflections on soft subglacial beds as a mosaic of deforming and stable spots. *Quaternary Science Reviews* 23: 993–1000. The position of the deforming spots on the bed changes spatially and temporally, producing overprinted deformation signatures in the subglacial material.

et al. (2001) who also document changes from sliding to deformation at specific locations. This can be reconciled with the spatial variability in sliding intensity proposed by Piotrowski and Kraus (1997) and the subglacial mosaic model of sliding and deformation proposed by Piotrowski et al. (1997) (Figure 7). Similarly, Lian and Hicock (2000) propose a change from lodgement to deformation to explain complex tills in the North American Cordillera.

Based upon the discussion presented above, it is clear that a combination of till transport and deposition processes is most likely in subglacial environments. A unifying model for subglacial till production must recognize the various processes observed in modern glacier systems and acknowledge the temporal and spatial variability in those processes at the glacier bed.



Figure 8 Highly fissile, compacted diamicton with bullet-shaped and faceted clasts, typical of subglacial lodgement till.

Subglacial Till Types

There are seven till types, some often found in association with stratified sediments, which have been proposed as the products of the processes discussed above, that is, lodgement till, melt-out till, sliding bed deposits, glacioteconite, communiton till, deformation till, and lee-side cavity fills. The geological signature of each of these till types is now reviewed in turn.

Lodgement Till

Lodgement till is defined as “sediment deposited by plastering of glacial debris from a sliding glacier sole due to the combined effects of pressure melting and frictional drag” (Chamberlin, 1895). It is either massive in appearance at outcrop or contains numerous subhorizontal joints which can be polished or slickensided and striated, indicating that they may be shear planes (Benn, 1994; Boulton, 1970; Figure 8). Grain-size characteristics and clast forms reflect the processes of grain crushing and

abrasion. Abundant stoss-and-lee forms are produced when clasts become lodged and then abraded and plucked, much like a miniature *roche moutonnée* (Boulton, 1978; Kruger, 1979, 1984, 1994; Sharp, 1982), although similar forms may be produced within deforming layers (Benn and Evans, 1996). Lodgement may produce clast clusters or pavements where lodged clasts cause the retardation of further debris (Boulton, 1975, 1982), although such pavements have also been interpreted as deformation products. Lodgement may also take place on a deformable bed and may be significant in the production of deformation tills whereby large clasts in the glacier sole plough through the substrate until their forward momentum is arrested; hence, the identification of a lodgement–deformation continuum. Modern lodgement tills clearly contain lodged particles with strong macrofabrics recording ice flow direction, ploughed particles, and shearing/deforming sediment (e.g., Benn, 1994; Kruger, 1994), the relative importance of lodgement and deformation in some instances varying with clast size

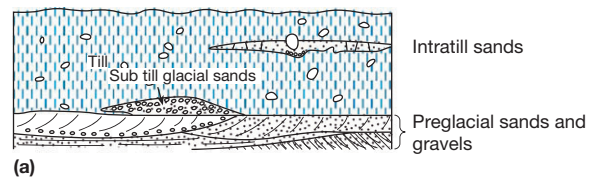
(e.g., Benn, 1994, 1995; Benn and Evans, 1996). Some significant problems of nomenclature have arisen from such observations (e.g., Ruszczyńska-Szenajch, 2001) and it is unclear what proportion of a subglacial diamicton must be lodged before it can be classified as a lodgement till rather than a deformation till. Unless some significant deformation and stacking of subglacial sediment occurs, it is difficult to envisage how the lodgement process *sensu stricto* can produce thick till sequences.

Subglacial Melt-Out Till

Melt-out till is defined as “sediment released by the melting of stagnant or slowly moving debris-rich glacier ice, and directly deposited without subsequent transport or deformation” (Benn and Evans, 1998). Few process studies have demonstrated melt-out till production (e.g., Ham and Mickelson, 1994; Lawson, 1979a,b, 1981), and consequently the characteristics of melt-out tills are largely based upon inferences from the geological record (e.g., Boulton, 1970, 1971; Haldorsen and Shaw, 1982; Mickelson, 1973; Ronnert and Mickelson, 1992; Shaw, 1982). Subglacial melt-out tills should inherit the foliation and structure of the parent ice but this is usually modified during the melt-out process, the least disturbed deposits being the product of sublimation (hence sublimation till of Shaw, 1977, 1988). Proposed examples of subglacial melt-out till display discontinuous and contorted layers and lenses; textural and compositional banding with indistinct contacts and internal flow structures; and undeformed, intratill soft sediment, or soft bedrock rafts (Lawson, 1979a,b; Shaw, 1979, 1982, 1983; Figure 9). Strong clast macrofabrics, recording the former glacier flow direction, are also regarded as diagnostic criteria (Lawson, 1979a,b). However, problems in differentiating ancient melt-out tills from the products of other subglacial processes persist, particularly where foliated or banded tills, interpreted by some as melt-out till (e.g., Clayton et al., 1989; Munro-Stasiuk, 2000; Ronnert and Mickelson, 1992), are regarded by others as the products of deformation (e.g., Alley, 1991; Boulton, 1996; Evans, 2000a).

Subglacial Sliding Bed Deposits

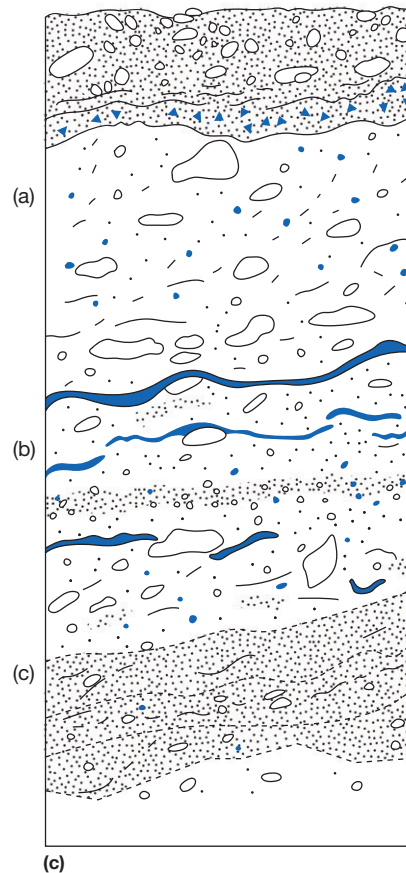
Occurrences of sub till and intratill stratified sediments have been interpreted as indicative of sliding glacier bed conditions (Figure 10), because they document deposition by a water film at the ice–bed interface (Piotrowski and Kraus, 1997; Piotrowski and Tulaczyk, 1999). This implies also that the tills found in association with the stratified units must be deposited by passive melt-out or that the stratified units record water films at the base of sliding tills (Hindmarsh, 1996). Undeformed, sub till stratified sediments are often found juxtaposed with deformed or mixed till–substrate contacts, suggesting that deforming layers and subglacial channels can coexist (Alley, 1992) and supporting the notion of a sliding and deforming glacier bed mosaic (Piotrowski and Kraus, 1997; Piotrowski et al., 2001, 2002) involving temporally varying sticky spots (Fischer and Clarke, 2001; Fischer et al., 1999). Therefore, till characteristics may change over short distances depending upon the predominant process at the time the glacial depositional system shut down. Stratified



(a)



(b)



(c)

Figure 9 Characteristic features of melt-out till: (a) soft sediment intratill lens, Alberta, Canada; (b) idealized vertical profile of typical features of proposed melt-out at the Matanuska Glacier, Alaska (reproduced from Lawson DE (1981) Distinguishing characteristics of diamictons at the margin of the Matanuska Glacier, Alaska. *Annals of Glaciology* 2: 78–84) – ‘a’ structureless pebbly, sandy silt; ‘b’ discontinuous laminae, stratified lenses, and pods of texturally distinct sediment in massive pebbly silt; ‘c’ layers of texturally, compositionally or color-contrasted sediment. Laminae or layers may drape over large clasts.

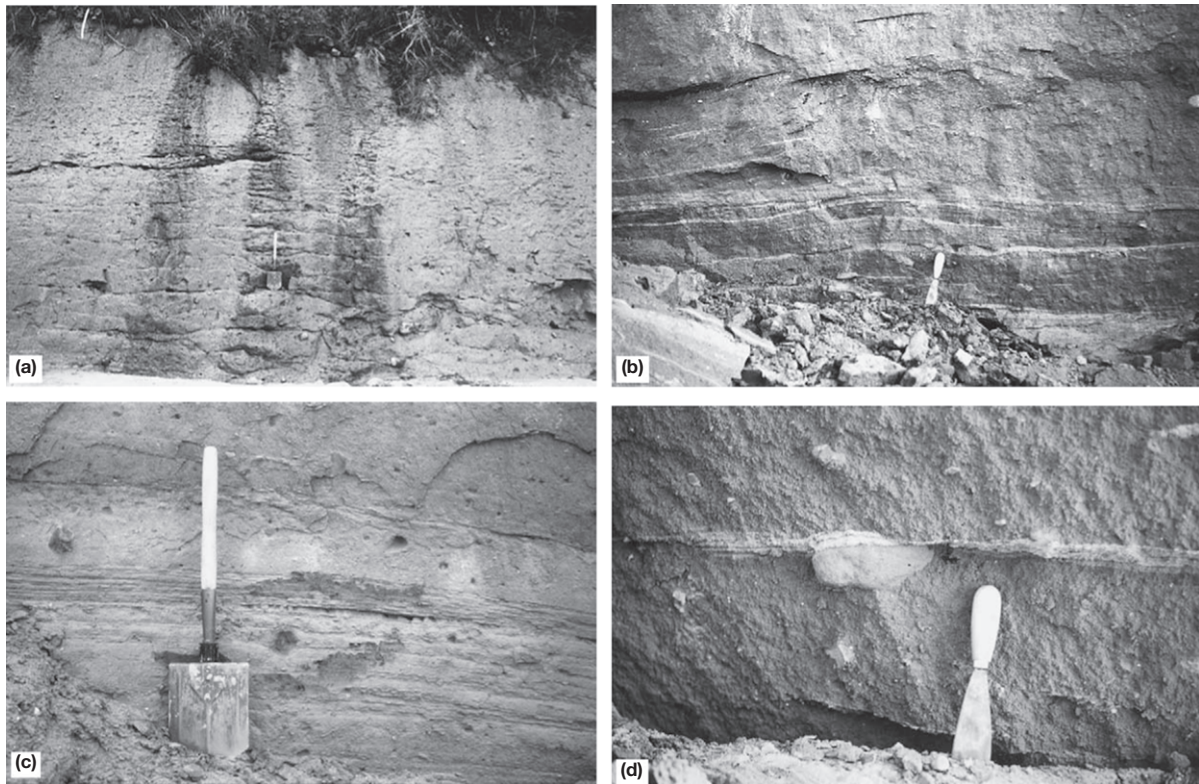


Figure 10 Examples of proposed ice–bed separation features. All features lack significant deformation and sand layers are interpreted as the product of thin water films at the ice–bed interface: (a) subhorizontal, often slickensided fissures, in places filled with syndepositional sorted sediments; (b) subhorizontal fissures (upper part) and thin stringers of sorted sediments interbedded with till matrix (lower part); (c) horizontally stratified till consisting of millimeter-thick sand layers intercalated with till matrix; (d) single horizontal stringer of stratified sand in till matrix. Reproduced from Piotrowski JA and Tulaczyk S (1999) Subglacial conditions under the last ice sheet in northwest Germany: Ice-bed separation and enhanced basal sliding? *Quaternary Science Reviews* 18: 737–751, with kind permission from J. Piotrowski.

sediments may also be interbedded with tills squirted into channels from nearby deforming layers or may record periods of subglacial meltwater sheetflow between phases of deformation till accretion. Further complications may then arise when such interbedded sequences are glacioteconized during later periods of ice–bed coupling (e.g., Boyce and Eyles, 2000; Evans et al., 1995; Eyles et al., 1982; Figure 11).

Glacioteconite

Glacioteconite (Banham, 1977; Pedersen, 1988) is defined as “rock or sediment that has been deformed by subglacial shearing (deformation) but retains some of the structural characteristics of the parent material” (Benn and Evans, 1998). The appearance of a glacioteconite reflects the nature of the parent material and its strain history, displaying evidence of either brittle or ductile deformation or a combination of the two processes (Figure 12). Inhomogeneous materials can display extremely complex deformation structures due to the variable response of individual sedimentary beds of differing rheology (e.g., Benn and Evans, 1996; Evans, 2000b; Evans et al., 1998; Hicock, 1992; Hicock and Fuller, 1995; Ó Cofaigh and Evans, 2001; Phillips et al., 2002). Benn and Evans (1996) identify Type A glacioteconites, which show evidence of penetrative deformation, and Type B glacioteconites, which have

undergone nonpenetrative deformation. The pervasive deformation of Type A glacioteconites to high cumulative strains can produce a tectonic foliation where primary structures are attenuated and rotated toward parallelism with the direction of shear (e.g., Hart, 1995; Hart and Boulton, 1991; Hart and Roberts, 1994) or transposed due to the imposition of a new foliation. Type B glacioteconites are produced by pervasive ductile deformation to small total cumulative strains and consequently, the predeformational sedimentary structures are still recognizable.

Comminution Till

Comminution till is defined as a “very dense till that appears to have been formed by abrasion of bedrock and the crushing of detritus dragged along underneath the ice, accompanied by a mixing process that results in the incorporation of rock powder produced by abrasion at the till–rock interface into the overlying glacial load” (Elson, 1988). Bedrock crushing and pulverizing reduces pore spaces and increases interparticle contacts, resulting in a grain size similar to that produced by industrial crushing and grinding (Benn and Evans, 1998; Elson, 1988). Grain crushing, fracturing, and attrition in the subglacial environment has been widely recognized (e.g., Boulton et al., 1974; Haldorsen, 1981; Hiemstra and van der Meer, 1997; Hooke

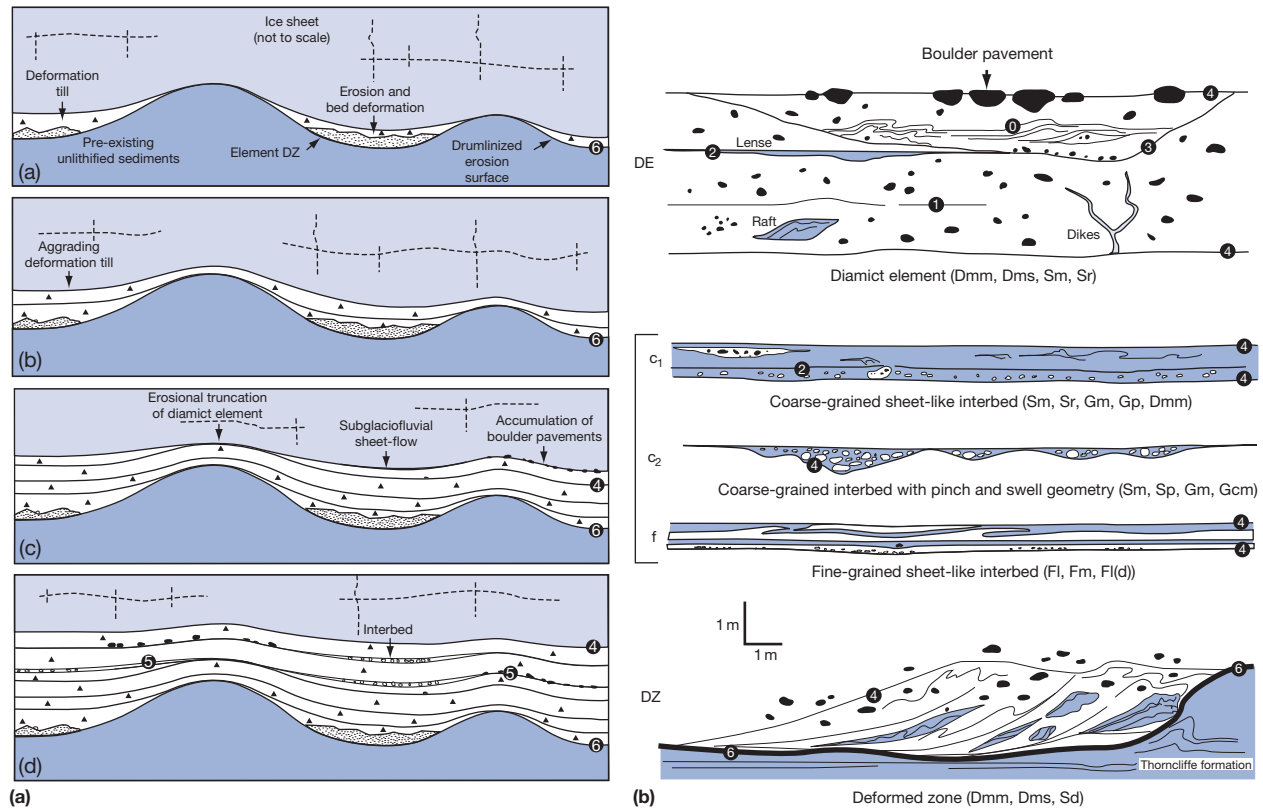


Figure 11 A model of till and interbed deposition (reproduced from Boyce *JI* and Eyles *N* (2000) Architectural element analysis applied to glacial deposits: Internal geometry of a late Pleistocene till sheet, Ontario, Canada. *Bulletin of the Geological Society of America* 112: 98–118): (a) the conceptual model involving: ‘a’ erosion and deformation of preexisting unlithified sediments to form drumlinized surface; ‘b’ conformable aggradation of deformation tills on drumlinized surface; ‘c’ subglacial fluvial reworking of diamictic till to form sheet-like interbeds and boulder pavements; ‘d’ continued aggradation of deformation tills and stratified interbeds; (b) schematic illustration of architectural element types with typical bounding surfaces as represented by interbeds. DE, diamict element, where tabular beds separated by planar to gently undulating bounding surfaces with occasional clast pavements are deposited by aggradation of deformation till; I = interbeds comprising coarse (I-c) and fine (I-f) ice-bed separation deposits; DZ, deformed zone comprising glacially deformed preexisting strata. Numbers in black dots refer to bounding surfaces whereby sixth-order surface, mappable stratigraphic unit; fifth-order surface, laterally continuous erosion surface; fourth-order surface, boundary between architectural elements; third-order surface, laterally discontinuous minor erosion surface; second-order surface, boundary defining dissimilar lithofacies; first-order surface, boundary separating similar lithofacies; and zero-order surface, laminae (see Boyce and Eyles, 2000 for details).

and Iverson, 1995; Yi Chaolu, 1997) but the production of thick comminution tills is most likely assisted by the displacement of relatively soft and/or heavily fractured bedrock slabs or thrust slices prior to crushing and disaggregation (Hiemstra et al., 2006, personal communication). Soft bedrock will be susceptible to *in situ* brecciation, thrusting, shearing, and plucking, and the formation of attenuated or boudinaged till intraclasts, thereby contributing to subglacial till matrixes. Evidence of the liberation of bedrock fragments from the substrate comes from descriptions of bedrock rafts in tills (e.g., Ruszczynska-Szenajch, 1987; Sauer, 1978; Stalker and Mac, 1973, 1976). Few examples of comminution till exist (e.g., Croot and Sims, 1996), possibly as a direct consequence of either their rarity and/or the overlap in the definitions of comminution till and glaciotectonite.

Deformation Till

First introduced by Elson (1961), deformation till is defined as “a rock or sediment that has been disaggregated and

completely or largely homogenized by shearing in a subglacial deforming layer” (Benn and Evans, 1998). Deformation is a permanent change in shape and orientation of a material from an initial to a final state (Passchier and Trouw, 1996). Such changes are visible in a material as structures like folds, faults, and/or folia (Figure 13). In the absence of such structures a till cannot unequivocally be interpreted as deformed. Subglacial tills are commonly massive and lack any obvious macroscopic and/or microscopic evidence of deformation but many have still been interpreted as deformation tills, the massive appearance being regarded as evidence for high cumulative strains and total homogenization of source materials (e.g., Benn and Evans, 1996; Boulton, 1987; Boulton et al., 2001; Evans, 2000a; Hart et al., 1990). This has led to controversy in till recognition because massive melt-out tills may be impossible to differentiate from massive deformation till (cf. Alley, 1991; Boulton and Jones, 1979; Clayton et al., 1989; Ronnert and Mickelson, 1992). However, the term deformation till is more confidently applied to stratigraphic successions in which layered, heterogeneous sediment grades into a deformed homogeneous till (Benn

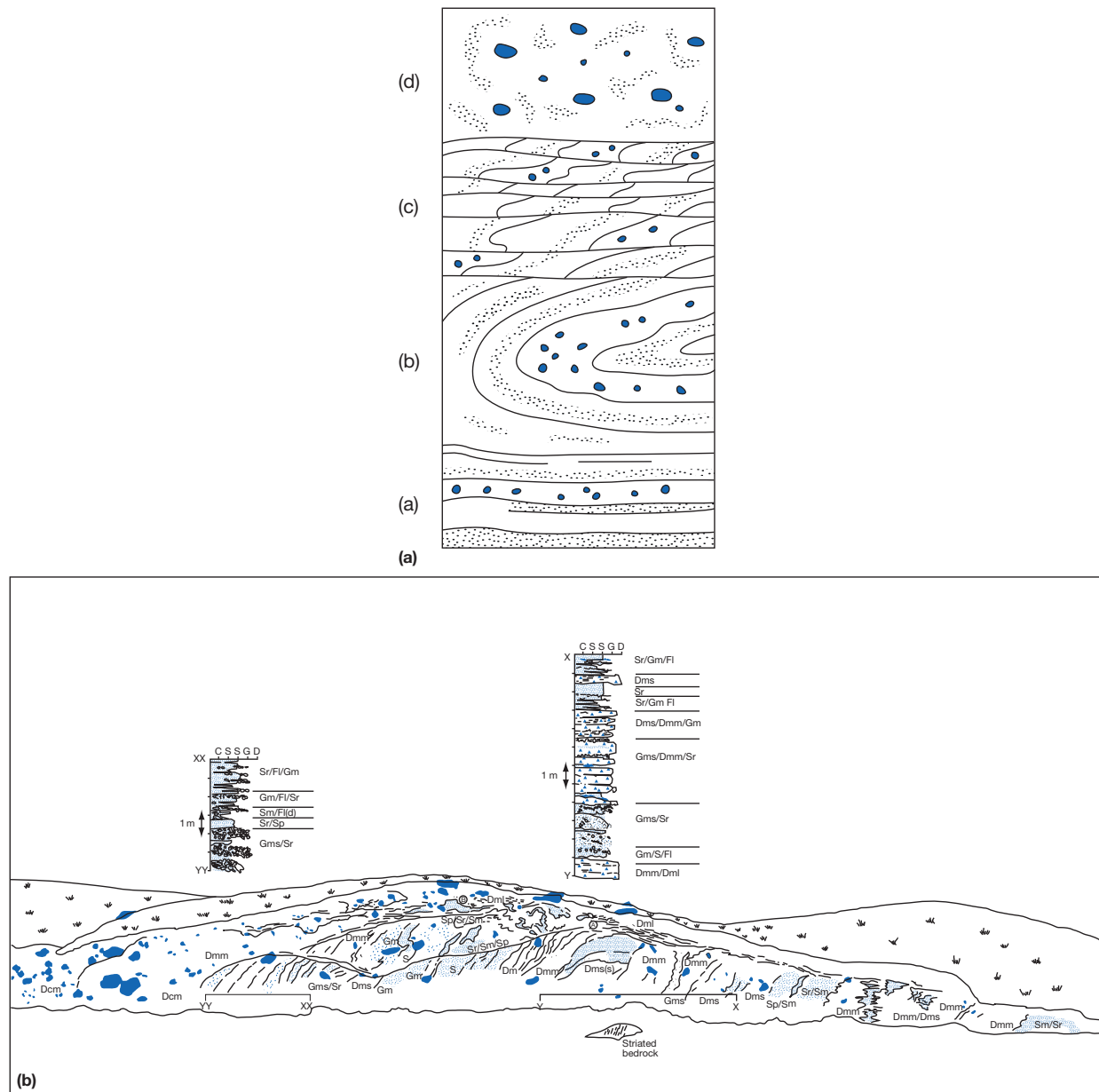


Figure 12 Glaciotectonite terminology and examples: (a) idealized vertical sequence/continuum of glacially deformed material (from Benn and Evans, 1996, after Banham, 1977). a – undeformed parent material (e.g., stratified sands and gravels); b – Type b glaciotectonite with nonpenetrative deformation (e.g., folded sands and gravels); c – Type a glaciotectonite with penetrative deformation (e.g., complexly folded, faulted, sheared, attenuated, and boudinaged sands and gravels); d – laminated or macroscopically massive till with sheared inclusions of sand and gravel (interpreted as a mixture of local glaciotectonite and far-traveled glacially transported sediment); (b) example of vertical continuum of glaciotectonites and subglacial till, Loch Quoich, Scotland (reproduced from Evans DJA, Rea BR, and Benn DI (1998) Subglacial deformation and bedrock plucking in areas of hard bedrock. *Glacial Geology and Geomorphology*. <http://ggg.qub.ac.uk/ggg/papers/full/1998/rp041998/rp04.html>). Folded glaciolacustrine sediments (Type b glaciotectonite) are exposed as vertically inclined beds at the base of the exposure (logs record the prefolded vertical sequence). Type a glaciotectonite at the center of the exposure lies in fault contact with the underlying Type b glaciotectonite and is capped by massive to laminated till.

and Evans, 1996). Therefore, unless evidence can be found in support of a till having been produced by homogenization during deformation, the term deformation till is best avoided.

Differentiation of melt-out and deformation tills is controversial even where sedimentary and structural features are present. The varying states of preservation of primary sedimentary structures may constitute evidence for depositional/

deformational continuums. The spatial and temporal mosaic of subglacial processes may also give rise to melt-out/deformation continuums where subglacial sediment released from the ice base by melting is then deformed in its uppermost layers, thereby preserving soft sediment intraclasts and structures to varying degrees in the resulting till sequence (Evans, 1994; Larsen et al., 2004).

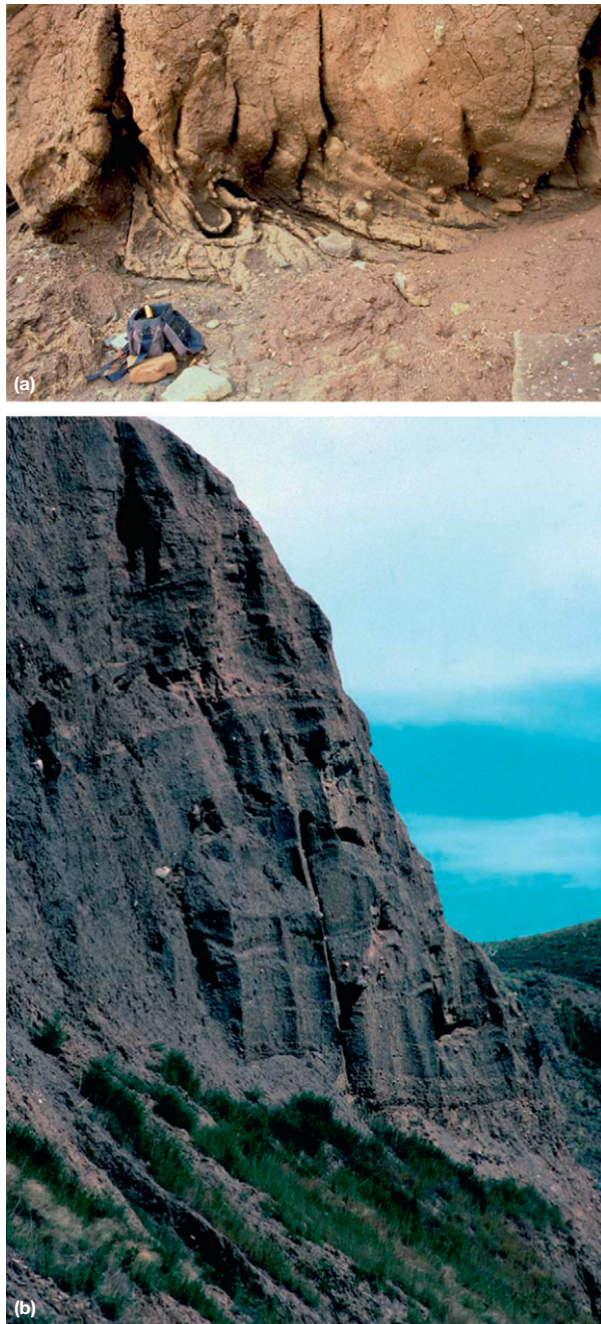


Figure 13 Examples of diamictic sediments interpreted as deformation till: (a) Skipsea Till at Filey (UK), clearly showing fold structures due to preferential weathering of sand-rich intratill wisps. In the absence of such weathering the till appears macroscopically massive; (b) intensely folded multiple tills at the Cameron Ranch section in southern Alberta (USA). Each till unit displays textural banding, thereby allowing the identification of strain signature.

The basal contacts of deformation tills with underlying sediments can vary due to rheological contrasts, porewater pressure differences, and the generation of perturbations such as folds. A sharp, erosional boundary, often marked by slickensided surfaces, suggests a shear plane (Boulton et al., 2001; Ehlers and Stephan, 1979). A shear zone may also develop,

where the two materials become mixed (Hooyer and Iverson, 2000). Other structures associated with the base of deforming-layer tills include incorporated or plucked intraclasts/smudges and highly attenuated, folded and boudinaged laminae of underlying sediment (e.g., Boulton, 1987; Evans, 2000b; Hicock and Dreimanis, 1989, 1992a,b). The laminae may be vertically stacked in the till and thereby represent tectonic slices produced during sequential phases of deformation till accumulation (Boulton et al., 2001) or they may be the product of large numbers of folding and attenuation events which have stretched them out to produce 'glaciotectonic laminae' (Figure 14). This stretching may continue until the highly strained and homogenized till appears massive until viewed under the microscope (Benn and Evans, 1996; Boulton, 1987; Hiemstra and Rijdsdijk, 2003; Menzies, 2000; van der Meer, 1993).

Lee-Side Cavity-Fill Deposits

On a rigid glacier bed covered with roches moutonnées, lee-side cavities will often open up and become the depository for sediments with a wide range of characteristics related to the processes of mass flowage, fall, melt-out, and fluvial reworking (Figure 6). The resulting down-ice thinning, wedge-shaped masses of massive-to-crudely stratified diamictons with steeply dipping lenses of water-sorted sediments are classified as lee-side cavity fills ('lee-side tills' of Hillefors, 1973) and constitute a landform termed a 'crag-and-tail.' Clasts are striated and possess a strong down-valley dipping fabric (Levson and Rutter, 1986, 1989), although other characteristics of ploughing and lodgement are absent. Once a cavity closes the sediment fill is subject to reworking, including overprinting and/or truncation by the subglacial processes of ploughing, lodgement, glaciotectonic disruption, and deformation (Levson and Rutter, 1986, 1989). This may explain the apparent stratification in some glaciotectonites and deformation tills.

Subglacial Till Nomenclature

Because the processes of deformation, melt-out, flowage, sliding, lodgement, meltwater reworking, and ploughing coexist subglacially, mobilizing and transporting sediment and depositing it as various end members that range from glaciotectonically folded and faulted stratified material to texturally homogeneous diamicton (e.g., Hicock, 1990), it is important to recognize the inherent problems in genetic labeling of Quaternary tills. Moreover, the dominance of a subglacial process varies both spatially and temporally, giving rise to complex till sequences that contain a superimposed signature of former transportation/deposition at the ice-bed interface. Consequently, Evans et al. (2006) have recommended that genetic labeling of tills should be less process specific. They propose a threefold classification scheme that incorporates the process and experimentally based observations on till production and the unequivocal process-form relationships identifiable in the geological record (Figure 15). The classifications are glaciotectonite, subglacial traction till, and melt-out till.

Glaciotectonite, as defined above, can be composed of an infinite variety of parent materials, from bedrock to previously deposited Quaternary sediments. Comminution till is included

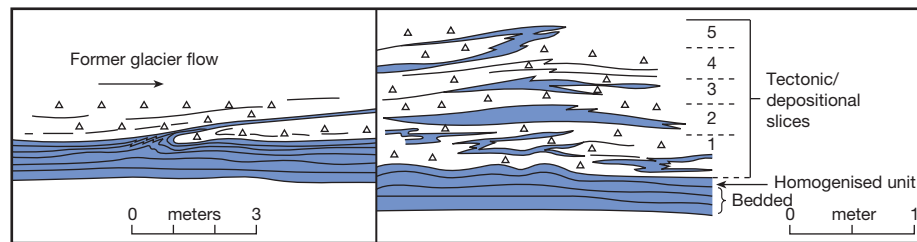


Figure 14 Examples of sedimentary structures produced by cumulative shear and the inclusion of substrate material in till due to subglacial deformation: Left: inclusion of sand lamina through the attenuation of a drag fold at the till–substrate interface. Adjacent sand wisps represent other laminae produced by earlier folding events; Right: multiple fold structures evident as attenuated, folded, and boudinaged sand laminae. Each structure is interpreted as a separate phase of deformation till accumulation, termed a tectonic/depositional slice. Reproduced from Boulton GS, Dobbie KE, and Zatzepin S (2001) Sediment deformation beneath glaciers and its coupling to the subglacial hydraulic system. *Quaternary International* 86: 3–28.

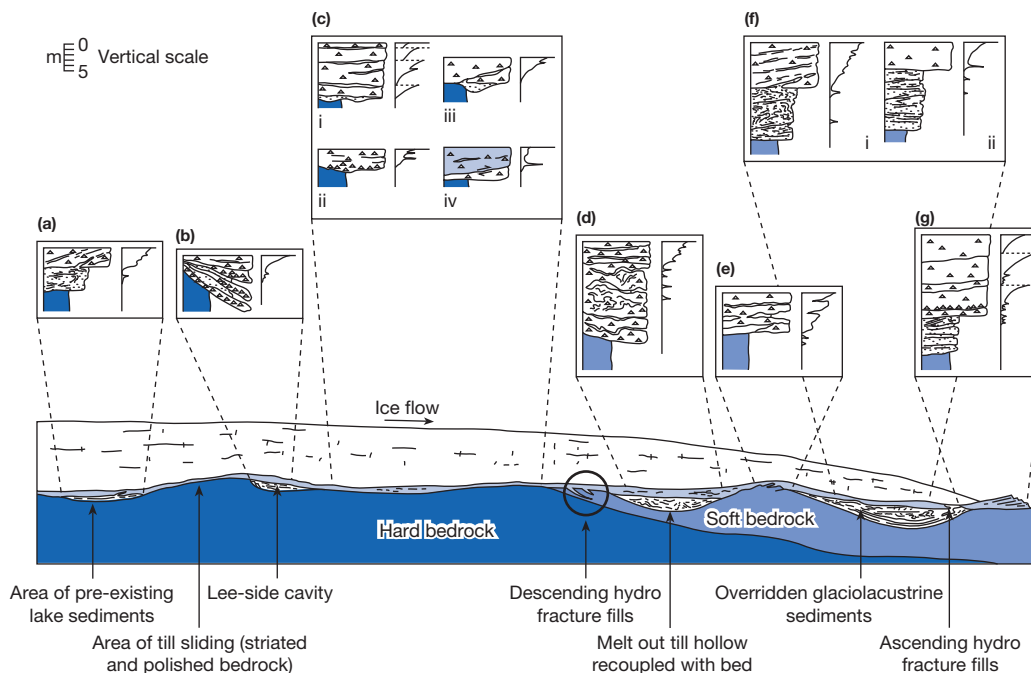


Figure 15 Conceptual model of the full spectrum of till production and depositional mechanisms resulting in a continuum of till formation. The processes depicted are those that would be expected beneath a large, predominantly temperate outlet glacier with a seasonally frozen snout. The shaded area of the glacier bed represents the subglacial traction till (STT). The final thickness of a till deposit does not reflect the thickness of the subglacial traction/deforming layer at any one point in time. Not all the process-form products would necessarily be coeval and signatures may be overprinted due to glacier advance and retreat cycles. The graphs to the right of each vertical profile log indicate the likely relative pattern of cumulative strain. The process-form products are: (a) preadvance lake sediments capped by STT; (b) lee-side cavity fill diamicts and stratified sediments capped by STT where ice has recoupled with the bed due to cavity closure. In (a) and (b), the tops of the lake sediment and cavity fill sequence are classified as glacioteconite, which is locally cannibalized and dragged into the base of the deforming till layer after folding/thrusting is induced by ice keel ploughing, clast ploughing, and/or migrations in the locus of failure planes in subglacial sediment aided by cyclical changes in porewater pressure. This produces a crudely laminated diamict; (c) sheared tills with partially to intensively glacioteconitized stratified intra- and interbeds produced over hard beds where, (i) STT and stratified interbeds (canal fills) are continually aggraded due to continuous replenishment of deformable sediment from up ice, (ii) large clasts are lodged to produce a clast pavement at the base of a thin STT in which finer grained material is continuously advected down ice, (iii) a single STT with associated glacioteconitized canal fill at a site where aggradation is restricted by poor till continuity, (iv) regelation of STT (dark gray shade) leads to the stacking of tills which later melt-out (note the similarity with sequences produced by marginal incremental thickening in (g)); (d) preadvance or remobilized melt-out till overlain by STT. The top of the melt-out till is classified as glacioteconite and is locally cannibalized and dragged into the base of the deforming till layer after folding/thrusting is induced at the STT–melt-out till interface and/or the loci of failure planes migrate through the subglacial sediment pile to produce a crudely laminated diamict; (e) glacioteconitized soft bedrock grades upwards into STT with bedrock rafts ('comminution till'); (f) glacioteconite (glacially overridden lake sediments) overlain by deforming STT where, (i) the glacioteconite has been locally cannibalized and dragged into the base of the deforming layer after folding/thrusting is induced at the STT–stratified sediment interface and/or the loci of failure planes migrate through the subglacial sediment pile to produce a crudely laminated diamict. This may be assisted by seasonal migration of the freezing front or cyclical changes in pore water pressures, (ii) the top of the glacioteconite is eroded by the base of the STT. The contact between the two sediment bodies is essentially a fault gouge; (g) glaciofluvial outwash or lake sediment, glacioteconitized in its upper layers and overlain by a stacked sequence of macroscopically massive STTs. Incremental stacking occurs where the glacier margin is stationary and slabs of STT freeze on and melt-out every year. Reproduced from Evans DJA, Rea BR, Hiemstra JF, and O Cofaigh C (2006) Genesis of glacial sediments and landforms in south-central Alberta, Canada: A critical assessment of mega-flood explanations. *Quaternary Science Reviews* 25: 1638–1667.

under this term because it is derived from the subglacial reworking of bedrock. The deformation structures displayed within glacioteconites reflect their strain history and the rheology of individual sedimentary beds. Glacioteconites can be subdivided into Type A, which display evidence of penetrative deformation, and Type B, which have undergone nonpenetrative deformation and so predeformational sedimentary structures are merely folded and faulted.

Subglacial traction till, because it is the product of a range of interrelated processes occurring in the traction zone at a glacier bed, is defined as “sediment deposited by a glacier sole either sliding over and/or deforming its bed, the sediment having been released directly from the ice by pressure melting and/or liberated from the substrate and then disaggregated and completely or largely homogenized by shearing. It includes localized lee-side cavity fills on hard beds and stratified sediments deposited by water films and channels at the ice–bed interface and within deforming subglacial sediment by Darcian and pipe flow” (Evans et al., 2006). This definition combines the inextricable subglacial process–form relationships associated with lodgement and deformation, includes the effects of substrate ploughing, and acknowledges that glacier beds are spatially and temporally changing mosaics of deformation and sliding and warm-based and cold-based conditions, implying that subglacial till end members are hybrids of the subglacial traction zone.

Melt-out till, as defined above, remains a theoretical construct and may rarely inherit and replicate the foliation and structure of its parent ice due to the severe limitations preservation potential. Melt-out tills probably occur in the geological record as both crudely stratified and macroscopically massive diamictites displaying a wide range of gravity induced disturbance structures. In situations where melt-out deposits have been overrun by glacier ice or recoupled with the glacier bed due to reactivation of ice flow, they will contain subglacial traction features and should therefore be classified as a glacioteconite.

See also: [Glacial Landforms, Ice Sheets: Evidence of Glacier and Ice Sheet Extent](#); [Glacial Landforms, Sediments: Clast Form Analysis](#); [Glaciogenic Lithofacies](#); [Micromorphology of Glacial Sediments](#); [Till Fabric Analysis](#).

References

- Alley RB (1989a) Water pressure coupling of sliding and bed deformation: 1. Water system. *Journal of Glaciology* 35: 108–118.
- Alley RB (1989b) Water pressure coupling of sliding and bed deformation: II. Velocity–depth profiles. *Journal of Glaciology* 35: 119–129.
- Alley RB (1991) Deforming-bed origin for southern Laurentide till sheets? *Journal of Glaciology* 37: 67–76.
- Alley RB (1992) How can low pressure channels and deforming tills coexist subglacially? *Journal of Glaciology* 38: 200–207.
- Alley RB, Blankenship DD, Bentley CR, and Rooney ST (1986) Deformation of till beneath ice stream B, West Antarctica. *Nature* 322: 57–59.
- Alley RB, Blankenship DD, Bentley CR, and Rooney ST (1987) Till beneath ice stream B. 3. Till deformation: Evidence and implications. *Journal of Geophysical Research* 92: 8921–8929.
- Alley RB, Blankenship DD, Rooney ST, and Bentley CR (1987) Till beneath ice stream B. 4. A coupled ice–till flow model. *Journal of Geophysical Research* 92: 8931–8940.
- Banham PH (1977) Glacioteconites in till stratigraphy. *Boreas* 6: 101–105.
- Benn DI (1994) Fluted moraine formation and till genesis below a temperate glacier: Slettmarkbreen, Jotunheimen, Norway. *Sedimentology* 41: 279–292.
- Benn DI (1995) Fabric signature of till deformation, Breidamerkurjökull, Iceland. *Sedimentology* 42: 735–747.
- Benn DI and Evans DJA (1996) The interpretation and classification of subglacially-deformed materials. *Quaternary Science Reviews* 15: 23–52.
- Benn DI and Evans DJA (1998) *Glaciers and Glaciation*. London: Arnold.
- Blake EW, Clarke GKC, and Gerin MC (1992) Tools for examining subglacial bed deformation. *Journal of Glaciology* 38: 388–396.
- Blankenship DD, Bentley CR, Rooney ST, and Alley RB (1986) Seismic measurements reveal a saturated porous layer beneath an active Antarctic ice stream. *Nature* 322: 54–57.
- Blankenship DD, Bentley CR, Rooney ST, and Alley RB (1987) Till beneath ice stream B. 1. Properties derived from seismic travel times. *Journal of Geophysical Research* 92: 8903–8911.
- Boulton GS (1970) On the deposition of subglacial and melt-out tills at the margins of certain Svalbard glaciers. *Journal of Glaciology* 9: 231–245.
- Boulton GS (1971) Till genesis and fabric in Svalbard, Spitsbergen. In: Goldthwait RP (ed.) *Till: A Symposium*, pp. 41–72. Columbus: Ohio State University Press.
- Boulton GS (1975) Processes and patterns of subglacial sedimentation: A theoretical approach. In: Wright AE and Moseley F (eds.) *Ice Ages: Ancient and Modern*, pp. 7–42. Liverpool: Seel House Press.
- Boulton GS (1978) Boulder shapes and grain size distributions of debris as indicators of transport paths through a glacier and till genesis. *Sedimentology* 25: 773–799.
- Boulton GS (1979) Processes of glacier erosion on different substrata. *Journal of Glaciology* 23: 15–38.
- Boulton GS (1982) Subglacial processes and the development of glacial bedforms. In: Davidson-Arnott R, Nickling W, and Fahey BD (eds.) *Research in Glacial, Glacio-fluvial and Glacio-lacustrine Systems*, pp. 1–31. Norwich: Geo Books.
- Boulton GS (1987) A theory of drumlin formation by subglacial deformation. In: Rose J and Menzies J (eds.) *Drumlin Symposium*, pp. 25–80. Rotterdam: Balkema.
- Boulton GS (1996) The origin of till sequences by subglacial sediment deformation beneath mid-latitude ice sheets. *Annals of Glaciology* 22: 75–84.
- Boulton GS and Caban PE (1995) Groundwater flow beneath ice sheets: Part II. Its impact on glacier tectonic structures and moraine formation. *Quaternary Science Reviews* 14: 563–587.
- Boulton GS, Dent DL, and Morris EM (1974) Subglacial shearing and crushing, and the role of water pressures in tills from south-east Iceland. *Geografiska Annaler* 56A: 135–145.
- Boulton GS and Dobbie KE (1998) Slow flow of granular aggregates: The deformation of sediments beneath glaciers. *Philosophical Transactions of the Royal Society of London A* 356: 2713–2745.
- Boulton GS, Dobbie KE, and Zatzepin S (2001) Sediment deformation beneath glaciers and its coupling to the subglacial hydraulic system. *Quaternary International* 86: 3–28.
- Boulton GS and Hindmarsh RCA (1987) Sediment deformation beneath glaciers: Rheology and geological consequences. *Journal of Geophysical Research* 92: 9059–9082.
- Boulton GS and Jones AS (1979) Stability of temperate ice sheets resting on beds of deformable sediment. *Journal of Glaciology* 24: 29–43.
- Boulton GS and Paul MA (1976) The influence of genetic processes on some geotechnical properties of tills. *Journal of Engineering Geology* 9: 159–194.
- Boyce JI and Eyles N (2000) Architectural element analysis applied to glacial deposits: Internal geometry of a Late Pleistocene till sheet, Ontario, Canada. *Bulletin of the Geological Society of America* 112: 98–118.
- Brown NE, Hallet B, and Booth DB (1987) Rapid soft bed sliding of the Puget glacial lobe. *Journal of Geophysical Research* 92: 8985–8997.
- Carlson AE (2004) Genesis of dewatering structures and its implications for melt-out till identification. *Journal of Glaciology* 50: 17–24.
- Chamberlin TC (1895) Recent glacial studies in Greenland. *Bulletin of the Geological Society of America* 6: 199–220.
- Chaolu Yi (1997) Subglacial comminution in till – Evidence from microfabric studies and grain-size distributions. *Journal of Glaciology* 43: 473–479.
- Clark PU and Hansel AK (1989) Clast ploughing, lodgement and glacier sliding over a soft glacier bed. *Boreas* 18: 201–207.
- Clark CD, Tulaczyk SM, Stokes CR, and Canals M (2003) A groove-ploughing theory for the production of mega-scale glacial lineations, and implications for ice-stream mechanics. *Journal of Glaciology* 49: 240–256.

- Clarke GKC (1987) Subglacial till: A physical framework for its properties and processes. *Journal of Geophysical Research* 92: 9023–9036.
- Clayton L, Mickelson DM, and Attig JW (1989) Evidence against pervasively deformed bed material beneath rapidly moving lobes of the southern Laurentide ice sheet. *Sedimentary Geology* 62: 203–208.
- Croot DG and Sims PC (1996) Early stages of till genesis: An example from Fanore, County Clare, Ireland. *Boreas* 25: 37–46.
- Dowdeswell JA and Sharp M (1986) Characterization of pebble fabrics in modern terrestrial glacial sediments. *Sedimentology* 33: 699–710.
- Dreimanis A (1989) Tills, their genetic terminology and classification. In: Goldthwait RP and Mutsch CL (eds.) *Genetic Classification of Glacigenic Deposits*, pp. 17–84. Rotterdam: Balkema.
- Ehlers J and Stephan HJ (1979) Forms at the base of till strata as indicators of ice movement. *Journal of Glaciology* 22: 345–355.
- Elson JA (1961) The geology of tills. In: Penner E and Butler J (eds.) *Proceedings of the 14th Canadian Soil Mechanics Conference*, pp. 5–36. Commission for Soil and Snow Mechanics. Technical Memoir 69.
- Elson JA (1988) Comment on glaciectonite, deformation till and comminution till. In: Goldthwait RP and Mutsch CL (eds.) *Genetic Classification of Glacigenic Deposits*, pp. 85–88. Rotterdam: Balkema.
- Evans DJA (1994) The stratigraphy and sedimentary structures associated with complex subglacial thermal regimes at the southwestern margin of the Laurentide Ice Sheet, southern Alberta, Canada. In: Warren WP and Croot DG (eds.) *Formation and Deformation of Glacial Deposits*, pp. 203–220. Rotterdam: Balkema.
- Evans DJA (2000a) Quaternary geology and geomorphology of the Dinosaur Provincial Park area and surrounding plains, Alberta, Canada: The identification of former glacial lobes, drainage diversions and meltwater flood tracks. *Quaternary Science Reviews* 19: 931–958.
- Evans DJA (2000b) A gravel outwash/deformation till continuum, Skálafellsjökull, Iceland. *Geografiska Annaler* 82A: 499–512.
- Evans DJA, Owen LA, and Roberts D (1995) Stratigraphy and sedimentology of Devensian (Dimlington Stadial) glacial deposits, east Yorkshire, England. *Journal of Quaternary Science* 10: 241–265.
- Evans DJA, Rea BR, and Benn DI (1998) Subglacial deformation and bedrock plucking in areas of hard bedrock. *Glacial Geology and Geomorphology*. <http://ggg.qub.ac.uk/ggg/papers/full/1998/rp041998/rp04.html>.
- Evans DJA, Rea BR, Hiemstra JF, and Ó Cofaigh C (2006) Genesis of glacial sediments and landforms in south-central Alberta, Canada: A critical assessment of mega-flood explanations. *Quaternary Science Reviews* 25: 1638–1667.
- Evans DJA and Twigg DR (2002) The active temperate glacial landsystem: A model based on Breidamerkjökull and Fjallsjökull, Iceland. *Quaternary Science Reviews* 21: 2143–2177.
- Eyles N and Boyce JI (1998) Kinematics indicators in fault gouge: Tectonic analog for soft-bedded ice sheets. *Sedimentary Geology* 116: 1–12.
- Eyles N, Sladen JA, and Gilroy S (1982) A depositional model for stratigraphic complexes and facies superimposition in lodgement tills. *Boreas* 11: 317–333.
- Fischer U and Clarke GKC (1997) Stick-slip sliding behaviour at the base of a glacier. *Annals of Glaciology* 24: 390–396.
- Fischer U and Clarke GKC (2001) Review of subglacial hydro-mechanical coupling: Trapridge Glacier, Yukon Territory, Canada. *Quaternary International* 86: 29–43.
- Fischer U, Clarke GKC, and Blatter H (1999) Evidence for temporally varying 'sticky spots' at the base of Trapridge Glacier, Yukon Territory, Canada. *Journal of Glaciology* 45: 352–360.
- Fischer U, Porter PR, Schuler T, Evans AJ, and Gudmundsson GH (2001) Hydraulic and mechanical properties of glacial sediments beneath Unteraargletscher, Switzerland: Implications for glacier basal motion. *Hydrological Processes* 15: 3525–3540.
- Fowler AC (2002) Rheology of subglacial till. *Journal of Glaciology* 48: 631–632.
- Fowler AC (2003) On the rheology of till. *Annals of Glaciology* 37: 55–59.
- Fuller S and Murray T (2000) Evidence against pervasive bed deformation during the surge of an Icelandic glacier. In: Maltman AJ, Hubbard B, and Hambrey MJ (eds.) *Deformation of Glacial Materials*, pp. 203–216. London: Geological Society. Special Publications 176.
- Haldorsen S (1981) Grain-size distribution of subglacial till and its relation to glacial crushing and abrasion. *Boreas* 10: 91–105.
- Haldorsen S and Shaw J (1982) The problem of recognizing melt out till. *Boreas* 11: 261–277.
- Ham NR and Mickelson DM (1994) Basal till fabric and deposition at Burroughs Glacier, Glacier Bay, Alaska. *Bulletin of the Geological Society of America* 106: 1552–1559.
- Hart JK (1995) Subglacial erosion, deposition and deformation associated with deformable beds. *Progress in Physical Geography* 19: 173–191.
- Hart JK and Boulton GS (1991) The inter-relation of glaciectonic and glaciodepositional processes within the glacial environment. *Quaternary Science Reviews* 10: 335–350.
- Hart JK, Hindmarsh RCA, and Boulton GS (1990) Styles of subglacial glaciectonic deformation within the context of the Anglian ice sheet. *Earth Surface Processes and Landforms* 15: 227–241.
- Hart JK and Roberts DH (1994) Criteria to distinguish between subglacial glaciectonic and glaciomarine sedimentation, I. Deformation styles and sedimentology. *Sedimentary Geology* 91: 191–213.
- Hicock SR (1990) Genetic till prism. *Geology* 18: 517–519.
- Hicock SR (1992) Lobal interactions and rheologic superposition in subglacial till near Bradville, Ontario, Canada. *Boreas* 21: 73–88.
- Hicock SR and Dreimanis A (1989) Sunnybrook drift indicates a grounded early Wisconsin glacier in the Lake Ontario basin. *Geology* 17: 169–172.
- Hicock SR and Dreimanis A (1992a) Sunnybrook drift in the Toronto area, Canada: Reinvestigation and reinterpretation. In: Clark PU and Lea PD (eds.) *The Last Interglacial–Glacial Transition in North America*, pp. 139–161. Boulder, CO: Geological Society of America. Special Paper 270.
- Hicock SR and Dreimanis A (1992b) Deformation till in the Great Lakes region: Implications for rapid flow along the south-central margin of the Laurentide Ice Sheet. *Canadian Journal of Earth Sciences* 29: 1565–1579.
- Hicock SR and Fuller EA (1995) Lobal interactions, rheologic superposition, and implications for a Pleistocene ice stream on the continental shelf of British Columbia. *Geomorphology* 14: 167–184.
- Hiemstra JF and Rijdsdijk KF (2003) Observing artificially induced strain: Implications for subglacial deformation. *Journal of Quaternary Science* 18: 373–383.
- Hiemstra JF and van der Meer JMM (1997) Porewater controlled grain fracturing as an indicator for subglacial shearing in tills. *Journal of Glaciology* 43: 446–454.
- Hillefors A (1973) The stratigraphy and genesis of stoss- and lee-side moraines. *Bulletin of the Geological Institute of the University of Uppsala* 5: 139–154.
- Hindmarsh RCA (1996) Sliding of till over bedrock: Scratching, polishing, comminution and kinematic wave theory. *Annals of Glaciology* 22: 41–48.
- Hooke RL and Iverson NR (1995) Grain-size distribution in deforming subglacial tills: Role of grain fracture. *Geology* 23: 57–60.
- Hooyer TS and Iverson NR (2000) Diffusive mixing between shearing granular layers: Constraints on bed deformation from till contacts. *Journal of Glaciology* 46: 641–651.
- Humphrey N, Kamb B, Fahnestock M, and Engelhardt H (1993) Characteristics of the bed of the lower Columbia Glacier, Alaska. *Journal of Geophysical Research* 98: 837–846.
- Iverson NR (1999) Coupling between a glacier and a soft bed: II. Model results. *Journal of Glaciology* 45: 41–53.
- Iverson NR, Baker RW, Hooke RL, Hanson B, and Jansson P (1999) Coupling between a glacier and a soft bed: I. A relation between effective pressure and local shear stress determined from till elasticity. *Journal of Glaciology* 45: 31–40.
- Iverson NR, Hanson B, Hooke RL, and Jansson P (1995) Flow mechanism of glaciers on soft beds. *Science* 267: 80–81.
- Iverson NR and Iverson RM (2001) Distributed shear of subglacial till due to Coulomb slip. *Journal of Glaciology* 47: 481–488.
- Iverson NR, Jansson P, and Hooke RL (1994) In situ measurements of the strength of deforming subglacial till. *Journal of Glaciology* 40: 497–503.
- Kruger J (1979) Structures and textures in till indicating subglacial deposition. *Boreas* 8: 323–340.
- Kruger J (1984) Clasts with stoss-lee form in lodgement tills: A discussion. *Journal of Glaciology* 30: 241–243.
- Kruger J (1994) Glacial processes, sediments, landforms and stratigraphy in the terminus region of Myrdalsjökull, Iceland. *Folia Geographica Danica* 21: 1–233.
- Larsen NK, Piotrowski JA, and Kronborg C (2004) A multiproxy study of a basal till: A time-transgressive accretion and deformation hypothesis. *Journal of Quaternary Science* 19: 9–21.
- Lawson DE (1979a) A comparison of the pebble orientations in ice and deposits of the Matanuska Glacier, Alaska. *Journal of Geology* 87: 629–645.
- Lawson DE (1979b) Sedimentological analysis of the western terminus region of the Matanuska Glacier, Alaska. *CRREL Report 79-9*, Hanover, NH.
- Lawson DE (1981) Distinguishing characteristics of diamictites at the margin of the Matanuska Glacier, Alaska. *Annals of Glaciology* 2: 78–84.
- Levson VM and Rutter NW (1986) A facies approach to the stratigraphic analysis of Late Wisconsinan sediments in the Portal Creek area, Jasper National Park, Alberta. *Geographie Physique et Quaternaire* 40: 129–144.
- Levson VM and Rutter NW (1989) A lithofacies analysis and interpretation of depositional environments of montane glacial diamictites, Jasper, Alberta, Canada. In: Goldthwait RP and Mutsch CL (eds.) *Genetic Classification of Glacigenic Deposits*, pp. 117–140. Rotterdam: Balkema.
- Lian OB and Hicock SR (2000) Thermal conditions beneath parts of the last Cordilleran Ice Sheet near its centre as inferred from subglacial till, associated sediments and bedrock. *Quaternary International* 68–71: 147–162.

- Mandl G and Harkness RM (1987) Hydrocarbon migration by hydraulic fracturing. In: Jones ME and Preston RMF (eds.) *Deformation of Sediments and Sedimentary Rocks*, pp. 39–53. London: Geological Society of London. Special Publication 29.
- Menzies J (2000) Micromorphological analyses of microfabrics and microstructures indicative of deformation processes in glacial sediments. In: Maltman AJ, Hubbard B, and Hambrey MJ (eds.) *Deformation of Glacial Materials*, pp. 245–257. London: Geological Society. Special Publication 176.
- Mickelson DM (1973) Nature and rate of basal till deposition in a stagnating ice mass, Burroughs Glacier, Alaska. *Arctic and Alpine Research* 5: 17–27.
- Munro-Stasiuk MJ (2000) Rhythmic till sedimentation: Evidence for repeated hydraulic lifting of a stagnant ice mass. *Journal of Sedimentary Research* 70: 94–106.
- Ng FSL (2000) Canals under sediment-based ice sheets. *Annals of Glaciology* 30: 146–152.
- Ó Cofaigh C and Evans DJA (2001) Sedimentary evidence for deforming bed conditions associated with a grounded Irish Sea glacier, southern Ireland. *Journal of Quaternary Science* 16: 435–454.
- Passchier CW and Trouw RAJ (1996) *Microtectonics*, p. 289. Springer: Berlin.
- Paul MA and Eyles N (1990) Constraints on the preservation of diamict facies (melt-out tills) at the margins of stagnant glaciers. *Quaternary Science Reviews* 9: 51–69.
- Pedersen SAS (1988) Glaciotectonite: Brecciated sediments and cataclastic sedimentary rocks formed subglacially. In: Goldthwait RP and Matsch CL (eds.) *Genetic Classification of Glacigenic Deposits*, pp. 89–91. Rotterdam: Balkema.
- Phillips ER, Evans DJA, and Auton CA (2002) Polyphase deformation at an oscillating ice margin following the Loch Lomond Readvance, central Scotland, UK. *Sedimentary Geology* 149: 157–182.
- Piotrowski JA, Doring U, Harder A, Qadir R, and Wenghofer S (1997) Deforming bed conditions on the Danischer World Peninsula, northern Germany: Comments. *Boreas* 26: 73–77.
- Piotrowski JA and Kraus AM (1997) Response of sediment to ice sheet loading in northwestern Germany: Effective stresses and glacier-bed stability. *Journal of Glaciology* 43: 495–502.
- Piotrowski JA, Mickelson DM, Tulaczyk S, Krzyszkowski D, and Junge FW (2001) Were deforming bed beneath past ice sheets really widespread? *Quaternary International* 86: 139–150.
- Piotrowski JA, Mickelson DM, Tulaczyk S, Krzyszkowski D, and Junge FW (2002) Reply to comments by G. S., Boulton, K. E., Dobbie, S., Zatsepin on: Deforming soft beds under ice sheets: How extensive were they? *Quaternary International* 97–98: 173–177.
- Piotrowski JA and Tulaczyk S (1999) Subglacial conditions under the last ice sheet in northwest Germany: Ice-bed separation and enhanced basal sliding? *Quaternary Science Reviews* 18: 737–751.
- Rijsdijk KF, Owen G, Warren WP, McCarroll D, and van der Meer JJM (1999) Clastic dykes in over-consolidated tills: Evidence for subglacial hydrofracturing at Killiney Bay, eastern Ireland. *Sedimentary Geology* 129: 111–126.
- Ronnert L and Mickelson DM (1992) High porosity of basal till at Burroughs Glacier, southeastern Alaska. *Geology* 20: 849–852.
- Ruszczynska-Szenajch H (1987) The origin of glacial rafts: Detachment, transport, deposition. *Boreas* 16: 101–112.
- Ruszczynska-Szenajch H (2001) 'Lodgement till' and 'deformation till'. *Quaternary Science Reviews* 20: 579–581.
- Sauer EK (1978) The engineering significance of glacier ice-thrusting. *Canadian Geotechnical Journal* 15: 457–472.
- Sharp MJ (1982) Modification of clasts in lodgement tills by glacial erosion. *Journal of Glaciology* 28: 475–481.
- Shaw J (1977) Tills deposited in arid polar environments. *Canadian Journal of Earth Sciences* 14: 1239–1245.
- Shaw J (1979) Genesis of the Sveg tills and Rogen moraines of central Sweden: A model of basal melt-out. *Boreas* 8: 409–426.
- Shaw J (1982) Melt out till in the Edmonton area, Alberta, Canada. *Canadian Journal of Earth Sciences* 19: 1548–1569.
- Shaw J (1983) Forms associated with boulders in melt-out till. In: Evenson EB, Schluchter C, and Rabassa J (eds.) *Tills and Related Deposits*, pp. 3–12. Rotterdam: Balkema.
- Shaw J (1988) Sublimation till. In: Goldthwait RP and Matsch CL (eds.) *Genetic Classification of Glacigenic Deposits*, pp. 141–142. Rotterdam: Balkema.
- Stalker A and Mac S (1973) The large interdrift bedrock blocks of the Canadian Prairies. *Geological Survey of Canada Paper* 75-1A: 421–422.
- Stalker A and Mac S (1976) Megablocks, or the enormous erratics of the Albertan Prairies. *Geological Survey of Canada Paper* 76-1C: 185–188.
- Thorsteinsson T and Raymond CF (2000) Sliding versus till deformation in the fast motion of an ice stream over a viscous till. *Journal of Glaciology* 46: 633–640.
- Tulaczyk S, Scherer RP, and Clark CD (2001) A ploughing model for the origin of weak tills beneath ice streams: A qualitative treatment. *Quaternary International* 86: 59–70.
- van der Meer JJM (1993) Microscopic evidence of subglacial deformation. *Quaternary Science Reviews* 12: 553–587.
- van der Meer JJM, Kjaer KH, and Kruger J (1999) Subglacial water escape structures and till structure, Slettjökull, Iceland. *Journal of Quaternary Science* 14: 191–205.
- Vivian R (1975) *Les Glaciers des Alpes Occidentales*. Grenoble, France: Allier.
- Walder JS and Fowler A (1994) Channelized subglacial drainage over a deformable bed. *Journal of Glaciology* 40: 3–15.
- Watts RJ and Carr SJ (2002) A laboratory simulation of the subglacial deforming bed (the deformation tank). *Quaternary Newsletter* 98: 1–9.
- Weertman J (1964) The theory of glacier sliding. *Journal of Glaciology* 5: 287–303.

Till Fabric Analysis

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Introduction

Till fabric analysis concerns the collection and processing of data on the orientation of particles in till deposits. The orientation of sand-size, or smaller, grains (matrix) is known as microfabric, and that of pebble-size or larger particles (clasts) is sometimes termed macrofabric by way of contrast. In general usage, however, 'fabric' is regarded as synonymous with macrofabric, and that sense is used here. There is a large literature on till fabric analysis, describing a wide range of analytical methods and results. Reviews of data collection techniques and analytical methods have been provided by Andrews (1971) and Benn (2004).

Data Collection

Unlithified Tills

Fabrics are usually determined in the field from a sample of clasts taken from a small region of a till deposit, using a compass-clinometer or similar instrument. The most commonly made measurements are the azimuth (compass bearing) and dip of particle long axes (L - or a -axes). However, this represents only part of the available information, and to uniquely define the orientation of any particle, it is necessary to know the azimuth and dip of one other axis or plane (Figure 1). The particle short axis (S - or c -axis) is at right angles to the a – b plane (i.e., it is the normal or pole to that plane), and provides useful information because the a – b plane, like the a -axis, is sensitive to forces experienced during deposition or deformation. For tills, the orientations of polished or fractured surfaces also provide valuable information on the origin of deposits. Taken together, measurements of a -axes, c -axes, and wear patterns can provide more insight into the origin of deposits than is possible with a -axis data alone (Benn, 1995; Hicock, 1991). However, the measurement of multiple fabric elements is time consuming and many researchers opt for the rapid collection of less comprehensive data. A sample size of 50 is usually chosen, but larger or smaller samples may be taken, depending on the aim of the investigation and the patience of the investigator. Care must be taken to avoid bias in sampling, so clasts are selected randomly from fresh vertical or horizontal exposures of the till. If the primary aim is to determine a -axis orientation, it is usual to choose only clasts for which the ratio between long and intermediate axis (a/b) is more than 1.5. This is because clasts with similar a - and b -axes will not be sensitive indicators of the forces and processes of interest.

Lithified Tills

For lithified tills, fabric measurement is complicated by the fact that the shape and orientation of clasts seen on an exposed rock

surface are not representative of the real three-dimensional (3D) fabric. The apparent long axis of a particle seen in section is not the true long axis, merely the longest dimension of the particular slice that happens to have been taken through the grain. It is possible to obtain some information on the true fabric by measuring 2D fabrics (i.e., azimuths or dips) on three polished sections oriented at right angles to each other. This approach is restricted to microfabric analysis because of the very large size of blocks required to obtain representative sample sizes for larger particles. 2D fabrics measured on bedding planes are likely to approximate true azimuth fabrics because clast long axes are commonly close to horizontal in tills, although clearly caution is recommended. If bedding planes are tilted, it is necessary to rotate the fabric data into the horizontal plane to convert them to azimuths.

Data Presentation

2D fabric data can be represented very effectively on rose diagrams. 3D data are usually plotted on stereonets or similar diagrams, in which particle axes or poles-to-planes plot as a point (Figure 2). Dips are represented by the distances of points from the center of the diagram, such that a vertically aligned particle axis plots at the exact center and a horizontally aligned axis plots on the rim. Azimuths are represented by the direction of the point relative to the center. Several variants of this type of diagram are in use, but the most common is constructed so that point densities on the diagram faithfully reflect, without distortion, data clustering in three dimensions, regardless of where on the diagram they occur. For this reason, they are known as equal-area stereonets. These diagrams may be intuitively less obvious than rose diagrams, but they communicate much more information to the trained user.

Summary Statistics

Quantitative comparison of fabric data from different sites or deposits can be done by using summary statistics. Numerous approaches have been proposed, but by far the most widely used statistics in till fabric analysis are based on the orientation tensor or eigenvalue method (Mark, 1973). This method resolves sets of azimuth and dip measurements into three orthogonal eigenvectors, V_1 , V_2 , and V_3 , where V_1 (the principal eigenvector) parallels the axis of maximum clustering in the data, and V_3 is normal to the preferred plane of the fabric. The degree of clustering of the data about the respective eigenvectors is given by the normalized vector magnitudes (eigenvalues) S_1 , S_2 , and S_3 , where $S_1 > S_2 > S_3$, and $S_1 + S_2 + S_3 = 1$. It is useful to visualize eigenvalues as three axes of ellipsoids that approximate the 'shape' of the data distribution in three dimensions.

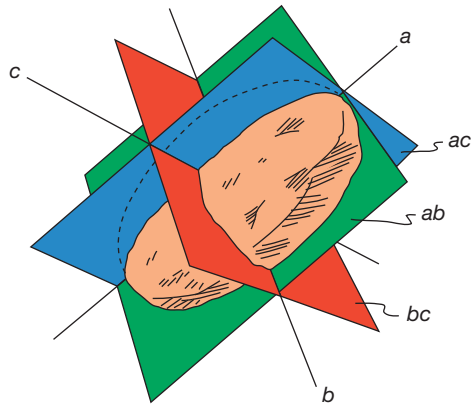


Figure 1 Definition of particle axes (a , b , c) and planes (ab , ac , bc). Note that the c -axis is at right angles to the ab plane, and so on. The orientation and dip of at least two axes and/or planes are required to uniquely define the alignment of any particle.

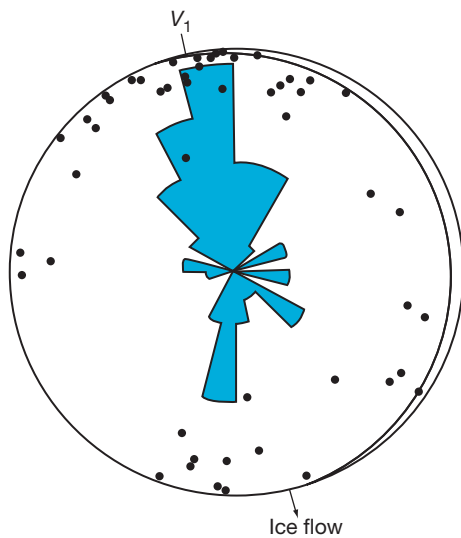


Figure 2 The same a -axis fabric data plotted in two ways: a 2D rose diagram (in blue) and an equal-area stereonet (point symbols). The rose diagram gives a more immediate impression of a -axis alignment, but the stereonet contains more information. Also indicated are the direction of the principal eigenvector, V_1 and the V_1 – V_2 plane.

Figure 3(a) shows the range of all possible eigenvalue combinations, plotted on a triangular diagram analogous to those used for particle shape data. Isotropic fabrics (with data points evenly distributed in all directions) have $S_1 \approx S_2 \approx S_3$; girdles (with points evenly distributed around a great circle) have $S_1 \approx S_2 \gg S_3$; and linear clusters (with all observations approximately parallel) have $S_1 \gg S_2 \approx S_3$. **Figure 3(a)** is scaled using an isotropy index $I = S_3/S_1$, and an elongation index $E = 1 - (S_2/S_1)$, which in combination can uniquely define any ellipsoidal fabric shape. Commercially available computer programs allow rapid calculation of eigenvectors and eigenvalues, in addition to plotting data on a choice of diagram. The mathematical basis of the orientation tensor method is clearly explained by Woodcock and Naylor (1983).

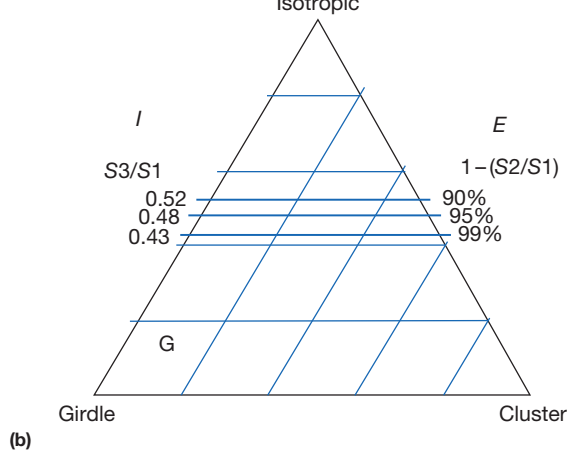
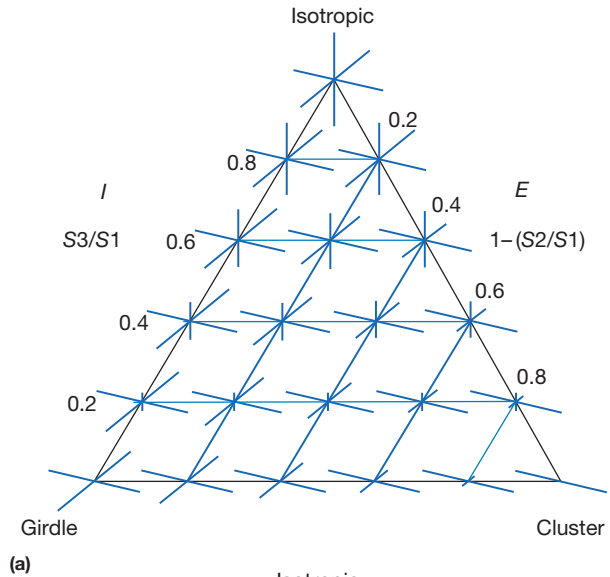


Figure 3 (a) Ratios between eigenvalues used to define a three-way continuum of fabric shapes; (b) graphical representation of the Woodcock–Naylor significance test. Reproduced from Benn DI (2004) *Macrofabric*. In: Evans DJA and Benn DI (eds.) *A Practical Guide to the Study of Glacial Sediments*, pp. 93–114. London: Arnold.

Statistical Analysis

Like all samples, till fabric data represent only one of the many possible sets of measurements that could have been taken from the parent population. Thus, some of the characteristics of samples (such as apparent preferred orientations) may be statistical artifacts, rather than inherent properties of the till. Statistical significance tests can be conducted to determine the probability that sample characteristics have arisen by chance sampling effects. This is relatively straightforward for 2D data (e.g., azimuths), which can be analyzed using standard statistical methods such as chi-square (Andrews, 1971). For 3D data, the problem is considerably more complex, and rigorous analytical methods have yet to be developed. One widely adopted test determines the probability that a fabric differs from a random (i.e., isotropic) distribution (Woodcock and Naylor, 1983). This has been misapplied by some researchers

to assess the 'significance' of preferred orientations, on the mistaken assumption that it refers to the direction of the principal eigenvector V_1 . A girdle fabric (G on Figure 3(b)) is clearly 'nonrandom' because the data are clustered in a preferred plane, although they have no preferred direction. The Woodcock–Naylor test is very useful for some applications, but it should not be used to determine whether apparent preferred orientations are significant.

A new approach to the problem of significance testing was developed by Ringrose and Benn (1997) who used bootstrapping methods to construct approximate 'confidence regions' for fabric eigenvalues. Bootstrapping methods are based on the analysis of multiple resamples from an original data set, and are so-called because they appear to generate information from nowhere (analogous to lifting yourself off the ground by your bootstraps). Irrespective of their theoretical basis, bootstrapping methods perform well, and can be applied to both synthetic and field data (Benn and Ringrose, 2001; Larsen and Piotrowski, 2003). Analyses of 'confidence regions' produced in this way shows that statistical variance is largest for isotropic fabrics, and smallest for strongly clustered fabrics, implying that large ($\gg 50$) samples are required to reduce the influence of statistical effects where fabrics are 'weak' (i.e., high isotropy or low elongation), but that relatively small samples (~ 30) may suffice when fabrics are 'strong' (i.e., low isotropy and high elongation).

Sample sizes are more often decided by the energy and attention span of the researcher than statistical considerations, and methods that allow objective sampling strategies to be developed should be welcomed. Another important statistical issue is that the calculation of eigenvalues rests on the assumption that the data have a 3D Gaussian distribution, such that they vary smoothly in a prescribed way and do not exhibit more than one cluster (mode). If this is not the case, the eigenvalue method will generate spurious results and eigenvectors may lie between two or more modes in the data. While this issue is widely acknowledged, the suitability of data for eigenvalue analysis is invariably assessed on a purely visual basis. Rigorous statistical tests for 3D normality are clearly needed.

Fabric Analysis and Till Formation

Fabric analysis has been used to address two main aims. The first, which dates back to the nineteenth century (Miller, 1884), is to use preferred a -axis orientations to reconstruct former ice flow directions, and it rests on the assumption that the two are parallel. While this is undoubtedly true for certain types of subglacial deposits (e.g., lodged boulders), in some tills, fabric maxima can deviate significantly from ice flow directions indicated by other criteria. With due caution, a -axis fabric maxima can provide indicators of general ice flow directions where other forms of evidence are lacking. The second aim is to use fabric characteristics to inform reconstructions of till genesis, and it has received particular attention since the 1980s.

Numerous studies have shown that fabrics taken from different till facies tend to have distinct ranges of eigenvalues, which plot as characteristic 'envelopes' on fabric shape diagrams (Dowdeswell et al., 1985; Benn, 1994; Figure 4). Such envelopes have then been used as control data to aid the interpretation of ancient deposits of unknown origin. This

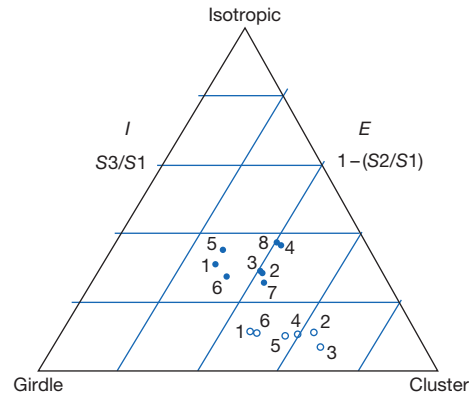


Figure 4 Eigenvalue plots of a -axis fabrics from two till facies at Breidamerkurjökull, Iceland. Solid circles: weak upper till; open circles: stiff lower till.

approach has been criticized by Bennett et al. (1999), who pointed out that there is often considerable overlap between eigenvalue envelopes for different facies, and therefore fabric eigenvalues do not constitute a unique genetic 'fingerprint.' This issue was already widely acknowledged, and does not in itself cause major problems if fabric data are interpreted in conjunction with other forms of evidence. A more fundamental problem with the use of fabric envelopes is that, for subglacial tills, the origin of even recently exposed examples is rarely well-known, and must be inferred from sedimentological characteristics (including fabric). This means that frequently there is no more information on the origin of recent tills than there is on ancient ones, so using them as analogs does not actually introduce any new information and initial misinterpretations will be perpetuated. This problem almost certainly besets the Bennett et al. data set as much as any other.

Uncertainties about the origin of control samples are compounded by the fact that many till fabrics are likely to have complex origins. Clast orientations can be affected by several processes during glacial transport and deposition, including englacial shear, drag between ice and the bed during ploughing and lodgement, and subglacial till deformation, and can be further modified by postdepositional processes such as dewatering and gravitational reworking. Most till fabrics, therefore, probably reflect complex histories, retaining memories of more than one process. Fabric overprinting is likely to be particularly important during deicing and dewatering of basal debris-rich ice, to the extent that unremoulded melt-out tills are probably very rare (Paul and Eyles, 1990). These issues clearly show that eigenvalue envelopes should not be regarded as simple templates allowing unambiguous categorization of any given deposit. Rather, they succinctly define the range of fabric shapes associated with a sediment facies, which can then be used to clarify thinking about its genesis, particularly in combination with other lines of evidence.

Significance of Till Fabrics

There has been considerable debate about the precise mechanisms that determine fabric development in subglacial environments. Two main processes have been proposed. The first,

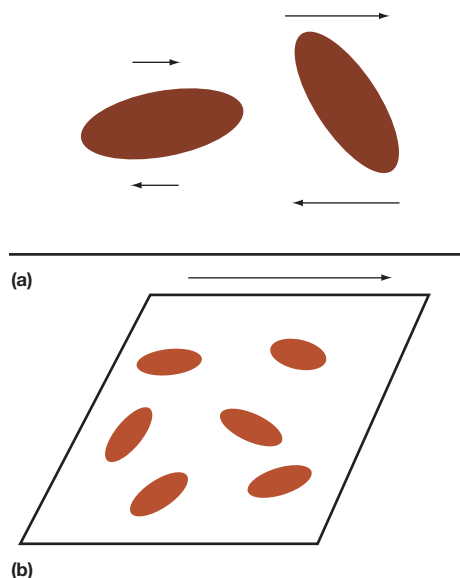


Figure 5 Schematic diagram illustrating the concepts of (a) Jeffery, and (b) March rotation. In (a) clasts are continuously rotated as the result of vertical velocity gradients, whereas in (b) clasts passively trace the deformation of the surrounding medium.

known as Jeffery rotation, invokes the rotation of clasts in response to vertical velocity gradients in a shearing medium, which may be either ice or till (Glen et al., 1957; Hicock and Dreimanis, 1992; Figure 5(a)). Because the medium above the clast moves faster than that below, the clast is subject to a force couple, which supposedly causes clasts to continuously roll about a transverse horizontal axis. Under this process, fabrics do not settle into a final steady state, but vary continuously with a weak to moderate mean preferred orientation down-flow. Collisions between clasts are thought to produce transverse preferred orientations under sustained shearing. The second mechanism, March rotation, regards clasts as passive markers, which progressively tilt in response to strain in the surrounding medium (Figure 5(b)). During both simple and pure shear, lines are tilted towards the horizontal, rapidly at first, then by decreasing increments at high strains. Under this mechanism, clasts quickly adopt a strong preferred orientation parallel to the direction of shear, then undergo little or no change thereafter. Force couples across low-dip clasts do not result in rotation, and the overpassing medium slips over the clast instead. These contrasting mechanisms of clast orientation have been invoked to support alternative models of basal till formation.

Many massive subglacial tills have weak to moderate *a*-axis fabrics, and some researchers have argued that this reflects Jeffery rotation in viscous, subglacial deforming layers (Hart, 1994; Hicock and Dreimanis, 1992). Alternatively, Benn (1995) and Benn and Evans (1996) proposed that weak fabrics might develop by unsteady March rotation in a deforming layer, in response to fluctuations in water pressure, clast interactions, or other processes. However, Hooyer and Iverson (2000) have shown that till samples deformed to high cumulative strains in ring-shear devices invariably develop strong *a*-axis fabrics by March rotation. This result has been used to argue that tills with weak or moderate fabrics cannot represent former deforming layers (Hooyer and Iverson, 2002), but

probably form by the 'ploughing' of passive till by overpassing ice. There are, therefore, conflicting interpretations of weak fabrics in massive tills, some of which regard them as evidence for pervasive deformation, and some that argue that they are evidence against pervasive deformation.

Concluding Remarks

While ring-shear devices are extremely valuable tools that have already yielded major insights into fabric evolution and till genesis, they cannot yet control for important variables such as water pressure and sediment heterogeneity, and so are not capable of replicating the range of conditions found in nature. Consequently, it is too early to say whether till deformation always results in strong fabrics, and it is important to avoid replacing one orthodoxy (massive tills with weak fabrics = deforming beds) with another (massive tills with weak fabrics = ploughed beds). It is not known how fabrics might develop during ploughing, but it can be suggested that it will cause localized deformation, and transient, spatially variable March rotation. Thus, the ploughing process may well account for relatively weak fabrics in massive tills. Indeed, it seems probable that relatively weak fabrics can develop in subglacial tills by a variety of processes, so it is unwise to use them as evidence for one particular process without careful weighing of other evidence. The current disagreement about the meaning of till fabrics reflects a wider debate about the interpretation of till facies (Piotrowski et al., 2002, 2005; van der Meer et al., 2003). With careful research design, fabric studies have the potential to contribute significantly to this debate, particularly if multiple fabric elements are analyzed.

See also: **Glacial Landforms:** Evidence of Glacier Recession.
Glacial Landforms, Ice Sheets: Evidence of Glacier and Ice Sheet Extent. **Glacial Landforms, Sediments:** Clast Form Analysis; Micromorphology of Glacial Sediments; Tills.

References

- Andrews JT (1971) Techniques of till fabric analysis. *British Geomorphological Research Group Technical Bulletin*, Geo Abstracts.
- Benn DI (1994) Fabric shape and the interpretation of sedimentary fabric data. *Journal of Sedimentary Research* A64: 910–915.
- Benn DI (1995) Fabric signature of till deformation, Breidamerkurjökull, Iceland. *Sedimentology* 42: 735–747.
- Benn DI (2004) Macrofabric. In: Evans DJA and Benn DI (eds.) *A Practical Guide to the Study of Glacial Sediments*, pp. 93–114. London: Arnold.
- Benn DI and Evans DJA (1996) The interpretation and classification of subglacially-deformed materials. *Quaternary Science Reviews* 15: 23–52.
- Benn DI and Ringrose T (2001) Random variation of fabric eigenvalues: Implications for the use of *a*-axis fabric data to differentiate till facies. *Earth Surface Processes and Landforms* 26: 295–306.
- Bennett MR, Waller RI, Glasser NF, Hambrey MJ, and Huddart D (1999) Glacigenic clast fabrics: Genetic fingerprint or wishful thinking? *Journal of Quaternary Science* 14: 125–135.
- Dowdeswell JA, Hambrey MJ, and Wu R (1985) A comparison of clast fabric and shape in Late Precambrian and modern glacigenic sediments. *Journal of Sedimentary Petrology* 55: 691–704.
- Glen JW, Donner JJ, and West RG (1957) On the mechanism by which stones in till become oriented. *American Journal of Science* 255: 194–205.

- Hart JK (1994) Till fabric associated with deformable beds. *Earth Surface Processes and Landforms* 19: 15–32.
- Hicock SR (1991) On subglacial stone pavements in till. *Journal of Geology* 99: 607–619.
- Hicock SR and Dreimanis A (1992) Deformation till in the Great Lakes region: Implications for rapid flow along the south-central margin of the Laurentide Ice Sheet. *Canadian Journal of Earth Sciences* 29: 1565–1579.
- Hooyer TS and Iverson NR (2000) Clast-fabric development in a shearing granular material: Implications for subglacial till and fault gouge. *Geological Society of America Bulletin* 112: 683–692.
- Hooyer TS and Iverson NR (2002) Flow mechanism of the Des Moines lobe of the Laurentide ice sheet. *Journal of Glaciology* 48: 575–586.
- Larsen NK and Piotrowski JA (2003) Fabric pattern in a basal till succession and its significance for reconstructing subglacial processes. *Journal of Sedimentary Research* 73: 725–734.
- Mark DM (1973) Analysis of axial orientation data, including till fabrics. *Geological Society of America Bulletin* 84: 1369–1374.
- Miller H (1884) On boulder-glaciation. *Proceedings of the Royal Physical Society of Edinburgh* 8: 156–189.
- Paul MA and Eyles N (1990) Constraints on the preservation of diamict facies (melt-out tills) at the margins of stagnant glaciers. *Quaternary Science Reviews* 9: 51–69.
- Piotrowski JA, Mickelson DM, Tulaczyk S, Krzyszkowski D, and Junge FW (2002) Deforming soft beds under ice sheets: How extensive were they? In: Boulton GS, Dobbie KE, and Zatsepin S (eds.) *Quaternary International* 97–98: 173–177.
- Piotrowski JA, Larsen NK, Menzies J, and Wysota W (2005) Formation of subglacial till under transient bed conditions: Deposition, deformation and basal decoupling under a Weichselian ice sheet lobe central Poland. *Sedimentology* 53: 83–106.
- Ringrose T and Benn DI (1997) Confidence regions for fabric shape diagrams. *Journal of Structural Geology* 19: 1527–1536.
- van der Meer JJM, Menzies J, and Rose J (2003) Subglacial till: The deforming glacier bed. *Quaternary Science Reviews* 22: 1659–1685.
- Woodcock NH and Naylor MA (1983) Randomness testing in three-dimensional orientation data. *Journal of Structural Geology* 5: 539–548.

Glacial Erratics and Till Dispersal Indicators

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Erratics

The occurrence of boulders in areas far removed from their bedrock source has been recognized as incontrovertible evidence for glacial transportation since the dawn of glacial theory. Very clear and widely celebrated examples include the Scottish Ailsa Craig microgranites in the glacial sediments surrounding the Irish Sea Basin; the Scandinavian erratics in the tills of eastern England, northern Germany, Denmark, and The Netherlands; and the quartzite blocks of the 'Foothills erratics train' lying on the Cretaceous bedrock prairie of southern Alberta more than 375 km from their source in Jasper National Park (Stalker, 1956, 1976). Many such erratic trains can be easily traced to their source outcrops, for example, the Assynt sandstone boulder trains of northwest Scotland, which extend down-ice from mountains of Torridonian Sandstone (Lawson, 1995; Figure 1), but their use in the reconstruction of former transportation pathways must be tempered by an appreciation that debris transport histories may involve several glacial cycles during which ice divides and ice-flow vectors may shift dramatically. This is well illustrated by reconstructions of the complex ice flow history of the central sector of the British-Irish Ice Sheet, wherein cross-cutting subglacial bedforms and erratic transport pathways record overprinting by radically changing ice flow vectors, driven by the migration of ice dispersal centers during the last glacial cycle (Evans et al., 2009; Livingstone et al., 2008); early southerly flows from Scotland were responsible for the transport of Scottish erratics through the Tyne Gap, up the Vale of Eden and over Stainmore, toward the eastern English coast, whereas later stages of ice flow were directed northward down the Vale of Eden and into the Solway Lowlands, making Scottish erratic transport pathways puzzling to early geologists.

Indicator erratics are those for which a definite source area is known, and indicator fans are regional fields over which the erratics from particular source rocks have been dispersed by glacier flow. Indicator fans encompass the range of transport directions produced by shifting ice divides and dispersal centers, and can be identified using a range of sediment sizes, from erratic blocks down to the fine-grained matrix of tills, their recognition coming through the analysis of till matrix rather than individual erratics, in a procedure known as till provenancing or till prospecting. When combined with other glacial landforms, such as subglacial lineations and ice-scoured bedrock, the dispersal of indicator erratics can highlight large paleo-ice streams. For example, the dispersal of granite erratics from the Canadian Shield into Massey Sound and Wellington Channel in the Canadian High Arctic demarcates the operational zone of a large ice stream in the Late Quaternary Inuitian Ice Sheet (Atkinson, 2003; O'Cofaigh et al., 2000). Similarly, the 'omars' of Prest et al. (2000) are diagnostic erratics that were dispersed by the Laurentide Ice Sheet from the Belcher Islands in Hudson Bay.

Indicator Fans and Till Dispersal Indicators

The concentrations of indicator erratics vary systematically along ice-flow lines. Within indicator outcrops, concentrations increase rapidly down-glacier reflecting the addition of new material from the glacier bed, but concentrations drop off rapidly down-ice of the outcrop margin (Gillberg, 1965; Perttunen, 1977). This simple distance decay effect may be complicated by shifting patterns of erosion and sediment accumulation during glacier advance and retreat, and in some cases, the peak in erratic concentration may be displaced down-ice from the boundary of the source outcrop (Boulton, 1996). The up-ice and down-ice limits of an indicator plume are known as the head and tail, respectively (Figure 2).

Research by DiLabio and Shilts (1979) on debris dispersal by modern glaciers has demonstrated that head and tail zones are identifiable in the lateral moraines of glaciers crossing from one bedrock type to another. The transport distances of the majority of indicator erratics are relatively short. A good index of transport distance is the half-distance value, or the distance from the nearest possible source to the point at which the frequency of indicator erratics is half of its maximum. The half-distance value varies with the material resistance to abrasion and crushing and the depth to the source rock beneath superficial sediment (Bouchard and Salonen, 1990). Salonen (1986; 1987) found that the average half-distances for particles in tills in Finland are only 5 km. Very far-traveled erratics (up to 1200 km in northern Europe) may have been transported during several glacial cycles and part of their journey may have been by iceberg rafting (e.g., Marcussen, 1973).

Dyke and Morris (1988) suggested a twofold classification scheme for erratic trains based upon examples in Canada (Figure 3): (i) 'The Dubawnt type' is named after the Dubawnt sandstone in central Keewatin, which is a relatively restricted source outcrop from which material was dispersed as a plume over the surrounding bedrock by ice moving at a uniform rate over an entire region. Another example of this type of dispersal train is the 40-km long plume located down flow from the Strange Lake alkalic intrusion in northern Labrador, Canada; (ii) 'The Boothia type', based on examples on the Boothia Peninsula, is characterized by debris plumes that extend from small parts of large source areas by zones of more rapid ice flow (ice streams) within the ice sheet. Vertical changes in till composition at a particular site may result from (i) the complete erosion or burial of some source outcrops, (ii) a shift in ice-flow direction so that material is transported from different source areas, or (iii) time-dependent patterns of erratic transport along a flowline. Stea et al. (1989) and Turner and Stea (1990) have linked till characteristics to ice divide migrations in Nova Scotia during the last glaciation. Early shifts in regional ice divides led to the influx of far-traveled debris to the area, forming the Hartlen Till and Lawrencetown Till, and later local

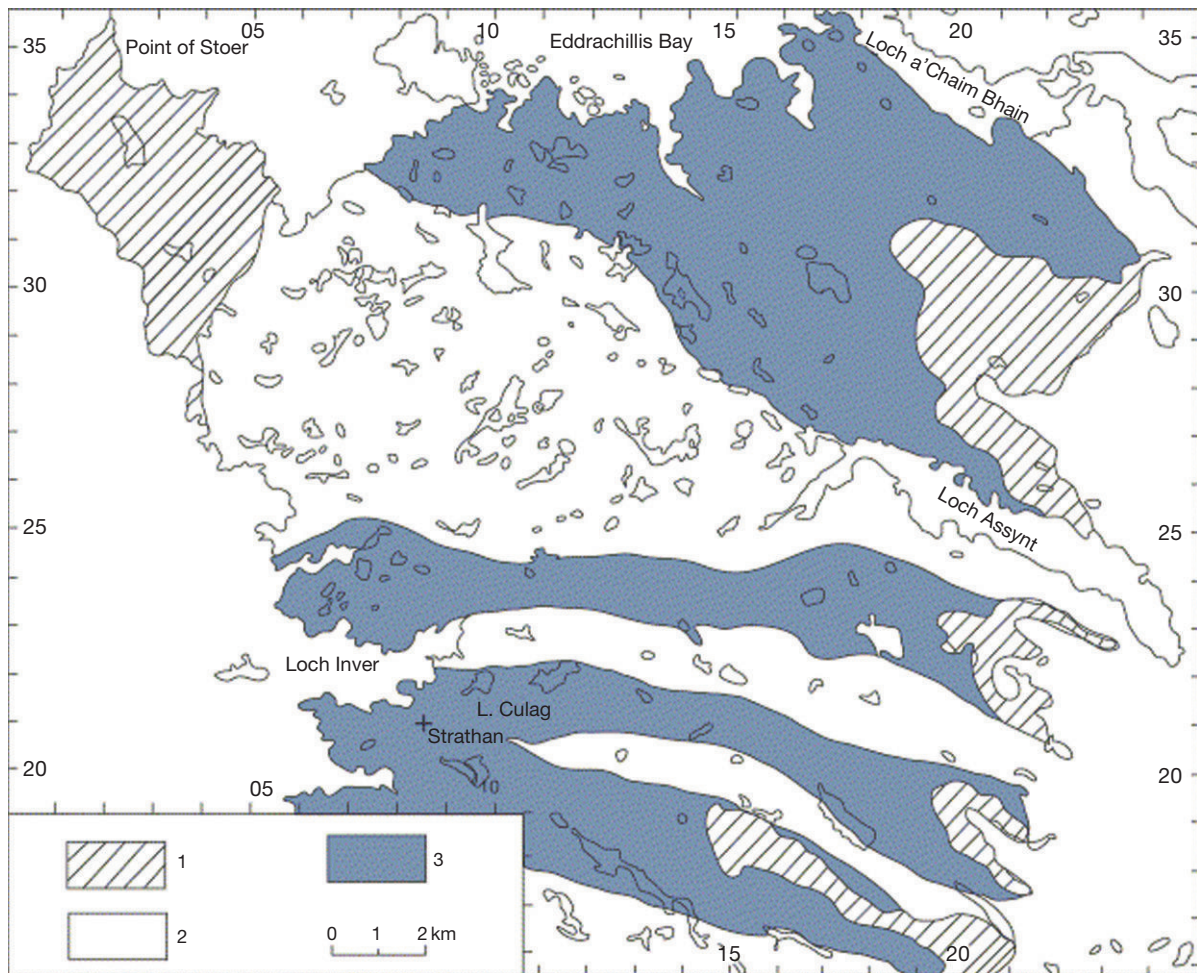


Figure 1 The boulder trains of Assynt, Scotland, indicating former ice-sheet flow from east to west. Plumes of Torridonian sandstone have been carried across neighboring bedrock surfaces. (1) Torridonian sandstone, (2) areas devoid of Torridonian sandstone boulders, and (3) areas containing Torridonian sandstone boulders. Reproduced from [Lawson TJ \(1995\)](#) Boulder trains as indicators of former ice flow in Assynt, N.W. Scotland. *Quaternary Newsletter* 75: 15–21.

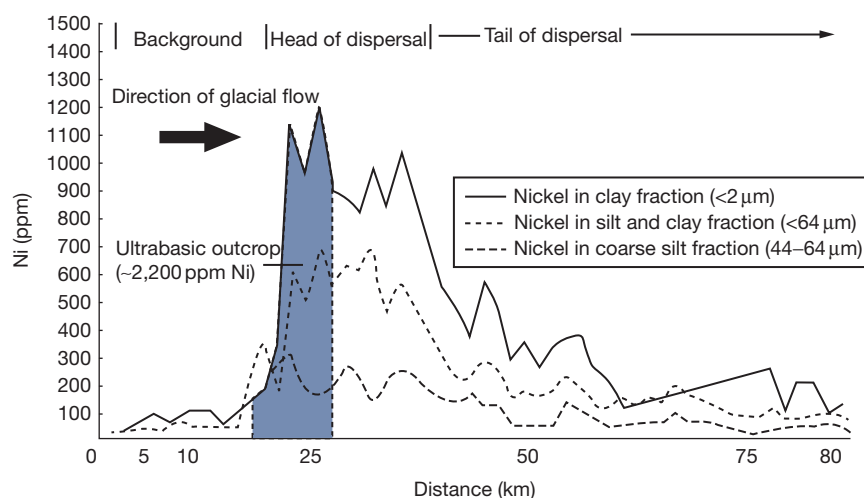


Figure 2 Dispersal curves for nickel concentrations in fine and coarse fractions of tills in a down-glacier direction from an ophiolitic complex at Thetford Mines, PQ, Canada. Reproduced from [Shilts WW \(1993\)](#) Geological Survey of Canada's contributions to understanding the composition of glacial sediments. *Canadian Journal of Earth Sciences* 30: 333–353.

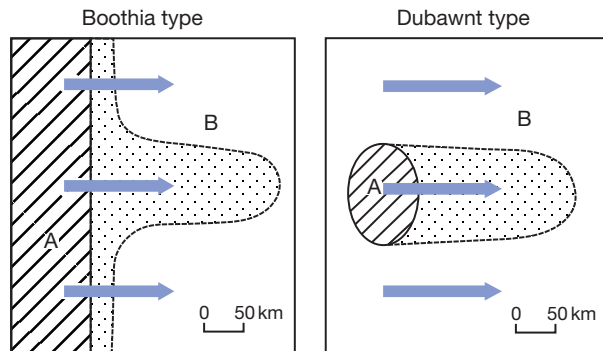


Figure 3 The twofold classification scheme for dispersal trains based upon the shape of the source outcrop and ice flow characteristics. Reproduced from Dyke AS and Morris TF (1988) Drumlin fields, dispersal trains and ice streams in arctic Canada. *Canadian Geographer* 32: 86–90. Two different rock types are marked A and B and the stippled area is dispersal plume.

ice dispersal resulted in the deposition of the Stony Till, with local lithological signatures.

The identification of paleo-ice lobes from till lithology has been attempted in a number of settings, and considerable success has been achieved in determining the source areas of individual till sheets in, for example, the Great Lakes region of North America (Gwyn and Dreimanis, 1979); the Canadian prairies (Shetsen, 1987); eastern England (Madgett and Catt, 1978); central Quebec and Labrador, Canada (Klassen and Thompson, 1989); and Finland (Hirvas and Nenonen, 1985), although Boston et al. (2010) have demonstrated that considerable intratill geochemical variability can arise from the stacking and cannibalization of till sheets at dynamic ice sheet margins. Dispersal patterns beneath former ice divides tend to be complex, and subglacial material can be transported in several different directions as a result of ice-divide migration. The resulting dispersal pattern consists of multiple plumes emanating from a source outcrop, termed an ‘amoeboid pattern’ by Shilts (1993). Indicator fans or dispersal fans provide important clues to the location of mineral outcrops or ore bodies in glaciated terrain, where elongate plumes of mineral-enriched tills extend down-ice from source outcrops (DiLabio, 1981; 1990; Shilts, 1976; 1993). The mineral content of the plumes is known as a mineral ‘float.’ Ore enrichment in tills can be assessed by pebble counts and/or geochemical and mineralogical analyses of the fine-grained matrix. Because ore bodies are often hidden by till, the geochemical or mineralogical characteristics of till samples, collected on a regular grid and then contoured, often reveal plume-shaped areas of metal enrichment or dispersal trains.

The largest dispersal trains can be hundreds of kilometers long but derived from very small ore bodies; they are hundreds to thousands of times larger than their source area, thus representing a much larger exploration target for prospectors than the ore body itself. Some dispersal trains can be identified visually on the basis of their color, such as the carbonate till plumes that stretch across Prince of Wales Island, Somerset Island, and Boothia Peninsula in the central Canadian Arctic archipelago (Dyke and Morris, 1988; Dyke et al., 1992) and central Baffin Island (Dredge, 2000; Dyke, 2008; Tippet, 1985).

Another excellent example is the variation in nickel concentration in till in Quebec, Canada (Shilts, 1993). The dispersal curves show only background levels of nickel up-ice from the source ultrabasic bedrock outcrop, and head and tail zones down-ice from the outcrop. The concentrations of nickel vary with grain size as a result of differential abrasion and crushing during subglacial transport. Each mineral has a different mode at which its chemical signature is strongest (‘chemical partitioning’), and this must be taken into account when analyzing till samples for their geochemical signatures (DiLabio, 1982; Nikkarinen et al., 1984; Shilts, 1991; Shilts and Wyatt, 1989).

Anomalous geochemical signals can be detected in till samples due to weathering and pedogenesis, the depth of which is dictated by such factors as the water table and grain size (e.g., Peuraniemi, 1984; Shilts, 1976; Shilts and Kettles, 1990). Therefore, till samples for geochemical analysis and indicator tracing must be taken from unweathered horizons in order to avoid postdepositional alterations such as the destruction of labile ore minerals and their depletion in till samples from above the water table.

See also: **Tills.**

References

- Atkinson N (2003) Late Wisconsinan glaciation of Amund and Ellef Ringnes islands, Nunavut: Evidence for the configuration, dynamics and deglacial chronology of the northwest sector of the Innuitian Ice Sheet. *Canadian Journal of Earth Sciences* 40: 351–363.
- Boston CM, Evans DJA, and O’Cofaigh C (2010) Styles of till deposition at the margin of the Last Glacial Maximum North Sea lobe of the British–Irish Ice Sheet: An assessment based on geochemical properties of glacial deposits in eastern England. *Quaternary Science Reviews* 29: 3184–3211.
- Bouchard MA and Salonen VP (1990) Boulder Transport in Shield Areas. In: Kujansuu R and Saarnisto M (eds.) *Glacier Indicator Tracing*, pp. 87–107. Rotterdam: Balkema.
- Boulton GS (1996) Theory of glacial erosion, transport and deposition as a consequence of subglacial sediment deformation. *Journal of Glaciology* 42: 43–62.
- Clark PU (1987) Subglacial sediment dispersal and till composition. *Journal of Geology* 95: 527–541.
- DiLabio RNW (1981) Glacial dispersal of rocks and minerals at the south end of Lac Mistassini, Quebec, with special reference to the Icon dispersal train. *Geological Survey of Canada Bulletin* 323: 46.
- DiLabio RNW (1982) *Gold and Tungsten Abundance vs Grain Size in Till at Waverly, Nova Scotia*, pp. 57–62. Geological Survey of Canada, Paper 82-1B.
- DiLabio RNW (1990) Glacial dispersal trains. In: Kujansuu R and Saarnisto M (eds.) *Glacier Indicator Tracing*, pp. 109–122. Rotterdam: Balkema.
- DiLabio RNW and Coker WB (eds.) (1989) *Drift Prospecting*, p. 169. Geological Survey of Canada, Paper 89-20.
- DiLabio RNW and Shilts WW (1979) Composition and dispersal of debris by modern glaciers, Bylot Island, Canada. In: Schluchter C (ed.) *Moraines and Varves*, pp. 145–155. Rotterdam: Balkema.
- Dredge LA (2000) Carbonate dispersal trains, secondary till plumes, and ice streams in the west Foxe Sector, Laurentide Ice Sheet. *Boreas* 29: 144–156.
- Dyke AS (2008) The Steensby Inlet ice stream in the context of the deglaciation on Northern Baffin Island, Eastern Arctic Canada. *Earth Surface Processes and Landforms* 33: 573–592.
- Dyke AS and Morris TF (1988) Drumlin fields, dispersal trains and ice streams in arctic Canada. *Canadian Geographer* 32: 86–90.
- Dyke AS, Morris TF, Green DEC, and England J (1992) *Quaternary Geology of Prince of Wales Island, Arctic Canada*, p. 142. Geological Survey of Canada, Memoir 433.
- Evans DJA, Livingstone SJ, Vieli A, and O’Cofaigh C (2009) The palaeoglaciology of the central sector of the British and Irish Ice Sheet: Reconciling glacial geomorphology and preliminary ice sheet modelling. *Quaternary Science Reviews* 28: 739–757.

- Gillberg G (1965) Till distribution and ice movements on the northern slopes of the south Swedish highlands. *Geologiska Föreningens i Stockholm Förhandlingar* 86: 433–484.
- Gwyn QHJ and Dreimanis A (1979) Heavy mineral assemblages in tills and their use in distinguishing glacial lobes in the Great Lakes region. *Canadian Journal of Earth Sciences* 16: 2219–2235.
- Hirvas H and Nenonen K (1985) The till stratigraphy of Finland. In: Kujansuu R and Saarnisto M (eds.) *INQUA Till Symposium*, pp. 49–63. Geological Survey of Finland Special, Paper 3.
- Jones MJ (ed.) (1973) *Prospecting in Areas of Glaciated Terrain*. London: Institution of Mining and Metallurgy.
- Klassen RA and Thompson FJ (1989) Ice flow history and glacial dispersal patterns, Labrador. In: DiLabio RNW and Coker WB (eds.) *Drift Prospecting*, pp. 21–29. Geological Survey of Canada, Paper 89-20.
- Kujansuu R and Saarnisto M (eds.) (1990) *Handbook of Glacial Indicator Tracing*. Rotterdam: Balkema.
- Lawson TJ (1995) Boulder trains as indicators of former ice flow in Assynt, N.W. Scotland. *Quaternary Newsletter* 75: 15–21.
- Livingstone SJ, O'Cofaigh C, and Evans DJA (2008) Glacial geomorphology of the central sector of the last British–Irish Ice Sheet. *Journal of Maps* 2008: 358–377.
- Madgett PA and Catt JA (1978) Petrography, stratigraphy and weathering of Late Pleistocene tills in east Yorkshire, Lincolnshire and north Norfolk. *Proceedings of the Yorkshire Geological Society* 42: 55–108.
- Marcussen I (1973) Stones in Danish tills as a stratigraphical tool. A review. *Bulletin of the Geological Institutions of Uppsala New Series* 5: 177–181.
- Nikkarinen M, Kallio E, Lestinen P, and Ayra M (1984) Mode of occurrence of Cu and Zn in till over three mineralized areas in Finland. In: Björklund AJ (ed.) *Geochemical Exploration – 1983. Journal of Geochemical Exploration* 21: 239–247.
- O'Cofaigh C, England J, and Zreda M (2000) Late Wisconsinan glaciation of southern Eureka Sound: Evidence for extensive Inuitian ice in the Canadian High Arctic during the Last Glacial Maximum. *Quaternary Science Reviews* 19: 1319–1341.
- Perttunen M (1977) The lithological relation between till and bedrock in the region of Hameenlinna, southern Finland. *Geological Survey of Finland Bulletin* 291: 68.
- Peuraniemi V (1984) Weathering of sulphide minerals in till in some mineralized areas in Finland. In: Jones MJ (ed.) *Prospecting in Areas of Glaciated Terrain*, pp. 127–135. London: Institution of Mining and Metallurgy.
- Prest VK, Donaldson JA, and Mooers HD (2000) The omar story: The role of omars in assessing glacial history of west-central North America. *Geographie Physique et Quaternaire* 54: 257–270.
- Salonen VP (1986) Glacial transport distance distributions of surface boulders in Finland. *Geological Survey of Finland Bulletin* 338: 57.
- Salonen VP (1987) *Observations on Boulder Transport in Finland*, pp. 103–110. Finland: Geological Survey of Finland, Special Paper 3.
- Shetsen I (1987) *Quaternary Geology, Southern Alberta*. Edmonton, AB: Alberta Research Council Map.
- Shilts WW (1976) Glacial till and mineral exploration. In: Leggett RF (ed.) *Glacial till*, pp. 205–224. Royal Society of Canada Special Publication 12.
- Shilts WW (1982) Glacial dispersal: Principles and practical applications. *Geoscience Canada* 9: 42–48.
- Shilts WW (1991) Principles of glacial dispersal and sedimentation. In: Coker WB (ed.) *Exploration Geochemistry Workshop*, pp. 2/1–2/42. Geological Survey of Canada Open File 2390.
- Shilts WW (1993) Geological Survey of Canada's contributions to understanding the composition of glacial sediments. *Canadian Journal of Earth Sciences* 30: 333–353.
- Shilts WW and Kettles IM (1990) Geochemical/mineralogical profiles through fresh and weathered till. In: Kujansuu R and Saarnisto M (eds.) *Handbook of Glacial Indicator Tracing*, pp. 187–216. Rotterdam: Balkema.
- Shilts WW and Wyatt PH (1989) *Gold and Base Metal Exploration using Drift as a Sample Medium, Kaminak Lake – Turquetal Lake Area, District of Keewatin*, p. 32. Geological Survey of Canada Open File 21.
- Stalker A MacS (1956) The erratics train, foothills of Alberta. *Geological Survey of Canada, Bulletin* 37: 31.
- Stalker A MacS (1976) *Megablocks, or the Enormous Erratics of the Albertan Prairies*, pp. 185–188. Canada: Geological Survey of Canada, Paper 76-1 C.
- Stea RR, Turner RG, Finck PW, and Graves RM (1989) Glacial dispersal in Nova Scotia: A zonal concept. In: DiLabio RNW and Coker WB (eds.) *Drift Prospecting*, pp. 155–169. Geological Survey of Canada, Paper 89-20.
- Tippett CR (1985) Glacial dispersal train of Paleozoic erratics, central Baffin Island, NWT, Canada. *Canadian Journal of Earth Sciences* 22: 1818–1826.
- Turner RG and Stea RR (1990) Interpretation of till geochemical data in Nova Scotia, Canada, using mapped till units, multi-element anomaly patterns and the relationship of till clast geology to matrix geochemistry. *Journal of Geochemical Exploration* 37: 225–254.

Glacial Sequence Stratigraphy

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Introduction

The concept of sequence stratigraphy (see Sequence stratigraphy) has recently been adapted for application to the study of glacial depositional systems in order to understand large-scale depositional cycles in formerly glaciated basins. This approach seeks to identify large-scale sequences of conformable, genetically related strata bounded by unconformities or other major discontinuities (van Wagoner et al., 1988; Walker, 1992). Sequences comprise strata that may be grouped into systems tracts, which are defined as linkages of contemporaneous depositional systems. Traditionally, this was tailored to the study of shallow marine environments, in which systems tracts were defined in terms of major erosional and depositional episodes associated with global cycles of sea-level change, but this is not always relevant to glacial depositional settings and, therefore, modifications to the sequence stratigraphy approach have been necessary to employ it in studies of glacial sedimentology.

The characteristics of glacial sediments are dictated by the process–form relationships of specific depositional zones, such as subglacial, supraglacial, subaqueous ice-proximal, and distal and proglacial (see Glaciogenic lithofacies) environments. Glaciation is typically recorded by a stratigraphy comprising the superimposition of the deposits of these zones, documenting cycles of glacial advance and retreat. During ice sheet advance, the ice-proximal and subglacial zones migrate, in turn, over the former ice-distal zone, depositing ice-contact facies associations on top of distal sediments. Such associations tend to be eroded, transported, or redeposited in the subglacial zone once glacier ice advances over the site. As the ice sheet retreats, the ice-proximal and distal zones migrate, in turn, over the former subglacial zone, commonly blanketing subglacial forms with a fining-upward drape in subaqueous settings or glaciofluvial outwash in subaerial settings. This simple sequence can be complicated by changes between subaqueous and subaerial conditions and/or ice-marginal oscillations.

Sequence Stratigraphy Related to Glacially Influenced Sea-Level Change

A significant problem associated with applying the traditional sequence stratigraphy approach to glacial depositional settings is the fact that patterns of local sea-level change in glaciated areas are complicated by glacioisostatic depression of the crust, which results in complex sea-level histories with sea-level trends varying across glaciated basins (Andrews, 1970; Bednarski, 1988; Boulton, 1990; Dyke, 1979; Quinlan & Beaumont, 1981). A typical sequence stratigraphic signature for a glaciated marine environment subject to glacioisostatic depression is presented by Bednarski (1988) for the Canadian High Arctic, based on a modification of Curry's (1964)

definitions of transgressive and regressive sedimentary cycles (Figure 1). This depicts a depositional sequence produced during a sea-level transgression–regression cycle following on from glaciation. During deglaciation, receding ice allows marine waters to flood the basin, forming an erosional transgression surface (VII) over glaciogenic sediments or till. A sequence of marine deposits is then laid down on top of the erosion surface, interfingering with terrestrial glaciogenic sediments at the landward end (VI and V). Subsequent regression can result in erosion, which truncates glaciomarine sediments deposited during transgression and highstand (I). Alternatively, glaciomarine sediment can be reworked into littoral features such as beaches, storm ridges, or tidal flats (II, III, and IV). These 'forced regressions' associated with glacioisostatic rebound can be interrupted by short-lived transgressions due to local ice oscillations (Helle, 2004). This means that the grouping of contemporaneous depositional systems into systems tracts on the basis of regional marine erosion surfaces is impractical (Eyles & Eyles, 1992). A further significant complication arises in glacial sedimentary sequences due to the common situation where the injection of sediment into the water column from a glacier tends to be below the water level rather than at its surface (Figure 2(a)). This results in the deposition of subaqueous fans or grounding line fans (Cheel & Rust, 1982; Powell, 1990; Rust & Romanelli, 1975), which may ultimately grade to the water level to produce ice-contact deltas if glacier margins remain quasi-stable and/or sediment loads are sufficiently high (e.g., Lønne, 1995; Lønne, 2001; Lønne et al., 2001; Figure 3). This effectively renders the sequence stratigraphic terms 'highstand' and 'lowstand' meaningless in ice-contact depositional environments.

Sequence Stratigraphy Related to Terrestrial Glaciogenic Deposits

Terrestrially based glacial depositional systems, such as those accumulating in lakes or below grounded glacier ice, are generally controlled by factors other than sea-level change and so, it is necessary to modify the traditional sequence stratigraphy approach significantly. An excellent example of this is the study of Brookfield and Martini (1999), which highlights the two main controls on sequence stratigraphy in nonglacial and glacial settings as (1) the water level relative to the depositional surface, and hence the accommodation space for sediment accumulation and (2) the point of sediment injection. In nonglacial settings, these controls will vary in tandem, allowing systems tracts to be defined by the rise and fall in water level. In contrast, in glaciated basins, the accommodation space and the injection points are controlled by the water level and the position of the glacier terminus (e.g., Winsemann et al., 2004). Figure 2 shows an idealized sequence stratigraphy related to changing lake levels and glacier input points.

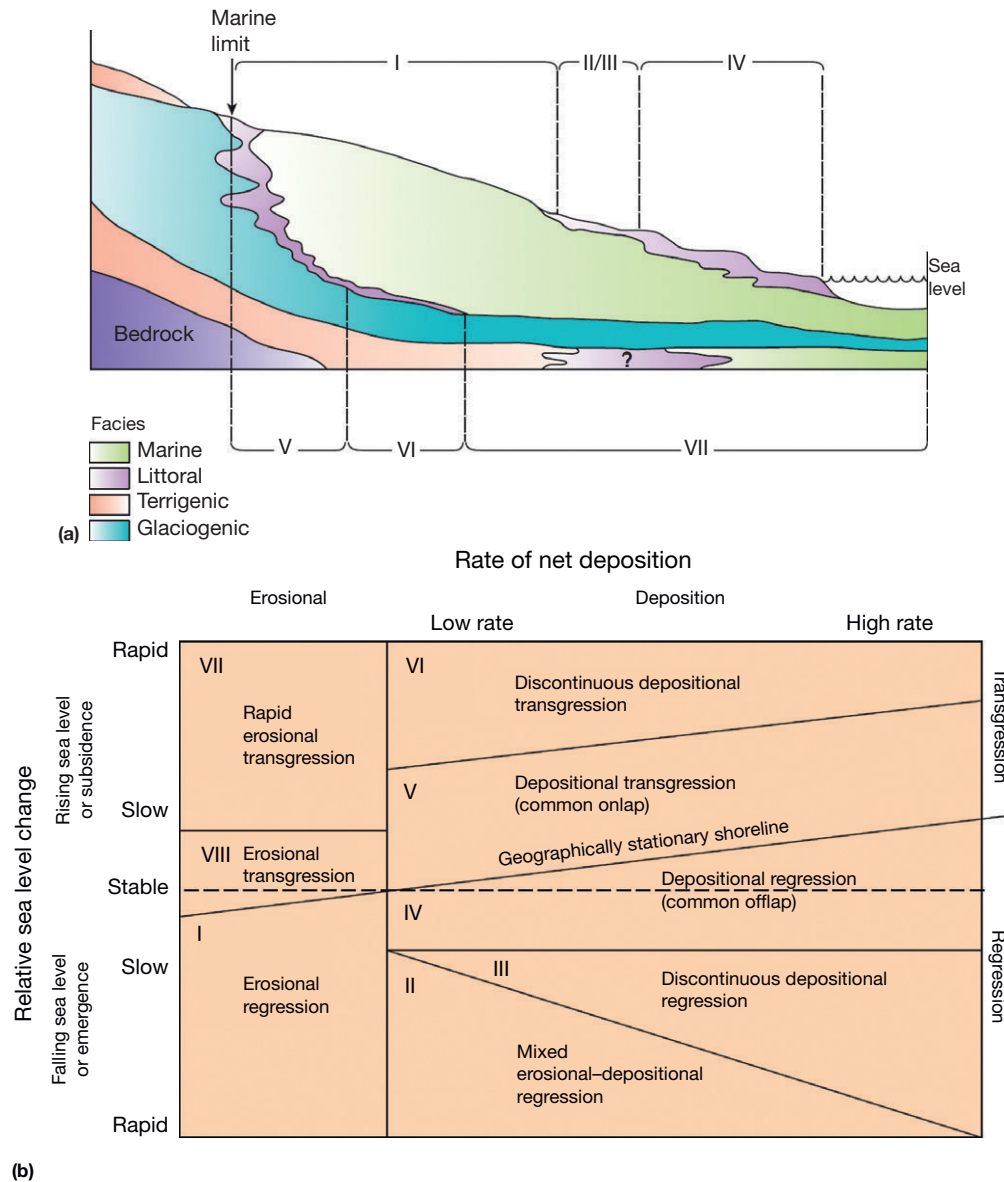


Figure 1 An idealized sequence stratigraphy of glaciomarine deposition on a deglaciated coastline: (a) idealized depositional sequences I–VII produced on a deglaciated coast subject to glacioisostatic depression and uplift during a glacial–deglacial cycle; (b) classification of transgressions and regressions based on the work of Curran (1964), with sedimentary sequences I–VII identified. Reproduced from Bednarski J (1988) The geomorphology of glaciomarine sediments in a high arctic fiord. *Geographie Physique et Quaternaire* 42: 65–74.

In Figure 2, a glacier snout and lake-level changes are related to the sequential change in localized depositional styles. This is translated in Figure 2(b) into spatial and temporal correlations of facies assemblages in the lake level system and the glacial input system and also shows systems tracts and bounding surfaces.

Nomenclature of Glacial Sequence Stratigraphy

Any attempt to compile a terminology that would recognize both water level and glacier input controls on glacial sequence stratigraphy would be unnecessarily complicated; hence, Brookfield and Martini (1999) have proposed the use of

allostratigraphy. This allows the construction of a descriptive and nongenetic sequence stratigraphy, wherein mappable sedimentary bodies are separated by discontinuities (Figure 2). The discontinuities or sequence boundaries can be erosional breaks caused by glacier readvance (Figure 2(b)) or the emptying of glacial lakes (e.g., Bennett et al., 2002). Glacial systems tracts can then be defined by the glacier advance-and-retreat cycle. Within the maximum glacier limits, depositional systems deposited during advance and retreat will be separated by a time-transgressive subglacial erosion surface or a wedge of subglacially deposited or deformed sediment (Berthelsen, 1978; Boulton, 1996a; Boulton, 1996b). Beyond the ice limit, complete depositional sequences may be preserved in the form of distal–proximal–distal successions, which, if they are

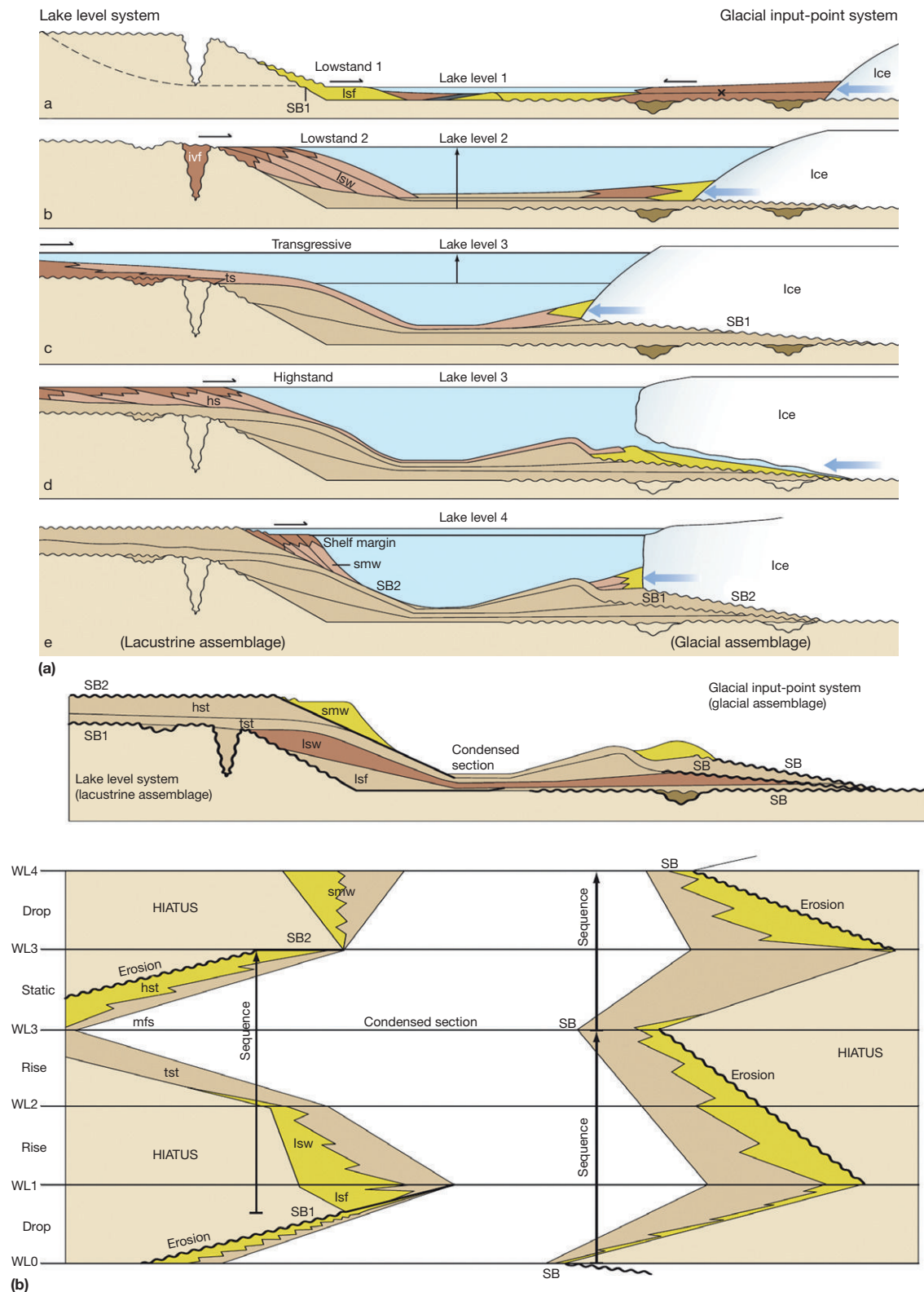


Figure 2 Idealized sequence stratigraphy explaining the correlation and complexity of facies assemblages due to changing lake levels and glacier input points: (a) localized control of successive units due to lake level change and ice front calving (lsf = lowstand fan; SB = sequence boundary; lsw = lowstand wedge; ts = transgressive surface; hs = highstand; smw = shelf margin wedge of shelf margin system tract; ivf = incised valley fill); (b) correlation of facies assemblages in the lake level system and the glacial input system showing systems tracts and bounding surfaces in space (A) and time (B) (WL = water level; hst = highstand systems tract; tst = transgressive systems tract; mfs = maximum flooding surface. Reproduced from Brookfield ME and Martini IP (1999) Facies architecture and sequence stratigraphy in glacially influenced basins: Basic problems and water-level/glacier input-point controls (with an example from the Quaternary of Ontario, Canada). *Sedimentary Geology* 123: 183–197.

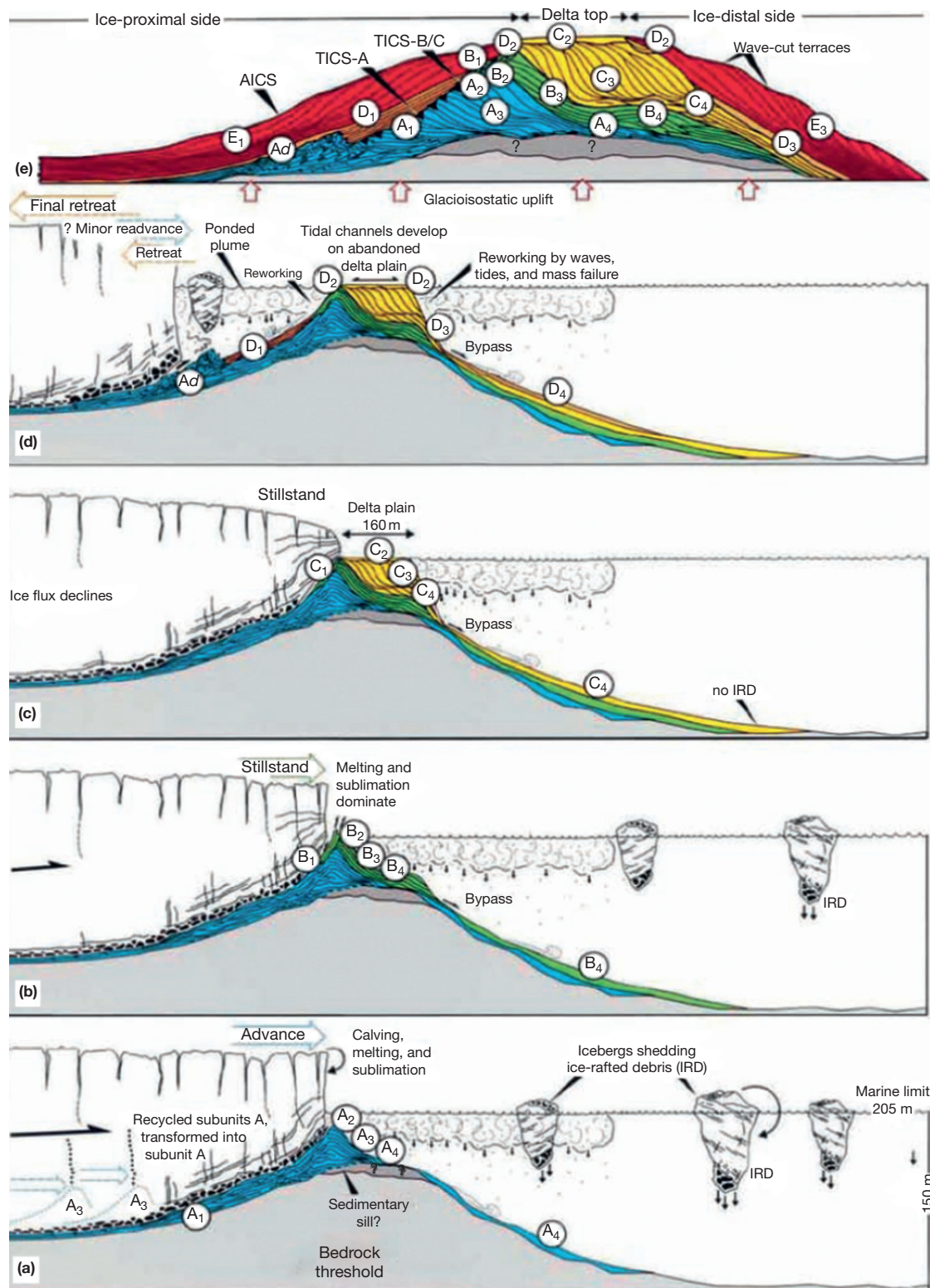


Figure 3 Reconstruction of the depositional history of the Mona moraine, southern Norway, showing the evolution of a grounding line fan and ice-contact delta and the deposition of letter-coded allostratigraphic units: (a) submarine ice-contact fan is formed on a bedrock threshold during the advance of a temperate glacier with a calving tidewater front; (b) glacier margin stabilizes and submarine fan aggrades to the sea surface; (c) ice-contact, Gilbert-type delta develops; (d) delta is abandoned due to a rapid ice front retreat by calving, and reworking by waves and slope failures begins; (e) moraine emerges due to regional glacioisostatic uplift and its slopes accumulate regressive foreshore deposits. AICS = apparent ice-contact surface. TICS = true ice-contact surface (indicated for units A–C). Ad = deformed unit A. Reproduced from [Lonne I \(1995\)](#) Sedimentary facies and depositional architecture of ice-contact glaciomarine systems. *Sedimentary Geology* 98: 13–43.

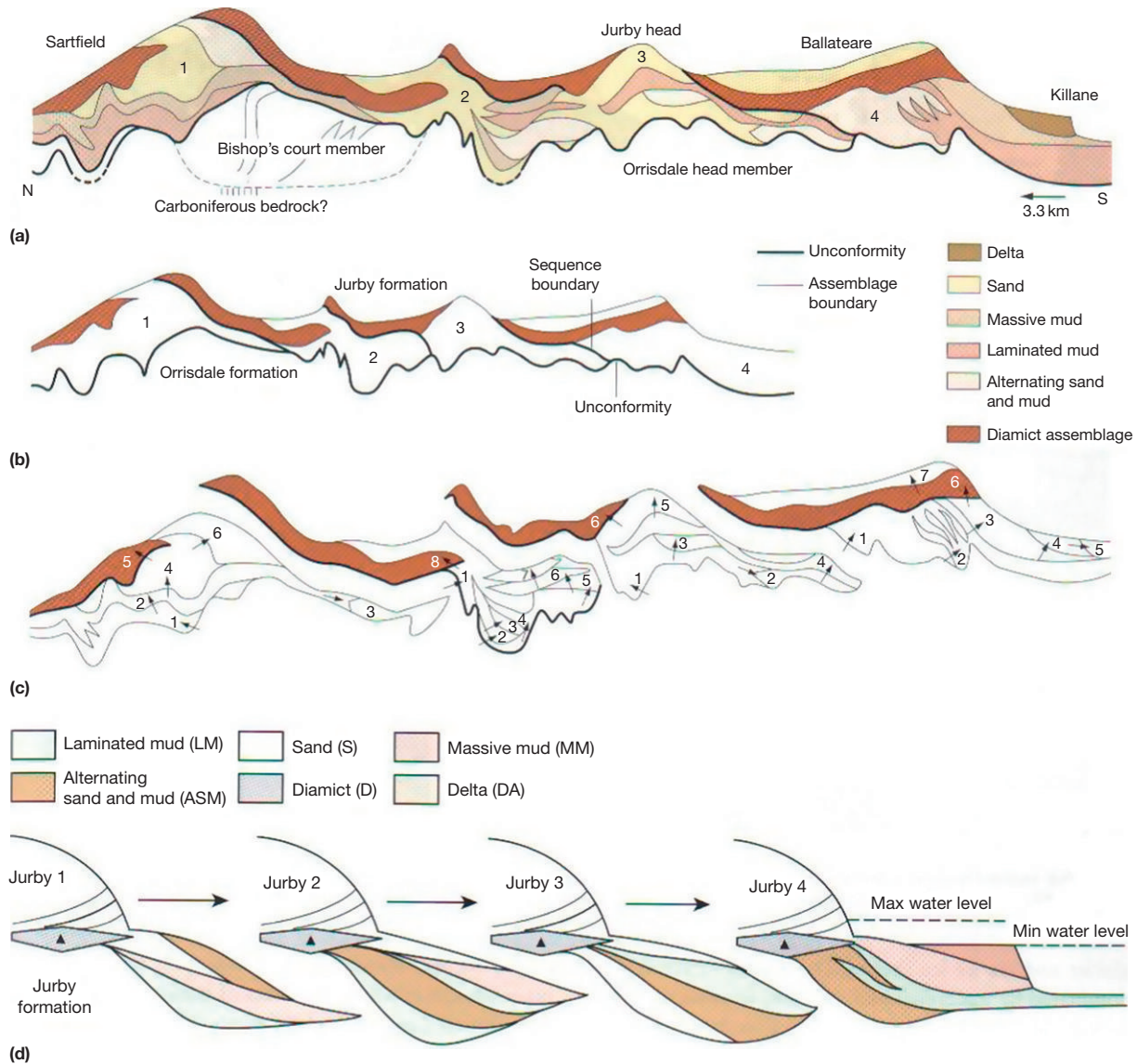


Figure 4 Sequence stratigraphy of glaciogenic deposits of the Jurby Formation, Isle of Man, including a diamicton assemblage interpreted as a subglacial till: (a) distribution of Jurby Formation lithofacies assemblages overlying the Bishop's Court and Orrisdale Head Members of the Orrisdale Formation, (b) summary of the four lithofacies assemblage packages, (c) exploded diagram of each package showing the order of deposition, and (d) simplified reconstruction of the marginal readvances recorded by the four packages. Reproduced from [Thomas GSP, Chiverrell RC, and Huddart D \(2004\)](#) Ice-marginal depositional responses to readvance episodes in the Late Devensian deglaciation of the Isle of Man. *Quaternary Science Reviews* 23: 85–106.

directly glacier fed, will record the migration of depositional systems in response to glacier advance and retreat ([Boulton, 1990](#); [Eyles & Eyles, 1992](#)). In glaciolacustrine settings, unconformities can arise through lake drainage, which may record short-lived events or more lengthy (i.e., interstadial, interglacial) episodes. The sequence stratigraphy/allostrostratigraphy approach to deciphering complex glaciogenic deposits is well illustrated by [Thomas et al. \(2004\)](#), who recognize four off-lapping packages of lithofacies assemblages, separated from one another by unconformities or sequence boundaries, in the Jurby Formation of the Isle of Man ([Figure 4](#)). These packages record four readvances during the overall recession of the British–Irish Ice Sheet.

See also: [Glacial Landforms, Sediments: Glaciogenic Lithofacies; Tills. Quaternary Stratigraphy: Morphostratigraphy/Allostrostratigraphy; Sequence Stratigraphy.](#)

References

- Andrews JT (1970) A Geomorphological Study of Postglacial Uplift with Particular Reference to arctic Canada. London: Institute of British Geographers. Special Publication 2, 156 p.
- Bednarski J (1988) The geomorphology of glaciomarine sediments in a high arctic fiord. *Geographie Physique et Quaternaire* 42: 65–74.

- Bennett MR, Huddart D, and Thomas GSP (2002) Facies architecture within a regional glaciolacustrine basin: Copper River, Alaska. *Quaternary Science Reviews* 21: 2237–2279.
- Berthelsen A (1978) The methodology of kineto-stratigraphy as applied to glacial geology. *Bulletin of the Geological Survey of Denmark* 27: 25–38.
- Boulton GS (1990) Sedimentary and sea level changes during glacial cycles and their control on glaciomarine facies architecture. In: Dowdeswell JA and Scourse JD (eds.) *Glaciomarine Environments: Processes and Sediments*, vol. 53, pp. 15–52. London: Geological Society Special Publication.
- Boulton GS (1996a) Theory of glacial erosion, transport and deposition as a consequence of subglacial sediment deformation. *Journal of Glaciology* 42: 43–62.
- Boulton GS (1996b) The origin of till sequences by subglacial sediment deformation beneath mid-latitude ice sheets. *Annals of Glaciology* 22: 75–84.
- Brookfield ME and Martini IP (1999) Facies architecture and sequence stratigraphy in glacially influenced basins: Basic problems and water-level/glacier input-point controls (with an example from the Quaternary of Ontario, Canada). *Sedimentary Geology* 123: 183–197.
- Cheel RJ and Rust BR (1982) Coarse grained facies of glaciomarine deposits near Ottawa, Canada. In: Davidson-Arnott R, Nickling W, and Fahey BD (eds.) *Research in Glaciofluvial and Glaciolacustrine Systems*, pp. 279–295. Norwich: Geobooks.
- Curry JR (1964) Transgressions and regressions. In: Miller RL (ed.) *Papers in Marine Geology*, pp. 175–203. New York: Macmillan.
- Dyke AS (1979) Glacial and sea level history of the southwestern Cumberland Peninsula, Baffin Island, NWT, Canada. *Arctic and Alpine Research* 11: 179–202.
- Eyles N and Eyles CH (1992) Glacial depositional systems. In: Walker RG and James NP (eds.) *Facies Models: Response to Sea-level Change*, pp. 73–100. Toronto: Geological Association of Canada.
- Helle SK (2004) Sequence stratigraphy in a marine moraine at the head of Hardangerfjorden, western Norway: Evidence for a high frequency relative sea level cycle. *Sedimentary Geology* 164: 251–281.
- Lønne I (1995) Sedimentary facies and depositional architecture of ice-contact glaciomarine systems. *Sedimentary Geology* 98: 13–43.
- Lønne I (2001) Dynamics of marine glacier termini read from moraine architecture. *Geology* 29: 199–202.
- Lønne I, Nemec W, Blikra LH, and Lauritsen T (2001) Sedimentary architecture and dynamic stratigraphy of a marine ice-contact system. *Journal of Sedimentary Research* 71: 922–943.
- Powell RD (1990) Glaciomarine processes at grounding-line fans and their growth to ice-contact deltas. In: Dowdeswell JA and Scourse JD (eds.) *Glaciomarine Environments: Processes and Sediments*, vol. 53, pp. 53–73. London: Geological Society Special Publication.
- Quinlan G and Beaumont C (1981) A comparison of observed and theoretical postglacial relative sea level in Atlantic Canada. *Canadian Journal of Earth Sciences* 18: 1146–1163.
- Rust BR and Romanelli R (1975) Late Quaternary subaqueous outwash deposits near Ottawa, Canada. In: Jopling AV and McDonald BC (eds.) *Glaciofluvial and Glaciolacustrine Sedimentation*, vol. 23, pp. 177–192. SEPM Special Publication.
- Thomas GSP, Chiverrell RC, and Huddart D (2004) Ice-marginal depositional responses to readvance episodes in the Late Devensian deglaciation of the Isle of Man. *Quaternary Science Reviews* 23: 85–106.
- van Wagoner JC, Posamentier HW, Mitchum RM, et al. (1988) An overview of the fundamentals of sequence stratigraphy and key definitions. In: Wilgus CK, et al. (ed.) *Sea Level Changes: An Integrated Approach* vol. 42, pp. 39–45. SEPM Special Publication.
- Walker RG (1992) Facies, facies models and modern stratigraphic concepts. In: Walker RG and James NP (eds.) *Facies Models: Response to Sea Level Change*, pp. 1–14. Toronto: Geological Association of Canada.
- Winsemann J, Aspöhn U, and Meyer T (2004) Sequence analysis of early Saalian glacial lake deposits (NW Germany): Evidence of local ice margin retreat and associated calving processes. *Sedimentary Geology* 165: 223–251.

GLACIAL LANDFORMS, TREE RINGS

Contents

Dendrogeomorphology

Dendroglaciology

Dendrogeomorphology

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Introduction

Geomorphology, a subfield of geosciences, deals with landforms and related processes ([Anderson and Anderson, 2010](#)), to reveal the complex interrelations between the origin of the landform and the dynamics of its alteration caused by such processes as erosion, mass movements, or tectonics. In this regard, dating and reconstructing past geomorphic processes, the determination of recurrence intervals, and their quantification are of special interest to reveal landform development and predict future changes in light of global change. The most accurate dating technique in geosciences is dendrochronology ([Schweingruber, 1993](#)), where the outermost ring of a living tree is used for dating. Furthermore, changes in environmental conditions as well as geomorphic processes cause changes in the growth and development of trees, which are datable to the year and even the season ([Gärtner and Heinrich, 2010](#)).

The response of trees to the influence of a changing environment has been a subject of research in botany and plant physiology since the middle of the nineteenth century. Research on trees at this time focused on descriptive analysis of the structure and variability of tree rings in an ecological context and the effects of external mechanical impacts were discussed ([Schweingruber, 1996](#)). This rather holistic approach was eclipsed when Andrew Ellicott Douglass invented the revolutionary crossdating technique for tree-ring research ([Parker et al., 1984](#)). For the first time, this method enabled the accurate dating of tree rings to the year and is regarded as the backbone of tree-ring research in all fields of its application ([Schweingruber, 1996](#)). In the following decades, crossdating formed the basis for the widespread use of tree rings, mainly in archaeology and climate reconstructions ([Fritts, 1976](#)). The first attempts of reconstructing glacial fluctuations were made by [Lawrence \(1950\)](#). Since then, new ecological aspects of tree-ring research have been integrated, especially in geosciences. Fundamental research studied the effect of environmental factors such as wind on the growth of trees ([Jacobs, 1954](#)). In the early stages, this research focused on growth reactions on a

macroscopic level, although single attempts were made on a microscopic level. In the 1960s, studies on erosion rates on slopes ([LaMarche, 1968](#)) and streams ([LaMarche, 1966](#)) were also conducted using annual growth rings in roots. In the following years, tree-ring analysis was established as a standard methodology in geosciences. [Alestalo \(1971\)](#) was the first to write a monograph on the use of tree rings in dating geomorphic processes and invented the term dendrogeomorphology ([Alestalo, 1971, p. 7](#)). Since then, in parallel to the development of computers, various types of dendroecological techniques have been developed in the subfields of geosciences. Within geomorphology, these techniques facilitated the dating of land surfaces in general ([Pierson, 2007](#)), the reconstruction of hydrological conditions ([Vrobesky and Yanosky, 1990](#)), the dating of volcanic eruptions ([Salzer and Hughes, 2007](#)), and the widespread effects of earthquakes ([Jacoby, 2010](#)). Additionally, special focus was placed on reconstructing the frequencies of geomorphic processes ([Filion and Gärtner, 2010](#); [Stoffel et al., 2010](#)) such as rock fall ([Perret et al., 2006](#)), landslides ([Corominas and Moya, 2010](#)), debris flows ([Stoffel et al., 2005](#)), or creeping slopes on permafrost ([Roer et al., 2007](#)). In this context, various species-specific reaction mechanisms in stems, roots, and branches due to mechanical stresses or environmental changes have been revealed ([Gärtner and Heinrich, 2010](#)). These reactions, which have been independently used, or combined with only small modifications over the last decades, are presented in this article.

Development of Tree Rings

The development of annual rings in trees is related to secondary growth ([Schweingruber, 2007](#)). Secondary growth is initiated by the cambium, a thin layer of undifferentiated, living cells producing xylem (wood) toward the inside and phloem (bark) toward the outside of the circumference of the stem ([Figure 1](#)). In a seasonal climate, tree growth is restricted to a certain vegetation period. This period is often limited by

minimum temperatures as a threshold for initiating growth in spring and leading to an end-of-cell development in late summer or autumn. Other potential limiting growth factors are day length and precipitation. The duration of this period always depends on the environmental parameters at the specific location (Schweingruber, 1996). In a seasonal climate, the resulting tree-ring increment formed during one single vegetation period can be ascribed unmistakably to the respective calendar year.

The growth rings of coniferous and deciduous trees can be separated into earlywood and latewood, produced within the first weeks of the vegetation period and toward the end of the vegetation period, respectively. This tissue differs between conifers and deciduous trees. Rings of conifers consist of big, thin-walled tracheids in earlywood and thick-walled, tangentially flattened tracheids in latewood. Rings of deciduous trees consist of big vessels for water conductance and a stabilizing tissue of tracheids and fibers (Gärtner and Heinrich, 2010). Trees most commonly used in dendrogeomorphology are

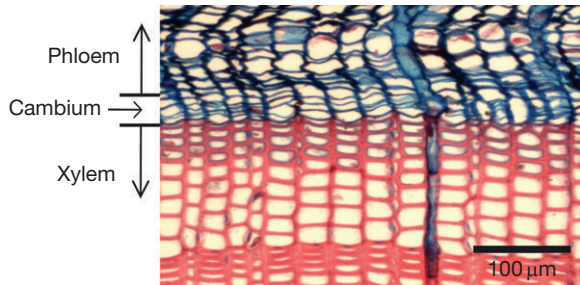


Figure 1 Microscopic thin section of wood showing the outermost ring of a European larch tree (*Larix decidua*) including the bark. The cambium is indicated between the wood (xylem) and the bark (phloem). The cambium is a layer of tissue between xylem and phloem in most vascular plants, producing xylem cells toward the inside and phloem cells toward the outside every year. As a consequence, this layer is responsible for the diameter growth of a tree.

conifers, which are the dominant tree species in alpine areas, where the potential of natural hazards is high. Another reason is the relatively simple structure of coniferous wood, which allows a straightforward analysis (Figure 2).

The width of the single rings, as well as the proportion of earlywood and latewood, depends on environmental factors (e.g., temperature and precipitation) dominantly controlling growth. These factors vary year by year, resulting in more or less distinct ring-width variations over time. If one environmental factor is the principal limiting factor, as is most often the case with temperature or precipitation, ring-width variations in most trees of a region show a similar pattern. This similarity enables the ring-width chronologies to be crossdated on the basis of the common trends of these variations and by comparing extremely narrow or wide rings obvious in the individual tree chronologies (Figure 3).

Reference Chronologies

In general, local environments change as a result of geomorphic processes. As soon as trees are in the range of these changes, or if they are directly influenced by the process, their growth adapts to the resulting conditions. With the onset of this adaptation, ring-width variations begin to differ from the common growth variations of trees not influenced by these changes. The presence of these so-called growth anomalies, as well as their initiation, needs to be accurately dated to a calendar year. For this, all samples taken for further analysis have to be crossdated (Cook and Kairiukstis, 1990) to check for such things as false or missing rings.

Depending on the dimension and spread of the geomorphic process, more or less local chronologies have to be built, representing common site-growth variations (reference chronologies) controlled by environmental factors such as temperature, precipitation, exposure, altitude, and soil properties. The sampling for reference material should be conducted as close as possible to the area where the event to be analyzed

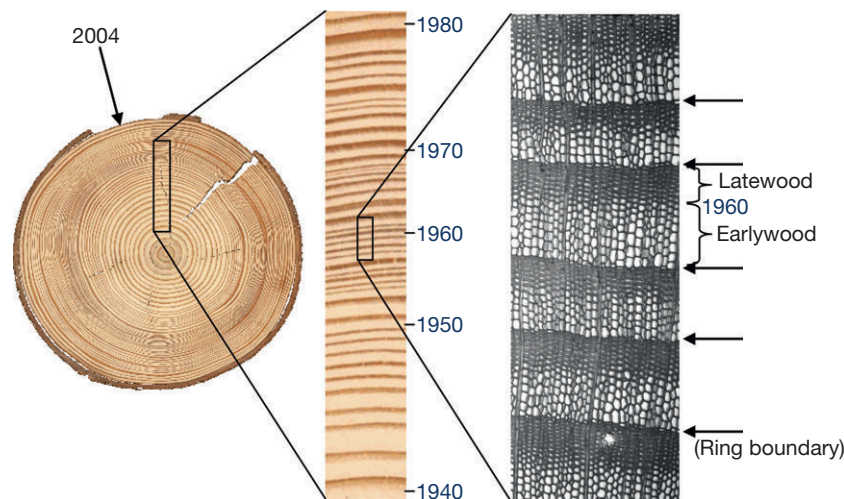


Figure 2 Macroscopic view (disc and core) of annual rings in conifers. The microscopic thin section on the right shows bigger, thin-walled cells of the earlywood and thick-walled, flattened cells of the latewood within an annual ring.

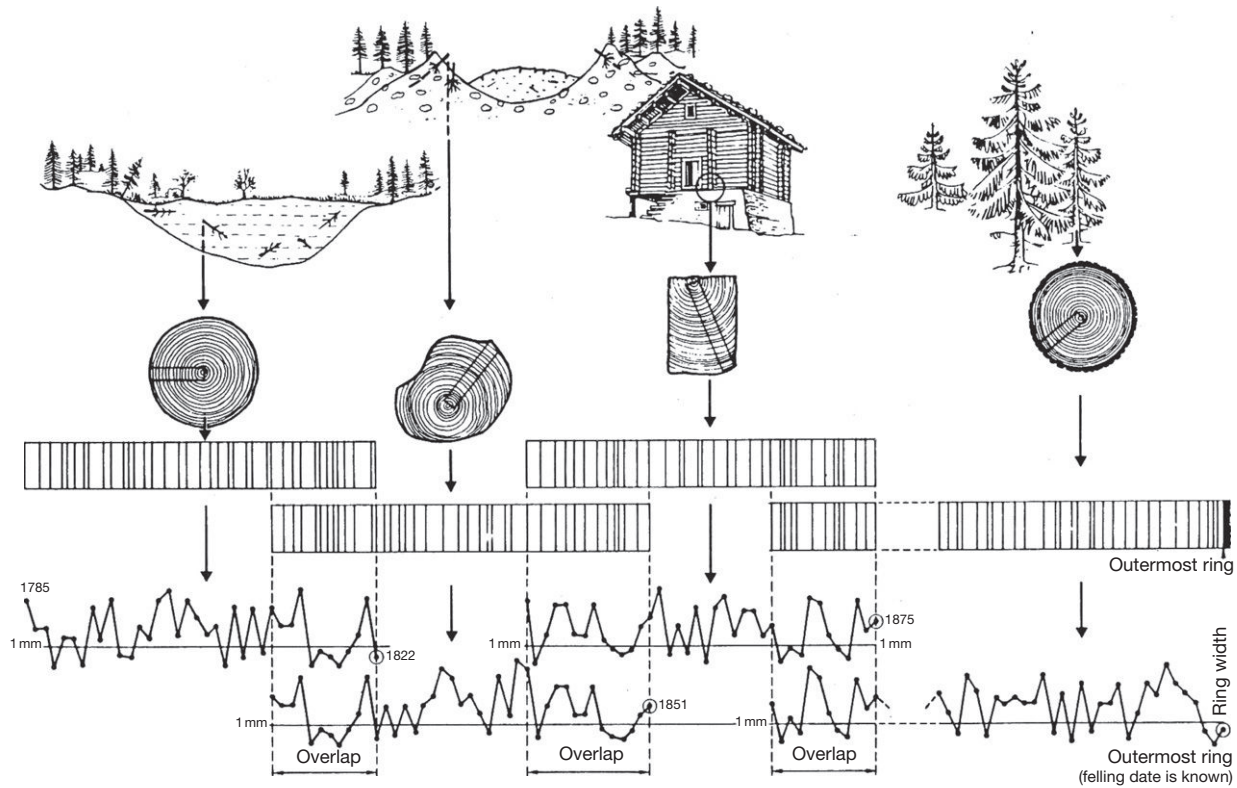


Figure 3 Principle of crossdating. Reproduced from Schweingruber FH (1983) *Der Jahrring: Standort, Methodik, Zeit und Klima in der Dendrochronologie*, p. 85. Berne: Paul Haupt.

happened, but definitely out of the range of its influence. This is especially important in alpine areas where environmental factors such as altitude, exposure, slope angle, precipitation, and even temperature vary over short distances.

The reference chronologies are used as a control to reveal all growth variations caused by the geomorphic process of interest. In this regard, the definition of pointer years and event years is of particular importance (Schweingruber et al., 1990). Pointer years represent extremely narrow or wide rings (compared to neighboring rings) which appear in the same calendar year in most trees of a site. They are caused by extremes of environmental factors (mostly temperature and/or precipitation), resulting in a negative (narrow ring) or positive (wide ring) effect on tree growth at the entire site. These pointer years, which should appear in the reference as well as in the disturbed chronologies, are then used to mark event years, which only occur in the chronologies disturbed by the process analyzed. Therefore, the definition of pointer years for a site or region is essential in dating geomorphic processes for two reasons: (1) they can aid during the crossdating procedure, and (2) they allow differentiation between site- and process-specific growth variations in single trees. Besides the advantages in the use of pointer years, there are also some disadvantages. It frequently happens that pointer years and process-related event years appear in the same calendar year. In these instances, the process cannot be accurately dated without further analysis of the specific structure of the respective rings (e.g., occurrence of reaction wood).

Process–Event–Response

Geomorphic processes exert various stresses on trees which alter 'normal' growth and lead to measurable growth disturbances, which can be observed macroscopically. These growth anomalies, their type, initiation, intensity, and duration permit the dating and reconstruction of such processes. Shroder (1980) coined the term 'process–event–response system' in dendrogeomorphology, that is, geomorphic processes cause specific events in the close vicinity of a tree, leading to measurable growth variations and changes in phenology as a response (Table 1). Since then, various authors have summarized possible process–event–response systems and have added new findings (Fantucci and McCord, 1995; Wiles et al., 1996). Most of these models are based on specific events with a list of the most common growth reactions likely to be found in disturbed trees.

In the following discussion, growth reactions and growth anomalies in trees are explained and related to possible processes, unlike those cited in the references mentioned earlier. This is done because a single process or event can cause various growth anomalies, depending on the way the tree is disturbed. Prior knowledge of possible growth variations as a likely response to specific geomorphic processes is of great value when decisions need to be made regarding sampling strategy, to guarantee further successful analysis. These decisions should only be reached after: (1) an accurate process

determination has been made, (2) geomorphic mapping has been done, (3) interpretation of the position of the disturbed tree relative to the related process has been made, and (4) screening is done for visible disturbances.

Growth Reactions in Trees

Growth Suppression

A more or less sudden decrease in ring width over a long period (growth suppression; **Figure 4**) is one of the most frequent features used in dendroecology to reconstruct environmental changes. Related to geomorphology, these suppressions can

Table 1 Event-related disturbances visible in tree phenology or the tree's direct environment and related growth responses

<i>Disturbance</i>	<i>Visible effect</i>	<i>Growth response</i>
Directed pressure	Tilting, bend of stem	Eccentricity (compression/tension wood)
Destabilizing ground	Tilting, bend of stem	Eccentricity (compression/tension wood), growth suppression, traumatic resin ducts
Nudation	Not visible (possibly remnants of trees in direct environment)	Growth release
Punctual impact	Scars (sometimes overgrown and for this not visible)	Callous tissue, growth suppression, traumatic resin ducts
Burial	Root collar not visible ^a	Growth suppression/growth release ^b adventitious roots
Erosion (A)	Roots partly exposed, living	Anatomical changes in annual rings of roots
Erosion (B)	Root tips exposed and mostly dead tilting of the stem, frequent	Growth suppression in stem (compare: pressure)
Water-table changes	Not visible	Growth release/suppression die-off

^aThe development of an adventitious root system may give the impression of an undisturbed root collar. In case of any doubts, the situation has to be checked by digging.

^bDepending on the material deposited around the stem.

be due to various processes leading to unfavorable conditions over a longer period, such as shearing processes affecting the root system caused by shallow landslides ([Schweingruber, 1996](#)). Mud and debris flows frequently deposit material around the lower portion of stems, which usually leads to suppressed or even missing rings, if growth is severely suppressed for 1 or more years. This is followed by a suppressed growth phase until the tree has adapted to the new situation. Avalanches also frequently lead to longer lasting growth suppressions due to such accidents as severed branches or a lifted and partly destroyed root system.

In the worst-case scenario, depending on the kind and intensity of the process, the affected trees are not able to adapt to the new conditions (e.g., deep burial of the trunk, complete shear of the root system, lowering of water table, or glacial advances) and they die. Dieback may occur abruptly after the event or start as a continuous process lasting for several years.

Succession

When destructive processes such as volcanic eruptions, avalanches, debris flows, or landslides cause a dieback or even total extinction of trees in the affected area, trees eventually recolonize these areas. This succession is always related to the recurrence intervals of the processes.

Commonly, trees start growing in the resulting open areas or on deposited material caused by singular events. Analysis of succession is also applied to glacial forefields and lateral moraines (**Figure 5**). In all cases, the age of the respective trees is of interest because process-related growth anomalies cannot be expected. To determine the age of a tree, samples have to be taken as low down the trunk as possible. In addition, the growth of the tree has to be analyzed to estimate the number of rings grown below the sample height ([Gutsell and Johnson, 2002](#)). Although age determination of these trees does not allow the precise dating of an event, it delivers a minimum age of depositional processes or the cessation of geomorphic disturbance in these areas.

Growth Release

In contrast to suppression, sudden growth releases occur as a result of environmental changes leading to more favorable

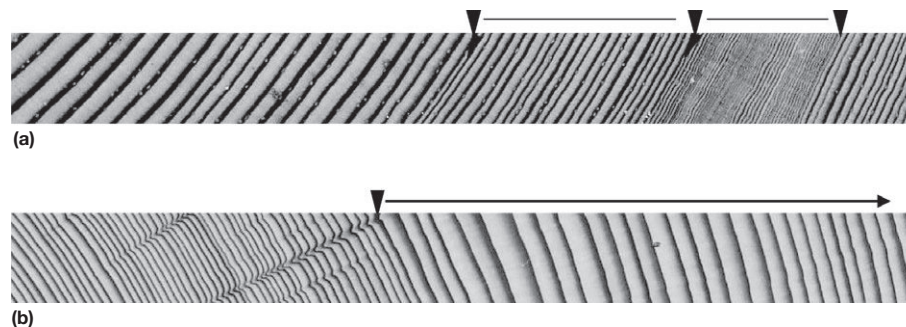


Figure 4 (a) Growth suppression. Black arrows indicate two phases: a weak suppression spanning 22 years, followed by a severe suppression lasting 64 years. (b) Growth release indicated by black arrows.



Figure 5 Glacier and forested lateral moraine (Karakorum, Pakistan). Photo: Thomas Reineke.



Figure 6 Left: Debris-flow material deposited in 2004, Val di Peijo, Italy. The bent trunks are the result of tilting caused by an avalanche several years before the debris-flow event. Right: Lateral levees along a debris-flow track burying and tilting trees, Turtmantal, Switzerland. Photo: Isabelle Gärtner-Roer.

conditions for the tree (compare [Figure 4](#)). The most common event causing a growth release is the elimination of neighboring trees ([Schweingruber, 1996](#)) due to various possible processes such as debris flows, avalanches, or rock fall, leading to a reduction of competitive stresses related to an increased availability of light, water, and nutrients. Thus, depositional processes (e.g., debris flows) do not always lead to suppression, but may also stimulate a growth release. This happens, depending on type and composition of the material deposited, at the former soil surface around the base of the tree, when the depositional process causes an enrichment of nutrients and a better availability of oxygen in the soil ([Strunk, 1991](#)).

Reaction Wood

Trees that are not perturbed by any mechanical disturbances tend to grow upright, that is, vertical against the force of gravity ([Chen et al., 1999](#)). In case of a mechanical impact or instability of the ground, trees can be tilted or even displaced and, as a result, moved out of their vertical position. To compensate for the resulting imbalance, reaction wood is formed on one side of the tree for as long as a new angle of growth is maintained. This one-sided growth results in the bend of the trunk ([Figure 6](#)). Dendrogeomorphological methods are commonly applied in alpine areas to date past processes using conifers as

the preferred indicator species. This is because conifers produce compression wood (Timell, 1986) which is recognizable without elaborate analysis, if intensely pronounced, also causing changes in ring width as well as anatomical changes. In contrast, deciduous trees have rarely been used to date and reconstruct past geomorphic processes (Gers et al., 2001; Heinrich and Gärtner, 2008a). This is due to the fact that their reaction wood (tension wood) is hardly detectable macroscopically (Côté and Day, 1965). The analysis of reaction wood is mostly restricted to trees because growth reactions to tilting or destabilization of the ground are concentrated in the trunk normally growing upright. Thus, the occurrence of reaction wood can be related to external stresses. Shrubs often do not have a single trunk, so when they are mechanically stressed, the resulting reaction wood is mostly not distinguishable from the commonly occurring reaction wood. Consequently, analyzing reaction wood in shrubs cannot be done without further analysis of other wood anatomical features. In most cases, this does not lead to the desired result.

Compression Wood

By definition, compression wood is characterized by round-shaped, thick-walled tracheid cells showing intercellular spaces (Timell, 1986). Further details, such as a missing S3 layer in the cell wall and the fact that the S2 layer is characterized by helical cavities (Timell, 1973), are not visible macroscopically. Thus, this kind of analysis requires the preparation of thin sections of wood and high-resolution microscopic analyses.

Although the definition discussed earlier is commonly used, it has been repeatedly shown that compression wood

can assume various forms, depending on the species in question, physiological status of the plants (season and water content), parts of the plant affected (trunk, branches, or roots), type and duration of impact, and climatic conditions (temperature and humidity). Gärtner et al. (2001) clearly illustrated that larch (*Larix decidua* Mill.) roots form visible compression wood but lack the usual large intercellular spaces. Compression wood can be classified in up to five severity grades (Yumoto et al., 1983), each class representing a combination of distinct characters with milder forms exhibiting a lack of some compression wood characteristics such as intercellular spaces. Such a classification can be used to calibrate wood anatomical reactions to growth stresses of various forms and strengths. Furthermore, Kennedy and Farrar (1965) found in their tilting experiments that the responses of the cells varied depending on the stage of cell maturation. For example, tilting encouraged cell division during the earliest period of cell development, while at later stages the cells showed altered shapes, wall thickness, or different degrees of lignification. Thus, typical compression wood tracheids were formed only when the differentiating cells were not induced before the growth stress occurred (Figure 7).

More recent experiments focusing on the onset of compression wood resulting from stem inclination in the dormant season also showed variability in the characteristics of the first cells produced after the event. According to this, the first cells of *Larix decidua* Mill. and *Picea abies* were macroscopically characterized as common earlywood tracheids (Heinrich and Gärtner, 2008b). Wood anatomical analysis of these cells revealed that they showed certain characteristics of compression wood cells, namely, helical cavities in the S2 layer and a

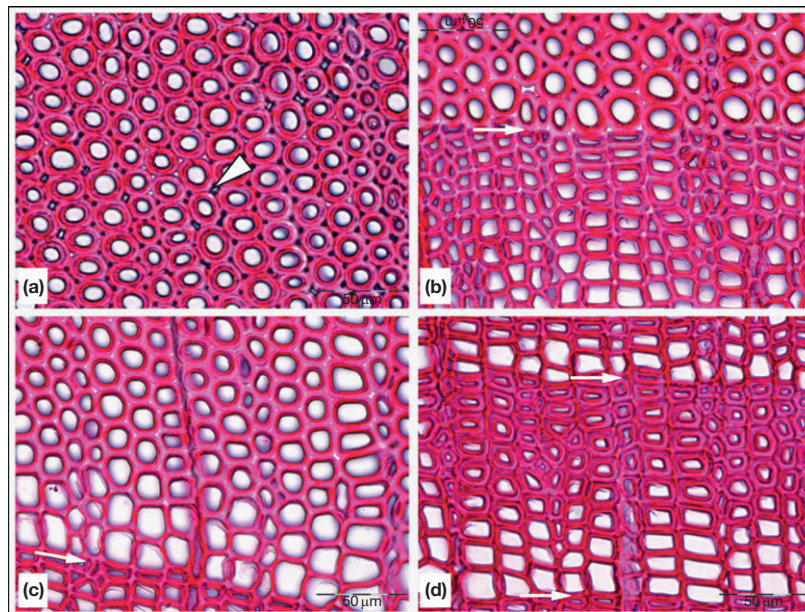


Figure 7 Development of compression wood in European larch (*Larix decidua* Mill.). White arrows indicate ring boundaries ((b)–(d)). (a) Typical structure of compression wood cells in conifers. Tracheids are rounded and thick walled. Between the cells there are intercellular spaces (white triangle). (b) Compression wood cells resulting from trunk tilting after the end of the vegetation period. The first cells of the new ring are round and thick walled (compression wood). (c) Trunk tilting after the beginning of the vegetation period. First earlywood cells of the ring show the common rectangular, rather thin-walled shape. Then there is a change toward more thick-walled, rounded cells showing intercellular spaces. (d) Trunk tilting in early summer. Only the latewood cells show rounded and thick-walled compression wood cells.

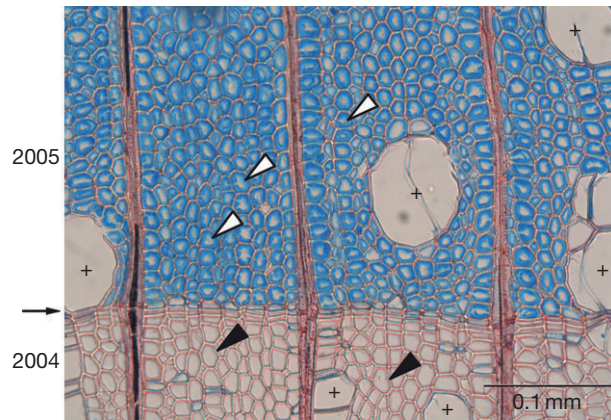


Figure 8 Microscopic thin section showing tension wood formation in *Fagus sylvatica* caused by an artificial treatment (tilting of the trunk) before the beginning of the vegetation period in 2005 (black arrow indicating ring boundary 2004/2005). The ring formed in 2004 shows the highly lignified cell walls of the fiber tissue (black triangles; red color = safranin). In 2005, lignification of fiber cell walls is reduced and the cells are filled with nonlignified gelatinous fibers (white triangles; blue color = astra blue) exerting a tensile stress to the stem. + = water-conducting tissue (vessels).

missing S3 layer of the cell wall. These characteristics are not definable solely by macroscopic analysis. Experiments such as this and subsequent wood anatomical analysis can help to date such destabilizing events to a higher precision than pure macroscopic analysis.

Tension Wood

In contrast to compression wood in gymnosperms, tension wood in angiosperms is only distinguishable from normal wood by a silvery sheen macroscopically visible on trunk cross sections (Barbacci et al., 2008; Côté and Day, 1965). Thus, tension wood can be easily missed during standard tree-ring analysis, especially when working on 5 mm core samples only. Tension wood of many species is characterized by the existence of a gelatinous layer (G-layer) attached to the secondary wall of the fiber cells, sometimes filling their entire lumen (Coutand et al., 2004).

The G-layer in tension wood is usually identified in microscopic examinations in combination with specific dyes such as astra blue and safranin staining cellulose blue and lignin red, respectively (Figure 8). Many angiosperms, however, do not show a G-layer in their tension wood. Rather, they show structural modifications of the secondary wall of the fibers (Qiu et al., 2008). Furthermore, these changes of the secondary wall frequently result in the development of dense fiber tissue (increased cell-wall thickness) and are also accompanied by structural changes to the entire anatomical structure of the ring. Tilting experiments performed on European alder (*Alnus glutinosa*) and beech (*Fagus sylvatica*) resulted in a reduced number of vessels and decreased vessel size with a G-layer present in alder and absent in beech (Heinrich and Gärtner, 2008a).

The characteristics of tension wood are generally highly variable, that is, species dependent and related to the physiological

status of the plants, the parts of the plant affected, the type and duration of the mechanical impact, and climatic conditions (Wardrop, 1964). The difficulty in detecting and identifying the different forms of tension wood without microscopic examination has resulted in a lack of studies using angiosperms for the reconstruction of geomorphic processes. Nevertheless, the resulting eccentricity is measurable and can also be determined statistically (e.g., Weiss, 1991). In addition, the direction of the growth of the reaction wood can be used to reconstruct the direction of the related process (Figure 9). The occurrence of reaction wood, mostly combined with other growth anomalies, has been used in reconstructing debris flows (Gärtner et al., 2003), avalanches (Butler and Sawyer, 2008; Herbertson and Jenkins, 2003), landslides (Fantucci and McCord, 1995), and even glacier fluctuations (Fleisher et al., 2006).

Wounding (Scars)

Scars result from wounds to the cambial layer of a trunk, branch, or root of woody plants. In the circumference of the destroyed cambial zone, living, undisturbed cambium cells start producing callous tissue, which is composed of irregular cells. These cells are only produced in the vicinity of the wound and they begin to expand the growth layer into the wounded zone. In cross section, the border zone of the scar is marked by this callous tissue layer, coincident with increased growth, resulting in a subsequent overgrowing of the wound by new growth layers (Figure 10).

The cambium cells of the callous zone return to normal shape within a short time, but continue to overgrow the wound until the wound is closed, that is, when the cambial layers of both sides come in contact again, forming a continuous ring. The duration of this overgrowing process depends on the size of the scar and the productivity of the tree, varying from only a few years to centuries. The occurrence of callous tissue has been demonstrated to be an immediate reaction of trees to wounding in many species (Schweingruber, 1996). Consequently, the analysis of the onset of the first callous cells facilitates an intra-annual dating of the wound by identifying the position of the first irregular cells in the vicinity of the wound. The sample material needs to include disks, wedges from the scarred region, or core samples using the technique described by Arno and Barrett (1988) to date the initiation of scars.

The features most commonly used in dendrochronology to reconstruct environmental impacts other than geomorphic processes are fire scars. This specific type of scar occurs when the bark and the cambium below are burned, resulting in an open or sometimes hidden wound. In contrast to scars caused by mechanical impacts, fire scars show a more regular shape and contain a charred region, that is, the burned remnants of cells already developed before the fire occurred. Like other scars, fire scars can be dated to an intra-annual resolution, depending on the position where the charred region occurs within the ring (Everett, 2008).

Resin Ducts

The occurrence of resin in conifers is related to their defense system to protect the plant against mechanical wounding (Fahn et al., 1979) as well as against herbivores and pathogens

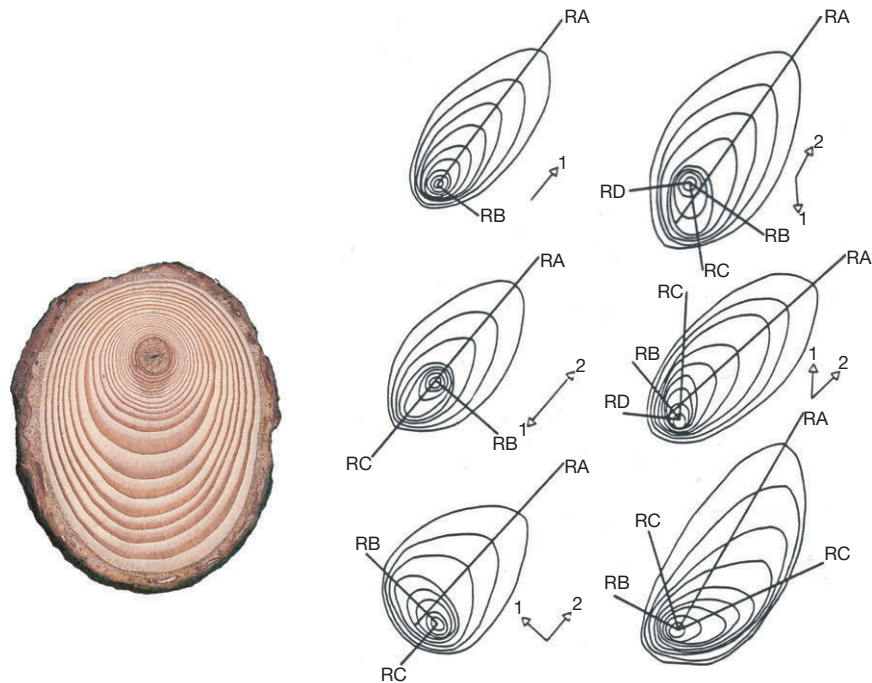


Figure 9 Disc of a coniferous tree showing eccentricity in the lower part of the cross section. Directions of compression wood are related to the direction of inclination. Using this, the direction of the process causing these reactions can be reconstructed. Arrows indicate direction of tilting; straight lines indicate optimal positions for core sampling. RA, RB, RC and RD indicate the radii needed to be analyzed to reconstruct the respective events precisely. Reproduced from Braam RR, Weiss EEJ, and Burrough PA (1987) Dendrogeomorphological analysis of mass movement. A technical note on the research methods. *Catena* 14: 585–589.

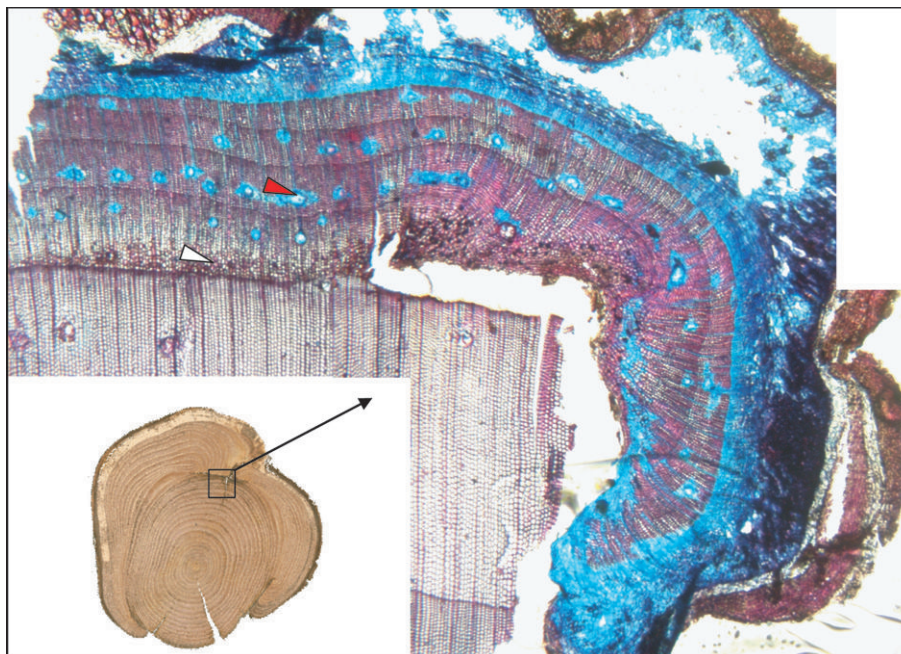


Figure 10 Stem disc indicating open and overgrown (closed) scars. Microscopic thin section: Cambium cells adjacent to the impact zone start producing irregular cells (callous tissue, white triangle), which tend to grow into the new wound followed by the succeeding process of ring growth to close the scar. Tangential rows of resin ducts are visible in the rings as a reaction to the impact process (blue color surrounding resin ducts indicates living epithelial cells producing resin; compare with [Figure 11](#)).

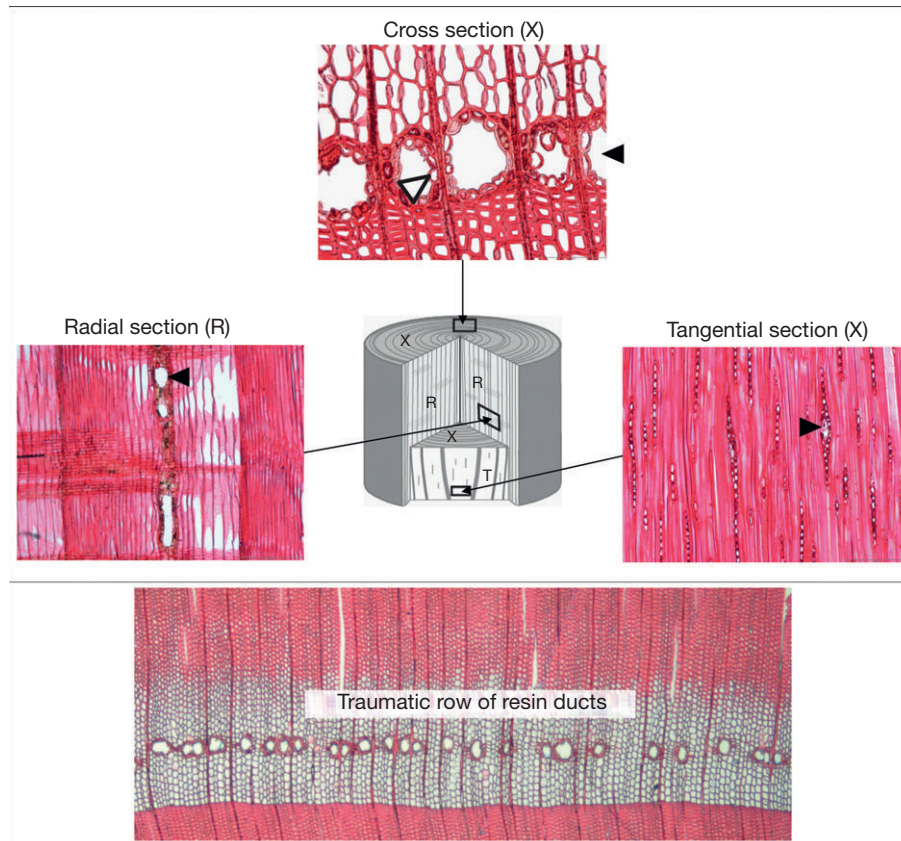


Figure 11 Upper part: Resin ducts and their three-dimensional distribution within a conifer trunk. The resin canals (black triangles) are surrounded by small epithelial cells (white triangle) producing the resin. This principle is true for all resin ducts (horizontal and vertical). Lower part: Microscopic thin section of a tangential row of resin ducts within the earlywood zone of an annual ring in a spruce tree.

(Nagy et al., 2005). Most conifer species produce single resin ducts within the xylem of their trunks, roots, or branches as part of this defense system, even without any stimulus from an environmental disturbance. The resin is produced in specialized cells (epithelial cells) framing the resin canals (Figure 11). The so-called resin ducts spread horizontally and vertically within the xylem, forming a three-dimensional network. In case of wounding, these ducts are concentrated in the area of the wound (Gärtner and Heinrich, 2009) forming rows of ducts directly in contact with each other, the so-called traumatic rows of resin ducts (TDs; Figure 11). Although their occurrence has sometimes been used to date geomorphic processes (Stoffel, 2008), most tree-ring related publications treat TDs as indicators for mechanical, insect-related, or other environmental disturbances, frequently describing them as a delayed response to environmental stress (Schweingruber, 2001, p. 198), not as an immediate reaction.

This supposition is mostly based on the fact that only sparse information exists on the spatial distribution of TDs within the disturbed annual ring of a mature conifer (Nagy et al., 2000). Most recent experiments have shown that TDs do not occur as an immediate reaction in Norway spruce (*P. abies*) when wounded in the dormant season. Furthermore, in European larch (*L. decidua* Mill.), they occur only as an immediate reaction in the vicinity of the wound, spreading toward the

latewood zone with increasing distance from the wound (Gärtner and Heinrich, 2009).

Various experiments conducted in recent decades have focused on the direct effect of insects or fungi and on the effect of hormones such as jasmonates, activating genes involved in the defense system of the plant (Martin et al., 2002). Since there is still no definite proof that TDs form immediately in the tissue of an annual ring after mechanical wounding in naturally grown conifers, without any immediate pathogenic or insect-related impact involved, other features such as the occurrence of callous tissue or reaction wood also need to be taken into account when dating geomorphic disturbances.

Adventitious Roots

Most tree species, especially flat-rooting conifers, are able to produce adventitious root layers when the lower part of the trunk is buried in accumulated material (Strunk, 1991). These accumulations are mostly related to flood, sheet-, mud-, or debris-flow events, and the burial of the base of the tree is easy to detect in the field, shortly after the event. Depending on the composition of the material, burial results in a reduced availability of oxygen to the root system. Therefore, buried trees initiate new roots, which, depending on the height and density of the accumulation, take over the function of the original root layer. In areas where tree bases are frequently

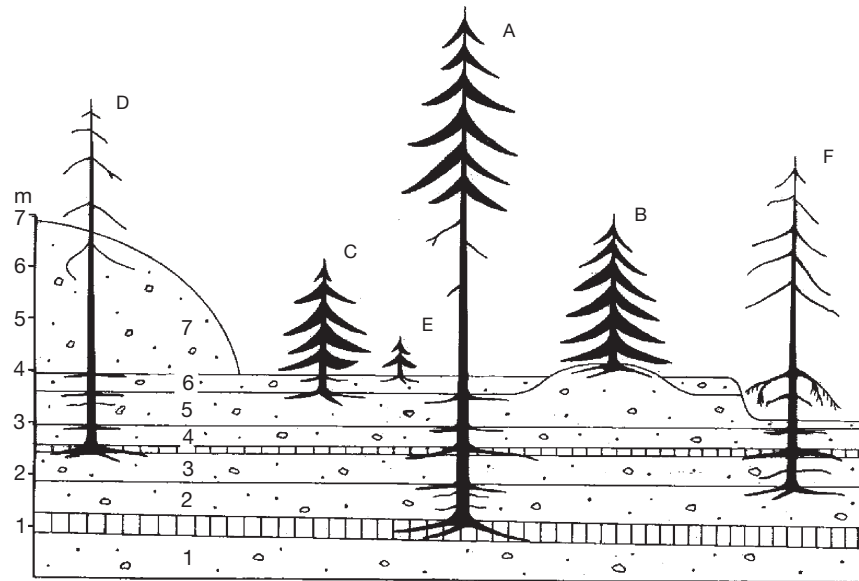


Figure 12 Schematic view of seven debris-flow layers influencing tree growth. Trees A, C, D, and F developed adventitious root layers after each burial event. Each layer can be dated by determining the year when the first root of a layer was initiated. Tree D died back as a result of the height of material covering the trunk. Tree F died after an erosional event exposed the uppermost adventitious root layer. Trees B and E grew up on the surface after the material was deposited. Reproduced from Strunk H (1995) *Dendrogeomorphologische Methoden zur Ermittlung der Murfrequenz und Beispiele ihrer Anwendung. Theorie und Forschung*, Band 317; *Geographie*, Band 1, p. 68. Roderer, Regensburg.

buried by geomorphic events, multiple adventitious root layers are grown (Figure 12). Spruce trees particularly tend to produce adventitious roots immediately after accumulation events happen. In other tree species, the initiation may be delayed for up to 5 years (Strunk, 1991). Consequently, at least the minimum age of depositional events can be determined. However, dating of the geomorphic event is not achieved by counting the rings of the roots at the transition to the stem, but rather by identifying the tree ring in the trunk containing the initiation of the adventitious root.

Growth Rings in Roots

The potential of analyzing radial root growth for dating purposes was already known in the first half of the last century and has been reconsidered in detail in the last decade (Krause and Morin, 1999). LaMarche (1968) determined the age of exposed roots by ring counting and then correlating the resulting age with the distance from the center of the exposed root to the recent soil surface to finally calculated erosion rates. Danzer (1996) revived the method of measuring the minimum depth of soil erosion from the degree of root exposure on datable trees. Carrara and Carroll (1979) added the idea of dating scars of exposed roots and proposed that this should be considered the first year of exposure. Despite all uncertainties, especially concerning the lack of a real cross-dating, this method is useful in areas where no data on soil erosion is available.

Anatomical Changes in Roots

It has been shown that it is possible to determine the age of adventitious roots and consequently the time of the trunk

burial caused by such things as debris-flow events. Furthermore, the age of exposed roots can be used to quantify soil erosion. Fayle (1968) studied anatomical features of exposed roots and showed the high variability of cell-structure changes due to changing environments (in this instance, contact with the atmosphere). Recent research took up this approach by focusing on the suitability of roots to date the time of exposure. Detailed analysis of structural variations in earlywood and latewood cells of coniferous trees allowed the accurate dating of the first year of exposure and also allowed differentiation between processes causing sudden or gradual exposure of coniferous roots (Gärtner et al., 2001). Specific anatomical features can be related to a sudden exposure of a root caused by an erosive event. This is an immediate 50% reduction of the size of earlywood cells and an increased number of thick-walled cells in latewood (Figure 13), resulting in a more stem-like shape of the ring. In the case of continuous denudation, cell-wall thickness and number of cells in latewood is increased first (Gärtner et al., 2001). Although the causal factors remain uncertain, these changes are not randomly distributed in the growth rings. The ability to determine the first year of exposure (Gärtner, 2007) facilitates a more detailed reconstruction of such details as sheet-erosion rates (Bodoque et al., 2005) and gully erosion (Malik, 2008).

Future Trends

The potential effects of global warming and the resulting environmental changes on the frequency and magnitude of mass movements has become an immediate problem in many mountainous regions of the world. To help address these

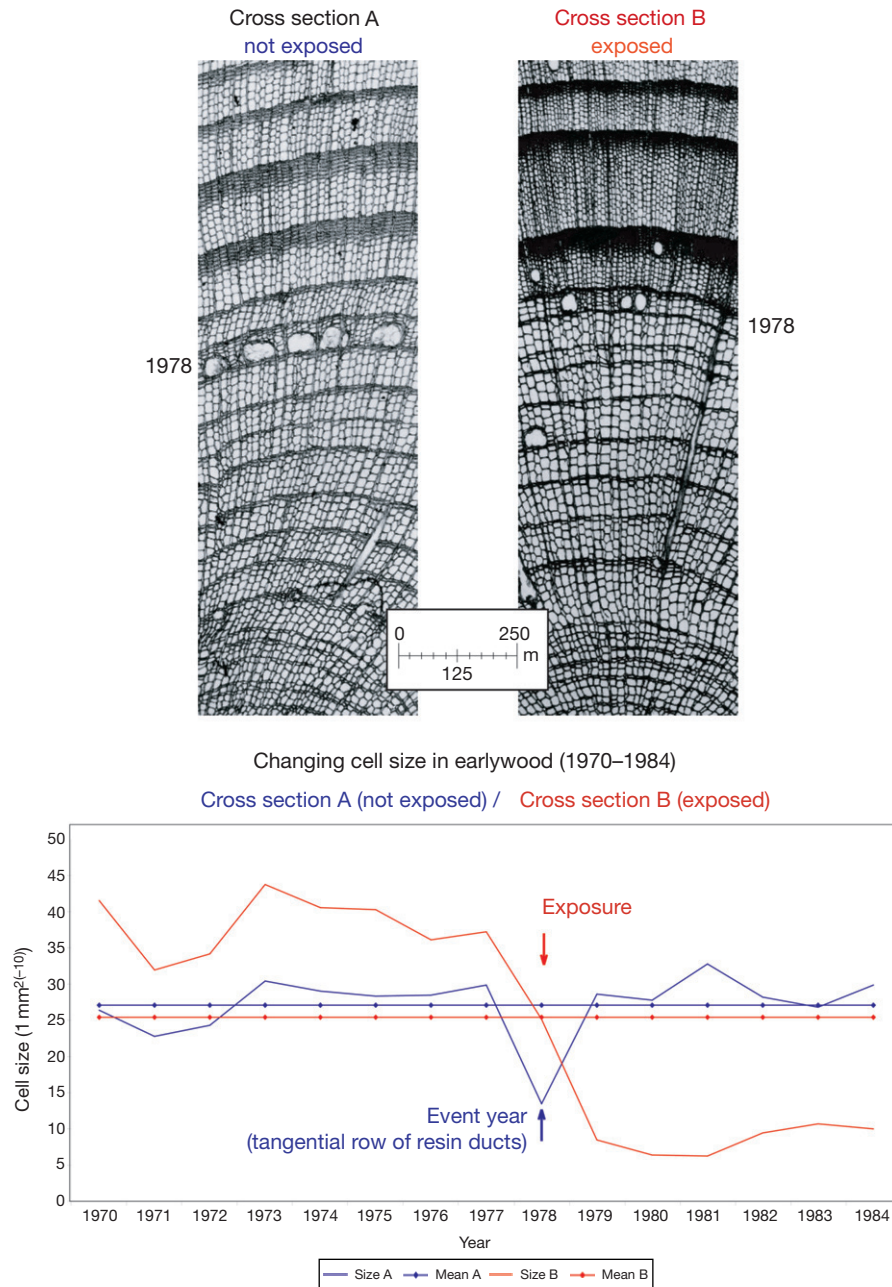


Figure 13 Microscopic thin sections of an unexposed (a) and an exposed (b) part of the same *Larix decidua* root due to a debris-flow event in 1978. Before 1978, both show the common cell structure of a buried root. After 1978: (a) Increasing number of earlywood tracheids but no significant change in latewood; common change in root structure near the surface. (b) Reduction of cell size in earlywood (compare cell size measurements, lower diagram), increasing thickness of cell walls in latewood, increasing number of cells. 1978: Tangential row of resin ducts.

issues, there is a need to develop new methods to record and quantify past and recent processes.

In recent years, microscopic examinations of wood samples have revealed a wide range of structural reactions of trees in response to different growth stresses. Wood anatomical investigations of the formation of compression wood, as well as tension wood, have illustrated the high variability of reaction-wood characteristics in proportion to different mechanical stresses. These results indicate the need for more

research into the calibration and verification of dendroecological methods used in geomorphology to ensure precise dating and reconstruction of different geomorphic events (Gärtner and Heinrich, 2010; Heinrich et al., 2007). There is also a need to combine millennia-long chronologies, mostly established for climate reconstructions, with dendrogeomorphic reconstructions to better understand environmental changes during the Holocene (Luckman, 2000; Solomina, 2002).

See also: Dendrochronology; Dendroclimatology. **Glacial Climates:** Volcanic and Solar Forcing. **Glacial Landforms, Tree Rings:** Dendroglaciology. **Plant Macrofossil Methods and Studies:** Dendroarchaeology.

References

- Alestalo J (1971) Dendrochronological interpretation of geomorphic processes. *Fennia* 105: 1–140.
- Anderson RS and Anderson SP (2010) *Geomorphology: The Mechanics and Chemistry of Landscapes*. Cambridge: Cambridge University Press.
- Arno SF and Barrett SW (1988) Increment borer methods for determining fire history in coniferous forests. General Technical Report INT-244. USDA Forest Service, Intermountain Research Station, p. 16.
- Barbacci A, Constant Th, Farré E, Harroué M, and Nepveu G (2008) Shiny beech wood is confirmed as an indicator of tension wood. *IAWA Journal* 29: 35–46.
- Bodoque JM, Díez-Herrero A, Martín-Duque JF, et al. (2005) Sheet erosion rates determined by using dendrogeomorphological analysis of exposed tree roots: Two examples from Central Spain. *Catena* 64: 81–102.
- Butler DR and Sawyer CF (2008) Dendrogeomorphology and high-magnitude snow avalanches: A review and case study. *Natural Hazards and Earth System Sciences* 8: 303–309.
- Carrara PE and Carroll TR (1979) The determination of erosion rates from exposed tree roots in the Piceance Basin, Colorado. *Earth Surface Processes* 4: 307–317.
- Chen R, Rosen E, and Masson PH (1999) Gravitropism in higher plants. *Plant Physiology* 120: 343–350.
- Cook ER and Kairiukstis LA (eds.) (1990) *Methods of Dendrochronology: Applications in the Environmental Science*. Dordrecht: Kluwer.
- Corominas J and Moya J (2010) Contribution of dendrochronology to the determination of magnitude–frequency relationships for landslides. *Geomorphology* 124: 137–149.
- Côté WA and Day AC (1965) Anatomy and ultrastructure of reaction wood. In: Côté WA (ed.) *Cellular Ultrastructure of Woody Plants*, pp. 391–418. New York: Syracuse University Press.
- Coutand C, Jeronimidis G, Chanson B, and Loup C (2004) Comparison of mechanical properties of tension and opposite wood in Populus. *Wood Science and Technology* 38: 11–24.
- Danzer SR (1996) Rates of slope erosion determined from exposed roots of ponderosa pine at Rose Canyon Lake, Arizona. In: Dean J, Meko DM, and Sewnam TW (eds.) *Tree Rings, Environment, and Humanity*. Radiocarbon 1996, Proceedings of the International Conference, Tucson, Arizona, 17–21 May 1994, pp. 671–678. Tucson, AZ: Department of Geosciences, The University of Arizona.
- Everett RG (2008) Dendrochronology-based fire history of mixed-conifer forests in the San Jacinto Mountains, California. *Forest Ecology and Management* 256: 1805–1814.
- Fahn A, Werker E, and Ben-Tzur P (1979) Seasonal effects of wounding and growth substances on development of traumatic resin ducts in Cedrus Libani. *New Phytologist* 82: 537–544.
- Fantucci R and McCord A (1995) Reconstruction of landslide dynamic with dendrochronological methods. *Dendrochronologia* 13: 1–22.
- Farley DFC (1968) Radial growth in tree roots – Distribution, timing, anatomy. Technical Report 9. Faculty of Forestry, University of Toronto, Toronto.
- Filion L and Gärtner H (2010) Dendrogeomorphologie. In: Payette S and Filion L (eds.) *La Dendroécologie: Principes, Méthodes et Applications*, pp. 537–572. Québec: Les Presses de l'Université Laval.
- Fleisher PJ, Lachniet MS, Muller EH, and Bailey PK (2006) Subglacial deformation of trees within overridden foreland strata, Bering Glacier, Alaska. *Geomorphology* 75: 201–211.
- Fritts HC (1976) *Tree Rings and Climate*. London: Academic Press.
- Gärtner H (2007) Tree roots – Methodological review and new development in dating and quantifying erosive processes. *Geomorphology* 86: 243–251.
- Gärtner H and Heinrich I (2009) The formation of traumatic rows of resin ducts in *Larix decidua* and *Picea abies* (Pinaceae) as a result of wounding experiments in the dormant season. *IAWA Journal* 30: 199–215.
- Gärtner H and Heinrich I (2010) Anatomie des cernes annuels chez les arbres et les arbustes. In: Payette S and Filion L (eds.) *La Dendroécologie : Principes, Méthodes et Applications*, pp. 33–60. Québec: Les Presses de l'Université Laval.
- Gärtner H, Schweingruber FH, and Dikau R (2001) Determination of erosion rates by analysing structural changes in the growth pattern of exposed roots. *Dendrochronologia* 19: 81–91.
- Gärtner H, Stoffel M, Lièvre I, and Monbaron M (2003) Tree ring analyses and detailed geomorphological mapping on a forested debris flow cone in Switzerland. In: Rickenmann D and Chen Ch (eds.) *Debris Flow Hazards Mitigation: Mechanics, Prediction, and Assessment*, vol. 1, pp. 207–217. Rotterdam: Millpress.
- Gers E, Florin N, Gärtner H, Glade T, Dikau R, and Schweingruber FH (2001) Application of shrubs for dendrogeomorphological analysis to reconstruct spatial and temporal landslide movement patterns – A preliminary study. In: Dikau R and Schmidt KH (eds.) *Mass Movements in South, West and Central Germany*. *Zeitschrift für Geomorphologie* (supplement 125): 163–175.
- Gutsell SL and Johnson EA (2002) Accurately ageing trees and examining their height growth rates: Implications for interpreting forest dynamics. *Journal of Ecology* 90: 153–166.
- Heinrich I and Gärtner H (2008a) Variations in tension wood of two broadleaved tree species in response to different mechanical treatments: Implications for dendrochronology and mass movement studies. *International Journal of Plant Sciences* 169: 928–936.
- Heinrich I and Gärtner H (2008b) Rekonstruktion von Naturereignissen mithilfe der Holzanatomie. *Schweizerische Zeitschrift für Forstwesen* 159: 58–65.
- Heinrich I, Gärtner H, and Monbaron M (2007) Tension wood formed in Fagus sylvatica and Alnus glutinosa after simulated mass movement events. *IAWA Journal* 28: 39–48.
- Herbertson EG and Jenkins MJ (2003) Historic climate factors associated with major avalanche years on the Wasatch Plateau, Utah. *Cold Regions Science and Technology* 37: 315–332.
- Jacobs MR (1954) The effect of wind sway on the form and development of Pinus radiata D. Don. *Australian Journal of Botany* 2: 35–51.
- Jacoby GC (2010) Application of tree-ring analysis to paleoseismology. In: Stoffel M, Bollschweiler M, Butler DR, and Luckman BH (eds.) *Tree Rings and Natural Hazards: A State-of-the-Art*, pp. 399–416. Berlin: Springer.
- Kennedy RW and Farrar JL (1965) Tracheid development in tilted seedlings. In: Côté WA (ed.) *Cellular Ultrastructure of Woody Plants*, pp. 419–453. New York: Syracuse University Press.
- Krause C and Morin H (1999) Root growth and absent rings in mature black spruce and balsam fir, Quebec, Canada. *Dendrochronologia* 16/17: 21–35.
- LaMarche VC (1966) An 800-year history of stream erosion as indicated by botanical evidence. U.S. Geological Survey Professional Paper 550-D, pp. D83–D86.
- LaMarche VC (1968) Rates of slope degradation as determined from botanical evidence, White Mountains, California. US Geological Survey Professional Paper 352-I, pp. 341–376.
- Lawrence DB (1950) Estimating dates of recent glacier advances and recession rates by studying tree growth layers. *Transactions of the American Geophysical Union* 31: 243–248.
- Luckman BH (2000) The Little Ice Age in the Canadian Rockies. *Geomorphology* 32: 357–384.
- Malik I (2008) Dating of small gully formation and establishing erosion rates in old gullies under forest by means of anatomical changes in exposed tree roots (Southern Poland). *Geomorphology* 93: 421–436.
- Martin D, Tholl D, Gershenzon J, and Bohlmann J (2002) Methyl jasmonate induces traumatic resin ducts, terpenoids resin biosynthesis, and terpenoids accumulation in developing xylem of Norway spruce stems. *Plant Physiology* 129: 1003–1018.
- Nagy NE, Franceschi VR, Solheim H, Krokene P, and Christiansen E (2000) Wound-induced traumatic resin duct development in stems of Norway spruce (Pinaceae). Anatomy and cytochemical traits. *American Journal of Botany* 87: 302–313.
- Nagy NE, Krokene P, and Solheim H (2005) Anatomical-based defense responses of Scots pine (*Pinus sylvestris*) stems to two fungal pathogens. *Tree Physiology* 26: 159–167.
- Parker ML, Jozsa LA, Johnson SG, and Bramhall PA (1984) Tree-ring dating in Canada and the northwestern U.S. In: Mahaney WC (ed.) *Quaternary Dating Methods*, pp. 211–225. Amsterdam: Elsevier.
- Perret S, Stoffel M, and Kienholz H (2006) Spatial and temporal rockfall activity in a forest stand in the Swiss Prealps – A dendrogeomorphological case study. *Geomorphology* 74: 219–231.
- Pierson TC (2007) Dating young geomorphic surfaces using age of colonizing Douglas fir in southwestern Washington and northwestern Oregon, USA. *Earth Surface Processes and Landforms* 32: 811–831.
- Qiu D, Wilson IW, Gan S, Washusen R, Moran GF, and Southerton SG (2008) Gene expression in Eucalyptus branch wood with marked variation in cellulose microfibril orientation and lacking G-layers. *New Phytologist* 179: 94–103.
- Roer I, Gärtner H, and Heinrich I (2007) Dendrogeomorphological analysis of alpine trees and shrubs growing on active and inactive rockglaciers. In: Haneca K, Verheyden A, Beekmann H, Gärtner H, and Schleser GH (eds.) *TRACE – Tree Rings*

- in *Archaeology, Climatology and Ecology*, vol. 5, pp. 248–258. Jülich: Forschungszentrum Jülich.
- Salzer MW and Hughes MK (2007) Bristlecone pine tree rings and volcanic eruptions over the last 5000 yr. *Quaternary Research* 67: 57–68.
- Schweingruber FH (1993) *Trees and Wood in Dendrochronology Morphological, Anatomical, and Tree-Ring Analytical Characteristics of Trees Frequently Used in Dendrochronology*, vol. VI. Berlin: Springer.
- Schweingruber FH (1996) *Tree Rings and Environment. Dendroecology. Swiss Federal Institute for Forest, Snow and Landscape Research*. Birmensdorf and Berne: Paul Haupt.
- Schweingruber FH (2001) *Dendroökologische Holzanatomie. Anatomische Grundlagen der Dendrochronologie*. Birmensdorf: Swiss Federal Institute for Forests, Snow and Landscape Research, Bern, Stuttgart, Vienna: Paul Haupt-Verlag.
- Schweingruber FH (2007) *Wood Structure and Environment. Springer Series in Wood Science*. Heidelberg: Springer.
- Schweingruber FH, Eckstein D, Serre-Bachet F, and Bräker OU (1990) Identification, presentation and interpretation of event years and pointer years in dendrochronology. *Dendrochronologia* 8: 9–38.
- Shroder JF (1980) Dendrogeomorphology: Review and new techniques of tree-ring dating. *Progress in Physical Geography* 4: 161–188.
- Solomina ON (2002) Dendrogeomorphology: Research requirements. *Dendrochronologia* 20: 233–245.
- Stoffel M (2008) Dating past geomorphic processes with tangential rows of traumatic resin ducts. *Dendrochronologia* 26: 53–60.
- Stoffel M, Bollschweiler M, Butler DR, and Luckman BH (eds.) (2010) *Tree Rings and Natural Hazards: A State-of-the-Art*. Berlin: Springer.
- Stoffel M, Lièvre I, Conus D, et al. (2005) 400 years of debris-flow activity and triggering weather conditions: Ritigraben, Valais, Switzerland. *Arctic, Antarctic, and Alpine Research* 37: 387–395.
- Strunk H (1991) Frequency distribution of debris flow in the Alps since the 'Little Ice Age'. *Zeitschrift für Geomorphologie N. F. Supplement-Band* 83: 71–81.
- Timell TE (1973) *Ultrastructure of the Dormant and Active Cambial Zones and the Dormant Phloem Associated with Formation of Normal and Compression Woods in Picea abies (L.) Karst.* Syracuse, New York: State University of New York, College of Environmental Science and Forestry. Technical Publication No. 96.
- Timell TE (1986) *Compression Wood in Gymnosperms 2*. In: *Springer Wood Series*, 3 vols., pp. 706–1338. Berlin: Springer.
- Vroblesky DA and Yanosky TM (1990) Use of tree-ring chemistry to document historical ground-water contamination events. *Ground Water* 28: 677–684.
- Wardrop AB (1964) Reaction anatomy of arborescent angiosperms. In: Zimmermann MH (ed.) *Formation of Wood in Forest Trees*, pp. 405–456. New York: Academic Press.
- Weiss E (1991) Reconstruction of mass movement activity by study of tree rings. Epoch lecture 26.
- Wiles GC, Calkin PE, and Jacoby GC (1996) Tree-ring analysis and Quaternary geology: Principles and recent applications. *Geomorphology* 16: 259–272.
- Yumoto M, Ishida S, and Fukazawa K (1983) Studies on the formation and structure of the compression wood cells induced by artificial inclination in young trees of *Picea glauca*. IV. Gradation of the severity of compression wood tracheids. *Research Bulletins of the College Experiment Forests* 40: 409–454. Hokkaido University.

Dendroglaciology

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Introduction

Dendroglaciology is a subdiscipline of dendrochronology that uses the tree-ring dating principles established by Douglass (1909, 1914) to assign ages to glacier-related sediments, landforms, and landscape processes (Koch, 2009; Luckman, 1988; Schweingruber, 1988). Common dendroglaciological methodologies include dating trees killed either directly or indirectly by glaciers, establishing the age of trees colonizing the surface of glacial deposits and landforms, and dating growth irregularities in ice-marginal forests affected by glaciers. The assigned annually resolved ages are related by various means to short- and long-term episodes of glacier expansion, retreat, and/or downwasting.

The application of tree-ring analyses to the study of glacier behavior began over 90 years ago (Cooper, 1916, 1923; Tarr and Martin, 1914). Dendroglaciological efforts have become increasingly effective as retreating glaciers expose the remains of forests overridden and buried beneath till deposits. The resulting temporal and spatial insights into Holocene glacier–climate relationships provide a framework for understanding glacier responses to future climate changes at local, regional, and global scales (Le Roy et al., 2009; Matthews and Briffa, 2005). Ring-width series contained in dendroglaciological samples have also been used to extend living tree-ring records through cross-dating (tree-ring-width pattern matching), allowing for the reconstruction of past climate conditions, and enabling the development of long-term proxy glacier mass balance records in remote areas where no instrumental data exist.

Dendroglaciological Applications

Glacially Killed Trees

Glaciers can kill trees either by expanding into standing forests, shearing their stems, and burying them beneath sediments (Allen and Smith, 2007; Luckman, 1995), or by partially burying them beneath distal proglacial outwash gravels (Garavaglia and Palfini, 2010; Wood and Smith, 2004). In recently deglaciated settings, glacially sheared stumps and boles are sometimes discovered in growth position with their roots embedded in a paleosol (Figure 1; Table 1). These remains are often found down-valley of protruding valley-side or valley-bottom bedrock outcrops, where low-pressure zones and pressed till protected them from glacial erosion and reworking (Koehler and Smith, 2011; Mathews, 1951).

Dendroglaciological remains are also commonly found within the eroded slopes and gullies of moraines where they can be deposited in various ways. In the case of composite lateral moraines composed of stacked till units deposited by successive glacier advances, the unit contacts often contain prominent wood mats comprising the remains of tree boles and stumps that colonized the moraine surface during interglacial episodes (Osborn et al., 2001; Rothlisberger et al.,

1980). While standing trees are often killed by debris spilling on distal moraine slopes, trees growing on proximal slopes are generally killed by the upward expansion of ice (Jackson et al., 2008). The latter process tends to crush and masticate trees, producing wood detritus buried at an angle reminiscent of the original moraine slope (Figure 2). Where these remains are pressed into a paleosol they can be interpreted as being near their original growth positions. In contrast, the distal spillage of rocks and boulders has a tendency to tilt and bury entire trees. Postglacial erosion of the proximal moraine face frequently distinguishes the latter as buried boles with projecting roots (Figure 3).

Intervals of glacial expansion can also be identified by dating the remains of trees that died after being submerged in glacially dammed lakes and bogs, or were buried by proglacial stream deposits. Expanding trunk valley glaciers frequently blockade unglaciated tributary valleys leading to ice-marginal ponding and flooding (Capps et al., 2011; Wiles et al., 2011) with the death date of the flooded tributary valley forests providing a convincing record of glacial expansion (Figure 4). In other situations, standing trees located downstream of glacier termini may be partially buried by outwash sediments (Garavaglia and Palfini, 2010) and their exposed apical stems sheared off by an advancing ice front (Figure 1; Wood and Smith, 2004).

Detrital tree remains and wood fragments are also commonly found in glacier forefield settings where they are encased in till, eroded from overlying moraine deposits, or scattered by glacial outwash (Allen and Smith, 2007; Osborn et al., 2007). Without stratigraphic context, however, these samples cannot be assumed to have been glacially killed as other processes (i.e., snow avalanches) may have caused their deaths (Table 1). Only where detrital remains have been pressed into an underlying paleosol and overlain by till can articulated remains be assumed to have been emplaced by an expanding glacier (Allen and Smith, 2007).

In general, bark and perimeter wood is generally absent or partially decayed on dendroglaciological samples, and minimum kill dates are assigned by cross-dating to a ‘master’ tree-ring chronology developed from living trees (Figure 5). While it is sometimes difficult to find long-lived living stands for cross-dating, multicentury and millennium-long master chronologies have been constructed using dendroglaciological samples (Barclay et al., 1999, 2009; Scuderi, 1987; Wiles et al., 2011). When cross-dating to living chronologies is not possible, samples are assigned relative ages by radiocarbon-dating (Koehler and Smith, 2011; Wood and Smith, 2004).

Surface Dating

Dendroglaciology is commonly used to assign ages to trees found growing on recently deposited surfaces (Figure 6). In these settings, tree age provides a minimum date of surface stabilization, which may in turn relate to glacier expansion

and/or retreat (Table 1; Allen and Smith, 2007; Koch, 2009; Koehler and Smith, 2011). To accurately assess the surface age, the oldest tree(s) present belonging to the first generation of colonizers must be sampled (Birks, 1980). These individuals are best located by sampling all, or as many trees as possible, at a site; particularly on well-forested surfaces where the oldest individuals may be difficult to locate (Table 2). Indicators such as species composition and the presence of coarse woody debris can help to distinguish older stands from first generation ones (Koch, 2009). In some cases, the oldest trees may be dead and these standing snags provide excellent opportunities for extending the temporal depth of calendar-dated tree-ring width chronologies (Koehler and Smith, 2011; Luckman, 1988).

On the surface of a recent deposit where only very young trees are found, stems can be cut at the tree base for ring counting (Figure 7), or an age may be assigned based on the number of annual whorls (Palmer and Miller, 1961). It is also important to consider the location of tree growth. For example,



Figure 1 Sheared stumps in growth position protruding from an ice proximal meltwater channel along the northern terminus of the Saskatchewan Glacier, Canadian Rocky Mountains, Alberta. The stumps were buried beneath the glacier until historical recession and stream incision exposed the 225- to 226-year-old stand of subalpine fir, Englemann spruce and whitebark pine trees. Cross-dating shows that all the subfossil stumps and boles exposed at this location were killed during a Neoglacial advance of the Saskatchewan Glacier in ca. 2850 ^{14}C years BP.



Figure 2 Detrital remains of tree boles and branches pressed into a paleosol and buried inside a prominent lateral moraine at Bromley Glacier, northern British Columbia Coast Mountains. The trees were killed as the glacier expanded up the proximal face of the moraine.



Figure 3 Large tree bole with attached roots eroding from the proximal face of the Franklin Glacier lateral moraine, central British Columbia Coast Mountains. The tree was killed in the early Little Ice Age as the glacier expanded down valley.

Table 1 Dendroglaciological dating methods

Type of evidence	Dating precision	Information provided	Potential limitations
Glacially killed trees	Exact calendar date of tree death	In growth position: kill date indicates glacier position at a specific time Detrital wood provides limiting date for glacial event	Requires cross-dating to determine tree age or kill date Dating precision for an event depends on preservation of wood and loss of outer rings
Surface dating	10 years or greater	Age of the oldest tree provides minimum estimate for surface age	Eccentric interval can be difficult to estimate Assumption that the oldest tree has been sampled
Abrupt change in ring symmetry as a result of tilting	Exact calendar date of tilting event	Date of eccentric or abnormal growth indicates onset of event	Event may be severe enough to kill the tree; the floating chronology would require cross-dating
Ice-contact scars	Exact calendar date of damage or ice-contact event	Damage date indicates glacier position at a specific time	Scarred, living trees are often difficult to find Dead trees require cross-dating.



Figure 4 Standing boles killed after being submerged in an ice-marginal pond during the Little Ice Age at Scimitar Glacier, central British Columbia Coast Mountains. The trees are rooted at several levels within the exposed sediments, indicating that the pond filled and drained several times as the surface of the glacier blockading the adjacent tributary valley expanded and downwasted.

trees may germinate on the distal side or crest of a moraine while ice remains in contact with the proximal side. Individuals sampled on the distal side therefore provide older and more accurate age estimates.

Samples for surface dating are best collected at the stem base where the maximum number of annual rings are present (**Figure 7**). Where destructive sampling is not possible, a tree



Figure 6 Recently deposited nested lateral moraine crests at Franklin Glacier, central British Columbia Coast Mountains. The age of the young subalpine fir trees growing on the crests provide a minimum date of surface stabilization and indicate when the glacier was no longer depositing sediment.

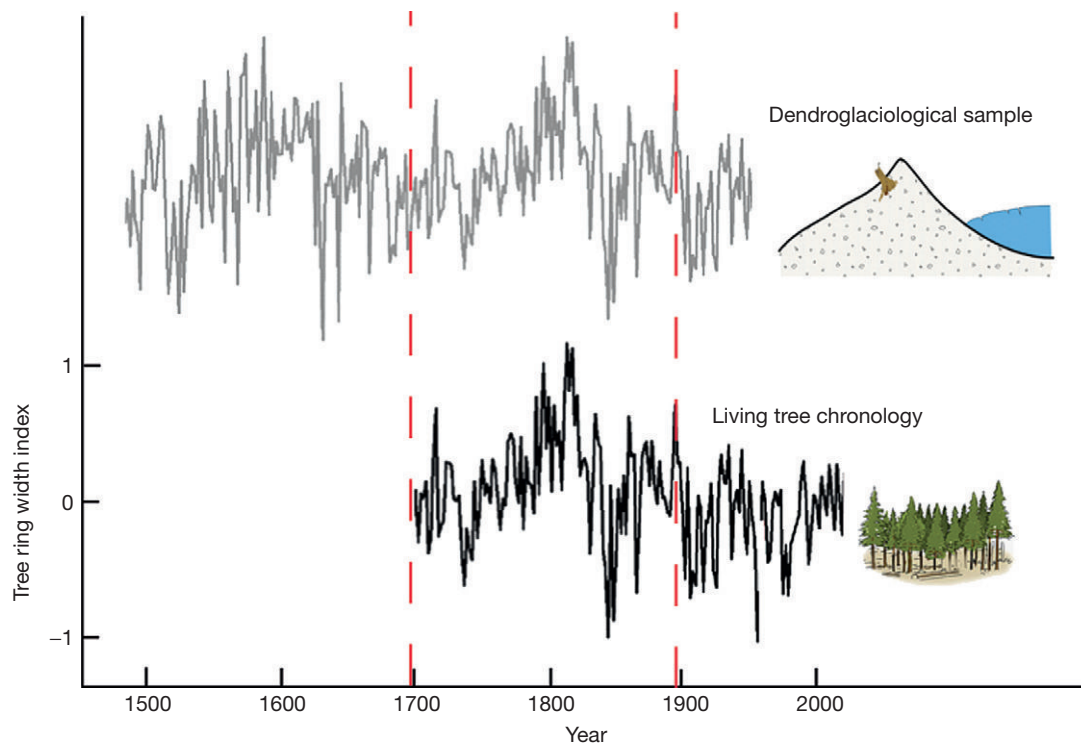


Figure 5 Schematic diagram illustrating how a calendar date for the growth rings of an undated wood sample (e.g., subfossil stump) can be determined by cross-dating with a dated living chronology. By anchoring the floating chronology to the living chronology, it is possible to determine the kill date of the stump and to temporally extend a chronology.

Table 2 Potential tree-ring dating errors and associated correction factors

Potential dating error	Error term	Method of calculation
OT = oldest tree	=0 if all trees are sampled	Sample all trees if possible
HE = height-age error	=0 if sampled at root crown = n , depends on height sampled above the root crown	Estimate time to grow to sampling height: – determine mean apical growth rate – sample at given intervals up the stem – factor in environmental stressors
PE = pith error	=0 if pith sampled = n , depends on years from pith	Estimate distance to pith in years based on ring curvature from similar aged trees
EI = ecesis interval	=1, . . . , n , depends on seed source, microclimate, and substrate	Calculate ecesis interval by subtracting seedling age from a known substrate age

An estimation for the minimum date of a glacial deposit (i.e., moraine) can be summarized by:
Landform date = Germination date + EI
Germination date = Age of the oldest tree = ER + HE + PE

ER = year of earliest ring.
HE = height error.
PE = pith error.

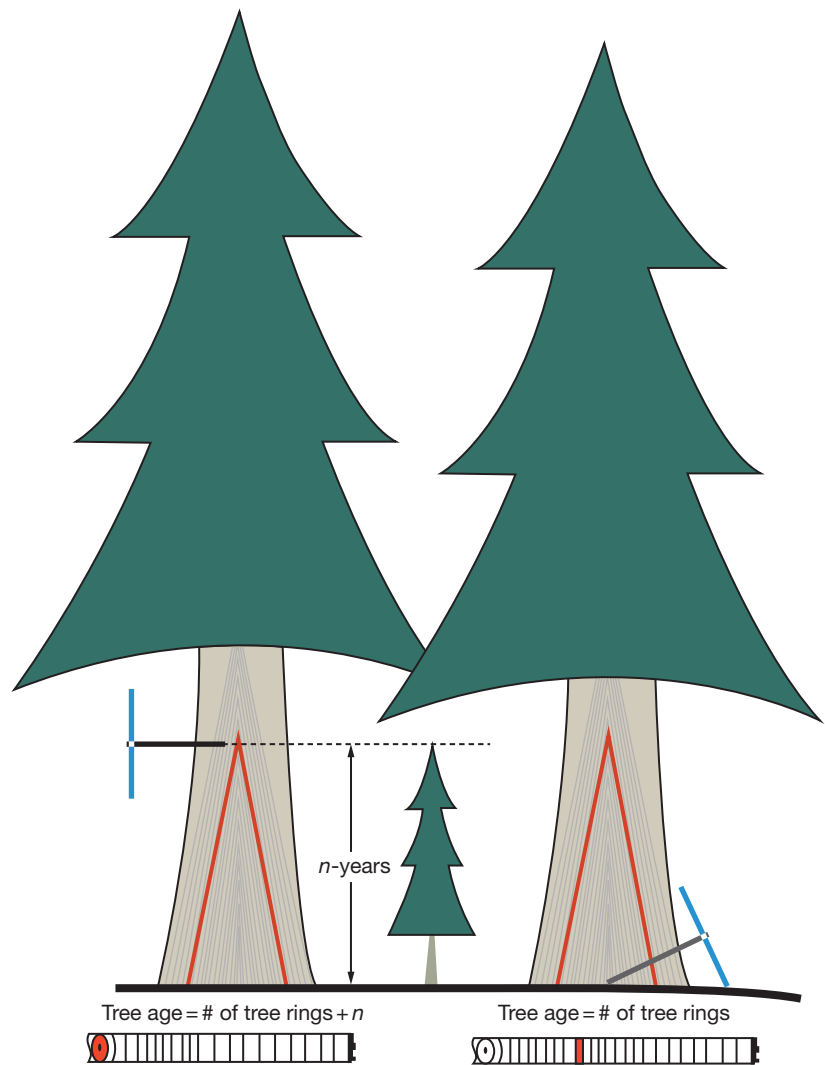


Figure 7 Schematic diagram illustrating the problems associated with coring above the root crown interface. The tree on the left is cored at a height of 1.5 m, the same height as the sapling in the middle, and the tree on the right is cored at the root crown interface. The red line in the trees and their associated cores below provides a marker year that illustrates the loss of rings (n) when determining tree age. If, for example, it took the sapling 30 years ($n = 30$) to reach the height at which the tree on the left was cored, the core taken from that tree would underestimate the age of the surface it is growing on by 30 years. The age and height of saplings growing adjacent to trees of interest is used to determine an apical growth rate (i.e., age-height correction) of 30-years/1.5 m (2 years cm^{-1}) for this example. Applied to the tree on the left, a more accurate surface date is obtained by multiplying the age-height correction by the height of the sample taken.

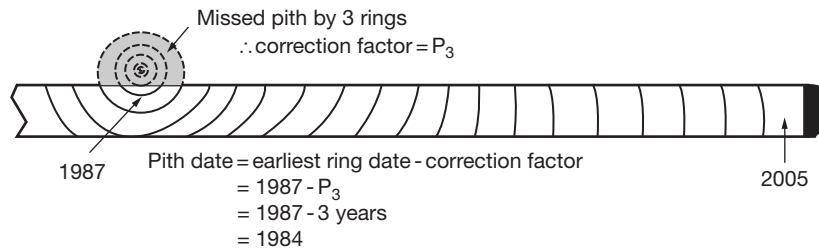


Figure 8 Schematic diagram illustrating the year-to-pith correction for core samples that were close to, but did not include, the pith. For an 18-year-old tree that missed the pith, a correction factor of 3 years was determined by tree-ring data (width, curvature, concentricity) of adjacent trees. For trees growing in stressed environments such as glacier forefields, the slow grow rates would likely produce missed-pith errors one order of magnitude larger.



Figure 9 Distal lateral moraine ridges at Beluga Glacier, southern British Columbia Coast Mountains. The standing snag cross-dates to a nearby tree-ring chronology. The tree died in AD 1738 shortly after deposition of the adjacent moraine. The living subalpine fir trees shown growing on the moraine ridges in the background are less than 60 years old and, accounting for an ecesis interval of 17 years (Koehler and Smith, 2011), indicate the glacier abutted this site within the last century.

core should be removed low on the stem within the root collar. These samples can be difficult to extract with the pith intact and may damage increment borers and chainsaws (Villalba and Veblen, 1997). Where early rings and/or pith cannot be extracted with repeat sampling, ring widths may be estimated using standard pith-correction methods (Table 2; Figure 8; Bosch and Gutiérrez, 1999; Villalba and Veblen, 1997). Where samples cannot be extracted at the base of a tree, various age–height corrections (Gutsell and Johnson, 2002; Wong and Lertzman, 2001), commonly based on a mean apical growth rate established from nearby trees (Table 2; McCarthy et al., 1991), may be applied to account for missing rings.

The lag time between surface stabilization and tree germination, or the ecesis interval, must also be accounted for (Table 2; Sigafos and Hendricks, 1969). Ecesis is usually assessed by establishing the number of years since a surface became ice-free or influenced by glacier processes and the point in time when trees began to germinate (Figure 9). The surface age is frequently established following examination of sequential historical images or aerial photographs (Garbarino et al., 2010; Masiokas et al., 2009; Wiles and Calkin, 1994; Winchester and Harrison, 2000). While it is recognized that ecesis is species-specific and dependent upon local geological, topographical, and microclimate controls (McCarthy and

Luckman, 1993), ecesis estimates established elsewhere or for different species are often applied to local study environments (Luckman, 1995; Smith et al., 1995; Wiles et al., 1999, 2002, 2003).

Under- or over-age estimation of a surface can occur as a result of false or locally absent tree rings. A false ring may develop if annual growth is temporarily interrupted as a result of a disturbance (Hoffer and Tardif, 2009), while radial growth may be circumferentially discontinuous or nonexistent if annual growth is suppressed (O'Neill, 1963). Although false rings can usually be recognized by their anatomical characteristics, missing rings are best identified by cross-dating comparison with established tree-ring chronologies.

Glacially Modified Trees

Dendrochronological methodologies can be used to identify instances when radial tree growth was either directly or indirectly affected by advancing glaciers. Trees record these events through various growth responses including scarring, changes in stem geometry, and changes in growth rate (Table 1; Luckman, 1988).

In some cases advancing glaciers push over and partially bury trees beneath slumped or bulldozed till deposits along their margins (Bachrach et al., 2004; Luckman, 1988). Apical stems that survive this tilting are capable of righting themselves by growing narrower upslope and/or wider downslope rings characterized by reaction wood (Figure 10). The year of inception of these rings identifies the date of the growth disturbance event (Table 1; Luckman, 1988; Villalba et al., 1990; Wiles et al., 1996). In some instances, new leader stems develop on glacially tilted trees and the age of this growth anomaly can be used to document glacial movement (Luckman, 1988). Radial growth abnormalities in trees found growing on debris-covered glaciers have also been used to monitor contemporary glaciological processes (Pelfini et al., 2007).

Glacier expansion into ice-marginal forests can result in surface abrasion and the development of callous tissue on tree stems. These scars may be dated by counting the number of rings between the year of scar formation and the subsequent growth that has occurred around the remaining circumference of the stem (Table 1; Rabassa et al., 1979). The glacial origin of these scars is particularly convincing when multiple trees exhibit scars of similar age on their ice-proximal sides (Villalba et al., 1990).

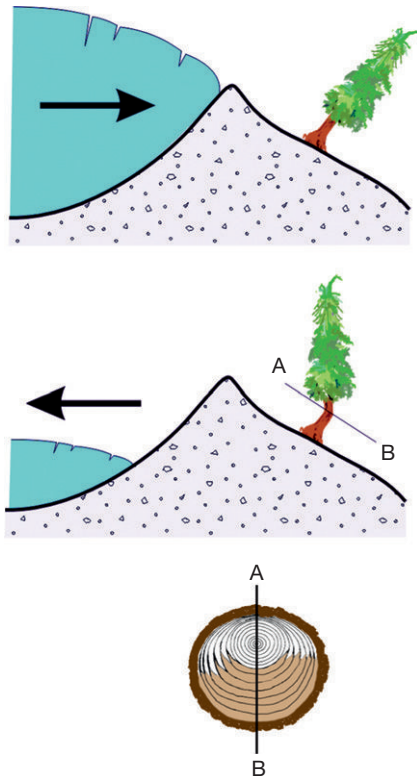


Figure 10 Schematic diagram illustrating the change in wood structure as a result of an advancing ice front tipping a tree growing in front of the glacier. The tilting of the tree results in the formation of both reaction wood and a change from concentric to distorted ring patterns on the underside of the tree. The year the tree was tilted provides a minimum date for the advance of the glacier at that location.

Changes in the growth patterns of trees that remain alive in close proximity to glaciers can reflect the timing of the arrival of ice or moraine development (Le Roy et al., 2009). The onset of cold conditions associated with encroaching ice can cause suppressed annual growth and/or the development of frost rings (Bray and Struik, 1963; Lawrence, 1950; Nicolussi and Patzelt, 1996; Villalba et al., 1990).

Climate and Glacier Mass Balance Reconstruction

Although traditional dendroglaciological techniques make it possible to date the timing of glacier advances and maximum glacier extent (Luckman, 1988), they provide little insight into the duration of glacier advances or the attendant climate conditions (Luckman, 2000; Porter, 1981). By contrast, applied dendroglaciological research has established linkages between glacier mass balance and tree-ring width variability, with Bray and Struik (1963) being among the first to relate tree-ring increments to glacial advances. LaMarche and Fritts (1971) and Matthews and Briffa (1977) later discerned that high winter snow accumulation and short, cool, and cloudy summers favored positive glacier mass balance conditions, but were detrimental to tree growth and resulted in narrow annual tree rings. Conversely, warm summer air temperatures result in

enhanced glacial ablation and the production of wider annual growth rings.

The inverse relationship between the response of climatically sensitive trees and the seasonal variables that drive glacier mass balance provides a means for developing proxy reconstructions of summer, winter, and net annual mass balance in remote areas where no such instrumental data exist (Borgaonkar et al., 2009; Brauning, 2006; Larocque and Smith, 2005; Lewis and Smith, 2004; Nicolussi and Patzelt, 1996; Watson and Luckman, 2004). In many cases these proxy records are consistent with independently constructed paleoclimate reconstructions and dated moraine sequences (Larocque and Smith, 2005; Villalba et al., 1990; Watson and Luckman, 2004; Wood et al., 2011). While inherent nonclimatic influences on radial growth (i.e., disturbance) and glacier mass balance (i.e., glacier size, geometry, debris cover) can be confounding, the temporal coherence of existing dendroglaciological proxy reconstructions highlights the influence of multiscale climate forcing mechanisms on both trees and glaciers (Watson and Luckman, 2004; Wood et al., 2011).

Despite their promise for reconstructing preinstrumental glaciological histories, tree ring derived proxy mass balance records remain limited in their applicability. Leonelli et al. (2011) indicated that tree-ring-based reconstructions of mass balance are most reliable for large, flat valley glaciers with ablation zones unobstructed by debris, also warning that they may not capture the magnitude of extreme ablation events. The paucity of historical mass balance data and high elevation records generally results in short calibration periods that limit the robustness of proxy data sets, a problem accentuated by the lagged responses of both glaciers and trees to climate shifts (Leonelli et al., 2011). The strength and direction of the relationships between both trees and glaciers to climate have also been shown to vary spatially according to elevation and other microclimate effects that are common in complex alpine environments (Leonelli and Pelfini, 2008; Splechtna et al., 2000). Recent reports of nonstationary climatic relationships between glacier behavior and ring-width growth introduce additional concerns (Leonelli et al., 2011; Wood et al., 2011).

Concluding Remarks

The use of dendroglaciological samples in the development of high-resolution, millennial-scale tree-ring records allows for the extension of both Holocene climate and glacier mass balance records. Combined with tree-ring dating these records provide a more complete temporal and spatial picture of paleoglacier–climate relationships, useful for understanding how glaciers may respond to future climates. Future research should focus on employing dendroglaciology to assess the impact of glacier size and geometry on glacier behavior over longer time intervals, closing the temporal gaps in existing dendroglaciological climate and mass balance reconstructions through targeted sampling, and addressing issues of nonstationarity and the complex relationships between tree rings, climate, and glacier mass balance in model building.

See also: Dendrochronology. **Glacial Landforms, Ice Sheets:** Evidence of Glacier and Ice Sheet Extent; Growth and Decay; Paleo-ELAs. **Glacial Landforms, Tree Rings:** Dendrogeomorphology.

References

- Allen SM and Smith DJ (2007) Late Holocene glacial activity of Bridge Glacier, British Columbia Coast Mountains. *Canadian Journal of Earth Sciences* 44: 1753–1773.
- Bachrach T, Jakobsen K, Kinney J, et al. (2004) Dendrogeomorphological assessment of movement at Hilda rock glacier, Banff National Park, Canadian Rocky Mountains. *Geografiska Annaler* 86A: 1–9.
- Barclay DJ, Wiles GC, and Calkin PE (1999) A 1119-year tree-ring-width chronology from western Prince William Sound, southern Alaska. *The Holocene* 9: 79–84.
- Barclay DJ, Wiles GC, and Calkin PE (2009) Holocene glacier fluctuations in Alaska. *Quaternary Science Reviews* 28: 2034–2048.
- Birks HJB (1980) The present flora and vegetation of the moraines of the Klutan Glacier, Yukon Territory, Canada a study in plant succession. *Quaternary Research* 14: 60–86.
- Borgaonkar HP, Ram S, and Sikder AB (2009) Assessment of tree-ring analysis of high-elevation *Cedrus deodara* D. Don from Western Himalaya (India) in relation to climate and glacier fluctuations. *Dendrochronologia* 27: 59–69.
- Bosch O and Gutiérrez E (1999) La sucesión en los bosques de *Pinus uncinata* del Pirineo. De los anillos decrecimiento a la historia del bosque. *Ecología* 13: 133–171.
- Brauning A (2006) Tree-ring evidence of 'Little Ice Age' glacier advances in southern Tibet. *The Holocene* 16: 369–380.
- Bray JR and Struik GJ (1963) Forest growth and glacial chronology in eastern British Columbia, and their relation to recent climatic trends. *Canadian Journal of Botany* 41: 1245–1271.
- Capps DM, Wiles GC, Clague JJ, and Luckman BH (2011) Tree-ring dating of the nineteenth-century advance of Brady Glacier and the evolution of two ice-marginal lakes. *The Holocene* 21: 641–649.
- Cooper WS (1916) Plant successions in the Mount Robson region. *Plant World* 19: 211–238.
- Cooper WS (1923) The recent ecological history of Glacier Bay, Alaska: The present vegetation cycle. *Ecology* 4: 223–246.
- Douglass AE (1909) Weather cycles in the growth of big trees. *Monthly Weather Review* 37: 225–237.
- Douglass AE (1914) A method for estimating rainfall by the growth of trees. *The American Geographical Society Bulletin* 46: 321–335.
- Garavaglia V and Palfini M (2010) Glacier stream activity in the proglacial area of debris covered glacier in Aosta Valley, Italy: An application of dendroglaciology. *Geografia fisica e dinamica quaternaria* 33: 15–24.
- Garbarino M, Lingua E, Nagel TA, Godone D, and Motta R (2010) Patterns of larch establishment following deglaciation of Ventina glacier, central Italian Alps. *Forest Ecology and Management* 259: 583–590.
- Gutsell SL and Johnson EA (2002) Accurately ageing trees and examining their height-growth rates: Implications for interpreting forest dynamics. *Journal of Ecology* 90: 153–166.
- Hoffer M and Tardif JC (2009) False rings in jack pine and black spruce trees from eastern Manitoba as indicators of dry summers. *Canadian Journal of Forest Research* 39: 1722–1736.
- Jackson SI, Laxton SC, and Smith DJ (2008) Dendroglaciological evidence for Holocene glacial advances in the Todd Icefield area, northern British Columbia Coast Mountains. *Canadian Journal of Earth Sciences* 45: 83–98.
- Koch J (2009) Improving age estimates for late Holocene glacial landforms using dendrochronology – Some examples from Garibaldi Provincial Park, British Columbia. *Quaternary Geochronology* 4: 130–139.
- Koehler L and Smith DJ (2011) Late Holocene glacial activity in Manatee Valley, southern Coast Mountains, British Columbia, Canada. *Canadian Journal of Earth Sciences* 48: 603–618.
- LaMarche VC and Fritts HC (1971) Tree rings, glacial advance, and climate in the Alps. *Zeitschrift für Gletscherkunde und Glazialgeologie* 7: 125–131.
- Larocque S and Smith DJ (2005) 'Little Ice Age' proxy glacier mass balance records reconstructed from tree rings in the Mt Waddington area, British Columbia Coast Mountains, Canada. *The Holocene* 15: 748–757.
- Lawrence DB (1950) Glacier fluctuations for six centuries in southeastern Alaska and its relation to solar activity. *Geographical Review* 40: 191–223.
- Le Roy M, Astrade L, Edouard JL, Miramont C, and Deline P (2009) La dendroglaciologie ou l'apport des l'étude des cernes d'arbres pour la reconstruction des fluctuations glacières holocènes. *Neige et Glace de Montagne: Reconstruction, Dynamiques, Pratiques* 8: 79–90.
- Leonelli G and Pelfini M (2008) Influence of climate and climate anomalies on Norway spruce tree-ring growth at different altitudes and on glacier responses: Examples from the central Italian Alps. *Geografiska Annaler* 90A: 75–86.
- Leonelli G, Pelfini M, D'Arrigo R, Haeberli W, and Cherubini P (2011) Non-stationary responses of tree-ring chronologies and glacier mass balance to climate in the European Alps. *Arctic, Antarctic, and Alpine Research* 43: 56–65.
- Lewis D and Smith DJ (2004) Dendrochronological mass balance reconstruction, Strathcona Provincial Park, Vancouver island, British Columbia, Canada. *Arctic, Antarctic, and Alpine Research* 36: 598–606.
- Luckman BH (1988) Dating the moraines and recession of Athabasca and Dome Glaciers, Alberta, Canada. *Arctic and Alpine Research* 20: 40–54.
- Luckman BH (1995) Calendar-dated, early 'Little Ice Age' glacier advance at Robson Glacier, British Columbia, Canada. *The Holocene* 5: 149–159.
- Luckman BH (2000) The Little Ice Age in the Canadian Rockies. *Geomorphology* 32: 357–384.
- Masiokas MH, Rivera A, Espíza LE, Villalba R, Delgado S, and Aravea JC (2009) Glacier fluctuations in extratropical South America during the past 1000 years. *Palaeogeography, Palaeoclimatology, Palaeoecology* 281: 242–268.
- Mathews WH (1951) Historic and prehistoric fluctuations of alpine glaciers in the Mount Garibaldi map-area, south-western British Columbia. *Journal of Geology* 59: 357–380.
- Matthews JA and Briffa KR (1977) Glacier and climatic fluctuations inferred from tree-growth variations over the last 250 years, central southern Norway. *Boreas* 6: 1–24.
- Matthews JA and Briffa KR (2005) The 'Little Ice Age': Re-evaluation of an evolving concept. *Geografiska Annaler* 87A: 17–36.
- McCarthy DP and Luckman BH (1993) Estimating ecesis for tree-ring dating of moraines: A comparative study from the Canadian Cordillera. *Arctic and Alpine Research* 25: 63–68.
- McCarthy DP, Luckman BH, and Kelly PE (1991) Sampling height-age error correction for spruce seedlings in glacial forefields, Canadian cordillera. *Arctic and Alpine Research* 23: 451–455.
- Nicolussi K and Patzelt G (1996) Reconstructing glacier history in Tyrol by means of tree-ring investigations. *Zeitschrift für Gletscherkunde und Glazialgeologie* 32: 207–215.
- O'Neill LC (1963) The suppression of growth rings in jack pine in relation to defoliation by the Swaine jack-pine sawfly. *Canadian Journal of Botany* 41: 227–235.
- Osborn G, Menounos B, Koch J, Clague JJ, and Vallis V (2007) Multi-proxy record of Holocene glacial history of the Spearhead and Fitzsimmons ranges, southern Coast Mountains, British Columbia. *Quaternary Science Reviews* 26: 479–493.
- Osborn GD, Robinson BJ, and Luckman BH (2001) Holocene and latest Pleistocene fluctuations of Stutfield Glacier, Canadian Rockies. *Canadian Journal of Earth Sciences* 38: 1141–1155.
- Palmer WH and Miller AK (1961) Botanical evidence for the recession of a glacier. *Oikos* 12: 75–86.
- Pelfini M, Santilli M, Leonelli G, and Bozzoni M (2007) Investigating surface movements of debris-covered Miage glacier, Western Italian Alps, using dendroglaciological analysis. *Journal of Glaciology* 53: 141–152.
- Porter SC (1981) Glaciological evidence of Holocene climatic change. In: Wigley TML, Ingram MJ, and Farmer G (eds.) *Climate and History*, pp. 82–110. Cambridge: Cambridge University Press.
- Rabassa J, Rubulis S, and Suarez J (1979) Rate of formation and sedimentology of (1976–1978) push-moraines, Frias Glacier, Mount Tronador (41°10'S–71°43'W), Argentina. In: Schluechter C (ed.) *Moraines and Varves*, pp. 65–79. Rotterdam: Balkema.
- Rothlisberger F, Haas P, Holzhauser H, Keller W, Bircher W, and Renner F (1980) Holocene climate fluctuations – Radiocarbon dating of fossil soils (fAh) and woods from moraines and glaciers in the Alps. *Geographica Helvetica* 35: 21–52.
- Schweingruber F (1988) Radiodensitometric-dendroclimatological conifer chronologies from Lapland (Scandinavia) and the Alps (Switzerland). *Boreas* 17: 559–566.
- Scuderi LA (1987) Glacier variations in the Sierra Nevada, California, as related to a 1200-year tree-ring chronology. *Quaternary Research* 27: 220–231.
- Sigafos RS and Hendricks EL (1969) The time interval between stabilization of alpine glacial deposits and establishment of tree seedlings. *US Geological Survey Professional Paper* 485-A, pp. 89–93.
- Smith DJ, McCarthy DP, and Colenutt ME (1995) Little Ice Age glacial activity in Peter Lougheed and Elk Lakes Provincial Parks, Canadian Rocky Mountains. *Canadian Journal of Earth Sciences* 32: 579–589.
- Splechna BE, Dobrys J, and Klinka K (2000) Tree-ring characteristics of subalpine fir (*Abies lasiocarpa* (Hook.) Nutt.) in relation to elevation and climatic fluctuations. *Annals of Forest Science* 57: 89–100.

- Tarr RS and Martin L (1914) *Alaskan Glacier Studies*, p. 498. Washington, DC: The National Geographic Society.
- Villalba R, Leiva JC, Rubulls S, Suarez J, and Lenzano L (1990) Climate, tree-ring, and glacial fluctuations in the Rio Frias Valley, Rio Negro, Argentina. *Arctic and Alpine Research* 22: 215–232.
- Villalba R and Veblen TT (1997) Regional patterns of tree population age structures in northern Patagonia: Climatic and disturbance influences. *Journal of Ecology* 85: 113–124.
- Watson E and Luckman BH (2004) Tree-ring-based mass-balance estimates for the past 300 years at Peyto Glacier, Alberta, Canada. *Quaternary Research* 62: 9–18.
- Wiles GC, Barclay DJ, and Calkin PE (1999) Tree-ring-dated 'Little Ice Age' histories of maritime glaciers from western Prince William Sound, Alaska. *The Holocene* 9: 163–173.
- Wiles GC and Calkin PE (1994) Late Holocene, high-resolution glacial chronologies and climate, Kenai Mountains, Alaska. *Geological Society of America Bulletin* 106: 281–303.
- Wiles GC, Calkin PE, and Jacoby GC (1996) Tree-ring analysis and Quaternary geology: Principles and recent applications. *Geomorphology* 16: 259–272.
- Wiles GC, Jacoby GC, Davi NK, and McAllister RP (2002) Late Holocene glacier fluctuations in the Wrangell Mountains, Alaska. *Geological Society of America Bulletin* 114: 896–908.
- Wiles GC, Lawson DE, Lyon E, Wiesenberger N, and D'Arrigo RD (2011) Tree-ring dates on two pre-Little Ice Age advances in Glacier Bay national Park and Preserve, Alaska, USA. *Quaternary Research* 76: 190–195.
- Wiles GC, McAllister RP, Davi NK, and Jacoby GC (2003) Eolian response to Little Ice Age climate change, Tana Dunes, Chugach Mountains, Alaska, USA. *Arctic, Antarctic, and Alpine Research* 35: 67–73.
- Winchester V and Harrison S (2000) Dendrochronology and lichenometry: Colonization, growth rates and dating of geomorphological events on the east side of the North Patagonian Icefield, Chile. *Geomorphology* 34: 181–194.
- Wong CM and Lertzman KP (2001) Errors in estimating tree age: Implications for studies of stand dynamics. *Canadian Journal of Forest Research* 31: 1262–1271.
- Wood C and Smith DJ (2004) Dendroglaciological evidence for a Neoglacial advance of the Saskatchewan Glacier, Banff National Park, Canadian Rocky Mountains. *Tree-Ring Research* 60: 59–65.
- Wood L, Smith DJ, and Demuth MN (2011) Extending the Place Glacier mass balance record to AD 1585, using tree-rings and wood density. *Quaternary Research* 76: 305–313.

Glacial–Interglacial Scale Fluvial Responses

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Introduction

Fluvial landforms and deposits are common features on the subaerial landscape, and fluvial deposits are now widely recognized on submerged continental shelves. Accordingly, fluvial landforms and deposits provide a readily studied record of terrestrial landscape evolution, and how surface processes respond to, and record, Quaternary climate change, sea-level change, and active uplift or subsidence. This article provides an overview of fluvial responses to Quaternary glacial–interglacial cycles.

History of Ideas

As recently as the 1950s and 1960s, the prevailing model for glacial–interglacial climate cycles was interwoven with, and in large part derived from, Penck and Brückner's (1909) study of terraces along tributaries to the Danube River in southern Germany. Prior to their work, Pleistocene glaciation was an accepted concept, but multiple glaciations were not yet firmly established. However, Penck and Brückner (1909) differentiated four moraines that could be traced to four separate terraces. Each terrace contained a deep weathering profile and was separated from the next younger terrace by a period of valley incision. From these data, they inferred four Pleistocene glacial–interglacial cycles, with each consisting of a long glacial, a long interglacial, and short transitions between end-member states. This 'square-wave' model for glacial–interglacial climate change (Figure 1(a)) remained dominant until the development of oxygen-isotope records from ocean basins in the 1950s–1980s.

Of equal importance, Penck and Brückner (1909) linked fluvial aggradation to glacial-period sediment supply, and incision with terrace formation to interglacial periods (Figure 1(b)). Many early studies focused on this 'glacial' versus 'interglacial' dichotomy, and assumed that present floodplains represent the Holocene, the first terrace was from the last Pleistocene glacial, the second terrace was from the previous glacial, and so on. This model was influential in the development of the climatic geomorphology popular among many twentieth century European scientists, their study established terraces and associated deposits as foci for the study of fluvial responses to climate change at the glacial–interglacial scale, and correlation of terraces to glacial periods can be found through the literature today.

Fisk's (1944) study of the Lower Mississippi Valley of the USA was similarly important. According to Fisk, the valley was deeply incised to a point some 1000 km upstream from the present shoreline due to increased channel gradients during sea-level fall and lowstand. Braided streams formed when gradients decreased during latest Pleistocene to middle Holocene

sea-level rise, with transition to a meandering regime during the late Holocene highstand (Figure 1(c)). Using this reasoning, he correlated terraces of the Lower Mississippi Valley and Gulf Coastal Plain to previous interglacial highstands. Fisk's model was 180° out of phase with Penck and Brückner (1909) in that deposition occurred during interglacials, and incision with terrace formation occurred during glacials. However, Fisk's model was equally influential in that many workers in coastal regions assigned successively older terraces to successively older interglacials.

During the 1960s–1970s, records of climate and sea-level change, and interpretations of changes in fluvial systems over time, have become increasingly independent. The development of ^{14}C dating in the 1950s provided an opportunity to develop chronological frameworks for climate change from paleobiological and other indicators, and, independently, for fluvial successions. Also in the 1950s, Leopold and Maddock's (1953) hydraulic geometry studies initiated a generation of efforts that developed quantitative relations between discharge regimes, sediment load, and measures of channel geometry. Development of ^{14}C dating and the hydraulic geometry paradigm made it possible to move beyond counting terraces and inferring correlations with glacial stages. It became clear that direct correlations between glacial–interglacial cycles and fluvial landforms were more complicated, and that fluvial systems were more sensitive to Holocene-scale climate changes than previously recognized (Knox, 1983).

The proliferation of paleobiological and geochemical data from ocean sediments in the 1960s–1980s, and acceptance of the Milankovitch theory for glacial–interglacial cycles, produced a paradigm shift in our understanding of the structure of glacial–interglacial cycles. However, in that same period, the temporal limits of ^{14}C -focused study on fluvial activity during the last 20–30 ka, the Last Glacial Maximum and the Holocene, which represents only the last 20–30% of a single 100-ka glacial–interglacial cycle. Re-examination of longer-term fluvial system evolution has been hampered by a paucity of geochronological methods that extend through a complete 100-ka glacial–interglacial cycle. Newer geochronological methods with longer time windows, for example, luminescence techniques and cosmogenic radionuclides, have been applied to fluvial landforms and deposits only in the middle to late 1990s (e.g., Blum and Price, 1998; Page et al., 1996; Phillips et al., 1997).

Glacial Versus Interglacial Worlds

Present understanding of Quaternary climate and sea-level change provides a set of global- to regional-scale boundary conditions for understanding fluvial system changes through time. For example, the new picture of glacial–interglacial cycles

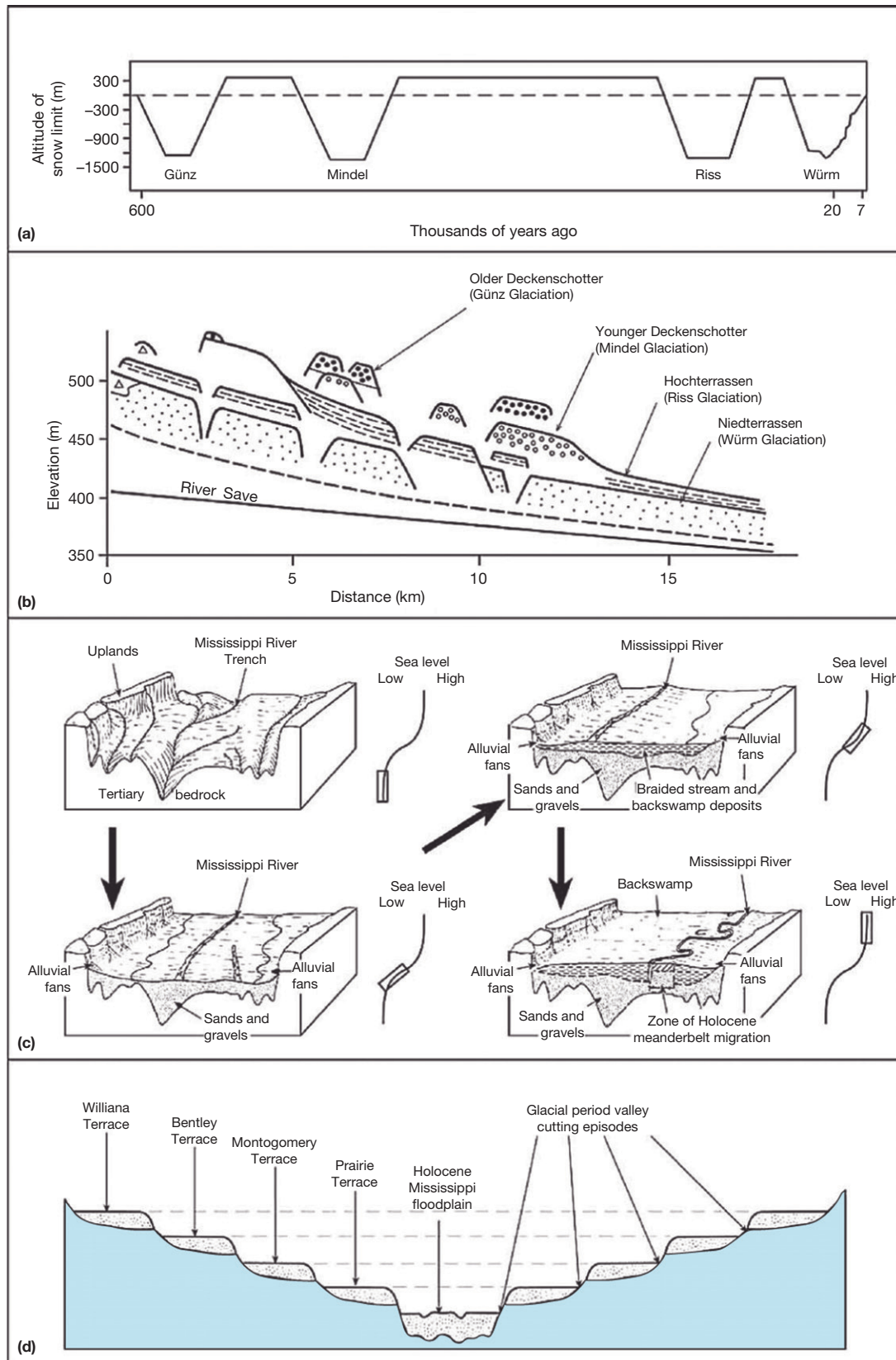


Figure 1 Historical ideas. (a) Rendition of original Penck and Brückner (1909) square-wave model for Quaternary glacial-interglacial cycles. (b) Penck and Brückner (1909) model for relationships between glacial moraines and outwash terraces in the Danube drainage of southern Germany. (c) Fisk (1944) model for Mississippi River response to sea-level change. (d) Fisk model for terraces of the Gulf of Mexico coast in Louisiana.

that emerged from studies of oxygen-isotope records in deep-sea sediments provides a model for global changes in climate and ice volume, and the glacioeustatic component of sea level over the last several millions of years (Figure 2(a)). More recently, work on ice cores and the marine record shows that significant globally coherent climate changes at timescales of 10^3 – 10^4 years are superimposed on the 100-ka glacial–interglacial cycles discussed above (Figure 2(b)).

Recognition of the 100-ka frequency for middle to late Pleistocene glacial–interglacial cycles, and recognition of the variability within each cycle, clearly rendered obsolete older models for climate and sea-level change, such as Penck and Brückner (1909) or Fisk (1944). Moreover, fluvial systems

respond to climate changes within their catchments and/or relative sea-level changes at their mouths, and it is clear that global-scale forcing of climate and sea-level change are manifested regionally in a variety of ways. For example, the last glacial period was cooler than today everywhere, but changes in precipitation and discharge regimes varied in direction or magnitude from region to region. Relative sea-level changes also depart from glacioeustasy in both rate and magnitude, and sometimes in direction.

Nevertheless, one concept that can be distilled concerns the nature of average Earth system conditions, and how they depart significantly from the old square-wave model, in that full interglacials like those of the Holocene are relatively short-

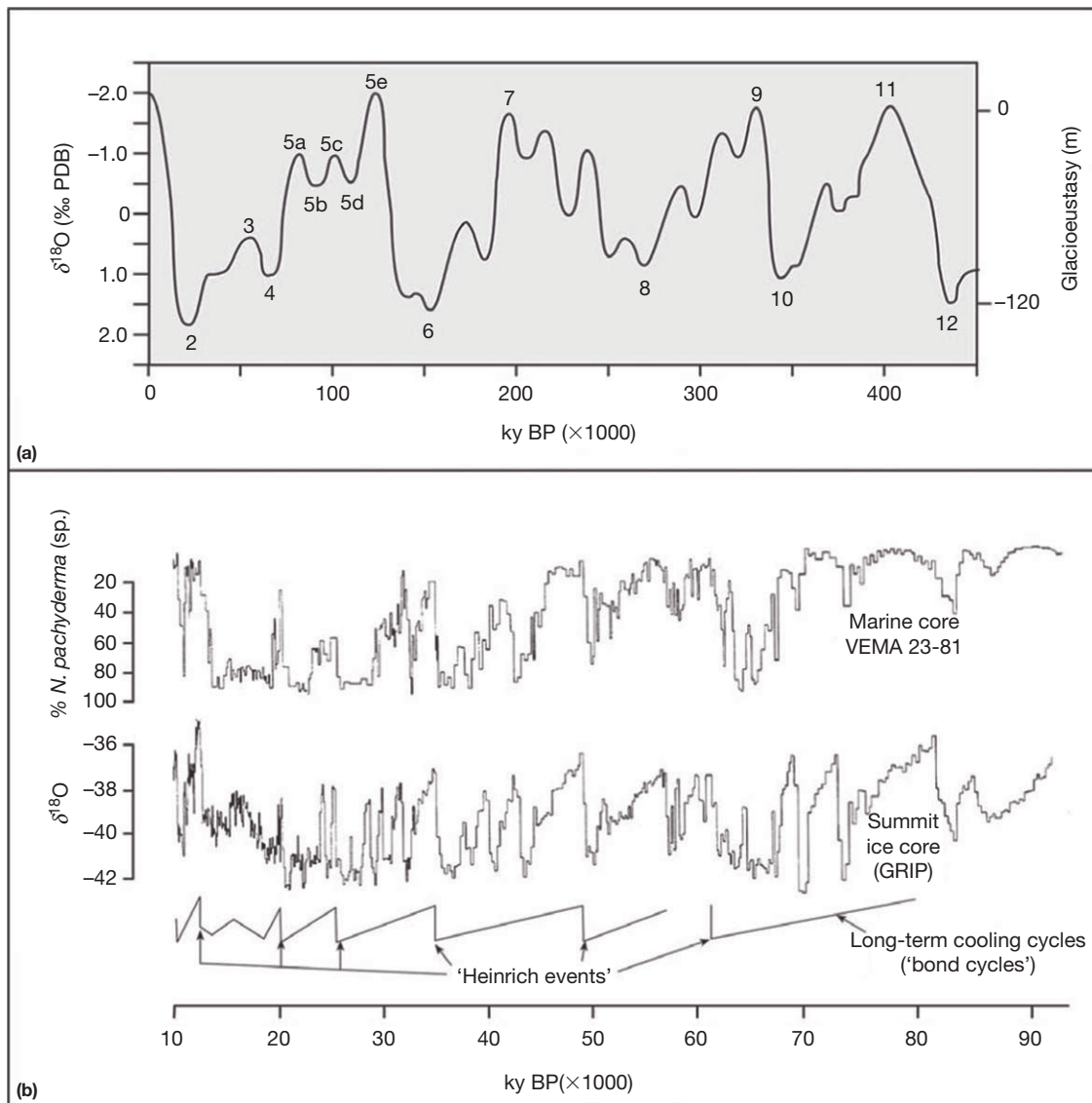


Figure 2 Quaternary climate and sea-level change. (a) Oxygen-isotope record from ocean basins for the last 400 ka, illustrating 100-ka cyclicality in glacial–interglacial cycles and inferred glacioeustatic component of sea-level change. (b) Comparison of the proportion of the planktonic coldwater foraminifer *Neogloboquadrina pachyderma* (sinistral) in marine core VM 23-81 with the $\delta^{18}\text{O}$ record from the Summit Ice Core (Greenland). Dansgaard–Oeschger cycles (D–O cycles) are the millennial-scale wiggles, whereas Bond cycles are the long-term cooling trends punctuated by abrupt warming associated with Heinrich events. Reproduced from Bond G, Broecker W, Johnsen S, et al. (1993) Correlations between climate records from North Atlantic sediments and Greenland ice. *Nature* 365: 143–147.

lived, as are full glacials. Up to 80% of any middle to late Pleistocene 100-ka cycle witnessed global temperatures that were significantly cooler than a full interglacial, but not as cold as a full glacial and eustatic sea level would have been at -15 to -65 m (Porter, 1989). Hence, an average world would be one that is significantly different from today, in terms of climate, vegetation, slope stability, and sediment production. Moreover, presently submerged continental shelves were previously exposed as subaerial extensions of present-day coastal plains, and shorelines to which rivers are graded resided in mid-shelf or farther basinward positions.

Glacial Versus Interglacial Fluvial Systems

Conceptual Framework

Theory, numerical and physical modeling, and empirical studies show that valley and channel gradients, channel size and shape, and the spatial distribution of erosional and depositional landforms reflect rates of uplift or subsidence, prevailing discharge regimes, and prevailing sediment loads (Figure 3(a) and 3(b)). Channels respond to changes in these controls through a series of morphological and sedimentary adjustments, and by storing or removing sediments from valley axes (Blum and Törnqvist, 2000; Bridge, 2003; Bull, 1991; Knox, 1983). Changes in discharge regimes and sediment load can be externally forced by climate change, glaciation, and relative sea-level change; however, a number of factors make linkages between external forcing and fluvial response less than straightforward (Figure 3(c)). These include (1) spatial and temporal scale, (2) convergence of responses to different forcing mechanisms, (3) divergence of response following similar external inputs, (4) differential sensitivity, (5) lag times between time of forcing and time of response, and (6) complexity of response due to internal system dynamics (Schumm, 1991).

The record of fluvial response to glacial–interglacial cycles varies by position within a river system due to local- to regional-scale rates of uplift or subsidence, the resultant ability of fluvial systems to incise, aggrade, and modulate sediment storage over various timescales, and proximity to sea-level and the influence of relative sea-level change (Figure 3). Steep upper reaches of drainage networks within rapidly uplifting continental interiors typically have high stream powers, incising bedrock channels with discontinuous floodplains, and short residence times for sediment as a whole. Records of fluvial responses to glacial–interglacial cycles are fragmentary, and may extend only through the last glacial–interglacial period. By contrast, lower gradient mixed bedrock–alluvial valleys (sensu Howard et al., 1994) are common to the vast relatively low relief continental interiors, and tend to have more continuous floodplains and longer sediment residence times. Outside of formerly glaciated regions, mixed bedrock–alluvial valleys typically contain flights of terraces and associated fluvial deposits (Figure 4(a)) that span multiple glacial–interglacial cycles.

Alluvial valleys and alluvial-deltaic plains are common to subsiding basins, for example, passive continental margins (e.g., the US Gulf of Mexico coast, or the east coast of China) or partially flooded foreland basins (e.g., the Po Valley and

Adriatic Sea or the Tigris–Euphrates Valley and the Persian Gulf). Sediments delivered to subsiding basins are preserved in the stratigraphic record when they subside below maximum depths of incision: until that time, they can be remobilized and transported out of the basin (continental interior) or elsewhere in the basin (both continental interior and continental margin basins). For continental margin systems, remobilization is important during glacial periods when river systems extend their courses to falling stage and lowstand shorelines, and develop incised valleys (Zaitlin et al., 1994; Figure 4(b) and 4(c)). As defined here, coastal-plain incised valleys extend from the highstand shoreline to the inland limits of sea-level change, and are filled with fluvial, estuarine, and deltaic deposits, whereas cross-shelf incised valleys extend from highstand to lowstand shorelines, and may include marine facies deposited during transgression and highstand.

Continental Interior Systems

As noted above, flights of terraces have served as focal points for the study of fluvial response to glacial–interglacial cycles since the classic study of Penck and Brückner (1909). It is now recognized that long-term, time-averaged rates of valley incision within many continental interiors closely reflect rates of rock uplift, and elevational differences between terraces with adequate geochronological controls are therefore used to calculate incision rates as a proxy for long-term time-averaged uplift rates (e.g., Pazzaglia and Brandon, 2001). However, the ubiquitous presence of terrace flights also indicates that bedrock incision is a highly unsteady process when considered over glacial–interglacial timescales, such that most geologic time may actually be represented by lateral migration, cutting of bedrock straths, and deposition of overlying fluvial sediments (Figure 5).

Qualitative reasoning has therefore long suggested that many flights of terraces reflect climate-controlled unsteadiness in discharge regimes and sediment supply, superimposed on trends of uplift-controlled bedrock valley incision (Bull, 1991). In a general sense, periods of bedrock valley incision reflect times when the concentration of sediments within channels is sufficiently low such that the entire sediment thickness can be mobilized during flood events, and bedrock can be incised by the mobile sediment load. Periods of lateral migration or aggradation, in contrast, represent times when sediment concentration is sufficiently high that all sediment within the channel is not mobilized, and bedrock straths are not subject to incision. Numerical modeling by Hancock and Anderson (2002) supports these qualitative inferences, as well as the view that climate change at the glacial–interglacial scale can produce flights of terraces.

Within mixed bedrock–alluvial valleys, flights of terraces therefore likely represent the primary response to glacial–interglacial cycles. A number of studies have begun to reexamine terraces in this context, and correlate long terrace sequences to marine isotope records (e.g., Antoine, 1994; Bridgland et al., 2004; Porter et al., 1992; Figure 6). However, there remains no general model, or well-established broadly applicable relationship, that links periods of specific river behavior to specific climate conditions or transitions. Many empirical records indicate the last major period of bedrock valley

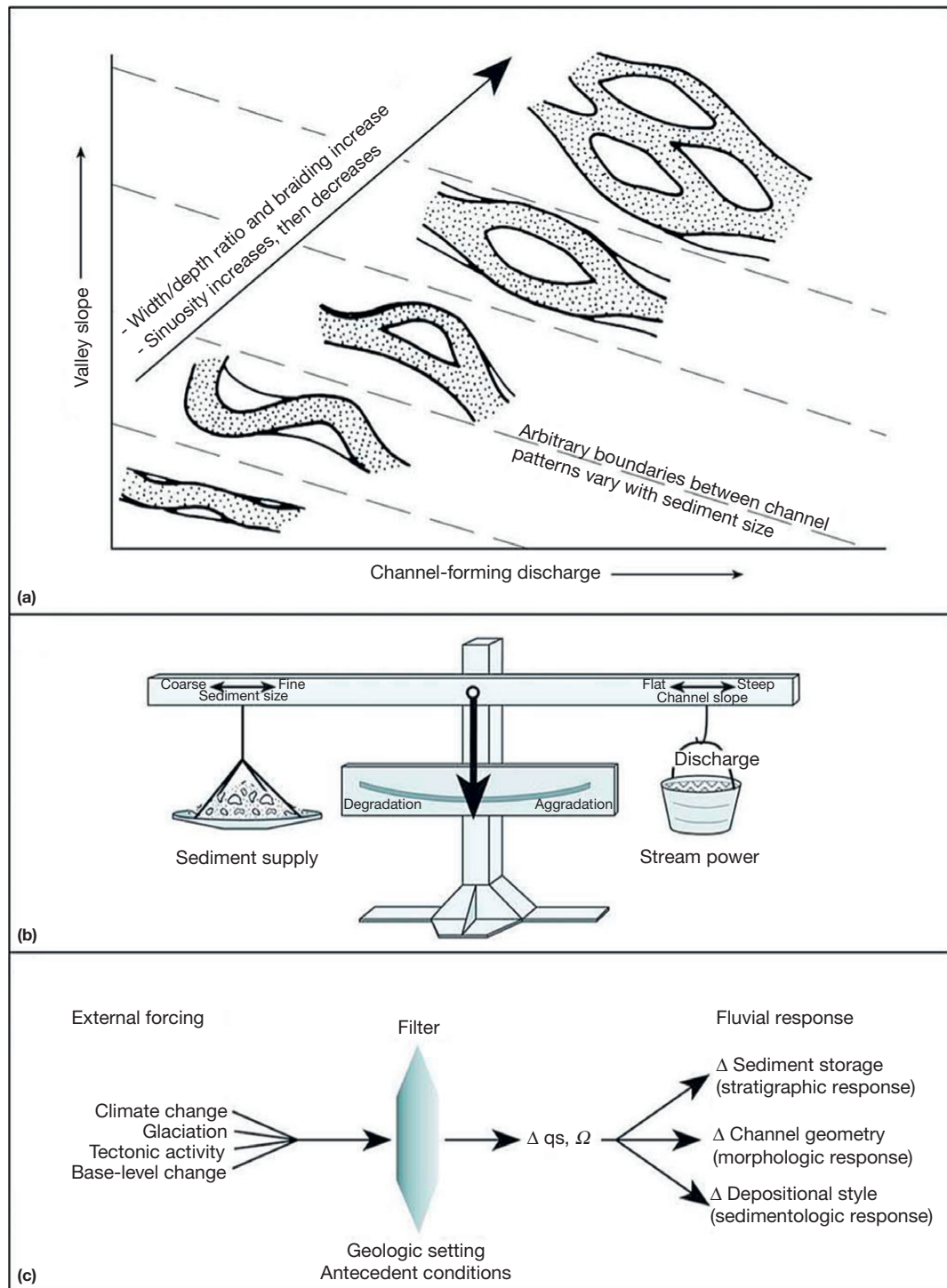


Figure 3 (a) Generalized relations between channel plan-view geometry, a dependent variable, and valley slope versus discharge. Adapted from Bridge JS (2003) *Rivers and Floodplains: Forms, Processes, and the Sedimentary Record*, p. 491. Oxford: Blackwell. (b) Balance model for aggradation and degradation of alluvial channels, emphasizing changes in the relationship between discharge and sediment supply. Channels aggrade when sediment supply exceeds transport capacity, and degrade when the reverse is true. Redrawn from a widely circulated but unpublished drawing by W. Borland of the USA Bureau of Reclamation, based on an equation by Lane (1955; Schumm SA and Ethridge FG, 1999 personal communication). Adapted from Blum MD and Straffin EC (2001) Fluvial response to external forcing: Examples from the Massif Central of France, the Texas Coastal Plain (USA), the Sahara of Tunisia, and the Lower Mississippi Valley (USA). In: Maddy D and Macklin MA (eds.) *River Basin Sediment Systems: Archives of Environmental Change*, pp. 195–228. Lisse, The Netherlands: Balkema Press. (c) Conceptual model for fluvial responses to external forcings. Forcing mechanisms, shown as independent controls on the left, are passed through a filter that defines how they are actually translated into changes in stream power (Ω) and sediment supply (q_s). Changes in these variables actually determine the direction and magnitude of fluvial stratigraphic, morphologic, and sedimentologic responses.

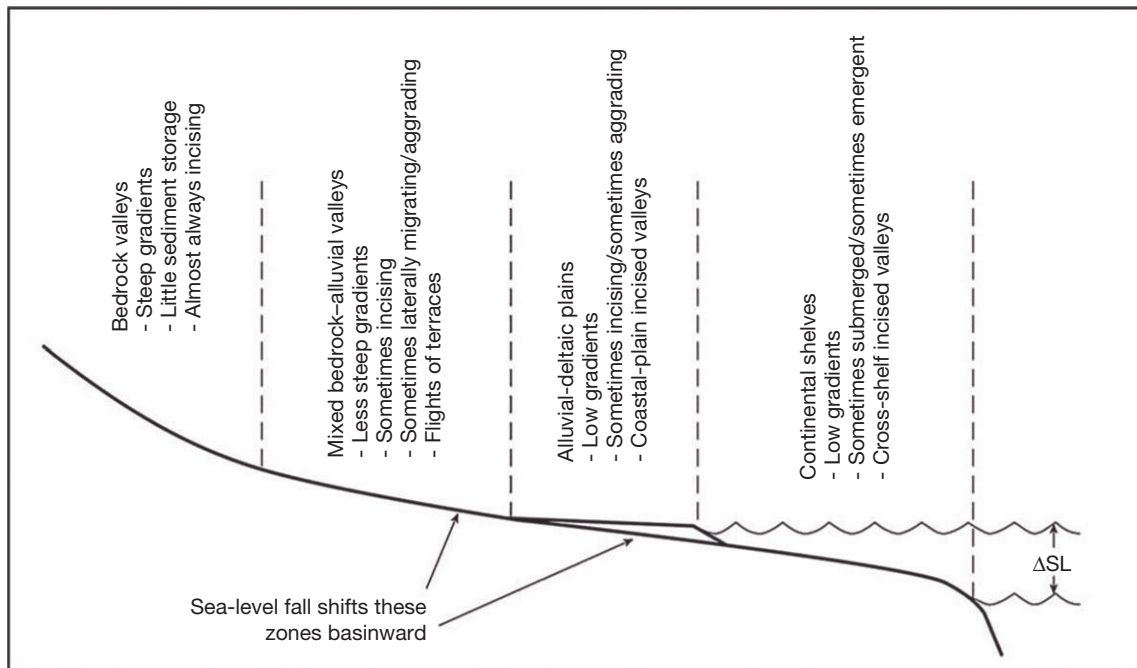


Figure 4 Long profile model illustrating spatial variability in river systems and the nature of their response to climate and sea-level change. In steep upper reaches, bedrock incision dominates, and terraces may be present, but the record will be fragmentary. Lower gradient mixed bedrock–alluvial valleys typically contain flights of terraces that may represent millions of years. Coastal-plain incised valleys and valley fills, as defined here, extend from the upstream limits of sea-level influence, to the highstand shoreline, whereas cross-shelf incised valleys extend from the highstand to lowstand shorelines.

incision, and terrace formation corresponds to the transition from the Last Glacial Maximum to the beginning of the Holocene interglacial: for time periods prior to the Last Glacial Maximum, there are few, if any, examples where the precise timing of lateral migration and deposition of fluvial deposits, or bedrock incision and terrace formation, is sufficiently constrained to make such linkages. Moreover, some studies interpret only one terrace from the last 100-ka glacial period, whereas other studies in the same or different areas identify multiple terraces. The record becomes less clear as one goes back in time to older glacial cycles, such that this remains a fertile topic for future research.

Regardless of the lack of precise correlations between terraces and specific climate periods, it is clear that within individual valleys, terrace surfaces commonly display differences in the morphology of formative channels. This is especially clear when comparing terrace surfaces from the last glacial period with channel patterns of modern interglacial rivers in the same valleys. Attention to morphological changes was a natural outcome of the hydraulic geometry paradigm, with early studies by Dury (1965) and Schumm (1968) discussing large-scale changes in channel geometry as a result of climate change in the US and Australia. Morphological contrasts between glacial and interglacial river systems are widely known from Europe (Figure 7(a); Starkel, 1991), where many present-day temperate regions experienced periglacial climates during glacial periods, or drained alpine glaciers or the Fennoscandian ice sheet. Many European rivers maintained braided planforms during the last glacial period, and sinuous single channels during the Holocene interglacial.

The Lower Mississippi Valley provides an example of this type of morphological response at the glacial–interglacial scale. The Mississippi was the primary conduit for sediments and meltwaters discharged from North American ice sheets, and, in some cases, routed through large ice-marginal lakes. The northern half of the 800-km-long alluvial valley contains terraces with clear braided stream patterns that stand in sharp contrast to the meanderbelts of the present day (Figure 7(b)). Braided channelbelts were interpreted by Fisk (1944) to represent the early Holocene rising stage of sea level. However, Saucier (1994) established that they were closely linked to glacial-period sediment loads and meltwater-dominated discharge regimes. Rittenour et al. (2005) provide chronological control on braided stream surfaces from luminescence dating, and show that the transition to braided stream patterns followed the isotope stage 5–4 transition, and that the reverse transformation occurred at the beginning of the Holocene with the loss of meltwaters and glacially derived sediments.

Another widespread morphological contrast between glacial and interglacial river systems can be found along the US Gulf of Mexico Coastal Plain. Many valleys in this region contain two to three distinct terraces from the last glacial period (Blum and Aslan, 2006). However, in contrast to the braided versus meandering dichotomy, terrace surfaces here display meander scars that are close to an order of magnitude larger than similar features from modern rivers in the same valleys (Figure 7(c)). Drainage basins were not glaciated, or subject to indirect effects of glaciation. Widely accepted process explanations are not yet available, but workers agree these

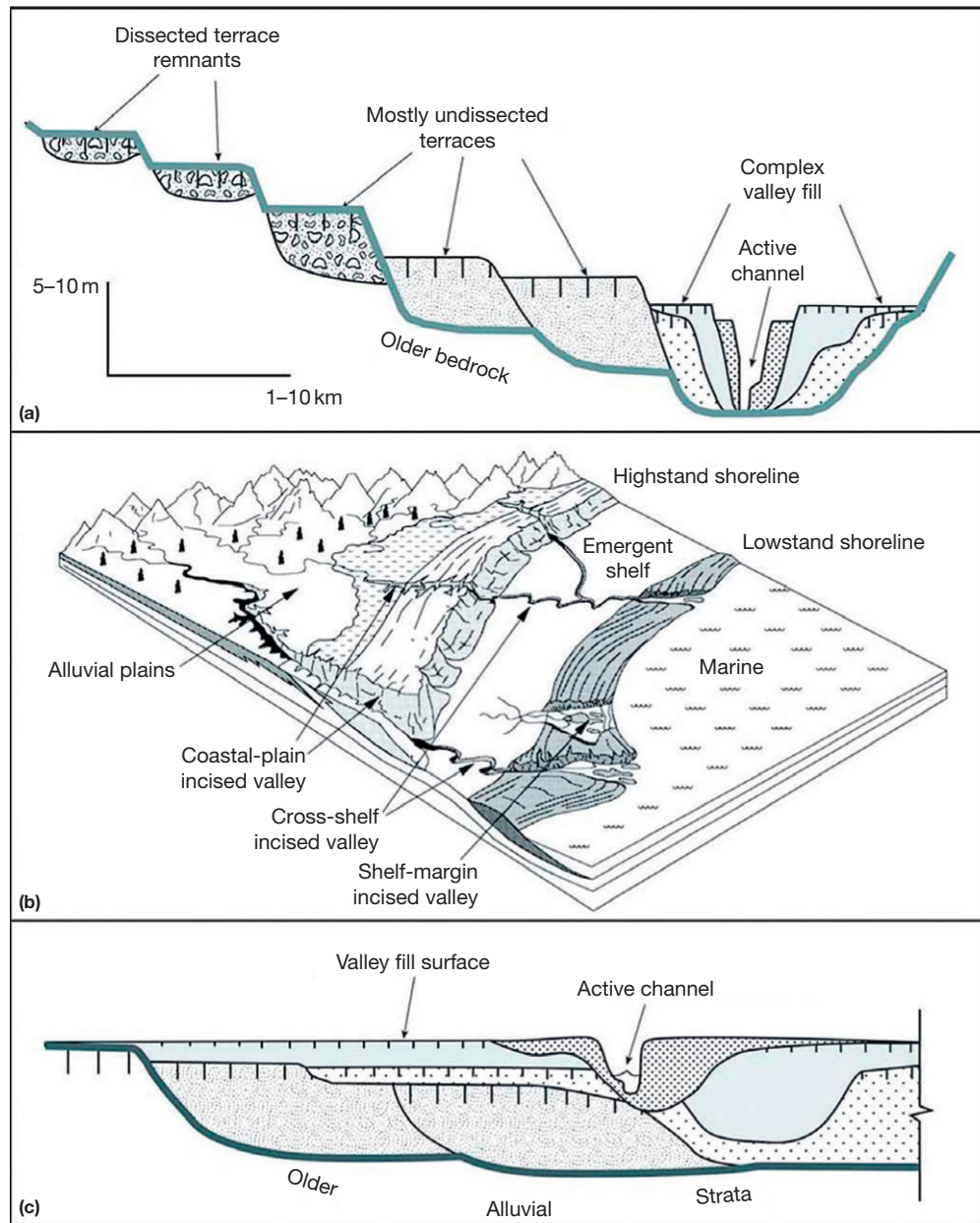


Figure 5 (a) Schematic model for mixed bedrock/alluvial valley setting, with flights of downward-stepping terrace surfaces and underlying fills, and a complex valley fill with multiple stratigraphic units. Older and higher terraces are commonly discontinuous and dissected, with significant thicknesses of strata removed; they also may represent long periods of time, and multiple lateral migration, aggradation, and incision events within a limited elevation range that are not readily differentiated due to degree of preservation. Younger mostly intact terraces and the complex valley fill may represent the last glacial cycle (enclosed by thick gray line). Vertical hachures represent soil profiles, with the time for surface stability and soil development represented by length of hachure. Adapted from Blum MD and Törnqvist TE (2000) Fluvial response to climate and sea-level change: A review and look forward. *Sedimentology* 47(supplement): 1–48. (b) Schematic continental margin setting, where sea-level fall results in development of a coastal-plain and cross-shelf incised valley system that extends through the highstand shoreline, and across the emergent shelf to the lowstand shoreline. Modified from Zaitlin BA, Dalrymple RW, and Boyd R (1994) The stratigraphic organization of incised-valley systems associated with relative sea-level change. In: Dalrymple RW, Boyd R, and Zaitlin BA (eds.) *SEPM Special Publication 51: Incised-Valley Systems: Origins and Sedimentary Sequences*, pp. 45–60. (c) Schematic model for a coastal-plain incised valley, where a cycle of sea-level fall, then rise, results in a succession of downward-stepping, then aggradational, stratigraphic units bounded by paleosols. The solid gray line encloses a single 100-ka glacial–interglacial cycle, and deposits above that line would correlate to deposits above the gray line in (a). Adapted from Blum MD and Törnqvist TE (2000) Fluvial response to climate and sea-level change: A review and look forward. *Sedimentology* 47(supplement): 1–48.

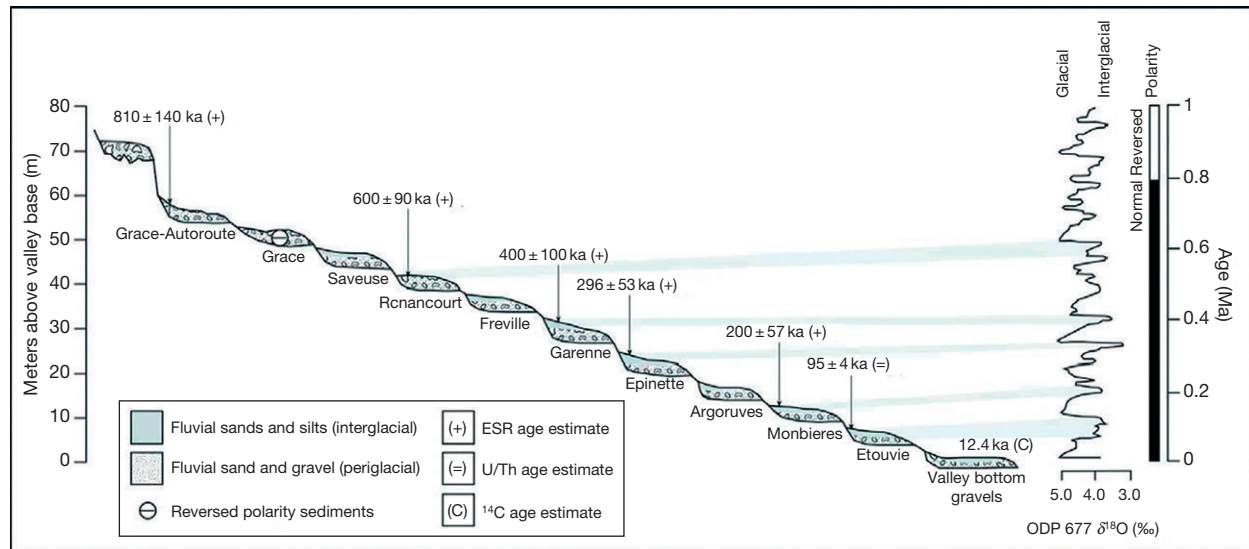


Figure 6 Idealized model for the flight of terraces in the Somme valley, northern France. Locations of key geochronological data as shown, along with suggested correlations between the terrace sequence and the marine record. In this example, each glacial–interglacial cycle in the middle to late Pleistocene has been correlated to an individual terrace, and glacial–interglacial contrasts are expressed as facies within the overall suite of deposits related to each terrace. Adapted and simplified from Bridgland D, Maddy D, and Bates M (2004). River terrace sequences: Templates for Quaternary geochronology and marine–terrestrial correlation. *Journal of Quaternary Science* 19: 203–218.

channels represent glacial-period discharge regimes and sediment loads that were significantly different than those of the present interglacial.

Sedimentological contrasts between glacial and interglacial river systems are common as well, and may ultimately provide the best indicators of changes in process through time. For example, mixed bedrock–alluvial valleys of the Loire River of central France and the Colorado River of the US Gulf of Mexico coast in Texas contain two to three terraces from the last glacial period (Blum and Straffin, 2001): the Loire was braided, whereas the Colorado River constructed the large meanders described above. Regardless of channel morphology, glacial-period deposits are dominated by coarse-grained channel-bar and channel-fill deposits, and systematically lack significant thicknesses of fine sandy and muddy overbank deposits (Figure 8(a) and 8(b)). This suggests that deep overbank flooding was not common and most floods were contained within bankfull channel perimeters during the long glacial period. By contrast, during the short Holocene, there has been little net valley incision in the Loire or Colorado systems, and minor adjustments in channel geometry. However, Holocene units contain appreciable thicknesses of overbank facies (Figure 8(c) and 8(d)), which indicates that deep overbank flooding has been important, and a significant volume of fine sediment is sequestered in overbank settings. Moreover, floodplain facies successions contain multiple weak paleosols that indicate century- to millennial-scale periods of overbank stabilization with soil development alternating with periods of deep overbank flooding and burial of paleosols. This type of high-frequency signature is lacking from the glacial period in either of these systems.

Continental Margin Systems

At some distance upstream from a highstand shoreline, fluvial systems become independent of high-frequency ($<10^6$ year)

sea-level change; mixed bedrock–alluvial valleys typical of continental interiors fall into this category. However, there can be no point moving in the downstream direction where the influence of climate-driven unsteadiness in discharge regimes and sediment load goes to zero; hence, continental margin systems must reflect the influence of both climate and sea-level change (Blum and Törnqvist, 2000).

Beginning in the 1970s and 1980s, examination of high-resolution seismic data showed that glacial period fluvial–deltaic systems had left behind an extensive record on the continental shelves (e.g., Anderson et al., 2004; Suter and Berryhill, 1985). Recent studies in a variety of settings further clarify the nature of the record in continental margin systems, and the role of different forcing mechanisms. From these studies, it is clear that the primary response to the high rates of sea-level change typical of the Quaternary is linked to regression and transgression of the shoreline to which rivers are graded: sea-level fall forces channel extension across newly emergent shelves, whereas sea-level rise forces channel shortening. Within this context, other responses are conditioned by alluvial valley, coastal plain, and shelf gradients (Figure 9), and by climate-controlled unsteadiness in discharge and sediment supply from the continental interior (Blum and Törnqvist, 2000; Ethridge et al., 1998).

At one end of a spectrum, many continental margin systems construct aggradational–progradational shorelines, or shallow subaqueous deltaic clinoforms, during sea-level highstand, through which rivers must incise when sea-level falls and channels extend across the shelves (Figure 9(a)). This type of geometry serves to partition coastal-plain and cross-shelf incised valleys, within which fluvial activity during the glacial period takes place. Valley filling during sea-level rise and highstand then produces a landward-tapering wedge of deposits that onlap fluvial deposits from the glacial period

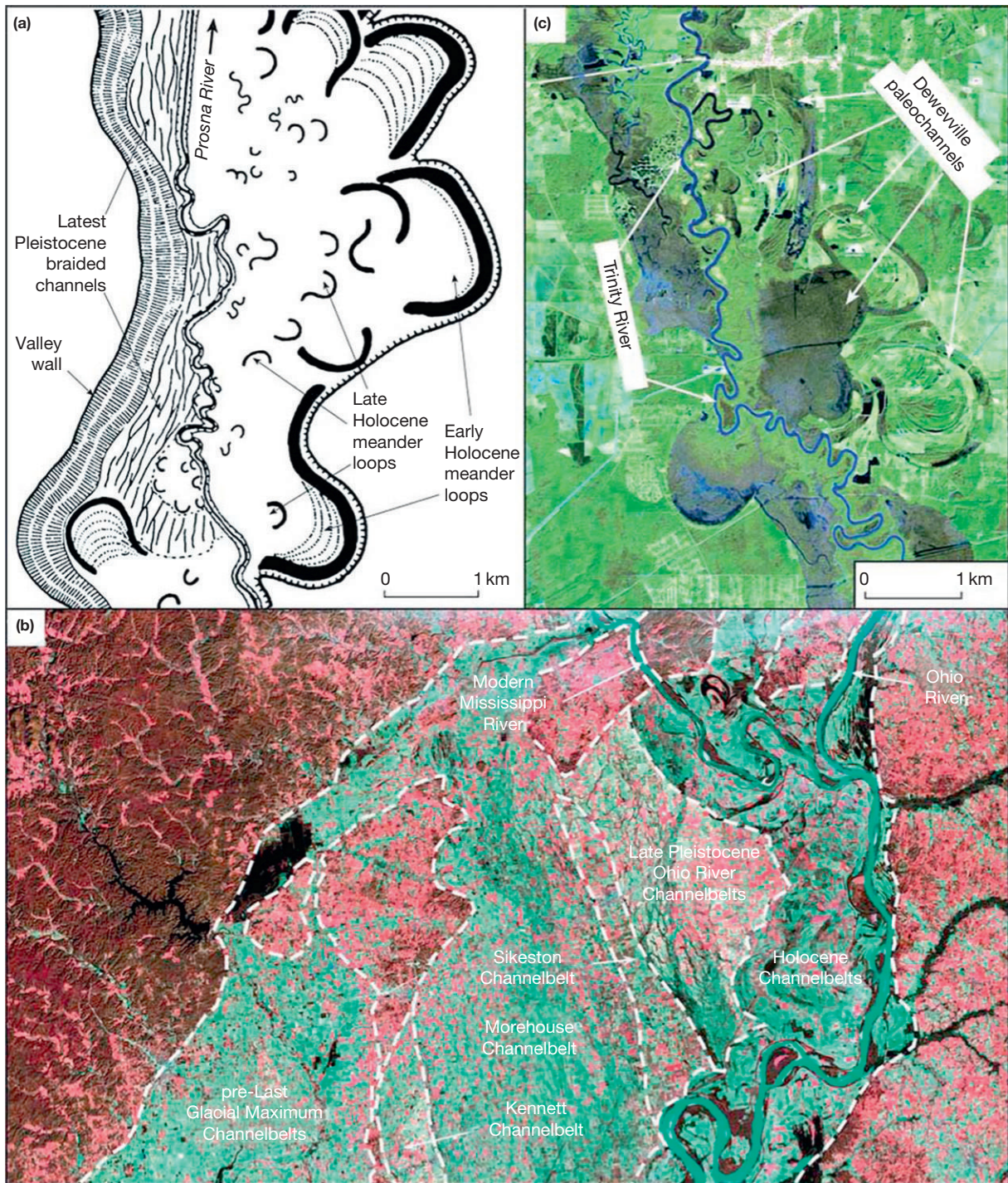


Figure 7 Channel changes in response to glacial versus interglacial climates. (a) Channel changes in plan-view interpreted for the Prosna River in Poland (modified from Blum MD and Törnqvist TE (2000) Fluvial response to climate and sea-level change: A review and look forward. *Sedimentology* 47(supplement): 1–48; original references therein), with late Pleistocene braided channels, large early Holocene meandering channels, and small late Holocene meandering channels. (b) LANDSAT Thematic Mapper image of the northern Lower Mississippi Valley, southcentral US in Missouri, illustrating the contrast between multiple braided channelbelts from the last glacial cycle (OIS 4–2) and the modern meandering channel (map unit designations based on Rittenour et al. (2005)). (c) LANDSAT Thematic Mapper image of the Trinity River, US Gulf Coastal Plain in Texas, illustrating the contrast between ‘Deweyville’ paleochannels from the last glacial cycle (OIS 4–2) and the modern channel from Blum MD and Aslan A (2006) Signatures of climate vs. sea-level change within incised valley fill successions: Quaternary examples from the Texas Gulf Coast. *Sedimentary Geology* 190: 177–211.

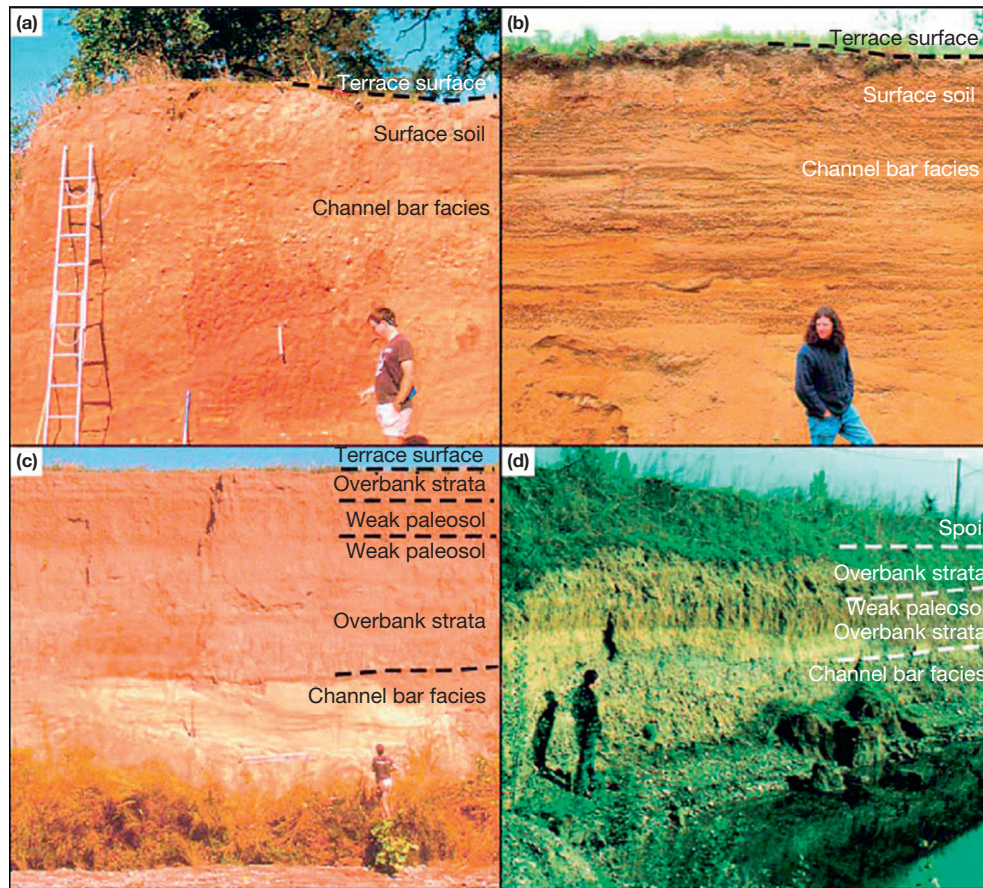


Figure 8 Sedimentological contrasts between glacial and interglacial river systems. (a) Typical exposure in late Pleistocene deposits of the Colorado River, US Gulf Coastal Plain in Texas, illustrating a complete lack of overbank facies. (b) Typical exposure in late Pleistocene deposits of the Loire River, Massif Central of France, illustrating a complete lack of overbank facies. (c) Typical exposure in late Holocene deposits of the lower Colorado River, illustrating thick overbank sand and mud with multiple weakly developed paleosols. (d) Photograph of typical exposure in late Holocene deposits of the Loire, illustrating by thick overbank facies resting on point bar gravel and punctuated by weakly developed soils. Adapted from Blum MD and Straffin EC (2001) *Fluvial response to external forcing: Examples from the Massif Central of France, the Texas Coastal Plain (USA), the Sahara of Tunisia, and the Lower Mississippi Valley (USA)*. In: Maddy D and Macklin MA (eds.) *River Basin Sediment Systems: Archives of Environmental Change*, pp. 195–228. Lisse, The Netherlands: Balkema Press.

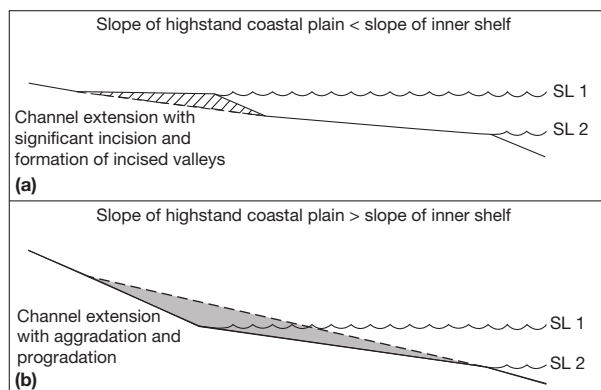


Figure 9 Model for fluvial response to sea-level fall as a function of highstand coastal-plain and inner-shelf gradients. (a) Incision through coastal prism (depositional shoreline break or delta clinoform) due to steeper gradients exposed on the inner shelf as sea-level falls; (b) aggradation and progradation due to sea-level fall across a shelf that has a gradient shallower than the coastal plain. Adapted from Blum MD and Törnqvist TE (2000) *Fluvial response to climate and sea-level change: A review and look forward*. *Sedimentology* 47(supplement): 1–48; original references therein.

(Figure 10). Incised valleys of this type are known from passive and other margins with low-gradient coastal plains, where the highstand depositional shoreline break is steeper than the coastal plain or shelf (e.g., Allen and Posamentier, 1993). At the opposite end of the spectrum, steep coarse-grained rivers that discharge onto a lower-gradient shelf (Figure 9(b)) will aggrade and prograde during sea-level fall, rather than partition a distinct incised valley. During transgression, these systems backstep as well; however, studies of the Canterbury Plains, New Zealand (Leckie, 1994) show that wave erosion of the aggradational and progradational falling stage and lowstand alluvial plain results in steepening of the channel profile, and incision that propagates headward. In this type of setting, valley incision can be related to transgression and highstand.

Sea-level changes also result in the stratigraphic juxtaposition of proximal fluvial and distal fluvial–deltaic–estuarine environments and facies. Allen and Posamentier (1993) illustrate this relationship for the Gironde estuary and Garonne River of SW France (Figure 10(a)). Studies of the Rhine–Meuse in The Netherlands show that glacial-period-braided stream surfaces are overlapped by the Holocene wedge, which

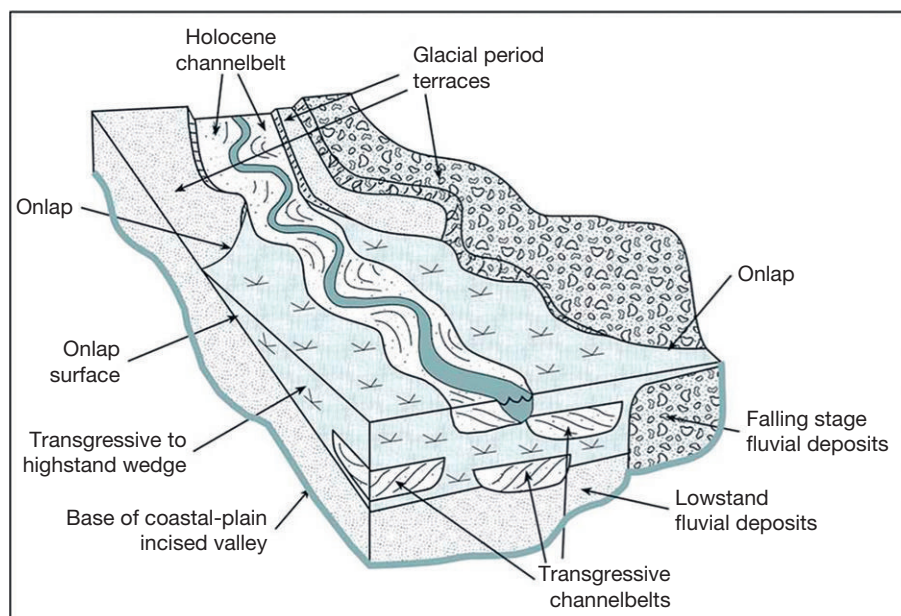


Figure 10 Model for the Gironde estuary and incised valley fill, illustrating channelbelts that represent the glacial period, when sea-level fell and channels extended across the shelf, and which are onlapped by estuarine, channel, and flood plain facies associated with the transgression and highstand. Note that channelbelts deposited during falling stage and lowstand occur as terraces updip from the onlap point, and are buried farther downdip. Modified from Allen GP and Posamentier HW (1993) Sequence stratigraphy and facies model of an incised valley fill: The Gironde Estuary, France. *Journal of Sedimentary Petrology* 63: 378–391.

is dominated by anastomosing, or high-sinuosity meandering, channel patterns, and associated mud-rich alluvial–deltaic deposits (Berendsen and Stouthamer, 2001; Törnqvist et al., 2003). In this case, the glacial period shoreline was hundreds of kilometers farther basinward in the English Channel, in contrast to the upper alluvial–deltaic plain setting today. Similarly, along the US Gulf of Mexico coast in Texas, major river systems cut large coastal-plain and cross-shelf incised valleys during the last glacial period and constructed large deltas on the shelf and at the shelf margin (Figure 11; Anderson et al., 2004; Blum and Aslan, 2006; Blum and Price, 1998). In this case, terrace surfaces with the large meanders described above extend downstream into the incised valleys, where they dip below, and are onlapped by the Holocene wedge. In either case, the common theme is juxtaposition of a more coarse-grained fluvial channelbelt from the glacial period, one that formed when channels were extended to a distant shoreline, with onlapping alluvial–deltaic and estuarine facies from the transgression and highstand.

Over multiple glacial–interglacial cycles, fluvial systems can respond to repeated regressions and transgressions by constructing laterally extensive highstand alluvial–deltaic plains. Blum and Price (1998) and Blum and Aslan (2006) show that alluvial plain surfaces of the Texas coastal plain consist of multiple cross-cutting incised valley fills, with each representing a 100-ka glacial–interglacial cycle (Figure 12). Each valley fill is interpreted to represent incision and extension during glacial period falling stage and lowstand, and valley filling and alluvial plain construction during transgression and highstand (Figure 13). Each valley fill contains multiple coarse-grained channelbelts from the falling stage and lowstand, when rivers were extended to shorelines in mid-shelf or farther basinward positions, which are onlapped and buried by more mud-rich fluvial, estuarine, and deltaic facies from transgressions and highstand.

Blum and Törnqvist (2000) compiled data to illustrate the upstream limits of sea-level influence during highstands, and the distance over which channels extend during lowstands. These data suggest the landward limit of onlap due to sea-level rise correlates to sediment supply, and is inversely related to the gradient of the glacial period floodplain. Accordingly, low-gradient systems with large sediment loads will be onlapped very far upstream: for the Mississippi, this distance exceeds 300 km upstream from the highstand shoreline, whereas distances of <30–40 km apply for steep-gradient, low-sediment-supply systems. For parts of the Canterbury Plains of New Zealand, the upstream limits of sea-level influence would be defined by the upstream limits of incision during highstand, since onlap may never take place.

Distances that channels extend during lowstand reflect shelf morphology. One extreme is the Rhine, which extended into the North Sea, then was diverted southwestward through the English Channel during the last lowstand, a distance of ~800 km, and the Tigris–Euphrates, which previously extended more than 600 km through the Persian Gulf to the Arabian Sea. Large Chinese river systems extended some >400 km, whereas the Po extended ~250 km into the Adriatic Sea, and rivers of the US Gulf of Mexico coast in Texas extended 100–150 km. The Mississippi River might represent the opposite extreme: if sea level were to fall to lowstand positions now, it would not extend at all, since it now discharges to the shelf margin; however, the coarse-grained braided channels characteristic of glacial periods would likely extend downstream as far as the present-day shelf.

An important corollary to channel extension during glacial periods is the merging of drainage basins. Glacial-period rivers like the Rhine and others become the axial stream for greatly enlarged drainage basins due to the addition of systems that, in a highstand world, discharge separately to the coastal oceans

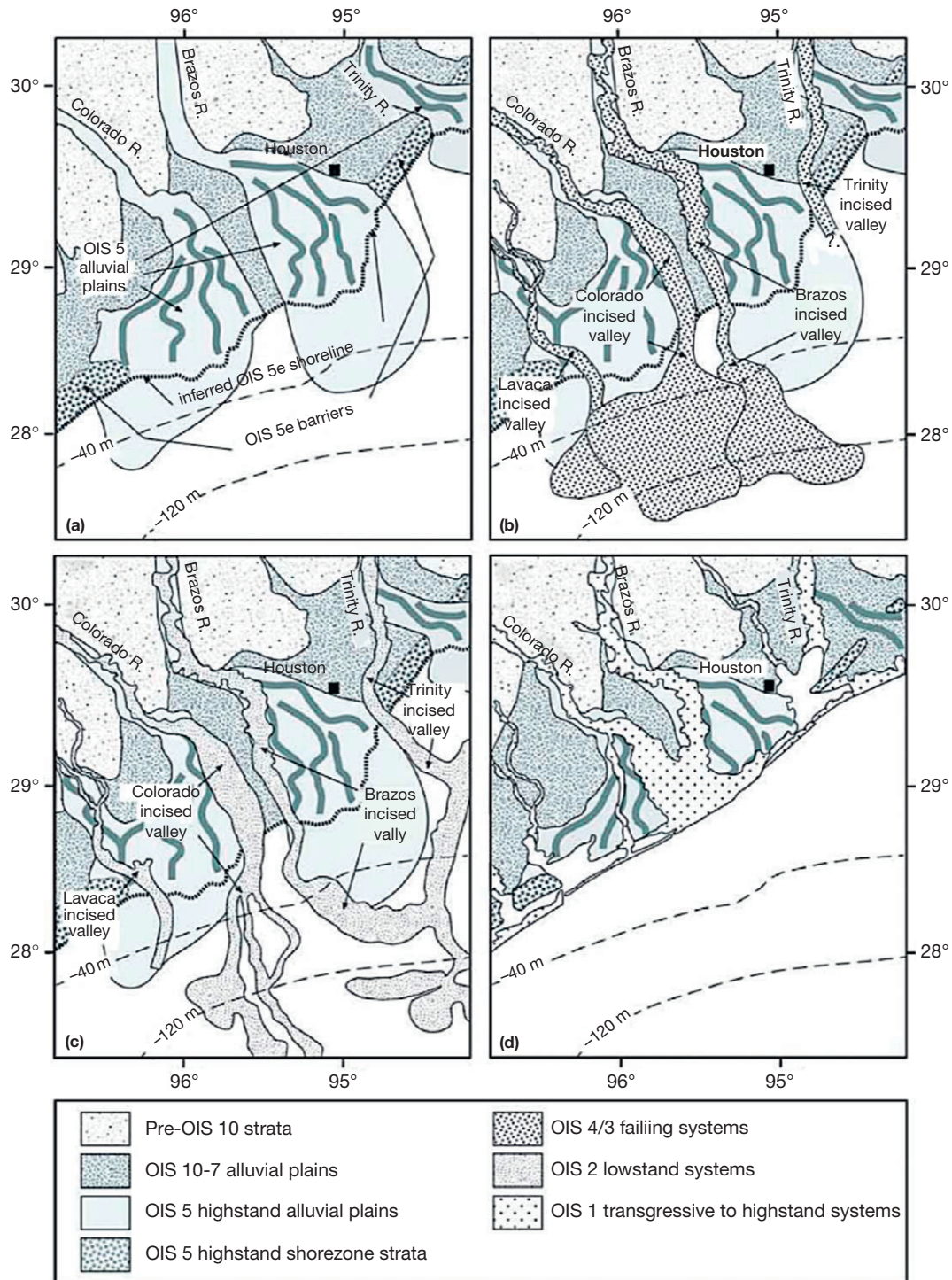


Figure 11 Map reconstructions of fluvial-deltaic systems of the US Gulf of Mexico coast and shelf in Texas. (a) Alluvial plains and deltas constructed during the last interglacial highstand, isotope stage 5. (b) Coastal-plain and cross-shelf incised valleys and deltas from the isotope stage 4–3 glacial period. (c) Coastal-plain and cross-shelf incised valleys and deltas from the isotope stage 2 full-glacial maximum lowstand. Note the merging of the Sabine, Trinity, and Brazos systems during the glacial period. (d) Holocene highstand alluvial-deltaic plains, estuaries, and barrier island systems. Adapted from Blum M and Aslan A (2006) Signatures of climate vs. sea-level change within incised valley fill successions: Quaternary examples from the Texas Gulf Coast. *Sedimentary Geology* 190: 177–211; Anderson JB, Rodriguez AB, Abdullah K, et al. (2004) Late Quaternary stratigraphic evolution of the northern Gulf of Mexico margin: A synthesis. In: Anderson JB and Fillon RH (eds.) *SEPM Special Publication 64: Late Quaternary Stratigraphic Evolution of the Northern Gulf of Mexico Margin*, pp. 1–23.

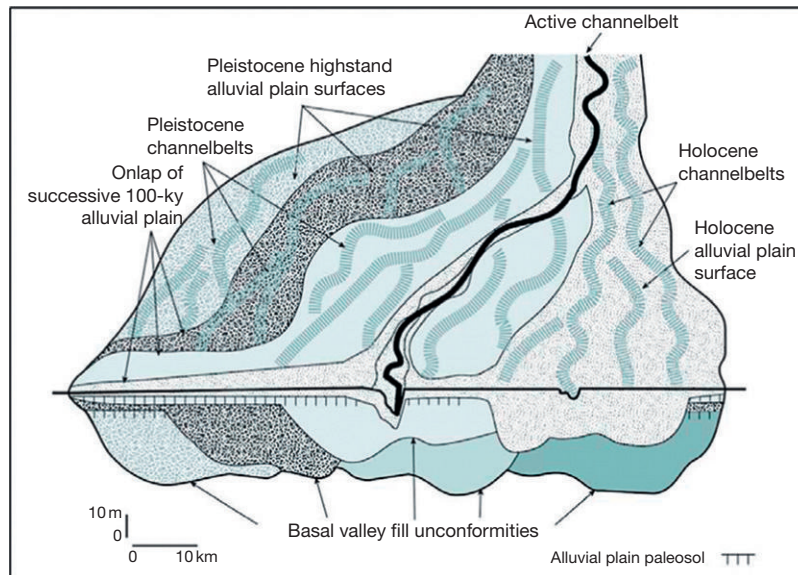


Figure 12 Schematic perspective view and schematic cross section of Pleistocene and Holocene alluvial plains of the Colorado River, Texas, illustrating large-scale stratigraphic architecture dominated by multiple cross-cutting paleovalley fills. Individual valley fills represent 100-ka glacial–interglacial cycles. The Holocene alluvial plain surface was recently abandoned when the channel avulsed to reoccupy an older Pleistocene meanderbelt. Modified from Blum MD and Törnqvist TE (2000) Fluvial response to climate and sea-level change: A review and look forward. *Sedimentology* 47(supplement): 1–48.

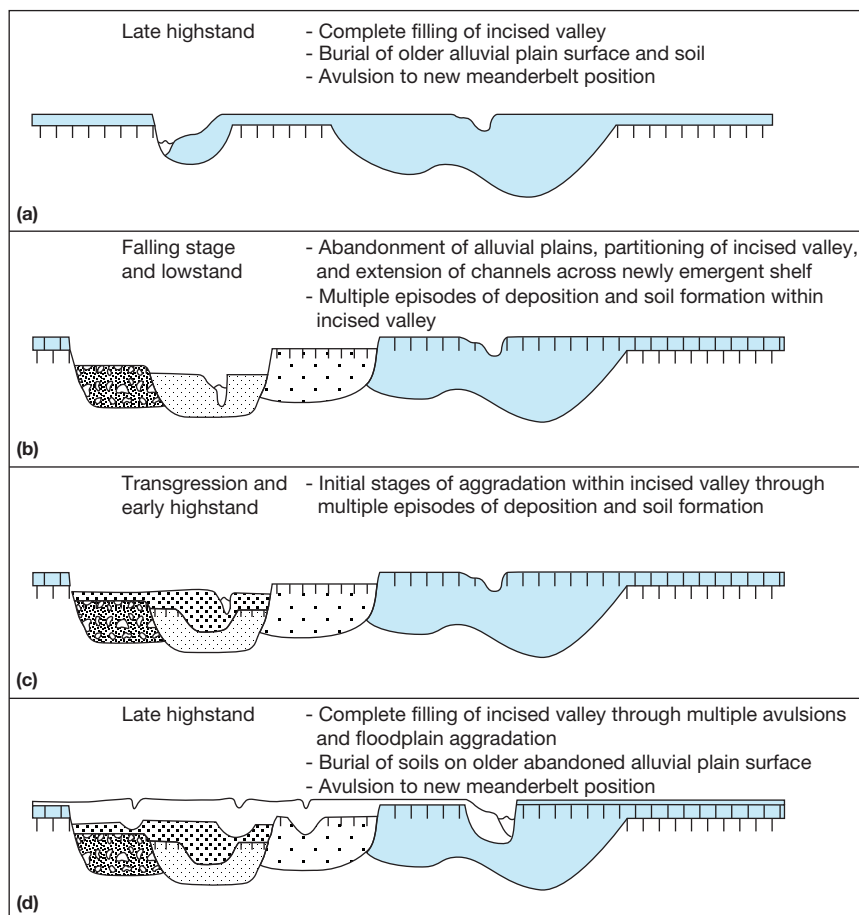


Figure 13 Schematic model for evolution of coastal-plain incised valley fills during a glacial–interglacial cycle, and development of alluvial plain successions. Reproduced from Blum MD and Price DM (1998) Quaternary alluvial plain construction in response to interacting glacioeustatic and climatic controls, Texas Gulf Coastal Plain. In: Shanley KW and McCabe PW (eds.) *SEPM Special Publication 59: Relative Role of Eustasy, Climate, and Tectonism in Continental Rocks*, pp. 31–48.

(Mulder and Syvitski, 1996). Given the nature of average climate and sea-level conditions during the Quaternary period, enlarged drainage basins such as these likely represent the norm, rather than the exception.

Conclusion

Quaternary glacial–interglacial cycles of climate and sea-level change have had a profound impact on fluvial systems. The record of this impact is spatially differentiated, such that continental interior systems reflect interactions between rates of uplift, and climate-driven unsteadiness in discharge regimes and sediment supply, whereas continental margin systems also reflect the influence of relative sea-level change. Studies of fluvial responses to glacial–interglacial cycles lag behind studies of climate and sea-level change per se, and studies in tectonic geomorphology, but nevertheless have a bright future. New geochronological methods and modeling approaches provide opportunities to more precisely define relationships between surface processes, landscapes, stratigraphic records, and glacial–interglacial cycles, quantify rates of change, and open up possibilities for developing true source-to-sink views on fluvial systems and landscape evolution.

See also: **Fluvial Environments:** Responses to Rapid Environmental Change; Sediments; Terrace Sequences. **Paleoceanography, Physical and Chemical Proxies:** Oxygen Isotope Composition of Seawater; Oxygen–Isotope Stratigraphy of the Oceans. **Sea Level Studies:** Eustatic Sea-Level Changes – Glacial–Interglacial Cycles; Eustatic Sea-Level Changes Since the Last Glacial Maximum.

References

- Allen GP and Posamentier HW (1993) Sequence stratigraphy and facies model of an incised valley fill: The Gironde Estuary, France. *Journal of Sedimentary Petrology* 63: 378–391.
- Anderson JB, Rodriguez AB, Abdullah K, et al. (2004) Late Quaternary stratigraphic evolution of the northern Gulf of Mexico margin: A synthesis. In: Anderson JB and Fillon RH (eds.) *Late Quaternary Stratigraphic Evolution of the Northern Gulf of Mexico Margin*, pp. 1–23. Tulsa, OK: SEPM. Special Publication 64.
- Antoine P (1994) The Somme valley terrace system (northern France): A model of river response to Quaternary climatic variations since 800,000 BP. *Terra Nova* 6: 453–464.
- Berendsen HJA and Stouthamer E (2001) *Palaeogeographic Development of the Rhine–Meuse Delta*. Assen, p. 268. The Netherlands: Van Gorcum.
- Blum M and Aslan A (2006) Signatures of climate vs. sea-level change within incised valley fill successions: Quaternary examples from the Texas Gulf Coast. *Sedimentary Geology* 190: 177–211.
- Blum MD and Price DM (1998) Quaternary alluvial plain construction in response to interacting glacioeustatic and climatic controls, Texas Gulf Coastal Plain. In: Shanley KW and McCabe PW (eds.) *Relative Role of Eustasy, Climate, and Tectonism in Continental Rocks*, pp. 31–48. Tulsa, OK: SEPM. Special Publication 59.
- Blum MD and Straffin EC (2001) Fluvial response to external forcing: Examples from the Massif Central of France, the Texas Coastal Plain (USA), the Sahara of Tunisia, and the Lower Mississippi Valley (USA). In: Maddy D and Macklin MA (eds.) *River Basin Sediment Systems: Archives of Environmental Change*, pp. 195–228. Lisse, The Netherlands: Balkema Press.
- Blum MD and Törnqvist TE (2000) Fluvial response to climate and sea-level change: A review and look forward. *Sedimentology* 47(supplement): 1–48.
- Bridge JS (2003) *Rivers and Floodplains: Forms, Processes, and the Sedimentary Record*, p. 491. Oxford: Blackwell.
- Bridgland D, Maddy D, and Bates M (2004) River terrace sequences: Templates for Quaternary geochronology and marine–terrestrial correlation. *Journal of Quaternary Science* 19: 203–218.
- Bull WB (1991) *Geomorphic Responses to Climate Change*. New York: Oxford University Press.
- Dury GH (1965) Principles of underfit streams. *United States Geological Survey Professional Paper 452-A*, Washington, pp. A1–A67.
- Ethridge FG, Wood LJ, and Schumm SA (1998) Cyclic variables controlling fluvial sequence development: Problems and perspectives. In: Shanley KW and McCabe PW (eds.) *Relative Role of Eustasy, Climate, and Tectonism in Continental Rocks*, pp. 17–29. Tulsa, OK: SEPM. Special Publication 59.
- Fisk HN (1944) *Geological Investigation of the Alluvial Valley of the Lower Mississippi River*, p. 78. Vicksburg, MS: Mississippi River Commission, United States Army Corps of Engineers.
- Hancock GS and Anderson RS (2002) Numerical modeling of fluvial strath-terrace formation in response to oscillating climate. *Geological Society of America Bulletin* 114: 1131–1142.
- Howard AD, Dietrich WE, and Seidl MA (1994) Modeling fluvial erosion on regional to continental scales. *Journal of Geophysical Research* 99: 13971–13986.
- Knox JC (1983) Responses of river systems to Holocene climates. In: Wright HE and Porter SC (eds.) *Late Quaternary Environments of the United States, Vol. 2: The Holocene*, pp. 26–41. Minneapolis: University of Minnesota Press.
- Lane EW (1955) The importance of fluvial morphology in hydraulic engineering. *American Society of Civil Engineering Proceedings* 81: 1–17.
- Leckie DA (1994) Canterbury Plains, New Zealand – Implications for sequence stratigraphic models. *American Association of Petroleum Geologists Bulletin* 78: 1240–1252.
- Leopold LB and Maddock T (1953) The Hydraulic Geometry of Stream Channels and Some Physiographic Implications. *United States Geological Survey Professional Paper 252*, Washington, p. 57.
- Mulder T and Syvitski JPM (1996) Climatic and morphologic relationships of rivers: Implications of sea-level fluctuations on river loads. *Journal of Geology* 104: 509–523.
- Page K, Nanson G, and Price D (1996) Chronology of Murrumbidgee River palaeochannels on the Riverine Plain, southeastern Australia. *Journal of Quaternary Science* 11: 311–326.
- Pazzaglia FJ and Brandon MT (2001) A fluvial record of long-term steady-state uplift and erosion across the Cascadia Forearc High, western Washington State. *American Journal of Science* 301: 385–431.
- Penck A and Brückner E (1909) *Die Alpen im Eiszeitalter*. Leipzig: Tauchnitz.
- Phillips FM, Zreda MG, Gosse JC, et al. (1997) Cosmogenic ^{36}Cl and ^{10}Be ages of Quaternary glacial and fluvial deposits of the Wind River Range, Wyoming. *Geological Society of America Bulletin* 109: 1453–1463.
- Porter SC (1989) Some geological implications of average Quaternary glacial conditions. *Quaternary Research* 32: 245–262.
- Porter SC, An ZS, and Zheng HB (1992) Cyclic Quaternary alluviation and terracing in a non-glaciated drainage basin on the north flank of the Qinling Shan, Central China. *Quaternary Research* 38: 157–169.
- Rittenour TM, Goble RJ, and Blum MD (2005) Development of an OSL chronology for Late Pleistocene channelbelts in the Lower Mississippi Valley, USA. *Quaternary Science Reviews* 24: 2539–2554.
- Saucier RT (1994) *Geomorphology and Quaternary Geologic History of the Lower Mississippi Valley*. Mississippi River Commission, p. 364. Vicksburg: Waterways Experiment Station, U.S. Army Corps of Engineers.
- Schumm SA (1968) *River Adjustment to Altered Hydrologic Regimen: The Murrumbidgee River and Palaeochannels, Australia*. United States Geological Survey Professional Paper 598.
- Schumm SA (1991) *To Interpret the Earth. Ten Ways to be Wrong*. Cambridge: Cambridge University Press.
- Starkel L (1991) Long-distance correlation of fluvial events in the temperate zone. In: Starkel L, Gregory KJ, and Thomas JB (eds.) *Temperate Palaeohydrology*, pp. 473–495. Chichester: Wiley.
- Suter JR and Berryhill HL (1985) Late Quaternary shelf-margin deltas, northwest Gulf of Mexico. *American Association of Petroleum Geologists Bulletin* 69: 77–91.
- Törnqvist TE, Wallinga J, and Busschers FS (2003) Timing of the last sequence boundary in a fluvial setting near the highstand shoreline – Insights from optical dating. *Geology* 31: 279–282.
- Zaitlin BA, Dalrymple RW, and Boyd R (1994) The stratigraphic organization of incised-valley systems associated with relative sea-level change. In: Dalrymple RW, Boyd R, and Zaitlin BA (eds.) *Incised-Valley Systems: Origins and Sedimentary Sequences*, pp. 45–60. Tulsa, OK: SEPM. Special Publication 51.

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GLACIATION, CAUSES

Contents

Tectonic Uplift and Continental Configurations

Astronomical Theory of Paleoclimates

Tectonic Uplift and Continental Configurations

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Introduction

Throughout Earth's history, climate has oscillated from times of extreme warmth with ice-free poles (greenhouse times) to times of extreme cold with massive continental ice sheets and polar ice caps (icehouse times). Lasting for many tens of millions of years, these icehouse and greenhouse times were separated by many tens of millions of years of transitional climate. Throughout the past ~620 Ma, Earth has experienced icehouse conditions during (1) the late Vendian, the 'snowball Earth' of Hoffman et al. (1998); (2) the late Ordovician; (3) the early Carboniferous to early Permian; and (4) the late Cenozoic, since ~36.5 Ma (Figure 1). Climate oscillates at frequencies ranging from 10^0 to 10^5 years during icehouse times. As a consequence, there are times of much more extensive glaciation (glacials), lasting $\sim 10^5$ years, and of significantly less glaciation (interglacials), lasting $\sim 10^4$ years. Shorter-duration oscillations of varying climate and glacial extent (stadial and interstadials) occur on a 10^3 – 10^4 -year timescale and these help characterize glacial times. Even shorter-duration climate oscillations on the centennial, decadal, and annual timescales occur in both glacial and interglacial times. Such climatic oscillations are to be expected because the primary forces that drive climate change, which include Earth's orbital parameters, plate tectonics, solar variability, and autocyclicity within the ocean's atmosphere and cryosphere, are also in perpetual motion. Determining and quantifying the relative importance of different forcing factors in driving climate change is a challenging task and has been a major focus of debate over the last few decades.

There is broad agreement that periodic and quasiperiodic oscillations in Earth's orbital parameters of eccentricity, obliquity and precession, which affect the distribution and amount of incident solar energy, are primarily responsible for climate change on a 10^4 – 10^5 -year timescale. The relative roles of the forcing factors that result in higher frequency climate change (10^0 – 10^4 years) include changes in sunspot activity and autocyclicity within the ocean–atmosphere–cryosphere systems. This article focuses on the forcing factors that drive

climate change on timescales that are very much greater than 10^5 years.

The usual culprits of climate change on timescales much greater than 10^5 years include plate tectonic and associated changes in biogeochemical cycles. The plate-tectonic mechanisms that help drive climate change are complex. These include changes in (1) volcanism and degassing of Earth's interior; (2) rates of seafloor spreading and the formation of oceanic crust with the associated sea level changes; (3) mountain building and subduction resulting in changing rates of metamorphism and weathering, leading to changes in the concentration of atmospheric CO_2 ; (4) the latitudinal distribution and sizes of the continents; and (5) oceanic circulation resulting from opening and closing of land bridges and seaways. In turn, each of these sets of changes affects the biogeochemical cycles, which leads to changes in the composition of the oceans, the atmosphere and the biosphere that help force climate change.

Role of Tectonics on Climate Change

The debate over the role of plate-tectonic processes in driving climate change is comparatively recent. This is because plate-tectonic theories began to be formulated only in the 1960s and 1970s, and it is only during the last few decades that detailed quantitative climatic records have been developed, most significantly as a consequence of the greater availability of high-quality sediment cores recovered by the Deep Sea Drilling Program (DSDP) and Ocean Drilling Program (ODP) and improved geochemical methods. Nevertheless, hypotheses invoking uplift effects on atmospheric circulation have a long history. Notable examples include discussions in papers by Dana (1856), Lyell (1875), and Ramsay (1924), which argued that uplift of mountain regions such as the European Alps or Apennines helped drive large-scale climate change. These regions are now known to have had at the most localized cooling effects and no larger-scale impact. Later researchers such as Flint (1957), Emiliani and Geiss (1959), and Hamilton

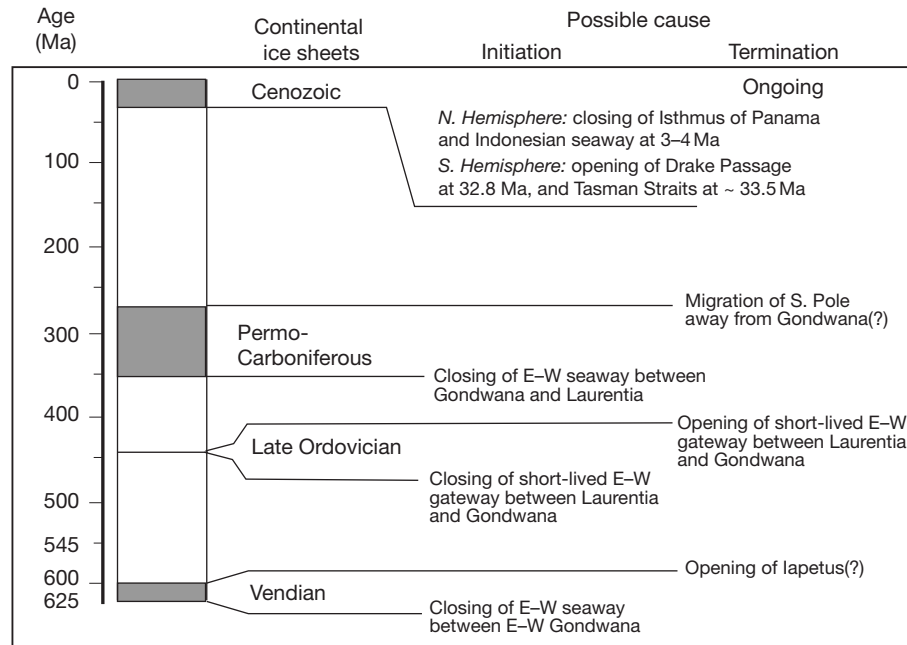


Figure 1 The durations of the main continental ice sheets for the past 620 million years. Reproduced from Smith AG and Pickering KT (2003) Oceanic gateways as a critical factor to initiate icehouse Earth. *Journal of the Geological Society, London* 160: 337-340. The possible causes for their initiation and termination are shown.

(1968) focused mainly on smaller scale uplift in high-latitude regions to suggest a potential mechanism for the initiation of the Pleistocene continental ice sheets.

Since the 1980s, however, more attention has been given to the larger scale (global) impact of tectonics on climate change, mainly in regard to the initiation of glaciation. Of note was the work by Birchfield and Weertman (1983) who argued that the geologically youthful uplift of the Tibetan Plateau had the potential to increase snow cover and enhance global cooling. Eyles (1993) followed with his seminal work on Earth's glacial record and its tectonic setting.

BLAG Hypothesis

The first significant modern analysis of the tectonic influences on climate suggested that the driving forces for icehouse and greenhouse conditions were the result of changes in global atmospheric CO₂ concentrations associated with metamorphism and volcanic degassing along subduction zones and mid-oceanic ridges (Berner et al., 1983). Berner et al.'s (1983) proposition became known as the BLAG hypothesis (an acronym based on the initials of the authors). Together with the changing latitudinal positions of continents, the BLAG hypothesis became the leading explanation for the evolution of icehouse and greenhouse conditions until new ideas emerged in the early 1990s.

Plateau Uplift and Climate Change

Suggestions that uplift of high mountain ranges and plateaus might be more important in driving climate challenged the

BLAG hypothesis (Raymo and Ruddiman, 1992; Raymo et al., 1988). The debate emerged mainly from studies of detailed stable isotopic records, largely $\delta^{18}\text{O}$, derived from the analysis of Cenozoic deep-sea sediments. These records suggested progressive cooling, together with the realization that high plateaus, such as Tibet and the American West, had progressively uplifted during the same period. In essence, it was argued that Cenozoic uplift of plateaus changed atmospheric circulations and atmospheric concentrations of greenhouse gases (Ruddiman, 1997). The discussion centered around the effects that a rising plateau would have on the blocking of regional air systems, and changes in the position of jet streams and air masses.

In the case of the Tibetan Plateau, it was suggested that plateau uplift would have forced cold polar air into key locations in the Northern Hemisphere, particularly North America and north-west Europe, which are the sites of ice sheet development (Ruddiman, 1997, and references therein). The elevated region would have enhanced temperature-driven atmospheric flows as higher and lower atmospheric pressures developed over the plateau during the winter and summer, respectively, creating and intensifying the Asian monsoon and the aridity across the Tibetan Plateau and central Asia. The increased aridity in Central Asia would have led to a higher annual and daily temperature ranges. In addition, the elevated Tibetan Plateau would have cooled as it rose, allowing glaciers to develop, increasing albedo, and leading to further cooling. Furthermore, the greater availability of newly exposed rock surfaces and detritus produced by denudation processes would have enhanced chemical weathering, which in turn would have drawn down atmospheric CO₂. It was argued that all these processes contributed to cooling, leading to icehouse conditions and ultimately the onset of the present Ice Age.

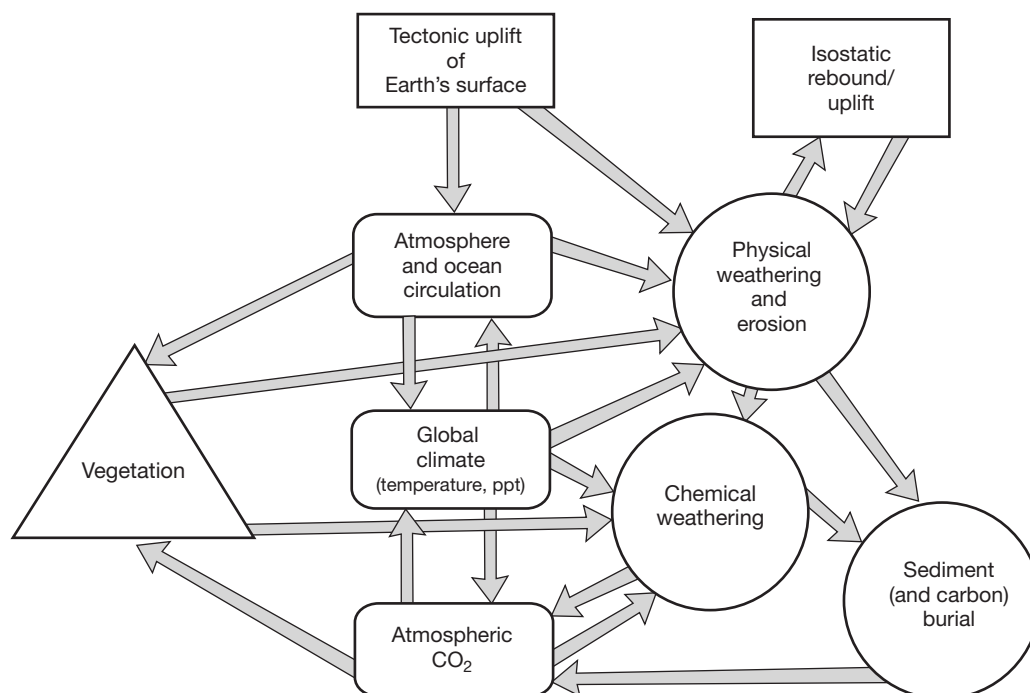


Figure 2 Potential interactions and feedbacks between tectonic uplift and Earth's climatic and environmental system. Reproduced from Ruddiman WF and Prell WL (1997) Introduction to uplift–climate connection. In Ruddiman WF (ed.) *Tectonic Uplift and Climate Change*, pp. 1–15. New York: Plenum Press. Rectangles indicate parts of Earth's system related to tectonic (uplift); rounded rectangles indicate parts of Earth's system related to atmospheric and oceanic circulation and climate; circles indicate parts of Earth's system related to weathering and erosion; and triangles indicate the vegetational component of Earth's system.

The relationship between plateau uplift and climate change is complex and involves multiple feedback mechanisms, as is schematically illustrated in [Figure 2](#). Despite the elegant models and plausible links between plateau uplift and climate change, there is considerable uncertainty as to whether the proposed mechanisms can explain progressive cooling throughout the Cenozoic and onset of the present Ice Age. This is partially because the timing of uplift of the Tibetan Plateau, and other plateaus such as in the American West, is poorly defined. In addition, the complex feedbacks are not fully understood and the rates of weathering and drawdown of atmospheric CO_2 are not well defined. Furthermore, there were many other tectonic events occurring elsewhere on the globe during the Cenozoic that could have helped drive climate change. These included North Atlantic rift volcanism; opening and widening of two Antarctic gateways, the Tasmanian and Drake Passages ([Carter et al., 1996](#); [Latimer and Filipelli, 2002](#)); uplift of Panama and the closure of the Central American Seaway ([Roth et al., 2000](#)); and changes in the topography of the Greenland-Scotland Ridge ([Wright and Miller, 1996](#)).

Complexities of the Cenozoic Cooling and Tectonics

The stepped Cenozoic cooling trend, based on the composite marine oxygen isotope curves, the timing and extent of glaciation, major biotic events, and a simplified history of the major Cenozoic tectonic events, is shown in [Figure 3](#). The progressive decrease in $\delta^{13}\text{C}$ (a measure of the ratio of ^{13}C and ^{12}C), based

on marine carbonates, throughout the Cenozoic reflects changes in the size of the organic carbon reservoir, representing a decrease in the burial of organic matter with time. This change in burial of organic matter likely caused additional CO_2 to be released into the oceans and the atmosphere ([Raymo and Ruddiman, 1992](#)). Oxygen is necessary to oxidize organic matter and results in the release of CO_2 . The decrease in buried organic carbon, therefore, possibly reflects an increase in dissolved oxygen concentrations in the oceans, which is necessary to oxidize organic matter. Oxygen solubility increases with decreasing temperatures and hence dissolution of oxygen into the oceans would, therefore, have increased as the oceans began to cool. This in turn, would have increased the oxidation of organic matter and led to a greater release of CO_2 into the atmosphere. As the residual time of CO_2 is <1 Ma and O_2 is ~ 10 Ma in the atmosphere, this would not lead to a perfect balance between these processes. Rather the balance of the organic subcycle would become less effective with time, leading to a more intense cooling in the late Cenozoic as atmospheric CO_2 becomes progressively depleted.

Throughout the early twenty-first century, new methods devised to reconstruct the CO_2 concentration of the atmosphere on longer time scales began to test the hypothesis that a decrease in CO_2 was driving Cenozoic climate change. Of note is the work by [Pearson and Palmer \(2000\)](#) who used boron-isotope ratios in Cenozoic planktonic Foraminifera shells to estimate the pH of surface-layer seawater, which they used to reconstruct atmospheric CO_2 concentrations. In addition, [Pagani et al. \(2005\)](#) presented results on stable carbon

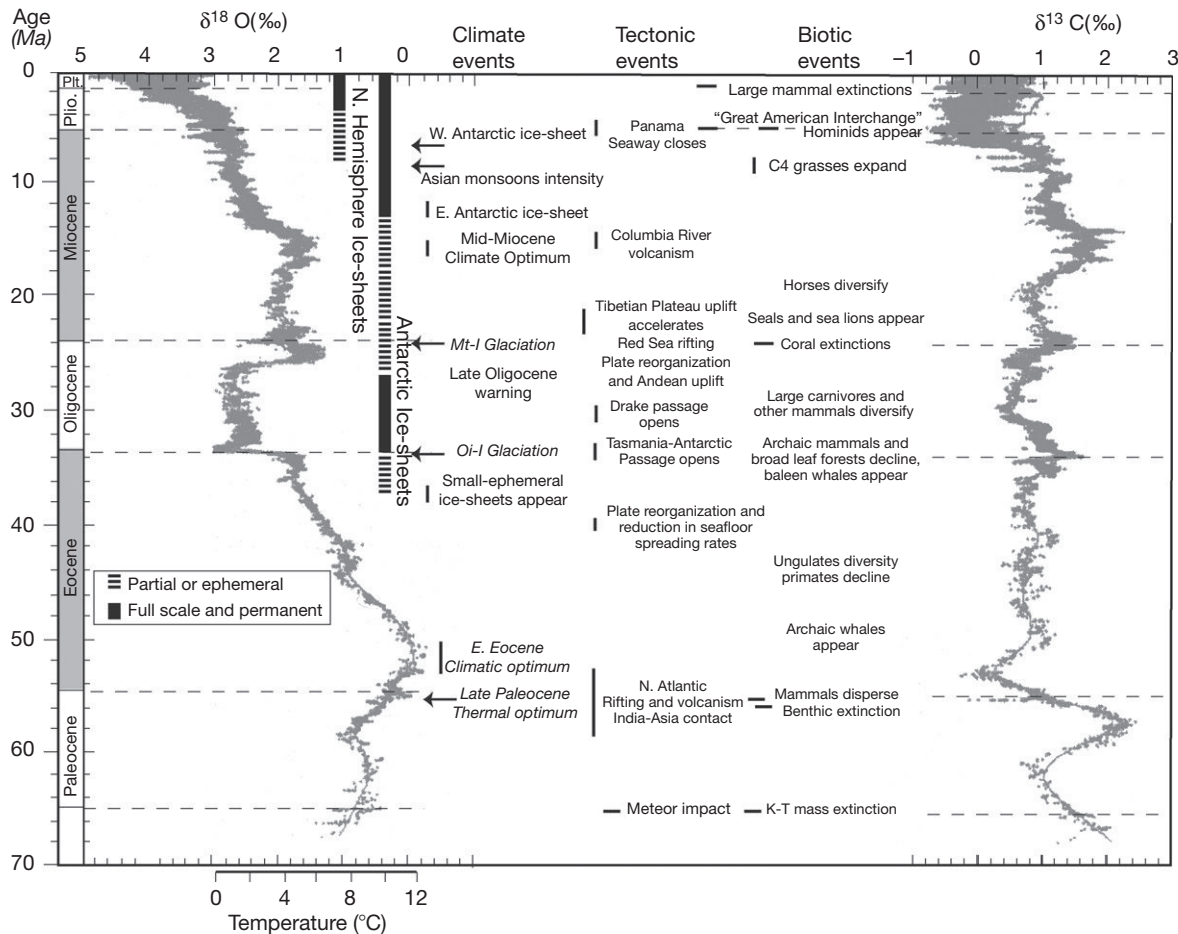


Figure 3 Global deep-sea oxygen and carbon isotope records based on data compiled from fine-grained carbonate-rich deep-sea sediments from more than 40 Deep Sea Drilling Program and Ocean Drilling Program sites. These are compared with the major climatic, tectonic, and biotic events. Adapted from Zachos J, Pagani M, Sloan L, Thomas E, and Billups K (2001) Trends, rhythms, and aberrations in global climate 65 Ma to present. *Science* 292: 686–693. The vertical bars provide a rough qualitative representation of volume in each hemisphere relative to the Last Glacial Maximum, with dashed bars representing periods of minimal ice ($\leq 50\%$) and full bars representing periods close to maximum ice coverage ($>50\%$ of present). The temperature scale is for bottom ocean-water temperatures based on the $\delta^{18}\text{O}$ values, but is only valid for the Cretaceous to the end of the Eocene; thereafter, the $\delta^{18}\text{O}$ values reflect a combined effect of ice volume and temperature.

isotopic values of di-unsaturated alkenones extracted from deep-sea cores to reconstruct atmospheric CO_2 concentrations from the middle Eocene to the late Oligocene ($\sim 45\text{--}25$ Ma).

The results from both methods suggested estimated atmospheric CO_2 concentrations of ~ 1000 ppm or more during much warmer climates tens of millions of years ago, compared with ice-core values of just 180–300 ppm during glacial times over the past 800 ka (Figure 4). As Ruddiman (2010) has pointed out, this long-term decrease in atmospheric CO_2 concentrations lends credence to the idea that CO_2 has been the long-term driver of global cooling; however, a closer look reveals major problems. By 22 Ma, for example, the alkenone and boron-isotope data both showed that estimated atmospheric CO_2 concentrations were already within the range typical of glacial times during the past 800 ka. Therefore, if CO_2 concentrations of 180–300 ppm have played an integral role in driving glacial cycles in the past 800 ka, one would have expected glacial cycles to have been initiated at ~ 22 Ma, rather than at ~ 2.8 Ma as suggested by the geologic evidence.

Moreover, if the average atmospheric CO_2 trend has not fallen in the past 22 Ma, then an alternative explanation is required to account for the substantial bipolar cooling during that time.

Ruddiman (2010) pointed out that the disparity between low atmospheric CO_2 and the cooling trend in the past 22 Ma could be the result of (1) something crucial having simply been overlooked; (2) one or more physical mechanisms previously proposed to account for global cooling having been much stronger than previously considered, for example, mountain uplift and/or northward movement of plates; and (3) the proxy methods to reconstruct atmospheric CO_2 concentrations having become invalid.

Ruddiman (2010) argued that the alkenone and boron-isotope proxies used between 45 and 25 Ma disagree by large amounts. For example, atmospheric CO_2 concentrations based on boron isotopes fall to very low values before those based on alkenone data. In addition, Ruddiman (2010) pointed out that there is disagreement with climate model stimulations, which suggest that atmospheric CO_2 concentrations below ~ 750 ppm

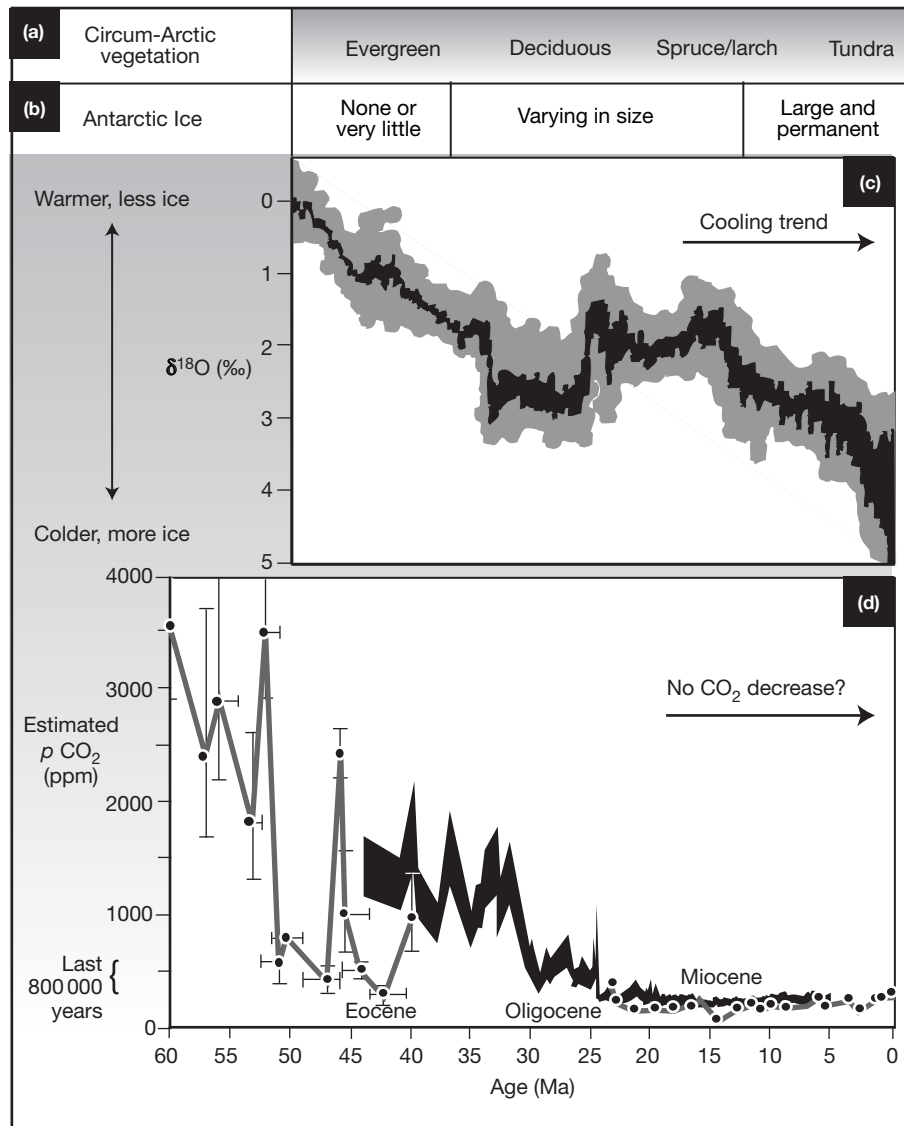


Figure 4 Indices of climate change in the past 50 Ma. Adapted from Ruddiman WF (2010) A paleoclimatic enigma? *Science* 328: 838–839. (a) Changes in circum-Arctic vegetation. (b) Changes in Antarctica ice sheet size. (c) Oxygen isotopic ($\delta^{18}\text{O}$) index of deep-ocean temperature and ice volume. (d) Estimates of past atmospheric CO_2 concentrations from alkenones (black sawtooth; data from Pagani et al., 2005) and boron-isotope ratios in Cenozoic planktonic Foraminifera shells (black circles and gray line; data from Pearson and Palmer, 2000). Paleoclimatic data from both poles indicate progressive cooling from ~50 Ma until the onset of Northern Hemisphere glaciation ~2.8 Ma, but reconstruction of atmospheric CO_2 concentrations from two prominent proxy models show very low values by 25 Ma and no subsequent decrease.

(based on boron isotopes) should cause an ice sheet to grow on Antarctica by 55–50 Ma, whereas alkenone data suggest that an ice sheet should have formed by 30 Ma, a few million years after the geologic evidence showed it to have formed. Moreover, new data presented by Tripathi et al. (2009) using B/Ca ratio predict atmospheric CO_2 concentrations of 350–450 ppm between 20 and 10 Ma (when alkenone and boron isotopes suggest 200–300 ppm). Ruddiman (2010) argued that the new studies suggest that atmospheric CO_2 did indeed decrease to drive climate and played a role in northern glacial inception, contrary to what alkenone and B-isotope proxy data suggest. Clearly, the debate is still ongoing and future studies will inevitably shed new light on this topic.

Molnar and England (1990) and England and Molnar (1990) challenged the uplift-climate hypotheses by proposing that Cenozoic cooling and increased glaciation drove mountain uplift. They argued that increased glaciation resulting in enhanced erosion results in denudational unloading and accelerated uplift of residual high mountain peaks. While this process may provide an alternative or complementary mechanism for uplift, geologic data suggest that much of the Himalayan-Tibetan orogen obtained its height before the onset of major glaciation. Nevertheless, this mechanism might be important in driving uplift in some mountain ranges.

A further twist on the role of tectonics and weathering on driving climate is provided by Willenbring and von

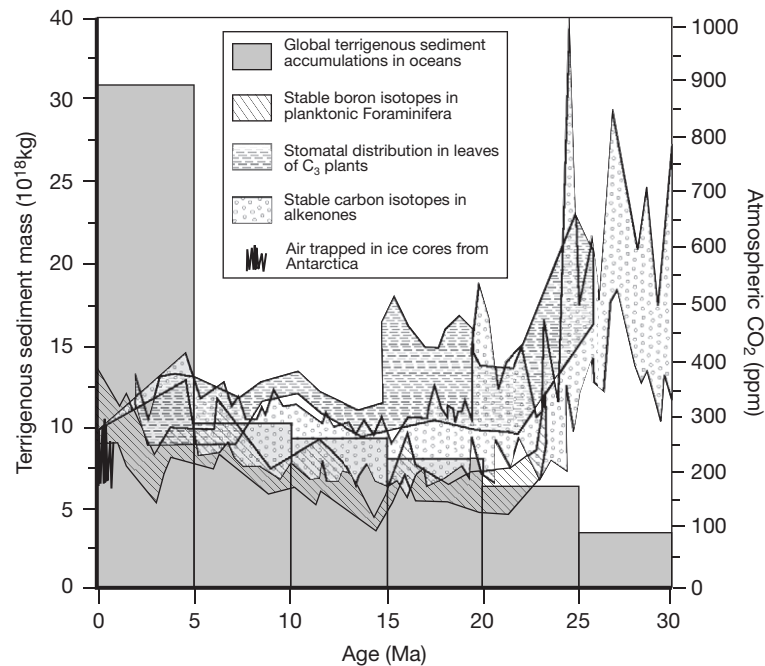


Figure 5 Terrigenous sediment input into the oceans through the late-Cenozoic era and atmospheric CO₂. Reproduced from Willenbring JK and von Blanckenburg F (2010) Long-term stability of global erosion rates and weathering during late-Cenozoic cooling. *Nature* 465: 211–214. Each band for CO₂ values spans the associated data points and their uncertainties. The values for terrigenous sediment accumulation in the oceans are separated into bins with an interval of 5 Ma and appear to increase abruptly during the past 5 Ma. The CO₂ data compilation is derived from a number of independent proxies and shows steady atmospheric concentrations from the mid-Miocene epoch to the present (preindustrial levels) despite the observed increase in terrigenous sediment accumulation.

Blanckenburg (2010). Until their paper, it was generally considered that there has been a fourfold increase in global sedimentation rates since ~5 Ma, a proxy for increased uplift and weathering resulting in a drawdown of atmospheric CO₂ (Figure 5). Using the ocean dissolved ¹⁰Be/⁹Be isotope system as a weathering proxy spanning the past 12 Ma, Willenbring and von Blanckenburg (2010) questioned whether this increase in global weathering and erosion actually occurred and whether the apparent increase in the sedimentation rate is due to observational biases in the sedimentary record. They argued that weathering fluxes during the late Cenozoic have been stable and that data show no clear evidence for increased erosion or any clear evidence for a pulse in weathered material into the ocean. Moreover, Willenbring and von Blanckenburg (2010) concluded that processes different from an increase in denudation caused Cenozoic global cooling, and furthermore, global cooling had no profound effect on spatially and temporally averaged weathering rates.

Matching tectonic events to global climate change is not simple. Stable marine isotope data show, for example, that Cenozoic climate change was not a gradual cooling trend, but was stepped and punctuated by periods of warming culminating in climatic optimums, which include the early Eocene Climatic Optimum, the late Oligocene Warming, and the Mid-Miocene Climatic Optimum (Figure 3). Furthermore, high-resolution δ¹⁸O data show aberrations superimposed on a long-term gradual trend that are well above the normal background variability, which last for ~10³–10⁵ years (Zachos et al., 2001). These aberrations represent changes in sea surface temperature by as much as 8 °C (Zachos et al., 2001). Carbon

isotope data suggest that the aberrations are usually accompanied by major perturbations in the global carbon cycle. The three largest ones occurred near epoch boundaries at ~55 Ma (late Paleocene Thermal Maximum aberration), 34 Ma (Oi-1 event), and 23 Ma (Mi-1 event). Changes near epoch boundaries suggest that these aberrations had a major impact on the biosphere; a point supported by accelerated rates of turnover and speciation in certain taxa groups at these times (Zachos et al., 2001). The Oi-1 and Mi-1 events are coincident with brief extremes in Antarctic ice volume and temperature, which are referred to as the Oi-1 and Mi-1 glaciations, respectively (Figure 3). The driving mechanisms for the aberrations are not known, but suggestions to explain aberrations such as the late Paleocene Thermal Maximum include the sudden release of methane from marine clathrates (compounds consisting of a lattice of water molecules with gas molecules occupying holes/cavities within the lattice; Dickens et al., 1997), abrupt deep-sea warming resulting from sudden changes in ocean circulation (Bralower et al., 1997), massive slope failure (Katz et al., 1999), and rapid release of greenhouse gases generated by heating of organic-rich sediments around intrusions in the NE Atlantic (Svensen et al., 2010). In addition, Zachos et al. (2001) suggest possible tectonic factors to explain the Oi-1 and Mi-1 events.

Drifting Continents and Oceanic Gateways

Smith and Pickering (2003) proposed a unifying explanation for the four major icehouses, which involves two sets of plate-

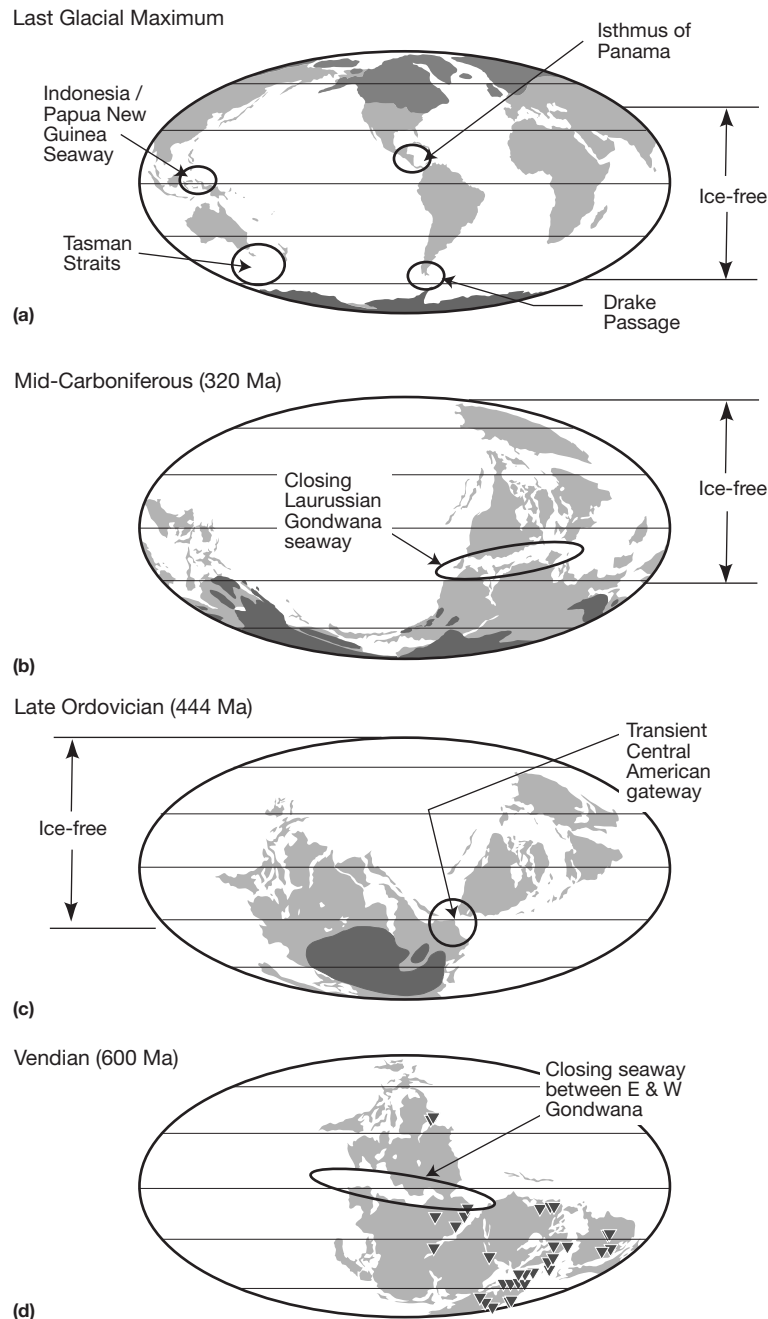


Figure 6 Maps showing the positions of the continents and oceans during icehouse times over the last 620 million years. Reproduced from Smith AG and Pickering KT (2003) Oceanic gateways as a critical factor to initiate icehouse Earth. *Journal of the Geological Society, London* 160: 337–340.

(a) The present-day world showing high-latitude gateways that have opened (the Drake Passage and the Tasman Straits) and the equatorial gateways that have closed (Isthmus of Panama and the seaway between Indonesia and Papua New Guinea) during the Cenozoic. The dark grey areas are the estimates of ice cover during the Last Glacial Maximum. (b) Reconstruction for the mid-Carboniferous time showing the major seaway between Laurussia and Gondwana, which closed when the supercontinents collided to form Pangaea. The dark grey areas are estimates of the ice cover for the entire Permo-Carboniferous glaciation. (c) Reconstruction for the Late Ordovician showing the estimates of ice cover (in dark grey). If Laurentia and Gondwana are brought into contact by sliding one or other supercontinent along paleolatitude lines, then NW South America joins Central America. The continents were farther apart before and after this time suggesting that an open gateway could have existed here just before and just after 444 Ma. The closure at 444 Ma might have caused effects analogous to the closing of the Isthmus of Panama changing the oceanic circulation in such a way as to initiate an ice cap, which melted once the gateway opened a million years later. (d) The Vendian map at 600 Ma. The collision of West and East Gondwana is one of the main tectonic events at this time, which eliminated a major east trending tropical seaway and may have initiated the formation of ice sheets as suggested by the Permo-Carboniferous ice age. The inverted triangles are locations of all glaciogenic deposits on Gondwana, Laurentia, Baltica, and Siberia that have been assigned to the younger Neoproterozoic.

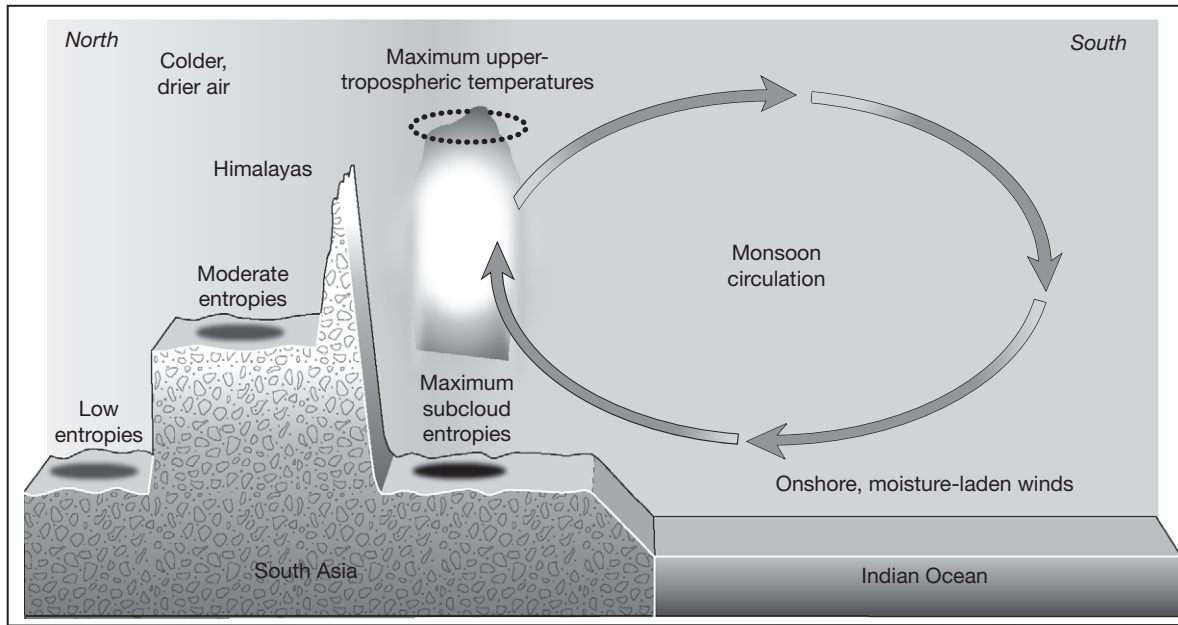


Figure 7 Schematic diagram showing Boos and Kuang's (2010) model for the role of moisture convection rather than heat absorbed and radiated by the Tibetan Plateau. Maximum subcloud moist entropy occurs south of the Himalayas, and the heat released as water vapor rises and condenses and is reflected in peak temperatures in the overlying upper troposphere. The Himalayas keep the moist warm air over South Asia separated from the colder, drier air to the north, so the high energy of this air mass is undiluted, remains favorable for moisture-driven convection, and underlies the strength of the South Asian monsoon. Reproduced from Cane MA (2010) A moist model monsoon. *Nature* 463: 163–164.

tectonic processes. The first set involved the movement of some continents into polar latitudes, while the second set involved the opening and closing of oceanic low latitude gateways. The latter set of tectonic processes changed global oceanic circulation from one with important circum-equatorial currents (greenhouse times) to one with inhibited circum-equatorial deepwater currents (icehouse times) and more restricted oceanic gyres. Smith and Pickering (2003) stressed that all four icehouse times occurred when one or both of the geographic poles lay within or less than 1000 km from a large continent. However, there have been periods when a polar continent, such as Antarctica during much of the Cretaceous greenhouse world, was at low latitudes but was not glaciated. Thus, Smith and Pickering (2003) argued that a polar or subpolar position for a continent appears to be a necessary condition, but is not a sufficient precondition for widespread glaciation.

Smith and Pickering (2003) suggested that during icehouse times, the distribution of continents inhibited circum-equatorial circulation, forcing faster oceanic circulation in the main ocean gyres. Whether increased circulation increases snow and ice accumulation in high-latitude regions depends in part on the strength of the contemporaneous circum-polar circulation. Smith and Pickering (2003) stressed that the problem is complex, but once oceanic circulation has changed, then all the isotopic effects follow. Only after the continental configurations, gateways, and associated ocean gyres are established do CO₂ concentrations, Milankovitch-forcing, and other processes become significant in driving glacial–interglacial events. Smith and Pickering (2003) also acknowledged that high topography may be an important factor in driving climate change, but do not attempt to quantify its impact. The major tectonic factors that are likely responsible for each of the four

icehouse times are summarized in Figure 1 and illustrated in Figure 6.

General Circulation Modeling

In recent years, much effort has focused on examining the complex links between tectonics and climate change using general circulation models (GCMs) (Cliff et al., 2010). GCMs involve simulation of atmospheric circulation using high-speed computers or supercomputers to linking a system of equations used to describe atmospheric and ocean-water motion, the heat exchange and fluxes within this system, and the consequences. GCMs allow inputs such as the configurations and shapes of the oceans, continents, and mountains to be varied to compute the possible resultant changes in climate. Early GCMs convincingly illustrated the effects of plateau uplift on climate and they helped initiate much research and debate (e.g., Ruddiman and Kutzbach, 1991). Subsequent models, such as those by Sloan et al. (2001) and DeConto and Pollard (2003) were more sophisticated and were able to examine detailed changes in ice sheet growth resulting from tectonic and climatic change. Of particular note is the recent work by Boos and Kuang (2010) on the uplift of the Himalayan–Tibetan region and the development of the South Asian monsoon. Their computer simulation questioned the generally accepted idea that the prime reason for the great strength of the South Asian monsoon is the heating effect of the Tibetan Plateau. Their model showed that the barrier posed by the mountains at the southern edge of the plateau is the driving factor for the monsoon, not the heating effect of the Tibetan Plateau (Figure 7). Future development of GCMs will clearly continue to shed light on the complex links and feedbacks

between tectonics and climate, and will probably initiate many new questions.

Conclusion

To summarize, long-term global climate change ($\gg 10^5$ years) is most likely driven by tectonic processes, notably, changes in the distribution and the configurations of continents, and their effect on oceanic circulation and biogeochemical cycles. The relative importance of various factors, such as plateau uplift, volcanism, and degassing and metamorphism, and their effects on biogeochemical cycles has yet to be adequately quantified. Moreover, defining the rates, magnitude, and nature of climate change and tectonic processes is a challenging proposition, and it is thus difficult to correlate between the tectonics and climate to test causal links and feedbacks. Nevertheless, new developments in analytical methods, especially with the development of high-precision geochronology and geochemical technologies, are beginning to allow the links and feedbacks between tectonics and climate to be tested at greater temporal detail and range.

See also: **Glaciation, Causes:** Astronomical Theory of Paleoclimates. **Paleoceanography, Physical and Chemical Proxies:** Carbon Cycle Proxies ($\delta^{11}\text{B}$, $\delta^{13}\text{C}_{\text{calcite}}$, $\delta^{13}\text{C}_{\text{organic}}$, Shell Weights, B/Ca, U/Ca, Zn/Ca, Ba/Ca); Oxygen-Isotope Stratigraphy of the Oceans. **Paleoclimate Reconstruction:** Sub-Milankovitch (DO/Heinrich) Events

References

- Berner RA, Lasaga AC, and Garrels RM (1983) The carbonate–silicate geochemical cycle and its effect on atmospheric carbon-dioxide over the past 100 million years. *American Journal of Science* 283: 649–683.
- Birchfield GE and Weertman J (1983) Topography, albedo-temperature feedback, and climate sensitivity. *Science* 219: 284–285.
- Boos WR and Kuang Z (2010) Dominant control of the South Asian monsoon by orographic insulation versus plateau heating. *Nature* 463: 218–222.
- Bralower TJ, Thomas DJ, Zachos JC, et al. (1997) High-resolution records of the late Paleocene thermal maximum and circum-Caribbean volcanism: Is there a causal link? *Geology* 25: 963–966.
- Cane MA (2010) A moist model monsoon. *Nature* 463: 163–164.
- Carter RM, Carter L, and McCave IN (1996) Current controlled sediment deposition from the shelf to the deep ocean: The Cenozoic evolution of circulation through the SW Pacific gateway. *Geologisches Rundschau* 85: 438–451.
- Clift PD, Tada R, and Zheng H (2010) *Monsoon Evolution and Tectonic-Climate Linkage in Asia*, vol. 342, p. 312. London: Geological Society Special Publication.
- Dana JD (1856) American geological history. *Proceedings of the American Association for the Advancement of Science* 22: 305–334.
- DeConto RM and Pollard D (2003) Rapid Cenozoic glaciation of Antarctica induced by declining atmospheric CO_2 . *Nature* 421: 245–249.
- Dickens GR, Castillo MM, and Walker JCG (1997) A blast of gas in the latest Paleocene: Simulating first-order effects of massive dissociation of oceanic methane hydrate. *Geology* 25: 259–262.
- Emiliani C and Geiss J (1959) On glaciations and their causes. *Geologische Rundschau* 46: 576–601.
- England P and Molnar P (1990) Surface uplift, uplift of rocks, and exhumation of rocks. *Geology* 18: 1173–1177.
- Eyles N (1993) Earth's glacial record and its tectonic setting. *Earth-Science Reviews* 35: 248.
- Flint RF (1957) *Glacial and Pleistocene Geology*. New York: Wiley.
- Hamilton W (1968) Cenozoic climatic change and its cause. *Meteorological Monographs* 8: 128–133.
- Hoffman PF, Kaufman AJ, Halverson GP, and Schrag DP (1998) A neoproterozoic snowball Earth. *Science* 281: 1342–1346.
- Katz ME, Pak DK, Dickens GR, and Miller KG (1999) The source and fate of massive carbon input during the latest Paleocene thermal maximum. *Science* 286: 1531–1533.
- Latimer JC and Filipelli GM (2002) Eocene to Miocene terrigenous inputs and export production; geochemical evidence from ODP Leg 177, Site 190. *Paleogeography, Paleoclimatology, Paleoecology* 182: 151–164.
- Lyell C (1875) *Principles of Geology*. London: Murray.
- Molnar P and England P (1990) Late Cenozoic uplift of mountain-ranges and global climate change – chicken or egg? *Nature* 346: 29–34.
- Pagani M, Zachos JC, Freeman KH, Tzippe B, and Bohaty S (2005) Marked decline in atmospheric carbon dioxide concentrations during the Paleogene. *Science* 309: 600–603.
- Pearson PN and Palmer MR (2000) Atmospheric carbon dioxide concentrations over the past 60 million years. *Nature* 406: 695.
- Ramsay W (1924) Climate problem in geology. *Geological Magazine* 61: 152–163.
- Raymo ME and Ruddiman WF (1992) Tectonic forcing and late Cenozoic climate. *Nature* 359: 117–122.
- Raymo ME, Ruddiman WF, and Froelich PN (1988) Influence of late Cenozoic mountain building on ocean geochemical cycles. *Geology* 16: 649–653.
- Roth JM, Droxler AW, and Kameo K (2000) The Caribbean carbonate crash at the middle to late Miocene transition; linkage to the establishment of the modern ocean conveyor. In: Leckie RM, Sigurdsson H, Acton GD, and Draper G (eds.) *Proceedings of the Ocean Drilling Program, Scientific Results* 165, pp. 249–273. College Station, TX: Ocean Drilling Program.
- Ruddiman WF (ed.) (1997) *Tectonic Uplift and Climate Change*, p. 535. New York: Plenum Press.
- Ruddiman WF (2010) A paleoclimatic enigma? *Science* 328: 838–839.
- Ruddiman WF and Kutzbach JE (1991) Plateau uplift and climatic change. *Scientific American* 264: 66.
- Ruddiman WF and Prell WL (1997) Introduction to uplift-climate connection. In: Ruddiman WF (ed.) *Tectonic Uplift and Climate Change*, pp. 1–15. New York: Plenum Press.
- Sloan LC, Huber M, Crowley TJ, Sewall JO, and Baum S (2001) Effect of sea surface temperature configuration on model simulations of “equable” climate in the early Eocene. *Paleogeography, Paleoclimatology, Paleoecology* 167: 321–335.
- Smith AG and Pickering KT (2003) Oceanic gateways as a critical factor to initiate icehouse Earth. *Journal of the Geological Society of London* 160: 337–340.
- Svensen H, Planke S, and Corfu F (2010) Zircon dating ties NE Atlantic sill emplacement to initial Eocene global warming. *Journal of the Geological Society, London* 167: 433–436.
- Tripathi AK, Roberts CD, and Eagle RA (2009) Coupling of CO_2 and ice sheet stability over major climate transitions of the last 20 million years. *Science* 326: 1394–1397.
- Willenbring JK and von Blanckenburg F (2010) Long-term stability of global erosion rates and weathering during late-Cenozoic cooling. *Nature* 465: 211–214.
- Wright D and Miller KG (1996) Control of North Atlantic deep water circulation by the Greenland-Scotland Ridge. *Paleoceanography* 11: 157–170.
- Zachos J, Pagani M, Sloan L, Thomas E, and Billups K (2001) Trends, rhythms, and aberrations in global climate 65 Ma to present. *Science* 292: 686–693.

Astronomical Theory of Paleoclimates

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Past Climates at Astronomical Timescales

Earth's history has been studded with periods when the climate was markedly colder than at others. These cold periods, the so-called Ice Ages, have, on the geological time scale, been relatively short. The most well-known are the Pre-Cambrian Ice Ages, the late Ordovician Ice Age, the Permo-Carboniferous Ice Age, and the most recent Ice Age. This most recent Ice Age, which the Earth entered 2 to 3 million years (My) ago is called the Quaternary Period and is composed of the Pleistocene and the Holocene. It was preceded by Antarctic ice-sheet formation from 25 to 10 My before present (BP) and a long cooling trend through all the Tertiary. It has been characterized by multiple switches of the global climate between glacials (with extensive ice sheets) and interglacials (with a climate warmer or less glaciated than the rest of the time). These past climates have been reconstructed from analysis of proxy data taken from deep-sea cores, ice cores, and land records. Throughout the last million years, these data have demonstrated that successive glaciation–deglaciation cycles have occurred with a dominant quasiperiodicity of 100 thousand years (ka) over which quasicycles of about 41 and 21 ka are superimposed. The astronomical theory of paleoclimates aims to explain the recurrence of these cycles (Berger, 1978, 1988). The last one covers the period from the Eemian interglacial centered about 125 ka BP to the present-day Holocene interglacial, which peaked around 6 ka BP, and includes the Last Glacial Maximum (LGM), which occurred at 21 ka BP. This LGM was documented in great detail by the Climate Long-Range Investigation Mapping and Prediction group (CLIMAP Project Members, 1976). In the Northern Hemisphere (NH), the LGM world differed strikingly from the present, having huge land-based ice sheets that reached approximately 2–3 km in thickness. Sea level was lower by at least 115 m, which represents roughly $50 \times 10^6 \text{ km}^3$ more ice than now, and the global average surface-air temperature was 5 °C below present (Gates, 1976). Aerosol loading was higher than present and carbon dioxide (CO₂) level was about 190 parts per million by volume (ppmv; Petit et al., 1999; much less than the 280 ppmv of the Pre-Industrial Revolution and the 392 ppmv of the year AD 2011).

The Astronomical Theory of Paleoclimates, a Historical Point of View

The astronomical theory is the oldest explanation of the existence of glacial periods found in the geological record of the Quaternary Ice Age. As expounded by Milankovitch (1941), this theory is currently the most popular. In broad terms, proponents of this theory claim that the changes in three of the Earth's orbital and rotational parameters have been

sufficiently large to induce significant changes in the seasonal and latitudinal distributions of incoming solar radiation, thus forcing the pattern of glacial and interglacial cycles to recur as seen in geological records.

Climatic variations at similar time scales also occurred during non-Ice-Age period (e.g., Shackleton et al., 1999) and are also tentatively explained by the astronomical theory.

Milankovitch Era

During the first decades of the twentieth century, Spitaler (1921) rejected Croll's theory (Croll, 1875). He adopted the opposite view, first put forward by Murphy (1876), that a long, cool summer and a short, mild winter are the most favorable for including a glacial. This diminution of heat during the summer season was later recognized by Brückner, Köppen, and Wegener (Brückner et al., 1925) as the decisive factor in glaciation. However, it was Milankovitch who first completed a full astronomical theory of Pleistocene ice ages, computing the numerical values of the astronomical elements and the subsequent changes in insolation and climate.

Milutin Milankovitch was a Serbian engineer–geophysicist–astronomer who was born in Dalj in 1879 and died in Beograd in 1958. Milankovitch's first book dates from 1920, but his massive special publication of the *Royal Serbian Academy of Science* was published in German in 1941, translated into English in 1969, and reedited in 1998. Milankovitch's main contribution was exploring the solar irradiances at different latitudes and seasons in great mathematical detail and relating them, in turn, with the planetary energy balance as determined by the albedo and by the reradiation in the infrared according to Stefan's law. The basis of Milankovitch's argument is that “under those astronomical conditions in which the heat budget around the summer solstice falls below average, so will summer melt, with uncompensated glacial advance being the result.”

This theory requires that the summer in northern high latitudes must be cold to prevent the winter snow from melting in such a way as to allow a positive value in the annual budget of snow and ice. This would initiate a positive feedback cooling over the Earth through a further extension of the snow cover and a subsequent increase in the surface albedo. On the assumption of a perfectly transparent atmosphere and of the northern high latitudes being the most sensitive to insolation changes, this hypothesis requires a minimum in the NH summer insolation at high latitudes (the so-called solar energy at the top of the atmosphere). The essential product of the Milankovitch theory is, therefore, his curve that shows how the intensity of summer sunlight varied over the past 600 ka at 65° N. Actually, he introduced the concept of the caloric season. These seasons are

exactly half-a-year long, which avoids the necessity of taking into account the variations in their length. This is not the case for the astronomical seasons whose lengths vary in time. The summer half-year comprises all the days during which a given mid- to high-latitude region receives more irradiation than any of the days of the winter half-year. On his curves, Milankovitch correlated certain low points in NH insolation with four European ice ages reconstructed 15 years earlier by Albrecht Penck and Eduard Brückner (Penck and Bruckner, 1909).

The Milankovitch Debate

However, when considering the Milankovitch insolation curve, we are left in no doubt that the success of the Milankovitch theory is only apparent, because the Quaternary has had many more glacial periods than was claimed during the first part of the twentieth century. In fact, until roughly 1970, the Milankovitch theory was largely disputed because the discussions were based on fragmentary geological sedimentary records and on inaccurate time scales. In addition, the climate was considered too resilient to react to 'such small changes' as those seen in the Milankovitch summer half-year caloric insolation.

Milankovitch Renaissance

In the late 1960s, judicious use of radioactive dating and other techniques gradually clarified the details of the time scale, better instrumental methods became available for the analysis of oxygen isotopes as proxies for ice-age temperatures, paleoecological analysis of sediment cores was greatly improved, global climates were reconstructed (CLIMAP Project Members, 1976), and atmospheric general circulation models (Gates, 1976) and climate models became available. Following these improvements, Hays et al. (1976) showed that quasiperiods of 100, 41, 23, and 19 ka are clearly marked in proxy records of past climate.

Independently, these periods were also found in the long-term variations of eccentricity, obliquity, and climatic precession using a new model of astronomical forcing (Berger, 1978). These results found simultaneously in geological and astronomical data constituted the most delicate proof that astronomical elements have a sizable effect on the slow evolution of climate.

These results sparked a revival of the popularity of the astronomical theory of paleoclimates. New studies were initiated in all four of the main elements of astronomical theory, namely, (i) the computation of the astronomical elements, (ii) the computation of the appropriate insolation parameters, (iii) the development of suitable climate models, and (iv) the analysis of geological data in the time and frequency domains in order to investigate the physical mechanisms that are responsible for the long-term climatic variations and to calibrate and validate the climate models.

Long-Term Variations of the Astronomical Parameters

As we are primarily interested in glacial–interglacial cycles, the focus here is only on the long-term variations of the

energy (insolation) received by the Earth from the Sun at the top of the atmosphere and at time scales of tens to hundreds of thousands of years.

The incoming solar radiation changes from day to day due to the Earth's elliptic motion around the Sun. But the seasonal and latitudinal distributions of daily insolation also have long-term variations that are related to the orbit of the Earth around the Sun and to the inclination of its axis of rotation. These involve three well-identified astronomical parameters (Figure 1): the eccentricity, e (a measure of the shape of the Earth's orbit around the Sun); the obliquity, ε (the tilt of the equator with respect to the plane of the Earth's orbit); and climatic precession, $e \sin \tilde{\omega}$, a measure of the Earth–Sun distance at the summer solstice ($\tilde{\omega}$, the longitude of the perihelion, a measure of the angular distance between the perihelion and the vernal equinox that are both in motion). The present-day value of e is 0.016. As a consequence, although the Earth's orbit is very close to a circle, the Earth–Sun distance and, consequently, the insolation vary by as much as 3.2% and 6.4%, respectively, over the course of one year. The obliquity, which defines the tropical latitudes and polar circles, is presently $23^\circ 27'$. The longitude of the perihelion is 102° , which means that the NH winter occurs when the Earth is closest to the Sun.

The secular variations of these elements of the Earth's orbit and rotation are due to the gravitational perturbations which the Sun, the other planets, and the moon exert on the Earth's orbit and axis of rotation. The solution of the system of equations that governs these motions allows the numerical values of precession, obliquity, and eccentricity to be expressed in trigonometrical form as quasiperiodic functions of time:

$$e \sin \tilde{\omega} = \sum P_i \sin(\alpha_i t + \eta_i) \quad [1]$$

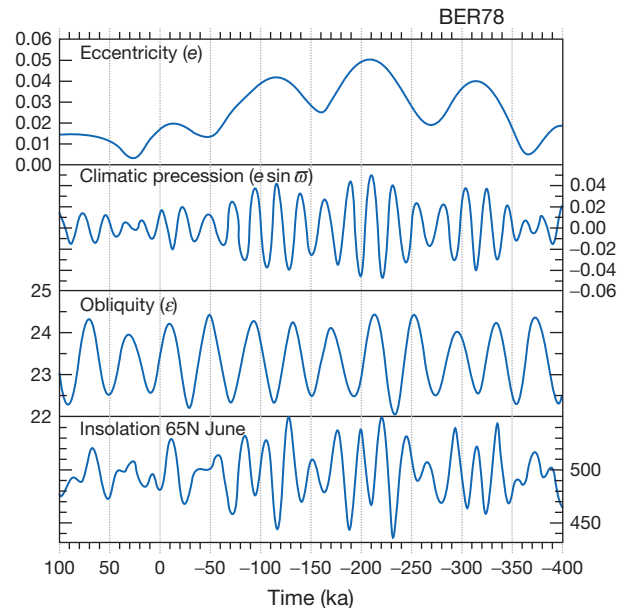


Figure 1 Long-term variations over the last 400 ka and the next 100 ka of eccentricity, climatic precession, obliquity, and 65° N insolation at the summer solstice in WM^{-2} (Berger, 1978). This insolation is shown as an example of Milankovitch (1941) principles. It must be stressed that all latitudes and days are used in the LLN models. LLN, Louvain-la-Neuve.

$$\varepsilon = \varepsilon^* + \sum A_i \cos(\gamma_i t + \zeta_i) \quad [2]$$

$$e = e^* + \sum E_i \cos(\lambda_i t + \phi_i) \quad [3]$$

where the amplitudes P_i , A_i , E_i , the frequencies α_i , γ_i , λ_i and phases η_i , ζ_i , ϕ_i have been computed by Berger (1978; Berger and Loutre, 1991; <ftp://ftp.elic.ucl.ac.be/berger/berger78>). They can be used over a few millions of years, but for more remote times, numerical solutions are necessary (Laskar, 1999).

Over the past 3 million years, the orbit varied between near circularity ($e=0$) and slight ellipticity ($e=0.07$), with a quasi-period of about 400 ka over which a period of around 100 ka is superimposed. The periods of the most important terms in the series expansion of e are actually 404, 95, 124, 99, 131, and 2380 ka (in decreasing order of amplitude). The tilt of the Earth's axis of rotation varied between about 22° and 25° at a period of nearly 41 ka. Although this period corresponds in eqn [2] to the amplitude that is by far the largest, there are two other important terms with periods of 54 and 29 ka. As far as precession is concerned, two phenomena must be considered. The first is the axial precession. The second is related to the fact that the elliptical figure of the Earth's orbit is itself rotating counterclockwise leading to an absolute motion of the perihelion. The two effects together result in what is known as the climatic precession of the equinoxes, the equinoxes and solstices shifting slowly around the Earth's orbit, relative to the perihelion, with a mean period of 21 ka. This period actually results from the existence in eqn [1] of four periods that are close to 23 and 19 ka.

The long-term variations of these three astronomical parameters over the past show that their periods are not constant in time (Berger et al., 1998). The average period of climatic precession varies between 15 and 30 ka. For eccentricity, the time interval between two successive maxima averages ~ 95 ka, but it varies with the 400 ka cycle, being remarkably short when the eccentricity vanishes, as at 27 ka after present (AP) when the Earth's orbit will be circular. Actually, the most important theoretical period of eccentricity, 400 ka, is weak before 1 My BP, and is particularly strong over the next 400 ka, with the strength of the components in the 100-ka band changing in the opposite way. It is worth pointing out that this weakening of the 100-ka period started about 900 ka ago when this period began to appear very strongly in paleoclimate records. This implies that the 100-ka period found in paleoclimatic records is definitely not linearly related to eccentricity.

For obliquity, the main period is essentially stable, but there is an amplitude modulation with a time duration of about 1.3 My. The spectra of both the amplitude and frequency modulations of obliquity display significant power at 171 and 97 ka. Although the 97 ka is close to the so-called 100-ka eccentricity period, they are not related.

Obliquity defines the seasons: boreal (austral) summer occurs when the Earth's axis of rotation and its North Pole (South Pole) are tilted toward the Sun. An increase of obliquity increases the solar energy in high latitudes of the summer hemispheres, increasing therefore the seasonal contrast in both the northern and southern hemispheres. On the other hand, precession modulates the intensity of the seasons but acts at the world scale instead of at the hemispheric one. For

example, ~ 10 ka ago, the whole Earth received more solar energy during boreal summer and less during boreal winter because boreal summer occurred at perihelion and boreal winter at aphelion, leading to a maximum seasonal contrast in the NH. Today the winter solstice occurs near perihelion and the summer solstice at aphelion, leading to a minimum seasonal contrast in the NH. The magnitude of this seasonal contrast changes every half-precession cycle and is moreover affected by eccentricity. The elliptical shape of the Earth's orbit around the Sun affects, indeed, the Earth–Sun distance through the year. When the Earth is at perihelion, the distance is the shortest, equal to $(1 - e)$, when it is at aphelion, it is the largest, equal to $(1 + e)$. Consequently, the larger e , the larger the difference between these two extreme positions and in particular, the larger the seasonal contrast in the NH when boreal summer occurs at perihelion (see also Yin and Berger, 2012). Moreover, because the length of the seasons varies in time according to Kepler's second law, the solstices and equinoxes occur at different calendar dates during the geological past and future. Presently in the NH, the longest seasons are spring (92.8 days) and summer (93.6 days), while autumn (89.8 days) and winter (89.0 days) are notably shorter. In AD 1250, spring and summer had the same length (as did autumn and winter) because the winter solstice was occurring at the perihelion. The daily insolation is primarily a function of precession, except for the latitudes and days close to the polar night where obliquity dominates (but where the energy is close to zero). On the contrary, the total energy received during any time interval of the year defined in terms of the true longitude is only a function of obliquity (Figure 2; Berger et al., 2010). However, in this case, the length in days of such intervals varies in time as a function of precession only (Berger and Loutre, 1994), leading to an average irradiance being a function of both precession and obliquity. This spectral characteristic is also seen in the total irradiance calculated over a period of time defined in terms of the calendar (Figure 2).

Milankovitch Theory and Past Climates

Given the Milankovitch conditions for a glacial to occur, the northern high latitudes would remain cool enough in summer to prevent snow and ice from melting, and (as pointed out by Berger, 1980) mild winters would allow a substantial evaporation in tropical and temperate latitudes as well. As a consequence, abundant snow would fall in polar latitudes, the moisture being supplied there by an intensified general circulation due to an intensified latitudinal gradient.

A simple linear version of the Milankovitch model therefore predicts that the total ice volume and climate vary with the same regular pattern as the insolation. If this is true, the proxy record of climate variations would contain the frequencies of the astronomical parameters that are responsible for changing the seasonal and latitudinal distributions of the incoming solar radiation. Investigations during recent decades have indeed demonstrated that the 19-, 23-, and 41-ka periodicities actually occur in long records of Quaternary climate (Hays et al., 1976); that the climatic variations observed in these frequency bands are linearly related to the orbital forcing (Imbrie et al., 1992); and that there is a fairly consistent phase relationship between

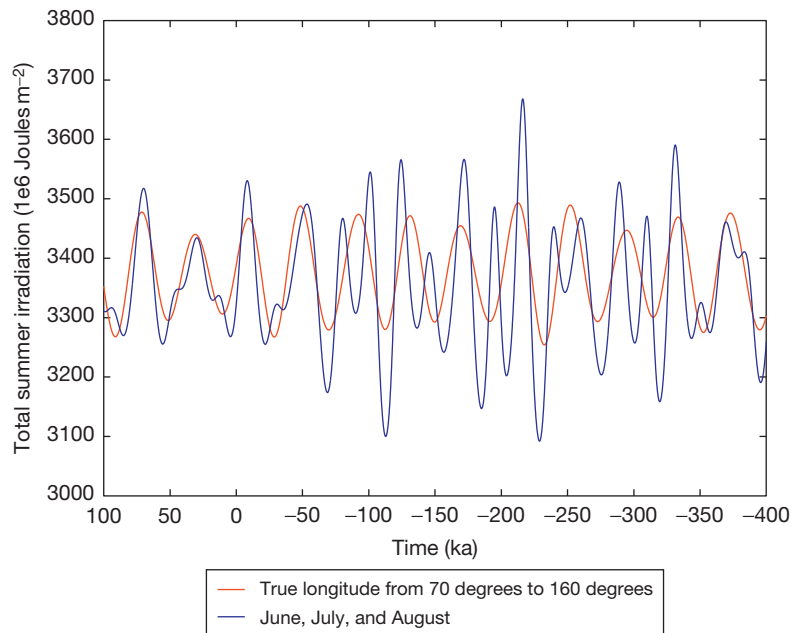


Figure 2 Long-term variations from 400 ka BP to 100 ka AP of the total irradiation received at 65°N during June, July and August (blue curve) and during the period going from 70° to 160° in true longitude (Berger et al., 2010).

insolation, sea-surface temperature, and ice volume (Imbrie et al., 1984). Moreover, the geological observation of the bipartition of the precessional peak, a bipartition also found independently in astronomical computations by Berger (1977), was one of the first most delicate and impressive tests of the Milankovitch theory.

However, despite the evidence in support of the Milankovitch hypothesis, the same investigations identified the largest climatic cycle as 100 ka. As this eccentricity cycle is very weak in the insolation, it cannot be related to the orbital forcing by any simple linear mechanism (Imbrie et al., 1993). Actually, the variance components centered near this 100-ka cycle seem to be in phase with the eccentricity cycle, but its exceptional strength in the climatic record demands a nonlinear amplification.

The identification of the periodicities found in the geological records is, however, not always as clear. The 100-ka cycle, the most dominant feature in the ice volume record for the late Pleistocene, does not exhibit a constant amplitude over the past 2–3 million years; it disappears before 1 My BP, at a time when the ice sheets were much less extensive over the Earth (Prell, 1982). The SPECMAP (Spectral Mapping Project) group has also shown that the nature of the paleoclimate reconstructions and the phase lags between proxy records and the climate response to orbital forcing depend upon the location of the deep-sea cores and the nature of the climatic parameters under consideration (Imbrie et al., 1992, 1993; Ruddiman, 2003a; Ruddiman and Raymo, 2003).

Modeling Past and Future Climates

Since the publication of Hays et al. (1976), a number of studies have attempted to explain the relationship between

astronomical forcing and climatic change and some have tried to relate different forcing parameters to the ice volume. Among the first, Kukla (1972) suggested the use of latitudinal gradient of insolation. More recently, Shackleton (2000) showed that the glacial–interglacial ice cycles lag temperature, CO₂, and orbital eccentricity significantly. Another approach is to use climate models: both general circulation models and models that take into account the different components of the climate system (for a partial review of such experiments see Berger, 1995). These models aim to explain the response of the climate system to external forcings taking into account, as much as possible, the complexity of the main feedback mechanisms. These mechanisms include albedo and water vapor feedbacks, the continental biospheric impacts on high-latitude surface albedo, the altitude and continental effects on precipitation over the ice sheets, the lithospheric response to the ice-sheet loading, and the rheology (deformation and flow) of ice itself.

Sensitivity analyses have tentatively confirmed that the astronomical parameters act as a pacemaker of the glacial–interglacial cycles, the insolation forcing being able to induce them, but only under low CO₂ concentrations (Berger and Loutre, 2004). Long-term variations of CO₂ alone are not sufficient to drive glacial–interglacial cycles (Loutre and Berger, 2000a). When both insolation and CO₂ variations are used, it is significant that the model of Louvain-la-Neuve (LLN), in particular, simulates almost correctly the entrance into glaciation around 2.75 My BP, the late Pliocene–early Pleistocene 41-ka cycle, the emergence of the 100-ka cycle around 850 ka BP, and the glacial–interglacial cycles of the last 600 ka (Berger and Loutre, 2004; Figure 3 for 200 ka BP to 130 ka AP).

This glacial–interglacial cycle about 100-ka long is one of the most striking features of the Quaternary and is also linked to the future of our climate. As each cycle is

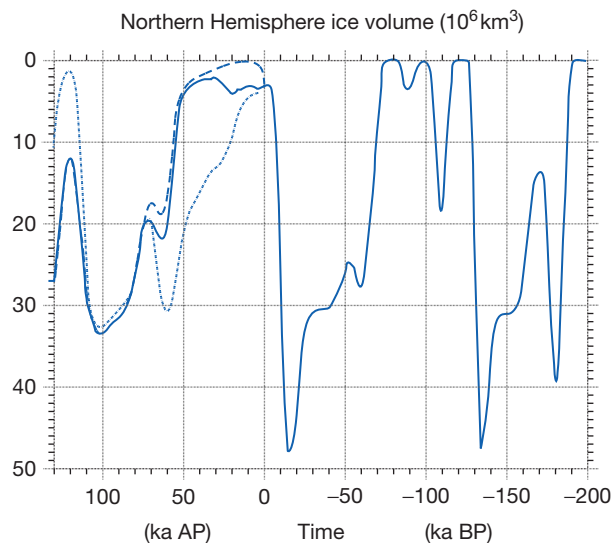


Figure 3 Simulated ice volume of the Northern Hemisphere using the two-dimensional LLN Northern Hemisphere climate model forced by insolation and CO₂ from 200 ka BP to 130 ka AP. The Vostok CO₂ concentration and Vostok scenario lead to the full line. The dotted line results from a simulation where CO₂ is kept equal to 210 ppmb over the next 130 ka, and the dashed curve provides the results of the 750 ppmv scenario. The ice volume increases downward (Loutre and Berger, 2000b). AP, after present; BP, before present; LLN, Louvain-la-Neuve; ppmv, parts per million by volume.

characterized by a long glacial period followed by a short interglacial (~10–15 ka long) and as our interglacial, the Holocene, is already 10 ka long, paleoclimatologists were naturally inclined to predict that the next Ice Age is quite close. Simulations using our climate model show, however, that the current interglacial will probably last much longer than any previous interglacial (Berger and Loutre, 2002) with an early entrance into glaciation being highly improbable. This is suggested to be related to the shape of the Earth's orbit around the Sun, which will be almost circular over the next tens of thousands of years. As this is primarily related to the 400-ka cycle of eccentricity, the best and closest analog for such a forcing is definitely Marine Isotopic Stage 11 (MIS-11), some 400 ka ago, not MIS-5e (Loutre and Berger, 2003). Modeling results with the LLN model (Berger and Loutre, 2002) and EPICA Community Members (2004) reconstruction confirm a similar behavior of the climate response during MIS-11 and MIS-1 with a length of 30 ka or more. An analysis of the interglacial modeling results (Yin and Berger, 2012) shows, however, that MIS-19 is an even better analog of MIS-1 than MIS-11. This is based on an analysis of the individual contributions of the greenhouse gas concentrations and insolation to the climate of the last 800 ka, showing that the latitudinal and seasonal distribution of insolation as well as the CO₂eq concentration of MIS-1 and MIS-19 are very much alike.

Because the latitudinal and seasonal distributions of insolation will not change significantly over the next tens of thousands of years, changes in greenhouse gas concentrations are expected to play an increasingly important role. This is particularly important in the context of the already exceptional present-day CO₂ concentrations (unprecedented over the past

1 million years) and, even more so, because of even larger values predicted to occur during the twenty-first century due to human activities (Houghton et al., 2001). According to experiments made with the two-dimensional LLN climate model, there seems to be a threshold in the CO₂ concentration (~700 ppmv) above which the Greenland ice-sheet melts and disappears totally within around 10 ka (Figure 3). Finally, sensitivity experiments using the LLN model show that only a CO₂ behavior similar to that which occurred during MIS-7 allows an early end of the present interglacial (Crucifix et al., 2005), thereby confirming that the anthropogenic greenhouse era might have begun thousands of years ago (Ruddiman, 2003b) but this is true only in this case.

However, many model refinements are still needed. For example, current models have still to explain why the variations of the oxygen isotope and EPICA records show a significantly lower amplitude before the mid-Brunhes event about 430 ka ago (Ganopolski and Calov, 2012; Yin and Berger, 2010), and why MIS-13 peak, according to Lisiecki and Raymo (2005), seems to respond to NH summer at aphelion rather than at perihelion.

See also: [Loess Deposits: Origins and Properties](#);
[Paleoceanography: Paleoceanography An Overview](#).
[Paleoclimate: Timescales of Climate Change](#). [Paleoclimate Modeling: Data–Model Comparisons](#); [The Last Interglacial](#).

References

- Adhemar JA (1842) *Revolutions des Mers, Deluges Periodiques*. Paris: Publication privee.
- Agassiz L (1838) Upon glaciers, moraines, and erratic blocks. *New Philosophical Journal Edinburgh* 24: 364–383.
- Berger A (1977) Support for the astronomical theory of climatic change. *Nature* 268: 44–45.
- Berger A (1978) Long-term variations of daily insolation and Quaternary climatic changes. *Journal of the Atmospheric Sciences* 35: 2362–2367.
- Berger A (1980) The Milankovitch astronomical theory of paleo-climates: A modern review. In: Beer A, Pounds K, and Beer P (eds.) *Vistas in Astronomy, an international review journal*, vol. 24 Part 2, pp. 103–122. Oxford: Pergamon Press.
- Berger A (1988) Milankovitch theory and climate. *Review of Geophysics* 26: 624–657.
- Berger A (1995) Modelling the response of the climate system to astronomical forcing. In: Henderson-Sellers A (ed.) *Future Climates of the World. A Modelling Perspective* 16: 21–69. Amsterdam: Elsevier.
- Berger A and Loutre MF (1991) Insolation values for the climate of the last 10 million years. *Quaternary Science Reviews* 10: 297–317.
- Berger A and Loutre MF (1994) Long-term variations of the astronomical seasons. In: Boudron C (ed.) *Topics in Atmospheric and Interstellar Physics and Chemistry*, pp. 33–61. Paris: Les Editions de Physique, Les Ulis.
- Berger A and Loutre MF (2002) An exceptionally long interglacial ahead? *Science* 297: 1287–1288.
- Berger A and Loutre MF (2004) Astronomical theory of climatic change. In: Boudron C (ed.) *From Indoor Air Pollution to the Search for Earth like Planets in the Cosmos*, vol. 6, pp. 1–36. EDP, Sciences, Les Ulis Cedex A, France. Also *Journal de Physique IV Proceedings*, vol. 121.
- Berger A, Loutre MF, and Mélice JL (1998) Instability of the astronomical periods over the last and next millions of years. *Paleoclimate Data and Modelling* 2(4): 239–280.
- Berger A, Loutre MF, and Yin QZ (2010) Total irradiation during any time interval of the year using elliptic integrals. *Quaternary Science Reviews* 29: 1968–1982.
- Brückner E, Köppen W, and Wegener A (1925) Über die Klimate der geologischen Vorzeit. *Zeitschrift für Gletscherkunde* 14: 149–169.
- CLIMAP Project Members (1976) The surface of the Ice-Age Earth. *Science* 191: 1131–1136.

- Croll J (1875) *Climate and Time in Their Geological Relations*, p. 577. New York/London: Appleton.
- Crucifix M, Loutre MF, and Berger A (2005) Commentary on 'The anthropogenic greenhouse era began thousands of years ago'. *Climatic Change* 69: 423–426.
- EPICA Community Members (2004) Eight glacial cycles from an Antarctic ice core. *Nature* 429: 623–628.
- Ganopolski A and Calov R (2012) Simulation of glacial cycles with an Earth system model. In: Berger A, Mesinger F, and Sijacki D (eds.) *Climate Change at the Eve of the Second Decade of the Century. Inferences from Paleoclimates and Regional Aspects, Milankovitch 130th Anniversary Symposium, Belgrade, 2009*, pp.49–55. Vienna: Springer-Verlag.
- Gates L (1976) Modeling the ice-age climate. *Science* 191: 1138–1144.
- Hays JD, Imbrie J, and Shackleton NJ (1976) Variations in the earth's orbit: Pacemaker of the ice ages. *Sciences* 194: 1121–1132.
- Houghton JT, Ding Y, Griggs DJ, et al. (2001) *Climate Change 2001: The Scientific Basis*. Cambridge: Cambridge University Press. Published for the Intergovernmental Panel on Climate Change.
- Imbrie J, Berger A, Boyle EA, et al. (1993) On the structure and origin of major glaciation cycles. 2. The 100,000-year cycle. *Paleoceanography* 8: 699–735.
- Imbrie J, Boyle EA, Clemens SC, et al. (1992) On the structure of major glaciation cycles. *Paleoceanography* 7: 701–738.
- Imbrie J, Hays J, Martinson DG, et al. (1984) The orbital theory of Pleistocene climate: Support from a revised chronology of the marine 180 record. In: Berger AL, Imbrie J, and Hays J, et al. (eds.) *Milankovitch and Climate*, pp. 269–305. Hingham, MA, USA: D Reidel.
- Kukla G (1972) Insolation and glacial. *Boreas* 1: 63–96.
- Laskar J (1999) The limits of Earth orbital calculations for geological time scale use. In: Shackleton NJ, McCave IN, and Weedon GP (eds.) *Astronomical (Milankovitch) Calibrations of the Geological Time Scale. Philosophical Transactions of the Royal Society of London, Series A: Mathematical, Physical and Engineering Sciences* 357: 1735–1759.
- Lisiecki LE and Raymo ME (2005) A Pliocene–Pleistocene stack of 57 globally distributed benthic delta 180 records. *Paleoceanography* 20(1): PA1003. <http://dx.doi.org/10.1029/2004001071>.
- Loutre MF and Berger A (2000a) No glacial-interglacial cycle in the ice volume simulated under a constant astronomical forcing and a variable CO₂. *Geophysical Research Letters* 27: 783–786.
- Loutre MF and Berger A (2000b) Future climatic changes: Are we entering an exceptionally long interglacial? *Climatic Change* 46: 61–90.
- Loutre MF and Berger A (2003) Stage 11 as an analogue for the present interglacial. *Global and Planetary Change* 36: 209–217.
- Milankovitch M (1941) *Kanon der Erdbastrahlung und seine Anwendung auf des Eiszeitenproblem*. Special Publication 132, Section of Mathematical and Natural Sciences, vol. 33, p. 633. Belgrade: Royal Serbian Sciences. [Canon of Insolation and the Ice Age Problem, English translation by Israël Program for the US Department of Commerce and the National Science Foundation, Washington DC, 1969, and by Zavod za Udzbenike I nastavna Sredstva in cooperation with Muzej nauke I tehnike Srpske akademije nauka I umetnosti, Beograd, 1998].
- Murphy JJ (1876) The glacial climate and the polar ice-cap. *Quarterly Journal of the Geological Society of London* 32: 400–406.
- Penck A and Bruckner E (1909) *Die Alpen in Eiszeitalter*. Leipzig: Tauchnitz.
- Petit JR, Jouzel J, Raynaud D, et al. (1999) Climate and atmospheric history of the past 420,000 years from Vostok ice core, Antarctica. *Nature* 399: 429–436.
- Prell WL (1982) Oxygen and carbon isotope stratigraphy for the Quaternary of Hole 502B: Evidence for two modes of iso-topic variability. *Initial Reports of the Deep Sea drilling Project* 68: 455–464.
- Ruddiman WF (2003a) Orbital insolation, ice volume and greenhouse gases. *Quaternary Science Review* 22: 1597–1629.
- Ruddiman WF (2003b) The anthropogenic greenhouse era began thousands of years ago. *Climatic Change* 61: 261–293.
- Ruddiman WF and Raymo ME (2003) A methane-based time scale for Vostok ice. *Quaternary Science Review* 22: 141–155.
- Shackleton NJ (2000) The 100,000-year Ice-Age Cycle identified and found to lag temperature, carbon dioxide and orbital eccentricity. *Science* 289: 1897–1902.
- Shackleton NJ, McCave IN, and Weedon GP (eds.) (1999) Astronomical (Milankovitch) calibration of the geological time scale. *Philosophical Transactions of the Royal Society of London, Series A: Mathematical, Physical and Engineering Sciences* 357: 1731–2007.
- Spitaler R (1921) *Das Klima des Eiszeitalters*. Prague: Selbstverlag.
- Yin QZ and Berger A (2010) Insolation and CO₂ contribution to the interglacial climate before and after the Mid-Brunhes Event. *Nature Geoscience* 3(4): 243–246.
- Yin QZ and Berger A (2012) Individual contributions of insolation and CO₂ to the interglacial climates of the past 800 ky. *Climate Dynamics* 38: 709–724.

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GLACIATIONS

Contents

Overview

Transition from Late Neogene to Early Quaternary Environments

Early Quaternary (Pleistocene) and Precursors

Middle Pleistocene in Eurasia

Mid-Quaternary in North America

Middle Pleistocene Glaciations in the Southern Hemisphere

Late Pleistocene Glacial Events in Beringia

Late Quaternary of the Southwest Pacific Region

Late Quaternary of Antarctica

Late Pleistocene in Eurasia

Late Quaternary in Highland Asia

Late Quaternary in North America

Late Pleistocene in South America

Neoglaciation in Europe

Neoglaciation in the American Cordilleras

Overview

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From the mid-nineteenth century, the Quaternary has been considered synonymous with extensive glaciation of the mid-latitudes. Although there were local precursors, significant glaciation began in the late Oligocene (~35 My) in eastern Antarctica. It was followed by glaciation in mountain areas through the Miocene (in Alaska, Greenland, Iceland, and Patagonia), later in the Pliocene (e.g., in the Alps, the Bolivian Andes, and possibly in Tasmania), and in the earliest Pleistocene (New Zealand, Iceland, and Greenland). Today, evidence from both the land and the ocean-core sequences demonstrates that the major continental glaciations, outside the polar regions, were markedly restricted to the last 1 My–800 ka or less. This is not to say that glaciation did not occur earlier; indeed, glaciation limited to higher latitudes or mountain massifs certainly occurred throughout the 2.6 My of the Quaternary, particularly in the Rocky Mountains. It also occurred in eastern North America, in the Andes, and in southern Australasia. Indeed, evidence of extensive ice rafting, an indication that glaciers had reached sea level, is found from the earliest cold stage – the Praetiglian (2.6–2.4 My) and its equivalents, in both the North Atlantic and the North Pacific oceans (cf. [Haug et al., 2005](#)). Even earlier ice-rafted debris is found in ocean-sediment cores from the North Atlantic region, including the Barents Sea, and areas adjacent to Norway, N and SE Greenland, Iceland, and northern North America, and in the Southern Ocean off Antarctica.

Earth's climate, which had been extremely warm during the Cretaceous 'hot-house,' continued to be both warm and without extreme fluctuations during the Paleogene. An overall, long-term trend of cooling global climates through the Tertiary culminated in the mid-latitudes with warm temperate climates, alternating with cool-temperate climates in the Late Neogene. Like the later Neogene, the early Pleistocene (2.6–0.78 Ma) was characterized by climatic fluctuations dominated by the 41 ka precession cycle (see [Pliocene Environments](#)), during which relatively few cold periods were sufficiently cold and long to allow the development of substantial ice sheets. They include the Plio–Pleistocene boundary events marine isotope stages (MIS) 104, 100, and 98, together with early Pleistocene MIS 82, 78, 68, 60, 58, 54, 52, 36, 34, 30, and 26. Their content of heavy oxygen isotopes in the sea water ($\delta^{18}\text{O}_{\text{ocean}}\text{‰}$) reached values of ~4.6–5. These relatively high values suggest that major quantities of water might have been converted to ice sheets. It was not until the transition in dominant orbital cyclicity to the 100 ka cycles, which began ~1.2 Ma and was fully established by about 800 ka, that the cold periods (glacials) were regularly cold and long enough to allow ice-sheet development on a continental scale, outside the polar regions. Here, MIS 22 (~870–880 ka) was the first of the 'major' cold events that reached critical values of ~5.5 or above $\delta^{18}\text{O}_{\text{ocean}}\text{‰}$ equivalent to substantial ice volumes that typify glaciations of the

later Pleistocene (i.e., MIS 16, 12, 10, 6, and 4–2). Potentially, therefore, it would seem likely that there were a minimum of 20 periods during which extensive glaciation could have developed during the last 2.6 Ma, with the most extensive (~5–6 periods) being limited to the last 900 ka ([Figure 1](#)).

Precisely where these glaciations occurred and how far they extended is very difficult to determine, given that the remnants of less extensive early glaciation tend to be obliterated and mostly removed by later, more extensive advances. Although this is so in all terrestrial areas, it is especially difficult in mountain regions where the preservation potential of older sequences rapidly diminishes with time and subsequent glaciation. However, examination of the frequency of glaciation through the Cenozoic ([Figure 2](#)) indicates that glaciation in the Southern Hemisphere having been established first, principally in Antarctica and southern South America, has occurred continually from the Early Neogene to the present day. By contrast, Northern Hemisphere glaciation, although initially somewhat restricted, increased markedly at the Plio–Pleistocene boundary, increasing again in frequency in the latest early Pleistocene and reaching very high levels in the middle–late Pleistocene. While this pattern is not unexpected, potentially a result of the relative distribution of land and land-locked seas on the earth’s surface, the striking increase in ice sheets through the Quaternary clearly emphasizes that worldwide glaciation is in effect a Northern Hemispheric phenomenon.

Examination of the evidence accumulated in the INQUA project ‘Extent and Chronology of Quaternary Glaciations’ ([Ehlers and Gibbard, 2004a,b,c; Figure 3](#)), supported by other published sources, demonstrates the current state of knowledge. When examining the resulting tables ([Tables 1–3](#)), it is important to bear in mind that the stratigraphical control between regions, beyond the range of radiocarbon dating, is weaker than might be desired. This is particularly a problem outside Europe where biostratigraphy is less well developed and the sheer extent of the unexplored regions makes future discoveries likely. Thus, the presentation below can only be a first step toward a comprehensive overview. The correlations applied here are based on the Global Correlation Chart ([Gibbard et al., 2004, 2005](#)).

Europe

Evidence of glaciation is widespread across Europe and Siberia from throughout the Quaternary and indeed the Neogene ([Table 1](#)). However, before the middle Pleistocene, glaciation appears to be represented only by ice-rafted clasts, outside the mountain or high-latitude regions (e.g., in the Netherlands, lowland Germany, European Russia, and Britain). In this context, the Dutch ‘Hattem Beds’ of Menapian age (1–1.2 Ma; ? MIS 34–36) are of particular significance in that they indicate substantial glaciation of the Baltic region late in the early Pleistocene. Iceland is an obvious exception to this pattern, with glaciation in the mountains beginning in the Miocene (23.8–5.3 Ma BP; see [Coral Records of Relative Sea-Level Changes](#)) and occurring regularly through the Pliocene to the present day (cf. Greenland, below, [Table 2](#)). Likewise, in Norway, its adjacent offshore, and the neighboring Barents Sea, glaciation is recorded from the early Miocene, early Pliocene, and Plio–Pleistocene, and periodically throughout the

remaining period. The absence of evidence from the early middle Pleistocene probably represents destruction of evidence by later glaciation rather than an interruption of the glacial history.

In Italy, particularly in the northern mountains, glaciation is identified possibly from the Plio–Pleistocene onward but becomes established in the Dolomites by MIS 22 ([Muttoni et al., 2003](#)). Comparable evidence is also found from north of the Alps in Switzerland and southern Germany, with records even from the Pliocene in the former. However, in many cases, truly glaciogenic deposits are scarce or lacking, and in the absence of long, continuous sequences, dating is often problematic. Further to the west, in the Pyrenees, the oldest glaciation currently identified is of late Cromerian age (MIS 16 or 14).

The till sheets of the major glaciations of the ‘glacial Pleistocene’ (see [Mid-Quaternary in North America](#)) are found throughout Europe, and especially in the lowlands. Widespread lowland glaciation began in the early middle Pleistocene shortly after the Brunhes/Matuyama paleomagnetic reversal (780 ka). The phases represented include the Weichselian (Valdaian, MIS 4–2), Saalian (Dniepr and Moscovian, MIS ?8/6 and 6, respectively), Elsterian (Okan, ? MIS 12), and the Donian (Narevian, MIS 16). More limited glaciation may also have occurred in the circum-Baltic region during the latest early Pleistocene (MIS 20 and 22). It is curious, however, that evidence for glaciation in the early middle Pleistocene is absent from the North Atlantic and Norway, while it is certainly present in Denmark, the Baltic region, and European Russia.

There can be no doubt that at least two major glaciations occurred in Europe after the Eemian (MIS 5e) and prior to the Last Glacial Maximum (MIS 2). An early glaciation (probably MIS 5b) covered major parts of the West Siberian lowlands and blocked the drainage of the Ob and Yenisei rivers. Another extensive ice sheet (MIS 4) reached nearly as far east, while the Last Glacial Maximum was limited to the circum-Scandinavian region and the Putorana Plateau (see [Neoglaciation in the American Cordilleras](#)).

North America

The North American record of glaciation is similar to its European counterpart, the longest sequences being restricted to Alaska and the adjacent North-West Territories of Canada, which, together with Greenland and the Rockies, preserve evidence of glaciation from the Neogene to the present (see [Middle Pleistocene in Eurasia](#)) ([Table 2](#)). In northern Canada and Alaska, the oldest till and the accompanying ice-rafted detritus in marine settings date from the early Miocene, with regionally widespread glaciation occurring in the Pliocene and regularly throughout the Pleistocene (cf. [Haug et al., 2005](#)). In adjacent British Columbia, a comparable sequence is found, particularly in the north. Similarly, in Greenland, like Iceland ([Table 1](#)), ice caps with glaciers reaching the sea probably developed in the north from the middle Miocene (~14 Ma). By the late Miocene, inland ice shields were periodically present, especially in the mountainous east, with ice reaching the sea in SE Greenland, while contiguous ice sheets have occurred since the earliest Pleistocene (see [Transition from Late Neogene to Early Quaternary Environments](#)). In the

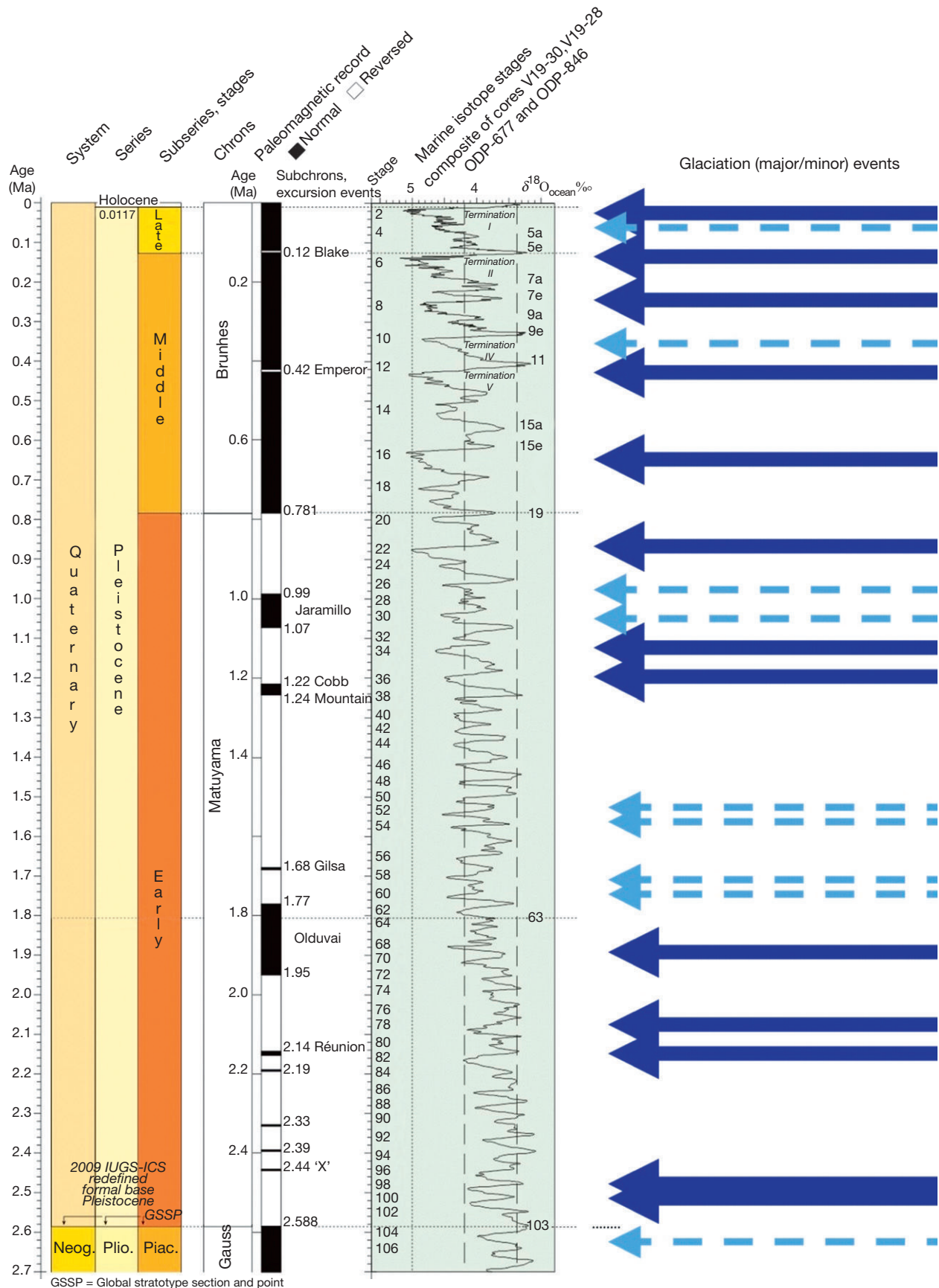


Figure 1 Distribution of major and minor glaciations through Quaternary time. Extract from Global Correlation Table by Gibbard et al. (2004, 2005) and Cohen and Gibbard (2011).

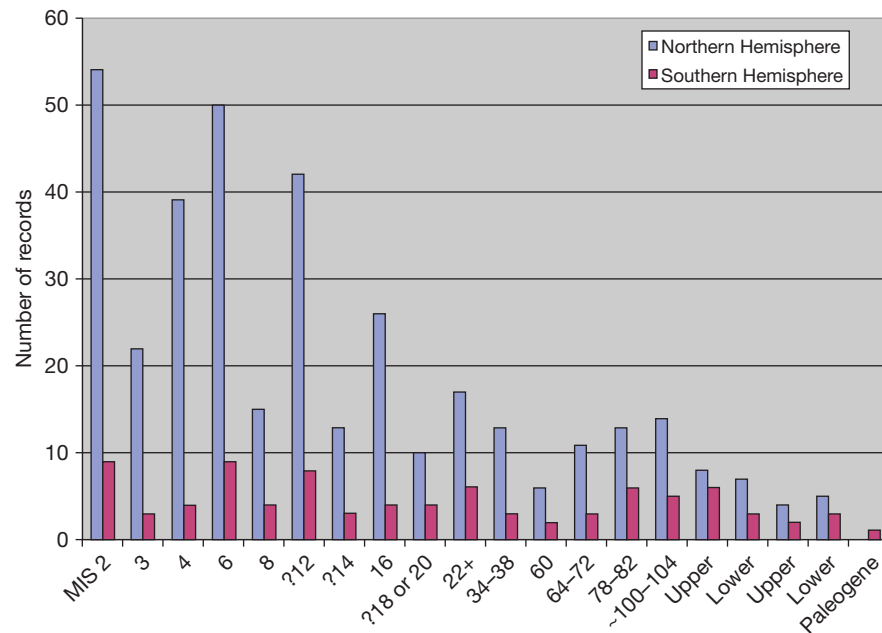


Figure 2 Cenozoic worldwide glaciation through time, showing the relative importance of glaciations as broadly indicated by the number of literature records cited in the compilation by Ehlers and Gibbard (2004a,b,c). MIS, Marine isotope stage.

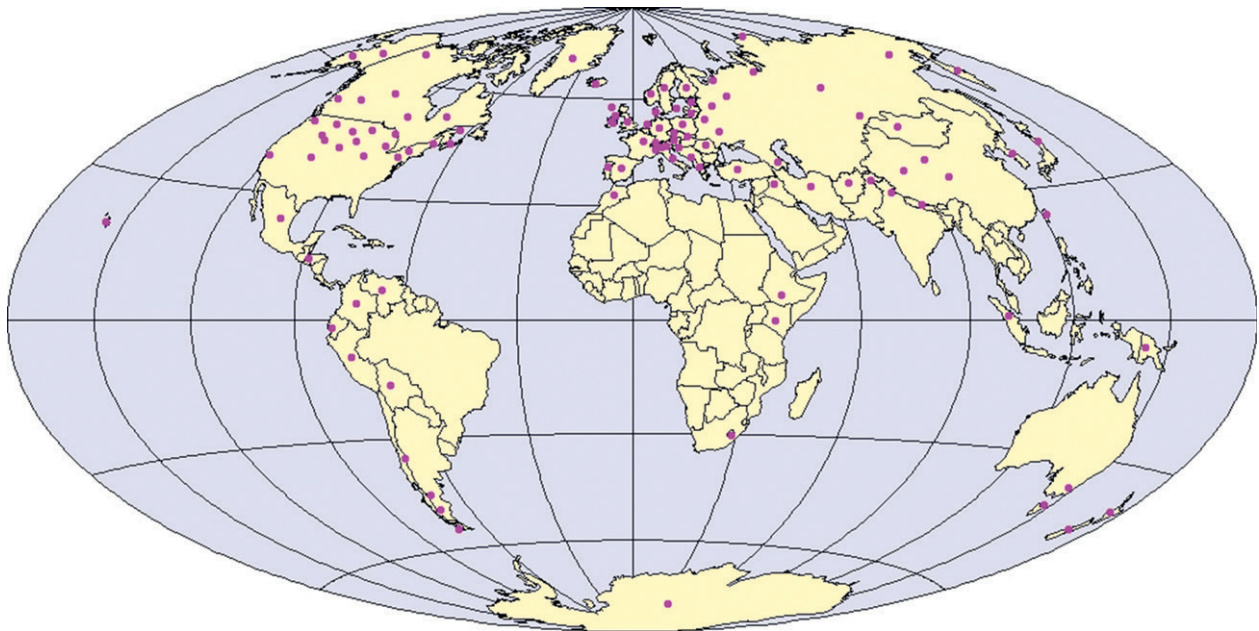


Figure 3 Overview map showing the regions covered by the INQUA project 'Extent and Chronology of Glaciations'. Each dot on the map represents one contribution to the project. Adapted from Ehlers J and Gibbard PL (2003) Extent and chronology of glaciations. *Quaternary Science Reviews* 22: 1561–1568.

United States, the eastern Rockies have a much shorter glacial sequence. Only from Montana and North Dakota is Pliocene-age till known, while in California, the glacial evidence begins at the Plio–Pleistocene boundary (MIS 98–104).

While evidence of early Pleistocene glaciation has been found in areas as distant as Pennsylvania, Wisconsin, and possibly Minnesota, widespread lowland glaciation is first seen again throughout the continent in MIS 22 (pre-Illinoian F),

immediately before the Brunhes/Matuyama paleomagnetic Chron boundary (cf. Europe, [Table 1](#)). From this point onward, major ice sheets are known to have covered large areas in the middle Pleistocene pre-Illinoian events during MIS 16 (D), 12 (B), 8 (A), and 6 (Illinoian *sensu stricto*) and the late Pleistocene MIS 2–4 (Wisconsinan) (see [Middle Pleistocene Glaciations in the Southern Hemisphere, Overview](#)). Evidence from East Greenland suggests that its quasi-permanent ice sheet may have

Table 2 Occurrence of glaciation in North America through the Cenozoic based on numbers of observations presented in contributions to the INQUA project 'Extent and Chronology of Quaternary Glaciations' (Ehlers, Gibbard and Hughes, 2011)

Stage/Age	MIS (approx)	Northern Canada and Alaska	Canadian Prairies	British Columbia	California	Great Basin	Illinois	Minnesota	Missouri	New England	New Brunswick	New York	Maritime Canada	Ohio	Ontario	Pennsylvania	Québec	Rocky Mountains	Washington	Wisconsin	Greenland
Wisconsinan	2	Δ	Δ	Δ	Δ	Δ	Δ		Δ	Δ	Δ	Δ	Δ	Δ	Δ	Δ	Δ	Δ	Δ	Δ	Δ
	3	Δ		?Δ						?Δ			?		Δ		Δ	Δ	?Δ	?	Δ
	4	Δ	Δ	Δ	Δ	Δ				?Δ	Δ	?Δ	?Δ			Δ	Δ	Δ	Δ	Δ	Δ
Illinoian	6	Δ	Δ	Δ	Δ	Δ	Δ	Δ	Δ	Δ	?		Δ	Δ	Δ	Δ	?	Δ	?Δ	Δ	Δ
Pre-Illinoian A	8	?Δ	Δ	?Δ				?Δ								Δ				?Δ	Δ
Pre-Illinoian B	?12	?Δ	?Δ	Δ			Δ	Δ	?Δ							Δ					Δ
Pre-Illinoian C	?14			Δ																	Δ
Pre-Illinoian D	16	Δ	?Δ	Δ	Δ		Δ	Δ								Δ					Δ
Pre-Illinoian E	?18 or 20	?Δ	?Δ	Δ					Δ												Δ
Pre-Illinoian F	M 22+	?Δ		Δ	Δ		Δ	?Δ		Δ				Δ		Δ		Δ	?Δ	?Δ	Δ
Pre-Illinoian G	34–38	Δ		?Δ	?Δ			?Δ	?Δ										?Δ	?Δ	Δ
Pre-Illinoian H	60	Δ		?Δ	Δ			?Δ													Δ
Pre-Illinoian I, J	64–72	Δ		?Δ				?Δ													Δ
	?78–82	Δ			?Δ			?Δ									?Δ				Δ
Plio/Pleistocene (pre-Illinoian K)	ca. 98–104	Δ			Δ			?Δ	Δ												Δ
Pliocene	G Upper	Δ																			Δ
	Lower	?Δ																			Δ
Miocene	Upper																				Δ
	Lower																				

come close to disappearance during the Eemian Stage interglacial (ca. MIS 5e) (Funder et al., 2011).

South America/Antarctica

As already noted above, the glacial record of the extreme Southern Hemisphere is the longest in the world during the Cenozoic, with ice already having formed in the late Eocene–early Oligocene in East Antarctica and built up in a step-like pattern through the Neogene (Table 3). Minor ice recession occurred in the youngest Miocene, and considerable fluctuations occurred during the Pliocene. The present polar conditions were already established by the early Pleistocene after 2.5 Ma. Similar Neogene glaciation is known from the Piedmont areas of Argentina and Chile, where, like Antarctica, substantial ice caps were established by 14 My ago. Widespread lowland glaciation between 2.05 and 1.86 Ma (ca. MIS 68–78), followed by the so-called 'Initioglacial' or 'Great Patagonian Glaciation,' took place at 1.15–1.00 Ma (ca. MIS 30–34), with subsequent less extensive glaciation occurring repeatedly up to the present.

Further north, the earliest glaciation recorded in the Bolivian Andes dates from 3.27 Ma, with extensive events at 2.2 Ma (ca. MIS 82), etc. In Columbia, the record begins at 2.5 Ma, with glacial outwash sequences extending well into the Bogota basin. As in Europe and North America, glaciation increased in

intensity throughout the Andean chain from 800 ka to the present day (see [Neoglaciation in Europe, Transition from Late Neogene to Early Quaternary Environments](#)). In Central America, the longest record is from Mexico, where oldest moraines on volcanoes have been dated to 195 ka and probably relate to a pre-MIS 6 glaciation.

Central Asia

Because of the relatively late uplift of High Asia, glaciation of Tibet and Tianshu is not recorded before the middle Pleistocene, of which MIS 12 was the most extensive (Table 3). In Tianshu, older glaciation (?MIS 16) may also have occurred. Subsequent events took place during MIS 8, 6, and 4–2 and continue even today in the highest peaks (see [Late Quaternary in Highland Asia](#)).

Australasia

Glaciation in Australasia, apart from that on some high volcanoes in Papua and Indonesia, occurred in both Tasmania and New Zealand (see [Eustatic Sea-Level Changes Since the Last Glacial Maximum](#)) (Table 3). The earliest records in both islands are from the Plio–Pleistocene (2.6 Ma: MIS 98–104), although it may be older in Tasmania. Only slightly younger is

Table 3 Occurrence of glaciation in the rest of the world through the Cenozoic based on numbers of observations presented in contributions to the INQUA project 'Extent and Chronology of Quaternary Glaciations' (Ehlers, Gibbard and Hughes, 2011)

Stage/Age		MIS (approx)	Antarctica and Southern Ocean	Argentina	Patagonia	Bolivia	Chile	China - Tibet	China - Tianshu	Colombia	East African mountains	Ecuador	Mexico	New Zealand	Tasmania
Weichselian/Wisconsinan		2	Δ↓	Δ	Δ	Δ	Δ	Δ	Δ	Δ	Δ	Δ	Δ	Δ	Δ
		3	Δ↓			Δ	Δ	Δ	Δ				Δ	Δ	
		4	Δ↓			?Δ	Δ	Δ	Δ		?Δ		Δ	Δ	
Saalian/Illinoian		6	Δ↓	Δ	Δ	Δ	Δ	Δ	Δ	Δ	Δ	Δ	Δ	Δ	Δ
		8	Δ↓			?Δ	Δ	Δ	Δ		?Δ		Δ	Δ	
Elsterian/pre-Illinoian B		?12	Δ↓	Δ		?Δ	Δ	Δ	Δ	Δ			Δ		?Δ
Early Middle Pleistocene		?14	Δ↓			Δ	Δ		Δ						
		16	Δ↓			?Δ	Δ	?Δ	Δ						
	B	?18 or 20	Δ↓		Δ	Δ			Δ						
Early Pleistocene	M	22+	Δ↓	Δ		Δ			?Δ	Δ					
		34–38	Δ↓	Δ		Δ									?Δ
		60	Δ↓			Δ									
		64–72	Δ↓	Δ		Δ									
		78–82	Δ↓		Δ	Δ				Δ			?Δ		
Plio/Pleistocene		c.100–104	Δ↓			Δ			?Δ				?Δ		
Pliocene	G	Upper	Δ↓		Δ	Δ									
		Lower	Δ↓	Δ		Δ									
Miocene		Upper	Δ↓			Δ									
		Lower	Δ↓	Δ		Δ									
Paleogene		Paleogene	Δ↓												

the oldest known glacial event in New Zealand (the Porika Glaciation). This is followed by a 1 My break before the glacial record continues in MIS 12, followed by MIS 6, 4, and 3. In Tasmania, an early middle Pleistocene event, possibly during MIS 16, is followed by glaciations during MIS 6 and 3.

Africa

The glacial record in Africa is restricted to the mountains of East Africa and the High Atlas (Table 3). On Mount Kilimanjaro, glaciation began at ~2.0 Ma (?ca. MIS 68), while a second glaciation is recorded before the Brunhes/Matuyama magnetic reversal. Younger glaciations appear to be broadly equivalent to those elsewhere, that is, during MIS 12, 6, and 2. There is evidence for probably three glaciations of the Atlas mountains, but their age remains equivocal at present. Very limited glaciation may also have occurred in South Africa (Hall and Meiklejohn 2011).

See also: **Glaciation, Causes:** Astronomical Theory of Paleoclimates. **Glaciations:** Early Quaternary (Pleistocene) and

Precursors; Late Pleistocene in South America; Late Quaternary in North America; Late Quaternary of Antarctica; Late Quaternary of the Southwest Pacific Region; Middle Pleistocene Glaciations in the Southern Hemisphere; Mid-Quaternary in North America.

Paleoceanography, Physical and Chemical Proxies: Oxygen-Isotope Stratigraphy of the Oceans.

References

- Cohen KM and Gibbard P (2011) *Global Chronostratigraphical Correlation Table for the Last 2.7 Million Years*. Cambridge, England: Subcommission on Quaternary Stratigraphy (International Commission on Stratigraphy. <http://www.quaternary.stratigraphy.org.uk/charts/>).
- Ehlers J and Gibbard PL (eds.) (2004a) Quaternary Glaciations – Extent and Chronology, Part I: Europe. *Developments in Quaternary Science*, vol. 2a. Amsterdam: Elsevier.
- Ehlers J and Gibbard PL (eds.) (2004b) Quaternary Glaciations – Extent and Chronology, Part II: North America. *Developments in Quaternary Science*, vol. 2b. Amsterdam: Elsevier.
- Ehlers J and Gibbard PL (eds.) (2004c) Quaternary Glaciations, Extent and Chronology – A Closer Look, Part III: South America, Asia, Africa, Australasia, Antarctica. *Developments in Quaternary Science*, vol. 2c. Amsterdam: Elsevier.

- Ehlers J, Gibbard PL, and Hughes PD (eds.) (2011) Quaternary Glaciations – Extent and Chronology, Part IV – A Closer Look. *Developments in Quaternary Science*, vol. 15. Amsterdam: Elsevier.
- Funder S, Kjellerup Kjeldsen K, Kurt Henrik Kjær KH, and Ó Cofaigh C (2011) The Greenland ice sheet during the last 300,000 years: A review. In: Ehlers J, Gibbard PL, and Hughes PD (eds.) *Quaternary Glaciations – Extent and Chronology, Part IV – A Closer Look. Developments in Quaternary Science*, vol. 15. Amsterdam: Elsevier.
- Gibbard PL, Boreham S, Cohen KM, and Moscardiello A (2004) *Global Chronostratigraphical Correlation Table for the Last 2.7 Million Years*. United Kingdom: Quaternary Palaeoenvironments Group, Godwin Institute of Quaternary Research, University of Cambridge.
- Gibbard PL, Smith AG, Zalasiewicz JA, et al. (2005) What status for the Quaternary? *Boreas* 34: 1–6.
- Hall K and Meiklejohn I (2011) Glaciation in southern Africa and in the sub-Antarctic. In: Ehlers J, Gibbard PL, and Hughes PD (eds.) *Quaternary Glaciations – Extent and Chronology, Part IV – A Closer Look. Developments in Quaternary Science*, vol. 15. Amsterdam: Elsevier.
- Haug HH, Ganopolski A, Sigman DM, et al. (2005) North Pacific seasonality and the glaciation of North America 2.7 million years ago. *Nature* 433: 821–825.
- Muttoni G, Carcano C, Garzanti E, et al. (2003) Onset of Pleistocene glaciations in the Alps. *Geology* 31: 989–992.

Transition from Late Neogene to Early Quaternary Environments

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Introduction

Definition of the Plio-Pleistocene Boundary

As published by Pratt et al. (2009), a strong majority of International Commission on Stratigraphy voting members approved the proposal that the base of Pleistocene Series/Epoch, likewise the base of the Quaternary System/Period, and thus the Neogene–Quaternary boundary are lowered such that the Pleistocene now includes the Gelasian Stage/Age and its base is defined by the Monte San Nicola GSSP in Sicily. This base also defines the base of the Gelasian. With these definitions, the Gelasian Stage/Age is transferred from the Pliocene Series/Epoch to the Pleistocene (Gibbard et al., 2009).

Accordingly, the base of the Quaternary now occurs near the end of a stepwise long-term climatic deterioration, at 2.61 Ma, which is coeval with marine isotope stage (MIS) 104 and the Gauss–Matuyama geomagnetic boundary (Figure 1). The deterioration had started about 3.0 Ma, subsequent to interglacial MIS G17 and glacial MIS G12, and was linked to the onset of major Northern Hemisphere glaciation (NHG), in particular, on Greenland. This event formed a major tipping point in the long-term benthic oxygen isotope ($\delta^{18}\text{O}$) history of the Cenozoic (Lisiecki and Raymo, 2005; Zachos et al., 2001). By contrast, the top of the Gelasian at ~ 1.8 Ma, formerly used to define the base of the Pleistocene, did not correspond to any major event in the Earth's climate history. In these discussions, evidence from ocean sediments is dominant, because the necessary stratigraphic control relies heavily on $\delta^{18}\text{O}$ records that were measured on Foraminifera in pelagic and hemipelagic sediments.

This article discusses various potential forcings and causal chains that may have triggered the great change in global environment, known as the onset of the Quaternary. Many lines of evidence and argumentation are directly adopted from Sarnthein et al. (2009) after including various findings published more recently.

Conceptual Models that Try to Explain the Onset of Major NHG

The potential forcings for the onset of major NHG, in particular, for the buildup of a continent-wide ice sheet on Greenland and the associated planetary cooling, do not form a merely academic topic. Rather, the forcings present a burning question, because the scenario of the warm upper Pliocene that preceded a full glaciation of Greenland may serve as an analog for the uncertain Earth's future climate in view of global warming in terms of albedo, sea-level rise, and continental aridity and/or humidity (Dowsett et al., 2010; Haywood, 2009). A number of hypotheses tend to explain the potential forcings that may have controlled the onset of 'major' NHG, as summarized by Sarnthein et al. (2009) and Seki et al. (2010). Willenbring and von Blanckenburg (2010) strongly questioned the model of globally increased rates of weathering and erosion that have triggered a late Neogene drawdown

of atmospheric CO_2 and successive cooling. In this article, four other theories are discussed in the context of paleoclimatic data sets. With regard to glaciations, we need to stress the term 'major,' because minor, regional ice sheets have already been registered in the Northern Hemisphere, in particular on Greenland, since the late Eocene (Eldrett et al., 2007) and in particular, since the late Miocene, as summarized in Paleogeography volume 23/3 (2008) and by De Conto et al. (2008). However, the major rise in the discharge of ice-rafted debris (IRD) to the ambient seas, which suggests the beginning of Quaternary-style full glaciations, only occurred near 3.0–2.7 Ma, as first outlined by Shackleton et al. (1984) and displayed further below.

The role of orbital forcing

Mere changes in orbital forcing of climate may form general thresholds and thus act as potential mechanism to initiate the buildup of major NHG, first proposed by Maslin et al. (1996) and recently by several other authors. Indeed, the amplitudes of orbital obliquity cycles increased significantly from ~ 3.0 to 2.5 Ma. This trend closely paralleled a major increase in global ice volume, as recorded in the benthic $\delta^{18}\text{O}$ signal (Figure 1). However, the increase in the amplitudes of obliquity was reversible (at 2.25–1.8 Ma; Laskar et al., 2004), whereas the onset of major NHG was not or only reversible to a minor degree and thus cannot be properly explained.

Huybers and Molnar (2007) established a more sophisticated approach to test the potential influence of orbital forcing on the onset of major NHG. They calculated the number of positive degree days (PDD) for North America $>50^\circ$ N versus the equator, obtained by applying temperature perturbations at high latitudes, expected to result from tropical anomalies and orbital variations (Figure 2(b)). After ~ 4.2 Ma, the PDD number decreased gradually down to a lowstand that persisted after 1.8 Ma. This trend preceded the start of the long-term cooling of sea-surface temperatures (SST) in the eastern equatorial Pacific by ~ 0.5 My (Figure 2(a)); likewise, it preceded the buildup of major NHG by more than 1 My (Figure 2(c)). This time span requires an extremely long-lasting internal memory effect in the climate system, difficult to conceive, thus leading the authors to the honest conclusion: "Understanding the puzzle of what caused the eastern equatorial Pacific to cool off on a timescale too long to invoke the atmosphere–ocean–cryosphere system by itself now appears all the more pressing."

The role of the Indonesian gateway systems

Two hypotheses invoke as a cause for the onset of Quaternary-style climates: either a significant constriction or closure of major ocean gateway systems, the Indonesian and the (former) Panamian/Central American seaways (CAS), which are fundamental for the reallocation of heat and sea-surface salinity (SSS) among the different tropical oceans.

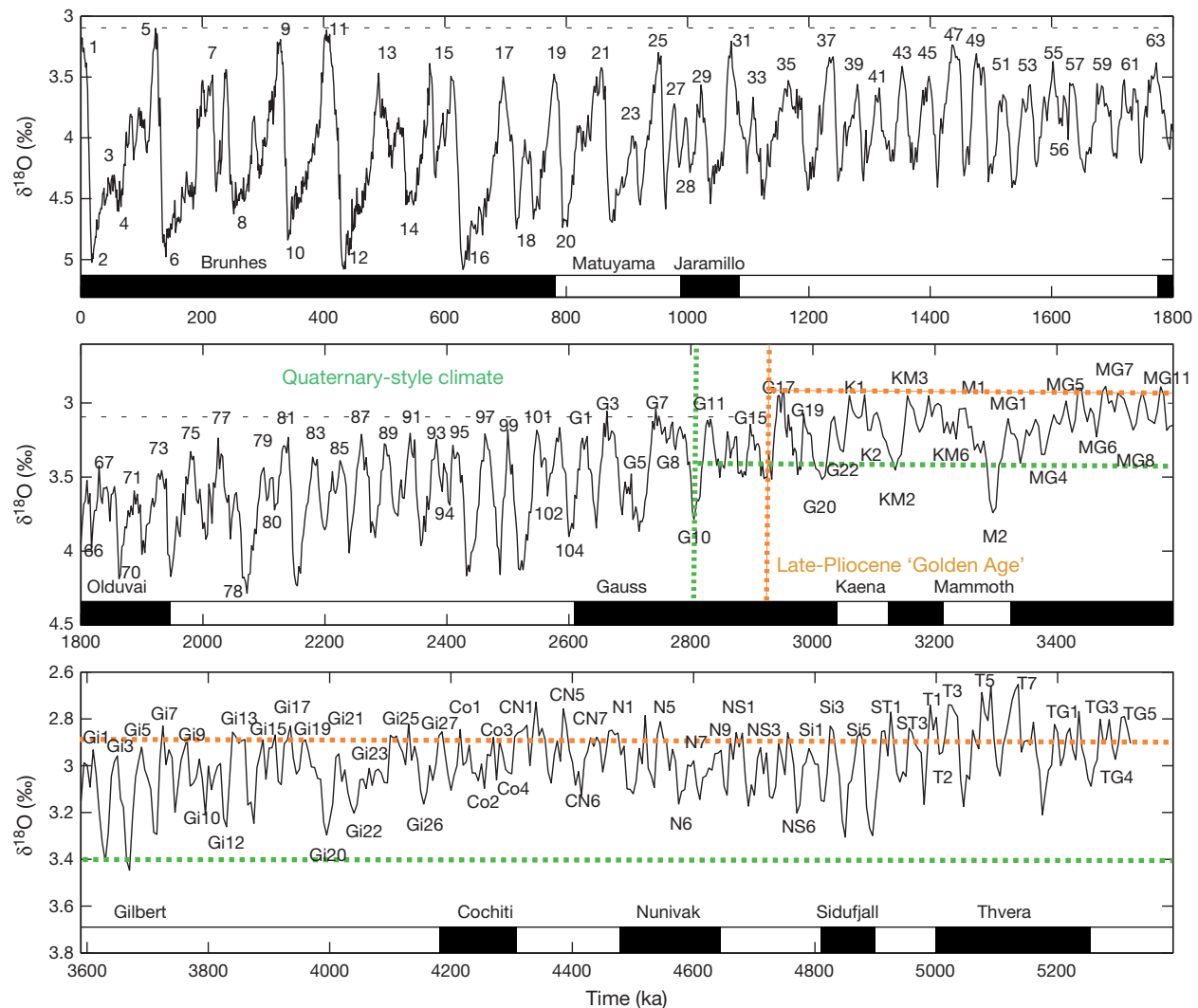


Figure 1 'LR04' benthic $\delta^{18}\text{O}$ record of global ice volume changes over the last 5 Myr (Lisiecki and Raymo, 2005; modified from Sarnthein M, Bartoli G, Prange M, et al. (2009) Mid-Pliocene shifts in ocean overturning circulation and the onset of Quaternary-style climates. *Climate of the Past* 5: 269–283). Transition from the Late Pliocene 'Golden Age' (2.9–3.4‰ $\delta^{18}\text{O}$) to Quaternary-style climates near 3.0–2.8 Ma. Note the variable $\delta^{18}\text{O}$ scales. The 'Golden Age' represents a state of small scale and fairly uniform climate oscillations that continued since 4.0 Ma, subsequent to a further 0.2–0.3‰ increase of cold $\delta^{18}\text{O}$ excursions only near 4.2–4.0 Ma in the middle early Pliocene. Numbers are marine isotope stages (MIS).

Cane and Molnar (2001) first proposed the gradual closure of the Indonesian seaways near 4 Ma as forcing important to set the stage for the onset of Quaternary-style climates. More precisely, they suggested that the northward plate-tectonic shift of the (previously little known) small and elongate island of Halmahera, northwest of New Guinea, finally barred the West Pacific Warm Pool (WPWP) from warming the Indonesian Throughflow (ITF) subsequently being fed by cooler subsurface waters from the subtropical Northern Pacific (Figure 3). On the other hand, near-surface heat export from the WPWP was then diverted to the northwest Pacific.

Karas et al. (2009) constrained a distinct cooling and freshening of subsurface waters in the Eastern Indian Ocean and, thus, of the ITF more closely, to 3.3–3.2 Ma, thus indeed suggesting a linkage to the onset of Quaternary-style climates. In particular, the progressive restriction of the ITF may have implied a cooling of West Indian equatorial surface and

upwelling waters and thus have reduced evaporation, which should result in a decrease in moisture of the African and Indian monsoon. However, East African lake records just reflect a reversed trend, a prominent lake maximum and humid event near 3.1 Ma (Trauth et al., 2005). Accordingly, the actual implications of the restriction of the Indonesian gateway system for large-scale climate change still need further clarification.

The Panama hypothesis

Various recent lines of evidence suggest direct linkages between the final closure of the CAS, more precisely at Panama, and several follow-up processes relevant to climate change in the North Atlantic, northern North Pacific, and Arctic Oceans (Sarnthein et al., 2009, and references summarized therein). The final closure of CAS presents the most recent plate-tectonic change of ocean configuration in geological history, an uplift process related to the subduction of the Galapagos aseismic

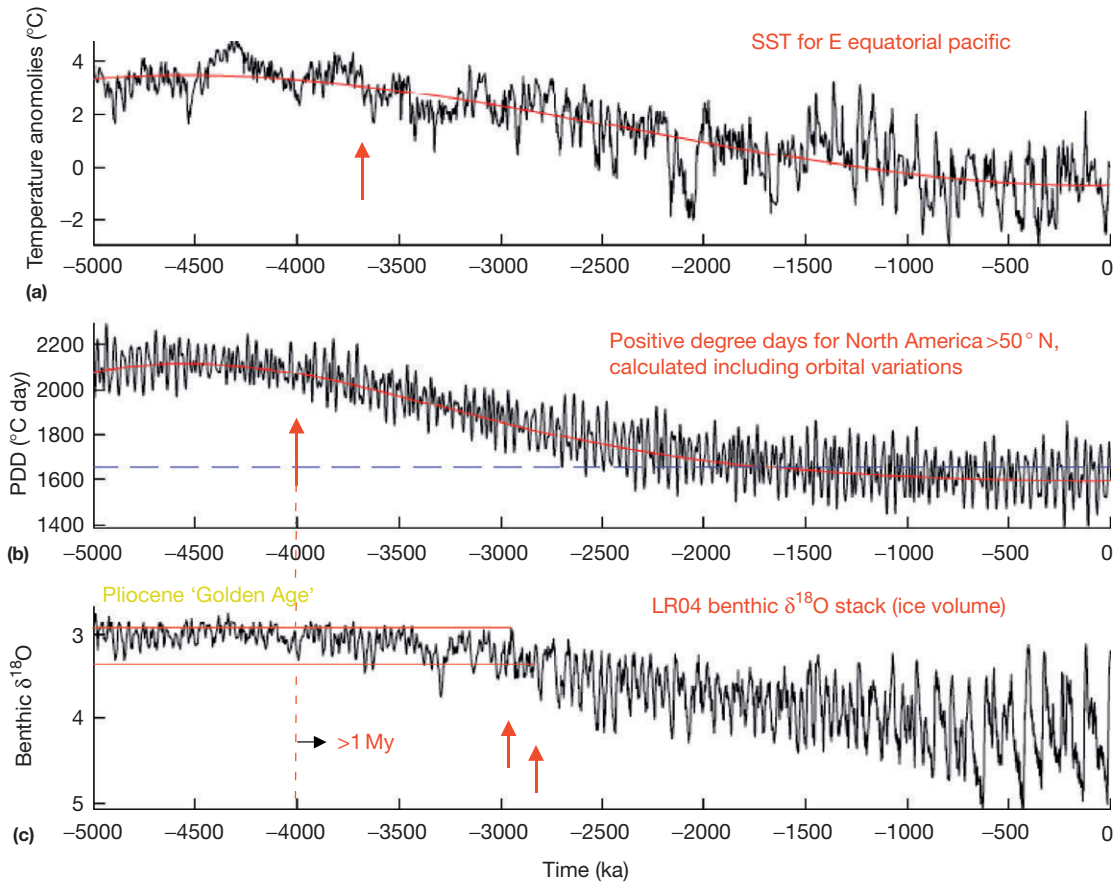


Figure 2 Model of Huybers and Molnar (2007, suppl.) on (b) North America's 'positive degree days' (PDD), compared with East Pacific cooling/temperature anomalies (Lawrence et al., 2006) (a), and the $\delta^{18}\text{O}$ record of growth of global ice volume (Lisiecki and Raymo, 2005) over the last 5 My (c). Vertical arrows mark major tipping points in climate records. The time span between the onset of reduced PDD and the end of the Pliocene 'Golden Age' amounts to >1 My. Modified from Sarnthein M, Bartoli G, Prange M, et al. (2009) Mid-Pliocene shifts in ocean overturning circulation and the onset of Quaternary-style climates. *Climate of the Past* 5: 269–283.

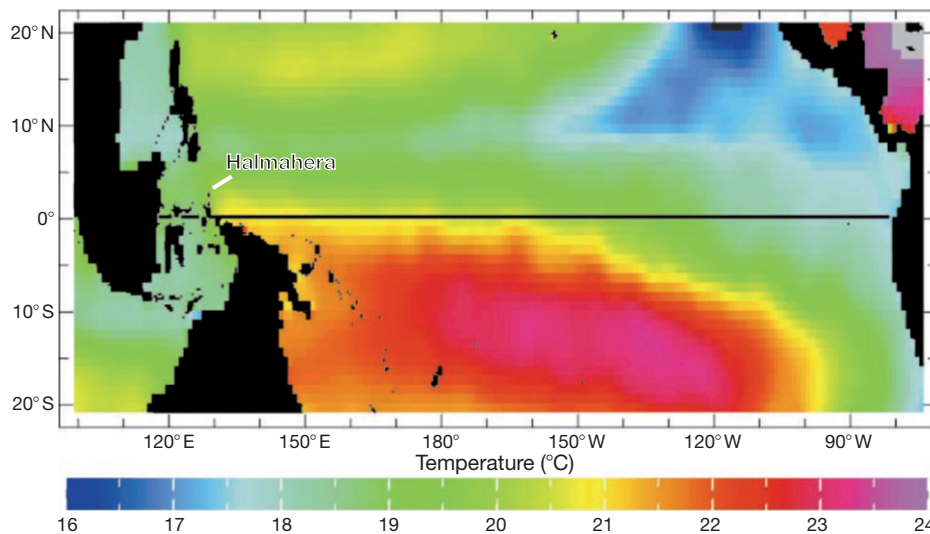


Figure 3 Location of the island Halmahera, today preventing the West Pacific Warm Pool from flowing into the Indonesian throughflow (Cane and Molnar, 2001). From Sarnthein M, Bartoli G, Prange M, et al. (2009) Mid-Pliocene shifts in ocean overturning circulation and the onset of Quaternary-style climates. *Climate of the Past* 5: 269–283.

ridge (Meschede and Barckhausen, 2001). Weyl (1968) and Keigwin (1982) first proposed this event as potential forcing for the onset of Quaternary-style climates. On the basis of diverging nannofossil communities, Kameo and Sato (2000) first succeeded in pinning down the final closure of CAS to an age close to 2.76 Ma (MIS G6; equal to 2.73 Ma on the revised timescale LR04; Lisiecki and Raymo, 2005), a timing that indeed comes close to the onset of major NHG as first derived by Shackleton et al. (1984). The closure may have triggered a rise in both the poleward heat and moisture transports and in meridional overturning circulation (MOC) of the Atlantic, which in turn have resulted in a lowering of atmospheric $p\text{CO}_2$; thus, the above phenomena may have finally been responsible for the full glaciation of Greenland and for the numerous and gradually intensifying Quaternary-style glaciations in Laurentia and northern Eurasia.

Driscoll and Haug (1998) assessed in detail the causal chains of climate change linked to the 'Panama Hypothesis,' trying to reconcile the effects of two actually counteracting forcings, increased poleward heat and moisture transports for the buildup of a continent-wide ice sheet on Greenland. Driscoll and Haug postulated that stronger heating of the northern North Atlantic will result (1) in enhanced evaporation and moisture transport via the westerlies to northwestern Eurasia, as also displayed by the model of Lunt et al. (2007), (2) in an increased river discharge from Siberia to the Arctic Ocean, in turn leading to (3) enhanced sea-ice formation and (4) an increase in Arctic albedo, an important feedback factor, and, therefore, to (5) more favorable conditions for the growth of Northern Hemisphere ice sheets.

On the other hand, Molnar (2008) listed extensive doubts on the proposed linkages between closing CAS and the onset of Quaternary-style Ice Age climates. In particular, he was concerned about the proposed late Pliocene age of the closure and the possibility of establishing an approximate simultaneity between the closure and the first major ice advance. Moreover, he challenged the feasibility of reconstructing a precise record of the progressive and final closure of CAS. Notably, he stressed the uncertainties in deriving the closure from an increase in (sediment proxy-based) SSS differences between the Caribbean and the eastern Equatorial Pacific. Also, he doubted the reliability of conclusions deduced from sediment records of concurrent low- and high-latitude changes. Finally, he pointed out the apparently differential evidence simulated by different general circulation models. In this article, we need to handle these questions with care.

The role of a decline in atmospheric CO_2

On the basis of a coupled general circulation model, Lunt et al. (2008) made a strong argument that a decline in the late Pliocene atmospheric CO_2 content from 400 to 280 ppmv needs to cause the buildup of major NHG at the onset of the Quaternary (Figure 4). This decline in $p\text{CO}_2$ did not affect the precipitation patterns over Greenland and the ambient sea regions but triggered a significant decrease in surface temperatures by 2–3 °C all over Greenland, thus promoting major ice growth. None the less, the origin of that reduction in the greenhouse effect still awaits a concise explanation.

Until recently, the CO_2 hypothesis suffered from the problem that published (marine and terrestrial) sediment records

provided hardly any satisfying evidence for the postulated late Pliocene drop in atmospheric CO_2 . Seki et al. (2010) now succeeded in generating two pertinent and strongly parallel high-resolution CO_2 records on the basis of two independent proxies in a tropical marine sediment core: (1) boron isotopes ($\delta^{11}\text{B}$) measured on the planktic surface dweller *Globigerinoides ruber* and (2) $\delta^{13}\text{C}$ of long-chain alkenones of haptophyte algae and planktonic Foraminifera calcite (ϵ_{p37-2}), values that were corrected for differential coccolith growth sizes. Both records reflect a Pliocene CO_2 level of 300–370 ppmv from ~4.9 to 3.1 Ma (Figure 4), which indeed slightly exceeded the preindustrial CO_2 level of 285 ppmv, characteristic of Quaternary interglacials. However, with the first onset of climatic deterioration after 3.1 Ma (MIS K1), the CO_2 values did not fall but rise markedly up to >400 ppmv – a prominent CO_2 maximum that terminated with a sudden drop near 2.85 Ma (immediately prior to MIS G10). This drop marked the onset of a largely constant CO_2 plateau at ~250–290 ppmv, characteristic of Quaternary interglacials.

These CO_2 records leave us with the open question as to which kind of change in the global carbon reservoirs, more specifically, which changes in ocean MOC, may have actually caused the fairly abrupt short-term rise and drop in atmospheric CO_2 during the latest Pliocene, that finally introduced the subsequent long-lasting CO_2 low of the Pleistocene. The CO_2 changes then may have served as a major amplifier mechanism of climate change, possibly similar to the CO_2 oscillations recorded for the orbital-scale climate changes of the last 400 ka (Kawamura et al., 2007). A top candidate mechanism to induce the outlined CO_2 changes may be linked to the final closure of the CAS, which triggered major changes in both the Atlantic MOC (AMOC) and a largely persistent sea-surface stratification of the subarctic North Pacific (Sigman et al., 2004), as outlined further below.

Final Closure of CAS as Trigger for the Onset of Quaternary Climates

Models, Concepts, and Implications

Maier-Reimer et al. (1990) first employed an ocean general circulation model to simulate the potential impact of the closure of CAS on ocean circulation and climate. Accordingly, the closure stopped the incursion of Pacific low-salinity and nutrient-enriched surface waters, diluting West Atlantic SSS in the Caribbean, here inducing a profound rise in poleward heat and salt transport and, in turn, an enhanced AMOC. Later, various more advanced general circulation model (GCM) tests basically confirmed these findings (e.g., Murdock et al., 1997). Schneider and Schmittner (2006) focused on testing the role of four subsequent Panamaian sill depths from –2000 up to 0 m in SSS and SST distributions in the adjacent sea regions. Different from previous concepts (Haug et al., 2001), they found that major Pacific-to-Atlantic contrasts in SSS at 50–125-m water depth emerged only at gateway sill depths of less than 500 m, whereas a barring of intermediate waters between 2000 and 700–500 m gateway depth did not produce this change. At sill depths of 300–250 m, only one-third of the final contrast in SSS was established, whereas at 130 m, two-thirds of the final contrasts

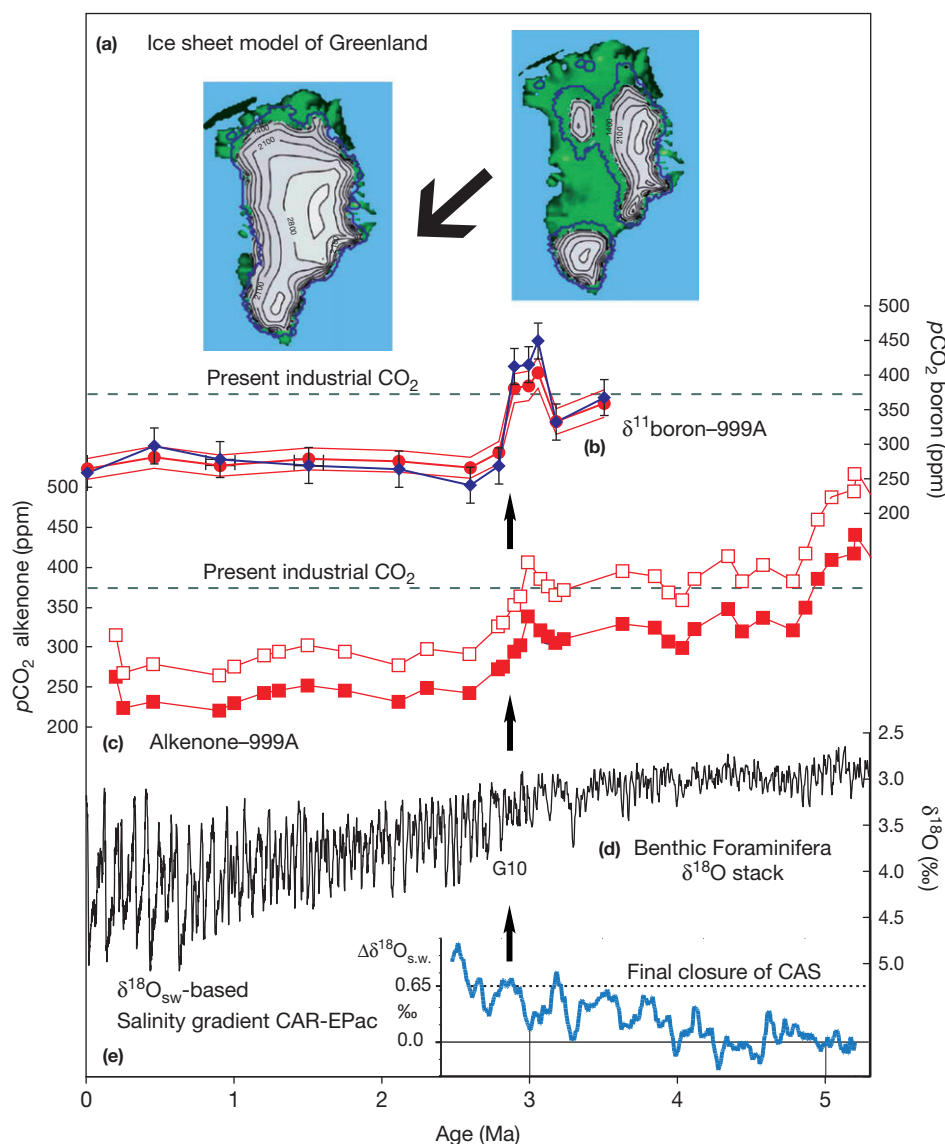


Figure 4 Plio-Pleistocene growth of Greenland ice sheet (a), merely induced by a decline in atmospheric $p\text{CO}_2$ from 400 to 280 ppmv (model simulation of Lunt et al., 2008). (b) $p\text{CO}_2$ evolution at tropical ODP Site 999 (modified from Seki O, Foster GL, Schmidt DN, Mackensen A, Kawamura K, and Pancost RD (2010) Alkenone and boron-based Pliocene $p\text{CO}_2$ records. Earth and Planetary Science Letters 292: 201–211), based on *G. ruber* boron isotope ($\delta^{11}\text{B}$) data, derived pH with $(\text{CO}_3^{2-})_{\text{modeled}}$ (blue diamonds), or modern total alkalinity (red cycles), and (c) based on alkenone-based estimates determined using ε_{p37-2} values with a size correction (red closed and open squares for $b=90$ and 108, respectively). (d) Growth of global ice volume reflected by global stacked benthic $\delta^{18}\text{O}$ record LR04 (Lisiecki and Raymo, 2005). (e) Salinity gradient between Caribbean and East Pacific, based on the $\delta^{18}\text{O}$ difference in surface waters (for further explanations, see Figure 7). Final full closures and a reopening of CAS are coeval with distinct drops and rises in atmospheric CO_2 and, finally, with glacial MIS G10.

in SSS (i.e., 1.2 psu) were already established (Figure 5). Accordingly, only the closing of the topmost 250–130-m depth of the Panamaian gateways finally turned out to be crucial for triggering major changes in AMOC and, thus, in the storage capacity of CO_2 in the sea and, accordingly, in climate.

The coupled and fully dynamic models of Klocker et al. (2005) and Lunt et al. (2007) in part confirmed the conceptual model of Driscoll and Haug (1998), claiming that the closure of Panamaian seaways led to a warming of the central-to-northern North Atlantic (Figure 6), more intense Atlantic thermohaline

circulation, and enhanced precipitation over Greenland and the Northern Hemisphere continents. On the other hand, these models suggest that these forcings in total were too small by an order of magnitude to overcome the coeval warming and produce the observed major ice-sheet growth that marked the onset of the Quaternary. Consequently, modelers stressed a necessary linkage to a significant decline in atmospheric CO_2 , as recently corroborated by Seki et al. (2010). However, the characterizations of the sea-ice albedo and the sea-ice rheology employed may present potential flaws in the model studies of Lunt et al. (2007, 2008; more details in Sarin et al., 2009).

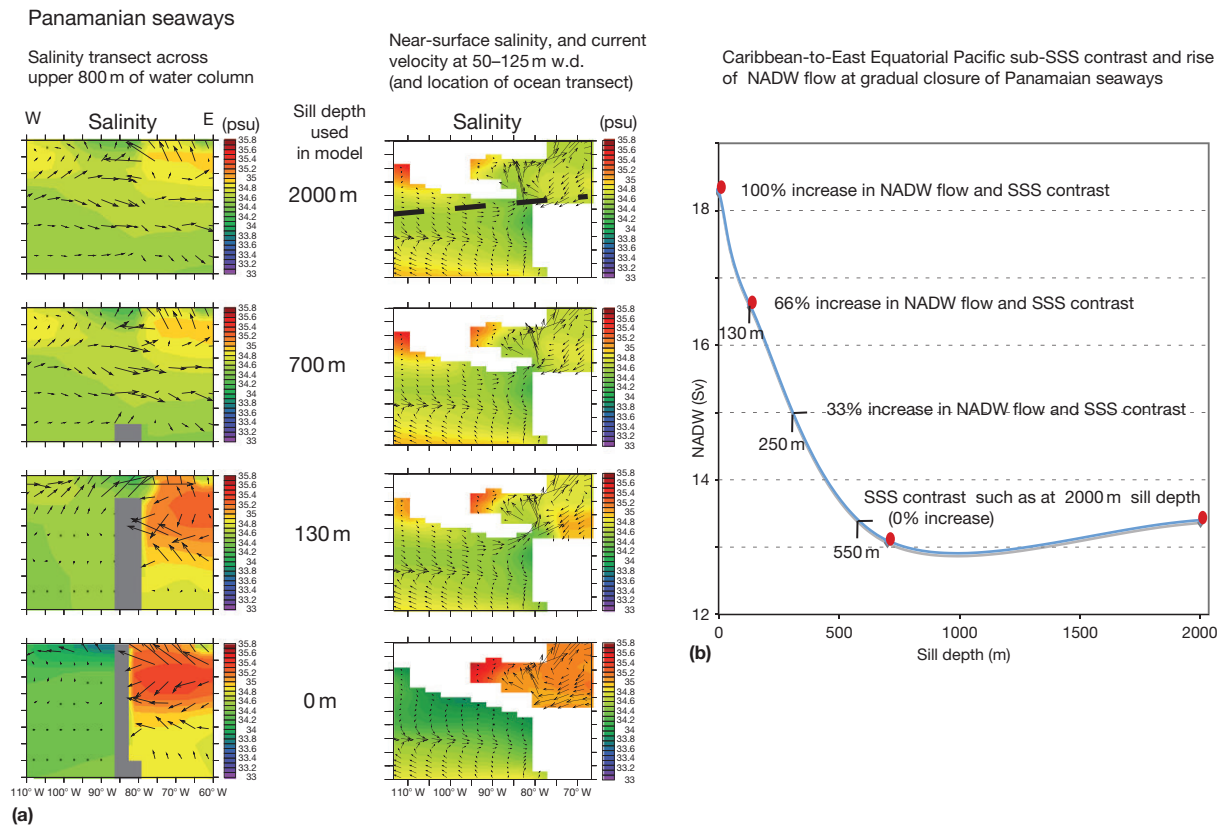


Figure 5 Impact of shoaling Panamanian seaways on Caribbean and equatorial east Pacific upper ocean salinities in a high-diffusion (HD) experiment with the UciV-ESCM model (Sarnthein et al., 2009; Schneider and Schmittner, 2006). (a) Model data for different sill depths plotted on vertical transect and map. (b) Increase in East-Pacific-to-Caribbean Sea-surface salinity (SSS) contrasts at 50–125 m water depth (w.d.) over different sill depths, deduced for waters ambient to Sites 999 and 1241 (site locations in Figure 8), and resulting increase in NADW flow, assuming a largely linear relationship with changes in SSS contrasts. Modified from Sarnthein M, Bartoli G, Prange M, et al. (2009) Mid-Pliocene shifts in ocean overturning circulation and the onset of Quaternary-style climates. *Climate of the Past* 5: 269–283.

Sediment Records on the Final Closing of CAS

The model-based SSS differences between the Caribbean and eastern Equatorial Pacific start to rise as soon as the Panamian sill depth has dropped to less than 250–300 m (Figure 5), here defined as impact of the final closure of the Panama seaways. The postulated rise in salinity anomalies of the mixed surface layer (at 40–80-m water depth; Steph et al., 2006) can be traced in continuous sediment records obtained from Ocean Drilling Program (ODP) sites 999 and 1241, immediately northeast and southwest of the Panama Isthmus (Bartoli et al., 2005; Groeneveld, 2005; Groeneveld et al., 2006). The records indeed show a major increase in the anomalies of sea water $\delta^{18}\text{O}$ over the late Pliocene, which reflects the expected rise in SSS differences (Figure 7).

The thermocline of the tropical Eastern Pacific was subject to a major shoaling already as early as 4.5–4.0 Ma (Steph et al., 2010). In contrast, the interoceanic SSS differences between the southwestern Caribbean and eastern Pacific remained close to zero until after 4 Ma. First but still reversible rises in the SSS gradient of the mixed layer occurred only near 3.75 and 3.5–3.3 Ma, when the gradient reached up to 0.5 psu and more. At 3.16 Ma (MIS KM3) it reached for a first time >0.8–1.0 psu. These short-lasting precursor signals of a major shallowing and/or closure of the Panama seaways may in part reflect short-term volcano-tectonic events and in part orbital ice and sea-level cycles. A largely irreversible SSS contrast began only after 2.95 Ma

(after MIS G17) and finally, at 2.6–2.5 Ma, when the difference values came close to the model-predicted SSS anomaly maximum. The latter two events suggest a final two-step closure of the shelf seaway induced by the uplift of the Panama Isthmus. At that time, the uplift factor obviously dominated any other superimposed orbital-scale sea-level fluctuations, which is also shown in Figure 7. Irrespective of some possible early diagenetic influences on the underlying Mg/Ca data (Groeneveld et al., 2008), the outlined record of SSS anomalies has a great advantage over various previously published discontinuous and in part poorly dated marine and biostratigraphical evidence on the closure of the Panama Isthmus (Molnar, 2008). This is particularly because of the clear orbital age control and the close resemblance to the modeled SSS trends (Figure 5). Various lines of land-based evidence from Panama likewise document a first full closure of the seaway, which occurred near 2.85 Ma only (Coates et al., 2004, and oral communication).

Northern Ocean Response to the Closing of CAS

North Atlantic Sediment Records Reflecting the Onset of Major NHG

Model-based evidence (Klocker et al., 2005; Lunt et al., 2007) suggests that the final closure of the CAS necessarily led to two counteracting climatic forcings with an opposite effect on

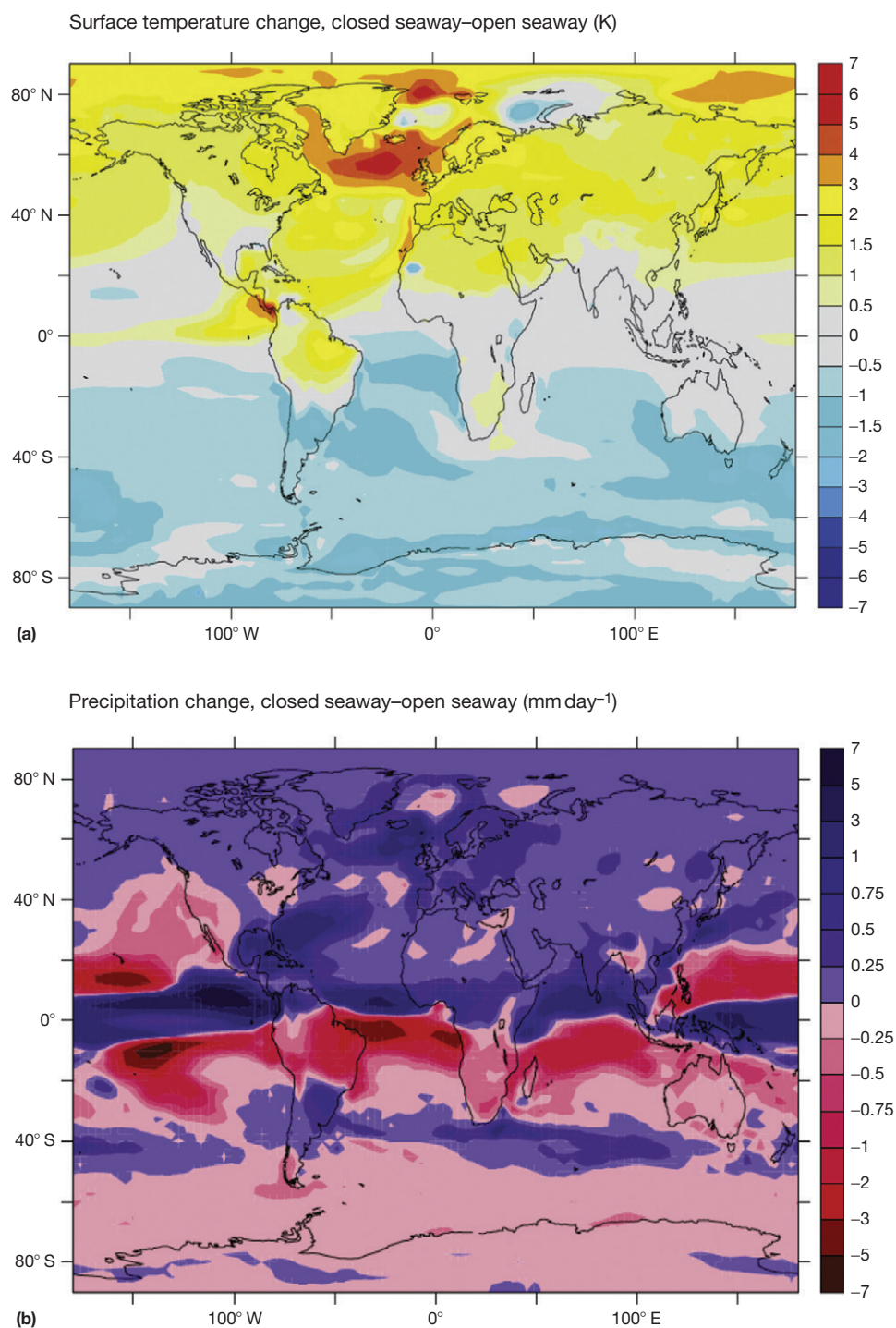


Figure 6 Changes in surface temperatures (a) and precipitation (b), obtained by comparing closed Panamanian seaways with open seaways (model experiments of Lunt et al., 2007). From Sarnthein M, Bartoli G, Prange M, et al. (2009) Mid-Pliocene shifts in ocean overturning circulation and the onset of Quaternary-style climates. *Climate of the Past* 5: 269–283.

high-latitude glaciation in Greenland (Figure 6). On the one hand, the closure strengthened the poleward heat transport that implied enhanced glacial melt. On the other hand, it promoted the poleward transport of atmospheric moisture to northern Eurasia, a freshening of the Arctic Ocean, and a series of consequential changes that resulted in snow and ice accumulation. Indeed, various sediment records from the northern North Atlantic provide independent evidence for

both transport processes and their linkages to the onset of major NHG together with the Neogene–Quaternary boundary.

The onset of major NHG is best documented at ODP Site 907 to the east of Greenland and the East Greenland Current (EGC; Figures 8 and 9). This position is ideal for monitoring the discharge of IRD from icebergs coming from northeast Greenland and the Arctic Ocean. A second IRD record measured at IODP Site 1307 near the southern tip of Greenland

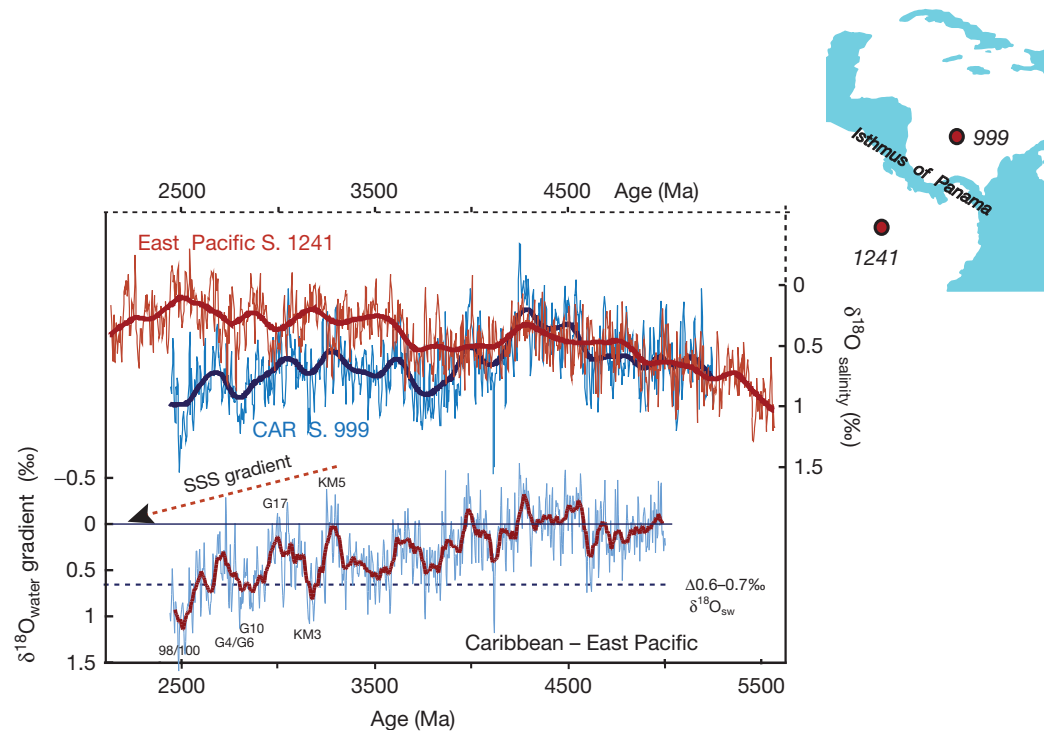


Figure 7 Late Pliocene increase in sea-surface salinity gradient between Caribbean (CAR) and eastern equatorial Pacific, based on paired *Globigerinoides sacculifer* Mg/Ca and $\delta^{18}\text{O}$ data for depth habitats of 40–80 m (records of Groeneveld, 2005; Steph et al., 2006). Modified from Sarnthein M, Bartoli G, Prange M, et al. (2009) Mid-Pliocene shifts in ocean overturning circulation and the onset of Quaternary-style climates. *Climate of the Past* 5: 269–283. A $\delta^{18}\text{O}_w$ gradient of 0.6–0.7‰ (hatched line; uncertainty of $\sim 0.07\text{‰}$ for the 10-point running mean of $\delta^{18}\text{O}_w$) equates to a SSS gradient of 1.2–1.8 psu, which corresponds to a full closure of CAS according to the modeled gradients shown in Figure 5(a). Age scale comes close to LR04 (Lisiecki and Raymo, 2005). Small letters and numbers at SSS gradient record (98/100, G6, etc.) label marine isotope stages.

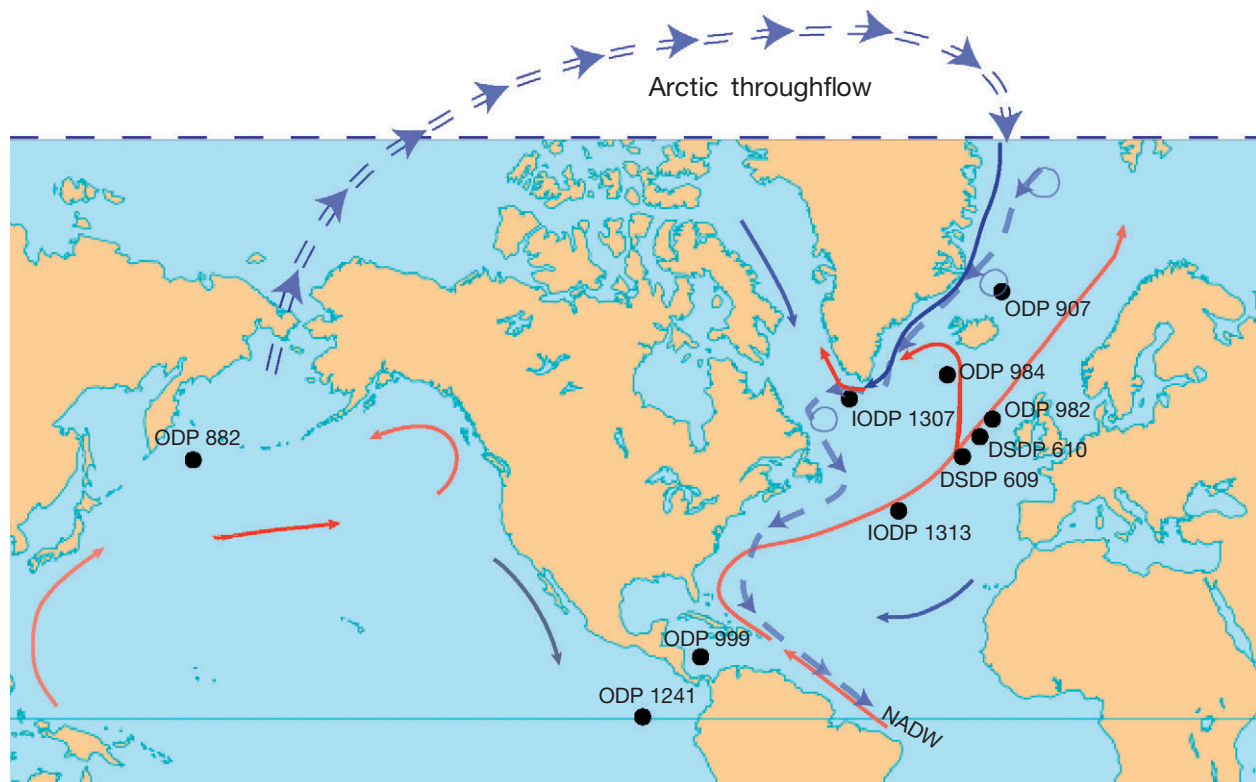


Figure 8 Deep Sea Drilling Project (DSDP), Ocean Drilling Program (ODP), and Integrated Ocean Drilling Program (IODP) site locations used in this article and various surfaces (warm, red; cold, blue) and deepwater currents (broken lines) in the North Atlantic and North Pacific.

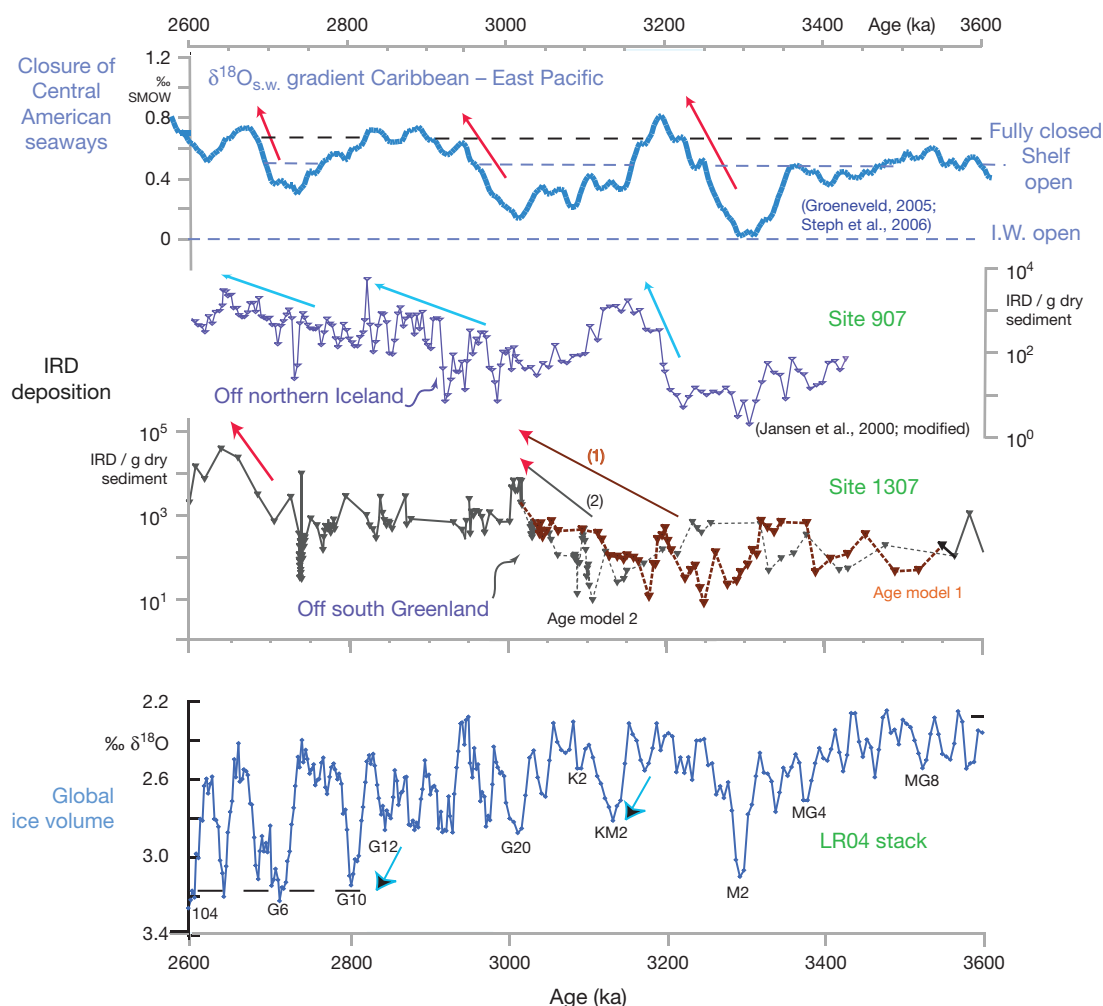


Figure 9 Closure of CAS (increasing $\delta^{18}\text{O}_{\text{sw}}$ differences between eastern tropical Pacific and southwest Caribbean, inverted from **Figure 7**) versus exponential changes in ice-rafted debris (IRD) at Site 907 (Bartoli et al., 2005; Jansen et al., 2000) and Site 1307 (plotted for two alternative age models) and the stacked benthic $\delta^{18}\text{O}$ record LR04 of global ice volume (Lisiecki and Raymo, 2005; small letters and numbers label marine isotope stages 104 – MG8). IW, intermediate water. The great IRD increase at Site 907 3.2–3.12 Ma matches a major cooling and freshening of the East Greenland Current (age model 1), shown in **Figure 12**. Modified from Sarnthein M, Bartoli G, Prange M, et al. (2009) Mid-Pliocene shifts in ocean overturning circulation and the onset of Quaternary-style climates. *Climate of the Past* 5: 269–283.

(**Figure 9**; Sarnthein et al., 2009) also includes IRD originating from mountain glaciers calving nearby into the Scoresby Sound and from southeast Greenland. Accordingly, IRD changes at Site 907 (Bartoli et al., 2005; Jansen et al., 2000) may best display the long term and major trends in the evolution of NHG, whereas IRD changes at Site 1307 rather reflect interesting aspects of short-term orbital-scale climate variability in southern Greenland. For comparison, the first IRD was deposited at the northeast Atlantic Site 552A to the west of Ireland at MIS G6, 2.72 Ma (Shackleton et al., 1984), approximately 110 ka prior to the official base of the Quaternary, defined at 2.61 Ma.

In harmony with the model of Driscoll and Haug (1998), the suite of changes in IRD at Site 907 from 3.4 to 2.6 Ma followed a closely similar suite of distinct steps in the evolution of CAS with great detail and a persisting lag of ~ 40 ka (**Figure 9**). This linkage applied, for example, to the closing

and opening events at 3.24–3.16, 3.13–2.97, and 2.96–2.87 Ma, except for the final episode of CAS closure after 2.7 Ma, when a major Greenland ice sheet had already preexisted from a preceding buildup. Each closing or opening of the CAS directly involved an exponential rise or drop in IRD abundance, respectively, with a final IRD extreme at MIS G10, 2.82 Ma.

However, the outlined coherence and phase lags of the IRD and CAS records alone do not yet provide adequate evidence to derive the origin of major NHG. Certainly, the persistent lag of changes in IRD may reflect significant but reasonable memory effects in the climate system. Moreover, the lags show that the oscillations in the buildup and melt of continental ice volume and resulting eustatic sea-level variations cannot have primarily controlled the closing events of CAS. Conversely, the CAS events need to have influenced the variations in IRD supply, in a way to be explored.

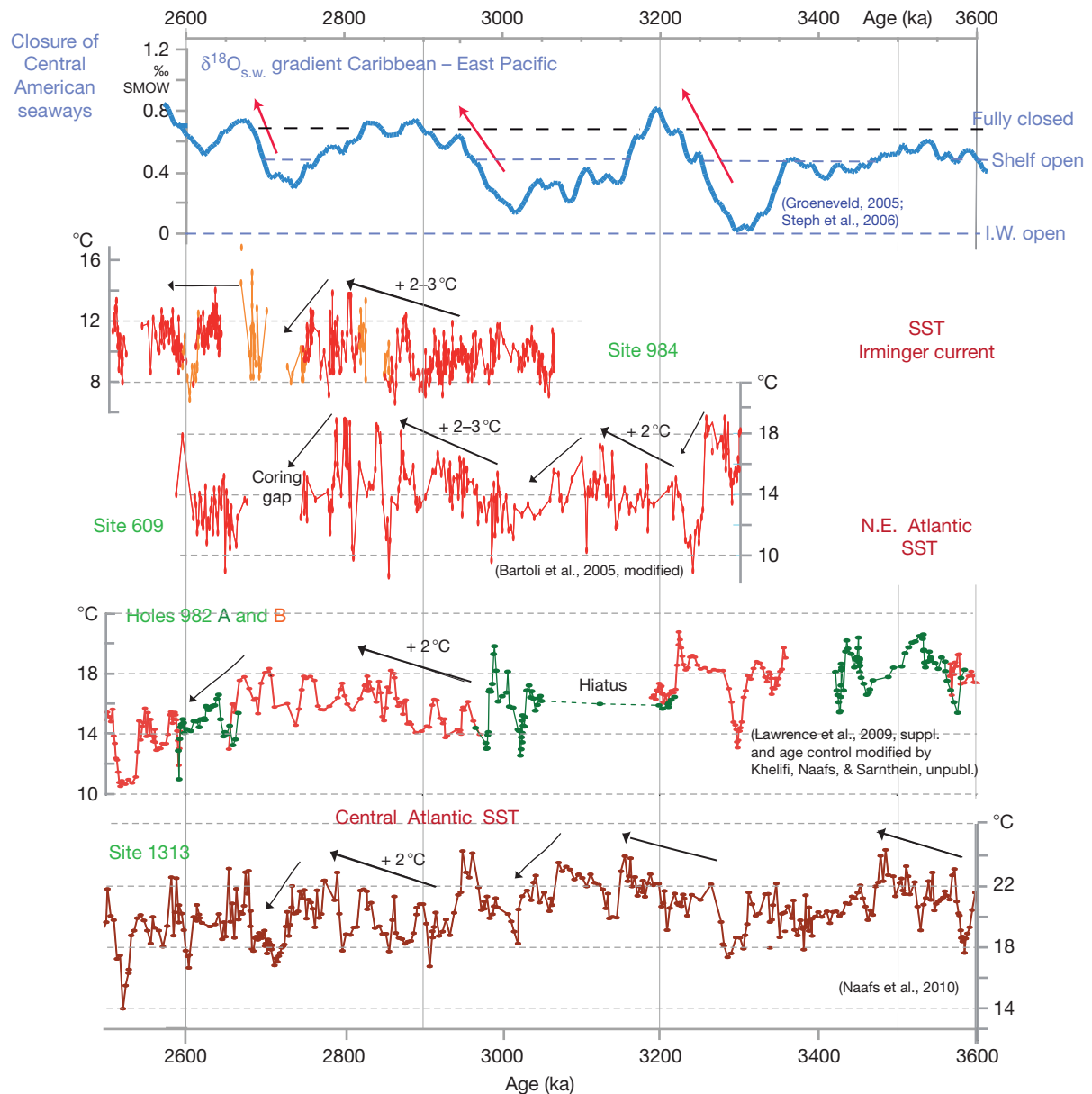


Figure 10 Closings of CAS ($\Delta\delta^{18}\text{O}_{\text{sw}}$ record inverted from [Figure 7](#)) and Plio-Pleistocene warmings at North Atlantic Sites 984 and 609 (based on Mg/Ca data; [Bartoli et al., 2005](#)), Site 982 (based on U^{37} ; [Lawrence et al., 2009](#), age scale modified and suppl. by [Khelifi et al., 2012](#)), and Site U1313 (based on U^{37} ; [Naafs et al., 2010](#)). Orbital age control of SST records may in part be slightly improved. Site locations are shown in [Figure 8](#).

Sediment Records of Increased Poleward Heat Transport, AMOC, and CO_2 Drawdown in the Ocean

Four continuous SST records with millennial-scale resolution were published from late Pliocene to early Pleistocene sediment sections in the northeastern and central North Atlantic ([Figures 9 and 10](#)); two of them were derived from Mg/Ca ratios of planktic foraminifers (S. 984 and S. 609) and two from the alkenone organic proxy Uk^{37} (S. 982, S. U1313). As predicted by modeling, all four SST records likewise reflect long-term coeval warmings and coolings. Each of them presented an increase or decrease in the poleward heat transport with the North Atlantic and

Irminger currents, coeval with an event of full closing or renewed opening of the CAS. For example, a distinct overall rise in SST by 2–3 °C occurred from 3.25 to 3.16 Ma and from 3.0/2.95 to 2.85/2.78 Ma, and conversely, a 4 °C drop occurred from 3.15 to 3.02 Ma. Each of these warmings and coolings includes up to four orbital-scale SST oscillations. The final closure at 2.73 Ma also induced a complete shut-down of the advection of nutrient-enriched surface waters from the eastern tropical Pacific to the North Atlantic, which in turn led to an abrupt productivity drop in the northern North Atlantic (reflected by the sudden appearance of the benthic foraminifer *Cassidulina teretis*; [Bartoli et al., 2005](#)).

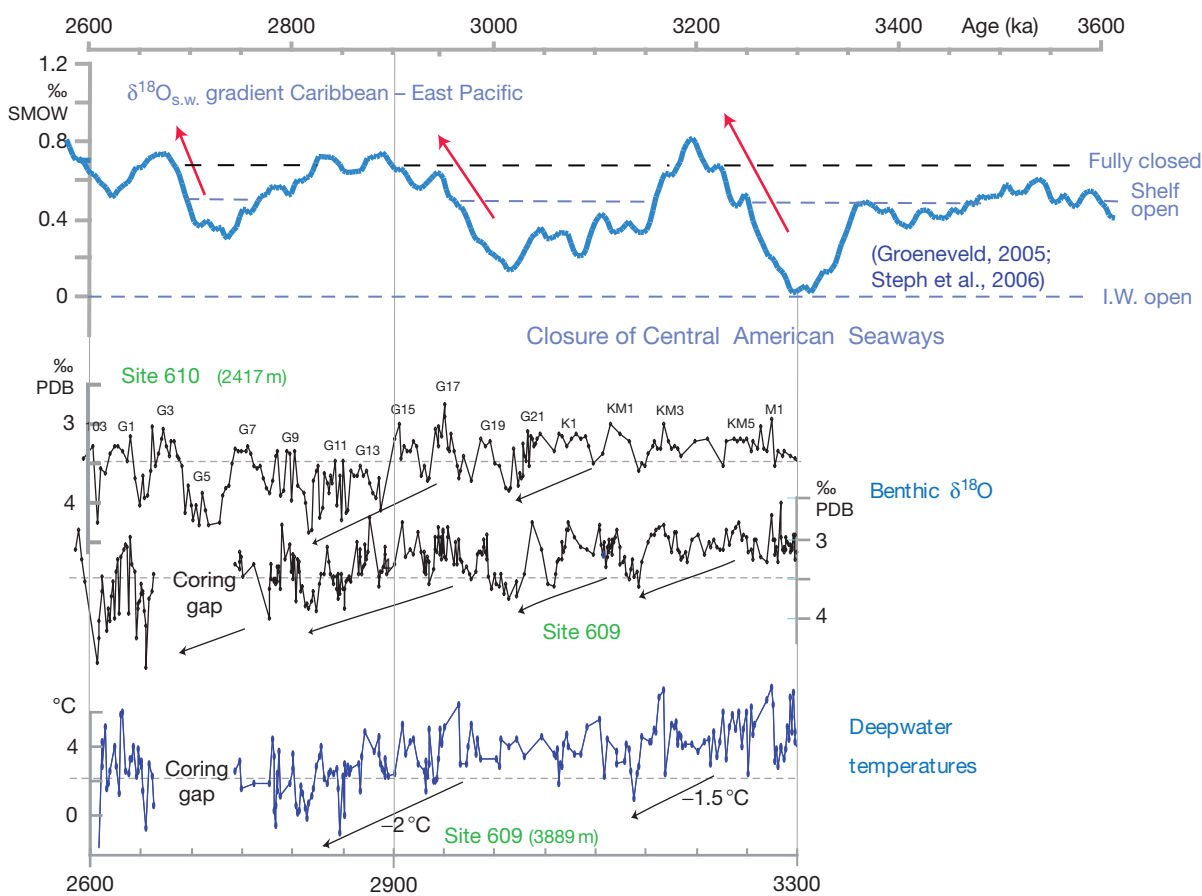


Figure 11 Closure of CAS ($\Delta\delta^{18}\text{O}_{\text{sw}}$ record inverted from [Figure 7](#)) and increased Atlantic MOC deduced from decreasing Mg/Ca-based bottom water temperatures at Site 609 and rising benthic $\delta^{18}\text{O}$ values at sites 609 and 610, suggesting a rise in deepwater density ([Bartoli et al., 2005](#), and references cited therein). Small letters and numbers (101–M1) label marine isotope stages. Age scale tuned to [Tiedemann et al. \(1994\)](#), coming close to LR04 ([Lisiecki and Raymo, 2005](#)). Site locations are shown in [Figure 8](#). Modified from [Sarnthein M, Bartoli G, Prange M, et al. \(2009\)](#) Mid-Pliocene shifts in ocean overturning circulation and the onset of Quaternary-style climates. *Climate of the Past* 5: 269–283.

Each warming of the North Atlantic Current matched an increase in SSS by 0.2–0.4 psu, as inferred from paired planktic $\delta^{18}\text{O}$ and Mg/Ca-based SST values ([Bartoli et al., 2005](#)). Accordingly, each closing of CAS, and in particular, the most effective closure at ~3.0–2.85 Ma, resulted in a significantly intensified convection of deepwater in the Nordic and Labrador Seas. The strengthened AMOC is reflected by a series of proxies, a coeval long-term temperature drop (–1.5 and –2 °C), and a slight rise in salinity and density (increased benthic $\delta^{18}\text{O}$ values) of deepwaters. In particular, it is suggested by a distinct rise in deepwater ventilation, which is documented by a 0.3‰ increase in the short-lasting interglacial maxima in epibenthic $\delta^{13}\text{C}$ at Sites 609 and 610 at 2400–3880-m water depth, that is, the core depth of North Atlantic Deep Water (NADW; [Bartoli et al., 2005](#); [Figure 11](#)). The rise in NADW ventilation at 3.0–2.85 Ma followed an earlier, similar rise interpreted for 4.7–4.0 Ma ([Steph et al., 2010](#)). From 2.8 to 2.73 Ma, however, the interglacial $\delta^{13}\text{C}$ values dropped back to the level prior to 3.0 Ma.

In summary, the outlined changes in North Atlantic circulation and hydrography clearly parallel the twofold exponential rise in IRD at Site 907 ([Figure 9](#)), which forms the onset of persisting high IRD contents and hence marks best the onset of

major NHG and Quaternary-style climates, in particular at 2.93–2.82 Ma (MIS G10). The parallel change of these records strongly suggests a close causal linkage between the buildup of a continental ice sheet on Greenland and an enhanced MOC.

Finally, the latter change was tied to a coeval significant reduction in atmospheric CO_2 , as displayed in two, independently generated CO_2 records ([Seki et al., 2010](#); [Figure 4](#)). Although still coarse, their ~100-ka resolution is sufficient to trace some distinct ups and downs in $p\text{CO}_2$ from 3.15 to 2.75 Ma, which indeed have directly paralleled the renewed opening and closing of the CAS and the coeval drops and rises in IRD deposition, SST, NADW temperatures, density, and ventilation and, thus, in MOC intensity. Since Atlantic deepwater formation provides a major ocean sink for atmospheric CO_2 ([Takahashi et al., 2009](#)), the close linkage between the final openings and closings of the CAS, acting as trigger, and the differential exchange between the oceanic and atmospheric CO_2 inventories becomes obvious. This link may finally have promoted an accelerated buildup of the ice sheet on Greenland, a conceptual model that needs further quantification by physically based modeling and awaits a further extension of the high-resolution CO_2 records over the interval 3.1–4.5 Ma. Once established, the continent-wide ice sheet developed a

long-lived climatic memory effect capable of bridging any further short-lasting orbital-scale reopenings of the CAS and changes in MOC intensity subsequent to MIS G10 (Figures 9 and 11).

Haug et al. (1999) and Sigman et al. (2004) outlined a further factor possibly highly important for the late Pliocene reduction in atmospheric CO₂, the two-step onset of massive surface stratification in the subpolar North Pacific, induced by the two-step incursion of low-saline surface waters at 2.85 and 2.73 Ma. This feature may have impeded the upwelling and outgassing of CO₂-enriched intermediate and deepwaters and persisted over most of the Quaternary Period. However, the initial trigger, that is, the origin of a major increase in the freshwater discharge to this sea region, still awaits explanation, a question that is discussed below.

Sediment Records of Strengthened Poleward Freshwater Transport

Model simulations (Klocker et al., 2005; Lunt et al., 2007) suggest that the closure of Panamayan seaways led to a minor increase in precipitation all over the Northern Hemisphere, in particular, over East Greenland, the northeastern North Atlantic, western and central Europe, and the Kuroshio Current in the northwest Pacific (Figure 6(b)). Today, the EGC and its low SSS of 32–33 psu and temperatures near 0 °C provide a summary record of the outflow of low-saline surface waters from the Arctic Ocean into the northwestern North Atlantic. Here, IODP Site 1307 occupies an ideal position off the southern tip of Greenland to monitor past changes in the EGC and its source region, the Arctic Ocean, including the postulated events of a late Pliocene freshening and cooling to test the ‘Panama Hypothesis.’

The Panamayan seaways closed for a first time around 3.2–3.16 Ma, although only for a short time, and reopened later. The sediments at Site 1307 indeed show a clear response to this closure, recording a significant and long-term change in the composition of EGC waters between 3.22 and 3.02 Ma (Figure 12; details of age control in Sarnthein et al., 2009). During that time, both SST maxima and minima decreased by 6 °C, synchronous with a major SSS drop by ~2 psu, as deduced from a 1‰ decrease in δ¹⁸O of subsurface waters (assuming that the depth habitat of the planktonic foraminifer *Neoglobobulimina atlantica* was equal to that of the modern species *N. pachyderma* s). The immense cooling, freshening, and density drop of the EGC greatly exceed the increase in poleward atmospheric moisture transport predicted by models (Klocker et al., 2005; Lunt et al., 2007). The freshening of the EGC led necessarily to an enhanced stratification of the northwesternmost North Atlantic and more instability of AMOC. Thus, it introduced a new ‘Quaternary’-style regime of the EGC at S. 1307, documented until 2.7 Ma (Figure 12). Note that the EGC freshening clearly preceded the main decrease in atmospheric CO₂ at 2.85–2.73 Ma by at least 0.3 My. Rather, it paralleled a major but short-lasting CO₂ rise, possibly the result of a transient reduction in AMOC intensity.

The pronounced freshening of the EGC at 3.22–3.02 Ma provides an immediate record of Arctic Sea-surface conditions and, thus, evidence for a coeval substantial expansion of perennial Arctic sea-ice and enhanced albedo, conditions that were crucial for promoting the onset of NHG as result of the

full closure of CAS. The EGC-based evidence concurs with that of an Arctic diatom record (Herman, 1974, cited in Stein, 2008) but is opposed to some low-resolution Arctic sediment records (e.g., Frank et al., 2008) that suggest a much earlier date for the onset of perennial Arctic sea-ice.

Linkages to Changes in North Pacific Oceanography

Obviously, the late Pliocene rise in poleward freshwater transport has by far overcompensated the coeval effect of the enhanced poleward heat transport. Until recent times, the question as to where the freshwater flow of the EGC actually originated from remained unsolved. The outlined modest rise in atmospheric poleward moisture transport, which GCM models suggest for vast parts of the Northern Hemisphere after a closure of the CAS (Figure 6(b)), indeed strengthened the runoff into the Arctic Sea. However, this increase appears by far too modest to explain the great freshening of the EGC and the distinct climatic changes that induced the onset of a major glacification of Greenland (Lunt et al., 2007, 2008).

Thus, we need to search for a more substantial forcing mechanism to explain the striking SST/SSS drop from MIS KM3 to G21. Recent model studies (Motoi et al. 2005; Steph et al., 2006; Prange in Sarnthein et al., 2009) suggest that the late Pliocene evolution of the hydrology in the subarctic North Pacific as result of the final closure of the Panamayan seaways should be inspected more closely. Both Prange’s simulation with the NCAR Community Climate System Model (CCSM2/T31x3a) and that of Motoi et al. agree that the termination of the Panamayan throughflow (originally possibly amounting to 12 Sv; Steph et al., 2006) implied a major buildup of a freshwater cap in the northern North Pacific. Here, it resulted in an enhanced stratification (Sigman et al., 2004) and a rise of the steric height by 15 cm, whereas the steric height of the North Atlantic dropped by 20 cm.

In total, these changes implied a doubling of the low-saline Arctic throughflow (~1 Sv today) from the North Pacific through the Bering Strait up to the EGC, then strongly intensified, when assuming a Holocene sea-level stand (Figure 13). During the late Pliocene interglacials (prior to MIS G10), it was possibly up to 35 m higher than that today, and during glacials, it was hardly more than 20 m lower (Dowsett et al., 2010).

The sediments of ODP Site 882 provided some first records that indeed reflect parts of the model-predicted late Pliocene freshening and increased stratification of surface waters in the subarctic Northwest Pacific. Two paired planktic δ¹⁸O records show an overall 1‰ decrease for surface and subsurface waters from 2.85 to 2.75 Ma (Maslin et al., 1996). This shift hardly reflects a 4° warming but rather a 2-psu freshening from MIS G12 to G7, an interval well known for a general cooling of the Northern Hemisphere (Figure 1) and coeval with the great drop in atmospheric CO₂ (Figure 4). Swann (2010) established a first record of SSS at the same Site 882 on the basis of δ¹⁸O of diatoms. It shows a distinct SSS drop by 3 psu a little later, at 2.75–2.7 Ma, but does not extend back beyond 2.85 Ma. Further, indirect evidence for a freshening and significantly enhanced stratification, which implies a suppressed North Pacific outgassing of CO₂, came from the pronounced two-step decrease in diatom productivity at ~3.0 and 2.72 Ma (Haug et al., 1999). However, no paleosalinity record was generated yet for the interval 3.2–3.0 Ma, which was crucial

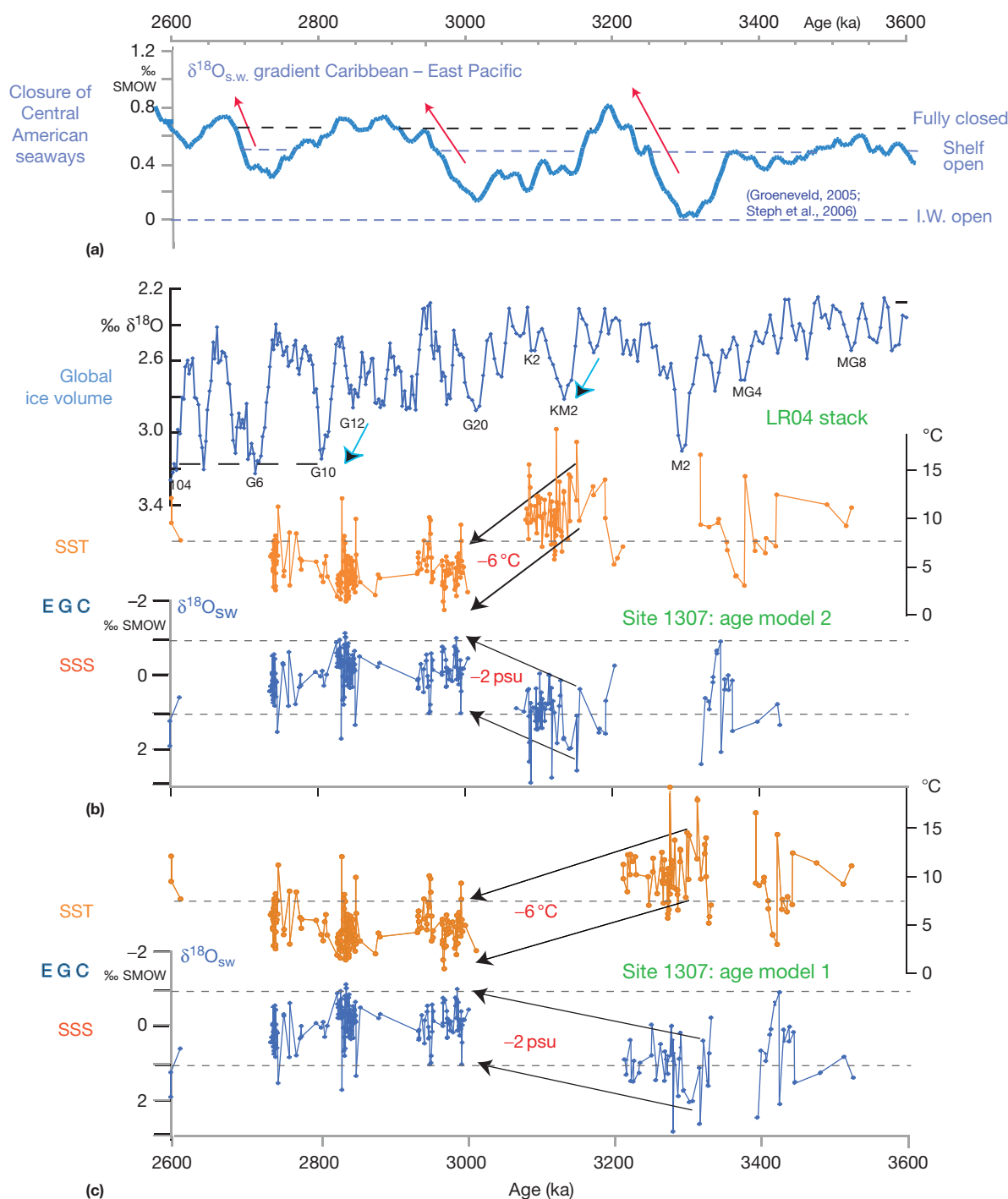


Figure 12 Closure of CAS ($\Delta\delta^{18}\text{O}_{\text{sw}}$ record inverted from Figure 7) (a) and stacked benthic $\delta^{18}\text{O}$ record LR04 of global ice volume (Lisiecki and Raymo, 2005; small letters and numbers label marine isotope stages 104 – MG8) (b) versus major subsea-surface temperature (SST) and salinity (SSS) drops of the EGC at Site 1307, 3.5–2.7 Ma, displayed for age models 2 and 1 (c) (Sarnthein et al., 2009). Sub-SST estimates are based on the Mg/Ca ratio of planktic subsurface species *Neogloboquadrina atlantica*. Two-psu shift in SSS is derived from a 1‰ shift in $\delta^{18}\text{O}_{\text{sea water}}$ ($\delta^{18}\text{O}_{\text{sw}}$; i.e., the $\delta^{18}\text{O}_{\text{carbonate}}$ record of *N. atlantica* after correction for SST), using a conversion factor of 1:2 as for modern EGC water. Modified from Sarnthein M, Bartoli G, Prange M, et al. (2009) Mid-Pliocene shifts in ocean overturning circulation and the onset of Quaternary-style climates. *Climate of the Past* 5: 269–283.

for testing the actual origin of the great freshening of the EGC at that time and thus remains a prime target for future studies.

Once low temperatures and low salinity of the EGC had been established, this current led to a robust thermal isolation of East Greenland from the counteracting poleward heat

transport further east through the North Atlantic and Norwegian currents. Over the last 3 My, the EGC formed an effective barrier to promote the growth of continental ice. Low-resolution ocean models may have problems in resolving this feature. Nevertheless, the simulations of Prange with the NCAR model indeed

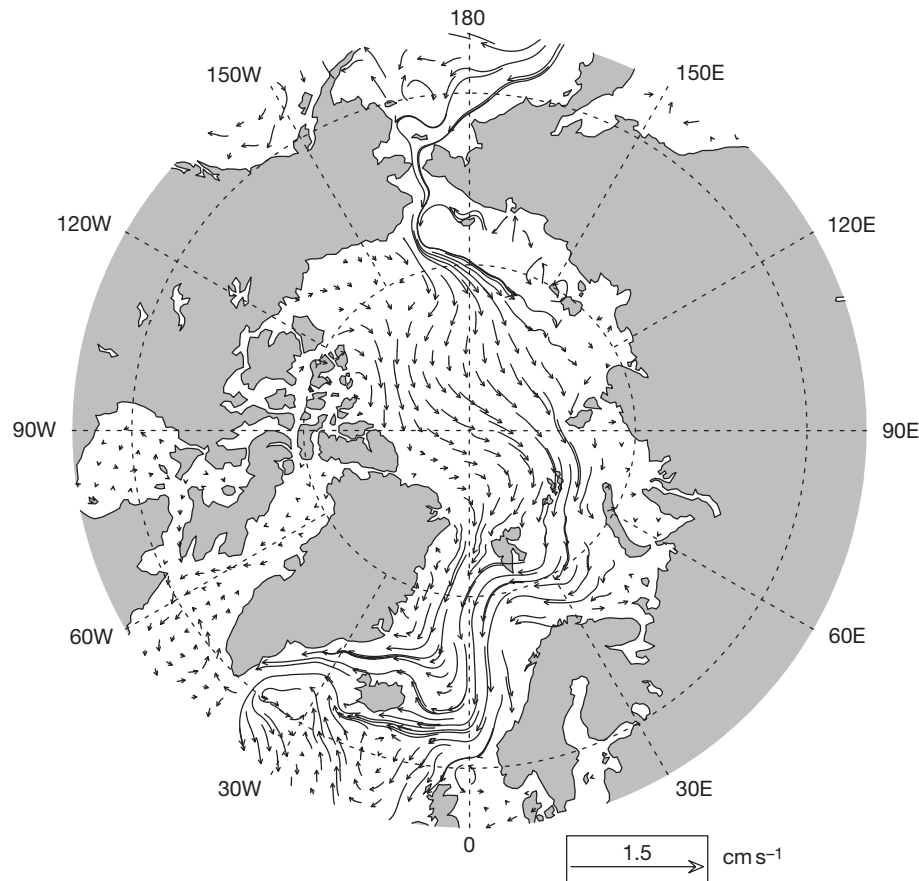


Figure 13 Doubled flow rate of low-saline surface waters through the Bering and Fram straits ('Arctic throughflow') as a result of the difference between two equilibrium model runs, one with closed and the other with open CAS (assumed sill depth of 800 m and aperture of ~200 km); both models are forced by modern boundary conditions (NCAR Community Climate System Model 2, version T31x3a; M. Prange in [Sarnthein et al., 2009](#)). Modified from Sarnthein M, Bartoli G, Prange M, et al. (2009) Mid-Pliocene shifts in ocean overturning circulation and the onset of Quaternary-style climates. *Climate of the Past* 5: 269–283.

show that an enhanced EGC led to a year-round regional cooling over Central and South Greenland by up to 2 °C and strengthened annual snowfall in South Greenland, while almost the entire Northern Hemisphere became warmer in response to the CAS closure.

The Greenland continental ice sheet, once established, strengthened the Greenland High and, in particular, northerly winds along its eastern margin (cf. [Toniazzi et al., 2004](#)), which in turn further strengthened the EGC flow. This positive feedback, a kind of long-term memory effect, may have been crucial in keeping the low-salinity and cold regime of the EGC going over interglacial stages K1, G21, and G19, when the CAS was shortly reopened until the renewed and final closure after 2.95 Ma ([Figure 12](#)). From this time on, pronounced northerly winds along the East Greenland coast were strong enough to drive a vigorous EGC that pushes the polar front across Site 1307, similar to today.

In summary, it appears that the actual ties between the full closure of CAS and the onset of a Quaternary-style NHG were finally established by two independent forcing mechanisms linked to the great increase in the subarctic North Pacific freshwater cap: (1) It probably promoted the reduction in atmospheric CO₂, in addition to the enhanced CO₂

drawdown by the increase in AMOC. (2) This cap, in turn, significantly strengthened the low-saline Bering Strait and Arctic throughflow, previously overseen ([Molnar, 2008](#)). This flow was far more effective for the poleward advection of moisture than the atmospheric moisture transport and clearly overcompensated the effects of enhanced poleward heat transport in the North Atlantic, also resulting from the final closing of CAS. Once a major Greenland ice sheet was established near 3.18–3.12 Ma and, in particular, after 2.9 Ma, it together with the outlined secondary feedback mechanisms formed an ongoing nucleus for Quaternary-style climates and glaciations that persisted in the Northern Hemisphere over the last 2.8–3.0 My.

Conclusions

The transition from late Neogene to Early Quaternary environments at ~3.2–2.7 Ma was linked to the onset of major NHG, forming the most recent major event in the Earth's climate history. Among various hypotheses and models that tend to explain the complex forcing processes, the gradual constriction of the ITF may have helped precondition the climate system, in particular, at 3.2–3.0 Ma. The stepwise, in part shortly

reversible, final closure of the Central America seaways (CAS) at sill depths of less than 250–500 m forms the basis of the most important ‘Panama Hypothesis.’ This closure is now constrained to 3.2–2.7 Ma on the basis of the growing gradient in SSS between the southwest Caribbean and eastern equatorial Pacific. Each closing matched times of an increased discharge of IRD to the Greenland Sea. Models and North Atlantic sediment records likewise suggest that the closure induced a significant rise in poleward oceanic heat transport, in AMOC and drawdown of atmospheric CO₂ and, less significant, in the poleward flux of atmospheric moisture, a process counteracting the poleward heat flux with regard to NHG. The pronounced two-step intensification of the freshwater cap in the subarctic North Pacific formed the perhaps crucial twofold link between the closure of the CAS and NHG. (1) It lowered the atmospheric CO₂ level by impeding the outgassing of CO₂ from upwelled North Pacific deepwaters. (2) It led to a doubling of the low-saline Arctic throughflow that promoted the spread of perennial Arctic sea-ice and albedo and helped stabilize the cold EGC and Polar Front in the Nordic Seas after 3.1 Ma and, in particular, after 2.85 Ma. This current, in turn, formed a long-term climatic feedback mechanism and memory that fostered the growth of continental ice in Greenland and helped to bridge short-lasting reopenings of the CAS.

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References

- Bartoli G, Sarnthein M, Weinelt M, Erlenkeuser E, Garbe-Schönberg D, and Lea D (2005) Final closure of Panama and the onset of northern hemisphere glaciation. *Earth and Planetary Science Letters* 237: 33–44.
- Cane MA and Molnar P (2001) Closing of the Indonesian seaway as precursor to east African aridification around 3–4 million years ago. *Nature* 411: 157–162.
- Coates AG, Collins LS, Aubry MP, and Berggren WA (2004) The geology of the Darien, Panama, and the later Miocene-Pliocene collision of the Panama arc with northwestern South America. *Geological Society Bulletin* 116: 1327–1344.
- De Conto RM, Pollard D, Wilson PA, Pálke H, Lear CH, and Pagani M (2008) Thresholds for Cenozoic bipolar glaciation. *Nature* 455: 652–656.
- Dowsett HJ, Robinson MM, Haywood AM, et al. (2010) The PRISM3D paleoenvironmental reconstruction. *Stratigraphy* 7(2–3): 123–139.
- Driscoll NW and Haug GH (1998) A short circuit in thermohaline circulation: A cause for Northern Hemisphere glaciation? *Science* 282: 436–438.
- Eldrett J, Harding IC, Wilson PA, Butler E, and Roberts AP (2007) Continental ice in Greenland during the Eocene and Oligocene. *Nature* 446: 176–179.
- Frank M, Backman J, Jacobsson M, et al. (2008) Beryllium isotopes in Central Arctic sediments over the past 12.3 million years: Stratigraphic and paleoclimatic implications. *Paleoceanography* 23: PA1S02.
- Gibbard P, Head MJ, and Walker M (2009) The International Subcommission on Quaternary Stratigraphy: Formal ratification of the Quaternary System/Period and the Pleistocene Series/Epoch with a base at 2.588 Ma. *Journal of Quaternary Science* 25: 96–102.
- Groeneveld J (2005) *Effect of the Pliocene Closure of the Panamanian Gateway on Caribbean and East Pacific Sea Surface Temperatures and Salinities by Applying Combined Mg/Ca and $\delta^{18}O$ Measurements (5.6–2.2 Ma)*. PhD Thesis, University of Kiel.
- Groeneveld J, Steph S, Tiedemann R, Garbe-Schönberg D, Nürnberg D, and Sturm A (2006) Pliocene mixed layer oceanography for Site 1241, using combined Mg/Ca and $\delta^{18}O$ analyses of *Globigerinoides sacculifer*. *Proceedings of the Ocean Drilling Program, Scientific Results*, 202, 27 pp. (Available at http://www.odp.tamu.edu/publications/202_SR/209/209.htm).
- Groeneveld J, Nürnberg D, Tiedemann R, et al. (2008) Foraminiferal Mg/Ca increase in the Caribbean during the Pliocene: Western Atlantic Warm Pool formation, salinity influence, or diagenetic overprint? *Geochemistry, Geophysics, Geosystems* 9(1): Q01P23.
- Haug GH, Sigman DM, Tiedemann R, Pedersen TF, and Sarnthein M (1999) Onset of permanent stratification in the subarctic Pacific Ocean. *Nature* 401: 779–782.
- Haug GH, Tiedemann R, Zahn R, and Ravelo AC (2001) Role of Panama uplift on oceanic freshwater balance. *Geology* 29: 447–450.
- Haywood AM (2009) Pliocene climates. In: Gornitz V (ed.) *Encyclopedia of Paleoclimatology and Ancient Environments*. Earth Sciences Series Dordrecht: Springer.
- Herman Y (ed.) (1974) *Marine Geology and Oceanography of the Arctic Seas*. New York: Springer-Verlag.
- Huybers P and Molnar P (2007) Tropical cooling and the onset of North American glaciation. *Climate of the Past* 3: 549–557.
- Jansen E, Fronval T, Ranck F, and Channell JET (2000) Pliocene-Pleistocene ice-rafting history and cyclicity in the Nordic Seas during the last 3.5 Myr. *Paleoceanography* 15: 709–721.
- Kameo K and Sato T (2000) Biogeography of Neogene calcareous nannofossils in the Caribbean and eastern equatorial Pacific – Floral response to the emergence of the Isthmus of Panama. *Marine Micropaleontology* 39: 201–218.
- Karas C, Nürnberg D, Gupta AK, Tiedemann R, Mohan K, and Bickert T (2009) Mid-Pliocene climate change amplified by a switch in Indonesian subsurface throughflow. *Nature Geoscience* 2: 434–438.
- Kawamura K, Parrenin F, Lisiecki L, et al. (2007) Northern Hemisphere forcing of climate cycles in Antarctica over the past 360,000 years. *Nature* 448: 912–917.
- Keigwin LD (1982) Isotopic paleoceanography of the Caribbean and East Pacific: Role of Panama uplift in Late Neogene time. *Science* 217: 350–353.
- Khelifi N, Sarnthein M, and Naafs BDA (2012) Technical note: Late Pliocene age control and composite depth at ODP Site 982, revisited. *Climate of the Past* 8: 79–87.
- Klocker A, Prange M, and Schulz M (2005) Testing the influence of the Central American Seaway on orbitally forced Northern Hemisphere glaciation. *Geophysical Research Letters* 32: L03703.
- Laskar J, Robutel P, Joutel F, Gastineau M, Correia AC, and Levrard B (2004) A long-term numerical solution for the insolation quantities of the Earth. *Astronomy and Astrophysics Manuscript No. La2004*, 1–26.
- Lawrence KT, Liu Z, and Herbert TD (2006) Evolution of the Eastern Tropical Pacific Through Plio-Pleistocene Glaciation. *Science* 312: 79–83.
- Lawrence K, Herbert T, Brown CM, Raymo ME, and Haywood AM (2009) High-amplitude variations in North Atlantic sea surface temperature during the Early Pliocene warm period. *Paleoceanography* 24: PA2218.
- Lisiecki LE and Raymo ME (2005) A Pliocene-Pleistocene stack of 57 globally distributed benthic $\delta^{18}O$ records. *Paleoceanography* 20: PA1003.
- Lunt DJ, Foster GL, Haywood AM, and Stone EJ (2008) Late Pliocene glaciation controlled by a decline in atmospheric CO₂ levels. *Nature* 454: 1102–1105.
- Lunt DJ, Valdes PJ, Haywood AM, and Rutt IC (2007) Closure of the Panama Seaway during the Pliocene: Implications for climate and Northern Hemisphere glaciation. *Climate Dynamics* 30: 1–18.
- Maier-Reimer E, Mikolajewicz E, and Crowley T (1990) Ocean general circulation model sensitivity experiment with an open Central American Isthmus. *Paleoceanography* 5: 349–366.
- Maslin MA, Haug G, Sarnthein M, and Tiedemann R (1996) The progressive intensification of Northern Hemisphere glaciation as seen from the northern Pacific. In: *International Journal of Earth Science – GR* 85: 452–465.
- Meschede M and Barckhausen U (2001) The relationship of the Cocos and Carnegie ridges – age constraints from palinspastic reconstructions. *International Journal of Earth Science* 90: 386–392.
- Molnar P (2008) Closing of the Central American Seaway and the Ice Age. *Paleoceanography* 23: PA2201.
- Motoi T, Chan W-L, Minobe S, and Sumata H (2005) North Pacific halocline and cold climate induced by Panamian Gateway closure in a coupled ocean–atmosphere GCM. *Geophysical Research Letters* 32: L10618.
- Murdock TQ, Weaver AJ, and Fanning AF (1997) Paleoclimatic response of the closing of the Isthmus of Panama in a coupled ocean–atmosphere model. *Geophysical Research Letters* 24: 253–256.
- Naafs D, Stein R, Hefter J, Khelifi N, De Schepper S, and Haug GH (2010) Late Pliocene changes in the North Atlantic Current. *Earth and Planetary Science Letters* 298: 434–442.
- Pratt B, Weisert H, Zalasiewicz J, and Petrizo MR (2009) On the rank, extent, definition, and/or redefinition of the Quaternary, Pleistocene, Neogene, and Pliocene. In: *International Subcommission on Stratigraphic Classification ISSC Newsletter* 15, 19.

- Sarnthein M, Bartoli G, Prange M, et al. (2009) Mid-Pliocene shifts in ocean overturning circulation and the onset of Quaternary-style climates. *Climate of the Past* 5(2): 269–283.
- Schneider B and Schmittner A (2006) Simulating the impact of the Panamanian closure on ocean circulation, marine productivity and nutrient cycling. *Earth and Planetary Science Letters* 246: 367–380.
- Seki O, Foster GL, Schmidt DN, Mackensen A, Kawamura K, and Pancost RD (2010) Alkenone and boron-based Pliocene pCO_2 records. *Earth and Planetary Science Letters* 292: 201–211.
- Shackleton NJ, Backman J, Zimmerman H, et al. (1984) Oxygen isotope calibration of the onset of ice-rafting and history of glaciation in the North Atlantic region. *Nature* 307: 620–623.
- Sigman DM, Jaccard SL, and Haug GH (2004) Cold climates and polar ocean stratification. *Nature* 428: 59–63.
- Stein R (2008) Arctic Ocean sediments – Processes, proxies, and paleoenvironment. *Developments in Marine Geology* 2: 1–592.
- Steph S, Tiedemann R, Prange M, et al. (2006) Changes in Caribbean surface hydrography during the Pliocene shoaling of the Central American Seaway. *Paleoceanography* 21: PA4221.
- Steph S, Tiedemann R, Prange M, et al. (2010) Early Pliocene increase in thermohaline overturning: A precondition for the development of the modern equatorial Pacific cold tongue. *Paleoceanography* 25: PA2202.
- Swann GEA (2010) Salinity changes in the North West Pacific Ocean during the late Pliocene/early Quaternary from 2.73 Ma to 2.52 Ma. *Earth and Planetary Science Letters* 297: 332–338.
- Takahashi T, Sutherland SC, Wanninkhof R, et al. (2009) Climatological mean and decadal change in surface ocean pCO_2 and net sea-air CO_2 flux over the global oceans. *Deep-Sea Research II* 56: 554–577.
- Tiedemann R, Sarnthein M, and Shackleton NJ (1994) Astronomic timescale for the Pliocene Atlantic $\delta^{18}O$ and dust flux records of ODP Site 659. *Paleoceanography* 9: 619–638.
- Toniazzo T, Gregory JM, and Huybrechts P (2004) Climatic impact of a Greenland deglaciation and its possible irreversibility. *Journal of Climate* 17: 21–33.
- Trauth MH, Maslin MA, Deino A, and Strecker MR (2005) Late Cenozoic moisture history of East Africa. *Science* 309: 2052–2053.
- Weyl PK (1968) The role of the oceans in climate change: A theory of the ice ages. *Meteorological Monographs* 8: 37–62.
- Willenbring JK and von Blanckenburg F (2010) Long-term stability of global erosion rates and weathering during late-Cenozoic cooling. *Nature* 465: 211–214.
- Zachos J, Dickens GR, and Zeebe RE (2001) An early Cenozoic perspective on greenhouse warming and carbon-cycle dynamics. *Nature* 411: 279–283.

Early Quaternary (Pleistocene) and Precursors

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Since the time when [Penck and Brückner \(1901/1909\)](#) published their classic account of the Alpine Ice Ages, their division of the Pleistocene into four cold stages with intervening warm stages became a model that has been applied globally for numerous local stratigraphies. Today, it is considered that there have been, for example, some 30–40 glaciations of Scandinavia, but only a few were large enough to reach beyond the Baltic Sea, and most of these large ones occurred during the last ca. 1 Ma.

The onset of the recent Ice Age dates back much further than the Quaternary. An overall, long-term trend of cooling global climates occurred throughout the Tertiary. Although there were local precursors, significant glaciation began in eastern Antarctica in the late Oligocene (ca. 34 Ma). It was followed by glaciation in mountain areas through the Miocene, for example, in Alaska and Iceland. In most cases, the traces of early glaciations are restricted to individual sites. Knowledge of the extent of the glaciers or ice sheets is rather limited, although there are a few exceptions. They include the Great Patagonian Glaciation in southern South America and the early glaciations in the Nebraska–Iowa area of the United States, in Alaska, and in northwestern Canada, where the ancient ice sheets reached beyond the limits of all later glaciations.

Antarctica

The nature and timing of the initial glaciation of Antarctica is not well known, and the evolution and nature of the West Antarctic Ice Sheet (WAIS) remains controversial ([Wilson, 1995](#)). The Antarctic Ice Sheet has existed for about 35–40 My, since the mid-Tertiary, and it is widely believed that in East Antarctica, it reached continental proportions by the latest Eocene–early Oligocene ([Ingólfsson, 2004](#)). Existing reconstructions of pre-Quaternary Antarctic ice volumes and ice extent rely heavily on the interpretation of Ocean Drilling Program (ODP) data from offshore cores, such as oxygen isotope ($\delta^{18}\text{O}$) concentrations, ice-rafted detritus (IRD) concentrations, sediment type and clay mineralogy, microfossil assemblages, and occurrences of hiatuses. The available data indicate that the Antarctic glacial system has evolved through time: during early periods (Oligocene–Miocene), it was largely temperate in the coastal areas. A transition to cooler conditions followed in Pliocene, and during much of the Quaternary, it has been a stable, cold system.

South America

Neogene glaciation is known from the Piedmont areas of Argentina and Chile, where, like Antarctica, substantial ice

caps were established before 14 Ma. The earliest references concerning the Patagonian Glaciation had already been made by [Darwin \(1842\)](#), who identified erratic boulders in the Magellan Straits region. Widespread lowland glaciation began between 2.05 and 1.86 Ma, followed by the so-called ‘Initioglacial’ or ‘Great Patagonian Glaciation’ that took place 1.15–1.00 Ma. As [Coronato et al. \(2004\)](#) point out, it is considered that valley and piedmont glaciers flowed from the mountain ice sheet at the Andean summit several hundred kilometers eastward, in the Extra-Andean plains domain. Several times during the early Pleistocene, they probably reached the present Atlantic submarine platforms. The ice sheets reached a size comparable to the Late Weichselian Scandinavian Ice Sheet.

The recognition of the maximum extent of the different glacial advances is still largely based on the outstanding work of the Swedish geologist [Caldenius, published in 1932](#), after 3 years of intensive field work in Patagonia, covering an area of roughly one million square kilometers. Later authors have provided new evidence in support of Caldenius’ basic model, modifying his original schemes. Although the currently identified boundaries of the different glacial advances are highly coincident with those mapped by Caldenius, the total number of glaciations and their chronological correlation have changed. Morphologically, six or seven terminal ice positions of different absolute or relative ages can be distinguished.

The early glaciation of South America was not limited to Patagonia, although it is best documented there. But some glacial deposits, for example, in the Bolivian Andes, have been dated back to 3.27 My, and in Columbia, 2.5-My-old glacial meltwater deposits have been reported from the Bogotá basin.

Australasia

New Zealand’s earliest glaciation – the Ross Glaciation – is recognized from till and glacial lake beds interbedded with folded late Pliocene gravels in north Westland ([Gage, 1945](#)). The ice limits are unknown. The age of this glaciation is thought to be 2.4–2.5 Ma. The Porika Glaciation is probably younger, according to glacial deposits that rest on basement rocks and Pliocene gravel. Following these glaciations, a gap of at least a million years in the glacial history probably resulted from the dominance of uplift and erosion, with the present drainage pattern becoming progressively established. In New Zealand, glaciation – and especially evidence of early glaciation – is largely restricted to the South Island.

Traces of glaciation in Tasmania are limited to the last million years. The oldest event, the Linda Glaciation, was first identified in the Linda valley. Unfortunately, age control is poor. The early glacial deposits contain strongly weathered clasts, and paleomagnetic investigations of related meltwater deposits suggest that they are older than the Brunhes/Matuyama magnetic reversal boundary.

Africa

Even in Africa, with its traces of Quaternary glaciation being limited to very few high mountains, there is evidence of early Pleistocene glaciation. Traces of two old glaciations have been found on Mount Kilimanjaro, the oldest of which has been dated to about 2.0 Ma (Osmaston, 2004).

North America

In North America, the longest sequences are found in Alaska and the adjacent Northwest Territories of Canada, which together with the Rockies preserve evidence of glaciation from the Neogene to the present. In Montana, for instance, three tills have been identified dating back to early Pleistocene Marine Isotope Stages (MIS) 96, 82, and 78 (Fullerton et al., 2004).

During the Matuyama Chron (ca. 2.6 Ma), ice appears to have been largely absent from large areas of the southern prairie provinces in Canada and the adjacent states of Montana and North Dakota, as well as from the Arctic Islands. However, in Late Matuyama time (ca. 1.0 Ma), a modest Keewatin ice center formed, delivering ice as far distant as Banks Island, NWT. In contrast, from the Labrador/Hudson Bay glaciation center, ice spread as far south as Topeka in Kansas during both the Early and Late Matuyama (Roy et al., 2003). Traces of Early Matuyama Chron (ca. 2.0 Ma) glaciation are found in easternmost South Dakota, Minnesota, and Wisconsin. Late Matuyama Chron (ca. 1.0 Ma) glaciation seems to have extended eastward into Illinois, Indiana, Ohio, and Pennsylvania. Ice caps appear for the most part to have been far more extensive during the Brunhes Chron (ca. 0.5–0.6 Ma) than in the Matuyama, and only in the southern Midwestern states did Brunhes-age ice not reach previous limits (Barendregt and Duk-Rodkin, 2004). Consequently, the most investigated sites are restricted to those areas.

Glaciation of Greenland seems to have started some 14 Ma in the mountains, with contiguous ice sheets established from the earliest Pleistocene (ca. 2.6 Ma). The surprisingly short and incomplete glacial record from Greenland reflects the fact that the ice sheet appears to have extended to the shelf break during each of the glacial maxima. The most complete records are therefore probably found on the adjacent continental slope.

Europe

The Scandinavian Ice Sheet

The impressive erosional landforms of Norway, such as fjords and valleys, have been formed during multiple glaciations

(Mangerud et al., 1996). Unfortunately, these features cannot be dated. The oldest dated trace of glaciation is ice-rafted detritus (IRD) dropped from icebergs into the deep sea beyond the coast. IRD pulses are found from 12.6 Ma in cores from the Vøring Plateau outside the Norwegian coast (Fronval and Jansen, 1996), suggesting that the Scandinavian Ice Sheet first reached sea level at that time. A synthesis of observations of IRD in cores from the Nordic seas (Kleiven et al., 2002) indicates that there have been marked expansions of the Greenland Ice Sheet 3.3 Ma and of the Scandinavian Ice Sheet from 2.75 Ma. A series of prograding wedges, named the Naust Formation, is the youngest formation on the Mid-Norwegian continental shelf in the area between 64 and 68° N, that is, north of the North Sea Fan. The age of the base is considered to be ca. 2.8 Ma (Ottesen et al., 2009), that is, close to the recorded major increase in IRD in the Norwegian Sea. The formation is more than a kilometer thick over a large area, and the shelf edge prograded up to 150 km during the accumulation of this formation (Dahlgren et al., 2005; Ottesen et al., 2009; Rise et al., 2005). Scientists agree that the Naust Formation reflects a number of glaciations that covered a large part of mainland Norway; some consider that the entire formation was deposited during ice sheet advances to the paleo-shelf edge, whereas others consider that the older part reflects smaller mountain-centered ice sheets and that the ice margin reached the shelf edge for the first time about 1.1 Ma. The latter interpretation is partly based on the fact that the Fedje Till, the oldest dated till on the Norwegian shelf, is assigned an age of about 1.1 Ma (=MIS 34–36; Sejrup et al., 2005).

Also, in the Barents Sea region, initial glacial growth is reflected in a pronounced increase in the general sedimentation rate along the Svalbard margin from around 2.3 to 2.5 Ma. The volume of thick wedges of glaciogenic sediments along the western continental slope suggests that there has been as much as 1000–1500 m of erosion over the Barents Sea region since the onset of glaciation. Consequently, during the oldest glaciations, the physiography of the Barents Sea region must have been very different from the present and the shelf area was probably above sea level (Elverhøi et al., 1998; Vorren et al., 1998).

Sedimentological data from the ODP site 986 drilled off western Svalbard reveal low IRD values during the interval 2.3–1.6 Ma, suggesting that glaciers in the Barents Sea region did not reach the sea (Butt et al., 2000). However, further extensive glaciation seems to have begun around 1.5–1.3 Ma, and after 1.3 Ma, ice sheets expanded to the shelf break several times. The seismic stratigraphy suggests that the ice sheet in the southwestern Barents Sea expanded to the outer shelf at least eight times (Laberg and Vorren, 1996). The last period of glacial deposition appears to have been dominated by aggradation of glacial sediments on the shelf margin rather than progradation. This may reflect a different style of glaciation as the shelf was glacially eroded and eventually became submerged. Accordingly, Solheim et al. (1998) suggest that the thick eroding ice sheets of the oldest glaciations were later replaced by thinner ice sheets with more rapidly flowing ice streams in the marginal areas.

Deposits of pre-Elsterian glaciations have only been sporadically reported from Denmark. At Harreskov (Jutland), Early middle Pleistocene interglacial deposits occur; they can

possibly be correlated with the Cromerian Complex Stage 'Interglacial II' (?MIS 19/17). Between these Harreskovian interglacial deposits and the underlying beds of Tertiary clay, glaciolacustrine, and glaciofluvial sediments, a clay-rich diamicton, interpreted as till, has been found. The age of this oldest record of glaciation in Denmark is unknown. It might date either from the Menapian Stage (?MIS 34–36) or, more likely, from an older part of the Cromerian Complex or Bavelian Complex stages (?MIS 22/20), when an ice sheet advanced southward from the Scandinavian mountains. There are other sites in Jutland where gravelly strata are found between the Tertiary deposits and the lowermost tills. They consist mainly of quartzite and quartzitic sandstone and also contain ventifacts. These strata might comprise residual till material dating from this pre-Elsterian glaciation (Houmark-Nielsen, 2004).

Further south, early glaciation appears to be represented only by ice-rafted clasts, outside the mountain or high-latitude regions (e.g., in the Netherlands, lowland Germany, northwestern France, and Britain). In the Late Tertiary, the sea withdrew from the Northwest European Basin and fluvial deposition prevailed. The deposits of the resulting 'Baltic River System' can be traced from the island of Sylt well into the Netherlands and beneath the North Sea. They contain lydites from the southern Central German uplands, as well as large blocks of sandstone, which must have been transported by drift ice from the eastern Baltic region. The Hattem Beds of the Menapian (MIS 34/36: ca. 1.2 Ma) are the oldest strata in the Netherlands and North Germany that carry major quantities of Scandinavian erratics from East Fennoscandian and central Swedish (Dalarna) source areas (Zandstra, 1983). They are exposed in the Ytterbeck-Uelsen push moraine in Niedersachsen (Lower Saxony). They are interpreted as evidence of the first major ice advance beyond the limits of the present Baltic basin (Ehlers, 1996). However, in the Netherlands, some Scandinavian rocks are also found in the Harderwijk Formation of Eburonian age, dating back to almost 1.8 Ma, and some consider that they were also transported by glaciers.

Iceland

In Iceland, the oldest tillites in the stratigraphic column indicate that mountain glaciation first occurred there in the Miocene, more than 7–6 Ma. The Icelandic sequence includes as many as 30 glacial horizons since then. There is evidence to suggest a progressive growth of glaciers during late Pliocene, and full-scale glacial–interglacial cyclicity, with a recurring growth of regional ice caps recorded since 2.6 Ma (Geirsdóttir and Eiríksson, 1994). The first full-scale glaciation of Iceland, with an ice sheet covering most of the island, dates back to 2.2–2.1 Ma. Glaciations occurred periodically through the Quaternary Period, but the record is fragmented by repeated destruction of stratigraphical evidence by subsequent glaciations. One of the key sections for the Pliocene–Pleistocene glacial and environmental development on Iceland is the Tjörnes sequence in northern Iceland. There, in a 600-m-thick sequence of interbedded lavas, tills, and marine and terrestrial sediments, at least 14 glaciations have been recorded since 2.5–2.0 Ma. This is probably the best

preserved terrestrial record anywhere for early and middle Pleistocene glaciations in the North Atlantic region (Figures 1 and 2).

Southern Europe

Early glaciation of the Alps is still poorly understood. Part of Penck and Brückner's (1901/1909) original Günz glacial deposits are known to be reverse magnetized. They must date therefore to the Matuyama Chron (i.e., pre-0.78 Ma). Early glaciations have also been reported from the French, Italian,

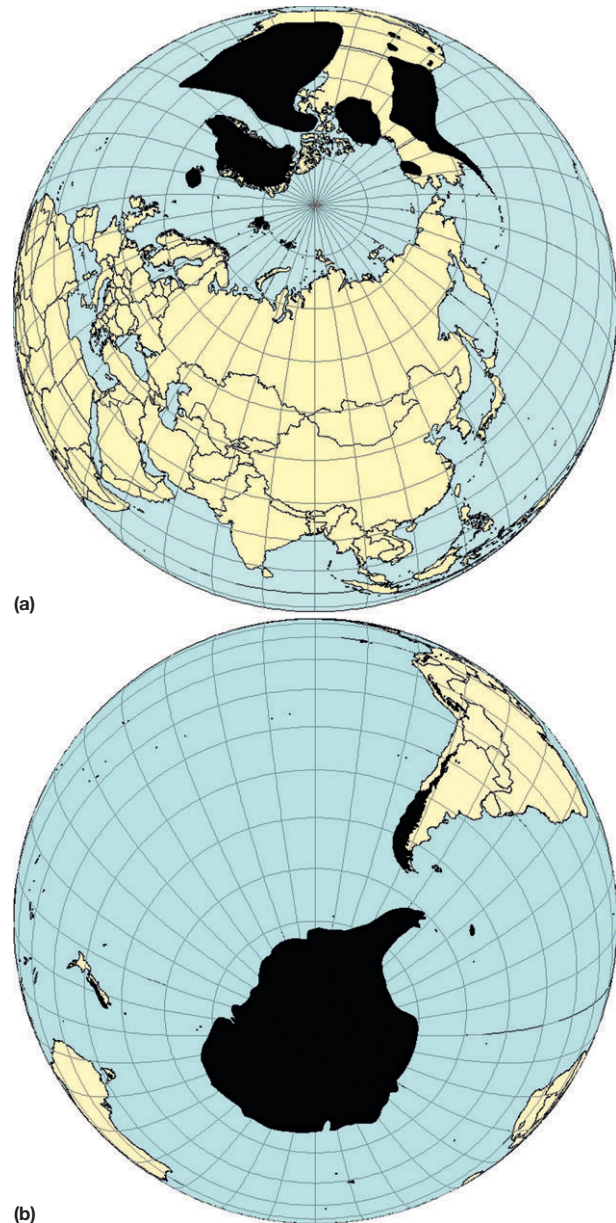


Figure 1 Synthetic maps, showing the extent of glaciations. (a) Late Matuyama Chron, Northern Hemisphere (ca. 1.0 My), (b) Late Matuyama Chron, Southern Hemisphere, (c) Middle Pleistocene glacial maximum, Northern Hemisphere, and (d) Middle Pleistocene glacial maximum, Southern Hemisphere. Maps are based on Ehlers and Gibbard (2004a,b,c).

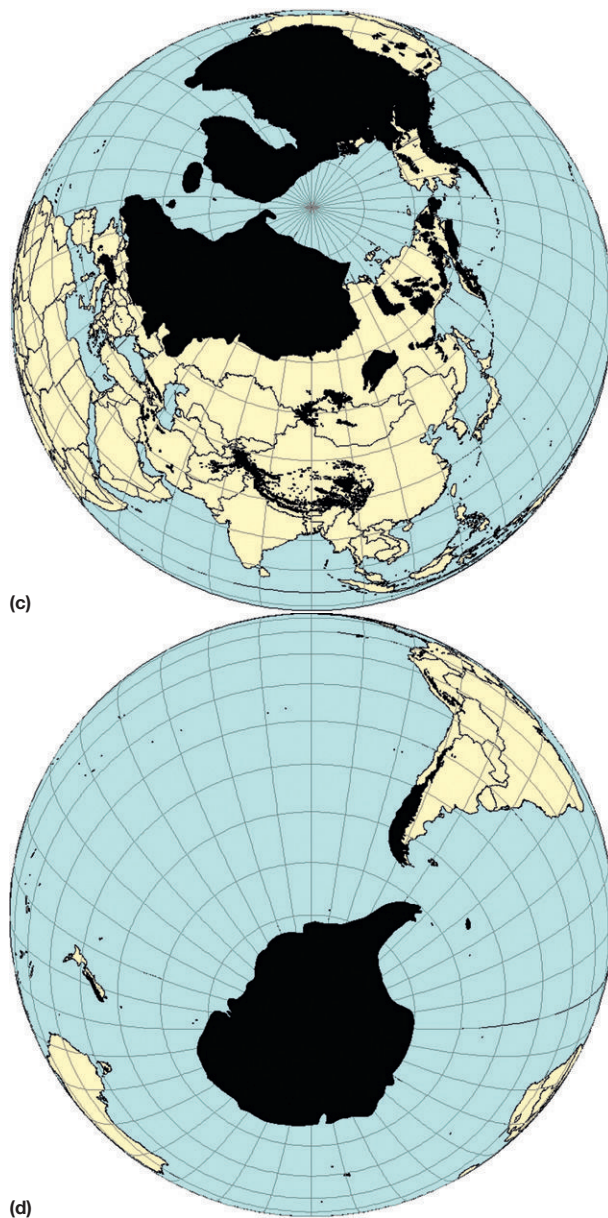


Figure 1 (Continued)

and Swiss parts of the mountains. However, in many cases, as with the northern Alpine Donau and Biber Stage deposits, much of the evidence is based on sediments indirectly of glacial origin, and the number and extent of early Alpine glaciations remain unknown.

See also: **Archaeological Records:** 2.7 Myr–300 000 Years Ago in Asia; Global Expansion 300 000–8000 Years Ago, Africa.

Glaciations: Overview; Transition from Late Neogene to Early Quaternary Environments. **Sea Level Studies:** Eustatic Sea-Level Changes Since the Last Glacial Maximum.

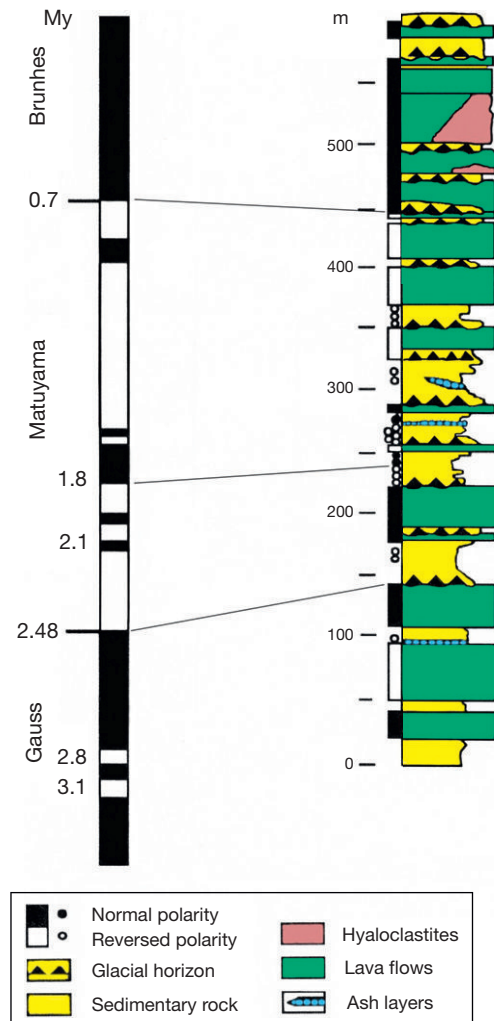


Figure 2 Composite stratigraphy of Tjörnes, Iceland, and correlations to the paleomagnetic timescale. Modified from Geirsdóttir Á and Eiríksson J (1994) Growth of an intermittent ice sheet in Iceland during the Late Pliocene and Early Pleistocene. *Quaternary Research* 42: 115–130.

References

- Barendregt RW and Duk-Rodkin A (2004) Chronology and extent of Late Cenozoic ice sheets in North America: A magnetostratigraphic assessment. In: Ehlers J and Gibbard PL (eds.) *Quaternary Glaciations – Extent and Chronology, Part II: North America. Developments in Quaternary Science*, vol. 2b, pp. 1–7. Amsterdam: Elsevier.
- Butt FA, Elverhøi A, Solheim A, and Forsberg CF (2000) Deciphering Late Cenozoic development of the western Svalbard Margin from ODP Site 986 results. *Marine Geology* 169: 373–390.
- Caldenius C (1932) *Las Glaciaciones Cuaternarias en Patagonia y Tierra del Fuego*, vol. 95(1), p. 148. Buenos Aires: Ministerio de Agricultura de la Nación. Dirección General de Minas y Geología.
- Coronato A, Martínez O, and Rabassa J (2004) Glaciations in Argentine Patagonia, southern South America. In: Ehlers J and Gibbard PL (eds.) *Quaternary Glaciations, Extent and Chronology, Part III: South America, Asia, Africa, Australasia, Antarctica. Developments in Quaternary Science*, vol. 2c, pp. 49–67. Amsterdam: Elsevier.
- Dahlgren KIT, Vorren TO, Stoker MS, Nielsen T, Nygard A, and Sejrup HP (2005) Late Cenozoic prograding wedges on the NW European continental margin: Their formation and relationship to tectonics and climate. *Marine and Petroleum Geology* 22: 1089–1110.

- Darwin Ch (1842) On the distribution of the erratic boulders and on the contemporaneous unstratified deposits of South America. *Transactions of the Geological Society London* 6: 415–431.
- Ehlers J (1996) *Quaternary and Glacial Geology*. Chichester: Wiley.
- Ehlers J and Gibbard PL (eds.) (2004a) *Quaternary Glaciations – Extent and Chronology, Part I: Europe. Developments in Quaternary Science*, vol. 2a. Amsterdam: Elsevier.
- Ehlers J and Gibbard PL (eds.) (2004b) *Quaternary Glaciations – Extent and Chronology, Part II: North America. Developments in Quaternary Science*, vol. 2b. Amsterdam: Elsevier.
- Ehlers J and Gibbard PL (eds.) (2004c) *Quaternary Glaciations, Extent and Chronology, Part III: South America, Asia, Africa, Australasia, Antarctica. Developments in Quaternary Science*, vol. 2c. Amsterdam: Elsevier.
- Elverhøi A, Hook RLB, and Solheim A (1998) Late Cenozoic erosion and sediment yield from the Svalbard-Barents Sea Region: Implications for understanding erosion of Glaciated Basins. *Quaternary Research* 44: 209–241.
- Fronval T and Jansen E (1996) Late Neogene paleoclimates and paleoceanography in the Iceland-Norwegian Sea: Evidence from the Iceland and Vøring plateaus. In: Thiede J, Myhre AM, Firth JV, Johnsen G, and Ruddiman W (eds.) *Proceedings of the Ocean Drilling Program, Scientific Results*, vol. 151, pp. 455–468. College Station, TX: Ocean Drilling Program.
- Fullerton DS, Colton RB, and Bush ChA (2004) Limits of mountain and continental glaciations east of the Continental Divide in northern Montana and north-western North Dakota, U.S.A. In: Ehlers J and Gibbard PL (eds.) *Quaternary Glaciations – Extent and Chronology, Part II: North America, Developments in Quaternary Science*, vol. 2b, pp. 131–150.
- Gage M (1945) The Tertiary and Quaternary geology of Ross, Westland. *Transactions of the Royal Society of New Zealand*, vol. 75, pp. 138–159. Amsterdam: Elsevier.
- Geirsdóttir Á and Eiríksson J (1994) Growth of an intermittent ice sheet in Iceland during the Late Pliocene and Early Pleistocene. *Quaternary Research* 42: 115–130.
- Houmark-Nielsen M (2004) The Pleistocene of Denmark: A review of stratigraphy and glaciation history. In: Ehlers J and Gibbard PL (eds.) *Quaternary Glaciations – Extent and Chronology, Part I: Europe. Developments in Quaternary Science*, vol. 2a, pp. 35–46. Amsterdam: Elsevier.
- Ingólfsson O (2004) Quaternary glacial and climate history of Antarctica. In: Ehlers J and Gibbard PL (eds.) *Quaternary Glaciations, Extent and Chronology, Part III: South America, Asia, Africa, Australasia, Antarctica. Developments in Quaternary Science*, vol. 2c, pp. 1–43. Amsterdam: Elsevier.
- Kleiven H, Jansen E, Fronval T, and Smith TM (2002) Intensification of Northern Hemisphere glaciations in the circum Atlantic region (3.5–2.4 Ma) – Ice-rafted detritus evidence. *Palaeogeography, Palaeoclimatology, Palaeoecology* 184: 213–223.
- Laberg JS and Vorren TO (1996) The Middle and Late Pleistocene evolution of the Bear Island Trough Mouth Fan. *Global and Planetary Change* 12: 309–330.
- Mangerud J, Jansen E, and Landvik J (1996) Late Cenozoic history of the Scandinavian and Barents Sea ice sheets. *Global and Planetary Change* 12: 11–26.
- Osmaston H (2004) Quaternary glaciations in the East African mountains. In: Ehlers J and Gibbard PL (eds.) *Quaternary Glaciations, Extent and Chronology, Part III: South America, Asia, Africa, Australasia, Antarctica. Developments in Quaternary Science*, vol. 2c, pp. 139–150. Amsterdam: Elsevier.
- Ottesen D, Rise L, Andersen ES, Bugge T, and Eidvin T (2009) Geological evolution of the Norwegian continental shelf between 61° N and 68° N during the last 3 million years. *Norwegian Journal of Geology* 89: 251–265.
- Penck A and Brückner E (1901/1909) *Die Alpen im Eiszeitalter*, 3 vols. Leipzig: Tauchnitz.
- Rise L, Ottesen D, Berg K, and Lundin E (2005) Large-scale development of the mid-Norwegian margin during the last 3 million years. *Marine and Petroleum Geology* 22: 33–44.
- Roy M, Clark PU, Barendregt RW, Glasman JR, Enkin RJ, and Baker J (2003) Glacial stratigraphy and paleomagnetism of late Cenozoic deposits of the north-central U.S. *Bulletin of the Geological Association of America* 116: 30–41.
- Sejrup HP, Hjelstuen BO, Dahlgren KIT, et al. (2005) Pleistocene glacial history of the NW European continental margin. *Marine and Petroleum Geology* 22: 1111–1129.
- Solheim A, Faleide JJ, Andersen ES, et al. (1998) Late Cenozoic seismic stratigraphy and glacial geological development of the East Greenland and Svalbard–Barents Sea Continental Margins. *Quaternary Science Reviews* 17: 155–184.
- Vorren TO, Laberg JS, Blaumme F, et al. (1998) The Norwegian-Greenland Sea continental margins: Morphology and late Quaternary sedimentary processes and environment. *Quaternary Science Reviews* 17: 273–302.
- Wilson GS (1995) The Neogene East Antarctic ice sheet: A dynamic or stable feature. *Quaternary Science Reviews* 14: 101–123.
- Zandstra JG (1983) Fine gravel, heavy mineral and grain-size analyses of Pleistocene, mainly glacial deposits in the Netherlands. In: Ehlers J (ed.) *Glacial Deposits in North-West Europe*, pp. 361–377. Balkema: Rotterdam.

Middle Pleistocene in Eurasia

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Northwest Europe

The first outline of the Quaternary stratigraphy was established by the end of the nineteenth century. At that time, the actual length of the Ice Age was still unknown. It was assumed that Northwest Europe, like the Alps, had been affected by three glaciations. The glaciations were named after rivers, in the catchment areas of which their deposits were well exposed. The names Elsterian, Saalian, and Weichselian for the Northwest European glaciations first appeared in 1911 on the 1:25 000 map sheets of the Prussian Geological Survey, and they gained general acceptance in the 1920s.

After it was discovered that the Alps had been glaciated (at least) four times instead of three, there have been many attempts to identify corresponding numbers of glaciations elsewhere. However, no convincing evidence was found in Britain, Ireland, or western central Europe. It was further to the east that traces of earlier glaciations were eventually identified.

Oldest Glaciations of Russia

Originally, it was assumed that the maximum glaciation on the East European Plain was the late middle Pleistocene Dnieper (Saalian) glaciation. Dnieper-age deposits overlie tills of an older glaciation, the Oka ice sheet, correlated to the Elsterian. However, in the 1970s, it was discovered that the till of the southernmost lobe of the middle Pleistocene glaciation in Central Russia, unlike that in the Ukraine, was overlain by Muchkap interglacial sediments bearing the fauna of the Tiraspol (Cromer) Complex and by four to five paleosols of interglacial character (Velichko et al., 2004). Accordingly, this so-called Don glaciation, presumably ca. 0.6-My old (postdating the Brunhes/Matuyama boundary), could no longer be correlated with the Saalian in Western Europe, the Dnieper glaciation in the Ukraine, or with the Samarovo glaciation of Siberia; it most probably relates to Marine Isotope Stage (MIS) 16 (Iosifova et al., 2006; Molodkov and Bolikhovskaya, 2006). The thick tripartite diamict sequence of the Don glaciation consists mostly of locally derived fine-grained materials with rare small clasts of Scandinavian origin. Pebbles of northeastern provenance and corresponding fabrics are more noticeable. This suggests limited influence of the Fennoscandian ice dispersal center and dominant ice flow from the Arctic (Velichko et al., 2011). In Poland, the maximum San 1 glaciation related to MIS 16 and Cromer glacial B, when the Scandinavian ice reached the Carpathians, left a till positioned below the Ferdinandovian interglacial formation (Lindner and Marks, 2008). A possible equivalent event has been suggested in Eastern England (Lee et al., 2004), but this is greatly disputed by local researchers.

Later on, traces of even older glaciations within the Brunhes Chron were identified buried deep within the boreholes of Moscow Region. The lowermost Likovo till contains no Scandinavian erratics. It is separated by the Akulovo interglacial strata, with 11% extinct floristic taxa from the more extensive Setun till containing 40–60% Scandinavian igneous rocks in the clast composition (Velichko et al., 2011). The fabric is mostly northwest–southeast whereas northeast–southwest fabrics dominate the Don till. The Setun and Likovo tills are correlated with the Cromerian Glacial A and MIS 18 (Iosifova et al., 2006). The same stratigraphic position is characteristic of the Nidanian till of Poland (Lindner and Marks, 2008).

Farther eastward, stratigraphic analogs of these ancient tills have not strictly been proven, although the maximum glaciation of the eastern Russian Plain has recently been mapped as Donian (Astakhov, 2004). In Siberia, a probable counterpart of the Don glacial sequence is the Mansi till, known from several deep boreholes of the Ob–Irtysh area (Volkova et al., 2002).

Elsterian Glaciation (Elsterian Stage)

The oldest glacial event that has been proved by the existence of tills throughout the North European glaciated area is the Elsterian glaciation. This glaciation reached as far south as the northern margin of the central German uplands. It was also the first time that ice sheets were extensive enough that Scandinavian and British ice sheets met at the center of the North Sea. As drainage toward the north was blocked, a large ice-dammed lake formed in the North Sea. Eventually, around 455 000 BP, the water breached the Weald–Artois threshold and started to drain southward toward the Gulf of Biscay. The Channel was formed at first only as a major river system, and in the course of the following warm and cold stages it gradually widened to its present size.

The history of the English Channel was difficult to reconstruct. The key was to be found south of the Channel, in the Bay of Biscay. Toucanne et al. (2009a,b) have studied the Deep Sea Core MD01-2448. This 28.8-m long core, drilled at 44°46,790 N, 11°16,470 S in a water depth of 3460 m on top of the Charcot Seamount, was ideally positioned to document the general course of the Pleistocene as well as the sedimentary influx from the Channel. The sedimentary sequence reaches back to MIS 35 – which means that it covers a period of 1.2 Ma. Comparison with the oxygen isotope curve from the LR04 stack (Lisiecki and Raymo, 2005) shows that the sedimentary sequence is complete, without any major gaps. Thus, it has been possible to determine an important correlation point between the terrestrial and the marine Quaternary stratigraphy. The Elsterian Stage is to be correlated with MIS 12.

The Elsterian glaciation was characterized by intensive glaciofluvial erosion. Deep subglacially formed channels of Elsterian age can be traced into the Northern Netherlands, beneath the southern North Sea, and into Eastern England. However, Elsterian till in the Netherlands seems to be restricted to a few thin layers. A major part of the Elsterian tills may have been eroded during subsequent cold stages, although continuous till sheets of this age occur north of the Weser River. In the southern parts of Niedersachsen and in Nordrhein-Westfalen, the Elsterian glaciation can in most cases only be inferred from Scandinavian erratics in contemporaneous river terrace deposits. As the thrust sequences of the end moraines west of the Weser River contain no Elsterian till, it is assumed that the ice advance did not extend beyond that area.

In Sachsen (Saxony) and Thüringen (Thuringia), where the Elsterian sediments were not overridden by the Saalian or any other ice sheet, an accurate reconstruction of the maximum extent of the Elsterian glaciation is possible. Here the southern limit of Nordic erratics, the so-called 'flint line,' is considered to represent the Elsterian ice sheet limit. It is found at an altitude of 300–480 m a.s.l., dipping toward the west. In the Saxonian-Thuringian area, especially in the vast lignite open-cast mines, two Elsterian tills have been found. These ice advances were apparently separated by only a short interstadial warm period, the Miltitz Interval. During this interval, the ice melted at least as far north as Mecklenburg, but no afforestation occurred (Eissmann, 1975).

The Elsterian tills of West Germany normally contain a relatively large proportion of western Scandinavian material. Rhomb porphyries and other indicator clasts from south Norwegian source areas are frequently found, together with flint conglomerate from the bottom of the Skagerrak. However, in no case do Norwegian rocks predominate; at best they make up a small percent of the (crystalline + sedimentary) indicator rock assemblage. Finds of rhomb porphyries related to the Anglian tills of Norfolk suggest a Norwegian glacial influence reaching beyond the North Sea and into Britain. Rhomb porphyries also occur far to the east, in Elsterian tills around Leipzig. This early Norwegian–West Swedish influence was later replaced by a North Swedish–Finnish ice stream. Consequently, in eastern Germany and Poland, the 'Baltic' facies of Elsterian till represents the normal lithofacies. This means that the tills tend to be rich in Paleozoic limestones and poor in flint.

In Denmark, Elsterian tills are known from exposures near Esbjerg, as well as from the Røgle cliff on Fyn. Borehole information suggests that a bipartite Elsterian till exists in Denmark. Again, an earlier ice advance from the north, characterized by rhomb porphyries from the Oslo area, seems to have been followed by a Baltic ice advance, which carried greater amounts of flint and Paleozoic limestones.

In Britain, the Anglian (Elsterian) tills are the first widespread unequivocal evidence of large-scale Pleistocene glaciation. The arrival of the Anglian ice sheet caused the burial of the Kesgrave Formation in all but the southernmost part of East Anglia. The Anglian ice sheet advanced into and overrode the contemporary valley of the River Thames in Hertfordshire and also advanced into its southern tributary valleys. Throughout the region, subglacial tunnel valleys were formed, radiating broadly in parallel with the ice movement. In northeastern

Norfolk, ice from both Scandinavia and the British glaciation centers interacted during the Anglian Stage. The ice sheet terminated in a large ice-dammed lake in the southern North Sea Basin. Although up to five different Anglian tills can be distinguished in parts of East Anglia, no evidence of a major intervening ice-free period has so far been discovered (Ehlers et al., 1991).

In eastern Poland, a counterpart of the Elsterian is represented by the postmaximum San 2 glaciation, which left surficial tills south of the Saalian ice limit (Lindner and Marks, 2008). However, farther eastward in Bielorrussia and central Russia, the contemporaneous tills of the Oka glaciation are overlain by tills of more extensive Dnieper and Moscow (Saalian) glaciations and therefore have never been described on the surface (Velichko et al., 2011).

In West Siberia, the thick tills of the two-stage Shaitan glaciation overlain by more extensive Saalian tills and local analogs of the Holsteinian have long been correlated with the double Elster glaciation (Volkova et al., 2002). The double Lebed till of the Yenisei valley buried well below sea level is also thought to be of Elsterian age, because it is overlain by fluvial and marine sediments of Tobol (Holsteinian) type. Its counterparts, as suggested by geologists, may form the ultimate drift limit in the east of the Central Siberian Uplands (Astakhov, 2004). This easternmost drift limit turns northward along the 114th meridian to get lost beyond the Arctic Circle because of the lack of data.

The North Sea Fan outside the mouth of the Norwegian Channel has been a depocenter of glacial sediments from southern Scandinavia, at least from MIS 12. Four pre-Weichselian glaciogenic debris-flow packages that were correlated with MIS 12, 10, 8, and 6, respectively, are identified in the fan (Hjelstuen et al., 2005; Nygård et al., 2005; Sejrup et al., 2005), indicating that the ice stream reached the mouth of the channel each time.

Along the coast of Norway, north of the Channel, there are till sheets in the continental shelf, showing that the ice sheet reached the shelf edge several times during the last 0.5 My. The tills cannot be directly dated or correlated with glaciations on the continent, but it is postulated that they were deposited during the Elsterian, Saalian, and other glaciations that expanded far south on the continent (Dahlgren et al., 2005; Hjelstuen et al., 2005; Ottesen et al., 2009; Rise et al., 2005).

Holsteinian Stage and Wacken/Dömnitz Substage (Saalian Stage)

Between the Elsterian and the Saalian glaciations, two or three interglacial events occurred, separated by intervening cold-climate intervals. During the Holsteinian, Northwest Europe experienced a major marine transgression, which was considerably more extensive than during the Eemian or the Holocene. In contrast to the Eemian, when the maximum sea-level highstand occurred only after the early part of the interglacial, the Holsteinian sea invaded parts of Jutland and North Germany by the end of the Elsterian. This early transgression may have been caused by strong isostatic depression by the Elsterian ice sheet. Besides glacio-isostatic adjustment, long-term crustal movements have to be taken into account. While in North

Germany the maximum of the Holsteinian transgression (uncorrected for later compaction) is situated at -13.5 m, the same level in East Anglia in the Nar Valley is located at $+23$ m, so a height difference of 36.5 m remains. Beyond the reaches of the Holsteinian sea, the interglacial is represented by fossil soils and lacustrine and organic strata. The Holsteinian was followed by a cold period in which permafrost developed in central Europe. This is referred to as the Fuhne cold substage (Saalian Stage).

The first evidence of an additional interglacial following the Holsteinian was discovered in the early 1960s, independently at Dömnitz (Mecklenburg-Vorpommern) and Wacken (Schleswig-Holstein), from where sequences with full forest development have been described. No marine transgression is known from this Wacken/Dömnitz Substage (Saalian Stage). The introduction of this new interglacial event was initially met with widespread skepticism. However, today it has been widely accepted that this Dömnitz Interglacial really exists (Litt et al., 2007). Recently, yet another interglacial, the Reinsdorf, has been introduced from the Schöningen site. Most importantly, in continuous sequences from boreholes in maar lakes in Massif Central in France, two warm interglacials are described between the Holsteinian and the Eemian, and these are correlated with the Reinsdorf and Dömnitz of Germany (Beaulieu et al., 2001).

The Holsteinian marine sediments with temperate fauna in Russia are tentatively described only in the Arctic, where they are known from boreholes penetrating below present sea level. They are thought to be synchronous with better-studied fluvial and lacustrine sequences of Likhvin (Holsteinian) type in European Russia (Velichko et al., 2004) and Tobol alluvial formation of West Siberia (Arkhipov et al., 1986). The Likhvin interglacial formation of European Russia contains the so-called *Brasenia* macroflora and pollen-reflecting

peculiar vegetation in which *Carpinus* coexisted with *Abies* and *Picea* with an admixture of exotic taxa of oceanic climate plants such as *Taxus*, *Ilex*, and *Pterocarya*. The Siberian Tobol formation contains forest fauna and shells of the central-asiatic freshwater mollusk *Corbicula tibetensis (fluminalis)*, shows normal magnetic polarity, and yields thermoluminescence dates in the range 260–390 ka BP. The Tobol alluvium, formed in a climate milder than the present, grades upward into glaciolacustrine facies of the subsequent Samarovo glaciation.

Saalian Glaciation (Saalian Stage)

Older Saalian Glaciation

The Saalian glaciation, like the Weichselian, began in North Germany, Denmark, and the Netherlands with a nonglacial cold phase of unknown, but probably rather long, duration. This was interrupted by several interstadials before the continental ice sheets invaded North Germany and the Netherlands.

The extent of what is called 'the Older Saalian' ice advance in Germany is shown in Figure 1. This maximum Saalian ice advance is correlated with the Odranian glaciation in Poland and the Dnieper glaciation in the Ukraine. In the west, it reached the Lower Rhine and left behind enormous pushed end moraines in the Netherlands (Koster, 2005) (Figures 2 and 3).

In the Netherlands, the Saalian glaciation (i.e., the 'Older Saalian glaciation') was originally subdivided into five different phases, all represented by push moraines, while only a single Saalian till sheet has ever been found. The Older Saalian till there occurs in a number of strikingly different lithofacies types. However, the lithologies involved do not represent deposits of different glaciations but a sequence of different ice masses within a single major ice advance.

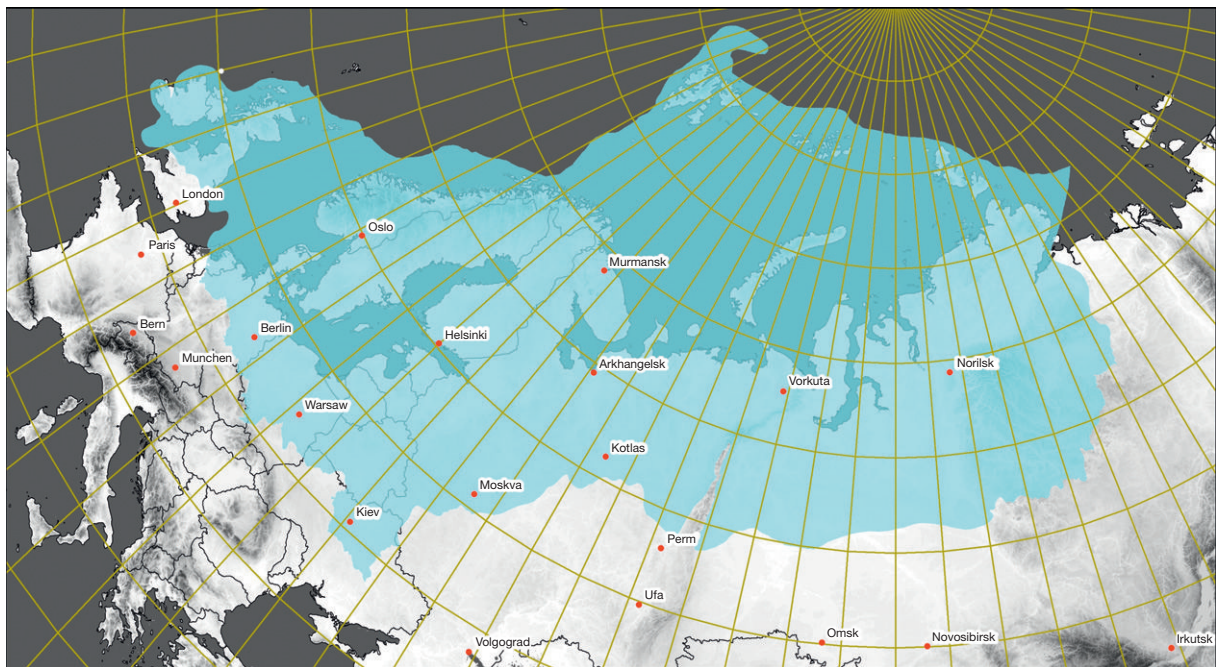


Figure 1 Extent of the Saalian glacial maximum. Reproduced from Ehlers J, Gibbard PL, and Hughes PD (eds.) (2011) *Quaternary Glaciations – Extent and Chronology, Part IV – A Closer Look, Developments in Quaternary Science*, vol. 15. Amsterdam: Elsevier.

In Denmark, three Saalian ice advances are distinguished; one of them is thought to be older than the German 'Older Saalian.' The oldest of these deposited a sandy, quartz-rich till, the Trelde Næs till, which contains abundant Norwegian rocks and other stones characteristic of an ice advance from a northerly direction. It is assumed that the ice crossed the border into Schleswig-Holstein. A second ice advance from a northeasterly direction followed, which – after accumulating fine-grained, coarsening-upward meltwater sands – deposited another sandy, quartz-rich till. This Ashoved till is characterized by a central Swedish indicator association (Kinne diabase) and Jurassic rocks incorporated at the margin of the Kattegat (Katholm erratics). This ice advance is probably the equivalent of the North German Older Saalian ice advance (Houmark-Nielsen, 2007).

In Britain, glaciation during what is almost certainly this phase is similar in its extent to that during the later Weichselian (Devensian) glaciation. Deposits from this Wolstonian Stage glaciation occur in the southern English Midlands, south of Birmingham, and in Yorkshire, Lincolnshire, and the Fenland Basin. The extent of ice lobes in the North Sea basin is poorly

known, but it is highly probable that British and Continental ice were confluent for a period during this phase.

At the end of the Older Saalian, the ice possibly melted back beyond the southern Baltic Sea coast before the Middle Saalian tills were deposited in Denmark and North Germany. The ice-free intervals between the individual ice advances of the Saalian in North Germany and adjoining areas are characterized by a lack of organic deposits, although all three advances left behind kettle holes that would have formed ideal sediment traps for the preservation of such deposits. Only during the subsequent Eemian Stage interglacial were those depressions foci of peat growth and gyttja, diatomite, or lake marl deposition. This seems to suggest that the ice-free periods were too short and/or too cold to allow extensive vegetational development.

Some striking intra-Saalian bleached loams on the Isle of Sylt (Figure 4) were most probably formed during an interstadial. The deep-penetrating wet bleach down to more than 10 m suggests pedogenesis under very moist conditions without permafrost. The pollen content of the bleached loam yielded evidence of open vegetation with grasses, *Calluna*, and other heathers (Ericaceae).

In Russia, for several decades, the Saalian was described as consisting of two substages: the maximum Dnieper correlated with Drenthe and the younger Moscow (or Warthe) glaciation separated by the Odintsovo (Roslavl) interglacial, presumably of MIS 7 interval. The age difference was supported by the difference in topography: only the area of the Moscow glaciation features spectacular push moraines and hummocky landscapes with boggy depressions containing Mikulino (Eemian) sediments. But now it is demonstrated beyond any doubt that the Roslavl strata with two deciduous pollen optima bear the Cromer fauna and flora and therefore the underlying till must be older than MIS 8 (Velichko et al., 2004).

Both Dnieper and Moscow glaciogenic sequences sandwiched between Likhvin (Holsteinian) and Mikulino (Eemian) interglaciations are currently related to MIS 6. The preceding Pechora (or Vologda) glaciation of northeastern provenance



Figure 2 Exposure in Saalian push moraine in The Netherlands.

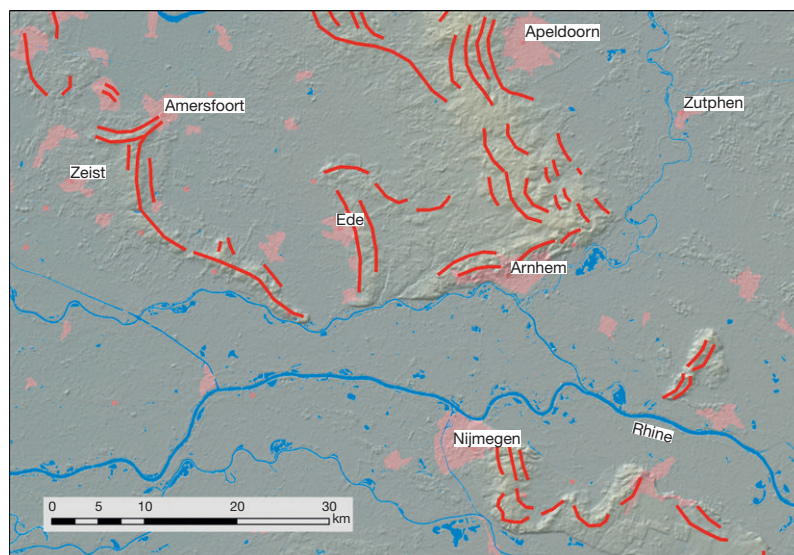


Figure 3 Saalian push moraines in the Lower Rhine area of Germany and The Netherlands.



Figure 4 Drenthe till with Saalian interstadial paleosol on Sylt, Germany.

was less extensive and did not reach the Moscow region. Formerly, this so-called ‘Novaya Zemlya till’ was correlated with the Dnieper till. Such correlation is now impossible because in the northeastern Russian Plain, thick peats with U/Th dates ca. 200–250 and OSL dates 170–200 ka BP are found between the Moscow and Pechora tills (Astakhov, 2004). This allows correlation of the Pechora till with tills of MIS 8 but not with any tills of MIS 6 time.

The Samarov glacial maximum of West Siberia is still considered part of the Saalian glaciation. This age is determined from apparently confirmable contact between the Samarov glaciolacustrine strata and the underlying Tobol (Holsteinian) interglacial alluvium. Only one horizon of interglacial soils and peats overlying the Samarov till has been described. Beneath this till, two pre-Holsteinian tills can be distinguished in the sedimentary sequences filling buried valleys, overdeepened to 200–300 b.s.l. These old tills have been found very close to the Samarov glacial limit, but not beyond it. However, considering the scarcity of geological information from this region, a pre-Holsteinian glaciation may well have been the most extensive in eastern Central Siberia.

A similar problem to that in Europe is encountered in Siberia, where the less extensive Taz till is thought to represent MIS 6, whereas the maximum Samarov glaciation, traditionally correlated with the Dnieper glaciation in Europe and Drenthe, is related to MIS 8. This correlation was apparently supported by rather conspicuous glacial topography of the Taz glaciation area in places covered by late Pleistocene interglacial sediments. Lately, however, one of the interglacial marine formations, previously correlated with the Eemian, has been proven to underlie the Taz till of MIS 6 on the Lower Ob river (Astakhov and Nazarov, 2010). This apparently confirms that the maximum Samarov glaciation cannot be younger than MIS 8.

Immediately west of the Urals, the Saalian ice limits are also uncertain. The Saalian limit was most probably positioned in the Volga–Pechora interfluvial area, just north of the maximum drift limit.

Middle Saalian Glaciation

The subsequent ice advance began in North Germany with the deposition of meltwater sands. In contrast to the glaciofluvial

deposits of the Elsterian, which in North Germany are largely concentrated in buried channels, the meltwater of the Middle Saalian accumulated vast outwash fans, the deposits of which can be several tens of meters thick. During the maximum phase of this advance, when the ice lay south of the present Elbe valley, the meltwater drained southward and then, via the Aller–Weser ice-marginal valley (‘urstromtal’), toward the North Sea. It is thought that the Middle Saalian ice advanced at least as far west as the Altenwalder Geest hills, south of Cuxhaven. Within the area covered by the Middle Saalian glaciation, a number of marked end moraine ridges are found. They include the hills of the Neuenkirchen and Falkenberg ice-marginal positions.

In Denmark, the Trelde Næs till and Ashoved till (both Older Saalian) are overlain by a widespread chalk-rich Saalian till (Lillebælt till), separated from the older tills by sands. It is found from the German border to Northern Jutland in the north (Skærumhede). Lithologically, it resembles the North German Middle Saalian till. However, as this is the uppermost Saalian till in Denmark, the Lillebælt till has been correlated with the North German Younger Saalian till. It probably represents both tills.

Younger Saalian Glaciation

After the Middle Saalian glaciation, the ice margin retreated far to the northeast and parts of the ice sheet stagnated and melted *in situ*. The active ice margin probably lay in the present Baltic Sea area. In the proglacial zone, widespread dead ice masses were covered by outwash during the subsequent readvance.

Indicator counts have demonstrated that the composition of the Younger Saalian till varies locally. In northeastern Niedersachsen, a red-brown, east Baltic facies of the Younger Saalian till is widespread. At the outermost ice margin in Schleswig–Holstein, tills that are rich in east Baltic indicator clasts (e.g., at Hohenwestedt and Osterrade) have also been found. In immediately adjoining areas, however, the east Baltic components are lacking. For example, in northwestern Schleswig–Holstein, the flint content of this till is higher, so it cannot be distinguished from the Middle Saalian till.

A line of marked end moraines in the northern Lüneburger Heide area has traditionally been associated with the Younger Saalian ‘Warthe’ ice advance. However, more recent investigations show that many assumed Younger Saalian end moraines in Niedersachsen often have a much older core.

In contrast to the Middle Saalian glaciation, outwash accumulation during the Younger Saalian seems to have been rather limited. Many areas that were previously interpreted as Younger Saalian sandur plains have been found to be covered by patches of Middle Saalian till.

The Middle Saalian and Younger Saalian glaciations seem to be closely related. They are both probably the equivalents of Woldstedt’s Warthe (Warta) Substage (Figure 5). Further to the east, the Moscow glaciogenic sequence, the equivalent of the Polish Warta glaciation, makes vast hummocky terrains with large push moraines in Bielorrussia and Central Russia. The end moraines are fringed by huge boggy sandurs locally called ‘polesye.’ Toward the east, its outer limit converges with that of the Older Saalian (Dnieper) glaciation, and it may even have advanced beyond the Dnieper glacial limit. In West

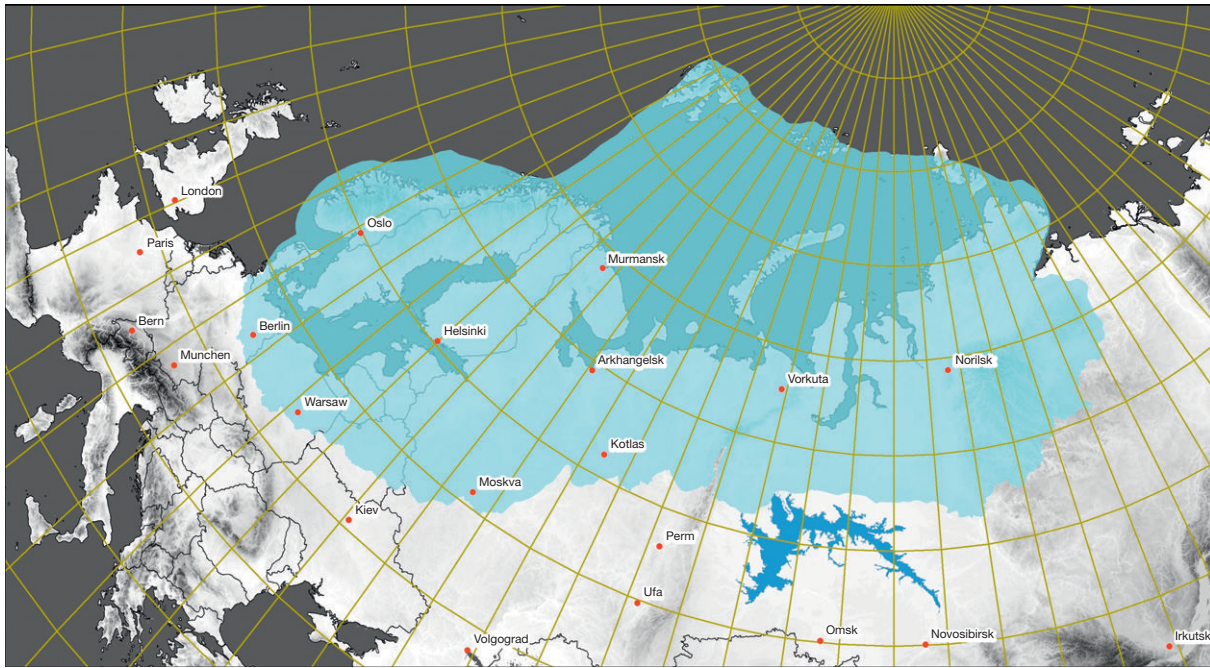


Figure 5 Extent of the Saalian Moscow glaciation. Reproduced from Ehlers J, Gibbard PL, and Hughes PD (eds.) (2011) *Quaternary Glaciations – Extent and Chronology, Part IV – A Closer Look. Developments in Quaternary Science*, vol. 15. Amsterdam: Elsevier.

Siberia, the Taz (or Yenisei) glaciation, traditionally correlated with the Moscow–Warta Substage, is supposed to be the final middle Pleistocene event, ca. 140–170 ka BP according to luminescence dating.

Whereas the Saalian tills of Central Russia are undoubtedly Scandinavian, east of 54° E, both Saalian glaciations were centered on the Kara Sea shelf as seen from the fabrics and orientation of push moraines. The ice sheets were up to 1-km thick even close to the ice limit, which is indicated by foreign tills, erratics, and striations on flat summits of the Urals and Central Siberian Uplands. The maximum calculated ice thickness in the lowland Yamal Peninsula is 3.5 km. This Antarctic thickness is amply manifested in pervading overthrusts and other disharmonic disturbances of the Cretaceous and Cenozoic substrates, in places penetrating 300–400 m beneath the surface. The ice sheets operated mostly on the perennially frozen substrate. This is evident from many intraclasts of stratified sand that could be transported only in a frozen state and which constitute a sizeable volume of lowland tills poor in crystalline pebbles.

Other signatures of cold-based glaciation are the lack of flutes, eskers, and ice retreat features. Siberian tills are expressed in hummocky terrains, mostly over hard Palaeozoic rocks, whereas in sedimentary lowlands, they are overlain by sand in the form of outwash plains or ice-disintegration features such as kames (Astakhov, 2004). Shallow proglacial lakes left varves, mostly in the ice-dammed river valleys. The thickest glaciolacustrine silty rhythmites occur mainly beyond the Arctic Circle, where they often laterally grade into meltout diamictons. This was probably the deepest isostatic trough where stagnant remnants of disintegrated Saalian ice slowly decayed awaiting the Eemian transgression.

In the east of the Russian Arctic, stagnant middle Pleistocene glaciers have survived until our day because the subsequent

interglacial sea had negative temperatures. Continuous strata of deformed basal glacial ice, which are glacial deposits *sensu stricto*, led to the sensational reconstruction of an ice sheet of the final middle Pleistocene on the shelf and islands of the East Siberian Sea. The ice sheet of a rather modest extent, ~600 × 400 km, was located between longitudes 141 and 159° E, that is, in the far eastern area previously thought unfavorable for ice accumulation (Anisimov et al., 2006; Basilyan et al., 2008). Its main signature is not till but dirty glacier ice buried under frozen marine clay with a distinct thaw contact (Figure 6). Prior to this discovery, fossil glacier ice in Russia was known only as a remnant of the late Pleistocene glaciers.

As mentioned earlier, there are tills of assumed Saalian age on most of the Norwegian continental shelf, but the tills cannot be correlated precisely with the different tills in Germany. Dowdeswell et al. (2006) identified an ice-stream path from the Saalian that had a different direction and also was longer than the corresponding ice stream during the Late Weichselian.

The Alpine Glaciations

In the Alps, a marked asymmetry is observed between the glaciation of the northern and southern slopes. On the southern side of the Alps, the ELA is about 200 m higher than in the north. This contrast is further enhanced by the different altitudes of the two Alpine forelands (about 500 m in the north as opposed to about 100 m in the south). As a consequence, Quaternary glaciers advanced far into the northern foreland of the Alps, whereas in Italy they halted as soon as they reached the foot of the mountains.

The ice streams of the Alps have reshaped a preexisting fluvial relief to a glacially molded landscape. One characteristic element of this relief is the U-shaped valley troughs. The upper

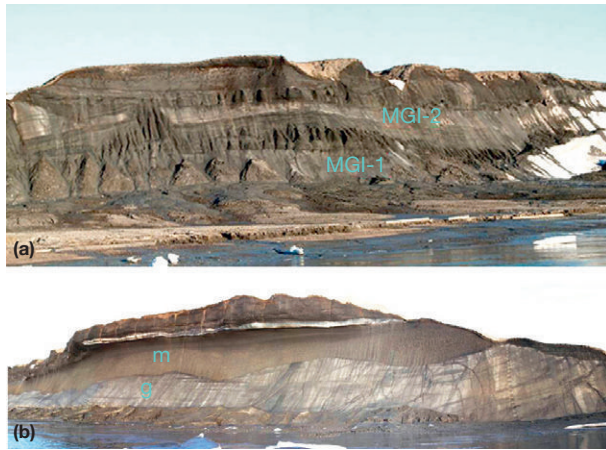


Figure 6 Easternmost fossil ice preserved from the middle Pleistocene on the New Siberian Islands. (a) Middle Pleistocene basal glacier ice (MGI-1) overlain by marine clay and by late Pleistocene fossil glacier (MGI-2). (b) Middle Pleistocene fossil glacier (g) overlain by marine clay (m) with distinct thaw contact. Photographs by V. Tumskoy and A. Basilyan.

limit of glacial erosion ('Schliffgrenze') is well developed, especially in the central Alps. Since glacial erosion was much stronger in the main valleys than in the tributaries, many of the latter today finish as hanging valleys, high above the modern main valley floor.

Glacial erosion largely modified the original longitudinal profile (thalweg) of the valleys. Numerous overdeepened basins were formed. Seismic investigations in the Inn valley have revealed that Quaternary sediments reach a thickness of about 900–1000 m there. Between Wattens and Schafteuau, the base of the Quaternary lies at a depth of 300–400 m b.s.l. Overdeepened rock basins are found, especially in narrow sections of the Alpine valleys, where the ice had to pass at particularly high velocities. Where valley glaciers merged, pronounced excavation resulted in the formation of confluence basins (e.g., near Innertkirchen in the Aare Valley). The most noteworthy overdeepenings still visible in the field, however, are found near the outer margins of the glaciations. Glacial tongue basins occupied by deep lakes are characteristic of both the southern and northern Alpine margins (e.g., Lake Constance and Lake Garda). Overdeepening of the lake basins at the southern Alpine margin reaches maximum values of 660 m at Lago di Como and 680 m at Lago Maggiore, with Lago di Como being 410 m and Lago Maggiore 372 m deep at present.

The classic Alpine Quaternary sequence is largely a morphostratigraphy (Jerz, 1993). It is based upon the concept that the so-called glacial series formed during any ice advance consists of a sequence of tongue basins with drumlins, moraine belts, and gravel spreads. In the northern foothills of the Alps, comprehensive morphological sequences of this type were initially found for only three glaciations; for the fourth, this evidence was produced much later through investigations in Upper Austria. From the early Pleistocene, the Biber and Donau glaciations, no such landform assemblages have been preserved.

Penck and Brückner (1901/1909) did not consider any glacial deposits to be older than the Günz cold stage. Therefore, all gravel occurrences at higher altitudes than the

Mindel Stage 'Jüngere Deckenschotter' were summarily referred to as 'Ältere Deckenschotter' and assigned to the Günz cold stage. Today, the 'Ältere Deckenschotter' sensu stricto are distinguished from the older gravel spreads of the Donau and Biber cold stages.

Originally, it had been assumed that the Günz was younger than the Matuyama/Brunhes boundary. However, on the Höchsten and near Heiligenberg (north of Lake Constance), tills and gravels of the Günz glaciation were found to be reversely magnetized. Consequently, at least part of the Günz Complex has to be assigned to the Matuyama Epoch. Near Rosshaupten in the Allgäu, the Matuyama/Brunhes reversal has also been found above the 'Ältere Deckenschotter' of the Günz.

According to the latest investigations, the Günz comprises the deposits of three cold stages. The older parts of this Günz Complex are reversely magnetized. The majority of the deposits, however, seem to be younger than the Brunhes/Matuyama boundary. End moraines of Günz age have been identified only in the area between Iller and Wartach glacier tongues. Apparently, the Günz glaciers advanced further into the foreland in this region than anywhere else (Habbe, 2007).

While the relative stratigraphy of the gravel terraces is undisputed, the chronostratigraphical determination of the various early cold stages remains very problematic. First attempts to date the older gravels by means of cosmogenic nuclides have yielded an age of 2.35 Ma for the Günz gravels, south of Memmingen, and for the Mindel gravels, an age of 680 000 years (Häuselmann et al., 2007).

The exact number of glaciations in the Alps prior to the Riss Stage is uncertain. Between the Günz and the Mindel, a separate Haslach glaciation has been postulated, based on the occurrence of distinct gravel units in the Württemberg Rottal (south of Ulm). Today, the Haslach glaciation is largely accepted as an independent unit.

The 'Jüngere Deckenschotter' of the Mindel valley and their association with moraines north of Obergünzburg caused Penck to name his second Alpine glaciation the Mindel cold stage.

The interglacial between the Mindel and Riss cold stages in the Alps was traditionally referred to as the 'Great Interglacial.' It is correlated with the Holsteinian interglacial of Northern Germany. The interglacial is characterized by the presence of *Pterocarya* (winged nut), *Fagus* (beech), and *Buxus* (yew). There are two well-studied sites that fit these characteristics, namely, Borehole Samerberg 2 (Grüger, 1983) and the Thalgut site in Switzerland (Welten, 1988).

When the glaciers advanced again after the Mindel–Riss Interglacial, they encountered a landscape that had already been sculpted by earlier ice advances. At the alpine margin, tongue basins already existed. But in the Riss Stage, the glaciers in many areas advanced well beyond the maximum extent of the previous glaciations. The maximum glaciation in the Emmental region (Switzerland) and the correlative Höhenschotter gravels based on U/Th datings are now correlated with the MIS 6 (Riss) (Preusser, 2010).

In southern Germany, theories of what should be referred to as the 'Riss cold stage' have also undergone a marked change in the course of time. When they originally established their Alpine Quaternary chronology, Penck and Brückner started from the assumption that each of their four cold stages

represented a single, uniform glaciation. However, within the Rhine Glacier region, the Riss glaciation can be subdivided into three ice advances. The Older Riss deposits are marked by a greater degree of weathering; in the Riss valley and neighboring valleys, they underlie the deposits of the Middle Riss from which they are separated by gravels. The extent of the intervening glacial retreat has not yet been assessed. The differences in weathering give reason to believe that the break might have been at least of interstadial status. The distribution of the till of this first ice advance is restricted to former glacier tongues; accordingly, the Middle Riss is referred to as the Zungen-Riss ('tongue Riss').

In the eastern part of the Rhine Glacier region (between Biberach and Leutkirch), the Middle Riss is confined by a double rampart of end moraines, the so-called Doppelwall-Riss ('double rampart Riss'). The end moraines are generally gravelly push moraines 10–30 m in height, formed at an interval of 1–3 km. They delineate the margin of the Middle Riss glaciation in the eastern region of the Rhine Glacier. In contrast to this pronounced landform assemblage, the end moraine of the Younger Riss is incompletely developed. It includes the moraine ridges north of Ingoldingen and south of Eberhardzell, at the eastern margin of the former Rhine Glacier.

The stratigraphy of the Alpine Riss cold stage is still far from solved – at least in major parts of the glaciated area. Neither the precise extent of the glaciers nor the number of ice advances, or their age, can be determined with certainty yet. Under these conditions, a direct correlation of the classic Riss or parts of it with the Saalian glaciation of Northern Europe is not yet possible.

See also: Plant Macrofossil Records: Arctic Eurasia. Glaciations: Late Pleistocene in Eurasia; Overview.

References

- Anisimov MA, Tumskoy VYe, and Ivanova VV (2006) Massive ice of the New Siberian Islands as a relict of ancient glaciation. *Materialy glatsiologicheskikh issledovaniy* 101: 143–145 (in Russian).
- Arkipov SA, Isayeva LL, Besspal'y VG, and Glushkova O (1986) Glaciation of Siberia and North-East USSR. *Quaternary Science Reviews* 5: 463–474.
- Astakhov V (2004) Middle Pleistocene glaciations of the Russian north. *Quaternary Science Reviews* 23(11–13): 1285–1311.
- Astakhov V and Nazarov D (2010) Correlation of Upper Pleistocene sediments in northern West Siberia. *Quaternary Science Reviews* 29: 3615–3629.
- Basilyan AE, Anisimov MA, and Nikolsky PA (2008) Pleistocene glaciation of the New Siberia Islands – No doubts left. *Information Bulletin IPY 2007/2008 News*, vol. 12, pp. 7–9. St. Petersburg: AARI (translated from Russian).
- Beaulieu J-L, Andrieu V, Reille M, et al. (2001) An attempt at correlation between the Velay pollen sequence and the Middle Pleistocene stratigraphy from central Europe. *Quaternary Science Reviews* 20: 1593–1602.
- Dahlgren KIT, Vorren TO, Stoker MS, Nielsen T, Nygård A, and Sejrup HP (2005) Late Cenozoic prograding wedges on the NW European continental margin: Their formation and relationship to tectonics and climate. *Marine and Petroleum Geology* 22: 1089–1110.
- Dowdeswell JA, Ottesen D, and Rise L (2006) Flow switching and large-scale deposition by ice streams draining former ice sheets. *Geology* 34: 313–316.
- Ehlers J, Gibbard P, and Rose J (eds.) (1991) *Glacial Deposits in Great Britain and Ireland*. Rotterdam, The Netherlands: Balkema.
- Eissmann L (1975) Das Quartär der Leipziger Tieflandsbucht und angrenzender Gebiete um Saale und Elbe. Modell einer Landschaftsentwicklung am Rand der europäischen Kontinentalvereisung. *Schriftenreihe für Geologische Wissenschaften* 2: 263.
- Grüger E (1983) Untersuchungen zur Gliederung und Vegetationsgeschichte des Mittelpleistozäns am Samerberg in Oberbayern. *Geologica Bavarica* 84: 21–40.
- Habbe KA (2007) Stratigraphische Begriffe für das Quartär des süddeutschen Alpenvorlandes. *Eiszeitalter und Gegenwart* 56(1/2): 66–83.
- Häuselmann Ph, Fiebig M, Kubik PW, and Adrian H (2007) A first attempt to date the original 'Deckenschotter' of Penck and Brückner with cosmogenic nuclides. *Quaternary International* 164–165: 33–42.
- Hjelstuen B, Sejrup H, Halldason H, Nygård A, Ceramicola S, and Bryn P (2005) Late Cenozoic glacial history and evolution of the Storegga Slide area and adjacent slide flank regions, Norwegian continental margin. *Marine and Petroleum Geology* 22: 57–69.
- Houmark-Nielsen M (2007) Extent and age of Middle and Late Pleistocene glaciations and periglacial episodes in southern Jylland, Denmark. *Bulletin of the Geological Society of Denmark* 55: 9–35.
- Iosifova Yul, Agadzhanian AK, Pisareva VV, and Semenov VV (2006) The Upper Don basin as a stratotypic area of the Middle Pleistocene of the Russian Plain. In: Zubakov VA (ed.) *Palinologicheskkiye, klimatostatigraficheskiye i geokologicheskkiye rekonstruktsii*, pp. 41–84. St. Petersburg: Nedra (in Russian, English summary).
- Jerz H (1993) *Das Eiszeitalter in Bayern-Erdgeschichte, Gesteine, Wasser, Boden, Geologie von Bayern, No. 2*. Stuttgart: Schweizerbart.
- Koster EA (ed.) (2005) *The Physical Geography of Western Europe*. Oxford: Oxford University Press.
- Lee JR, Rose J, Hamblin RJO, and Moorlock BSP (2004) Dating the earliest lowland glaciation of eastern England: A pre-MIS 12 early Middle Pleistocene Happisburgh glaciation. *Quaternary Science Reviews* 23: 1551–1566.
- Lindner L and Marks L (2008) Pleistocene stratigraphy of Poland and its correlation with stratotype sections in the Volhynian Upland (Ukraine). *Geochronometria* 31: 31–37.
- Lisiecki LE and Raymo ME (2005) A Pliocene–Pleistocene stack of 57 globally distributed benthic $\delta^{18}\text{O}$ records. *Paleoceanography* 20: PA1003.
- Litt T, Behre K-E, Meyer K-D, Stephan H-J, and Wansa S (2007) Stratigraphische Begriffe für das Quartär des norddeutschen Vereisungsgebietes. *Eiszeitalter und Gegenwart* 56(1/2): 7–65.
- Molodkov AN and Bolikhovskaya NS (2006) Long-term palaeoenvironmental changes recorded in palynologically studied loess–palaeosol and ESR-dated marine deposits of Northern Eurasia: Implications for sea–land correlation. *Quaternary International* 152–153: 48–58.
- Nygård A, Sejrup HP, Halldason H, and Bryn P (2005) The glacial North Sea Fan, southern Norwegian Margin: Architecture and evolution from the upper continental slope to the deep-sea basin. *Marine and Petroleum Geology* 22: 71–84.
- Ottesen D, Rise L, Andersen ES, Bugge T, and Eidvin T (2009) Geological evolution of the Norwegian continental shelf between 61° N and 68° N during the last 3 million years. *Norwegian Journal of Geology* 89: 251–265.
- Penck A and Brückner E (1901/1909) *Die Alpen im Eiszeitalter*. Leipzig, Germany: Tauchnitz.
- Preusser F (2010) Stratigraphische Gliederung des Eiszeitalters in der Schweiz (Exkursion E am 8 April 2010). *Jahresberichte und Mitteilungen des Oberrheinischen Geologischen Vereins, N.F.* 92: 83–98.
- Rise L, Ottesen D, Berg K, and Lundin E (2005) Large-scale development of the mid-Norwegian margin during the last 3 million years. *Marine and Petroleum Geology* 22: 33–44.
- Sejrup HP, Hjelstuen BO, Dahlgren KIT, et al. (2005) Pleistocene glacial history of the NW European continental margin. *Marine and Petroleum Geology* 22: 1111–1129.
- Toucanne S, Zaragosi S, Bourillet JF, et al. (2009a) Timing of massive 'Fleuve Manche' discharges over the last 350 kyr: Insights into the European Ice Sheet oscillations and the European drainage network from MIS 10 to 2. *Quaternary Science Reviews* 28: 1238–1256.
- Toucanne S, Zaragosi S, Bourillet JF, et al. (2009b) A 1.2 Ma record of glaciation and fluvial discharge from the West European Atlantic margin. *Quaternary Science Reviews* 28: 2974–2981.
- Velichko AA, Faustova MA, Gribchenko YuN, Pisareva VV, and Sudakova NG (2004) Glaciations of the East European Plain – Distribution and chronology. In: Ehlers J and Gibbard PL (eds.) *Quaternary Glaciations – Extent and Chronology. Part 1: Europe*, pp. 337–354. Amsterdam: Elsevier.
- Velichko AA, Faustova MA, Pisareva VV, Gribchenko YuN, Sudakova NG, and Lavrentiev NV (2011) Glaciations of the East European Plain - distribution and chronology. In: Ehlers J, Gibbard PL and Hughes PD (eds.) *Quaternary Glaciations - Extent and Chronology. A Closer Look. Developments in Quaternary Science* Vol. 15, pp. 337–359.
- Volkova VS, Arkipov SA, Babushkin AYe, et al. (2002) *Stratigrafia neftegazonosnykh basseinov Sibiri. Kainozoi Zapadnoi Sibiri (Stratigraphy of Oil and Gas Basins of Siberia. Cenozoic of West Siberia)*, p. 246. Novosibirsk: Siberian Branch RAS, Dept. 'GEO' (in Russian).
- Welten M (1988) Neue pollenanalytische Ergebnisse über das Jüngere Quartär des nördlichen Alpenvorlandes der Schweiz (Mittel- und Jungpleistozän). *Beiträge zur Geologischen Karte der Schweiz N.F.* 162: 40.

Mid-Quaternary in North America

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Contrary to the fourfold stratigraphic framework developed for Pleistocene terrestrial glacial sediments in central North America (Wisconsinan, Illinoian, Kansan, and Nebraskan; [Flint, 1957](#)), the record of global ice volume contained in the oxygen isotope ($\delta^{18}\text{O}$) record of deep-sea sediments provides an elaborate and more complete history of glaciations ([Figure 1](#); [Hays et al., 1976](#); [Imbrie et al., 1984](#); [Mix et al., 1995](#); [Ruddiman et al., 1989](#)). The beginning of middle Pleistocene time has been set at the last major magnetic reversal, 780 thousand years ago (ka), the transition from Matuyama Reversed Chron to the Brunhes Normal Chron ([Renne et al., 1994](#); [Shackleton et al., 1990](#); [Spell and McDougall, 1992](#)). Middle and late Pleistocene glaciations therefore are entirely within the Brunhes Normal Chron. Soon after the Brunhes–Matuyama reversal, the global ice volume increased significantly beginning with Marine Isotope Stage (MIS) 16, based on the magnitude of the $\delta^{18}\text{O}$ isotopic excursion, which is related to the volume of water transferred from oceans to land and stored as ice. It was also at this time that the period between glacial maxima became longer, increasing from approximately 41 000 years during the early Pleistocene to 100 000 years during the Middle and late Pleistocene ([Imbrie et al., 1993](#); [Pisias and Moore, 1981](#); [Ruddiman et al., 1989](#)).

The volume of continental ice inferred for peak middle Pleistocene glaciations dictates that the global extent of glaciation during these times was similar to the extent in the most recent glacial cycle. Limited stratigraphic preservation and exposure on land, however, have affected the ability to confirm the character and extent of these interpreted glacial events by mapping alone, except in places where the records of older glaciations were more extensive than younger events. Statistical simulations that follow the simple rule of ‘obliterative overlap’ – that is, that glaciations more extensive than prior glaciations obliterate the record of less extensive, older glaciations – determine the probability of moraine survival. Only 3 out of 10 random-size glaciations would be preserved at the surface if these very simplified rules are followed ([Gibbons et al., 1984](#)). A natural corollary is that the oldest glaciation preserved is also the most areally extensive record at the surface and that progressively younger glaciations are progressively less extensive. If this statistical argument is applied to the geological record, the mappable remnants of glaciations would be expected to represent the larger isotopic excursions of the middle Pleistocene – that is, MIS 16, 12, and 6. The basis for the original fourfold chronology in the midcontinent may therefore simply be a result of the statistical likelihood of moraine, till, and erratic survival at the surface.

In regions such as the midcontinent of North America, where older glaciations extend beyond the Late Wisconsinan ice margin, they have been described, at least qualitatively, for more than a century. Of these, the Illinoian (MIS 6), the only

pre-Wisconsinan glaciation that retains its original name, is the best documented ([Aber, 1999](#); [Hallberg, 1980a](#); [Williman and Frye, 1970](#)). The MIS 6 Lake Michigan lobe reached farther south than any other glaciation in southern Illinois ([Williman and Frye, 1970](#)), but farther west along the ice margin, the Illinoian ice was much less extensive than older glaciations. On the west side of the Driftless Area in Wisconsin, there are extensive areas of pre-Illinoian till at the surface; on the east side (e.g., in the Baraboo Hills of Wisconsin), the Late Wisconsinan MIS 2 ice reached as far as any older glaciation ([Farrand et al., 1984](#)). In the Rockies, there is variability in ice extent, with pre-MIS 6 (pre-Bull Lake) and/or MIS 6 (Bull Lake) till extending well beyond the MIS 2 (Pinedale) moraines in some areas (e.g., in the southwestern Wind River Range) but not detectable at all in other places ([Dahms, 2004](#); [Hall, 1999](#)).

While acknowledging the obvious limitation of the fragmentary nature of the terrestrial record of glaciations, correlation and dating of continental glacial records of possible middle Pleistocene age remains a priority to confirm, to the extent possible, the timing, extent, and character of the glacial cycles. Recent advances toward this goal have been due, in part, to progress in magnetostratigraphy and cosmogenic exposure and burial dating methods.

Correlation of Glaciated Landscapes

Where middle Pleistocene deposits outcrop, these landscapes have been separated from younger landscapes on the basis of soil development, depth of leaching, and degradation of glacial landforms (e.g., Cincinnati, Ohio; [Szabo and Chanda, 2004](#)). middle Pleistocene landscapes may also be eroded by wind and thereby display a faceted stone ‘lag’ or may be buried by loess or sand, especially where located near the Late Wisconsinan ice margin. These weathering and erosion horizons, where preserved, make the task of recognizing these older landscapes easier. However, these landscapes may have also been eroded by hillslope processes, especially those that allowed the movement of the active layer on low slopes when permafrost was present. This makes the task of recognizing them more difficult because of the removal of well-developed soils, as on the ‘Iowan Erosion Surface’ ([Ruhe, 1969](#)).

In mountain glacier systems, boulder weathering, pitting, grussification, and degradation of moraine form are commonly used as relative dating methods and have been supported by $^{36}\text{Cl}/^{10}\text{Be}$ exposure ages, for example, for the Bull Lake moraines in the Rockies, dated as being >130–95 ka ([Dahms, 2004](#)). In California, the Mono Basin glaciation was originally assigned to the middle Pleistocene based on these relative dating methods and has been confirmed as being MIS 6 based on ^{36}Cl exposure dates of 80–60 ka ([Gillespie and](#)

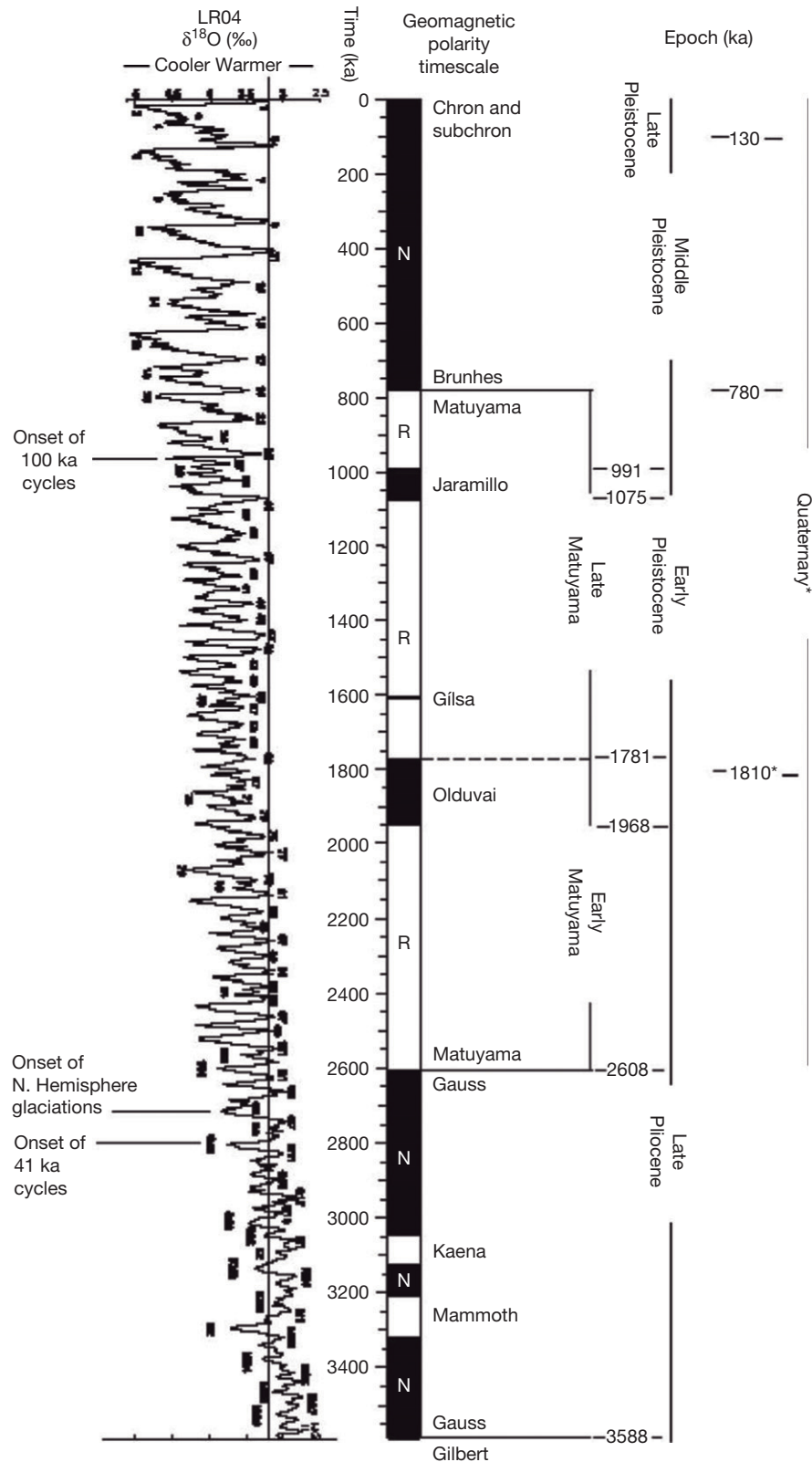


Figure 1 Geomagnetic polarity timescale for LR04 benthic and $\delta^{18}\text{O}$ paleotemperature profile. Black (white) areas represent normal (reversed) polarity. Marine Isotope Stages (MIS) are labeled on LR04 (even numbers represent glacials and odd numbers interglacials). MIS scheme follows Ruddiman et al. (1989) and Raymo et al. (1989) from the present to MIS 104 and Shackleton et al. (1995) in the Gauss Chron. Arrow marks Holocene mean $\delta^{18}\text{O}$. *Gradstein et al. (2004) and Pillans and Naish (2004). Reproduced from Lisiecki LE and Raymo ME (2005) A Pliocene–Pleistocene stack of 57 globally distributed benthic $\delta^{18}\text{O}$ records. *Paleoceanography* 20, with permission.

Zehfuss, 2004). Alaska has a particularly well-preserved middle Pleistocene record owing to the limited extent of Wisconsin ice. Glacial units were qualitatively dated by relative weathering, paleosol formation, and tectonic evolution and further bracketed in time by magnetic chronology of underlying sediment and overlying basalts and amino-acid racemization dates on underlying shells (Hamilton, 2001; Manley et al., 2001). There are exceptions, however, where surfaces that were assumed to belong to 'early' glaciations because of their relative weathering and discontinuous preservation are actually younger than expected. One such landscape, represented by scattered granitoid erratics southwest of the Missouri River in North Dakota, was tentatively revised as being early Late Wisconsin in age, between 27.8 and 28.8 ka, based on ^{36}Cl exposure dating (Manz, 2005). Similarly, on Baffin Island, young erratics indicated that heavily weathered upland surfaces had been overrun by latest Pleistocene ice that was cold based (Bierman et al., 1999; Marsella et al., 2000).

In the Midwest, particularly in Iowa and Missouri, the pre-Illinoian tills have good continuity in the subsurface and are distinct enough lithologically to correlate over broad areas (Hallberg, 1980a,b; Rovey and Kean, 1996). However, this is not the case everywhere with pre-Illinoian units at the surface. If the till sheets are discontinuous, it can be difficult to extend the exposed units into the subsurface, especially if tills of different age are lithologically similar and lack identifiable horizons. For example, pre-Illinoian deposits exposed in a limited area at the surface in New Jersey have not been correlated to subsurface units anywhere in New England (Stone et al., 2002). In some places, it is even difficult to distinguish Pleistocene units from remnants of pre-Pleistocene units in the subsurface (e.g., Nova Scotia; Stea, 2004).

In some areas, there are no older ice margins on land that extend beyond the Late Wisconsin margin (e.g., Maine; Borns et al., 2004) or they extend for only a short distance (e.g., Pennsylvania, Braun, 2004; Montana, Fullerton et al., 2004). Additionally, in many places outside the midwestern United States, subglacial erosion during the last glaciation has removed all but a few patches of glacial sediment attributed to glaciations prior to MIS 2 (e.g., New York, Cadwell and Muller, 2004; Nova Scotia, Stea, 2004). This appears to be more commonly reported in the east, where crystalline substrates produced sandy tills that are more easily reworked during subsequent glaciations.

In areas where thick glacial sequences are preserved (e.g., South Dakota, Minnesota, Michigan, Iowa, and Missouri), the till is derived from fine-grained sedimentary rocks (Soller, 2004) and is therefore more cohesive and resistant to erosion. However, the lack of subsurface information away from natural exposures even in the areas where thick sequences are preserved makes distinguishing the remnants of older glaciations very difficult. Most geological surveys have simply lumped similar glacial units or 'counted down from the top,' attributing units in the same stratigraphic position to the same glaciation. For example, in New Hampshire, units older than Wisconsin glaciation in the subsurface have traditionally been treated as a single older unit of Illinoian age (Koteff and Pessl, 1985). In Minnesota, tills in the subsurface that are presumed to be pre-Late Wisconsin have only recently been distinguished and correlated regionally, mainly on the

basis of details in texture and lithology rather than on dated horizons (Meyer, 1997, 2005). In exposures or cores, the distinctions are problematic if soil profiles are truncated or if the time between glaciations did not allow recognizable weathering profiles to develop.

Dating Tools

Dating tools that have been used to confirm or refute suggested middle Pleistocene ages include amino-acid racemization; optically stimulated luminescence of sediment formerly exposed to light (e.g., loess); uranium–thorium dating of shells where they are incorporated or of calcrete soils that cap glacial units; radionuclide dating of volcanic ashes or lava where they intercalate glacial deposits; magnetostratigraphy; and cosmogenic isotope dating. The latter two techniques rely less on fortunate occurrences of datable material and are being more widely applied.

Magnetostratigraphy can place normally magnetized sediment of the middle Pleistocene within the Brunhes Normal Chron (Barendregt and Duk-Rodkin, 2004) and is especially useful if transition from the Matuyama Reversed Chron is captured in a sequence because it marks a unique event in time. In Pennsylvania, large proglacial lakes record the transition from Matuyama to Brunhes (Braun, 2004). In Missouri, the early Pleistocene, reversed Moberly Formation is distinguished from three overlying normally magnetized middle Pleistocene units of the McCredie Formation (Colgan, 1999; Rovey and Kean, 1996). The same sequence is present in Iowa; the reversed Alburnett Formation is overlain by three normal tills of the Wolf Creek Formation (Baker and Stewart, 1984; Rovey et al., 2002). Findings from northeastern Kansas and northwestern Missouri (Aber, 1991) restricted the 'classical Kansas' till (now Independence Formation) to the age range ~600–700 ka based on the normal polarity of the till and a dated volcanic ash. The Independence Formation is therefore related to MIS 16 (or possibly 18), the most extensive glaciation with a record preserved on land in the central United States. Also for the midcontinent, Roy et al. (2004a,b) presented magnetostratigraphical data together with dated units that are constrained by three volcanic ashes (0.6, 1.2, and 2.0 Ma) derived from eruptions of the Yellowstone Caldera present in the sections (Boellstorff, 1978; Izett, 1981). They presented a summary map of the limit of Early and middle Pleistocene ice in eastern North America, in which a large fraction of the glacial deposits are now assigned to the middle Pleistocene (Figure 2; Roy et al., 2004a,b). The southernmost border of the eastern portion of the ice sheet extends from northeastern Kansas through Missouri, southern Illinois, the southern tip of Indiana, the northern part of Kentucky, includes all of Ohio, crosses the middle of West Virginia, and includes all of New York and the northern part of New Jersey.

A great deal of paleomagnetic work has been conducted in northwestern North America. An isolated and uncorrelated section in southwestern Saskatchewan includes till that is interpreted as being middle Pleistocene in age based on magnetic polarity. This interpretation is strengthened by the age of an underlying tephra and an underlying assemblage of mammal fossils (Barendregt et al., 1991). In Montana and Alberta,

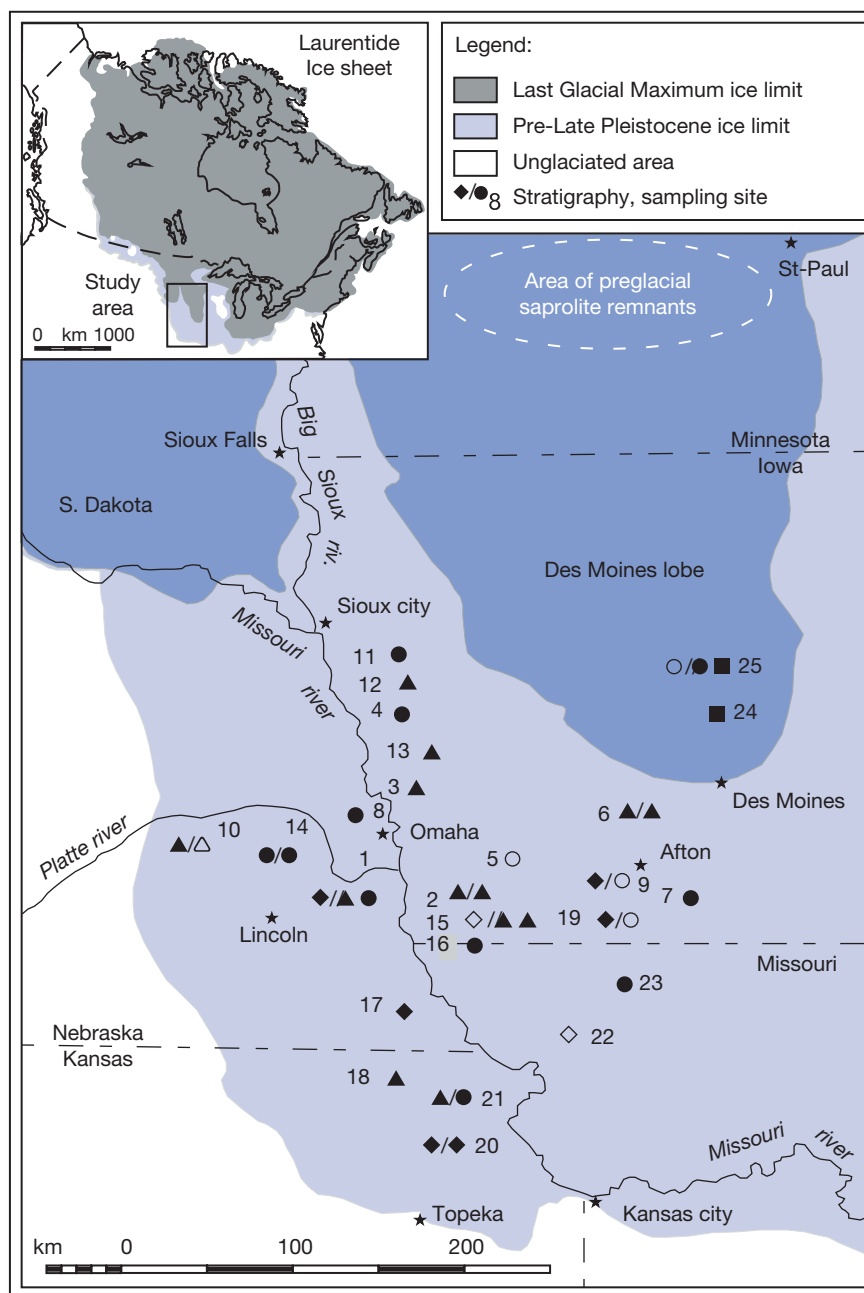


Figure 2 Location of the stratigraphic sections investigated in relation to the maximum ice limit of late Cenozoic glaciations. Symbols correspond to the till stratigraphy exposed at each site: solid diamonds, late Pliocene R2 tills (~ 2.7 – 1.3 My); solid circles, middle Pleistocene N tills (~ 0.8 – 0.25 My); solid squares, late Pleistocene LGM tills (~ 17 ka). Age constraints are provided by three volcanic ashes and the Brunhes–Matuyama magnetic reversal. Current chronological methods do not allow the dating of individual ice advances, and each till group represents multiple ice advances. Open symbols correspond to till units with uncertain magnetic polarity but of lithological composition similar to their corresponding till group category. Volcanic ash localities: sites 4, 8, 9, and 10. Reproduced from Roy M, Clark PU, Raisbeck GM, and Yiou F (2004) Geochemical constraints on the regolith hypothesis for the Middle Pleistocene transition. *Earth and Planetary Science Letters* 227: 281–296, with permission.

paleomagnetism was used to decide between two age estimates for the Kennedy Drift, composed of tills and paleosols. The inferred rates of soil development placed the paleosols within the middle Pleistocene (younger than 800 ka) or in the Pliocene (>2600 ka). Paleomagnetism demonstrated the presence of six glacial and six interglacial episodes that span that time, beginning in the Gauss Normal Chron and continuing through

the Matuyama Reversed Chron and ending during the Brunhes Normal Chron (Cioppa et al., 1995). In the Pacific Northwest, a middle Pleistocene flood deposit is recognized by its normal polarity and capping calcrete soil with a 200 to >400 ka Th/U date, whereas earlier flood deposits in the sequence had a reversed polarity (Bjornstad et al., 2001). On Banks Island in the Canadian Arctic Archipelago, the strata exposed in the



Figure 3 Proposed distribution and timing of Brunhes/middle Pleistocene ice sheets in northwest North America. Montane glaciations for northern and northwestern Alaska are not included. Reproduced from Barendregt RW and Duk-Rodkin A (2004) Chronology and extent of Late Cenozoic ice sheets in North America: A magnetostratigraphic assessment. In Ehlers J and Gibbard PL (eds.) *Quaternary Glaciations – Extent and Chronology, Part II: North America*, pp. 1–7. Amsterdam: Elsevier.

Nelson River Bluffs and nearby outcrops span the Pleistocene. They record the Brunhes–Matuyama boundary within an interglacial deposit and preserve the deposits of an unnamed middle Pleistocene glaciation (<780 ka), unnamed interglacials (one

or more) between 780 and 130 ka, and the previously recognized middle Pleistocene (Illinoian) Thomsen glaciation (>130 ka). The latter is constrained by the inferred Sangamonian age of the capping paleosol (Barendregt et al., 1998).

A long sequence in the Mackenzie Mountains in the Northwest Territories begins in the late Tertiary (Gauss-normal); has two reversed tills (Matuyama) and, as its upper units, three normal tills (Brunhes); and is further constrained by means of ^{36}Cl dating of exposed boulders and multiple paleosols (Duk-Rodkin et al., 1996, 2004).

The collective results of the magnetostratigraphic work in the west have been interpreted to support a reconstruction of paleo-ice centers and the number of middle Pleistocene glaciations related to these proposed centers: Cordilleran Ice Sheet, three glaciations; the Horton Ice Cap/Keewatin Ice Sheet, three glaciations. These ice centers are proposed in areas near the late Pleistocene, Laurentide ice centers but the ice sheets were apparently less extensive (Figure 3; Barendregt and Duk-Rodkin, 2004; Barendregt and Irving, 1998).

Other refinements in the dating of middle Pleistocene glaciations have been the result of the application of cosmogenic exposure dating. Exposed, striated quartzite lying beyond the Late Wisconsinan glacial limit in southwestern Minnesota revealed a complex burial history that indicated the striations were between 640 and 740 ka (MIS 16; Bierman et al., 1999). Stratigraphic interpretations do not have this region affected by glaciation after this time. In some places, outcrops that have been glaciated during Wisconsinan time preserve an inherited cosmogenic nuclide abundance that is more a signature of limited Late Wisconsinan erosion, owing perhaps to cold-based glaciation, than the true exposure age of the rock (Baffin Island, Bierman et al., 1999; Wisconsin, Colgan et al., 2002; Puget Lowland, Washington, Briner and Swanson, 1998; and Arctic Canada, Briner et al., 2003).

The use of the disequilibrium of two radioactive, cosmogenic nuclides that have decayed since the burial of previously exposed sediment can constrain the age of that surface (Balco et al., 2005). In southwestern Minnesota and eastern South Dakota, the minimum limiting age for tills previously known only as pre-Wisconsinan was determined to be >500 ka (Balco et al., 2005b). A core in eastern Nebraska revealed that the Loveland Loess (135–150 ka) (Forman and Pierson, 2002) overlay an unnamed loess with a best-fit burial age of MIS 14–16. This loess overlaid the youngest till in eastern Nebraska (Balco et al., 2005a,b), most likely of MIS 16. In general, paleosols with long exposure periods produced the best burial ages, and this technique is therefore more effective in the southern tier of states, where the time between glaciations was longer (Balco et al., 2005c).

See also: **Glaciations: Late Quaternary in North America; Overview.**

References

- Aber JS (1991) The glaciation of northeastern Kansas. *Boreas* 20: 297–314.
- Aber JS (1999) Pre-Illinoian glacial geomorphology and dynamics in the central United States, west of the Mississippi. In: Mickelson DM and Attig JW (eds.) *Glacial Processes, Past and Present*, pp. 113–119. Boulder, CO: Geological Society of America. Special Paper No. 337.
- Baker JL and Stewart RA (1984) Paleomagnetic study of glacial deposits at Conklin Quarry and other locations in southeast Iowa. In: Bunker BJ and Hallberg GR (eds.) *Underburden–Overburden, an Examination of Paleozoic and Quaternary Strata at the Conklin Quarry near Iowa City, Guidebook 41*, pp. 63–69. Iowa City: Iowa Geological Survey.
- Balco G, Stone JOH, and Mason J (2005a) Numerical ages for Plio–Pleistocene glacial sediment sequences by $^{26}\text{Al}/^{10}\text{Be}$ dating of quartz in buried paleosols. *Earth and Planetary Science Letters* 232: 179–191.
- Balco G, Stone JOH, and Jennings C (2005b) Dating Plio–Pleistocene glacial sediments using the cosmic-ray-produced radionuclides ^{10}Be and ^{26}Al . *American Journal of Science* 305: 1–41.
- Balco G, Rovey CW II, and Stone JOH (2005c) The first glacial maximum in North America. *Science* 307(5707): 222.
- Barendregt RW and Duk-Rodkin A (2004) Chronology and extent of Late Cenozoic ice sheets in North America: A magnetostratigraphic assessment. In: Ehlers J and Gibbard PL (eds.) *Quaternary Glaciations – Extent and Chronology, Part II: North America*, pp. 1–7. Amsterdam: Elsevier.
- Barendregt RW and Irving E (1998) Changes in the extent of North American ice sheets during the late Cenozoic. *Canadian Journal of Earth Sciences* 35: 504–509.
- Barendregt RW, Thomas FF, Irving E, et al. (1991) Stratigraphy and paleomagnetism of the Jaw Face section, Wellsch Valley site, Saskatchewan. *Canadian Journal of Earth Sciences* 28: 1353–1364.
- Barendregt RW, Vincent J-S, Irving E, and Baker J (1998) Magnetostratigraphy of Quaternary and late Tertiary sediments on Banks Island, Canadian Arctic Archipelago. *Canadian Journal of Earth Sciences* 35: 147–161.
- Bierman PR, Marsella KA, Patterson C, Davis PT, and Caffee M (1999) Mid-Pleistocene cosmogenic minimum-age limits for pre-Wisconsinan glacial surfaces in southwestern Minnesota and southern Baffin Island: A multiple nuclide approach. *Geomorphology* 27: 25–39 (comment and reply, *Geomorphology* 39(3–4): 255–261, 2001).
- Bjornstad BN, Focht KR, and Pluhar CJ (2001) Long history of pre-Wisconsinan, ice age cataclysmic floods: Evidence from southeastern Washington State. *Journal of Geology* 109: 695–713.
- Boellstorff J (1978) North American Pleistocene stages reconsidered in the light of probable Pliocene–Pleistocene continental glaciation. *Science* 202: 305–307.
- Borns HW, Doner LA, Dorion CC, et al. (2004) The deglaciation of Maine. In: Ehlers J and Gibbard PL (eds.) *Quaternary Glaciations – Extent and Chronology, Part II: North America*, pp. 89–109. Amsterdam: Elsevier.
- Braun DD (2004) The glaciation of Pennsylvania, USA. In: Ehlers J and Gibbard PL (eds.) *Quaternary Glaciations – Extent and Chronology, Part II: North America*, pp. 237–242. Amsterdam: Elsevier.
- Briner JP, Miller GH, Davis PT, Bierman PR, and Caffee M (2003) Last Glacial Maximum ice sheet dynamics in Arctic Canada inferred from young erratics perched on ancient tors. *Quaternary Science Reviews* 22: 437–444.
- Briner JP and Swanson TW (1998) Using inherited cosmogenic ^{36}Cl to constrain glacial erosion rates of the Cordilleran ice sheet. *Geology* 26(1): 3–6.
- Cadwell DH and Muller EH (2004) New York glacial geology, USA. In: Ehlers J and Gibbard PL (eds.) *Quaternary Glaciations – Extent and Chronology, Part II: North America*, pp. 201–206. Amsterdam: Elsevier.
- Cioppa MT, Karlstrom ET, Irving E, and Barendregt RW (1995) Paleomagnetism of tills and associated paleosols in southwestern Alberta and northern Montana: Evidence for Late Pliocene–Early Pleistocene glaciations. *Canadian Journal of Earth Sciences* 32: 555–564.
- Colgan PM (1999) Early Middle Pleistocene glacial sediments (780,000–620,000 BP) near Kansas City, northeastern Kansas and northwestern Missouri, USA. *Boreas* 28(4): 477–489.
- Colgan PM, Bierman PR, Mickelson DM, and Caffee M (2002) Variation in glacial erosion near the southern margin of the Laurentide Ice Sheet, south-central Wisconsin, USA: Implications for cosmogenic dating of glacial terrains. *Geological Society of America Bulletin* 114(12): 1581–1591.
- Dahms DE (2004) Glacial limits in the middle and southern Rocky Mountains, USA, south of Yellowstone Ice Cap. In: Ehlers J and Gibbard PL (eds.) *Quaternary Glaciations – Extent and Chronology, Part II: North America*, pp. 275–288. Amsterdam: Elsevier.
- Duk-Rodkin A, Barendregt RW, Froese DG, et al. (2004) Timing and extent of Plio–Pleistocene glaciations in north-western Canada and east-central Alaska. In: Ehlers J and Gibbard PL (eds.) *Quaternary Glaciations – Extent and Chronology, Part II: North America*, pp. 347–362. Amsterdam: Elsevier.
- Duk-Rodkin A, Barendregt RW, Tarnocai C, and Phillips FM (1996) Late Tertiary to late Quaternary record in the Mackenzie Mountains, Northwest Territories, Canada: Stratigraphy, paleosols, paleomagnetism, and chlorine-36. *Canadian Journal of Earth Sciences* 33: 875–895.
- Farrand WR, Mickelson DM, Cowan RW, and Goebel JE (1984) Quaternary geologic map of the Lake Superior 4° × 6° Quadrangle, United States and Canada. In: Richmond GM and Fullerton DS (eds.) *Quaternary Geologic Atlas of the United States, Misc. Invest. Series Map I-1420 (NL-16)*. Reston, VA: US Geological Survey.

- Flint RF (1957) *Glacial and Pleistocene Geology*. New York: Wiley.
- Forman SL and Pierson J (2002) Late Pleistocene luminescence chronology of loess deposition in the Missouri and Mississippi River valleys, United States. *Palaeogeography, Palaeoclimatology, Palaeoecology* 186: 25–46.
- Fullerton DS, Colton RB, and Bush CA (2004) Limits of mountain and continental glaciations east of the Continental Divide in northern Montana and north-western North Dakota, USA. In: Ehlers J and Gibbard PL (eds.) *Quaternary Glaciations – Extent and Chronology, Part II: North America*, pp. 131–150. Amsterdam: Elsevier.
- Gibbons AB, Pierce KL, and Megeath JD (1984) Probability of moraine survival in a succession of glacial advances. *Geology* 12: 327–330.
- Gillespie AR and Zehfuss PH (2004) Glaciations of the Sierra Nevada, California, USA. In: Ehlers J and Gibbard PL (eds.) *Quaternary Glaciations – Extent and Chronology, Part II: North America*, pp. 51–62. Amsterdam: Elsevier.
- Gradstein FM, Ogg JG, Smith AG, et al. (2004) *~500A Geologic Time Scale 2004*.
- Hall RD (1999) Effects of climate change on soils in glacial deposits, Wind River Basin, Wyoming. *Quaternary Research* 51: 248–261.
- Hallberg GR (1980a) *Illinoian and Pre-Illinoian Stratigraphy of Southeast Iowa and Adjacent Illinois*. Iowa City: Iowa Geological Survey Technical Information Series No. 11.
- Hallberg GR (1980b) *Pleistocene Stratigraphy in East-Central Iowa*. Iowa City: Iowa Geological Survey Technical Information Series No. 10.
- Hamilton TD (2001) Quaternary glacial, lacustrine, and fluvial interactions in the western Noatak basin, northwest Alaska. *Quaternary Science Reviews* 20: 371–391.
- Hays JD, Imbrie J, and Shackleton NJ (1976) Variations in the Earth's orbit: Pacemaker of the ice ages. *Science* 194: 1121–1132.
- Imbrie J, Berger A, Boyle EA, et al. (1993) On the structure and origin of major glaciations cycles: 2. The 100,000-year cycle. *Paleoceanography* 8: 699–735.
- Imbrie J, Hays JD, Martinson DG, et al. (1984) The orbital theory of Pleistocene climate: Support for a revised chronology of marine $\delta^{18}\text{O}$ record. In: Berger AL, Hays J, Kukla G, and Salzman B (eds.) *Milankovitch and Climate, Part 1*, pp. 269–305. Norwell, Dordrecht: Reidel.
- Izett GA (1981) Volcanic ash beds: Recorders of Upper Cenozoic silicic pyroclastic volcanism in the western United States. *Journal of Geophysical Research* 86: 10200–10222.
- Koteff C and Pessl F Jr. (1985) Till stratigraphy in New Hampshire: Correlations with adjacent New England and Quebec. In: Borns HW Jr., LaSalle P, and Thompson WB (eds.) *Late Pleistocene History of Northeastern New England and Adjacent Quebec*, pp. 1–12. Boulder, CO: Geological Society of America. Special Paper No. 197.
- Lisiecki LE and Raymo ME (2005) A Pliocene–Pleistocene stack of 57 globally distributed benthic $\delta^{18}\text{O}$ records. *Paleoceanography* 20: .
- Manley WF, Kaufmann DS, and Briner JP (2001) Pleistocene glacial history of the southern Ahklun Mountains, southwestern Alaska: Soil-development, morphometric, and radiocarbon constraints. *Quaternary Science Reviews* 20: 353–370.
- Manz LA (2005) Cosmogenic ^{36}Cl dating of Quaternary glacial deposits in southwestern North Dakota. *Abstracts with Programs, North-Central Section, Geological Society of America Meeting* 37(5): 91.
- Marsella K, Bierman PR, Davis T, and Caffee M (2000) Deglacial dynamics and timing, Pangnirtung Fjord and Kolic Valley, Baffin Island, Canada. *Geological Society of America Bulletin* 112: 1296–1312.
- Meyer GN (1997) Pre-Late Wisconsinian till stratigraphy of north-central Minnesota. *Minnesota Geological Survey Report of Investigations* 48: 67.
- Meyer GN (2005) The location of the Labrador and Keewatin ice centers shifted from one glaciation to the next. *Abstracts with Programs, North-Central Section, Geological Society of America Meeting* 37(5): 91.
- Mix AC, Pisias NG, Rugh W, et al. (1995) Benthic foraminifer stable isotope record from site 849 (0–5 Ma): Local and global climate changes. *Proceedings of the Ocean Drilling Program, Scientific Results* 138: 371–1342.
- Pillans B and Naish T (2004) Defining the Quaternary. *Quaternary Science Reviews* 23: 2271–2282.
- Pisias NG and Moore TC Jr. (1981) The evolution of the Pleistocene climate: A time series approach. *Earth and Planetary Science Letters* 52: 450–458.
- Raymo ME, Ruddiman WF, Backman J, Clement BM, and Martinson DG (1989) Late Pliocene variation in Northern Hemisphere ice sheets and North Atlantic Deep Water circulation. *Paleoceanography* 4: 413–446.
- Renne P, Deino A, Walter R, et al. (1994) Intercalibration of astronomical and radioisotopic time. *Geology* 22: 783–786.
- Rovey CW II, Bettis EA III, and Kean WF (2002) Position of the Matuyama–Brunhes paleomagnetic datum within the pre-Illinoian Alburnett Formation, eastern Iowa, USA. *GSA Abstracts with Programs* 34: .
- Rovey CW II and Kean WF (1996) Pre-Illinoian glacial stratigraphy in north-central Missouri. *Quaternary Research* 45: 17–29.
- Roy M, Clark PU, Barendregt RW, Glasman JR, and Enkin RJ (2004) Glacial stratigraphy and paleomagnetism of late Cenozoic deposits of the north-central United States. *Geological Society of America Bulletin* 226(1–2): 30–41.
- Roy M, Clark PU, Raisbeck GM, and Yiou F (2004) Geochemical constraints on the regolith hypothesis for the Middle Pleistocene transition. *Earth and Planetary Science Letters* 227: 281–296.
- Ruddiman WF, Raymo ME, Martinson DG, Clement BM, and Backman J (1989) Pleistocene evolution: Northern Hemisphere ice sheets and North Atlantic Ocean. *Paleoceanography* 4: 353–412.
- Ruhe RV (1969) *Quaternary Landscapes in Iowa*. Ames: Iowa State University Press.
- Shackleton NJ, Berger A, and Peltier W (1990) An alternative astronomical calibration of the lower Pleistocene timescale based on ODP Site 677. *Transactions of the Royal Society of Edinburgh: Earth Sciences* 81: 25–261.
- Shackleton NJ, Crowhurst S, Hagelberg T, Pisias NG, and Schneider DA (1995) A new late Neogene time scale: Application to Leg 138 sites. *Proceedings of Ocean Drilling Program, Scientific Results* 138: 73–101.
- Soller DR (2004) Thickness and character of Quaternary sediments in the glaciated United States east of the Rocky Mountains. In: Ehlers J and Gibbard PL (eds.) *Quaternary Glaciations – Extent and Chronology, Part II: North America*, pp. 363–372. Amsterdam: Elsevier.
- Spell T and McDougall I (1992) Revisions to the age of the Brunhes–Matuyama boundary and the Pleistocene geomagnetic polarity time scale. *Geophysics Research Letters* 19: 1181–1184.
- Stea RR (2004) The Appalachian glacier complex in Maritime Canada. In: Ehlers J and Gibbard PL (eds.) *Quaternary Glaciations – Extent and Chronology, Part II: North America*, pp. 213–222. Amsterdam: Elsevier.
- Stone B, Stanford SD, and Witte RW (2002) *Surficial geologic map of northern New Jersey (1:100,000 scale)*. Boulder, CO: US Geological Survey Miscellaneous Investigation Series, Mpa, I-2540-C, 3 sheets and 1 pamphlet, pp. 41.
- Szabo JP and Chanda A (2004) Pleistocene glaciation of Ohio, USA. In: Ehlers J and Gibbard PL (eds.) *Quaternary Glaciations – Extent and Chronology, Part II: North America*, pp. 233–236. Amsterdam: Elsevier.
- Willman HB and Frye JC (1970) Pleistocene stratigraphy of Illinois. *Illinois State Geological Survey Bulletin* 94: 204.

Middle Pleistocene Glaciations in the Southern Hemisphere

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Introduction

Middle Pleistocene glaciations are understood to be one of the consequences of climatic cooling forced by solar radiation variation during the 100-ka period cycles driven by Earth's orbit eccentricity.

The marine isotope stage (MIS) curves reflect high frequency and intensive periods of low $\delta^{18}\text{O}$ since 0.88 Ma (MIS 22) and throughout the subsequent Brunhes Normal magnetic polarity chron. According to Ehlers and Gibbard (2007), MIS 22 is the first of the colder events in which critical $\delta^{18}\text{O}$ values ($\sim 5.55\text{‰}$) were reached, which allowed continental ice cap formation in mid-latitudes, thus defining the critical conditions during the following middle Pleistocene times. This time spans the period between 780 ka, when the magnetic polarity changed from reverse to normal, and ~ 130 ka, when the last interglacial period began.

Although there are fewer land masses in the Southern Hemisphere in comparison to the Northern Hemisphere, evidence for the middle Pleistocene glaciations has been recognized, mostly through geomorphological mapping and cosmogenic exposure dating. The latter uses ^{10}Be , ^{26}Al , and ^{36}Cl , as well as ^{39}Ar – ^{40}Ar incremental heating techniques on basaltic rocks related to till units. The dates obtained have been correlated with the $\delta^{18}\text{O}$ MIS or with Antarctic temperature curves, depending on the author's decisions. This work shows how the cosmogenic and ^{39}Ar – ^{40}Ar dates presented herein correspond to the cold periods in the MIS curve produced by Lisiecki and Raymo (2005), in which the benthic $\delta^{18}\text{O}$ records of 57 globally distributed sites were averaged.

Knowledge of these glaciations has been improved by recent work in the Central Andes (Bolivia and Peru), as well as on the eastern slope of the Southern Andes (Patagonia, Argentina), in the New Zealand Alps, and in the western ranges of Tasmania. As this contribution is an updated review of Coronato and Rabassa (2007), only post-2005 findings and the literature not previously considered is presented herein. For previous work and other references, the reader should consult the authors' article in the first edition of the *Encyclopedia of Quaternary Science*.

South America

Intensive research has been performed during the last few years in the Central Andes in Bolivia and Peru, mainly on the basis of geomorphological surveying using digital terrain models (DTM), accompanied by satellite imagery and chronological studies using cosmogenic exposure dating techniques on moraine boulders.

Pre-Last Glacial Maximum (pre-LGM) moraines have been identified in the Cojup and Queshgue valleys in Cordillera

Blanca (9° – 10° S), Perú (Farber et al., 2005). Assuming that there has been no subsequent erosion of the boulder surfaces, ^{10}Be exposure ages of ~ 120 , 225, and 440 ka were obtained. These dates roughly correlate with MIS 4, the end of the MIS 8, and MIS 12 cooler periods in the ocean sequences (Figure 1), respectively. As a consequence of the poor constraint of the boulder surface erosion rate, the authors consider that these ages should be regarded as only tentative at this stage. Nevertheless, geomorphological relationships confirm that moraines carrying these boulders predate the LGM as well as the last interglacial period. Consequently, they could be considered as of middle Pleistocene age. Farber et al. (2005) suggested that maximum tropical alpine ice volume occurred during the middle Pleistocene at MIS 12 or later. The ice was considerably less extensive during the LGM.

In the Central Peruvian Andes, ^{10}Be and ^{26}Al cosmogenic radionuclide dating was undertaken in order to understand the pre-LGM glacial chronology. Here, Smith et al. (2005) obtained a set of dates from over a hundred boulders on moraines developed in valleys bordering the Junín Plains. The resulting chronology spans multiple glacial cycles, including boulder ages exceeding 1 Ma. This suggests that boulder erosion has been very low through this period. Two major groups of ages, within the set of no erosion/no uplift exposure ages on 10 dated moraines, were distinguished in the ranges ~ 440 – 325 and ~ 280 – 50 ka. Although it is not possible to establish clear morphological and chronological relationships, the data obtained suggest that middle Pleistocene glacial advances occurred repeatedly in the Junín Plain during MIS 12 at the end of MIS 10 and during MIS 6. This compares well with the global MIS (cf. Lisiecki and Raymo, 2005) sequence (Figure 1).

The history of middle and late Pleistocene regional glaciations has been interpreted from limnological studies in Lake Titicaca in the Bolivian Highlands (Altiplano). Lake-level variations were interpreted from magnetic susceptibility, total organic carbon (TOC), total inorganic carbon (TIC), calcium carbonate content, benthic and saline diatom assemblages, and $\delta^{13}\text{C}$ analyses, assuming that lake high-stands represent glaciations in the tropical Andes triggered by a combination of low temperatures and higher precipitation (Fritz et al., 2007). Thus, the lower part of the sequence was interpreted by these authors to indicate that regional glaciations could have expanded during 370–324, 300–238, 230–213, and 188–132 ka BP, perhaps during the colder events MIS 10, 8, and 6 (Figure 1). The proposed ages were obtained from an age model based on bulk organic carbon samples distributed into the acceptable confidence range (Hughen et al., 2004). The record of Lake Titicaca fresher and deeper conditions offer a middle Pleistocene paleoclimatic reference for geomorphological and cosmogenic studies in the glaciated valleys of the tropical Andes from which superficial absolute ages could potentially be controlled.

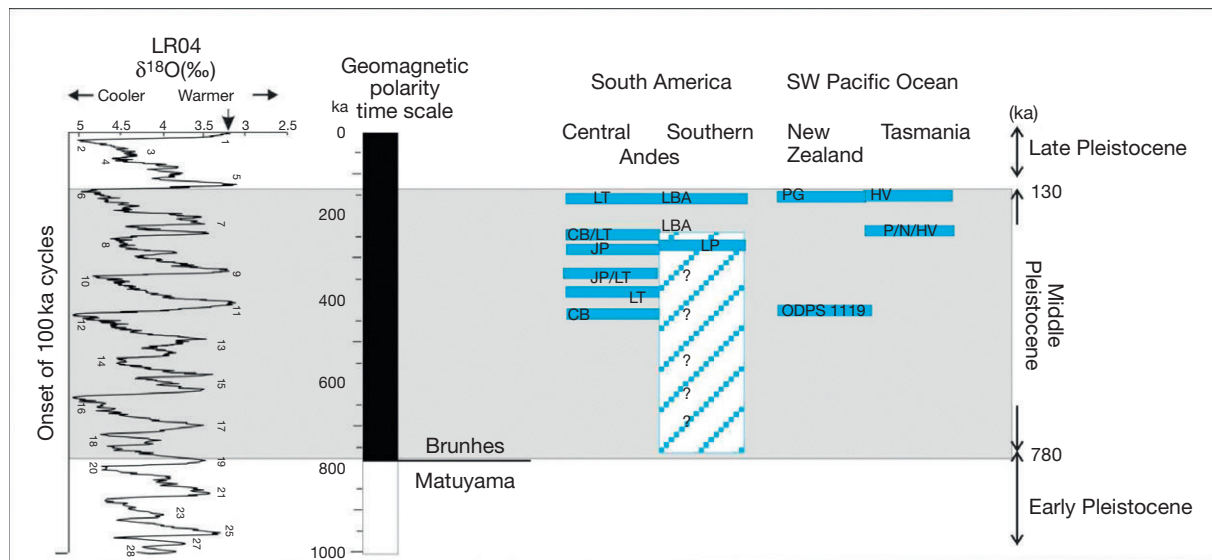


Figure 1 Middle Pleistocene glaciations recognized in the Central and Southern Andes, South America, New Zealand, and Tasmania, from the works cited in this article. The solid light blue lines indicate the absolute ages of moraines, mainly obtained by cosmogenic radionuclide determinations. The dashed light blue lines indicate the age of lava flows constraining the pre-Last Glacial Maximum sequence of six moraine belts. Lake Titicaca and ODPS 1119 provide sedimentary evidence of cold stages from core analyses, while the other localities give geomorphological glacial evidence. Core ODPS 1119 is located at 44° S–171° E, its paleoclimatic results derived from natural gamma-ray emission records that represent clay content variations corresponding to changes in the supply rate of glacial rock flour from the Pleistocene ice cap located immediately to the west, in the South Island of New Zealand (Carter and Gammon, 2004). CB, Cordillera Blanca (Peru); LT, Lago Titicaca (Bolivia); JP, Junín Plain (Peru); LBA, Lago Buenos Aires (Argentina); LP, Lago Pueyrredón (Argentina); PG, Poplar Gully in the Hope Valley (New Zealand); ODPS 1119, Ocean Drilling Project Site, in the NW Pacific Ocean; HV, Hesperus valley (Tasmania); P/N/HV, Pluto, Neptuno and Hayes valleys (Tasmania).

The glacial chronostratigraphy, formerly established for the Central Andes at latitude 36° in the Río Mendoza upper basin (eastern slope of the Andes, Argentina), has been recently rejected by Fauqué et al. (2009). Strong field and laboratory evidence supports a new interpretation for glacial deposits previously known as the Punta de Vacas, Penitentes, Horcones, Almacenes, and Confluencia drifts (cf. Coronato and Rabassa, 2007, for details). Most of these deposits were interpreted as mass-wasting deposits by Fauqué et al. (2009), who demonstrated that debris flow and mega-landslides, including true glacial boulders, occurred several times along the middle and late Pleistocene. These gravity flow deposits overlaid previous glacial deposits and reshaped them as mass-wasting accumulation landforms at the bottom of the Horcones and Cuevas valleys. Glacial deposits at 100–150 m above the valley bottom represent the middle Pleistocene Penitentes drift which was unaffected by the debris flow, but the lower glacial deposits along the main valleys would have thus been reworked. Fauqué et al. (2009) argued, on the basis of ^{36}Cl cosmogenic exposure dating and radiocarbon ages, that mass-wasting processes had a stronger impact on the landscape modeling than did the middle to late Pleistocene glaciers.

Comprehensive reviews of middle Pleistocene glaciations in Patagonia, between 39° S and 46° S are given by Glasser et al. (2008), Rabassa (2008), and Rabassa and Coronato (2009). In addition, Coronato and Rabassa (2011) and Martínez et al. (2011) have offered detailed information for most of the transverse glacial valleys of the Patagonian Andes.

The Lago Buenos Aires (46° 30' S) early to late Pleistocene glacial sequence (Kaplan et al., 2005) is so far the most

completely dated sequence. ^{10}Be , ^{26}Al , and ^{36}Cl exposure surface dating of morainic boulders as well as $^{40}\text{Ar}/^{39}\text{Ar}$ dating of basaltic rocks associated with till units or glacial landforms were applied. A detailed chronostratigraphy is still required for the six middle Pleistocene morainic belts recognized. The Moreno and Deseado moraines, deposited during ice expansion onto the Patagonian plains, were dated to between 760 and 200 ka. This means not only that they predate the LGM but also that they postdate the greatest Patagonian glaciation (GGP), the latter occurring at 1.1–1.0 Ma (cf. Coronato et al., 2004; Rabassa, 2008, for details). The maximum age of these moraines is given by $^{40}\text{Ar}/^{39}\text{Ar}$ dating of the Arroyo Page volcanic rocks, whereas their minimum age is given by cosmogenic exposure ages of boulders on the Moreno moraines. This broad period of time roughly correlates with the MIS 8–18 intervals. Between 150 and 140 ka, another expansion of glacial ice occurred forming the younger Moreno moraines, during MIS 6. As the moraines corresponding to MIS 4 have not yet been found, it was assumed by Kaplan et al. (2005) that they would have probably been overrun and removed by the ice during subsequent LGM times. Likewise, MIS 8 moraines (if they existed) may have been obliterated during the MIS 6 glacial advance.

The age of the last middle Pleistocene glaciations, proposed by Kaplan et al. (2005), was constrained by Phillips et al. (2006) using $^{230}\text{Th}/^{234}\text{U}$ analyses on soil carbonate formed in outwash gravel plains, correlated with the Moreno II terminal moraines. The results indicated that calcic pedogenesis began between 170 ± 8.3 and 146 ± 4.8 ka (MIS 6), with a subsequent period of pedogenesis constrained between

80.3 ± 1.8 and 38.9 ± 1.9 ka. This broadly corresponds to the MIS 4 cold period.

To the south, in the Lago Pueyrredón valley ($47^\circ 30' \text{ S}$), ^{10}Be cosmogenic surface exposure dating was performed on cobbles from old outwash terraces by Hein et al. (2009). Three pre-LGM moraine limits and related outwash plains were mapped by the authors and named, mostly following Caldenius (1932), from younger to older; the Hatcher, Caracoles, and Gorra de Poivre glaciations. On that basis and according to previous geomorphological and paleomagnetic studies, Hein et al. (2009) proposed a mean age of 260 ka (MIS 8) for the Hatcher glaciation. This interpretation does not correlate with its geomorphological equivalent glacial limit at Lago Buenos Aires where moraines in a similar topographic position with respect to the LGM limits were dated close to MIS 6 (Kaplan et al., 2005). In contrast to the Lago Buenos Aires sequence, glacial deposits related to cold events during MIS 6 and 4 were not found in the Lago Pueyrredón valley. This might result from erosion during the later LGM when the glaciers were confined to the valley at the same topographic position as during previous glaciations but lower than during the MIS 8 glaciation.

In view of the fact that the number of middle Pleistocene glacial limits currently identified in Patagonian valleys varies from two to three between the Great Patagonian Glaciation (GPG) and the LGM, as noted by Coronato et al. (2004), and in order to favor regional correlations, these middle Pleistocene events were referred to as post-GGP one to three glaciations.

Southern Pacific Ocean Islands

Based on detailed stratigraphy, depositional models, and optical dating chronology, pre-LGM glacial deposits have been identified in the Hope Valley (42° S) on the South Island of New Zealand (Rother et al., 2010). As these glaciations occurred in a maritime perhumid climate, where runoff erosion is exacerbated in the mountains, it is quite difficult to find deposits older than those belonging to the LGM and the MIS 5 interglacial period. Nevertheless, the Popplar Gully sedimentary sequence in the Hope Valley shows that the LGM glaciers extended down valley but nested into the main valleys, without eroding former upslope glacial deposits. Rother et al. (2010) demonstrated that the majority of the local valley infillings were not deposited during the LGM, but on the contrary, during MIS 5, 4, and 3 based on thermoluminescence (TL) dates of $95.7\text{--}32.1$ ka. These authors suggested that the deposits were laid down during periods of MIS 5–3 and the colder period MIS 4. Glacial sediments of MIS 6 or even earlier were deposited on the upper slopes of the Hope Valley and throughout the valley. They were overridden and reshaped during the last glacial advance. As the late middle Pleistocene glacial deposits identified are more voluminous than those of the LGM cycle, it is thought that the MIS 6 glaciers excavated larger and deeper valleys than during the younger glacial advances. Moreover, the aggradation surface during MIS 6 was calculated at least 120 m above the LGM surface.

New *in situ* cosmogenic ^{10}Be and ^{26}Al exposure ages were recently reported by Kiernan et al. (2010) from moraine boulders in the Tasmanian western ranges. Several of those

moraines were previously interpreted as of LGM in age. On the contrary, Kiernan et al. (2010) presented evidence that they were deposited during middle–late Pleistocene glaciations. They also confirmed that glaciers at that time were significantly larger than during the LGM. The MIS 6 glaciation was interpreted from the average ^{10}Be erosion-corrected cosmogenic exposure ages of 149 ± 14 and 127 ± 12 ka in the Hesperus glacier valley (Western Arthur range, 42° S). Similar ages were obtained in the nearby valley of Neptune glacier. The MIS 8–10 glaciations were interpreted as being represented by lateral and interlobate moraines developed in the Pluto and Neptuno–Hayes glacier valleys based upon ^{10}Be erosion-corrected cosmogenic exposure average ages of 220 ± 12 , 232 ± 13 , and 220 ± 12 ka. Figure 1 shows that the proposed ages for the Hesperus valley fully match those from the end of MIS 6, but the age of the interlobate moraines of Pluto and Neptuno/Hayes fall into the end of MIS 8, instead of MIS 10. They are very close to the beginning of the warm trend represented by MIS 7.

Although the association of moraine formation, based on boulder exposure ages, and MIS cold events is not completely definitive, the close match between boulder ages and the $\delta^{18}\text{O}$ scale suggests that glaciers advanced down to elevations of $\sim 750\text{--}320$ m a.s.l. in the western Tasmanian mountains. The ages and geometry of moraines indicate that successive glaciations of the Western Arthur Range have become progressively less extensive through the middle Pleistocene cycles.

Severe cooling during the middle Pleistocene has been recorded in the ODP Site 1119 sediment core (Southwestern Pacific) at ~ 430 ka, that is, MIS 12 (Carter and Gammon, 2004). A clear relationship on millennial-scale climatic variations between Antarctica, the mid-latitude New Zealand Southern Alps, and the oceanic oxygen isotopic record during the past 370 ka has been proposed by these authors.

Conclusion

Significant advances in knowledge of the middle Pleistocene glaciations in South America, New Zealand, and Tasmania have been achieved during recent years. Geomorphological mapping, mostly based on DTM tools, as well as cosmogenic radionuclide dating, was undertaken in several regions and then correlated with limnological studies from the Andean lakes, with marine $\delta^{18}\text{O}$ isotope curves, as well as with Antarctic temperature curves. In Patagonia, volcanic sequences interbedded with lava flows have proved useful for constraining the glacial chronologies with the use of $^{39}\text{Ar}\text{--}^{40}\text{Ar}$ dating techniques. The new radionuclide cosmogenic dating techniques applied by Hein et al. (2009) on cobbles from pre-LGM outwash plains, instead of eroded boulders, offer huge opportunities for the establishment of chronologies in those areas where surface boulders, lava flows, or organic matter are unavailable. Concentration depth profiles supported the cosmogenic ages and suggest erosion rates on the terraces, as well as the moraine degradation rates when the results are compared with surface boulders age.

Kaplan et al. (2007) have questioned the suitability of cosmogenic nuclide dating as a chronometer for pre-LGM deposits in southern South America and Tierra del Fuego. They have suggested that this dating technique is not so effective in

the Lago Buenos Aires region. As most of the exposure surface ages obtained from boulders on middle Pleistocene moraines gave LGM ages, they hypothesized that higher rates of exhumation, boulder weathering, and landscape change had occurred further to the south during the Quaternary. This was perhaps influenced by the more oceanic and subantarctic climates. The effect of the hyperhumid climate-derived erosion in the mountains of New Zealand has also been used to explain the relative scarcity of pre-LGM glacial features. It seems that aridity in the tropical and central Andes, and on the eastern slope of the Southern Andes, has helped to preserve glacial landforms and to restrict chemical weathering on the exposed glacial boulders. This explains the suitability of the cosmogenic radionuclide techniques for constraining middle Pleistocene chronologies in those particular environments.

During the last 0.8 Ma, glacial advances have been triggered in the Southern Hemisphere in the high mountains and close to present sea level in subantarctic latitudes. Although the full correspondence between glacial advances during the oldest middle Pleistocene glaciations is not yet clear, glacial evidence from MIS 12 to 4 is clearly identified in several glaciated valleys of South America and the Southern Pacific Islands. Although many similar cosmogenic ages were found for glacial sequences in Tasmania and Lago Buenos Aires, according to Rother et al. (2010), the resulting depositional glacial sequences in the maritime-climate New Zealand Alps may differ from those of drier, higher latitudes and higher elevation glacier systems in the Southern Hemisphere. A more complete dataset is still required to compare those environmentally similar mountain regions in the Southern Hemisphere. Moreover, the evidence available indicates that the middle Pleistocene glaciations occurred in phase with those in the Northern Hemisphere during the period dominated by the 100 ka orbital cyclicity period.

See also: **Glaciations:** Late Pleistocene in South America; Late Quaternary of the Southwest Pacific Region; Overview.

References

- Caldenius C (1932) Las Glaciaciones Cuaternarias en Patagonia y Tierra del Fuego. Ministerio de Agricultura de la Nación. *Dirección General de Minas y Geología* 95: 1–148. Buenos Aires.
- Carter R and Gammon P (2004) New Zealand maritime glaciation: Millennial-scale southern climate change since 3.9 Ma. *Science* 304: 1659–1662.
- Coronato A, Martínez O, and Rabassa J (2004) Pleistocene glaciations in Argentine Patagonia, South America. In: Ehlers J and Gibbard PL (eds.) *Quaternary Glaciations – Extent and Chronology. Part III*, pp. 49–67. Amsterdam: Elsevier.
- Coronato A and Rabassa J (2007) Mid-Quaternary glaciations in the Southern Hemisphere. In: Elias S (ed.) *Encyclopedia of Quaternary Science*, vol. 2, pp. 1051–1056. Amsterdam: Elsevier.
- Coronato A and Rabassa J (2011) Pleistocene glaciations in Southern Patagonia and Tierra del Fuego. In: Ehlers J, Gibbard PL and Hughes PD (eds.) *Extent and Chronology of Quaternary Glaciations – a Closer Look. Developments in Quaternary Science*, 15. Chapter 51: 715–727. Amsterdam: Elsevier.
- Ehlers J and Gibbard P (2007) The extent and chronology of Cenozoic Global Glaciation. *Quaternary International* 164–165: 6–20.
- Farber D, Hancock G, Finkel R, and Rodbell T (2005) The age and extent of tropical alpine glaciation in the Cordillera Blanca, Perú. *Journal of Quaternary Science* 20(7–8): 759–776.
- Fauqué L, Hermanns R, Hewitt K, et al. (2009) Mega-deslizamientos de la pared sur del Cerro Aconcagua y su relación con depósitos asignados a la Glaciación Pleistocena. *Revista de la Asociación Geológica Argentina* 65(4): 691–712.
- Fritz SC, Baker P, Seltzer G, et al. (2007) Quaternary glaciation and hydrologic variation in South American tropics as reconstructed from the Lake Titicaca drilling project. *Quaternary Research* 68: 410–420.
- Glasser N, Jansson K, Harrison S, and Kleman J (2008) The glacial geomorphology and Pleistocene history of South America between 38° S and 56° S. *Quaternary Science Reviews* 27: 365–390.
- Hein A, Hulton N, Dunai T, et al. (2009) Middle Pleistocene glaciations in Patagonia dated by cosmogenic-nuclide measurements on gravels. *Earth and Planetary Science Letters* 286: 184–197.
- Hughen K, Lehman J, Overpeck J, Marchal O, Herring C, and Turnbull J (2004) Carbon-14 activity and global carbon cycle changes over the past 50,000 yr. *Science* 303: 202–207.
- Kaplan M, Coronato A, Hulton N, Rabassa J, Kubik P, and Freeman S (2007) Cosmogenic nuclide measurements in southernmost South America and implications for landscape change. *Geomorphology* 87: 284–301.
- Kaplan M, Douglass D, Singer B, Ackert R, and Caffee M (2005) Cosmogenic nuclide chronology pre-last glacial maximum moraines at Lago Buenos Aires, 46 °S, Argentina. *Quaternary Research* 63: 301–315.
- Kiernan K, Fink D, Greig D, and Mifud C (2010) Cosmogenic radionuclide chronology of the pre-last glacial cycle moraines in the western Arthur range, Southwest Tasmania. *Quaternary Science Reviews* 29: 3286–3297.
- Lisiecki L and Raymo M (2005) A Pliocene-Pleistocene stack of 57 globally distributed benthic delta 18O records. *Paleoceanography* 20: PA 1003 1–17.
- Martínez O, Coronato A, and Rabassa J (2011) Pleistocene glaciations in Northern Patagonia, Argentina: An updated review. In: Ehlers J, Gibbard PL and Hughes PD (eds.) *Quaternary Glaciations – Extent and Chronology – A Closer Look. Developments in Quaternary Science*, 15, Chapter 52 :729–734. Amsterdam: Elsevier.
- Phillips R, Sharp W, Singer B, and Douglass D (2006) Utilising U-Series disequilibrium of calcic soils to constrain the surface age of Quaternary deposits: A comparison with ¹⁰Be, ²⁶Al age data from Patagonian glacial moraines. *Geochimica et Cosmochimica Acta* 70(18): A491.
- Rabassa J (2008) Late Cenozoic glaciations in Patagonia and Tierra del Fuego. In: Rabassa J (ed.) *Late Cenozoic of Patagonia and Tierra del Fuego. Developments in Quaternary Science* 11: pp. 151–204. Amsterdam: Elsevier.
- Rabassa J and Coronato A (2009) Glaciations in Patagonia and Tierra del Fuego during the Ensenadan Stage (early Pleistocene – Earliest middle Pleistocene). *Quaternary International* 210: 18–36.
- Rother H, Shulmeister J, and Rieser U (2010) Stratigraphy, optical dating chronology (ISRL) and depositional model of pre-LGM glacial deposits in the Hope Valley, New Zealand. *Quaternary Science Reviews* 29: 576–592.
- Smith J, Finkel R, Farber D, Rodbell D, and Seltzer G (2005) Moraine preservation and boulder erosion in the tropical Andes: Interpreting old surface exposure ages in glaciated valleys. *Journal of Quaternary Science* 20(7–8): 735–738.

Late Pleistocene Glacial Events in Beringia

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Late Pleistocene glaciations in Beringia included rapid expansion of valley glaciers directly after the last interglacial interval (marine isotope stage (MIS) 5e). Glaciation during MIS 4 is referred to as the Early Wisconsin in North America and the Zyryan in northeastern Russia. With the exception of the Alaska Range and Chugach/St. Elias Mountains, the MIS 4 glacial ice caps and mountain glacier complexes throughout much of Beringia were considerably more extensive than during the subsequent Late Wisconsin/Sartan glaciation in MIS 2.

Introduction

The region of the western Arctic stretching from the Lena River in northeast Russia to the Mackenzie River in Canada is geographically known as Beringia. From a glaciological standpoint, Beringia, including the vast continental shelves of the Bering and Chukchi Sea, has presented a great anomaly throughout the Cenozoic. While nearly all other high-latitude regions of the Northern Hemisphere were covered by ice sheets during each glaciation, the lowlands of Beringia remained free of ice, thus providing a refuge for high-latitude flora and fauna. Owing to the buildup of continental-sized ice sheets over mountain ranges in North America, Greenland, parts of Eurasia, and other regional mountain ranges during Pleistocene glaciations, global sea level repeatedly dropped significantly, exposing large portions of the continental shelves. For instance, during the last glaciation, global sea-level dropped by 120–130 m (Clark et al., 2009; Lambeck et al., 2002), exposing the continental shelf regions of the Bering and Chukchi seas between Siberia and Alaska. These vast shelves lie only about 30–110 m below modern sea level. Therefore, as continental ice sheets grew, sea-level dropped, and the shelf regions between NE Russia and Alaska became dry land forming the Bering land bridge. By mapping the 120 m contour of the sea floors for the Bering and Chukchi seas (without compensation for gravitational, deformational, and rotational perturbations in sea level, see Milne and Mitrovica, 2008), it is possible to roughly trace the outline of the former land bridge region (Figure 1). This land bridge was no narrow isthmus between two continents. Rather, from north to south, the land bridge extended more than 1000 km and may have existed for more than 10 000 years (from 32 000 to 18 000 years ago) during the last glacial advance. From east to west, the land bridge was narrow at the center (in the Bering Strait region) but quite broad at both the north and south ends. The area of the land bridge was larger than the state of Texas.

When it existed, the Bering Land Bridge cut off circulation between the waters of the North Pacific and Arctic oceans with global influences on thermohaline circulation and climate (Hu et al., 2010). Regionally, the presence of this vast land bridge increased continentality and diminished the inland

flow of relatively warm, moist air masses from the North Pacific. Beringia was also influenced by being downwind of large ice sheets in Scandinavia and the Eurasian north, which themselves created widespread aridity during full glacial conditions (Siegert et al., 2001; Figure 1). Moisture stripped from the westerlies by these ice sheets left little but strong dry winds to sweep the landscapes of Beringia. Consequently, despite being cold enough to accumulate glacial ice, the desiccating environment of the land bridge restricted glaciation to isolated mountain ranges and kept glacial ice out of the lowlands of Alaska, the Yukon, and eastern Siberia, plus the land bridge itself. The largest of the alpine glacier complexes was, in fact, the westernmost extension of the Cordilleran Ice Sheet that once occupied western North America. This ice sheet covered the Alaska Range, and extended south and west to the Alaskan Peninsula and the continental shelf regions of Prince William Sound and the Gulf of Alaska (Kaufman et al., 2011; Mann and Hamilton, 1995). The Beringian lowlands remained ice-free throughout the Pleistocene.

Beringia is most conveniently partitioned for discussion into three geographic sectors: western Beringia, central Beringia, and eastern Beringia. Western Beringia is comprised of the regions of northeast Russia, from the Verkhoyansk Mountains in the west, to the Bering and Chukchi Sea coasts in the east, and the Sea of Okhotsk in the south. The Verkhoyansk Mountains lie along the eastern margin of the lower Lena River Basin and form the western limit of some characteristic Beringian plant taxa. The central sector of Beringia is that area of the Bering Land Bridge that is now submerged beneath the Bering and Chukchi seas, forming what was once a broad, relatively flat landscape with only a few highlands (the modern islands of the northern Bering Sea region). Eastern Beringia is defined as the unglaciated regions of Alaska and the Yukon. The eastern boundary of Beringia has been considered to be the lower Mackenzie River in the Canadian Northwest Territories (Hopkins, 1982). This places the eastern edge of Beringia along the margin of the Laurentide ice sheet at its maximum extent.

The vast regions of Beringia contained enormous topographic diversity, spanning over more than 20° of latitude and 80° of longitude (Figure 1). One of the principal differences between western and eastern Beringia is that the former region is topographically more complex and rugged than the latter (Brigham-Grette et al., 2004). The northward-draining river basins of the Lena, Indigirka, and Kolyma are broken by steep mountain ranges, as well as broad uplands. The geography of Eastern Beringia is marked by three sets of mostly east-west trending mountain ranges. From south to north, these comprise (1) the Pacific coastal ranges that continue southwest along the Alaska Peninsula, (2) the Alaska Range, about 100 km inland from the Gulf of Alaska, and (3) the Brooks Range, separating the interior lowlands from the North Slope

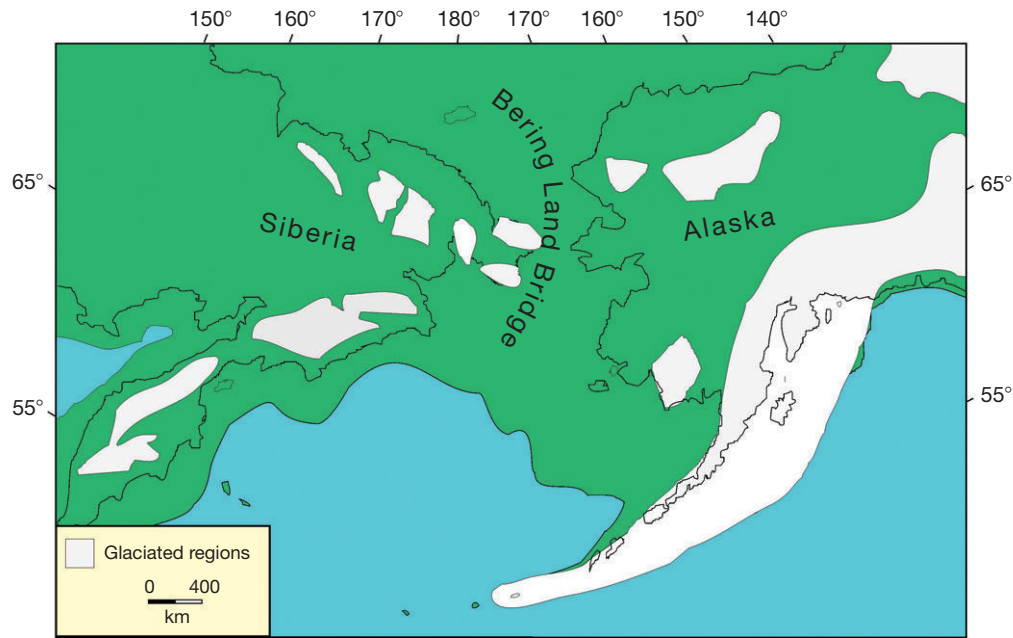


Figure 1 Map of northeast Asia and northwestern North America, showing the regions of western and eastern Beringia, and the Bering Land Bridge.

region. The interior regions of Alaska and the Yukon lie in the rain shadow of two high mountain ranges, which block Pacific moisture.

Contrasts in Paleoclimate Across Eastern and Western Beringia

During the past 20 years, cooperation between Russian and North American scientists has added significantly to our understanding of Beringian paleoenvironments, especially concerning the last glaciation. For instance, we have learned that there were significant differences between the climate and vegetation of western versus eastern Beringia during the marine isotopic stage 3 interstadial (MIS 3, ca. 48–28 ka). In western Beringia, temperatures apparently rose to nearly modern levels, allowing treeline to migrate northward and into mountain ranges to virtually its modern limit (Anderson and Lozhkin, 2001). In contrast to this, however, very little evidence exists for the expansion of coniferous forest in eastern Beringia during MIS 3. Based on fossil beetle evidence, there appears to have been different degrees of warming in the northern and southern sectors of eastern Beringia during this interstadial (Elias, 2001). The Arctic regions experienced mean summer temperatures as much as 1.5 °C warmer than modern levels at about 35 ka, whereas the synchronous warming in subarctic regions remained about 2 °C cooler than modern levels.

As in other high-latitude regions, the Last Glacial Maximum (LGM) was an interval of climatic cooling. Paleotemperature estimates based on fossil beetle assemblages (Elias, 2001) suggest that average summer temperatures (T_{MAX}) were as much as 7.5 °C cooler at 22 ka than they are today in the Yukon. However, it appears that this cooling of summer temperatures was far from uniform across Beringia. On the Seward Peninsula, for instance, T_{MAX} was depressed by a maximum of 2.9 °C.

Most eastern Beringian paleotemperature estimates based on LGM beetle data fall within a range of 3–5 °C colder than modern T_{MAX} . Mean January temperatures (T_{MIN}) in eastern Beringia were probably depressed by very little, if at all (Elias, 2001). The fossil beetle evidence (Alfimov and Berman, 2001) indicates that LGM summer temperatures in western Beringia were essentially as high as they are today in many regions, and T_{MIN} values remained close to modern levels as well. However, not all researchers agree on these temperature reconstructions. For instance, periglacial features that developed during the LGM indicate that mean annual temperatures dropped significantly (Hopkins, 1982). Most paleoclimatologists agree that on a very broad scale, Beringia was relatively dry and cold with cooler summers during the LGM. More mesoscale patterns indicate east to west trends in temperature and moisture gradients with colder and drier conditions dominant over eastern Beringia (Brigham-Grette et al., 2004). Guthrie (2001) postulated that during the LGM, western Beringia was kept arid by a huge, stable, high-pressure system north of the Tibetan Plateau. North Atlantic sea-ice cover and the Scandinavian/Eurasian ice sheets would have cut off the main source of westerly moisture, enhancing aridity in Asiatic steppe regions. Lowered sea level and the Cordilleran Ice Sheet would have reduced the moisture available to interior regions of eastern Beringia.

Sea-Level History

Six marine raised beaches have been documented along the more stable parts of the Alaskan coastline providing evidence that sea level was higher than that during some earlier interglacial warm periods. The oldest four high sea-level events occurred during the late Pliocene to middle Pleistocene and are significant with respect to paleoclimate (Kaufman and Brigham-Grette, 1993). The two late Pleistocene high stands

Table 1 Timing of major late Pleistocene cold events and marine transgressions in Beringia

Cold event	W. Beringian terminology	E. Beringian terminology	Age (calibrated year BP $\times$ 1000)	Marine transgression event
Middle MIS 5			125 to >100 100–88 <88 to >70	Pelukian Simpsonian
MIS 4	Zyryan glaciation	Early Wisconsinan glaciation	70–55	
MIS 2	Sartan glaciation	Late Wisconsinan glaciation	25–15	

are known as the Pelukian transgression, dated between 125 000 and 115 000 years BP (MIS 5e), and the Simpsonian transgression, dated between about 88 000 and 70 000 years BP (MIS 5a; [Table 1](#); [Brigham-Grette and Hopkins, 1995](#); [Brigham-Grette et al., 2001](#)).

The last interglacial, or MIS 5, broadly includes the period from about 130 000 to 75 000 years. The time period is normally characterized as including the peak of the last interglacial (MIS 5e) followed by two mild and two cooler intervals (MIS 5d through a) which eventually terminated in the first of the two major glacial intervals, MIS 4 ([Figure 2](#)). The peak of the last interglacial, MIS 5e with insolation at +11% of modern values, is remarkable across Beringia because the Bering Strait was submerged; the relative sea level along the Bering Strait coast stood roughly 6–7 m higher than it is today, and terrestrial climates were generally warmer. In the Bering and Chukchi seas, it has been estimated that for much of this interval, winter maximum sea-ice limits were as much as 800 km further north, well north of the narrow Bering Strait; summer sea ice was also less extensive and marine mollusks from the Bering Sea extended their range to Barrow, Alaska ([Brigham-Grette and Hopkins, 1995](#); [CAPE Last Interglacial Project Members, 2006](#); [Polyak et al., 2010](#)). Similar range extensions have been documented in marine sediments on the Russian coast and confirmed with the expansion of species of foraminifer from northern Japan as far north as eastern Chukotka ([Brigham-Grette et al., 2001](#)). Treeline spread to the Russian coast across most of western Beringia with range extensions among some tree species of 600–1000 km ([Lozhkin and Anderson, 1995a](#)). At the same time, in Alaska, the northward shift of treeline was more modest but extended at least 50 km north of the Brook Range ([Brubaker et al., 2005](#)). Temperatures estimates based on pollen and fossil beetle assemblages across the interior of Alaska suggest that summer temperatures were similar to today, but conditions were considerably wetter ([CAPE Last Interglacial Project Members, 2006](#); [Muhs et al., 2001](#)). In general, across Beringia, summer temperatures were 0–2 °C warmer, winters a few degrees cooler, and conditions were generally wetter.

Rapid Onset of Glaciation During MIS 5

Numerous workers have now documented that parts of the high arctic can become rapidly glaciated at the end of interglacials while eustatic sea level is still high. The Beringian region is no exception. Various researchers, including [Pushkar et al. \(1999\)](#), demonstrated that the mountains to the north and west of Kotzebue Sound had to have been rapidly glaciated

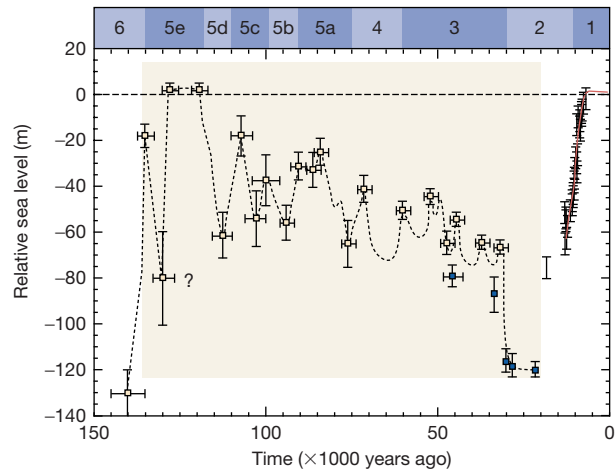


Figure 2 Relative sea-level curve for the last glacial cycle for Huon Peninsula, Papua New Guinea, supplemented with observations from Bonaparte Gulf, Australia. Error bars define the upper and lower limits. Time scale is based on U-series ages of corals older than about 30 ka and on calibrated ^{14}C ages of the younger corals and sediments. Reproduced from [Lambeck K, Esat TM, and Potter E-K \(2002\) Links between climate and sea levels for the past three million years. *Nature* 419: 199–206.](#)

near the end of MIS 11 (possibly stage 9). Such an explanation was necessary to explain the continuous deposition of extensive prodeltaic interglacial marine sediments overlain by glacially deformed glaciomarine muds. Because these deposits lie some 400 km from the edge of the shallow continental shelf, eustatic sea level had to have remained high as a consequence of widespread glaciomarine sedimentation in Kotzebue Sound in front of thin but extensive valley glaciers emanating from the western Brooks Range.

Research in Chukotka ([Brigham-Grette and Hayden, 2002](#); [Brigham-Grette et al., 2001](#)) and parts of southwest Alaska ([Kaufman et al., 1996, 2001](#)) has led to the realization that these regions also experienced rapid valley glacier expansion at times directly after the peak of MIS 5e. On outer Chukotka near the narrow Bering Strait, [Brigham-Grette et al. \(2001\)](#) re-assigned the Pinakul Formation, once thought to be of early Pleistocene age, to MIS 5a–e. The depositional sequence consists of an upward shallowing interglacial marine sequence characterized by warm faunas (the lower Pinakul Formation of MIS 5) overlain by a thin coarse gravel buried beneath a thick glaciomarine sequence containing classic drop stones (upper Pinakul Formation). The entire sequence is capped by glacial till, dated to MIS 4. [Brigham-Grette et al. \(2001\)](#) interpreted the section to represent 5e marine deposits overlain by an unconformity and then the rapid deposition of glaciomarine

sediments in MIS 5d. Deposition of such a thick section of glaciomarine mud in the shallowest coastal reaches of the Bering Strait requires that eustatic sea level remained near modern levels as rapidly expanding valley glaciers in coastal mountains reached the sea. Dating of the deposits is based on amino acid age estimates, radiocarbon, U/Th ages on concretions, and electron spin resonance (ESR) ages.

Valley glaciers responded in a similar fashion at roughly the same time in the Bristol Bay region of southwestern Alaska. Kaufman et al. (2001; Kaufman and Manley, 2004) demonstrated that mountain glaciers extended as much 100 km from their cirques and intersected the sea before eustatic sea level had dropped. Like the Pinakul sequence, this advance likely occurred either at the MIS 5e/5d transition or the MIS 5a/4 transition, based on stratigraphic relationships with the underlying Old Crow Tephra (ca. 140 ka) and a thermoluminescence (TL) age estimate of $70 \text{ ka} \pm 10$ on basalt overrun by advancing glacier ice. Calibrated amino acid age estimates and infrared-luminescence ages, however, offer the best geochronological control.

Stauch and Gualtieri (2008) used infrared stimulated luminescence (IRSL) dating of overlying cover sediments and ^{10}Be and ^{26}Al ages from moraine boulders in an east–west transect of Western Beringia, from the Verkhoyansk Mountains in the west to the Anadyr Uplands and Wrangel Island in the east. They documented glacial advances in the Verkhoyansk Mountains in MIS 5d, 5b, and possibly, MIS 4. The most recent glaciation in this region took place earlier than 50 000 years ago.

The rapid glacierization in Beringia at the end of MIS 5 was presumably driven by a precipitous drop in insolation while oceanic temperatures remained warm and eustatic sea level remained relatively high. However, changes in oceanic circulation also contributed to the onset of widespread Arctic glaciation in sensitive regions. Marine evidence suggests that major climate shifts such as the MIS 5e/5d transition also were driven by changes in deep water flux (Adkins et al., 1997; Kageyama et al., 2010). There is evidence for a rapid reorganization of deepwater production at the 5e/5d transition over a period of less than 400 years (Adkins et al., 1997). Hemispheric changes in the ocean and atmospheric system likely contributed to the rapid buildup of the Eurasian Ice Sheet at the 5e/5d transition (Krinner et al., 2011). Reconstructions of the Barents and Kara Ice sheets at this time suggest that the glaciers came onto shore and dammed large proglacial lakes in the Ural Mountains as early as 85–90 ka cal years BP (Mangerud et al., 2001). Ice sheets then retreated northward and readvanced in stage 4 to the Markhida Moraine by about 60 ka cal years BP.

Changes in the height and extent of the Scandinavian and Barents/Kara sea-ice sheets likely had a significant influence on the temporal and spatial response of the eastern Siberia and western Beringia (NE Siberia) to hemispheric scale climate change. Ice sheet and general circulation modeling (GCM) of the Scandinavian and Eurasian ice sheets by Siebert and Marsait (2000) suggests the extent to which changes in the size of these ice sheets likely diminished the temperature and precipitation influence of the North Atlantic eastward across the Russian arctic. Fine-tuned models of ice sheet size for parts of the late Pleistocene (Siebert et al., 2001) allow for more realistic assessments of how the physical stratigraphy of

western Beringia may have been influenced by ‘downwind’ effects while being upwind of the maritime influences of the Bering Strait and the sea-ice conditions in the Bering Sea.

The timing of ice sheet growth in Eurasia and Beringia fits well with the notion for early ice buildup over parts of the Canadian Arctic at this same time (Brigham-Grette and Hopkins, 1995). In Northern Alaska along the Beaufort Sea coast, the Flaxman glaciomarine deposits are dated to substage 5a (Carter et al., 1991). It is also well known and mapped below 7 m altitude on the Alaskan Arctic Coastal Plain (Brigham-Grette and Hopkins, 1995) recording both a high sea-level event that flooded the Bering Straits and the *Heinrich-like* collapse into the Beaufort Sea via Amundsen Gulf of a large ice sheet building over the Canadian Arctic islands, presumably starting in stage 5d. Erratics in the Flaxman beds are exclusively of Canadian provenience, reaching all the way to the Chukchi Sea coast of Alaska. TL ages on sediments, several ESR ages on shells, and a uranium-series age on whale bone, all place the age of the sediments at about 75 ka BP. The presence of the large whale bones in these deposits east of Barrow is especially significant since it requires that the Bering Strait was submerged enough (sea level above –55 m) to allow for the seasonal migration of bowhead and gray whales from the open Pacific, as they do today.

Flooding of the Bering Strait in MIS 5a, coupled with high insolation, may have contributed to the collapse of these Canadian and Eurasian ice sheets. The collapse of these ice sheets may have then allowed the penetration of moisture across the continent, permitting the expansion of valley glaciers in the Bering Straits region and elsewhere in the high latitudes. Afterward, increasing continentality, sea-ice cover, and the expansion of the Scandinavian/Eurasian ice sheet began to limit the available moisture supply across most of Beringia.

Early Wisconsin/Zyryan Glaciation (MIS 4)

Glaciation during MIS 4 is referred to as the Early Wisconsin in North America and the Zyryan in northeastern Russia, using traditional time stratigraphic terms. With the exception of the Alaska Range and Chugach/St. Elias Mountains facing the Gulf of Alaska, the Early Wisconsin/Zyryan glacial ice caps and mountain glacier complexes throughout much of Beringia were considerably more extensive than during the subsequent Late Wisconsin/Sartan glaciation in MIS 2 (Figure 3). Satellite and aerial photo mapping followed up by field studies of moraine limits across much of Chukotka indicates that Zyryan glaciers advanced 50–200 km from their source areas during MIS 4 (Heiser and Roush, 2001). Glushkova (1995, 2011) suggested that nearly 40% of Chukotka was covered with ice during this interval and in general, glacial advances were up to three times more extensive than during the Sartan glaciation. On the southeast corner of Chukotka, mountain glacier complexes coalesced into a piedmont lobe of sufficient size that the glacier terminated on St Lawrence Island leaving a moraine belt mappable on the sea floor (Brigham-Grette et al., 2001). Stauch and Gualtieri (2008) concluded that the penultimate glaciation had been the most extensive throughout Western Beringia. In some regions (i.e., the Koryak Mountains), this glacial advance apparently took place in MIS 4, while in the Verkhoyansk Mountains, and in the Pekulney Mountains, the

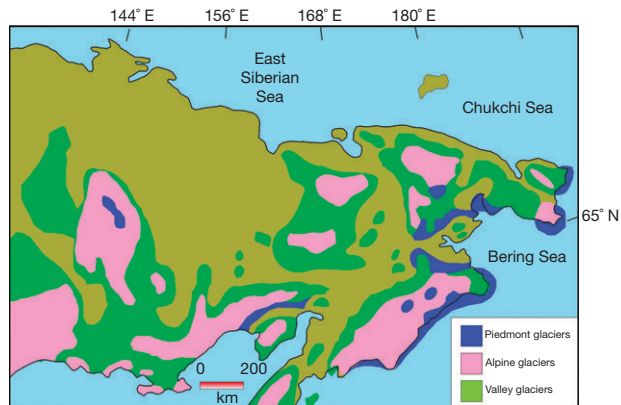


Figure 3 Map of western Beringia, showing the extent of glaciers during the Zyryan glaciation. Reproduced from Brigham-Grette J, Lozhkin AV, Anderson PM, and Glushkova OY (2004) Paleoenvironmental conditions in Western Beringia before and during the Last Glacial Maximum. In: Madsen D.B. (ed.) *Entering America. Northeast Asia and Beringia Before the Last Glacial Maximum*, pp. 29–61. Salt Lake City: University of Utah Press.

Tanyurer River Valley, and in Kamchatka, the penultimate glaciation apparently took place in MIS 5 and MIS 6, respectively.

Correlation between Zyryan glacial events in western Beringia and Early Wisconsinan glacial events in eastern Beringia can be difficult (Heiser and Roush, 2001) because of dating problems. The ages of MIS 4 glacial advances in Beringia are most commonly based on morphostratigraphic relationships coupled with a variety of relative age estimate criteria, stream terrace stratigraphy, and most recently, cosmogenic isotopic dating.

New sediment records from Lake El'gygytyn ('Lake E') in the heart of Chukotka, NE Russia, provide evidence of time-continuous paleoclimate change over the past 350 000 years (Brigham-Grette et al., 2007). It is located 100 km north of the Arctic Circle in Chukotka, NE Russia (Figure 4), and was produced by a meteorite impact 3.6 million years ago during the warm mid-Pliocene geological period. When the meteorite hit, the Arctic was exceedingly different from today, with mean annual temperatures $\sim 12^\circ\text{C}$ warmer and global averages $3\text{--}4^\circ\text{C}$ warmer than today (Lunt et al., 2009). At that time, most of the Arctic borderlands were heavily forested, the Arctic Ocean lacked permanent sea ice, and the Greenland Ice Sheet did not exist, at least not in its present form (Brigham-Grette and Carter, 1992; Miller et al., 2010). In fact after the impact, a million years passed before the onset of the first major glaciation of the Northern Hemisphere. The El'gygytyn Crater is unique, having formed in the center of what was to become Beringia – the largest contiguous landscape in the Arctic to escape continental scale glaciation. Pilot cores from Lake E document regional changes in paleoclimate reflected by changes from cold-dry to cold wet episodes during glaciation to warm-wet interglacial intervals (Melles et al., 2007). The last interglacial (MIS 5e) at this site is exceptionally warm, consistent with the idea that tundra was nearly absent from western Beringia during this time (Lozhkin and Anderson 1995a; Lozhkin et al., 2007). This nonglacial record provides the first continuous record of multiple glacial and interglacial cycles from the western Arctic (Brigham-Grette et al., 2007 and papers therein;

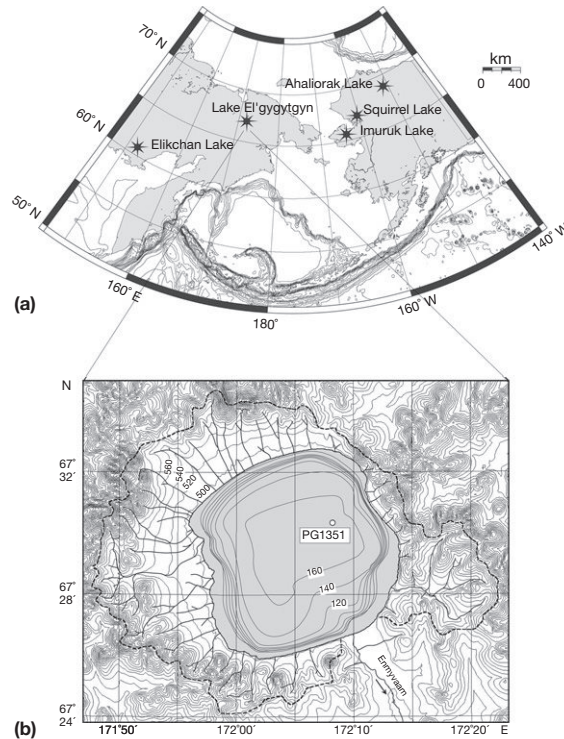


Figure 4 Location of Lake El'gygytyn, 100 km north of the Arctic Circle in northeastern Russia. (a) Location of Lake El'gygytyn relative to other key lacustrine basins in Beringia with long sediment records including Elikchan Lake (65 ka; Lozhkin and Anderson, 1995b); Imuruk Lake (130 ka; Colinvaux, 1964); Squirrel Lake (180 ka; Berger and Anderson, 2000); and Ahaliork Lake (130 ka; Eisner and Colinvaux, 1990; Anderson and Brubaker, unpublished). (b) Topography of El'gygytyn crater and bathymetry of the modern lake system showing the location of pilot core PG 1351. Reproduced from Brigham-Grette J, Melles M, Minyuk P, and Scientific Party (2007) Overview and significance of a 250 ka paleoclimate record from El'gygytyn Crater Lake, NE Russia. *Journal of Paleolimnology* 37(1): 1–16.

Figure 5). Glacial intervals suggest variable cold and dry conditions consistent with tundra steppe conditions.

Several glacial advances have been documented in Alaska between the middle and late Pleistocene. Among the most distinctive features of these glaciations are the Salmon Lake moraines of the Kigluaik Mountains of the Seward Peninsula, Alaska (Kaufman and Calkin, 1988; Figure 6). The moraines of the Itkilik I stage in the Brooks Range are also thought to represent MIS 4 glaciation (Hamilton, 1986). Ice advanced onto St. Lawrence Island, north of the Bering Strait, some time in the late Pleistocene. A stratigraphic study of sedimentary sequences on the island indicates that this glaciation occurred after the last interglacial and before the Late Wisconsinan glaciation, so it seems most likely that it occurred in MIS 4 (Heiser and Roush, 2001).

Cosmogenic exposure ages of moraine boulders from sites in the Alaska Range and Ahklun Mountains that are associated with the penultimate glaciation have shown that this advance culminated between 60 000 and 50 000 years ago (Briner and Kaufman, 2008). In the northern Alaska Range, ^{10}Be ages from moraine boulders indicate that the MIS 4 glaciation stabilized between 58 000 and 52 000 years ago (Kaufman et al., 2011).

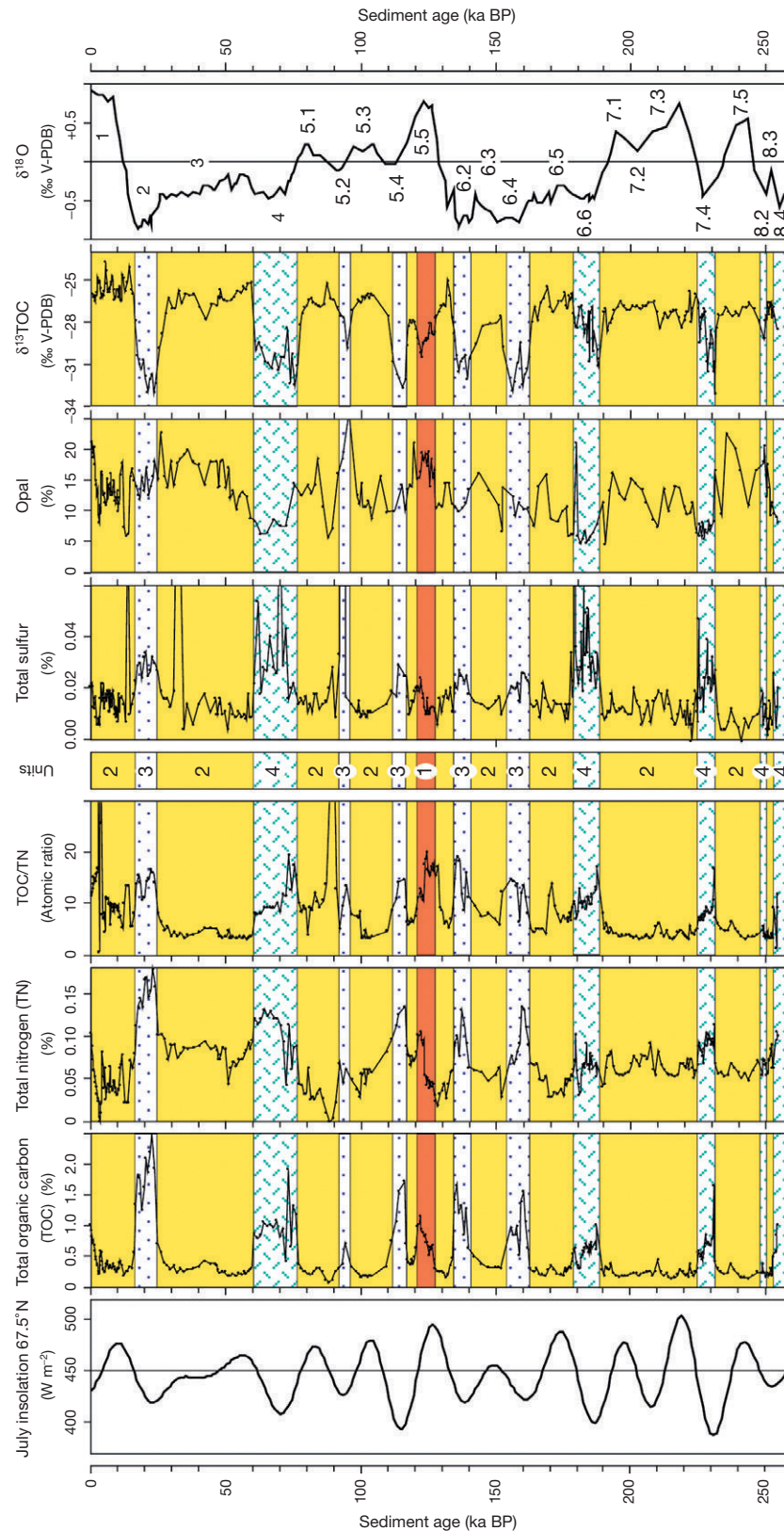


Figure 5 July insolation at 67.5° N (Paillard et al., 1996) and stacked marine oxygen isotope record (Martinson et al., 1987) with isotopic stages and substages for the past 260 ka, plotted versus the major biogeochemical proxies in core PG1351 from Lake El'gygytyn according to the age model presented in Nowaczyk et al. (2007). The peak warm (unit 1), normal warm (unit 2), cold and dry (unit 3), and cold but more moist (unit 4) climate modes. Reproduced from Melles M, Brigham-Grette J, Glushkova O.Yu, Minyuk PS, Nowaczyk NR, and Hubberten HW (2007) Sedimentary geochemistry of a pilot core from Lake El'gygytyn – A sensitive record of climate variability in the East Siberian Arctic during the past three climate cycles. *Journal of Paleolimnology* 37: 89–104.

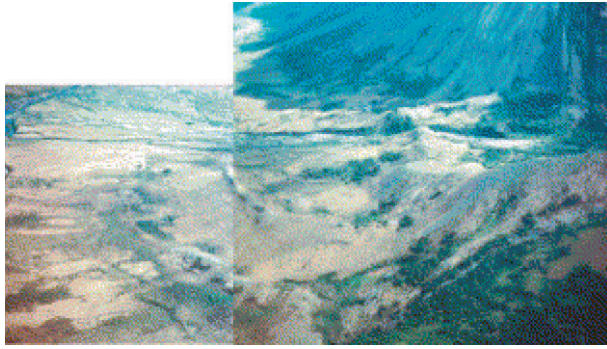


Figure 6 Ridge of the Early Wisconsin Salmon Lake moraine exiting Cobblestone Creek, Seward Peninsula. Photo by Brigham-Grette.

The ages of penultimate glacial moraines in the Brooks Range have not yet been dated, but likely are of similar age as they postdate the Old Crow tephra, which is now thought to have been deposited in MIS 6 (Hamilton, 2001). But in the Bristol Bay region of southwest Alaska, the penultimate advance appears to have culminated in late MIS 5 (Kaufman and Thompson, 1998).

¹⁰Be ages from moraine boulders from high-elevation sites in the Anvil Range and Pelly Mountains of the Yukon indicate that these areas were ice-covered before the McConnell (Late Wisconsin) glaciation and that they became ice-free ~30 000 years before the present (Stroeven et al., 2010). This places a likely MIS 4 age for this glacial advance. Ward et al. (2007) also found evidence for an MIS 4 glacial advance in the Aishihik area of southwestern Yukon. Cosmogenic ages on boulders on the penultimate drift ranged from about 54 000 to 51 000 years ago. They proposed that this glacial advance be termed the 'Gladstone glaciation' in the Yukon. This correlated well with the Icefield glaciation defined from glacial deposits in the nearby St. Elias Mountains of southeastern Alaska.

Paleotemperature estimates for MIS 4 in Eastern Beringia are indicative of cold summers. T_{MAX} cooling of 6 °C below modern levels has been estimated from beetle assemblages in Alaska and the Yukon (Elias, 2001). However, T_{MIN} values were estimated to be near modern levels. This seasonal temperature profile would aid the growth of mountain glaciers, as colder summer temperatures would inhibit summer melting of snowpack, and winter temperatures at these high latitudes have been sufficiently low to allow glacial ice to form, throughout the Pleistocene.

Middle Wisconsin/Karaginsky Interstadial

The Beringian interstadial warming associated with MIS 3 is known as the Karaginsky interstadial in western Beringia and as the Middle Wisconsin interstadial in eastern Beringia. During this interval, glacial margins retreated throughout Beringia. As discussed earlier, the paleobotanical evidence from western Beringia indicates warmer than modern conditions there during this interval, especially in the Arctic. Fossil beetle evidence (Sher et al., 2002) indicates that T_{MAX} was 1–3 °C warmer than modern in the Arctic and that T_{MIN}

was lower than modern, pointing toward increased continentality during the Karaginsky interval. This level of interstadial summer warming apparently did not occur in eastern Beringia. The fossil beetle evidence suggests that early MIS 3 environments had T_{MAX} values about 3 °C colder than modern and T_{MIN} values about 3 °C warmer than modern. This reconstruction fits well with both the summer and winter insolation pattern at the beginning of stage 3, about 60 000 years BP. Younger (stratigraphically higher) stage 3 samples from the Yukon yielded T_{MAX} closer to modern values, with T_{MIN} values remaining warmer than present. Eastern Beringian samples dated between 32 400 and 29 600 years BP are indicative of mean summer temperatures that were 1–4 °C colder than today.

Paleobotanical evidence from eastern Beringia shows some expansion of coniferous forests in interior regions, but not approaching modern northern treeline in Alaska. The warmest interval of the interstadial documented in the vegetation comes from interior Alaska. The 'Fox Thermal Event,' dated from 35 000 to 30 000 ¹⁴C years BP, is associated with expansion of spruce-forest tundra (Anderson and Lozhkin, 2001).

Late Wisconsin/Sartan Glaciation

The last Pleistocene glaciation in western Beringia is called the Sartan glaciation. The lack of significant moisture across Beringia during the LGM prevented the growth of large ice masses. As shown in Figures 7 and 8, the Sartan and Late Wisconsin ice limits were well inside those of some previous glaciations. LGM glaciation in western Beringia was essentially limited to valley and cirque glaciation in some mountain ranges (Brigham-Grette et al., 2003). Glacial extent decreased from west to east in western Beringia, probably because of increasing aridity toward the Bering Strait region. The Verkhoyansk Mountain region developed glacial ice that extended downslope almost to the Lena River (Figure 8; Stauch and Gualtieri, 2008; Zamoruyev, 2004). Radiocarbon dating of wood and organic matter, combined with cosmogenic isotope surface exposure ages, indicates that LGM ice in western Beringia reached its maximum extent between 27 000 and 20 000 cal years BP (Brigham-Grette et al., 2003). A study of fossil diatom assemblages from Bering Sea sediment cores (Brigham-Grette et al., 2003) suggests that the Bering Sea experienced sea-ice cover for as much as 9 months per year during the LGM; Caissie et al. (2010) have refined this estimate to suggest the presence of perennial ice in the eastern Bering Sea during the LGM. The combination of Bering Sea ice cover and the 1000-km-long land bridge in the center of Beringia meant that Pacific moisture was far removed from the interior regions of Beringia during the LGM.

Late Wisconsin glaciers occupied about 726 000 km² of Alaska (Kaufman et al., 2011). The transition from MIS 3 to MIS 2 is imprecisely dated in eastern Beringia, but by about 32 000 cal years BP, the climate changed from interstadial to glacial, as evidenced by changes in regional floras and faunas. Apparently, the glacial advances in eastern Beringian mountain ranges were not synchronous. The Late Wisconsin advance began in the mountains of southwestern Yukon by 31 000 cal years BP (Hamilton and Fulton, 1994). The Cordilleran Ice

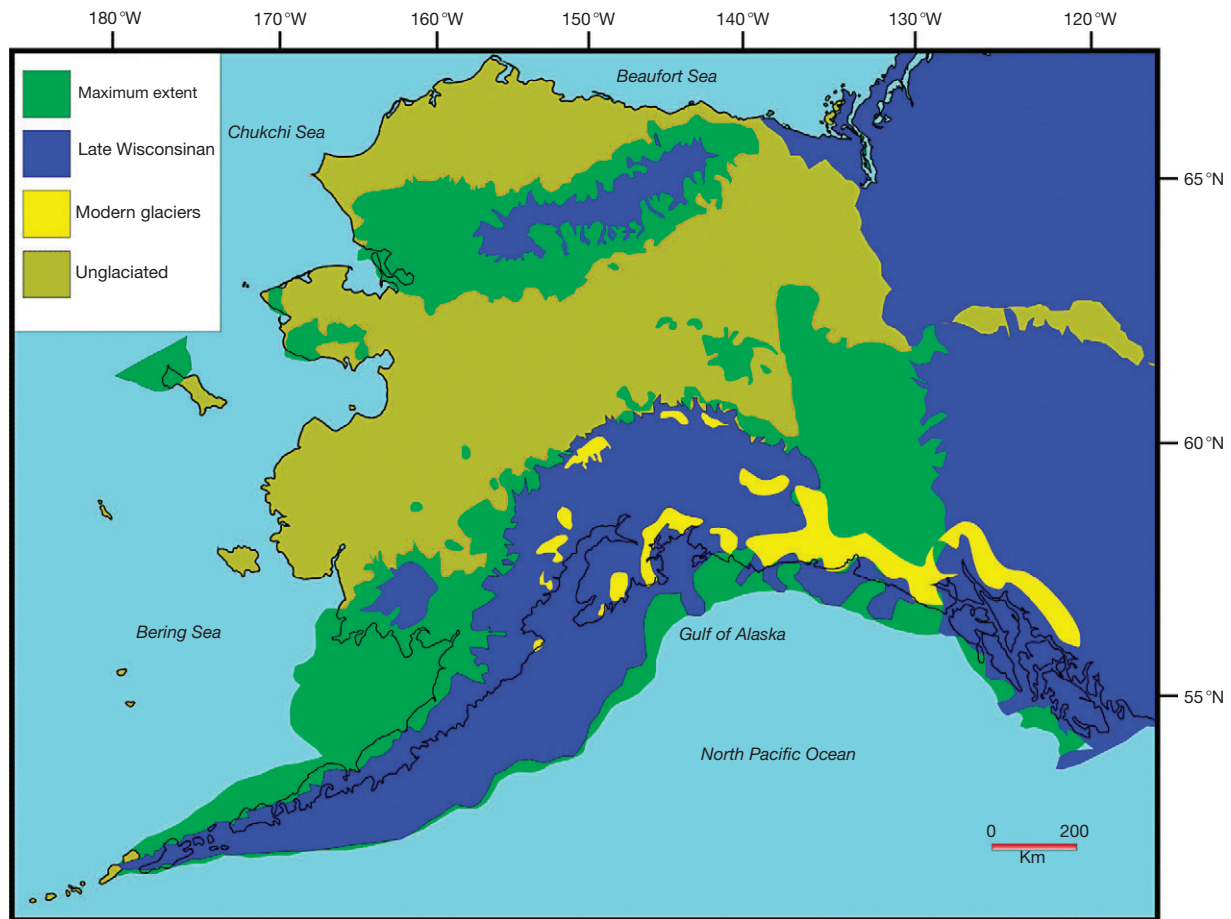


Figure 7 Map of eastern Beringia, showing the extent of glaciers during the Early and Late Wisconsinan glaciations. Reproduced from Manley WF and Kaufman DS (2002) *Alaska PaleoGlacier Atlas*. Institute of Arctic and Alpine Research, University of Colorado; Duk-Rodkin A (1999) Glacial limits map of Yukon. *Geological Survey of Canada*, Open File Report 3694.

Sheet advanced throughout the coastal regions of the Pacific Northwest about 28 000 cal years BP. At that time, ice advanced in the Brooks Range, signaling the beginning of the Itkilik II glaciation (Hamilton and Ashley, 1993). Ice advances in the highlands of the Seward Peninsula are termed the Mount Osborne glaciation (Figure 9). In Denali National Park (Alaska Range), glaciers neared their late Wisconsin limit about 22 000 years ago and began retreating by about 19 000 years ago (Kaufman et al., 2011). LGM glaciers reached their maximum size in the Ahklun Mountains of southwestern Alaska about 24 000 years ago (Briner and Kaufman, 2008) and began retreating by about 20 000 years ago. In contrast to these chronologies, LGM ice margins in the Brooks Range were already retreating by 25 000 years ago. Likewise, evidence for stabilization of sand sheets and dunes in the Kobuk Dunes region of northwestern Alaska after 22 000 cal years BP suggests climatic amelioration that may correlate with the Port Moody Interstade documented from southwestern British Columbia (Clague et al., 2004).

During the LGM, the Cordilleran Ice Sheet covered southern and central Yukon and parts of southern Alaska (Figure 7), where it formed ice fields and large valley and piedmont glaciers, flowing out of the St. Elias Mountains, the Chugach Range, and the Alaska Range. The ice extended out onto the

continental shelf to the south and into the broad, low country drained by the Yukon River and its tributaries to the north (Mann and Hamilton, 1995). Ice expanded from the coastal mountains of British Columbia into the Alexander Archipelago of southeastern Alaska, reaching the outer continental shelf in some regions (Carrara et al., 2002).

One of the earliest dates for deglaciation following the LGM in eastern Beringia comes from the Hungry Creek site, Yukon. Hughes (1983) suggested that the western limit of Cordilleran ice began retreating here by about 19 000 years BP. Mann and Hamilton (1995) suggest that ice retreat was time-transgressive in southern Alaska, with northern shelf areas becoming ice-free by 16 000 years BP. Much of southeastern Alaska was deglaciated by about 15 000 cal years BP (Clague et al., 2004). Deglaciation in the Brooks Range began by at least 13 400 cal years BP (Hamilton, 1983). Postglacial environmental change throughout eastern Beringia brought about wholesale changes in vegetation, the regional extinction of much of the megafauna, and the entrance of *Homo sapiens*.

The Younger Dryas oscillation first documented in Northwest Europe and in the Greenland ice core stable isotope records was expressed to greater or lesser degrees in the Alaska glaciological record. Briner et al. (2002) dated glacial advances in the Ahklun Mountains to the interval 11 700–11 000 years

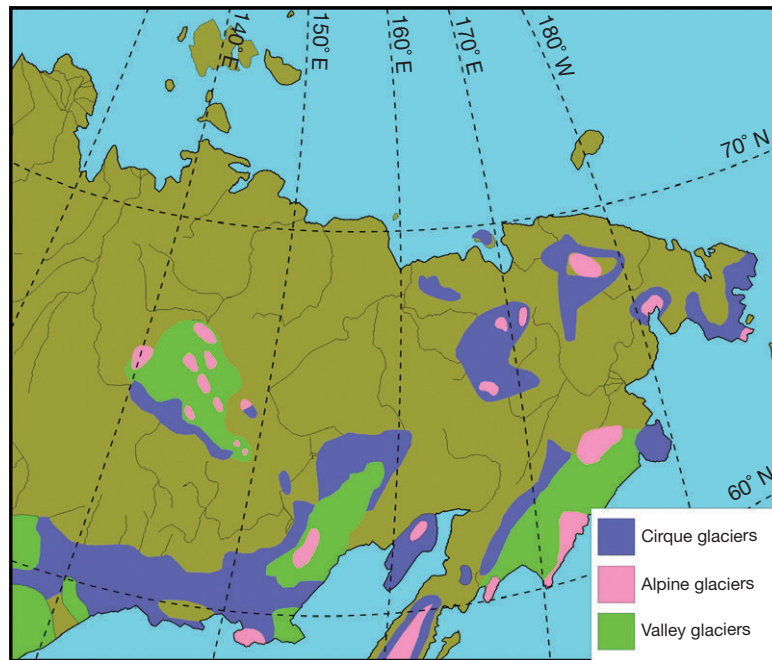


Figure 8 Map of western Beringia, showing the extent of glaciers during the Sartan glaciation. Reproduced from Brigham-Grette J, Lozhkin AV, Anderson PM, and Glushkova OY (2004) Paleoenvironmental conditions in Western Beringia before and during the Last Glacial Maximum. In: Madsen DB (ed.) *Entering America. Northeast Asia and Beringia Before the Last Glacial Maximum*, pp. 29–61. Salt Lake City: University of Utah Press.



Figure 9 Late Wisconsinan Mt. Osborn moraines exiting a small cirque into the valley of Cobblestone Creek, Seward Peninsula. Photo by Brigham-Grette.

ago. Matmon et al. (2006) obtained ^{10}Be ages on moraines along the northern front of the Alaska Range that also date to this interval. But in other regions, such as the Brooks Range, there is little or no evidence for a Younger Dryas glacial advance (Hamilton, 2003).

See also: **Glaciations:** Late Pleistocene in Eurasia; Late Quaternary in North America; Overview. **Plant Macrofossil Records:** Arctic North America. **Vertebrate Records:** Late Pleistocene Megafaunal Extinctions.

References

- Adkins JF, Boyle EA, Keigwin L, and Cortijo E (1997) Variability of the North Atlantic thermohaline circulation during the last interglacial period. *Nature* 390: 154–156.
- Alfimov AV and Berman DI (2001) Beringian climate during the Late Pleistocene and Holocene. *Quaternary Science Reviews* 20: 127–134.
- Anderson PM and Lozhkin AV (2001) The Stage 3 interstadial complex (Karginskii/middle Wisconsinan interval) of Beringia: Variations in paleoenvironments and implications for paleoclimatic interpretations. *Quaternary Science Reviews* 20: 93–125.
- Berger GW and Anderson PM (2000) Extending the geochronometry of arctic lake cores beyond the radiocarbon limit by using thermoluminescence. *Journal of Geophysical Research* 105(D12): 15439–15455.
- Brigham-Grette J and Carter J (1992) Pliocene marine transgressions of northern Alaska: Circumarctic correlations and paleoclimatic interpretations. *Arctic* 45: 74–89.
- Brigham-Grette J, Gualtieri LM, Glushkova OY, Hamilton TD, Mostoller D, and Kotov A (2003) Chlorine-36 and ^{14}C chronology support a limited last glacial maximum across central Chukotka, northeastern Siberia, and no Beringian ice sheet. *Quaternary Research* 59: 386–398.
- Brigham-Grette J and Hayden T (2002) Land/sea correlations: Separating climate events at the 5e/d and 5a/4 transitions in Chukotka and across the western Arctic. *Geological Society of America Annual Meeting, Program and Abstracts*, 145–5.
- Brigham-Grette J and Hopkins DM (1995) Emergent marine record and paleoclimate of the last interglaciation along the northwest Alaskan coast. *Quaternary Research* 43: 159–173.
- Brigham-Grette J, Hopkins DM, Ivanov VF, et al. (2001) Last Interglacial (isotope stage 5) glacial and sea-level history of coastal Chukotka Peninsula and St. Lawrence Island, Western Beringia. *Quaternary Science Reviews* 20: 419–436.
- Brigham-Grette J, Lozhkin AV, Anderson PM, and Glushkova OY (2004) Paleoenvironmental conditions in Western Beringia before and during the Last Glacial Maximum. In: Madsen DB (ed.) *Entering America: Northeast Asia and Beringia Before the Last Glacial Maximum*, pp. 29–61. Salt Lake City: University of Utah Press.
- Brigham-Grette J, Melles M, and Minyuk P (2007) Scientific Party: Overview and significance of a 250 ka paleoclimate record from El'gygytyn Crater Lake, NE Russia. *Journal of Paleolimnology* 37(1): 1–16.

- Briner JP and Kaufman DS (2008) Late Pleistocene mountain glaciation in Alaska: Key chronologies. *Journal of Quaternary Science* 23: 659–670.
- Briner JP, Kaufman DS, Werner A, et al. (2002) Glacier readvance during the late glacial (Younger Dryas?) in the Ahklun Mountains, southwestern Alaska. *Geology* 30: 679–682.
- Brubaker LB, Anderson PM, Edwards ME, and Lozhkin AV (2005) Beringia as a glacial refugium for boreal trees and shrubs: New perspectives from mapped pollen data. *Journal of Biogeography* 35: 833–848.
- Caissie BE, Brigham-Grette J, Lawrence K, Herbert T, and Cook MS (2010) Last Glacial Maximum to Holocene sea surface condition at Umnak Plateau, Bering Sea as inferred from diatom, alkenone, and stable isotope records. *Paleoceanography* 25: PA1206. <http://dx.doi.org/10.1029/2008PA001671>.
- CAPE Last Interglacial Project Members (2006) Last Interglacial Arctic warmth confirms polar amplification of climate change. *Quaternary Science Reviews* 25: 1383–1400.
- Carrara PE, Ager TD, Baichtal JF, and Van Sistine D (2002) Late Wisconsin glacial limits in southern Southeastern Alaska, as indicated by a new bathymetric map. In: *32nd International Arctic Workshop*, Program and Abstracts, pp. 41–43. Institute of Arctic and Alpine Research, University of Colorado, Boulder.
- Carter LD, Hamilton TD, and Galloway JP (1991) Late Cenozoic history of the interior basins of Alaska and the Yukon. U.S. Geological Survey Circular 1026.
- Clague JJ, Mathewes RW, and Ager TD (2004) Environments of northwestern North America before the Last Glacial Maximum. In: Madsen DB (ed.) *Entering America: Northeast Asia and Beringia Before the Last Glacial Maximum*, pp. 63–94. Salt Lake City: University of Utah Press.
- Clark PU, Dyke AS, Shakun JD, et al. (2009) The Last Glacial Maximum. *Science* 325: 710–714.
- Colinvaux PA (1964) The environment of the Bering Land Bridge. *Ecological Monographs* 34: 297–329.
- Eisner WR and Colinvaux PA (1990) A long pollen record from Ahaliarak Lake, Alaska. *Review of Palaeobotany and Palynology* 63: 35–52.
- Elias SA (2001) Mutual climatic range reconstructions of seasonal temperatures based on late Pleistocene fossil beetle assemblages in Eastern Beringia. *Quaternary Science Reviews* 20: 77–91.
- Glushkova OY (1995) Paleogeography of Late Pleistocene glaciation of North-Eastern Asia. In: Simakov KV and Thurston DK (eds.) *Proceedings of the International Conference of Arctic Margins*, p. 350. Russian Academy of Sciences, Far East Branch, Northeast Science Center, Magadan.
- Glushkova OY (2011) Late Pleistocene Glaciations in North-East Asia. In: Ehlers J, Gibbard PL, and Hughes PD (eds.) *Quaternary Glaciations – Extent and Chronology, A closer look. Developments in Quaternary Science* 15: 865–875.
- Guthrie RD (2001) Origins and causes of the mammoth steppe: A story of cloud cover, woolly mammoth tooth pits, buckles, and inside-out Beringia. *Quaternary Science Reviews* 20: 549–574.
- Hamilton TD (1983) Glaciation of the Brooks Range. In: Thorson RM and Hamilton TD (eds.) *Glaciation in Alaska*, pp. 35–41. Fairbanks: Alaska Quaternary Center, University of Alaska.
- Hamilton TD (1986) Late Cenozoic glaciation of the central Brooks Range. In: Hamilton TD, Reed KM, and Thorson RM (eds.) *Glaciation in Alaska*, pp. 9–49. Anchorage: Alaska Geologic Society.
- Hamilton TD (2001) Quaternary glacial, lacustrine, and fluvial interactions in the western Noatak basin, northwest Alaska. *Quaternary Science Reviews* 20: 371–391.
- Hamilton TD (2003) Surficial geologic map of parts of the Misheguk Mountain and Baird Mountains quadrangles, Noatak National Preserve, Alaska. U.S. Geological Survey Open File Report 2003–367.
- Hamilton TD and Ashley GM (1993) Epiguruk: A late Quaternary environmental record from northwestern Alaska. *Geological Society of America Bulletin* 105: 583–602.
- Hamilton TD and Fulton RJ (1994) Middle and Late Wisconsin environments of Eastern Beringia. In: Bonnicksen R, Frison GC, and Turnmire K (eds.) *Ice-Age Peoples of North America*. College Station, TX: Texas A&M University Press.
- Heiser PA and Roush JJ (2001) Pleistocene glaciations in Chukotka, Russia: Moraine mapping using satellite synthetic aperture radar (SAR) imagery. *Quaternary Science Reviews* 20: 393–404.
- Hopkins DM (1982) Aspects of the paleogeography of Beringia during the late Pleistocene. In: Hopkins DM, Matthews JV, Schweger CE, and Young SB (eds.) *Paleoecology of Beringia*, pp. 3–28. New York: Academic Press.
- Hu A, Mehl GA, Otto-Bliessen BL, et al. (2010) Influence of the Bering Strait flow and North Atlantic circulation on glacial sea level changes. *Nature Geoscience* 3: 118–121.
- Hughes O (1983) Quaternary geology, Yukon Territory and western District of Mackenzie. In: Thorson RM and Hamilton TD (eds.) *Glaciation in Alaska*, pp. 51–56. Fairbanks: Alaska Quaternary Center, University of Alaska.
- Kageyama M, Paul A, Roche DM, and Van Meerbeeck CJ (2010) Modelling glacial climatic millennial-scale variability related to changes in the Atlantic meridional overturning circulation: A review. *Quaternary Science Reviews* 29: 2931–2956.
- Kaufman DS and Brigham-Grette J (1993) Aminostratigraphic correlations and paleotemperature implications, Pliocene-Pleistocene high-sea-level deposits, northwestern Alaska. *Quaternary Science Reviews* 12: 21–33.
- Kaufman DS and Calkin PE (1988) Morphometric analysis of Pleistocene glacial deposits in the Kigluak Mountains, northwestern Alaska, USA. *Arctic and Alpine Research* 20: 272–284.
- Kaufman DS, Forman SL, Lea PD, and Wobus CW (1996) Age of pre-late-Wisconsin glacial-estuarine sedimentation, Bristol Bay, Alaska. *Quaternary Research* 45: 59–72.
- Kaufman DS and Manley WF (2004) Pleistocene maximum and Late Wisconsin glacier extents across Alaska, U.S.A. In: Ehlers J and Gibbard PL (eds.) *Quaternary Glaciations – Extent and Chronology, Part II: North America. Developments in Quaternary Science*, vol. 2, pp. 9–27. Amsterdam: Elsevier.
- Kaufman DS, Manley WF, Forman SL, and Lauer P (2001) Pre-late-Wisconsin glacial history, coastal Ahklun Mountains, southwestern Alaska – New amino acid, thermoluminescence, and $^{40}\text{Ar}/^{39}\text{Ar}$ results. *Quaternary Research* 45: 59–72.
- Kaufman DS and Thompson CH (1998) Re-evaluation of pre-late-Wisconsin glacial deposits, lower Naknek valley, southwestern Alaska. *Arctic and Alpine Research* 30: 142–151.
- Kaufman DS, Young NE, Briner JP, and Manley WF (2011) Alaska Palaeoglacier Atlas version 2. In: Ehlers J and Gibbard PL (eds.) *Quaternary Glaciations Extent and Chronology, Part IV: A Closer Look. Developments in Quaternary Science*, pp. 427–445. Amsterdam: Elsevier.
- Krinner G, Diekmann B, Colleoni F, and Stauch G (2011) Global, regional and local scale factors determining glaciation extent in Eastern Siberia over the last 140,000 years. *Quaternary Science Reviews* 30: 821–831.
- Lambeck K, Esat TM, and Potter E-K (2002) Links between climate and sea levels for the past three million years. *Nature* 419: 199–206.
- Lozhkin AV and Anderson PM (1995a) The last interglaciation in northeast Siberia. *Quaternary Research* 43: 147–158.
- Lozhkin AV and Anderson PM (1995b) A late Quaternary pollen record from Elikchan 4 Lake, northeast Siberia. *Geology of the Pacific Ocean* 14: 18–22.
- Lozhkin AV, Anderson PM, Matrosova TV, and Minyuk P (2007) Vegetation and climate histories of El'gygytyn Lake, Northeast Siberia. *Journal of Paleolimnology* 37: 135–153.
- Lunt DJ, Haywood AM, Schmidt GA, Salzmann U, Valdes PJ, and Dowsett HJ (2009) Earth system sensitivity inferred from Pliocene modeling and data. *Nature Geoscience* 3: 60–64.
- Mangerud J, Aasthkov V, Jakobsson M, and Svendsen JI (2001) Huge ice-age lakes in Russia. *Journal of Quaternary Science* 16: 773–777.
- Mann DH and Hamilton TD (1995) Late Pleistocene and Holocene paleoenvironments of the north Pacific coast. *Quaternary Science Reviews* 14: 449–471.
- Martinson DG, Pisias NG, Hays JD, Imbrie J, Moore TC Jr., and Shackleton NJ (1987) Age dating and the orbital theory of the ice ages: Development of a high-resolution 0 to 300,000-year chronostratigraphy. *Quaternary Research* 27: 1–29.
- Matmon A, Schwartz DP, Haeussler PJ, et al. (2006) Denali fault slip rates and Holocene-Late Pleistocene kinematics of central Alaska. *Geology* 34: 645–648.
- Melles M, Brigham-Grette J, Glushkova OY, Minyuk PS, Nowaczyk NR, and Hubberten HW (2007) Sedimentary geochemistry of a pilot core from Lake El'gygytyn – A sensitive record of climate variability in the East Siberian Arctic during the past three climate cycles. *Journal of Paleolimnology* 37: 89–104.
- Miller G, Alley R, Brigham-Grette J, et al. (2010) Arctic amplification: Can the past constrain the future? *Quaternary Science Reviews* 29: 1779–1790.
- Milne GA and Mitrovica JX (2008) Searching for eustasy in deglacial sea-level histories. *Quaternary Science Reviews* 27: 2292–2302.
- Muhs DR, Ager TA, and Begét JE (2001) Vegetation and paleoclimate of the last interglacial period, central Alaska. *Quaternary Science Reviews* 20: 41–62.
- Nowaczyk NR, Melles M, and Minyuk P (2007) A revised age model for core PG1351 from Lake El'gygytyn, Chukotka, based on magnetic susceptibility variations tuned to northern hemisphere insolation variations. *Journal of Paleolimnology* 37: 65–76.
- Paillard D, Labeyrie L, and Yiou P (1996) Macintosh program performs time-series analysis. *Eos Transactions, American Geophysical Union* 77: 379.
- Polyak L, Alley RB, Andrews JT, et al. (2010) History of sea ice in the Arctic. *Quaternary Science Reviews* 29: 1757–1778.
- Pushkar VS, Roof SR, Cherepanova MV, Hopkins DM, Ivanov VF, and Brigham-Grette J (1999) Paleogeographic and paleoclimatic significance of diatoms from Middle Pleistocene marine and glaciomarine deposits on Baldwin Peninsula, northwestern Alaska. *Paleogeography, Paleoclimatology, Paleoecology* 152: 67–85.
- Sher A, Alfimov A, Berman D, and Kuzmina S (2002) Attempted reconstruction of seasonal temperature for the Late Pleistocene arctic lowlands of north-eastern

- Siberia based on fossil insects. In: *Sixth Quaternary Environment of the Eurasian North (QUEEN) Workshop, Programme and Abstracts*, 50.
- Siegert MJ, Dowdeswell JA, Hald M, and Svendsen J (2001) Modelling the Eurasian ice sheet through a full (Weichselian) glacial cycle. *Global and Planetary Change* 31: 367–385.
- Siegert MJ and Marsait I (2000) Numerical reconstructions of LGM climate across the Eurasian Arctic. *Quaternary Science Reviews* 20: 1595–1606.
- Stauch G and Gualtieri L (2008) Late Quaternary glaciations in northeastern Russia. *Journal of Quaternary Science* 23: 545–558.
- Stroeve AP, Fabel D, Codilean AT, et al. (2010) Investigating the glacial history of the northern sector of the Cordilleran Ice Sheet with cosmogenic ^{10}Be concentrations in quartz. *Quaternary Science Reviews* 29: 3630–3643.
- Ward BC, Bond JD, and Gosse JC (2007) Evidence for a 55–50 ka (early Wisconsin) glaciation of the Cordilleran ice sheet, Yukon Territory, Canada. *Quaternary Research* 68: 141–150.
- Zamoruyev V (2004) Quaternary glaciation of north-eastern Asia. In: Ehlers J and Gibbard PL (eds.) *Quaternary Glaciations – Extent and Chronology, Part III: South America, Asia, Africa, Australasia, Antarctica*, pp. 321–323. Amsterdam: Elsevier.

Late Quaternary of the Southwest Pacific Region

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Introduction

The main land masses of the southwest Pacific region are the archipelago of New Zealand and the southeastern parts of Australia. To the south is the Southern Ocean, within which are several small sub-Antarctic islands. Southern parts of the region lie within the Southern Hemisphere belt of middle-latitude westerly winds, while northern parts lie within the influence of the subtropical high pressure belt. Glaciers formed on parts of New Zealand and southernmost Australia in the late Pleistocene during episodes of cold climate (glaciations). In the high mountain areas of New Zealand, much-reduced glaciers persisted into the Holocene and some of these glaciers still survive today.

The late Quaternary Period spans the last 130 000 years and comprises the late Pleistocene and Holocene epochs (Table 1). The Last Glaciation occurred between about 74 000 and 18 000 years ago and brought generally cooler conditions than prevail today. The coldest episodes (stadials) occurred in the early and late parts of the glaciation, separated by an interstadial period of generally less severe climate (Barrell, 2011; Suggate, 1990; Table 1). During stadials, glaciers formed in mountain areas of New Zealand, Tasmania, and the Snowy Mountains of southeastern Australia (Colhoun and Shulmeister, 2007). The Last Glaciation was succeeded by the Last Glacial–Interglacial Transition. This period of transition saw a steady improvement in temperatures, major adjustments of mountain and mountain-valley landscapes to the melting away of glaciers, the withdrawal of alpine and tundra vegetation to higher altitudes on mountain ranges, and the progressive expansion of forests. By the start of the Holocene, about 11 700 years ago, temperatures in New Zealand had risen to levels close to those of today (Alloway et al., 2007), and glaciers had retreated to the vicinities of the highest mountains (Gellatly et al., 1988). The climatic picture is less clear in Australia, due to difficulties in distinguishing between the effects of temperature and variable aridity on the biological indicators of climate (Turney et al., 2006).

This article describes the late Pleistocene glaciers of southern parts of Australia, and then examines the glaciers of New Zealand during the late Pleistocene and the Holocene. Brief mention is made of late Pleistocene glaciation on the main sub-Antarctic islands of the Australasian region; Auckland Island, Campbell Island and Macquarie Island (Figure 1).

In regard to the late Pleistocene and Holocene glacial history of the southwest Pacific, it is important to consider the dynamics of the plates that form the Earth's crust in this wide region. The boundary between the Indo-Australian Plate (commonly called the Australian Plate), and the Pacific Plate, lies beneath the Southwest Pacific Ocean, and passes through New Zealand on a northeast–southwest trend. As a result, New Zealand has a dynamic tectonic setting, with active volcanoes

in the central part of the North Island, recently uplifted ranges, and ongoing deformation and uplift of the Earth's crust, particularly in the South Island. In contrast, southern Australia and Tasmania lie on a continental platform, with landscapes that are much more stable than those of New Zealand. In the sub-Antarctic part of the region, Auckland and Campbell islands lie on a stable continental platform, away from the main locus of the active plate boundary through the New Zealand mainland islands. In contrast, Macquarie Island lies on a recently uplifted fragment of ocean floor crust, alongside the plate boundary.

Australia

The Snowy Mountains

The mainland of Australia, being relatively far north of the southern polar region, of relatively low altitude and, away from the coast, relatively dry, is an environment not conducive to the formation of glaciers, even during the coldest parts of the Last Glaciation. Only at Mt Kosciuszko in the Snowy Mountains (Figure 1) is there unmistakable evidence for late Pleistocene glaciers on the Australian mainland.

Mt Kosciuszko (2228 m; latitude 36°27'S) is Australia's highest mountain and the highest point on a broad range (massif) about 2000 m high and extending for about 120 km². The massif has ice-eroded cirque basins, some of which contain lakes, and there are also some short, ice-scoured valleys. In a few locations there are well-formed moraine loops marking the margins of former glaciers, notably at Blue Lake and Lake Cootapatamba (Figure 2).

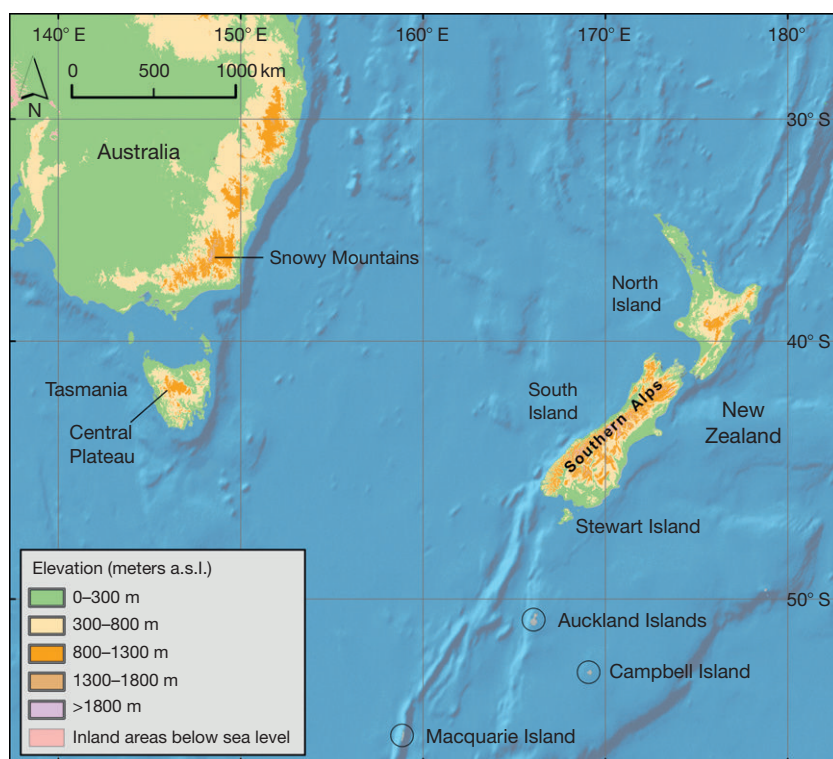
Since the nineteenth century, scientists have offered differing opinions on the extents of ice-age glaciers on the Kosciuszko massif, and when they reached their maximum size(s) (David et al., 1901; Galloway, 1963). Barrows et al. (2001) made a geomorphological map of glacial landforms on the Kosciuszko massif, which showed that cirque and short valley glaciers, at their maximum extent as indicated by well-defined glacial landforms, covered only about 15 km². The map also highlighted evidence for at least two glaciations having affected the massif. In order to determine when these glacier advances took place, Barrows et al. (2001) measured the ages of 24 boulders resting on the surfaces of moraines at Blue Lake and Lake Cootapatamba, using the terrestrial cosmogenic nuclide (TCN) dating method. Based on chemical changes produced by cosmic radiation that penetrates the Earth's atmosphere, the TCN dating method involves measuring the amounts of cosmogenic isotopes in minerals from rock surfaces, in this case beryllium-10 (¹⁰Be) in quartz. The TCN ages given in this article are as reported in the original papers that are cited, and may not accord to present calibration standards. In contrast, radiocarbon calendar ages are recalculated according to

Table 1 Late Pleistocene and Holocene glacier activity in the SW Pacific region

<i>Geological epoch</i>	<i>Climate event</i>	<i>MIS stage</i>	<i>Approximate calendar age ($\times 1000$ years)</i>	<i>New Zealand</i>		<i>Snowy Mountains</i>	<i>Tasmania</i>
Holocene	Postglacial	1	0–11.7	Aranui Postglacial	'Neoglacial' events near modern glaciers	No glaciers	No glaciers
Late Pleistocene	Last Glacial–Interglacial Transition	2	11.7 to ~18.0		Late-Glacial ice advances		
	Last Glaciation	Late	~18.0 to ~28	Otira Glaciation	Late Otira Glaciation	Late Kosciuszko Glaciation	Episodes of glacier activity, no formal regional names applied
		3	~28 to ~59		Mid-Otira interstadial		
		Early	~59 to ~74		Early Otira Glaciation	Early Kosciuszko Glaciation	
Middle Pleistocene	Last Interglacial	5	~74 to ~130	Kaihinu Interglacial – indications of ice advance during cold stadials		No evidence preserved for glacier activity	Uncertain
	Penultimate Glaciation	6	Older than ~130	Waimea Glaciation – glaciers more extensive than in late Pleistocene			Glaciers more extensive than in late Pleistocene

The Marine Oxygen Isotope Scale (MIS) stages are from [Martinson et al. \(1987\)](#).

Source: Barrell DJA (2011) Quaternary Glaciers of New Zealand. In: Ehlers J, Gibbard PL, and Hughes PD (eds.) *Quaternary Glaciations – Extent and Chronology: A Closer Look. Developments in Quaternary Science* 15: 1047–1064. Amsterdam: Elsevier; Barrows TT, Stone JO, Fifield LK, and Cresswell RG (2001) Late Pleistocene Glaciation of the Kosciuszko Massif, Snowy Mountains, Australia. *Quaternary Research* 55: 179–189; Barrows TT, Stone JO, Fifield LK, and Cresswell RG (2002) The timing of the Last Glacial Maximum in Australia. *Quaternary Science Reviews* 21: 159–173; Colhoun EA, Hannan D, and Kiernan KW (1996) Late Wisconsin Glaciation of Tasmania. *Papers and Proceedings of the Royal Society of Tasmania* 130: 33–45.

**Figure 1** Location map. Bathymetry is the WorldSat Color Shaded Relief Model, topographic relief is from GNS Science digital archives.

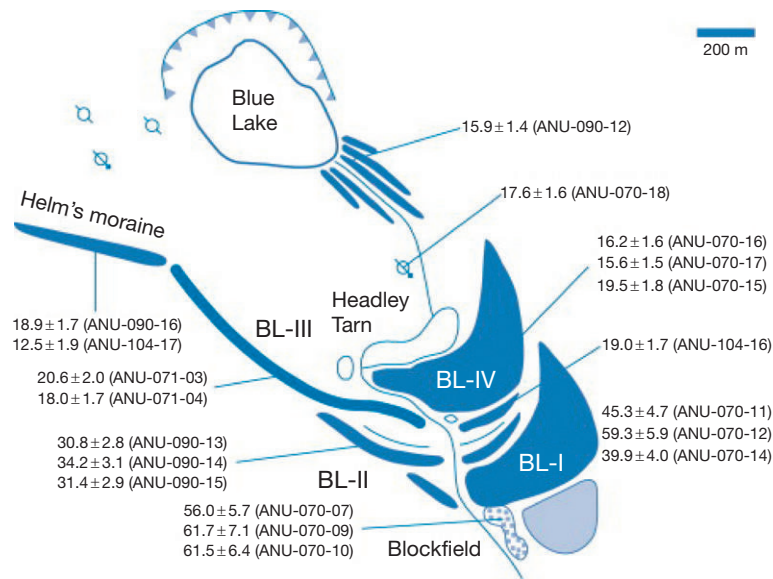


Figure 2 Last Glaciation moraines at Blue Lakes, Mt Kosciuszko area, Snowy Mountains, Australia. Reproduced from Barrows TT, Stone JO, Fifield LK, and Cresswell RG (2001) Late Pleistocene Glaciation of the Kosciuszko Massif, Snowy Mountains, Australia. *Quaternary Research* 55: 179–189, with permission (Figure 8 of Colhoun and Shulmeister, 2007).

the 2009 IntCal09 calibration (Reimer et al., 2009). The mapping together with the TCN dating indicates that the first and also the most extensive glaciation, known locally as the Early Kosciuszko Glaciation, produced a single glacier advance (Snowy River Advance) sometime between ~65 000 and ~55 000 years ago. The Late Kosciuszko Glaciation comprised three glacier advances, progressively less extensive. The first one, named the Headley Tarn Advance, occurred sometime between ~35 000 and ~30 000 years ago. Next was the Blue Lake Advance, sometime between ~21 700 and ~17 500 years ago, and last was the Mt Twynam Advance, sometime between ~18 200 and ~15 400 years ago (Barrows et al., 2001). Radiocarbon (^{14}C) dating of lake-floor sediments in Blue Lake shows that sometime between 16 770 and 15 060 calendar years ago, the ice had gone and the lake was in existence (from Barrows et al., 2001; recalculated). These results indicate that glaciers disappeared promptly from the Kosciuszko massif at the end of the Last Glaciation, and that Kosciuszko has remained ice-free ever since (Barrows et al., 2001, 2002).

Tasmania

Tasmania today is a large island separated from the Australian mainland by Bass Strait. The strait is shallow and was dry land during glaciations, when sea level was as much as 120 m lower than it is today. At such times, Tasmania was a large cape of the Australian mainland. Tasmania lies in the path of the mid-latitude westerly wind belt, and has a temperate maritime climate. The prevailing westerly winds bring wetter and cloudier conditions to the southwestern part of the island, and drier rain-shadow conditions to the northeast. The landscape is hilly with a few mountain ranges, and an extensive tableland, the Central Plateau, in northern central Tasmania. The highest peak is Mt Ossa (1617 m) on the western edge of the Central Plateau.

During the Last Glaciation, small ice caps, cirque glaciers and small valley glaciers formed on and around the highest mountains and ranges in Tasmania, and a small ice sheet grew on the western side of the Central Plateau (Figure 3). The extent of late Pleistocene ice was considerably less than during middle and early Pleistocene glaciations (Augustinus, 1999; Colhoun, 2004; Colhoun and Fitzsimons, 1990; Kiernan, 1990). In contrast to the Snowy Mountains, there is not an established set of names for glacial and interglacial events across Tasmania – instead, local names have been applied in areas where detailed studies have been done. For this reason, only the generalized terms early Last Glaciation and late Last Glaciation are used in the description below.

In southwestern Tasmania, late Pleistocene glacial features, such as ice-smoothed topography, cirques and moraines, abound in the vicinity of peaks that are more than about 1000 m above sea level, the southernmost examples being at Pinders Peak (latitude 43°30'S) in Southwest National Park. Northeastward across Tasmania, the altitude of mountains with associated glacial features becomes progressively higher (Colhoun et al., 1996; Peterson and Robinson, 1969). In northeastern Tasmania, the Ben Lomond plateau, at about 1400 m above sea level, lacks glacial features, except for a small cirque moraine at Tranquil Tarn (41°37'S) in a shaded south-east facing position at the southern tip of the plateau (Barrows et al., 2002). The eastward rise in altitude of glacial features mimics the present-day gradient of westerly-wind precipitation that decreases eastwards across Tasmania.

TCN dating is advancing the state of knowledge considerably with regard to late Pleistocene Tasmanian glaciers. It is turning out that, similarly to the Snowy Mountains, late Pleistocene ice in Tasmania was more extensive during early parts of the Last Glaciation. For example, in the eastern part of the Central Plateau, morainic boulders lying outside of well-defined late Last Glaciation moraines returned ages ranging

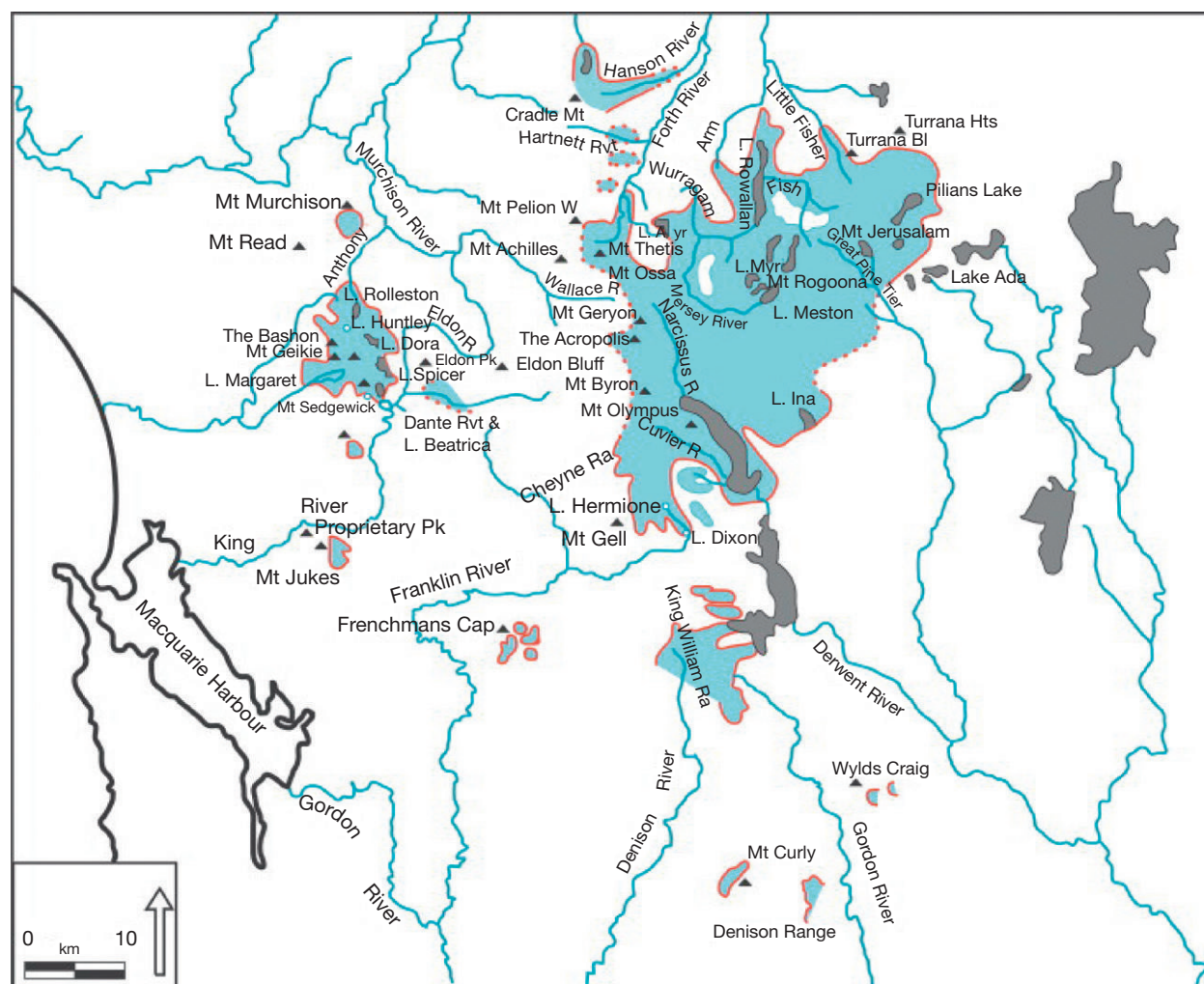


Figure 3 The extent of glaciers in west-central Tasmania during the latter part of the Last Glaciation. Lakes are in gray, glaciers are in light blue, and ice limits are in red, dashed where uncertain. Reproduced from Colhoun EA, Hannan D, and Kiernan KW (1996) Late Wisconsin Glaciation of Tasmania. *Papers and Proceedings of the Royal Society of Tasmania* 130: 33–45, with permission (Figure 7 of Colhoun and Shulmeister, 2007).

from ~53 000 to ~24 000 years (Fink et al., 2000). In southern Tasmania, a study by Mackintosh et al. (2006) at Broad Valley in Mt Field National Park obtained ages of between ~44 000 and ~40 000 years for moraine boulders beyond the late Last Glaciation moraines, while at Schnells Ridge on Mt Anne in southwestern Tasmania, Kiernan et al. (2004) reported ages of between ~141 000 and ~39 000 years for extensive moraines outboard of the late Last Glaciation moraines. These findings are in keeping with interpretations from Chamouni in the King River valley in the West Coast Range, where Fitzsimons et al. (1993) obtained a ^{14}C age of 48 700 years for wood overlying the Chamouni glacial deposits, implying that the glacial sediments were deposited during early phases of the Last Glaciation, or during a preceding glaciation.

Dating has also provided some surprises regarding the moraine sequences formed late in the Last Glaciation. Near Lake Margaret in the West Coast Range, the largest moraine ridge in Tasmania, known as the Hamilton End Moraine, was generally thought to mark the late Last Glaciation ice limit (Colhoun, 1985), but TCN dating by Barrows et al. (2002) has returned

ages ranging from ~350 000 to ~190 000 years. This moraine therefore formed in the middle Pleistocene, whereas the late Pleistocene ice limit was about 1 km up-valley at the Poets Hill moraines, which are between ~19 000 and ~16 000 years old based on TCN measurements by Barrows et al. (2002).

In the West Coast Range at Dante Rivulet in the King River catchment, a buried soil horizon that is ~21 000 ^{14}C years old is overlain by glacial outwash sands and gravels containing wood fragments ~19 000 ^{14}C years old (Gibson et al., 1987). This indicates that late Last Glaciation ice appeared in this catchment between ~25 000 and ~22 000 calendar years ago. At Lake St Clair on the Central Plateau, TCN dating of late Last Glaciation terminal moraines enclosing the lake gave a wide spread of ages ranging from ~30 000 to ~17 000 years (Barrows et al., 2002). These ages are difficult to reconcile with ^{14}C dating of lake bed sediments from Lake St Clair near the terminal moraine complex, which indicates that the glacier retreated from its moraine complex, initiating the lake, ~18 450 ^{14}C years ago (~22 000 calendar years ago; Hopf et al., 2000). These

conflicting results indicate that the timing of late Pleistocene events at Lake St Clair needs further investigation.

TCN dating of several glacier systems across Tasmania indicates that late Last Glaciation terminal moraines formed between $\sim 20\,000$ and $\sim 17\,000$ years ago, with a few instances of outer parts of the moraine complexes dating from as long ago as $\sim 45\,000$ years (Barrows et al., 2002; Kiernan et al., 2004; Mackintosh et al., 2006). This suggests that in addition to the main late Last Glaciation ice maxima between $\sim 25\,000$ and $\sim 18\,000$ years ago, there may have been ice advances of similar extent in the middle part of the Last Glaciation, about $45\,000$ years ago. The TCN dating results also indicate that the main deglaciation occurred between $\sim 18\,000$ and $\sim 16\,000$ years ago, and that no ice survived in cirques after

about $15\,000$ years ago. In keeping with the results from the Kosciuszko massif, it appears that Tasmanian glaciers melted away rapidly at the end of the Last Glaciation, and Tasmania has been ice-free ever since (Barrows et al., 2002).

New Zealand

In New Zealand, small modern glaciers occur on Mt Ruapehu in the North Island at about latitude $39^{\circ}17'S$. In central to southern parts of the South Island's Southern Alps, modern glaciers occur from about latitude $42^{\circ}90'S$ to $45^{\circ}02'S$, with a few small perennial ice patches farther to the north and south (Chinn, 1995; Figure 4).

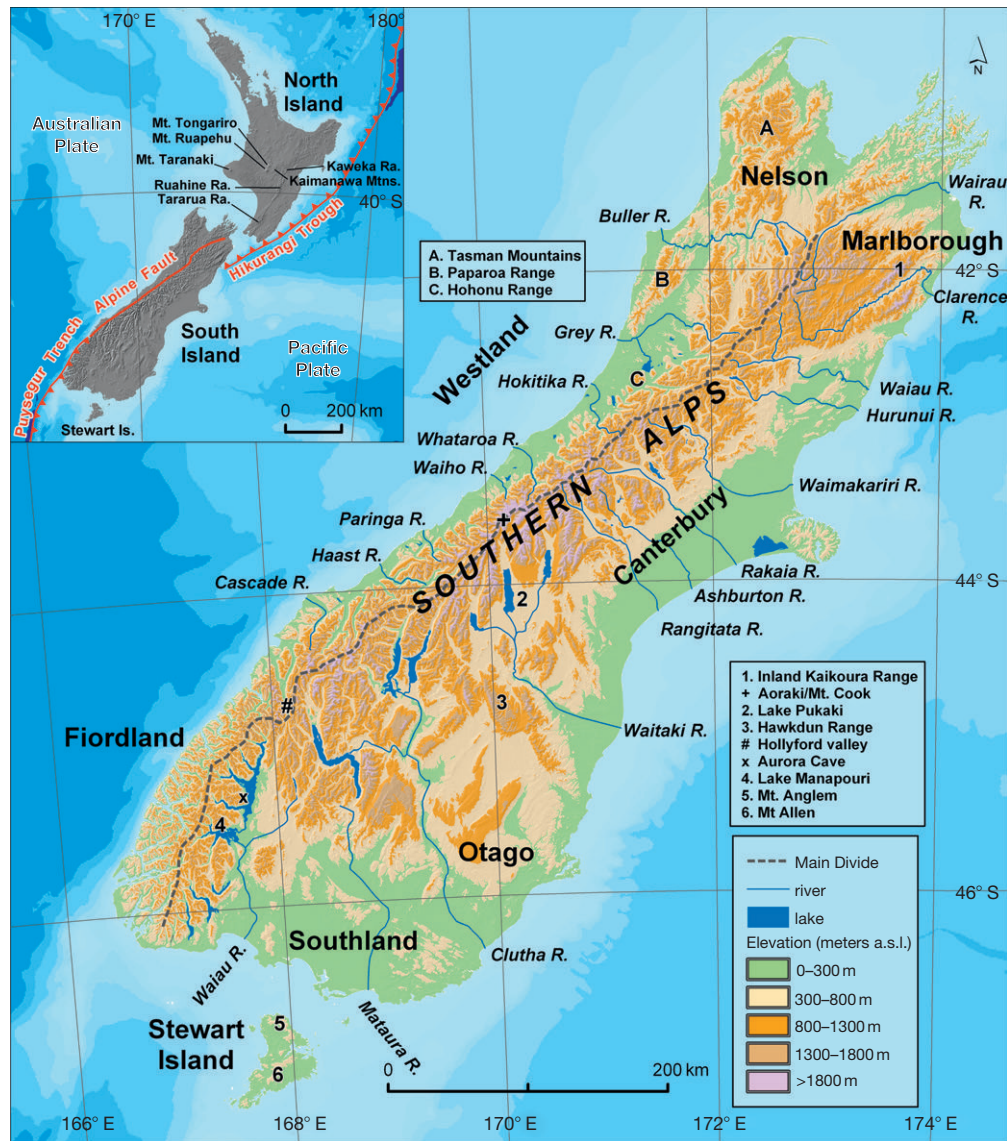


Figure 4 New Zealand location map (inset) and present-day topography of the South Island, showing rivers that were important conduits for ice-age meltwater, and locations mentioned in this article. The Australia–Pacific plate boundary (inset) comprises the Puysegur and Hikurangi subduction thrusts (teeth point down-dip) and the strike-slip Alpine Fault. Bathymetry is from GNS Science digital archives, with color changes at depths of 250, 750, 1500, 2500, and 4000 m. The onland digital elevation model is generated from 20 m-interval contours and spot heights from Land Information New Zealand digital data. Crown copyright reserved.

Like Tasmania, New Zealand lies in the path of the mid-latitude westerly-wind belt. The interaction between the prevailing westerly winds and the mountain ranges of New Zealand produces a strong orographic precipitation pattern, especially across the Southern Alps. As a result, the climate is very wet on the western side of mountains and much drier to the east. The modern glaciers occur where the mountains are high and the annual precipitation is relatively large.

During late Pleistocene glaciations, immense ice fields formed in the Southern Alps, and glaciers developed on many of the taller ranges of the South Island. There were also small glaciers on the higher ranges of the North Island. In New Zealand, the Last Glaciation is known as the Otira Glaciation (Suggate, 1990). A major ice advance early in the glaciation, approximately 65 000 years ago, is known as the Early Otira Glaciation, while later resurgences of the glaciers, between about 30 000 and 18 000 years ago, mark the Late Otira Glaciation (Alloway et al., 2007; Suggate and Almond, 2005; Figure 5). There is evidence suggesting that there may also have been major ice advances during the mid-Otira period, between about 50 000 and 30 000 years ago, preserved in Aurora Cave in Fiordland (Williams, 1996; see Figure 4), and in South Westland (Almond et al., 2001). Evidence has also been presented for major ice advances about 80 000 years ago, towards the end of the Last (Kaihinu) Interglacial period (Sutherland et al., 2007).

North Island

At the southern end of the Taupo Volcanic Zone are three tall, active volcanoes, from northeast to southwest, Mounts Tongariro (1967 m), Ngauruhoe (2287 m) and Ruapehu (2797 m) (Figure 4). Only Mt Ruapehu, the North Island's highest mountain, carries glaciers today, but on the mid to lower flanks of both Ruapehu and Tongariro are paired moraines formed by valley glaciers that flowed from ice caps on these mountains during late Pleistocene glaciations (Lee et al., 2011; Mathews, 1967; McArthur and Shepherd, 1990; Townsend et al., 2008; see Figure 5).

Although there is little direct information on the ages of the moraines themselves, radiometric dating of the volcanic rocks shows that much of Mt Tongariro is older than about 65 000 years (Hobden et al., 1996), and so has been in existence through the early and later phases of the Otira Glaciation. Accordingly, it has spectacular glacial valleys and lateral moraines, though somewhat modified by post-glacial volcanic activity. Mt Ngauruhoe is a near-perfect volcanic cone that has formed in the head of a formerly glaciated valley of the Tongariro massif. Ngauruhoe itself shows no sign of having been glaciated, which is consistent with its age of late Holocene (less than 6500 years; Lee et al., 2011). Mt Ruapehu has several well expressed glacial valleys on its eastern flank (Figure 6). The rocks forming the volcano's southern and eastern slopes are mostly between ~150 000 and ~120 000 years old (Gamble et al., 2003). This indicates that the glacial features, which must be younger than the rocks on which they rest, are probably no older than late Pleistocene. The western flanks of Ruapehu are mainly formed of lava flows younger than 50 000 years, but with many paired glacial moraines on the lowermost slopes. The extent to which these moraines were

formed by glaciers that flowed over these lava sheets during the Late Otira Glaciation, or whether some were formed earlier, and lava flows encroached to their margins subsequently, is yet to be established.

To the west, Mt Taranaki, also known as Mt Egmont (2518 m), is a near-perfect volcanic cone. The oldest rocks exposed on its slopes are about 18 000 years old, indicating that the volcano in its present form is primarily a post-glacial feature. No evidence for past glaciers has been found (Townsend et al., 2008).

Based on elevation and aspect, localized glaciers may possibly have formed on parts of the North Island axial ranges (Figure 5) during the Otira Glaciation, such as the highest parts of the Kaweka Range (~1700 m), the Kaimanawa Mountains (~1700 m), and the Ruahine Range (~1700 m), but no well-defined glacial landforms are evident.

On the Tararua Range, farther southwest along the North Island axial range, the existence of late Pleistocene ice-age glaciers is indicated by u-shaped valleys and cirque-like amphitheater basins in valleys draining the higher parts of the range (Adkin, 1911). The Park River valley, draining south off a peak called Arete (1505 m; 40°45'S), contains the best known example of a lateral moraine in the Tararua Range (Brook, 2009a).

South Island

Geographically, the name Southern Alps is applied to the main belt of high mountains that runs from southeast Nelson/west Marlborough to the Hollyford valley (Figure 4). Strictly speaking, mountain ranges around the periphery of the Southern Alps are regarded as entities that are separate from the Southern Alps, as is the Fiordland massif that forms the southwestern corner of the island.

Ranges in the northern parts of the South Island

Late Pleistocene glaciers formed on many of the taller mountain ranges in the northern part of the South Island, lying at the periphery of the Southern Alps, in northern Westland, Nelson, and Marlborough (Figures 4 and 5).

In the northwestern corner of the South Island are the Tasman Mountains, whose highest peaks range from 1500 to as much as 1800 m. Although ice-free today, the Tasman Mountains have numerous u-shaped valleys and cirque basins. These indicate that during glaciations, minor ice caps formed and glaciers extended down many of the mountain valleys (Figure 5). Terminal moraines are rarely preserved, and only tentative estimates can be made of how far the glaciers extended (Thackray et al., 2009). Recent studies have placed some limits on the extents of Otira Glaciation ice, both early in the glaciation at about 65 000 years ago, as well as during its late phase, culminating about 18 000 years ago (McCarthy et al., 2008), and the subsequent ice recession (Shulmeister et al., 2001, 2005).

Near the west coast, several of the higher ranges have glacial topography in the form of southeast-facing cirques and localized u-shaped valleys, such as in the Paparoa Range, where the highest peaks range from 1200 to 1500 m. An example from the Hohonu Range is illustrated in Figure 7, where the highest peaks are between 1200 and 1300 m.

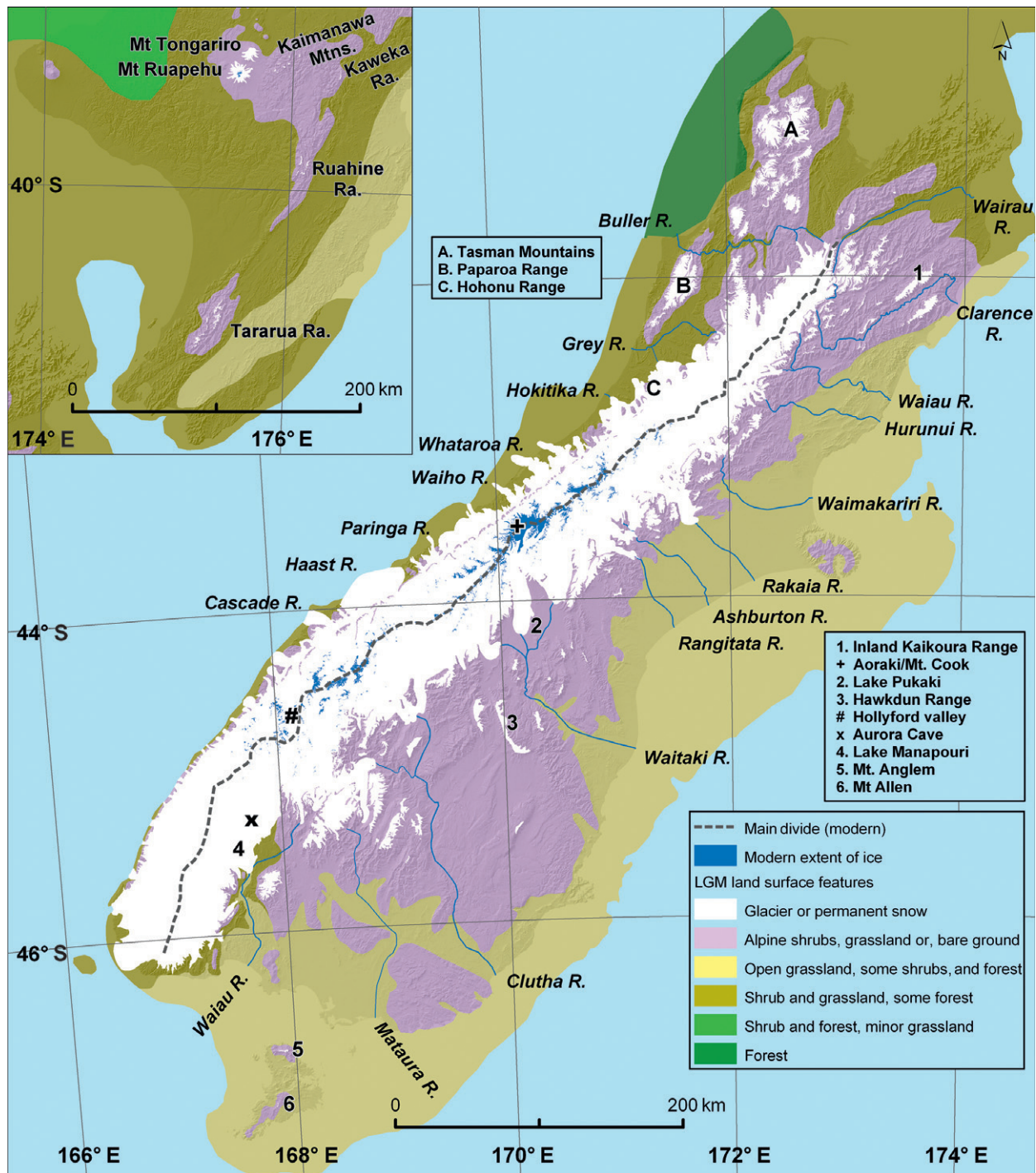


Figure 5 An interpretation of the distribution of ice and the main vegetation zones at the maximum of the Late Otira Glaciation, between about 20 000 and 22 000 calendar years ago. The land extent approximates the 120 m isobath on the modern sea floor, reflecting lower global sea level at that time. Vegetation zonation, modified from Alloway et al. (2007), was originally based on McKinnon (1997) and Newnham et al. (1999), adjusted in detail to the topography. Late Otiran ice extent has been modified slightly from Alloway et al. (2007), as described by Barrell (2011). The ice map is generalized in its interior; it is likely that more bare rock was exposed than is depicted here. Modern ice is from Land Information New Zealand 1:50 000 scale digital data, and the present-day digital elevation model is generated from 20 m-interval contours and spot heights from Land Information New Zealand digital data, Crown copyright reserved.



Figure 6 Mt Ruapehu, the North Island's highest mountain, is an active volcano although much of its bulk dates from the Middle Pleistocene. As a result, Ruapehu carries many imprints of late Pleistocene glaciation. Just to the right of center, paired lateral moraines of the Wahianoa valley, probably of Late Otiran age, descend southeast from the upper slopes of the mountain. Although obscured by winter snow at the time this photograph was taken, small modern glaciers survive on the highest parts of the mountain. Mt Ngauruhoe, also covered in snow, lies at upper right. Photo: D.B. Townsend, GNS Science.



Figure 7 Ice-sculpted glacier cirque basins developed on granitic bedrock in the Hohonu Range, northern Westland. Photo: D.L. Homer, GNS Science CN10184/24.

In the ranges of central to eastern Marlborough, scattered cirque-like features near the crests of the highest peaks suggest local ice cover during glaciations. Cirque-like and moraine-like features are best developed in the Inland Kaikoura Range, where peak heights range from about 1800 m to a maximum of 2885 m at Tapuae-o-Uenuku (Bacon et al., 2001; Figure 5).

The Southern Alps

The Southern Alps run from southeast Nelson southwest to the Hollyford valley, at the northern margin of Fiordland (Figure 4). Along the axis of the Southern Alps is the Main Divide, which defines the boundary between westward- and eastward-draining catchments. Today, glaciers are confined to

the vicinity of higher parts of the Southern Alps, where many peaks exceed 2000 m. The largest modern glaciers occur near Aoraki/Mt Cook (3754 m; [Figure 5](#)).

During late Pleistocene glaciations, ice formed on the Southern Alps, with glacier tongues draining to the northwest and southeast off the mountain chain ([Figure 5](#)). Many of the northwest-flowing glaciers extended out into the lowlands of Westland, producing well-defined glacier troughs rimmed by belts of moraine. In contrast, many of the southeast-flowing glaciers terminated within mountain valleys or basins floored by hard bedrock and, as a result, glacier troughs are not so widely developed and moraines are not so extensively preserved ([Barrell et al., 2011](#)). An outstanding exception is the Pukaki glacier system ([Figures 4 and 5](#)), which drained southeast from the highest part of the Southern Alps into a broad inland basin. Very well preserved sets of moraines and associated meltwater outwash plains surround the Pukaki glacier trough ([Figures 8 and 9](#)). The good preservation of these landforms, and the fact that the moraines are very bouldery and therefore well suited to TCN dating methods, has meant that much effort has been directed to dating the Pukaki moraine sequences using ^{10}Be .

Examining the Pukaki glacier system in more detail, the ages of moraines measured by TCN dating using ^{10}Be indicate that the Late Otira Glaciation ended here about 18 300 years ago as the Pukaki glacier began a sustained retreat ([Putnam et al., 2010a; Schaefer et al., 2006](#)). A resurgence of the glacier is recorded by the Birch Hill moraine ([Figure 10](#)), one of the better expressed late-glacial moraines of New Zealand ([McGregor, 1967; Porter, 1975](#)). Formation of this moraine belt culminated approximately 13 000 years ago, based on TCN measurements, followed by rapid recession ([Putnam et al., 2010b](#)). The age of the Birch Hill moraines coincides with the climax of the Antarctic Cold Reversal climatic event recorded in core samples from the Antarctic ice sheet ([Putnam et al., 2010b](#)).

The Pukaki catchment contains several of the largest modern glaciers in New Zealand ([Figure 11](#)), notably the Tasman, Mueller, and Hooker glaciers. Belts of moraines around the terminus of each glacier have been investigated using a variety of dating methods. Attempts have been made to date these moraines using relative indicators of age, such as the sizes of lichens on rocks (lichenometry; [Burrows, 1973; Lowell et al., 2005; Winkler, 2004](#)), thickness of weathering rinds on rocks (weathering-rind dating; [Birkeland, 1982; Chinn, 1981](#)), and rock hardness determined by the Schmidt hammer method ([Winkler, 2005](#)). These methods are hindered by difficulties in calibrating the measurements to a geological clock, and the variations in results have led to much debate ([Gellatly et al., 1988; Winkler and Matthews, 2010](#)). There has also been success in obtaining radiometric dates from these moraine sequences. Deposits within the lateral moraine belts were exposed as the glaciers shrank during the late twentieth century, revealing fossil soils buried in the moraines. These soils provide a prehistoric record of glacier recession and resurgence. Soil formed on the moraines during ice recession, and was subsequently buried by debris during resurgence of the glacier. Dating of the soils using ^{14}C has provided a chronology of glacier fluctuations ([Burrows, 1989; Röthlisberger, 1987](#)). More recently, TCN dating of large boulders on the surface of the moraines using ^{10}Be has added to the picture of advance and retreat of these glaciers ([Schaefer et al., 2009](#)). The outermost preserved Holocene moraines of the Tasman and Mueller glaciers are about 6000 years old, and lie only about 0.8 km beyond historical ice limits (in New Zealand the historical period begins about 1860 A.D.). As these glaciers were respectively about 28.5 and 13.7 km long when they were at their historical maxima, this means that these glaciers have been no more than between 3% and 6 % longer during the last 6000 years. Thus, the picture that is emerging for the modern glaciers of the Pukaki catchment is that during the last 6000 years, the glacier termini have fluctuated back and forth in about their present locations.



Figure 8 Moraines and meltwater outwash channels on the eastern (true left) margin of the Pukaki glacier trough. Compare to the geomorphological map in [Figure 9](#). Note the distinctively subdued form of the 'Balmoral' (Early Otiran) moraines to the right, compared to the more sharply expressed 'Mt John' (Late Otiran) moraines and outwash terraces in the center. The rather featureless belt of 'Tekapo' moraine close to the lake represents the final advance of the Last Glaciation and subsequent glacier melt-back that began about 18 300 years ago ([Putnam et al., 2010a](#)). Photo: G.H. Denton.

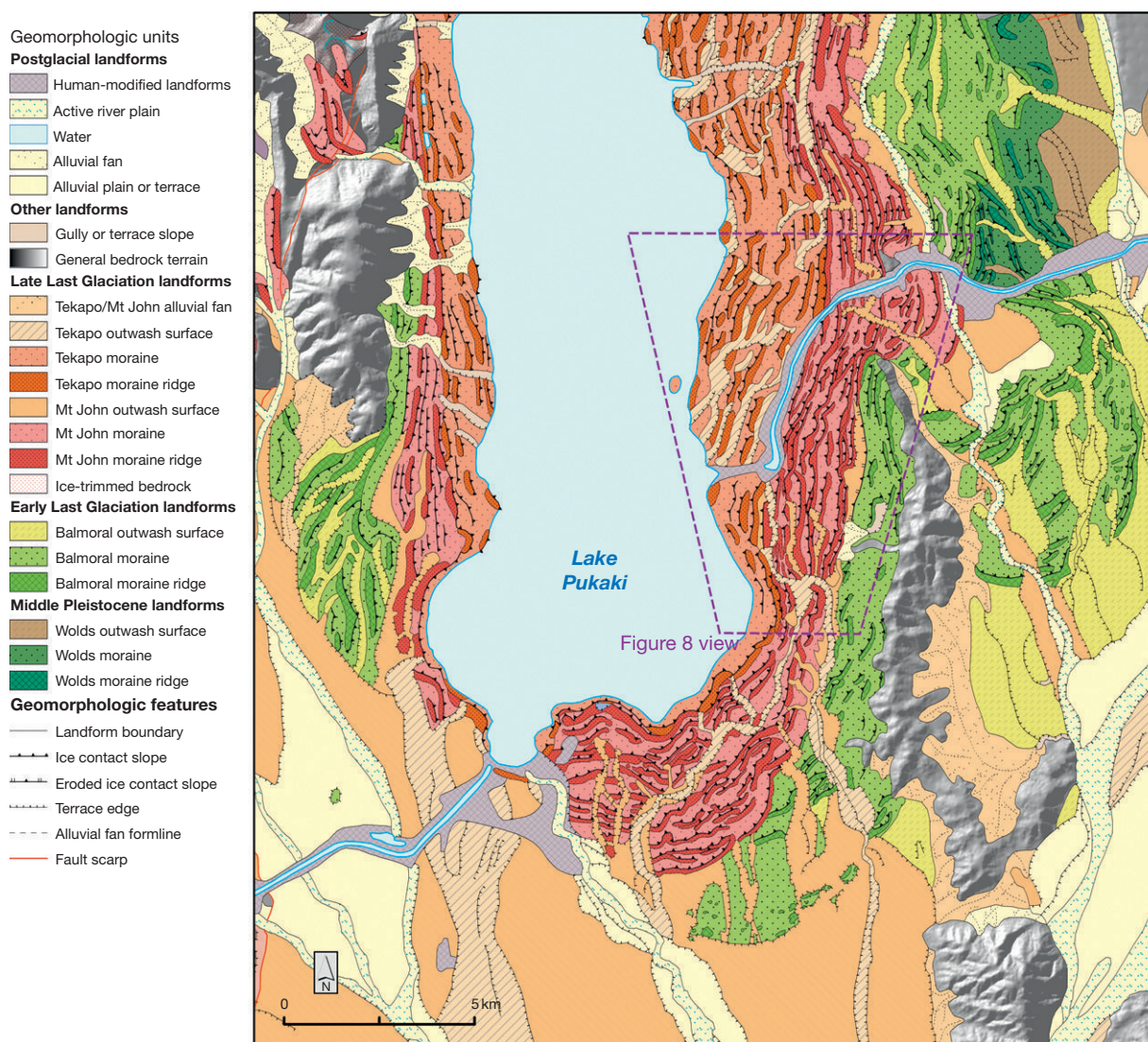


Figure 9 Geomorphology map of the lower reaches of the Lake Pukaki basin, which is an ice-age glacier trough. The map is generated from a Geographic Information System geomorphological map dataset produced for the central part of the South Island (Barrell et al., 2011). The map was compiled for presentation at a scale of 1:100 000, so many details of the landforms are simplified in order to be legible at that scale. A feature of this map is the detailed depiction of landforms, such as ridges of glacier moraine and the locations of river-cut terrace edges. These are factual observations whereas the colors on the map depict an interpretation of the ages of the landforms. The hypotheses that are represented by the coloring on the map can be tested by applying dating methods to the moraines.

Fiordland

The Fiordland massif forms the southwestern corner of the South Island. It is for the most part formed in hard crystalline plutonic and gneissic rocks, and is deeply dissected by steep-sided u-shaped valleys headed by innumerable cirques, with an array of lakes and fiords around the margins of the massif. The massif is highest towards its northern end, with many peaks taller than 1800 m, and numerous present-day cirque glaciers (Augustinus, 1992). On its western (seaward) side, the massif has no coastal lowland and the seafloor descends steeply down the continental slope. In contrast, to the south offshore is a broad continental shelf, and to the east are piedmont basins and valleys (Figure 4).

It is likely that ice on Fiordland approximated a cap during the maxima of the Otira Glaciation (Figure 5). To the west,

many of the outlet glaciers would have reached the sea at the times of glacial maxima/sea level minima, with glacier trough morphologies and submarine outwash fan features evident in fiord and continental slope bathymetries (Barnes, 2009). To the south and east, the glaciers terminated on land. Many of the Late Otiran glacier troughs are occupied by lakes, bordered by moraine and outwash sequences (Figure 12). In places on the south coast, these moraine and outwash sequences are intersected by uplifted coastal terraces (Turnbull et al., 2010).

Little is known of late-glacial or Holocene ice limits because relatively few moraines are preserved in the mid- to upper parts of the catchments. For the most part, the late Pleistocene glacial landforms of Fiordland are dominated by ice-sculpted bedrock topography (Figure 12).



Figure 10 On the lowest flanks of the northeast side of the Tasman River valley, lateral ridges of the Birch Hill moraine (lower right) lie about 500 m below the Late Otiran moraines (center right). The Birch Hill moraines are about 13000 ± 300 years old (Putnam et al., 2010b). Photo: B.G. Andersen.

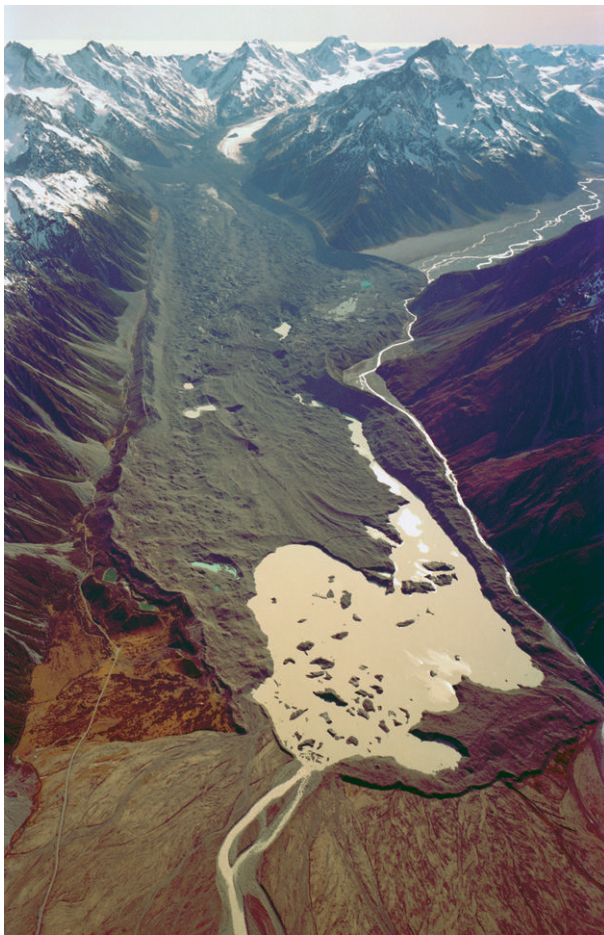


Figure 11 The Tasman Glacier is New Zealand's longest glacier, its lower reaches covered in a veneer of rock debris. The glacier shrank in height through the early to middle parts of the twentieth century, and Tasman Lake began to form in the late 1970s. Intricately braided outwash plains emanate from the moraine complex. Light-colored outwash areas

Southeastern ranges of the South Island

Several of the highest mountain ranges to the east of the Southern Alps carry geomorphological indicators of former ice cover. These include southeast-facing cirques and cirque-like basins (Figure 13), and over-widened, in some cases u-shaped, upper reaches of main valleys in the ranges. In most cases, it is likely that any such ice-related features were last occupied during the Late Otira Glaciation, though undoubtedly most of these features were initiated during earlier glaciations (e.g., Brook et al., 2008). Notable examples can be seen in the Torlesse Range ($43^{\circ}15'S$), southern Two Thumb Range ($43^{\circ}50'S$), Benmore Range ($44^{\circ}25'S$), Hawkdun Range ($44^{\circ}45'S$), Garvie Mountains ($45^{\circ}30'S$), and Takitimu Mountains ($45^{\circ}42'S$).

Stewart Island

Stewart Island rises from a shallow area of the continental shelf and would have comprised hills surrounded by a coastal plain at times of lower sea levels. No glaciers exist today, but small cirques lie on the southeast sides of two of the highest peaks. Mt Anglem ($46^{\circ}45'S$; ~ 980 m) has two cirques, each containing a small pond impounded by moraine (Allibone and Wilson, 1997). Assigned a Holocene age by Turnbull and Allibone (2003), a Late Otiran age seems more likely, judging by a lack of well-preserved moraines farther downslope. At Mt Allen ($47^{\circ}05'S$; 750 m), a well-defined ridge on its southeast face is thought to be a moraine (Allibone and Wilson, 1997; Brook, 2009b), and is probably of Late Otiran age.

have been part of the active Tasman River floodplain within historic time (since the late 1800s). At center left the Blue Lakes are enclosed within vegetated moraine ridges that range in age from about 900 years (inner ridges) to 1800 years (outer ridges) (Schaefer et al., 2009). Photo: D.L. Homer, GNS Science CN22243, taken April 1992.



Figure 12 A view from the rugged mountains of Fiordland southeast across the Manapouri glacial trough. The lake is studded with islands of ice-shorn bedrock and enclosed by Late Otiran moraines and outwash middle left to right. Photo: D.L. Homer, GNS Science CN10405/5.



Figure 13 Cirque basins at the crest of the Hawkdun Range, North Otago, eastern South Island. These basins have been scooped out by the action of ice during glaciations. The well-defined morphology of the basin floors suggests that glaciers were last active here during the Late Otira Glaciation. Photo: D.L. Homer, GNS Science CN42857/16.

Sub-Antarctic Islands

The largest sub-Antarctic islands of the SW Pacific region are the Auckland Islands, Campbell Island and Macquarie Island. Administratively, the first two are territories of New Zealand, while Macquarie Island is an Australian territory. There is no resident human population on these islands, but the respective governments maintain bases on these islands for purposes that include conservation work and scientific research.

The Auckland Islands ($50^{\circ}50'S$), about 626 km^2 in extent, are the deeply eroded remnants of an Oligocene–Miocene basalt

shield volcano. Many hills rise to between 400 and 600 m above sea level and the highest point is Mt Dick at 705 m. Mean annual temperature near sea level is between 7.5° and 8.1° (McGlone, 2002). The island has abundant evidence of past glacial activity in the form of u-shaped valleys, hanging valleys, moraine-dammed lakes, and cirques (McGlone, 2002). The full extents of ice have yet to be established. There remains the question of whether some of these ice-shaped landforms may have originated during middle Pleistocene glaciations.

Campbell Island/Motu Ihupuku, as it is officially named, lies at $52^{\circ}32'S$ and is the eroded remains of a Miocene volcano.

The island is 113 sq km² in area. Several peaks rise to between 400 and 500 m, with Mt Honey marking the island's highest point at about 570 m. Close to sea level, mean annual temperature is 6.8° (McGlone, 2002). The island has extensive smoothed slopes and broad-floored valleys, the lower reaches of which are now coastal inlets. The landscape has hallmarks of having been shaped by glaciers, although this interpretation has been a subject of debate amongst previous researchers (McGlone, 2002).

Macquarie Island (54°40'S) is long and thin, aligned north-south on the emergent crest of the submarine Macquarie Ridge, and is 129 sq km² in area. Geologically, it is a recently uplifted slice of ocean floor rocks, lying along the Australia-Pacific plate boundary. The island would have been much longer to the north and south, but not much wider, under lowered sea levels during glaciations. Macquarie Island has an irregular plateau-like top, between 200 and 300 m above sea level. Holocene raised beaches around much of the coast of the island indicate continuing uplift. It is not known when the island rose sufficiently high to be emergent above interglacial sea levels, such as today. It is not inconceivable that the island as we see it today is entirely a late Pleistocene feature. The setting of the island, in the Southern Ocean to the south of the Sub-Antarctic Front, results in a cold climate, with mean annual temperature of 4.8°. Opinions have varied on the past extent of glaciations. As summarized by McGlone (2002), earlier workers favored ice cover to some extent on the island, while later workers have favored little or no development of glaciers. The debate revolves around the interpretation of landforms, as to whether their origin is glacial or due to other processes, and is yet to be resolved.

Discussion

The highest mountains of New Zealand intersect the modern snowline, resulting in glaciers that survive in the present-day interglacial climate. During cold episodes during the late Pleistocene, notably in the Last Glaciation, snowlines dropped sufficiently low to allow glaciers to form on many mountainous areas of New Zealand, Tasmania, and the highest part of the Snowy Mountains in southeast Australia. Glaciers are responsive indicators of climatic conditions that leave distinctive landforms marking their former extents. The ability to determine the sizes that glaciers extended to in the past, and to measure when they were that large by geological dating methods, provides a valuable record of prehistoric climate change. In some respects the records of glacier activity from tectonically stable continental areas such as Tasmania and the Snowy Mountains are particularly valuable because there is greater certainty about the height and form of the mountain catchments of those glaciers, over long time periods. By contrast, in tectonically and volcanically active areas such as New Zealand, ongoing uplift of mountain ranges, or the episodic destruction and rebuilding of volcanoes, are complicating factors in the climatic interpretation of glacier landforms.

The development of high-precision TCN dating is a real break-through in the quest to obtain detailed climate change histories from glacier landforms. Much remains to be learned about the detail of prehistoric glacier variations in New

Zealand and Tasmania, and in so doing gaining a better understanding of the climate systems in the SW Pacific region.

See also: Dating Techniques. **Cosmogenic Nuclide Dating:** Exposure Geochronology; Methods; Overview. **Glacial Landforms:** Introduction. **Glaciation, Causes:** Tectonic Uplift and Continental Configurations. **Glaciations:** Early Quaternary (Pleistocene) and Precursors; Middle Pleistocene Glaciations in the Southern Hemisphere; Overview. **Ice Core Methods:** Overview. **Introduction:** Understanding Quaternary Climatic Change. **Paleoceanography, Records:** Postglacial South Pacific; **Pollen Records, Late Pleistocene:** Australasia. **Radiocarbon Dating:** Calibration of the ¹⁴C Record.

References

- Adkin GL (1911) The discovery and extent of former glaciation in the Tararua Ranges, North Island, New Zealand. *Transactions of the New Zealand Institute* 44: 308–316.
- Allibone AH and Wilson S (1997) Evidence of glacial activity at Mt Allen, southern Stewart Island, New Zealand. *New Zealand Journal of Geology and Geophysics* 40: 151–155.
- Alloway BV, Lowe DJ, Barrell DJA, et al. (2007) Towards a climate event stratigraphy for New Zealand over the past 30,000 years (NZ-INTIMATE Project). *Journal of Quaternary Science* 22: 9–35.
- Almond PC, Moar NT, and Lian OB (2001) Reinterpretation of the glacial chronology of south Westland. *New Zealand Journal of Geology and Geophysics* 44: 1–15.
- Augustinus PC (1992) Outlet glacier trough size-drainage area relationships, Fiordland, New Zealand. *Geomorphology* 4: 347–361.
- Augustinus PC (1999) Reconstruction of the Bulgobac Glacial System, Pieman River Basin, western Tasmania. *Australian Geographical Studies* 37: 24–36.
- Bacon SN, Chinn TJ, Van Dissen RJ, Tillinghast SF, Goldstein HL, and Burke RM (2001) Paleo-equilibrium line altitude estimates from late Quaternary glacial features in the Inland Kaikoura Range, South Island, New Zealand. *New Zealand Journal of Geology and Geophysics* 44: 55–67.
- Barnes PM (2009) Postglacial (after 20 ka) dextral slip rate of the offshore Alpine fault, New Zealand. *Geology* 37: 3–6.
- Barrell DJA (2011) Quaternary Glaciers of New Zealand. In: Ehlers J, Gibbard PL, and Hughes PD (eds.) *Quaternary Glaciations – Extent and Chronology: A Closer Look. Developments in Quaternary Science* 15: 1047–1064. Amsterdam: Elsevier.
- Barrell DJA, Andersen BG, and Denton GH (2011) Glacial geomorphology of the central South Island, New Zealand. *GNS Science Monograph*, vol. 27, 81 p + map (5 sheets). Lower Hutt, New Zealand: GNS Science.
- Barrows TT, Stone JO, Fifield LK, and Cresswell RG (2001) Late Pleistocene Glaciation of the Kosciuszko Massif, Snowy Mountains, Australia. *Quaternary Research* 55: 179–189.
- Barrows TT, Stone JO, Fifield LK, and Cresswell RG (2002) The timing of the Last Glacial Maximum in Australia. *Quaternary Science Reviews* 21: 159–173.
- Birkeland PW (1982) Subdivision of Holocene glacial deposits, Ben Ohau Range, New Zealand, using relative-dating methods. *Geological Society of America Bulletin* 93: 433–449.
- Brook MS (2009a) Lateral moraine age in Park Valley, Tararua Range, New Zealand. *Journal of the Royal Society of New Zealand* 39: 63–69.
- Brook MS (2009b) Glaciation of Mt Allen, Stewart Island (Rakiura): The southern margin of LGM glaciation in New Zealand. *Geografiska Annaler* 91A: 71–81.
- Brook MS, Kirkbride MP, and Brock BW (2008) Temporal constraints on glacial valley cross-profile evolution: Two Thumb Range, central Southern Alps, New Zealand. *Geomorphology* 97: 24–34.
- Burrows CJ (1973) Studies on some glacial moraines in New Zealand. 2. Ages of moraines of the Hooker, Mueller and Tasman Glaciers. *New Zealand Journal of Geology and Geophysics* 16: 831–855.
- Burrows CJ (1989) Aranuiian radiocarbon dates from moraines in the Mount Cook region, New Zealand. *New Zealand Journal of Geology and Geophysics* 32: 205–216.
- Chinn TJH (1981) Use of rock weathering rind thickness for Holocene absolute age-dating in New Zealand. *Arctic and Alpine Research* 13: 33–45.
- Chinn TJH (1995) Glacier fluctuations in the Southern Alps of New Zealand determined from snowline elevations. *Arctic and Alpine Research* 27: 187–198.

- Colhoun EA (1985) Glaciations of the West Coast Range, Tasmania. *Quaternary Research* 24: 39–59.
- Colhoun EA (2004) Quaternary glaciations of Tasmania and their ages. In: Ehlers J and Gibbard PL (eds.) *Quaternary Glaciations – Extent and Chronology, Part III: South America, Asia, Africa, Australasia, Antarctica. Developments in Quaternary Science* 2: 353–360. Amsterdam: Elsevier.
- Colhoun EA and Fitzsimons SJ (1990) Late Cainozoic Glaciation in Western Tasmania, Australia. *Quaternary Science Reviews* 9: 199–216.
- Colhoun EA, Hannan D, and Kiernan KW (1996) Late Wisconsin Glaciation of Tasmania. *Papers and Proceedings of the Royal Society of Tasmania* 130: 33–45.
- Colhoun EA and Shulmeister J (2007) Glaciations/Late Pleistocene of the SW Pacific Region. In: Elias SA (ed.) *Encyclopaedia of Quaternary Science*, 1st edn., pp. 1066–1077. Amsterdam: Elsevier.
- David TWE, Helms R, and Pittman EF (1901) Geological notes on Kosciusko, with special reference to evidences of glacial action. *Proceedings of the Linnean Society of New South Wales* 26: 26–74.
- Fink D, McKelvey B, Hannan D, and Newsome D (2000) Cold rocks, hot sands: In-situ cosmogenic applications in Australia at ANTARES. *Nuclear Instruments and Methods in Physics Research B* 172: 838–846.
- Fitzsimons SJ, Colhoun EA, van de Geer G, and Pollington MP (1993) The Quaternary geology and glaciation of the King Valley, western Tasmania. *Bulletin Geological Survey of Tasmania* 68: 57.
- Galloway RW (1963) Glaciation in the Snowy Mountains: A re-appraisal. *Proceedings of the Linnean Society of New South Wales* 88: 180–198.
- Gamble JA, Price RC, Smith IEM, McIntosh WC, and Dunbar NW (2003) $^{40}\text{Ar}/^{39}\text{Ar}$ geochronology of magmatic activity, magma flux and hazards at Ruapehu volcano, Taupo Volcanic Zone, New Zealand. *Journal of Volcanology and Geothermal Research* 120: 271–287.
- Gellatly AF, Chinn TJH, and Röthlisberger F (1988) Holocene glacier variations in New Zealand: A review. *Quaternary Science Reviews* 7: 227–242.
- Gibson N, Kiernan KW, and Macphail MK (1987) A fossil plant bolster from the King River, Tasmania. *Papers and Proceedings of the Royal Society of Tasmania* 121: 35–42.
- Hobden BJ, Houghton BF, Lanphere MA, and Nairn IA (1996) Growth of the Tongariro volcanic complex: New evidence from K–Ar age determinations. *New Zealand Journal of Geology and Geophysics* 39: 151–154.
- Hopf FVL, Colhoun EA, and Barton CB (2000) Late-glacial and Holocene record of vegetation and climate from Cynthia Bay, Lake St Clair, Tasmania. *Journal of Quaternary Science* 15: 725–732.
- Kiernan KW (1990) The extent of Late Cenozoic glaciation in the central highlands of Tasmania, Australia. *Arctic and Alpine Research* 22: 341–354.
- Kiernan K, Fifield LK, and Chappell J (2004) Cosmogenic nuclide ages for Last Glacial Maximum moraine at Schnells Ridge, southwest Tasmania. *Quaternary Research* 61: 335–338.
- Lee JM, Bland KJ, Townsend DB, and Kamp PJJ (compilers) (2011). Geology of the Hawke's Bay area. Lower Hutt, New Zealand: GNS Science. Institute of Geological & Nuclear Sciences 1:250,000 Geological Map 8.
- Lowell TV, Schoonenberger K, Deddens JA, et al. (2005) Rhizocarpon calibration curve for the Aoraki/Mount Cook area of New Zealand. *Journal of Quaternary Science* 20: 313–325.
- Mackintosh AN, Barrows TT, Colhoun EA, and Fifield LK (2006) Exposure dating and glacial reconstruction at Mt. Field, Tasmania, Australia, identifies MIS 3 and MIS 2 glacial advances and climatic variability. *Journal of Quaternary Science* 21: 363–376.
- Martinson DG, Pisias NG, Hays JD, Imbrie J, Moore TC Jr., and Shackleton ND (1987) Age dating and the orbital theory of the ice ages: Development of a high-resolution 0 to 300,000-year chronostratigraphy. *Quaternary Research* 27: 1–29.
- Mathews WH (1967) A contribution to the geology of the Mount Tongariro massif, North Island, New Zealand. *New Zealand Journal of Geology and Geophysics* 10: 1027–1039.
- McArthur JL and Shepherd MJ (1990) Late Quaternary glaciation of Mt Ruapehu, North Island, New Zealand. *Journal of the Royal Society of New Zealand* 20: 287–296.
- McCarthy A, Mackintosh A, Rieser U, and Fink D (2008) Mountain glacier chronology from Boulder Lake, New Zealand, indicates MIS 4 and MIS 2 ice advances of similar extent. *Arctic, Antarctic, and Alpine Research* 40: 695–708.
- McGlone MS (2002) The Late Quaternary peat, vegetation and climate history of the Southern Oceanic Islands of New Zealand. *Quaternary Science Reviews* 21: 683–707.
- McGregor VR (1967) Holocene moraines and rock glaciers in the central Ben Ohau Range, South Canterbury, New Zealand. *Journal of Glaciology* 6: 737–748.
- McKinnon M (ed.) (1997) *New Zealand Historical Atlas*, 290 p. Auckland: Bateman. Plate 7: Glaciation and climate change.
- Newnham RM, Lowe DJ, and Williams PW (1999) Quaternary environmental change in New Zealand: A review. *Progress in Physical Geography* 23: 567–610.
- Peterson JA and Robinson G (1969) Trend surface mapping of cirque floor levels. *Nature* 222: 75–76.
- Porter SC (1975) Equilibrium-line altitudes of late Quaternary glaciers in the Southern Alps, New Zealand. *Quaternary Research* 5: 27–47.
- Putnam AE, Denton GH, Schaefer JM, et al. (2010) Glacier advance in southern middle-latitudes during the Antarctic Cold Reversal. *Nature Geoscience* 3: 700–704.
- Putnam AE, Schaefer JM, Barrell DJA, et al. (2010) In situ cosmogenic ^{10}Be production-rate calibration from the Southern Alps, New Zealand. *Quaternary Geochronology* 5: 392–409.
- Reimer PJ, Baillie MGL, Bard E, et al. (2009) IntCal09 and Marine09 radiocarbon age calibration curves, 0–50,000 years cal BP. *Radiocarbon* 51: 1111–1150.
- Röthlisberger F (1987) *10 000 Jahre Gletschergeschichte der Erde*, 348 pp. Aarau: Verlag Sauerländer.
- Schaefer JM, Denton GH, Barrell DJA, et al. (2006) Near-synchronous interhemispheric termination of the Last Glacial Maximum in mid-latitudes. *Science* 312: 1510–1513.
- Schaefer JM, Denton GH, Kaplan M, et al. (2009) High-frequency Holocene glacier fluctuations in New Zealand differ from the northern signature. *Science* 324: 622–625.
- Shulmeister J, Fink D, and Augustinus PC (2005) A cosmogenic nuclide chronology of the last glacial transition in North-West Nelson, New Zealand – New insights in Southern Hemisphere climate forcing during the last deglaciation. *Earth and Planetary Science Letters* 233: 455–466.
- Shulmeister J, McKay R, Singer C, and McLea W (2001) Glacial geology of the Cobb Valley, north-west Nelson. *New Zealand Journal of Geology and Geophysics* 44: 55–62.
- Suggate RP (1990) Late Pliocene and Quaternary glaciations of New Zealand. *Quaternary Science Reviews* 9: 175–197.
- Suggate RP and Almond PC (2005) The Last Glacial Maximum (LGM) in western South Island, New Zealand: Implications for the global LGM and MIS 2. *Quaternary Science Reviews* 24: 1923–1940.
- Sutherland R, Kim K, Zondervan A, and McSaveney M (2007) Orbital forcing of mid-latitude Southern Hemisphere glaciation since 100 ka inferred from cosmogenic nuclide ages of moraine boulders from the Cascade Plateau, southwest New Zealand. *Geological Society of America Bulletin* 119: 443–451.
- Thackray GD, Shulmeister J, and Fink D (2009) Evidence for expanded middle and late Pleistocene glacier extent in northwest Nelson, New Zealand. *Geografiska Annaler* 91A: 291–311.
- Townsend D, Vonk A, and Kamp PJJ (compilers) (2008) Geology of the Taranaki area. Lower Hutt, New Zealand: GNS Science. Institute of Geological & Nuclear Sciences 1:250,000 Geological Map 7.
- Turnbull IM and Allibone AH (compilers) (2003) Geology of the Murihiku area. Lower Hutt, New Zealand: GNS Science. Institute of Geological & Nuclear Sciences 1:250,000 Geological Map 20.
- Turnbull IM, Allibone AH, and Jongens R (compilers) (2010) Geology of the Fiordland area. Lower Hutt, New Zealand: GNS Science. Institute of Geological & Nuclear Sciences 1:250,000 Geological Map 17.
- Turney CSM, Haberle S, Fink D, et al. (2006) Integration of ice core, marine and terrestrial records for the Australian Last Glacial Maximum and Termination: A contribution from the OZ-INTIMATE group. *Journal of Quaternary Science* 21: 751–761.
- Williams PW (1996) A 230 ka record of glacial and interglacial events from Aurora Cave, Fiordland, New Zealand. *New Zealand Journal of Geology and Geophysics* 39: 225–241.
- Winkler S (2004) Lichenometric dating of the 'Little Ice Age' maximum in Mt Cook National Park, Southern Alps, New Zealand. *The Holocene* 14: 911–914.
- Winkler S (2005) The Schmidt hammer as a relative-age dating technique: Potential and limitations of its application on Holocene moraines in Mt Cook National Park, Southern Alps, New Zealand. *New Zealand Journal of Geology and Geophysics* 48: 105–116.
- Winkler S and Matthews JA (2010) Holocene glacier chronologies: Are 'high-resolution' global and interhemispheric comparisons possible? *The Holocene* 20: 1137–1147.

Relevant Websites

- www.aqua.org.au – Australasian Quaternary Association.
- www.ga.gov.au – Geoscience Australia.
- www.gsnz.org.nz – Geoscience Society of New Zealand.
- www.gns.cri.nz – GNS Science.
- www.paleoclimate.org.nz – NZ paleoclimate research collective.

Late Quaternary of Antarctica

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Glossary

Cenozoic era: The most recent of the four major subdivisions of geologic time based on the history of animal evolution. The other three are the Proterozoic, Paleozoic, and Mesozoic. The Cenozoic spans about 65 My, from the end of the Cretaceous period to the present.

Deglaciation: This term describes ice retreat. An area is said to have deglaciated when it becomes ice free.

EAIS: Acronym for the East Antarctic Ice Sheet, or the part of the Antarctic glacial system that is east of the Transantarctic Mountains.

Eocene epoch: The Eocene epoch is part of the Tertiary period in the Cenozoic era, and lasted from about 55 to 34 Ma.

Glacial–interglacial: The term ‘glacial’ or ‘glaciation’ refers to a period of colder climate and extended glaciers compared to the present, and conversely ‘interglacial’ or ‘interglaciation’ refers to a period characterized by climate and glacial expansion similar to the present.

Grounding line: The grounding line of a glacier/ice sheet entering the sea describes the position where the ice loses contact with the sea bed and starts to float. Inside the grounding line, subglacial deposition or erosion takes place; outside it, marine and meltwater processes control sedimentation.

Holocene: The term Holocene refers to the present geological time, the past 10 ka since the end of the Pleistocene.

Last Glacial Maximum (LGM): This term refers to the time of global maximum extent of ice sheets during the last glaciation, some 20–18 ka.

Little Ice Age (LIA): A period of cooling and advancing glaciers, lasting approximately from the fourteenth to the late nineteenth centuries.

Marine oxygen-isotope stage (MIS): Oxygen-isotope ratios ($^{18}\text{O}/^{16}\text{O}$) measured in shells of marine microfossils are the basis for recognizing oxygen-isotope stages. The last 900 ka are divided into 23 stages, where even-numbered stages indicate glacial conditions (high $\delta^{18}\text{O}$); odd stages represent interglacial periods (low $\delta^{18}\text{O}$). The present interglacial

period, the Holocene, is numbered MIS-1, and the LGM occurred during MIS-2. Marine oxygen-isotope stages have become a global standard for interpreting Quaternary paleoclimate and chronology.

Marine $\delta^{18}\text{O}$ record: There are three isotopes of oxygen: ^{18}O , ^{17}O , and ^{16}O . The marine $\delta^{18}\text{O}$ record describes the relative abundance of the heavy ^{18}O isotope through time. It increases during glacial periods, when more of the lighter ^{16}O isotope is captured in glacial ice on land.

Miocene epoch: The Miocene epoch is a part of the Tertiary period in the Cenozoic era. It lasted from about 24 to 5.3 Ma.

Neoglaciation: This is a ‘Northern Hemisphere’ term, originally describing a Stone Age cooling occurring after the Holocene climatic optimum, later than 3–2.5 ka BP. Here it is used in a more general sense, meaning late Holocene cooling and growing glaciers.

Oligocene epoch: The Oligocene epoch is part of the Tertiary period in the Cenozoic era, and lasted from about 34 to 24 Ma.

Pleistocene epoch: The Pleistocene epoch covers the bulk of the Quaternary period, between 1.8 Ma and 10 ka.

Pliocene epoch: The Pliocene epoch is the last part of the Tertiary Period, and lasted from about 5.3 to 1.8 Ma.

Quaternary period: The Quaternary period covers the span of time from 1.8 Ma to the present day. It is subdivided into the Pleistocene and Holocene epochs.

Stratigraphic data: The term ‘stratigraphic data’ refers to any geological data that are retrieved from stratified deposits, that is, have bearings on spatial and temporal arrangements of rocks.

Tertiary period: The Tertiary period of geological time spans most of the Cenozoic era, between 65 and 1.8 Ma. It is divided into five epochs: Paleocene, Eocene, Oligocene, Miocene, and Pliocene. In the geological time column, the Tertiary follows the Cretaceous period, and is succeeded by the Quaternary period.

WAIS: This is an acronym for the West Antarctic Ice Sheet, or the huge glacial system that occurs west of the Transantarctic Mountains.

Introduction

Glaciers and ice sheets have existed intermittently in Antarctica for more than 30 My. The Antarctic glacial system can be divided into three main components (Figure 1):

- *The East Antarctic Ice Sheet (EAIS).* It is predominantly terrestrial, relatively stable, and has limited discharge to surrounding ice shelves.
- *The West Antarctic Ice Sheet (WAIS).* It is largely marine based, very dynamic, and discharges huge volumes of ice to floating ice shelves through large ice-stream systems.

- *Ice caps and glaciers on the Antarctic Peninsula.* This system is very dynamic and discharges ice to a number of small ice shelves.

The overall volume of glacial ice in Antarctica is estimated to be about 25 million km^3 , potentially containing 57 m of sea-level equivalent (Lythe et al., 2001). The Antarctic glacial system is an important factor in regulating, modifying, and forcing the global climate and oceanographic system. For the past >30 My, it has driven global sea-level stands and deep-ocean circulation, and acted as a regulator of global climate. Large-scale glaciations in Antarctica are not restricted to the

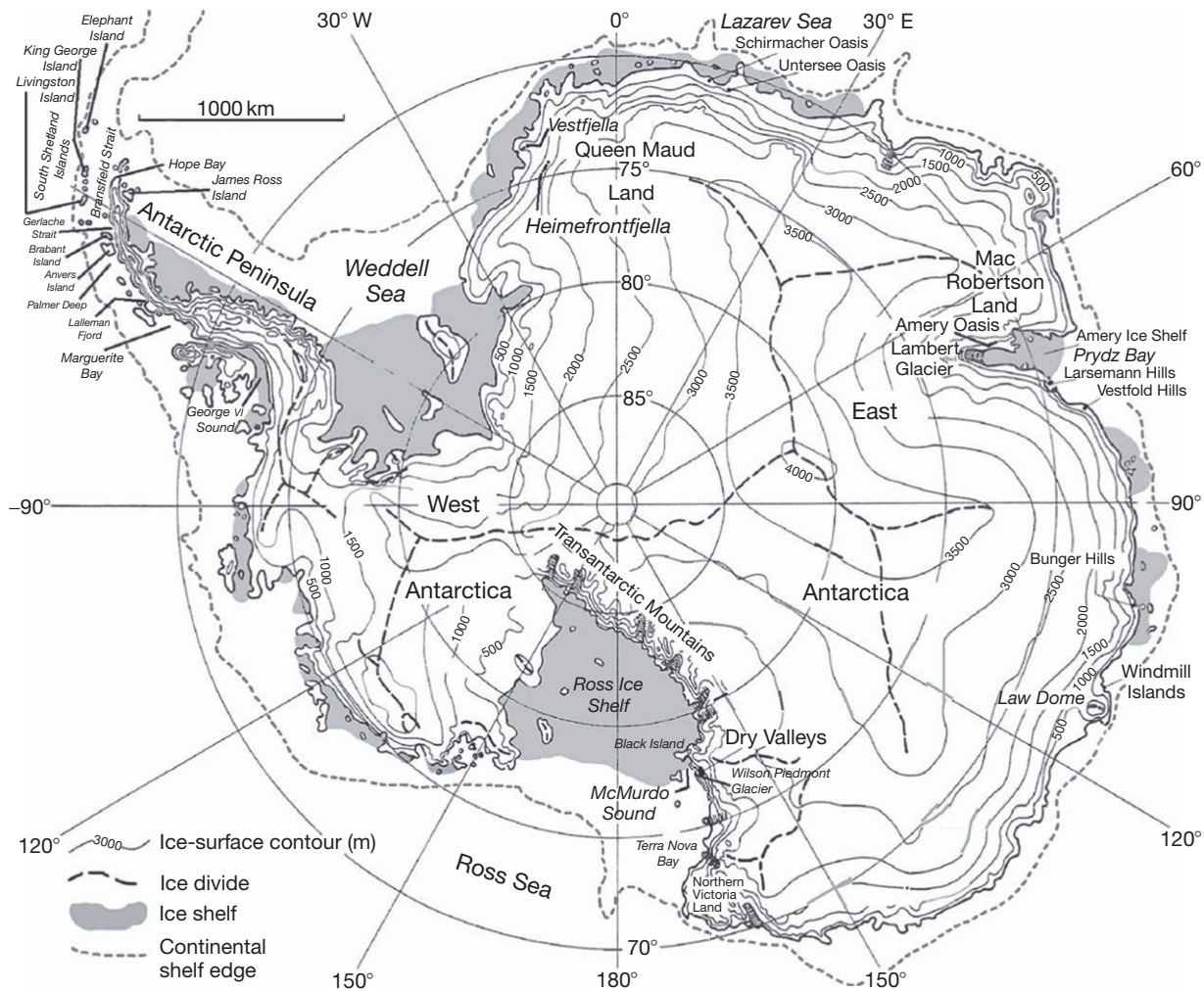


Figure 1 Map of Antarctica, with sites mentioned in the text.

Quaternary period, and for appreciating the late Quaternary glacial history, some understanding of the Cenozoic development of the Antarctic Ice Sheet is necessary. This review first presents a brief summary of the pre-Quaternary history, and then proceeds with a general overview of the Pleistocene record before relating more detailed reviews of the late Quaternary glacial development.

Pre-Quaternary Glacial History of Antarctica

The following review is a summary of overviews by Denton et al. (1991), Wilson (1995), Anderson (1999) and Hambrey et al. (2002).

Onset of Antarctic Glaciation

The initiation of cooling and glaciation in Antarctica was a consequence of plate-tectonic poleward drift and the isolation of Antarctica due to development of ocean passages and currents around the continent. Increases in deep-sea $\delta^{18}\text{O}$ concentrations show global cooling during the Eocene and Oligocene epochs, and there is compelling evidence (seismic profiles

and glacial diamictons in sediment cores) for an ice sheet on East Antarctica at the Eocene–Oligocene transition (34 Ma). Ice spread into the western Ross Sea by the late Oligocene. Stepwise increase in deep-sea $\delta^{18}\text{O}$ concentrations during the Miocene and into the Pliocene, and stratigraphical data (marine sediment cores, seismic records, and terrestrial sequences) provide evidence for significant middle-to-late Miocene ice-sheet expansion in Antarctica. It has been suggested that the ice sheet over East Antarctica developed into a major and permanent feature about 15 Ma. The data suggest repeated shifts from temperate to subpolar or polar climates throughout the Miocene and expansion of glaciers from the Transantarctic Mountains toward and beyond the Victoria Land coastline. Large ice caps existed on the Antarctic Peninsula, with ice expanded over the continental shelf areas by middle–late Miocene.

Pliocene and the Transition to the Pleistocene

The deep-sea record from around Antarctica suggests significant Pliocene sea-surface temperature fluctuations, and global oxygen-isotope records indicate considerable sea- and ice-volume fluctuations. As Northern Hemisphere ice sheets

were not developed at this stage, it is thought that global sea-level changes were driven by oscillations in Antarctic ice volumes. Seismic records from the Ross Sea region and Antarctic Peninsula continental shelf show a number of shelfwide unconformities bounding till sheets, indicating waxing and waning of continental glaciers during the Pliocene. Sediment cores from the western Ross Sea have meltwater deposits interbedded with till sheets, signifying repeated glacier oscillations in the Victoria Land region.

The Pliocene terrestrial record of Antarctic glaciations is fragmentary and difficult to interpret, and different opinions exist as to the Miocene–Pliocene stability of the EAIS. Reconstructions range from drastic fluctuations, with the ice sheet oscillating between extensive collapse and very extensive buildup and mountain overriding, to suggestions that it has existed close to its present configuration for the past ~14 My (Barrett, 1997; Denton et al., 1993). The greatest controversy focuses around different interpretations of the fossil evidence from the so-called Sirius Group. It has not been resolved whether the ice sheet underwent sufficient massive changes in ice volumes to allow the interior East Antarctica to become ice free, alternating with very extensive glacial overriding of the Transantarctic Mountains (Anderson, 1999; Barrett, 1997; Denton et al., 1991; Wilson, 1995). The Pliocene history of the WAIS is scantily known, but seismic records from the Antarctic Peninsula continental shelf indicate that there were repeated fluctuations in ice volume during the Pliocene (Anderson, 1999). The combined evidence suggests that ice sheets advanced to the edge of the continental shelf around Antarctica during the late Pliocene, with an ice extent similar to or greater than during the Last Glacial Maximum (LGM), and that those glaciations were interrupted by interglacial episodes that may have been warmer than the present.

Pleistocene Glacial Cycles in Antarctica

The Pleistocene record of glacial events in Antarctica is fragmentary due to the lack of long Pleistocene sediment sequences (marine core or sections on land) with good time resolutions. Stratigraphic data from sites in the Transantarctic Mountains record a major cooling in the early Pleistocene, and that the EAIS overrode the mountains during several episodes of extensive Pleistocene glaciations. The West Antarctic continental record is similarly fragmentary, but available data suggest that the WAIS was considerably thicker than at present on several occasions during the early and middle Pleistocene, and probably grounded on the shelf edge. Seismic profiles and core data from many shelf area locations off Antarctica show numerous stacked till sheets interbedded with glaciomarine and meltwater sediments. The controlling mechanisms for waxing and waning of the Antarctic Ice Sheets are not well understood. Hollin (1962) concluded that there is an overall in-phase relationship between glacial cycles in the Northern and Southern hemispheres, perhaps driven by falling and rising global sea levels regulated by buildup and decay of Northern Hemisphere ice sheets.

The Pleistocene stability of the marine-based WAIS has been discussed since Mercer (1968) suggested that it had repeatedly collapsed during Pleistocene interglacials. A model

study by MacAyeal (1992) showed that sporadic collapse of the ice sheet occurred throughout the past 1 My, and data have been presented (Scherer et al., 1998) that suggest that the WAIS collapsed at least once during the late Pleistocene. This remains to be tested through long sedimentary sequences highlighting the structure and timing of interglacial and glacial cycles. The deep-sea $\delta^{18}\text{O}$ record suggests that during the last interglacial (130–115 ka BP) there was less global ice volume than at present. It has been suggested that the collapse of the WAIS caused global sea levels to be 5–6 m higher than present during isotope stage 5e. If this episode of high global sea-level stand signifies collapse or decay of the WAIS (Mercer, 1968), if the water source was the Greenland Inland Ice (Cuffey and Marshall, 2000), or if it came from melting of the Northern Hemisphere ice sheets and partial melting of the Antarctic Ice Sheet (Denton et al., 1971) are issues that remain unresolved.

The available data indicate that the Antarctic glacial system has evolved through time: during early periods (Oligocene–Miocene), it was largely temperate in the coastal areas. A transition to cooler conditions followed in the Pliocene, and during much of the Quaternary it has been a stable, cold system.

Last Glacial Cycle and the LGM in Antarctica

Studies on late Quaternary glacial and climate changes in Antarctica have relied heavily on ice-core records (EPICA community members, 2004; Petit et al., 1999) and marine records (Anderson, 1999; Anderson et al., 2002; Bentley and Anderson, 1998). The evidence from ice-free areas in Antarctica is primarily based on stratigraphic and morphologic investigations, studies of lake sediment and moss-bank cores, and of fossil penguin rookeries (Berkman et al., 1998; Ingólfsson et al., 1998). While Antarctic ice cores depict glacial cycles in the form of oxygen-isotope and atmospheric compositions for the past ~800 ka (EPICA community members, 2004), this information is not easily translated into glacial history. There are large discrepancies between reconstructions of Antarctic late Quaternary glacial and climate history based on the geological data and ice-core records (Ingólfsson et al., 1998, 2003; Masson et al., 2000). The following summary focuses on the terrestrial and offshore geological records as the most reliable proxies for the glacial history of Antarctica. All ^{14}C ages are given in the text as uncalibrated ^{14}C years BP, reported as ka BP. The review relies on previous overviews by Denton et al. (1991), Bentley and Anderson (1998), Berkman et al. (1998), Ingólfsson et al. (1998), Anderson (1999), and Anderson et al. (2002).

Little is known about the extent of glaciation prior to the LGM during the last glacial cycle. Bentley and Anderson (1998) suggested that ice volumes in the Antarctic Peninsula region were considerably greater before 35 ka BP than at the 20–18 ka BP LGM. Hjort et al. (2003) suggested that ice cover during the earlier part of the last glacial cycle was more extensive than during the LGM. That earlier glaciation was followed by an interstadial, probably during marine oxygen isotope stage 3 (MIS 3), and thereafter by the LGM glaciation. There are a number of coastal sites around Antarctica where fossil-bearing pre-LGM marine interstadial or interglacial sediments occur (Berkman et al., 1998). A number of radiocarbon dates from a few sites constrain a possible interstadial between 30 and

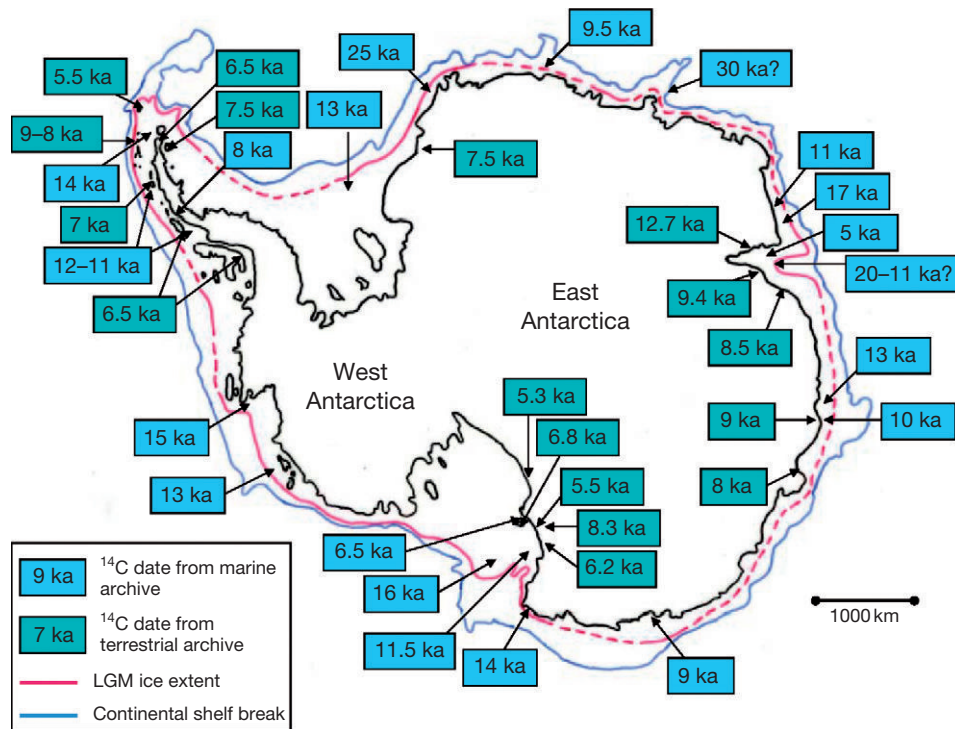


Figure 2 Extent of ice over Antarctica at the Last Glacial Maximum (LGM) and constraining minimum ages for ice retreat and deglaciation. Radiocarbon dates are given as ka BP. Data compiled from [Anderson et al. \(2002\)](#) and [Ingólfsson \(2004\)](#).

45 ka BP, but the sites are too few and too poorly constrained to draw conclusions as to the pre-LGM fluctuation pattern of the Antarctic Ice Sheets.

Most models suggest that during the LGM, the ice was considerably thicker and more extensive than present (see [Figure 2](#)). There are data showing that some coastal areas remained wholly or partly ice free, and there were numerous nunataks protruding above the ice surface in coastal and inland mountain ranges. LGM ice extended at least onto the mid-shelf areas off the western Antarctic Peninsula, the Weddell Sea, and East Antarctica, whereas it probably grounded close to the continental shelf edge off the Ross Sea. The timing of the last maximum glacial expansion around Antarctica is poorly constrained by ^{14}C dates, but most authors assume it to have coincided with the timing of lowest global sea levels during MIS 2, around 20–18 ka BP.

Deglaciation and Holocene Development of Glaciation and Climate in the Antarctic Peninsula Region

Deglaciation of the Shelf and Near-Shore Areas

The oldest ^{14}C dates, constraining ice retreat from the outer and middle shelf areas west of the Antarctic Peninsula, come from the Bransfield Strait and the shelf areas west of Anvers Island. These indicate ice retreat by 14–13 and 12–11 ka BP, respectively. The Palmer Deep area was ice free around 11 ka BP. Most dates from the inner shelf areas and fjords and bays on the peninsula constrain deglaciation to as late as 8–6 ka BP. Glaciomarine sedimentation in the central part of the Gerlache Strait began after 8 ka BP, providing a constraining

minimum date for deglaciation and shelf-ice retreat in the strait. Similarly, the onset of glaciomarine sedimentation in Lalleman Fjord occurred ca. 8 ka BP. The inner shelf areas around Anvers Island and in Marguerite Bay were ice free at 7–6 ka BP.

Deglaciation of Presently Ice-Free Coastal Areas

The deglaciation of coastal areas that are presently ice free is constrained by the minimum ages obtained from ^{14}C -dating of subfossil mollusk shells from raised marine deposits, peat deposits in moss-banks on the islands off the Antarctic Peninsula, organics from lake sediments, and remains from abandoned penguin colonies (see [Figure 3](#)). The oldest ^{14}C dates give minimum ages for deglaciation on the South Shetland Islands as 9–8 ka BP. Northern James Ross Island deglaciated prior to ca. 7.4 ka BP. In the south, George VI Sound became free of shelf ice 6.5–5.7 ka BP. Data from fossil penguin rookeries in Marguerite Bay suggest initial colonization in the period 6.5–5.5 ka BP ([Emslie, 2001](#)), which confirms minimum ages for deglaciation of 7–6 ka BP provided by marine-geological studies. The marine limit, which regionally lies between 30 and 55 m a.s.l., dates to ca. 7 ka BP and conforms with relatively late coastal deglaciation.

Once the glaciers were inside the present coastline, glacial retreat and ice disintegration was slow. On King George Island, glaciers were at or within their present limits by ca. 6 ka BP, and on northern James Ross Island prior to 4.7 ka BP, but parts of Livingston Island deglaciated as late as 5–3 ka BP. A minimum date for deglaciation on Elephant Island is provided by the onset of moss-bank formation on the steep slopes by 5.5 ka



Figure 3 A pair of Gentoo penguins nesting on the Antarctic Peninsula. The earliest Holocene penguin occupation in the Antarctic Peninsula region dates back to ~6.5 ka BP, which is an indication of relatively late deglaciation. Photo by Ó. Ingólfsson, 1987.

BP. There are indications from a number of sites that glaciers readvanced during the period <6–4 ka BP (Ingólfsson et al., 2003). Advances have been described from King George Island (5–4 ka BP), Brabant Island (later than 5.3 ka BP), James Ross Island (around 4.6 ka BP), and Hope Bay (around 4.7 ka BP). It is also known that the George VI Sound ice shelf expanded after 5.7 ka BP.

Holocene Climatic Optimum

Björck et al. (1996) presented a paleoclimatic synthesis for the Holocene development in the Antarctic Peninsula region, based on a compilation of numerous stratigraphic variables in lake sediments and moss-bank peat. Their reconstruction indicates that a regional warming began around 4.2 ka BP and reached an optimum around 3 ka BP, followed by a distinct and rapid climatic deterioration. A marine record from Lallemand Fjord recognizes a climatic optimum, reflected by high productivity and minimal sea ice in the fjord, between 4.2 and 2.7 ka BP.

Neoglaciation 3–0.1 ka: Cooling, Expanding Glaciers, and Increased Sea Ice

Since 3 ka BP, the climate has fluctuated between cold-dry and cool-humid, but remaining colder compared to the situation during the climatic optimum. Glaciers have oscillated and expanded somewhat in the Antarctic Peninsula region during the past ~2.5 ka, and glacial advances on the South Shetland Islands have been correlated to the Little Ice Age (LIA) expansion of glaciers in the Northern Hemisphere. The marine records suggest cooling expressed as decreased fjord/near-shore biologic productivity and more extensive and seasonally persistent sea ice after 3 ka BP.

The Last 50 Years: A Distinct Warming

The western Antarctic Peninsula region has moved to an increasingly warm climate regime during the past 50 years (Smith et al., 1999). Ecosystem research indicates that ecological transitions have occurred in response to this climate

change. A number of the Antarctic Peninsula ice shelves have been retreating with increasing speed since the late 1980s.

Deglaciation and Holocene Development of Glaciation and Climate in East Antarctica

Data highlighting the dynamics and patterns of deglaciation along the vast East Antarctic coastline, from the Weddell Sea to the Ross Sea, is extremely sparse. Ice-free areas in East Antarctica are few and far in between, and there is a lack of well-constrained marine records. This review relies on previous overviews by Denton et al. (1991), Berkman et al. (1998), Ingólfsson et al. (1998), Anderson (1999), and Anderson et al. (2002).

Queen Maud Land

There are very few constraints on the deglaciation in the eastern sector of the Weddell Sea and Queen Maud Land. Grounding line retreat in the Weddell Sea was underway about 13 ka BP. Post-LGM lowering of the ice-sheet profile at Vestfjella and Heimefrontfjella has been estimated to 700 and 200 m, respectively. The oldest avian occupation (proxy for deglaciation) there has been dated to 7.4 ka BP, but details as to glacial and climate fluctuations are lacking. Grounding line retreat on the Lazarev Sea shelf was underway prior to 9.5 ka BP, but reliable deglaciation chronologies for the Schirmacher and Untersee oases are lacking.

MacRobertson Land, Amery Oasis, and Amery Ice Shelf

Data from marine cores off MacRobertson Land suggest that deglaciation of the outer shelf occurred prior to 10.8–10 ka BP. The Amery oasis deglaciated prior to 12.7 ka BP, signifying an ice-surface lowering of the Lambert Glacier. A study of lake sediments revealed two postglacial periods of relatively warm conditions: an early period peaking around 7.8 ka BP and a later peaking around 3 ka BP (Wagner et al., 2004). The large Amery Ice Shelf has been present throughout the Holocene.

Larsemann Hills and Vestfold Hills

A number of radiocarbon dates from sediment cores in the Larsemann Hills give mid-Holocene ages for the deglaciation of the oasis. The last deglaciation of the Vestfold Hills has been determined by ^{14}C dates on mollusks from raised marine deposits and moraines, on marine algal sediments from numerous lake basins, and on fossil mosses. The oldest ^{14}C dates, giving minimum ages for the initial deglaciation and the incursion of marine water onto coastal areas, as well as for the initiation of aquatic moss growth, are between 8.6 and 8.4 ka BP. Relative sea level stood at the marine limit in the Vestfold Hills at ~9 m a.s.l., at ca. 6 ka BP. A number of studies indicate that a climate optimum can be recognized between ca. 4.7 and 2.5 ka BP. A late Holocene glacial expansion (the Chelnok glaciation) occurred between 2.5 and 1 ka BP. There is evidence for minor late Holocene (post 0.7 ka BP) ice marginal advances, in the form of a discontinuous series of ice-cored moraine ridges, which may be an LIA event. A study of marine

sediment cores from the Vestfold Hills area recognizes a significant cooling and increase in sea-ice cover after 2.5 ka BP.

Bunger Hills

Numerous ^{14}C dates indicate that the initial deglaciation of the Bunger Hills oasis dates back to between 10 and 8 ka BP. The glacial retreat was partly controlled by the rise of sea level, which caused a relatively rapid collapse of the ice-sheet margin. Breeding colonies of snow petrels expanded continuously following ice retreat and downwasting of dead ice, with the most intense phases of colonization between 6.7 and 4.7 ka BP. Before 5.6 ka BP, when the sea stood at the marine limit at between 9 and 7 m, glaciers were at or behind their present margins. A glacier advance, forming moraines called the Older Edisto Moraines, is bracketed between 6.2 and 4.7 ka BP. A high and stepwise increasing biogenic production in lakes between 3.5 and 2.5 ka BP has been taken to indicate increased temperatures and a climate optimum. It was followed by an abrupt and dramatic cooling that took place within less than 200 years. The so-called Younger Edisto Moraines postdate 1.1 ka BP and have been suggested to be an LIA glacial event.

Windmill Islands

The Windmill Islands deglaciated prior to 8 ka BP. A recent study by Roberts et al. (2004) found indications of a climate optimum in mid-Holocene. There is evidence for a readvance of the Law Dome ice-sheet margin onto part of the Windmill Islands some time between 4 and 1 ka BP. The readvance has been attributed to a positive mass balance on the Law Dome caused by high-precipitation rates during the Holocene.

Deglaciation and Holocene Environmental Development in Ross Sea and Victoria Land

This review relies on previous overviews by Denton et al. (1991), Berkman et al. (1998), Ingólfsson et al. (1998), Anderson (1999), Conway et al. (1999), Hall and Denton (2000), and Anderson et al. (2002).

Deglaciation and Grounding Line Retreat

A number of ^{14}C dates constrain the marine-based ice-sheet retreat in the Ross Sea from the LGM grounding line after 17 ka BP. The minimum age for grounding line retreat south of Terra Nova Bay is ca. 11.4 ka BP. The breakup of the shelf ice was probably considerably delayed. The minimum age for the deglaciation of coastal Terra Nova Bay is some time before 6.2 ka BP, when the coasts there became free of ice and marine habitats could begin to develop. The oldest penguin rookeries in Terra Nova Bay date to 5.8 ka BP.

Dating of penguin rookeries on Ross Island and in the McMurdo Sound area give minimum ages for the deglaciation in the southern Ross Sea, being 6.8–5.7 ka BP, and marine fossils collected from debris bands on the McMurdo Ice Shelf date the grounding line recession of the Ross Sea Ice Sheet to a position south of Black Island by 6.3 ka BP. ^{14}C dates on fossil marine mollusks from raised beaches along the southern Victoria Land coast, McMurdo Sound, and the islands in the

southwestern Ross Sea, as well as additional dates on penguin occupation in McMurdo Sound, all give minimum ages for the deglaciation as ca. 6.3–6.1 ka BP. A relative sea-level curve for the southern Victoria Land coast indicates unloading of grounded ice by 6.5 ka BP and provides one minimum estimate for the deglaciation (Hall and Denton, 1999).

Mid-late Holocene Glacier Oscillations

After the deglaciation in Terra Nova Bay, the ice shelves entering the bay were less extensive than today. The ice margins stood 2–5 km inside their present margins between 6.2 and 5.3 ka BP, but readvanced some time after 5.3 ka BP. After a withdrawal, moraine ridges containing fossil marine shells show a glacial readvance in Terra Nova Bay after 0.5 ka BP, which has been interpreted to correspond to the LIA glacial expansion in the Northern Hemisphere. The Wilson Piedmont glacier retreated inside the coast between 5.7 and 5 ka BP, but readvanced in late Holocene times, and at a number of locations it advanced across raised beaches that date back to 5.4–5 ka BP.

Climatic Optimum in Victoria Land

The best information on environmental development in Victoria Land during the later part of the Holocene comes from the spatial and temporal distribution of penguin rookeries (Baroni and Orombelli, 1994). They documented continuous presence of Adélie penguins after ca. 7 ka BP, but the greatest diffusion of rookeries occurred between 3.6 and 2.6 ka BP. Baroni and Orombelli concluded that this 'penguin optimum' was a period of particularly favorable environmental conditions. It was followed by a sudden decrease in the number of penguin rookeries in southern Victoria Land shortly after 2.6 ka BP, attributed to an increase in the sea ice. Raised beaches formed after 5 ka BP, with rounded cobbles indicating high-wave energy at the time of their formation, have been described from a number of sites in southern Victoria Land (Figure 4). At many of these sites today, the ice foot rarely breaks up and recent beaches are of low-energy type.



Figure 4 Raised beach ridge of middle-late Holocene age in southern Victoria Land. Well-rounded cobbles and boulders suggest high-wave energy at the time of beach-ridge formation. At this site today, the ice foot rarely breaks up and recent beaches are of low-energy type. Photo by Ó. Ingólfsson, 1994.

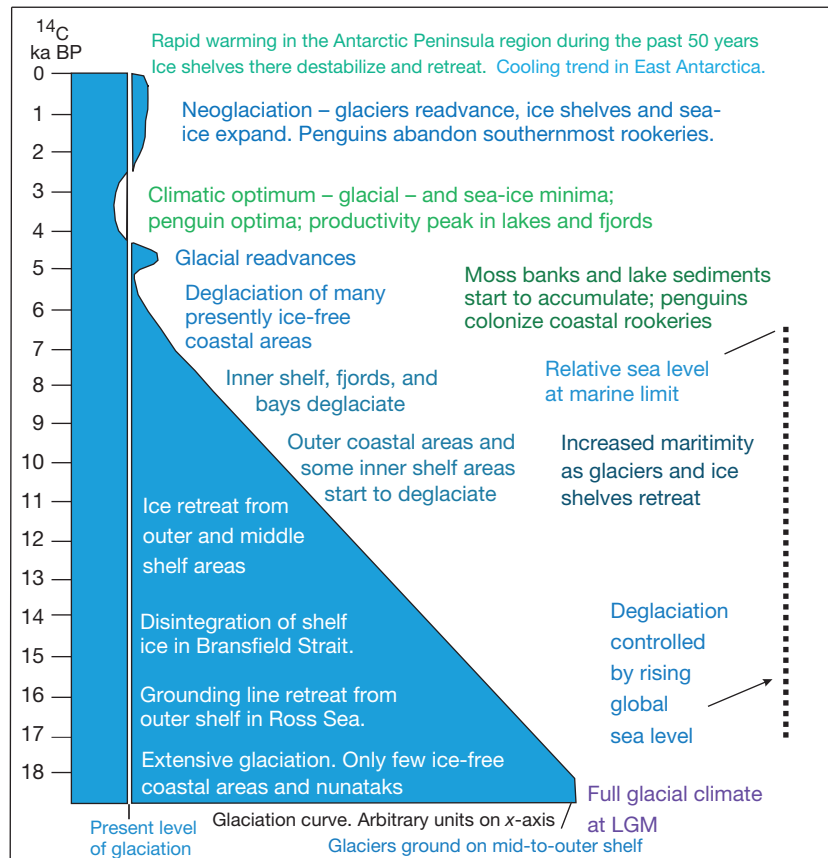


Figure 5 A broad synthesis of the glacial and environmental history of Antarctica since the LGM.

Summary of the Late Quaternary Glacial History of Antarctica

The following broad pattern describing the late Quaternary glacial history of Antarctica can be recognized (Figure 5):

1. Little is known about late Quaternary interglacials–interstadials older than the LGM. There are indications of deglaciation 30–45 ka BP and subsequent LGM glaciation.
2. LGM glaciers from Antarctica generally grounded on the middle to outer shelf, and ice retreat from the LGM positions was underway by 17–14 ka BP.
3. Deglaciation occurred mainly during the time period >14–6 ka BP. Outer- and middle-shelf areas deglaciated between 14 and 8 ka BP, while most inner shelf areas, fjords, bays, and, most presently, ice-free coastal land areas deglaciated prior to 8–6 ka BP.
4. The transition from glacial to interglacial conditions in Antarctica, characterized by an ice extent similar to or less than at present, the onset of lake-sediment accumulation in ice-free basins, moss-bank growth on the islands off the Antarctic Peninsula, and penguin occupation of coastal rookeries, was broadly completed by 6 ka BP.
5. Mid-Holocene glacial expansion events have been described from a number of sites.
6. A circum-Antarctic Holocene climate optimum occurred in the period 4.5–2.5 ka BP, followed by a distinct cooling. Many outlet glaciers and ice shelves expanded during the

late Holocene, suggesting a possible LIA glacial event in Antarctica.

7. A significant warming has occurred in the Antarctic Peninsula region for the past 50 years.

See also: **Beetle Records:** Late Tertiary and Early Quaternary Records. **Diatom Methods:** $\delta^{18}\text{O}$ Records. **Glacial Landforms, Sediments:** Glaciomarine Sediments and Ice-Rafted Debris. **Glaciations:** Early Quaternary (Pleistocene) and Precursors; Overview. **Ice Core Methods:** Chronologies. **Ice Cores:** Dynamics of the East Antarctic Ice Sheet; Dynamics of the West Antarctic Ice Sheet. **Paleoceanography, Physical and Chemical Proxies:** Oxygen-Isotope Stratigraphy of the Oceans. **Sea Level Studies:** Overview.

References

- Anderson JB (1999) *Antarctic Marine Geology*. Cambridge: Cambridge University Press.
- Anderson JB, Shipp SS, Lowe AL, Wellner JS, and Mosola AB (2002) The Antarctic Ice Sheet during the Last Glacial Maximum and its subsequent retreat history: A review. *Quaternary Science Reviews* 21: 49–70.
- Baroni C and Orombelli G (1994) Abandoned penguin rookeries as Holocene paleoclimatic indicators in Antarctica. *Geology* 22: 23–26.
- Barrett PJ (1997) Antarctic paleoenvironment through Cenozoic times: A review. *Terra Antarctica* 3: 103–119.

- Bentley MJ and Anderson JB (1998) Glacial and marine geological evidence for the ice sheet configuration in the Weddell Sea-Antarctic Peninsula region during the Last Glacial Maximum. *Antarctic Science* 10: 309–325.
- Berkman PA, Andrews JT, Björck S, et al. (1998) Circum-Antarctic coastal environmental shifts during the Late Quaternary reflected by emerged marine deposits. *Antarctic Science* 10: 345–362.
- Björck S, Håkansson H, Olsson S, Ellis-Evans C, Humlum O, and Lirio JM (1996) Late Holocene palaeoclimatic records from lake sediments on James Ross Island, Antarctica. *Palaeogeography, Palaeoclimatology, Palaeoecology* 113: 195–220.
- Conway H, Hall BL, Denton GH, Gades AM, and Waddington ED (1999) Past and future grounding-line retreat of the West Antarctic Ice Sheet. *Science* 286: 280–283.
- Cuffey KM and Marshall SJ (2000) Substantial contribution to sea level rise during the last interglacial from the Greenland Ice Sheet. *Nature* 404: 591–594.
- Denton GH, Armstrong RL, and Stuiver M (1971) The late Cenozoic glacial history of Antarctica. In: Flint RF and Turekian KK (eds.) *The Late Cenozoic Glacial Ages*, pp. 267–306. New Haven: Yale University Press.
- Denton GH, Prentice ML, and Burckle LH (1991) Cainozoic history of the Antarctic Ice Sheet. In: Tingey RJ (ed.) *Geology of Antarctica*, pp. 365–433. Oxford: Oxford University Press.
- Denton GH, Sugden DE, Marchant DR, Hall BL, and Wilch TI (1993) East Antarctic Ice Sheet sensitivity to Pliocene climate change from a Dry Valleys perspective. *Geografiska Annaler A* 75: 155–204.
- Emslie SD (2001) Radiocarbon dates from abandoned penguin colonies in the Antarctic Peninsula region. *Antarctic Science* 13: 289–295.
- EPICA Community Members (2004) Eight glacial cycles from an Antarctic ice core. *Nature* 429: 623–628.
- Hall BL and Denton GH (1999) New relative sea-level curves for the southern Scott Coast, Antarctica: Evidence for Holocene deglaciation of the Ross Sea. *Journal of Quaternary Science* 14: 641–650.
- Hall BL and Denton GH (2000) Radiocarbon chronology of Ross Sea drift, Eastern Taylor Valley, Antarctica: Evidence for a grounded ice sheet in the Ross Sea at the Last Glacial Maximum. *Geografiska Annaler* 82: 305–336.
- Hambrey MJ, Barrett PJ, and Powell RD (2002) Late Oligocene and early Miocene glacial marine sedimentation in the SW Ross Sea, Antarctica: The record from offshore drilling. In: Dowdeswell JA and O'Cofaigh C (eds.) *Special Publication of the Geological Society of London 203: Glacier-Influenced Sedimentation in High-Latitude Continental Margins*, pp. 105–128. Bath: Geological Society Publishing House.
- Hjort C, Ingólfsson Ó, Bentley MG, and Björck S (2003) The Late Pleistocene and Holocene glacial and climate history of the Antarctic Peninsula region as documented by the land and lake sediment records – A review. *American Geophysical Union Antarctic Research Series* 79: 95–102.
- Hollin JT (1962) On the glacial history of Antarctica. *Journal of Glaciology* 4: 173–195.
- Ingólfsson Ó (2004) Quaternary glacial and climate history of Antarctica. In: Ehlers J and Gibbard PL (eds.) *Quaternary Glaciations – Extent and Chronology, Part III*, pp. 3–43. Amsterdam: Elsevier.
- Ingólfsson Ó, Hjort C, Berkman P, et al. (1998) Antarctic glacial history since the Last Glacial Maximum: An overview of the record on land. *Antarctic Science* 10: 326–344.
- Ingólfsson Ó, Hjort C, and Humlum Ó (2003) Glacial and climate history of the Antarctic Peninsula since the Last Glacial Maximum. *Arctic, Antarctic, and Alpine Research* 35: 175–186.
- Lythe M and Vaughan DG, BEDMAP Consortium (2001) BEDMAP: A new ice thickness and subglacial topographic model of Antarctica. *Journal of Geophysical Research* 106: 11335–11352.
- MacAyeal D (1992) Irregular oscillations of the West Antarctic Ice Sheet. *Nature* 359: 29–32.
- Masson V, Vimeux F, Jouzel J, et al. (2000) Holocene climate variability in Antarctica based on 11 ice-core isotopic records. *Quaternary Research* 54: 348–358.
- Mercer JH (1968) Antarctic ice and Sangamon sea level. *Commission of Snow and Ice Reports and Discussions*, pp. 217–225. Publication no. 79 de l'Association Internationale d'Hydrologie Scientifique. Gentbrugge: International Association of Hydrological Sciences.
- Petit JR, Jouzel J, Raynaud D, et al. (1999) Climate and atmospheric history of the past 420,000 years from the Vostok ice core, Antarctica. *Nature* 399: 429–436.
- Roberts D, McMinna A, Cremer H, Gore DB, and Melles M (2004) The Holocene evolution and palaeosalinity history of Beall Lake, Windmill Islands (East Antarctica) using an expanded diatom-based weighted averaging model. *Palaeogeography, Palaeoclimatology, Palaeoecology* 208: 121–140.
- Scherer RP, Aldahan A, Tulaczyk S, Possnert G, Engelhardt H, and Kamb B (1998) Pleistocene collapse of the West Antarctic Ice Sheet. *Science* 281: 82–85.
- Smith RC, Ainley D, Baker K, et al. (1999) Marine ecosystem sensitivity to climate change. *Bioscience* 49: 393–404.
- Wagner B, Cremer H, Hultsch N, Gore DB, and Melles M (2004) Late Pleistocene and Holocene history of Lake Terrasovoje, Amery Oasis, East Antarctica, and its climatic and environmental implications. *Journal of Paleolimnology* 32: 321–339.
- Wilson GS (1995) The Neogene East Antarctic Ice Sheet: A dynamic or stable feature. *Quaternary Science Reviews* 14: 101–123.

Late Pleistocene in Eurasia

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Eemian Stage Interglacial

The late Pleistocene started with a warm stage, named the Eemian in NW Europe. This article considers only those glaciogenic sediments that obviously postdate the terrestrial or marine Eemian-age formations.

Correlation of the Eemian interglacial Stage *sensu stricto* with the Marine Isotope Stage (MIS) stratigraphy was not perceived as a problem until recently, as it has generally been assumed that boundaries identified using a variety of proxies on land were precisely coeval with those seen in ocean sediments. Yet, the different response rates of different proxies to climate changes and these changes themselves may be time transgressive. Off Portugal, [Shackleton et al. \(2003\)](#) demonstrated that the MIS 6/5 boundary is not coeval with the Saalian/Eemian Stage boundary on land, as previously assumed. There is a gap of 6000 years between the beginning of the Eemian (126 ka), defined on the basis of palynology, and MIS 5e (132 ka), defined on the basis of oxygen isotope determinations from the same sediments. Similar problems may arise at the end of the event. Thus, land and marine sequences must be correlated with considerable caution as it cannot be assumed that the boundaries recognized in different situations are indeed coeval.

While the Eemian has been traditionally correlated with MIS 5e Substage, the Sangamonian of North America is normally equated with all of MIS 5, which includes the early, very warm Weichselian interstadials (Brörup, Odderade) of Europe. It also includes the cold Marine Isotope Substages 5b and 5d, during which, as pointed out later, major ice sheets may have developed. However, some North American scientists use the Sangamonian for only Marine Isotope Substage 5e.

Eemian marine deposits are well known from western Norway, Denmark, northern Germany, Poland, Finland, and along the northern margin of European Russia and Siberia ([Figure 1](#)). The Eemian probably included the last marine transgression into the Baltic Sea basin before the Holocene. The Eemian Sea is traceable around Fennoscandia into the White Sea basin, where it is described as the Boreal Transgression, and farther eastward across the Russian Arctic ([Funder et al., 2002](#)). For a century, the Boreal formation, with its characteristic Atlantic fauna, has been known as a major stratigraphical landmark of northern Russia. In West Siberia and on the Taimyr Peninsula, the Kazantsevo Formation, which includes the extinct mollusk *Cyrtodaria angusta*, was traditionally accepted as its counterpart (e.g., Kind, 1974). However, lately it has been geochronometrically demonstrated that a much younger Karginsky Formation corresponds to the Eemian ([Astakhov and Nazarov, 2010](#)). The greater extent of the Eemian Sea, as compared to that of the Holocene, has been

explained by the major extent and thickness of the Saalian ice sheets and the different deglaciation history of the Saalian and Weichselian ice sheets.

Besides the marine sediments, an important stratigraphic marker separating the deposits of the last glaciation from older strata is provided by terrestrial formations with a high organic content that fill in numerous depressions overlying the middle Pleistocene glaciogenic sequences.

Weichselian Cold Stage

Northern Russia

Originally, on the basis of mapped boulder trains and the occurrence of fresh-looking hummock-and-lake landscapes, the latest ice sheets of the West Siberian and Pechora Lowlands were thought to have been confined to the Arctic and fed from upland ice domes of the Urals, Novaya Zemlya, Putorana Plateau of Central Siberia, and Byrranga Mountains in the Taimyr Peninsula. It was only through the widespread use of remote sensing data in the 1970s that many push moraines were detected, the configuration of which in most places contradicted the mountain glaciation hypothesis and testified to upslope ice flow from the coastal lowlands of the Kara Sea toward the high areas of the Urals and Taimyr. The mapped pattern of ice-pushed ridges could form only if the ice sheet was centered on the Barents and Kara Sea floors. During the most extensive glaciations, the northbound drainage was blocked by the advancing glaciers, forming huge ice-dammed lakes ([Astakhov, 1977, 1998](#)). Apart from ice-marginal features belonging to the vast Kara Ice Sheet, only traces of small alpine glaciers were mapped in the Polar Urals. The late Pleistocene ice sheet was thinner than that during the middle Pleistocene, but was sufficiently thick to push morainic ridges up to 560 a.s.l. on the northern slope of the Urals. This trimline suggests that ice was at least 1 km thick on the Kara shelf.

During an early phase of the last Ice Age, very large morainic loops, overlying interglacial marine sediments, were deposited in the southern part of the Taimyr Peninsula. These morainic assemblages, together with boulder trains, pointed to a former ice divide north of the Byrranga Mountains, that is, in the north eastern Kara Sea region ([Makaveyev et al., 1982](#)).

When it was demonstrated that large ice domes were located on the Kara Sea shelf, the former paleogeographical paradigm changed and totally different ideas of Weichselian ice limits in the Arctic were envisioned. The determination of several finite radiocarbon dates from beneath the youngest till tipped the scale of the discussion toward the idea of a very extensive last glaciation, comparable to the style of the well-studied Late Wisconsinan Laurentide and Late Weichselian



Figure 1 Extent of the Eemian Sea. Reproduced from Ehlers J (2011) *Das Eiszeitalter*, p. 363. Heidelberg: Spektrum.

Fennoscandian glacial systems. Grosswald (1980), the most ardent proponent of the maximalist model, maintained that Weichselian ice sheets in northern Russia ca. 20 ka extended far south of the Arctic Circle, close to the middle Pleistocene ice limit. However, subsequent testing of the maximalist concept in West Siberia and in European Russia has led to the rejection of a Late Weichselian age for non-Scandinavian glaciers. It turned out that finite radiocarbon dates from sediments overlying the uppermost till greatly outnumbered those from beneath the till. The assumption that the Late Weichselian ice sheet was much smaller than the penultimate glaciations was substantiated by radiocarbon dates obtained from frozen mammoth carcasses and other remains demonstrating that these northern areas were not covered by glacier ice after the animals' death 35–40 ^{14}C ka. It is now clear that most of the old conventional ^{14}C dates from beneath the uppermost till are spurious (Astakhov, 1998, 2006; Astakhov and Nazarov, 2010). This is why the last Siberian interglacial and the Kargin'sky marine transgression were previously erroneously correlated with MIS 3 (e.g., Kind, 1974).

During the last decade, studies of the glacial history of the Barents and Kara Sea regions were carried out within the framework of the European Science Foundation programme Quaternary Environments of the Eurasian North (QUEEN) and by several multinational subprojects. Marginal formations of the Weichselian glaciation maximum were mapped using photogeological interpretation and dated by radiocarbon and luminescence methods. This work has established that the first major glacial advance from the Barents and Kara Seas into the

Russian mainland occurred during an early phase of the Weichselian Stage, some 80–100 ka (Svendsen et al., 2004). During this glaciation, a huge ice-dammed lake, named Lake Komi, formed over the Pechora Lowland in the European part of Northern Russia (Mangerud et al., 2004). It is probable that an even larger ice-dammed lake formed over the West Siberian Lowland (Astakhov, 2006). Later, the northern rim of the continent was affected by a younger advance during the period 70–50 ka. It deposited the Markhida moraine north of the Arctic Circle in the Pechora drainage basin (Svendsen et al., 2004). Recent research has also confirmed the absence of sizeable Late Weichselian mountain glaciers in the Urals, on the Taimyr Peninsula, and even in the higher East Siberian mountains (Hjort et al., 2004; Mangerud et al., 2008; Stauch and Gualtieri, 2008).

The main signature of the last ice sheet in the West Siberian Plain is fine-grained diamicton with blocks of frozen sand, laterally replaced by huge flat-lying fields of dead ice 5–60 m thick, which are barely covered by melted-out silts (Figure 2).

The massive ground ice that is normally banded and deformed, sometimes with suspended boulders, is believed to be mostly remnants of the Early to Middle Weichselian basal glacier ice, preserved in the thick Pleistocene permafrost. The number of sites, the degree of their preservation, and probable age increase eastward from the northeastern corner of the European Arctic toward the East Siberian Sea.

This direct signature of former glaciation is instrumental in delineating the Weichselian ice margin in the central lowland between the Ob and Yenisei river valleys. On the Gydan and

Yamal peninsulas, the massive buried ice is overlain in places by cold-water marine silts with *Portlandia arctica*. However, it is very unlikely that buried glacial ice, normally found at low altitudes, could survive the Eemian marine transgression which, judging from the Atlantic fauna, had bottom temperatures well above zero. Thus, the area with known sites of massive buried ice is thought to have been covered by an ice sheet that is younger than the Eemian marine inundation but beyond the range of the radiocarbon method (more than 40 ka) (Astakhov, 1998, 2006).



Figure 2 Late Pleistocene dead glacier ice covered by melted-out silt on the Gydan Peninsula in West Siberia, 70° N/75° E. This banded and deformed ice is a counterpart of Scandinavian basal tills. Photograph by V. Astakhov.

Along the eastern border of the Plain, a hummock-and-lake landscape continues to develop by glaciokarst of stagnant ice bodies. A flat plain at 45–55 m a.s.l., composed of thick glaciolacustrine rhythmities, intrudes into this hummocky terrain of the Yenisei valley close to the Arctic Circle. The rhythmities are overlain by alluvium, radiocarbon dated to 34–42 ka at Farkovo, and by sinkhole silts containing frozen logs with dates from 35 ka to infinite at Igarka (Kind, 1974). Based on comprehensive investigations, it has now been demonstrated that during the Last Glacial Maximum (LGM), a major ice sheet did exist in the Barents Sea–Svalbard region (Landvik et al., 1998), but it did not expand into the Russian mainland. This ice sheet limit is now identified well offshore in the Barents Sea, based on marine cores and seismic data (Svendsen et al., 2004).

No observations indicate that there was an ice-dammed lake along the northern rim of the Russian Lowland during the Late Weichselian, suggesting that northbound drainage was not blocked by a coherent ice sheet (Astakhov, 2006). Accordingly, the glacier advance that reached the NW coast on Taimyr may have been unconnected with the Barents–Kara Ice Sheet, so there may have been an ice-free corridor on the shelf (Mangerud et al., 2002).

Thus, on the basis of modern data, the LGM was preceded by at least two major Weichselian glaciations. The Early Weichselian ice sheet (Figure 3) that went far inland and blocked the major West Siberian rivers was followed by an independent ice advance from the northeast (Figure 4) that did not dam the Yenisei river (Svendsen et al., 2004). Further to the NE, the Middle Weichselian ice limit of the Yenisei region can only be traced by morainic arcs, which block deep valleys dissecting the Putorana Plateau. Traditionally, these moraines, locally

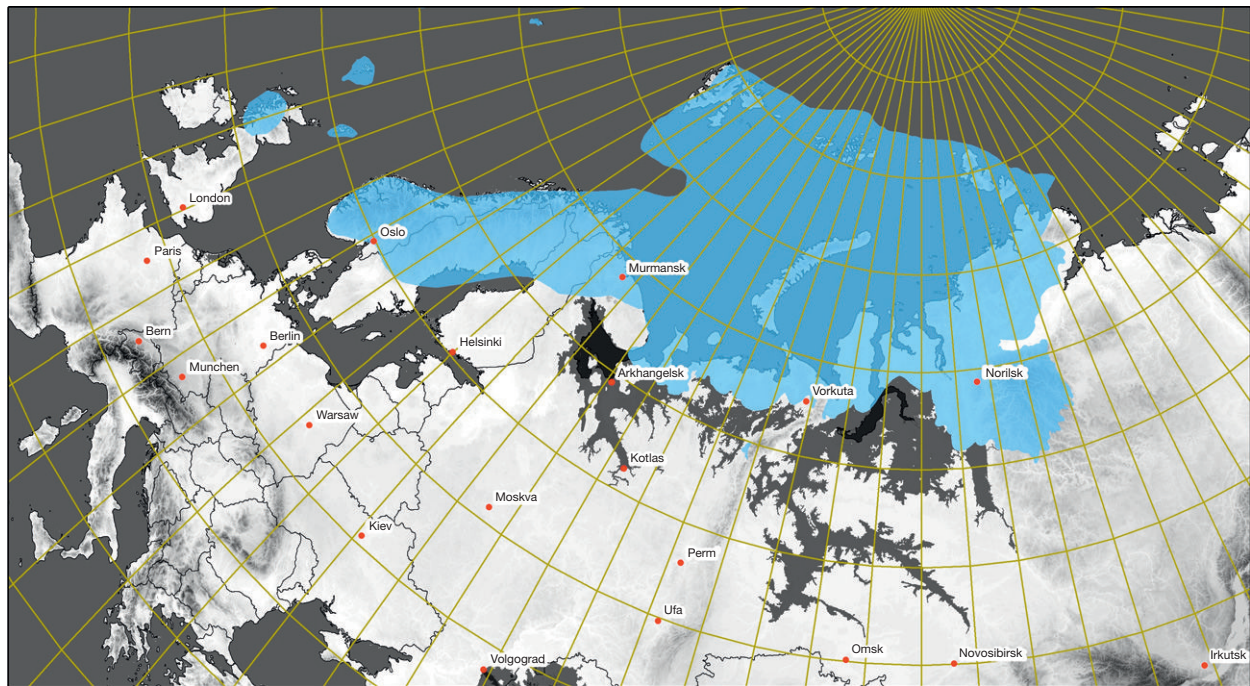


Figure 3 Extent of the Early Weichselian glaciation. Reproduced from Ehlers J, Gibbard PL and Hughes PD (eds.) (2011) *Quaternary Glaciations – Extent and Chronology, Part IV – A Closer Look. Developments in Quaternary Science*, vol. 15. Amsterdam: Elsevier.

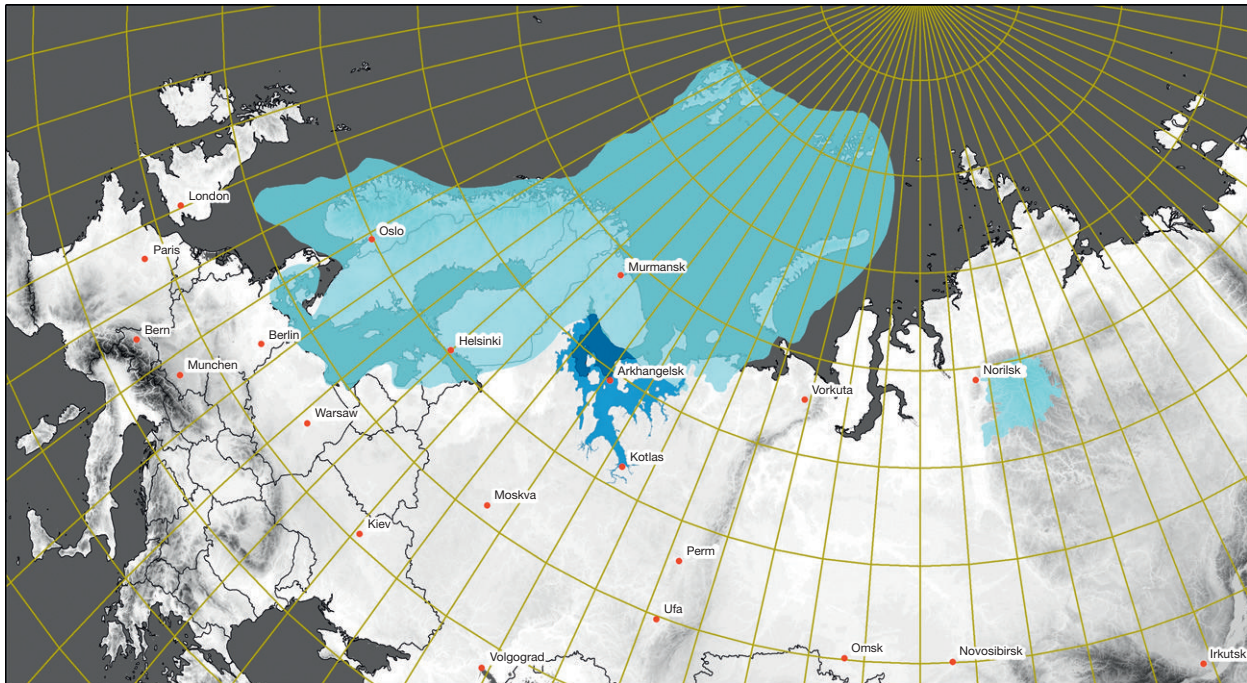


Figure 4 Extent of the Middle Weichselian glaciation. Reproduced from Ehlers J, Gibbard PL, and Hughes PD (eds.) (2011) *Quaternary Glaciations – Extent and Chronology, Part IV – A Closer Look. Developments in Quaternary Science*, vol. 15. Amsterdam: Elsevier.

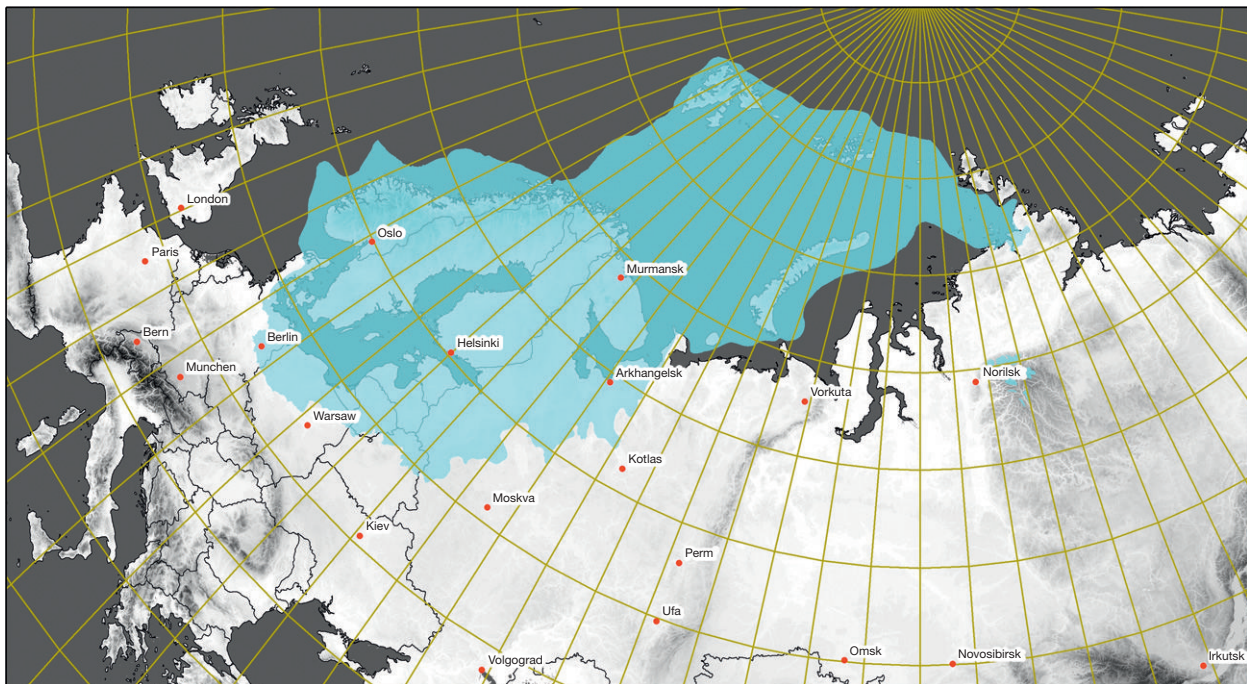


Figure 5 Extent of the Late Weichselian glaciation. Reproduced from Ehlers J, Gibbard PL, and Hughes PD (eds.) (2011) *Quaternary Glaciations – Extent and Chronology, Part IV – A Closer Look. Developments in Quaternary Science*, vol. 15. Amsterdam: Elsevier.

called the Norilsk Stade, were attributed to the Late Weichselian, and even to the Younger Dryas Stadial (Kind, 1974). However, more recent investigations of the bottom sediments of Lake Lama, situated on the proximal side of the Norilsk Moraine, suggest that the lacustrine sedimentation in this

basin began before the LGM (Svendsen et al., 2004). If correct, this may imply that the last coherent ice cap on the Putorana Plateau is older and that the Late Weichselian glaciation was more restricted than suggested by the radiocarbon dates from the submill sediments mentioned above (Figure 5).

In the perennially frozen Russian Arctic, postglacial time began at ca. 50 ka BP when present-day river valleys developed across morainic landscapes and the bottoms of drained proglacial lakes. The valleys provided sufficient vegetation to allow the first Paleolithic hunters to penetrate beyond the Arctic Circle as early as ca. 40 ka BP (Svendsen et al., 2010). Most finds of mammoth fauna belong to this large and cold, basically treeless interstadial ca. 50–25 ka BP called the Byzovaya in European Russia and Varyakha in West Siberia. The LGM time on the Russian arctic mainland is represented by a mantle of loess-like silts with ice wedges, and rare remains of a mammoth fauna (Astakhov and Nazarov, 2010).

Scandinavian Ice Sheet

In contrast to the spectacular changes regarding the Weichselian ice-sheet configuration in northern Russia over the last decade, much less new information is available from the Scandinavian Ice Sheet area (Figure 6). The maximum ice advances of the last cold stage both in Britain and continental Europe did not occur before the last part of the Weichselian (Devensian) Stage. Nevertheless, there can be no doubt that

extensive glaciers existed in northern Europe also during the Early Weichselian substage (Mangerud, 2004). The exact extent of the Early Weichselian glaciations in Scandinavia, the deposits of which have been identified in Norway, Finland, and northern and central Sweden, is so far unknown. It is certain, however, that it was less extensive than the Late Weichselian glaciation. Finnish Lapland has traditionally been considered ice-covered between the Eemian Stage and the Early Weichselian Brørup Interstadial. This is suggested by the fact that deposits of the two periods are always found separated from each other by a till bed (Hirvas, 1991). Recently, it has been shown that this first ice cover (at locality Sokli) was during isotope stage 5b (Helmens et al., 2000). However, farther to the south, in Finland, there is no evidence of Early Weichselian glaciation. Thermoluminescence (TL)- and optically stimulated luminescence (OSL)-dated Early Weichselian buried podsols and overlying sands seem to indicate that central and western Finland remained ice free (Lunkka et al., 2004).

Major ice-sheet expansion in the southern Baltic region induced an ice stream to spread fanlike across Denmark (Houmark-Nielsen, 2010), and eventually glaciers covered most of the country early in the Middle Weichselian MIS 3

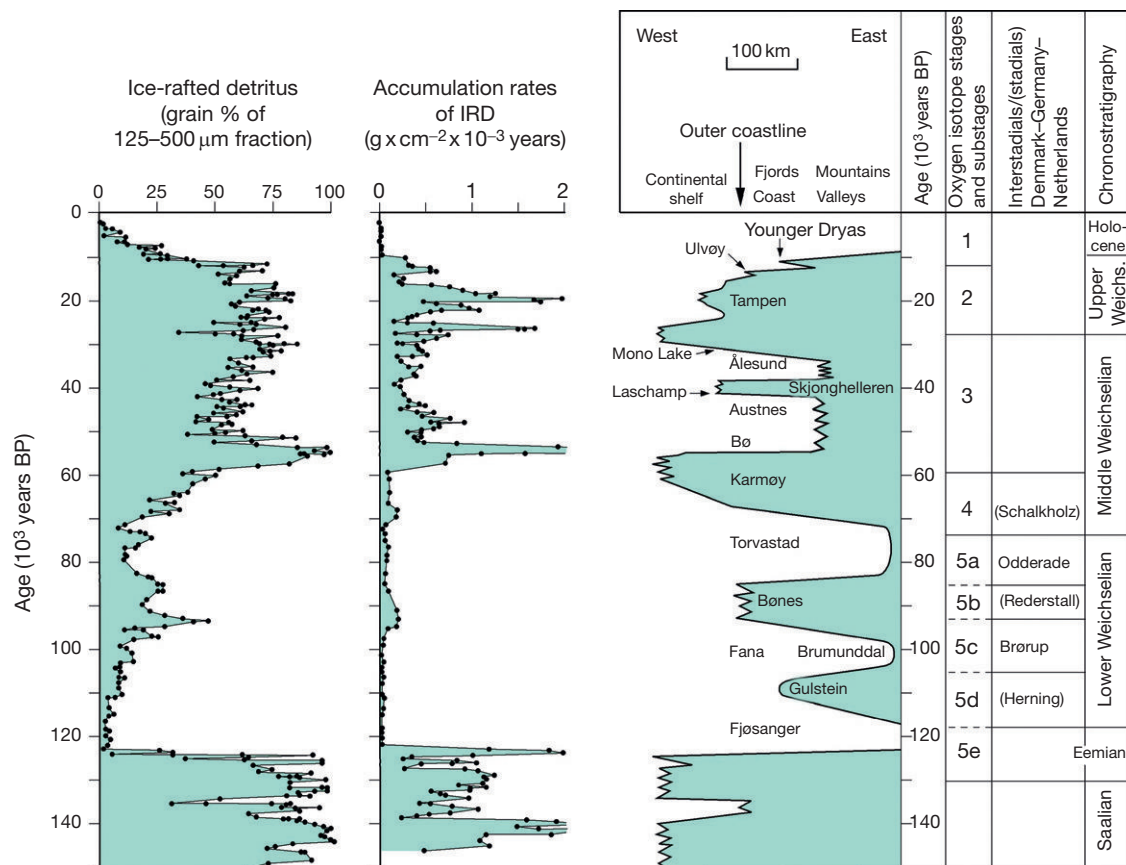


Figure 6 The right-hand panel shows a schematic glacialization curve for the southwestern flank of the Scandinavian Ice Sheet through the Weichselian. Reproduced from Mangerud J, Gyllencreutz R, Lohne Ø, and Svendsen JI (2011) Glacial history of Norway. In: Ehlers J, Gibbard PL, and Hughes PD (eds.) *Quaternary Glaciations – Extent and Chronology. Part IV. A Closer Look*. Amsterdam: Elsevier. The curve shows how the glaciers expanded from the mountains, out from the valleys and fjords, and finally onto the continental shelf. Names on the curve represent geological sites. The Laschamp and Mono Lake paleomagnetic excursions are marked with arrows. The curve to the left is a stacked record for grain % of IRD in five cores, and the middle curve a stacked record of accumulation rates in three cores; all cores collected from the Vøring Plateau (Baumann et al., 1995). The time scale is (in principle) in calendar years, but the curves are partly dated and correlated with different methods.

between 55 and 48 ka. For North Germany, [Stephan \(2003\)](#) claims a first Weichselian ice advance around 60 ka BP (Ellund Advance). [Müller \(2004\)](#) even postulates that the early ice advance reached the Weichselian maximum limit (Warnow Phase).

In northern Denmark and southwestern Sweden, deposits of the Kattegat ice lake were overridden by the Kattegat Ice Stream about 30–27 ka. The latter reached its maximum across central Denmark and in northwest Skåne and deposited till of Norwegian provenance ([Houmark-Nielsen, 2003; Larsen et al., 2009](#)), while periglacial conditions with Arctic treeless vegetation were maintained in southern Denmark. Deglaciation caused the Kattegat ice lake to reopen; however, it was still dammed off from the Atlantic Ocean by glacier ice flowing out of the Norwegian Channel ([Houmark-Nielsen, 2011](#)).

In Norway and Sweden, there was a major deglaciation after MIS 4 and much, probably even most, of Scandinavia was ice free during MIS 3 interstadials, especially during the Ålesund Interstadial ([Mangerud et al., 2010; Wohlfarth, 2010](#)).

Directly following the Middle Weichselian (MIS 3) sub-stage, the ice margin advanced to its maximum extent at the Brandenburg moraine and other moraines on the continent and to the Main Stationary Line in Denmark in MIS 2 around 22–20 ka BP ([Houmark-Nielsen, 2004](#)).

The major ice sheets of the Pleistocene cold stages spread very rapidly. This is best documented for the Late Weichselian glaciation. As late as 24 000 BP the major Alpine valleys were still all ice free. The same is true of southern Sweden at least as far north as Göteborg. If the glacial maximum in Europe was reached at about 20 000 BP, the ice sheet must have spread at a rate of no less than 75 m year^{-1} . This calculation only takes into account the progradation of the ice margin from Sweden to North Germany. If one considers that the erratic composition of the Late Weichselian tills indicates that ice from the glaciation center advanced to the areas south of the Baltic Sea, and that melting occurred at the margin, velocities of at least $100\text{--}150 \text{ m year}^{-1}$ must be assumed. Based on varve analyses, ice advance rates of 250 m year^{-1} were reconstructed for the first Elsterian and 140 m year^{-1} for the second Saalian ice advances in the Leipzig region, Germany ([Junge, 1998](#)).

Whereas a rapid advance of the Alpine glaciers seems plausible because of the short distances and the special character of the glaciation (ice-stream network), the rapid advances of the north European and North American ice sheets are hard to comprehend using traditional explanations. Instantaneous glacierization ([Ives et al., 1975](#)) and deformable beds (after [Boulton and Jones, 1979](#)) may have played a role.

The maximum extent of the Late Weichselian ice sheet south of the Baltic Sea is rather well known. The Weichselian ice did not cross the River Elbe. For example, the Eemian deposits at Lauenburg on the River Elbe are not overlain by till. The chalk-rich tills of the Lüneburg area further to the south are of Saalian age. In eastern Germany, the ice advanced more than 200 km inland. The classic morphostratigraphical subdivision of the Weichselian Glaciation was defined in that area. Here, the Brandenburg, Frankfurt, and Pomeranian 'Stages' were originally distinguished, each of which could be subdivided into several belts of end moraines. This morphological subdivision, however, is not reflected in the stratigraphy. Till sheets of three Late Weichselian ice advances can be

distinguished in northeast Germany. The maximum Brandenburg Advance is assumed to have taken place earlier than 32 000 BP, but no precise dates are available. The Pomeranian Advance occurred about 17 600 cal BP, while the Mecklenburg Advance, which is equivalent to the Fehmarn Advance of Schleswig-Holstein, and the Young Baltic Advance in Denmark are assigned to the period around 15 000 BP. It thus coincides with the Oldest Dryas Chronozone. Its deposits are overlain by sediments of the subsequent Meindorf Interstadial. Varve counts from the Meerfelder Maar have shown that this interstadial started at 14 450 varve years BP ([Litt et al., 2007; Stephan, 2003](#)).

The present cover of unconsolidated sediments on the landscape of Scandinavia was largely formed during the last glaciation. However, in places, pre-Late Weichselian sediments have been preserved, including, for instance, sediments underlying till in Gudbrandsdalen, Norway ([Bergersen and Garnes, 1981](#)) or the Veiki moraines in northern Sweden ([Lagerbäck and Robertsson, 1988](#)). It is thought that cold-based ice in the central parts of the ice sheet helped to preserve both pre-Weichselian landforms and deposits ([Kleman et al., 2008](#)).

The Lower Vistula region of Poland is the type region of the Vistulian (Weichselian) Glaciation ([Ehlers et al., 1995](#)). Five Vistulian till units have been identified here, overlying Eemian-age marine deposits. In Poland, the maximum advance of the Vistulian Glaciation extended about 50 km south of Poznań. The ice front advanced in several lobes, the most conspicuous of which was the Vistula lobe, which extended upstream to the Plock region. In the Suwałki region, an advance of narrow ice tongues caused valley-side tectonic deformation. Subglacial drainage resulted in the formation of a dense network of meltwater channels. Proglacial outwash accumulated sandur plains (formed by glacial meltwater), the sedimentary facies of which largely resemble the North German examples. The Vistulian ice advances are dated to about 15–20 ka BP according to thermoluminescence dates or ca. 14.5–20 ka BP by radiocarbon.

The eastern margin of the last Scandinavian Ice Sheet was first mapped and studied in the 1920s–1950s. The ice limit was correlated with the German Late Weichselian maximum by means of varve chronology, which was later supported by radiocarbon dating. The ice dome discharged in the form of four large ice streams distinguished by fabrics and pebble counts. The maximum ice advance in northern Belarus occurred after the deposition of nonglacial sediments dated by radiocarbon to ca. 18 ka BP ([Velichko and Faustova, 1986](#)). Further northward in the Vologda area, the ice advance terminated ca. 18 calendar ka BP, according to accelerator mass spectrometry (AMS) radiocarbon and OSL chronology ([Lunkka et al., 2001](#)). In the far north of the Russian Plain, in the Arkhangelsk region, including the northern tip of the Kanin Peninsula in the Barents Sea, the same methods indicated that the ice front progressed eastward until 17 ka BP ([Demidov et al., 2006](#)). Thus, in Russia, the maximum ice advance appears delayed in comparison to that in Germany and the British Isles.

Traces of earlier late Pleistocene ice advances by Scandinavian ice are dubious. Since the 1950s, there has been a school professing a separate Early Weichselian glaciation ice sheet that reached the northern part of the Moscow region ([Velichko and Faustova, 1986](#)). Tills of the so-called Kalinin (or Early Valdai) Glaciation, predating the classical, less-

extensive Late Valdai Glaciation, were pictured in many maps by the Russian Geological Survey. However, this concept proved erroneous as the mass of stratigraphical descriptions show that the Mikulino (Eemian) organic sediments in the area in question either lie on the surface or are overlain only by thin soliflucted diamicton without other members of true glaciogenic deposits (Gey and Malakhovsky, 2004; Velichko and Faustova, 1986).

Moreover, no till has been found in numerous boreholes in St. Petersburg, 60°N, intervening between the Eemian marine clay and the uppermost diamicton. The Late Weichselian till is instead underlain by lacustrine sands and silts containing pollen of spruce, pine, and dwarf birch. Radiocarbon dates have given ages from 46 to 40 ka BP suggesting, if correct, that the deposits represent the first warming of the Middle Valdai 'megainterstadial.' The start of a cooling with a periglacial herb flora was dated to ca. 39 ¹⁴C ka BP. Farther south, similar subglacial sediments show a couple of additional warming events with an intermediate cooling indicated by pollen of birch parkland with xerophytic herbs (Arslanov et al., 1981). The radiocarbon dates 29–25.5 ka BP from the final warming suggest its correlation with the Denekamp Interstadial of western Europe and with the cryoturbated Bryansk paleosol, which is found to be widespread in periglacial Russia. Velichko et al. (2004) identify three relatively warm episodes within the Middle Valdai interval of 50–24 ka BP and two more interstadials within the Early Valdai Substage, all separated by cool episodes with periglacial vegetation. All short warming episodes (also the Allerød Interstadial) are marked by the pollen of boreal coniferous-birch forests. From Russia, no tills predating the Late Valdai have been described south of the Baltic Shield.

The ice recession is punctuated by several morainic assemblages located in a ~600-km-wide belt between the Valdai Upland and the Finnish border. The rank and origin of the distal hummocky agglomerations (Veps, Krestsy) is uncertain because significant readvances have not strictly been proven and radiocarbon dating of the interstadials so far is insufficient (Gey and Malakhovsky, 2004).

The most active was the Luga stadial readvance that reached up to 100–150 km south of St. Petersburg. It occurred just prior to 14.2 calendar ka BP according to paleomagnetic, varve, and AMS radiocarbon chronologies. The next phase, the Neva Phase, is dated to 13.3 calendar ka BP. After that, during the Allerød interval, the ice margin retreated rapidly from St. Petersburg to the Finnish border (some 350–500 km in 1500 years) leaving behind the largest European lakes Ladoga and Onega at the place of former major ice lobes. The last Younger Dryas Stadial readvance stood at the double Salpausselkä End Moraine complex between 12.25 and 11.6 cal ka BP (Saarnisto and Saarinen, 2001).

British/Irish Ice Sheet

Information on the extent of Early or Middle Devensian (Weichselian=MIS 4) ice in Britain is still very scarce. In England, concrete evidence for glaciation during the mid-Devensian is found only in the Lake District. Here, new evidence from boreholes indicates that valley glaciers formed in Wasdale and other valleys. At Wasdale, the ice advanced

periodically into a proglacial lake in the Ravensdale, Seascale, and Lower Wasdale areas. Early-Devensian Substage glaciation has also been reconstructed for NE Scotland, based on amino acid determinations. In Ireland, poorly dated glacial deposits may also belong to limited Middle Devensian ice sheets (Ehlers et al., 1991).

In both Britain and Ireland, the main Devensian glaciation occurred in the later part of the cold stage (MIS 2). In Britain, the Late Devensian glaciation is referred to as the Dimlington Stadial. Its glaciers advanced far south along the east coast of Britain, touching the northern coast of Norfolk. An ice tongue, following the Vale of York, advanced into Lincolnshire, damming the Trent to form proglacial Lake Humber. In the Pennines, ice flow generally followed the major valleys. The ice sheet also intruded into the Wash, damming up a lake in the southern Fenland. In the North Sea, the Dogger Bank remained ice free, but major parts of central and northern North Sea were glaciated (Bradwell et al., 2008; Clark et al., 2012).

In the Irish Sea, ice advanced far to the south, touching the south Wales coast and possibly reaching as far as the Isles of Scilly. However, many details of the glacial history are still far from clear. Not only are there problems of connecting the British and Irish stratigraphic sequences with those of continental Europe, but also with one another. The last glaciation of Ireland, the Midlandian or Fenitian Glaciation, consisted of a major Irish Ice Sheet and a smaller Kerry/Cork Ice Sheet in the southwest. The flow lines and boundaries of the ice sheets have been reconstructed from landforms, striae, and indicator erratics. For example, the distribution of Galway Granite erratics clearly demonstrates the southern extent of an Irish Ice Sheet, but this seems to have been older than the last glaciation. The distribution of periglacial features suggests that a corridor more than 50 km wide between the two Irish ice sheets remained unglaciated (Knight et al., 2004).

The dynamics of the Irish ice sheets are still hotly debated. Landforms along the west coast seem to indicate discharge of a major ice stream via Sligo Bay into the North Atlantic, forming a calving bay. Isostatic depression under the weight of the ice sheet is also demonstrated, for instance by the occurrence of an ice-contact glaciomarine delta in the Upper Carey Valley, Northern Ireland, and by shallow marine deposits at Portballintrae. Isostatic downwarping in the Irish Sea basin is also thought to have resulted in marine contact and glaciomarine deposition. The widespread occurrence of what is referred to as 'Irish Sea Till,' a shelly diamicton found on the western shores of the Irish Sea, provides the evidence for this advance. This was formerly interpreted as a product of shelf ice. In Counties Dublin and Wicklow, however, it extends to altitudes of 300 m O.D., clearly beyond any marine limit; therefore, terrestrial ice must also have been involved. As a result of glacioisostatic depression, development of a calving bay may have resulted in rapid disintegration of the combined British/Irish ice sheet at the end of the glaciation.

In northern Scotland, the Hebrides were not completely crossed by mainland ice during the Devensian Stage. There is evidence of an older glaciation that twice covered the north-western Hebridean shelf to the shelf break, but the age of these ice advances is still uncertain. The Outer Hebrides supported a local ice cap during the Late Devensian. The Inner Hebrides

and the Orkney Islands were overridden by Scottish ice during the Late Devensian, whereas Shetland and the Færøer Islands supported local ice caps. It has been postulated that parts of Scotland, mainly Caithness and Buchan, remained ice free during the last glaciation. Buchan is an area where deep pre-Quaternary weathering products have been preserved. However, more recent investigations seem to indicate that both areas were probably covered by Late Devensian ice. As on the continent, the unglaciated parts of Britain and Ireland underwent strong periglacial reworking (Ballantyne and Harris, 1994).

The extent of the Late Weichselian ice sheet in the North Sea is still a matter of debate. The distribution of buried subglacial channels on the North Sea floor suggests that the last British and Scandinavian Ice Sheets did meet. Recently, the investigation of sediment cores has shown that both ice sheets were in contact in the northern North Sea. AMS radiocarbon dates show that this event took place rather early, between 29 400 and ca. 22 000 BP. This ice advance thus predates the Late Devensian Dimlington Stadial.

The Alps

At the northern shore of Lake Mondsee (Austria), a thick sequence of limnic deposits was exposed, both over- and underlain by till. Palynological investigations revealed that the lake deposits comprised the last interglacial plus four Early Weichselian interstadials. The Eemian vegetational succession closely resembles that of Samerberg. The climatic optimum is represented by a mixed oak forest, which is rich in *Ilex* and *Taxus*. This suggests that the mean annual temperature in the Eemian was some 2–3 °C warmer than during the Holocene climatic optimum. At the end of the Eemian, similar to northern Germany, complete deforestation followed. The Early Weichselian interstadials saw a reimmigration of trees. The interstadial forest was dominated by *Picea* and, to a lesser degree, also by *Quercus* (oak) and *Taxus* (ash) (Drescher-Schneider, 2000).

In the Alps, initially a single undivided Würmian Glaciation was envisaged, subdivided only by a series of recessional end moraines. However, today it is quite clear that the Alps, like northern Europe and North America, experienced a Mid-Würmian ice advance. Only its extent is still a matter of debate.

Based on stratigraphical investigations at Thalgut, Jaberg, Thungschneit, and Cossonay, it can be said that the last cold stage in Switzerland was represented by two major ice advances, the glaciers of which reached well into the alpine foreland, the earlier having been less extensive than the later ice advance. The sections at Gossau and at Walensee indicate that, from 60 000 to 28 000 BP, an ice-free period occurred between the two ice advances. In this period, three interstadials occurred, all characterized by coniferous forest (Schlüchter, 2004). At Wangen on the River Aare, an erratic block on the Würmian moraine could be dated by cosmogenic nuclides at $20\,100 \pm 1000$ years BP (Ivy-Ochs et al., 2004). Also, in the French Alps, the main advance of the Würmian ice has been dated to around 20 000 BP (Buoncristiani and Campy, 2004).

In contrast to the results from France and Switzerland, the Samerberg (Bavaria) and Mondsee (Austria) profiles show an uninterrupted sequence of lake sediments between unequivocal deposits of the Riss/Würm Interglacial (Eemian Stage) and the Late Würmian till (approximately 20 000 BP). Also, in the Austrian Traun valley, traces of only one Würmian ice advance have been found. Here, there is no evidence of a possible advance during the Early or Middle Würmian Substages. The main ice advance of the last glaciation accordingly started only after 25 000–24 000 BP and reached its maximum around 20 000 BP as in northern Europe (van Husen, 2011). The extent of the Würmian glacial maximum is shown in Figure 7.

Within the area of the last glaciation, extensive accumulations of lake sediments that are assigned to the Middle Würmian by means of radiocarbon dating have been preserved in some places. The most prominent occurrence of pre-Late Würmian lacustrine deposits are the varved clays of Baumkirchen, 12 km east of Innsbruck, which seem to postdate the Middle Würmian interstadial. During their deposition, the



Figure 7 Extent of the Würmian Alpine glaciation. Reproduced from Ehlers J (2011) *Das Eiszeitalter*, p. 363. Heidelberg: Spektrum.

climate was already rather cool: in the vicinity of the lake shrub, tundra was dominant. The sequence of varved clays, over a hundred meters thick, is overlain by ~70 m of outwash gravels, which are, in turn, covered by a thick till. The varved clays were deposited in a shallow lake, as indicated by trace fossils of fish on the bedding planes. The sedimentation rate was constantly high, reaching an average of 5 cm per year. At Baumkirchen, the boundary between the Middle and Late Würmian Substages has been defined. Radiocarbon dates on wood (*Pinus*, *Alnus*, and *Hippophae*) from the varved clays have yielded ages of between $31\,600 \pm 1300$ and $26\,800 \pm 1300$ years, giving a maximum age for the Late Würmian glaciation (Fliri, 1973).

In contrast to the Scandinavian glaciation area, where the thickness of the ice can be determined only at very few points because of the almost total absence of nunataks, the maximum height of the Alpine ice streams can be precisely reconstructed by mapping the upper limits of erratics. During the Würmian glacial maximum, transfluence occurred at various points of the ice-stream network. In the Inn valley, for instance, around Landeck, five tributary glaciers merged to form one major valley glacier, advancing eastward down the Inn valley. At the same time, a major tributary glacier from the south had already blocked the Inn valley around Innsbruck. Congestion resulted in a rise of the ice surface until the threshold to the north was finally crossed, and ice from the Inn valley flowed over the Fernpass and the Seefelder Senke due north into the Isar and Loisach drainage systems. Such events are reflected in the supply of gravel and major erratics. In the foreland of the Isar–Loisach Glacier, Alpine crystalline rocks constitute up to 35% of the total, in the 20–31.5 mm fraction. In contrast, in the Tölz Glacier lobe, adjoining the east, the contents range only between 0 and 1%, as no ice crossed over from the Inn valley (Dreesbach, 1985).

Both north and south of the Alpine main ranges, the Late Würmian glacial maximum was followed by a long stillstand during which the ice margin stagnated within a few hundred meters behind the outermost moraine. During this halt, a morphologically more distinct end moraine than that during the Würmian maximum accumulated, and the main gravel body of the Lower Terrace also originated from the position of this moraine. Glaciers rapidly collapsed before 17 500 BP. A series of minor readvances occurred during the Late Glacial, but the glaciers barely reached the main Alpine valleys. The last of these advances formed the Egesen moraine at about 11 800 BP, that is, during the Younger Dryas Chronozone (Jerz, 1993).

Other Areas

There are other formerly glaciated areas, such as the Pyrenees, the Carpathians, and the Caucasus, or the high mountains of Italy, Greece, Turkey, and the Balkans, which provide interesting information on former glaciation levels and equilibrium line altitudes.

Many of the other Asian mountain ranges, such as the Verkhoyansk Mountains, the Jablonovyi, or the mountains of Kamchatka, have not yet been mapped in any detail, and comparison of the available maps with satellite imagery or a

Digital Elevation Model makes it immediately clear that much more work is needed in those areas. Remote sensing provides excellent tools to start such investigations, with imagery up to a scale of over 1:5000 available through the QuickBird and Ikonos satellites. However, apart from all the technical progress, the decisive work to answer the open questions will still need to be done in the field.

Deglaciation

The end of the Weichselian cold stage is characterized by a number of climatic oscillations, which for a long time have been used for stratigraphical subdivision (Mangerud et al., 1974). Especially the mild Bølling and Allerød interstadials, and the cold Younger Dryas stadial are seen in a number of proxy records such as pollen, beetles, and marine fossils, in addition to the glacial fluctuations.

Bock et al. (1985) also distinguished the Meiendorf Interstadial before the Bølling. According to the distribution of *Hippophae* in North Germany, the Meiendorf Interstadial might have had climatic conditions similar to the Allerød Interstadial, although there were no trees. During the Allerød, North Germany and adjoining areas were covered by a mixed forest of birch and pine.

A consistent mapping of the pattern of the deglaciation has turned out to be difficult. One reason for this is that the deglaciation was driven by different forces. Certainly, warmer summers were the main factor driving the ice retreat and the dominant factor in determining where the ice sheet ended on land. However, much of the ice margin was located in lakes, particularly the Baltic Ice Lake, and along the western margin in the sea, and in these areas calving and sea-water temperature were more important factors. It is a common feature that the ice in the fjords broke up before the adjacent land was ice free.

The extent of the Younger Dryas readvance is well known (Figure 8). The Younger Dryas moraines are the only moraines that can be mapped continuously around the entire Scandinavian Ice Sheet (Andersen et al., 1995a). The interpretation of the main pattern is that summers became colder and thus the retreat halted, or there was a readvance. A number of differences in reactions are, however, superimposed on this main pattern. In western Norway, there was a readvance that started during the late Allerød and continued to the very end of the Younger Dryas (Bondevik and Mangerud, 2002; Lohne et al., 2007), whereas in Trøndelag, some 400 km farther north, the ice margin retreated during the Younger Dryas (Andersen et al., 1995b). This means that in some places around Scandinavia, the outermost Younger Dryas moraines were formed at the beginning of the Younger Dryas, whereas in other areas they were formed at the very end. The Baltic Ice Lake was drained and the Baltic connected to the sea at the end of the Younger Dryas (Lundqvist, 2004) to form the Yoldia Sea (Figure 9).

See also: **Glaciations:** Late Quaternary in Highland Asia; Middle Pleistocene in Eurasia; Overview.



Figure 8 Extent of the Younger Dryas ice sheet in Scandinavia. Reproduced from Ehlers J and Gibbard PL (eds.) (2011) *Quaternary Glaciations – Extent and Chronology, Part IV – A Closer Look. Developments in Quaternary Science*, vol. 15. Amsterdam: Elsevier.



Figure 9 Extent of the Baltic Yoldia Sea. Reproduced from Ehlers J (2011) *Das Eiszeitalter*, p. 363. Heidelberg: Spektrum.

References

- Andersen BG, Lundqvist J, and Saarnisto M (1995) The Younger Dryas margin of the Scandinavian Ice Sheet – An introduction. *Quaternary International* 28: 145–146.
- Andersen BG, Mangerud J, Sørensen R, et al. (1995) Younger Dryas ice-marginal deposits in Norway. *Quaternary International* 28: 147–169.
- Arslanov KhA, Breslav SL, Zarrina YeP, et al. (1981) The climatostratigraphy and chronology of the Middle Valdai in the northwestern and central Russian Plain. In: Velichko AA and Faustova MA (eds.) *Pleistotsenovye oledneniya Vostochno-Yevropeiskoi ravniny (Pleistocene Glaciations of the East European Plain)*, pp. 12–27. Moscow: Nauka (in Russian).
- Astakhov VI (1977) The Kara centre of the Pleistocene glaciation as reconstructed from ancient moraines of West Siberia. *Materialy glatsiologicheskikh issledovaniy (Data of Glaciological Studies)* 30: 60–69 (in Russian, English summary).
- Astakhov V (1998) The last ice sheet of the Kara Sea: Terrestrial constraints on its age. *Quaternary International* 45/46: 19–29.
- Astakhov VI (2006) Evidence of Late Pleistocene ice-dammed lakes in West Siberia. *Boreas* 35: 607–621.
- Astakhov V and Nazarov D (2010) Correlation of Upper Pleistocene sediments in northern West Siberia. *Quaternary Science Reviews* 29: 3615–3629.
- Ballantyne CK and Harris C (1994) *The Periglaciation of Great Britain*. Cambridge: Cambridge University Press.
- Baumann K-H, Lackschewitz KS, Mangerud J, et al. (1995) Reflection of Scandinavian Ice Sheet fluctuations in Norwegian Sea sediments during the last 150,000 years. *Quaternary Research* 43: 185–197.
- Bergersen O and Garnes K (1981) Weichselian in central South Norway: The Gudbrandsdal Interstadial and following glaciation. *Boreas* 10: 315–322.
- Bock W, Menke B, Strehl E, and Ziemus H (1985) Neuere Funde des Weichselspätglazials in Schleswig-Holstein. *Eiszeitalter und Gegenwart* 35: 161–180.
- Bondevik S and Mangerud J (2002) A calendar age estimate of a very late Younger Dryas ice sheet maximum in western Norway. *Quaternary Science Reviews* 21: 1661–1676.
- Boulton GS and Jones AS (1979) Stability of temperate ice caps and ice sheets resting on beds of deformable sediment. *Journal of Glaciology* 24: 29–43.
- Bradwell T, Stoker MS, Gollidge NR, et al. (2008) The northern sector of the last British Ice Sheet: Maximum extent and demise. *Earth-Science Reviews* 88: 207–226.
- Buoncrisiani J-F and Campy M (2004) The palaeogeography of the last two glacial episodes in France: The Alps and Jura. In: Ehlers J and Gibbard PL (eds.) *Quaternary Glaciations – Extent and Chronology, Part I, Europe Developments in Quaternary Science*, vol. 2a, pp. 101–110. Amsterdam: Elsevier.
- Clark CD, Hughes ALC, Greenwood SL, Jordan C, and Sejrup HP (2012) Pattern and timing of retreat of the last British–Irish Ice Sheet. *Quaternary Science Reviews* 44: 112–146.
- Demidov IN, Houmark-Nielsen M, Kjær KH, and Larsen E (2006) The last Scandinavian Ice Sheet in northwestern Russia: Ice flow patterns and decay dynamics. *Boreas* 35: 425–443.
- Dreesbach R (1985) *Sedimentpetrographische Untersuchungen zur Stratigraphie des Würmglazials im Bereich des Isar-Loisachgletschers*, p. 176. Dissertation, München.
- Drescher-Schneider R (2000) Die Vegetations- und Klimaentwicklung im Riß/Würm-Interglazial und im Früh- und Mittelwürm in der Umgebung von Mondsee. *Mitteilungen der Kommission für Quartärforschung der Österreichischen Akademie der Wissenschaften* 12: 37–92.
- Ehlers J, Gibbard P, and Rose J (eds.) (1991) *Glacial Deposits in Great Britain and Ireland*. Rotterdam: Balkema.
- Ehlers J, Kozarski S, and Gibbard PL (eds.) (1995) *Glacial Deposits in North-East Europe*. Rotterdam: Balkema.
- Fliri F (1973) Beiträge zur Geschichte der alpinen Würmvereisung: Forschungen am Bänderton von Baumkirchen (Inntal, Nordtirol). *Zeitschrift für Geomorphologie N. F., Supplement-Band* 16: 1–14.
- Funder S, Demidov I, and Yelovicheva Y (2002) Hydrography and mollusc faunas of the Baltic and the White Sea–North Sea seaway in the Eemian. *Palaeogeography, Palaeoclimatology, Palaeoecology* 184: 275–304.
- Gey VP and Malakhovsky DB (2004) On the age and extent of the maximum Late Pleistocene ice advance along the Baltic-Caspian watershed. In: Ehlers J and Gibbard PL (eds.) *Quaternary Glaciations – Extent and Chronology, Part 1: Europe Development in Quaternary science*, vol. 2, pp. 355–358. Amsterdam: Elsevier.
- Grosswald MG (1980) Late Weichselian ice sheets of Northern Eurasia. *Quaternary Research* 13: 1–32.
- Helmens KF, Räsänen ME, Johansson PW, Jungner H, and Korjonen K (2000) The last interglacial–glacial cycle in NE Fennoscandia: A nearly continuous record from Sokli (Finnish Lapland). *Quaternary Science Reviews* 19: 1605–1623.
- Hirvas H (1991) Pleistocene stratigraphy of Finnish Lapland. *Geological Survey of Finland Bulletin* 354: 123.
- Hjort C, Möller P, and Alexanderson H (2004) Weichselian glaciation of the Taymyr Peninsula, Siberia. In: Ehlers J and Gibbard PL (eds.) *Quaternary Glaciations – Extent and Chronology, Part I: Europe Developments in Quaternary Science*, vol. 2a, pp. 359–367. Amsterdam: Elsevier.
- Houmark-Nielsen M (2003) Signature and timing of the Kattegat Ice Stream: Onset of the LGM-sequence in the southwestern part of the Scandinavian Ice Sheet. *Boreas* 32: 227–241.
- Houmark-Nielsen M (2004) The Pleistocene of Denmark: A review of stratigraphy and glaciation history. In: Ehlers J and Gibbard PL (eds.) *Quaternary Glaciations – Extent and Chronology, Part I: Europe Developments in Quaternary Science*, vol. 2a, pp. 35–46. Amsterdam: Elsevier.
- Houmark-Nielsen M (2010) Extent, age and dynamics of Marine Isotope Stage-3 glaciations in the south western Baltic Basin. *Boreas* 39: 343–359.
- Houmark-Nielsen M (2011) Pleistocene glaciations in Denmark: A closer look at chronology, ice dynamics and landforms. In: Ehlers J, Gibbard PL, and Hughes PD (eds.) *Quaternary Glaciations – Extent and Chronology, Part IV – A Closer Look Developments in Quaternary Science*, vol. 15, pp. 47–58. Amsterdam: Elsevier.
- Ives JD, Andrews JT, and Barry RG (1975) Growth and decay of the Laurentide Ice Sheet and comparisons with Fennoscandia. *Naturwissenschaften* 62: 118–125.
- Ivy-Ochs S, Schäfer J, Kubik PW, Synal H-A, and Schlüchter C (2004) Timing of deglaciation on the northern alpine foreland (Switzerland). *Eclogae Geologicae Helvetiae* 97: 47–55.
- Jerz H (1993) *Das Eiszeitalter in Bayern – Erdgeschichte, Gesteine, Wasser, Boden*. Schweizerbart: Geologie von Bayern, Bd. 2.
- Junge FW (1998) Die Bändertone Mitteldeutschlands und angrenzender Gebiete. *Altenburger Naturwissenschaftliche Forschungen* 9: 210.
- Kind NV (1974) *Late Quaternary Geochronology According to Isotopes Data (Geokhronologia pozdnego antropogena po izotopnym dannym)*, p. 255. Moscow: Nauka (in Russian).
- Kleman J, Stroeven AP, and Lundqvist J (2008) Patterns of Quaternary ice sheet erosion and deposition in Fennoscandia and a theoretical framework for explanation. *Geomorphology* 97: 73–90.
- Knight J, Coxon P, McCabe AM, and McCarron SG (2004) Pleistocene glaciations in Ireland. In: Ehlers J and Gibbard PL (eds.) *Quaternary Glaciations – Extent and Chronology, Part I: Europe Developments in Quaternary Science*, vol. 2a, pp. 183–191. Amsterdam: Elsevier.
- Lagerbäck R and Robertsson A-M (1988) Kettle holes – Stratigraphical archives for Weichselian geology and palaeoenvironment in northernmost Sweden. *Boreas* 17: 439–468.
- Landvik JY, Bondevik S, Elverhøi A, et al. (1998) The Last Glacial Maximum of Svalbard and the Barents Sea area: Ice sheet extent and configuration. *Quaternary Science Reviews* 17: 43–75.
- Larsen NK, Knudsen KL, Krohn CF, Kronborg C, Murray AS, and Nielsen OB (2009) Late Quaternary ice sheet, lake and sea history of southwest Scandinavia – A synthesis. *Boreas* 38: 732–761.
- Litt T, Behre K-E, Meyer K-D, Stephan H-J, and Wansa S (2007) Stratigraphische Begriffe für das Quartär des norddeutschen Vereisungsgebietes. *Eiszeitalter und Gegenwart* 56(1/2): 7–65.
- Lohne ØS, Bondevik S, Mangerud J, and Svendsen JI (2007) Sea-level fluctuations imply that the Younger Dryas ice-sheet expansion in western Norway commenced during the Allerød. *Quaternary Science Reviews* 26: 2128–2151.
- Lundqvist J (2004) Glacial history of Sweden. In: Ehlers J and Gibbard PL (eds.) *Quaternary Glaciations – Extent and Chronology, Part I: Europe Developments in Quaternary Science*, vol. 2a, pp. 401–412. Amsterdam: Elsevier.
- Lunkka JP, Johansson P, Saarnisto M, and Sallasmaa O (2004) Glaciation of Finland. In: Ehlers J and Gibbard PL (eds.) *Quaternary Glaciations – Extent and Chronology, Part I: Europe Developments in Quaternary Science*, vol. 2a, pp. 93–100. Amsterdam: Elsevier.
- Lunkka JP, Saarnisto M, Gey V, Demidov I, and Kiselova V (2001) Extent and age of the last glacial maximum in the southeastern sector of the Scandinavian Ice Sheet. *Global and Planetary Change* 31(1–4): 407–425.
- Makaveyev AN, Aseyev AA, Astakhov VI, et al. (1982) Geomorphological criteria for the reconstruction of Late Pleistocene ice caps in Northern Eurasia. *Polar Geography and Geology* VI(1): 25–46.
- Mangerud J (2004) Ice sheet limits on Norway and the Norwegian continental shelf. In: Ehlers J and Gibbard PL (eds.) *Quaternary Glaciations – Extent and Chronology, Part I: Europe Developments in Quaternary Science*, vol. 2a, pp. 271–294. Amsterdam: Elsevier.

- Mangerud J, Andersen ST, Berglund BE, and Donner JJ (1974) Quaternary stratigraphy of Norden, a proposal for terminology and classification. *Boreas* 3: 109–128.
- Mangerud J, Astakhov V, and Svendsen J-I (2002) The extent of the Barents–Kara Ice Sheet during the Last Glacial Maximum. *Quaternary Science Reviews* 21(1–3): 111–119.
- Mangerud J, Gosse J, Matiouchkov A, and Dolvik T (2008) Glaciers in the Polar Urals, Russia, were not much larger during the Last Global Glacial Maximum than today. *Quaternary Science Reviews* 27: 1047–1057.
- Mangerud J, Gulliksen S, and Larsen E (2010) ^{14}C -dated fluctuations of the western flank of the Scandinavian Ice Sheet 45–25 kyr BP compared with Bølling–Younger Dryas fluctuations and Dansgaard–Oeschger events in Greenland. *Boreas* 39: 328–342.
- Mangerud J, Jakobsson M, Alexanderson H, et al. (2004) Ice-dammed lakes and rerouting of the drainage of Northern Eurasia during the last glaciation. *Quaternary Science Reviews* 23: 1313–1332.
- Müller U (2004) Weichsel–Frühglazial in Nordwest-Mecklenburg. *Meyniana* 56: 81–115.
- Saarnisto M and Saarinen T (2001) Deglaciation chronology of the Scandinavian Ice Sheet from the Lake Onega to the Salpausselkä End Moraines. *Global and Planetary Change* 31(1–4): 387–405.
- Schlichter C (2004) The Swiss glacial record – A schematic summary. In: Ehlers J and Gibbard PL (eds.) *Quaternary Glaciations – Extent and Chronology. Part I: Europe Developments in Quaternary Science*, vol. 2, pp. 413–418. Amsterdam: Elsevier.
- Shackleton NJ, Sanchez-Goni MF, Pailler D, and Lancelot Y (2003) Marine isotope Substage 5e and the Eemian interglacial. *Global and Planetary Change* 36: 151–155.
- Stauch G and Gualtieri L (2008) Late Quaternary glaciations in northeastern Russia. *Journal of Quaternary Science* 23: 545–558.
- Stephan H-J (2003) Zur Entstehung der eiszeitlichen Landschaft Schleswig-Holsteins. *Schriften des naturwissenschaftlichen Vereins für Schleswig-Holstein* 67: 101–118.
- Svendsen JI, Alexanderson H, Astakhov VI, et al. (2004) Late Quaternary ice sheet history of Northern Eurasia. *Quaternary Science Reviews* 23: 1229–1271.
- Svendsen JI, Heggen HP, Hufthammer AK, Mangerud J, Pavlov P, and Roebroeks W (2010) Geo-archaeological investigations of Palaeolithic sites along the Ural Mountains: On the northern presence of humans during the last Ice Age. *Quaternary Science Reviews* 29: 3138–3156.
- van Husen D (2011) Quaternary glaciations in Austria. In: Ehlers J, Gibbard PL, and Hughes PD (eds.) *Quaternary Glaciations – Extent and Chronology, Part IV – A Closer Look. Developments in Quaternary Science*, vol. 15, 15–28. Amsterdam: Elsevier.
- Velichko AA and Faustova MA (1986) Glaciations in the East European region of the USSR. *Quaternary Science Reviews* 5: 447–461.
- Velichko AA, Faustova MA, Gribchenko YuN, Pisareva VV, and Sudakova NG (2004) Glaciations of the East European Plain – Distribution and chronology. In: Ehlers J and Gibbard PL (eds.) *Quaternary Glaciations – Extent and Chronology. Part 1: Europe Developments in Quaternary Science*, vol. 2, pp. 337–354. Amsterdam: Elsevier.
- Wohlfarth B (2010) Ice-free conditions in Sweden during Marine Oxygen Isotope Stage 3? *Boreas* 39: 377–398.

Late Quaternary in Highland Asia

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Introduction

Central Asia is home to the world's greatest topography that includes the mountains of the Himalaya, Tibet, Pamir, Tien Shan, and Altai, which together constitute Highland Asia (Figure 1). Within Highland Asia, the Himalaya, the Trans Himalaya, and the Tibetan Plateau are by far the most striking and stretch E–W for ~2000 km from Burma to Afghanistan and northward for ~1500 km from India to central China, with Tibet having an average elevation of ~5000 m above sea level (asl). The average elevation for the Himalaya and the Trans Himalaya is less, but they contain numerous peaks that exceed 6000 m asl and all of the world's 8000 m high peaks. The region also has the land's greatest relative relief, which commonly exceeds several kilometers. Together, the mountains of Highland Asia form the most glaciated track outside of the Polar Region and they have a profound influence on regional and global atmospheric circulation, and the hydrology of some of the world's greatest rivers, which include the Indus, the Ganges, the Yellow River, and the Yangtze (Lehmkuhl and Owen, 2005).

Four major climatic systems influence the region, which include (1) the mid-latitude westerlies, (2) the south Asian monsoon, (3) the Mongolian high-pressure system, and (4) the El Niño Southern Oscillation (ENSO). The relative importance of these climate systems varies throughout the region and has likely varied significantly, both spatially and temporally throughout the Quaternary (Lehmkuhl and Owen, 2005). Presently, the southern slopes of the Himalaya and the high mountains of eastern Tibet receive snowfall mainly in the summer monsoon season. In contrast, northern and western Tibet, and ranges such as the Karakoram Mountains, Pamir, Tien Shan, and Altai, receive heavy snowfall during the winter with moisture supplied from the mid-latitude westerlies and the Mongolian high-pressure system. This snow supports ice caps and valley glaciers throughout Tibet and the bordering mountains. The regional snowline varies significantly throughout the region; for example, it is 4600–5600 m asl and 6000–6200 m asl south and north of Everest, respectively, 5200–5700 m asl in the Garhwal Himalaya, 4500–4700 m asl around Nanga Parbat, 4600–5300 m asl on the south side of the Karakoram Mountains, and 5200–5700 m asl on the northern side of the Karakoram (Owen and Benn, 2005 and references therein). The glacial and hydrological systems in these mountains provide water for hundreds of millions of people in central Asia and understanding the spatial and temporal variations in climate, glaciation, and hydrology clearly has important socioeconomic implications.

In recent years, there has been increasing recognition of the geologic, climatic, and anthropogenic importance of the region. This has resulted in an increasing number of studies on the nature, extent, and precise timing of glaciation. Many of these studies have involved detailed fieldwork, the application

of newly developing remote sensing technologies, and extensive programs of numerical dating. Yet, there is much more to be known about Late Quaternary glaciation, and the associated hydrological and climatic responses are presently very poorly defined. This is partially due to the logistical and political inaccessibility of the region; it also reflects the continued need to develop more precise ways to reconstruct the extent and to define the timing of former glacial events.

Intense fluvial and glacial erosion within the mountains of High Asia often destroys diagnostic morphologies of glacial and mass movement landforms, making their identification difficult. In addition, diamictos that comprise mass movement and glacial landforms look very similar and have similar particle size distributions and particle shapes. The misinterpretation of glacial and mass movement landforms and sediments within many of the ranges in Highland Asia has led to erroneous reconstructions of former glacier extent (Hewitt, 1999). In some cases, this has resulted in overestimates of glacial extent, and in an extreme case has helped perpetuate the theory that an extensive ice sheet covered Tibet during the last glacial cycle (Kuhle, 1985). The existence of extensive ice sheet on Tibet was hotly debated during the 1990s, but is no longer considered plausible based on the geologic evidence now available (Owen et al., 1998, 2005; Rutter, 1995; Schäfer et al., 2002; Seong et al., 2008). To provide a framework for accurate glacial reconstructions in Highland Asia, Benn and Owen (2002) provided a review and presented new studies on the sedimentology and geomorphology of Himalayan glaciers to aid and help emphasize the need for detailed analysis of landforms using modern sedimentological techniques.

The lack of organic material preserved in glacial and associated landforms and sediments throughout much of Highland Asia has inhibited the use of radiocarbon dating to help define the timing of glaciation, and where present, most organic matter is Holocene in age (Owen, 2009). However, with the rapid development and application of optically stimulated luminescence (OSL) and terrestrial cosmogenic nuclide (TCN) surface exposure dating throughout the early part of the twenty-first century, many of the glacial successions throughout Highland Asia have been dated (e.g., Barnard et al., 2006; Dortch et al., 2010; Finkel et al., 2005; Hedrick et al., 2011; Owen et al., 2001, 2002b, 2003a,b,c, 2005, 2006a,b, 2008, 2009, 2010; Phillips et al., 2000; Richards et al., 2000a,b; Schäfer et al., 2002, 2008; Scherler et al., 2010; Seong et al., 2007, 2009; Tsukamoto et al., 2002; Zech et al., 2005, 2009). Although both techniques have inherent and associated problems (Fuchs and Owen, 2008; Owen et al., 2008), careful sampling design (Benn and Owen, 2002) and combining the two sets of methods have enabled robust chronologies to be developed (e.g., Owen et al., 2009).

Schäfer et al. (2008), for example, illustrated some of the challenges in applying TCN methods to dating moraines in their study of the glacial successions in Nyalam County in the

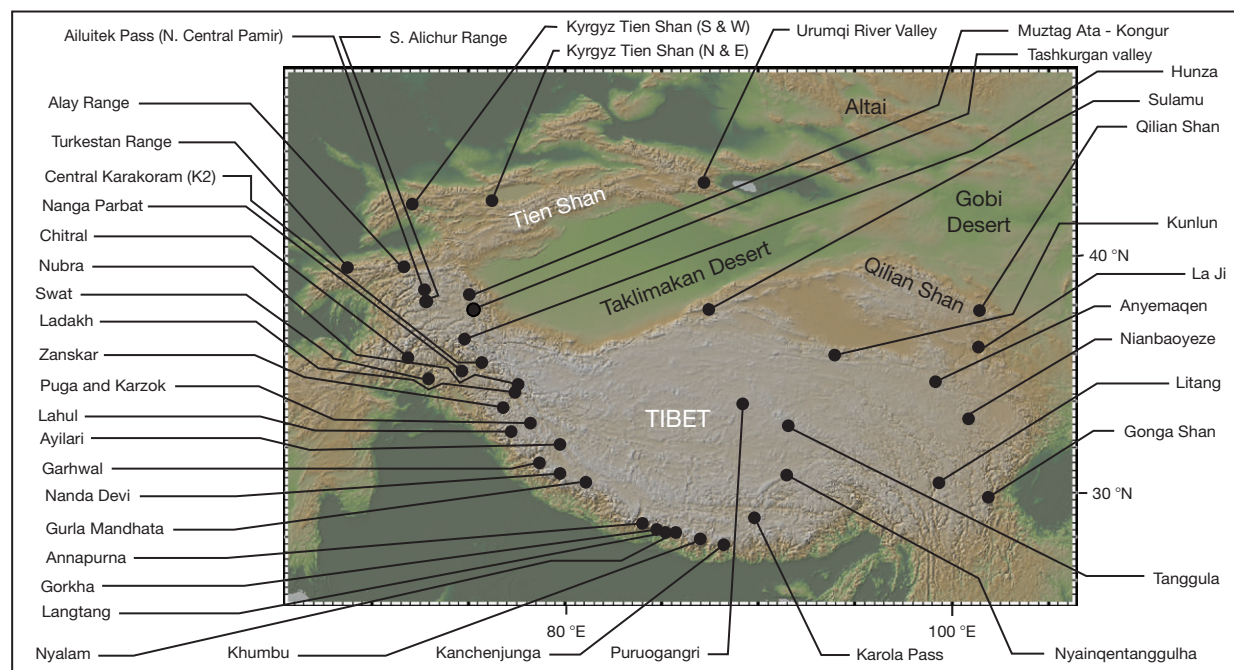


Figure 1 Digital elevation model of Highland Asia showing the main regions where modern detail glacial chronologies have been developed and that are discussed in this article.

monsoonal Himalaya, and suggested that the inconsistencies in their ages reflected times of enhanced weathering. Owen et al. (2008) emphasized that limits to the precision and accuracy of available dating methods and, more importantly, the geological uncertainty imposed by both processes of moraine formation and alteration conspire to limit the time resolution on which correlations can be made to Milankovitch timescales. To put all studies on a common scale, Owen et al. (2009) reevaluated well-dated sites and recalculated all the published ^{10}Be TCN ages for moraine boulders and glacially eroded surfaces throughout the Himalaya and Tibet. The study showed that locally there are considerable variations in the extent of glaciation from one region to the next during a glaciation. Moreover, glaciers throughout monsoon-influenced Tibet, the Himalaya, and the Trans Himalaya were likely synchronous both with climate change resulting from oscillations in the South Asian monsoon and Northern Hemisphere cooling cycles. In contrast, Owen et al. (2008) showed that glaciers in the Pamir in the far western regions of the Himalayan–Tibet orogen advanced asynchronously relative to the other regions that are monsoon-influenced, and glaciers in the Pamir appear to be mainly in phase with Northern Hemisphere cooling cycles.

Compilations of research on the Quaternary glaciation of Highland Asia are presented in several volumes of *Quaternary International* (Owen and Lehmkuhl, 2000; Owen and Zhou, 2002; Schwab, 2010; Yi and Owen, 2006; Yi et al., 2011), reviewed by Lehmkuhl and Owen (2005), Owen et al. (2008), and Owen (2009), and glacial mapping and chronologies are summarized for the Global Mapping Project of INQUA in Ehlers and Gibbard (2004, 2011). The best summary map of the extent of glaciation throughout Highland Asia during the last glacial was produced by Shi (1992; Figure 2). To summarize the current knowledge on the nature of

Quaternary glaciation in Highland Asia, this article considers each major region in turn.

Himalaya and Trans Himalaya

Owen et al. (1998, 2002a, 2008), Lehmkuhl and Owen (2005), and Owen (2009) provide reviews of the Quaternary glacial history of the Himalaya, and Ehlers and Gibbard (2004, 2011) provide further descriptions of the Quaternary glacial history of each Himalayan region. Simplified summaries to show the timing of glacial advances for each of the main regions in Highland Asia that have been studied using modern numerical dating methods are shown in Figure 3.

Glacial geologic studies throughout the Himalaya and the Trans Himalaya during the last decade have been quite detailed, and numerous glacial advances have been dated in selected regions. Of note is the work of Seong et al. (2007) in the Central Karakoram, where four glacial stages are defined using ^{10}Be TCN dating to >0.7 Ma, marine oxygen isotope stage (MIS) 6 or older, MIS 2, and the Holocene. Glaciers advanced several times during each glacial stage. Glacial advances are not well defined for the oldest glacial stages, but the younger glacial advances likely occurred at ~16, ~11–13, ~5, and ~0.8 ka. Scherler et al. (2010) provided ^{10}Be TCN ages to define five glacial episodes at ~16, ~11–12, ~8–9, ~5, and <1 ka in the Tons, Thangro, and Pin valleys in the Garhwal Himalaya. These glacial advances were attributed to glacial–interglacial oscillations related to waxing and waning of the Northern Hemisphere ice sheets, while the timing of millennial-scale oscillations was related more directly to monsoon strength. Similarly, around Annapurna, Zech et al. (2009) applied ^{10}Be TCN dating to define the landsliding and alluviation to help

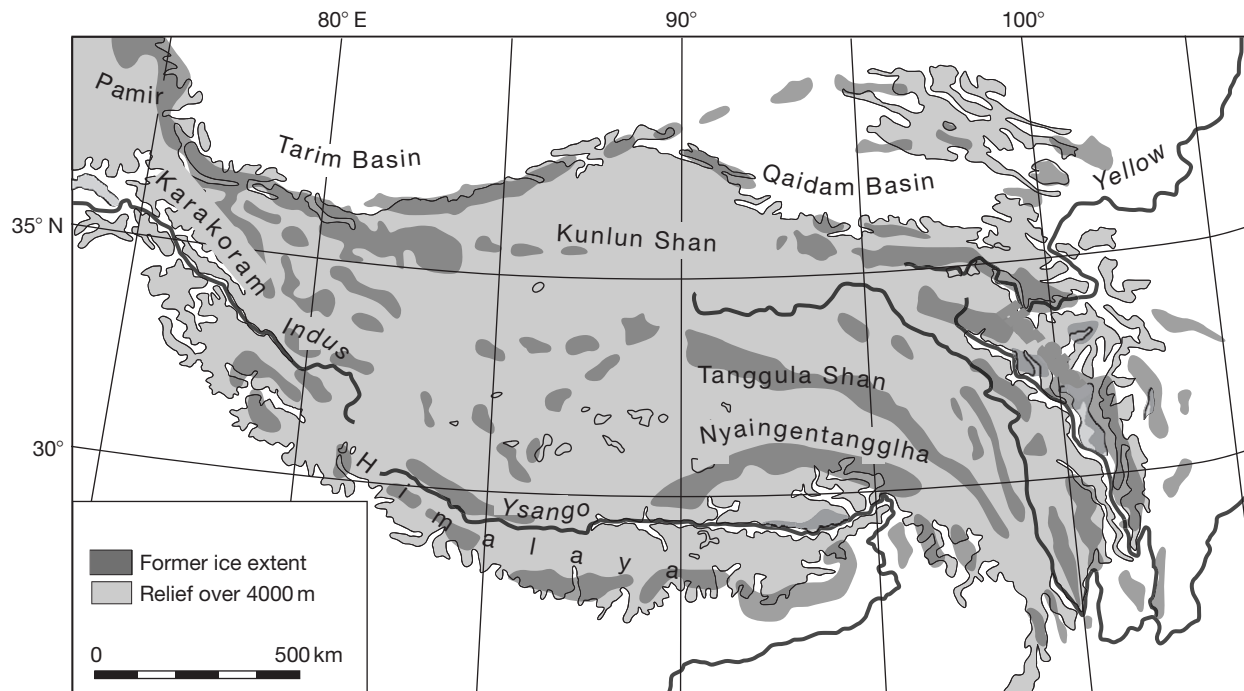


Figure 2 Reconstruction for the former extent of glaciation across Tibet and the bordering mountains for the maximum extent of glaciation during the last glacial (after Shi, 1992 from Owen, 2009). This reconstruction was based on detailed field mapping of glacial and associated landforms and sediments.

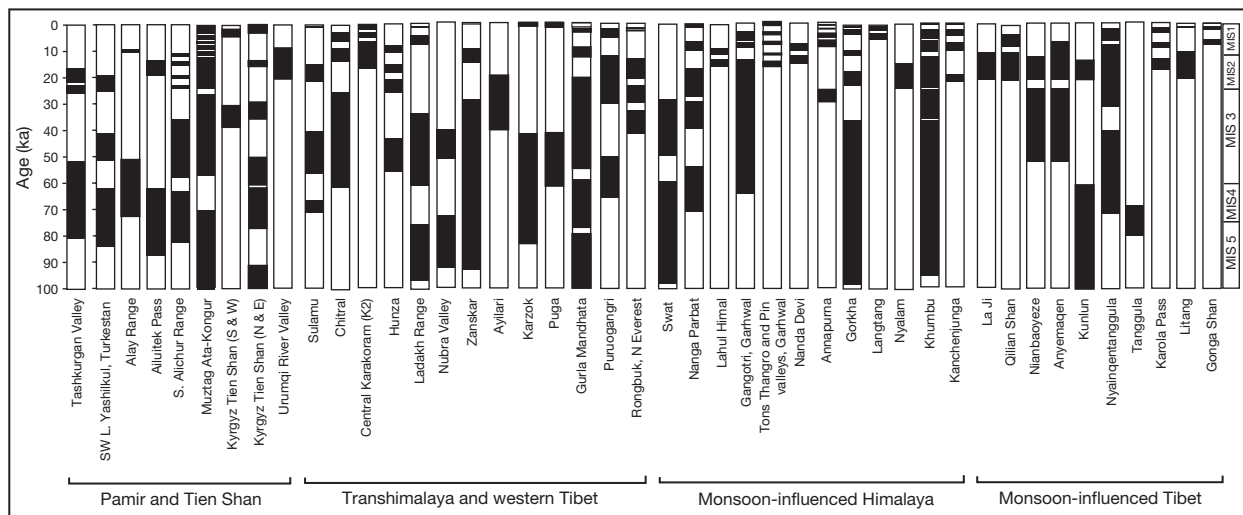


Figure 3 Simplified glacial chronologies for the Late Quaternary (back to 100 ka) that have been numerically dated throughout the Himalaya and Tibet. The black bars represent times of glacial advance (adapted from Owen LA, Caffee MW, Finkel RC, and Seong BY (2008) Quaternary glaciations in the Himalayan-Tibetan orogen. *Journal of Quaternary Science* 23: 513–532; Owen LA (2009) Latest Pleistocene and Holocene glacier fluctuations in the Himalaya and Tibet. *Quaternary Science Reviews* 28: 2150–2164; and data from Abramowski et al., 2006; Barnard et al., 2004a,b, 2006a,b; Chevalier et al., 2005; Colgan et al., 2006; Dortch et al., 2010; Finkel et al., 2003; Meriaux et al., 2004; Owen et al., 2001, 2002b,c, 2003a,b,c, 2005, 2006a,b; Phillips et al., 2000; Richards et al., 2000a,b; Schäfer et al., 2002, 2008; Scherler et al., 2010; Seong et al., 2007, 2008; Sharma and Owen, 1996; Shiraiwa, 1993; Spencer and Owen, 2004; Tsukamoto et al., 2002; Yi et al., 2002; Zech et al., 2003, 2005, 2009). See Figure 1 for the locations of each study area. The marine oxygen isotope stages (MISs) are shown to the far right.

constrain the timing of deglaciation. They suggested that some glaciers advanced in-phase with Northern Hemisphere ice sheets, whereas others reached their maximum extent at times of increased monsoonal precipitation during MIS 3 and the early Holocene.

In a study on Mount Everest, Owen et al. (2009) showed using OSL and ^{10}Be TCN dating that glaciations across the Everest massif were broadly synchronous. However, unlike the southern slopes of Mount Everest, no early Holocene glacier advance was recognized on its northern slopes. Owen et al.

(2009) suggested that the southern slopes may have received increased monsoon precipitation in the early Holocene to help increase positive glacier mass balances, while the northern slopes were too sheltered to receive monsoon moisture during this time and glaciers could not advance. Comparing equilibrium-line altitude (ELA) depressions for glacial stages across Mount Everest, Owen et al. (2009) revealed asymmetric patterns of glacier retreat that likely reflect greater glacier sensitivity to climate change on the northern slopes, possibly due to precipitation starvation (Figure 4). This illustrates some of the sharp contrasts in the degree of glaciation across relatively short geographical distances.

To generalize, glaciation in many regions throughout the northern areas of the Himalaya and the Trans Himalaya was very much restricted during the global Last Glacial Maximum (LGM), with glaciers generally advancing <10 km from contemporary ice margins. The local LGM, however, occurred during the early part of the last glacial cycle, and in many regions, this was during MIS 3 (e.g., Dortch et al., 2010; Finkel et al., 2003; Owen et al., 2009; Richards et al., 2000b; Seong et al., 2007). In contrast, in the southern, more monsoon-influenced areas of the Himalaya, glacier advances during the LGM were more extensive and often it is difficult to find evidence for pre-LGM glaciation (e.g., Barnard et al., 2006).

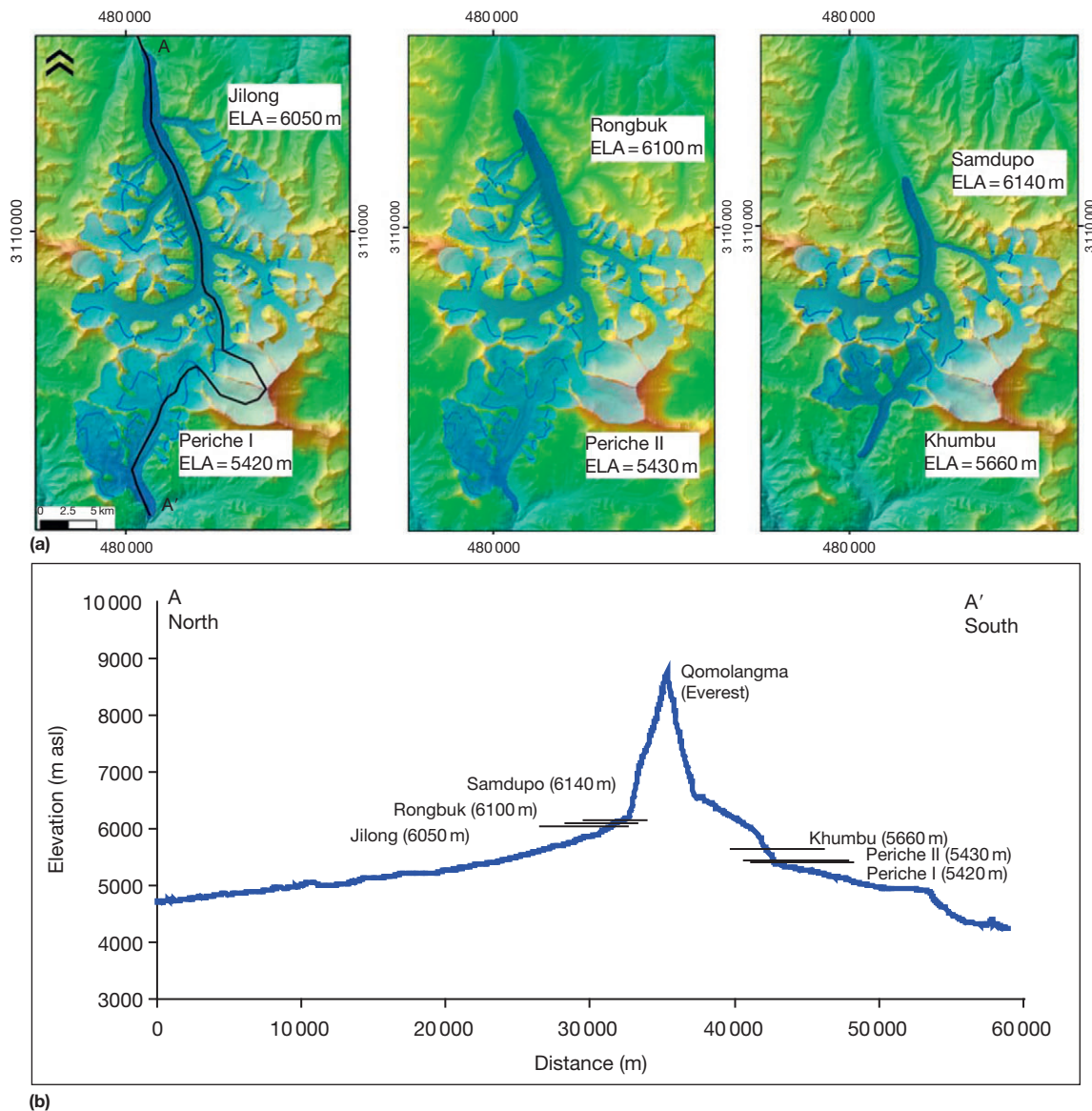


Figure 4 Example of modern glacial reconstructions and chronologies for a mountain mass in Highland Asia, provided by Mount Everest. Reproduced from Owen LA, Robinson R, Benn DI, et al. (2009) Quaternary glaciation of Mount Everest. *Quaternary Science Reviews* 28: 1412–1433. (a) Reconstructions of the glacial extent across Mount Everest plotted on a digital elevation model for (left to right): the Jilong ($\sim 24.3 \pm 3.8$ ka) and Periche I (23.9 ± 2.5 ka) glacial stages, Rongbuk (16.6 ± 4.1 ka) and Periche II (17.0 ± 1.7 ka) glacial stages, and the Samdupo I (7.1 ± 0.5 ka) and Chhukung (10.1 ± 0.4 ka) glacial stages. Ages shown are based on ^{10}Be TCN dates except for Samdupo I, which is based on OSL dating. (b) Profile across Everest showing ELA reconstructions for the Jilong, Periche I, Rongbuk, Periche II, Samdupo I, and Chhukung glacial stages (the profile line is shown in part A).

The complexity in the timing and style of glaciation across Highland Asia supports the view of [Benn and Owen \(1998\)](#) that glaciation throughout much of the region was asynchronous with the Northern Hemisphere ice sheets and oceans, with the maximum glacial advances occurring during MIS 3 or MIS 5a to 5d. The timing of these advances is attributed to increased cloud cover and precipitation as snow at high altitudes due to a strengthened monsoon that penetrated further north into the Himalaya and Trans Himalaya during times of increased insolation ([Finkel et al., 2003](#); [Owen et al., 2002b, 2005](#)). During times of lower insolation, particularly during the global LGM (MIS 2), the influence of the monsoon was reduced in the northern areas of the Himalaya and the Trans Himalaya, which in turn resulted in lower snowfall and snow accumulation, and hence less extensive glaciation. In contrast, in the southern regions of the Himalaya, significant moisture and cloud cover enabled glaciers to grow in times of lower insolation and reduced temperatures.

Many glaciers throughout the Himalayan and Trans Himalayan regions have been retreating throughout the last century ([Fujita et al., 1997](#); [Mayewski and Jeschke, 1979](#); [Mayewski et al., 1980](#)), but the extent of retreat has yet to be adequately quantified. However, in some regions such as Karakoram, glaciers have been remarkably stable and some glaciers have thickened and surged over the last decade or so ([Copland et al., 2011](#); [Mayer et al., 2006](#)). This may be coincident with a period of increased precipitation and positive glacier mass balance in the Karakoram. Possible widespread deglaciation, particularly the uncertainty in the degree of likely change, poses a serious threat to water resources on the Indian subcontinent and may result in hazards such as glacial lake outburst floods ([Richardson and Reynolds, 2000](#)).

Tibet

The existence of an extensive ice sheet was one of the most contentious glacial issues concerning the Quaternary glaciation of Tibet during the latter part of the twentieth century. The existence of an ice sheet was mainly stimulated by the work of Kuhle (e.g., [Kuhle, 1985](#)); however, numerous publications discuss the evidence against an extensive ice sheet, and it is now generally accepted that a large ice sheet did not exist on the Tibetan Plateau during at least the last few glacial cycles ([Lehmkuhl and Owen, 2005](#); [Owen et al., 1998, 2005](#); [Rutter, 1995](#); [Schäfer et al., 2002](#); [Seong et al., 2008](#)). [Liu et al. \(1988\)](#), [Li et al. \(1991\)](#), and [Shi \(1992\)](#) provide the most widely accepted regional reconstructions of the former extent of glaciers across the Tibetan Plateau ([Figure 2](#)).

The general style of glaciation in Tibet appears to have changed over the last several glacial cycles from expanded ice caps, to piedmont-style glaciers, and then to entrenched valley glaciers ([Lehmkuhl et al., 2002](#); [Owen et al., 2003, 2005, 2010](#); [Schäfer et al., 2002](#)). This is well illustrated by [Owen et al. \(2010\)](#) who examined the evidence for glaciation in and around Gurla Mandhata (Naimona'nyi), an isolated massif in southern Tibet, and developed one of the most well-defined numerically dated glacial successions for the region. [Owen et al. \(2009\)](#) were able to show that glaciation changed from an expanded ice cap that stretched ~5 km into the foreland of

the Gurla Mandhata massif, likely during MIS 10 or an earlier glacial cycle, to piedmont-type glaciation probably during the early part of the last glacial cycle or the penultimate glacial cycle, after which glaciers became restricted to entrenched valley types. Impressive laterofrontal moraines at the mouths of most valleys date to the early part of the last glacial cycle and MIS 3. A succession of six sets of moraines within the valleys and up to the contemporary glaciers shows that glaciers advanced during the Late Glacial, early Holocene, Neoglacial, and possibly, Little Ice Age. [Owen et al. \(2010\)](#) suggested that the change of style of glaciation throughout the latter part of the Quaternary, from expanded ice caps to deeply entrenched valley glaciers, might reflect climatic controls with reduced moisture supply to the region over time and/or tectonic controls reflecting increased basin subsidence due to detachment faulting and enhanced valley incision.

As in the Himalaya, the local LGM in Tibet appears to have been early in the last glacial cycle ([Owen et al., 2003, 2005, 2010](#); [Schäfer et al., 2002](#)). This might reflect increased moisture flux, and cloudiness during these times created higher precipitation in the form of snow at high altitude, which in turn led to positive glacial mass balances and glacial advances. However, although precipitation would have been reduced during the insolation minima of MIS 2 (global LGM), temperatures were low enough to lead to positive glacier mass balances, allowing glaciers to advance, albeit not as very far as to the local LGM limits.

The few published studies on Late Glacial and Holocene glacier fluctuations in Tibet suggest that glaciers advanced at 2–15 ka in the West Kunlun and in the mountainous areas surrounding the Qaidam Basin, and advanced during the early Holocene advances in the Anyemaqen, Gurla Mandhata, on the Karola Pass, and in the Gongga Shan during the early Holocene ([Lehmkuhl and Owen, 2005](#); [Owen et al., 2005, 2006b, 2010](#)). Little Ice Age and Neoglacial moraines are defined by dates on wood and branches incorporated in moraines for several regions ([Lehmkuhl and Owen, 2005](#) and references therein). [Su and Shi \(2002\)](#) discuss the nature of glacial fluctuations since the Little Ice Age (seventeenth to nineteenth century) and show that the mean temperature of monsoonal temperate glaciers in China has increased by 0.8 °C with a decrease in modern glacier area of ~30%, a loss of some 4000 km².

Tien Shan and Altai Mountains

Until recently, there was much debate concerning the extent of late Pleistocene glaciation in the Tien Shan and Altai Mountains. [Grosswald et al. \(1994\)](#), for example, argued that an extensive ice sheet existed in these regions during the last glacial cycle, with late Pleistocene glaciers extending to the foothills of the Tien Shan. However, other arguments and recent studies provide evidence that glaciers were restricted to the mountain range and did not reach the foothills ([Heuberger and Sgibnev, 1998](#); [Koppes et al., 2008](#); [Zech et al., 1996](#)).

The first comprehensive study of moraine successions in the Tien Shan was undertaken by [Koppes et al. \(2008\)](#) who used numerical ¹⁰Be TCN dating methods to define the ages of glacial advances. [Koppes et al. \(2008\)](#) showed that glaciers in

the north and east of the Kyrgyz Tien Shan last advanced to their maximum positions during MIS 5 and again during MIS 4, while in the south and west of the range, glacier advances occurred during MIS 3. In contrast to maritime Europe and North America, there was no evidence of a major glacial advance during MIS 2, and glacier advances during MIS 2 and since were restricted to the vicinity of modern glaciers. [Koppes et al. \(2008\)](#) also support the view that an extensive ice sheet did not exist in the region.

In the NW Tien Shan, [Kong et al. \(2009\)](#) presented ^{10}Be ages ranging from 9 to 21 ka on glacial boulders at the head of the Urumqi River and suggested that glacial advance and retreat commenced at the global LGM and most likely abated by the early Holocene.

Valley glaciers reached Lake Teleski at 430 m asl in the Russian Altai, with the extent of glaciation being greatest in the western Russian Altai ([Lehmkuhl and Owen, 2005](#), and references therein). In the eastern region, valley glaciers stretched down to the Kuray and Chuya Basins damming the main rivers to form ice-dammed lakes, which ultimately produced the largest glacial lake outburst floods in the world ([Baker et al., 1993](#); [Rudoy, 2002](#)). In the eastern part of the Russian Altai and the Mongolian Altai, [Lehmkuhl \(1998\)](#), [Klinge et al. \(2003\)](#), and [Lehmkuhl et al. \(2004\)](#) provide evidence for two major Pleistocene glaciations of a similar extent. The limited extent of the present and Pleistocene glaciers in the western part of the Russian Altai and in the Mongolian Altai is the result of reduced precipitation from west to east, which causes a rise of present and Pleistocene ELAs toward the east. Although there is a lack of numerical dating of glacial sediments and landforms in this region, [Lehmkuhl et al. \(2004\)](#) argued that most late Pleistocene glacier advances in Mongolia and in the Russian Altai took place during MIS 2 and 4.

Pamir

The nature of Late Quaternary glaciation in the Pamir of Tajikistan has been examined by [Zech et al. \(2005\)](#) who mapped moraines southwest of Lake Yashilkul and applied ^{10}Be TCN surface exposure methods. [Zech et al. \(2005\)](#) recognized three main glacial advances, which included an extensive glaciation at ~ 60 ka, a less extensive glacial advance at or before 47 ka, and an even more restricted glacial advance during the latter part of the last glacial. The pattern of progressively limited glacial extent during the late Pleistocene was attributed to increasing aridity in the Pamir ([Zech et al., 2005](#)).

In the easternmost Pamir, [Seong et al. \(2009\)](#) studied the glacial record of two massifs, Muztag Ata and Kongur Shan. Using remote sensing, field mapping, and ^{10}Be TCN methods, [Seong et al. \(2009\)](#) showed that glaciers advanced at least 12 times during at least the last two glacial cycles. Over this time, the style of glaciation changed progressively from one that produced ice caps to one that produced less extensive and more deeply entrenched valley glaciers. [Seong et al. \(2009\)](#) suggested that this might reflect increasing aridity over time, which might be influenced by tectonics. The timing of the two earliest glaciations, however, was poorly defined, but likely occurred prior to the penultimate glacial cycle, the early part

of the last glacial cycle, or during the penultimate glacial cycle. The younger glacial advances were relatively well defined showing quasiperiodical oscillations on millennial timescales at ~ 17.1 , ~ 13.7 , ~ 11.2 , ~ 10.2 , ~ 8.4 , ~ 6.7 , ~ 4.2 , ~ 3.3 , ~ 1.4 ka, and a few hundred years before the present. [Seong et al. \(2009\)](#) argued that since the global LGM, the glaciers in the easternmost Pamir and westernmost Tibet were likely responding to Northern Hemisphere climate oscillations (rapid climate changes), with minor influences from the south Asian monsoon.

Equilibrium-Line Altitudes

Reconstructions of past ELAs for glaciers throughout Highland Asia are summarized in [Owen and Benn \(2005\)](#). [Owen and Benn \(2005\)](#) and [Benn et al. \(2005\)](#) stressed the need to use multiple methods when reconstructing past and present ELAs for high-altitude glaciers, which are strongly influenced by topography and have extensive supraglacial debris cover. Even though much numerical dating has been undertaken in recent years, there are few studies that combine chronostratigraphies and ELA studies. This makes it particularly difficult to reconstruct patterns of ELAs across large areas of Highland Asia. Using the two best dated moraine successions in the Himalaya and Tibet, namely the Hunza Valley and the Khumbu Himal, [Owen and Benn \(2005\)](#) and [Owen et al. \(2009\)](#) calculated the ELA for MIS 2 (global LGM) to show that for these regions, the ELA depression during MIS 2 was ~ 200 – 300 m and ≤ 100 m, respectively. This contrasts sharply with earlier estimates for the ELA depression for these regions, which were estimated to be > 500 m, and it helps highlight how restricted glacial advances were during the global LGM for large areas of Highland Asia. [Owen et al. \(2009\)](#) also provide ELA reconstructions across Mount Everest for different numerically dated glacial advances during the Late Quaternary ([Figure 4](#)).

Broad regional ELA gradients can be reconstructed across Highland Asia even though there are significant problems correlating glacial successions and reconstructions of ELA depressions. [Lehmkuhl and Owen \(2005\)](#) provide illustrations that show present and Pleistocene glaciers and ELAs, which crudely quantify the nature of glaciation in some of the high mountains of Highland Asia ([Figure 5](#)). These show that ELAs dropped by 300–500 m in the drier parts and by 600 to more than 1000 m in the wetter parts of the Tibetan Plateau and the Himalaya sometime during the last glacial cycle, but in some areas, these ELA depressions might actually be for a previous glacial cycle as many are not defined by numerical dating.

Conclusion

An understanding of the nature, timing, and extent of late Quaternary glaciation in Highland Asia is rapidly advancing, but the vastness and inaccessibility of the region still presents significant challenges. With the development and application of new dating techniques, specifically TCN and OSL dating, the number of glacial successions that have been dated during the last decade is encouraging. The numerical dating studies suggest that on millennial timescales glaciation is broadly

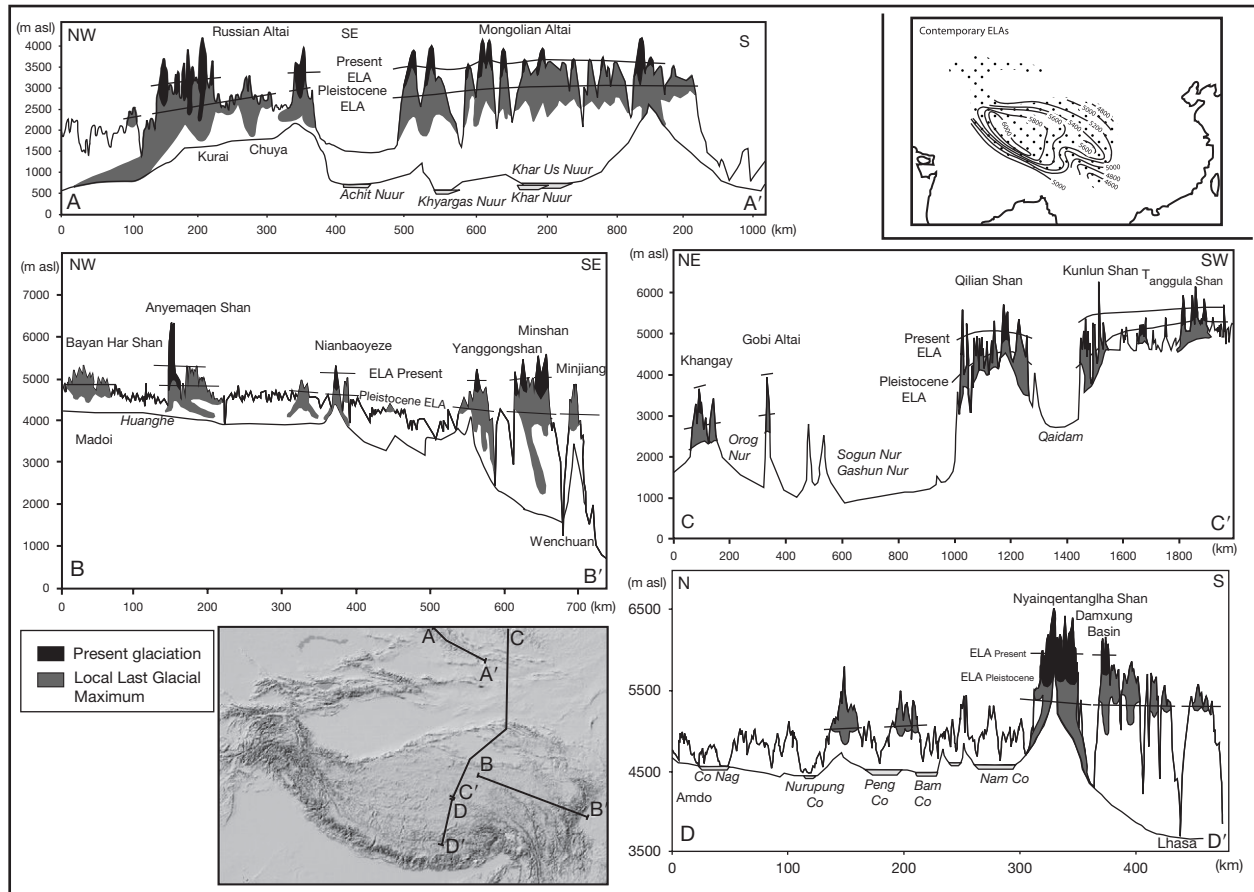


Figure 5 Selected cross-sections of Central and High Asia, including the distribution of present and local Last Glacial Maximum ice and the ELAs. Adapted from Lehmkuhl F and Owen LA (2005) Late Quaternary glaciation of Tibet and the bordering mountains: A review. *Boreas* 34: 87–100. Cross-section A–A' – the Russian Altai to the Mongolian Altai. Cross-section B–B' – eastern fringe of the Tibetan Plateau. Cross-section C–C' – Khangay and Gobi Altai to the Tibetan Plateau. Cross-section D–D' – southern Tibet. The inset map in the top right corner shows the variation in contemporary regional snowlines across Tibet and the bordering mountains.

synchronous throughout Highland Asia, but the exact timing and extent of glaciation varies considerably within and between different mountain ranges within Highland Asia. The complex patterns of glaciation that are emerging imply that the glaciation throughout High Asia reflects varying, temporally and spatially, forcing by the different climatic systems that dominate the region. As such, the style and timing of glaciation appears to be strongly controlled by climatic gradients, but topography may also have an important control. The extent of glaciation has become increasingly restricted throughout the Late Quaternary in the more arid regions of Highland Asia, which has helped preserve very old (pre-last glacial cycle) glacial landforms. In contrast, in regions that are very strongly influenced by the monsoon, the preservation potential of old moraine (pre-Late Glacial) is poor. This may be because Late Glacial and Holocene glacial advances were most extensive, destroying the geomorphic or sedimentary evidence of earlier glaciations, together with the intense fluvial and mass movement denudation that is characteristic of the wetter regions. Neoglacial, Little Ice Age, and historical moraines testify to a progressive retreat of glaciers during the twentieth century and warn of a continued retreat during the coming century.

See also: [Paleoclimate Relevance to Global Warming.](#)

Luminescence Dating: [Optical Dating.](#) **Paleoclimate Modeling:** [Paleo-ENSO.](#)

References

- Abramowski U, Bergau A, Seebach D, et al. (2006) Pleistocene glaciations of Central Asia: Results from ^{10}Be surface exposure ages of erratic boulders from the Pamir (Tajikistan) and the Alay-Turkestan range (Kyrgyzstan). *Quaternary Science Reviews* 25: 1080–1096.
- Baker VR, Benito G, and Rudoy AN (1993) Palaeohydrology of late Pleistocene superflooding, Altay Mountains, Siberia. *Science* 259: 348–350.
- Barnard PL, Owen LA, and Finkel RC (2004a) Style and timing of glacial and paraglacial sedimentation in a monsoonal influenced high Himalayan environment, the upper Bhagirathi Valley, Garhwal Himalaya. *Sedimentary Geology* 165: 199–221.
- Barnard PL, Owen LA, and Finkel RC (2006a) Quaternary fans and terraces in the Khumbu Himal south of Mount Everest: their characteristics, age and formation. *Journal of Geological Society of London* 163: 383–399.
- Barnard PL, Owen LA, Finkel RC, and Asahi K (2006a) Landscape response to deglaciation in a high relief, monsoon-influenced alpine environment, Langtang Himal, Nepal. *Quaternary Science Reviews* 25: 2162–2176.
- Barnard PL, Owen LA, Finkel RC, and Asahi K (2006b) Landscape response to deglaciation in a high relief, monsoon-influenced alpine environment, Langtang Himal, Nepal. *Quaternary Science Reviews* 25: 2162–2176.

- Barnard PL, Owen LA, Sharma MC, and Finkel RC (2004b) Late Quaternary landscape evolution of a monsoon-influenced high Himalayan valley, Gori Ganga, Nanda Devi, NE Garhwal. *Geomorphology* 61: 91–110.
- Benn DI and Owen LA (1998) The role of the Indian summer monsoon and the mid-latitude westerlies in Himalayan glaciation: Review and speculative discussion. *Journal of the Geological Society of London* 155: 353–364.
- Benn DI and Owen LA (2002) Himalayan glacial sedimentary environments: A framework for reconstructing and dating former glacial extents in high mountain regions. *Quaternary International* 97–98: 3–26.
- Benn DI, Owen LA, Osmaston HA, Seltzer GO, Porter SC, and Mark B (2005) Reconstruction of equilibrium-line altitudes for tropical and sub-tropical glaciers. *Quaternary International* 138(139): 8–21.
- Chevalier M-L, Ryerson FJ, Tapponnier P, et al. (2005) Slip-rate measurements on the Karakoram Fault may imply secular variations in fault motion. *Science* 307: 411–414.
- Colgan PM, Munroe JS, and Zhou S (2006) Cosmogenic radionuclide evidence for the limited extent of last glacial maximum glaciers in the Tanggula Shan of the central Tibetan Plateau. *Quaternary Research* 65: 336–339.
- Copland L, Sylvestre T, Bishop MP, et al. (2011) Expanded and recently increased glacier surging the Karakoram. *Artic, Alpine and Antarctic Research* 43: 503–516.
- Dortch JM, Owen LA, and Caffee MW (2010) Quaternary glaciation in the Nubra and Shyok valley confluence, northernmost Ladakh, India. *Quaternary Research* 74: 132–144.
- Ehlers J and Gibbard P (eds.) (2004) Quaternary glaciations – extent and chronology, Part III: South America, Asia, Africa, Australia, Antarctica. In: *Developments in Quaternary Science*, vol. 2, p. 380. Amsterdam: Elsevier.
- Ehlers J, Gibbard PL, and Hughes PD (eds.) (2011) Quaternary glaciations, extent and chronology – A closer look. In: *Developments in Quaternary Science*, vol. 16. Amsterdam: Elsevier.
- Finkel RC, Owen LA, Barnard PL, and Caffee MW (2003) Beryllium-10 dating of Mount Everest moraines indicates a strong monsoonal influence and glacial synchronicity throughout the Himalaya. *Geology* 31: 561–564.
- Fuchs M and Owen LA (2008) Luminescence dating of glacial and associated sediments: Review, recommendations and future directions. *Boreas* 37: 636–659.
- Fujita K, Nakawo M, Fujii Y, and Paudyal P (1997) Changes in glaciers in Hidden Valley, Mukut Himal, Nepal Himalayas, from 1974 to 1994. *Journal of Glaciology* 43: 583–588.
- Grosswald MG, Kuhle M, and Fastook JL (1994) Würm glaciation of Lake Issyk-Kul Area, Tian Shan Mts.: A case study in glacial history of central Asia. *Geojournal* 33: 273–310.
- Hedrick KA, Seong YB, Owen LA, Caffee MC, and Dietsch C (2011) Towards defining the transition in style and timing of Quaternary glaciation between the monsoon-influenced Greater Himalaya and the semi-arid Transhimalaya of Northern India. *Quaternary International* 236: 21–33.
- Heuberger H and Sgibnev VV (1998) Paleoglaciological studies in the Ala-Archa National Park, Kyrgyzstan, NW Tian-Shan mountains, and using multitextural analysis as a sedimentological tool for solving stratigraphic problems. *Zeitschrift für Gletscherkunde und Glazialgeologie* 34: 95–123.
- Hewitt K (1999) Quaternary moraines vs catastrophic rock avalanches in the Karakoram Himalaya, Northern Pakistan. *Quaternary Research* 51: 220–237.
- Klinge M, Böhner J, and Lehmkühl F (2003) Climate patterns, snow- and timberline in the Altai Mountains, Central Asia. *Erdkunde* 57: 296–308.
- Kong P, Fink D, Na C, and Huang F (2009) Late Quaternary glaciation of the Tianshan, Central Asia, using cosmogenic ¹⁰Be surface exposure dating. *Quaternary Research* 72: 229–233.
- Koppes M, Gillespie AR, Burke RM, Thompson SC, and Stone J (2008) Late Quaternary glaciation in the Kyrgyz Tien Shan. *Quaternary Science Reviews* 27: 846–866.
- Kuhle M (1985) Glaciation Research in the Himalayas: A new ice age theory. *Universitas* 27: 281–294.
- Lehmkuhl F (1998) Quaternary glaciations in central and western Mongolia. *Quaternary Proceedings* 6: 153–167.
- Lehmkuhl F, Klinge M, and Lang A (2002) Late Quaternary glacier advances, lake level fluctuations and aeolian sedimentation in Southern Tibet. *Zeitschrift für Geomorphologie N.F.* 126: 183–218.
- Lehmkuhl F, Klinge M, and Stauch G (2004) The extent of Late Pleistocene glaciations in the Altai and Khangai Mountains. In: Ehlers J and Gibbard PL (eds.) *Quaternary Glaciations – Extent and Chronology, Part III: South America, Asia, Africa, Australia, Antarctica*, pp. 243–254. Oxford: Elsevier.
- Lehmkuhl F and Owen LA (2005) Late Quaternary glaciation of Tibet and the bordering mountains: A review. *Boreas* 34: 87–100.
- Li B, Li J, and Cui Z (eds.) (1991) Quaternary glacial distribution map of Qinghai-Xizang (Tibet) Plateau 1:3,000,000. Shi Y (Scientific Advisor). *Quaternary Glacier, and Environment Research Center, Lanzhou University, China*.
- Liu Z, Jiao S, Zhang Y, et al. (1988) Geological Map of the Qinghai-Xizang (Tibet) Plateau and adjacent areas (1:1,500,000). Explanatory note. Chengdu Institute of Geology Resources, Chinese Academy of Geological Sciences, Chengdu, China.
- Mayer C, Lambrecht A, Belò M, Smiraglia C, and Diolaiuti G (2006) Glaciological characteristics of the ablation zone of Baltoro glacier, Karakoram, Pakistan. *Annals of Glaciology* 43: 123–131.
- Mayewski PA and Jeschke PA (1979) Himalayan and Trans-Himalayan glacier fluctuations since AD 1812. *Arctic and Alpine Research* 11: 267–287.
- Mayewski PA, Pregent GP, Jeschke PA, and Ahmad N (1980) Himalayan and Trans-Himalayan glacier fluctuations and the south Asian monsoon record. *Arctic and Alpine Research* 12: 171–182.
- Meriaux A-S, Ryerson FJ, Tapponnier P, et al. (2004) Rapid slip along the central Altyn Tagh Fault: Morphochronologic evidence from Cherchen He and Sulamu Tagh. *Journal of Geophysical Research* 109(B6): B06401.
- Owen LA (2009) Latest Pleistocene and Holocene glacier fluctuations in the Himalaya and Tibet. *Quaternary Science Reviews* 28: 2150–2164.
- Owen LA and Benn DI (2005) Equilibrium-line altitudes of the Last Glacial Maximum for the Himalaya and Tibet: An assessment and evaluation of results. *Quaternary International* 138(139): 55–78.
- Owen LA, Caffee MW, Bovard KR, Finkel RC, and Sharma MC (2006a) Terrestrial cosmogenic nuclide surface exposure dating of the oldest glacial successions in the Himalayan orogen: Ladakh Range, northern India. *Geological Society of America Bulletin* 118: 383–392.
- Owen LA, Caffee MW, Finkel RC, and Seong BY (2008) Quaternary glaciations of the Himalayan-Tibetan orogen. *Journal of Quaternary Science* 23: 513–532.
- Owen LA, Derbyshire E, and Fort M (1998) The Quaternary glacial history of the Himalaya. In: Owen LA (ed.) *Mountain Glaciation Quaternary Proceedings* 6: 91–120.
- Owen LA, Finkel RC, Barnard PL, et al. (2005) Climatic and topographic controls on the style and timing of Late Quaternary glaciation throughout Tibet and the Himalaya defined by ¹⁰Be cosmogenic radionuclide surface exposure dating. *Quaternary Science Reviews* 24: 1391–1411.
- Owen LA, Finkel RC, and Caffee MW (2002a) A note on the extent of glaciation throughout the Himalaya during the global Last Glacial Maximum. *Quaternary Science Reviews* 21: 147–157.
- Owen LA, Finkel RC, Caffee MW, and Gualtieri L (2002b) Timing of multiple glaciations during the Late Quaternary in the Hunza Valley, Karakoram Mountains, Northern Pakistan: Constrained by cosmogenic radionuclide dating of moraines. *Geological Society of America Bulletin* 114: 593–604.
- Owen LA, Finkel RC, Ma H, and Barnard PL (2006b) Late Quaternary landscape evolution in the Kunlun Mountains and Qaidam Basin, Northern Tibet: A framework for examining the links between glaciation, lake level changes and alluvial fan formation. *Quaternary International* 154–155: 73–86.
- Owen LA, Finkel RC, Ma H, et al. (2003a) Timing and style of Late Quaternary glaciations in NE Tibet. *Geological Society of America Bulletin* 11: 1356–1364.
- Owen LA, Gualtieri L, Finkel RC, Caffee MW, Benn DI, and Sharma MC (2001) Cosmogenic radionuclide dating of glacial landforms in the Lahul Himalaya, Northern India: Defining the timing of Late Quaternary glaciation. *Journal of Quaternary Science* 16: 555–565.
- Owen LA, Kamp U, Spencer JQ, and Haserot K (2002c) Timing and style of Late Quaternary glaciation in the eastern Hindu Kush, Chitral, northern Pakistan: A review and revision of the glacial chronology based on new optically stimulated luminescence dating. *Quaternary International* 97(98): 41–55.
- Owen LA and Lehmkühl F (eds.) (2000) Late quaternary glaciation and paleoclimate of the Tibetan plateau and bordering mountains. In: *Quaternary International*, vol. 65/66, p. 212.
- Owen LA, Ma H, Derbyshire E, et al. (2003b) The timing and style of Late Quaternary glaciation in the La Ji Mountains, NE Tibet: Evidence for restricted glaciation during the latter part of the Last Glacial. *Zeitschrift für Geomorphologie* 130: 263–276.
- Owen LA, Robinson R, Benn DI, et al. (2009) Quaternary glaciation of Mount Everest. *Quaternary Science Reviews* 28: 1412–1433.
- Owen LA, Spencer JQ, Ma H, et al. (2003c) Timing of Late Quaternary glaciation along the southwestern slopes of the Qilian Shan, Tibet. *Boreas* 32: 281–291.
- Owen LA, Yi C, Finkel RC, and Davis N (2010) Quaternary glaciation of Gurla Mandata (Naimon'anyi). *Quaternary Science Reviews* 29: 1817–1830.
- Owen LA and Zhou S (eds.) (2002) Glaciation in Monsoon Asia. In: *Quaternary International*, vol. 97–98, p. 179.
- Phillips WM, Sloan VF, Shroder JF Jr., Sharma P, Clarke ML, and Rendell HM (2000) Asynchronous glaciation at Nanga Parbat, northwestern Himalaya Mountains, Pakistan. *Geology* 28: 431–434.

- Richards BWM, Benn D, Owen LA, Rhodes EJ, and Spencer JQ (2000) Timing of Late Quaternary glaciations south of Mount Everest in the Khumbu Himal, Nepal. *Geological Society of America Bulletin* 112: 1621–1632.
- Richards BWM, Owen LA, and Rhodes EJ (2000) Timing of Late Quaternary glaciations in the Himalayas of northern Pakistan. *Journal of Quaternary Science* 15: 283–297.
- Richardson SD and Reynolds JM (2000) An overview of glacial hazards in the Himalayas. *Quaternary International* 65(66): 31–48.
- Rudoy AN (2002) Glacier-dammed lakes and geological work of glacial superfoods in the Late Pleistocene, Southern Siberia, Altai Mountains. *Quaternary International* 87: 119–140.
- Rutter NW (1995) Problematic ice sheets. *Quaternary International* 28: 19–37.
- Schäfer JM, Oberholzer P, Zhao Z, et al. (2008) Cosmogenic beryllium-10 and neon-21 dating of late Pleistocene glaciations in Nyalam, monsoonal Himalayas. *Quaternary Science Reviews* 27: 295–311.
- Schäfer JM, Tschudi S, Zhizhong Z, et al. (2002) The limited influence of glaciations in Tibet on global climate over the past 170,000 years. *Earth and Planetary Science Letters* 194: 287–297.
- Scherler D, Bookhagen B, Strecker MR, von Blanckenburg F, and Rood D (2010) Timing and extent of late Quaternary glaciation in the western Himalaya constrained by ^{10}Be moraine dating in Garhwal, India. *Quaternary Science Reviews* 29: 815–831.
- Schwab A (ed.) (2010) Climate evolution and environmental response on the Tibetan plateau. In: *Quaternary International* 218: 1–202.
- Seong YB, Owen LA, Bishop MP, et al. (2007) Quaternary glacial history of the Central Karakoram. *Quaternary Science Reviews* 26: 3384–3405.
- Seong YB, Owen LA, Bishop MP, et al. (2008) Reply to comments by Matthias Kuhle on 'Quaternary glacial history of the central Karakoram'. *Quaternary Science Reviews* 27: 1656–1658.
- Seong YB, Owen LA, Yi C, and Finkel RC (2009) Quaternary glaciation of Muztag Ata and Kongur Shan: Evidence for glacier response to rapid climate changes throughout the Late Glacial and Holocene in westernmost Tibet. *Geological Society of America Bulletin* 121: 348–365.
- Sharma MC and Owen LA (1996) Quaternary glacial history of the Garhwal Himalaya, India. *Quaternary Science Reviews* 15: 335–365.
- Shi Y (1992) Glaciers and glacial geomorphology in China. *Zeitschrift für Geomorphologie, Supplement-Band* 86: 19–35.
- Shiraiwa T (1993) *Glacier fluctuations and cryogenic environments in the Langtang valley, Nepal Himalaya*. Contributions from the Institute of Low Temperature Science. Sapporo: Institute of Low Temperature Sciences, Hokkaido University.
- Spencer JQ and Owen LA (2004) Optically stimulated luminescence dating of Late Quaternary glaciogenic sediments in the upper Hunza valley: Validating the timing of glaciation and assessing dating methods. *Quaternary Sciences Reviews* 23: 175–191.
- Su Z and Shi Y (2002) Response of monsoon temperate glaciers to global warming since the Little Ice Age. *Quaternary International* 97–98: 123–131.
- Tsukamoto S, Asdahi K, Watanabe T, Kondo R, and Rink WJ (2002) Timing of past glaciation in Kanchenjunga Himal, Nepal by optically stimulated luminescence dating of tills. *Quaternary International* 97(98): 57–67.
- Yi C, Li X, and Qu J (2002) Quaternary glaciation of Puruogangri: The largest modern ice field in Tibet. *Quaternary International* 97(98): 111–123.
- Yi C and Owen LA (eds.) (2006) Quaternary paleoenvironmental change in Tibet and the bordering mountains. In: *Quaternary International*, vol. 154–155, p. 157.
- Yi C, Xhou L, Dortch J, and Owen LA (eds.) (2011) Quaternary landscape evolution and paleoenvironmental change: Himalaya-Karakorum-Tibet. In: *Quaternary International* 236: 1–166.
- Zech R, Abramowski U, Glaser B, Sosin P, Kubik PW, and Zech W (2005) Late Quaternary glacial and climate history of the Pamir Mountains derived from cosmogenic ^{10}Be exposure ages. *Quaternary Research* 64: 212–220.
- Zech W, Bäumler R, Sovoskul O, and Sauer G (1996) Zur Problematik der pleistozänen und holozänen Vergletscherung des Westlichen Tianshan (Usbekistan). *Eiszeitalter und Gegenwart* 46: 144–151.
- Zech W, Glaser B, Abramowski U, Dittmar C, and Kubik PW (2003) Reconstruction of the Late Quaternary Glaciation of the Macha Khola valley (Gorkha Himal, Nepal) using relative and absolute (^{14}C , ^{10}Be , dendrochronology) dating techniques. *Quaternary Science Reviews* 22: 2253–2265.
- Zech R, Zech M, Kubik PW, Kharki K, and Zech W (2009) Deglaciation and landscape history around Annapurna, Nepal, based on ^{10}Be surface exposure dating. *Quaternary Science Reviews* 28: 1106–1118.

Late Quaternary in North America

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The structure of late Pleistocene glaciations is now well-known at the global scale because of the intimate association between global sea levels and ice volumes. Measurements of the stable isotopic composition of benthic Foraminifera and surveys of raised marine terraces from tectonically uplifting areas have provided this template (Chappell et al., 1996). However, the extent to which individual ice sheets or continental ice complexes mirror the global pattern is unclear, especially for those events that exceed the reliable limits of radiocarbon dating. Here, we take that limit to be in the range of 40–50 cal ka BP, but dating the interval between ~20 and 50 cal ka BP is often problematic because the samples are more than four half-lives old and contamination by small amounts of younger carbon results in apparent ages that are much too young. The global pattern of changes in ice volume has been correlated with variations in the solar insolation in the Northern Hemisphere, often represented by changes at 65°N, and attributed to changes in the so-called Milankovitch orbital forcings (Budd and Smith, 1981; Imbrie et al., 1993) (Figure 1).

At the global scale, using the Marine Isotope Stage (MIS) designations, the last interglacial (MIS substage 5e) was followed by a series of glacial (MIS substages 5b and 5d) and interstadial events (MIS substages 5a and 5c), which were followed in turn by a long period of ice sheet growth during MIS 4. MIS 3 is an interval of 'interstadial' conditions positioned between the major glacial events of MIS 4 and MIS 2 (Figure 1). At the termination of the latest glacial cycle, Northern Hemisphere insolation rose rapidly, and to this rise is attributed the disappearance of many of the world's large ice sheets and glaciers—all except the Greenland and Antarctica ice sheets.

The view of late Pleistocene glaciations embodied in Figure 1, or in deep-sea benthic isotope curves, is dominated by the relatively slow and predictable Milankovitch forcings at ~20, 40, and 100 ka. Ice sheet behavior is thus seen as predictable and driven largely by changes in insolation and associated changes in temperature and precipitation. However, the earth's climate is and has been spatially variable. Changes in ice volumes are linked to the balance between accumulation and ablation, and there is no reason *a priori* to expect synchronicity of changes in mass balance for all parts of large ice sheets that span 45° of latitude. Furthermore, as ice sheets and outlet glaciers encroach onto continental shelves, and especially as they move seaward toward the shelf break, their mass balance changes will be due to nonclimatic forcings, such as relative sea level, which will be controlled by regional rates of glacial isostatic depression or rebound. For example, calving rates are strongly dependant on water depth at the ice margin and the slope of the sea floor (Hughes, 1987; Clarke et al., 1999). The same would also apply as ice sheets advanced into or retreated from large glacial lakes. These calving events will

clearly influence the mass balance of entire ice drainage basins and thus significantly control the position of ice margins.

Although this article pertains to 'North America,' the focus is on the large ice sheets that covered much of the land north of 40°N during MIS 2, which includes the Cordilleran, Laurentide, and Innuitian ice sheets (Figure 2). These contained ~99% of the North American ice volume during the last glaciation, the remainder being in the Alpine glaciers of Alaska and the American Cordillera. The terms used (Figure 1) follow those in the last compilation of the glacial records of Canada (Fulton, 1989).

Early Glacial Initiation—MIS Substages 5b and 5d

A rapid increase in global ice volume occurred during MIS substage 5d. The initial growth of ice over North America probably first occurred on the broad uplands and mountains of the eastern Canadian Arctic, and possibly the central Canadian Arctic, at elevations above 400 m (Andrews and Mahaffy, 1976; Marshall, 2002). Computer simulations of three-dimensional glaciological models indicate that with greatly enhanced accumulation the rate of global sea level lowering during MIS substage 5d can be matched by initial growth of the ice sheet (Andrews and Mahaffy, 1976). All such models indicate that ice expands across Hudson Strait, coalescing from Baffin Island and Labrador and damming a lake in Hudson Bay (Adam, 1976). The lacustrine sediment below the Adams till in the Hudson Bay lowlands may have been deposited in such a lake (Fulton, 1989, p. 231).

However, it is unclear what changes in climatic conditions would drive such a dramatic increase in snow accumulation. One possibility is that there was early ice growth across the Canadian high Arctic islands and channels, thus terminating the outflow of cold, fresh Arctic waters from the Arctic Ocean into Baffin Bay. If this happened, then the modified Atlantic waters of the West Greenland Current could expand and fill the surface of Baffin Bay, resulting in more open water and increased snow accumulation over the long polar winters. However, firm ages for MIS substages 5a–5d events are lacking, and the chronology is based on stratigraphy, climate proxies, and amino acid racemization, which is dependent on both time and site temperature. Nevertheless, Baffin Island outlet glaciers seem to have extended to the present coastline soon after the Last Interglacial MIS substage 5e.

Insights into the onset of the last glaciation across the eastern Canadian Arctic come from the examination of marine sediments from the Labrador Sea, especially cores from the SW Greenland rise and Orphan Knoll (east of Newfoundland) (Hillaire-Marcel et al., 1994) (Figure 2). The chronology is based principally on the oxygen isotope stratigraphy of

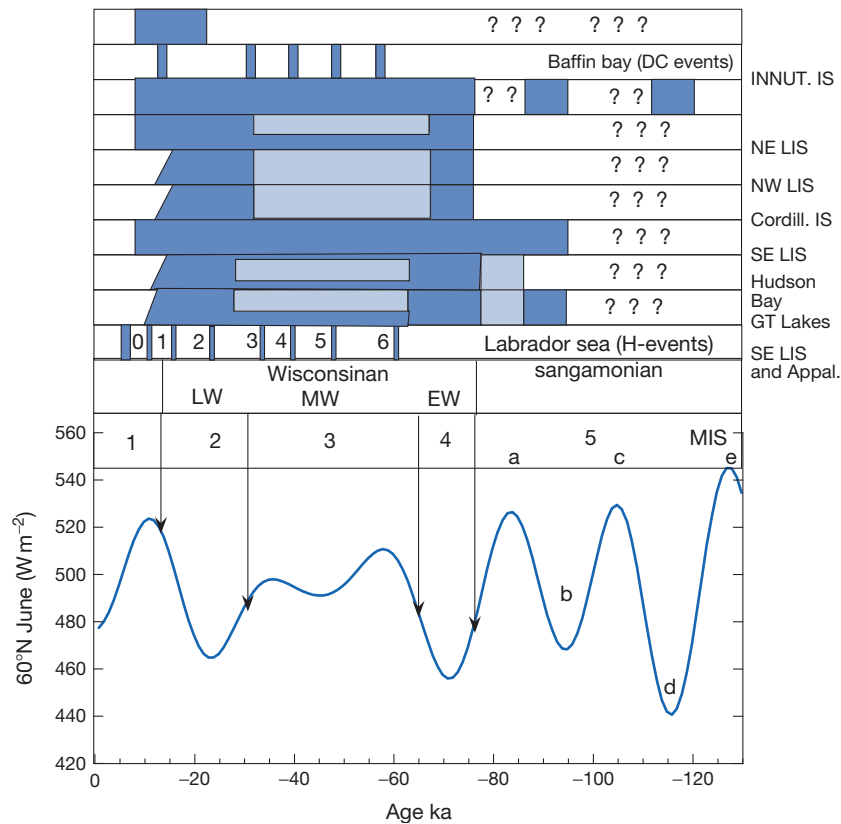


Figure 1 Main Marine Isotope Stages (MIS) during the last glacial cycle and the variations in ice sheet history at various points around the main North American ice sheets and the timing of the major detrital carbonate (Heinrich events) in the Labrador Sea and Baffin Bay (see [Figure 2](#)).



Figure 2 Map of the outlines of the main North American Late Pleistocene ice sheets and mountain glacier complexes (the latter mainly in the western United States). IIS, Innuitian Ice Sheet. The locations of marine cores noted in the text are as follows: 1, core from the SW Greenland Rise; and 2, core on Orphan Knoll, off Newfoundland.

planktonic Foraminifera, radiocarbon dates for events ≤ 40 ka BP, and correlation with Heinrich events (Andrews, 1998; Bond et al., 1992). Substage 5e is noteworthy for the low $\delta^{18}\text{O}$ values and an absence of coarse ice-rafted sediments (ice rafted debris (IRD), $>125\ \mu\text{m}$). On the Greenland Rise, IRD increases during MIS substages 5b–5d to consistent but low values. At Orphan Knoll, the record does not extend back beyond MIS substage 5a but coarse sand (IRD) varies between 1 and 20% by weight, suggesting that icebergs were being released from Greenland and/or the Canadian Arctic.

Early Wisconsinan—MIS Stage 4

MIS 4 lies beyond the limits of radiocarbon dating but there seems to be general consensus that the decline in high-latitude Northern Hemisphere insolation at ca. 70–80 cal ka BP resulted in a major glacial interval across the high-latitude and high-altitude areas of North America (Clark and Lea, 1992). However, the chronological placement of early Wisconsinan glacial sediments is largely *ad hoc*, being based on their positions above a last interglacial deposit and below glacial or interstadial deposits of MIS 2 or 3. The last major compilation of Canadian Quaternary data (Fulton, 1989) indicated no clear certainty as to which glacial deposits fall within MIS 5 and 4, and our compilation of coeval events (Figure 1) is largely based on the premise of ‘least surprise.’ From British Columbia evidence of extensive MIS 4 glaciation has been reported (Clague and Ward, 2011). It appears reasonable, but not proven, that MIS 4 is represented by glacial sediments across most formerly glaciated regions of North America.

The general consensus in the literature is that there was a widespread and extensive MIS 4 glaciation across North America. If the initial glacial coverage during MIS 5 was limited to the uplands of the Canadian Arctic, then the move toward a full-fledged North American ice sheet must have been extremely dramatic because the interval between MIS substage 5a and the insolation trough of MIS 4 was only 12 000 years (Figure 1). This may indicate that the growth of the ice sheet was associated with more open-water conditions in the Labrador Sea and Baffin Bay, resulting in enhanced mass accumulations in order to account for the rapid decrease in global sea level. On Baffin Island, the faunas and sedimentology of large ice-proximal glacial marine deltas suggested that the Kogalu stade of assumed MIS 4 age was associated with relatively mild conditions, hence standing in contrast with the cold-based late Wisconsinan (MIS 2) glacial cover (Briner et al., 2003).

In the Labrador Sea, MIS 4 includes Heinrich event 6 (H-6) (Hillaire-Marcel et al., 1994) (Figure 1), the first of a series of massive carbonate and IRD events associated with the large ice stream extending along Hudson Strait, and possibly other ice streams extending into the Labrador Sea and Baffin Bay (Marshall and Clarke, 1996; Andrews et al., 1998). The assumption is that the ice sheet expanded from its areas of nucleation on the uplands of the northeastern Canadian Arctic and sub-Arctic and extended southwards; in doing so, a large fraction of the ice sheet extended across areas below sea level (as in the case today of the West Antarctic Ice Sheet). Many of these areas are floored by more easily eroded

Paleozoic carbonates, which may have had a major influence on the ice sheet thickness and dynamics (Fisher et al., 1985; Clark, 1994).

Middle Wisconsinan—MIS 3

The middle Wisconsinan extends from about the limits of radiocarbon dating to ca. 27 cal ka BP. It is a period of moderate insolation spanning between the insolation troughs of MIS 4 and 2 (Figure 1). Global sea level is poorly defined for this interval because of inherent problems in obtaining reliable numerical ages; thus, this interval has been difficult to understand because of the ambiguities of radiocarbon dates >30 cal ka BP. However, several of the massive sedimentary strata associated with the collapse of the Laurentide Ice Sheet (LIS), the Heinrich (H)-events, are dated within this interval, notably H-5, H-4, and H-3 (Andrews, 1998). These are prominent strata in sediments in the Labrador Sea (Hesse et al., 1997) and there may be coeval sediments within Baffin Bay, associated with the history of the large ice streams that fed into its northern end (Andrews, 1998; Dyke et al., 2002).

At the MIS 4/3 transition, rising insolation appears to have led to retreat of the LIS from the Canadian prairie provinces and from the Great Lakes region to deglaciation of the entire Cordilleran region from British Columbia to the Yukon and Alaska, and perhaps unexpectedly to substantial recession of glaciers in the High Arctic Islands. Interstadial deposits and redeposited fossils of MIS 3 age, including extinct megafaunal taxa, are widespread in these regions. Although a precise boundary of the remnant ice sheet is not known, ice appears to have been confined to the ‘hard bed’ of the Precambrian shield rocks (Andrews, 1987; Dyke et al., 2002). Massive iceberg discharges, possibly accompanied by subglacial outburst floods occurred in the vicinity of Hudson Strait, resulting in carbonate-rich sediments being dispersed from the Labrador Sea eastwards to the Portuguese and British continental margins. Heinrich events 6, 5, 4, and 3 all occur within the middle Wisconsinan MIS 3 and require a large ice sheet centered on Hudson Bay and draining through Hudson Strait to provide these ice-rafted and meltwater plume sediments.

During MIS 3, Baffin Bay also received discrete layers of detrital carbonate, principally dolomite from Paleozoic outcrops on the floors of channels that carried the major ice streams entering the north end of the bay from the Innuitian (Figure 1, INNUT. IS; Figure 2, IIS) and Greenland ice sheets (Andrews et al., 1998). Cosmogenic dates on moraines from east-central Baffin Island indicate that outlet glaciers terminated within the fiords and did not extend onto the continental shelf.

Late Wisconsinan—MIS 2/1

Globally, the maximum volume of ice sheets and glaciers since MIS substage 5e occurred close to the trough of Northern Hemisphere summer insolation during MIS 2 (Figure 1). Within this interval, radiocarbon dating is generally reliable. Although few details of the MIS 3/2 transition are known, it seems that ice sheets and glaciers expanded or regrew rapidly but not necessarily simultaneously. It was during this

transition that the Cordilleran and Innuitian ice sheets became reestablished, and the LIS expanded to limits that generally exceeded previous late Pleistocene, and in places all-time Pleistocene, limits (e.g. Jackson et al., 2011). These limits were reached in most places between 25 and 22 ^{14}C ka BP (27–25 cal ka BP) and maintained until approximately 18 ^{14}C ka BP (21.4 cal ka BP). Interesting exceptions are the southwestern Cordilleran Ice Sheet margin, which did not reach its maximum position until approximately 14.5 ^{14}C ka BP (17.3 cal ka BP), and the southeastern Innuitian Ice Sheet margin, which in places lingered near modern ice margins until 17–18 ^{14}C ka BP (Dyke et al., 2002).

Field evidence and theory have made it clear that bed topography and complex ice divide configurations inevitably lead to the flow within large ice sheets being organized into drainage basins (ice-flow basins?) centered on large ice streams (Marshall and Clarke, 1996). Heinrich events recorded in Labrador Sea indicate that the Hudson Strait ice stream was a, if not the, premier ice stream of the LIS, but also that its discharge behavior was highly erratic. Within the late Wisconsin, H-2 was approximately synchronous with the Last Glacial Maximum (LGM). Another massive event took place at ~16.5 cal ka BP (H-1), and H-0 is represented by an advance of ice from

Labrador across Hudson Strait approximately coeval with the Younger Dryas cold climatic event. In Baffin Bay, glacial activity in the High Arctic Channels is recorded by detrital carbonate events, of which the largest event occurred at approximately 14.5–15 cal ka BP and thus lagged behind H-1 (Figure 1).

By 1969, only a decade after the general advent of radiocarbon dating as a geologic tool, sufficient dates had been obtained to map the isochrons on the retreat of the North American ice sheets (Bryson et al., 1969; Prest, 1969). These data and interpretations were updated in 1987 and again more recently (Dyke and Prest, 1987; Dyke et al., 2002; Dyke, 2004). In the 35 years since the first compilations, there has been an enormous increase in the number of dates, but the overall pattern of retreat between ca. 22 and 6 cal ka BP has held remarkably constant (Figure 3). The overall retreat is strongly asymmetrical, with early and rapid retreat along southern, western, and northwestern margins, possibly associated with a combination of thin ice and calving into glacial lakes and seas (Andrews, 1973), and limited retreat along the eastern margin (Dyke, 2004) (Figure 3). However, the demise of the LIS was accompanied by the development of large glacial lakes in the Mackenzie, Hudson Bay, and St. Lawrence drainage basins and by the incursion of marine waters into Hudson



Figure 3 Isochrons on retreat of the Laurentide, Cordilleran, and Innuitian ice sheets. The dates are based on the radiocarbon (ka) timescale.

Bay via Hudson Strait. The final drainage of the glacial lakes through Hudson Strait occurred at ca. 8.2 cal ka BP (Barber et al., 1999; Clarke et al., 2004). This resulted in the final massive discharge of carbonate-rich sediments onto the Canadian eastern continental margin (Figure 1). The last marine-based portion of the LIS to become ice-free was Foxe Basin (ca. 7.5 cal ka BP), and during the past 7 cal ka BP the remnant of the LIS has retreated across Baffin Island but is 'preserved' in the basal ice of the Barnes Ice Cap.

Of particular interest to the paleoclimate community has been the export of freshwater from the margins of the LIS to the North Atlantic Ocean and attempts to link these outburst floods to specific Late Glacial climate events, such as during the Younger Dryas and 8.2 cal ka BP cold events (Barber et al., 1999; Clark et al., 2001; Clarke et al., 2003; Teller et al., 2005). However, despite decades of research on the chronology and routing of the various large glacial lakes, there is still considerable uncertainty concerning the precise timing of critical events (Teller et al., 2005), even though the links to resultant changes in the global thermohaline circulation appear probable.

New insights into the location of ice margins during late Wisconsinan (MIS 2) have come from the extensive application of cosmogenic exposure dating to areas of the Eastern Canadian Arctic, the High Canadian Arctic, the Canadian Rockies, and mountain glaciers in Alaska, Canada, and the continental United States. The application of this method has resulted in at least partial resolution of the 'Big Ice' versus 'Little Ice' controversy of the 1970s (Denton and Hughes, 1981). Specifically, evidence indicates that an Innuitian Ice Sheet merged with the Greenland Ice Sheet at the head of Baffin Bay and with the Laurentide Ice Sheet to the south (Dyke et al., 2002), that cold-based ice covered the forelands of eastern Baffin Island with retreat starting 12–13 cal ka BP, and that the Saglék Moraines of the Torngat Mountains (Clark et al., 2003) are post-LGM (i.e., Late Glacial) in age (~13 cal ka BP).

See also: **Glaciations: Mid-Quaternary in North America; Overview.** **Lake Level Studies: North America.**

References

- Adam DP (1976) Hudson Bay, Lake Zissaga, and the growth of the Laurentide Ice Sheet. *Nature* 261(5562): 679–680.
- Andrews JT (1973) The Wisconsin Laurentide Ice Sheet: Dispersal centers, problems of retreat, and climatic implications. *Arctic and Alpine Research* 5: 185–199.
- Andrews JT (1987) The late Wisconsin glaciation and deglaciation of the Laurentide Ice Sheet. In: Ruddiman WF and Wright HEJ (eds.) *North America and Adjacent Oceans during the Last Deglaciation*, pp. 13–37. Boulder, CO: Geological Society of America.
- Andrews JT (1998) Abrupt changes (Heinrich events) in late Quaternary North Atlantic marine environments: A history and review of data and concepts. *Journal of Quaternary Science* 13: 3–16.
- Andrews JT and Mahaffy MA (1976) Growth rates of the Laurentide Ice Sheet and sea level lowering. *Quaternary Research* 6: 167–183.
- Andrews JT, Kirby ME, Aksu A, Barber DC, and Meese D (1998) Late Quaternary detrital carbonate (DC) events in Baffin Bay (67°–74°N): Do they correlate with and contribute to Heinrich events in the North Atlantic? *Quaternary Science Reviews* 17: 1125–1137.
- Barber DC, Dyke A, Hillaire-Marcel C, et al. (1999) Forcing of the cold event of 8200 years ago by catastrophic drainage of Laurentide lakes. *Nature* 400: 344–348.
- Bond G, Heinrich H, Broecker WS, et al. (1992) Evidence for massive discharges of icebergs into the glacial Northern Atlantic. *Nature* 360: 245–249.
- Briner JP, Miller GH, Davis PT, Bierman PR, and Caffee M (2003) Last Glacial Maximum ice sheet dynamics in Arctic Canada inferred from young erratics perched on ancient tors. *Quaternary Science Reviews* 22: 437–444.
- Bryson RA, Wendland WM, Ives JD, and Andrews JT (1969) Radiocarbon isochrones on the disintegration of the Laurentide Ice Sheet. *Arctic and Alpine Research* 1: 1–14.
- Budd WF and Smith IN (1981) The growth and retreat of ice sheets in response to orbital radiation changes. *International Association of Hydrological Science* 131: 369–409.
- Chappell J, Omura A, Esat T, et al. (1996) Reconciliation of late Quaternary sea levels derived from coral terraces at Huon Peninsula with deep sea oxygen isotope records. *Earth and Planetary Science Letters* 141: 227–236.
- Clague JJ and Ward B (2011) Pleistocene Glaciation of British Columbia. In: Ehlers J, Gibbard PL, and Hughes PD (eds.) *Quaternary Glaciations – Extent and Chronology. A closer look*, pp. 563–573. Amsterdam: Elsevier.
- Clark PU (1994) Unstable behavior of the Laurentide Ice Sheet over deforming sediment and its implications for climate change. *Quaternary Research* 41: 19–25.
- Clark PU and Lea PD (1992) *The Last Interglacial–Glacial Transition in North America*. Boulder, CO: Geological Society of America.
- Clark PU, Marshall SJ, Clarke GKC, et al. (2001) Freshwater forcing of abrupt climate change during the last glaciation. *Science* 293: 283–287.
- Clark PU, Brook EJ, Raisbeck GM, Yiou F, and Clark J (2003) Cosmogenic ¹⁰Be ages of the Saglék Moraines, Torngat Mountains, Labrador. *Geology* 31: 617–620.
- Clarke G, Leverington D, Teller J, and Dyke A (2003) Superlakes, megafloods, and abrupt climate change. *Science* 301: 922–923.
- Clarke GKC, Marshall SJ, Hillaire-Marcel C, Bilodeau G, and Veiga-Pires C (1999) A glaciological perspective on Heinrich events. In: Clark PU, Webb RS, and Keigwin LD (eds.) *Mechanisms of Global Climate Change at Millennial Time Scales*, pp. 243–262. Washington, DC: American Geophysical Union.
- Clarke GKC, Leverington DW, Teller JT, and Dyke AS (2004) Paleohydraulics of the last outburst flood from Glacial Lake Agassiz and the 8200 BP cold event. *Quaternary Science Reviews* 23: 389–408.
- Denton GH and Hughes TJ (1981) *The Last Great Ice Sheets*. New York: Wiley.
- Dyke AS (2004) An outline of North American deglaciation with emphasis on central and northern Canada. In: Ehlers J and Gibbard PL (eds.) *Quaternary Glaciations—Extent and Chronology Part II*, pp. 373–424. New York: Elsevier.
- Dyke AS and Prest VK (1987) Late Wisconsinan and Holocene history of the Laurentide Ice Sheet. *Geographie Physique et Quaternaire* 41: 237–263.
- Dyke AS, Andrews JT, Clark PU, et al. (2002) The Laurentide and Innuitian ice sheets during the Last Glacial Maximum. *Quaternary Science Reviews* 21: 9–31.
- Fisher DA, Reeh N, and Langley K (1985) Objective reconstruction of the late Wisconsinan Laurentide Ice Sheet and the significance of deformable beds. *Geographie Physique et Quaternaire* 39: 229–238.
- Fulton RJ (ed.) (1989) *Quaternary Geology of Canada and Greenland*. Boulder, CO: Geological Society of America.
- Hesse R, Klauke I, Ryan WBF, et al. (1997) Ice-sheet sourced juxtaposed turbidites systems in Labrador Sea. *Geoscience Canada* 24: 3–12.
- Hillaire-Marcel C, de Vernal A, Bilodeau G, and Wu G (1994) Isotope stratigraphy, sedimentation rates, deep circulation, and carbonate events in the Labrador Sea during the last ~200 kyr. *Canadian Journal of Earth Sciences* 31: 63–89.
- Hughes T (1987) Ice dynamics and deglaciation models when ice sheets collapsed. In: Ruddiman WF and Wright HEJ (eds.) *North America and Adjacent Oceans during the Last Deglaciation*, pp. 183–220. Boulder, CO: Geological Society of America.
- Imbrie J, Berger A, and Shackleton NJ (1993) Role of orbital forcing: A two million-year perspective. In: Eddy JA and Oeschger H (eds.) *Global Change in the Perspective of the Past*, pp. 263–277. New York: John Wiley and Sons Ltd.
- Jackson LE, Andriashek LD, and Phillips FM (2011) Limits of Successive Middle and Late Pleistocene Continental Ice Sheets, Interior Plains of Southern and Central Alberta and Adjacent Areas. In: Ehlers J, Gibbard PL, and Hughes PD (eds.) *Quaternary Glaciations – Extent and Chronology. A closer look*, pp. 575–589. Amsterdam: Elsevier.
- Marshall SJ and Clarke GKC (1996) Geologic and topographic controls on fast flow in the Laurentide and Cordilleran ice sheets. *Journal of Geophysical Research* 101: 17,827–17,839.
- Marshall SM (2002) Modelled nucleation centres of the Pleistocene ice sheets from an ice sheet model with subgrid topographic and glaciologic parameterizations. *Quaternary International* 95–96: 125–137.
- Prest VK (1969) Retreat of Wisconsin and Recent Ice in North America. *Geological Survey of Canada*. Ottawa: Queen's Printer.
- Teller JT, Bouyd M, Yang Z, Kor PSG, and Fard AM (2005) Alternative routing of Lake Agassiz overflow during the Younger Dryas: New dates, paleotopography, and re-evaluation. *Quaternary Science Reviews* 24: 1890–1905.

Late Pleistocene in South America

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Introduction

An updated review of the current state of the knowledge of late Pleistocene glaciations in South America, previously published in the first edition of this Encyclopedia (Coronato and Rabassa, 2007), is presented here. As this article is a revision of the previous work, only the contributions published during the last 5 years are discussed.

The glacial sequences of the Andean Cordillera are presented here separately for the Northern, Central, and Southern Andes.

The chronology of glacial advances has been refined in recent years, mostly through the use of different dating techniques, mainly cosmogenic radionuclides (^{10}Be , ^{26}Al) applied in several geographical areas such as the tropical Andes of Bolivia, the arid mountains of northern Chile and northwestern Argentina, Patagonia, the Magellan Straits, and Tierra del Fuego. In the Patagonian eastern side of the Andes, the occurrence of volcanic flows related to the different drifts favored the use of $^{39}\text{Ar}/^{40}\text{Ar}$ incremental heating dating technique as an option. Radiocarbon dating has also been undertaken on many lake-sediment cores and peat and organic sediments related to glacial deposits. In addition to advances in glacial chronology, precise regional mapping has improved and geomorphological knowledge has increased, mainly through the use of digital terrain models (DTMs) and satellite imagery as mapping tools. In addition, paleoseismic, stratigraphical, and sedimentological analyses have been undertaken on sediment cores from tropical and Patagonian lakes and fjords, which have contributed to the understanding of the late Pleistocene climate changes.

Detailed reviews of the Last Glacial Maximum (LGM) in Patagonia are given by Rabassa (2008), Coronato and Rabassa (2011), Martínez et al. (2011), and Rabassa et al. (2011), while detailed information on the glacial advances during the Wisconsinan Late Glacial and the Holocene in the Central Andes is presented by Rodbell et al. (2009).

Ice Advances in the Northern Andes (11° N–9° S, Higher Elevations at ~5000 masl)

Following the review by Coronato and Rabassa (2007), further advances in the late Pleistocene of the Northern Andes glaciations have been presented mainly by Mahaney et al. (2008) and Stansell et al. (2010).

In contrast to previous information, the regional equilibrium line altitude (ELA) depression during the LGM has been calculated in a range from 500 to 1625 m and a mean of $\sim 1200 \pm 300$ m (Lachniet and Vázquez-Selem, 2005). The wide range of elevation indicates that the ELA, as a paleoclimatic indicator of ice volume, should be analyzed and compared carefully, considering factors such as elevation, slope

orientation, insolation rate, and precipitation gradient. In the Eastern Mérida Andes of Venezuela (8° N–71° W), the early phases of the LGM are represented in a sequence of till, proglacial sediments, and peat and paleosols of the Pedregal sequence more than 40 m thick (Dierszowsky et al., 2005). The overlying sediments of the Pedregal Interstadial were dated at ca. 60 ka BP. Full-glacial and Late-glacial conditions returned between ~ 23 and 19 ka BP.

Late-glacial advances in the Venezuelan Andes were reported by Mahaney et al. (2008) based on geomorphology, sedimentology, and soil analyses on the glacial and glaciofluvial deposits that overlie older tills and peat beds. The authors proposed that ice had advanced ca. 12.4 cal ka BP in the Humboldt Massif and ca. 13.7–13.3 cal ka BP in catchments of the Eastern Cordillera. They correlated these advances with the Northern Hemisphere Younger Dryas cooling event. These discoveries were later confirmed by paleolimnological studies in the Mérida Andes by Stansell et al. (2010), who postulated that glaciers advanced at ca. 12.8 cal ka BP, reaching their maximum extent at ca. 12.6 cal ka BP. The ELA elevations were calculated by these authors between 360 and 480 masl, while the temperature was calculated as 2.2–2.9 °C colder than modern values. These authors also suggested that this Late-glacial advance reflects climate variations in good agreement with paleoclimatic records from marine sediment records of the Cariaco Basin.

A synthesis of new numerical ages presented here is shown in Figure 1.

Ice Advances in the Central Andes (9°–39° S, Higher Elevations at ~7000 masl)

The Central Andes is nowadays mainly affected by two systems of atmospheric circulation. While the north is influenced by the southeasterly South American Seasonal Monsoon (SASM) which brings humidity from Amazonia during the summer, the southern mountains receive rain from the Westerlies during the winter. This explains the climatic variability of the region, which had a major impact on glacier dynamics and landscape evolution. As a consequence, the debate about the interpretation of the information obtained from proxy records is still ongoing. According to Kull et al. (2008), several research groups have pointed out that an increase in summer precipitation occurred between 14 and 10 ka BP, forcing a Late-glacial ice advance in the western Altiplano, whereas the LGM was forced by a temperature decrease in the Eastern Cordillera at 16°–23° S. Moreover, pre- and post-LGM (32–14 ka BP) ice advances occurred at 30° S, related to increased precipitation and reduced temperature.

Millennial-scale climatic variations between 60 and 20 ka BP have been recognized in the tropical Andes in sediment

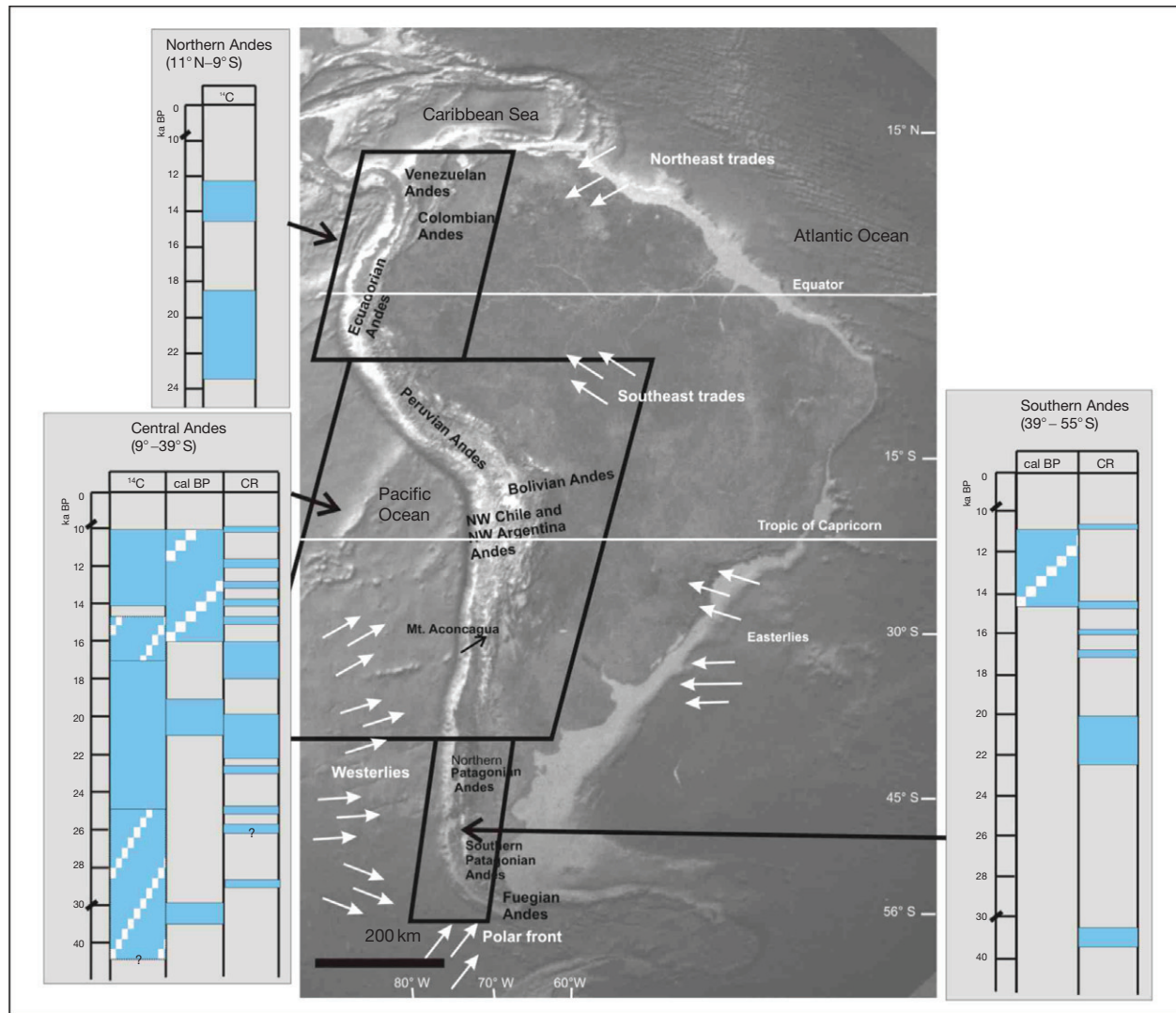


Figure 1 Distribution of the Andes Cordillera in South America and the influence of global atmosphere moist air movement. The ages constraining glacial and cold climatic periods are shown in the light blue boxes; they are differentiated by dating methods. The white lines in the boxes show that glacial advances have occurred through a weakly bracketed period. A question mark indicates where maximum ages are unknown.

cores from Lake Titicaca. Very distinctive sedimentary, geochemical, and paleobiological characteristics have allowed the recognition of wet regional climates. These events have been correlated to cold events recorded in the Greenland ice cores, indicating that near-global scale climate excursions controlled the increase of the SASM during the late Pleistocene (Fritz et al., 2010).

Glacial models of the northwestern Argentinian Andes (22°S) allowed Kull et al. (2008 and papers cited therein) to propose that a massive temperature fall of 4.5–8 °C, combined with a moderate precipitation increase, was required to force glaciers to expand during marine isotope stage (MIS) 2 (25–18 ka BP). This fact shows that the more humid slopes of the Andes were sensitive areas for glaciations during the LGM. This arose from temperature decrease, whereas on the more arid slopes, the glacier advances were forced by precipitation increase later during Wisconsinan Late-glacial times (Kull et al., 2008 and papers cited therein). The atmospheric

mechanisms resulting in humidity increase have been sought in a southward displacement of the Intertropical Convergence Zone (ITCZ) and the presence of La Niña climatic conditions (Zech et al., 2010).

In their regional synthesis, Zech et al. (2008) postulated that glaciers in the northern parts of the tropical Andes were temperature sensitive, while those located in the southern and western parts were precipitation sensitive. Little evidence of LGM ice advances has been found. These late cooling stages were found to be in phase with high Andean lake transgressions.

Clastic sediment flux analyses in alpine lakes in the tropical Andes have been used by Rodbell et al. (2009) as a proxy indicator for regional ice extent. They proposed that high clastic influx, triggered by increased precipitation, began before ~40 ka BP and culminated between 29 and 20 ka BP. This observation compares closely with the LGM in the high Bolivian and Peruvian valleys. Another regional clastic influx to the lakes

was between 16 and 12 ka BP, again corresponding to the Late-glacial period, in which several sets of moraine have been recorded in many valleys.

Pre-LGM (Before 25 ka BP)

Magnetic susceptibility analysis of radiocarbon-calibrated sediments from Lake Titicaca has suggested that the LGM began at ca. 60 ka BP, during the last part of MIS 4 in this region. Calcium carbonate and diatom-content records have suggested that the rise of lake level, probably reflecting precipitation increase, took place preceding glacial advances in the mountains (Fritz et al., 2010). In contrast to the results from the well-known Funza site (in the Northern Andes) where glacial climates were identified as cold and dry phases, these authors concluded that the glacial phases were cold and wet on the Bolivian Altiplano.

Likewise, further south, in the northern/central Chilean Andes, it has been argued that the most extensive glaciation occurred earlier than 32 ka, based on ^{10}Be surface-exposure dating and the degree of morainic erosion (Zech et al., 2011). Nevertheless, the scarcity of available geomorphological and cosmogenic exposure data does not allow the dating of the climatic deterioration which forced glaciers to grow.

Climatic conditions that were more humid during pre-LGM times than at present were determined from multidisciplinary lacustrine investigations of the Lake Tagua-Tagua basin (36° S, Chile), at the foot of the Andes (Valero-Garcés et al., 2005). High-stand levels, which occurred before 43.5 cal ka BP and particularly between 40 and 21.5 cal ka BP, were interpreted as resulting from precipitation increase in response to enhanced westerly wind activity. The available information supports Smith et al. (2005) who suggested that the local LGM might have occurred late in MIS 3, rather than close to 21 ka BP. Similar results were recently presented by Zech et al. (2011), who provided evidence that full-glacial conditions occurred ca. 39 ka BP in the southern Central Andes (Ruca Choroí valley, 39° S) and that further glacial advances did not occur there until the Late Glacial. Based on ^{10}Be boulder exposure ages on moraines, they suggested that increased precipitation arising from a northward shift of the Westerlies was the reason for glacier enlargement along the Central Andes, from northern Chile and Argentina (22° S) to around 39° S (middle Neuquén province, Argentina).

LGM (25–18 ka BP)

The LGM in the Cordillera Blanca (9° S–77° W, Peruvian Andes) was studied by Farber et al. (2005). They reported that the Rurec moraine includes boulders that yield surface-exposure ages of between ca. 20 and 30 ka BP and an average age of ca. 23.4 ka BP. They considered this moraine the indicator of the LGM limit in the Cordillera Blanca. However, cosmogenic ages of ca. 29 ka BP suggest that they represent the beginning of glacier advances from the Western Cordillera before the LGM developed fully.

A few valleys of the Cordillera Huayhuash (10° S, 76° W, ~4800 masl) in the Peruvian Andes preserve geomorphological evidence of glacial advances. ^{10}Be exposure surface ages and radiocarbon dating from peat bogs indicate that the LGM advance occurred there at 20–22 ka and before 26 ka

(Hall et al., 2009). This is consistent with the LGM chronology proposed by other authors in other regions of the tropical Andes. The similar geomorphological character and ^{10}Be surface-exposure dates of moraines in the middle and upper Peruvian Andes valleys allowed Smith et al. (2005) to conclude that LGM conditions were established in this region at ca. 32–21 ka BP.

Maximum glacial conditions were also studied using the ^{10}Be surface-exposure technique on morainic boulders from the Cordillera Real and the Cordillera Cochabamba by Zech et al. (2007). Their conclusions correspond closely with previous studies by Smith et al. (2005) and Farber et al. (2005), who also presented radiocarbon dates and lake-sediment analyses. These results indicate that maximum ice extent occurred ca. 22–25 ka BP. After testing a climatic model, Kull et al. (2008) concluded that a massive temperature depression of about 6.4 °C had allowed the glaciers to advance during this time.

Glacial deposits are well exposed in the filling of the Tarija–Padcaya basin, on the eastern side of the Bolivian Cordillera. Based on geomorphology, stratigraphy, and sedimentological analysis, Coltorti et al. (2010) reported MIS 4-age glacial landforms and deposits in this region for the first time. The sedimentary record in this basin ended at ca. 22 ka BP, but erosional glacial landforms present suggest that a major glacial advance occurred in the basin. This new evidence allowed these authors to hypothesize that glaciers flowed from the Bolivian Cordillera to the lower surrounding basins both during MIS 4 and 2.

In close correspondence with the Peruvian–Bolivian records previously presented by several authors, Zech et al. (2010) obtained results from ^{10}Be analyses of boulder exposure surfaces and radiocarbon dating of an 8-m sediment core from a glacial lake in the Sierras de Santa Victoria region (NW of Argentina, 22° S). Here, the LGM ice advance is represented by moraines dated older than 22.3 and 25.6 ka BP. These dates correlate with the lake-sediment records from the Uyuni Basin and Sajsi paleolake in Bolivia (Placzek et al., 2006). The glaciers' advance was forced by increased moisture advection onto the Altiplano (located to the west) resulting from the southward position of the ITCZ (Zech et al., 2010).

The Encierro (29° S) and Cordón Doña Rosa valleys (31° S), which are located along the 'Arid Diagonal' in Chile, were studied by Zech et al. (2008) who applied ^{10}Be exposure surface dating to morainic boulders. Here, the LGM advance could not be securely dated in both valleys because of the absence of prominent moraines and scarcity of boulders for dating or landscape instability and weathering. Further south, and also along the western slope of the Andes, the Laguna Tagua-Tagua (34° S, Chile) sediment core showed records that imply much higher precipitation during the full-glacial period compared to today, particularly during 40–21 cal year BP (Valero-Garcés et al., 2005), which could provide sufficient snow for glacier formation on the Andean summits during a global period of temperature decrease.

At almost the same latitude, but on the eastern slope of the Central Andes (36°–39° S, Neuquén, Argentina), glacial landforms and glaciofluvial deposits have been recognized between 1400 and 2900 masl and interpreted as LGM in age, although there are few numerical dates to confirm this. A detailed review of the alpine glacial advances during the late Pleistocene in this

region has been presented by Rabassa (2008) and Rabassa et al. (2011).

As at least two frontal moraine suites occur in several valleys of their eastern slopes, it is thought that lobes might have spread out from the volcanic peaks once or twice, but the detail remains uncertain. Cirques, hanging valleys, and frontal moraines can also be seen in the southern Central Andes, along the Chachil and Catan Lil ranges (500–2700 masl). Because of their elevation above the LGM equilibrium line (1500–2000 masl) and the low degree of erosion they display, they have been interpreted as belonging to the last glacial advance (Rabassa et al., 2011). Although detailed geomorphological and chronostratigraphic studies are required to support a glaciation model of this region of the Central Andes, it is thought that glaciers would have been shorter and thinner than those of the Southern Andes as a result of the lower availability of moisture from the Pacific Ocean.

Late Glacial (18–10 ka BP)

At the Cordillera Blanca (9° S) in the Peruvian Andes, boulder surface-exposure ages indicate that ice advances ranging from 18 to 16 ka BP with an average age of 16.6 ka BP occurred during the Wisconsin Late Glacial (Farber et al., 2005). They were interpreted as the second glacial advance during MIS 2, which retreated very quickly because no major moraine suites are preserved up the valley.

Geomorphological and sedimentary evidence in the high Central Peruvian Andean valleys show that Late-glacial advances occurred on both slopes of the Huayhuash Cordillera (10° S). Hall et al. (2009) found that glaciers advanced between 16 and 10 ka BP and postulated that temperature drop was the dominant control on glacier extent. These data corroborate the findings of Smith et al. (2005) from the Peruvian Andes, which indicated that glacial advances occurred between 20–15 and 14–12 ka BP.

At the Cordillera Cochabamba (17° S, Bolivia), glacial advances were recognized during the Heinrich Event 1 and Younger Dryas events by using ^{10}Be surface-exposure dating techniques. Zech et al. (2008) have obtained ^{10}Be exposure surface ages indicating that Late-glacial advances occurred at ca. 15 and 12 ka BP in the Cordillera Real and the Cordillera Cochabamba glaciers (16°–17° S). In addition, recent work has shown that the most extensive glaciation was recorded by exposure ages of 14.2 and 13.6 ka, while an inner lateral morainic arc was dated at 11.8–11.7 ka. The innermost moraine front yielded an exposure age of between 10.1 and 10.2 ka BP (Zech et al., 2010). This led these authors to propose that moisture advection, precipitation increase in the mountains, and glacial advances were forced by El Niño/Southern Oscillation (ENSO)-like conditions. Further south, on the northern slope of the Nevado Illimani (Cordillera Real, Bolivia), the age of glacial advances was also estimated by Smith et al. (2010) with the use of ^{10}Be cosmogenic nuclide surface dating. These authors found that glaciers retreated from moraines at 15.5–13.0 ka for the last time during the Late Glacial, suggesting that western-derived precipitation increased during this time, and a marked fluctuation has occurred in the east–west ELA gradient, located almost 500 m higher today.

At 22° S, in the Sierras de Santa Victoria (NW Argentina), glacier advances are known from boulder surface-exposure dating (^{10}Be) to have occurred at up to 14.5 and 13.8–12 ka BP (Zech et al., 2010). This coincides with lake transgression phases on the Altiplano, emphasizing the importance of precipitation for the positive ice mass balance in the arid regions of the Central Andes.

In the Encierro valley and the Cordón Doña Rosa (29°–30° S, Northern–Central Chile), boulder surface-exposure dating by ^{10}Be analyses performed on very prominent moraines indicates that extensive Late-glacial advances occurred ca. 14.7–11.6 ka BP (Zech et al., 2008). Further south, at Lago Tagua-Tagua (36° S), at the foot of the Andes, evidence of more humid climates was presented by Valero-Garcés et al. (2005). Here, humid intervals were detected in the lake-sediment record between 19.5–17 and 13.5–11.5 cal ka BP. Although no evidence of glacial landforms was related to these climatic models, they quite closely correspond to the younger group of moraines found in the Encierro valley and the Cordón Doña Rosa.

Eastward, along the Río Mendoza valley (2500 masl), close to the eastern slope of Mount Aconcagua (Figure 1), the ‘Horcones Drift,’ previously interpreted as pre-LGM glacial deposits, was reinterpreted as distal flows of rock avalanches by Fauqué et al. (2009). Based on the mineralogy, sedimentology, and cosmogenic radionuclide dating, these authors postulated that deposits at the bottom of the valley are remnants of megalandslides that occurred on the southern slope of Mount Aconcagua, rejecting their interpretation as moraine deposits. The landslides occurred upon the Horcones paleoglacier and the sediments were carried by the ice to their final deposition down valley when the ice melted, during the Late Glacial. This new interpretation changed the regional glacial model, as proposed by previous authors, and emphasized that ancient mass-wasting process products are much more frequent than previously thought.

New Late-glacial cosmogenic radionuclide dates were provided by Zech et al. (2010) in the Ruca Choroí valley, southern Central Andes (39° S). Based on ^{10}Be exposure ages on boulder ages of between 17.9 and 14.8 ka BP, they postulated that the glacier was still occupying the upper part of the valley during Late-glacial times, but that most of the valley had become ice free by ca. 15.5 ka BP. They also suggested that minor ice readvances have occurred only in the cirques, where one exposure age of ca. 12.2 ka BP was obtained. Although these are quite interesting conclusions, more statistically based mean ages are required to establish a robust chronology for this region.

A synthesis of the new numerical ages presented in this article is shown in Figure 1.

Ice Advances in the Southern Andes (39°–55° S, Higher Elevations at 2000 masl)

LGM

The LGM is represented by the Nahuel Huapi, Segunda Angostura, and Moat drifts in the Northern and Southern Patagonian and Fuegian Andes, respectively.

Detailed information concerning the LGM extent and chronology in the more than 20 transverse eastern Andean valleys from 39° to 52° S were presented by Rabassa (2008), Martínez

et al. (2011), and Coronato and Rabassa (2011). These authors agree that the LGM limit is represented by one or more morainic arcs located close to the eastern heads of the glacial lakes and the marine channels.

In the Northern Patagonian Andes (39° S), the eastern frontal limit of the LGM glaciers was blocked by the end of the Central Andes ranges and volcanic tablelands. In most cases, piedmont glaciers were formed when alpine glaciers joined outside the main valleys, at the foot of the ranges and tablelands. No numerical ages are available for the LGM chronology in this region because of the lack of suitable material for dating. It is accepted that the glaciation had occurred in phase with that in southern Patagonia, that is, ca. 25 cal ka (Coronato and Rabassa, 2011; Rabassa, 2008), and perhaps simultaneously on both slopes of the Andes.

Further south, in the valley of Lago Buenos Aires (46° S, Argentina), ^{10}Be and ^{26}Al surface-exposure ages on erratic boulders from moraines were obtained by Douglass et al. (2006). They suggested that the LGM occurred here between 22.7 and 19.9 ka BP, broadly corresponding to the proposed ages of 27–25 and 23–22 ka BP found in the Lago Pueyrredón valley (47° S) by Hein et al. (2010). This closely matches the Magellan Straits (53° S) chronology. In this latter area, the LGM limits have been dated by cosmogenic nuclides at 25 ka BP (Kaplan et al., 2008). This number is well corroborated with other cosmogenic and radiocarbon dating formerly published by different authors.

Glasser et al. (2011) mapped several moraine belts along the Lago San Martín glacial valley (49° S) and performed ^{10}Be cosmogenic exposure ages on boulders. The results indicate that the moraine suite enclosing the lake formed at 22.4 ka BP. This date agrees with that for the Fenix Moraines of the northern Lago Buenos Aires region and with the southern Río Blanco moraines in Lago Pueyrredón. In addition, exposure surface ages of ca. 34–37 ka BP for boulders from an eastward morainic belt indicate that the maximum extent of ice was reached earlier, demonstrating that this valley was less covered by ice during the LGM than previous studies had suggested.

The LGM age is still not well bracketed in Tierra del Fuego (53°–55° S), although geomorphological mapping has shown very conspicuous moraine belts along all the main and tributary valleys (Coronato and Rabassa, 2011). Recent work in the Lago Fagnano area (54° S) has shown that two moraine suites occur 10–20 km apart; here they were interpreted as two LGM positions before general deglaciation (Coronato et al., 2009). Although the outer moraine system has been dated by thermoluminescence (TL) to 25.7 ka BP, and outer glaciofluvial deposits were dated at 16.2 ka BP, these ages should be considered preliminary. Along the Beagle Channel (55° S), terminal moraines have been mapped, but an appropriate chronology is yet to be established here (Coronato and Rabassa, 2011). The end of the LGM cycle has been identified by several authors at ca. 17 ka BP.

Late Glacial

Several glacial readvances have been proposed for the Wisconsin Late Glacial in the Chilean Lake District in particular, and for the eastern side of the Southern Andes in general (Coronato and Rabassa, 2007; Rabassa, 2008). Along the northern and southern shores of major glacial lakes, frontal moraines that descend

beneath the lake waters are quite visible. They are always located westward from the LGM terminal moraines and at much lower elevations than those within the cirques. Based on the radiocarbon age of basal peat, cosmogenic dating, and seismic surveys, the moraines were interpreted to be younger than those located at the heads of the lakes. Seismic surveys allowed Waldmann et al. (2010) to identify additional, previously undescribed, buried moraine suites (Coronato et al., 2009).

It is accepted that southern Patagonia and Tierra del Fuego were affected by a climatic deterioration that occurred in phase with the Antarctic Cold Reversal (ACR), whereas the cooling would have occurred later in the Northern Patagonian Andes, much closer to the Northern Hemisphere Younger Dryas Chron.

Glacial advances that occurred in the Lago Buenos Aires basin after the LGM have been identified at 17, 15.8, and 14.4 ka BP by Douglass et al. (2006) using ^{10}Be exposure dating techniques on morainic boulders.

Further south, in the Lago San Martín valley (49° S), Late-glacial-age moraines dammed the lake arm named 'Brazo Maipú' and provide evidence of past high-standing lake levels. A single boulder on the inner moraine crest yielded a ^{10}Be age of 14.3 ka BP, which closely corresponds to the age of other boulders. Collectively, these boulders indicate the age of the innermost moraines formed along this valley.

On the western side of the southernmost Andes (Gran Campo Nevado and Sound Skyring region, 53° S, Chile), a slowdown of the fjord-glacier retreat was noted by Kilian et al. (2007) between 14 and 11 ka BP, which was followed by complete ice retreat. Although they are close to each other, evidence in these valleys does not reflect deglaciation and Late-glacial advances as much as in the Magellan Straits. This is probably due to the different mountain ice caps and the calving effect at the ice fronts.

Along the Lago Fagnano (54° S) and the Beagle Channel (55° S), minimum basal radiocarbon ages from peat bogs show that glaciers definitively receded from the main valleys between 14.6 and 10 ka BP, but a detailed chronology of recessional moraines is not yet available.

According to Douglass et al. (2006), the 14.4 ka BP advance clearly reflects the impact of the ACR instead of a YD influence, thus indicating asynchrony with the Northern Patagonian Andes and with the Northern and Central Andes during the Late Glacial. This is potentially a result of the northern limit of the Westerlies. New evidence confirming this was presented by Moreno et al. (2009) from the western side of the Southern Patagonian Andes (50° S), where evidence for glacier readvances and colder episodes was bracketed by radiocarbon dates of between 14.8 and 12.6 ka BP. On the contrary, at the same latitude but on the eastern side of the Andes, boulder surface-exposure ages of the outermost Late-glacial moraines of Lago Argentino suggested glacier advances toward the end of the Younger Dryas Chron (Ackert et al., 2008). These advances were attributed to higher precipitation rather than temperature decrease. The ^{10}Be and ^{36}Cl mean exposure age of 10.8 ka BP, found by these authors, is inconsistent with previous results from the area but coherent with higher precipitation and resulting lake high-stands reported for other northern regions of semiarid Patagonia following the severe drought of the ACR times.

A synthesis of new numerical ages presented in this article is shown in [Figure 1](#).

Final Remarks

Knowledge of late Pleistocene glaciations in South America has improved greatly over the last 20 years. However, the chronological scheme and the paleoclimatic interpretation are still far from complete, with much work still being needed in certain areas. For instance, the late Pleistocene glaciations in the Equatorial Andes and Northern Andes are still poorly constrained. Apart from sediment-core results from the highest lake in the Chorreras valley ([Rodbell et al., 2009](#)) and research results previous to 2007, no new information has been found in the recent literature.

It is very difficult to establish only one setting for late Pleistocene glaciations that would cover all the Andean latitudinal belt. This is because of the highly different summit elevations (from 7000 to 1000 masl) and broadly varying atmospheric conditions. For example, past glaciers should have responded to atmospheric changes arising from the SASM and ITCZ circulation patterns in the Central Andes, but depending upon the slope orientation of the mountains.

It is clear that in the southern Central Andes, as well as in the Patagonian Andes, the westerly winds have controlled the moisture supply for glacial ice accumulation, independent of mountain elevation.

Several core drilling projects performed in the Central and Southern Andes lakes have provided excellent information on glacial sediment and water discharge and allowed paleoclimatic reconstructions, suggesting that these paleolimnological techniques are perhaps the most promising methodology for the near future.

The extreme aridity of the Central Andes, together with the lack of contact between the ice and the Patagonian forest, are responsible for the scarcity of organic material appropriate for radiocarbon dating. In recent years and with increasing frequency, ^{10}Be and other cosmogenic isotope surface-exposure dating techniques have provided chronological information about glacier advances and retreats. However, sufficient field evidence is still lacking to provide an adequate framework to the available numerical dates. Likewise, methodological uncertainties of cosmogenic exposure dating techniques conspire against regional correlations in South America, where the intervals between ice advances in different regions seem to be shorter than the usual method errors.

Late-glacial advances occurred all along the Andes, mostly in two phases, the younger close to the ACR and YD cooling events, the older very close to 19–16 cal ka BP. These circumstances are particularly noted in the high Central Andes.

In the southern latitudes of South America (beyond 46° S), the atmospheric dynamics and glacial growth were influenced not only by the Westerlies but also by the Antarctic Frontal Zone. Late-glacial ice readvances after the general LGM deglaciation seem to have taken place under the control of the ACR. This suggests a different response of glacier advances during the Late-glacial period in the Northern–Central Andes in relation to the Southern Andes, and a precise, reliable chronology for both areas is still needed.

The discrepancies between the northern and southern Patagonian response to millennial-scale climatic variation and how these regions are in-phase with the ACR episode or to the Northern Hemisphere YD event is a current research topic. Somewhere between the Magellan Straits region and the Northern Patagonian Andes, there was a switch in atmospheric influence from one condition to the other, but its location is still unknown.

Much scientific work is still urgently required and many areas are still blank, but significant progress has been achieved recently and a breakthrough in the knowledge of South American glaciations is clearly achievable within the next decade, thanks to multiple, interdisciplinary efforts from Colombia to Cape Horn.

See also: [Glaciations: Late Quaternary in North America; Overview.](#)

References

- Ackert R, Becker R, Singer B, Kurz M, Caffee M, and Mickelson D (2008) Patagonia glacier response during the Late Glacial–Holocene transition. *Science* 321: 392–395.
- Coltorti M, Pieruccini P, and Paredes RF (2010) Late Pleistocene stratigraphy, sedimentology and paleoenvironmental evolution of the Tarija–Padcaya basin (Bolivian Andes). *Proceedings of the Geologists Association* 121: 162–179.
- Coronato A and Rabassa J (2007) Late Quaternary glaciations in South America. In: Scott E (ed.) *Encyclopedia of Quaternary Science*, vol. 2, pp. 1101–1108. Amsterdam: Elsevier.
- Coronato A and Rabassa J (2011) Pleistocene glaciations in southern Patagonia and Tierra del Fuego. In: Ehlers J, Gibbard PL, and Hughes PD (eds.) *Quaternary Glaciations - Extent and Chronology, Part IV - A Closer Look*, vol. 15. Amsterdam: Elsevier.
- Coronato A, Seppälä M, Ponce F, and Rabassa J (2009) Glacial geomorphology of the Pleistocene Lake Fagnano ice lobe, Tierra del Fuego, southern South America. *Geomorphology* 112: 67–81.
- Dirszowsky R, Mahaney W, Hodder K, et al. (2005) Lithostratigraphy of the Mérida (Wisconsinan) glaciation and Pedregal Interstade, Mérida Andes, northwestern Venezuela. *Journal of South American Earth Science* 19: 525–536.
- Douglass D, Singer B, Kaplan M, Mickelson D, and Caffee M (2006) Cosmogenic nuclide surface exposure dating on boulders on last-glacial and late-glacial moraines and paleoclimatic implications. *Quaternary Geochronology* 1: 43–58.
- Farber D, Hancock G, Finkel R, and Rodbell T (2005) The age and extent of tropical alpine glaciation in the Cordillera Blanca, Perú. *Journal of Quaternary Science* 20(7–8): 759–776.
- Fauqué L, Hemanns R, Hewitt K, et al. (2009) Mega-deslizamientos de la pared sur del Cerro Aconcagua y su relación con depósitos asignados a la Glaciación Pleistocena. *Revista de la Asociación Geológica Argentina* 65(4): 691–712.
- Fritz SC, Baker P, Ekdahl E, Seltzer G, and Stevens L (2010) Millennial-scale climate variability during the last glacial period in the tropical Andes. *Quaternary Science Reviews* 29: 1017–1024.
- Glasser N, Jansson K, Goodfellow B, de Angelis H, Rodnight H, and Rood D (2011) Cosmogenic nuclide exposure ages for moraines in the Lago San Martín Valley, Argentina. *Quaternary Research* 75(3): 636–646. <http://dx.doi.org/10.1016/j.yqres.2010.11.005>.
- Hall SR, Farber D, Ramage J, et al. (2009) Geochronology of Quaternary glaciations from the Tropical Cordillera Huayhuash, Peru. *Quaternary Science Reviews* 28: 2991–3009.
- Hein A, Hulton N, Dunai T, Sudgen D, Kaplan M, and Xu Sh (2010) The chronology of the Last Glacial Maximum and deglacial events in central Argentine Patagonia. *Quaternary Science Reviews* 29: 1212–1227.
- Kaplan M, Fogwill C, Sudgen D, Hulton N, Kubik P, and Freema S (2008) Southern Patagonian glacial chronology for the last glacial period and implications for Southern Ocean climate. *Quaternary Science Reviews* 27: 284–294.

- Kilian R, Schneider C, Koch J, et al. (2007) Paleoeological constraints on Late Glacial and Holocene ice retreat in the Southern Andes (53° S). *Global and Planetary Change* 59: 49–66.
- Kull C, Imhof S, Grosjean M, Zech R, and Veit H (2008) Late Pleistocene glaciation in Central Andes: Temperature versus humidity control – A case study from the eastern Bolivian Andes (17° S) and regional synthesis. *Global and Planetary Change* 60: 148–164.
- Lachniet M and Vázquez-Selem L (2005) Last Glacial Maximum equilibrium line altitudes in the circum-Caribbean (Mexico, Guatemala, Costa Rica, Colombia and Venezuela). *Quaternary International* 138–139: 129–144.
- Mahaney W, Milner M, Kalm V, Dirszowsky R, Hancock R, and Beukens R (2008) Evidence for a Younger Dryas glacial advance in the Andes of northwestern Venezuela. *Geomorphology* 96: 199–211.
- Martínez O, Coronato A, and Rabassa J (2011) Pleistocene glaciations in Northern Patagonia, Argentina: An updated review. In: Ehlers J, Gibbard PL, and Hughes PD (eds.) *Quaternary Glaciations – Extent and Chronology – A Closer Look, Developments in Quaternary Science*, vol. 15, Chapter 52: 729–734. Amsterdam: Elsevier.
- Moreno P, Kaplan M, Villa-Martínez R, Moy C, Stern C, and Kubik P (2009) Renewed glacial activity during the Antarctic Cold reversal and persistence of cold conditions until 11.5 ka in southwestern Patagonia. *Geological Society of America Bulletin* 37(4): 375–378.
- Placzek C, Quade J, and Patchett PJ (2006) Geochronology and stratigraphy of late Pleistocene lake cycles on the southern Bolivian Altiplano: Implications for causes of tropical climate change. *Bulletin of the Geological Society of America* 118: 515–532.
- Rabassa J (2008) Late Cenozoic glaciations in Patagonia and Tierra del Fuego. In: Rabassa J (ed.) *The Late Cenozoic of Patagonia and Tierra del Fuego. Developments in Quaternary Science*, vol. 11, pp. 151–204. Amsterdam: Elsevier.
- Rabassa J, Coronato A, Ponce JF, Schlieder G, and Martínez O (2011) Geoformas y depósitos glaciogénicos en la Provincia de Neuquén. In: Leanza H (ed.) *Relatorio de la Geología y Recursos Naturales de la Provincia de Neuquén*, pp. 295–314. Buenos Aires: Asociación Geológica Argentina.
- Rodbell D, Smith J, and Mark B (2009) Glaciations in the Andes during the Late Glacial and Holocene. *Quaternary Science Reviews* 28: 2165–2212.
- Smith CA, Lowell T, Owen L, and Caffee M (2010) Late Quaternary glacial chronology on Nevado Illimani, Bolivia, and the implications for paleoclimatic reconstructions across the Andes. *Quaternary Research* 75(1): 1–10. <http://dx.doi.org/10.1016/j.yqres.2010.07.001>.
- Smith J, Seltzer G, Rodbell D, and Klein A (2005) Regional synthesis of last glacial maximum in the Tropical Andes, South America. *Quaternary International* 138–139: 145–167.
- Stansell N, Abbott M, Rull V, Rodbell D, Bezada M, and Montoya E (2010) Abrupt Younger Dryas cooling in the northern tropics recorded in lake sediments from the Venezuelan Andes. *Earth and Planetary Science Letters* 293: 154–163.
- Valero-Garcés B, Jenny B, Rondanelli M, et al. (2005) Paleohydrology of Laguna Tagua-Tagua (34°30'S) and moisture fluctuations in Central Chile for the last 46000 yr. *Journal of Quaternary Science* 20(7–8): 625–641.
- Waldmann N, Aristegui D, Anselmetti F, Coronato A, and Austin J Jr. (2010) Geophysical evidence of multiple glacier advances in Lago Fagnano (54° S), southernmost Patagonia. *Quaternary Science Reviews* 29(9–10): 1188–1200.
- Zech R, Kull Ch, Kubik PW, and Veit H (2007) LGM and Late Glacial glacier advances in the Cordillera Real and Cochabamba (Bolivia) deduced from ¹⁰Be surface exposure dating. *Climate of the Past* 3: 623–635.
- Zech R, May J-H, Kull Ch, Ilgner J, Kubik P, and Veit H (2008) Timing of the late Quaternary glaciation in the Andes from ~15° to 40° S. *Journal of Quaternary Science* 23(6–7): 653–674.
- Zech R, Zech J, Kull C, Kubik PW, and Veit H (2011) Early last glacial maximum in the Southern Central Andes reveals northward shift of the westerlies at ~39 ka. *Climate of the Past* 7: 41–46.
- Zech J, Zech R, May JH, Kubik P, and Veit H (2010) Late Glacial and early Holocene glaciation in the tropical Andes caused by La Niña-like conditions. *Palaeogeography, Palaeoclimatology, Palaeoecology* 293: 248–254.

Neoglaciation in Europe

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Introduction

In Europe, the term neoglaciation tends to be used in two ways. First, it is used for the recrudescence or regrowth of glaciers in the mid-Holocene following partial or complete deglaciation in the early Holocene. Neoglaciation in this conventional sense tends to be viewed as a response to gradual (multimillennial) climatic deterioration after an early Holocene thermal maximum or 'climatic optimum.' Second, it is used with reference to shorter term, possibly abrupt, century- to millennial-scale glacier expansion episodes whenever they occurred within the Holocene. Usage has shifted toward this second model of neoglacial events, also known as 'Little Ice Age'-type events, as more evidence has come to light on early Holocene glacial activity, the complexity of Holocene glacier variations has become better known, and interest has grown in climate variability on century-to-millennial timescales. The methods used to reconstruct Holocene glacier variations in Europe, and the detailed evidence relating to neoglaciation that is now available, are reviewed here with particular reference to Scandinavia and the Alps.

Reconstructing Neoglaciation

A wide range of methods has been used to reconstruct and date neoglaciation in Europe. These can be classified into four main approaches: (1) historical and documentary evidence of glacier variations, (2) moraine-ridge stratigraphy and the dating of glacier expansion episodes, (3) dating glacier retreat phases using dendroglaciology and archaeoglaciology, and (4) reconstructing continuous records using glaciolacustrine and glaciofluvial stratigraphy. The principles and strengths and weaknesses of each approach are outlined in the following sections.

Historical and Documentary Evidence of Glacier Variations

Historical and documentary sources, which range from written accounts, etchings, paintings, and sketches to photographs, maps, and annual retreat measurements, provide the most accurate means of reconstructing the most recent glacier variations and may also be used for calibrating other methods. Such evidence has proved invaluable for building up a detailed record of glacier variations over the last ~500 years in the Alps and for shorter periods in many other glacierized areas of Europe, including Scandinavia and Iceland. This evidence is available, however, only for selected locations and recent centuries, and requires the application of high standards of historical analysis and data verification.

Moraine-Ridge Stratigraphy and the Dating of Glacier Expansion Episodes

For earlier in the Holocene, and/or in the absence of historical evidence, moraine-ridge stratigraphy has been the most widely used approach to the reconstruction of glacier variations.

Moraine-ridge sequences or moraine complexes characterize the marginal zones of many glaciers throughout Europe. In some places, the timing of the glacier advances that produced the moraines has been estimated by radiocarbon dating of buried organic material, by tephrochronology (especially in Iceland), or by the application of rock-surface exposure-age dating techniques, including lichenometric dating, terrestrial cosmogenic-nuclide dating (TCND), and Schmidt-hammer exposure-age dating (SHD).

This approach has been most successful where either (1) moraine ridges that are well separated horizontally and can be clearly identified or (2) where complex lateral moraine sequences formed by accretion or superimposition can be investigated in vertical sections. The latter situation is illustrated in [Figure 1](#), which shows the lateral moraine complex at Simonykees, a glacier in the Venediger massif of the Austrian Alps. Historically dated moraine ridges and calibrated radiocarbon dates are linked via peat, soil profiles, and associated pollen stratigraphy. Radiocarbon dates provide minimum limiting ages for the relatively old moraines, while the pollen profiles provide indirect evidence of the younger glacier variations based on the association between glacier advances, low values of arboreal pollen (AP), and high values of nonarboreal pollen (NAP, dominated by Gramineae), which is a characteristic feature of such proximal mires.

There are at least four limitations of the moraine-stratigraphic approach including, first, the evidence relates primarily to times of moraine deposition at the maxima of glacier advances or highstands. The approach sheds little light on the extent or timing of the intervening phases of glacier recession. Second, the likelihood that evidence will have been destroyed by more extensive later advances is high. This is a major limitation in southern Norway, where the 'Little Ice Age' glacier advance was normally the most extensive of the mid-to-late Holocene. Third, radiocarbon dates associated with wood, peat, and buried paleosols, may be characterized by large calibration or interpretive errors. In the case of well-developed buried paleosols, for example, large apparent mean residence times (AMRT) may lead to radiocarbon dates that are up to several thousand years older than the moraines that buried them. Fourth, organic material suitable for radiocarbon dating may be sparse or absent at high-altitude and high-latitude sites.

These problems have led to increasing use of rock-surface exposure-age techniques. Lichenometry has been very successful for dating 'Little Ice Age' recessional moraines but its effective temporal range does not extend beyond the last ~500 years. Despite limitations on the precision of the technique, recent applications of TCND have demonstrated potential in both Scandinavia and the Alps, especially in relation to early Holocene moraines dating from the Preboreal period and, possibly, the '8200 event.' After being limited for several decades to relative age dating, SHD has been developed recently in southern Norway as a calibrated-age dating technique. Under optimum conditions, SHD appears capable of dating

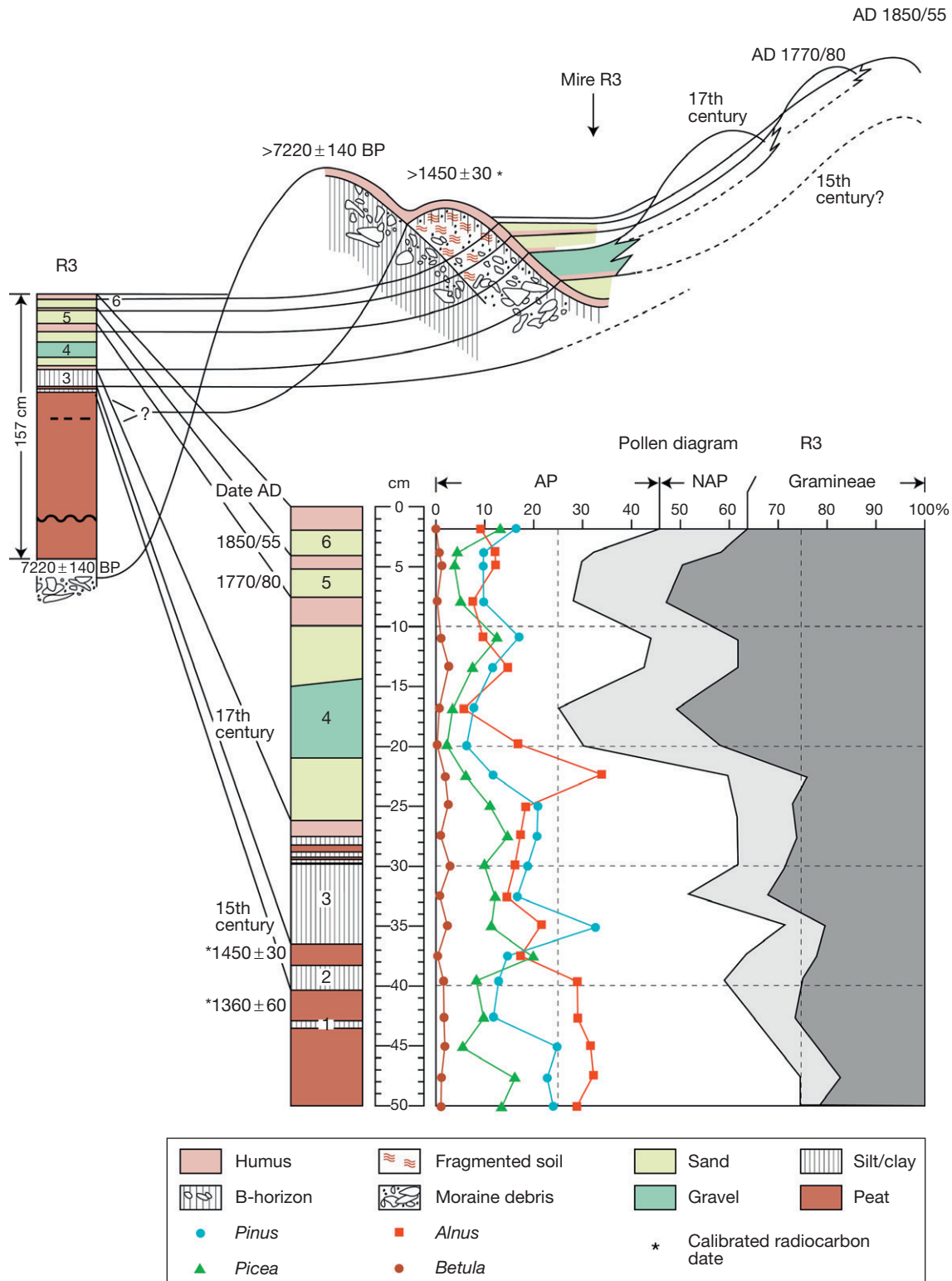


Figure 1 Moraine-stratigraphic evidence of Holocene glacier variations at Simonykees, Austrian Alps. Interrelationships between the moraine ridges are shown schematically; insets show the stratigraphic units (numbered 1–6) that are related to glacier advances, and the associated pollen diagram from the mire (R3) that lies distal to the historically dated seventeenth–nineteenth century lateral moraines; stars indicate calibrated radiocarbon dates (AD) from the mire. Adapted from Patzelt G and Bortenschlager S (1973) Die postglazialen Gletscher- und Klimaschwankungen in der Venedigergruppe (Hohe Tauern, Ostalpen). *Zeitschrift für Geomorphologie NF, Supplement Band 16*: 25–72.

with a higher level of precision than is currently available using TCND. This is demonstrated in Figure 2, which shows the 95% confidence intervals associated with the predicted ages of two pre-'Little Ice Age' moraine ridges according to SHD and TCND.

Dating Glacier Retreat Phases Using Dendroglaciology and Archaeoglaciology

Reconstruction of the duration and timing of phases of glacier recession, and the position of the glaciers when at their minimum extents, has been more difficult. Knowledge of these aspects is essential to gain a complete picture of Holocene glacier variations and hence neoglaciation. Unprecedented glacier retreat during the late twentieth and early twenty-first centuries has provided new evidence for a dendroglaciological approach to phases of reduced glacier extent in the Alps. This is based on the radiocarbon dating of pieces of wood exposed on glacier forelands. The setting indicates that trees grew up-valley of present glacier snouts and hence that the glaciers were

previously smaller than their present shrunken state. The age range of the wood samples also permits reconstruction of the duration of the main retreat phases for the whole Holocene and shows that the glaciers were smaller than today for much of the early to mid-Holocene. In some cases, *in situ* stumps enable a minimum extent of the glacier recession to be determined precisely. Up to 12 phases of reduced glacier extent have been defined using this approach in the Swiss Alps (Figure 3(a)).

A further important dimension of dendroglaciology is the establishment of tree-ring chronologies derived from the sub-fossil wood, which indicate climatic fluctuations that are correlated with glacier variations. As the proximity of the glacier to growing trees varies, the 'glacier climate' may accentuate and modify the climatic signal expressed in the tree rings. This approach has considerable potential for reconstructing high-resolution glacier variations.

Following the hot summer of 2003, reduction in size of the Schnidejoch ice field in the Swiss Alps uncovered over 100 archaeological artifacts (hunting gear, fur, leather, woollen

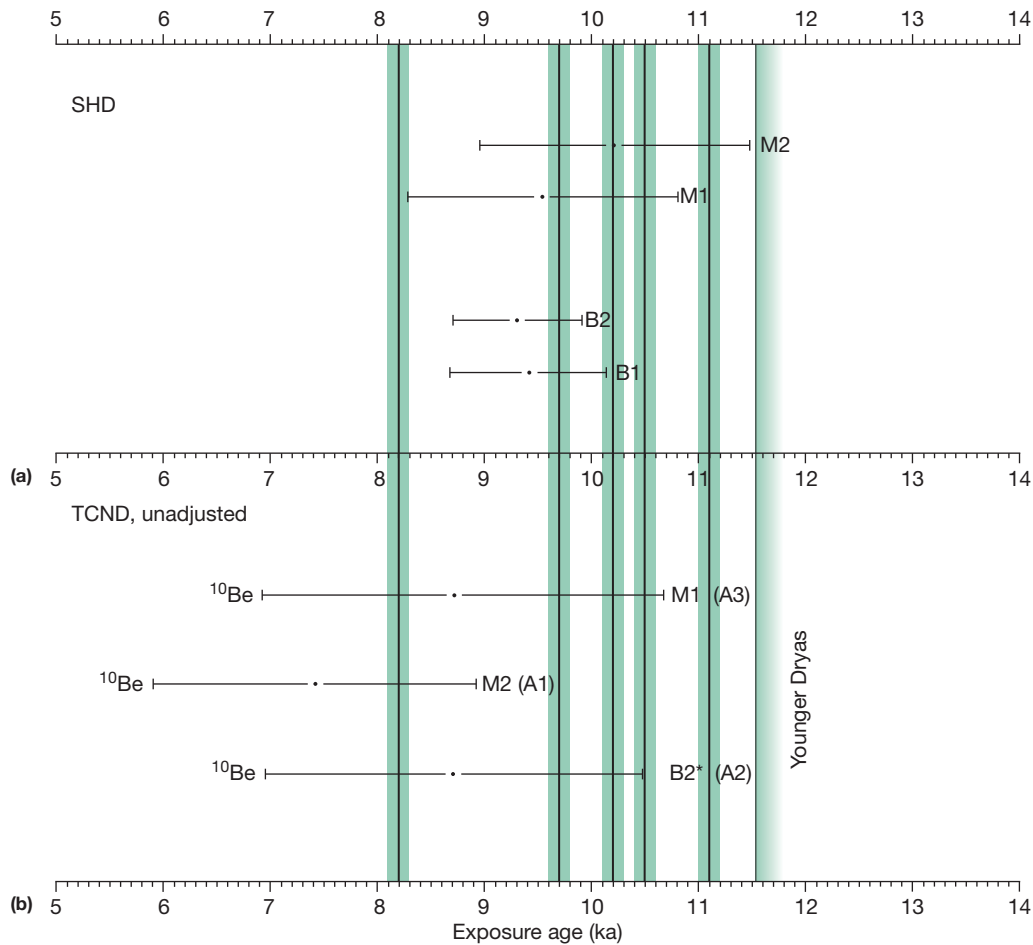


Figure 2 (a) Four Schmidt-hammer exposure-age dates (SHD) for bedrock (B1, B2) and boulder (M1, M2) samples associated with two moraine ridges (M1, M2) at Austanbotnbreen, Jotunheimen, Norway, derived from various calibration curves. (b) Three terrestrial cosmogenic-nuclide dates (TCND using ^{10}Be) for boulders associated with the same moraines. The 95% confidence intervals shown for each date are narrower for SHD, the more precise estimates of which are capable of rejecting an 8.2 ka age for the moraines. Gray bands indicate known southern Norwegian neoglacial events. Adapted from Matthews JA and Winkler S (2011) Schmidt-hammer exposure-age dating (SHD): Application to early Holocene moraines and a reappraisal of the reliability of terrestrial cosmogenic-nuclide dating (TCND) at Austanbotnbreen, Jotunheimen, Norway. *Boreas* 40: 256–270.

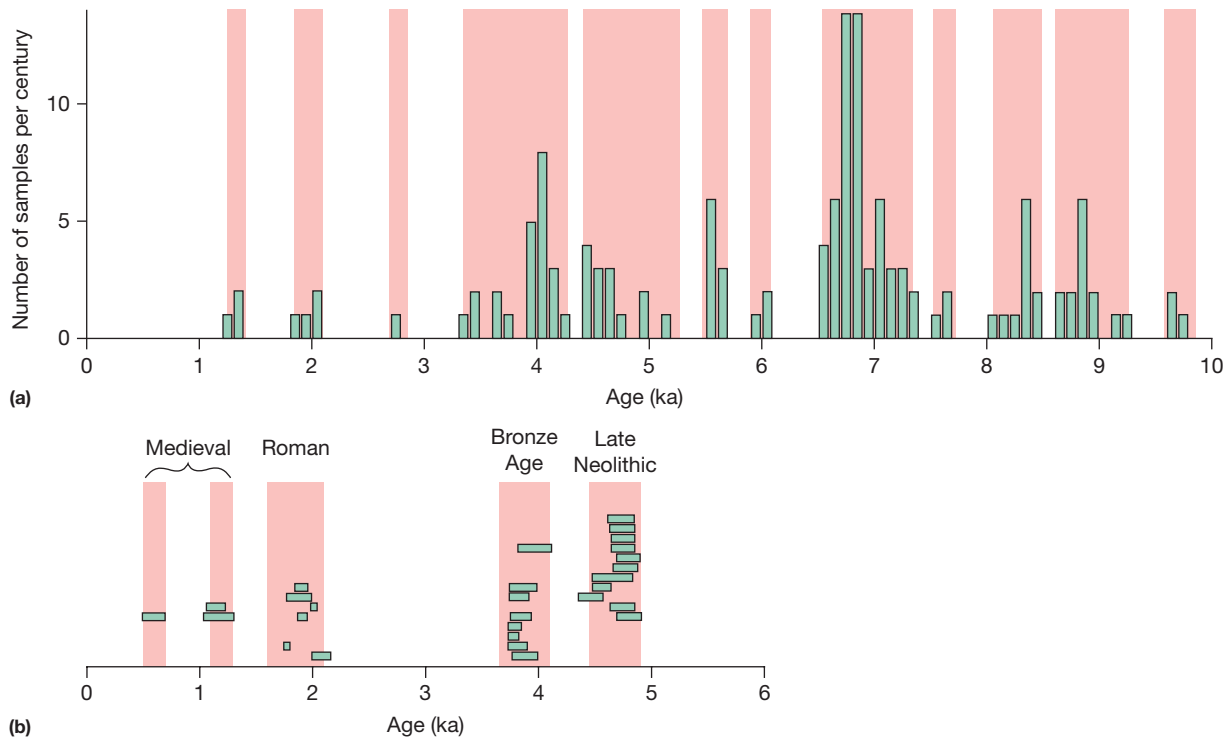


Figure 3 (a) Frequency of radiocarbon dates on fossil wood from glacier forelands indicative of glacier recession in the Swiss Alps: 12 recession phases are indicated by green bands. Adapted from Joerin UE, et al. (2006) Multi-century glacier fluctuations in the Swiss Alps during the Holocene. *The Holocene* 16: 697–704. (b) Radiocarbon dates (95% probability bars) on archaeological finds from Schnidejoch icefield. Four phases of glacier recession, which correspond with the Late Neolithic, Bronze Age, Roman and Medieval Warm Periods, are indicated by green bands. Adapted from Grosjean M, et al. (2007) Ice-borne prehistoric finds in the Swiss Alps reflect Holocene glacier fluctuations. *Journal of Quaternary Science* 22: 203–207.

clothing, tools, shoe nails, and coins, which have been dated by radiocarbon dating and archaeological means (Figure 3(b)). The ages of the finds fall into four distinct clusters, which indicate periods when glacier retreat was sufficient to open a transalpine route connecting northern Italy to the northern Alps: namely, 4900–4450 cal years BP, 4100–3650 cal years BP, first to third century AD, and eighth to fifteenth century AD (the last corresponding with the Medieval Warm Period). The preservation of Neolithic leather indicates, moreover, more-or-less permanent ice cover at that site since ~4900 cal years BP. This date is quite close to that of the ‘Alpine Iceman’ or ‘Ötzi,’ the remains of which were found in a similarly remarkable state of preservation after the melting-out of glacier ice close to the border between Austria and Italy. Both sets of finds are consistent with the onset of Neoglaciation, in the conventional sense, at or before ~5300 cal years BP (the date of the iceman himself and his associated artifacts).

Reconstructing Continuous Records Using Glaciolacustrine and Glaciofluvial Stratigraphy

Whereas all the approaches derived from the glacier foreland depend on piecing together fragmentary evidence, two types of distal sedimentary sequences have the potential for reconstructing records that are both continuous and high resolution. This approach has been widely applied in Scandinavia. In the case of glacier-fed lakes, glacier history is reflected in the minerogenic content of the lacustrine sediments. Strategically

located lakes that are sufficiently distal from this sediment source, or the second or third lake in a chain of lakes, have been found particularly useful. Upstream lakes act as traps for coarse sediment, whereas lower lakes contain a record of variations in the deposition of suspended sediment. Consequently, glacier presence and relatively large glacier extent are recorded by various sedimentological indicators such as visible silt layers, varves, high X-ray density, low weight on ignition (organic content), high silt content, high magnetic susceptibility, and lithological variations.

An example of variations in a glaciolacustrine sediment core is shown in Figure 4. Twenty-three lithostratigraphic units are recognized, which fall broadly into a tripartite structure. Organic-rich sediments, dating from ~10 000 to 2000 cal years BP, are sandwiched between silt-rich minerogenic materials deposited when glaciers were relatively extensive. The sedimentological indicators record short-term variations in glacier size, such as the rapid glacier expansion around 8200 cal years BP, and a longer-term neoglaciation buildup of glaciers after ~2000 cal years BP. In this case, high calcium carbonate content is indicative of relatively small glaciers because the glaciers within this catchment are located on noncalcareous bedrock.

Similar indicators can be used in the context of glaciofluvial stream-bank mires, which are flooded episodically by melt-water when overbank deposition of suspended sediment occurs. The example in Figure 5 shows the lithostratigraphy in two such mires where glacier expansion episodes are recorded by predominantly silty minerogenic units. During episodes of

Bøvertunsvatnet cores 1 and 2, jotunheimen

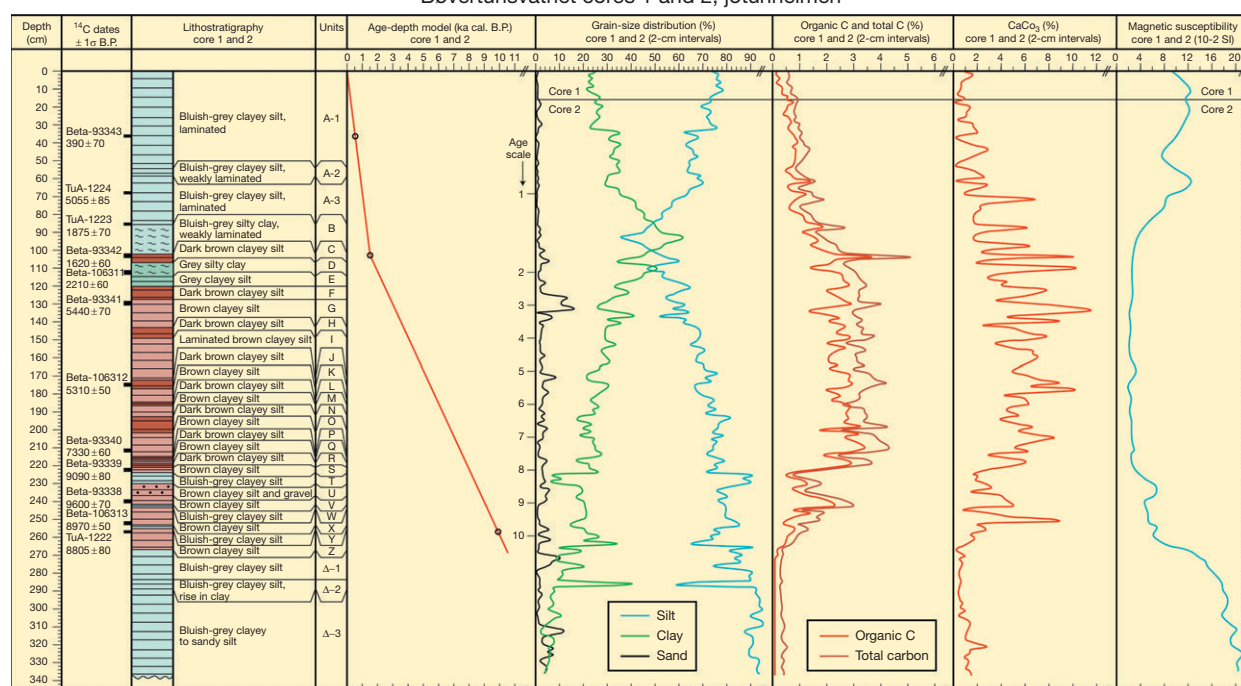


Figure 4 Glaciolacustrine evidence of century- to multimillennial-scale Holocene glacier variations at Bøvertunsvatnet, Jotunheimen, Norway. (Left to right) Lithostratigraphy, primarily of core 2, with uncalibrated radiocarbon dates; age–depth curve in calibrated radiocarbon years BP; silt, clay, and sand content (%); organic carbon and total carbon content (%); calcium carbonate content (%); and magnetic susceptibility (10^{-2} SI). The interpretation is shown in **Figure 6(b)**. Adapted from Matthews JA, et al. (2000) Holocene glacier variations in central Jotunheimen, southern Norway based on distal glaciolacustrine sediment cores. *Quaternary Science Reviews* 19: 1625–1647.

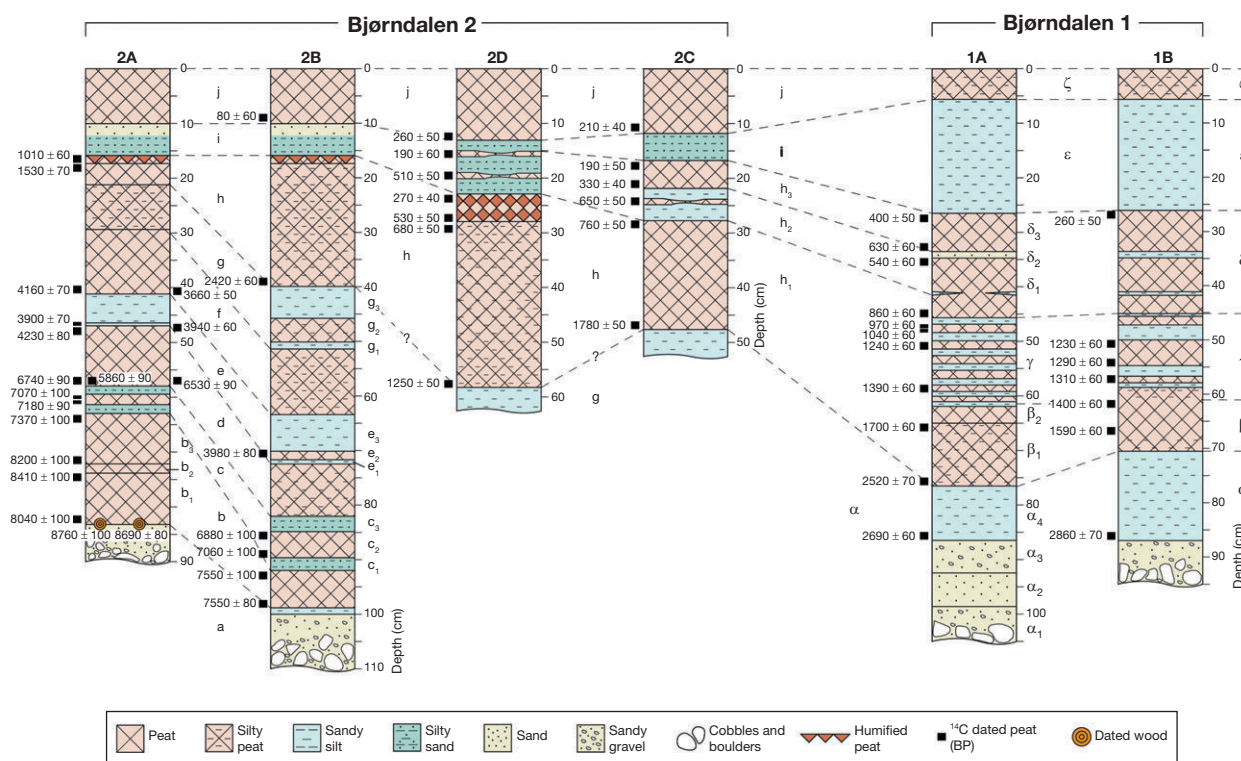


Figure 5 Glaciofluvial evidence of century- to millennial-scale Holocene glacier variations from stream-bank mires downstream of Bjørnbreen, Jotunheimen, Norway. Lithostratigraphy and uncalibrated radiocarbon dates are shown based on six excavated profiles in the two mires, Bjørndalen 1 and 2. The interpretation is shown in [Figure 6\(a\)](#). Adapted from Matthews JA, et al. (2005) Holocene glacier history of Bjørnbreen and climatic reconstruction in central Jotunheimen, Norway, based on proximal glaciofluvial stream-bank mires. *Quaternary Science Reviews* 24: 67–90.

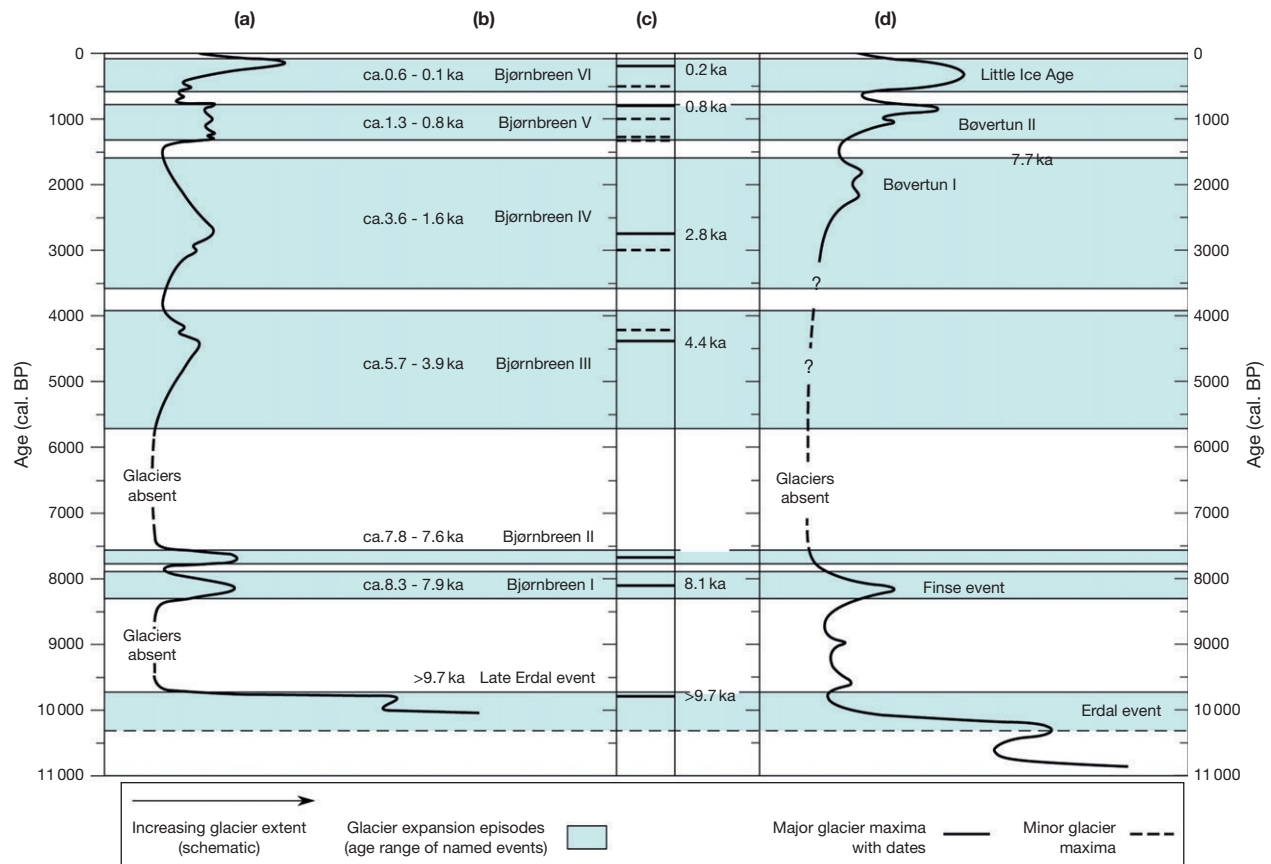


Figure 6 Reconstructed Holocene glacier variations from sites on opposite sides of the Smørstabbtindan massif, Jotunheimen, Norway. (a) Variations of Bjørnbreen (based on the evidence shown in Figure 5); (b) named local neoglacial events (also shown as horizontal shaded bands); (c) the age of major and minor glacier maxima; and (d) variations of Leirbreen/Bøverbreen on the west side of the massif (based on the evidence shown in Figure 4). All dates are in calibrated radiocarbon years. Adapted from Matthews JA, et al. (2005) Holocene glacier history of Bjørnbreen and climatic reconstruction in central Jotunheimen, Norway, based on proximal glaciofluvial stream-bank mires. *Quaternary Science Reviews* 24: 67–90.

reduced glacier extent, peat accumulation proceeds with little interruption. Depending on availability, glaciolacustrine and glaciofluvial sites can provide alternative and complementary strategies. The main limitations of these approaches are (1) differentiating the glacial signal from possible flood layers and colluvial inputs (which are generally coarser in texture), and (2) in the case of glaciofluvial sites, the possibility of changing patterns of deposition and erosion over the mire surface associated with discharge variations, shifting stream patterns, and the infilling of depositional basins. These limitations can generally be minimized by not relying on single sites and by taking full account of the geomorphological setting of the sites.

Pattern and Timing of Neoglaciation

Application of the aforementioned approaches, often in combination, has produced evidence of Holocene glacier variations from several regions of Europe, which have been summarized by Grove (2004). Here, emphasis is given to two regions – southern Scandinavia and the Alps – where the evidence has permitted reconstruction of the pattern and timing of neoglaciation in greatest detail.

Scandinavia

A near-continuous record of Holocene glacier variations in the Smørstabbtindan massif of Jotunheimen, southern Norway, which has been reconstructed from the glaciolacustrine and glaciofluvial evidence presented in Figures 4 and 5, is summarized in Figure 6. The results of the two approaches agree in most respects, despite relating to different glaciers on opposite sides of the massif. Using the conventional definition, the onset of neoglaciation can be recognized at ~5700 cal years BP and at least seven local neoglacial events can be identified.

The pattern for the Smørstabbtindan massif, taking account of results from additional sites now available (Figure 7(a)), is compared with similar detailed reconstructions from three other subregions of southern Scandinavia in Figure 7(b)–7(d). The four records indicate a fluctuating buildup of glaciers after the onset of neoglaciation at ~6500–6000 cal years BP, which culminated in the ‘Little Ice Age’ maximum (around the mid-eighteenth century in southern Norway). There is also good agreement on the absence of glaciers for prolonged intervals during the early to mid-Holocene, especially around 7000 cal years BP. Differences between the subregions can be accounted for by differences in glacier size, geometry, and

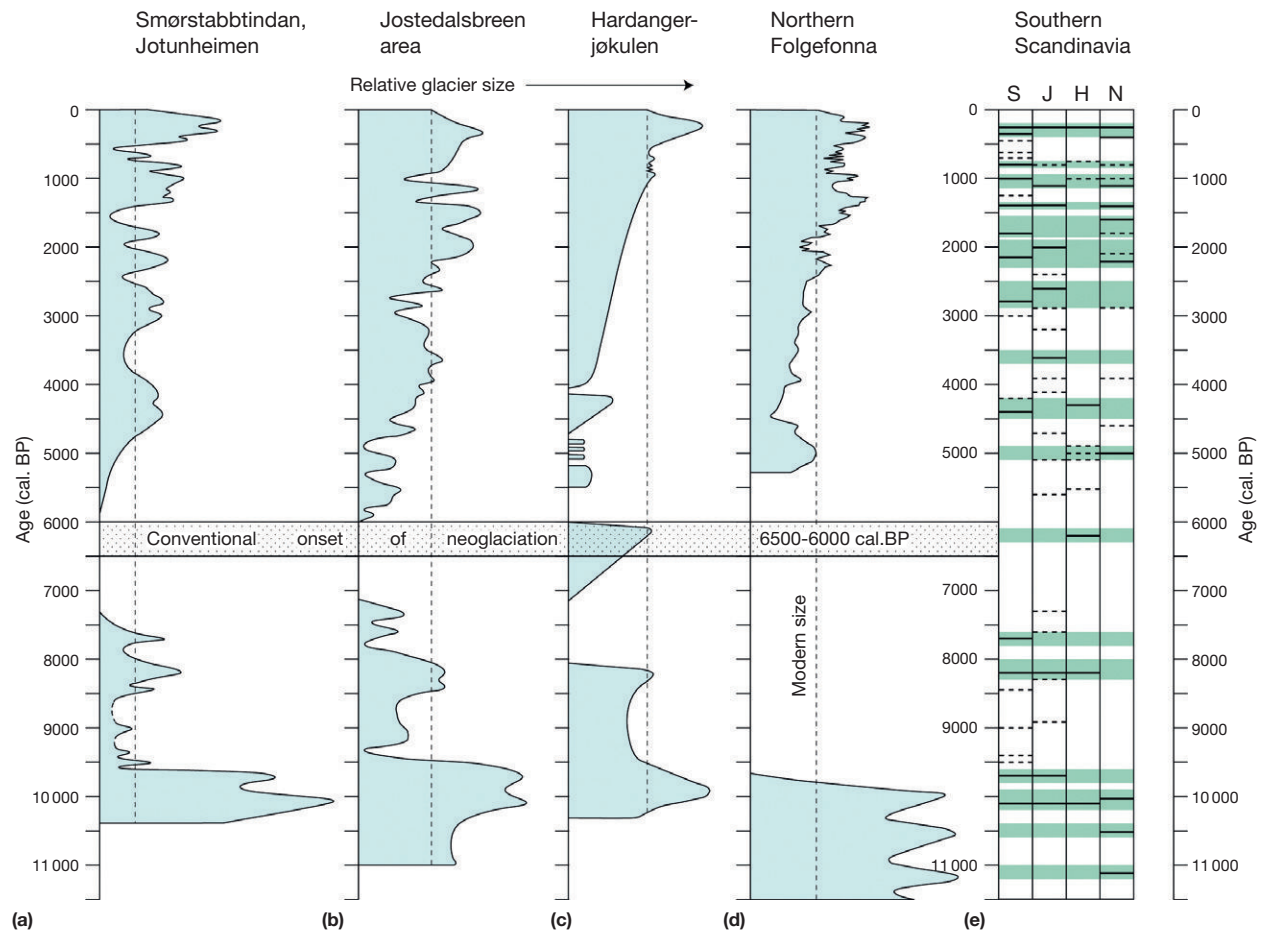


Figure 7 Holocene glacier variations in southern Scandinavia. (a–d) Estimated glacier size variations relative to modern for four glacierized subregions of southern Norway, and the conventional onset of neoglaciation (horizontal shaded band): (a) Smørstabbtindan, Jotunheimen (S; Matthews and Dresser, 2008); (b) Jostedalssbreen area (J; Nesje et al., 2000, 2001); (c) Hardangerjøkulen (H; Dahl and Nesje, 1996); (d) Nordre Folgefonna (N; Bakke et al., 2005b,c); and (e) southern Norwegian synthesis (shaded horizontal bands are the regional neoglaciation maxima shown as horizontal bars). Adapted from Matthews JA and Dresser PQ (2008) Holocene glacier variation chronology of the Smørstabbtindan massif, Jotunheimen, southern Norway, and the recognition of century- to millennial-scale European Neoglaciation Events. *The Holocene* 18: 181–201.

location, as well as dating uncertainties and the existence or not of a residual ice sheet at the beginning of the Holocene. The completeness of these records justifies the recognition of 17 regional neoglaciation events in southern Norway (Figure 7(e)), each defined by at least one major glacier maximum at the subregional scale. Ten of these major glacier maxima have been detected in at least two subregional records. Minor maxima, which are of limited scale or based on limited evidence, have not been used in defining the regional events.

Alps

A similar compilation for the Central and Eastern Alps is shown in Figure 8(a)–8(f). Up to 11 neoglaciation events were recognized as early as the 1970s, based on moraine stratigraphy (Figure 8(a)). More recently, dendroglaciology has revealed a similar number of glacier recession phases in Austria (Figure 8(b)) and in Switzerland (Figure 8(c)). A complementary record of glacier activity based on median grain size from

Lake Silvaplana in the Upper Engadine subregion of the south-eastern Swiss Alps (Figure 8(d)) indicates glaciers of restricted size between about 10 600 and 3500 cal years BP. After the latter date, continuous deposition of varves, together with relatively small median grain sizes, reflects an increase in glaciogenic silt and clay associated with conventional neoglaciation. This record is in good agreement with radiocarbon-dated peat from the glacier foreland of Rutor Glacier in the Italian Alps, which indicates that this glacier was smaller than today from at least 10 200 to 5800 cal years BP.

The summary of length variations of Swiss glaciers (Figure 8(e)) implies that, although neoglaciation events became more frequent in the late Holocene, each was of similar extent to the 'Little Ice Age' maximum of ~AD 1850. The Alpine synthesis shown in Figure 8(f) recognizes 23 Alpine regional neoglaciation events based on the glacier maxima recognizable in one or more of the individual records, while taking additional account of the relatively accurate glacier chronology for Swiss glaciers over the last 3500 years.

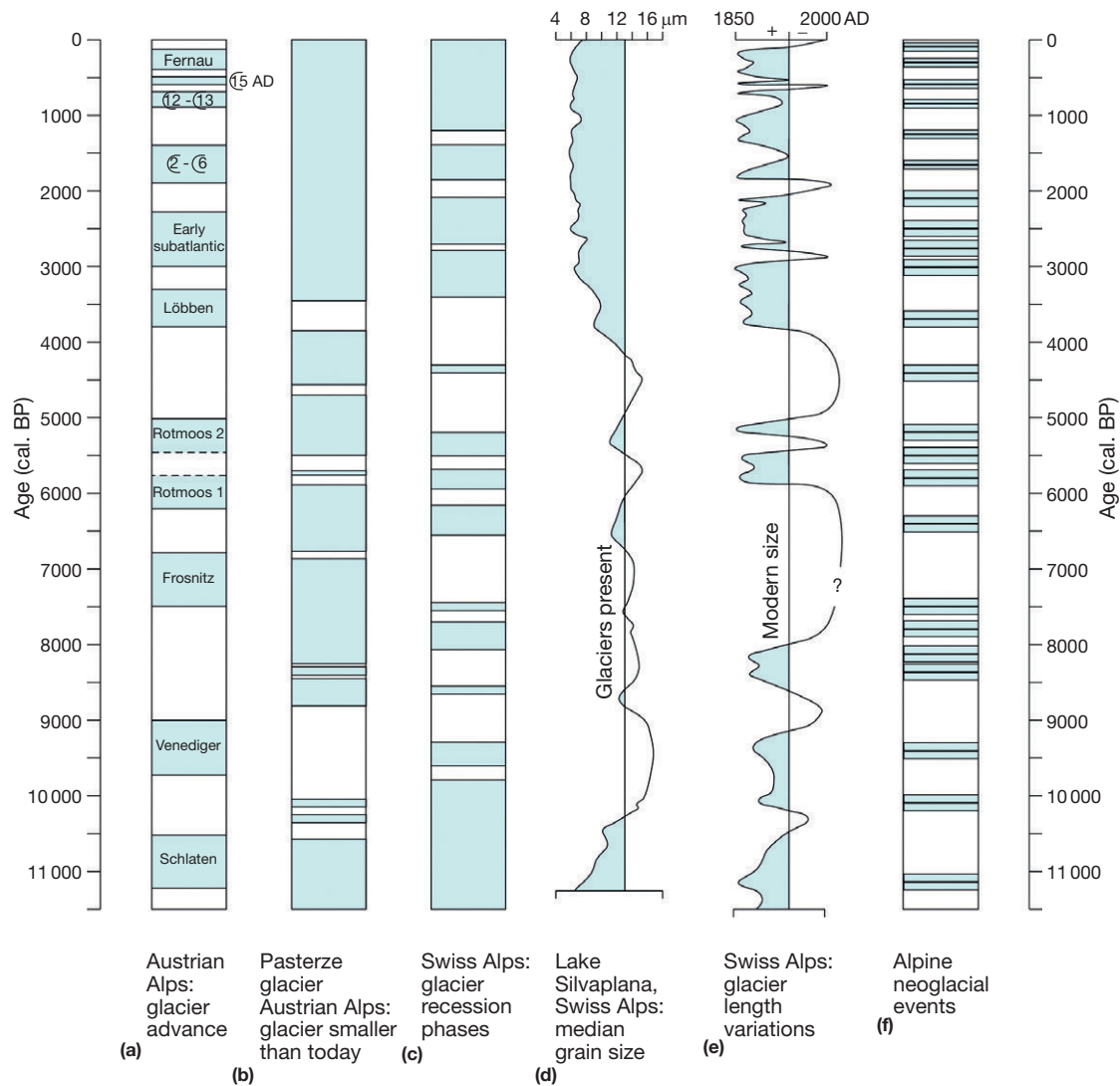


Figure 8 Holocene glacier variations in the Central and Eastern Alps. (a) Phases of glacier expansion in the Austrian Alps (Patzelt and Bortenschlager, 1973); (b) phases of glacier recession (white) of Austrian glaciers (Nicolussi and Patzelt, 2000b); (c) glacier recession phases (white) in the Swiss Alps (Hormes et al., 2001; Joerin et al., 2006); (d) median grain size variation in Lake Silvaplana, Swiss Alps (Leemann and Niessen, 1994); (e) glacier length variations in the Swiss Alps (Maisch et al. in Tinner and Ammann, 2001); and (f) Alpine synthesis (shaded bands represent major maxima ± 50 or ± 100 years). Adapted from Matthews JA and Dresser PQ (2008) Holocene glacier variation chronology of the Smørstabbtindan massif, Jotunheimen, southern Norway, and the recognition of century- to millennial-scale European Neoglacial Events. *The Holocene* 18: 181–201.

Models of Neoglaciation and Geographical Variation in Glacier Behavior

The evidence suggests that the two current models of neoglaciation extant in Europe are both oversimplifications of reality. First, the conventional model of a distinct neoglacial period after a long, glacier-free mid-Holocene is only applicable at the broadest level of generalization. Second, the more recent model of a largely undifferentiated Holocene punctuated by numerous abrupt glacier expansion episodes of similar extent is equally difficult to verify when the evidence is scrutinized in detail. Alternatively, elements of both models seem to vary in their applicability to particular regional or subregional records, and particular glaciers buck the trend.

Throughout Europe, therefore, the spatial and temporal patterns of Holocene glacier variations require a new, hybrid model of neoglaciation to supercede both current models. Such a model, in which long- and short-term variations are recognized, the magnitude and frequency of century- to millennial-scale glacier variations vary through time, and regional and subregional differences can be accommodated, is suggested by Figures 7 and 8. In general, the early Holocene appears to have been characterized by relatively large glaciers and infrequent, high-magnitude neoglacial events. Neoglacial events remained infrequent during the mid-Holocene when neoglacial events were relatively few in number and glaciers tended to be small and melted away for some of the time. Following the onset of neoglaciation, the area glacierized

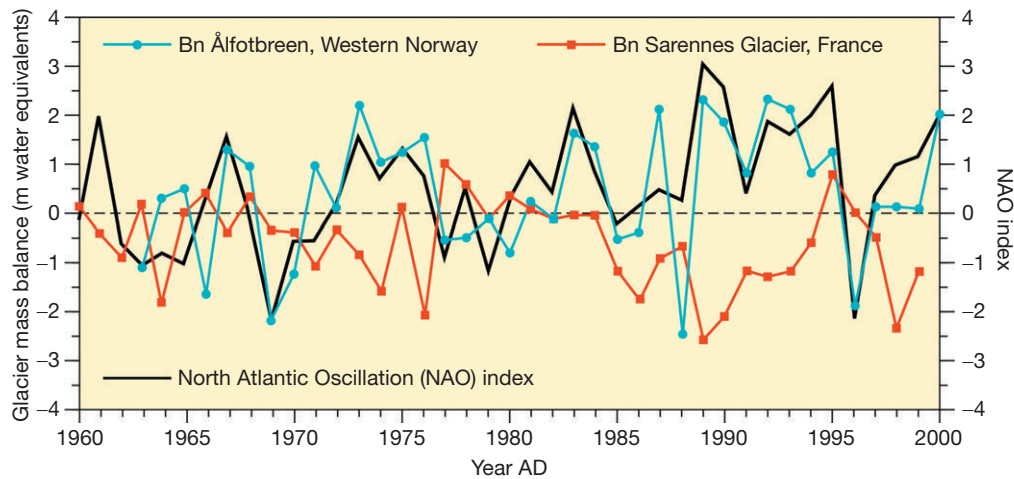


Figure 9 Variations in the annual net mass balance of Ålfotbreen, southern Norway (open circles), and the Sarennes Glacier, southeastern France, Western Alps, related to the North Atlantic Oscillation Index (thick line), AD 1960–2000. Bn, Net balance. Adapted from Nesje A and Dahl SO (2003) The ‘Little Ice Age’ – Only temperature? *The Holocene* 13: 139–145.

increased considerably and the frequency of neoglaciation events increased, culminating in the Little Ice Age.

Apparent differences between glacier behavior within and between Scandinavia and the Alps can be related to glacier size, geometry, and location in relation to atmospheric circulation patterns. According to the records from southern Norway, the shortest glacier-free interval (~7000–6000 cal years BP) is associated with the largest ice body, Jostedalsgreen (Figure 7(b)), while the relatively small glaciers of Smørstabbtindan, Jotunheimen, appear more sensitive to the abrupt ~8200 cal years BP event. This is likely to reflect the rapid response of small glaciers to short-term climatic variations. Another factor to be taken into account is the differential sensitivity of glacier variations to seasonal aspects of climate, especially winter precipitation and summer temperature. In southern Norway, the importance of winter precipitation (relative to summer temperature) in explaining recent frontal variations decreases with continentality from west to east. This is because the winter balance dominates the annual net balance on the maritime (western) glaciers. Differences in the pattern and timing of neoglaciation events between northern and southern Scandinavia are likely to reflect shifts in the position and strength of the North Atlantic westerlies and other changes in the general circulation of the atmosphere, including the frequency of blocking anticyclones.

Atmospheric circulation patterns also help to explain some of the differences in glacier behavior between Scandinavia and the Alps. This is illustrated by the relationship between the annual net balance of Ålfotbreen (southwestern Norway), the Sarennes Glacier (southwestern Alps), and the North Atlantic Oscillation (NAO) Index (Figure 9). The antiphase relationship between the net mass balance of the two glaciers contrasts with the positive correlation ($r = +0.77$) between the net balance of the Norwegian glacier and the NAO Index. This correlation is consistent with the dominant effect of the winter balance on this maritime glacier and reflects the association between positive-NAO-Index winters and above-normal precipitation in

Iceland and adjacent northwestern Europe (while below-normal precipitation is received throughout southern Europe and the Mediterranean).

European Neoglaciation Events and Climatic Forcing Factors

The extent to which it is possible currently to recognize a Europe-wide pattern of neoglaciation events, and their climatic implications, can be assessed with reference to Figure 10. Here, 13 continental scale, European Neoglaciation Events (Figure 10(c)) are identified as the areas of overlap between the previously defined Southern Norwegian (Figure 10(b)) and Alpine (Figure 10(a)) regional events. It can therefore be argued that, notwithstanding the existence of regional and sub-regional differences, a majority of the regional events extend throughout Europe.

One of the several potential climatic forcing factors that may contribute to synchronous European Neoglaciation Events is low solar irradiance, represented in Figure 10(d) by periods of above average residual $\Delta^{14}\text{C}$. These periods correlate strongly with above average drift-ice rafting in the North Atlantic, derived from stacked ocean cores (Figure 10(e)), and less strongly with Central European cold-humid phases defined on the basis of terrestrial paleoecological indicators of timberline depression. It is perhaps not surprising, given the dependence of glacier variations on both winter precipitation and summer temperatures, that there is only partial agreement between continental-scale neoglaciation events and low solar irradiance. Several recent studies in the Alps have, however, suggested stronger correlations at the regional scale.

The implication of these data is that variation in solar irradiance, which may be periodic in character, is an important driver of century- to millennial-scale glacier and climatic variability in Europe. At least three other forcing factors are probably involved in explaining either the general multimillennial pattern or the timing of particular neoglaciation events. These are,

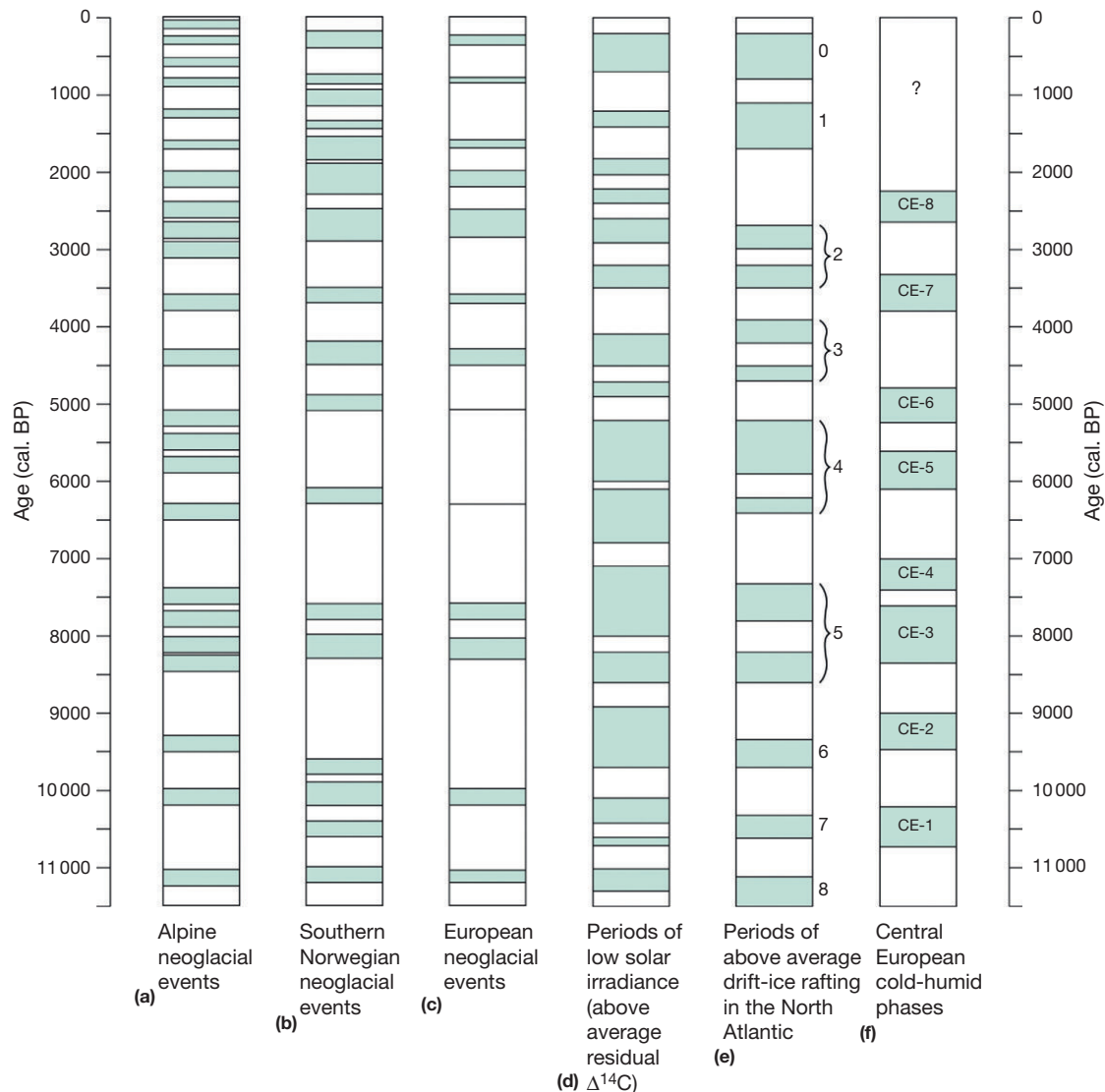


Figure 10 European Neoglacial Events, regional neoglacial events and related records. (a) regional Alpine Neoglacial Events (from Figure 8(f)); (b) regional Southern Norwegian Neoglacial Events (from Figure 7(e)); (c) 16 European Neoglacial Events defined as periods of overlap between the regional events in (a) and (b); (d) periods of above average residual $\Delta^{14}\text{C}$ and hence low solar irradiance (Bond et al., 2001); (e) periods of above average drift-ice rafting in the North Atlantic (Bond et al., 2001) using Bond's numbering system for nine major Holocene ice-rafting events; and (f) cold-humid phases in Central Europe (CE-1 to CE-8) based on tree pollen and plant macrofossils (Haas et al., 1998). Adapted from Matthews JA and Dresser PQ (2008) Holocene glacier variation chronology of the Smørstabbtindan massif, Jotunheimen, southern Norway, and the recognition of century- to millennial-scale European Neoglacial Events. *The Holocene* 18: 181–201.

first, the well-known reduction in solar radiation of around 8% that has affected the Northern Hemisphere during the Holocene as a result of changes in the Earth's orbital parameters (Milankovitch forcing). This may account for much of the general increase in the magnitude and frequency of neoglacial events toward the present and also for a conventional mid-to-late Holocene neoglaciation where this occurs. Second, volcanic forcing may make an effective contribution to both the general trend and to the timing of particular neoglacial events. Temporal clustering of large volcanic eruptions that introduce significant quantities of volcanic aerosols into the upper atmosphere provides the potential mechanism in this case. The third forcing factor is ocean forcing by freshwater outburst events

into the North Atlantic and Arctic Oceans from the Laurentide Ice Sheet and the Baltic Ice Lake with concomitant effects on European climate. This factor was a potential influence on neoglacial events only prior to about 8000 cal years BP.

Conclusion

- (i) Historical and documentary evidence, moraine-ridge stratigraphy, dendroglaciology, archaeoglaciology, and evidence from distal glaciolacustrine and glaciofluvial sediments, have been combined to reconstruct and compare

detailed, continuous records of Holocene glacier variations from both northern and southern Europe.

- (ii) Records from southern Norway suggest a conventional onset of neoglaciation at 6500–6000 cal years BP. Before this, Jostedalbreen, the largest ice cap on mainland Europe, melted away. Subsequently, southern Norwegian glaciers reached their late Holocene maximum extent in the 'Little Ice Age.' In the Alps, many glaciers advanced to similar positions during expansion episodes at various times during the Holocene. Consequently, it is more difficult to define a conventional onset to neoglaciation in the Alps.
- (iii) Throughout Europe, a large number of neogacial events, defined as century- to millennial-scale glacier expansion episodes separated by phases of glacier recession, have been recognized. At least 13 European Neogacial Events, defined as century- to millennial-scale events that occurred in both southern Norway and the Alps, can be identified. A hybrid model of neoglaciation is therefore supported in which the magnitude and frequency of European Neogacial Events tend to increase through the late Holocene, but differences exist between glaciers at regional and sub-regional scales, which may be accounted for largely in terms of atmospheric circulation patterns.
- (iv) It is likely that the pattern and timing of Holocene glacier variations in Europe will eventually be explained by a combination of four climatic forcing factors. Milankovitch forcing potentially explains residual glaciers in the early Holocene following deglaciation, mid-Holocene glacier demise, conventional neoglaciation, and the late Holocene buildup of glaciers. Solar forcing appears to account for the tendency toward periodic neogacial events. These gradual and regular changes seem to have been moderated by volcanic forcing from injections of volcanic aerosols into the atmosphere, and by ocean forcing (before about 8000 cal years BP) triggered by freshwater outbursts into the North Atlantic and Arctic Oceans.

See also: Dendrochronology; Lichenometry. **Cosmogenic Nuclide Dating:** Exposure Geochronology; Methods; Overview. **Glaciations:** Neoglaciation in the American Cordilleras. **Radiocarbon Dating:** AMS Radiocarbon Dating; Conventional Method; Sources of Error.

References and Further Reading

Bakke J, Dahl SO, Paasche Ø, Løvlie R, and Nesje A (2005a) Utilising physical sediment variability in glacier-fed lakes for continuous glacier reconstruction during the Holocene, northern Folgefonna, western Norway. *The Holocene* 15: 161–176.

Bakke J, Dahl SO, and Nesje A (2005b) Late glacial and early Holocene paleoclimatic reconstruction based on glacier fluctuations and equilibrium-line altitudes at northern Folgefonna, Hardanger, western Norway. *Journal of Quaternary Science* 20: 179–198.

Bakke J, Lie Ø, Nesje A, Dahl SO, and Paasche Ø (2005c) Glacier fluctuations, equilibrium-line altitudes and paleoclimate in Lyngen, northern Norway, during the Late glacial and Holocene. *The Holocene* 15: 387–409.

Bakke J, Lie Ø, Dahl SO, Nesje A, and Bjune AE (2008) Strength and spatial patterns of the Holocene wintertime westerlies in the NE Atlantic region. *Global and Planetary Change* 60: 28–41.

Baroni C and Orombelli G (1996) The Alpine 'Iceman' and Holocene climatic change. *Quaternary Research* 34: 78–83.

Bond G, et al. (2001) Persistent solar influence on North Atlantic climate during the Holocene. *Science* 294: 2130–2136.

Dahl SO and Nesje A (1996) A new approach to calculating Holocene winter precipitation by combining glacier equilibrium-line altitudes and pine-tree limits: A case study from Hardangerjøkulen, central-southern Norway. *The Holocene* 6: 381–398.

Dahl SO, Bakke J, Lie Ø, and Nesje A (2003) Reconstruction of former equilibrium-line altitudes based on proglacial sites: An evaluation of approaches and selection of sites. *Quaternary Science Reviews* 22: 275–287.

Deline P and Orombelli G (2005) Glacier fluctuations in the western Alps during the Neoglaciation, as indicated by the Miage moraine amphitheatre (Mont Blanc massif, Italy). *Boreas* 34: 456–467.

Dugmore AJ (1989) Tephrochronological studies of Holocene glacier variations in Iceland. In: Oerlemans J (ed.) *Glacier Fluctuations and Climatic Change*, pp. 37–55. Amsterdam: Kluwer.

Frenzel B, Boulton GS, Gläser B, and Huchriede U (1997) *Glacier Fluctuations During the Holocene*. Stuttgart: Fischer.

Grosjean M, Suter PJ, Trachsel M, and Wanner H (2007) Ice-borne prehistoric finds in the Swiss Alps reflect Holocene glacier fluctuations. *Journal of Quaternary Science* 22: 203–207.

Grove J (2004) *Little Ice Ages: Ancient and Modern*. Oxford: Routledge 2 vols, 718 pp.

Haas JN, Richoz I, Tinner W, and Wick I (1998) Synchronous Holocene climatic oscillations recorded on the Swiss Plateau and at timberline in the Alps. *The Holocene* 8: 301–309.

Holzhauser H, Magny M, and Zumbühl HJ (2005) Glacier and lake-level variations in west-central Europe over the last 3500 years. *The Holocene* 15: 791–803.

Hormes H, Müller BH, and Schlüchter C (2001) The Alps with little ice: Evidence for eight Holocene phases of reduced glacier extent in the Central Swiss Alps. *The Holocene* 11: 255–265.

Hormes H, Beer J, and Schlüchter C (2006) A geochronological approach to understanding the role of solar activity on Holocene glacier length variability in the Swiss Alps. *Geografiska Annaler* 88A: 281–294.

Ivy-Ochs S, Kerschner M, Maisch M, Christl M, Kubik PW, and Schlüchter C (2009) Latest Pleistocene and Holocene glacier variations in the European Alps. *Quaternary Science Reviews* 28: 2137–2149.

Joerin UE, Stocker TF, and Schlüchter C (2006) Multi-century glacier fluctuations in the Swiss Alps during the Holocene. *The Holocene* 16: 697–704.

Kerschner H and Ivy-Ochs S (2009) Palaeoclimate from glaciers: Examples from the Eastern Alps during the Alpine Lateglacial and early Holocene. *Global and Planetary Change* 60: 58–71.

Kerschner H, Hertl A, Gross G, Ivy-Ochs S, and Kubik PW (2006) Surface exposure dating of moraines in the Kromer valley (Silvretta Mountains, Austria): Evidence for glacial response to the 8.2 ka event in the Eastern Alps. *The Holocene* 16: 7–15.

Leemann A and Niessen F (1994) Holocene glacial activity and climatic variations in the Swiss Alps: Reconstructing a continuous record from proglacial lake sediments. *The Holocene* 4: 259–268.

Matthews JA (2005) 'Little Ice Age' glacier variations in Jotunheimen, southern Norway: A study in regionally controlled lichenometric dating of recessional moraines with implications for climate and lichen growth rates. *The Holocene* 15: 1–19.

Matthews JA and Briffa KR (2005) The 'Little Ice Age': A re-evaluation of an evolving concept. *Geografiska Annaler* 87A: 17–36.

Matthews JA and Dresser PQ (2008) Holocene glacier variation chronology of the Smørstabbtindan massif, Jotunheimen, southern Norway, and the recognition of century- to millennial-scale European Neoglaciation Events. *The Holocene* 18: 181–201.

Matthews JA and Winkler S (2011) Schmidt-hammer exposure-age dating (SHD): Application to early Holocene moraines and a reappraisal of the reliability of terrestrial cosmogenic-nuclide dating (TCND) at Austanbotnbreen, Jotunheimen, Norway. *Boreas* 40: 256–270.

Matthews JA, Dahl SO, Nesje A, Berrisford MS, and Andersson C (2000) Holocene glacier variations in central Jotunheimen, southern Norway based on distal glaciolacustrine sediment cores. *Quaternary Science Reviews* 19: 1625–1647.

Matthews JA, et al. (2005) Holocene glacier history of Bjørnbreen and climatic reconstruction in central Jotunheimen, Norway, based on proximal glaciolacustrine stream-bank mires. *Quaternary Science Reviews* 24: 67–90.

Matthews JA, Shakesby RA, Schnabel C, and Freeman S (2008) Cosmogenic ¹⁰Be and ²⁶Al ages of Holocene moraines in southern Norway I: Testing the method and confirmation of the date of the Erdalen Event (c.10 ka) at its type-site. *The Holocene* 18: 1155–1164.

Nesje A (2009) Latest Pleistocene and Holocene alpine glacier fluctuations in Scandinavia. *Quaternary Science Reviews* 28: 2119–2136.

- Nesje A and Dahl SO (2003) The 'Little Ice Age' – Only temperature? *The Holocene* 13: 139–145.
- Nesje A and Johannessen T (1992) What were the primary forcing mechanisms of high-frequency Holocene climate and glacier variations? *The Holocene* 2: 79–84.
- Nesje A, Dahl SO, Andersson C, and Matthews JA (2000) The lacustrine sedimentary sequence in Syneskdvatnet, western Norway: A continuous, high-resolution record of the Jostedalsgreen ice cap during the Holocene. *Quaternary Science Reviews* 19: 1047–1065.
- Nesje A, Matthews JA, Dahl SO, Berrisford MS, and Andersson C (2001) Holocene glacier fluctuations of Flatebreen and winter-precipitation changes in the Jostedalsgreen region, western Norway, based on glaciolacustrine sediment records. *The Holocene* 11: 267–280.
- Nesje A, Dahl SO, and Bakke J (2004) Were abrupt Late glacial and early Holocene climatic changes in northwest Europe linked to freshwater outbursts to the North Atlantic and Arctic Oceans? *The Holocene* 14: 299–310.
- Nesje A, Bakke J, Dahl SO, Lie Ø, and Matthews JA (2008) Norwegian mountain glaciers in the past, present and future. *Global and Planetary Change* 60: 10–27.
- Nicolussi K and Patzelt G (2000a) Discovery of early Holocene wood and peat on the foreland of Pasterze Glacier, Eastern Alps, Austria. *The Holocene* 10: 191–199.
- Nicolussi K and Patzelt G (2000b) Untersuchungen zur holozänen Gletscherentwicklung von Pasterze und Gepatschferner (Ostalpen). *Zeitschrift für Gletscherkunde und Glazialgeologie* 36: 1–87.
- Patzelt G and Bortenschlager S (1973) Die postglazialen Gletscher- und Klimaschwankungen in der Venedigergruppe (Hohe Tauern, Ostalpen). *Zeitschrift für Geomorphologie NF, Supplement-Band* 16: 25–72.
- Shakesby RA, Matthews JA, and Schnabel C (2008) Cosmogenic ^{10}Be and ^{26}Al ages of Holocene moraines in southern Norway I: Evidence for individualistic responses of high-altitude glaciers to millennial-scale climatic fluctuations. *The Holocene* 18: 1165–1177.
- Six D, Reynaud L, and Letréguilly A (2001) Bilans de masse des glaciers alpins et scandinaves, leur relations avec d'oscillation du climat de l'Atlantique nord. *Sciences de la Terre et des Planètes* 333: 693–698.
- Tinner W and Ammann B (2001) Timberline paleoecology in the Alps. *PAGES News* 9: 9–11.
- Wanner H, Holzhauser H, Pfister C, and Zumbühl H (2000) Interannual to century scale climatic variability in the European Alps. *Erdkunde* 54: 62–69.
- Wanner H, et al. (2008) Mid- to late- Holocene climate change: An overview. *Quaternary Science Reviews* 27: 1791–1828.
- Winkler S and Matthews JA (2010) Holocene glacier chronologies: Are 'high-resolution' global and inter-hemispheric comparisons possible? *The Holocene* 20: 1137–1147.
- Zumbühl HJ (1980) *Die Schwankungen der Grindelwaldgletscher in den Historischen Bild- und Schriftquellen des 12. bis 19. Jahrhunderts*. Basel: Birkhäuser.

Neoglaciation in the American Cordilleras

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Concept of Neoglaciation

By the middle of the past century, glacial geologists working in the high mountains of western North America had recognized that moraines near the termini of glaciers, far upvalley from those of the last (Wisconsin) glaciation, represented small-scale fluctuations of postglacial climate. In the 1930s and 1940s, Francois Matthes of the US Geological Survey published reports concerning his studies in the Sierra Nevada of California and proposed that Sierran glaciers had disappeared at the end of the Wisconsin glaciation during a time of mild climate, but then were reborn and reached their greatest size after the end of the Pleistocene (Matthes, 1942) (Figures 1 and 2). Previously, it had been widely assumed that modern glaciers were merely shrunken remnants of former large valley glaciers and ice caps and that fresh-looking moraines beyond modern glacier termini were recessional drift bodies left behind as the remaining ice wasted away. Matthes recognized that the large coastal glaciers of Alaska and Canada, as well as large valley glaciers on high Cascade volcanoes, likely persisted throughout the mild postglacial interval, but he inferred that smaller cirque and valley glaciers in the Cascades, Rocky Mountains, and Sierra Nevada disappeared as the climate warmed in postglacial time. He supported his inference by calling attention to a study of Owens Lake, at the eastern foot of the Sierra. In this study, it was estimated that 4000 years were required for the salt content of the lake to reach its modern concentration following the lake's desiccation during a mild mid-postglacial interval and its subsequent regeneration as the climate became cooler and wetter. Matthes postulated that the rebirth of Owens Lake and the nearby Sierran glaciers were responses to the same pattern of postglacial climate change and that the modern Sierran glaciers, therefore, were, in all probability, somewhat less than 4000 years old.

Recognition of a 'Little Ice Age'

Matthes recognized two sets of young moraines fronting the Sierran glaciers. Those adjacent to the glaciers were unstable and typically ice cored, whereas those beyond were smaller, concentric, and stable. He surmised that the former were comparable to moraines in the Alps that formed in the nineteenth century, whereas the latter were several centuries old. Matthes concluded that glacier oscillations of the past few centuries had been among the greatest during the past 4000 years of renewed glacier activity and marked the culmination of a 'Little Ice Age,' a name that had been proposed by a 'clever journalist.'

In the six decades since Matthes published his Sierran studies, many additional glacial-geological investigations have clarified the sequence and chronology of Holocene glacier variations, both in the Sierra and throughout the world. The concept of a Little Ice Age is now widely accepted and the term has been formalized by capitalizing it. Nevertheless, its duration

and detailed character remain subjects of continuing debate. Although it is widely agreed that this interval was confined to the past millennium, some place its beginning in the seventeenth century, whereas others place it as early as the thirteenth century. Furthermore, accumulating evidence has shown that this interval did not always include the culminating advance or advances of the past 4000 years.

The Hypsithermal Interval

By the mid-nineteenth century, European pollen analysts had identified a series of pollen zones spanning Holocene time, which they designated zones IV–IX. Deevey and Flint (1957) recognized an interval (zones V–VIII) of generally mild climate within this series and named it the Hypsithermal interval. The Hypsithermal followed the cool climate of the earliest Holocene time (pollen zone IV = Preboreal) and preceded an interval of cool late Holocene climate (pollen zone IX = Subatlantic). The ages of the zonal boundaries are based on radiocarbon dating (Mangerud, 1982).

Definition of Neoglaciation

A middle to late Holocene interval marked by repeated episodes of glacier advance has been given various names (e.g., Hypothermal, Katathermal, Medithermal, and Neoglaciation). As a result of its increasing use in the North American Cordillera, Porter and Denton (1967) proposed to define Neoglaciation as 'the climatic episode characterized by rebirth and/or growth of glaciers following maximum shrinkage during the Hypsithermal interval.' They argued that the maximum shrinkage probably coincided with the maximum Hypsithermal warmth, which, based on evidence available to them, appeared to have occurred between ~8000 and 5000 years ago. Further, they called attention to pollen studies in western North America that implied a regionally nonsynchronous transition from Hypsithermal to post-Hypsithermal climate, implying that this boundary was latitudinally time transgressive rather than isochronous. They attempted to resolve this problem by defining Neoglaciation as a geological-climate unit (a unit that may lack isochronous boundaries) and suggested that evidence of the initial Neoglacial advance in a given area might fall at any time within the past five millennia. In other words, in some areas, the early part of Neoglaciation may overlap the later part of the Hypsithermal interval. The Little Ice Age constituted the youngest, and sometimes the only, recognized Neoglacial episode. At the time their initial paper was written, Porter and Denton documented a pre-Little Ice Age Neoglacial advance approximately 2800 ¹⁴C years ago. Subsequent work suggested a possible additional Neoglacial advance approximately 5000 ¹⁴C years ago (Denton and Karlén, 1973; Denton and Porter, 1970).

Establishing Neoglacial Chronologies

The establishment of a reliable chronology of Neoglaciation depends on accurate dating. Whereas in Europe and Iceland, historical data (e.g., written observations, geographic and topographic maps, paintings, sketches, lithographs, and photographs) that record the position of accessible glacier margins span hundreds of years, in the Western Hemisphere such records seldom extend back more than a century (Porter, 1981). For many inaccessible glaciers, historical records began only in



Figure 1 Francois Matthes (US Geological Survey) was the first to propose the concept of a Little Ice Age.

the era of aerial photography and satellite imaging. Therefore, geologists have had to rely substantially on several physical and biological dating methods to develop chronologies, and these have varying degrees of reliability.

Radiocarbon dates commonly provide only maximum or minimum limiting ages for alpine moraines and glacial drift. However, a stacked succession of moraines may include buried soil, peat, or wood that provides bracketing ages for successive moraines. Chronologies can also be developed by dating organic matter in lakes and bogs lying on, behind, or beyond a moraine. The temporal variation in atmospheric radiocarbon means that for Holocene radiocarbon dates to be useful for correlation they should be calibrated using standard calibration curves (Stuiver et al., 1998). Radiocarbon ages in this article are expressed in ^{14}C years BP and calibrated ages in ^{14}C cal years BP.

Dendrochronology has been used to date late Holocene deposits in northwestern North America where glaciers terminated in forested areas. The life span of most forest trees is generally less than a millennium, so tree-ring dating based on the oldest tree is mainly applicable for dating middle to late Little Ice Age moraines. Minimum ages for advances have been obtained by coring the oldest trees growing on moraines. Sources of potential error include uncertainty as to whether the oldest tree has been found, the time between moraine construction and germination of the first seedlings, and the time for a seedling to grow to the height at which the tree was cored. Although these factors may lead to underestimating the age of a morainal substrate, in most cases the error range probably is no more than a decade or two. In some circumstances, the culmination of a glacier advance may be determined by coring trees at the ice margin that were tilted but not killed and noting a marked change in the pattern of ring growth due to tree disturbance.

Lichenometry has been the most useful dating method in the Brooks Range and the Canadian Arctic. Because of the slow



Figure 2 Cirque moraines, like those pictured here near Mt. Whitney in the southern Sierra Nevada, led Matthes to envision a rebirth of glaciers in the middle Holocene following their disappearance at the end of the last glaciation.

growth rate of the ubiquitous species *Rhizocarpon geographicum* at high latitudes, the method is applicable to most of the Neoglacial records (albeit with substantial error ranges at the older end of the timescale). In lower latitude mountains (e.g., the Rockies and Cascades), growth rates are more rapid, permitting high-resolution dating for the past century or two, but the effective range of lichenometry is generally placed within the Little Ice Age. The method has proved less useful in South America, where long radiocarbon-controlled growth curves are difficult to construct.

Tephrochronology has been used in dating Neoglacial moraines on and downwind from active volcanoes (e.g., in southern Alaska, the Cascade Range, and the Andes). Holocene tephra layers have been dated by means of bracketing radiocarbon dates, and their presence or absence on and/or beneath moraines permits limiting ages to be assigned to glacier advances. In Mt. Rainier National Park, for example, Holocene tephras from eruptions of Mt. Rainier, Mt. St. Helens, and Mt. Mazama provide bracketing ages for Neoglacial moraines (Figure 3).

Cosmogenic isotopes (e.g., ^3He , ^{10}Be , ^{26}Al , and ^{36}Cl) are being widely used to obtain surface-exposure ages of alpine moraine boulders, especially those of the last glaciation. Although cosmogenic isotopes can also be used to date Neoglacial deposits, few such dates have been reported. A study on Mt. Rainier in the Cascade Range, however, points to a potential problem. A large Holocene volcanic mudflow deposit at ~ 1800 m altitude is ~ 5000 years old, according to ^{14}C dating of incorporated wood. However, large boulders at the surface have ^{36}Cl surface-exposure ages half as old. This discrepancy is apparently due to a thick and persistent snow cover during successive Neoglacial events that shielded the mudflow boulders and adjacent Neoglacial moraines from cosmic radiation

(Mt. Rainier has had post-Little Ice Age annual snowfall totals that exceed 30 m). Thus, only moraines minimally affected by Neoglacial snow cover are likely to provide reliable exposure ages.

Neoglacial Chronology of North and South America

Most chronologies of Neoglaciation in American mountain ranges focus on selected areas, ranging from Alaska's Brooks Range to the southern Andes. Only the most important are reviewed here. In low latitudes, where the snow line is very high, modern glaciers and their Neoglacial predecessors terminated on land and were confined to high cirques and valley heads, mostly well above 4000 m. In middle latitudes, small valley and cirque glaciers lay at altitudes of 3000–4000 m, with the largest glaciers typically heading on high stratovolcanoes. At higher southern latitudes, many outlet glaciers of the large North and South Patagonian ice caps terminated in lake basins or in tidewater and experienced advance/retreat cycles like those of tidewater glaciers in southcoastal Alaska.

Brooks Range

Numerous cirque glaciers and small valley glaciers are located in the highest sectors of the Brooks Range. All lie above or beyond the treeline in a landscape of herbaceous tundra vegetation. Although radiocarbon dates have been obtained for some middle to late Neoglacial moraines, the glacial chronology is based mainly on lichenometric data obtained near the margin of 97 glaciers, mostly in the central part of the range (Calkin, 1988). A lichen growth curve that is controlled by radiocarbon dates back to 1300 years BP was extrapolated

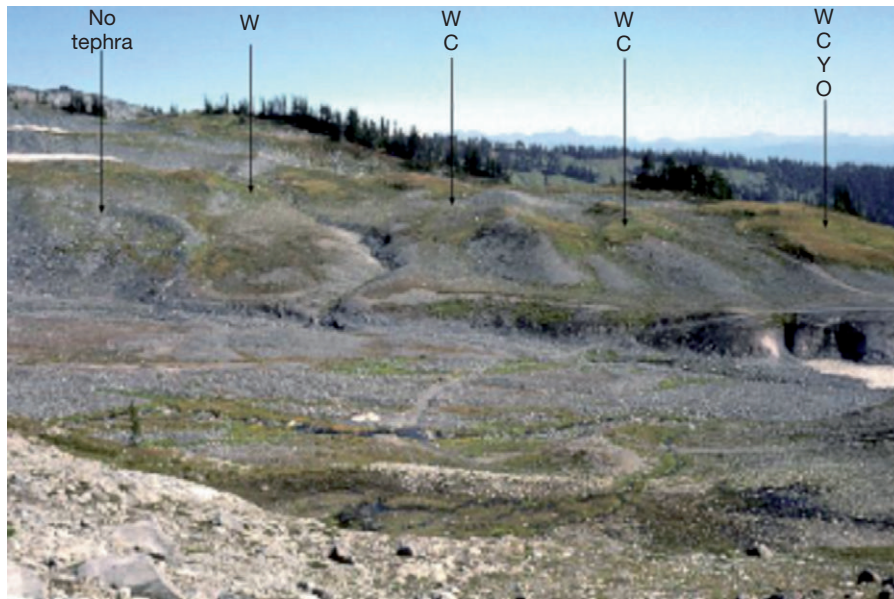


Figure 3 Neoglacial moraines of Paradise Glacier (out of view to left) on Mt. Rainier volcano in the Cascade Range that are dated by tephrochronology. Tephra layers W (450 years), C (2200 years), Y (3400 years), and O (Mazama, 6800 years) are used to bracket the age of moraines. A grassy slope beyond the moraines is mantled by the entire suite of layers, demonstrating that it has not been covered by glacier ice since at least 6800 years ago. The two outer moraines are less than 3400 years old and more than 2200 years old, whereas the next moraine is less than 2200 and more than 450 years old. The innermost (late Little Ice Age) moraine is younger than 450 years.

linearly to 8000 years BP, thus spanning all of Neoglaciation. The estimated error range of the dates is $\pm 20\%$ (i.e., ± 1000 years for the earliest Neoglacial time). Moraines postdating 5000 years BP tend to be closely nested and often lie within less than 100 m of glacier margins (Figure 4). Early Neoglacial moraines (4200–4600 and 3400–3700 years old) have been found in the west-central Brooks Range. Pre-Little Ice Age Neoglacial moraines (~ 2700 –2900, 2200–2500, and 1600–2000 years old), as well as clusters averaging 1400–1600 and 1000–1300 years, are prominent in the central and eastern sectors of the range. Little Ice Age moraines that border nearly every glacier terminus have ages that cluster prominently at 600–900, 300–500, and 90 years BP.

Southern Alaska and Yukon Territory

This most extensively glacierized sector of the Americas includes thousands of small to large land-terminating glaciers, as well as numerous large tidewater-calving glaciers. Local Holocene chronologies are based mainly on radiocarbon and tree-ring dating. In the Kenai Mountains and other ranges that border the Gulf of Alaska, land-terminating glaciers experienced major periods of expansion during the Little Ice Age in the middle thirteenth, early fifteenth, middle seventeenth, and last half of the nineteenth centuries; earlier, pre-Little Ice Age advances are dated between approximately 1550 and 1300 years BP (Calkin et al., 2001; Wiles et al., 1999).

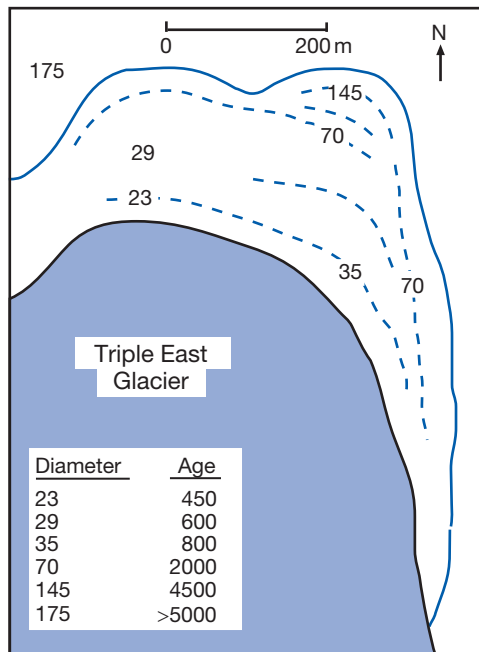


Figure 4 Moraines of a cirque glacier in the Brooks Range that border the ice margin have been dated by lichenometry. Using a growth curve extrapolated beyond 1300 years, the diameter (in mm) of the largest *Rhizocarpon geographicum* growing on each moraine provides an approximate age for a Neoglacial advance. Adapted from Ellis JM and Calkin PE (1984) Chronology of Holocene glaciation, central Brooks Range, Alaska. *Geological Society of America Bulletin* 95: 897–912, Figure 9B.

By contrast, large calving tidewater glaciers in this region experience advance/retreat cycles that differ from glacier to glacier: some glaciers may be advancing while others are retreating or are stabilized. Nevertheless, on the scale of millennia, a gross correspondence can be seen among phases of expansion, which may last nearly 1000 years, and intervals of rapid retreat (Calkin et al., 2001; Figure 5). The record of calving glacier activity has mainly been reconstructed from wood- and peat-bearing deposits exposed by ice recession (Wiles et al., 1999). Rooted tree stumps, sheared off by advancing ice, have been used to develop ring chronologies that help assess the rate of glacier advance and time of glacier maxima. Around the Gulf of Alaska, Neoglacial episodes of calving glacier expansion may have begun as early as 4500 years BP in Glacier Bay and ~ 3600 –3000 years BP and 1300–1400 years BP for other coastal glaciers. A major episode of advance coincided with the Little Ice Age (Figure 5). By the end of the nineteenth century, many of these large calving glaciers were at or near their Little Ice Age maxima, but within several decades they had begun to retreat. By the end of the twentieth century, most had receded to fjord heads and their termini were either stabilized or once again advancing. Columbia Glacier was the last of the major calving glaciers to begin retreating and its terminus is now receding rapidly upfjord.

Neoglacial advances in the northeastern St. Elias Mountains of Yukon Territory culminated ~ 2800 , 2600, and 1230–1050 years BP and were followed by Little Ice Age advances (Denton and Karlén, 1977). The Kaskawulsh Glacier was advancing 500 years BP and subsequently attained its greatest extent of the past ca. 10 000 years.

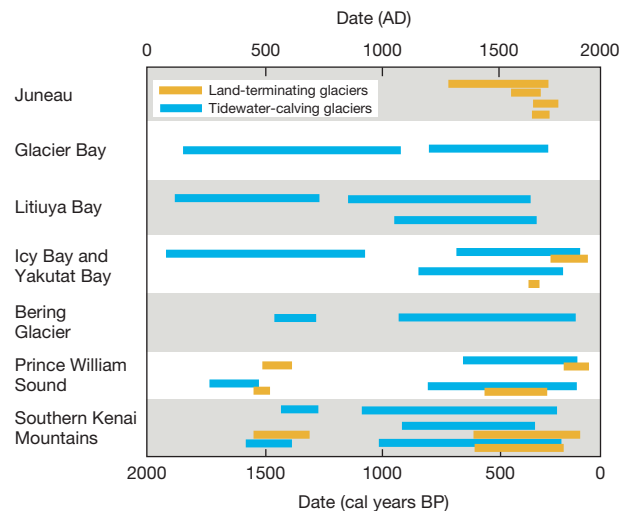


Figure 5 A record showing that intervals of advance for calving and land-terminating glaciers in southcoastal Alaska span the past two millennia. For tidewater glaciers, intervals of advance vary widely but nevertheless group very broadly in two intervals centered on ~ 500 and 1500 years ago (Little Ice Age and pre-Little Ice Age). Land-terminating advances fall mostly in the fifteenth to nineteenth centuries, but three glaciers also advanced close to 1500 years ago. Adapted from Calkin PE, Wiles GC, and Barclay DJ (2001) Holocene coastal glaciation of Alaska. *Quaternary Science Reviews* 20: 449–461, Figure 2.

Coast Mountains of British Columbia

In the southern Coast Mountains of British Columbia, the earliest advance is dated from glacially overridden tree stumps in growth position at Garibaldi Park (~5300 years BP). The apparent absence of moraines of this age implies that this early advance was less extensive than later ones. In the northern Coast Mountains, a middle Neoglacial advance that culminated ~2300 years BP was equal to or more extensive than Little Ice Age advances, whereas an advance ~1900 years BP was as extensive. The Little Ice Age glacier expansion began shortly before 900 years BP and culminated in the eighteenth or nineteenth century when most glacier termini reached their maximum Neoglacial limit (Osborn and Luckman, 1988; Ryder and Thomson, 1986).

Cascade Range

Many of the hundreds of glaciers in the Cascade Range terminate in vegetated terrain where radiocarbon dating of buried wood and soils, lichenometry, and tree-ring dating are feasible methods for dating Neoglacial deposits. Tephrochronology is also applicable, and in a few cases historical data extend back to the middle to late nineteenth century. An early advance (~4700–4900 years BP) is inferred from *in situ* glacier-sheared trees that were exposed by the recent retreat of the South Cascade Glacier (Miller, 1969). No moraines of this advance have been identified beyond the Little Ice Age moraines, implying that this advance was less extensive than those of recent centuries. Ring counts show that a tree growing on a nearby Chickamin Glacier moraine germinated before AD 1280; this moraine may record the onset of the Little Ice Age in the northern Cascades. Most Neoglacial Cascade moraines date to the sixteenth, nineteenth, and twentieth centuries. Tephra layers broadly bracket earlier Neoglacial moraines at Mt. Rainier between 3400 and 2200 years ago and 2200 and 450 years ago (Crandell and Miller, 1965; Figure 3). At Nisqually Glacier, a 3400-year-old tephra is found immediately adjacent to the outermost Little Ice Age moraine, which marks the most extensive Neoglacial advance (ca. AD 1825).

Canadian and US Rocky Mountains

No direct evidence of Neoglacial glacier advances older than 4500 years BP has been reported from the Canadian Rockies. Evidence of advances from 4400 to 3000 and from 2300 to 1800 years BP has been found at Bugaboo Glacier in the Purcell Mountains (Osborn and Luckman, 1988). Little Ice Age advances, dated by tree-ring records, began soon after 900 years BP and culminated mainly in the eighteenth or nineteenth century (Luckman, 1993); for most investigated glaciers, these were the most extensive Holocene advances in the Canadian Rockies.

Because of a paucity of radiocarbon dates, the Neoglacial chronology of the US Rocky Mountains relies mainly on lichenometry and an array of relative dating methods (e.g., soils and relative rock-weathering rates) that are useful in distinguishing presumed Neoglacial cirque moraines and rock glaciers from early Holocene or late-glacial deposits (Burke and Birkeland, 1983). The youngest deposits are assigned to the Little Ice Age.

Rock glacier lobes in the central Colorado Rockies dated by lichenometry have ages that cluster at ~3100, 2100, and 1200 cal years ago. In the San Juan Mountains of the southern Rockies, where Neoglacial moraines are lacking, many cirques have been ice free for at least 14 000 years.

Sierra Nevada

Although the Sierra Nevada might well be considered the 'classic' area of the Little Ice Age and Neoglaciation, its geological record of Holocene glaciation is sparse, not yet well studied, and rather poorly dated (Clark and Gillespie, 1997). The known moraine record is confined mainly to the Little Ice Age, although cores from the Conness Lakes taken downvalley from cirque moraines suggest that Neoglacial glacier activity may have begun by ~3200 years BP (Konrad and Clark, 1998; i.e., remarkably close to Matthes' original estimate). Moraines of the Little Ice Age are, appropriately, referred to as Matthes-age moraines; on the basis of lichenometry, radiocarbon dating, and tephrochronology, they date to the past ~700 years. Lichen ages of ~970 and 1100 years BP for moraines in the east-central Sierra were reported more than three decades ago, but they have not been confirmed or replicated elsewhere in the range.

Northern and Central Andes

Moraines built during the early and middle phases of Neoglaciation in the northern Andes of Colombia and Ecuador are yet to be identified and dated. In Colombia, Little Ice Age advances are inferred to have overridden and buried any older Holocene moraines. The available evidence suggests that the maximum Neoglacial advances occurred in the sixteenth and seventeenth centuries.

Numerous studies of Neoglacial deposits in the lofty tropical to subtropical Peruvian Andes suggest a somewhat abbreviated record of ice advances (Clapperton, 1993). Some studies are contradictory. For example, in the Cordillera Blanca and Cordillera de Raura, glaciers apparently were in advanced positions ~4300 and soon after 2400 years ago, whereas in the Cordillera Vilcanota/Quelcaya area, radiocarbon dates of organic matter between superposed moraines imply no glacier advances between ~10 000 and 3000 years ago. Most dated moraines in this sector of the Andes have only minimum or maximum ages that may not be closely limiting. Ubiquitous Little Ice Age moraines beyond existing glaciers are as old as 600 years.

Southern Andes

Moraines of the southern Andes that predate the Little Ice Age have been dated mainly by radiocarbon analysis of *in situ* or glacier-transported trees in superposed moraine sequences and of basal peat in bogs on moraines or in meltwater channels. Tree-ring chronologies have been developed for some Little Ice Age moraine sequences.

Middle to late Holocene moraines beyond large outlet glaciers of the Northern and Southern Patagonian ice caps fall into several distinct groups based on limiting radiocarbon ages (Clapperton, 1993; Mercer, 1976). An early Neoglacial

advance between 4700 and 4200 years BP was the most extensive for many maritime glaciers, but glaciers to the east were generally less extensive during this early advance phase than during a subsequent middle Neoglacial interval between 2700 and 2000 years ago. Little Ice Age moraines are found at all the glaciers investigated and fall within the range of 900–100 years BP.

A majority of available dates provide only minimum limiting ages, and the assumption commonly made is that these ages are closely limiting. However, some dates give only broad age brackets (e.g., sometime within a 1500-year interval or sometime after ~3300 years ago), making close age assignment and reliable correlations impossible. Moraines deposited by tidewater- or lake-calving glaciers may be out of phase with the regional pattern, as is the case with such glaciers in coastal Alaska.

Onset of Neoglaciation

Evidence bearing on the initial advance of glaciers during Neoglaciation is most abundant in the middle southern latitudes (i.e., southern Andes and New Zealand Alps; Porter, 2000). Limiting and bracketing ages of the earliest Neoglacial moraines in the southern Andes point to advances that culminated ~4600–4400 years ago (5400–4900 cal years ago) (Figure 6). Only in the Brooks Range have cirque moraines approaching this age been reported in North America. The Cascades of Washington and British Columbia and Glacier Bay, Alaska, are the only sites in North America where glacially overridden evidence of such early advances is known. It seems likely that glaciers in other regions of the cordillera north of the southern Andes also experienced early glacier activity at that time, but the evidence likely has been obscured by later, more extensive advances.

Temporal Pattern of Glacier Variations

Dates for Neoglacial advances tend to cluster broadly in three time intervals (Figure 7). The youngest (late Neoglaciation) includes the Little Ice Age (~800–100 years ago), which is prominent in most glacial records along the American cordilleras. An earlier event (pre-Little Ice Age, ~1500–1100 years ago) is represented by glacier advances in the Brooks Range, coastal Alaska and British Columbia, Yukon Territory, and possibly the Cascades and Sierra Nevada. Middle Neoglacial events (~3300–2700 and 2400–2000 years ago) are recorded in the Brooks Range, southcoastal Alaska, Yukon Territory, the Canadian and US Rocky Mountains, and likely in the Cascades and Sierra Nevada. Evidence of early Neoglacial activity (4800–4100 years ago) is found in the Brooks Range, coastal Alaska, Canadian Rockies, Cascades, and central Andes. The disparate nature of most Neoglacial records reflects limited research on and dating of pre-Little Ice Age moraine sequences, as well as the obliteration or burial of earlier deposits during Little Ice Age advances. An exception is the Brooks Range, where cold, slow-moving cirque glaciers produced nested and superposed moraines that occasionally include early Neoglacial deposits.

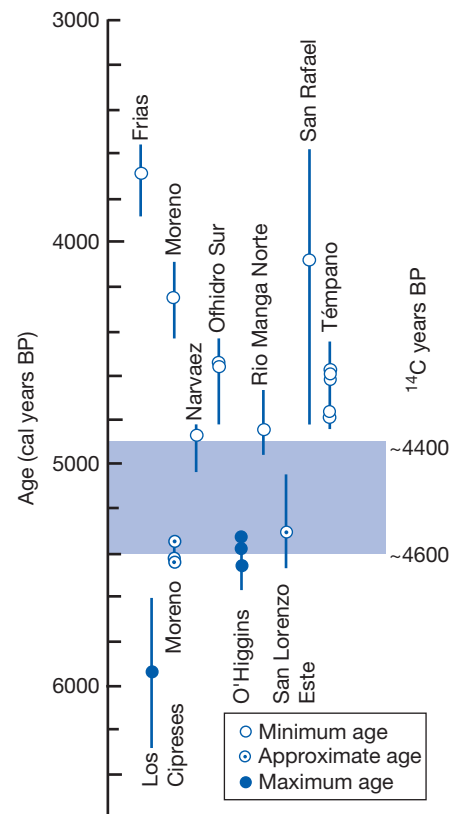


Figure 6 The onset of Neoglaciation in the southern Andes is recorded by maximum, minimum, and approximate ages for advances of 11 glaciers that narrow an initial expansion to the interval of ~4600–4400 years BP (5400–4900 cal years BP). Adapted from Porter SC (2000) Onset of Neoglaciation in the Southern Hemisphere. *Journal of Quaternary Science* 15: 395–408, Figure 5.

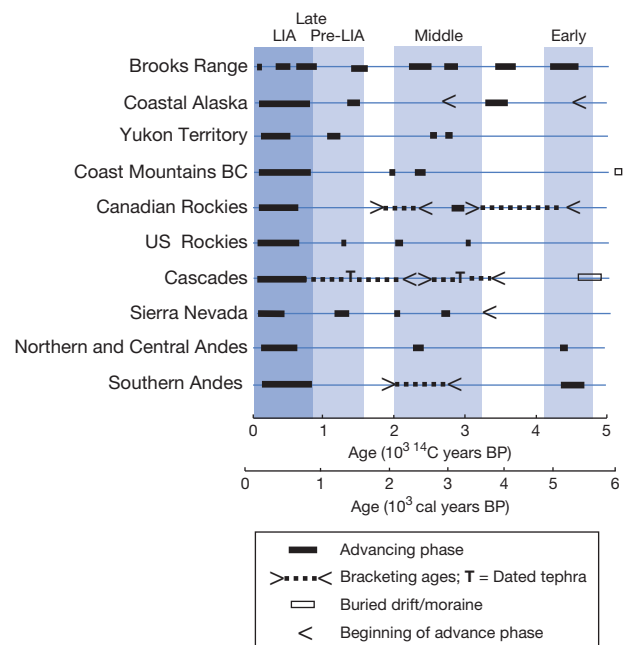


Figure 7 Plot of dates related to Neoglacial events (shaded intervals) recorded along the North and South American cordilleras.

The timing of quasicyclic Neoglacial glacier variations on millennial timescales has been attributed to variations in solar output (Denton and Karlén, 1973). On decadal and shorter scales, however, glaciers may respond to large explosive volcanic eruptions that affect climate and glacier mass balance on a hemispheric or global scale (Porter, 1986). In addition, calving dynamics can affect the advance/retreat pattern of tidewater and lacustrine glacier termini.

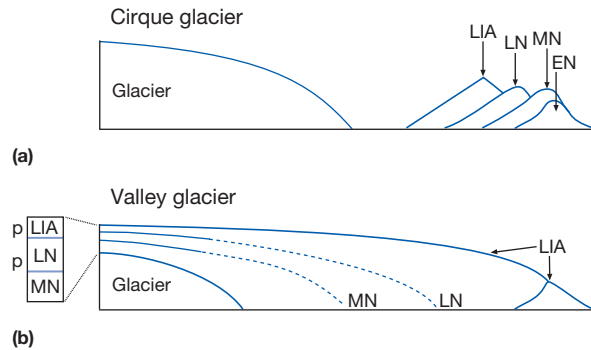


Figure 8 Moraine records of cirque and valley glaciers. (a) Neoglacial moraine succession of a small cirque glacier in which the oldest moraine (EN, early Neoglacial) has acted as a barrier to the glacier during later advances. Advancing ice may run up against older moraine barriers, developing a stacked sequence, and it may override them, burying the older deposits. MN, middle Neoglacial; LN, late Neoglacial; LIA, Little Ice Age. (b) Successively more extensive advances of a valley glacier override earlier moraines (dashed lines) along the valley floor, with the Little Ice Age advance being the greatest of Neoglacial. Evidence of earlier advances is preserved in lateral-moraine stratigraphy in which buried soils (p) separate deposits of successive advances.

Hemispheric Contrasts in Relative Glacier Extent

The pattern of relative glacier extent during Neoglacial episodes varies with latitude along the Americas. In the central and southern Andes, valley glaciers reached their greatest extent during early Neoglacial time, with successive subsequent limits diminishing in extent. In the Cascades, the oldest moraines date to middle Neoglacial, although limited evidence indicates more restricted early Neoglacial activity. Farther north, Little Ice Age moraines represent the most extensive advances. The limited older record in the Brooks Range may reflect occasional preservation near the margin of stacked and closely nested moraine sequences (Figures 4 and 8).

This latitudinal variation in relative glacier extent is likely related to the hemispheric contrast in insolation (incident solar radiation). In southern South America, summer insolation (and summer temperature) was lowest at the onset of Neoglacial, when glaciers reached their maximum limits, but it became progressively higher approaching the Little Ice Age when glaciers were smaller (Figure 9). By contrast, in the Northern Hemisphere, insolation was initially high during early Neoglacial and fell toward the Little Ice Age. The maximum Neoglacial ice extent would be predicted to occur during the Little Ice Age, which is generally the case.

See also: Dating Techniques; Dendrochronology; Lichenometry. Radiocarbon Dating: ^{14}C of Plant Macrofossils; Charcoal.

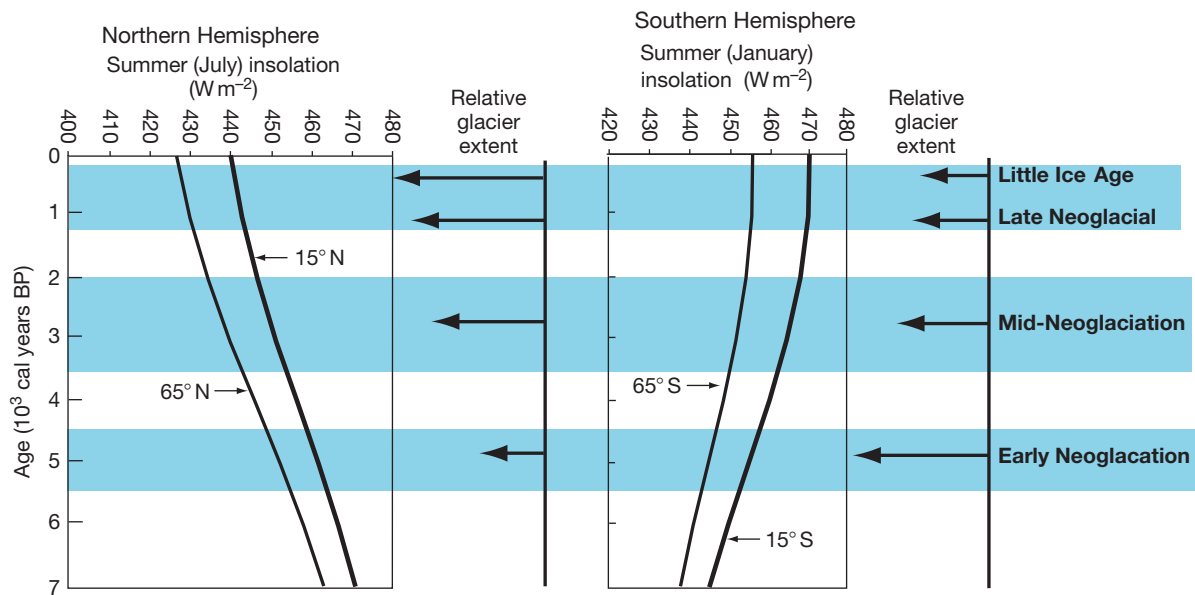


Figure 9 Relative extents of Neoglacial moraines in the North and South American cordilleras compared with the summer insolation received at different latitudes. In northern North America, insolation was lowest and glaciers were most extensive during late Neoglacial time. Early Neoglacial records are rarely found, all but a very few having been overridden by middle and late Neoglacial advances, which were often of nearly equal extent. In southern South America, glaciers were most extensive early in Neoglacial when insolation was lowest, and their limits shrank as insolation increased toward the Little Ice Age.

References

- Burke RM and Birkeland PW (1983) Holocene glaciation in the mountains of western United States. In: Wright HE Jr. (ed.) *The Holocene Late-Quaternary Environments of the United States*, vol. 2, pp. 3–11. Minneapolis: University of Minnesota Press.
- Calkin PE (1988) Holocene glaciation of Alaska (and adjoining Yukon Territory, Canada). *Quaternary Science Reviews* 7: 159–184.
- Calkin PE, Wiles GC, and Barclay DJ (2001) Holocene coastal glaciation of Alaska. *Quaternary Science Reviews* 20: 449–461.
- Clapperton C (1993) *Quaternary Geology and Geomorphology of South America*. New York: Elsevier.
- Clark DH and Gillespie AR (1997) Timing and significance of late-glacial and Holocene glaciation in the Sierra Nevada, California. *Quaternary International* 38/39: 21–38.
- Crandell DR and Miller RD (1965) Post-Hypsithermal Glacier Advances at Mount Rainier, Washington. Menlo Park, CA: U.S. Geological Survey. U.S. Geological Survey Professional Paper 501-D, pp. D110–D114.
- Deevey ES and Flint RF (1957) Postglacial Hypsithermal interval. *Science* 125: 182–184.
- Denton GH and Karlén W (1973) Holocene climatic variations – Their pattern and possible cause. *Quaternary Research* 3: 155–205.
- Denton GH and Karlén W (1977) Holocene glacial and tree-line variations in the White River Valley and Skolai Pass, Alaska and Yukon Territory. *Quaternary Research* 7: 63–111.
- Denton GH and Porter SC (1970) Neoglaciation. *Scientific American* 222: 101–110.
- Konrad SK and Clark DH (1998) Evidence for an early Neoglacial advance from rock glaciers and lake sediments in the Sierra Nevada, California, USA. *Arctic and Alpine Research* 30: 272–284.
- Luckman BH (1993) Glacier and tree-ring records for the last millennium in the Canadian Rockies. *Quaternary Science Reviews* 12: 441–450.
- Mangerud J (1982) The chronostratigraphical subdivision of the Holocene in Norden: A review. *Striae* 16: 65–70.
- Matthes FE (1942) Glaciers. In: Meinzer OE (ed.) *Hydrology*, pp. 149–219. New York: Dover.
- Mercer JH (1976) Glacial history of southernmost South America. *Quaternary Research* 6: 125–166.
- Miller CD (1969) Chronology of Neoglacial moraines in the Dome Peak area, North Cascade Range, Washington. *Arctic and Alpine Research* 1: 49–66.
- Osborn G and Luckman BH (1988) Holocene glacier fluctuations in the Canadian Cordillera (Alberta and British Columbia). *Quaternary Science Reviews* 7: 115–128.
- Porter SC (1981) Glaciological evidence of Holocene climatic change. In: Wigley TML, Ingram MJ, and Farmer G (eds.) *Climate and History*, pp. 82–110. Cambridge: Cambridge University Press.
- Porter SC (1986) Pattern and forcing of Northern Hemisphere glacier variations during the last millennium. *Quaternary Research* 26: 27–48.
- Porter SC (2000) Onset of Neoglaciation in the Southern Hemisphere. *Journal of Quaternary Science* 15: 395–408.
- Porter SC and Denton GH (1967) Chronology of Neoglaciation. *American Journal of Science* 165: 177–210.
- Ryder JM and Thomson B (1986) Neoglaciation in the southern Coast Mountains of British Columbia: Chronology prior to the late Neoglacial maximum. *Canadian Journal of Earth Sciences* 23: 273–287.
- Stuiver M, Reimer PJ, Bard E, et al. (1998) INTCAL 98 radiocarbon calibration, 24,000–0 cal BP. *Radiocarbon* 40: 1041–1083.
- Wiles GC, Barclay DJ, and Calkin PE (1999) Tree-ring-dated ‘Little Ice Age’ histories of maritime glaciers from western Prince William Sound, Alaska. *The Holocene* 9: 163–173.

Global Warming *see* Paleoclimate Relevance to Global Warming

H

Heinrich Events *see* Paleoclimate Reconstruction: Paleodroughts and Society

Historical Climate Records *see* Paleoclimate Reconstruction: Historical Climatology; The Last Millennium

Holocene Environments *see* Dendroclimatology; **Archaeological Records:** Postglacial Adaptations; **Beetle Records:** Postglacial Europe; Postglacial North America; **Chironomid Records:** Postglacial Europe; Postglacial Southern Hemisphere; **Diatom Methods:** Use in Archaeology; **Glaciations:** Neoglaciation in Europe; Neoglaciation in the American Cordilleras; **Ice Cores:** Dynamics of the Greenland Ice Sheet; Dynamics of the West Antarctic Ice Sheet; Dynamics of the East Antarctic Ice Sheet; **Paleoceanography, Records:** Postglacial Indian Ocean; Postglacial North Atlantic; Postglacial North Pacific; Postglacial South Pacific; **Paleoclimate:** Timescales of Climate Change; **Paleoclimate Modeling:** Paleo-ENSO; **Paleoclimate Reconstruction:** Historical Climatology; The Last Millennium; **Plant Macrofossil Methods and Studies:** Treeline Studies; **Plant Macrofossil Records:** Holocene North America; **Pollen Records, Postglacial:** Africa; Australia and New Zealand; Northeastern North America; Northwestern North America; Southeastern North America; Southwestern North America; South America; Northern Asia; Northern Europe; Southern Europe

Human Evolution *see* Archaeological Records: Overview; 2.7 Myr–300 000 Years Ago in Africa; 2.7 Myr–300 000 Years Ago in Asia; 1.9 Myr–300 000 Years Ago in Europe; Human Evolution in the Quaternary

I

ICE CORE METHODS

Contents

Overview

Biological Material

Borehole Temperature Records

Chronologies

CO₂ Studies

Conductivity Studies

Glaciochemistry

Methane Studies
Microparticle and Trace Element Studies
Stable Isotopes
¹⁰Be and Cosmogenic Radionuclides in Ice Cores
Studies of Firm Air

Overview

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Introduction

Cores through the polar ice sheets provide a remarkable record of past environmental conditions, on timescales of thousands of millennia to the last few decades. Ice-core records are unique in paleoclimatology, primarily because of the very high ice accumulation rates (providing excellent time resolution), the preservation of multiple proxies of past environmental conditions, and the fact that small samples of the ancient atmosphere are entombed in bubbles.

Ice coring evolved as a science in the 1950s and 1960s through the pioneering efforts of scientists from several nations and is now conducted by a variety of national and international research teams, in the Arctic, Antarctic, and glaciated mountain regions of the world. Existing ice-core data inform us about human impacts on the polar regions and the global atmosphere as well as the long march of glacial–interglacial climate changes over the last 800 000 years; they also provide spectacular, if only partly understood, data about abrupt climate change during the last Ice Age and the stability of climate over the last 10 000 years. Future ice cores will take the records further back in time and also provide more detailed spatiotemporal information about the evolution of climate and environmental change. Summaries of a number of specific areas of ice-core research are available elsewhere in this encyclopedia.

Ice-Core Basics

Snow accumulates on the polar plateaus of Antarctica and Greenland, and at high elevations in lower latitude mountain regions, in a regularly ordered fashion that preserves a stratigraphic sequence accessible by drilling vertical borings in the ice. Collecting these cores is a specialized engineering challenge. The drilling equipment used for ice coring was developed in the late 1950s and early 1960s, and has since been continuously improved upon. Because ice is a deformable material and flows, ice-coring locations must be chosen carefully to extract the most reliable paleoclimate information. The best drilling sites are generally on or near ice divides ([Figure 1](#)), where ice deposited at or near the same elevation as the current surface elevation can be obtained, and the distortion of the age versus depth curve due to thinning can be reliably modeled.

Ice cores of various lengths are collected for different purposes. Shallow cores (generally <100–200 m) can be collected relatively easily with lightweight equipment and generally cover a few hundred years of history, depending on accumulation rates. Networks of shallow cores have been collected to study spatial variations in climate on these timescales (e.g., the Program for Arctic Regional Climate Assessment (PARCA) and International Trans-Antarctic Scientific Expedition (ITASE) programs in Greenland and Antarctica, respectively). Deeper cores require larger equipment and fluid-filled boreholes to avoid collapse of the hole. Drilling fluids are typically petroleum-derived fluids such as refined kerosene, but other organic liquids (e.g., butyl acetate) have also been used. The physical characteristics of the fluid, particularly freezing point and viscosity (and viscosity–temperature characteristics), are critical parameters and can make the choice of appropriate fluids a challenge. The fluid-filled hole also provides access for measuring ice sheet temperature profiles, which are useful for paleothermometry. The deepest cores (>3000 m) in Greenland and Antarctica have been collected during multiyear campaigns from temporary or permanent scientific camps.

Perhaps the most unique aspect of ice cores is that they trap ancient atmosphere when surface snowfall compacts to ice. This happens at the base of the firm, or unconsolidated snow layer, typically 50–100 m below the surface in polar regions, and shallower at high accumulation sites with warmer temperatures.

Ice-Core Chronology

Accurate chronology is obviously key to using ice-core records to study past environmental changes, particularly to compare features of ice-core records with data from other locations. Annual layer counting is the most powerful and best known technique, but a variety of other methods are also employed.

Layer counting provides the most accurate dating. Annual layering can be discerned visually because of seasonal variations in snow properties ([Figure 2](#)), but annual signals are also present in a variety of geochemical parameters, including stable isotopes, electrical conductivity, and a variety of soluble chemical species. Recent work uses several of these parameters at the same time to construct chronologies.

Annual layers are not discernable at all sites, and thinning makes them difficult to discern in older sections of deep ice

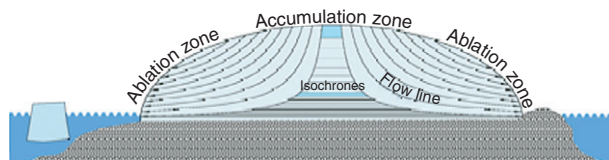


Figure 1 Schematic cross-section of an ice sheet. Ideal ice-core drilling sites are at or near the 'ice divide' at the center of the ice sheet where a core recovers ice deposited at approximately the same location on the surface.

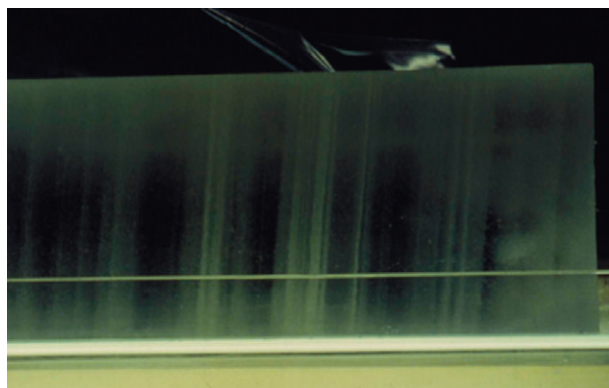


Figure 2 Annual layers in the GISP2 ice core. This section of ice core is ~30 cm long. The horizontal line is the edge of a plastic sample tray.

cores. In many cases, other techniques are needed to establish the ages of deep ice cores.

Radiometric dating of ice has not been widely applied so far. A number of isotopic systems have been investigated for this purpose, but the low levels of the radioactive species in question have hampered progress. Improvements in analytical techniques may change this situation in the future (Aciego et al., 2011) and a robust radiometric dating technique for ice is highly desirable.

Another common strategy is to use the fairly well understood physics of ice flow and thinning to estimate the depth–age relationship for an ice core. Such estimates can be checked or constrained by assuming the ages of key climate transitions (e.g., glacial–interglacial transitions) recorded in ice-core isotope (or other) records, or can be correlated to chronologies of ocean sediment records. The deepest Antarctic records have been dated in this way (e.g., EPICA Community Members, 2004).

Gas records in ice cores also provide dating techniques. One approach is to use globally synchronous variations in trapped atmospheric gases as tools to create precise relative chronologies between ice-core records (Bender et al., 1994; Blunier and Brook, 2001). This allows two ice cores (e.g., one in Greenland and another in Antarctica) to be placed on common chronologies, allowing precise comparisons between them. The isotopic composition of atmospheric oxygen in the ice-core record can also be correlated with the marine isotope record and its established chronology (Sowers et al., 1993). More recently, Bender (2002) showed that the O_2/N_2 ratio of trapped air in the Vostok ice core varies in a way that appears to be coherent with insolation variations on timescales associated with orbital precession, and tuning O_2/N_2 records to local insolation can provide chronological control (Kawamura et al., 2007).

Correlation of various parameters in ice cores to a variety of other independently dated records can also assist in creating ice-core chronologies. Variations in cosmogenic isotopes (e.g., ^{10}Be) in the ice core can be correlated to well-dated oscillations in the tree-ring ^{14}C record (Taylor et al., 2004). Stable isotope records from cave deposits are dated accurately using U-series dating and show variations that are apparently synchronous with fluctuations in Greenland ice-core records (Wang et al., 2001). Peaks in volcanic sulfate in ice cores can be correlated to dates of historically known volcanic eruptions, providing an extremely important check on layer-counting chronologies (Meese et al., 1994).

Proxy and Other Measurements in Ice Cores

A variety of chemical and physical measurements are performed on ice core samples, intended to infer something about past environmental conditions. Some of the more notable environmental 'proxies' are reviewed briefly below. Space limitations prevent mention of all the areas of study, and the reader is referred to other sections of this encyclopedia for more detailed information.

Visual Stratigraphy and Other Physical Measurements

The thickness of annual layers provides a first-order record of snowfall variations (Figure 2), after corrections for thinning due to glacier flow, which can be made fairly accurately in many situations (Cuffey and Clow, 1997). Past precipitation rates are an extremely important paleoenvironmental parameter and are often correlated with temperature in the polar regions. Layer counting is laborious but extremely informative. New techniques combine visual stratigraphy with seasonal variations in other parameters to provide multiparameter dating and accumulation rate estimates, and algorithms have been developed to automatically count annual layers in some of these variables (Taylor et al., 2004).

The distribution of summer melt layers in ice cores is another visual measurement that provides climate information (more melt layers associated with warmer summer temperatures; Alley and Anandakrishnan, 1995). Melt layers are formed when surface snow melts and the meltwater percolates in to the snow-pack and collects in a layer at shallow depth, probably along high-density snow surfaces. Melt layers are visible as bubble-free ice layers, and their distribution can be counted as a function of time.

Other physical properties of the ice accessible by visual observation or basic physical measurements include the orientation and size of ice crystals, size and distribution of bubbles, disappearance of bubbles as clathrate forms at depth, and ice density. Careful measurements of these properties provide the context for interpreting climate records as well as understanding stratigraphic discontinuities, anomalous ice flow, and ice sheet dynamics in general (i.e., Gow et al., 1997).

Stable Isotopes

The stable isotopic composition of ice ($\delta^{18}O$ and δD) has long been used as a paleotemperature indicator, with original work on this subject begun in the 1960s and 1970s. The temperature

dependence of the isotope ratios arises as a result of fractionation during transport from a moisture source to the ice-core site. The processes involved are complex, but the basic concept is that lower temperatures allow a greater isotopic ‘distillation’ as moisture travels to the ice-core site. Measurements of isotope ratios in high-latitude precipitation show the expected trend of lighter ratios with lower temperature. The modern relationship can be used to calculate past temperature change from ice-core isotope records, but numerous recent studies employing bore-hole temperatures and nitrogen and argon isotopes show that the modern ‘spatial’ slope is probably not always appropriate for past conditions. These studies suggest that, in Greenland, past isotope variations have a greater sensitivity to temperature change than indicated by the spatial slope (Severinghaus and Brook, 1999), but in Antarctica, the spatial and temporal sensitivities appear to be similar.

The ratios $\delta^{18}\text{O}$ and δD have been measured somewhat interchangeably in the history of ice-core research. There are small differences between the responses of these two ratios to environmental conditions though, expressed as a parameter called ‘deuterium excess.’ Its primary use is as a monitor of the conditions (primarily temperature) at the moisture source. This is important for understanding the impact of moisture source changes on the isotope–temperature calibration, as well as low-latitude climate change.

In contrast to the polar regions, the systematics of isotope records in mid and tropical latitudes is more complex and inferences of temperature change more difficult.

Greenhouse and Biogenic Gases

The most unique aspect of ice-core records is the archive of ancient atmosphere that they provide. Air is trapped at the base of the firn layer where compacting snow finally turns to ice and the trapped air is present in small bubbles (Figure 3).

The depth of this transition varies with temperature and accumulation rate but is ~50–100 m below the surface at polar sites. For this reason, there is an offset between the age of the air and the age of the ice, which is accounted for with glaciological models of firn densification and gas trapping. Air in



Figure 3 Air bubbles in a thin section of ice from ~140 m depth at the GISP2 ice coring site in Greenland.

bubbles is extracted by melting, crushing, or grating ice in vacuum. Dry techniques of crushing or grating are required for some gases, such as carbon dioxide.

Detailed records of the three main biogenic greenhouse gases, namely, carbon dioxide, methane, and nitrous oxide, are available from ice cores from Greenland and Antarctica. These records cover time periods as recent as the rise of the levels of these gases over the last few centuries to long-term glacial–interglacial cycles going back over 800 000 years. The overall reliability of the gas records can be demonstrated by the overlap of ice-core records with instrumental measurements that started in the 1950s and agreement between ice cores from different locations. Analytical or preservation artifacts impact these data in some cases, but are fairly well recognized. Early carbon dioxide records from Greenland are compromised by melt layers, which have elevated carbon dioxide levels because of the high solubility of carbon dioxide in water. As a result, only carbon dioxide records from colder locations in Antarctica are deemed reliable. Methane records do not suffer from this problem, but, very occasionally, high levels are found in basal ice from ice cores, possibly related to the presence of microbes associated with particle-rich layers (Tung et al., 2005). Nitrous oxide has proved somewhat more problematic. Elevated levels of nitrous oxide are associated with elevated dust levels in several ice cores (Schilt et al., 2010; Sowers, 2001; Spahni et al., 2005) and the controls on these elevated levels are not completely understood. Nonetheless, reliable data can apparently be obtained by selecting samples with low particulate concentrations (Spahni et al., 2005).

More recent work expands this field by using the stable isotopic composition of the three major biogenic gases to constrain sources and sinks (Elsig et al., 2009; Ferretti et al., 2005; Indermühle et al., 1999; Laurantou et al., 2010; Smith et al., 1999; Sowers et al., 2003) and by investigating the possibility of reconstructing levels of rarer gases that are important components of global biogeochemical cycles, for example, carbonyl sulfide, methyl bromide, and methyl chloride (e.g., Aydin et al., 2004; Montzka et al., 2004; Saltzman et al., 2004).

Finally, trace gas records from firn air deserve mention. The firn is the unconsolidated snow pack. Gas diffusion in the firn is sufficiently slow so that firn contains air with a mixture of gases that increase in age with depth. The significance of this is that large volumes of air can be pumped from the firn for measurements, which would otherwise be impossible with traditional ice samples (e.g., Fain et al., 2009; Montzka et al., 2004; Trudinger et al., 2002).

N, O, Ar, and Other Noble Gases

Although greenhouse gas records from ice cores receive attention with respect to the modern global warming problem, equally interesting paleoenvironmental information is available from the major gases in air trapped in ice (O_2 , N_2 , and Ar), other noble gases, and the isotopes of these gases.

Atmospheric O_2 is important in ice-core records because its stable isotopic composition ($\delta^{18}\text{O}$) is slaved to that of the ocean through global photosynthesis and respiration. Ocean $\delta^{18}\text{O}$ varies as the ice sheets grow and shrink through glacial–

interglacial cycles. The turnover time of O_2 in the ocean-atmosphere system is long with respect to the atmospheric mixing time, so $\delta^{18}O$ of O_2 variations are synchronous everywhere on the Earth, providing stratigraphic markers for ice-core chronologies (Bender et al., 1994).

The isotopic composition of nitrogen ($\delta^{15}N$ of N_2) is influenced by gravitational fractionation in the firm, and can be used to correct other isotopes for this fractionation (atmospheric $\delta^{15}N$ of N_2 is thought to be constant on million-year timescales). More importantly, nitrogen and argon isotope ratios have also been used together as paleothermometers, taking advantage of a phenomenon known as ‘thermal diffusion,’ which causes fractionation in a temperature gradient. Nitrogen and argon data can be used together to deconvolve gravitational and thermal signals and calculate the magnitude of temperature change associated with fast changes in climate recorded in ice-core records (Grachev and Severinghaus 2003a; Grachev and Severinghaus 2003b; Severinghaus and Brook, 1999; Severinghaus et al., 1998), thus providing an important independent paleothermometer.

These gases also have other uses. Ratios of O_2 , N_2 , Ar, Xe, and Kr provide important information on gas loss during bubble close-off or storage (Severinghaus and Battle, 2006) (because of the size dependence of loss processes) and on the possible influence of melting on the gas content of ice samples (because of the differences in solubility). Studies of these gases in firm air provide insight into processes that transport air in the firm and their significance in the interpretation of the ice-core record (Kawamura et al., 2006).

Borehole Temperatures

The vertical temperature profile of an ice sheet can contain information about the past surface temperature history. Borehole temperature profiles are measured by suspending temperature sensors in fluid-filled boreholes and recording the temperature as a function of depth using highly accurate and precise instrumentation. This is an important area of study because it can provide independent estimates of temperature change that complement the stable isotopes of oxygen and hydrogen in ice and the nitrogen and argon isotopic work described above. The basic idea is that the ice sheet remembers past surface temperature changes because these propagate through the ice as a result of heat transport. If heat transport properties of the ice, as well as the geothermal heat flux, are known, the temperature profile can be inverted to provide a record of surface temperature change, which is independent of proxy calibration. Analogous work is done with continental boreholes to measure warming trends over the past few centuries. Such records are smoothed representations of more abrupt changes due to the nature of heat transport, but provide extremely important information, for example, the first estimate of very large glacial–interglacial temperature change in Greenland, revealing complications with the stable isotope thermometer (Cuffey et al., 1995).

Glaciochemistry

The chemical composition of ice-core samples provides information about past changes in atmospheric circulation, dust loading, volcanic eruptions, biological activity, human impact

on the environment, and a wealth of other information. In general, this area of research includes measurements of soluble species (those released to solution when ice samples are melted), as opposed to studies of insoluble particles (below). The distinction between soluble species and very small particles is operational but not significant for the purposes of this discussion.

Chemical measurements at a variety of sites reveal the human impact on the atmosphere. For example, records of lead concentration in Greenland ice (Figure 4) reveal the large impact of industrial lead, and show how it has changed with regulation of leaded gasoline and other global political–environmental changes (McConnell et al., 2002a,b). Many other pollutant species are present in the ice as well.

On longer timescales, soluble chemistry records the impacts of major volcanic eruptions through peaks in volcanic sulfate deposition. These data provide information about the frequency of eruptions and the climate impact of large volcanic eruptions, and, when the age of the eruption event is known (i.e., Zielinski et al., 1994), they provide independent tests of ice-core chronology.

The major ion chemistry of ice cores is useful in a more general way as a tracer of the input of chemical species from continental dust, sea salt, and biogenic processes. For example, calcium concentrations are believed to largely reflect the dustiness of the atmosphere, and sodium and chloride are largely derived from the ocean and their concentrations may reflect the proximity to open water (which may change as climate changes), other aspects of meteorology that influence the transport of marine air to ice-core sites, or more complex processes related to the behavior of salts on sea ice surfaces (Rankine and Wolff, 2002). Major ions commonly measured in ice include Na^+ , Mg^{2+} , Ca^{2+} , K^+ , NH_4^+ , Cl^- , NO_3^- , and SO_4^{2-} , and organic acids such as methylsulphonate and CH_3SO^- , and multivariate statistical analysis has been used to attempt to describe the relationship between the deposition of these species and meteorological parameters. Many of these species are measured by ion chromatography using continuous melter systems that provide a continuous sample stream, creating very high resolution datasets.

More recently, the advent of high-sensitivity Inductively coupled plasma (ICP) mass spectrometers has opened a new window in glaciochemistry, allowing for measurements of many trace species in ice at very high resolution (McConnell et al., 2002a,b).

Insoluble Particles

The insoluble particulate content of ice cores includes mineral dust, volcanic ash, and extraterrestrial material (discussed below), and potentially contains important information about the dustiness of the atmosphere in past times, and, through the particle size distribution, possibly information about past wind speed. Particle abundances in most ice cores are very low, and optical or electrical sensing methods on solid or liquid samples have been used to produce continuous or semicontinuous records of particle numbers and sizes. Bulk dust data generally show the strong glacial–interglacial cycles, with higher levels of dust in cold glacial periods. For example, a recent long dust record from the East Antarctic EPICA Dome C core clearly shows glacial–interglacial dust cycles (Figure 5; Jouzel et al., 2007). The amplitude of the cycles is too large to

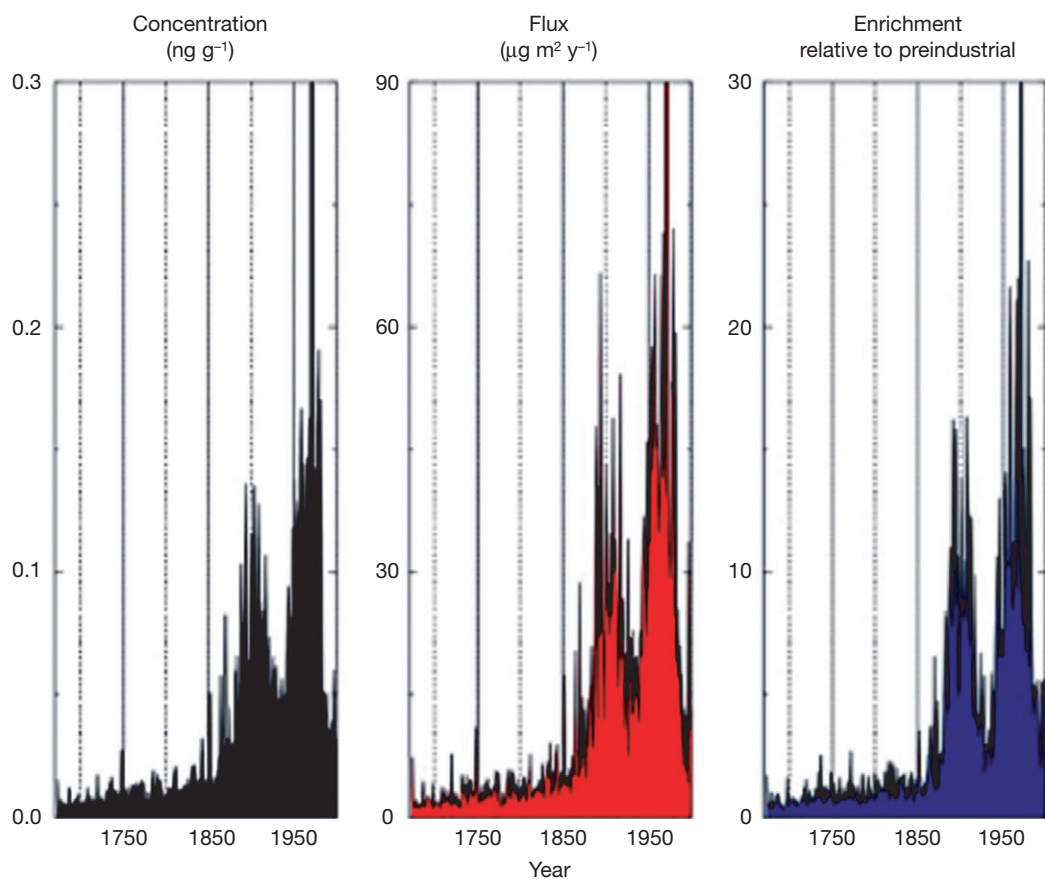


Figure 4 A 330-year continuous ice-core record of Pb concentration, flux, and relative enrichment at the D5 ice core site (68.5°N , 42.9°W) in west central Greenland. Reproduced from McConnell JR, Lamorey GW, and Hutterli MA (2002a) A 250-year high-resolution record of Pb flux and crustal enrichment in central Greenland. *Geophysical Research Letters* 29: 2130; McConnell JR, Lamorey GW, Lambert SW, and Taylor KC (2002b) Continuous ice-core chemical analyses using inductively coupled plasma mass spectrometry. *Environmental Science and Technology* 36: 7–11.

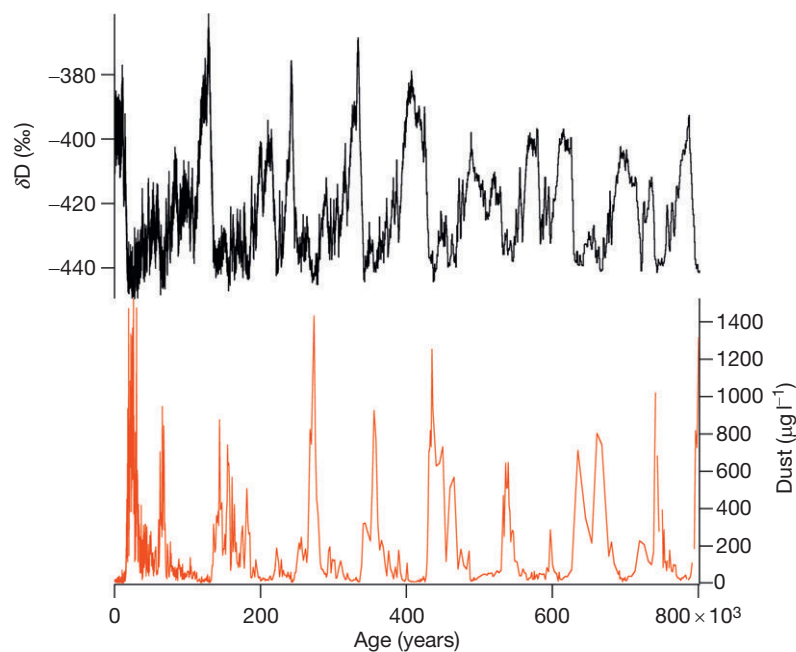


Figure 5 The dust and δD records from the EPICA Dome C ice core in Antarctica (EPICA Community Members, 2004; Jouzel et al., 2007; Lambert et al., 2008).

be explained by changes in ice accumulation rate alone, implicating a dustier, drier atmosphere, or changes in atmospheric circulation patterns that bring dust to this high-latitude site.

The chemical and isotopic composition of insoluble dust can provide important information about the source of dust to remote ice-coring sites. This work is difficult because of the low levels of dust involved and the sample size requirements for some of the techniques. Isotopic fingerprinting using lead, strontium, and neodymium isotope ratios suggests, for example, that most of the dust at central Greenland sites is derived from low-latitude regions of east Asia (Biscaye et al., 1997), and that the dust in east Antarctica originates mainly from the Patagonian region of South America, with possible contributions from other areas (Delmonte et al., 2004).

Particles of volcanic origin (glass shards or tephra) can sometimes be identified uniquely in ice-core samples, providing a history of volcanic activity and, in some cases, datable time horizons useful for ice-core chronology.

Cosmogenic Isotopes and Extraterrestrial Material

Ice cores contain several kinds of information about extraterrestrial phenomena. Cosmogenic isotopes (produced by cosmic rays) produced in the atmosphere are deposited on the ice sheet, providing a record of the processes that produce and modulate the cosmic ray flux, and influence their deposition rates, including changes in solar activity and the Earth's magnetic field (e.g., Muscheler et al., 2005). ^{10}Be (half-life 1.5×10^6 years) and ^{36}Cl (3.0×10^5 years) are the most commonly measured cosmogenic isotopes in ice. ^{10}Be data record the influence of a geomagnetic excursion (the Laschamp excursion) around 41 000 years ago, and detailed records show variations related to the modulating effects of the solar wind on cosmic ray flux at high latitudes (Finkel and Nishiizumi, 1997). In addition to revealing the modulating effects, these features provide chronological ties that can be used to correlate ice cores and constrain ice chronologies. These data have relevance beyond ice-core studies because of the significant influence of these modulation processes on the production of radiocarbon in the atmosphere and the resulting implications for the radiocarbon timescale. ^{10}Be data have also been used as accumulation rate constraints, assuming a constant ^{10}Be flux and fraction of dry versus wet deposition (e.g., Raisbeck et al., 1987; Steig et al., 1998), although in detail this assumption may not be correct.

Ice cores also contain meteoritic material, albeit in small quantities given the limited sample availability (micrometeorites have been extracted from large surface samples of melted snow and ice for some time, but given the limited size of ice samples, this type of work is not practical for deep ice cores). Isotopic and elemental measurements of species enriched in meteoritic material, including iridium (Gabrielli et al., 2004; Karner et al., 2003) and ^3He (Brook et al., 2000, 2009), can be used to examine possible changes in flux of extraterrestrial dust to the earth, although this field is in its infancy.

Biological Material in Ice

Although the existence of life in harsh environments has been a topic of research for quite some time, studies of life in ice have flourished in the last decade or so. This work is difficult due to potential contamination issues and low levels of biological

material preserved in ice. Current attention focuses on the possible viability of microorganisms in ice; characterization and preservation of ancient DNA and RNA; preservation of fungi, spores, and other organisms; and implications for life in subglacial lakes and in ice on other planets.

Major Highlights of Ice Core Research

This section is a brief review of some of the major patterns and findings of ice-core research. Space limits this review to a few key issues – the remainder of the ice-core section of this encyclopedia provides much more relevant details.

Glacial–Interglacial Cycles

Deep ice cores from Antarctica are the only ice-core records to penetrate beyond the end of the last interglacial period. In Greenland, there are tantalizing hints (Landais et al., 2003; Suwa et al., 2006) of this penultimate interglacial near the base of the ice sheet, but no complete sections have yet been obtained. Until recently, the oldest Antarctic record was that of the east Antarctic core from Vostok station, extending back through four glacial–interglacial cycles to ~440 000 years before present (Petit et al., 1999).

An ice core on Dome C drilled by the European Program for Ice Coring in Antarctica (EPICA) collaboration recovered a core with at least 800 000 years of interpretable record, almost doubling the Vostok achievement (EPICA Community Members, 2004; Jouzel et al., 2007). Records from EPICA Dome C (Figure 6) clearly show the strong 100 000 year cyclicity in climate well known from marine sediments. One of the major achievements of this drilling program is the complementary carbon dioxide, methane, and nitrous oxide records (Loulergue et al., 2008; Lüthi et al., 2008; Siegenthaler et al., 2005; Spahni et al., 2005), which clearly show how closely the concentrations of these gases follow Antarctic climate patterns and how unusually the modern values of these gases are viewed in the context of over 800 000 years of Earth's history.

The relationship between the gas records, particularly carbon dioxide, and Antarctic and global temperatures is a subject of intense research, and has been so since the original publication of long Antarctic carbon dioxide records in the 1980s. A major question concerns the timing of carbon dioxide and temperature change on these timescales. Most studies of this issue have concluded that there is a lag of several hundred to several thousand years between temperature change and carbon dioxide change (temperature leads carbon dioxide), suggesting that warming, triggered by other factors, causes a change in the carbon cycle that increases carbon dioxide, which is then a positive feedback on warming. Another major question concerns the ultimate cause of the carbon dioxide variations. It is widely believed that they are caused by changes in the ocean carbon cycle, either as a result of changes in partitioning of carbon between the deep and surface ocean due to changes in ocean circulation, or because of changes in biological productivity that transfer carbon from the surface to the deep ocean (Sigman and Boyle, 2000). Despite several decades of work, the ultimate cause of the carbon dioxide cycles remains elusive.

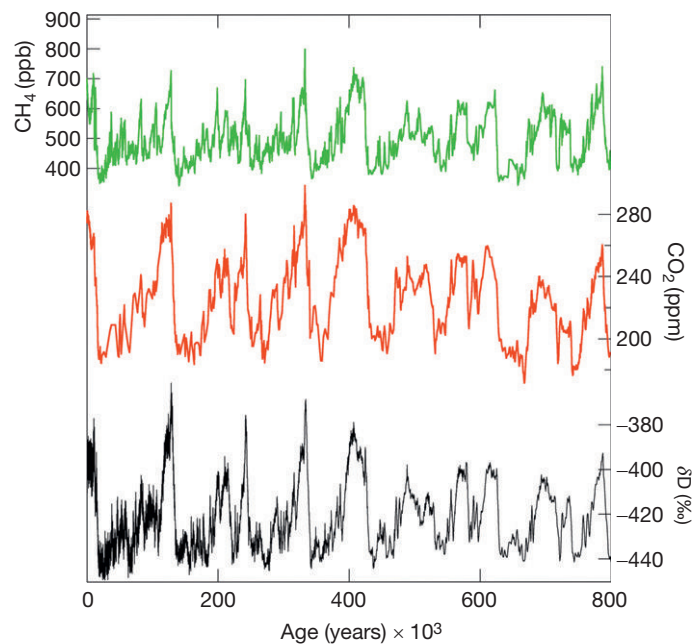


Figure 6 Composite records of atmospheric carbon dioxide and methane from the EPICA Dome C, Vostok, and Taylor Dome ice cores in Antarctica (Louergue et al., 2008; Lüthi et al., 2008; Siegenthaler et al., 2005; Spahni et al., 2005) and the δD record (EPICA Community Members, 2004; Jouzel et al., 2007).

Abrupt Change, the Last Ice Age, and the Holocene

In contrast to the long march of long-term climate cycles seen in the longest Antarctic ice cores, cores from central Greenland have revealed a remarkable pattern of more rapid climate change during the last Ice Age (Figure 7). This more abrupt style of climate change is often referred to as ‘millennial-scale climate change,’ although the abrupt warmings and coolings in the Greenland records are just specific examples of change on this timescale.

Although the first deep ice cores drilled in Greenland showed these abrupt climate shifts, it was not until the recovery of the parallel Greenland Ice Core Project (GRIP) and Greenland Ice Sheet Project 2 (GISP2) ice cores from central Greenland in the early 1990s that the significance of these abrupt shifts was recognized more widely outside.

The stable isotope records from central Greenland deep cores (Figure 7) show a sequence of more than 20 abrupt warming and cooling events during the last Ice Age (from ~120 000 to 10 000 years ago). These events, called ‘Dansgaard-Oeschger events’ after their discoverers Hans Oeschger and Willi Dansgaard, are remarkable for their frequency, size, and abruptness. Detailed studies of the isotopic variations and associated changes in layer thickness indicate that the cold-warm transitions are very abrupt (changes in mean accumulation rate take place over decades or less; Alley et al., 1993), and represent large temperature changes (~10–15°). These observations have led to a large amount of research dedicated to understanding the causes of abrupt climate change, their manifestation in other parts of the world, and the possibility of their recurrence in the future.

An intriguing aspect of millennial-scale climate change is that Antarctic ice cores do not record a pattern similar to that of Greenland. Instead, Antarctic isotope records seem to show a

‘see-saw’ relationship, where large warming events in the Northern Hemisphere are preceded in the South by more gradual warming that terminates at the time of abrupt warming in the North. The pattern has been attributed to changes in ocean heat transport associated with the D–O events (Knutti et al., 2004 and references therein), although other explanations have been suggested (Wunsch, 2003).

Following the last abrupt climate shift recorded in Greenland (the end of the Younger Dryas period), climate variation was significantly muted compared to the last Ice Age, with one notable abrupt cool event at about 8.2 ka (Alley et al., 1997). Other aspects of the Greenland records are not as uniform in the Holocene. For example, Mayewski et al. (2004) discuss oscillations in major ion concentrations that they relate to polar cooling and tropical drying. Atmospheric methane levels dropped dramatically during the early Holocene, and then rose after ~5000 years ago. Atmospheric CO₂ rose from an early Holocene low at ~8000 years ago to the late Holocene value of ~280 ppm by 1000 years ago. Antarctic stable isotope records also do not indicate large, abrupt changes during the Holocene, but temporal trends related to changes in ice sheet elevation, an early Holocene climate optimum 11 500–9000 years ago, and shorter-term millennial events are apparent (Masson et al., 2000).

Recent and Anthropogenic Change

As mentioned above, ice cores record in great detail the anthropogenic impact on the remote atmosphere. Records of pollutants such as lead and sulfate are preserved in annual resolution, and the rise of greenhouse gases in the ice core records provides fundamental information relevant to the modern global warming problem (Figure 8).

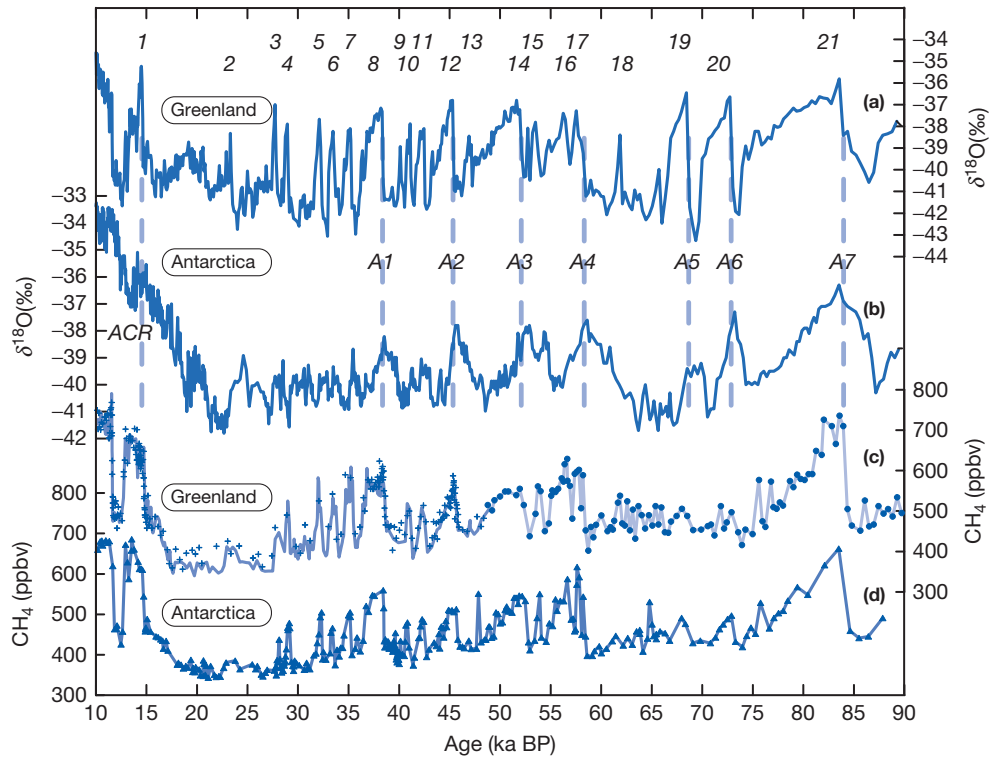


Figure 7 Records of millennial-scale variations in stable isotope records from Greenland (GISP2 ice core) and Antarctica (Byrd ice core). Reproduced from [Blunier T and Brook EJ \(2001\)](#) Timing of millennial-scale climate change in Antarctica and Greenland during the last glacial period. *Science* 291: 109–112.

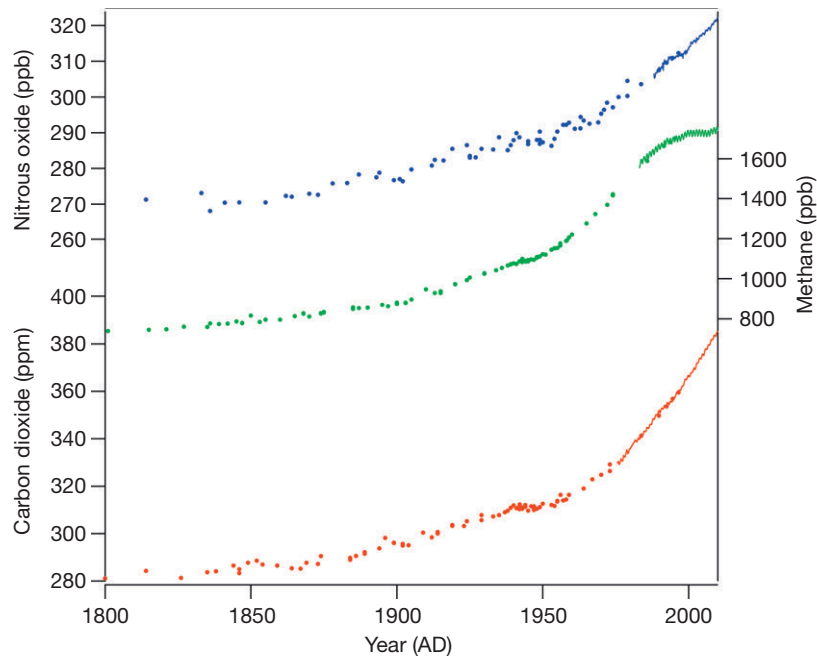


Figure 8 Records of the greenhouse gases carbon dioxide, methane, and nitrous oxide, showing the anthropogenic increase over the last two centuries. Data from ice cores ([MacFarling Meure et al., 2006](#)) and direct atmospheric measurements, compiled and updated on the basis of publicly available data provided by the National Oceanic and Atmospheric Administration Earth Systems Research Laboratory.

New data on a variety of trace gas species and their isotopes in firn air and ice promise to improve our understanding of anthropogenic change, as do high-resolution studies of trace elements. Equally important are new networks of annually resolved records that will allow for greater understanding of the spatial distribution of climate change at high latitude and altitudes, through programs such as ITASE and PARCA. These regional assessments are critical for adequately assessing the influence of greenhouse warming at high latitudes.

Future of Ice Coring

Data from ice cores underpin much of global change science, and in the last two decades have provided more detailed and longer views of atmospheric and environmental change in the polar regions than perhaps originally envisioned. The oldest records are 800 000 years old, and records from areas of high snow accumulation provide annual data for tens of millennia. Cores from high elevation sites in the tropics and mid-latitudes, obtained by heroic measures, provide important information about climate change in these regions, unavailable from other means.

The future of ice coring involves both collecting more cores, to examine spatial variations in environmental change, and collecting older cores, to extend the record as far back as possible. A broad-based international program called International Partners in Ice Coring Sciences is involved in planning major new coring programs. For example, cores penetrating the last glacial transition and Greenland and Antarctica show significant inter-hemispheric differences, as well as differences in the character of deglacial climate change within Antarctica. Fully understanding the nature of this large change in the earth system requires a detailed and well-dated spatial map of the progression of the deglaciation; a network of deep ice cores in Greenland and Antarctica could accomplish this. On shorter timescales, a key global change question concerns separating anthropogenically forced climate change from natural variability. For this question, a network of cores in both polar regions and other parts of the world covering the last 2000 years is desirable. For older cores, a major goal in Antarctica is an ice-core record back to 1.2 or even 1.5 My. Such a core would extend through the mid-Pleistocene transition, a time period where Earth's climate shifted from 40 000 to 100 000-year climate cycles. This core would allow investigation of the role of greenhouse gases in this transition, as well as provide a detailed land-based record of the evolution of temperature and other environmental variables over this time period. In Greenland, a core in a location that will allow recovery of the Eemian, or last interglacial, period, is a major goal.

See also: **Ice Core Methods:** Biological Material; Borehole Temperature Records; Chronologies; CO₂ Studies; Conductivity Studies; Glaciochemistry; Methane Studies; Microparticle and Trace Element Studies; Stable Isotopes. **Ice Core Records:** Africa; Antarctic Stable Isotopes; Chinese, Tibetan Mountains; Correlations Between Greenland and Antarctica; Greenland Stable Isotopes; Ice Margin Sites; South America; Thermal Diffusion Paleotemperature Records. **Ice Cores:** Dynamics of the Greenland Ice Sheet; History of Research, Greenland and Antarctica.

References

- Aciego S, Bourdon B, Schwander J, Baur H, and Forieri A (2011) Toward a radiometric ice clock: Uranium ages of the Dome C ice core. *Quaternary Science Reviews* 30: 2389–2397.
- Alley RB and Anandakrishnan S (1995) Variations in melt-layer frequency in the GISP2 ice core: Implications for Holocene summer temperatures in central Greenland. *Annals of Glaciology* 40: 341–349.
- Alley R, Mayewski P, Sowers T, Stuiver M, Taylor K, and Clark P (1997) Holocene climatic instability: A prominent, widespread event 8200 yr ago. *Geology* 25: 483–486.
- Alley RB, Meese D, Shuman CA, et al. (1993) Abrupt accumulation increase at the Younger Dryas termination in the GISP2 ice core. *Nature* 362: 527–529.
- Aydin M, DeBruyn WJ, Saltzman ES, Montzka SA, and Butler JH (2004) Atmospheric variability of methyl chloride during the last 300 years from an Antarctic ice core and firn air. *Geophysical Research Letters* 31: L0219. <http://dx.doi.org/10.1029/2003GL018750>.
- Bender ML (2002) Orbital tuning chronology for the Vostok climate record supported by trapped gas composition. *Earth and Planetary Science Letters* 204: 275–289.
- Bender M, Sowers T, Dickson ML, et al. (1994) Climate correlations between Greenland and Antarctica during the past 100,000 years. *Science* 372: 663–666.
- Biscaye PE, Grousset FE, Revel M, et al. (1997) Asian provenance of glacial dust (stage 2) in the Greenland Ice Sheet Project 2 ice core, Summit, Greenland. *Journal of Geophysical Research* 102: 26765–26781.
- Blunier T and Brook EJ (2001) Timing of millennial-scale climate change in Antarctica and Greenland during the last glacial period. *Science* 291: 109–112.
- Brook E (2005) Tiny bubbles tell all. *Science* 310: 1285–1287.
- Brook EJ, Kurz MD, and Curtice J (2009) Flux and size fractionation of ³He in interplanetary dust from Antarctic ice core samples. *Earth and Planetary Science Letters* 286(3–4): 565–569.
- Brook EJ, Kurz MD, Curtice J, and Cowburn S (2000) Accretion of interplanetary dust in polar ice. *Geophysical Research Letters* 27: 3145–3148.
- Cuffey KM and Clow GD (1997) Temperature, accumulation, and ice sheet elevation in central Greenland through the last deglacial transition. *Journal of Geophysical Research* 102: 26383–26396.
- Cuffey KM, Clow GD, Alley RB, Stuiver M, Waddington ED, and Saltus RW (1995) Large Arctic temperature change at the Wisconsin–Holocene glacial transition. *Science* 270: 455–458.
- Delmonte B, Basile-Doelsch I, Petit J-R, et al. (2004) Comparing the EPICA and Vostok dust records during the last 220,000 years: Stratigraphical correlation and provenance in glacial periods. *Earth-Science Reviews* 66: 63–87.
- Elsig J, Schmitt J, Leuenberger D, et al. (2009) Stable isotope constraints on Holocene carbon cycle changes from an Antarctic ice core. *Nature* 461: 507–510.
- EPICA Community Members (2004) Eight glacial cycles from an Antarctic ice core. *Nature* 429: 623–628.
- Fain X, Ferrari CP, Dommergue A, et al. (2009) Polar firn air reveals large-scale impact of anthropogenic mercury emissions during the 1970s. *Proceedings of the National Academy of Sciences* 106(38): 16114–16119.
- Ferretti DF, Miller JB, White JWC, et al. (2005) Unexpected changes to the global methane budget over the past 2000 years. *Science* 309: 1714–1716.
- Finkel RC and Nishiizumi K (1997) Beryllium 10 concentrations in the Greenland Ice Sheet Project 2 ice core from 3–40 ka. *Journal of Geophysical Research* 102: 26699–26706. <http://dx.doi.org/10.1029/97JC01282>.
- Gabrielli P, Barbante C, Plane JMC, et al. (2004) Meteoric smoke fallout over the Holocene epoch revealed by iridium and platinum in Greenland ice. *Nature* 432: 1011–1014.
- Gow AJ, Meese DA, Alley RB, et al. (1997) Physical and structural properties of the Greenland Ice Sheet Project 2 ice core: A review. *Journal of Geophysical Research* 102: 26559–26576.
- Grachev AM and Severinghaus JP (2003a) Determining the thermal diffusion factor for ⁴⁰Ar/³⁶Ar in air to aid paleoreconstruction of abrupt climate change. *Journal of Physical Chemistry A* 107: 4636–4642.
- Grachev AM and Severinghaus JP (2003b) Laboratory determination of thermal diffusion constants for ²⁹N₂/²⁸N₂ in air at temperatures from -60 to 0°C for reconstruction of magnitudes of abrupt climate changes using the ice core fossil-air paleothermometer. *Geochim Cosmochim Acta* 67: 345–360.
- Grachev A and Severinghaus J (2005) A revised +10 ± 4 °C magnitude of the abrupt change in Greenland temperature at the Younger Dryas termination using published GISP2 gas isotope data and air thermal diffusion constants. *Quaternary Science Reviews* 24: 513–519.
- Indermühle A, Stocker TF, Joos F, et al. (1999) Holocene carbon-cycle dynamics based on CO₂ trapped in ice at Taylor Dome, Antarctica. *Nature* 398: 121–126.
- Jouzel J, Masson-Delmotte V, Cattani O, et al. (2007) Orbital and millennial Antarctic climate variability over the past 800,000 years. *Science* 317(5839): 793.

- Karner DB, Muller RA, Levine J, Asaro F, Ram M, and Stolz MR (2003) Extraterrestrial accretion from the GISP2 ice core. *Geochimica et Cosmochimica Acta* 67: 751–763.
- Kawamura K, Parrenin F, Lisiecki L, et al. (2007) Northern Hemisphere forcing of climatic cycles in Antarctica over the past 360,000 years. *Nature* 448(7156): 912–916.
- Kawamura K, Severinghaus JP, Ishido S, et al. (2006) Convective mixing of air in firn at four polar sites. *Earth and Planetary Science Letters* 244: 672–682.
- Knutti R, Flückiger J, and Timmermann A (2004) Strong hemispheric coupling of glacial climate through freshwater discharge and ocean circulation. *Nature* 430: 851–856.
- Lambert F, Delmonte B, Petit JR, et al. (2008) Dust-climate couplings over the past 800,000 years from the EPICA Dome C ice core. *Nature* 452(7187): 616–619.
- Landais A, Chappellaz J, Delmotte M, et al. (2003) A tentative reconstruction of the last interglacial and glacial inception in Greenland based on new gas measurements in the Greenland Ice Core Project (GRIP) ice core. *Journal of Geophysical Research* 108: 4563.
- Loulergue L, Schilt A, Spahni R, et al. (2008) Orbital and millennial-scale features of atmospheric CH₄ over the past 800,000 years. *Nature* 453(7193): 383–386.
- Lourantou A, Lavric JV, Kohler P, et al. (2010) Constraint of the CO₂ rise by new atmospheric carbon isotopic measurements during the last deglaciation. *Global Biogeochemical Cycles* 24: GB2015.
- Lüthi D, Le Floch M, Bereiter B, et al. (2008) High-resolution carbon dioxide concentration record 650,000–800,000 years before present. *Nature* 453(7193): 379–382.
- MacFarling Meure C, Etheridge D, Trudinger C, et al. (2006) Law Dome CO₂, CH₄ and N₂O ice core records extended to 2000 years BP. *Geophysical Research Letters* 33(14): L14810.
- Masson V, et al. (2000) Holocene climate variability in Antarctica based on 11 ice core isotopic records. *Quaternary Research* 54: 348–358.
- Mayewski PA, Rohling E, Stager C, et al. (2004) Holocene climate variability. *Quaternary Research* 62: 243–255.
- McConnell JR, Lamorey GW, and Hutterli MA (2002a) A 250-year high-resolution record of Pb flux and crustal enrichment in central Greenland. *Geophysical Research Letters* 29: 2130.
- McConnell JR, Lamorey GW, Lambert SW, and Taylor KC (2002b) Continuous ice-core chemical analyses using inductively coupled plasma mass spectrometry. *Environmental Science and Technology* 36: 7–11.
- Meese DA, Alley RB, Fiacco RJ, et al. (1994) Preliminary depth-age scale of the GISP2 ice core. *Special CRREL Report* 94-1, US.
- Montzka SA, Aydin M, Butler JH, et al. (2004) A 350-year atmospheric history for carbonyl sulfide inferred from Antarctic firn air and air trapped in ice. *Journal of Geophysical Research* 109: D22302. <http://dx.doi.org/10.1029/2004JD004686>.
- Muscheler R, Beer J, Kubik PW, and Synal HA (2005) Geomagnetic field intensity during the last 60,000 years based on ¹⁰Be and ³⁶Cl from the Summit ice cores and ¹⁴C. *Quaternary Science Reviews* 24(16–17): 1849–1860.
- Petit JR, et al. (1999) Climate and atmospheric history of the past 420,000 years from the Vostok ice core, Antarctica. *Nature* 399: 429–436.
- Raisbeck GM, Yiou F, Bourles D, Lorius C, Jouzel J, and Barkov NI (1987) Evidence for two intervals of enhanced ¹⁰Be deposition in Antarctic ice during the last glacial period. *Nature* 326: 273–277.
- Rankine AM and Wolff E (2002) Frost flowers: Implications for tropospheric chemistry and ice core interpretation. *Journal of Geophysical Research* 107: 4683. <http://dx.doi.org/10.1029/2002JD002492>.
- Saltzman ES, Aydin M, DeBruyn WJ, King DB, and Yvon-Lewis SA (2004) Methyl bromide in preindustrial air: Measurements from an Antarctic ice core. *Journal of Geophysical Research* 109: D05301. <http://dx.doi.org/10.1029/2003JD004157>.
- Schilt A, Baumgartner M, Blunier T, et al. (2010) Glacial–interglacial and millennial-scale variations in the atmospheric nitrous oxide concentration during the last 800,000 years. *Quaternary Science Reviews* 29(1): 182–192.
- Severinghaus J and Battle M (2006) Fractionation of gases in polar ice during bubble close-off: New constraints from firn air Ne, Kr, and Xe observations. *Earth and Planetary Science Letters* 244: 474–500.
- Severinghaus J and Brook E (1999) Simultaneous tropical-abrupt climate change at the end of the last glacial period inferred from trapped air in polar ice. *Science* 286: 930–934.
- Severinghaus JP, Sowers T, Brook EJ, Alley RB, and Bender ML (1998) Timing of abrupt climate change at the end of the Younger Dryas interval from thermally fractionated gases in polar ice. *Nature* 391: 141–148.
- Siegenthaler U, Stocker TF, Monnin E, et al. (2005) Stable carbon cycle–climate relationship during the late Pleistocene. *Science* 310: 1313–1317.
- Sigman D and Boyle EA (2000) Glacial/interglacial variations in atmospheric carbon dioxide. *Nature* 407: 859–869.
- Smith HJ, Fischer H, Wahlen M, Mastroianni D, and Deck B (1999) Dual modes of the carbon cycle since the Last Glacial Maximum. *Nature* 400: 248–250.
- Sowers T (2001) The N₂O record spanning the penultimate deglaciation from the Vostok ice core. *Journal of Geophysical Research* 106: 31903–31914.
- Sowers T, Alley RB, and Jubenville J (2003) Ice core records of atmospheric N₂O covering the last 106,000 years. *Science* 301: 945–948.
- Sowers T, Bender M, Labeyrie L, et al. (1993) 135,000 year Vostok-SPECMAP common temporal framework. *Paleoceanography* 6: 867–875.
- Spahni R, et al. (2005) Atmospheric methane and nitrous oxide of the Late Pleistocene from Antarctic ice cores. *Science* 310: 1317–1321.
- Steig EJ, Brook EJ, White JWC, et al. (1998) Synchronous climate changes in Antarctica and the North Atlantic during the last glacial–interglacial transition. *Science* 282: 92–95.
- Suwa M, Von Fischer JC, Bender ML, Landais A, and Brook EJ (2006) Chronology reconstruction for the disturbed bottom section of the GISP2 and the GRIP ice cores: Implications for Termination II in Greenland. *Journal of Geophysical Research* 111: D02101.
- Taylor KC, Alley RB, Meese DA, et al. (2004) Dating the Siple Dome, Antarctica ice core by manual and computer interpretation of annual layering. *Journal of Glaciology* 50: 453–461.
- Trudinger C, Etheridge D, Rayner P, Enting I, Sturrock G, and Langenfelds R (2002) Reconstructing atmospheric histories from measurements of air composition in firn. *Journal of Geophysical Research* 107: 4780. <http://dx.doi.org/10.1029/2002JD002545>.
- Tung HC, Bramall NE, and Price PB (2005) Microbial origin of excess methane in glacial ice and implications for life on Mars. *Proceedings of the National Academy of Sciences* 102: 18292–18296.
- Wang YJ, Cheng H, Edwards RL, et al. (2001) A high-resolution absolute-dated Late Pleistocene monsoon record from Hulu Cave, China. *Science* 294: 2345–2348.
- Wunsch C (2003) Greenland–Antarctic phase relations and millennial time-scale climate fluctuations in the Greenland ice-cores. *Quaternary Science Reviews* 22: 1631–1646.
- Zielinski GA, Mayewski PA, Meeker LD, et al. (1994) Record of volcanism from the GISP2 ice core since 7000 B.C. and implications for the volcano-climate system. *Science* 264: 948–952.

Biological Material

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Introduction

Snow falling on polar and high-altitude regions has formed $\sim 1.6 \times 10^7$ km² of glacial ice that provides an invaluable archive of past conditions on Earth. The expansive ice sheets of Greenland and Antarctica cover $\sim 10\%$ ($> 1.5 \times 10^7$ km²) of the Earth's terrestrial surface with ice and contain $\sim 70\%$ of the freshwater on the planet (Paterson, 1994). Temperate glaciers cover more than 5×10^5 km² and comprise $\sim 3.5\%$ of the glacial ice on our planet (Table 1). The present volume of the Earth's glacier ice, if totally melted, represents about 80 m in potential sea-level rise with 91%, 8%, and 1% represented by the East and West Antarctic Ice Sheets, the Greenland Ice Sheet, and mountain glaciers, respectively (Hambrey and Alean, 2004).

Research on ice cores from polar and temperate glaciers has focused primarily on the reconstruction of the paleoclimate record to determine the mechanisms responsible for ice-sheet mass balance, associated sea-level change, and the processes leading to the transition between glacial and interglacial periods (e.g., Petit et al. (1999) and Alley (2002)). Ice cores collected from Greenland before the 1990s provided important evidence showing persistent climate instability over the last glacial cycle. Data from the Antarctic Vostok ice core have shown that over the past 400 ka, there was a clear correlation between temperature and greenhouse gases, implying that greenhouse gases contributed to the temperature observations during this period. A recent ice core collected at Dome C as part of the European Project for Ice Coring in Antarctica (EPICA) has extended this record to 650 ka and may ultimately produce a gas record 900 ka old (EPICA, 2004; Brook, 2005; Siegenthaler et al., 2005; Spahni et al., 2005). Information derived from the ice-core record allows predictions of future changes in climate and provides important data to anticipate how these changes will impact future societal issues on our planet.

Atmospheric impurities deposited in glacial ice include aerosols emitted by the oceans and the continents, and anthropogenic activity. Figure 1 shows the process of glacial deposition together with the global distribution of glaciers on our planet. The cold, dry conditions typical of glacial periods reduced the hydrological cycle and precipitation rate. The reduced precipitation increased the residence time of aerosols and dust allowing them to disperse to great distances. This paradigm is revealed in the dust record from the Vostok ice core, which shows that dust concentrations during glacial periods were ~ 50 times greater than for interglacial periods (Petit et al., 1999; Figure 2). The isotopic (Sr and Nd) composition of dust in the Vostok core further shows that it originated in the Patagonian region of South America (Delmonte et al., 2004). Unfortunately, the biogenic nature of the eolian dust particles in glacial ice has received relatively little attention despite its known role in microbial transport (Griffin et al., 2003).

The biological contents of ice cores were first investigated in the 1980s by Abyzov (Abyzov, 1993; Abyzov et al., 2001), who used microscopic observations and traditional cultivation methods to show that microorganisms were present in the Vostok ice core and were positively correlated with the concentration of dust particles. An example of dust and associated microbiota in the Sajama ice core, Bolivia, is shown in Figure 3. Abyzov's results, although controversial, motivated other microbiologists to examine the biotic content of ice cores collected from many icy regions of our planet (e.g., Priscu et al. (1998, 1999), Karl et al. (1999), Christner et al. (2000, 2001, 2003), and Castello et al. (2005)). Anomalies in the concentration and isotope ratio in biogenic trace gases (e.g., CH₄, N₂O) trapped in temperate and polar glacial ice have led glaciogeochemists to suggest enzymatic alteration within the ice by *in situ* metabolism (Sowers, 2001; Campen et al., 2003; Tung et al., 2005). These trace gas anomalies imply that microorganisms are actually metabolizing in solid ice following deposition. Collectively, these data infer that ice cores represent 'ice museums' containing novel records of evolution and habitat variability on our planet. Clearly, biologists should be included in future ice-coring efforts if we are to produce a comprehensive record of past conditions on Earth. We present an overview of recent investigations that focus on biogenic matter in ice cores collected from polar ice sheets (Antarctica and Greenland) and from low-latitude mountain glaciers. We conclude by discussing the metabolic and evolutionary potential of ice-bound microorganisms.

Table 1 Aerial coverage of selected glaciated areas on Earth

Region	Surface area (km ²)	Percent of world total
<i>Polar</i>		
Antarctica	13,593,310	87.5
Greenland	1,726,400	10.88
<i>Polar total</i>	15,319,710	96.58
<i>Temperate</i>		
Africa	10	<0.01
Asia and Eastern Europe	185,211	1.17
Australasia (i.e., New Zealand)	860	0.01
Europe (Western)	53,967	0.34
North America excluding Greenland	276,100	1.74
South America	25,908	0.16
<i>Temperate total</i>	542,056	3.42
<i>World total</i>	15,861,766	

Data from the World Glacier Monitoring Service, 1989.

Biogenic Matter in Polar Ice Sheets and Temperate Mountain Glaciers

The study of biogenic material in ice cores is in its infancy and, except for a few reports, was unheard of a decade ago. We now know that microorganisms and associated nonliving organic matter are present in virtually all ice samples examined for biological properties (Figure 4). The past decade has seen biological research in ice cores go from a novelty to a focused area of scientific research.

The Vostok Ice Core

Vostok Station is centrally located on the East Antarctic Ice Sheet (78°27'51" S 106°51'57" E; altitude 3,448 m a.s.l.) and sits over ~3,743 m of ice. Nearly 20 yr after Vostok Station was constructed for the purpose of paleoclimatic ice-core research, airborne radio-echo surveys and satellite images revealed that a

very large lake (Subglacial Lake Vostok) exists beneath the ice sheet (Kapitsa et al., 1996; Studinger et al., 2004) (Figure 5). A 3,623-m core (designated 5 G) was recovered from the ice sheet by an international drilling team in 1998 before drilling was terminated ~120 m from the ice-water interface to prevent contamination of the lake environment.

The upper 3310 m of the Vostok ice core has provided a detailed paleoclimate record spanning the past 420 ka (Petit et al., 1999). However, the deepest portion of the ice core (3539–3623 m) has a chemistry, isotopic composition, and crystallography distinctly different from the overlying glacial ice (Jouzel et al., 1999; Figure 6). Geochemical and physical data from this deep ice indicate that it originated from the freezing (accretion) of subglacial lake water to the underside of the ice sheet. Microbiological studies of the Vostok glacial and accreted ice have indicated that low, but detectable, concentrations of prokaryotic cells (34–430 cells ml⁻¹; Christner et al., 2006) and DNA are present (Priscu et al., 1999; Christner et al., 2001; Bulat et al., 2004), and a portion

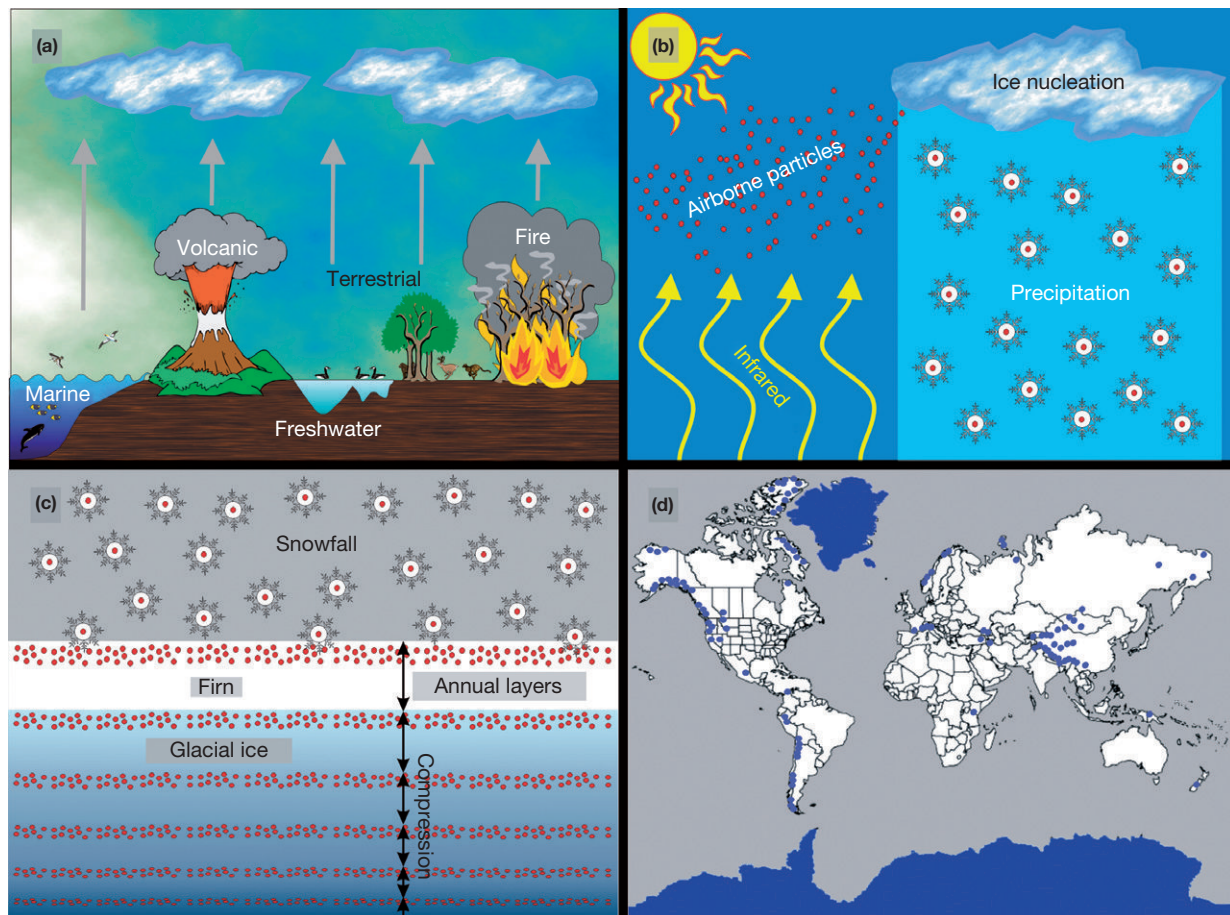


Figure 1 History of entrapped particles in glacial ice and the present-day distribution of glaciers. (a) Schematic illustrating the range of source environments that contribute particles to the atmosphere. (b) Advective currents, created by solar generated infrared radiation, inject surface derived aerosols high in the atmosphere. Such aerosols (red dots) may serve as primary ice nuclei in clouds, and are subsequently precipitated in snowfall or rain. (c) In geographical locations where the annual temperature remains cold enough for snowfall to accumulate annually, particles from the atmosphere are archived in a chronological sequence in firn and glacial ice. (d) Global locations of present-day ice sheets and mountain glaciers (in blue). Each glacial environment is unique, as the nearest ecosystems that would most likely contribute the majority of airborne biological particles are very different. Distribution data based in part on Satellite Image Atlas of Glaciers of the World (US Geological Survey, (2002) Satellite Image Atlas of Glaciers of the World).

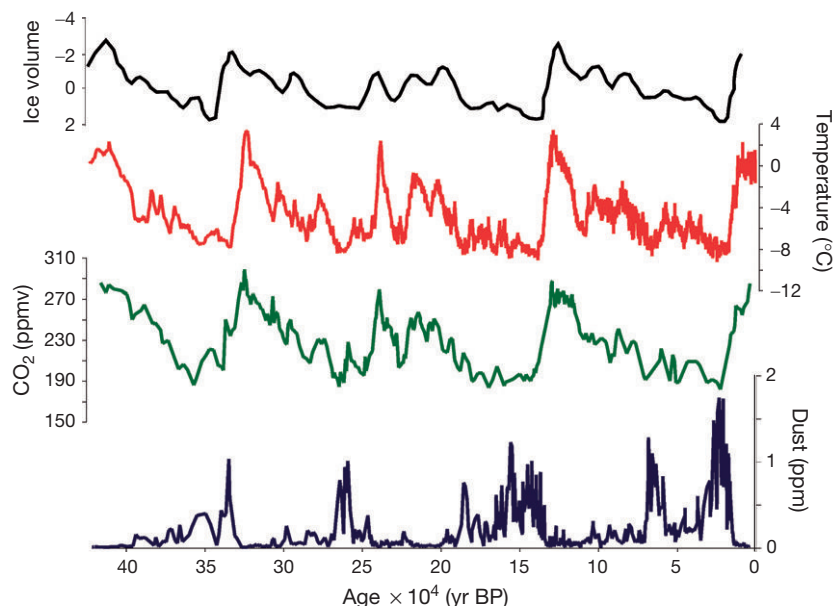


Figure 2 The climatic record over the last 420 ka deduced from the first 3310 m of the Vostok ice core (adapted from Petit et al. (1999)). From top to bottom: Global ice volume (black, in relative units, redrawn from the data of Petit et al. (1999)) as deduced from the marine sediment record; temperature (orange, difference with the present surface temperature) deduced from the stable isotope composition of the ice; records of CO₂ (green, ppmv) are deduced from entrapped air bubbles; profile of continental dust concentration (blue, ppm).

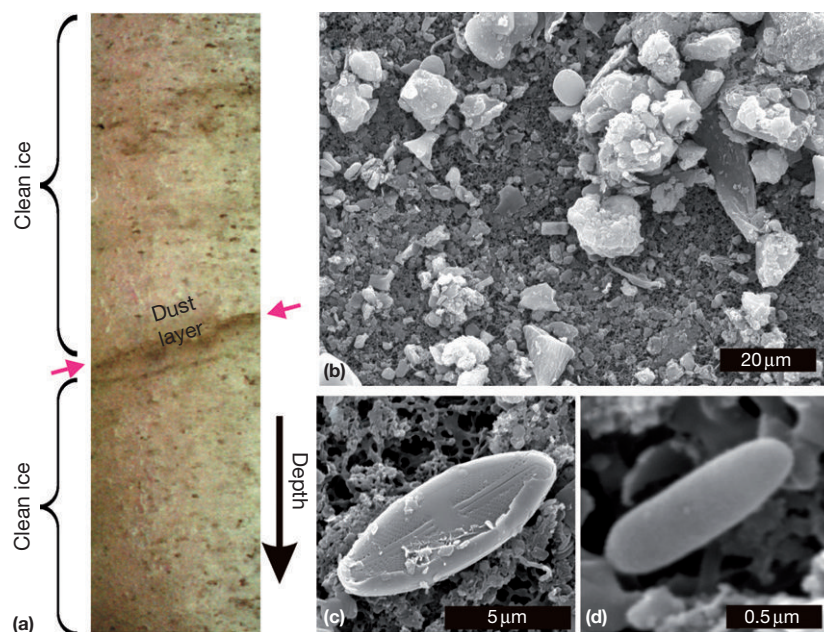


Figure 3 Dust layer, core particles, and associated microorganisms. (a) Cross-sectional view of ice core recovered 112 m (deposited, ~12 ka BP) below the surface from the Sajama ice cap, Bolivia. A large number of entrapped gas bubbles and macroscopic particles are visible. A prevalent dust layer is shown (pink arrows), characteristic of the increased concentrations of airborne particles associated with dry climatic conditions. (b) Low-resolution scanning electron micrograph illustrating the range of biological and inorganic particles entrapped in this ice-core sample. (c) Pennate diatom that most likely originated from one of the many saline lakes and salt flats in the area. (d) Rod-shaped bacteria preserved within the ice.

of the entrapped microbial assemblage is metabolically active when melted and exposed to nutrients (Karl et al., 1999; Christner et al., 2001). Many of the bacterial cells are associated with nonliving organic and inorganic particulate matter (Figure 7). Molecular identification of microbes within the accreted ice (by both culturing and direct 16S rDNA

amplification) show close agreement with present-day surface microbiota that classify within the phyla Proteobacteria (α , β , and γ), low G+C Gram-positives, Actinobacteria, and the Cytophaga-Flavobacterium-Bacteroides line of descent (Priscu et al., 1999; Christner et al., 2001; Bulat et al., 2004). Recent data from the Vostok ice core, using electron backscatter

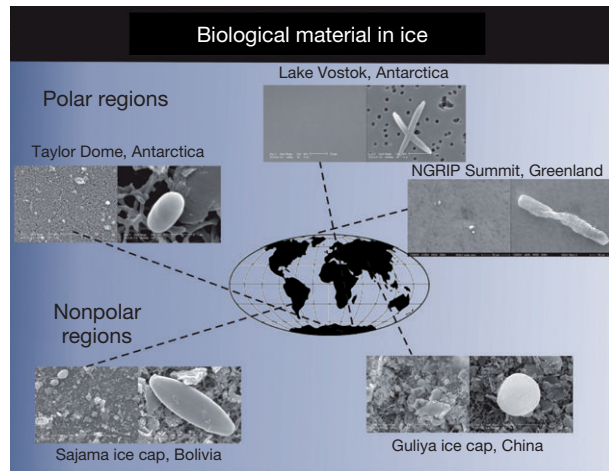


Figure 4 Scanning electron micrographs of particles from polar and nonpolar regions on Earth. Images in the left panel of each set are low-resolution micrographs ($1000\times$) illustrating the elevated concentrations of particles from nonpolar regions. The right panels show selected images of prokaryotic cells (Taylor Dome and Lake Vostok), an organic fiber (NGRIP), a diatom (Sajama), and a pollen grain (Guliya).

scanning electron microscopy, have also shown that up to 40% of the particles in glacial ice from the Vostok core represents nonliving biogenic matter; this value reaches 60% for accretion ice (Figure 8; Royston-Bishop; unpublished data). Dissolved organic carbon (DOC) in the Vostok core ranges from 0.2×10^{-5} M to 2.7×10^{-5} M in glacial ice and increases to 10.1×10^{-5} M in the underlying accretion ice. The dissolved and particulate organic fractions are positively correlated, inferring that these constituents covaried in the atmosphere when deposited on the ice sheet or were produced within the glacial ice by complimentary *in situ* metabolic activity following deposition. The positive relationship between these fractions in accretion ice could result from complimentary *in situ* metabolism in either the ice itself or within Lake Vostok before the ice accreted to the bottom of the ice sheet.

There has been speculation that geothermal input into Lake Vostok may fuel chemolithoautotrophic (organisms that use CO_2 as their carbon source and reduction/oxidation reactions for energy) microbial communities (Bulat et al., 2004). More importantly, it is still possible for a chemolithoautotrophic-based ecosystem to exist without invoking geothermal activity. A range of energy-rich reduced compounds (e.g., HS^- , S^0 , and Fe^{2+}) are supplied to the lake by the ice sheet itself, which alone can supply the energy and carbon required to support metabolic activity within the lake. Organisms trapped in the ice sheet can also provide the microbial seed to initiate the lake population (Priscu et al., 1999; Karl et al., 1999). Although geothermal input is not required to support life within the lake, it can provide an additional energy source for microbial growth in the lake if present. While other views of life in Lake Vostok exist, they all imply that microorganisms are present in this subglacial environment, despite high pressure, constant cold, low nutrient input, and an absence of sunlight (e.g., Bulat et al. (2004), Petit et al. (2005), Priscu et al. (1999), and Karl et al. (1999)). The exact nature of the biology in Lake Vostok awaits direct sampling of the lake water.

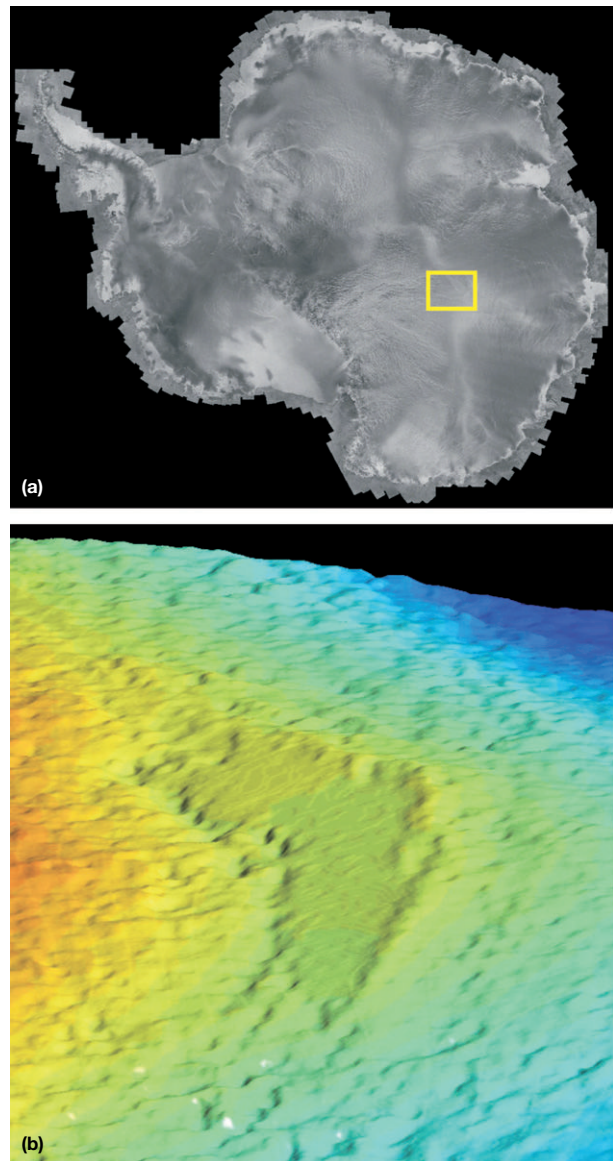


Figure 5 (a) Satellite image of Antarctica showing the location of Lake Vostok (yellow box). Digital mosaic compiled by the Canadian Space Agency (Alaska SAR Facility) using data from RADARSAT-1. (b) Image shows a perspective view of the ice surface above Lake Vostok compiled from ERS-1 radar altimeter data. Lake Vostok is the flat, featureless area formed where the glacial ice overrides the actual lake. Image courtesy of M. Studinger (Lamont Doherty Environmental Observatory, Columbia University, New York) using RAMP elevation data provided by the National Snow and Ice Data Center (NSIDC).

Ice Cores from Greenland

Biological studies of glacial ice cores have focused primarily on Antarctic cores and nonpolar, high-elevation glaciers (Priscu and Christner (2004) and references within). However, new studies on ice cores drilled in Greenland (i.e., GISP2, Gow et al. (1997), and NGRIP, Anderson et al. (2004)) have yielded important data on the nature of biogenic matter within and below the Greenland Ice Sheet. Culturable bacteria were recovered from the 'silty' ice in the basal portion of the GISP2 core,

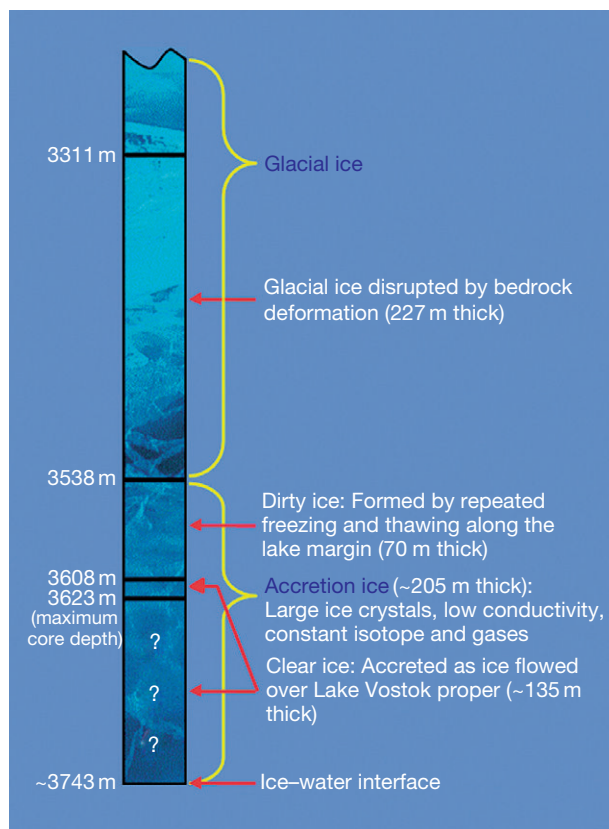


Figure 6 Schematic of the Vostok ice core showing the region of transition between glacial and accretion ice. The accretion ice is represented by (dirty) ice containing numerous sediment inclusions and clear ice, which was formed over the lake proper and is relatively free of sediment. The ice below 3623 m has recently been collected but has yet to be analysed; it is thought to have formed over the lake proper. See Jouzel et al. (1999) for details of the transition zone.

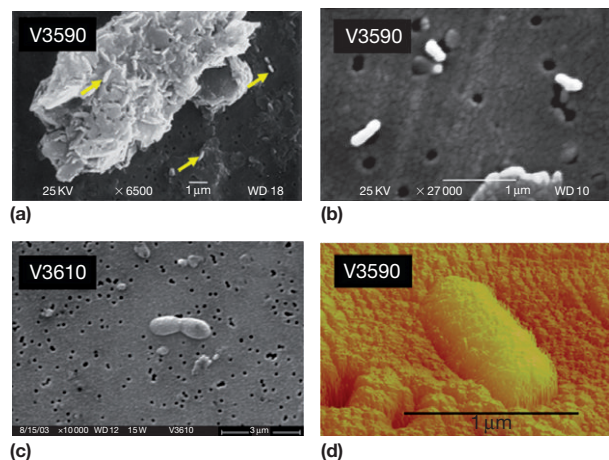


Figure 7 (a,b,c) Scanning electron and (d) atomic force microscope images of bacterial cells from Vostok accretion ice showing their close association with organic particles. The large particle in (a) was shown to be organic using backscattered electron imaging. The numbers on each panel refers to the depth from which the ice cores were collected. Samples for microscopy were prepared as outlined by Priscu et al. (1999). Arrows in 'a' denote the location of bacterial cells.

with reported total cell concentrations $>10^7$ cells ml^{-1} of meltwater (Sheridan et al., 2003; Miteva et al., 2004). The bacteria entrapped in the deepest sediment-laden portion of the GISP2 core could have originated from both the overlying 120 ka-old glacial ice and the underlying glacial till. Phylogenetic analysis of 16S rRNA gene sequences in the silty ice revealed that 36% of the bacterial isolates obtained were related to genera previously documented in glacial and permanently cold environments (i.e., *Methylobacterium*, *Rhodococcus*, *Mycobacterium*, *Sphingomonas*, *Arthrobacter*, and *Frigoribacterium*; Miteva et al. (2004)), providing support for the notion that members of these genera are particularly adept at surviving freezing and persist under cold and low or nongrowth conditions.

Refrozen, sediment-laden ice-core samples from the subglacial environment in Greenland were recently obtained from the NorthGRIP borehole when pressurized, 'pink'-colored basal water at the base of the ice sheet entered the lower 45 m of the borehole and froze (Anderson et al., 2004; Figure 9). Before this discovery, the existence of large amounts of water at the base of the Greenland Ice Sheet was not anticipated. Although cell concentrations in this refrozen basal water were 10 times greater than in the overlying glacial ice (basal ice = 1.6×10^3 cells ml^{-1} , glacial ice = 1.6×10^2 cells ml^{-1} ; Christner et al., 2006), the basal material was extensively contaminated with the hydrocarbon fluid used for ice-core drilling, making conclusions about the microbiology of the subglacial environment tentative. This subglacial debris contains minerals (e.g., sulfides and iron, which produced the pinkish color) and organic material from the bedrock, which could fuel microbially mediated chemical weathering reactions (e.g., Tranter et al. (2002)) and support a community isolated from direct input from the surface. The data collected so far show that Greenland, like Antarctica, contains microorganisms within glacial ice and in the liquid subsurface environment.

Temperate Mountain Glaciers

Glacial ice cores from nonpolar, low-latitude glaciers in the Andes, Himalayas, and New Zealand generally contain more recoverable bacteria (i.e., colony-forming units) and a greater variety of species than those from polar ice cores (Christner et al., 2000; Christner et al., 2003; Foght et al., 2004; Xiang et al., 2005). This geographic difference is consistent with the closer proximity of mountain glaciers to locations with substantial vegetation and exposed soils, which serve as major sources of atmospheric particles. Despite great differences in the environments contributing biological particles to polar and nonpolar glaciers, phylogenetic analysis based on the phylogeny of small subunit rRNA sequences reveals that many of the bacteria characterized from different geographical locations belong to the same genera and possibly the same species (Priscu and Christner, 2004). One striking result from biological studies of ice cores is that there is no consistent, monotonic decrease in the number of recoverable bacteria with increasing age of the ice core (Christner et al., 2000). Rather, the numbers of recoverable bacteria at different depths appear to be a reflection of the prevalent climate and individual events that occurred at the time of deposition.

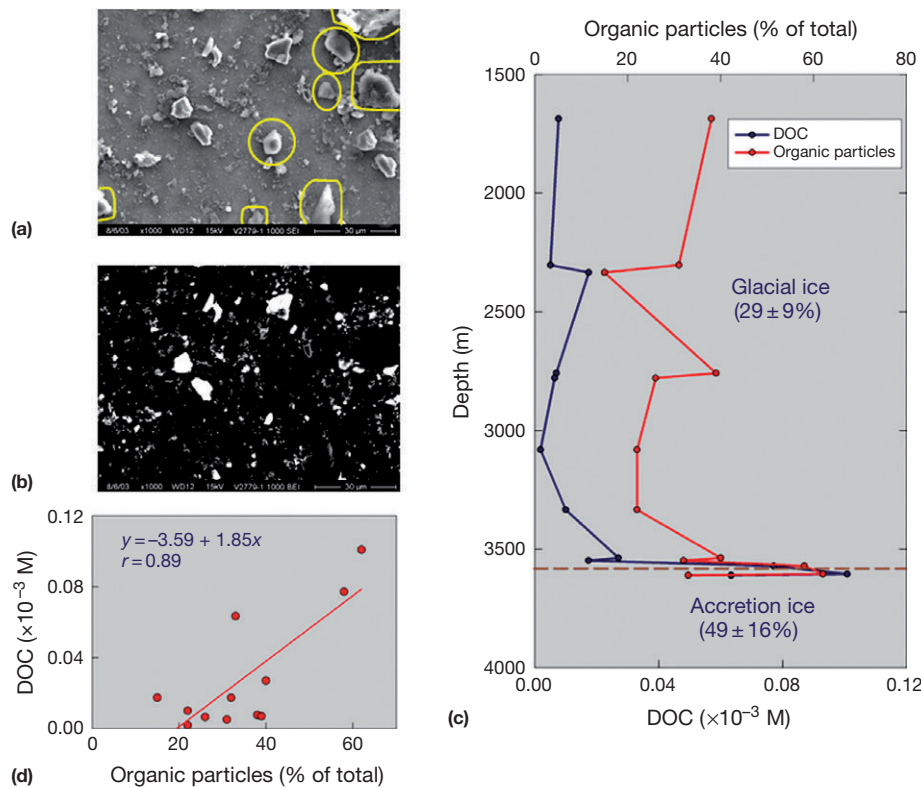


Figure 8 Scanning electron microscope images showing the relative amounts of particulate mineral and organic matter in a Vostok glacial ice core collected from 2779 m below the surface. Backscattered electron imaging was used to view the particles in panel (b); only mineral particles are detectable using electron backscatter. Organic particles not detected by electron backscatter are denoted by the yellow outlines in panel (a). The percentage of organic particles in ice cores below 1600 m, relative to the total particle count, is presented in panel (c) along with profiles of dissolved organic carbon (DOC). The positive relationship between particulate organic matter and DOC is shown in panel (d). DOC and scanning electron microscope methodology are described in [Priscu et al. \(1999\)](#).

Cells revived from ice-core samples have endured desiccation, high solar irradiation while at the surface, freezing, an extended period of frozen storage, and eventual thawing. Therefore, it is not surprising that a large number of the isolates recovered belong to bacterial groups that form spores (e.g., *Bacillus* and *Actinomyces*) or have thick cell walls and polysaccharide capsules. These structures would help overcome the stresses associated with water loss, namely, increased intracellular solute concentrations, decreased cell size, a weakened cell membrane, and physical cell rupture caused by freezing and thawing. The high frequency of pigment production in recovered isolates from temperate ice cores ([Christner et al., 2000](#)) is consistent with the need to absorb harmful solar irradiation, which can cause lethal DNA damage. Even though the surviving cells may have resistant structures and protective pigments, they must still incur some radiation and chemical damage during extended periods of inactivity. Long periods (30–180 days) of incubation were often necessary before visible colonies appear during culture efforts, yet the vast majority of the isolates could subsequently be subcultured to the stationary phase in <1 week. This observation indicates that time is required to repair accumulated cellular damage before growth and reproduction could begin.

A significant number of bacterial isolates characterized from a global range of ice cores are phylogenetically related to species recovered from Antarctic lake mats, sea ice, polar ice

cores, and other predominantly cold environments ([Priscu and Christner, 2004](#)). In terms of geographical differences between ice-core sites, *Bacillus* and *Paenibacillus* relatives of strains prevalent in soils were most commonly isolated from nonpolar glacial ices, and species of *Sphingomonas*, *Methylobacterium*, *Acinetobacter*, and *Arthrobacter* were ubiquitous and recovered from both polar and nonpolar ice-core locations. It is noteworthy that close relatives of the radiation-resistant type strains *Methylobacterium radiotolerans* and *Acinetobacter radioresistens* were also commonly isolated. The recovery of related microorganisms from many geographically diverse, but predominantly frozen environments, argues that these species probably have features that confer resistance to freezing and survival under frozen conditions.

Methods for Measuring Biotic Matter in Ice Cores

Particles in ice cores are typically measured on discrete core segments (usually at a depth resolution of several meters) using instruments based on the Coulter principle (e.g., [Delmonte et al. \(2004\)](#)). These investigations focused on the size, concentration, and source of the particles but failed to discriminate between biotic and abiotic forms, collectively referring to them as ‘dust’. Most of the data on the abundance of microorganisms in ice cores are from light microscopy, epifluorescence

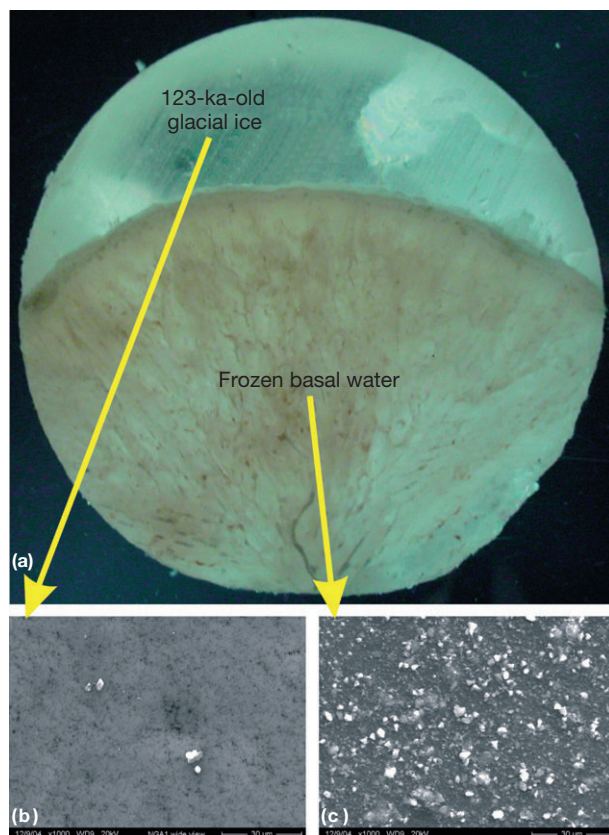


Figure 9 Images of an ice core collected at a depth of ~ 3310 m from the NGRIP (Greenland) drilling site (Anderson et al., 2004). The ice core contains both colored frozen basal water and glacial ice (a). Low-resolution scanning electron micrographs of particles in glacial ice and adjacent frozen basal water are shown in b and c, respectively.

microscopy, and electron microscopy. Although these microscopic methods allow morphological and physiological characterization of the microorganisms, they are tedious and subject to large errors, particularly when cell density is low. Hence, microscopic analysis does not lend itself to high resolution, routine determinations of biogenic matter in ice cores.

Borehole optical loggers have been used successfully to quantify the *in situ* abundance of dust in several systems and modifications are underway to incorporate filter fluorimeters on the loggers to detect biota and biomolecules (Bay et al., 2005). Flow cytometers offer another promising method to detect and characterize biogenic matter rapidly in ice cores retrieved from selected sites. Flow cytometers were developed for use in the medical field to rapidly enumerate, characterize, and sort cells from blood and human tissues. More recently, they have been used to characterize viruses, bacteria, and phytoplankton in natural marine environments (e.g., Minor and Nallathamby (2004)). Flow cytometers examine particles in a focused stream of fluid directed individually in front of an optical beam. Morphological characterization and enumeration of the particles (biotic and abiotic) can be made by their light-scattering properties. The physiological properties of the microbial component of the 'dust' can be determined by either autofluorescence or fluorescence induced by the addition of specific fluorescence probes (Bay et al., 2005).

Autofluorescence or induced fluorescence allows microorganisms to be enumerated separately from abiotic particles, and can provide detailed physiological measurements such as DNA content, cell membrane integrity, and metabolic potential.

Our laboratory has begun to use flow cytometry to characterize particulate matter in ice cores from the Antarctic and Greenland. Importantly, these ice cores all followed strict preparation and decontamination protocols outlined by Christner et al. (2005) before cytometric analyses to ensure samples were not contaminated. A Microcyte™ flow cytometer was placed in a class 100 environmental chamber (less than 100 particles larger than $0.5 \mu\text{m}$ in each cubic foot of air space; equivalent to an ISO class 5 clean room), and calibrated with fluorescent beads of known size and fluorescence quantum yield. The fluorescent DNA probe Syto 60 (Molecular Probes Inc.) was added to melted ice cores, which were then incubated in the dark for 5 min before cytometric analysis. Forward light scatter was used to measure the size and abundance of the particles, while fluorescence counts discriminated the biotic from the abiotic particles in the samples. Initial tests on Vostok glacial (1686 and 2334 m) and accretion ice (3612 m) showed that counts of total particles and DNA-containing bacterial cells can be successfully determined (Figure 10). Data obtained from these samples by flow cytometry were verified by epifluorescence microscopy of samples stained with the DNA probe SYBR Gold and by scanning electron microscopy (Christner et al., 2005). Two previous studies have also employed flow cytometers to enumerate microbes in glacial ice cores. Karl et al. (1999) used a flow cytometer to show that there were $500\text{--}700 \text{ cells ml}^{-1}$ in Vostok accretion ice from 3,603 m, and Miteva et al. (2004) found concentrations of bacterial cells ranging from 6.1 to $9.1 \times 10^7 \text{ cells ml}^{-1}$ in ice cores taken from the GISP2 (Greenland) "silty" ice (~ 10 m above bed-rock). In both of these studies, as well as our own, small cells ($\sim 1 \mu\text{m}$) dominated the populations found within ice cores.

Metabolic and Evolutionary Potential of Ice-Bound Microorganisms

The question of whether the microorganisms in glacial ice are metabolically active has fundamental significance to both glacial microbiologists and geochemists. Microbiologists are most interested in the diversity, evolution, and survival mechanisms of microorganisms, whereas glacial geochemists are concerned that *in situ* microbial activity consumes or liberates trace gases, thereby affecting the absolute concentration and isotopic content of the gases under concern. If the gases within the ice core are significantly altered by biogenic activity, a reassessment of ice core paleogas records may be necessary (Campen et al., 2003; Sowers, 2001).

Priscu and Christner (2004) estimated that the polar ice sheets of Greenland and Antarctica contain a combined total of 9.6×10^{25} prokaryotic cells, which equates to 2.7×10^{-3} petagrams ($1 \text{ Pg} = 10^{15} \text{ g}$) of organic carbon. This value represents a previously unrecognized carbon pool and approaches the bacterial carbon contained in all surface freshwater on our planet (Whitman et al., 1998).

Paleoclimate studies have accurately dated layers within the ice caps of both Antarctica (Petit et al., 1999; EPICA, 2004) and

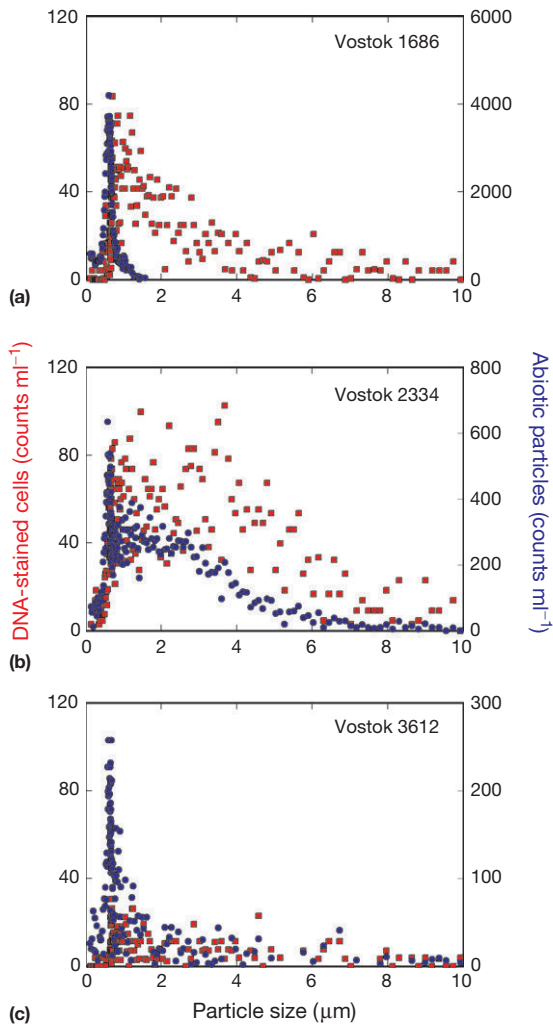


Figure 10 Flow cytometer data showing the relative abundance and size distribution of cellular and abiotic particles in Vostok glacial ice (a = 1686 m; b = 2334 m) and accretion ice (c = 3612 m). Cellular particles were characterized by their fluorescence induced with the DNA stain Syto 60. Red and blue symbols denote DNA-stained cells and abiotic particles, respectively.

Greenland (Alley, 2002), providing an important stratigraphic age record for these microbial ice repositories. This reservoir of ancient microorganisms, which can include fungi, bacteria, and viruses (Castello and Rogers, 2005), provides an unexplored frontier for the study of microbial variability over at least eight glacial cycles (~1 million yr). Data collected from selected habitats on our planet shows that microbial cells can be revived after extended entrapment in geological materials (Table 2). Willerslev et al. (2004a and b) further showed that DNA can remain viable for several million years in icy environments and concluded that natural ice provides an excellent repository for maintaining stable DNA.

A global assessment of prokaryotic diversity in glacial ice, based on 16S rRNA gene sequences, revealed that many of the isolates are related to psychrophilic (capable of growth at 5 °C or below, maximal temperature is approximately 15 °C and minimal temperature is 0 °C or lower) and psychrotolerant

Table 2 Reports of viable microorganisms retrieved from icy habitats representing a range of temporal isolation

Investigators	Location	Age (ka)
Sheridan et al. (2003); Miteva et al. (2004)	Glacial ice: GISP2, Greenland	120
Abyzov (1993); Christner et al. (2005)	Glacial ice: Vostok, Antarctica	420
Christner et al. (2003)	Glacial ice: Guliya, China	750
Shi et al. (1997)	Permafrost	3000

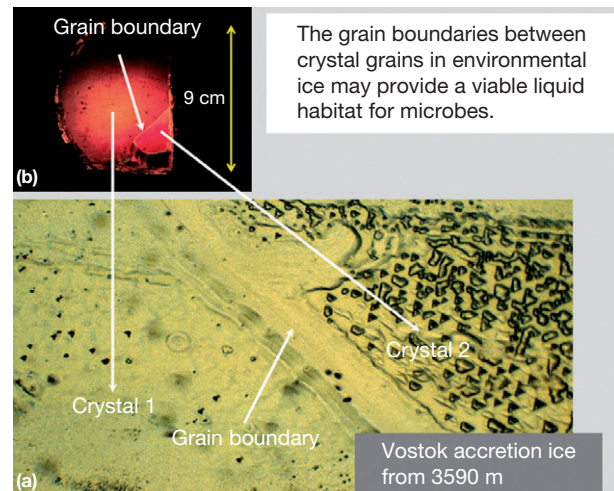


Figure 11 (a) Cross-polarized image of a Vostok accretion ice core from 3590 m below the surface showing the grain boundary between two crystals. (b) Microscopic view of the same vein following treatment with an epoxy resin, which highlights the grain boundary. The grain boundaries in polycrystalline ice, particularly where they form triple junctions, have been proposed as a site for life in solid ice (Price, 2000).

(capable of growing at 5 °C or below, with maximum growth temperature often between 25 °C and 30 °C) species originating from sites ranging from aquatic and marine ecosystems to terrestrial soils, with little in common except that all are permanently cold or frozen. The distribution of related species in geographically diverse glacial environments implies that members of these bacterial genera evolved under cold conditions and likely possess similar strategies to survive freezing and metabolize at low temperatures. This paradigm supports the contention of others (e.g., Price (2000), Campen et al. (2003), and Sowers (2001)) that the organisms are actually metabolizing in 'solid ice', presumably within grain boundaries (Figure 11). If microorganisms are indeed metabolizing and growing within the ice itself, a completely new set of selection pressures must be considered.

Rogers et al. (2004) postulated that selected pathogenic microbes survive and are recycled through ice in a process they call 'genome recycling' (Figure 12). Their contention is that organisms trapped in ice for hundreds of thousands to millions of years are eventually released when glaciers calve and the ice melts, allowing the flow of genetic material from trapped microorganisms to contemporary microbial populations. The mixing of ancient and modern genotypes may affect mutation rates, fitness, survival, pathogenicity, and other

characteristics through a change in allele proportions in the populations. Genome recycling is dependent on the revival and establishment of organisms once they emerge from the ice sheets and glaciers and transfer their genetic information to extant populations, which are then reincorporated into ice sheets and glaciers via eolian processes. Rogers et al. (2004) estimate that 10^{17} – 10^{21} viable microorganisms are released annually from melting glacial ice.

Castello et al. (2005) documented the recovery and identification of viable phage (a virus for which the natural host is a bacterial cell) from the bacterium *Bacillus subtilis*, and also have amplified genomic segments of tomato mosaic tobamovirus from ice cores assigned dates exceeding 100 ka BP. Our laboratory has

also observed free viral particles in both meteoric and accretion ice from the Vostok ice core (Figure 13), implying that viruses, like bacteria, are prevalent in glacial ice. These viral data, in concert with the concept of genome recycling proposed by Rogers et al. (2004), have led Smith et al. (2004) to postulate that glacial ice may provide a reservoir for pathogenic human viruses, particularly caliciviruses, influenza viruses, and enteroviruses. These viral groups occur in great abundance, are readily transported in the atmosphere, and may participate in ongoing disease cycles once released from the ice. Smith et al. (2004) contend that icy reservoirs may explain cyclic calicivirus events and the decades-long disappearances and subsequent reappearances of influenza-A subtypes. Although further research on ice cores is required, these preliminary data infer that viral reservoirs provided by glacial ice should be considered in eradication efforts for pathogenic human viruses.

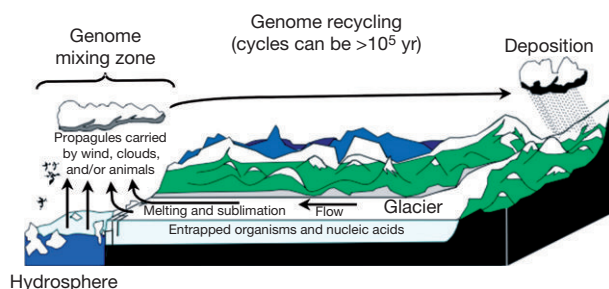


Figure 12 General scheme of genome recycling through glacial ice proposed by Rogers et al. (2004). Organisms immured in the ice for millennia are eventually released as the glaciers calve or melt, releasing them to the environment where their genes can mix with contemporary gene pools. Organisms forming the new gene pools are eventually trapped as dust particles in the ice where they remain isolated until the next cycle begins. Courtesy of S. Rogers, Bowling Green State University.

Conclusions

Glacial ice covers a large portion of Earth and contains information of past climate extending back to at least eight glacial cycles. In addition to important records of paleoclimate, recent investigations have shown that glacial ice contains an important record of microorganisms on our planet that may provide new information on biogeochemical processes, microbial evolution and physiology, and biomedical information on pathogens that occurred during past glacial and interglacial periods. Data on the biological nature of ice cores also provides us with an important Earthly analogue for the study of life on Mars and other icy worlds.

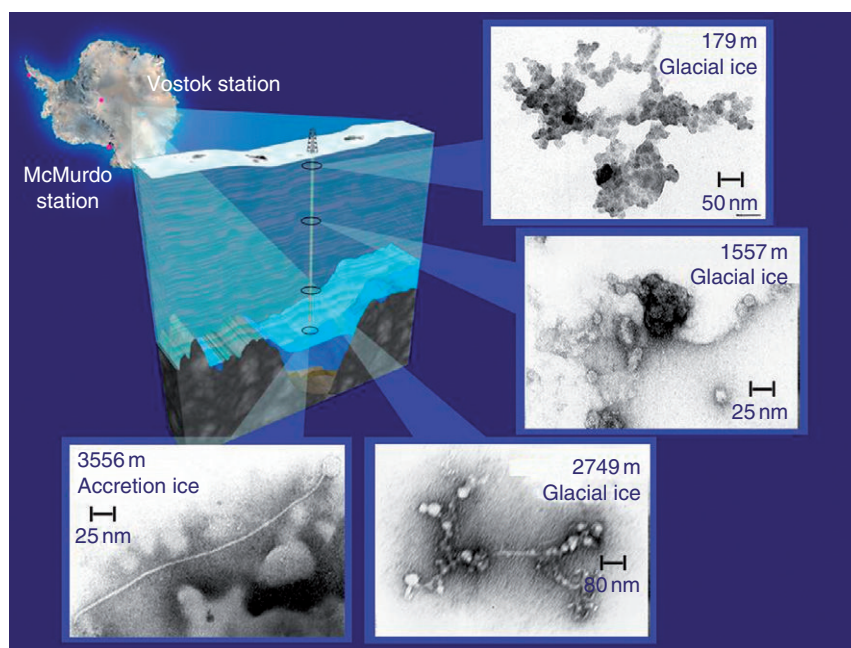


Figure 13 Viral particles observed in the Vostok ice core from selected depths in the glacial and accretion ice. All images were obtained by transmission electron microscopy. Courtesy of M. Young, Montana State University at Bozeman.

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See also: **Ice Core Methods: Overview.** **Ice Core Records:** Antarctic Stable Isotopes; Correlations Between Greenland and Antarctica; Greenland Stable Isotopes. **Ice Cores:** Dynamics of the East Antarctic Ice Sheet; Dynamics of the Greenland Ice Sheet; Dynamics of the West Antarctic Ice Sheet; History of Research, Greenland and Antarctica.

References

- Abyzov SS (1993) Microorganisms in the Antarctic ice. In: Friedmann EI (ed.) *Antarctic Microbiology*, pp. 265–295. New York: Wiley-Liss.
- Abyzov SS, Mitskevitch IN, Poglazova MN, et al. (2001) Microflora in the basal strata at Antarctic ice core above the Vostok lake. *Advances in Space Research* 28: 701–706.
- Alley RB (2002) *The Two-Mile Time Machine: Ice Cores, Abrupt Climate Change, and Our Future*. Princeton: Princeton University Press.
- Anderson KK, Azuma N, Barnola J-M, et al. (2004) High-resolution record of Northern Hemisphere climate extending into the last interglacial period. *Nature* 431: 47–51.
- Bay R, Bramall N, and Price PB (2005) Search for microbes and biogenic compounds in polar ice using fluorescence. In: Castello JD and Rogers SO (eds.) *Life in Ancient Ice*, pp. 268–276. New Jersey: Princeton University Press.
- Brook EJ (2005) Tiny bubbles tell all. *Science* 310: 1285–1287.
- Bulat SA, Alekhina IA, Blot M, et al. (2004) DNA signature of thermophilic bacteria from the aged accretion ice of Lake Vostok, Antarctica: Implications for searching for life in extreme icy environments. *International Journal of Astrobiology* 3: 1–12.
- Campan KR, Sowers T, and Alley RB (2003) Evidence for microbial consortia metabolizing within a low latitude mountain glacier. *Geology* 31: 231–234.
- Castello JD and Rogers SO (2005) *Life in Ancient Ice*. Princeton: Princeton University Press.
- Castello JD, Rogers SO, Smith JE, Starmar WT, and Zhao Y (2005) Plant and bacterial viruses in the Greenland Ice Sheet. In: Castello JD and Rogers SO (eds.) *Life in Ancient Ice*, pp. 196–207. New Jersey: Princeton University Press.
- Christner BC, Royston-Bishop G, Foreman CM, Arnold BR, Tranter M, Welch KA, Lyons WB, Tsapin AI, and Priscu JC (2006) Limnological Conditions in Subglacial Lake Vostok, Antarctica. *Limnology and Oceanography*. In press.
- Christner BC, Mikucki JA, Foreman CM, Denson J, and Priscu JC (2005) Glacial ice cores: A model system for developing extraterrestrial decontamination protocols. *Icarus* 174: 572–584.
- Christner BC, Mosley-Thompson E, Thompson LG, and Reeve JN (2001) Isolation of bacteria and 16S rDNAs from Lake Vostok accretion ice. *Environmental Microbiology* 3: 570–577.
- Christner BC, Mosley-Thompson E, Thompson LG, and Reeve JN (2003) Bacterial recovery from ancient ice. *Environmental Microbiology* 5: 433–436.
- Christner BC, Mosley-Thompson E, Thompson LG, Zagorodnov V, Sandman K, and Reeve JN (2000) Recovery and identification of viable bacteria immured in glacial ice. *Icarus* 144: 479–485.
- Delmonte B, Basile-Doelsch I, Petit JR, et al. (2004) Comparing the Epica and Vostok dust records during the last 220,000 years: Stratigraphical correlation and origin in glacial periods. *Earth-Science Reviews* 66: 63–87.
- EPICA community members (56 authors) (2004) Eight glacial cycles from an Antarctic ice core. *Nature* 429: 623–629.
- Foght J, Aislaikie J, Turner S, Brown CE, Rykurn J, Saul DJ, and Lawson W (2004) Culturable bacteria in subglacial sediments and ice from two southern hemisphere glaciers. *Microbial Ecology* 47: 329–340.
- Gow AJ, Meese DA, Alley RB, and Fitzpatrick JJ (1997) Physical and structural properties of the Greenland Ice Sheet Project 2 ice core: A review. *Journal of Geophysical Research* 102: 26559–26575.
- Griffin DW, Kellogg CA, Garrison VH, Lisle JT, Borden TC, and Shinn EA (2003) Atmospheric microbiology in the northern Caribbean during African dust events. *Aerobiology* 19: 143–157.
- Hambrey MJ and Alean J (2004) *Glaciers*, 2nd ed. Cambridge: Cambridge University Press.
- Jouzel J, Petit JR, Souchez R, et al. (1999) More than 200 meters of lake ice above subglacial Lake Vostok, Antarctica. *Science* 286: 2138–2141.
- Kapitsa AP, Ridley JK, Robin G, deQ Siegert MJ, and Zotikov IA (1996) A large deep freshwater lake beneath the ice of central East Antarctica. *Nature* 381: 684–686.
- Karl DM, Bird DF, Bjorkman K, Houlihan T, Shackelford R, and Tupas L (1999) Microorganisms in the accreted ice of Lake Vostok, Antarctica. *Science* 286: 2144–2147.
- Minor EC and Nallathambay PS (2004) Cellular vs. Detrital POM: A preliminary study using fluorescent stains, flow cytometry, and mass spectrometry. *Marine Chemistry* 92: 9–21.
- Miteva VI, Sheridan PP, and Brenchley JE (2004) Phylogenetic and physiological diversity of microorganisms isolated from a deep Greenland glacier ice core. *Applied and Environmental Microbiology* 70: 202–213.
- Paterson WSB (1994) *The Physics of Glaciers*, 3rd ed. Tarrytown: Elsevier.
- Petit JR, Alekhina I, and Bulat S (2005) Lake Vostok, Antarctica: Exploring a subglacial lake and searching for life in an extreme environment. In: Gargaud M, Barbier B, Martin H, and Reisse J (eds.) *Lectures in Astrobiology. Series: Advances in Astrobiology and Biogeophysics*, vol. 1, pp. 227–288. Berlin/Heidelberg: Springer.
- Petit J-R, Jouzel J, Raynaud D, et al. (1999) Climate and atmospheric history of the past 420,000 years from the Vostok ice core, Antarctica. *Nature* 399: 429–436.
- Price BP (2000) A habitat for psychrophiles in deep Antarctic ice. *Proceedings of the National Academy of Sciences* 97: 1247–1251.
- Priscu JC, Adams EE, Lyons WB, et al. (1999) Geomicrobiology of subglacial ice above Lake Vostok, Antarctica. *Science* 286: 2141–2144.
- Priscu JC and Christner BC (2004) Earth's icy biosphere. In: Bull AT (ed.) *Microbial Diversity and Bioprospecting*. American Society for Microbiology, pp. 130–145. DC: Washington.
- Priscu JC, Fritsen CH, Adams EE, et al. (1998) Perennial Antarctic lake ice: An oasis for life in a polar desert. *Science* 280: 2095–2098.
- Rogers SO, Starmar WT, and Castello JD (2004) Recycling of pathogenic microbes through survival in ice. *Medical Hypotheses* 63: 773–777.
- Sheridan PP, Miteva VI, and Brenchley JE (2003) Phylogenetic analysis of anaerobic psychrophilic enrichment cultures obtained from a Greenland glacier ice core. *Applied and Environmental Microbiology* 69: 2153–2160.
- Shi T, Reeves RH, Gilichinsky DA, and Friedmann EI (1997) Characterization of viable bacteria from Siberian permafrost by 16S rDNA sequencing. *Microbial Ecology* 33: 169–179.
- Siegenthaler U, Stocker TF, Monnin E, et al. (2005) Stable carbon cycle-climate relationship during the late Pleistocene. *Science* 310: 1313–1317.
- Smith AW, Skilling DE, Castello JD, and Rogers SO (2004) Ice as a reservoir for pathogenic human viruses: Specifically calciviruses, influenza viruses, and enteroviruses. *Medical Hypotheses* 63: 560–566.
- Sowers T (2001) The N₂O record spanning the penultimate deglaciation from the Vostok ice core. *Journal of Geographical Research* 106: 31903–31914.
- Spahni R, Chappellaz J, Stocker TF, et al. (2005) Atmospheric methane and nitrous oxide of the late Pleistocene from Antarctic ice cores. *Science* 310: 1317–1321.
- Studinger M, Bell RE, and Tikku AA (2004) Estimating the depth and shape of subglacial Lake Vostok's water cavity from aerogravity data. *Geophysical Research Letters* 31: 12401–12405.
- Tranter M, Sharp MJ, Lamb HR, Brown GH, Hubbard BP, and Willis IC (2002) Geochemical weathering at the bed of Haut Glacier d'Arolla, Switzerland- a new model. *Hydrological Processes* 16: 959–993.
- Tung HC, Bramall NE, and Price PB (2005) Microbial origin of excess methane in glacial ice and implications for life on Mars. *Proceedings of the National Academy of Sciences* 102: 18292–18296.
- Whitman WB, Coleman DC, and Wiebe WJ (1998) Prokaryotes: The unseen majority. *Proceedings of the National Academy of Sciences* 95: 6578–6583.
- Willerslev E, Hansen AJ, and Poinar HN (2004a) Isolation of nucleic acids and cultures from fossil ice and permafrost. *Trends in Ecology and Evolution* 19: 141–144.
- Willerslev E, Hansen AJ, Røn R, et al. (2004b) Long-term survival of bacterial DNA. *Current Biology* 14(No.1): R9–R10.
- Xiang S, Yao T, An L, Xu B, and Wang J (2005) 16S rRNA sequences and differences in bacteria isolated Muztag Ata glacier at increasing depths. *Applied and Environmental Microbiology* 71: 4619–4627.

Borehole Temperature Records

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General Concepts

Temperatures vary as a function of depth within the ice sheets. The temperature–depth profiles (symbolized $T(z)$, z being the common symbol for the vertical coordinate in glaciology) are controlled by surface climate, ice flow and thickness, and heat flow from the underlying crust (reviewed by Paterson (1994)). In the thick central regions of ice sheets, tens of millennia are required for the temperature at depth to adjust to major climate changes, including changes of surface temperature and snowfall rate. The $T(z)$, measurable as the temperature profile in boreholes, thus contains a legacy of climate history that is an important source of paleoclimate information. In particular, analyses of borehole temperatures yield the most convincing estimates for long-term averages of past surface temperatures at ice core sites, and hence for the magnitude of major transitions between persistent climate states (in particular the Wisconsin–Holocene transition, and the termination of the Holocene Climatic Optimum). The most comprehensive and accurate multimillennial reconstructions of temperature history in the late Quaternary are achieved by combining borehole temperature analyses with analyses of stable isotopes of ice and of gases (see review by Cuffey and Brook (2000)). Borehole analyses also yield an estimate for the geothermal flux, a parameter of great interest in studies of ice dynamics and tectonics.

The central reason for the value of the borehole thermometry method is that the measured variable, the temperature, is a direct remnant of the variable being reconstructed, not a proxy for it. The remarkable purity of the ice and absence of liquid flow at the deep core sites greatly assist the analyses; in this setting the heat flow problem is rather simple compared to that in most geologic media. The flow of the ice itself does introduce a major complication, but one that can be handled effectively and even used to advantage. Knowing the depth–age scale at the borehole site (from analyses on the ice core – see *Chronologies*) is the key to circumventing this problem, as it provides a strict constraint on the net displacement of ice layers over time (Cuffey et al., 1995). The ice flow then actually becomes beneficial because it increases the spatial wavelength of transient temperature signals, reducing their diffusional damping. Heat flow physics severely limits the time resolution of the borehole temperature reconstructions, and ensures that this resolution decreases drastically as one probes backward in time (Clow, 1992). The maximum possible time resolution varies substantially from site to site, but central Greenland provides a useful illustration of its general character; here the reconstructed temperature for ages of 0.2, 5, and 20 ka is an average over approximately 0.05, 1, and 10 ka, respectively.

As with many aspects of ice core analysis, borehole temperature reconstructions work best in the thick, cold, central ice divide regions of the ice sheets, where liquid water is negligible and horizontal flow is slow. Boreholes that are kilometers deep

are essential for gaining any information about the Wisconsin–Holocene transition. Analyses for sites with significant summer melt generation may be useful, but the impact of refreezing water on the energy budget confounds paleotemperature reconstruction (Paterson and Clarke, 1978).

Borehole temperature analyses specifically give information about the temperature of the ice sheet surface averaged over time. This is in contrast to the stable isotopes of ice, which are most strongly related to temperatures in the overlying atmosphere during condensation (see *Stable Isotopes*).

Heat Transfer in Ice

General Physics

The evolution of the $T(z)$ profile obeys the conservation of energy, and is calculated using the traditional heat equation (e.g., Paterson and Clarke (1978)). The rate of change of temperature over time at a given coordinate within the ice mass is governed by diffusion (the divergence of conductive heat fluxes), advection (heat transport by the moving ice), and heat generation due to the work of ice deformation (strain heating). The thermal parameters, the conductivity and heat capacity, are both temperature dependent (Yen, 1981). The thermal diffusivity for ice is approximately $10^{-6} \text{ m}^2 \text{ s}^{-1}$, implying the effective length scale for diffusive propagation of temperature signals over a time interval t is $0.001 t^{1/2}$, or roughly 200 m in 1 ka, and 600 m in 10 ka. The downward vertical ice flow governing heat advection is greatest at the surface and declines with depth in the ice sheet, to a value of zero at the bedrock interface. Characteristic values in the upper part of the ice column are comparable to the accumulation rate: decimeters per year in Greenland and West Antarctica, but only centimeters per year in dry East Antarctica (see *Dynamics of the Greenland Ice Sheet*, *Dynamics of the East Antarctic Ice Sheet* and *Dynamics of the West Antarctic Ice Sheet*). Advection and diffusion thus have comparable influence on the $T(z)$ in the former, but diffusion dominates $T(z)$ in the latter. The conditions in central Greenland and West Antarctica (both with ice 2–3 km in thickness) are fortuitously ideal for using borehole temperature analyses to probe the Wisconsin–Holocene transition and persistent trends within the Holocene.

Characteristics of Profiles

Ice sheets are effective thermal insulators, so the geothermal flux significantly warms the ice sheet bed relative to the surface. The temperature difference through the ice thickness is more than 50 °C in central East Antarctica, and more than 20 °C in central Greenland (Figure 1). In accumulation regions, the downward advection of cold surface ice reduces the temperature at all depths (Robin, 1955). In East Antarctica, this cooling

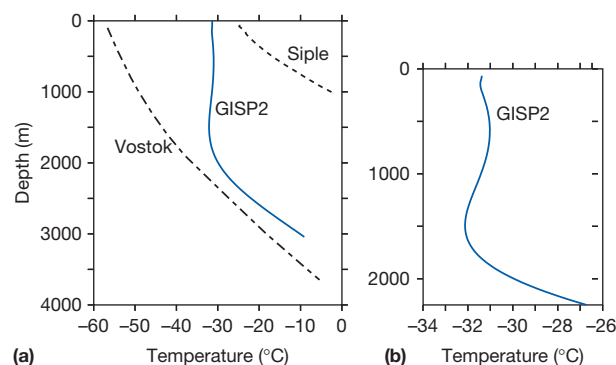


Figure 1 Measured borehole temperature profiles from Greenland and Antarctica: (a) The profiles through the entire ice thickness at three locations: Vostok Station, East Antarctica (Tsyganova and Salamatin, 2004), the GISP2 site, central Greenland (Clow et al., 1996), and Siple Dome, West Antarctica (Engelhardt, 2004). The GISP2 profile illustrates the strong cooling influence of vertical heat advection (see *Dynamics of the Greenland Ice Sheet*). Both Antarctic sites have much drier climate and the profiles are therefore conduction dominated (see *Dynamics of the East Antarctic Ice Sheet* and *Dynamics of the West Antarctic Ice Sheet*). Siple Dome is a warm site with relatively thin ice, located near the margin of the West Antarctic Ice Sheet, whereas Vostok is in the thick, cold interior of East Antarctica. (b) The upper portion of the GISP2 temperature profile, showing the remnants of past climates; cool ice centered at 1600 m depth reflects the last glacial climate, whereas warm ice centered near 700 m depth reflects the Holocene climatic optimum.

is quantitatively significant but does not cause the form of $T(z)$ to deviate much from the linear (Figure 1). In Greenland and West Antarctica, the advection is strong enough to make $T(z)$ strongly concave (Figure 1), with nearly isothermal ice in the upper half of the ice sheet above basal layers that warm strongly with increasing depth (Gow et al., 1968; Cuffey et al., 1995).

The paleoclimate signal in $T(z)$ is typically a small deviation of temperature from the steady profile, tenths of a degree Centigrade in magnitude. The borehole temperature measurements must therefore have high accuracy.

Measurement of the Temperature–Depth Profile

The thermal mass of fluid in a borehole is trivial compared to that of the surrounding ice sheet. Further, horizontal variations in temperature are negligible over kilometers. Thus, the temperature in a borehole accurately represents the temperature at same depth in the surrounding ice sheet, provided that the thermal disturbance from drilling has been allowed to decay, and that spontaneous convection of the borehole fluid is limited (Clow et al., 1996). In locations with very strong temperature gradients and relatively wide boreholes, it may be desirable to reduce such convection by inserting baffles.

The borehole temperature is measured using sensors, usually thermistors, suspended on a cable connected to a recording device at the surface. For the kilometers-long deep ice sheet boreholes, the great length of cable presents a major logistical challenge, as the winch system must be large and heavy, and hence difficult to deploy at remote locations. Proper circuit

design allows the electrical resistance of the long cable to be cancelled so that all of the instrumentation except the sensor can remain in a controlled environment at the surface, allowing for very accurate measurements. This is the design of the United States Geological Survey system (Clow et al., 1996), the best in the world, with an accuracy of more than 0.002°C . Less accurate, but nonetheless useful, measurements are made with circuits that attach to drill cables and use their communication wires to return numbers to the surface (e.g., Gow et al., 1968).

Where the fluid filling the borehole is liquid (as in all deep boreholes), the thermal contact between sensor and fluid is strong and the measurement can be made by steadily lowering the probe into the borehole, if a fast-response sensor is used. Otherwise, the sensor must be held at a given depth until equilibration with the surroundings, a time-consuming process.

The Inverse Problem

The goal of the ‘paleothermometry problem’ is to use borehole temperature measurements to infer the history of climatic surface temperature, effectively reversing nature’s heat flow process. This is a difficult problem of inverse theory (Clow et al., 1992; MacAyeal et al., 1993). One would like to simply use the mathematical description for temperature evolution and calculate backward in time to construct the answer. But this is essentially impossible because diffusion acts as an averaging process, so its reversal is unstable.

There are two distinct major problems. The first is that the time resolution of the answer is limited. The reconstructed temperature history cannot accurately represent the high-frequency content of the true climatic forcing, as the diffusion process removes this information by averaging. As an extreme example, this means that the method cannot reveal the magnitude of diurnal temperature cycles centuries or millennia in the past. More to the point, the method cannot be used to directly learn the magnitudes of the Dansgaard–Oeschger events (roughly a 1.5-ka period) or demonstrate that they are truly temperature changes. And for a climate transition (an abrupt increase in temperature, for example) the method cannot reveal its true rapidity; a warming spanning one decade and a warming spanning five centuries will be essentially indistinguishable a few millennia later. The second problem is that the inverse calculation can greatly magnify errors in the measurement of $T(z)$. Forcing a perfect match between model and measured $T(z)$ will always reconstruct a temperature history with unrealistic large and meaningless oscillations.

Successful temperature reconstructions use, in a variety of forms, one of two approaches for coping with the inherent difficulties of the analysis. First, one can impose limitations on the effective frequency content of the reconstruction, in accordance with the resolving power of the method for the available data at a given site. A simple way to impose such limitations is to reconstruct not a continuous temperature history but a small number of discrete temperatures representing mean values or extrema (e.g., Dahl-Jensen and Johnsen (1986)). The resolving power itself can be derived rigorously from numerical calculations of the influence of past impulsive

temperature forcings on the present $T(z)$ (MacAyeal et al., 1991; Clow 1992; Cuffey et al., 1995), but such results are seldom included in publications on borehole paleoclimatology. Second, one can force (MacAyeal et al., 1991) the temperature reconstruction to adopt some aspects of a specified climate history (called the preconception). This is referred to as ‘regularization.’ For example, the reconstruction can be made proportional to the ice isotopic time series (e.g., Cuffey et al., 1995 – see [Stable Isotopes](#)).

All methods for solving the inverse problem use the heat equation in conjunction with an ice flow model to calculate the modern $T(z)$ from a specified surface temperature history (this is the ‘forward problem’). This calculated $T(z)$ is then compared to the measured borehole temperatures, and the mismatch used to guide adjustments to the imposed surface temperature history. This process is repeated until an optimal surface temperature model is found. An ‘optimal’ model is one which minimizes a chosen performance index. This index contains a measure of the summed squares of mismatch between calculated and measured $T(z)$, and may include additional constraints (MacAyeal et al., 1991). The optimization itself can be achieved with a variety of methods, including iterative singular value decomposition, Levenberg–Marquardt minimization, and Monte Carlo analysis (the testing of a large number of possible solutions to compile a distribution of acceptable results). Application of control theory can maximize the efficiency of the optimization computation (MacAyeal et al., 1991).

The vertical heat advection is a crucial term in the forward calculations. Vertical flow rate is a function of ice sheet geometry, which evolves as a lagged function of changes in accumulation rate, ice margin position, and ice temperature (Paterson, 1994). The most complete analyses of borehole temperature are ones that perform simultaneous reconstructions of the surface temperature and accumulation rate histories, using borehole temperature data and the measured depth–age scale in conjunction with other information about ice sheet history (Cuffey et al., 1995; Cuffey and Clow, 1997).

Specific Strategies

Within this general framework, most borehole paleothermometry analyses use one of the following methodologies. Note that in all of these, the geothermal flux is found as part of the optimization, in addition to parameters characterizing the surface temperature history.

1. *Sparse reconstruction.* The reconstructed surface temperature history consists of a small number of optimized values, linked by simple interpolation to form a continuous function of time (e.g., Dahl-Jensen and Johnsen (1986) and Alley and Koci (1990)).
2. *Monte Carlo reconstruction.* A large number of temperature histories are found, by guided random perturbation, that predict the measured $T(z)$ to within uncertainties (Dahl-Jensen et al., 1998). Each of these histories consists of values at specified times, separated by an interval that increases back in time. The reconstructed surface temperature then consists of a frequency distribution of temperatures at each time.

3. *Spectral composite.* The reconstruction is a sum of periodic components, in which the contribution from higher frequencies is forced to decrease backward in time in accordance with the physical limitations on the resolution (Clow and Waddington, 1999).
4. *Isotopic calibration.* The reconstructed surface temperature history is proportional to the stable isotopic composition (see [Stable Isotopes](#)) of ice of a given age (Cuffey et al., 1995; Johnsen et al., 1995). In this case, the optimized parameters are not temperature values but rather sensitivity coefficients describing the magnitude of isotopic variation per variation in temperature (a single set of optimal values may be found for the entire history, or they may be allowed to change between specified intervals). The dominating influence of climatic temperature and its gradients on isotopic composition of polar precipitation justifies this method, *a priori*, on physical grounds; the isotopes are a strong correlate of temperature for thermodynamic reasons (Dansgaard, 1964; Cuffey and Brook, 2000). The connection of stable isotope records with temperature profiles was made first by Robin (1976), Johnsen (1977), and Paterson and Clarke (1978).

The risk with this method is that it may fail completely to provide an acceptable reconstruction if the isotopes are not in fact correlated with temperature. This has never proved to be a problem in the polar regions, however (Alley and Cuffey, 2001). The benefits of this method are considerable. It immediately generates an estimate for the climatic temperature variation at most frequencies, from periods of a few years to tens of millennia. The sensitivity coefficients themselves contain additional information about paleoclimate that is testable using climate models equipped with isotopic tracers. It should be remembered, however, that the method does not evaluate whether the high-frequency variations are accurately reconstructed or not; this must be done using some other method like gas isotopes (see [Thermal Diffusion Paleotemperature Records](#)).

Major Results

The achievements of borehole temperature analysis have increased significantly from the mid-1990s. Though valuable, older studies faced serious limitations (Weertman, 1968; Gow et al., 1968; Dahl-Jensen and Johnsen, 1986) because the ice core sites were chosen for logistical rather than scientific reasons, and because accuracy of the temperature measurements was not very high. This recency explains why no linked borehole–ice core studies comparable to those of central Greenland have yet been completed in central West Antarctica. Further, it was only in the early years of the twenty-first century that the gas isotopic methods for assessing rapid temperature changes became available (Severinghaus et al., 1998 – see [Thermal Diffusion Paleotemperature Records](#)). These provided compelling evidence that the borehole temperature-based isotope calibrations are quantitatively applicable to high frequency events.

Greenland

Analyses from central and southern Greenland (see [Greenland Stable Isotopes](#)) show that this region did experience a

Medieval Warm Period (culminating around AD 1000), a Little Ice Age cool period (AD 1500–1900), and warming to the mid-twentieth century (Alley and Koci, 1990; Cuffey et al., 1994; Dahl-Jensen et al., 1998). In contrast to some other regions of the Arctic, such as northern Alaska and Siberia, Greenland did not have renewed warming through the latter decades of the twentieth century. The Medieval Warm Period was approximately 1 °C warmer than present, and the Little Ice Age 0.6 °C cooler than present, in central Greenland.

These also reveal a persistent Climatic Optimum through the early–mid Holocene with temperatures at least 2 °C warmer than present (in central Greenland), terminating in cooling between 4 and 2 ka BP (Cuffey and Clow, 1997; Dahl-Jensen et al., 1998). All of these Holocene variations appear to have been larger in southern than in central Greenland (Dahl-Jensen et al., 1998).

The Wisconsin–Holocene deglacial transition was very large in central Greenland (Figure 2), at least 15 °C between the mean glacial climate and the mean Holocene, with the Last Glacial Maximum being at least 20 °C colder than present (Cuffey et al., 1995; Johnsen et al., 1995; Cuffey and Clow, 1997; Dahl-Jensen et al., 1998). Such extremely low surface temperatures appear to be consistent with the large expansion of continental ice sheets and sea ice extent during the glacial climate, coupled with the global reduction in atmospheric greenhouse gas concentrations.

The isotopic sensitivity across the deglacial transition was approximately 0.4 per mil $^{\circ}\text{C}^{-1}$ (when corrected for sea water composition changes), 0.25 per mil $^{\circ}\text{C}^{-1}$ across the mid–late Holocene cooling, and 0.5 per mil $^{\circ}\text{C}^{-1}$ at the termination of the Little Ice Age (Cuffey et al., 1995). These low values may be explained (see citations in Alley and Cuffey, 2001) by changes in the seasonal timing of precipitation, by the effects of evaporative recharge from nearby ocean surfaces, or by changes in

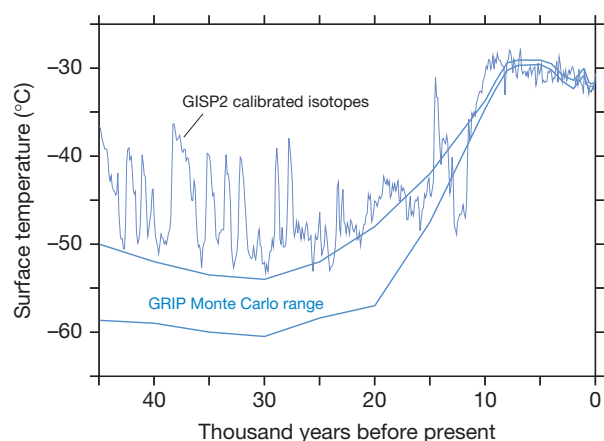


Figure 2 Reconstructed surface temperature history in central Greenland, from calibration of the isotopic record (see [Dynamics of the Greenland Ice Sheet](#)) and borehole temperature measurements at GISP2 (Cuffey et al., 1995), and the Monte Carlo reconstruction from borehole temperature measurements at GRIP (Dahl-Jensen et al., 1998), shown as the curves for plus/minus one standard deviation from the mean reconstructed temperature. The GRIP results are shifted to colder values for reasons that are not resolved but probably are related to a change in the deformation field deep in the ice beneath GRIP during the last deglacial transition (Marshall and Cuffey, 2000).

structure of cyclones. Some researchers have assumed that the isotopic sensitivity should be higher (>0.6 per mil $^{\circ}\text{C}^{-1}$) during the interglacial climate because there are fewer such correlated changes. This assumption is clearly contradicted by the available borehole temperature data (Cuffey et al., 1995; Dahl-Jensen et al., 1998).

The geothermal flux in central Greenland is approximately 50 mW m^{-2} (Dahl-Jensen et al., 1998).

West Antarctica

Analyses of a borehole at Siple Dome (Engelhardt, 2004) show that the geothermal flux is very high here ($\sim 70 \text{ mW m}^{-2}$). This is not surprising because the region is tectonically active, but it is very important because this heat source generates basal melt that lubricates the flow of ice streams draining West Antarctica (Paterson, 1994 – see [Dynamics of the West Antarctic Ice Sheet](#)).

East Antarctica

Temperature analyses for the deep boreholes in the dry interior of East Antarctica are challenged by the dominance of diffusional heat transport. The famous site at Vostok presents further challenges because it is not situated on an ice divide. Nonetheless, it has been shown here that the Wisconsin–Holocene deglacial surface temperature increase was greater than 10 °C and probably closer to 15 °C (Salamatin et al., 1998 – see [Antarctic Stable Isotopes](#)).

Holocene temperature changes have been examined using boreholes in thinner local flow centers close to the ice sheet margin. At Law Dome, the late Holocene features two thermal minima at AD 1250 and 1850, and a 0.7 °C temperature increase over the most recent 150 yr (Dahl-Jensen et al., 1999). At the dry Taylor Dome site, it appears that late Holocene temperature changes are anti-correlated with those from the Northern Hemisphere (Clow and Waddington, 1999).

Locations off the Main Ice Sheets

Borehole analyses from the much thinner ice masses apart from the great ice sheets have produced information about recent climate change. For example, a dramatic warming appears to have occurred in Svalbard during the twentieth century (Van de Wal et al., 2002).

See also: [Ice Core Methods: Chronologies](#); [Stable Isotopes](#). [Ice Core Records](#): [Antarctic Stable Isotopes](#); [Greenland Stable Isotopes](#); [Thermal Diffusion Paleotemperature Records](#). [Ice Cores](#): [Dynamics of the East Antarctic Ice Sheet](#); [Dynamics of the Greenland Ice Sheet](#); [Dynamics of the West Antarctic Ice Sheet](#).

References

- Alley RB and Koci BR (1990) Recent warming in central Greenland? *Annals of Glaciology* 14: 6–8.
- Alley RB and Cuffey KM (2001) Oxygen and Hydrogen-isotopic ratios of water in precipitation: Beyond paleothermometry. *Reviews in Mineralogy and Geochemistry* 43: 527–553.

- Clow GD (1992) The extent of temporal smearing in surface-temperature histories derived from borehole temperature measurements. *Global and Planetary Change* 98: 81–86.
- Clow GD, Saltus RW, and Waddington ED (1996) A new high-precision borehole-temperature logging system used at GISP2, Greenland, and Taylor Dome, Antarctica. *Journal of Glaciology* 42: 576–584.
- Clow GD and Waddington ED (1999) Quantification of natural climate variability in central Greenland and East Antarctica using borehole paleothermometry. *IUGG XXII General Assembly, abstract B249*.
- Cuffey KM, Alley RB, Grootes PM, Bolzan JM, and Anandakrishnan S (1994) Calibration of the delta 18-oxygen isotopic paleothermometer for central Greenland, using borehole temperatures. *Journal of Glaciology* 40: 341–349.
- Cuffey KM, Clow GD, Alley RB, Stuiver M, Waddington ED, and Saltus RW (1995) Large Arctic temperature change at the Wisconsin-Holocene glacial transition. *Science* 270: 455–458.
- Cuffey KM and Clow GD (1997) Temperature, accumulation, and ice sheet elevation through the last deglacial transition in central Greenland. *Journal of Geophysical Research* 102: 26383–26396.
- Cuffey KM and Brook EJ (2000) Ice sheets and the ice core record of climate change. In: Jacobsen MC, Charlson RJ, Rodhe H, and Orians GH (eds.) *Earth System Science: From Biogeochemical Cycles to Global Change*, pp. 459–497. Academic Press.
- Dahl-Jensen D and Johnsen SJ (1986) Paleotemperatures still exist in the Greenland Ice Sheet. *Nature* 320: 250–252.
- Dahl-Jensen D, Mosegaard K, Gundestrup N, et al. (1998) Past temperatures directly from the Greenland Ice Sheet. *Science* 282: 268–271.
- Dahl-Jensen D, Morgan VI, and Elcheikh A (1999) Monte Carlo inverse modelling of the Law Dome (Antarctica) temperature profile. *Annals of Glaciology* 29: 145–150.
- Dansgaard W (1964) Stable isotopes in precipitation. *Tellus* 16: 436–447.
- Engelhardt H (2004) Ice temperature and high geothermal flux at Siple Dome, West Antarctica, from borehole measurements. *Journal of Glaciology* 50: 251–256.
- Gow AJ, Ueda HT, and Garfield DE (1968) Antarctic ice sheet: Preliminary results of first core hole to bedrock. *Science* 161: 1011–1013.
- Johnsen SJ (1977) Stable isotope profiles compared with temperature profiles in firn with historical temperature records. *International Association of Hydrologic Science Publication* 118: 388–392. Symposium at Grenoble 1975 – Isotopes and Impurities in Snow and Ice.
- Johnsen SJ, Dahl-Jensen D, Dansgaard W, and Gundestrup NS (1995) Greenland paleotemperatures derived from GRIP bore hole temperature and ice core isotope profiles. *Tellus B* 47: 624–629.
- MacAyeal DR, Firestone J, and Waddington E (1991) Paleothermometry by control methods. *Journal of Glaciology* 37: 326–338.
- MacAyeal DR, Firestone J, and Waddington E (1993) Paleothermometry redux. *Journal of Glaciology* 39: 423–431.
- Marshall SJ and Cuffey KM (2000) Peregrinations of the Greenland Ice Sheet divide in the last glacial cycle: Implications for central Greenland ice cores. *Earth and Planetary Science Letters* 179: 73–90.
- Paterson WSB (1994) *The Physics of Glaciers*, 3rd edn. Pergamon.
- Paterson WSB and Clarke GKC (1978) Comparison of theoretical and observed temperature profiles in Devon Island ice cap, Canada. *Geophysical Journal of the Royal Astronomical Society* 55: 615–632.
- Robin GdQ (1955) Ice movement and temperature distribution in glaciers and ice sheets. *Journal of Glaciology* 2: 523–532.
- Robin GdQ (1976) Reconciliation of temperature-depth profiles in polar ice sheets with past surface temperatures deduced from oxygen-isotope profiles. *Journal of Glaciology* 16: 9–22.
- Salamatin AN, Lipenkov VY, Barkov NI, Jouzel J, Petit JR, and Raynaud D (1998) Ice core age dating and paleothermometer calibration based on isotope and temperature profiles from deep boreholes at Vostok Station (East Antarctica). *Journal of Geophysical Research* 103(D8): 8963–8977.
- Severinghaus JP, Sowers T, Brook ET, Alley RB, and Bender ML (1998) Timing of abrupt climate change at the end of the Younger Dryas interval from thermally fractionated gases in polar ice. *Nature* 391: 141–146.
- Tsyganova EA and Salamatin AN (2004) Non-stationary temperature field simulation along the ice flow line Ridge B – Vostok Station, East Antarctica. *Mater. Glyatsiol. Issled* 97: 57–70. [Data of Glaciological Studies].
- Van de Wal RSW, Mulvaney R, Isaksson E, et al. (2002) Reconstruction of the historical temperature trend from measurements in a medium-length borehole on the Lomonosovfonna plateau, Svalbard. *Annals of Glaciology* 35: 371–378.
- Weertman J (1968) Comparison between measured and theoretical temperature profiles of the Camp Century, Greenland, borehole. *Journal of Geophysical Research* 73: 2691–2700.
- Yen Y-C (1981) Review of thermal properties of snow, ice and sea ice. *Cold Regions Research Engineering Laboratory Report* 81–110.

Introduction

In the ideal case, the snow deposited on a glacier or ice sheet is buried by the following snowfall, building an undisturbed chronological precipitation record. After accessing this record by core drilling into this natural ice archive, one of the first and most important tasks is to establish the ice-core chronology as precisely as possible. How well this can be done depends strongly on the climatic and ice-flow conditions at the drilling site. In this article, only cold glaciers and ice sheets, with ice body temperatures constantly below freezing, are discussed. In these ice bodies, no mass is lost by internal melting. As ice is plastically deformed under its own weight, it is subjected to increasing thinning with depth and age (Figure 1). The upper 50–150 m of an ice sheet consists of consolidated snow, commonly referred to as ‘firn.’ Air is continuously trapped in the lower 10% of the firn (firn–ice transition), where interstitial pores around the firn grains are progressively closed off until the air is permanently isolated from the overlying atmosphere. Because the firn is permeable, the air above the firn–ice transition region can exchange with the atmosphere via diffusion and convection. As a consequence, the age of the air in a sample of an ice core differs from the age of the ice.

The main dating methods for ice cores are layer counting, ice-flow modeling, radioactive dating, and synchronization.

Layer Counting

Owing to the seasonal variation of temperature, sunlight, or atmospheric circulation, concentrations of many trace substances or isotopes show cyclic variations in ice cores (Figure 2). If these variations occur in a regular mode and are sufficiently resolved by measurements, annual layers can be counted back many thousands of years. Depending on the conditions in the ice (temperature, thinning of layers, etc.), variations are smoothed by diffusion or altered by chemical reactions. Diffusion is especially strong in the porous firn, where the transport is enhanced through the gas phase. The problem of diffusion, which is, for example, the predominant process in the case of water isotopes, can be corrected mathematically to a certain degree by a deconvolution algorithm (Johnsen, 1977), which extends the depth range for layer counting.

The classical method for tracing seasonal variations in ice cores is the ratio of the stable water isotopes $^{18}\text{O}/^{16}\text{O}$ and $^2\text{H}/^1\text{H}$. As this ratio is strongly affected by the condensation temperature, it reflects the seasonal variation of temperature. Seasonal variations are preserved beyond the firn–ice transition if the annual accumulation rate is at least 200 kg m^{-2} . In the consolidated ice, the self-diffusion of the water molecules is much smaller than diffusive transport in the firn, and seasonal variations in the water isotopes are preserved for many

thousands of years. Smoothing by self-diffusion is favored by thin annual layers (low accumulation rate, thinning by horizontal strain) and by high temperatures, for example, in the layers closer to bedrock, which are warmed by geothermal heat flux.

Trace substances are generally less affected by diffusion, especially when they are present in particulate form. Yet, there is always a lower limit of the annual accumulation rate, below which the annual variations are lost. One reason for this is that wind mixes the snow on the surface by scouring and relocation. Thus, annual layers are no longer preserved or are sometimes even missing. Another reason is that the annual variations are already smoothed to such a degree that when they are compacted to solid ice at the firn–ice transition, the deconvolution algorithm cannot reproduce an unequivocal record.

The best accuracy for layer counting is obtained by combining different tracers (Führer et al., 1993; Johnsen et al., 1992; Rasmussen et al., 2006). Ideally, the counting uncertainty is on the order of $\pm 1\%$.

Flow Models

Rheology of Ice

Ice is a non-Newtonian fluid with an empirical relation between strain rate $\dot{\epsilon}$ and stress τ , namely, $\dot{\epsilon} = A\tau^n$, known as Glen’s law (Glen, 1955). The coefficient A is temperature dependent, and the stress exponent n is on the order of 3 for intermediate stresses as they occur in glaciers. Given the knowledge of all material properties and boundary conditions, the plastic deformation of a glacier or ice sheet could be modeled. However, these boundary conditions and properties are generally poorly known. The bedrock topography is not resolved in sufficient detail, and past surface conditions (temperature, precipitation) can only be indirectly estimated. The same holds for the parameters that determine flow properties, such as ice temperature, crystal orientation, impurities, and basal sliding. Therefore, only rough ice chronologies can be constructed using only three-dimensional (3D) flow models.

For ice sheets, the so-called shallow ice approximation is often used: the ice flow is locally treated by simple shear, which allows calculation of horizontal flow velocities. With the assumption of a constant total ice thickness, the thinning function becomes very simple. The horizontal deformation profile has been approximated by Johnsen and Dansgaard (1992) with linear segments (Figure 3). The horizontal strain rate $\dot{\epsilon}_x = \dot{\epsilon}_{x0}$ is constant for the upper part of the ice sheet. Below is a shear layer of thickness h with horizontal strain rate $\dot{\epsilon}_x$ decreasing linearly, and the bottom of the ice sheet sliding with the velocity qv_{x0} , where v_{x0} is the horizontal flow velocity at the surface. The vertical strain ϵ_z is readily derived by observing continuity (x and z are horizontal and vertical coordinates,

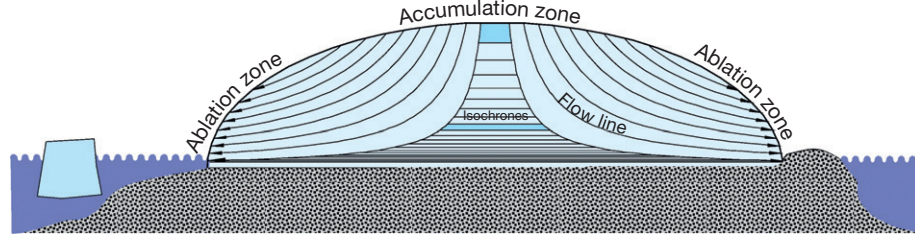


Figure 1 Schematic cross-section through an ice sheet. Layers are stretched and thinned when sinking. The two darker layers comprise roughly the same number of annual layers.

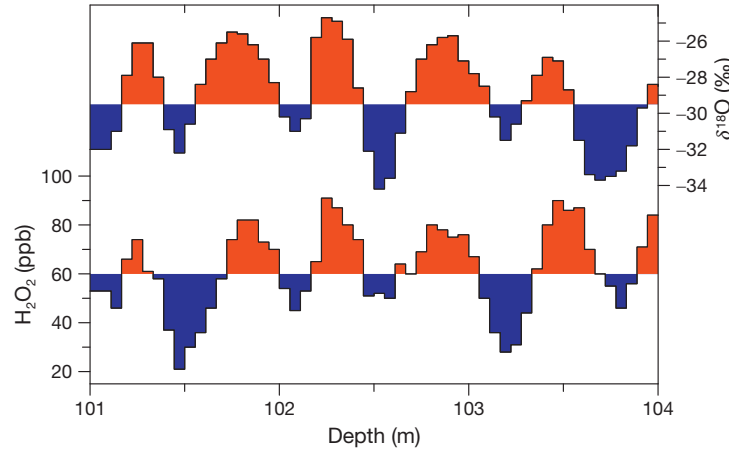


Figure 2 Annual variations of stable isotopes (temperature proxy) and chemical trace substances (here, e.g., hydrogen peroxide (sensitive to sunlight)). Summers: red, winters: blue.

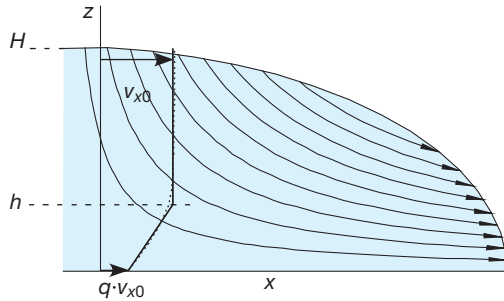


Figure 3 Johnsen–Dansgaard flow model: simplified horizontal flow velocity profile (thick solid line) for the location where the z -axis is drawn. A more realistic profile is indicated by the dotted line.

respectively, where z is zero at the bed rock and positive z point upwards):

$$\varepsilon_z = \begin{cases} 1 - k(H - z) & h \leq z \leq H \\ kz \left(q + \frac{1-q}{2h} z \right) & 0 < z < h \end{cases}$$

with

$$k = \frac{2}{2H - h(1 - q)}$$

where H is ice-sheet thickness (ice equivalent), and h is on the order of $H/3$ to $H/2$. The reported values for q are on the order of 0.15 (Johnsen and Dansgaard, 1992).

In the model, the ice becomes infinitely old at the base if no basal melting occurs, yet with annual layer thicknesses approaching zero. It is not known whether ice from the initial glaciation (30 Ma in Antarctica) is still present at the base of the large ice sheets, but two factors prevent the ice records from being extremely old: disordered ice flow due to roughness of bedrock topography, and basal melting. The oldest ice yet recovered from a continuous core is the one from Dome Concordia, Antarctica, with an estimated age of over 800 ka (European Project for Ice Coring in Antarctica (EPICA) Community Members, 2004).

Ice cores are often drilled on ice divides or local domes because of the relatively simple ice flow. At such sites, all ice retrieved in a core originates from snow deposited in the vicinity of the drill site. If the drill site is not located on an ice-divide, ice from deeper layers originates from snow that has fallen upstream of the drill site. Often, the simple approach can still be applied in such cases if one takes into account the upstream effect on the precipitation rate. In the case of long upstream trajectories and changing bedrock topography, the effect on thinning of the deeper layers can be handled only with a full flow model (Ritz et al., 2001). The same applies if variations in total ice thickness need to be taken into account. Yet, estimates show that the surface elevation of the polar plateaus has in general remained rather stable.

Special sites are the so-called domes, which are preferred drilling sites because of the expected large age of the ice near the bottom. These are places with zero surface velocity at the center of a divergent surface flow pattern.

Raymond Bumps

Underneath domes and divides, the effective deviatoric stresses are low and, because of the power-law dependence of strain with a stress exponent >1 , this situation results in a higher ice viscosity. Consequently, the deformation of the ice is slower, leading to upward arching of isochrones over the divide (**Figure 4**). These are the so-called Raymond bumps (**Raymond, 1983; Vaughan et al., 1999**). However, the importance of this effect depends on the thickness of the ice sheet, the accumulation rate, and the temperature (**Pettit and Waddington, 2003**). Raymond bumps are more important at divides of small ice caps, but less so for the large polar ice sheets.

Past Accumulation-Rate Estimates

If the thinning (vertical strain ε_z) at any depth is assessed with a model, then one can construct a chronology of the ice if past accumulation rates $a(z)$ are known:

$$\text{age}(z) = \int_z^H \frac{dz'}{\varepsilon_z' \cdot a(z')}$$

Past accumulation rates are often assumed to be proportional to the derivative of the mean saturation vapor pressure at the inversion layer with respect to temperature (**Jouzel et al., 1987**). The inversion layer temperature T_{inv} is related to the surface temperature T_s by $T_{\text{inv}} = T_{\text{inv}0} + c_1(T_s - T_{s0})$, where the index 0 refers to modern values. The mean surface temperature is estimated from the stable isotope ratio in the ice: $T_s = T_{s0} + c_2(\delta - \delta_0)$. c_1 and c_2 are constants and δ is the measured per mille deviation of the isotopic ratio $^2\text{H}/^1\text{H}$ or $^{18}\text{O}/^{16}\text{O}$ in the ice core relative to mean ocean-surface water, corrected for past changes in isotopic ratio and temperature. c_1 and c_2 result from calibrating the model with age reference points (**Schwander et al., 2001**).

An alternative approach for reconstructing paleo-accumulation rates is based on the concentration of trace substances in the ice. The net flux (atmosphere $\rightarrow$ ice) of any impurity is the product of the accumulation rate and the concentration in the ice. Past accumulation rates can thus be

computed by measuring the concentrations and estimating the past fluxes. By this means, **Wagner et al. (2001)** have assessed the accumulation rates for Central Greenland, using the cosmogenic radionuclides ^{10}Be and ^{36}Cl .

Optimized Combined Modeling

The estimate of past accumulation rates represents one of the largest uncertainties in ice-core chronologies. Other parameters that are often poorly constrained are basal melting, basal sliding, and upstream effects. In order to improve the accuracy of the age model, these poorly known parameters can be constrained by assigning a series of well-dated time horizons to corresponding depths. Such events include glacial terminations or minima of the Earth's magnetic field (see Section 'Age Horizons and Synchronization') that have been dated with best possible accuracy on other paleoarchives. It is sensible to assign the time horizons to within a certain tolerance (probabilistic approach). In order to guarantee a physically reasonable thinning and climatically realistic accumulation rate history, inverse modeling is used to match all time markers by searching the optimal parameters of combined ice flow and firn densification models. Mostly, this approach has been used for individual ice cores (**Parrenin et al., 2007**), with the consequence of leaving some discrepancies between different records, which must be considered carefully when investigating, for example, climatic phase lags between drilling sites. **Lemieux-Dudon et al. (2010)** have suggested an approach to reduce these chronology discrepancies by extending this method simultaneously to one Greenlandic (North Greenland Ice Core Project (NGRIP)) and four Antarctic ice cores (EPICA Dome C (EDC), EPICA Dronning Maud Land (EDML) Vostok).

Radioactive Dating

Small amounts of radioactive trace substances are incorporated in the ice, either as deposits in the snow or trapped in air bubbles during the transition from firn to ice. As ice is a very tight container for nearly all elements (exceptions are gases with small atomic radii, e.g., helium), inclusions in cold ice are extremely well preserved. Therefore, the age can be calculated from the radioactive decay law if the initial amount of radioactive material is known. Tracers used for radioactive dating of ice are given in **Table 1**.

Table 1 Radioactive isotopes used for dating ice cores

Isotope	Half-life (years)
^3H (tritium)	12.26
^{210}Pb	22.3
^{32}Si	172
^{39}Ar	269
^{85}Kr	10.77
^{14}C	5730
^{81}Kr	229 000
^{36}Cl	301 000
^{10}Be	1510 000
^{238}U	4 468 000 000

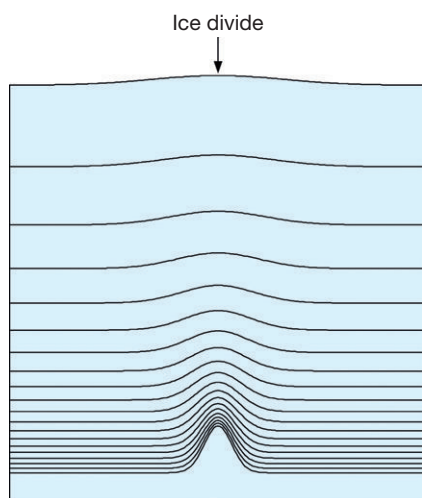


Figure 4 Raymond bump: isochrones arch up underneath ice divides and domes.

Fission Products

^{137}Cs , ^{90}Sr , ^{14}C , tritium, etc. resulting from nuclear weapons testing in the atmosphere during the 1950s and 1960s caused worldwide radioactive fallout and a twofold increase in atmospheric ^{14}C concentration, peaking in around 1963. This 'bomb peak' can be easily detected by total beta-activity measurements. It has often been used for a quick estimate of accumulation rates, allowing for a rough age scale for the firn layer.

^{210}Pb

^{210}Pb is a decay product of ^{222}Rn . Its half-life of 22 years makes it suitable to date relatively young ice. As the initial content of ^{210}Pb in snow is not very constant, the method is used only when no other dating options are available.

^{32}Si

A half-life of 172 years makes this isotope very suitable for dating historic age. But compared to other methods, it has never gained much importance because of its low and variable concentrations.

^{85}Kr , ^{39}Ar , ^{81}Kr

Gases are occluded in bubbles only at a certain depth, and the age of the air is less than the age of the ice. While ^{85}Kr has been used to date the air within the permeable firn layer (Schwander et al., 1993), ^{39}Ar would be appropriate to date the air below the firn-ice transition. However, its abundance is too small to obtain accurate ages of the air in the ice using conventional counting techniques (Loosli, 1983). This problem may be overcome by new techniques such as resonance ionization mass spectrometry or atom trap trace analysis, which could potentially extend the dating range for the occluded air using ^{81}Kr .

^{14}C

Ice contains carbon in the form of CO_2 in the bubbles and dissolved or particulate matter (dust, pollen). Because of the small amount of ^{14}C , only measurements with accelerator mass spectrometry (AMS) are feasible with the available amount of ice from core drilling. A few measurements have been done on 10-kg samples of 6–11-ka-old ice from the Dye3 core, with an accuracy of between 5 and 10% (Andree et al., 1986), which is significantly less than the accuracy of other methods.

Modern radiocarbon AMS dating can be done on very small samples, for example, a few hundred pollen grains (Vasil'chuk et al., 2005) and ultimately probably even on single larger pollen grains. The method has been successfully applied on organic carbon filtrates of as little as 5 μg from an Alpine ice core (Jenk et al., 2009).

For using ^{14}C for dating of ice, one must consider *in situ* production of ^{14}C by cosmogenic neutrons and muons, which interact with the oxygen of the water molecules (Lal et al., 1987; Van Der Kemp et al., 2002) yielding extra ^{14}CO and

$^{14}\text{CO}_2$. The amount of *in situ* produced ^{14}C depends on cosmic ray intensity, exposure time, as well as losses by diffusion and decay. The penetration depth of neutrons and muons in firn is a few meters and a few tens of meters, respectively. At low accumulation sites with firn depths of around 100 m, most *in situ* produced ^{14}C has time to escape to the atmosphere before bubbles are closed off (de Jong et al., 2004). Postcoring *in situ* production of ^{14}C may also be important, especially when ice cores are stored at high altitudes and latitudes, where the cosmic ray intensity is considerable.

^{10}Be and ^{36}Cl

These isotopes are produced by cosmic rays and modulated by solar activity and Earth's magnetic field. Except for the latest drilled ice cores from Antarctica, most ice cores that have been recovered so far are significantly younger than the half-life of these isotopes. Fluctuations in the initial contents limit the dating accuracy. Since both isotopes are produced and transported by similar mechanisms, one could expect that the initial $^{10}\text{Be}/^{36}\text{Cl}$ ratio should be rather constant, and that this ratio, with an apparent half-life of 376 ka, could be adequate for dating ice. However, even this ratio fluctuates by a factor of 2–5, probably because of the differential chemical behavior. Therefore, the main purpose of ^{10}Be and ^{36}Cl measurement on ice cores has been to study variations in solar activity and the Earth's magnetic field. However, in the case of disturbed layering of very old remnant ice near the bedrock, dating with these isotopes could provide rough estimates of the period during which ice was formed (Baumgartner et al., 1997).

Uranium Series

The measurement of ^{238}U and its daughter products has been used to determine the age of blue ice from Allen Hills (Antarctica; Goldstein et al., 2004), though with relatively large uncertainties. Daughter products in the ice originate predominantly from the recoiling out of dust particles into the ice after alpha decay. The amount of the daughters depends on the length of time the particles have been incorporated in the ice. Aciego et al. (2011) have significantly improved the method and validated it on well-dated ice samples from the Dome C ice core, and provided a first age estimate for the basal ice. The age range of the U-series method is limited to about 1 My, where the daughter products of ^{238}U approach equilibrium concentrations. The accuracy is about 10% for intermediate ages (200–500 ka).

Age Horizons and Synchronization

Since the uncertainty of direct absolute dating of ice is considerable (mainly due to the small amount of datable material), ice-core chronologies often rely on indirect dating, that is, by comparison with other better dated paleoarchives, including other ice cores. In the ideal case, the well-dated record is annually laminated (e.g., the marine sediments from the Cariaco basin (Hughen et al., 1998), or the lake sediments from Holzfelder Maar and Holzmaar (Zolitschka, 1998)). Unfortunately, these varved sediments are often affected by hiatuses or

do not extend to the present, but they can provide valuable information on the duration of climate events. Other archives are well suited to the application of absolute dating methods, such as U/Th-dated speleothems or corals.

The prerequisite for a correlation with ice-core records is that events occurring in the record are unequivocally assignable to a synchronous event in the dated record. Such events include volcanic eruptions, Milankovitch (orbital) variations, abrupt climate variations (Dansgaard–Oeschger (D–O) events), variations in the Earth’s magnetic field, variations in solar activity, or the impact of extraterrestrial objects.

Volcanic Eruptions

Acid precipitation (mainly sulfuric acid) from volcanic eruptions can easily be identified by electric conductivity measurements and ionic analyses. Large volcanic eruptions from the equatorial zone produce a global signal (e.g., Tambora, Indonesia, AD 1815; [Figure 5](#)). Volcanoes at higher latitudes usually produce only hemispheric signals (e.g., Laki, Iceland, AD 1873). Volcanic eruptions in the historic era have been used extensively to date younger ice with a 1-year accuracy and to determine mean modern accumulation rates ([Udisti et al., 2000](#)). Older volcanic eruptions are very useful for correlating different ice cores. However, the correlation between Northern and Southern ice cores has been successful only in a few cases, because correlation of the volcanic layers is often ambiguous. Potentially, ice cores can be synchronized by tephrochronology if volcanic layers can be attributed unequivocally to a specific eruption, for example, by its chemical signature ([Grönvold et al., 1995](#); [Narcisi et al., 2005](#)).

Tuning to Milankovitch Cycles

In the second half of the nineteenth century, James Croll proposed that the Pleistocene Ice Ages are caused by periodic changes in the Earth’s orbit around the Sun. These quasicyclic changes are a result of the gravitational interaction of the Earth with the Sun, the Moon, and the planets. Milutin Milankovitch improved the astronomical theory, leading eventually to the presently generally accepted causality between Earth’s orbital parameters and climate variations. Orbital computations have

been further developed, for example, by Berger and Laskar. The accuracy in absolute timing is about 1 ka in 1 My.

Tuning is accomplished by changing the chronology of a record such that its correlation with the target curve is increased (e.g., by maximizing the spectral power at the orbital bands). The target curve (e.g., total ice volume for tuning marine benthic $\delta^{18}\text{O}$ records) is calculated from the astronomical parameters (e.g., insolation at a given latitude) using a suitable model. A shortcoming of the method is the problem of over-tuning. For example, it has been shown that the spectral power at the Milankovitch band can be readily generated by tuning an arbitrary record of white noise. Recent orbital tuning work has therefore been as conservative as possible ([Karner et al., 2002](#); [Lisiecki and Raymo, 2005](#)). The uncertainty in the phasing between paleorecords and the astronomical cycles is on the order of 5 ka.

Ice-core records can be correlated to the Milankovitch cycles either directly or indirectly via synchronization, for example, with marine sediment records ([EPICA Community Members, 2004](#)). Direct tuning includes temperature-sensitive stable isotope ratios, the $\delta^{18}\text{O}$ of oxygen in the trapped air, and the O_2/N_2 ratio.

Stable isotope ratios (δD , $\delta^{18}\text{O}$) in precipitation reflect predominantly local temperature and are therefore directly climate sensitive ([Johnsen et al., 1995](#)). The oxygen in air bubbles ($\delta^{18}\text{O}_{\text{atm}}$) largely follows the $\delta^{18}\text{O}$ content of the ocean water and is thus a proxy for total ice volume, but it depends also on other aspects of the water cycle and biosphere productivity ([Landais et al., 2010](#)). The transfer from the ocean $\delta^{18}\text{O}$ to the air happens through the biosphere, with a time lag of about 2 ka and an offset of 23.5‰ (Dole effect). [Figure 6](#) shows the correlation between $\delta^{18}\text{O}_{\text{atm}}$ and Northern insolation ([Petit et al., 1999](#)). Because $\delta^{18}\text{O}_{\text{atm}}$ is a global parameter, it can also be used to synchronize the gas records of ice cores from both hemispheres. However, an even better candidate for this purpose is methane, because of its more distinct variations (see section ‘Abrupt Climate Variations’).

The O_2/N_2 ratio of the air extracted from the ice core is slightly less than the atmospheric ratio because O_2 is preferentially excluded at the firn–ice transition during the gas-trapping process. Additionally, there is usually some preferential loss of O_2 from the outer part of the ice core during storage. [Bender \(2002\)](#) has observed a high correlation between the O_2/N_2

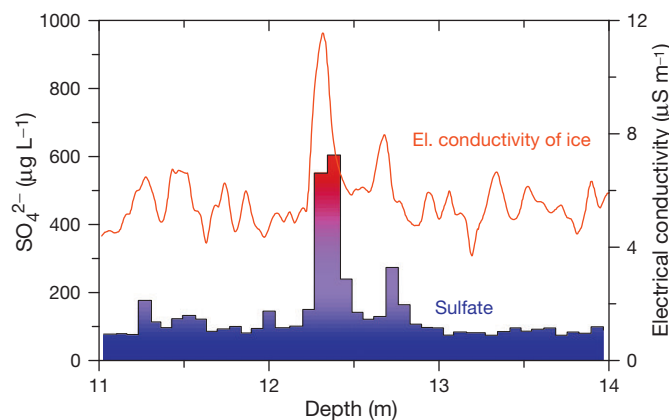


Figure 5 The Tambora AD 1815 eruption recorded in the ice core from Dome C, Antarctica.

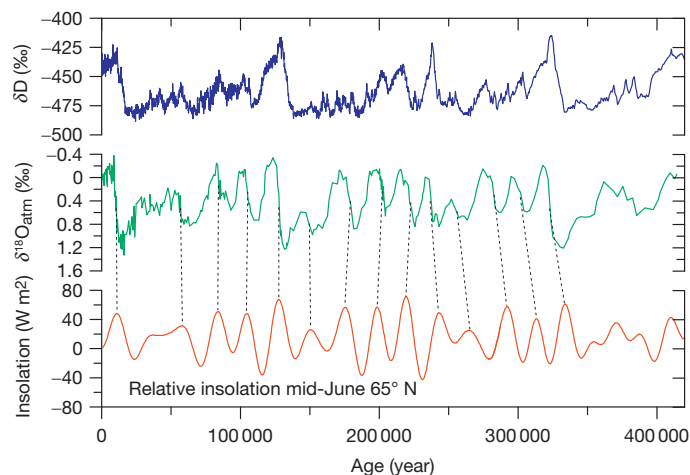


Figure 6 The $^{18}\text{O}/^{16}\text{O}$ isotope ratio of the air in the Vostok ice core correlating with the insolation at 65°N . The stable isotope ratio of the ice (δD) is a proxy for local temperature.

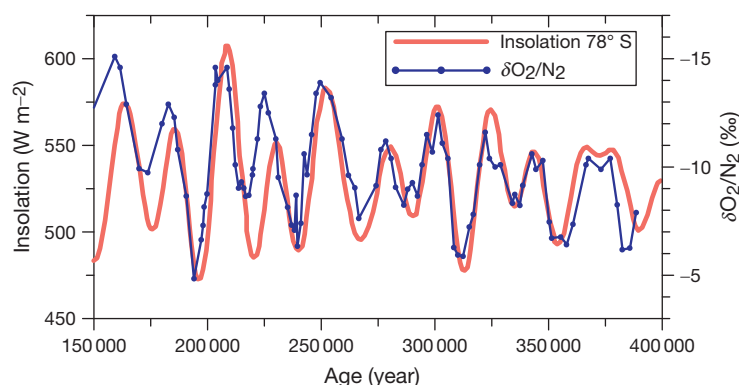


Figure 7 Correlation of the O_2/N_2 ratio with local insolation in the Vostok ice core.

ratio measured in the Vostok ice core and the local summer insolation (**Figure 7**). Bender assumes that the summer climate influences the physical properties of the firn grains, which eventually affects exclusion of O_2 at the firn-ice transition. Although the mechanism is not well understood yet, it is most promising because of the direct link with local insolation.

Abrupt Climate Variations

Abrupt warmings during glacial times (the so-called D–O events) are observed in the Greenlandic ice cores. Most probably, they reflect rapid changes in the thermohaline circulation, with reduced deep-water formation in the North Atlantic. Temperature increases of 10°C have occurred within decades. Such rapid events are therefore most suitable for synchronization with other paleorecords of similar sensitivity to these abrupt changes. Shackleton et al. (2004) have proposed a chronology for Greenland ice cores based on a correlation with planktonic $\delta^{18}\text{O}$ from a marine core taken from the Iberian margin off southern Portugal, which shows the same abrupt variations. D–O events are accompanied by abrupt variations of the atmospheric methane concentration. Since the mixing time of Earth's atmosphere is of the order of a few

years, these variations are observed in both hemispheres without considerable delay and are therefore ideal to synchronize the gas records from ice cores of the Northern and Southern Hemispheres (Blunier and Brook, 2001).

Variations of the Earth's Magnetic Field

The magnetic field of the Earth has undergone substantial variations in the past, including changes in overall strength, movement of the magnetic poles, and magnetic reversals. The Laschamp event, a prominent period with substantially reduced magnetic field intensity, centered around 41 ka BP. A weak magnetic field causes a lower shielding of the cosmic rays, thereby leading to an increase in cosmogenic production of radionuclides. Peaks in ^{10}Be and ^{36}Cl in polar ice cores have been found to coincide with the Laschamp event (Wagner et al., 2000).

Variations of the Solar Activity

The activity of the sun shows several quasicyclic variations. The most prominent one is the 11-year sunspot cycle (Schwabe cycle; half period of magnetic reversal) and the associated

22-year Hale cycle (full period of magnetic reversal). Other suggested cycles are Gleissberg (88 years), Suess (200 years), and Hallstatt (2.3 ka). There are also periods with exceptionally low solar activity (sunspot minima): these include the Dalton minimum (AD 1790–1820), the Maunder minimum (AD 1645–1715), the Spörer minimum (AD 1450–1550), and the Wolf minimum (AD 1280–1350). Changes in solar wind associated with changes in solar activity affect the magnetic field of the Earth and thus the production of cosmogenic radionuclides. The sunspot minima and the Schwabe cycle are clearly reflected in the ^{10}Be and ^{36}Cl content of ice-core records. Whether the 11-year cycle could eventually serve as a dating tool depends on the stability of its period, a question that is presently still open. An attempt to use the Schwabe cycle for dating has been made for the Tayler Dome ice core (Steig et al., 1998).

Impact of Extraterrestrial Objects

Impacts of large objects on the surface of the Earth are very distinct events and have the potential to be used as time horizons. Attempts have been made to find effects of the 1908 Tunguska Event. The term ‘Tunguska Event’ refers to a cosmic phenomenon observed on 30 June 1908 in Central Siberia over the Krasnoyarsk Territory, Irkutsk Region, when a large forest area was destroyed by the impact of a stellar object. The search for traces of iridium in a Greenland ice core from a large meteorite impact that may have caused the Tunguska event has failed (Rasmussen et al., 1995).

Age Scale for the Air Occluded in the Ice

The upper 50–120 m of an ice sheet is made of consolidated snow (firn) which is permeable to air (Figure 8). Air occluded at the firn–ice transition is therefore substantially younger than the ice. This age difference can be estimated with densification/diffusion models (Schwander et al., 1997). The air at the firn–ice transition is on the order of one to a few decades. The age of the ice at the firn–ice transition spans a larger range of a few tens to several thousand years under present conditions, mainly depending on temperature and accumulation rates. Locations with high accumulation rates and relatively high annual temperatures show the smallest ages, while sites in central Antarctica with accumulation rates of few centimeters per year show the largest ages. The difference between the age of the ice and air has not remained constant over time. Under glacial conditions, this age difference was generally several times larger than at present.

In ice-core records with sharp climate transitions (e.g., D–O events), the age model for the air can be verified by comparing $\delta^{15}\text{N}$ of the air with $\delta^{18}\text{O}$ or δD of the ice. Both $\delta^{15}\text{N}$ and the stable isotopes of the ice are sensitive to temperature. $\delta^{15}\text{N}$ is fractionated in the firn by thermal diffusion and records fast temperature changes with a short delay in the air trapped at the firn–ice transition, while the stable isotopes of the surface snow are direct proxies for the temperature. Thus, temperature changes are quasimultaneously recorded at two different depths. This depth offset can be reconstructed with the ice-thinning model for rapid climate changes in the past, and can

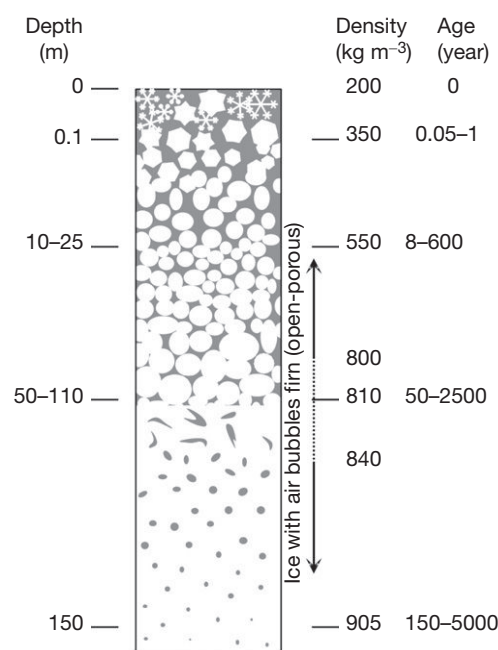


Figure 8 Firn layer of polar ice sheets: typical modern range of density and age for the top 150 m. Air bubbles are normally closed off in the density range 800–840 kg m⁻³.

be used to validate the firn models (Lang et al., 1999; Severinghaus and Brook, 1999).

See also: Dating Techniques; Geomagnetic Excursions and Secular Variations. **Cosmogenic Nuclide Dating:** Methods. **Glaciation, Causes:** Astronomical Theory of Paleoclimates. **Ice Core Methods:** ^{10}Be and Cosmogenic Radionuclides in Ice Cores; Conductivity Studies; Glaciochemistry; Methane Studies; Stable Isotopes; Studies of Firn Air. **Paleoclimate:** Timescales of Climate Change.

References

- Aciego SM, Bourdona B, Schwander J, Baur H, and Forierie A (2011) Toward a radiometric ice clock: Uranium ages of the dome C ice core. *Quaternary Science Reviews* 30(19–20): 2389–2397.
- Andree M, Beer J, Loetscher HP, et al. (1986) Dating polar ice by ^{14}C accelerator mass spectrometry. *Radiocarbon* 28: 417–423.
- Baumgartner S, Beer J, Suter M, et al. (1997) Chlorine 36 fallout in the Summit Greenland Ice Core Project ice core. *Journal of Geophysical Research* 102: 26659–26662.
- Bender ML (2002) Orbital tuning chronology for the Vostok climate record supported by trapped gas composition. *Earth and Planetary Science Letters* 204: 275–289.
- Blunier T and Brook EJ (2001) Timing of millennial-scale climate change in Antarctica and Greenland during the last glacial period. *Science* 291: 109–112.
- de Jong AFM, Alderliesten C, van der Borg K, van der Veen C, and van De Wal RSW (2004) Radiocarbon analysis of the EPICA Dome C ice core: No in situ C-14 from the firn observed. *Nuclear Instruments and Methods in Physics Research Section B: Beam Interactions with Materials and Atoms* 223: 516–520.
- EPICA Community Members (2004) Eight glacial cycles from an Antarctic ice core. *Nature* 429: 623–628.
- Fuhrer K, Neff A, Anklin M, and Maggi V (1993) Continuous measurements of hydrogen peroxide, formaldehyde, calcium and ammonium concentrations along the new GRIP ice core from Summit, Central Greenland. *Atmospheric Environment* 27A: 1873–1880.

- Glen JW (1955) The creep of polycrystalline ice. *Proceedings of the Royal Society of London Series A: Mathematical and Physical Sciences* 228: 519–538.
- Goldstein SJ, Murrell MI, Nishiizumi K, and Nunn AJ (2004) Uranium-series chronology and cosmogenic Be-10–Cl-36 record of Antarctic ice. *Chemical Geology* 204: 125–143.
- Grönvold K, Óskarsson N, Johnsen SJ, et al. (1995) Ash layers from Iceland in the Greenland GRIP ice core correlated with oceanic and land sediments. *Earth and Planetary Science Letters* 135: 149–155.
- Hughen KA, Overpeck JT, Lehmann SJ, et al. (1998) Deglacial changes in ocean circulation from an extended radiocarbon calibration. *Nature* 391: 65–68.
- Jenk TM, Szidat S, Bolius D, et al. (2009) A novel radiocarbon dating technique applied to an ice core from the Alps indicating late Pleistocene ages. *Journal of Geophysical Research-Atmospheres* 114: D14305.
- Johnsen SJ (1977) In: Proc. of Symp. on Isotopes and Impurities in Snow and Ice, Int. Ass. of Hydrol. Sci., Commission of Snow and Ice, International Union of Geodesy and Geophysics XVI, General Assembly, Grenoble Aug. Sept., 1975. International Association of Hydrological Sciences - Association Internationale des Sciences Hydrologiques (IAHS-AISH) Publication, pp. 210–219.
- Johnsen SJ, Clausen HB, Dansgaard W, Gundestrup NS, Hammer CU, and Tauber H (1995) The Eem stable isotope record along the GRIP ice core and its interpretation. *Quaternary Research* 43: 117–124.
- Johnsen SJ, Clausen HB, Dansgaard W, et al. (1992) Irregular glacial interstadials recorded in a new Greenland ice core. *Nature* 359: 311–313.
- Johnsen SJ and Dansgaard W (1992) In: Bard E and Broecker WS (eds.) *The Last Deglaciation: Absolute and Radiocarbon Chronologies*. NATO ASI Series, pp. 13–24. New York: Springer Verlag.
- Jouzel J, Lorius C, Petit JR, et al. (1987) Vostok ice core: A continuous isotope temperature record over the last climatic cycle (160,000 years). *Nature* 329: 403–408.
- Karner DB, Levine J, Medeiros BP, and Muller RA (2002) Constructing a stacked benthic delta O-18 record. *Paleoceanography* 17: 1–16.
- Lal D, Nishiizumi K, and Arnold JR (1987) In situ cosmogenic H-3, C-14, and Be-10 for determining the net accumulation and ablation rates of ice sheets. *Journal of Geophysical Research-Solid Earth and Planets* 92: 4947–4952.
- Landais A, Dreyfus G, Capron E, et al. (2010) What drives the millennial and orbital variations of delta O-18(atm)? *Quaternary Science Reviews* 29: 235–246.
- Lang C, Leuenberger M, Schwander J, and Johnsen S (1999) 16 °C rapid temperature variation in Central Greenland 70000 years ago. *Science* 286: 934–937.
- Lemieux-Dudon B, Blayo E, Petit JR, et al. (2010) Consistent dating for Antarctic and Greenland ice cores. *Quaternary Science Reviews* 29: 8–20.
- Lisiecki LE and Raymo ME (2005) A Pliocene–Pleistocene stack of 57 globally distributed benthic delta O-18 records. *Paleoceanography* 20(1–17): PA1003.
- Loosli HH (1983) A dating method with ³⁹Ar. *Earth and Planetary Science Letters* 63: 51–62.
- Narcisi B, Petit JR, Delmonte B, Basile-Doelsch I, and Maggi V (2005) Characteristics and sources of tephra layers in the EPICA-Dome C ice record (East Antarctica): Implications for past atmospheric circulation and ice core stratigraphic correlations. *Earth and Planetary Science Letters* 239: 253–265.
- Parrenin F, Barnola JM, Beer J, et al. (2007) The EDC3 chronology for the EPICA dome C ice core. *Climate of the Past* 3: 485–497.
- Petit JR, Jouzel J, Raynaud D, et al. (1999) Climate and atmospheric history of the past 420,000 years from the Vostok ice core, Antarctica. *Nature* 399: 429–436.
- Pettit EC and Waddington ED (2003) Ice flow at low deviatoric stress. *Journal of Glaciology* 49: 359–369.
- Rasmussen SO, Andersen KK, Svensson AM, et al. (2006) A new Greenland ice core chronology for the last glacial termination. *Journal of Geophysical Research-Atmospheres* 111: D06102. <http://dx.doi.org/10.1029/2005JD006079>.
- Rasmussen KL, Clausen HB, and Kallemeyn GW (1995) No iridium anomaly after the 1908 Tunguska impact: Evidence from a Greenland ice core. *Meteoritics* 30: 634–638.
- Raymond CF (1983) Deformation in the vicinity of ice divides. *Journal of Glaciology* 29: 357–373.
- Ritz C, Rommelaere V, and Dumas C (2001) Modeling the evolution of Antarctic ice sheet over the last 420,000 years: Implications for altitude changes in the Vostok region. *Journal of Geophysical Research-Atmospheres* 106: 31943–31964.
- Schwander J, Barnola J-M, Andrié C, et al. (1993) The age of the air in the firn and the ice at Summit Greenland. *Journal of Geophysical Research* 98: 2831–2838.
- Schwander J, Jouzel J, Hammer CU, Petit J-R, Udisti R, and Wolff E (2001) A tentative chronology for the EPICA Dome Concordia ice core. *Geophysical Research Letters* 28: 4243–4246.
- Schwander J, Sowers T, Barnola J-M, Blunier T, Malaizé B, and Fuchs A (1997) Age scale of the air in the summit ice: Implication for glacial–interglacial temperature change. *Journal of Geophysical Research* 102: 19483–19494.
- Severinghaus JP and Brook EJ (1999) Abrupt climate change at the end of the last glacial period inferred from trapped air in polar ice. *Science* 286: 930–934.
- Shackleton NJ, Fairbanks RG, Chiu TC, and Parrenin F (2004) Absolute calibration of the Greenland time scale: Implications for Antarctic time scales and for Delta C-14. *Quaternary Science Reviews* 23: 1513–1522.
- Steig EJ, Morse DL, Waddington ED, and Polissar PJ (1998) Using the sunspot cycle to date ice cores. *Geophysical Research Letters* 25: 163–166.
- Udisti R, Becagli S, Castellano E, et al. (2000) Holocene electrical and chemical measurements from the EPICA-Dome C ice core. *Annals of Glaciology* 30: 20–26.
- Van Der Kemp WJM, Alderliesten C, Van Der Borg K, et al. (2002) In situ produced C-14 by cosmic ray muons in ablating Antarctic ice. *Tellus Series B: Chemical and Physical Meteorology* 54: 186–192.
- Vasil'chuk A, Kim JC, and Vasil'chuk Y (2005) AMS C-14 dating of pollen concentrate from Late Pleistocene ice wedges from the Bison and Seyaha sites in Siberia. *Radiocarbon* 47: 243–256.
- Vaughan DG, Corr HFJ, Doake CSM, and Waddington ED (1999) Distortion of isochronous layers in ice revealed by ground-penetrating radar. *Nature* 398: 323–326.
- Wagner G, Beer J, Laj C, et al. (2000) Chlorine-36 evidence for the Mono Lake event in the Summit GRIP ice core. *Earth and Planetary Science Letters* 181: 1–6.
- Wagner G, Laj C, Beer J, et al. (2001) Reconstruction of the paleoaccumulation rate of central Greenland during the last 75 kyr using the cosmogenic radionuclides Cl-36 and Be-10 and geomagnetic field intensity data. *Earth and Planetary Science Letters* 193: 515–521.
- Zolitschka B (1998) A 14,000 year sediment yield record from western Germany based on annually laminated lake sediments. *Geomorphology* 22: 1–17.

Introduction

Analyses of air bubbles extracted from selected polar ice cores are an accurate and straightforward method of reconstructing the evolution of atmospheric CO₂ concentrations over the past few hundred thousand years. Only ice cores from locations where summer melting can be excluded are suited for such analyses. The reproducibility, determined by measuring the CO₂ concentration on closely neighboring samples of the same ice core, is between 1.5 and 3 ppm, depending on extraction and analytical methods, but also on origin and depth of the ice cores. However, the accuracy of the measurements could be slightly affected by the production or depletion of CO₂ in the air bubbles, as will be discussed in the last section. By comparing CO₂ records from ice cores from different drilling sites with different impurity concentrations and different surface temperature at the drilling site, it can be estimated that the uncertainty is below about 5 ppm. Meanwhile, progress in the analytical techniques to measure $\delta^{13}\text{C}$ of CO₂ has been made in recent years, reducing the analytical uncertainties to 0.05–0.07‰. $\delta^{13}\text{C}$ –CO₂ allows further investigations of the processes responsible for atmospheric CO₂ variations.

Antarctic ice core records presently reach an age of 800 000 years BP (before present, where present is normally defined as 1950) (Jouzel et al., 2007). Air gets enclosed in bubbles at the firn–ice transition at 70–120 m depth below the surface, depending on temperature and accumulation rates. This means that air in the bubbles is younger than the surrounding ice by a few decades for high accumulation drilling sites, and up to more than 5000 years for low accumulation drilling sites. For a given drilling site the age difference varies with temperature and accumulation rate changes in the past. Locations, surface temperature, and accumulation rates for some ice core drilling sites are given in Table 1.

The enclosure of bubbles occurs gradually in a depth interval of about 10 m. The mixing of air in the firn and the gradual enclosure process are responsible for the fact that the air in an ice sample does not have a well-defined age, but rather represents a mixture of atmospheric air within a certain age range. The possible time resolution depends on temperature and especially on annual accumulation rates. In ice cores from high accumulation rate sites the time resolution is only a few years, but very long ice core records are only available from low accumulation drilling sites, where the time resolution is less precise. For the EPICA (European Project for Ice Coring in Antarctica) Dome C ice core it was shown that a deviation of a gas concentration lasting 150 years is still detectable but its amplitude is reduced to about 50% of the original. Indirect CO₂ measurements from $\delta^{13}\text{C}$ in peat and stomatal density and stomatal index measurements on fossil leaves have a better time resolution and suggest century or even decadal variations of atmospheric CO₂ concentration in certain time intervals.

However, it was shown that despite the limited time resolution of the Dome C record, it would still detect the reported short-term fluctuations of the stomatal records (Blunier et al., 2005). The discrepancy can, therefore, not be explained by the time resolution of ice core CO₂ records.

In this article, CO₂ records of different time intervals, starting with the last millennium, are presented and discussed in individual sections. Where available, results from CO₂ isotope measurements ($\delta^{13}\text{C}$) are also included. Analytical methods and possible natural artifacts are discussed in the last two sections. Most of the CO₂ data presented in the different figures can be found on the website of the World Data Centre for Paleoclimatology: <http://www.ncdc.noaa.gov/paleo/data-list.html>. The CO₂ concentration is given in ppm (parts per million), which refers to the mole fraction of CO₂. $\delta^{13}\text{C}$ of CO₂ is reported as per mil values relative to Vienna Pee Dee Belemnite (the international reference material).

The Last Millennium

Continuous direct atmospheric CO₂ concentration measurements started in 1958. For earlier times, our knowledge about the development of the atmospheric concentration depends on the ice core archive and air in the firn, the top open porous layers of an ice sheet. The general trend based on various ice core records shows a consistent picture of an almost constant level of the concentration during the first eight centuries of the last millennium, of about 280 ppm, followed by the dramatic anthropogenic increase during the last two centuries. Ice core, firn air, and atmospheric measurements overlap smoothly as shown in Figure 1. Results from Greenland ice cores, shown with gray symbols, diverge from results from Antarctic ice cores with significantly higher values especially in the first six centuries.

The higher values are considered to be an artifact due to higher temperatures and higher impurity concentrations in Greenland ice cores. Due to these artifacts, only results from Antarctic ice cores are discussed in the following.

Law Dome firn and ice core measurements (Etheridge et al., 1996) cover the entire time period from AD 1000 and overlap with instrumental measurements. The reproducibility from these CO₂ results is 1.2 ppm, and the individual results represent a short-time interval due to the high accumulation rate at the coastal Law Dome. The record suggests an increase from about 279 ppm at the beginning of the millennium to about 282 ppm at AD 1100, relatively constant concentration until about AD 1600, followed by a decrease to a concentration of 277 ppm which lasted for about one and a half centuries. These natural variations are very important for our understanding of the carbon cycle. However, Law Dome is a coastal site with a relatively high surface temperature of –22 °C and the

Table 1 Characteristics of important ice core drilling sites

Drilling site	Coordinates	Surface temperature (°C)	Accumulation rate (kg m ⁻² year ⁻¹)
Byrd Station	79°59' S, 120°01' W	−28	160
Vostok	78°28' S, 106°48' E	−57	22
Law Dome (DSS)	66°46' S, 112°48' E	−22	600
Taylor Dome	77°48' S, 158°43' E	−42	70
EPICA Dome C	75°06' S, 123°21' E	−54.5	25
EPICA DML	75°00' S, 00°04' E	−44.6	64
Camp Century	77°11' N, 61°09' W	−24	340
GRIP Summit	72°34' N, 37°37' W	−32	200

DML, Dronning Maud Land; EPICA, European Project for Ice Coring in Antarctica; GRIP, Greenland Ice Core Project.

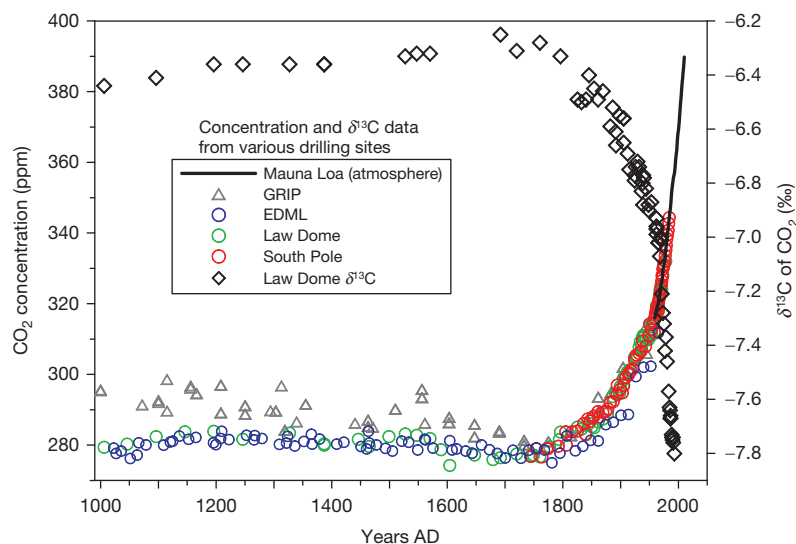


Figure 1 CO₂ concentrations and $\delta^{13}\text{C}$ of CO₂ from various Antarctic sites over the last millennium. The results from the Greenland ice core (gray triangles; Barnola et al., 1995) are considered to be too high due to the production of CO₂ in the ice. The solid line represents direct CO₂ concentration measurements on atmospheric air from the Mauna Loa (Tans and Keeling).

impurity concentrations are larger compared to drilling sites from the Antarctic Plateau. This and the fact that older, low-resolution Antarctic records were not able to confirm the small variations in the Law Dome record triggered remeasuring the last millennium CO₂ variations on an inland ice core. Recently, new detailed measurements along the EPICA Dronning Maud Land ice core confirmed the Law Dome record if the smoothing due to the lower accumulation rate is taken into account (Siegenthaler et al., 2005a; Figure 1).

It has been suggested that the lower values between AD 1600 and 1750 are related to the so-called Little Ice Age. However, it would be surprising if this climatic event, with a globally very small temperature anomaly, would have exerted such a distinct effect on global atmospheric CO₂ concentration. The Dronning Maud Land record strongly supports the variations suggested by the Law Dome record, but it does not constitute a final proof.

Over the past 200 years the $\delta^{13}\text{C}$ record of atmospheric CO₂ (Francey et al., 1999) is drifting to lighter values in parallel to the anthropogenic CO₂ increase. This reflects the input of isotopically light carbon into the atmosphere by the use of fossil fuel and deforestation with $\delta^{13}\text{C}$ values of about −30‰ and −25‰, respectively.

The Holocene

Two high-resolution CO₂ records exist that cover the entire Holocene. The Taylor Dome ice core from coastal Antarctica was drilled 1993/94. The Holocene section is found from about 120 to 350 m depth. A total of 417 samples from 69 different depth levels have been measured on this ice core (Indermühle et al., 1999). The Taylor Dome results have been confirmed by measurements along the EPICA Dome C ice core drilled on the East Antarctic plateau (Flückiger et al., 2002). A total of 700 samples from 117 different short-depth intervals have been measured between 99 and 416 m depth below surface. The two CO₂ records from Taylor Dome and Dome C are in good agreement, after the age scales are synchronized based on the CO₂ results (Monnin et al., 2004). Both records are shown versus the synchronized time scale (Figure 2).

The records show that the carbon cycle was not in a steady state during the Holocene. CO₂ concentrations decrease from 268 ppm at 10 500 years BP to 260 ppm at about 8200 years BP. The decrease was followed by an almost linear increase from 260 ppm to about 280 ppm at the beginning of the last millennium.

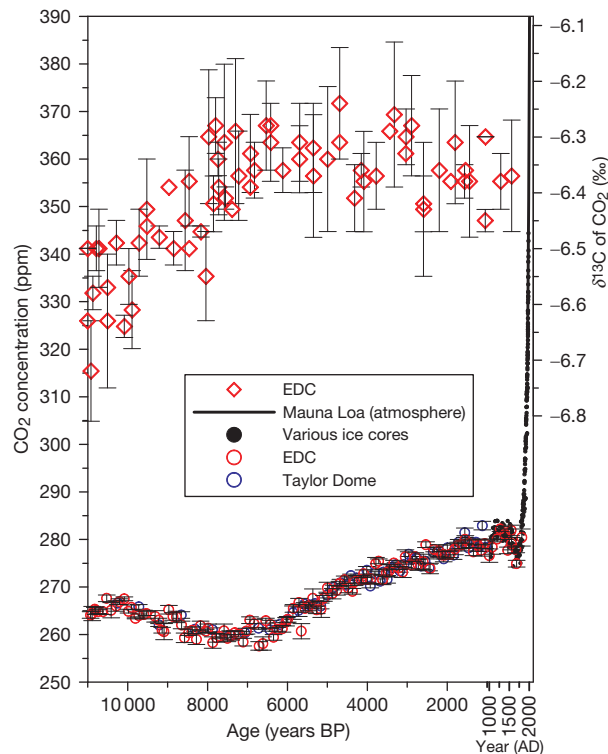


Figure 2 CO₂ data spanning the Holocene from Taylor Dome and Dome C and $\delta^{13}\text{C}$ data from Dome C. Data are plotted against the age scale of Dome C after the Taylor Dome record was synchronized with the one from Dome C, based on the CO₂ records (Monnin et al., 2004). The error bars characterize 1σ uncertainties. In black Antarctic CO₂ data over the last millennium from Figure 1 are plotted together with the atmospheric data from Mauna Loa. Note that other than in Figure 1 the $\delta^{13}\text{C}$ record stops before the anthropogenic change.

Without further information it is difficult to interpret the CO₂ record. Fortunately, it has been possible in recent years to finally obtain reliable $\delta^{13}\text{C}$ data over the Holocene (Elsig et al., 2009). The main features of the EPICA Dome C $\delta^{13}\text{C}$ record are an almost linear increase from -6.58‰ to -6.33‰ from 11 to 6 ka BP and a small decrease of 0.05‰ after 6 ka BP. This data revises earlier low-resolution data from Taylor Dome showing higher isotope variability over the Holocene.

The main sources and sinks of atmospheric CO₂ on a millennial time scale are the terrestrial biosphere and the ocean. Carbon in the terrestrial biosphere is significantly depleted in the heavy carbon isotope ^{13}C , while oceanic processes like carbonate compensation or coral reef growth have small fractionations versus atmosphere.

The increase of $\delta^{13}\text{C}$ to 8 ka BP can be explained by the growing terrestrial biosphere after the Ice Age. The Holocene record as a whole is seen as a combination of terrestrial uptake and changes in the marine carbonate cycle (Elsig et al., 2009).

It has been suggested that the CO₂ increase from 8 ka BP is related to emissions from early human activities related to land-use change. A release of terrestrial carbon should have led to a depletion of the atmospheric $\delta^{13}\text{C}$ values. However, the data reveals that this is not the case, strongly arguing against such a scenario (Stocker et al., 2011).

The Transition from the Last Glacial Epoch to the Holocene

The transition from the last glacial epoch to the Holocene (Termination I) was accompanied by an increase of atmospheric CO₂ concentrations of about 40%. This large, naturally induced CO₂ increase was observed by analyses of the first two ice cores reaching bedrock in Greenland and Antarctica: Camp Century and Byrd station. However, measurements on more recently drilled cores taken from more suitable drilling sites provide better core quality and improved analytical techniques. Thus, the newer sites have yielded more reliable and more detailed results concerning changes in atmospheric CO₂ concentrations during this transition.

Four hundred and thirty-two samples from 72 different short-depth intervals have been measured from the EPICA Dome C ice core, in a core interval between 350 and 580 m depth, covering the period 22 000–9000 years BP. The increase in CO₂ concentration from a mean value of 189 ppm between 18 100 and 17 000 years BP to a mean value of 265 ppm between 11 100 and 10 500 years BP, occurred in four clearly distinguishable intervals (Figure 3; Monnin et al., 2001).

The CO₂ increase starts at about 17 000 years BP and is almost linear in interval I. In interval II the rate of increase is reduced, but at the end of the interval a sudden increase of about 8 ppm is observed. During a cooling in Antarctica, the Antarctic cold reversal, the CO₂ concentration decreases slightly. In interval IV the increase continues at about the same rate as in interval I. At the end of the interval a sudden increase is observed again.

At the end of the last glacial epoch temperature increased in Antarctica and Greenland, but the magnitude and especially the evolution is quite different as shown (Figure 4). The CO₂ increase follows in its general trend more closely the temperature increase in Antarctica, suggesting that the Southern Ocean played a key role causing the CO₂ increase. However, the sudden increases at the end of interval II and IV coincide with sudden temperature increases in Greenland, indicating a significant influence of the northern hemisphere as well.

The start of the CO₂ increase is delayed by 800 ± 600 years (Fischer et al., 1999; Monnin et al., 2001), compared to the temperature increase at Dome C. This small delay, compared to the transition period of 6000 years, does not call into question the important amplification effect of the CO₂ increase for global warming during this climatic transition.

There are many suggestions for the possible causes of the CO₂ increase. The main suggestions are:

- Reduced solubility of CO₂ in the Southern Ocean surface waters due to a temperature increase synchronous with the one observed from Dome C.
- Enhanced air–sea exchange around Antarctica due to a reduced sea ice extent, leading to an increase in deep sea ventilation.
- Decreasing bioactivity in the Southern Ocean due to a reduction of aeolian dust deposition, leading to reduced iron fertilization.
- Effects in connection with the sudden temperature increases in the North Atlantic region connected with changes in deep water formation in the North Atlantic.

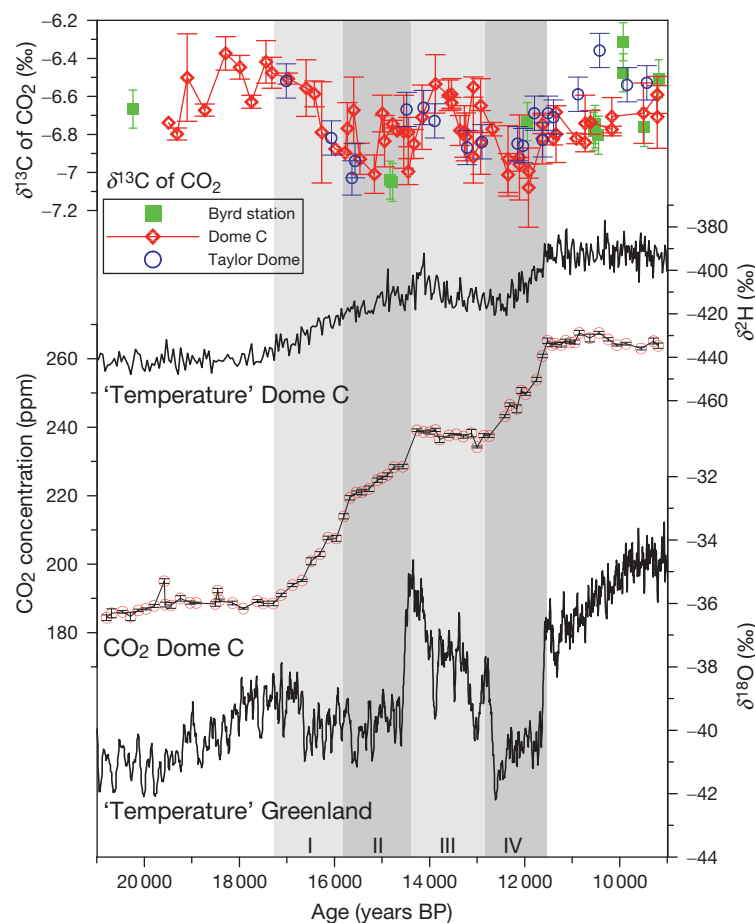


Figure 3 CO₂ and $\delta^{13}\text{C}$ record spanning the transition from the last glacial epoch to the Holocene, compared with temperature proxies from Antarctica (Dome C) and Greenland. The four intervals are discussed in the text. The error bars characterize 1σ uncertainties.

The section of ice discussed above has been remeasured for CO₂ by a different laboratory, using a different analytical method (Lourantou et al., 2010b). The agreement between the CO₂ data discussed above and this new dataset is remarkable ($R^2=0.996$). The new measurements also included analysis of $\delta^{13}\text{C}$ of CO₂, significantly increasing the handful of data points over the last glacial–interglacial transition previously available. The extended data from Dome C (Lourantou et al., 2010b) are generally in agreement with older data from the Byrd station (Leuenberger et al., 1992) and Taylor Dome (Smith et al., 1999) ice cores. The new results show a complicated w-shape pattern over the transition from glacial to interglacial. Despite the progress in $\delta^{13}\text{C}$ analyses and the increasing amount of available data, it is not currently possible to quantify the contribution of the potential causes presented above to explain the observed 76 ppm increase in atmospheric CO₂ concentration. A reasonable guess is that all four causes contributed to the increase, counterbalanced to some extent by the uptake of the growing terrestrial biosphere.

Initial investigations comparable to the ones presented above for Termination I have been made over the previous glacial–interglacial transition (Termination II). Compared to Termination I, the exact pattern and absolute size differ for both, the CO₂ concentration increase and its $\delta^{13}\text{C}$ variations.

However, they are attributed to the same main processes discussed above although individual contributions and timing vary due to the complex nature of climate–carbon cycle interactions (Lourantou et al., 2010a).

The Last Glacial Epoch

Drastic climatic variations on the millennial scale in Greenland and surroundings are characteristic for the last glacial epoch. They are related to variations in deep water formation in the North Atlantic. These variations have parallels in the Antarctic, but the amplitudes of the temperature variations are much smaller and they are not in phase with the ones in Greenland. Are there variations in atmospheric CO₂ concentration either synchronous with those in Greenland or with the Antarctic temperature variations? – Greenland ice core records are not well suited for the reconstruction of atmospheric CO₂ concentrations. Therefore, this question could only be answered after it became possible to synchronize the age scales of different ice cores through comparison of methane variations. A first reconstruction based on the comparison of the Byrd with the Greenland Ice Core Project (GRIP) Summit ice core records showed that there are variations in the CO₂ concentration on the order

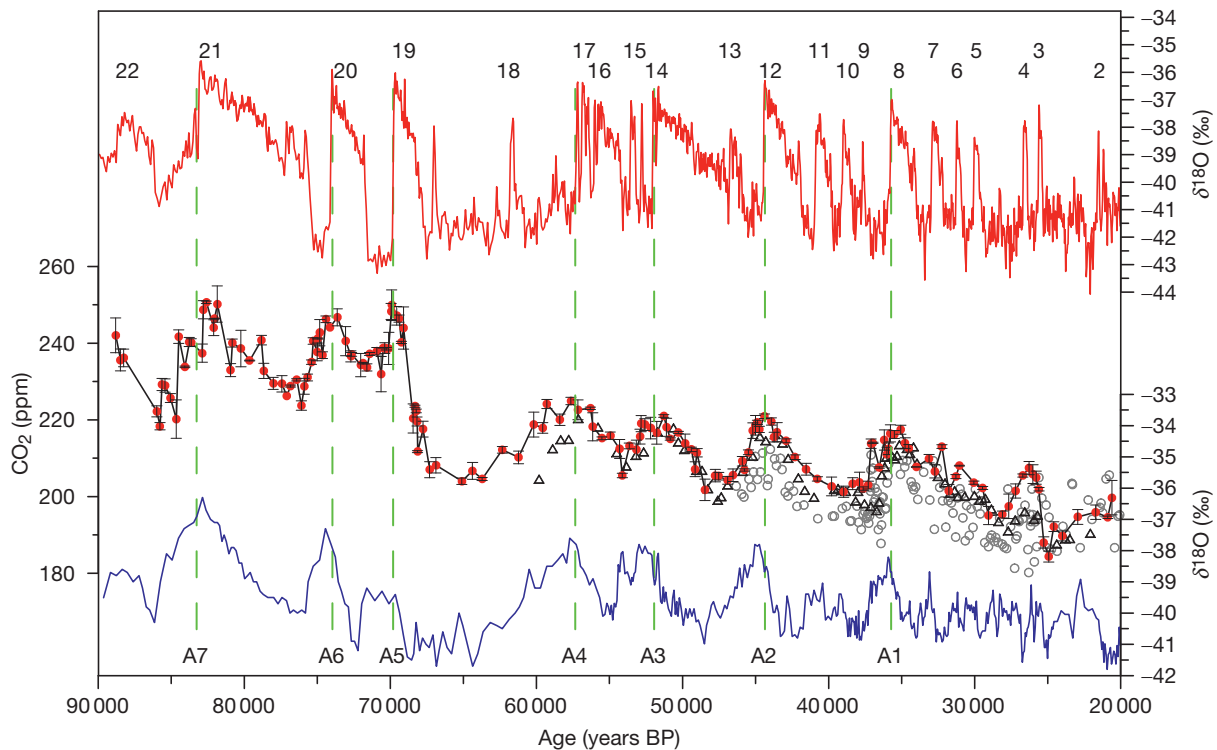


Figure 4 Comparison of CO₂ variations during the last glacial epoch with proxies of Antarctic and Greenland temperatures. CO₂ measurements from the Byrd station ice core (gray circles, Stauffer et al., 1998; red dots, Ahn and Brook, 2008) and the Taylor Dome ice core (black open triangles). The error bars for the new Byrd station data characterize 1 σ uncertainties. The time scales for temperature proxies ($\delta^{18}\text{O}$ of the precipitation water) from the Greenland GRIP core and the Byrd station core have been synchronized by methane data. The green dashed lines mark rapid temperature changes in Greenland with a correspondence in Antarctica.

of 20 ppm (Stauffer et al., 1998). This finding was confirmed by measurements taken from the Taylor Dome ice core (Indermühle et al., 2000). The 1968 Byrd core has been remeasured and extended in higher resolution and better precision (Ahn and Brook, 2008). The Byrd and Taylor Dome records are presented in Figure 4 together with stable isotope records from Antarctica (Byrd ice core) and Greenland (GRIP ice core). The Greenland and Antarctic records have been synchronized by methane data (Blunier and Brook, 2001).

CO₂ variations mirror Antarctic temperature oscillations during the last glacial. In detail, the CO₂ excursions are related to the longer Greenland interstitials with a prominent Antarctic counterpart. In general, the CO₂ concentration increases in parallel to the Antarctic temperature, but its decrease is slower, starting 1–2 millennia after the temperature has started to decrease. This may be a consequence of a reduced/increased North Atlantic Deep Water (NADW) formation resulting in a reduced/increased stratification of the southern ocean leading to a release/uptake of CO₂ (Ahn and Brook, 2008). Alternatively, the Antarctic temperature variations are accompanied by drastic variations in the nonsea salt calcium deposits (nss Ca²⁺) originating from aeolian dust, so that the iron fertilization by such dust deposits in the Southern Ocean varied considerably. This variable level of fertilization could be responsible for the variations in atmospheric CO₂ concentration. Because the amplitude of nss Ca²⁺ variations during the last glacial epoch were of a similar magnitude as the decrease at the end of the last glacial epoch, it was concluded that the iron

fertilization change during the transition and, therefore, the biological activity in the Southern Ocean could at most be responsible for 20 ppm of the total 76 ppm increase of the CO₂ concentration (Röthlisberger et al., 2004) at the end of the last glacial epoch.

Besides an Antarctic imprint the CO₂ record also shows features which seem to be directly linked to events in the Greenland. As during Termination I it is observed that at times of rapid temperature increases in Greenland the CO₂ concentration raises on the order of 10 ppm. This effect is seen accompanying Dansgaard–Oeschger (DO) events 21–19. There may be a link to higher sea level stands during this period opposed to later in the glacial. The observed drop in the mean CO₂ concentration at the end of DO 19 may also be related to a drop in sea level stands or potentially to a rapid increase of dust flux in the equatorial Pacific and Antarctica around that time (Ahn and Brook, 2008).

CO₂ Records over Eight Glacial Cycles

In 1998, the ice core drilling at Vostok reached a depth of 3623 m. The stratigraphy is undisturbed down to a depth of 3310 m, and the ice at this depth has an age of about 420 000 years. The records cover, therefore, about four glacial–interglacial cycles. There is a close correlation between the CO₂ record and the $\delta^2\text{H}$ record of precipitation water which is a proxy for the Antarctic temperature, confirming

the clear interplay between this important greenhouse gas and climate (Petit et al., 1999).

The record shows that the main trend of atmospheric CO₂ levels is similar for each glacial cycle. During glacial–interglacial transitions the concentrations rise from about 180 to 300 ppm. The increase is about synchronous with the temperature increase, but the start of the CO₂ increase probably lags the beginning of the temperature increase by a few centuries (Caillon et al., 2003; Fischer et al., 1999). The decrease in CO₂ concentration at the end of an interglacial is slower than its previous increase and it lags the temperature decrease by up to several thousand years (Fischer et al., 1999).

The EPICA Dome C ice core drilled to a depth of 3260 m (EPICA Community Members, 2004) extends the Vostok record. Analyses on a wide range of constituents have been performed down to a depth of 3190 m, at which the age of the ice is estimated to be about 800 000 years old. Below 3190 m, the stratigraphy becomes disturbed once again. The Vostok CO₂ record back to 420 000 years BP (Petit et al., 1999) and the Dome C record from 420 000 to 800 000 years BP (Lüthi et al., 2008; Siegenthaler et al., 2005b) are shown compared with the $\delta^2\text{H}$ record of the EPICA Dome C core (Figure 5). The CO₂ records show that atmospheric CO₂ concentration was never higher than 300 ppm during the past 800 000 years, compared to a modern concentration of 390 ppm (2010). During interglacial epochs before 450 000 years BP, showing lower mean temperature but longer duration, CO₂ concentration was also lower. This is further evidence for a close relation between global, climatic cycles, and atmospheric CO₂ variations.

Analytical Methods

To measure the CO₂ concentration of air enclosed in air bubbles of polar ice samples, the air has first to be extracted and then analyzed in a second step.

Most laboratories are using a dry extraction method to extract the air, in order to avoid any interaction of melt water

with ice included carbonates potentially resulting in the production of surplus CO₂. Air bubbles can either be opened mechanically by cracking, milling, or grating ice samples or be released by sublimation of the surrounding ice.

One method is to crack ice samples of about 10 g in a needle cracker which is cooled to -30°C and evacuated after the ice was filled to the cracker. The escaping air from the opened bubbles directly expands over a cold trap into the absorption cell of a laser absorption spectrometer. The infrared laser is tuned several times over the absorption line of a vibration–rotation transition of the CO₂ molecule to determine the concentration (Indermühle et al., 1999). A variant of this technique uses gas chromatography and a cryogenic trapping step (Ahn et al., 2009).

A second method is milling ice samples of about 40 g in a kind of ball mill. The ice sample and stainless steel balls are placed in a small container which is evacuated and then shaken. Temperature during shaking is kept at -50°C to keep water vapor pressure low. Air is released directly to the inlet system of a gas chromatograph (Barnola et al., 1995).

A third method (Etheridge et al., 1996) is to place ice samples weighing 500–1500 g in a vacuum-tight stainless steel container with a kind of stainless steel ‘cheese grater’ inside. By shaking the closed container after evacuation, the ice is grated in small pieces. The escaping air is extracted by condensation at -255°C in an electropolished stainless steel trap. The extracted air is analyzed with a gas chromatograph. The advantage of this method is that the extracted air is not exposed to moving metal parts. For many measurements the large sample size is a disadvantage, but if enough ice is available and does not restrict the desired depth resolution, it has the advantage of having a very small analytical uncertainty (1σ error of 1.2 ppm).

A fourth method applied (Gürlük et al., 1998; Schmitt et al., 2011) is based on the sublimation of an ice sample below the triple point of water. This guarantees a high extraction efficiency of close to 100% which cannot be achieved by the mechanical dry extraction methods. As long as all air is enclosed in bubbles the extraction efficiency is not crucial.

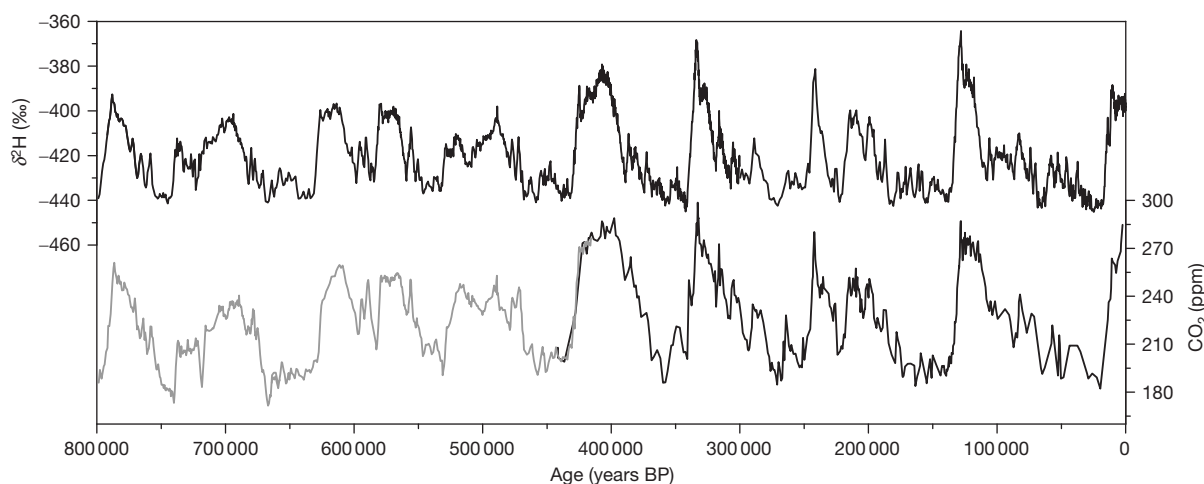


Figure 5 CO₂ records from Vostok back to 430 000 years BP (black line) and from Dome C from 430 000 to 800 000 years BP (gray line), compared with the $\delta^2\text{H}$ record from Dome C (EPICA Community Members, 2004), a proxy for Antarctic temperature.

However, at great depth in the ice sheet the pressure is high and air becomes enclosed in clathrates. Clathrates are a crystalline, cage-like structure formed by water molecules, with air molecules trapped in the cages. If air is enclosed partly in bubbles and partly in clathrates, a fractionation of different air components may occur if the extraction is not complete. With the high extraction efficiency achieved by the sublimation technique such fractionation is avoided independent of how the air is trapped in the ice (i.e., in bubbles, enclosed in clathrates or even dissolved in ice). A small disadvantage might be the fact that this technique is more time-consuming than the mechanical ones.

For bubbly ice, no difference has been observed between results derived by the mechanical and the sublimation method. The reproducibility (1σ) for the sublimation method using about 30 g of ice is ~ 2 ppm for CO₂ and very good precision of around 0.05‰ is achieved for the analysis of $\delta^{13}\text{C}$ of CO₂ (Schmitt et al., 2011). For the mechanical methods, the precision for bubbly ice can be as low as 1.5 ppm for CO₂ and around 0.07‰ for $\delta^{13}\text{C}$ using ice samples of around 8 g and 5–6 g, respectively (Elsig et al., 2009; Lüthi et al., 2008).

Reliability of CO₂ Data from Ice Cores

The CO₂ records from several ice cores overlap with direct atmospheric measurements and show good agreement. However, this is no guarantee that the CO₂ concentration in air bubbles from older ice reflects exactly the true atmospheric CO₂ at the time of ice formation. The following processes could alter the CO₂ concentration in air bubbles of ice:

- CO₂ could be produced by chemical reactions of impurities contained in the ice (e.g., acid-carbonate reaction or oxidation of organic compounds).
- CO₂ could be enclosed in so-called microbubbles or dissolve in the ice at the time of ice formation and then over a long period of time slowly diffuse into the air bubbles.
- CO₂ in ice samples stored in cold rooms could get lost by diffusion from clathrates or bubbles under high pressure of the surrounding atmosphere.

The likelihood of enrichment of CO₂ in air bubbles by the first two processes is small in Antarctic ice, because records from different ice cores with different impurity concentrations and from drilling sites with different temperatures and accumulation rates provide consistent results. Large scale changes like the industrial increase or the variations of CO₂ concentration within glacial cycles can, therefore, be taken for granted. However, with the improvement of analytical techniques smaller and more detailed variations have been detected. The best test to confirm such variations is the measurement of samples from additional ice cores. This is sometimes difficult, because the synchrony of the age scale of two different ice cores is often not sufficiently precise to allow a comparison of short-term CO₂ variations.

One way of determining whether a single ice core record is reliable or not, is to perform measurements with a high spatial resolution. If CO₂ is measured along a core interval representing only a few years, so that a single sample represents only one

or a fraction of an annual layer, the scatter of the individual results should fall within the analytical uncertainty. Erratic fluctuations along a core section representing a few years cannot represent variations of the atmospheric CO₂ concentration due to the gradual enclosure of air in bubbles. Extensive tests like the ones discussed above show that CO₂ concentration in bubbles can be altered not only in ice from Greenland but also in ice from Antarctica (Stauffer et al., 2003). But while the alterations are large in Greenland ice they are much smaller in Antarctic ice. In general, it can be concluded that the uncertainty is at most 1% or 5 ppm.

While they remain in ice sheets, air trapped in bubbles or clathrates is under high pressure. After recovery of the ice cores the pressure decreases only slowly and the question arises, whether air components could get lost by diffusion from the ice sample to the surrounding atmosphere, if ice is stored for several years in cold rooms. Diffusion of air components, except helium and neon, is so small that it cannot be measured directly. Different estimates give varying results concerning the solubility and diffusion constants for CO₂ and other air components. However, the diffusion of N₂, O₂, and CO₂ is so small that gas losses can be neglected. Indeed it has been shown that for CO₂ no effect can be found due to long-term storage given that a layer of 2–3 mm from the outside of an ice sample is removed prior to analyses (Bereiter et al., 2009).

See also: **Ice Core Methods:** Chronologies; Methane Studies; Studies of Firn Air. **Ice Core Records:** Correlations Between Greenland and Antarctica. **Paleoclimate Modeling:** Quaternary Environments. **Paleoclimate Reconstruction:** Sub-Milankovitch (DO/Heinrich) Events. **Plant Macrofossil Methods and Studies:** CO₂ Reconstruction from Fossil Leaves.

References

- Ahn J and Brook EJ (2008) Atmospheric CO₂ and climate on millennial time scales during the last glacial period. *Science* 322: 83–85.
- Ahn J, Brook EJ, and Howell K (2009) A high-precision method for measurement of paleoatmospheric CO₂ in small polar ice samples. *Journal of Glaciology* 55: 499–506.
- Barnola J-M, Ankin M, Porcheron J, et al. (1995) CO₂ evolution during the last millennium as recorded by Antarctic and Greenland ice. *Tellus* 47B: 264–272.
- Bereiter B, Schwander J, Lüthi D, et al. (2009) Change in CO₂ concentration and O₂/N₂ ratio in ice cores due to molecular diffusion. *Geophysical Research Letters* 36: L05703.
- Blunier T and Brook EJ (2001) Timing of millennial-scale climate change in Antarctica and Greenland during the last glacial period. *Science* 291: 109–112.
- Blunier T, Monnin E, and Barnola J-M (2005) Atmospheric CO₂ data from ice cores: Four climatic cycles. In: Ehleringer JR, Cerling TE, and Dearing MD (eds.) *A History of Atmospheric CO₂ and its Effects on Plants, Animals, and Ecosystems*, pp. 62–82. New York: Springer.
- Caillon N, Severinghaus JP, Jouzel J, et al. (2003) Timing of atmospheric CO₂ and Antarctic temperature changes across Termination III. *Science* 299: 1728–1731.
- Elsig J, Schmitt J, Leuenberger D, et al. (2009) Stable isotope constraints on Holocene carbon cycle changes from an Antarctic ice core. *Nature* 461: 507–510.
- EPICA Community Members (2004) Eight glacial cycles from an Antarctic ice core. *Nature* 429: 623–628.
- Etheridge DM, Steele LP, Langenfelds RL, et al. (1996) Natural and anthropogenic changes in atmospheric CO₂ over the last 1000 years from air in Antarctic ice and firn. *Journal of Geophysical Research* 101: 4115–4128.
- Fischer H, Wahlen M, Smith J, et al. (1999) Ice core records of atmospheric CO₂ around the last three glacial terminations. *Science* 283: 1712–1714.

- Flückiger J, Monnin E, Stauffer B, et al. (2002) High resolution Holocene N₂O ice core record and its relationship with CH₄ and CO₂. *Global Biogeochemical Cycles* 16: 1010. <http://dx.doi.org/10.1029/2001GB001417>.
- Francey RJ, Allison CE, Etheridge DM, et al. (1999) A 1000-year high precision record of $\delta^{13}\text{C}$ in atmospheric CO₂. *Tellus Series B: Chemical and Physical Meteorology* 51: 170–193.
- Güllük T, Slemr F, and Stauffer B (1998) Simultaneous measurements of CO₂, CH₄ and N₂O in air extracted by sublimation from Antarctica ice cores: Confirmation of the data obtained using other extraction techniques. *Journal of Geophysical Research* 103: 15971–15978.
- Indermühle A, Monnin E, Stauffer B, et al. (2000) Atmospheric CO₂ concentration from 60 to 20 kyr BP from the Taylor Dome ice core, Antarctica. *Geophysical Research Letters* 27: 735–738.
- Indermühle A, Stocker TF, Fischer H, et al. (1999) Holocene carbon-cycle dynamics based on CO₂ trapped in ice at Taylor Dome, Antarctica. *Nature* 398: 121–126.
- Jouzel J, Masson-Delmotte V, Cattani O, et al. (2007) Orbital and millennial Antarctic climate variability over the past 800 000 years. *Science* 317(5839): 793–796.
- Leuenberger M, Siegenthaler U, and Langway CC (1992) Carbon isotope composition of atmospheric CO₂ during the last ice age from an Antarctic ice core. *Nature* 357: 488–490.
- Lourantou A, Chappellaz J, Barnola JM, et al. (2010a) Changes in atmospheric CO₂ and its carbon isotopic ratio during the penultimate deglaciation. *Quaternary Science Reviews* 29: 1983–1992.
- Lourantou A, Lavric JV, Kohler P, et al. (2010b) Constraint of the CO₂ rise by new atmospheric carbon isotopic measurements during the last deglaciation. *Global Biogeochemical Cycles* 24: GB2015.
- Lüthi D, Le Floch M, Bereiter B, et al. (2008) High-resolution carbon dioxide concentration record 650,000–800,000 years before present. *Nature* 453: 379–382.
- Monnin E, Indermühle A, Dällenbach A, et al. (2001) Atmospheric CO₂ concentrations over the last glacial termination. *Science* 291: 112–114.
- Monnin E, Steig EJ, Siegenthaler U, et al. (2004) Evidence for substantial accumulation rate variability in Antarctica during the Holocene, through synchronization of CO₂ in the Taylor Dome, Dome C and DML ice cores. *Earth and Planetary Science Letters* 224: 45–54.
- Petit JR, Jouzel J, Raynaud D, et al. (1999) Climate and atmospheric history of the past 420,000 years from the Vostok ice core, Antarctica. *Nature* 399: 429–436.
- Röthlisberger R, Bigler M, Wolff EW, et al. (2004) Ice core evidence for the extent of past atmospheric CO₂ change due to iron fertilisation. *Geophysical Research Letters* 31: L16207.
- Schmitt J, Schneider R, and Fischer H (2011) A sublimation technique for high-precision measurements of $\delta^{13}\text{C}$ CO₂ and mixing ratios of CO₂ and N₂O from air trapped in ice cores. *Atmospheric Measurement Techniques* 4: 1445–1461.
- Siegenthaler U, Monnin E, Kawamura K, et al. (2005a) Supporting evidence from the EPICA Dronning Maud Land ice core for atmospheric CO₂ changes during the past millennium. *Tellus Series B: Chemical and Physical Meteorology* 57: 51–57.
- Siegenthaler U, Stocker TF, Monnin E, et al. (2005b) Stable carbon cycle–climate relationship during the late Pleistocene. *Science* 310: 1313–1317.
- Smith HJ, Fischer H, Wahlen M, et al. (1999) Dual modes of the carbon cycle since the Last Glacial Maximum. *Nature* 400: 248–250.
- Stauffer B, Blunier T, Dällenbach A, et al. (1998) Atmospheric CO₂ concentration and millennial-scale climate change during the last glacial period. *Nature* 392: 59–62.
- Stauffer B, Flückiger J, Monnin E, et al. (2003) Discussion of the reliability of CO₂, CH₄ and N₂O records from polar ice cores. In: Shoji H and Watanabe O (eds.) *Global Scale Climate and Environment Study Through Polar Deep Ice Cores*. Proceeding of the International Symposium on Dome Fuji Ice Core and Related Topics, 27–28 February 2001, Tokyo, pp. 139–152. Tokyo: National Institute of Polar Research.
- Stocker BD, Strassmann K, and Joos F (2011) Sensitivity of Holocene atmospheric CO₂ and the modern carbon budget to early human land use: Analyses with a process-based model. *Biogeosciences* 8: 69–88.
- Tans P and Keeling R. NOAA Earth System Research Laboratory, Global Monitoring Division Website: Trends in Atmospheric Carbon Dioxide. www.esrl.noaa.gov/gmd/ccgg/trends/.

Conductivity Studies

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Background

Snow falling on ice sheets comprises ice crystals that have formed in the cloud, or as clear-air precipitation nucleating on soluble or insoluble particles and capturing further aerosol material on the surfaces of the crystals as they pass through the atmosphere. Particularly in the polar regions (Greenland and Antarctica), and also on cold nonpolar glaciers, there is little surface melting, and the snow builds up, layer upon layer, compacting over time to form ice sheets. In Antarctica, the ice sheet in places exceeds 4000 m in thickness, and may contain ice that is over 1 My old in some places (though not necessarily where the ice is deepest as the insulating character of the ice traps geothermal heat from the bedrock causing the deepest ice to melt). In contrast, the Greenland ice sheet, although reaching over 3000 m, seems to contain continuous layers of ice only as old as around 130 000 years because of melting of the older ice at the base (partly due to the geothermal heat flux, and partly because the surface temperature of central Greenland is higher than that of Antarctica). As these deep ice sheets build up, air percolating around the ice crystals in the upper tens of meters (the firn layer) eventually becomes trapped in bubbles as compaction turns the porous firn into nonporous ice. Thus the ice becomes an archive of precipitation, atmospheric aerosol, and air. This archive can be recovered by drilling into the ice sheets and collecting ice cores, typically around 100 mm in diameter and recovered in lengths of 1–4 m. A profile of past climate and atmospheric change can be constructed by analysis of the isotopic composition of the water molecules, the chemistry of soluble and insoluble particles, and the gases trapped in the air bubbles. Since the early days of ice core research in the 1960s, about two dozen deep ice cores have penetrated through to the base of the ice sheets in Greenland and Antarctica, while considerably more shallow and intermediate depth cores have been recovered spanning the full geographic range of the polar regions as well as from many nonpolar glaciers. The deepest cores may span up to 800 000 years in Antarctica, and to around 130 000 years in Greenland, while shallower cores may only capture the past few decades or centuries.

Many analytical techniques have been brought to bear on ice-core samples, from measuring the concentration and isotopic composition of the trapped gases; the isotopic ratio of the hydrogen and oxygen atoms in the water itself; bulk and trace elements and ions in the meltwater; and the number, size, and elemental composition of the insoluble dust particles. The time resolution of the analyses depends on the depth resolution of the sampling, on the surface accumulation rate (which can be as little as a few centimeters in central Antarctica, to more than 1 m in the coastal regions), and on the layer thickness at depth, which decreases exponentially from the surface to the bed. The potential depth resolution of sampling has

improved over the years as instrumental techniques have improved, and particularly as detection limits of modern analytical instruments have become lower; but there are limits to the time resolution possible due to the mixing of snow during deposition by wind action and by subsequent metamorphism of the snow grains, and due to diffusion of water vapor and some chemical components in the ice. The majority of analytical measurements made on ice cores are essentially destructive – the core needs to be cut into samples and melted for introduction into the instrument or, in the case of air, crushed to release the trapped gases. This, of course, consumes the ice core, and inevitably leads to some compromises between the analytical aims and the availability of ice – particularly in sections of the core that are climatically significant such as transitions between cold and warm periods. There is also a trade-off between the considerable effort required to process and analyze the ice and the resolution (time or depth) desired. A technique that is both nondestructive and requires relatively little effort, yet is potentially capable of high depth resolution, is found in the electrical measurement of ice. Because the electrical logging of cores is usually fast, does not consume ice, offers potentially high depth and time resolution, and responds to both seasonal and climatic changes in the chemistry of the ice, most of the major ice cores drilled in the past three decades have undergone some form of electrical logging. The two main techniques are usually referred to as ‘electrical conductivity measurement’ (ECM) and ‘dielectric profiling’ (DEP).

The electrical properties of ice are dependent on the trapped impurities, as well as on the temperature and density of the ice. The air passages or bubbles in the ice are poor conductors but affect the electrical properties via the density of ice, while the dust particles tend to modify the electrical properties via their effect in neutralizing the acidity of ice rather than conducting significant electrical current. While there were many early studies of the electrical properties of pure laboratory-grown ice (Hobbs, 1974), it was the discovery, by the pioneers of the radar surveys of the ice sheets, that radar-reflecting horizons were visible within the ice sheet which suggested variations in the dielectric properties of natural *in situ* ice (e.g., Paren and Robin, 1975). Hobbs (1974) describes in detail the physics underlying the dielectric dispersion in ice: at low frequencies, typically <100 Hz, the polarization of the water molecule is similar to that of ice, with a relative permittivity of about 90, while at frequencies above several hundred kilohertz, the dipole moment is randomly oriented to the exciting field, the relative permittivity is about 3, and ice has very little absorption. The frequency at which dielectric absorption is greatest is known as the ‘relaxation frequency,’ and is around 12 kHz (Moore and Paren, 1987). Wolff (2000) summarizes the physics of the electrical properties of ice relevant to ice-core measurements, and notes that the properties commonly measured

are the real permittivity and the conductivity (which is itself related to the imaginary permittivity). Above the firn–ice transition, the permittivity is strongly controlled by the density of the ice, while below that depth, chemical impurities, temperature, and the fabric influence the permittivity. For conductivity, the temperature is important, and to some extent the density, but it is the chemical impurities in the ice that have the major influence. To reduce the temperature variability on the permittivity and conductivity, ice cores are normally measured at a temperature of around -20°C , having been stored at a similar and stable temperature for a few days prior to measurement in order to equilibrate the ice. Further, the ice temperature, either at the surface or, more preferably, internally with a temperature probe inserted into a small drilled hole in the core, is measured and a small correction made to the electrical measurements.

DC Conductivity and the ECM Technique

Early work on DC conductivity of ice was often carried out by making low-frequency AC measurements on ice disks in a flat-plate capacitor. This technique is inherently slow, and not appropriate to the continuous electrical logging of long sections of ice. Claus Hammer of the Copenhagen ice-core group made a significant breakthrough in creating a practical technique when he developed a method of continuously measuring the resistance between two electrodes carrying a high DC voltage applied to the surface of ice (Hammer, 1980). This technique became known as the ‘electrical conductivity method,’ commonly referred to as ECM. The technique requires a freshly cut and preferably planed surface, which in some senses implies that the technique is not strictly ‘nondestructive,’ but the common approach is to make the measurement on the first longitudinal cut surface on the core once the initial ice processing begins. A pair of electrodes (often brass) about 10 mm apart is brought into contact with the ice, and a DC voltage of about 1 kV applied across the electrodes. The electrodes are then drawn along the ice at a rate of typically $50\text{--}100\text{ mm s}^{-1}$, and the current flowing (the ECM value typically up to around $50\text{ }\mu\text{A}$) between the electrodes recorded along with the position along the core taken from some form of encoder. Although the ECM is not a direct measurement of the true DC conductivity of ice, there is a close relationship between the two, provided the measurement is carried out on timescales slower than the relaxation frequency but faster than the time for space charges to build up in the ice and reduce the current (Wolff, 2000). Hammer et al. (1980) suggest that this time is of the order of $0.1\text{--}0.5\text{ s}$, with a minimum speed of movement of 20 mm s^{-1} . It is a tribute to the simplicity of the technique and the utility of the ECM data that the Hammer technique is essentially used even today to measure ice-core electrical stratigraphy. Instrument versions might use higher voltages for lower conducting ice, and differing electrode material, size, and spacing, which makes direct quantitative comparison of the magnitude of the ECM signal between cores difficult, but the qualitative shape of the signal in response to ice impurities is remarkably similar. A more recent development has been the multitrack ECM method with a number of electrodes contacting the ice across the width of the core to

construct a two-dimensional image of conductivity, which has been used to determine whether ice layers are dipping or perhaps folded (Taylor and Alley, 2004).

Wolff (2000) summarizes the theory governing DC conductivity in ice, and concludes that in solid ice it can be assumed to be controlled by the displacement of ionic H_3O^+ defects in the lattice. He points out that probably only H^+ , NH_4^+ (or NH_3), and Cl^- (or HCl) are small enough to enter the ice lattice substitutionally and present in sufficient quantities to modify conductivity through the lattice. Larger ions commonly present as impurities in ice, such as NO_3^+ and SO_4^{2-} , may not enter the lattice, and therefore play little role in the conduction mechanism of solid ice. Wolff and Paren (1984) suggested a two-phase model for DC conductivity in natural polar ice, including conduction along liquid acid veins at triple grain boundaries which form a continuous matrix through the ice structure along with solid-phase lattice defect conduction. The later discovery of strong sulfuric acid located at triple grain boundaries in ice (Mulvaney et al., 1988) supports the liquid vein conduction mechanism, particularly because the eutectic temperature of a sulfuric acid/water mixture is well below the coldest polar ice sheet temperature. Indeed, it is commonly observed that the ECM signal recorded in ice (Figure 1) is strongly correlated to the acidity of ice, particularly evident when compared to measurements of the sulfate ion (which may not enter the lattice) which has origins in the seasonal acidic sulfate from the oxidation of marine biogenic sulfurous gases commonly ascribed to dimethyl sulfide and from the sporadic acidic sulfate arising from the oxidation of sulfur dioxide emitted

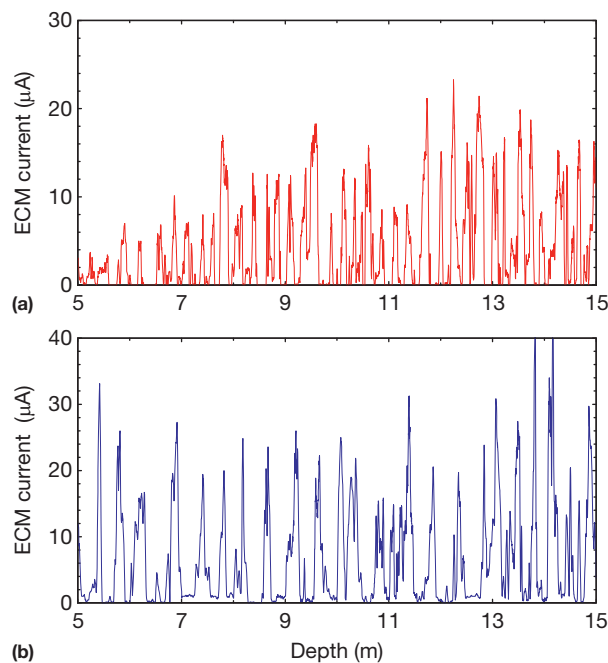


Figure 1 Short sections of electrical conductivity method (ECM) profiles from two ice cores drilled approximately 160 km apart from (a) the south dome and (b) north dome of Berkner Island, Antarctica, showing clear seasonal cycles. The greater number of peaks in the south dome core indicates a lower accumulation rate than the north dome core (data source: Mulvaney et al., 2002).



Figure 2 The electrodes of the ECM instrument in use in the field on a section of the deep ice core from the NEEM project, Greenland. Photograph by R. Mulvaney.

by volcanoes. But, the observations that the relationship between the measured acidity (measured as H^+) and DC current is not linear (e.g., see Figure 5 in Moore et al., 1992b) and that fluctuations in ECM current persist even in single large crystals in deep ice suggest that solid-phase conduction must also contribute significantly to the measured current flow.

Regardless of the precise physics behind the measured ECM conductivity, the fact that the technique responds well both to local seasonal acidity and to regional or global sporadic acidic events of volcanic origin has proved a remarkable utility for dating ice cores, and it has therefore become a common technique on many ice-core drilling projects. An example of a recent application can be seen in the photograph in Figure 2: the ECM signal is measured using a simple handheld double brass electrode drawn across the planed surfaces of freshly cut lengths of ice cores drilled by the NEEM project in Greenland (<http://neem.dk/>). Since Hammer's (1980) first description of the ECM technique, all the deep ice cores in Greenland drilling projects coordinated by the Copenhagen group have been measured in a similar fashion, allowing dating by counting seasonal layers in the ECM signal (usually alongside parallel seasonal chemical signals measured by a 'continuous flow analysis' technique). Perhaps the best example of multicomponent seasonal layer counting is the establishment of the Greenland GICC05 age scale, which now extends to 60 ka BP (Svensson et al., 2008). The appearance of sporadic larger peaks in ECM conductivity of volcanic origin allows for checks on the seasonal dating where the date of the volcanic eruption is known, and additionally allows for cross-correlation of the dating between cores by matching the pattern of the large conductivity peaks (e.g., Taylor et al., 1993).

Dielectric Profiling

While the DC conductivity of pure ice appears to be related to H_3O^+ defects in the lattice and conduction through the grain boundary and vein network, Moore et al. (1992b) attributes the high-frequency AC dielectric conductivity to movements of Bjerrum L- and D-defects in the ice, and associates these by experimental observation to the H^+ , NH_4^+ , and Cl^- ions (most other major ions such as NO_3^+ and SO_4^{2-} are too large to enter

the lattice and to contribute to the dielectric conductance). As with the ECM technique, the early AC dielectric properties of ice were traditionally measured on disks of ice sandwiched between two guarded parallel plates. The first practical system for routinely measuring the DEP of ice and relating the data to the bulk chemistry of the core was described by Moore and Paren (1987) and Moore et al. (1989), who demonstrated a method that was not only nondestructive but also did not require either a fresh surface or contact between the electrodes and the ice, implying a polythene sleeve could be left on the ice to reduce contamination. In Moore's instrument, the ice acts as a capacitor between two curved aluminum electrodes, each subtending an arc of 108° and at least as long as the ice core to be measured. The lower electrode was continuous along the core, and connected to the high-voltage side of a measuring device (an LCR meter measuring inductance L , conductance C , and resistance R at a range of frequencies from hertz to megahertz). The upper electrode consisted of a number of similarly curved but separated electrodes each of 50 mm length along the core, while two earthed flat-plate aluminum side guard electrodes ran the whole length of the electrode assembly to confine the field to be close to parallel across the core. No electrical contact between the electrodes and the ice was required for the technique, which was achieved by anodizing the aluminum and keeping a plastic sleeve around the ice and a small air gap with the upper electrodes. The whole electrode assembly was then housed in an aluminum case to isolate it from electromagnetic interference. A switching device connected each of the narrow upper electrodes in turn to the low-voltage side of the measuring device, and the sequence of capacitance and conductance measurements was made at a range of frequencies from 20 Hz to 300 kHz along a 1-m length of core in 30 min. A development model of this device was used successfully on the 3029-m deep Greenland Ice Core Project (GRIP) and on the 3270-m deep ice core drilled by the EPICA project at Dome C. In this instrument (Figure 3), 120 electrodes of 20 mm width were arranged along the length of the instrument, and a switching device connected each electrode in turn to the measuring device, giving an effective depth resolution of 20 mm. This instrument produced a DEP record from the full core, demonstrating the glacial-interglacial contrast in electrical properties of the ice clearly (Figure 4). Improvements to the early instrumentation followed from a better understanding of the geometry of the electrodes (Wilhelms et al., 1998). Most modern DEP instruments use a single, narrow (a few millimeters) central electrode with earthed guard electrodes either side, which is moved by a stepper motor along the length of a core which sits on a continuous lower electrode (e.g., see Figure 2 in Hofstede et al., 2004), while an LCR measuring device records the conductance and impedance at frequencies up to 1 MHz. As noted earlier, temperature has an effect on all electrical measurements, which is therefore controlled by buffering the ice in the field or laboratory to achieve a uniform temperature throughout the core section, and the results corrected for the actual temperature measured of the ice. The spatial resolution of the DEP tends to be limited to about 10–20 mm but, in a further development of the DEP, Fujita et al. (2002) uses a hybrid technique applying a high-frequency current to a pair of electrodes dragged across the surface of the core, which is claimed to give a resolution more similar to that of the ECM.



Figure 3 A section of 2.2 m of EPICA Dome C ice core loaded onto the DEP instrument in the field in Antarctica: the ice core is marked up into 55 cm 'bags' during the processing, and here bags 173–176 (94.6–96.8 m) are being made ready for measurement. Photograph by R. Mulvaney.

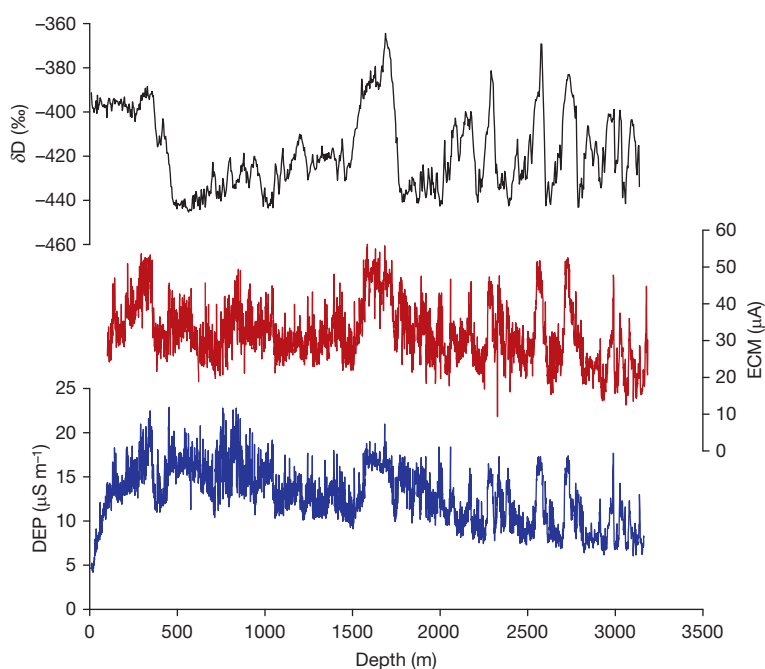


Figure 4 The EPICA Dome C stable water isotope and electrical data: (a) deuterium data showing eight glacial–interglacial climate cycles; (b) ECM conductivity; (c) 100-kHz DEP conductance (data source: [EPICA Community Members, 2004](#); [Stauffer et al., 2004](#)).

The DEP responds strongly to acidity, but has also been shown to respond more weakly to sea salt ions, particularly Cl^- ([Moore et al., 1992a](#)) and NH_4^+ ([Moore et al., 1994](#)) which are probably dispersed within the fabric of the ice. In contrast to its DC conductivity, pure ice has a significant high frequency conductivity, which can be seen as a background response even when concentrations of impurities are low. There is a significant dependence on the density of the ice (effectively the quantity of ice between the electrodes and the dielectric capacitance of air trapped between ice grains), and this is clearly seen in the DEP response from the upper 100 m of ice-core records (e.g., [Barnes et al., 2002](#)), which follows the densification curve of the ice ([Figure 5](#)). [Wolff et al. \(1995\)](#)

demonstrate the dependence of the DEP signal on the concentrations of acidity, ammonium, and chloride in the ice over a complete glacial to interglacial climate cycle in a Greenland core, and note that there is no evidence for any changes in the DEP (and ECM) electrical response to the depth or age of the ice, other than via the changes in chemistry.

Uses of Electrical Measurements

Because of their nondestructive nature, electrical measurements tend to be applied early in the analysis of many ice cores. They are often made in the field at the drilling site,

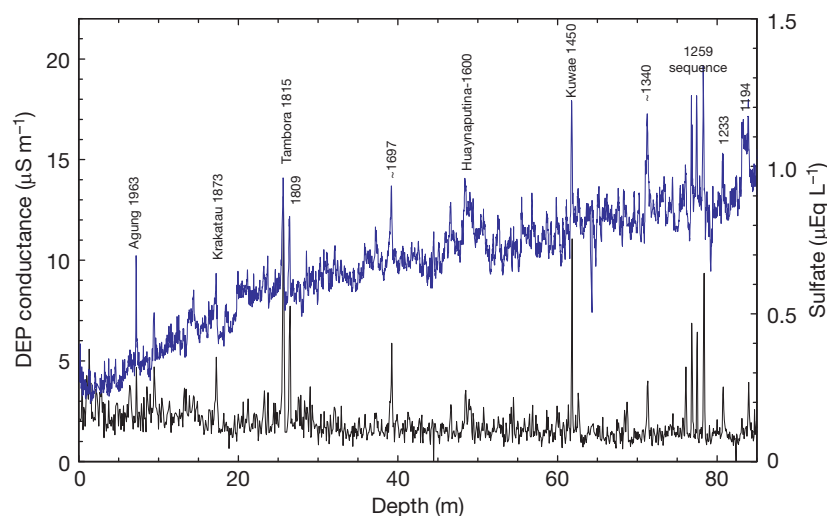


Figure 5 Comparison of (a) the DEP 50-kHz conductance (blue curve) with (b) the nonsea-salt sulfate impurities (black curve) in a core from Dronning Maud Land, Antarctica. The nonsea-salt sulfate is calculated from the total sulfate measured in the ice minus the contribution from sea salt calculated from a conservative sea-salt marker such as sodium, and is the sulfate arising from oxidation of marine biogenic and volcanic gases. The slope on the DEP signal is due to the densification of firn to ice.

offering an early indication of the stratigraphy of the ice through the recognition of seasonal layers and well-dated volcanic horizons in cores from high accumulation sites to patterns of volcanic horizons and the contrast between glacial and interglacial ice in the deep cores from low accumulation sites. The DEP response increases with the volume of ice between the electrodes, so it is commonly the first measurement technique carried out on the ice and can even be applied while the ice remains in the polythene sleeve used to protect the ice during transport to the laboratory. Most of the major deep ice-core drilling projects have made these measurements in the field shortly after drilling the ice (after allowing the ice to reach a thermal equilibrium, generally at a temperature of around -20°). In these projects, the ECM is then usually carried out on the first lengthways-cut surface of the ice, though a planed surface on the outer edge of the core is normally effective too.

For shallow and intermediate depth cores taken from relatively high accumulation sites, and particularly in near-coastal regions, the ECM signal often shows a clear seasonal cycle which can be used to date the core by counting annual layers (Figure 1). This layer counting technique has reached a pinnacle of achievement in the dating of many of the Greenland ice cores which, in contrast to the deep Antarctic cores, are commonly recovered in areas of sufficiently high accumulation for the seasonal cycles in acidity and, therefore, the ECM response to be retained even at great depth. Thus the ECM signal and continuous chemical analysis defining seasonal layers in other impurities have been used together to date the cores with some degree of reliability to as much as 60 ka (e.g., Svensson et al., 2008). Even where seasonal layers in the ECM are not resolved, the pattern of discrete peaks resulting from volcanic deposition of acidity can be used to cross-correlate between ice cores from different sites. Clausen et al. (1997) compares 4000 years of volcanic horizons between the Greenland GRIP and Dye 3 cores. Vinther et al. (2008) synchronizes several Canadian high Arctic cores with the Renland, GRIP, and NGRIP cores from Greenland. Taylor et al. (1993) shows a particularly

convincing match in the pattern of ECM peaks in two separate 10-m sections taken from depths of more than 2300 m in the Greenland Ice Sheet Project 2 (GISP2) and GRIP cores. Taylor et al. (1993, 1997) goes on further to demonstrate that the ECM signal changes by a factor of 40 between stadial and interstadial periods, with the strongest ECM response in the warm periods, and a low signal from the cold periods where a high concentration of alkaline dust tends to neutralize the acidity of the ice suppressing the ECM response. This pattern of high and low ECM signals is easily matched between the two cores and, taken together with the recognizable patterns in the volcanic peaks, the ECM profiles form a powerful method of tying the age of the ice between cores. Few of the very deepest (>3000 m) cores from Antarctica have sufficient accumulation at the surface to avoid mixing of the snow by surface drifting, effectively eliminating the seasonal signal in impurities. However, for the first of the deep Antarctic ice cores from Byrd Station, where the present-day accumulation is around half that of central Greenland, the seasonal signal appear to be preserved. Hammer et al. (1994) were still able to readily identify some seasonal cycles in the Byrd core ECM to a depth of 1750 m, or 40 000 years, and could pick out some layers even deeper than this. While not able to count seasonal cycles continuously from the surface, the layer thickness at various depths does contribute significantly to the age–depth model. At Siple Dome, Antarctica, Taylor et al. (2004) included an ECM record in their seasonal layer counting across the full Holocene period.

In central Antarctica, away from the coastal region with its particularly strong seasonal acidic signal from seasonal biogenic sources and marine sea salt input to the ice, the DEP profiles from ice cores have proven particularly useful in picking out the acidity from volcanic events. Over the most recent 1000 years, many of these volcanic eruptions are historically recorded events and therefore well dated, and can be used as reference horizons to date cores. Karlof et al. (2000) in fact have demonstrated that the contrast between the volcanic peaks and the background seasonal signal is greater in the

DEP than in the ECM, making DEP probably the more useful technique, at least in regions closer to the coast. In a DEP record from a 100-m core from the Mizuho Plateau in Dronning Maud Land, Moore et al. (1991) picked out horizons attributed to the eruptions of Agung (1963), Krakatau (1873), Tambora (1815), and the eruptions of 1601 and 1259. These horizons can be picked out in the majority of DEP profiles from Antarctic ice cores, and, indeed, the sequence of four volcanic peaks around the year 1259 is quite distinctive (Figure 5). Hofstede et al. (2004) used DEP profiles from six shallow cores (up to 150 m, but all spanning at least 1000 years) from Dronning Maud Land to identify the principal volcanic peaks, and then derived the mean accumulation rate experienced at each site between successive horizons to determine the spatial and temporal pattern of accumulation in this region over the past millennium. Udisti et al. (2000) used ECM, DEP, and chemistry from continuous flow instruments to pick out volcanic horizons, contributing to the dating of the Holocene section of the Dome C deep ice core.

Electrical Properties of the Ice Sheet

The discovery of internal reflecting horizons in the early radar sounding of ice sheets suggested a wide variation in the dielectric properties of natural ice (e.g., Paren and Robin, 1975), which eventually led to the continuous electrical measurement techniques on ice cores. Radar layers originate from contrast in the real and imaginary permittivity of ice layers, and they arise ultimately from contrasts in the density, fabric, and chemistry of the ice. Now that the electrical profiles of deep ice cores are routinely measured, interest in modeling the radar response of the ice column has increased. Hempel et al. (2000) compared the radar-reflecting horizons in the Greenland ice sheet with the ECM trace from the GRIP core. After accounting for the density profile, and correcting for the wave velocity of radar in ice, they concluded that the acidic layers of volcanic origin show up as radar horizons. Indeed, it has been clear for a long time that there is a clear contrast between the numerous radar returns from the acidic Holocene ice and the paucity of returns from the dust-laden glacial ice where the acidity is neutralized, punctuated with strong reflectors thought to originate from the acidic interstadial layers. Using the DEP record from the same GRIP core, Miners et al. (2002) constructed a synthetic radargram that managed to pick out many of the larger reflecting horizons below 500 m (where, presumably, radar returns from density contrasts are less significant than chemical contrasts in the ice), and nearly all the features below 1500 m, including some of the interstadial events in the glacial ice. In Antarctica, Eisen et al. (2003) have deconvoluted DEP records from a 150-m core to produce a synthetic radargram to compare with high-frequency shallow radar profiles and shown that only some reflections in the upper layers of the firn correlate to acidic horizons. The use of DEP profiles from the deeper Antarctic ice cores offers a refinement of the synthetic radargram method, which has application in dating and tracing isochronal layers in radar profiles between ice-core sites across the ice sheet (Eisen et al., 2006).

See also: **Ice Cores: History of Research, Greenland and Antarctica.**
Ice Core Methods: Chronologies; Glaciochemistry; Overview.

References

- Barnes PRF, Wolff EW, Mulvaney R, et al. (2002) Effect of density on electrical conductivity of chemically laden polar ice. *Journal of Geophysical Research-Solid Earth* 107: 2029. <http://dx.doi.org/10.1029/2000jb000080>.
- Clausen HB, Hammer CU, Hvidberg CS, et al. (1997) A comparison of the volcanic records over the past 4000 years from the Greenland Ice Core Project and Dye 3 Greenland ice cores. *Journal of Geophysical Research-Oceans* 102: 26707–26723.
- Eisen O, Wilhelms F, Nixdorf U, and Miller H (2003) Identifying isochrones in GPR profiles from DEP-based forward modeling. *Annals of Glaciology* 37: 344–350.
- Eisen O, Wilhelms F, Steinhage D, and Schwander J (2006) Improved method to determine radio-echo sounding reflector depths from ice-core profiles of permittivity and conductivity. *Journal of Glaciology* 52: 299–310.
- EPICA Community Members (2004) Eight glacial cycles from an Antarctic ice core. *Nature* 429: 623–628.
- Fujita S, Azuma N, Motoyama H, et al. (2002) Electrical measurements on the 2503 m Dome F Antarctica ice core. *Annals of Glaciology* 35: 313–320.
- Hammer CU (1980) Acidity of polar ice cores in relation to absolute dating, past volcanism, and radio echoes. *Journal of Glaciology* 25: 359–372.
- Hammer CU, Clausen HB, and Dansgaard W (1980) Greenland ice sheet evidence of post-glacial volcanism and its climatic impact. *Nature* 288: 230–235.
- Hammer CU, Clausen HB, and Langway CC (1994) Electrical conductivity method (ECM) stratigraphic dating of the Byrd Station ice core, Antarctica. *Annals of Glaciology* 20: 115–120.
- Hempel L, Thyssen F, Gundestrup N, Clausen HB, and Miller H (2000) A comparison of radio-echo sounding data and electrical conductivity of the GRIP ice core. *Journal of Glaciology* 46: 369–374.
- Hobbs PV (1974) *Ice Physics*. Oxford: Clarendon Press.
- Hofstede CM, van de Wal RSW, Kaspers KA, et al. (2004) Firn accumulation records for the past 1000 years on the basis of dielectric profiling of six cores from Dronning Maud Land, Antarctica. *Journal of Glaciology* 50: 279–291.
- Karolof L, Winther JG, Isaksson E, et al. (2000) A 1500 year record of accumulation at Amundsenisen western Dronning Maud Land, Antarctica, derived from electrical and radioactive measurements on a 120 m ice core. *Journal of Geophysical Research-Atmospheres* 105: 12471–12483.
- Miners WD, Wolff EW, Moore JC, Jacobel R, and Hempel L (2002) Modeling the radio echo reflections inside the ice sheet at Summit, Greenland. *Journal of Geophysical Research-Solid Earth* 107: 2172. <http://dx.doi.org/10.1029/2001jb000535>.
- Moore JC, Mulvaney R, and Paren JG (1989) Dielectric stratigraphy of ice – A new technique for determining total ionic concentrations in polar ice cores. *Geophysical Research Letters* 16: 1177–1180.
- Moore JC, Narita H, and Maeno N (1991) A continuous 770-year record of volcanic activity from east Antarctica. *Journal of Geophysical Research-Atmospheres* 96: 17353–17359.
- Moore JC and Paren JG (1987) A new technique for dielectric logging of Antarctic ice cores. *Journal de Physique* 48: C1/155–C1/160.
- Moore JC, Paren JG, and Oerter H (1992a) Sea salt dependent electrical conduction in polar ice. *Journal of Geophysical Research* 97: 19803–19812.
- Moore JC, Wolff EW, Clausen HB, and Hammer CU (1992b) The chemical basis for the electrical stratigraphy of ice. *Journal of Geophysical Research* 97: 1887–1896.
- Moore JC, Wolff EW, Clausen HB, Hammer CU, Legrand MR, and Fuhrer K (1994) Electrical response of the Summit-Greenland ice core to ammonium, sulfuric-acid, and hydrochloric-acid. *Geophysical Research Letters* 21: 565–568.
- Mulvaney R, Oerter H, Peel DA, et al. (2002) 1000 year ice-core records from Berkner Island, Antarctica. *Annals of Glaciology* 35(1): 45–51.
- Mulvaney R, Wolff EW, and Oates K (1988) Sulphuric acid at grain boundaries in Antarctic ice. *Nature* 331: 247–249.
- Paren JG and Robin G de Q (1975) Internal reflections in polar ice sheets. *Journal of Glaciology* 14: 251–259.
- Stauffer B, Fluckiger J, Wolff E, and Barnes P (2004) The EPICA deep ice cores: First results and perspectives. *Annals of Glaciology* 39: 93–100.
- Svensson A, Andersen KK, Bigler M, et al. (2008) A 60 000 year Greenland stratigraphic ice core chronology. *Climate of the Past* 4: 47–57.
- Taylor KC and Alley RB (2004) Two-dimensional electrical stratigraphy of the Siple Dome (Antarctica) ice core. *Journal of Glaciology* 50: 231–235.

- Taylor KC, Alley RB, Lamorey GW, and Mayewski P (1997) Electrical measurements on the Greenland Ice Sheet Project 2 core. *Journal of Geophysical Research-Oceans* 102: 26511–26517.
- Taylor KC, Alley RB, Meese DA, et al. (2004) Dating the Siple Dome (Antarctica) ice core by manual and computer interpretation of annual layering. *Journal of Glaciology* 50: 453–461.
- Taylor KC, Hammer CU, Alley RB, et al. (1993) Electrical-conductivity measurements from the GISP2 and GRIP Greenland ice cores. *Nature* 366: 549–552.
- Udisti R, Becagli S, Castellano E, et al. (2000) Holocene electrical and chemical measurements from the EPICA-Dome C ice core. *Annals of Glaciology* 30: 20–26.
- Vinther BM, Clausen HB, Fisher DA, et al. (2008) Synchronizing ice cores from the Renland and Agassiz ice caps to the Greenland ice core chronology. *Journal of Geophysical Research-Atmospheres* 113: D08115. <http://dx.doi.org/10.1029/2007jd009143>.
- Wilhelms F, Kipfstuhl J, Miller H, Heinloth K, and Firestone J (1998) Precise dielectric profiling of ice cores: A new device with improved guarding and its theory. *Journal of Glaciology* 44: 171–174.
- Wolff E (2000) *Electrical Stratigraphy of Polar Ice Cores: Principles, Methods and Findings*. Sapporo: Hokkaido University Press.
- Wolff EW, Moore JC, Clausen HB, Hammer CU, Kipfstuhl J, and Fuhrer K (1995) Long-term changes in the acid and salt concentrations of the Grip Greenland ice core from electrical stratigraphy. *Journal of Geophysical Research* 100: 16249–16264.
- Wolff EW and Paren JG (1984) A two-phase model for electrical conduction in polar ice sheets. *Journal of Geophysical Research* 89: 9433–9438.

Introduction

Ice cores recovered from polar and high-elevation regions provide a unique and valuable archive of past atmospheric conditions, in large part because of the numerous physical and chemical measurements that can be performed at any given depth, and because the transformation of snow to firn to ice on a glacier or ice sheet provides two basic sets of atmospheric information. First, when firn turns into ice, bubbles are formed that trap ambient air whose composition provides information on atmospheric trace gases (e.g., CO₂, CH₄) at the time of ice formation. Second, aerosols and water-soluble gas species can be trapped during precipitation, as well as deposited directly on the snow surface. While the term ‘glaciochemistry’ is broad and could conceivably encompass all chemical measurements made on ice core samples, in practice, it generally represents the soluble mineral and organic components originally deposited on the snow surface (Legrand and Mayewski, 1997). Measurement of the suite of major ions (Na⁺, Mg²⁺, Ca²⁺, K⁺, NH₄⁺, Cl[−], NO₃[−], and SO₄^{2−}) represents approximately 95% of the soluble composition of the atmosphere, and therefore, provides a powerful tool for investigating modern and past changes in atmospheric chemistry. Ions from soluble organic acids, such as formate (HCOO[−]), acetate (CH₃COO[−]), and oxylate (C₂O₄[−]), methylsulfonate (CH₃SO₃[−]; MS[−]), and other volatile species such as hydrogen peroxide (H₂O₂) and formaldehyde (HCHO) have been measured at various ice core sites and used to study atmospheric and biogeochemical processes. However, the major ion suite is most commonly measured in ice cores, and there is an extensive dataset available from a globally distributed set of ice core sites. Because of the varied geographic positions (e.g., elevation, distance from ocean or arid regions, etc.) of these sites, major ion (plus MS[−]) concentrations in ice cores show a wide range of values based on temporal changes in chemical source strength, climate dynamics, and biogeochemical processes. This article therefore contains a review of current issues in the field of ice core glaciochemistry (restricted here to major ions plus MS[−]), and summarizes information related to chemical sources, transport, deposition, and interpretation in terms of paleoclimate and paleoenvironmental change. For a more thorough understanding of the range of chemical species studied in ice cores, see elsewhere in this encyclopedia.

Glaciochemical Background and Measurement

The composition of precipitation in polar and alpine regions contains various soluble impurities, introduced to the atmosphere either directly as primary aerosols (e.g., dust entrained from terrestrial surfaces), or produced within the atmosphere (secondary aerosols) during the oxidation of trace gases

involved in the sulfur, nitrogen, halogen, or carbon cycles. Some soluble ion species have multiple sources (Legrand and Mayewski, 1997); for example, SO₄^{2−} can be linked to both primary aerosols (sea salt and dust) as well as secondary aerosols (from the oxidation of sulfur gases produced by volcanic activity, anthropogenic [human] activities, and the biosphere). It is therefore necessary to study all soluble species present in ice core samples in order to understand and reconstruct the original association of the ions. The ionic budget (Σ ; sum of all ion species present) in polar and alpine snow can generally be written as follows (concentrations in microequivalents per liter, $\mu\text{eq L}^{-1}$):

$$\begin{aligned} \Sigma = & [\text{Na}^+] + [\text{NH}_4^+] + [\text{K}^+] + [\text{H}^+] + [\text{Ca}^{2+}] + [\text{Mg}^{2+}] \\ & + [\text{F}^-] + [\text{Cl}^-] + [\text{NO}_3^-] + [\text{SO}_4^{2-}] + [\text{CH}_3\text{SO}_3^-] \\ & + [\text{HCOO}^-] + [\text{CH}_3\text{COO}^-] \end{aligned}$$

In most situations, the concentrations of F[−] and the light carboxylates are insignificant. While measurements of MS[−] are important in the study of the atmospheric sulfur cycle, concentrations of the compound are usually low. Therefore, ΔC , the imbalance between cations and anions, can usually be simplified to:

$$\begin{aligned} \Delta C = & [\text{Na}^+] + [\text{NH}_4^+] + [\text{K}^+] + [\text{H}^+] + [\text{Ca}^{2+}] + [\text{Mg}^{2+}] \\ & - [\text{Cl}^-] + [\text{NO}_3^-] + [\text{SO}_4^{2-}] \end{aligned}$$

Likewise, the balance achieved between cations and anions ($\Delta C=0$), a useful measure for evaluating original chemical deposition forms, can be calculated as:

$$\begin{aligned} & [\text{Na}^+] + [\text{NH}_4^+] + [\text{K}^+] + [\text{H}^+] + [\text{Ca}^{2+}] + [\text{Mg}^{2+}] \\ & = [\text{Cl}^-] + [\text{NO}_3^-] + [\text{SO}_4^{2-}] \end{aligned}$$

Deposition of primary and secondary aerosols to the glacier or ice sheet surface occurs via several mechanisms, and therefore using soluble ion data to reconstruct past atmospheric chemical concentrations requires some knowledge of these processes. In the case of wet deposition, the aerosol falls within or is attached to a snowflake after either serving as a condensation nucleus or being scavenged by the falling snowflake. During dry deposition, the air-to-surface transfer occurs without any associated water transfer. To account for these processes when using ice core glaciochemical data, two procedures are common: (1) ice core time series of ion concentrations (e.g., parts per billion [ppb] or $\mu\text{eq L}^{-1}$) are interpreted directly, and (2) the estimated chemical flux (e.g., $\text{ng cm}^{-2} \text{ year}^{-1}$) is calculated and interpreted. Chemical fluxes are calculated by multiplying the sample concentration by the estimated H₂O flux (accumulation). In some cases, a qualitative examination of either concentration or flux time series yields similar results. For example, the deposition of Ca²⁺ on the Antarctic polar plateau and coastal sites was clearly higher during the Last Glacial Maximum (LGM) when either ice core

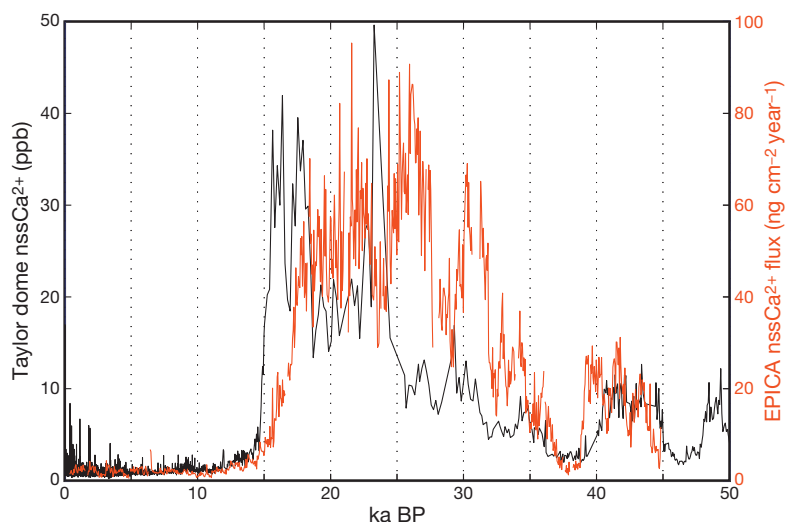


Figure 1 Nonsea salt (nss) Ca^{2+} concentrations (black) from the Taylor Dome, Antarctica ice core (Mayewski et al., 1996), and 50-year average nssCa^{2+} flux values (red) from the EPICA Dome C (EDC), Antarctica ice core (Rothlisberger et al., 2002).

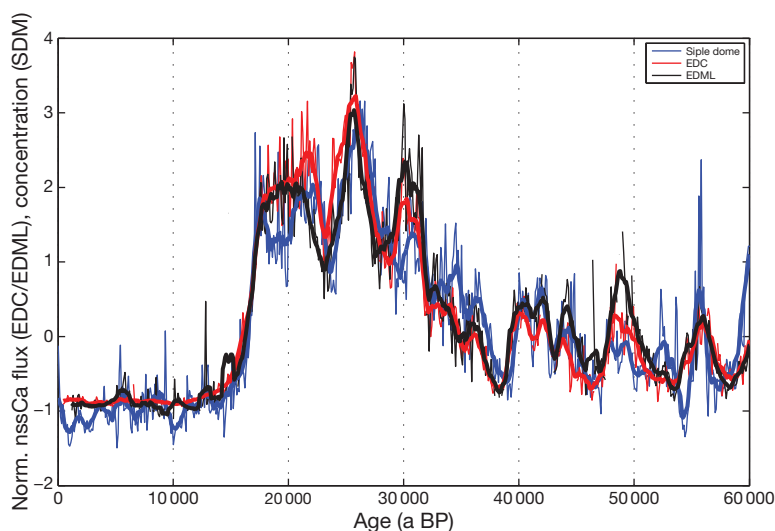


Figure 2 nssCa^{2+} fluxes from the EPICA Dome C (Fischer et al., 2007) and EPICA Dronning Maud Land (EDML; Fischer et al., 2007) ice cores, and nssCa^{2+} concentrations from the Siple Dome ice core, West Antarctica (Mayewski et al., 2009).

concentration or flux estimates are viewed (Figures 1 and 2). However, detailed quantitative analyses often require a more rigorous knowledge of site-specific air-snow-surface transfer functions, and justification for using either concentration or flux time series is necessary (e.g., Meeker et al., 1997). In addition, any postdepositional modification of the original chemical signal needs to be taken into account when reconstructing past atmospheric concentrations (e.g., Barnes et al., 2003).

Glaciochemical measurement techniques have evolved over the past ~30 years; however, the majority of soluble ion data have been and continue to be produced by ion chromatography (IC; e.g., Buck et al., 1992; Curran and Palmer, 2001; Legrand et al., 1984). Advances in IC technology have allowed much lower concentration measurements to be made with high precision, such that data produced from the most remote and pristine polar and alpine locations are now reasonably

routine. Analytical challenges still occur in certain situations, such as with samples collected from arid regions that have very high ion concentrations and particulate material that impede IC column performance. Decontamination of recovered ice cores is crucial to producing reliable glaciochemical data, and is as important as the chemical measurement itself. Early methods (still frequently used at high-accumulation sites) involved removing the outer portion of the core by rinsing with ultra-pure water, or by the physical removal of the outer core by scraping. More recently, several international groups have developed continuous melting systems that utilize a heated melting head that produces samples from the inner and outer portions of the core, thereby avoiding contamination from the outer core and providing the opportunity for high-resolution (small depth interval) measurements. Samples are either collected discretely (e.g., the melt water stream is collected into

individual sample vials at a specified interval; Osterberg et al., 2006), or injected directly into an instrument during continuous flow analysis (CFA; e.g., Fuhrer et al., 1993; Kaufmann et al., 2008; Rothlisberger et al., 2000; Sigg et al., 1994). When samples are collected discretely, standard IC techniques are used to analyze a full suite of ion species. In CFA, reagents are added to the sample stream, and measurement is typically via fluorometric (Na^+ , NH_4^+ , Ca^{2+}) or absorption (Na^+ , NO_3^-) techniques. A third analytical method has recently been developed known as fast ion chromatography (FIC), and is designed specifically for high-resolution ice core measurements, where a portion of the steady sample stream is injected into an IC at a specified time interval (Cole-Dai et al., 2006; Traversi et al., 2009; Wolff et al., 2006). FIC and CFA techniques currently have limitations in terms of the number of ion species that can be analyzed; however, as techniques evolve, other species will likely be added. Good agreement is found when ion concentrations measured on the same samples with the three different methods are compared (Littot et al., 2002). Soluble ion concentrations are typically reported in either mass-by-mass units (e.g., ppb), or molar units ($\mu\text{mol L}^{-1}$ or $\mu\text{eq L}^{-1}$).

Glaciochemical Sources and Spatial/Temporal Variability

The sources of soluble ion species in polar and alpine ice cores can be grouped into several general categories based on the original aerosol formation process. Because several ion species have multiple sources, statistical techniques are often applied to differentiate compounds responsible for ion deposition. In particular, multivariate techniques have been developed to deal with multiple chemical sources in a more quantitative fashion, in addition to accounting for factors inherent in ice core glaciochemical studies such as nonuniform sampling, nonstationary processes, and sporadic events (Meeker et al., 1995; Peixoto and Oort, 1992).

Sea Salt Aerosol

Sea salt aerosols are one of the major sources of impurities in polar ice cores, and contribute a significant portion of several species (notably Na^+ , Cl^- , SO_4^{2-} , and Mg^{2+}) to the ion balance. Originally, the source of the sea salt aerosol reaching coastal and inland plateau sites was thought to be primarily bubbles bursting over open ocean water. Because ion species associated with sea salt typically peak during winter, and also show elevated concentrations during glacial periods, the role of sea ice was not clear. Increased sea ice extent during winter and glacial periods should presumably lead to a decrease in the amount of sea salt aerosol reaching a particular site. One possibility for explaining this discrepancy is that increased storminess over the ocean and enhanced transport of sea salt aerosols inland during winter and glacial periods more than offsets the greater distance (Petit et al., 1999). Another possibility is that sea salt aerosol (particularly ssNa) is produced from newly formed sea ice, where highly saline brine and fragile frost flowers form a very effective source of aerosol in the winter (Wolff et al., 2003). A sea ice source is supported by the chemical signature of sea salt aerosols (e.g., depletion of SO_4^{2-}) during winter (e.g., Rothlisberger et al., 2010). Both of these factors (atmospheric dynamics and sea ice) likely contribute to sea salt deposition at a particular polar site, with each process being relatively more important on different timescales. Correlation of annually dated sea salt time series in Antarctica (Kreutz et al., 2000) and Greenland (Meeker and Mayewski, 2002) with sea level pressure records suggests a strong transport influence at these sites and timescales, while an ice core sea salt record from the Penny Ice Cap, Baffin Island is significantly correlated with local sea ice extent (Grumet et al., 2001). On glacial–interglacial timescales, sea salt records from Greenland (Figure 3; Mayewski et al., 1997) and Antarctica (Rothlisberger et al., 2002; Wolff et al., 2010) display significant concentration increases during glacial periods, suggesting the influence of both enhanced atmospheric transport

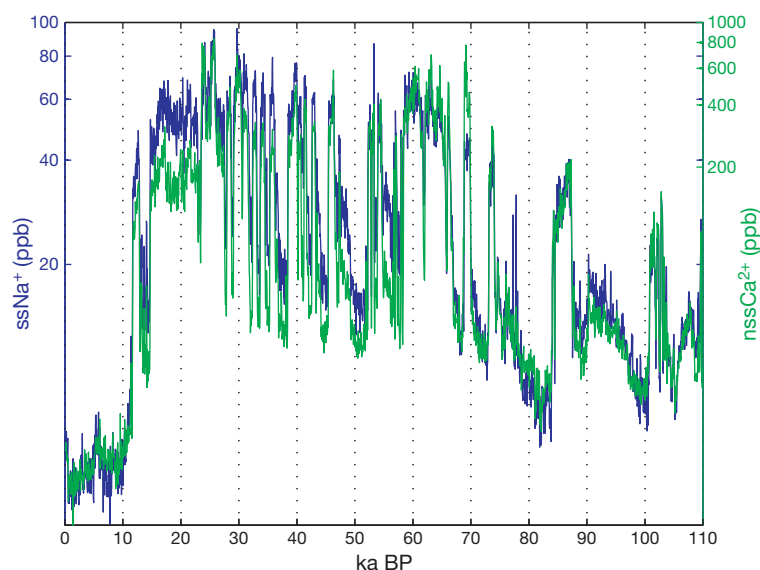


Figure 3 Sea salt (ss) Na^+ concentrations and nssCa^{2+} concentrations from the GISP2 ice core (Mayewski et al., 1997). Both datasets are 50-year averages.

and sea ice extent. At peak glacial conditions, however, ice core sea salt records may not be sensitive to sea ice variability; thus, incorporation of marine sediment core data can be useful (Rothlisberger et al., 2010). For high-elevation sites in mid- and low-latitude regions, the contribution of sea salt aerosol varies greatly depending on location relative to an ocean source. In Asia, most ice coring sites are located far from a marine source, and therefore, sea salt contributes little to the overall ion budget (Kang et al., 2004). In coastal ranges (e.g., St. Elias Mountains in Alaska/Yukon) or sites relatively close to the ocean (e.g., the Andes), the relative contribution of sea salt aerosol depends on elevation and accumulation rate (Thompson et al., 1998; Yalcin et al., 2006).

The calculation of the percentage of sea salt in any particular sample is an important step in determining the relative contribution to various species. The simplest calculation is based on the assumption that one ion species is derived solely from sea salt, and therefore standard seawater ion ratios can be used to calculate the sea salt (ss) and nonsea salt (nss) contribution for other species. In Antarctica, Na^+ is generally thought to have no other significant source, and after deposition behaves conservatively. In this case, calculation of nssCa^{2+} , for example, would be:

$$[\text{nssCa}^{2+} = \text{Ca}^{2+}] - [\text{Na}^+] \times [R_m]$$

where R_m is the $\text{Ca}^{2+}/\text{Na}^+$ ratio (0.038) found in marine aerosols. A slightly more sophisticated approach is to recognize that some Na^+ derives from other sources, and therefore, solve two equations simultaneously:

$$[\text{ssNa}^+] = [\text{Na}^+] - \frac{[\text{nssCa}^{2+}]}{R_c}$$

$$[\text{nssCa}^{2+}] = [\text{Ca}^{2+}] - [\text{ssNa}^+] \times R_m$$

where R_c is the $\text{Ca}^{2+}/\text{Na}^+$ ratio (1.78) found in average crust. Finally, models that estimate ss and nss components by an iterative process in which each sample is tested to determine which species is the most conservative (limiting) are useful for time-series calculations where the relative importance of sea salt aerosols may change (O'Brien et al., 1995). All of these methods assume that fractionation of sea salt aerosols during transport via reaction with acids, and possible loss after deposition, does not occur (Rothlisberger et al., 2003).

Terrestrial Dust

Dust records from polar and alpine ice cores have been widely used to infer past changes in atmospheric circulation and climatic/environmental conditions in dust source regions. Dust is important in the climate system because of its scattering effect on incoming solar radiation, and dust particles also serve as cloud condensation nuclei. Dust is also an important source of the trace nutrient iron (Fe) to the world's oceans, where it can have a fertilizing effect on primary production and nitrogen fixation (e.g., Martin et al., 1990; Moore et al., 2009). Thus, ice core measurements of dust allow us to estimate the magnitude of these effects through time. Measurement of dust (mineral aerosols) in polar ice cores can be accomplished in several ways, including both direct and proxy measurements. The most commonly used dust proxy is nssCa^{2+} , as the major portion of

nssCa^{2+} detected in ice cores comes from continental dust (Legrand and Mayewski, 1997) and it can be measured in a straightforward manner using IC. While it has been shown to record major changes in dust deposition on glacial–interglacial timescales (Figures 1–3) nssCa^{2+} varies in its reproducibility of insoluble particle measurements by up to a factor of 2 between low (interglacial) and high (glacial) dust concentrations (Ruth et al., 2008). This may indicate variable dust composition, and in fact, work in East Antarctica has found evidence of calcium carbonates at coastal sites (Sala et al., 2008). Ruth et al. (2008) suggest that quantitative reconstruction of dust deposition depends on the analysis method used.

In addition to nssCa^{2+} , the nss fractions of several other species (including Mg^{2+} , Na^+ , K^+ , SO_4^{2-} , NO_3^- , Cl^-) can be linked to insoluble silicate or soluble carbonate or evaporite (e.g., gypsum, halite) minerals derived from terrestrial sites. Snow samples and ice cores collected from sites located close to arid regions therefore have a correspondingly high dust load. High-elevation sites in Asia, either on or near the Tibetan Plateau, show elevated concentrations of all measured ions related to dust input (Thompson et al., 1997), as do European alpine sites influenced by Saharan dust (Schwikowski et al., 1999). Likewise, South American sites near the dry Altiplano region have enhanced ion concentrations during periods of lake desiccation (Thompson et al., 1998). In modern polar snow, the terrestrial dust contribution is usually small due to the large distance from available sources (Legrand and Mayewski, 1997), and in general dust only contributes a measurable nss fraction to Ca^{2+} and potentially Mg^{2+} and K^+ . Exceptions exist in ice-free regions such as the Dry Valleys, Antarctica, where exposed soils provide a source of soluble particulate material (Williamson et al., 2007). During the Holocene, changes in dust transport to Greenland are reflected in ice core nssCa^{2+} , nssMg^{2+} , and nssK^+ concentrations (O'Brien et al., 1995), and correlation with instrumental data as well as trace element isotopic work indicates that an Asian source is most probable (Bory et al., 2003; Meeker and Mayewski, 2002).

During the LGM, nssCa^{2+} concentrations show dramatic increases on the Antarctic (Figures 1 and 2) and Greenland (Figure 3) polar plateaus, as well as significant millennial-scale variability (Mayewski et al., 1996, 1997; Petit et al., 1999; Rothlisberger et al., 2002). In general, atmospheric dust loadings globally were a factor of 2–4 higher during glacial periods than during interglacials, with the northern hemisphere producing the majority of dust on a global scale (Fischer et al., 2007; Kohfeld and Harrison, 2002; Lambert et al., 2008; Maher et al., 2010). Modeling studies have suggested that changes in LGM climate (increased surface wind speed, reduced atmospheric and soil moisture, higher aerosol transport efficiency) and vegetation were largely responsible for the observed increases in dust deposition (e.g., Mahowald et al., 2006), while glacial activity may have been an important local factor (Sugden et al., 2009). However, a synthesis of empirical and modeling results that tests these various factors points to wind gustiness as a primary first-order driver of dust emissions during the LGM (McGee et al., 2010).

Volcanic Aerosols

Volcanic eruptions emit large amounts of particulate matter and gases (mainly SO_2) to different heights in the atmosphere

depending on eruption intensity and magma composition. Sulfate aerosols formed from the atmospheric oxidation of SO_2 and gas-to-particle conversions can be transported on regional to global scales, particularly when injected into the stratosphere (upper atmosphere), and are deposited on alpine glaciers and polar ice sheets. Sulfate ice core stratigraphy can be used to interpret the impact of individual eruptions on the atmospheric aerosol load, and any associated climate impacts. In addition to SO_4^{2-} , volcanic aerosols can also contribute other species to the overall ionic budget, including Cl^- and F^- (Herron, 1982). In the Northern Hemisphere, several ice core locations record both global and local/regional scale eruptions, particularly from the Icelandic and North Pacific regions (Zielinski et al., 1994). In the Southern Hemisphere, ice core records from coastal (Talos Dome) and West Antarctic sites (Kurbatov et al., 2006; Langway et al., 1995) also show a mixture of global and local/regional eruptions. In contrast, East Antarctic plateau ice cores generally are removed from any local-scale volcanic signals and therefore provide complementary estimates of explosive global-scale eruptions for use in bipolar comparisons (Cole-Dai et al., 2000). Several methods are used to estimate the contribution of volcanic nssSO_4^{2-} to the total SO_4^{2-} deposition, including smoothing and residual analysis, multivariate statistics, and definitions based on signals above a standard deviation cutoff. Isotopic analysis of sulfate can be used to differentiate among sources; for example, in West Antarctica, volcanic and stratospheric sources of sulfate dominate, while marine sources are more important in East Antarctica (e.g., Kunasek et al., 2010; Pruetz et al., 2004). Ice core-based reconstructions of past volcanic eruption frequency and intensity have, for example, identified an increase in volcanic activity in the early Holocene (Zielinski et al., 1994; Figure 4), possibly related to climatic cooling, and in the late Holocene (Castellano et al., 2005). In addition, ice core volcanic eruption histories are often used in climate forcing estimates input to climate models.

Biogenic Emissions

Biogenic emissions play a large role in the atmospheric sulfur cycle. In the remote unpolluted marine atmosphere, oxidation of dimethylsulfide (DMS) released from marine organisms represents a major source of nssSO_4^{2-} , both on local and global scales. In addition, the oxidation of DMS also produces methanesulfonic acid (measured as methylsulfonate; MS^-), and this formation pathway is the only known source of MS^- . Therefore, the measurement of both nssSO_4^{2-} and MS^- in ice cores represents a potential tool for deconvolving marine biogenic from nonbiogenic sulfur sources, and potentially for reconstruction of past changes in DMS emissions and hence ocean productivity (Legrand, 1995). Given that MS^- and nssSO_4^{2-} both exist as submicron aerosols, and therefore should have similar atmospheric transport and deposition processes, the ratio of $\text{MS}^-/\text{nssSO}_4^{2-}$ (R) in ice cores may provide a tool for estimating marine emissions. Such measurements are particularly relevant in Antarctica, given the surrounding open and ice-covered ocean regions, and to a somewhat lesser extent in Northern Hemisphere sites that are located near coastal regions. Recent sulfur aerosol and surface snow studies and modeling results have shown that R values vary spatially in Antarctica; seasonal values vary based on transport and chemical source region, and inland Antarctic ice cores sites and R values may provide large-scale estimates of bioproductivity. Therefore, with careful study of Antarctic snow and ice cores, R values may lead to an improved understanding of the oxidative capacity of the atmosphere, the temperature of the atmospheric oxidation of DMS, and the aging of marine air masses during their transport from source regions to Antarctica. Because the oxidation of DMS is the only known source of MS^- , several authors have used ice core MS^- concentrations to investigate past variability in the sulfur cycle, particularly the relationships among productivity, climate, and sea ice. MS^- concentrations from the Vostok ice core (Legrand et al.,

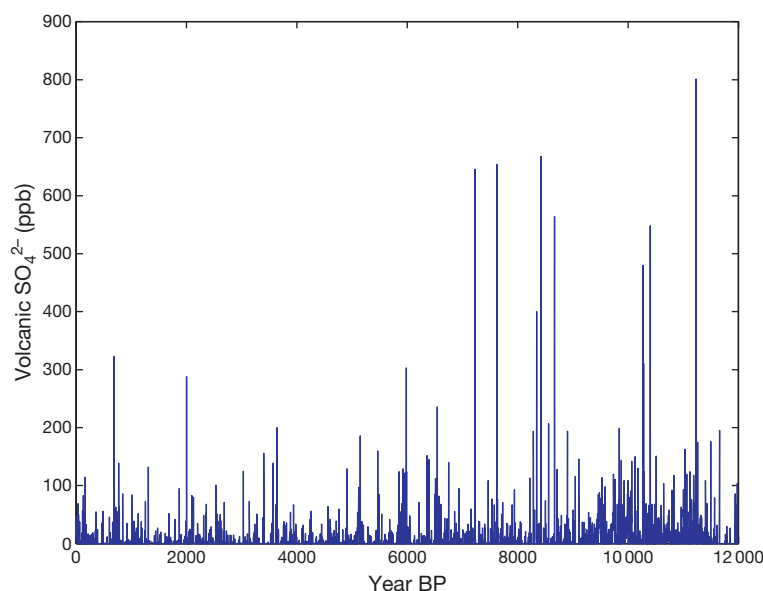


Figure 4 Volcanic SO_4^{2-} concentrations from the GISP2 ice core (Zielinski et al., 1994). The volcanic fraction of total SO_4^{2-} concentration are residuals calculated using a low-tension robust spline method.

1991) display an increase during the LGM, suggesting enhanced ocean productivity during periods of colder temperatures and decreased greenhouse gas concentrations. On shorter (interannual) timescales, correlations between South Pole ice core MS^- concentrations and the El Niño–Southern Oscillation have been found (Meyerson et al., 2002), indicating a link between low and high-latitude ocean/atmosphere circulation and marine productivity. Finally, several authors have found correlations between snow and ice core MS^- concentrations and sea ice extent (Abram et al., 2007; Curran et al., 2003; Rhodes et al., 2009), suggesting that algal productivity within and on top of sea ice may contribute to atmospheric DMS and hence influence ice core MS^- values.

Ice core NH_4^+ concentrations primarily arise from biological emissions of ammonia from plants, soils, and animals, bacterial decomposition, burning of biological materials (forest and grass fires), and potentially marine biological emissions. In the high-latitude Northern Hemisphere, NH_4^+ is present in significant amounts in precipitation, and ice core time series have been useful for reconstructing the history of forest fires from the boreal zone (Savarino and Legrand, 1998). On longer (multidecadal to glacial–interglacial) timescales, NH_4^+ concentrations in Greenland are mainly related to continental biogenic emissions from soils, and no evidence has been found for a significant marine contribution. A study of NH_4^+ concentrations in the Greenland Ice Sheet Project 2 (GISP2) core (Meeker et al., 1997) found that orbital parameters (summer forcing associated with the precessional cycle) and ice volume exerted a strong control on continental biogenic emissions over the past 110 000 years (Figure 5). In Antarctica, NH_4^+ concentrations in coastal aerosol and snow are highly variable and likely related to ornithogenic (penguin) soils, while oceanic emissions appear to play a minor role. Therefore, the low concentrations of NH_4^+ from inland sites remain

difficult both to measure and to interpret. In alpine regions, industrial and agricultural emissions likely impact snow and ice core NH_4^+ concentrations.

Nitrogen Cycling

To study past changes in atmospheric NO_x (NO and NO_2) concentration, nitrate (NO_3^-), an oxidation product of NO_x , has often been measured in polar ice cores (e.g., Wolff, 1995). Some features in the NO_3^- records can be readily explained, such as the clear increase since 1940 in Greenland ice cores attributed to NO_x emissions from industrialized countries (Mayewski et al., 1986). However, the interpretation of NO_3^- records beyond the anthropogenic era in Greenland, and in Antarctica in general, remains difficult. Several minor sources of NO_x have been suggested, including meteorite impact, supernovae, and solar modulation (sunspot cycle, solar proton events). The main sources are considered to be NO_x production in the stratosphere and troposphere by lightning. Recent studies of atmospheric NO_y (NO_x , HNO_3 , N_2O_5 , particulate and organic nitrates) have shown considerable amounts of organic nitrate, further complicating the interpretation of ice core NO_3^- time-series records. Studies have demonstrated that depositional and postdepositional processes have a strong influence on NO_3^- concentrations preserved in snow and ice cores (Wolff, 1995; Wolff et al., 2010 and references therein). In particular, it appears that changes in climate (temperature and accumulation rate) and atmospheric chemistry (Ca^{2+} concentrations) had a strong impact on the preservation of NO_3^- during the Holocene and LGM in the polar regions. At sites where potential NO_3^- sources may be close, changes in forest cover and local terrestrial biogeochemistry may explain glacial–interglacial NO_3^- concentration changes (Thompson et al., 1998).

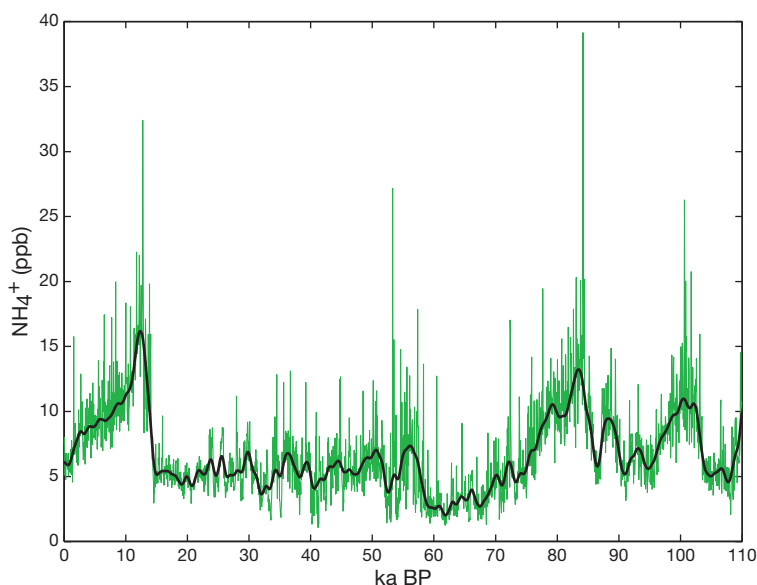


Figure 5 Ammonium (NH_4^+) concentration from the GISP2 ice core (Meeker et al., 1997). The green line is 50-year averaged data, and the black line is a low-tension robust spline chosen to highlight millennial-scale variability.

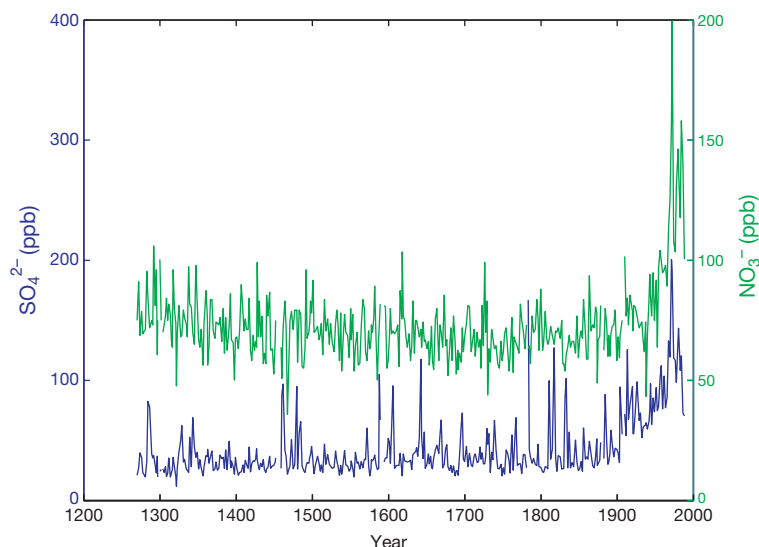


Figure 6 Sulfate (SO_4^{2-} ; blue) and nitrate (NO_3^- ; green) concentrations from the GISP2 ice core (Mayewski et al., 1986, 1990), showing a clear increase in the twentieth century due to anthropogenic emissions.

Anthropogenic Emissions

Emissions of SO_2 and NO_x from industrial processes and fossil fuel use have been a source for SO_4^{2-} and NO_3^- since the end of the nineteenth century in Greenland, the Canadian Arctic, Svalbard, and the European Alps (Schwikowski et al., 1999). For example, trends of increasing SO_4^{2-} and NO_3^- over the 50–100 years leading up to the 1970s are evident in data from South Greenland (Figure 6; Mayewski et al., 1986, 1990). In the Alps, population centers are located close to ice core sites, and therefore determining the sources responsible for SO_4^{2-} and NO_3^- is relatively straightforward (Schwikowski et al., 1999). In the Arctic, long-range transport of pollutant aerosol makes source identification more difficult due to mixing. Different approaches have been used to distinguish source contributions, including statistical analysis and time-series correlation of spatially distributed records (Goto-Azuma and Koerner, 2001). Results suggest that it may be possible to separate multiple anthropogenic sources from different regions in a single ice core record through the use of stable isotopic analysis. Long-term records of possible anthropogenic impact on atmospheric SO_4^{2-} and NO_3^- in Asia do not exist, in part because of the masking effect from large terrestrial dust inputs. However, stable sulfur isotope ratios in Asian snow suggest that anthropogenic SO_4^{2-} can be distinguished (Pruett et al., 2004). At the present time, there is no convincing evidence for an anthropogenic impact on SO_4^{2-} or NO_3^- concentrations in Antarctica.

See also: Ice Cores: History of Research, Greenland and Antarctica. Ice Core Methods: Biological Material; Chronologies; CO_2 Studies; Conductivity Studies; Methane Studies; Microparticle and Trace Element Studies; Overview; Stable Isotopes.

References

- Abram N, Mulvaney R, Wolff E, and Mudelsee M (2007) Ice core records as sea ice proxies: An evaluation from the Weddell Sea region of Antarctica. *Journal of Geophysical Research* 112: D15101.
- Barnes PRF, Wolff EW, Mader HM, Udisti R, Castellano E, and Rothlisberger H (2003) Evolution of chemical peak shapes in the Dome C: Antarctic, ice core. *Journal of Geophysical Research* 108: ACH 17-1–ACH 17-14.
- Bory A, Biscaye P, Piotrowski A, and Steffensen J (2003) Regional variability of ice core dust composition and provenance in Greenland. *Geochemistry, Geophysics, Geosystems* 4: 1107. <http://dx.doi.org/10.1029/2003GC000627>.
- Buck CF, Mayewski PA, and Whitlow SI (1992) Determination of major ions in snow and ice by ion chromatography. *Journal of Chromatography. A* 594: 225–228.
- Castellano E, Becagli S, Hansson M, et al. (2005) Holocene volcanic history as recorded in the sulfate stratigraphy of the European Project for Ice Coring in Antarctica Dome C (EDC96) ice core. *Journal of Geophysical Research* 110: D06114. <http://dx.doi.org/10.1029/2004JD005259>.
- Cole-Dai J, Budner D, and Ferris D (2006) High speed, high resolution, and continuous chemical analysis of ice cores using a melter and ion chromatograph. *Environmental Science & Technology* 40: 6764–6769.
- Cole-Dai J, Mosley-Thompson E, Wight S, and Thompson LG (2000) A 4100-year record of explosive volcanism from an East Antarctica ice core. *Journal of Geophysical Research* 105: 24431–24441.
- Curran MAJ and Palmer A (2001) Suppressed ion chromatography methods for the routine determination of ultra low level anions and cations in ice cores. *Journal of Chromatography A* 919: 107–113.
- Curran MAJ, Van Ommen T, Morgan V, Phillips KL, and Palmer AS (2003) Ice core evidence for Antarctic sea ice decline since the 1950's. *Science* 302: 1203–1206.
- Fischer H, et al. (2007) Reconstruction of millennial changes in dust emissions, transport, and regional sea ice coverage using the deep EPICA ice cores from the Atlantic and Indian Ocean sector of Antarctica. *Earth and Planetary Science Letters* 260: 340–354.
- Fuhrer K, Neff A, Ankin M, and Maggi V (1993) Continuous measurements of hydrogen peroxide, formaldehyde, calcium, and ammonium concentrations along the new GRIP ice core from Summit, Central Greenland. *Atmospheric Environment* 27A: 1873–1880.
- Grumet N, Wake C, Mayewski P, Zielinski G, Koerner R, and Fisher D (2001) Variability of sea-ice extent in the Baffin Bay over the last millennium. *Climatic Change* 49: 129–145.
- Herron MM (1982) Impurity sources of F^- , Cl^- , NO_3^- and SO_4^{2-} in Greenland and Antarctic precipitation. *Journal of Geophysical Research* 87: 3052–3060.
- Kang S, Mayewski P, Qin D, Sneed S, Ren J, and Zhang D (2004) Seasonal differences in snow chemistry from the vicinity of Mt. Everest, central Himalayas. *Atmospheric Environment* 38: 2819–2829.

- Kaufmann PR, Federer U, Hutterli MA, et al. (2008) An improved continuous flow analysis system for high-resolution field measurements on ice cores. *Environmental Science & Technology* 42: 8044–8050.
- Kohfeld KE and Harrison RM (2002) DIRTMAP: The geological record of dust. *Earth-Science Reviews* 54(1–3): 81–114.
- Kreutz KJ, Mayewski PA, Pittalwala II, Meeker LD, Twickler MS, and Whitlow SI (2000) Sea-level pressure variability in the Amundsen Sea region inferred from a West Antarctic glaciochemical record. *Journal of Geophysical Research* 105: 4047–4059.
- Kunasek S, Alexander B, Steig E, et al. (2010) Sulfate sources and oxidation chemistry over the past 250 years from sulfur and oxygen isotopes of sulfate in a West Antarctic ice core. *Journal of Geophysical Research* 115: D18313. <http://dx.doi.org/10.1029/2010JD013846>.
- Kurbatov A, Zielinski G, Dunbar N, et al. (2006) A 12,000 year record of explosive volcanism in the Siple Dome ice core, West Antarctica. *Journal of Geophysical Research* 111: D12307. <http://dx.doi.org/10.1029/2005JD006072>.
- Lambert F, Delmonte B, Petit J, et al. (2008) Dust-climate couplings over the past 800,000 years from the EPICA Dome C ice core. *Nature* 452: 616–619.
- Langway CCJ, Osada K, Clausen HB, Hammer CU, and Shoji H (1995) A 10-century comparison of prominent bipolar volcanic events in ice cores. *Journal of Geophysical Research* 100: 16241–16247.
- Legrand M (1995) Sulphur-derived species in polar ice: A review. In: Delmas R (ed.) *Ice Core Studies of Global Biogeochemical Cycles*, pp. 91–120. New York: Springer.
- Legrand M, De Angelis M, and Delamas RJ (1984) Ion chromatographic determination of common ions at ultratrace levels in Antarctic snow and ice. *Analytica Chimica Acta* 156: 181–192.
- Legrand M, Feniet-Saigne C, Saltzman ES, Germain C, Barkov NI, and Petrov VN (1991) Ice-core record of oceanic emissions of dimethylsulphide during the last climate cycle. *Nature* 350: 144–146.
- Legrand M and Mayewski PA (1997) Glaciochemistry of polar ice cores: A review. *Reviews of Geophysics* 35: 219–243.
- Littot GC, Mulvaney R, Rothlisberger R, et al. (2002) Comparison of analytical methods used for measuring major ions in the EPICA Dome C (Antarctica) ice core. *Annals of Glaciology* 35: 299–305.
- Maier BA, Prospero J, Mackie D, Gaiero D, Hesse P, and Balkanski Y (2010) Global connections between aeolian dust, climate and ocean biogeochemistry at the present day and at the last glacial maximum. *Earth-Science Reviews* 99: 61–97.
- Mahowald N, Muhs D, Levis S, et al. (2006) Change in atmospheric mineral aerosols in response to climate: Last glacial period, preindustrial, modern, and doubled carbon dioxide climates. *Journal of Geophysical Research* 111: D10202. <http://dx.doi.org/10.1029/2005JD006653>.
- Martin JH, Gordon RM, and Fitzwater SE (1990) Iron in Antarctic waters. *Nature* 345: 156–158.
- Mayewski PA, Lyons WB, Spencer MJ, Twickler MS, Buck CF, and Whitlow S (1990) An ice-core record of atmospheric response to anthropogenic sulfate and nitrate. *Nature* 346: 554–556.
- Mayewski PA, Lyons WB, Spencer MJ, et al. (1986) Sulfate and nitrate concentrations from a South Greenland ice core. *Nature* 232: 975–977.
- Mayewski PA, Meredith MP, Summerhayes CP, et al. (2009) State of the Antarctic and Southern Ocean climate system. *Reviews of Geophysics* 47: RG1003, 38 pp.
- Mayewski PA, Meeker LD, Twickler MS, et al. (1997) Major features and forcing of high latitude northern hemisphere atmospheric circulation over the last 110,000 years. *Journal of Geophysical Research* 102: 26345–26366.
- Mayewski PA, Twickler MS, Whitlow SI, et al. (1996) Climate change during the deglaciation in Antarctica. *Science* 272: 1636–1638.
- McGee D, Broecker W, and Winckler G (2010) Gustiness: The driver of glacial dustiness? *Quaternary Science Reviews* 29: 2340–2350.
- Meeker LD and Mayewski PA (2002) A 1400 year long record of atmospheric circulation over the North Atlantic and Asia. *The Holocene* 12: 257–266.
- Meeker LD, Mayewski PA, and Bloomfield P (1995) A new approach to glaciochemical time series analysis. In: Delmas RJ (ed.) *Ice Core Studies of Global Biogeochemical Cycles*. Berlin: Springer-Verlag.
- Meeker LD, Mayewski PA, Twickler MS, Whitlow SI, and Meese D (1997) A 110,000-year history of change in continental biogenic emissions and related atmospheric circulation inferred from the Greenland Ice Sheet Project ice core. *Journal of Geophysical Research* 102: 26489–26504.
- Meyerson E, Mayewski PA, Kreutz KJ, Meeker LD, Whitlow SI, and Twickler MS (2002) The polar expression of ENSO and sea-ice variability as recorded in a South Pole ice core. *Annals of Glaciology* 35: 430–436.
- Moore CM, Mills MM, Achterberg EP, et al. (2009) Large-scale distribution of Atlantic nitrogen fixation controlled by iron availability. *Nature Geoscience* 2: 861–871.
- O'Brien SR, Mayewski PA, Meeker LD, Meese DA, Twickler MS, and Whitlow SI (1995) Complexity of Holocene climate as reconstructed from a Greenland ice core. *Science* 270: 1962–1963.
- Osterberg EO, Handley M, Sneed S, Mayewski P, and Kreutz K (2006) Continuous ice core melter system with discrete sampling for major ion, trace element, and stable isotope analysis. *Environmental Science & Technology* 40: 3355–3361.
- Peixoto J and Oort A (1992) *Physics of Climate*, 520 pp. New York, USA: American Institute of Physics.
- Petit JR, Jouzel J, Raynaud D, et al. (1999) Climate and atmospheric history of the past 420,000 years from the Vostok ice core, Antarctica. *Nature* 399: 429–436.
- Pruett L, Kreutz K, Mayewski P, and Wadleigh M (2004) Sulfur isotopic measurements from a West Antarctic ice core: Implications for sulfate source and transport. *Annals of Glaciology* 39: 161–165.
- Rhodes RH, Bertler N, Baker J, Sneed S, Oerter H, and Arrigo K (2009) Sea ice variability and primary productivity in the Ross Sea, Antarctica, from methylsulphonate snow record. *Geophysical Research Letters* 36: L10704, 5pp. <http://dx.doi.org/10.1029/2009GL037311>.
- Rothlisberger R, Bigler M, Hutterli M, Sommer S, and Stauffer B (2000) Technique for continuous high-resolution analysis of trace substances in firn and ice cores. *Environmental Science & Technology* 34: 338–342.
- Rothlisberger R, Crosta X, Abram N, Armand L, and Wolff E (2010) Potential and limitations of marine and ice core sea ice proxies: An example from the Indian Ocean sector. *Quaternary Science Reviews* 29: 296–302.
- Rothlisberger R, Mulvaney R, Wolff EW, et al. (2002) Dust and sea salt variability in central East Antarctica (Dome C) over the last 45 kyr and its implications for southern high-latitude climate. *Geophysical Research Letters* 29: 1963–1966.
- Rothlisberger R, Mulvaney R, Wolff E, et al. (2003) Limited dechlorination of sea-salt aerosols during the last glacial period: Evidence from the European Project for Ice Coring in Antarctica (EPICA) Dome C ice core. *Journal of Geophysical Research* 108: 4256 <http://dx.doi.org/10.1029/2003JD003604>.
- Ruth U, et al. (2008) Proxies and measurement techniques for mineral dust in Antarctic ice cores. *Environmental Science & Technology* 42: 5675–5681.
- Sala M, Delmonte B, Frezzotti M, et al. (2008) Evidence of calcium carbonates in coastal (Talos Dome and Ross Sea area) East Antarctica snow and firn: Environmental and climatic implications. *Earth and Planetary Science Letters* 271: 43–52.
- Savarino J and Legrand M (1998) High northern latitude forest fires and vegetation emissions over the last millennium inferred from the chemistry of a central Greenland ice core. *Journal of Geophysical Research* 103: 8267–8279.
- Schwikowski M, Brüttsch S, Gäggeler HW, and Schotterer U (1999) A high-resolution air chemistry record from an Alpine ice core: Fiescherhorn glacier, Swiss Alps. *Journal of Geophysical Research* 104: 13709–13719.
- Sigg A, Fuhrer K, Anklin M, Staffelbach T, and Zurmühle D (1994) A continuous analysis technique for trace substances in ice cores. *Environmental Science & Technology* 28: 204–209.
- Sugden D, McCulloch R, Bory A, and Hein A (2009) Influence of Patagonian glaciers on Antarctic dust deposition during the last glacial period. *Nature Geoscience* 2: 281–285.
- Thompson LG, Davis ME, Mosley-Thompson E, et al. (1998) A 25,000-year tropical climate history from Bolivian ice cores. *Science* 282: 1858–1864.
- Thompson LG, Yao T, Davis ME, et al. (1997) Tropical climate instability: The last glacial cycle from a Qinghai-Tibetan ice core. *Science* 276: 1821–1825.
- Traversi R, et al. (2009) Sulfate spikes in the deep layers of EPICA-Dome C ice core: Evidence of glaciological artifacts. *Environmental Science & Technology* 43: 8737–8743.
- Williamson B, Kreutz K, Mayewski P, et al. (2007) A coastal transect of McMurdo Dry Valleys (Antarctica) snow and firn: Marine and terrestrial influences on glaciochemistry. *Journal of Glaciology* 53: 681–685.
- Wolff EW (1995) Nitrate in polar ice. In: Delmas RJ (ed.) *Ice Core Studies of Global Biogeochemical Cycles*, NATO ASI Series. New York: Springer-Verlag.
- Wolff E, Rankin AM, and Rothlisberger R (2003) An ice core indicator of Antarctic sea ice production? *Geophysical Research Letters* 30: 2158. <http://dx.doi.org/10.1029/2003GL01854>.
- Wolff E, et al. (2006) Southern Ocean sea-ice extent, productivity, and iron flux over the past eight glacial cycles. *Nature* 440: 491–496.
- Wolff E, et al. (2010) Changes in environment over the last 800,000 years from chemical analysis of the EPICA Dome C ice core. *Quaternary Science Reviews* 29: 285–295.
- Yalcin K, Wake C, Kang S, and Kreutz K (2006) Seasonal and spatial variability in snow chemistry at Eclipse Icefield, Yukon, Canada. *Annals of Glaciology* 43: 200–204.
- Zielinski GA, Mayewski PA, Meeker LD, et al. (1994) Record of volcanism since 7000 B.C. from the GISP2 Greenland ice core and implications for the volcano-climate system. *Nature* 264: 948–952.

Methane Studies

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Introduction

Current Understanding of the CH₄ Cycle

Methane (CH₄) is a trace compound of the Earth's atmosphere with a globally averaged mixing ratio of 1800 ppbv, representing a total load of 5000 Tg (1 Tg = 10¹² g). It is removed from the atmosphere mainly (85–90%) through oxidation with the hydroxyl radical OH. The reaction rate between the two gases is slow enough so that the lifetime of CH₄ (time required to reduce its atmospheric mixing ratio by a factor *e*) today is 9.6 years. Therefore, differences of tropospheric mixing ratios as a function of latitude and longitude exist but are not more than typically 10% of the average. Five to ten percent of CH₄ is oxidized in the stratosphere and another 5% is oxidized in soils by methanotrophic archaea.

Under natural conditions, methane is mainly produced in the continental biosphere, through decomposition of organic matter under anoxic conditions. These conditions are met in wetlands, during the enteric fermentation in ruminants and in termites. In addition, CH₄ is produced during incomplete combustion of terrestrial biomass and by degassing of natural gas reservoirs (from coal and gas deposits, volcanoes and hydrates in permafrost). The world oceans are small contributors of CH₄ to the atmosphere (at the most 5–10 Tg year⁻¹). Methane then comes either from anoxic layers of the oceans or from hydrate degassing at the sediment/water interface. Altogether, the natural methane sources represent between 180 and 250 Tg year⁻¹, depending on the estimates. It was shown that CH₄ can also be directly produced by plants under oxic conditions (Keppler et al., 2006), but there is ongoing debate on the global strength of this source. A thorough review of the CH₄ cycle is provided in the IPCC AR4 report (Denman et al., 2007).

Reconstructing CH₄ Levels During the Quaternary

Although in the case of CO₂, other media – such as carbon isotopic ratios of organic compounds in bogs or boron isotopes in ocean sediments – are proposed to reconstruct its Quaternary trend, for CH₄ the only way to access its past mixing ratio is to measure it in bubbles trapped in polar ice sheets. Measurements can also be performed on ice from tropical glaciers but their interpretation is less straightforward due to local artifacts. Summer melting and percolation of the meltwater can modify the gas composition by solubility effect; and there are indications that methanogenesis by methanogenic archaea probably transported and deposited with atmospheric aerosols can also take place in these conditions (Campen et al., 2003).

Polar ice-core CH₄ records encompass now the last 800 000 years (Louergue et al., 2008). The time resolution of the records ranges from a decade (industrial increase) to a couple of millennia (oldest glacial–interglacial cycles in the available records). Different analytical techniques have been

applied for extracting the gas from the ice and quantifying the CH₄ content as well as its carbon and hydrogen isotopic ratios (see section 'Analytical Techniques'). The typical reproducibility obtained on replicate measurements of neighboring samples lies between 4 and 10 ppbv, that is, 1–3% of the maximum amplitude of mixing ratio changes. It usually amounts between 0.1 and 0.3‰ for ¹³C/¹²C and to 4‰ for D/H, that is, 5–25% of the full scale of CH₄ isotopic ratio changes measured so far.

The characteristics of the main ice-core drilling sites used for CH₄ studies are given elsewhere in this encyclopedia. Here, extra details on the firn densification process and gas diffusion are provided. They are crucial phenomenon to determine (1) the phasing between atmospheric composition changes, recorded in the bubbles, and the climate changes, recorded in the ice matrix, and (2) the smoothing of atmospheric variations during the gas trapping in the ice.

Ice forms in polar regions by compacting and then sintering of snow grains, under the weight of subsequently fallen snow. The speed of densification depends primarily on snow temperature and on accumulation rate at the surface. This leads to a firn/ice transition ranging between 60 and 120 m of depth below the surface and a Δ-age (mean difference of age between the trapped gas and the surrounding ice) of 30 years (present-day conditions at the high accumulation rate site of Law Dome, Antarctica) to more than 7000 years (very dry glacial conditions at Vostok, Antarctica). Based on current knowledge of the physics of snow and firn densification, it has been possible to develop models simulating the firn densification and Δ-age under various conditions. Uncertainties on past temperature and accumulation rates are the main factors limiting our ability to accurately calculate Δ-age in the past. At Greenland Ice Core Project (GRIP) and Greenland Ice Sheet Project 2 (GISP2) in Greenland, the uncertainty around Δ-age calculations is about 100 years. At EPICA Dome C (EDC, Antarctica) or Vostok, it amounts to about 1000 years. Empirical Δ-age evaluations at EDC and EPICA Dronning Maud Land (EDML, Antarctica) under glacial conditions suggest that the densification models tend to slightly overestimate the true Δ-age (Louergue et al., 2007), thus also questioning our understanding of physical processes at work during firnification under drier and colder conditions than today in the Antarctic plateau.

An analytical method developed in the 1990s provided an evaluation of the Δ-age independently from the firn densification models: it relies on the anomalies of permanent gas stable isotopic ratios (¹⁵N/¹⁴N of N₂, ⁴⁰Ar/³⁶Ar) in the bubbles, generated by abrupt temperature changes at the surface of the ice sheet. Temperature gradients between the atmosphere and the firn/ice transition produce gas thermal diffusion and fractionation of these isotopes with depth, as gases diffuse about ten times faster than heat in snow. It thus provides a climate record in the gas phase, shifted by Δ-age with the same record in the ice phase (water isotopes). The method

has been applied with success in Greenland where rapid and large temperature changes took place in the past (e.g., Landais et al., 2004; Severinghaus and Brook, 1999), but its application to Antarctica remains so far sporadic and difficult, due to the lack of rapid (decadal to centennial) and ample temperature changes in the past.

When bubbles are formed within an ice density range of 0.78–0.84 g cm⁻³, the age of the gas inside the bubble is not zero; air molecules must diffuse through the thick porous snow and firn column before getting trapped. It requires from 10 to 100 years for CH₄ molecules to diffuse from the atmosphere down to the closing bubbles in the firn. The diffusion process is mainly driven by the gradient of gas mixing ratio as a function of depth, by the tortuosity of the firn structure, and by temperature gradient between the surface and the firn/ice transition (Schwander, 1996). It is also influenced by gravitation, enriching the deepest portion of the firn with the heaviest gas compounds. For CH₄, gravitation reduces its mixing ratio at the base of the firn by only 0.7% at maximum compared with the atmosphere. This effect is thus neglected in most studies.

Air diffusion smooths out part of the atmospheric composition variability. On the Antarctic plateau for instance, the age distribution of air molecules (with respect to the first time when they crossed the atmosphere/surface snow boundary) at the base of the firn typically covers several decades. The progressive enclosure of bubbles in the density range 0.78–0.84 is another process smoothing atmospheric variations. This transition of density requires from 10 to 1000 years, depending on the accumulation rate. As a consequence, CH₄ records associated with rapid atmospheric changes and obtained from ice cores with different accumulation rates can significantly differ from each other. For instance a comparison between CH₄ records from the GRIP and EDC ice cores covering the methane drop associated with the rapid 8.2-ka BP cooling showed that the latter core attenuated the amplitude of the CH₄ drop by about 50% (Spahni et al., 2003). A precise quantification of this phenomenon under glacial conditions has yet to be performed.

The Last Millennium

Since the late 1980s, several ice-core records of the CH₄ mixing ratio have been obtained from Greenland and Antarctica covering the last millennium, depicting a relatively stable level around 700 ppbv followed by the dramatic anthropogenic increase over the last two centuries. As in the case of CO₂, using the most detailed and accurate record of Law Dome in Antarctica, ice-core measurements also cover the most recent time up to the late 1970s (MacFarling Meure et al., 2006), providing an excellent overlap with mixing ratios obtained on the Cape Grim air archive (pressurized tanks), and nicely extending the direct atmospheric measurements at South Pole. The overlap is confirmed by measurements of CH₄ in interstitial firn air at the same site. This is a clear indication that polar ice cores accurately record the atmospheric CH₄ mixing ratio.

The ~1100 ppbv increase of atmospheric CH₄ is accompanied by an increase of its ¹³C/¹²C stable isotopic ratio by about 2‰, and by about 15‰ for D/H of CH₄ (Figure 1). It traces the

increased ratio of average anthropogenic sources (rice cultivation, ruminants, landfills, biomass burning, coal and gas extraction, and distribution) over natural sources (Sowers et al., 2005).

The time period between AD 1000 and 1800 shows small but significant fluctuations of the CH₄ mixing ratio, both over Greenland and over Antarctica, with an amplitude of ~50 ppbv. They seem to be roughly correlated with the medieval warm period and the Little Ice Age, two climatic phenomena which could have directly affected natural sources such as biomass burning and wetlands, but also early anthropogenic sources. Measurements of the ¹³C/¹²C isotopic ratio of CH₄ over the last 2000 years reveal an enrichment by about 2‰ between AD 0 and 1000 compared with the time period AD 1700–1900, whereas the CH₄ mixing ratio does not show concomitant and large changes (Ferretti et al., 2005; Mischler et al., 2009). A parallel enrichment of D/H (by about 10‰) is also observed (Mischler et al., 2009). This is interpreted as an increased contribution of biomass burning emissions to the CH₄ budget (and contemporary reduced wetland emissions) during the earlier period. Agricultural sources of CH₄ would have progressively increased with time. An alternative explanation of the parallel evolution of CH₄ and its ¹³C/¹²C isotopic ratio could be related to the proposed CH₄ production by plants and changes of primary productivity during the last two millennia (Houweling et al., 2008; Kepler et al., 2006). However, recent investigations of carbon monoxide mixing ratio and isotopic ratio changes (tracing different sources including biomass burning) recorded in Antarctic ice cores over the last millennium reinforce the scenario of large biomass burning emission changes controlling the millennial CH₄ evolution, at least in the Southern hemisphere (Wang et al., 2010).

The Holocene

High-resolution CH₄ records have been obtained over the Holocene from many Antarctic (D47, Byrd, Taylor Dome, EPICA cores, Talos Dome) and Greenland (GRIP, GISP2) ice cores (e.g., Blunier et al., 1995; Brook et al., 1999; Chappellaz et al., 1997; Schilt et al., 2010). They all show a much larger amplitude of change than during the last millennium: methane peaks at an average of ~700 ppbv during the early Holocene (10 000–11 500 years BP) and again during the last millennium. In between, it shows lower levels, with a minimum of ~570 ppbv centered at 5200 years BP (Figure 2). In addition, a decrease by ~80 ppbv is observed 8200 years ago, lasting about 300 years and synchronous with an abrupt cooling event recorded at the same time in the northern hemisphere (Spahni et al., 2003).

The comparison between Greenland and Antarctic records allows one to quantify the inter polar CH₄ difference. This difference results from the uneven distribution of CH₄ sources with latitude and from the fact that the residence time of methane in the atmosphere is only about ten times longer than the average atmospheric exchange time between the two hemispheres. Over the Holocene, on average, the inter polar difference amounts to 44 ± 7 ppbv but it also shows significant variations of the order of 10 ppbv. When translated into a

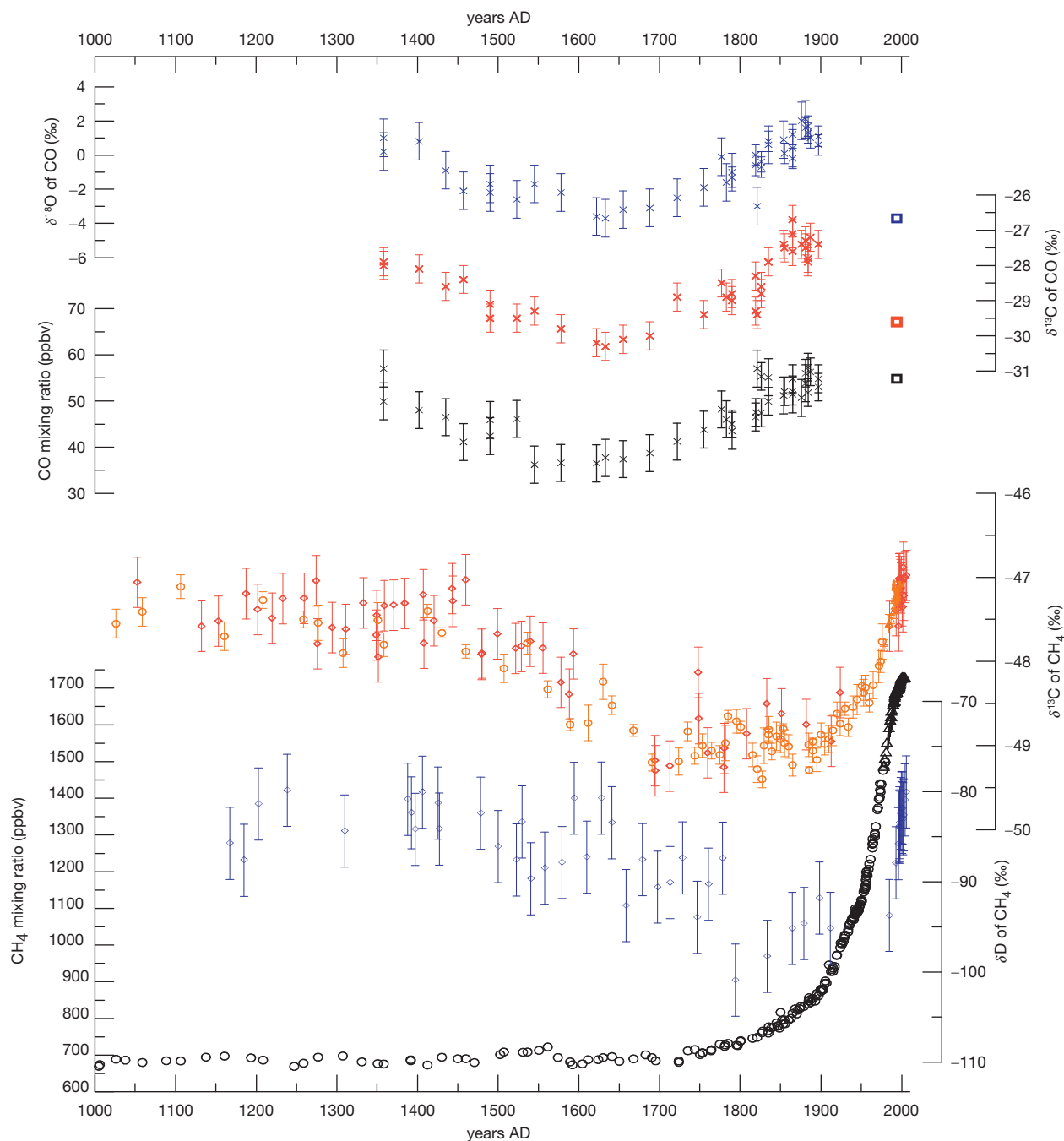


Figure 1 Atmospheric CH₄ evolution over the last millennium from the Law Dome (Antarctica) ice cores (black circles; MacFarling Meure et al., 2006). Black triangles correspond to observations on pressurized tanks of the Cape Grim air archive; the black line runs through NOAA/CMDL atmospheric measurements performed at the South Pole. Blue and red diamonds indicate D/H and ¹³C/¹²C changes in atmospheric CH₄, respectively, measured along the WAIS Divide ice core (Antarctica; Mischler et al., 2009). Orange circles are ¹³C/¹²C changes measured on the Law Dome ice cores (Ferretti et al., 2005). The upper three graphs show the concomitant evolution of the CO mixing ratio in Antarctica (black symbols) as well as its ¹³C/¹²C (red symbols) and ¹⁸O/¹⁶O (blue symbols) changes (Wang et al., 2010). 1994 values are indicated with the squared symbol. CO affects the oxidative capacity of the atmosphere (and thus the CH₄ budget) and shares biomass burning as a common source with CH₄. Error bars are 1σ uncertainties. Each record is on its own time scale.

simple three-box model of the atmosphere, the Holocene variations of the CH₄ inter-pole difference suggest that tropical sources (possibly tropical wetlands and biomass burning) dominated the Holocene CH₄ budget and that northern

sources would have contributed to the higher mixing ratios of the early Holocene (Brook et al., 1999; Chappellaz et al., 1997). The CH₄ increase occurring mainly during the last 2500 years, combined with a reduced inter-pole gradient during

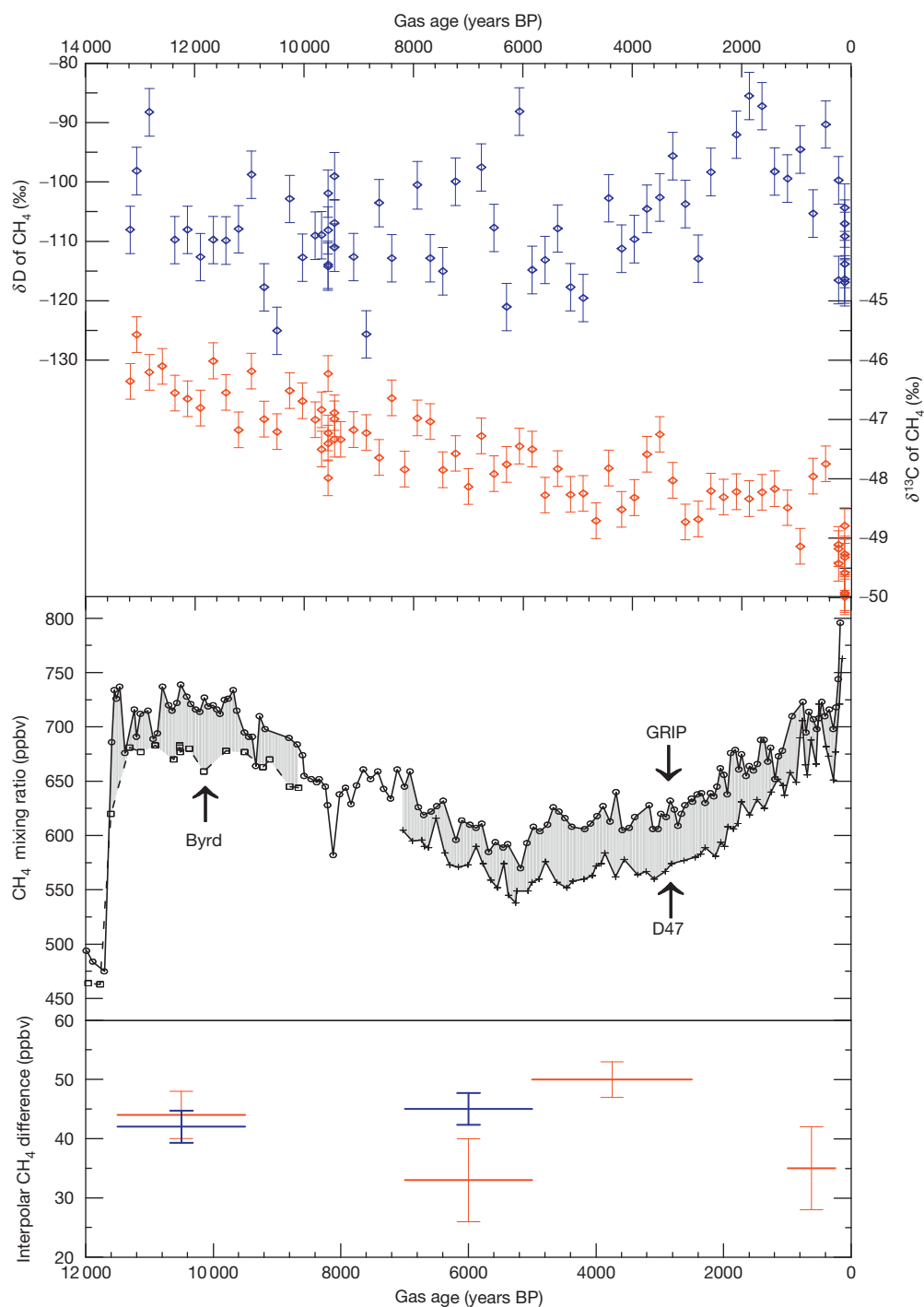


Figure 2 CH₄ mixing ratio over the Holocene, measured on Greenland (GRIP) and Antarctic (D47 and Byrd) ice cores. The shaded area depicts the inter-polar CH₄ gradient. Its average evolution through time is shown in the lower panel with 1 σ error bars (red: Chappellaz et al., 1997; blue: Brook et al., 1999, using the Greenlandic GISP2 and Antarctic Taylor Dome ice cores). Upper panel: blue and red diamonds indicate D/H and ¹³C/¹²C changes in atmospheric CH₄, respectively, measured along the GISP2 ice core (Greenland; Sowers, 2010). Error bars are 1 σ uncertainties. Each record is on its own time scale.

the last millennium, suggests that early anthropogenic emissions (through rice cultivation, livestock and human-induced biomass burning) could have already significantly contributed to the preindustrial CH₄ budget (Chappellaz et al., 1997; Ruddiman and Thomson, 2001).

Recent investigations of the CH₄ stable isotopic ratios over Greenland and covering the Holocene period provide an additional story about the possible mechanisms involved. A progressive decrease of the ¹³C/¹²C isotopic ratio (by about 2‰) with time could reflect a growing contribution of Arctic

ecosystems relatively depleted in ^{13}C , and/or a progressive shift from C4 to C3 dominating plants in wetlands (Sowers, 2010). The D/H evolution of atmospheric CH_4 bears more sample-to-sample variability but mostly shows a $\sim 20\%$ positive shift between 4000 and 1800 years BP (Figure 2). Potential candidates to explain this trend are a shift to stronger CH_4 emissions from the tropics in recent times, or a gradual release of marine clathrates enriched in D versus H (Sowers, 2010).

The Last Glaciation and Deglaciation

Very detailed CH_4 records (with a time resolution reaching two or three decades at specific periods for some cores) have been obtained from the GRIP, GISP2, and NorthGRIP ice cores in Greenland, and the Byrd, EDC, EDML, Vostok, Dome Fuji, Taylor Dome, Siple Dome, and Talos Dome in Antarctica, covering the last glaciation and deglaciation. The warming of the last deglaciation is associated with the largest observed increase of atmospheric CH_4 since the last glaciation. It doubles over Greenland, starting from an average of 365 ± 7 ppbv between 17 000 and 20 000 years BP, and reaching an average of 718 ± 3 ppbv between 11 500 and 9500 years BP. No significant inter polar gradient (-3 ± 4 ppbv) is observed during the minimum of the Last Glacial Maximum (Dällenbach et al., 2000).

The four distinct intervals punctuating the CO_2 increase during the deglaciation find counterparts in the CH_4 trend. Methane levels start to increase at 17 000 years BP, in synchronicity with CO_2 . After a slow increase of ~ 100 ppbv until 14 500 years BP, its mixing ratio suddenly increases by another ~ 200 ppbv within 100 years. It then remains at ~ 650 ppbv during the entire Bolling/Allerød warm period. The following time interval, corresponding to the Younger Dryas cold period, is accompanied by a CH_4 drop of comparable amplitude and speed as the first rapid increase; the Younger Dryas ends with again a ~ 200 ppbv CH_4 increase within 100 years, leading to the early Holocene maximum of ~ 700 ppbv. A direct time comparison of the rapid CH_4 changes of the deglaciation with thermal anomalies in permanent gas isotopes has shown that the start of the CH_4 and Greenland climate events is synchronous or with a small delay of CH_4 by a maximum of 30 years (Severinghaus and Brook, 1999). On the other hand, the duration of the change is longer for methane than for Greenland temperature.

The remarkable correspondence between atmospheric CH_4 changes and Greenland climate fluctuations during the last deglaciation is indeed a feature characterizing the last glaciation (Figure 3). Each of the Dansgaard/Oeschger events recorded in Greenland ice has a counterpart in atmospheric CH_4 levels, although with a variable relationship in terms of amplitude (Brook et al., 1996; Chappellaz et al., 1993b). The CH_4 variations cover a range from 50 to 200 ppbv. As in the

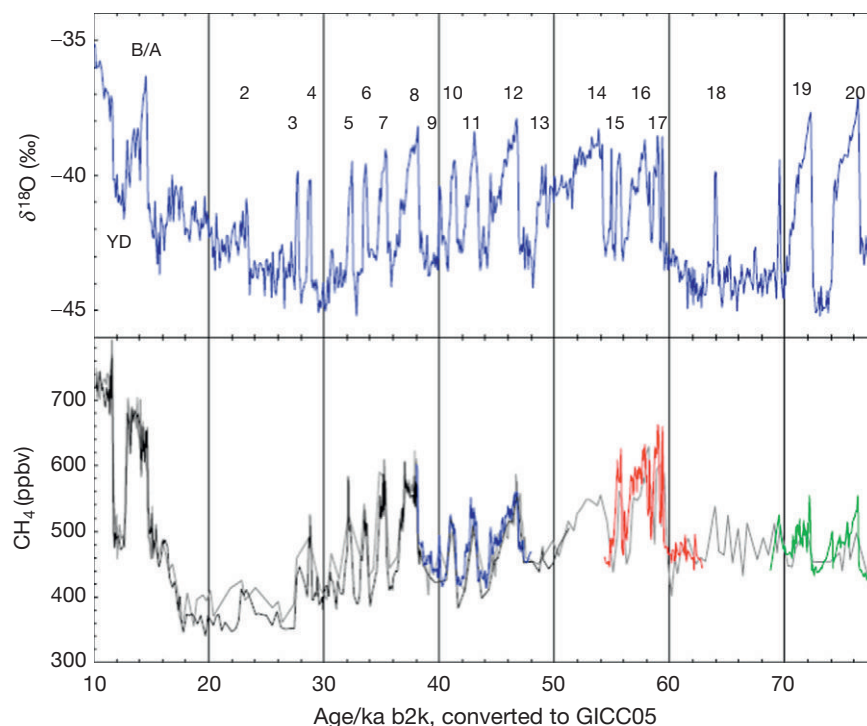


Figure 3 Lower curve: CH_4 mixing ratio recorded in Greenland ice, over the last deglaciation and back to 78 000 years BP. Methane data are from GISP2 (gray: Blunier and Brook, 2001), GRIP (black: Blunier and Brook, 2001, compiling data from earlier papers; and green: Flückiger et al., 2004), and NorthGRIP (blue: Flückiger et al., 2004 and red: Huber et al., 2006). Ages have been transferred approximately to the NorthGRIP GICC05 age scale using matches between GRIP and GISP2, and between GRIP and NorthGRIP (Wolff et al., 2010). Upper curve: climatic record in Greenland reconstructed from the water isotopes. The numbers indicate Dansgaard/Oeschger events. YD means Younger Dryas. B/A means Bolling/Allerød. Adapted from Wolff EW, Chappellaz J, Blunier T, Rasmussen S, and Svensson A (2010) Millennial-scale variability during the last glacial: The ice core record. *Quaternary Science Reviews* 29 (21–22): 2828–2838.

case of the last deglaciation, each Dansgaard/Oeschger warming which has been investigated both for CH₄ and thermal anomalies of permanent gas isotopes indicates that the start of each phenomenon is in phase.

The modulation of CH₄ average amplitude throughout the last glaciation bears resemblance with the summer insolation curve at northern tropical latitudes (Flückiger et al., 2004). On the other hand, the average inter polar CH₄ gradient calculated over multiple Dansgaard/Oeschger events suggest that some of these events involved an increased CH₄ emission at boreal latitudes (Dällenbach et al., 2000). Few attempts have been made to quantify the CH₄ budget of the last glaciation maximum and the Holocene, and none during Dansgaard/Oeschger events. Two estimates of the global wetland extent and associated CH₄ emissions reach conflicting conclusions: a reduction of their extent by 50% (Chappellaz et al., 1993a) or an increase by 15% (Kaplan, 2002; Kaplan et al., 2006) during the Last Glacial Maximum compared to the preindustrial Holocene. But they reach a better agreement on the CH₄ emissions, with a reduction of 35 and 24%, respectively. Both studies conclude that other mechanisms than wetland extent and emission changes have probably contributed to the methane variability at glacial–interglacial time scale. One of the candidates is an increased oxidative capacity of the atmosphere during glacial times, reducing the CH₄ residence time in the atmosphere. Calculations of the OH concentration, taking into account the reduced sources of CH₄ and other organic compounds emitted by the biosphere, suggest that OH levels were 25% larger during glacial times (Kaplan et al., 2006; Valdes et al., 2005).

As for younger time periods, CH₄ stable isotopic ratio changes are a powerful means to evaluate different scenarios able to explain the CH₄ doubling during the deglaciation or its large changes associated with Dansgaard/Oeschger events. Combining ¹³C/¹²C, D/H and the inter polar gradient information, a plausible scenario is that the boreal wetlands were essentially switched-off during the Last Glacial Maximum and considerably developed during the deglaciation. On the other hand, biomass burning would have remained nearly constant at global scale throughout the glacial–interglacial transition (Fischer et al., 2008). For the few investigated Dansgaard/Oeschger events, methane stable isotopes suggest that boreal wetlands were again the main trigger of the concentration increase (Bock et al., 2010). The hypothesized catastrophic degassing of methane hydrates suggested from marine records seems to be ruled out by the evolution of D/H of atmospheric CH₄ (Bock et al., 2010).

As methane is well recorded both in Greenland and Antarctic ice, shows rapid changes, and is a signal of global significance (i.e., its changes are synchronous at a global scale), it provides an excellent tool to put ice cores from different places on a common relative time scale. Provided that the Δ -age is calculated with sufficient precision, the phase relationship between climate records from different ice cores can then be determined with a precision reaching in the best case about 200 years. This method, pioneered by Blunier et al. (1997) when they showed that the Antarctic Cold Reversal was synchronous with the Bolling/Allerod warm period in Greenland, has been applied to the Dansgaard/Oeschger events of the last glacial (Blunier and Brook, 2001; Blunier et al., 1998; Brook

et al., 2005; EPICA Community Members, 2006). It reveals that each major Dansgaard/Oeschger warming recorded in Greenland is associated with the start of a cooling trend in Antarctica. This important finding for climate dynamics is a clear demonstration of the existence of the bipolar seesaw phenomenon of energy redistribution between the two poles associated with the strengthening or weakening of the North Atlantic thermohaline circulation.

The Last Eight Glacial–Interglacial Cycles

The first ice core giving access to the atmospheric methane variability in the course of several glacial–interglacial cycles was the Vostok one (Chappellaz et al., 1990; Delmotte et al., 2004; Petit et al., 1999). Covering the last 420 000 years, it indicates that CH₄ levels naturally fluctuated in a range between 320 and 800 ppbv, that is, largely below its present-day level of 1800 ppbv. The methane trends are highly correlated with Antarctic temperature changes, with highest mixing ratios during interglacial times and lowest ones during the glacial maxima. All deglaciations depict a similar trend of the CH₄ increase, in two steps: after a slow increase over a few millennia, a rapid increase over one to three centuries takes place. On the other hand, methane closely follows Antarctic temperature during the glacial inception and does not show a lag as observed for CO₂. The periodicities characterizing insolation changes at the Earth's surface are imprinted in the CH₄ record: the largest variance is associated with the 100 000-year periodicity but with significant parts of the variance being associated with the 40 000-year and 23 000/19 000-year periodicities. The calculated direct radiative impact of the glacial–interglacial CH₄ changes implies a mean global temperature increase (including chemical feedbacks on methane, i.e., tropospheric ozone and stratospheric water vapor, but excluding other climate feedbacks) of ~ 0.10 °C.

In addition to the low-frequency variability, the Vostok methane record also depicts millennial-scale changes during previous glaciations, comparable with variations observed during the Dansgaard/Oeschger events of the last glaciation (Delmotte et al., 2004). This suggested that abrupt climatic changes during glaciations are a robust feature of the late Quaternary.

The Vostok CH₄ record has been confirmed over the last three climatic cycles with the Japanese Dome Fuji ice core. In 2005, the newly available EDC ice core has allowed one to extend the Vostok record back to 800 000 years, that is, four additional glacial–interglacial cycles (Louergue et al., 2008; Spahni et al., 2005). Before 420 000 years BP, the EDC climate record is characterized by milder and longer interglacials compared with the more recent ones. The atmospheric CH₄ mixing ratio indicates a similar trend, with relatively lower (~ 600 ppbv) and longer maxima during the two older interglacials (Figure 4). The extended record shows that the precessional component of orbitally-induced CH₄ variability bore a stronger contribution during the four more recent climatic cycles. It also reveals that millennial-scale CH₄ variability is ubiquitous of the last eight glaciations. Although a quantitative and mechanistic link between climate and methane changes on such time scale is not yet available, the observation of a similar relation of proportionality throughout eight climatic cycles between Antarctic

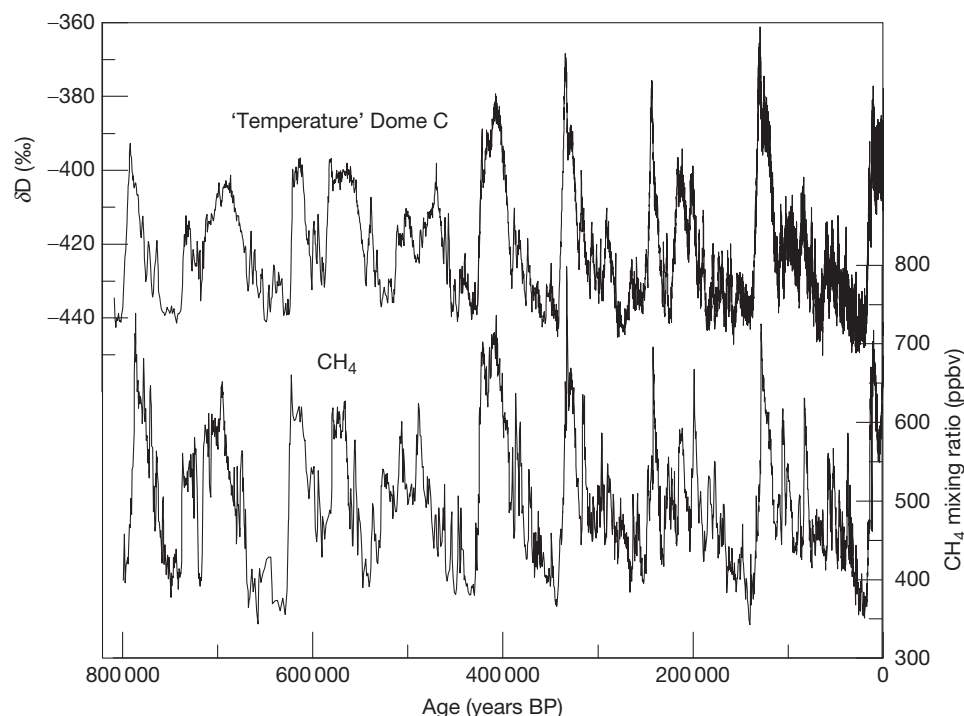


Figure 4 Evolution of the CH₄ mixing ratio over the last 800 000 years, compared with the D/H Antarctic temperature proxy. Data from Loulergue et al. (2008) and Jouzel et al. (2007).

temperature and atmospheric CH₄ levels suggests that the mechanisms involved are persistent and robust.

Analytical Techniques

To measure CH₄ in ice cores, the extraction of trapped gases can be performed using a dry extraction method, provided that there is no friction between metal parts within the extracting device. These frictions usually degas significant and variable amount of methane. Only the 'cheese grater' technique has proved to deliver CH₄ mixing ratios with negligible contamination from the extraction chamber (Etheridge et al., 1998).

Instead, the most popular technique consists in melting the ice sample (typically 40–50 g) under vacuum and subsequently refreezing the melt water from the bottom part of the container, in order to expel the dissolved gases. Containers are usually made of glass, but stainless steel is also used (Sowers et al., 1997).

A sublimation technique has also been developed and applied to CH₄ in ice cores (e.g., Güllük et al., 1998). It gives good agreement with the melting–refreezing method but its complexity does not justify a routine application for methane alone.

Then the gas is expanded in a sample loop and injected in the analytical instrument, or eventually purified on cold traps where only CH₄ is retained.

The detection and quantification of CH₄ in the gas stream from the sample loop or cold traps is obtained by gas chromatography, using a flame-ionization detector. Calibration of the instrument is performed for each extraction against standard gases of known CH₄ mixing ratio.

Recently, a new method has been developed where CH₄ is measured by gas chromatography downstream from a

continuous-flow analysis system initially built for chemical analyses of ice cores (Schüpbach et al., 2009). It provides a nearly continuous methane record along ice cores, with a much larger sample throughput than discrete sample techniques, but with a lower accuracy. The best results are now expected to come soon from the coupling between continuous-flow extraction techniques and new optical detectors to accurately measure the CH₄ mixing ratio in air.

Stable isotopic ratios of CH₄ are measured using gas chromatography, followed by on-line combustion of CH₄ to CO₂ under helium flow, and introduction into a magnetic-sector isotopic ratio mass spectrometer. Due to the small signal to decipher, it requires larger ice sample size: typically a few hundred grams for each measurement.

See also: **Carbonate Stable Isotopes:** Speleothems; **Glacial Climates:** Biosphere Feedbacks; **Glaciation, Causes:** Astronomical Theory of Paleoclimates; **Ice Core Methods:** Chronologies; CO₂ Studies; Overview; **Stable Isotopes:** **Ice Core Records:** Correlations Between Greenland and Antarctica; Greenland Stable Isotopes; Thermal Diffusion Paleotemperature Records; **Ice Cores:** History of Carbon Monoxide and Ultra-Trace Gases from Ice Cores; **Paleoceanography, Physical and Chemical Proxies:** Carbon Cycle Proxies ($\delta^{11}\text{B}$, $\delta^{13}\text{C}_{\text{calcite}}$, $\delta^{13}\text{C}_{\text{organic}}$, Shell Weights, B/Ca, U/Ca, Zn/Ca, Ba/Ca); **Paleoceanography, Records:** Late Pleistocene North Atlantic; **Paleoclimate:** The Younger Dryas Climate Event; **Paleoclimate Reconstruction:** Sub-Milankovitch (DO/Heinrich) Events; The Last Millennium; Younger Dryas Oscillation, Global Evidence; **Permafrost and Periglacial Features:** Active Layer Processes.

References

- Blunier T and Brook EJ (2001) Timing of millennial-scale climate change in Antarctica and Greenland during the last glacial period. *Science* 291: 109–112.
- Blunier T, Chappellaz J, Schwander J, Stauffer B, and Raynaud D (1995) Variations in atmospheric methane concentration during the Holocene epoch. *Nature* 374: 46–49.
- Blunier T, Chappellaz J, Schwander J, et al. (1998) Asynchrony of Antarctic and Greenland climate during the last glacial period. *Nature* 394: 739–743.
- Blunier T, Schwander J, Stocker T, et al. (1997) Timing of the Antarctic Cold Reversal and the atmospheric CO₂ increase with respect to the Younger Dryas event. *Geophysical Research Letters* 24: 2683–2686.
- Bock M, Schmitt J, Möller L, Spahni R, Blunier T, and Fischer H (2010) Hydrogen isotopes preclude marine hydrate CH₄ emissions at the onset of Dansgaard–Oeschger events. *Science* 328: 1686–1689.
- Brook EJ, Harder S, Severinghaus JP, and Bender M (1999) Atmospheric methane and millennial-scale climate change. In: Clark PU, Webb RS, and Keigwin LD (eds.) *Mechanisms of Global Climate Change at Millennial Time Scales. Geophysical Monograph Series*, vol. 112, pp. 165–176. San Francisco, CA: American Geophysical Union.
- Brook EJ, Sowers T, and Orchard J (1996) Rapid variations in atmospheric methane concentration during the past 110,000 years. *Science* 273: 1087–1093.
- Brook E, White JWC, Schilla A, et al. (2005) Timing of millennial-scale climate change at Siple Dome, West Antarctica, during the last glacial period. *Quaternary Science Reviews* 24: 1333–1343.
- Campan RK, Sowers T, and Alley RB (2003) Evidence of microbial consortia metabolizing within a low-latitude mountain glacier. *Geology* 31: 231–234.
- Chappellaz J, Barnola J-M, Raynaud D, Korotkevich YS, and Lorius C (1990) Ice-core record of atmospheric methane over the past 160,000 years. *Nature* 345: 127–131.
- Chappellaz J, Blunier T, Barnola J-M, Raynaud D, Schwander J, and Stauffer B (1993a) Synchronous changes in atmospheric CH₄ and Greenland climate between 40 and 8 kyr BP. *Nature* 366: 443–445.
- Chappellaz J, Blunier T, Kints S, et al. (1997) Changes in the atmospheric CH₄ gradient between Greenland and Antarctica during the Holocene. *Journal of Geophysical Research* 102: 15987–15997.
- Chappellaz JA, Fung IY, and Thompson AM (1993b) The atmospheric CH₄ increase since the Last Glacial Maximum. 1. Source estimates. *Tellus* 45B: 228–241.
- Dällenbach A, Blunier T, Flückiger J, Stauffer B, Chappellaz J, and Raynaud D (2000) Changes in the atmospheric CH₄ gradient between Greenland and Antarctica during the Last Glacial and the transition to the Holocene. *Geophysical Research Letters* 27: 1005–1008.
- Delmotte M, Chappellaz J, Brook E, et al. (2004) Atmospheric methane during the last four glacial–interglacial cycles: Rapid changes and their link with Antarctic temperature. *Journal of Geophysical Research* 109: D12104. <http://dx.doi.org/10.1029/2003JD004417>.
- Denman KL, Brasseur G, Chidthaisong A, et al. (2007) Couplings between changes in the climate system and biogeochemistry. In: Solomon S, Qin D, Manning M, Chen Z, Marquis M, Averyt KB, Tignor M, and Miller HL (eds.) *Climate Change 2007: The Physical Science Basis*. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change, Cambridge, UK: Cambridge University Press.
- EPICA Community Members (2006) One-to-one coupling of glacial climate variability in Greenland and Antarctica. *Nature* 444: 195–198.
- Etheridge DM, Steele LP, Francey RJ, and Lagenfelds RJ (1998) Atmospheric methane between 1000 A.D. and present: Evidence of anthropogenic emissions and climatic variability. *Journal of Geophysical Research* 103: 15979–15993.
- Ferretti DF, Miller JB, White JWC, et al. (2005) Unexpected changes to the global methane budget over the Past 2000 Years. *Science* 309: 1714–1717.
- Fischer H, Behrens M, Bock M, et al. (2008) Changing boreal methane sources and constant biomass burning during the last termination. *Nature* 452: 864–867.
- Flückiger J, Blunier T, Stauffer B, et al. (2004) N₂O and CH₄ variations during the last glacial epoch: Insight into global processes. *Global Biogeochemical Cycles* 18: GB1020. <http://dx.doi.org/10.1029/2003GB002122>.
- Gülliük T, Slemr F, and Stauffer B (1998) Simultaneous measurements of CO₂, CH₄ and N₂O in air extracted by sublimation from Antarctica ice cores: Confirmation of the data obtained using other extraction techniques. *Journal of Geophysical Research* 103: 15971–15978.
- Houweling S, van der Werf GR, Goldewijk KK, Rockmann T, and Aben I (2008) Early anthropogenic CH₄ emissions and the variation of $\delta^{13}\text{C}_{\text{CH}_4}$ over the last millennium. *Global Biogeochemical Cycles* 22: GB1002. <http://dx.doi.org/10.1029/2007GB002961>.
- Huber C, Leuenberger M, Spahni R, et al. (2006) Isotope calibrated Greenland temperature record over Marine Isotope Stage 3 and its relation to CH₄. *Earth and Planetary Science Letters* 243: 504–519.
- Jouzel J, Masson-Delmotte V, Cattani O, et al. (2007) Orbital and Millennial Antarctic Climate Variability over the Past 800,000 Years. *Science* 317: 793–796.
- Kaplan JO (2002) Wetlands at the Last Glacial Maximum: Distribution and methane emissions. *Geophysical Research Letters* 29(6): 1079. <http://dx.doi.org/10.1029/2001GL013366>.
- Kaplan JO, Folberth G, and Hauglustaine DA (2006) Role of methane and biogenic volatile organic compound sources in late glacial and Holocene fluctuations of atmospheric methane concentrations. *Global Biogeochemical Cycles* 20: GB2016. <http://dx.doi.org/10.1029/2005GB002590>.
- Keppler F, Hamilton JTG, Braß M, and Röckmann T (2006) Methane emissions from terrestrial plants under aerobic conditions. *Nature* 439: 187–191.
- Landais A, Barnola J-M, Masson-Delmotte V, et al. (2004) A continuous record of temperature evolution over a sequence of Dansgaard–Oeschger events during Marine Isotopic Stage 4 (76 to 62 kyr BP). *Geophysical Research Letters* 31: L22211. <http://dx.doi.org/10.1029/2004GL021193>.
- Loulergue L, Parrenin F, Blunier T, et al. (2007) New constraints on the gas age–ice age difference along the EPICA ice cores, 0–50 kyr. *Climate of the Past* 3: 527–540.
- Loulergue L, Schilt A, Spahni R, et al. (2008) Orbital and millennial-scale features of atmospheric CH₄ over the last 800,000 years. *Nature* 453: 383–386.
- MacFarling Meure C, Etheridge D, Trudinger C, et al. (2006) Law Dome CO₂, CH₄ and N₂O records extended to 2000 years BP. *Geophysical Research Letters* 33: L14810. <http://dx.doi.org/10.1029/2006GL026152>.
- Mischler JA, Sowers TA, Alley RB, et al. (2009) Carbon and hydrogen isotopic composition of methane over the last 1000 years. *Global Biogeochemical Cycles* 23: GB4024. <http://dx.doi.org/10.1029/2009GB003460>.
- Petit JR, Jouzel J, Raynaud D, et al. (1999) Climate and atmospheric history of the past 420,000 years from the Vostok Ice Core, Antarctica. *Nature* 399: 429–436.
- Ruddiman WF and Thomson JS (2001) The case for human causes of increased atmospheric CH₄ over the last 5000 years. *Quaternary Science Reviews* 20: 1769–1777.
- Schilt A, Baumgartner M, Schwander J, et al. (2010) Atmospheric nitrous oxide during the last 140,000 years. *Earth and Planetary Science Letters* 300: 33–43.
- Schüpbach S, Federer U, Kaufmann P, et al. (2009) A new method for high-resolution methane measurements on polar ice cores using continuous flow analysis. *Environmental Science and Technology* 43: 5371–5376. <http://dx.doi.org/10.1021/es9003137>.
- Schwander J (1996) Gas diffusion in firn. In: Wolff EW and Bales RC (eds.) *Chemical Exchange Between the Atmosphere and Polar Snow*, pp. 527–539. Berlin: Springer Verlag.
- Severinghaus JP and Brook EJ (1999) Abrupt climate change at the end of the Last Glacial Period inferred from trapped air in polar ice. *Science* 286: 930–937.
- Sowers T (2010) Atmospheric methane isotope records covering the Holocene period. *Quaternary Science Reviews* 29: 213–221.
- Sowers T, Bernard S, Aballain O, Chappellaz J, Barnola J-M, and Marik T (2005) Records of the $\delta^{13}\text{C}$ of atmospheric CH₄ over the last two centuries as recorded in Antarctic snow and ice. *Global Biogeochemical Cycles* 19: GB2002. <http://dx.doi.org/10.1029/2004GB002408>.
- Sowers T, Brook E, Blunier T, et al. (1997) An inter-laboratory comparison of techniques for extracting and analysing trapped gases in ice cores. *Journal of Geophysical Research* 102: 26527–26538.
- Spahni R, Chappellaz J, Stocker TF, et al. (2005) Atmospheric methane and nitrous oxide of the late Pleistocene from Antarctic ice cores. *Science* 310: 1317–1321.
- Spahni R, Schwander J, Flückiger J, Stauffer B, Chappellaz J, and Raynaud D (2003) The attenuation of fast atmospheric CH₄ variations recorded in polar ice cores. *Geophysical Research Letters* 30(11): 1571. <http://dx.doi.org/10.1029/2003GL017093>.
- Valdes PJ, Beerling DJ, and Johnson CE (2005) The ice age methane budget. *Geophysical Research Letters* 32: L02704. <http://dx.doi.org/10.1029/2004GL021004>.
- Wang Z, Chappellaz J, Park K, and Mak JE (2010) Large variations in Southern hemisphere biomass burning during the last 650 years. *Science* 330: 1663–1666.
- Wolff EW, Chappellaz J, Blunier T, Rasmussen SO, and Svensson A (2010) Millennial-scale variability during the last glacial: The ice core record. *Quaternary Science Reviews* 29: 2828–2838.

Microparticle and Trace Element Studies

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Introduction

Mountain glaciers, small ice caps, and the two large ice sheets in Greenland and Antarctica arguably contain the highest temporal resolution and most direct paleoenvironmental archives available for the Quaternary. Impurities scavenged from the atmosphere by falling snow and deposited directly by gravitational settling comprise a highly detailed record of past changes in climate and anthropogenic impacts – including atmospheric composition (e.g., Mayewski et al. (1993) and BOUTRON et al. (1994)), oceanic and atmospheric circulation (e.g., Mayewski et al. (1994) and Delmonte et al. (2004)), meteorology (e.g., Alley et al. (1993)), explosive volcanism (e.g., Robock and Free (1996)), marine and terrestrial biosphere emissions (e.g., Legrand et al. (1991)), biomass burning (e.g., Holdsworth et al. (1996)), and industrial pollution (e.g., Davidson et al. (1991), BOUTRON et al. (1994), and Wolff and Suttie (1994)). The challenge is for Earth scientists to strategically sample and analyze the unique glaciochemical archive as well as comprehensively interpret the history contained therein.

Ice cores collected from glaciers, ice caps, and ice sheets are used to investigate the glaciochemical record. Cores from high-snowfall regions reveal annual layers that are easily distinguished by their chemical characteristics (e.g., McConnell et al. (2002)). Sampling of cores from these regions is often designed to yield glaciochemical records with 20 or more samples per year. These high-resolution records span recent decades to a few millennia and have been developed from cores collected in high-snowfall regions in Greenland, Antarctica, and a number of mountain glaciers and ice caps (e.g., Kreutz et al. (2000) and Correia et al. (2003)). The longest glaciochemical records – possibly spanning up to a million years – are available only from low-snowfall, high-latitude regions of Antarctica. Because of the low snowfall at these sites, temporal resolution is limited and annual layering is not preserved in the longest records.

Extent of the Glaciochemical Archive

Glaciers and small ice caps outside of Antarctica have a combined area of about 0.54 million km² (Intergovernmental Panel on Climate Change, 2001). While located at most latitudes, they are found only at very high elevations in the low and mid-latitudes. Glaciochemical records developed from cores drilled to bedrock in these regions typically span hundreds to thousands of years; some records are as long as 15 ka (e.g., Thompson et al. (1995); see South America; Chinese, Tibetan Mountains; Africa.

The Greenland ice sheet covers 1.71 million km² and lies between 60° and 82° N latitude (Intergovernmental Panel on

Climate Change, 2001). Because snowfall rates are relatively high over much of Greenland, ice cores drilled to bedrock typically span tens of thousands to hundreds of thousands of years. To date, five ice cores have been collected to bedrock in Greenland. The longest of these Greenland records are from the ~3200-m-long European Greenland Ice Sheet Project (GRIP, 72.6° N, 37.6° W) and US Greenland Ice Sheet Project 2 (GISP2; 72.6° N, 38.5° W) cores which were drilled near the summit of the ice sheet in the early 1990s. These high-resolution records span more than 250 ka, with annual layering identified for the past 110 ka.

The Antarctic ice sheet has an area of 12.37 million km² (Intergovernmental Panel on Climate Change, 2001). With an area half the size of North America, the Antarctic ice sheet and smaller glaciers and ice caps in Antarctica extend from 67° S at the tip of the Antarctic Peninsula to 90° S at the South Pole. Several cores reaching bedrock have been collected in Antarctica including the Vostok (78.5° S, 106.8° E), EPICA Dome C (81.7° S, 148.8° W), and Dome Fuji (77.3° S, 39.7° W) cores in East Antarctica and the Siple Dome core (78.5° S, 106.8° W) in West Antarctica. The longest record is from Dome Fuji, which, although not yet confirmed, may span 1000 ka.

Aerosols and Microparticles

Impurities contained in glacier ice include gases trapped in air bubbles, soluble gases, and acids adsorbed or dissolved in the ice, and soluble and insoluble impurities from aerosols – particulate matter suspended in the air. The focus here is on impurities from atmospheric aerosols which form soluble and insoluble microparticles when found in glacier ice. Natural sources of aerosols include sea spray, volcanic emissions, forest and grassland fires, mineral dust, and biogenic emissions. Aerosols also result directly from human activities including burning of fossil fuels and biomass burning to clear land, and indirectly from land-use changes which expose soil to wind erosion (e.g., Tegen et al. (1996) and Prospero et al. (2002)).

Aerosols are an important part of the Earth system, influencing critical environmental processes such as biogeochemical cycling and climate forcing. For example, soluble iron (Fe) in dust is an important fertilizer of the remote oceans with increased dust flux linked to new phytoplankton production and carbon sequestration in Fe-limited regions such as the Southern Ocean and northern Pacific (Martin, 1990). Since aerosols also affect the Earth's radiation balance (e.g., Intergovernmental Panel on Climate Change (2001), Menon et al. (2002), and Tegen et al. (1996)), investigating how aerosol characteristics and concentrations vary in the atmosphere today and how they might have varied in the past is crucial to

understanding the Earth system and how it will evolve in the future. As with modern studies of airborne aerosols (e.g., Rahn and Lowenthal (1985)), rock type and chemical composition of insoluble microparticles found in different layers of ice within a glacier provide important information on the origin of these microparticles.

Microparticles in glacier ice are both soluble and insoluble, with most being soluble. Soluble microparticles include sea salts, sulfate aerosols from volcanic and marine biogenic emissions, and some terrestrial dusts such as carbonates from evaporite deposits. Insoluble microparticles are composed of crustal materials including aluminosilicate rock dust, with lesser amounts from volcanic shards and extraterrestrial material. The ratio of soluble to insoluble fractions of elements measured in glacier ice varies with the location and depth of the ice cores. For example, calcium (Ca) found in central Greenland ice comes primarily from continental dust, with the soluble form associated with sea salt, calcium carbonate, and dolomite dust. The insoluble fraction is primarily from apatite and plagioclase minerals. Laj et al. (1997) found that the soluble fraction of Ca in the GRIP core varied from 30% to almost 100% during the past 110 ka, with a high fraction of soluble Ca during cold periods when total Ca concentrations were high.

Sources of Microparticles in Glacier Ice

Most elements in microparticles archived in glacier ice have multiple sources. For example, magnesium (Mg) is found in large quantities in sea salt as well as evaporite and aluminosilicate dust. It also may occur in lesser amounts in volcanic and forest fire ash, as well as pollution aerosols. Similarly, lead (Pb) is found in rock dust but is generally heavily enriched even in preindustrial aerosols, probably as a result of adsorption of Pb from quiescent volcanic emissions (Matsumoto and Hinkley, 2001) and early metal smelting (Boutron et al., 1994). Since industrialization, Pb in the atmosphere and glacier ice has been dominated by industrial pollution. Fully interpreting an elemental record of Mg or Pb requires simultaneous measurement of a range of other elements or chemical compounds to elucidate all possible sources.

Ice-Core Research Results

The number of glaciochemical studies of microparticles in ice cores that have been conducted since the 1980s is very large, particularly those focused on soluble microparticles (e.g., Mayewski et al. 1993, 1994 and De Angelis et al. (1997)). A complete review is far beyond the scope of this contribution. Instead, the following addresses selected ice-core research results organized into two overlapping themes: (1) dust provenance in Greenland and Antarctica and (2) heavy metal pollution in central Greenland. These themes, and the studies selected to illustrate them, reflect the potential of ice-core research to address critical issues in climate change and anthropogenic impacts. Although not exhaustive, the examples reveal new methods' development, creative approaches to interpreting data, and, perhaps most significantly, the complexity and

interrelationships that must be considered in interpreting results of ice-core studies.

Fingerprinting Microparticle Provenance in Central Greenland and East Antarctica

The mineralogy of rocks as well as elemental and isotopic ratios in dust derived from these rocks varies naturally from region to region around the world. Elemental and isotopic ratios in industrial emissions and volcanic tephra vary as well. Together, these often-subtle variations provide a means to fingerprint the source region (or provenance) of dust, volcanic tephra, and pollution aerosols. Knowledge of provenance leads to a better understanding of atmospheric circulation and how it may have changed in the past.

Satellite-based studies indicate that topographic lows with large accumulations of fine particles easily eroded by wind are the primary sources of dust aerosols today. Many of these alluvial sediment deposits were formed during the late Pleistocene or Holocene, and most are found in low- to mid-latitude regions such as northern Africa, central Asia, and Australia, with higher-latitude sources in southern South America and southern Africa (Prospero et al., 2002). Dust aerosols found in most polar ice cores have been transported over very large distances since the cores are usually located in the interior of the ice sheets, well away from any local dust sources. Conversely, dust aerosols found in Alpine cores are often a combination of proximal and distal sources. Distinguishing the source can be difficult, with one method – use of isotopic ratios – applied in some studies.

Isotopic ratios have been used extensively to determine the provenance of dust in deep ice cores and provide a demonstration of the use of trace element analyses of microparticles for determining provenance. Building on earlier studies in Antarctica, Biscaye et al. (1997) analyzed Pb, strontium (Sr), and neodymium (Nd) isotopes in microparticles filtered from GISP2 ice-core samples to determine the source region for dust deposited during the last glacial period. The measured ratios were compared to similar measurements of particles less than 5 μm in diameter from possible source areas (PSAs) in North America, Asia, and Africa. Based on the isotopic ratios, Biscaye and his colleagues were able to exclude mid-continental North America and Africa as source regions, and concluded that eastern Asia was the likely source of terrestrial dust deposited in central Greenland during the last glacial period.

Basile et al. (1997) used similar measurements of Sr and Nd isotopes in the Vostok and old Dome C ice cores to evaluate the provenance of dust during the last three glacial periods in East Antarctica. Comparisons to similar measurements from PSA in Antarctica, New Zealand, Australia, South America, and southern Africa indicated that South America – specifically, Patagonian loess – was the source. Delmonte et al. (2004) expanded this research to include the new EPICA Dome C core and far more samples of dust from possible source areas of the Southern Hemisphere. In agreement with the earlier study, Delmonte and her colleagues concluded that a broader region in South America – including Patagonia but also the Pampas and possibly the continental shelf of Argentina exposed by lower sea level – was the provenance of the dust at the two

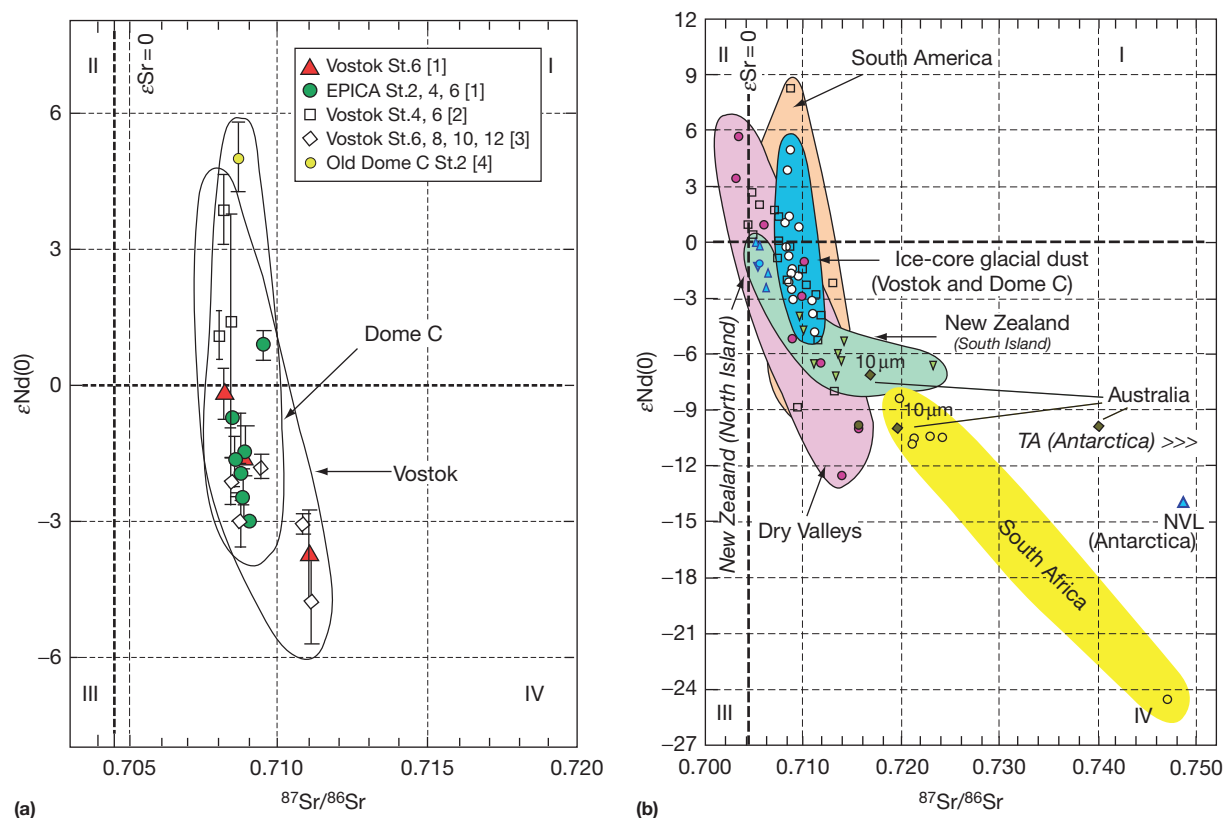


Figure 1 $^{87}\text{Sr}/^{86}\text{Sr}$ versus $\epsilon_{\text{Nd}}(0)$ fields for (a) dust measured in the Dome C and Vostok ice cores; and (b) as in (a) but with similar measurements of samples from possible source areas throughout the Southern Hemisphere. Note the difference in the vertical axes. Samples from possible sources areas were limited to $<5 \mu\text{m}$ since this size range is subject to long-range transport in the atmosphere. Reprinted from Delmonte B, Basile-Doelsch I, Petit J-R, et al. (2004) Comparing the Epica and Vostok dust records during the last 220 000 years: Stratigraphical correlation and provenance in glacial periods. *Earth-Science Reviews* 66: 63–87, Copyright 2004, with permission from Elsevier.

ice-core sites. Similarities to the Sr and Nd isotopic signatures for Antarctica (Dry Valleys) and New Zealand, however, indicate that these regions also may be sources for some of the dust found in East Antarctica (Figure 1).

Evaluation of Heavy Metal Pollution in Central Greenland

Microparticles deposited in glacier ice reflect aerosol composition of the local atmosphere as well as aerosol composition and atmospheric chemistry along the trajectory of air parcels arriving at an ice-core site, particularly moisture-bearing air parcels (Davidson et al., 1991). Many ice-core chemistry studies are aimed at reconstructing how atmospheric composition has changed in the past as a result of climate change, volcanic eruptions, or industrial pollution. For example, Mayewski et al. (1993) used measurements of soluble major ions in the GISP2 ice core in Greenland to evaluate changes in atmospheric composition around the Younger Dryas, a period of rapid climate change which occurred during the transition from the last glacial period to the Holocene some 12–13 ka BP.

Because long-range transport results in widespread contamination, many studies have used measurements of heavy metals in well-dated glacier ice to document changes in industrial pollution of the remote Arctic from recent decades to several centuries ago (Boutron et al., 1994; Davidson et al.,

1991). In most of these studies, total concentrations were measured, including both soluble and insoluble fractions. Lead was the focus of many of the early studies, in part because of the very dramatic increases in emissions associated with leaded gasoline use in automobiles, coordinated government efforts to eliminate Pb emissions (e.g., the U.S Clean Air Act implemented in the early 1970s), and well-defined concerns for human and ecosystem health. Ice-core studies document a tenfold increase of Pb deposition in parts of Antarctica and up to a 200-fold increase in central Greenland (Boutron et al., 1994; Wolff and Suttie, 1994).

Heavy metal concentrations in Antarctic snow and ice are exceedingly low, often in the fg g^{-1} to pg g^{-1} levels ($1 \text{ fg} = 10^{-15} \text{ g}$; $1 \text{ pg} = 10^{-12} \text{ g}$; $1 \text{ ng} = 10^{-9} \text{ g}$; $1 \mu\text{g} = 10^{-6} \text{ g}$). With recent advances in analytical capabilities, however, the number and scope of ice-core studies of heavy metal pollution have increased dramatically. For example, no recent trends were observed in some metal concentrations – including manganese (Mn), cobalt (Co), and vanadium (V) – in ice-core samples from Coats Land, Antarctica. Recent increases were observed in the same samples, however, for chromium (Cr), copper (Cu), zinc (Zn), Pb, and uranium (U). Planchon et al. (2002) attributed these increases to nonferrous mining and smelting in South America, southern Africa, and Australia.

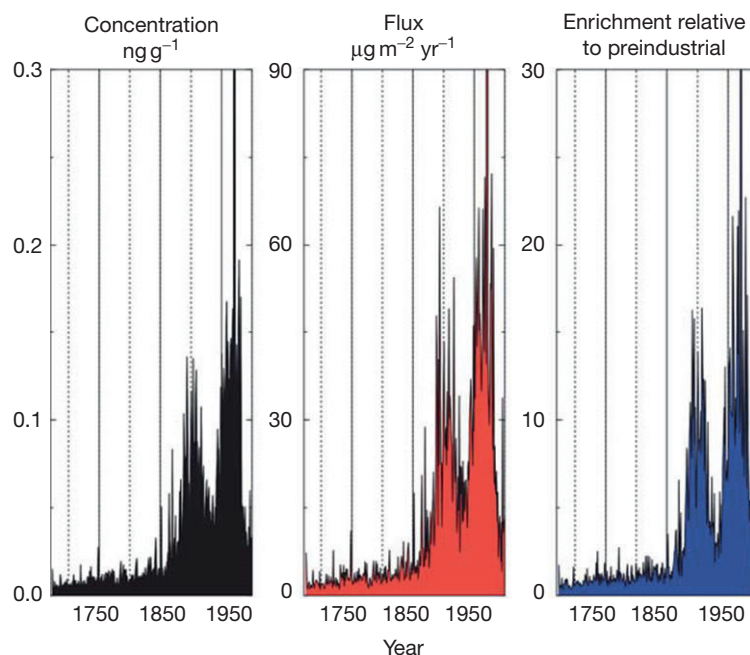


Figure 2 A 330-yr continuous ice-core record of Pb concentration, flux, and relative enrichment at the D5 ice-core site (68.5° N, 42.9° W) in west central Greenland. Shown are annual average values determined from high-resolution measurements (>40 samples yr^{-1}). Enrichment was calculated using core-registered measurements of the crustal dust tracer Al.

Related studies have been conducted in Alpine regions (Kreutz et al., 2000; Correia et al., 2003) and Greenland as well. For example, Barbante et al. (2001) investigated pollution in Greenland from platinum group metals (PGMs) that are emitted from aging catalytic converters on automobiles. They reported concentrations 40 to 120 times higher in 1990–95 snow than in ice deposited ~ 7 ka, suggesting that PGM pollution is indeed widespread. Also, a high-resolution ice-core record of Pb concentrations, fluxes, and crustal enrichments (relative to aluminum, a crustal dust tracer) demonstrates how Pb pollution has varied in west central Greenland during the past 330 yr (Figure 2).

Specifically, Pb concentration and flux were low but slowly rising prior to widespread industrialization in the mid-1800s. Fluxes increased from $\sim 1.5 \mu\text{g m}^{-2} \text{yr}^{-1}$ in 1670 to $\sim 5 \mu\text{g m}^{-2} \text{yr}^{-1}$ in 1860, and enrichment rose slowly as well. Increases in Pb deposition lasting from a few months to a few years resulted from fallout from explosive volcanism. Beginning in ~ 1865 , Pb flux increased dramatically and reached $\sim 37 \mu\text{g m}^{-2} \text{yr}^{-1}$ by 1895, more than a sevenfold increase from pre-1865 levels. Following the sharp, well-defined drop during the Great Depression when fluxes declined to $\sim 17 \mu\text{g m}^{-2} \text{yr}^{-1}$, fluxes increased rapidly from 1939 to a peak in 1970 of more than $120 \mu\text{g m}^{-2} \text{yr}^{-1}$, a 24-fold increase from pre-1870 levels. By 1985, Pb deposition had declined to $\sim 20 \mu\text{g m}^{-2} \text{yr}^{-1}$ in response to government-mandated emission reductions. Mean annual fluxes from 1985 to 2002, however, were still 200% of those observed prior to 1865 and 700% of those observed in 1670. Relative enrichment of Pb with respect to the crustal dust tracer aluminum (Al) provides a similar story since Al concentrations have not changed

significantly during this time period. While the drop in Pb pollution from ~ 1970 to ~ 2000 closely matches similar declines in estimated US industrial emissions (McConnell et al., 2002), industrial emissions were orders of magnitude greater during the era of leaded gasoline use in automobiles than before. As Figure 2 clearly shows, however, Pb pollution in west central Greenland during the leaded gasoline era was only about twice that observed at the beginning of the twentieth century, some three decades before the widespread use of leaded gasoline. It is clear that much of the Pb pollution found in Greenland ice cores comes from other sources such as smelting and coal combustion, although the exact sources remain unknown.

Conclusions and Future Directions

Glaciochemical records developed from ice cores collected from mountain glaciers and the polar ice sheets are a cornerstone of environmental and climate change research. Chemical, elemental, and isotopic measurements of soluble and insoluble microparticles in ice cores can be used to develop highly detailed records of change in atmospheric composition, atmospheric circulation, and source areas during the past hundreds of thousands of years. Advances in analytical methods mean that even more detailed environmental records will be obtained from ice cores in the future, leading to a better understanding of the Earth system and the impacts of human activities. Future emphasis on ice-core arrays to better define spatial distribution patterns and application of high-resolution analytical methods will increase the validity of ice-core

research results and support more sophisticated interpretation of atmospheric and ocean circulation patterns.

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See also: [Ice Core Methods: Overview](#). [Ice Core Records: Antarctic Stable Isotopes; Correlations Between Greenland and Antarctica; Greenland Stable Isotopes](#). [Ice Cores: Dynamics of the East Antarctic Ice Sheet; Dynamics of the Greenland Ice Sheet; Dynamics of the West Antarctic Ice Sheet; History of Research, Greenland and Antarctica](#).

References

- Alley RB, et al. (1993) Abrupt increase in Greenland snow accumulation at the end of the Younger Dryas event. *Nature* 362: 527–529.
- Barbante C, et al. (2001) Greenland snow evidence of large scale contamination for platinum, palladium, and rhodium. *Environmental Science and Technology* 35: 835–839.
- Basile I, Grousset FG, Revel M, Petit J-R, Biscaye PE, and Barkov NI (1997) Patagonian origin of glacial dust deposited in East Antarctica (Vostok and Dome C) during glacial stages 2, 3, and 6. *Earth and Planetary Science Letters* 146: 573–589.
- Biscaye PE, Grousset FE, Revel M, et al. (1997) Asian provenance of glacial dust (stage 2) in the Greenland Ice Sheet Project 2 ice core, Summit, Greenland. *Journal of Geophysical Research* 102(C12): 26765–26781.
- Boutron CF, Candelone J-P, and Hong S (1994) Past and recent changes in large-scale tropospheric cycles of lead and other heavy metals as documented in Antarctic and Greenland snow and ice: A review. *Geochimica et Cosmochimica Acta* 58(15): 3217–3225.
- Correia A, Freydier R, Delmas RJ, et al. (2003) Trace elements in South America aerosol during the 20th century inferred from a Nevado Illimani ice core, Eastern Bolivian Andes (6350 m asl). *Atmospheric Chemistry and Physics* 3: 1337–1352.
- Davidson CI, Jaffrezo JL, and Mayewski PA (1991) Arctic air pollution as reflected in snowpits and ice cores. In: Sturges WT (ed.) *Pollution of the Arctic Atmosphere*, pp. 43–95. Elsevier.
- De Angelis M, Steffensen JP, Legrand M, Clausen H, and Hammer C (1997) Primary aerosol (sea salt and soil dust) deposited in Greenland ice during the last climatic cycle: Comparison with east Antarctic records. *Journal of Geophysical Research* 102(C12): 26681–26698.
- Delmonte B, Basile-Doelsch I, Petit J-R, et al. (2004) Comparing the Epica and Vostok dust records during the last 220,000 years: Stratigraphical correlation and provenance in glacial periods. *Earth-Science Reviews* 66: 63–87.
- Ferretti DF, Miller JB, White JWC, et al. (2005) Unexpected changes to the global methane budget over the past 2000 years. *Science* 309: 1714–1717.
- Holdsworth G, et al. (1996) Historical biomass burning: Late 19th century pioneer agriculture recolonization in northern hemisphere ice core data and its atmospheric interpretation. *Journal of Geophysical Research* 101(D18): 23371–23384.
- Intergovernmental Panel on Climate Change (2001) *Climate Change 2001: The Scientific Bases*. Cambridge University Press.
- Kreutz KJ and Sholkovitz ER (2000) Major element, rare earth element, and sulfur isotopic composition of a high-elevation firn core: Sources and transport of mineral dust in central Asia. *Geochemistry, Geophysics, Geosystems* 1: <http://dx.doi.org/10.1029/2000GC00082>.
- Laj P, Ghermandi G, Cecchi R, et al. (1997) Distribution of Ca, Fe, K, and S between soluble and insoluble material in the Greenland Ice Core Project ice core. *Journal of Geophysical Research* 102(C12): 26615–26623.
- Legrand M, Feniet-Saigne C, Saltzman ES, Germain C, Barkov NI, and Petrov VN (1991) Ice-core record of oceanic emissions of dimethyl sulphide during the last climate cycle. *Nature* 350: 144–146.
- Martin JH (1990) Glacial-interglacial CO₂ change: The iron hypothesis. *Paleoceanography* 5: 1–13.
- Matsumoto A and Hinkley TK (2000) Trace metal suites in Antarctic pre-industrial ice are consistent with emissions from quiescent degassing of volcanoes worldwide. *Earth and Planetary Science Letters* 186: 33–43.
- Mayewski PA, et al. (1993) The atmosphere during the Younger Dryas. *Science* 261: 195–197.
- Mayewski PA, et al. (1994) Changes in atmospheric circulation and ocean ice cover over the North Atlantic during the last 41,000 years. *Science* 263: 1747–1751.
- McConnell JR, Lamorey GW, and Hutterli MA (2002) A 250-year high-resolution record of Pb flux and crustal enrichment in central Greenland. *Geophysical Research Letters* 29(23): 2130.
- Menon S, Hanesen J, Nazarenko L, and Luo Y (2002) Climate effects of black carbon aerosols in China and India. *Science* 297: 2250–2253.
- Planchon FAM, et al. (2002) Changes in heavy metals in antarctic snow from Coats Land since the mid-19th to late-20th century. *Earth and Planetary Science Letters* 200: 207–222.
- Prospero JM, Ginoux P, Torres O, Nicholson SE, and Gill TE (2002) Environmental characterization of global sources of atmospheric soil dust identified with the NIMBUS 7 total ozone mapping spectrometer (TOMS) absorbing aerosol product. *Reviews of Geophysics* 40(2–11): 2–31.
- Rahn KA and Lowenthal DH (1984) Elemental tracers of distant regional pollution aerosols. *Science* 223: 132–139.
- Robok A and Free M (1996) The volcanic record in ice cores for the past 2000 years. In: Jones P, Bradley R, and Jouzel J (eds.) *Climatic Variations and Forcing Mechanisms or the Last 2000 Years*, pp. 533–546. Springer.
- Simoneit BRT (2002) Biomass burning – A review of organic tracers for smoke from incomplete combustion. *Applied Geochemistry* 17: 129–162.
- Tegen I, Lacis AA, and Fung I (1996) The influence on climate forcing of mineral aerosols from disturbed soils. *Nature* 380: 419–422.
- Thompson LG, Mosley-Thompson E, Davis ME, et al. (1995) Late glacial stage and Holocene tropical ice core records from Huascaran, Peru. *Science* 269: 46–50.
- Wolff EW and Suttie ED (1994) Antarctic snow record of southern hemisphere lead pollution. *Geophysical Research Letters* 21: 781–784.

Stable Isotopes

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Introduction

Stable isotope ratios measured in ice ($^{18}\text{O}/^{16}\text{O}$ and D/H) provide some of the most fundamental data sets for ice-core paleoclimatology. This field was pioneered in Greenland in the 1960s by Dansgaard and colleagues and the classic paper outlining why stable isotopes in ice are affected by temperature change was published by Dansgaard in 1964. Since then stable isotope records have been measured for virtually all ice cores collected, providing a rich data set with which to understand the controls on the isotope ratios and what they can tell us about past environmental conditions. This is a large field, and this review attempts to summarize the important points necessary for understanding isotopic time series from polar ice-core records. Processes affecting ice-core records from tropical and temperate glaciers can be significantly different than for polar sites, and are discussed in separate articles on cores from such sites.

Standard Notation and Measurements

The stable isotope pairs normally measured in ice cores are $^{18}\text{O}/^{16}\text{O}$ and D/H of water. The majority of natural water molecules contain the lighter isotopes (e.g., H_2^{16}O). Heavier forms are present in small but measurable amounts. H_2^{18}O is approximately 0.2% of the total in natural waters, and HD^{16}O is less than 0.1%, but varies widely due to the influence of the large relative mass difference between D and H on isotopic fractionation between liquid, vapor, and solid (Faure, 1986). D_2O is present, but in very small quantities, as are forms containing ^{17}O , which is not normally measured in this application. In recent ice-core literature $^{18}\text{O}/^{16}\text{O}$ and D/H of ice are sometimes referred to as 'water isotopes' to distinguish them from isotopes of gases in air bubbles. These ratios are measured by standard dual inlet gas source isotope ratio mass spectrometry. Oxygen isotopes are normally measured by equilibrating the water with carbon dioxide (CO_2) with known oxygen isotope ratio. Through isotope exchange reactions the CO_2 takes on the isotope ratio of the water in a quantifiable way, and the CO_2 gas is inlet to the mass spectrometer for measurement. D/H ratios are measured after converting the water to hydrogen gas by reduction over hot metal. In both cases the sample ratios are repeatedly measured relative to a standard to obtain the high precision needed. The normal standard used for virtually all ice-core measurements is maintained by the International Atomic Energy Agency and called Vienna Standard Mean Ocean Water (V-SMOW). The D/H ratio for V-SMOW is 155.75×10^{-6} and the $^{18}\text{O}/^{16}\text{O}$ ratio is 2005.2×10^{-6} .

Natural water and ice vary by small amounts relative to these values. For this reason isotope ratios are expressed in

'delta' notation, in units of parts per thousand, or 'per mil' (‰) as follows:

$$\delta^{18}\text{O} = \frac{(^{18}\text{O}/^{16}\text{O}_{\text{sample}}) - (^{18}\text{O}/^{16}\text{O}_{\text{standard}})}{(^{18}\text{O}/^{16}\text{O}_{\text{standard}})} \times 1000,$$

or

$$\delta\text{D} = \frac{(\text{D}/\text{H}_{\text{sample}}) - (\text{D}/\text{H}_{\text{standard}})}{(\text{D}/\text{H}_{\text{standard}})} \times 1000$$

In general, polar ice is depleted in the heavy isotope with respect to seawater, for reasons outlined below, making delta values for ice-core samples negative. δD and $\delta^{18}\text{O}$ have been measured somewhat interchangeably over the history of ice-core research, primarily as a matter of convenience, but the combination of the two measurements provides additional information, as described in the section below on deuterium excess.

Basic Theory and Observations of Modern Isotope Distribution in Polar Regions

As mentioned above, stable isotope records have utility in ice-core studies for both paleothermometry and stratigraphy. The dependence of $\delta^{18}\text{O}$ and δD values on temperature in polar regions is commonly accepted, but the origin of this dependence is indirect, rather than arising from a simple temperature dependence of equilibrium isotopic fractionation in a chemical reaction, and is complicated by several factors. To start with, a simple picture, derived from the work of Dansgaard et al. (1973), serves to illustrate the origin of the temperature dependence (Figure 1).

Water evaporated from lower latitude oceans and continents is transported poleward as vapor. The vapor pressure of H_2^{16}O is slightly higher than that of H_2^{18}O (true also for H_2^{16}O versus HD^{16}O) and as a result vapor evaporated from seawater is depleted in the heavy isotope relative to the seawater source. Cooling of this vapor as it travels poleward causes condensation, removing water from the vapor phase. The condensed water is enriched in the heavy isotope (because the vapor pressure of water molecules containing the heavier isotope is lower than for water molecules containing the light isotope, slightly favoring condensation). The vapor parcel is thus progressively depleted in the heavier isotope as it moves toward the pole, and as a result $\delta^{18}\text{O}$ and δD values for polar snow and ice are highly negative. This is the underlying reason the oxygen isotope composition of the ocean varies with time – as the ice sheets grow they progressively remove isotopically depleted water from the ocean, leaving ocean water isotopically enriched. Because of the rather complex processes that lead to depletion of $\delta^{18}\text{O}$ and δD over the ice sheets these isotope

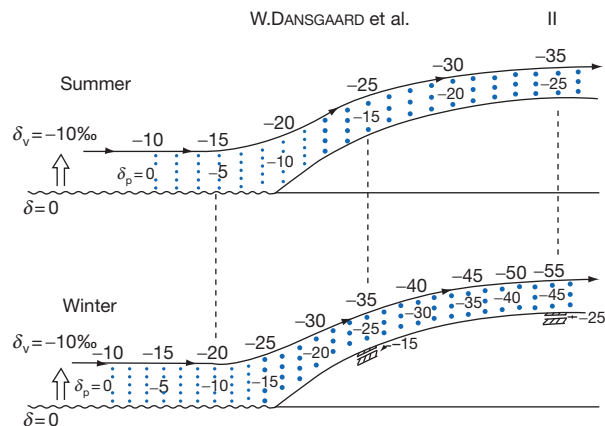


Figure 1 Schematic illustration of isotope fractionation ($\delta^{18}\text{O}$) from source in tropical ocean water to deposition site on an ice sheet. $\delta^{18}\text{O}$ of water vapor and precipitation is shown for summer and winter. Cooler winter temperatures lead to condensation and removal of a larger amount of water vapor during transport of moisture poleward, and a corresponding greater depletion of the heavy isotope in precipitation finally reaching the polar site. In an analogous way colder climates lead to greater isotopic depletion relative to warmer climates. The isotope values on this figure are primarily for heuristic purposes and the depiction neglects the temperature dependence of isotopic fractionation between liquid, vapor, and solid, which enhances the temperature dependence of isotope ratios at high (cold) latitudes (Cuffey and Brook, 2000). From Dansgaard et al. (1973).

ratios are not directly linked just to the surface temperature at an ice-core site. Instead, the simple picture outlined above suggests that $\delta^{18}\text{O}$ and δD are controlled by the temperature difference between the site and the moisture source. However, because temperature changes at high latitudes are generally larger than those at low latitudes a relationship between surface temperature and $\delta^{18}\text{O}$ and δD is expected (Patterson, 1994). The temperature dependence of isotopic fractionation is also important in controlling isotopic changes from source to ice-core site, and a simple mathematical description of the influence of both factors in a basic Rayleigh fractionation model is provided by Cuffey and Brook (2000). This simple picture is complicated by several factors which are discussed further below.

Measurements of $\delta^{18}\text{O}$ and δD in snow collected at various locations in Greenland and Antarctica validate this expectation and show a strong linear relationship with a slope of $\sim 0.6\text{--}0.7\text{‰}/^\circ\text{C}$ (Figure 2). δD values are generally $\sim 8 \times \delta^{18}\text{O}$; deviations from this relationship are caused by secondary processes that become important in interpreting isotope data and are discussed below. The figure of $0.6\text{--}0.7\text{‰}/^\circ\text{C}$ for oxygen isotopes is often called the 'spatial slope' to denote its origin in the spatial variation of isotope values at sites with different mean annual temperatures. The spatial slope has often been used as a calibrating isotopic thermometer and applied to ice-core records for reconstructing past temperature. This approach, substituting space for time in a climate reconstruction, is common in paleoclimatology, but as discussed below, is not entirely accurate for reconstructing temperature from ice cores.

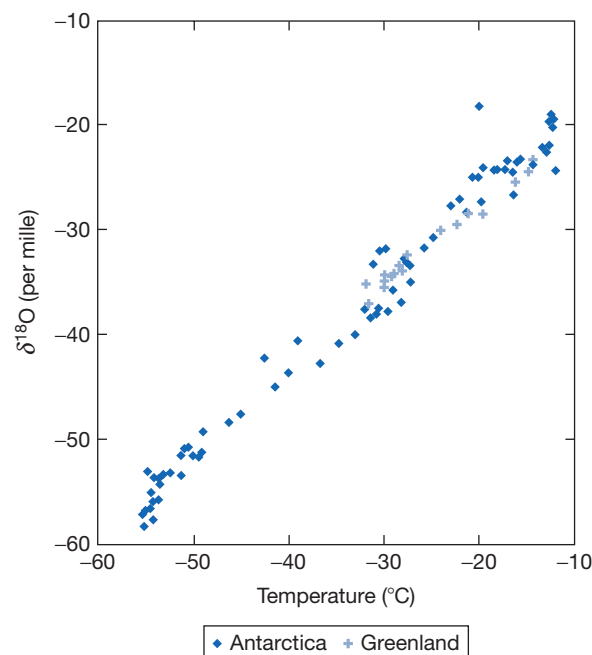


Figure 2 Relationship between $\delta^{18}\text{O}$ of snow in Greenland and Antarctica and surface temperature. From Johnsen et al. (1989) and Dahe et al. (1994). Reproduced from Cuffey and Brook (2000).

One further issue regarding isotope records is relevant. Early work on stable isotopes in polar regions revealed large seasonal variations in the isotopic composition of snow but found that those variations were damped after deposition in the firn (snowpack). This is due to diffusion and redeposition of water vapor in the firn. Seasonal variations in $\delta^{18}\text{O}$ and δD are damped to the extent that at low deposition rate sites (e.g., deep ice-coring sites in Antarctica with accumulation rates $\sim 2\text{ cm/year}$) seasonal cycles in isotopic composition are not preserved. Even at high accumulation sites (e.g., deep drilling sites in Greenland with accumulation rates $>20\text{ cm/year}$) seasonal variations in isotope ratios are damped (Johnsen et al., 2000). Johnsen et al. traced the history of this problem and describe modeling with the goal of correcting isotope profiles for diffusion and reconstructing seasonal cycles. They also showed that diffusion in solid ice further smooths seasonal variations. Sampling ice cores at seasonal resolution is common for shallow cores covering the last few centuries and such studies must consider the impact of smoothing due to diffusion. Sampling of deeper cores generally averages over many seasonal cycles. Seasonality in isotope ratios is an important issue, however, because bias in isotope records can be created by changes in the timing and amount of precipitation throughout the year.

Reliability of the Isotope Thermometer

There are a variety of reasons to think that the scenario outlined in Figure 1 oversimplifies vapor transport to the extent that the reliability of calibration based on modern measurements (spatial slope) for paleotemperature reconstruction

can be questioned. Rather than depending on a single well-understood temperature-dependent isotopic equilibrium, like some paleothermometers, the isotopic composition of polar precipitation is dictated by the influence of processes at the moisture source, during transport, and at deposition. At the source, one problem is that the whole ocean water isotopic composition changes with time as the ice sheets wax and wane. This can be accounted for using isotope records from calcareous fossils ocean sediments as proxies for changing ocean isotopic composition (see Jouzel et al., 2003 and Kavanaugh and Cuffey, 2003 for recent discussion of correction techniques), but these contain a mixed signal of changes in water isotopic composition and the temperature dependence of isotope fractionation on the formation of CaCO_3 skeletons, creating some uncertainty in the corrections. Recent work on independent temperature estimates for the deep sea using Mg/Ca ratios will help in this regard.

A second problem at the source is that the location, and therefore possibly the temperature and isotopic composition, of the moisture source may change with time. This might arise due to changes in storm tracks (Charles et al., 1994) or other large-scale meteorological parameters. A third is that positively correlated changes in moisture source temperature and ice-core site temperature will damp inferred temperature changes from the ice-core isotope record. For example, Boyle (1997) suggested that a 5°C cooling of the tropics and subtropics during glacial times would change the apparent isotope-temperature sensitivity in such a way that the glacial-interglacial temperature change would be underestimated by using the spatial slope.

Complications also potentially arise during vapor transport. 'Recycling' of water vapor between the ocean and atmosphere complicates the simple picture of Figure 1, though the significance diminishes at high latitudes (Kavanaugh and Cuffey, 2003; Hendricks et al., 2000). Kinetic fractionation during snow formation in clouds, the possible influence of impurities, and other processes in clouds and precipitation have also been considered (Jouzel et al., 1997; Alley and Cuffey, 2000; Kavanaugh and Cuffey, 2003).

Site processes are also significant. First, seasonal and shorter-term meteorological variations at the ice-core site may be important (Cuffey and Brook, 2000). Seasonal temperature variations at high latitudes are very large, creating seasonal variations in the isotopic composition of snow. Shifts in the timing and amount of precipitation within the year could bias isotope records (Steig, 1994; Denton et al., 2005) as could shifts in the location of storm tracks. This topic has been discussed extensively for Greenland (for discussion see Denton et al., 2005), and may contribute to the discrepancy between spatial slope and isotope-temperature relationships derived from other methods (see below). For Antarctica the influence of changes in seasonality is thought to be of secondary importance (Jouzel et al., 2003). Finally, since snowfall rates are higher under warmer conditions in the polar regions the isotopic composition of snow may conceivably be biased toward shorter term events caused by the poleward advection of warm air (Cuffey and Brook, 2000).

Another site phenomenon to consider is the strength of the temperature inversion (warmer air aloft than at the surface) that can exist over an ice sheet. The isotopic composition of snowfall depends on the temperature where precipitation

forms in the atmosphere (Jouzel et al., 1997) rather than the ice-sheet surface temperature. A change in the strength of the inversion (e.g., between glacial and interglacial times) could bias the reconstruction of temperature from the isotope record. Krinner et al. (1997) examined this effect for Greenland in a global circulation model (GCM) study and found it to be relatively small.

Finally, deep ice-core isotope records are typically interpreted in terms of changes in temperature with respect to modern climate at the ice-core site (whether it be surface or air temperature aloft). The dynamics of ice flow can complicate this simple scenario, however. If ice in deeper sections of a core was deposited at higher altitudes than the core site (possible if the core is not drilled on an ice divide) the isotopic composition of that ice would reflect both differences in climate and the impact of elevation change (every 100 m change in elevation results in a $\sim 0.6^\circ\text{C}$ change in temperature). This problem is important for interpreting some well-known ice-core records, e.g., those from Camp Century and Dye 3 sites in Greenland, and to a lesser extent the Vostok and Byrd sites in Antarctica.

In light of the complications of the temperature dependence of isotope ratios in polar precipitation described above several studies have used more complex atmospheric models incorporating isotopes to examine the reliability of paleotemperature reconstructions using isotopes. In general, GCM studies that incorporate water isotopes have confirmed the temperature dependence of isotope ratios in high latitude snowfall (Jouzel et al., 1997; Delaygue et al., 2000; Werner et al., 2001; Kavanaugh and Cuffey, 2003) although such studies have also shown how the phenomena discussed above can influence the isotope ratio-temperature calibration and reveal spatial variations in the isotope ratio versus temperature relation. GCM studies for Antarctic precipitation generally confirm the validity of using the spatial slope for determining temporal temperature trends, though tend to suggest temporal slopes are slightly lower than spatial slopes (Jouzel et al., 2003). GCM and other modeling studies for Greenland tend to confirm other work (below) that indicates that the spatial slope is not an accurate calibration for determining temperature from Greenland ice cores (Hendricks et al., 2000; Krinner et al., 1997; Delaygue et al., 2000).

Independent Tests of the Isotope Thermometer

As a result of the problems raised in the previous section, a number of recent studies have attempted to test the application of the isotopic thermometer with independent temperature estimates. For example, in Greenland, Shuman et al. (1995) compared isotope ratios in snow pits at Summit Greenland with satellite-derived and direct measurements of temperature over several seasonal cycles. They obtained excellent agreement between temperature and isotopic trends. They inferred an isotope-temperature sensitivity of $\sim 0.4\text{--}0.5\text{‰}/^\circ\text{C}$, similar to, but slightly lower than, the spatial sensitivity. Groote and Stuiver (1997) compared surface temperature and the isotopic composition of newly fallen snow, frost, and water vapor. They found that water vapor had a sensitivity higher than the spatial slope, $0.83\text{‰}/^\circ\text{C}$, and suggested that water vapor exchange and other processes during vapor transport may have

influenced this result. Similar work in Antarctica is described in the chapter entitled 'Stable isotope records from Antarctic Ice Cores'.

Instrumental calibrations are useful but clearly cannot test the accuracy of the isotopic thermometer for large climate changes in the more distant past. Two independent approaches have provided such a test, but with very different temporal resolution. The first independent test of the isotope thermometer is derived from borehole temperature measurements. The temperature profile in an ice sheet is controlled by the surface temperature history, heat transfer in the ice, and geothermal heat flux. As described in the chapter entitled 'Borehole temperature records and inferences of temperature change', under appropriate conditions the modern temperature profile measured in a fluid-filled borehole can be inverted to reveal a temperature history. This temperature history is a smoothed version of the true temperature history due to heat transport in the ice. Nonetheless, temperature changes over a variety of timescales (Little Ice Age, Holocene, glacial–interglacial) are accessible from borehole records, although information for times prior to the last glacial period is not preserved. Cuffey et al. (1995) describe an explicit test of the isotope thermometer with borehole data from the Greenland Ice Sheet Project-2 (GISP2) ice core. They used an inverse method in which they assumed the temperature history was proportional to the relative variations in the GISP2 isotope record without employing the calibration between isotope ratios and temperature. This method suggests varying isotopic sensitivity for different time periods. The glacial–interglacial transition has a sensitivity of about $0.33\text{‰}/^{\circ}\text{C}$ (Cuffey et al., 1995), while the sensitivity to mid-late Holocene cooling was smaller, $0.25\text{‰}/^{\circ}\text{C}$. At the termination of the Little Ice Age the sensitivity was $0.5\text{‰}/^{\circ}\text{C}$. Similar results were obtained using slightly different methods for the Greenland Icecore Project (GRIP) borehole record (Dahl-Jensen et al., 1998).

An alternate test of the isotope thermometer is provided by records of the stable isotopic composition of nitrogen and argon in air bubbles in Greenland ice cores. In the firn layer of an ice sheet gases are mobile and subject to both gravitational settling and thermal diffusion. Thermal diffusion impacts the isotopic composition of these gases when a temperature gradient is created in the firn, for example during abrupt temperature changes in Greenland or during seasonal temperature changes (Severinghaus et al., 2001). The result is an isotopic anomaly in nitrogen and argon isotopes that both marks the time of abrupt warming in the gas record and can be used to reconstruct temperature change. In the late 1990s, Jeff Severinghaus showed that by measuring nitrogen and argon isotopes, which have the same sensitivity to gravitational fractionation but different thermal diffusion sensitivities, one could determine temperature change independent of the water isotope record, at least for abrupt events (Severinghaus et al., 1998; Severinghaus and Brook, 1999). In the original work on this subject Severinghaus et al. demonstrated the potential of the paired argon and nitrogen isotope data, but uncertainties were too large to allow highly accurate temperature change estimates. They were able to estimate temperature change at the end of the Younger Dryas using the data in an alternate way, however. Using the nitrogen isotope anomaly as a stratigraphic marker they were able to estimate the age

difference between gas bubbles and the surrounding ice by counting the number of annual layers between the gas anomaly and the onset of warming recorded in the ice isotope record. This offset is temperature dependent, and allowed them to estimate a temperature of -46°C for the Younger Dryas, approximately 7°C colder than the estimate based on the spatial slope, and close to the value inferred from the temporal slope derived from borehole temperatures (Severinghaus et al., 1998). In later work, Severinghaus and Brook (1999) were able to use nitrogen and argon isotope data for the warming at the onset of the Bølling period to reveal an abrupt warming of $9 \pm 3^{\circ}\text{C}$ and an isotope-temperature slope of $0.38\text{‰}/^{\circ}\text{C}$, again close to the borehole value. (Subsequent work based on updated diffusion parameters adjusted the temperature change to $11 \pm 3^{\circ}\text{C}$; Severinghaus et al., 2003.) Later work on other abrupt changes revealed variable sensitivities (Landais et al., 2004a, b; Lang et al., 1999) but all with isotope-temperature sensitivity lower than those derived from the spatial slope model.

In Antarctica, borehole temperature reconstructions and the firn gas thermometer are more difficult to apply due to damping by heat diffusion for the borehole temperature case and the lack of abrupt temperature changes and high accumulation rate sites for the latter. Salamatin et al. (1998) analyzed Vostok borehole temperatures and concluded that spatial slopes underestimate temperature changes based on δD . Significant uncertainties limit the accuracy of the results (Jouzel et al., 2003). Caillon et al. (2003) studied a rapid warming in the Vostok record at about 107–108 thousand years (ka) using the N and Ar isotope technique and concluded that the spatial slope underestimated temperature change by $\sim 20\%$ for that event.

In summary, calibration and tests of the isotopic thermometer support the notion that δD and $\delta^{18}\text{O}$ values for ice samples qualitatively reflect temperature change, but also indicate that the sensitivity of the isotope ratios to temperature may change with time, and that substitution of space for time is not an appropriate calibration technique without independent testing. The problems are more apparent in Greenland records, where a variety of evidence implies a potential sensitivity range as seen by different methods of a factor of two. In Antarctica, however, available information suggests only a $\pm 20\%$ variation in isotope-temperature sensitivity. The latter conclusion is most relevant to the cold high elevation sites in East Antarctica (Jouzel et al., 2003). A number of reasons have been proposed to explain why the spatial and temporal slopes do not match in Greenland. The likely explanations include changes in seasonality, cooling of the moisture source, and changes in temperature inversion strength (Alley and Cuffey, 2000).

Deuterium Excess

The relationship between δD and $\delta^{18}\text{O}$ in polar regions is highly linear with a slope of approximately eight (Cuffey and Brook, 2000). The intercept of this relationship is not zero, however, and this offset, which arises due to kinetic isotope fractionation at the moisture source and fractionation during vapor transport (Cuffey and Brook, 2000; Kavanaugh and Cuffey, 2003) depends on climatic properties of interest including source

temperature and humidity. As a result, ice-core geochemists have defined this as deuterium excess, or d , as $d = \delta D - 8 \cdot \delta^{18}O$, and use it to examine source-related processes, primarily variations in source temperature (see [Kavanaugh and Cuffey, 2003](#), for a detailed discussion) which are of central concern for evaluating ice-core isotope records. Determining d requires that δD and $\delta^{18}O$ are measured on the sample. Recent examples of the use of deuterium excess in interpreting ice-core records include reconstructing source water temperatures for East Antarctic (Stenni et al., 2003) and Greenland ice cores ([Masson-Delmotte et al., 2005](#)) and correcting isotopic temperature records for source temperature changes ([Cuffey and Vimeaux, 2001](#); [Masson-Delmotte et al., 2005](#)).

Ice-Core Isotopic Stratigraphy

In this section the major features of ice-core stable isotope records from Greenland and Antarctica are reviewed. The reader is referred to separate articles on Antarctic records, Greenland records, and various articles on other ice-core records for more details. In general, isotope records from both hemispheres show the well-known cycling between cold glacial and warm interglacial states. In common with marine and other sedimentary climate records over these time periods the records have a strong 100 000 year cyclicity ([EPICA Dome community members, 2004](#)), as well as variation on 20 000 and 40 000 year timescales linked to Milankovitch cycles, at least as far back as the oldest ice-core record (720 000 years from the EPICA Dome C Core).

Deep ice cores from Greenland provide a detailed picture of climate change over the last 110 000–120 000 years, covering the Holocene and the last glacial period. Ice from the end of the last interglacial period is present in the North GRIP core ([NGRIP members, 2004](#)) and possibly in other Greenland cores ([Johnsen et al., 2001](#)) and older ice also appears to be present, but disturbed by ice flow and no longer in stratigraphic order ([Chappellaz et al., 1997](#); [Landaï et al., 2003](#); [Suwa et al. in press](#)). The most striking aspect of the Greenland records are the large isotopic fluctuations during the Ice Age ([Figure 3](#)) often called Dansgaard-Oeschger events. As discussed above, firn gas thermometry verifies that these represent large temperature changes. The chapter entitled ‘Methane Studies’ discusses the relationship between the abrupt events in Greenland and millennial-scale climate change in Antarctica and the chapter entitled ‘SubMilankovitch (DO/Heinrich) events’ discusses the abrupt events in a broader context.

Antarctica provides longer records due to colder conditions and lower accumulation rates, and the oldest Antarctic record, from EPICA Dome C, is plotted in [Figure 4](#). This record extends to 720 000 years, and may be lengthened by further drilling and analysis. The dominant 100 000 year cycle is evident as well as shorter term Milankovitch cyclicity mentioned above. The similarity of the Antarctic δD record to the marine stable isotope record, which reflects global glacial–interglacial conditions, is readily apparent, reinforcing the conclusion that the Antarctic ice-core record is an appropriate indicator of long-term trends in global climate.

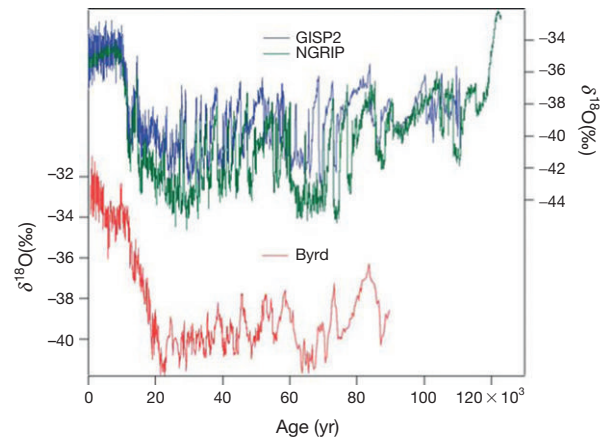


Figure 3 Stable isotope records for the last glacial period from Greenland (top) and Antarctica (bottom). GISP2 record from [Grootes et al. \(1993\)](#), NGRIP record from [NGRIP members \(2004\)](#), Byrd record from [Blunier and Brook \(2001\)](#).

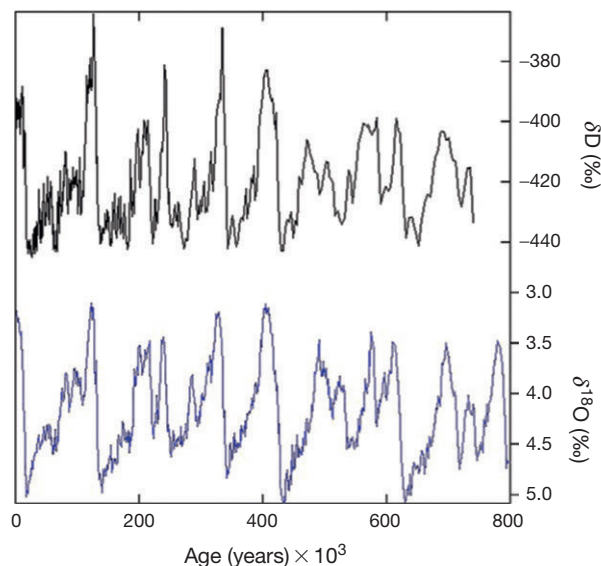


Figure 4 Top: Stable isotope record from the EPICA Dome C ice core for the last 720, 000 years ([EPICA community members, 2004](#)). Bottom: Oceanic benthic isotope stack from [Lisiecki and Raymo \(2005\)](#). The similarity between the two records underscores the global nature of the ice-core isotope variations.

See also: **Glaciation, Causes:** Astronomical Theory of Paleoclimates. **Ice Core Methods:** Borehole Temperature Records; Methane Studies. **Ice Core Records:** Africa; Antarctic Stable Isotopes; Chinese, Tibetan Mountains; Correlations Between Greenland and Antarctica; Greenland Stable Isotopes; Ice Margin Sites; South America; Thermal Diffusion Paleotemperature Records. **Ice Cores:** Dynamics of the West Antarctic Ice Sheet. **Paleoceanography, Physical and Chemical Proxies:** Mg/Ca and Sr/Ca Paleothermometry from Calcareous Marine Fossils; Oxygen-Isotope Stratigraphy of the Oceans. **Paleoceanography, Records:** Postglacial North Atlantic. **Paleoclimate Reconstruction:** Sub-Milankovitch (DO/Heinrich) Events.

References

- Alley RB and Cuffey KM (2000) Oxygen and hydrogen isotopic ratios of water in precipitation: Beyond paleothermometry. In: Valley J and Cole D (eds.) *Stable Isotope Geochemistry, Reviews in Mineralogy and Geochemistry*, vol. 34.
- Blunier T and Brook EJ (2001) Timing of millennial-scale climate change in Antarctica and Greenland during the last glacial period. *Science* 291: 109–112.
- Boyle EA (1997) Cool tropical temperatures shift the global $\delta^{18}\text{O}$ -T relationship: An explanation for the ice core $\delta^{18}\text{O}$ borehole thermometry conflict? *Geophysical Research Letters* 24: 273–276.
- Caillon N, Severinghaus JP, Barnola JM, Chappellaz JC, Jouzel J, and Parrenin F (2001) Estimation of temperature change and of gas age-ice age difference, 108 kyr BP, at Vostok, Antarctica. *Journal of Geophysical Research* 106: 31,893–31,901.
- Chappellaz J, Brook E, Blunier T, and Malaize B (1997) CH_4 and $\delta^{18}\text{O}$ of O_2 records from Greenland and Antarctic ice: A clue for stratigraphic disturbance in the bottom part of the GRIP and GISP2 ice cores. *Journal of Geophysical Research* 102: 26,547–26,557.
- Charles C, Rind D, Jouzel J, Koster R, and Fairbanks R (1994) Glacial-interglacial changes in moisture sources for Greenland: Influences on the ice core record of climate. *Science* 261: 508–511.
- Cuffey K and Brook EJ (2001) Chapter 18: Ice core paleoclimate records. In: Jacobsen (ed.) *Global Biogeochemical Cycles*. New York: Academic Press.
- Cuffey KM and Vimeux F (2001) Covariation of carbon dioxide and temperature from the Vostok ice core after deuterium-excess correction. *Nature* 421: 523–527.
- Cuffey KM, Clow GD, Alley RB, Stuiver M, Waddington ED, and Saltus RW (1995) Large Arctic temperature change at the Wisconsin-Holocene glacial transition. *Science* 270: 455–458.
- Dahe Q, Petit JR, Jouzel J, and Stievenard M (1994) Distribution of stable isotopes in surface snow along the route of the 1990 International Trans-Antarctic Expedition. *Journal of Glaciology* 40: 107–118.
- Dahl-Jensen D, Mosegaard K, Gundestrup N, et al. (1998) Past temperatures directly from the Greenland ice sheet. *Science* 282: 268–271.
- Dansgaard W (1964) Stable isotopes in precipitation. *Tellus* 16: 436–468.
- Dansgaard W, Johnsen SJ, Clausen HB, et al. (1973) Stable isotope glaciology. *Meddelelser om Grønland* 197: 1–53.
- Delaygue G, Jouzel J, Masson V, Koster RD, and Bard E (2000) Validity of the isotopic thermometer in central Antarctica: Limited impact of glacial precipitation seasonality and moisture origin. *Geophysical Research Letters* 27: 2677–2680.
- Denton G, Alley R, Comer G, and Broecker W (2005) The role of seasonality in abrupt climate change. *Quaternary Science Reviews* 24: 1159–1182.
- EPICA community members (2004) Eight glacial cycles from an Antarctic ice core. *Nature* 429: 623–628.
- Faure G (1986) *Principles of Isotope Geology*, 2nd Edition London: John Wiley & Sons.
- Groote PM and Stuiver M (1997) Oxygen 18/16 variability in Greenland snow and ice with 10–3 – to 105 – year time resolution. *Journal of Geophysical Research* 102: 26,455–26,470.
- Groote PM, Stuiver M, White JWC, Johnsen S, and Jouzel J (1993) Comparison of oxygen isotope records from the GISP2 and GRIP Greenland ice cores. *Nature* 366: 552–554.
- Hendricks MB, De Paolo DJ, and Cohen RC (2000) Space and time variation of δD and $\delta^{18}\text{O}$ in precipitation: Can paleotemperature be estimated from ice cores? *Global Biogeochemical Cycles* 14: 851–861.
- Johnsen SJ, Clausen HB, Cuffey KM, Hoffmann G, Schwander J, and Creyts T (2000) Diffusion of stable isotopes in polar firn and ice: The isotope effect in firn diffusion. In: Boutron C (ed.) *Physics of Ice Core Records*, pp. 121–140. Sapporo, Japan: Hokkaido University Press.
- Johnsen SJ, Dansgaard W, and White JW (1989) The origin of Arctic precipitation under present and glacial conditions. *Tellus Series B* 41: 452–469.
- Jouzel J, Alley RB, Cuffey KM, et al. (1997) Validity of the temperature reconstruction from water isotopes in ice cores. *Journal of Geophysical Research* 102: 26,471–26,487.
- Jouzel J, Vimeux F, Caillon N, et al. (2003) Magnitude of the isotope/temperature scaling for interpretation of central Antarctic ice cores. *Journal of Geophysical Research* 108: 1029–1046.
- Kavanaugh J and Cuffey K (2003) Space and time variation of $\delta^{18}\text{O}$ and δD in Antarctic precipitation revisited. *Global Biogeochemical Cycles* 17: 1017.
- Krinner G, Genthon C, and Jouzel J (1997) GCM analysis of local influences on ice core δ signals. *Geophysical Research Letters* 24: 2825–2828.
- Landais A, Barnola JM, Masson-Delmotte V, et al. (2004a) A continuous record of temperature evolution over a whole sequence of Dansgaard-Oeschger during Marine Isotopic Stage 4 (76 to 62 kyr BP). *Geophysical Research Letters* 31: L22211.
- Landais A, Caillon N, Jouzel J, et al. (2004b) A method for precise quantification of temperature change and phasing between temperature and methane increases through gas measurements on Dansgaard-Oeschger event 12 (–45 kyr). *Earth and Planetary Sciences Letters* 225: 221–232.
- Landais A, Chappellaz J, Delmotte MJ, et al. (2003) A tentative reconstruction of the last interglacial and glacial inception in Greenland based on new gas measurements in the Greenland Ice Core Project (GRIP) ice core. *Journal of Geophysical Research* 108: D18.
- Lang C, Leuenberger M, Schwander J, and Johnsen S (1999) 16°C rapid temperature variation in central Greenland 70,000 years ago. *Science* 286: 934–937.
- Liesicki LE and Raymo ME (2005) A Pliocene-Pleistocene stack of 57 globally distributed benthic $\delta^{18}\text{O}$ records. *Paleoceanography* 20.
- Masson-Delmotte V, Jouzel J, Landais A, et al. (2005) GRIP Deuterium Excess Reveals Rapid and Orbital Scale Changes in Greenland Moisture Origin. *Science* 309: 118–121.
- NGRIP Members (2004) High resolution record of northern hemisphere climate extending in to the last interglacial period. *Nature* 431: 147–1141.
- Paterson WSB (1994) *The Physics of Glaciers*, 3rd edn. Oxford: Pergamon.
- Salamatin AN, Lipenkov VY, Barkov NI, Jouzel J, Petit JR, and Raynaud D (1998) Ice core age dating and paleothermometer calibration based on isotope and temperature profiles from deep boreholes at Vostok Station (East Antarctica). *Journal of Geophysical Research* 103: 8963–8977.
- Severinghaus J, Grachev A, Luz B, and Caillon N (2003) A method for precise measurement of argon 40/36 and krypton/argon ratios in trapped air in polar ice with applications to past firn thickness and abrupt climate change in Greenland and at Siple Dome, Antarctica. *Geochimica et Cosmochimica Acta* 67: 325–343.
- Severinghaus JP and Brook EJ (1999) Abrupt climate change at the end of the last glacial period inferred from trapped air in polar ice. *Science* 286: 930–934.
- Severinghaus JP, Sowers T, Brook E, Alley RB, and Bender ML (1998) Timing of abrupt climate change at the end of the Younger Dryas interval from thermally fractionated gases in polar ice. *Nature* 391: 141–148.
- Steig EJ, Groote P, and Stuiver M (1994) Seasonal precipitation timing and ice core records. *Science* 266: 1885–1886.
- Stenni B, Masson V, Johnsen SJ, et al. (2001) An oceanic cold reversal during the last deglaciation. *Science* 293: 2074–2077.
- Suwa M, von Fischer JC, Bender ML, Landais A, and Brook E (in press) Chronology reconstruction for the disturbed bottom section of the GISP2 and the GRIP ice cores: Implications for Termination II in Greenland. *Journal of Geophysical Research*.
- Werner M, Heimann M, and Hoffmann G (2001) Isotopic composition and origin of polar precipitation in present and glacial climate simulations. *Tellus Series B* 53: 53–71.

¹⁰Be and Cosmogenic Radionuclides in Ice Cores

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Introduction

Studies of cosmogenic radionuclides in ice cores combine two distinct advantages. On the one hand, ice cores represent the most direct natural archive of climate change since they directly preserve past precipitation, its chemical and isotopic composition, and the ambient atmosphere. On the other, cosmogenic radionuclides represent a relatively simple system with a well-known production in the atmosphere and an even better-constrained sink–radioactive decay. Therefore, it is possible to reconstruct the production history and the geochemical behavior of cosmogenic radionuclides in great detail. Examples of this research include solar activity reconstructions, geomagnetic field reconstructions, dating/correlation of ice cores, and the study of atmospheric circulation and depositional processes.

There is a long list of cosmogenic radionuclides but only a few have suitable half-lives and concentrations to be of interest for ice-core studies. ¹⁰Be with a half-life of about 1 390 000 years (Chmeleff et al., 2010; Korschinek et al., 2010) is the most commonly measured cosmogenic radionuclide in ice cores. ³⁶Cl ($T_{1/2}$ = 301 000) is also often part of major ice-core drilling programs. These nuclides can be routinely measured in ice cores but their low concentrations usually lead to the requirement of large sample sizes (in the order of 100 g of ice) and sophisticated measurement techniques such as accelerator mass spectrometry (Litherland, 1980). Other cosmogenic radionuclides that have been measured in ice cores include ³H, ³²Si, ¹⁴C, ²⁶Al, and ⁸¹Kr (Lal and Jull, 1992, and references therein). In this article, the focus lies on cosmogenic radionuclides although the production of some of these nuclides by anthropogenic means (nuclear weapons and reactors, primarily) can be relevant (such as the bomb peak in ³⁶Cl). In the first section of this article, the processes influencing cosmogenic radionuclides from production to deposition in ice cores will be discussed. In the second section, some of the most important results of cosmogenic radionuclide studies in ice cores will be presented.

Cosmogenic Radionuclide Basics

Three factors influence cosmogenic radionuclides in ice cores: production, deposition, and postdepositional processes in the ice (Figure 1). The importance of these processes for the interpretation of radionuclides in ice cores is examined.

Production

Cosmogenic radionuclides, as the name indicates, are produced by the interaction of galactic cosmic rays (GCRs) with the atoms in the atmosphere and on the Earth's surface. High-energy cosmic rays originate in the galaxy and their origin is probably connected to supernova explosions. Different processes are responsible for modulation of the cosmic ray flux

reaching the Earth's atmosphere and, therefore, the varying cosmogenic radionuclide-production rates.

Solar modulation

GCRs mainly consist of charged particles. The nuclear component consists of 87% protons, 12% α -particles, and 1% heavier nuclei (Simpson, 1983). In the heliosphere, they interact with the solar magnetic field that is carried by the solar wind. The amount of GCRs reaching the Earth's atmosphere depends on the strength of the magnetic field and the energy of the GCR particles. Lower energy GCR particles get preferentially deflected with higher solar shielding (Lal and Peters, 1967). Low solar activity implies less solar magnetic shielding, more cosmic rays reaching the Earth, and higher production rates of cosmogenic radionuclides. Without going into the details it should be noted that the drift of the GCRs in the heliosphere also depends on the polarity of the solar magnetic field. For example, the well-known solar 11-year cycle is a 22-year polarity cycle as sunspots indicate strong magnetic fields on the Sun but not the polarity. However, for most applications, the so-called force-field approximation (Gleeson and Axford, 1968), where the solar modulation process can be described with one single parameter, the so-called solar modulation function, is sufficient. Low solar activity implies a low solar modulation function and higher fluxes of GCRs.

Geomagnetic modulation

A similar shielding effect occurs around the Earth due to the geomagnetic field. The higher the field intensity, the larger is the shielding and the fewer GCRs can enter the atmosphere. This process is strongly dependent on the geomagnetic latitude. No modulation occurs at the geomagnetic poles due to vertical magnetic field lines (parallel to the incoming GCRs), while there is a maximal shielding effect at the equator (e.g., Masarik and Beer, 1999). An absent geomagnetic field would lead to an increase in the radionuclide-production rates of about a factor of 2.

When high-energy GCRs enter the atmosphere, they produce a cascade of reactions (which produces the so-called secondary cosmic rays) and in some instances these produce cosmogenic radionuclides. Most cosmogenic radionuclides are produced via spallation reactions on heavier atmospheric particles (e.g., ¹⁶O = >¹⁰Be, Ar = >³⁶Cl, etc.). However, radiocarbon (¹⁴C) is mainly produced via capture of thermal (slow) neutrons by nitrogen. Due to the different production processes, the ¹⁰Be and ¹⁴C production rates react slightly differently to solar and geomagnetic field changes (Figure 2).

GCR variations

In most studies, it is assumed that the GCR flux was constant over the periods of interest in ice-core studies. From meteorite investigations and additional terrestrial evidence, it can be concluded that the GCR intensity has been constant within

10% during the past 10 million years (e.g., [Wielers et al., 2011](#)). In addition to the meteorite evidence, there is presently no need to invoke GCR variations as an explanation for ice-core cosmogenic radionuclide records. For example, one of the most prominent and reproducible ^{10}Be changes is an increase by about a factor of 2 about 40 000 years ago. In early studies, a GCR change was suggested as a potential (but unlikely) explanation ([Raisbeck et al., 1987](#)). Later on, it became clear that this event can be fully explained by a geomagnetic field minimum ([Baumgartner et al., 1998](#)).

Solar proton events

In general, solar cosmic rays, that is, particles ejected from the Sun, are not energy-rich enough to produce significant

amounts of cosmogenic radionuclides. Nevertheless, during solar proton events (SPEs), when the Sun emits a large amount of higher-energy particles, solar particles might produce detectable increases in radionuclide-production rates on short timescales. During such an event, the number of particles penetrating the atmosphere at the polar regions can be several orders of magnitude higher than the flux of GCRs ([Lal and Peters, 1967](#)). However, the energy of the solar particles is, on average, much lower than that of the GCRs. Therefore, the solar proton contribution to the radionuclide-production rates depends strongly on the energy thresholds for the production of the different radionuclides ([Lal and Peters, 1967](#)). For ^{10}Be , it has been suggested that SPEs contribute about 1–2% of the total average ^{10}Be -production rate ([Usoskin et al., 2006](#)). There are some studies that discuss the possibility of detectable SPEs in ice-core ^{10}Be records (e.g., [Pedro et al., 2009](#); [Usoskin et al., 2006](#)). However, the studies are rather vague on the reliability of the solar proton signal in the data as weather and climate ‘noise’ potentially masks the small effects due to SPEs. This is illustrated in a publication by [Pedro et al. \(2011\)](#), which shows that a ^{10}Be increase, that was previously suggested to be of SPE origin, did not stand out as a clear signal in an extended ^{10}Be record. The discussion concentrates on the GCR-related production processes and the results that can be drawn from them.

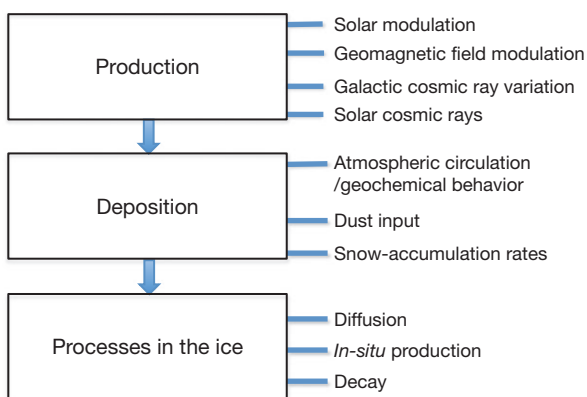


Figure 1 Illustration of the processes that are important for the interpretation of cosmogenic radionuclides in ice cores.

Transport and Deposition

The majority of the cosmogenic radionuclides are produced in the stratosphere (e.g., [Heikkilä et al., 2009a](#); [Masarik and Beer, 1999](#)). Therefore, the exchange processes between the stratosphere and troposphere influence the deposition rates

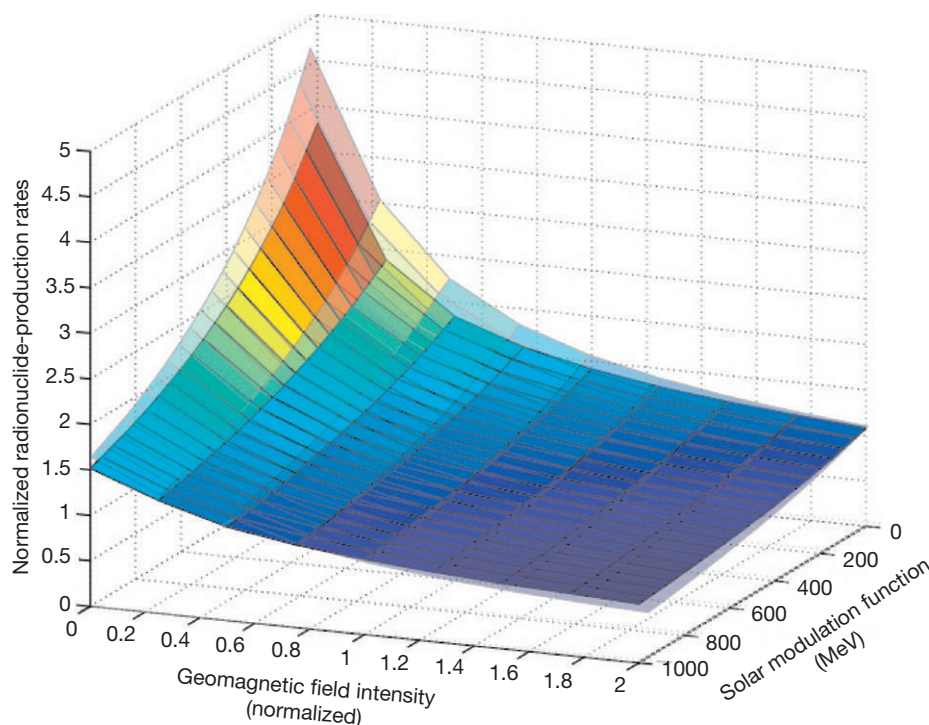


Figure 2 Globally averaged radionuclide-production rates according to [Masarik and Beer \(1999\)](#) depending on geomagnetic field intensity and solar-modulation function. The opaque surface shows the ^{10}Be -production rates and half-transparent ^{14}C -production rates.

of cosmogenic radionuclides. After the production, ^{10}Be attaches to aerosols (similar to ^7Be and ^{26}Al) and follows their pathways down to the Earth surface. ^{36}Cl , on the other hand, stays in a more gaseous form as HCl . ^{14}C oxidizes to CO_2 and enters the global carbon cycle. These differences influence the atmospheric lifetime, pathways, and deposition of the different cosmogenic radionuclides. ^{10}Be and ^7Be (similar to ^{36}Cl and ^{26}Al) are deposited within an average period of about 1 year (McHargue and Damon, 1991; Raisbeck et al., 1981a) while ^{14}C stays much longer (~ 5 years) in the atmosphere. The atmospheric pathways of ^{10}Be have been modeled with general circulation climate models (GCMs) (Field et al., 2006; Heikkilä et al., 2008a). The results show that the radionuclide-deposition rates depend on the precipitation rates and the preferred locations and season for the stratosphere–troposphere exchange in the mid-latitudes.

An important issue for polar ice-core records is whether they show a polar bias, that is, whether they deviate from the globally average production signal. As mentioned, there is a strong asymmetry of the cosmogenic radionuclide-production rates from the poles to the equator due to geomagnetic shielding. This effect also implies that the relative variations in cosmogenic radionuclide-production rates induced by a variable solar shielding are much larger at the poles than at the equator. Therefore, it has been suggested that this effect has to be taken into account when using ^{10}Be as a proxy for solar activity variations (Bard et al., 1997; Mazaud et al., 1994). GCM model results do not agree on this topic. While Field et al. (2006) see such a bias in their results, Heikkilä et al. (2009a) do not see such evidence in their higher resolution model. Comparison of ice-core ^{10}Be with ^{14}C data suggests a small polar bias, but the picture is not conclusive as this possible bias is not constant through time (Muscheler and Heikkilä, 2011).

There is ample evidence that the snow-accumulation rates have a strong effect on the ^{10}Be concentrations. The higher the accumulation rates, the lower the ^{10}Be concentrations. This applies to ^{10}Be deposition in space (e.g., differences between different locations; Delaygue and Bard, 2010) and in time (differences between different climate states; Wagner et al., 2001b). Alley et al. (1995) proposed separating the ^{10}Be signal into two components, where component I could best be explained with ‘wet deposition.’ For wet deposition one does not expect that the ^{10}Be concentrations depend on the snow-accumulation rates (more snow would imply proportionally more ^{10}Be deposition). Component II can be described as ‘dry deposition’ where additional snowfall would directly dilute the ^{10}Be signal and lead to lower ^{10}Be concentrations. In reality, the majority of the ^{10}Be is deposited with precipitation, with the exceptions of very low accumulation sites (Heikkilä et al., 2008a), but the wet deposition might indeed have a signature that is typical for dry deposition. For example, if atmospheric ^{10}Be gets depleted during the course of a precipitation event, the ^{10}Be concentration in the ice would depend on the precipitation rate. Another process that has to be considered at some locations is the input of ‘recycled’ cosmogenic radionuclides attached to dust. A nearby dust source could increase the input of cosmogenic radionuclides considerably and, therefore, perturb the signal if one is interested in the direct meteoric input of cosmogenic radionuclides. This problem is likely to be relevant for ice cores that are close to areas

that are significant sources for dust. In the center of the two large polar ice sheets, this effect is considered to be of minor importance (Baumgartner et al., 1997).

Postdepositional Processes

The most obvious process relevant for radionuclides in ice cores is radioactive decay. Depending on the half-life, it leads to a depletion of cosmogenic radionuclides with age. However, ice-core radionuclide studies are usually done with radionuclides that have a ‘suitable’ half-life, that is, such studies are usually focused on ^{10}Be and sometimes also on ^{26}Al and ^{36}Cl but hardly on shorter-lived radionuclides such as ^7Be . Nevertheless, there are other processes in the ice that might alter the radionuclide concentrations or the availability of the radionuclides for measurements.

Dust-borne ^{10}Be

Dust-borne (recycled) ^{10}Be is usually a factor that one wishes to exclude when interpreting polar ice-core records in terms of production-rate changes. This component can be separated by filtering the meltwater before sample preparation. However, measurements of ^{10}Be in meltwater and on filters indicate that the older the ice, the more ^{10}Be can be found on the dust (Baumgartner et al., 1997). The process is not satisfactorily understood, but the data could be explained with a time-dependent process that leads to an attachment of ^{10}Be to the dust grains (that are removed by filtering the meltwater through a fine-meshed filter). Processes within the ice matrix, such as recrystallization, might be responsible for the transport of ^{10}Be to the crystal borders where it attaches to the dust. However, an expected correlation with dust amount cannot be found. Similarly, Raisbeck et al. (2006) find large spikes in ^{10}Be concentrations in very old ice from the European Project for Ice Coring in Antarctica (EPICA) ice core drilled at DomeC (East Antarctic Plateau) that cannot always be confirmed by remeasuring ice of the same depth/age. This could indicate that, by coincidence, one can sample ice with dust that accumulated ^{10}Be and leads to considerable ^{10}Be increases. However, ice from the same depth does not necessarily contain the same amount of dust and ^{10}Be attached to it and some samples might be depleted of ^{10}Be .

^{36}Cl mobility

In contrast to ^{10}Be , ^{36}Cl stays in a gaseous phase in the form of HCl . Mobility of HCl in the firm could lead to a smoothening of the ^{36}Cl signal and a loss of ^{36}Cl back to the atmosphere. Based on the bomb- ^{36}Cl peak profile, it has been shown that radionuclide concentrations can, after their deposition, be altered by exchange processes within the firm. Delmas et al. (2004) investigated the ^{36}Cl bomb signal in the firm at the Vostok Station in central East Antarctica. The profile suggests that HCl is mobile in firm and that diffusion processes can alter the atmospheric input signal in the ice. This may seriously question the usefulness of the $^{10}\text{Be}/^{36}\text{Cl}$ dating method but this process is likely to be less important when accumulation rates are higher and the acidity of the ice is low (e.g., during ice-age conditions; Delmas et al., 2004).

In situ production

In situ production of cosmogenic radionuclides in the ice has been shown to be a significant process for ^{14}C (Fireman and Norris, 1982; Lal et al., 1990). This means that the cosmic ray flux at the top of an ice sheet can produce cosmogenic radionuclides in measurable amounts in the uppermost layers of the ice sheet. ^{14}C is produced via spallation reactions on the oxygen in the ice. Even below ~ 10 m, there could be a significant production of ^{14}C due to the muons in the GCR shower (Van Der Kemp et al., 2002). The *in situ* contribution is often considered negligible for ice-core ^{10}Be studies due to the relatively large amounts of meteoric ^{10}Be . However, especially at low accumulation sites, this process could add a ^{10}Be signal that cannot be neglected (Horiuchi et al., 2008).

Applications

In spite of the uncertainties in our knowledge regarding the production, transport, and deposition processes, there is a wealth of information that can be gained from ice-core radionuclide records. Some of the most important applications and results are presented here.

Solar Activity Reconstructions

The solar modulation of the GCR flux to the Earth's atmosphere leads to a solar signal in the radionuclide-production rates. Especially, ^{10}Be records have been used to reconstruct the history of the Sun. Beer et al. (1990) were the first to trace the solar 11-year cycle with high-resolution ^{10}Be measurements in an ice core from Southern Greenland. Major solar minima, such as the Maunder Minimum from about AD 1645 to 1715, are usually clearly visible also in lower resolution ^{10}Be data. Figure 3 shows a compilation of ice-core ^{10}Be records for the last 1000 years. The data show a common signal that can be attributed to a series of solar minima (Oort (AD ~ 1020 –1070), Wolf (AD ~ 1270 –1340), Spörer (AD ~ 1400 –1540), Maunder (AD ~ 1615 –1715), and Dalton (AD ~ 1800 –1830)), but there are also differences when it comes to the amplitude and timing in the different records. The differences can most likely be attributed to (1) climate and weather influences on the ^{10}Be transport and deposition, (2) measurement quality issues, and (3) dating uncertainties. Such unresolved uncertainties are usually not a problem for qualitative solar activity reconstructions, but they can lead to important differences when such records are used to provide a solar forcing record for the past. This is best illustrated by two widely different conclusions about past changes in solar activity. Bard et al. (2000) conclude that solar activity was similar to present levels around AD 1200, while Usoskin et al. (2003) suggest that solar activity reached a maximum in the second part of the twentieth century that was significantly higher than during the preceding 1000 years. The main difference is the underlying ^{10}Be data set that the authors based their conclusions on. The ^{10}Be record from the South Pole, which Bard et al. (2000) include in their analysis (Figure 3(c)), does not show a similarly strong decreasing trend during the twentieth century compared to the ^{10}Be record from Southern Greenland (Figure 3(e)), which Usoskin

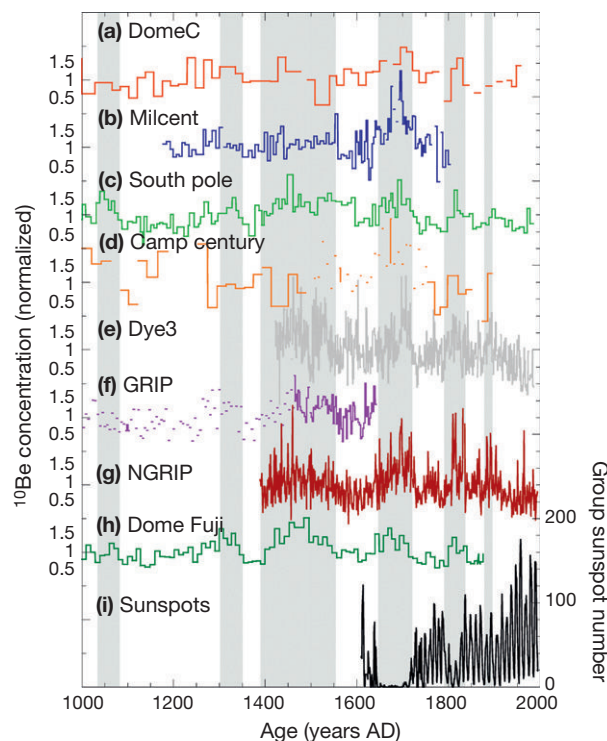


Figure 3 Several published ^{10}Be records for the last 1000 years from Greenland and Antarctica (Beer et al., 1988; Beer et al., 1990; Beer et al., 1991a; Berggren et al., 2009; Horiuchi et al., 2008; Muscheler et al., 2004; Raisbeck et al., 1990; Yiou et al., 1997) together with the group sunspot record (Hoyt and Schatten, 1998). Major solar minima are highlighted with the gray shading. Dots indicate noncontinuous measurements.

et al. (2003) prefer in their analysis (Muscheler et al., 2007; Raisbeck and Yiou, 2004). It is obvious that the implications for the inferred sun–climate link are very different depending on the solar-forcing reconstruction.

Going further back in time, there is one published high-resolution ^{10}Be record for approximately the last 9300 years (Muscheler et al., 2004; Vonmoos et al., 2006; see Figure 4). The ^{10}Be variations agree relatively well with the changes in the atmospheric ^{14}C levels as reconstructed from tree-ring ^{14}C measurements (Vonmoos et al., 2006). This implies that the major Maunder and Spörer Minimum-type solar minima can be reliably seen in ice-core ^{10}Be records over millennia back in time. However, the absolute levels of solar activity remain uncertain mainly because of differences between the different radionuclide records and uncertainties in the necessary geomagnetic field correction (Vonmoos et al., 2006).

The problem of disentangling the solar signal from ice-core ^{10}Be records gets even more challenging if one goes further back in time. The deglaciation and the end of the last Ice Age (about 11 600 years ago) implied changes in the atmospheric circulation and snow-deposition rates. The consequences for the ^{10}Be deposition are unclear. A simple subtraction of the geomagnetic influence on the GRIP ice-core ^{10}Be record would suggest high solar activity levels during the last Ice Age since the ^{10}Be data are lower than expected from the geomagnetic field data (Muscheler et al., 2004). However, without a better

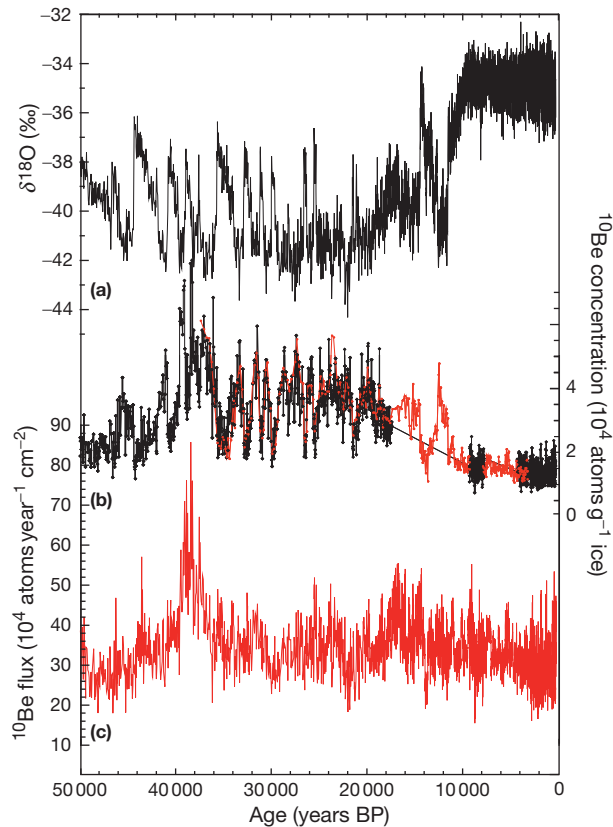


Figure 4 ^{10}Be concentration (panel (b)) and fluxes (panel (c)) from the GRIP and GISP2 ice cores from Central Greenland (Finkel and Nishiizumi, 1997; Muscheler et al., 2004; Vonmoos et al., 2006; Yiou et al., 1997) together with the GRIP $\delta^{18}\text{O}$ record (Johnsen et al., 1997) (panel (a)). The Dansgaard/Oeschger events during the last Ice Age imply accumulation-rate changes in the order of a factor of 2.

knowledge of the climate influence on the ^{10}Be deposition, such a conclusion cannot be drawn. Nevertheless, it is possible to concentrate on known solar changes even during periods of a highly varying climate. With this approach, it is possible to extract a limited but more reliable solar signal. Wagner et al. (2001c) found the solar 207-year cycle, which clearly stands out in the Holocene radionuclide records (Suess, 1982), in the ^{10}Be data from the GRIP ice core from 50 000 to 25 000 years before present supporting the solar origin for part of the ^{10}Be variability in the glacial section of this deep ice core from Central Greenland.

Geomagnetic Field Reconstructions

One of the most prominent features of ice-core ^{10}Be records is a doubling of the ^{10}Be concentration dating to about 40 000 years ago (Figure 4). Possible explanations were a climate influence, snow-accumulation rate changes, changes in primary cosmic ray intensity (e.g., due to a nearby supernova explosion), a geomagnetic field intensity minimum or low solar activity (Raisbeck et al., 1981b; Raisbeck et al., 1987). With improved data and better timescales, it could be shown that timing and amplitude of the radionuclide peak fit with the so-called Lascamp geomagnetic field minimum (Baumgartner

et al., 1998). In addition, for the last 60 000 years, there is a generally good agreement between independent geomagnetic field reconstructions and cosmogenic radionuclide fluxes to the GRIP ice core in Central Greenland (Muscheler et al., 2005; Wagner et al., 2000). Furthermore, Raisbeck et al. (2006) found the ^{10}Be signature of the last geomagnetic field reversal about 800 000 years in ice from the DomeC ice core.

From the ^{10}Be records from Central Greenland it appears that the geomagnetic modulation is the dominant process for controlling the cosmogenic radionuclide flux on timescales of 3000 years and longer. However, it is difficult to draw the line where solar modulation might be less important and where the geomagnetic modulation becomes dominant. Such investigations depend strongly on the quality of the independent geomagnetic field reconstructions. By comparing ice-core ^{10}Be records with independent geomagnetic field reconstructions, one can infer that the geomagnetic shielding cannot be neglected on timescales of 500 years and longer (Snowball and Muscheler, 2007).

^{14}C and Ice Cores – *In Situ* Production and Dating

^{14}C is the most important tool to date geological archives for the last 50 000 years. However, ^{14}C dating of polar ice cores is difficult as ^{14}C in air trapped in ice cores shows not only the atmospheric signal but is also strongly influenced by *in situ* production of ^{14}C in the snow and firn, mainly via interaction of secondary cosmic rays with oxygen (Lal et al., 2005; Wilson and Donahue, 1992) and via interactions of muons with oxygen (Nesterenok and Naidenov, 2011). The *in situ* production depends on snow-accumulation rates (i.e., how quickly the snow and firn get shielded) and cosmic ray intensity (geographic latitude, height, and solar activity). The additional *in situ* signal can dominate the measured ^{14}C concentrations.

Lal et al. (2005 and references therein) estimated the solar shielding based on the *in situ* ^{14}C signal in the GISP2 ice core. This method has the potential to add valuable information about long-term changes in solar activity. However, several processes complicate the interpretation. These consist of accumulation-rate changes that influence the shielding (Lal et al., 2005), variations in the local cosmic ray flux that are not related to solar activity (e.g., elevation changes; Lal et al., 2005), separation of the *in situ* and atmospheric signal, input of dust-borne carbon, and loss of ^{14}C from the firn (Smith et al., 2000). Nevertheless, Lal et al. (2005) conclude that it is possible to pinpoint periods of elevated and lower solar activity with this method.

^{14}C in low-latitude ice cores, where input from terrestrial carbon sources is more abundant, has been successfully used to date the ice (Jenk et al., 2009; Sigl et al., 2009).

Dating and Correlation

As mentioned, ^{14}C dating is possible in ice cores from low latitudes, but for polar ice-cores cosmogenic radionuclides provide additional possibilities to date the ice.

On the one hand, the radionuclide concentrations are negatively correlated to accumulation-rate changes and, therefore, allow us to estimate changes in accumulation rates (Wagner et al., 2001b) that then are a key parameter for age modeling.

On the other, the globally synchronous variations in the radionuclide-production rates allow us to synchronize different environmental archives that have detectable cosmogenic radionuclide variations due to solar or geomagnetic field variability (e.g., Finkel and Nishiizumi, 1997; Muscheler et al., 2000; Raisbeck et al., 2007; Yiou et al., 1997).

Finally, the radioactive decay of ^{10}Be and ^{36}Cl can be used to estimate the age of the lowermost/oldest ice of ice cores (Willerslev et al., 2007). The climate/accumulation rate influence on radionuclide records makes it difficult to use single radionuclide records for dating. However, it is possible to improve this method by analyzing radionuclide ratios under the assumption that climate influences have a similar impact on the radionuclides. The different half-lives of ^{36}Cl (301 000 years) and ^{10}Be (approx. 1 390 000 years (Chmeleff et al., 2010; Korschinek et al., 2010)) then lead to a time-dependent change in the $^{10}\text{Be}/^{36}\text{Cl}$ ratio ($^{10}\text{Be}/^{36}\text{Cl}$ half-life' = 384 000 years). With this approach, it was possible to constrain the age of the lowermost ice in the GRIP ice core to much older values than previously thought (Willerslev et al., 2007).

Accumulation-Rate Reconstructions

The influence of the variable-accumulation rate on radionuclide concentrations is a matter of current debate. The concept of separating the signal in a dry and wet deposition component was presented in the section 'Transport and Deposition.' For the GRIP ice core, the ^{10}Be concentration indicates that the signal can almost completely be explained by 'dry deposition.' For example, an accumulation-rate increase of a factor of two leads to roughly a halved ^{10}Be concentration (Figure 4). However, this effect is rather caused by an atmospheric depletion of ^{10}Be during periods of increased precipitation rather than by a real dry deposition process. Based on the varying ^{10}Be concentrations and known ^{10}Be production variations, Wagner et al. (2001b) estimated changes in past accumulation rates in Central Greenland during the strong climate fluctuations of the last Ice Age. This radionuclide-based estimate agrees well with layer thickness-inferred reconstructions of the snow-accumulation rate (Johnsen et al., 1995; Svensson et al., 2006; Wagner et al., 2001b).

Sun/Cosmic Ray–Climate Relation

The relation between solar activity changes and climate is a controversially discussed topic in climate change. Cosmogenic radionuclide records from ice cores have the advantage that solar/cosmic ray proxy data and climate data can be inferred from the same geological archive with virtually no errors in the relative dating. There are some examples where a solar influence on climate has been inferred from ice cores for the Holocene period (e.g., Grootes and Stuiver, 1997; O'Brien et al., 1995). However, the comparison of radionuclide and climate data for selected events suggests a complex relation between solar activity and climate change at the beginning of the Holocene (Björck et al., 2001). Furthermore, the hypothesis of a cosmic ray influence on climate can be tested with ice-core radionuclide and climate data. The doubling of the cosmic ray intensity during the geomagnetic field minimum around 40 000 years BP had no discernable effect on climate (Wagner et al., 2001a).

Atmospheric Circulation/Deposition

Atmospheric transport and deposition changes and their influence on cosmogenic radionuclides complicate the interpretation of these records in terms of production-rate changes. However, the major source region in the stratosphere and the deposition processes can provide complementary information about past circulation and climate changes and, therefore, be the subject of research in itself.

There are several studies where high-resolution (subannual) ^{10}Be records are presented and interpreted in terms of a combined signal of atmospheric circulation, stratosphere–troposphere exchange, and production. Pedro et al. (2011), for example, discuss several monthly resolution ^{10}Be records from Law Dome in East Antarctica. Besides the solar signal, there is a strong atmospheric signal probably related to a maximum stratosphere–troposphere exchange during the Southern Hemisphere late summer to early autumn. Beer et al. (1991b) report a bimodal (spring and fall) enhancement of ^{10}Be deposition during the year in Greenland. They suggest that both might be caused due to injection of stratospheric air. The autumn peak was proposed to be due to delayed transport of ^{10}Be from low to high Northern latitudes. Heikkilä et al. (2008b) discuss high-resolution data from a snow pit in Central Greenland and compare it to model results. They see maximum ^{10}Be concentrations during boreal summer. The differences highlight the fact that the atmospheric transport and weather/climate influences are still one of the biggest uncertainties in interpreting radionuclide records quantitatively.

The atmospheric transport can be studied with another cosmogenic radionuclide with a strong anthropogenic signal caused by the atmospheric nuclear weapon tests in the 1950s and early 1960s. These introduced comparably large amounts of ^{36}Cl into the atmosphere, visible as an increase by about a factor of 1000 in some archives (Heikkilä et al., 2009b). Combined model-data studies help to constrain the atmospheric ^{36}Cl input and test model performances with respect to transport and deposition of cosmogenic radionuclides.

Results from Low-Latitude Ice Cores

In general, low-latitude radionuclide ice-core records are more complicated to interpret. For example, the topography in mountainous areas leads to a more variable microclimate than on the large ice sheets on Greenland and Antarctica. In addition, low-latitude ice cores are closer to radionuclide sources other than the atmospheric production. For example, recycled ^{10}Be on dust cannot be neglected in Alpine ice cores. Radionuclide data from low latitudes has mainly been used for dating. This includes bomb-peak dating using ^{36}Cl or ^{137}Cs (Green et al., 2000; Thompson et al., 2002) or the ^{14}C dating of carbon from terrestrial sources mentioned earlier (Jenk et al., 2009; Thompson et al., 2005).

Conclusion

Radionuclide measurements in ice cores allow us to trace a variety of different natural and human-induced processes. These range from emission of anthropogenic sources of

radioactivity to dating of ice cores. An important field is the reconstruction of the solar and geomagnetic history over the past millennia and the relationship to climate change. Climate influences on radionuclide transport and deposition can complement climate reconstructions from independent natural tracers. The globally synchronous production-rate changes provide numerous marker horizons that allow us to correlate natural archives (e.g., ¹⁰Be in ice cores and ¹⁴C in tree rings). The low abundance of natural radionuclides in ice cores makes the measurements expensive and ice consuming. However, the wealth of information that can be gained from such records made radionuclide studies an integral part of the major deep drilling projects in polar areas.

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See also: **Cosmogenic Nuclide Dating: Exposure Geochronology. Ice Core Methods: Chronologies; Stable Isotopes. Radiocarbon Dating: ¹⁴C of Plant Macrofossils; AMS Radiocarbon Dating; Calibration of the ¹⁴C Record; Causes of Temporal ¹⁴C Variations; Charcoal; Conventional Method; Sources of Error; Variations in Atmospheric ¹⁴C.**

References

- Alley RB, Finkel RC, Nishiizumi K, et al. (1995) Changes in continental and sea-salt atmospheric loadings in central Greenland during the most recent deglaciation: Model-based estimates. *Journal of Glaciology* 41(139): 503–514.
- Bard E, Raisbeck GM, Yiou F, and Jouzel J (1997) Solar modulation of cosmogenic nuclide production over the last millennium: Comparison between ¹⁴C and ¹⁰Be records. *Earth and Planetary Science Letters* 150: 453–462.
- Bard E, Raisbeck GM, Yiou F, and Jouzel J (2000) Solar irradiance during the last 1200 years based on cosmogenic nuclides. *Tellus* 52B: 985–992.
- Baumgartner S, Beer J, Masarik J, Wagner G, Meynadier L, and Synal H-A (1998) Geomagnetic modulation of the ³⁶Cl flux in the GRIP ice core, Greenland. *Science* 279(5355): 1330–1332.
- Baumgartner S, Beer J, Wagner G, et al. (1997) ¹⁰Be and dust. *Nuclear Instruments and Methods in Physics Research B* 123: 296–301.
- Beer J, Raisbeck GM, and Yiou F (1991a) Time variations of ¹⁰Be and solar activity. In: Sonett P, Giampapa MS, and Matthews MS (eds.) *The Sun in Time*, pp. 343–359. Tucson: University of Arizona Press.
- Beer J, Siegenthaler U, Bonani G, et al. (1988) Information on past solar activity and geomagnetism from ¹⁰Be in the Camp Century ice core. *Nature* 331: 675–679.
- Beer J, et al. (1990) Use of ¹⁰Be in polar ice to trace the 11-year cycle of solar activity. *Nature* 347(6289): 164–166.
- Beer J, et al. (1991b) Seasonal variations in the concentration of ¹⁰Be, Cl⁻, NO₃⁻, SO₄²⁻, H₂O₂, ²¹⁰Pb, ³H, Mn and d¹⁸O in Greenland snow. *Atmospheric Environment* 25A(5/6): 899–904.
- Berggren A-M, Beer J, Possnert G, et al. (2009) A 600-year annual ¹⁰Be record from the NGRIP ice core, Greenland. *Geophysical Research Letters* 36: L11801. <http://dx.doi.org/10.1029/2009GL038004>.
- Björck S, Muscheler R, Kromer B, et al. (2001) High-resolution analyses of an early Holocene climate event may imply decreased solar forcing as an important climate trigger. *Geology* 29(12): 1107–1110.
- Chmeleff J, von Blanckenburg F, Kossert K, and Jakob D (2010) Determination of the ¹⁰Be half-life by multicollector ICP-MS and liquid scintillation counting. *Nuclear Instruments and Methods in Physics Research Section B* 268: 192–199.
- Delaygue G and Bard E (2010) An Antarctic view of beryllium-10 and solar activity for the past millennium. *Climate Dynamics* 36(11–12): 2201–2218. <http://dx.doi.org/10.1007/s00382-010-0795-1>.
- Delmas RJ, Beer J, Synal H-A, Muscheler R, Petit J-R, and Pourchet M (2004) Bomb-test ³⁶Cl measurements in Vostok snow (Antarctica) and the use of ³⁶Cl as a dating tool for deep ice cores. *Tellus* 56B: 492–498.
- Field CV, Schmidt GA, Koch D, and Salyk C (2006) Modeling production and climate-related impacts on ¹⁰Be concentration in ice cores. *Journal of Geophysical Research* 111: D15107. <http://dx.doi.org/10.1029/2005JD006410>.
- Finkel RC and Nishiizumi K (1997) Beryllium 10 concentrations in the Greenland ice sheet project 2 ice core from 3–40 ka. *Journal of Geophysical Research* 102(C12): 26699–26706.
- Fireman EL and Norris TL (1982) Ages and composition of gas trapped in Allan-Hills and Byrd Ice Core. *Earth and Planetary Science Letters* 60(3): 339–350.
- Gleeson LJ and Axford WI (1968) Solar modulation of galactic cosmic rays. *The Astrophysical Journal* 154: 1011–1026.
- Green JR, DeWayne CL, Synal H-A, et al. (2000) Chlorine-36 and cesium-137 in ice-core samples from mid-latitude glacial sites in the Northern Hemisphere. *Nuclear Instruments and Methods in Physics Research Section B* 172: 812–816.
- Groote PM and Stuiver M (1997) Oxygen 18/16 variability in Greenland snow and ice with 10⁻³ to 10⁵-year time resolution. *Journal of Geophysical Research* 102(12): 26455–26470.
- Heikkilä U, Beer J, and Feichter J (2008a) Modeling cosmogenic radionuclides ¹⁰Be and ⁷Be during the Maunder Minimum using the ECHAM5-HAM general circulation model. *Atmospheric Chemistry and Physics* 8: 2797–2809.
- Heikkilä U, Beer J, and Feichter J (2009a) Meridional transport and deposition of atmospheric ¹⁰Be. *Atmospheric Chemistry and Physics* 9: 515–527.
- Heikkilä U, Beer J, Feichter J, et al. (2009b) ³⁶Cl bomb peak: Comparison of modeled and measured data. *Atmospheric Chemistry and Physics* 9: 4145–4156.
- Heikkilä U, Beer J, Jouzel J, Feichter J, and Kubik P (2008b) ¹⁰Be measured in a GRIP snow pit and modeled using the ECHAM5-HAM general circulation model. *Geophysical Research Letters* 35: L05817. <http://dx.doi.org/10.1029/2007GL033067>.
- Horiuchi K, Uchida T, Sakamoto Y, et al. (2008) Ice core record of (10)Be over the past millennium from Dome Fuji, Antarctica: A new proxy record of past solar activity and a powerful tool for stratigraphic dating. *Quaternary Geochronology* 3(3): 253–261.
- Hoyt DV and Schatten KH (1998) Group sunspot numbers: A new solar activity reconstruction. *Solar Physics* 179: 189–219.
- Jenk TM, Szidat S, Bolius D, et al. (2009) A novel radiocarbon dating technique applied to an ice core from the Alps indicating late Pleistocene ages. *Journal of Geophysical Research* 114: D14305. <http://dx.doi.org/10.1029/2009JD011860>.
- Johnsen SJ, Dahl-Jensen D, Dansgaard W, and Gundestrup N (1995) Greenland palaeotemperatures derived from GRIP bore hole temperature and ice core isotope profiles. *Tellus* 47B: 624–629.
- Johnsen SJ, et al. (1997) The d¹⁸O record along the Greenland Ice Core Project deep ice core and the problem of possible Eemian climatic instability. *Journal of Geophysical Research* 102: 26397–26410.
- Korschinek G, et al. (2010) A new value for the half-life of ¹⁰Be by heavy-ion elastic recoil detection and liquid scintillation counting. *Nuclear Instruments and Methods in Physics Research Section B* 268: 187–191.
- Lai D and Jull AJT (1992) Cosmogenic nuclides in ice sheets. *Radiocarbon* 34(2): 227–233.
- Lai D, Jull AJT, Donahue DJ, Burtner D, and Nishiizumi K (1990) Polar ice ablation rates measured using in situ cosmogenic ¹⁴C. *Nature* 346: 350–352.
- Lai D, Jull AJT, Pollard D, and Vacher L (2005) Evidence for large century time-scale changes in solar activity in the past 32 Kyr, based on in-situ cosmogenic ¹⁴C in ice at Summit, Greenland. *Earth and Planetary Science Letters* 234: 335–349.
- Lal D and Peters B (1967) Cosmic ray produced radioactivity on the earth. In: Flüggé S (ed.) *Handbuch für Physik*, pp. 551–612. Berlin: Springer.
- Litherland AE (1980) Ultrasensitive mass spectrometry with accelerators. *Annual Review of Nuclear and Particle Science* 30: 437–473.
- Masarik J and Beer J (1999) Simulation of particle fluxes and cosmogenic nuclide production in the Earth's atmosphere. *Journal of Geophysical Research* 104(D10): 12099–12111.
- Mazaud A, Laj C, and Bender M (1994) A geomagnetic chronology for Antarctic ice accumulation. *Geophysical Research Letters* 21(5): 337–340.
- McHargue LR and Damon PE (1991) The global beryllium 10 cycle. *Reviews of Geophysics* 29(2): 141–158.
- Muscheler R, Beer J, Kubik PW, and Synal H-A (2005) Geomagnetic field intensity during the last 60,000 years based on ¹⁰Be & ³⁶Cl from the Summit ice cores and ¹⁴C. *Quaternary Science Reviews* 24: 1849–1860.

- Muscheler R, Beer J, Wagner G, and Finkel RC (2000) Changes in deep-water formation during the Younger Dryas cold period inferred from a comparison of ^{10}Be and ^{14}C records. *Nature* 408: 567–570.
- Muscheler R, Beer J, Wagner G, et al. (2004) Changes in the carbon cycle during the last deglaciation as indicated by the comparison of ^{10}Be and ^{14}C records. *Earth and Planetary Science Letters* 219: 325–340.
- Muscheler R and Heikkilä U (2011) Constraints on long-term changes in solar activity from the range of variability of cosmogenic radionuclide records. *Astrophysics and Space Sciences Transactions* 7: 355–364. <http://dx.doi.org/10.5194/astra-7-355-2011>.
- Muscheler R, Joos F, Beer J, Mueller SA, Vonmoos M, and Snowball I (2007) Solar activity during the last 1000 yr inferred from radionuclide records. *Quaternary Science Reviews* 26: 82–97. <http://dx.doi.org/10.1016/j.quascirev.2006.07.012>.
- Nesterenko AV and Naidenov VO (2011) Radiocarbon in polar ice: The mechanism by which radiocarbon is preserved in firn grains. *Geomagnetism and Aeronomy* 51(3): 421–428.
- O'Brien SR, Mayewski PA, Meeker LD, Meese DA, Twickler MS, and Whitlow SI (1995) Complexity of Holocene climate as reconstructed from a Greenland ice core. *Science* 270: 1962–1964.
- Pedro JB, Smith AJ, Duldig M, et al. (2009) ^{10}Be concentrations in snow at Law Dome Antarctica following the 29 October 2003 and 20 January 2005 solar cosmic ray events. In: Duldig M (ed.) *Advances in Geosciences: Solar Terrestrial (2007)*, pp. 285–303. London: World Scientific Publishing Company, Imperial College Press.
- Pedro JB, Smith AM, Simon KJ, Van Ommen TD, and Curran MAJ (2011) High-resolution records of the beryllium-10 solar activity proxy in ice from Law Dome East Antarctica: Measurement, reproducibility and principal trends. *Climate of the Past* 7: 707–721. <http://dx.doi.org/10.5194/cp-7-707-2011>.
- Raisbeck GM and Yiou F (2004) Comment on 'Millennium scale sunspot number reconstruction: Evidence for an unusually active sun since the 1940s'. *Physical Review Letters* 92(19): 211101.
- Raisbeck GM, Yiou F, Bourles D, Lorius C, Jouzel J, and Barkov NI (1987) Evidence for two intervals of enhanced ^{10}Be deposition in Antarctic ice during the last glacial period. *Nature* 326: 273–277.
- Raisbeck GM, Yiou F, Cattani O, and Jouzel J (2006) ^{10}Be evidence for the Matuyama–Brunhes geomagnetic reversal in the EPICA Dome C ice core. *Nature* 444: 82–84. <http://dx.doi.org/10.1038/nature05266>.
- Raisbeck GM, Yiou F, Fruneau M, Loiseaux JM, Lieuvain M, and Ravel JC (1981a) Cosmogenic $^{10}\text{Be}/\text{Be}$ as a probe of atmospheric transport processes. *Geophysical Research Letters* 8(9): 1015–1018.
- Raisbeck GM, Yiou F, Fruneau M, et al. (1981b) Cosmogenic ^{10}Be concentrations in Antarctic ice during the past 30,000 years. *Nature* 292: 825–826.
- Raisbeck GM, Yiou F, Jouzel J, and Petit JR (1990) ^{10}Be and $\delta^2\text{H}$ in polar ice cores as a probe of the solar variability's influence on climate. *The Philosophical Transactions of the Royal Society London, A* 330: 65–72.
- Raisbeck GM, Yiou F, Jouzel J, and Stocker TF (2007) Direct north–south synchronization of abrupt climate change record in ice cores using Beryllium 10. *Climate of the Past* 3: 541–547.
- Sigl M, et al. (2009) Towards radiocarbon dating of ice cores. *Journal of Glaciology* 55(194): 985–996.
- Simpson JA (1983) Elemental and isotopic composition of the galactic cosmic rays. *Annual Review of Nuclear and Particle Science* 33: 323–381.
- Smith AM, Levchenko VA, Etheridge DM, et al. (2000) In search of in-situ radiocarbon in Law Dome ice and firn. *Nuclear Instruments and Methods in Physics Research Section B* 172: 610–622.
- Snowball I and Muscheler R (2007) Palaeomagnetic intensity data: An Achilles heel of solar activity reconstructions. *The Holocene* 17(6): 851–859.
- Suess HE (1982) The radiocarbon record in tree rings of the last 8000 years. *Radiocarbon* 22(2): 200–209.
- Svensson A, et al. (2006) The Greenland ice core chronology 2005, 15–42 ka. Part 2: Comparison to other records. *Quaternary Science Reviews* 25: 3258–3267. <http://dx.doi.org/10.1016/j.quascirev.2006.08.003>.
- Thompson LG, Davies ME, Mosley-Thompson E, Lin P-N, Henderson KA, and Mashiotta TA (2005) Tropical ice core records: Evidence for asynchronous glaciation on Milankovitch timescales. *Journal of Quaternary Science* 20(7–8): 723–733.
- Thompson LG, et al. (2002) Kilimanjaro ice core records: Evidence of Holocene climate change in tropical Africa. *Science* 298: 589–593.
- Usoskin IG, Solanki SK, Kovaltsov GA, Beer J, and Kromer B (2006) Solar proton events in cosmogenic isotope data. *Geophysical Research Letters* 33: L08107. <http://dx.doi.org/10.1029/2006GL026059>.
- Usoskin IG, Solanki SK, Schüssler M, Mursula K, and Alanko K (2003) A millennium scale sunspot number reconstruction: Evidence for an unusually active sun since the 1940s. *Physical Review Letters* 91(21): 211101–211104.
- Van Der Kemp WJM, Alderliesten C, Van Der Borg K, et al. (2002) In situ produced C-14 by cosmic ray muons in ablating Antarctic ice. *Tellus Series B: Chemical and Physical Meteorology* 54(2): 186–192.
- Vonmoos M, Beer J, and Muscheler R (2006) Large variations in Holocene solar activity: Constraints from ^{10}Be in the Greenland Ice Core Project ice core. *Journal of Geophysical Research* 111: A10105. <http://dx.doi.org/10.1029/2005JA011500>.
- Wagner G, Beer J, Masarik J, et al. (2001a) Presence of the solar de Vries cycle (205 years) during the last ice age. *Geophysical Research Letters* 28(2): 303–306.
- Wagner G, Laj C, Beer J, et al. (2001b) Reconstruction of the paleoaccumulation rate of central Greenland during the last 75 kyr using the cosmogenic radionuclides ^{36}Cl and ^{10}Be and geomagnetic field intensity data. *Earth and Planetary Science Letters* 193: 515–521.
- Wagner G, Livingstone DM, Masarik J, Muscheler R, and Beer J (2001c) Some results relevant to the discussion of a possible link between cosmic rays and the Earth's climate. *Journal of Geophysical Research* 106: 3381–3387.
- Wagner G, Masarik J, Beer J, et al. (2000) Reconstruction of the geomagnetic field between 20 and 60 kyr BP from cosmogenic radionuclides in the GRIP ice core. *Nuclear Instruments and Methods in Physics Research B* 172: 597–604.
- Wieler R, Beer J, and Leya I (2011) The galactic cosmic ray intensity over the past 10^6 – 10^9 years as recorded by cosmogenic nuclides in meteorites and terrestrial samples. *Space Science Reviews* pp. 1–13. <http://dx.doi.org/10.1007/s11214-011-9769-9>.
- Willerslev E, et al. (2007) Ancient biomolecules from deep ice cores reveal a forested southern Greenland. *Science* 317: 111–114.
- Wilson AT and Donahue DJ (1992) AMS radiocarbon dating of ice: Validity of the technique and the problem of cosmogenic in-situ production in polar ice cores. *Radiocarbon* 34(3): 431–435.
- Yiou F, et al. (1997) Beryllium 10 in the Greenland Ice Core Project ice core at Summit, Greenland. *Journal of Geophysical Research* 102(C12): 26783–26794.

Studies of Firn Air

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Introduction

Firn is the intermediate stage between snow and glacial ice, which constitutes the upper 40–120 m of the accumulation zone of ice sheets. Within the firn a vast network of interconnected pores exists, which exchanges air with the overlying atmosphere. At the bottom of the firn column, air is permanently trapped within the ice matrix, storing an atmospheric sample from the time of pore closure. In this way, air bubbles in glacial ice have been used to reconstruct changes in atmospheric composition up to 800 ka back in time (Louergue et al., 2008; Lüthi et al., 2008; Petit et al., 1999).

If one thinks of glacial ice as a flask for storing samples of ancient atmosphere, then the firn is the valve that gradually closes the flask. But far from an ideal valve, the firn alters the composition of air as it travels downward, through, for example, gravitational separation (Craig et al., 1988; Schwander, 1989), diffusive smoothing (Spahni et al., 2003), and thermal fractionation (Severinghaus et al., 2001). An understanding of firn air transport processes is therefore essential for interpretation of ice core gas records. Continued exchange with the atmosphere furthermore keeps firn air at the firn–ice transition younger than the surrounding ice, resulting in an age difference (Δ age) between glacial ice and the air bubbles it contains (Schwander and Stauffer, 1984).

The firn itself also acts as an archive of old air, preserving a continuous record of atmospheric composition up to a century back in time (Kaspers et al., 2004). Sampling of firn air allows reconstruction of the recent atmospheric history of many trace gas species and their isotopic composition.

Firn Air Sampling

Firn Air Sampling Device

The primary means of studying firn transport is by sampling of firn air (Schwander et al., 1993). The principle is illustrated in Figure 1(a). The firn air sampling device (FASD) is lowered into a borehole drilled to the desired sampling depth. By inflating the bladder, firn air is isolated from the overlying atmosphere. The borehole is then purged from contamination by pumping on the purge and sampling lines. CO₂-mixing ratios are typically monitored on the air flow to assess whether the hole has been sufficiently purged. A baffle plate (Figure 1(b)) is often used to separate the purge and sampling lines. The purge flow has a higher risk of being contaminated by air leaking past the bladder, and is discarded. Air from the sampling line is normally collected in flasks, which are shipped to the laboratory for analysis. Field deployment of FASD and ice drill is shown in Figure 2.

In the deep firn, density layering causes air to flow preferentially in the horizontal plane during sampling. Sturrock et al. (2002) observed that the mixing ratios of CH₄, CO₂, N₂O, H₂,

and CO do not change systematically with time or air volume extracted during sampling at each level. They interpret this as evidence that the air from each sampling depth originates from a narrow horizontal firn layer, rather than from regions significantly above or below the FASD inlet. In shallow strata with higher permeability, the collection region is presumably closer to a sphere or ellipsoid.

At several sites, air sampled from the deepest firn strata was found to be slightly depleted in heavy gases and isotopologues such as δ Kr/N₂, δ Xe/N₂, δ^{15} N₂, and δ^{40} Ar (Severinghaus and Battle, 2006). The authors interpret this as a mass-dependent collection artifact caused by the large pressure gradients that the sampling procedure induces in these low-permeability layers.

Overview of Firn Sampling Sites

Firn air has been sampled from sites on both hemispheres, covering a wide range of climatic conditions. Table 1 gives an overview of sampling sites. The Law Dome DE08-2 site is characterized by a very high accumulation rate, which allows for reconstructing atmospheric variations with high temporal resolution. At sites on the Antarctic plateau (Dome C, Dome Fuji, Megadunes, Vostok), on the other hand, the oldest air can be found.

Physical Firn Structure and Transport Properties

Firn Density and Porosity

Polar firn is a porous medium that is densified gradually under the weight of overlying precipitation, until its density ρ approaches the pure ice density ρ_{ice} of around 920 kg m⁻³. The interstitial space between ice crystals is referred to as the porosity s , which is defined as the volume fraction not occupied by ice, or $s = 1 - \rho/\rho_{\text{ice}}$. It can be divided into open (s_{op}) and closed (s_{cl}) porosity; the former refers to pores that are still connected to the overlying atmosphere, whereas the latter refers to pores that have already been closed-off. Owing to firn densification, the porosity decreases with depth and air is gradually occluded in bubbles. At Siple station, Schwander and Stauffer (1984) found that 80% of the bubble volume is formed in the density range $795 < \rho < 830$ kg m⁻³. The average density $\bar{\rho}_{\text{co}}$ at which the bubbles are closed-off is found in this interval. This mean close-off density was found to be independent of accumulation rate, and is mostly a function of the site temperature (Martinier et al., 1992, 1994):

$$\bar{\rho}_{\text{co}} = \left(\frac{1}{\rho_{\text{ice}}} + 6.95 \times 10^{-7} T - 4.3 \times 10^{-5} \right)^{-1} \quad [1]$$

where T is the mean annual site temperature in K. The mean close-off density in eqn [1] was determined by total air content measurements in mature ice. It is defined through the porosity,

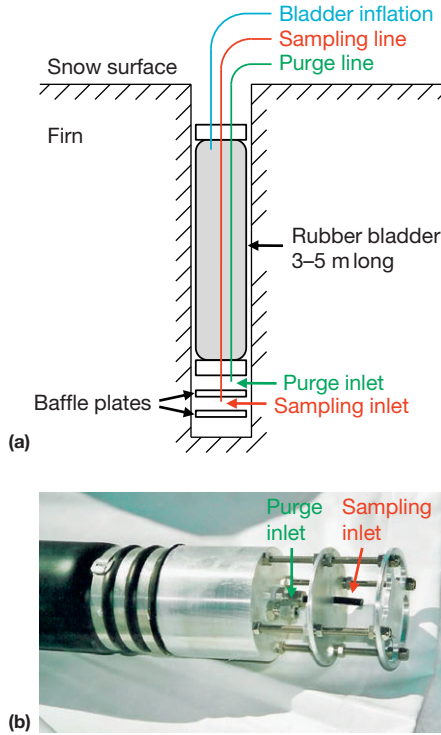


Figure 1 (a) The principle of firm air sampling (not to scale). The FASD is suspended in the borehole from a steel cable. (b) Sampling head of the University of Bern FASD. Photograph courtesy of Jakob Schwander.

which, at the atmospheric pressure of the site, matches the measured air content. Apart from temperature, the close-off density (and thereby the total air content) is also influenced by wind strength (Martinerie et al., 1994) and insolation (Raynaud et al., 2007).

A second useful definition is the full close-off depth z_{cod} (with corresponding ρ_{cod}), which is the depth where all air is occluded ($s_{\text{op}}=0$). The full close-off is deeper than the mean close-off depth of eqn [1]. The deepest point at which firm air can be sampled in the field is just above z_{cod} .

Several s_{cl} parameterizations exist. Schwander (1989) gives the following empirical relation based on measurements of bubble volume in ice samples from Siple station, Antarctica:

$$s_{\text{cl}} = \begin{cases} s \exp[75(\rho/\rho_{\text{cod}} - 1)] & \text{for } \rho \leq \rho_{\text{cod}} \\ s & \text{for } \rho > \rho_{\text{cod}} \end{cases} \quad [2]$$

It must be noted that measurements of bubble volume tend to underestimate closed porosity, as pores can be reopened by sample cutting. Severinghaus and Battle (2006) adapted eqn [2] for Summit, Greenland, to have a more gradual close-off.

An alternative parameterization is given by Goujon et al. (2003), based on density and total air content measurements from three Greenland and Antarctic sites:

$$s_{\text{cl}} = 0.37s \left(\frac{s}{\bar{s}_{\text{co}}} \right)^{-7.6} \quad [3]$$

where $\bar{s}_{\text{co}}$ is the mean close-off porosity $\bar{s}_{\text{co}} = 1 - \rho_{\text{co}}/\rho_{\text{ice}}$. Equation [3] is designed to be consistent with eqn [1] and indicates that 37% of the porosity is closed for $\bar{\rho}_{\text{co}}$. Both parameterizations are shown in Figure 3(a).

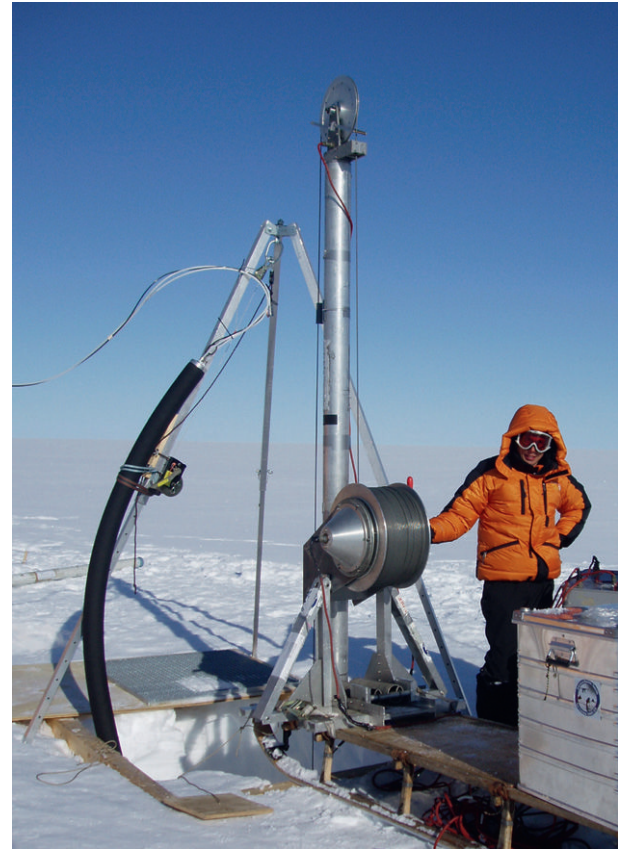


Figure 2 Firm air sampling at North Greenland Eemian Ice Drilling (NEEM) site, July 2009. Vasilli Petrenko with the University of Copenhagen shallow electromechanical drill (3 inch core) and the CSIRO FASD bladder assembly. Photograph courtesy of Anne Solgaard.

Open Porosity Air Flow and Transport Properties

In the firm–ice transition, air bubbles are trapped and advected downward with the ice. Conservation of mass demands airflow in the open pores to replace the air thus removed. Assuming the firm to be in steady state with respect to accumulation and densification, the downward velocity of air in the open pores can be derived from the porosity ratio and mass conservation (Buizert et al., 2012):

$$w_{\text{air}} = \frac{A\rho_{\text{ice}}}{s_{\text{op}}^*P_0} \left(\frac{s_{\text{cl}}(z_{\text{COD}})P_{\text{cl}}(z_{\text{COD}})}{\rho_{\text{COD}}} - \frac{s_{\text{cl}}(z)P_{\text{cl}}(z)}{\rho(z)} \right) \quad [4]$$

where A is the accumulation rate in m year^{-1} ice equivalent, $s_{\text{op}}^* = s_{\text{op}} \exp(M_{\text{air}}gz/RT)$, is the effective open porosity, P_0 the barometric pressure, $P_{\text{cl}}(z)$ the enhanced pressure in closed bubbles due to firm compaction, M_{air} the molar mass of air in kg mol^{-1} , g the gravitational acceleration, and R the molar gas constant. The result is plotted in Figure 3(b), together with the downward velocity of the ice layers $w_{\text{ice}} = A\rho_{\text{ice}}/\rho$. The ice layers descend faster than the air does, that is, the air in the open porosity is moving upward relative to the ice matrix. This back flow is due to compression of open pores by the firm densification process (Rommelaere et al., 1997).

Pore geometry and connectivity dictate the firm air transport properties. Gas diffusivity and permeability in firm samples

Table 1 Overview of firn air sampling sites and characteristics

Site	Location		Altitude (m a.s.l.)	P (hPa)	T (°C)	A (cm ice year ^{−1})	CZ depth (m)	z _{CO2} (m)	Year(s) sampled	Main references
Northern Hemisphere										
Devon Isl.	75.3° N	82.1° W	1929	792 ^a	−23	30	~0 ^b	59	1998	FIRETRACC, Clark et al. (2007)
NEEM	77.4° N	51.1° W	2484	745	−28.9	22	4	78	2008, 2009	Buizert et al. (2012)
NGRIP	75.1° N	42.3° W	2917	691	−31.1	19	1–2 ^c	78	2001 ^d	CRYOSTAT, Kawamura et al. (2006)
Summit	72.6° N	38.4° W	3214	665	−31.4	23	~0 ^b	80	1989, 2006	Schwander et al. (1993), Witrant et al. (2011)
Southern Hemisphere										
Berkner Isl.	79.6° S	45.7° W	900	895 ^e	−26	13	<2 ^c	64	2003	CRYOSTAT
DE08-2	66.7° S	113.2° E	1250	850	−19	120	0 ^b	85	1993	Etheridge et al. (1996)
DML ^f	77.0° S	10.5° W	2176	757 ^e	−39	7	<5 ^c	74	1998	FIRETRACC
Dome C	75.1° S	123.4° E	3233	658 ^e	−54	^c 2.7–3.2	2 ^c	100	1999	FIRETRACC
Dome Fuji	77.3° S	39.7° E	3810	600	−57.3	^c 2.3–2.8	9	104	1998	Kawamura et al. (2006)
DSSW20K	66.8° S	112.6° E	1200	850	−21	16	4	52	1998	Trudinger (2001)
H72	69.2° S	41.1° E	1241	857	−20.3	33	2	65	1998	Kawamura et al. (2006)
Megadunes	80.8° S	124.5° E	2880	677	−49	~0 ^g	23	68	2004	Severinghaus et al. (2010)
Siple Dome	81.7° S	148.8° W	620	940	−25.4	13	<2 ^b	57	1996, 1998	Severinghaus et al. (2001)
South Pole	90.0° S		2840	681	−51	8	<2 ^b	123	1995–2008 ^h	Severinghaus et al. (2001)
Vostok	78.5° S	106.8° E	3471	632	−56	2.4	13 ^b	100	1996	Fabre et al. (2000)
WAIS-D	79.5° S	112.1° W	1766	780	−31	22	3	76.5	2005	Battle et al. (2011)
YM85	71.6° S	40.6° E	2246	730	−34	17	14 ⁱ	68	2002	Kawamura et al. (2006)

Unless indicated otherwise, values can be found in the main references or references therein (last column).

^aHuber et al. (2006).

^bSeveringhaus et al. (2010).

^cLandais et al. (2006).

^dThere were two separate NGRIP firn air campaigns in 2001 (J. Schwander, personal communication, 2011).

^eCalculated from the altitude using the pressure–altitude relationship over Antarctica from Stone (2000).

^fAlso referred to as BAS depot (Landais et al., 2006). Note that the location differs from that of the EDM1 ice core.

^gThe long-term average accumulation estimate is 2.5 cm ice year⁻¹. Sampling was done during an accumulation hiatus.

^hFirn air was sampled in 1995, 1998, 2001, and 2008.

ⁱHigh wind speeds at the site cause the unusually thick CZ.

have been determined using direct measurements (Albert et al., 2000; Fabre et al., 2000; Schwander et al., 1988) and by modeling of gas transport in reconstructed pore geometries (Courville et al., 2011; Freitag et al., 2002). However, gas diffusivities determined on individual firn samples do not represent the transport properties of the firn as a whole (Fabre et al., 2000), showing that the lateral dimensions of the diffusive path exceed that of a typical firn core sample.

Density Layering

Polar firn is a layered medium that exhibits large density variations around the mean caused by seasonal changes in climatic conditions and precipitation density, as well as wind and insolation features that are preserved in the densification process. As high-density layers reach the close-off density first, they can form impermeable layers that inhibit vertical gas transport (Martinerie et al., 1992). Such sealing layers are often invoked to explain the presence of a non-diffusive zone, or lock-in zone

(LIZ) just above the bubble close-off depth (Battle et al., 1996; Landais et al., 2006). Firn air can be pumped from the LIZ, meaning a large (laterally) connected open porosity still exists in the low-density layers.

The high-density sealing layers have been linked to winter precipitation through measurements of water stable isotopes (Martinerie et al., 1992). Recent studies of firn microstructure indicate a density cross-over at $\rho \sim 600\text{--}650\text{ kg m}^{-3}$, suggesting that high density LIZ layers originate as low density layers at the surface (Freitag et al., 2004). Boreholes separated by as little as 65 m were found to have different firn air transport properties caused by lateral variability in the firn stratigraphy (Buizert et al., 2012).

Although the dense layers impede vertical transport, they do not completely seal off the air below. The air content implied by such fully sealing layers is incompatible with measurements in mature ice. Furthermore, at many sites, firn air transport models require finite gas diffusivity in the LIZ to reproduce the measured mixing ratios of trace gases

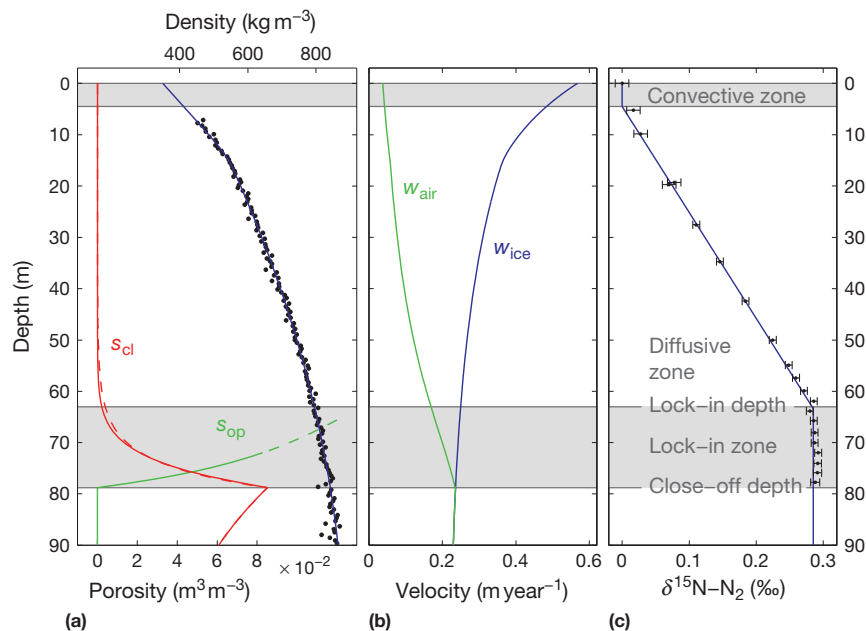


Figure 3 Firn characteristics at NEEM. (a) Firn density and porosity using the parameterizations of Schwander (1989) (solid line) and Goujon et al. (2003) (dashed line). (b) Downward velocity of firn layers (w_{ice}) and of air in the open porosity (w_{air}) calculated using eqn [4]. (c) Zonal division based on gravitational enrichment of $\delta^{15}\text{N}-\text{N}_2$. Data are corrected for thermal fractionation.

(Severinghaus et al., 2010; Witrant et al., 2011). The back flow of air due to pore compression can be facilitated by (micro) cracks and channels in the dense layers.

Trace Gas Transport in the Firn

Zonal Description of Transport

The firn column is commonly divided into three zones, based on the gravitational isotopic enrichment of molecular nitrogen with depth (Sowers et al., 1992). The zones correspond to different modes of firn air transport, as depicted in Figure 3(c). The convective zone (CZ) refers to the upper few meters of the firn column, which are vigorously ventilated. The air in the CZ is essentially of current atmospheric composition, and consequently it has the same $\delta^{15}\text{N}_2$ as the atmosphere (which is zero by definition). Below the CZ we find the diffusive zone (DZ), where mass transfer is dominated by molecular diffusion. In diffusive equilibrium, gravitational separation gives enrichment in heavy isotopes with depth, as is evident from the presented $\delta^{15}\text{N}_2$ measurements. The lock-in depth is defined as the depth at which gravitational enrichment stops. Continued pore compaction leads to a decreasing effective diffusivity with depth; the lock-in depth corresponds to the point where diffusivity becomes effectively zero. At this depth the air gets isolated from the atmosphere, and therefore the lock-in depth determines the Δ_{age} in mature ice. Below the lock-in depth, we find the LIZ or non-diffusive zone, where advection with the ice matrix dominates the transport. Since advection does not discriminate between isotopologues, gravitational separation does not occur in the LIZ. The existence of the LIZ has been linked to the formation of high density sealing layers that inhibit vertical diffusion (see Section ‘Density Layering’).

Overview of Mass Transfer Mechanisms

Diffusion

Spatial variations in gas concentration or partial pressure lead to isobaric mass transfer along the gradient, from regions of higher to lower concentration. In porous media, such as polar firn or unsaturated soil, two types of diffusion should be considered. Knudsen diffusion occurs when the pore diameter is sufficiently small, or pressure sufficiently low, and molecules collide predominately with the walls rather than with other gas molecules. The second type is molecular diffusion, which represents the opposite regime of predominantly intermolecular collisions. In polar firn, the molecular free mean path of $\lambda \sim 100$ nm is about four orders of magnitude smaller than typical pore sizes of ~ 1 mm (Freitag et al., 2004; Kipfstuhl et al., 2009). Consequently, firn air diffusion is of the molecular type. In soils, particle and pore sizes are generally much smaller and Knudsen diffusion cannot always be neglected. Fick’s law is commonly used to describe molecular diffusion in firn. The validity of this approach has been questioned in the case of soil diffusion (Thorstenson and Pollock, 1989), in particular for studies of soil respiration (Freijer and Leffelaar, 1996). When modeling trace gas transport, Fick’s law provides a good approximation; it is not suited to describe transport of main components of air such as O_2 .

The effective diffusivity of gas X in the open porosity is given by (Dullien, 1975):

$$D_X = s_{op} \frac{D_X^0}{\tau} = s_{op} \gamma_X \frac{D_{\text{CO}_2}^0}{\tau} \quad [5]$$

where D_X^0 is the free-air molecular diffusion coefficient for gas X at the pressure and temperature of the site, τ is the tortuosity of the pore geometry and $\gamma_X = D_X/D_{\text{CO}_2}$ is the diffusivity of gas

X relative to that of CO_2 . Recommended values for γ_X are given in the supplement of Buizert et al. (2012). Although parameterizations of τ as a function of s_{op} do exist (Schwander et al., 1988), diffusivities measured on firn samples do not represent the diffusive behavior of the entire firn (Fabre et al., 2000). Best results are obtained when $\tau(z)$ is reconstructed site-specifically using measurements of reference tracers (see Section ‘Reconstructing the Effective Diffusivity’).

Diffusive transport leads to a separation of the gas mixture by gravity, resulting in a gradual enrichment in both heavy molecules and isotopologues with depth. In diffusive equilibrium, the enrichment relative to the atmosphere is given in delta notation by:

$$\delta_{\text{grav}} = \left[\exp\left(\frac{gz\Delta M}{RT}\right) - 1 \right] \times 10^3\text{‰} \cong \frac{gz}{RT} \Delta M \times 10^3\text{‰} \quad [6]$$

where ΔM can be the molar mass deviation from air in kg mol^{-1} ($\Delta M = M_X - M_{\text{air}}$) or the mass difference between two isotopologues when considering isotopic enrichment. Macroscopic transport processes, such as convection and advection, can locally drive the air column out of diffusive equilibrium, reducing the gravitational separation below the equilibrium slope given by eqn [6].

Temperature gradients in the firn also lead to diffusive separation of the gas mixture. Thermal diffusion results in seasonal isotope fractionation in the firn in response to the seasonal temperature cycle (Severinghaus et al., 2001; Weiler et al., 2009), as well as isotope excursions in the ice core record during abrupt climate changes (see Section ‘Thermal Fractionation and Gas Thermometry’). Thermal diffusion can be described as the tendency of heavier molecules to preferentially diffuse toward colder regions. This is illustrated in Figure 4 for firn air sampled at Siple Dome, Antarctica. The seasonal

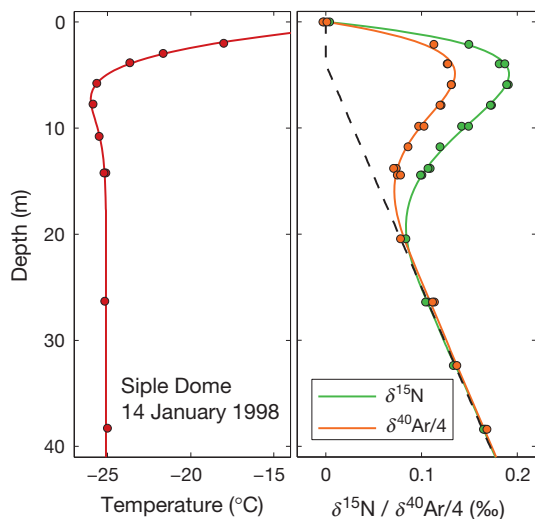


Figure 4 Thermal fractionation by seasonal temperature gradients in the upper firn. Data are from Severinghaus et al. (2001); temperature measurements and firn air sampling were both done on 14 January 1998 at Siple Dome, Antarctica. Solid curves are a guide to the eye; the dashed line gives the equilibrium gravitational slope of eqn [6]. The $\delta^{40}\text{Ar}/^{36}\text{Ar}$ data have been divided by 4 to show the gravitational enrichment per unit mass difference.

temperature minimum around 5 m depth leads to a local enrichment in heavy isotopologues of N_2 and Ar.

Convection

In the firn air literature, convection is used as a generic term for processes that ventilate the upper firn, such as wind pumping (Colbeck, 1989), pressure variations due to seasonality and synoptic scale weather systems (Waddington and Cunningham, 1996), and convective flow driven by (seasonal) temperature gradients. Ventilation affects ice core and firn gas records, as well as postdepositional processes of volatile chemical species. The CZ is typically 1–15 m deep (Kawamura et al., 2006; Landais et al., 2006); an unusually thick CZ of 23 m was observed at the Megadunes site in central Antarctica, caused by a combination of strong winds and an accumulation hiatus giving rise to deep vertical cracks that act as conduits for air exchange (Severinghaus et al., 2010). CZ thicknesses are listed in Table 1. Typically, deep CZs are found at sites with low accumulation rates (e.g., Vostok) or high wind speeds (e.g., YM85). A clear negative correlation between accumulation rate and estimated CZ thickness was found along the EPICA Dome C deep ice core (Dreyfus et al., 2010).

The CZ thickness can be determined empirically using the barometric line-fitting method (Bender et al., 1994), in which the barometric slope is fit to isotopic data in the lower DZ, where the gravitational separation is undisturbed by temperature gradients and ventilation. The intercept with the depth axis gives the CZ thickness, as shown in Figures 3(c) and 4.

In numerical modeling of firn air transport, convection can be included in several ways. The most rudimentary way is to use the bottom of the CZ as the upper boundary and assume the air above to be of atmospheric composition. A more advanced approach is to include an eddy diffusion term that equally affects all gases and isotopologues (Kawamura et al., 2006):

$$D_{\text{eddy}}(z) = D_{\text{eddy}}^0 \exp\left(-\frac{z}{H}\right) \quad [7]$$

where D_{eddy}^0 is the eddy diffusivity at the surface and H is a characteristic depth scale. The eddy diffusion terms prevent gravitational separation by overwhelming molecular diffusion.

Advection

The bulk motion of air in the open porosity (see Section ‘Open Porosity Air Flow and Transport Properties’) gives rise to downward advective mass transfer of trace gases. Firn air models deal with advection in different ways. It can be included by dividing the firn column into boxes of equal air content, and shuffling them downward after a given time interval has elapsed (Schwander et al., 1993). An advantage of this approach is that it minimizes numerical diffusion artifacts. Alternatively, advection can be included as a continuous flux using a calculated air velocity in the open pores (Rommelaere et al., 1997; Sugawara et al., 2003). A third approach is to work in a Lagrangian reference frame that moves downward with descending ice layers (Trudinger et al., 1997). It must be noted that this approach overestimates the advective transport, unless the backflow due to pore compaction is explicitly taken into account.

Dispersive mixing

Equation [4] describes the average air velocity; in reality, a nonuniform velocity distribution exists between air expelled upward by pore compaction and air in isolated pockets, which is advected downward at the velocity of descending ice layers. Furthermore, atmospheric pressure variations and firn densification induce pressure gradients in the firn, which force viscous air flow. All these macroscopic flow patterns lead to dispersive mixing throughout the firn column, although it is overwhelmed by molecular diffusion in the DZ. Soil air studies have shown that viscous flow from atmospheric pressure fluctuations can induce mixing down to tens of meters, and that the dispersive flux can exceed the diffusive one (Massmann and Farrier, 1992). Firn air models do not explicitly include these flow patterns, and dispersive mixing is mostly neglected. Some studies include dispersion phenomenologically through an eddy diffusion term in the deepest firn (Buizert et al., 2012; Severinghaus et al., 2010).

Dating of Firn Air and Reconstructing Recent Atmospheric Composition

Reconstructing the Effective Diffusivity

Numerical firn air transport models are essential for dating and interpretation of firn air records. The largest uncertainty in the transport description is how the effective diffusivity of gases changes with depth (eqn [5]). Best results are obtained when the diffusivity profile is reconstructed through an inverse method (Rommelaere et al., 1997; Trudinger et al., 1997). The method consists of forcing a firn air model with a reference tracer of known atmospheric history (usually CO_2 or CH_4) and optimizing the fit to the measurements through adjusting the diffusivity profile. The procedure is illustrated in Figure 5(a) for the South Pole (SPO) and Northern Greenland Eemian Ice Drilling (NEEM) sites. The dashed line shows the modeled CO_2 -mixing ratio using the tortuosity parameterization of (Schwander et al., 1988). The solid lines show the fit to the data after the optimization procedure.

Several studies have used multiple reference tracers simultaneously to improve the diffusivity reconstruction (Buizert et al., 2012; Trudinger et al., 2002).

Age Distribution

Firn air does not have a single age, but is a mixture of air with different ages. Figure 5(b) shows for SPO how the CO_2 age distribution $G(z,t)$ progresses with depth. The mean age $\Gamma(z)$ of the gas mixture (first moment of the distribution) increases with depth, as the gases need time to be transported into the firn. Diffusion broadens the distribution with depth. Figure 5(c) compares age distributions at the bottom of the firn column for NEEM and SPO. The latter has a longer firn column and low accumulation rate, resulting in older air at the bottom of the firn as well as a wider distribution.

The gas age distribution, which is different for each gas, provides a complete description of the firn transport properties. Age distributions can be calculated by forcing a numerical firn air transport model at the atmospheric boundary with a

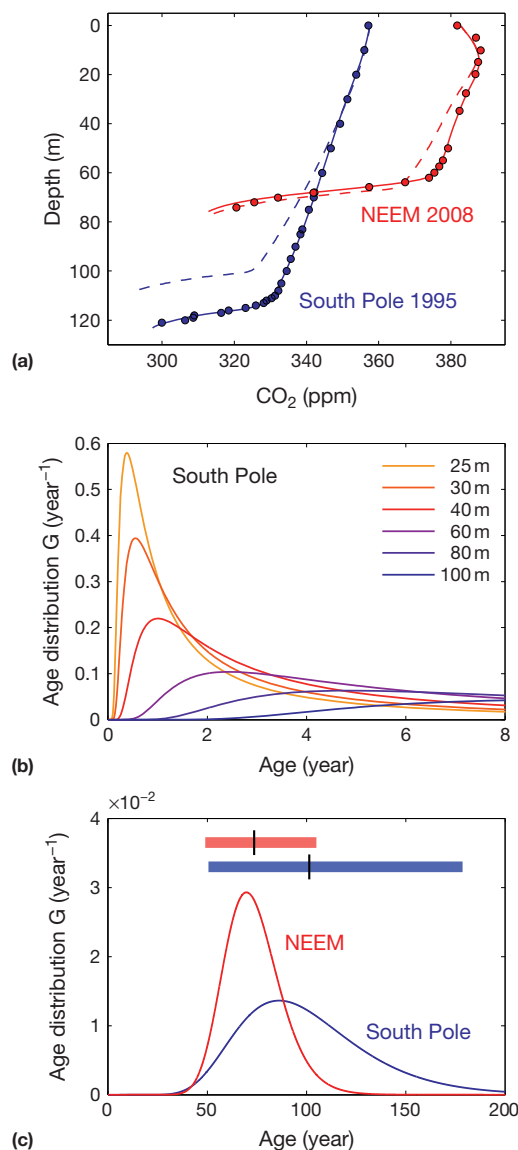


Figure 5 (a) Tuning procedure for transport modeling. Dashed lines show model solutions using the diffusivity parameterization of Schwander et al. (1988); solid lines show the optimized fit using CO_2 as a reference tracer. The Northern Hemisphere CO_2 seasonal cycle is clearly visible in the upper firn at NEEM. (b) CO_2 age distributions for selected depths at South Pole. (c) Comparison of NEEM and South Pole age distributions at the close-off depth (78 and 122 m, respectively). Top bars show the 95% age interval and mean CO_2 age.

rectangular pulse of short duration. Alternatively, they have been inferred experimentally by measuring the spread of the $\Delta^{14}\text{C}$ - CO_2 'bomb spike' caused by atmospheric testing of thermonuclear weapons, which peaks in the early to mid-1960s (Levchenko et al., 1996).

The gradual nature of the trapping process further broadens the age distribution. Air bubbles found at a single depth have not all been formed at the same time and represent a mixture of ages. Within the LIZ, where the majority of bubbles form, gas diffusivities are small and air in the open pores is advected downward at nearly the same velocity as the ice (Figure 3(b)).

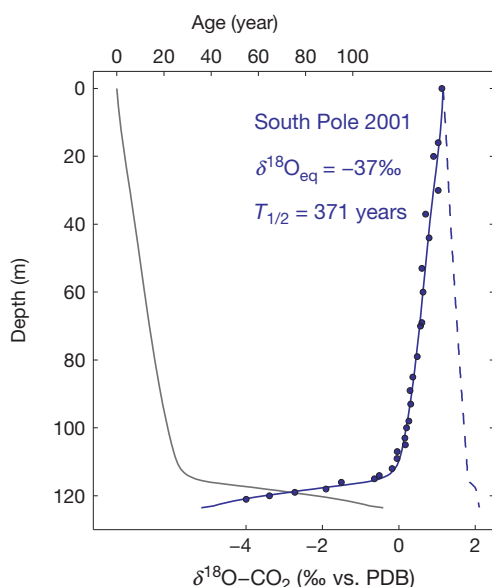


Figure 6 Oxygen exchange between CO_2 and firn results in ^{18}O depletion with depth (lower axis). The dashed line shows the expected $\delta^{18}\text{O}$ signal in the absence of exchange; the solid line gives the best fit to the data based on the modeled ages (gray curve, upper axis) and the parameters listed. The reaction half-time $T_{1/2}$ is the only parameter that is tuned in the data fitting.

Consequently, the broadening due to trapping is small compared to the diffusive broadening described above (Blunier and Schwander, 2000).

Firn Air Dating with $\delta^{18}\text{O}$ of CO_2

An alternative firn air-dating method based on the isotopic composition of CO_2 has been presented by Clark et al. (2007). Atmospheric $\delta^{18}\text{O}-\text{CO}_2$ is relatively stable over time (Allison and Francey, 2007). Within the firn, atomic oxygen exchange with water molecules gradually alters the oxygen isotopic composition of CO_2 , until equilibrium with the strongly depleted $\delta^{18}\text{O}$ of glacial ice is reached (Siegenthaler et al., 1988). This principle is illustrated in Figure 6 for SPO firn.

Depending on the site temperature, the time required for the magnitude of the exchange to reach half its final magnitude ($T_{1/2}$) ranges from tens to hundreds of years depending on temperature (Assonov et al., 2005). Knowing the rate of O-exchange, the amount of time the CO_2 has been in contact with the ice (i.e., the age) can be calculated. The main advantage of the technique is its applicability to warm sites with summer melt layers that complicate conventional dating. The drawback is that $T_{1/2}$ and the equilibrium fractionation must be well known. Clark et al. (2007) solve this issue by using the $\Delta^{14}\text{C}-\text{CO}_2$ signal, which peaks around 1963, as an absolute age marker to calibrate $T_{1/2}$.

Reconstructing Recent Atmospheric Composition

The firn air itself preserves a continuous record of atmospheric composition which can be sampled to reconstruct trace gas-mixing ratios and isotopes for the recent atmosphere. Each firn

air sample represents a mixture of gas ages, and consequently mixing ratios measured in the firn cannot be directly mapped onto a timescale.

Once the firn air transport is well characterized using reference tracers, the forward diffusive problem is well described by the modeled age distribution alone. Subsequently, the most probable atmospheric history of a trace gas of interest can be reconstructed from firn air measurements through an inverse method (Rommelaere et al., 1997; Sugawara et al., 2003; Trudinger et al., 2002). For a more rudimentary dating, mixing ratios of a trace gas of interest can also be directly compared to reference gas measurements at identical sampling depths (Montzka et al., 2004).

Firn air has thus been used to reconstruct the recent atmospheric mixing ratios and/or isotopic composition of, for example, carbon dioxide (e.g., Etheridge et al., 1996), methane (e.g., Braunlich et al., 2001), ethane (Aydin et al., 2011), nitrous oxide (e.g., Ishijima et al., 2007), carbon monoxide (Assonov et al., 2007), several halocarbons (e.g., Butler et al., 1999; Sturges et al., 2001a), and carbonyl sulfide (e.g., Sturges et al., 2001b).

Effects of Firn Air Transport on Ice Core Records

The firn layer complicates the interpretation of ice core records as it alters the atmospheric composition prior to bubble closure. At the same time, the peculiarities of firn air transport give rise to additional signals that can be used for dating and climate reconstruction. In this section, we discuss the consequences of firn air transport for the interpretation of ice core records.

Delta Age and Gravitational Fractionation

Continued exchange with the atmosphere keeps firn air at the lock-in depth younger than the surrounding ice, resulting in an age difference Δage between glacial ice and the air bubbles it contains (Schwander and Stauffer, 1984). This is arguably the most important artifact of the firn layer. Accurate Δage estimates are pivotal in investigating the relative timing of abrupt changes in temperature and greenhouse gas concentrations (Barnola et al., 1991; Fischer et al., 1999), interhemispheric synchronization of ice core records (Blunier and Brook, 2001; Morgan et al., 2002), and deriving consistent ice core time-scales (Lemieux-Dudon et al., 2010). Figure 7 shows Δage estimates as a function of accumulation rate and temperature, assuming stationary conditions. The figure also indicates modern-day conditions of several firn-sampling and ice-coring sites.

The Δage is fixed at the lock-in depth, where gas diffusion effectively ceases and the stagnant air is advected downward with the ice. The modern-day Δage can be estimated accurately for sites that are well characterized using reference tracers; the largest uncertainty being the fraction of bubbles formed above the lock-in depth. To estimate Δage back in time, a combination of several methods can be used. First, past changes in lock-in depth can be estimated from accumulation and temperature variations, using firn densification models (Barnola et al., 1991; Goujon et al., 2003). Second, climatic changes recorded

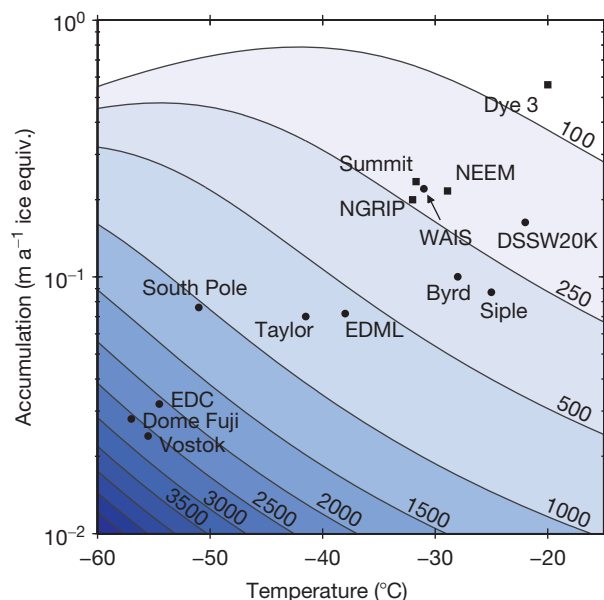


Figure 7 Contour plot of Δage estimates as a function of temperature and accumulation rate. Indicated are modern-day conditions of several ice-coring sites. Density profiles were calculated using the steady state Herron-Langway model (Herron and Langway, 1980). Following Blunier and Schwander (2000), we assume gas lock-in at a density of 14 kg m^{-3} below the mean close-off density of eqn [1]. The age of the air was parameterized following Schwander et al. (1997).

in both the gas record and in precipitation stable isotopes allow determination of Δdepth (Figure 8), which in turn constrains Δage (Caillon et al., 2001, 2003; Severinghaus et al., 1998). Third, records of, for example, CH_4 or $\delta^{18}\text{O}-\text{O}_2$ from other well-dated ice cores, can be used for synchronization of chronologies (Capron et al., 2010; Loulergue et al., 2007).

The lock-in depth calculated by densification models can be compared to measurements of the gravitational enrichment in $\delta^{15}\text{N}-\text{N}_2$, which records the diffusive column length. Note that the diffusive column length does not directly give the lock-in depth, as the thickness of the CZ is unknown. Models predict a deeper lock-in depth during colder glacial conditions, implying a larger $\delta^{15}\text{N}-\text{N}_2$. This is indeed observed in Greenland ice cores (Schwander et al., 1997). Antarctic ice cores (with the exception of the coastal Byrd and WAIS-Divide sites) show a pronounced model-data mismatch with lower $\delta^{15}\text{N}-\text{N}_2$ (i.e., shorter diffusive column) during glacial conditions (Landais et al., 2006). An increased CZ thickness is often invoked to explain the mismatch, where the Megadunes site, with an unusually thick CZ of $>23 \text{ m}$, could provide a modern analog for central Antarctic glacial conditions (Severinghaus et al., 2010).

Measurements of gravitational enrichment in $\delta^{15}\text{N}-\text{N}_2$ are also used to correct gas records for the effect of gravity.

Thermal Fractionation and Gas Thermometry

Thermal diffusion causes isotopic fractionation in the presence of temperature gradients (see Section ‘Diffusion’). Seasonal temperature variations and the associated isotope effect occur

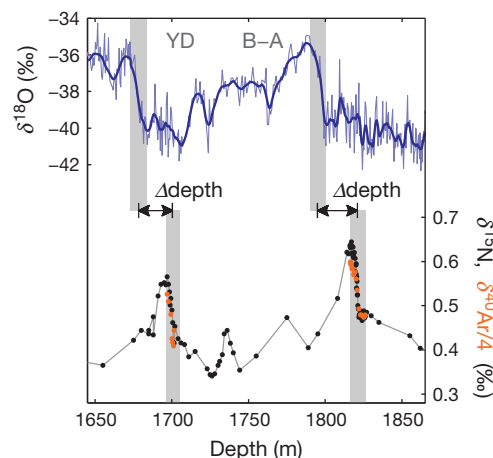


Figure 8 GISP2 oxygen isotopes of precipitation with 200-year running average (upper curve, left axis) and transient thermal fractionation signals in N_2 and Ar caused by rapid warming over Greenland (lower data, right axis). Data from Severinghaus et al. (1998) and Severinghaus and Brook (1999). The depth difference between the observed warming in the ice and gas records constrains Δage .

only in the upper firn (Figure 4). They could potentially affect the deep firn (and thereby the ice core record) through seasonality in transport properties, such as convective mixing and temperature-dependent diffusion coefficients. No evidence for such rectifier effects was found (Severinghaus et al., 2001); seasonal thermal fractionation should have little or no impact on the ice core record.

Climate-induced changes in mean annual surface temperature, on the other hand, can cause isotopic signals in the ice core record if the temperature change is sufficiently large and rapid. This principle is illustrated in Figure 8 for the GISP2 core in central Greenland. Abrupt warming events during the last deglaciation are accompanied by positive $\delta^{15}\text{N}$ and $\delta^{40}\text{Ar}$ excursions as the deep, colder firn gets enriched in heavy isotopes relative to the warmer surface. As the temperature sensitivity of $\delta^{15}\text{N}$ and $\delta^{40}\text{Ar}/4$ is different, the signals can be used in combination to infer the amplitude of the temperature change (Landais et al., 2004; Leuenberger et al., 1999; Severinghaus and Brook, 1999). The excursions furthermore provide clear time markers in the gas record that can be used to determine Δage more accurately than densification models allow. The method does not work well for Antarctica, where temperature changes are more gradual.

Molecular Size Fractionation at Close-Off

Within the LIZ, a systematic enrichment is observed for gas species with a small molecular diameter, such as He , Ar , Ne , and O_2 (Huber et al., 2006; Severinghaus and Battle, 2006). This has been explained by the preferential exclusion of these species from closing bubbles, causing them to accumulate in the open porosity. Closed bubbles are pressurized through continued firn compaction, which increases gas partial pressures in the bubbles relative to the open porosity. This gradient drives selective permeation of gas through the ice lattice. Huber et al. (2006) find a strong dependence of the fractionation magnitude on the collision diameter of the molecule, suggesting

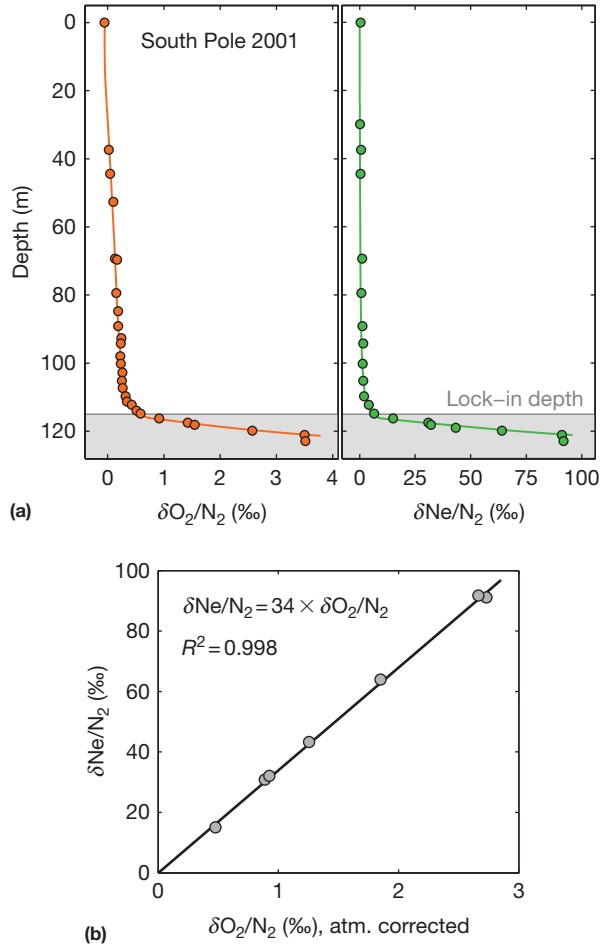


Figure 9 Molecular size fractionation during bubble close-off, data and model output are from [Severinghaus and Battle \(2006\)](#). (a) Enrichment in O₂ and Ne versus N₂, expressed in delta notation using the modern atmosphere as the reference. Data are corrected for the effect of gravity. Note the slope in the DZ, indicative of a diffusive upward flux. (b) After δO₂/N₂ has been corrected for the atmospheric trend (mostly due to anthropogenic influence), there is a linear relation between δNe/N₂ and δO₂/N₂, indicating their preferential expulsion from closing bubbles occurs with a constant ratio.

a critical size of 3.6 Å. Recently, [Battle et al. \(2011\)](#) reported that the O₂ excluded from the closing bubbles is isotopically depleted in δ¹⁸O–O₂, showing the permeation through the ice lattice to be mass dependent. The mechanisms causing close-off fractionation are similar to those responsible for post-coring gas loss during ice sample storage ([Bender et al., 1995](#)).

The close-off fractionation is shown in [Figure 9](#) for SPO firn. The enrichment is strongest in the LIZ where the vast majority of bubbles are occluded, and the low effective diffusivity prevents diffusion of the signal. Note that a gradient exists in the DZ, indicative of an upward diffusive gas flux. Hereby some O₂ (Ne) is lost to the atmosphere, leading to a correspondingly depleted δO₂/N₂ (δNe/N₂) in gas bubbles.

It has been found that δO₂/N₂ can be used as a local summer insolation proxy, allowing orbital tuning of ice core chronologies ([Bender, 2002](#); [Kawamura et al., 2007](#)). This is thought to be controlled by a chain of events. Summer

insolation influences physical properties of surface ice grains, which in time will determine physical properties of deep firn strata after densification and metamorphism. [Bender \(2002\)](#) suggested that physical properties of grains near the firn–ice transition directly influence the magnitude of the O₂/N₂ fractionation. More recently, [Fujita et al. \(2009\)](#) proposed a model in which deep firn air transport is modulated, thus controlling the amount of O₂ lost to the atmosphere. Note that, though measured in the gas phase, δO₂/N₂ dating constrains the Ice Age and not the gas age.

Isotopic Diffusive Fractionation

A secular variation in the atmospheric mixing ratio of a gas species results in isotopic fractionation in the firn, even in the absence of a changing atmospheric isotopic composition ([Trudinger et al., 1997](#)). As an example, consider a rise in atmospheric CO₂. The lighter ¹²CO₂ isotopologue diffuses faster into the firn than ¹³CO₂, giving rise to isotopic depletion with depth. This diffusive fractionation can be approximated by

$$\varepsilon_{\text{DF}} = \begin{cases} -\frac{k}{C} \left(\frac{D}{D_i} - 1 \right) \Gamma(z) & \text{for } z \leq z_{\text{lid}} \\ -\frac{k}{C} \left(\frac{D}{D_i} - 1 \right) \Gamma(z_{\text{lid}}) & \text{for } z > z_{\text{lid}} \end{cases} \quad [8]$$

where C is the atmospheric mixing ratio (ppm), k the rate of increase (ppm year^{−1}), $D_i(D)$ the diffusion coefficient of the minor (major) isotopologue, Γ the mean gas age (year), and z_{lid} the lock-in depth.

[Figure 10](#) shows the diffusive fractionation calculated with eqn [8] and six numerical firn air models for a hypothetical exponential CO₂ increase of $k/C = 2.5 \times 10^{-3}$ (equivalent to $k = 1$ ppm year^{−1} at $C = 400$ ppm), while keeping δ¹³CO₂ fixed at 0‰. Within the DZ, the profiles agree well. Within the LIZ, the model solutions diverge, as the models use different parameterizations of LIZ transport. More work is needed to elucidate the nature of LIZ transport, and thereby the true magnitude of the diffusive fractionation.

Firn air and ice core records need to be corrected for the effect of diffusive fractionation, although often the effect will be small. Some corrections for the recent anthropogenic increase found in literature are 0.1‰ for ¹³CO₂ at Law Dome and 1.2‰ for ¹³CH₄ at WAIS-Divide ([Francey et al., 1999](#); [Mischler et al., 2009](#)). The effect is larger for CH₄ as the relative mass difference between the isotopologues, and thereby the ratio of diffusivities, is larger.

Smoothing by Diffusion and Bubble Trapping

The firn causes smoothing of the atmospheric signal, limiting the temporal resolution at which atmospheric variations are recorded in the ice. One can think of the firn as a low-pass filter to the atmospheric signal; high-frequency variations, such as the annual cycles of trace gases, are not preserved. [Spahni et al. \(2003\)](#) studied the attenuation of the rapid CH₄ excursion associated with the 8.2 ka event, and found it to be attenuated by about 13% in the GRIP core and 44% in the EPICA Dome C

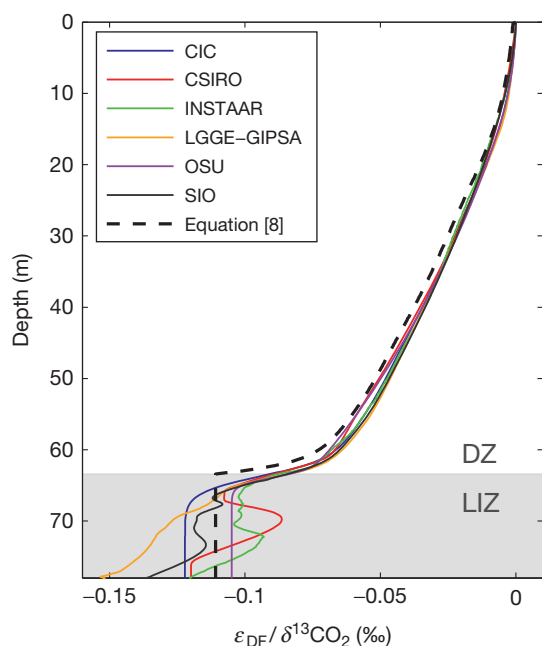


Figure 10 Isotopic diffusive fractionation in the NEEM firn for a hypothetical atmospheric CO₂ scenario. See Buizert et al. (2012) for details and model acronyms.

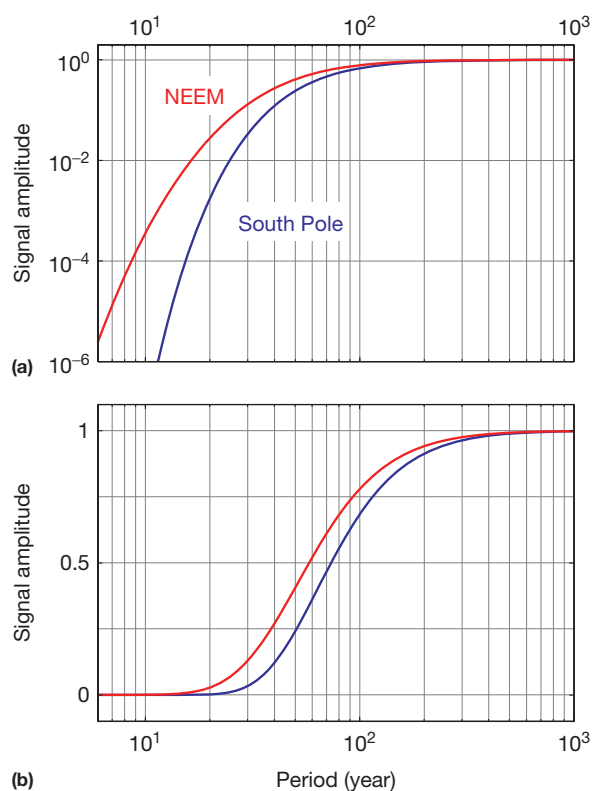


Figure 11 Bode plot of the firn transfer function on a log–log scale (a) and semi-log scale (b), obtained by fast Fourier transform (FFT) of the modeled CO₂ age distributions halfway the LIZ. The transmitted signal amplitude is plotted as a function of the period of the oscillation ($P=1/f$ with f the frequency in year^{−1}).

core. The low accumulation rate and deeper lock-in depth at the latter site are responsible for the stronger attenuation.

By taking the Fourier transform of the age distribution, we can evaluate how atmospheric variations at different time scales are attenuated by the firn column (Figure 11). As the bubble trapping occurs predominately throughout the LIZ, the age distribution halfway the LIZ is taken as a measure of the signal recorded in the air bubbles. Additional smoothing by the trapping process is not accounted for. Atmospheric variations on time scales of several hundreds of years or longer are recorded with their full amplitude; variations faster than ~50 years are attenuated so strongly they cannot be observed within the measurement noise. Variations at intermediate time scales (such as the aforementioned 8.2 ka event) are recorded with reduced amplitude. In theory, the true atmospheric history could be reconstructed from high-resolution records with deconvolution techniques also used for water stable isotope records (Johnsen, 1977).

Summary

Firn air studies are important for interpretation of ice core gas records and reconstructing recent atmospheric variations. Following $\delta^{15}\text{N}-\text{N}_2$ gravitational enrichment, the firn column is divided in the convective, diffusive, and lock-in zones, where gas transport is dominated by surface ventilation, diffusion, and advection, respectively. Transport properties are determined by pore geometry and connectivity, as well as density layering. Diffusivity measurements on finite samples do not represent the entire firn, and diffusivity needs to be reconstructed using reference tracers. Gravitational separation, Δage , diffusive smoothing, and diffusive isotopic fractionation are firn effects that need to be considered when interpreting ice core records. Thermal fractionation and close-off fractionation give rise to new proxies that can be used for temperature reconstruction and dating.

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References

- Albert MR, Schultz EF, and Perron FE Jr. (2000) Snow and firn permeability at Siple Dome, Antarctica. *Annals of Glaciology* 31: 353–356.
- Allison CE and Francey RJ (2007) Verifying Southern Hemisphere trends in atmospheric carbon dioxide stable isotopes. *Journal of Geophysical Research – Atmospheres* 112(D21): 27.
- Assonov SS, Brenninkmeijer CAM, and Jockel P (2005) The O-18 isotope exchange rate between firn air CO₂ and the firn matrix at three Antarctic sites. *Journal of Geophysical Research – Atmospheres* 110(D18): 15.
- Assonov SS, Brenninkmeijer CAM, Jockel PJ, Mulvaney R, Bernard S, and Chappellaz J (2007) Evidence for a CO increase in the SH during the 20th century based on firn air samples from Berkner Island, Antarctica. *Atmospheric Chemistry and Physics* 7: 295–308.

- Aydin M, Verhulst KR, Saltzman ES, et al. (2011) Recent decreases in fossil-fuel emissions of ethane and methane derived from firm air. *Nature* 476(7359): 198–201.
- Barnola JM, Pimienta P, Raynaud D, and Korotkevich YS (1991) CO₂-climate relationship as deduced from the Vostok ice core – A reexamination based on new measurements and on a reevaluation of the air dating. *Tellus Series B: Chemical and Physical Meteorology* 43(2): 83–90.
- Battle M, Bender M, Sowers T, et al. (1996) Atmospheric gas concentrations over the past century measured in air from firm at the South Pole. *Nature* 383(6597): 231–235.
- Battle MO, Severinghaus JP, Sofen ED, et al. (2011) Controls on the movement and composition of firm air at the West Antarctic Ice Sheet Divide. *Atmospheric Chemistry and Physics Discussions* 11: 11007–11021.
- Bender ML (2002) Orbital tuning chronology for the Vostok climate record supported by trapped gas composition. *Earth and Planetary Science Letters* 204(1–2): 275–289.
- Bender ML, Sowers T, Barnola JM, and Chappellaz J (1994) Changes in the O₂/N₂ ratio of the atmosphere during recent decades reflected in the composition of air in the firm at Vostok Station, Antarctica. *Geophysical Research Letters* 21(3): 189–192.
- Bender M, Sowers T, and Lipenkov V (1995) On the concentrations of O-2, N-2, and AR in trapped gases from ice cores. *Journal of Geophysical Research – Atmospheres* 100(D9): 18651–18660.
- Blunier T and Brook EJ (2001) Timing of millennial-scale climate change in Antarctica and Greenland during the last glacial period. *Science* 291(5501): 109–112.
- Blunier T and Schwander J (2000) Gas enclosure in ice: Age difference and fractionation. In: Hondoh T (ed.) *Physics of Ice Core Records*, pp. 307–326. Sapporo: Hokkaido University Press.
- Braunlich M, Aballanin O, Marik T, et al. (2001) Changes in the global atmospheric methane budget over the last decades inferred from C-13 and D isotopic analysis of Antarctic firm air. *Journal of Geophysical Research – Atmospheres* 106(D17): 20465–20481.
- Buizert C, Martinierie P, Petrenko VV, et al. (2012) Gas transport in firm: Multiple-tracer characterisation and model intercomparison for NEEM, Northern Greenland. *Atmospheric Chemistry and Physics* 12: 4259–4277.
- Butler JH, Battle M, Bender ML, et al. (1999) A record of atmospheric halocarbons during the twentieth century from polar firm air. *Nature* 399(6738): 749–755.
- Caillon N, Severinghaus JP, Barnola JM, et al. (2001) Estimation of temperature change and of gas age – Ice age difference, 108 kyr B.P., at Vostok, Antarctica. *Journal of Geophysical Research* 106(D23): 31893–31901.
- Caillon N, Severinghaus JP, Jouzel J, Barnola J-M, Kang J, and Lipenkov VY (2003) Timing of atmospheric CO₂ and Antarctic temperature changes across termination III. *Science* 299(5613): 1728–1731.
- Capron E, Landais A, Lemieux-Dudon B, et al. (2010) Synchronising EDM and NorthGRIP ice cores using [delta]18O of atmospheric oxygen ([delta]18O_{atm}) and CH₄ measurements over MIS5 (80–123 kyr). *Quaternary Science Reviews* 29(1–2): 222–234.
- Clark ID, Henderson L, Chappellaz J, et al. (2007) CO₂ isotopes as tracers of firm air diffusion and age in an Arctic ice cap with summer melting, Devon Island, Canada. *Journal of Geophysical Research* 112: 7.
- Colbeck SC (1989) Air movement in snow due to windpumping. *Journal of Glaciology* 35(120): 209–213.
- Courville Z, Hörhold M, Hopkins M, and Albert M (2011) Lattice-Boltzmann modeling of the air permeability of polar firm. *Journal of Geophysical Research* 115(F4): F04032.
- Craig H, Horibe Y, and Sowers T (1988) Gravitational separation of gases and isotopes in polar ice caps. *Science* 242(4886): 1675–1678.
- CRYOSPHERIC STUDIES OF ATMOSPHERIC TRENDS IN STRATOSPHERICALLY AND RADIATIVELY IMPORTANT GASES (CRYOSTAT): <http://badc.nerc.ac.uk/data/cryostat> (access August 2011).
- Dreyfus GB, Jouzel J, Bender ML, Landais A, Masson-Delmotte V, and Leuenberger M (2010) Firm processes and [delta]15N: Potential for a gas-phase climate proxy. *Quaternary Science Reviews* 29(1–2): 28–42.
- Dullien FAL (1975) Single phase flow through porous media and pore structure. *Chemical Engineering Journal* 10(1): 1–34.
- Etheridge DM, Steele LP, Langenfelds RL, Francey RJ, Barnola JM, and Morgan VI (1996) Natural and anthropogenic changes in atmospheric CO₂ over the last 1000 years from firm air in Antarctic ice and firm. *Journal of Geophysical Research* 101(D2): 4115–4128.
- Fabre A, Barnola JM, Arnaud L, and Chappellaz J (2000) Determination of gas diffusivity in polar firm: Comparison between experimental measurements and inverse modeling. *Geophysical Research Letters* 27(4): 557–560.
- Firm Record of Trace Gases Relevant to Atmospheric Chemical Change over 100 yrs (FIRETRACC/100): <http://badc.nerc.ac.uk/data/firetracc> (access August 2011).
- Fischer H, Wahlen M, Smith J, Mastoianni D, and Deck B (1999) Ice core records of atmospheric CO₂ around the last three glacial terminations. *Science* 283(5408): 1712–1714.
- Francey RJ, Allison CE, Etheridge DM, et al. (1999) *A 1000-Year High Precision Record of Delta C-13 in Atmospheric CO2*, pp. 170–193. Copenhagen: Munksgaard Int Publ Ltd.
- Freijer JI and Löffelaar PA (1996) Adapted Fick's law applied to soil respiration. *Water Resources Research* 32(4): 791–800.
- Freitag J, Dorbrindt U, and Kipfstuhl J (2002) A new method for predicting transport properties of polar firm with respect to gases on the pore-space scale. *Annals of Glaciology* 35: 538–544.
- Freitag J, Wilhelms F, and Kipfstuhl J (2004) Microstructure-dependent densification of polar firm derived from X-ray microtomography. *Journal of Glaciology* 50: 243–250.
- Fujita S, Okuyama J, Hori A, and Hondoh T (2009) Metamorphism of stratified firm at Dome Fuji, Antarctica: A mechanism for local insolation modulation of gas transport conditions during bubble close off. *Journal of Geophysical Research* 114: 9.
- Goujon C, Barnola JM, and Ritz C (2003) Modeling the densification of polar firm including heat diffusion: Application to close-off characteristics and gas isotopic fractionation for Antarctica and Greenland sites. *Journal of Geophysical Research – Atmospheres* 108(D24): 18.
- Herron MM and Langway CC (1980) Firm densification – An empirical model. *Journal of Glaciology* 25(93): 373–385.
- Huber C, Beyerle U, Leuenberger M, et al. (2006) Evidence for molecular size dependent gas fractionation in firm air derived from noble gases, oxygen, and nitrogen measurements. *Earth and Planetary Science Letters* 243(1–2): 61–73.
- Ishijima K, Sugawara S, Kawamura K, et al. (2007) Temporal variations of the atmospheric nitrous oxide concentration and its delta N-15 and delta O-18 for the latter half of the 20th century reconstructed from firm air analyses. *Journal of Geophysical Research – Atmospheres* 112(D3): 12.
- Johnsen SJ (1977) Stable isotope homogenization of polar firm and ice. In: *Proceedings of the Symposium on Isotopes and Impurities in Snow and Ice, IUGG XVI, General Assembly, Grenoble* 210–219.
- Kaspers KA, van de Wal RSW, van den Broeke MR, Schwander J, van Lipzig NPM, and Brenninkmeijer CAM (2004) Model calculations of the age of firm air across the Antarctic continent. *Atmospheric Chemistry and Physics* 4(5): 1365–1380.
- Kawamura K, Parrenin F, Lisiecki L, et al. (2007) Northern Hemisphere forcing of climatic cycles in Antarctica over the past 360,000 years. *Nature* 448(7156): 912–916.
- Kawamura K, Severinghaus JP, Ishidoya S, et al. (2006) Convective mixing of air in firm at four polar sites. *Earth and Planetary Science Letters* 244(3–4): 672–682.
- Kipfstuhl J, Faria SH, Azuma N, et al. (2009) Evidence of dynamic recrystallization in polar firm. *Journal of Geophysical Research* 114(B5): B05204.
- Landais A, Barnola JM, Kawamura K, et al. (2006) Firm-air delta N-15 in modern polar sites and glacial-interglacial ice: A model-data mismatch during glacial periods in Antarctica? *Quaternary Science Reviews* 25(1–2): 49–62.
- Landais A, Caillon N, Goujon C, et al. (2004) Quantification of rapid temperature change during DO event 12 and phasing with methane inferred from air isotopic measurements. *Earth and Planetary Science Letters* 225(1–2): 221–232.
- Lemieux-Dudon B, Blayo E, Petit J-R, et al. (2010) Consistent dating for Antarctic and Greenland ice cores. *Quaternary Science Reviews* 29(1–2): 8–20.
- Leuenberger MC, Lang C, and Schwander J (1999) Delta(15)N measurements as a calibration tool for the paleothermometer and gas-ice age differences: A case study for the 8200 BP event on GRIP ice. *Journal of Geophysical Research – Atmospheres* 104(D18): 22163–22170.
- Levchenko VA, Francey RJ, Etheridge DM, et al. (1996) The ¹⁴C ‘bomb spike’ determines the age spread and age of CO₂ in Law Dome firm and ice. *Geophysical Research Letters* 23(23): 3345–3348.
- Loulergue L, Parrenin F, Blunier T, et al. (2007) New constraints on the gas age-ice age difference along the EPICA ice cores, 0–50 kyr. *Climate of the Past* 3(3): 527–540.
- Loulergue L, Schilt A, Spahni R, et al. (2008) Orbital and millennial-scale features of atmospheric CH₄ over the past 800,000 years. *Nature* 453(7193): 383–386.
- Lüthi D, Le Floch M, Bereiter B, et al. (2008) High-resolution carbon dioxide concentration record 650,000–800,000 years before present. *Nature* 453(7193): 379–382.
- Martinierie P, Lipenkov VY, Raynaud D, Chappellaz J, Barkov NI, and Lorius C (1994) Air content paleo record in the Vostok ice core (Antarctica) – A mixed record of climatic and glaciological parameters. *Journal of Geophysical Research – Atmospheres* 99(D5): 10565–10576.
- Martinierie P, Raynaud D, Etheridge DM, Barnola JM, and Mazaudier D (1992) Physical and climatic parameters which influence the air content in polar ice. *Earth and Planetary Science Letters* 112(1–4): 1–13.
- Massmann J and Farrier DF (1992) Effects of atmospheric pressures on gas transport in the vadose zone. *Water Resources Research* 28(3): 777–791.
- Mischler JA, Sowers TA, Alley RB, et al. (2009) Carbon and hydrogen isotopic composition of methane over the last 1000 years. *Global Biogeochemical Cycles* 23.

- Montzka SA, Aydin M, Battle M, et al. (2004) A 350-year atmospheric history for carbonyl sulfide inferred from Antarctic firm air and air trapped in ice. *Journal of Geophysical Research – Atmospheres* 109(D22): 9–11.
- Morgan V, Delmotte M, van Ommen T, et al. (2002) Relative timing of deglacial climate events in Antarctica and Greenland. *Science* 297(5588): 1862–1864.
- Petit JR, Jouzel J, Raynaud D, et al. (1999) Climate and atmospheric history of the past 420,000 years from the Vostok ice core, Antarctica. *Nature* 399(6735): 429–436.
- Raynaud D, Lipenkov V, Lemieux-Dudon B, Duval P, Loutre MF, and Lhomme N (2007) The local insolation signature of air content in Antarctic ice. A new step toward an absolute dating of ice records. *Earth and Planetary Science Letters* 261(3–4): 337–349.
- Rommelaere V, Arnaud L, and Barnola JM (1997) Reconstructing recent atmospheric trace gas concentrations from polar firm and bubbly ice data by inverse methods. *Journal of Geophysical Research – Atmospheres* 102(D25): 30069–30083.
- Schwander J (1989) The transformation of snow to ice and the occlusion of gases. In: Oeschner H and Langway CC (eds.) *The Environmental Record in Glaciers and Ice Sheets*, pp. 53–67. New York: Wiley.
- Schwander J, Barnola JM, Andrieu C, et al. (1993) The age of the air in the firm and the ice at Summit, Greenland. *Journal of Geophysical Research – Atmospheres* 98(D2): 2831–2838.
- Schwander J, Sowers T, Barnola JM, Blunier T, Fuchs A, and Malaize B (1997) Age scale of the air in the summit ice: Implication for glacial–interglacial temperature change. *Journal of Geophysical Research – Atmospheres* 102(D16): 19483–19493.
- Schwander J and Stauffer B (1984) Age difference between polar ice and the air trapped in its bubbles. *Nature* 311(5981): 45–47.
- Schwander J, Stauffer B, and Sigg A (1988) Air mixing in firm and the age of the air at pore close-off. *Annals of Glaciology* 10: 141–145.
- Severinghaus JP, Albert MR, Courville ZR, et al. (2010) Deep air convection in the firm at a zero-accumulation site, central Antarctica. *Earth and Planetary Science Letters* 293(3–4): 359–367.
- Severinghaus JP and Battle MO (2006) Fractionation of gases in polar ice during bubble close-off: New constraints from firm air Ne, Kr and Xe observations. *Earth and Planetary Science Letters* 244(1–2): 474–500.
- Severinghaus JP and Brook EJ (1999) Abrupt climate change at the end of the last glacial period inferred from trapped air in polar ice. *Science* 286(5441): 930–934.
- Severinghaus JP, Grachev A, and Battle M (2001) Thermal fractionation of air in polar firm by seasonal temperature gradients. *Geochemistry, Geophysics, Geosystems* 2: Art. no. 2000GC000146.
- Severinghaus JP, Sowers T, Brook EJ, Alley RB, and Bender ML (1998) Timing of abrupt climate change at the end of the Younger Dryas interval from thermally fractionated gases in polar ice. *Nature* 391(6663): 141–146.
- Siegenthaler U, Friedli H, Loetschewer H, et al. (1988) Stable-isotope ratios and concentration of CO₂ in air from polar ice cores. *Annals of Glaciology* 10: 151–156.
- Sowers T, Bender M, Raynaud D, and Korotkevich YS (1992) Delta-N-15 of N₂ in air trapped in polar ice – A tracer of gas-transport in the firm and a possible constraint on ice age-gas age-differences. *Journal of Geophysical Research – Atmospheres* 97(D14): 15683–15697.
- Spahni R, Schwander J, Flückiger J, Stauffer B, Chappellaz J, and Raynaud D (2003) The attenuation of fast atmospheric CH₄ variations recorded in polar ice cores. *Geophysical Research Letters* 30(11): 1571.
- Stone JO (2000) Air pressure and cosmogenic isotope production. *Journal of Geophysical Research – Solid Earth* 105(B10): 23753–23759.
- Sturges WT, McIntyre HP, Penkett SA, et al. (2001a) Methyl bromide, other brominated methanes, and methyl iodide in polar firm air. *Journal of Geophysical Research* 106(D2): 1595–1606.
- Sturges WT, Penkett SA, Barnola J, Chappellaz JM, Atlas E, and Stroud V (2001b) A long-term record of carbonyl sulfide (COS) in two hemispheres from firm air measurements. *Geophysical Research Letters* 28(21): 4095–4098.
- Sturrock GA, Etheridge DM, Trudinger CM, Fraser PJ, and Smith AM (2002) Atmospheric histories of halocarbons from analysis of Antarctic firm air: Major Montreal Protocol species. *Journal of Geophysical Research – Atmospheres* 107(D24): 14.
- Sugawara S, Kawamura K, Aoki S, Nakazawa T, and Hashida G (2003) Reconstruction of past variations of delta C-13 in atmospheric CO₂ from its vertical distribution observed in the firm at Dome Fuji, Antarctica. *Tellus Series B: Chemical and Physical Meteorology* 55(2): 159–169.
- Thorntson DC and Pollock DW (1989) Gas transport in unsaturated zones: Multicomponent systems and the adequacy of Fick's laws. *Water Resources Research* 25(3): 477–507.
- Trudinger CM (2001) *The Carbon Cycle over the Last 1000 Years Inferred from Inversion of Ice Core Data*. PhD Thesis.
- Trudinger CM, Enting IG, Etheridge DM, et al. (1997) Modeling air movement and bubble trapping in firm. *Journal of Geophysical Research – Atmospheres* 102(D6): 6747–6763.
- Trudinger CM, Etheridge DM, Rayner PJ, Enting IG, Sturrock GA, and Langenfelds RL (2002) Reconstructing atmospheric histories from measurements of air composition in firm. *Journal of Geophysical Research – Atmospheres* 107(D24): 13.
- Waddington ED and Cunningham J (1996) The effects of snow ventilation on chemical concentrations. In: Wolff EW and Bales RC (eds.) *NATO ASI Series*. Berlin: Springer-Verlag.
- Weiler K, Schwander J, Leuenberger M, et al. (2009) Seasonal variations of isotope ratios and CO₂ concentrations in firm air. In: Hondoh T (ed.) *Low Temperature Science, Supplement Issue: Physics of Ice Core Records II*, pp. 247–272. Sapporo: Institute of Low Temperature Science, Hokkaido University.
- Witrant E, Martinerie P, Severinghaus JP, et al. (2011) A new multi-gas constrained model of trace gas non-homogeneous transport in firm: Evaluation and behavior at eleven polar sites. *Atmospheric Chemistry and Physics Discussions* 11: 23029–23080.

ICE CORE RECORDS

Contents

Africa

Chinese, Tibetan Mountains

South America

Antarctic Stable Isotopes

Greenland Stable Isotopes

Correlations Between Greenland and Antarctica

Ice Margin Sites

Thermal Diffusion Paleotemperature Records

Africa

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Long ice-core records from Africa are confined to the ice fields on the top of Kilimanjaro, which were drilled in 2000. Other glaciers such as the Lewis Glacier in Kenya and the Speke Glacier in Uganda have been investigated and their current retreat has been documented, but Kilimanjaro at 5893 m above sea level contains the highest and coldest ice fields on the continent, and thus, was determined to have the greatest potential to yield long, high quality records of climatic and environmental variation.

Located in northern Tanzania along its border with Kenya (**Figure 1(a)**), Kilimanjaro ($3^{\circ}04'S$, $37^{\circ}21'E$) is the highest mountain in Africa and the largest of a group of volcanoes that continued to be active until at least the end of the Pleistocene (**Downie and Wilkinson, 1972**). It is a massive and relatively undissected volcano, which rises 5895 m above sea level from a $95 \text{ km} \times 65 \text{ km}$ -wide base. Glaciers are believed to have existed on Kilimanjaro, as well as on Mount Kenya and in the Ruwenzori Range, throughout the Holocene (**Rosqvist, 1990**). In fact, glaciers existed on Kilimanjaro during several periods within the Quaternary, and evidence exists for three glaciations before 100 ka. This would suggest that no major eruptions have occurred in at least the last 500 ka.

A drilling program conducted in 2000 yielded six ice cores from three ice fields on the top of Kilimanjaro (**Thompson et al., 2002**). Kibo, the 2.5 km-wide caldera at the summit, contains several small ice fields (**Figures 1(b), 2(a), and 2(b)**), which are remnants of a former, more extensive glacier. Three ice cores were drilled on the largest, the Northern Ice Field (NIF), which was $\sim 50 \text{ m}$ thick in 2000. The Southern Ice Field (SIF), from which two cores were obtained, was $\sim 20 \text{ m}$ thick in 2000, while a small body of ice in the center of Kibo, the Furtwängler Glacier (FWG) was 10 m thick and was drilled

for one core. In addition, the remnants of the Eastern Ice Field are still visible, often as isolated thin spires of ice (**Figure 2(c)**).

Meteorology of Equatorial East Africa

The climate records from the Kilimanjaro ice cores reflect variations in monsoon intensity. The annual climatic regime over East Africa is very similar to that of southern Asia, where the surface airflow is determined by the position of the leading edge of a monsoon trough. Along the east coast of Africa, the intertropical convergence zone (ITCZ) moves southward from about 2°S in October to 12°S by the end of December. It remains in this extreme southern position until the end of January, then starts slowly back on its northern journey until the end of April, when it is again located at 2°S . At this stage, the wind convergence zone moves north of the equator and forces the southwest monsoon northwards. As the ITCZ moves southward and then northward, the rains also move with it. During its southward movement, rains occur over Kenya, Uganda, and Tanzania from mid-October to mid-December. This period is locally known as the season of 'Short Rains.' This area receives rain again from about mid-March to the beginning of June during the northward migration of the ITCZ, during the season of 'Long Rains' (**Asnani, 1993**). During the 'Short Rains,' northeasterlies and northerlies replace the southeasterlies in the lower troposphere along the west coast of Africa in association with the build-up of the anticyclonic ridge over the Arabian peninsula. During the 'Long Rains,' southeasterlies push from the south in the lower troposphere along the coast until the low-level jet is established by the end of June.

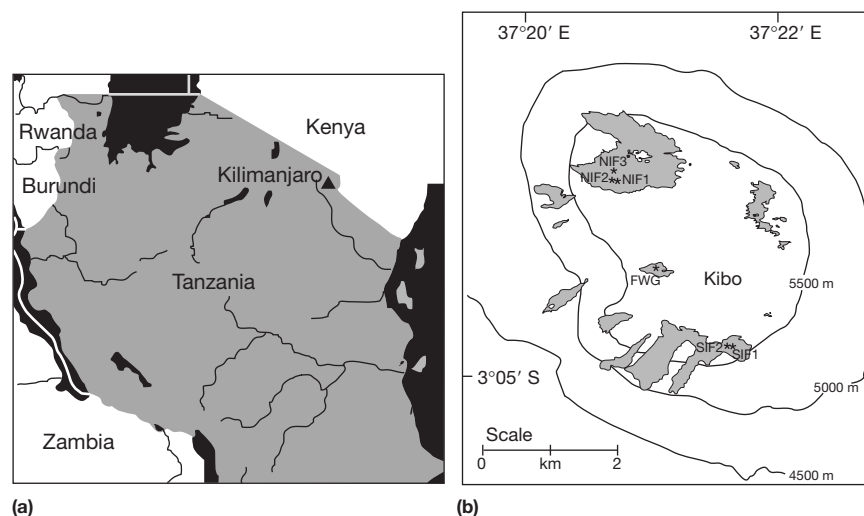


Figure 1 (a) Map of equatorial East Africa showing the location of Kilimanjaro in Tanzania. (b) The Kibo caldera on the top of Kilimanjaro, showing the ice fields in 2007 (shaded) with the locations of the six 2000 ice core drill sites.



Figure 2 (a) Kilimanjaro as seen from the south, with the southern lobes of the Southern Ice Field visible. (b) The Northern Ice Field (background), and the Furtwängler Glacier (foreground), as seen from the Uhuru peak, the summit of Kilimanjaro. (c) A remnant of the Eastern Ice Field.

The Ice Core Records

Five of the six cores drilled were analyzed for stable isotopic ratios of oxygen (^{18}O and ^{16}O). Figure 3(a)–3(e) shows these records as 10-year averages ending at AD 1950, which is close to where a dated tritium signal was located (in 1951/52). The tops of the ice fields are rapidly disappearing; therefore, dating above the known chronology marker is difficult. Figure 3 shows the marked reproducibility of the climate signal from ice cores drilled from one end of Kibo to the other. The shortest

time series, that from the FWG (Figure 3(e)), indicates that its ice began accumulating around the middle of the seventeenth century, which was the beginning of the Maunder Minimum as well as a period of very high water levels in Lake Naivasha in Kenya (Verschuren et al., 2000; Figure 3(f)). This suggests that this small glacier in the middle of Kibo originated and grew during a pluvial phase in the East African climate history, which was also the onset of the coldest part of the recent neoglacial period, called the 'Little Ice Age.' This is also marked by ^{18}O depletion (more negative $\delta^{18}\text{O}$) in the records from the

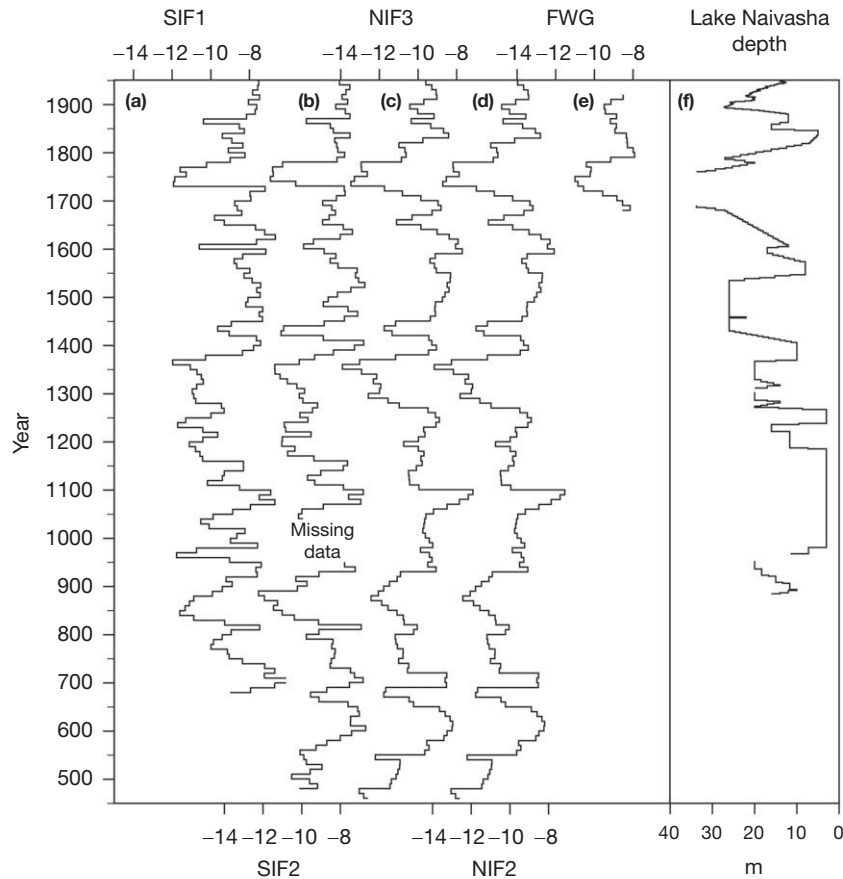


Figure 3 Ten-year averages of isotopic ratios of ^{18}O – ^{16}O ($\delta^{18}\text{O}$) from (a) Southern Ice Field Core 1, (b) Southern Ice Field Core 2, (c) Northern Ice Field Core 3, (d) Northern Ice Field Core 2, (e) the Furtwängler Glacier, and (f) water levels from Lake Naivasha in Kenya over the last 1100 years (Verschuren et al., 2000; data archived at the World Data Center for Paleoclimatology, Boulder, CO, USA).

SIF and the NIF (Figure 3(a)–3(d)). The records from the SIF extend back to AD 500–700, suggesting that this ephemeral glacier began to form during this interval.

A high-resolution record of Holocene climate conditions for eastern Africa is provided by NIF Core 3 (NIF3). Fifty-year averages of soluble aerosols, insoluble dust, and $\delta^{18}\text{O}$ are shown for the past 11700 years (Figure 4). Lower aerosol values (chloride, nitrate, ammonium, potassium, magnesium, and sulfate) after ~1800 years ago suggest wetter conditions, which may have contributed to the accumulation of the SIF. The Kilimanjaro ice fields therefore appear to be sensitive to climatic changes in East Africa, responding relatively quickly to regional changes in effective moisture balances.

Major Climatic Events Recorded in the Kilimanjaro Record

The African Humid Period

During the African Humid Period (11–4 ka), generally warmer and wetter monsoon conditions prevailed in response to the precession-driven increase in solar radiation (Nicholson and Flohn, 1980; Street and Grove, 1979). Lakes in the region rose as much as 100 m above today's levels (Street-Perrott and Perrott, 1990). For example, Lake Chad expanded from 17000 km² to cover an area comparable to that of the Caspian

Sea today (Broström et al., 1998; Grove and Warren, 1968; Street and Grove, 1979). Calculations of the hydrological balance (Kutzbach, 1980) suggest that annual precipitation over the Lake Chad Basin was as high as 650 mm, compared with 350 mm today, and mean precipitation over the Ziway-Shala Basin in Ethiopia was also calculated to be 47% higher than today. The Magadi Natron basin on the border between Tanzania and Kenya contained a lake that was 50 m deeper than today and had an area of 1600 km² in the early Holocene (Hillaire-Marcel and Casanova, 1987). After ~4000 years, African lake levels dropped as conditions became cooler and drier.

In the ice-core record, the African Humid Period is documented by isotopic enrichment (less negative $\delta^{18}\text{O}$), at least until 6.5 ka, and lower concentrations of species such as magnesium, calcium, sulfate, and nitrate (Figure 4). Contemporaneously, fluoride and sodium were at their highest levels, which is interpreted as intermittent periods of rapidly fluctuating lake levels within the longer period of overall high lake stands. Sodium and fluoride rich alkaline volcanic rocks in the East African Rift Valley are the source for these anions (Hecky and Kilham, 1973), and when the rocks are weathered, the products that are deposited in lakes form fluoride salt crusts during evaporation, known as trona (Nanyaro et al., 1984), or

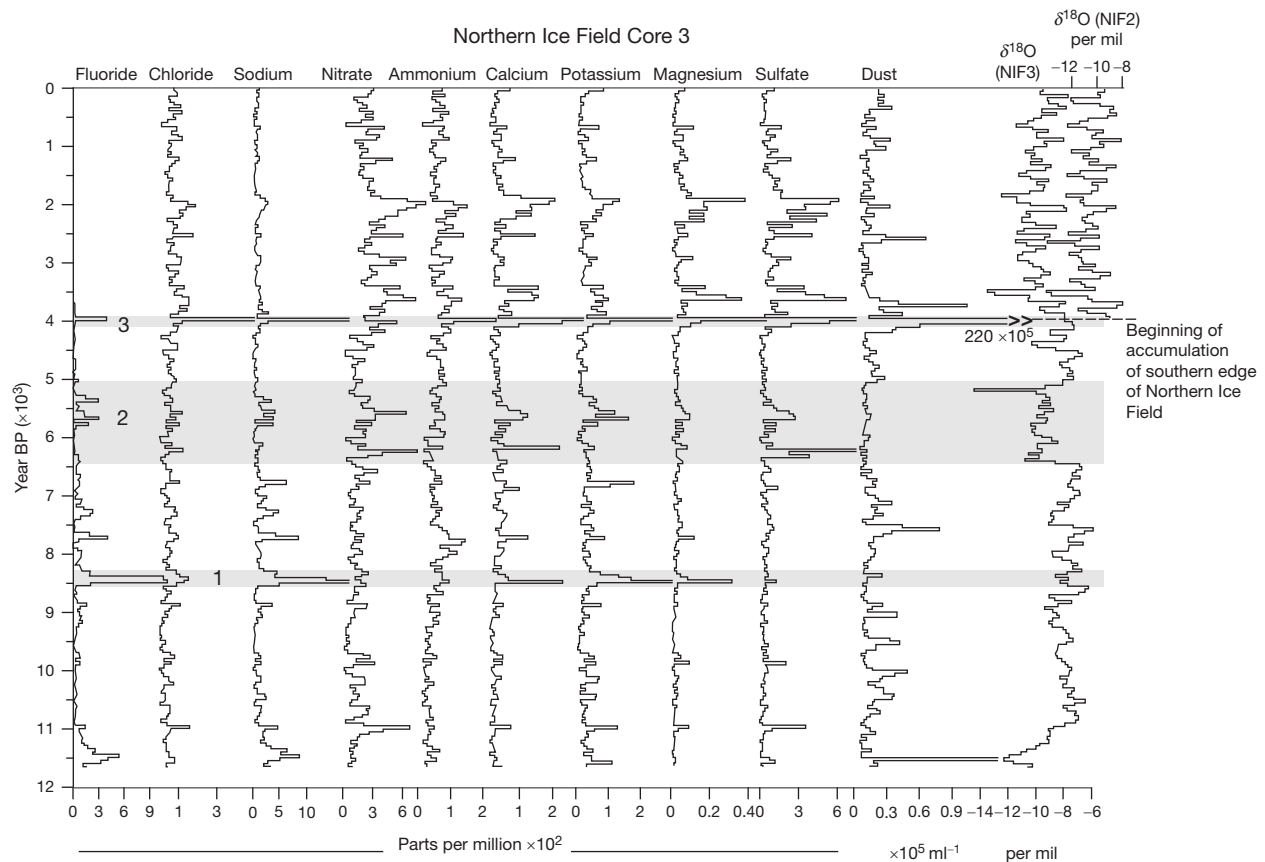


Figure 4 Fifty-year averages of concentrations of ion species and dust from NIF3, and stable isotopic ratios of ^{18}O – ^{16}O in NIF3 and NIF2. Abrupt climate events are marked by gray bars and events 1, 2, and 3 are discussed in the text.

magadi in the local vernacular. Periods of high runoff concentrate sodium and fluoride in the lakes, and when lake levels fall due to evaporation, trona precipitates and provides an aerosol source for the Kilimanjaro ice fields.

Abrupt climate events

Early Holocene

The fluoride concentrations between 8.4 and 8.2 ka are the highest in the Kilimanjaro record (marked as '1' in Figure 4), and are in fact the highest yet reported from any ice core. These high levels are nearly coincident with the large, abrupt reduction in methane (CH_4) in the European Greenland Ice Core Program (GRIP) core (Blunier et al., 1995), and with the largest ^{18}O depletion in the Holocene portion of both the GRIP and the Greenland Ice Sheet Project (GISP) 2 records (Alley et al., 1997). The peak fluoride and sodium levels suggest abrupt and intense lake level decreases accompanying a regional desiccation event. It is possible that the relatively short-term global CH_4 decrease was linked, and perhaps even preceded, by tropical hydrological changes leading to aridity, particularly in Africa.

Middle Holocene

An abrupt decrease in $\delta^{18}\text{O}$ from 6.5 to 5.3 ka ('2' in Figure 4), culminating in another sharp isotopic depletion at 5.2 ka, is

suggestive of a millennium-long climatic cooling. The interval of more negative ^{18}O values from 6.5 to 5 ka is nearly contemporaneous with a 'second humid period' (6.5–4.5 ka) when conditions were wetter than today, but drier than the early Holocene 'first humid period' (Nicholson and Flohn, 1980). During this time, the Ziway-Shala Basin attained its highest levels before dropping rapidly to near modern low levels by 4 ka (Street-Perrott and Perrott, 1990). The abrupt cooling event at the end of this second humid period about 5.2 ka is recorded as the largest ^{18}O depletion in the NIF record. This climatic change recorded in Kilimanjaro ice is contemporaneous with the largest ^{18}O enrichment in the Soreq cave speleothem record in Israel (Bar-Matthews et al., 1999), a decline in both lake levels in northern Africa (Sirocko et al., 1993; Street and Grove, 1976; Street-Perrott et al., 1990) and vegetative cover in India (Swain et al., 1983), and the lowest Holocene CH_4 concentrations in the GRIP ice core (Blunier et al., 1995). Documentary evidence ranging from northern Africa to Arabia (Weiss, 2000) also confirms this abrupt climate change. Around 5300 years ago, hierarchical societies formed in the Nile Valley and Mesopotamia, while Neolithic settlements in the inner desert of Arabia were abandoned (Uerpmann, 1990). This century-scale rapid cooling and drying event correlates with the Late Uruk abrupt climate change around 5.2 ka that is suggested to have altered environments outside the Mesopotamian lowlands (Weiss, 2000).

The second abrupt climate event recorded in the middle Holocene, dated around 4200 years ago ('3' in Figure 4), is associated with a visible dust layer (30 mm thick) and very high aerosol concentrations in NIF3. This is interpreted to represent a century-scale hiatus in ice accumulation due to a prolonged period of weaker monsoon activity. During this time, evidence indicates that the NIF decreased in size and that its southern ice wall retreated substantially. The $\delta^{18}\text{O}$ record from NIF2 actually begins at the chronological level of this event (Figure 4), which means that this part of the ice field began to reform only after the end of the drought. The ice field was most likely much larger by the end of the African Humid Period, but contracted very quickly during this dry interval. Thus, it appears that the major dust layer in NIF3 was formed during extremely dry conditions either before or during the retreat of the NIF margin.

This drought appears to have been widespread throughout the tropics, especially in those regions affected by monsoon circulation, and its climatic effects were experienced throughout northern and tropical Africa, the Middle East, and western Asia (Gasse and Van Campo, 1994; Pachur and Hoelzmann, 2000; Street-Perrott and Perrott, 1990). Water levels dropped in tropical east African lakes (Street-Perrott and Perrott, 1990), there was a very large influx of carbonate dust that was transported from Syria into the Gulf of Oman (Cullen et al., 2000), and the event is marked by a sharp increase in carbon isotopic ratios in the Soreq cave speleothem (Bar-Matthews et al., 1999). The drought was so severe that it has been considered instrumental in the collapse of a number of civilizations from North Africa to India (see articles in Dalfes

et al., 1997). The possibility that the drought extended beyond Africa, Asia and the Middle East into South America is suggested by the dust record in an ice core from the Andes of Northern Peru (Davis and Thompson, 2006). As the drought ameliorated, the NIF began to accumulate mass and reestablish itself over a larger portion of the crater rim.

Recent Retreat of East African Glaciers

All of the ice fields and glaciers in tropical East Africa are currently undergoing rapid retreat. Since ~1850 the Lewis Glacier on Mt. Kenya has decreased from 0.69 to 0.20 km², losing 71% of its surface area (Allison and Peterson, 1976), and from 1963 to 1992 alone it lost 40% of its mass (Hastenrath and Kruss, 1992). In fact a 1978 expedition to the Lewis Glacier, which sought to ascertain its suitability for drilling, confirmed that the ice was too warm (due to its low elevation and the recent warming) to preserve a useful climate record (Thompson, 1979). The surface area of the Speke glacier in the Ruwenzori Range in Uganda has decreased 74% from 6.51 km² in 1906 to 1.67 km² in 1990 (Kaser, 1998). Likewise, the ice cover on Kibo has retreated 85% (~12 to ~2.6 km²) from 1912 (Hastenrath and Greichar, 1997) to 2007 (Figure 5), and 26% of the ice that was present in 2000 has disappeared (Thompson et al., 2009). The ice on the Northern and Southern Ice Fields thinned about 3.6 and 24%, respectively from 2000 to 2007, while the ice on the FWG in the center of the crater thinned 50% from 2000 to 2009. At this rate, the FWG will disappear by 2018 and the remaining ice fields over the following decade, making the top of Kilimanjaro completely ice free for the first time in 11 ka.

There is no definitive consensus for the cause of the recent glacier retreat in East Africa. The discussion has been conducted in the scientific literature (e.g., Kaser et al., 2004; Mölg and Hardy, 2004; Mölg et al., 2009; Thompson et al., 2009, 2011) as well as in the media. The vertical lifting of the 0 °C isotherm resulting from increasing atmospheric warming has been cited, as has decreasing moisture in the region. The linear relationship, close to the meteoric water line, between $\delta^{18}\text{O}$ and deuterium isotopic ratios (δD) in the ice cores indicates that the disappearance of the ice fields on the top of Kilimanjaro is driven by melting (Thompson et al., 2011) rather than by sublimation (Cullen et al., 2007). In the Ruwenzori Mountains, which fall on the border between Uganda and D.R. Congo, both field work and satellite imagery show that the glaciers have decreased by ~50% in aerial extent from 1987 to 2003 (Taylor et al., 2006). This was synchronous with an increasing trend in air temperature of ~0.5 °C per decade over the last four decades. Limnological studies in this region confirm that the beginning of the modern glacier recession occurred with a warming regional climate and not with hydrological changes (Eggermont et al., 2010a,b; Russell et al., 2008). However, regardless of the cause(s), the disappearance of the ice fields in East Africa, particularly on Kibo, will have negative economic implications for the local populations as the ice is important for both the tourist industry and for the agricultural activities on the lower slopes.

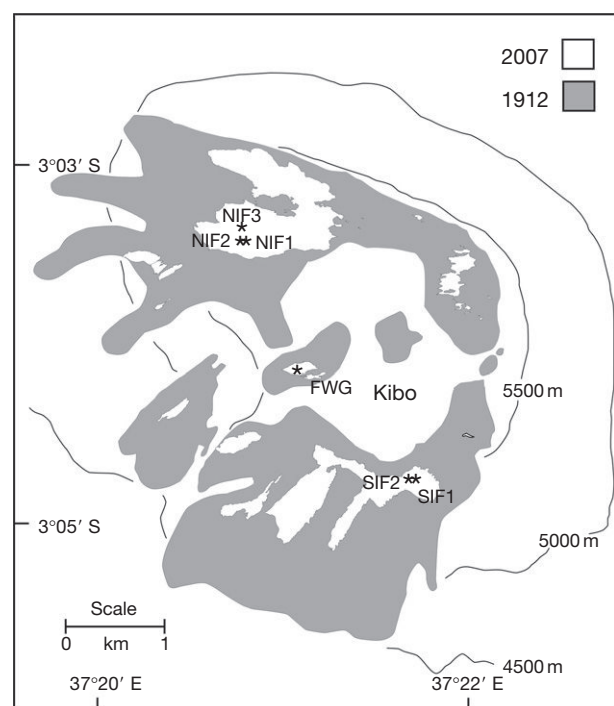


Figure 5 Extent of the Kibo ice fields in 1912 (shaded) and 2007 (white). The 1912 ice extent is from Hastenrath and Greichar (1997), the 2007 ice extent is from Thompson et al. (2009).

References

- Alley RB, Mayewski PA, Sowers T, Stuiver M, Taylor KC, and Clark PU (1997) Holocene climatic instability: A prominent, widespread event 8200 yr ago. *Geology* 25: 483–486.
- Allison I and Peterson JA (1976) Ice areas on Mt. Jaya. Their extent and recent history. In: Hope GS, Peterson JA, Radok U, and Allison I (eds.) *The Equatorial Glaciers of New Guinea. Results of the 1971–1973 Australian Universities' Expeditions to Irian Jaya: Survey, Glaciology, Meteorology, Biology and Palaeoenvironments*, p. 27. Rotterdam: A.A. Balkema.
- Asnani GC (1993) *Tropical Meteorology*, vol. 1, p. 375. Pune, India: Noble Printers Pvt. Ltd.
- Bar-Matthews M, Ayalon A, Kaufman A, and Wasserburg GJ (1999) The Eastern Mediterranean paleoclimate as a reflection of regional events: Soreq Cave, Israel. *Earth and Planetary Letters* 166: 85–95.
- Blunier T, Chappellaz J, Schwander J, Stauffer B, and Raynaud D (1995) Variations in atmospheric methane concentration during the Holocene epoch. *Nature* 374: 46–49.
- Broström A, Coe M, Harrison SP, et al. (1998) Land surface feedbacks and palaeomonsoons in northern Africa. *Geophysical Research Letters* 25: 3615–3618.
- Cullen HM, deMenocal PB, Hemming S, et al. (2000) Climate change and the collapse of the Akkadian empire: Evidence from the deep sea. *Geology* 28: 379–382.
- Cullen NJ, Mölg T, Kaser G, Steffen K, and Hardy DR (2007) Energy-balance model validation on the top of Kilimanjaro, Tanzania, using eddy covariance data. *Annals of Glaciology* 46: 227–233.
- Dafes HN, Kukla G, and Weiss H (eds.) (1997) *Third Millennium BC Climate Change and the Old World Social Collapse*. Heidelberg: Springer.
- Davis ME and Thompson LG (2006) An Andean ice-core record of a Middle Holocene mega-drought in North Africa and Asia. *Annals of Glaciology* 43: 34–41.
- Downie C and Wilkinson P (1972) *The Geology of Kilimanjaro; the Results of the Joint Sheffield University-Tanzania Geological Survey Expeditions in 1953 and 1957*. Sheffield: University of Sheffield, Department of Geology, and the Geological Survey of Tanzania.
- Eggermont H, Heiri O, Russell J, Vuille M, Audenaert L, and Verschuren D (2010a) Paleotemperature reconstruction in tropical Africa using fossil Chironomidae (insect: Diptera). *Journal of Paleolimnology* 43(3): 413–435.
- Eggermont H, Verschuren D, Audenaert L, et al. (2010b) Limnological and ecological sensitivity of Rwenzori mountain lakes to climate warming. *Hydrobiologia* 648: 123–142.
- Gasse F and Van Campo E (1994) Abrupt post-glacial climate events in West Asia and North Africa monsoon domains. *Earth and Planetary Science Letters* 126: 435–456.
- Grove AT and Warren A (1968) Quaternary landforms and climate on the south side of the Sahara. *The Geographical Journal* 134: 194–208.
- Hastenrath S and Greicher L (1997) Glacier recession on Kilimanjaro, East Africa, 1912–89. *Journal of Glaciology* 43: 455–459.
- Hastenrath S and Kruss PD (1992) The dramatic retreat of Mount Kenya's glaciers between 1963 and 1987: Greenhouse forcing. *Annals of Glaciology* 16: 127–133.
- Hecky RE and Kilham P (1973) Diatoms in alkaline, saline lakes: Ecology and geochemical implications. *Limnology and Oceanography* 18: 53–71.
- Hillaire-Marcel C and Casanova J (1987) Isotopic hydrology and paleohydrology of the Magadi (Kenya)–Natron (Tanzania) basin during the Late Quaternary. *Palaeogeography, Palaeoclimatology, Palaeoecology* 58: 155–181.
- Kaser G (1998) A review of the modern fluctuations of tropical glaciers. *Global and Planetary Change* 22: 93–103.
- Kaser G, Hardy DR, Mölg T, Bradley RS, and Hyera TM (2004) Modern glacier retreat on Kilimanjaro as evidence of climate change: Observations and facts. *International Journal of Climatology* 24: 329–339.
- Kutzbach JE (1980) Estimates of past climate at Paleolake Chad, North Africa, based on a hydrological and energy balance model. *Quaternary Research* 14: 210–223.
- Mölg T, Cullen NJ, Hardy DR, Winkler M, and Kaser G (2009) Quantifying climate change in the tropical midtroposphere over East Africa from glacier shrinkage on Kilimanjaro. *Journal of Climate* 22: 4162–4181.
- Mölg T and Hardy DR (2004) Ablation and associated energy balance on a horizontal glacier surface on Kilimanjaro. *Journal of Geophysical Research* 109: D16104. <http://dx.doi.org/10.1029/2003JD004338>.
- Nanyaro JT, Aswathanarayana U, Mungure JS, and Lahermo PW (1984) A geochemical model for the abnormal fluoride concentrations in water in Northern parts of Tanzania. *Journal of African Earth Sciences* 2: 129–140.
- Nicholson S and Flohn H (1980) African environmental and climatic changes and the general circulation in late Pleistocene and Holocene. *Climatic Change* 2: 176–179.
- Pachur HJ and Hoelzmann P (2000) Late quaternary palaeoecology and palaeoclimates of the eastern Sahara. *Journal of African Earth Sciences* 30: 929–939.
- Rosqvist G (1990) Quaternary glaciations in Africa. *Quaternary Science Reviews* 9: 281–297.
- Russell JM, Eggermont H, and Verschuren D (2008) Paleolimnological records of recent glacier recession in the Rwenzori Mountains, Uganda–D.R. Congo. *Journal of Paleolimnology* 41(2): 253–271.
- Sirocko F, Sarnthein M, Erlenkeuser H, Lange H, Arnold M, and Duplessy JC (1993) Century-scale events in monsoonal climate over the 24,000 years. *Nature* 364: 322–325.
- Street FA and Grove AT (1976) Environmental and climatic implications of Late Quaternary lake-level fluctuations in Africa. *Nature* 261: 385–390.
- Street FA and Grove AT (1979) Global maps of lake-level fluctuations since 30,000 yr BP. *Quaternary Research* 12: 83–118.
- Street-Perrott FA, Mitchell JFB, Marchand DS, and Brunner JS (1990) Milankovitch and albedo forcing of the tropical monsoons: A comparison of geological evidence and numerical simulations for 9000 yr BP. *Transactions of the Royal Society of Edinburgh* 81: 407–427.
- Street-Perrott FA and Perrott RA (1990) Abrupt climate fluctuations in the tropics: The influence of the Atlantic Ocean circulation. *Nature* 343: 607–612.
- Swain AM, Kutzbach JE, and Hastenrath S (1983) Estimates of Holocene precipitation for Rajasthan, India, based on pollen and lake-level data. *Quaternary Research* 19: 1–17.
- Taylor RG, Mileham L, Tindimugaya A, Majugu A, Muwanga A, and Nakileza B (2006) Recent glacial recession in the Rwenzori Mountains of East Africa due to rising air temperature. *Geophysical Research Letters* 33: L10402. <http://dx.doi.org/10.1029/2006GL025962>.
- Thompson LG (1979) Ice core studies from Mt. Kenya, Africa, and their relationship to other ice core studies. *Sea Level, Ice, and Climatic Change (Proceedings of the Canberra Symposium, December, 1979)*, pp. 55–62. IAHS Publ. No. 131.
- Thompson LG, Brecher HH, Mosley-Thompson E, Hardy DR, and Mark BG (2009) Glacier loss on Kilimanjaro continues unabated. *Proceedings of the National Academy of Sciences* 106: 19770–19775.
- Thompson LG, Mosley-Thompson E, Davis ME, and Mountain K (2011) A paleoclimatic perspective on the 21st-century glacier loss on Kilimanjaro, Tanzania. *Annals of Glaciology* 52: 60–68.
- Thompson LG, Mosley-Thompson E, Davis ME, et al. (2002) Kilimanjaro ice core records: Evidence of Holocene climate change in Tropical Africa. *Science* 298: 589–593.
- Uerpmann HP (1990) Radiocarbon dating of shell middens in the Sultanate of Oman. *PACT* 29: 335–347 (in German).
- Verschuren D, Laird KR, and Cumming BF (2000) Rainfall and drought in equatorial east Africa during the past 1,100 years. *Nature* 403: 410–413.
- Weiss H (2000) Beyond the Younger Dryas: Collapse as adaptation to abrupt climate change in ancient west Asia and the eastern Mediterranean. In: Bawden G and Reyrcraft RM (eds.) *Environmental Disaster and the Archeology of Human Response*, p. 75. Albuquerque, NM: Maxwell Museum of Anthropology.

Further Reading

- Thompson LG, Mosley-Thompson E, and Davis ME (2011) A paleoclimatic perspective on the 21st Century glacier loss on Kilimanjaro. *Annals of Glaciology* 52: 60–68.
- Thompson LG, Mosley-Thompson E, Davis ME, and Brecher HH (2011) Tropical glaciers, recorders and indicators of climate change, are disappearing globally. *Annals of Glaciology* 52(59): 23–34.
- Thompson LG (2010) Climate change: The evidence and our options. *The Behavior Analyst* 33: 1153–1170.
- Kehrwald NM, Thompson LG, Tandong Y, Mosley-Thompson E, Schotterer U, Alfimov V, Beer J, Eikenberg J, and Davis ME (2008) Mass loss on Himalayan glacier endangers water resources. *Geophysical Research Letters* 35: L22503.
- Thompson LG, Mosley-Thompson E, Brecher H, Davis ME, Leon B, Les D, Mashiotta TA, Lin P-N, and Mountain K (2006) Evidence of abrupt tropical climate change: past and present. *Proceedings of the National Academy of Sciences* 103 (28): 10536–10543.

Chinese, Tibetan Mountains

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Introduction

The Tibetan Plateau is defined by several mountain ranges that occupy the heart of central Asia (Figures 1 and 2). The plateau occupies an area extending roughly 2500 km (east–west) by 1000 km (north–south) and has an average elevation of over 4500 m above sea level (a.s.l.). The mountain ranges on the southern margin of the Tibetan Plateau – the Himalayas – contain the highest mountains in the world. Central Asian mountain ranges also contain a large number of glaciers (Figure 3) which together cover an area greater than 120 000 km² (Table 1), the highest concentration of glaciers outside of polar regions. A few of these glaciers provide sites that are suitable for the recovery of ice cores that extend back thousands of years and allow for the development of detailed records of climatic and environmental change in the region.

While the majority of ice-core investigations have been undertaken in the polar regions, chemical and physical analyses of ice cores recovered from carefully selected accumulation zones of central Asian glaciers allow for the development of detailed, high-resolution (i.e., seasonal to decadal) paleoclimate records. These records are especially important, given the critical role that the highlands of central Asia play in the development and intensity of the Asian monsoon, and the importance of the summer monsoon in providing life-sustaining rains to more than half the world's population. Central Asian ice cores provide records that document change in the monsoon on a variety of temporal and spatial scales. These records also provide the basis for the investigation of the linkages among the Asian monsoon and other coupled ocean–atmosphere circulation systems (e.g., El Niño/Southern Oscillation (SO) and the North Atlantic Oscillation (NAO)) as well as to assess possible forcing functions that cause climate to change and thereby improve our understanding of the Earth's climate system, especially at latitudes where the majority of Earth's human population live.

Asian Versus Polar Ice-Core Records

Compared to ice cores recovered from polar regions, the geographical setting of central Asian glaciers creates some additional challenges for the interpretation of central Asian ice-core records. First, while the relatively remote polar regions are distant from significant local sources of wind-blown dust, biogenic emissions, and anthropogenic emissions, central Asian glaciers lie adjacent to many of these large sources. For example, there are several large sources of wind-blown dust in central Asia which include the Taklamakan and Gobi deserts, and the arid regions east of the shrinking Aral Sea. This wind-blown dust is rich in calcium, magnesium, potassium, sodium, and chloride and is the dominant determinant of the spatial and seasonal variation of snow chemistry on most central

Asian glaciers (Wake et al., 1993, 2004). Glaciers which lie adjacent to large arid regions (such as those on the northern and western margins of the Tibetan Plateau) display very high concentrations of dust-related major ions, due to the influx of dust from local sources. Glaciers more distant from major dust source regions show intermediate levels of major ions and dust, while glaciers on the southern slopes of the eastern Himalayas are relatively free from the influence of desert dust (Figure 4). Anthropogenic emissions from Urumqi and the Indian subcontinent are also recorded in precipitation chemistry in the Bogda Shan and the Himalayas, respectively (Shrestha et al., 2002; Wake et al., 1992). In addition, biogenic emissions from agricultural activities in the Himalayas and southern Tibetan Plateau influence snow chemistry in the region (Kang et al., 2002c; Shrestha et al., 2002).

Second, the stable isotope measurements that have been central to the climatic records developed from polar ice cores do not necessarily provide a readily decipherable temperature proxy for central Asian ice cores. Several studies have shown that in the northern regions of the Tibetan Plateau, stable isotopes measured from individual precipitation events are well correlated with temperature as measured at nearby meteorological stations (Tian et al., 2003; Yao et al., 1996), and oxygen isotope records developed from ice cores in this region likely provide a proxy for temperature change, although the records are also influenced by changes in sources of moisture during different seasons or climatic periods (Hoffman and Heinmann, 1997; Kreutz et al., 2003). Stable isotopes measured in precipitation in the southern half of the Tibetan Plateau are influenced by the 'amount effect' and show the largest depletion of the heavier isotope (¹⁸O) during the summer (Figure 5; e.g., Kang et al., 2002a; Tian et al., 2003; Wushiki, 1977), and are also influenced by changes in source regions of moisture (e.g., Indian Ocean, Atlantic Ocean, Mediterranean Sea). These multiple influences on stable isotope ratios in precipitation in the Himalayas and southern Tibetan Plateau complicate the climatic interpretation of stable isotope records from this region and makes deciphering a proxy temperature signal difficult.

Third, several central Asian ice cores have been drilled in areas with significant meltwater percolation during the summer and in some cases, where accumulation occurs primarily as superimposed ice (ice exposed at the glacier surface that was formed by the refreezing of melted snow; commonly found in the accumulation area of a glacier). For example, at 5850 m on a small glacier on the south side of Nangapi Gosum in the central Himalayas, 20 m borehole temperatures were 0 °C and water was found in the borehole just below the firn–ice transition at 20 m. Stable isotope records show a seasonal signal for only the most recent year of accumulation: below this the seasonal signals have been homogenized as a result of meltwater percolation. Water was present in shallow boreholes drilled at 5350 m on the Fedchenko glacier in the Pamirs.

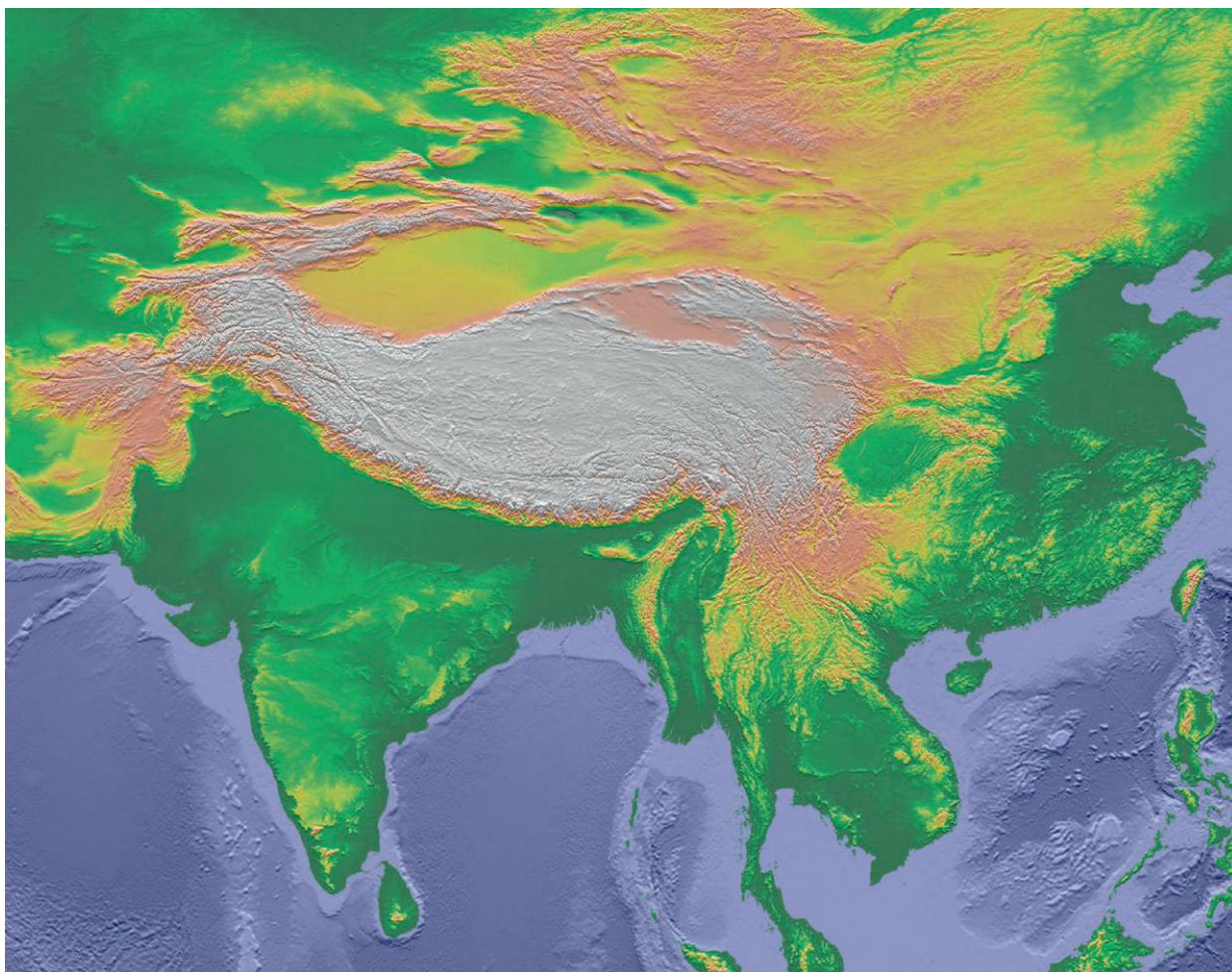


Figure 1 Topography of central Asia from the Global Land One-km Base Elevation (GLOBE) Project digital elevation model. Images available from the US National Oceanic and Atmospheric Administration (NOAA) National Climatic Data Center online at: <http://www.ngdc.noaa.gov>.

The effects of meltwater percolation have also been noted at 5180 m on Yala glacier (Watanabe et al., 1984). Despite this, glaciers in which melting occurs in the upper portion of the record do show seasonal signals and have provided valuable paleoclimate records discussed in detail below (e.g., Dundu, Guliya, Far East Rongbuk glacier), presumably because the melting is restricted to upper layers of central Asian glaciers in response to recent warming in the region (e.g., Liu and Chen, 2000).

Central Asian Ice Cores

Early Asian Ice Cores

Early efforts at recovering and analyzing ice cores from the mountains of central Asia began in the 1980s with a 60 m core recovered from the Yala glacier (5180 m a.s.l.), Nepal Himalayas (Watanabe et al., 1984), and a 17 m core recovered from the Sentik glacier (4908 m a.s.l.) in the Nun Kun massif in northwest India (Mayewski et al., 1984). Both cores were recovered in areas with relatively high snow accumulation (e.g., 0.9 m per year at Sentik) and were dated using the 1963

peak in total beta-activity resulting from the fallout of atmospheric nuclear weapons testing. Due to meltwater infiltration in summer, no seasonal signals were preserved in the Yala glacier core. Conversely, the Sentik core retained distinct seasonal layering and chemically differentiable shifts that were related to variability in monsoon circulation on seasonal to multiannual scales, indicating that climate-related records were preserved in at least some of the glaciers on the southern slopes of the Himalayas.

Deep Asian Ice Cores

Since the late 1980s, several long ice cores have been recovered from the mountains of central Asia (Table 2 and Figure 3) and provide records to investigate climate change on timescales ranging from decades to millennia. Three surface-to-bedrock cores were recovered from the Dundu ice cap on the north-east margin of the Tibetan Plateau in 1987 (Thompson et al., 1989). The $\delta^{18}\text{O}$, microparticle, and anion records developed from these cores provide a record of climate change for the northern regions of the Tibetan Plateau for the Holocene and late Pleistocene. The relatively high anion and microparticle

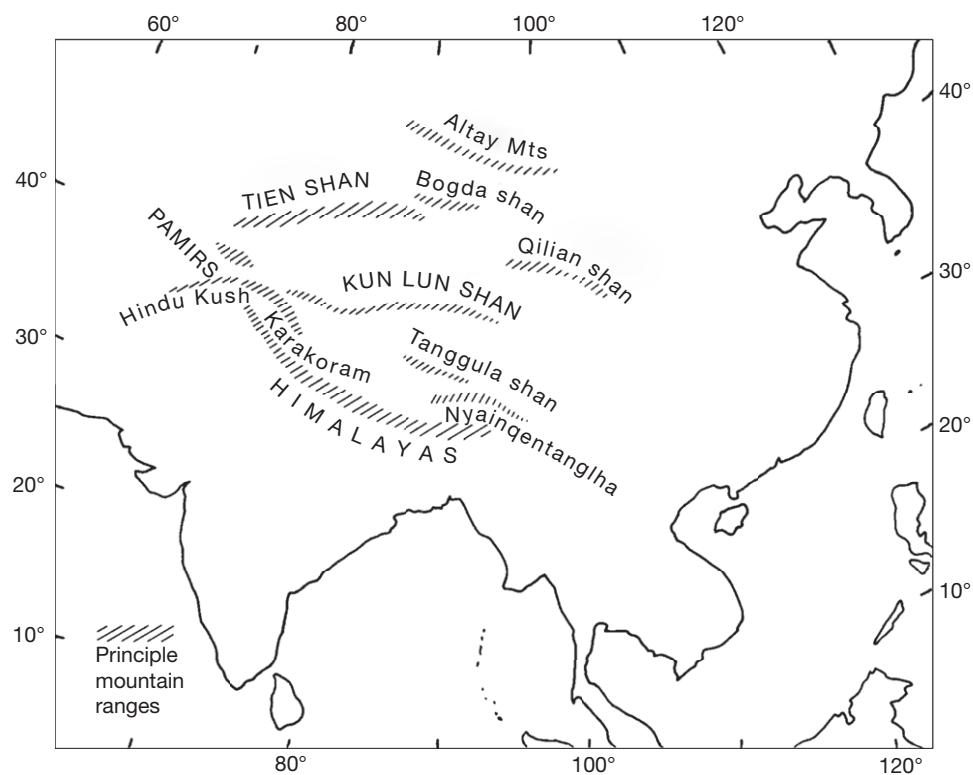


Figure 2 Principle mountain ranges in central Asia.

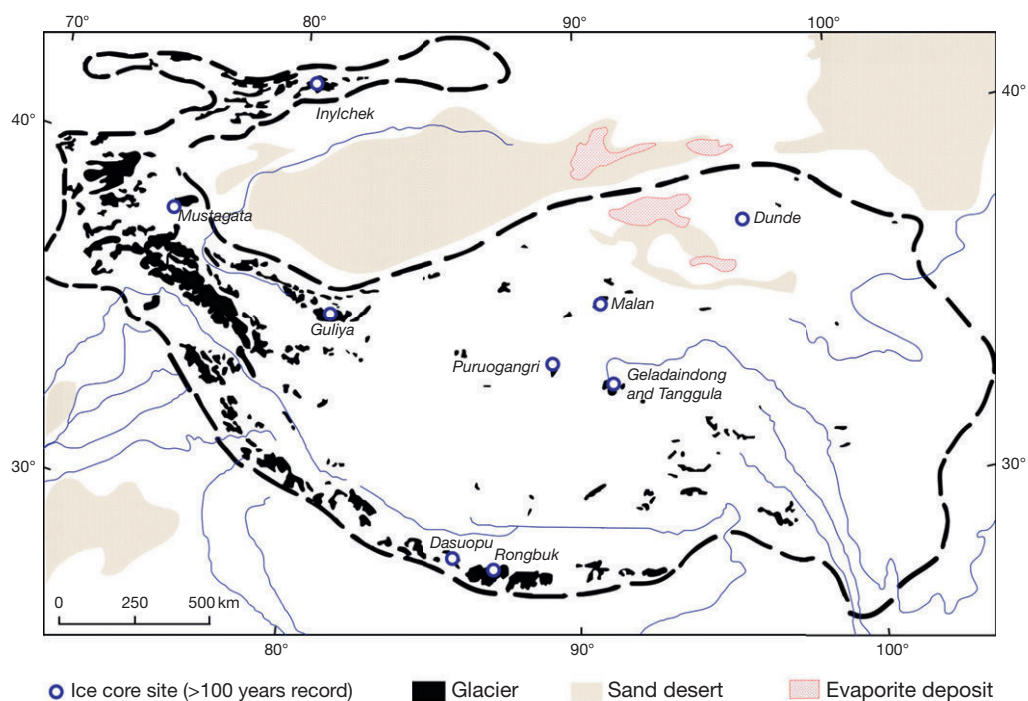


Figure 3 Location map for deep ice cores (>100 years of record) recovered from the mountains of central Asia. The heavy dashed line represents the 3000 m contour line and outlines the Tibetan Plateau. Landscape and topographic information were derived from a map (Tibet and the Mountains of Central Asia) published by the Royal Geographic Society (UK).

Table 1 Glacierized surface area in central Asia

Country	Area (km ²)
China	56 500
Pakistan and India	40 000
Kazakhstan, Tajikistan, and Kyrgyzstan	18 200
Nepal and Bhutan	7500
Afghanistan	4000

Source: Haeberli W, Bosch H, Scherler K, Ostrom G, and Wallen CC (1989) World Glacier Inventory: Status 1988. Wallingford, England: IAHS – UNEP – UNESCO (compiled by the World Glacier Monitoring Service).

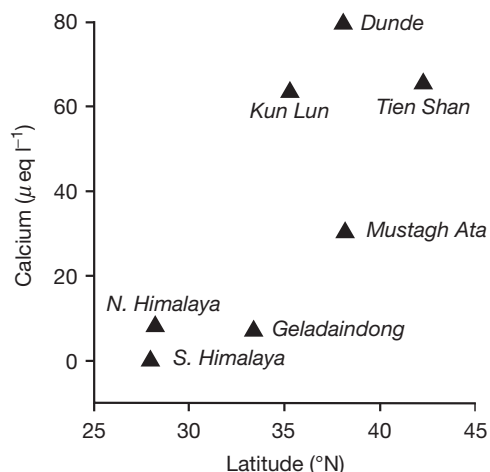


Figure 4 Variation of calcium concentrations (microequivalents per liter) in central Asian snow and ice with latitude. Details for data for each site provided in Wake et al. (1993, 2004). Note the very low calcium concentration on the southern slopes of the Himalayas (S. Himalayas) and the very high concentrations in the northern regions of the Tibetan Plateau.

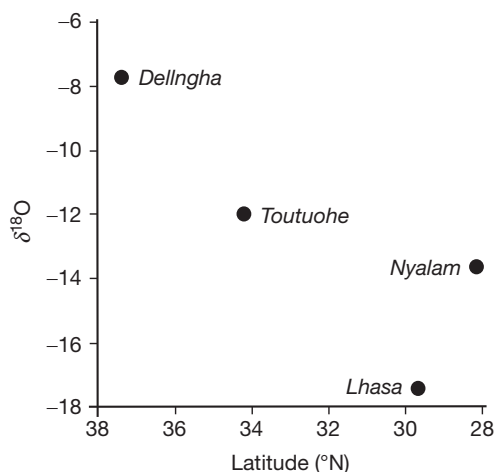


Figure 5 Variation in mean $\delta^{18}\text{O}$ values in snow and precipitation by latitude from the northern Tibetan Plateau to the Himalayas. Period of record for each site ranges from 4 to 9 years over the time period 1991–99 (Tian et al., 2003). Note how the more southern locations have the most negative $\delta^{18}\text{O}$ value. This highlights the influence of the effect on $\delta^{18}\text{O}$ in precipitation in the Himalayas and the southern half of the Tibetan Plateau.

concentrations measured in ice cores from the Dunde ice cap primarily reflect the influx of dust from the large desert basins (e.g., Taklamakan and Qaidam deserts) on the northern margin of the Tibetan Plateau. The glacial–interglacial transition is marked by a significant shift in many of the ice-core records. Holocene ice is characterized by a decrease in fine particles (<2 mm diameters), less negative $\delta^{18}\text{O}$, and an increase in soluble anions (chloride, nitrate, and sulfate). Enhanced dust deposition during the Late Glacial stage may have resulted from increased wind strength during this period, consistent with data from polar ice cores. The less negative $\delta^{18}\text{O}$ values suggest a shift to a warmer climate in the early Holocene, also consistent with other paleoclimate data. The increasing concentration of anions in the Dunde core during the Holocene suggests an increase of soluble aerosol originating from the Qaidam Basin, possibly a result of the formation and exposure of extensive evaporite deposits in the region, reflecting less precipitation and/or more evaporation in the late Pleistocene/early Holocene, although this cannot be verified by ice core accumulation time series as annual layers in the ice core could not be counted till 4.55 ka. Two of the three cores also show a trend to less negative $\delta^{18}\text{O}$ values over the last 60 years, suggesting a recent warming trend.

In 1992, a 309-m ice core was recovered from the Guliya Ice Cap, located in the West Kunlun Mountains (Thompson et al., 1997). Much of the ice recovered in the core appears to have accumulated since the beginning of the Pleistocene glaciation and provides details of the last glacial cycle in the western regions of the Tibetan Plateau. The $\delta^{18}\text{O}$ record from the Guliya core reveals a sequence of stadial and interstadial events similar to those recorded in polar ice cores and suggests that climate in this region was cool and variable during the last glacial cycle. In addition, between 25 and 33 ka, the $\delta^{18}\text{O}$ record contains abrupt high amplitude changes with an average length of 200 years; less negative $\delta^{18}\text{O}$ signals are associated with higher levels of dust, ammonium, and nitrate. These events are much shorter than the Dansgaard–Oeschger oscillations and may be related to short-term changes in snow-cover on the Tibetan Plateau. The transition from glacial to interglacial climate recorded in the Guliya core is not as sharp as the transition in polar ice cores, suggesting a slower change from glacial to interglacial climate in western China. Investigation of the stable isotope and calcium records suggests that the coldest periods in both the Guliya and Dunde ice cores occurred between AD 1000 and 1200. As was found in the Dunde ice core, the Guliya ice core revealed that the Little Ice Age was not as cold as expected. The Guliya $\delta^{18}\text{O}$ record also shows a trend toward less negative values over the last 200 years, suggestive of a region-wide warming trend (Thompson et al., 2003) and consistent with instrumental records of warming over the Tibetan Plateau over the past five decades (Liu and Chen, 2000).

Since 1997, many ice cores have been recovered from glaciers across central Asia. The locations of these new cores ranges from the eastern Himalayas to the central region of the Tibetan Plateau to the Altay in the north and to the Tien Shan in the northwest (Figure 3 and Table 2). This recent set of cores primarily provides records over the late Holocene; however, the relatively high accumulation rates allow for the development of high-resolution (i.e., annual to decadal) records of climate change over a broad region in central Asia.

Table 2 Chronological summary of ice cores (>70 years of record) that have been recovered from central Asian glaciers

Glacier	Elevation (m)	Latitude (N)	Longitude (E)	Maximum core length (m)	Age (years)	Year drilled	References
Dunde	5325	38.10	96.40	139	10 000	1987	Thompson et al. (1989)
Guliya	6710	35.28	81.48	309	100 000+	1992	Thompson et al. (1997)
Dasuopu	7200	28.38	85.71	168	1000	1997	Thompson et al. (2000)
Far East Rongbuk	6500	28.10	86.96	41	182	1997	Wake et al. (2001)
East Rongbuk I	6500	28.02	86.97	80	153	1998	Qin et al. (2002)
Malan	5680	90.67	35.83	102	112	1999	Wang et al. (2003)
Inilcheck	5100	42.27	80.42	167	100	2000	Kreutz et al. (2003)
Puruogangri	5970	33.90	89.17	215	7000	2000	Thompson et al. (2006)
Belukha I	4062	49.80	86.58	140	750	2001	Olivier et al. (2003, 2004, 2006)
Belukha II	4115	49.80	86.57	171	1000?	2002	Joswiak (2008)
East Rongbuk II	6518	28.03	86.97	108	1500	2002	Kaspari et al. (2007, 2008, 2009a,b)
Muztagata	7010	38.28	75.10	50	90	2003	Zhao et al. (2011)
Geladaindong	5720	33.58	91.18	147	70	2005	Grigholm et al. (2009) and Kang et al. (2010)
Tanggula	5746	33.12	92.09	131	??	2005	Zheng et al. (2010)

Himalayas

In the late 1990s, several cores were recovered from the central Himalayas in the Mt. Everest and Xixabangma regions (Figure 3 and Table 2) and provide records extending back hundreds of years. Cores recovered from the FER glacier in 1997 and the East Rongbuk (ER) glacier in 1998 have been analyzed for oxygen isotopes and a complete suite of major ions. Depth-age relationships were developed via counting of annual layers in the glaciochemical records, combined with identification of the 1963 and 1954 peaks in beta activity. Comparison of the two records suggests that the last 15 years of the FER may have been compromised as a result of increased ablation that coincides with a reduction in snow accumulation during the 1960s in the region (Qin et al., 2002; Thompson et al., 2000) and a rapid retreat of the ER glacier over the last 30 years (Qin et al., 2000). An increasing trend in sulfate, nitrate, calcium, and magnesium is apparent from the ER core beginning in the 1940s. The dominance of dust-related sources of major ions in this region (Wake et al., 2004) suggests that this increase reflects changes in atmospheric circulation and/or desertification in the region. The ER core also shows a doubling of ammonium over the last 50 years that likely reflects changes in anthropogenic activities, including enhanced agricultural activities in the region and atmospheric acidification (Hou et al., 2003; Kang et al., 2002c). Increasing temperatures in the southern regions of the Tibetan Plateau over the past several decades (e.g., Shrestha et al., 1999) may also have contributed to an increase in natural biological emissions of ammonia. Major ion records from both the ER and FER cores have been used to track interannual to decadal changes in atmospheric circulation in central Asia (Kang et al., 2002b; Wake et al., 2001). Increases in dust deposition over the past several decades in both cores record an intensification of winter circulation in central Asia.

More recently, a 1500-year record of the Asian monsoon has been developed from multiparameter analysis of a 108-m core on the ER glacier (Kaspari et al., 2007). Via comparison to NOAA National Centers for Environmental Prediction (NCEP) surface pressure records over the past 50 years, increases in calcium and decreases in sodium deposition over the last 400

years have been linked to an increase in the influence of continental air masses and decrease in the influence of marine air masses in the Himalayas. This, combined with a decrease in annual accumulation and increase in δD over the same time period, suggests a southward shift in the mean summer position of the inter tropical convergence zone over the past 400 years. High-resolution records of major and trace elements from the ER core show that concentrations and enrichment factors for bismuth, uranium, cesium, sulfur, and calcium have increased over recent decades, likely a result of increase in industrial activities and land use changes (Kaspari et al., 2009b). The major and trace element analyses also provide a record of variability in the amount and composition of dust deposited on the glacier (Kaspari et al., 2009a). Significant correlations of Everest dust concentrations with Total Ozone Mapping Spectrometer aerosol index during the winter (December, January, February) suggest that the dominant winter sources of dust are the Arabian Peninsula, Thar Desert, and northern Sahara. While the record was examined for recent increases in dust deposition associated with anthropogenic activities, no conclusive relationship was found. Temporal variations in the number and size of insoluble particles over the past 500 years were used to examine a link among the NAO (a proxy for strength of westerly circulation), precipitation and accumulation in central Asia, and dust deposition during the spring (Xu et al., 2009). Hong et al. (2009) developed an 800-year record from the ER core of select trace elements that show pronounced increase in levels of arsenic, molybdenum, tin, and antimony over the past century, which they link to rising levels of atmospheric pollution in response to rapid economic growth in central Asia.

A 168-m core recovered from Dasuopu glacier on Mt. Xixabangma in 1997 provides a record extending back 1000 years (Thompson et al., 2000). Peaks in the dust and chloride records in 1876–77 and 1790–96 are thought to represent reductions in the intensity of the summer monsoon that resulted in devastating drought across much of India and considerable loss of life. The 1876 signal was also observed in the ER core (Hou et al., 2003). The Dasuopu core also shows recent increases in dust, nitrate and, sulfate that are attributed to increased aridity, increased agricultural activity, and/or

increasing temperatures in the region. The oxygen isotope record also shows a trend to less negative values over the past 200 years. Based on comparison of this increasing trend with similar trends in oxygen isotope records from the Guliya and Dundee cores, and with similar trends in Northern Hemisphere temperature records, Thompson et al. (2000, 2003) have interpreted the Dasuopo oxygen isotope record as a proxy for increasing temperatures in the region (Figure 6). However, observational and modeling studies focused on the $\delta^{18}\text{O}$ signal in Himalayan precipitation have repeatedly concluded that the amount effect exerts significant control on the seasonal variation in stable isotope ratios in precipitation in the region, with the most isotopically depleted precipitation (and therefore most negative $\delta^{18}\text{O}$ values) in summer monsoon-derived precipitation (Araguas-Araguas et al., 1998; Hoffman and Heinmann, 1997; Tian et al., 2003). These studies also show that the temporal variability in stable isotopes in Himalayan precipitation is influenced by a complex set of processes involving changes in the sources of moisture, changes in atmospheric transport histories, and changes in temperature. This will require additional field and modeling studies to better decipher the climatic signals preserved in stable isotope records from Himalayan glaciers. The increasing trend in $\delta^{18}\text{O}$ recorded in Himalayan cores (Dasuopo, ER) over the past several decades does, however, indicate that rising temperatures in the region identified from instrumental data (Shrestha et al., 1999) are, not surprisingly, linked to changes in the regional hydrological cycle. Davis et al. (2005) investigated relationships among recent snow accumulation and $\delta^{18}\text{O}$ records from the Dundee and Dasuopo, the NAO, monsoon-derived precipitation records, and Northern Hemisphere temperature records. They concluded that increased snow accumulation during the latter stages of the Little Ice Age in central Asia was influenced by North Atlantic and Eurasian sources of moisture, while there has been a greater tropical influence on snow accumulation since 1880.

Central Tibetan Plateau

Preliminary results are available from several of these ice cores. The upper 24 m of the 102 m Malan ice core has been dated, providing a 112-year stable isotope record that reflects changes in summertime temperatures in the region (Wang et al., 2003).

Comparison of the stable isotope record with indices representing the NAO and SO suggests that warm season air temperatures over the Tibetan Plateau are influenced by the phase and strength of NAO and SO, most likely via changes in the strength of westerlies and the Indian monsoon, although the lack of any supporting analysis of regional meteorological data makes the results speculative.

New ice cores recovered from the Puruogangri Ice Cap in the central region of the Tibetan Plateau provide a record extending back through the mid-Holocene (Thompson et al., 2006). The record since ~1920 suggests a warmer and wetter climate in the central region of the Tibetan Plateau. The longer term record shows several events of high dust and major ions over the past 7000 years and a significant increase in calcium at 1200 years before present. The latter likely reflects local environmental changes. Analysis of changes in accumulation recorded in ice cores on a southwest-to-northeast transect across the Tibetan Plateau from the eastern Himalayas to the Dundee ice cap show a broadly consistent regional pattern of gradual decrease in accumulation over two to three centuries followed by an increase beginning in the late 1800s/early 1900s, and then a decrease in the 1990s (Kaspari et al., 2008).

A 70-year soluble ice-core dust record from the north side of Mt. Geladaindong in the central Tibetan Plateau suggests a teleconnection between dust deposition in central Asia and the Pacific Decadal Oscillation (PDO), as a threefold decrease in dust concentrations coincides with the 1976–77 shift in the PDO (Grigholm et al., 2009). Kang et al. (2010) linked the shift in dust deposition in the 1970s recorded in the Geladaindong core to a strengthening of atmospheric circulation over the Tibetan Plateau.

A 131 m surface-to-bedrock core was recovered from Mt. Tanggula (adjacent to Mt. Geladaindong) in 2005. The upper 14.5 m spans a period from 1950 to 2004 shows an increase in sulfate, nitrate, and ammonium that are attributable to anthropogenic emissions and increasing temperatures enhancing biological activity (Zheng et al., 2010).

Altay, Pamirs, and Tien Shan

The upper 86 m of a 140 m core from the Belukha glacier in the Altai provides a record extending back two centuries (Olivier et al., 2003, 2006). Ammonium and formate time series

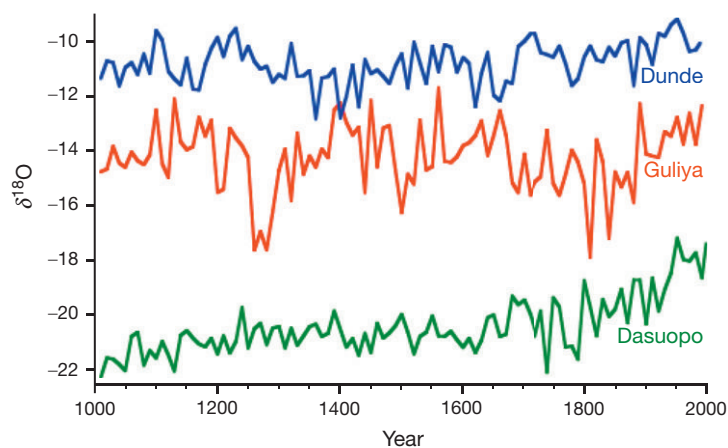


Figure 6 Decadal averages of $\delta^{18}\text{O}$ values for the last millennium from three Tibetan ice cores (Dunde, Guliya, and Dasuopo; Thompson et al., 2003).

provide a record of Siberian forest fire activity (no trends). Sulfate shows a sharp increase after 1950 that peaked in 1970, and then decreased during the 1980s, reflecting variations in sulfur dioxide emissions in Siberia and Kazakhstan. Rising nitrate concentrations since 1960 reflect emissions from motor vehicles and/or enhanced fertilizer production, while rising ammonium since 1950s is related to enhanced agricultural activity. Analysis of large samples from the upper portions of the core has provided a record of global fallout of plutonium over the past five decades (Olivier et al., 2004). Henderson et al. (2006) found no long-term trends in accumulation in the Belukha ice-core record, but did suggest that the two temperature proxies ($\delta^{18}\text{O}$ and melt percent profiles) show strong warming over the past century. Okamoto et al. (2011) developed a record of summer temperature via analysis of melt features and annual accumulation on the upper 48 m of a second ice core retrieved from Mt. Belukha in the Siberian Altai mountains. The ice-core record shows substantial increase in summertime temperatures that began in the 1980s.

Three ice cores have been recovered from the Inylchek glacier in the Tien Shan in Kyrgyzstan. Preliminary analysis shows significant interannual variability in dust deposition during the 1990s as well as a strong anthropogenic ammonium signal, while deuterium excess displays a strong seasonal cycle related to seasonal changes in the moisture sources, transport history, and moisture recycling in the Caspian/Aral Sea region (Kreutz et al., 2003). Analyses of major element, rare earth element, and stable isotopes ($\delta^{18}\text{O}$ and $\delta^{34}\text{S}$) indicate three distinct sources for dust (loess, calcium carbonate, and gypsum) and two distinct sources for sulfur (evaporite deposits and anthropogenic emissions; Kreutz and Sholkovitz, 2000; Preutt et al., 2004). Analysis of the amount of summer melt in the ice core (as determined by the amount of clear ice) indicates a significant increase in surface melting over the past few decades, consistent with observations of retreating glaciers in the Tien Shan over the years.

A 90-year record of major ions developed from a 55 m core recovered from Mt. Mustagata in the Pamirs shows an increase in sulfate and nitrate since the late 1970s suggesting an anthropogenic source (Zhao et al., 2011). Analysis of heavy metals on a separate core from Mustagata shows an increase in antimony and bismuth in the 1960s, followed by a decrease in the late 1990s (Li et al., 2006). These trends were linked to changes in anthropogenic activities in central Asia.

Preliminary results from glaciochemical analysis of snowpit and shallow core samples from the Fedchenko glacier in the Pamir (38.25° N, 72.25° E) show well-defined seasonal layering appearing in stable isotope and trace element content, suggesting the potential for developing a deep core from this site in the future (Aizen et al., 2009).

Final Comments

Analysis of the chemical and physical characteristics of ice cores recovered from several different glaciers in the mountains of central Asia over the past 25 years have significantly improved our understanding of climatic and environmental change in the region. These ice-core records have documented oscillations in the climate system on several different

timescales, identified abrupt changes in the climate system, revealed recent warming trends and changes in the hydrological cycle, and documented the influence of anthropogenic emissions on precipitation chemistry. Several new ice cores have been recovered over the past decade that provide additional details on the spatial and temporal variability of the Asian monsoon and the westerly jet stream, and additional records detailing the anthropogenic influence on atmospheric chemistry in this remote region. Despite this, there remain considerable gaps in our understanding of the regional details of climate change in this vast area. Detailed synthesis of existing records (e.g., Kaspari et al., 2008) are required to better understand the spatial and temporal changes in the environment and climate of this vast region. Additional ice cores will be required to fill in these details. However, the recent warming across much of the region, recorded in instrumental and ice-core records and evidenced by retreating glaciers, indicates that time is of essence. Many of the unique archives locked up in central Asian glacier ice appear destined to disappear in a world warmed by anthropogenic greenhouse gases.

See also: **Glaciations: Late Quaternary in Highland Asia. Ice Core Methods: Borehole Temperature Records; Chronologies; Glaciochemistry; Overview; Stable Isotopes.**

References

- Aizen VB, Mayewski PA, Aizen EM, et al. (2009) Stable-isotope and trace element time series from Fedchenko glacier (Pamirs) snow/firn cores. *Journal of Glaciology* 55(190): 275–291.
- Araguas-Araguas L, Froehlich K, and Rozanski K (1998) Stable isotope composition of precipitation over southeast Asia. *Journal of Geophysical Research* 103(22): 28721–28742.
- Davis ME, Thompson LG, Yao T, and Wang N (2005) Forcing of the Asian monsoon on the Tibetan plateau: Evidence from high resolution ice core and tropical coral records. *Journal of Geophysical Research* 110: D04101. <http://dx.doi.org/10.1029/2004JD004933>.
- Grigholm B, Mayewski PA, Kang S, et al. (2009) Atmospheric soluble dust records from a Tibetan ice core: Possible climate proxies and teleconnection with the Pacific Decadal Oscillation. *Journal of Geophysical Research* 114: D20118. <http://dx.doi.org/10.1029/2008JD011242>.
- Haeberli W, Bosch H, Scherler K, Ostrem G, and Wallen CC (1989) *World Glacier Inventory: Status 1988*. Wallingford, England: IAHS – UNEP – UNESCO (compiled by the World Glacier Monitoring Service).
- Henderson KA, Laube A, Gäggeler HW, Olivier S, Papina T, and Schwikowski M (2006) Temporal variations of accumulation and temperature during the past two centuries from Belukha ice core, Siberian Altai. *Journal of Geophysical Research* 111: D03104. <http://dx.doi.org/10.1029/2005JD005819>.
- Hoffman G and Heinmann M (1997) Water isotope modeling in the Asian monsoon region. *Quaternary International* 37: 115–128.
- Hong S, Lee K, Hou S, et al. (2009) An 800-year record of atmospheric As, Mo, Sn, and Sb in central Asia in high-altitude ice cores from Mt. Qomolangma (Everest), Himalayas. *Environ Sci Technol* 43(21): 8060–8065.
- Hou S, Qin D, Zhang D, Kang S, Mayewski PA, and Wake CP (2003) A 154 a high resolution ammonium record from the Rongbuk Glacier, north slope of Mt. Qomolangma (Everest), Tibet–Himal region. *Atmospheric Environment* 37(5): 721–729.
- Joswiak DR (2008) Changes in atmospheric chemical composition determined from ice core records in southwestern Siberia during the 20th century. PhD Thesis, University of Idaho.
- Kang S, Kreutz KJ, Mayewski PA, Qin D, and Yao T (2002a) Stable-isotopic composition of precipitation over the northern slope of the central Himalaya. *Journal of Glaciology* 48(163): 519–526.

- Kang S, Mayewski PA, Qin D, et al. (2002b) Glaciochemical records from a Mt. Everest ice core: Relationship to atmospheric circulation over Asia. *Atmospheric Environment* 36: 3351–3361.
- Kang S, Mayewski PA, Qin D, et al. (2002c) Twentieth century increase of atmospheric ammonia recorded in Mt. Everest ice core. *Journal of Geophysical Research* 107(21D): 4595.
- Kang S, Zhang Y, Zhang Y, et al. (2010) Variability of atmospheric dust loading over the central Tibetan Plateau based on ice core glaciochemistry. *Atmospheric Environment* 44: 2980–2989. <http://dx.doi.org/10.1016/j.atmosenv.2010.05.014>.
- Kaspari S, Hooke R, Mayewski P, Kang S, Hou S, and Qin D (2008) Snow accumulation rate on Mt. Everest: Synchronicity with sites across the Tibetan Plateau on 50–100 year timescales. *Journal of Glaciology* 54(185): 343–352.
- Kaspari S, Mayewski PA, Handley M, et al. (2007) Reduction in northward incursions of the South Asian Monsoon since 1400 AD inferred from a Mt. Everest ice core. *Geophysical Research Letters* 34: L16701. <http://dx.doi.org/10.1029/2007GL030440>.
- Kaspari S, Mayewski PA, Handley MJ, et al. (2009a) A high-resolution record of atmospheric dust variability and composition since 1650 AD from a Mt. Everest ice core. *Journal of Climate* 22: 3910–3925.
- Kaspari S, Mayewski PA, Handley MJ, et al. (2009b) Recent increases in atmospheric concentrations of Bi, U, Cs, Ca and S from a 350-Year Mt. Everest ice core record. *Journal of Geophysical Research*. <http://dx.doi.org/10.1029/2008JD011088>.
- Kreutz KJ and Sholkovitz ER (2000) Major element, rare earth element, and sulfur isotopic composition of a high-elevation firn core: Sources and transport of mineral dust in Central Asia. *Geochemistry, Geophysics, Geosystems* 1: 2000GC000082.
- Kreutz KJ, Wake CP, Aizen VB, Cecil LD, and Synal HA (2003) Seasonal deuterium excess in a Tien Shan ice core: Influence of moisture transport and recycling in Central Asia. *Geophysical Research Letters* 30(18): 1922. <http://dx.doi.org/10.1029/2003GL017896>.
- Li Y, Yao T, Wang N, et al. (2006) Recent changes of atmospheric heavy metals in a high elevation ice core from Muztagh Ata in East Pamirs: Initial results. *Annals of Glaciology* 43: 154–159.
- Liu X and Chen B (2000) Climatic warming in the Tibetan Plateau during recent decades. *International Journal of Climatology* 20: 1729–1742.
- Mayewski PA, Lyons B, Ahmad N, Smith G, and Pourchet M (1984) Interpretation of the chemical and physical time-series retrieved from Sentik Glacier, Ladakh Himalaya, India. *Journal of Glaciology* 30(104): 66–76.
- Okamoto S, Fujita K, Narita H, et al. (2011) Reevaluation of the reconstruction of summer temperatures from melt features in Belukha ice cores, Siberian Altai. *Journal of Geophysical Research: Atmospheres* 116: D2.
- Olivier S, Bajo S, Fifield LK, et al. (2004) Plutonium from global fallout recorded in an ice core from the Belukha Glacier, Siberian Altai. *Environmental Science and Technology* 38(24): 6507–6512. <http://dx.doi.org/10.1021/es0492900>.
- Olivier S, Blaser C, Brütsch S, et al. (2006) Temporal variations of mineral dust, biogenic tracers, and anthropogenic species during the past two centuries from Belukha ice core, Siberian Altai. *Journal of Geophysical Research: Atmospheres* 111: D5.
- Olivier S, Schwikowski M, Brütsch S, et al. (2003) Glaciochemical investigation of an ice core from Belukha Glacier, Siberian Altai. *Geophysical Research Letters* 19: <http://dx.doi.org/10.1029/2003GL018290>.
- Pruett LE, Kreutz KJ, Wadleigh M, and Aizen V (2004) Assessment of Sulfate Sources in High-Elevation Asian Precipitation Using Stable Sulfur Isotopes. *Environmental Science and Technology* 38: 4728–4733.
- Qin D, Hou S, Zhang D, et al. (2002) Preliminary results from the chemical records of an 80.4 m ice core recovered from the East Rongbuk Glacier, Mt. Qomolangma (Everest). *Annals of Glaciology* 35: 278–284.
- Qin D, Mayewski PA, Wake CP, et al. (2000) Evidence for recent climate change from ice cores in the central Himalaya. *Annals of Glaciology* 31: 153–158.
- Shrestha AB, Wake CP, Dibb JE, and Whitlow SI (2002) Aerosol and precipitation chemistry at a remote Himalayan site in Nepal. *Aerosol Science and Technology* 36: 441–456.
- Shrestha A, Wake CP, Mayewski PA, and Dibb J (1999) Maximum temperature trends in the Himalaya and vicinity: An analysis based on temperature records from Nepal for the period 1971–94. *Journal of Climate* 12: 2775–2786.
- Thompson LG, Mosley-Thompson E, Davis ME, et al. (1989) Holocene–Late Pleistocene climatic ice core records from Qinghai–Tibetan Plateau. *Science* 246: 474–477.
- Thompson LG, Mosley-Thompson E, Davis ME, et al. (2003) Tropical glacier and ice core evidence of climate change on annual to millennial time scales. *Climatic Change* 59: 137–155.
- Thompson LG, Yao T, Davis M, et al. (2006) Holocene climate variability archived in the Puruogangri ice cap on the central Tibetan Plateau. *Annals of Glaciology* 43: 61–69. <http://dx.doi.org/10.3189/172756406781812357>.
- Thompson LG, Yao T, Davis ME, et al. (1997) Tropical climate instability: The last glacial cycle from a Qinghai–Tibetan ice core. *Science* 276: 1821–1825.
- Thompson LG, Yao T, Mosley-Thompson E, Davis ME, Henderson BA, and Lin P-N (2000) A high-resolution millennial record of the South Asian Monsoon from Himalayan ice cores. *Science* 289: 1916–1919.
- Tian L, Yao T, Schuster PF, et al. (2003) Oxygen-18 concentrations in recent precipitation and ice cores on the Tibetan Plateau. *Journal of Geophysical Research* 108(D9): 4293. <http://dx.doi.org/10.1029/2002JD002173>.
- Wake CP, Mayewski PA, and Kang S (2004) Climatic interpretation of the gradient in glaciochemical signals across the crest of the Himalaya. In: Cecil LD, Green JR, and Thompson L (eds.) *Earth Paleoenvironments: Records Preserved in Mid- and Low-Latitude Glaciers. Developments in Paleoenvironmental Research*, vol. 9, pp. 81–94. Dordrecht: Kluwer.
- Wake CP, Mayewski PA, Qin D, et al. (2001) Changes in atmospheric circulation over the south-eastern Tibetan Plateau over the last two centuries from a Himalayan ice core. *PAGES News* 9(3): 14–16.
- Wake CP, Mayewski PA, Wang P, Yang Q, Han J, and Xie Z (1992) Anthropogenic sulfate and Asian dust signals in snow from Tien Shan, northwest China. *Annals of Glaciology* 16: 45–52.
- Wake CP, Mayewski PA, Xie Z, Wang P, and Li Z (1993) Regional variation of monsoon and desert dust signals recorded in Asian glaciers. *Geophysical Research Letters* 20: 1411–1414.
- Wang N, Thompson LG, Davis ME, Mosley-Thompson E, Yao T, and Pu J (2003) Influence of variations in NAO and SO on air temperature over the northern Tibetan Plateau as recorded by $\delta^{18}\text{O}$ in the Malan ice core. *Geophysical Research Letters* 30(22): 2167. <http://dx.doi.org/10.1029/2003GL018188>.
- Watanabe O, Takenaka S, Iida H, Kamiyama K, Thapa K, and Mulmi D (1984) First results from Himalayan glacier boring project in 1981–1982. *Bulletin of Glacier Research* 2: 7–23.
- Wushiki H (1977) Deuterium content in Himalayan precipitation at Khumbu district observed in 1977/1975. *Seppyo* 39: 50–56.
- Xu J, Hou S, Qin D, et al. (2009) An 108.83 m ice core record of atmospheric dust deposition at Mt. Qomolangma (Everest), central Himalayas. *Quaternary Research* 73: 33–38.
- Yao T, Thompson LG, Mosely-Thompson E, Yang Z, Wang Y, and Lin P-N (1996) Climatological significance of $\delta^{18}\text{O}$ in north Tibetan ice core. *Journal of Geophysical Research* 101: 29531–29537.
- Zhao H, Xu B, Yao T, Tian L, and Li Z (2011) Records of sulfate and nitrate in an ice core from Mount Muztagata, central Asia. *Journal of Geophysical Research: Atmospheres* 116: D13.
- Zheng W, Yao T, Joswiak DR, Xu B, Wang N, and Zhao H (2010) Major ions composition records from a shallow ice core on Mt. Tanggula in the central Qinghai–Tibetan Plateau. *Atmospheric Research* 97: 70–79.

South America

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Introduction

Tropical and subtropical glaciers are restricted to high mountains in the Andes, on the Tibetan Plateau, in East Africa, and in a few isolated locations. Currently, ~500 000 km² of the world's ice is frozen in glaciers and ice sheets outside the polar regions (Benn and Evans, 1998). Although the tropics (30° N to 30° S) contain only a fraction of all the Earth's ice, they encompass 50% of the Earth's surface area and are home to 70% of the world's inhabitants. In regions such as the Andes of South America, populations are critically dependent on these glaciers, as they are the source of water for agriculture and hydroelectric power.

Many of the measurements made on polar ice cores are also performed on cores from the Andean mountain glaciers. Researchers can utilize an ever expanding ice-core database of multiple proxy information (i.e., stable isotopes, insoluble dust, major and minor ion chemistry, precipitation reconstruction) that spans the globe in spatial coverage and is of the highest possible temporal resolution. These parameters potentially can yield information on regional histories of variations in temperature, precipitation, moisture source, aridity, vegetation changes, etc. Many of these physical and chemical constituents have wet and dry/cold and warm seasonal signals in the ice, which allow the years to be counted, much as tree rings.

The ice-core records from the Earth's Alpine ice caps and glaciers provide a wealth of data that contribute to a spectrum of critical scientific questions. These range from the reconstruction of high-resolution climate histories to help explore the oscillatory nature of the climate system, to the timing, duration, and severity of abrupt climate events, to the relative magnitude of twentieth-century global climate change and its impact on the cryosphere. The longest of these records have revealed the nature of climate variability since the Last Glacial Maximum (LGM), 18 000 to 20 000 years ago, and even beyond. The more recent sections, which are of annual and even seasonal resolution, have yielded data on the temporal variations in the occurrence and intensity of coupled ocean-atmosphere phenomena of global significance such as El Niño, which are most strongly expressed in the tropics. This is particularly valuable information since meteorological observations from much of this region are scarce and of short duration.

The History of Ice-Core Drilling in the Andes

Glaciological and ice-coring programs have been conducted throughout the Andes for over three decades (Table 1). Figure 1 illustrates the sites in Peru and Bolivia from where the longest ice-core records have been obtained. The earliest efforts to retrieve climate records from mountain glaciers included programs from 1974 to 1979 involving surface sampling on the

Quelccaya ice cap (13° 56'S, 70° 50'W; 5,670 m above sea level, or m a.s.l.) in southern Peru (Thompson et al., 1979). The results from this preliminary research, conducted by the Institute of Polar Studies (now the Byrd Polar Research Center, or BPRC) at The Ohio State University, paved the way for the first tropical high-altitude deep-drilling program on Quelccaya in 1983, which recovered two cores to bedrock that yielded precisely-dated, annually-resolved 1500-year climate records (Thompson et al., 1985; 1986). The Quelccaya program marked a series of 'firsts' in the history of ice-core drilling, in that it produced the first long high-altitude cores from the tropics, and was the first in which solar power was exclusively used for drilling (Figure 2).

The 1990s was a decade of increased ice drilling activity in the central Andes. In 1993, two cores were drilled to bedrock in the col of Nevado Huascarán (9° 07'S, 77° 37'W, 6050 m a.s.l.) in the Cordillera Blanca range of northern Peru. The climate records from these cores extend back to 19 000 years to the end of the LGM, and provided the first glacier evidence of substantial ice-age cooling in the tropics (Thompson et al., 1995). Four years after the Huascarán program, the summit of Nevado Sajama (18° 06'S, 68° 53'W, 6540 m a.s.l.), an extinct volcano on the Bolivian Altiplano, was drilled. The two cores drilled to bedrock provided a 25 000-year climate history, which confirmed the ice-age cooling in the tropics that was seen in Huascarán (Thompson et al., 1998). The time series from Sajama was the first to be reconstructed from an ice core with the aid of radiocarbon (¹⁴C) dates. In 1999, the Institute of Research and Development (IRD) in France led an effort to recover two cores from the Illimani ice cap (16° 37'S, 67° 47'W, 6350 m a.s.l.), which like Huascarán and Quelccaya, is located along the eastern edge of the Andes. The record from this glacier also extends back to the end of the last glacial stage (Ramirez et al., 2003).

In 2003, two decades after it was first drilled, a second program on the Quelccaya ice cap resulted in the recovery of two cores to bedrock on adjacent domes. That same year three cores were retrieved from Nevado Coropuna, located in the Western Andean Cordillera (15° 32'S, 72° 39'W). Cores 1 and 2 were drilled on the crater rim at the summit, while Core 3 was drilled in the central crater (Table 1). These records are currently being developed.

The Physical and Meteorological Setting of Tropical South America

The topography and meteorology of tropical South America have strong influences on the climate record preserved in the Andean ice cores. Although the South American continent covers 68° of latitude between 12° N and 56° S and encompasses a variety of climatic regimes from tropical rainforest in

Table 1 Ice-cores sites in the Peruvian and Bolivian Andes

Site name	Location	Elevation(m asl)	Year drilled	Annual net accumulation m water eq.	Depth of boreholes (m)	Length of record (yrs)
Huascarán	9°06'S, 77°37'W	6050	1993	1.3	160, 166	19 000
Quelccaya	13°56'S, 70°50'W	5670	1983	1.4	155, 164	1500
			2003		170, 129	
Illimani ^a	16°37'S, 67°47'W	6350	1999	0.58	137, 139	18 000
Coropuna	15°32'S, 72°39'W	6425	2003	0.50	34, 34, 146	To be determined
Sajama	18°06'S, 68°53'W	6540	1997	0.44	132, 133	25 000

Synopsis of the major ice-core drilling programs in the South American Andes Mountains. All the cores listed were drilled and analyzed by the Byrd Polar Research Center at the Ohio State University with the exception of Illimani^a, which was drilled and analyzed for isotopes by the Institute of Research and Development (IRD) in France (Ramírez et al., 2003). Analysis of the Quelccaya and Coropuna cores drilled in 2003 are currently underway.

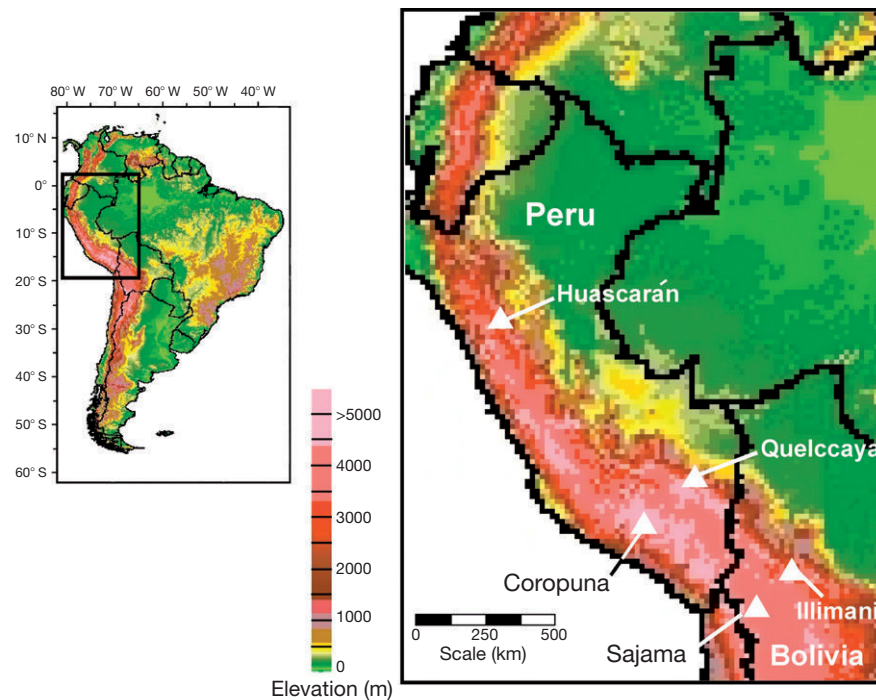


Figure 1 Map of South America, with the central Andes region expanded to show locations of sites in the Andes of South America where major ice-core drilling programs have been conducted.



Figure 2 The drilling operation on the summit of Quelccaya in 1983, during which solar panels were used to power the drill.

the north to polar tundra at its southern tip, its meteorology is simpler than expected for so large a landmass. Because it is much longer than it is wide and is virtually surrounded by oceans, it is highly affected by surrounding maritime air masses and pressure systems. The most prominent physical features include the Andes Mountain complex that runs along the length of the continent parallel to the west coast, the Amazon Basin which covers most of the central region north of 10°S, and the high (3500 to 4000 m a.s.l), arid Altiplano of southern Peru and Bolivia. From 15°S to 22°S, the Altiplano divides the Andes into the main range of the Cordillera Occidental to the west and the Cordillera Central to the east.

The Andes Mountains and the Amazon rainforest are especially vital influences on South American climates and environments. The extensive meridional length and the great altitude of the Andes are responsible for the division of

the subtropical high-pressure belt in the Southern Hemisphere into two major cells that flank the continent on both the Pacific and Atlantic sides. From December to February (the austral summer), these cells strengthen and move slightly southward (Figure 3a). A summertime thermal low, called the Chaco Low, develops in the center of the continent as an extension of the southward-dipping equatorial trough. During the austral summer, which is also the wet season in the Amazon Basin and along the eastern flanks of the Andes, the equatorial trough dips to 25° S and the northeasterly trade winds are channeled into the continental interior by the southern limb of the Azores high-pressure cell in the subtropical North Atlantic. The vegetation in the Amazon rainforest recycles moisture that is carried by the trade winds as they move across the continent to the Andes. An upper level closed anticyclone, the Bolivian High (not shown), also forms during

the summer as a result of latent heat release over the Amazon Basin (Lenters and Cook, 1997).

In the austral winter (June to August), the equatorial trough moves northward to around 10° N, and the circulation around the South Atlantic high-pressure system becomes more dominant (Figure 3b). The South Pacific and South Atlantic high-pressure cells weaken and move slightly to the north, and the high-pressure systems in the Northern Hemisphere strengthen and move northward along with the equatorial trough. Thus, moisture from the North Atlantic is effectively blocked to the northern interior of the continent. The Bolivian High and the Chaco Low disappear in the winter, and northerly winds diminish over the Amazon Basin and the northern and central Andes.

Subsiding air from the South Pacific high pressure system, along with a northward advecting cold water current along the coast, produce stable, dry conditions along the west coast from 30° S to 14° S. The result is the Atacama Desert of Northern Chile and Southern Peru, one of the driest places on Earth (Vuille and Ammann, 1997). South of ~20° S the prevailing westerly winds dominate upper and lower tropospheric circulation throughout the year, and they channel 'cut-off lows' (Vuille, 1999; Marengo and Rogers, 2000) that provide the coastal areas with a humid climate, while on the lee side of the mountains the land is dominated by the Patagonian desert in southern Argentina.

Precipitation over Tropical South America

The ultimate source for the precipitation that falls on the central and northern Andes is the subtropical North Atlantic Ocean (Taljaard, 1972; Grootes et al., 1989) via the Amazon Basin. Precipitation in the Amazon is strongly influenced by the strength and position of the Intertropical Convergence Zone/equatorial trough (ITCZ/ET) and by the Azores high-pressure system (Figure 3a), which advects maritime tropical air westward in a shallow, cool, and humid layer. In the austral summer, when the equatorial trough is at its southernmost position around 25° S over the continent, this shallow layer encounters the Chaco Low just to the east of the Andes (between 15 and 30° S and 60 to 68° W) and rises to ~4000 m a.s.l., resulting in humid, unstable air masses that diverge at the 500-hPa level. The upper level air flows westward over the Andes, releasing large amounts of precipitation through convection. The water vapor that travels from the Atlantic accounts for only part of the rainfall over the Amazon Basin (Marques et al., 1977), and much of the remainder is supplied by evapotranspiration of the vegetation over the rain forest canopy. Although November to April is the wet season in the southern tropics of South America, some snow also falls in the winter in the northern and central Andes, but the precipitation is sporadic and is the product of local orogenic effects.

Stable Isotopic Ratios of Oxygen ($\delta^{18}\text{O}$) in Andean Ice Cores

Isotopic ratios of oxygen ($\delta^{18}\text{O}$) and hydrogen (δD) in water are among the most widespread and important of the measurements made on ice cores (see Stable Isotopes). The information provided by $\delta^{18}\text{O}$ and δD is based on the fractionation of the oxygen and hydrogen atoms, respectively, into

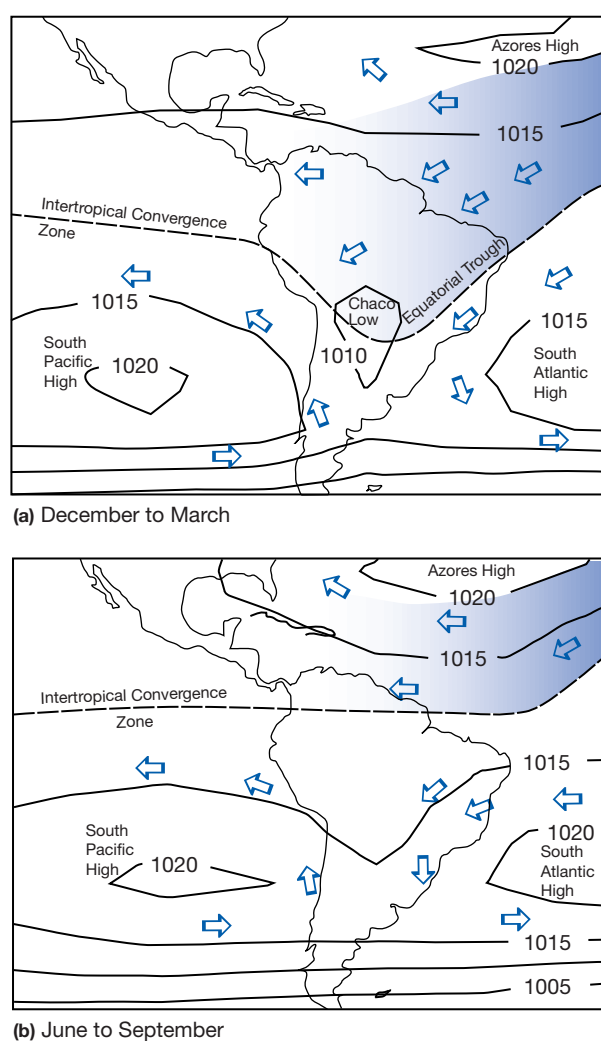


Figure 3 Major surface pressure cells and prevailing winds (shown by open arrows) over South America and the adjacent oceans for (a) December to March (the austral summer) and (b) June to September (the austral winter). Sea level pressure values are in hectopascals (hPa), the metric equivalent of millibars. The shading over the North Atlantic illustrates regions in which the northeasterly trade winds dominate the surface air circulation.

their light and heavy isotopes (^{16}O and ^{18}O , H and ^2H) and on the higher vapor pressure of H_2^{16}O over H_2^{18}O , and $^2\text{H}_2^{16}\text{O}$. Early work on these 'stable' (i.e., as opposed to unstable, or radioactive) isotopes in polar ice cores resulted in the development of transfer functions between the isotopic ratios and the water temperature at the moisture source and the atmospheric temperature as the water vapor condenses into precipitation (Dansgaard, 1964). The use of $\delta^{18}\text{O}$ as a temperature proxy for polar ice is now widely accepted, but is, however, still a source of controversy for low-latitude cores. Some who have studied the problem believe that it is precipitation, rather than temperature, that exerts the strongest influence on $\delta^{18}\text{O}$ in the tropics (Rozanski et al., 1993; Baker et al., 2001), while others contend that temperature is the dominant factor, at least over intra-annual and longer timescales (Thompson et al., 2003).

The relatively large annual accumulation rates and the high-resolution sampling of the ice cores from two of the Andean sites, Huascarán and Quelccaya, allow wet/dry season partitioning of the dust and oxygen isotopic records. Dust and other aerosols such as salts are concentrated in the dry season (austral winter) layers and, in contrast to polar ice-core records, the summer snow is depleted in ^{18}O (resulting in more negative $\delta^{18}\text{O}$ values). A model by Grootes et al. (1989) presents a possible mechanism for this summertime isotopic depletion, which involves biological (evapotranspiration) processes over the Amazon Basin and adiabatic cooling during the uplift of air when it comes in contact with the Andes.

The Atlantic influence on interannual variations in $\delta^{18}\text{O}$

The spatial distribution of sea surface temperature anomalies (SSTAs) in the western tropical Atlantic has an effect on the oxygen isotopic fractionation in precipitation via the strength of the 500-hPa zonal easterlies over the Amazon Basin (Henderson et al., 1999). More enriched isotopes in the Huascarán record correspond to lower Azores sea-level pressure (SLP), and they are linked through the influence of the strength of the NE trade winds that compose the southern limb of that high pressure cell. During austral summers in which the Atlantic ITCZ is weaker than average, the precipitation tends to be less ^{18}O -depleted (i.e., higher $\delta^{18}\text{O}$ values). In the equatorial Atlantic the austral summer SLPs are positively correlated with the wet season $\delta^{18}\text{O}$, meaning that more ^{18}O -enriched water vapor occurs with higher SLPs in this low-pressure zone. A weaker equatorial low implies weaker low-level convergence and less convective precipitation that is channeled into the Amazon Basin, leaving the isotopic composition of the water vapor over the rainforest more influenced by the unfractionated product of evapotranspiration and resulting in a trend toward ^{18}O enrichment.

All these mechanisms combine into a reasonable model that links the Azores high pressure system with the isotopic composition of the precipitation over the Andes between 9°S and 14°S . Weaker 500-hPa easterlies and reduced precipitation in the northern Amazon Basin and northern Peru tend to be consequences of strong El Niño Southern Oscillation (ENSO) events, and less isotopic fractionation of the prevailing air masses results in ^{18}O -enriched precipitation that ultimately falls on the Andes glaciers. Several studies in addition to Henderson et al. (1999) show that there is a tropical Atlantic

sea surface temperature analog to the ENSO patterns in the Pacific (Weare, 1977; Hastenrath, 1978; Lough, 1986; Servain and Legler, 1986; Servain, 1991).

Climate Records from Andean Glaciers

Millennial-Scale Variations

Three of the records from the Andes (Huascarán in northern Peru and Sajama and Illimani in Bolivia, Figure 4a) extend to the last glacial stage and confirm, along with other climate proxy records (Stute et al., 1995; Guilderson et al., 1994), that the LGM was much colder in the tropics and subtropics (Thompson et al., 1995; Thompson et al., 1998; Ramirez et al., 2003) than previously believed (CLIMAP, 1976). Although this period was consistently colder, it was not as consistently arid throughout the lower latitudes as it was in the polar regions. For example, the dust records from Huascarán and Sajama (Figure 4b) indicate that the effective moisture (precipitation minus evaporation) along the axis of the Andes Mountains during the LGM was variable, being much drier in the north (as suggested by the high dust concentrations in the early part of the Huascarán record) than in the Altiplano region in the central part of the range (shown by the low dust concentrations in the Sajama ice-core record).

Ice cores from the Andes can also contribute to our understanding of past environmental and climatic conditions in the Amazon Basin. The extent of biological activity in the Amazon rainforest during the LGM is controversial (Colinvaux et al., 2000), and the nitrate concentration record from the Huascarán ice core (Figure 4c) has been cited in the argument (Thompson et al., 1995). Pollen studies from the Amazon Basin suggest that the extent of the rainforest has not changed much between the LGM and the Holocene, however, proponents of the 'refugia' theory (Clapperton, 1993) assert that the cold, dry climate in the tropics caused a major retreat of the rainforest flora into small, geographically isolated areas, leaving most of the Basin covered by grasslands. In the Huascarán record, the nitrate concentration profile is similar to that of the $\delta^{18}\text{O}$ throughout most of the record (Figures 4a, and c), and the very low concentrations of nitrate that are concurrent with very depleted $\delta^{18}\text{O}$ suggest that biological activity upwind of the Cordillera Blanca was impeded by the cold and dry climate $\sim 19\,000$ years ago. Nitrate variations in the Sajama ice core record also reflect changes in effective moisture, however, the link is through aerosols derived from the numerous salars (salt deposits) on the Altiplano, rather than just through biological emissions.

Many of the deep cores from the low latitudes that extend through the Holocene show spatial variations in climate, even between records from the same region. For example, the Holocene stable isotope from Huascarán and Illimani, while similar to each other in that they show early Holocene isotopic enrichment, are different from the Sajama isotopic record, which shows little trend through the last 10 000 years (Figure 4a). Although Sajama and Illimani are geographically close to each other, they lie on opposite sides of the Andes Mountains, with Illimani located in the eastern range. Like Huascarán far to the north, it has received most of its precipitation from the northeast after it has been recycled through the Amazon Basin. Sajama, which is located on the high, dry

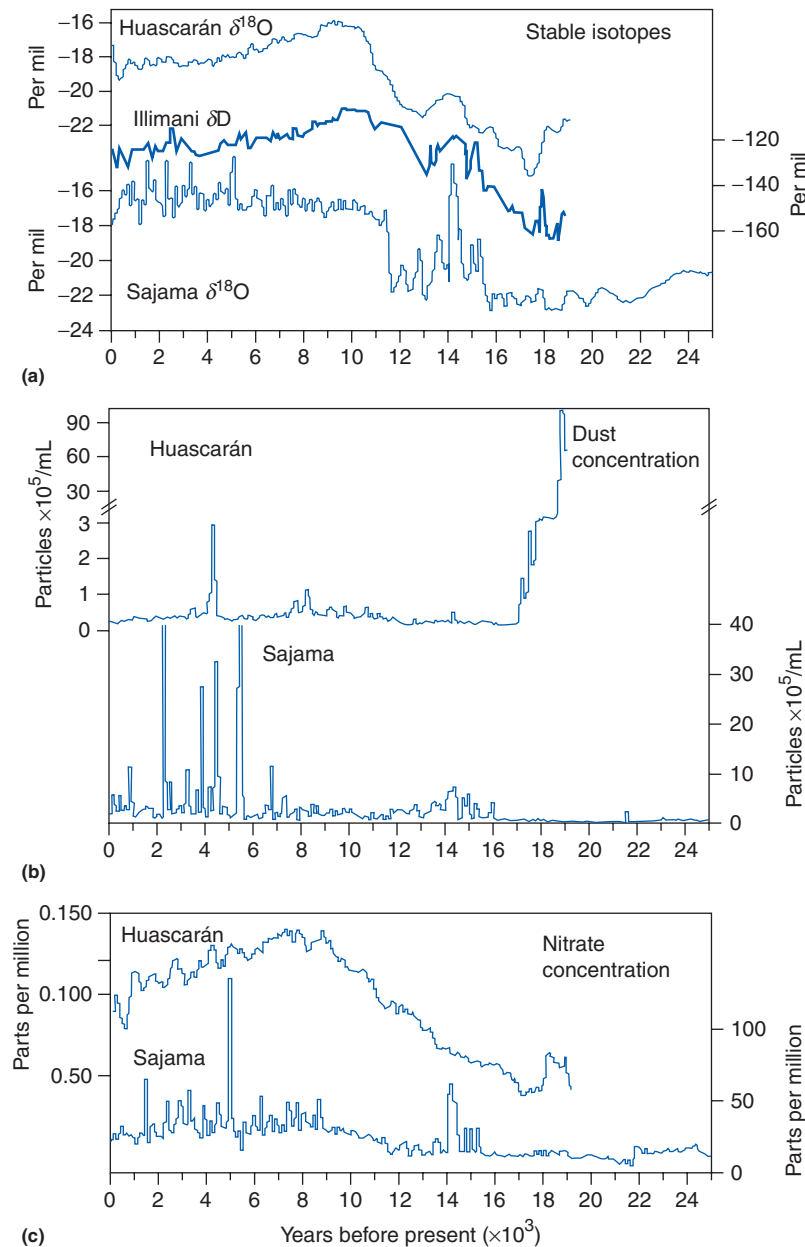


Figure 4 Long-term records of (a) stable isotopic ratios from the Huascarán, Illimani, and Sajama ice cores, (b) concentrations of dust (0.63 to 16.0 microns in diameter) from Huascarán and Sajama, and (c) concentrations of nitrate in the Huascarán and Sajama ice cores. The Huascarán and Sajama records are shown as 100-year averages. The Illimani δD record is adapted from [Ramírez et al. \(2003\)](#), with permission.

Altiplano, is more subject to Pacific influences and local hydrological effects. Also, because its precipitation rate is much lower than that for the eastern Andes cores ([Table 1](#)), it is more prone to snow sublimation at the glacier surface, which contributes to isotopic enrichment.

The existence of high-altitude glaciers from glacial to interglacial stages has struck a balance between temperature and orbital precession-driven precipitation ([Thompson et al., 2005](#)). The ice cap on top of Nevado Sajama began to grow $\sim 25\,000$ years ago in response to insolation-driven precipitation, although the temperatures remained cold. The ice on Nevado Huascarán, 1370 km to the northeast on the east side of the Andes, began to grow $\sim 19\,000$ years ago,

about 6000 years after the beginning of the formation of the Sajama ice cap. Approximately 23 000 years ago summer insolation had just reached its maximum at 9°S and precipitation intensity was beginning to increase from the LGM aridity in the tropical latitudes. At the time that Huascarán began to grow, precipitation over the Altiplano where Sajama is located was at its highest since the LGM ([Thompson et al., 1998](#)). Unlike the glaciers on Huascarán and Sajama, which are frozen to the bedrock, the base of the much lower Quelccaya ice cap is at the pressure melting point. Thus, thousands of years of the ice-core record have been removed from the bottom, leaving only a 1500-year climate history.

Century-Scale Variations

Abrupt climate events

Holocene ice-core records from mountain glaciers around the world show evidence of major climatic disruptions such as droughts and abrupt cold events during this period, which previously was believed to have been stable. A major dust event beginning between 4200 and 4500 years ago and lasting several hundred years is observed in the Huascarán record (Figure 4b), and its timing and character are similar to drought-related signals seen in a marine core record from the Gulf of Oman (Cullen et al., 2000) and a speleothem $\delta^{13}\text{C}$ record from Israel (Bar-Matthews et al., 1997). This dry period is also documented in several other proxy climate records throughout Asia and Africa (Dalfes et al., 1997; Thompson et al., 2002), and its occurrence in those regions and in tropical South America may have been due to a prolonged period of anomalously frequent and/or intense ENSO conditions (Davis and Thompson, 2006).

Several short-term dry periods are visible in the Sajama dust and nitrate records (Figures 4b, and c) after about 5500 years ago. Other proxy climate data from South America (Rodbell et al., 1999) and Australasia (Schulmeister, 1999) suggest that it was at this time in the middle Holocene that the frequency and/or intensity of El Niño increased from early Holocene levels. Garreaud and Aceituno (2001) explained the relationship between interannual variation in precipitation over the Altiplano and ENSO by linking the warming of the tropical troposphere during the negative phase of ENSO with the strengthening of the westerlies, which sets up lower precipitation and more arid conditions. If this relationship between Altiplano aridity and ENSO holds true, then the Sajama aerosol records also confirm this middle Holocene change in tropical oceanic/atmospheric coupling.

The 'Little Ice Age'

High-frequency climatic variations over the last millennium are recorded in decadal averages of $\delta^{18}\text{O}$ from the Quelccaya ice core (Figure 5a), which unlike the other records is annually dated back 1500 years (Thompson et al., 1985). From these

records came some of the first evidence for a Southern Hemisphere climatic cooling, indicated by lower $\delta^{18}\text{O}$, from the mid-fifteenth century to the mid-nineteenth century (Thompson et al., 1986). This was coincident with the 'Little Ice Age' previously seen in other parts of the world, as discussed in several papers in Bradley and Jones (1992). The reconstructed net accumulation rate (a indicator of snow precipitation) fluctuated widely during this period (Figure 5b), increasing to a maximum in the mid-seventeenth century, then decreasing rapidly to a minimum that persisted from the mid-eighteenth to the mid-nineteenth centuries. This changing relationship between $\delta^{18}\text{O}$ and the accumulation is at odds with the argument that stable isotopes in the tropics are primarily influenced by precipitation amount over decadal and longer time scales. If this were the case, the net accumulation would be consistently higher in that part of the record corresponding to the Little Ice Age, when $\delta^{18}\text{O}$ is consistently lower.

The Modern Climate Warming and its Effects on the Andes Mountain Glaciers

Meteorological data from around the world suggest that the Earth's globally averaged temperature has increased 0.6°C since 1950. The El Niño year of 1998 saw the highest globally averaged temperatures on record, while 2002 (a non-El Niño year) was the second warmest, followed by 2003 and then 2004 (a La Niña year). The recent warming of the past century, which has been accelerating in the last two decades, is recorded in alpine glaciers both within the ice-core records and by the rapid retreat of many of the ice fields themselves. In the Andes this climate change has left its mark. The twentieth-century shrinking of the Andean ice fields, when placed within a maximum 25 000-year perspective provided by the ice core records, shows that the current wasting of these ice fields is unprecedented in the Holocene despite evidence of intense and abrupt climatic disruptions in the past.

The ongoing retreat of Andean glaciers is now well-documented (Ames, 1998; Thompson et al., 2003; Thompson, 2004) and can be placed within a longer time perspective.

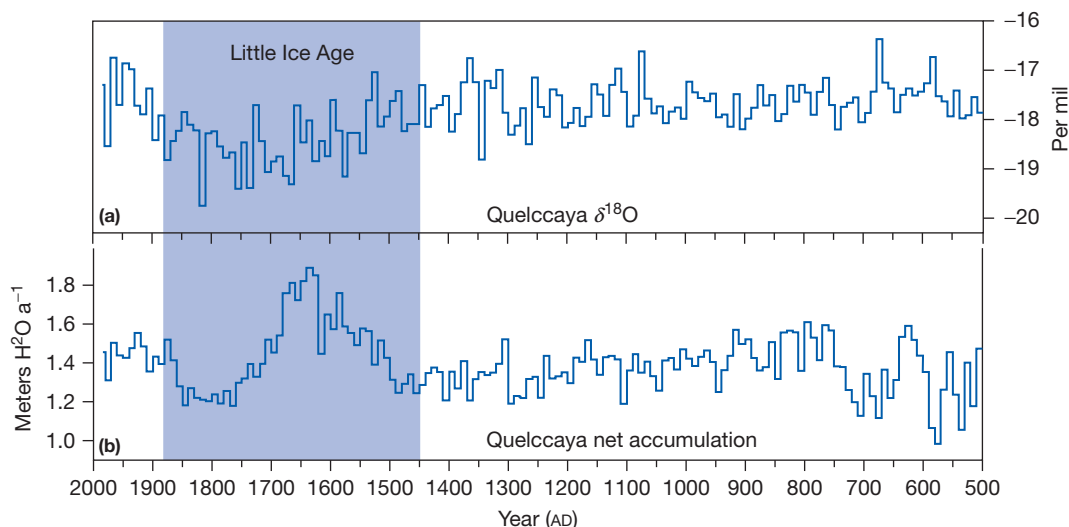


Figure 5 Decadal averages from AD 500 to 1980 from the Quelccaya ice-core record of (a) oxygen isotopic ratios and (b) reconstructed net accumulation, a proxy for precipitation. The blue bar outlines the 'Little Ice Age'.

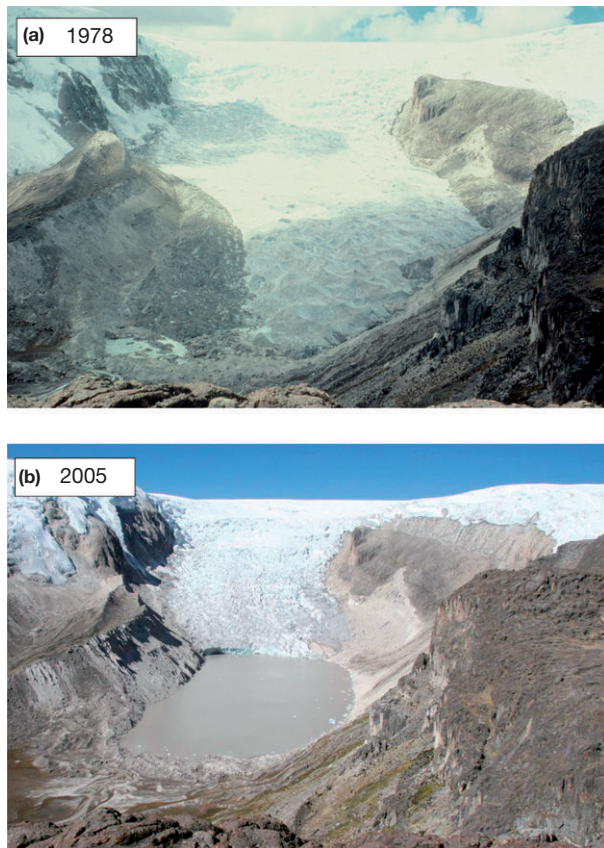


Figure 6 Photographs of the Qori Kalis outlet glacier on the Quelccaya ice cap. Between (a) 1978 and (b) 2005 the glacier has retreated 1100 meters.

The retreat of Qori Kalis, the largest outlet glacier from the Quelccaya ice cap, has been studied by terrestrial photogrammetry since 1978 and is the best documented recent history of any tropical glacier (Figure 6). The rate of its retreat from 1983 to 1991 (14 m/yr) was more than two times that between 1963 and 1983 (6 m/yr), and in the 2000/2001 year it reached 205 m/yr. There were periods when the retreat actually stopped (1991 to 1993 and 2002 to 2003) only to resume again in subsequent years. Overall, the rapid and accelerating wastage, in which the glacier front was receding between 1991 and 2005, was about 10 times faster than it was during the initial measurement period from 1963 to 1978.

This warming in the Andes is affecting the climatic signals in the ice-core records. The distinct seasonal oscillations in $\delta^{18}\text{O}$ that were visible at the top of the Huascarán ice cores (6050 m a.s.l.) in 1993 were obliterated by percolating meltwater in nearby, lower-elevation glaciers in the Cordillera Blanca such as Hualcán (5250 m a.s.l.), Pucahirca (5120 m a.s.l.), and Copap (5060 m a.s.l.) (Davis et al., 1995). The rate of melting is very fast; for example, the seasonal signals that were visible in the top of the Pucahirca glacier in 1984 were completely smoothed by 1990.

Future Priorities

Ice cores from mountain glaciers, especially those from the tropical latitudes where most of the world's population is concentrated, provide unique and valuable archives of climate

information because they are able to record variations in atmospheric chemistry and conditions. Since 1982, the ENSO phenomenon has gained worldwide attention as populations and governments have come to realize the extent of its widespread and often devastating effects on weather. As we begin to understand how this coupled atmospheric-oceanic process works, we also see its linkages with other important systems such as the Asian/African monsoons. Because both of these tropical systems influence precipitation and temperature over large regions, their effects are also recorded in the chemistry and the net accumulation of snow that falls on alpine glaciers. Seasonal and annual resolution of chemical and physical parameters in ice-core records from the Andes Mountains have allowed reconstruction of the variability of the ENSO phenomenon over several hundred years (Thompson et al., 1992; Henderson et al., 1999). Because the effects of El Niño and La Niña events are spatially variable, future ice core records from the northernmost (Columbia) and southernmost (Patagonia) reaches of the Andes Mountains will help further resolve their frequency and intensity, along with temperature variations long before human documentation. This will aid in placing the modern climate changes and the modern ENSO into a more comprehensive perspective.

At this time all the Earth's mountain glaciers are in retreat (Meier et al., 2003), a process which in many places has been accelerating over the past few decades. This has been linked to a marked and sustained rise in the global temperatures since the 1970s that has been superimposed on the more gradual warming that has occurred over the twentieth century. Some of the most dramatic evidence for the major climate warming that is underway today comes from the drastic decrease in both total area and total volume of the mountain glaciers. This rapid retreat causes concern for two reasons. First, these glaciers are the world's 'water towers', and their loss threatens water resources necessary for hydroelectric power production, crop irrigation, and municipal water supplies for many nations. The ice fields constitute a 'bank account' that is drawn upon during dry periods to supply populations downstream. The current melting is cashing in on that account, which was built up over thousands of years but is not currently being replenished. The second concern that is brought about by the disappearance of these ice fields is that they contain paleoclimatic histories that are unattainable elsewhere, and as they melt, the records preserved within them are lost forever. These records are needed to discern how climate has changed in the past in these regions and to assist in predicting future changes.

See also: [Ice Core Methods: Microparticle and Trace Element Studies; Stable Isotopes.](#) [Ice Core Records: Africa; Chinese, Tibetan Mountains.](#)

References

- Ames A (1998) A documentation of glacier tongue variations and lake developments in the Cordillera Blanca. *Zeitschrift für Gletscherkunde und Glazialgeologie* 34: 1–26.
- Baker PA, Seltzer GO, Fritz SC, et al. (2001) The history of South American tropical precipitation for the past 25,000 years. *Science* 291: 640–643.
- Bar-Matthews M, Ayalon A, Kaufman A, and Wasserburg GJ (1999) The Eastern Mediterranean paleoclimate as a reflection of regional events: Soreq Cave, Israel. *Earth and Planetary Science Letters* 166: 85–95.
- Benn DI and Evans DJA (1998) *Glaciers and Glaciation*. Arnold.
- Bradley RS and Jones PD (eds.) (1992) *Climate Since AD 1500*. Routledge.

- Clapperton C (1993) *Quaternary Geology and Geomorphology of South America*. Elsevier.
- CLIMAP Project Members (1976) The surface of the Ice-Age Earth. *Science* 191: 1131–1137.
- Colinvaux PA, De Oliveira PE, and Bush MB (2000) Amazonian and neotropical plant communities on glacial time-scales: The failure of the aridity and refuge hypotheses. *Quaternary Science Reviews* 19: 141–170.
- Cullen HM, deMenocal PB, Hemming S, et al. (2000) Climate change and the collapse of the Akkadian empire: Evidence from the deep sea. *Geology* 28: 379–382.
- Dalfes HN, Kukla G, and Weiss H (1997) *Third Millennium BC Climate Change and the Old World Social Collapse*. Springer.
- Dansgaard W (1964) Stable isotopes in precipitation. *Tellus* 16: 436–468.
- Davis ME, Thompson LG, Mosley-Thompson E, Lin P-N, Mikhalevko VN, and Dai J (1995) Recent ice-core climate records from the Cordillera Blanca, Peru. *Annals of Glaciology* 21: 225–230.
- Davis ME and Thompson LG (2006) An Andean ice-core record of a Middle Holocene mega-drought in North Africa and Asia. *Annals of Glaciology* 43: (in press).
- Garreaud RD and Aceituno P (2001) Interannual rainfall variability over the South American Altiplano. *Journal of Climate* 14: 2779–2789.
- Grootes PM, Stuvier M, Thompson LG, and Mosley-Thompson E (1989) Oxygen isotope changes in tropical ice, Quelccaya, Peru. *Journal of Geophysical Research* 94: 1187–1194.
- Guilderson TP, Fairbanks RG, and Rubenstone JL (1994) Tropical temperature variations since 22,000 years ago: modulating inter-hemispheric climate change. *Science* 263: 663–665.
- Hastenrath S (1978) On modes of tropical circulation and climate anomalies. *Journal of the Atmospheric Science* 35: 2222–2223.
- Henderson KA, Thompson LG, and Lin P-N (1999) Recording of El Niño in ice core ^{18}O records from Nevado Huascarán, Peru. *Journal of Geophysical Research* 104: 31,053–31,065.
- Lenters JD and Cook KH (1997) On the origin of the Bolivian High and related circulation features of the South American climate. *Journal of the Atmospheric Science* 54: 656–677.
- Lough JM (1986) Tropical Atlantic sea surface temperatures and rainfall variations in Sub-Saharan Africa. *Monthly Weather Review* 114: 561–570.
- Marengo JA and Rogers JC (2000) Polar air outbreaks in the Americas: Assessments and impacts during modern and past climates. In: Markgraf V (ed.) *Interhemispheric Climate Linkages*, p. 31. Academic Press.
- Marques J, dos Santos JM, Villa Nova NA, and Salati E (1977) Precipitable water and water vapor flux between Belem and Manaus. *Acta Amazonia* 7: 355–362.
- Meier MF, Dyurgerov MB, and McCabe GJ (2003) The health of glaciers: Recent changes in glacier regime. *Climatic Change* 59: 123–135.
- Ramírez E, Hoffmann G, Taupin JD, et al. (2003) A new Andean deep ice core from Nevado Illimani (6350 m), Bolivia. *Earth and Planetary Science Letters* 212: 337–350.
- Rodbell D, Seltzer GO, Anderson DM, Enfield DB, Abbott MB, and Newman JH (1999) A high-resolution ~15,000 year record of El Niño driven alluviation in southwestern Ecuador. *Science* 283: 516–520.
- Rozanski K, Araguás-Araguás L, and Gonfiantini R (1993) Isotopic patterns in modern global precipitation. In: Swart PK, Lohman KC, McKenzie J, and Savin S (eds.) *Climate Change in Continental Isotopic Records – Geophysical Monograph*, vol. 78, p. 1. American Geophysical Union.
- Servain J (1991) Simple climatic indices for the tropical Atlantic Ocean and some applications. *Journal of Geophysical Research* 96: 15,137–15,146.
- Servain J and Legler DM (1986) Empirical orthogonal function analysis of tropical Atlantic sea surface temperature and wind stress: 1964–1979. *Journal of Geophysical Research* 91: 14,181–14,191.
- Shulmeister J (1999) Australasian evidence for mid-Holocene climate change implies precessional control of Walker circulation in the Pacific. *Quaternary International* 57(58): 81–91.
- Stute M, Forster M, Frischkorn H, et al. (1995) Cooling of tropical Brazil (5 °C) during the last glacial maximum. *Science* 269: 379–383.
- Taljaard JJ (1972) Synoptic meteorology of the southern hemisphere. In: Newton CW (ed.) *Meteorology of the Southern Hemisphere*, p. 13. American Meteorological Society.
- Thompson LG (2004) High-Altitude, Mid and Low-Latitude Ice Core Records: Implications for our future. In: Cecil LD, Green JR, and Thompson LG (eds.) *Earth Paleoenvironments: Records Preserved in Mid- and Low-Latitude Glaciers*, p. 3. Kluwer.
- Thompson LG, Davis ME, Mosley-Thompson E, et al. (1998) A 25,000 year tropical climate history from Bolivian ice cores. *Science* 282: 1858–1864.
- Thompson LG, Davis ME, Mosley-Thompson E, Lin P-N, Henderson KA, and Mashiotta MA (2005) Tropical ice core records: Evidence for asynchronous glaciation on Milankovitch timescales. *Journal of Quaternary Science* 20: 723–733.
- Thompson LG, Hastenrath S, and Arnao BM (1979) Climatic ice core records from the tropical Quelccaya ice cap. *Science* 203: 1240–1243.
- Thompson LG, Mosley-Thompson E, Bolzan JF, and Koci BR (1985) A 1500 year record of climate variability recorded in ice cores from the tropical Quelccaya Ice Cap. *Science* 229: 971–973.
- Thompson LG, Mosley-Thompson E, Dansgaard W, and Grootes PM (1986) The 'Little Ice Age' as recorded in the stratigraphy of the tropical Quelccaya ice cap. *Science* 234: 361–364.
- Thompson LG, Mosley-Thompson E, Davis ME, et al. (1995) Late Glacial Stage and Holocene tropical ice core records from Huascarán, Peru. *Science* 269: 46–50.
- Thompson LG, Mosley-Thompson E, Davis ME, et al. (2002) Kilimanjaro ice core records: Evidence of Holocene climate change in tropical Africa. *Science* 298: 589–593.
- Thompson LG, Mosley-Thompson E, Davis ME, Lin P-N, Henderson KA, and Mashiotta TA (2003) Tropical glacier and ice core evidence of climate change on annual to millennial time scales. *Climatic Change* 59: 137–155.
- Thompson LG, Mosley-Thompson E, and Thompson PA (1992) Reconstructing interannual climate variability from tropical and subtropical ice-core records. In: Diaz HF and Markgraf V (eds.) *El Niño: Historical and Paleoclimatic Aspects of the Southern Oscillation*, p. 295. Cambridge: Cambridge University Press.
- Vuille M (1999) Atmospheric circulation over the Bolivian Altiplano during dry and wet periods and extreme phases of the Southern Oscillation. *International Journal of Climatology* 19: 1579–1600.
- Vuille M and Ammann C (1997) Regional snowfall patterns in the high arid Andes. *Climatic Change* 36: 413–423.
- Weare BC (1977) Empirical orthogonal function analysis of Atlantic surface temperatures. *Quarterly Journal of the Royal Meteorological Society* 103: 467–478.

Antarctic Stable Isotopes

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Introduction

An intensive monitoring of precipitation isotopic composition has been operating since the 1950s (e.g., Dansgaard, 1953). Natural waters formed mostly of H_2^{16}O molecules (99.7%) also exhibit less frequent stable isotopic forms, including about 0.2% of H_2^{18}O and 0.03% of HD^{16}O (D stands for deuterium ^2H). Isotopic concentrations are expressed in per mil δ units (δD and $\delta^{18}\text{O}$) with respect to Vienna standard mean ocean water (V-SMOW) (the Vienna standard mean ocean water, with D/H and $^{18}\text{O}/^{16}\text{O}$ atomic ratios of 155.76×10^{-6} and 2005.2×10^{-6} , respectively).

Measurements of modern precipitation revealed the potential of past precipitation for reconstructing paleoclimatic changes, owing to the existence of a relationship between δD or $\delta^{18}\text{O}$ and surface temperature at the middle and high latitudes. This relationship is very consistent and linear over Antarctica with an isotope content/surface temperature slope of $\sim 6\text{‰}/\text{C}$ for δD and of $\sim 0.75\text{‰}/\text{C}$ for $\delta^{18}\text{O}$ (Lorius and Merlivat, 1977). This ice sheet, which offers the widest reservoir of preserved past precipitation, has provided a wealth of ice-core isotopic records. Deuterium and ^{18}O measurements allow us (1) to reconstruct past temperature changes, (2) to access to past accumulation changes, related to water vapor pressure and thus to past surface temperatures, and (3) to derive information on conditions prevailing in the oceanic regions providing precipitation to Antarctica. The latter approach is based on the deuterium-excess parameter, $d = \delta\text{D} - 8\delta^{18}\text{O}$, which is influenced by the temperature of oceanic moisture sources and to a lesser degree by its relative humidity and prevailing wind speed.

In this article, we first examine the spatial distribution of water isotopes in Antarctica for present-day precipitation and assess the use of these observations (the present-day spatial slope) to derive temperature records back in time at a given site (the temporal slope). We then discuss paleoclimatic reconstructions based on this isotopic paleothermometer. Water isotope records are now available from a variety of Antarctic ice cores (Figure 1) both for the current interglacial period, the Holocene, and at the glacial–interglacial timescale with the longest record covering eight climatic cycles back to 800 ka BP.

Modern Precipitation Isotopic Composition and the Isotopic Paleothermometer

Surface snow sampled along traverse routes from coastal stations to inland sites can be analyzed for its isotopic composition (Masson-Delmotte et al., 2008). The linear relationship observed with temperature (Figure 2) can be modeled using a simple isotopic model as well as by isotopic general circulation models (GCMs) (atmospheric general circulation models

equipped with the explicit modeling of water isotopes). This well-explained ‘spatial slope’ is the basis for the use of stable isotope profiles measured along deep ice cores for temperature estimates (isotopic paleothermometry).

In order to assess the persistence of this spatial relationship in time (temporal slope), several studies have explored sites where long meteorological records are available together with isotopic records. Law Dome, a coastal site with particularly high accumulation rates (70 cm of ice equivalent per year at the Dome Summit South (DSS) ice-core site), offers the possibility of comparing the spatial slope, the seasonal slope (Delmotte et al., 2000; van Ommen and Morgan, 1997), and the decadal slope (Masson-Delmotte et al., 2003). At Vostok, due to the low accumulation rate (2.3 cm of ice equivalent per year) postdepositional effects such as wind scouring only permit to evaluate decadal slopes (Ekaykin et al., 2004).

Spatial or seasonal relationships between the site temperature and the isotopic content of the snow must however be regarded with caution. As reflected by the spatial fluctuations of the deuterium excess (Figure 2), they integrate the effect of both local condensation temperature and changes in moisture origin and distillation history. Isotopic modeling allows us to disentangle the relative impact of changes in evaporation conditions versus condensation conditions on isotopic distribution. Such an approach must be implemented for modern data (spatial variations, seasonal cycle with simple models; inter-annual to decadal variability with isotopic GCMs). In particular, simple isotopic models require an adjustment of the parametrizations controlling the type of distillation (closed vs. open cloud) or the dependency of supersaturation over ice crystals with respect to temperature. The latter parameter is also tuned in isotopic GCMs (Jouzel et al., 1991). For instance, at the Dome C site, we obtain the following relationships from linear regressions conducted on multiple simulations based on a simple isotopic model forced with varying evaporation and condensation conditions (in the range of glacial to interglacial changes):

$$\Delta\delta\text{D} = 7.6\Delta T_s - 3.6\Delta T_o \quad [1]$$

$$\Delta d = -0.5\Delta T_s - 1.3\Delta T_o \quad [2]$$

where T_s is the site temperature, T_o is the ocean surface temperature, and $\delta^{18}\text{O}_{\text{sw}}$ is the ocean water isotopic composition; the symbol Δ is used for anomalies between two time periods (Masson-Delmotte et al., 2004). From this modeling approach, one can in principle access both the site and source temperatures, once both δD and $\delta^{18}\text{O}$ are measured, as is now systematically done on Antarctic ice cores.

The local temperature record inferred from this set of equations, $\Delta T_s = 0.161 \times \Delta\delta\text{D} + 0.446 \times \Delta d$, is indeed very close (i.e., within $\pm 10\%$) to the one calculated using the present-day spatial slope as illustrated (Stenni et al., 2003) from the

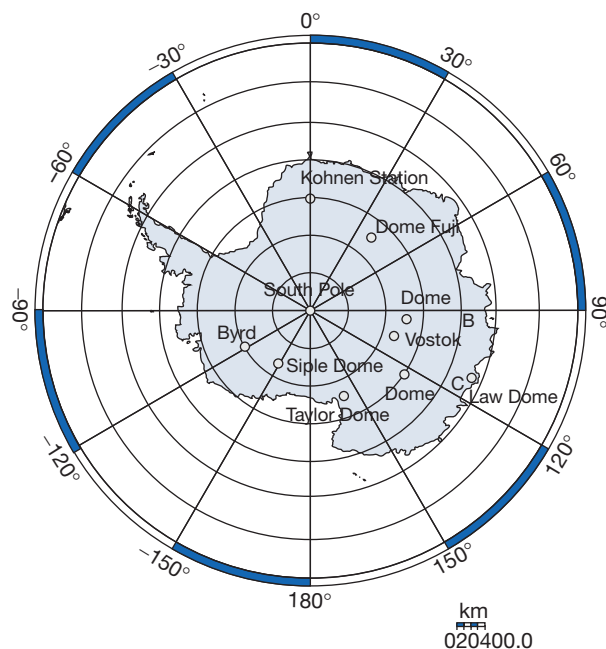


Figure 1 Map showing the location of deep Antarctic ice cores.

temperature record reconstructed at the Dome C site from coisotopic δD and $\delta^{18}O$ determinations (Figure 2).

The validity of this simple model approach, in which the spatial slope is used as a surrogate of the temporal slope to estimate the surface temperature at the site, has been independently confirmed from isotopic GCMs. Comparison of simulations performed for glacial and present-day climates confirms that the temporal slope is close to the modern spatial isotope-temperature slope of inland Antarctica (Delaygue et al., 2000), (Werner et al., 2001). Unlike for Greenland, which is characterized by drastic glacial-interglacial changes in precipitation seasonality, Antarctic water-stable isotopes appear to provide reliable quantitative information on past temperature changes (within e.g., $\pm 20\%$), at least at the glacial-interglacial scale (Figure 3). Further confirmation of the validity of the isotopic paleothermometer comes from chronological constraints or from estimates of the gas age-Ice Age difference (Jouzel et al., 2003), (Blunier et al., 2004).

Holocene Isotopic Records

Last 200 Years

High-resolution isotopic measurements conducted on ice cores from high accumulation coastal sites have been used to reconstruct accumulation and temperature fluctuations during recent decades. In inland locations, deposition and postdeposition effects prevent such decadal-scale analyses unless numerous pits are stacked to provide a record of precipitation isotopic composition. Such an effort has been undertaken for Vostok (Ekaykin et al., 2004). At Vostok, δD multidecadal fluctuations appear to occur during the past two centuries, consistent with accumulation changes and Pacific decadal oscillation changes; these fluctuations, when translated into temperature changes, appear to remain within $1^\circ C$ of modern parameters. A small

increasing trend is detected over the past two centuries. In contrast to this, a high accumulation site such as Law Dome can provide an isotopic record with a subannual resolution (Morgan and Van Ommen, 1997; Schneider et al., 2006). Combined measurements of δD and $\delta^{18}O$ at Law Dome (Masson-Delmotte et al., 2003) revealed both that coastal Antarctic water-stable isotopes do capture multidecadal changes in temperatures (a local temperature increase since the 1960s) and that abrupt reorganizations in the advection of moisture toward Law Dome took place both in the 1940s and the 1970s, identified as 'steps' in Law Dome deuterium excess.

The diversity of locations where shallow ice cores have been extracted and analyzed for isotopic changes should enable the synthesis of the evolution of Antarctic snow isotopic composition over the past centuries with a decadal resolution. However, it must be stressed that coastal and inland snowfalls do not undergo similar distillation histories, and do not have similar moisture sources, as reflected by changes in deuterium excess. Systematic comparisons of decadal-scale deuterium excess fluctuations over the last centuries should bring new insights into the recent dynamics of water cycle and climate change. These aspects are also important for determining the mass balance of the ice sheet.

The Last Millennium

A comparison of isotopic records spanning this time period has been recently presented by Goose et al. (2004), who constructed a stack of East Antarctic deuterium records from South Pole, Taylor Dome, Talos Dome, EPICA Dome C, and Law Dome. This stack suggests that there is an apparent phase lag between the so-called Medieval warm period reconstructed in the Northern Hemisphere and the Antarctic warmest period that occurred one to two centuries later. Temperature fluctuations, reconstructed using the spatial slope, remained within $1^\circ C$ of modern parameters. This preliminary synthesis is limited by the availability of records (here only in a portion of east Antarctica), the lack of a common timescale, and the limited deuterium excess information.

Holocene

For the current interglacial period, a synthesis of 11 ice-core isotopic records has been achieved for δD (Masson et al., 2000). Although limited by the lack of a common age scale (based, for instance, on volcanic eruptions), this comparison revealed a consistent long-term isotopic pattern for Eastern Antarctica, with an early Holocene optimum of $5\text{--}10\text{‰}$ occurring $12\text{--}8$ ka BP. Following this optimum, different sites undergo different isotopic trends, possibly partly associated with the evolution of the ice-core site elevation; most of the sites exhibit a small decreasing trend. Superimposed on these long-term trends, the various ice cores reveal an organized millennial-scale periodicity associated with eight well-marked warm peaks and three well-marked cold peaks, the last one taking place during the last millennium. Spectral analyses conducted on all the records show several common intervals of significant periodicities in the multidecadal to centennial modes (from 70 to 240 yr periodicities). Unfortunately, the meteorological

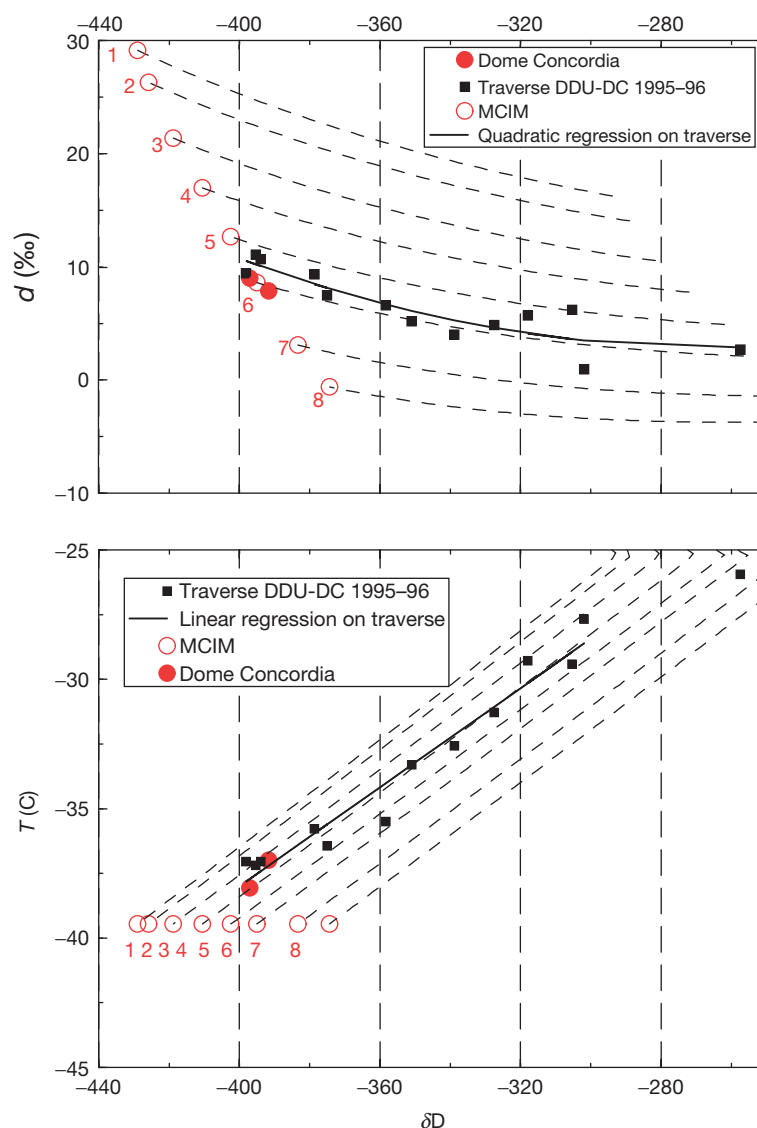


Figure 2 Spatial variability of water-stable isotopes of modern Antarctic snow along traverse routes from Dumont d'Urville to Dome C (square points), interpolation (solid line) and comparison with distillation histories calculated by a cloud isotopic model in response to varying initial conditions (dashed lines). Adapted from Masson-Delmotte V, et al. (2004) Common millennial scale variability of Antarctic and southern ocean temperatures during the past 5000 years reconstructed from EPICA Dome C ice core. *The Holocene* 14: 145–151.

records in Antarctica are too short to independently verify presence of such long-term oscillations.

The Holocene deuterium excess fluctuations recorded on four of these records (Vostok, Dome B, Dome C, and Taylor Dome) were compared by Vimeux et al. (2001a). The common warming in the first half of the Holocene is attributed to a progressive warming of the oceanic moisture sources. Available paleoceanographic records reflect a progressive warming of the tropics together with a progressive cooling of the high latitudes. Holocene changes in obliquity, redistributing annual mean insolation changes at low versus high latitudes, could account for the observed deuterium excess increase, attributed to an increasing contribution of low latitudes to Antarctic precipitation during the Holocene. Again, although the long-term trends are common to all Antarctic records, millennial and high-frequency variability of the deuterium excess profiles

appear to be related to local processes and are not common to all the sites.

On EPICA Dome C, $\delta^{18}O$ and δD analyses conducted at a ~ 20 -yr resolution (Masson-Delmotte et al., 2004) were combined to quantify the compatible changes in moisture source and site temperatures. This approach suggests that the early Holocene optimum appeared first in Antarctic temperatures, followed 800 yr later by an optimum in the moisture source, and revealed a specific out-of-phase behavior relating site and source temperature occurring about 8 ka, similar to the situation taking place during the Antarctic cold reversal (see 'Antarctic Isotopic Records Over Glacial-Interglacial Cycles'). The reconstructed site and source temperature strongly covary at the centennial scale during the last 5 ka, suggesting the setup of a coupling mechanism between ocean moisture source and Antarctic temperature, possibly related to the onset of modern

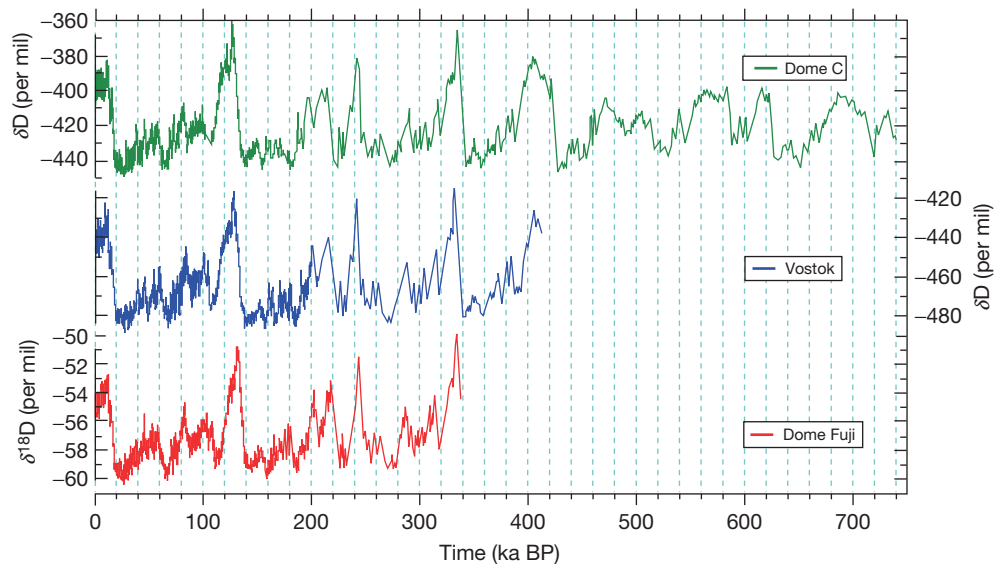


Figure 3 Stable isotope records from deep Antarctic ice cores: Vostok, Dome F, Dome C.

El Niño southern oscillation (ENSO). Finally, this approach allows us to estimate that the peak warmth at the beginning of the Holocene was $\sim 1.5^\circ\text{C}$ warmer than today.

Antarctic Isotopic Records over Glacial–Interglacial Cycles

The Byrd ice core, drilled in West Antarctica ($80^\circ 01' \text{S}$, $119^\circ 31' \text{W}$), provided a record of precipitation $\delta^{18}\text{O}$ spanning almost one glacial climatic cycle ($\sim 80 \text{ ka}$) (Johnsen et al., 1972). Several other ice cores now cover the last deglaciation, on the western (Siple Dome) (Brook et al., 2005) and eastern (Taylor Dome) (Steig et al., 1998) sides of the Ross Sea, on the East Antarctic coast in the Indian Ocean sector (Law Dome) (Morgan et al., 2002), on the East Antarctic plateau (Komsomolskaia, Vostok, Dome B, Dome F, EPICA Dome C) (Lorius et al., 1979; Jouzel et al., 2001; Petit et al., 1999; Watanabe et al., 2003) and on the coast of the Atlantic sector of East Antarctica (EPICA Kohnen Station) (Figure 1). Moreover, three of these deep Antarctic ice cores cover more than one climatic cycle: Dome F spans three climatic cycles (Watanabe et al., 2003), Vostok four climatic cycles (Petit et al., 1999) and EPICA Dome C eight climatic cycles (EPICA-community-members, 2004).

Glacial–Interglacial Temperature Change

All inland east Antarctic deep ice cores show a consistency in their glacial to interglacial isotopic fluctuations with a typical amplitude of δD changes of $\sim 50\text{‰}$. The use of the spatial isotope-temperature slope to estimate the associated temperature range first requires a correction for the imprint of changes in ocean water isotopic composition associated with the buildup of glacial land ice masses. The impact of ocean surface isotopic composition, $\delta^{18}\text{O}_{\text{sw}}$, on Antarctic precipitation isotopic composition can be calculated explicitly, by correcting $\delta^{18}\text{O}$ from the factor $\delta^{18}\text{O}_{\text{sw}} \times (1 + \delta^{18}\text{O}) / (1 + \Delta\delta^{18}\text{O}_{\text{sw}})$

and similarly by correcting δD from the factor $8 \times \delta^{18}\text{O}_{\text{sw}} \times (1 + \delta^{18}\text{O}) / (1 + 8 \times \Delta\delta^{18}\text{O}_{\text{sw}})$ (Jouzel et al., 2003). The application of the spatial isotope temperature to inland Antarctic sites such as Dome C suggests a glacial to Holocene surface temperature change of $8\text{--}9^\circ\text{C}$ (Figure 4).

However, the change between glacial and modern conditions also involves profound reorganizations of the water cycle, from evaporation conditions to deposition conditions in interior Antarctica. Available deuterium excess records from inland sites (Jouzel et al., 1982; Stenni et al., 2003; Vimeux et al., 2001b; Watanabe et al., 2003) consistently show a 3‰ smaller deuterium excess at the Last Glacial Maximum (LGM). Taken together, the deuterium excess and δD constraints can be quantified in terms of changes in both source and site temperatures. In this case, the LGM–present-day estimated temperature change is only marginally affected ($\sim 0.5^\circ\text{C}$). Of course, many factors other than just surface temperature or source temperature can alter the isotope-temperature relationships, such as the factors controlling evaporation kinetics (relative humidity, wind speed), changes in cloud microphysics when atmospheric dust levels vary, changes in moisture advection history, changes in condensation versus surface temperature, or changes in the seasonality or intermittency of snow deposition.

Millennial Climate Variability in Antarctica

The water-stable isotope profiles from Antarctic ice cores do not undergo as rapid a variability during the Ice Ages as that recorded in Greenland, where this rapid variability reaches two-thirds of the full glacial–interglacial isotopic amplitude (Section Greenland). However, superimposed on the glacial–interglacial cycles, central Antarctic ice cores exhibit a succession of higher-frequency fluctuations reflected in many parameters analyzed in ice cores, including stable isotopes.

The last deglaciation as reflected in Antarctic water-stable isotopes does not appear as a continuous process but was interrupted by the so-called ‘Antarctic cold reversal’ (Jouzel et al., 1995) from about 14 to 13 ka BP. This cold reversal,

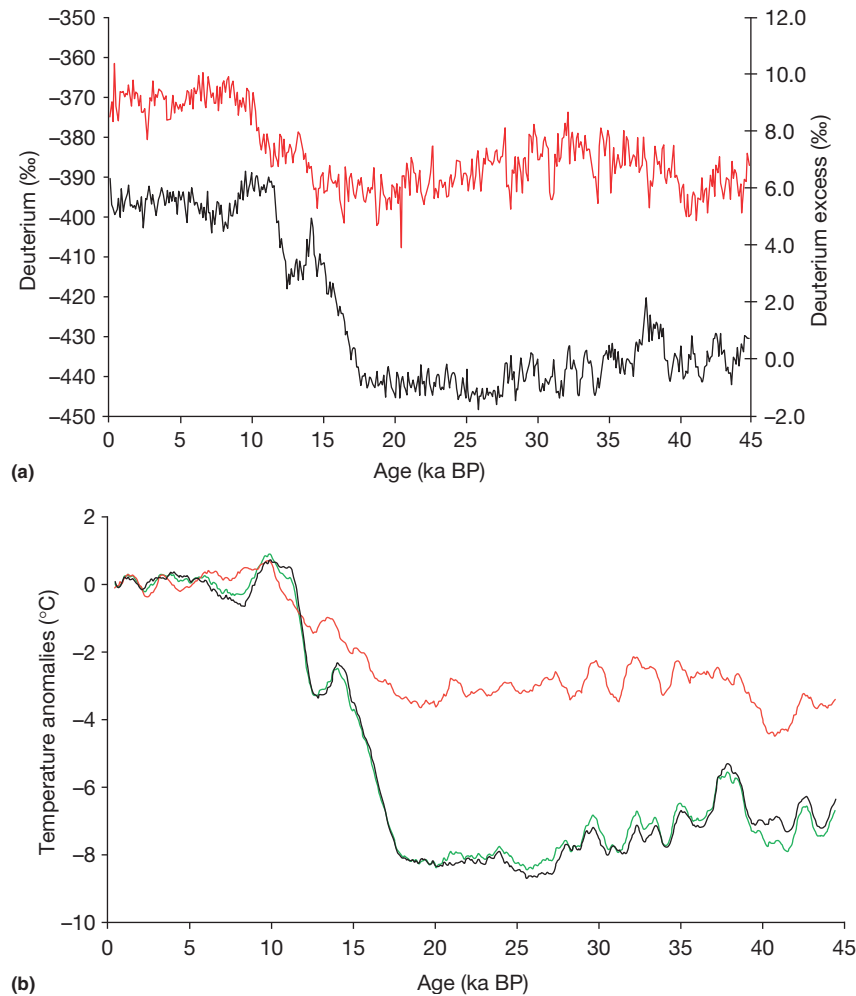


Figure 4 (a) Dome C water-stable isotope records (black, δD on the left axis; red, deuterium excess on the right axis) and (b) reconstructed temperatures during the past 45 ka (black, site temperature from the use of the spatial slope after correction for sea water isotopic composition; green, site temperature after inversion taking into account deuterium excess and sea water isotopic composition; red, source temperature after inversion taking into account deuterium excess and sea water isotopic composition). Adapted from Stenni *B. et al.* (2003) A high resolution site and source Late Glacial temperature record derived from the EPICA Dome C isotope records (East Antarctica). *EPSL* 217: 183–195.

identified in all Antarctic ice cores, could be the Antarctic counterpart of the abrupt warming corresponding to the Bölling-Alleröd recorded in Greenland ice cores, owing to the synchronization of bipolar ice cores using the atmospheric composition changes recorded in air from ice cores (Blunier *et al.*, 1997) (Morgan *et al.*, 2002) (Pedro *et al.*, 2011).

The last Ice Age is punctuated by a series of deuterium events, occurring at a pace of ~4–10 ka and with typical amplitudes of one-fourth to one-half of the full glacial-interglacial amplitude, with progressive increases and decreases over several millenia (Blunier and Brook, 2001; EPICA-community-members, 2006). These so-called ‘A’ events (labeled A1–A7 on Byrd ice core) (Blunier and Brook, 2001) have been placed on a common chronological framework with Greenland rapid events, using a synchronization based on methane rapid fluctuations. It appears that Antarctic temperatures increase progressively when Greenland is in a cold or cooling phase, reach their peak warmth when Greenland undergoes an abrupt warming associated with a Dansgaard-Oeschger event,

and then progressively decrease. This behavior appears to be related to a seesaw mechanism involving reorganizations of the ocean circulation and heat transport, particularly marked during the largest Dansgaard-Oeschger events.

High-resolution deuterium and deuterium excess measurements conducted on EPICA Dome C ice core and spanning the last 45 ka have allowed us to propose a quantitative reconstruction of these rapid events; at the resolution achieved here, all Greenland rapid events appear to have Antarctic counterparts (Stenni *et al.*, 2003). Taking into account deuterium excess fluctuations reveals that rapid deuterium fluctuations (ACR, A1 event, millennial-scale fluctuations of smaller amplitude) cannot be explained by changes in source conditions, and correspond to Antarctic temperature fluctuations of ~1 to ~2 °C. During the past 45 ka, a strong correlation is observed between the reconstructed site to source temperature gradient and the aerosol fluxes derived from chemical analyses of the ice core, revealing a strong link between meridional temperature gradients and atmospheric transport of aerosols to Antarctica.

For the most rapid deuterium fluctuation recorded in Vostok ice, the 5d/5c transition that took place ~ 107 ka BP (GT4 age scale), [Caillon et al. \(2001\)](#) applied the gas-fractionation method recently developed as an alternative paleothermometry method for Greenland rapid events. This method enables the precise assessment of the phase lag between the Antarctic temperature increase (as seen in deuterium and $\delta^{15}\text{N}$ and $\delta^{40}\text{Ar}$ of the air trapped in the ice) and Northern Hemisphere temperatures (as recorded in Vostok methane concentration). The temperature signal inferred from $\delta^{15}\text{N}$ and $\delta^{40}\text{Ar}$ is attributed to changes in a gravitational fractionation resulting from changes in firn thickness and close-off depths. This change in close-off depths enables the assessment of changes in temperatures. This method confirms again that, for central Antarctica (here Vostok), the classical use of water-stable isotopes remains valid within 20%, not only for glacial–interglacial changes but also during rapid events.

Finally, the first rapid events of the last Ice Age have recently been analyzed from a bipolar perspective owing to a preliminary synchronization of NorthGRIP ice core (Greenland) and Vostok ice. This common age scale was constructed based on the 5d/5c transition identified in methane from both poles and also on the air $\delta^{18}\text{O}$ from both sites ([Landais et al., 2005](#)). It appears that the onset of the last Ice Age began by a common bipolar cooling that lasted at least 5 ka. In Greenland, this progressive cooling was interrupted by a first Dansgaard–Oeschger event (D/O 25), which does not have a clear Antarctic counterpart as seen in the Vostok deuterium profile. However, the following D/O events 24 and 23 are clearly associated with Antarctic counterparts, similar to the rapid events occurring during full glacial conditions.

Glacial–Interglacial Cycles

Three deep Antarctic ice cores span several climatic cycles: Dome F (three climatic cycles) ([Watanabe et al., 2003](#)), Vostok (four climatic cycles) ([Petit et al., 1999](#)), and Dome C (eight climatic cycles) ([EPICA-community-members, 2004](#)). Over the overlap period, the deuterium records of these three sites exhibit very similar fluctuations, confirming both the spatial representativity of stable isotope records from inland East Antarctica. A comparison of ice-core deuterium records with marine global ice volume records ([Liesicki and Raymo, 2005](#)) shows also a large similarity ([EPICA-community-members, 2004](#)), revealing the imprint of 100 ka cycles and periodicities associated with obliquity and precession ([Petit et al., 1999](#)). Glacial maximum levels appear rather stable over time, although Antarctic ice-core deuterium fluctuations depict a variety of warm periods. A classical use of the spatial isotope-temperature relationship suggests that temperatures up to $\sim 5^\circ\text{C}$ above present-day levels occurred during marine isotopic stages 5 and 9 ([Watanabe et al., 2003](#)).

The longest interglacial period occurred about 400 ka, and corresponding with marine isotopic stage 11. As expected from considerations of thresholds in 65°N mid-June insolation resulting from the low eccentricity ([Berger and Loutre, 2002](#)), this interglacial period lasted ~ 28 ka in the first Dome C age scale. Its peak warmth corresponds to a temperature $\sim 2.5^\circ\text{C}$ above present-day level, using the spatial isotope-temperature relationship.

Deuterium excess measurements have been conducted on Vostok and Dome F ice cores over several climatic cycles ([Watanabe et al., 2003](#); [Vimeux et al., 1999](#); [Vimeux, 2001b](#)). These measurements have revealed the stability of evaporation conditions during glacial–interglacial cycles; reconstructed source temperature fluctuations remain within $\sim 3^\circ\text{C}$ of modern parameters. The impact of changes in source conditions remains generally limited in deuterium fluctuations, apart from glacial inception. During glacial inceptions, deuterium excess levels remained high when deuterium levels decreased. Over the past 250 ka, the Vostok and Dome F deuterium-excess profiles were strongly influenced by changes in the obliquity of the Earth's orbit. It is suggested that small obliquities induce warmer tropics, colder high latitudes, larger low to high latitude gradients, and a more intense transport of moisture from warm low latitude areas to the poles. All these effects combine to generate a high deuterium excess during periods of small obliquities, and a systematic antiphase behavior between deuterium excess of Antarctic ice and obliquity. The marked high levels of deuterium excess at the onset of Ice Ages result from this general link with obliquity and point to large reorganizations of the water cycle during glacial inceptions, providing the moisture required in the high Northern Hemisphere latitudes to build up large ice caps ([Vimeux et al., 1999](#)) ([Vimeux, 2001b](#)).

Conclusions and Perspectives

When ice cores are drilled in Antarctica, water-stable isotope records are now among the essential analyses to quantify past changes in climate and hydrological cycle, and to date the ice cores. The state-of-the-art of the paleothermometry method and climate modeling suggest that on both glacial–interglacial and rapid event timescales they provide reliable and quantitative indications of past climate change ([Figure 5](#)). Further progress on the quantification of uncertainty sources such as changes in evaporation conditions may arise from $\delta^{17}\text{O}$ analyses of Antarctic snow and ice, and the combination of isotopic modeling and analyses of atmospheric back-trajectories, together with an improved characterization of the modern spatial distribution in unexplored regions of Antarctica. Obtaining more information on past climate changes from water-stable isotopes in Antarctic ice cores requires analyses at the highest possible temporal resolution, together with the collection of new ice cores both at high accumulation sites (to reconstruct climate change during the last millennia) and very low accumulation sites (to extend the records before 1 Myr). Deep ice cores are now being planned at the international level within the International Partnership for Ice Core Science (IPICS) umbrella.

See also: [Glaciations: Late Quaternary of Antarctica](#). **Ice Core Methods:** [Chronologies](#); [CO₂ Studies](#); [Overview](#). **Ice Core Records:** [Greenland Stable Isotopes](#). **Ice Cores:** [History of Research, Greenland and Antarctica](#). **Paleoceanography, Physical and Chemical Proxies:** [Oxygen-Isotope Stratigraphy of the Oceans](#). **Paleoclimate Modeling:** [Last Glacial Maximum GCMs](#). **Paleoclimate Reconstruction:** [Sub-Milankovitch \(DO/Heinrich\) Events](#).

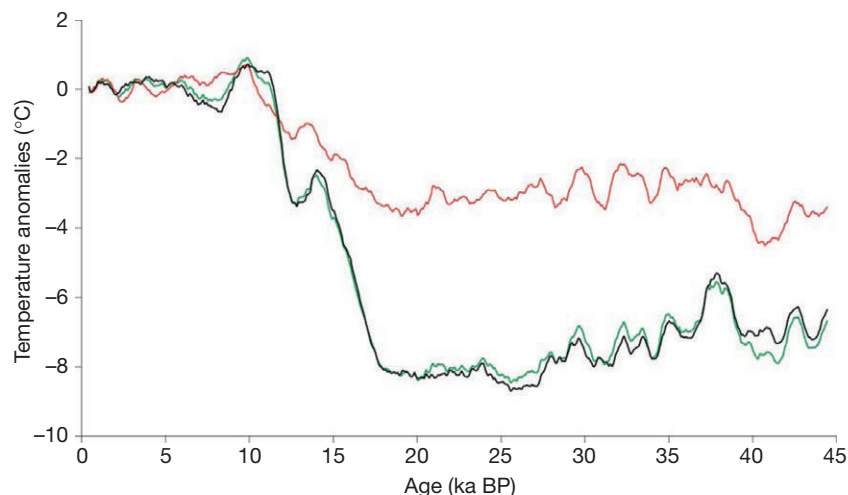


Figure 5 Spatial isotope–temperature slopes simulated for a variety of boundary conditions by the ECHAM atmospheric general circulation model equipped with the explicit modeling of water-stable isotopes (Hoffmann, personal communication). Modern: control present-day simulation; preindustrial simulation; simulations for boundary conditions representative of 6, 11, 14, 16, 21, and 175 ka.

References

- Berger A and Loutre MF (2002) An exceptionally long interglacial ahead? *Science* 297: 1287–1288.
- Blunier T, et al. (1997) Timing of the Antarctic Cold Reversal and the atmospheric CO₂ increase with respect to the Younger Dryas event. *Geophysical Research Letters* 24: 2683–2686.
- Blunier T, et al. (2004) What was the surface temperature in central Antarctica during the Last Glacial Maximum? *Earth and Planetary Science Letters* 218: 379–388.
- Blunier T and Brook E (2001) Timing of millennial-scale climate change in Antarctica and Greenland during the last glacial period. *Science* 109–112.
- Brook EJ, et al. (2005) Timing of millennial-scale climate change at Siple Dome, West Antarctica, during the last glacial period. *Quaternary Science Reviews* 24: 1333–1343.
- Caillon N, et al. (2001) Estimation of temperature change and of gas age–ice age difference, 108 kyr BP, at Vostok, Antarctica. *Journal of Geophysical Research* 106: 31893–31901.
- Dansgaard W (1953) The abundance of ¹⁸O in atmospheric water and water vapour. *Tellus* 5: 461–469.
- Delaygue G, et al. (2000) The origin of the Antarctic precipitation: A modelling approach. *Tellus* 27: 19–36.
- Delmotte M, et al. (2000) Seasonal deuterium excess signal at Law Dome, coastal Eastern Antarctica: Observations and modelling. *Journal of Geophysical Research* 105: 7187–7197.
- Ekaykin A, et al. (2004) The changes in isotope composition and accumulation of snow at Vostok station, East Antarctica, over the past 200 years. *Annals of Glaciology* 39: 569–575.
- EPICA-Community-Members (2004) Eight glacial cycles from an Antarctic ice core. *Nature* 429: 623–628.
- EPICA-Community-Members (2006) One-to-one coupling of glacial climate variability in Greenland and Antarctica. *Nature* 444: <http://dx.doi.org/10.1038/nature05301>. 9 November 2006.
- Goosse H, et al. (2004) A delayed medieval warm period in the Southern Ocean? *Geophysical Research Letters* 31: 1029–1032.
- Johnsen SJ, et al. (1972) Oxygen isotope profiles through the Antarctic and Greenland ice sheets. *Nature* 235: 429–434.
- Jouzel J, et al. (1982) Deuterium excess in an East Antarctic ice core suggests higher relative humidity at the oceanic surface during the last glacial maximum. *Nature* 299: 688–691.
- Jouzel J, et al. (1991) Simulations of the HDO and H₂18O atmospheric cycles using the NASA GISS general circulation model: Sensitivity experiment for present day conditions. *Journal of Geophysical Research* 96: 7495–7507.
- Jouzel J, et al. (1995) The two-step shape and timing of the last deglaciation in Antarctica. *Climate Dynamics* 11: 151–161.
- Jouzel J, et al. (2001) A new 27 Ky high resolution East Antarctic climate record. *Geophysical Research Letters* 28: 3199–3202.
- Jouzel J, et al. (2003) Magnitude of the Isotope/Temperature scaling for interpretation of central Antarctic ice cores. *Journal of Geophysical Research* 108: 1029–1046.
- Landais A, et al. (2005) The glacial inception recorded in the NorthGRIP Greenland ice core: Information from air isotopic measurements. *Climate Dynamics* <http://dx.doi.org/10.1007/s00382-005-0063-y>.
- Liesicki LE and Raymo ME (2005) A Pliocene–Pleistocene stack of 57 globally distributed benthic δ¹⁸O records. *Paleoceanography* 20: <http://dx.doi.org/10.1029/2004PA001071>.
- Lorius C, et al. (1979) A 30,000-Yr isotope record from Antarctic ice. *Nature* 280: 644–648.
- Lorius C and Merlivat L (1977) Distribution of mean surface stable isotope values in East Antarctica. Observed changes with depth in a coastal area. In: IAHS (ed.) *Isotopes and impurities in snow and ice*, pp. 125–137. Proceedings of the Grenoble Symposium Aug./Sep. 1975, Vienna: IAHS.
- Masson V, et al. (2000) Holocene climate variability in Antarctica based on 11 ice-core isotopic record. *Quaternary Research* 54: 348–358.
- Masson-Delmotte V, et al. (2003) Recent southern Indian Ocean climate variability inferred from a Law Dome ice core: New insights for the interpretation of coastal Antarctic isotopic records. *Climate Dynamics* 21: 151–166.
- Masson-Delmotte V, et al. (2004) Common millennial scale variability of Antarctic and southern ocean temperatures during the past 5000 years reconstructed from EPICA Dome C ice core. *The Holocene* 14: 145–151.
- Masson-Delmotte V, et al. (2008) A review of Antarctic surface snow isotopic composition: Observations, atmospheric circulation, and isotopic modeling. *Journal of Climate* <http://dx.doi.org/10.1175/2007JCLI2139.1>.
- Morgan V, et al. (2002) Relative timing of Deglacial Climate events in Antarctica and Greenland. *Science* 297: 1862–1864.
- Morgan V and Van Ommen T (1997) Seasonality in late holocene climate from ice core records. *The Holocene* 7(3): 351–354.
- NorthGRIP-community-members (2004) High resolution climate record of the northern hemisphere reaching into last interglacial period. *Nature* 431: 147–151.
- Pedro JB, van Ommen TD, Rasmussen SO, et al. (2011) The last deglaciation: Timing the bipolar seesaw. *Clim. Past* 7: 671–683. <http://dx.doi.org/10.5194/cp-7-671-2011>.
- Petit JR, et al. (1999) Climate and Atmospheric History of the Past 420000 years from the Vostok Ice Core, Antarctica. *Nature* 399: 429–436.
- Schneider DP, et al. (2006) *Geophysical Research Letters*, vol. 33, L16707, <http://dx.doi.org/10.1029/2006GL027057>, 2006.
- Steig EJ, et al. (1998) Synchronous climate changes in Antarctica and the North Atlantic. *Science* 282: 92–95.
- Stenni B, et al. (2003) A high resolution site and source late glacial temperature record derived from the EPICA Dome C isotope records (East Antarctica). *EPSL* 217: 183–195.
- van Ommen TD and Morgan V (1997) Calibrating the ice core paleothermometer using seasonality. *Journal of Geophysical Research* 102: 9351–9357.

- Vimeux F, et al. (1999) Glacial-interglacial changes in ocean surface conditions in the Southern Hemisphere. *Nature* 398: 410–413.
- Vimeux F, et al. (2001a) Holocene hydrological cycles changes in southern hemisphere documented in Antarctic deuterium-excess records. *Climate Dynamics* 17: 503–513.
- Vimeux F, Masson V, Delaygue G, Jouzel J, Petit J-R, and Stievenard M (2001b) A 420,000 year deuterium excess record from East Antarctica: Information on past changes in the origin of precipitation at Vostok. *Journal of Geophysical Research* 106: 31863–31873.
- Watanabe O, et al. (2003) Homogeneous climate variability across East Antarctica over the past three glacial cycles. *Nature* 422: 509–512.
- Werner M, et al. (2001) Isotopic composition and origin of polar precipitation in present and glacial climate simulations. *Tellus* 53B: 53–71.

Greenland Stable Isotopes

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The Early Isotope Work

By doing stable isotope investigations of frontal precipitation over Copenhagen, Denmark, in the early 1950s, Willi Dansgaard discovered a correlation between condensation temperatures and isotopic ratios. Subsequently, Dansgaard proposed in a groundbreaking paper that paleoclimate could be studied by doing stable isotope measurements on Greenland ice cores (Dansgaard, 1954). At this time, no ice cores were available from Greenland for such investigations, and the immense scientific scope of this proposal was hardly realized even 15 years later when the first results from the Camp Century core became available. Carl Benson of the US Army Snow, Ice, and Permafrost Research Establishment (SIPRE) was, however, already doing pit studies in Northwest and Central Greenland in the early and mid-1950s (Benson, 1962). In cooperation with Epstein and Sharp, he discovered the very strong annual isotopic cycle in the Greenland snow pack. Eventually, this cycle became most useful for measuring accumulation rates and dating ice cores. When the Site 2 ice core (see Figure 1), which was drilled by SIPRE in preparation for the International Geophysical Year's Antarctic activities in 1957, became available in 1955, Chester C Langway, also from SIPRE, initiated the modern era of multidisciplinary ice-core studies including detailed isotope profiles deep down in the core (Langway, 1967). The Site 2 isotope data (Epstein and Sharp, 1959) showed a clear annual cycle below the firn layers, but smoothed compared with the isotope data from Benson's pit studies (Langway, 1967). Thus, Langway discovered the effect of firn diffusion discussed below.

Willi Dansgaard continued his studies in Greenland and discovered the isotopic latitude effect for Greenland and the strong correlation between mean annual temperatures and isotopic values on the ice cap (Dansgaard, 1961). He also used the Rayleigh condensation model to improve the understanding of the physical basis for the climatic information contained in natural precipitation. Most of his earlier results are summarized in his classic Tellus paper (Dansgaard, 1964) where he also provided the first results from the International Atomic Energy Agency (IAEA) precipitation network and introduced the concept of deuterium excess.

The Camp Century deep ice core was recovered by Lyle Hansen and his coworkers from the US Army Cold Region Research and Engineering Laboratories in 1966. Two years later, Dansgaard suggested that Langway do a comprehensive isotope study on the core, a proposal that Langway accepted. This most fruitful cooperation was later extended with the group of Hans Oeschger from Bern, Switzerland, and became the basis for the Greenland Ice Sheet Program (GISP) that managed to recover a deep core from Dye-3 South Greenland with a new Danish deep drill in 1981. When the first Dye-3 isotopic results were published (Dansgaard et al., 1982), the immense impact of the isotopic paleostudies on Greenland

deep ice cores was firmly established. In the early 1970s, the Danish group also built a unique inlet system for running 512 daily mass-spectrometric oxygen-isotope measurements. The original aim was to date the Dye-3 core by counting annual isotopic cycles through most of the Holocene by measuring a total of 80 000 samples. However, this system greatly strengthened the isotope work done with subsequent Greenland ice cores and over one million samples were analyzed using it.

Shallow Ice Cores – The Role of Diffusion

A large number of ice cores reaching depths of 100–200 m have been drilled at various locations on the Greenland ice sheet. Such cores offer a time span ranging from a few centuries to thousands of years, depending on core depth and local accumulation rates. As instrumental climatic observations are limited to the past two centuries for most of the Northern Hemisphere, the records of $\delta^{18}\text{O}$ from shallow ice cores are a valuable asset when investigating relationships between climate and $\delta^{18}\text{O}$. Due to the multiple cores available, it is possible to stack the records and thereby reduce the glaciological noise in the $\delta^{18}\text{O}$ data. Using cores from various locations, it is also possible to resolve local climatic variations affecting only parts of the ice sheet. The investigation of $\delta^{18}\text{O}$ records from the upper part of the ice cores is complicated by water vapor diffusion in the pores of snow and firn in the top 60–70 m of the ice sheet. This diffusion tends to smooth the high-frequency oscillations of the $\delta^{18}\text{O}$. Figures 2 and 3 show the diffusion-related decay of the annual $\delta^{18}\text{O}$ cycle for a central Greenland shallow core.

The Climate Signal in Seasonal $\delta^{18}\text{O}$

The close relationship between $\delta^{18}\text{O}$ and site temperature makes $\delta^{18}\text{O}$ a valuable proxy both for Greenland climate and Northern Hemisphere atmospheric circulation patterns. To exploit the potential of Greenland $\delta^{18}\text{O}$ records fully Vinther et al. (2003, 2010) have shown that it is essential to consider seasonal oscillations in the $\delta^{18}\text{O}$ records.

In the vast areas of the Greenland ice sheet where annual accumulation exceeds 20 cm of ice equivalent, seasonal oscillations of $\delta^{18}\text{O}$ can be retrieved through most of the Holocene. In areas of lower accumulation, the seasonal $\delta^{18}\text{O}$ cycles are obliterated by the diffusion in the upper snow and firn layers.

As can be seen in Figures 2 and 3, the seasonal cycle of $\delta^{18}\text{O}$ is influenced by diffusion even in areas with accumulation rates larger than 20-cm ice equivalent per year. In the top 60–70 m of the ice sheet, a marked decay in amplitude of the $\delta^{18}\text{O}$ cycle can be observed. If a climatic interpretation of seasonal $\delta^{18}\text{O}$ values is to be pursued, it is necessary to

understand the diffusion process that masks the seasonal oscillations in the $\delta^{18}\text{O}$ signal.

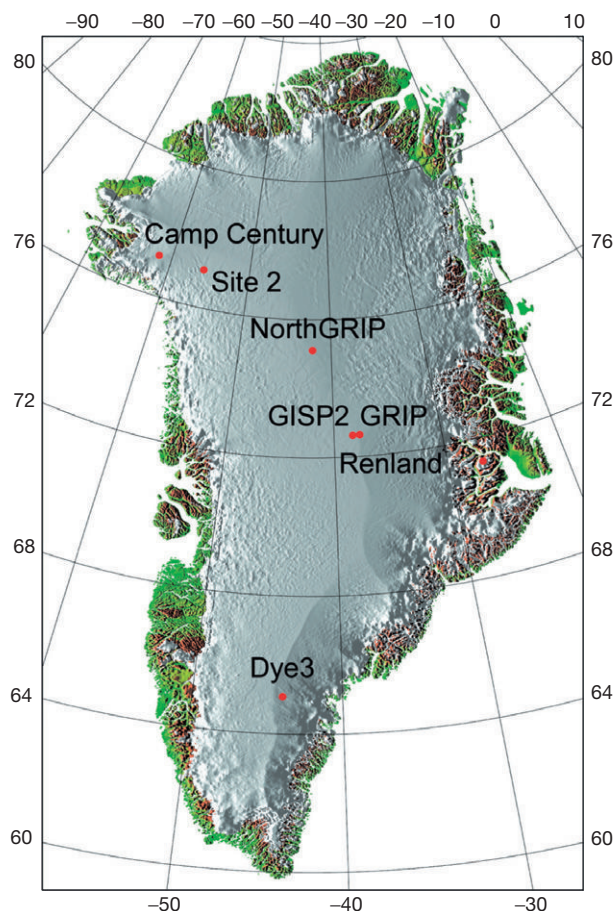


Figure 1 Locations of selected Greenland ice cores. The Site 2 core was the first ice core drilled in Greenland. The Camp Century, Dye-3, GISP2, GRIP, NGRIP, and Renland cores all contain ice from the Holocene, Glacial, and Eemian periods.

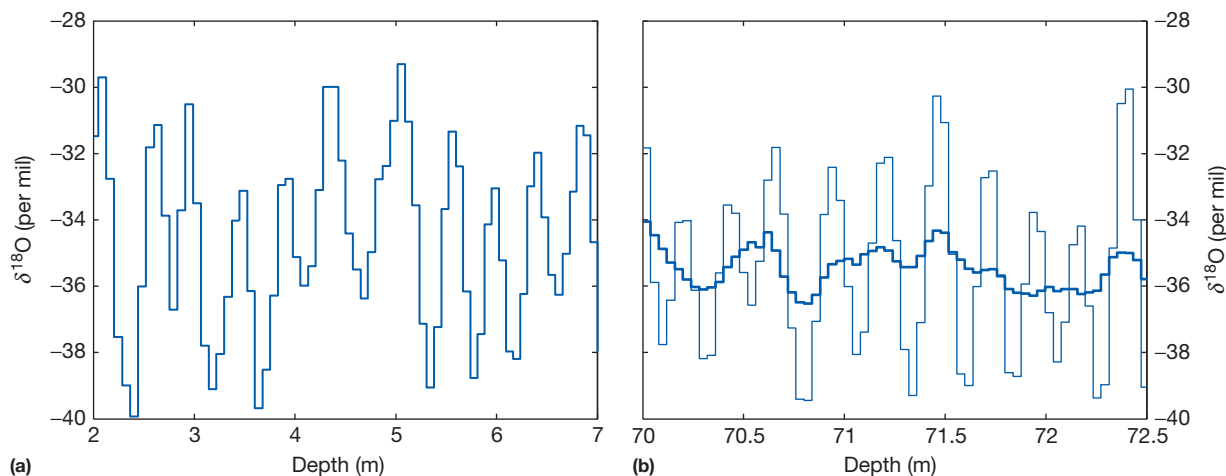


Figure 2 Measurements of $\delta^{18}\text{O}$ from different depths in the Site E central Greenland ice core. Left: near the surface before firn diffusion dampens the annual cycle. Right: below the diffusion zone, the annual cycle is significantly dampened. The thin curve is diffusion-corrected data. Annual accumulation at Site E amounts to 21 cm of ice equivalent.

In a publication by Johnsen (1977), the first step was taken in comprehending the physics behind the diffusion process. Later papers by Merlivat and Jouzel (1979), Herron and Langway (1980), and Schwander et al. (1988) led to the Johnsen et al. (2000) paper in which a physical model covering both $\delta^{18}\text{O}$ and δD diffusion was presented and validated. The Johnsen et al. (2000) model predictions of the diffusion-driven dampening of the annual $\delta^{18}\text{O}$ cycle can be seen in Figure 3.

Using the Johnsen et al. (2000) diffusion model and the mathematical formulation of the diffusion problem presented in Johnsen (1977), it is possible to correct $\delta^{18}\text{O}$ series for diffusion. An example of diffusion-corrected $\delta^{18}\text{O}$ data can be seen in Figure 2(b). Having diffusion-corrected data, the underlying seasonally resolved climatic signal can be retrieved from ice-core $\delta^{18}\text{O}$ records.

It is easy to demonstrate the importance of correcting the seasonal $\delta^{18}\text{O}$ records for diffusion before any climatic interpretation is pursued. Four different central Greenland ice cores were drilled in 1984/1985 about 500 km from the coastal Tasilaq meteorological station. Table 1 demonstrates that correlations between winter season $\delta^{18}\text{O}$ and December-to-March temperatures at the Tasilaq station (Laursen, 2003) are greatly improved by diffusion correcting the records. Furthermore, it can be seen that the ice-core records from the areas of modest accumulation (Sites E and G) benefit the most from the diffusion correction.

Using the approach of diffusion correction on several Greenland ice cores, Vinther et al. (2003, 2010) demonstrated that winter season $\delta^{18}\text{O}$ from Greenland ice cores contained both a distinct Greenland temperature signal and the signature of the North Atlantic Oscillation (NAO), a large-scale atmospheric flow pattern. In fact, Greenland annual average temperatures were found to be better captured by winter season $\delta^{18}\text{O}$ data than annual average $\delta^{18}\text{O}$ data.

A linear regression between the multi-ice core winter $\delta^{18}\text{O}$ signal and Northern Hemisphere December-to-March sea level pressure (slp) is seen in Figure 4. The NAO pattern with opposite pressure trends over Iceland and the Azores is clearly seen in Figure 4. The regression was carried out for the years

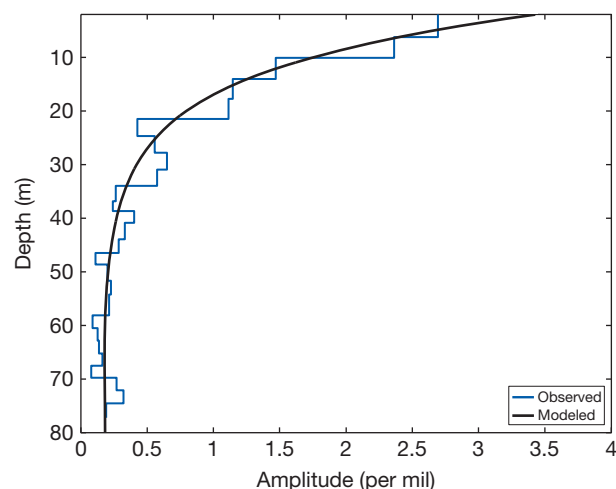


Figure 3 Amplitude of the annual $\delta^{18}\text{O}$ cycle in the Site E central Greenland ice core and model estimate.

Table 1 Correlations (r) of winter $\delta^{18}\text{O}$ with Tasiilaq December-to-March temperatures (1895–1982)

Ice core	Site A	Site B	Site E	Site G	Mean ABEG
Accumulation (ice equivalent; cm)	29	31	21	25	–
r (before diffusion correction)	0.29	0.43	0.25	0.21	0.36
r (after diffusion correction)	0.33	0.50	0.41	0.37	0.52

1872–1970 and the slp-data are from the HadSLP1 dataset (Basnett and Parker, 1997).

Deep Ice Cores – Climate During 125 Millennia

During the past 40 years a total of five deep ice cores have been drilled all the way through the Greenland ice sheet. All five records contain ice from the entire Holocene, the last glaciation, and the Eemian. A sixth ice core drilled through the small east coast Renland ice cap also reaches back into the Eemian (for drill site locations see Figure 1). Measurements of $\delta^{18}\text{O}$ have been carried out on the full length of all six ice-core records. From Figure 5 it is clear that the six $\delta^{18}\text{O}$ records are very similar for the past 40 000 years. Below 40 000 years, the Dye-3 record is disturbed by ice flow, as is the Renland core below 60 000 years. The Greenland Icecore Project (GRIP) and GISP2 records are affected by flow-induced foldings below 105 000 years. The North Greenland Icecore Project (NGRIP) core presents the only $\delta^{18}\text{O}$ record known to be stratigraphically continuous into the Eemian. However, Eemian ice is found in all six ice cores.

Timescales for Deep Ice Cores

A prerequisite for the climatic interpretation of the deep ice cores is having a well-established depth–age relation. For most

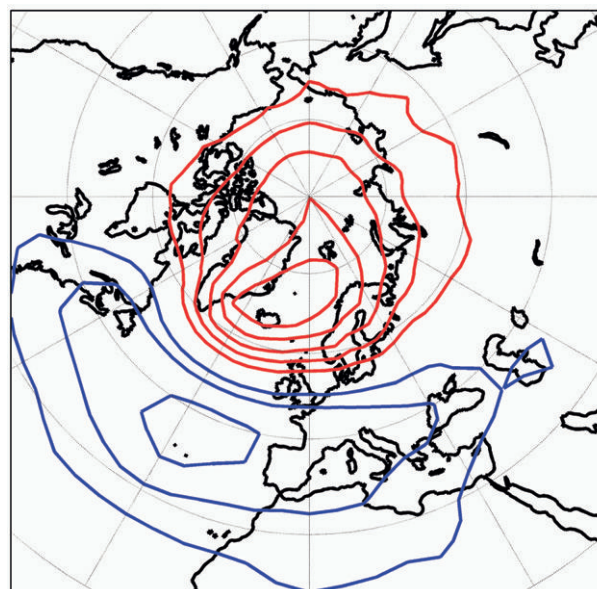


Figure 4 December–March slp-response to a one-sigma increase in winter $\delta^{18}\text{O}$ in a series based on multiple Greenland ice cores. Contour interval is 0.4 hPa, red contours are positive while blue are negative.

of the Holocene, counting annual layers using the seasonal oscillations in $\delta^{18}\text{O}$ is a very robust way of obtaining a time-scale for the Greenland ice cores. During the glacial, generally lower accumulation rates lead to diffusion-caused obliteration of the annual $\delta^{18}\text{O}$ cycles, rendering annual layer counting impossible.

The first approach to date the glacial part of the ice cores relied on analytical 2D flow model estimates of the progressive thinning of annual layers (Dansgaard and Johnsen, 1969). The flow model presented by Dansgaard and Johnsen (1969) (D–J model) proved to be very adaptive and Johnsen and Dansgaard (1992) refined the analytical D–J model to account for both increased shear near the bottom of the ice sheet and sliding of the ice sheet over the bedrock. The D–J model can also be modified to include flow due to bottom melting affecting the ice sheet.

To include temporal variations of the accumulation rate, a numerical version of the D–J model was required. The first effort to create such a model was made by Johnsen et al. (1992). Numerical D–J flow models have been used to provide timescales to both the GRIP and the NGRIP ice cores (NGRIP members, 2004). These models require past accumulation rates as an input parameter. This requirement can be fulfilled by using the relationships between $\delta^{18}\text{O}$ and accumulation rate at the drill sites. Figure 6 shows the Dye-3, GRIP, and NGRIP $\delta^{18}\text{O}$ profiles on timescales derived from numerical D–J flow models. As the bottom-most part of the Dye-3 core is disturbed by ice flow, only the past 40 000 years of the Dye-3 profile is shown. The similarity of the three profiles is striking.

The first effort to count annual layers during the glacial was carried out on the GISP2 core (Meese et al., 1997). Seasonally changing features in the visual stratigraphy of the GISP2 core were used for counting the layers. Surprisingly, the glacial GISP2 timescale derived from this methodology did not show a persistent relationship between the $\delta^{18}\text{O}$ and the

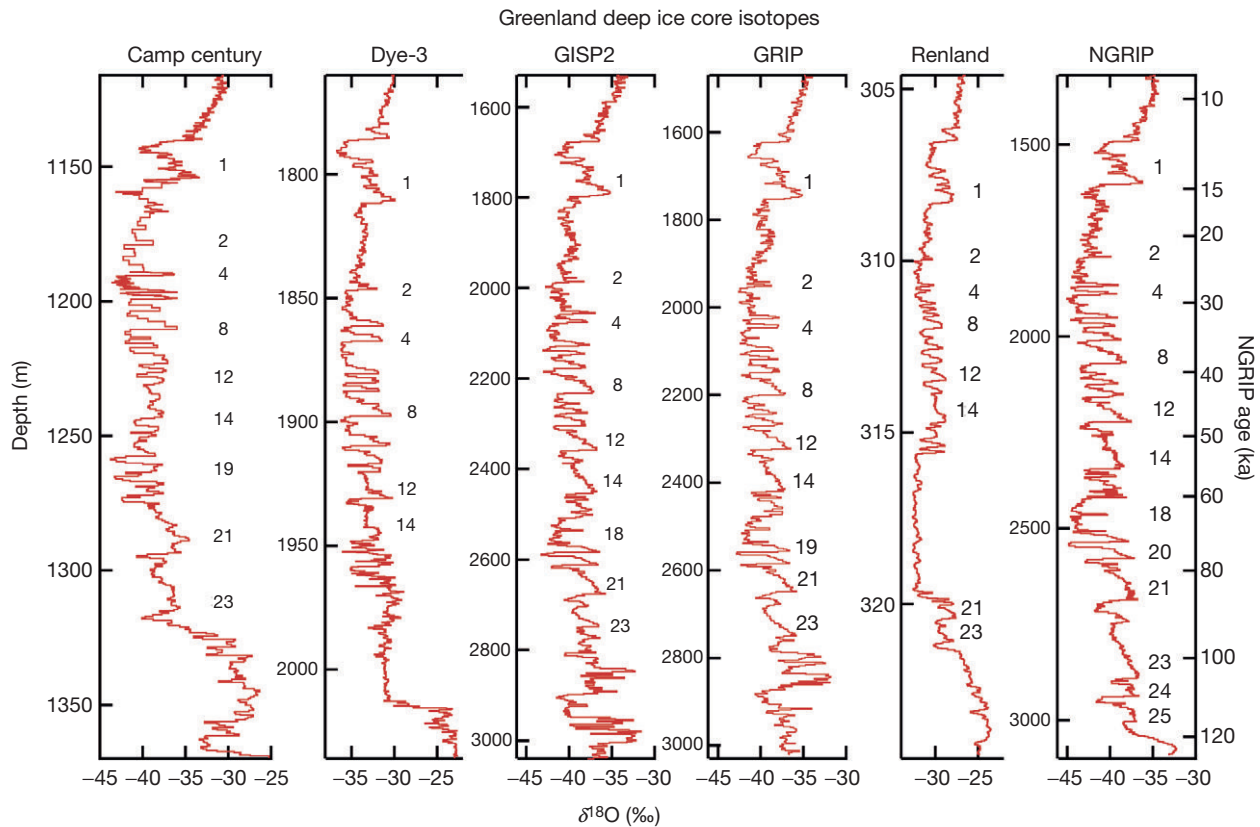


Figure 5 Early Holocene, Glacial, and Eemian $\delta^{18}\text{O}$ records from six Greenland ice cores are shown. The cores are all on depth scales. The NGRIP timescale is shown, and the numbers on the left side of the scale give the most prominent interstadials.

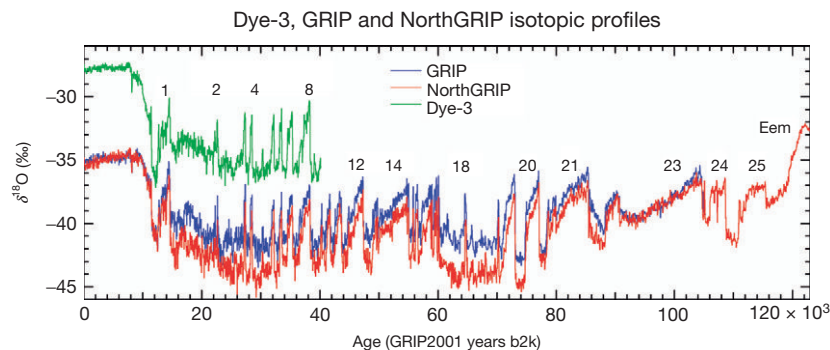


Figure 6 Profiles of $\delta^{18}\text{O}$ from the Dye-3, GRIP, and NGRIP cores, on the ss09sea flow-modeled timescale. Numbers are given for the most prominent interstadials. b2k, before AD 2000.

accumulation rate back in time. Such a relation is underpinned both by modern observations and the following simple argument: a change in $\delta^{18}\text{O}$ is associated with a change in site air temperature, and as the air temperature determines the maximum water vapor content of the air, the amount of moisture reaching the site will be related to observed $\delta^{18}\text{O}$ values.

Based on a new effort to date the glacial part of the NGRIP core using both seasonal oscillations in chemical impurities and visual stratigraphy, it is now believed that the apparent breakdown of the $\delta^{18}\text{O}$ /accumulation relationship observed in the GISP2 core is an artifact caused by erroneous

interpretation of the GISP2 visible strata (Andersen et al., 2006; Svensson et al., 2006).

The Holocene

The major part of the deep ice cores consists of Holocene ice. Looking at Figure 7, it is obvious that millennial scale trends in the ice core records are different during the course of the Holocene. Vinther et al. (2009) have shown that these differences are most likely due to a combination of past Greenland ice sheet elevation change and upstream effects (i.e., that ice in

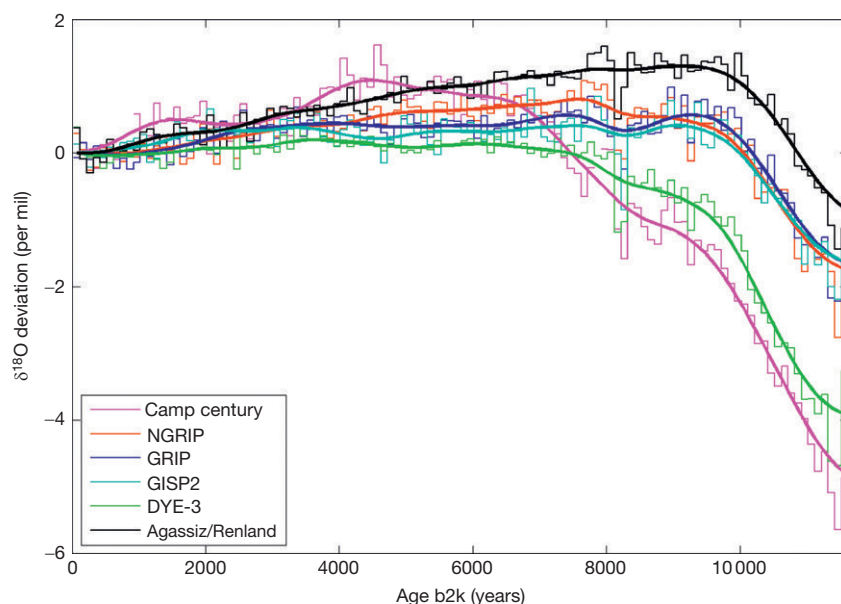


Figure 7 Holocene profiles of $\delta^{18}\text{O}$ deviations with respect to present. Heavy lines show millennial scale trends in the data. All profiles are shown on the GICC05 time scale, except for GISP2 $\delta^{18}\text{O}$ which is shown on the original GISP2 time scale. b2k, before AD 2000.

the lower strata originates from higher elevation positions upstream from the drill site). Furthermore, [Vinther et al. \(2009\)](#) demonstrated, that the average $\delta^{18}\text{O}$ from the two small ice caps of Renland (eastern Greenland) and Agassiz (Ellesmere Island north west of Greenland) can be used to establish the Holocene climatic $\delta^{18}\text{O}$ evolution, if corrected for post-glacial rebound (the corrected average $\delta^{18}\text{O}$ is shown as the black curve in [Figure 7](#)).

The differences between the Renland/Agassiz $\delta^{18}\text{O}$ average and $\delta^{18}\text{O}$ from the other five ice cores shown in [Figure 7](#) can thus be converted directly to past changes in depositional elevation at the five locations using the observed $\delta^{18}\text{O}$ -elevation relationship for Greenland.

A conversion of the Renland/Agassiz $\delta^{18}\text{O}$ average to past Greenland temperatures show a climatic optimum 2–3 °C warmer than present ([Vinther et al., 2009](#)) in good agreement with results from [Dahl-Jensen et al. \(1998\)](#) showing that bore-hole temperature measurements at GRIP and Dye-3 imply that temperatures on the ice sheet were, 3 °C higher than present during the climatic optimum. All but one of the $\delta^{18}\text{O}$ records shown in [Figure 7](#) are plotted on the Greenland Ice Core Chronology 2005 (GICC05). GICC05 is a newly counted time-scale that synchronizes the ice cores using common volcanic reference horizons ([Rasmussen et al., 2006; Vinther et al., 2006, 2008](#)). Exploiting the synchronization of the $\delta^{18}\text{O}$ records ([Rasmussen et al., 2007](#)) showed that the only significant common events in the early Holocene Dye-3, GRIP, and NGRIP records are the decreases in $\delta^{18}\text{O}$ some 8200, 9300, and 11400 years ago.

These events are known as the 8.2 ka and 9.3 ka cold events and the Preboreal oscillation, and they are observed in ocean sediment cores, tree ring data, and stalagmite records from all over the Northern Hemisphere. The events are believed to be associated with very large freshwater outbursts from the decaying Laurentide ice sheet. The large amount of freshwaters would temporarily weaken the North Atlantic overturning

circulation thereby diminishing the oceanic heat transport towards the polar region.

The Last Glaciation

The abrupt shifts in $\delta^{18}\text{O}$ leading to the warm Greenland interstadials ([Dansgaard et al., 1993; Johnsen et al., 1992](#)) are often referred to as Dansgaard–Oeschger events. These events are the most prominent features observed in the glacial parts of all Greenland ice cores (see [Figures 5](#) and [6](#)). The shifts are most probably due to large-scale reorganizations of the atmospheric and oceanic circulations during the glacial. Hence, full glacial (stadial) conditions are probably related to a shutdown of the North Atlantic overturning circulation, while the interstadials are warmed by the return of the overturning circulation. Changes from stadial climatic conditions to the 10–15 °C warmer interstadial climate are believed to have happened within a decade or two. The glacial climate is therefore characterized by a degree of instability that cannot be found during the Holocene. This instability is likely to be caused by the N–S shifts of the atmospheric polar front that follows the oceanic polar front ([Ruddiman and McIntyre, 1981](#)) in the glacial. The $\delta^{18}\text{O}$ at the polar front is close to –8‰ [Dansgaard and Tauber \(1969\)](#), which is also likely valid in the glacial. The Greenland $\delta^{18}\text{O}$ values are mainly driven by the latitudinal gradient of –0.54‰ $\delta^{18}\text{O}/\text{lat}$, the distance to the polar front and elevation changes [Dansgaard \(1961\)](#) and [Johnsen et al. \(1989\)](#).

[Johnsen et al. \(1997\)](#) estimated the $\delta^{18}\text{O}$ short-term annual RMS (Root Mean Square) variability from the GRIP ice core ([Figure 5](#) in that paper). The records of annual variability ([Figure 5\(c\)](#)) and the $\delta^{18}\text{O}$ ([Figure 5\(a\)](#)) are in anticorrelation. The RMS variability for Holocene, interstadials, and stadials are 0.9‰, 1.8‰, and 3.4‰, respectively. The increased variability is partly due to lower accumulation rate, in the interstadials and stadials, and partly due to the rapid N–S shifts of the polar front during these cold periods.

Table 2 Average values and differences of $\delta^{18}\text{O}$ during different climatic periods (all values are given in ‰)

Drill site	Present (P)	Stadial (S)	Eemian (E)	P–S	E–P
Camp century	–29.5	–41.9	–27.0	12.4	2.5
Dye-3	–27.6	–36.1	–22.9	8.5	4.7
GISP2	–35.0	–42.2	–32.3	7.2	2.7
GRIP	–35.1	–411.9	–31.6	6.6	3.5
NGRIP	–35.4	–44.5	–32.3	9.1	3.1
Renland	–27.3	–32.1	–23.8	5.0	3.5

Johnsen et al. (1992) found that an intercomparison between glacial $\delta^{18}\text{O}$ values observed in different Greenland ice cores shows important differences in the size of the $\delta^{18}\text{O}$ increase associated with the shift from stadial to Holocene climatic conditions. The Camp Century core shows the largest $\delta^{18}\text{O}$ changes while the changes at Renland are the most modest (see Table 2 and Figure 5). It is most likely that the large changes in Camp Century (and NGRIP) $\delta^{18}\text{O}$ are related to elevation changes, but the changes could also be partly due to a strongly depleted northwesterly source of precipitation during the stadials at those sites. Such a source would stem from a split in the polar jet-stream, with branches both north and south of the Laurentide ice sheet. Precipitation arriving from the north-west would not be able to reach the Renland ice cap as it is situated east of the main ice sheet, and therefore, is shielded from any such precipitation. Hence, the relatively modest decrease in Renland $\delta^{18}\text{O}$ values during the stadials can be understood in terms of the Renland ice cap not receiving precipitation from the depleted northern branch of a split jet-stream.

Recent atmospheric modeling efforts suggesting a split jet-stream during the glacial in the winter (i.e., Bromwich et al., 2004) therefore enjoy significant support from observations of stadial $\delta^{18}\text{O}$ in Greenland ice cores. The whole glacial period in Camp Century is $\sim 3\text{‰}$ colder compared to the present than seen in NGRIP. This difference is most likely caused by $\sim 500\text{--}600$ m higher elevation at the Camp Century compared to the NGRIP elevation.

The Eemian

Ice from the last interglacial, the Eemian, has been found near the bedrock in all deep ice cores retrieved from the Greenland ice sheet (see Figure 5). It is, therefore, safe to conclude that most of Greenland was covered by ice during the Eemian. By using the $\delta^{18}\text{O}$ values observed in Eemian ice, it is possible to further quantify the changes in the size of the Greenland ice sheet during the Eemian.

Comparing present day $\delta^{18}\text{O}$ values to those observed in the Eemian ice, it is clear that the Eemian had 2.5–3.5‰ higher $\delta^{18}\text{O}$ values than presently observed all across Greenland (Table 2 and Figure 5) except for at the southern Dye-3 site. At Dye-3, Eemian $\delta^{18}\text{O}$ was 4.7‰ higher than recent values. The observed $\delta^{18}\text{O}$ value at a site on the Greenland ice sheet is dependent on elevation (the gradient is approximately -0.6‰ per 100 m altitude according to Johnsen et al., 1989). It can therefore be inferred from the rather uniform ($\sim 3\text{‰}$) increase

in Eemian $\delta^{18}\text{O}$, that the differences in elevation between the Camp Century, GRIP, GISP2, NGRIP, and Renland sites during the Eemian were almost equal to those presently observed. Furthermore, the elevated Eemian $\delta^{18}\text{O}$ values at the Dye-3 site indicate that the southern part of the Greenland ice sheet was lowered by some 400 m relative to the rest of the ice sheet (the 1.5‰ extra change at Dye-3 added to a 1‰ change due to ice flow induced extra lowering of the Eemian $\delta^{18}\text{O}$). It is more difficult to access the absolute elevation of the main ice sheet during the Eemian. The general $\sim 3\text{‰}$ increase in $\delta^{18}\text{O}$ could be attributed both to changes in elevation and changes in temperature, as the Greenland climate during the Eemian was significantly warmer than the present climate.

The fact that the small Renland ice cap experienced almost the same increase in $\delta^{18}\text{O}$ as most of the Greenland ice sheet does however put constraints on the possible changes in elevation. The Renland ice cap is only about 330 m thick, and due to its position on a high elevation plateau, it cannot grow any thicker. Furthermore, the fact that a more than 3-m-thick layer of Eemian ice has been found close to bedrock at Renland (see Figure 5) does suggest that the Renland ice cap was well preserved during the Eemian. The ice flow-driven thinning rate in the Renland ice cap estimated by Johnsen and Dansgaard (1992), probably could not have allowed the preservation of a 3-m thick slab of Eemian ice, had the Renland ice cap been significantly thinner during the Eemian than today.

It is, therefore, reasonable to assume that the Eemian thickness of the Renland ice cap was not very far from the 330 m observed today, thus imposing narrow constraints on thickness changes at Camp Century, GISP2, GRIP, and NGRIP as well. Given this information, it is obvious that most (if not all) of the 2.5–3.5‰ Eemian increase in $\delta^{18}\text{O}$ should be attributed to temperature change; not to elevation change. Hence, it is probably only the southern part of the Greenland ice sheet that experienced significant decreases in thickness during the Eemian. In fact it could be argued that a slight increase in the thickness of the northern part of the Greenland ice sheet is supported by the $\delta^{18}\text{O}$ data, as NGRIP and Camp Century Eemian $\delta^{18}\text{O}$ seems to have increased a bit less than the Renland $\delta^{18}\text{O}$.

See also: Ice Core Methods: Borehole Temperature Records; Chronologies; Overview; Stable Isotopes. Ice Core Records: Antarctic Stable Isotopes; Correlations Between Greenland and Antarctica. Ice Cores: Dynamics of the Greenland Ice Sheet; History of Research, Greenland and Antarctica.

References

- Andersen KK, Svensson A, Rasmussen SO, et al. (2006) The Greenland ice core chronology 2005, 15–42 kyr, Part 1: Constructing the time scale. *Quaternary Science Reviews* 25(23–24): 3246–3257. Shackleton special issue.
- Basnett T and Parker D (1997) *Development of the Global Mean Sea Level Pressure Data Set GMSLP2*. UK: Hadley Centre. Climate Research Technical Note 79.
- Benson CS (1962) *Stratigraphic Studies in the Snow and Firn of the Greenland Ice Sheet*, p. 63. SIPRE Research Report 70.
- Bromwich DH, Toracinta ER, Wei H, Oglesby RJ, Fastook JL, and Hughes TJ (2004) Polar MM5 simulations of the winter climate of the Laurentide Ice Sheet at the LGM. *Journal of Climate* 17: 3415–3433.
- Dahl-Jensen D, Mosegaard K, Gundestrup N, et al. (1998) Past temperatures directly from the Greenland ice sheet. *Science* 282: 268–271.
- Dansgaard W (1954) The O^{18} -abundance in fresh water. *Geochimica et Cosmochimica Acta* 6: 241–260.
- Dansgaard W (1961) The isotopic composition of natural waters. *Meddelelser om Grønland* 165: 120.
- Dansgaard W (1964) Stable isotopes in precipitation. *Tellus* 16: 436–468.
- Dansgaard W, Clausen HB, Gundestrup N, et al. (1982) A new Greenland deep ice core. *Science* 218: 1273–1277.
- Dansgaard W and Johnsen SJ (1969) A flow model and a time scale for the ice core from Camp Century, Greenland. *Journal of Glaciology* 8: 215–223.
- Dansgaard W, Johnsen SJ, Clausen HB, et al. (1993) Evidence for general instability of past climate from a 250-kyr ice-core record. *Nature* 364: 218–220.
- Dansgaard W and Tauber H (1969) Glacier oxygen-18 content and Pleistocene ocean temperatures. *Science* 166(3904): 499–502.
- Epstein S and Sharp RP (1959) Oxygen isotope studies. *Transactions, American Geophysical Union* 40: 81–84.
- Herron MM and Langway CC Jr. (1980) Firn densification: An empirical model. *Journal of Glaciology* 25: 373–385.
- Johnsen SJ (1977) Stable isotope homogenization of polar firn and ice. In: *Proceedings on the Symposium on Isotopes and Impurities in Snow and Ice*, pp. 210–219. I.U.G. G. XVI, General Assembly, Grenoble, 1975, Washington, DC.
- Johnsen SJ, Clausen HB, Cuffey KM, Hoffmann G, Schwander J, and Creyts T (2000) Diffusion of stable isotopes in polar firn and ice: The isotope effect in firn diffusion. In: Hondoh T (ed.) *Physics of Ice Core Records*, pp. 121–140. Sapporo: Hokkaido University Press.
- Johnsen SJ, Clausen HB, Dansgaard W, et al. (1992) Irregular glacial interstadials recorded in a new Greenland ice core. *Nature* 359: 311–313.
- Johnsen SJ and Dansgaard W (1992) On flow model dating of stable isotope records from Greenland ice cores. *NATO ASI Series I* 2: 13–24.
- Johnsen SJ, Dansgaard W, and White JWC (1989) The origin of Arctic precipitation under present and glacial conditions. *Tellus* 41B: 452–468.
- Johnson KS, Gordon RM, and Coale KH (1997) What controls dissolved iron in the world ocean?. *Marine Chemistry* 57: 137–161.
- Langway CC Jr. (1967) *Stratigraphic Analysis of a Deep Ice Core from Greenland, CRREL*. Research Report 77, p. 130.
- Laursen EV (2003) *DMI Monthly Climate Data, 1873–2002, Contribution to Nordic Arctic Research Programme (NARP)*. Technical Report 03–25 from the Danish Meteorological Institute.
- Meese DA, Gow AJ, Alley RB, et al. (1997) The Greenland Ice Sheet Project 2 depth-age scale: Methods and results. *Journal of Geophysical Research* 102: 26411–26423.
- Merlivat L and Jouzel J (1979) Global climatic interpretation of the deuterium-oxygen 18 relationship for precipitation. *Journal of Geophysical Research* C8: 5029–5033.
- North Greenland Ice Core Project members (2004) High-resolution record of Northern Hemisphere climate extending into the last interglacial period. *Nature* 431: 147–151.
- Rasmussen SO, Andersen KK, Svensson AM, et al. (2006) A new Greenland ice core chronology for the last glacial termination. *Journal of Geophysical Research* 111: D06102. doi:10.1029/2005JD006079.
- Rasmussen SO, Vinther BM, Clausen HB, and Andersen KK (2007) Early Holocene oscillations recorded in three Greenland ice cores. *Quaternary Science Reviews* 26: 1907–1914.
- Ruddiman WF and McIntyre A (1981) The North Atlantic Ocean during the last deglaciation. *Palaeogeography, Palaeoclimatology, Palaeoecology* 35: 145–214.
- Schwander J, Stauffer B, and Sigg A (1988) Air mixing in firn and the age of the air at pore close-off. *Annals of Glaciology* 10: 141–145.
- Svensson A, Andersen KK, Bigler M, et al. (2006) The Greenland Ice Core Chronology 2005, 15–42 kyr, Part 2: Comparison to other records. *Quaternary Science Reviews* 25(23–24): 3258–3267. Shackleton special issue.
- Vinther BM, Buchardt SL, Clausen HB, et al. (2009) Holocene thinning of the Greenland ice sheet. *Nature* 461: 385–388.
- Vinther BM, Clausen HB, Johnsen SJ, et al. (2006) A synchronized dating of three Greenland ice cores throughout the Holocene. *Journal of Geophysical Research* 111: D13102. doi:10.1029/2005JD006921.
- Vinther BM, Clausen HB, Johnsen SJ, et al. (2008) Reply to comment on: A synchronized dating of three Greenland ice cores throughout the Holocene. *Journal of Geophysical Research* 113: D12306. doi:10.1029/2007JD009083.
- Vinther BM, Johnsen SJ, Andersen KK, Clausen HB, and Hansen AW (2003) NAO signal recorded in the stable isotopes of Greenland ice cores. *Geophysical Research Letters* 30: 1387.
- Vinther BM, Jones PD, Briffa KR, et al. (2010) Climatic signals in multiple highly resolved stable isotope records from Greenland. *Quaternary Science Reviews* 29: 522–538.

Correlations Between Greenland and Antarctica

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Introduction

Stable isotope records from ice cores in Greenland reveal a repeated pattern of abrupt warmings and coolings during the last Ice Age ([Figure 1](#)). These cycles are widely known as 'Dansgaard-Oeschger' (D-O) oscillations, named after two pioneers in the study of ice core climate records. While these abrupt changes were known from earlier data, the publication of high-quality long records from the Greenland Ice Sheet Project 2 (GISP2) and Greenland Ice Core Project (GRIP) ice cores in the early 1990s made their significance clear, and spawned large-scale efforts to document such changes elsewhere, and understand their underlying cause.

The warm/cool D-O cycles are characterized by abrupt (taking place over decades; [Alley et al., 1993](#)) and large ($>10^{\circ}\text{C}$; [Severinghaus and Brook, 1999](#)) warming, generally followed by one to several thousand years of declining temperatures, and finally abrupt cooling ([Figure 1](#)). Temperature reconstructions based on the stable isotope records are bolstered by other techniques that verify the very large temperature shifts associated with the D-O events ([Cuffey et al., 1995](#); [Severinghaus and Brook, 1999](#)).

One way to gain insight into the dynamics of the D-O events is to understand their manifestation in other parts of the globe. Doing so requires obtaining climate proxy data with time resolution appropriate for examining climate variations on time scales of at least centuries, and preferably shorter, to approach the detail provided by the ice core records. Although such records are not abundant, they do exist. Accurate dating is equally important. Comparison of ice cores with marine terrestrial sediment records can be difficult because different dating methods are used, with varying accuracy and precision, making it difficult to be sure of the temporal phasing of events in different records. Some paleoclimate archives can be dated quite accurately, for example, corals or speleothems with U-Th dating, or varved ice and sedimentary deposits, but such records are relatively rare and not always continuous.

However, because the relative timing between records is often of most interest, another set of techniques can be used. Establishing relative chronologies only requires that demonstrably synchronous stratigraphic markers be available. Oxygen isotope records from Foraminifera shells in marine sediments have been used for this purpose for some time although precision is limited by the relatively slow pace of change of the isotope signal, and the influence of local temperature on the isotopic composition of the shells. More recently, records of atmospheric gases provide a similar tool for ice cores. This article reviews correlations of polar ice core records and the results of comparing records from the Northern and Southern Hemispheres on common chronologies, primarily using gas records.

Techniques

Correlation techniques discussed here rely primarily on records of relatively long-lived atmospheric gases that have globally synchronous variations. Gases with lifetimes longer than the mixing time of the atmosphere become homogenized, and substantial changes in sources or sinks result in synchronous concentration variations. Gases used as stratigraphic markers must also exhibit variations on time scales useful for chronology – constant or very slowly varying concentrations would be of limited use. The isotopic composition of atmospheric oxygen ($\delta^{18}\text{O}_{\text{atm}}$) and the concentration of atmospheric methane ([Bender et al., 1994](#); [Blunier and Brook, 2001](#); [EPICA Community Members et al., 2006](#)) both fulfill these criteria. Many other gases might be used for additional chronological constraints but have not yet been applied in this context.

Creating a stratigraphic framework using gas records involves several steps. The starting point is an accepted chronology for one core; in most recent work the GISP2, GRIP, or North GRIP (NGRIP) records from Greenland are used. New results from the NEEM ice core in northwest Greenland will add to this data set in the next few years. These Greenland chronologies provide an age versus depth curve for the ice, with the most accurate age scales based on counting annual layers. Trapped gases are younger than the surrounding ice due to the fact that air diffuses through the firn layer (snowpack) of polar ice sheets and is trapped at depth. Glaciological models are used to account for this difference (called delta age), allowing calculation of a gas age versus depth curve. These models require estimates of temperature and snow accumulation rate; these are typically derived from proxies like stable isotopes, layer thickness measurements, ^{10}Be concentrations, and other glaciological data. Correlation or 'curve matching' of this reference gas record is then made with the analogous record in a core from another location, providing a gas age versus depth curve. This has been done by manually picking control points for correlation, or by various numerical algorithms. To obtain a chronology for parameters measured in the ice phase of this second core a delta age calculation must again be applied to translate the new curve of gas age versus depth to Ice Age versus depth.

The steps taken depend on the specific problem at hand. In some cases, a gas chronology forms the primary basis for an ice core chronology, as is the case for dating the Vostok ice core through correlations of atmospheric oxygen isotopes with the marine isotope record ([Sowers et al., 1993](#)) and subsequent dating of the 50–110,000 year section of the GISP2 core by tying it to Vostok ([Bender et al., 1994](#)). Dating Antarctic ice cores through correlation with Greenland has used the Greenland ice chronologies as a starting point ([Blunier and Brook, 2001](#); [Blunier et al., 1998](#); [Brook et al., 2005](#); [EPICA Community Members et al., 2006](#); [Steig et al., 1998](#)).

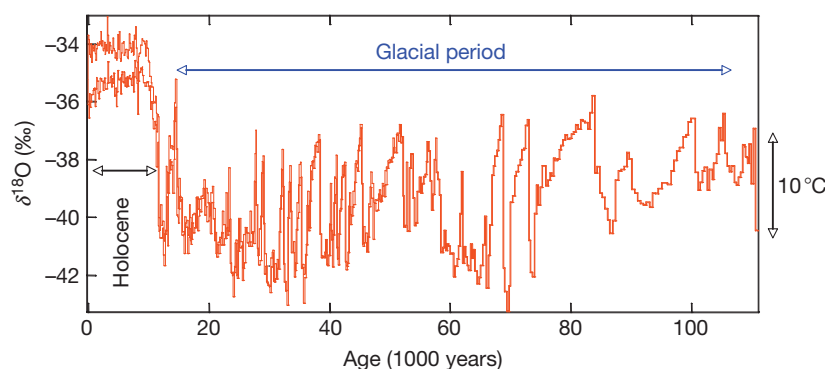


Figure 1 The water–oxygen isotope record from the GISP2 ice core (Grootes et al., 1993). The isotope ratio is expressed as per mil deviation from a standard and is a temperature proxy. Higher values indicate warmer temperatures.

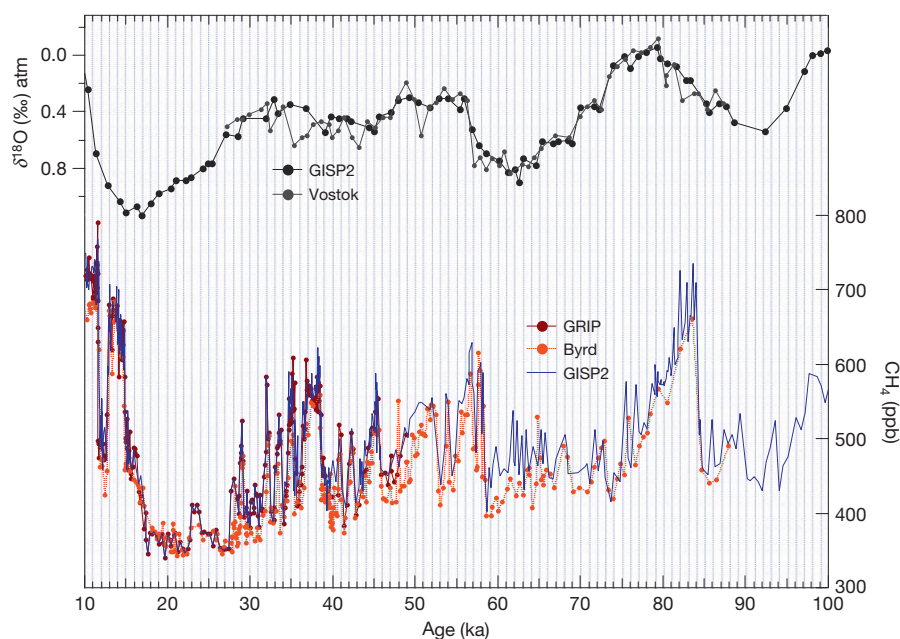


Figure 2 Records of atmospheric methane and $\delta^{18}\text{O}_{\text{atm}}$ from the GISP2 (Greenland), GRIP, Byrd (Antarctica), and Vostok (Antarctica) ice cores. The figure shows how variations in globally distributed gas signals can be matched between the hemispheres. Methane data from Brook et al. (1996), Brook et al. (2000), Blunier and Brook (2001), and Chappellaz et al. (1993); oxygen isotope data from Bender et al. (1999).

Each step in the process of translating a chronology from one core to another involves linked uncertainties, which must be incorporated in the final uncertainty estimate for the new chronology. The first is that of the absolute age scale for the ice record used as the primary chronology. This uncertainty is not relevant if only relative chronology between cores is of interest. Additional relevant uncertainties include uncertainty in delta age models and correlation uncertainties related to sample resolution and variability in the original signal – periods of higher variability create a greater number of targets for matching records. The importance of the various uncertainties depends on the nature of the ice core records and the gas parameter used for correlation. Delta age is relatively small and well constrained for warm, high accumulation rate sites

like central Greenland today (where delta age is ~ 200 years), but at cold low accumulation sites like Vostok or Dome C in east Antarctica, Delta age is slightly greater than 3000 years today and may have reached 6000–8000 years during the Last Glacial Maximum. Uncertainties in delta age depend on both the accuracy of densification models and the accuracy of parameters needed to use the models. Uncertainty due to correlation can be very low for high-resolution records with abrupt transitions like the methane variations related to D–O events (Figure 2), but records with very low-frequency variability and/or low resolution can result in larger correlation errors.

One other specific correlation tool uses the accumulation of cosmogenic ^{10}Be in polar ice. ^{10}Be (half-life = 1.36×10^6 years) is produced in the stratosphere by interactions of cosmic ray

neutrons with atmospheric molecules. ^{10}Be is particle reactive and quickly scavenged and removed to the earth surface. The flux of ^{10}Be is controlled by the primary cosmic ray flux (assumed constant on the time scales of interest), and modulation of the flux by changes in the earth's magnetic field strength and solar activity. It is expected that excursions in the modulation effects create globally synchronous changes in ^{10}Be flux to polar regions. One well-known excursion that has been used for correlation is the '40 ka' event, an interval of significantly enhanced ^{10}Be deposition found in Antarctic and Arctic ice cores, associated with the Laschamp geomagnetic excursion (Blunier et al., 1998; Raisbeck et al., 2002). Smaller scale variations in ^{10}Be associated with solar variability have also proved useful (Taylor et al., 2004).

Comparing Greenlandic and Antarctic Ice Core Records on Common Chronologies

In the first study of this type Bender et al. (1994) compared stable isotope records from GISP2 (Greenland) and Vostok (Antarctica) on a common chronology constructed using $\delta^{18}\text{O}_{\text{atm}}$ as a stratigraphic correlation tool. $\delta^{18}\text{O}_{\text{atm}}$ varies with time because the oxygen isotopic composition of the atmosphere is controlled by the ocean isotopic composition through the cycling of oxygen between the atmosphere and ocean by photosynthesis and respiration. $\delta^{18}\text{O}_{\text{atm}}$ is also secondarily influenced by processes in the terrestrial biosphere. The lifetime of oxygen in the atmosphere with respect to these processes is ~ 1000 years (Bender et al., 1994) – as a result $\delta^{18}\text{O}_{\text{atm}}$ variations are globally synchronous, but relatively slow (Figure 2). Bender et al. estimated that the precision of their comparison between GISP2 and Vostok was ~ 3000 years, due to analytical uncertainties, uncertainty in correlation, and uncertainty in delta age (primarily for Vostok). Their main observation was that long D–O events in Greenland had counterparts in Antarctica, but that uncertainties in their study were too large to discern if these events were temporally in or out of phase. A later study (Bender et al., 1999) revisited the GISP2–Vostok comparison with a higher precision data set. This study confirmed the link between long Greenland D–O and Antarctic warm events, and suggested that shorter Greenland D–O events also had Antarctic counterparts. As in the previous study, temporal phasing was difficult to establish, but they suggested that the timing of peak warming in both hemispheres was similar, a conclusion supported by later work (see below).

Sowers and Bender (1995) performed a similar study for the deglacial section of the Byrd (Antarctica) ice core and showed that the Antarctic cold reversal (ACR), a cool interval that interrupted deglacial warming about 15 000 years ago, occurred at the time of abrupt warming (onset of the Bølling period) in Greenland, and that the subsequent cooling in Greenland (the Younger Dryas) corresponded with warming in Antarctica. Ice accumulation rates at Byrd are $\sim 5\times$ than at Vostok, making delta age smaller, and these conclusions are more robust than those for the Greenland–Vostok comparison.

Correlation using $\delta^{18}\text{O}_{\text{atm}}$ has the disadvantage that the low-amplitude variability creates relatively large uncertainties for some time intervals. Methane variations are more rapid (Brook et al., 1996) and this recognition led to studies using

methane as a correlation tool. For example, Blunier et al. (1998) used methane for correlating stable isotope records from the GRIP (Greenland) and Byrd ice cores over the time period from 10 000 to 50 000 years. The high-frequency methane variations over this time period (Figure 2), and lower delta age at Byrd versus Vostok (due to higher temperatures and accumulation rates), allowed a more precise comparison of isotope records between the hemispheres. Blunier et al. found that two prominent interstadial events in Greenland (D–O events 8 and 12) were preceded by warming in Antarctica by 2–3 ka, and over the 23 000–47 000 year time period the average phase relationship was Antarctica leading Greenland by 1000–2500 years. As these phase differences are large with respect to the various relative chronological uncertainties, they are robust. Blunier et al. also compared the Vostok record with GRIP, and came to similar conclusions as those based on the Byrd record. As discussed above, delta age is very large at Vostok, making definitive conclusions about phasing more difficult. However, Blunier et al. were able to verify their Vostok chronology at $\sim 40 000$ years by showing that ^{10}Be records from GRIP and Vostok matched for this time.

Blunier and Brook (2001) followed this work with a more comprehensive comparison of GISP2 and GRIP with Byrd, also based on methane, for the period 10 000–90 000 years. This comparison (Figure 3) showed that all seven of the larger/longer interstadial events in Greenland over this time period were preceded by warming events in Antarctica (numbered A1–A7) (Figure 3).

The total uncertainty in the relative chronology between the Byrd and Greenland records is between 300 and 500 years. Close visual comparison of the two records provides some further support for the suggestion that the 'out of phase' relationship between the Greenlandic and Antarctic records extends to the smaller D–O events and corresponding warmings in Antarctica. Blunier and Brook noted this relationship but cautioned that chronological uncertainties made it difficult to interpret. Taking advantage of new ice core records, EPICA Community Members et al. (2006) undertook an analogous study using the NGRIP (Greenland) and EPICA Dronning Maud Land (EDML) (Antarctica) ice cores, covering the last 120 000 years. Those authors concluded that not only did the onset of the 'A' events precede warming in Greenland, but they suggested that every Greenland interstadial event was preceded by a warm event in Antarctica.

One of the open questions from the above work concerns the precise timing of the onset of warming in Greenland versus the maximum warmth in Antarctica. 'Seesaw' models based on changes in North Atlantic ocean circulation and deep water formation predict that the maximum Antarctic warmth (and subsequent start of cooling) will lag or be coincident with abrupt warming in the south. This relationship is difficult to test as it requires very precise age control. Morgan et al. (2002) attempted such a test using data from the Law Dome ice core in Antarctica, and the GRIP core in Greenland, for the last deglaciation (between 11 000 and 17 000 years ago). Law Dome, in coastal east Antarctica, is an excellent location for this work due to high snow accumulation rates and relatively warm temperatures. During the time period covered by this study, Antarctic warming started at about 17 000 years ago, and as described above was interrupted at about 14 000–15 000 years by the

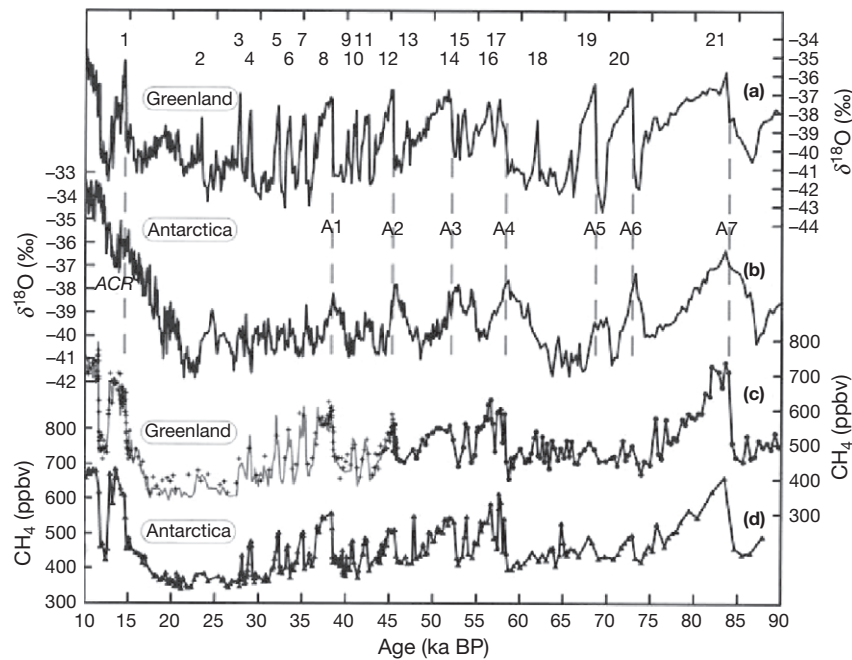


Figure 3 Comparison of stable isotope records from the GISP2 and Byrd ice cores on a common chronology created using methane concentration variations (also shown). The large oscillations in isotope ratios in Greenland represent temperature changes of $>10^{\circ}\text{C}$. The more modest Antarctic isotopic fluctuations probably represent a few degrees of temperature change. Reproduced from [Blunier T and Brook EJ \(2001\)](#) Timing of millennial-scale climate change in Antarctica and Greenland during the last glacial period. *Science* 291: 109–112.

ACR ([Figure 3](#)). In the standard seesaw model, the ACR corresponds to the warming during the Bølling and Ållerød periods in the Northern Hemisphere, and the cold Younger Dryas that followed occurred at a time of strong warming in Antarctica ([Figure 3](#)). Using methane as a stratigraphic marker, Morgan et al. inferred that the start of the ACR preceded the onset of the Bølling period, a relationship that is inconsistent with the standard seesaw concept. At the limits of their uncertainties, the Morgan et al. data allow the onset of the ACR onset to be synchronous with Bølling warming, but they viewed this scenario as unlikely. Later work creating a composite Antarctic temperature record ([Pedro et al., 2011](#)) suggests that regional climate variability contributed to the early cooling at Law Dome, and the composite record suggests that the ACR is synchronous with the Younger Dryas.

[Brook et al. \(2005\)](#) examined the timing of Antarctic events in the Siple Dome ice core (west Antarctica) and compared Siple Dome to both Byrd and GISP2 for the 8700–56 000-year period. Conditions at Siple Dome are similar to that at Byrd, resulting in a record with comparable time resolution and precision of comparison. The results are similar to those of [Blunier and Brook \(2001\)](#) for the larger D–O/Antarctic event pairs in this time interval. No strongly consistent relationships for the smaller D–O events and Antarctic climate is evident, although the seesaw pattern is apparent for some time intervals.

In a variant of the above studies, [Delmotte et al. \(2004\)](#) used a high-resolution 440 000-year methane record from the Vostok ice core as a proxy for Greenland climate variations, and compared the timing of methane variations with the timing of temperature change in Antarctica via the Vostok stable isotope record. This approach is justified based on several studies that show an extremely close relationship between methane

variations and Greenland temperature variations (e.g., [Severinghaus and Brook, 1999](#); [Severinghaus et al., 1998](#)). Using a lagged correlation analysis, Delmotte et al. showed that methane consistently lagged the δD temperature proxy in the Vostok record, by an average of 1100 ± 200 years, which suggests that the seesaw pattern in Antarctic and Greenlandic temperatures is persistent feature of late Quaternary climate change. [Loulergue et al. \(2008\)](#) reported a more detailed record and analysis of similar variations covering the last 800 000 years.

One study has suggested an exception to the seesaw pattern. [Steig et al. \(1998\)](#) used methane and $\delta^{18}\text{O}_{\text{atm}}$ to create a chronology for the Taylor Dome ice core in east Antarctica for the last deglaciation. In that chronology, a small cooling trend at Taylor Dome seems to correlate best with the Younger Dryas period in Greenland, and the main Antarctic deglacial warming seems to correlate best with the abrupt warming at the start of the Bølling period in Greenland. In all other Antarctic records, deglacial warming started at about 17 000 years (see [Figure 3](#)) well before the main Greenland warming. Taylor Dome is glaciologically unusual though, and [Mulvaney et al. \(2000\)](#) suggested that extremely low accumulation rates at Taylor Dome at the end of the Ice Age, unrecognized by Steig et al., might explain an apparent synchrony between Taylor Dome and Greenland records.

A weakness of the gas-based chronologies is reliance on models for delta age, although models are improving in sophistication ([Goujon et al., 2003](#)). In some cases, verification of the delta age model is possible through the use of nitrogen and argon isotope proxies for temperature change ([Caillon et al., 2003](#); [Huber et al., 2006](#); [Severinghaus and Brook, 1999](#); [Severinghaus et al., 1998](#)). Abrupt temperature changes cause a detectable fractionation of nitrogen and argon

isotopes due to thermal diffusion in the snowpack. In favorable circumstances, this signal is preserved in the gas phase in ice cores, and provides a record of the timing of abrupt temperature change that can be directly compared to changes in other gases (e.g., methane). One main conclusion from these studies has been that abrupt changes in temperature in Greenland coincide very closely with abrupt changes in atmospheric methane. Delta age models are generally able to reproduce the timing indicated by the thermal fractionation studies, reducing uncertainty in this aspect of the inter-polar comparison between ice cores. Unfortunately, thermal fractionation records from Antarctic ice cores are rare, though not unknown (Caillon et al., 2003) because temperature changes are often small and/or slow to be recorded by this proxy.

Statistics of Inter-polar Comparisons

Conceptual and numerical models attribute warming of high southern latitudes at times when Greenland cooled to decreases in the North Atlantic overturning circulation, which diminished heat transport to high northern latitudes while leaving more heat in the Southern Hemisphere (e.g., Crowley, 1992; Ganopolski and Rahmstorf, 2001; Knutti et al., 2004; Menviel et al., 2008; Schmittner et al., 2003; Stocker, 2002; Stocker et al., 1992; Vellinga and Wood, 2002). The Antarctic–Greenland comparison has been used to support this mechanism of abrupt climate change.

A variety of quantitative analyses have been applied to the data sets discussed here with varying conclusions. Steig and Alley (2002) used lag correlation analysis of filtered data to examine the relationships between Byrd and GISP2. They concluded that the data are consistent with either a Southern Hemisphere lead of 1000–1600 years or a 400–800-year Southern Hemisphere lag, and emphasized that neither is a complete description of the data. Schmittner et al. (2003), also using correlation analysis of filtered data, argued that temperature changes in Greenland led changes of the opposite sign in Antarctica. Their analysis also revealed high correlation for Antarctica leading Greenland by ~1300 years, but they emphasized that this observation need not require a trigger for abrupt climate change in the Southern Hemisphere. Huybers (2004), commenting on Schmittner et al., and using a different model analysis, argued that a small Antarctic lead was statistically superior to a Greenland lead. Wunsch (2003) analyzed the records in the frequency domain, and concluded that at periods longer than about 2500 years Antarctic temperature changes led those in Greenland by 1400–1800 years, but suggested that evidence for mechanistically significant relationships between the records on shorter time scales is weak or absent. Stocker and Johnsen (2003) developed a simple ‘thermal bipolar seesaw’ model which allowed them to predict the Byrd isotope curve fairly accurately from the Greenland (GRIP in this case) record by incorporating a heat reservoir (the Southern Ocean), which delays and integrates the impact of a Northern Hemisphere climate changes on the Southern Hemisphere. They argued that almost all D–O events between 25 and 65 ka have a recognizable counterpart in the Byrd record, an argument further developed by EPICA Community Members et al. (2006). Knutti et al. (2004) elaborated on this

approach with a more sophisticated model, and reached similar conclusions. Roe and Steig (2004) argued for a significant relationship between the Byrd and GISP2 records, but only for the largest and longest of the millennial-scale fluctuations, consistent with the visual interpretation of the data described above. In summary, there is consensus that the largest of the D–O events were preceded by Antarctic warming, and that the Antarctic warming terminated at about the time of abrupt warming in Greenland. Cooling then followed in both locations (Figure 3). The Byrd, Vostok, and, EDML records appear to show similar patterns for shorter events, but not all approaches to analyzing the statistical robustness of this observation come to the same conclusion. Long Antarctic records with higher accumulation rate than the Byrd ice core would be helpful. A suitable new core in West Antarctica (WAIS Divide ice core) was collected in the period from 2006 to 2011 and may help to address these questions once analyzed.

Implications and Directions for Future Work

As described previously the seesaw pattern observed for the larger D–O/Antarctic event pairs has been used as evidence for a major role for changes in thermohaline circulation in the origin of D–O events. In fact, the concept (if not the name) of the bipolar seesaw predates the detailed comparison of ice core records from Greenland and Antarctica, and arose originally from theoretical considerations of how reductions in the Atlantic deep overturning circulation should cause warming in southern latitudes (Crowley, 1992). Broecker (1998) coined the ‘seesaw’ term in explaining a mechanism for linking northern cooling and southern warming.

A number of these modeling studies reproduce the seesaw pattern by forcing changes in Atlantic Ocean meridional overturning circulation through a variety of means, and thereby reducing the northward transfer of heat in the North Atlantic region. Models applied to this problem range from simple conceptual models (Stocker and Johnsen, 2003) to coupled ocean atmosphere general circulation models (GCMs) (Vellinga and Wood, 2002). The result is often, though not universally, a warming in the Southern Hemisphere in the region of the Antarctic continent, although (consistent with the data) the response is not of equal magnitude to cooling in the north Atlantic.

To the extent that these models can reproduce the seesaw pattern they support the role of changes in North Atlantic overturning circulation in abrupt climate change. However, the exact mechanisms involved are not clear, and may not be represented properly in all models. For example, some studies have emphasized the possible role of key mechanisms in the southern ocean that might control the seesaw pattern, including sea ice formation and eddy mixing processes (Keeling and Stevens, 2001; Keeling and Visbeck, 2005).

Further understanding of the bipolar seesaw and related phenomenon requires work in several areas. New ice core records from sites with high accumulation rates are particularly important, as these reduce uncertainty in delta age, and may allow definitive constraints on the timing of the ‘tipping’ point of the seesaw mechanism. A wider spatial distribution of cores in Antarctica is also desirable since it is not clear where the

fulcrum of the seesaw is. Locating the fulcrum is important as it can help to distinguish between various mechanisms that cause Antarctic warming in models.

The chronologies used to interpret the bipolar comparison of ice core records can also be improved in a variety of ways. One is by the use of multiple constraints (various gases, ^{10}Be , etc.) in correlation schemes. Objective error analysis, for example using Monte Carlo simulations is also desirable. Lemieux-Dudon et al. (2010) present an example of this type of comprehensive approach. Delta age models are improving, and increasing our understanding of their reliability over the broad spectrum of accumulation/temperature conditions of glacial–interglacial climate cycles is important. Higher resolution, and higher precision, gas data will also be critical as uncertainty simply related to sample spacing and analytical error limits conclusions of some recent studies. New measurement techniques that can provide nearly continuous gas data will be particularly useful. Statistical analysis of these records can also be improved, and attention should be paid to developing objective analysis of the timing of Greenland warming/Antarctic cooling.

See also: Ice Core Methods: ^{10}Be and Cosmogenic Radionuclides in Ice Cores; Chronologies; Methane Studies. Ice Core Records: Antarctic Stable Isotopes; Greenland Stable Isotopes. Paleoclimate Reconstruction: Sub-Milankovitch (DO/Heinrich) Events.

References

- Alley RB, Meese D, Shuman CA, et al. (1993) Abrupt accumulation increase at the Younger Dryas termination in the GISP2 ice core. *Nature* 362: 527–529.
- Bender ML, Malaize B, Orchardo J, Sowers T, and Jouzel J (1999) High precision correlations of Greenland and Antarctic ice core records over the last 100 kyr. In: Clark PU, et al. (ed.) *Mechanisms of Global Climate Change at Millennial Time Scales*. *Geophysical Monograph*, vol. 112, pp. 149–164. Washington, DC: American Geophysical Union.
- Bender M, Sowers T, Dickson ML, et al. (1994) Climate correlations between Greenland and Antarctica during the past 100,000 years. *Science* 372: 663–666.
- Blunier T and Brook EJ (2001) Timing of millennial-scale climate change in Antarctica and Greenland during the last glacial period. *Science* 291: 109–112.
- Blunier T, Chappellaz JA, Schwander J, et al. (1998) Asynchrony of Antarctic and Greenland climate change during the last glacial period. *Nature* 393: 739–743.
- Broecker WS (1998) Paleocene circulation during the last deglaciation: A bipolar seesaw? *Paleoceanography* 13: 119–121.
- Brook EJ, Harder S, Severinghaus JP, Steig E, and Sucher C (2000) On the origin and timing of rapid changes in atmospheric methane during the last glacial period. *Global Biogeochemical Cycles* 14: 559–572.
- Brook EJ, Sowers T, and Orchardo J (1996) Rapid variations in atmospheric methane concentration during the past 110,000 years. *Science* 273: 1087–1091.
- Brook E, White JWC, Schilla A, et al. (2005) Timing of millennial-scale climate change at Siple Dome West Antarctica, during the last glacial period. *Quaternary Science Reviews* 24: 1333–1343.
- Caillon N, Jouzel J, Severinghaus JP, Chappellaz J, and Blunier T (2003) A novel method to study the phase relationship between Antarctic and Greenland climate. *Geophysical Research Letters* 30: 1899. <http://dx.doi.org/10.1029/2003GL017838>.
- Chappellaz JA, Blunier T, Raynaud D, Barnola JM, Schwander J, and Stauffer B (1993) Synchronous changes in atmospheric methane and Greenland climate between 40 and 8 kyr BP. *Nature* 366: 443–445.
- Crowley TJ (1992) North Atlantic deep water cools the southern hemisphere. *Paleoceanography* 7: 489–497.
- Cuffey KM, Clow GD, Alley RB, Stuiver M, Waddington ED, and Saltus RW (1995) Large Arctic temperature change at the Wisconsin-Holocene glacial transition. *Science* 270: 455–458.
- Delmotte M, Chappellaz J, Brook E, et al. (2004) Atmospheric methane during the last four glacial–interglacial cycles: Rapid changes and their link with Antarctic temperature. *Journal of Geophysical Research* 109: D12104. <http://dx.doi.org/10.1029/2003JD004417>.
- EPICA Community Members, Barbante C, Barnola JM, et al. (2006) One-to-one coupling of glacial climate variability in Greenland and Antarctica. *Nature* 444(7116): 195–198.
- Ganopolski A and Rahmstorf S (2001) Rapid changes of glacial climate simulated in a couple climate model. *Nature* 409: 153–158.
- Goujon C, Barnola J-M, and Ritz C (2003) Modeling the densification of polar firn including heat diffusion: Application to close-off characteristics and gas isotope fractionation for Antarctic and Greenland sites. *Journal of Geophysical Research* 108: 4792. <http://dx.doi.org/10.1029/2002JD003319>.
- Groote PM, Stuiver M, White JWC, Johnsen S, and Jouzel J (1993) Comparison of oxygen isotope records from the GISP2 and GRIP Greenland ice cores. *Nature* 366: 552–554.
- Huber C, Leuenberger M, Spahni R, et al. (2006) Isotope calibrated Greenland temperature record over Marine Isotope Stage 3 and its relation to CH_4 . *Earth and Planetary Science Letters* 243: 504–519.
- Huybers P (2004) Comments on 'Coupling of the hemispheres in observations and simulations of glacial climate change' by A. Schmittner, O.A. Saenko, and A.J. Weaver. *Quaternary Science Reviews* 23: 207–212.
- Keeling R and Stevens B (2001) Antarctic sea ice and the control of Pleistocene climate instability. *Paleoceanography* 16: 112–131.
- Keeling R and Visbeck M (2005) Northern ice discharges and Antarctic warming, could ocean eddies provide the link? *Quaternary Science Reviews* 24: 1809–1820.
- Knutti R, Flückiger J, and Timmermann A (2004) Strong hemispheric coupling of glacial climate through freshwater discharge and ocean circulation. *Nature* 430: 851–856.
- Lemieux-Dudon B, Blayo E, Petit JR, et al. (2010) Consistent dating for Antarctic and Greenland ice cores. *Quaternary Science Reviews* 29(1–2): 8–20.
- Loulergue L, Schilt A, Spahni R, et al. (2008) Orbital and millennial-scale features of atmospheric CH_4 over the past 800,000 years. *Nature* 453(7193): 383–386.
- Menviel L, Timmermann A, Mouchet A, and Timm O (2008) Meridional reorganizations of marine and terrestrial productivity during Heinrich events. *Paleoceanography* 23(1): PA1203.
- Morgan V, Delmotte M, van Ommen T, et al. (2002) Relative timing of deglacial climate events in Antarctica and Greenland. *Science* 297: 1862–1864.
- Mulvaney R, Rothlisberger R, Wolf EW, et al. (2000) The transition from the last glacial period in inland and near-coastal Antarctica. *Geophysical Research Letters* 27: 2673–2676.
- Pedro J, van Ommen T, Rasmussen S, et al. (2011) The last deglaciation: Timing the bipolar seesaw. *Climate of the Past* 7: 671–683.
- Raisbeck G, Yiou F, and Jouzel J (2002) ^{10}Be as a high-resolution correlation tool for climate records. *Geochimica et Cosmochimica Acta* 66: A623.
- Roe J and Steig E (2004) Characterization of millennial-scale climate variability. *Journal of Climate* 17: 1929–1944.
- Schmittner A, Saenko OA, and Weaver AJ (2003) Coupling of the hemispheres in observations and simulations of glacial climate change. *Quaternary Science Reviews* 22: 659–671.
- Severinghaus JP and Brook EJ (1999) Abrupt climate change at the end of the last glacial period inferred from trapped air in polar ice. *Science* 286: 930–934.
- Severinghaus JP, Sowers T, Brook EJ, Alley RB, and Bender ML (1998) Timing of abrupt climate change at the end of the Younger Dryas interval from thermally fractionated gases in polar ice. *Nature* 391: 141–148.
- Sowers TA and Bender M (1995) Climate records during the last deglaciation. *Science* 269: 210–214.
- Sowers T, Bender M, Labeyrie L, et al. (1993) 135,000 year Vostok-SPECMAP common temporal framework. *Paleoceanography* 6: 867–875.
- Steig EJ and Alley RB (2002) Phase relationships between Antarctic and Greenland climate records. *Annals of Glaciology* 35: 451–456.
- Steig EJ, Brook EJ, White JWC, et al. (1998) Synchronous climate changes in Antarctica and the North Atlantic during the last glacial–interglacial transition. *Science* 282: 92–95.
- Stocker TF (2002) North–south connections. *Science* 297: 1814–1815.
- Stocker TF and Johnsen SJ (2003) A minimum thermodynamic model for the bipolar seesaw. *Paleoceanography* 18: 1087. <http://dx.doi.org/10.1029/2003PA000920>.
- Stocker TF, Wright DG, and Broecker WS (1992) The influence of high-latitude surface forcing on the global thermohaline circulation. *Paleoceanography* 7: 529–541.
- Taylor KC, Alley RB, Meese DA, et al. (2004) Dating the Siple Dome (Antarctica) ice core by manual and computer interpretation of annual layering. *Journal of Glaciology* 50(170): 453–461.
- Vellinga M and Wood RA (2002) Global climate impacts of a collapse of the Atlantic thermohaline circulation. *Climatic Change* 54: 251–267.
- Wunsch C (2003) Greenland–Antarctic phase relations and millennial time-scale climate fluctuations in the Greenland ice-cores. *Quaternary Science Reviews* 22: 1631–1646.

Ice Margin Sites

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Introduction

The large polar ice sheets are rich archives of information on the climate and environment during the past ~800 000 years as demonstrated by the results of the deep ice core drilling programs in central Greenland (e.g., Andersen et al., 2004) and Antarctica (e.g., Augustin et al., 2004). The ancient ice found at depth in the central regions of the ice sheets can also be retrieved at the surface at certain locations along the ice sheet margin (e.g., Lorius, 1968; Reeh et al., 1991). This is illustrated by Figure 1, which shows particle paths in a cross-section of an ice sheet. The higher the ice was originally deposited in the accumulation area, the deeper into the ice sheet it will sink, and the nearer to the ice edge it will resurface in the ablation area. If basal temperatures never reached the melting point, and if basal ice was never subject to folding or faulting, a complete sequence of all deposited layers would occur at the surface in the ablation area with the oldest ice closest to the ice edge. In this way, collecting a 'horizontal ice core' at the ice-margin surface becomes possible.

Several different types of ice-margin sites from which records have been collected can be identified. The Greenland horizontal ice-coring sites (Figure 2) are land-based ice-margin sectors within the ice ablation zone where ice flow stagnates (typically due to a topographical obstacle, such as a ridge). Reeh et al. (2002) present a comprehensive overview of Greenland ice-margin sampling sites. In the summer, the surface of such ice margins consists of glacier ice that ablates by melting. The annual ablation rate at such sites ranges from near zero at the high altitude sites in North-East Greenland (e.g., Grandjean Fjord and Dronning Louise Land) to 1–3 m of ice at the sites at lower latitude and elevation (e.g., Pakitsoq). Surface irregularities with wavelength and wave height of typically 2–5 and 0.5 m, respectively, are a common occurrence. Supraglacial streams and lakes are also common. A photo of a typical Greenland ice-margin site is shown in Figure 3.

Another type of ice-margin site, known as a blue ice area (BIA), occurs in Antarctica (Figure 4). As the name suggests, BIAs are areas where bare glacier ice is exposed at the surface, often making the surface appear blue. The ice dynamics and climatology of Antarctic BIAs has been described in detail by Bintanja (1999). In contrast to the Greenland sites, ice ablates by sublimation rather than by melting. Ablation rates depend on local climatology and range between near zero and 35 cm water equivalent per year. BIAs most often exist in the lee of mountain ranges and nunataks, which act as barriers for snowdrift and cause stagnating ice flow as well. BIAs have substantially lower albedo, are warmer, and typically have gustier winds than the adjacent snow-covered areas (Bintanja, 1999). As a result of sublimation, the surface of BIAs is generally smooth but often rippled. Figure 5 shows a typical Antarctic BIA.

One final type of ice-margin site discussed here is an Antarctic valley glacier in which bare ice is exposed at the surface. They are Type II BIAs according to the classification of Bintanja (1999). On these glaciers, strong katabatic winds remove snow from the surface and induce ice sublimation. In contrast to the BIAs described in the above paragraph, the ice flow in these valley glaciers is not stagnated by topographic obstacles. Several such glaciers exist in the Dry Valleys region of Antarctica (Figure 6).

The Norwegian–British–Swedish Antarctic expedition of 1949–52 was the first to investigate ice-margin areas. These glaciological and geomorphological investigations took place in Dronning Maud Land (Bintanja, 1999 and references therein). In the mid-1960s, the first two shallow ice cores from an ice-margin area were drilled by a French expedition near the Dumont D'Urville station in the Terre Adelie coastal region (Lorius, 1968). Analyses of these ice cores were interpreted to suggest that some of the near-surface ice at this location was formed far from the coast. In 1969, the first meteorites were discovered on the surface of the Yamato Mountains BIA, dramatically increasing scientific interest in ice-margin sites (Yoshida et al., 1971). By the 1970s, the first water isotope, and even gas, records were being produced from Arctic ice-margin sites as well (Hooke, 1976).

Scientific interest in records from ice-sheet margins continues to increase, for several reasons. First, and most importantly, ice margins expose ancient ice at, or near, the surface. Deep ice-core drilling has the advantage of producing a stratigraphically ordered paleoenvironmental record that can be dated in a reliable way. However, deep ice-core drilling is very expensive and logistically challenging. Ice margins offer an opportunity to collect ancient ice much more easily and with orders of magnitude less expense. In addition, ice-margin outcrops can offer a virtually unlimited amount of ancient ice of a given age, making possible analyses of ultra-trace components that are not feasible in a conventional ice core. Second, Antarctic BIAs have proven to be the best meteorite collection sites in the world. Third, ice-margin records can offer insight into ice-sheet dynamics. Finally, ice-margin sites may contain the oldest ice on Earth.

Meteorites

More than 25 000 meteorites have been found on Antarctic BIAs (Harvey, 2003). Meteorites that fall in the snow accumulation zone of the BIAs are incorporated in the ice and carried with the ice flow to the BIAs. Since at many BIAs the ice flow stagnates and the ice is removed only by sublimation of the surface, the embedded meteorites are exposed and accumulate on the surface (Whillans and Cassidy, 1983). Meteorites from Antarctic BIAs have provided us with a wealth of scientific

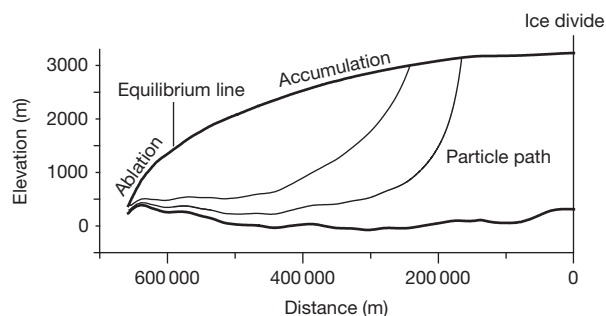


Figure 1 Cross-section of ice sheet, showing particle paths connecting snow deposition sites in the accumulation zone with locations where the ice resurfaces in the ablation zone. Reproduced from Reeh N, Oerter H, and Thomsen HH (2002) Comparison between Greenland ice-margin and ice-core oxygen-18 records. *Annals of Glaciology* 35: 136–144, with permission.



Figure 3 An aerial view of the Pakitssoq ice-margin site in West Greenland. The orange laboratory tent (~6 m long) provides a length scale. Photo courtesy of Jeffrey Severinghaus.



Figure 2 Location of Greenland ice-margin sampling sites (dots) and deep ice-core drilling sites (crosses). Reproduced from Reeh N, Oerter H, and Thomsen HH (2002) Comparison between Greenland ice-margin and ice-core oxygen-18 records. *Annals of Glaciology* 35: 136–144, with permission.

information on extraterrestrial material, such as its geochemical composition and the size distribution of material reaching the Earth's surface, as well as estimates of the influx rate of extraterrestrial material (e.g., Harvey, 2003; Whillans and Cassidy, 1983; Yoshida et al., 1971). Meteorites may also be able to provide information about the age of ice exposed at the BIAs (e.g., Curzio et al., 2008; Nishiizumi et al., 1983 and

references therein). For example, cosmogenic isotope dating of a layer of debris thought to be of meteoritic origin within the ice at the Allan Hills BIA suggested an age as old as 2.8 My (Harvey et al., 1998). While a more detailed discussion of meteorites at Antarctic BIAs is outside the scope of this article, the interested reader could find a review of the topic in Harvey (2003).

Volcanic Tephra Layers

Many Antarctic BIAs contain well-defined outcropping tephra layers (e.g., Curzio et al., 2008; Nishio et al., 1985). There are a number of active volcanoes located in West Antarctica as well as in Victoria Land. Ash from eruptions can be deposited anywhere in Antarctica, become embedded in the ice and advected with the ice flow. Tephra layers have been found in all Antarctic deep ice cores (e.g., Dunbar et al., 2003; Fujii et al., 1999; Palais, 1985). Ice advection to the BIAs can result in exposure of tephra layers at the surface at these sites. Tephra layers have been documented at Mount Moulton (Dunbar et al., 2008; Wilch et al., 1999), Frontier Mountain and Lichen Hills (Curzio et al., 2008; Perchiazzi et al., 1999), Yamato Mountains (Katsushima et al., 1984; Koeberl, 1990; Nishio et al., 1985), Allan Hills (Harvey et al., 1998; Marvin, 1986), Lewis Cliff (Koeberl et al., 1988), and Skelton Neve and Kempe Glacier (Keys et al., 1977).

Studies of tephra layers at ice-margin sites can yield a variety of valuable information. Grain size can be used to infer distance from the source eruption (e.g., Nishio et al., 1985). Geochemical composition of the ash can be used to identify the volcanic area of provenance or sometimes even the individual volcano (e.g., Perchiazzi et al., 1999; Wilch et al., 1999). The age of the ash layer can be determined using the $^{40}\text{Ar}/^{39}\text{Ar}$ technique (e.g., Curzio et al., 2008; Dunbar et al., 2008; Wilch et al., 1999). In this way, ice-margin tephra layers can potentially be used to reconstruct the volcanic history of a region (Curzio et al., 2008).

With the application of the $^{40}\text{Ar}/^{39}\text{Ar}$ technique, tephra layers can also provide absolute age stratigraphic markers for the outcropping ice itself. While ice-margin sites present many advantages over ice cores as described above (e.g., cost-effective, unlimited ice amount), one of the main

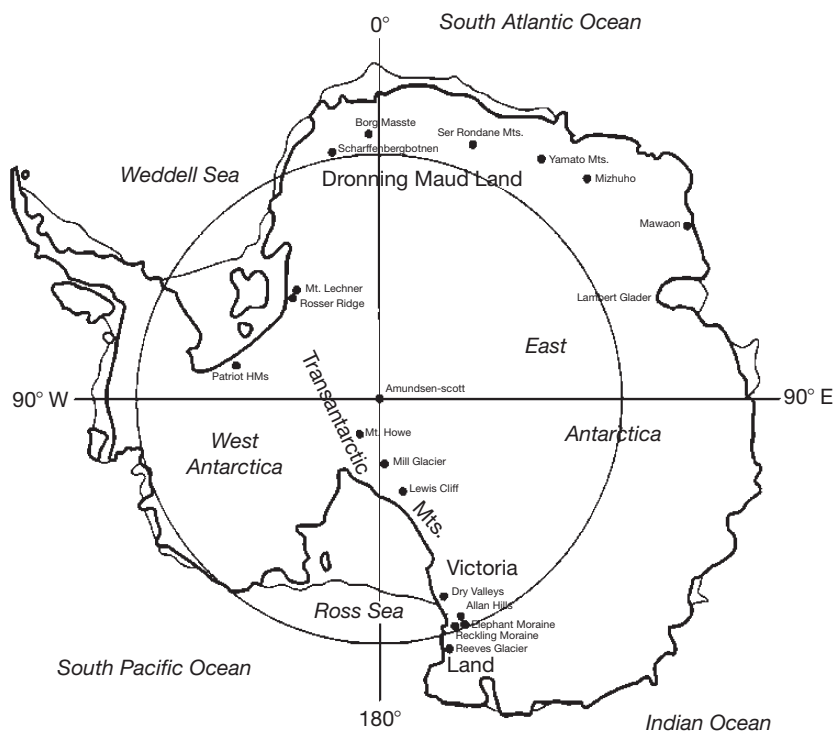


Figure 4 Map showing locations of Antarctic blue ice areas (BIAs). Reproduced from Bintanja R (1999) On the glaciological, meteorological, and climatological significance of Antarctic blue ice areas. *Reviews of Geophysics* 37: 337–359, with permission.

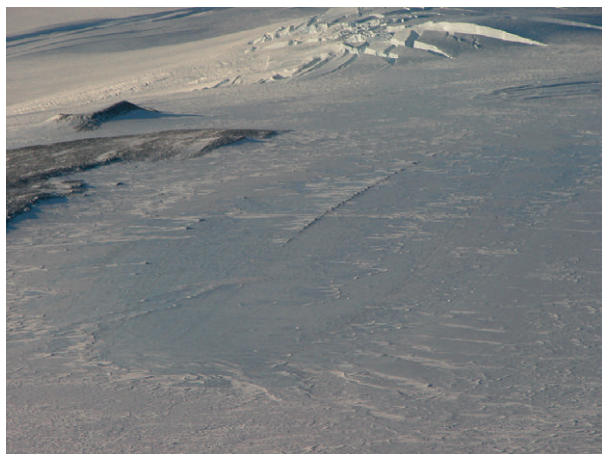


Figure 5 An aerial view of the Mount Moulton BIA in West Antarctica. Photo courtesy of Nelia Dunbar.



Figure 6 Shallow ice-core drilling on Taylor Glacier, Dry Valleys, Antarctica. Photo courtesy of Hinrich Schaefer.

disadvantages is that dating the ice in these outcrops is much more challenging. There is no well-defined ‘top’ or ‘bottom,’ there are no layers to count, and using a flow model to trace ice parcels back to the accumulation zone involves more uncertainties than modeling ice thinning in the deeper part of a conventional ice core. Multiple tephra layers outcropping at an ice-margin site allow this problem to be resolved.

Dunbar et al. (2008) and Wilch et al. (1999) studied the outcrop pattern of tephra layers at Mount Moulton, West Antarctica (Figure 7). $^{40}\text{Ar}/^{39}\text{Ar}$ dates on several tephra layers showed that ice between ~ 10 and 496 ka old was outcropping in stratigraphic order at this site. The $^{40}\text{Ar}/^{39}\text{Ar}$ dates showed

that the ice stratigraphy becomes progressively more compressed with age, similar to what is observed in conventional ice cores. The oldest tephra layer was dated at 495.6 ± 9.7 ka, making this the oldest known ice in West Antarctica, and thus, a very valuable archive for paleoenvironmental studies.

Curzio et al. (2008) analyzed $^{40}\text{Ar}/^{39}\text{Ar}$ in samples from two tephra layers at Frontier Mountain in northern Victoria Land, finding that the oldest ice at this BIA appears to be ~ 100 ka old, while $>90\%$ of the outcropping ice is younger than ~ 49 ka.

In addition to providing the absolute ages of the ice layers, tephra layers allow for easy visual identification and tracing

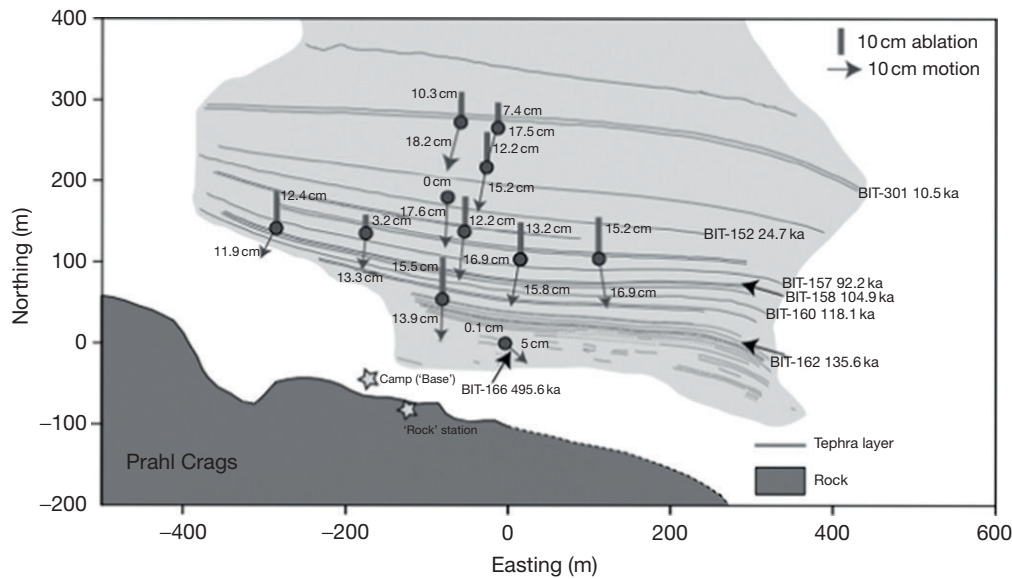


Figure 7 A map of the Mount Moulton BIA showing ablation rates, ice motion, and the outcropping tephra layers (gray lines) with $^{40}\text{Ar}/^{39}\text{Ar}$ dates for some of the layers. Reproduced from Dunbar NW, McIntosh WC, and Esser RP (2008) Physical setting and tephrochronology of the summit caldera ice record at Mount Moulton, West Antarctica. *Geological Society of America Bulletin* 120: 796–812, with permission.

of stratigraphic layers, even in areas where these layers have undergone deformation and folding (Figure 8).

Water Stable Isotope Records

$\delta^{18}\text{O}$ and δD of ice respond mostly to variations in local temperature, moisture sources, and seasonality of precipitation, and have always been some of the main and most informative measurements made in ice cores (e.g., Jouzel et al., 1997; Petit et al., 1999). A number of studies have measured $\delta^{18}\text{O}$ and δD at ice-margin sites, either in horizontal surface profiles or in shallow cores, providing information about ice provenance, age, and flow.

Antarctic Sites

The first water stable isotopic measurements at an ice margin were described by Lorius (1968) and came from two shallow (~100 m) cores collected in 1964–66 near the Dumont D’Urville station in Terre Adelie, East Antarctica (66°40’S, 140°01’E). It was found that δD showed near-surface values of ~–130‰ for the top ~1/3 of the core, then a transition to much more negative values (~–370‰) around mid-core, with low values continuing to the bottom. The isotopically light values in the lower part of these two cores were initially interpreted to be a result of ice that formed far inland and at much higher elevations. Subsequent work and further shallow core drilling in this area (Tison et al., 1993 and references therein), however, indicated that the very negative δD values were also the result of this ice having been deposited during the last glacial period, and much of the observed shift in δD with depth was a record of the glacial–interglacial climatic transition.

Faure and Buchanan (1987) collected ice surface samples each 100 m in a profile in the ice flow direction at Allan Hills (76°42.3’S; 159°24’E). The $\delta^{18}\text{O}$ values indicate that ice from both glacial and interglacial periods is represented along the profile (Figure 9). Groote (1990) reports results of $\delta^{18}\text{O}$ analyses of surface ice samples from three Antarctic BIAs. At Reckling Moraine (76°15’S; 158°40’E), near-surface ice samples were collected every 100 m in a transect across the moraine and the surrounding blue ice. Fluctuations in $\delta^{18}\text{O}$ of about 5‰ over short horizontal distances suggested that ice exposed at the surface originated in different glacial and interglacial periods. At Lewis Cliff Ice Tongue (84°17’S; 161°05’E), a 200-m transect was sampled every 10 m. The most negative $\delta^{18}\text{O}$ value of –58.8‰ would, under the present conditions, require the origin of the ice to be in the highest central part of East Antarctica, whereas a glacial period origin of the ice would allow a source area that was closer. From Allan Hills (76°43’S; 159°40’E) only four 1-m core samples were analyzed, that is, too few to allow conclusions about the origin of the ice in this area.

Taylor Glacier in the Dry Valleys area has an 80-km long ablation zone, exposing potentially the longest and highest-resolution ice-margin record. Aciego et al. (2007) collected surface ice samples along a 28-km long profile following approximately the center flow line on the glacier. Samples were collected at a resolution of 25–50 samples per kilometer and subsequently analyzed for $\delta^{18}\text{O}$ and δD . Figure 10 shows the Aciego et al. (2007) δD results from the Taylor Glacier profile together with the $\delta^{18}\text{O}$ data from the Taylor Dome ice core (Steig et al., 2000). Taylor Glacier ice has Taylor Dome as its source, and the two are thus expected to contain the same climatic record. Aciego et al. (2007) used a comparison of variations in their water isotopic record with the Taylor Dome record as well as a few measurements of $\delta^{18}\text{O}$ of atmospheric O_2 in the Taylor Glacier profile to construct an approximate age scale for the

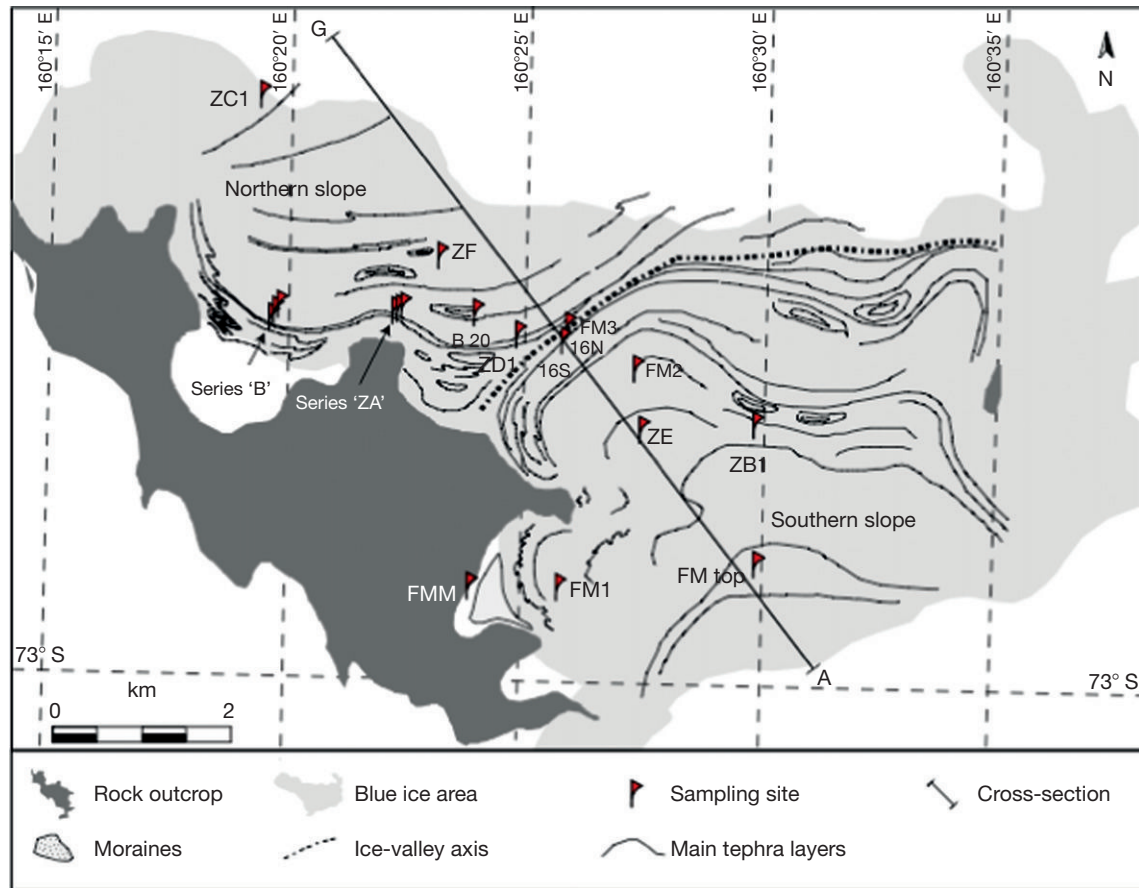


Figure 8 A map showing the main outcropping tephra layers at the Frontier Mountain BIA, Antarctica. Reproduced from Curzio P, Folco L, Laurenzi MA, Mellini M, and Zeoli A (2008) A tephra chronostratigraphic framework for the Frontier Mountain blue-ice field (northern Victoria Land, Antarctica). *Quaternary Science Reviews* 27: 602–620, with permission.

sampled section, as shown in Figure 10. Based on this age scale, Taylor Glacier exposes ice with ages ranging from Holocene to >70 ka.

Greenland Sites

The vast majority of ice-margin water stable isotope records have been retrieved from Greenland sites. The first sampling campaign on the west Greenland ice margin (69.7° N, 50.1° W) took place in 1977–78, and two shallow cores (<100 m) and a 400-m surface profile oriented perpendicular to the ice margin were collected. The $\delta^{18}\text{O}$ results of the study (Clausen and Stauffer, 1988) showed that ice with low $\delta^{18}\text{O}$ characteristic of the cold glacial periods was outcropping nearest to the ice margin, while further away from the margin ice had higher $\delta^{18}\text{O}$ values characteristic of the warm interglacial periods. Both shallow cores contained the low- $\delta^{18}\text{O}$ ice near the bottom and the high- $\delta^{18}\text{O}$ ice in the upper sections. The $\delta^{18}\text{O}$ transition in all three records was similar in magnitude to that observed in Greenland deep ice cores. This transition occurred deeper in the core that was situated further away from the ice margin. These results were as expected for an ice-margin site with relatively undisturbed flow, where the oldest ice flows closest to bedrock and emerges nearest to the margin.

Between 1985 and 1994, the $\delta^{18}\text{O}$ distribution of the surface ice was studied at 15 different locations on the Greenland ice-sheet margin (Reeh et al., 2002; Figure 11). Ice samples were collected in profiles perpendicular to the ice margin. At some locations, two or more records were obtained along closely spaced parallel sampling profiles in order to study the reproducibility of the records. The sampling frequency ranged from spot sampling at distances between 1 and 100 m to continuous sampling (20 cm sample length) over a distance of more than 450 m. All together, 28 $\delta^{18}\text{O}$ -records were established, giving an overview of the origin of the ice presently exposed at the Greenland ice margin.

Pleistocene ice is missing at 7 of the 15 $\delta^{18}\text{O}$ -sampling sites: Hans Tausen ice cap and Flade Isblink in eastern North Greenland, Tuto Ramp and Nuna Ramp in Northwest Greenland, Dronning Louise Land in central East Greenland, and Sermilik and Nansen Halvø in Southeast Greenland (for locations, see map in Figure 2). The absence of Pleistocene ice at these sites can be explained in two ways: (1) the Pleistocene ice was removed by basal melting; (2) the ice margin retreated considerably during the Holocene Climatic Optimum, and the later readvance or build-up to the present configuration consisted of ice of local Holocene origin. Reeh et al. (2002) argue that, very likely, the latter explanation applies to Tuto Ramp, Hans Tausen ice cap, and Flade Isblink ice cap, whereas low

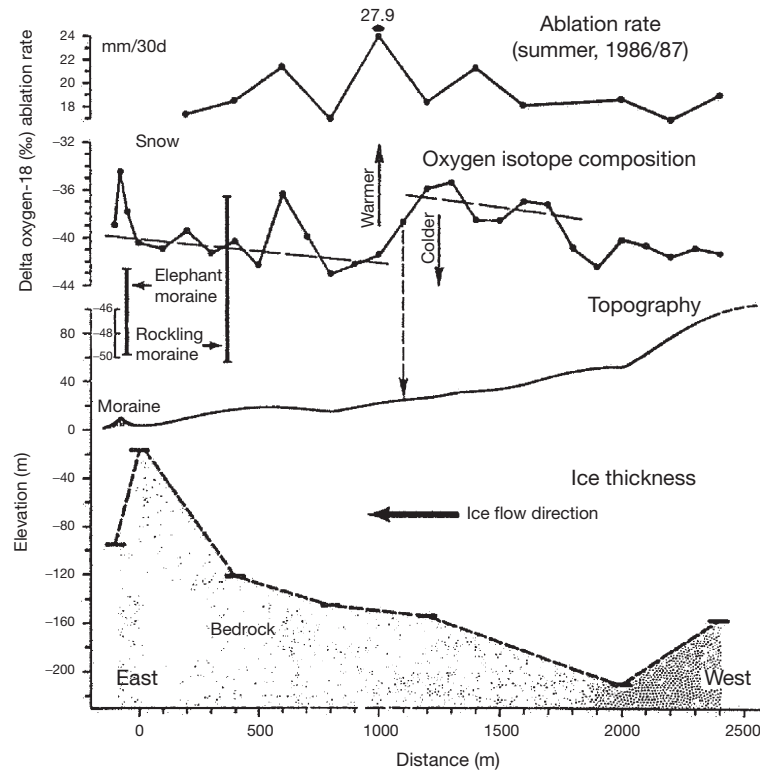


Figure 9 Summary of results along a traverse parallel to the local flow direction of the ice sheet near Allan Hills, Antarctica. Reproduced from Faure G and Buchanan D (1987) Glaciology of East Antarctic ice sheet at the Allan Hills: A preliminary interpretation. *Antarctic Journal of the United States* XXII: 74–75, with permission.

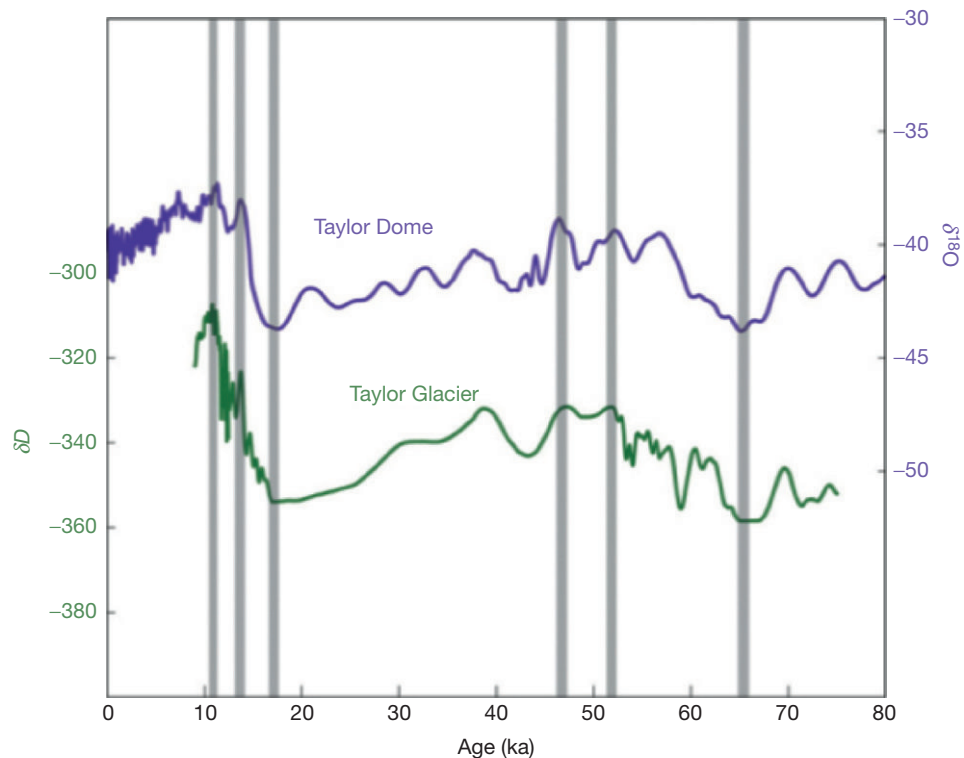


Figure 10 A comparison of Taylor Glacier and Taylor Dome water stable isotope records, with the Taylor Glacier record on an age scale that was inferred as described in the text. Reproduced from Aciego SM, Cuffey KM, Kavanaugh JL, Morse DL, and Severinghaus JP (2007) Pleistocene ice and paleo-strain rates at Taylor Glacier, Antarctica. *Quaternary Research* 68: 303–313, with permission.

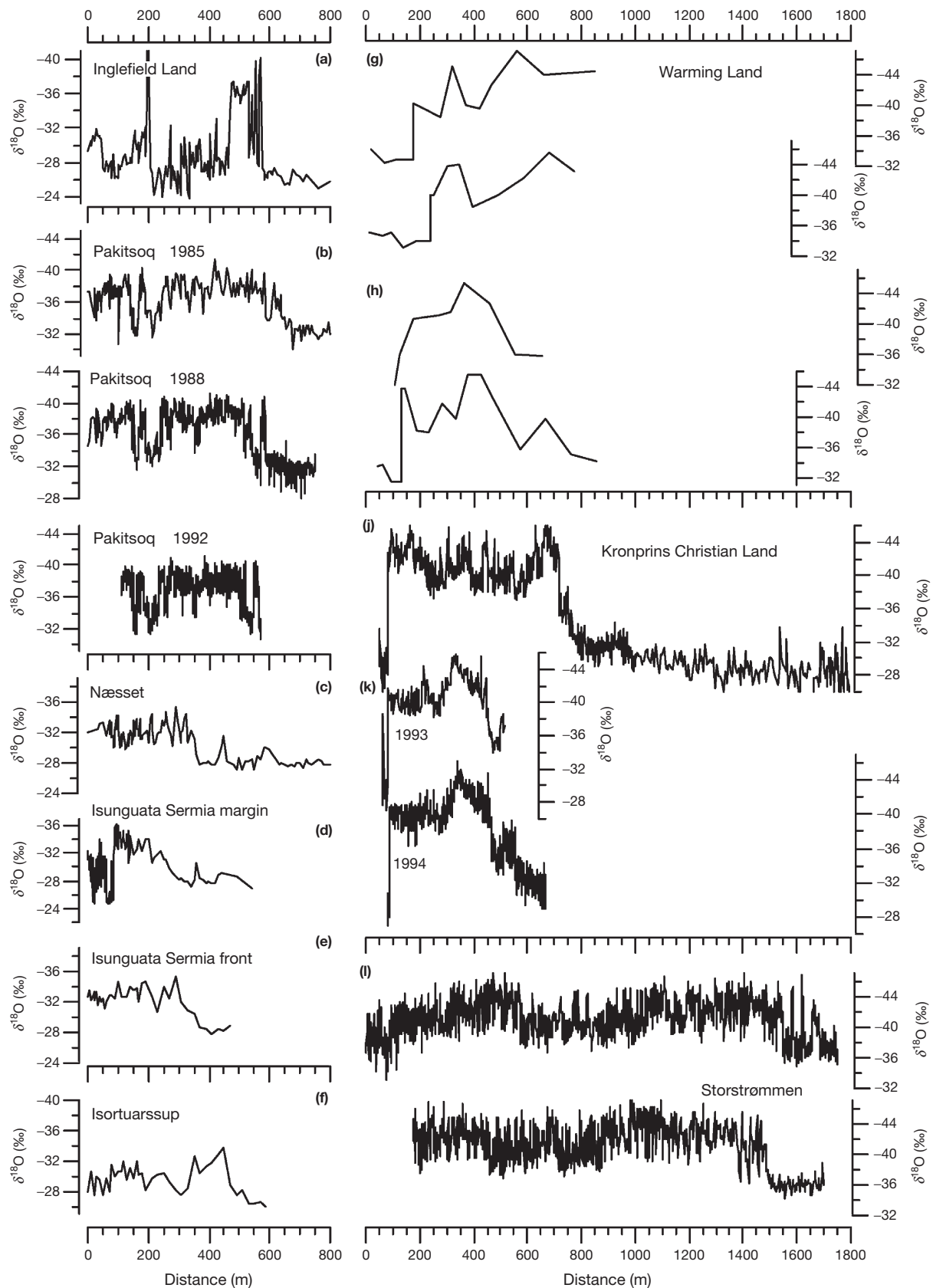


Figure 11 $\delta^{18}\text{O}$ -records from the Greenland ice-sheet margin dating back into the Pleistocene. (a) Inglefield Land; (b) Pakitsoq; (c) Næsset; (d) Isunguata Sermia margin; (e) Isunguata Sermia front; (f) Isortuarssup; (g and h) Warming Land; (j and k) Kronprins Christian Land; and (l) Storstrømmen. For locations, see map in [Figure 2](#). Reproduced from [Reeh N, Oerter H, and Thomsen HH \(2002\) Comparison between Greenland ice-margin and ice-core oxygen-18 records. *Annals of Glaciology* 35: 136–144, with permission.](#)

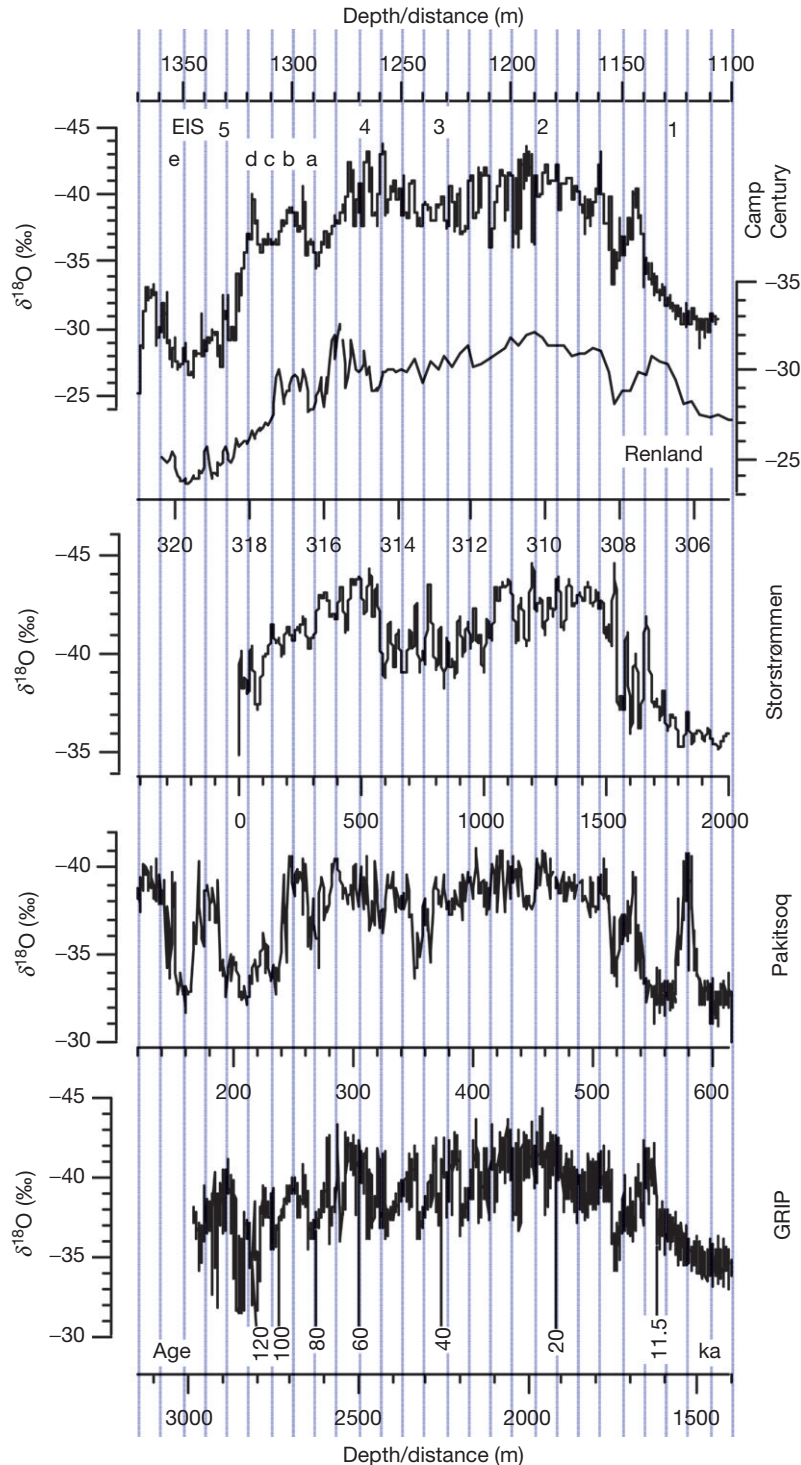


Figure 12 Comparison of ice-margin $\delta^{18}\text{O}$ records from Storstrømmen and Pakitsq with ice-core $\delta^{18}\text{O}$ records from Camp Century, Renland and GRIP Summit. Emiliani isotopic stages (EIS) are shown at the top of the figure. For locations, see [Figure 2](#). Reproduced from [Reeh N, Oerter H, and Thomsen HH \(2002\)](#) Comparison between Greenland ice-margin and ice-core oxygen-18 records. *Annals of Glaciology* 35: 136–144, with permission.

accumulation rates in North Greenland favoring basal melting may explain the missing Pleistocene ice in some records from North Greenland ice sheet margins. At the remaining eight Greenland ice-margin sites, a 350–1700 m-wide band of Pre-Holocene ice occurs next to the ice edge.

Several attempts have been made to correlate the Greenland ice-margin $\delta^{18}\text{O}$ records with deep ice-core records based on the comparison of characteristic $\delta^{18}\text{O}$ events in the ice margin and ice-core records (e.g., [Reeh et al., 1991, 1993](#)). A postulated chronology along the ice-margin records was then

established by using the ice-core timescale. An example is shown in [Figure 12](#) in which $\delta^{18}\text{O}$ records from the Pakitsoq and Storstrømmen ice margins are compared to the $\delta^{18}\text{O}$ records from Camp Century on the Northwest Greenland ice sheet, GRIP in the Summit area, and the local Renland ice cap in central East Greenland. A more reliable age scale has since then been established for the Pakitsoq site and it will be discussed in more detail in the section on gas measurements below.

Barnes Ice Cap

Several studies that included water stable isotopic measurements were conducted on the margin of the Barnes Ice Cap, Baffin Island, Canada. [Hooke \(1976\)](#) reported measurements on several samples collected at the surface or from shallow cores. The Barnes Ice Cap margin exposes a distinctive band of white, bubble-rich ice, which is both underlain and overlain by darker blue-gray ice layers. The white ice band yielded very low $\delta^{18}\text{O}$ values that were similar to glacial-period values from the Camp Century ice core, while the darker ice layers gave values more characteristic of the Holocene. [Hooke and Clausen \(1982\)](#) provided further $\delta^{18}\text{O}$ measurements confirming earlier results and suggested that the unusually large $\delta^{18}\text{O}$ shift observed between the postulated glacial and Holocene ice bands was due in part to a lower accumulation altitude for the Holocene-age ice. [Zdanowicz et al. \(2002\)](#) measured $\delta^{18}\text{O}$ in a 2000 m-long surface transect from the margin, capturing a well-resolved glacial-interglacial transition.

Gas Records

Early Work

Possibly the first systematic gas composition analyses on any type of glacier ice were carried out on margin ice from the Storbreen glacier in Norway ([Coachman et al., 1956](#)), although some even earlier attempts are mentioned by these authors. Ice was analyzed by dry extraction (ice shaving under vacuum) for nitrogen (N_2), oxygen (O_2) and carbon dioxide (CO_2) content. Subsequent work by some of the same authors ([Coachman et al., 1958a,b](#); [Scholander et al., 1962](#)) included analyses of argon (Ar) and even carbon-14 of CO_2 ($^{14}\text{CO}_2$) in very large samples from the Storbreen glacier ablation zone as well as in Greenland icebergs.

[Fireman and Norris \(1982\)](#) analyzed CO_2 , $^{14}\text{CO}_2$, O_2 , Ar, N_2 , $^{15}\text{N}/^{14}\text{N}$ of N_2 , and $^{40}\text{Ar}/^{36}\text{Ar}$ in samples from the Allan Hills. While the major gas (O_2 , Ar, and N_2) analyses yielded values very close to ambient atmospheric levels, there was considerable scatter in the CO_2 and $^{14}\text{CO}_2$ results.

In Situ Cosmogenic ^{14}C

^{14}C is produced from ^{16}O in near-surface ice by cosmic rays. Early efforts to measure ^{14}C (in CO_2) in ice from margin sites (e.g., [Scholander et al., 1962](#); [van de Wal et al., 1990](#)) did not take this into account and focused instead on trying to use $^{14}\text{CO}_2$ to date the ice. The ages obtained by this method were too young because of the cosmogenic ^{14}C contribution.

[Lal et al. \(1990\)](#) and [Lal et al. \(1987\)](#) were the first to systematically investigate *in situ* cosmogenic production of

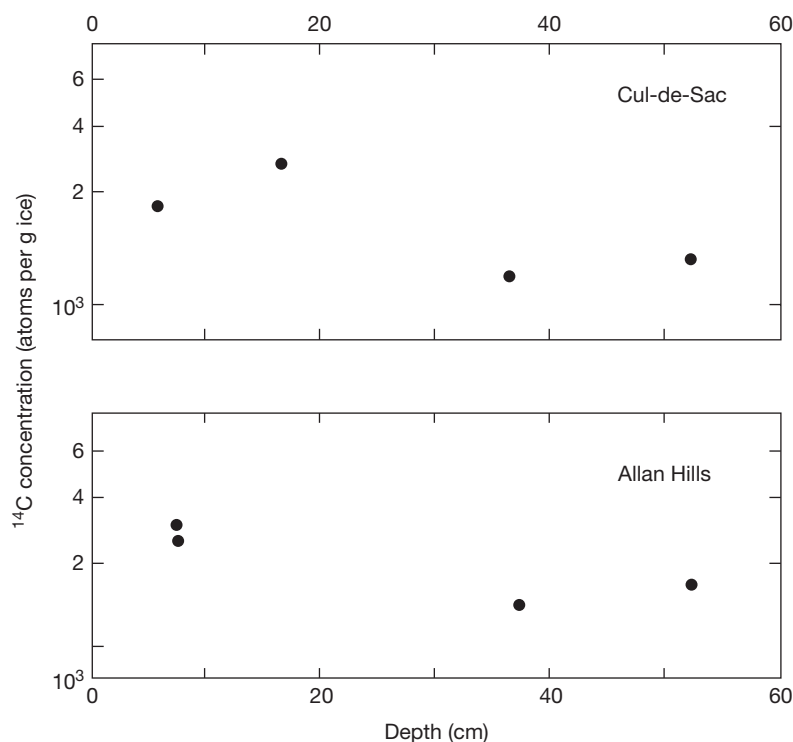


Figure 13 ^{14}C content of ice samples from two locations within the Allan Hills BIA. Reproduced from [Lal D, Jull AJT, Donahue DJ, Burtner D, and Nishiizumi K \(1990\)](#) Polar ice ablation rates measured using *in situ* cosmogenic C-14. *Nature* 346: 350–352, with permission.

^{14}C in ice by fast neutrons. Ice samples from the Allan Hills BIA with depths between 0 and 60 cm below surface were analyzed for both $^{14}\text{CO}_2$ and ^{14}C of CO (^{14}CO), Figure 13. They found that the ^{14}C content of the samples was as expected based on theoretical *in situ* production estimates and decreased with depth as expected due to shielding from overlying ice. The cumulative amount of *in situ* ^{14}C at the surface depends on the ice ablation rate. The authors were therefore able to use their ^{14}C measurements to estimate the ablation rates at the two sites and found their estimates to be consistent with conventional stake measurements.

The Scharffenbergbotnen BIA ($74^\circ 34' 40''\text{S}$, $11^\circ 02' 58''\text{W}$) in Queen Maud Land, East Antarctica, has been the location of several *in situ* cosmogenic ^{14}C studies. This work also explored the possibility of using *in situ* ^{14}C to estimate ablation rates,

suggesting the use of ^{14}CO to correct for *in situ* $^{14}\text{CO}_2$, in the hope of determining the true paleoatmospheric $^{14}\text{CO}_2$ signal to date the ice (van Roijen et al., 1994, 1995). Later work at this site also investigated *in situ* ^{14}C production by negative muon capture (van der Borg et al., 2001; van der Kemp et al., 2002). Like fast neutrons, muons are also secondary cosmic particles that are produced in the Earth's atmosphere and have the potential to interact with ^{16}O nuclei. However, muons are able to penetrate considerably deeper into the ice than neutrons. The authors found evidence that ^{14}C in ice is indeed produced by both neutrons and muons.

While the pioneering *in situ* ^{14}C work described above demonstrated cosmogenic ^{14}C production in ice by both neutrons and muons, much remains to be understood. For example, there is considerable disagreement with regard to the

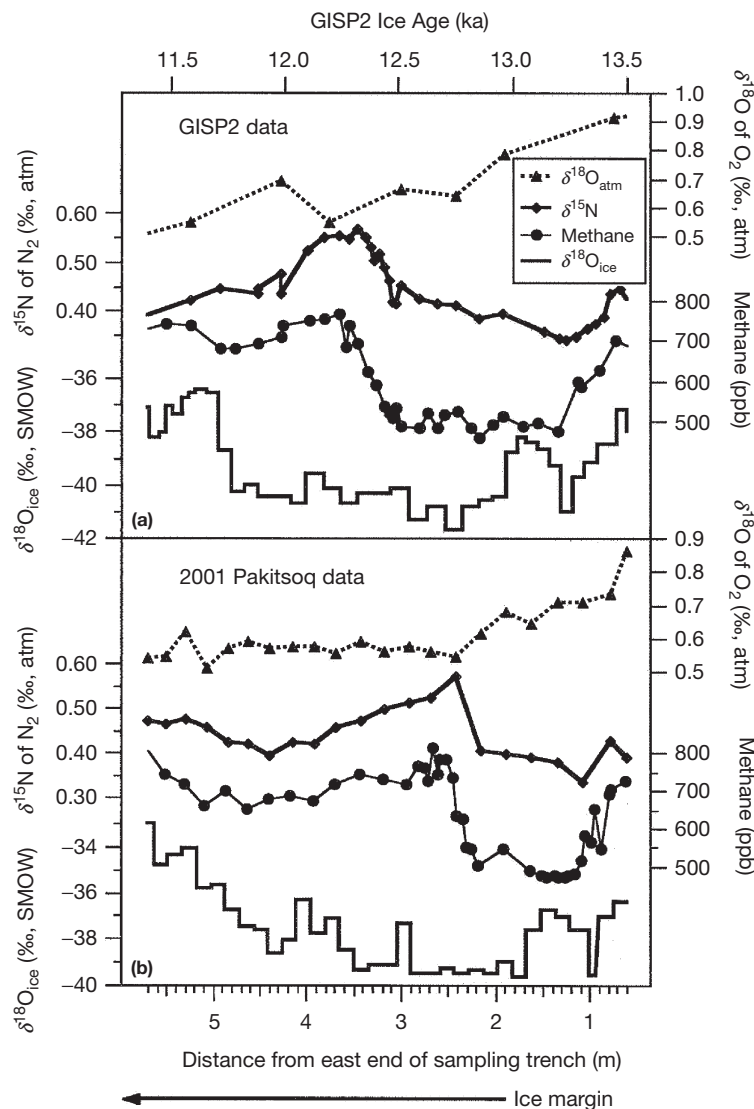


Figure 14 Comparison of GISP2 ice-core data with 2001 data from the Pakitsq ice margin, Greenland. (a) GISP2 $\delta^{18}\text{O}_{\text{ice}}$, $[\text{CH}_4]$, $\delta^{18}\text{O}_{\text{atm}}$, and $\delta^{15}\text{N}$ are shown on the ice timescale for comparison with Pakitsq ice. (b) Record from the 2001 trench sample profile at Pakitsq. Reproduced from Petrenko VV, Severinghaus JP, Brook EJ, Reeh N, and Schaefer H (2006) Gas records from the West Greenland ice margin covering the Last Glacial Termination: A horizontal ice core. *Quaternary Science Reviews* 25: 865–875, with permission.

partitioning of this ^{14}C between the CO and CO_2 phases. In addition, large uncertainties remain in estimates of ^{14}C production rates in ice.

The Pakitsq Ice-Margin Site

The ice-margin location that up to now has been studied in most detail is the Pakitsq ice margin in West Greenland (Figure 3; for location, see map in Figure 2). Studies at this site have been performed since 1985 (e.g., Reeh et al., 1987, 1991, 2002), providing information on $\delta^{18}\text{O}$ of surface ice, ice ablation, surface and bottom topography, and ice dynamics. The studies demonstrated that ice older than ~ 11.5 ka (Pre-Holocene ice) forms a ~ 500 m-wide band adjacent to the ice edge (Figure 11). Since 1992, the width of the Pre-Holocene ice band has diminished by $7\text{--}8$ m year^{-1} . In the same period, the ice thickness has decreased by almost 1 m year^{-1} , because ice ablation (~ 3 m year^{-1}) presently exceeds the vertical ice velocity supplying ice to the surface (~ 2 m year^{-1}). Visual inspection of shallow trenches cut into the ice indicate a $50\text{--}70^\circ$ dip of the stratigraphy.

In 2001, a major program was launched at the Pakitsq ice margin aimed at investigating the potential of this site as a source for recovery of very large-volume samples of uncontaminated, well-dated ancient ice (Petrenko et al., 2006). The main aim of the program was to obtain measurements of ^{14}C of methane ($^{14}\text{CH}_4$) in the ice and use these measurements to better understand changes in the global methane budget during the last glacial–interglacial transition. Meaningful $^{14}\text{CH}_4$ results required that the sampled ice be very well dated and that CH_4 in the ice was well preserved.

Dating of Pakitsq surface ice utilized a combination of $\delta^{18}\text{O}_{\text{ice}}$, $\delta^{18}\text{O}$ of O_2 ($\delta^{18}\text{O}_{\text{atm}}$), $\delta^{15}\text{N}$ of N_2 , and methane concentration ($[\text{CH}_4]$) measurements in near-surface sample profiles. These records were compared to the GISP2 ice-core record from Greenland Summit. An ice-sheet flow-line model (Reeh et al., 2002) predicts that Pakitsq ice dating to the last glacial termination originates near Greenland Summit, about 190 km south and 40 km west of the GISP2 site, making this a meaningful comparison.

Figure 14 compares $\delta^{18}\text{O}_{\text{ice}}$, $\delta^{18}\text{O}_{\text{atm}}$, $\delta^{15}\text{N}$ of N_2 , and $[\text{CH}_4]$ in samples collected in a short transect in 2001 at Pakitsq to GISP2 for the Younger Dryas (YD)–Preboreal (PB) transition. Figure 15 shows a record of $\delta^{15}\text{N}$ of N_2 and $[\text{CH}_4]$ from trench sample profiles taken in 2001–2003. As concluded by Petrenko et al. (2006), the unique combination of the four geochemical tracers presented in Figure 14 is evidence that the Pakitsq section contains a well-preserved YD–PB transition. Figure 15 demonstrates the ability to re-establish the position of YD ice each season, as well as the fact that Pakitsq gas records are reproducible from year to year, despite the ~ 3 m year^{-1} ablation rate.

The same four tracers as well as visual appearance of surface ice were also used to interpret Pakitsq surface ice stratigraphy and infer the three-dimensional structure of the layers (Schaefer et al., 2009). Figure 16 shows measurements on samples taken in three consecutive years in a longer horizontal transect at the Pakitsq site. The measurements showed that this transect crossed a syncline–anticline pair fold that contains multiple exposures of the YD and other climatic intervals dating to the glacial–interglacial transition. Measurements in another near-surface transect revealed that a layer of suspected

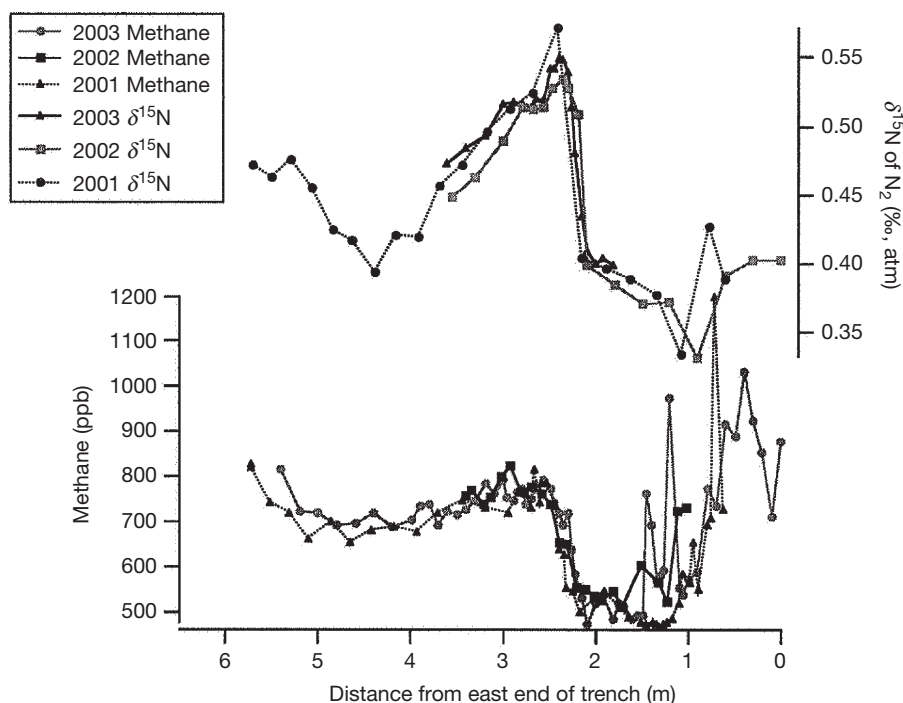


Figure 15 $\delta^{15}\text{N}$ of N_2 and $[\text{CH}_4]$ for 2001, 2002, and 2003 trench sample profiles from Pakitsq. Reproduced from Petrenko VV, Severinghaus JP, Brook EJ, Reeh N, and Schaefer H (2006) Gas records from the West Greenland ice margin covering the Last Glacial Termination: A horizontal ice core. *Quaternary Science Reviews* 25: 865–875, with permission.

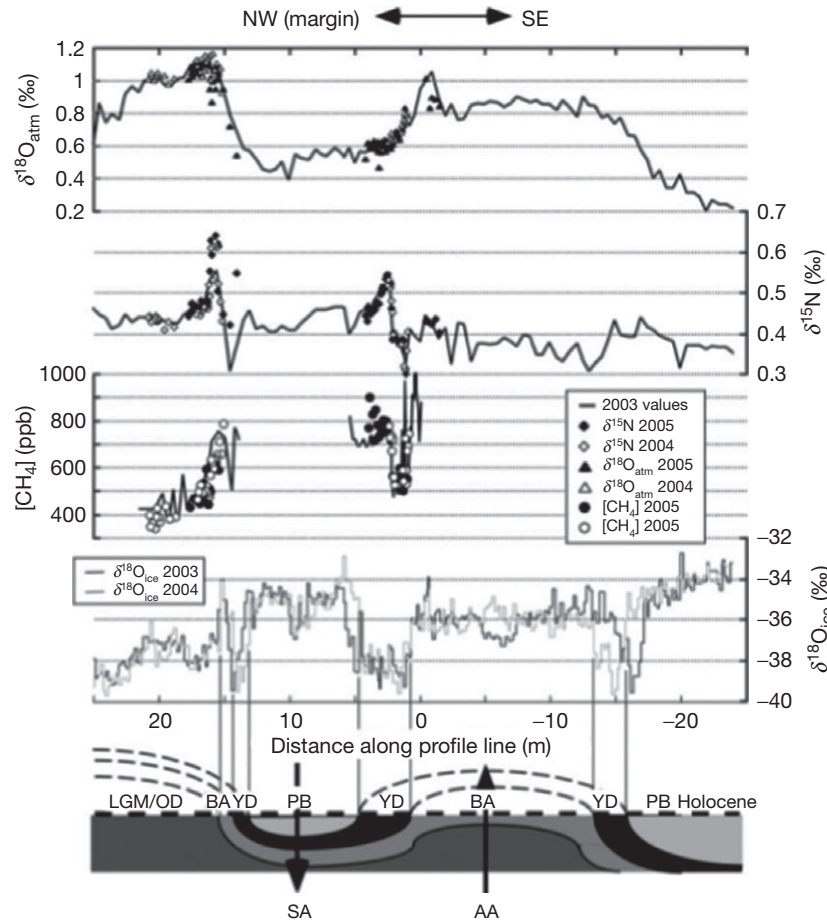


Figure 16 $\delta^{18}\text{O}_{\text{ice}}$ profiles and gas records from the main study area at Pakitsoq. The Last Glacial Maximum (LGM) to Bolling-Allerod (BA) sequence (at ~15–20 m) shows increasing age toward the margin (to NW), whereas the BA to YD to PB sequence (at ~0–5 m) is reversed and increases in age away from the margin (to SE). The cartoon at the bottom schematically illustrates the folding of the ice layers and indicates approximate positions of syncline axis (SA) and anticline axis (AA), as well as age of surface ice. Reproduced from [Schaefer H, Petrenko VV, Brook EJ, et al. \(2009\)](#) Ice stratigraphy at the Pakitsoq ice margin, West Greenland, derived from gas records. *Journal of Glaciology* 55: 411–421.

Eemian ice immediately adjacent to the ice edge was actually of Holocene origin and part of a large-scale fold that exposed another outcrop of the glacial-interglacial transition.

$[\text{CH}_4]$ records from multiple transects at Pakitsoq revealed that $[\text{CH}_4]$ was well preserved in many ice sections, but was elevated in others ([Petrenko et al., 2006](#); [Schaefer et al., 2009](#)). Using ice from sections with well-preserved CH_4 , [Schaefer et al. \(2006\)](#) produced the first record of $\delta^{13}\text{C}$ of CH_4 spanning the YD–PB climatic transition ([Figure 17](#)). The large and rapid YD–PB warming was associated with a ~50% increase in atmospheric CH_4 , and there has been considerable debate regarding the causes of the CH_4 increase. The [Schaefer et al. \(2006\)](#) record provided evidence that the increase was likely due to increased CH_4 emissions from wetlands rather than marine methane clathrates. Using very large volumes of ice (~1000 kg per sample) from the same YD–PB section, [Petrenko et al. \(2009\)](#) measured $^{14}\text{CH}_4$ over this climatic transition. The measurements showed no large changes in ^{14}C of CH_4 over the transition despite the large increase in CH_4 concentration, arguing strongly against

major involvement of ‘old’ CH_4 sources such as marine clathrates.

Other Studies

Ice-margin records also contain information about the dynamics of the ice sheet regions upstream of the ice-margin sites from which the records are retrieved. This is because geochemical tracers in the ice, besides showing large temporal variations reflecting climate change, also display large, systematic spatial variations over the surface of the accumulation region. These tracers measured on ice samples from the ice margin can therefore be used to link the ice sample to a specific site in the accumulation region and a specific time of deposition, and thereby provide information on ice sheet dynamics along the travel path of the ice parcel. The idea of using trace substances to obtain the time and site of origin of ice-margin ice was first exploited by [Dansgaard \(1961\)](#), who related $\delta^{18}\text{O}$ of West Greenland icebergs to the $\delta^{18}\text{O}$

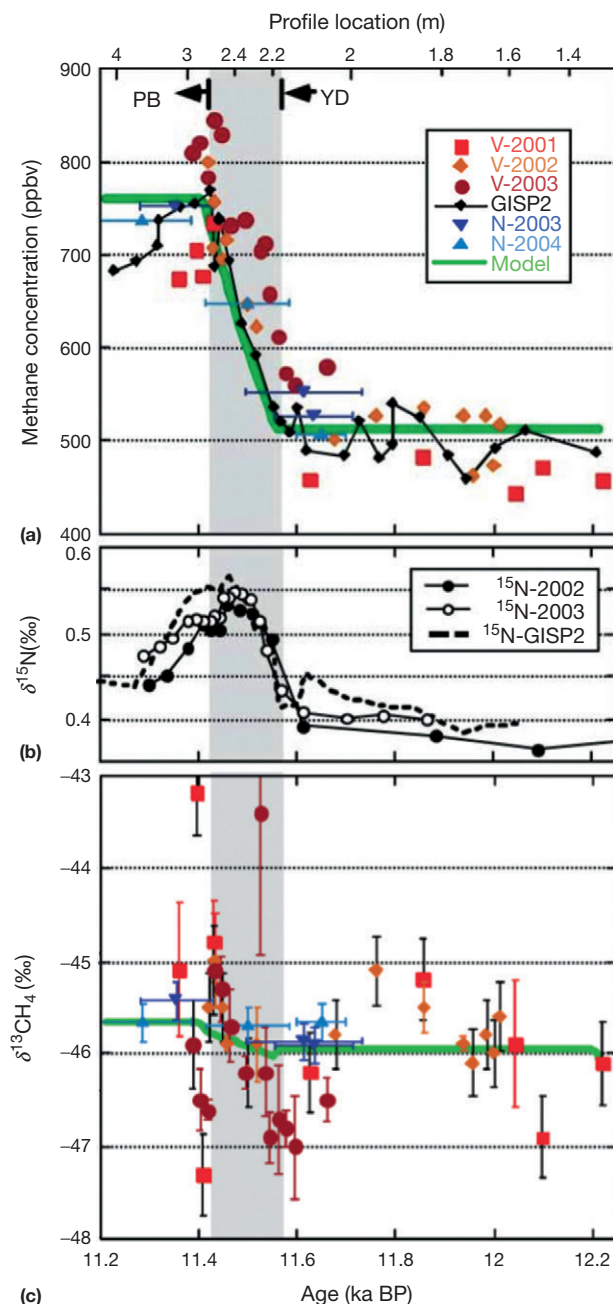


Figure 17 $[\text{CH}_4]$ (a), $\delta^{15}\text{N}$ of N_2 (b), and $\delta^{13}\text{CH}_4$ (c) from a Pakitsoq transect spanning the YD–PB transition (same transect as shown in Figures 14 and 15). Reproduced from Schaefer H, Whiticar MJ, Brook EJ, Petrenko VV, Ferretti DF, and Severinghaus JP (2006) Ice record of delta C-13 for atmospheric CH_4 across the Younger Dryas–Preboreal transition. *Science* 313: 1109–1112.

distribution of the surface snow in the accumulation area feeding the calving glaciers from which the icebergs had been detached. Reeh (1988) and Reeh et al. (2002) further advanced this work by combining isotope measurements with ice-flow model calculations to determine the location on the ice sheet where ice in the ice-margin record was originally deposited as snow.

It seems likely that ice-margin sites may contain the oldest glacial ice preserved on Earth. One possible example of this was mentioned briefly above for the Allan Hills BIA (Harvey et al., 1998). A more famous example is Beacon Valley in the Dry Valleys region of Antarctica, which contains ice that is covered by a thick layer of glacial till. The till surface contains contraction-crack polygons, and volcanic ash deposits have been found filling some of the contraction cracks (Sugden et al., 1995 and references therein). $^{40}\text{Ar}/^{39}\text{Ar}$ dating of these ash deposits yields ages near 8 My, suggesting that the ice may date as far back as the Miocene (Sugden et al., 1995). Cosmogenic nuclide dating of the surface till material has also suggested a very old minimum age for the buried ice of ~ 2.3 My (Schafer et al., 2000). However, other studies have suggested that the ice may be younger (e.g., Ng et al., 2005). It is hypothesized that the buried Beacon Valley ice was originally derived from an expanded Taylor Glacier. The till layer above the ice formed from entrained material that became exposed as the ice sublimated. As the till layer thickened, it dramatically reduced the ice sublimation rates, allowing this ice to persist for millions of years.

The extremely old margin ice has been of interest to biologists studying microbial survival in extreme environments. Bidle et al. (2007) analyzed five samples of ancient ice from Beacon Valley and the adjacent Mullins Valley for cell viability and state of community DNA. They found that, surprisingly, even ice thought to be ~ 8 My in age contained some viable microbes. The study found, however, that viability and DNA are compromised with age and suggested that average community DNA declines exponentially with a half-life of ~ 1.1 My. Another study included examination of a single sample of Beacon Valley ice, and found no evidence of bacterial survival (Johnson et al., 2007).

Outlook and Ongoing Work

The full potential of ice-margin sites as archives of easily accessible ice paleoenvironmental records is just beginning to be realized. Several exciting projects are in various stages of progress as of the time of this writing.

A horizontal ice-core record of $\delta^{18}\text{O}_{\text{ice}}$ has been recently measured for Mount Moulton (Popp et al., 2004). This record is constrained by radioisotopic $^{40}\text{Ar}/^{39}\text{Ar}$ dates on several ash layers in the ice, and includes the complete penultimate glacial–interglacial transition for the first time for West Antarctic ice.

A project to date the stratigraphy of the ~ 80 -km long ablation zone of Taylor Glacier is underway in the Dry Valleys, Antarctica. One of the main aims of the project is to establish a well-dated, easily accessible ice archive with Ice Ages between Holocene and >100 ka. This project also aims to use Taylor Glacier ice for constraining *in situ* cosmogenic ^{14}C production rates in ice, as well as for $^{13}\text{CH}_4$ and $^{14}\text{CH}_4$ paleoatmospheric reconstructions.

A similar effort is underway at the Allan Hills BIA, Antarctica. This project aims to establish a surface ice paleoarchive that could extend as far as 2.5 My back in time. A further project is attempting to more reliably date the buried ice in the Dry Valleys region, with the hope of extending the ice paleoarchive even further back in time.

See also: [Ice Core Methods: Stable Isotopes](#). [Ice Core Records: Antarctic Stable Isotopes](#); [Correlations Between Greenland and Antarctica](#); [Greenland Stable Isotopes](#). [Ice Cores: Dynamics of the East Antarctic Ice Sheet](#); [Dynamics of the Greenland Ice Sheet](#); [Dynamics of the West Antarctic Ice Sheet](#); [History of Research, Greenland and Antarctica](#).

References

- Aciego SM, Cuffey KM, Kavanaugh JL, Morse DL, and Severinghaus JP (2007) Pleistocene ice and paleo-strain rates at Taylor Glacier, Antarctica. *Quaternary Research* 68: 303–313.
- Andersen KK, Azuma N, Barnola JM, et al. (2004) High-resolution record of Northern Hemisphere climate extending into the last interglacial period. *Nature* 431: 147–151.
- Augustin L, Barbante C, Barnes PRF, et al. (2004) Eight glacial cycles from an Antarctic ice core. *Nature* 429: 623–628.
- Bidle KD, Lee S, Marchant DR, and Falkowski PG (2007) Fossil genes and microbes in the oldest ice on Earth. *Proceedings of the National Academy of Sciences of the United States of America* 104: 13455–13460.
- Bintanja R (1999) On the glaciological, meteorological, and climatological significance of Antarctic blue ice areas. *Reviews of Geophysics* 37: 337–359.
- Clausen H and Stauffer B (1988) Analyses of two ice cores drilled at the ice-sheet margin in West Greenland. *Annals of Glaciology* 10: 23–27.
- Coachman LK, Enns T, and Scholander PF (1958a) Gas loss from a temperate glacier. *Tellus* 10: 493–495.
- Coachman LK, Hemmingsen E, and Scholander PF (1956) Gas enclosures in a temperate glacier. *Tellus* 8: 415–423.
- Coachman LK, Hemmingsen E, Scholander PF, Enns T, and Vries HD (1958b) Gases in glaciers. *Science* 127: 1288–1289.
- Curzio P, Folco L, Laurenzi MA, Mellini M, and Zeoli A (2008) A tephra chronostratigraphic framework for the Frontier Mountain blue-ice field (northern Victoria Land, Antarctica). *Quaternary Science Reviews* 27: 602–620.
- Dansgaard W (1961) The isotopic composition of natural waters. *Meddelelser om Grønland* 165: 1–120.
- Dunbar NW, McIntosh WC, and Esser RP (2008) Physical setting and tephrochronology of the summit caldera ice record at Mount Moulton, West Antarctica. *Geological Society of America Bulletin* 120: 796–812.
- Dunbar NW, Zielinski GA, and Voisins DT (2003) Tephra layers in the Siple Dome and Taylor Dome ice cores, Antarctica: Sources and correlations. *Journal of Geophysical Research-Solid Earth* 108: JB002056.
- Faure G and Buchanan D (1987) Glaciology of East Antarctic ice sheet at the Allan Hills: A preliminary interpretation. *Antarctic Journal of the United States* XXII: 74–75.
- Fireman EL and Norris TL (1982) Ages and composition of gas trapped in Allan-Hills and Byrd core ice. *Earth and Planetary Science Letters* 60: 339–350.
- Fuji Y, Kohno M, Motoyama H, et al. (1999) Tephra layers in Dome Fuji (Antarctica) deep ice core. *Annals of Glaciology* 29: 126–130.
- Grootes P (1990) ^{18}O results from blue-ice areas: 'Old' ice at the surface? In: Cassidy WA and Whillans IM (eds.) *Workshop on Antarctic Meteorite Stranding Surfaces*. Lunar and Planetary Institute Technical Report, pp. 67–69.
- Harvey R (2003) The origin and significance of Antarctic meteorites. *Chemie Der Erde-Geochemistry* 63: 93–147.
- Harvey RP, Dunbar NW, McIntosh WC, et al. (1998) Meteoritic event recorded in Antarctic ice. *Geology* 26: 607–610.
- Hooke RL (1976) Pleistocene ice at the base of the Barnes Ice Cap Baffin Island, N.W.T., Canada. *Journal of Glaciology* 17: 49–59.
- Hooke RL and Clausen HB (1982) Wisconsin and Holocene Delta-0-18 variations, Barnes Ice Cap, Canada. *Geological Society of America Bulletin* 93: 784–789.
- Johnson SS, Hebsgaard MB, Christensen TR, et al. (2007) Ancient bacteria show evidence of DNA repair. *Proceedings of the National Academy of Sciences of the United States of America* 104: 14401–14405.
- Jouzel J, Alley RB, Cuffey KM, et al. (1997) Validity of the temperature reconstruction from water isotopes in ice cores. *Journal of Geophysical Research-Oceans* 102: 26471–26487.
- Katsushima T, Nishio F, Ohmae H, Ishikawa M, and Takahashi S (1984) Composition of dirt layers in the bare ice areas near the Yamato Mountains in Queen Maud Land and the Allan Hills in Victoria Land, Antarctica. *Memoirs of National Institute of Polar Research* 34: 174–187 (special issue).
- Keys JR, Anderton PW, and Kyle PR (1977) Tephra and debris layers in Skelton Neve and Kempe Glacier, south Victoria Land, Antarctica. *New Zealand Journal of Geology and Geophysics* 20: 971–1002.
- Koeberl C (1990) Dust bands in blue ice fields in Antarctica and their relationship to meteorites and ice. In: Cassidy WA and Whillans IM (eds.) *Workshop on Antarctic Meteorite Stranding Surfaces*. Lunar and Planetary Institute Technical Report, pp. 70–73.
- Koeberl C, Yanai K, Cassidy WA, and Schutt JW (1988) Investigation of dust bands from blue ice fields in the Lewis Cliff (Beardmore) area, Antarctica: A progress report. *Proceedings of NIPR Symposium on Antarctic Meteorites* 1: 291–309.
- Lal D, Jull AJT, Donahue DJ, Burtner D, and Nishiizumi K (1990) Polar ice ablation rates measured using in situ cosmogenic C-14. *Nature* 346: 350–352.
- Lal D, Nishiizumi K, and Arnold JR (1987) In situ cosmogenic H-3, C-14, and Be-10 for determining the net accumulation and ablation rates of ice sheets. *Journal of Geophysical Research-Solid Earth and Planets* 92: 4947–4952.
- Lorius C (1968) A physical and chemical study of the coastal ice sampled from a core drilling in Antarctica. *IAHS Publication* 79: 141–148.
- Marvin U (1986) Components of dust bands in ice from Cul de Sac, Allan Hills region, Antarctica. *Meteoritics* 21: 442–443.
- Ng F, Hallet B, Sletten RS, and Stone JO (2005) Fast-growing till over ancient ice in Beacon Valley, Antarctica. *Geology* 33: 121–124.
- Nishiizumi K, Arnold JR, Elmore D, Ma X, Newman D, and Gove HE (1983) Cl-36 and Mn-53 in Antarctic meteorites and Be-10–Cl-36 dating of Antarctic ice. *Earth and Planetary Science Letters* 62: 407–417.
- Nishio F, Katsushima T, and Ohmae H (1985) Volcanic ash layers in bare ice areas near the Yamato Mountains, Dronning Maud Land and the Allan Hills, Victoria Land, Antarctica. *Annals of Glaciology* 7: 34–41.
- Palais J (1985) Particle morphology, composition and associated ice chemistry of tephra layers in the Byrd ice core: Evidence for hydrovolcanic eruptions. *Annals of Glaciology* 7: 42–48.
- Perchiazzi N, Folco L, and Mellini M (1999) Volcanic ash bands in the Frontier Mountain and Lichen Hills blue-ice fields, northern Victoria Land. *Antarctic Science* 11: 353–361.
- Petit JR, Jouzel J, Raynaud D, et al. (1999) Climate and atmospheric history of the past 420,000 years from the Vostok ice core, Antarctica. *Nature* 399: 429–436.
- Petrenko VV, Severinghaus JP, Brook EJ, Reeh N, and Schaefer H (2006) Gas records from the West Greenland ice margin covering the Last Glacial Termination: A horizontal ice core. *Quaternary Science Reviews* 25: 865–875.
- Petrenko VV, Smith AM, Brook EJ, et al. (2009) (CH_4) -C-14 measurements in Greenland ice: Investigating last glacial termination CH_4 sources. *Science* 324: 506–508.
- Popp T, Sowers T, Dunbar N, McIntosh W, and White JWC (2004) Radioisotopically dated climate record spanning the last interglacial in ice from Mount Moulton, West Antarctica. *Poster Presented at the AGU Fall Meeting*, San Francisco: AGU December 2004.
- Reeh N (1988) A flow-line model for calculating the surface profile and the velocity, strain-rate, and stress-fields in an ice-sheet. *Journal of Glaciology* 34: 46–54.
- Reeh N, Oerter H, Letreguilly A, Miller H, and Hubberten HW (1991) A new, detailed ice-age 0-18 record from the ice-sheet margin in central West Greenland. *Global and Planetary Change* 90: 373–383.
- Reeh N, Oerter H, and Miller H (1993) Correlation of Greenland ice-core and ice-margin $\delta(^{18}\text{O})$ records. In: Peltier WR (ed.) *Ice in the Climate System*. NATO ASI Series I: *Global Environmental Change*, vol. 12, pp. 481–497. Berlin: Springer.
- Reeh N, Oerter H, and Thomsen HH (2002) Comparison between Greenland ice-margin and ice-core oxygen-18 records. *Annals of Glaciology* 35: 136–144.
- Reeh N, Thomsen HH, and Clausen HB (1987) The Greenland ice-sheet margin – A mine of ice for paleo-environmental studies. *Palaeogeography, Palaeoclimatology, Palaeoecology* 58: 229–234.
- Schaefer H, Petrenko VV, Brook EJ, et al. (2009) Ice stratigraphy at the Pakitsq ice margin, West Greenland, derived from gas records. *Journal of Glaciology* 55: 411–421.
- Schaefer H, Whiticar MJ, Brook EJ, Petrenko VV, Ferretti DF, and Severinghaus JP (2006) Ice record of delta C-13 for atmospheric CH_4 across the Younger Dryas–Preboreal transition. *Science* 313: 1109–1112.
- Schäfer H, Baur H, Denton GH, et al. (2000) The oldest ice on Earth in Beacon Valley, Antarctica: New evidence from surface exposure dating. *Earth and Planetary Science Letters* 179: 91–99.
- Scholander PF, de Vries H, Dansgaard W, Coachman LK, Nutt DC, and Hemmingsen E (1962) Radiocarbon age and oxygen-18 content of Greenland icebergs. *Meddelelser om Grønland* 165: 1–26.
- Steig EJ, Morse DL, Waddington ED, et al. (2000) Wisconsin and Holocene climate history from an ice core at Taylor Dome, western Ross Embayment, Antarctica. *Geografiska Annaler Series A-Physical Geography* 82A: 213–235.

- Sugden DE, Marchant DR, Potter N, et al. (1995) Preservation of Miocene glacier ice in East Antarctica. *Nature* 376: 412–414.
- Tison JL, Petit JR, Barnola JM, and Mahaney WC (1993) Debris entrainment at the ice-bedrock interface in sub-freezing temperature conditions (Terre-Adelie, Antarctica). *Journal of Glaciology* 39: 303–315.
- van de Wal RSW, van der Borg K, Oerter H, Reeh N, Dejong AFM, and Oerlemans J (1990) Progress in C-14 dating of ice at Utrecht. *Nuclear Instruments and Methods in Physics Research Section B: Beam Interactions with Materials and Atoms* 52: 469–472.
- van der Borg K, van der Kemp WJM, Alderliesten C, de Jong AFM, and Lamers RAN (2001) In-situ radiocarbon production by neutrons and muons in an Antarctic blue ice field at Scharffenbergbotnen: A status report. *Radiocarbon* 43: 751–757.
- van der Kemp WJM, Alderliesten C, van der Borg K, et al. (2002) In situ produced C-14 by cosmic ray muons in ablating Antarctic ice. *Tellus Series B: Chemical and Physical Meteorology* 54: 186–192.
- van Roijen JJ, Bintanja R, van der borg K, van den Broeke MR, Dejong AFM, and Oerlemans J (1994) Dry extraction of (CO₂)-C-14 and (CO)-C-14 from Antarctic ice. *Nuclear Instruments and Methods in Physics Research Section B: Beam Interactions with Materials and Atoms* 92: 331–334.
- van Roijen J, van der Borg K, DeJong A, and Oerlemans J (1995) A correction for in-situ C-14 in Antarctic ice with (CO)-C-14. *Radiocarbon* 37: 165–169.
- Whillans IM and Cassidy WA (1983) Catch a falling star – Meteorites and old ice. *Science* 222: 55–57.
- Wilch TI, McIntosh WC, and Dunbar NW (1999) Late Quaternary volcanic activity in Marie Byrd Land: Potential Ar-40/Ar-39-dated time horizons in West Antarctic ice and marine cores. *Geological Society of America Bulletin* 111: 1563–1580.
- Yoshida M, Ando H, Omoto K, Naruse R, and Ageta Y (1971) Discovery of meteorites near Yamato Mountains, East Antarctica. *Antarctic Record* 39: 62–65.
- Zdanowicz CM, Fisher D, Clark ID, and Lacelle D (2002) An ice-marginal $\delta^{18}\text{O}$ record from Barnes Ice Cap, Baffin Island, Canada. *Annals of Glaciology* 35: 145–149.

Thermal Diffusion Paleotemperature Records

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Introduction

It came as a surprise, when in 1917, Chapman realized from some theoretical derivations, that there was a phenomenon of ‘thermal diffusion’, wherein diffusion is driven not by a concentration gradient, but by a temperature gradient (Chapman, 1917). A few years before that Enskog independently arrived at the same conclusion, however, being more mathematically oriented, he did not realize the importance of the new discovery for physics (Enskog, 1917). Chapman and Dootson performed an experiment in which they maintained a mixture of two gases in a temperature difference which made it separate slightly as predicted (Chapman and Dootson, 1917). Many researchers followed this pursuit, and measured thermal diffusion constants for various gas mixtures. They noted that predictions of the theory by Chapman and Enskog worked especially well for mixtures of noble gases. Overall, the theory was found to be very successful, and pointed to the severity of collisions between molecules as the underlying driver of thermal diffusion (Mason et al., 1966).

In 1998, Severinghaus and colleagues realized that heavy gas-isotopic enrichments in ice cores immediately following abrupt climate warmings were caused by thermal diffusion. Moreover, this previously unexplained small isotopic artifact provided a powerful means to reconstruct magnitudes of climate changes (Severinghaus et al., 1998; Severinghaus and Brook, 1999). If constants of thermal diffusion for species of interest in air are known, the magnitude of temperature difference that separated them can be reconstructed. These workers were able to place, for the first time, thermal diffusion-based quantitative constraints on the prominent Younger Dryas-Preboreal, and Bølling transitions manifested in fossil air from the Greenland Ice Sheet Project 2 (GISP2) ice core.

The importance of the thermal diffusion ice-core paleotemperature-change recorder was emphasized by Jouzel (1999) and Alley (2000), as the difficulties associated with the conventional water isotope thermometer in ice ($\delta^{18}\text{O}_{\text{ice}}$) called for an independent check of temperature-change estimations. The empirical calibration of the conventional $\delta^{18}\text{O}$ thermometer was likely variable, especially at times of abrupt transitions, when other factors affecting it were changing (seasonality of snowfall, changing moisture sources, etc.; Jouzel et al., 1997). The problems with the conventional thermometer were partly uncovered by the so-called borehole $\delta^{18}\text{O}$ calibration, which makes use of the known cold pulse measured today deep inside the ice sheet, that reflects how cold the ice sheet was during the last glacial period (Cuffey et al., 1995). Subsequent work proved that the thermal diffusion explanation of the isotope artifact was correct (e.g., Severinghaus et al., 2001), and allowed documentation of the magnitudes of a few additional prominent abrupt climate changes, such as those in

Greenland ~70 000 years before present (BP; Lang et al., 1999); 45 000 years BP (Landaï et al., 2004), and ~8000 years BP (Leuenberger et al., 1999).

Controls on Isotopic Composition of Ice-Core Air

Current atmosphere is preserved in the Greenland ice sheet as the interstitial spaces in densified snow at ~100-m depth are closed off, as has occurred previously. Prior to becoming entrapped in newly formed ice, the air is free to diffuse up to the surface and back. The diffusive environment allows for two important fractionation processes to occur: thermal diffusion (Severinghaus et al., 1998) and ‘gravitational settling’ (Craig et al., 1988).

Thermal diffusion takes place if the top boundary (surface) is warmer (or colder) than the bottom boundary (close-off depth). The nonequal temperature, in turn, originates from climate change, which affects the top boundary first. The composition of the enclosed air is then slightly offset from atmospheric air, because of the thermal diffusion process. In addition, there is a slight tendency of heavier species in air to settle down, hence the term gravitational settling (Schwander, 1989). The composition offset of the enclosed air caused by the two processes is typically small (<1%), and is best noticed in its stable isotopes, which can be measured with marked sensitivity by mass spectrometry.

Measuring the Thermal Diffusion Trace

To measure a trace of thermal diffusion, a piece of ice from known depth is obtained (e.g., from 1820 m depth at Greenland summit, which corresponds to 14 650 years of age for the enclosed gas), and fossil air is removed under vacuum (Severinghaus and Brook, 1999). The resulting air sample is passed through a mass spectrometer to measure stable isotope ratios, most notably nitrogen (N_2) and argon (Ar) that are abundant and unchanging in the atmosphere (the measured ratios are $^{29}\text{N}_2/^{28}\text{N}_2$ and $^{40}\text{Ar}/^{36}\text{Ar}$).

If the fossil air has undergone some gravitational fractionation of its isotopes, but has not been affected by thermal diffusion, any anomalies in nitrogen (29/28) will always be exactly equal to those in argon (40/36) divided by four. That is explained by the mass difference governing gravitational settling being the same in both cases:

$$\Delta M(^{29}\text{N}_2, ^{28}\text{N}_2) = \Delta M(^{40}\text{Ar}, ^{36}\text{Ar})/4 = 1 \text{ a.m.u.} \quad [1]$$

However, if thermal diffusion has played a role in fractionating isotopes in air, the anomaly in nitrogen (29/28) will be bigger than that in argon (40/36) divided by four. In fact, the difference of the two terms, usually close to zero unless thermal

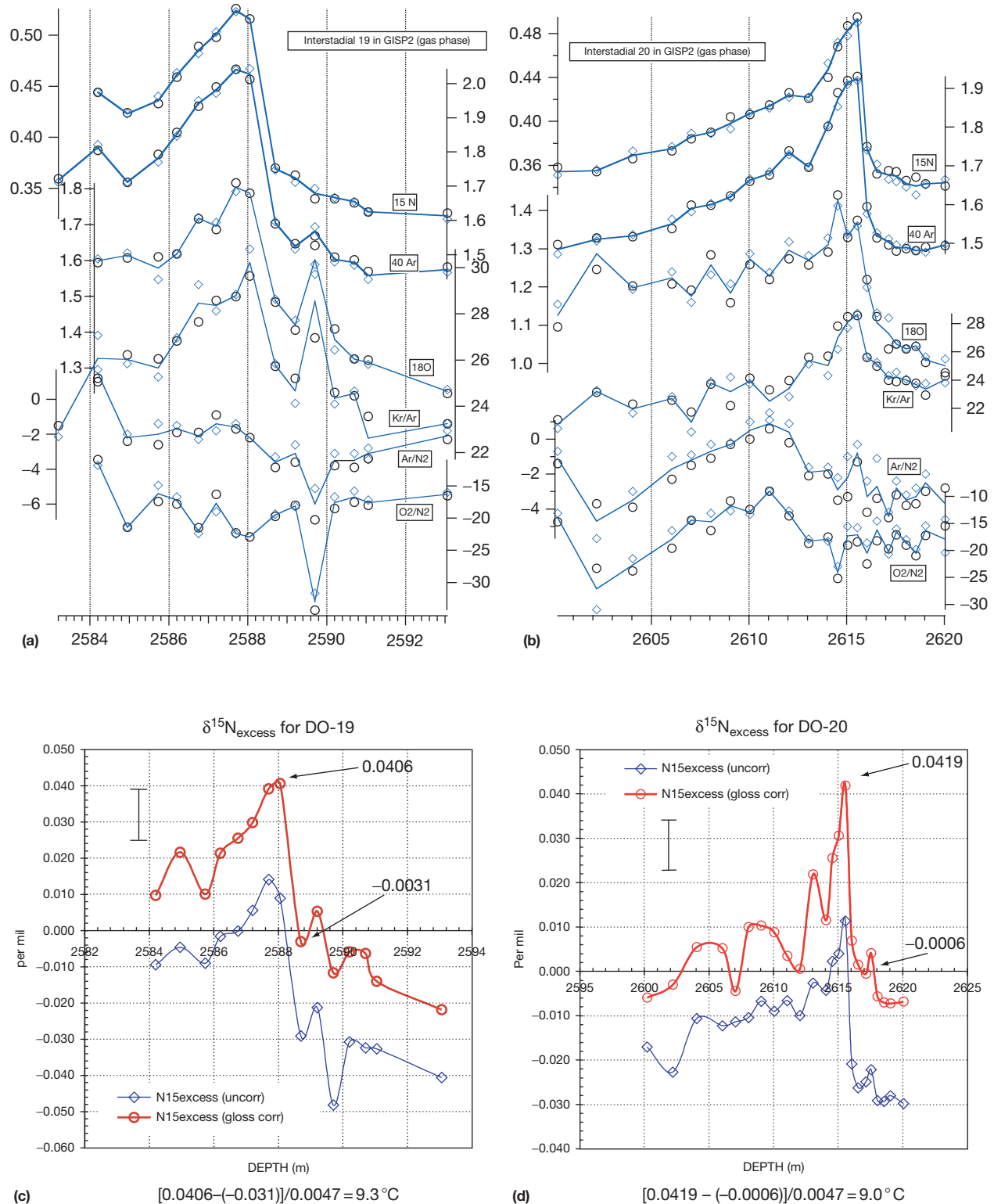


Figure 1 (a) Measured profile of $\delta^{15}\text{N}$, $\delta^{40}\text{Ar}$, and other gases encompassing the depth interval 2584–2593 m in the GISP2 ice core (~69 000 years BP). (b) Measured profile of $\delta^{15}\text{N}$, $\delta^{40}\text{Ar}$, and other gases encompassing the depth interval 2600–2620 m in the GISP2 ice core (~73 000 years BP). (c) Derived profile of $\delta^{15}\text{N}_{\text{excess}}$ for the depth interval 2584–2593 m, constructed using the $\delta^{15}\text{N}$ and $\delta^{40}\text{Ar}$ data of (a). A small correction for gas loss is made to account for the fact that argon is slightly fractionated during the bubble close-off process and storage, hence 'gloss corr', or gas loss-corrected values (Grachev, 2004). (d) Derived profile of $\delta^{15}\text{N}_{\text{excess}}$ for the depth interval 2600–2620 m constructed using the $\delta^{15}\text{N}$ and $\delta^{40}\text{Ar}$ data of (b).

diffusion played a role, even if less important ‘bubble close-off’ fractionation processes have occurred (Severinghaus and Battle, 2006), serves as a marker of purely thermal diffusion fractionation:

$$\delta^{15}\text{N}_{\text{excess}} = \delta^{15}\text{N}_{\text{measured}} - \delta^{40}\text{Ar}_{\text{measured}}/4, \quad [2]$$

(Severinghaus and Brook, 1999), where the delta notation is employed: $\delta^{15}\text{N}_{\text{measured}} = [(^{29}\text{N}_2/^{28}\text{N}_2)_{\text{sample}}/(^{29}\text{N}_2/^{28}\text{N}_2)_{\text{reference}} - 1] \times 1000\text{‰}$, (one per mil unit (‰) is equal to 0.1% (reference may be defined as the measured ratio in current atmosphere)).

Laboratory experiments have shown that the sensitivity of $\delta^{15}\text{N}$ to thermal diffusion is $\sim 14.8 \times 10^{-3} \text{‰/°C}$, while that of argon is $40.4 \times 10^{-3} \text{‰/°C}$ (at an average temperature of -30°C ; Grachev and Severinghaus, 2003a,b). In other words, for example, the cold side of a vessel with air subjected to a 10°C temperature difference will be enriched by $\sim 0.0404\text{‰}$ in $\delta^{40}\text{Ar}$ relative to the warm side. Hence, the thermal diffusion sensitivity of the combined quantity $\delta^{15}\text{N}_{\text{excess}}$ of Eq. 2 can be found as follows:

$$\begin{aligned} \Omega(\delta^{15}\text{N}_{\text{excess}}) &= 14.8 \times 10^{-3} - 40.4 \times 10^{-3}/4 \\ &= 4.7 \times 10^{-3} \text{‰/°C}, \end{aligned} \quad [3]$$

which is in essence the calibration of the thermal diffusion paleotemperature-change recorder, constant through time.

Interpretation of Observed Thermal Diffusion Trace

From the preceding section it follows, that if only the $\delta^{15}\text{N}$ of an ice sample is measured, it is not possible to determine how much of the measured fractionation is due to thermal diffusion. Alternatively, if both $\delta^{15}\text{N}$ and $\delta^{40}\text{Ar}$ are measured, the calculated nonzero $\delta^{15}\text{N}_{\text{excess}}$ is purely the result of thermal diffusion. Moreover, if, for example, a value of $\delta^{15}\text{N}_{\text{excess}}$ on an ice sample of a certain age of $47 \times 10^{-3}\text{‰}$ is determined, it can be inferred that a climate change occurred at that time, with a deduced magnitude of:

$$\begin{aligned} \Delta T &= \delta^{15}\text{N}_{\text{excess}}/\Omega(\delta^{15}\text{N}_{\text{excess}}) \\ &= 47 \times 10^{-3} \text{‰}/4.7 \times 10^{-3} \text{‰/°C} = 10 \text{°C} \end{aligned} \quad [4]$$

Usually the isotopes are measured on a sequence of ice pieces, for example between 2584 m and 2593 meters depth in the Greenland GISP2 ice core, which allows a time profile of $\delta^{15}\text{N}_{\text{excess}}$ to be constructed (Figures 1a and b; Grachev, 2004). Such a profile serves several purposes: (i) it allows us to check whether the background value of $\delta^{15}\text{N}_{\text{excess}}$ is at its ‘relaxed’ zero value, or is somewhat offset, perhaps not having fully recovered from a preceding climate change; (ii) the maximum of the $\delta^{15}\text{N}_{\text{excess}}$ peak can be found; (iii) it can be determined at exactly what depth the abrupt warming sets in. Relating to the last point, an important comparison of $\delta^{15}\text{N}_{\text{excess}}$ with CH_4 measurements in ice fossil air allows us to decipher which region was first to change directly: the tropics, which presumably generate large proportions of atmospheric methane at glacial times, or Greenland temperature, suggesting mechanisms of abrupt climate change (Figure 2; Severinghaus and Brook, 1999).

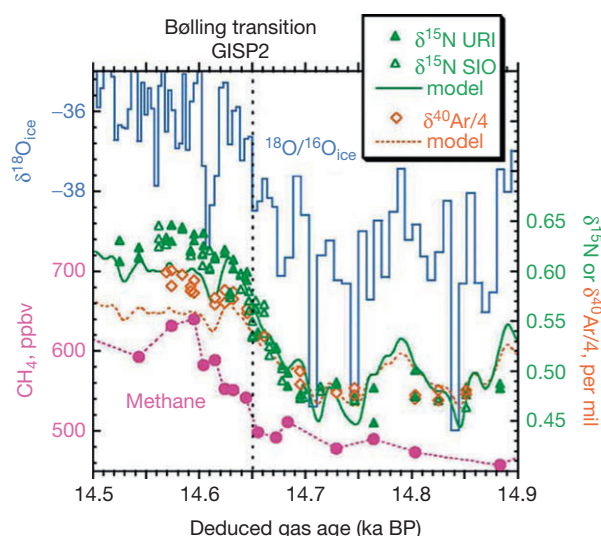


Figure 2 Climate warming as reflected in N and Ar isotopes versus change in CH_4 at the Bölling transition ($\sim 14\,650$ years BP) (from Severinghaus and Brook, 1999; reproduced with permission). Direct comparison of the timing of change is possible, because the air age scale is the same for these indicators. The conventional ice water-isotope thermometer in the top panel serves well for reference however its age scale has to be adjusted (the so-called ‘gas age-ice age difference’ problem), and its empirical sensitivity coefficient to surface temperature has been found to vary markedly (Jouzel, 1999).

Direct Climate Change Reconstruction

A direct way of obtaining a climate-change reconstruction is to determine the magnitude of change in $\delta^{15}\text{N}_{\text{excess}}$ between the maximum of the peak and the baseline (Figures 1c and d). When the result is divided by the thermal diffusion sensitivity of $\delta^{15}\text{N}_{\text{excess}}$ of $4.7 \times 10^{-3} \text{‰/°C}$, it yields the approximated magnitude of the climate change. The actual climate change is usually best estimated at $\sim 1\text{--}3 \text{°C}$ above that value for central Greenland, because of the slow warming of the base of the firm that occurs in concert with thermal diffusion (e.g., Grachev and Severinghaus, 2005).

A direct thermal diffusion ice core reconstruction of an abrupt climate change can be performed according to the equation:

$$\begin{aligned} (\Delta T)_{\text{climate}} &= [\delta^{15}\text{N}_{\text{excess}}(\text{max}) \\ &\quad - \delta^{15}\text{N}_{\text{excess}}(\text{baseline})]/\Omega(\delta^{15}\text{N}_{\text{excess}}) \\ &\quad + (\sim 1 - 3 \text{°C heat correction}) \end{aligned} \quad [5]$$

(Figures 1c and d). It should be noted that the thermal diffusion recorder has been shown to work well for fast changes, such as the Dansgaard-Oeschger events of the last glacial period (Dansgaard et al., 1993). The reason is due to the requirement of a sustained temperature difference in the top portion of the ice sheet, where thermal diffusion proceeds and fractionates the air that subsequently becomes fossilized.

See also: K/Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ Dating. **Ice Core Methods:** Stable Isotopes. **Ice Core Records:** Antarctic Stable Isotopes; Correlations Between Greenland and Antarctica; Greenland Stable Isotopes.

References

- Alley RB (2000) The Younger Dryas cold interval as viewed from central Greenland. *Quaternary Science Reviews* 19: 213–226.
- Chapman S (1917) The kinetic theory of simple and composite monatomic gases: Viscosity thermal conduction, and diffusion. *Kinetic Theory* vol. 3, p. 102. New York: Pergamon. Reprinted in Brush S. G. (1972).
- Chapman S and Dootson FW (1917) A note on thermal diffusion. *Philosophical Magazine* 33: 248–253.
- Craig H, Horibe Y, and Sowers T (1988) Gravitational separation of gases and isotopes in polar ice caps. *Science* 242: 1675–1678.
- Cuffey KM, Clow GD, Alley RB, Stuiver M, Waddington ED, and Saltus RW (1995) Large Arctic Temperature Change at the Wisconsin-Holocene Glacial Transition. *Science* 270: 455–458.
- Dansgaard W, Johnsen SJ, Clausen HB, et al. (1993) Evidence for general instability of past climate from a 250-kyr ice-core record. *Nature* 364: 218–220.
- Enskog D (1917) Kinetic theory of processes in dilute gases. 3. In: *Kinetic Theory*, pp. 102. New York: Pergamon Reprinted in: Brush S. G. (1972).
- Grachev AM (2004) Laboratory-determined air thermal diffusion constants applied to reconstructing the magnitudes of past abrupt temperature changes from gas isotope observations in polar ice cores (Greenland). PhD Thesis, p. 297. San Diego: University of California.
- Grachev AM and Severinghaus JP (2003a) Laboratory determination of thermal diffusion constants for $^{29}\text{N}_2/^{28}\text{N}_2$ in air at temperatures from -60 to 0 °C for reconstruction of magnitudes of abrupt climate changes using the ice core fossil-air paleothermometer. *Geochimica et Cosmochimica Acta* 67: 345–360.
- Grachev AM and Severinghaus JP (2003b) Determining the thermal diffusion factor for $^{40}\text{Ar}/^{36}\text{Ar}$ in air to aid paleoreconstruction of abrupt climate change. *Journal of Physical Chemistry* 23A: 4636–4642.
- Grachev AM and Severinghaus JP (2005) A revised $+10 \pm 4$ °C magnitude of the abrupt change in Greenland temperature at the Younger Dryas termination using published GISP2 gas isotope data and air thermal diffusion constants.
- Jouzel J, Alley RB, Cuffey KM, et al. (1997) Validity of the temperature reconstruction from water isotopes in ice cores. *Journal of Geophysical Research* 102: 26,471–26,487.
- Jouzel J (1999) Calibrating the isotopic paleothermometer. *Science* 286: 910–911.
- Landais A, Caillon N, Goujon C, et al. (2004) Quantification of rapid temperature change during DO event 12 and phasing with methane inferred from air isotopic measurements. *Earth and Planetary Science Letters* 225: 221–232.
- Lang C, Leuenberger M, Schwander J, and Johnsen S (1999) 16 °C rapid temperature variation in Central Greenland 70 000 years ago. *Science* 286: 934–937.
- Leuenberger MC, Lang C, and Schwander J (1999) Delta(15)N measurements as a calibration tool for the paleothermometer and gas-ice age differences: a case study for the 8200 BP event on GRIP ice. *Journal of Geophysical Research* 104: 22,163–22,170.
- Mason EA, Munn RJ, and Smith FJ (1966) Thermal diffusion in gases. *Advances in Atomic and Molecular Physics* 2: 33–91.
- Schwander J (1989) The transformation of snow to ice and the occlusion of gases. In: Oeschger H and Langway CC Jr. (eds.) *The Environmental Record in Glaciers and Ice Sheets*, pp. 53–67. New York: Wiley.
- Severinghaus JP and Brook EJ (1999) Abrupt climate change at the end of the last glacial period inferred from trapped air in polar ice. *Science* 286: 930–934.
- Severinghaus JP and Battle MO (2006) Fractionation of gases in polar ice during bubble close-off: New constraints from firn air Ne, Kr and Xe observations. *Earth and Planetary Science Letters* 244: 474–500.
- Severinghaus JP, Sowers T, Brook EJ, Alley RB, and Bender ML (1998) Timing of abrupt climate change at the end of the Younger Dryas interval from thermally fractionated gases in polar ice. *Nature* 391: 141–146.
- Severinghaus JP, Grachev A, and Battle M (2001) Thermal fractionation of air in polar firn by seasonal temperature gradients. *Geochemistry, Geophysics, Geosystems* 2: article #2000GC000146.

ICE CORES

Contents

History of Research, Greenland and Antarctica

Dynamics of the Greenland Ice Sheet

Dynamics of the West Antarctic Ice Sheet

Dynamics of the East Antarctic Ice Sheet

History of Carbon Monoxide and Ultra-Trace Gases from Ice Cores

History of Nitrous Oxide from Ice Cores

History of Research, Greenland and Antarctica

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Introduction

To predict future changes to the Earth's climate, we need to understand the processes that control climate. Likewise, to understand how changes in the size of the Earth's ice sheets will influence sea level, we need to understand the processes that control the characteristics of the ice sheets. One way to improve our understanding of how these processes will operate in the future is to study how they operated in the past. To do this we need information about how and why climate and ice sheets have changed over time. Ice sheets and large glaciers contain an archive of ancient ice, dust, and trapped atmospheric gases. By analyzing these inclusions in ancient ice, we can determine how and why climate and ice sheets have changed in the past.

As snow falls on to the surface of an ice sheet, it compacts the snow below it, forming ice. Over millions of years the snow has built up into sheets over 4 km thick. The ice within the ice sheets flows both vertically and horizontally in response to gravity. In some locations, seasonal variations in the chemical and physical properties of the ice are preserved, allowing individual years to be identified as far back as 100 000 years ago. Ancient ice is most commonly collected by drilling into the ice and recovering ice cores, which are collected in metal cylinders from 3 to 30 cm in diameter ([Figure 1](#)). In rare locations, ancient ice is pushed to the surface with minimal distortion of the stratigraphy such that it can be directly sampled by trenching along the surface. The analysis and interpretation of such ice is done in the same manner as ice collected by coring.

Among the many contributions of ice core research to Quaternary Science, two stand out as having had the most

influence on our understanding of the climate system. The first demonstrates the tight link between the concentration of carbon dioxide in the atmosphere and surface temperature during the last 800 000 years ([Lüthi et al., 2008](#); [Petit et al., 1999](#); [Siegenthaler et al., 2005](#)). The second demonstrates that large climate change (e.g., change in surface temperature of 12 °C and 200% changes in precipitation) can occur in periods of less than a few decades ([Alley et al., 1993](#); [Siegenthaler et al., 1993](#); [Taylor et al., 1997](#)).

This article does not mention every ice-coring project; instead it focuses on efforts that were either pioneering in their approach or had a large influence on our understanding of climate ([Figure 2](#)). This article also does not include significant ice core research that has occurred outside of Greenland and Antarctica ([Thompson et al., 2003, 2005](#)). In reviewing the history of ice core science in Greenland and Antarctica, it is useful to divide the review into several periods.

Early Ice Core Projects

Prior to the early 1980s, ice core projects were not common and it was difficult to predict their outcome. The location of many of the early large ice-coring projects was selected to simplify logistics instead of optimizing the science. Despite this, the field was sufficiently new that such pioneering cores still produced significant results.

In 1954, Willi Dansgaard showed how the isotopic concentration of the oxygen entrapped in the ice could be used to determine the temperature history of the drill site ([Dansgaard, 1954](#)). In 1966, the Camp Century ice core was drilled in

northwest Greenland at an American military research site; yielding a low time resolution temperature record extending far enough into the past to show that the Greenland ice sheet did not completely melt during the last interglacial period (Dansgaard et al., 1969). Low time resolution, uncertainties about the dating of the core, and poor-understood ice flow limited the scientific value of the core. However, it demonstrated the potential of ice cores for climate research, thus enabling future projects to proceed.

The first deep Antarctic ice core was recovered at the American Byrd Station in 1968, which showed strong similarities with Northern Hemisphere records (Epstein et al., 1970). This core was also used in conjunction with later Greenland cores to investigate the relationship between climate changes in the Northern and Southern Hemisphere (Blunier et al., 1998).

In 1981, in a collaboration between Danish, Swiss, and American researchers, the Greenland ice sheet project (GISP), completed recovery of an ice core at Dye-3 in Southern Greenland (Langway and Oeschger, 1985). This well-dated record covered the last 90 000 years, and had sufficient time resolution to show warm interstadial events (referred to as Dansgaard/Oeschger events) of several thousand years' duration during the glacial period and abrupt changes in climate of 10–12 °C in less than a century (Dansgaard et al., 1989).



Figure 1 An ice core from a depth of 300 m from Siple Dome, Antarctica.

Concerns about the influence of ice flow and the lack of a replicate record reduced the widespread acceptance that this research deserved.

Recent Ice Core Projects

The previous ice cores led to wide acceptance of using ice core records to study climate phenomena. This enabled subsequent projects to obtain the resources needed to conduct field operations at sites selected to optimize the scientific value of the core, instead of at sites that were selected in part for their favorable logistics.

The evidence of rapid climate changes generated sufficient interest that two related Greenland ice core projects (GRIPs) were conducted to further investigate the Dye-3 results. After careful consideration, the optimal location for these projects was determined. The GRIP (involving eight European nations) and Greenland ice sheet project 2 (GISP2, involving 19 US institutions), completed field operations in central Greenland in 1992 and 1993, respectively (Hammer et al., 1998). Both of these projects used unique electromechanical drills that were raised and lowered with a wire line in an uncased hole. A petroleum-based drilling fluid was used to equilibrate the hydrostatic and pressure of the hole to prevent its collapse. The length of core section recovered on each drill run varied from 1 to 6 m. Designs based on the Danish ISTUK drill were developed for subsequent projects by European and other nations. The heavier US drill, which recovered a larger diameter core, was used on subsequent US projects.

The major outcome of the GRIP and GISP2 projects was the unquestionable confirmation of the Dansgaard/Oeschger events, and that many abrupt climate changes occurred during the glacial period, with temperature changes of up to 12 °C within a few decades (Siegenthaler et al., 1993) and factor of two changes in the ice accumulation rates (Alley et al., 1993). The Dansgaard/Oeschger events recorded in the ice cores were easily correlated to lower time resolution records of the same events in ocean cores (Bond et al., 1993). This greatly strengthened the concept that rapid reorganization of ocean circulation in the North Atlantic was a driving force of climate change.

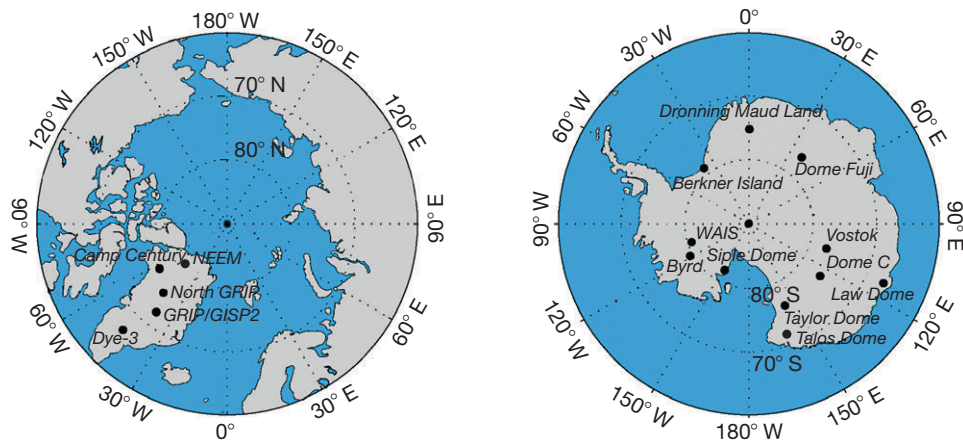


Figure 2 Location of ice cores discussed in text.

In parallel with these Greenland projects, a continuing Russian effort of the Vostok coring site in Antarctic, with assistance in later years by France and the United States, extended the record of atmospheric greenhouse gases to 420 000 years. This record showed a tight link between changes in the atmospheric concentration of greenhouse gases and surface temperatures, and that the current concentrations and rate of increase of greenhouse gases is greater than at any time during the last 420 000 years (Petit et al., 1999). In 2002, the Japanese Dome Fuji core recovered a record similar to Vostok, confirming that the major characteristics of climate change for at least the last 340 000 years are similar throughout the East Antarctic Plateau (Watanabe et al., 2003). This project ultimately obtained a record that is about 720 000 years old (Motoyama, 2007).

Recovery of the Danish-lead North GRIP (NGRIP) was completed in 2003. This core contained discernable annual layers all the way to the bedrock, covering the last 100 000 years. The record shows that the last glacial period ended gradually over a period of 7000 years (North Greenland Ice Core Project Members, 2004).

Many intermediate-age cores have also been drilled in Antarctica. The Austrian-lead Law Dome project provided an exceptional record of atmospheric carbon dioxide during the last 2000 years (Etheridge et al., 1996; MacFarling Meure et al., 2006). The US-lead projects at Taylor Dome (Steig et al., 1998) and Siple Dome (Taylor et al., 2004) both yielded records extending to the beginning of the last glacial period, indicating large regional differences. However, ice flow makes their interpretation controversial and complete analysis of these records will require more ice cores. The British-lead Berkner Island project, in a region where climate is strongly influenced by the Atlantic Ocean, has been drilled and the initial results are available (Ruth et al., 2004). The International Trans-Antarctic Expedition has drilled an array of ice cores focusing on the last 200 years, with some cores extending to 2000 years BP. These short cores provide insights into the spatial patterns of climate change (Genthon et al., 2005).

Recovery of a core from Dome C in Antarctica was completed in 2005 (EPICA Community Members, 2004) by the European projects for ice coring in Antarctica (EPICA) consortium. Results from the upper portion of the core, spanning 740 000 years, show that the interglacial periods between 740 000 and 430 000 were cooler than more recent interglacials, and that a greater portion of each glacial cycle was spent in the warm mode. Marine Isotope Stage 11, which is often suggested as an analog to the current interglacial, persisted for 28 000 years, suggesting that current climate conditions will continue for an extended time unless altered by significant human activities. The core also showed that the strong relationship between greenhouse gases and Antarctic climate extends back through at least the last 800 000 years (Lüthi et al., 2008; Siegenthaler et al., 2005).

Recovery of a core from Dronning Maud Land, Antarctica, was completed in 2006 by the EPICA consortium, which has a well-dated chronology for the last 150 000 years (Ruth et al., 2007). This record will be exceptionally well suited for comparison with Greenland records, because the climate of both were strongly influenced by the Atlantic Ocean (EPICA Community Members, 2006). The Talos Dome deep ice core in East

Antarctica was drilled during the same period (2004–2007). It provided a precise, high-resolution chronology for the last 50 000 years (Buiron et al., 2011) and provides data through the last interglacial period. These new ice core records allow comparisons with other ice core records from Antarctica and Greenland at submillennial time scales (Masson-Delmotte et al., 2011; Stenni et al., 2011).

The US-led West Antarctic ice sheet (WAIS) divide ice core started field operations in 2005 and is expected to recover a record extending back about 60 000 years. This core will probably be the best Antarctic core for comparison with the Greenland cores, in determining the temporal relationship between Arctic and Antarctic climate changes. It will also provide an exceptionally high-resolution record of greenhouse gases extending into the last glaciation. A new drill that brings together many proven features from previous drills, and adds some new ones, was developed at the University of Wisconsin, Madison. This drill has separate down-hole motors for the drill cutters and a high-volume pump that collects the drill cuttings, high borehole travel speeds, and improved down-hole sensors and automatic control systems. The project will continue for several more years although the main coring was completed in 2011.

The Future of Polar Ice Core Research

In 2004, scientists from all the nations involved in major ice-coring programs, met to form the International Partners in Ice Core Sciences (IPICS), in order to develop plans for future international ice-coring efforts. Four projects had wide international support and are expected to proceed.

For the last 1 My, the climate has alternated between glacial and warmer conditions, approximately every 100 000 years. Prior to 1 Ma these shifts occurred approximately every 40 000 years. The cause of this change is probably related to changes in the carbon cycle. To improve our understanding of how the carbon cycle controls climate, we need to have a record of greenhouse gases and climate that extends at least 1.2 My into the past. This record can only be obtained from an Antarctic ice core drilled at a carefully selected location. It is envisioned that at least two cores will be drilled at locations that are likely to yield the longest possible ice core climate records from Antarctica. Preliminary site selection work indicates these sites are in East Antarctica and site selection activities started in 2007.

The last interglacial period ended 122 000 years ago. To understand how the climate might change at the end of the current interglacial, we need to have a better understanding of what happened when the last interglacial ended. Unfortunately, the existing Greenland ice cores do not contain a complete, high-quality record of this transition. The best Northern Hemisphere record of climate change can be developed from an ice core collected in northwest Greenland. Drilling for the North Greenland Eemian ice drilling (NEEM) project started in 2007 and reached bedrock in 2010. The first results from the NEEM project are already available (Steen-Larsen et al., 2011). Unlike the NGRIP core, this core may provide a complete, high-resolution Northern Hemisphere record of the last interglacial.

To understand the climatic teleconnections between different regions, we need a global array of climate records. One array will investigate the spatial response of climate to the changes that occurred from the end of the last glacial period to the present. This array will consist of existing cores and new ones. It is anticipated that several new cores, mostly in Antarctica, will be recovered to fill in this array. A new US deep-drilling project is ongoing at the West Antarctic ice sheet divide (WAIS-D) and the borehole reached 3330 m depth in 2011 after four drilling seasons. This WAIS-D deep ice core is expected to yield an annual-layer-counted chronology for the last 40 000 years and an undisturbed depth–age relationship that likely extends back about 60 000 years (Laird et al., 2010). Shallow ice cores drilled as part of the WAIS-D project have already yielded high-resolution records of atmospheric changes during the late Holocene (Kunasek et al., 2010; Lamarque et al., 2011; Mitchell et al., 2011).

To place the current climate change problem in perspective, and to understand the phenomena that influence today's climate, we need a better understanding of the spatial interaction of recent climate phenomena on timescales of decades and centuries. This can be accomplished by a global array of ice cores that span the last several millennia with annual or near-annual resolution. This array will consist of many existing cores, and it is anticipated that many new cores will be collected in polar and nonpolar regions to complete this array.

See also: **Ice Core Methods:** Borehole Temperature Records; Chronologies; Overview; Stable Isotopes. **Ice Core Records:** Africa; Antarctic Stable Isotopes; Chinese, Tibetan Mountains; Correlations Between Greenland and Antarctica; Greenland Stable Isotopes; Ice Margin Sites; South America. **Ice Cores:** Dynamics of the East Antarctic Ice Sheet; Dynamics of the Greenland Ice Sheet; Dynamics of the West Antarctic Ice Sheet.

References

- Alley RB, Meese DA, Shuman CA, et al. (1993) Abrupt increase in Greenland snow accumulation at the end of Younger Dryas event. *Nature* 362: 527–529.
- Blunier T, Chappellaz J, Schwander J, Dallenbach A, and Stauffer B (1998) Asynchrony of Antarctic and Greenland climate change during the last glacial period. *Nature* 394: 739–743.
- Bond GC, Broecker W, Johnsen S, McManus J, and Labeyrie L (1993) Correlations between climate records from North Atlantic sediments and Greenland ice. *Nature* 365: 143–147.
- Buiron D, Chappellaz J, Stenni B, et al. (2011) TALDICE-1 age scale of the Talos Dome deep ice core, East Antarctica. *Climate of the Past* 7(1): 1–16.
- Dansgaard W (1954) The O¹⁸-abundance in fresh water. *Geochimica et Cosmochimica Acta* 6: 241–260.
- Dansgaard W, Johnsen SJ, Møller J, Oersted HC, and Langway CC Jr. (1969) One thousand centuries of climatic record from Camp Century on the Greenland Ice Sheet. *Science* 166: 377–381.
- Dansgaard W, White J, and Johnsen S (1989) The abrupt termination of the Younger Dryas climate event. *Nature* 339: 532–534.
- EPICA Community Members (2004) Eight glacial cycles from an Antarctic ice core. *Nature* 429: 623–628.
- EPICA Community Members (2006) One-to-one coupling of glacial climate variability in Greenland and Antarctica. *Nature* 444: 195–198.
- Epstein S, Sharp RP, and Gow AJ (1970) Antarctic ice sheet: Stable isotope analyses of Byrd station cores and interhemispheric climatic implications. *Science* 168: 1570–1572.
- Etheridge DM, Steele LP, Langenfelds RL, Francey RJ, Barnola J-M, and Morgan VI (1996) Natural and anthropogenic changes in atmospheric CO₂ over the last 1000 years from air in Antarctic ice and firn. *Journal of Geophysical Research* 101(D2): 4115–4128.
- Genthon C, Kaspari S, and Mayewski PA (2005) Interannual variability of the surface mass balance of West Antarctica from ITASE cores and ERA40 re-analyses, 1958–2000. *Climate Dynamics* 24: 759–770.
- Hammer C, Mayewski P, Peel D, and Stuiver M (eds.) (1998) *Journal of Geophysical Research* 2(12): 26315–26886 (ed. Greenland Summit Ice Cores).
- Kunasek SA, Alexander B, Steig EJ, et al. (2010) Sulfate sources and oxidation chemistry over the past 230 years from sulfur and oxygen isotopes of sulfate in a West Antarctic ice core. *Journal of Geophysical Research* 115: D18313.
- Laird CM, Blake WA, Matsuoka K, et al. (2010) Deep ice stratigraphy and basal conditions in central West Antarctica revealed by coherent radar. *IEEE Geoscience and Remote Sensing Letters* 7: 246–250.
- Lamarque JF, McConnell JR, Shindell DT, Orlando JJ, and Tyndall GS (2011) Understanding the drivers for the 20th century change of hydrogen peroxide in Antarctic ice cores. *Geophysical Research Letters* 38: L04810.
- Langway CC, Oeschger H, and Dansgaard W (1985) *Greenland Ice Core: Geophysics, Geochemistry, and the Environment*. *Geophysical Monograph*, vol. 33, pp. 1–118. Washington, DC: AGU.
- Lüthi D, Le Floch M, Bereiter B, et al. (2008) High-resolution carbon dioxide concentration record 650,000–800,000 years before present. *Nature* 453: 379–382.
- MacFarling Meure C, Etheridge D, Trudinger C, et al. (2006) Law Dome CO₂, CH₄, and N₂O ice core records extended to 2,000 years BP. *Geophysical Research Letters* 33: L14810. <http://dx.doi.org/10.1029/2006GL026152>.
- Masson-Delmotte V, Buiron D, Ekaykin A, et al. (2011) A comparison of the present and last interglacial periods in six Antarctic ice cores. *Climate of the Past* 7: 397–423.
- Mitchell LE, Brook EJ, Sowers T, McConnell JR, and Taylor K (2011) Multidecadal variability of atmospheric methane, 1000–1800 C.E.. *Journal of Geophysical Research* 116(6): G02007.
- Motoyama H (2007) The second deep ice coring project at Dome Fuji, Antarctica. Scientific Drilling No. 5, pp. 41–43. <http://dx.doi.org/10.2204/iodp.sd.5.05.2007>.
- North Greenland Ice Core Project Members (2004) High-resolution record of northern hemisphere climate extending into the last interglacial period. *Nature* 431: 147–151.
- Petit JR, Jouzel J, Raymond D, Barkov NI, and Barnola JM (1999) Climate and atmospheric history of the past 420,000 years from the Vostok ice core, Antarctica. *Nature* 399: 429–436.
- Ruth U, Barnola J-M, Beer J, et al. (2007) 'EDML1': A chronology for the EPICA deep ice core from Dronning Maud Land, Antarctica, over the last 150 000 years. *Climate of the Past* 3: 475–484.
- Ruth U, Wagenbach D, Mulvaney R, et al. (2004) Comprehensive 1000 year climatic history from an intermediate-depth ice core from the south dome of Berkner Island, Antarctica: Methods, dating and first results. *Annals of Glaciology* 39: 146–154.
- Siegenthaler U, Stocker T, Monnin E, et al. (1993) Abrupt increase in Greenland snow accumulation at the end of the Younger Dryas event. *Nature* 362: 527–529.
- Siegenthaler U, Stocker TF, Monnin E, et al. (2005) Stable carbon cycle–climate relationship during the last Pleistocene. *Science* 310: 1313–1317.
- Steen-Larsen HC, Masson-Delmotte V, Sjolte J, et al. (2011) Understanding the climatic signal in the water stable isotope records from the NEEM shallow firn/ice cores in northwest Greenland. *Journal of Geophysical Research* 116: D06108.
- Steig EJ, Brook EJ, White JWC, et al. (1998) Synchronous climate changes in Antarctica and the North Atlantic. *Science* 282: 92–95.
- Stenni B, Buiron D, Frezzotti M, et al. (2011) Expression of the bipolar sea-saw in Antarctic climate records during the last deglaciation. *Nature Geoscience* 4: 46–49.
- Taylor KC, Mayewski PA, Alley RB, et al. (1997) The Holocene/Younger Dryas transition recorded at Summit Greenland. *Science* 278: 825–827.
- Taylor KC, White JWC, Severinghaus JP, et al. (2004) Abrupt late glacial climate change in the Pacific sector of Antarctica. *Quaternary Science Reviews* 23: 7–17.
- Thompson LG, Davis ME, Mosley-Thompson E, Lin P-N, Henderson KA, and Mashiotta TA (2005) Tropical ice core records: Evidence for asynchronous glaciation on Milankovitch timescales. *Journal of Quaternary Science* 20: 723–733.
- Thompson LG, Mosley-Thompson E, Davis ME, Lin PN, Henderson K, and Mashiotta TA (2003) Tropical glacier and ice core evidence of climate change on annual to millennial time scales. *Climatic Change* 59(1–2): 137–155.
- Watanabe O, Jouzel J, Johnsen S, Parrenin F, and Shoji H (2003) Homogeneous climate variability across East Antarctica over the past three glacial cycles. *Nature* 422: 509–512.

Dynamics of the Greenland Ice Sheet

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Introduction

The Greenland ice sheet is the largest ice sheet of the Northern Hemisphere (Figure 1). It covers about 80% of Greenland and has an area of $1.71 \times 10^6 \text{ km}^2$ and an average thickness of about 1.6 km. It covers a latitude span from 84° N near the northern margin in Peary Land, to 60° N at the southern tip: a length of about 2700 km. In Central Greenland at 72° N , its width from west to east is about 900 km. It contains a total volume of $2.95 \times 10^6 \text{ km}^3$ ice (Thomas and PARCA Investigators, 2001), enough to raise Earth's sea level by 6–7 m if it melted away completely. The Greenland ice sheet is the second largest ice sheet on Earth, exceeded only by the Antarctic ice sheet which has a volume that is about ten times larger than that of the Greenland ice sheet.

The Greenland ice sheet plays an important role in climate research. The ice sheet interacts dynamically with the atmosphere and ocean circulation in the North Atlantic, a region thought to play a key role in Earth's climate system. The Greenland ice sheet contains a unique record of past climate changes, and ice more than 100 000 years old has been recovered from deep ice cores drilled through the Greenland ice sheet (e.g., NorthGRIP Members, 2004). Because snow accumulation rates in the central areas have been sufficiently high, annual layers can be resolved far back into the last glacial period (Svensson et al., 2005), providing high-resolution climate records.

The Greenland ice sheet continues to evolve in response to changes in Earth's climate. Snow deposited in the interior is gradually transformed to ice, and layers of ice are thinned and stretched by ice sheet flow. Ice flows from the interior toward the margin where it is lost by surface melting or by discharge of ice into the ocean. Current observations of retreat along the margins and acceleration of outlet glaciers suggest that the ice sheet is currently reacting to a changing climate. Understanding the dynamics of the Greenland ice sheet is crucial for interpretation of the climate records obtained from ice cores, and for understanding the dynamical response of the ice sheet to climate change.

Surface and Basal Topography

The Greenland ice sheet is elongated in the north–south direction and configured with a central ridge running from south to north to about 77° N , connecting the two domes in South Greenland and in Central Greenland. The highest dome, Summit in Central Greenland, has an elevation of approximately 3200 m (near the deep drilling locations GRIP (Greenland Ice Core Project) and GISP2 (Greenland Ice Sheet Program 2)), and the other dome in South Greenland has an elevation of 2850 m (Bamber et al., 2001a; Figure 1). The saddle between the domes forms a distinct ridge with elevations of around 2500 m. North of Summit, the ridge widens as it continues

northwest. Near 77° N , it forks into a northern and western branch. About two-thirds of the Greenland ice sheet has surface elevations above 2000 m.

Overall, the bedrock is bowl shaped, with low central areas near sea level and coastal mountains with elevations of 1–3 km above sea level almost entirely surrounding it. The bedrock is mountainous in South Greenland and particularly in East Greenland where coastal mountains reach above 3 km above sea level. In West Greenland, the coastal mountains are lower than in the East. In South Greenland, bedrock elevations in the central areas are generally above 200 m above sea level. North of Summit, the bedrock in the central areas is generally smooth and the ice sheet is grounded near or below sea level, with widespread areas below sea level (Bamber et al., 2001b).

The ice thickness reaches a maximum of 3000–3300 m near the central north–south ridge between 70° and 75° N (Bamber et al., 2001b). The loading of ice has depressed the bedrock through isostatic adjustment of the lithosphere. If the ice were removed, and the bedrock allowed to rebound isostatically, the elevation of the central basin would be 600–800 m above sea level (Bamber et al., 2001b). The ice sheet fills the low central areas completely and flows toward the coast. The coastal mountains are cut by deep fjords where outlet glaciers reach the ocean and icebergs break off from floating glacier tongues. In North-West Greenland, the ice sheet extends to the ocean over a longer distance.

Surface and Basal Temperature

Over most of the ice sheet, the mean annual surface temperature is far below the freezing point. The high-elevation central areas have a mean annual temperature below -25° C , and surface melting occurs only rarely (dry-snow zone). At Summit, the mean annual temperature is -32° C . July is the warmest month, with a mean temperature of -12° C , and February is the coldest month, with a mean temperature of -42° C . In South Greenland, surface temperatures near the north–south ridge are -10 to -20° C . At Dye3, the mean annual temperature is -19° C , and surface melting occurs often during summer. At Dye3, meltwater percolates down into the firn and refreezes entirely to form ice layers in the firn (percolation zone). At lower elevations along the margin, surface meltwater forms drainage systems, and is eventually lost by runoff through surface streams and subglacial drainage systems (e.g., Cuffey and Paterson, 2010).

The ice temperature below the surface is determined by the downward advection of cold ice that is gradually heated by several processes. The geothermal heat flux is heating up the ice from beneath. The near surface firn layers may be heated by surface meltwater that percolates down into the firn and refreezes. The basal ice layers are heated by internal deformation of the ice and by frictional heating at the base if there is

basal sliding. The basal temperature near the north–south ridge is controlled by ice thickness, geothermal heat flux, and past surface temperature and accumulation that have determined the downward advection of cold ice due to flow through time. Under Summit, central Greenland, the basal temperature was measured to be -9.5°C in the GRIP borehole. The basal temperature is still affected by the cold glacial climate, and if the current climate had prevailed it would have been several degrees higher, maybe reached the melting point. The effect is complicated, as the glacial temperatures were colder, but at the same time accumulation rates were lower, thereby decreasing the downward advection. The ice free surface of Greenland consists mostly of old rocks with a low geothermal heat flux of $30\text{--}50\text{ mW m}^{-2}$ (e.g., Van der Veen, 1999). Using this geothermal heat flow, estimated basal temperatures are -5 to -10°C in the central areas. In Northern Greenland, widespread areas with basal melting have been identified by analyses of radar data (see section ‘Internal Stratigraphy’ below). These areas have a basal temperature at the pressure melting

point of -2 to -3°C (beneath 3 km of ice) (Dahl-Jensen et al., 2003), and the basal melting is due to a higher geothermal heat flux in these areas. In the NorthGRIP deep drilling borehole at the north-west ridge in North Greenland, the basal temperature was at the pressure melting point of -2.4°C , and the borehole drilling reached water-saturated sediments at the base beneath 3185 m of ice. Basal melting has been estimated to be 7 mm a^{-1} , and the geothermal heat flux has been estimated to be 130 mW m^{-2} (NorthGRIP Members, 2004).

Surface Mass Balance

A compilation of the accumulation rate for 1971–90 was presented by Bales et al. (2009; Figure 2). The average annual

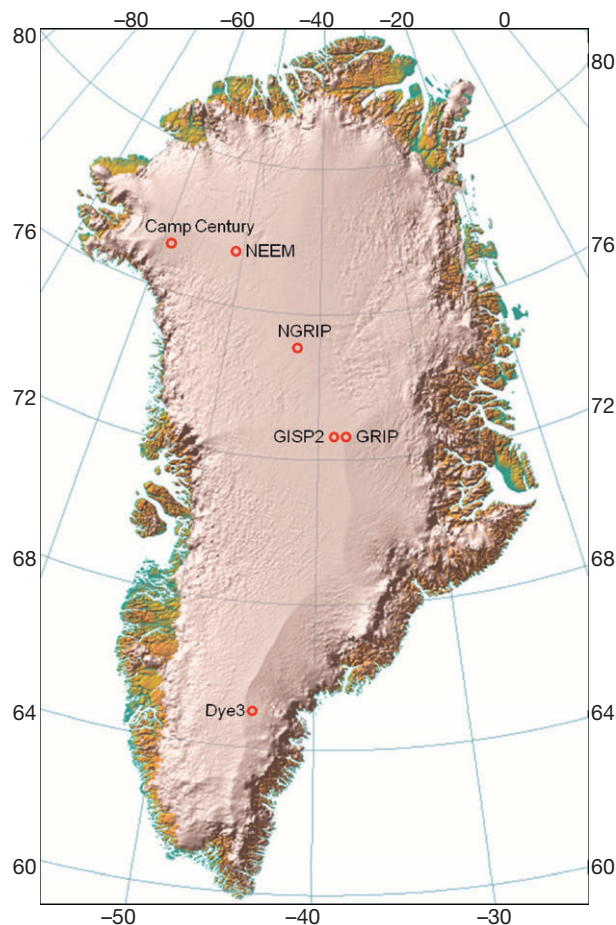


Figure 1 Shaded relief plot of Greenland surface topography (Bamber et al., 2001a). White areas are ice-covered. Deep drilling sites are indicated, where ice cores have been drilled through the Greenland ice sheet to the bedrock. These sites are: Camp Century, Dye3, GISP2, GRIP, NorthGRIP, NEEM (North Greenland Eemian Ice Drilling) (for a historical overview, see Dansgaard, 2004; about NEEM, see Buchard and Dahl-Jensen, 2008). Summit, the highest dome on the Greenland ice sheet, is near the GISP2 and GRIP sites.

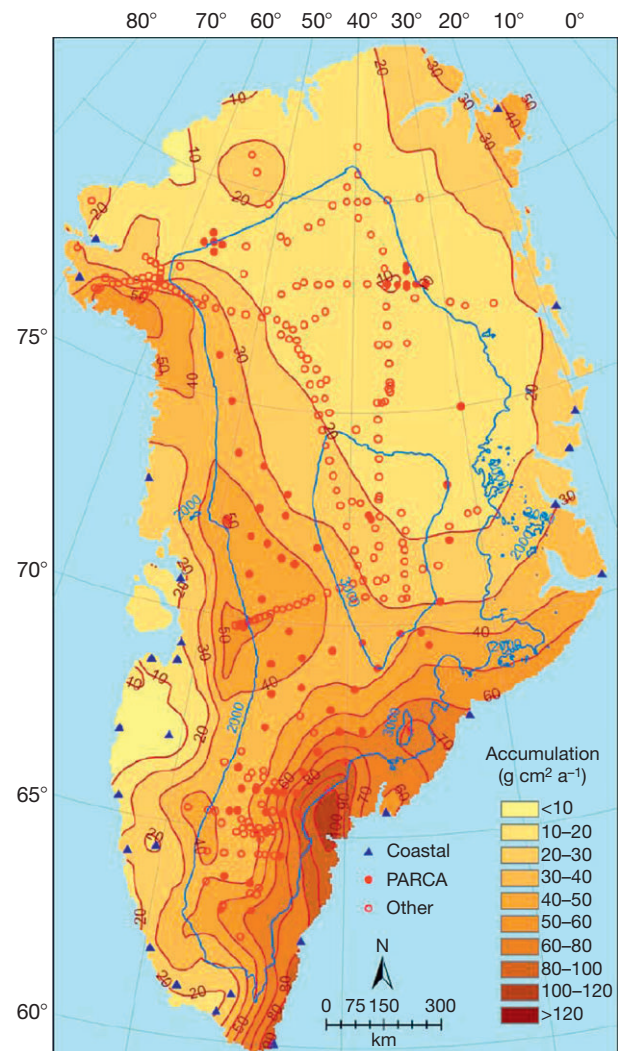


Figure 2 Accumulation map of Greenland. Reproduced from Bales RC, Shen D, Du G, et al. (2009) Annual accumulation for Greenland updated using ice core data developed during 2000–2006 and analysis of daily coastal meteorological data. *Journal of Geophysical Research, Atmospheres* 114: D06116, <http://dx.doi.org/10.1029/2008JD011208>. Contour and shading show accumulation rate in mm of water equivalent a^{-1} . Also shown for reference are the 1000, 2000, and 3000 m elevation contours.

accumulation rate is $0.3 \text{ m of water equivalent a}^{-1}$. The most important trend in accumulation rate is the south to north decrease with variations ranging from almost $1 \text{ m of water equivalent a}^{-1}$ in the south to below $0.1 \text{ m water equivalent a}^{-1}$ in the northeast. The north-south ridge forms a topographic barrier with significant effect on the accumulation rate pattern. In the south, weather systems traveling along the North-Atlantic storm track carry moisture and precipitation to the Greenland ice sheet. North of Summit, the accumulation rate gradually decreases with elevation on the western flank, but drops rather abruptly with a factor of two across the ridge.

Surface melting is widespread and occurs over about 50% of the Greenland ice sheet (Thomas and PARCA Investigators, 2001), but only a few measurements are available over both time and space. In addition to direct observations of surface melt and runoff, passive microwave data observed from satellite have been used to infer melt extent and duration. Ablation by melting and runoff has also been modeled by the positive degree day (PDD) approach or by energy balance models of the upper firn layer. PDD methods have been widely used to estimate surface melting, but it is based on an empirical relationship between surface temperature and melt that is based on present-day observations. Energy balance models consider the key physical processes, but require a large set of meteorological data as forcing (e.g., Van der Veen, 1999).

The line that separates the net accumulation zone from the net ablation zone is called the equilibrium line. In Greenland, the total accumulation of snow above the equilibrium line is approximately balanced by surface melting in the ablation zone and discharge of ice into the ocean. At the end of the summer, when fresh snow from the previous year has been removed by surface melting in the ablation zone, the surface of the ice sheet will consist of progressively younger ice inward from the margin toward the center. The oldest surface ice is found at the margin, and inland of the equilibrium line, the surface consists of snow from the previous year.

Discharge of Ice into the Ocean

Discharge of ice into the ocean constitutes an equally important contribution to the mass loss of the Greenland ice sheet as surface melting (van den Broeke et al., 2009). The discharge consists partly of calving of icebergs, and partly of melting from beneath floating ice margins. Near the margin, ice flow rates vary significantly spatially. The ice sheet drains through many fast-flowing outlet glaciers and ice streams in valleys in the coastal mountains, and these fast-flowing areas are separated by areas of slow flow (Figure 3; Joughin et al., 2010). Several glaciers drain into fiords, and produce icebergs from floating glacier tongues. The ocean terminating outlet glaciers produce the highest discharge rates of the Greenland ice sheet; they are all directed by descending bedrock elevations, and their mouths are associated with deep low channels carved into the bedrock (Bamber et al., 2001b). Ice streams form upstream of the ice fiords, extending far inland and with significantly increased surface velocities. The most important ocean terminating outlet glacier is the Jakobshavn Isbrae that drains into the Jakobshavn ice fiord in the west with a maximum surface velocity of about 12.6 km a^{-1} at the terminus (Joughin et al.,

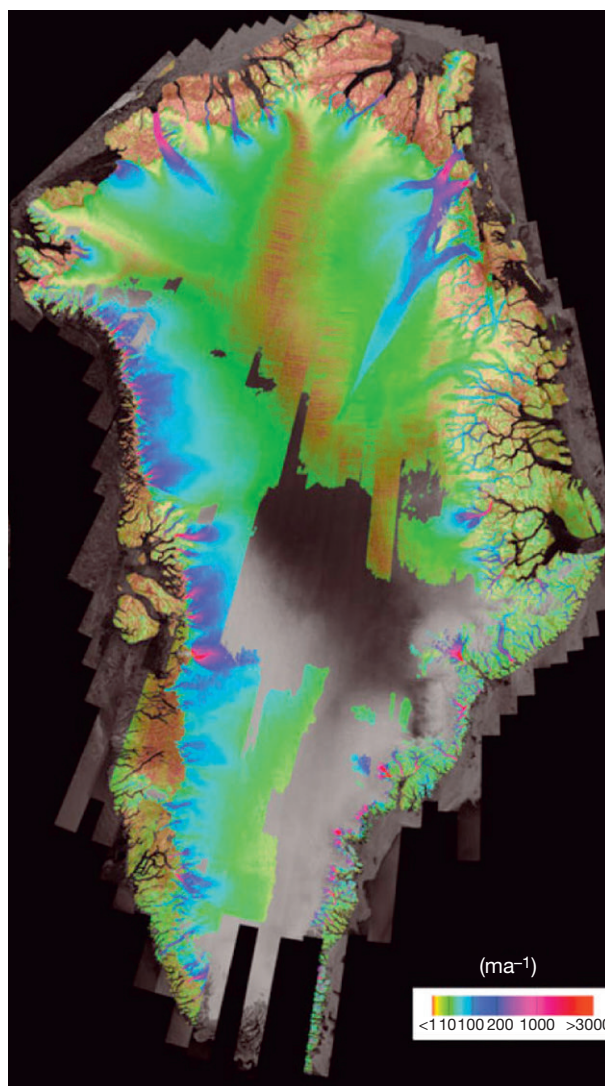


Figure 3 Map of surface flow speed for winter 2005/2006 displayed over radar mosaics (©Canadian Space Agency, 2001). Reproduced from Joughin I, Smith BE, Howat IM, Scambos T, Moon T (2010) Greenland flow variability from ice-sheet-wide velocity mapping. *Journal of Glaciology* 56(197): 415–430.

2004). Another example is the north-east ice stream that drains into the 79-Fjorden in the north-east. The ice stream extends hundreds of km inland and originates near an area with high basal melt rates near the central ice divide north-west of Summit (Fahnestock et al., 2001).

Ice Deformation and Large-Scale Flow

Glacier ice flows due to gravity. In a grounded ice sheet the flow is driven by the surface slope, and glacier ice is a slowly flowing viscous material.

Large-Scale Flow Pattern

The general flow pattern of an idealized, steady-state, cross-section of the Greenland ice sheet from west to east is

illustrated in [Figure 4](#). The ice sheet accumulates snow in the high-elevation central areas (the accumulation zone), and it loses mass in the low-elevation marginal areas (the ablation zone). If the ice sheet is in steady state, the spreading of the ice sheet due to ice flow is balanced by the surface mass balance, and the shape remains constant in time. A snow particle, originally deposited at the surface in the central part, will gradually sink down and be carried with the flow to the margin. The further inland the ice is formed, the deeper it will sink before it eventually reaches the marginal areas after thousands of years and emerges at the surface, where it will be lost by surface melt or by calving of icebergs. Ice loss may also occur as basal melting and draining through subglacial tunnels, or from melting beneath floating glacier tongues. [Figure 4](#) also illustrates how an annual layer deposited near the ice divide (the central north–south running ridge that separates east flowing ice from west flowing ice) is thinned and stretched by flow as it sinks down. Ice cores are often drilled close to the ice divide, where horizontal flow is minimal, and no upstream corrections are needed.

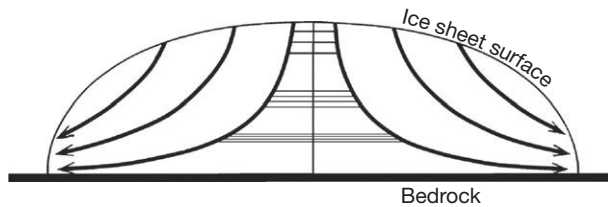


Figure 4 General flow pattern of an idealized, two-dimensional ice sheet. The arrows indicate particle paths. The vertical line in the center is the so-called ice divide, because it separates ice flowing to the left from ice flowing to the right (similar to watershed). The horizontal lines in the center indicate annual layers (their thickness is greatly exaggerated), and illustrates how annual layers are thinned and stretched with depth. The optimal site for ice core drilling is near the ice divide, where ice movement is vertical, and no up-stream corrections are needed.

Transformation of Snow to Ice

As snow is deposited in the central areas of an ice sheet and buried by subsequent layers of snow, it is gradually transformed into glacier ice. The term *firn* is usually used to describe the material in the intermediate stages of transformation. How snow transforms into ice, and the time the transformation takes, depends on the temperature ([Cuffey and Paterson, 2010](#)). In the interior of the Greenland ice sheet, there is no melting, and the firn–ice transition depth is 65–80 m. Below the firn–ice transition, the ice density remains practically constant at 917 kg m^{-3} . Air bubbles contain samples of the past atmosphere. Due to the densification process, the age of air contained in bubbles is younger than the age of the surrounding ice.

Ice Crystals and Deformation

Glacier ice is a polycrystalline material. Below the firn–ice transition, the ice is composed of small densely packed crystals in the mm to cm size range ([Figure 5](#)). The ice crystals evolve and interact dynamically: As ice slowly descends into the ice sheet, the crystals grow and rotate, they deform and fragment, and new crystals come into existence. The average crystal size depends exponentially on the temperature of the ice, and near the bottom of the ice sheet, where temperatures generally are high, the crystals can be up to tens of cm across. The crystal size also depends on the amount of impurities in the ice. In the glacial ice with its relatively high load of impurities, the crystals are smaller than in the cleaner interglacial ice (e.g., [Cuffey and Paterson, 2010](#)).

Each crystal is built up of stacked layers of water molecules (or rather oxygen atoms), so-called basal planes. The *c*-axis is perpendicular to the basal planes and defines the orientation of the crystal ([Cuffey and Paterson, 2010](#)). It requires much less energy to deform an ice crystal in a direction parallel to the

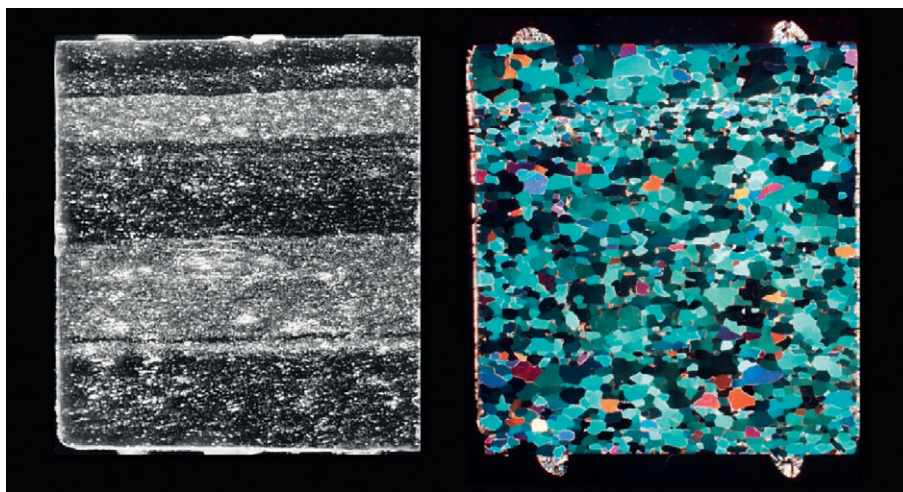


Figure 5 The left panel shows a $10 \times 10 \text{ cm}^2$ vertical sample from a depth of 2020 m in the GRIP ice core, where the ice is about 27 ka old. As in [Figure 7](#), white layers appear where the ice has a high content of impurities (and dust). The right panel shows the same sample placed between two crossed polarizers whereby individual ice crystals become visible. Each colored field in the sample is an ice crystal, and the crystals different coloring is due to their different orientations. The impurity content and the crystal size are related, such that more impurities imply smaller crystals ([Hansen et al., 2002](#)).

basal planes than in the direction perpendicular to the basal planes. Therefore, ice primarily deforms by basal slip and the orientation of an ice crystal determines the 'hardness' of the ice with respect to deformation in a given direction. Whereas the average crystal orientation in the upper, younger part of the ice sheet is close to random, the crystals will eventually rotate and orient themselves according to the main direction of compression, so that in the deeper regions of the ice sheet, the average crystal orientation exhibit a preferred orientation. Because the average crystal orientation influences the hardness of the ice, this anisotropy of the ice plays an important role for the large-scale ice flow which must be taken into account by ice flow models.

Glacier ice deforms in a way that is intermediate between Newtonian viscous and perfectly plastic behavior (Cuffey and Paterson, 2010). The deformation is described by a flow law that relates the strain rates with the applied stresses. Laboratory experiments have shown that the deformation of glacier ice is practically independent of the hydrostatic pressure, and established a flow law in which strain rates are expressed in terms of deviatoric stresses (deviation of stresses from the hydrostatic pressure). The flow law is a nonlinear power law, where the power is usually taken to be 3. The flow law rate factor depends on many factors, most notably the temperature of the ice (increasing temperatures soften the ice), but also crystal orientation (fabric) and impurity content, as mentioned above (Cuffey and Paterson, 2010).

In all Greenland ice cores, the crystals develop a strong vertical single maximum with depth, that is, all crystal *c*-axes gradually become oriented with *z*-axes pointing vertically upward. This fabric is favorable for shear, which is the most important stress component to determine the overall flow of the grounded ice sheet, and ice sheet models have been using an enhancement factor of 3 to take this anisotropy into account (e.g., Cuffey and Paterson, 2010).

Surface Velocities

The surface velocity is normally oriented in the direction of the steepest surface slope. Surface velocities are generally low in the interior, and increasing toward the margin. Surface velocity and strain rates are measured directly from repeated observations of surface stakes by global positioning system (GPS). In the interior parts where ice movement is slow, an observation time of several years is needed in order to obtain a sufficient accuracy (e.g., Hvidberg et al., 2002). This has been done during drilling projects; otherwise, data coverage of the interior parts is sparse. Surface velocity maps are also derived from radar satellite data, and satellite radar interferometry (inSAR) has become a major tool for measuring ice motion over large areas (Joughin et al., 2010; Figure 3).

Thinning of Annual Layers with Depth

At deep drilling sites, a detailed knowledge of how flow and thinning of layers varies with depth is obtained from ice core analysis and geophysical surveys of the bore hole. The depth dependence of the horizontal velocity is measured directly only in deep bore holes. When a deep ice core is being drilled, a nonfreezing liquid in the bore hole prevents it from closing

by flow. After the drilling is terminated, the liquid-filled bore hole continues to move with the ice flow while it becomes more inclined with time. Depth variations of strain rates and horizontal flow velocities may be determined from repeated surveys of bore hole inclination. Surveys of the Dye3 bore hole in South Greenland have shown that strain rates were high in the glacial ice located in the bottom 300 m (out of a total thickness of 2000 m ice), and that the glacial ice deformed about three times faster than interglacial ice (Dahl-Jensen and Gundestrup, 1987). A simple and commonly used approximation to the depth dependence of the horizontal velocity is to assume a constant velocity down to a certain depth, and then a linear decrease to zero at the base, the so-called Dansgaard-Johnsen model (e.g., Cuffey and Paterson, 2010).

Once the ice core is dated and annual layers are identified in the ice core, timescales and vertical thinning of annual layers are known (Figure 6). Typically, the annual layer thickness decreases linearly in the upper part of the ice, and then decreases to zero at the base if there is no melting. At Summit, the ice sheet is frozen at the bed; annual layers in the GRIP ice core were 0.23 m a^{-1} of ice equivalent at the surface but decreased to fractions of a mm near the base where the ice is several hundreds of thousands of years old. At NorthGRIP, the ice is melting at the bed, with basal melt rates of 7 mm a^{-1} , and annual layers decrease from 0.20 m a^{-1} of ice equivalent at the surface to a thickness of about 7 mm a^{-1} at the base, corresponding to the basal melt rate. The basal melting at NorthGRIP has removed ice older than 123 000 years. The basal melting has prevented the basal layers from further thinning and folding, and kept the layers stratigraphically undisturbed to the bottom. It is possible to identify annual layers more than 100 ka back in time, and the NorthGRIP ice core has provided ice core records with unprecedented resolution (NorthGRIP Members, 2004).

Internal Stratigraphy

The Greenland ice sheet has built up by layers of snow deposited through tens of thousands of years. Although the ice sheet is flowing, the stratigraphic structure is preserved to great depths on both large and fine scales. The large-scale stratigraphy is known from radar soundings of internal reflecting layers, while the fine scale stratigraphy is known from ice cores. This section presents some stratigraphic data sets that are relevant for understanding the dynamics of the Greenland ice sheet.

Visible Stratigraphy in Ice Cores

Inspection of deep ice cores shows that the internal layering of ice in the glacial part of the Greenland ice sheet is clearly visible to the naked eye (Figure 7). The visible layers are caused by alternating layers of ice with variable amounts of impurities, which may consist of small dust particles or low concentrations of acids or sea salt spray. In the interglacial periods the impurity content is generally too low to be visible. In recent times, the impurities are mainly deposited during spring and summer, whereas the winter and fall snow is cleaner. Assuming a similar seasonality was active during the glacial, the visible

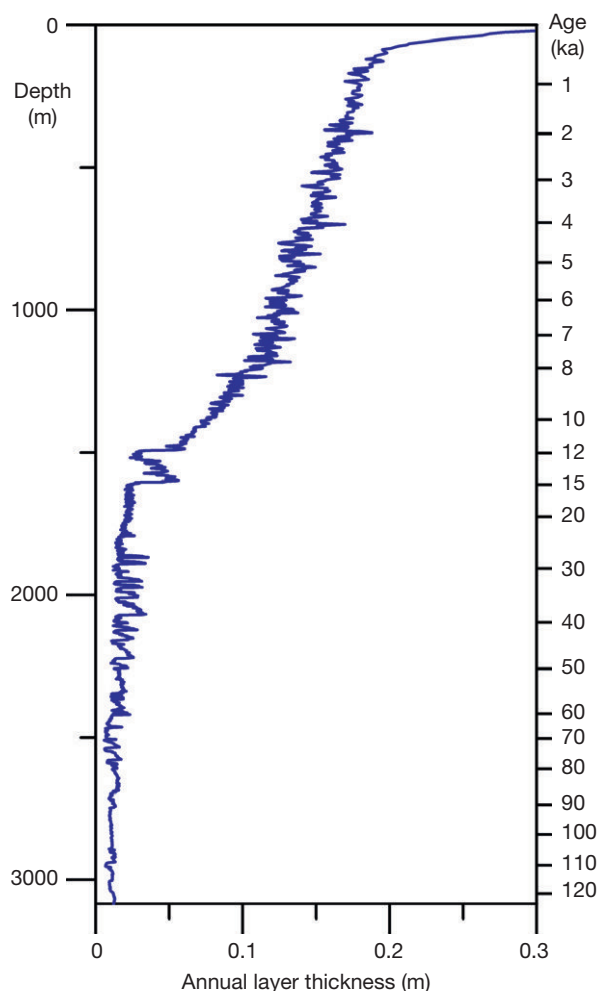


Figure 6 Thinning of annual layers with depth in the NorthGRIP deep ice core. The curve is based on annual layer counting and ice flow modeling (NorthGRIP Members, 2004). The annual layer thickness decreases with depth due to firn compression (only in the top 100 m) and due to flow, and reflects past variations in accumulation rate. The glacial–interglacial transition at 11 703 years b2k (Rasmussen et al., 2006) was found at a depth of about 1500 m. The abrupt shifts in annual layer thickness below 1500 m depth are due to changes in accumulation rate during the last glacial period. There is basal melting at NorthGRIP, and as a result, annual layer thickness does not decrease further below 2500 m depth.

layers in the glacial ice can be regarded as annual layers. Unlike tree rings, however, the ice core annual layers are not unambiguous because the typical ice sheet annual layer is composed of multiple precipitation events and thus contains several visible layers.

At great depths, the visible stratigraphy becomes less well defined as a result of significant thinning of annual layers. The layers become gradually blurred, and cm-scale folds emerge. Evidence of small-scale folds has been observed in several ice cores in the bottom few hundred meters (GRIP, GISP2, NorthGRIP) (Alley et al., 1997; Svensson et al., 2005). Folds are a result of local flow instabilities, thought to form due to alternating soft and hard ice, because of variations in impurity content.

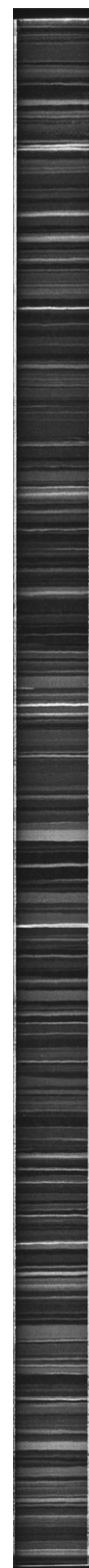


Figure 7 The internal layering in the Greenland ice sheet is clearly visible to the naked eye. This shows an example of a 1.65 m-long ice core section from 1870 m depth in the NorthGRIP ice core, where the ice is about 28 ka old. The image is obtained with indirect illumination of the ice core so that black areas correspond to clear ice, whereas white layers contain visible impurities (Svensson et al., 2005).

Radar Observations of Internal Layers

Polar ice and snow is transparent to electromagnetic waves in the very high frequency (VHF) and ultra high frequency (UHF) bands, and since the early sixties, radio-echo sounding (RES) has been used to investigate the Greenland ice sheet (historical overview given by, e.g., Gogineni et al., 1998). Several data sets have been obtained in Greenland using both airborne and ground-based radars. An example of an RES image from Greenland is shown in Figure 8.

In addition to the ice surface and bedrock, RES surveys reveal internal reflection layers that can be traced over large distances. Some of the most important sources of the reflections from the internal layers are changes in density, orientation of ice crystals, and acidity (Hammer, 1980; Harrison, 1973). Many of the layers have been identified as large volcanic horizons or abrupt climate events. The layers therefore represent former depositional surfaces, and are isochrones.

The shape of the isochrones is determined by climate (especially accumulation), the amount of basal melting, and the dynamic history of the ice sheet. If the RES layers can be tracked to a deep ice core drill site, they can be dated by the timescale established for the ice core. Thus, the internal structure of the ice sheet, as it appears on RES images, contains invaluable information about the basal conditions and ice flow, and studies of internal layers may provide time scales and other information in areas where no deep core has yet been drilled. This method was, for example, used in the selection of the NorthGRIP ice core drill site (Dahl-Jensen et al., 1997).

Basal Melting

The shape of the isochrones may reveal locations with basal melting. Basal melting drags isochrones downward, because thinning of annual layers is reduced when ice is removed at the base by basal melting and draining (Dahl-Jensen et al., 1997). Ice flow models have been used together with RES data to identify locations with basal melting and to estimate the

basal melt rates in North and Central Greenland (Dahl-Jensen et al., 2003; Fahnestock et al., 2001). It is established that basal melting is widespread under the central and northern part of the Greenland ice sheet. The basal melting constitutes a significant, but not well determined part of the total mass balance of the Greenland ice sheet. The drainage system for the generated basal melt water is not well known, but the presence of liquid water at the base may play a significant role in accelerating the rate of flow of the fast-flowing ice streams. The large northeast flowing ice stream in Northeast Greenland appears to originate at a location in Central Greenland with high basal melt rates (Fahnestock et al., 2001).

History of the Greenland Ice Sheet

About 5 Ma, Northern Hemisphere ice sheets are thought to have formed as a result of a generally cooling climate (Zachos et al., 2001). Glaciers started to form in Greenland 2–3 Ma in the mountainous areas along the east coast and gradually advanced to the central plains (Souchez, 1997). Glacial and interglacial periods have since then alternated with periods of growth and decay of Northern Hemisphere ice sheets.

During the previous interglacial, the Eemian (130–110 ka BP), interpretation of the ice core isotope records suggests a warmer climate with temperatures over Greenland about 5 °C higher than at present (NorthGRIP Members, 2004). Existence of Eemian ice in all Greenland ice cores suggests that the Greenland ice sheet survived the Eemian in the northern and central part, but its extent in the southern part during the Eemian is uncertain. All deep ice cores in Greenland contain Eemian ice, including the Dye3 ice core in South Greenland, suggesting that the Greenland ice sheet was present during the Eemian. Ice sheet modeling supports that the Greenland ice sheet may have been considerably smaller during the Eemian (Cuffey and Marshall, 2000; Otto-Bliesner et al., 2006). The estimated extent and volume of the ice sheet during the Eemian depends, however, critically on the assumed temperature and accumulation rate history which is not well constrained during the previous interglacial.

At the Last Glacial Maximum, 20 ka BP, vast areas of the Northern Hemisphere were covered by ice, with large ice sheets in North America (Laurentide ice sheet), in Greenland, and in Northern Europe (Fennoscandian ice sheet). The high quality of the climate chronology archived in the Central and North Greenland ice cores indicates that the interior elevation of Greenland ice sheet and the overall flow pattern have generally been stable during the last glacial. Model simulations support the theory that a high-elevation dome was present during the glacial–interglacial cycle, and also that the divide position may have shifted with up to 150 km near Summit (Marshall and Cuffey, 2000). The suggested shifts in divide position have implied variable shear or nonshear conditions at Summit, and they could be responsible for the observed disagreements in the ice core records in the bottom 200 m of the two Summit cores, GRIP and GISP2 (NorthGRIP Members, 2004) due to folding of near-basal layers.

The transition from the last glacial to the present interglacial occurred 11 703 years b2k (Rasmussen et al., 2006). The effect on the Greenland ice sheet was complex and included a

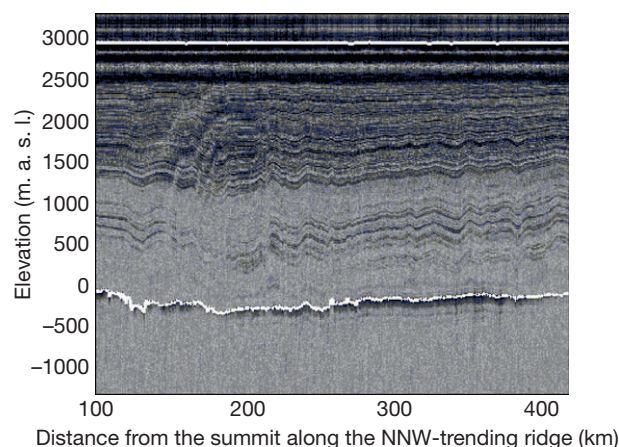


Figure 8 RES image collected by the University of Kansas along the NNW ice ridge in Northern Greenland (Gogineni et al., 2001). The surface and the bed are indicated by the thick white lines. The internal layers are isochrones. The region of highly disturbed layers about 200 km from the summit is coincident with the suggested origin of the large northeast flowing ice stream (Fahnestock et al., 2001).

number of processes with opposite effects and with different response times up to tens of thousands of years. Increased surface-melt rates and runoff in the margin areas made the ice sheet margins retreat. At the glacial–interglacial transition, accumulation rates doubled in the central areas, and led to a build-up in the interior that has gradually spread by flow to the margin. Downward advection of warmer surface temperatures gradually softens the ice through the thermo-mechanical coupling (expressed by the temperature-dependent flow law). The evolution of the Greenland ice sheet during the present interglacial has been derived from ice core records (Vinther et al., 2009). The ice sheet has thinned by hundreds of meters at Dye3 and Camp Century that are sites near the margin, while the central areas have remained stable with little change in ice thickness. A model study considered transient effects from the last glacial to interglacial transition, on the present ice sheet, and concluded that the Greenland ice sheet is still responding to the climatic transition (Huybrechts, 1994).

Stability and Current Mass Balance

Today, the Greenland ice sheet is the largest remaining ice sheet in the Northern Hemisphere. Its stability during the present interglacial is thought to be due to a combination of high precipitation rates over Greenland, high surface elevation over the interior parts, and the surrounding coastal mountains. The proximity of the Greenland ice sheet to the Atlantic Ocean is important, because it strongly affects the climate in Greenland and provides ample supplies of precipitation. Weather systems traveling northeast across the North Atlantic carry moisture to the Greenland ice sheet. Precipitation rates in Central Greenland are of the order of tens of cm of water equivalent a^{-1} , about ten times the annual accumulation in East Antarctica (e.g., Van der Veen, 1999).

The present mass balance of the Greenland ice sheet is of particular interest because of its relation to the global sea level. The amount of water stored as ice in the Greenland ice sheet corresponds to a rise in global sea level of 6–7 m, about the same as the West Antarctic ice sheet, and the Greenland ice sheet is thought to be an important contributor to future sea-level increase.

The net mass balance is defined as the change in mass over a certain time period, and it is the result of accumulation and ablation processes. Previous estimates from before 1990 (Reeh et al., 1999) suggested that the Greenland ice sheet was close to steady state with surface accumulation of roughly 550 Gt year^{-1} , balanced equally by surface ablation and calving of ice bergs. The net mass balance was estimated by considering each term in the mass budget. However, the uncertainties were large, and data coverage was sparse.

In the last decade, several studies have reported an increasingly negative net mass balance, primarily due to a large increase of surface melt and discharge of ice, which has led to a thinning of the margin areas in Greenland (Rignot and Kanagaratnam, 2006). Repeated airborne laser altimetry surveys reported in 2000 a loss of volume of the Greenland ice sheet of about 50 Gt a^{-1} , corresponding to 0.13 mm a^{-1} , approximately 7% of the observed global sea level rise (Kravill et al., 2000). By 2003–2008, the rate of mass loss had increased to

more than 200 Gt a^{-1} , according to a compilation of satellite laser altimetry data (Sørensen et al., 2011). Some of the marginal areas are lowering with more than 1 m a^{-1} (Joughin et al., 2004) and the fast-flowing ice streams in north, west, and northeast are thinning and retreating rapidly. Over the same time period, the interior parts have been relatively stable (Johannessen et al., 2005). Gravity measurements from the GRACE satellite mission have provided the rate of mass loss from the Greenland ice sheet from 2002 to 2009 (Velicogna, 2009). During this time, the rate of mass loss has accelerated from 137 Gt a^{-1} in 2002–2003 to 286 Gt a^{-1} in 2007–2009, and the corresponding contribution to sea level rise from the Greenland ice sheet has been continuously and rapidly growing to $>0.5 \text{ mm a}^{-1}$ in 2009.

The Greenland ice sheet has been thought to respond gradually to climate changes through changes in its surface mass balance. Several new findings have challenged this view. The ice streams in north, west and northeast are associated with deep bedrock valleys, and their onset appears to be related to areas with basal melting. These glaciers drain a significant fraction of the Greenland ice sheet, but it is not yet known how changes at the margin propagate inward at the ice sheet. Recently, it was shown that near the equilibrium line in West Greenland, ice flow accelerates during periods of summer melting, and decelerates after the melting ceases (Zwally et al., 2002). This suggests a coupling between surface melting and ice flow that involves rapid migration of surface meltwater to the base and provides a mechanism for rapid large-scale dynamic response of the Greenland ice sheet to climate warming. Model simulations of the future evolution of the Greenland ice sheet and its response to a warmer climate predict increased melting and runoff in a warmer climate, but the dynamic response of the ice sheet to the changes at the margin is not yet fully understood and incorporated into ice sheet models.

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See also: **Glacial Landforms, Ice Sheets:** Evidence of Glacier and Ice Sheet Extent; Growth and Decay. **Glaciations:** Late Quaternary in North America. **Ice Core Methods:** Borehole Temperature Records; Chronologies; Overview. **Ice Core Records:** Greenland Stable Isotopes; Ice Margin Sites. **Ice Cores:** Dynamics of the East Antarctic Ice Sheet; Dynamics of the West Antarctic Ice Sheet; History of Research, Greenland and Antarctica. **Sea Level Studies:** Eustatic Sea-Level Changes Since the Last Glacial Maximum.

References

- Alley RB, Gow AJ, Meese DA, Fitzpatrick JJ, Waddington ED, and Bolzan JF (1997) Grain-scale processes, folding and stratigraphic disturbance in the GISP2 ice core. *Journal of Geophysical Research* 102(C12): 26315–26316.
- Bales RC, Shen D, Du G, et al. (2009) Annual accumulation for Greenland updated using ice core data developed during 2000–2006 and analysis of daily coastal meteorological data. *Journal of Geophysical Research-Atmospheres* 114: D06116. <http://dx.doi.org/10.1029/2008JD011208>.

- Bamber JL, Ekholm S, and Krabill WB (2001a) A new, high-resolution digital elevation model of Greenland fully validated with airborne laser altimeter data. *Journal of Geophysical Research* 106(B4): 6733–6745.
- Bamber JL, Layberry RL, and Gogineni SP (2001b) A new ice thickness and bed data set for the Greenland ice sheet. 1. Measurement, data reduction, and errors. *Journal of Geophysical Research* 106(D24): 33773–33780.
- Buchard SL and Dahl-Jensen D (2008) At what depth is the Eemian layer expected to be found at NEEM? *Annals of Glaciology* 48: 100–102.
- Cuffey KM and Marshall SJ (2000) Substantial contribution to sea-level rise during the last interglacial from the Greenland ice sheet. *Nature* 404: 591–594.
- Cuffey KM and Paterson WSB (2010) *The Physics of Ice*, 4th edn. Oxford: Elsevier.
- Dahl-Jensen D and Gundestrup NS (1987) *Constitutive Properties of Ice at Dye 3, Greenland*. Oxfordshire: IAHS Publication No. 170, pp. 31–43.
- Dahl-Jensen D, Gundestrup N, Gogineni SP, and Miller H (2003) Basal melt at NorthGRIP modeled from borehole, ice-core and radio-echo sounder observations. *Annals of Glaciology* 37: 207–212.
- Dahl-Jensen D, Gundestrup NS, Keller K, et al. (1997) A search in north Greenland for a new ice-core drill site. *Journal of Glaciology* 43(144): 300–306.
- Dansgaard W (2004) *Frozen Annals*. Copenhagen, p. 122. Denmark: Niels Bohr Institute, University of Copenhagen.
- Fahnestock M, Abdalati W, Joughin I, Brozena J, and Gogineni P (2001) High geothermal heat flow, basal melt, and the origin of rapid ice flow in Central Greenland. *Science* 294: 2338–2342.
- Gogineni S, Chuah T, Allen C, Jezek K, and Moore RK (1998) An improved coherent radar depth sounder. *Journal of Glaciology* 44(148): 659–669.
- Gogineni S, Tammana D, Braaten D, et al. (2001) Coherent radar ice thickness measurements over the Greenland ice sheet. *Journal of Geophysical Research* 106(D24): 33761–33772.
- Hammer CU (1980) Acidity of polar ice cores in relation to absolute dating, past volcanism, and radio-echos. *Journal of Glaciology* 25(93): 359–372.
- Hansen KM, Svensson A, Wang Y, and Steffensen JP (2002) Properties of GRIP ice crystals from around Greenland interstadial 3. *Annals of Glaciology* 35: 531–537.
- Harrison CH (1973) Radio echo sounding of horizontal layers in ice. *Journal of Glaciology* 12(66): 383–397.
- Huybrechts P (1994) The present evolution of the Greenland ice sheet: An assessment by modeling. *Global and Planetary Change* 9: 39–51.
- Hvidberg CS, Keller K, and Gundestrup N (2002) Mass balance and ice flow along the NNW ridge of the Greenland Ice Sheet at NorthGRIP. *Annals of Glaciology* 35: 521–526.
- Johannessen OM, Khvorostovsky K, Miles MW, and Bobylev LP (2005) Recent ice-sheet growth in the interior of Greenland. *Science* 310: 1013–1016.
- Joughin I, Abdalati W, and Fahnestock M (2004) Large fluctuations in speed on Greenland's Jakobshavn Isbrae glacier. *Nature* 432: 608–610.
- Joughin I, Smith BE, Howat IM, Scambos T, and Moon T (2010) Greenland flow variability from ice-sheet-wide velocity mapping. *Journal of Glaciology* 56(197): 415–430.
- Krabill W, Abdalati W, Frederick E, et al. (2000) Greenland ice sheet: High-elevation balance at peripheral thinning. *Science* 289: 428–430.
- Marshall SJ and Cuffey KM (2000) Peregrinations of the Greenland ice sheet divide in the last glacial cycle. Implications for central Greenland ice cores. *Earth and Planetary Science Letters* 179: 73–90.
- North Greenland Ice-Core Project (NorthGRIP) Members (2004) High resolution climate record of the northern hemisphere reaching into the last glacial–interglacial period. *Nature* 431: 147–151.
- Otto-Bliesner BL, Marshall SJ, Overpeck JT, Miller GH, and Hu ACAPE Last Interglacial Project Members (2006) Simulating Arctic climate warmth and icefield retreat in the last interglaciation. *Science* 311: 1751–1753. <http://dx.doi.org/10.1126/science.1120808>.
- Rasmussen SO, Andersen KK, Svensson AM, et al. (2006) A new Greenland ice core chronology for the last glacial termination. *Journal of Geophysical Research* 111: D06102. <http://dx.doi.org/10.1029/2005JD006079>.
- Reeh N, Mayer C, Miller H, Thomson HH, and Weidick A (1999) Present and past climate control on fjord glaciations in Greenland: Implications for IRD-deposition in the sea. *Geophysical Research Letters* 26: 1039–1042.
- Rignot E and Kanagaratnam P (2006) Changes in the velocity structure of the Greenland Ice Sheet. *Science* 311: 5763. <http://dx.doi.org/10.1126/science.1121381>.
- Sørensen LS, Simonsen SB, Nielsen K, et al. (2011) Mass balance of the Greenland ice sheet (2003–2008) from ICESat data – The impact of interpolation, sampling and firn density. *The Cryosphere* 5: 173–186. <http://dx.doi.org/10.5194/tc-5-173-2011>.
- Souchez R (1997) The buildup of the ice sheet in central Greenland. *Journal of Geophysical Research* 102(C12): 26317–26323.
- Svensson A, Nielsen SW, Kipfstuhl S, et al. (2005) Visual stratigraphy of the North Greenland Ice Core Project (NorthGRIP) ice core during the last glacial period. *Journal of Geophysical Research* 110: D02108. <http://dx.doi.org/10.1029/2004JD005134>.
- Thomas RHPARCA Investigators (2001) Program for arctic regional climate assessment (PARCA): Goals, key findings, and future directions. *Journal of Geophysical Research* 106(D24): 33691–33705.
- van den Broeke M, Bamber J, Ettema J, et al. (2009) Partitioning recent Greenland mass loss. *Science* 326: 984–986. <http://dx.doi.org/10.1126/science.1178176>.
- Van der Veen CJ (1999) *Fundamentals of Glacier Dynamics*. Rotterdam: Balkema.
- Velicogna I (2009) Increasing rates of ice mass loss from the Greenland and Antarctic ice sheets revealed by GRACE. *Geophysical Research Letters* 36: L19503. <http://dx.doi.org/10.1029/2009GL040222>.
- Vinther BM, Buchardt SL, Clausen HB, et al. (2009) Holocene thinning of the Greenland ice sheet. *Nature* 461: 385–388. <http://dx.doi.org/10.1038/nature08355>.
- Zachos J, Pagani M, Sloan L, Thomas E, and Billups K (2001) Trends, rhythms, and aberrations in global climate 65 Ma to present. *Science* 292: 686–693.
- Zwally HJ, Abdalati W, Herring T, Larson K, Saba J, and Steffen K (2002) Surface melt-induced acceleration of Greenland ice-sheet flow. *Science* 297: 218–222.

Dynamics of the West Antarctic Ice Sheet

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Overview

Ice cores collected on highly dynamic ice sheets acquire additional complexities in interpretation; however, they return to the glaciologist additional constraints that help resolve the history of the ice sheet itself. The paleoclimatic record of such ice cores must be extracted in the context of changes in flow and shape of the ice sheet. Without knowing these changes, the paleoclimate record will contain errors.

The West Antarctic ice sheet has been intensively studied by Antarctic researchers for the past quarter century (Alley and Bindshadler, 2001). It is the type of ice sheet capable of changing suddenly and could make its demise felt worldwide by releasing enough ice to raise sea level by an average of 6 m. This might happen in centuries or longer, or perhaps not at all. The fastest changes in ice volume and sea level will be driven by changes in the flow of the ice sheet rather than changes in the rate of snow accumulating on the ice sheet or melting at the margins. Predicting such changes requires an understanding of the dynamics of this ice sheet.

Geographic and Geologic Setting

The West Antarctic ice sheet is a distinct region of the larger Antarctic ice sheet that lies predominantly in the Western Hemisphere. Its 1.9 million square kilometer area covers about 10% of the Antarctic continent. It is bounded on the south by the 3300 km long Transantarctic Mountain chain and on the north by the Weddell, Bellingshausen, Amundsen, and Ross Seas. The Antarctic Peninsula is usually not considered part of West Antarctica. West Antarctica contains the two largest ice shelves in the world: the Ross and Ronne Ice Shelves, as well as a perimeter of smaller ice shelves along the Bellingshausen and Amundsen Sea coasts (Figure 1).

The ice cradled between the Transantarctic Mountains and the coastal mountains fills a deep oceanic basin that averages more than a kilometer deep. This ocean basin includes an extensional rift basin similar to the Basin and Range region of the western United States. The thinner ocean crust leads to higher rates of geothermal heating that warm the base of the ice sheet to or near the melting point over most of the region. Warmer ice is softer and more easily deformed. Numerous volcanoes hint at a restless geology, the details of which lie mainly hidden beneath an icy mantle.

During warmer periods in Earth's past, there was little or no ice in West Antarctica and this basin was filled with a living ocean. Expired marine organisms fell to the ocean floor building up a layer of marine sediment on which the ice sheet now sits. Water, produced by the geothermal heat, weakens this marine sediment making it difficult for a thick ice sheet to find a foothold on such a soft slippery bed.

For these reasons, the West Antarctic ice sheet is thinner and more dynamic than the larger, thicker, and colder East Antarctic ice sheet. This interaction of ice, water, and sediment is the essential interplay that dictates the dynamics of the West Antarctic ice sheet. This type of ice sheet is sometimes referred to as a 'marine-based' ice sheet (Figures 2(a) and (b)).

Thin is a relative term, however, and West Antarctic ice averages over 1 km thick and in places fills the valleys of rugged subglacial mountain ranges to an overall ice depth of 2500 m. These thicknesses are more than enough to make contact with the submarine bed. As the ice flows, it becomes thinner, eventually thinning enough to float but still hundreds of meters thick. This grounded-to-floating transition occurs at the 'grounding line' and marks the beginning of a floating ice shelf. West Antarctica contains the two largest ice shelves: the Ross Ice Shelf (850 000 km²) and the Ronne Ice Shelf (257 000 km²) on either side of the West Antarctic ice sheet. These ice shelves begin with ice up to 2000 m thick at the grounding line and thin to a few hundred meters thick at the seaward perimeter of the ice sheet where tidal flexure, wind, and wave action eventually fracture the ice, releasing icebergs.

West Antarctica can be subdivided into three large regions. Identified by the receiving waters, these are the Ross Sea sector, the Amundsen/Bellingshausen Sea sector, and the Weddell Sea sector. In the Ross Sea sector, the ice is discharged via a series of ice streams, rapidly moving rivers of ice set within nearly stagnant ice ridges. These ice streams, named from south to north, Mercer, Whillans, Kamb, Bindshadler, MacAyeal, and Echelmeyer Ice Streams, are about 1 km thick, move at speeds of many hundreds of meters per year, extend for a few hundred kilometers, and are tens of kilometers wide. More elevated ice ridges of similar size lie between the ice streams, but move at speeds of less than 10 m yr⁻¹.

The Amundsen/Bellingshausen Sea sector differs from the Ross Sea sector in two significant ways. First, the ice is confined by a rim of coastal mountains through which the ice flows. The ice overrides the lower elevation peaks, creating a few rapidly moving outlet glaciers, Pine Island and Thwaites glaciers being the largest. They are similar in size to ice streams. The use of the term glacier rather than ice stream is more historical than scientific. These coastal glaciers were observed by early Antarctic explorers long before ice streams were known. The elevated terrain probably diminishes the amount of marine sediment in this area, but the subglacial setting observations are insufficient to decide how the dynamics of this region differ from that of the Ross Sea sector. The second difference with the Ross Sea sector is that the ice shelves fed by these outlet glaciers are much smaller.

The Weddell Sea sector has features somewhat intermediate to the other two sectors. It contains a large mountain range, the Ellsworth Mountains, that strongly influences the ice flow, but the major outlet features, the Rutford and Evans Ice Streams,

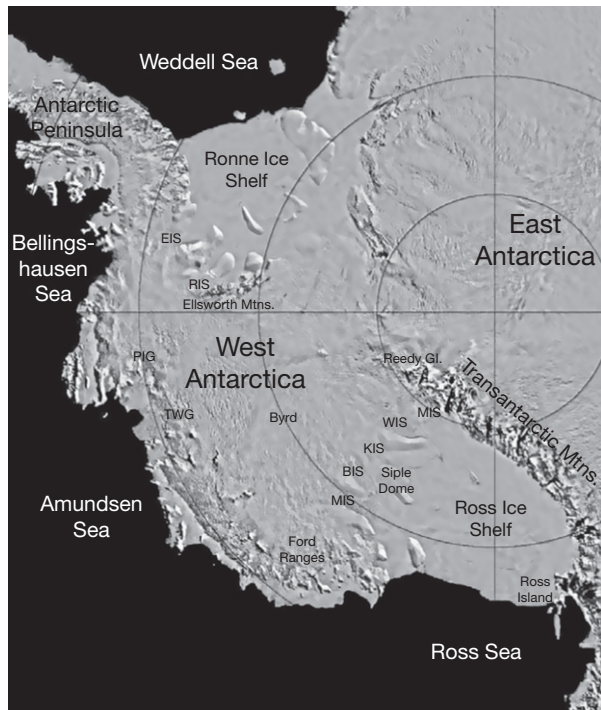


Figure 1 Satellite image of the West Antarctic ice sheet with various features identified. Ice streams and outlet glaciers are indicated by initials. Satellite image is a composite of data from NASA's Moderate Resolution Imaging Sensor (MODIS). Courtesy of National Snow and Ice Data Center.

resemble ice streams in form as well as name and their thick ice at the grounding line feed the large Ronne Ice Shelf.

Theories of Marine-Based Ice Sheet Collapse

The sudden demise of marine-based portions of the Laurentide and Fennoscandian ice sheets, in Canada and Scandinavia, respectively, during the last glacial period (roughly from 125 ka to 20 ka), caused episodes of rapid sea-level rise. Rates of rise exceeded 1 m per decade, more than 50 times the average rate during the twentieth century (Fairbanks, 1989). Today West Antarctica is the only marine-based ice sheet on this planet. Its modern existence poses a vexing quandary: does the West Antarctic ice sheet still exist because it is the last ice sheet of this type to suffer this inevitable fate or, conversely, may it contain some unique counterbalancing feature that causes it to avoid the fate of its Northern Hemisphere counterparts.

It has been argued that because floating ice thins faster than grounded ice of equal thickness, marine-based ice sheets may be inherently unstable (Weertman, 1974). Another argument for instability posits that ice shelves buttress the grounded ice that feeds them and any reduction of this floating portion will weaken its damming effect and lead to an increased discharge of the buttressed ice into the ocean, raising sea levels (Thomas, 1973). A deeper understanding of the dynamics of marine-based ice sheets is required to answer this disturbing question.

West Antarctic Ice-Sheet Dynamics

Ice Streams and Outlet Glaciers

The key to understanding the dynamics of the West Antarctic ice sheet is understanding the processes responsible for the observed behavior of ice streams and outlet glaciers. Nearly all the snow that falls in West Antarctica is eventually funneled through these critical features. By virtue of their speed, they are the fastest responding elements of the ice sheet and dictate whether the ice sheet grows or shrinks. In the remaining discussion, reference will be made to ice streams and not outlet glaciers. In West Antarctica, the differences are probably not significant; however, the dynamics of outlet glaciers have not received as much attention as that of ice streams.

Ice streams move fast because of the properties of the material upon which they sit and the lubrication between the ice and substrate. West Antarctic ice streams rest on marine till, a mixture of marine sediment, deposited during times when West Antarctica was open ocean, and glacial till, created by the abrasive motion of ice and encased rocks sliding across an erodable bed. The lubricating fluid is water, produced by the geothermal heat rising through the bed, and by additional friction of the ice's motion against its bed.

An ice stream's rapid motion is predominantly the result of sliding over its bed and not deformation within either the ice or the till. The weight of the ice, acting under the force of gravity, forces the ice to flow downhill. This force is counteracted primarily by resistance at the base and the sides of the ice stream (Figure 3).

Side Shear

When the bed is well lubricated (an ice shelf is the extreme case), the shearing of ice at the lateral boundaries provides resistance to flow. The work done during shearing generates heat that warms the margin, making the ice softer and prone to deform even more quickly. This is offset by the cooling effect of the slow-moving and colder ridge ice flowing across the margin into the ice stream. Modeling studies of the margin indicate they can be expected to migrate at rates of tens of meters per year (Jacobson and Raymond, 1998). Observations have confirmed that margin migration does occur in some locations at roughly these rates.

Mechanically, the resistive forces associated with side shear are transferred through the body of the slower-moving ridge to the bed. Surprisingly, specific field studies to detect differences in bed conditions across an active ice-stream margin showed no difference in water content. The tentative conclusion is that the transition from the drier bed margin to a wet till under the central ice stream is not sharp.

Basal Stress

At the ice-stream bed, water is a key ingredient. It is a lubricating fluid and its inclusion in marine till makes it extremely soft. Existing water also moves along the ice-bed interface, controlled by hydraulic pressure gradients. The pressure of this water system has been measured as being very close to the weight of the overlying ice. As the water moves, these pressures fluctuate, sometime exceeding the ice stream's weight, causing

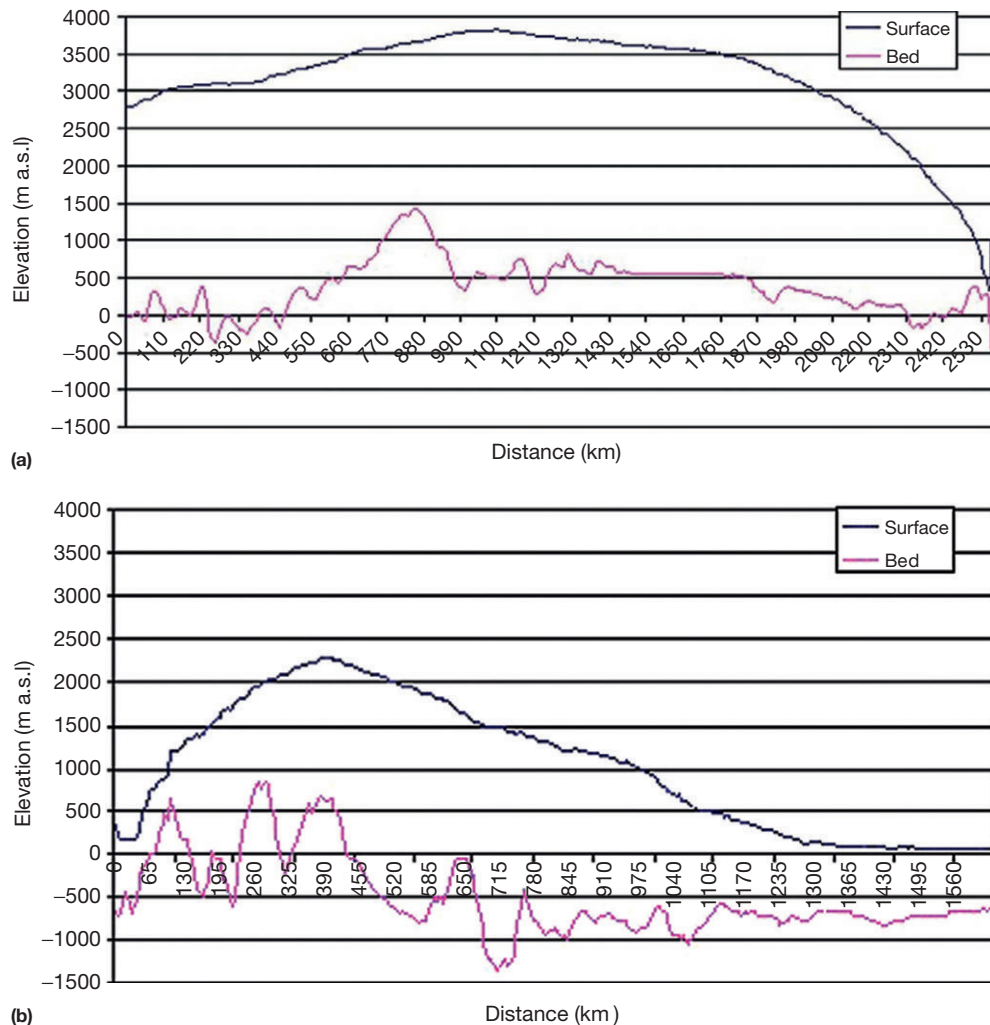


Figure 2 Two longitudinal profiles contrasting the surface and bed shape of East and West Antarctica. The East Antarctic ice sheet is higher, thicker, and is grounded on a bed mostly above sea level. The West Antarctic ice sheet is lower, thinner, and is grounded on a bed below sea level. The East Antarctic profile begins at the South Pole and follows the 90° E longitude to the ice-sheet edge. The West Antarctic profile begins on the Ronne Ice Shelf, runs obliquely across the ice sheet, follows an ice stream, and ends at the edge of the Ross Ice Shelf.

the ice to be fully supported by the water and the basal stress to drop to zero. The surface topography influences the subglacial hydrology more than the basal topography, but both are relevant, so the pattern of subglacial water flow can be different from the ice-flow pattern and changes in the surface can alter the flow patterns of ice and water differently.

As with the sides, heat flow is a critical factor in determining the influence of the bed on ice flow. An excess amount of heat can melt basal ice and keep the bed lubricated, whereas a heat deficit will cause water to freeze, removing the lubricating fluid, drying the till, and making it stiffer.

There is lingering controversy as to how much the deformation of the till contributes to ice-stream motion (Kamb, 2001). A till in front of active ice streams shows foresetted beds that indicate downstream transport and redeposition of till (Alley et al., 1989). In addition, ice penetrating radar measurements of the bed beneath active ice streams show long furrows in the till aligned with the flow direction, suggesting the active sculpting of the till by the ice stream. There is

probably an active layer of very deformable till immediately beneath the ice stream. The depth of this active layer depends on the thermal and hydrologic conditions and as these factors change, so will the characteristics of the active layer (Figure 4).

Transitions and Tributaries

At the ends of an ice stream, a third resistive force becomes important. When the ice stream enters the ice shelf, basal resistance is lost and the nearby margins are replaced by margins much farther away. These sudden changes would cause a dramatic jump in speed, were it not for longitudinal stress acting along the direction of motion smoothing this transition. Another transition occurs at the upstream end where ice streams start. This transition is also gradual and takes place within well defined geologic troughs that allow thicker ice to accelerate. The accelerating ice defines 'tributaries' that coalesce into larger, faster flows (Joughin et al., 1999). The tributaries can extend a few hundred kilometers upstream of the ice stream's origin, or

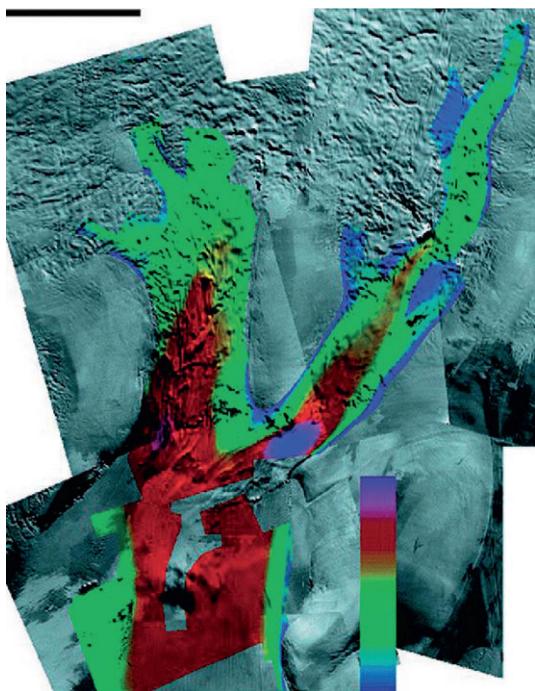


Figure 3 Mosaic of Landsat images and color map of surface velocity of Bindschadler (right) and MacAyeal (left) Ice Streams. Flow is from top to bottom and velocities range from near zero (blue) to 700 m yr^{-1} (purple). Central portion of each ice stream flows fastest. Slower ice at the margins is wider on Bindschadler Ice Stream. Data coverage of surface velocity is not continuous over Ross Ice Shelf, toward bottom of figure, where two ice streams merge. Black bar at top represents a distance of 100 km.



Figure 4 Dr. Barclay Kamb examining the first subglacial till removed from beneath a West Antarctic ice stream. Photo by H. Englehardt.

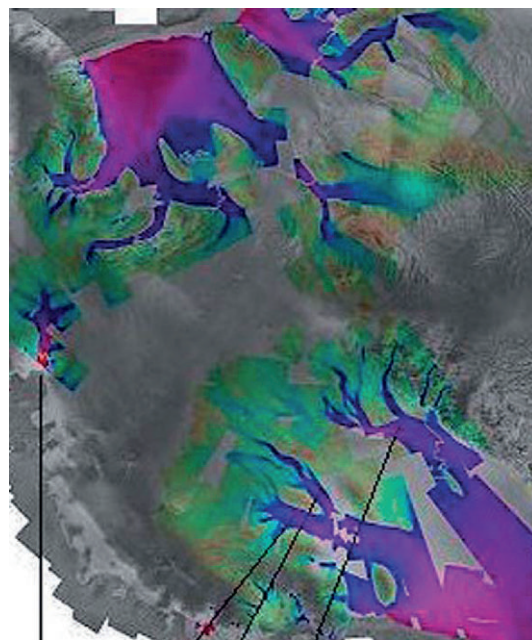


Figure 5 Pattern of West Antarctic ice surface speed shows the network of tributaries feeding ice streams and outlet glaciers. Slowest speeds are in brown, then yellow, with fastest speeds of over 500 m yr^{-1} in magenta. Colors are superimposed on radar mosaic of West Antarctica. Image courtesy of I. Joughin.

‘onset’. Basal sliding is likewise the dominant process of ice motion. When ice speeds reach about 100 m yr^{-1} , increased side-shear stresses begin to form crevasses (Bindschadler et al., 2001). The positions of onsets may be dictated by the presence of the marine till because they closely match the paleoshoreline of an ice-free West Antarctic after accounting for the full recovery of the earth’s crust following removal of the ice load (Blankenship et al., 2001). Alternatively, the gradual increase in basal heat, caused by increased sliding, may determine when the supply of subglacial water is sufficient to sustain an ice stream’s rapid motion. The ice streams exhibit a generally straighter flow direction and are steered less by the subglacial topography than the tributaries that feed the ice streams. The paths of ice streams sometime disregard the topography altogether, occupying one side of a deep subglacial valley and not the other side, or straddling a portion of a valley and the surrounding elevated terrain. This reinforces the fact that the margins of ice streams can migrate (Figure 5).

Important Timescales

One of the most confusing aspects of West Antarctic ice-sheet behavior is that change has been observed on many different timescales. This can confound scientific study and make comparison of data difficult.

Glacial Timescale

Over the past million years, Earth has experienced waxing and waning ice sheets at a general 100 ka periodicity: glacials

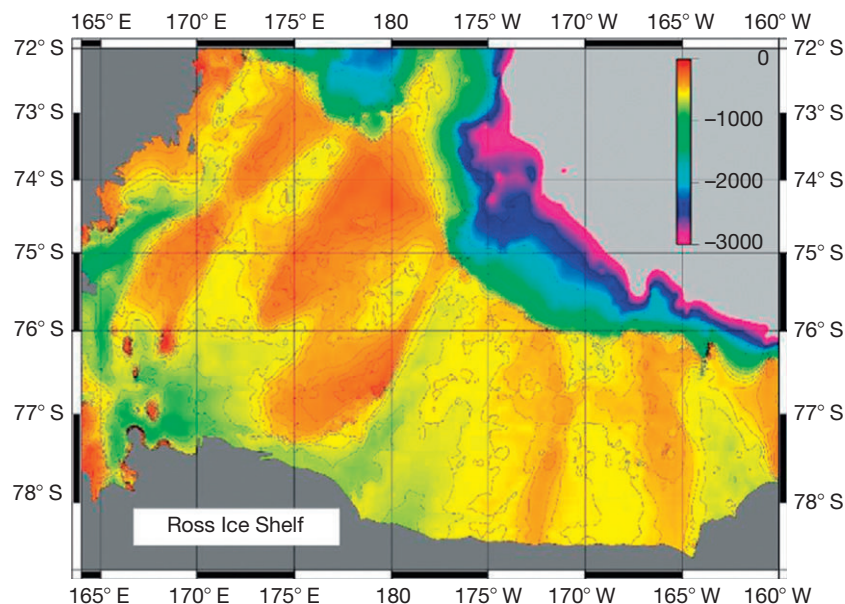


Figure 6 Bathymetry of the Ross Sea shows sequence of ridges (red) and troughs (yellow) trending north–northeast. Troughs were occupied by paleo-ice streams of a former, larger West Antarctic ice sheet. Map courtesy of J. Anderson.

lasting about 90 ka separated by interglacial periods typically lasting 10 ka. The West Antarctic ice sheet is known to have been largely absent at least once during the last 400 ka based on ocean-dwelling diatoms found in subglacial sediments recovered from beneath the center of the ice sheet (Scherer, 1991). It has not been successfully determined whether the West Antarctic ice sheet survived the last interglacial (that peaked about 125 ka). If not, it underscores the importance of the question of why this ice sheet persists when the present interglacial has lasted twice as long as the average.

It is easier to map the extent of ice sheets during their more extensive glacial periods because the postglacial retreat reveals the former subglacial beds and margins. The deep-carved troughs that stretch to nearly the edge of the continental shelf around West Antarctica confirm both the former extent of the ice sheet and the former existence of ice streams. Ice covering the Ross Sea began its retreat about 11.400 ka, later than the retreat initiation 18 ka in the Northern Hemisphere and in parts of East Antarctica. Portions of the retreat were rapid, with a retreat hiatus occurring about 7 ka when the front was in the vicinity of Ross Island (Conway et al., 1999). This hiatus may be associated with a widespread mid-Holocene warm period evidenced throughout much of Antarctica.

Thinning, associated with this retreat, has been followed in the near-coastal Ford Ranges. The time when previously ice-covered rocks are first exposed can be determined by measuring the decay rate of certain chemical isotopes. The highest summits emerged about 10.4 ka and the lower summits as recently as 3.8 ka. Thinning is presently continuing. Average exposure time of the Ford Range summits over the more distant past is less than 50% while the lower elevations that are most recently exposed have been exposed less than 1–5% of the time. This means that the ice extent in this region is at an extreme minimum and probably less than during the extended and very warm interglacial 400 ka.

More interior sites experienced lesser amounts of thinning. Isotopic analysis of ice cored from Byrd Station indicates it was no more than 400 m higher during the Last Glacial Maximum (Steig et al., 2001). Reedy Glacier, at the very southern extreme of West Antarctica, too was only a few hundred meters higher and experienced a more gradual drawdown during the past 10 ka. Internal layering revealed by ice-penetrating radar and dated by an ice core has been used to deduce that Siple Dome was likely overrun by the glacial ice sheet, did not begin to thin until 7 ka, and emerged as an ice dome about 3 ka, supporting the theory of unsteady retreat (Figure 6; Nereson and Raymond, 2001).

Millennial Timescale

Not only is the size of the West Antarctic ice sheet ephemeral, the ice streams within it are short-lived as well. The fast motion of active ice streams creates crevassed margins and rough surfaces. Past margins have been identified by ice-penetrating radars and can be seen in satellite imagery. The region of slow-moving ice between Whillans Ice Stream and Mercer Ice Stream is now relatively slow moving, but its rough surface and buried crevasses suggest it flowed at ice-stream speeds sometime in the past.

Distinct paleomargins identify and define a former ice stream, unofficially dubbed ‘Siple Ice Stream’, that flowed northwesterly across the upstream side of Siple Dome. It was active until about 450 yr ago and may have directed the upstream region of Kamb Ice Stream into the downstream region of Bindshadler Ice Stream (Gades et al., 2000).

Some former margins are so clearly defined it is unreasonable to expect that they migrated gradually to their present positions. Thus, the dynamics of ice streams must include provision for margins to jump distances of many kilometers. It is not known how suddenly this would occur and whether a

brief period of inactivity would be required. The self-sustaining aspect of streaming flow, the possible difficulties of restarting a stagnant ice stream, and the observation that near an active margin the subglacial till underneath the ice stream is relatively dry, make it likely that margins can at least jump inward without needing to stop streaming flow.

Surface ridges and troughs that stretch along the length of fast-moving ice streams tracing out the direction of flow aid in disclosing past changes in flow direction. Called 'flowstripes', 'flow traces', or 'streaklines', these features are created by irregularities of either basal topography or basal lubrication. Similar to medial moraines on mountain glaciers, these features align with the velocity field when flow is steady for long periods of time. Misalignment indicates nonsteady flow.

Flowstripes being transported across the large Ross and Ronne ice shelves provide a means to look back at the flow pattern emanating from the grounded ice streams for the past few centuries. Ice presently near the seaward edge of the Ross Ice Shelf entered the ice shelf about 1 ka. Large bowed structures are indisputable evidence of major changes in flow having occurred in the region now fed by the Whillans and Mercer Ice Streams. The analysis of this large bowed pattern, other less emphatic flowstripe deviations, and crevassing patterns indicate that there were some major flow changes in the West Antarctic ice streams feeding the Ross Ice Shelf about 500 yr ago (Fahnestock et al., 2000). By contrast, the flowstripes discharged from the East Antarctic ice sheet, through the Transantarctic Mountains into the Ross Ice Shelf and the general pattern across the Ronne Ice Shelf align well with present flow, indicating steady flow direction (Figure 7).

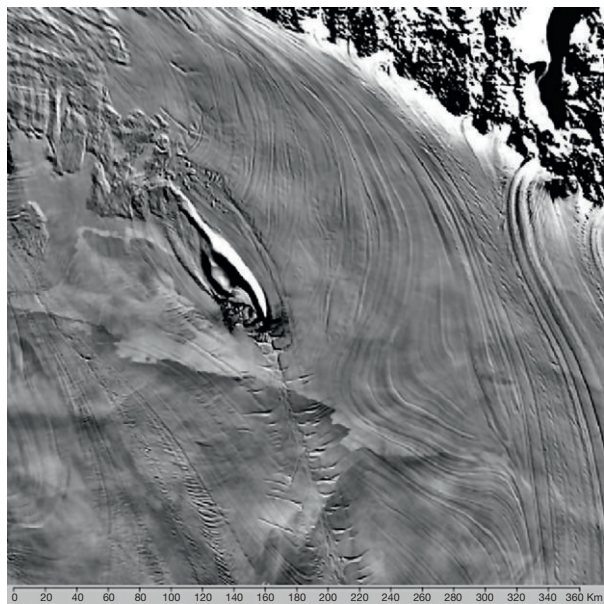


Figure 7 Satellite image of a portion of the Ross Ice Shelf shows bowed flowstripes indicating a large change in flow occurred upstream in the past. Transantarctic Mountains are in the upper right and ice flow is from upper left to lower right. As ice flows around Crary Ice Rise (grounded ice in the mouth of Whillans Ice Stream indicated by the bright and dark feature left of center), crevasses are formed that are carried along the ice shelf.

Centennial Timescale

The most recent known stagnation event of the West Antarctic ice sheet took place on Kamb Ice Stream. The amount of snow covering its once active ice margins indicates it stopped suddenly over most of its length approximately 150 yr ago (Smith et al., 2002). The tributary system feeding the ice stream remains active, forming a bulge of ice growing at about half a meter per year and unable to exit the ice sheet. How the growth of this bulge will evolve and whether the lower ice stream will reinitiate, are ongoing research questions motivating sustained field observations.

There has also been speculation about what caused the stagnation. The present geometry is diverting subglacial water from flowing underneath Kamb Ice Stream to flowing south into the still-active Whillans Ice Stream. One theory claims that this diversion caused stagnation of Kamb Ice Stream. Another theory disputes this, contending that the hydrologic diversion was a consequence, rather than a cause, of the stagnation and that the ice stream thinned enough to freeze basal water, stiffen the till, and stop the stream, forcing upstream water to go elsewhere (Anandakrishnan et al., 2001).

Decadal

Field studies of the flow of the West Antarctic ice sheet began in the early 1980s and satellite data extend to the early 1960s, allowing direct measurements of change on the decadal timescale. Acceleration, deceleration, and steady flow are all observed. Curiously, these three different regimes occur distinctly in the three different geographical regions of West Antarctica. Acceleration is most commonly observed in the Amundsen Sea region. Pine Island and Thwaites Glaciers and some of the much smaller outlet glaciers are accelerating at rates of a few percent per year (Rignot and Schmelz, 2002). These rates cannot have been sustained for decades without there being many far-ranging and observable changes elsewhere. Thus, they are likely a recent occurrence.

There has been a decrease in the velocities across the entire Whillans Ice Stream at the typical rate of $1\text{--}2\%$ yr^{-1} since the 1980s (Joughin et al., 2002). Slightly higher rates have been observed downstream where basal freezing is likely, implying that the removal of water is making the till stiffer, increasing the basal resistance to flow. It is possible this deceleration will eventually result in stagnation of the ice stream.

In the Weddell Sea region, no sustained changes in flow speed have been observed over this same multidecadal period.

Twenty-five years of surface elevation measurements by satellite altimeters show a pattern of change consistent with the spatial pattern of velocity change: the Amundsen Sea sector is thinning; the Ross Sea sector is thickening; and the Weddell Sea sector is roughly in equilibrium (Wingham et al., 1998). The thickening in the Ross Sea sector is primarily the result of the stagnant Kamb Ice Stream. Based on the most recent laser altimetry results, the decelerating Whillans Ice Stream is now contributing to this thickening, while the other active ice streams are thinning slightly. In the Amundsen Sea sector, there are major changes taking place in many of the outlet glaciers (discussed below). However, Thwaites Glacier stands out as a major outlet whose current flow is considerably

different from what accumulation and geometry suggest should be the discharge pattern of this glacier.

Daily

Ocean tides have a surprisingly large effect on ice stream motion. Precise Global Positioning System (GPS) positions indicate that where ice streams enter both the Ronne and Ross Ice Shelves, ice flows about 10% faster during periods of intense tides. In the Ross Sea sector, there is even more dramatic sensitivity: most of the lower ice streams accelerate to flow 50% above their mean speed during falling tides with a symmetric deceleration to speeds approximately 50% below the average during rising tides (Anandakrishnan et al., 2003). The notable exception is Whillans Ice Stream which exhibits stick-slip motion: most of the motion occurs during two brief episodes lasting 20–30 min during falling tide. During these slip events, speed corresponds to that expected for a completely lubricated bed and the acceleration and deceleration require no more than one to a few minutes (Bindschadler et al., 2003). In all cases of tidal modulation, the phenomenon decays gradually upstream, but can still be detected more than 100 km from the grounding line.

The timing of the stick-slip events can be reproduced by a simple, plastic-bed model where the slip occurs after a threshold strain is exceeded. It is probably important that the only ice stream exhibiting stick-slip is also the ice stream experiencing strong basal freezing of basal water. It is unlikely that the tidal variations actually transport water significant distances under the ice. The tidal effects are probably transmitted either via pressure through the subglacial hydraulic system or by longitudinal stresses within the ice. A more complex model is needed to explain the gradational response of the other ice streams.

A View of the Future

Satellite sensors have revealed the multiple facets of a rapidly shrinking ice sheet in the Amundsen Sea region where Pine Island and Thwaites glaciers are the major outlets. Both they and their neighboring glaciers are accelerating and thinning most rapidly nearest the coast (Shepherd et al., 2004). This is the pattern expected of an ice-sheet collapse. Recent airborne observations establish that thinning is accelerating (Thomas et al., 2004). The consistent and widespread pattern suggests external forcing. Nearly none of the former ice shelf remains and the extent of the fringing sea-ice cover is decreasing, as well. These factors may be significant because the larger magnitude changes at the coast suggest that they are at least partly driven by the influx of warmer waters up onto the continental shelf, where currents lead them to interact with the underside of the ice shelf and other floating ice (Figure 8).

Extensive warming on the Antarctic Peninsula has led to the sudden disintegration of individual ice shelves along both east and west coasts, driven by intense summer melting. In one monitored case, ice upstream of a removed ice shelf accelerated in agreement with the theory contending that the grounded portion of the West Antarctic ice sheet is buttressed by the ice shelf and that removal or thinning of the shelf will lead to acceleration and thinning of the grounded ice (Scambos et al., 2000). The present Amundsen Sea behavior also conforms to this pattern. Presently, the Ross and Ronne Ice Shelves have experienced only occasional summer melting. Many more degrees of warming will be required to increase the frequency and the intensity of summer melting on these ice shelves.

Research on the dynamics of the West Antarctic ice sheet is intended to produce knowledge of the important processes so that predictions of the future behavior of the ice sheet can be

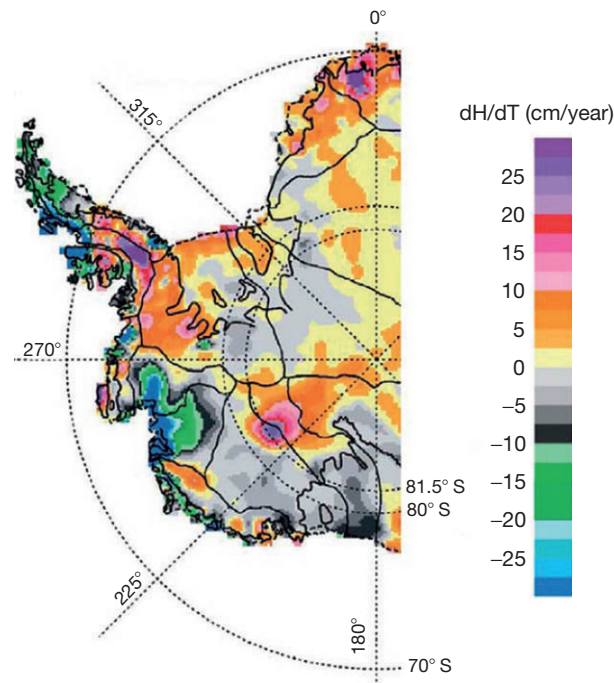


Figure 8 Composite figure showing elevation changes taking place in West Antarctica. Results are processed from satellite radar altimeter data. Figure courtesy of H.J. Zwally.

made well in advance of dramatic change to the ice sheet and global sea level. Numerical models are the tools to make these predictions of the future. Models of the West Antarctic ice sheet with variable grids can include adequate resolution to capture the important features but they do not yet include all the complex interactions between ice, till, and water in ways that reproduce the observed and inferred behavior on the variety of timescales known from observations. Recent models have been able to simulate ice-stream stagnation and reactivation through thermal triggering and are also becoming better at matching the variety of field data of motion, flow history, and temperature structure (Hulbe and Payne, 2001). The compelling importance of knowing the heat budget requires these models to be fully thermodynamic. Also, the fundamental nature of water in a variety of ways also requires calculation of not only water balance, but also subglacial water flow, and its effect on till strength. These are challenges that will be met as knowledge of the dynamics of the West Antarctic ice sheet improves.

See also: **Glaciations:** Early Quaternary (Pleistocene) and Precursors; Late Quaternary of Antarctica. **Ice Core Methods:** Stable Isotopes. **Ice Cores:** Dynamics of the East Antarctic Ice Sheet; Dynamics of the Greenland Ice Sheet. **Introduction:** Societal Relevance of Quaternary Research.

References

- Alley RB and Bindschadler RA (2001) The West Antarctic ice sheet and sea-level change, in the West Antarctic ice sheet: Behavior and environment. *Antarctic Research Series* 77: 1–11.
- Alley RB, Blankenship DD, Rooney ST, and Bentley CR (1989) Sedimentation beneath ice shelves – The view from ice stream B. *Marine Geology* 85: 101–120.
- Anandakrishnan S, Alley RB, Jacobel RW, and Conway H (2001) The flow regime of ice stream C and hypotheses concerning its recent stagnation, in the West Antarctic ice sheet: Behavior and environment. *Antarctic Research Series* 77: 283–296.
- Anandakrishnan S, Voigt DE, Alley RR, and King MA (2003) Ice stream D flow speed is strongly modulated by the tide beneath the Ross Ice Shelf. *Geophysical Research Letters* 30(7): 1361 Art.no.
- Bindschadler RA, Bamber J, and Anandakrishnan S (2001) Onset of streaming flow in the Siple Coast region, West Antarctica. *Antarctic Research Series* 77: 123–136.
- Bindschadler RA, King M, Alley RA, Anandakrishnan S, and Padman L (2003) Tidally controlled stick-slip discharge of a West Antarctic ice stream. *Science* 301: 1087–1089.
- Blankenship DD, Morse DL, Finn CA, et al. (2001) Geologic controls on the initiation of rapid basal motion for West Antarctic ice streams: A geophysical perspective including new airborne radar sounding and laser altimetry results. In: Alley RB and Bindschadler R (eds.) *Antarctic Research Series: The West Antarctic Ice Sheet: Behavior and Environment*, 105–121.
- Conway H, Hall BL, Denton GH, Gades AM, and Waddington ED (1999) Past and future grounding-line retreat of the West Antarctic ice sheet. *Science* 286: 280–283.
- Fahnestock MA, Scambos TA, Bindschadler RA, and Kvaran G (2000) A millennium of variable ice flow recorded by the Ross Ice Shelf, Antarctica. *Journal of Glaciology* 46(155): 652–664.
- Fairbanks RG (1989) A 17,000-year glacio-eustatic sea level record: Influence of glacial melting rates on the Younger Dryas event and deep-ocean circulation. *Nature* 342: 637–642.
- Gades AM, Raymond CF, Conway H, and Jacobel RW (2000) Bed properties of Siple dome and adjacent ice streams, West Antarctica, inferred from radio-echo sounding measurements. *Journal of Glaciology* 46(152): 88–94.
- Hulbe CL and Payne AJ (2001) The contribution of numerical modeling to our understanding of the West Antarctic ice sheet. *Antarctic Research Series* 77: 201–219.
- Jacobson HP and Raymond CF (1998) Thermal effects on the location of ice stream margins. *Journal of Geophysical Research* 103(B6): 12,111.
- Joughin IL, Gray R, Bindschadler S, et al. (1999) Tributaries of West Antarctic ice streams revealed by Radarsat interferometry. *Science* 286: 283–286.
- Joughin I, Tulaczyk S, Bindschadler RA, and Price SF (2002) Changes in West Antarctic ice stream velocities. *Journal of Geophysical Research* 107(B11): 2289. <http://dx.doi.org/10.1029/2001JB001029>.
- Kamb B (2001) Basal zone of the West Antarctic ice streams and its role in lubrication of their rapid motion. In: Alley RB and Bindschadler R (eds.) *Antarctic Research Series: The West Antarctic Ice Sheet: Behavior and Environment*, vol. 77, 157–199.
- Nereson NA and Raymond CF (2001) The elevation history of ice streams and the spatial accumulation pattern along the Siple Coast of West Antarctica inferred from ground-based radar data from three inter-ice-stream ridges. *Journal of Glaciology* 47(157): 303–313.
- Rignot E and Schmelz M (2002) Acceleration of Pine Island and Thwaites Glaciers, West Antarctica. *Annals of Glaciology* 34: 189–194.
- Scambos TA, Hulbe C, Fahnestock M, and Bohlander J (2000) The link between climate warming and breakup of ice shelves in the Antarctic Peninsula. *Journal of Glaciology* 46(154): 516–530.
- Scherer RP (1991) Quaternary and Tertiary microfossils from beneath ice stream B: Evidence for a dynamic West Antarctic ice sheet history. *Palaeogeography, Palaeoclimatology, Palaeoecology* 90(4): 395–412.
- Shepherd A, Wingham DJ, and Rignot E (2004) Warm ocean is eroding West Antarctic ice sheet. *Geophysical Research Letters* 31(23) Art. no. L23402 DEC 9 2004.
- Smith BE, Lord NE, and Bentley CR (2002) Crevasse ages on the northern margin of ice stream C, West Antarctica. *Annals of Glaciology* 34: 209–216.
- Steig EJ, Fastook JL, Zweck C, et al. (2001) West Antarctic ice sheet elevation changes. In: Alley RB and Bindschadler R (eds.) *Antarctic Research Series: The West Antarctic Ice Sheet: Behavior and Environment*, 75–90.
- Thomas RH (1973) The creep of ice shelves: Theory. *Journal of Glaciology* 12(64): 45–53.
- Thomas R, and 17 others (2004) Accelerated sea-level rise from West Antarctica. *Science* 306(5694): 255–258, October 8 2004.
- Weertman J (1974) Stability of the junction of an ice sheet and an ice shelf. *Journal of Glaciology* 13(67): 3–11.
- Wingham DJ, Ridout AJ, Scharroo R, Arthern RJ, and Shum CK (1998) Antarctic elevation change from 1992 to 1996. *Science* 282: 456–458.

Dynamics of the East Antarctic Ice Sheet

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The Antarctic Ice Sheet ([Figure 1](#)), Earth's largest glacier, forms the highest-elevation continental surface on Earth. The East Antarctic portion of this ice sheet is located primarily in the eastern hemisphere between longitudes 30° W and 165° E; it flows into and is contiguous with the smaller and lower-elevation West Antarctic Ice Sheet, which is located primarily in the western hemisphere. These two components of the ice sheet are largely dynamically independent, because they are separated by the Transantarctic Mountains. The West Antarctic Ice Sheet is a 'marine ice sheet', grounded well below sea level. (See the article on [Dynamics of the West Antarctic Ice Sheet](#) by Robert Bindshadler in this encyclopedia.) By contrast, the much larger East Antarctic Ice Sheet is grounded on continental crust that is mostly, although not entirely, above sea level.

The East Antarctic Ice Sheet reaches elevations exceeding 4000 m above sea level at Dome Argus near its center ([Figure 2a](#)), has an area of 9 855 570 km², an estimated volume of 25 920 000 km³, a mean thickness of ca. 2630 m, and a maximum thickness of 4780 m ([Drewry, 1983](#)). The ice sheet rests on continental crust ([Figure 3](#)) that includes large subglacial basins and several mountain ranges, some of which are entirely subglacial. Some regions of the bed are below sea level, because the weight of the ice sheet has caused depression of the lithosphere, forcing viscous rock in the asthenosphere to flow out from beneath the ice load. [Figure 4](#) shows the distribution of ice thickness. Many of the details of ice flow remain unknown, but satellite remote sensing is yielding steadily-increasing information (e.g., [Bamber et al. \(2000\)](#)). Here, we describe some general features of the flow field of the ice sheet that can be inferred from glaciological principles.

The East Antarctic Ice Sheet, like all glaciers, flows in response to gravity. 'Ice-sheet dynamics' refers to the ways in which the ice sheet deforms, flows, and changes through time in response to changes in accumulation associated with changing climate, to changes in ocean temperatures, and to changes in relative sea-level at the margins and grounding lines. In a depth-averaged sense, ice flows down the surface slopes, from higher to lower surface elevations. Thus, to a good approximation, a map of ice-flow directions can be made from a surface elevation map because the ice flows at 90 degrees to the elevation contours.

At any given depth in an ice sheet, the ice above flows faster than the ice below. This vertical gradient in the flow velocity causes shear deformation ([Figure 5a](#)). The magnitude of this shear deformation is maximum at and immediately above the bed, and decreases with height above the bed. The flow velocity thus increases most rapidly immediately above the bed. It increases more slowly at greater heights above the bed, reaching a maximum at the surface. Like other viscous fluids, glacier ice flows faster when its surface slope is steeper, and when its thickness is greater.

The constitutive properties or 'stiffness' of ice are described by a flow law, which quantifies the relationship between deformation rates and the stress field in three dimensions. This relationship is strongly nonlinear; ice sheets are modeled as power-law fluids. The flow law of ice is also strongly temperature-dependent; ice is softer and thus deforms more rapidly at higher temperatures.

Ice sheets can also slide over their substrates if the basal ice is at the pressure-melting temperature and if sufficient water is present, from basal melting, to lubricate the interface. If the substrate is relatively smooth and soft; for instance, water-saturated subglacial till or mud, or a subglacial lake, then basal motion can be enhanced further. [Figure 5b](#) shows typical trajectories followed by ice particles in a vertical section as they flow from the ice-sheet interior toward the coast. Hooke (2004) and [Paterson \(1994\)](#) offer in-depth mathematical and theoretical treatments of both ice deformation and basal motion.

The ice sheet is formed from precipitated snow. The snow compacts to form glacial ice over time scales of hundreds of years, due to the overlying weight of accumulating snow. East Antarctica is a climate desert; the net accumulation of snow, expressed as ice-equivalent thickness, is less than 10 cm annually over large areas ([Figure 6](#)). On the steeper slopes near the coast, orographic lifting of maritime air leads to higher precipitation rates. A year or more can pass without snowfall in the driest regions. Much snow is drifted by the wind and some is blown from the ice sheet. The small amounts of snow that fall are largely preserved, because low air temperatures, even in summer, preclude melting on the surface of the East Antarctic Ice Sheet, with the exception of some small regions near the margins. [Figure 7](#) shows the mean annual surface temperatures on the ice sheet. In Antarctica, as elsewhere, Earth's crust is heated by decay of radioactive elements. This geothermal energy from the solid Earth must ultimately be conducted upward through the ice, from the warm rock toward the colder atmosphere, and this heat transfer requires a temperature gradient. This geothermal gradient causes temperatures to be much warmer at the base of the ice sheet than at its upper surface. Additional heat can be generated by shear deformation rates within the ice. Shear deformation rate is generally greatest in the basal ice ([Figure 5a](#)). The basal ice is warmer and is therefore softer than the ice above; this can further enhance shear deformation in the basal ice. If the combination of geothermal energy and deformational heat causes the basal ice to reach the pressure-melting temperature, this can, in turn, lead to sliding, or to basal motion through deformation of a soft water-saturated substrate. The frictional heat of basal motion can in turn generate more melting, further enhancing basal motion. The highest rates of basal melting occur beneath the Antarctic ice shelves, which are floating seaward extensions

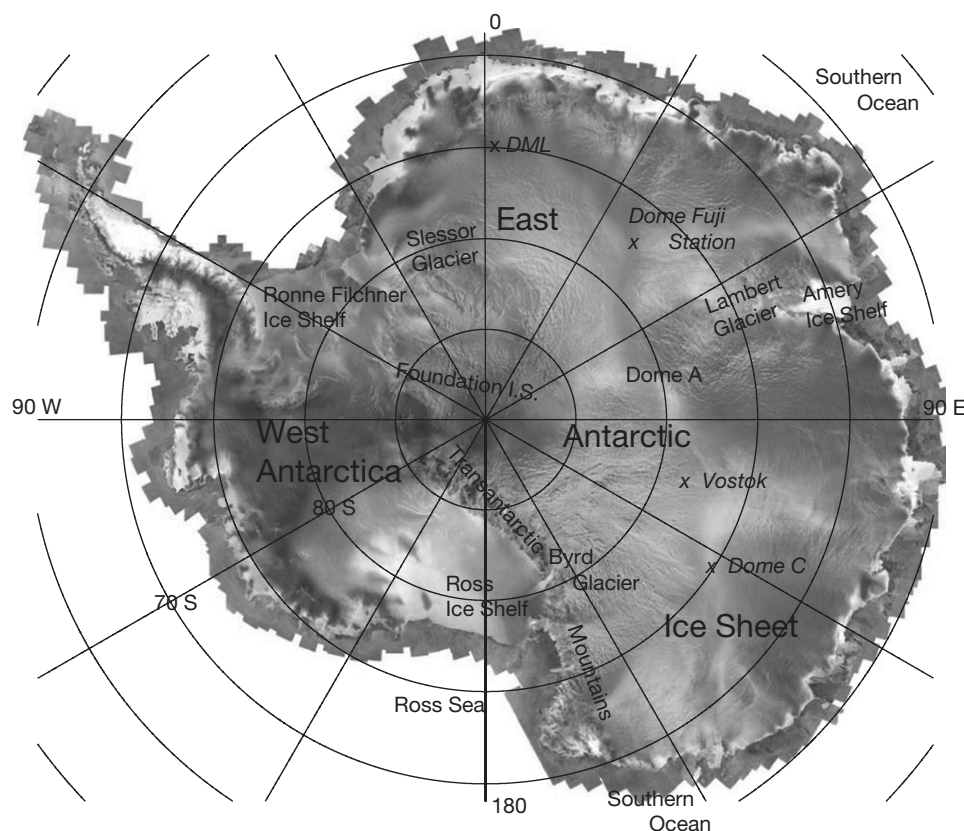


Figure 1 The East Antarctic Ice Sheet covers the area between longitudes 165°E and 30°W and is bounded by the Southern Ocean and the Transantarctic Mountains. Byrd Glacier, Lambert Glacier, Slessor Glacier, and Foundation Ice Stream are examples of major outlet glaciers. Ross, Amery, and Ronne-Filchner are the major ice shelves. DML (Kohnen Station, Dronning Maud Land), Dome Fuji Station, Vostok, and Dome C are sites of deep ice cores. (Antarctic image from Radar Antarctic Mapping Program (RAMP), courtesy of Alaska Satellite Facility.)

of the grounded ice sheet, where the oceanic heat flux can be much higher than the geothermal flux.

The East Antarctic Ice Sheet has been present for ca. 40 Myr; however, ice continually cycles through the ice sheet on much shorter time scales. As illustrated in [Figure 5b](#), ice accumulates in the high interior, flows toward the margins, and leaves the continent as icebergs, which calve from coastal glaciers or from ice shelves. Over time, the ice sheet has adjusted its shape to establish a near-equilibrium state, in which its thickness, slope, and internal and basal temperatures combine to cause basal motion and flow by internal deformation that can transport, in an average year, just the correct amount of ice to balance the total amount of accumulation that falls upstream in an average year. Glaciologists often define a flow band as the portion of the ice sheet bounded by a pair of flow lines on the surface, as shown in [Figure 2b](#). Ice is unlikely to cross through the vertical surfaces beneath each flow line. In three dimensions (3-D) ([Figure 8](#)), ice accumulates on the upper surface of the flow band by snowfall, and ice flow carries ice through the vertical section between the two flow lines. This vertical surface can be called a “gate”. In a steady state, the volume of ice transported through the gate in 1 yr is equal to the volume of ice accumulated upstream on the flow-band surface in 1 yr. The average velocity of ice passing through the gate is called the balance velocity U . Because the largest velocities are generally found at

the surface ([Figure 5a](#)), the balance velocity is smaller than the actual ice velocity that would be measured at the ice-sheet surface; however, their relationship can be estimated using ice-flow theory to estimate the contribution of basal motion, and the rates of internal deformation (see [Hooke \(2005\)](#) and [Paterson \(1994\)](#)). [Figure 9](#) shows the balance-velocity distribution on the ice sheet. Although we have very little information on the actual velocities due to internal deformation and sliding in East Antarctica, the balance velocity can generally be used as a good approximation to their depth-averaged sum. Clearly, the flow is slow in the interior, but speeds up dramatically as ice is funneled into a few major outlet glaciers such as Byrd Glacier ([Figure 1](#)) where the ice flows rapidly through gaps in the Transantarctic Mountains. Even where the ice sheet abuts the Southern Ocean, flow is concentrated into a few large fast outlet glaciers, because fast flow causes erosion of deep bedrock troughs, and these troughs in turn encourage faster flow. Some ice is also discharged through ice streams, which are fast-flowing currents within the more sluggish mass of the surrounding ice sheet. Foundation Ice Stream ([Figure 1](#)) is one example.

The dynamics of the ice sheet also control both the distribution of age within the ice, and the increasing degree of vertical thinning and horizontal stretching that layers of ice have experienced by the time that they arrive at each depth, at

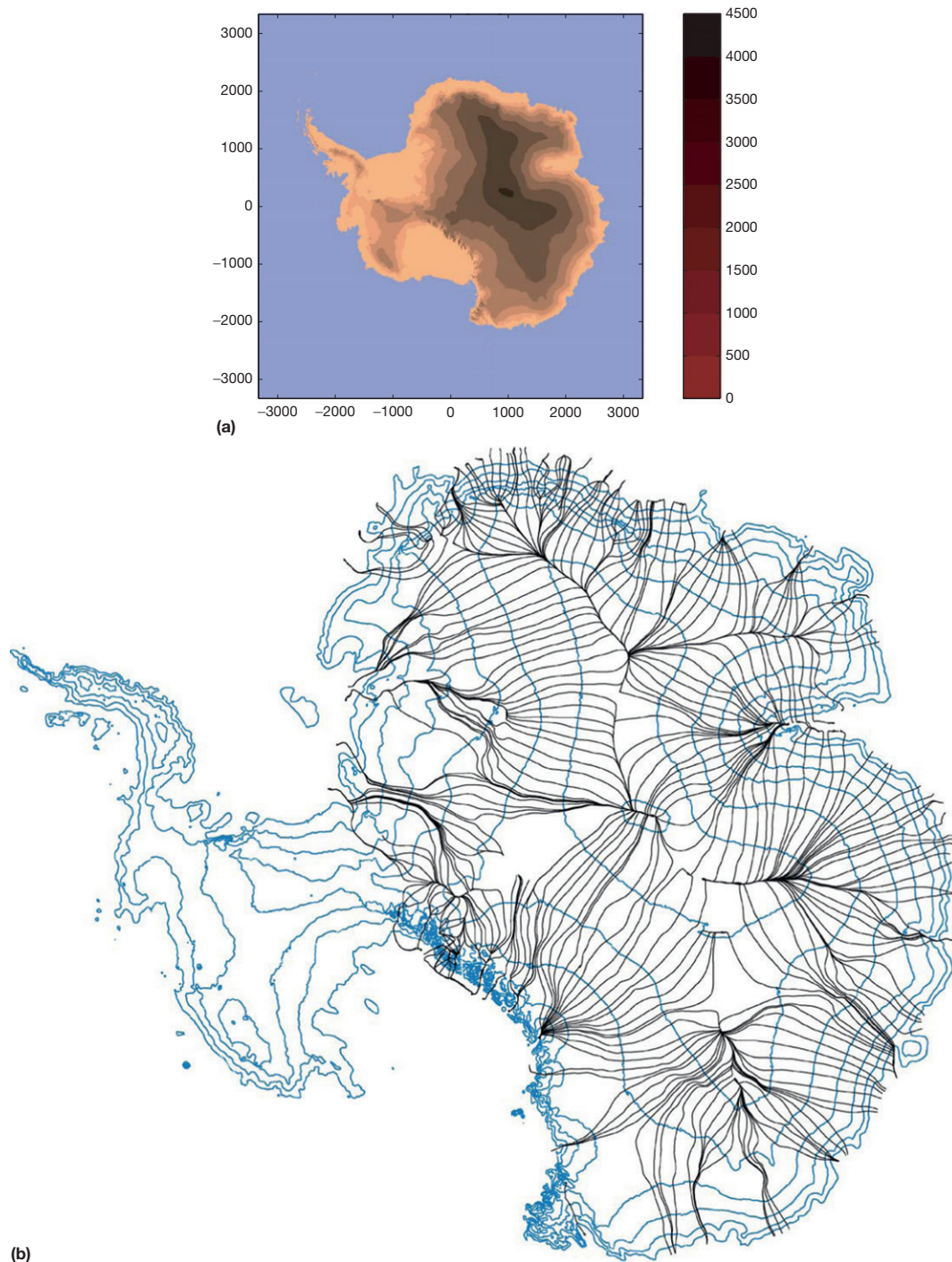


Figure 2 (a) Surface topography of Antarctica, in meters. The East Antarctic Ice Sheet reaches an elevation greater than 4000 m at Dome A. (Figure courtesy of J. Kallen-Brown. Data set courtesy of D. Vaughan.) (b) Flow lines (black) from the interior to the coast are approximated by steepest-descent paths, at 90° to the elevation contours (blue). Most of the ice discharge from East Antarctica is funneled into fast-moving outlet glaciers. (Figure courtesy of I. Joughin.)

each location on the ice sheet. It is possible to estimate the age distribution with depth that might be expected in an ice sheet, following a simple thought experiment. First, imagine building up an ice sheet to thickness H from nothing, by depositing successive layers of ice with thickness b , one layer each year. Furthermore imagine that these layers do not flow or deform.

As **Figure 10a** illustrates, the time to build the ice sheet is then $\tau = H/b$. For central East Antarctica, H is ca. 3 km, and b is ca. 0.05 m yr^{-1} or less when averaged over 100-ka Ice Age cycles. Under these conditions, this characteristic time τ is ca. 60 ka.

Because the ice sheet is actually flowing, and because the speed increases with distance from an ice divide, stretching

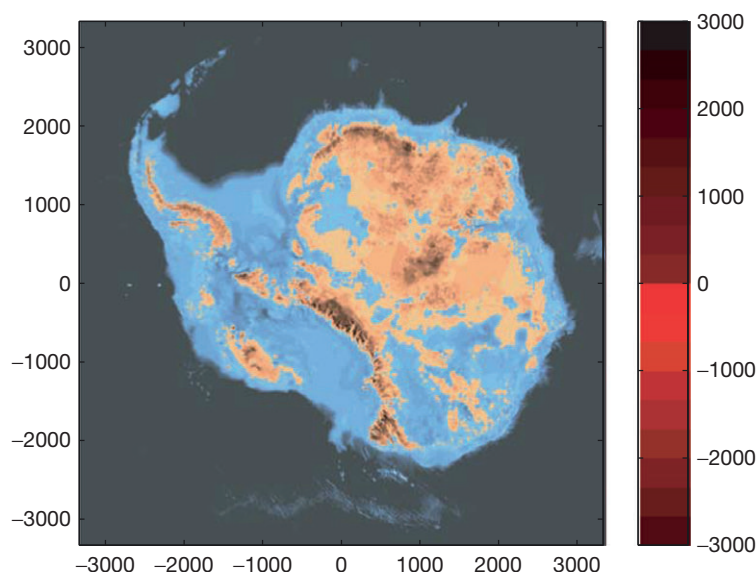


Figure 3 Bed topography under the Antarctic ice sheets, in meters above sea level. While several large basins are currently below sea level, peaks of the Gambertsev Subglacial Mountains below Dome A (Figure 1) approach 3 km above sea level. (Figure courtesy of J. Kallen-Brown. Data set courtesy of D. Vaughan.)

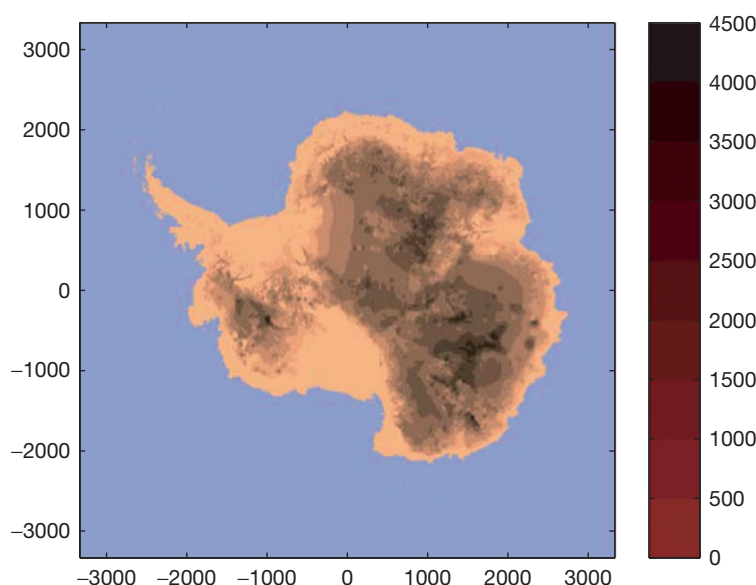


Figure 4 Ice thickness map, (ice thickness in meters). The most extensive areas of very thick ice (in excess of 4 km) are found west of Dome C (Figure 1). (Figure courtesy of J. Kallen-Brown. Data set courtesy of D. Vaughan.)

layers along the direction of flow, a layer gets progressively thinner with time (Figure 10b), and typically an order of magnitude more layers can be stacked into the depth H , suggesting that ice cores should recover 600-ka climate records. Cores from Vostok, Dome C, Dome Fuji Station, and Dronning Maud Land (Figure 1) all recovered climate records going back on the order of a half million years, confirming the validity of this simple estimate based on ice dynamics.

Because of its extreme cold climate and limited melting, the East Antarctic Ice Sheet has been more stable than other Earth's other ice sheets. Ice sheets can lose mass rapidly if accelerated

flow of ice streams and outlet glaciers significantly exceeds upstream accumulation. This dynamical thinning can draw down the inland ice much faster than passive melting alone could achieve. This appears to be the way that former ice sheets have disappeared during times of warming climate. Ice shelves and floating glacier tongues seaward of outlet-glacier grounding lines appear to provide a buttressing restraint. They also protect the fronts of those glaciers from warm seawater. Acceleration of some outlet glaciers appears to be initiated primarily by thinning and breakup of such ice shelves and floating glacier tongues (Scambos et al., 2000). In many outlet glaciers,

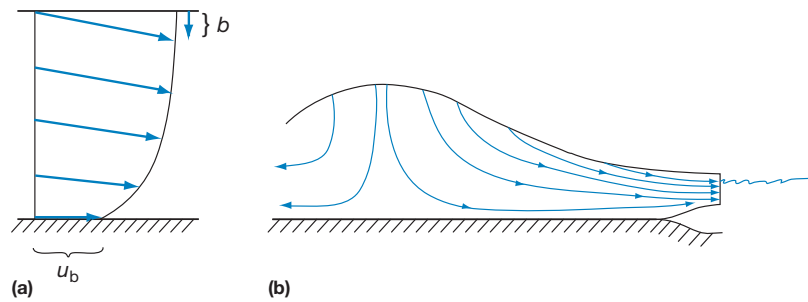


Figure 5 (a) Pattern variation of ice-flow velocity with depth in the East Antarctic Ice Sheet. The ice may slide along the basal substrate (bedrock or soft sediments) as indicated by u_b . Ice also experiences internal deformation or shearing flow, in which the ice above any level flows faster than the ice below; the rate of shearing is greater at greater depths. Because the ice sheet is close to a steady state, the ice at the surface at any site also moves downward at a rate that balances the annual accumulation rate b . The downward velocity diminishes with depth; at the bed, the bed-convergent motion just balances the basal melting rate, which typically varies between zero and a few millimeters annually. (b) Flow pattern along a typical flow line from the ice sheet interior to the coast. Where the water depth exceeds 9/10 of the ice thickness, the ice floats and is called an ice shelf.

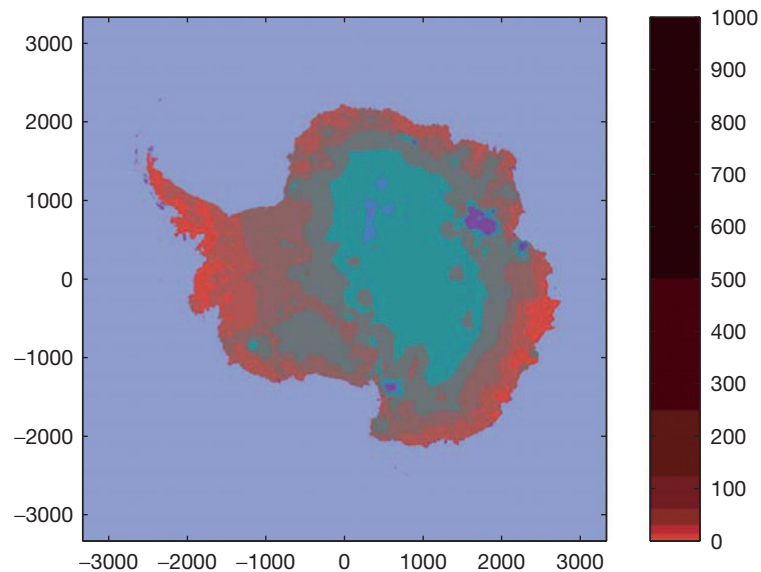


Figure 6 Accumulation map (accumulation rate (ice equivalent) in units of mm per year). The interior of East Antarctica is a desert. (Figure courtesy of J. Kallen-Brown. Data set courtesy of D. Vaughan.)

fast flow is due primarily to fast basal motion. Ice discharge through these glaciers can increase rapidly if faster flow then produces more water at the bed. Because of this positive feedback, outlet glaciers control the future of ice sheets.

Some ice shelves have been lost on the Antarctic Peninsula in recent decades (e.g. Scambos et al., 2000). Some outlet glaciers on the Greenland Ice Sheet and the West Antarctic Ice Sheet (see Bindshadler, *this Encyclopedia*) are currently undergoing dramatic accelerations in response to warming climate. Although the East Antarctic Ice Sheet today appears to be close to steady-state, it also has buttressing ice shelves, large fast outlet glaciers, and large areas where the bed is below sea level. If increased basal melting of ice shelves, for example due to increasing temperatures in the Southern Ocean, were to cause thinning and breakup of the ice shelves, the resulting reduction of buttressing effect could cause acceleration of the ice streams

and outlet glaciers that feed the ice shelves. These processes, if they were to occur, could cause dynamic thinning of the inland ice sheet when ice discharge exceeded replacement by accumulation. The East Antarctic Ice Sheet too could respond rapidly to climate change, if its ice shelves should fail in a warmer world.

Future measurements promise to produce a more-detailed picture of the dynamics of the East Antarctic Ice Sheet. Satellite remote sensing using interferometric Synthetic Aperture Radar (inSAR) is producing progressively better maps of ice-flow velocities at the surface. Satellite altimeters (both laser and radar) are revealing transient flow and vertical motions related to changes in the accumulation pattern (climate), and to changes in basal conditions such as transient water storage (e.g. Wingham et al., 2006). Subtle changes in the gravitational field over Antarctica based on orbits of tandem satellites such

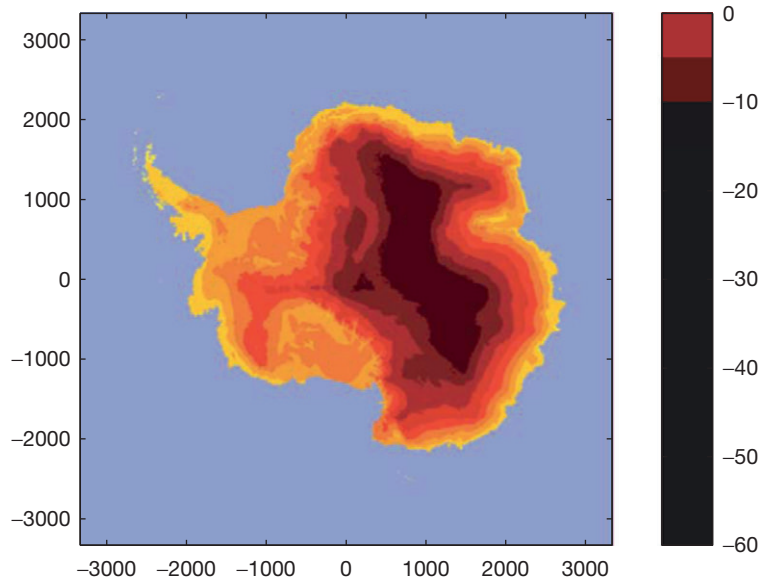


Figure 7 Mean annual temperature at the ice-sheet surface (in °C). The interior of East Antarctica is the coldest region on Earth. (Figure courtesy of J. Kallen-Brown. Data set courtesy of D. Vaughan.)

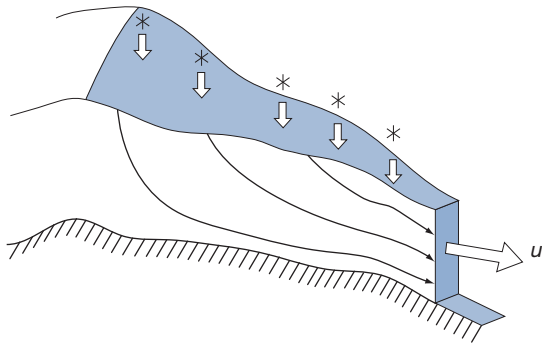


Figure 8 When a gate is defined as a vertical section between two flow lines on the ice-sheet surface, and transverse to the ice flow, the balance velocity U is the velocity needed to carry a volume of ice through that gate in 1 yr that is equal to the volume of ice accumulating upstream from the gate and between those two flow lines in 1 yr.

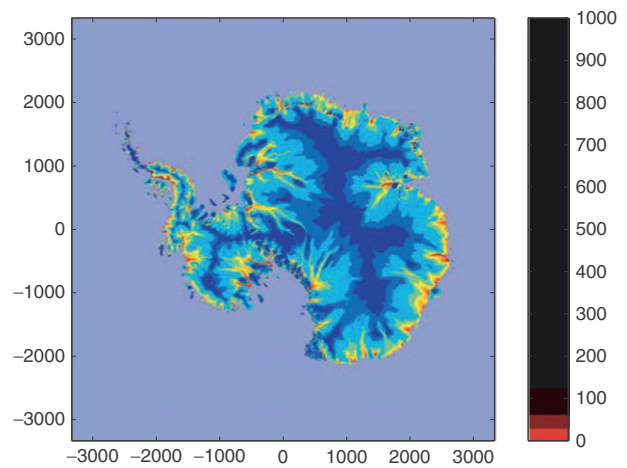


Figure 9 Balance-velocity map (in m per year). Ice in the interior flows slowly, but as it approaches the coast it is funneled into fast-moving outlet glaciers such as Byrd Glacier, Lambert Glacier, Slessor Glacier, and Foundation Ice Stream (see Figure 1). (Figure courtesy of J. Kallen-Brown. Data set courtesy of D. Vaughan.)

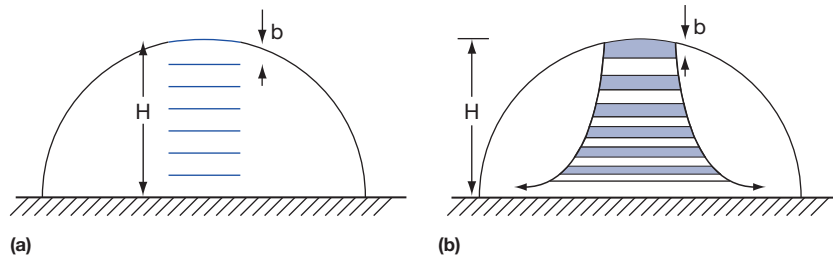


Figure 10 (a) By considering how many years it would take to build an ice sheet of thickness H by stacking up the annual accumulation b in the absence of flow, a characteristic time scale $\tau = H/b$ for growth of a polar ice sheet can be derived. (b) Real ice sheets flow, and the layers are stretched and thinned as they get older. The same time scale $\tau = H/b$ is characteristic of ice deformation, giving a rough estimate of the time required for a layer to thin by 50%. An ice core can be expected to recover a paleoclimate record spanning approximately 10τ .

as the GRACE mission (Wahr et al., 2000) can detect changes in ice mass; however the short temporal baseline and ongoing isostatic rebound are complications still being resolved. Future measurements in the ice sheet itself will also improve knowledge and understanding of the ice dynamics. For example, there is considerable interest in development of a coiled-tubing hydraulically-driven drill system to make low-cost access holes. Holes could be drilled in a few days, and kept open indefinitely with appropriate environmentally friendly, low-density anti-freeze fluids. Such holes would permit measurements of geothermal flux and temperature distribution within the ice sheet, and direct measurements of the pattern of internal deformation and sliding (Figure 5a). By offering direct access to the subglacial bed, holes would also permit measurements of the properties of subglacial materials that control the dynamics of basal motion.

See also: **Glaciations:** Early Quaternary (Pleistocene) and Precursors; Late Quaternary of Antarctica. **Ice Core Records:**

Greenland Stable Isotopes. **Ice Cores:** Dynamics of the Greenland Ice Sheet; Dynamics of the West Antarctic Ice Sheet.

References

- Bamber JL, Vaughan DG, and Joughin I (2000) Widespread complex flow in the interior of the Antarctic Ice Sheet. *Science* 287: 1248–1250.
- Bindschadler, R. A. (this Encyclopedia). Dynamics of the West Antarctic Ice Sheet.
- Drewry DJ (1983) *Antarctica: Glaciological and Geophysical Folio*. Cambridge, UK: Scott Polar Research Institute.
- Hooke R and Le B (2005) *Glacier Mechanics*, 2nd edn. Cambridge UK: Cambridge U.P.
- Paterson WSB (1994) *The Physics of Glaciers*, 3rd edn. Oxford, UK: Pergamon.
- Scambos TA, Hulbe C, Fahnestock M, and Bohlander J (2000) The link between climate warming and breakup of ice shelves in the Antarctic Peninsula. *Journal of Glaciology* 46(154): 516–530.
- Wahr J, Wingham D, and Bentley C (2000) A method of combining ICESat and GRACE satellite data to constrain Antarctic mass balance. *Journal of Geophysical Research* 105(7): 16,279–16,294.
- Wingham DJ, Siegert MJ, Shepherd A, and Muir AS (2006) Rapid discharge connects Antarctic subglacial lakes. *Nature* 440(4660): 1033–1036.

History of Carbon Monoxide and Ultra-Trace Gases from Ice Cores

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Introduction

Air trapped in polar ice cores has proven to be a valuable archive for determining paleoatmospheric levels of trace gases such as carbon dioxide (CO₂), methane (CH₄), and nitrous oxide (NO₂). The ultra-trace gases (e.g., carbon monoxide, nonmethane hydrocarbons, carbonyl sulfide, methyl chloride) have weaker emissions and are removed faster from the atmosphere, resulting in abundances as low as a few parts per trillion (ppt) in the most extreme cases. Despite very low atmospheric abundances, ultra-trace gases have a profound influence on the chemistry of the atmosphere and impact global climate, drawing interest from a wide range of scientific disciplines. The ambient mixing ratios of many ultra-trace gases are routinely measured as part of atmospheric sampling networks (e.g., <http://www.esrl.noaa.gov/gmd>). Establishing the past atmospheric variability of ultra-trace gases will contribute to our understanding of the feedback mechanisms between the global biogeochemical systems and the climate, as well as providing opportunities to observe past variability in the chemistry of the atmosphere.

Ice core ultra-trace gas measurements present many analytical challenges. Ice cores contain only ~10% air by volume. Using large amounts of ice is not feasible because of both practical limitations on extraction techniques and the fact that these procedures require destructive sampling of rare and valuable ice cores. In most cases, ice core gas analyses are carried out with considerably smaller samples than typically used in atmospheric measurements. As a result, even minor levels of contamination can present significant problems and lead to experimental artifacts. Contaminants can be introduced during drilling, transport, storage, sampling, extraction, and analysis phases. Implementing clean handling procedures and using the right materials for the application throughout these phases are often the most critical components in ultra-trace gas measurements.

Potential causes of measurement artifacts are not limited to external contaminants. Chemically reactive gases can be produced or lost while still in the snow, in the firn before getting trapped in ice cores, or within the ice core matrix. Such *in situ* effects can be very difficult to identify. Direct verification of the ice core measurements with contemporaneous real-time atmospheric data can be difficult to achieve. Shallow ice cores from high accumulation rate sites contain air from the second half of the twentieth century and represent the only means for direct verification of ice core measurements. Often, confirming the fidelity of the ice core ultra-trace gas records requires an indirect approach, in which determining the atmospheric composition of the recent past through analysis of firn air plays a crucial role. Firn air gas records can overlap with both the real-time atmospheric measurements and the youngest (most recent) time periods that can be studied with the ice core records. Ultimate confirmation requires agreement between multiple firn air and ice core records.

So far, only a handful of ultra-trace gases have been successfully measured in ice cores. The following sections include a review of the published literature on ice core measurements of four such ultra-trace gases: carbon monoxide, carbonyl sulfide, methyl chloride, and methyl bromide. The research is still in its preliminary stages. In this context, successful measurements refer primarily to confidence in the analytical procedures and the data thus generated. Much progress is needed on the verification front. There is reasonably good evidence that the data presented in this chapter represent the true composition of the past atmosphere; however, this issue is an open question and will remain so until the existing records can be reliably reproduced with measurements from multiple ice cores.

Carbon Monoxide

Carbon monoxide (CO) is one the most important ultra-trace gases in the atmosphere. It is readily removed from the atmosphere via oxidation with the hydroxyl radical (OH) and is considered a short-lived gas. OH is the most important oxidant in the atmosphere and CO oxidation is the primary removal mechanism for it. The oxidation chemistry of CO also affects other atmospheric oxidants such as ozone (O₃) and hydroperoxy radical (HO₂). Consequently, atmospheric CO levels have considerable influence on the overall oxidative capacity of the atmosphere and indirectly impact the abundance of most trace and ultra-trace gases (Bakwin et al., 1994; Novelli et al., 1998).

CO is emitted from a variety of anthropogenic and natural sources, which include fossil fuel combustion, oxidation of methane and nonmethane hydrocarbons, biomass burning and biofuel use, and to a lesser degree, plants and oceans (Duncan et al., 2007; Khalil and Rasmussen, 1988, 1990). Anthropogenic emissions, which comprise more than half of the total CO emissions today, occur mostly in the northern hemisphere. The uneven distribution of sources coupled with a short lifetime results in large interhemispheric gradients. At the end of the twentieth century, the annual mean CO mixing ratios in the high latitudes of the northern and southern hemispheres were ~120 and ~50 ppb, respectively (Novelli et al., 1998, 2003). In the absence of anthropogenic emissions, source-driven variations in atmospheric CO levels would likely be due to changes in biomass burning emissions and atmospheric CH₄ levels. Any change in the oxidative capacity of the atmosphere would also impact the abundance of CO in the preindustrial atmosphere.

CO has been measured successfully in polar firn air from Berkner Island, Antarctica (Figure 1). The top 40 m of the firn column at Berkner Island is impacted by seasonal CO variability at the surface. Below 40 m, the CO levels drop from ~52 ppbv (parts per billion by volume) at 50 m to ~46 ppbv at 59 m. It is necessary to use a firn air model to develop an atmospheric history based on these measurements. Based on the modeling results, the Berkner Island data imply that the CO levels over

Antarctica were likely lower than 40 ppbv at the beginning of the twentieth century and rose to above 50 ppbv by AD 2000 (Assanov et al., 2007). The magnitude and the timing of inferred changes are in agreement with what is known about the sources of atmospheric CO. Overall, the Berkner Island firn data establish a link between the real-time measurements of the present day atmosphere and the ice core measurements. There is no indication that CO levels are altered via *in situ* chemistry in the firn, suggesting that ice cores likely contain a true atmospheric record.

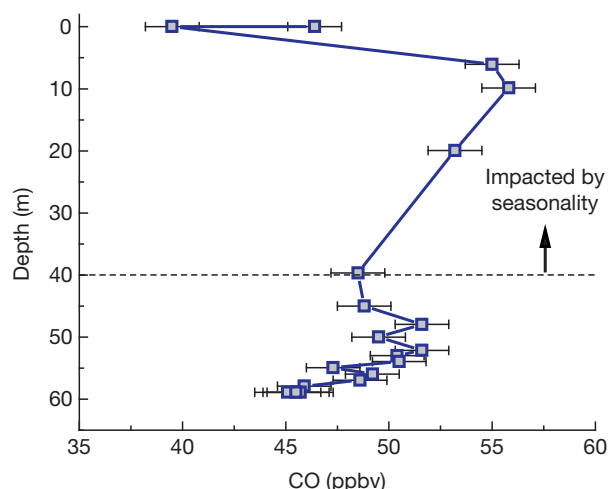


Figure 1 CO measurements in firn air from Berkner Island, Antarctica (Assanov et al., 2007). Data from the top 40 m are impacted by seasonal CO variations at the surface. The data between 40 and 60 m store information on how the annual mean CO levels changed over Antarctica during the twentieth century.

CO has been measured in multiple ice cores from Greenland and Antarctica, using a 'wet' extraction technique (Haan and Raynaud, 1998; Haan et al., 1996). The Greenland measurements (Figure 2 – top panel) are all from the dry-drilled (no drilling liquid) Eurocore from Summit (72.6° N, 37.5° W). During the time period before AD 1650, the measurements display high scatter and remain above 100 ppbv. After AD 1650, CO levels remain at 90–100 ppbv until 1850, then gradually rise to 110 ppbv in AD 1950. There is considerably less scatter in the data for the AD 1650–1950 period and the CO rise between AD 1850 and 1950 is consistent with the timing of the methane increase due to increasing anthropogenic activities. Of the two discrete groups in the Greenland CO record, only the period since 1650 appears to reflect true atmospheric variations. The ice core samples with elevated CO levels contain more total organic material than the ones with no excess CO. The apparently successful recovery of a true atmospheric record from the shallower depths of this core with identical experimental procedures suggests that oxidation of organic matter contained within the ice is the likely cause of the CO elevation (Haan and Raynaud, 1998).

The Antarctic CO measurements are from three different cores dry-drilled at two different sites in East Antarctica (Figure 2). The younger samples for years AD 1850–1920 are from the D47 ice core (67.2° S, 154.0° E). The older samples that predate AD 1400 are from two different Vostok ice cores (78.5° S, 106.9° E). The CO mixing ratio in the Antarctic samples mostly stays within 50–60 ppbv, with averages of 57 ± 1 ppbv ($\pm 1\sigma$) and 51 ± 2 ppbv in D47 and Vostok samples, respectively (Haan and Raynaud, 1998). Low sample-to-sample variability and the close agreement between the two data sets suggest the Antarctic CO record reflects real atmospheric variations, although it is not yet possible to rule

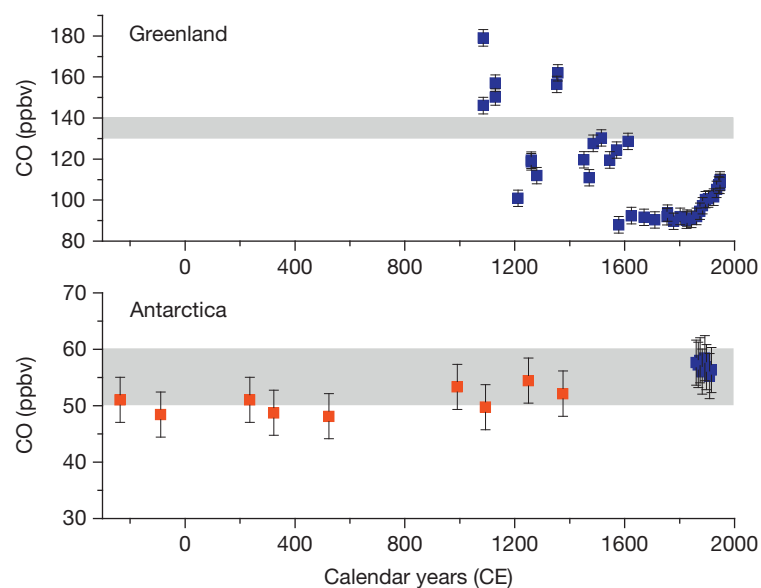


Figure 2 CO measurements in polar ice cores from Greenland (upper panel) and Antarctica (lower panel). Upper panel – CO data (blue squares) from the Eurocore ice core dry-drilled at Summit, Greenland (Haan and Raynaud, 1998; Haan et al., 1996). The gray band represents the CO levels in the modern-day high-latitude northern hemisphere. Lower panel – CO measurements in the D47 ice core from East Antarctica (blue squares – Haan et al., 1996) and in two different ice cores drilled at Vostok, East Antarctica (red squares – Haan and Raynaud, 1998). The gray band represents the CO levels in the modern-day high-latitude southern hemisphere.

out artifactual enhancement. Antarctic ice cores generally contain fewer impurities (organic and inorganic) than the Greenland ice cores because Antarctica is more isolated from other continents. Consequently, Antarctic ice cores are better candidates for measurements of ultra-trace gases such as CO.

Considering the substantial rise in atmospheric CH₄ during the twentieth century, it is interesting that the mean CO levels in the Antarctic record are mostly within the range measured over Antarctica today. As discussed previously, the CO mixing ratio measured at the bottom of the firn at Berkner Island (Figure 1) is 45 ± 2 ppbv, which implies that the actual atmospheric levels were less than 40 ppbv around AD 1900. The difference between the firn air results and the D47 measurements for the same period is about 20 ppbv. This is larger than the reported precision of the measurements and may suggest that the ice core CO levels in the D47 core are elevated, albeit at a much lower level than what is observed in the Eurocore from Greenland. Haan and Raynaud (1998) compared the impurity levels in their Greenland and Antarctic samples and calculated that the CO mixing ratios in their Antarctic samples could contain ~ 10 ppbv of excess CO.

New ice core CO measurements are forthcoming. A recent record from the Law Dome ice cores in Antarctica (Ferretti et al., 2005) shows CO mixing ratios at ~ 50 ppbv between AD 1500 and 1700, in apparent agreement with the D47 and Vostok records. However, the CO levels in Law Dome ice cores stay above 60 ppbv between AD 0 and 1500, exceeding 80 ppbv for 400 years around AD 800 (Ferretti et al., 2005). The Vostok record is fairly coarse resolution and does not contain any samples between AD 600 and 900 (Figure 2); however, the difference between the means of these two records appears to be significant. Overall, there is circumstantial evidence for low-level artifactual CO enhancement in the Antarctic ice cores. Contemporaneous measurements from multiple Antarctic sites with different impurity levels are needed to resolve this outstanding issue.

There is every reason to expect more extensive ice core records of CO. Such measurements will likely lead to reconstruction of true CO variations in the Antarctic atmosphere during the Holocene. Most recently, a 650-year-long record of CO and stable isotopes of CO ($\delta^{13}\text{C}$ and $\delta^{18}\text{O}$) has been recovered from Antarctic ice cores, suggesting that CO changes in the preindustrial atmosphere have been dominated by large variations in biomass burning (Wang et al., 2010). The prospects of developing a continuous northern hemisphere CO record may not be as good because of the high impurity levels in Greenland ice cores. However, it is possible that a partial record can be retrieved from the Greenland ice cores, providing a baseline for the northern hemispheric mean CO levels in the preindustrial atmosphere.

Carbonyl Sulfide

Carbonyl sulfide (COS) is the most abundant sulfur gas in the troposphere. Its tropospheric lifetime is 2–4 years, allowing for mixing between hemispheres. The mean COS mixing ratio was ~ 480 ppt in the southern and 490 ppt in the northern hemispheres based on atmospheric measurements over the 5 years between 2000 and 2005 (Montzka et al., 2007). In

the stratosphere, COS oxidizes to form sulfate aerosols, which reduce the amount of solar radiation that reaches the earth. It is estimated that COS contributes about half of the background stratospheric sulfate during periods of volcanic inactivity (Weissenstein et al., 2006), implying that large variations in the atmospheric COS levels would have an impact on global climate.

Natural COS sources include terrestrial and oceanic emissions. Anthropogenic emissions are linked to industrial production of synthetic fibers. COS is oxidized by OH in the troposphere and taken up by soils but the most significant removal process is uptake by terrestrial vegetation (Kettle et al., 2002; Montzka et al., 2007). The removal of COS by terrestrial biota has recently drawn considerable scientific interest (Campbell et al., 2008; Sandoval-Soto et al., 2005; Seibt et al., 2010; Xu et al., 2002). COS is taken up by plants during photosynthesis alongside CO₂. This process represents a net COS uptake, because unlike CO₂, COS is not respired back into the atmosphere. Consequently, ice core COS records could potentially reveal atmospheric COS variability caused by changes in global photosynthetic activity and therefore help constrain the causes of past CO₂ variations in the atmosphere.

First data on atmospheric COS variability during the periods that predate the instrumental record were obtained from polar firn air measurements at two sites in Antarctica and one in the Canadian Arctic (Sturges et al., 2001b). This study was followed up by additional firn air measurements at South Pole, Antarctica and the first COS measurements in a shallow ice core from Siple Dome, Antarctica that displayed consistency with each other and with the available instrumental data during periods of overlap (Montzka et al., 2004). These observations offer evidence that mean COS levels in the preindustrial atmosphere were 300–350 ppt and that they peaked at ~ 550 ppt during the latter half of the twentieth century, presumably as a result of rising anthropogenic emissions. In agreement with these results, first measurements in shallow Greenland ice cores indicate that the COS abundance in the northern hemisphere atmosphere between 1700 and 1900 was in the 300–350 ppt range (Aydin et al., 2007).

More recent measurements from a South Pole ice core provide a continuous COS record for the last 2000 years when combined with the earlier Siple Dome record (Figure 3). The Siple Dome and South Pole data display consistency where they overlap and most neighboring data points within both data sets exhibit agreement. These are strong indicators that the COS levels measured in these two ice cores from Antarctica reflect true atmospheric mixing ratios over the last 2000 years. There are a few cases when the measured COS mixing ratio in individual samples is elevated with respect to the overall trend and the neighboring data points. It is unlikely these outliers reflect true atmospheric levels. Instead, they may be due to intermittent contamination introduced during extraction and analysis of the ice core samples or *in situ* production of COS via a yet unidentified chemical mechanism.

Excluding the outliers, the average COS mixing ratio in the South Pole record is 310–350 ppt, with an overall increasing trend of ~ 1.8 ppt per 100 years (Aydin et al., 2008). One major concern about COS ice core measurements has been potential *in situ* loss of COS via hydrolysis in liquid-like layers

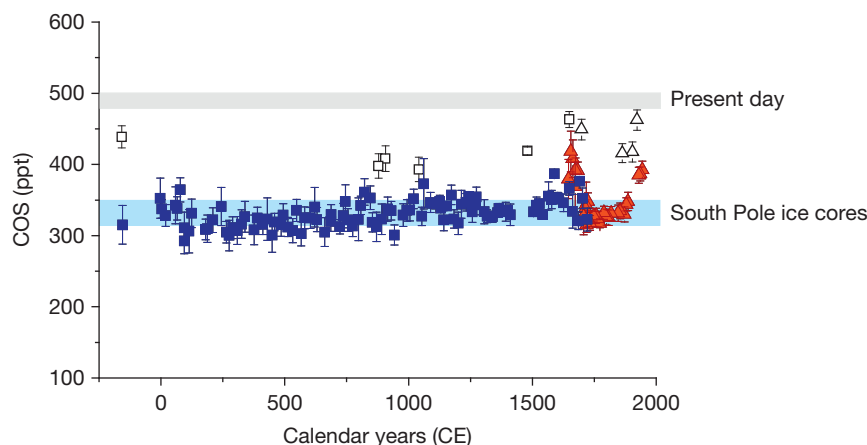


Figure 3 COS measurements in polar ice cores. The Siple Dome (red triangles) and the South Pole (blue squares) data are consistent during the period of overlap. There are outliers in both the South Pole (open squares) and Siple Dome (open triangles) data sets. The gray band represents the mean COS mixing ratio in the southern hemisphere between CE 2000 and 2005 (Montzka et al., 2007). The blue band is the mean of the South Pole record, excluding the outliers. The thicknesses of the gray and blue bands represent the estimated uncertainty in the reported means.

within the ice matrix (Assonov et al., 2005). It is possible that this long-term increasing trend is linked to slow hydrolysis of COS in the ice sheet. However, when this trend is extrapolated to the present day, the calculated COS increase between the preindustrial times and the end of the twentieth century is much smaller than what is apparent in the ice core and firn air records, providing strong evidence that anthropogenic emissions during the twentieth century had a profound impact on the atmospheric mixing ratio of this gas. The South Pole record also displays higher than average COS levels between CE 1000–1200 and CE 1500–1700, and lower than average mixing ratios between CE 100 and 700. Such anomalies may be due to changes in the biogeochemical cycling of COS as a response to variations in the global climate. Longer-term measurements from different ice cores will help determine whether these really are periods of COS variations in the atmosphere that are climatically driven. The currently available data suggest COS can be measured in ice cores from both hemispheres over millennial time scales. Extending the record to longer time scales will require more detailed studies on the hydrolysis of COS in ice cores, which is strongly temperature dependent. Measurements in contemporaneous samples with different temperature histories will likely reveal the significance of *in situ* COS loss to hydrolysis in ice cores.

Methyl Chloride

Methyl chloride (CH_3Cl) is an abundant halocarbon with a mean tropospheric mixing ratio of ~ 550 ppt. It is the largest source of chlorine in the stratosphere, thus considered an important ozone-depleting gas (Montzka et al., 2003b). Emissions from tropical vegetation and biomass burning are considered to be the two most important sources of CH_3Cl . Direct emissions linked to industrial activities are deemed insignificant and oceanic emissions are large but balanced with oceanic uptake that is almost equally large, resulting in a relatively small net source. Oxidation with OH is the most important loss pathway for CH_3Cl , with additional removal by oceanic

uptake and degradation in soils, resulting in a tropospheric lifetime of just over 1 year (Hamilton et al., 2003; Khalil and Rasmussen, 1999; Tokarczyk et al., 2003; Yokouchi et al., 2002).

There have been several CH_3Cl firn air studies at various sites in Antarctica (Aydin et al., 2004; Butler et al., 1999; Kaspers et al., 2004; Trudinger et al., 2004). The firn air records consistently displayed evidence that atmospheric CH_3Cl levels rose about 40–50 ppt during the twentieth century, despite relatively small discrepancies regarding the mean levels measured at different Antarctic sites (Trudinger et al., 2004). First measurements in a shallow ice core from Siple Dome, Antarctica, showed that mean CH_3Cl levels were ~ 500 ppt between CE 1700 and 1900, in agreement with the inferred CH_3Cl abundance in the preindustrial atmosphere based on the firn air measurements (Aydin et al., 2004).

A longer CH_3Cl record was obtained from the same South Pole ice core (Williams et al., 2007) that was used for the COS measurements shown in Figure 3. These measurements provide a continuous CH_3Cl record for the last 2000 years when combined with the earlier Siple Dome observations (Figure 4). There is consistency between the two data sets during the time of overlap between CE 1700 and 1800. There are a few outliers in the South Pole data set; the remaining measurements exhibit good sample-to-sample agreement between neighboring data points within the uncertainties, suggesting that the impact of potential experimental artifacts is minimal. The average CH_3Cl mixing ratio in the South Pole ice core is ~ 460 ppt, or 70–80 ppt lower than the annual mean levels observed in Antarctica today. The South Pole ice core data exhibit a positive trend of 3 ppt per 100 years (Williams et al., 2007). This slow increase may be a real atmospheric trend or may indicate slow degradation of CH_3Cl within the ice cores. Measurements from different sites are required to confirm the observed long-term increasing trend. The South Pole data also reveal centennial scale variability that appears to be correlated with climatic shifts during the last millennia, suggesting a warming global climate may lead to increased CH_3Cl levels in the atmosphere (Williams et al., 2007).

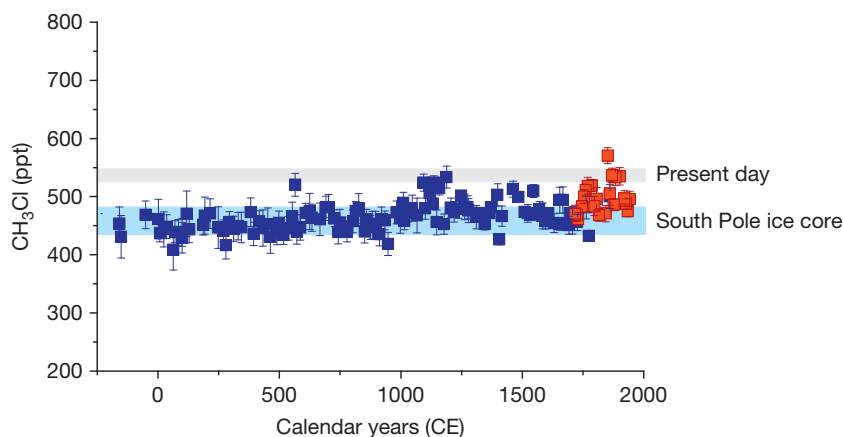


Figure 4 CH_3Cl measurements in polar ice cores. The Siple Dome (red squares) and the South Pole (blue squares) data are consistent during the period of overlap. The gray band represents the annual mean CH_3Cl mixing ratio in the southern hemisphere today (Montzka et al., 2007). The blue band represents the average of CH_3Cl measurements in the South Pole ice core. The thicknesses of the gray and blue bands represent the estimated uncertainty in the reported means.

CH_3Cl has also been measured in two deep ice cores from Siple Dome, West Antarctica, (Saltzman et al., 2009) and Dome Fuji, East Antarctica (Saito et al., 2007). These are the first ultra-trace gas records that extend through the entire Holocene and into the last glacial period (Figure 5). During the Holocene, the mean CH_3Cl level in the two deep cores is 500–600 ppt, generally exhibiting good agreement where they overlap, although there are only a few data points from the Dome Fuji ice core. When compared with the shorter late Holocene record from the South Pole ice core (Figure 4), the CH_3Cl data from the deep Siple Dome ice core are more scattered and display 50–150 ppt elevation in most samples that overlap with the South Pole record. Intermittent CH_3Cl production in basal firn at the warmer Siple Dome site may be the reason (Saltzman et al., 2009). Surprisingly, parts of a short preindustrial record measured in two shallow Greenland ice cores also indicated *in situ* production of CH_3Cl that likely occurred in the firn (Aydin et al., 2007). It is not clear whether the CH_3Cl ‘excess’ observed in Holocene ice from Greenland and Siple Dome, Antarctica, are causally linked. While Greenland ice cores contain high levels of impurities, Siple Dome is the warmest site. The potential impacts of site temperature and ice impurity levels on CH_3Cl production in the firn need to be examined with additional firn air and ice core measurements.

There are large discrepancies between the Siple Dome and Dome Fuji measurements from the last glacial period, with marked enhancements apparent in the Dome Fuji data (Figure 5). It has been shown that the excess CH_3Cl in the entrapped bubbles is correlated with elevated dust levels in the enclosing ice from the last glacial period at the Dome Fuji site (Saito et al., 2007). There is no indication of a correlation between measured CH_3Cl mixing ratio and the dust content of the enclosing ice at the Siple Dome site except possibly at the height of the Last Glacial Maximum (~25 ka BP) when the dust levels peak (Mayewski and Maasch, 2006; Saltzman et al., 2009). It should be noted that the Siple Dome ice core contains considerably less dust than the Dome Fuji ice core. Earlier during the last glacial period between 30 and 70 ka BP, Siple

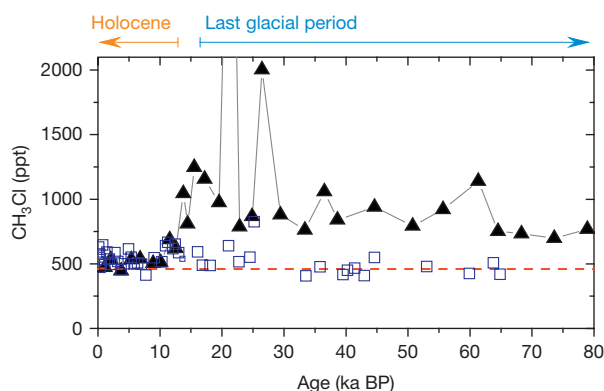


Figure 5 CH_3Cl measurements in polar ice cores. The Siple Dome (blue squares) and the Dome Fuji (black triangles) data from the Holocene are consistent during periods of overlap. The Dome Fuji CH_3Cl data from the last glacial period are considerably elevated with respect to the contemporaneous Siple Dome measurements. The red dashed line represents the mean CH_3Cl mixing ratio in the South Pole ice core shown in Figure 4.

Dome CH_3Cl measurements display low variability between adjacent samples and stay mostly at or below the late Holocene levels measured in the South Pole ice cores (Figure 5). Collectively, these results suggest *in situ* production of CH_3Cl might occur when the dust content of the ice cores exceeds a certain threshold, although the mechanisms are unclear. It cannot be ruled out that the Siple Dome measurements from 30 to 70 ka BP include a small bias due to dust related *in situ* production of CH_3Cl as the entire Antarctic continent experienced enhanced dust deposition during the last glacial period. The CH_3Cl mixing ratio measured in the Siple Dome glacial ice cores during 30–70 ka BP may represent only an upper limit.

Methyl Bromide

Methyl bromide (CH_3Br) contributes significantly to ozone depletion in the stratosphere because it is an abundant

bromine (a halogen) containing gas. It is produced naturally in oceanic and terrestrial biogeochemical systems, and during biomass burning. In addition, it has been industrially manufactured to be used as a fumigant in agriculture and other industries during the twentieth century (Clerbaux et al., 2006). It is taken up by oceans and soils, and oxidized in the atmosphere by OH, resulting in a mean atmospheric lifetime of ~ 0.7 years (Yvon-Lewis and Butler, 1997). Following the recognition of its prominent role in depletion of stratospheric ozone layer, it was included in the Montreal Protocol and its amendments. Restrictions on CH_3Br use as a fumigant started in 1998. Consistent with the timing of the regulatory actions, the atmospheric abundance of CH_3Br peaked in 1998–1999 at 10–11 ppt in the northern hemisphere and 8–8.5 ppt in the south (Montzka et al., 2003a). Since then, its atmospheric mixing ratio declined considerably, especially in the north where most of the anthropogenic emissions used to occur (Montzka et al., 2003a; Yokouchi et al., 2002; Yvon-Lewis et al., 2009).

CH_3Br measurements in firn air at four different sites in Antarctica revealed that the atmospheric mixing ratio of this gas increased by about 3 ppt during the twentieth century, with 2 ppt of the increase occurring after the 1950s. In contrast, measurements in firn air from the Canadian Arctic and Greenland displayed increasing CH_3Br mixing ratios in deeper firn, suggesting *in situ* CH_3Br production close to the base of the firn (Butler et al., 1999; Sturges et al., 2001a). These results indicate that it could be possible to develop a southern hemisphere CH_3Br record with measurements in Antarctic ice cores, but it was unlikely that the same could be done in the northern hemisphere.

The first ice core CH_3Br measurements from a shallow core from Siple Dome, Antarctica, displayed first direct evidence that the CH_3Br abundance in the pre-twentieth century atmosphere was below 6 ppt (Saltzman et al., 2004). In contrast, CH_3Br mixing ratios from two Greenland shallow ice cores were higher than the peak northern hemisphere levels in all the samples, displaying high scatter with little or no consistency between neighboring data points (Aydin et al., 2007). These first sets of measurements suggest it is highly unlikely that a northern hemisphere CH_3Br record will be recovered using ice cores from Greenland and the Canadian Arctic.

CH_3Br was analyzed in the same South Pole ice core that was used in COS and CH_3Cl measurements. The South Pole data provide a continuous CH_3Br atmospheric history for the last 2000 years when combined with the shorter Siple Dome record from Antarctica (Figure 6). The mean CH_3Br mixing ratio in the South Pole ice core is 5.4 ± 0.6 ppt ($\pm 1\sigma$), with no discernable long-term trend but variations at 10–20% level are evident over centennial time scales (Saltzman et al., 2008). This centennial variability could be driven by changes in climate or may be related to artifactual CH_3Br enhancement due to unknown factors. The lack of a long-term trend in the 2000-year-long record is strong evidence against a time-dependent chemical loss or production within the ice cores. The stability of the CH_3Br mixing ratios over the measurement periods provides a strong constraint on how low we can expect atmospheric CH_3Br levels to drop as a result of the measures taken under the Montreal Protocol. Longer records from Antarctic ice core records are needed to investigate the impacts of significant changes in global climate on the atmospheric CH_3Br levels and the state of the stratospheric ozone layer.

Ice Core Air Extraction Methods

A ‘wet’ extraction technique was used for the CO measurements (Haan and Raynaud, 1998; Haan et al., 1996). Prior to the extraction, the outer surface of an ice cores sample (~ 50 g) is shaved off for eliminating potential contaminants on the surface. The sample is then placed under vacuum in a glass extraction vessel. During the first stage of the extraction, the ice core sample is melted by immersing the glass vessel in a hot water bath. While the sample melts, the air trapped inside the sample collects in the headspace inside the glass vessel. Once melting is complete, the hot bath is replaced with a cold one to refreeze the sample. Slow refreezing from bottom upward results in bubble-free ice and ensures close to 100% recovery of the ice core air.

COS, CH_3Cl , and CH_3Br measurements were conducted using a ‘dry’ extraction method, in which the entrapped air is liberated in a cold vacuum chamber without melting the ice core samples (Aydin et al., 2007). After cleaning the outer

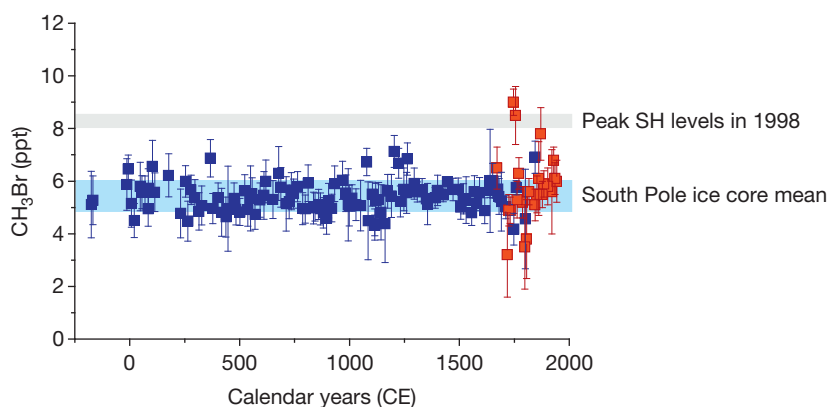


Figure 6 CH_3Br measurements in polar ice cores. The Siple Dome (red squares) and the South Pole (blue squares) data are consistent during the period of overlap. The gray band represents the peak annual mean CH_3Br mixing ratios observed in the southern hemisphere during 1997–1999 (Montzka et al., 2007). The blue band represents the mean CH_3Br mixing ratio in the South Pole ice core. The thicknesses of the gray and blue bands represent the estimated uncertainty in the reported means.

surfaces, ice core samples (400–500 g) are placed on a sharp cutting surface welded inside a stainless steel vacuum chamber. Once the chamber is isolated and pumped out, it is put into linear motion device that results in the sample sliding back and forth over the cutting surface. In 15 min, most of the ice core sample is shredded and the liberated air is collected in the vacuum chamber. This method results in ~70% recovery of the air trapped in the ice core.

Regardless of the extraction technique that has been used, very small amounts of air are available for measurements, requiring highly sensitive analytical instrumentation. The COS, CH₃Cl, and CH₃Br have been measured in a gas chromatogram–mass spectrometer (GC–MS) system that makes use of high-resolution mass spectrometry for improved sensitivity, which enables ppt level measurements in ice core air samples that often do not exceed 20–30 ml STP (standard temperature and pressure) in size (Aydin et al., 2007). The CO measurements are generally carried out with a GC equipped with a reduction gas detector.

Considerations for Future Work

The potential of the ice cores in preserving a true archive of past atmospheric ultra-trace gas variations remains largely unexplored. In addition to the compounds covered in this article, ethane and carbon disulfide (Aydin et al., 2007), and stable isotopic ratios of CO (Wang et al., 2010) have been measured in ice core air. It is safe to assume that there will be attempts to add other ultra-trace gases (e.g., nonmethane hydrocarbons) to this list, and clearly more work is required on the verification of the already existing data with multiple ice cores from a variety of sites. Broader participation by the international scientific community and involvement of additional laboratories are required for faster progress.

See also: **Ice Core Methods:** CO₂ Studies; Methane Studies; Microparticle and Trace Element Studies; Overview; Studies of Firm Air. **Ice Cores:** History of Nitrous Oxide from Ice Cores.

References

- Assanov SS, Brenninkmeijer CAM, Jöckel P, Mulvaney R, Bernard S, and Chappellaz J (2007) Evidence for a CO increase in the SH during the 20th century based on firm air samples from Berkner Island, Antarctica. *Atmospheric Chemistry and Physics* 7: 295–308. www.atmos-chem-phys.net/7/295/2007/.
- Assanov SS, Brenninkmeijer CAM, and Jöckel P (2005) The ¹⁸O isotope exchange rate between firm air CO₂ and the firm air matrix at three Antarctic sites. *Journal of Geophysical Research* 110: D18310. <http://dx.doi.org/10.1029/2005JD005769>.
- Aydin M, Saltzman ES, Bruyn WJD, Montzka SA, Butler JH, and Battle M (2004) Atmospheric variability of methyl chloride during the last 300 years from an Antarctic ice core and firm air. *Geophysical Research Letters* 31(2): L02109. <http://dx.doi.org/10.1029/2003GL018750>.
- Aydin M, Williams MB, and Saltzman ES (2007) Feasibility of reconstructing paleoatmospheric records of selected alkanes, methyl halides, and sulfur gases from Greenland ice cores. *Journal of Geophysical Research* 112: D07312. <http://dx.doi.org/10.1029/2006JD008027>.
- Aydin M, Williams MB, Tatum C, and Saltzman ES (2008) Carbonyl sulfide in air extracted from a South Pole ice core: A 2,000 year record. *Atmospheric Chemistry and Physics* 8: 7533–7542. www.atmos-chem-phys.net/8/7533/2008/.
- Bakwin PS, Tans PP, and Novelli PC (1994) Carbon monoxide budget in the Northern Hemisphere. *Geophysical Research Letters* 21(6): 433–436.
- Butler JH, Battle M, Bender ML, et al. (1999) A record of atmospheric halocarbons during the twentieth century from polar firm air. *Nature* 399: 749–755.
- Campbell JE, Carmichael GR, Chai T, et al. (2008) Photosynthetic control of atmospheric carbonyl sulfide during the growing season. *Science* 322: 1085–1088.
- Clerbaux E, et al. (2006) Long-lived compounds. In: W.M. Organization (ed.) *Scientific Assessment of Ozone Depletion: 2006*. Global Ozone Research and Monitoring Project – Report No. 50. Geneva: World Meteorological Organization.
- Duncan BN, Logan JA, Bey I, et al. (2007) Global budget of CO, 1988–1997: Source estimates and validation with a global model. *Journal of Geophysical Research* 112: D22301. <http://dx.doi.org/10.1029/2007JD008459>.
- Ferretti DF, Miller JB, White JWC, et al. (2005) Unexpected changes to the global methane budget over the past 2000 years. *Science* 309(5741): 1714–1717.
- Haan D, Martinerie P, and Raynaud D (1996) Ice core data of atmospheric carbon monoxide over Antarctica and Greenland during the last 200 years. *Geophysical Research Letters* 23(17): 2235–2238.
- Haan D and Raynaud D (1998) Ice core record of CO variations during the last two millennia: Atmospheric implications and chemical interactions within the Greenland ice. *Tellus* 50B: 253–262.
- Hamilton JTG, Colin McRoberts W, Keppler F, Kalin RM, and Harper DB (2003) Chloride methylation by plant pectin: An efficient environmentally significant process. *Science* 301: 206–209.
- Kaspers KA, van de Wal RSW, de Gouw JA, et al. (2004) Analyses of firm gas samples from Dronning Maud Land, Antarctica: Study of nonmethane hydrocarbons and methyl chloride. *Journal of Geophysical Research* 109: D02307. <http://dx.doi.org/10.1029/2003JD003950>.
- Kettle AJ, Kuhn U, von Hobe M, Kesselmeier J, and Andreae MO (2002) Global budget of atmospheric carbonyl sulfide: Temporal and spatial variations of the dominant sources and sinks. *Journal of Geophysical Research* 107(D22): 4658. <http://dx.doi.org/10.1029/2002JD002187>.
- Khalil MAK and Rasmussen RA (1988) Carbon monoxide in the Earth's atmosphere: Indications of a global increase. *Nature* 332: 242–245.
- Khalil MAK and Rasmussen RA (1990) The global cycle of carbon monoxide: Trends and mass balance. *Chemosphere* 20: 227–242.
- Khalil MAK and Rasmussen RA (1999) Atmospheric methyl chloride. *Atmospheric Environment* 33: 1305–1321.
- Mayewski PA and Maasch KA (2006) Recent warming inconsistent with natural association between temperature and atmospheric circulation over the last 2000 years. *Climate of the Past Discussions* 2: 1–29.
- Montzka SA, Aydin M, Battle MO, et al. (2004) A 350-year atmospheric history for carbonyl sulfide inferred from Antarctic firm air and air trapped in ice. *Journal of Geophysical Research* 109: D22302. <http://dx.doi.org/10.1029/2004JD004686>.
- Montzka SA, Butler JH, Hall BD, Mondeel DJ, and Elkins JW (2003) A decline in tropospheric organic bromine. *Geophysical Research Letters* 30(15): 1826. <http://dx.doi.org/10.1029/2003GL017745>.
- Montzka SA, Calvert P, Hall BD, et al. (2007) On the global distribution, seasonality, and budget of atmospheric carbonyl sulfide (COS) and some similarities to CO₂. *Journal of Geophysical Research* 111: D09302. <http://dx.doi.org/10.1029/2006JD007665>.
- Montzka SA, Fraser PJ, et al. (2003) Controlled substances and other source gases. ch. 1 In: Organization WM (ed.) *Scientific Assessment of Ozone Depletion: 2002*. Global Ozone Research and Monitoring Project – Report No. 47, pp. 1.5–1.83. Geneva: World Meteorological Organization.
- Novelli PC, Masarie KA, and Lang PM (1998) Distributions and recent changes of carbon monoxide in the lower troposphere. *Journal of Geophysical Research* 103: 19015–19033.
- Novelli PC, Masarie KA, Lang PM, Hall BD, Myers RC, and Elkins JW (2003) Reanalysis of tropospheric CO trends: Effects of the 1997–1998 wildfires. *Journal of Geophysical Research* 108(D15): 4464. <http://dx.doi.org/10.1029/2002JD003031>.
- Saito T, Yokouchi Y, Aoki S, Nakazawa T, Fujii Y, and Watanabe O (2007) Ice-core record of methyl chloride over the last glacial – Holocene climate change. *Geophysical Research Letters* 34: L03801. <http://dx.doi.org/10.1029/2006GL028090>.
- Saltzman E, Aydin M, DeBruyn J, King DB, and Yvon-Lewis SA (2004) Methyl bromide in preindustrial air: Measurements from an Antarctic ice core. *Journal of Geophysical Research* 109(D5): D05301. <http://dx.doi.org/10.1029/2003JD004157>.
- Saltzman ES, Aydin M, Tatum C, and Williams MB (2008) 2,000-year record of atmospheric methyl bromide from a South Pole ice core. *Journal of Geophysical Research* 113: D05304. <http://dx.doi.org/10.1029/2007JD008919>.
- Saltzman ES, Aydin M, Williams MB, and Verhulst KR (2009) Methyl chloride in a deep ice core from Siple Dome, Antarctica. *Geophysical Research Letters* 36: L03822. <http://dx.doi.org/10.1029/2008GL036266>.

- Sandoval-Soto L, Stanimirov M, von Hobe M, et al. (2005) Global uptake of carbonyl sulfide (COS) by terrestrial vegetation: Estimates corrected by deposition velocities normalized to the uptake of carbon dioxide (CO₂). *Biogeosciences* 2: 125–132. <http://www.biogeosciences.net/2/125/2005/>.
- Seibt U, Kesselmeier J, Sandoval-Soto L, Kuhn U, and Berry J (2010) A kinetic analysis of leaf uptake of COS and its relation to transpiration, photosynthesis and carbon isotope fractionation. *Biogeosciences* 7: 333–341. www.biogeosciences.net/7/333/2010/.
- Sturges WT, McIntyre HP, Penkett SA, et al. (2001) Methyl bromide, other brominated methanes and methyl iodide in polar firn air. *Journal of Geophysical Research* 106 (D2): 1595–1606.
- Sturges WT, Penkett SA, Barnola J-M, Chappellaz J, Atlas E, and Stroud V (2001) A long-term record of carbonyl sulfide (COS) in two hemispheres from firn air measurements. *Geophysical Research Letters* 28: 4095–4098.
- Tokarczyk R, Saltzman ES, and Moore RM (2003) Biological degradation of methyl chloride in coastal seawater. *Global Biogeochemical Cycles* 17(2): 1057. <http://dx.doi.org/10.1029/2002GB001949>.
- Trudinger CM, Etheridge DM, Sturrock GA, Fraser PJ, Krummel PB, and McCulloch A (2004) Atmospheric histories of halocarbons from analysis of Antarctic firn air: Methyl bromide, methyl chloride, chloroform, and dichloromethane. *Journal of Geophysical Research* 109(D22): D22310. <http://dx.doi.org/10.1029/2004JD004932>.
- Wang Z, Chappellaz J, Park K, and Mak JE (2010) Large variations in southern hemisphere biomass burning during the last 650 years. *Science* 330(6011): 1663–1666.
- Weissenstein DK, Bekki S, Mills M, et al. (2006) In: Thomason L and Peter T (eds.) *Modeling of Stratospheric Aerosols, Stratospheric Processes and Their Role in Climate (SPARC)*. Geneva, Switzerland: World Climate Research Programme. Report 4.
- Williams MB, Aydin M, Tatum C, and Saltzman ES (2007) 2000 year atmospheric history of methyl chloride from a South Pole ice core: Evidence for climate-controlled variability. *Geophysical Research Letters* 34(7): L07811. <http://dx.doi.org/10.1029/2006GL029142>.
- Xu X, Bingemer HG, and Schmidt U (2002) The flux of carbonyl sulfide and carbon disulfide between the atmosphere and a spruce forest. *Atmospheric Chemistry and Physics* 2: 171–181. <http://www.atmos-chem-phys.net/2/171/2002/>.
- Yokouchi Y, Toon-Saunty D, Yazawa K, Inagaki T, and Tamaru T (2002) Recent decline of methyl bromide in the troposphere. *Atmospheric Environment* 36(32): 4985–4989.
- Yvon-Lewis SA and Butler JH (1997) The potential effect of oceanic biological degradation on the lifetime of atmospheric CH₃Br. *Geophysical Research Letters* 24: 1227–1230.
- Yvon-Lewis SA, Saltzman ES, and Montzka SA (2009) Recent trends in atmospheric methyl bromide: Analysis of post-Montreal Protocol variability. *Atmospheric Chemistry and Physics* 9: 5963–5974. www.atmos-chem-phys.net/9/5963/2009/.

History of Nitrous Oxide from Ice Cores

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Introduction

As a trace gas in the atmosphere, nitrous oxide (N_2O) is part of the global nitrogen cycle (Gruber and Galloway, 2008) and important because it acts as a strong greenhouse gas and is involved in the destruction of stratospheric ozone. The atmospheric burden of N_2O dramatically increased over the last centuries, resulting in a current atmospheric concentration of 322 parts per billion by volume (ppbv). This $\sim 20\%$ increase with respect to preindustrial levels of ~ 270 ppbv contributes $\sim 6\%$ to the anthropogenic radiative forcing of all long-lived greenhouse gases. The increase originates from anthropogenic N_2O emissions from expanding and fertilized agricultural lands, burning of fossil fuels and biomass, and industrial processes. Further, human activities (human excreta, landfills, and atmospheric deposition) enhance the input of nitrogen compounds into various ecosystems, leading to significant anthropogenic N_2O emissions from, for example, coastal and open oceans, rivers, and estuaries. Anthropogenic N_2O emissions currently account for ~ 6.7 Tg of N per year, while natural N_2O sources emit ~ 11 Tg of N per year (IPCC, 2007). Nitrification and denitrification are the fundamental microbiological processes responsible for N_2O production in terrestrial and marine ecosystems, while minor N_2O emissions derive from chemical processes in the atmosphere. About two thirds of the total natural N_2O emissions stem from soils, where the emission strength not only depends on the water-filled pore space and, thus, oxygen availability, but also on nitrogen input, temperature, and soil structure. About one third of the total natural N_2O emissions stems from marine sources, which are highest in equatorial and coastal upwelling regions, the Southern Ocean ($40\text{--}60^\circ\text{S}$), the North Pacific, and the Arabian Sea. The major sink of N_2O is its destruction by photodissociation, followed by chemical reactions with excited oxygen. Both processes take place only in the stratosphere, making N_2O an inert gas in the troposphere. Consequently, N_2O has a relatively long atmospheric life time of ~ 120 years, and is globally well mixed. The strength of the stratospheric sink depends on the amount of solar UV radiation, which is needed for the photodissociation of N_2O , and on the exchange rate of air between troposphere (where N_2O is produced) and stratosphere (where N_2O is destroyed). Since there are no indications for substantial changes in either the solar irradiance or the exchange rate of air between troposphere and stratosphere, the strength of the N_2O sink is commonly considered as remaining constant, an assumption that is also supported by modeling studies (Martinerie et al., 1995).

In the accumulation zones of polar ice sheets, snow is stored layer by layer and advected downward. In the firn column (the upper 50–120 m of an ice sheet), atmospheric air can circulate through the pore space. Below, where the densification process of snow to ice is completed, air bubbles are enclosed in the ice. Accordingly, the past evolution of the

atmospheric composition is recorded in the ice and can directly be reconstructed by analyses of air extracted from ice cores recovered from the polar ice sheets of Greenland and Antarctica. To date, a composite record consisting of measurements from various studies and ice cores provides a complete insight into the evolution of atmospheric N_2O over the last 140 000 years (140 ka) (Schilt et al., 2010b). Interglacial N_2O records even reach back to as far as 800 ka before present (BP) (Spahni et al., 2005; Schilt et al., 2010a). An overview of N_2O records is also provided in Wolff and Spahni (2007).

Analytical Techniques

The analytical techniques applied to measure past N_2O concentrations in air extracted from polar ice cores are very similar to the techniques used for carbon dioxide (CO_2 , see [CO₂ Studies](#)) and methane (CH_4 , see [Methane Studies](#)). Often, the air extracted from one single sample is analyzed for more than one trace gas. In a first step, an ice sample (40–600 g) is cut with a band saw and put in a sealed glass or stainless steel container, which is then connected to a vacuum system in order to remove ambient air. Different extraction techniques have provided reproducible results for N_2O concentrations. These techniques include (i) dry extraction with a mill or cheese grater system (e.g., Machida et al., 1995; Sowers, 2001), (ii) sublimation extraction (Güllük et al., 1998), (iii) melt extraction by subsequent refreezing of the melt water (e.g., Chappellaz et al., 1997; Flückiger et al., 1999). Following the extraction, the N_2O concentration is determined using either a laser spectrometer or, up to now more commonly applied, a gas chromatograph equipped with an electron capture detector. Measurements are calibrated using standard gases of known concentrations. The obtained N_2O concentrations are corrected based on measurements with bubble-free ice and standard gases, ensuring that potential contamination and solubility effects do not bias the results. Typically, the reproducibility of N_2O concentration measurements is 5.6 ppbv (as determined by measuring neighboring samples with the melt extraction, Schilt et al., 2010b).

Reliability of N_2O Records from Polar Ice Cores

Consistent results have been obtained for past N_2O concentrations on centennial, millennial, and glacial–interglacial time scales with different extraction techniques and analysis systems, located in different labs. Nevertheless, N_2O records measured along polar ice cores are occasionally affected by *in situ* production of N_2O in the ice. Measurements affected by *in situ* production show concentrations above the atmospheric value at the time of air trapping. Such artifacts are mainly observed in ice from cold (glacial) time intervals, and may originate from

microbiological activities in the ice (Miteva et al., 2007; Rohde et al., 2008). Precautions are needed in order to ensure that all measurements affected by *in situ* production are excluded from the atmospheric records. The applied methods for the exclusion of artifacts include algorithms iteratively detecting outliers in highly resolved records (Flückiger et al., 2004, see also Figure 4), as well as a dust criterion broadly excluding measurements along ice samples with large amounts of impurities (Spahni et al., 2005; Schilt et al., 2010a). Although the exclusion of measurements affected by *in situ* production is done in an empirical way, the remaining atmospheric records from different ice cores are in very good agreement. Such agreement of N₂O records from different ice cores (recovered from both Greenland and Antarctica) is a strong and required validation of the atmospheric origin of the measured concentrations, based on the fact that N₂O is globally well mixed in the atmosphere.

Another issue regarding a reliable reconstruction of atmospheric N₂O concentrations from ice cores is the gravitational enrichment of N₂O in the firn column. However, this effect typically accounts for only ~2 ppbv and is therefore not considered in most studies.

The Last Two Millennia

While many N₂O records from different ice cores and investigators confirm the general trend in atmospheric N₂O concentrations over the last two millennia (Figure 1), the most accurate N₂O record covering this time interval may currently stem from Law Dome (MacFarling Meure et al., 2006). The available records from firn air and ice cores from Greenland and Antarctica are confirmed and extended to present by direct atmospheric measurements, which already record more than the last 30 years. Between AD 0 and 1850, atmospheric N₂O concentrations show a rather stable level with some minor variations of up to ~10 ppbv, for example, reflected in an increase between AD 675 and 810 (MacFarling Meure et al., 2006). Numerous studies indicate N₂O concentrations of ~270 ppbv between AD 1750 and 1850. After AD 1850, the

atmospheric N₂O concentrations drastically increase to today's concentration of 322 ppbv, with a higher increase rate during the second half of the twentieth century. Contemporaneous with this concentration increase, measurements of the isotopic composition of N₂O indicate a decrease in the relative amount of ¹⁸O, ¹⁵N, and position dependent ¹⁵N of N₂O (Sowers et al., 2002; Bernard et al., 2006; Ishijima et al., 2007). This evolution of the isotopic composition is in agreement with an increasing importance of anthropogenic N₂O emissions from agricultural activities since AD 1850.

The Holocene

The first accurate high-resolution N₂O record covering the Holocene (current interglacial) has been measured along the EPICA Dome C (EDC) ice core (Flückiger et al., 2002). This record has then been confirmed by the Greenland Ice Sheet Project 2 (GISP2) ice core (Sowers et al., 2003), as well as by the EPICA Dronning Maud Land (EDML) and Talos Dome ice cores (Schilt et al., 2010a,b), indicating that all records reveal atmospheric values (Figure 2). Although the N₂O concentrations remain rather stable during the preindustrial Holocene, some significant variations are observed. In the early Holocene, the N₂O concentrations start to decrease from ~270 ppbv and reach minimal Holocene levels of below ~260 ppbv between about 10 and 6 ka BP, followed by a slow increase to the preindustrial level. Some short-term variations may be present, but they are poorly resolved with the current resolution and analytical precision of the measurements.

Assuming that the sink of N₂O has remained constant, the minimal concentrations during the Holocene correspond to a decrease of the total preindustrial source of 3.7%. Assigning the whole decrease to one source type only would either require a decrease of 5.6% or 11.2% of the terrestrial or marine source, respectively (Flückiger et al., 2002). At this point, however, the causes for the observed variations in N₂O concentrations during the Holocene are not yet completely understood.

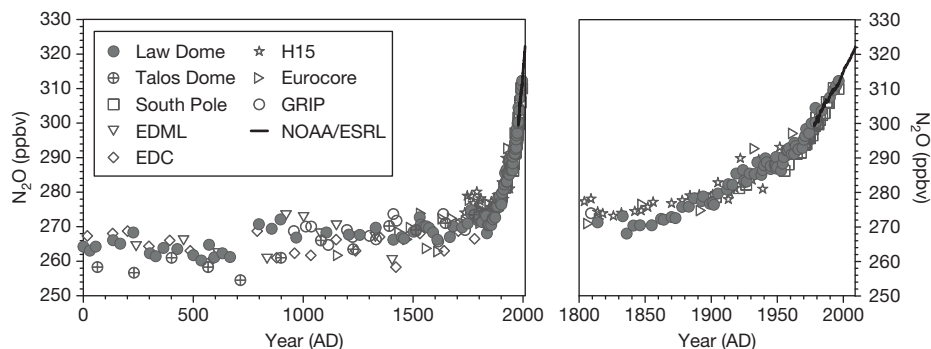


Figure 1 The evolution of atmospheric N₂O concentrations during the last two millennia. Shown are selected results from various firn air and ice cores (Machida et al., 1995; Battle et al., 1996; Flückiger et al., 1999, 2002; MacFarling Meure et al., 2006; Schilt et al., 2010a,b). Other investigations (not shown here) have also clearly revealed the anthropogenic N₂O increase (e.g., Pearman et al., 1986; Etheridge et al., 1988; Khalil and Rasmussen, 1988; Leuenberger and Siegenthaler, 1992; Dibb et al., 1993; Sowers, 2001). However, the latter studies somewhat differ from the results shown here, either due to calibration inconsistencies, larger scatter, and/or too high preindustrial concentrations. The black line shows a combined data set of direct atmospheric N₂O measurements from the NOAA/ESRL global monitoring division.

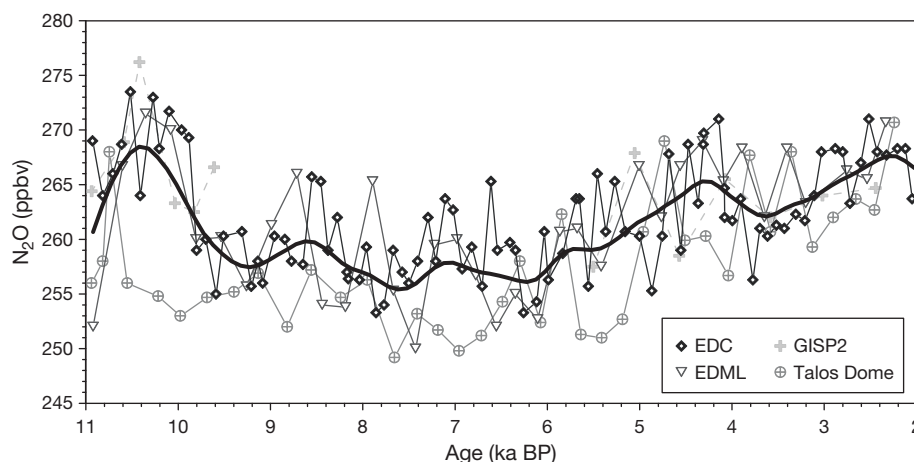


Figure 2 High-resolution N_2O records covering the Holocene as measured along the EDC (Flückiger et al., 2002), EDML (Schilt et al., 2010a), Talos Dome (Schilt et al., 2010b), and GISP2 (Sowers et al., 2003) ice cores. Differences, in particular around 10 ka BP, may at least partly be the result of uncertainties in the time scales of the different ice cores. The black line represents a spline with a cutoff period of 1000 years through all given records.

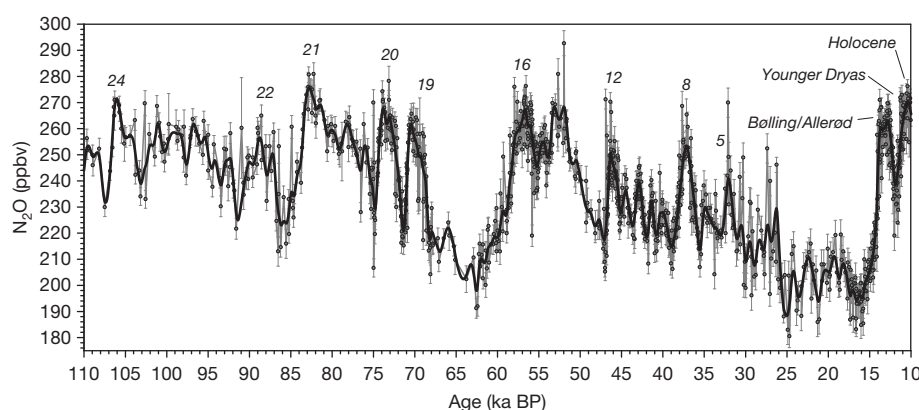


Figure 3 The evolution of atmospheric N_2O concentrations during the last glacial and the transition to the Holocene. Shown is a composite record consisting of measurements along the following ice cores: EDC (Flückiger et al., 1999; Stauffer et al., 2002; Spahni et al., 2005), EDML (Schilt et al., 2010a), NGRIP (Flückiger et al., 2004; Schilt et al., 2010b), Talos Dome (Schilt et al., 2010b), GRIP (Flückiger et al., 1999), and GISP2 (Sowers et al., 2003). The records of different ice cores have been put on a common timescale using CH_4 as a synchronization tool (Schilt et al., 2010b), and a spline with a cutoff period of 1000 years has been calculated through all N_2O measurements (black line). Numbers denominate Dansgaard/Oeschger events.

The Transition from the Last Glacial to the Holocene

The transition from the last glacial to the Holocene marks one of the most extreme warmings in the Earth's climate during the Late Quaternary. First measurements along Antarctic ice cores have already identified a substantial N_2O increase in concert with the termination of the last glacial (Zardini et al., 1989). The N_2O increase has then been quantified to roughly 30% (Leuenberger and Siegenthaler, 1992). More recent measurements, however, are much more precise and yield a detailed and highly resolved insight into the N_2O evolution over the transition from the last glacial to the Holocene (Figure 3).

Around 16 ka BP, N_2O concentrations start to increase from an extremely low level of ~ 200 ppbv. The increase accelerates at 14.5 ka BP, at the same time when CH_4 concentrations jump into the Bolling/Allerød warm period (see Methane Studies). During the latter time interval, N_2O concentrations remain fairly constant at ~ 260 ppbv. In response to the

subsequent Younger Dryas cold period, N_2O concentrations decrease by ~ 30 ppbv. In contrast to CH_4 , the N_2O evolution in response to the Younger Dryas cold period is much smoother, and the strength of N_2O sources always remains above typical glacial values (Flückiger et al., 1999). For the Younger Dryas cold period, Goldstein et al. (2003) have simulated a concentration decrease of ~ 10 ppbv with a coupled climate-biogeochemical model of reduced complexity. The model suggests that a reduced Atlantic thermohaline circulation leads to a stronger stratification within the ocean. As a consequence, total marine N_2O emissions decrease because N_2O production decreases and N_2O storage increases. Since the model only takes into account marine processes, terrestrial emissions may be responsible for the remaining ~ 20 ppbv of the concentration decrease in response to the Younger Dryas cold period. Following the Younger Dryas cold period, N_2O concentrations increase to an early Holocene value of ~ 270 ppbv.

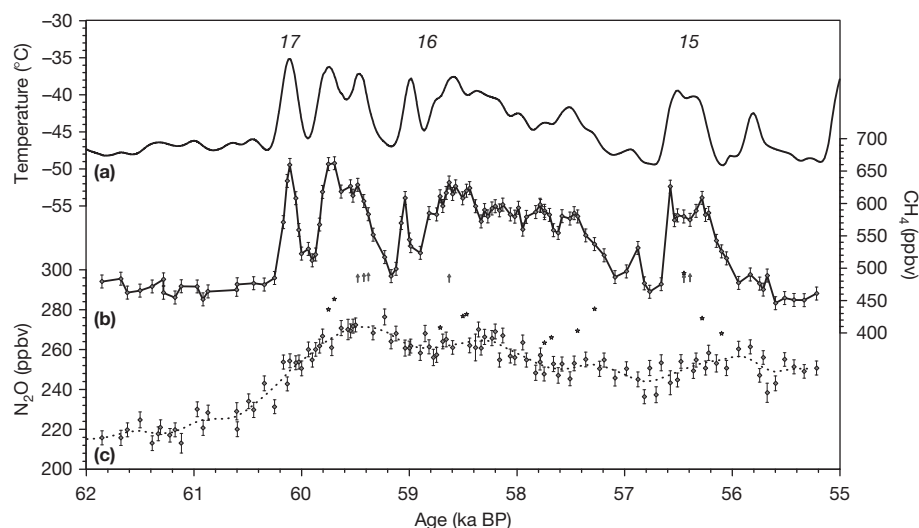


Figure 4 Greenland temperature, CH₄, and N₂O concentrations during the Dansgaard/Oeschger events 17–15 as measured along the NGRIP ice core. (a) Reconstruction of Greenland temperature (Huber et al., 2006). (b) CH₄ concentrations (Huber et al., 2006). (c) N₂O concentrations (Schilt et al., 2010b). Stars remark measurements which are affected by *in situ* production of N₂O in the ice, arrows indicate the location of such artifacts exceeding the limits of the y-axis. Figure adapted from Schilt et al. (2010b).

Since N₂O emitted from terrestrial and marine sources differs significantly in its characteristic isotopic composition, a change in the isotopic composition of atmospheric N₂O can help to decipher the relative contribution of terrestrial to marine sources. For the last 33 ka, measurements along the Taylor Dome ice core indicate that both $\delta^{15}\text{N}$ and $\delta^{18}\text{O}$ of N₂O remained fairly constant (Sowers et al., 2003). This suggests that the ratio of terrestrial to marine emissions varied by less than 16%. Additional measurements of the isotopic composition of N₂O will further help to understand the biogeochemical processes responsible for the observed variations in atmospheric N₂O concentrations.

The Last Glacial Epoch

During the last glacial, temperatures are generally reduced, but reconstructions of Greenland surface temperature show rapid oscillations of up to 15 °C (Huber et al., 2006). These so-called Dansgaard/Oeschger events take place on millennial time scales and have counterparts in the southern hemisphere (see [Correlations Between Greenland and Antarctica, Greenland Stable Isotopes](#)). In general, N₂O concentrations are low during cold time intervals and strongly increase during Dansgaard/Oeschger events, occasionally reaching concentrations as high as during the Holocene. This is visible in a composite N₂O record, which consists of measurements along numerous ice cores and which covers without any gaps the last glacial–interglacial cycle (Figure 3). Further, high-resolution N₂O records have revealed numerous and detailed characteristics of the N₂O evolution during Dansgaard/Oeschger events. (i) The N₂O amplitude in response to Dansgaard/Oeschger events increases approximately linearly with the duration of the events. In contrast to CH₄, however, the amplitude of N₂O during Dansgaard/Oeschger events does not show any relationship with solar insolation (Flückiger et al.,

2004). (ii) For long-lasting Dansgaard/Oeschger events, N₂O concentrations start to increase several hundred years before the rapid rise in Greenland temperature and CH₄ concentrations (one example is shown in Figure 4), while the available records for shorter events are less clear (Flückiger et al., 2004). A modeling study has suggested that the early increase results from a long-term adjustment of the nitrate and oxygen inventories in the upper ocean following a substantial reduction of the Atlantic Meridional Overturning Circulation (Schmittner and Galbraith, 2008). (iii) Although starting earlier, the N₂O increase lasts significantly longer than the CH₄ increase at the beginning of Dansgaard/Oeschger events, and is completed within some hundred years. The longer-lasting increase can mostly be explained by the differences in atmospheric life times, which are ~120 and ~10 years for N₂O and CH₄, respectively (Flückiger et al., 2004). On the other hand, it appears that the reduction of the N₂O source at the end of the Dansgaard/Oeschger event 8 is slower than the reduction of the CH₄ source (Flückiger et al., 1999). (iv) In the course of the Dansgaard/Oeschger events 17–15, Greenland temperature shows large variations on centennial time scales (Figure 4). While CH₄ concentrations closely follow the temperature evolution, N₂O concentrations are not affected by the oscillations in Greenland temperature. Given that the atmospheric life times of the two gases can explain only parts of the differences, parallel emission changes in the sources of CH₄ and N₂O can be excluded for these centennial timescale variations (Schilt et al., 2010b).

Explaining the causes of the observed changes in atmospheric N₂O concentrations remains a challenging task. Some insight may come from the analysis of marine sediment cores. For instance, considerable enhancement of the water column denitrification in the Arabian Sea corresponds to the N₂O variations reconstructed in response to Dansgaard/Oeschger events (Suthhof et al., 2001; Pichevin et al., 2007). Since denitrification in the Arabian Sea substantially contributes to

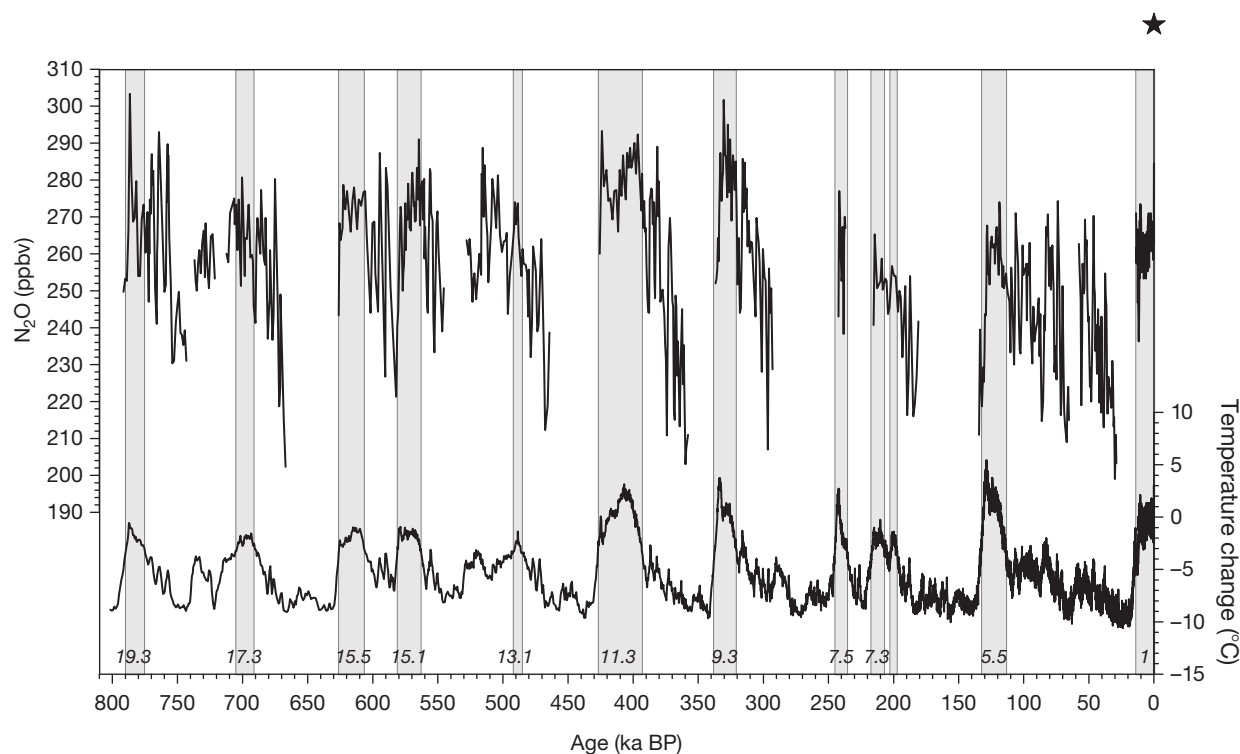


Figure 5 Atmospheric N_2O concentrations (Flückiger et al., 2002; Stauffer et al., 2002; Spahni et al., 2005; Schilt et al., 2010a) and a proxy of temperature changes over Antarctica (relative to the average of the last 1000 years, Jouzel et al., 2007) from the EDC ice core covering the last 800 ka. Along the EDC ice core, glacial time intervals are affected by *in situ* production of N_2O and, thus, do not allow for a reconstruction of atmospheric N_2O concentrations. The star outside the graph indicates the current N_2O concentration of 322 ppbv. Gray shadows mark interglacials, which are arbitrarily defined as time intervals with a temperature not more than 3°C colder than the last 1000 years. Numbers denote Marine Isotope Stages.

the global marine N_2O source, such changes could contribute to the observed variations in atmospheric N_2O concentrations. However, the large variations in CH_4 concentrations, which are driven by terrestrial sources only, point to substantial changes in the terrestrial biogeochemical cycles as well.

Previous Interglacials

To date, there exist only a few ice cores containing ice from previous glacial–interglacial cycles. All of them are unfortunately affected to various degrees by *in situ* production of N_2O in most of the available glacial ice, which so far precludes a detailed insight into the evolution of atmospheric N_2O concentrations during glacial time intervals beyond the last glacial. Nevertheless, atmospheric N_2O concentrations are available for all interglacials of the last 800 ka (Figure 5).

As in the course of the last glacial–interglacial cycle, N_2O concentrations vary in line with climate, showing high concentrations in warm phases and low concentrations in cold phases. Today's N_2O concentration of 322 ppbv considerably exceeds the natural variation range of ~ 200 – 300 ppbv as determined from the reconstructions of the last 800 ka. While the greenhouse gases CO_2 and CH_4 show lower values for the cooler interglacials between 440 and 800 ka BP compared to the interglacials of the last 440 ka, N_2O concentrations show increased levels. At the end of most interglacials, N_2O

concentrations remain on interglacial levels several thousand years longer than CH_4 concentrations, the latter starting to decrease synchronously with Antarctic temperature (Schilt et al., 2010a). Although the EDC record suffers from large gaps in the glacial N_2O record due to *in situ* production of N_2O in the ice, millennial timescale N_2O variations are observed throughout the last 800 ka. Similar to the Dansgaard/Oeschger events observed during the last glacial, these N_2O events occur contemporaneously with variations in CH_4 concentrations and show a linear relationship between amplitude and duration. These similarities suggest that the underlying mechanism driving the Dansgaard/Oeschger events of the last glacial may also be active throughout the last 800 ka (Schilt et al., 2010a).

See also: **Ice Core Methods:** Chronologies; CO_2 Studies; Methane Studies; Overview. **Ice Core Records:** Correlations Between Greenland and Antarctica; Greenland Stable Isotopes. **Paleoceanography, Physical and Chemical Proxies:** Nutrient Proxies.

References

- Battle M, Bender M, Sowers T, et al. (1996) Atmospheric gas concentrations over the past century measured in air from firn at the South Pole. *Nature* 383: 231–235.

- Bernard S, Röckmann TR, Kaiser J, et al. (2006) Constraints on N₂O budget changes since pre-industrial time from new firn air and ice core isotope measurements. *Atmospheric Chemistry and Physics* 6: 493–503.
- Chappellaz J, Blunier T, Kints S, et al. (1997) Changes in the atmospheric CH₄ gradient between Greenland and Antarctica during the Holocene. *Journal of Geophysical Research – Atmospheres* 102: 15987–15997.
- Dibb JE, Rasmussen RA, Mayewski PA, and Holdsworth G (1993) Northern hemisphere concentrations of methane and nitrous oxide since 1800 – Results from the Mt. Logan and 20D ice cores. *Chemosphere* 27: 2413–2423.
- Etheridge D, Pearman GI, and de Silva F (1988) Atmospheric trace-gas variations as revealed by air trapped in an ice core from Law Dome, Antarctica. *Annals of Glaciology* 10: 28–33.
- Flückiger J, Dällenbach A, Blunier T, et al. (1999) Variations in atmospheric N₂O concentration during abrupt climatic changes. *Science* 285: 227–230.
- Flückiger J, Monnin E, Stauffer B, et al. (2002) High-resolution Holocene N₂O ice core record and its relationship with CH₄ and CO₂. *Global Biogeochemical Cycles* 16: 1010. <http://dx.doi.org/10.1029/2001GB001417>.
- Flückiger J, Blunier T, Stauffer B, et al. (2004) N₂O and CH₄ variations during the last glacial epoch: Insight into global processes. *Global Biogeochemical Cycles* 18: GB1020. <http://dx.doi.org/10.1029/2003GB002122>.
- Goldstein B, Joos F, and Stocker TF (2003) A modeling study of oceanic nitrous oxide during the Younger Dryas cold period. *Geophysical Research Letters* 30: 1092. <http://dx.doi.org/10.1029/2002GL016418> 64.1-64.4.
- Gruber N and Galloway JN (2008) An Earth-system perspective of the global nitrogen cycle. *Nature* 451: 293–296.
- Güllük T, Slemr F, and Stauffer B (1998) Simultaneous measurements of CO₂, CH₄, and N₂O in air extracted by sublimation from Antarctica ice cores: Confirmation of the data obtained using other extraction techniques. *Journal of Geophysical Research* 103: 15971–15978.
- Huber C, Leuenberger M, Spahni R, et al. (2006) Isotope calibrated Greenland temperature record over Marine Isotope Stage 3 and its relation to CH₄. *Earth and Planetary Science Letters* 243: 504–519.
- IPCC (2007) *Climate Change 2007: The Physical Science Basis, Contribution of Working Group I to the Fourth Assessment Report of the IPCC*.
- Ishijima K, Sugawara S, Kawamura K, et al. (2007) Temporal variations of the atmospheric nitrous oxide concentration and its delta N-15 and delta O-18 for the latter half of the 20th century reconstructed from firn air analyses. *Journal of Geophysical Research – Atmospheres* 112: D03305. <http://dx.doi.org/10.1029/2006JD007208>.
- Jouzel J, Masson-Delmotte V, Cattani O, et al. (2007) Orbital and millennial Antarctic climate variability over the past 800,000 years. *Science* 317: 793–796.
- Khalil MAK and Rasmussen RA (1988) Nitrous oxide: Trends and global mass balance over the last 3000 years. *Annals of Glaciology* 10: 73–79.
- Leuenberger M and Siegenthaler U (1992) Ice-age atmospheric concentration of nitrous oxide from an Antarctic ice core. *Nature* 360: 449–451.
- MacFarling Meure CM, Etheridge D, et al. (2006) Law Dome CO₂, CH₄ and N₂O ice core records extended to 2000 years BP. *Geophysical Research Letters* 33: L14810. <http://dx.doi.org/10.1029/2006GL026152>.
- Machida T, Nakazawa T, Fujii Y, Aoki S, and Watanabe O (1995) Increase in the atmospheric nitrous oxide concentration during the last 250 years. *Geophysical Research Letters* 22: 2921–2924.
- Martinerie P, Brasseur GP, and Granier C (1995) The chemical composition of ancient atmospheres – A model study constrained by ice core data. *Journal of Geophysical Research – Atmospheres* 100: 14291–14304.
- Miteva V, Sowers T, and Brenchley J (2007) Production of N₂O by ammonia oxidizing bacteria at subfreezing temperatures as a model for assessing the N₂O anomalies in the Vostok ice core. *Geomicrobiology Journal* 24: 451–459.
- Pearman GI, Etheridge D, Desilva F, and Fraser PJ (1986) Evidence of changing concentrations of atmospheric CO₂, N₂O and CH₄ from air bubbles in Antarctic Ice. *Nature* 320: 248–250.
- Picévin L, Bard E, Martinez P, and Billy I (2007) Evidence of ventilation changes in the Arabian Sea during the late Quaternary: Implication for denitrification and nitrous oxide emission. *Global Biogeochemical Cycles* 21: GB4008. <http://dx.doi.org/10.1029/2006GB002852>.
- Rohde RA, Price PB, Bay RC, and Bramall NE (2008) In situ microbial metabolism as a cause of gas anomalies in ice. *Proceedings of the National Academy of Sciences of the United States of America* 105: 8667–8672.
- Schilt A, Baumgartner M, Blunier T, et al. (2010a) Glacial–interglacial and millennial-scale variations in the atmospheric nitrous oxide concentration during the last 800,000 years. *Quaternary Science Reviews* 29: 182–192.
- Schilt A, Baumgartner M, Schwander J, et al. (2010b) Atmospheric nitrous oxide during the last 140,000 years. *Earth and Planetary Science Letters* 300: 33–43.
- Schmittner A and Galbraith ED (2008) Glacial greenhouse-gas fluctuations controlled by ocean circulation changes. *Nature* 456: 373–376.
- Sowers T (2001) N₂O record spanning the penultimate deglaciation from the Vostok ice core. *Journal of Geophysical Research* 106: 31903–31914.
- Sowers T, Rodebaugh A, Yoshida N, and Toyoda S (2002) Extending records of the isotopic composition of atmospheric N₂O back to 1800 AD from air trapped in snow at the South Pole and the Greenland Ice Sheet Project II ice core. *Global Biogeochemical Cycles* 16: 1129. <http://dx.doi.org/10.1029/2002GB001911>.
- Sowers T, Alley RB, and Jubenville J (2003) Ice core records of atmospheric N₂O covering the last 106,000 years. *Science* 301: 945–948.
- Spahni R, Chappellaz J, Stocker TF, et al. (2005) Atmospheric methane and nitrous oxide of the late Pleistocene from Antarctic ice cores. *Science* 310: 1317–1321.
- Stauffer B, Flückiger J, Monnin E, Schwander J, Barnola JM, and Chappellaz J (2002) Atmospheric CO₂, CH₄ and N₂O records over the past 60,000 years based on the comparison of different polar ice cores. *Annals of Glaciology* 35: 202–208.
- Suthhof A, Ittekkot V, and Gaye-Haake B (2001) Millennial-scale oscillation of denitrification intensity in the Arabian Sea during the late Quaternary and its potential influence on atmospheric N₂O and global climate. *Global Biogeochemical Cycles* 15: 637–649.
- Wolff E and Spahni R (2007) Methane and nitrous oxide in the ice core record. *Philosophical Transactions of the Royal Society* 365: 1775–1792.
- Zardini D, Raynaud D, Scharffe D, and Seiler W (1989) N₂O measurements of air extracted from Antarctic ice cores – Implication on atmospheric N₂O back to the last glacial–interglacial transition. *Journal of Atmospheric Chemistry* 8: 189–201.

Ice Sheets, Pleistocene *see* **Glacial Landforms, Ice Sheets:** Growth and Decay; Evidence of Glacier and Ice Sheet Extent; Trimlines and Paleonunataks; **Glaciations:** Early Quaternary (Pleistocene) and Precursors; Middle Pleistocene in Eurasia; Mid-Quaternary in North America; Middle Pleistocene Glaciations in the Southern Hemisphere; Late Pleistocene Glacial Events in Beringia; Late Quaternary of the Southwest Pacific Region; Late Quaternary of Antarctica; Late Pleistocene in Eurasia; Late Quaternary in Highland Asia; Late Quaternary in North America; Late Pleistocene in South America; **Ice Cores:** Dynamics of the Greenland Ice Sheet; Dynamics of the West Antarctic Ice Sheet; Dynamics of the East Antarctic Ice Sheet

Ice Wedges, Ice Wedge Casts *see* **Permafrost and Periglacial Features:** Ice Wedges and Ice-Wedge Casts

K

K/Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ Dating

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Introduction

Radioisotopic dating techniques in the earth sciences are used for the determination of absolute ages for events and processes in the history of the Earth and the solar system. These techniques are based on the decay of radioactive isotopes that occur in our natural environment. Radioactive isotopes that are useful as geochronometers can be split into two groups: radioactive isotopes with long half-lives that have been present since the origin of our solar system (half-lives $>\sim 500$ million years) and radioactive isotopes with short half-lives (generally half-lives $<\sim 10$ million years). The radionuclides in the latter group are steadily produced on the Earth's surface. Examples of these include the intermediate daughter products of the uranium and thorium decay series and the stable lead isotopes that are used in U-series dating, as well as radioactive nuclides formed by interaction of cosmogenic rays in the atmosphere and uppermost layers of the lithosphere which are used in terrestrial cosmogenic nuclide dating, including ^{14}C . These are important techniques for dating Quaternary processes and are described elsewhere in this encyclopedia.

Isotope geochronometers based on isotopes with long half-lives, including the K/Ar method, rely on the *accumulation of radiogenic daughter isotopes* and consequently lose sensitivity rapidly toward very young rocks because of the low accumulation rates of radiogenic daughter isotopes. Most dating techniques are *radioisotopic* or *isotopic* techniques that are based on the determination of the ratio of the radioactive parent isotope to the radiogenic daughter isotope as a function of time by isotope-ratio mass spectrometry. By contrast, measurement of the parent radioisotope, as in the ^{14}C technique, gives the

highest sensitivity in the youngest materials. While ^{14}C dating is an excellent method for the youngest ~ 25 ka and an adequate technique in the 25–50 ka age range, virtually all radioisotope methods based on isotopes with long half-lives struggle in the youngest part of the timescale (i.e., in the late Quaternary).

Most radioisotopes with long half-lives occur in trace elements in natural substances, for example, ^{87}Rb , ^{235}U , ^{238}U , ^{147}Sm , ^{187}Re , usually in ppm concentration levels. The radioactive isotope of potassium (^{40}K ; half-life 1.25×10^9 years) is the only long-lived radioisotope occurring in a major element of the continental crust (the average continental crust has a K_2O content of 1.1% and the radioactive isotope ^{40}K has an abundance, $^{40}\text{K}/\text{K}_{\text{total}}$, of 0.01167%; [Garnier et al., 1975](#)). Because of the relatively high ^{40}K concentrations in rocks and minerals, the radiogenic daughter isotope ^{40}Ar , commonly designated $^{40}\text{Ar}^*$, grows more rapidly and hence can be more readily detected even in very young rocks. Since the 1950s, the detection limit and the analytical precision of the K/Ar method have increased as the result of technical advances, which now allow for the dating of K-bearing volcanic and magmatic rocks spanning most of the Quaternary.

Historical Development of K/Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ Dating Methods

Radioactivity of potassium has been known since the first decade of the twentieth century ([Campbell and Wood, 1906](#)), and the anomalously high abundance of ^{40}Ar in the Earth's atmosphere was first associated with the radioactivity of natural potassium in the 1930s ([von Weizsaecker, 1937](#)). ^{40}Ar

anomalies in K-bearing geological materials were first reported in the late 1940s, providing the basis for the development of the K/Ar dating method (Aldrich and Nier, 1948). Continuing developments in vacuum and mass spectrometer technology, in particular in Minnesota and Berkeley (Nier, 1950; Reynolds, 1956), allowed the measurement of minute ^{40}Ar anomalies associated with radioactive decay in Quaternary minerals and rocks since the late 1950s. This early interest in dating Quaternary processes is illustrated by the publication of 1.6–1.7 Ma ages for Sutter Butte rhyolite in the Sacramento Valley (Curtis et al., 1956), and for basalt from East African hominid sites at Olduvai Gorge (e.g., Curtis and Evernden, 1962). In the early 1960s, geochronologists at the US Geological Survey, Menlo Park, and at the Australian National University, Canberra, published the first absolute calibrations of the geomagnetic reversal timescale (e.g., Cox et al., 1963; McDougall and Tarling, 1963), thus providing the geology community for the first time with a widely applicable method for the determination of absolute timescales of the Earth's history. This K/Ar calibrated geomagnetic reversal timescale provided the foundations for the scientific proof of sea floor spreading at the mid-oceanic ridges (Vine and Matthews, 1963) and hence the first independent confirmation of the theory of plate tectonics.

The majority of modern K/Ar dating studies are based on the derivative $^{40}\text{Ar}/^{39}\text{Ar}$ technique, where ^{39}Ar is measured as a proxy for ^{40}K , allowing the analysis of smaller single aliquot samples. This approach was developed at the UC Berkeley in the mid-1960s (Merrihue and Turner, 1966). The initial focus of the $^{40}\text{Ar}/^{39}\text{Ar}$ method was to resolve the age information of partially disturbed samples, whereas the exploitation of the potential for increased analytical precision came later (e.g., McDougall, 1981). By the mid-1980s, the University of Toronto team started to use continuous-wave lasers for sample heating instead of furnaces, resulting in a substantial reduction in system blanks. In addition, they replaced the commonly used Faraday collectors with secondary electron multiplier (SEM) detectors, yielding higher sensitivity and thus allowing the detection of 100–1000 times smaller ion-beam intensities. The potential of these new developments for dating the Quaternary was demonstrated in case histories for the Massif Central and Eifel volcanic fields in Europe (Lobello et al., 1987; Van der Bogaard et al., 1987). From the early 1990s, several laboratories acquired the capability to measure young Pleistocene samples with the laser fusion $^{40}\text{Ar}/^{39}\text{Ar}$ method and even explored Holocene volcanism (Hora et al., 2007; Renne et al., 1997; Wijbrans et al., 2011). The most recent advances involve the development of multicollector technology for noble-gas mass spectrometry (Coble et al., 2011; Mark et al., 2009; Rivera et al., 2011), potentially allowing yet another substantial increase in sensitivity.

Independently, French teams have adopted a different approach for measuring extremely small radiogenic ^{40}Ar anomalies by perfecting the conventional K/Ar method (Cassignol and Gillot, 1982). This technique relies on very careful determination of the enrichment in radiogenic ^{40}Ar of unknowns against a standard amount of air argon and has been applied with success to late Pleistocene and Holocene problems (e.g., Quidelleur et al., 2001).

Principle of the Method and Age Calculation

In isotope geochronology, ages can be calculated from the general geochronological age equation. This equation is derived from the observation that the change in the amount of radioactive parent isotope P with time ($-dP/dt$) is proportional to the original amount present (P_0), with λ representing the constant of proportionality ($-dP/dt = \lambda P_0$). In a closed system, the original amount of the parent isotope (P_0) at any time equals the sum of the currently present amounts of parent P and daughter D , that is, $P_0 = P + D$. Solving these equations for t yields the general age equation

$$t = (1/\lambda) \times \ln(D/P + 1) \quad [1]$$

where P is the radioactive parent isotope, D is the radiogenic daughter isotope, t is the time passed since the system became closed in years (*annus*, abbreviated as 'a'), and λ is the decay constant specific for the activity of radioactive isotope P (unit *per annum*, a^{-1} ; Holden et al., 2011).

The decay of ^{40}K (λ) is an example of a dual-decay system, as both ^{40}Ca (~89% by positron (β^+) emission) and ^{40}Ar (~11% by electron capture (e.c.) reactions) are formed as radiogenic daughter isotopes. This means that $P = ^{40}\text{K}$ and $D = (^{40}\text{Ca} + ^{40}\text{Ar})$ in eqn [1]. The total activity of ^{40}K , from which the decay constant is calculated, is the summed activity caused by the e.c. and β^+ emission reactions, and the decay constants thus can be summed with $\lambda = \lambda_{\text{e.c.}} + \lambda_{\beta^+}$. As it is analytically challenging to distinguish the radiogenic ^{40}Ca from the nonradiogenic ^{40}Ca component (natural abundance ^{40}Ca is 96.9% of total Ca) and impractical to measure radiogenic ^{40}Ca and ^{40}Ar simultaneously, the branch yielding ^{40}Ar as radiogenic daughter product ($^{40}\text{Ar}^*$) provides the basis for the K/Ar method.

Taking into account only the $^{40}\text{K} \rightarrow ^{40}\text{Ar}$ decay branch, the general age equation, eqn [1], can be rewritten for the $^{40}\text{K}/^{40}\text{Ar}$ system as

$$t = (1/\lambda) \times \ln[^{40}\text{Ar}/^{40}\text{K} \times \lambda_{\text{e.c.}}/(\lambda_{\text{e.c.}} + \lambda_{\beta^+}) + 1] \quad [2]$$

In the conventional K/Ar technique, ^{40}Ar and ^{40}K are measured independently on different aliquots of the sample. ^{40}K is commonly determined from the total K concentration as measured by analytical chemical techniques, for example, by flame photometry, atomic absorption spectroscopy, optical inductively coupled plasma spectroscopy or X-ray fluorescence spectroscopy techniques, using the natural abundance of ^{40}K to derive an amount of ^{40}K in moles ($^{40}\text{K}/\text{K}_{\text{total}} = 0.01167\%$; Garner et al., 1975). For small samples or in cases of low K concentrations, direct determination of ^{40}K by isotope dilution using a thermal ionization mass spectrometry technique would be appropriate. ^{40}Ar is determined by isotope-ratio mass spectrometry using static analyses in a noble-gas mass spectrometer. Classically, most laboratories used isotope dilution with ^{38}Ar as a tracer isotope of known concentration, but recently laboratories have started to use peak height comparison approaches measuring $^{40}\text{Ar}^*$ beam intensities of unknowns against a mineral standard with a known $^{40}\text{Ar}^*$ content or against ^{40}Ar from an air argon aliquot that comes from a pipette system that has been absolutely calibrated.

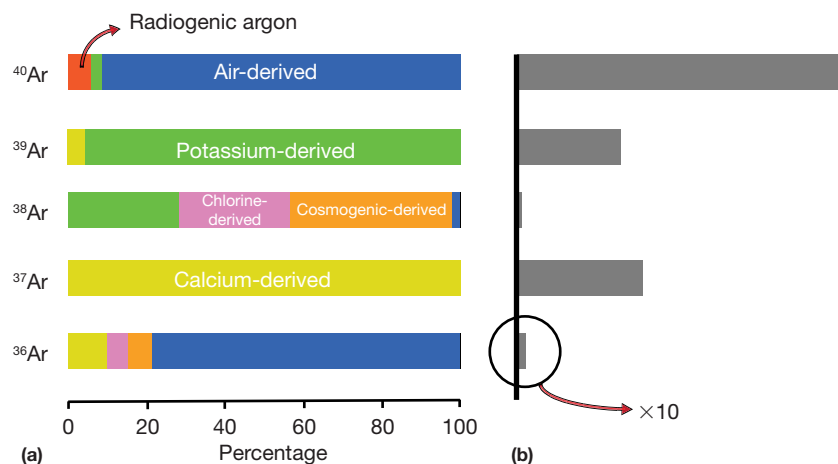


Figure 1 (a) Relative contributions of radiogenic, nucleogenic, and air-contaminant argon isotopes shown for an arbitrary Quaternary sample irradiated for $^{40}\text{Ar}/^{39}\text{Ar}$ age determination. (b) Absolute beam intensity differences in an arbitrary Quaternary sample argon mass spectrum. Note that the ^{36}Ar beam is very small and most challenging to measure accurately.

The $^{40}\text{Ar}/^{39}\text{Ar}$ technique requires a neutron activation step for the reaction of ^{39}K nuclei in the sample with fast (2–7 MeV) neutrons to form ^{39}Ar in an (n,p) reaction. Given that the variation of the $^{40}\text{K}/^{39}\text{K}$ ratio in natural substances is small (<0.2%; Garner et al., 1975), the ^{39}Ar signal after neutron activation becomes a proxy for the radioactive parent isotope, ^{40}K . Thus isotopic ages can be calculated from the measured $^{40}\text{Ar}/^{39}\text{Ar}$ ratios because this ratio contains information on the radiogenic daughter, D (from ^{40}Ar), and radioactive parent isotope, P (from ^{39}Ar), which are required for solving the age equation (eqn [1]).

The $^{40}\text{Ar}/^{39}\text{Ar}$ technique has several important advantages with respect to the conventional K/Ar technique because, in the K/Ar technique, absolute concentration determinations for ^{40}K and for ^{40}Ar have to be carried out on two different splits of a sample, whereas with the $^{40}\text{Ar}/^{39}\text{Ar}$ technique the isotope ratio determination can be obtained from a single sample aliquot. The benefit of this procedure is twofold: (1) isotope ratios can be measured much more precisely than absolute amounts, and with increasing instrumental sensitivity increasingly smaller amounts of sample can be dated; and (2) because the D/P ratio information required for solving the age equation is obtained from isotopes of a gas, argon, incremental heating strategies become possible, such as extraction by stepwise degassing of the samples. Incremental heating can be used to test the homogeneity of radiogenic ^{40}Ar in a mineral when partial disturbance of the signal is suspected. It can be used also to extract information on the local K–Ca–Cl chemistry of individual reservoirs contributing to the degassing of a sample at any given temperature, which is of particular importance when whole rock or groundmass samples are measured.

In the $^{40}\text{Ar}/^{39}\text{Ar}$ technique, the age of unknowns is measured against a standard of known age. The irradiation conditions for the production of ^{39}Ar can be accurately and precisely monitored by the standard if the standard and sample are irradiated together in a carefully defined geometry. An irradiation parameter, commonly designated J , can be defined as

$$J = (e^{\lambda t} - 1) / (^{40}\text{Ar}^* / ^{39}\text{Ar}_K) \quad [3]$$

where t is the absolute age of the standard as determined by other techniques such as conventional K/Ar, U/Pb, or cyclostratigraphic dating based on the expression in the geological rock record of climate variation caused by oscillations in the Earth's orbital parameters, $^{40}\text{Ar}^*$ is the amount of radiogenic daughter isotope in the standard, and $^{39}\text{Ar}_K$ is the amount of ^{39}Ar derived from ^{39}K during irradiation. J depends on the duration of the irradiation, the production efficiency of ^{39}Ar from ^{39}K , and the branching ratio. The J value of the sample is determined by interpolation using the J values of the standards positioned around the sample during irradiation, and eqn [2] can now be rewritten as

$$t = (1/\lambda) \times \ln [J \times ^{40}\text{Ar}^* / ^{39}\text{Ar}_K + 1] \quad [4]$$

where t is the age of the unknown, J is derived from measurement of the $^{40}\text{Ar}^* / ^{39}\text{Ar}_K$ ratio of the standards, and $^{40}\text{Ar}^* / ^{39}\text{Ar}_K$ represents the parent/daughter isotope ratio of the unknown and is measured on a noble-gas mass spectrometer.

Some additional corrections are required. These corrections are graphically represented in Figure 1. For both K/Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ methods, the measured ^{40}Ar intensity needs to be corrected for the nonradiogenic ^{40}Ar component derived from air argon. This correction is very important in Quaternary samples, since the enrichment in radiogenic ^{40}Ar is often relatively low. The correction is generally performed by multiplying the ^{36}Ar intensity by the $^{40}\text{Ar}/^{36}\text{Ar}$ air atmospheric ratio (296, Nier, 1950; 298.7 Lee et al., 2006) and subtracting this value from the ^{40}Ar signal. Unnoticed and/or uncorrected isobaric interferences on the ^{36}Ar beam and the recent discussion on the true value of the $^{40}\text{Ar}/^{36}\text{Ar}_{\text{air}}$ atmospheric ratio (Lee et al., 2006; Valkiers et al., 2010) need to be addressed for precise and accurate application of the $^{40}\text{Ar}/^{39}\text{Ar}$ dating technique in the Quaternary. For the $^{40}\text{Ar}/^{39}\text{Ar}$ method, additional corrections need to be applied for the production of undesired isotopes of argon from K, Ca, and Cl isotopes during neutron activation (Figure 1). Production factors are commonly derived by measuring the isotope ratios of pure K, Ca, and Cl phases such as pure CaF_2 , K_2SO_4 , synthetic anorthite, and K-feldspar glass. The

correction factors for interfering isotopes of Ca tend to be quite stable in most reactors, but the correction factor for the production of ^{40}Ar from ^{40}K during neutron irradiation is more dependent on the fast to thermal neutron ratio and may vary between individual irradiations. Therefore this correction factor needs to be monitored more frequently.

Strengths and Weaknesses

One of the main strengths of the method lies in the abundance of ^{40}K in nature: because this abundance is quite high when compared to other naturally occurring radioactive isotopes, it is possible to apply the K/Ar method over much of the Quaternary. For minerals and rocks high in potassium, the detection limit lies well below 50 000 a, and specialist laboratories are able to reliably measure materials in the 1000–10 000 a range (Quideleur et al., 2001; Renne et al., 1997; Wijbrans et al., 2011), potentially opening the way for intercalibration with ^{14}C , U-series, and optically stimulated fluorescence dating techniques.

The strongest point of the conventional K/Ar method is its relative simplicity and rapid turnaround time, which has made it one of the standard methods for industry contract work. The most significant weakness of the K/Ar technique is that it relies on measurement of an absolute amount of ^{40}Ar that needs to be extracted quantitatively from a rock sample by melting a carefully weighed amount of rock or mineral fragments in a vacuum extraction system. Rock fragments of common basalt need to be carefully screened for alteration, altered glass, and phenocrysts of plagioclase, pyroxenes, and olivine. While alteration removes radiogenic argon, phenocryst phases commonly have significant amounts of extraneous (excess or

inherited) ^{40}Ar . Both effects will lead to biased results, which are often difficult to quantify with the K/Ar method.

The strong points of the $^{40}\text{Ar}/^{39}\text{Ar}$ method are that smaller amounts of sample can be measured more precisely and that, by applying the incremental heating technique, it allows for internal checks for ^{40}Ar inhomogeneity due to alteration, recoil loss of ^{39}Ar , or the presence in the sample of extraneous argon. The main drawback of the technique is the neutron activation step: it adds significantly to the per-sample analysis costs; the turnaround time is increased significantly; undesired neutron-produced argon isotopes interfere with the measurement of the $^{40}\text{Ar}^*/^{39}\text{Ar}_\text{K}$ ratio and have to be quantified; and finally the laboratory operation needs to be certified for handling radioactivity. Further, for absolute dating, the introduction of a mineral standard with known age introduces extra uncertainty. However, for most applications, the advantages of the $^{40}\text{Ar}/^{39}\text{Ar}$ technique outweigh the disadvantages.

New Perspectives

While researchers have used the K/Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ dating techniques with considerable success for many years now, at this moment a number of initiatives have the potential to substantially improve the reliability and the precision of the technique. Both the K/Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ techniques are routinely calibrated against mineral standards of accurately known age. Only a relatively few mineral standards (e.g., USGS biotite BS1, ANU biotite GA1550) have been calibrated absolutely against a carefully determined amount of air argon (Lanphere and Dalrymple, 2000; McDougall and Wellman, 2011). The ^{40}Ar contents of most other standards have been

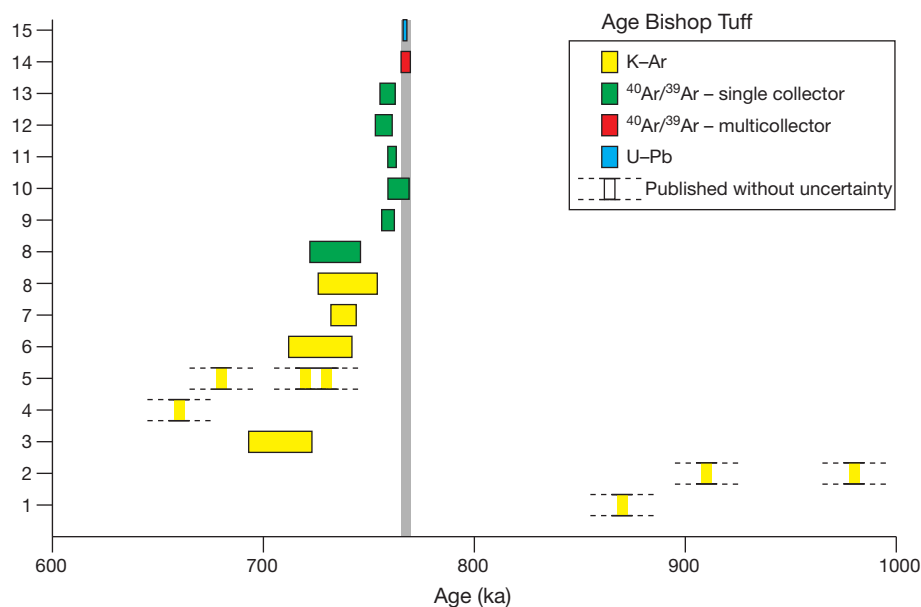


Figure 2 Summary of published ages for the Quaternary Bishop Tuff ash fall. Ages are shown as published in the original publication and not recalculated to currently accepted decay constants or in case of $^{40}\text{Ar}/^{39}\text{Ar}$ dates currently accepted standard ages. Width of the bar represents 2 sigma uncertainty of the determination by Rivera et al. (2011). Data are from Evernden et al. (1957), Evernden et al. (1964), Dalrymple et al. (1965), Huber and Rinehart (1967), Hildreth (1977), Mankinen and Dalrymple (1979), Izett et al. (1988), Hurford and Hammerschmidt (1985), Pringle et al. (1992), Izett and Obradovich (1992), Bogaard and Schirnick (1994), Lueddecke (1998), Sarna-Wojcicki et al. (2000), Rivera et al., 2011, and Crowley et al. (2007).

determined by intercalibration against such 'primary' standards (Fish Canyon sanidine (FCs), Taylor Creek Rhyolite sanidine (TCR, TCR2), Drachenfels sanidine (DRA), Alder Creek, and biotites USGS standard biotite (SB3), Mount Dromedary biotite (MD-2) as examples). Although there is no reason to question the fundamental validity of these calibrations, they were carried out in the 1960s and 1970s and are still generally accepted without considering subsequent improvements in analytical protocols. As a consequence, the accuracies of the $^{40}\text{Ar}^*$ content determinations are substantially larger than the current analytical precisions with which we can determine the K/Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ ages ($\sim 2.0\%$ accuracy against $\sim 0.1\text{--}0.5\%$ analytical precision in favorable cases). Recently, a new initiative was started to accurately determine the absolute $^{40}\text{Ar}^*$ content of mineral standards against a carefully determined absolute amount of pure ^{40}Ar , providing a new metrologically sound basis for the techniques (Morgan et al., 2011). Parallel to this effort, common K/Ar standards have been accurately calibrated against tephra sanidine for which the age of extrusion was determined from the astronomical age of the cyclically bedded sediments into which the tephra was deposited (Kuiper et al., 2008; Rivera et al., 2011), and independent U/Pb data for selected tephra horizons have confirmed this result (Crowley et al., 2007; Wotzlaw et al., 2010). These initiatives combined will help to much improve the accuracy of the technique.

A second exciting prospect is in the development of new instrumentation. During the last 60 years, isotopic determinations were mainly carried out in the single-collector (Faraday cup, or SEM) mode or in some cases dual-collector mode, where the larger signals were measured on a Faraday collector channel and the smaller beams on an SEM. Recently, new mass spectrometer designs have become available with several SEM channels (Nu Instruments Noblesse, ThermoFisher Helix MC) or Faraday collector channels (ThermoFisher Argus-VI). These new instruments will combine faster data acquisition and improved sensitivity, resolution, and analytical precision because individual ion beams will be measured for longer time intervals. Several groups are working on testing data acquisition protocols that will fully exploit these intrinsic improvements (Coble et al., 2011; Jicha et al., 2010; Mark et al., 2009). As already mentioned, enrichment in radiogenic ^{40}Ar in Quaternary minerals is extremely small and its accurate determination relies on precise determination of $^{36}\text{Ar}_{\text{air}}$. Thus, a significant source of error, namely analytical uncertainty and bias, lies in the precision with which we can measure ^{36}Ar . With some of the new multi-collector instruments, the resolution on mass 36 is good enough to partially resolve molecular isobaric interferences such as $^1\text{H}^{35}\text{Cl}$ and $^{12}\text{C}_3$ from ^{36}Ar , and thus derive a better and more accurate determination of the ^{36}Ar beam intensity on mass 36.

As an illustration of technical improvements dating Pleistocene samples, we have plotted ages obtained for the Bishop Tuff in Figure 2. The Bishop Tuff is the product of a great volcanic eruption near Bishop, Long Valley caldera, California, and serves as an important chronostratigraphic marker for correlation and age control in the Quaternary of western United States. This tuff was originally dated by radioisotopic methods in the early 1960s using conventional K/Ar, and a recent study (in 2011) is the first to apply a multicollector data acquisition $^{40}\text{Ar}/^{39}\text{Ar}$ approach to the Bishop Tuff (Rivera et al., 2011).

In summary, these technical developments hold a promise for improvements in dating the late Pleistocene and Holocene. The $^{40}\text{Ar}/^{39}\text{Ar}$ method will be a reliable geochronometer that can be used for dating a wide range of materials in the youngest part of the timescale when these new improvements all have been implemented and tested in the coming years.

See also: Dating Techniques; Fission-Track Dating; U-Series Dating. Radiocarbon Dating: Conventional Method.

References

- Aldrich LT and Nier AO (1948) Argon 40 in potassium minerals. *Physics Reviews* 74: 876–877.
- Bogaard Pvd, and Schirnick C (1994) Single crystal $^{40}\text{Ar}/^{39}\text{Ar}$ ages of quartz protocrysts in the 0.761 Ma Bishop Tuff rhyolite (Long Valley, USA), ICOG-8. *US Geological Survey Circular*, 1107, 32 pp.
- Campbell NR and Wood A (1906) The radioactivity of alkali metals. *Proceedings of the Cambridge Philosophical Society* 14: 15–21.
- Cassinol C and Gillot P-Y (1982) Range and effectiveness of unspiked potassium – Argon dating: Experimental groundwork and applications. In: Odin GS (ed.) *Numerical Dating in Stratigraphy*, pp. 159–179. Chichester: Wiley.
- Coble MA, Grove M, and Calvert AT (2011) Calibration of Nu-Instruments Noblesse multicollector mass spectrometers for argon isotopic measurements using a newly developed reference gas. *Chemical Geology* 290: 75–87. <http://dx.doi.org/10.1016/j.chemgeo.2011.09.003>.
- Cox A, Doell RR, and Dalrymple GB (1963) Geomagnetic polarity epochs and Pleistocene geochronometry. *Nature* 198: 1049–1051.
- Crowley JL, Schoene B, and Bowring SA (2007) U–Pb dating of zircon in the Bishop Tuff at the millennial scale. *Geology* 35: 1123–1126.
- Curtis GH and Evernden JF (1962) The age of basalt underlying Bed I, Olduvai Gorge. *Nature* 194: 610–612.
- Curtis GH, Lipson J, and Evernden JF (1956) Potassium–argon dating of Plio-Pleistocene intrusive rocks. *Nature* 178: 1360.
- Dalrymple GB (1991) The GLM continuous laser system for $^{40}\text{Ar}/^{39}\text{Ar}$ dating: Description and performance characteristics. *USGS Bulletin* 1890: 89–98.
- Dalrymple GB, et al. (1965) *Geological Society of America Bulletin* 76: 665–674.
- Evernden JF, Curtis GH, and Kistler R (1957) Potassium–argon dating of Pleistocene volcanics. *Quaternaria* IV: 13–17.
- Evernden JF, Savage DE, Curtis GH, and James GT (1964) Potassium–argon dates and the Cenozoic mammalian chronology of North America. *American Journal of Science* 262: 145–198.
- Garner EL, Murphy TJ, Gramlich JW, Paulsen PJ, and Barnes IL (1975) Absolute isotopic abundance ratios and atomic weight of a reference sample of potassium. *Journal of Research of the National Bureau of Standards, Section A-Physics and Chemistry* 79: 713–725.
- Hildreth EW (1977) The magma chamber of the Bishop Tuff: Gradients in temperature, pressure, and composition PhD thesis, 328 pp. University of California, Berkeley.
- Holden NE, Bonardi ML, de Bièvre P, Renne PR, and Villa IM (2011) IUPAC-IUGS common definition and convention on the use of the year as a derived unit of time (IUPAC Recommendations 2011). *Pure and Applied Chemistry* 83: 1159–1162. <http://dx.doi.org/10.1351/PAC-REC-09-01-22>.
- Hora JM, Singer BS, and Wörner G (2007) Volcano evolution and eruptive flux on the thick crust of the Andean Central Volcanic Zone: $^{40}\text{Ar}/^{39}\text{Ar}$ constraints from Volcan Paríncota, Chile. *Geological Society of America Bulletin* 119: 343–362.
- Huber NK, Rinehart CD (1967) Cenozoic volcanic rocks of the Devils Postpile quadrangle, eastern Sierra Nevada, California. *US Geological Survey Professional Paper*, 554-D, pp. D1–D21.
- Hurford AJ and Hammerschmidt K (1985) $^{40}\text{Ar}/^{39}\text{Ar}$ and K/Ar dating of the Bishop and Fish Canyon tuffs: Calibration ages for fission-track dating standards. *Chemical Geology* 276: 257–268.
- Izett GA and Obradovich JD (1991) Dating of the Matuyama-Brunhes boundary based on $^{40}\text{Ar}/^{39}\text{Ar}$ ages of the Bishop Tuff and Cerro San Luis rhyolite. *Geological Society of America Abstracts* 23: A106.
- Izett GA, Obradovich JD, and Mehnert HH (1988) The Bishop ash bed (middle Pleistocene) and some older (Pliocene and Pleistocene) chemically similar ash beds in California, Nevada, and Utah. *USGS Bulletin* 1675: 37.

- Jicha BR, Sobol P, and Singer BS (2010) The UW-Madison 5-collector mass spectrometer for high-precision $^{40}\text{Ar}/^{39}\text{Ar}$ geochronology. AGU Fall meeting, abstract #V31A-2299.
- Kuiper KF, Deino A, Hilgen FJ, Krijgsman W, Renne PR, and Wijbrans JR (2008) Synchronizing rock clocks of earth history. *Science* 320: 500–504.
- Lanphere MA and Dalrymple GB (2000) *First-Principles Calibration of ^{39}Ar Tracers: Implications for the Ages of $^{40}\text{Ar}/^{39}\text{Ar}$ Fluence Monitors*. Menlo Park, CA: US Geological Survey. Professional paper 1621.
- Lee JY, Marti K, Severinghaus JP, et al. (2006) A redetermination of the isotopic abundances of atmospheric Ar. *Geochemica et Cosmochimica Acta* 70: 4507–4512. <http://dx.doi.org/10.1016/j.gca.2006.06.1563>.
- Lobello Ph, Feraud G, Hall CM, York D, Lavina P, and Bernat M (1987) $^{40}\text{Ar}/^{39}\text{Ar}$ step heating and laser fusion dation of a Quaternary pumice from Neschers, Massif Central France: Defeat of the xenocrystic contamination. *Chemical Geology (Isotope Geoscience)* 66: 61–71.
- Lueddecke SB, Pinter N, and Gans P (1998) Plio-Pleistocene Ash Falls, Sedimentation, and Range-Front Faulting along the White-Inyo Mountains Front, California. *Journal of Geology* 106: 511–522.
- Mankinen EA and Dalrymple GB (1979) Revised geomagnetic polarity time scale for the interval 0–5 m.y. B.P. *Journal of Geophysical Research* 84: 615–626.
- Mark DF, Barfod D, Stuart FM, and Imlach J (2009) The ARGUS multicollector noble gas mass spectrometer: Performance for $^{40}\text{Ar}/^{39}\text{Ar}$ geochronology. *Geochemistry, Geophysics, Geosystems* 10: Q0AA02.
- McDougall I (1981) $^{40}\text{Ar}/^{39}\text{Ar}$ age spectra from the KBS Tuff, Koobi Fora formation. *Nature* 294: 120–124.
- McDougall I and Tarling DH (1963) Dating of polarity zones in the Hawaii Islands. *Nature* 200: 54–56.
- McDougall I and Wellman P (2011) Calibration of GA1550 biotite standard for K/Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ dating. *Chemical Geology* 280: 19–25.
- Merrihue C and Turner G (1966) Potassium Argon dating by activation with fast neutrons. *Journal of Geophysical Research* 71: 2852–2857.
- Morgan LE, Postma O, Kuiper KF, et al. (2011) A metrological approach to measuring $^{40}\text{Ar}^*$ concentrations in geological materials. *Geochemistry, Geophysics, Geosystems* 12: 17 pp, A0AA20. <http://dx.doi.org/10.1029/2011GC003719>.
- Nier AO (1950) A redetermination of the relative abundances of the isotopes of carbon, nitrogen, oxygen, argon and potassium. *Physics Reviews* 79: 450–454.
- Pringle MS, McWilliams M, Houghton BF, Lanphere MA, and Wilson CJN (1992) $^{40}\text{Ar}/^{39}\text{Ar}$ dating of Quaternary feldspar: Examples from the Taupo volcanic zone, New Zealand. *Geology* 20: 531–534.
- Quidelleur X, Gillot PY, Soler V, and Lefevre JC (2001) K/Ar dating extended into the last millennium: Application to the youngest effusive episode of the Teide volcano (Spain). *Geophysical Research Letters* 28: 3067–3070. <http://dx.doi.org/10.1029/2000GL012821>.
- Renne PR, Deino AL, Walter RC, et al. (1994) Intercalibration of astronomical and radioisotopic time. *Geology* 22: 783–786.
- Renne PR, Sharp WD, Deino AL, Orsi G, and Civetta L (1997) $^{40}\text{Ar}/^{39}\text{Ar}$ dating into the historical realm: Calibration against Pliny the Younger. *Science* 277: 1279–1280.
- Reynolds JH (1956) High sensitivity mass spectrometer for rare gas analysis. *Review of Science Instruments* 27: 928.
- Rivera TA, Storey M, Zeeden C, Hilgen FJ, and Kuiper K (2011) A refined astronomically calibrated $^{40}\text{Ar}/^{39}\text{Ar}$ age for fish Canyon sanidine. *Earth and Planetary Science Letters* 311: 420–426.
- Sarna-Wojcicki, et al. (2000) *Journal of Geophysical Research* 105(B9): 21431–21443.
- Valkiers S, Vendelbo D, Berglund M, and De Podesta M (2010) Preparation of argon primary measurement standards for the calibration of ion current ratios measured in argon. *International Journal of Mass Spectrometry* 291: 41–47. <http://dx.doi.org/10.1016/j.ijms.2010.01.004>.
- Van der Bogaard P, Hall CM, Schmincke HU, and York D (1987) $^{40}\text{Ar}/^{39}\text{Ar}$ Laser dating of single grains – Ages of Quaternary tephra from the East Eifel volcanic field, FRG. *Geophysical Research Letters* 14: 1211–1214.
- Vine FJ and Matthews DH (1963) Magnetic anomalies over oceanic ridges. *Nature* 199: 947–949.
- von Weizsaecker CF (1937) Ueber die möglichkeit eines dualen Beta-Zerfalls von Kalium. *Physikalische Zeitschrift* 38: 623.
- Wijbrans J, Schneider B, Kuiper K, et al. (2011) $^{40}\text{Ar}/^{39}\text{Ar}$ geochronology of Holocene basalts: examples from Stromboli, Italy. *Quaternary Geochronology* 6: 223–232.
- Wotzlaw J, Schaltegger U, Kuiper K, and Guenther D (2010) Deriving accurate eruption ages from complex zircon populations: Insights from zircon trace element chemistry and intercalibration with astronomical time. AGU Fall Meeting 2010, abstract #V31A-2312.



LAKE LEVEL STUDIES

Contents

Overview

Africa during the Late Quaternary

Asia

Australia

Latin America

North America

West-Central Europe

Modeling

Overview

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Introduction

The main focus of paleoclimate research over the last 50 years has been temperature change reconstruction, driven by the dramatic results produced by ocean and ice-core research. Paleohydrological research in comparison has received significantly less attention. As a result, the range of proxy techniques available to the paleohydrologist is significantly less than that available to those interested in past temperature change. This situation is perplexing when one considers how much more important is water availability than temperature to the vast majority of the world's population. Over the past 10 years, this situation has begun to change, due in part to the needs of the modeling community, which increasingly requires hydrological data sets (e.g., precipitation, evaporation, and surface moisture status) for climate model development and validation.

Lake-level research represents a critical paleohydrological archive. Throughout the Quaternary period, lakes have

fluctuated in size in response to climatically induced changes in hydrological balance over the lake–catchment system. The ubiquity of lake basins across the globe provides researchers with an opportunity to study paleohydrological data over a wide variety of spatial (local–continental–hemispheric) and temporal (10^1 – 10^6 years) scales. This spatial dimension is particularly important as the ability to distinguish patterns of lake-level change at the continental scale represents a starting point for elucidating the mechanisms of climate change (e.g., [Steffen et al., 2003](#)).

The roots of lake-level research probably lie in the investigation of the vast pluvial lakes that used to intermittently dominate the landscape of southwest United States, intertropical Africa, and Australia throughout the Quaternary. The existence of these ‘prehistoric’ lakes has been well documented for over a century. In Australia, detailed surveying recorded the full extent of Lake Eyre by the 1870s. In the 1840s, in southwest United States, G. K. Gilbert of the USGS began work on a

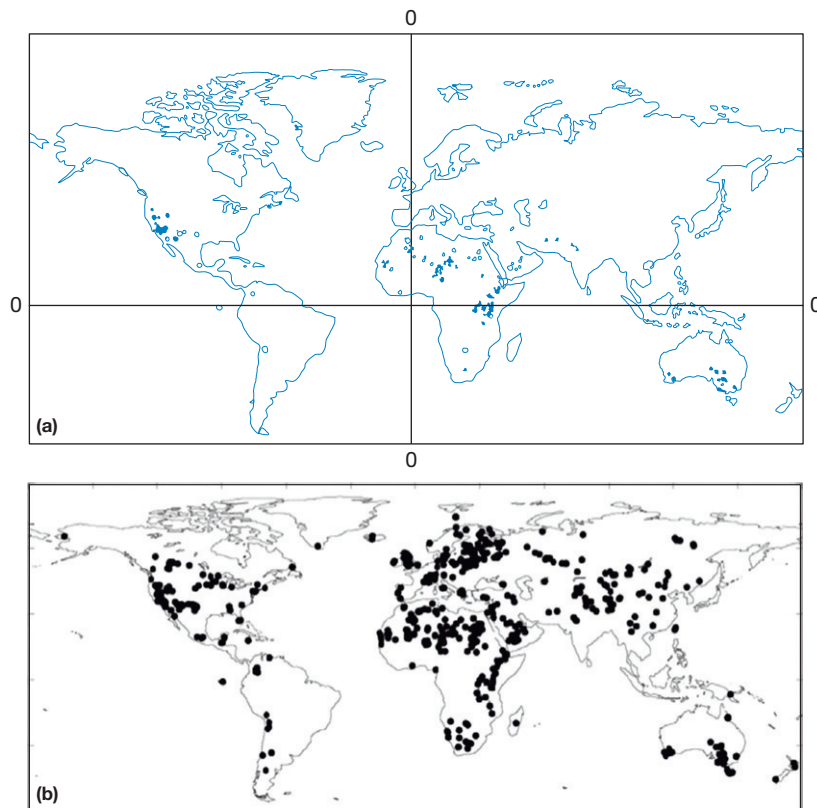


Figure 1 Comparison of the distribution of data points for (a) the first global lake status database published in 1979 and (b) the more recent situation in 1999. In 30 years, the database has quadrupled in size and continues to grow (GSLDB, 1999; Street and Grove, 1979).

sequence of ancient lake sediments deposited across northwest Utah, northwest Nevada, and southeast Idaho. In his final report to the USGS in 1890, he accurately calculated the area (20 000 square miles) and depth (1000 ft) of the ancient lake, the dramatic lowering through catastrophic flood, and also determined that crustal rebound must have been responsible for the elevated shorelines he observed. He named the site Lake Bonneville after an earlier explorer in the region. In 1949, Ernst Antevs was among the first to suggest that pluvial phases in the Great Basin lakes of southwest United States were linked to the presence of continental ice sheets over North America. It was not until the late 1970s, however, that the connection between pluvial and interpluvial phases was linked to Quaternary glacial cycles.

Advances in ^{14}C dating and drilling technology in the 1960s and 1970s saw an increase in lake-level studies, and slowly, a continental picture of events began to emerge. The first synthesis of continental data was produced for Africa in the late 1970s by Street and Grove (1976), followed 3 years later by the publication of the first global synthesis of lake-level data (Street and Grove, 1979). Data from 141 closed basin records spanning the period 1958–79 were collated and synthesized and a series of global maps were produced, which provided the first global overview of changing lake status for the last 30 ka. For the first time, it was possible to see how climate change occurred on the global scale.

Though the original 1979 paper was pitched as a ‘global’ synthesis, this title was somewhat misleading as data for large parts of the globe were missing (Figure 1(a)). In the case of

Russia and China, this was due to political reasons, but for much of the mid- and high latitudes the dominance of open lake systems posed a problem. Such problems were soon overcome with the development of new methods for open lake-level reconstruction such as the ‘Digerfeldt technique’ outlined below. The global database continued to expand and develop, culminating in the global lake status database (GSLDB; e.g., Kohfeld and Harrison, 2000), with a resource of over 600 sites spread across the globe (Figure 1(b)). Detailed continental scale syntheses are now available for most parts of the globe. There has been a marked improvement in temporal resolution as well, highlighted by the recent publication of the first high-resolution study of lake-level variation in west-central Europe (e.g., Magny et al., 2004; Figure 2). By combining the lake-level data with pollen, they were able to quantitatively reconstruct various climatic parameters, including mean annual temperature and annual precipitation, for the Holocene.

Next, we present an overview of the approach to lake-level reconstruction, focusing on the various sedimentary archives and reconstructive techniques that can be used in such studies.

Reconstructing Past Lake-Level Change

Background

The use of past records of lake level as a proxy for paleohydrological change is based on the close relationship that exists between lake volume/area and the balance of precipitation:

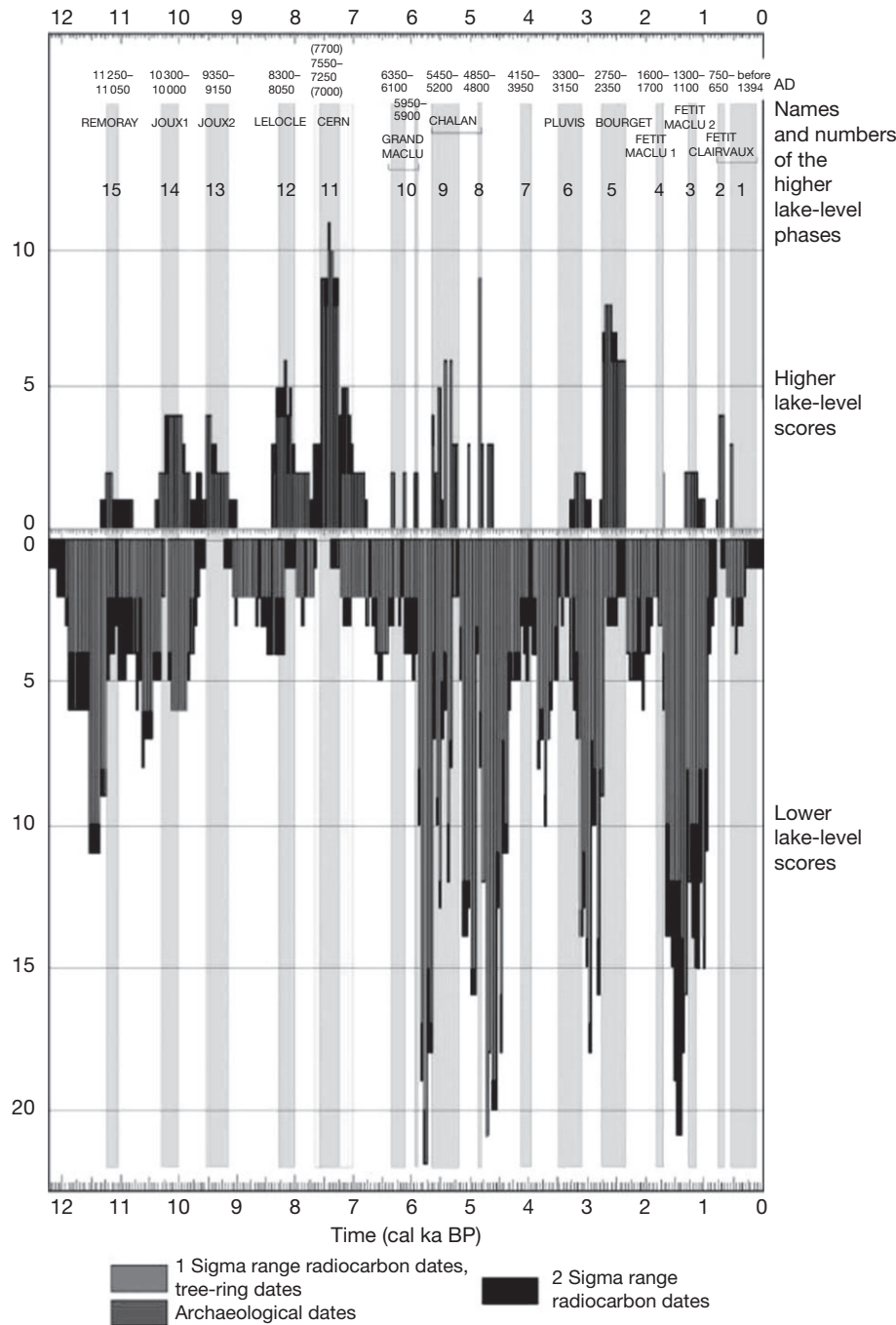


Figure 2 High-resolution Holocene lake-level record for the Jura mountains, the northern French pre-Alps, and the Swiss Plateau by Magny (2004). The vertical scales represent the number of dates for successive 50-year intervals between 12.25 and 0 cal ka BP.

evaporation ($P-E$) across a catchment. In closed (endorheic) 'nonoverflowing' lakes, particularly in semiarid regions, lake area represents the equilibrium between runoff from the catchment and the water deficit over the lake. This can be summarized in the following equation:

$$P + R + G_1 = E + O + G_0 \pm \Delta S$$

where P is direct precipitation onto the lake surface, R is catchment runoff, G_1 is groundwater input, E is evaporitic loss from the lake surface, O is surface outflow, G_0 is groundwater outflow, and ΔS is the change in storage (Street-Perrott

and Harrison, 1985). The situation is more complex with respect to open (exorheic) systems, which have a permanent inflow/outflow and tend to occur in regions where runoff dominates. Under such circumstances, a positive change in water balance may simply result in increased discharge at the outflow. Open systems are therefore likely to be much more insensitive to hydrological change, particularly over shorter time periods. Nevertheless, relict high shorelines around currently overflowing lakes in temperate regions suggest that increased discharge at the outflow is not always sufficient to compensate for changes in water balance (Harrison et al., 1991).



Figure 3 (a) Satellite image of the Lake Erhai catchment, northern Yunnan. The lake is bordered to the West by the Dali Plain (A), which is composed of old lake sediment deposited during an early Pleistocene highstand phase. The lake is tectonic in origin and the basin is currently subsiding in relation to the Diangchang Mountains, which abut the Dali Plain. The north of the lake is dominated by the Miju Delta (B), testament to the rapid infilling of the lake due to the high rates of erosion in the catchment. A selection of smaller lakes are scattered across the northern catchment, being further evidence of the previous highstand (C). Two former lake basins can be observed to the south of Erhai (D); both of these have been drained for agricultural land. (b) Xierhe river/rock bar looking upstream to Xiaguan and Erhai. The original outflow relates to a period of high lake level during the early Pleistocene. The present outflow is man made – another example of anthropogenic modification of the catchment hydrology. (c) Ceramic irrigation pipes and a decorative plate showing wet field irrigation; both are found in the catchment. Dating to the early Han dynasty (206 BCE–CE 220), they illustrate that even at this early stage a well-organized extensive wet field irrigation system was established in the catchment, drawing much of its water from the lake and the 18 streams which feed of the mountains. (d) Map of the Dali plain sent to the British Government by an envoy of the Muslim Suliman ca. 1871. The Miju Delta, a prominent feature of the modern lake, is conspicuous by its absence, suggesting it as a very recent phenomenon. A detailed environmental history of the Erhai catchment can be found in [Elvin et al. \(2002\)](#).

Long-term changes in 'lake status' (lake level, area, and depth) will affect a range of physical and biological processes in a lake and so can be reconstructed from geomorphic and lithostratigraphic records ([Harrison and Digerfeldt, 1993](#); [Street-Perrott and Harrison, 1985](#)). Water-level fluctuations affect space and habitat availability, the extent of the photic zone (the zone below the surface which allows enough light penetration in order for

photosynthesis to take place), water chemistry, stratification, and mixing regimes. In shallow lakes, falling lake levels may prevent thermal stratification and increase salt concentrations, in turn forcing a change in lake pH. Turbulence may also increase, leading to the reworking of exposed marginal sediments into the lake.

Disentangling and interpreting the complex set of processes that led to the preservation of a climatic signature within a

sediment record is a complex process. Such complexity arises from the fact that 'signals' preserved within the sedimentary record could be the result of a variety of internal and external factors unrelated to climate-driven changes in water balance. Significant changes in lake level can be triggered by a variety of nonclimatic factors, including natural infilling, land cover change, stream channel development, and anthropogenic activity. If an accurate interpretation of a sedimentary record is to be achieved, a clear understanding of the hydrological and sedimentological process at work across in the lake–catchment system is critical. A working knowledge of the geological history of an area is particularly important, especially in tectonically active areas. Tectonic/neotectonic processes are a relatively common cause of lake-level change, either directly through subsidence/uplift or indirectly through earthquake-induced rockfalls.

For Holocene-based studies, an understanding of past anthropogenic activity is critical, particularly in areas where there has been a long history of land-use change and modification. [Figure 3\(a\)–3\(d\)](#) highlights the potential role tectonic and human activity have played in influencing past lake levels at Erhai Lake, Yunnan province, China, and in doing so reinforces the importance of undertaking a detailed site survey prior to sampling.

Site Selection

For any given region, it is likely that only a small percentage of lake–catchment systems will be suitable targets for lake-level research. The ideal site will be the one that strikes the right balance between 'sensitivity,' 'longevity,' and the 'temporal resolution' of the sedimentary record. The approach taken will be largely dependent on whether one is dealing with a 'freshwater' or a 'saline' lake system as the marked differences that exist between their hydrological and sedimentological processes often warrant a very different approach to reconstruction.

The vast majority of freshwater lakes are located in the 'mid–high latitudes' where evaporation rates are comparatively low and precipitation is high. Most of the lakes in these regions are open systems, which may be much less sensitive to hydrological change than closed systems, but are less likely to suffer from desiccation, therefore increasing the chances of obtaining a complete sedimentary record. The sedimentary record for most of these lakes tends to be clastic dominated and comparatively organic rich. Freshwater carbonates are also common in karstic areas.

Most mid–high-latitude lakes are glacial in origin, and as such the sediment record normally only extends back to the last Pleistocene glaciation. Comparatively high rates of sedimentation, particularly in the mid-latitudes, can provide an excellent temporal resolution for this period. As one moves into the higher latitudes, sedimentation rates tend to decline, the colder climate and seasonal ice cover limiting biological activity and hence autochthonous sedimentation ([Lamoueux and Gilbert, 2004](#)). A number of freshwater sequences that extend beyond 18 ka, such as the European Maar lake sequences, which typically extend back to 100 ka, have been identified. Significantly older sedimentary records can be found in some of the world's greatest lakes, such as Lake Titicaca, formed ~2 Ma, and Lake Baikal, which is thought to

be the world's oldest lake with a sediment record thought to extend over the last 7 My.

In comparison, saline lakes are mainly found in the sub-humid, semiarid/arid regions (e.g., North American Great Plains, southeast Australia, and Africa) and at altitudes (e.g., Qinghai–Tibetan plateau and Altiplano of South America) where evaporation rates are high and precipitation is infrequent or very strongly seasonal. Most saline lakes are located in fault-bounded, 'closed' basins. Seasonal changes in precipitation often result in dramatic shifts in lake status between enlarged perennial lakes (wet) and playa deflation basins (dry). Consequently, the sedimentary record is often quite complex, consisting of a mix of evaporites, geochemical precipitates, catchment-derived clastic material, and eolian deposits.

Located in unglaciated areas, many saline lakes have the potential to provide long sedimentary records stretching back through the last glacial–interglacial cycle and beyond. Such long-core records reveal a paleohydrological record for the Quaternary characterized by repeated phases of lake contraction and expansion, reflecting their sensitivity to global climate change events. Owing to the unique hydrological and sedimentary processes that characterize saline-lake systems, they are widely regarded as the most sensitive of sites for paleohydrological research (e.g., [Street-Perrott and Harrison, 1985](#)). Unfortunately, owing to the high rates of evaporation and limited precipitation, saline lakes are comparatively rare.

Site Survey

For sites with an extant lake, a site survey should begin with an assessment of the basin topography, either manually using a weight and line or, if available, a digital echo-sounder. Shallow-sided basins are the preferred choice, as any change in lake level will be accompanied by a much more significant shift in surface area than that in a steeper sided basin. Slumping of marginal sediments is also much more problematic in steep-sided basins, particularly in tectonically active areas where earthquakes are common.

If possible, a subsurface survey of the sedimentary deposits should be undertaken and a variety of seismic reflection/acoustic profiling techniques can be used depending on the size of the lake basin under investigation and the required balance between resolution and penetration can be determined by the nature of the project. Though the equipment is expensive and often unwieldy, it can provide a detailed picture of the subsurface stratigraphy. Such information can be used to help characterize the depositional process at work (e.g., sediment focusing and faulting), which can then be used to identify the best coring site. If access to such equipment is not possible, a good estimate of depositional trends can be gained through the application of wave theory. All that is required is an understanding of basin hydrology and local meteorological conditions, particularly wind speed and direction ([Dearing, 1997](#)).

A detailed limnological survey of the site should also be carried out to provide an insight into the hydrological and sedimentological processes at work in the basin. Sedimentary processes in the lake can be significantly altered by a variety of factors unrelated to water balance, including wind strength (frequency and exposure), gravity, and heat transfer, all of which would be expected to vary over time ([Dearing, 1997](#)).

A detailed overview of the current situation can therefore aid in future interpretation of the sediment record.

Techniques for Lake-Level Reconstruction

It is now well established that, when working with lake sediments, any interpretation of lake-level changes must be based on multiple and supporting lines of evidence (e.g., [Harrison and Digerfeldt, 1993](#)). The wide diversity of sites precludes the development of a single sampling protocol, and the choice of archives to be analyzed and techniques to be used will ultimately be site specific. The majority of lake-level studies are based on a combination of 'geomorphological' (catchment sediments) and 'lithological' (lacustrine sediments) archives, which can be analyzed using a combination of 'lithostratigraphic' and 'biostratigraphic' reconstructive techniques. A summary of the different approaches to lake-level reconstruction is presented below with reference to freshwater and saline lakes. The marked differences in origin, hydrology, and sedimentary processes within such lakes often require a very

different approach to lake-level reconstruction and as such serve as a useful way of highlighting the various approaches available.

Geomorphic Archives

Historically, fluctuations in lake level were first noted from the presence of abandoned shorelines formed during periods of lake highstands. As highlighted by [Smith and Street-Perrott](#)



Figure 4 Images of the closed basin lake Dangqiong Cuo, Tibetan Plateau, China. Clearly visible in the foreground in both (a) and (b) are a series of concentric arcuate shorelines, geomorphic evidence of former higher lake levels. In the far distance, again in both (a) and (b), are a series of long-abandoned shorelines or terraces that drape the lower parts of the surrounding mountains, evidence of much higher paleolake levels. Photographs from NIGL, China.

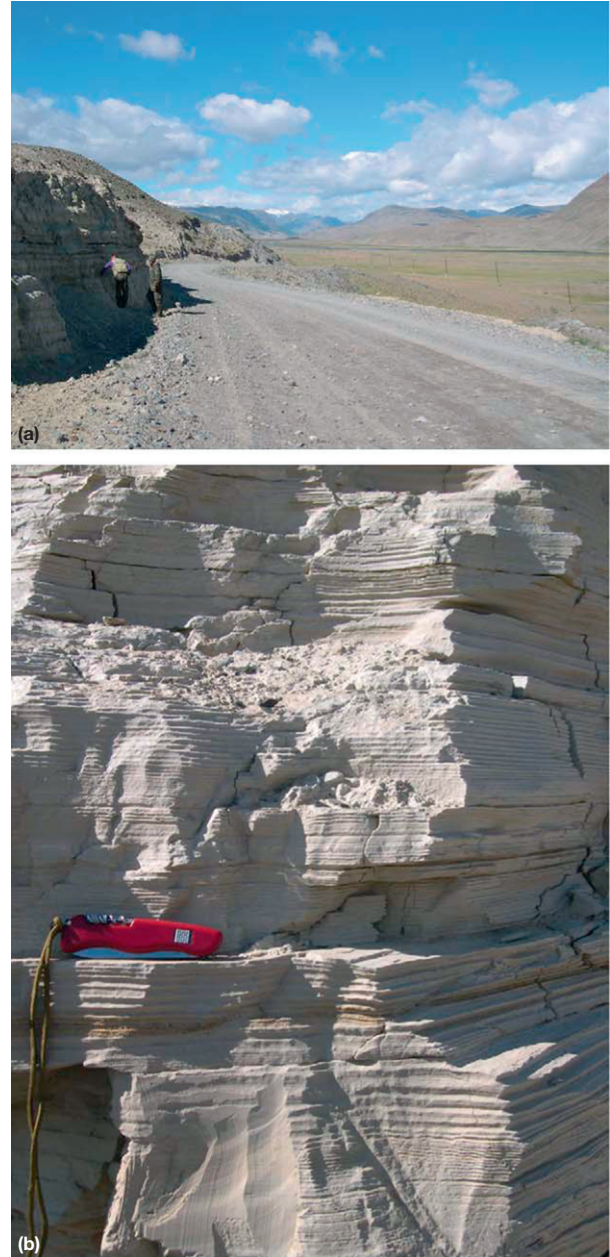


Figure 5 Altai Region, Siberia, current lake visible in far distance. (a) Roadside exposure of lacustrine sediments deposited during glacial highstand, with geomorphic evidence pointing toward catastrophic jökulhlaup causing drop in lake level. (b) Close-up of roadside varved sediments, a characteristic deposit of glacial lakes in the region. Photographs courtesy of Dr. Philipp Allen.

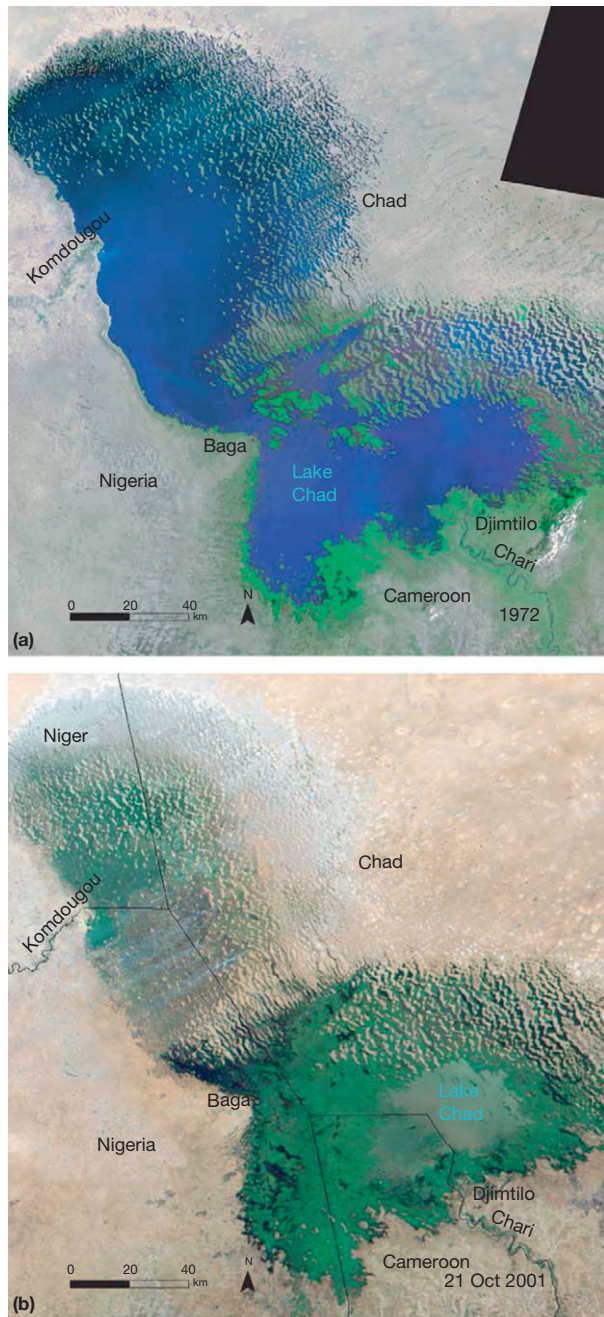


Figure 6 Satellite images of Lake Chad taken in 1972 and 2001 showing the reduction in lake level and the further exposure of dune fields in the northeast. The shoreline of Mega Lake Chad is clearly visible. The images are taken from *The African Lakes Atlas*, UNEP, 2005.

(1983) and Sack (2001), landforms are broadly represented by the deposition of sediment or remnants of the erosion of sediment; therefore, geomorphology and sedimentology were key to early lake-level studies.

The interpretation of geomorphological evidence for changes in lake level is still as important as it ever was, but it is now universally augmented by the use of other lithological techniques discussed elsewhere in this article. The idea of a multiproxy approach to lake-level studies should not be

diminished; however, an understanding of the various geomorphic lines of evidence is needed in order to produce something of a baseline for further study. During the initial site investigation and mapping phase, geomorphic features should be identified. This stage may involve maps, aerial photographs, and remote/satellite imagery. The next stage would involve the detailed surveying of all features to a known datum so that altitudinal changes can be inferred. If applicable to the feature, sedimentological work and dating should also be carried out.

The geomorphology of a lake basin and especially the former shorelines is effectively a measure of the interaction between the water body and the landscape it is situated in. Sack (2001) provides an extended explanation of the use of shorelines and basin configuration for paleolimnological reconstruction. The presence or lack of an outflow in hydrographically open or closed lake basins, respectively, serves to control the geomorphological response to increased or decreased water input to the system. Open basin lakes respond to increases in water input by varying discharge via flow velocity and outflow width, whereas closed basin lakes express changes in input as alterations in lake-level and surface-area fluctuations. The result of this is that closed basin lakes tend to preserve fluctuations in lake level in the form of discrete shorelines (Figure 4(a) and 4(b)), whereas open basin systems are less likely to produce such identifiable geomorphological features. The catastrophic draining of a lake, possibly as a Jökulhlaup, often leaves behind remnant features such as stranded clifflines, shorelines, and raised lake sediments (Figure 5(a) and 5(b)). Places where modern lake levels are low enough, the exposure of other geomorphological features may also occur. The Chad Basin, North Africa, is a well-documented example of large-scale lake-level change (Grove and Warren, 1968). The basin now contains extensive dune systems within the shoreline of the former 'Mega Lake Chad.' Modern satellite imagery of Lake Chad shows the increased reduction in the extent of the lake and clearly highlights the dune fields present within the modern lake basin (Figure 6(a) and 6(b)).

In general, geomorphic features are discontinuous in terms of the entire perimeter of a lake. The mapping of paleoshorelines rarely allows entire lake levels to be reconstructed. However, with the introduction of digital elevation models and their incorporation into a GIS, shorelines can be extrapolated and 'fit' to existing maps. An example of this is shown by DeVogel et al. (2004) who combined the mapping and surveying of Lake Eyre, central Australia, with a digital elevation model, which in turn was used to produce a number of maps of the paleolevels of the entire lake basin (Figure 7).

Lithological Archives

Freshwater clastic lake sediments

The use of lake sediments from open freshwater lake systems for lake-level reconstruction has always been somewhat controversial. Dearing and Foster (1986), for example, consider that the level of a lake with a permanent inflow is unlikely to respond to climatic changes except for a change in the rate of discharge at the outflow. Owing to a paucity of targeted lake-level studies, many of the early open lake submissions to the GLSDB were reinterpretations of published data. Very few of these studies were targeted lake-level reconstructions, the vast

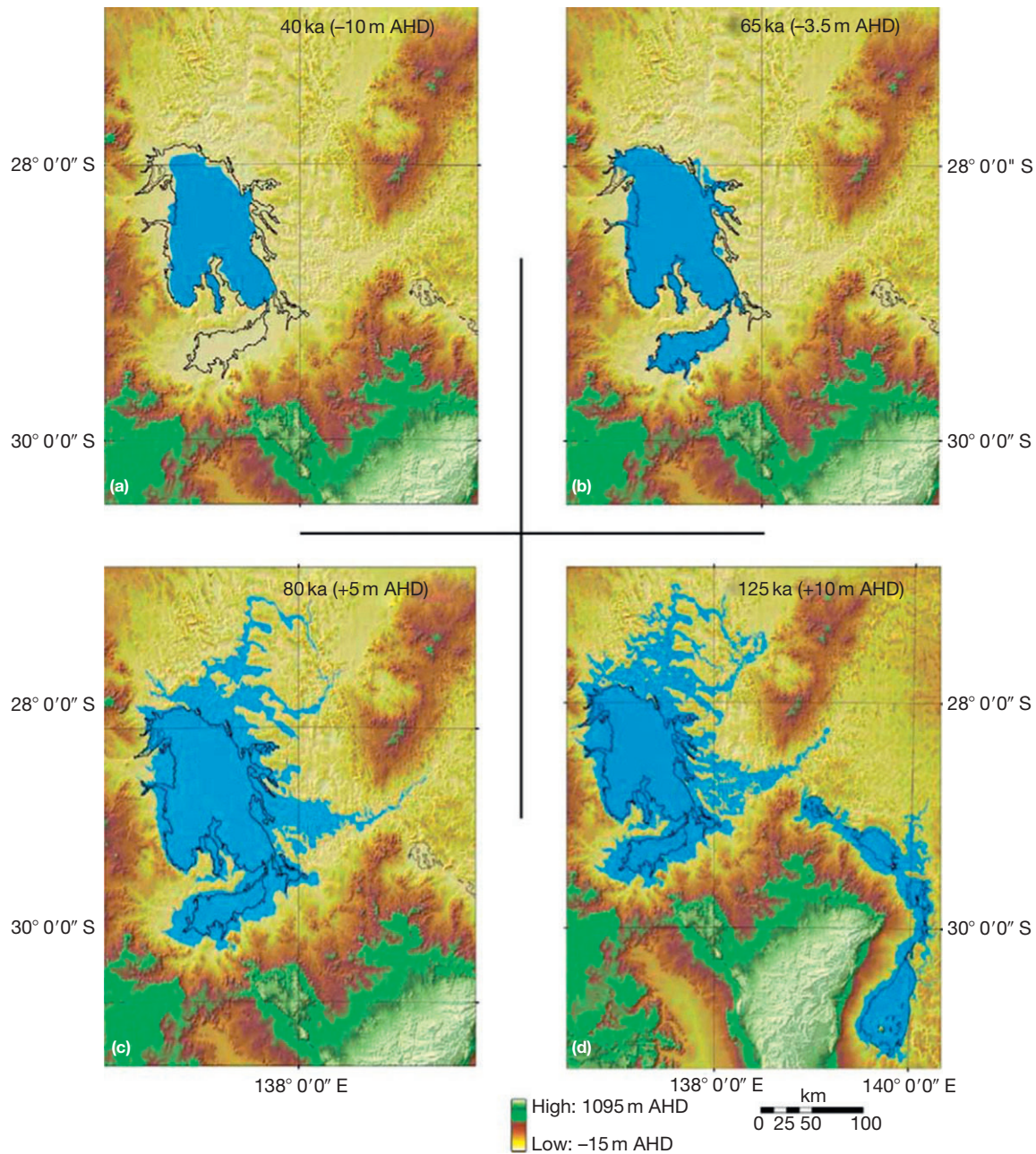


Figure 7 Maps depicting the outline of Lake Eyre at known paleolevels. The modern playas are outlined in black with the paleolake shown in blue. At the peak highstand in 125 ka (stage 5e), Lake Eyre would have covered a total area of $\sim 34\,000\text{ km}^2$, holding up to 430 km^3 of water. Reproduced from DeVogel SB, Magee JW, Manley WF, and Miller GH (2004) A GIS-based reconstruction of late Quaternary paleohydrology: Lake Eyre, arid central Australia. *Palaeogeography, Palaeoclimatology, Palaeoecology* 204: 1–13.

majority being pollen-based paleoenvironmental studies. The situation was further compounded by the fact that very little information was available regarding catchment hydrology and sedimentation processes. With runoff being a dominant factor controlling lake level, a detailed understanding of catchment hydrology and sedimentary processes is critical.

The situation improved significantly in the early 1980s with the development of a new technique for lake-level reconstruction designed by Gunnar Digerfeldt, specifically for use on open lake systems (Figure 8; Digerfeldt, 1986, 1988). The key

to the ‘Digerfeldt’ approach lay in the fact that it combined three main lines of stratigraphic data: (1) changes in the sediment limit, (2) sediment composition and type, and (3) distribution of littoral vegetation.

During a period of stable lake level, a ‘sediment limit’ will be established, above which fine sediment and organic matter cannot accumulate due to wind and wave action. Above this limit, the sediment record tends to be dominated by coarse sediments ($>0.125\text{ mm}$). A future drop in lake level should result in the establishment of a new sediment limit, the

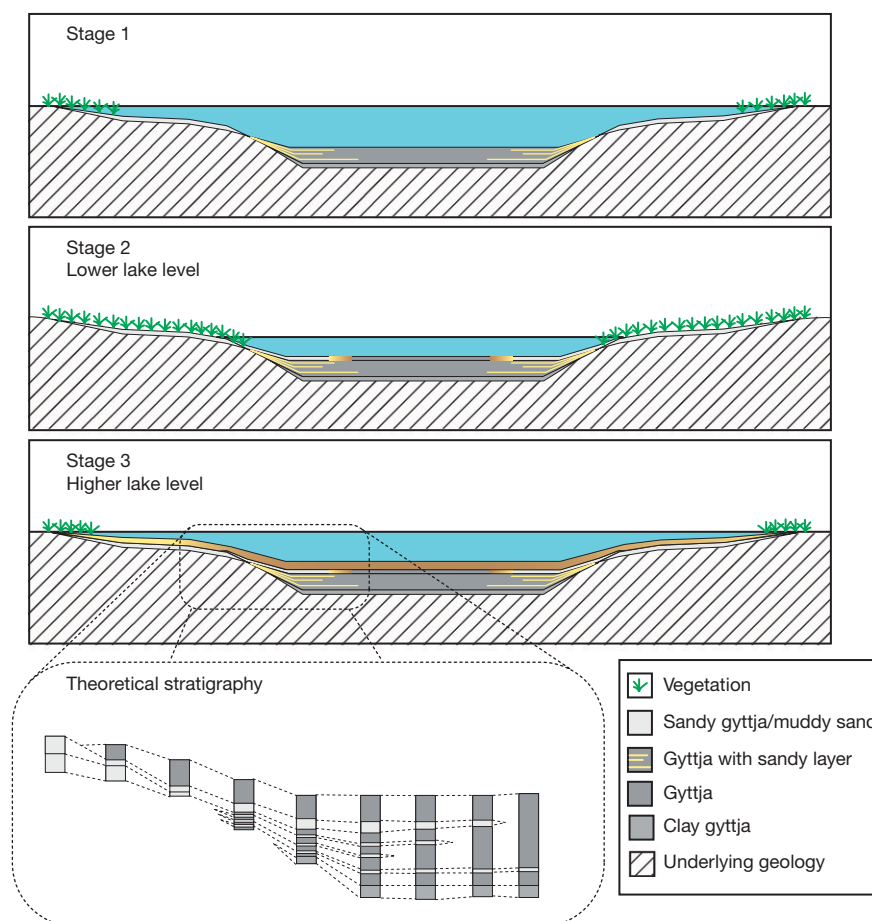


Figure 8 Diagrammatic representation of the lake-level reconstruction approach of Digerfeldt (1986). The three temporal stages shown document the fall and then the subsequent rise in lake level, sedimentation, and the associated movement in the sediment limit (see main text for explanation). The effective movement of the sediment limit is preserved in the stratigraphic transect shown at the base of the diagram by the recorded presence or absence of coarse sandy gyttja/muddy sand which in turn corresponds to periods of low lake level or high lake level, respectively.

reworking of marginal sediments, and a hiatus of sedimentation in that area. This will be reflected in the sediment record by a coarsening of grain size lakeward. Any subsequent rise in lake level will reset the sediment limit to a higher position, with renewed sedimentation burying the coarse layers deposited during the preceding lowstand.

Changes in lake level will also have an effect on the spatial distribution of littoral macrophytes. Macrophyte distribution in lakes is primarily a function of light intensity and water depth. Under natural conditions, the littoral margins of a lake would typically be dominated by emergent macrophytes such as *Carex* and *Phragmites*. As one begins to move away from the littoral zone with the increasing water depth, the submerged/floating macrophytes such as *Potamogeton* and *Myriophyllum* increase in importance; they themselves are then replaced in the deeper sublittoral zone by low-light taxa (*Chara* and *Nitella*), which tend to dominate here (Cohen, 2003). Within the sediment record, an increase in lake level should therefore be reflected by an abrupt increase in aquatic taxa as the littoral zone expands, which will be countered by a decline in terrestrial macrofossils. As lake levels continue to rise, the more marginal taxa will be steadily replaced by deep-water taxa, and vice versa.

Macrofossil analysis was already a well-established technique for lake-level reconstruction prior to Digerfeldt's work but was normally applied to a single-cored sequence. This was most often taken from the center of the lake, away from the littoral zone where the vast majority of macrophytes lived. Determining whether an observed shift in the macrofossil record is climate driven or a response to catchment-based processes is very difficult from a single-cored sequence, due in part to a lack of repeatability and information on the nature of the sedimentary record (Hannon and Gaillard, 1997). Macrophyte distribution can be affected by a host of other factors including changing water chemistry or nutrient supply, turbidity, and competition. Significant changes in down-core macrophytes can occur as a result of hydrosere activity around a lake margin. Differentiating *in situ* macrofossils from inwashed material is also a problem, particularly in the shallows of high-energy lakes. By replicating macrofossil counts across multiple littoral cores, an accurate interpretation of the macrofossil record can be made.

The required data are collected from a transect of cores running from the margins toward the lake center (Figure 8). The distance between each core will vary from site to site but as a general rule the depth increase between adjoining cores

should not exceed 1 m. Core correlation is achieved through a combination of visual stratigraphy and lithological characteristics (particle size, mineral magnetic analysis, loss on ignition). The stratigraphic nature of the technique means that it can provide a quantitative picture of lake-level change, something that had never before been achieved. The correlated sequence is then dated using a palynostratigraphic framework, based on local pollen zones, supported by a sequence of targeted AMS ^{14}C dates.

A major advantage of the Digerfeldt approach is that it is broadly compatible with a whole range of other proxy techniques, particularly diatoms and chironomids, which are discussed next.

Diatoms

The use of diatoms for aquatic environmental reconstruction is well established. They are extremely sensitive to changes in certain environmental characteristics such as pH and salinity. They have lifestyles that are invariably planktonic (free floating), benthonic (pertaining to the water body bed), epiphytic (attached to plants), or perhaps tychoplanktonic (carried into the plankton from the benthos during periods of increased turbulence or water-column changes).

In terms of lake studies, the most obvious and clear diatom habitat changes would occur with isostatic uplift in a coastal area which would effectively 'isolate' a sedimentary basin, thus altering the once brackish water to freshwater over time. This

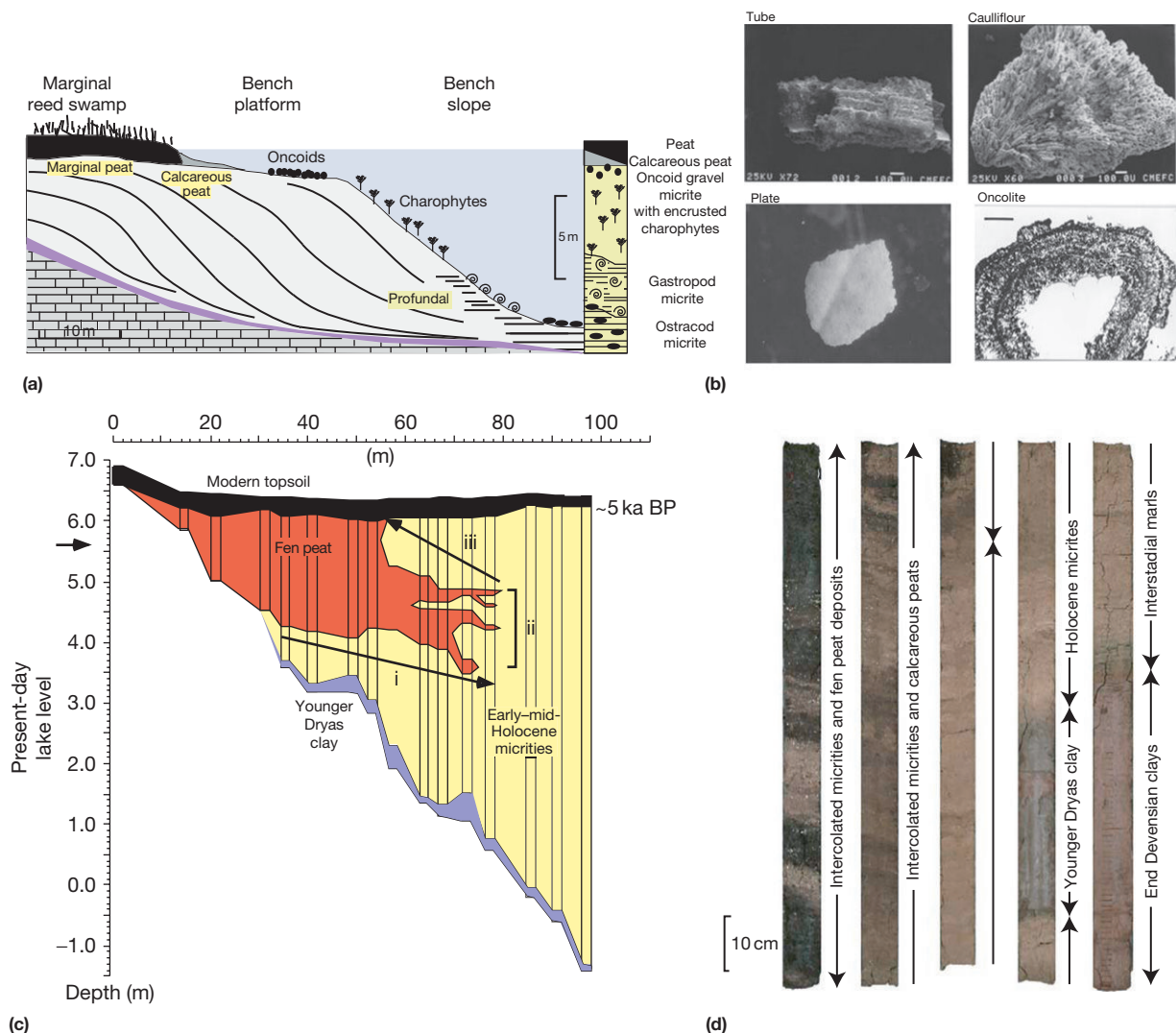


Figure 9 Carbonate lakes. (a) Theoretical model for a freshwater carbonate bench, highlighting zonal distribution of macrophytes (Cohen, 2004). Redrawn from [Murphy DH and Wilkinson BH \(1980\) *Sedimentology* 27: 123–135](#). (b) Morphological variation in carbonate precipitates sampled across the carbonate. Bar = 100 μm . Redrawn from [Magny M \(1998\) *Reconstruction of Holocene lake-level changes in the Jura \(France\): Methods and results*](#). In: Harrison SP, Frenzel B, Huckried U, and Weis M (eds.) *Palaeohydrology as Reflected in Lake-Level Changes as Climatic Evidence for Holocene Times*, *Paläoklimaforschung* 25, pp. 67–85. Stuttgart: Gustav Fischer Verlag. (c) Marginal carbonate bench stratigraphy for Hawes Water, northwest England. Spatial variation in the peat/marl contact reveals a detailed history of lake-level change at the site. An early phase of hydrosere encroachment (i) is followed by a period of rapid fluctuation (ii) and then a final phase of increasing levels (iii). The upper portion of the record has been lost through marginal peat cutting. (d) Down-core stratigraphy through carbonate bench spanning Late Glacial to mid-Holocene (Urswick Tarn, NW England).

salinity gradient is essentially 'missing' from freshwater lake systems (exceptions being closed basin lakes that encourage the concentration of mineral salts and therefore can exhibit saline and indeed hypersaline conditions), and instead the use of diatom life-form (ratios between deep-water benthos and shallow-water plankton) becomes a key to reconstructing former lake levels. A detailed review of the use of diatoms as indicators of water-level changes in freshwater lakes is given by Wolin and Duthie (1999), who identify a number of key lake sediment/diatom indicators that can and have been used, including life-form, turbulence, nutrient environment, and pH.

The majority of freshwater diatom studies make use of the changes in life-form, in particular the ratios between planktonic and nonplanktonic species. The association between open/deeper water and planktonic species and the links between low lake levels and the anticipated increase in macrophytes and therefore benthos and epiphytic species form the basis for diatom life-form studies. Bertzen (1987) and Lotter (1988) used life-form to reconstruct actual lake depths. The Digerfeldt (1986) method of lake-level reconstruction combined with diatom life-form is highlighted in Wolin (1996), where the signal from Lower Herring Lake, Michigan, showed that the transition period associated with both falling and rising lake levels was accompanied by an increase in benthic and epiphytic diatom species.

Chironomids and Cladocera

Chironomids and Cladocera are invertebrate remains that can be found preserved in most lake sediments and have both been used successfully as proxy methods for reconstructing paleoenvironments. Nonbiting midges (Insecta: Diptera: Chironomidae) have been used to study the effects of eutrophication and acidification and most significantly for paleotemperature reconstruction. Indeed, so promising is the use of chironomids for paleotemperature studies that they are likely to replace Coleoptera as the premier method for temperature reconstruction in temperate and arctic environments. Until recently, Cladocera have mainly been used for trophic and qualitative temperature reconstructions. A characteristic feature of both groups is that they occur in two distinct communities, shallow littoral and pelagic. Both groups have been used to reconstruct past lake levels but only in a qualitative manner. Korhola et al. (2000, 2005) have utilized these depth-species relationships to produce the first combined Cladocera/chironomid transfer function for quantitative lake-depth reconstruction. The transfer function is based on surface-sediment analysis from 53 lakes distributed across the northern Fennoscandian timberline. The transfer function appears to work well but has yet to be applied to other lakes.

Freshwater carbonate lake sediments

Carbonate sediments from small, hard-water lakes represent one of the most sensitive archives for midlatitude paleohydrological research. Carbonates are typically precipitated in the upper water column during the summer months, under the influence of the photosynthesis of both macro- and microphytes. Precipitation rates tend to be fastest in the littoral zone, where rates of biological productivity are highest. Over time, continued littoral deposition leads to the formation of a carbonate bench, which progrades lakeward. The autochthonous marginal nature of the carbonate bench deposits makes them exceptionally

sensitive to changes in lake level. The landward edge of a bench commonly supports a marginal fen peat complex, which typically encroaches lakeward in step with the prograding bench (Figure 9). Stratigraphically, the basal contact between the marginal peats and the carbonate sediments is very important as it defines the lake margin. Any subsequent change in lake level will be reflected by a lakeward/basinward displacement of this contact (Figure 9).

Following Digerfeldt's original method, spatial variation in the peat-carbonate contact can be traced using a series of cored transects. One of the advantages of working with carbonate lakes is that in the majority of cases the carbonate bench has become terrestrialized, making the process easier to execute. The correlated stratigraphic sequence can then be dated using a combination of pollen and ^{14}C AMS dates. Depending on the purity of the carbonate sediments, uranium-series dating may also be an option.

In a novel twist to the use of macrofossils, Magny (1998) has proposed a method that utilizes the morphology of calcite grains, the morphological characteristics of which mimic the macrophyte they were precipitated onto. If one can identify the range of calcite morphotypes associated with each macrophyte, it should be possible to use down-core variation in the type of calcite formations to reconstruct lake levels in the same way one would if one used aquatic macrofossils (Figure 9). Obviously, all the problems associated with macrofossil analysis are equally applicable to this technique. A site-specific survey of the relationship between macrophyte distribution and associated calcite morphotypes may help to alleviate at least some of the associated problems.

Most hard-water lakes are silica deficient due to the karstic nature of the catchment and the alkaline water chemistry. Consequently, diatom analysis is not always an option as the frustules are often simply redissolved into the water column at the end of the life cycle. Both chironomids and Cladocera are normally present in carbonate lake sediments, although the lake-level training set discussed above is yet to be applied to carbonate lakes.

Saline lakes

Over multimillennial timescales, down-core variation in the sediment record for the majority of saline lakes will be a reflection of long-term hydrological expansion and contraction in response to changing climatic conditions. During dry periods, when evaporation exceeds inflow, evaporitic concentration of the lake water occurs, increasing salinity, leading to the precipitation of evaporite minerals in the water column and on the lake floor. If the arid climatic conditions persist, desiccation will lead to the formation of a deflation basin and a break in the continuity of the sediment record. With the onset of a pluvial (wet) phase, the volume of the lake will increase with increasing inflow. Salinity will decrease and clastic sedimentation will increase in significance and previously deposited evaporites may be redissolved back into the water column. The exact nature of the sedimentary record is very site specific, and it is difficult to try and predict what one may find without a detailed knowledge of the hydrology and geochemistry of a site. A good example of the way in which climate can affect sedimentary processes in a saline lake is given by the depositional model for Lake Bogoria, Kenya, produced by Renaut and Tiercelin (1994) (Figure 10). Over shorter timescales, sedimentation will be influenced by a whole host of factors. Local differences in hydrology, topography, and

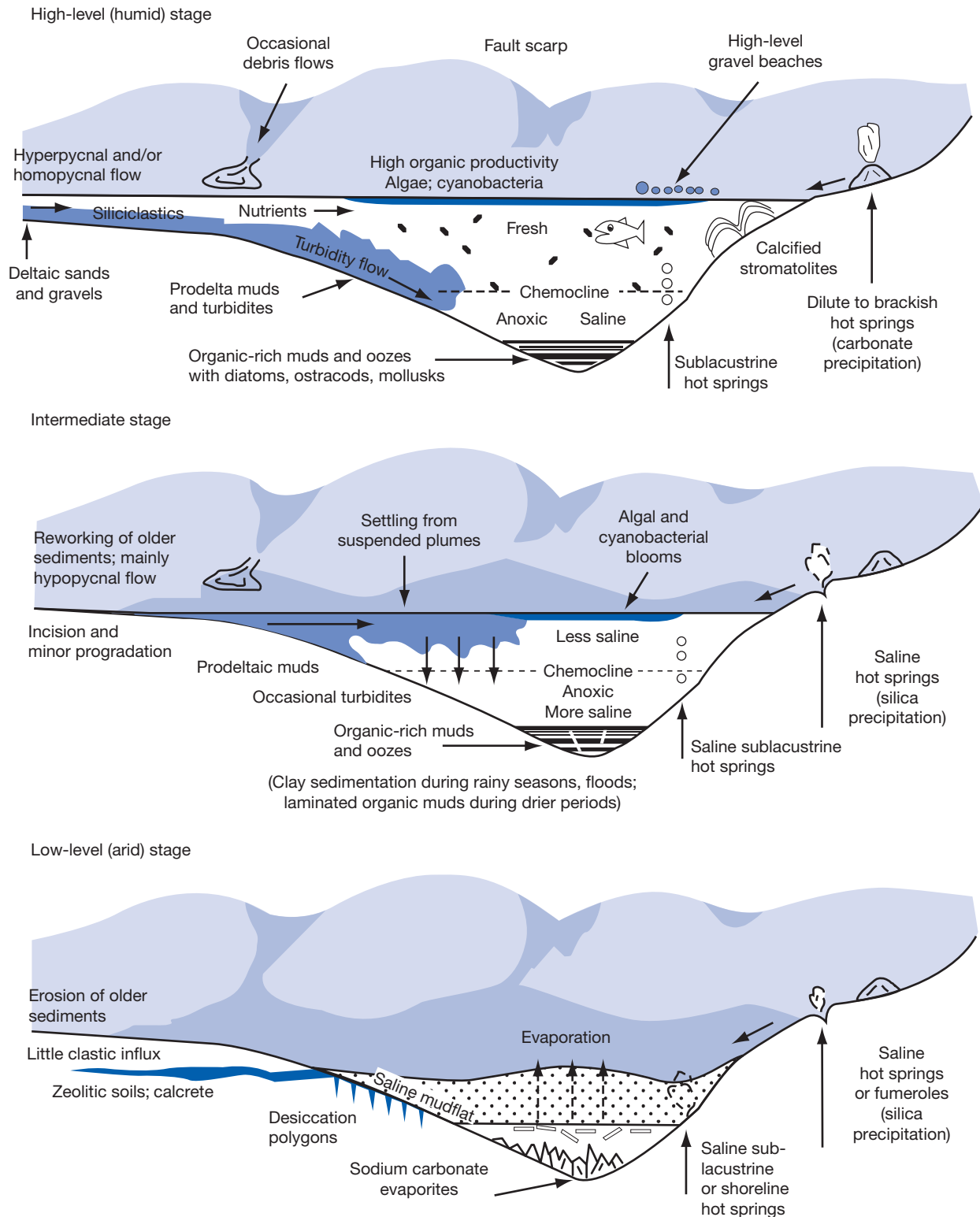


Figure 10 Theoretical model of sedimentation in Lake Bogoria, Kenya, highlighting the influence of climate change on sedimentary processes in saline-lake systems. Reproduced from Renault RW and Tiercelin J-J (1994) Lake Bogoria, Kenya Rift Valley: A sedimentological overview. In: Renault RW and Last WM (eds.) *Sedimentology and Geochemistry of Modern and Ancient Saline Lakes*, SEPM Special Publication No. 50, pp. 101–124, with permission from SEPM (Society for Sedimentary Geology).

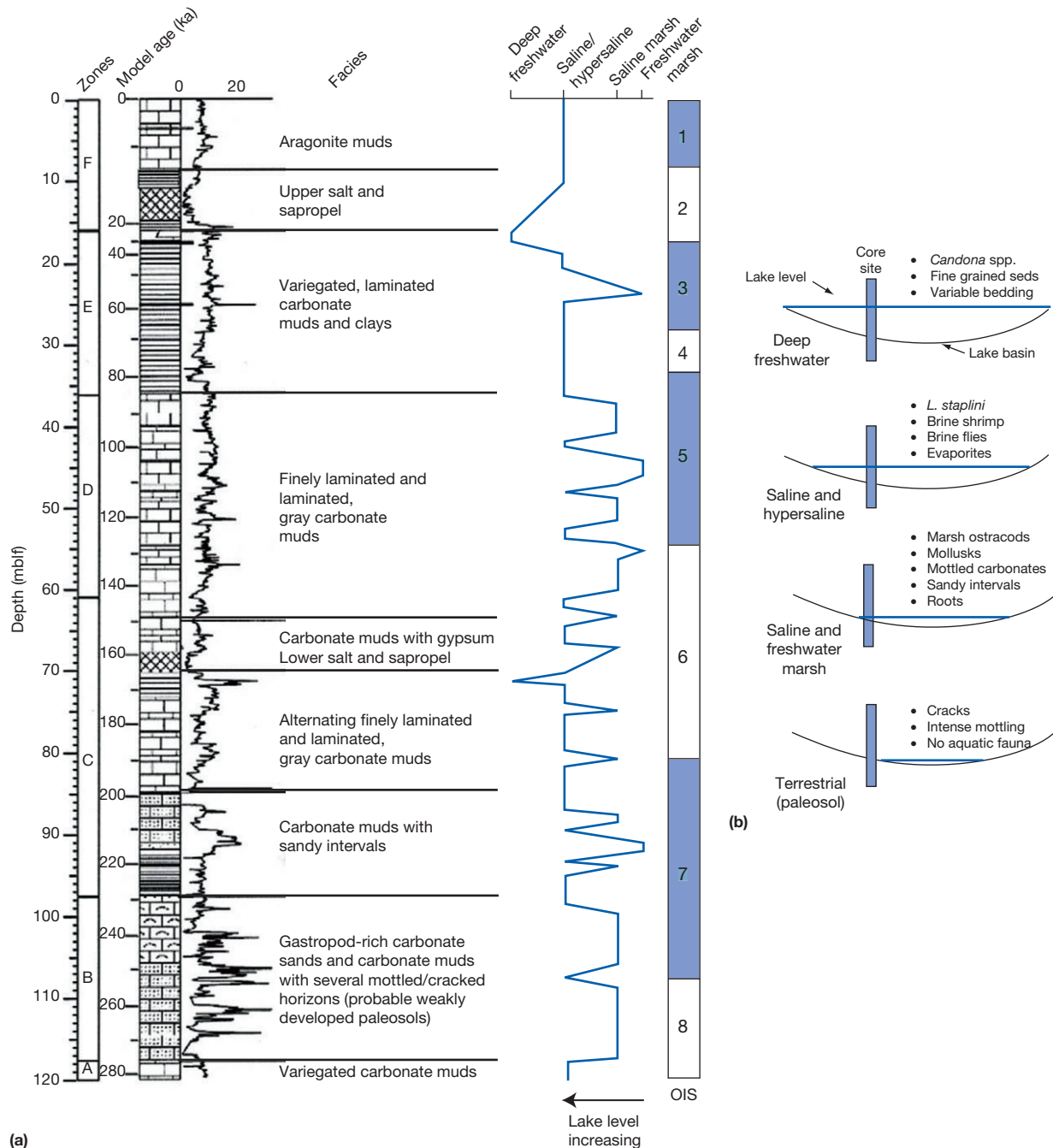


Figure 11 Composite diagram from Great Salt Lake. The record was reconstructed using a wide range of reconstructive techniques. Redrawn from Balch DP, Cohen AS, Schnurrenberger DW, et al. (2005) Ecosystem and paleohydrological response to Quaternary climate change in the Bonneville Basin, Utah. *Palaeogeography, Palaeoclimatology, Palaeoecology* 221(1–2): 99–122.

limnological processes can cause significant spatial variation in water chemistry across a site, which will clearly have implications for sedimentation patterns across a basin and must be considered when interpreting the long-core record.

Core Recovery and Sediment Classification

The harsh environment conditions and the remote location of many saline lakes make fieldwork a difficult and expensive

proposition. Specialist drilling equipment is needed due to the depth of preserved sediment: >100 m is commonplace. The long duration of sedimentation has obvious implications for the integrity of the record. The older the sediment, the more likely it would have been altered by postdepositional processes such as diagenesis. Sampling resolution is often a serious issue in terms of defining the temporal resolution of the reconstructed record and also logistically with respect to sample numbers. Analyzing 100 m of sediment at a 1-cm resolution is clearly

impractical. Too coarse a resolution, however, and one could miss an entire interstadial!

Just as with freshwater lakes, a variety of reconstructive techniques are employed to characterize the sediment sequence in terms of 'lithofacies' and 'biofacies.' Lithofacies are normally defined on the basis of multiple lines of evidence, including TOC, TIC, X-ray diffraction, particle size, and mineral magnetics. Whereas in the past these analyses would have been carried out in the laboratory by an overworked post-doctoral student, the development of automated systems for data acquisition has made the whole process much less painful. Multisensor core loggers can now simultaneously run a whole suite of measurements (e.g., gamma-ray attenuation, magnetic susceptibility, and compressional wave velocity), while the cores are still intact. The process is rapid (up to 12 m h⁻¹) and more importantly nondestructive. Once the various lithofacies have been identified, a tentative environmental interpretation is then drawn up with reference to modern analogs and existing sediment classification schemes (e.g., Schnurrenberger et al., 2003). The down-core biofacies are then determined and a comparison is made with the lithostratigraphy. From this comparison, a composite stratigraphy is drawn up, which, once dated, will form the basis of the environmental interpretation. A recent study by Balch et al. (2005) on a 121-m core from Great Salt Lake, Utah, provides an excellent case study of how such studies should be undertaken (Figure 11).

Prior to the mid-1980s, ¹⁴C was the only viable dating option for saline-lake-sediment studies. Owing to a dearth of organic matter, chronologies were typically based on ¹⁴C-dated authigenic/biogenic carbonates. This approach was often problematic as contamination by 'old' carbon was difficult to prevent, further compounded by the increased mobility of CaCO₃ in arid conditions. As coring technology improved and deeper cores were recovered, the limited age range of ¹⁴C (45 ka BP) also became problematic. A variety of alternative dating techniques now exist that are more suited to saline-lake sediments. Radiometric techniques include U-series dating of authigenic and biogenic carbonates (e.g., Lin et al., 1998) and ⁴⁰Ar/³⁹Ar dating of K-bearing minerals (e.g., Trauth et al., 2003). Optical dating is now widely used on both lake and catchment sediments (e.g., Rich et al., 2003), whereas paleomagnetic secular variation has been employed successfully to date a number of sequences (Lund, 1996). Depending on the volcanic history of a region, tephrochronology is particularly useful not only as a dating technique but also for correlation purposes (e.g., Whitlock et al., 2000).

The vast majority of proxy techniques used to reconstruct past lake-level histories from freshwater sequences are also applicable to saline lakes, although the way they are interpreted will often be different. Therefore, only a brief overview of the techniques most relevant to saline sediments is presented here.

Oxygen-Isotope Analysis

Oxygen-isotope analysis ($\delta^{18}\text{O}$) provides a unique source of hydrological and paleohydrological data for saline-lake studies. Analyses can now be routinely measured on a variety of components found within the sediment record, including authigenic (e.g., marls and micritite) and biogenic (Ostracods and bivalves) carbonates, biogenic silica, and sulfates.

The stable isotope composition of saline-lake carbonates ($\delta^{18}\text{O}_{\text{Carbonate}}$) is primarily a reflection of the isotopic composition of the lake water at the point of precipitation. The isotopic composition of the lake water will in turn be a reflection of several interconnecting factors including atmospheric temperature and the hydrological balance of the lake catchment. During periods of increased evaporation and/or decreased precipitation, $\delta^{18}\text{O}_{\text{lake}}$ will become enriched in ¹⁸O as the lighter ¹⁶O is preferentially lost to the atmosphere. With the transition to a pluvial phase, increasing lake volume will result in a decrease in $\delta^{18}\text{O}_{\text{lake}}$ as the input of isotopically lighter water sources (atmospheric, groundwater, and fluvial) exceeds the loss of lighter isotopes through evaporation (Leng et al., 2005).

Significant fluctuations in $\delta^{18}\text{O}$ have been attributed to the changing balance between precipitation and evaporation at a number of closed sites in Africa (Gasse, 2000), southwest United States (Benson, 2004), and elsewhere. $\delta^{18}\text{O}_{\text{lake}}$ can be influenced by a wide range of factors and when one is working in a system that is prone to rapid and pronounced changes in hydrology, a detailed understanding of the modern hydrology is crucial. If possible, an overview of the isotopic systematics for the whole catchment should be produced. The $\delta^{13}\text{C}_{\text{TDIC}}$ record (measured in tandem with $\delta^{18}\text{O}$) is particularly useful under such circumstances as closed basin lakes commonly show covariation between $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values for carbonates (e.g., Talbot, 1990). The complexities involved in disentangling the signal have to date precluded a quantitative interpretation of $\delta^{18}\text{O}$ data. Nevertheless, close parallels between long-core $\delta^{18}\text{O}$ records from saline-lake basins and global records of climate change such as ice-core records and SPECMAP illustrate the importance of this technique for paleohydrological research.

Ostrocods

Ostrocods are one of the most versatile fossils for paleoenvironmental research, particularly in saline systems. They are common to a wide variety of lakes and their robust nature means they are readily preserved. Their taxonomy and life history/ecology are well understood, and they have a long fossil record stretching back beyond the Quaternary. Analysis of down-core fossil assemblages provides a wealth of information on past limnology in the lake and changing climatic conditions (Griffiths and Holmes, 2000). Mourguiart and Carbonel (1994) have used quantitative changes in Ostracod assemblages to reconstruct paleolake levels in the Altiplano of Bolivia. $\delta^{18}\text{O}$ analysis of Ostracod shells for paleoclimate research is a well-established technique (Griffiths and Holmes, 2000). In most instances, it is also possible to examine the trace element composition of the valves; ratios of Mg/Ca and Sr/Ca have come under particular scrutiny as they are thought to be a proxy for salinity (Chivas et al., 1986).

Diatoms

In contrast to freshwater systems, the use of diatoms to reconstruct lake levels in saline lakes relies on the relative changes in conductivity from the concentration of salts due to fluctuations in moisture availability. Principally, deeper water is synonymous with reduced concentrations of mineral salts and therefore reduced conductivity.

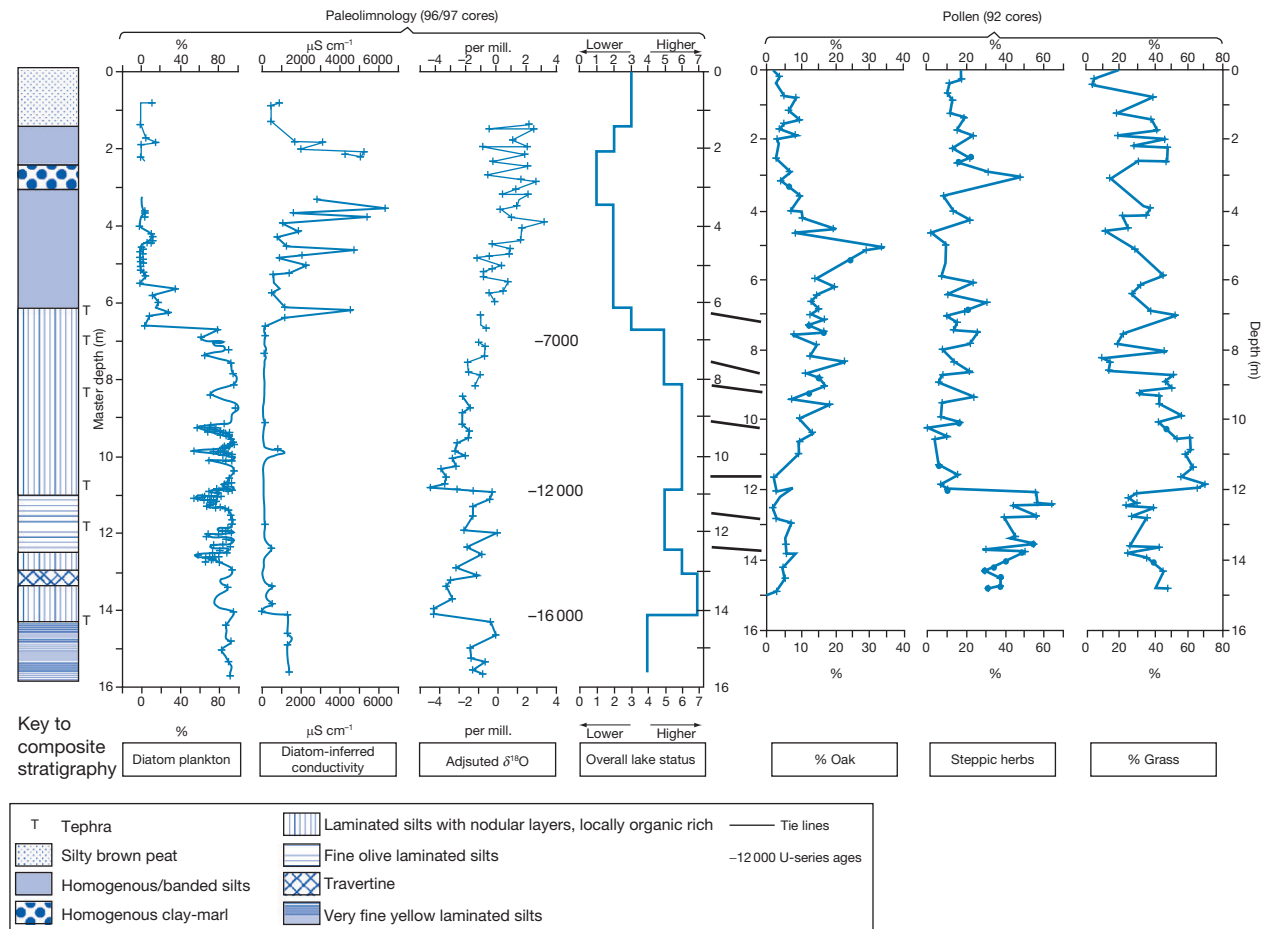


Figure 12 Composite diagram of late Pleistocene–Holocene environmental changes at Eski Acigöl. Past lake status was estimated using a combination of diatom and $\delta^{18}\text{O}$ analysis. Reproduced from Roberts N, et al. (2001) *The Holocene* 11(6): 721–736.

In a closed basin, falling lake levels will cause a concentration of salts and therefore an increase in conductivity. This will be reflected in the diatom assemblage. Fluctuations in closed basin lakes have been shown by Reed et al. (2001) for the Laguna de Medina, Cádiz, Spain, where a complete Holocene lake-level change record is compared with North Atlantic Bond cycles (Bond et al., 1997) and the tropical African lake-level record summarized by Gasse (2000). A study from the closed basin lakes in the Konya Basin, Turkey, by Reed et al. (1999) compared stable isotope data with diatom data. It was stated that the diatom data provided the corroborative evidence needed for the interpretation of changes found in the oxygen-isotope ratios.

As with the freshwater lake equivalents, a fall in lake level would also be shown as changes in diatom life-form, with lower levels favoring benthonic and epiphytic species. Roberts et al. (2001) used a multiproxy approach to reconstructing lake levels in central Turkey. In addition, using diatom-inferred conductivity, diatom life-form was also utilized. The study allowed seven stages of lake level to be inferred from 1 (shallow, mesopolysaline) to 7 (deep, stratified, dilute; Figure 12).

Conclusions

The investigation of lake levels over time can be carried out in a number of different ways using a growing number of methods.

Historically, geomorphology and sedimentology formed the prime methodological tools for the reconstruction and inference of paleolake levels. Since the work of Gunnar Digerfeldt in the 1980s, developments in the use of lithostratigraphy to incorporate biostratigraphy, mineral magnetics, water chemistry, isotopic data, and a variety of different dating methods have increased to further both the resolution and the wholesale understanding of lake sedimentation and ultimately lake-level movements. The incorporation of transfer functions, statistical modeling, and GIS/DTM technology has allowed the subject to advance well beyond that imagined by the early scientists who chose to map the extent of ancient lake shorelines. As more pioneering studies emerge using new techniques, and as further studies adopt these techniques and refine them, the multiproxy approach to studying paleolake levels will undoubtedly get more and more accurate. As with any study, the planning stages are important, as they should outline what techniques are best suited to the basin being investigated and to what extent the project budget will allow these investigations to be performed. Without doubt, the use of a multiproxy approach utilizing mapping, surveying, sedimentology, biostratigraphy, and chemical analysis is the ideal, which should all aim to provide accurate data for the reconstruction of former lake levels; it is these studies that will help to increase the accuracy and depth of our understanding.

See also: Diatom Introduction. **Chironomid Records:** Chironomid Overview. **Diatom Records:** Large Lakes. **Lake Level Studies:** Africa during the Late Quaternary; Asia; Latin America; Modeling; North America; West-Central Europe. **Paleolimnology:** Cladocera; Physical Properties of Lake Sediments. **Quaternary Stratigraphy:** Continental Biostratigraphy; Lithostratigraphy.

References

- Balch DP, Cohen AS, Schnurrenberger DW, et al. (2005) Ecosystem and paleohydrological response to Quaternary climate change in the Bonneville Basin, Utah. *Palaeogeography, Palaeoclimatology, Palaeoecology* 221(1–2): 99–122.
- Battarbee RW (1986) Diatom analysis. In: Berglund BE (ed.) *Handbook of Holocene Palaeohydrology*, pp. 527–570. Chichester: Wiley.
- Benson LV (2004) Western lakes. In: Rose J (ed.) *The Quaternary Period in the United States Development in Quaternary Science*, pp. 185–204.
- Bertzen G (1987) Diatomeenanalytische Untersuchungen an spätpleistozänen und holozänen Sedimenten des Tegeler Sees. *Berliner Geographische Abhandlungen* 45: 1–151.
- Bond G, Showers W, Cheseby M, et al. (1997) A pervasive millennial-scale cycle in North Atlantic Holocene and glacial climates. *Science* 278: 1257–1266.
- Chivas AR, DeDeckker P, and Shelly JMG (1986) Magnesium content of nonmarine Ostracod shells: A new paleosalinometer and paleothermometer. *Palaeogeography, Palaeoclimatology, Palaeoecology* 54(1–4): 43–61.
- Cohen AV (2003) *Palaeoclimatology: The History and Evolution of Lake Systems*, 528 pp. Oxford: Oxford University Press.
- Dearing JA (1997) Sedimentary indicators of lake-level changes in the humid temperate zone: A critical review. *Journal of Paleolimnology* 18(1): 1–14.
- Dearing JA and Foster IDL (1986) Lake sediments and paleohydrological studies. In: Berglund BE (ed.) *Handbook of Holocene Palaeohydrology*, pp. 67–90. New York: Wiley.
- DeVogel SB, Magee JW, Manley WF, and Miller GH (2004) A GIS-based reconstruction of late Quaternary paleohydrology: Lake Eyre, arid central Australia. *Palaeogeography, Palaeoclimatology, Palaeoecology* 204: 1–13.
- Digerfeldt G (1986) Studies on past lake-level fluctuations. In: Berglund BE (ed.) *Handbook of Holocene Palaeohydrology*, pp. 127–143. Chichester: Wiley.
- Digerfeldt G (1988) Reconstruction and regional correlation of Holocene lake-level fluctuations in Lake Bysjön, South Sweden. *Boreas* 17: 237–263.
- Elvin M, et al. (2002) *East Asian History* 23: 1–60.
- Gasse F (2000) Hydrological changes in the African tropics since the Last Glacial Maximum. *Quaternary Science Reviews* 19: 189–211.
- Griffiths HI and Holmes JA (2000) *Non-Marine Ostracods and Quaternary Palaeoenvironments*. London: Quaternary Research Association. QRA Technical guide No. 8.
- Grove AT and Warren A (1968) Quaternary landforms and climate on the south side of the Sahara. *Geographical Journal* 134: 194–208.
- Hannon GE and Gaillard MJ (1997) The plant-macrofossil record of past lake-level changes. *Journal of Paleolimnology* 18(1): 15–28.
- Harrison SP and Digerfeldt G (1993) European lakes as paleohydrological and paleoclimatic indicators. *Quaternary Science Reviews* 12(4): 233–248.
- Harrison SP and Dodson J (1993) Climates of Australia and New Guinea since 18,000 yr BP. In: Wright HE, et al. (ed.) *Global Climates Since the Last Glacial Maximum*, pp. 265–292. Minneapolis: University of Minnesota Press.
- Harrison SP, Saarse L, and Digerfeldt G (1991) Holocene changes in lake levels as climate proxy data in Europe. *Paläoklimaforschung* 6: 159–170.
- Kohfeld KE and Harrison SP (2000) How well can we simulate past climates? Evaluating the models using global palaeoenvironmental datasets. *Quaternary Science Reviews* 19(1–5): 321–346.
- Korhola A, Olander H, and Blom T (2000) Cladoceran and chironomid assemblages as qualitative indicators of water depth in subarctic Fennoscandian lakes. *Journal of Paleolimnology* 24(1): 43–54.
- Korhola A, Tikkanen M, and Weckström J (2005) Quantification of Holocene lake-level changes in Finnish Lapland using a cladocera – Lake depth transfer model. *Journal of Paleolimnology* 34(2): 175–190.
- Lamoueux SF and Gilbert R (2004) Physical and chemical properties and proxies of high latitude lake sediments. In: Pienitz R, Douglas MSV, and Smol JP (eds.) *Long-Term Environmental Change in Arctic and Antarctic Lakes*, vol. 8. Developments in Palaeoenvironmental Research.
- Leng MJ (2006) Isotopes in Palaeoenvironmental Research. *Developments in Palaeoenvironmental Research*, vol. 10. Dordrecht: Kluwer Academic Publishers.
- Lin JC, Broecker WS, Hemming SR, et al. (1998) A reassessment of U–Th and C-14 ages for late-glacial high-frequency hydrological events at Searles Lake, California. *Quaternary Research* 49(1): 11–23.
- Lotter A (1988) Past water level fluctuations at Lake Rotsee (Switzerland), evidenced by diatom analysis. *Proceedings of Nordic Diatomist Meeting*. Stockholm, 1987, pp. 47–55. Stockholm: University of Stockholm. Department of Quaternary Research (USDQR) Report 12.
- Lund SP (1996) A comparison of Holocene paleomagnetic secular variation records from North America. *Journal of Geophysical Research – Solid Earth* 101(B4): 8007–8024.
- Magny M (1998) Reconstruction of Holocene lake-level changes in the Jura (France): Methods and results. In: Harrison SP, Frenzel B, Huckried U, and Weis M (eds.) *Palaeohydrology as Reflected in Lake-Level Changes as Climatic Evidence for Holocene Times*, *Paläoklimaforschung* 25, pp. 67–85. Stuttgart: Gustav Fischer Verlag.
- Magny M (2004) Holocene climate variability as reflected by mid-European lake-level fluctuations and its probable impact on prehistoric human settlements. *Quaternary International* 113: 65–79.
- Mourguiart P and Carbonel P (1994) A quantitative method of palaeolake-level reconstruction using ostracod assemblages: An example from the Bolivian Altiplano. *Hydrobiologia* 288: 183–193.
- Murphy DH and Wilkinson BH (1980) *Sedimentology* 27: 123–135.
- Reed JM, Roberts N, and Leng MJ (1999) An evaluation of the diatom response to late Quaternary environmental change in tow lakes in the Konya Basin, Turkey, by comparison with stable isotope data. *Quaternary Science Reviews* 18: 631–646.
- Reed JM, Stevenson AC, and Juggins S (2001) A multi-proxy record of Holocene climatic change in southwestern Spain: The Laguna de Medina, Cádiz. *The Holocene* 11(6): 707–709.
- Renaut RW and Tiercelin J-J (1994) Lake Bogoria, Kenya Rift Valley: A sedimentological overview. In: Renaut RW and Last WM (eds.) *Sedimentology and Geochemistry of Modern and Ancient Saline Lakes*, pp. 101–124. SEPM Special Publication No. 50.
- Rich J, Stokes S, Wood W, et al. (2003) Optical dating of tufa via in situ aeolian sand grains: A case example from the Southern High Plains, USA. *Quaternary Science Reviews* 22(10–13): 1145–1152.
- Roberts N, et al. (2001) *The Holocene* 11(6): 721–736.
- Sack D (2001) Shoreline and basin configuration techniques in palaeolimnology. In: Last WM and Smol JP (eds.) *Tracking Environmental Change Using Lake Sediments, Vol. 1: Basin Analysis, Coring, and Chronological Techniques*, pp. 49–71. Dordrecht: Kluwer Academic Publishers.
- Schnurrenberger D, Russell J, and Kelts K (2003) Classification of lacustrine sediments based on sedimentary components. *Journal of Paleolimnology* 29: 141–154.
- Smith GI and Street-Perrott FA (1983) Pluvial lakes of the western United States. In: Porter SC (ed.) *Late Quaternary Environments of the United States*, pp. 190–212. Minneapolis: University of Minnesota Press.
- Steffen W, et al. (2003) *Global Change and the Earth System: A Planet Under Pressure*. The IGBP Series.
- Street FA and Grove AT (1976) Environmental and climatic implications of Late Quaternary lake-level fluctuations in Africa. *Nature* 261(5559): 385–390.
- Street FA and Grove AT (1979) Global maps of lake-level fluctuations since 30,000 yr BP. *Quaternary Research* 12(1): 83–118.
- Street-Perrott FA and Harrison SP (1985) Lake levels and climate reconstruction. In: Hect AD (ed.) *Paleoclimate Analysis and Modeling*, pp. 291–340. New York: Wiley.
- Talbot MR (1990) A review of the paleohydrological interpretation of carbon and oxygen isotopic-ratios in primary lacustrine carbonates. *Chemical Geology* 80(4): 261–279.
- Trauth MH, Deino AL, Bergner AGN, et al. (2003) East African climate change and orbital forcing during the last 175 kyr BP. *Earth and Planetary Science Letters* 206(3–4): 297–313.
- Whitlock C, Sarna-Wojcicki AM, Bartlein PJ, et al. (2000) Environmental history and tephrostratigraphy at Carp Lake, southwestern Columbia Basin, Washington, USA. *Palaeogeography, Palaeoclimatology, Palaeoecology* 155(1–2): 7–29.
- Wolin JA (1996) Late Holocene lake-level fluctuations in Lower Herring Lake, Michigan, USA. *Journal of Paleolimnology* 15: 19–45.
- Wolin JA and Duthie HC (1999) Diatoms as indicators of water level change in freshwater lakes. In: Stroemer EF and Smol JP (eds.) *The Diatoms: Applications for the Environmental and Earth Sciences*, pp. 183–202. Cambridge: Cambridge University Press.
- Yu G and Harrison SP (1995) Lake status records from Europe: Data base documentation. Boulder, CO: World Data Centre-A for Palaeoclimatology, NOAA Palaeoclimatology Program. Palaeoclimatology Publications Series Report No. 3.

Africa during the Late Quaternary

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Introduction

Africa is one of the most arid continents under today's climate regime, yet it also contains many of the world's largest lakes. It spans the equator, and much of its land lies within the tropics, its extremities just extending to the warm temperate zones of both hemispheres ([Figure 1](#)). That many African lakes have changed size dramatically has long been appreciated ([North America](#)), and studies on African lakes can contribute significantly to our understanding of Quaternary climate dynamics in the lower latitudes ([Gasse, 2000](#); [Street and Grove, 1979](#)). Lakes are focal points within landscapes and ecosystems and, particularly in arid/semiarid regions, are centers of productivity and biodiversity. Reconstructions of dramatic changes in lake size and conditions, coupled with an understanding of current species distributions and genetic relationships, have recently provided biologists with new insights into evolutionary processes such as speciation (e.g., [Sturmbauer et al., 2001, 2003](#)). Lakes are key resources for human societies – evidence of the long story of human evolution is embedded in the deposits associated with East African lakes. Reconstructions of effective moisture over the past ~2 My in East Africa suggest that humid phases (which interrupted a Quaternary drying trend) may have been linked to key events in hominid evolution ([Trauth et al., 2005](#)). Throughout the Holocene, changes in society appear to have been linked to moisture changes reflected by lake dynamics (e.g., [Hoelzmann et al., 2001](#); [Verschuren et al., 2000](#)). The sources of the river Nile are collected in two great lakes, Tana and Victoria, which lie near the equator, and over many millennia the people of the Nile valley have monitored and responded to changes in its flow. Today, in many regions of Africa, water is a critical resource, and human populations have suffered under the impact of extended droughts, such as the Sahelian drought of the 1970s, 1980s, and 1990s and the drought affecting the Horn of Africa in the first decade of the twenty-first century. Concerns about detrimental changes in African moisture regimes in relation to global warming are increasing. In the light of paleolake data, fluctuations experienced in recent decades are less extreme than those experienced over the period from the LGM through the Holocene, but on the other hand, previous centuries have seen megadroughts far more intense than any experienced in the past 100 years ([Russell and Johnson, 2007](#); [Wolff et al., 2011](#)).

The focus of this article is on the tropics and the Sahara, where the most dramatic lake level changes are recorded. Key findings of lake studies over millennial and century time-scales are briefly reviewed. It is by no means comprehensive, and the reader is directed to the review of [Gasse \(2000\)](#) for tropical African paleohydrology; the volumes edited by [Johnson and Odada \(1996\)](#) and [Lehman \(1998\)](#) on East African lakes and [Gasse \(2002\)](#) on the Sahara-Sahel region provide more information. Few records extend back more than

one glacial–interglacial cycle, and many are intermittent. Most information is known for the period from the LGM to present. Millennial-scale changes in African lakes reflect the rhythms of orbital cycles and the status of sea surface temperatures in the Atlantic and Indian Oceans ([Barker and Gasse, 2003](#); [Gasse, 2000](#); [Liu et al., 2004](#)). However, the balance of forcing has changed with time. At the LGM and during deglaciation, events at high latitudes and a widespread drop in SSTs appear to have had a strong influence on African moisture levels. In the Holocene, in the absence of ice sheets and major thermohaline circulation (THC) fluctuations, orbital forcing has been predominant. Superimposed on long-term changes during the Holocene are more rapid, high-amplitude fluctuations that suggest another faster-acting control over moisture levels. Long-term and short-term variations in African lakes are addressed in the next two sections.

African Lake Dynamics and Millennial-Scale Climate Change

Africa's climate zones range from tropical–equatorial to Mediterranean. The equator bisects the continent, and seasonal shifts in the inter-tropical convergence zone (ITCZ) enhance the rains twice yearly in equatorial tropical zones. Further from the equator, the rains occur once a year in a monsoonal pattern ([Figure 2](#)); monsoon strength depends in part upon summer insolation values ([Liu et al., 2004](#); [Data-Model Comparisons](#)). The Indian Ocean monsoon strongly affects much of East Africa, and the West African monsoon brings moisture from the Atlantic to the Sahel zone south of the Sahara desert. A weaker monsoonal flow brings precipitation in the austral summer to areas south of the equator. In Southern Africa, cold near-shore waters in the Atlantic create an arid climate in the West, but warm Indian Ocean waters bring moist conditions to the East. In tropical East Africa, upwelling and cool sea surface temperatures along the western edge of the Indian Ocean suppress evaporation, making the Horn of Africa relatively arid, and the orography of the East African mountains and Ethiopian highlands adds further complexity to rainfall patterns.

Time series of isotopic variations and proxies such as diatoms, physical and chemical sediment properties of single cores and offshore transects, and geomorphological approaches such as mapping and dating of paleoshorlines have been successfully applied to obtain Late Quaternary records of changes in African lakes. Many lakes are carbonate systems, and consequently radiocarbon dating can be challenging; thus, chronologies may be imprecise ([Gasse, 2000](#)).

A Green Sahara

In Michael Ondaatje's novel 'The English Patient,' the protagonists find themselves deep in the Sahara Desert exploring a

place known as the Cave of Swimmers. On the cave walls are pictographs representing humans swimming in a lake surrounded by a landscape teeming with game. Archaeological evidence of human occupation earlier in the Holocene, including pictographs and charcoal hearths, has been reported from across the Saharan region (Neumann, 1989). Ancient drainages and lake basins, eroding subaerial lake sediments, and exhumed bones of crocodiles provide evidence that large regions of the now hyperarid Sahara were once moister and the semidesert regions to the south supported productive savanna vegetation (e.g., de Smet, 1998; Gasse, 2002). The vast changes recorded in the Chad basin (Leblanc et al., 2006; North America; Overview) and evidence of large lakes in other now dry basins underscore that over the Holocene, moisture regimes in Saharan and sub-Saharan Africa have fluctuated dramatically, together with landscapes and biota.



Figure 1 Map of Africa showing places mentioned in the text.

The early Holocene greening of the Sahara is largely attributable to Milankovitch orbital variations (**Astronomical Theory of Paleoclimates**). Today, moisture reaches the southern fringes of the Sahara via the monsoonal circulation (**Figure 2** (July/August)). The onshore flow is driven in part by heating of the continental interior during the boreal summer. Between 12 000 and 6000 ^{14}C years BP, Northern Hemisphere summer insolation values were greater than present, the continent heated more strongly, and a stronger onshore flow developed. The rains penetrated far further north, filling Lake Chad and other lake systems and greening the landscape.

The dynamics of the North African monsoon were investigated in an early data model comparison (**Data–Model Comparisons; Paleophytogeography**), which tested hypotheses about the causes of past climate change by comparing a paleoclimate model simulation with the observed proxy data (**COHMAP, 1988; Street-Perrott and Perrott, 1993**). This study was one of the first to use a large lake level database as a proxy record of continental-scale paleoclimate. At the Last Glacial Maximum (18 000 ^{14}C years BP), lake levels were extremely low. At 9000 and 6000 ^{14}C years BP, lakes were high, compared with the present (**Figure 3**). In the early Holocene simulation, stronger heating generated greater and more extensive precipitation. In contrast, the LGM was characterized by weak summer insolation, and this, coupled with a generally cooler and drier global climate, led to conditions at least as arid as those of today in the region.

A newer study by Lézine et al. (2011) makes use of over 1500 dated Holocene paleohydrological records from the Sahara and Sahel zones and new model simulations. The signal of maximum rainfall lagged the early Holocene summer insolation peak, and this is likely due to the effects of residual northern ice sheets and meltwater pulses to the North Atlantic, which in turn caused the ITCZ to shift southward, reducing precipitation. Furthermore, the signal for maximum open-water extent lagged precipitation, and this is interpreted to reflect the time required for aquifer recharge, which varied spatially, depending upon topography and groundwater hydrology; the interpretation is supported by the record from Lake Bosumtwi (**Figure 1**), which lies on the southern edge of

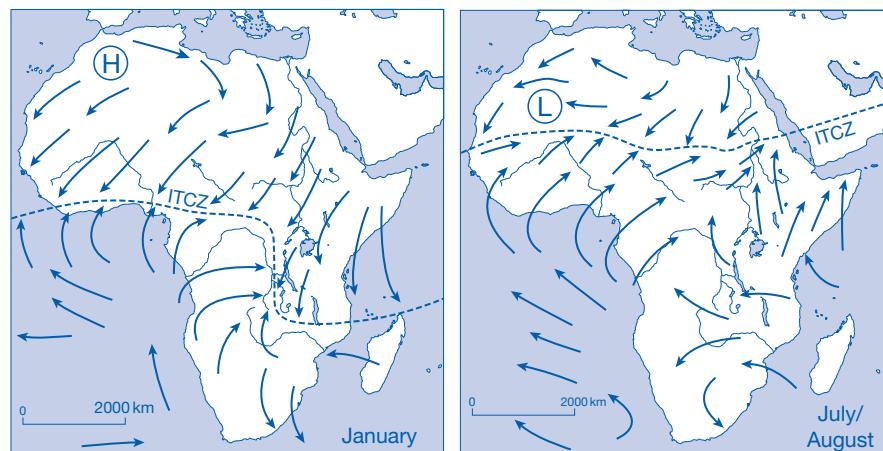


Figure 2 African monsoon circulation. Reproduced from Gasse F (2000) Hydrological changes in the African tropics since the Last Glacial Maximum. *Quaternary Science Reviews* 19: 189–211; Figure 3.

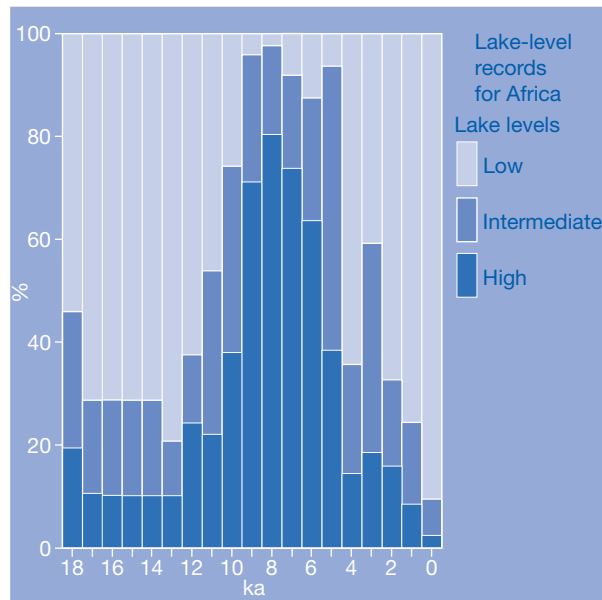


Figure 3 Graph shows the percent of lakes in one of three categories (low, intermediate, high) on a timescale of 18–0 ^{14}C ka. Modified after COHMAP members (1988) Climatic changes of the last 18 000 years: Observations and model simulations. *Science* 241: 1043–1052.

the Sahel zone in Ghana and has a lake level record that is thought to mirror local precipitation variation (Shanahan et al., 2006), and which leads the Saharan pattern.

The dramatic climate changes in the Saharan and Sahelian zones (Figure 1) during the Holocene are linked with prolonged human occupation of now arid regions. Hoelzmann et al. (2001) describe a sequence of cultures existing in association with a large paleolake basin in Western Nubia between 9400 and 3800 ^{14}C years BP and suggest that communication routes between the Nile valley and Lake Chad region existed during this time. Gasse (2002) also highlights the asynchrony of change and points out that the varied interactions of water and geology within the aquifers underlying these regions, plus differences in patterns of atmospheric deposition, led to variations in the chemistry of water bodies in both space and time. This in turn led to differing chemical composition of evaporate deposits associated with lakes. Certain deposits are preferred as a source of salt, and the ancient routes of salt caravans, still in use today, reflect this interaction of lake dynamics and geology (Gasse, 2002).

Late Quaternary Lake Dynamics in Eastern and Southern Africa

East Africa is a complex and dynamic region tectonically. The Great Rift Valley is the visible manifestation of a long-term process of continental rifting, whereby the easternmost portion of Africa is detaching from the rest of the continent. The region is characterized by earthquakes and volcanism, and the rift system is dotted with strings of lakes (Figure 1). The highest mountain complex in Africa borders the Rift Valley, adding complexity to the regional climate. The dynamics of East African lakes, particularly those within the rift, are thus affected by a complex set of climate factors, with tectonism a further

influence. Many of the Great Lakes are deep, subject to currents and turbidities, and have large sediment packages in their depocenters, meaning that drilling and core retrieval and interpretation are challenging.

Pre-Holocene records come from, among others, the Great Lakes Victoria, Tanganyika, and Malawi; Lakes Massoko and Rukwa, both south of the equator in Tanzania; and Lake Challa on the east slope of Mt. Kilimanjaro (Figure 1). Moernaut et al. (2010) use seismic-reflection data from a ~200-m sediment package from Lake Challa to interpret a lake level history over the last ~140 000 years (older ages based on an extrapolation of well-dated upper sediments). Low stands are linked to features identified as ponded basin sediments and high stands to units of draped lacustrine sediment. Greatest aridity, according to the age model, occurred during the period just prior to ~128 ka BP and between ~114 and ~97 ka BP; aridity was less extreme during the LGM. Several short periods of aridity identified in the record appear to match Heinrich events (Sub-Milankovitch (DO/Heinrich) Events), but otherwise much of the last glacial cycle was characterized by relatively stable and moist conditions. These data are supported by long records from Lake Malawi (Cohen et al., 2007; Scholz et al., 2011) that indicate severe aridity between 135 and ~70 ka BP; at these times, lake levels were far lower than during the LGM. After ~70 ka BP, sediment records from Lakes Malawi, Tanganyika, and Bosumtwi (equatorial West Africa, Figure 1) indicate that the period from ~70 ka BP until the LGM was relatively moist and stable across the continent, possibly due to a reduction of strong precessional influence in the orbital drivers of climate change (Scholz et al., 2011).

In contrast, at the LGM, as in the Sahara–Sahel region, records from many lakes in East Africa indicate low moisture levels (Barker and Gasse, 2003; Gasse, 2002). Lake Victoria was so low that at times it dried out, as demonstrated by seismic reflectors and subaerial soil horizons in cores (North America; Stager and Johnson, 2008; Stager et al., 2002; Talbot and Livingstone, 1989). In the Rift Valley, Lake Tanganyika was lower than today by at least 300 m (Gasse, 2002), and other rift lakes also record evidence of low stands (Figure 4; Felton et al., 2007; Gasse, 2000). Diatom data suggest that Lake Malawi was also at least as low as present during the LGM (Barker and Gasse, 2003). At Lake Rukwa, a 23 000-year record of terrestrial and aquatic vegetation indicates cool and dry conditions at the LGM (Vincens et al., 2005). At Lake Massoko (Barker et al., 2003), diatom-inferred lake levels are low at the LGM.

Relatively few lake records are available from further south in Africa, and records to date are somewhat equivocal (Gasse, 2000; Liu et al., 2004). A 200 000-year reconstruction of rainfall from the Pretoria Saltpan in Southern Africa (Partridge et al., 1997) records a cyclical pattern of rainfall that matches precessionally driven insolation variations that modulate the strength of the monsoon. However, rainfall is less well correlated with insolation over the past two cycles. In particular, the Pretoria Saltpan records dry conditions at the LGM, whereas orbital forcing would suggest an enhanced southern monsoonal effect. Dry conditions can be partly attributed to lower sea surface temperatures in tropical oceans, which led to significantly reduced evaporation and less moisture transport to the African continent thus reducing monsoonal effects (Gasse, 2000).

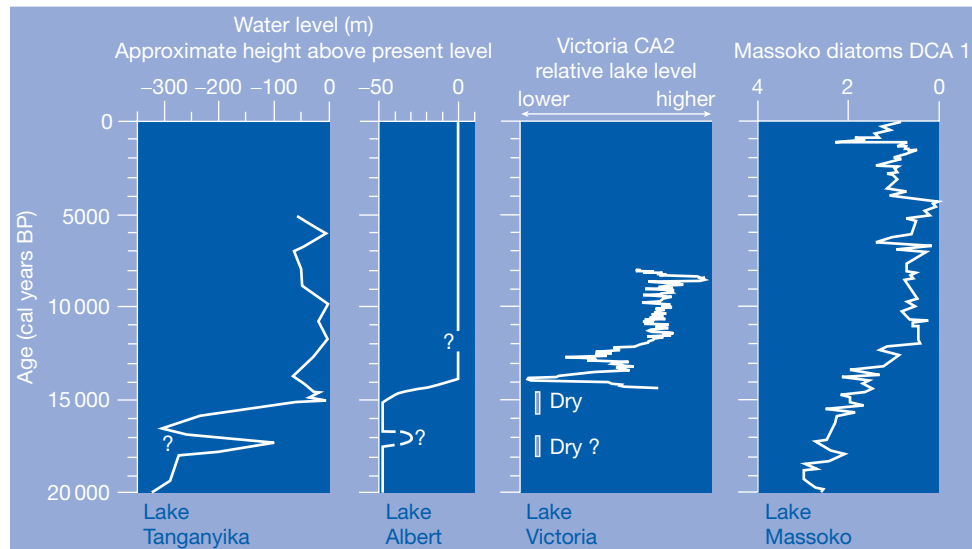


Figure 4 Late Quaternary records from the Great Lakes. Based on Barker P, Williamson D, Gasse F, and Gibert E (2003) Climatic and volcanic forcing revealed in a 50,000-year diatom record from Lake Massoko, Tanzania. *Quaternary Research* 60: 368–376; Gasse F (2000) Hydrological changes in the African tropics since the Last Glacial Maximum. *Quaternary Science Reviews* 19: 189–211; and Stager JC, Mayewski PA, and Meeker LD (2002) Cooling cycles, Heinrich event 1, and the desiccation of Lake Victoria. *Palaeogeography, Palaeoclimatology, Palaeoecology* 183: 169–178.

With the onset of deglaciation, lake levels generally rose (Figure 4), but rising trends were punctuated by regressions (see below). The detailed record from Lake Rukwa (Vincens et al., 2005) shows a deglacial warming, but indicators of a swamp environment at the coring locality suggest continuing low lake levels. The period ca. 12 000–5500 cal years BP saw low aquatic pollen totals, indicating a higher lake level. Subsequently, the return of swamp indicators shows lake level dropping and the return of more arid conditions. Relatively high Holocene levels are evident in lakes across East Africa as far as $\sim 7\text{--}8^\circ$ S, suggesting major trends in lake level largely reflect the patterns of ITCZ-driven precipitation in the equatorial and northern tropical regions.

Further south, evidence from Lake Malawi ($16\text{--}9^\circ$ S) from carbonate stratigraphy and diatoms in an offshore transect and from seismic studies (Finney and Johnson, 1991; Scholz and Finney, 1994; North America) suggests a moderate lowering of 100–150 m between 10 000 and 6000 years BP in an otherwise extended high stand extending from the Late Glacial. This is possibly the imprint of the Southern Hemisphere monsoon, which should have weakened in the Holocene (Liu et al., 2004; see below). The Holocene behavior of Malawi may reflect special subregional conditions related to its latitudinal span and its large and mountainous catchment. Further south, the Pretoria Saltpan records continuing low levels in the Holocene (Partridge et al., 1997).

Liu et al. (2004) modeled all the monsoon climates for the mid-Holocene and showed that insolation and ocean feedback are both important to the evolution of monsoon circulations. In particular, the mid-Holocene North African monsoon is enhanced by a reduction in upwelling and warmer sea surface temperatures north of the equator in the Atlantic. The Southern African monsoon weakens in response to lower austral summer insolation; the signal is reduced slightly in southernmost regions by warmer waters along the western coast of Southern

Africa. These results are broadly consistent with the lake level data, which indicate that the northern monsoon dominates even south of the equator. Further lake data from Southern Africa are needed to clarify the complex Late Quaternary forcing and response in this region.

High-Frequency (Centennial-Scale) Dynamics in African Lakes

Across the continent, millennial trends in lake level are punctuated by shorter events that are sometimes of considerable magnitude. With the onset of deglaciation, increases in moisture were not monotonic but pulsed, and arid intervals have punctuated the Holocene, even within generally wet periods such as the early and middle Holocene. Lake records from the late Holocene indicate that recent millennia have also seen severe episodes of drought, and these events can be related to the cultural record.

For the deglacial period, Gasse (2000) reports many lakes in equatorial Africa filling from as early as $>17\,000$ cal yr BP and certainly by 15 000 cal yr BP. Stager et al. (2002) dated desiccation events in Lake Victoria to $\sim 18\text{--}17\,000$ cal yr BP and some time between 15 900 and 14 200 years BP. These may be consistent with low stands in other lakes, although accurate dating of low stands is particularly problematic. A less severe low stand in Lake Victoria correlates with the Younger Dryas stage of the North Atlantic deglaciation ($\sim 12\,500\text{--}11\,500$ cal years BP). Arid events during the general trend to moister conditions in the Sahel and Sahara regions are also evident (Gasse, 2002; Street-Perrott and Perrott, 1990). Several authors have pointed out correlations between African lake behavior and events in, for example, the North Atlantic during deglaciation (Stager et al., 2002; Street-Perrott and Perrott, 1990). It appears that cooling of Atlantic SSTs, via glacial melt events and THC slowdown, for

example, affected large parts of the African continent, overriding regional controls over precipitation and evaporation, including the precessionally driven monsoon (see above). As in the full glacial, African moisture regimes remained sensitive to global-scale and ocean-driven events during the glacial–interglacial transition.

Such links can explain arid events into the early Holocene, but marked arid events are nevertheless recorded across large parts of the continent throughout Holocene. So far, this high-frequency variability over decades and centuries is not well understood, yet we need to understand it to plan for future climate change. There has been considerable discussion as to what the main forcing factors may be. For the Holocene, Lamb et al. (1995) recognize a synchronicity in the nature and timing of dry events in the warm temperate Atlas mountain region, which receives predominantly winter precipitation, with those in tropical Africa, where summer rains dominate, and they suggest the link may be cooler SSTs in the North Atlantic that could affect both monsoonal and westerly moisture levels.

Spectral analysis has been applied to high-resolution late Holocene records from several of the great lakes of East Africa. Over the past 4000 years, Lake Turkana (Kenya) laminated deposits show several submillennial periodicities, including 22, 18.6, and 11 years, which Halfman et al. (1994) have pointed out matches short-term solar variability (see [Volcanic and Solar Forcing](#)). A high-resolution record of calcite elemental composition and biogenic silica for the past 5400 years from Lake Edward, Uganda ([Russell and Johnson, 2005](#)), indicates century-scale droughts at ~4100, 2000, and 850 years BP and numerous lesser droughts. At Lake Edward, [Russell and Johnson \(2005\)](#) demonstrate marked fluctuations over past 1500 years, with three arid phases characterizing the majority of the period. The past two centuries have been unusually moist based on the whole period of record, but an extreme arid event is indicated between AD 1700 and 1800. [Bessemis et al. \(2008\)](#) report sedimentological evidence of a severe drought about 200 years ago from several shallow lakes, two

in Uganda and one in Kenya. A record from Naivasha ([Verschuren et al., 2000](#)) based on sedimentology, diatom assemblages, and fossil midge assemblages indicates drought for the period AD 1000–1270; wetter conditions followed, but these were punctuated by three more long dry events that appear more extreme than any twentieth century droughts (AD 1380 ± 1420 , 1560 ± 1620 , and 1760 ± 1840). Furthermore, records of precolonial history indicate alternating phases of famine, migration, and unrest on the one hand and periods of prosperity and stability on the other; the periods of unrest map almost exactly onto the drought periods in the Naivasha record ([Figure 5](#); [Verschuren et al., 2000](#)).

The coherence of the AD ~1750–1850 drought signal across Eastern Equatorial Africa suggests a major climate anomaly. However, while high stands occurred in the several centuries prior to the nineteenth century in Eastern Equatorial Africa, further south at Lake Malawi, levels were lower at this time ([Russell and Johnson, 2007](#)). This meridional difference does not seem explainable by an ITCZ shift, but it may reflect the position of the Congo Air Boundary, which marks a convergence zone between Atlantic and Indian Ocean air masses, and thus may represent a teleconnection to the ENSO ([Russell and Johnson, 2007](#)).

The role of ENSO in short-term variability is also invoked for the interannual scale in a study by [Wolff et al. \(2011\)](#). Currently, yearly precipitation in Eastern Equatorial Africa is partly influenced by ENSO, with rain/floods linked with El Niño and drought with La Niña years. At Lake Challa, dark-light varve couplets are formed from amorphous organic matter deposited in the moist season and diatom frustules that characterize the dry season, respectively; the latter control varve thickness. When varve thickness is compared with the past 150 years with ENSO index, thick varves occurred in La Niña years and thin ones in El Niño years. Thicker varves in LGM sediments relate to the generally drier climate then (see [Moernaut et al., 2010](#)), but interannual variability in varve thickness over the past 3000 years is greater than that in LGM; some varves

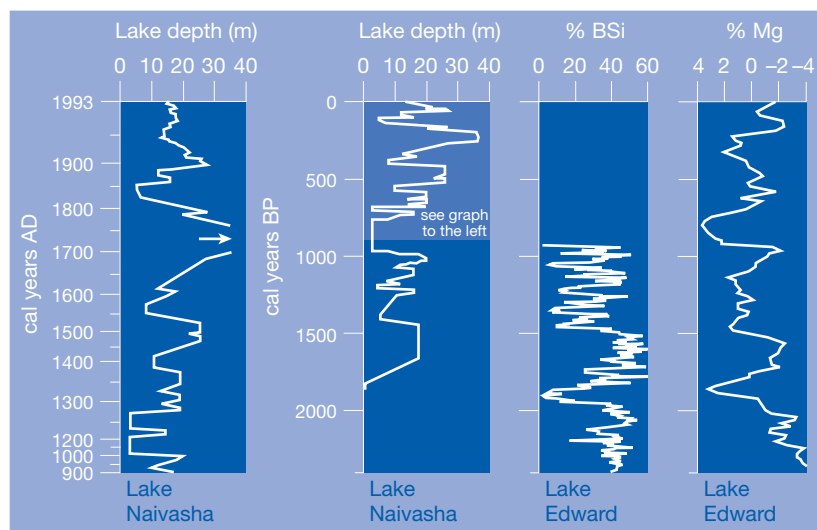


Figure 5 High-resolution records from East Africa. Based on Russell JM and Johnson TC (2005) A high-resolution geochemical record from Lake Edward, Uganda Congo and the timing and causes of tropical African drought during the late Holocene. *Quaternary Science Reviews* 24: 1375–1389; Verschuren D, Laird KR, and Cumming BF (2000) Rainfall and drought in equatorial east Africa during the past 1100 years. *Nature* 403: 410–414.

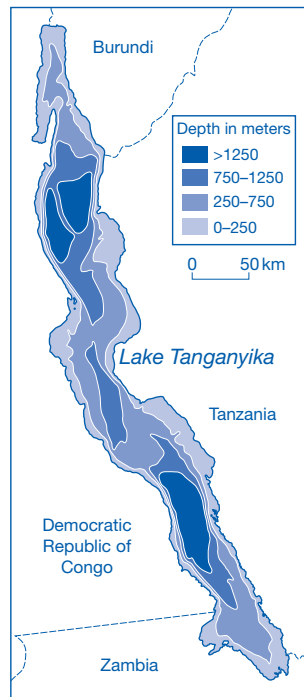


Figure 6 Bathymetric map of Lake Tanganyika.

exceed those of the LGM in thickness, suggesting years of extreme drought. If this association can be extrapolated to the future, warmer temperatures in Eastern Equatorial Africa may also be associated with higher interannual variability and hence extreme drought years.

Dating uncertainties still mean that the precise details of such short-term variability need further investigation. Even so, the general pattern shows droughts (that are likely regionally synchronous) superimposed onto a general late Holocene trend of increased aridity that reversed somewhat in the past two millennia. The conclusion is clear and somewhat daunting: century-scale droughts of great magnitude have occurred naturally under late Holocene conditions and we can assume they are likely to occur again, and intermittent extreme drought years are also likely. This is clearly of great concern, given the increasing pressures on water resources generally in East Africa (Russell and Johnson, 2005). Irrespective of the impact of future global warming, water management strategies should be flexible enough to cope with extended and/or droughts in the future.

Lake Level Change and Evolution: The Cichlid Fishes of African Lakes

Water level changes in the order of several hundred meters are recorded in great lakes such as Malawi and Tanganyika during the Late Quaternary, and Lake Victoria may have completely desiccated (see Section 'Late Quaternary Lake Dynamics in Eastern and Southern Africa'). Such large fluctuations in lake area and volume may have led to the temporary fragmentation of large lakes into several isolated basins and subsequent coalescence. Changes in the extent and continuity of habitats are

important factors driving evolution, as they offer potential environments for processes of speciation and radiation to occur; gene flow among subpopulations is prevented by intermittent barriers – in this case, changing lake levels.

The organisms in question are the highly diverse cichlid fishes. Molecular phylogenies have been developed for so-called species flocks in several of the great lakes (Verheyen et al., 1996; Sturmbauer et al., 2001, 2003). The mapping of genes across the populations shows evolutionary patterns developing in response to water level changes at two scales. For example, in Lake Tanganyika, three clades (ancestrally related groupings) map onto three sub-basins that during Quaternary low stands would have been isolated from each other (Figure 6; Verheyen et al., 1996). However, as the fish tend to inhabit shallow regions of the lakes, major increases in lake level also establish barriers to gene flow among populations by interposing stretches of deep water (Sturmbauer et al., 2003). Observed genetic patterns suggest that increases in lake level can lead to explosive radiations within isolated populations and that these often show parallelism in morphologies and adaptations. New genetic studies on populations of cichlid fish in shallow water areas of Lake Tanganyika show the dynamics of population shifts that underlie the speciation events. Reconstructed population histories largely match with known major lake level changes, with deeper splits occurring >100 ka BP and others during the LGM low stand ca. 20 ka BP (Koblmüller et al., 2011).

That Lake Victoria probably dried completely in the late Pleistocene suggests that the latest species flock became established and radiated extremely rapidly (Johnson et al., 1996; Stager and Johnson, 2008). With the advent of molecular phylogenetic techniques, the cichlid fish in African lakes promise to be a model system for investigating evolution in strongly fluctuating environments.

See also: [Lake Level Studies: Asia](#); [Modeling](#); [Overview](#).

References

- Barker P and Gasse F (2003) New evidence for a reduced water balance in East Africa during the last glacial maximum: Implication for model-data comparison. *Quaternary Science Reviews* 22: 823–837.
- Barker P, Williamson D, Gasse F, and Gibert E (2003) Climatic and volcanic forcing revealed in a 50,000-year diatom record from Lake Massoko, Tanzania. *Quaternary Research* 60: 368–376.
- Bessemers ID, Verschuren JM, Russell J, Hus FM, and Cumming BF (2008) Palaeolimnological evidence for widespread late 18th century drought across equatorial East Africa. *Palaeogeography, Palaeoclimatology, Palaeoecology* 259: 107–120.
- Cohen AS, Stone JR, Beuning KRM, et al. (2007) Ecological consequences of early Late Pleistocene megadroughts in tropical Africa. *PNAS* 104: 16422–16427. <http://dx.doi.org/10.1073/pnas.0703873104>.
- COHMAP members (1988) Climatic changes of the last 18,000 years: Observations and model simulations. *Science* 241: 1043–1052.
- de Smet K (1998) Status of the Nile crocodile in the Sahara desert. *Hydrobiologia* 391: 81–86.
- Felton AA, Russell JM, Cohen AS, et al. (2007) Paleolimnological evidence for the onset and termination of glacial aridity from Lake Tanganyika, Tropical East Africa. *Palaeogeography, Palaeoclimatology, Palaeoecology* 252: 405–423.
- Finney BP and Johnson TC (1991) Sedimentation in Lake Malawi (East Africa) during the past 10,000 years – A continuous paleoclimatic record from the Southern tropics. *Palaeogeography, Palaeoclimatology, Palaeoecology* 85: 351–366.

- Gasse F (2000) Hydrological changes in the African tropics since the last glacial maximum. *Quaternary Science Reviews* 19: 189–211.
- Gasse F (2002) Diatom-inferred salinity and carbonate oxygen isotopes in Holocene waterbodies of the Western Sahara and Sahel (Africa). *Quaternary Science Reviews* 21: 737–767.
- Halfman JD, Johnson TC, and Finney BP (1994) New AMS dates, stratigraphic correlations and decadal climatic cycles for the past 4-ka at Lake Turkana, Kenya. *Palaeogeography, Palaeoclimatology, Palaeoecology* 111: 83–98.
- Hoelzmann P, Kedig B, Berke H, Kropelin S, and Kruse H-J (2001) Environmental change and archaeology: Lake evolution and human occupation in the eastern Sahara during the Holocene. *Palaeogeography, Palaeoclimatology, Palaeoecology* 169: 193–217.
- Johnson TC and Odada EO (eds.) (1996) *The Limnology, Climatology and Paleoclimatology of the East African Lakes*. Amsterdam: Gordon and Breach.
- Johnson TC, Scholz CA, Talbot MR, et al. (1996) Late Pleistocene desiccation of Lake Victoria and rapid evolution of Cichlid fishes. *Science* 273: 1091–1093.
- Koblmüller S, Salzburger W, Obermüller B, Eigner E, Sturmbauer C, and Sefc KM (2011) Separated by sand, fused by dropping water: Habitat barriers and fluctuating water levels steer the evolution of rock-dwelling cichlid populations in Lake Tanganyika. *Molecular Ecology* 20: 2272–2290.
- Leblanc MJ, Leduc C, Stagnitti F, et al. (2006) Evidence for Megalake Chad, North-Central Africa, during the late Quaternary from satellite data. *Palaeogeography, Palaeoclimatology, Palaeoecology* 230: 230–242.
- Lehman JT (ed.) (1998) *Environmental change and response in East African Lakes*. Dordrecht: Kluwer.
- Lézine AM, Hely C, Grenier C, Braconnot P, and Krinner G (2011) Sahara and Sahel vulnerability to climate changes, lessons from Holocene hydrological data. *Quaternary Science Reviews* 30: 3001–3012.
- Liu Z, Harrison SP, Kutzbach J, and Otto-Bliesner B (2004) Global monsoons in the mid-Holocene and oceanic feedback. *Climate Dynamics* 22: 157–182.
- Moernaut J, Verschuren D, Charlet F, et al. (2010) The seismic–stratigraphic record of lake-level fluctuations in Lake Challa: Hydrological stability and change in equatorial East Africa over the last 140 kyr. *Earth and Planetary Science Letters* 290: 214–223.
- Neumann K (1989) Holocene vegetation of the Eastern Sahara: Charcoal from prehistoric sites. *African Archaeological Review* 7: 97–116.
- Partridge TC, Demenocal PB, Lorentz SA, Paiker MJ, and Vogel JC (1997) Orbital forcing of climate over South Africa: A 200,000-year rainfall record from the Pretoria Saltpan. *Quaternary Science Reviews* 16: 1125–1133.
- Russell JM and Johnson TC (2005) A high-resolution geochemical record from Lake Edward, Uganda Congo and the timing and causes of tropical African drought during the Late Holocene. *Quaternary Science Reviews* 24: 1375–1389.
- Russell JM and Johnson TC (2007) Little ice age drought in equatorial Africa: Intertropical convergence zone migrations and El Niño–Southern oscillation variability. *Geology* 35: 21–24.
- Scholz CA and Finney BP (1994) Late quaternary sequence stratigraphy of Lake Malawi (Nyasa), Africa. *Sedimentology* 41: 163–179.
- Scholz CA, Johnson TC, Cohen AS, et al. (2011) East African megadroughts between 135 and 75 thousand years ago and bearing on early-modern human origins. *PNAS* 104: 16416–16421.
- Shanahan TM, Overpeck JT, Wheeler CW, et al. (2006) Paleoclimatic variations in West Africa from a record of Late Pleistocene and Holocene lake level stands of Lake Bosumtwi, Ghana. *Palaeogeography, Palaeoclimatology, Palaeoecology* 242: 287–302. <http://dx.doi.org/10.1016/j.palaeo.2006.06.007>.
- Stager JC and Johnson TC (2008) The Late Pleistocene desiccation of Lake Victoria and the origin of its endemic biota. *Hydrobiologia* 596: 5–16.
- Stager JC, Mayewski PA, and Meeker LD (2002) Cooling cycles, Heinrich event 1, and the desiccation of Lake Victoria. *Palaeogeography, Palaeoclimatology, Palaeoecology* 183: 169–178.
- Street FA and Grove AT (1979) Global maps of lake-level fluctuations since 30,000 yr BP. *Quaternary Research* 12(1): 83–118.
- Street-Perrott FA and Perrott RA (1990) Abrupt climate fluctuations in the tropics: The influence of Atlantic ocean circulation. *Nature* 343: 607–611.
- Street-Perrott FA and Perrott RA (1993) Holocene vegetation, lake levels and climate in Africa. In: Wright HE Jr., Kutzbach JE, Webb T III, Ruddiman WF, Street-Perrott FA, and Bartlein PJ (eds.) *Global Climates Since the last glacial maximum*, pp. 318–356. Minneapolis: University of Minnesota Press.
- Sturmbauer C, Baric S, Salzburger W, Ruber L, and Verheyen E (2001) Lake level fluctuations synchronize genetic divergences of cichlid fishes in African lakes. *Molecular Biology And Evolution* 18: 144–154.
- Sturmbauer C, Hainz U, Baric S, Verheyen E, and Salzburger W (2003) Evolution of the tribe Tropheini from Lake Tanganyika: Synchronized explosive speciation producing multiple evolutionary parallelism. *Hydrobiologia* 500: 51–64.
- Talbot MR and Livingstone DA (1989) Hydrogen index and carbon isotopes of lacustrine organic matter as lake-level indicators. *Palaeogeography, Palaeoclimatology, Palaeoecology* 80: 283–300.
- Trauth MH, Maslin MA, Deino A, and Strecker MR (2005) Late Cenozoic Moisture history of East Africa. *Science* 309: 2051–2053.
- Verheyen E, Ruber L, Snoeks J, and Meyer A (1996) Mitochondrial phylogeography of rock-dwelling cichlid fishes reveals evolutionary influence of historical lake level fluctuations of Lake Tanganyika, Africa. *Philosophical Transactions of the Royal Society of London. Series B, Biological Sciences* 351: 797–805.
- Verschuren D, Laird KR, and Cumming BF (2000) Rainfall and drought in equatorial east Africa during the past 1100 years. *Nature* 403: 410–414.
- Vincens A, Buchet G, Williamson D, and Taieb M (2005) A 23,000 yr pollen record from Lake Rukwa (88S, SW Tanzania): New data on vegetation dynamics and climate in Central Eastern Africa. *Review of Palaeobotany and Palynology* 137: 147–162.
- Wolff C, Haug GH, Timmermann A, et al. (2011) Reduced interannual rainfall variability in East Africa during the last ice age. *Science* 333: 743–747.

Asian Lake-Level Records

Lake level reflects water-volume changes in a lake and thus is a hydrological measure that changes largely in response to local water budget (precipitation–evaporation (P–E)) (Street-Perrott and Harrison, 1985). In Asia, lake-level changes have been sensitive to climate changes, particularly to extreme droughts in arid and semiarid catchments (e.g., lakes in the Aral and Tarim Basins), to widespread catchment flooding in coastal plains (e.g., the mid and lower reaches of Changjiang/Yangtze River), and to the development of saline waters on highlands (e.g., lakes of the Tibetan Plateau; Hu, 2001; Pan and Chen, 2010; Qin, 2008). Tracing long-term lake-level change aids scientific prediction of catchment floods and droughts, rational utilization of lake water resources, and protection of lake environments in Asia (Shen et al., 2010; Yu and Shen, 2009).

As no instrumental records date from pre-Industrial time, all long-term P–E reconstructions rely on geological lake-level records, with various indicators derived from studies of geomorphology, sedimentology, geochemistry, paleobiology, and archaeology of lakes (Harrison and Digerfeldt, 1993). Reconstructions of lake-level changes involve different timescales. Under the setting of tectonic subsidence, structurally controlled lakes can persist from one to ten million years: for example, Qinghai Lake, which has continuous lacustrine deposits dating from the Oligocene (Zheng et al., 1989). Many other lakes provide Holocene or late-Quaternary histories over many thousands of years. On the other hand, more ephemeral lakes formed by geomorphologic processes, such as ox-bows, lagoons (e.g., Xihu/West Lake in Eastern China), and proglacial lakes (e.g., Tianchi/Heaven Lake of Altai Mountains; Wang and Dou, 1998) normally only last hundreds to a few thousand years.

Dramatic lake-level changes on a late-Quaternary timescale have been recorded in Asian lakes with extreme characteristics, such as the Aral Sea and Qarhan Salt Lakes (Hu, 2001), the hypersaline Dead Sea and Aiding Lake, two of the lowest elevation lakes, and Sumuxi Co and Num Co (Zhu et al., 2003), two of the highest elevation lakes. Records of lake-level changes and syntheses of lake status from Asia have become an important tool for reconstructing paleoclimate at a continental to global scale (e.g., An et al., 2011; Harrison et al., 1996). Assessments of lake-level changes using a multiple-category scheme (very high, high, ..., very low, etc.) have provided a flexible coding basis to evaluate precipitation and P–E simulations in paleoclimate modeling (e.g., Qin et al., 1998; Yu et al., 2007). Lake-level databases (Street-Perrott et al., 1989; Tarasov et al., 1996; Yu et al., 2001a,b) hold about 150 lake-level records for Asia, ~54% with a recording length of the Holocene and 26% extending to 20 ka BP (Figure 1).

Historical Review of Asian Lake-Level Studies

From Closed to Overflowing Lakes

Investigations of Quaternary Asian lakes began early in the twentieth century. Some of these studies documented changes in lake level in response to climate changes (e.g., Hedin, 1922). With investigations of minerals or water resources, more lake-level studies were achieved up to the 1980s (e.g., Nanjing Institute of Geography and Limnology (NIGL), 1989; Zheng et al., 1989). Due to the assumption that overflowing lakes respond to changes in regional water budgets solely through changes in discharge (Dearing and Foster, 1986), most studies during this period were focused on closed lakes, for example, studies of the Dead Sea (Kaufman, 1971) and Qaidam Basin (Chen and Bowler, 1986). The recognition that overflowing lakes also respond to climate changes (e.g., Mason et al., 1994) has led to more Asian lake data being used to reconstruct past-regional climate changes, for example, Hulun Lake and Xingkai/Khanka Lake in Northeast China (Qiu et al., 1988; Wang et al., 1995), and Erhai and Fuxian Lake in Yunnan Plateau (NIGL, 1989). These studies mean that many Asian areas are well represented, and they extend study areas from closed basins that are mostly in the arid–semiarid interior into open basins that are mostly in the Asian monsoonal lowlands where overflowing lakes are dominant.

Basis of Lake-Level Reconstruction

Like studies in Europe and North America (Harrison and Digerfeldt, 1993), Asian lake-level reconstructions for most of the basins are based on changes in the nature of the sediments and sedimentary structures, as revealed in sediment cores, which can qualitatively reconstruct lake-level changes (NIGL, 1989; Zheng et al., 1989). Evidence of geomorphology, archaeology and historical records has been used to reconstruct the absolute levels and areas of lakes (Chen and Bowler, 1986; Pachur et al., 1995; Qiu et al., 1988). In arid–semiarid areas of Asia, changes in water salinity can also be used as an indicator of changes in relative water depth and hence lake-level status. Qualitative estimates of salinity (fresh, brackish, and saline) can be made based on sediment mineralogy and geochemistry. As the evaporative concentration of the lake water increases, there is a shift in the mineralogy/geochemistry of the sediments. The general sequence is from carbonate, borate, mirabilite, and nitrate to halite with increasing water salinity (Zheng et al., 1989). Paleoecological evidence, specifically changes in the species assemblages of aquatic plants, diatoms, and ostracodes, has been used to provide corroborating evidence for the reconstructed changes in water depth.

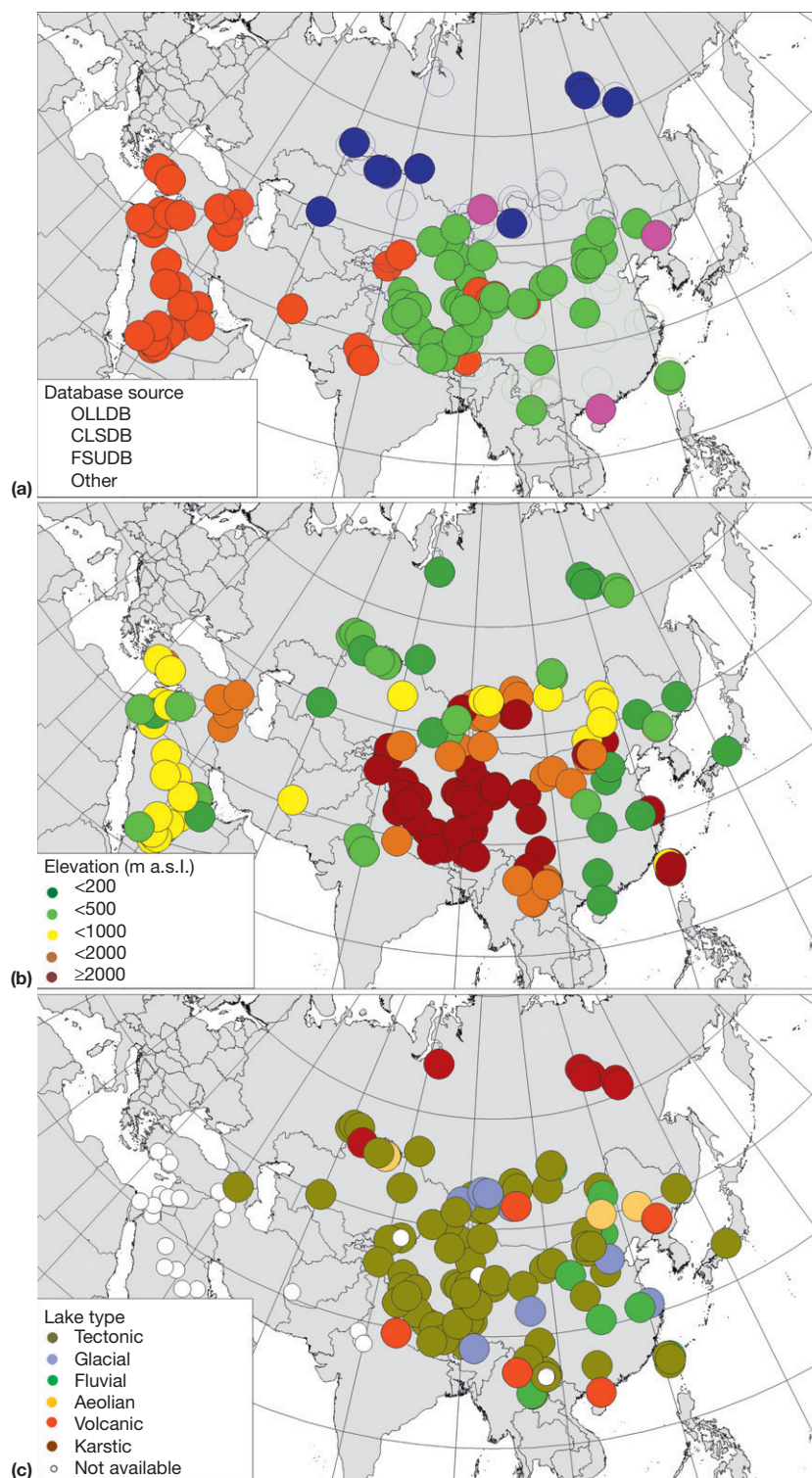


Figure 1 (a)–(f) Information on lake-level studies in Asia. (a) Database source: OLLDB (from [Street-Perrott et al., 1989](#)), CLSDB (from [Yu et al., 2001a,b](#)), FSUDB (from [Tarasov et al., 1996](#)), with lake hydrological type; (b) lake elevation in meters a.s.l.; (c) lake origin types; (d) and (e) lake record length (ka BP); (f) indication of numbers of ^{14}C dates in each lake study. Open circles = overflowing lakes; closed circles = closed lakes.

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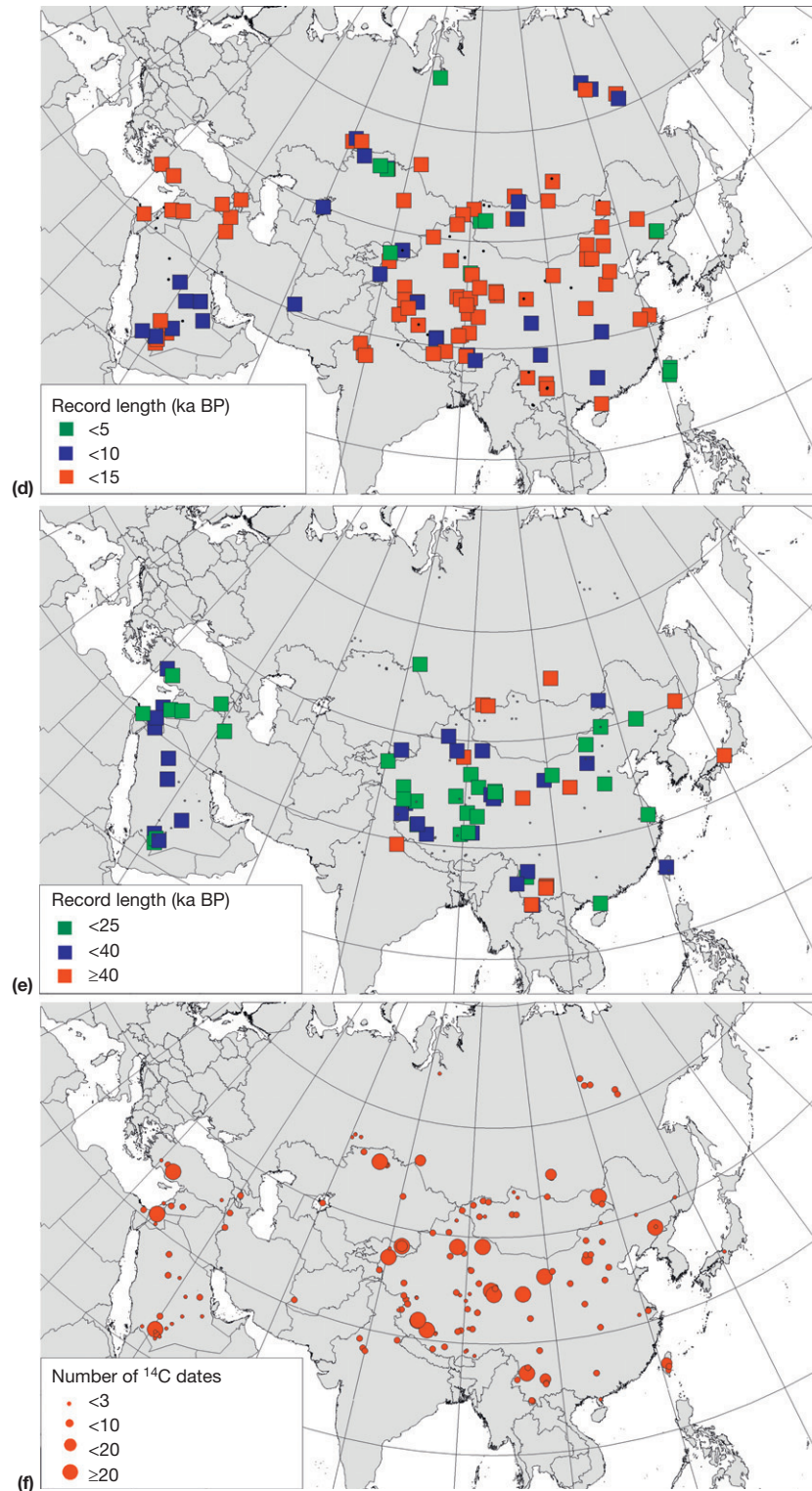


Figure 1 (Continued)

Studies of the aquatic communities in modern Asian lakes provide the basis to infer water salinity from species assemblages. Such relationships of water salinity and depth ranges of ostracodes/diatoms are shown in [Figure 2\(a\)](#) and [2\(b\)](#), respectively.

Deep-Lake Drilling in Asia

Asia has deep-lake basins with thick lacustrine deposits that are ideal sites for exploring lake-level history and long-term climate change. Considerable progress has been made toward

reconstructing past climate in Asia, due in part to deep-lake drilling programs (e.g., [Baikal Drilling Project, 1997](#)). Deep-lake drilling has taken place in the Dead Sea, with a record back to 400 ka BP ([Stein et al., 2005](#)); Lake Baikal in Russia, which produced a 200 m-long core with a 150-ka BP record ([Oberhänsli and Mackay, 2005](#)); Lake Biwa in Japan, a 911 m-long core dating from 430 ka BP ([Takemura et al., 2000](#)); Cuoe Lake in China, a 197 m-long core with a 2800-ka BP record ([Shen et al., 2004](#)); and Heqing Paleolake in China, a 737 m-long core with a 2600-ka BP record ([Hu et al., 2005](#)).

The deep-drilling core from Heqing Paleolake on Yunnan Plateau, southwestern China, was sampled at 10–50 cm

intervals, with 900–2000-year resolution ([An et al., 2011](#)). The record, which covers 2.6 Ma, has demonstrated the importance of inter-hemispheric forcing in driving Indian summer monsoon (ISM) variability at the glacial–interglacial time scale. The Heqing ISM index, global ice volume of benthic $\delta^{18}\text{O}$, and the Antarctic deuterium record show a variable relationship of Heqing ISM minima with global ice-volume maxima during three intervals over the past 2.6 My. In the middle interval (1.82–0.92 Ma), most ISM minima are coincident with ice-volume maxima. In the younger (<0.92 Ma) interval, most ISM minima lead ice-volume maxima. In the older (2.60–1.82 Ma) record, half of ISM minima lead

Ostracode	Water depth		Water salinity			Measured salinity range (g l ⁻¹)
<i>Candoniella leatea</i>						>1
<i>C. allicans</i>						
<i>C. mirabilis</i>						
<i>Candona neglecta</i>						
<i>C. xizangensis</i>						
<i>C. candida</i>						<2
<i>C. rausoni</i>						
<i>C. gyirongensis</i>						
<i>Cyprideis torosa</i>						<1 to 120
<i>C. littoralis</i>						2–5
<i>Cyprinotus</i> spp.						
<i>Cundimella mirabilis</i>						20–30
<i>Dolerocypris fasciata</i>						
<i>Eucypris gyirongensis</i>						<5
<i>E. inflata</i>						8.5–256.7
<i>E. pigna</i>						
<i>E. imglata</i>						
<i>Ilyocypris biplicate</i>						0.51–280.07
<i>I. bradi</i>						
<i>I. gibba</i>						1.78
<i>I. fadyi</i>						
<i>Leucocythere mirabilis</i>						0.487–256.733
<i>L. postilirata</i>						38574.00
<i>L. bispinosa</i>						0.67–0.72
<i>L. binoda</i>						
<i>L. dorsotuberosa</i>						1.78
<i>Leucocytherella sinensis</i>						0.74–12.90
<i>Limnocythere dubiosa</i>						1.50–33.72
<i>L. sancti-patricii</i>						
<i>L. inopinata</i>						0.5–25
<i>Limnocytherellina binoda</i>						80.00
<i>L. bispinosa</i>						0.6–172.95
<i>L. trispinosa</i>						20.00
<i>L. kunlunensis</i>						
<i>Cypridopsis ofesa</i>						
<i>Potamocypris</i>						
	Deep	Shallow	Fresh (<1g l ⁻¹)	Brackish (1–35 g l ⁻¹)	Saline (>35 g l ⁻¹)	

(a)

Figure 2 Water depth and salinity indicated by ostracodes (a) and by diatoms (b). Review of data sources from [Yu and Ke \(2007\)](#).

(Continued)

Diatom	Ecology			Water salinity	
<i>Achnanthes exigua</i>					
<i>A. lanceolata</i>					
<i>Amphora ajajensis</i>					
<i>A. delicatissima</i>					
<i>A. mexicana</i>					
<i>A. ovalis</i>					
<i>A. perpusilla</i>					
<i>Anomoeoneis sphaerophora</i>					
<i>Aulacoseira granulata</i>					
<i>Cocconeis placentula</i>					
<i>C. disculus</i>					
<i>Coscinodiscus lacustris</i>					
<i>Cyclostephanos dubis</i>					
<i>Cyclotella comta</i>					
<i>C. ocellata</i>					
<i>C. commensis</i>					
<i>C. meneghiana</i>					
<i>C. kutzingiana</i>					
<i>C. quadrijuncta</i>					
<i>C. stelligera</i>					
<i>C. spp.</i>					
<i>Cymbella affinis</i>					
<i>C. minuta</i>					
<i>C. cymbiformis</i>					
<i>C. cistula</i>					
<i>C. gracilis</i>					
<i>C. hustedtii</i>					
<i>C. parva</i>					
<i>C. helvetia</i>					
<i>C. lanceolata</i>					
<i>C. leptocera</i>					
<i>C. lata</i>					
<i>C. stelligera</i>					
<i>C. turgida</i>					
<i>C. ventricosa</i>					
<i>C. aequalis</i>					
<i>C. lacustris</i>					
<i>C. amphicephala</i>					
<i>C. laevis</i>					
<i>Cymato pleura</i>					
<i>C. solea</i>					
<i>Diploneis elliptica</i>					
<i>D. parma</i>					
<i>Epithemia intermedia</i>					
<i>E. sorex</i>					
<i>E. zebra</i>					
<i>Eunotia arcus</i>					
<i>E. argus</i>					
<i>E. zebra</i>					
<i>E. sorex</i>					
<i>E. turgida</i>					
<i>E. adnata</i>					
<i>E. reicheltii</i>					
<i>E. proboscidea</i>					
<i>Fragilaria lapponica</i>					
<i>F. brevistriata</i>					
<i>F. construens</i>					
<i>F. bicapitata</i>					
<i>F. intermedia</i>					
	Planktonic	Epiphytic	Benthic	Fresh	Brackish

Diatom	Ecology			Water salinity	
<i>Fragilaria zeilleri</i>					
<i>F. pinnata</i>					
<i>F. virscens</i>					
<i>Gomphonema constrictum</i>					
<i>G. acuminatum</i>					
<i>G. intricatum</i>					
<i>G. subtile</i>					
<i>G. longiceps</i>					
<i>G. angrenensis</i>					
<i>G. olivaceum</i>					
<i>Gyrosigma attenuatum</i>					
<i>G. scalproides</i>					
<i>Hantzschia amphioxys</i>					
<i>Melosira granulata</i>					
<i>Mastogloia elliptica</i>					
<i>M. smithii</i>					
<i>M. streptoraphe</i>					
<i>Nitzschia denticula</i>					
<i>N. frustulum</i>					
<i>N. sinuate</i>					
<i>N. angustata</i>					
<i>N. hantzschiana</i>					
<i>Navicula oblonga</i>					
<i>N. radiosa</i>					
<i>N. notha</i>					
<i>N. stroemii</i>					
<i>N. exilis</i>					
<i>N. schonfeldii</i>					
<i>N. diluviana</i>					
<i>N. tuscula</i>					
<i>N. pupula</i>					
<i>N. mutica</i>					
<i>N. rhynchocephala</i>					
<i>N. scutelloides</i>					
<i>N. lanceolata</i>					
<i>N. caroliniana</i>					
<i>N. gracilis</i>					
<i>N. cincta</i>					
<i>N. cryptocephala</i>					
<i>Neidium bisulcatum</i>					
<i>Opephora vulgura</i>					
<i>Pinnularia borealis</i>					
<i>P. cuneata</i>					
<i>Rhopalodia gibba</i>					
<i>Rhoicosphenia curvata</i>					
<i>R. parallela</i>					
<i>R. currata</i>					
<i>Stauroneis acuta</i>					
<i>S. phoenicenteron</i>					
<i>Stephanodiscus astraea</i>					
<i>S. rotula</i>					
<i>Synedra capitata</i>					
<i>S. ulna</i>					
<i>S. intermedia</i>					
<i>S. radians</i>					
<i>S. amphicephala</i>					
<i>Tabellaria flocculosa</i>					
<i>T. fenestrata</i>					
	Planktonic	Epiphytic	Benthic	Fresh	Brackish

(b)

Figure 2 (Continued)

ice-volume maxima. In contrast, interglacial ISM maxima are correlated with global ice volume minima and associated with an enhanced Indian low. These patterns suggest that feedbacks from ice sheet and ocean dynamics can modify effects of orbitally forced insolation (An et al., 2011). Such records can be compared with deep-sea drilling studies to describe long-term cycles of climate change in Asia and thus improve understanding of ocean-continent linkages, and the roles of orbital forcing and inter-hemispheric forcing in the Earth system.

From Individual Study to Regional Synthesis

The recognition that lake data could be used to reconstruct past regional climate changes led to the construction of the African lake-level database, and thus to a global database: the Oxford Lake Level Data Base (OLLSDB: Street-Perrott et al., 1989). However, the OLLDB only featured closed lake basins. The recognition that open lakes also showed climate-induced changes led to efforts to synthesize lake data for more overflow lakes than the close-basin lakes and areas: the European Lake Status Data Base, the Former Soviet Union and Mongolian Lake Status Data Base (FSUDB: Tarasov et al., 1996) and the Chinese Lake Status Data Base (CLSDB: Yu et al., 2001a,b) have now been established. As more individual Asian lake-level studies are carried out, it has become possible to synthesize regional lake-level changes and identify changing patterns at a continental scale (Fang, 1991; Farrera et al., 1999; Harrison et al., 1996; Shi et al., 2001). The changing patterns have been interpreted in terms of changes in interplay of the westerly and the Asian monsoons during the late Quaternary. These lake-level records and databases provide a systematic basis for mapping lake-level changes at a variety of spatial scales, identifying the patterns of hydrological and atmospheric circulation changes, and evaluating precipitation and P-E fields of paleoclimate simulations (see below).

From Reconstruction to Determining Causes of Lake-Level Change

Lake-level changes are controlled by two fundamental factors: lake-basin change and change in water input. Changes in lake-basin volume can, for example, result from tectonism, siltation/scouring, and changed river routes. Changes of water input reflect the discharge into the lake, which can be a direct consequence of climate change in the catchment: variation of strength and intensity of precipitation and/or variation in evaporation reflecting air temperature and humidity levels. Indirectly, climate influences land-surface conditions, which influence evapotranspiration and runoff.

Mapping modern lake status onto P-E shows that modern lake-level changes in China are mainly controlled by regional P-E (Figure 3(a)). Among 487 large natural lakes (surface area greater than 10 km²) there are 57.1% saline lakes and 42.9% freshwater lakes. Most freshwater lakes are found at the locations where there is a positive water balance ($P > E$) and mean annual precipitation is >600 mm. Where the water balance is negative ($P < E$) and mean annual precipitation is <600 mm, saline lakes dominate. Comparing this pattern with the modern lake-level status (0 ka BP) reconstructed from geological evidence, (Yu et al., 2001a) show lakes with high and

intermediate levels are mostly distributed in Southern China and those with low levels are in northern and northwestern China and the Tibetan Plateau (Figure 3(b)). The boundary line between the high-intermediate and the low lake-level regions is located at 35–30° N, virtually coincident with the 0-mm contour of annual P-E.

Nowadays, lake-level studies are not only focused on reconstructing processes and regional patterns but are also linked with paleoclimate modeling, which allows the exploration of mechanisms. Possible causes of Asian lake-level changes were studied through data and modeling comparisons (COHMAP Members, 1988; Qin et al., 1998; Yu et al., 2003). These studies have revealed that the lake-level changes are the results of different mechanisms at different scales: direct response to catchment P-E changes on a regional scale, consequence of changes in atmospheric circulations and land-surface conditions on subcontinental and continental scales, and fundamental changes in the global driving forces of climate: insolation, glaciation, and land-sea dynamics during the late Quaternary.

Overview of Regional Asian Lake-Level Changes Through the Late Quaternary

Based on lake-level databases of the OLLDB, FSUDB, and CLSDB, we can systematically view time sequences of Asian lake-level changes. For the convenience of presentation, the Asian region is divided into four subregions based on their geographical features: the Middle East (20–55° E and 10–50° N), Central Asia (55–104° E and 40–70° N), the Tibetan Plateau and surrounding area (70–104° E and 10–40° N), and East Asia (104–150° E and 20–70° N). For each subregion, six to eight lake-level curves in part (a) of Figure 4(a)–4(d) and histograms in part (b) of Figure 4(a)–4(d) are plotted, showing a general temporal sequence of the grouped lake behaviors.

Middle East

There are 30 lake-level records in the subregion with recording lengths from <10 ka BP to >30 ka BP (Figure 4(a)). High lake levels dominated (30–80%) before 17 ka BP, low lake levels dominated (70–20%) after 17 ka BP, suggesting a long-term trend toward drying from 30 to 0 ka BP (currently 70% of lakes are at low levels).

Central Asia

There are 34 lake records (Figure 4(b)). Four peaks of high lake levels occurred at 30–25, 17–15, 7–6, and 4–3 ka BP. Low lake levels (40–60%) occurred at the Last Glacial Maximum (LGM) and in the early Holocene (~ 8 ka BP).

Tibetan Plateau and Surrounding Areas

Forty-one lake-level records are registered for the Tibetan Plateau subregion (Figure 4(c)). Three periods of high lake levels occurred at 30–24, 22–20, and 8–5 ka BP, respectively, with each high stage less prominent, indicating a long-term trend

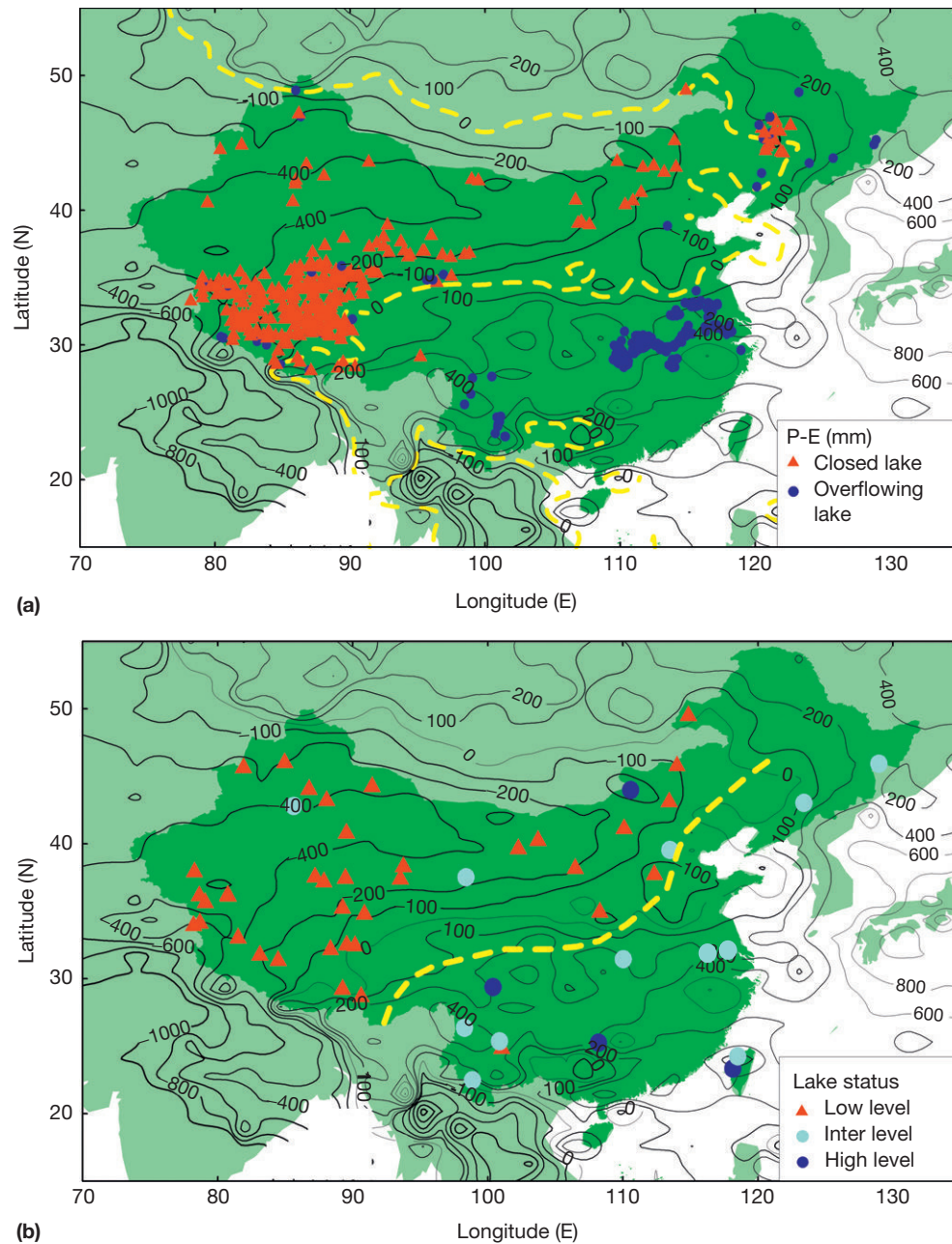


Figure 3 (a) and (b) Comparison of modern annual P-E and lake status (closed or overflowing) in China. (a) Contours of mean annual P-E (mm) where $P-E=0$ is shown by the dashed line. All 459 lakes plotted in the figure have surface areas greater than 10 km^2 . (b) Levels for 55 Chinese lakes at the present time (0 ka BP), where the boundary between low and high-intermediate lake status is indicated by the dashed line. Data sources from Wang and Dou (1998) and Yu et al. (2001b).

toward a more arid climate from 35 ka BP to the present. In fact, most modern lakes in the region are smaller now than at any time during the past 35 ka BP.

East Asia

There are 27 lake-level records (Figure 4(d)) from the east-Asian subregion. There are no records for the interval 23–18 ka BP. Evidence of sedimentary hiatuses or loess deposits suggests that this period was extremely arid and most lakes were dried

up. High lake levels occurred in the early and mid-Holocene (intermediate and high lake levels 95%).

Spatial Patterns of Asian Lake-Level Changes

Mapping of lake-level data makes it possible to identify patterns of hydrological and climatic change. Here, we introduce spatial patterns for three time slices (Figure 5), which represent key climate periods during the late Quaternary: the

mid-Holocene (~ 6 ka BP; e.g., [Qin et al., 1998](#)), the LGM (~ 18 ka BP in ^{14}C years or equivalent to 21 ka BP in calendar years; e.g., [Farrera et al., 1999](#)), and the late Marine Isotope Stage 3 (MIS-3) dated ~ 30 – 35 ka BP (e.g., [Shi et al., 2001](#)). The data sources are same as in section 'Historical Review of Asian Lake-Level Studies' above, and the climatic interpretations are based on the relevant literature.

Mid-Holocene

Reconstructed lake levels at 6 ka BP show conditions wetter than those of the present across much of the Asian continent, including the Russian Far East, northeastern north China, the Mongolian Plateau, the interior of Xinjiang, the Tibetan Plateau, the northern Indian mountains, and the Arabian Plateau. This wet belt was more than $\sim 10,000$ km long from east to west,

and $\sim 3,000$ km wide from north to south ([Figure 5\(a\)](#)). On both sides of this wet belt, lake levels were lower than the present, with a northern dry area surrounding the Caspian Sea, and the southern dry area in Southwestern and Southern China.

This pattern indicates a major change compared with the present. The key difference is the enhancement of Asian summer monsoon, forced by higher summer insolation that would have increased precipitation in South, Central and East Asia ([Kutzbach et al., 1998](#)). The melting of the last glacial ice sheets in the Northern Hemisphere resulted in northward-shifted and weakened Westerlies, leading to less precipitation over the eastern Mediterranean region ([Harrison et al., 1996](#)). The drying of southern China represented the northward and westward shift of the subtropical high-pressure system; conditions were dominated by offshore (land-sea) winds, low summer precipitation and high-annual aridity ([Qin et al., 1998](#)).

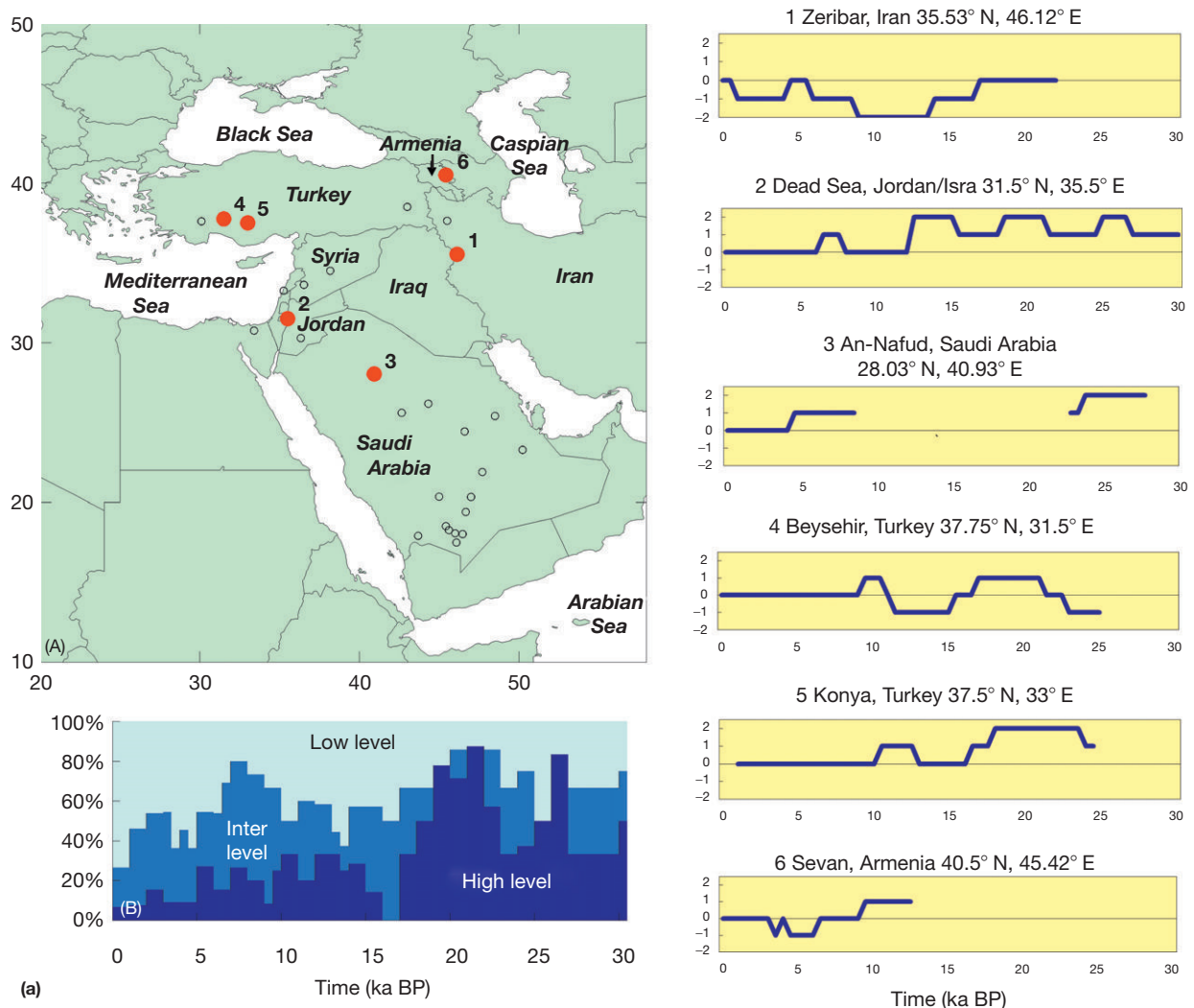


Figure 4 (a)–(d) Asian lake-level changes since 30 ka BP in four subregions. For each subregion there is a lake location map, a set of individual lake-level curve diagrams, and a histogram showing the numbers of lakes of high, intermediate, and low level through time. The lake locations are marked by circles; lakes coded as 1, 2, etc. have a matching lake-level curve. Vertical marks on the level curves (2, 1, 0, –1, –2) represent much higher, higher, no change, lower, much lower lake levels than today, respectively. Sources for histograms: OLLDB, FUSDB, and CLSDB. (a) Middle East; (b) Central Asia; (c) Tibetan Plateau and surrounding area; (d) East Asia.

(Continued)

LGM

Conditions drier than those of today characterized most areas east of 100° E, including the Yunnan Plateau, the Mongolian Plateau, the Eastern China Plains, and Taiwan. The extremely dry conditions meant that most lakes dried up; no lacustrine deposits have been found to date. Loess and mixed terrestrial deposits representing this period were found at Taihu Lake and Hulun Lake, where the modern water areas of these two lakes are 2428 and 2339 km², respectively (Yu et al., 2003). This pattern (Figure 5(b)) indicates that there was a stronger winter monsoon and weaker summer monsoon than present. This resulted from the dominance of a glacial anticyclone induced

by increased ice and snow coverage and the development of a highly continental air mass in the mid and high latitudes of Eurasia (Farrera et al., 1999; Harrison et al., 1996).

Most of the Tibetan Plateau and western China remained relatively wetter than today, and the Middle East was characterized by high lake levels (Figure 5(b)). There appears to have been a wet belt from the eastern Mediterranean, through Central Asia to western China. The high lake levels were interrupted at $100\text{--}105^{\circ}$ E, that is, Qinghai Lake (100.5° E) and Baijian Paleolake (104.2° E; Pachur et al., 1995). This pattern suggests that they may have resulted from a southward shift and strengthening of the Westerlies under the influence of the high-latitude

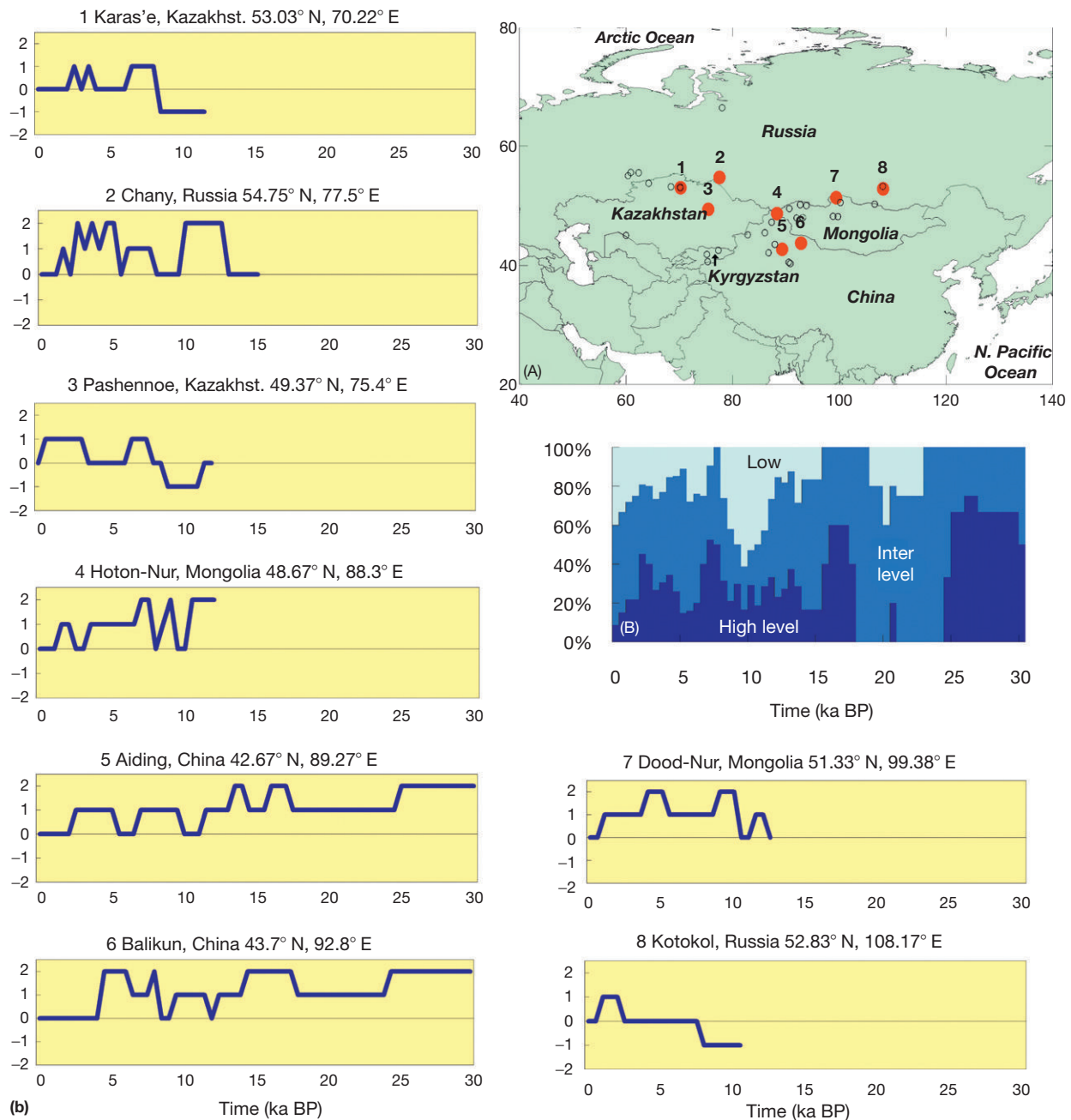


Figure 4 (Continued)

ice sheet and very low evaporation over an extremely cold plateau surface (Farrera et al., 1999; Qin and Yu, 1998).

Late MIS-3

The records of the period are mainly concentrated in the Tibetan Plateau, Yunnan Plateau, and western China, and they are generally characterized by wetter conditions than today (Figure 5(c)). Individual lake studies on the Tibetan Plateau have provided several high lake-level records during the interval 35–30 ka BP, such as Bangong Co (79 m higher than today), Selin Co (30 m higher), Tianshuihai Lake (40–60 m higher), Qarhan Salt Lake (29 m higher), Gaxun-Sogo Co (22–30 m higher), Nam Co (18–26 m higher) and Baijian

Paleolake (30 m higher) (Hu, 2001; Li, 2000; Pachur et al., 1995; Zheng et al., 1989; Zhu et al., 2003). In contrast, two sites from the Arabian Plateau and Taiwan registered a more negative P-E balance than today.

The large extent of wet conditions in Asia suggests a major increase in annual precipitation, possibly resulting from an enhanced southwest monsoon. This circulation change may have coincided not only with a large land–sea temperature difference between the equatorial Indian Ocean and the Tibetan Plateau, but also a strong Australian high-pressure system, which strengthened the south Asia monsoon via the trans-equatorial air mass: an air flow in the southern hemisphere that passes across the Equator towards the Asian subtropical areas of the northern hemisphere in the summer season (Shi et al., 2001).

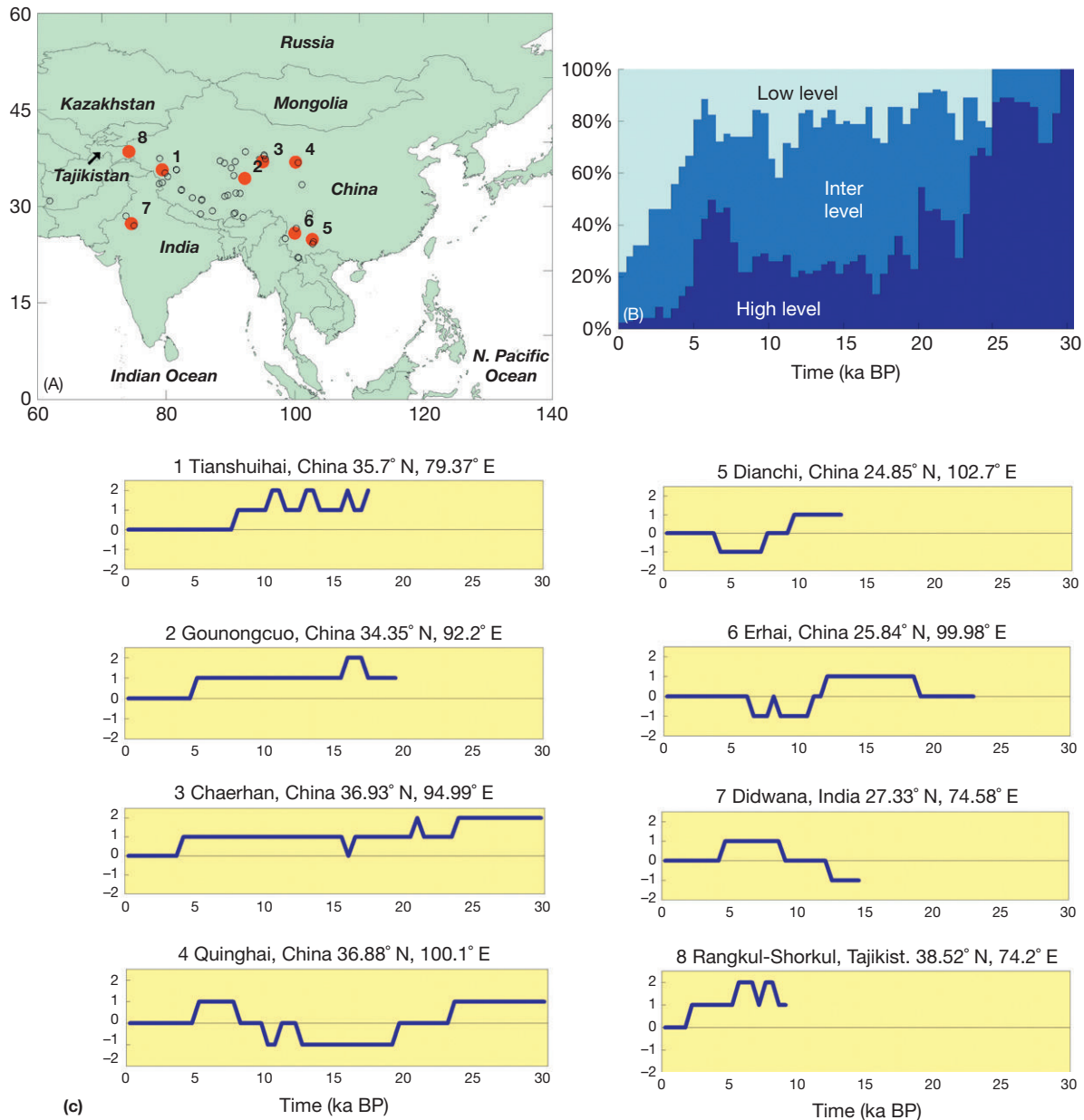


Figure 4 (Continued)

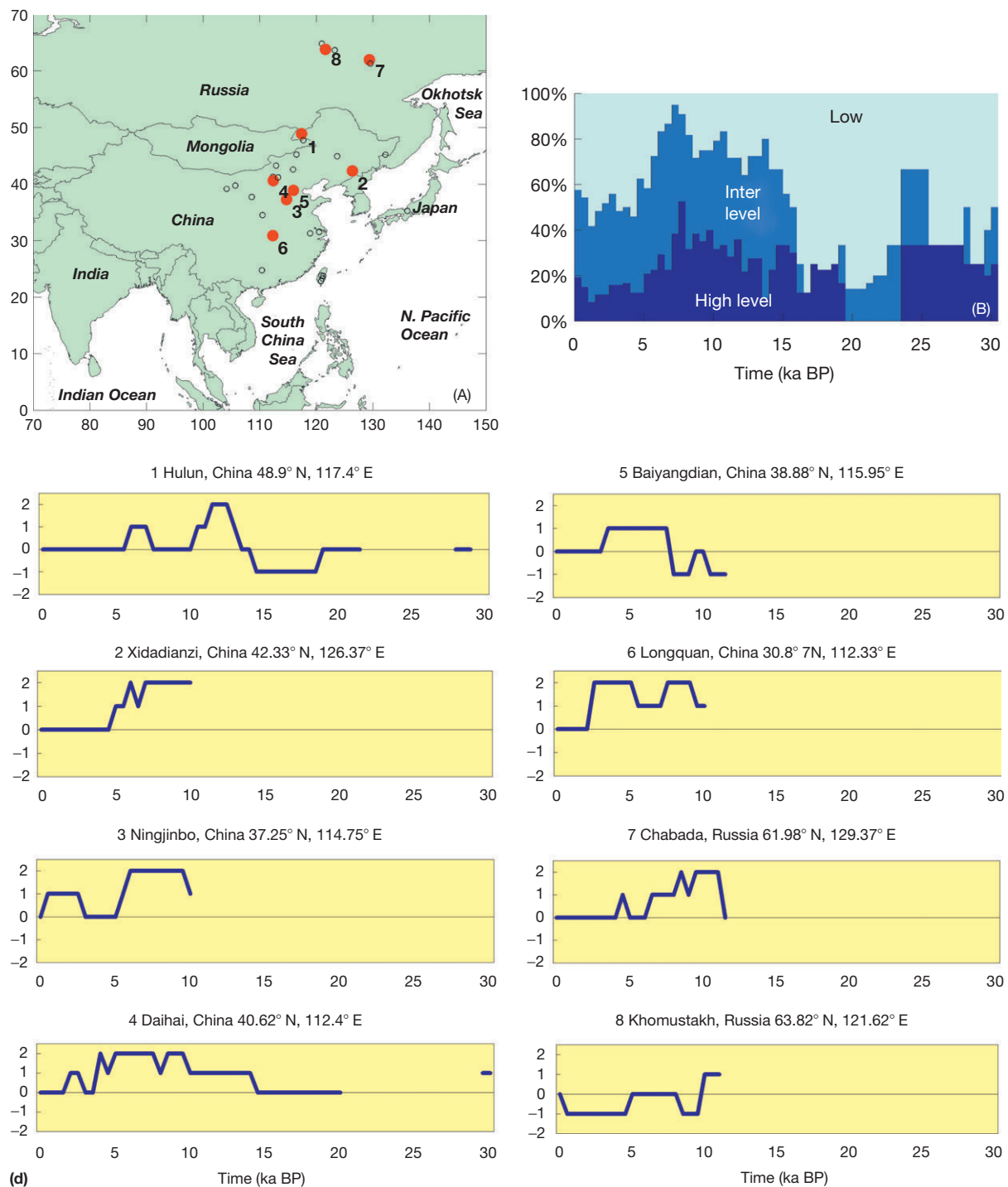


Figure 4 (Continued)

Lake-Level Change and Paleoclimate Modeling

On a late-Quaternary timescale Asian lake-level changes are clearly linked with climate change. The observed data can indicate the magnitude, direction, and timing of climate changes. However, it has proved harder to develop an

understanding of the mechanisms underlying the climate changes. The advent of paleoclimate simulations using physically based numerical models (e.g., Jousaume et al., 1999; Kutzbach et al., 1998) represents a major advance in our understanding of the underlying climate mechanisms. General circulation models (GCM) and regional climate models can

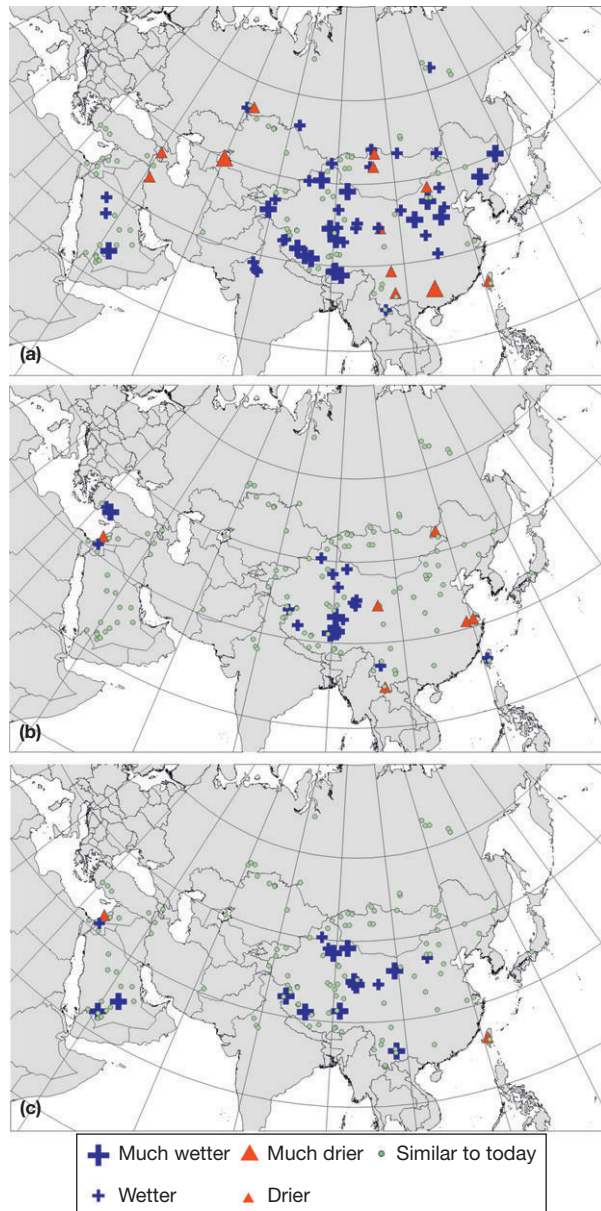


Figure 5 (a)–(c) Spatial patterns of lake-level changes in Asia at: (a) 6 ka BP; (b) 21 ka BP; and (c) 35 ka BP (sources: OLLDB, FUSDB, and CLSDB). Large and small crosses represent much wetter and wetter than today, and large and small triangles represent much drier and drier than today. Circles represent similar to today.

simulate atmospheric precipitation and effective moisture, and simulations can be compared with the character and spatial distribution of past lake-level changes.

A paleoclimate simulation reflects the status of key physical parameters that drive climate change. Many climate simulations for 6 and 21 ka BP use the Paleoclimate Modeling and Intercomparison Project (Joussaume et al., 1999) values for insolation, ice cover, sea-surface temperatures, and atmospheric CO₂ concentrations. Simulations for 35 ka BP (Yu et al., 2007) reflect a positive insolation anomaly and reduced extent of the Laurentide and Fennoscandian ice sheets (50–70% of those during the LGM). Furthermore, Asian vegetation cover

reconstructed from paleoecological records has been used in paleoclimate simulations in order to examine land-surface effects on climate (e.g., Yu et al., 2001b). The results of paleoclimate simulations for the key periods of 6, 21 and 35 ka BP and their implications for our understanding of the mechanisms of Asian lake-level changes are introduced below.

The Climate at 6 ka BP

Paleoclimate simulations for 6 ka BP produce significantly warmer summers in the low and mid-latitudes of Asia (Joussaume et al., 1999; Qin et al., 1998; Yu et al., 2001b); vegetation feedback reversed orbitally-forced winter cooling and resulted in a warmer winter (and higher mean annual temperature) than present (Figure 6(a)). Simulated increases of precipitation in Arabian, Indian, and Tibetan plateaus were induced by a significantly enhanced Asian monsoon. The reduction in precipitation over southern China (Figure 6(b)) is linked with adjustments in the position and strength of the Pacific Subtropical High induced by orbital forcing.

A climate simulation for 6 ka BP using the regional climate model RegCM2 has generated a high-resolution (120 × 120 km² per gridpoint) climatic image at the scale of Asia (Zheng et al., 2004). The increase in Northern Hemisphere summer insolation at 6 ka BP led to a significant temperature increase in the mid- and high latitudes in Asia and enlargement of the sea–land thermal contrast, which in turn caused the strengthening of Asian summer monsoon and hence the west- and northward expansion of its rain belt. The 6-ka BP P-E simulation (Figure 6(c)) shows a prominent increase over most of Asia, and the low- and mid-latitude precipitation belts extended westward and northward. The 850-hPa summer southerly wind components show that along with the strengthening of the Asian summer monsoon, the north border of southerly winds advanced markedly northward and westward, which favors intrusion of warm and wet flow from the South China Sea and West Pacific into the interior of Asia. The 850-hPa winter northerly wind is weakened, and the advection cooling into East Asia correspondingly reduced. Consequently, there are annual convergences over Mongolia, north-northeast China, Tibet, and India, indicating more moisture and frequently wet conditions. Conversely, divergences over southern China contribute to dry conditions (Figure 6(d)).

The Climate at 21 ka BP

GCM simulations (e.g., COHMAP Members, 1988; Kutzbach et al., 1998; Yu et al., 2003) show global annual temperatures, including those for Asia, generally lower than today. Across most latitudes globally, P-E values were also lower (with the exception of North America south of the ice sheet and the Mediterranean, where southerly displacement of the Westerlies enhanced precipitation). Simulations of less precipitation over most areas of Asia (Figure 7(b)) and negative P-E anomalies in the Southeast Asian tropical area and East Asian lowlands are linked to a weakened Asian summer monsoon and a persistent winter Mongolian High. The simulated increases in annual precipitation and P-E in west China were due to a southward and eastward shift in the position of the Westerlies plus a decrease in evaporation in the cool conditions that prevailed

across Asia (temperatures simulated at 4–16 °C lower than today: [Figure 7\(a\)](#)). A regional climate model simulation of the annual P-E pattern for the LGM ([Figure 7\(c\)](#)) reveals two patterns of moisture flux increase over the Tibetan Plateau in the annually averaged 850 hPa map: one from the East China

Sea and the other from the northern Bay of Bengal ([Zheng et al., 2004](#)). These patterns are consistent with the observed lake-level changes ([Figure 7\(d\)](#); the east-Asian region and adjacent seas were areas of net moisture output and substantially reduced precipitation).

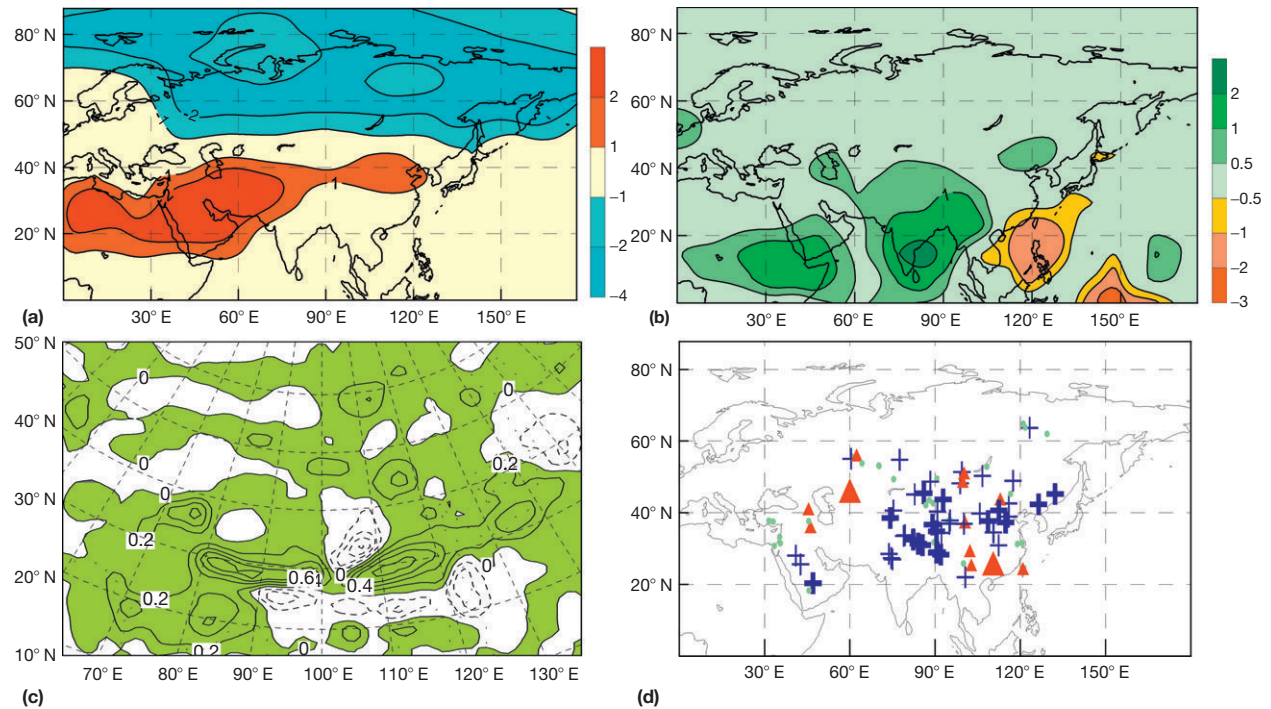


Figure 6 (a)–(c): GCM-derived paleoclimate simulations for 6 ka-BP shown as anomaly maps (departures from present values). (a) mean annual temperature (°C); (b) precipitation (mm d⁻¹); (c) P-E (mm d⁻¹). Values derived from [Yu et al. \(2001b\)](#) and the regional climate model simulations of [Zheng et al. \(2004\)](#). Figure 6(d) shows moisture conditions indicated by the lake-level records for comparison (symbols the same as [Figure 5](#)).

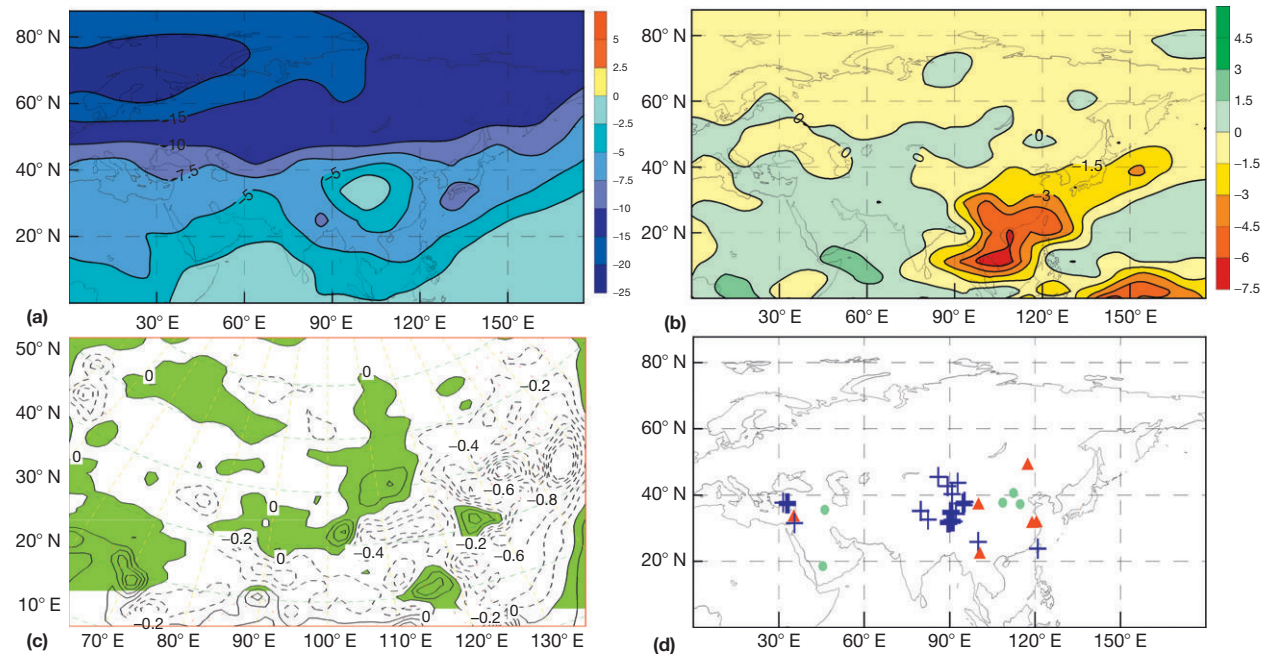


Figure 7 (a)–(d) As in [Figure 6](#), but for 21 ka BP.

The Climate at 35 ka BP

Regarding the period 30–35 year BP (late MIS3), a GCM simulation of the Asian climate focused on 35-ka BP, the period of lake high-stands in Tibet and northwest China, shows higher temperatures compared with present in most areas of low- and mid-latitude Asia (**Figure 8(a)**), reflecting forcing by increased insolation (Yu et al., 2007). Lower temperatures than present at high latitudes mainly reflect the cooling effects of the Quaternary ice sheets. The simulated monsoonal precipitation belt is shifted northwards (**Figure 8(b)**); the annual P-E simulation shows conditions wetter than today in the Indian Peninsula, Tibetan Plateau and northern China, but drier in southern China (**Figure 8(c)**). Changes in surface heating patterns (i.e., larger differences in insolation between high and low latitudes) and circulation (an increase in meridional exchange but a decrease in east–west exchange) and land-surface conditions (extension of southern forests in East Asia) led to wet conditions in inland areas that are dry today.

The atmospheric circulation changes over Asia reveal the likely mechanisms behind lake-level change at 35 ka BP (**Figure 8(d)**). Strong precession over low latitudes compared with modern orbital conditions and Quaternary ice sheets at high latitudes were important factors affecting the Asian 35 ka BP climate: the temperature decreased at mid-high latitudes but increased at low latitudes, which generated larger south–north temperature gradients compared with today. In turn, moisture transport from low to high latitudes was enhanced, and monsoonal precipitation belts were shifted northwards to Northwest China, while dry conditions developed in South China. This climatic pattern at 35 ka BP is linked to the occurrence of larger lake areas and higher lake levels than today in Tibet and Northwest China (Li, 2000; Pachur et al., 1995; Shi et al., 2001; Zheng et al., 1989; Zhu et al., 2003).

Recent Progress and Future Outlook

Asian lake-level studies have provided a scientific context for understanding environmental changes in the future, both as regional histories of hydrology and climate in terms of processes, frequency, and trends, or as benchmarks for paleoclimate model evaluation and testing. The accuracy of model-based predictions of future climate cannot be directly assessed, but the models used to generate such predictions can be tested by simulating the climates of times in the past and comparing these simulations with paleoclimate reconstructions. This is one of the important goals of lake-level studies. However, there remain considerable uncertainties about the past lake-level changes, which will in turn affect model evaluations and projections for the future.

During the last ten years, additional lake-level studies from Asia have been carried out using multiple proxies from lacustrine sediments (**Table 1; Figure 9**). There has been progress toward the filling of key geographical gaps in data, in applying tools to derive quantitative estimates of lake-level change, in the use of high-resolution sampling to achieve time series resolved at 10^1 – 10^2 years, and in deep-lake drilling to obtain million-year records. For example, at Qinghai Lake (No. 15 in **Figure 9**) on the Tibetan Plateau sediment cores have been taken to reconstruct a long lake history, with high-resolution sampling and many dates on core material. The 795-cm long core was sampled at 1–4 cm intervals (22–90 years resolution; Shen et al., 2005a,b). Such high-resolution studies have revealed short-term climatic oscillations, for examples, low lake levels coincided with the Younger Dryas stadial, and sharp drying lasting ca. 100 years was centered on the 8.2 ka BP the cooling event, with other lake level fluctuations occurring between 8.5–7.8 ka BP.

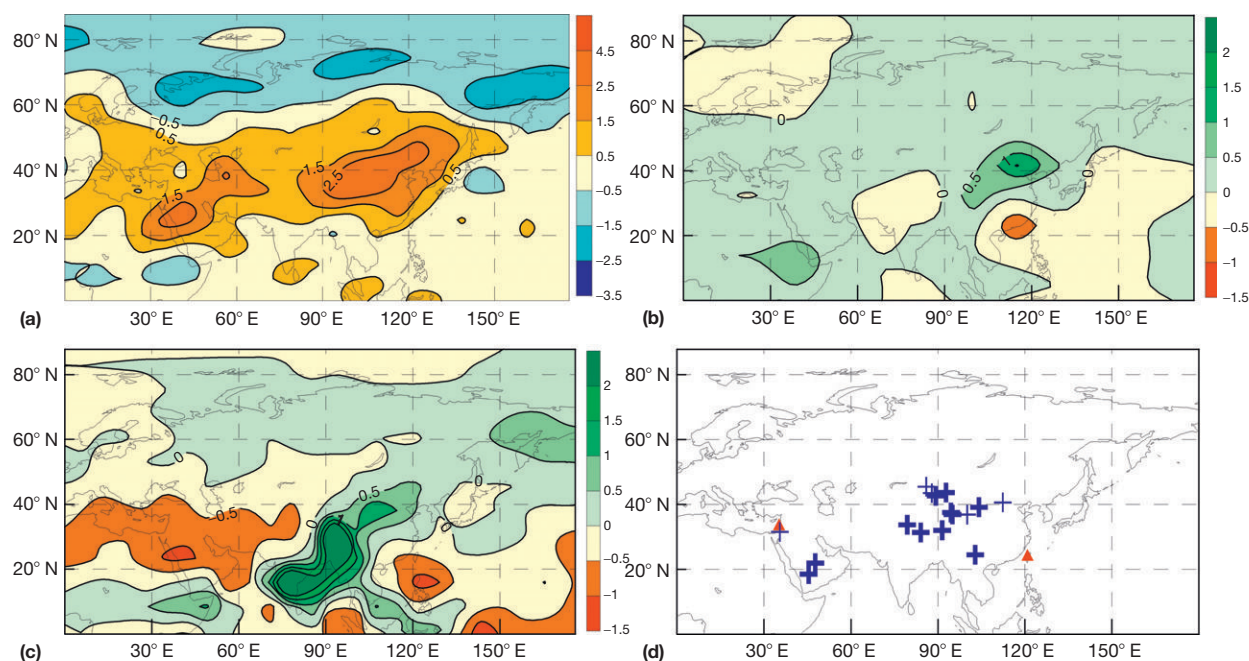


Figure 8 (a)–(d) As in **Figure 6**, but for 35 ka BP.

Table 1 Information on new lake-level studies since 2000

No.	Name	Lat. (°N)	Long. (°E)	Elev. (m a.s.l.)	No. ¹⁴ C dates	Record length (ka BP)	Major references
1	Aibi Lake	45.00	82.80	194	3	20	Wu et al. (2003)
2	Peiku Co	28.83	85.33	4590	3	45	Yu (2008)
3	Zarinanmu Co	30.95	85.50	4653	1 + ESR10	12	Yu (2008)
4	Angren Lake	29.30	87.18	4300	6	11	Peng and Li (2003)
5	Wulungu Lake	47.00	87.50	478	7	8.2	Liu et al. (2008)
6	Bosten Lake	42.15	90.00	1047	6	12	Huang et al. (2009); Mischke and Wünnemann (2006)
7	Chen Co	28.85	90.52	4429	7	20	Zhu et al. (2004)
8	Nam Co	30.70	90.65	4718	10 + ESR3 + 23Us	116	Zhu et al. (2003); Zhu et al. (2010)
9	Cuoe Lake	31.50	91.50	4532	9 + TL4 + Mag	2800	Shen et al. (2004)
10	Kusai Lake	35.71	92.91	4475	7	4	Liu et al. (2009)
11	Sugan Lake	38.85	93.9	2800	5	2.7	Zhang et al. (2010)
12	Huahai	40.43	98.07	1195	14	13.8	Wang et al. (2010)
13	Heqing Paleolake	26.56	100.17	2190	17 + Mag	2600	An et al. (2011); Hu et al. (2005)
14	Caohai	26.60	100.20	1800	7	32	Lin and Wei (2000)
15	Qinghai Lake	36.90	100.70	3246	10 + OSL17	18	Shen et al. (2005b)
16	Ximen Co	33.38	101.67	4020	7	35	Pu et al. (2011)
17	Poyang Lake	29.05	116.20	25	6	8	Ma et al. (2004)
18	Gucheng Lake	31.20	118.80	5.5	5 + OSL3	10	Yao et al. (2007)
19	Taihu Lake	31.30	119.80	121	6	12	Zong et al. (2011)
20	Dajiu Lake	31.50	110.50	1700	10	15.75	Zhu et al. (2006)
21	Nansi Lake	34.89	116.96	33	4	2.5	Shen et al. (2008)
22	Bahannao	39.32	109.27	1278	15	24	Guo et al. (2007)
23	Daihai	40.45	112.45	1255	12	20	Sun et al. (2009)
24	Huangqihai	40.80	113.45	1264.1	4 + OSL2	40	Shen et al. (2005a)
25	Angulinao	41.35	114.42	1312	3	8.5	Jiang et al. (2004)
26	Juyanhai	42.12	102.07	820	6	2.6	Chen et al. (2010)
27	Sihailongwan Maar Lake	42.28	126.6	791	16	18.5	Liu et al. (2005)
28	Dabusu Lake	45.00	123.65	140	4	10.4	Tareq et al. (2006)
29	Lake Baikal	53.25	108.1	456	18	150	Oberhänsli and Mackay (2005)
30	Lake Biwa	35.25	136.07	85.6	3 + 28Tep	430	Takemura et al. (2000)
31	Dead Sea	31.5	35.5	−417	Us	400	Stein et al. (2005)

OSL, TL, ESR, Us, Tep, and Mag represent dating by optical stimulated luminescence, thermoluminescence, electron spin resonance, uranium-series, tephrochronology, and magnetostratigraphic chronology, respectively.

Lake level changes have recently been reconstructed via quantitative approaches. For example, at Nam Co (No. 8 in [Figure 9](#)) on the Tibetan Plateau, which has a lake area of 1920 km², past water depth was inferred from an ostracode-based transfer function derived from 25 surface sediment samples taken between 47 and 93.5 m depth ([Zhu et al., 2010](#)). The percentages of four dominant species were used to infer paleo water-depth values, of which the maximum inferred water depth was about 63 m and the minimum 15.7 m. Water depth during the last 8400 years has been reconstructed from a 332-cm long sediment core. Several stages are apparent: in the early mid-Holocene water depth was much lower than today, dropping from an estimated 50 m to 20–30 m between 8400 and 6800 years BP. The lake water depth remained at this low stand between 6800 and 5700 years BP but then gradually deepened to around 50 m during 5700 and 2900 year BP. A cold-dry event occurred between 1700 and 1500 years BP, when lake depth decreased slightly, but the lake then continuously deepened to 60 m after 1500 years BP. The study shows that ostracode assemblages are not only indicative of

cold-warm conditions, but may also be used to imply the dry-wet status of a lake by inferring water-depth variations.

Many Asian countries are currently facing severe problems in connection with lacustrine environments and water resources, and these are expected to increase in the twenty-first century: diminishing lake areas, salinization, and eutrophication. To meet these challenges, lake-level studies should pay more attention to the historical period in order to understand both natural evolution and anthropologic impacts. Some studies have attempted to apply lake-level records to scenarios of regional hydrological changes in lake basins. Because catchment water balance and lake water levels in arid and semiarid areas of China are sensitive to direct and indirect changes related to precipitation and temperature, lake-level records not only show lake responses to climate changes in the past, but they can be used as analogs to estimate lake changes with future climate change. Four large lakes: Aibi Lake (542 km²), Qinghai Lake (4635 km²), Daihai Lake (134 km²), and Hulun Lake (2339 km²) which lie on a west-to-east transect across northern China (locations see No. 1, 15, 23 and 32 in [Figure 9](#)), have

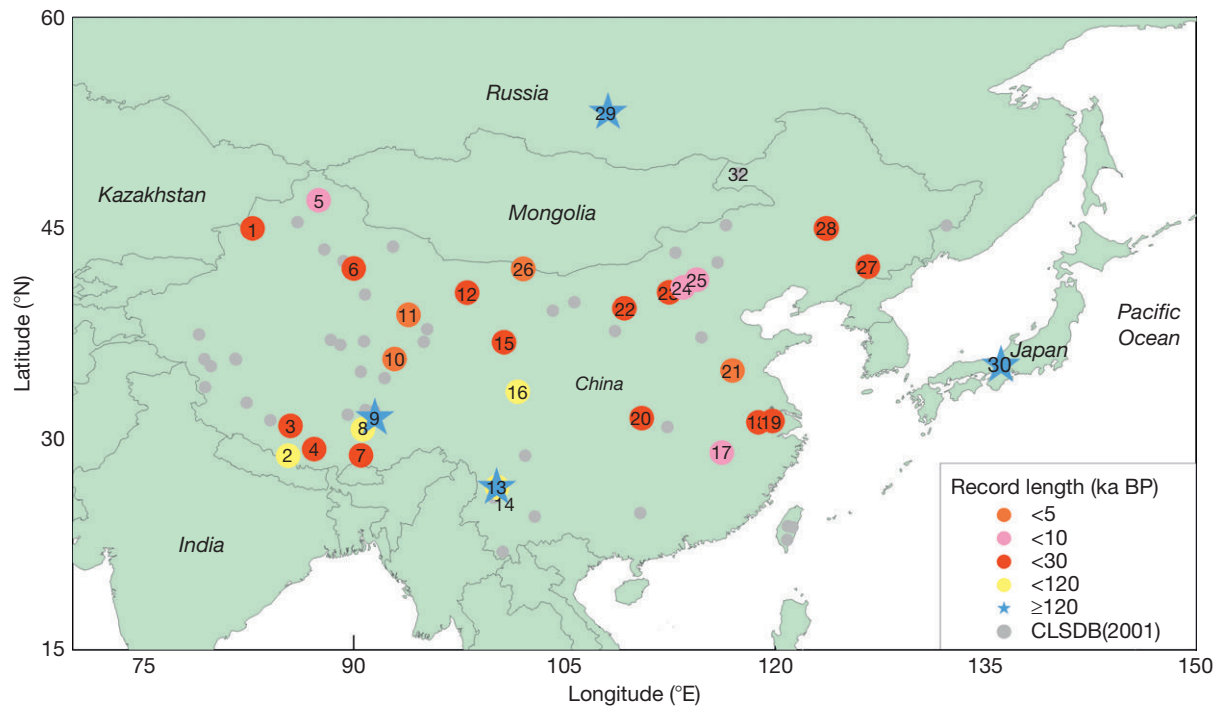


Figure 9 Lake locations and the record lengths updated for lake level studies since 2000 (lake numbers and references same as Table 1) from the CLSDB (Yu et al., 2001a). Star sites are deep-lake drilling cores in Asia. Data sources listed in Table 1.

been used for such studies (Yu and Shen, 2009). Lake-level time series derived from the gaged period of the past 50 years and from reconstructions of the past 30 ka BP provide frequencies and amplitudes of variation at different time scales; these are used in simulating the probability of future water-level changes. GCM climate simulations based on Intergovernmental Panel on Climate Change (IPCC) scenarios were applied to the four lake basins. Given a ~95% chance of a temperature increase in the coming 50 years, the predicted twenty-first century variation in water level is about $\pm 50\%$ in Qinghai Lake, $+25$ to $+70\%$ in Daihai Lake, $+30\%$ to -20% in Hulun Lake and $\pm 20\%$ in Aibi Lake (Yu et al. 2009). The scenarios can be used to evaluate the probability of changes in lake volume occurring over the next 50 years, and they will provide an important basis upon which to assess uncertainties in future hydrological changes.

Acknowledgments

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See also: **Lake Level Studies:** Africa during the Late Quaternary; Modeling; Overview; West-Central Europe. **Paleoclimate Modeling:** Data–Model Comparisons; Quaternary Environments. **Paleoclimate Reconstruction:** Approaches. **Paleolimnology:** Overview of Paleolimnology.

References

- An ZS, Clemens SC, Shen J, et al. (2011) Glacial–Interglacial Indian summer monsoon dynamics. *Science* 333(719): 719–723.
- Baikal Drilling Project BDP-96 (Leg II) Members (1997) A continuous record of climate changes for the last five million years from the bottom sediment of Lake Baikal. *Russian Geology and Geophysics* 39: 135–154.
- Chen KZ and Bowler JM (1986) Late Pleistocene evolution of salt lakes in the Qaidam Basin, Qinghai Province China. *Palaeogeography, Palaeoclimatology, Palaeoecology* 54(1–4): 87–104.
- Chen HF, Song SR, Lee TQ, et al. (2010) A multiproxy lake record from Inner Mongolia displays a late Holocene teleconnection between Central Asian and North Atlantic climates. *Quaternary International* 227(2): 170–182.
- COHMAP Members (1988) Climatic changes of the last 18,000 years: Observations and model simulations. *Science* 241: 1043–1052.
- Dearing JA and Foster IDL (1986) Lake sediments and palaeohydrological studies. In: Berglund B (ed.) *Handbook of Holocene Palaeoecology and Palaeohydrology*, pp. 67–90. New York: John Wiley and Sons.
- Fang J-Q (1991) Lake evolution during the past 30,000 years in China, and its implications for environmental changes. *Quaternary Research* 36: 37–60.
- Farrera I, Harrison SP, Prentice IC, et al. (1999) Tropical climates at the last glacial maximum: A new synthesis of terrestrial palaeoclimate data. I. Vegetation, lake-levels and geochemistry. *Climate Dynamics* 15: 823–856.
- Guo LL, Feng ZD, Li XQ, et al. (2007) Holocene climatic and environmental changes recorded in Baahar Nuur Lake core in the Ordos Plateau, Inner Mongolia of China. *Chinese Science Bulletin* 52(7): 959–966.
- Harrison SP and Digerfeldt G (1993) European lakes as palaeohydrological and palaeoclimatic indicators. *Quaternary Science Reviews* 12: 233–248.
- Harrison SP, Yu G, and Tarasov P (1996) Late Quaternary lake-level records from northern Eurasia. *Quaternary Research* 45: 138–159.
- Hedin S (1922) *Formation of Pangong-tso, Southern Tibet*. VII, Stockholm. pp. 511–525.
- Hu DS (2001) *Study on the Salt Lake of Qarhan*, pp. 296+26 Plates (in Chinese). Changsha: Press of Hunan Normal University of China.
- Hu SY, Goddu SR, Appel E, et al. (2005) Palaeoclimatic changes over the past 1 million years derived from lacustrine sediments of Heqing basin (Yunnan), China. *Quaternary International* 136(1): 123–129.

- Huang XZ, Chen FH, Fan YX, et al. (2009) Dry late-glacial and early Holocene climate in arid central Asia indicated by lithological and palynological evidence from Bosten Lake China. *Quaternary International* 194(1–2): 19–27.
- Jiang JM, Wu JL, and Shen J (2004) Lake sediment records of climatic and environmental change in Angulino Lake. *Scientia Geographica Sinica* 24(3): 346–351 (in Chinese).
- Joussaume S, Taylor KE, Braconnot P, et al. (1999) Monsoon changes for 6000 years ago: Results of 18 simulations from the Palaeoclimate Modeling Intercomparison Project (PMIP). *Geophysical Research Letters* 26: 859–862.
- Kaufman A (1971) U-series dating of Dead Sea basin carbonates. *Geochemistry Cosmochim Acta* 35: 1269–1281.
- Kutzbach J, Gallimore R, Harrison SP, et al. (1998) Climate and biome simulation for the past 21,000 years. *Quaternary Science Reviews* 17: 473–506.
- Li BY (2000) The last greatest lakes on the Xizang (Tibetan) Plateau. *Acta Geographica Sinica* 55: 174–182 (in Chinese).
- Lin RF and Wei KQ (2000) A $\delta^{13}\text{C}$ record of the organic matter in lacustrine sediments of the core ZHJ from Lake Caohai and its palaeoclimate implications. *Geochimica* 29(4): 390–396 (in Chinese).
- Liu XQ, Dong HL, Yang XD, et al. (2009) Late Holocene forcing of the Asian winter and summer monsoon as evidenced by proxy records from the northern Qinghai–Tibetan Plateau. *Earth and Planetary Science Letters* 280(1–4): 276–284.
- Liu XQ, Herzschuh U, Shen J, et al. (2008) Holocene environmental and climatic changes inferred from Wulungu Lake in northern Xinjiang, China. *Quaternary Research* 70(3): 412–425.
- Liu Q, Liu JQ, Chen XY, et al. (2005) Stable carbon isotope record of bulk organic matter from the Sihailongwan Maar Lake, Northeast China during the past 18.5 ka. *Quaternary Sciences* 25(6): 711–721 (in Chinese).
- Ma ZX, Huang JH, Wei Y, et al. (2004) Organic carbon isotope records of the Poyang Lake sediments and their implications for the palaeoclimate during the last 8 ka. *Geochimica* 33(3): 279–285 (in Chinese).
- Mason IM, Guzikowska MAJ, Rapley CG, et al. (1994) The response of lake levels and areas to climatic change. *Climatic Change* 27: 161–197.
- Mischke S and Wünnemann B (2006) The Holocene salinity history of Bosten Lake (Xinjiang, China) inferred from ostracod species assemblages and shell chemistry: Possible palaeoclimatic implications. *Quaternary International* 154–155: 100–112.
- Nanjing Institute of Geography and Limnology (NIGL) (1989) *Environment and Sedimentology of Fault Lakes, Yunnan Province*, pp. 515+32 plates (in Chinese). Beijing: Science Press.
- Oberhänsli H and Mackay AW (2005) Introduction to Progress towards reconstructing past climate in Central Eurasia, with special emphasis on Lake Baikal. *Global and Planetary Change* 46: 1–7.
- Pachur HJ, Wünnemann B, and Zhang HC (1995) Lake Evolution in the Tengger Desert, Northwestern China During the Last 40000 years. *Quaternary Research* 44: 171–180.
- Pan AD and Chen BS (2010) *Late Quaternary Palaeoenvironments of Gahai Lake in Qaidam Basin*, pp. 126+15 plates (in Chinese). Beijing: China Meteorological Press.
- Peng JL and Li SF (2003) Ostracoda from the Holocene Limnic Laminiae in Angren Lake, Tibet. *Science Technology and Engineer* 3(4): 349–355 (in Chinese).
- Pu Y, Zhang HC, Wang YL, et al. (2011) Climatic and environmental implications from *n*-alkanes in glacially eroded lake sediments in Tibetan Plateau: An example from Ximen Co. *Chinese Science Bulletin* 56(14): 1503–1510.
- Qin BQ (ed.) (2008) *Lake Taihu, China – Dynamics and Environmental Change*, p. 339. Beijing: Springer (in Chinese).
- Qin BQ, Harrison SP, and Kutzbach J (1998) Evaluation of modelling regional water balance using lake status data: A comparison of 6 ka simulations with the NCAR CCM. *Quaternary Science Reviews* 17: 535–548.
- Qin BQ and Yu G (1998) Implications of lake level fluctuations at 6ka and 18ka in mainland Asia. *Global and Planetary Change* 18: 59–72.
- Qiu SW, Wang EP, and Wang PF (1988) Changes in shorelines of Khanka Lake and discovering of the origins of Songecha River. *Chinese Science Bulletin* 12: 937–940.
- Shen HY, Jia YL, and Wei L (2005a) Palaeoprecipitation reconstruction during the interstadial of the Last Glacial (40–22 ka BP) in Huangqihai Lake, Inner Mongolia. *Acta Sedimentologica Sinica* 23(3): 523–530 (in Chinese).
- Shen J, Liu XQ, Wang SM, et al. (2005b) Palaeoclimatic changes in the Qinghai Lake area during the last 18,000 years. *Quaternary International* 136(1): 131–140.
- Shen J, Lu HY, Wang SM, et al. (2004) A 2.8 Ma record of environmental evolution and tectonic events inferred from the Cuoe core in the middle of Tibetan Plateau. *Science in China Series D: Earth Sciences* 47(11): 1025–1034.
- Shen J, Xue B, Wu JL, et al. (2010) *Lake Sediments and Environmental Evolution*, pp. 448–473. Beijing: Science Press (in Chinese).
- Shen J, Zhang ZL, Yang LY, et al. (2008) *Nansihu Lake – Study on Environments and Resources*, pp. 284. Beijing: Seismal Press of China (in Chinese).
- Shi YF, Yu G, Liu XD, et al. (2001) Reconstruction for 30–40 ka B.P. enhanced Indian monsoon based on geological records from the Tibetan Plateau. *Palaeogeography, Palaeoclimatology, Palaeoecology* 169: 69–83.
- Stein M, Agnon A, Avraham ZB, et al. (2005) The Dead Sea as a Global Palaeo-Environmental Archive (Prospects of Scientific Deep Drill).
- Street-Perrott FA and Harrison SP (1985) Lake levels and climate reconstruction. In: Hecht AD (ed.) *Palaeoclimate Analyses and Modelling*, pp. 291–340. New York: John Wiley & Sons.
- Street-Perrott FA, Marchand DS, Roberts N, et al. (1989) Global Lake-Level Variations from 18,000 to 0 years ago, A Palaeoclimatic Analysis. U.S. DOE/ ER/ 60304-H1 TR046, *Technical Report*, pp. 63. Washington DC: U.S. Department of Energy.
- Sun QL, Wang SM, Zhou J, et al. (2009) Lake surface fluctuations since the late glaciation at Lake Daihai, North central China: A direct indicator of hydrological process response to East Asian monsoon climate. *Quaternary International* 194(1–2): 45–54.
- Takemura K, Hayashida A, Okamura M, et al. (2000) Stratigraphy of multiple piston core sediments from the last 30,000 yrs from Lake Biwa, Japan. *Journal of Palaeolimnology* 23(2): 185–199.
- Tarasov PE, Pushenko MYa, Harrison SP, et al. (1996) Lake Status Records from the Former Soviet Union and Mongolia: Documentation of the Second Version of the Database. *Palaeoclimatology Publications Series Report No. 5*, pp. 224. Boulder, USA.
- Tareq SM, Handa N, and Tanoue E (2006) A lignin phenol proxy record of mid Holocene palaeovegetation changes at Lake DaBuSu, northeast China. *Journal of Geochemical Exploration* 88(1–3): 445–449.
- Wang SM and Dou HS (1998) *Lake Documentations of China*, pp. 580. Beijing: Science Press (in Chinese).
- Wang SM, Ji L, Yang XD, et al. (1995) *Hulun Lake-Palaeolimnology Study*, pp. 125. Hefei: China Science and Technology University Press (in Chinese).
- Wang NA, Li ZL, Cheng HY, et al. (2010) Younger Dryas event recorded by the mirabilite deposition in Huahai Lake, Hexi Corridor, NW China. *Quaternary International* 1–7.
- Wu JL, Shen J, and Wang SM (2003) Characters of climate and environment of the early Holocene in Aibi Lake, Xinjiang Province. *Science in China: Series D* 33(6): 569–575.
- Yao SC, Wang XL, and Xue B (2007) Preliminary study of deposition pattern of Gucheng Lake at Jiangsu during Holocene. *Quaternary Sciences* 27(3): 365–370 (in Chinese).
- Yu J (2008) *The Quaternary Environmental Evolution of Typical Lakes in Central Tibetan Plateau*, pp. 85. MsD Thesis, Graduate University of Chinese Academy of Geological Sciences (in Chinese).
- Yu G, Gui F, Shi YF, et al. (2007) Late Marine Isotope Stage 3 palaeoclimate for East Asia: Data and modelling comparison. *Palaeogeography, Palaeoclimatology, Palaeoecology* 250: 167–183.
- Yu G and Ke X (2007) Lake Level Studies / Asia. In: Scott AE (ed.) *Encyclopedia of Quaternary Science*, 1st edn., pp. 1343–1359. Amsterdam: Elsevier.
- Yu G and Shen HD (2009) Lake water changes in response to climate change in northern China: Simulations and uncertainty analysis. *Quaternary International* 212: 44–56.
- Yu G, Xue B, Liu J, et al. (2001a) *Lake Records from China and the Palaeoclimate Dynamics*, pp. 196. Beijing: China Meteorological Press (in Chinese).
- Yu G, Harrison SP and Xue B (eds.) (2001b) Lake Status Records from China: Data Base Documentation. *Technical Report No. 4*, pp. 243. Jena: Max-Planck-Institute for Biogeochemistry of Germany.
- Yu G, Xue B, Liu J, et al. (2003) LGM lake records from China and an analysis of the climate dynamics using a modelling approach. *Global and Planetary Change* 38: 223–256.
- Zhang K, Zhao Y, Yu ZC, et al. (2010) A 2700-year high resolution pollen record of climate change from varved Sugan Lake in the Qaidam Basin, northeastern Tibetan Plateau. *Palaeogeography, Palaeoclimatology, Palaeoecology* 297(2): 290–298.
- Zheng MP, Xiang J, Wei XJ, et al. (1989) *Saline lakes on the Qinghai-Xizang (Tibet) Plateau*, pp. 431. Beijing: Scientific and Technical Publishing House (in Chinese).
- Zheng YQ, Yu G, Wang SM, et al. (2004) Simulations of Palaeoclimates over the East Asia at 6 ka and 21 ka B.P. by a Regional Climate Model. *Climate Dynamics* 23: 513–529.
- Zhu C, Ma CM, Zhang WQ, et al. (2006) Pollen record from Daijihu basin of Shennongjia and environmental changes since 15.753 ka B.P. *Quaternary Sciences* 26(5): 814–826 (in Chinese).

- Zhu LP, Peng P, Xie MP, et al. (2010) Ostracod-based environmental reconstruction over the last 8,400 years of Nam Co Lake on the Tibetan plateau. *Hydrobiologia* 648(1): 157–174.
- Zhu LP, Wang JB, Chen L, et al. (2004) 20,000-year Environmental change reflected by multidisciplinary lake sediments in Chen Co, Southern Tibet. *Acta Geographica Sinica* 59(4): 514–524 (in Chinese).
- Zhu DG, Zhao XT, Meng XG, et al. (2003) Late Pleistocene lacustrine-beach rock around the Nam Co in Xizang. *Geological Review* 49(4): 432–439 (in Chinese).
- Zong YQ, Innes JB, Wang ZH, et al. (2011) Mid-Holocene coastal hydrology and salinity changes in the east Taihu area of the lower Yangtze wetlands, China. *Quaternary Research* 76(1): 69–82.

Introduction

Australian lakes span a wide range of physiographic and climatic regions and are generally characterized by shallow depths, saline water, and discontinuous records. Coastal lagoons with continuous, spasmodic, or groundwater connections to the ocean are common but will not be considered in this article. Small fresh-water perennial lakes occur in a very limited area of the mainland around Mt. Kosciuszko, which was glaciated during Pleistocene cold phases, and are larger and more abundant in the more widespread glaciated regions of Tasmania. Most lake-level studies have focused on a variety of basins of the semiarid and arid interior, crater lakes in Quaternary volcanic regions, particularly in Western Victoria and also in northeastern Queensland, and some long-lived structural basins formed by drainage disruption due to tectonic uplift of the Eastern Highlands.

Australia is dominated by arid and semiarid environments in the interior. Higher rainfall zones are restricted to the southern, eastern, and northern margins and are seasonal with northern rainfall highly seasonal. Precipitation is derived in the north from the summer monsoon and in the south from winter-westerly frontal systems. A broad zone of weak or mixed seasonal dominance covers much of arid and semiarid Australia and extends from the northwest to the southeast (Figure 1). Precipitation can also be derived from the east. High-magnitude events recorded east of the Eastern Highlands divide are associated with stationary offshore cyclonic systems that are derived from either tropical or winter-westerly circulation. In the modern system, significant rainfall from these events rarely occurs west of the divide and does not influence arid-zone hydrology. Aridity in Australia is not due to an absolute lack of water vapor in the air masses, but rather due to a lack of a mechanism for triggering of precipitation by rising air masses due to either convective or orographic processes. Therefore, climate change that results either in changes in latitudinal position of monsoon or winter-westerly-derived precipitation or in enhanced convection can significantly affect lake levels.

Bowler (1981, 1986) examined the spatial variability and hydrologic evolution of Australian lakes and classified them according to catchment area/lake area ratios. These are related to functions of climate and runoff, and they elegantly illustrate the factors controlling the paleoclimatic sensitivity of lakes. Lakes close to the equilibrium line in this relationship have water budgets almost in balance and tend to be ephemeral, or nearly so. These lakes are the most sensitive to climate change: even small changes of climate can have a dramatic impact on sedimentary processes and hence the preserved sediment records. On the other hand, lakes at the wet end of the spectrum (lakes with extremely large catchments and those occurring in high rainfall zones or both) have strongly positive water budgets and will only respond to extremes of climate change, if at

all. Bowler's classification recognized a hydrologic continuum from wet to dry that operates both spatially, as climate varies across the continent, and through time, as climate has varied during Quaternary glacial–interglacial cycles. Under full fresh-water conditions, wind-induced wave erosion and sediment transport produce smooth shorelines and quartz-rich beaches and foredunes on the downwind margins; sediments are terrigenous and rich in biological carbonates. With negative hydrological budgets, groundwater and salinity dominate. Salts (particularly gypsum, halite, and sodium sulfates) crystallize displacively in the lake sediments, disrupting them and allowing deflation of sand-sized sediment aggregates and salts by the dominant wind. Clay-rich dunes or lunettes form on the downwind shoreline, and finer dust material is spread further down basin (Bowler, 1973, 1983; Magee, 1991). At the extreme dry end of the spectrum, surface-water cover is extremely rare and groundwater processes dominate, producing systems with highly irregular outlines and multiple islands. These groundwater playas are also insensitive to climate change because, lacking the necessary amplification of a large catchment area, evaporation always exceeds inflow. Thus, they will not change their state even in response to a considerable wetting of the climate.

The features described above mean that Australian lakes do not have long, continuous high-resolution records, but they are nevertheless very often highly sensitive to climate change. Extraction of paleoclimatic records rarely requires use of complex proxies dependent on sophisticated, expensive, and time-consuming analyses. Sequences from many Australian lakes have so much variation in their sedimentary record that a simple log of deep- or shallow-water clastics, evaporites, macrofossils, color, soils, deflation events, etc., can provide a rich paleoclimatic record. Wide variations in sedimentology and biota also undoubtedly extend the opportunities for applying different dating techniques and obtaining a reliable chronology.

Eastern Highland Tectonic Basins

Lake George is a closed basin in a fault angle depression astride the Eastern Highlands divide. It lies at 35° S at an elevation of 670 m about 100 km from the coast (Figure 1). The modern lake has an excellent historic record back to 1820 of variations from dry to 7.3 m water depth, which correlates with climate variability. Shoreline evidence has been studied by Galloway (1965) and Coventry (1976) and central basin sediments and pollen by Singh et al. (1981). Paleomagnetic reversal dating has established a firm chronology for the basin fill back to about 3 My and an extrapolated age for the origin of sedimentation as middle Miocene or older (Singh et al., 1981). Lacustrine deposition has not been continuous, though drier episodes are marked by pedogenesis rather than the deflation

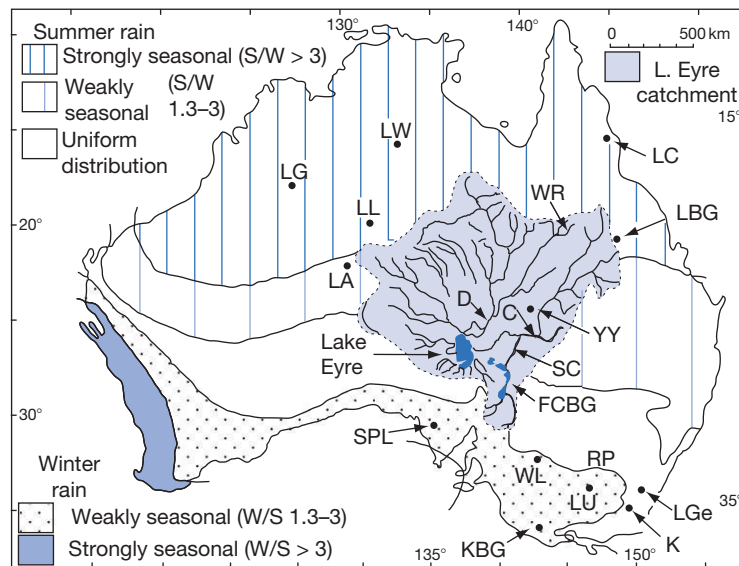


Figure 1 Australian rainfall seasonality and location of sites mentioned in text. Modified from Magee JW, Miller GH, Spooner NA, and Questiaux D (2004) Continuous 150 k.y. monsoon record from Lake Eyre, Australia: Insolation-forcing implications and unexpected Holocene failure. *Geology* 32(10): 885–888. S/W = summer/winter rainfall ratio; W/S = winter/summer rainfall ratio. Rainfall seasonality data modified from Gentili J (1986) Climate. In: Jeans DN (ed.) *The Natural Environment – Vol. 1: Australia, A Geography*, pp. 14–48. Sydney: Sydney University Press. Site location key: LG, Lake Gregory; LW, Lake Woods; LL, Lake Lewis; LA, Lake Amadeus; D, Diamantina River; C, Cooper Creek; WR, Western River; LC, Lynches Crater; LBG, Lakes Buchanan and Gallilee; YY, Lake Yamma Yamma; SC, Strzelecki Creek; FCBG, Lakes Frome, Callabonna, Blanche, and Gregory; WL, Willandra Lakes; RP, Riverine Plain; KBG, Lakes Keilambete, Bullenmerri and Gnotuk; LU, Lake Urana; K, Mt. Kosciuszko; LGe, Lake George.

of lake sediment that characterizes arid-zone basins. Deep-water lacustrine conditions extend from the base of the cored sequence at 72 m (between 4.1 and 7 Ma) to a major discontinuity at 51.5 m, where evidence of deep weathering and slope mantle deposition indicate a long hiatus in lake deposition. Lacustrine conditions returned to the basin resulting in increased depositional rates above 21.5 m (0.94 Ma) and dominated the Brunhes magnetic chron (last 790 ka) with regular cycles over the past 400 ka. The late Quaternary shoreline evidence indicates overflow (~36 m water depth) between 27 and 21 ka BP and subsequent, lower highstands (between overflow and modern levels) at 15, 8, 5, and 3 ka BP. However, this radiocarbon chronology is based on early conventional dates and, with an abundance of nearshore and beach sandy sediments available, it should be revised by a combination of optically stimulated luminescence (OSL) and accelerator mass spectrometry ^{14}C dates that target reliable microfossils with rigorous pretreatment.

Western Victorian Volcanic Lakes

The newer volcanic region of Western Victoria has widespread Quaternary eruptive centers and numerous lakes that formed predominantly as crater lakes and maars. These lakes vary in age, depth, and salinity, and are mostly closed basins with respect to surface-water hydrology in the modern setting. Water depth and salinity have varied through time in response to Quaternary climate changes, and lake levels and past climates have been deduced from (1) shoreline and lacustrine sediments and (2) paleosalinity estimates based on various biological and chemical proxies. Many basins have been studied, and the

best lake-level records come from Lakes Bullenmerri, Gnotuk, and Keilambete, particularly the last (Figure 1). Here, close agreement exists between shoreline and sedimentary facies (Bowler, 1981), pollen analysis (Dodson, 1974), ostracod species abundance (De Deckker, 1982), and ostracod shell trace element and isotope geochemistry (Chivas et al., 1993). Keilambete (Figure 2) was dry prior to about 10 ka BP with a soil developed on ancient lacustrine sediments. Filling began about 10 ka BP, and more sustained rising levels after 8.5 ka BP resulted in overflow (at slightly <40 m water depth) from about 7.5 to 5.5 ka BP. Lake levels then declined and reached a minimum between 3.2 and 3.6 ka BP, subsequently oscillating at generally low levels until rising again at about 2 ka BP. A relatively constant water depth was maintained at about 25 m until a sustained drop recorded historically since about AD 1850. The modern lake lies at depths as low as anything seen since filling began at 10 ka BP.

Bowler (1981) recognized that the Western Victorian crater lakes were climatically very sensitive, and with their simplified hydrology they had the potential to be virtual rain-gauge records – if the P/E relationship could be unraveled. Jones et al. (2001) constructed a water-balance model for Lakes Bullenmerri, Gnotuk, and Keilambete, calibrated by detailed historical lake-level records and historical climate records (rendered reliable by extensive investigative reduction of dubious data). This water-balance model allowed a detailed record of P/E for the record of the three lakes (Figure 2), since 16 ka BP for Bullenmerri, and since 10 ka BP for Keilambete and Gnotuk (Jones et al., 1998).

Coring techniques used in the crater lakes, suitable for the soft sediments deposited in the postglacial high-water events, have not been able to penetrate the compacted/cemented soil

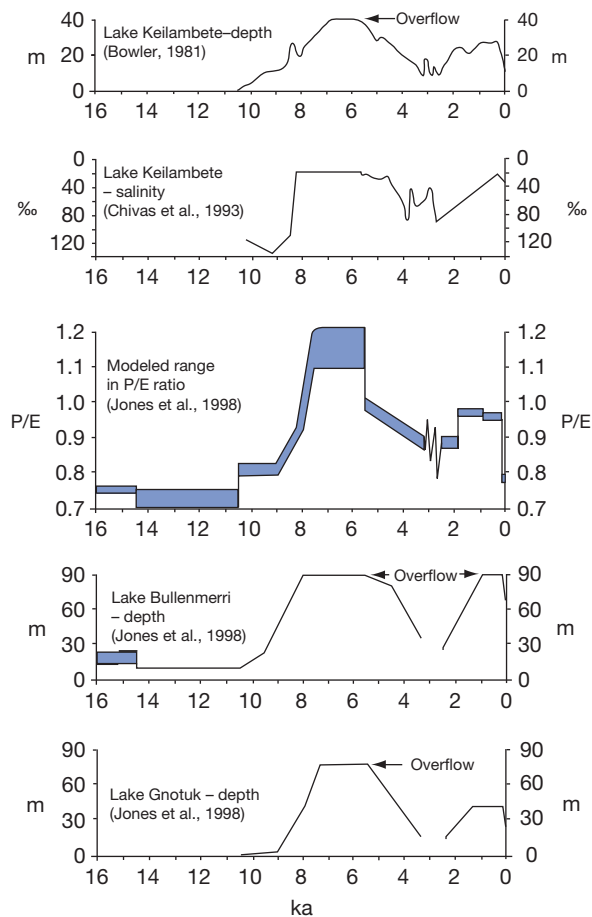


Figure 2 Lake-level records from Lakes Keilambete, Bullenmerri, and Gnotuk, volcanic crater lakes in western Victoria. Modified from Jones RN, Bowler JM, and McMahon TA (1998) A high-resolution Holocene record of P/E ratio from closed lakes, western Victoria, Australia. *Journal of Hydrology* 246: 159–180.

horizons formed during the Last Glacial Maximum (LGM) dry period. Many of the craters extend back through a number of glacial cycles and have the potential for high-quality long, though discontinuous, lake-level records. Many of the older craters are filled with sediment and maintain swamp rather than lacustrine environments. Cores from these craters, some with long records, have been analyzed for pollen but not studied specifically for lake levels (D'Costa and Kershaw, 1997). Similarly, a number of crater lakes and sediment-filled swamp craters in north eastern Queensland have been cored but only analyzed for pollen sequences (e.g., Kershaw, 1986; Turney et al., 2001).

Arid-Zone Lakes

Australia's arid desert region is a broad zone of moderately low, highly variable rainfall with mixed seasonal dominance, with the semiarid margins extending into the monsoonal summer-rainfall zone to the north and the westerly winter-rainfall zone to the south. Flood-out, overflow or terminal lakes, dominated by surface-water hydrology, are associated with coordinated

drainage of the very large Murray–Darling and Lake Eyre Basins of the southeast and northeast, respectively, with large systems such as Lake Woods (NT) and Lake Gregory (WA), and with smaller systems such as Lakes Buchanan and Gallilee (Qld; Figure 1).

Murray–Darling Basin

Coordinated external drainage in the arid zone is restricted to the Murray–Darling Basin, where moisture sources extend from winter westerlies for the southern Murray tributaries in Victoria to summer tropical systems for the upper Darling tributaries in Queensland. Some of the tributaries in this basin can be influenced by easterly derived moisture, and many rise in the highlands of Victoria and southern and northern NSW, which are subject to complexities of snow pack and periglacial conditions, much expanded in the past (Barrows et al., 2004). Much of the Murray–Darling system is characterized by very low-gradient fluvial plains with abundant ephemeral lakes, of variable size and origin. Hydrologically, most of these are flood-out lakes; they lie adjacent to fluvial systems with connecting channels and are filled by passing floods then partially emptied as the flood peak passes and flow reverses in the feeder channel. Evaporation then lowers lake level until the next flood replenishes the lake, and deflation of the bed during dry periods prevents filling by sediment. Flood-out lakes can thus contain long but discontinuous lake-level records that reflect the flood history of the adjacent fluvial system. In contrast to the flood-out lakes, the Willandra Lakes are an overflow system of now-dry lakes fed by a distributary of the Lachlan River, the Willandra Billabong Creek, which rises in the Eastern Highlands north of Canberra. In common with most Murray–Darling Basin lakes the Willandra Lakes are characterized by lunettes, lee-side eolian accumulations derived from the lake basins. These are high-quality archives of the hydrologic state of the lake, with quartz foredune sediments representing lake-full conditions, gypseous, clay pellet-rich sediments reflecting saline groundwater domination, and soils forming when deposition is not occurring. When full, the lakes of the system provided rich resources for human occupation, and archaeological evidence is abundant in the lunette depositional sites adjacent to the lakes.

Relatively recent severe erosion of some of the Willandra Lakes lunettes has exposed their internal stratigraphy, and long-term sedimentary and chronological studies (summarized in Bowler (1998) and Bowler et al. (2003)) have elucidated a complex history of wetting and drying cycles between 60 and 15 ka (Figure 3) and provided an environmental and chronological context for the associated archaeology. Older phases of lake activity are preserved as mostly buried sedimentary units on lake floors and as lunettes capped by well-developed soils. These ancient wet phases are generally poorly dated. Some luminescence dates from the youngest of them, the Goolgol lunette unit, indicate an age in the range 100–130 ka. By 60 ka, at the latest, the lakes had refilled and the system was full, fresh, and overflowing; quartz-rich beach and foredune lunette sediments of the Lower Mungo unit were deposited throughout the system at this time. At about 42 ka, a significant drying episode began, marked by a transition to Upper Mungo

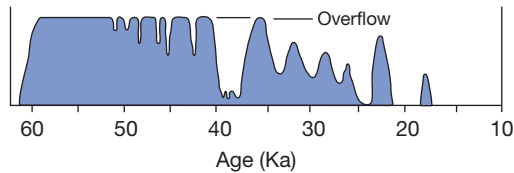


Figure 3 Lake-level record from the Willandra Lakes, Murray Basin, southwestern New South Wales. Modified from Bowler JM (1998) Willandra Lakes revisited: Environmental framework for human occupation. *Archaeology in Oceania* 33: 120–155; Bowler JM, Johnston H, Olley JM, et al. (2003) New ages for human occupation and climatic change at Lake Mungo, Australia. *Nature* 421: 837–840.

unit sediments, which are rich in clay pellets derived via deflation that is controlled by saline groundwater. The Mungo unit is followed by the Arumpo and Zanci units, which also record wetter, surface-water-dominated conditions followed by lake-floor deflation under drier conditions. The trend is toward increasingly effective dry phases and culminates in a thick sequence of gypseous pelletal clay deposited during the Zanci unit saline groundwater-deflation stage (22–19 ka). Following this intense dry phase at the glacial maximum, there is a brief return to wetter conditions that brought water to the northern lakes of the system (Mulurulu and Garpung) and deposited the Mulurulu unit at about 18 ka. Since then the lakes have remained dry, with regionally low water tables, and soils have formed over the landscape. Page et al. (1996) report a thermoluminescence (TL) chronology for a stratigraphic sequence of fluvial units associated with paleochannels of the Murrumbidgee River on the Riverine Plain, south of the Lachlan system. There is general accord with the Willandra Lakes sequence in the chronology of the Yanco, Gum Creek, and Kerarbury units in the 50–15 ka range, but an earlier phase, the Coleambally Unit, dated to late marine isotope stage (MIS) 5 (80–105 ka) has no known equivalent units in the Willandra Lakes.

Lake Eyre Basin

Until luminescence dating became available, little was known about the Quaternary paleohydrology of fluvial and lacustrine systems that are northern monsoon-fed with coordinated internal drainage. The lacustrine records from Lakes Eyre, Woods, and Gregory, and the fluvial records from the Lake Eyre catchment are now dated back to 2–300 ka, with the last 130 ka relatively well constrained. All these records indicate wetter conditions associated with the warm interglacial phases and drier conditions associated with the cold glacial phases.

Lake Eyre has a complex suite of deep-water, nearshore, beach, lunette, and inflowing stream sediments that provide the most complete and best dated monsoon-derived lake-level record over the past 150 ka (Figure 4; Magee and Miller, 1998; Magee et al., 1995, 2004). Prior to 130 ka, in MIS 6, the lake was dry, and deflation had lowered the basin floor to levels lower than today's playa floor and lower than that of the LGM. Early in MIS 5, at around 130 ka, there was an abrupt change to lacustrine conditions and establishment of a salinity-stratified perennial water body, 25 m deep with a shoreline at about 10 m Australian height datum (AHD). The lake was fed by enhanced flow from the Diamantina and Cooper systems.

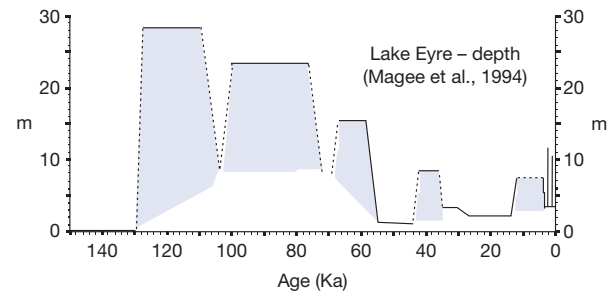


Figure 4 Lake-level record from Lake Eyre, northern South Australia. Modified from Magee JW, Miller GH, Spooner NA, and Questiaux D (2004) Continuous 150 k.y. monsoon record from Lake Eyre, Australia: Insolation-forcing implications and unexpected Holocene failure. *Geology* 32(10): 885–888.

This included overflow from a combined Frome–Callabonna–Blanche–Gregory megalake at 18 m AHD, fed via the Strzelecki Creek, a distributary from the Cooper (DeVogel et al., 2004). While the hydrological response to climate change in such an extremely large and complex catchment is unlikely to be linear, the conditions of early MIS 5 must indicate an enhanced Australian monsoon. After a brief drying, lower-level perennial lacustrine conditions returned later in MIS 5, and there was considerable variation in water level and salinity. Drying and deflation of the basin occurred again, followed at about 65 ka by the last, relatively short, high-level lacustrine conditions, which constructed a beach at –3 m AHD. Between 60 and 50 ka, the lowering of water table associated with the transition from the previous high-lake events of stage 5 resulted in a major deflation event that effectively excavated the modern Lake Eyre playa basin. A low-level (–10 m AHD), poorly dated lacustrine event occurred in stage 3 and was followed by prolonged dry conditions with episodic deflation from about 35 to 14 ka. At about 12 ka low-level, semiperennial lacustrine conditions (below –10 m AHD) returned to the basin, indicating renewed enhancement of the monsoon, but less effectively than in the previous interglacial (early MIS 5). The modern ephemerally flooded playa conditions were established at about 3–4 ka, indicating a change to a drier monsoon regime marked by more extreme events.

Older deposits sedimentologically and stratigraphically similar to those of the last glacial cycle occur east and northeast of the current playa and have luminescence dates indicating that similar monsoon enhancement occurred in MIS 7 and probably earlier interglacials. However, the migration of the depocenter through time and the destruction of pre-MIS 5 shorelines and beach ridges preclude the possibility of comparing the lake area through time and consequently of assessing long-term trends in monsoon strength.

The fluvial sequences of the middle and upper reaches of the Lake Eyre Basin have been extensively studied and dated by luminescence, mostly TL by Nanson et al. (1991, 1992). They have a record in accord with that from Lake Eyre, showing enhanced activity in interglacials. Nanson et al. (1992) report the greatest number of MIS 5 dates centered on about 109 ± 4 ka. They interpret this as indicating that the peak of fluvial activity occurred somewhat later in the previous interglacial than indicated at Lake Eyre. An alternative explanation

is that the fluvial dates are biased toward the end of a phase of activity, as the river continually reworks sediments within the valley tract. In good agreement with the Lake Eyre record, from an upstream Diamantina tributary the Western River, Nanson et al. (1992) reports a minor enhanced regime in stage 3, and a Holocene enhancement much less significant than the last interglacial. The longer record from the rivers provides strong indications of a trend of decreasing monsoon strength through successive interglacials. Maroulis (2000) reports that the middle Cooper had reworked virtually the whole of its valley tract in MIS 7, less than half in MIS 5, and very little in the Holocene, when suspended-load streams contrasted strongly with the sand-transporting systems of earlier interglacials. Flood-out lakes are commonly associated with many of the Lake Eyre Basin rivers, including the very large Lake Yamma Yamma in southwest Queensland. These are sparsely studied but should contain lake-level records reflecting the stream flood histories.

Other Monsoon-Fed Terminal Lakes

Lake Woods and Lake Gregory are both moderately large modern lakes in Northern Australia, which vary in accord with the strength of monsoon events, particularly cyclonic activity in their catchment streams, Newcastle Waters and Sturt Creek, respectively. Both modern lakes lie within the basin of much larger megalakes. Large sand-dominated lunette beach ridges occur on the downwind western margins, and longitudinal dunes are developed on the megalake floors inside the outer shorelines. The Lake Woods outer shoreline is a single ridge with complex internal stratigraphy for most of its length, but it splits to multiple crests at its northern extremity. Bowler et al. (1998) report luminescence dates indicating a high-level lacustrine event early in MIS 5, a lower-level inset event in MIS 3, and a much lower lake, similar to the modern water body, in the Holocene. They also indicate that an earlier high-level MIS 7 lacustrine phase occurred, but the stratigraphic and chronologic definition is not sufficient to identify a longer-term trend in lake area and monsoon strength. At Lake Gregory, a series of shoreline ridges between the large outer shoreline and the modern water bodies suggests a successive contraction of the size of the megalake through time. Bowler et al. (2001) report TL dates from shoreline ridges and longitudinal dunes on the megalake floor, which indicate an age of about 300 ka for the outer shoreline, 200 ka (MIS 7) for an intermediate shoreline, and MIS 5 ages for the shoreline just outside the modern water bodies. This suggests a clear long-term waning trend in megalake size and consequently in monsoon strength. However, the relatively poor quality of the TL chronology, as indicated by the extremely low precision of the dates, suggests caution in interpreting these results. Recent redating by OSL (Miller and Magee, unpublished data) of the same shoreline ridge site dated to 200 ka by TL, indicates a much younger age.

Groundwater Lake/Playa Systems

Much of the arid zone, particularly in central and western parts of the continent, is characterized by uncoordinated drainage

and numerous large and small playa lakes, which are mostly groundwater-fed systems often occupying now-defunct former valleys and drainage lines of Tertiary or earlier age. Many of these systems lack effective catchments and never act as surface-water lakes even in wetter climate phases. These systems, termed boinkas by Macumber (1991), generally do not provide lake-level records, but where wetter phases result in enhanced groundwater inflow, deposits that are indicative of wetter conditions may occur, such as the gypsum-rich shoreline marginal dunes of Lake Amadeus (Chen et al., 1991). Some of the groundwater-dominated playas have presently defunct catchment drainage and may have sedimentary sequences and shorelines indicating former surface-water phases; these have the potential to provide lake-level records.

Few of the groundwater playa system records have been analyzed. In Central Australia, the large groundwater-dominated playas Amadeus and Lewis both lie within the zone of mixed rainfall seasonal dominance, but their hydrologies probably chiefly reflect a monsoon signal, particularly that of Lake Lewis, which lies north of the McDonnell Ranges. Both playa sequences have basal clay-rich units indicating much wetter lacustrine conditions; these are constrained to the lower Quaternary by paleomagnetic dating (Chen et al., 1993; English et al., 2001). A later transition to playa environments is indicated by evaporites and deflation events in both basins, but the chronology of these events is poor at Amadeus due to lack of luminescence dating and at Lake Lewis because of high-dose rates that lead to saturated dates.

Paleoclimatic Summary

Bowler et al. (1998) examined the paleoclimatic implications of the enlarged Lake Woods using the P/E and runoff efficiency relationship established earlier (Bowler, 1981, 1986), and they concluded that enhanced monsoon precipitation was required to explain the lake enlargement. Chappell (1991) also considered the hydroclimatology of the megalakes and concluded that the enlarged lakes could not be explained by reduced evaporation due to temperature reduction. Rather, enhanced precipitation or runoff efficiency or a combination of both was required. He pointed to empirical evidence of catchment runoff and rainfall gauging from the modern northern Australian Daly River system, which demonstrates significant increase in runoff efficiency as precipitation increases. Chappell's (1991) review of paleoclimate records in eastern and central Australia was prepared just before the luminescence chronologies became available. At that time, the available evidence pointed to the central and northern Australian megalakes being coincident with early stage 3 lacustrine events in the Murray Basin. Chappell pointed out the conundrum of megalake phases requiring enhanced precipitation/runoff at a time when sea level and temperatures were lower, and when pollen records from Lynch's Crater in the northeast (Kershaw, 1986) suggested somewhat lower precipitation than in the Holocene, when there were no megalakes. Subsequent luminescent dating of the megalakes to the last interglaciation has essentially resolved this issue.

Monsoon-fed systems clearly indicate a strong enhancement of the monsoon during warm interglacial periods, a

weaker enhancement in interstadials, and inactivity during cold glacial phases. There is evidence from the rivers of the Lake Eyre Basin of a long-term waning trend in monsoon strength through successive interglaciations, but evidence for this trend is equivocal or lacking at other sites. All records indicate that during the Holocene monsoon enhancement is much less effective than during the previous interglaciation. It is not clear if this relative failure of the Holocene monsoon is anomalous or part of a long-term trend.

On the southern desert margin, the best winter-westerly records come from the lakes and rivers of the Murray Basin. The winter-westerly fed fluvial systems show an MIS 5 enhancement that appears to be later than that of the MIS 5 monsoon systems. Both lacustrine and fluvial winter-westerly fed systems show strong enhancement in MIS 3. The Murray Basin sequences are the best available records for winter-westerly influenced arid-zone hydrology, but they undoubtedly are subject to other influences. The catchments of the Lachlan, and to a lesser extent the Murrumbidgee, can both be influenced by summer rainfall and perhaps by an easterly derived component. Both catchments, especially the Murrumbidgee, are at relatively high altitude and are likely to have experienced periglacial and snow pack conditions during cold glacial stages, which must have enhanced seasonality of flow regimes. It is difficult to interpret a clear precipitation signal from these systems, especially the streams closer to the highland sources. This complexity of catchment processes is a probable explanation for the apparently anomalous high LGM lake level at Lake Urana (Page et al., 1994; Figure 1) when compared to the Willandra record. Ideally, future investigation of winter-westerly precipitation should target lakes at the desert margins of Eyre Peninsula or the South West of Western Australia, where no other precipitation sources exist and other catchment complexities are minimized. Preliminary results from the Scrubby Peak Lake, northern Eyre Peninsula (Magee et al., unpublished data) indicate an MIS 5 wet phase (100–110 ka) and LGM deflation, but no MIS 3 wet phase, which is difficult to reconcile with the Murray Basin records.

Widespread aridity, dune activation, and low or dry lakes are associated with the LGM. Evidence from Northern Australia (Wyrwoll and Miller, 2001) suggests reactivation of the monsoon by 14 ka, and by 10–12 ka, lakes across the continent were rising. Central and northern Australian lakes existed at relatively low levels before falling to modern dry or ephemeral levels from about 3 ka. The highly sensitive Western Victorian volcanic lakes demonstrate a more complex Holocene record with a significant maximum, when many overflowed, centered on about 6 ka, and lower levels from about 3 ka. Since just prior to the arrival of Europeans, there has been a significant lowering related to a shift in the climate regime.

In summary, there are indications that the two climatic systems that influence Australian precipitation, the monsoon and the winter westerlies, have both been enhanced and contracted during different parts of the glacial–interglacial cycles and that these changes may have been out of phase, with westerly enhancement occurring later than the monsoon in MIS 5 and being much stronger than the monsoon in MIS 3. Holocene monsoon enhancement is much less than during the previous interglaciation, and it is not yet certain if this is anomalous or part of a longer-term trend. If anomalous,

it implies a change in boundary conditions, and a possible explanation is that the arrival of humans (50 ± 5 ka) and consequent changed fire regime, caused sufficient ecological change in monsoon catchments to cross a threshold in the hydrological response (Magee et al., 2004; Miller et al., 2005). To investigate these questions advances in a number of areas are required: stratigraphic control of lacustrine and fluvial sequences, accuracy and precision of chronologies, hydrologic modeling of large complex nonlinear catchments, recognition of catchment response thresholds, our understanding of ecophysiological vegetation and soil-moisture factors, and our understanding of landscape evolution and the role played by inheritance and antecedent conditions. Important questions to be resolved include the significance of a long-term drying trend in monsoon lake records over the past 300 ka and the resolution of unequivocal winter-westerly records through stage 5. This latter question would be addressed by long records in lakes of southwestern Australia and extended records below the LGM paleosols of Western Victorian crater lakes.

See also: Carbonate Stable Isotopes: Lake Sediments. Lake Level Studies: Africa during the Late Quaternary; Asia; Latin America; Modeling; North America; Overview; West-Central Europe. Luminescence Dating: Electron Spin Resonance Dating; Optical Dating; Thermoluminescence. Pollen Records, Late Pleistocene: Australasia.

References

- Barrows TT, Stone JO, and Fifield LK (2004) Exposure ages for Pleistocene periglacial deposits in Australia. *Quaternary Science Reviews* 23: 697–708.
- Bowler JM (1973) Clay dunes: Their occurrence, formation and environmental significance. *Earth-Science Reviews* 9: 315–338.
- Bowler JM (1981) Australian salt lakes: A palaeohydrologic approach. In: Williams WD (ed.) *Developments in Hydrobiology 5: Salt Lakes*, pp. 431–444. The Hague: Junk Publishers.
- Bowler JM (1983) Lunettes as indices of hydrologic change: A review of Australian evidence. *Proceedings of the Royal Society of Victoria* 95: 147–168.
- Bowler JM (1986) Spatial variability and hydrologic evolution of Australian lake basins: Analogue for Pleistocene hydrologic change and evaporite formation. *Palaeogeography, Palaeoclimatology, Palaeoecology* 54: 21–42.
- Bowler JM (1998) Willandra Lakes revisited: Environmental framework for human occupation. *Archaeology in Oceania* 33: 120–155.
- Bowler JM, Duller GAT, Perret N, Prescott J, and Wyrwoll K-H (1998) Hydrologic changes in monsoonal climates of the last glacial cycle: Stratigraphy and luminescence dating of Lake Woods, N.T., Australia. *Palaeoclimates* 3: 179–207.
- Bowler JM, Johnston H, Olley JM, et al. (2003) New ages for human occupation and climatic change at Lake Mungo, Australia. *Nature* 421: 837–840.
- Bowler JM, Wyrwoll K-H, and Lu Y (2001) Variations of the northwest Australian monsoon over the last 300,000 years: The paleohydrological record of the Gregory (Mulan) Lakes System. *Quaternary International* 83–85: 63–80.
- Chappell JMA (1991) Late Quaternary environmental changes in eastern and Central Australia, and their climatic interpretation. *Quaternary Science Reviews* 10: 377–390.
- Chen XY, Bowler JM, and Magee JW (1991) Aeolian landscapes in central Australia: Gypseous and quartz dune environments from Lake Amadeus. *Sedimentology* 38: 519–538.
- Chen XY, Bowler JM, and Magee JW (1993) Late Cenozoic stratigraphy and hydrologic history of Lake Amadeus, a central Australian playa. *Australian Journal of Earth Sciences* 40: 1–14.
- Chivas AR, De Deckker P, Cali JA, Chapman A, Kiss E, and Shelley JMG (1993) Coupled stable-isotope and trace-element measurements of lacustrine carbonates as palaeoclimatic indicators. In: Swart PK (ed.) *Geophysical Monograph: Climate Change in Continental Isotopic Records*, vol. 78. pp. 113–121. Washington: American Geophysical Union.

- Coventry RJ (1976) Abandoned shorelines and the Late Quaternary history of Lake George, New South Wales. *Journal of the Geological Society of Australia* 23: 249–273.
- D'Costa D and Kershaw AP (1997) An expanded recent pollen database from South-eastern Australia and its potential for refinement of palaeoclimatic estimates. *Australian Journal of Botany* 45: 583–605.
- De Deckker P (1982) Holocene ostracods, other invertebrates and fish remains from cores of four maar lakes in southeastern Australia. *Proceedings of the Royal Society of Victoria* 94: 183–219.
- DeVogel SB, Magee JW, Manley WF, and Miller GH (2004) A GIS-based reconstruction of late Quaternary paleohydrology: Lake Eyre, arid central Australia. *Palaeogeography, Palaeoclimatology, Palaeoecology* 204: 1–13.
- Dodson JR (1974) Vegetation and climate history near Lake Keilambete, western Victoria. *Australian Journal of Botany* 22: 709–717.
- Engliish P, Spooner NA, Chappell JMA, Questiaux DG, and Hill NG (2001) Lake Lewis basin central Australia: Environmental evolution and OSL chronology. *Quaternary International* 83–85: 81–102.
- Galloway RW (1965) Late Quaternary climates in Australia. *Journal of Geology* 73: 603–618.
- Gentili J (1986) Climate. In: Jeans DN (ed.) *The Natural Environment – Vol. 1: Australia, A Geography*, pp. 14–48. Sydney: Sydney University Press.
- Jones RN, Bowler JM, and McMahon TA (1998) A high-resolution Holocene record of P/E ratio from closed lakes, western Victoria, Australia. *Journal of Hydrology* 246: 159–180.
- Jones RN, McMahon TA, and Bowler JM (2001) Modelling historical lake levels and recent climate change at three closed lakes, western Victoria, Australia. *Journal of Hydrology* 246: 159–180.
- Kershaw AP (1986) The last two glacial-interglacial cycles from northeastern Australia: Implications for climate change and Aboriginal burning. *Nature* 322: 47–49.
- Macumber PG (1991) *Interaction Between Ground Water and Surface Systems in Northern Victoria*, 345 pp. Victoria: Department of Conservation and Environment.
- Magee JW (1991) Late Quaternary lacustrine, groundwater, aeolian and pedogenic gypsum in the Prungle Lakes, southwestern New South Wales, Australia. *Palaeogeography, Palaeoclimatology, Palaeoecology* 84: 3–42.
- Magee JW, Bowler JM, Miller GH, and Williams DLG (1995) Stratigraphy, sedimentology, chronology and palaeohydrology of Quaternary lacustrine deposits at Madigan Gulf, Lake Eyre, South Australia. *Palaeogeography, Palaeoclimatology, Palaeoecology* 113: 3–42.
- Magee JW and Miller GH (1998) Lake Eyre palaeohydrology from 60 ka to the present: Beach ridges and glacial maximum aridity. *Palaeogeography, Palaeoclimatology, Palaeoecology* 144: 307–329.
- Magee JW, Miller GH, Spooner NA, and Questiaux D (2004) Continuous 150 k.y. monsoon record from Lake Eyre, Australia: Insolation-forcing implications and unexpected Holocene failure. *Geology* 32(10): 885–888.
- Maroulis J (2000) *Stratigraphy and Mid to Late Quaternary Chronology of the Cooper Creek Floodplain*, Southwest Queensland, Australia. PhD Thesis, Wollongong: University of Wollongong.
- Miller GH, Mangan J, Pollard D, Thompson S, Felzer B, and Magee JW (2005) Sensitivity of the Australian Monsoon to insolation and vegetation: Implications for human impact on continental moisture balance. *Geology* 33(1): 65–68.
- Nanson GC, Price DM, and Short SA (1992) Wetting and drying of Australia over the past 300 ka. *Geology* 20: 791–794.
- Nanson GC, Price DM, Short SA, Young RW, and Jones BG (1991) Comparative uranium–thorium dating and thermoluminescence dating of weathered Quaternary alluvium in the tropics of northern Australia. *Quaternary Research* 35: 347–366.
- Page KJ, Dare-Edwards AJ, Nanson GC, and Price DM (1994) Late Quaternary evolution of Lake Urana, southeastern Australia. *Journal of Quaternary Science* 9: 47–57.
- Page KJ, Nanson GC, and Price D (1996) Chronology of Murrumbidgee River palaeochannels on the Riverine Plain, southeastern Australia. *Journal of Quaternary Science* 11: 311–326.
- Singh G, Opdyke ND, and Bowler JM (1981) Late Cainozoic stratigraphy, palaeomagnetic chronology, and vegetational history from Lake George, NSW. *Journal of the Geological Society of Australia* 28: 435–452.
- Turney CSM, Kershaw AP, Moss P, et al. (2001) Redating the onset of burning at Lynch's Crater (North Queensland): Implications for human settlement in Australia. *Journal of Quaternary Science* 16: 767–771.
- Wyrwoll K-H and Miller GH (2001) Initiation of the Australian summer monsoon 14,000 years ago. *Quaternary International* 83–85: 119–128.

This article deals with lake-level studies from South and Central America, as far north as Mexico (i.e., Latin America). This vast area extends from the mid-latitudes of the Southern Hemisphere (more than 50° S) to the subtropics of the Northern Hemisphere (about 30° N; [Figure 1](#)). At its core lie the Neotropics, which today are dominated by summer rainfall regimes associated with the seasonal migration of the intertropical convergence zone (ITCZ) and monsoon-type circulations such as the South American summer monsoon (SASM) and the Mexican (or North American) monsoon. At higher latitudes, the influences of the subtropical high and the midlatitude Westerlies both become important. The region is also strongly influenced by the El Niño-Southern Oscillation (ENSO) driven by changes in sea surface temperatures and pressures in the tropical Pacific Ocean. Climate change studies have focused on identifying the importance of insolation forcing (millennial timescales), Northern Hemisphere conditions (ice sheets, temperatures in the North Atlantic which might lead to synchronous climate change), and the tropical Pacific (ENSO). The impact of such forcings on the location of the ITCZ, the strength of monsoons, and the position of the Westerlies all underpin the interpretation of lake-level records in paleoclimatic terms. There has also been increasing realization that climatic change and cultural change may be interlinked.

Background

Much of the early work in South America focused on the history of the Amazon rain forest, in particular whether the diversity of species in the basin reflected climatic stability or variability. A key to this was to understand the nature of climatic change during the Last Glacial Maximum (LGM; 21 000 cal years BP or 18 000 ¹⁴C years BP) and whether this change had fundamentally altered the distribution of humid tropical forest. All the early studies of lake sediment sequences in tropical South America were palynological, using vegetation change as an indicator of climate. As a result of the dominance of pollen analysis, the reconstruction of lake-level change as an indicator of changes in water balance (precipitation–evaporation) was, at best, a secondary goal and often not considered directly. Early interpretations of a dry glacial Amazon were based on geomorphological evidence, but [Liu and Colinvaux \(1985\)](#) published the first pollen record from the Amazon (Mera, Ecuador) with a full glacial radiocarbon age. These authors suggested that the glacial Amazon was significantly cooler and drier, but not arid. In contrast, the Dutch group of Thomas Van der Hammen and his collaborators, working in Colombia, also used pollen analysis to suggest that the Amazon was significantly drier and that much of the tropical forest had been replaced by savannas (e.g., [Van de Hammen, 1974](#)). Palynology has continued to dominate studies of the Amazon basin and the surrounding dry forest areas,

although now more often complemented by other proxies such as pigment analysis, carbon isotopes ($\delta^{13}\text{C}$), and diatom analysis, although siliceous microfossils are often poorly preserved in Amazonian sediments. Palynology has also dominated paleoenvironmental studies in Central America, where a number of Holocene records have been interpreted as reflecting human impact rather than climatic change.

More traditional lake-level studies, based on paleolimnological methods, have become more of a focus since the 1980s. It is also worth noting that the increasing use of core loggers and scanners is enabling data collection at unprecedented resolution, allowing the identification of high temporal-scale variability and cyclicity in records based on large datasets. Across Latin America as a whole, however, intensive lake-level studies have been limited to relatively few areas such as the Bolivian and Chilean Altiplanos, the Yucatan peninsula (Mexico and Guatemala), and the central highlands of Mexico ([Figure 1](#)). Glacial and Holocene lake-level records across the Americas were reviewed by [Bradbury et al. \(2001\)](#) and [Fritz et al. \(2001\)](#). [Vimeux et al. \(2009\)](#) provide a wide-ranging review of climate change in South America since the LGM, including lake-level data. Bradbury et al.'s review confirmed the picture of generally dry (and cool) conditions in tropical latitudes around the LGM and into the Late Glacial. In low-lying areas such as the Yucatan peninsula, this climatic drying was exacerbated by low glacial sea levels which resulted in a fall in regional groundwater levels and hence desiccation of a number of lakes dependent on groundwater inputs to maintain them ([Hodell et al., 1995](#)). The summer rainfall regimes, especially in the Northern Hemisphere tropics, were significantly diminished ([Leyden et al., 1994](#); [Metcalfe et al., 2000](#)). Reduced tropical sea surface temperatures, a lack of summer heating over continental North America (much of which was covered by the Laurentide Ice sheet), a cold North Atlantic, and reduced Northern Hemisphere summer insolation, all played a part. At a number of sites, the Late Glacial appears to have been drier than the full glacial (see Section 'Regional Examples'). The magnitude of precipitation change in the Southern Hemisphere tropics, particularly toward their southern margin where the effect of maximum summer insolation at the LGM would have been strong, is less clear. This is discussed further in the Section 'The Bolivian Altiplano'. Furthermore, long speleothem records from southern Brazil have made a major contribution to our understanding. In the Northern Hemisphere, the Laurentide ice sheet apparently split the jet stream, driving the Westerlies south, bringing increased winter rainfall to the subtropics and parts of the tropics. This resulted in the formation of extensive paleolakes in the southern part of the Basin and Range province, which extended south at least into parts of central Mexico ([Bradbury et al., 2001](#)). The exact extent of this southward migration of the Northern Hemisphere Westerlies, or increased winter precipitation due to more frequent cold fronts, is still unclear, but it may even have reached as far as Guatemala

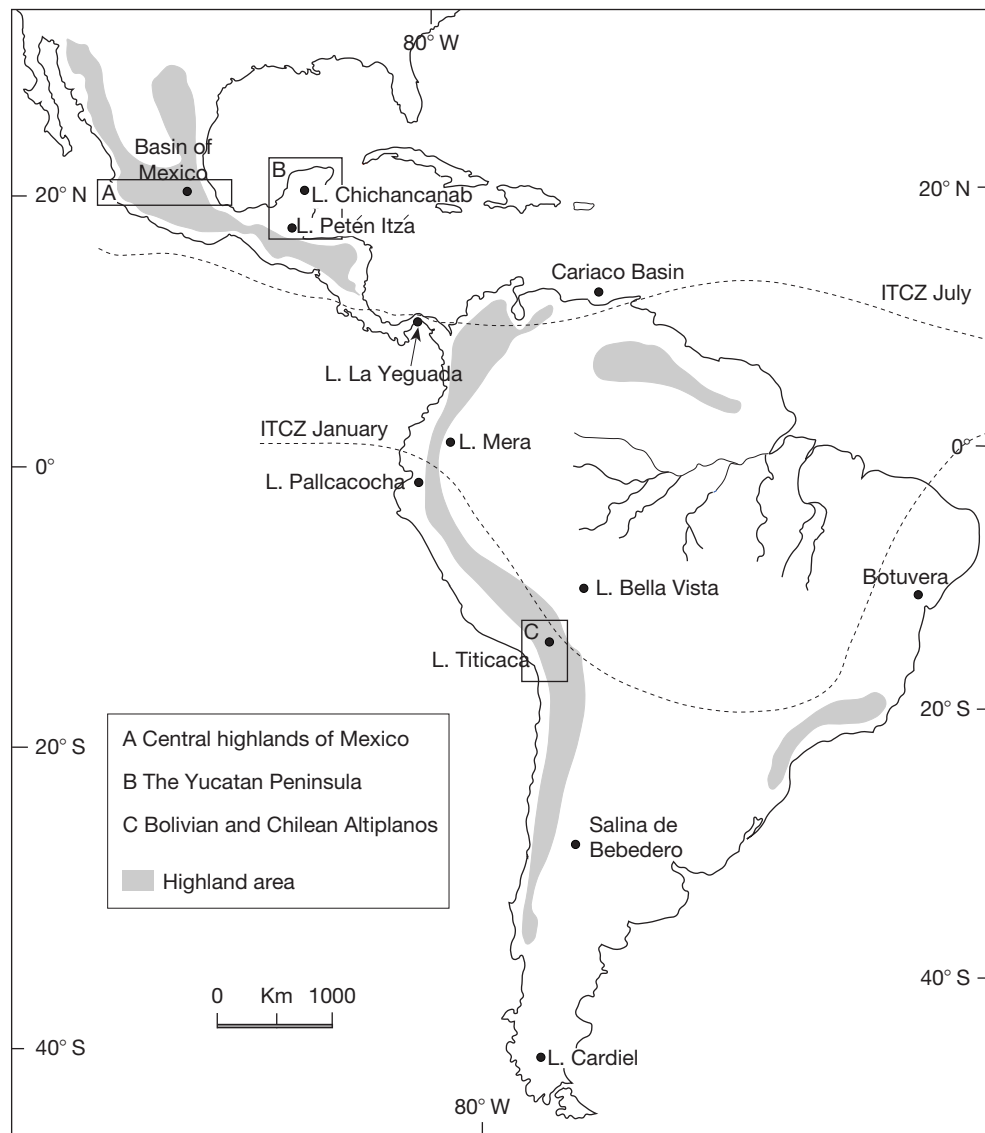


Figure 1 The study area showing the present-day average of locations of the ITCZ in July and January, the locations of the regional examples, and other key sites referred to in the text.

(Hodell et al., 2008). In the Southern Hemisphere, whether the Westerlies tracked poleward or equatorward has been the subject of intensive debate. A review of lake data from South America south of the Tropic of Capricorn (Piovano et al., 2009) highlights the presence of two climatic regimes in southern South America, separated by the ‘Arid Diagonal,’ with the influence of the SASM to the north and east (the area of the Pampean Plain), and the Southern Westerlies to the south and west (Patagonia). The authors report that results from Salina de Bebedero (33° S, currently within the Pampean plain climatic regime) indicate several Late Glacial high stands in response to the equatorward displacement of the Westerlies and increased snow melt from the Andes. From about 13 500 cal BP into the early to mid-Holocene, high lake levels across the Pampean Plains are attributed to a strengthening of the SASM (see Section ‘The Bolivian Altiplano’).

Holocene lake-level records for Latin America are more numerous and generally better dated than those extending back to the full glacial. Sites in the Northern Hemisphere show a broadly coherent pattern of rising levels in the early Holocene as strong summer precipitation became established in response to insolation forcing, the final collapse of the Laurentide ice sheet, and sea-level rise. There are a few exceptions, however, such as La Yeguada in Panama (Bush et al., 1992). At many sites, wet conditions persisted throughout the mid-Holocene, with drying and more variable conditions setting in about 3000 years BP. In the late Holocene, the impacts of both climate and humans on lake catchments have to be taken into account. Some of the most detailed Holocene records come from the Yucatan peninsula and these are discussed below. In the Southern Hemisphere, there are fewer records, and specific lake-level studies are highly concentrated

in the region of the south central Andes (the Altiplano areas of Bolivia, southern Peru, and northern Chile). There are an increasing number of paleoenvironmental reconstructions from areas such as northeastern Brazil, but these are largely based on pollen records. In the Altiplano, the early Holocene was wet with marked drying then setting in through the mid-Holocene (e.g., Wirrmann and de Oliveira Almeida, 1987). This view of dry mid-Holocene conditions has been challenged in the literature both for the south central Andes and for southern South America more generally and is discussed further in the Section 'The Bolivian Altiplano'. By about 3000 ^{14}C years BP, however, there is evidence for wetter conditions in the Altiplano and southern Amazonia, which have been attributed to the insolation-driven southward migration of the ITCZ, the reverse of conditions in the Northern Hemisphere. There are fewer lake-level records from southern South America, but data from Lago Cardiel in Argentinian Patagonia indicate that the modern southern Westerlies became established at this latitude (49°S) by the early Holocene, following a dry, very Late Glacial (Markgraf et al., 2003). Although there is evidence for changes in the intensity of the Westerlies, Lago Cardiel does not seem to have fallen below its present level over the Holocene.

Within the broad picture outlined above, individual studies, or numbers of studies within the same area, raise particular questions or highlight specific issues. The regional examples below illustrate some of these.

Regional Examples

The Central Highlands of Mexico

Volcanic activity, mainly during the Quaternary, resulted in the formation of the central highlands of Mexico (at about 19°N) and a large number of lake basins, many of which still contain water. Lake sediment studies began quite early (Hutchinson et al., 1956), but were not undertaken extensively until the 1980s. Data from Mexico were reviewed by Metcalfe et al. (2000). Most of the records cover only the Late Glacial and Holocene and are of relatively low resolution. In this area, conditions at the LGM are still not clearly established and there are no records covering a full glacial cycle. Tephra deposition and long-term human impact also complicate the interpretation of the records.

The Basin of Mexico lies within the central highlands at about $19^\circ30'\text{N}$, at an elevation of 2240 m. It is surrounded by mountains, with peaks over 5000 m high; the volcanoes Popocatepetl and Iztaccíhuatl have permanent snow and ice fields today. The basin was formerly occupied by a series of lakes draining toward the central Lake Texcoco at about 2236 m. It has a long history of human settlement and was the site of the Aztec city of Tenochtitlan, now Mexico city. The hydrology of the basin has been completely disrupted by human activity including deep drainage, excavation of an artificial outlet, and abstraction of groundwater to supply Mexico city. The basin has been the focus for a number of lake-level studies, but these are generally poorly dated, discontinuous, or difficult to interpret due to occasional inputs of glacial meltwater (in the south of the basin), periodic desiccation, and apparently conflicting evidence from different proxies (Bradbury, 1989; Caballero et al., 1999). Prior to the LGM

(~35 000), the basin was apparently occupied by an extensive but shallow lake with marshlands. Conditions were drier around the LGM, although the intensity of this is not clear. Whether increased winter westerly precipitation reached the Basin of Mexico is disputed, but now seems unlikely. The Late Glacial was, however, very dry, with the subbasins becoming isolated and showing individualistic behavior. Texcoco and Tecocomulco (in the north of the basin) apparently both desiccated. Dry conditions persisted into the early Holocene and there is only modest evidence for increased moisture in the south of the basin around 6000 BP. Late Holocene sediments are not preserved as a result of deflation, so recent change is not recorded.

There are no well-dated, long lake-level records from Mexico, although diatom records have been published from long cores from the Basin of Mexico (Bradbury, 1971) and from Lake Cuitzeo (to the west of the Basin of Mexico). Recent deep drilling of the Basin of Mexico should provide more detailed records of long-term climate change in due course. The Cuitzeo record may extend back more than 120 000 years, but only the top 9 m of the sequence is dated. The last interglacial was probably wetter than present, as was the mid-Pleistocene. Around 35 000 BP, the lake was very low, but it then deepened around the LGM, in contrast to the Basin of Mexico. Conditions in the Late Glacial are unknown, as this part of the sequence is missing. Very low lake level is indicated around 5000 BP, which is part of a North American pattern of mid-Holocene aridity that is not evident further south. The lake has been shallow and increasingly saline over the last 3000 years, a trend which is also apparent in $\delta^{18}\text{O}_{\text{calcite}}$ data. Relatively few of the lakes in the central highlands deposit carbonate, which is suitable for the conventional $\delta^{18}\text{O}$ analysis done at Cuitzeo, but some sites deposit very diatom-rich sediment which is suitable for $\delta^{18}\text{O}_{\text{diatom}}$. Unfortunately, results to date using $\delta^{18}\text{O}_{\text{diatom}}$ have been difficult to interpret and are not discussed further here.

There has been increasing focus on recent change, using a variety of proxies to identify both human and climatic effects and attempts to integrate paleolimnological records with information from the Mexico's large Hispanic archives and more limited instrumental records (e.g., Metcalfe et al., 2010). Increased climatic variability over the last 3000–4000 years is evident in many records from the central highlands and there is emerging evidence that dry periods may show the same periodicities of Bond cycles in the North Atlantic. A major dry period is recorded around 1000 BP (the Late Classic of Mesoamerican archaeology). In the central highlands, however, this dry period was less severe than the mid-Holocene. The application of a wider range of dating methods (such as ^{210}Pb) has also allowed investigation of more recent climatic variability, including the impact of the 'Little Ice Age' (LIA). It is apparent that the expression of this period is complex, with evidence from the central highlands of conditions both wetter and drier than present, although drier seems the more widespread response.

The Yucatan Peninsula

This lowland area of predominantly limestone geology includes parts of Mexico, Belize, and Guatemala (Figure 1). Work here was pioneered by Cowgill et al. (1966) and

continued by Edward S. Deevey, who spent many years trying to find lake deposits that extended back into the LGM (see Section 'Background') and to establish how climate, humans (in this case the Maya), and environment had interacted (Deevey et al., 1979). Before accelerator mass spectrometry (AMS) dating, obtaining reliable radiocarbon dates from bulk sediments was also problematic. Palynology was the main tool of early studies, but stable isotope analysis has produced the most detailed lake-based records for this region. So far, only three lakes have yielded records extending back to the LGM, all in Guatemala: Quexil, Salpeten (Leyden et al., 1994), and Petén Itzá. Lake Petén Itzá (16°55'N), with a maximum water depth of 160 m, is the deepest lake in the region. Seismic survey has revealed a shoreline 56 m below the modern lake that formed at the end of the last glaciation, indicating an 87% reduction in water volume (Anselmetti et al., 2006). This magnitude of lake level fall explains why few other sites in this region contained water in the Late Glacial. Petén Itzá clearly has the potential to produce the longest records from this part of the Northern Hemisphere American tropics and has been cored using the GLAD800 drilling system (<http://www.dosecc.org>). Hodell et al. (2008) have published a record from Petén Itzá covering the last 85 000 years, based on core logging for density and magnetic susceptibility, pollen analysis, and a significant number of radiocarbon dates over the last 40 000 years. Unlike previous studies, this work indicates that the period of the LGM was marked by the presence of a deep lake and pine-oak vegetation, indicative of cool and moist conditions. Reduced evaporation and an increase in winter precipitation may be the explanation. Generally, conditions in the Peten were dry when the ITCZ was displaced south of its current position. On short timescales, such a southward migration of the ITCZ occurs during Heinrich events, when the North Atlantic cools abruptly and the Atlantic meridional overturning circulation slows down. Heinrich events 1–4 (H1–H4) and the Younger Dryas (sometimes referred to as H0) are recorded over the last 40 000 years of the record.

A number of detailed reconstructions of late Holocene lake-level change have been published from the Yucatan peninsula. Lake Chichancanab (Mexico) has yielded records of increasing resolution from the original stable isotope study of Covich and Stuiver (1974) to a high-resolution study of the last 2800 years. The pattern of late Holocene change in Chichancanab has been identified as showing periodicities (208 years) consistent with solar forcing (Hodell et al., 2001). Similar solar cycles have been identified in marine cores from the Cariaco basin off northern Venezuela. One of the strongest features of the Yucatan records is a series of major droughts at the end of late Classic (ca. AD 900), which coincided with the collapse of the Maya civilization. This is discussed in more detail in Section 'Climate and Culture'. In this region, it appears that the LIA was drier than present, consistent with the effect of North Atlantic cooling recorded during the LGM and Heinrich events.

The Bolivian Altiplano

The Bolivian (and Peruvian) Altiplano is a high-altitude valley (average elevation about 4000 m) within the central Andes (Figure 1). The valley is occupied by a series of lakes or salt flats (salars) which have shown varying degrees of hydrological

integration over the Pleistocene. Lake Titicaca is the highest of the basins (3810 m) and most northerly, overflowing to the south through the Rio Desaguadero into Lake Poopo. The Salar de Uyuni (3653 m) is the most southerly of the lake basins, where drilling has revealed a long sequence of alternating lacustrine clays (wet) and salt deposits (dry). The site of the Quelccaya ice core lies to the north, while that of the Sajama ice core is the western Cordillera between Titicaca and the salars (for details see <http://www.ncdc.noaa.gov/paleo/ice-core.html>). The integration of glacial, ice core, and lake records has been an important element of work in the Altiplano. A number of papers on the area have been brought together by Seltzer et al. (2003). Early studies reported the existence of a major deep lake phases in the Late Glacial, together called the Taucu phase, when Titicaca was deep and overflowing and Taucu was also deep (up to 142 m according to Fornari et al., 2001). The timing of the start of the Taucu phase was extended back to before the LGM (to ~26 000), for details see Baker et al. (2001). The identification of wet conditions at the LGM sets this area apart from the Northern Hemisphere tropical Americas and also implies that Amazonia must have been wet as it (with the Atlantic) is the main source of precipitation for the Altiplano. The implication is that the SASM remained strong, driven by the southward displacement of the ITCZ and stronger summer insolation. A link between wet conditions on the Altiplano and cold sea surface temperatures in the North Atlantic has been proposed. The enhancement of the SASM during the LGM, and coincident with Heinrich events in the North Atlantic, has been supported by long speleothem records from southern Brazil, primarily from Caverna Botuverá (Wang et al., 2007). As described above, however, the records from Amazonia generally point to drier conditions, or perhaps little change from present. There are very few records from southern Amazonia, but Mayle et al.'s (2000) pollen records from lowland eastern Bolivia (Laguna Bella Vista and Laguna Chaplin) indicate dry conditions in the full and Late Glacial.

Although the transition from the glacial into the Holocene was complex (Baker et al., 2001), lake levels fell in the early Holocene, and by 6000–5000 BP Titicaca was apparently some 85 m below its present level, the lowest of the last 25 000 years. This pattern of mid-Holocene aridity was apparently widespread in the Bolivian and Chilean Altiplano as well as in southern Amazonia. The pattern of drying from the early Holocene is the opposite of that in the Northern Hemisphere tropics and can be explained by the reduced summer insolation and the northward displacement of the ITCZ.

As described in the Section 'Background', however, the ubiquity of a dry mid-Holocene across southern South America has been challenged. It has been suggested that one of the reasons for the differences of opinion may relate to the varying response times of different proxies (Grosjean et al., 2003). However, the relative sparseness of records across a very large area must be considered, and the more abundant records from the northern tropical Americas illustrate the spatial and temporal complexity of Holocene change. Lake levels in the Altiplano recovered in the late Holocene and Titicaca has been generally high, except for four drier intervals, over the last 3000 years.

The Altiplano is starting to yield some long records of change due to deep drilling in the Salar de Uyuni, the valley of the Desaguadero, and Titicaca itself, using the GLAD800 coring system (<http://www.dosecc.org>). These have confirmed

the occurrence of multiple wet phases on the Altiplano over the mid- and late Pleistocene. There seem to be different trends in the northern and southern Altiplano, with the lake levels during wet periods getting progressively higher in the south and lower in the north, possibly due to changes in the height of the outlet of the Desaguadero. It also appears that the conditions of the LGM and Holocene are not typical of previous glacial and interglacial stages, which has implications for the identification of climatic forcings (Fritz et al., 2004).

ENSO Records

As described above, the modern climate of South America is particularly strongly affected by ENSO, which brings much wetter than normal conditions to areas of the west coast (e.g., coastal Peru and Ecuador) and drier than normal conditions to areas such as north east Brazil. (For more details regarding ENSO, see <http://www.esrl.noaa.gov/psd/ENSO/>.) To date, there are few lake sediment records from South or Central America (including Mexico) of sufficiently high resolution (annual or subannual) to capture ENSO signals; our understanding of ENSO is based largely on records from corals, high-resolution marine cores, and tree rings. There are ENSO records from laminated (or mainly laminated) lake sediments, from Laguna Pallcacocha in the Andes of Ecuador and Bainbridge and El Junco crater lakes in the Galapagos Islands. In both areas, El Niños are marked by increased precipitation. At Pallcacocha, El Niño events are marked by layers of clastic sediments produced by increased runoff (Rodbell et al., 1999), while, at Bainbridge Crater, clastic layers are thought to represent very intense El Niños and carbonate layers moderate events which lower the lake's salinity but do not generate a lot of catchment erosion. A similar distinction between rainfall amount (silt/clay) and intensity (sand) is made at El Junco. The records show that the frequency and strength of ENSO events increased over the Holocene, reaching modern periodicities about 7000 BP and very clearly from 5000 to 4000 BP. Strong El Niños have become more common. Pallcacocha and Bainbridge also record millennial-scale variability in ENSO, which may be the result of the system's internal dynamics rather than external forcing (Moy et al., 2002). It seems likely that there are other lake basins in Latin America depositing lake sediments which could yield paleo-ENSO records. The impact of ENSO-type variability, particularly over decadal to centennial timescales, is of increasing interest, although identifying the forcing mechanisms for these is complex.

Climate and Culture

Although one has to be cautious about attributing cultural change to climatic variability, there is paleolimnological evidence from a number of sites in South and Central America to suggest that persistent drought conditions have been important in driving change in human societies (Brenner et al., 2001). A key area for this has been the Yucatan peninsula (see Section 'The Yucatan Peninsula') which lay at the heart of the pre-Hispanic Maya culture. Around AD 900, major building projects apparently ceased and a number of large settlements were abandoned, which has come to be called

the Maya 'collapse.' Given the problems of water storage and access in limestone geology and a variable climate, it had been hypothesized that drought may have played a part in the late Classic collapse. Hodell et al.'s (1995) paper on Chichancanab (see Section 'The Yucatan Peninsula') provided the first unequivocal evidence for drought in the late Classic, although the sampling and dating resolution were not high. A more detailed record from Lake Punta Laguna (Mexico) confirmed the late Classic drought, but also indicated that the droughts had multiple peaks and had become progressively more intense (Curtis et al., 1996). Prior to the collapse, an earlier series of droughts could be associated with the Maya hiatus. The complexity of the drought signal has been confirmed both by the marine Cariaco core (see Section 'The Yucatan Peninsula') and by a higher resolution study from Chichancanab based on bulk density data (Hodell et al., 2005). It should not be assumed, however, that all lake records in the Yucatan record a major drought in the late Classic; some lakes do not, either because they are large enough for their water balance not to be significantly affected, or because the spatial extent of the drought was probably variable. There is some evidence for late Classic drought in the central highlands of Mexico (see Section 'The Central Highlands of Mexico'), but here it was apparently less intense except perhaps on the drier northern margin of the highlands where some catchments may have been abandoned.

There is also lake-level evidence for drought affecting pre-Hispanic cultures in the central Andes. The best documented case relates to the Tiwanaka culture that occupied the area around Titicaca from about AD 600, but collapsed about AD 1150 at the same time as drought conditions are recorded in a core from Titicaca (and in the Quelccaya ice cap). This drought period has also been recorded in the area around Cuzco (Peru), where many pre-Hispanic drought periods correspond to important cultural changes (Chepstow-Lusty et al., 2003 or <http://antiquity.ac.uk/ant/085/ant0850570.htm>).

See also: **Diatom Methods:** Salinity and Climate Reconstructions from Continental Lakes. **Glaciations:** Late Pleistocene in South America; Late Quaternary in North America. **Ice Core Records:** South America. **Lake Level Studies:** Modeling; North America. **Paleoclimate:** Introduction; Timescales of Climate Change. **Paleoclimate Modeling:** Data-Model Comparisons; Last Glacial Maximum GCMs. **Pollen Records, Late Pleistocene:** South America; Western North America. **Pollen Records, Postglacial:** South America; Southwestern North America.

References

- Anselmetti FS, Ariztegui D, Hodell DA, et al. (2006) Late Quaternary climate-induced lake level variations in Lake Petén Itzá, Guatemala, inferred from seismic stratigraphic analysis. *Palaeogeography, Palaeoclimatology, Palaeoecology* 230: 52–69.
- Baker PA, Seltzer GO, Fritz SC, et al. (2001) The history of South American tropical precipitation for the past 25,000 years. *Science* 291: 640–643.
- Bradbury JP (1971) Paleolimnology of Lake Texcoco, Mexico. Evidence from diatoms. *Limnology and Oceanography* 16: 180–200.
- Bradbury JP (1989) Late Quaternary lacustrine paleoenvironments in the Cuenca de Mexico. *Quaternary Science Reviews* 8: 75–100.
- Bradbury JP, Grosjean M, Stine S, and Sylvestre F (2001) Full and late glacial lake records along the PEP1 transect: Their role in developing interhemispheric

- paleoclimate interactions. In: Markgraf V (ed.) *Interhemispheric Climate Linkages*, pp. 265–291. San Diego, CA: Academic Press.
- Brenner M, Hodell DA, Curtis JH, Rosenmeier MF, Binford MW, and Abbott MB (2001) Abrupt climate change and pre-Columbian cultural collapse. In: Markgraf V (ed.) *Interhemispheric Climate Linkages*, pp. 87–103. San Diego, CA: Academic Press.
- Bush MB, Piperno DR, Colinvaux PC, De Oliveira Krissek LA, Miller MC, and Rowe WE (1992) A 14,300 yr palaeoecological profile of a lowland tropical lake in Panama. *Ecological Monographs* 62: 251–275.
- Caballero M, Lozano S, Ortega B, Urrutia J, and Macias J-L (1999) Environmental characteristics of Lake Tecocomulco, northern Basin of Mexico, for the last 50,000 years. *Journal of Paleolimnology* 22: 399–411.
- Chepstow-Lusty A, Frogley MR, Bauer BS, Bush MB, and Tupayachi-Herrera A (2003) A late Holocene record of arid events from the Cuzco region, Peru. *Journal of Quaternary Science* 18: 491–502.
- Covich A and Stuiver M (1974) Changes in oxygen 18 as a measure of long-term fluctuations in tropical lake levels and molluscan populations. *Limnology and Oceanography* 19: 682–691.
- Cowgill UM, Goulden C, Hutchinson GE, Patrick R, Recek AA, and Tsukada M (1966) The history of Laguna de Pentexil, a small lake in northern Guatemala. *Memoirs of the Academy of Arts and Sciences XVII*: 1–126.
- Curtis JH, Hodell DA, and Brenner M (1996) Climate variability on the Yucatan peninsula (Mexico) during the past 3500 years, and implications for Maya cultural evolution. *Quaternary Research* 46: 37–47.
- Deevey ES, Rice DS, Rice PM, Vaughan HH, Brenner M, and Flannery MS (1979) Mayan urbanism: Impact on a tropical karst environment. *Science* 206: 298–306.
- Fornari M, Risacher F, and Feraud G (2001) Dating of paleolakes in the central Altiplano of Bolivia. *Palaeogeography, Palaeoclimatology, Palaeoecology* 172: 269–282.
- Fritz SC, Baker PA, Lowenstein TK, et al. (2004) Hydrologic variation during the last 170,000 years in the southern hemisphere tropics of South America. *Quaternary Research* 61: 95–104.
- Fritz SC, Metcalfe SE, and Dean W (2001) Holocene climate patterns in the Americas inferred from paleolimnological records. In: Markgraf V (ed.) *Interhemispheric Climate Linkages*, pp. 241–263. San Diego, CA: Academic Press.
- Grosjean M, Cartajena I, Geyh MA, and Nunez L (2003) From proxy data to paleoclimate interpretation: The mid-Holocene paradox of the Atacama desert, northern Chile. *Palaeogeography, Palaeoclimatology, Palaeoecology* 194: 247–258.
- Hodell DA, Anselmetti FS, Aritzegui D, et al. (2008) An 85-ka record of climate change in lowland Central America. *Quaternary Science Reviews* 27: 1152–1165.
- Hodell DA, Brenner M, and Curtis JH (2005) Terminal Classic drought in the northern Maya lowlands inferred from multiple sediment cores in Lake Chichancanab (Mexico). *Quaternary Science Reviews* 24: 1413–1427.
- Hodell DA, Brenner M, Curtis JH, and Guilderson T (2001) Solar forcing of drought frequency in the Maya lowlands. *Science* 292: 1367–1370.
- Hodell DA, Curtis JH, and Brenner M (1995) Possible role of climate in the collapse of Classic Maya civilization. *Nature* 375: 391–394.
- Hutchinson GE, Patrick R, and Deevey ES (1956) Sediments of Lake Patzcuaro, Michoacan, Mexico. *Bulletin of the Geological Society of America* 67: 1491–1504.
- Leyden BW, Brenner M, Hodell DA, and Curtis JH (1994) Orbital and internal forcing of climate on the Yucatan peninsula for the past ca. 36 ka. *Palaeogeography, Palaeoclimatology, Palaeoecology* 109: 193–210.
- Liu K-B and Colinvaux P (1985) Forest change in the Amazon basin during the last glacial maximum. *Nature* 318: 556–557.
- Markgraf V, Bradbury JP, Schwalb A, et al. (2003) Holocene palaeoclimates of southern Patagonia: Limnological and environmental history of Lago Cardiel, Argentina. *The Holocene* 13: 597–607.
- Mayle FE, Burbridge R, and Killeen TJ (2000) Millennial-scale dynamics of southern Amazonian rain forests. *Science* 290: 2291–2294.
- Metcalfe SE, Jones MD, Davies SJ, Noren A, and MacKenzie A (2010) Climate variability over the last two millennia in the North American Monsoon region, recorded in laminated sediments from Laguna de Juanacatlan, Mexico. *The Holocene* 20: 1195–1206.
- Metcalfe SE, O'Hara SL, Caballero M, and Davies SJ (2000) Records of late Pleistocene–Holocene climatic change in Mexico – A review. *Quaternary Science Reviews* 19: 699–721.
- Moy CM, Seltzer GO, Rodbell DT, and Anderson DM (2002) Variability of El Niño/Southern Oscillation activity at millennial timescales during the Holocene epoch. *Nature* 420: 162–165.
- Piovano EL, Ariztegui D, Vimeux F, Cioccale M, and Silvestre F (2009) Hydrological variability in South America below the Tropic of Capricorn (Pampas and Patagonia, Argentina) during the last 13.0 Ka. In: Silvestre F, Sylvestre F, and Khodri M (eds.) *Past Climate Variability in South America and Surrounding Regions*, pp. 323–351. Berlin: Springer.
- Rodbell DT, Seltzer GO, Anderson DM, Abbott M, Enfield DB, and Newman JH (1999) An 15,000-year record of El Niño-driven alluviation in southwestern Ecuador. *Science* 283: 516–520.
- Seltzer GO, Rodbell DT, and Wright HE (eds.) (2003) *Late-Quaternary paleoclimates of the southern tropical Andes and adjacent regions*. *Palaeogeography, Palaeoclimatology, Palaeoecology* 194. (special issue).
- Van de Hammen T (1974) The Pleistocene changes of vegetation and climate in tropical South America. *Journal of Biogeography* 1: 3–26.
- Vimeux F, Sylvestre F, and Khodri M (eds.) (2009) *Past Climatic Variability in South America and the Surrounding Regions*. *Developments in Paleoenvironmental Research*, vol. 14. New York: Springer.
- Wang X, Auler AS, Edwards RL, et al. (2007) Millennial-scale precipitation changes in southern Brazil over the past 90,000 years. *Geophysical Research Letters* 34: L23701. <http://dx.doi.org/10.1029/2007GL031149>.
- Wirrmann D and de Oliveira Almeida LF (1987) Low Holocene level (7700 to 3650 years ago) of Lake Titicaca (Bolivia). *Palaeogeography, Palaeoclimatology, Palaeoecology* 59: 315–323.

North America

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Introduction

Stratigraphic records of lake-level change are important for reconstructing the hydrologic budget of the past. They provide a clearer picture of long-term hydroclimate variability than instrumental records, which in North America are typically shorter than 150 years, and they complement other long-term records of changes in terrestrial moisture budgets, such as dendroclimatological, palynological, and charcoal records. In western North America, dramatic changes in the climate and landscape during the late Pleistocene and Holocene produced variations in lake level of 10s to 100s of meters (Benson, 2003). The geological evidence of this variation includes former lake shorelines where no lakes currently exist, deltas evident in subsurface geophysical surveys where rivers previously entered vast lakes, and massive flood-formed landforms that were the result of the release of water from ice-dammed lakes in regions currently without continental ice sheets. From the spectacular records of the release of meltwater from glacial ice-dammed Lakes Agassiz (Teller et al., 2002) and Missoula (Peterson et al., 2011) to the rapid drainage of the extensive pluvial Lakes Lahontan and Bonneville (Benson, 1978; Oviatt, 1997), to the more subtle variation embedded in the sediments of small glacial lakes (Williams et al., 2010), records of lake-level change have played an important role in identifying and developing our understanding of Quaternary climate change in North America (Figure 1; Table 1).

Each lake is defined by a unique set of physical, chemical, and biological parameters that determine how it responds to the dynamics of external drivers, such as changes in climate. A substantial change in any of the major climatic variables that affect a lake's water budget, for example, a change in precipitation, runoff, evaporation rate, or groundwater flow can effect a change in lake level (Fritz, 2008). The influence of these external climate drivers is modulated by the size and morphology of a lake basin and its watershed, which determine the rate and magnitude of the lake's response to changes in hydrologic variation and the nature of the climate signature recorded in the basin. Small lakes typically function somewhat like individual climate stations tracking local conditions that are superimposed on long-term patterns of climate change, whereas larger lakes tend to integrate the influence of the hydrologic budget of a broader region (Shuman et al., 2002; Williams et al., 2010). In addition, some regions are more sensitive to certain aspects of climate variability, such that even small changes in effective moisture can produce profound changes in lake level.

Inferring the drivers of lake-level change and the magnitude of their variation from paleolimnological records can be challenging because of the complex influences of hydrogeomorphic setting on a lake's hydrologic budget (Fritz, 2008).

Many paleolakes have shown marked fluctuations in lake level during the Pleistocene and Holocene in response to capture or diversion of influent rivers rather than climate (Benson and Paillet, 1989; Bradbury et al., 1989; Kaufman et al., 2009; Kowalewska and Cohen, 1998). Landscape position also affects the nature of lake-level change in response to external drivers. For example, in kettle lakes connected by a shallow aquifer, the magnitude of lake-level lowering can be strongly influenced by the distance between the lake and the base level of the river system that drains the watershed (Digerfeldt et al., 1993); therefore, regionally, lake-level lowering in response to aridity might be greater in lakes farther from rivers. In dune-dominated landscapes, hydrologic response to drought can be quite complex. In some regions of the Nebraska Sand Hills, blockage of paleovalleys by migrating dune sand during the late-Glacial and Holocene periods caused a rise in the regional water table (Loope et al., 1995; Mason et al., 1997). In interdunal wetlands above the dune dams, water levels aggraded, producing lake-level rise during intervals of extreme aridity.

To some extent, a network of sites can depict the dominant patterns of lake-level variation in response to climate and can circumvent the variable influences of hydrogeomorphic setting. A North American lake-level database, largely derived from sedimentological and palynological data, has been used to examine patterns and rates of lake-level change in the eastern and Midwestern parts of the continent following the retreat of the Laurentide Ice Sheet (Shuman et al., 2002; Williams et al., 2010). In the Midwest, for example, the impact of site-specific factors is clearly evident in the varied lake status in any given time slice (Figure 2), yet the central tendency was one of declining lake levels as climate dried from the early to mid-Holocene. At the majority of sites, the rate of change was most rapid at 8 cal ka BP (Williams et al., 2010), which suggests the combined influence of the collapse of the Laurentide Ice Sheet and high summer insolation in driving lake-level decline.

Regional Lake-Level Histories

The Northern Great Plains and Midwestern North America

Glacial Lake Agassiz is undoubtedly the most significant lacustrine lake-level record from north-central North America. Lake Agassiz was an enormous ice-dammed lake that formed at the southern margin of the Laurentide ice sheet during the last deglaciation. At ~13 cal ka BP, it was the largest lake in North America (NA) (>400,000 km²), and changes in the routing of its drainage are posited to have played a major role in affecting both regional and global processes (Teller et al., 2002). The lake-level history is delineated by several strandlines formed at high lake stages separated by intervals of lowered lake level. The lowered levels resulted from rapid drainage

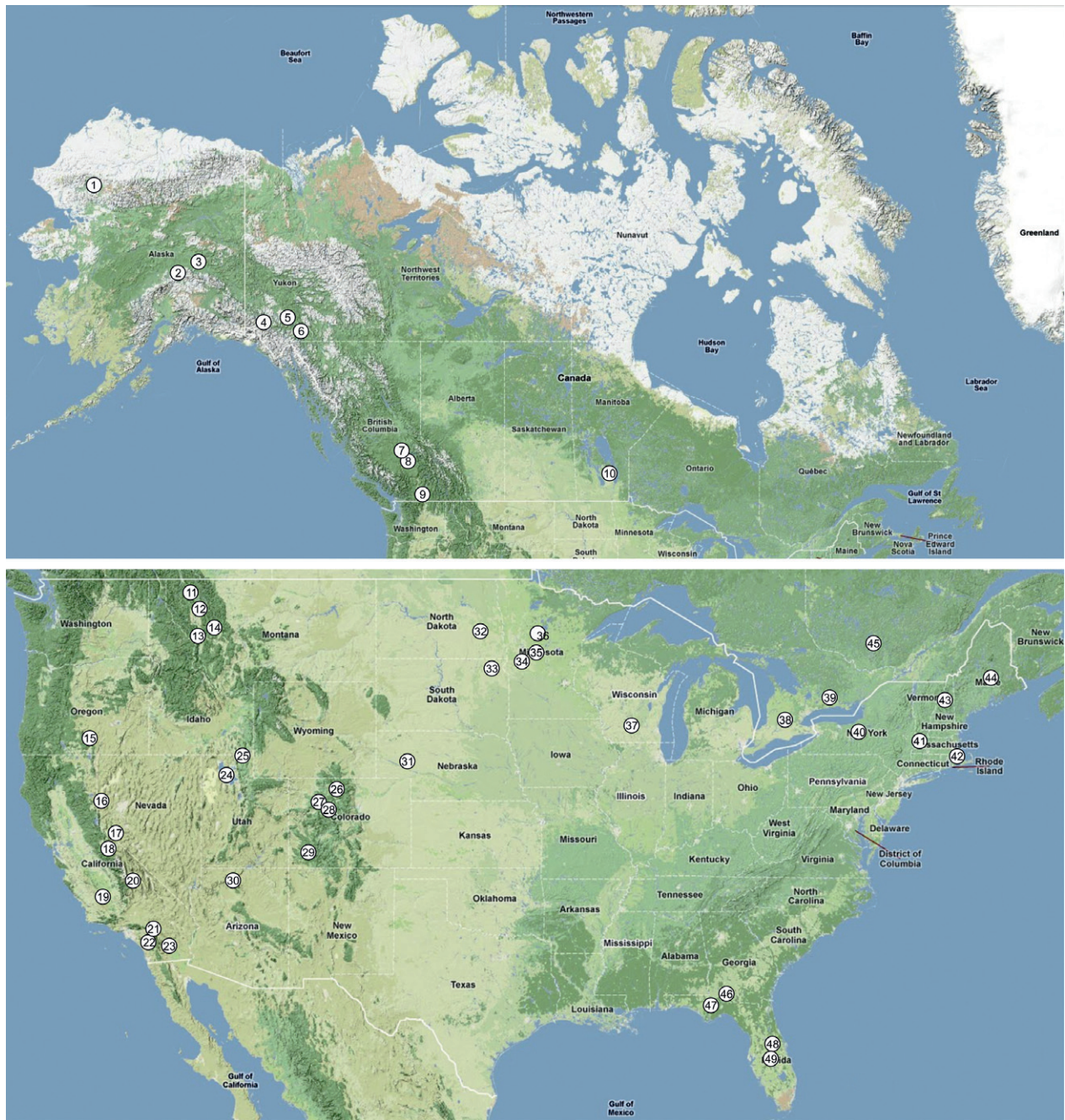


Figure 1 Sites of lake-level studies in North America discussed in the text. Number locations are keyed to the references listed in [Table 1](#).

of large volumes of water as the southern margin of the Laurentide ice sheet retreated and exposed new outlets. Rapid outbursts of glacial meltwater into the North Atlantic have been hypothesized to have affected meridional overturning circulation and produced widespread and rapid climatic change, including the Younger Dryas interval, the Preboreal Oscillation, and the 8.2 climate event (Clark et al., 2001). However, the chronology of drainage, the specific locations of outlets, and hence the role of Lake Agassiz in affecting the observed climate events is controversial (Fisher et al., 2008; Murton et al., 2010). The Laurentian Great Lakes themselves also

show a dramatic history of lake-level change during deglaciation and the Holocene that reflects of the complex interplay of climate, glacial meltwater inputs, and isostatic adjustment (Clark et al., 2011; Colman et al., 1994). Higher frequency lake-level variation in the late Holocene interval may be more clearly linked to climate than the fluctuations of prior periods (Wilcox et al., 2007).

The majority of other lacustrine records in the northern tier ($>40^{\circ}\text{N}$) of the NA continental interior are from smaller lakes that were formed by retreat of the Laurentide ice sheet; thus stratigraphic records from these sites span the terminal

Table 1 Relevant lake-level reconstruction sites and references discussed in the text. Map numbers index to numbers in **Figure 1**

Map #	Site	Reference
1	Burial Lake, AK	Abbott et al. (2010)
2	Windmill Lake, AK	Bigelow and Edwards (2001)
3	Birch Lake, AK	Abbott et al. (2000)
4	Kluane Lake, YT	Brahney et al. (2008)
5	Jellybean Lake, YT	Anderson et al. (2005a)
6	Marcella Lake, YT	Anderson et al. (2005b)
7	Big Lake, BC	Walker and Pellatt (2008)
8	Valentine Lake, BC	Walker and Pellatt (2008)
9	Burnell, Mahoney, Kilpoola Lakes, BC	Walker and Pellatt (2008)
10	Glacial Lake Agassiz	Teller et al. (2002), Fisher et al. (2008), and Murton et al. (2010)
11	Foy Lake, MT	Stone and Fritz (2004, 2006), Stevens et al. (2006), and Shuman et al. (2009)
12	Flathead Lake, MT	Hofman et al. (2006)
13	Glacial Lake Missoula	Pardee (1910) and Peterson et al. (2011)
14	Jones Lake, MT	Shapley et al. (2009)
15	Summer Lake, OR	Cohen et al. (2000)
16	Pyramid Lake, NV/ Glacial Lake Lahontan	Russell (1885), Benson (1978), and Benson and Thompson (1987)
17	Walker Lake, NV/ Glacial Lake Lahontan	Bradbury et al. (1989), Benson and Thompson (1987), Benson and Paillet (1989), and Adams (2007)
18	Mono Lake, CA	Stine (1990, 1994)
19	Glacial Tulare Lake	Negrini et al. (2006)
20	Owens Lake, CA	Bacon et al. (2006)
21	Dry Lake, CA	Bird et al. (2010)
22	Elsinore Lake, CA	Kirby et al. (2010)
23	Salton Sea, CA	Li et al. (2008)
24	Great Salt Lake, UT/ Glacial Lake Bonneville	Gilbert (1890), Oviatt et al. (1992), Oviatt (1997), Kowalewska and Cohen (1998), Balch et al. (2005), and Patrickson et al. (2010)
25	Bear Lake, UT/ID	Smoot and Rosenbaum (2009) and Kaufman et al. (2009)
26	Hidden Lake, CO	Anderson (2011)
27	Bison Lake, CO	Anderson (2011)
28	Cottonwood Pass Pond, CO	Fall (1997)
29	Little Molas Lake, CO	Shuman et al. (2009)
30	Kaibab Plateau (Fracas & Bear Lakes), AZ	Weng and Jackson (1999)
31	Crescent & Swan Lakes, NE	Loope et al. (1995) and Mason et al. (1997)
32	Moon Lake, ND	Laird et al. (1996)
33	Pickrel Lake, SD	Smith et al. (2002)
34	Elk Lake, MN	Schwalb and Dean (1998)
35	Parkers Prairie, MN	Digerfeldt et al. (1993)
36	Shingobee Lake, MN	Filby et al. (2002)
37	Mendota Lake, WI	Winkler et al. (1986)
38	Ontario	Lavoie and Richard (2000)
39	Rice Lake, ON	Yu and McAndrews (1994) and Yu et al. (1997)
40	Owasco Lake, NY	Dwyer et al. (1996)
41	Crooked Pond, MA	Shuman et al. (2001)

(Continued)

Table 1 (Continued)

Map #	Site	Reference
42	Makepeace Cedar Swamp, MA	Newby et al. (2000, 2011)
43	New Hampshire	Shuman et al. (2005)
44	Mansell Pond, ME	Almquist et al. (2001)
45	Quebec	Muller et al. (2003)
46	Scott Lake, GA	Shuman et al. (2002)
47	Camel Pond, FL	Watts et al. (1992)
48	Buck Lake, FL	Shuman et al. (2002)
49	Lake Tulane, FL	Grimm et al. (1993, 2006)

Pleistocene and the Holocene. In the northern Great Plains, a grassland region to the east of the Rocky Mountains, lake-level variation commonly has been inferred from reconstructions of salinity, ionic composition, and isotopic variation rather than proxies of lake level itself. Here, freshwater conditions persisted from the onset of lake formation into the early Holocene, suggesting high lake levels and abundant precipitation. High salinity in the mid-Holocene was followed by lowered salinity in the late Holocene, suggesting a shift from low lake levels and arid climate to higher lake levels during the cooler, drier conditions of the last 3–4 cal ka (Laird et al., 1996; Schwalb and Dean, 1998; Smith et al., 2002). In the forested areas of the Midwest, from the modern forest-prairie border to the western Great Lakes region, lake-level reconstructions have been derived from paleolimnological, pollen, and sedimentological data, including transects of cores (Figure 3; Digerfeldt et al., 1993; Filby et al., 2002; Harrison and Metcalfe, 1985; Winkler et al., 1986). Here Holocene patterns of lake-level and inferred climate variation are similar to those inferred from the northern Great Plains, although the timing of onset of mid-Holocene aridity is commonly somewhat later than in sites further west (Fritz et al., 2001). The high lake levels and effective moisture of the early Holocene throughout the northern Great Plains and Midwest, at a time when summer insolation also was very high, likely reflects the downstream effects of the remnant Laurentide ice sheet, which persisted in northern Canada until ~8 cal ka BP. Similarly, the onset of mid-Holocene aridity in the interior is probably linked to ice-sheet collapse and the associated abrupt drainage of Lake Agassiz at ~8.4 cal ka BP (Williams et al., 2010).

The Great Basin

Throughout the Pleistocene, the Great Basin was occupied by dozens of large pluvial lakes (Benson and Thompson, 1987; Gilbert, 1890; Russell, 1885), whereas today just a few permanent lakes remain. The two largest of these lake systems were Lake Bonneville, which at its maximum extent covered more than 50,000 km² and was more than 370 m deep, and Lake Lahontan, which covered more than 22,000 km² and was more than 275 m deep. Early work focused on the timing and correlation of physical evidence of shorelines from former Lakes Bonneville and Lahontan in the subsurface or exposed in outcrops (Benson, 1978; Oviatt et al., 1992; Smith and Street-Perrot, 1983), while subsequent research has coupled this approach with continuous high-resolution stratigraphic records of past lake level from persistent or remnant lakes, inferred

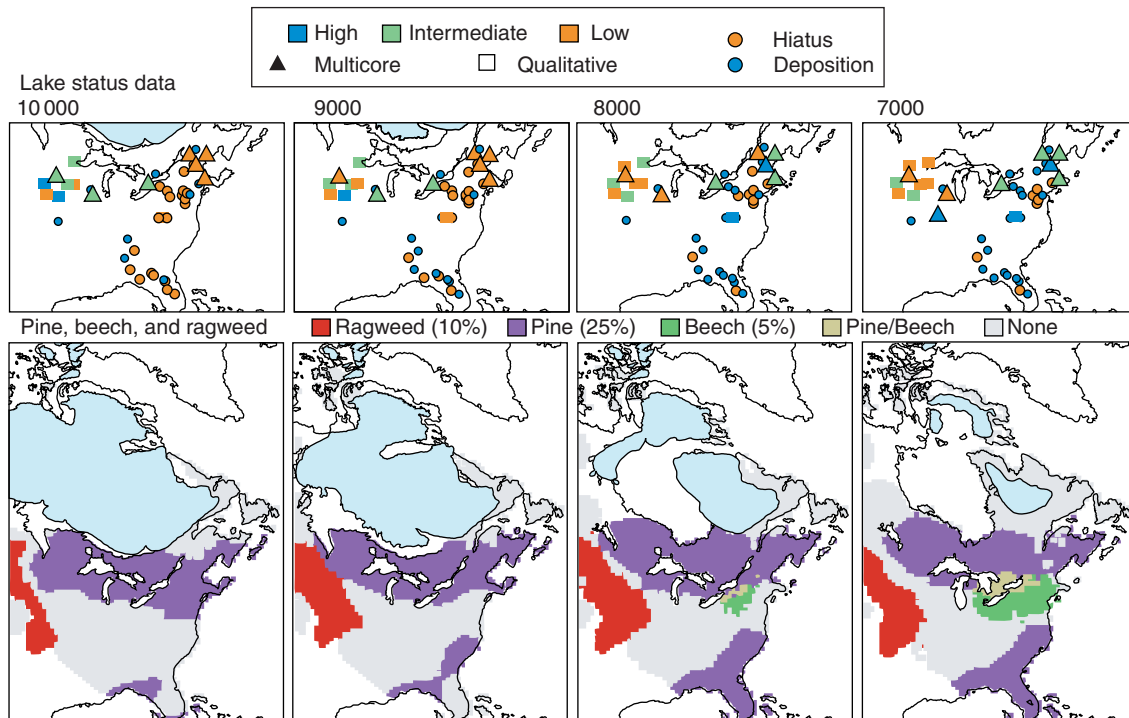


Figure 2 Lake-level sites and plant taxa distributions for eastern North America. Lake levels dropped in the central US between 10 and 7 ka, but rose in other regions, indicating that moisture-balance variations differed by region. The lake-level data generally match with palynological evidence for an eastward advance of the prairie–forest boundary during the early Holocene and increased abundances of mesophytic (relatively moisture-loving) taxa (beech, *Fagus*, hemlock, *Tsuga*, and southern species of pine, *Pinus*) in the eastern US; northern species of pine (*Pinus*) prefer drier conditions than beech and hemlock. Note that symbols in the upper panels indicate whether lake-level changes were inferred from core transect studies, qualitative interpretations of data from single cores, or from sedimentary hiatuses in single cores. Data from Shuman et al. (2002).

from changes in fossil diatom and ostracode assemblages, pollen and macrophyte assemblages, sedimentology, isotopic composition, other geochemical variables, and grain size (Adams, 2007; Bacon et al., 2006; Benson, 2003; Benson and Thompson, 1987; Bradbury et al., 1989; Cohen et al., 2000; Kaufman et al., 2009; Li et al., 2008; Negrini et al., 2006).

The longest records from the lakes of the Great Basin express a rich lake-level history through the Quaternary, with multiple prolonged periods of lake expansion and contraction. Often change was asynchronous among basins as lakes spilled over, downcut through topographic divides, and coalesced (Benson and Thompson, 1987). The influx of moisture for lakes in the Great Basin is strongly influenced during the cool seasons by the polar jet stream, which carries moisture from air masses sourced in the North Pacific. Throughout the late Pleistocene, persistent changes in the inferred position and intensity of the polar jet stream were associated with changes in the global ice volume, providing a teleconnective link between glacial–interglacial cycles, large-scale ocean–atmospheric processes, and observed long-term regional patterns in lake level from Pyramid, Owens, Summer, Mono, Great Salt, and Bear Lakes (Balch et al., 2005; Benson et al., 2003; Kaufman et al., 2009; Oviatt, 1997; Patrickson et al., 2010). Leading into the Last Glacial Maximum (LGM), the records from most of the lakes of the Great Basin indicate a gradual rise in lake level, reaching the most recent highstands between 16 and 13 cal ka BP (Benson and Thompson, 1987). Between 13 and 10 cal ka BP, the lake-level response to climate was much more variable among basins, but overall the records indicate declining lake levels.

By 12–10 cal ka BP, most of the pluvial lakes of the Great Basin were largely reduced to remnant lakes or had disappeared entirely.

Similar to the record from the Northern Great Plains, the early Holocene was a period of widespread increase in effective moisture, with a broadly coherent pattern of increasing the lake level throughout most of the Great Basin. Many modern lakes in the region reached their Holocene maximum highstands when summer insolation was high (Figure 4), between 10 and 7.5 cal ka BP (Bacon et al., 2006; Bird et al., 2010; Kirby et al., 2010; Smoot and Rosenbaum, 2009), but declined rapidly in the mid-Holocene to near-desiccation levels in a period centered around 5 cal ka BP. After 3.5 cal ka BP, most lakes in the Great Basin began to rise again, although fluctuating around lake levels persistently lower than the early Holocene. Highly detailed physical records of lake-level change from truncated deltas and former shorelines exposed in outcrop, sedimentary structures in sediment cores, and death dates on drowned trees have provided unequivocal evidence of significant lake-level change throughout the late Holocene from the Mono, Tulare, and Walker Lake basins (Adams, 2007; Negrini et al., 2006; Stine, 1990, 1994), including prolonged drought during Medieval times.

The Rocky Mountains and Southern Canadian Cordillera

The mountainous western part of North America is an arid to sub-humid region speckled with numerous small and moderate-sized, shallow to deep lakes, many in

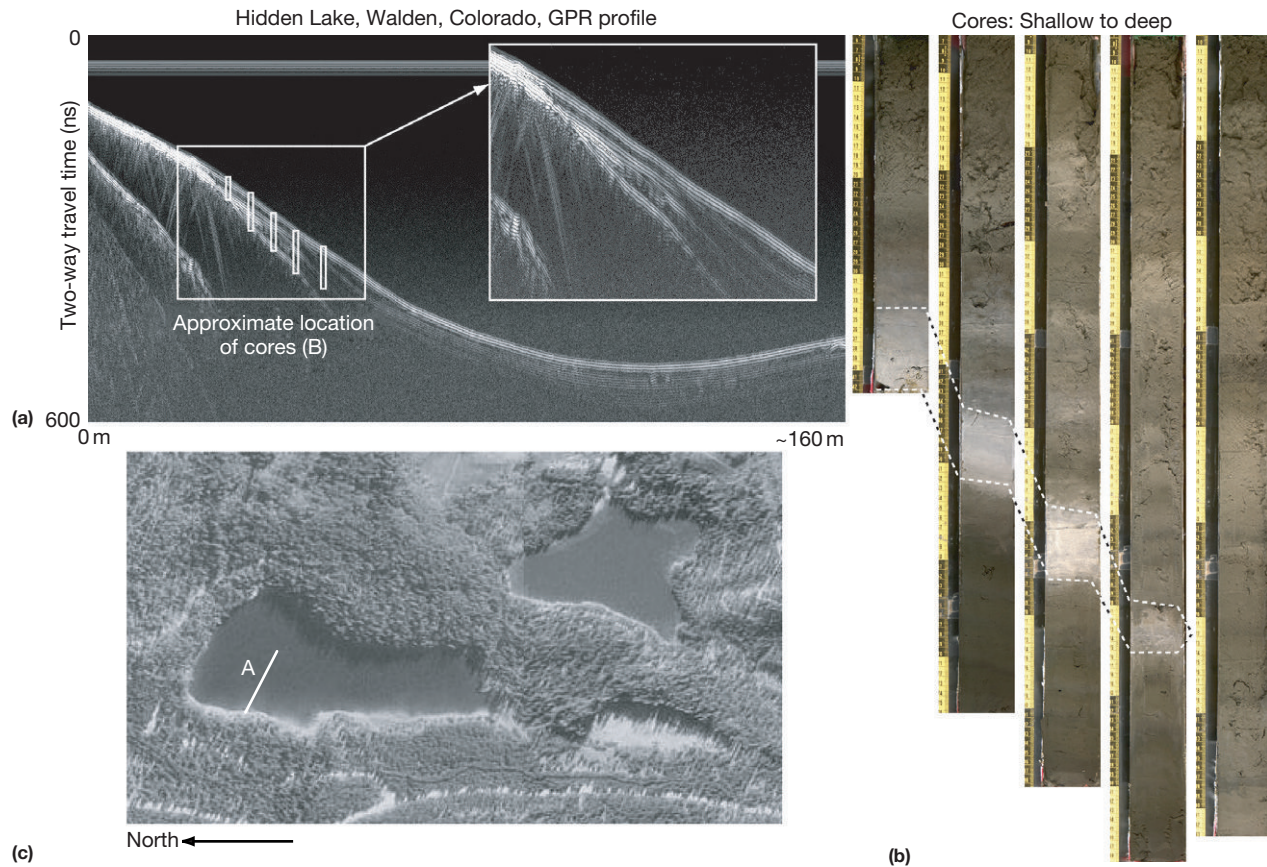


Figure 3 (a) Ground-penetrating radar (GPR) profile and (b) images of transect cores from Hidden Lake in northern Colorado. (c) An aerial photo shows the location of the GPR profile, which contains a bright reflector indicative of a paleoshoreline (and sedimentary unconformity) near the western shore of the lake. Dashed lines highlight the same feature in the core images. The stratigraphic evidence indicates that water levels were likely several meters lower than today during the mid-Holocene. Data from [Shuman et al. \(2009\)](#).

hydrographically closed basins. However, the heterogeneous terrain and large elevational gradients among sites often result in spatially complex patterns of climate. At the time of the LGM, most of this region was overlain by continental ice sheets and mountain glaciers, so many lake archives in the region typically span the last 16 cal ka BP or shorter, depending upon the timing of regional deglaciation. Prior to deglaciation, classic ice-dammed lake sites, such as Glacial Lake Missoula, underwent multiple sequences of substantial changes in lake level as a result of outburst flooding ([Pardee, 1910](#); [Peterson et al., 2011](#)).

[Walker and Pellatt \(2008\)](#) produced a comprehensive analysis of 61 lake records from the area around the Columbia Basin, which included evidence of lake-level changes from Big, Mahoney, Kilpoola, Burnell, Fir, and Valentine Lakes inferred from pollen and paleosalinity reconstructions from fossil chironomid and diatom data. The analysis showed that the timing and direction of lake-level change varied greatly between individual lake basins. For example, the records from Big and Burnell Lakes indicate significant low stands between 9 and 6 cal ka BP, whereas nearby Mahoney Lake has a nearly diametric pattern, with a substantial drop in salinity centered at 7 cal ka BP.

In the Northern Rockies of the United States, changes in lake level and effective moisture have been reconstructed from

fossil diatom assemblages, oxygen isotopes, and multiple-core transect stratigraphy coupled with subsurface geophysical surveys. Evidence from a 14-cal ka BP record from Foy Lake, Montana ([Figure 5](#)) indicates that lake levels fluctuated substantially during the deglaciation and into the early Holocene, yet were lower than today ([Shuman et al., 2009](#); [Stone and Fritz, 2006](#)). A substantial drop in lake level occurred between 7 and 3 cal ka BP, evidenced by changes in fossil diatom and sedimentology, and lakes rose substantially after 3 cal ka BP, reaching the highest stage in the last 1.5 cal ka BP ([Shuman et al., 2009](#); [Stevens et al., 2006](#); [Stone and Fritz, 2004](#)). This pattern matches regional changes in effective moisture reconstructed from oxygen isotopic records from Jones Lake, Montana in the neighboring Ovando Valley of Montana ([Shapley et al., 2009](#)) and the broad pattern of the changing lake level from extensive multiple core and seismic profile studies of nearby Flathead Lake ([Hofman et al., 2006](#)).

Mid-Holocene lowstands also have been identified in the Central and Southern US Rockies. The records of Hidden and Little Molas Lakes in Colorado ([Figure 5](#)) show low lake levels from 7 to 4 cal ka BP ([Shuman et al., 2009](#)), matching the general pattern of low effective moisture during this period reconstructed from oxygen isotope record from Bison Lake, Colorado ([Anderson, 2011](#)), as well as records from multiple other lakes in Arizona, Colorado, and Wyoming that show

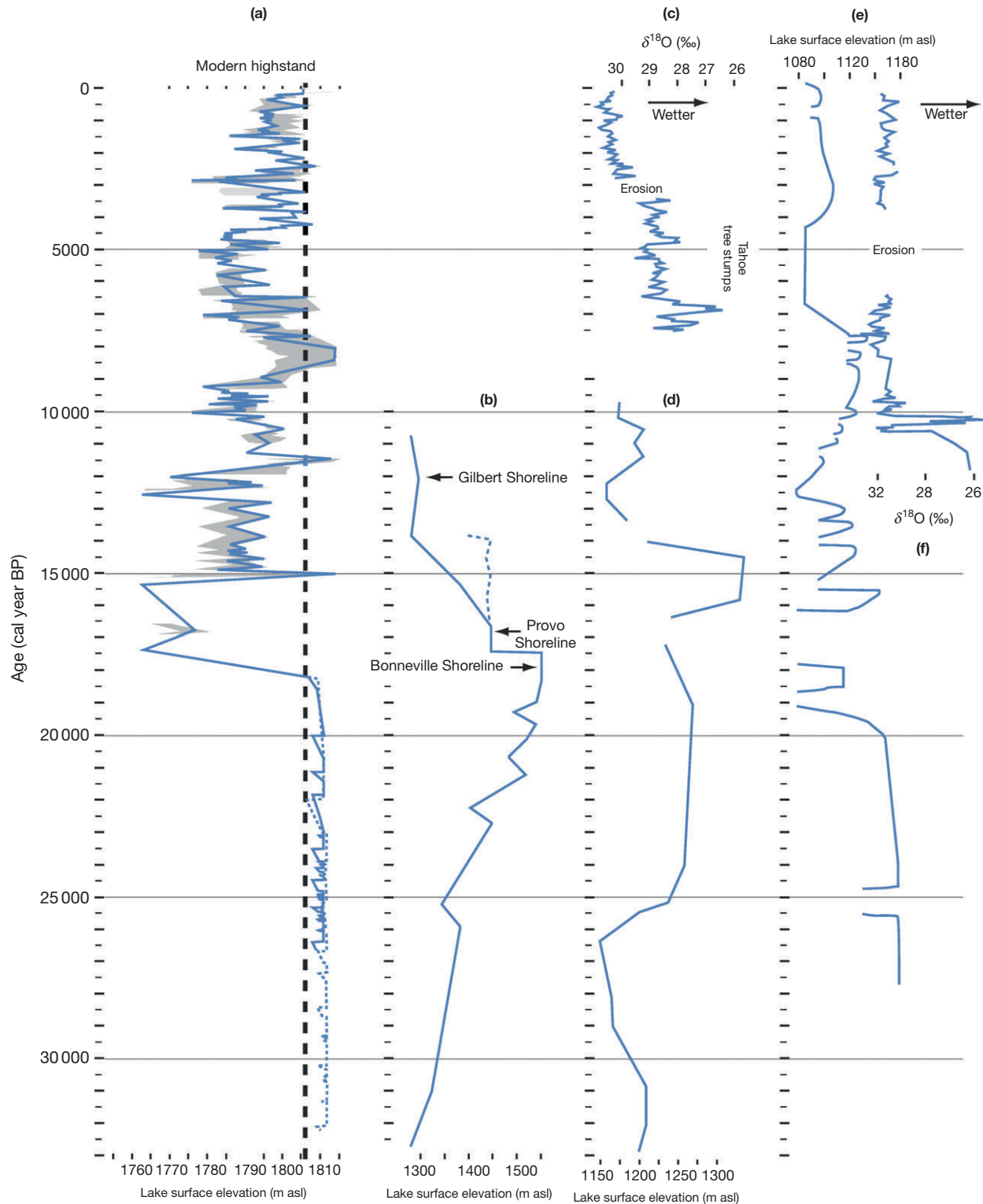


Figure 4 Lake-surface elevation history and moisture balance proxies for lakes in the Great Basin region. (a) Lake-surface elevations for Bear Lake, UT/ID. (b) Lake-surface elevations for Lake Bonneville, UT. (c) Oxygen-isotope measurements from sediments of Pyramid Lake, NV. (d) Lake-surface elevations for Pyramid Lake, NV. (e) Lake-surface elevations for Owens Lake, CA. (f) Oxygen-isotope measurements from sediments of Owens Lake, CA. Gray area shows range of possible interpretations. Dashed lines indicate alternative models. Modified from Smoot JP and Rosenbaum JG (2009) *Sedimentary constraints on late Quaternary lake-level fluctuations at Bear Lake, Utah and Idaho*. In: Rosenbaum JG and Kaufman DS (eds.) *Paleoenvironments of Bear Lake, Utah and Idaho and its Catchment*, pp.263–290. Boulder: Geological Society of America. Geological Society of America Special Paper 450.

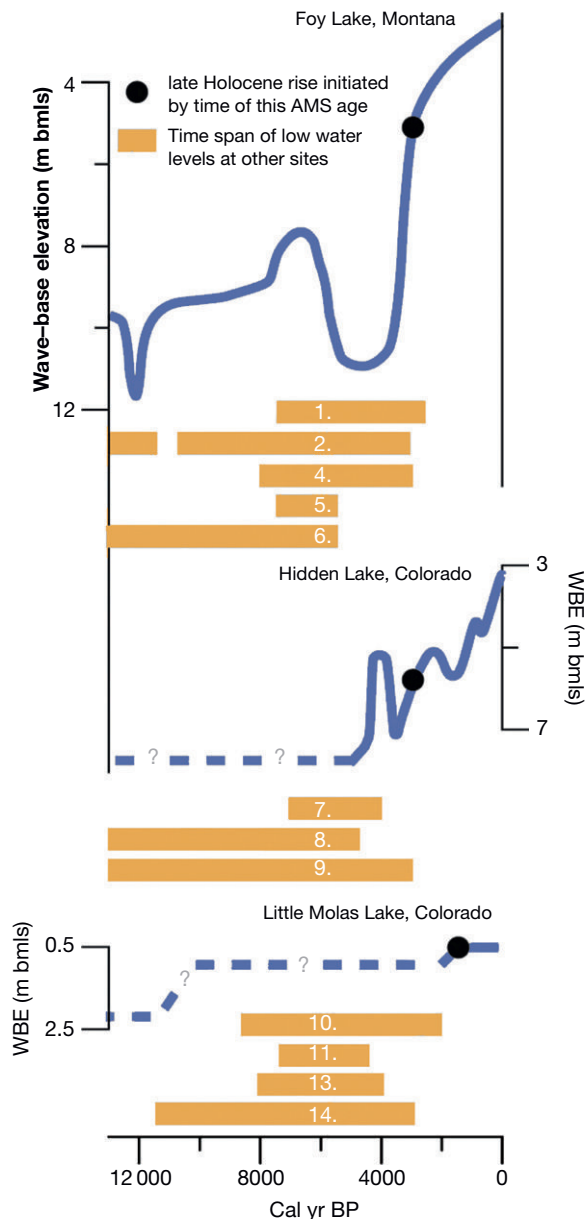


Figure 5 Summary of regional lake level for the Rocky Mountain region. Blue lines indicate inferred changes in the lake levels at Foy, Hidden, and Little Molas Lakes; dashed lines indicate when lower-than-modern levels are inferred but poorly constrained. Numbered bars indicate the timing of low-lake stands at other locations: (1) Flathead Lake, MT; (2) Jones Lake, MT; (4) Park Pond 2, WY; (5) Forest Pond 2, WY; (6) Park Pond 1, WY; (7) Cottonwood Pass Pond, CO; (8) Red Lady Fen, CO; (9) Red Well Fen, CO; (10) Bear Lake, AZ; (11) Fracas Lake, AZ; (13) Montazuma Well, AZ; (14) Potato Lake, AZ. Modified from Shuman B, Henderson A, Colman SM, et al. (2009) Holocene lake-level trends in the Rocky Mountains, USA. *Quaternary Science Reviews* 28: 1861–1879.

evidence for low lake levels during the mid-Holocene. Weng and Jackson (1999) report unconformities in cores from the center of Fracas and Bear lakes in northern Arizona. The unconformities indicate lake desiccation between 8.5 and 2 cal ka BP at Fracas Lake and between 7.4 and 4.23 cal ka BP at Bear

Lake. In Colorado, Cottonwood Pass Pond shows evidence of transition to a mire without open water from ca. 7 to 4 cal ka BP, and several nearby fens only formed in the last 4.5 cal ka BP (Fall, 1997).

Similar to the Great Plains and Great Basin, Holocene changes in summer insolation played an important role in the regional pattern of effective moisture balance in the Rocky Mountains. Particularly in the southern Rocky Mountains, summer precipitation rates in the early Holocene may have increased under the effect of high summer insolation on the southwestern 'monsoon' (Harrison et al., 2003). The small number of high-resolution (decadal to centennial scale) Holocene records of effective moisture balance from the northern and central Rocky Mountain regions (Anderson, 2011; Shapley et al., 2009; Stone and Fritz, 2006) also show higher-frequency (decadal to centennial) changes in lake level, probably driven by ocean-atmospheric dynamics expressed in climate mode indices, such as the Pacific Decadal Oscillation and the North Pacific Index (Barron and Anderson, 2011).

Alaska and the Yukon Territory

The semi-arid interior regions of Alaska and the Yukon have low precipitation rates compared to the relatively high rates of evaporative loss (Barber and Finney, 2000), thus lakes in the region have the potential to record significant changes in hydrologic balance. Lake-level fluctuations in this region have been inferred from sedimentary evidence of desiccation, changes in pollen stratigraphy, and more recently, multiple-core transect stratigraphy (e.g., Abbott et al., 2000, 2010; Anderson et al., 2005b; Bigelow and Edwards, 2001; Brahney et al., 2008; Mann et al., 2002).

Much of Alaska and the Yukon was not glaciated during the Quaternary. Long lacustrine records from Alaska commonly indicate that the last glacial period was characterized by a widespread decline in lake level, typically expressed as an extensive hiatus between 35 and 23 cal ka BP (Abbott et al., 2010). Lake levels started to rise between 23 and 17 cal ka BP, before a widespread increase in effective moisture, evidenced by substantial changes in vegetation, resulted in a regional lake-level rise by 14–15 cal ka BP (Abbott et al., 2000; Ager, 2003; Bigelow and Edwards, 2001; Edwards et al., 2001). Lakes fluctuated substantially during the late-Glacial period, with significant lowstands; some of these are closely associated with the Younger Dryas event (Mann et al., 2002) in the western part of Alaska. Pollen and lake records for the eastern interior of Alaska suggest a gradual increase in moisture from the LGM through the mid-Holocene, reaching near-modern levels by 6 cal ka BP (Edwards et al., 2001). Records throughout the western region of Alaska suggest that lakes remained lower than modern into the early to mid-Holocene (Figure 6), then rose to reach near-modern levels by 4 cal ka BP (Abbott et al., 2000; Anderson et al., 2005b; Barber and Finney, 2000).

As with other regions of North America, aridity during the early Holocene may be related to greater summer insolation (Abbott et al., 2000; Finney et al., 2005). Several recent studies have suggested that ocean-atmosphere linkages are further mediated by solar forcing and together these features have influenced effective moisture in the Alaska and Yukon region, particularly with respect to abrupt climate transitions since the

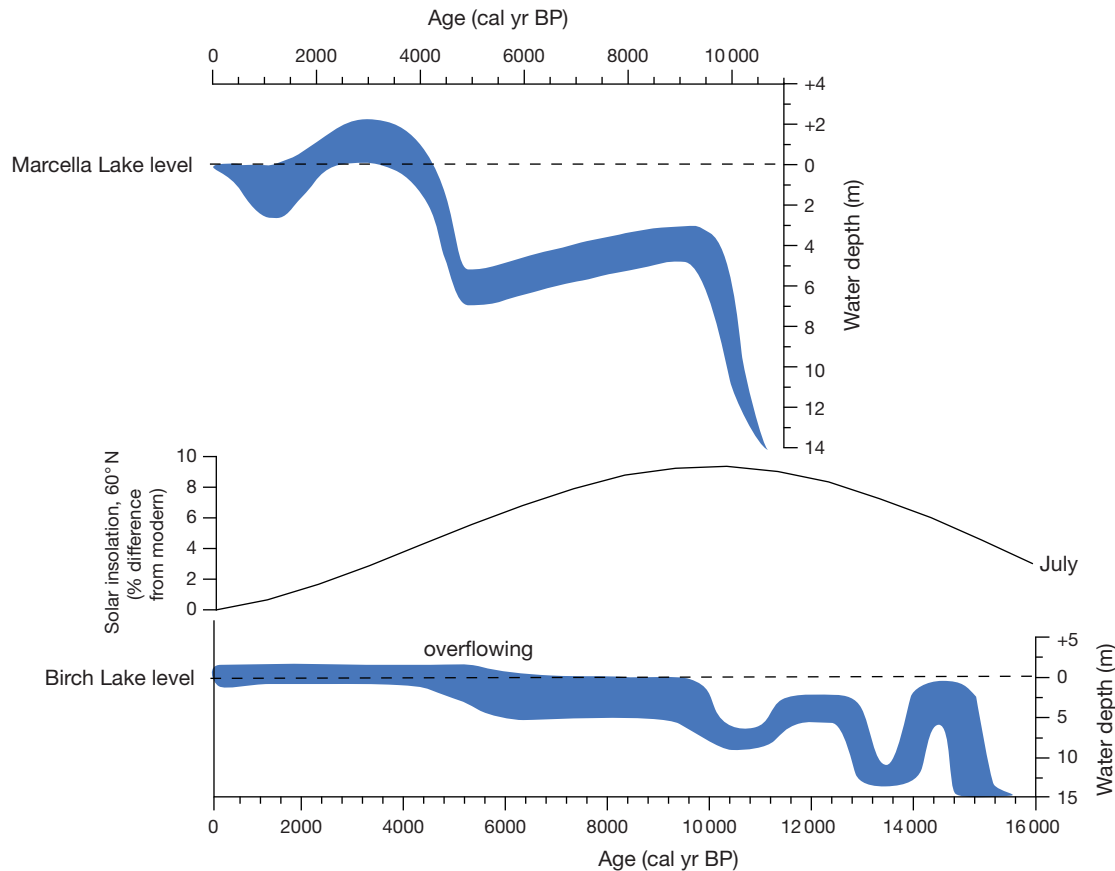


Figure 6 Comparisons of lake-level reconstructions for lakes in interior Alaska and the Yukon Territory. Lakes were low prior to ca. 15 cal ka BP, and rose with fluctuations during the late deglacial. An abrupt lake-level rise occurred in the early Holocene (9.5 cal ka BP), though lakes remained lower than modern until the mid-Holocene. Modified from Anderson L, Abbott MB, Finney BP, and Edwards ME (2005) Paleohydrology of the southwest Yukon Territory, Canada, based on multi-proxy analyses of lake sediment cores from a depth transect. *The Holocene* 15: 1–12; The Birch Lake, Alaska, record is from Abbott et al. (2000).

LGM (Hu et al., 2003, 2006). Increased continentality resulting from the large areas of shallow adjacent continental shelves exposed during lower sea levels has also been considered as a factor contributing to lower lake levels during this period (Abbott et al., 2000). During the deglacial to early Holocene transition, changes in atmospheric circulation related to effects of the Laurentide ice sheet may have been important (Abbott et al., 2000; Finney et al., 2005). The amount of precipitation reaching the interior is strongly affected by changes in winter storm tracks, primarily driven by the semi-permanent low pressure center known as the Aleutian Low. On decadal to millennial timescales, persistent shifts in the intensity and position of the Aleutian Low throughout the Holocene may have had a profound effect on lake levels (Anderson et al., 2005a; Clegg and Hu, 2010; Edwards et al., 2001; Kaufman et al., 2010).

Northeastern North America

Lakes in the northeastern part of the continent also were formed primarily by glaciation and contain records that span the terminal Pleistocene and Holocene periods. Lake-level reconstructions across the region, derived from sedimentological data, show considerable regional variation in temporal trends, reflecting the complex interplay of multiple large-scale

influences, including the northward advection of tropical moisture, westerly storm tracks, the Great Lakes, and changes in aquifer height linked to sea-level rise, coupled with more localized factors. Multiple sites from New York north to Quebec (Dwyer et al., 1996; Lavoie and Richard, 2000; Newby et al., 2000; Shuman et al., 2001, 2005) show low lake levels in the early Holocene, but with considerable variation in the onset and termination of dry conditions. In Quebec, lake level continued to be low until ~3 cal ka BP, except for a brief interval ~6 cal ka BP (Lavoie and Richard, 2000; Muller et al., 2003). In Maine, lake level also was low in the mid-Holocene (8–6 cal year BP) but began to rise to modern levels after ~6 cal ka BP (Almquist et al., 2001). The pattern of mid-Holocene lake level in sites immediately to the south in New Hampshire (Shuman et al., 2005), Massachusetts (Newby et al., 2000, 2011; Shuman et al., 2001), and New York (Dwyer et al., 1996) differ from those in Maine and Quebec, with high lake levels and inferred wet conditions through much of the mid-Holocene, followed by lake-level lowering between ~5.5 and 3 cal ka BP, spanning the period of the eastern hemlock decline. On the western edge of Lake Ontario (Yu and McAndrews, 1994; Yu et al., 1997), high lake levels in the early Holocene resemble the patterns in the Midwest, but the onset of mid-Holocene aridity does not occur until ~6 cal ka BP, like the sites directly to the east in

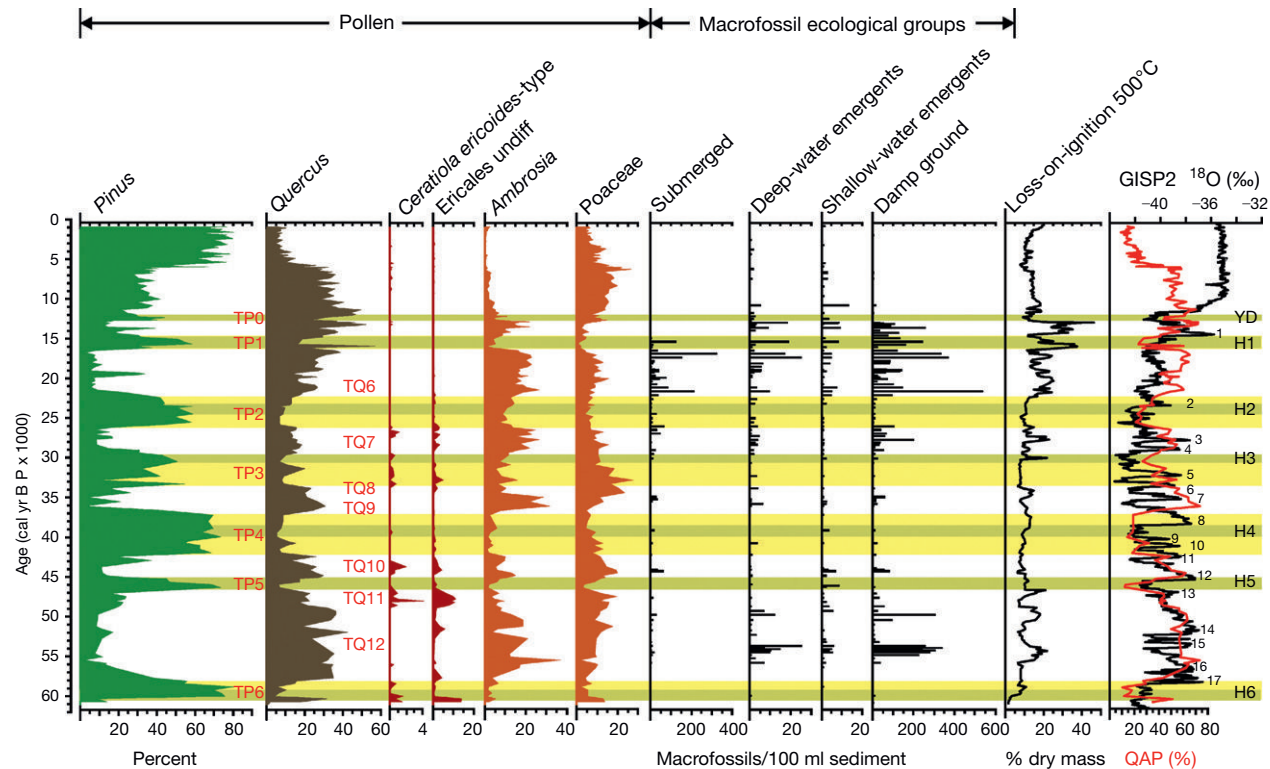


Figure 7 Summary pollen, macrofossil, and loss-on-ignition diagram for Lake Tulane, FL. The black curve (right) shows the GISP2 $\delta^{18}\text{O}$ curve; the red curve (right) is the sum of *Quercus*, *Ambrosia*, and *Poaceae* pollen. The interstadials are numbered to the right of the $\delta^{18}\text{O}$ graph. The darker bars indicate the ages and time spans of Heinrich events; the yellow bars indicate the Tulane *Pinus* phases. Reproduced from Grimm EC, Watts WA, Jacobson GL Jr., Hansen BCS, Almquist HR, and Diffenbacher-Krall AC (2006) Evidence for warm wet Heinrich events in Florida. *Quaternary Science Reviews* 25: 2197–2211.

southern New England. High lake levels, in some cases the highest of the Holocene, are characteristic of the last 3–2 cal ka BP in all of northeastern NA (Webb et al., 1993a).

The low lake levels and dry conditions in northeastern NA in the early Holocene are attributed to both the high summer insolation and the influence of the glacial anticyclone at the margin of the remnant Laurentide ice sheet (Webb et al., 1993a). As the ice sheet collapsed, the higher lake levels and moist conditions in southern New England may be a product of the reduced influence of the glacial anticyclone and the resultant enhanced northward advection of subtropical moisture (Shuman et al., 2002). Increased effective moisture and lake-level rise across the northeast in the late Holocene was associated with insolation-driven increases in winter precipitation and reductions in summer temperature; the drivers of the ~5.5–3 cal ka BP period of lowered lake levels from southeastern Ontario through southern New England are less clear, but may be related to higher frequency changes in Atlantic sea-surface temperatures that affected the seasonality of precipitation (Newby et al., 2011; Yu et al., 1997).

Southeastern United States

No studies in the southeastern United States have explicitly studied lake-level history, but numerous studies of fossil pollen data have produced evidence suggesting that lake levels have changed during the late Quaternary. The region was not

glaciated, and many lakes were either sinkholes or Carolina bays, which are circular depressions of uncertain origin. The southeastern US is the one of the most humid portions of North America, but many lakes – especially in Florida – show evidence of desiccation (e.g., Watts et al., 1992). Lake drying during glacial times likely reflects a lowered groundwater table associated with low sea level. Lake basins began to fill throughout the region in the early Holocene, probably because of the rising sea level and aquifer height, possibly augmented by increased precipitation (Shuman et al., 2002). The macrophyte and pollen evidence from Lake Tulane, Florida, which extends back 60 ka, also provides compelling evidence of earlier lake-level fluctuations driven by large-scale ocean–atmospheric teleconnections (Grimm et al., 1993, 2006), marked by alternating periods of domination by oak and pine (Figure 7). Sharp increases in fossil pine pollen abundances occurred during Heinrich events and the latter half of the Holocene. In periods dominated by oak forest, enhanced deposition of organic matter occurred and shallow emergent and damp-ground macrophytes were more abundant. This suggests very shallow water and a close proximity of the core site with the shoreline.

Continental-Scale Climate Variation

The prevailing pattern throughout North America, particularly west of the Great Lakes, was high lake levels during the late

Pleistocene and early Holocene, low lake levels during the mid-Holocene, and a return to relatively high levels more recently. These long-term trends have been attributed to two major factors. First, the presence of the Laurentide ice sheet dramatically influenced regional energy budgets and atmospheric circulation patterns prior to ca. 8 cal ka BP (COHMAP, 1988; Webb et al., 1993b). The ice sheet shifted the position of the polar jet stream southward, delivering high amounts of precipitation; cool conditions associated with the ice sheet also reduced evapotranspiration in neighboring regions. However, the southward-displaced jet, plus high summer insolation, left northwestern areas dry in the late Pleistocene and early Holocene. As the ice sheet collapsed, its climatic influence diminished – decreasing precipitation as storm tracks changed and increasing evapotranspiration as conditions warmed. Second, changes in insolation strongly influenced regional energy budgets as well as atmospheric circulation patterns. Insolation became the dominant control after the influence of the ice sheet waned, and probably contributed to the dry mid-Holocene conditions through heightened evaporation rates (Webb et al., 1993b) and through continental-scale atmospheric dynamics (e.g., Harrison et al., 2003). As orbital configuration changed, insolation levels have progressively reached the low seasonal contrast of today and conditions became progressively wetter than before. The influence of higher frequency ocean–atmosphere interactions on climate and lake-level variation, particularly variation associated with Pacific Ocean sea-surface temperatures, also are evident in the small number of high-resolution (decadal) studies from western North America.

Dry early Holocene conditions in eastern North America from New England to Florida have also been explained as the result of greater-than-modern summer insolation and the atmospheric influence of the Laurentide ice sheet (Shuman et al., 2001, 2002; Webb et al., 1993a). Moisture levels increased in the southeast and then in the northeast, as indicated by lake-level rises and plant species migrations, while conditions simultaneously dried in the mid-continent. As the ice sheet collapsed, high summer insolation probably increased the influence of the Bermuda subtropical high on eastern North America, causing the northward advection of subtropical moisture (Shuman et al., 2002), while simultaneously decreasing annual moisture balance in the mid-continent.

See also: **Diatom Methods:** Salinity and Climate Reconstructions from Continental Lakes. **Glacial Climates:** Volcanic and Solar Forcing. **Glaciations:** Late Quaternary in North America. **Lake Level Studies:** Modeling; Overview. **Pollen Records, Postglacial:** Northeastern North America; Northwestern North America; Southeastern North America; Southwestern North America.

References

- Abbott MB, Edwards ME, and Finney BP (2010) A 40,000-yr record of environmental change from Burial Lake in Northwest Alaska. *Quaternary Research* 74: 156–165.
- Abbott MB, Finney BP, Edwards ME, and Kelts KR (2000) Lake-level reconstructions and paleohydrology of Birch Lake, Central Alaska, based on seismic reflection profiles and core transects. *Quaternary Research* 53: 154–166.
- Adams KD (2007) Sedimentary environments and lake-level fluctuations at Walker Lake, Nevada. *Geological Society of America Bulletin* 119: 126–139.
- Ager T (2003) Late Quaternary vegetation and climate history of the central Bering land bridge from St. Michael Island, western Alaska. *Quaternary Research* 60: 19–32.
- Almquist H, Dieffenbacher-Krall AC, Flanagan-Brown R, and Sanger D (2001) The Holocene record of lake levels at Mansell Pond, central Maine, USA. *The Holocene* 11: 189–201.
- Anderson L (2011) Holocene record of precipitation seasonality from lake calcite $\delta^{18}\text{O}$ in the central Rocky Mountains, United States. *Geology* 39: 211–214.
- Anderson L, Abbott MB, Finney BP, and Burns S (2005a) Regional atmospheric circulation change in the North Pacific during the Holocene inferred from lacustrine carbonate oxygen isotopes, Yukon Territory, Canada. *Quaternary Research* 64: 21–35.
- Anderson L, Abbott MB, Finney BP, and Edwards ME (2005b) Paleohydrology of the southwest Yukon Territory, Canada, based on multi-proxy analyses of lake sediment cores from a depth transect. *The Holocene* 15: 1–12.
- Bacon SN, Burke RM, Pezzopane SK, and Jayko AS (2006) Last glacial maximum and Holocene lake levels of Owens Lake, eastern California, USA. *Quaternary Science Reviews* 25: 1264–1282.
- Balch DP, Cohen AS, Schnurrenberger DW, et al. (2005) Ecosystem and paleohydrological response to Quaternary climate change in the Bonneville Basin, Utah. *Palaeogeography, Palaeoclimatology, Palaeoecology* 221: 99–122.
- Barber VA and Finney BP (2000) Late Quaternary paleoclimatic reconstructions for interior Alaska based on paleolake-level data and hydrologic models. *Journal of Paleolimnology* 24: 29–41.
- Barron JA and Anderson L (2011) Enhanced Late Holocene ENSO/PDO expression along the margins of the eastern North Pacific. *Quaternary International* 235: 3–12.
- Benson LV (1978) Fluctuation in the level of pluvial Lake Lahontan during the last 40,000 years. *Quaternary Research* 9: 300–318.
- Benson LV (2003) Western lakes. In: Gillespie A, Porter S, and Atwater B (eds.) *Developments in Quaternary Science. The Quaternary Period in the United States*, vol. 1, pp. 185–204. Amsterdam: Elsevier.
- Benson LV, Lund S, Negrini R, Linsley B, and Zic M (2003) Response of North American Great Basin lakes to Dansgaard–Oeschger oscillations. *Quaternary Science Reviews* 22: 2239–2251.
- Benson LV and Paillet FL (1989) The use of total lake-surface area as an indicator of climatic change: Examples from the Lahontan Basin. *Quaternary Research* 32: 262–275.
- Benson LV and Thompson RS (1987) The physical record of lakes in the Great Basin. In: Ruddiman WF and Wright HE Jr. (eds.) *North America and Adjacent Oceans During the Last Glaciation*, vol. K-3, pp. 241–260. Boulder: Geological Society of America.
- Bigelow NH and Edwards ME (2001) A 14,000 yr paleoenvironmental record from Windmill Lake, Central Alaska: Lateglacial and Holocene vegetation in the Alaska Range. *Quaternary Science Reviews* 20: 203–215.
- Bird BW, Kirby ME, Howat IM, and Tulaczuk S (2010) Geophysical evidence for Holocene lake-level change in southern California (Dry Lake). *Boreas* 39: 131–144.
- Bradbury JP, Forester RM, and Thompson RS (1989) Late Quaternary paleolimnology of Walker Lake, Nevada. *Journal of Paleolimnology* 1: 249–267.
- Brahney J, Clague JJ, Menounos B, and Edwards TWD (2008) Timing and cause of water level fluctuations in Kluane Lake, Yukon Territory, over the past 5000 years. *Quaternary Research* 70: 213–227.
- Clark JA, Befus KM, and Sharman GR (2011) A model of surface water hydrology of the Great Lakes, North America during the past 16,000 years. *Physics and Chemistry of the Earth* <http://dx.doi.org/10.1016/j.pce.2010.12.005>.
- Clark PU, Marshall SJ, Clarke GKC, Hostetler SW, Licciardi JM, and Teller JT (2001) Freshwater forcing of abrupt climate change during the last glaciation. *Science* 293: 283–287.
- Clegg BF and Hu FS (2010) An oxygen-isotope record of Holocene climate change in the Brooks Range, Alaska. *Quaternary Science Reviews* 29: 928–939.
- Cohen AS, Palacios-Fest MR, Negrini RM, Wigand PE, and Erbes DB (2000) A paleoclimate record for the past 250,000 years from Summer Lake, Oregon, USA: II. Sedimentology, paleontology and geochemistry. *Journal of Paleolimnology* 24: 151–182.
- COHMAP Members (1988) Climatic changes of the last 18,000 years: Observations and model simulations. *Science* 241: 1043–1052.
- Colman SM, Forester RM, Reynolds RL, et al. (1994) Lake-level history of Lake Michigan for the past 12,000 years: The record from deep lacustrine sediments. *Journal of Great Lakes Research* 20: 73–92.
- Digerfeldt G, Almendinger JE, and Björck S (1993) Reconstruction of past lake levels and their relation to groundwater hydrology in the Parkers Prairie sandplain, west-central Minnesota. *Palaeogeography, Palaeoclimatology, Palaeoecology* 94: 99–118.
- Dwyer TR, Mullins HT, and Good SC (1996) Paleoclimatic implications of Holocene lake-level fluctuations, Owasco Lake, New York. *Geology* 24: 519–522.
- Edwards ME, Mock CJ, Finney BP, Barber V, and Bartlein PJ (2001) Potential analogues for paleoclimatic variations in eastern interior Alaska for the past 14,000 years:

- Atmospheric circulation controls of regional temperature and moisture responses. *Quaternary Science Reviews* 20: 189–202.
- Fall PL (1997) Timberline fluctuations and late Quaternary paleoclimates in the southern Rocky Mountains, Colorado. *Geological Society of America Bulletin* 109: 1306–1320.
- Filby SK, Locke SM, Person MA, et al. (2002) Mid-Holocene hydrologic model of the Shingobee Watershed, Minnesota. *Quaternary Research* 58: 246–254.
- Finney BP, Rühland KM, Smol JP, and Fallu MA (2005) Paleolimnology of the North American subarctic. In: Pienitz R, Douglas MSV, and Smol JP (eds.) *Long-Term Environmental Change in Arctic and Antarctic Lakes*, pp. 269–318. Dordrecht: Kluwer.
- Fisher TG, Yansa C, Lowell TV, Lepper K, Hajdas I, and Ashworth A (2008) The chronology, climate, and confusion of the Moorhead Phase of glacial Lake Agassiz: New results from the Ojata Beach, North Dakota, USA. *Quaternary Science Reviews* 27: 1124–1135.
- Fritz SC (2008) Deciphering climate history from lake sediments. *Journal of Paleolimnology* 39: 5–16.
- Fritz SC, Metcalfe SE, and Dean W (2001) Holocene climate patterns in the Americas inferred from paleolimnological records. In: Markgraf V (ed.) *Interhemispheric Climate Linkages*, pp. 241–263. San Diego: Academic Press.
- Gilbert GK (1890) Lake Bonneville. *US Geological Survey Monographs* 1: 1–275.
- Grimm EC, Jacobson GL Jr., Watts WA, Hansen BCS, and Maasch KA (1993) A 50,000-year record of climate oscillations from Florida and its temporal correlation with the Heinrich events. *Science* 261: 198–200.
- Grimm EC, Watts WA, Jacobson GL Jr., Hansen BCS, Almquist HR, and Diffenbacher-Krall AC (2006) Evidence for warm wet Heinrich events in Florida. *Quaternary Science Reviews* 25: 2197–2211.
- Harrison SP and Digerfeldt G (1993) European lakes as palaeohydrological and palaeoclimatic indicators. *Quaternary Science Reviews* 12: 233–248.
- Harrison SP, Kutzbach JE, Liu Z, et al. (2003) Mid-Holocene climates of the Americas: A dynamical response to changed seasonality. *Climate Dynamics* 20: 663–688.
- Harrison SP and Metcalfe SE (1985) Variations in lake levels during the Holocene in North America: An indication of changes in atmospheric circulation patterns. *Geographie physique et Quaternaire* 39: 141–150.
- Hofman MH, Hendrix MS, Moore JN, and Sperazza M (2006) Late Pleistocene and Holocene depositional history of sediments in Flathead Lake, Montana: Evidence from high-resolution seismic reflection interpretation. *Sedimentary Geology* 184: 111–131.
- Hu FS, Kaufman DS, Yoneji S, et al. (2003) Cyclic variation and solar forcing of Holocene climate in the Alaskan subarctic. *Science* 301: 1890–1893.
- Hu FS, Nelson DM, Clarke GH, et al. (2006) Abrupt climatic events during the last glacial–interglacial transition in Alaska. *Geophysical Research Letters* 33: L18708.
- Kaufman DS, Anderson RS, Hu FS, Berg E, and Werner A (2010) Evidence for a variable and wet Younger Dryas in southern Alaska. *Quaternary Science Reviews* 29: 1445–1452.
- Kaufman DS, Bright J, Dean WE, et al. (2009) A quarter-million years of paleoenvironmental change at Bear Lake, Utah and Idaho. In: Rosenbaum JG and Kaufman DS (eds.) *Paleoenvironments of Bear Lake, Utah and Idaho and its Catchment*, pp. 311–351. Boulder: Geological Society of America Geological Society of America Special Paper 450.
- Kirby ME, Lund SP, Patterson WP, et al. (2010) A Holocene record of Pacific Decadal Oscillation (PDO)-related hydrologic variability in Southern California (Lake Elsinore, CA). *Journal of Paleolimnology* 44: 819–839.
- Kowalewska A and Cohen AS (1998) Reconstruction of paleoenvironments of the Great Salt Lake Basin during the late Cenozoic. *Journal of Paleolimnology* 20: 381–407.
- Laird KR, Fritz SC, Grimm EC, and Mueller PG (1996) Century-scale paleoclimatic reconstruction from Moon Lake, a closed-basin lake in the northern Great Plains. *Limnology and Oceanography* 41: 890–902.
- Lavoie M and Richard JH (2000) Postglacial water-level changes of a small lake in southern Quebec, Canada. *The Holocene* 10: 621–634.
- Li HC, Xu XM, Ku TL, You CF, Buchheim HP, and Peters R (2008) Isotopic and geochemical evidence of paleoclimate changes in Salton Basin, California, during the past 20 kyr: 1. $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ records in lake tufa deposits. *Palaeogeography, Palaeoclimatology, Palaeoecology* 259: 182–197.
- Loope DB, Swinehart JB, and Mason JP (1995) Dune-dammed paleovalleys of the Nebraska Sand Hills: Intrinsic versus climatic controls on the accumulation of lake and marsh sediments. *Geological Society of America Bulletin* 107: 396–406.
- Mann DH, Peteet DM, Reanier RE, and Kunz ML (2002) Responses of an Arctic landscape to Lateglacial and early Holocene climatic changes: The importance of moisture. *Quaternary Science Reviews* 21: 997–1021.
- Mason JP, Swinehart JB, and Loope DB (1997) Holocene history of lacustrine and marsh sediments in a dune-blocked drainage, Southwestern Nebraska Sand Hills, USA. *Journal of Paleolimnology* 17: 67–83.
- Muller S, Richard PJH, Guiot J, de Beaulieu JL, and Fortin D (2003) Postglacial climate in the St. Lawrence lowlands, southern Quebec: Pollen and lake-level evidence. *Palaeogeography, Palaeoclimatology, Palaeoecology* 193: 51–72.
- Murton JB, Bateman MD, Dallimore SR, Teller JT, and Yang Z (2010) Identification of Younger Dryas outburst flood path from Lake Agassiz to the Arctic Ocean. *Nature* 464: 740–742.
- Negrini RM, Wigand PE, Draucker S, et al. (2006) The Rambla highstand shoreline and the Holocene lake-level history of Tulare Lake, California, USA. *Quaternary Science Reviews* 25: 1599–1618.
- Newby PE, Killoran P, Waldorf MR, Shuman BN, Webb RS, and Webb T (2000) 14,000 years of sediment, vegetation, and water-level changes at the Makepeace Cedar Swamp, southeastern Massachusetts. *Quaternary Research* 53: 352–368.
- Newby P, Shuman BN, Donnelly JP, and MacDonald D (2011) Repeated century-scale droughts over the past 13,000 yr near the Hudson River watershed, USA. *Quaternary Research* 75: 523–530.
- Oviatt CG (1997) Lake Bonneville fluctuations and global climate change. *Geology* 25: 155–158.
- Oviatt CG, Currey DR, and Sack D (1992) Radiocarbon chronology of Lake Bonneville, eastern Great Basin, USA. *Paleogeography, Paleoclimatology, Paleocology* 99: 225–241.
- Pardee JT (1910) *The Glacial Lake Missoula*. Chicago: University of Chicago Press.
- Patrickson SJ, Sack D, Brunelle AR, and Moser KA (2010) Late Pleistocene to early Holocene lake level and paleoclimate insights from Stansbury Island, Bonneville basin, Utah. *Quaternary Research* 73: 237–246.
- Peterson CD, Minor R, Peterson GL, and Gates EB (2011) Pre- and post-Missoula flood geomorphology of the pre-Holocene ancestral Columbia River Valley in the Portland forearc basin, Oregon and Washington, USA. *Geomorphology* 129: 276–293.
- Russell IC (1885) Geology of Lake Lahontan: A quaternary Lake of Northern Nevada. *US Geological Survey Monographs* 11: 288.
- Schwab A and Dean WE (1998) Stable isotopes and sediments from Pickerel Lake, South Dakota, USA: A 12 ky record of environmental changes. *Journal of Paleolimnology* 20: 15–30.
- Shapley MD, Ito E, and Donovan JJ (2009) Lateglacial and Holocene hydroclimate inferred from a groundwater flow-through lake, Northern Rocky Mountains, USA. *The Holocene* 19: 523–535.
- Shuman B, Bartlein PJ, Logar N, Newby P, and Webb TW (2002) Parallel climate and vegetation responses to the early Holocene collapse of the Laurentide Ice Sheet. *Quaternary Science Reviews* 21: 1793–1805.
- Shuman B, Bravo J, Kaye J, Lynch JA, Newby P, and Webb T (2001) Late Quaternary water-level variations and vegetation history at Crooked Pond, Southeastern Massachusetts. *Quaternary Research* 56: 401–410.
- Shuman B, Henderson A, Colman SM, et al. (2009) Holocene lake-level trends in the Rocky Mountains, USA. *Quaternary Science Reviews* 28: 1861–1879.
- Shuman B, Newby P, Donnelly JP, Tarbox A, and Webb T (2005) A record of late-Quaternary moisture balance change and vegetation response in the White Mountains, New Hampshire. *Annals of the Association of American Geographers* 95: 237–248.
- Smith AJ, Donovan JJ, Ito E, Engstrom DR, and Panek VA (2002) Climate-driven hydrologic transients in lake sediment records: Multiproxy record of mid-Holocene drought. *Quaternary Science Reviews* 21: 625–646.
- Smith GI and Street-Perrot FA (1983) Pluvial lakes of the western United States. In: Wright HE Jr. (ed.) *Late-Quaternary Environments of the United States*, pp. 190–212. Minneapolis: University of Minnesota Press.
- Smoot JP and Rosenbaum JG (2009) Sedimentary constraints on late Quaternary lake-level fluctuations at Bear Lake, Utah and Idaho. In: Rosenbaum JG and Kaufman DS (eds.) *Paleoenvironments of Bear Lake, Utah and Idaho and its Catchment*, pp. 263–290. Boulder: Geological Society of America Geological Society of America Special Paper 450.
- Stevens LR, Stone JR, Campbell J, and Fritz SC (2006) A 2200-yr record of hydrologic variability from Foy Lake, MT inferred from diatom and geochemical data. *Quaternary Research* 65: 264–274.
- Stine S (1990) Late Holocene fluctuations of Mono lake, eastern California. *Palaeogeography, Palaeoclimatology, Palaeoecology* 78: 333–381.
- Stine S (1994) Extreme and persistent drought in California and Patagonia during medieval time. *Nature* 369: 546–549.
- Stone JR and Fritz SC (2004) Three-dimensional modeling of lacustrine diatom habitat areas: Improving paleolimnological interpretation of planktic:benthic ratios. *Limnology and Oceanography* 49: 1540–1548.
- Stone JR and Fritz SC (2006) Multi-decadal drought and Holocene climate instability in the Rocky Mountains. *Geology* 34: 409–412.
- Teller JT, Leverington DW, and Mann JD (2002) Freshwater outburst to the oceans from glacial Lake Agassiz and their role in climate change during the last deglaciation. *Quaternary Science Reviews* 21: 879–887.

- Walker IR and Pellatt MG (2008) Climate change and ecosystem response in the northern Columbia River basin — A paleoenvironmental perspective. *Environmental Reviews* 16: 113–140.
- Watts WA, Hansen BCS, and Grimm EC (1992) Camel Lake: A 40,000-yr record of vegetational and forest history from northwest Florida. *Ecology* 73: 1056–1066.
- Webb RS, Anderson KH, and Webb T (1993a) Pollen response-surface estimates of late-Quaternary changes in the moisture balance of the northeastern United States. *Quaternary Research* 40: 213–227.
- Webb T, Bartlein PJ, Harrison SP, and Anderson KH (1993b) Vegetation, lake levels, and climate in eastern North America for the past 18,000 years. In: Wright HE Jr. (ed.) *Global Climates Since the Last Glacial Maximum*, pp. 415–467. Minneapolis: University of Minnesota Press.
- Weng C and Jackson ST (1999) Late-glacial and Holocene vegetation and climate history of the Kaibab Plateau, northern Arizona. *Palaeogeography, Palaeoclimatology, Palaeoecology* 153: 179–201.
- Wilcox DA, Thompson TA, Booth RK, and Nicholas JR (2007) Lake level variability and water level variability in the Great Lakes. *US Geological Survey Circular* 1311: 1–25.
- Williams JW, Shuman B, Bartlein PJ, Diffenbach NS, and Webb T (2010) Rapid, time-transgressive, and variable responses to early Holocene mid-continental drying in North America. *Geology* 38: 135–138.
- Winkler M, Swain A, and Kutzbach J (1986) Middle Holocene dry period in the northern Midwestern United States, lake levels and pollen stratigraphy. *Quaternary Research* 25: 235–250.
- Yu Z and McAndrews JH (1994) Holocene water levels at Rice Lake, Ontario, Canada: Sediment, pollen and plant-macrofossil evidence. *The Holocene* 4: 141–152.
- Yu Z, McAndrews JH, and Eicher U (1997) Middle Holocene dry climate caused by change in atmospheric circulation patterns: Evidence from lake levels and stable isotopes. *Geology* 25: 251–254.

Relevant Website

<http://goo.gl/maps/pwzYY> — Lake-level study sites for North America.

Introduction

The water cycle is a crucial factor in climate change; variations in the hydrological cycle are a key issue in the perspective of present global warming and concerns over water resources availability. Most data collected to reconstruct past Holocene climatic oscillations preferentially document past changes in temperature. In addition to the reconstruction of past variations in fluvial discharge and water table in mires, lake basins provide interesting archives since the lake-level records make it possible to document past changes in the water budget in relation to climatic variations. After the pioneering investigations by T. Nilsson and G. Digerfeldt in Sweden, lake-level studies developed in Europe against the backdrop of (1) the IGCP Project 158B (Berglund et al., 1983; Gaillard, 1985; Ralska-Jasiewiczowa and Starkel, 1988), (2) the construction of a European Lake Status Database (Harrison et al., 1998; Yu and Harrison, 1995), and (3) the archaeological controversies on the prehistoric lake dwellings in the sub-Alpine area (Ammann, 1982; Jacomet, 1985; Lüdi, 1935; Magny, 2004a). This article presents (1) the method used to reconstruct past lake-level changes and (2) a synthesis of results obtained over the last two decades in west-central Europe to establish a regional pattern of Holocene lake-level changes. Finally, it examines their paleoclimatic significance in terms of atmospheric circulation, climatic parameters, and forcing factors.

Study Area and Method of Lake-Level Reconstruction

The study area is composed of the Jura Mountains, the northern French Pre-Alps, and the Swiss Plateau (Figure 1). This region in west-central Europe is well placed for the study of changes in atmospheric circulation patterns, particularly the position of the Atlantic Westerly Jet, which determines the latitudinal extension of the Hadley cell and the midlatitude storm tracks.

The lake-level data presented here are a compilation of records obtained from 26 lakes (Figure 1). Lake levels are influenced by climatic parameters affecting both evaporation and precipitation, but they can also be induced by a variety of local, nonclimatic factors such as geomorphological and hydrographic phenomena. In Europe, another factor may be human impact on the vegetation cover in the catchment, which could have occurred relatively early, that is, ca. 8.5–8 ka BP. The replication of records from the same area, although time-consuming, allows one to observe synchronous changes in several lakes within a region and to testify to their climatic origin (Digerfeldt, 1988; Magny, 2004b). Most of the lakes studied are characterized by a nivo-pluvial regime (highstands during autumn and spring due to rain and/or snow melting); some have a small proportion of their water furnished by glacier meltwater in summer.

Since the 1980s, different methods have been used to reconstruct past variations in lake levels in Europe, that is, stable isotope analyses (Hammarlund et al., 2003; Valero-Garcés et al., 1998), changes in assemblages of mollusks (Clerc et al., 1989), diatoms and cladocera (Hyvärinen and Alhonen, 1994; Korhola et al., 2005), and plant macrofossils (Digerfeldt, 1988; Hannon and Gaillard, 1997). Archaeological data also provide evidence of past lake levels, though this is often questionable due to uncertain interpretations of architectural remains, capacities of past societies to adapt, and their ability to alter environmental conditions (Menotti, 2004). Lake-level records based on natural systems are better. The lake-level data presented here were obtained using a specific sedimentological method based on multiples lines of evidence, as follows:

- *Changes in the sediment structure.* Coarser deposits correspond to near-shore areas (shallower water and higher hydrodynamism).
- *Changes in the lithology.* Organic deposits (detritus gyttja, peat) often characterize shallow water or late stages of infilling (shallow residual basins). Lake marl is found in deeper water.
- *Changes in the assemblages of carbonate concretions.* The coarser fractions (>0.2 mm) of lake marl consist mainly of various carbonate concretion morphotypes of biochemical origin from bacterial (Cyanophyceae) or algal (Characeae; Haas, 1994) activity and from phenomena associated with the photosynthetic activity of aquatic vegetation. Modern analogs from ten regional lakes (Magny, 1998) highlight that each morphotype shows a specific spatial-depth distribution from the shore to the extremity of the littoral platform (Figure 2; see Overview), with the successive domination of oncolithes (near-shore areas with high-energy environment), cauliflower-like forms (littoral platform), plate-like concretions (formed on leaves of aquatic plants from the Potamogeto-Nupharetum belt), and finally tube-like concretions (stem encrustations from the Characeae belt of the platform slope). Thus, where the carbonate status of the lake allows, this sedimentological approach is similar to that based on plant macrofossils (Digerfeldt, 1988) but offers the advantage of requiring relatively small samples (~1–2 cm³).
- *Sediment hiatuses.* These mark either erosion or nondeposition. They result from a lowering of the sediment limit associated with a lowering of lake level (see Overview) for an explanation of the sediment limit). Sediment hiatuses are marked by unconformities between sediment layers (erosion surfaces; Digerfeldt, 1988), and they may be identified on the basis of observations along core transects perpendicular to the shore.
- *Contiguous high-resolution subsampling of sediment sequences identifies short-lived paleohydrological events and abrupt changes in lake level.*

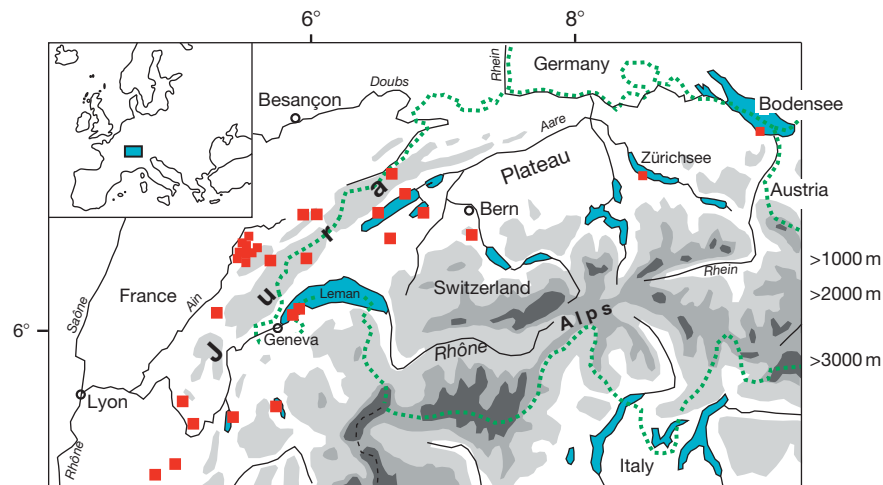


Figure 1 Geographical location of the study sites.

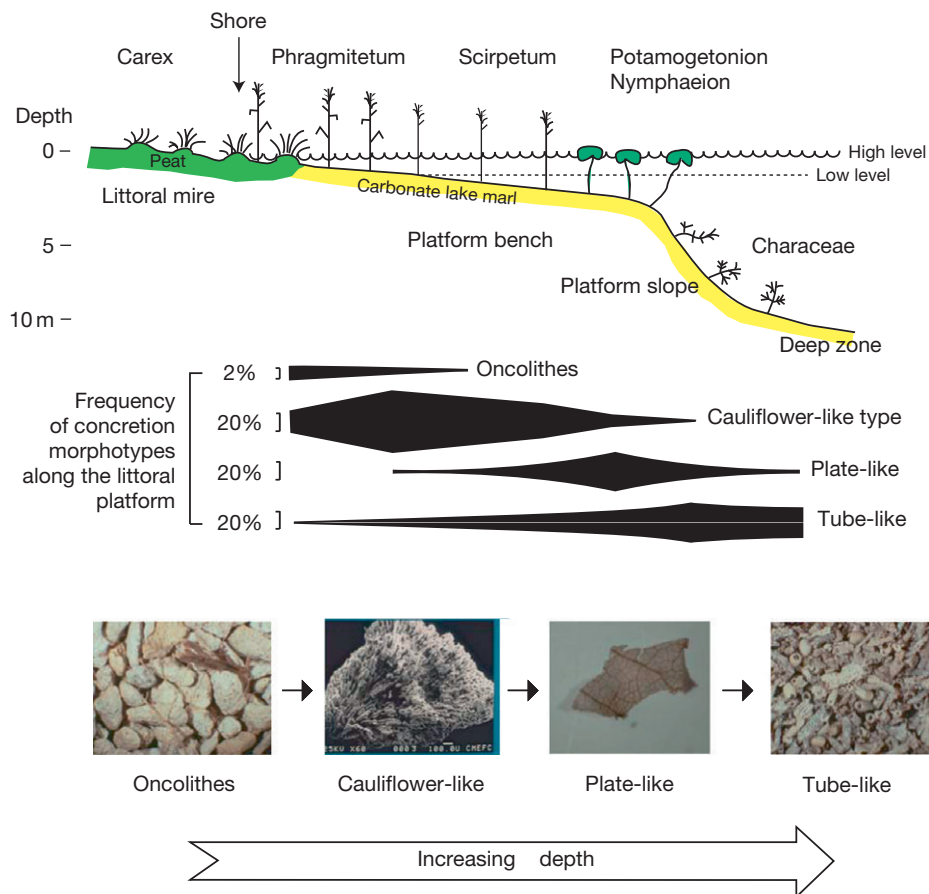


Figure 2 Section showing the distribution of carbonate concretion morphotypes along the littoral platform of Lake Clairvaux (Jura Mountains) in relation to the morphology and the vegetation zonation. The relative frequency of each morphotype is indicated as a percentage of the sample components.

The chronology of the lake-level fluctuations reconstructed from sediment analyses is based on three methods:

- **Radiocarbon dates.** This method uses terrestrial macrofossils to avoid errors due to the hard-water effect.
- **Tree-ring dates.** The Holocene sediment sequences of many lakes in the European sub-Alpine area include anthropogenic organic layers interbedded in lacustrine deposits (Menotti, 2004). These layers consist of vegetal remains

accumulated by inhabitants of Neolithic and Bronze Age lake dwellings. Thanks to permanent humid conditions, wooden posts used by prehistoric people for house construction are well preserved in these lake-shore archaeological sites, and they provide precise tree-ring dates for lake-level changes reconstructed from sediment sequences sampled on the archaeological sites (see [Dendrochronology](#)). Thus, in the sub-Alpine area the period 6–2.8 cal ka BP is well, and precisely, documented thanks to the remains of Neolithic and Bronze Age lacustrine villages.

- *Archaeological dates.* Additional chronological control with a time resolution of ca. 50 years comes from artifacts found on archaeological sites (or related to other sites) dated by the tree-ring method, or identified with reference to already well-dated historical material (such as Roman amphoras).

The Regional Pattern of Holocene Lake-Level Fluctuations

Theoretically, lake-level records obtained from 26 sites in the same region should allow the construction of a robust regional pattern of paleohydrological changes for the Holocene. However, as more data become available, problems arise in interpreting and reconciling multiple records, most often due to uncertainties in the chronology. Sediment sequences may be poorly dated, or even when better dated, key ages may be derived by extrapolation between two radiocarbon-dated horizons assuming a regular sedimentation rate. The synthesis presented in [Figure 3](#) was constructed on the basis of more than 180 radiocarbon, tree-ring, and archaeological dates that gave the ages of the sequence of higher and lower lake-level events ([Magny, 2004b](#)).

The two histograms displayed in [Figure 3](#) show the distribution of the collected dates over the whole Holocene for the lower lake-level episodes versus the higher lake-level ones, with a time resolution of 50 years. A clear disparity appears between the two histograms: the lower lake-level episodes are better documented than the higher ones. This difference directly reflects changes in the lithology. Low lake levels often correspond to an extension of peat and organic detritus deposition in the near-shore areas, thus providing a source of datable material, whereas high stands are characterized by the accumulation of more minerogenic sediments. In addition, the archaeological sites often provide well-dated sediment sequences because of the potential for tree-ring dating. This reliance on archaeological dates may also partly explain that even the lower lake-level episodes during the first half of the Holocene appear to be more weakly documented than those of the later Holocene. This disparity also reflects the fact that early Holocene deposits often have a more offshore and deeper stratigraphic position in a lake than younger deposits. It should be borne in mind, therefore, that the patterns in both histograms are not related to the magnitude of the recorded events, but rather reflect the present state of documentation, and possibly the duration of the events (in that longer lasting events are more likely to leave an imprint).

One may observe in [Figure 3](#) that the collected dates are not distributed randomly over the Holocene, but that they cluster in successive distinct time windows. This suggests an

alternation of lake-level phases. Due to the standard deviation of the radiocarbon dates, calibration time windows of ages obtained for low and high lake levels tend to partially overlap. Moreover, except when using the ^{14}C -wiggle-matching method (see [Calibration of the \$^{14}\text{C}\$ Record](#)), radiocarbon dating does not have the precision to pinpoint the age and the duration of a short-lived event. A possible solution to overcome these difficulties was to focus on 1-sigma calibration time windows (and tree-ring dates when available). Thus, keeping in mind the limitations discussed above, 15 successive episodes of higher lake level may be identified at 11 250–11 050, 10 300–10 000, 9550–9150, 8300–8050, 7550–7250, 6350–5900, 5650–5200, 4850–4800, 4150–3800, 3500–3100, 2750–2350, 1800–1700, 1300–1100, 750–650 cal years BP, and after ca. 450 cal years BP.

Contrasting Patterns of Hydrological Changes in Europe

As pointed out previously ([Magny, 1993b](#); [Magny et al., 2002](#)), the pattern of alternative phases of lower and higher lake level observed from the data set presented in [Figure 3](#) shows a general agreement with other sequences of environmental and climatic changes established in the mid-European latitudes. The major phases of high lake level in west-central Europe coincided with glacier advance and timberline declines in the Alps and with increasing discharge in the Upper Vistula (Poland) and in the middle Durance basin (southern French Alps; [Magny, 1993b](#)). This convergence among paleoenvironmental records implies the mid-European lake-level record has climatic significance, and the records suggest continual variability over the Holocene.

A comparison of the hydrological signal given by lake levels in west-central Europe with other hydrological data obtained in Europe for the Holocene suggests that there were contrasting patterns of hydrological changes in Europe in response to known cooling phases ([Magny et al., 2003](#)). As an example, [Figure 4](#) shows the general situation in Western Europe for two distinct climatic reversals: the Preboreal oscillation (PBO) at ca. 11.2 cal years BP and the 8.2 ka event. The dry conditions to the north and south probably reflect weaker evaporation due to lower sea-surface temperatures. Furthermore, in the north a southward extension of perennial sea-ice coverage in the Nordic seas would have promoted drier conditions. The resulting stronger thermal gradient between high and low latitudes and southward displacement of the Atlantic westerly jet would have reinforced cyclonic activity over mid-Europe, leading to increasing moisture in this region.

If the maps established for the PBO and the 8.2 ka event are compared, it can be seen that during the PBO, the middle zone marked by wetter conditions appears to have been larger, extending further northward than during the 8.2 ka event. This may reflect a weaker, more meandering Atlantic Westerly Jet in response to a weaker thermal gradient between high and low latitudes. If this were the case, the climate reversal was less pronounced in Europe during the PBO than during the 8.2 ka event. A thorough understanding of the mechanisms behind these hydrological contrasts requires further investigation,

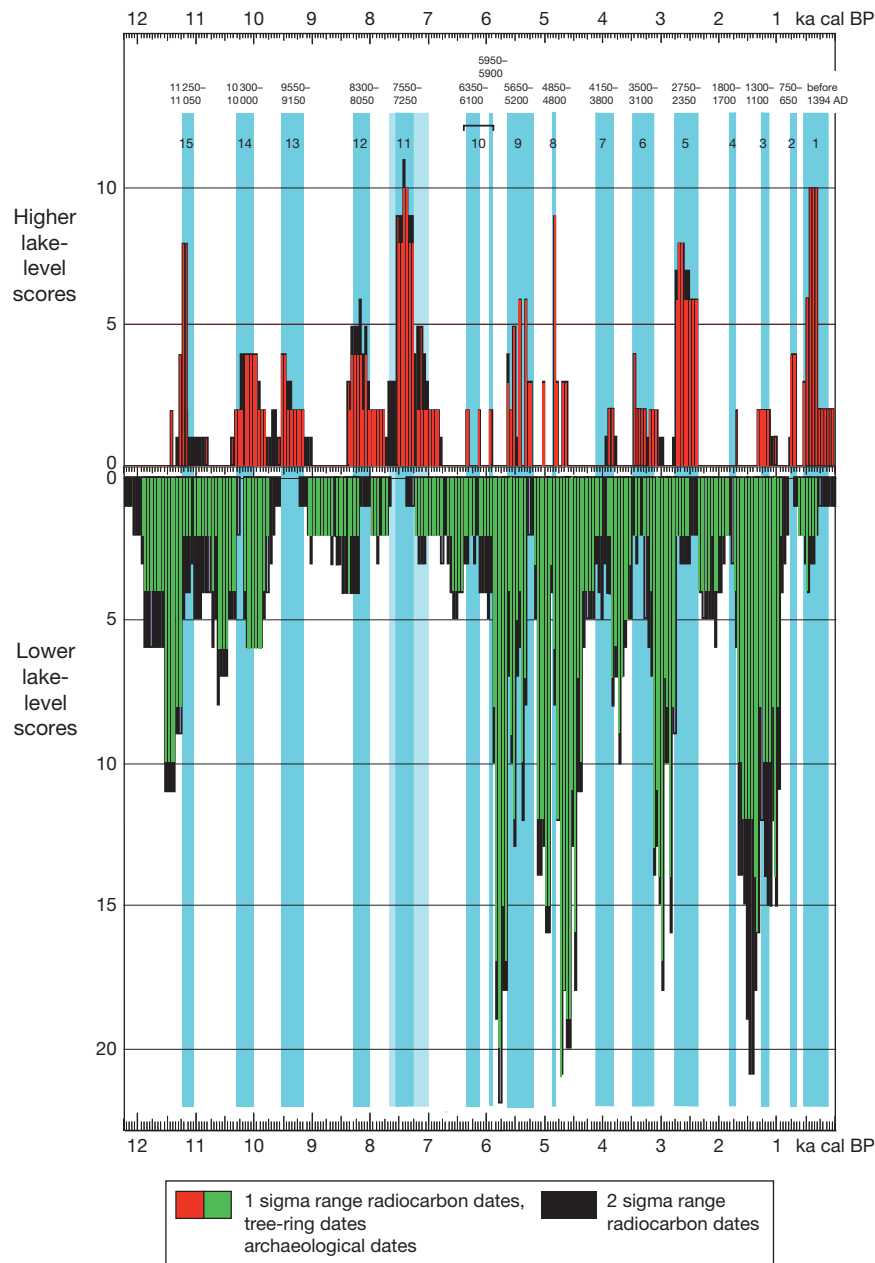


Figure 3 Distribution of the dates of higher and lower lake-level events reconstructed in the Jura Mountains, the northern French Pre-Alps and the Swiss Plateau over the Holocene period. After calibration (Stuiver et al., 1998), the dates were divided into two groups, that is, lower versus higher lake-level events, and placed on the timescale to construct two distinct histograms as follows. The temporal resolution chosen to construct the histograms was successive 50-year periods depicted as bars from 12 ka BP to present day. Every time a 50-year period was documented in a 1 or 2 sigma time window defined by the calibration of radiocarbon dates listed, it was given 2 or 1 point(s) respectively. Hence, the length of bars is proportional to the total of points accumulated by each 50-year period. Parts of the bars colored in red and green correspond with ages documented in 1 sigma calibration time windows; those in black correspond with ages documented in 2 sigma calibration time windows.

including the documentation of the role of seasonality (e.g., Hammarlund et al., 2005).

Variations in Climatic Variables Reflected by Changes in Lake Level

The quantitative reconstruction of climatic oscillations punctuating the present interglacial is one of the major challenges in paleoclimatic studies. These phenomena present a challenge to

climate modeling, and they are central to understanding the possible impact of climate change on ancient agricultural societies. Until now, most paleoclimatic data collected in Europe relate to paleotemperature, whereas reconstructions of paleoprecipitation are still uncommon. Pollen data offer the advantage of furnishing information not only on temperature but also on precipitation. However, problems arise when reconstructing past precipitation because, in Europe, the moisture of the growing season is rarely the main limiting factor for

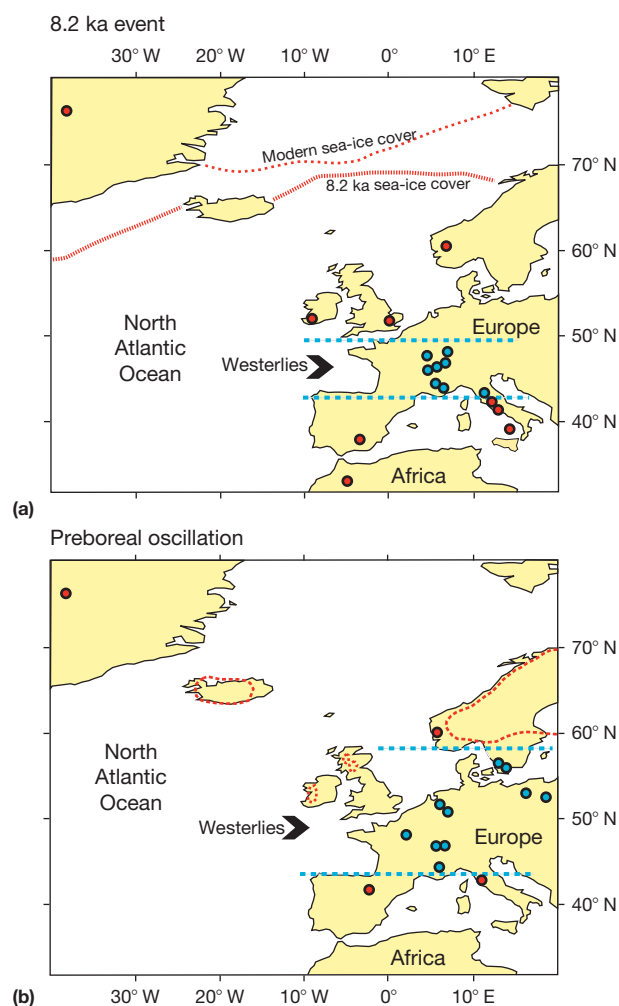


Figure 4 Comparison of hydrological signals associated with the 8.2 ka event (a), and the PBO (b) in Western Europe. Blue broken lines: northern and southern limits of the midlatitude zone characterized by wetter climatic conditions.

vegetation on a regional scale, and modern analogs tend to span a wide range of precipitation estimates. As pointed out by Guiot et al. (1993), lake-level records can be used to refine the paleoclimatic reconstructions based only on pollen data.

This approach has been extensively described in Magny et al. (2001) (see also Approaches). The principle is to find, for each fossil pollen assemblage, several similar pollen spectra (modern analogs). The degree of similarity (or distance) is estimated between fossil assemblages and spectra from a modern reference pollen database (today including more than 1300 surface-pollen samples from Europe), among which eight to ten modern analogs are selected. The climate of these analogs is averaged to provide an estimate of the fossil assemblage climate. An additional constraint is provided by the lake-level reconstruction. Analogs giving a climate that was not compatible with the lake-level status were rejected. The climate reconstruction is therefore based only on analogs that are coherent with lake levels.

As an example, Figure 5 presents results obtained for the early and mid-Holocene periods from two sites: Lake Le Locle in the Jura Mountains and Montilier, on the south shore of

Lake Morat on the Swiss Plateau (Magny et al., 2001, 2005). Climatic parameters reconstructed here suggest that phases of higher lake level coincided with an increase in annual precipitation, a decrease in summer temperature and a shortening of the growing season. Conversely, periods of low lake level corresponded to a decrease in annual precipitation, an increase in summer temperature, and a longer growing season. This general pattern could have resulted from alternate southward-northward displacements of the Atlantic Westerly Jet and from changes in its strength in response to variations in the thermal gradient between high and low latitudes. Summer temperature estimates show fluctuations with a $\sim 2^\circ\text{C}$ amplitude, while the changes in annual precipitation reached 100/150 mm on the Swiss Plateau and ~ 200 mm at higher altitude on the Jura Mountains. However, the high variability in estimates of annual precipitation obtained at Le Locle for the Preboreal period probably reflect poor pollen analogs for that period, during which dynamic vegetation change was characterized by the reestablishment of forest cover.

Taken as a whole, these results concur with mechanisms put forward to explain changes in tree-limit altitude in the Alps during the Holocene. Observations and experiments have shown that a tree-limit decline was caused not by severe winters but chiefly by a lowering of summer temperature and a shortening of the growing season (Bortenschlager, 1977). Precipitation and summer conditions are also factors driving glacier variations in the Alps (Holzhauser et al., 2005), and variations in the production of *Najas flexilis* seeds observed in Swiss lakes (Haas et al., 1998).

As regards the possible impact of the climatic oscillations on past agricultural societies, the variations in climatic parameters reconstructed at Montilier suggest an influence of these climatic oscillations on prehistoric groups not only for the location of their villages (Figure 6), but also for their socio-economic equilibrium. Phases of cooler and wetter climatic conditions may have provoked harvest failures, more or less dramatic depending on the subsistence strategy and population density. Archaeological studies of Neolithic lake-shore villages at Lakes Chalain and Clairvaux in the Jura Mountains (5.7 and 4.5 cal ka BP) provide evidence of stronger sensitivity to bad weather conditions and increased instability when ancient agricultural communities chose to develop cereal production and a crop subsistence strategy in response to the growth of the regional population to the detriment of grazing (Arbogast et al., 1996). These observations suggest complex interactions between climatic and human factors. In agreement with the quantitative estimates of climatic variables reconstructed from pollen and lake-level data in west-central Europe, Tinner et al. (2003) have also pointed to the fact that, during climatic reversals, increasing precipitation probably had a more decisive impact than temperature decrease on former settlements in marginal elevated areas such as the Alps.

Toward a Better Understanding of the Forcing Factors

Finally, what do we know of the possible forcing factors driving the climatic oscillations reflected by Holocene changes in lake levels in west-central Europe? Several causes have been postulated, including changes in the oceanic thermohaline circulation (THC), freshwater outbursts from proglacial lakes due to

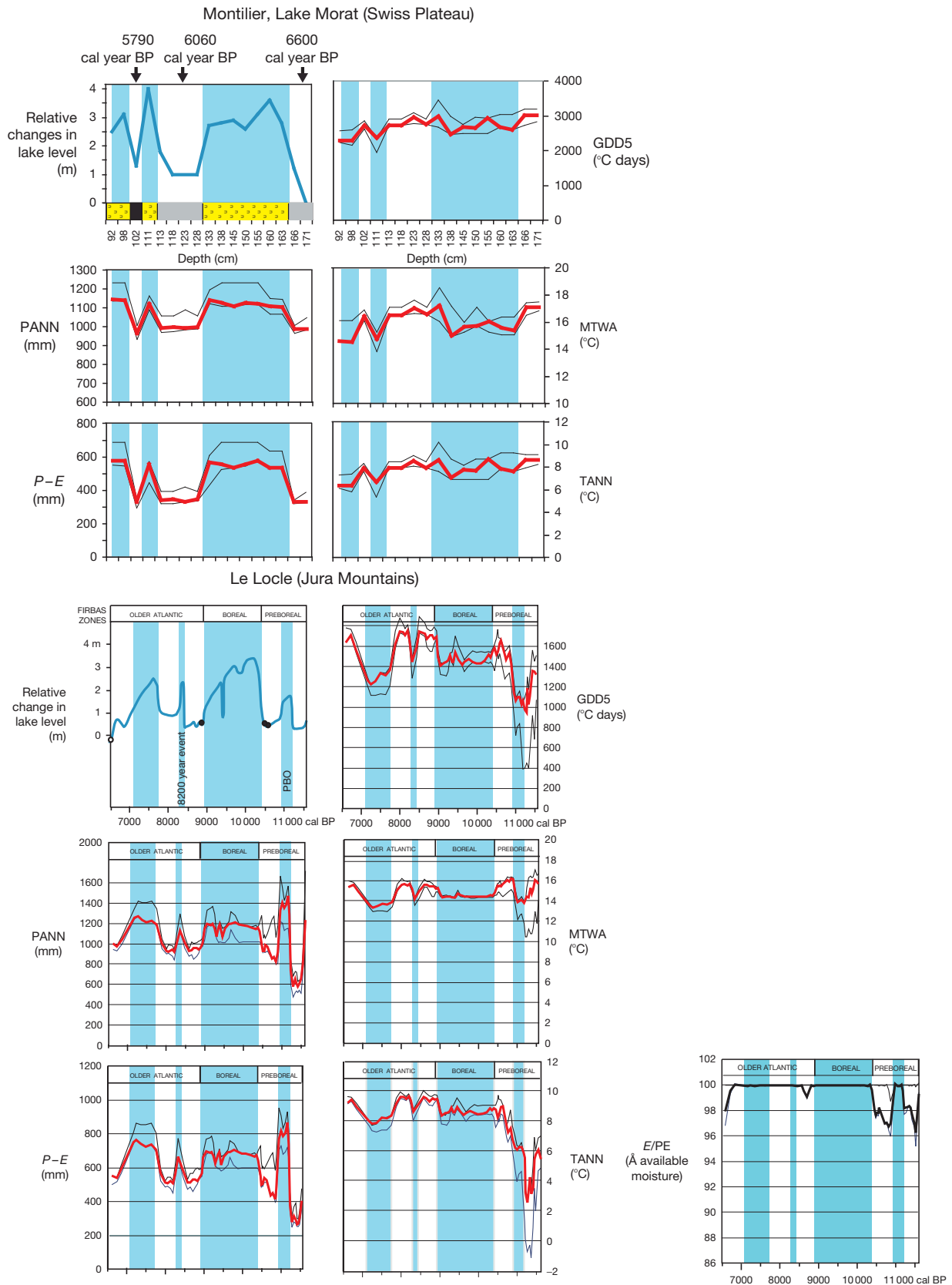


Figure 5 Climatic variables with confidence interval reconstructed from the sediment sequences of Montilier, Lake Morat (429 m, Swiss Plateau) and Le Locle (915 m, Jura Mountains). MTWA, mean temperature of the warmest month; MTCO, mean temperature of the coldest month; TANN, mean annual temperature; GDD5, growing degrees above 5 °C × days; $P-E$, precipitation minus evaporation; PANN, annual precipitation; E/PE , actual evapotranspiration/potential evapotranspiration.

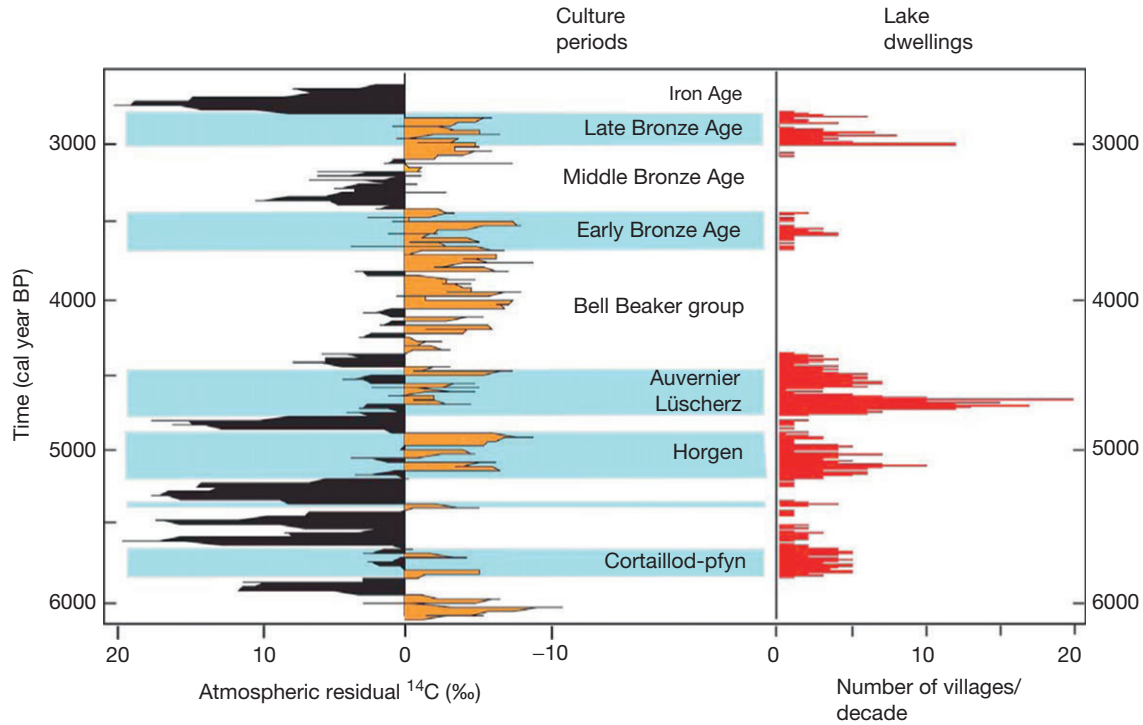


Figure 6 Comparison between the atmospheric residual ^{14}C variations (Stuiver et al., 1998) and the frequency of Neolithic and Bronze Age lake-shore villages in France and in Switzerland.

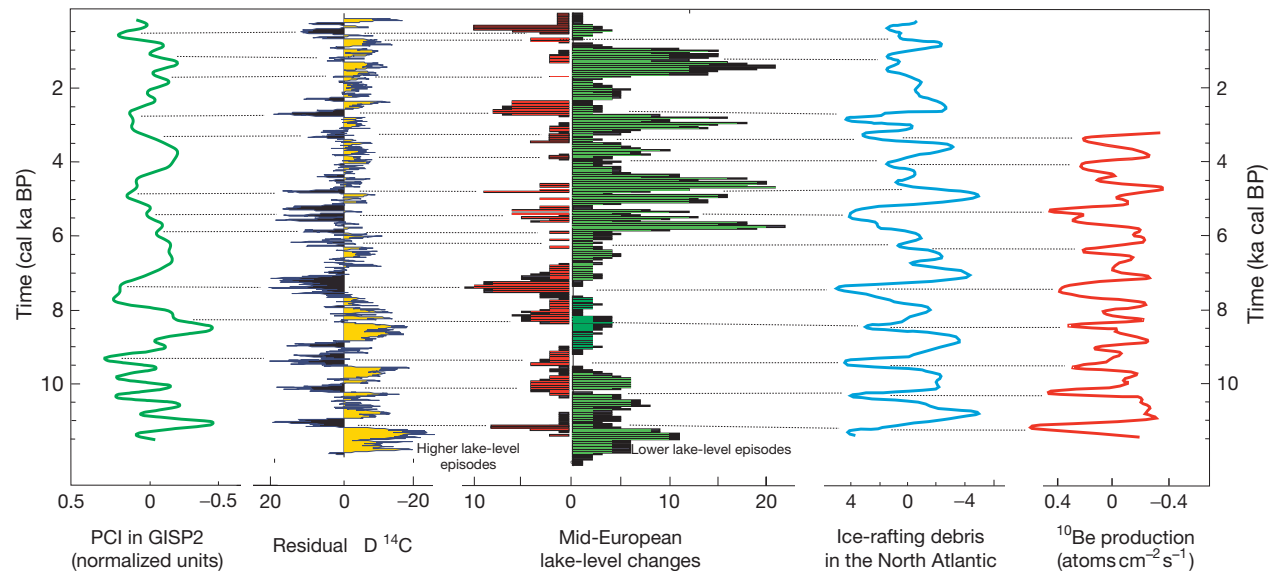


Figure 7 Comparison of the Holocene record of the Polar Circulation Index (PCI) at GISP2 (Mayewski et al., 1997), the atmospheric residual ^{14}C variations (Stuiver et al., 1998), the mid-European phases of higher lake level (see Figure 3), the ice-rafting debris (IRD) events in the North Atlantic Ocean (Bond et al., 2001), and the Greenland ^{10}Be record (Bond et al., 2001).

the final steps of deglaciation in the early Holocene, and variations in solar activity (Björck et al., 1996; Bond et al., 2001; Magny, 1993a; van Geel et al., 1996).

Figure 7 shows a comparison of the mid-European lake-level record with (1) the Polar Circulation Index (PCI)

established from the GISP2 glaciochemical series (Mayewski et al., 1997); (2) the atmospheric ^{14}C residual series (Stuiver et al., 1998); and (3) the North Atlantic ice-rafting debris (IRD) record (Bond et al., 2001). The PCI is thought to provide a relative measure of the average size and intensity of polar

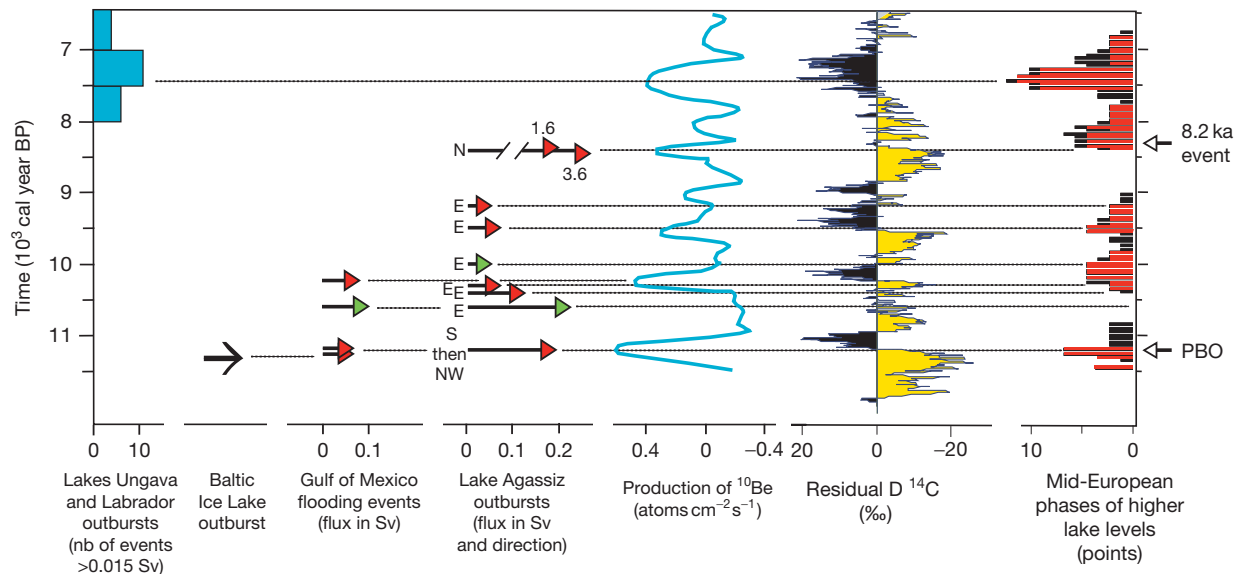


Figure 8 Comparison of early Holocene cooling phases in west-central Europe as reflected by highstands of lake levels (see Figure 3) with the atmospheric ^{14}C residual record (Stuiver et al., 1998), the Greenland ^{10}Be record (Bond et al., 2001), and freshwater outbursts from north-European and North American proglacial lakes (Jansson and Kleman, 2004; Teller et al., 2002).

atmospheric circulation. The radiocarbon curve may be considered as a proxy record of past changes in solar activity while IRD peaks indicate cooling events punctuating the whole Holocene in response to variations in the solar irradiance. In general, as pointed out by Magny (1999), phases of higher lake level in west-central Europe coincided with IRD events, and ^{14}C and PCI maxima. This supports the hypothesis that changes in solar activity were a major driving factor of Holocene climatic oscillations.

As illustrated by Figure 6, another indirect indication of a possible link between solar activity and mid-European lake-level fluctuations may be observed for the period 6–2.8 cal ka BP, when there are close correlations between the atmospheric ^{14}C residual record and the frequency of the Neolithic and Bronze Age lake dwellings well dated by dendrochronology.

Finally, Figure 8 compares the phases of higher lake level in west-central Europe with the atmospheric ^{14}C and Greenland ^{10}Be records, which here are proxies for past fluctuations of solar activity, and freshwater outbursts from different proglacial lakes that punctuated the deglaciation in Europe and North America. These meltwater flooding events could have induced decreases in THC (Björck et al., 1996; Jansson and Kleman, 2004; Teller et al., 2002).

Two kinds of situations may be distinguished from data illustrated in Figure 8. First, meltwater pulses at 11300–11250–11200–11170, 10400–10300–10200, 9500, 9200, 8400, and 7500–7000 cal years BP happened at times characterized by minima in solar activity or immediately preceded them. They led to widespread cooling events in the North Atlantic area and are marked by high lake levels in west-central Europe. Second, the freshwater outbursts at 10600–10000, 8000–7500, and 7000–6400 cal years BP resulted neither in a widespread cooling event nor in a rise in mid-European lake levels. This second type of meltwater event occurred during periods of solar activity maxima.

These contrasting patterns suggest that solar activity may have played a crucial role in amplifying or damping the possible effects of meltwater flooding events on the climate (Magny and Bégeot, 2004).

See also: [Lake Level Studies: Modeling; Overview.](#)
[Paleoclimate: Introduction.](#) [Plant Macrofossil Methods and Studies: Mire and Peat Macros.](#) [Radiocarbon Dating: Calibration of the \$^{14}\text{C}\$ Record.](#)

References

- Ammann B (1982) Säkulare Seespiegelschwankungen: Wo, wie, wann, warum? *Mitteilungen der Naturforschungs Gesellschaft, Bern, Neue Folge* 39: 97–106.
- Arbogast R-M, Magny M, and Pétrequin P (1996) Climat, cultures céréalières et densité de population au Néolithique: le cas des lacs du Jura français de 3500 à 2500 av. J.C. *Archäologisches Korrespondenzblatt* 26: 121–144.
- Berglund BE, Aaby B, Digerfeldt G, et al. (1983) Palaeoclimatic changes in Scandinavia and on Greenland – A tentative correlation based on lake and bog stratigraphical studies. *Quaternary Studies in Poland* 4: 27–44.
- Björck S, Kromer B, Johnsen S, et al. (1996) Synchronized terrestrial-atmospheric deglacial records around the North-Atlantic. *Science* 274: 1155–1160.
- Bond G, Kromer B, Beer J, et al. (2001) Persistent solar influence on North Atlantic climate during the Holocene. *Science* 294: 2130–2136.
- Bortenschlager S (1977) Ursachen und Ausmass postglazialer Waldgrenzschwankungen in den Ostalpen. In: Frenzel B (ed.) *Dendrochronologie und Postglaziale Klimaschwankungen in Europa*, pp. 260–266. Wiesbaden: Steiner Verlag.
- Clerc J, Magny M, and Mouthon J (1989) Histoire d'un milieu lacustre du Bas-Dauphiné: le Grand-Lemps. *Revue de Paléobiologie* 8: 1–19. Etude palynologique des remplissages tardiglaciaires et holocènes et mise en évidence de fluctuations lacustres à l'aide d'analyses sédimentologiques et malacologiques.
- Digerfeldt G (1988) Reconstruction and regional correlation of Holocene lake-level fluctuations in Lake Bysjön, South Sweden. *Boreas* 17: 165–182.
- Gaillard M-J (1985) Postglacial palaeoclimatic changes in Scandinavia and central Europe. A tentative correlation based on studies of lake-level fluctuations. *Ecologia Mediterranea* 11: 159–175.

- Guiot J, Harrison SP, and Prentice IC (1993) Reconstruction of Holocene pattern of moisture in Europe using pollen and lake-level data. *Quaternary Research* 40: 139–149.
- Haas JN (1994) First identification key for charophyte oospores from central Europe. *Journal of Phycology* 29: 227–235.
- Haas JN, Richoz I, Tinner W, and Wick L (1998) Synchronous Holocene climatic oscillations recorded on the Swiss Plateau and at the timberline in the Alps. *The Holocene* 8: 301–304.
- Hammarlund D, Björck S, Buchardt B, Israelson C, and Thomsen CT (2003) Rapid hydrological changes during the Holocene revealed by stable isotope records of lacustrine carbonates from Lake Igelsjön, southern Sweden. *Quaternary Science Reviews* 22: 353–370.
- Hammarlund D, Björck S, Buchardt B, and Thomsen CT (2005) Limnic responses to increased effective humidity during the 8200 cal yr BP cooling event in southern Sweden. *Journal of Paleolimnology* 34: 471–480.
- Hannon GE and Gaillard MJ (1997) The plant-macrofossil record of past lake-level changes. *Journal of Paleolimnology* 18: 15–28.
- Harrison S, Frenzel B, Huckriede U, and Weiss MM (1998) *Paläoklimaforschung*. Palaeohydrology as Reflected in Lake-Level Changes as Climatic Evidence for Holocene Times, vol. 25, p. 180. Stuttgart: Gustav Fischer Verlag.
- Holzhauser H, Magny M, and Zumbühl H (2005) Glacier and lake-level variations in west-central Europe over the last 3500 years. *The Holocene* 15: 789–801.
- Hyvärinen H and Alhonen P (1994) Holocene lake-level changes in the Fennoscandian tree-line region, western Finnish Lapland: Diatom and cladoceran evidence. *The Holocene* 4: 251–258.
- Jacomet S (1985) *Botanische Makroreste aus den Sedimenten des neolithischen Siedlungsplatzes AKAD-Seehofstrasse am untersten Zürichsee*. Zürcher Studien zur Archäologie, p. 94. Zürich: Verlag Juris Druck + Verlag AG.
- Jansson KN and Kleman J (2004) Early Holocene glacial lake meltwater injections into the Labrador Sea and Ungava Bay. *Paleoceanography* 19: 1–12.
- Korhola A, Tikkanen M, and Weckström J (2005) Quantification of Holocene lake-level changes in Finnish Lapland using a cladocera-lake depth transfer model. *Journal of Paleolimnology* 34: 175–190.
- Lüdi W (1935) *Das Grosse Moos im westschweizerischen Seelande und die Geschichte seiner Entstehung*, p. 344. Bern: Verlag Hans Huber.
- Magny M (1993a) Solar influences on Holocene climatic changes illustrated by correlations between past lake-level fluctuations and the atmospheric ^{14}C record. *Quaternary Research* 40: 1–9.
- Magny M (1993b) Holocene fluctuations of lake levels in the French Jura and Subalpine ranges and their implications for past general circulation pattern. *The Holocene* 3: 306–313.
- Magny M (1998) Reconstruction of Holocene lake-level changes in the Jura (France): Methods and results. In: Harrison SP, Frenzel B, Huckriede U, and Weiss MM (eds.) *Paläoklimaforschung 25: Palaeohydrology as Reflected in Lake-Level Changes as Climatic Evidence for Holocene Times*, pp. 67–85. Stuttgart: Gustav Fischer Verlag.
- Magny M (1999) Lake-level fluctuations in the Jura and French subalpine ranges associated with ice-rafting events in the North Atlantic and variations in the polar atmospheric circulation. *Quaternaire* 10: 61–64.
- Magny M (2004a) The contribution of palaeoclimatology to the lake-dwellings. In: Menotti F (ed.) *Living on the Lake in Prehistoric Europe. 150 Years of Lake-Dwelling Research*, pp. 132–143. London: Routledge.
- Magny M (2004b) Holocene climatic variability as reflected by mid-European lake-level fluctuations, and its probable impact on prehistoric human settlements. *Quaternary International* 113: 65–79.
- Magny M and Bégeot C (2004) Hydrological changes in the European midlatitudes associated with freshwater outbursts from Lake Agassiz during the Younger Dryas event and the early Holocene. *Quaternary Research* 61: 181–192.
- Magny M, Bégeot C, Guiot J, and Peyron O (2003) Contrasting patterns of hydrological changes in Europe in response to Holocene climate cooling phases. *Quaternary Science Reviews* 22: 1589–1596.
- Magny M, Guiot J, and Schoellhammer P (2001) Quantitative reconstruction of Younger Dryas to mid-Holocene paleoclimates at Le Locle, Swiss Jura, using pollen and lake-level data. *Quaternary Research* 56: 170–180.
- Magny M, Miramont C, and Sivan O (2002) Assessment of climate and anthropogenic factors on Holocene Mediterranean vegetation in Europe on the basis of palaeohydrological records. *Palaeogeography, Palaeoclimatology, Palaeoecology* 186: 47–59.
- Magny M, Peyron O, Bégeot C, and Guiot J (2005) Quantitative reconstruction of mid-Holocene climatic variations in the northern Subalpine zone. A comparative study of two sites from lakes Morat (Swiss Plateau) and Annecy (French Pre-Alps). *Boreas* 34: 1–13.
- Mayewski PA, Meeker LD, Twickler MS, Whitlow S, Yang Q, and Prentice M (1997) Major features and forcing of high latitude northern hemispheric atmospheric circulation using a 110 000 year long glaciochemical series. *Journal of Geophysical Research* 102: 26345–26366.
- Menotti F (2004) *Living on the Lake in Prehistoric Europe. 150 Years of Lake-Dwelling Research*, p. 186. London: Routledge.
- Ralska-Jasiewiczowa M and Starkel L (1988) Records of the hydrological changes during the Holocene in the lake, mire and fluvial deposits of Poland. *Folia Quaternaria* 57: 91–127.
- Stuiver M, Reimer PJ, Bard E, et al. (1998) INTCAL 98 radiocarbon age calibration, 24000–0 cal BP. *Radiocarbon* 40: 1041–1083.
- Teller JT, Leverington DW, and Mann JD (2002) Freshwater outbursts to the oceans from glacial Lake Agassiz and their role in climate change during the last deglaciation. *Quaternary Science Reviews* 21: 879–887.
- Tinner W, Lotter AF, Ammann B, et al. (2003) Climatic change and contemporaneous land-use phases north and south of the Alps 2300 BC to 800 AD. *Quaternary Science Reviews* 22: 1447–1460.
- Valero-Garcés B, Zeraoui E, and Kelts K (1998) Arid phases in the western Mediterranean region during the Last Glacial Cycle reconstructed from lacustrine records. In: Benito G, Baker VR, and Gregory KJ (eds.) *Palaeohydrology and Environmental Change*, pp. 67–80. Chichester: Wiley.
- van Geel B, Buurman J, and Waterbolk HT (1996) Archaeological and palaeoecological indications of an abrupt climate change in The Netherlands, and evidence for climatological teleconnections around 2650 BP. *Journal of Quaternary Science* 11: 451–460.
- Yu G and Harrison SP (1995) Holocene changes in atmospheric circulation patterns as shown by lake status changes in northern Europe. *Boreas* 24: 260–268.

Modeling

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Introduction

Lakes have been recognized as paleoclimatic indicators for a long time. Lakes respond to changes in the local water balance by changing in depth and area. Lake water balance is in equilibrium with climate changes and therefore can be used as an indicator of past climate. Changes in the lake water balance affect a variety of biological, limnological, and sedimentological processes within the lake and changes in these processes are often recorded in the lake sediments. Therefore, investigations of lake sediments can be used to reconstruct past lake-level changes and hence the past climate changes.

Closed-basin lakes are most sensitive to climate changes; these are widespread today in arid and semiarid climate areas (Street-Perrott and Harrison, 1985). Lake sediments and the position of the past shorelines indicate that lake water levels in those areas were much higher (in the magnitude of 10–100 m) in the past. The water balance of open-basin lakes has been relatively less studied, because of the assumption that they are insensitive to climate changes and should respond to changes in water balance through changes in discharge rather than in lake level. However, Digerfeldt (1986) developed a method for reconstruction of past lake-level changes using detailed biostratigraphic investigations. He found that lakes in southern Sweden had experienced significant lake-level lowering (in magnitude of up to 10 m) during the Holocene. Many other investigators used Digerfeldt's approach and found that significant lake-level changes in open-basin lakes in humid climates have occurred in the past. So the geological records of both closed-basin and overflowing lakes have been used successfully to reconstruct lake-level changes during Quaternary. However, the geological record indicates whether the climate was generally wetter or drier, but tells us little about the specific nature of the climate changes involved. Empirically, it was understood how water-level fluctuations in arid and semiarid areas were driven by climate changes, but the low lake levels in humid climates were difficult to explain only by climate changes. A possible approach to understanding the interaction between climate change and lake-level fluctuations is through modeling.

The Basis of Lake-Level Modeling

Lake models can be described as either deterministic or stochastic (Henderson-Sellers, 1984). Deterministic models assume that variables can be well specified and simulated, while stochastic models allow for some random variation in these variables. Deterministic models are common but stochastic ones are very rare. Further, models can be either conceptual or empirical. Conceptual models are based on physical reasoning while empirical models are largely statistical fits to

observational data. Deterministic–empirical relationships were commonly used in the earliest attempts to understand past lake-level fluctuations. A direct statistical correlation was assumed between evaporation (or lake level) and some meteorological variable, such as air or water temperature. However, empirical estimates of evaporation oversimplified the interaction between a lake and the overlying atmosphere. Both evaporation and lake water temperature are actually determined by the lake energy balance (Henderson-Sellers, 1984). Lakes with different morphometries and physical characteristics have different thermal and water balance regimes, and empirical relationships are usually valid only for the specific lake for which the relationship was developed. Therefore, empirical relationships are not ideal for predicting the likely impact of climate changes (either past or future) on other lakes with different morphometries and in different climates.

Models for paleoclimate studies should necessarily be conceptual or physically based (i.e., based on the underlying physics or chemistry of the phenomena being described), in order to eliminate questions regarding the applicability of the model in the past (or similarly, in the future) when statistical relationships among variables may have significantly differed from those at present. Conceptual lake models, which explicitly represent the physical processes governing the energy and water balances of the lake, offer a more robust way to predict climate-induced changes in lake water volume, level, and discharge. Such models allow an assessment of the relative importance of individual climate controls and the interactions among them and testing of specific paleoclimatic hypotheses.

Lake Water Balance

The response of lake volume, and hence level, to climate changes is given by the lake water balance equation (Mason et al., 1994; Street-Perrott and Harrison, 1985):

$$\frac{\Delta V}{\Delta t} = (R + P + G_1) - (D + E + G_0)$$

where the change in lake volume (ΔV) with time (t) is the difference between water inputs and outputs. Water inputs to the lake are runoff (R) from the catchment, precipitation (P) over the lake surface, and groundwater influx (G_1). Water outputs from the lake are outflow or discharge (D), evaporation (E), and groundwater discharge (G_0) (Figure 1). Most paleomodeling studies assume that groundwater fluxes are negligible ($G_1 - G_0 \approx 0$), and changes in lake volume (and level) can be derived from the water fluxes over the lake and catchment, providing that it is possible to calculate amount of discharge – although models do exist that take groundwater fluxes into account (Cardille et al., 2004). When groundwater fluxes are negligible, evaporation, runoff, and discharge need to be

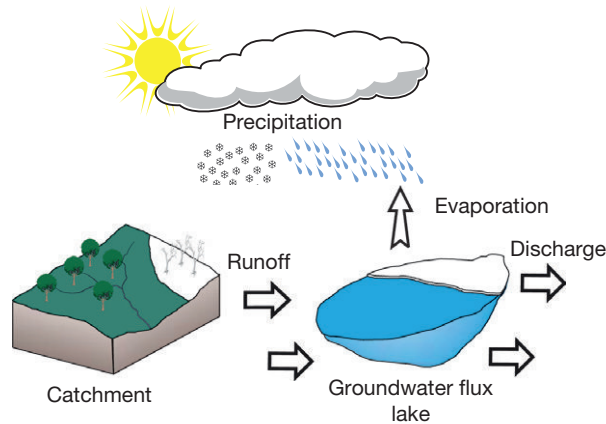


Figure 1 Main components of the lake water balance.

calculated. Lake physical models derive evaporation from the lake energy balance and runoff from the soil moisture balance.

Lake Energy Balance

Evaporation is one of the important factors controlling the lake water volume and level, especially in closed-basin lakes. Evaporation occurs when a vapor pressure deficit between a lake surface and the overlying atmosphere exists and sufficient energy is available. Evaporation is affected by meteorological factors, such as solar radiation, air and water temperatures, wind speed, and humidity, but also by physical characteristics of the lake such as the size of the lake surface area and water quality. Many formulas exist to calculate evaporation, and they require lake water energy-balance modeling (e.g., [Burman and Pochop, 1994](#); [Henderson-Sellers, 1984](#)).

Lake Surface Energy Balance

The lake water temperature can be derived from the lake water surface energy balance (SEB), where the SEB is the sum of all energy fluxes across the air–water interface of a lake ([Figure 2](#)). However, a significant amount of heat is stored in the lake water during the heating period (e.g., this is summer in the Northern Hemisphere extratropics) and is released during the cooling period (autumn in the Northern Hemisphere extratropics). This thermal inertia causes lake heating to be out of phase with solar radiation and air temperature forcing ([Hostetler and Bartlein, 1990](#)). Therefore, to calculate lake water temperature and evaporation, some stratification (vertical heat transfer) is required in the model.

Two types of stratification models have been used: eddy-diffusion (or differential) and mixed-layer (or integral) models ([Henderson-Sellers, 1984](#)). Comparison between these two types of models suggests that they are complementary rather than competitive. Mixed-layer models have been widely used in lake studies, although they have some limitations because of instability when run for long periods ([Henderson-Sellers and Davies, 1989](#)). Eddy-diffusion models are numerically more stable over long integrations and have been successfully used

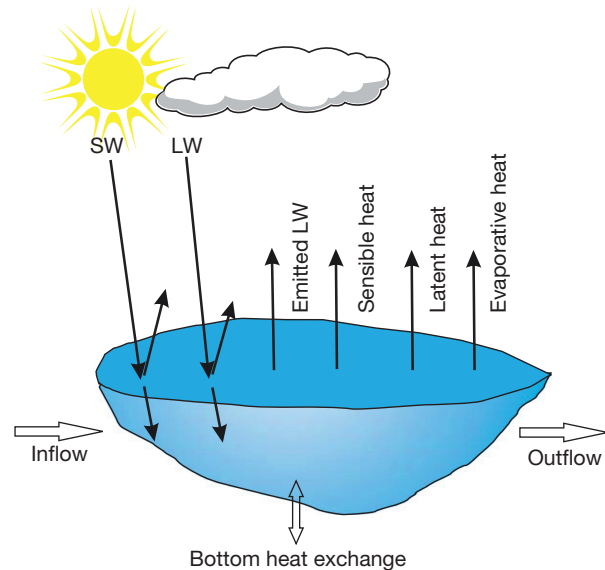


Figure 2 Main components of the lake energy balance model. SW is shortwave radiation and LW longwave radiation, both of which are partly reflected and/or absorbed by the lake water surface.

in a variety of lake studies (e.g., [Hostetler and Benson, 1990](#); [Vassiljev, 1998](#); [Vavrus et al., 1996](#)).

Lake Ice Models

During cold periods, lakes in the mid- and high latitudes will ice over; this increases the surface albedo (reflectivity) and limits the amount of solar radiation penetrating into the water of the lake. This ice cover will significantly alter the lake energy balance (and hence evaporation). Therefore, the models used for lakes in cold climates incorporate some type of thermodynamic ice model to simulate the accretion and ablation of ice and the snow covering the ice ([Hostetler, 1991](#); [Patterson and Hamblin, 1988](#); [Vassiljev et al., 1994](#); [Vavrus et al., 1996](#)).

Soil Moisture Balance

Runoff from the catchment is an important source of water inputs to lakes, especially in humid climates. Changes in runoff are related to climatic factors, such as precipitation and actual evapotranspiration, but they are also affected by some catchment-specific characteristics, such as soil texture and vegetation cover ([Figure 3](#)). Soil texture determines the water-holding capacity of the soil, which in turn determines the extent to which moisture is carried through into drier seasons. When the soil moisture storage is large, there can be a significant time lag between precipitation events and runoff. The density of the vegetation and the type of vegetation (e.g., grasses or trees) affects the rate of actual evapotranspiration and hence the amount of water that can be removed from the soil.

The physically based models currently used to predict runoff are basically soil moisture balance models ([Evans and Trevisan, 1995](#); [Harrison et al., 1993](#); [Haxeltine et al., 1996](#);

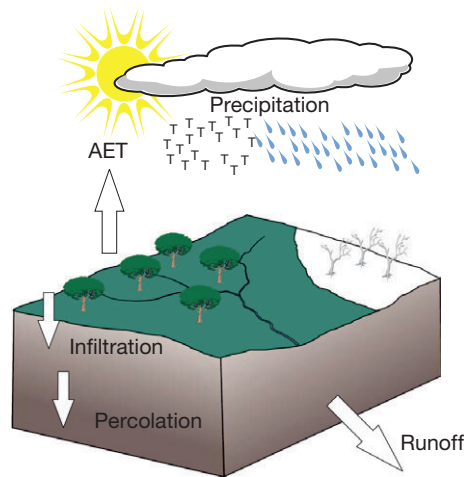


Figure 3 Simplified principles of the soil moisture balance models. Soil receives moisture from precipitation (in the form of rain or snow) and releases it through actual evapotranspiration (determined by vegetation type and by soil water and energy supply) and runoff. Precipitation infiltrates into the soil at a rate determined by soil texture. Moisture in the soil percolates at a rate determined by soil texture and moisture content, and when the soil moisture is at field capacity runoff occurs.

Prentice et al., 1992). The models have been developed from the so-called Budyko 'bucket' model, as reformulated by Manabe (1969), in which the soil was considered as a single simple water store of fixed depth. Although some soil moisture models preserve this concept (Evans and Trevisan, 1995; Harrison et al., 1993; Prentice et al., 1992), there are more complex models wherein the soil is subdivided into two (Haxeltine et al., 1996) or even more (Nikolaidis et al., 1993) compartments. The incorporation of multiple soil layers allows a more explicit differentiation between the rate of initial infiltration into the soil and the rate of transmission of water through the soil column. More importantly, it permits differences in the rooting depths of different plant types (and hence in their ability to compete for water) to be taken into account explicitly. Although all physically based soil moisture models include a vegetation, the treatment of the vegetation varies in complexity. While the simplest models implicitly consider only the presence/absence of vegetation, more complex models allow for a number of different vegetation types, which differ in factors such as albedo, phenology, and rooting depth.

Discharge

Lakes in humid climates usually have a surface outlet in the form of a stream or river. The form of this outlet can be changed by erosion and infilling, except in the case where the outlet is cut through solid bedrock (Dearing and Foster, 1986). It is difficult to determine the size and the shape of the outlet at specific times in the past. However, since it is necessary to allow for possible changes in outflow volume, then some approximation of past outflow parameters must be made. It is usually assumed that the shape of the channel outlet and basin have not changed substantially through time. The simplest way to

calculate the volume of outflowing water is as a function of the outlet cross-sectional area and the velocity of the flow. The flow velocity at the outlet can be calculated using, among the others, the Manning equation (Bras, 1990). In this way, only the cross-sectional area of the outlet and the slope of the outflow channel need to be specified. These parameters can be obtained, for example, from detailed contour maps.

Input Data

The models are usually driven by monthly meteorological data including solar radiation, air temperature, vapor pressure, cloudiness, wind speed, and precipitation. Some basin-specific parameters, such as lake bathymetry, are also required. Most of the models run with a daily time step.

Modeling Strategies

Modeling consists of three main aspects: model development, which has been described above, validation, and application phase. Models must be validated under modern climate conditions, preferably using long instrumental records of climate and hydrology (Hostetler and Benson, 1990; Vassiljev, 1998). The validation shows how correctly the model simulates the observed behavior of the lake (e.g., water level) and indicates the weaknesses of the model that need to be improved. One of the most extensively validated models, developed by Hostetler and Bartlein (1990), was successfully validated with observed evaporation and surface temperature data from three North American lakes: Salton Sea (years 1961–62, California), Pretty Lake (years 1963–64, Indiana), and Pyramid Lake (years 1987–88, Nevada). Later the modified model (Vassiljev et al., 1994, 1998) was successfully validated against observed lake-level data for lakes in northern Europe: Lakes Viljandi (years 1940–90) and Karujärv (years 1976–87, Estonia), and Lake Bysjön (years 1973–77, southern Sweden).

In the application phase, two modeling strategies are used to examine the response of a lake to a change in climate forcing: simulation and sensitivity analyses.

Simulation Analysis

In simulation mode, known changes in climate forcings at a given time are used to drive the model. A number of climate parameters, such as solar radiation, temperature, precipitation, vapor pressure, cloudiness, and wind speed, are required. Many types of paleoenvironmental data can be used to reconstruct the required climate parameters for model input. For example, there are well-known changes in solar radiation (insolation) during the Holocene (Berger and Loutre, 1991) and pollen-based reconstructions of temperature and precipitation. Unfortunately, not all parameters can be reconstructed from proxy data directly. For example, there are limited possibilities to reconstruct changes in wind speed and cloudiness from proxy data (although see Adams and Tetzlaff, 1984; Harrison et al., 1993). Models use the present-day mean or extreme values for such parameters. Other possible climate data sources used in the modeling are general circulation

models (GCMs), which can provide all required data (e.g., Hostetler and Benson, 1990).

Sensitivity Analysis

In sensitivity mode, the climate forcing is changed systematically, possibly but not necessarily within a range determined by actual, observed changes. This allows the testing of the importance of every climate parameter, individually or in some combination, for the lake water balance.

Modeling Results

The Effect of Changes in Solar Radiation

Variations in the Earth's orbital parameters (precession, tilt, and eccentricity) are known to have caused significant changes to the seasonal and latitudinal distribution of incoming solar radiation (insolation) during the Holocene (Berger and Loutre, 1991). For example at 9 ¹⁴C ka BP, when perihelion was in June rather than in January as in the present, summer insolation in the Northern Hemisphere was ~6–16% greater than today (depending on latitude), but winter insolation was reduced by 7–16% (Table 1). Earth's orbital speed is greatest near perihelion and least at aphelion. The shift from perihelion in June at 9 ¹⁴C ka BP to perihelion in Jan today has resulted in changes in the lengths of the climatological winter and summer seasons (Kutzbach and Gallimore, 1988): the Northern Hemisphere summer was shorter by ~6 days, and winter accordingly longer (Table 2). These orbitally induced anomalies gradually attenuated through the Holocene. In most lake models, solar radiation is modified according to orbital changes and formulas suggested by Berger and Loutre (1991). In some models (e.g., Harrison et al., 1993; Vassiljev et al., 1998), the lengths of each

month under different orbital configurations are also modified using a procedure proposed by Kutzbach and Gallimore (1988), in which the beginning and end of each celestial month is defined by celestial longitude.

Modeling studies show that orbitally induced changes in solar radiation and in the lengths of the climatological seasons affect the lake water balance in both open- and closed-basin lakes. However, the changes are small and cannot explain either the magnitude or pattern of observed lake-level changes during the Holocene.

Simulation Analyses

The earliest paleomodeling studies were mainly concentrated on closed-basin lakes in arid and semiarid climates. Only from the 1990s have open-basin paleolakes in humid climates been modeled. These studies showed the changes in climate that were required to produce the observed patterns of lake-level change.

Paleolake Chad

One of the first water balance modeling studies was for Paleolake Chad (northern Africa), where the magnitude of climate changes required to explain lake fluctuations was estimated. A combined hydrological and energy-balance model was first applied to Paleolake Chad by Kutzbach (1980) to estimate climate conditions between 10 and 5 ka BP, when the water level of the lake was 40 m higher than present. Kutzbach found that such a high water level could exist only when the annual precipitation was at least 600–650 mm (double that of the present). However, these precipitation values are minimum estimates, because Paleolake Chad had risen to the overflow level and the volume of discharge was unknown. Later Adams and Tetzlaff (1984) simulated climate conditions for the Last Glacial Maximum (18 ka BP). Their simulations suggested that south of the lake air temperature was 4 °C lower than in present times, while the north of the lake was 6 °C lower. Lower temperatures were paralleled by decreases in annual humidity, and annual precipitation was 75 mm (five times lower than present). Lower humidity reduced cloud cover by 10% and increased solar radiation by 10%. These climate parameters were used to drive the model; results showed that the northern part of Paleolake Chad would have dried up and only a very shallow lake with water temperatures around 40 °C would have existed in the southern part. This too probably disappeared in some years.

Table 1 Example of insolation values at the top of the atmosphere during the Holocene at 55° N according to orbital parameters given by Berger and Loutre (1991)

Time (¹⁴ C years BP)	Perihelion (°)	Tilt (°)	Eccentricity	Insolation (W m ⁻²)	
				July	January
0	282.04	23.446	0.016724	457	64
3000	231.08	23.820	0.017824	468	61
6000	164.28	24.167	0.018911	491	57
9000	114.82	24.227	0.019419	500	56

Table 2 Changes in month length during the Holocene as a result of changes in the timing of perihelion

Time (¹⁴ C years BP)	Length of month (number of days)											
	Jan	Feb	Mar	Apr	May	Jun	Jul	Aug	Sep	Oct	Nov	Dec
0	31	28	31	30	31	30	31	31	30	31	30	31
3000	32	30	31	31	31	30	30	30	29	30	30	31
6000	33	31	32	31	31	29	29	29	28	30	30	32
9000	33	31	31	30	30	28	29	29	29	31	31	33

Paleolake Lahontan

High lake-level stands in Paleolake Lahontan (USA, Nevada) between about 13.5 and 11 ^{14}C ka BP were modeled using an energy flux balance model (Benson, 1981). High stands of Lake Lahontan were achieved by increasing historical values of cloudiness by 10% and by reducing the mean air temperature by at least 10 $^{\circ}\text{C}$, so that evaporation rates were an order of magnitude lower than present times. This was sufficient to support water levels 150 m higher. Hostetler and Benson (1990) conducted experiments based on the simulations of atmospheric GCMs, which indicate that during the late Pleistocene, the polar jet stream was split around the North American ice sheet. The position of the southern branch of the jet stream around 18 ^{14}C ka BP was $\sim 33^{\circ}\text{N}$ during winter and $\sim 45^{\circ}\text{N}$ during summer (currently 42°N during winter and 58°N in summer). They found that such a change in the position of the polar jet stream brought precipitation further south than at present and led to enhanced wetness in the area of Paleolake Lahontan ca. 20–13.5 ^{14}C ka BP. After 13.5 ^{14}C ka BP the jet stream shifted northward and aridity increased. The enhanced effective moisture was thus a result of increased precipitation combined with a reduction in evaporation due to increased summer cloudiness. Evaporation was estimated at 42% lower than today and the required precipitation was 540 mm, 1.8 times higher than historical maximum. The lake-level drop of 135 m from the maximum stand to Holocene levels could have occurred within 300 years. Later Hostetler (1991) showed that ice cover could have existed up to 4–4.5 months on the lake, leading to significant decreases in annual evaporation (about 60% from present value). Under these conditions, the precipitation requirement to maintain the high stand of the lake was similar to the historical maximum value (300 mm) from the year 1983.

Lake Bysjön

Biostratigraphical studies showed that Lake Bysjön (southern Sweden) experienced several episodes of major lake-level lowering during the Holocene, with the maximum around 9 ^{14}C ka BP, when water level dropped by 7 m (Figure 4). Modeling was used to investigate if it was possible to reproduce the observed lake-level changes through the Holocene by combining the precipitation changes with known changes in radiation and pollen-based temperature reconstructions (Vassiljev et al., 1998). Temperature reconstructions showed that the mean annual temperature was higher than today (up to 2 $^{\circ}\text{C}$) during the Holocene, caused mainly by increased winter temperatures, except in the period around 7 ^{14}C ka BP when winters were about 1.2 $^{\circ}\text{C}$ colder than today. Summers were slightly cooler (0.6–1.6 $^{\circ}\text{C}$) most of the time, except around 7 ^{14}C ka BP, when they were warmer by about 1.8 $^{\circ}\text{C}$.

Modeling showed that to produce a 7-m water level drop at 9 ^{14}C ka BP, changes in radiation and temperature (January warmer 2.3 $^{\circ}\text{C}$, July cooler 0.6 $^{\circ}\text{C}$ than today) must be accompanied by a decrease in annual precipitation of 75 mm (13% less than the present value of 587 mm; Figure 4). When winter temperatures were warmer than today, a reduction of runoff into the lake occurred. Together with relatively small changes

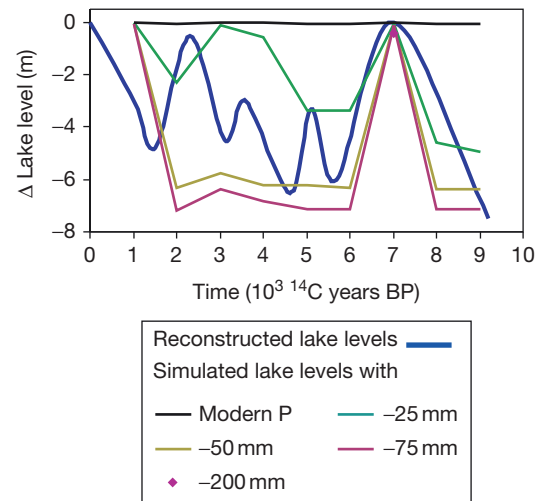


Figure 4 Simulated lake-level changes for Lake Bysjön (southern Sweden) during the Holocene in response to changes in radiation and pollen-based temperatures, together with different annual precipitation values (modern precipitation 587 mm and annual precipitation reduced by 25, 50, 75, and 200 mm). Blue line shows the water-level fluctuations reconstructed by biostratigraphical studies, 0 m corresponds to the present-day lake level.

in precipitation, this caused the lake to become closed, and the water level dropped significantly.

The simulations also indicated that an observed high water level around 7 ^{14}C ka BP, when the winter was colder and summer warmer than today, could have occurred even when there was a significant reduction in annual precipitation. For example, decreases in annual precipitation of 200 mm could cause only 0.4 m drop in the lake level (Figure 4). This apparent contradiction relates to winter temperatures colder than present, which significantly reduced the actual evapotranspiration. Decreases in runoff from spring snowmelt were relatively small, and the total volume of the runoff was sufficient to maintain the high water level. Experiments showed that all the observed lowerings occurred when winter temperatures were warmer than today. Thus, even though changes in precipitation appear to be the driver of observed lake-level changes, the trigger seems to be changes in winter temperature. This study demonstrated that lake levels in humid temperate climates are sensitive to changes in winter climates rather than changes in summer climates.

Sensitivity Analyses

As the examples above illustrate, changes in lake water balance can be caused by a range of different climate variables. Harrison et al. (1993) used a simple water balance model to change a number of climate parameters in accordance with paleo-observations. They showed that known changes in solar radiation, temperature, and cloudiness were unable to produce observed changes in lake level in Europe during the Holocene. Prentice et al. (1992) and Evans and Trevisan (1995) found that changes in the seasonality of precipitation could explain observed changes in lake levels in southern

Europe. [Prentice et al. \(1992\)](#) showed that a shift in precipitation to the winter period without changes in the total annual precipitation, was sufficient to maintain high lake levels at Lake Ioannina (northern Greece) at the Last Glacial Maximum. Moreover, they found that a shift in precipitation seasonality, combined with colder temperatures, could explain the coexistence of high lake levels and steppe vegetation in the eastern Mediterranean. [Evans and Trevisan \(1995\)](#) obtained similar results at the Lagoni peat-bog (northern Italy), showing that mesic conditions indicated by pollen data and humidity conditions suggested by lake level are not in conflict, but could be caused only by changes in the rainfall seasonality.

[Vassiljev \(1998\)](#) investigated the sensitivity of lakes to changes in precipitation. Systematic changes in annual and seasonal precipitation were applied to three lakes from northern Europe (Lake Bysjön, southern Sweden, and Lakes Karujärv and Viljandi in Estonia; [Figure 5](#)). Simulations showed that lake level is more sensitive to decreases than increases in precipitation. In cold/temperate humid climates, actual evapotranspiration is close to potential evapotranspiration, and increased precipitation leads to increased runoff. However, increased runoff is compensated for by increased discharge

and lake-level change is small. Reduction in winter precipitation had a much bigger impact on the lake levels than changes in summer precipitation. In cold/temperate humid climates an important component of runoff is snowmelt, and therefore to produce significant changes in runoff, unless there are changes in the winter climate are required.

Effect of Lake Morphometry

Since lakes vary in their morphology, sedimentation, and water balance type ([Street-Perrott and Harrison, 1985](#)), it is plausible that different responses of lakes to the same change in climate could reflect basin parameters. Some studies ([Dearing and Foster, 1986](#); [Magny, 1992](#); [Vassiljev, 1998](#)) suggested that lake level is sensitive to the ratio of the lake area to the catchment area. The behavior of 156 overflowing artificial lakes (in northern Europe) was examined to evaluate the influence of lake area and catchment area on responses to precipitation changes under modern climate conditions ([Harrison et al., 2002](#)). The results show that lake response to prescribed precipitation changes is determined by the ratio of the lake area to the catchment area. However, the limiting value of the ratio varies according to the magnitude of the prescribed change. For example, a decrease in mean annual precipitation of 200 mm (or 34%) produces a significant decrease in lake level when the catchment area is less than ten times larger than the lake area. A decrease in precipitation of 100 mm (or 17%) can produce a significant decrease in lake level only when the catchment area is less than two times larger than the lake area. [Cardille et al. \(2004\)](#) showed that groundwater-dominated lakes in North America behave in a similar way.

The results of these experiments help to explain earlier empirical studies that classified the response of lakes to climate change by lake water balance type. [Street-Perrott and Harrison \(1985\)](#) identified three kinds of lakes: atmosphere-controlled lakes (precipitation–evaporation), amplifier lakes (runoff–evaporation), and reservoir lakes (runoff–discharge). The main source of water input for atmosphere-controlled lakes is precipitation and for amplifier lakes it is runoff; both types of lakes lose water through evaporation. The main water input for reservoir lakes is runoff and water is removed mainly by discharge. They found that reservoir lakes are of little use as paleoclimatic indicators and closed-basin amplifier lakes are the most sensitive to climate changes. The results of [Harrison et al. \(2002\)](#) and [Cardille et al. \(2004\)](#) indicate that lake water balance type is related to the ratio of lake area to catchment area. Lakes with relatively large catchments tend to be reservoir lakes, and lakes with relatively small ones tend to be atmosphere-controlled lakes. The modeling results suggest that water level of the open-basin lakes with relatively small catchments, where runoff is less important, is sensitive to climate change, because their outlets close easily. Only amplifier and atmosphere-controlled lakes can be closed-basin lakes (water output is only through evaporation), because their water level is most sensitive to climate change. However, reservoir lakes can also be used for climate change studies but only if some other water quality parameters that are not dependent on the water level are used.

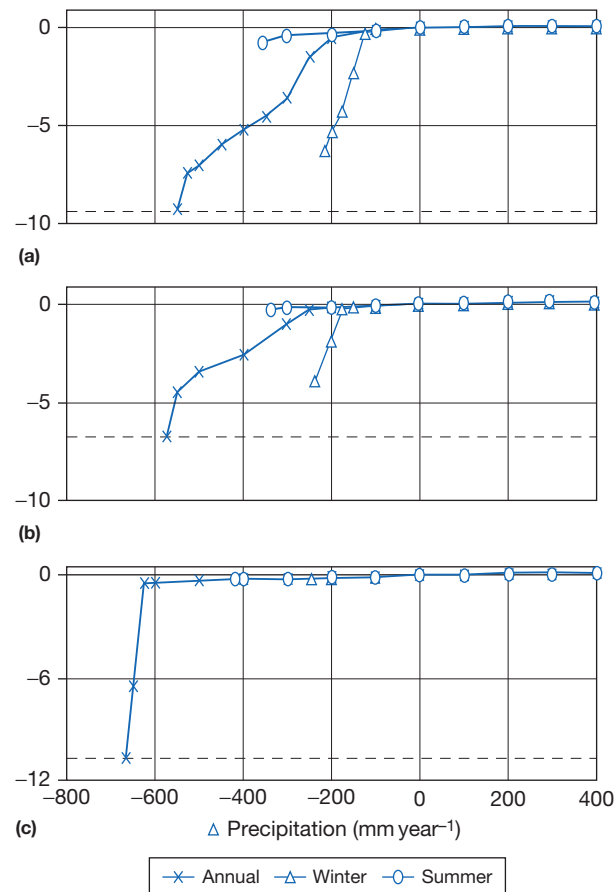


Figure 5 Simulated changes in lake level at Lake Bysjön (a), Lake Karujärv (b), and Lake Viljandi (c) in response to changes in annual winter or summer precipitation. The changes are expressed as anomalies from the present; dashed lines mark the level at which the lake is dry.

Conclusions

Lake-level modeling studies have successfully simulated the balance of closed- and open-basin lakes. The models have helped to explain the climatic factors behind lake-level fluctuations. Lake level of closed-basin lakes in arid climates is more sensitive to climate variables that control evaporation from the lake, such as temperature and cloudiness. The water level of open-basin lakes in humid climates is related to runoff, which is controlled by precipitation. Lake-level fluctuations can be simulated using realistic ranges of climate variability, often with more than one climate variable changing at the same time.

Simulations also explain why not all lakes respond to climate change in the same way: the sensitivity of lake level to climate change is determined by lake morphometry, especially by the ratio between lake and catchment.

See also: Diatom Introduction; **Glaciation, Causes:** Astronomical Theory of Paleoclimates; **Lake Level Studies:** Africa during the Late Quaternary; Overview; West-Central Europe; **Paleoclimate Modeling:** Last Glacial Maximum GCMs; The Last Interglacial; **Pollen Methods and Studies:** Use of Pollen as Climate Proxies.

References

- Adams LJ and Tetzlaff G (1984) Did Lake Chad exist around 18,000 yr BP? *Archives for Meteorology, Geophysics, and Bioclimatology* 34: 299–308.
- Benson LV (1981) Paleoclimatic significance of lake-level fluctuations in the Lahontan Basin. *Quaternary Research* 16: 390–403.
- Berger A and Loutre MF (1991) Insolation values for the climate of the last 10 million years. *Quaternary Science Reviews* 10: 297–317.
- Bras RL (1990) *Hydrology: An Introduction to Hydrologic Science Reading*. Reading, MA: Addison-Wesley.
- Burman R and Pochop LO (1994) *Evaporation, Evapotranspiration and Climate Data*. Amsterdam: Elsevier.
- Cardille JA, Coe MT, and Vano J (2004) Impacts of climate variation and catchment area on water balance and lake type in groundwater-dominated systems: A generic lake model. *Earth Interactions* 8: 1–24.
- Dearing JA and Foster IDL (1986) Lake sediments and palaeohydrological studies. In: Berglund BE (ed.) *Handbook of Holocene Palaeoecology and Palaeohydrology*, pp. 67–90. New York: Wiley.
- Digerfeldt G (1986) Studies on past lake-level fluctuations. In: Berglund BE (ed.) *Handbook of Holocene Palaeoecology and Palaeohydrology*, pp. 127–143. New York: Wiley.
- Evans SP and Trevisan M (1995) A soil water-balance 'bucket' model for paleoclimatic purposes. 2: Model application to late Holocene climate. *Ecological Modelling* 82: 131–138.
- Harrison SP, Prentice IC, and Guiot J (1993) Climatic controls on Holocene lake-level changes in Europe. *Climate Dynamics* 8: 189–200.
- Harrison SP, Yu G, and Vassiljev J (2002) Climate changes during the Holocene recorded by lakes from Europe. In: Wefer G, Berger WH, Behre K-E, and Jansen E (eds.) *Climate Development and History of the North Atlantic Realm*, pp. 191–204. Berlin: Springer.
- Haxeltine A, Prentice IC, and Cresswell ID (1996) A coupled carbon and water flux model to predict vegetation structure. *Journal of Vegetation Science* 7: 651–666.
- Henderson-Sellers B (1984) *Engineering Limnology*. London: Pitman.
- Henderson-Sellers B and Davies AM (1989) Thermal stratification modeling for oceans and lakes. In: Tien CL and Chawla TH (eds.) *Annual Review of Numerical Fluid Mechanics and Heat Transfer*, pp. 86–156. New York: Hemisphere.
- Hostetler SW (1991) Simulation of lake ice and its effect on the late Pleistocene evaporation rate of Lake Lahontan. *Climate Dynamics* 6: 43–48.
- Hostetler SW and Bartlein PJ (1990) Simulation of lake evaporation with application to modeling lake level variations of Harney-Malheur Lake, Oregon. *Water Resources Research* 26: 2603–2612.
- Hostetler SW and Benson LV (1990) Paleoclimatic implications of the high stand of Lake Lahontan derived from models of evaporation and lake level. *Climate Dynamics* 4: 207–217.
- Kutzbach JE (1980) Estimates of past climate at Paleolake Chad, North Africa, based on a hydrological and energy-balance model. *Quaternary Research* 14: 210–223.
- Kutzbach JE and Gallimore RG (1988) Sensitivity of a coupled atmosphere/mixed layer ocean model to changes in orbital forcing at 9000 years BP. *Journal of Geophysical Research* 93: 803–821.
- Magny M (1992) Holocene lake-level fluctuations in Jura and the northern subalpine ranges, France: Regional pattern and climatic implications. *Boreas* 21: 319–334.
- Manabe S (1969) Climate and the ocean circulation. 1: The atmospheric circulation and the hydrology of the Earth's surface. *Monthly Weather Review* 97: 739–774.
- Mason IM, Guzkowska MAJ, Rapley CG, and Street-Perrott FA (1994) The response of lake levels and areas to climatic change. *Climatic Change* 27: 161–197.
- Nikolaidis NP, Hu H-L, Escedy C, and Lin JD (1993) Hydrologic response of freshwater watersheds to climatic variability: Model development. *Water Resources Research* 29: 3317–3328.
- Patterson JC and Hamblin PF (1988) Thermal simulation of a lake with winter ice cover. *Limnology and Oceanography* 33: 323–338.
- Prentice IC, Guiot J, and Harrison SP (1992) Mediterranean vegetation, lake levels and palaeoclimate at the Last Glacial Maximum. *Nature* 360: 658–660.
- Street-Perrott FA and Harrison SP (1985) Lake levels and climate reconstruction. In: Hecht AD (ed.) *Paleoclimate Analysis and Modeling*, pp. 291–340. New York: Wiley.
- Vassiljev J (1998) The simulated response of lakes to changes in annual and seasonal precipitation: Implication for Holocene lake-level changes in northern Europe. *Climate Dynamics* 14: 791–801.
- Vassiljev J, Harrison SP, and Guiot J (1998) Simulating the Holocene lake-level record of Lake Bysjön, southern Sweden. *Quaternary Research* 49: 62–71.
- Vassiljev J, Harrison SP, Hostetler S, and Bartlein PJ (1994) Simulation of long-term thermal characteristics of three Estonian lakes. *Journal of Hydrology* 163: 107–123.
- Vavrus SJ, Wynne RH, and Foley JA (1996) Measuring the sensitivity of southern Wisconsin lake ice to climate variations and lake depth using a numerical model. *Limnology and Oceanography* 41: 822–831.

Origins of Lichenometry

Lichenologist Roland Beschel is widely recognized as being the first to show how lichens could be used to estimate the timing of glacier retreat and other geomorphic activity (e.g., Webber and Andrews, 1973). While doing lichenological research in cemeteries, Beschel noticed that larger lichens were found on older gravestones. It was clear to him that individuals of certain lichen species were slow-growing and long-lived and grew radially outward to form crust-like, almost circular patches on the rocks. When he saw many of these same crustose lichen species growing downvalley from retreating glaciers, Beschel reasoned that the diameter of these lichens could be used as a relative measure of the amount of time that glacial debris had been exposed, stable, and available for lichen colonization and growth. Though Beschel died before he could fully explore the potential of his lichen dating technique, his early work (e.g., Beschel, 1950, 1961) has sparked considerable interest among scientists who work in lichen-rich polar and alpine areas. Following Beschel's lead, hundreds of earth scientists have used lichenometry to estimate the timing of geomorphic events on such features as raised beaches, debris flow and rockfall deposits, and morainic ridges, as well as to estimate the age of archaeological features. Potential users of lichenometry typically learn about the technique by studying the detailed reviews that trace the historical development of lichenometry and describe a variety of sampling designs that users can adapt to fit their particular application and field circumstances (e.g., Innes, 1985, 1988; Locke et al., 1979; Matthews, 1994; McCarthy, 2002; Noller and Locke, 2000; Webber and Andrews, 1973).

Lichenometry is now a widely accepted and much used dating technique that continues to evolve. Prior to the 1980s, thallus measurement was often done with rulers, and the location and size of a few of the largest thalli were reported. Since the 1980s, lichenometry has evolved from being a descriptive report of thallus sizes and tentative age estimates to reports that assign ages and error bars using approaches that increasingly rely on trend fitting software, polynomial equations, and assumptions that defy articulation. Initially, the shift toward statistical manipulation was seen in paleoseismic applications (e.g., Bull and Brandon, 1998; Smirnova and Nikonov, 1990). Now, few studies use the sort of descriptive lichenometry that was commonplace in the 1980s. Bradwell (2010) has rightly noted that this statistical shift has led to less transparency in the way lichenometry is practiced. It should also be noted that today there is no scientific debate, and little research focused on the fundamental assumptions of lichenometry. This is evidenced by the small number of papers that have offered constructive criticism of lichenometry (e.g., Jochimsen, 1966; Kirkbride and Dugmore, 2001; McCarthy, 1999; Schuhwerk, 1992).

The following sections provide a general overview of lichenometry with a special focus on the rationale, assumptions, and especially the strengths and limitations of the more

common approaches to lichenometry. The article concludes with some practical suggestions and a brief critique of lichenometry. Potential users are advised to consult specific references for more detailed information about sampling designs and data analysis.

Rationale and Assumptions

Lichenometry is based on several key assumptions.

- (a) A surface can be no younger than the oldest individual lichen (thallus) growing on it. Thus, while the age of an individual lichen is usually unknown, the lichenometric age is a minimum estimate of surface age.
- (b) Thalli with circular or nearly circular outlines are individuals that grew radially outward from a central point. Thalli that do not have a circular outline are not suitable for lichenometry because they may have developed by coalescence of neighboring lichens.
- (c) The radial growth of lichens is sporadic, but is assumed to be constant when averaged over the long term.
- (d) The largest lichen is not necessarily the oldest, but it is among the first to have become established after all others were stripped from the surface by the last major disturbance (e.g., glacial activity).
- (e) The largest lichen occupies an optimal site for growth and is the fastest growing in the community.
- (f) The diameter of the largest or several of the largest lichen thalli with oval or circular thallus outlines can be used to estimate the time elapsed since colonization began.

These basic assumptions were broadly outlined in Beschel's early publications. Some are not used in certain approaches to lichenometry and some have been challenged by critics as being biologically untenable (e.g., McCarthy, 1999). The assumptions are seldom stated and are occasionally modified by users so as to permit the use of certain statistical tests. Additional assumptions about errors in search, thallus measurement, or species identification are also implicit. These potential sources of error are of significant concern especially if field work is done by a team of people and/or involves *Rhizocarpon geographicum* (L.) DC. and some of its look-alikes that may not grow at similar rates. While users may be reassured if a lichenologist identifies voucher samples of the general types of lichens that were measured in the field, it is difficult to evaluate the thoroughness of the field search, the representativeness of the voucher samples, the accuracy and precision of the measurements, or the closeness of the age determinations. As few, if any, studies use quantitative testing to assess the impact of the potential sources of error on their age estimates, there is little reason to place trust in the reproducibility and accuracy of lichenometric ages. At best, lichenometric ages are minimum estimates of surface age.

Approaches to Lichenometry

Unlike many other dating techniques, lichenometry does not use a single standardized approach. Consequently, almost anyone – regardless of their understanding of lichen biology – can generate and manipulate lichen size data sets so as to produce statistically reproducible age estimates. Unfortunately, statistical reproducibility (e.g., Matthews, 1974, 1975) does not guarantee accuracy or precision, and the many potential errors associated with data collection, interpretation, and growth rate calibration cast doubt as to how closely some lichenometric ages match the true age of the surface (e.g., Kirkbride and Dugmore, 2001). Given such concerns, it is wise to clearly report ages generated by lichenometry as ‘lichenometric ages’ and place more trust in age estimates that are supported by independent dating techniques.

Each new user of lichenometry seems to adopt a slightly different approach to sampling and data interpretation, but most studies use a variant of two or three ‘schools’ of lichenometry. The following sections outline these basic ‘schools’ or approaches.

Relative Dating Using Lichens

Lichens can be used to provide either a relative or a calendric (semiabsolute) age for a surface. For example, where lichen cover has been disturbed by a flood or other geomorphic event so that there is a sharp demarcation between lichen-free and lichen-rich areas (i.e., a lichen trimline), then younger and older surfaces can be readily identified (Figure 1). Assigning relative ages in such settings does not require knowledge of lichen growth rate and assumes that the dominant geomorphic process(es), not microclimate or other factors, are solely responsible for obvious differences in lichen cover. This sort of relative age dating with lichens is easily done and is widely accepted, but it is best restricted to situations where there are clear lichen trimlines (e.g., Timoney and Marsh, 2004). If relative age dating is to be attempted where there are only

slight differences in lichen cover, it is both necessary to accurately measure lichen cover in a systematic way and to find clear evidence that changes in percentage cover are due to specific geomorphic events. In general, percentage lichen cover and relative age dating using lichens have a greater chance of success on silicate rocks where lichen populations can be quite large and uniformly distributed. However, these are not well suited for use on carbonates where lichens are sparsely distributed, and colonization and growth rates can be strongly influenced by chemical weathering, microtopography, and nitrogen enrichment at bird perches and rodent burrows (McCarthy and Smith, 1995). The use of percentage lichen cover as an index of age usually involves subjective judgment about where to sample and how to accurately measure cover (e.g., Innes, 1986; McCarthy and Zaniewski, 2001). As lichen cover can vary due to imperceptible differences in microclimate, geochemical differences in substrates, and many factors other than geomorphic disturbance or surface age, the use of percentage cover as a dating tool is not widely trusted or used (e.g., Noller and Locke, 2000, p. 266). Nevertheless, relative age determinations can sometimes be accurately estimated if the successional indicator status of certain lichen species can be established (e.g., Hill, 1994; Schuhwerk, 1992).

Lichen Growth Rate Calibration

A calendar year can be estimated for the stabilization of a moraine or other lichenized substrate if a quantitative link can be made between some measure of lichens at the site and the age of the lichenized substrate. This relationship is often represented by a scatterplot that has traditionally been called a ‘lichen growth curve.’ This is an unfortunate misnomer. The line plotted through the points is not always a curve, and indices of lichen size (e.g., a radius of an imperfect circle) are plotted against surface age. The change depicted by such a plot is a change in the relationship between the substrate age and the lichen size index. This is not equivalent to a lichen



Figure 1 A lichen trimline was formed on these quartzite boulders when the lichen-covered moraine was eroded by a stream. Historical photographs and tree ring dating show that lichen cover on this moraine originated in the late 1890s when the Illecillewaet Glacier retreated from this position. Since *Rhizocarpon geographicum* agg. takes ca. 15 years to become established on recently exposed rocks in the region, the lichen-free area in the abandoned stream channel is less than ca. 15 years old.

growth rate. True lichen growth 'curves' use repeated measurements (e.g., at exactly the same lobe) of the exact same thallus (thalli) over a certain period of time. Workers who rely on indices of thallus or population size (e.g., largest inscribed diameter or largest five or ten thalli at a site) have not tracked historical changes in marked thalli and have a tenuous basis on which to define lichen growth rates or infer biological process from the shape of thallus size distributions. However, it is commonplace in lichenometric dating to report that the lichenometric 'growth curve' is an estimate of the long-term radial growth rate of lichens that has been 'calibrated' at that site. The radial growth trend of crustose lichens is typically represented by a scatterplot that pairs the diameter or some other measure of lichen size (e.g., longest axis, shortest axis, or diameter of the largest inscribed circle) with the approximate age of the substrate surface. As ages estimated by lichenometry are an approximation of the amount of time elapsed since lichens first colonized the surface, the true age of the surface may be older.

Lichenologists have long used repeated measurement of marked thalli to estimate lichen growth rates and document short-term changes in lichen communities (e.g., Smith, 1995). Direct measurement of lichen growth can be done using calipers, pen and ink tracings on acetate overlays, photogrammetry and/or digital photographs, and Geographic Information Systems (e.g., Benedict, 1990, 2008; Brabyn et al., 2005; Hooker and Brown, 1977; Proctor, 1983). In each case, changes in lichen size are measured relative to painted markers or rock crystals. As the radial growth of epilithic lichens is often slow (e.g., $<0.5 \text{ mm year}^{-1}$) and uneven, direct measurement must be done using repeated measurements to acceptable tolerances (e.g., $\pm 0.01 \text{ mm}$), and they should ideally be done along several radii at many thalli. The slight swelling that occurs when thalli are wet and the distortion associated with photographic lenses, parallax error, and problems in exactly repositioning a camera for repeat photography are important potential sources of error that can be quantitatively assessed and minimized. However, this is tedious and exacting work and few workers use this approach for the purpose of assigning lichenometric ages. Even when detailed records of radial growth are compiled, very few workers have attempted to compare annual growth rate measurements to indirect estimates of long-term growth trends. Thus, it is unclear whether modern growth rates are similar to those in the past. In one recent study, McCarthy (2003) showed that while annual growth varied within and between *Rhizocarpon* thalli, linear extrapolation of mean annual growth rates collected over a 4-year period closely matched century-long growth as estimated by a lichen growth curve that was developed from measurements on historical and tree-ring dated surfaces. This finding shows that while tedious and time consuming, repeated measurement and linear extrapolation of growth rates can, in some cases, provide a reasonably accurate way to calibrate a lichen growth curve.

Workers should also note that measurements of growth rate derived by one approach (e.g., direct measurement to painted markers) are not interchangeable with measurements done using another approach (e.g., repeated measurement of change along random radii). Figure 2 and Tables 1 and 2 illustrate this. Despite this and other concerns (e.g., small sample sizes, accuracy and precision of the measurement method), repeated measurement of lichen growth and the establishment of

long-term monitoring plots are the only way to generate true growth rate data that can be used to address some of the substantial questions that remain about the nature of growth rates at various stages in the lichen life cycle (e.g., Armstrong, 1983; Armstrong and Smith, 1987). Direct measurement may also be the only way to calibrate lichen growth curves in remote regions where dating controls are unavailable.

Lichen 'Growth Curves'

The term 'growth curve' is an unfortunate misnomer. It was introduced to lichenometry in Roland Beschel's early work to describe the long-term radial growth trend in crustose lichens. Beschel estimated this growth trend as being a curved line on a lichen size-age scattergraph. This curved line is readily produced using 'traditional' and some statistical approaches to lichenometry (e.g., the fixed quadrat or the fixed-area largest lichen (FALL) approach to lichenometry: Bull and Brandon, 1998; Savoskul and Zech, 1997). However, lichen growth rates estimated by some statistical approaches to lichenometry tend to plot as straight lines. Linear growth 'curves' result when only older control points (e.g., >100 years) are used or if researchers use assumptions underlying randomized sampling designs and linear regression analysis (e.g., Matthews, 1974; McCarroll, 1994).

In general, lichenometrically useful species begin growth after a colonization lag interval (ecesis) that can be as short as a decade or longer than a century at some micro sites. Initially, there is a phase of rapid radial growth (sometimes described as logarithmic growth or the 'great period,' e.g., Beschel, 1950). This phase lasts for a few decades. A prolonged period of steady, linear growth follows and may last for centuries. While it is unknown how long an epilithic lichen thallus can live, growth curves calibrated using radiocarbon ages of buried organics are sometimes used to estimate lichen growth at the surface of deposits that are several thousand years old (e.g., Miller and Andrews, 1972). Growth curves constructed in this way give the false impression that individual lichens can live for millennia and that even in old age, thallus growth rates never drop to zero. Instead, the finding of large thalli with nearly circular outlines on old surfaces should be seen as circumstantial evidence of lichen life spans. Some exceptionally large lichen thalli are genetically complex mosaics. The weathering and loss of the central part of large lichen thalli is common in old age and such areas can be recolonized in such a way that new and old parts of the thallus become indistinguishable. While old and potentially useful as indicators of substrate age, these thalli may not be individuals in a genetic or biological sense. Similarly, extrapolation of growth rates beyond control points and/or from short-term measurements is a questionable way to 'model' growth rates. However, the apparent scarcity of acceptable alternatives calls attention to the need for greater empirical research in this regard.

The 'Traditional' Approach to Lichenometry

Roland Beschel and early practitioners of lichenometry used a simple search and measurement strategy to indirectly estimate lichen growth rate. This 'traditional' approach involves a

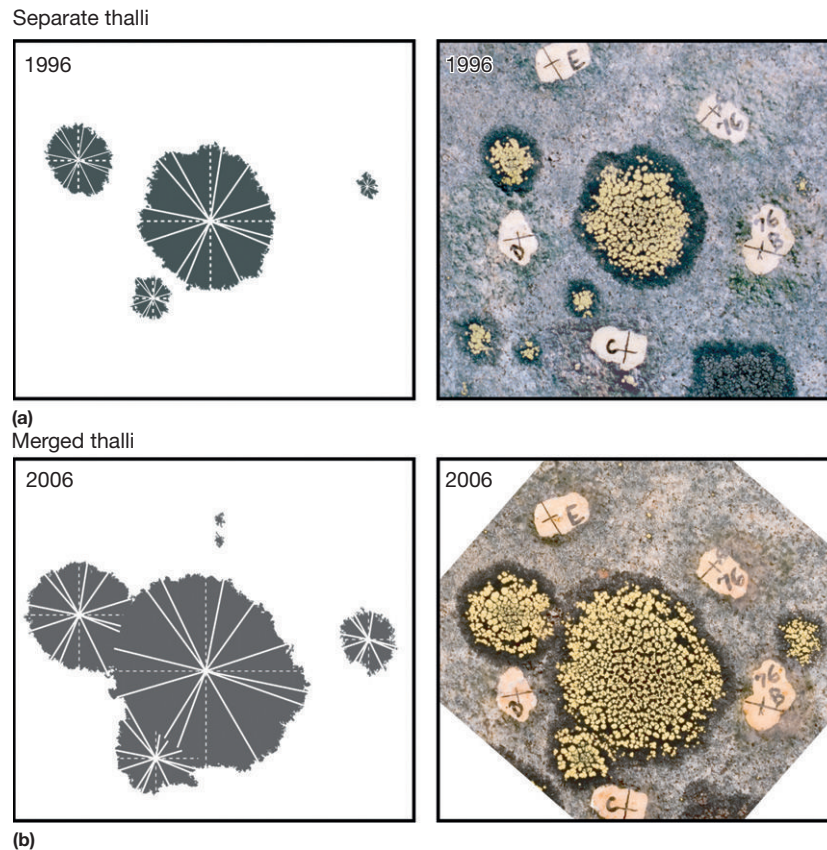


Figure 2 Comparative photographs show marked lichen thalli in the summers of 1996 (a) and 2006 (b). The original photographs (right) have been paired with a bitmap of each photograph (left). Painted crosshairs surround thallus #76 (A–E, see above right) and provide fixed points where digital calipers were used to measure distances between the crosshairs and the thallus margin. The lines shown on the bitmaps (a and b, see above left) show six randomly chosen diameters (12 radii) where change in thallus size was photogrammetrically measured. These illustrations and the data in [Tables 1](#) and [2](#) show that the shape and growth of these thalli is not well represented by change in a single radius or diameter. Use of the mean of 12 radii or five control points should ensure that the growth indices are representative (see [Tables 1](#) and [2](#)). Users of lichenometry should, but do not, quantify thallus circularity, thallus area or thallus perimeter.

complete search of all exposed surfaces on a deposit and the measurement of the diameter of a few of the largest lichens in the search area. In some cases, parts of a landform may be searched and analyzed as separate units. This is especially common in the dating of moraine complexes at valley glaciers where individual morainic ridge segments have been crosscut by streams or are suspected of having proximal or distal surfaces that became ice-free and stable at different times.

In order to construct a lichen growth curve using the traditional approach, several growth ‘control points’ are developed using scatterplots that match the sizes of the largest lichen thalli with the known ages of their substrates. The substrates are considered to be growth control surfaces and their ages are known by independent dating methods (e.g., tree-ring dating, archival documents or photographs, etc.). The resulting data (control points) are used to calibrate/estimate the radial growth trend over the longer term. A growth trend line is then fit to the points by eye so that accepted control points fall on the line. Subjective judgments rather than statistical tests are used to decide which points are anomalous and which best reflect the maximum growth rate of thalli in mesic or xeric habitats ([Figure 3](#)). As the likelihood of finding a larger thallus tends to increase as microhabitat variety and search area

size increases, greater trust is generally placed in larger control surfaces. Tombstones, buildings, and monuments are sometimes used as lichen growth control points, but they are often at lower elevations than where lichenometry will be used and may not closely reflect rates of lichen establishment and growth in the alpine. Accordingly, authors usually provide careful descriptions and evaluations of each control point and explain why they accept or reject them. The traditional approach to lichenometry involves a thorough (complete) search of all surfaces on a landform unit. Lichens in moist and mossy sites are either not measured or are measured for comparative purposes and excluded as anomalous. In the traditional and other approaches to lichenometry, thalli with circular outlines are assumed to be individual lichens that arose by radial growth from a single point. Despite the fact that most epilithic lichens infrequently have perfectly circular thallus outlines, users of lichenometry assume that noncircular thallus outlines are ‘irregular’ and must have formed by the merger of neighboring thalli, violating at least one of the basic assumptions of lichenometry.

Since Beschel’s first use of lichenometry in glacier forefields, the search and measurement strategies and the data reporting formats used in the traditional approach to lichenometry have

Table 1 Raw measurement data obtained for one thallus (#76) shown in [Figure 2](#). A random number generator was used to select six angles at which the radius would be measured in the upper half of the thallus. These radial lines were extended into the lower half of the image to make a total of 12 radii where measurements were taken. The distance along each radius was first measured in pixels with the Ruler tool in Adobe® Photoshop® CS2. Distances in pixels were converted to millimeters by dividing pixel distances by the resolution of this image (944.882 pixels per cm) and multiplying by 10. As this image was at 1/3 life size, the raw values for radius distance, area, and perimeter were then multiplied by this scale factor (3) to convert to life size. These data are for illustrative purposes only. Space limitations do not permit a full articulation of the methods used to eliminate distortion and validate measurement accuracy and precision. Note that the 1999 data are used because, as [Figure 2](#) shows, thallus #76 had merged with nearby thalli in 2006

Year	Random no. (angles)	Radius distance (pixels)	Raw radius distance (mm)	True radius distance (mm)	Area (cm ²)	Perimeter (cm)	Circularity	Scale factor
1996	12	269.87	2.86	8.57	0.84	14.64	0.15	3
	51	290.02	3.07	9.21				
	62	305.53	3.23	9.70				
	98	290.06	3.07	9.21				
	125	337.36	3.57	10.71				
	162	265.96	2.81	8.44				
	193	266.93	2.83	8.48				
	202	269.63	2.85	8.56				
	238	286.62	3.03	9.10				
	292	270.93	2.87	8.60				
	302	253.87	2.69	8.06				
	340	271.09	2.87	8.61				
1999	12	337.67	3.57	10.72	1.55	20.61	0.05	3
	51	336.09	3.56	10.67				
	62	351.77	3.72	11.17				
	98	340.97	3.61	10.83				
	125	372.86	3.95	11.84				
	162	296.21	3.13	9.40				
	193	310.26	3.28	9.85				
	202	304.61	3.22	9.67				
	238	312.01	3.30	9.91				
	292	318.10	3.37	10.10				
	302	336.94	3.57	10.70				
	340	317.39	3.36	10.08				

Table 2 Comparison of thallus growth data photogrammetrically measured using ImageJ software at 12 radii and measurements done in the field using digital calipers and five fixed points. The means are similar, but there is a wide scatter in the data. The growth (of thallus #76, [Figure 2](#)) is not well represented by any one radius or by measurement to any one fixed point. Note that these data cover a 3-year growth interval, not the 10-year interval shown in [Figure 2](#)

Calipers			ImageJ		
Number of measurement points (fixed points)	Mean growth (mm)	Standard deviation (mm)	Number of measurement points (random sample of radii)	Mean growth (mm)	Standard deviation (mm)
5	1.29	0.43	12	1.47	0.51

become somewhat standardized. By the mid-1970s, it became common practice to report the mean size of five or more of the largest thalli on individual moraine segments and size differences of 10–20% between the largest thalli became an accepted way to identify/accept/reject thalli (e.g., [Benedict, 1967](#); [Webber and Andrews, 1973](#)). This continues today with flexible plastic rulers or calipers used to measure the maximum diameter, minimum diameter, and the longest and/or the shortest axis of thalli as the index of age. Over the years, it has also become accepted practice to measure thalli that belong

to broad, morphologically similar, aggregated species groups (e.g., *Rhizocarpon geographicum* agg.; [Innes, 1985](#)). This is clearly a disturbing source of error if some of the aggregated species grow at different rates and are indistinguishable in the field. However, the use of species aggregates is often a practical necessity because positive identification of crustose lichens to the species level involves the use of chemical stains and microscopic examination of spores ([Benedict, 1988](#)).

Typically, lichen size data are summarized using tables and maps that show where the size data were collected on each



Figure 3 Nitrogen from the feces of rodents and birds has created a dense lichen cover of *Xanthoria elegans* (Link) Th. Fr. and other nitrophilous lichen species on these granitic rocks in Canada's Northwest Territories. Most other rocks in this area have only one or two lichen thalli. Such sites are easily identified in the field and are excluded from lichenometric searches.

landform (e.g., showing lichen sizes on the proximal vs. the crest or distal slope of a moraine). Growth curves derived using this traditional approach seldom extend for more than three centuries. The curved (nonlinear) growth curve, the complete search strategy, the use of a single largest thallus as a control point, and simple descriptive statistics are hallmarks of the traditional approach.

It would seem that when carefully executed, the traditional and statistical approaches to lichenometry provide equally acceptable and accurate age estimates (e.g., [Karlén and Black, 2002](#)). However, each approach has its advantages and limitations. For example, the traditional approach is not well suited to situations that require an exhaustive search of large areas or where there is a need to establish minimum ages for episodic or sequential events that intermix old and newer materials (e.g., as occurs in seismically triggered rockfalls and in debris flows). However, the traditional approach is well suited to the study of small- or moderate-sized morainic ridge complexes at alpine glaciers and perhaps the only approach that has been successfully used in carbonate terrain.

Statistical Approaches to Lichenometry

Statistical approaches to lichenometry use parametric statistics not only to characterize, but also to infer thallus sizes with reference to an ideal (statistically normal or normalized) distribution of thallus size data. Many researchers have used

random, stratified random or plot-based sampling designs and parametric statistical tests to develop lichen growth curves and assign lichenometric ages. These approaches generate reproducible and, in some cases, highly accurate age estimates with well-defined errors of the estimate (e.g., [Bull and Brandon, 1998](#)).

There are many strengths and potential advantages of using a statistical approach to lichenometry. For example, the use of replicated plots (e.g., 10 m × 10 m plots) makes it possible to collect representative samples for large areas and facilitate comparisons within and between study sites. Statistical approaches have also seen considerable use in the dating of large glacier forefields (e.g., [Larocque and Smith, 2004](#)), development of growth curves representative of broad regions (e.g., [Solomina and Calkin, 2003](#)), and paleoseismic applications (e.g., [Bull and Brandon, 1998](#)) where individual blocks were displaced by earthquake-generated rockfalls. While assumptions about thallus growth are generally unchanged from those used in the traditional approach to lichenometry, some assumptions are adjusted or added in order to satisfy the requirements of statistical tests (e.g., each thallus has an equal opportunity of being measured, a normal distribution of measurement error exists across the population, etc.).

Most statistical approaches to lichenometry have used linear regression to plot growth curves, but methodologies can widely differ in their sampling designs. For example, estimates of deglaciation dates have been derived using a stratified random sampling design with or without nested quadrats

(e.g., Matthews, 1994), and transects and plots are sometimes located along the center line of glacier retreat (e.g., Larocque and Smith, 2004). In paleoseismic work with the FALL sampling design (e.g., Bull and Brandon, 1998), sampling is restricted to smooth rock surfaces on boulders of a fixed size. This reduces the amount of variance that may be due to microtopography and microhabitat. The FALL method also assigns a subjective rank to thallus shape/regularity and assigns statistical confidence intervals so as to define the upper and lower range of the FALL estimated ages. The assumption that lichen communities (cohorts) usually have a unimodal distribution of thallus sizes allows users of the FALL approach to statistically smooth (normalize) lichen size distributions and assign ages to thallus size distributions. Unfortunately, these assumptions have not yet been supported by robust biological data sets.

Practical Suggestions and Concerns About Lichenometry

The credibility of lichenometry as a dating technique is being eroded as author after author ignores the need to validate the accuracy and precision of thallus measurements, species identifications, and sampling thoroughness. Analytical chemists routinely use certified reference materials and repeated tests to calibrate their analyses. Why do users of lichenometry not use data and simple tests to quantitatively define their measurement accuracy and precision, and verify the certainty of their species identifications? Lacking accuracy and precision data, how would we know if our ages, models, and assumptions were precisely wrong? Perhaps the biggest challenge we face in lichenometry is the lack of scientific rigor. It would seem that authors/editors and readers do not ask or seek answers for such basic questions as: How reproducible is this sampling method? Has the author and/or journal reported and archived enough information to ensure that another worker can easily find and resample/rephotograph these thalli and independently test the findings? Has the author provided sufficient data to demonstrate the accuracy and precision of thallus measurements? Has the author adequately demonstrated that (e.g., 90%) of the lichens were properly identified to species or group level? There are many ways by which we might improve the scientific rigor of lichenometric work. For example, certified reference material could be circulated and used to determine how well individual workers can positively identify lichens to species or group level, thalli of known dimensions could be used to ascertain the accuracy and reproducibility of photogrammetric work, and data repositories could be used to archive complete data sets and facilitate follow-up studies by independent workers. With the possible exception of archival storage, these things are not yet part of our research culture. However, workers could, at the very least, do repeated measures of marked thalli on different days and use descriptive statistics to show the variability in measurements within and between operators. Species identification success rate could also be assessed if workers collected and identified more than a few voucher specimens and reported the success of their species identification as compared to the identifications made by a paid, professional lichenologist. If these suggestions are followed, we might learn whether

workers in some areas have greater success or are more skilled in their identification of lichen species, and/or if measurement and identification errors vary with thallus size or on landforms of different age. Statistical procedures should also be turned inward so that workers evaluate and report on the likelihood that untested and unknown errors invalidate their findings.

My final observations focus on the use and misuse of the words 'growth curve.' The only true growth curves are plots that depict changes in the size of thalli that have been repeatedly measured in the same way at the same locations over successive years. As diametric growth is not uniform along any axis in nonplacoidial thalli (e.g., the yellow-green *Rhizocarpon*s), our ability to find and precisely remeasure a diameter (or radius) is challenged by shape changes at the thallus margin. If workers claim that they have accurately measured a diameter, then they must use data and tests to show that this same diameter measure can be reproduced to an acceptable accuracy and precision (e.g., on other days). Long-term studies face major challenges. They must prove that their diameter measure is highly reproducible in the short term and that the measure is accurate and highly reproducible at the decadal scale. Reliance on one or two thallus diameters as an index of thallus growth is tenuous and foolhardy. Diameter data are meaningless if we do not know how circular the thallus outline is/was at the time of measurement. Lacking this information, how can we determine the representativeness of diametric data and how can a worker justify its use? Ultimately, workers, readers, and editors must realize that the use of unvalidated and unrepresentative data sets can lead to precisely the wrong conclusions.

Future Directions in Lichenometry

Lichenometry is an inexpensive, widely adaptable, and invaluable tool for use in estimating surface ages in lichen-dominated landscapes. Its continued use is virtually guaranteed as is the likelihood that users will ignore critics and continue to use unvalidated measurements to generate unvalidated ages. In this digital age, it seems reasonable to suggest that the future will see increased use of digital imaging and/or customized computer software to assist in the development of 'growth curves.' Perhaps with concomitant advances in lichenology, we may gain better insights into some questions: Why are growth curves curved? What is the maximum life span of a *Rhizocarpon* thallus? Does the presence/absence of certain lichen species clearly indicate surface age? Are modern growth rates identical to those in the recent past? Unfortunately, history suggests that users of lichenometry will not soon provide substantive answers to these basic questions.

See also: [Glacial Landforms: Evidence of Glacier Recession; Moraine Forms and Genesis. Glacial Landforms, Tree Rings: Dendrogeomorphology; Dendroglaciology. Glaciations: Neoglaciation in Europe; Neoglaciation in the American Cordilleras. Introduction: History of Dating Methods. Permafrost and Periglacial Features: Rock Glaciers and Protalus Forms.](#)

References

- Armstrong RA (1983) Growth curve of the lichen *Rhizocarpon geographicum*. *New Phytologist* 94: 619–622.
- Armstrong RA and Smith SN (1987) Development and growth of the lichen *Rhizocarpon geographicum*. *Symbiosis* 3: 287–300.
- Benedict JB (1967) Recent glacial history of an alpine area in the Colorado Front Range, USA. 1. Establishing a lichen-growth curve. *Journal of Glaciology* 6: 817–832.
- Benedict JB (1988) Techniques in lichenometry: Identifying the yellow *Rhizocarpons*. *Arctic and Alpine Research* 20: 285–291.
- Benedict JB (1990) Experiments on lichen growth. 1. Seasonal patterns and environmental controls. *Arctic and Alpine Research* 22: 244–254.
- Benedict JB (2008) Experiments on lichen growth, III. The shape of the age–size curve. *Arctic, Antarctic, and Alpine Research* 40: 15–26.
- Beschel RE (1950) *Flechten als Altersmaßstab rezenter Moränen*. Zeitschrift für Gletscherkunde und Geologie. N.F. 1: 152–162. (Translated by W. Barr as 'Lichens as a measure of the age of recent moraines.' *Arctic and Alpine Research* 5: 303–309).
- Beschel RE (1961) Dating rock surfaces by lichen growth and its application to glaciology and physiography (lichenometry). In: Raasch GO (ed.) *Geology of the Arctic*, vol. 2, pp. 1044–1062. Toronto: University of Toronto Press.
- Brabyn L, Green A, Beard C, and Seppelt R (2005) GIS goes nano: Vegetation studies in Victoria Land, Antarctica. *New Zealand Geographer* 61: 139–147.
- Bradwell T (2010) Preface: Lichenometry in subpolar environments. *Geografiska Annaler* 92: 1–2.
- Bull WB and Brandon MT (1998) Lichen dating of earthquake-generated regional rockfall events, Southern Alps, New Zealand. *Geological Society of America Bulletin* 110: 60–84.
- Hill DJ (1994) The succession of lichens on gravestones: A preliminary investigation. *Cryptogamic Botany* 4: 179–186.
- Hooker TN and Brown DH (1977) A photographic method for accurately measuring the growth of crustose and foliose saxicolous lichens. *Lichenologist* 9: 65–75.
- Innes JL (1985) Lichenometry. *Progress in Physical Geography* 9: 187–254.
- Innes JL (1986) The use of percentage cover values in lichenometric dating. *Arctic and Alpine Research* 18: 209–216.
- Innes JL (1988) The use of lichens in dating. In: Galun M (ed.) *Handbook of Lichenology*, pp. 75–92. Boca Raton, FL: CRC Press.
- Jochimsen M (1966) Ist die Grösse des Flechten thallus wirklich ein brauchbarer Massstab zur Datierung von glazialgeomorphologischen Relikten? *Geografiska Annaler* 48A: 157–164 (Translated by W. Barr as 'Does the size of lichen thalli really constitute a valid measure for dating glacial deposits?' *Arctic and Alpine Research* 5: 417–424).
- Karlén W and Black JL (2002) Estimates of lichen growth rate in northern Sweden. *Geografiska Annaler* 84A: 225–232.
- Kirkbride MP and Dugmore AJ (2001) Can lichenometry be used to date the 'Little Ice Age' glacial maximum in Iceland? *Climatic Change* 48: 151–167.
- Larocque SJ and Smith DJ (2004) Calibrated *Rhizocarpon geographicum* growth curve for the Mount Waddington area, British Columbia Coast Mountains, Canada. *Arctic, Antarctic, and Alpine Research* 36: 407–418.
- Locke WW III, Andrews JT, and Webber PJ (1979) A Manual for Lichenometry. London: British Geomorphological Research Group.
- Matthews JA (1974) Families of lichenometric dating curves from the Storbreen gletschervorfeld, Jotunheimen, Norway. *Norsk Geografisk Tidsskrift* 28: 215–235.
- Matthews JA (1975) Experiments on the reproducibility and reliability of lichenometric dates, Storbreen gletschervorfeld, Jotunheimen, Norway. *Norsk Geografisk Tidsskrift* 29: 97–109.
- Matthews JA (1994) Lichenometric dating: A review with particular reference to 'Little Ice Age' moraines in southern Norway. In: Beck C (ed.) *Dating in Exposed and Surface Contexts*, pp. 185–212. Albuquerque, NM: University of New Mexico Press.
- McCarroll D (1994) A new approach to lichenometry: Dating single age and diachronous surfaces. *The Holocene* 4: 383–396.
- McCarthy DP (1999) A biological basis for lichenometry? *Journal of Biogeography* 26: 379–386.
- McCarthy DP (2002) Lichenometry. In: Nimis PL, Scheidegger C, and Wolseley PA (eds.) *Monitoring with Lichens – Monitoring Lichens*, pp. 379–383. The Netherlands: Kluwer Academic Publishers.
- McCarthy DP (2003) Estimating lichenometric ages by direct and indirect measurement: A case study of *Rhizocarpon agg.* at the Illecillewaet Glacier, British Columbia. *Arctic, Antarctic, and Alpine Research* 35: 203–213.
- McCarthy DP and Smith DJ (1995) Growth curves for calcium-tolerant lichens in the Canadian Rocky Mountains. *Arctic and Alpine Research* 27: 290–297.
- McCarthy DP and Zaniewski K (2001) Digital analysis of lichen cover: A technique for use in lichenometry and lichenology. *Arctic, Antarctic, and Alpine Research* 33: 107–113.
- Miller GH and Andrews JT (1972) Quaternary history of northern Cumberland Peninsula, east Baffin Island, N.W.T. Canada. Part VI. Preliminary lichen growth curve for *Rhizocarpon geographicum*. *Geological Society of America Bulletin* 83: 1133–1138.
- Noller JS and Locke WW (2000) Lichenometry. In: Noller JS, Sowers JM, and Lettis WR (eds.) *Quaternary Geochronology: Methods and Applications. AGU Reference Shelf*, vol. 4, pp. 261–272. Washington DC: American Geophysical Union.
- Proctor MCF (1983) Sizes and growth-rates of thalli of the lichen *Rhizocarpon geographicum* on the moraines of the Glacier de Valsorey, Valais, Switzerland. *Lichenologist* 15: 249–261.
- Savoskul OS and Zech W (1997) Holocene glacier advances in the Topolovaya Valley, Bystrinskiy Range, Kamchatka, Russia, dated by tephrochronology and lichenometry. *Arctic and Alpine Research* 29: 143–155.
- Schuhwerk F (1992) Die Berücksichtigung der Ökologie in der Lichenometrie: Datierung mit Sukzessionsstadien von Flechtengesellschaften. *Stuttgarter Geographische Studien* 117: 161–175.
- Smirnova TY and Nikonov AA (1990) A revised lichenometric method and its application dating great past earthquakes. *Arctic and Alpine Research* 22: 375–388.
- Smith RIL (1995) Colonization by lichens and the development of lichen-dominated communities in the maritime Antarctic. *Lichenologist* 27: 473–483.
- Solomina O and Calkin PE (2003) Lichenometry as applied to moraines in Alaska, U.S.A., and Kamchatka, Russia. *Arctic, Antarctic, and Alpine Research* 35: 129–143.
- Timoney KP and Marsh J (2004) Lichen trimlines in northern Alberta: Establishment, growth rates, and historic water levels. *Bryologist* 107: 429–440.
- Webber PJ and Andrews JT (1973) Lichenometry: A commentary. *Arctic and Alpine Research* 5: 295–302.

Loess Deposits: Origins and Properties

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Introduction

Loess is an eolian (windblown) sediment that is an important archive of Quaternary climate changes. It may provide one of the most complete terrestrial records of interglacial–glacial cycles. Loess is unusual as a record of Quaternary climate change because it is one of the few sediments that is deposited directly from the atmosphere. Thus, it is a geologic deposit that contains a record of atmospheric circulation and can be used to reconstruct synoptic-scale paleoclimatology. Loess is also unusual in that it can be dated directly using ‘trapped electron’ or luminescence methods that require only the sediment itself. Commonly, loess deposits are not homogenous sediments, but most contain buried soils, or paleosols. It is the combination of both unaltered loess deposits and intercalated paleosols that gives this sedimentary record much of its richness as a Quaternary paleoclimate record.

Definition of Loess

Loess can be defined as sediment that has been entrained, transported, and deposited by the wind and is dominated by silt-sized (50–2 μm diameter) particles. Most loess deposits are not composed completely of silt, but also contain measurable amounts of sand (>50 μm) and clay (<2 μm). Nevertheless, loess typically has 60–90% silt-sized particles. To distinguish loess from fine-grained (aerosolic) dust that may have a subtle presence within soils or sediments, loess should be recognizable in the field as a distinctive sedimentary body. It commonly forms a mantle or cover on preexisting landscapes and can be anywhere from a few centimeters to several hundred meters in thickness.

Unlike eolian sands, fluvial sediments, or marine sediments, primary structures in loess are subtle. Some loess deposits have primary bedding structures, such as faint, horizontal laminations and, less commonly, cross-bedding. Many loess deposits are characterized by a massive (as opposed to ‘loose’ or structured) condition. Interparticle binding by clays and/or carbonates often results in considerable material strength and explains the ability of loess deposits to form vertical faces along river or stream banks and road cuts. In fact, some European researchers regard this weak cementation by carbonate as a process called ‘loessification,’ a prerequisite for a sediment to be considered true loess. Most other researchers do not have such a restricted definition of loess, however, and regard all eolian silt that forms

a distinct sedimentary body as loess (Muhs and Bettis, 2003; Pye, 1987). Secondary structures in loess are more common than primary structures, and consist of fractures, burrows, rhizoliths (root casts composed of iron oxides or carbonate), carbonate nodules or concretions, oxidation or reduction streaks or bands, and paleosols. Although mammal fossils are sometimes found in loess deposits, the most common fossils are shells of land snails, which can be powerful paleoclimatic tools (Rousseau and Kukla, 1994).

Spatial Distribution of Loess

Loess covers a significant amount of the Earth’s land surface, perhaps as much as 10% (Pye, 1987). Because of its widespread distribution and favorable texture and mineralogy, it forms some of the world’s most important agricultural soils. In the Eastern Hemisphere, loess is abundant over much of Eurasia (Figure 1). Most loess in Eurasia is distributed in a latitudinal belt between about 40° and 60° N, covering areas south of the limits of continental or mountain glaciers of Quaternary age in western and central Europe, Russia, and Central Asia (Frechen et al., 2003; Pye, 1987; Rozycki, 1991). An important exception is China, where loess covers large areas at lower latitudes that were not near continental ice sheets, mountain ice caps, or valley glaciers (Liu, 1988). Loess is largely absent from subtropical and tropical latitudes of Eurasia.

In the Western Hemisphere, loess is present in both North and South America (Figures 2 and 3). In South America, there are two major loess belts, the Pampas loess in central Argentina and the Chaco loess in northern Argentina; less-certain loess deposits may be located in Bolivia, Paraguay, Uruguay, and Brazil (Zárate, 2003). In North America, loess is found in Alaska and adjacent Yukon Territory (Péwé, 1975), the Palouse area of eastern Washington and adjacent Oregon (Busacca et al., 2004), the Snake River Plain and adjacent uplands of Idaho, the Great Plains region east of the Rocky Mountains, and the greater Mississippi River drainage basin (Bettis et al., 2003). Loess bodies of the North American midcontinent appear to be continuous at small scale. At a larger scale, it is apparent that these loess bodies have very different thickness trends that are not part of a larger regional trend (Bettis et al., 2003; Muhs and Bettis, 2003). The variability of loess thickness over a landscape is, however, one of the most powerful tools in using this sediment for paleoclimatic reconstructions.

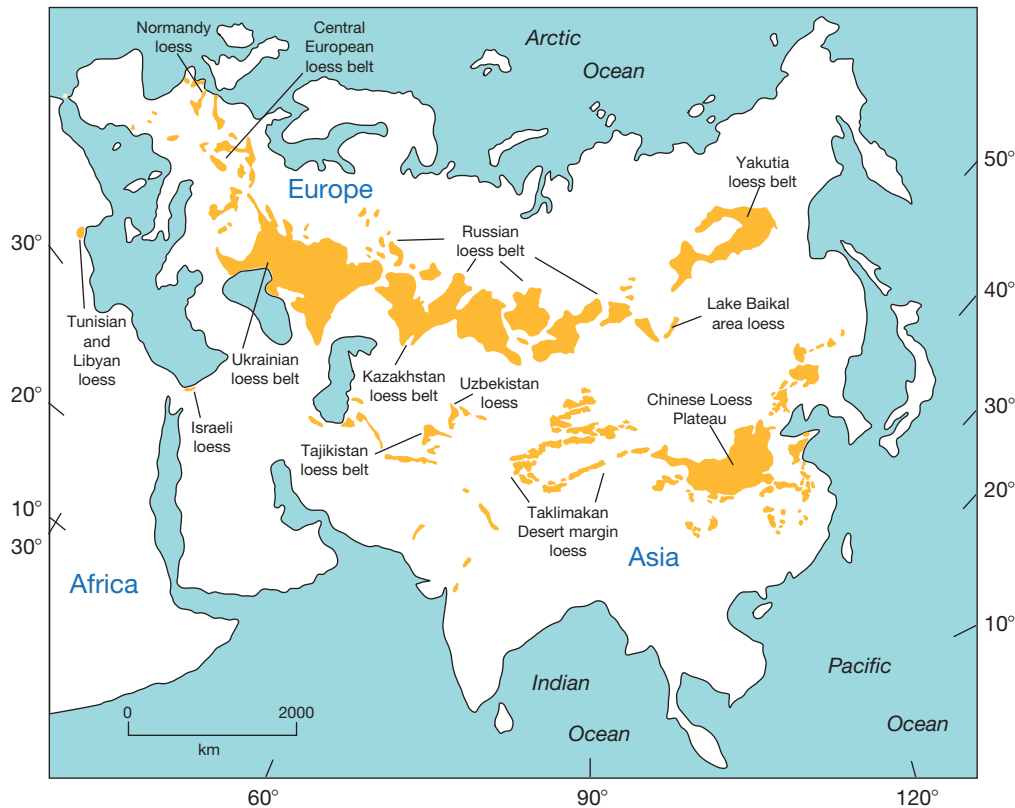


Figure 1 Map showing the distribution of loess in Eurasia and localities or regions referred to in the text. Compiled from Rozycki (1991) and Liu (1988), and redrawn from figure in Muhs and Bettis (2003).

Loess is not extensive over Africa, nor is it widespread in adjacent subtropical parts of the Middle East. There are, however, well-documented, but geographically limited areas of loess in Tunisia, Libya, Nigeria, Namibia, and Israel (Figure 1). Loess is also largely absent in Australia, although there are areas of fine-grained eolian mantles and considerable evidence of exotic quartz in soils (Hesse and McTainsh, 2003). However, loess is found over much of New Zealand, where its stratigraphy (e.g., Berryman, 1993) and distribution (Eden and Hammond, 2003) have been studied in considerable detail.

Mineralogy and Geochemistry of Loess

Most loess deposits have a mineralogy that includes quartz, plagioclase, K-feldspar, mica, calcite (and sometimes dolomite), and phyllosilicate clay minerals (smectite, chlorite, mica, and kaolinite). Heavy minerals are usually present, but in small amounts. Bulk geochemical studies show that the dominant constituent in loess is SiO_2 , which ranges from ~45% to 75%, but is typically 55–65%. The high SiO_2 contents of loess deposits reflect a dominance of quartz, but smaller amounts of feldspars and clay minerals also contribute to this value. Plots of SiO_2 versus Al_2O_3 show that most loess has a composition that falls between that of average shale and quartz-dominated sandstone (Muhs and Bettis, 2003). Loess

with high clay mineral content has more Al_2O_3 , Fe_2O_3 , and TiO_2 ; loess with higher carbonate (calcite and dolomite) content has more CaO and MgO .

Loess Origins: Processes of Silt Particle Formation

A traditional view of loess is that silt-sized particles are produced mostly by glacial grinding of crystalline rocks, deposited in till, reworked by fluvial processes as outwash, and finally entrained, transported, and deposited by wind (Figure 4). This classical model of loess formation has led to the view that loess deposits are primarily markers of continental-scale (global) glacial periods. The model is supported by observations of the geographic proximity of loess bodies to the southern limits of the Laurentide ice sheet in North America (Ruhe, 1969) and the Fennoscandian ice sheet in Europe (Frechen et al., 2003), as well as smaller glaciers in Asia (Dodonov, 1991) and South America (Zárate, 2003). In the 1950s and 1960s, widespread application of radiocarbon dating showed that the youngest loess deposits in North America coincided with the ages of the last major expansion of the Laurentide ice sheet (Ruhe, 1969). More recently, many luminescence ages show that the youngest loess dates to the last glacial period in Europe as well (Frechen et al., 2003).

Despite the long-term support for the classical 'glacial' concept of loess formation, there have been challenges to this

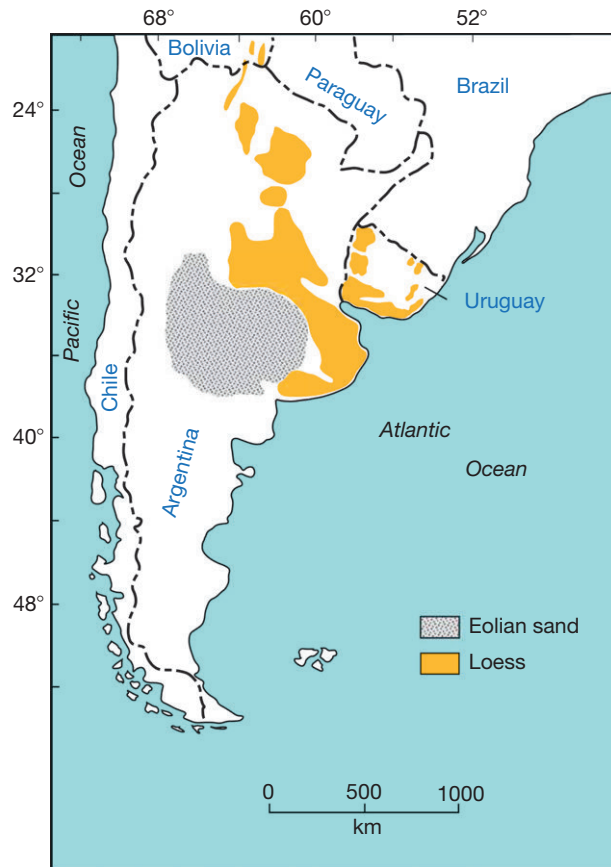


Figure 2 Map showing the distribution of loess and eolian sand in southern South America. Redrawn from Zárate (2003) and sources therein.

model going back at least 50 years. The debate has continued to this day and centers on the issue of 'glacial' loess versus 'desert' loess (Muhs and Bettis, 2003; Smalley, 1995; Wright, 2001). 'Desert' loess is a term used loosely to describe eolian silt generated in and derived from arid or semiarid regions that were not glaciated. The debate on desert loess versus glacial loess centers on whether silt-sized particles can be produced by mechanisms other than glacial grinding; specifically, whether or not they can be produced in deserts. A variety of mechanisms can, in principle, produce silt-sized particles in arid regions and these are summarized in a highly simplified model (Figure 5). These processes include frost shattering, comminution (particle size reduction by crushing or grinding) by fluvial and mass-movement transport, chemical weathering, salt weathering, eolian abrasion, and ballistic impacts.

China is the region most often cited as the best example of a long-term and spatially extensive nonglacial (or 'desert') loess record. The observations cited for a desert origin are loess thickness and particle size trends that show decreases downwind of desert basins. In addition, modern dust storms originate in the same desert basins. However, loess in China may have, as its ultimate source, glacially derived silt. The mountains surrounding the largest desert basins in China have glaciers at present and were more extensively glaciated in the past. It is possible that much primary silt production

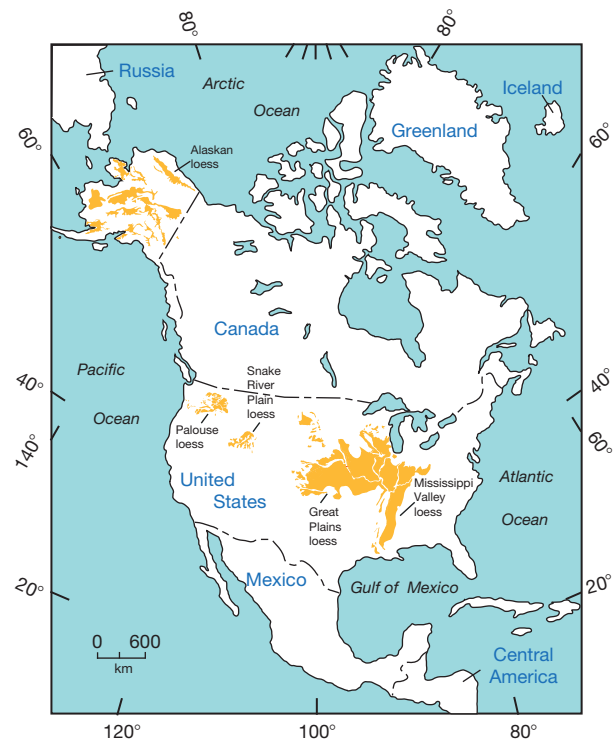


Figure 3 Map showing the distribution of loess in North America, from compilations in Péwé (1975), Bettis et al. (2003), and Busacca et al. (2004), and references therein.

takes place by glacial grinding (as well as other processes) in the mountains, and outwash may carry the silts into the basins (Smalley, 1995). The silts are then transported from the desert basins by wind. If this model is correct, then the arid basins are simply reservoirs for particle storage and have little to do with silt production itself, a concept supported by geochemical and isotopic data (Sun, 2002).

Despite continuing controversy over whether loess in China is of 'glacial' or 'desert' origin (or both), it does appear that the main periods of loess sedimentation in China correspond to glacial periods, whereas periods of soil formation (when loess deposition appears to cease) correspond to interglacials or interstadials (Kukla and An, 1989; Porter, 2001; Shen et al., 1992). Indeed, the loess-paleosol record of China can be reasonably well correlated with the deep-sea sediment oxygen-isotope record of glacials and interglacials (Figure 6). Nonglacial loess in the Great Plains of North America is also dated to the last glacial period (Aleinikoff et al., 1999; Maat and Johnson, 1996; Muhs et al., 1999; Roberts et al., 2003). These observations suggest that even in regions where glacial sediment may not have been abundant as a loess source, glacial periods may be favorable times for eolian silt availability, entrainment, transportation, and deposition. Glacial-period factors that may have favored loess entrainment and transportation include decreased vegetation cover, increased aridity, increased wind strength, and a decreased intensity of the hydrological cycle, such that silt can stay in suspension for longer distances (Kohfeld and Harrison, 2000; Mahowald et al., 1999).

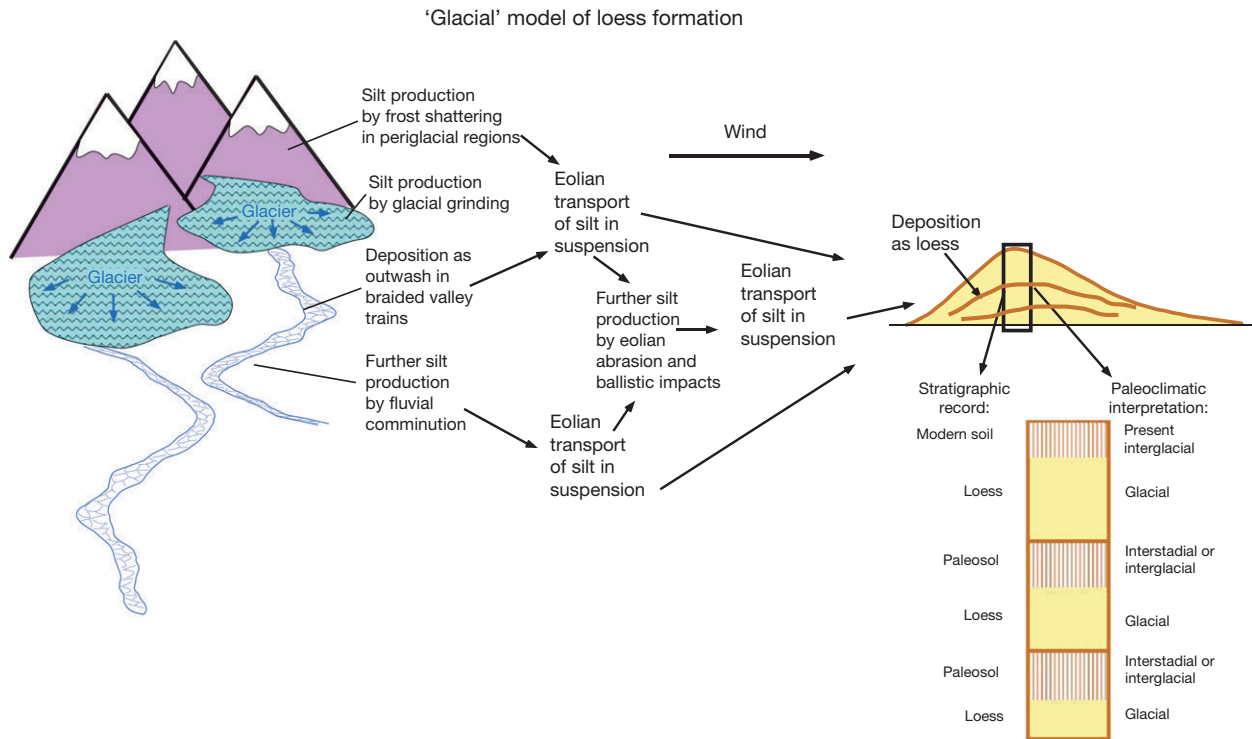


Figure 4 Classical model of ‘glacial’ loess formation wherein silt-sized particles are produced primarily by glacial grinding, delivered to outwash streams and finally entrained by wind. Redrawn from [Muhs DR and Bettis III EA \(2003\)](#) Quaternary loess-paleosol sequences as examples of climate-driven sedimentary extremes. *Geological Society of America Special Paper* 370: 53–74.

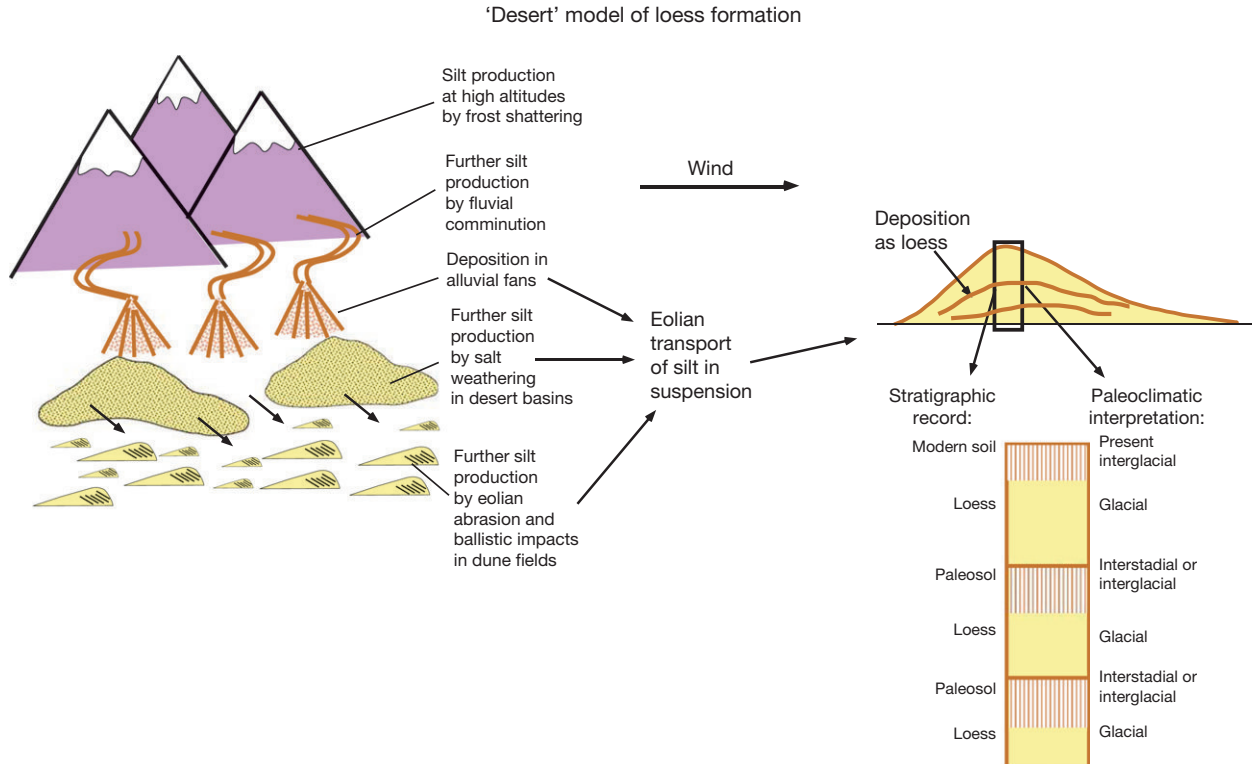


Figure 5 Model of ‘desert’ loess formation wherein silt-sized particles are produced by a variety of nonglaciogenic processes before eventual entrainment by wind. Redrawn from [Muhs DR and Bettis III EA \(2003\)](#) Quaternary loess-paleosol sequences as examples of climate-driven sedimentary extremes. *Geological Society of America Special Paper* 370: 53–74.

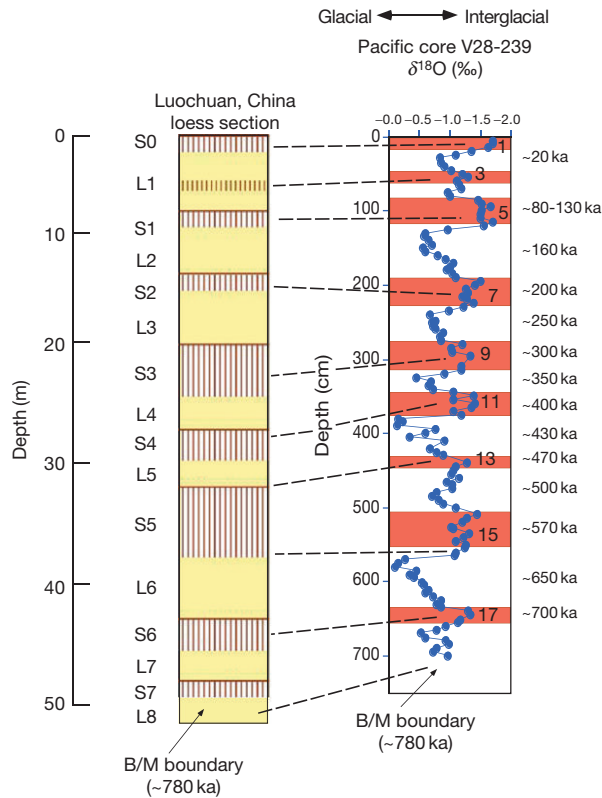


Figure 6 Loess stratigraphy at Luochuan, China (Kukla and An, 1989) and correlation with the deep-sea oxygen-isotope ($\delta^{18}\text{O}$) record of equatorial Pacific core V28-239 (Shackleton and Opdyke, 1976). Periods dominated by loess deposition ('L' units) occur during interglacials and periods dominated by soil formation ('S' units) occur during glacials. Correlations are based on ^{10}Be accumulation in loess and paleosols at Luochuan (redrawn from Shen C, Beer J, Liu T, et al. (1992) ^{10}Be in Chinese loess. *Earth and Planetary Science Letters* 109: 169–177) and identification of the Brunhes–Matuyama (B/M) boundary at 50.5 m at Luochuan and at 726 cm in V28-239. Approximate ages in V28-239 are based on the depth of the B/M boundary (~780 ka) and an assumed long-term average sedimentation rate of $\sim 0.93 \text{ cm ka}^{-1}$.

Loess Stratigraphy

Loess stratigraphy is rarely simple. Although loess is sometimes conceptualized as a thick accumulation of unaltered, massive silt, in reality it is usually sediment that has experienced varying degrees of weathering and pedogenesis (soil-forming processes). Some workers have pointed out that loess sedimentation and soil formation are essentially competing processes: when loess sedimentation rates are high, pedogenic processes cannot keep up and relatively unaltered sediment accumulates (Muhs et al., 2004; Verosub et al., 1993). When loess sedimentation rates are low, soil-forming processes extend deeper into previously deposited loess, but soils may also continue to accumulate small amounts of eolian sediment during pedogenesis. Thus, loess sequences should not be viewed in the same way as other Quaternary records, such as deep-sea or lacustrine sediments, where a better case can be made for more or less continuous sedimentation.

One distinct advantage of loess over many other Quaternary sediments is that it can be dated directly, using luminescence methods (Aitken, 1998). Because of the decay of naturally occurring radioelements in surrounding sediment, electrons are trapped in defect areas in minerals such as quartz and feldspar. These electron traps are effectively emptied, or 'zeroed' when sediment is entrained by wind and exposed to sunlight. In the case of optically stimulated luminescence, the traps are emptied and therefore the signal is zeroed usually in a matter of seconds to minutes. After deposition and sediment burial, radiation, largely from naturally occurring elements (potassium (K), rubidium (Rb), uranium (U), thorium (Th)) within the sediment itself, excites electrons and causes them to be trapped again within the crystal structures and defect areas of minerals. Subsequent stimulation in the laboratory by optical or thermal sources causes luminescence as the trapped electrons are released. The amount of luminescence is a function of the total radiation dose to which the mineral has been exposed. Because the natural annual environmental dose rate can be calculated from the concentrations of K, Rb, U, and Th within the sediment, it is therefore possible to determine the time the sediment was last exposed to sunlight. Luminescence dating is ideal for loess deposits because unlike other dating methods, no fossils or organic materials are required; the minerals of choice (quartz and feldspar) are common in loess; being eolian sediment, it is effectively zeroed; and dating is possible back to 60–100 thousand years ago (ka). Errors are typically of the order of 5–10% (e.g., Roberts et al., 2003), but tend to increase as the limit of the technique is approached.

Loess, as a representation of the glacial record, has been studied extensively in Europe and North America. Decades of study and dating have established a clear link between the loess record and glacial–interglacial cycles. The modern soils of much of Europe have been developing since the end ($\sim 13 \text{ ka}$) of the last ('Weichshelien' or 'Würm') glacial period. Hundreds of luminescence ages confirm that the uppermost loess in Europe was deposited during the Weichshelien–Würm glacial period, from ~ 28 to $\sim 13 \text{ ka}$ (Frechen et al., 2003). Loess from this period is well displayed as relatively unaltered sediment in many exposures, such as the classical section near Kesselt, Belgium (Figure 7), described in detail by Vandenbergh et al. (1998) and Van den Haute et al. (1998). This last glacial loess has an equivalent in North America that is called Peoria Loess (Figures 8 and 9). Weichshelien loess is, in turn, underlain by a minimally developed paleosol (LH, or 'Limon Humifère') that may have its equivalent in North America as the Farmdale 'Geosol' (=paleosol) and Gilman Canyon Formation (the upper part of which contains a paleosol). In Europe, a well-developed soil, the Rocourt paleosol, is found below the LH paleosol and formed during the last interglacial period. This soil has its equivalent in North America as the Sangamon Geosol. Both the Rocourt paleosol and Sangamon Geosol probably developed over tens of thousands of years during a long interglacial period with little loess sedimentation. The Rocourt paleosol, in turn, is developed in loess of the penultimate glacial period ('Saalien'), which has its equivalent in North America as the Loveland Loess. The age of penultimate glacial loess on both continents may be on the order of around 160–140 ka (Forman and Pierson, 2002; Maat and Johnson, 1996).

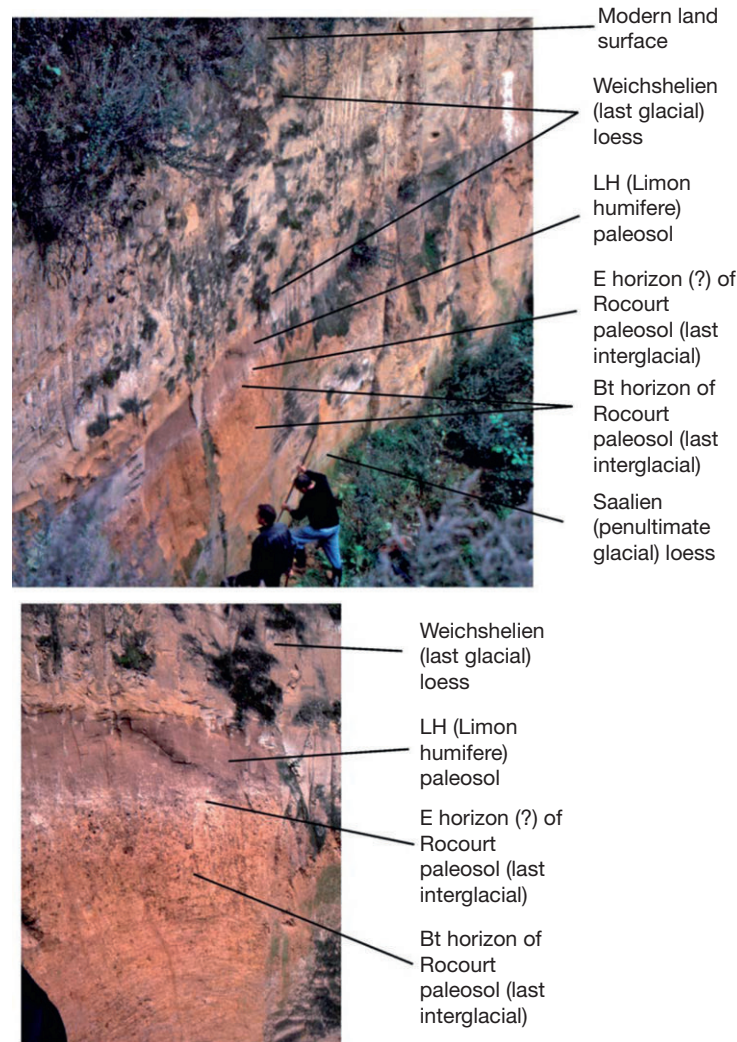


Figure 7 Photograph of loess and paleosols representing the last interglacial–glacial cycle exposed in brickyard near Kesselt, Belgium (see [Van den Haute et al., 1998](#) and [Vandenberghe et al., 1998](#)). Photograph by D.R. Muhs.

In nonglacial environments, loess stratigraphy is generally similar to that of glacial regions. Thus, even where glacial silt is not the primary sediment source, loess deposition occurs primarily during glacial periods and soil formation occurs primarily during interglacial periods. In Central Asia ([Dodonov, 1991](#)) and in the Great Plains region of North America ([Aleinikoff et al., 1999](#)), loess is derived from both nonglaciogenic and glaciogenic sources. Nevertheless, many of the same stratigraphic units have been identified and correlated with loess sequences near continental ice. In Central Asia, for example, the main episodes of loess deposition were during the last two glacial periods ([Frechen and Dodonov, 1998](#)). In the Great Plains of North America, Peoria Loess, a Farmdale-equivalent soil (a paleosol in the Gilman Canyon Formation), the Sangamon Geosol, and Loveland Loess have all been identified ([Figure 9](#)). Last glacial (Peoria) loess in North America also has its equivalent as the ‘Malan’ or ‘L1’ loess in China, although Malan Loess spans a longer period of time and includes what in North America would be called the Roxana Silt or the Gilman Canyon Formation. The last

interglacial soil complex (the Sangamon Geosol of North America) is called the ‘S1’ paleosol in China; it is developed in the ‘L2’ loess, which is roughly equivalent to Loveland Loess in North America.

There are, however, exceptions to the generalization of loess deposition occurring only in glacial periods. For example, there is well-documented evidence of Holocene loess in China (see [Roberts et al., 2001](#), for a recent example) and in North America, both in the Great Plains (e.g., [Miao et al., 2005](#)) and Alaska (e.g., [Muhs et al., 2004](#)). In other regions, it is difficult to correlate loess deposition and soil formation with specific glacial and interglacial periods. For example, in Israel, desert loess and fine-grained dust are thought to have been deposited over much of the Quaternary ([Dan and Yaalon, 1971](#)), probably during both glacials and interglacials. Soil formation seems to take place syndepositionally, that is, concurrently with loess deposition. In this area, relatively unaltered loess is, therefore, not apparent in most stratigraphic sections; clay-rich (Bt) horizons form in the loess and calcic (Bk) horizons form below the Bt horizons. Nevertheless, stratification is



Figure 8 Photograph of loess and paleosols representing the last interglacial–glacial cycle exposed on the north end of Crowley's Ridge, Arkansas, USA (see Markewich et al., 1998). Photograph by D.R. Muhs.

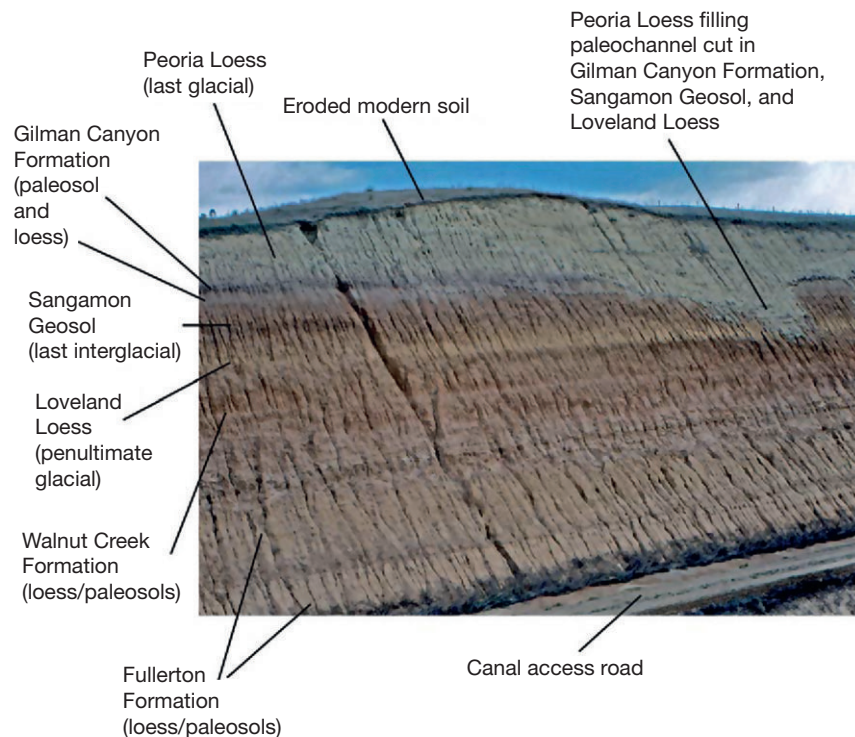


Figure 9 Photograph of loess and paleosols exposed in the Elba canal cut, eastern Nebraska, USA (see May et al., 1995). Note paleochannel cut into the Gilman Canyon Formation, Sangamon Geosol, and Loveland Loess, filled by Peoria Loess. Photograph by D.R. Muhs.

apparent in these sections, and several cycles of loess deposition/soil formation can be seen in outcrops (Figure 10).

The stratigraphy of loess–paleosol sequences can sometimes be used as a relative dating tool. In China, the sequence of loesses (S0, L1, S1, L2, etc.) is so readily identifiable, it can be used as a relative dating or correlation tool for stream terraces when these landforms are mantled with loess (Porter

et al., 1992). In New Zealand, loess deposits, tephras, and paleosols have an increasingly complex record on successively older marine terraces (Berryman, 1993). The oldest marine terraces have the most complete succession of loess deposits, tephras, and paleosols (Figure 11). Younger marine terraces have only the upper part of the record, and the youngest marine terrace is not overlain by loess at all.

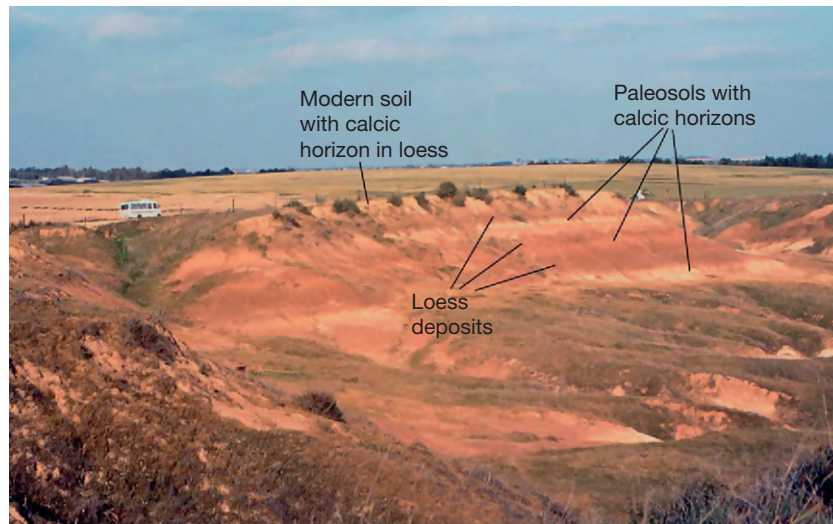


Figure 10 Loess deposits with calcic-horizon-bearing paleosols, northern Negev Desert, Israel (see [Dan and Yaalon, 1971](#)). Photograph by D.R. Muhs.

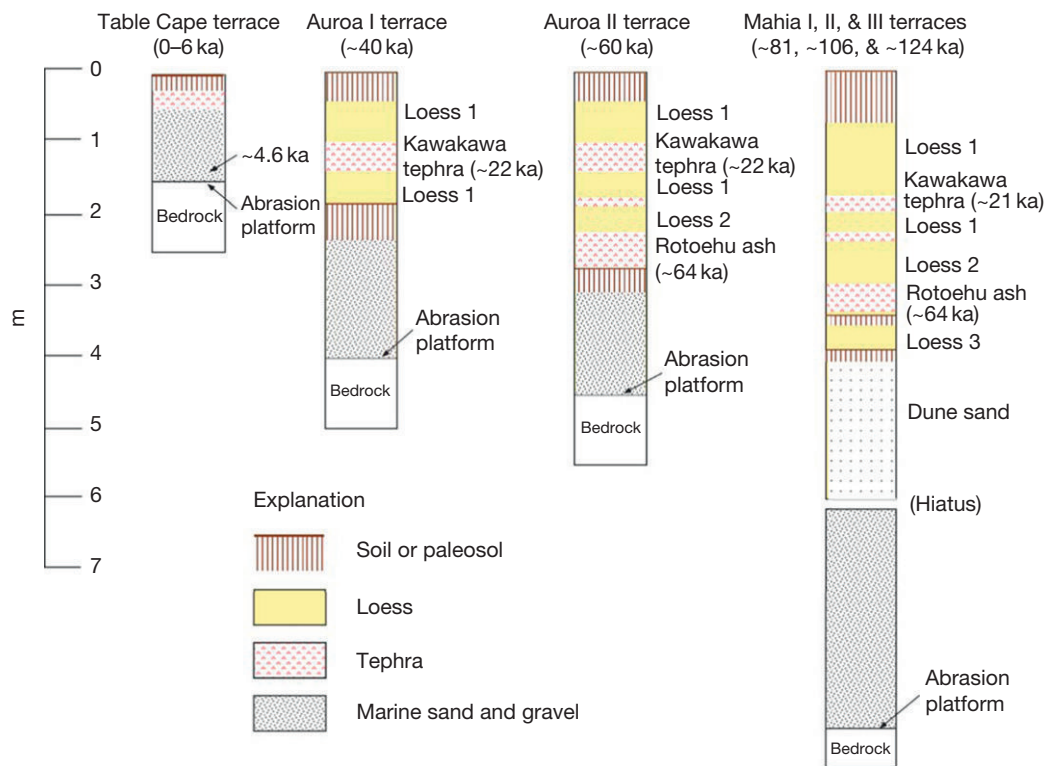


Figure 11 Loess, paleosol, and tephra stratigraphy on marine terraces on the Mahia Peninsula, New Zealand. Modified from Berrryman (1993).

Quaternary Paleoenvironmental Information from Loess Sequences

There is a wide variety of paleoenvironmental information that can be obtained from loess–paleosol sequences. Loess itself, as its properties change over a landscape, can yield important

clues about the paleowind that deposited it. Loess thickness, particle size, and carbonate content generally decrease away from the source ([Bettis et al., 2003](#); [Liu, 1988](#); [Mason, 2001](#); [Muhs and Bettis, 2003](#); [Muhs et al., 2004](#); [Porter, 2001](#); [Ruhe, 1969](#); [Smith, 1942](#)). The decrease in loess thickness reflects a reduction of sediment load downwind from the source, the

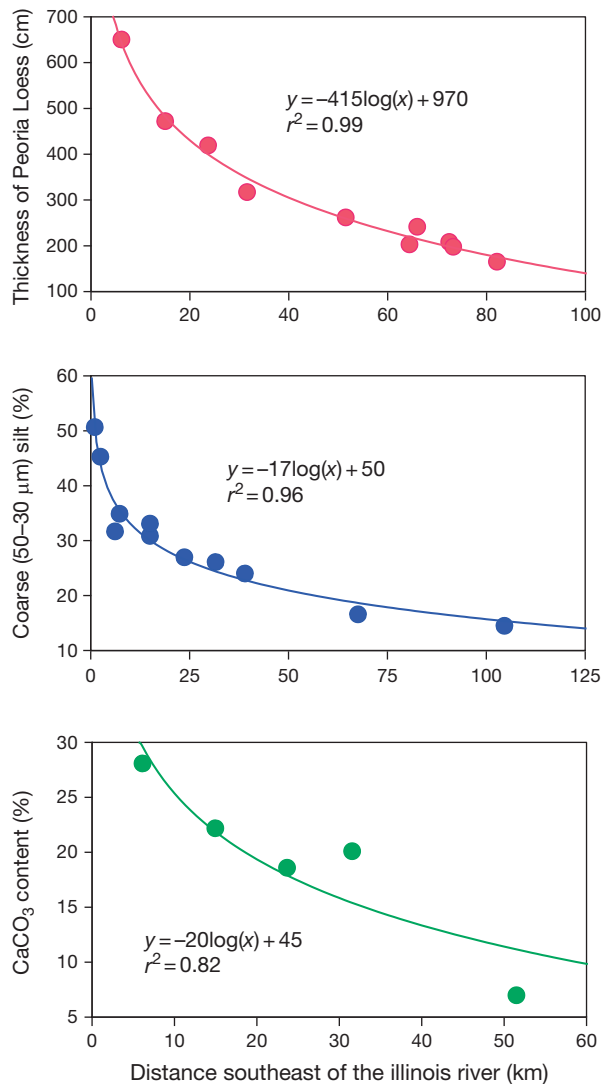


Figure 12 Thickness, coarse silt content, and carbonate content of Peoria (last glacial) Loess as a function of distance southeast of the Illinois River, IL, USA. Data from Smith (1942); regression equations computed by D.R. Muhs.

decrease in mean particle size reflects winnowing of the coarse load, and the decrease in carbonate content reflects syndepositional leaching downwind, where deposition rates are lower. If a loess source were a north-to-south-trending river valley, a decrease in loess thickness and mean particle size to the east of the river would imply northwesterly, westerly, or southwesterly paleowinds (Figure 12), at least during those periods when winds were strong enough to transport silt-sized particles.

Loess lacks many of the Quaternary paleoecological indicators that are commonly used in lacustrine or marine sediments, such as pollen, diatoms, Ostracoda, Radiolaria, or Foraminifera. Furthermore, it is rare for mammal fossils to be preserved in loess. However, it is common for the shells of land snails to be preserved in loess, and they are abundant in China, Europe, and North America. Most or all of these snails are extant species, and their modern zoogeography is reasonably well established.

Thus, it is possible to infer past climates during the times of loess deposition by identification of extralimital taxa, that is, those species that do not presently live at a locality where they are found as fossils. In North America, the upper part of last glacial loess of the central Great Plains contains several extralimital boreal or Cordilleran species of snails (Figure 13). The presence of these northern-forest and mountain-forest species implies a much cooler-than-present last glacial climate with forest vegetation, as opposed to the present, temperate grassland of the region (Rousseau and Kukla, 1994).

Whereas sedimentologic and paleontologic data in loess give information about glacial periods, paleosols within loess deposits yield information about interglacial or interstadial periods. Without question, the most common method applied to loess-derived paleosols since the 1980s has been the measurement of magnetic susceptibility and other mineral magnetic properties (Kukla and An, 1989; Maher et al., 1994; Porter, 2001; Porter et al., 2001; Verosub et al., 1993). Magnetic susceptibility is essentially a measurement of the abundance of magnetic minerals such as magnetite, a primary, rock-forming mineral, and maghemite, a secondary, pedogenic mineral. Other minerals, such as hematite, have very low but measurable magnetic susceptibility that can be differentiated from magnetite and maghemite. Measurement of these properties is rapid, inexpensive, and highly reproducible. Numerous studies have shown that Chinese loess has relatively low magnetic susceptibility and intercalated paleosols have high susceptibility. This makes the technique highly valued as a section-to-section correlation tool (Kukla and An, 1989) and many researchers have extended the method to correlation with deep-sea oxygen-isotope records. Other researchers have attempted to develop transfer functions, correlating magnetic susceptibility in modern soils with climate parameters, such as mean annual precipitation (Maher et al., 1994). If the assumptions are valid in this approach, then it would, in principle, be possible to estimate past climate from magnetic susceptibility of paleosols in loess. However, several problems arise with this approach. One is that magnetic susceptibility in modern, loess-derived soils in China is partly a function of particle size and sediment accumulation rate (as a dilution effect), as well as climate. Both these factors are spatially variable but highly correlated with one another (and climate) across the Chinese Loess Plateau (Porter et al., 2001). Another problem is that transfer functions of magnetic susceptibility and climate assume that soils quickly reach a steady state with regard to pedogenic magnetic mineral production. Soil chronosequence studies, however, show that magnetic susceptibility in soils continues to increase over time (Singer et al., 1992). Finally, in some regions, magnetic susceptibility trends are the reverse of what they are in China. In Siberia and Alaska, for example, susceptibility is highest in loess and lowest in paleosols, due to high-magnetite loess sources and little or no production of secondary maghemite in soils (Begét et al., 1990; Chlachula, 2003). Undoubtedly, part of the attraction of using magnetic susceptibility in studying loess and paleosols is the relative ease, rapidity, and low cost of analysis. Nevertheless, it is important to remember that in addition to the problems just summarized, magnetic minerals constitute a very small portion of the mineral suites of loess and its paleosols.

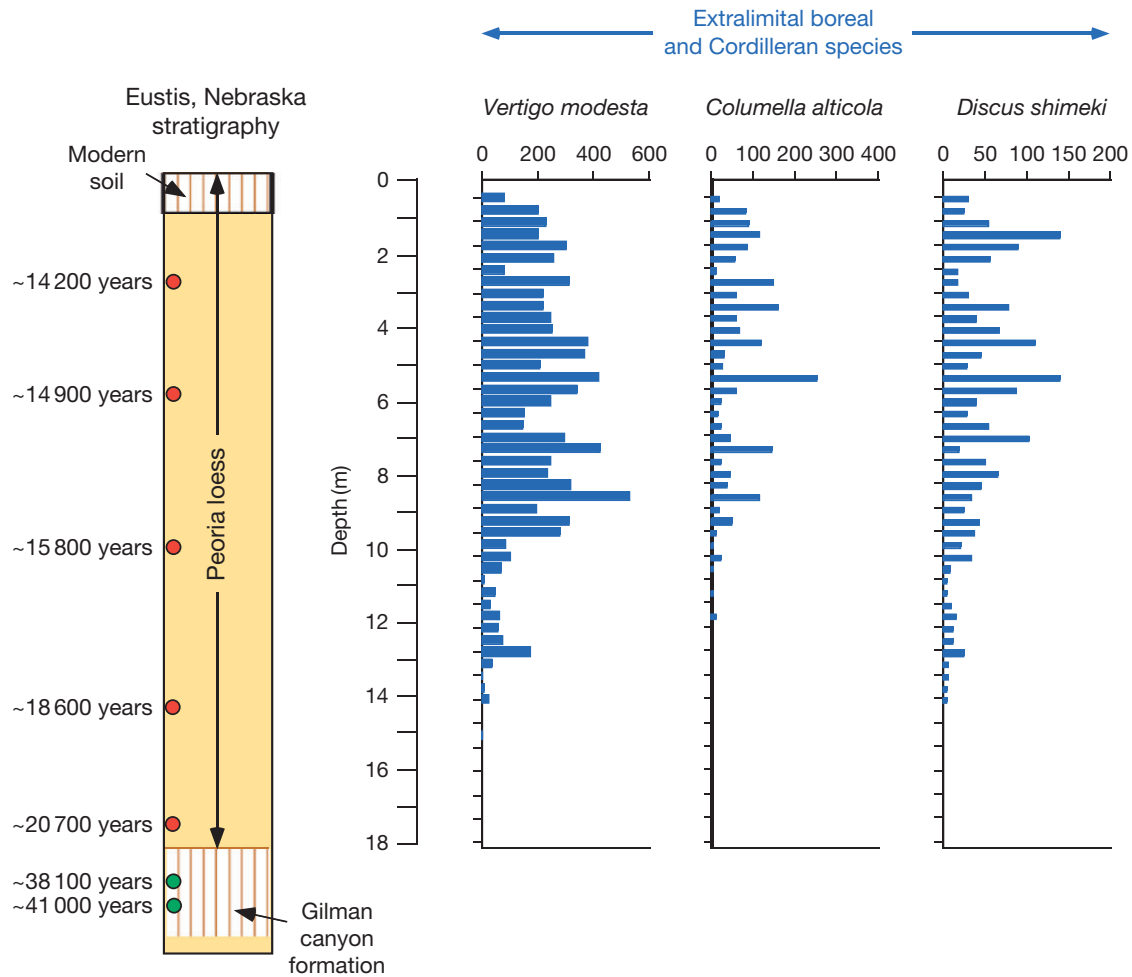


Figure 13 Stratigraphy of last glacial (Peoria) loess at Eustis, Nebraska, USA, with optically stimulated luminescence (OSL) ages and calibrated radiocarbon ages (data from Roberts et al., 2003 and Maat and Johnson, 1996). Also shown are absolute abundances of extralimital boreal and Cordilleran species of land snails (data from Rousseau and Kukla, 1994). Note that the section thickness as given by the latter authors differs slightly from that in Roberts et al. (2003); thus, depths of OSL ages in Peoria Loess are approximate.

Less-controversial methods have also been utilized in studying loess-derived paleosols, although not as commonly as magnetic susceptibility. In the Mississippi River valley of Illinois, soil morphology, particle size, and mineralogical and geochemical data show that paleosols differ in the amount of development and chemical weathering they have experienced (Grimley et al., 2003). For example, the Sangamon Geosol and an older paleosol, called the Yarmouth paleosol, are redder and more clay-rich than the Farmdale Geosol or the modern soil. Proxies for plagioclase depletion ($\text{Na}_2\text{O}/\text{TiO}_2$), apatite depletion ($\text{P}_2\text{O}_5/\text{TiO}_2$), and silicate mineral depletion in general ($\text{SiO}_2/\text{Al}_2\text{O}_3$) show that the Yarmouth paleosol has experienced more weathering than the Sangamon Geosol, which in turn has experienced more weathering than the Farmdale Geosol or modern soil (Figure 14). These data can be interpreted to mean that past interglacials were warmer and more humid than the present one, with enhanced chemical weathering in soils. An alternative interpretation is that past interglacials had a longer duration

than what the present one has experienced, or that past interglacials were warmer, more humid, and longer than the present.

Conclusion

Loess–paleosol sequences are one of the most important terrestrial records of Quaternary climate change. Loess is dominantly silt sized, windblown sediment, typically composed of quartz, feldspars, micas, carbonates, and clay minerals. These deposits usually blanket preexisting landscapes and can be centimeters to hundreds of meters thick. Loess is found over a large portion of the Earth's surface, in the central and northwestern United States, Alaska, Argentina, Europe, Russia, Central Asia, China, and New Zealand. It has distinct advantages over other Quaternary sediments in that it is a direct record of atmospheric circulation and can be dated directly, using luminescence methods. Many loess records

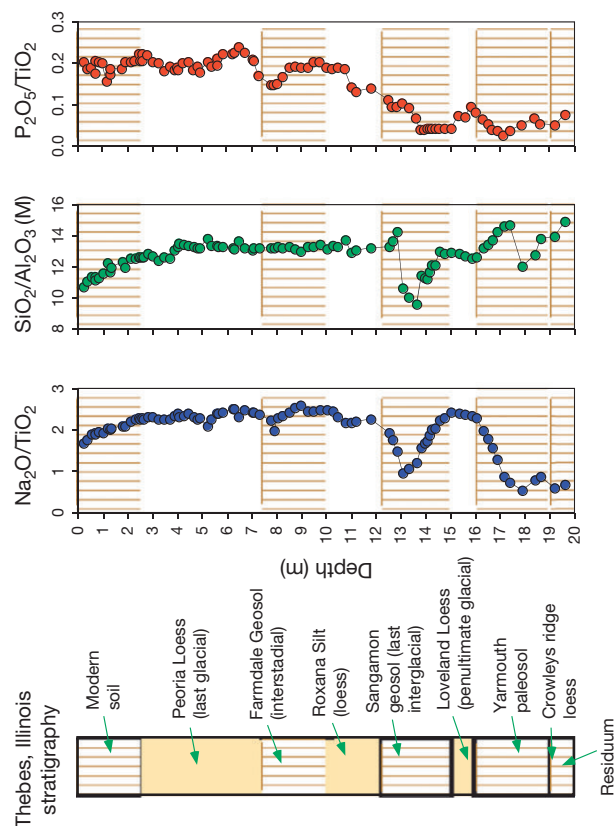


Figure 14 Stratigraphy and geochemistry of loess near Thebes, IL, Mississippi River valley, USA. Stratigraphy and $\text{Na}_2\text{O}/\text{TiO}_2$ data are from Grimley et al. (2003); other data, previously unpublished, are courtesy of D.A. Grimley, Illinois State Geological Survey.

span much or all of the Quaternary and thus represent a terrestrial analog to the deep-sea sediment record. In most regions, loess was deposited during glacial periods, and soils formed during interglacial periods. There are exceptions to this, however, and in some areas, such as China, neither loess deposition nor soil formation ever cease completely. Loess can be used to reconstruct paleowinds using spatial trends of thickness, particle size, and carbonate content. Although many paleoecological indicators are rare or absent in loess, shells of land snails are common and can be a powerful tool in reconstructing paleoclimate. Finally, paleosols in loess sequences have many characteristics that can be used to estimate past climates.

Acknowledgments

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See also: **Loess Records:** Central Asia; China; Europe; North America; South America. **Luminescence Dating:** Thermoluminescence. **Paleoceanography, Physical and Chemical Proxies:** Oxygen-Isotope Stratigraphy of the Oceans. **Paleosols and Wind-Blown Sediments:** Mineral Magnetic Analysis; Nature of Paleosols; Soil Morphology in Quaternary Studies; Weathering Profiles.

References

- Aitken MJ (1998) *An Introduction to Optical Dating*. Oxford: Oxford University Press.
- Aleynikov JN, Muhs DR, Sauer R, and Fanning CM (1999) Late Quaternary loess in northeastern Colorado, Part II – Pb isotopic evidence for the variability of loess sources. *Geological Society of America Bulletin* 111: 1876–1883.
- Begét JE, Stone DB, and Hawkins DB (1990) Paleoclimatic forcing of magnetic susceptibility variations in Alaskan loess during the Quaternary. *Geology* 18: 40–43.
- Berryman KR (1993) Distribution, age, and deformation of Late Pleistocene marine terraces at Mahia Peninsula, Hikurangi subduction margin, New Zealand. *Tectonics* 12: 1365–1379.
- Bettis EA III, Muhs DR, Roberts HM, and Wintle AG (2003) Last glacial loess in the conterminous USA. *Quaternary Science Reviews* 22: 1907–1946.
- Busacca AJ, Begét JE, Markewich HW, Muhs DR, Lancaster N, and Sweeney MR (2004) Eolian sediments. In: Gillespie AR, Porter SC, and Atwater BF (eds.) *The Quaternary period in the United States*, pp. 275–309. Amsterdam: Elsevier.
- Chlachula J (2003) The Siberian loess record and its significance for reconstruction of Pleistocene climate change in north-central Asia. *Quaternary Science Reviews* 22: 1879–1906.
- Dan J and Yaalon DH (1971) On the origin and nature of the paleopedological formations in the coastal desert fringe areas of Israel. In: Yaalon DH (ed.) *Paleopedology: Origin, Nature and Dating of Paleosols*, pp. 245–260. Jerusalem: Israel Universities Press.
- Dodonov AE (1991) Loess of Central Asia. *GeoJournal* 24(2): 185–194.
- Eden DN and Hammond AP (2003) Dust accumulation in the New Zealand region since the last glacial maximum. *Quaternary Science Reviews* 22: 2037–2052.
- Forman SL and Pierson J (2002) Late Pleistocene luminescence chronology of loess deposition in the Missouri and Mississippi River valleys, United States. *Palaeogeography, Palaeoclimatology, Palaeoecology* 186: 25–46.
- Frechen M and Dodonov AE (1998) Loess chronology of the Middle and Upper Pleistocene in Tadzhikistan. *Geologische Rundschau* 87: 2–20.
- Frechen M, Oches EA, and Kohfeld KE (2003) Loess in Europe – Mass accumulation rates during the Last Glacial Period. *Quaternary Science Reviews* 22: 1835–1857.
- Grimley DA, Follmer LR, Hughes RE, and Solheid PA (2003) Modern, Sangamon and Yarmouth soil development in loess of unglaciated southwestern Illinois. *Quaternary Science Reviews* 22: 225–244.
- Hesse PP and McTainsh GH (2003) Australian dust deposits: Modern processes and the Quaternary record. *Quaternary Science Reviews* 22: 2007–2035.
- Kohfeld KE and Harrison SP (2000) How well can we simulate past climates? Evaluating the models using global palaeoenvironmental datasets. *Quaternary Science Reviews* 19: 321–347.
- Kukla G and An Z (1989) Loess stratigraphy in central China. *Palaeogeography, Palaeoclimatology, Palaeoecology* 72: 203–225.
- Liu T (1988) *Loess in China*, 2nd edn. Beijing: China Ocean Press. Berlin: Springer-Verlag.
- Maat PB and Johnson WC (1996) Thermoluminescence and new ^{14}C age estimates for late Quaternary loesses in southwestern Nebraska. *Geomorphology* 17: 115–128.
- Maier BA, Thompson R, and Zhou LP (1994) Spatial and temporal reconstructions of changes in the Asian palaeomonsoon: A new mineral magnetic approach. *Earth and Planetary Science Letters* 125: 461–471.
- Mahowald N, Kohfeld K, Hansson M, et al. (1999) Dust sources and deposition during the last glacial maximum and current climate: A comparison of model results with paleodata from ice cores and marine sediments. *Journal of Geophysical Research* 104: 15895–15916.
- Markewich HW, Wysocki DA, Pavich MJ, et al. (1998) Paleopedology plus TL, ^{10}Be , and ^{14}C dating as tools in stratigraphic and paleoclimatic investigations, Mississippi River Valley, USA. *Quaternary International* 51/52: 143–167.
- Mason JA (2001) Transport direction of Peoria Loess in Nebraska and implications for loess sources on the central Great Plains. *Quaternary Research* 56: 79–86.

- May DW, Swinehart JB, Loope D, and Souders V (1995) Late Quaternary fluvial and eolian sediments: Loup River Basin and the Nebraska Sand Hills. In: Diffendal RF Jr. and Flowerday CA (eds.) *Geologic field trips in Nebraska and adjacent parts of Kansas and South Dakota, Guidebook No. 10*, pp. 13–31. Lincoln, NE: Conservation and Survey Division, Institute of Agriculture and Natural Resources, University of Nebraska-Lincoln.
- Miao X, Mason JA, Goble RJ, and Hanson PR (2005) Loess record of dry climate and aeolian activity in the early- to mid-Holocene, central Great Plains, North America. *The Holocene* 15: 339–346.
- Muhs DR, Aleinikoff JN, Stafford TW Jr., et al. (1999) Late Quaternary loess in northeastern Colorado: Part I – Age and paleoclimatic significance. *Geological Society of America Bulletin* 111: 1861–1875.
- Muhs DR and Bettis EA III (2003) Quaternary loess-paleosol sequences as examples of climate-driven sedimentary extremes. *Geological Society of America Special Paper* 370: 53–74.
- Muhs DR, McGeehin JP, Beann J, and Fisher E (2004) Holocene loess deposition and soil formation as competing processes, Matanuska Valley, southern Alaska. *Quaternary Research* 61: 265–276.
- Péwé TL (1975) Quaternary geology of Alaska. *US Geological Survey Professional Paper* 835: 1–145.
- Porter SC (2001) Chinese loess record of monsoon climate during the last glacial–interglacial cycle. *Earth-Science Reviews* 54: 115–128.
- Porter SC, An Z, and Zheng H (1992) Cyclic Quaternary alluviation and terracing in a nonglaciated drainage basin on the north flank of the Qinling Shan, central China. *Quaternary Research* 38: 157–169.
- Porter SC, Hallet B, Wu X, and An Z (2001) Dependence of near-surface magnetic susceptibility on dust accumulation rate and precipitation on the Chinese Loess Plateau. *Quaternary Research* 55: 271–283.
- Pye K (1987) *Aeolian Dust and Dust Deposits*. San Diego, CA: Academic Press.
- Pye K (1995) The nature, origin and accumulation of loess. *Quaternary Science Reviews* 14: 653–657.
- Roberts HM, Muhs DR, Wintle AG, Duller GAT, and Bettis EA III (2003) Unprecedented last glacial mass accumulation rates determined by luminescence dating of loess from western Nebraska. *Quaternary Research* 59: 411–419.
- Roberts HM, Wintle AG, Maher BA, and Hu M (2001) Holocene sediment-accumulation rates in the western Loess Plateau, China, and a 2500-year record of agricultural activity, revealed by OSL dating. *The Holocene* 11: 477–483.
- Rousseau DD and Kukla G (1994) Late Pleistocene climate record in the Eustis loess section, Nebraska, based on land snail assemblages and magnetic susceptibility. *Quaternary Research* 42: 176–187.
- Rozycki SZ (1991) *Loess and Loess-Like Deposits*. Warsaw: Ossolineum Press, Polish Academy of Sciences.
- Ruhe RV (1969) *Quaternary Landscapes in Iowa*. Ames, IA: Iowa State University Press.
- Shackleton NJ and Opdyke ND (1976) Oxygen-isotope and paleomagnetic stratigraphy of Pacific core V28-239 late Pliocene to latest Pleistocene. *Geological Society of America Memoir* 145: 449–464.
- Shen C, Beer J, Liu T, et al. (1992) ^{10}Be in Chinese loess. *Earth and Planetary Science Letters* 109: 169–177.
- Singer MJ, Fine P, Verosub KL, and Chadwick OA (1992) Time dependence of magnetic susceptibility of soil chronosequences on the California coast. *Quaternary Research* 37: 323–332.
- Smalley IJ (1995) Making the material: The formation of silt-sized primary mineral particles for loess deposits. *Quaternary Science Reviews* 14: 645–651.
- Smith GD (1942) Illinois loess: Variations in its properties and distribution, a pedologic interpretation. *University of Illinois Agricultural Experiment Station Bulletin* 490: 139–184.
- Sun J (2002) Provenance of loess material and formation of loess deposits on the Chinese Loess Plateau. *Earth and Planetary Science Letters* 203: 845–859.
- Van den Haute P, Vancraeynest L, and de Corte F (1998) The Late Pleistocene loess deposits and palaeosols of eastern Belgium: New TL age determinations. *Journal of Quaternary Science* 13: 487–497.
- Vandenberghe J, Huijzer BS, Múcher H, and Laan W (1998) Short climatic oscillations in a western European loess sequence (Kesselt, Belgium). *Journal of Quaternary Science* 13: 471–485.
- Verosub KL, Fine P, Singer MJ, and TenPas J (1993) Pedogenesis and paleoclimate: Interpretation of the magnetic susceptibility record of Chinese loess-paleosol sequences. *Geology* 21: 1011–1014.
- Wright JS (2001) 'Desert' loess versus 'glacial' loess: Quartz silt formation, source areas and sediment pathways in the formation of loess deposits. *Geomorphology* 36: 231–256.
- Zárate MA (2003) Loess of southern South America. *Quaternary Science Reviews* 22: 1987–2006.

LOESS RECORDS

Contents

Central Asia

China

Europe

North America

South America

Central Asia

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Central Asia can be regarded as one of the classical regions of loess distribution. During the Quaternary period, accumulation of wind-blown dust formed a widespread and thick loess cover at the piedmonts of the mountains in this region. Any traveler crossing Central Asia remembers the dust blown along desert and semidesert areas. The dust consists of clayey, silty, and fine sandy material derived from sandy deserts, from glacialfluvio/fluviol deposits of the river floodplains and from weathered and frost-shattered rock surfaces in high mountains where glaciers exist (Figures 1 and 2). The desert basins and the piedmonts of Central Asian mountains are arid to semiarid with the average annual air temperature of 10 °C or higher, and average annual precipitation between 200 and 600–800 mm, depending on the local climatic conditions. The plain regions, intramontane basins, and mountains to the north of the Hindu Kush are characterized by a dry, continental climate with moist, mild winters in the southern part and colder frosty winters in the north. Westerlies to the north of Hindu Kush provide a moist air-mass transfer during the winter–spring seasons, whereas a hot season occurs in the summer and early autumn period complemented by active eolian processes. Areas to the south of Hindu Kush and Himalayas belong to a warmer climatic belt, where Indian monsoon, carrying the moist air mass, occurs during the summer, while the winter and spring are a period of aridification when dust deflation is intensified.

Loess forms thick mantles which follow the shape of the present-day relief, smoothing it out or creating steep scarps and loess ‘cirques’ (Figures 3–5). Due to high porosity and permeability, water infiltration is pervasive in loess deposits, leading to widespread fluvial erosion, landsliding, collapsing, and redeposition (Figures 6 and 7). These processes create exposures for

studies of loess–paleosol strata on vertical loess walls, although trenches are usually prepared as well (Figure 8(a)–8(c)).

There are many hypotheses about the origin of loess. Loess experts such as Kriger (1965), Pécsi (1993), and Smalley et al. (2001) have systematically reviewed these hypotheses. One of the dominant models, the eolian theory of loess deposition, was proposed by the prominent nineteenth-century German scientist von Richthofen (1882), who had studied Chinese loess. A fundamental contribution to the eolian theory was made by Obruchev (1933) through his famous work on Central Asian loess from the end of the nineteenth to mid-twentieth century. A ‘proluvial’ hypothesis (deposition in the temporary water streams) of loess origin was proposed by Pavlov (1903), who studied loess and loess-like deposits in the Tashkent region. One of the independent hypotheses of loess formation was developed by Berg (1947), suggesting that loess results from *in situ* weathering and soil formation, called ‘soil-eluvial theory.’ This hypothesis was supported by some pedologists, such as Gerasimov (1976), who drew significant attention to pedogenic processes in loess formation; he considered that pedogenic hypothesis should not be in contradiction with the eolian idea of loess deposition. The idea of pedogenic formation and weathering was modified as a process called ‘loessification.’ A similar approach was elaborated by Minervin (1984), who used an *in situ* weathering model to explain the origin of collapsibility in loess strata by cryogenic processes. However, it should be emphasized that lack of cryogenic textures in Central Asian loesses does not allow acceptance of this model. The role of eolian sedimentation, explaining the uncompacted nature of loess and its formation in arid climatic zones such as Central Asia, was redeveloped by Kriger (1986). In addition, however, a polygenetic model of loess/loess-like formation should be mentioned, which results from a combination of different processes during deposition of loess and loess-like material (Mavlyanov, 1958). At present, the majority

[†]Deceased.



Figure 1 High mountain periglacial zone in the Eastern Pamirs where abundant moraine and glaciofluvial material serve as a source of clayey and silty sediments, which become a part of the system of fluvial runoff in mountain valleys.



Figure 4 Loess 'cirque' (bowl-shaped landsliding cavity) more than 100 m height, Chashmanigar section in southern Tajikistan.



Figure 2 On the high mountain plateau of the Eastern Pamirs, physical weathering processes occur as a function of temperature fluctuations and eolian abrasion. These processes disintegrate large boulders on the moraine surface, producing a large amount of fine-grained material.



Figure 5 Loess scarp in the Khonako reference section, 140 m in height and more than 2 km in length. Reproduced from Dodonov AE (2002) *Quaternary of Middle Asia: Stratigraphy, Correlation Paleogeography*, 250 p. Moscow: GEOS (in Russian); Figure 4, with permission from GEOS.

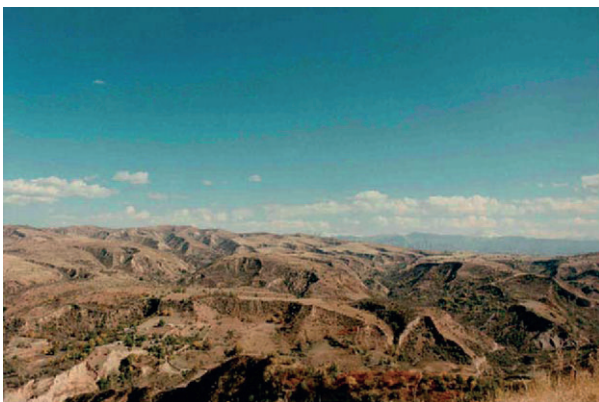


Figure 3 Loess plateau ~2000 m a.s.l. in the Western Pamirs piedmont.



Figure 6 Erosion in the loess mantle.



Figure 7 Damaged loess mantle resulting in collapsing and landsliding.



Figure 8 The Lakhuti loess reference section (a) cut at the top by a deep trench; (b) preparation of the trench; (c) vertical trench in the loess stratum.

of scientists accept that primary loess has an eolian origin, but most also recognize that in some areas, loess has been intensively reworked, and redeposited by slope processes and running water that results in what are called 'loess-like' deposits.

Central Asian deserts and piedmont plains have been subjected to deflation processes for many hundred thousands of years in the Quaternary. Vast amounts of silt material were formed from frost and thermal weathering as well as glacial grinding in high mountain areas that have glaciers or had them previously. Silt was washed out to become a part of the fluvial system in mountain valleys, piedmont fans, deserts and semi-deserts, extending over vast areas with an arid climate. Many millions of tons of dust from the arid deflation zone have been transported by dust storms and have accumulated on river terraces, plateaus, and mountain slopes (Figure 9). In the piedmonts of high mountain ranges, loess distribution is usually limited to elevations of 2000–2500 m, although fragments of thin loess cover occur even higher. For instance, in the Pamirs thin loess mantles can be encountered at an elevation of 4000 m, and in the northern Kunlun mountains, loess deposits are present at 4500 m.

The Central Asian loess region, to the north and west of the Pamirs, covers parts of several countries, including Uzbekistan, Kyrgyzstan, southern Kazakhstan, Tajikistan, Turkmenistan, and northern Afghanistan (Figure 10). The very arid territory of Xinjiang is east of the Pamirs. Loess is characteristic for the marginal parts of great Central Asian deserts such as the Takla Makan, Gobi, Karakumy, Kyzylkum, and Muyunkum (Figure 11). In the southern part of Central Asia, to the south of the Hindu Kush and Himalayas, loess is found in the intra-montane Kashmir valley and along the Siwalik foothills, to the north of the Thar desert, in the Punjab, the Peshawar basin, and the Potwar plateau. The sources of eolian dust for loess regions are the deflated sands and silts in deserts, fluvial and lacustrine deposits, and alluvial (proluvial) fans along

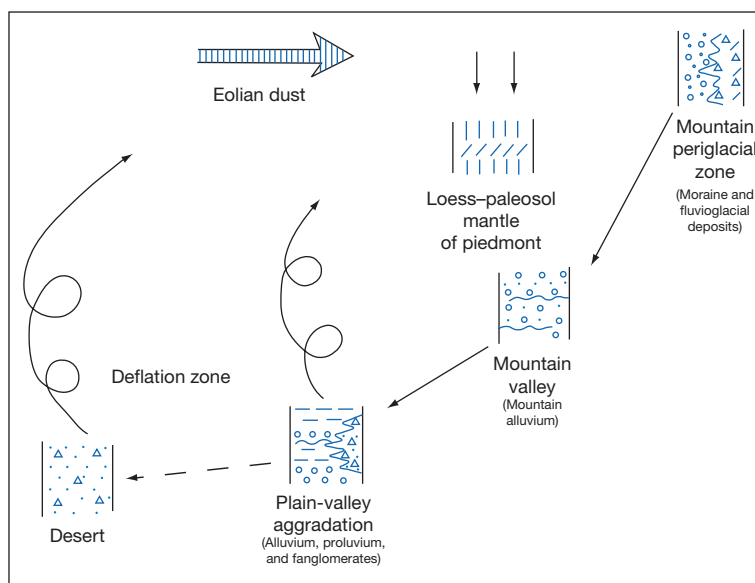


Figure 9 General scheme of silt material movement and loess formation processes in Central Asia. Reproduced from Dodonov AE (2002) *Quaternary of Middle Asia: Stratigraphy, Correlation Paleogeography*, p. 250. Moscow: GEOS (in Russian); Figure 70, with permission from GEOS.

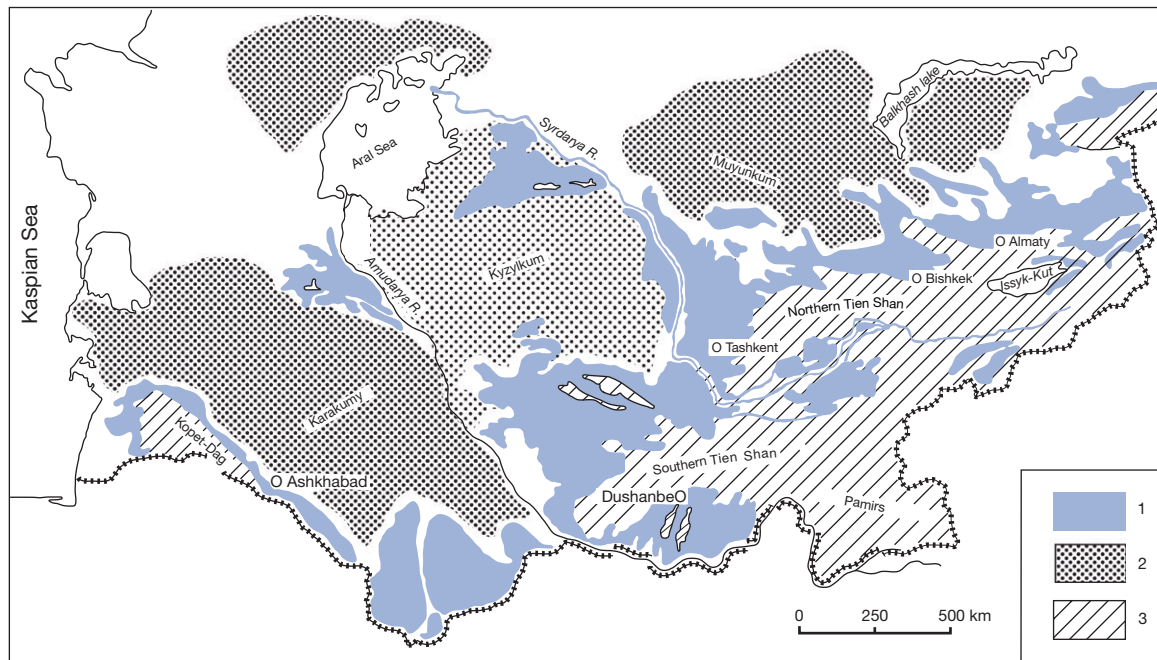


Figure 10 Schematic map of loess cover and sandy deserts in FSU Central Asia: (1) loess and loess-like deposits, (2) sandy deserts, (3) mountain bedrocks. Reproduced from Dodonov AE (2002) *Quaternary of Middle Asia: Stratigraphy, Correlation Paleography*, p. 250. Moscow: GEOS (in Russian); Figure 26, with permission from GEOS.

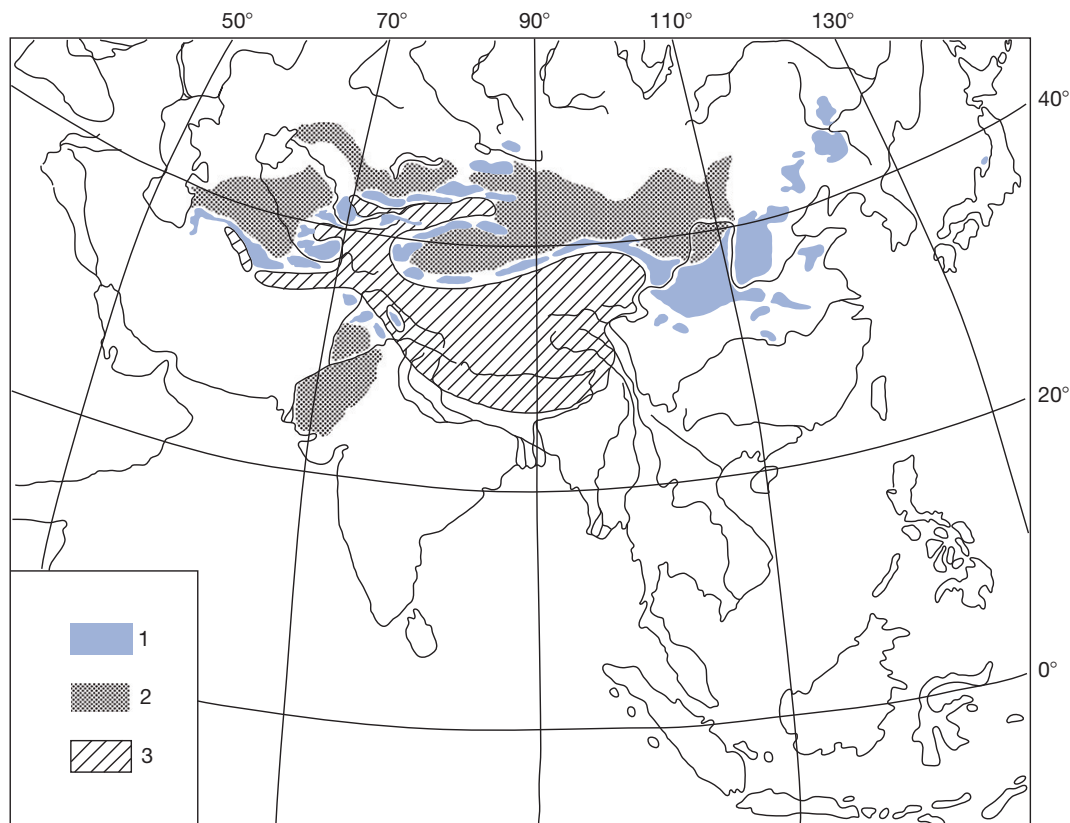


Figure 11 Sketch map of loess deposits and sandy desert distribution in Central Asia with adjacent areas of northern China (1) loess cover, (2) sandy deserts, (3) the uplifted regions, higher than 3000 m a.s.l. Reproduced from Dodonov AE (2002) *Quaternary of Middle Asia: Stratigraphy, Correlation Paleography*, p. 250. Moscow: GEOS (in Russian); Figure 44, with permission from GEOS.

mountain fronts. In the southern part of Central Asia, to the south of the Hindu Kush, the thickness of loess cover is variable, from a few meters to 20–30 m. In the northern part of Central Asia, in key sections of the Tien Shan and Western Pamirs piedmonts, the loess mantle is thicker, particularly in southern Tajikistan and Uzbekistan, where it attains a thickness of 100–200 m.

Paleoclimatic fluctuations are manifested in loess strata as the alternation of soil and loess horizons. Many scientists assume that loess horizons in Central Asia were formed during the glacial/cold epochs, whereas formation of soils corresponded to interglacials or warm phases. According to another viewpoint, paleosols are associated with pluvial events correlated with glaciations, while loess horizons correspond to interglacials. As mentioned above, the low-latitude loesses of this arid zone lack cryogenic features, which is the reason they are called 'warm loess.' Nevertheless, apart from this characteristic, 'warm loess' is not really associated with interglacial periods. Loess-forming processes correlate with the increasing aridization and continental conditions associated with glacial periods, accompanied by considerable expansion of desert and desert-steppe landscapes in the lower piedmont zone. The available age control, data on paleosols, and paleontological and archaeological evidence indicate that regionally manifested soil-forming intervals are definitely related to interglacial events (An et al., 1991; Bronger and Heinkelé, 1989; Dodonov et al., 1999; Liu et al., 1985).

On the basis of analytical data and observation of loess in many localities of Central Asia, such as Tajikistan, Uzbekistan, and Kyrgyzstan, loess horizons can be described as well-sorted grayish-pale-yellow, homogeneous, porous clay-silt with a carbonate content up to 15–20%. Abundance of root-like calcite forms and amorphous concentrations of pelitic calcite are characteristic. In the north of Central Asia, the coarse silt fraction (0.05–0.01 mm) is dominant, ranging from 35 to 45% of the total and nearly 40% on average. There is a high content of the medium silt fraction (6–20 μ m, 30–40%) reported by Bronger et al. (1987) from Kashmir loess. A significant amount of medium silt suggests that the dispersed material was at least partially transported from remote territories located beyond the bounds of Kashmir valley. On the other hand, according to the data of Lodha et al. (1988), the average concentrations of elements such as K, Ti, and Fe in the loess–paleosol sequences of Kashmir correlate well with the rock mineralogy of the region, showing a dominantly local source of loess material. Particle size studies of Potwar loess in the Middle Soan valley (Rendell, 1988) reveal that the loess comprises 87–90% silt (silt-clay, according to Rendell) and 15–16% clay-size particles, and calcium carbonate contents in the range of 14–18%. The composition of loess in the Peshawar basin is very similar to this.

The regionally well-developed paleosols are gray, brown, and reddish brown in color and are decalcified to various degrees. They have an essentially clayey composition, the clay (<0.001 mm) content is up to 20–30%. Carbonate content varies from 0–3% in B horizons to 30–45% in underlying illuvial carbonate horizons. Most fossil soils are polygenetic, with interbedded zones of carbonate concretions. The genetic type of paleosols is generally recognized as the Serozem and Cinnamon types in the Russian soil classification (Egorov

et al., 1977), which correspond to the Haplic Calcisol and Castanozem types in the international soil classification (World Reference Base, 1998). A majority of the middle and late Pleistocene paleosols, in southern Tajikistan, are similar to the surface Holocene soils, formed at altitudes of ~1000–2200 m a.s.l. under a subtropical climate with a predominantly seasonal precipitation and vegetation characterized by broadleaf forest, or open forest-steppe, associated with grasses and shrubs. In the loess–paleosol sequences of Central Asia, the paleosols are frequently grouped into pedocomplexes (PC). They are defined as comprising two or more soils ('pedomembers'), which overlap each other or are separated by thin deposits within one member, and are overlain and underlain by greater thicknesses of loess separating different pedomembers/PC.

The loess-soil record in Central Asia suggests a long-term trend toward colder and drier conditions during the Quaternary. This trend is related to the uplift of Tibet, and the Tien Shan, Pamirs, and Himalayas mountain ranges. The uplift of mountains resulted in the extension of mountain glaciation and the appearance of a periglacial zone. Glacial and periglacial processes produced abundant silt material that was carried over by rivers to the piedmont plains, where it was deflated. Furthermore, the uplift of high mountain ranges created barriers for the transfer of moist air masses from the westerlies and the Indian monsoon. This resulted in a general desiccation of intramontane depressions and plateaus. Due to the enlargement of arid areas and more favorable conditions for accumulation of the fine-grained material, the broadening of loess regions in the middle and late Pleistocene resulted in the addition of the Kashmir valley, Peshawar basin, and Potwar plateau to the Central Asian loess province.

The loess–paleosol formations of Central Asia provide one of the most significant climatostratigraphic records on the continent. Paleomagnetic investigations and paleontological studies in Central Asia, especially in southern Tajikistan, have suggested an age of ~2.5 Ma for the oldest loess horizons. The most complete loess sequences in the Tajik Depression provide a detailed record of climate changes for the last 2.5 My. Over 40 paleosol horizons are interpreted to have formed during this period. A major shift toward a cooler and drier climate occurred at about 1–0.8 Ma, that corresponds to the interval between the Jaramillo Suchron and the Matuyama–Brunhes reversal. At this level, these changes are definitely marked by an increase in the thickness of loess horizons. The Matuyama–Brunhes reversal (~780 ka) is located in the loess horizon between the tenth and ninth PC, or at the base of PC9.

The oldest loess corresponds to the lower part of the Matuyama paleomagnetic interval, below the Olduvai Subchron. According to paleomagnetic markers used for estimating the duration of loess + paleosol pedosedimentary cycles, in the early Pleistocene, one loess–paleosol cycle lasted 35–50 ka; in the middle and late Pleistocene, each cycle had a duration of about 100 ka. These estimates reveal a correspondence with the general trend of paleoclimatic cyclicity reflected in the oxygen-isotope record for the past 2.5 Ma (Ruddiman et al., 1986; Shackleton, 1995).

Stratigraphic data obtained from loess sequences in the Kashmir valley, the Potwar plateau, and the Peshawar basin

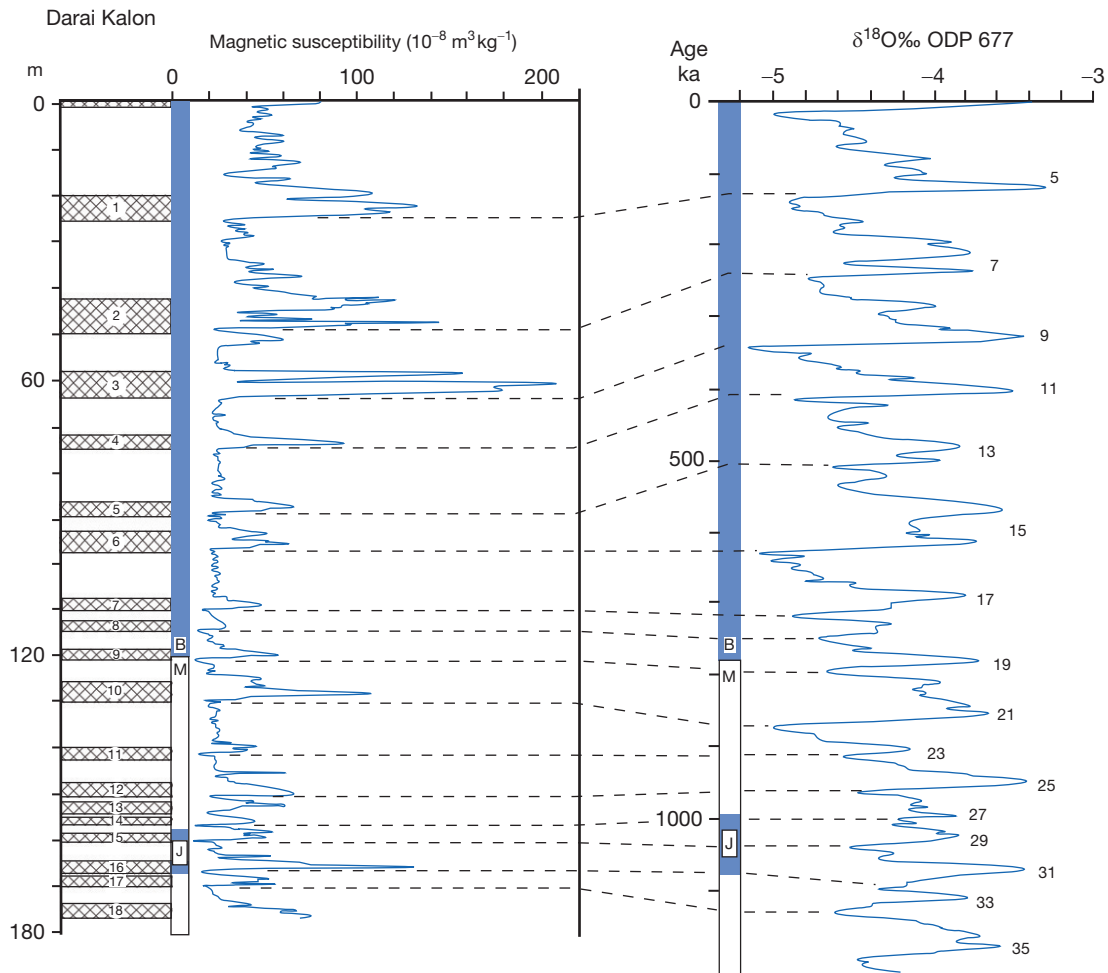


Figure 12 Magnetic susceptibility records on loess–paleosol succession in Darai Kalon, southern Tajikistan, and its correlation with oxygen-isotope curve. Reproduced from Dodonov AE (2002) *Quaternary of Middle Asia: Stratigraphy, Correlation Paleography*, p. 250. Moscow: GEOS (in Russian); Figure 57, with permission from GEOS.

show that the initial stage of loess formation in southern Central Asia took place much later in comparison with northern regions, depending on the rate of aridification and sedimentary environment. In the Kashmir valley, a 20–30 m thick loess cover contains the last interglacial soil and below it there are at least four mostly polygenetic paleosol horizons corresponding to interglacials (Bronger et al., 1987). According to paleomagnetic measurements and thermoluminescence (TL) dating, their estimated age is not older than the middle part of the middle Pleistocene, although TL ages greater than ~100 ka have an uncertain reliability. Loess covers of the Potwar plateau and the Peshawar basin are thinner and apparently lack paleosols. Their TL age is mainly in the range of the late Pleistocene, with one earlier TL date of 170 ka (Rendell, 1988).

In the loess sequence of southern Tajikistan, the Brunhes paleomagnetic epoch includes nine PC. A similar sequence has been documented in the Tashkent region of Uzbekistan. The magnetic susceptibility record for this interval shows pronounced peaks corresponding to paleosols, which correlate reasonably well with those of the warm stages in the marine oxygen-isotope record (Figure 12). The geochronology for

the upper pedocomplex PC1 of Central Asia has been confirmed by TL dating (Frechen and Dodonov, 1998; Zhou et al., 1995). TL measurements yield ages in the range of 130–70 ka for PC1, establishing its correlation with stage 5 of the marine isotope record. Age markers such as PC1, corresponding to the last interglacial, and the position of Matuyama–Brunhes reversal at 780 ka have been used as the reference age levels for the general calibration of the Central Asian loess–paleosol succession as well as for correlation with oxygen-isotope stages. For the Brunhes Chron, the correlation of the loess–paleosol succession of Tajikistan and northern China as well as with the oxygen-isotope curve is illustrated in Figure 13.

Geochronology for the upper paleosols is very important for the reconstruction of early human evolution, because many Paleolithic sites have been discovered in pedocomplexes PC6, PC5, PC4, PC3, PC2, PC1 (Dodonov, 1995, 2002; Lazarenko and Ranov, 1977; Ranov and Schäfer, 2000). It has been established that PC6, PC5, PC4, and PC3, which contain stone artifacts of a pebble culture, span the time range between ~650 and ~300 ka. Younger pedocomplexes, PC2 and PC1, with stone implements of the middle

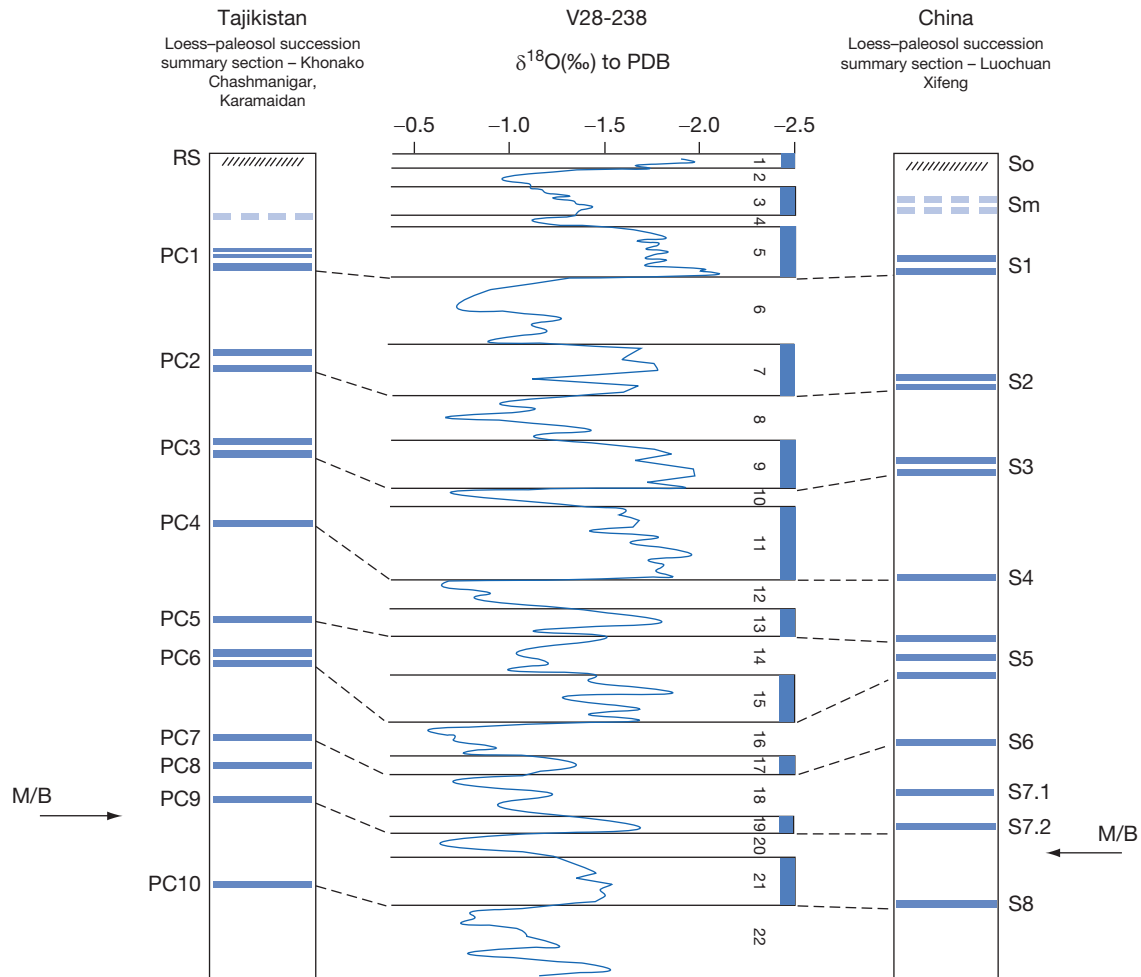


Figure 13 Stratigraphic correlation of Tajikistan summary loess–paleosol section with oxygen-isotope curve and summary loess–paleosol section of China. B/M – position of the Matuyama–Brunhes paleomagnetic reversal. Reproduced from Dodonov AE (2002) *Quaternary of Middle Asia: Stratigraphy, Correlation Paleography*, p. 250. Moscow: GEOS (in Russian); Figure 74, with permission from GEOS.

Paleolithic, are assumed to be in the range of ~250 to ~70 ka. The earliest stone artifacts have been found below the Matuyama–Brunhes paleomagnetic reversal in paleosols 11 and 12, which are estimated to be around 900 ka (Ranov and Dodonov, 2003). The most recent studies of loess–paleosol sequences demonstrate that the relatively wide expansion of Paleolithic people in the loess province of the Tien Shan and Pamirs piedmonts took place during the middle Pleistocene, whereas the earliest, stratigraphically proved, appearance of ancient man occurred in the late early Pleistocene.

It should be emphasized that numerous Paleolithic finds in the northern part of Central Asia have been recovered from well-developed paleosol horizons. The paleoclimatic conditions during the soil-forming epochs favored the habitation of ancient people. During loess-forming intervals, deteriorating climatic conditions, with severe winters and frequent dust storms, yielded abundant dust accumulation, and caused human migration away from zones of intensive loess formation to other areas. It is worth noting that ancient people returned to the same loess regions during later periods of soil formation. According to paleoenvironmental reconstructions, the archaeological sites in loess sections were located at river

banks, on slopes along river valleys and on river divides. Ancient people used pebbles from alluvial gravels for making tools, and water from the springs in gorges. The loess watersheds were a good place for hunting. The Paleolithic artifacts are mostly found in Bt horizons of the interglacial PC and are attributed to climatic optima. The habitats of hominids were probably forest-steppe and steppe, based on paleoclimatic reconstructions. Notably, the paleosol-related pebble tools do not permit the identification of distinct cultural horizons. The stone artifacts are generally dispersed throughout the paleosol profile. This situation may be accounted for by the accumulation of dust material during the soil-forming epochs, that is, it corresponds to a syndepositional model of soil development. Another factor in the vertical distribution of artifacts in paleosols is the recurrent occupation of ancient people at the same site. All these factors contributed to the significant depth range of artifacts and is a very characteristic pattern for loess Paleolithic sites in Central Asia.

In the southern part of the Central Asian loess region, the presence of the Late Soan stone industry in the loess cover on the Soan river terrace (T_2), 30–40 m height, as well as in the basal alluvial gravels has been reported by de Terra and Paterson

(1939) and supported by Movius (1944). It has been inferred from the geoarchaeological context in the Soan valley that occupation by the Soan people covered both the pluvial and interpluvial phases. Therefore, in this region, Paleolithic humans likely witnessed dust storms during the arid periods.

The pollen spectra obtained from loess–paleosol sequences have a complex, mixed composition, which does not usually show distinct differences between the loess horizons and paleosols. Nevertheless, the palynological data from one of the reference sections, Darai Kalon in southern Tajikistan, demonstrate a rather good agreement with the pedological features. Arboreal pollen dominates in the well-developed, typical and leached Cinnamon Soils of PC4, PC3, PC2, and PC1. The lower paleosol members in these pedocomplexes are the most well developed and show soil formation under a more humid climate, whereas the upper soils are less developed and reflect drier environments. Pollen assemblages of PC1 correlate well with the paleopedological data, demonstrating the upward decrease of arboreal pollen and increase of grasses. The upper paleosol of PC1, assigned to the Serozem soil type, is characterized by a predominance of herb and shrub pollen.

The mammalian fossils contained in the loess–paleosol sequences and associated sediments reveal the regional Quaternary evolutionary history of mammals and provide useful biostratigraphic data, as well as reflecting the ecological conditions in loess regions of Central Asia. The paleoecological study of the late Pliocene–early Pleistocene fauna in the northern Central Asian localities, together with palynological data, shows the mosaic-like structure of paleolandscape and distinct altitudinal zonation. For instance, at the Kuruksay locality (~2.4–2.0 Ma), in southern Tajikistan, the inhabitants of dry, open landscapes such as hyenas (*Pachycrocuta*, *Chasmaporthetes*), antelopes (*Proteryx*, *Damalops*), camels, and cheetahs are characteristic, while the presence of bears, cervids, and primates (*Paradolichopithecus*) suggests a landscape with mixed shrub and bush areas and forest belt in the high piedmont (Sotnikova et al., 1997). At that time, loess had just started to be deposited, covering the watersheds and slopes, and reworked by pedogenic processes, resulting in the formation of pedosediment. The mammalian fauna from the next interval, at the end of the early Pleistocene, as demonstrated in the thoroughly studied Lakhuti locality of southern Tajikistan, yielded an open steppe fauna, such as hyenas, camels, and horse, as well as cervids (*Sinomegaceros* and *Praemegaceros*) connected with woodland patches. Among small mammals, inhabitants of the arid zone, *Ellobius*, *Meriones*, and *Cricetulus*, were also present. Carnivores such as the small wolf, *Xenocyon*, the giant hyena *Pachycrocuta brevirostris*, and felids were also common elements (Sotnikova et al., 1997). It is worth noting that loess accumulation significantly increased from the beginning of the middle Pleistocene. During the periods of loess sedimentation along the Siwalik foothills, there were dry plains and plateaus inhabited by horse, bison, camel, and wolf.

The geochronological data and paleomagnetic markers are accepted as a basis for evaluating the mean loess sedimentation rates during the Quaternary, using the thickness of loess–paleosol series. An approximation of loess sedimentation rates for key loess sections in southern Tajikistan results in a trend of loess accumulation in a range from 0.05 to 0.16 mm year⁻¹

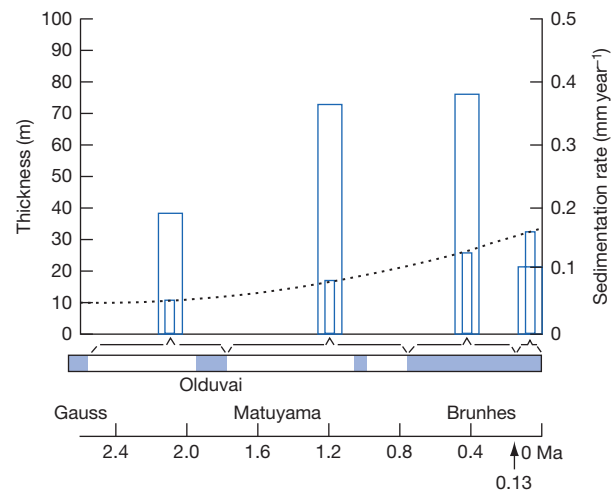


Figure 14 Variations in the thickness of loess–paleosol strata, marked by wide columns, and mean sedimentation rate, narrow columns, during the last 2.5 Ma. Age boundaries in the loess reference sections of southern Tajikistan, Chashmanigar, Karamaidan, Khonako, Darai Kalon, Kairubak, Lakhuti, and Karatau, dated by paleomagnetic and thermoluminescence methods have been used for estimation of loess sedimentation rate. Reproduced from Dodonov AE (2002) *Quaternary of Middle Asia: Stratigraphy, Correlation Paleogeography*, p. 250. Moscow: GEOS (in Russia); Figure 75, with permission from GEOS.

during the last 2.5 Ma (Figure 14). This estimation of rates does not take into account the duration of soil formation, when dust sedimentation was steadily reduced. The mean sedimentation rate of loess itself, excluding the paleosols, attained an average rate of 0.2–0.3 mm year⁻¹ and even up to 0.4 mm year⁻¹ (Figure 15). On the basis of TL ages, the sedimentation rate varied from 0.08 to 0.27 mm year⁻¹ in the Potwar plateau and the Peshawar basin for the late Pleistocene (Rendell, 1988). It should be noted that proportional changes of sedimentation rates have been calculated for different loess provinces, for example, southern Tajikistan and Loess Plateau of China, as illustrated in Figure 14. This demonstrates the general regularities in development of eolian processes in different regions of Asia, affected by uniform paleoclimatic changes.

In the current conditions of Central Asia, to the north of Hindu Kush, the summer–autumn period is favorable for dust storms, owing to minimal precipitation and intense drying of soil covered by scarce vegetation. The number of dust-storm days per year ranges from five in the Pamir–Alai and Tien Shan valleys to 50 and more in Turkmenistan, Uzbekistan, and other areas of sandy grounds not covered by vegetation (Figure 16). In the Takla Makan desert, dust storms occur on more than 30 days per year (Pye and Zhou, 1989). Today, intense deflation and dust accumulation are observed in the high mountain valleys in the glacial zones of the Eastern Pamirs and the Central Tien Shan, at the altitude of 3500–4000 m a.s.l. To the south of the Hindu Kush and Himalayas, the Central Asian loess province is located within the limits of the Indian monsoon, where the bulk of precipitation occurs mainly in the second half of the summer and the early autumn. Here, intense eolian dust falls during the dry season, which lasts up to 9 months. More than 30 dust storms per year occur in northern Pakistan.

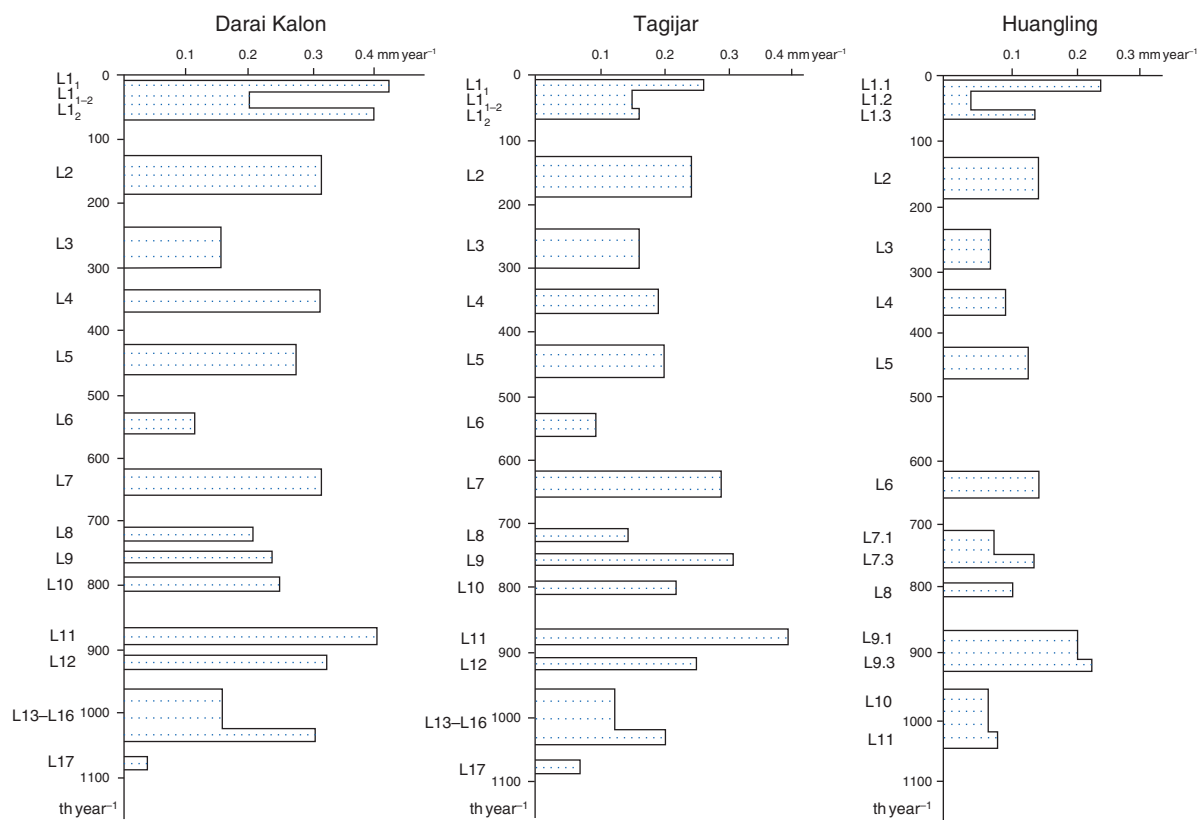


Figure 15 Loess sedimentation rate for the last 1 Ma excluding the time span of paleosol formation. The data on the loess reference sections Darai Kalon (after Dodonov) and Tagijar (after Dodonov and Haesaerts) in southern Tajikistan, and the Huangling section in northern China (after Haesaerts). Reproduced from Dodonov AE (2002) *Quaternary of Middle Asia: Stratigraphy, Correlation Paleogeography*, p. 250. Moscow: GEOS (in Russia); Figure 30, with permission from GEOS.

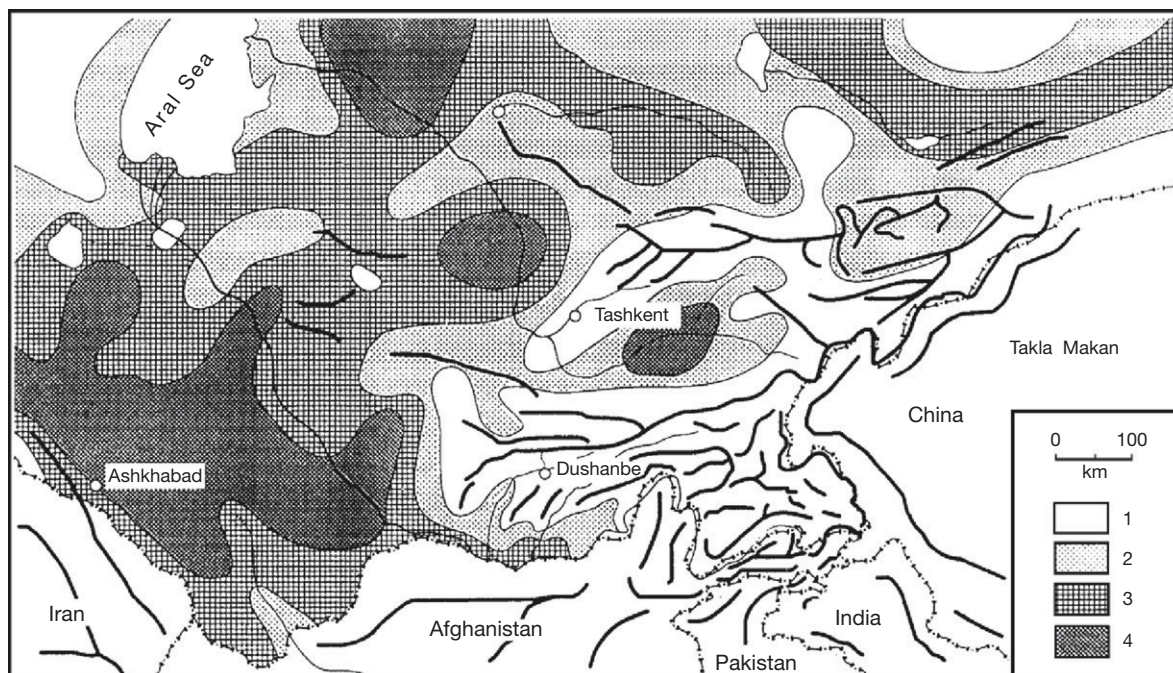


Figure 16 The annual frequency of dust storms in Central Asia to the north of the Hindu Kush. Reproduced from Dodonov AE, Shackleton N, Zhou LP, Lomov SP, and Finaev AF (1999) *Quaternary loess-paleosol stratigraphy of Central Asia: Geochronology, correlation and evolution of paleoenvironments. Stratigraphy and Geological Correlation* 7(6): 66–80.

Loess areas of Central Asia are highly populated regions, with high intensity of agriculture and industrial development. Thus, these areas are very sensitive to geological changes and geotechnical impact. One of the problems regarding the exploitation of loess-mantled landscapes is collapsibility. The studies carried out in the former Soviet Central Asia indicate that in the areas with thick loess cover, the thickness of the collapsible zone attains a depth of 30–40 m from the surface. Collapse properties are explained by the lack of compact quality of loess as a result of cementation of loosely deposited sediments under arid climate. This characteristic, called Denisov's principle, was considered in depth by Kriger (1965, 1986) and his collaborators. The model is in good agreement with the concepts of the eolian genesis of Central Asian loess. Another problem connected with changing of the water balance over vast loess areas is the rational supply of water resources in Central Asia. Areas covered by permeable loess-derived soils are often exposed to underflooding. The highly populated cities of Central Asia such as Tashkent, Dushanbe, and others suffer greatly from the underflooding processes, creating collapsibility and settlement of soil foundations, deformation of buildings, etc. Many geological processes, which are active in Central Asian loess regions, such as erosion, suffusion, collapsing, landsliding, and underflooding, are intensified in connection with human activities. Another very serious factor in Central Asia is seismic activity. Loess mantles are highly susceptible to seismic movements. In loess areas, seismic subsidence, landslides, and slope earthflow are characteristic results of earthquakes. We are now in an epoch of construction and very progressive development of Central Asian countries. Industrial and civil buildings, highways, hydrotechnical and irrigation systems are gradually modifying and removing the very sensitive loess cover.

See also: Loess Deposits: Origins and Properties. **Loess Records:** China.

References

- An Z, Wu X, Wang P, et al. (1991) Changes in the monsoon and associated environmental changes in China since the last interglacial. In: Liu T (ed.) *Loess, Environment and Global Change*, pp. 1–29. Beijing: Science Press.
- Berg LS (1947) *Climate and Life*, 2nd edn., p. 356. Moscow: Geografiz (in Russian).
- Bronger A and Heinkel Th (1989) Paleosol sequences as witnesses of Pleistocene climatic history. *Paleopedology, Catena Supplement* 16: 163–186.
- Bronger A, Pant RK, and Singhvi AK (1987) Pleistocene climatic changes and landscape evolution in the Kashmir basin, India: Paleopedologic and chronostratigraphic studies. *Quaternary Research* 27: 167–181.
- de Terra H and Paterson TT (1939) *Studies of the Ice Age in India and Associated Human Cultures*, p. 354. Washington: Carnegie Institute Publications.
- Dodonov AE (1995) Geoarchaeology of palaeolithic sites in loesses of Tajikistan (Central Asia). In: Johnson E (ed.) *Ancient Peoples and Landscapes*, pp. 127–136. Lubbock: Museum of Texas Tech University.
- Dodonov AE (2002) *Quaternary of Middle Asia: Stratigraphy, Correlation, Paleogeography*, p. 250. Moscow: GEOS (in Russian).
- Dodonov AE, Shackleton N, Zhou LP, Lomov SP, and Finaev AF (1999) Quaternary loess-paleosol stratigraphy of Central Asia: Geochronology, correlation and evolution of paleoenvironments. *Stratigraphy and Geological Correlation* 7(6): 66–80.
- Egorov VV, Fridland VM, Ivanova EN, Rozov NN, Nosin VA, and Fraev TA (1977) *Classification and Diagnostics of Soils of USSR*, p. 223. Moscow: Kolos Press (in Russian).
- Frechen M and Dodonov AE (1998) Loess chronology of Middle and Upper Pleistocene in Tadjikistan, Central Asia. *Geologische Rundschau* 87: 2–20.
- Gerasimov IP (1976) *Genetic, Geographical and Historical Problems of Recent Pedology*, p. 298. Moscow: Nauka (in Russian).
- Kriger NI (1965) *Loess, its Characteristic and Relation to the Geographical Environment*, p. 295. Moscow: Nauka (in Russian).
- Kriger NI (1986) *Loess – Origin of Hydrocompaction Properties*, p. 132. Moscow: Nauka (in Russian).
- Lazarenko AA and Ranov VA (1977) The earliest palaeolithic site of Middle Asia loesses. *Bulletin of the Commission on the Quaternary Research* 48: 45–57 (in Russian).
- Liu T, et al. (1985) *Loess and the Environment*, p. 251. Beijing: China Ocean Press.
- Lodha GS, Sawhney KJ, and Razdan H (1988) Characterization of loess-paleosol sequences in Kashmir (India) valley using multi-element concentration data. In: Jain DVS (ed.) *Palaeoclimatic and Paleoenvironmental Changes in Asia During the Last 4 Million Years*, pp. 33–45. New Delhi: Indian National Science Academy.
- Mavlyanov GA (1958) *Genetical Types of Loesses and Loess-Like Rocks in the Central and Southern Parts of Central Asia and Their Engineering-Geological Properties*, p. 609. Moscow: Izd-vo Acad. Nauk Uzbek SSR (in Russian).
- Minervin AV (1984) Cryogenic processes in loess formation in Central Asia. In: Velichko AA (ed.) *Late Quaternary Environments in the Soviet Union*, pp. 133–140. London: Longman.
- Movius HL (1944) Early man and Pleistocene stratigraphy in Southern and Eastern Asia. In: *Papers of the Peabody Museum of American Archaeology and Ethnology*, vol. XIX(3), p. 125. Cambridge, MA: Peabody Museum of American Archaeology and Ethnology, Harvard University.
- Obruchev VA (1933) The problem of loess. *The Transactions of the 2nd International Congress of Association on Quaternary Research* 2: 115–137.
- Pavlov AP (1903) On Turkestan and European loess. *Byulleten' Moskovskogo Obshchestva Ispytatelej Prirody* 4: 43–67 (in Russian).
- Pécsi M (1993) Quaternary and loess research (Summary). In: Pécsi M (ed.) *Negyedcoreás löszkutatás*, pp. 293–332. Budapest: Akadémiai Kiadó.
- Pye K and Zhou LP (1989) Late Pleistocene and Holocene eolian dust deposition in North China and the Northwest Pacific Ocean. *Palaeogeography, Palaeoclimatology, Palaeoecology* 73: 11–23.
- Ranov VA and Dodonov AE (2003) Small instruments of the Lower Palaeolithic site Kuldara and their geoarchaeological meaning. In: Burdukiewicz JM and Ronen A (eds.) *Lower Palaeolithic Small Tools in Europe and the Levant*, pp. 133–147. Oxford: Archaeopress/British Archaeological Reports. BAR International Series 1115.
- Ranov VA and Schäfer J (2000) The Palaeolithic of the late Middle Pleistocene in Central Asia, 400–100 ka ago. In: Ronen A and Weinstein-Evron M (eds.) *Toward Modern Humans – The Yabroudian and Micoquian, 400–50 kyears ago*, pp. 77–92. Oxford: Archaeopress/British Archaeological Reports. BAR International Series 850.
- Rendell HM (1988) Environmental changes during the Pleistocene in the Potwar Plateau and Peshawar Basin, Northern Pakistan. In: Jain DVS (ed.) *Palaeoclimatic and Paleoenvironmental Changes in Asia During the Last 4 Million Years*, pp. 58–66. New Delhi: Indian National Science Academy.
- Ruddiman WF, Raymo M, and McIntyre A (1986) Matuyama 41,000-year cycles: North Atlantic Ocean and northern hemisphere ice sheets. *Earth and Planetary Science Letters* 80: 117–129.
- Shackleton NJ (1995) New data on the evolution of Pliocene climatic variability. In: Vrba ES, Denton GH, and Patridge TC, et al. (eds.) *Palaeoclimate and Evolution with Emphasis on Human Origins*, pp. 242–248. New Haven: Yale University Press.
- Smalley IJ, Jefferson IF, Dijkstra TA, and Derbyshire E (2001) Some major events in the development of scientific study of loess. *Earth-Science Reviews* 54: 5–18.
- Sotnikova MV, Dodonov AE, and Pen'kov AV (1997) Upper Cenozoic bio-magnetic stratigraphy of Central Asian mammalian localities. *Palaeogeography, Palaeoclimatology, Palaeoecology* 133: 243–258.
- von Richthofen F (1882) On the mode of origin of the loess. *Geological Magazine* 9(7): 293–305.
- WRB (World Reference Base) (1998) World Reference Base for soil resources. In: *World Soil Resources Reports* 84, p. 88. Rome: Food and Agricultural Organization of the United Nations.
- Zhou LP, Dodonov AE, and Shackleton NJ (1995) Thermoluminescence dating of the Orkutsay loess section in Tashkent Region, Uzbekistan, Central Asia. *Quaternary Science Reviews* 14: 721–730.

China

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Introduction

Loess consists largely of silt-size mineral grains that have been deposited by wind to form homogeneous nonstratified sediment of local to regional extent. Widespread loess is now found mainly in temperate latitudes of central Asia, northwestern and central USA, central Europe, and Argentina. Loess also is present in high northern latitudes, including northern Russia and interior Alaska, and reportedly occurs in part of tropical South America. Most deposits are of Pleistocene age and typically formed under arid to semiarid conditions. Major loess regions lie downwind from source areas that include deserts, periglacial zones, and the floodplains of melt-water streams that drain large piedmont glaciers and ice sheets. Loess typically forms a blanket across the landscape. Where thin it mirrors the underlying topography, but where thick (≥ 100 m) it may form a constructional surface that obscures underlying bedrock terrain. Because it consists of fine-grained, little-weathered minerals, in humid-to-semihumid regions loess generates rich soils that support crops to feed millions of people.

Loess in China

Central China is one of the world's most extensive and important loess regions. The loess caps a broad dissected plateau, the Loess Plateau, which lies largely within the great northward bend of the Yellow River (Huang He) as it emerges from the Tibetan Plateau near Lanzhou (Figure 1). It forms distinctive landscapes, to which the Chinese have given an elaborate series of specific names (e.g., yuan (high, flat tableland), ta (rounded hills), liang (ridges)). The flat-topped yuan are extensively farmed, as are nearly all gentle slopes in the loess region. Deep valleys cut into the coherent loess deposits often have steep cliffs in which can be seen alternating bands of yellowish loess and reddish-brown buried soils (paleosols) that form a record of changing climate conditions through time (Figure 2).

Monsoon Climate of Eastern Asia

The history of Chinese loess deposition is closely linked to the East Asian monsoon climate system (Figure 3). The monsoon is driven by differential heating of the land and adjacent ocean. During the annual transition from winter to summer, northern hemisphere insolation slowly increases. The land warms more rapidly than the ocean, thereby strengthening the temperature contrast between the warm Asian continent and the cooler Pacific Ocean. The resulting pressure gradient between ocean and land produces a steady flow of moist marine air onto the continent during the summer and early autumn. Interaction along the monsoon front between this warm, moist Pacific air and cooler, drier continental air produces a belt of heavy rain

that migrates inland. With the approach of winter, insolation declines, summer monsoon conditions weaken, and cold, dry northwesterly air flows southeastward from a large region of high pressure that expands over north-central Asia. As the winter monsoon strengthens, dust is deflated from arid source regions and is spread across China's loess region and beyond across the North Pacific Ocean. Loess accumulation is greatest during winter and spring when the winter monsoon is strongest, and soils mainly develop under moist, warm summer monsoon climate.

Accretionary Paleosols

Paleosols found in most continental loess sequences represent times when dust sedimentation ceased and soil-forming processes operated on stable land surfaces. In monsoonal China, however, dust deposition is essentially continuous, although variable in rate. Soil formation also takes place continuously as dust accumulates, but at variable rates and intensities. The resulting soils are accretionary soils; when buried by younger loess deposits they become accretionary paleosols. Such paleosols do not record breaks in the stratigraphic record. Instead, they represent intervals when soil-forming processes intensified while dust accumulation proceeded at a slow rate.

Loess/Paleosol Stratigraphy

Stratigraphic studies of the loess and paleosol sequence in the Loess Plateau have used lithology (e.g., color, relative thickness, grain size, and mineralogy) and soil characteristics, as well as magnetic properties as primary criteria for classification and correlation. Magnetic susceptibility (SUS) has proved especially useful in this regard (Figure 4). The primary susceptibility signal originates from *in situ* production of fine-grained magnetic particles (maghemite) by soil bacteria. SUS values are highest for warm, mild interglacial conditions that promote soil formation, and lowest for glacial times when cold, dry conditions prevailed.

The standard stratigraphic succession has been divided into primary numbered loess and paleosol couplets and secondary units within some loess or paleosol units (Figure 5). The letter L is used to designate loess units and S for paleosols. Successive units are given higher numbers with increasing age. Some 32 primary couplets have been identified in a classic loess succession near Baoji in the central loess region, the oldest units being designated L32 and S32 (Figure 6). Particular attention has been given to deposits of the last interglacial/glacial cycle (S1 through L1) and the thin overlying late Holocene ('Potou') loess and early-to-middle Holocene paleosol (S0) (Figure 5). The L1 loess, also called Malan Loess, includes two primary loess units (L1LL1 and L1LL2) and an intervening weak soil or

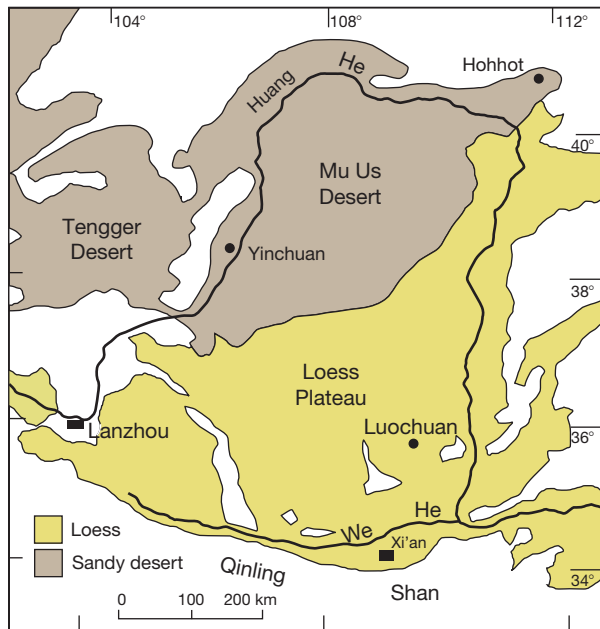


Figure 1 Map of the Loess Plateau and adjacent sandy deserts in central China.

soil complex (L1SS1). The S1 paleosol complex below the Malan Loess consists of three soils separated by two loess units.

Chronology of Loess and Paleosol Units

Ages for eolian strata have been obtained based on several complementary methods:

- radiocarbon dating of the middle and upper Malan Loess (i.e., the last ~50 000 years);
- luminescence dating of the younger part of the classic loess sequence (the last ~200 000 years);
- cosmogenic-isotope dating using the beryllium-10 (^{10}Be) isotope;
- paleomagnetic dating based on interpolation between polarity reversals of known age assuming (i) a nearly constant long-term accumulation rate of dust and weighting the thickness of a layer by its SUS, or (ii) using the concentration of magnetic minerals as an approximation of geologic time;
- grain-size data (i) 'orbitally tuned' to the marine oxygen-isotope record obtained by the SPECMAP (Mapping Spectral Variability in Global Climate Project), following correlation of loess and paleosol boundaries with isotope-event boundaries; (ii) correlated with the SPECMAP record to derive a sedimentation rate for each unit, its duration (ratio of layer thickness to sedimentation rate), and its age (by summing the duration of deposition of all overlying units), or (iii) using the coarse silt ($>40\ \mu\text{m}$) fraction, or the median diameter of the chemically isolated quartz fraction as a proxy for dust deposition rate, ages being assigned based on an accumulation-rate model linked to the SPECMAP record. In the latter approach, the estimated error of ages over the last glacial-interglacial cycle (the last ~130 000 years) is about 1000–3000 years.



Figure 2 A steep stratigraphic section at Luochuan on the central Loess Plateau shows alternating yellowish-brown loess and reddish-brown paleosols (SCP photograph 94-310).

Comparison with the Marine Isotope Record

The basic stratigraphy of the classic Chinese loess succession is strikingly similar to variations in the marine oxygen-isotope stages (MIS; **Figure 6**). Paleosols correlate with odd-numbered isotope stages (interglaciations) and loess units with even-numbered stages (glaciations). The pattern of variations in SUS of the loess and paleosol succession is essentially identical to and correlates with isotope variations in the marine record over at least the last 2.5 My. Both vary in phase with Milankovitch Earth-orbital cycles, implying that the monsoon climate of east-central China has fluctuated synchronously with variations in solar radiation reaching the Earth's surface. Clearly discernible in the loess record is a ~40 000-year (obliquity) cycle that is dominant prior to about 800 000 years ago. It was replaced by a dominant 100 000-year (eccentricity) cycle that extends to the present and also is a major feature of the marine isotope record.

Variations in Grain Size and Thickness

The loess reaches a thickness of 300–400 m in the central and western part of the Loess Plateau, and thins progressively toward the southeast. Very few exposed sections include more than the upper half of the loess sequence, and so the regional pattern of thickness of the oldest units cannot be determined with confidence. Instead, the thickness of the Malan Loess

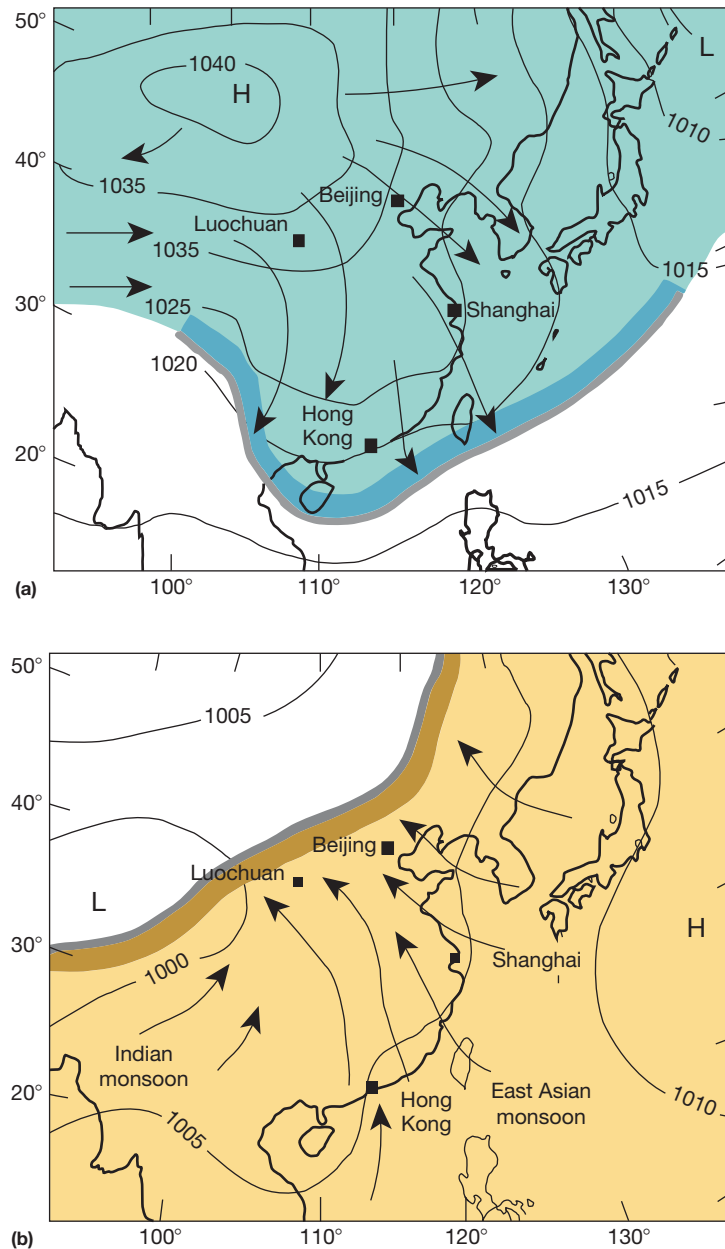


Figure 3 Map of eastern Asia showing seasonal conditions of the East Asian monsoon system. (a) Mean January sea-level pressure field (hPa) generates a southward flow of cold, dry air (arrows) across China's loess region from a zone of high pressure centered over southern Siberia. (b) During July, a reversal of wind flow brings warm, moist air inland across the loess region from the south China Sea (after Xiao et al., 1995, Figure 1).

(L2–L4), which is easily measured at numerous sites, has been used to determine regional thickness variations (Figure 7). This unit is thickest (≥ 30 m) in a belt that extends northeastward from Lanzhou to the southern margin of the Mu Us Desert. Isopachs (lines of equal thickness) decrease in magnitude southeastward toward the junction of the Wei River and Yellow River where the thickness averages about 5 m.

Measurements of the median diameter of loess at the ground surface in transects across the Loess Plateau demonstrate a southeastward decrease in particle size from ~ 80 μm near the desert margin to 20 μm at the southeastern limit of the plateau. Comparable measurements for the Malan and older

thick loess units are difficult to assess, for their grain size varies not only horizontally but vertically as well (a response to changing climate). Therefore, unless the same (thin) stratigraphic level within a loess unit can be sampled at each site, a difficult task due to dating uncertainties, systematic regional trends in grain size are difficult to define accurately.

Potential Upwind Source Regions

The exponential southeastward decrease in thickness of Malan Loess and a decrease in mean particle size of surface loess in

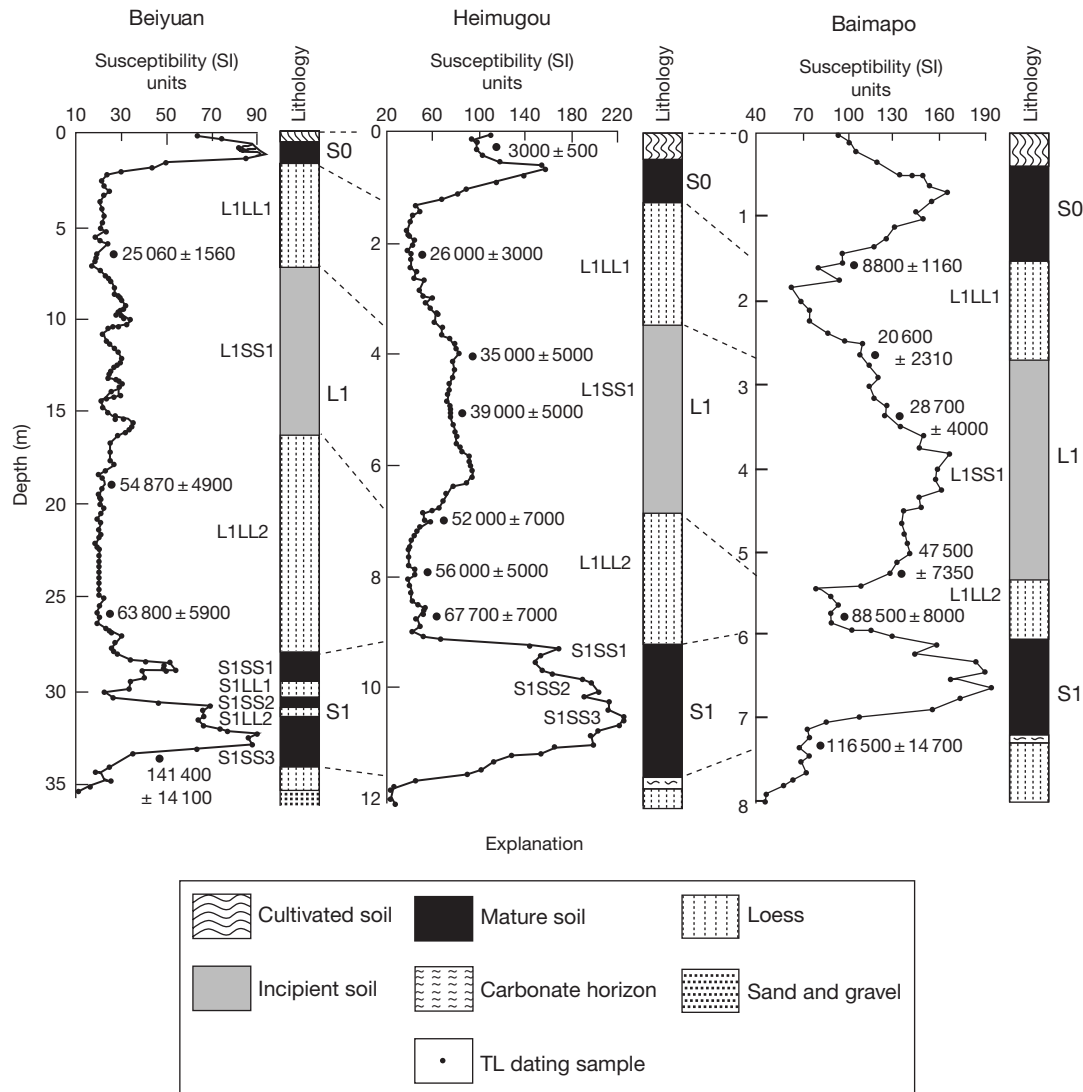


Figure 4 Magnetic susceptibility (SUS) profiles through loess and paleosols spanning the last 130 000 years at three sites on the western (Beiyuan), central (Heimugou), and southeastern (Baimapo) Loess Plateau. The chronology is based on luminescence ages. Reproduced from [An Z, Kukla G, Porter SC, and Xiao J \(1991\)](#) Magnetic susceptibility evidence of monsoon variation on the Loess Plateau of central China during the last 130,000 years. *Quaternary Research* 36: 29–36, Figure 2.

that direction imply that the dust was derived from a source or sources lying northwest of the Loess Plateau. Sandy deserts have long been considered potential sources. Mongolia's Gobi Desert, an oft-cited possible source, is both arid and a source of modern dust storms, some of which have crossed the Loess Plateau. Other likely source regions are desert basins in China west and north of the Loess Plateau (e.g., Mu Us, Tengger, and Badain Jaren deserts; [Figure 7](#)). The vast Takla Makan sandy desert in the Tarim Basin, the dry surface of which is covered by extensive dune fields, playas, and large marginal alluvial fans, is also a potential source ([Figure 8](#)). Glacial silt transported by meltwater streams from surrounding high mountains to alluvial fans and playa surfaces in these basins during successive glacial ages could be lofted by strong westerly windstorms and carried in the direction of the loess region. Mineralogic, isotopic, and geochemical studies in

which silt from these deserts is compared with loess on the Loess Plateau support the likelihood of multiple loess source areas.

Downwind Transport and Deposition

Chinese loess thins rapidly beyond the Loess Plateau, and farther downwind the fine-silt fraction is detectable in sediment cores from Lake Biwa, Japan, as well as in deep-sea sediments across the North Pacific Ocean ([Figure 9](#)). In the Hawaiian Islands, Chinese dust has been identified from its fine-grained quartz component, for quartz does not occur naturally on these mid-ocean islands. Its presence in Greenland ice means that the fine dust must also have been spread across North America as well, but its very fine grain size, extreme thinness and

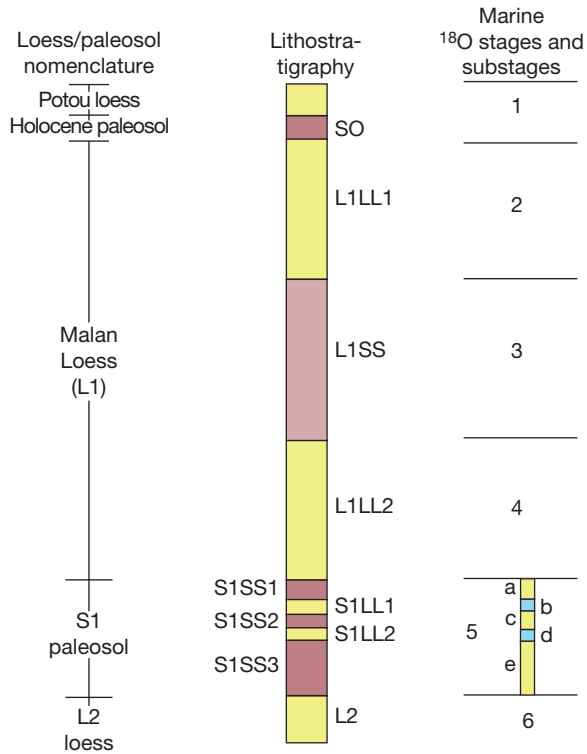


Figure 5 Stratigraphic subdivision and nomenclature of loess and paleosols on the Loess Plateau that span the last ~130 000 years, and their correlation with the standard marine oxygen-isotope stages and substages. Subdivisions are made on the basis of physical, chemical, and biological properties. Reproduced from Porter SC (2001) Chinese loess record of monsoon climate during the last glacial-interglacial cycle. *Earth Science Reviews* 54: 115–128, Figure 2.

erodibility, and nondistinctive mineralogy on the continent make its recognition highly unlikely.

Rates of Dust Influx

It is apparent from the loess and paleosol stratigraphy that the rate of dust accumulation has varied through time: glacial-age loess units are thick compared to interglacial accretionary paleosols. Quantitative estimates of dust influx rates can be obtained from well-dated sections, but these are rare. The section at Luochuan in the central Loess Plateau is among the best studied and most carefully dated. Based on a dust-flux model, the average dust flux during the last 120 000 years was $0.13 \pm 0.02 \text{ mm year}^{-1}$ (Figure 10). During the height of the last glaciation (i.e., loess L1LL1), the average full-glacial rate was $0.30\text{--}0.35 \text{ mm year}^{-1}$, but during the last interglaciation (paleosol S1) it was $<0.10 \text{ mm year}^{-1}$. Along the western margin of the plateau and along the Yellow River near Zhengzhou, loess of the last glaciation is far thicker and influx rates locally were as high as 1 mm year^{-1} .

In the North Pacific Ocean, far downwind from the Loess Plateau, the rate of dustfall has been calculated from deep-sea sediments, the mineral content of which is predominantly air-fall dust from western China.

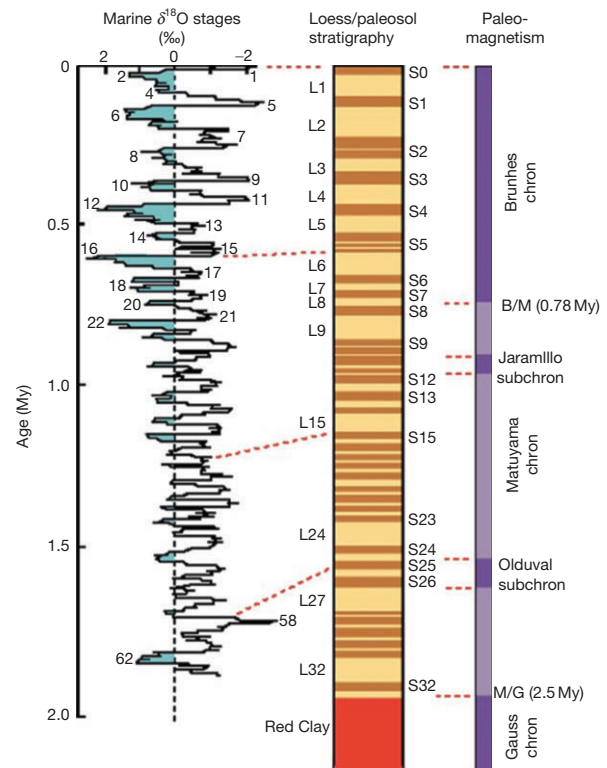


Figure 6 Correlation of the loess-paleosol succession on the Loess Plateau spanning the last 2.5 My with the marine oxygen-isotope record and the paleomagnetic reversal chronology. Reproduced from Porter SC (2001) Chinese loess record of monsoon climate during the last glacial-interglacial cycle. *Earth Science Reviews* 54: 115–128, Figure 1.

Late Quaternary Loess/Paleosol Record

The SUS of loess has been used as a proxy for summer monsoon strength, for it is regarded as being strongly related to seasonal precipitation and temperature. The intensity of SUS at the land surface diminishes northwestward across the Loess Plateau, reflecting a decrease in summer monsoon strength and in related soil-forming activity (enhanced production of maghemite) in that direction (Figure 4). In most loess profiles that span late Quaternary time, SUS closely resembles the standard MIS pattern, which leads to the conclusion that both records reflect a response of climate to changing Earth-orbital (i.e., Milankovitch) conditions. SUS was high during the present (Holocene) interglaciation (MIS 1) and the last interglaciation (MIS 5), but low during coldest parts of the last glaciation (MIS 2 and 4; Figure 11). In other words, low SUS is consistent with cold, dry glacial conditions while high values correlate with warm, moist interglacial conditions. Intermediate values characterize MIS 3.

Unlike the SUS signal, loess particle size displays high-frequency, high-amplitude variations through time (Figure 11). Intervals marked by coarse-grained loess are inferred to reflect climates in which strong winter monsoon conditions are dominant and strong winds were able to transport larger particles and greater amounts of dust. They also likely reflect increased

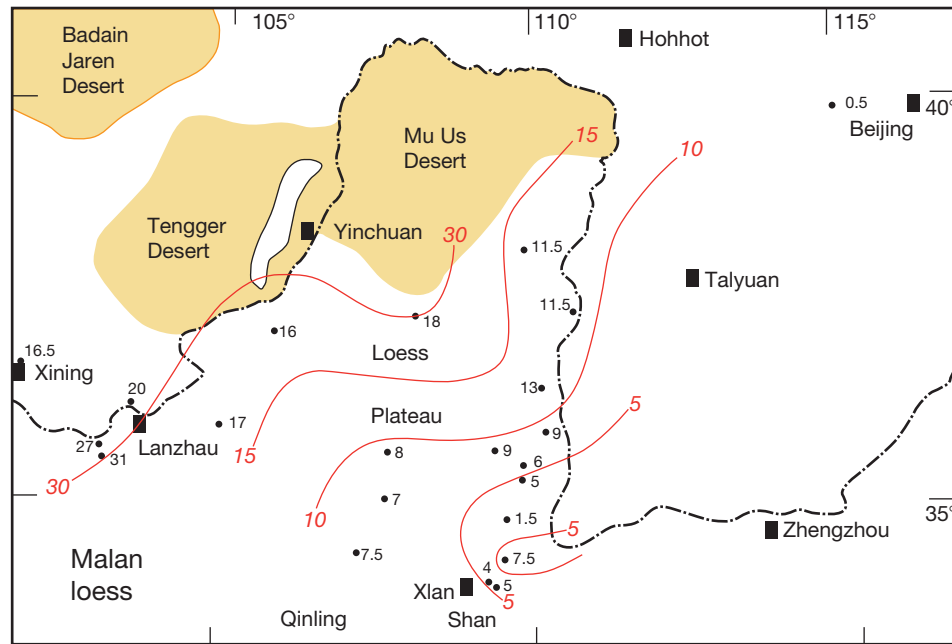


Figure 7 Map of Chinese Loess Plateau and adjacent sandy deserts showing regional variation in thickness (in m) of Malan Loess (L1). A logarithmic decrease in thickness across the plateau suggests that dust-bearing winds flowed southeastward from dust source areas to the northwest. Reproduced from Porter SC (2001) Chinese loess record of monsoon climate during the last glacial–interglacial cycle. *Earth Science Reviews* 54: 115–128, Figure 3.

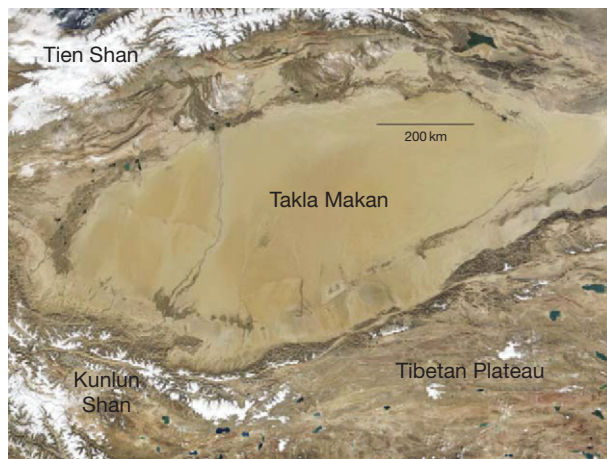


Figure 8 Satellite image of the Takla Makan Desert in the Tarim Basin showing extent of dune sands on desert floor, basin-margin alluvial fans, and glacierized mountain ranges bordering the tectonic depression. Fine-grained meltwater sediments are transported to playas on the desert floor from which they are entrained by winds that carry them to the loess region downwind to the east (NASA image A2001300).

frequency and intensity of dust storms that followed surface storm tracks moving across dust source areas. Despite the high-frequency variability in the grain-size record, if the data are smoothed, the resulting curve rather closely resembles the SUS record. Not surprisingly, Late Pleistocene dust flux on the central Loess Plateau has a pattern that resembles both SUS and the grain-size data (Figure 9).

Glacial–Interglacial Dust/Paleosol Model

A simplified model showing the relationship between changes in summer monsoon climate, dust accumulation, and soil formation illustrates how the loess/paleosol stratigraphy develops (Figure 12). During peak cold, dry phases (stades) of glaciations, the average rate of dust accumulation is high and the rate of soil development is low. Warm, moist interglacial times are characterized by a threefold or greater reduction in dust influx and by intensified soil formation. Interstades, which are warmer and moister intervals within glaciations, show intermediate values between these more extreme conditions. During interglaciations, dust accumulation proceeds at a much slower rate, while during glaciations the intensity of soil development is greatly reduced.

High-Frequency Variations in the Dust Record

The prominent grain-size peaks that appear in loess records at frequent intervals record short-term (millennial-scale) changes of climate that are not detectable in the SUS data. A record of variation in the size of wind-transported quartz grains in the central Loess Plateau (Figure 11) displays prominent peaks at about 67 500, 50 500, 43 500, 33 500, 26 000, 21 500, and 15 000 years ago that must be related to intervals of pronounced strengthening of the winter monsoon. These also were times when large-scale ice rafting occurred in the North Atlantic Ocean related to partial disintegration of bordering continental ice sheets and the generation of exceptionally cold surface ocean waters in this region. Known as Heinrich Events, the ice-rafting episodes are linked to changes in climate in the North Atlantic and in many other parts of the world.

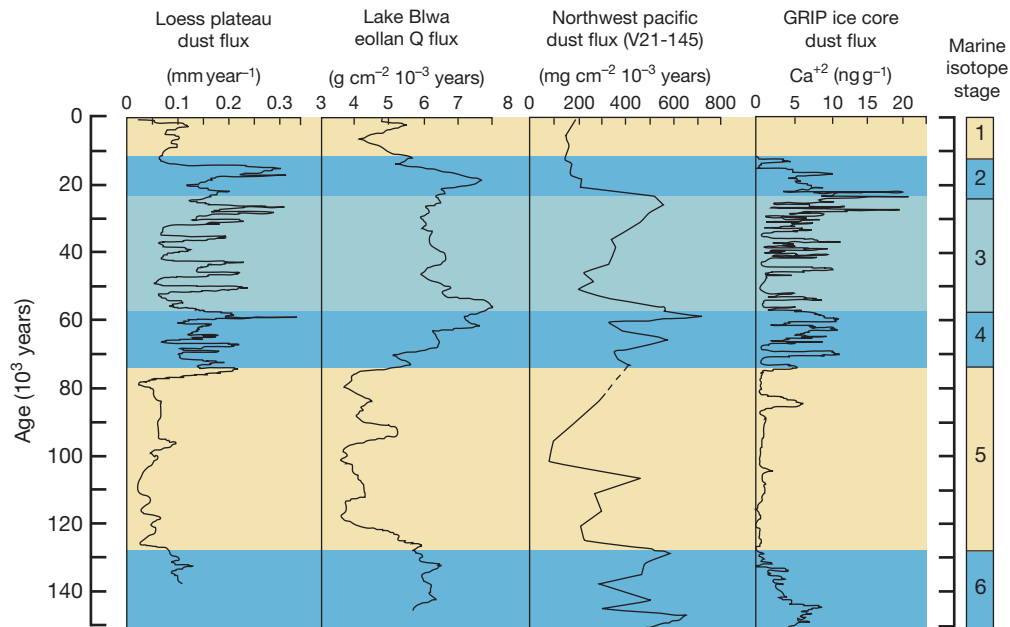


Figure 9 Variations in dust flux during the last ~150,000 years along a transect from the Loess Plateau to Japan (Xiao et al., 1997), the North Pacific Ocean (Hovan et al., 1991), and the Greenland Ice Sheet (Greenland Ice-Core Project Members, 1993). Intervals of greatest and least dust flux are similar among these records, but obvious differences in detail likely reflect differences in the parameters that were measured, the spacing of samples, and chronological uncertainties. Reproduced from Porter SC (2001) Chinese loess record of monsoon climate during the last glacial-interglacial cycle. *Earth Science Reviews* 54: 115–128, Figure 9.

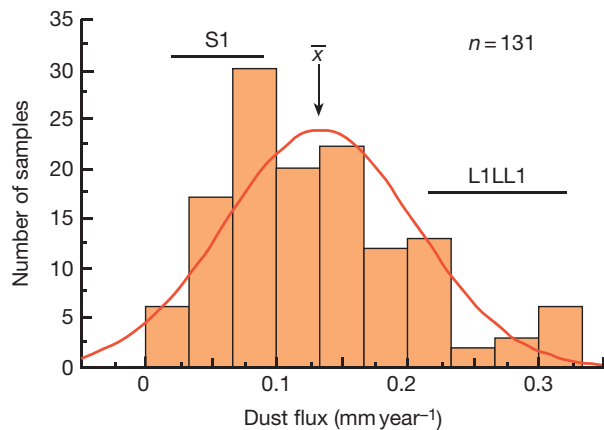


Figure 10 Dust-flux values for the last glacial-interglacial cycle at the Luochuan section in the central Loess Plateau. The mean value is 0.13 ± 0.02 mm year⁻¹. During deposition of loess L1LL1, during full-glacial time (=MIS 2). The rate averaged 0.30–0.35 mm year⁻¹ and during the last interglaciation when the S1 paleosol formed (=MIS 5), the rate fell to <0.10 mm year⁻¹. Reproduced from Porter SC (2001) Chinese loess record of monsoon climate during the last glacial-interglacial cycle. *Earth Science Reviews* 54: 115–128, Figure 6.

The recognition of apparently correlative climatic events in the Chinese loess record in the form of increased dust-fall activity is intriguing, and a common cause of the implied climate linkage deserves further study. A similar high-frequency variation in grain size, although of lower amplitude, has been detected in loess records that span the last interglaciation (MIS 5). Their ages compare closely with surface cooling events detected in North Atlantic deep-sea sediments.

The 'Red Clay' and the Onset of Loess Formation

Throughout the Loess Plateau, the primary Quaternary loess and paleosol section rests on reddish silt and clay. This unit, known informally as the 'Red Clay,' has long attracted attention because of its rich content of Tertiary mammal fossils. Paleomagnetic measurements indicate that it is older than 2.5 My (Figure 13). Recent investigations have shown that the Red Clay consists of alternating layers of reddish eolian silts and carbonate-rich dark-reddish paleosols similar in character to those in the overlying loess/paleosol section. Sedimentary, geochemical, and mineralogical studies confirm that the formation is mainly loess. At the Lingtai section in the middle of the Loess Plateau, the entire eolian section is nearly 300 m thick, with the Red Clay component measuring 125 m. The upper part of the Red Clay is strongly weathered, as indicated by its intense reddish color and the structure of its soils. A cyclic pattern is apparent, with more than 100 discrete couplets of weathered loess and accretionary soil having been identified. The average accumulation rate of the Red Clay section was about half of that of the overlying Quaternary section.

It is now apparent that accumulation of eolian dust on the Loess Plateau was not restricted to the last 2.5 My, as previously thought. At several sites in the middle of the Loess Plateau, the base of the Red Clay dates back 6.5–7.2 My – well into the late Miocene. Paleomagnetically dated eolian sediments in thick sections near Lanzhou extend back 22 My, meaning that the onset of loess accumulation and of the monsoon climate likely dates back to the early Miocene. A progressive increase in summer monsoon strength and sedimentation rate over this long time interval has been attributed to progressive uplift of the Tibetan Plateau and the aridification of central Asia, which

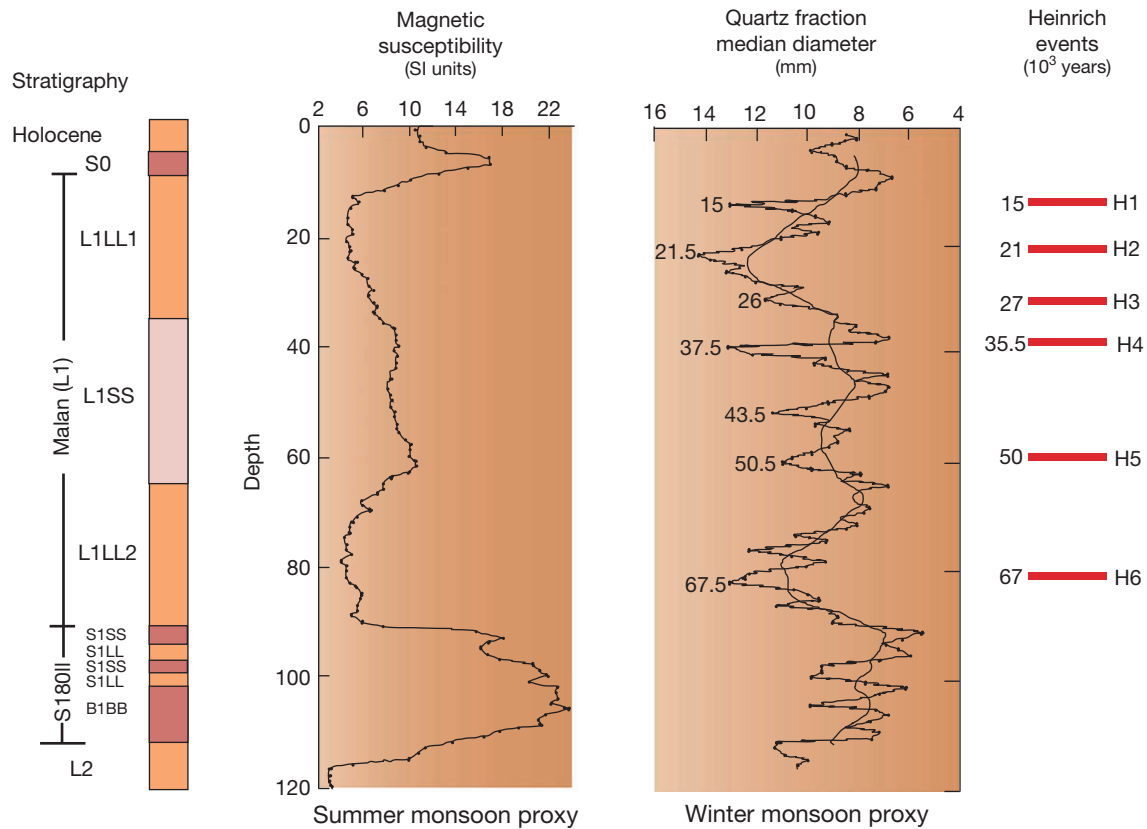


Figure 11 SUS and grain-size profiles through loess and paleosols of the last glacial-interglacial cycle. These parameters serve as climate proxies for summer and winter monsoon variations, respectively. The latter record shows higher variability at the millennial timescale, but the smoothed curve (bold line) bears a general similarity to the curve of SUS. Peaks in grain size occur at times of extensive iceberg discharge into the North Atlantic Ocean (Heinrich events), which apparently had a strong effect on global climate. Reproduced from [Porter SC \(2001\)](#) Chinese loess record of monsoon climate during the last glacial-interglacial cycle. *Earth Science Reviews* 54: 115–128, Figure 7.

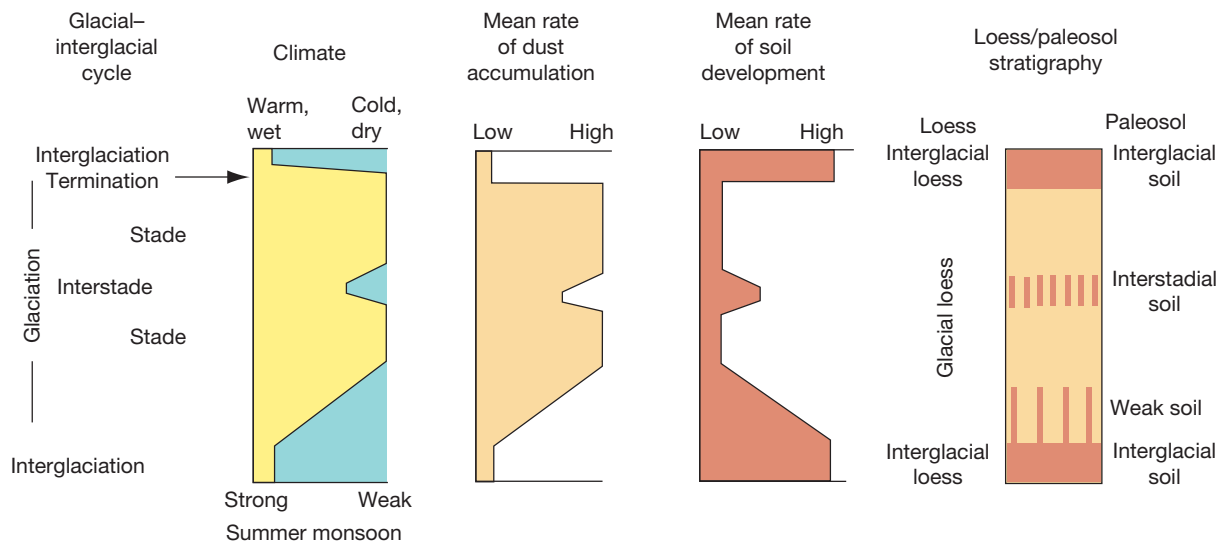


Figure 12 Inferred relationship between changing monsoon climate and the formation of loess and paleosol sequences in the loess region of China. During cold, dry glaciations, dust transport and deposition reached a maximum, while during interglaciations soil formation occurred under warm, moist conditions. Reproduced from [Porter SC \(2001\)](#) Chinese loess record of monsoon climate during the last glacial-interglacial cycle. *Earth Science Reviews* 54: 115–128, Figure 8.

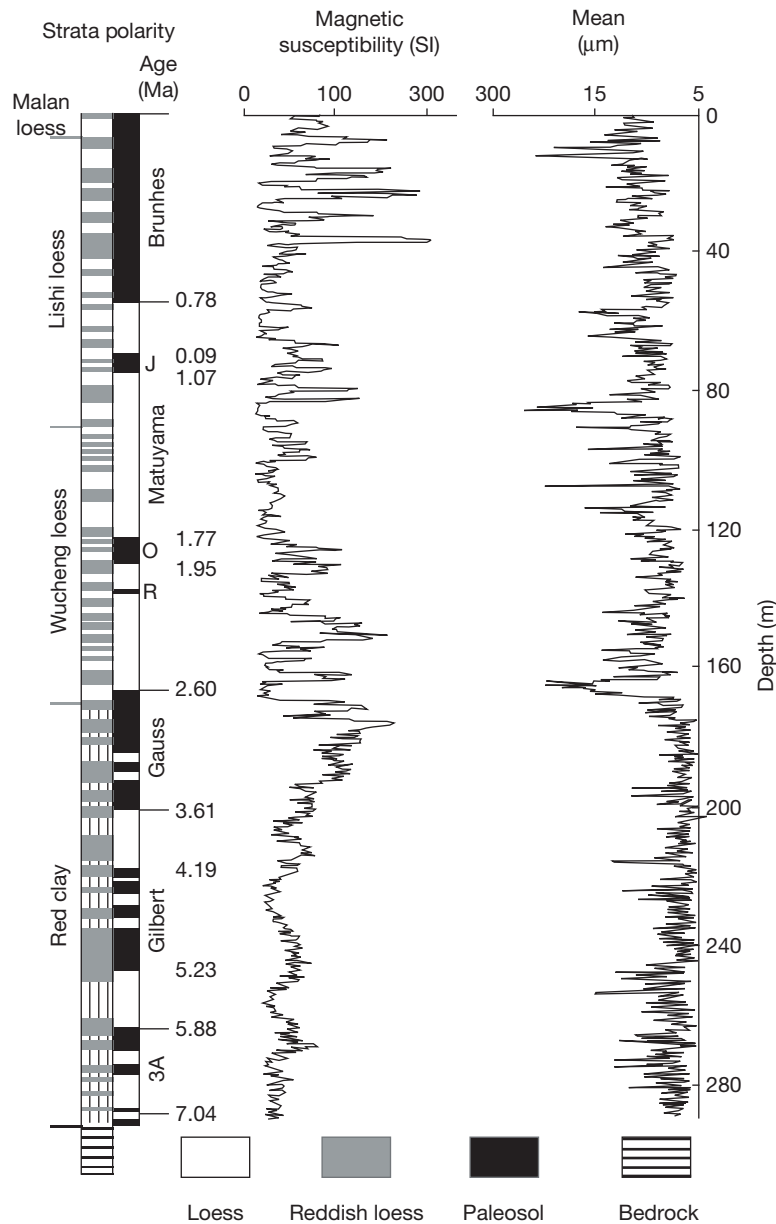


Figure 13 Paleomagnetic polarity, SUS, and grain-size record of the Lingtai site on the Loess Plateau. The 'classic' loess section extends back about 2.7 Ma, but loess deposition and soil formation also characterized the Red Clay, which in this profile extends back more than 7 Ma, through the Pliocene and into the late Miocene. Reproduced from [An ZS \(2000\)](#) The history and variability of the East Asian paleomonsoon climate. *Quaternary Science Reviews* 19: 171–187, Figure 4.

were primary factors leading to the development of the East Asian monsoon system. SUS values show a long-term overall increase between 7.0 and 2.7 Ma.

Erosion of Loess and Loss of Topsoil

China has one of the world's highest rates of soil erosion, which affects 19% of the country's land area. On the Loess Plateau, where the soil loss is particularly devastating, 70% of the land is subjected to erosion. The unconsolidated to semi-consolidated nature of loess makes it highly susceptible to

erosive processes, and vast amounts of loess and soils are eroded from the landscape annually. Gully erosion on steep hillslopes is widespread and constitutes a primary source of fine sediment entering tributary streams. Each year the Yellow River erodes some $16\text{--}18.8 \times 10^8$ tons of loess and topsoil, which imparts a dark yellowish-brown color to the turbid water. The sediment discharge of the Yangtze River exceeds that of the world's two largest rivers, the Nile and Amazon. Between 1949 and 1990, sedimentation along river channels reduced their navigable length by more than 50%. Large quantities of sediment are deposited in reservoirs behind artificial dams along the main river courses and their tributaries. Dam

construction is motivated by the need for flood control and hydroelectric power, but rapid deposition of fine-grained sediment in reservoirs results in progressive reduction in their storage capacity and effective lifespan. The removal of topsoil by fluvial erosion in the loess region is a serious problem because this is prime agricultural land that supports many tens of millions of people.

Environmental Hazards in the Loess Region

A dense human population, intensive agriculture, and natural geologic processes operating in loess-covered terrain can lead to environmental hazards that impact thousands to millions of people.

Landslides

Under normal conditions, loess is stable and can stand as abrupt steep cliffs along valley sides. However, it is very susceptible to landsliding if stability is perturbed. Although the loess is permeable, paleosols are less so. Downward and lateral movement of groundwater along the top of less-permeable paleosols can destabilize the overlying loess and lead to slope failure. The contact between basal Quaternary loess and the less-permeable Red Clay is also a locus of groundwater movement that can lead to massive failure of the overlying loess.

The southern Loess Plateau is susceptible to large earthquakes that can generate extensive slope failures. A most remarkable example is the great Huaxian earthquake of 1556 that occurred along the southern margin of the Wei He graben east of Xi'an. The huge earthquake is estimated to have been of magnitude 8.0 and it generated many landslides. The human casualty toll of this quake was more than 800 000 people. Many of the fatalities likely were caused not only by landslides along steep loess slopes but also by collapse of cave houses excavated in the loess, which are and were numerous in this region (Figure 14).



Figure 14 Cave house excavated in loess. The interior extends back about 10 m, the floor being cut across a reddish paleosol and the walls in the overlying loess unit. Collapse of such dwellings in the loess region likely contributed to the high death toll associated with the great Huaxian earthquake of 1556 (SCP photo 92-441).

Endemic Diseases of the Loess Region

Several important endemic diseases affect the inhabitants of China's loess region. Keshan disease, a frequently fatal impairment of the cardiac muscles, is prevalent in rural villages lying within the humid-to-arid transition zone that crosses the central Loess Plateau. Recent studies conclude that the disease is related to water, soil, and geology. It is largely restricted to hilly regions, rather than to flat yuan terrain or river floodplains. The disease always occurs where strong leaching has removed important nutrients from soils, especially selenium. Adding this element to table salt appears to control this disease.

Kaschin-Beck disease has a similar distribution and causes painful changes in joints and bones. Studies have shown that it, too, is related to selenium deficiency in the diet.

Endemic fluorosis is caused by fluorine excess (poisoning) that leads to aching bones, kidney stones, deformed limbs, and eventual paralysis. Fluorine can reach high concentrations in the loess, meaning that persons living in the loess region can be easily susceptible to this disease. Prevention and successful cures involve extraction of drinking water from deep wells, and water treatment.

Other diseases also apparently are specifically related to widespread surficial loess. For example, Keding disease and goiter are dysfunctions of the thyroid gland related to lack of iodine and are widespread in the loess region where iodine content of groundwater is very low; prevention and treatment involve adding iodine to table salt. Cancer of the esophagus, the third most prevalent cancer in China, is especially serious in the southeastern Loess Plateau. Although the mortality associated with this disease correlates closely with loess particle size, the specific cause of the relationship is still unknown.

See also: [Loess Deposits: Origins and Properties](#). [Loess Records: Central Asia](#). [Paleoclimate Reconstruction: Sub-Milankovitch \(DO/Heinrich\) Events](#). [Paleosols and Wind-Blown Sediments: Mineral Magnetic Analysis; Nature of Paleosols](#).

References

- An Z, Kukla G, Porter SC, and Xiao J (1991) Magnetic susceptibility evidence of monsoon variation on the Loess Plateau of central China during the last 130,000 years. *Quaternary Research* 36: 29–36.
- An ZS (2000) The history and variability of the East Asian paleomonsoon climate. *Quaternary Science Reviews* 19: 171–187.
- An ZS and Porter SC (1997) Millennial-scale climatic oscillations during the last interglaciation in central China. *Geology* 25: 603–606.
- Derbyshire E (2001) Geological hazards in loess terrain, with particular reference to the loess regions of China. *Earth-Science Reviews* 54: 231–260.
- Greenland Ice-Core Project (GRIP) Members (1993) Climate instability during the last interglacial period recorded in the GRIP ice core. *Nature* 364: 203–207.
- Guo ZT, Ruddiman WF, Hao QZ, et al. (2002) Onset of Asian desertification by 22 myr ago inferred from loess deposits in China. *Nature* 416: 159–163.
- Hovan SA, Rea DK, Pisias NG, and Shackleton NJ (1989) A direct link between the China loess and marine $\delta^{18}\text{O}$ records: Aeolian flux to the north Pacific. *Nature* 340: 296–298.
- Kukla G (1987) Loess stratigraphy in central China. *Quaternary Science Reviews* 6: 191–219.
- Liu JG and Diamond J (2005) China's environment in a globalizing world. *Nature* 435: 1179–1186.
- Liu TS (1985) *Loess and the Environment*. Beijing, China: China Ocean Press.
- Martinson DG, Pisias NG, Hays JD, et al. (1987) Age dating and the orbital theory of the ice ages: Development of a high-resolution 0 to 300,000-year chronostratigraphy. *Quaternary Research* 27: 1–29.

- Porter SC (2001) Chinese loess record of monsoon climate during the last glacial–interglacial cycle. *Earth-Science Reviews* 54: 115–128.
- Porter SC and An ZS (1995) Correlation between climate events in the North Atlantic and China during the last glaciation. *Nature* 375: 305–308.
- Porter SC, An ZS, and Zheng HB (1992) Cyclic Quaternary alluviation and terracing in a nonglaciated drainage basin on the north flank of the Qinling Shan, central China. *Quaternary Research* 38: 157–169.
- Porter SC, Hallet B, Wu XH, and An ZS (2001) Dependence of near-surface magnetic susceptibility on dust accumulation rate and precipitation on the Chinese Loess Plateau. *Quaternary Research* 55: 271–283.
- Rutter N, Ding Z, Evans ME, and Wang Y (1991) Magnetostratigraphy of the Baoji loess–paleosol section in the north-central China Loess Plateau. *Quaternary International* 7: 1–22.
- Sun JM (2002) Provenance of loess material and formation of loess deposits on the Chinese Loess Plateau. *Earth and Planetary Science Letters* 203: 845–859.
- Sun JZ (1988) Environmental geology in loess areas of China. *Environmental Geology and Water Science* 12: 49–61.
- Xiao JL, Inouchi Y, Kumai H, et al. (1997) Eolian quartz flux to Lake Biwa, central Japan over the past 145,000 years. *Quaternary Research* 48: 48–57.

Europe

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Introduction

The loess of Europe makes up the western end of the most extensive and voluminous loess belt on Earth that stretches some 10 000 km eastward to China's Pacific coast. Owing to its eolian origin, loess occurs on the landscape as relatively thin drapes, a few meters thick, on mountain foot slopes and on river terraces. Loess from five to several tens of meters in thickness is mainly found in basins, major river valleys, and on plateaus and extensive plains, as shown in the influential map of loess distribution in Europe published by [Grahmann \(1932\)](#). European loess generally lies outside the limits of the last Fennoscandian and Alpine glaciations ([Figure 1](#)), with extensive mantles in southern Russia, northern Ukraine and Belarus in the east, northern France and southern England in the west, and the Po Basin of northern Italy in the south. Although loess is described as an eolian sediment in Chinese texts more than 2 ka old, the process link between dust transport and loess deposits was not widely accepted in Europe until the German scientist [Richtofen](#) published his work on the Chinese loess, which he considered very similar to the loess of eastern Europe ([Pye, 1995; von Richtofen, 1882](#)). For a definition of loess, see [Pye \(1984\)](#).

The discovery and first scientific description of European loess is attributed to Karl Caesar von Leonhard, who, in approximately 1820, noted pale yellow, unstratified sediment, containing snail shells and fossil root channels, in the Neckar River Valley east of Heidelberg, Germany. He called it 'loess,' a local word used to indicate a yellow lime-rich soil ([Smalley et al., 2001](#)). A number of years later, von Leonhard showed the loess outcrop to Charles Lyell who was so impressed that he included substantial text on loess in his 'Principles of Geology' (1833), which certainly led to increasing recognition of loess by the world's geologists.

Loess of Europe: The Material

Mineralogy of European Loess

The most common mineral in most European loess is quartz (~40–80%), the principal ancillary minerals being the feldspars (predominantly K-feldspar), carbonates, and clay minerals. Exceptions include some Carpathian Basin loess, in which phyllosilicates are the dominant mineral group (up to 34%). Heavy minerals, although constituting only a small percentage (<5%) by weight, have proved useful as indicators of provenance and degree of pedogenesis in both loess and paleosol units ([Maruszczak and Wilgat, 1995](#)). While the calcite–dolomite ratio is fairly consistent at about 3:1, carbonate content varies widely within Europe, for example, <11% in

Poland, 12–15% in The Netherlands, 5–20% in east Kent (UK), <12 to >20% in northern France, and up to 25% at some sites in western Germany. Carbonate is present in both clastic and secondary forms, the latter occurring as concretions ('loess dolls'), pore linings, encrustations, and intergranular cements.

The finest fractions (clay grade: <2 µm) of loess and paleosols are rich in clay minerals, and also include varying amounts of lithic and biogenic quartz. Kaolinite and illite are the most common clay minerals, together with chlorite, vermiculite, smectite, and several mixed-layer clays. Clay species vary regionally in response to source-area rock composition, sorting during transport, and postdepositional weathering. Smectite/illite is prominent in the loess of southern Poland (50–80% of the clay grade, the highest values occurring in the older units), with illite (up to 40%) and minor kaolinite (2–5%; [Grabowska-Olszewska, 1988](#)). The smectite group (montmorillonite, nontronite, and beidellite) is also prominent in loess and paleosols in parts of the Ukrainian plains, together with hydromicas, mixed-layer hydromica–montmorillonite minerals, kaolinite and halloysite. Ancillary minerals include chlorite, goethite, calcite, gypsum, and quartz ([Perederij, 2001](#)).

Morphology and Particle Size Distribution of European Loess

Although frequently described as a homogeneous sediment, the bulk properties of loess show important variations with age (i.e., the depth below the surface), location, site, source area, topography, and depositional and weathering history. For example, in regions such as Europe that were subjected to multiple glaciations in the Quaternary, glacial grinding produced abundant 'rock flour' that was deposited by meltwater, and reworked by the wind, as well as by periglacial braided fluvial systems such as the Danube, Dnieper, and Rhine rivers. The mainly mechanical breakage of rock particles into silt-sized (2–63 µm diameter) dust susceptible to deflation yielded particles that are dominantly of tabular and blade shape ([Smalley, 1966](#)). As most of these silt particles are transported in suspension, they lack the edge-rounding and 'frosted' surface texture so characteristic of wind-blown sand grains. Partly rounded particles in loess tend to be nonquartz components, including grains arising from certain secondary (chemical) postdepositional processes. The quartz grains in some of the loess in Normandy, France, and the Channel Islands are subangular in shape, but appear rounded because they are almost completely mantled in a clay coating rich in Si, Al, and Fe.

Sediments described as loess in Europe show a wide range of particle sizes. Leaving aside intraregional variations arising from distance from sources, silt content of loess by weight is generally between 60 and 80%, with <20% clay grade, and

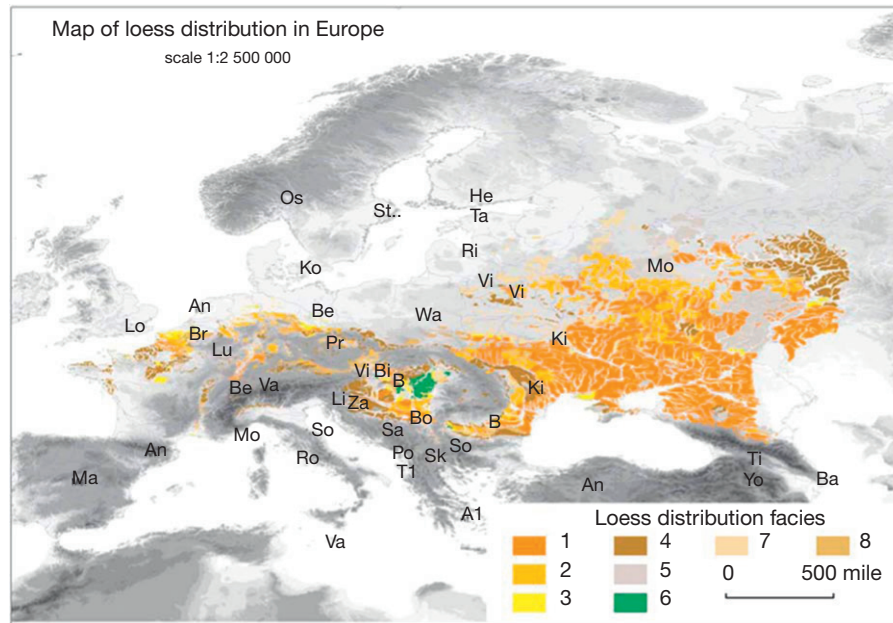


Figure 1 Loess distribution in Europe. Key to symbols: 1, loess >5 m; 2, loess >2 m; 3, sandy loess; 4, loess derivatives; 5, loess and loess derivatives in fragmentary distribution; 6, alluvial loess; 7, eolian sands; 8, loess thickness not differentiated. Reproduced from Haase G, Haase D, Ruske R, Jäger KD, and Altermann M (2006) Loess in Europe – Spatial distribution in a scale 1:2,500,000. *Quaternary Science Reviews* 26: 1301–1312.

sand grades of <15% across a swathe of western European countries. In the loess of the Russian plain and Ukraine, the content of fine silt and clay (<5 μm) in the south is approximately twice than that found further north, the particle size gradients being north–south in the west and northwest–southeast further to the east. Higher clay contents are generally found between the Volga and the Ural Mountains (Rozycki, 1991). A southeast-to-northwest particle size gradient, resulting from the northwestward advection of fine dust from the Caspian–Aral depression, is evident in southeast Europe; this transition zone runs northeastward from Bucharest, and marks the border between the Caspian–Black Sea loess zone and the western European loess supplied from dominantly Atlantic westerly (proglacial) and subsidiary southerly (Saharan) dust sources (Figure 2; Rozycki, 1991). Stratigraphically important volcanic tephra are present on both sides of this transition zone – the Eifel tephra, on the west side, indicating transport by the (south) westerlies, and those from the Caucasus carried by southeasterly winds. Other tephra have been identified in European loess series as, for example, the volcanic ash fallout of the Bag tephra in Hungarian loess during marine isotope stages (MISs) 10 and 8, and the tephra layer identified in the MIS 3 loess in the Paks sequence (Frechen et al., 1997; Horvath, 2001). More recently, in Germany, the loess sequences of the Rhine Valley (Nussloch) have been shown to record the Etlviller tuff (Semmel, 1967) fallout during loess deposition in late MIS 2 (Antoine et al., 2001) contributing greater precision to the discussed age of this ash layer (Juvigné and Wintle, 1988; Zöller et al., 1988).

The loess of western Europe varies with age and geographical location. Loosely packed coarse to medium angular silts, with little clay mineral content and limited, often localized cementation (comparable to much Asian loess), can be found at many sites in continental Europe. Thus, fabric varies with

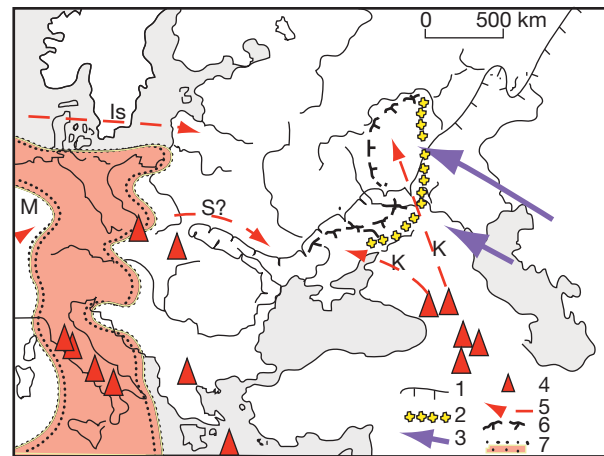


Figure 2 Reach of dust of various origins transported by long-distance eolian transport in eastern and central Europe. Key to symbols: 1, northwesterly extent of western Asiatic dust storms; 2, border between Atlantic and western Asian systems; 3, recorded transport paths of salts; 4, volcanoes active in the Quaternary; 5, volcanic dust pathways (K, north Caucasian; Is, Iceland; S, Sudetic (assumed); M, from Rhine maars); 6, approximate extent of Caucasian volcanic ash; approximate margin of red dust precipitation from the northern Sahara. Reproduced from Rozycki SZ (1991) *Loess and Loess-Like Deposits*, p. 187. Wrocław: Ossolineum-Polish Academy of Sciences.

climatic type and variation through time, not only with changes between soil saturation and desiccation as well as the freeze-drying associated with cryoturbation (Van Vliet-Lanoë and Coutard, 1984), but with a suite of other processes including bioturbation, leaching and redeposition, snow meltwater infiltration, mineral weathering, natural loading and unloading, and reworking by running water and mass movements on

slopes. Preferred fabric trends (anisotropic fabrics) are quite common in European loess. They range from visible lamination (as in alluvially redeposited loessic silts; Derbyshire and Mellors, 1988) to the strongly parallel particle fabrics generated *in situ* by cyclic freezing and thawing. *Limon à doublets* is a distinctive, noncalcareous loess facies found from the Channel Islands off the Normandy coast in the west to the Russian Plain in the east. Its distinctive fabric consists of thin, gently dipping alternating laminae of brown, clay-rich and gray, clay-poor layers, between 1 mm and >1 cm thick (Derbyshire et al., 1988). The 'doublets' features are widely regarded as postsedimentary in origin, having been interpreted as thin layers enriched by pedological clay overprinted on previous grain size discontinuities, the concentration of clay and silt particles on lamellar freeze-thaw features having followed rapid decay of the permafrost at the end of the last glacial.

Calculation of mass accumulation rates (MARs) for loess of the last glacial period (~28–13 ka BP) at over 30 sites across Europe (Frechen et al., 2003) has indicated variable accumulation rates ranging from ~100 to 7000 g m⁻² year⁻¹ along an increasingly continental east–west transect. Making allowances for variation in regional and local silt sources as well as precipitation and wind patterns, which tend to result in the highest individual accumulation rates occurring on terraces of major rivers such as the Rhine, a general pattern emerges in which the lowest mean rates occur on the northwestern periphery of Europe in France and Belgium (~100–600 g m⁻² year⁻¹), with higher accumulation rates evident toward the southeast, in the Czech Republic, Austria, and Hungary (800–3200 g m⁻² year⁻¹).

Loess of Europe: The Origin

Atmospheric dynamics in Europe during the last glacial period were probably very different from those of today because of the presence of extensive and thick ice sheets. Thus, air masses were channeled into a west-to-east trending corridor along the 50° N parallel that broadly corresponds to the main loess-deposition belt in Europe. To the east of the Carpathians, in contrast, elevated terrain gives way to the western lowlands of Ukraine and Tajikistan. Here, reduced topographic constraints on air-flow resulted in a mode of loess deposition quite different from that in the west.

Based on the comparison between the loess zones and mineralogical data (heavy minerals and silts), it is possible to locate the main source of Weichselian (last glacial age) loess in Europe. These studies show that loess in northwestern Europe originated from the North Sea, the central part of the English Channel, and Brittany (eastern English Channel, westward of the Cotentin peninsula). On the other hand, there were also local contributions, linked to the deflation of the main alluvial plains (Seine, Oise, Aisne, Marne, and Somme), which were mixed with particles from the bedrock substratum (frost-shattered chalk and sand). There were also contributions from the dried-out and extended paleoestuaries of the Seine and Somme rivers, and from the wide alluvial braided floodplains of periglacial valleys.

According to Lautridou (1985), eolian deflation during the Weichselian should have prevailed in the large estuaries located 20–30 m below present sea level in the English Channel, which were partially preserved during the maximum lowering of sea level at the Last Glacial Maximum (LGM). In this context, loess sedimentation in northern France should have been supplied by deflation, which prevailed in the paleoestuary of the Somme River, and on the dense paleochannel system of the eastern Channel (Antoine et al., 2003b; Auffret et al., 1982). In addition to these sources of loess, there were more local contributions, originating from the Oise, Marne, Meuse, or Rhine Valleys. In Belgium, study of heavy minerals has shown that the main source of the Weichselian loess was the floor of the North Sea, which, at that time, consisted of sets of braided channels carrying the outwash of the Fennoscandian Ice sheet (Juvigné, 1985). Quite apart from any climatic factors, loess sedimentation in Europe was controlled by sources of available dust. Thus, the main zones of deflation identified in Europe are the dried-out plains (paleoestuaries) of the English Channel and the North Sea where it was exposed by sea-level lowering. Southward, the northern part of the Adriatic Sea at the mouth of the Po River played a similar role. Other sources include the alluvial plains occupied by braided channels during the Pleniglacial phases (times of maximum ice extent, roughly from 70 until 12 ka BP) of the last glacial period. In these fluvial systems typical of periglacial environments, the numerous sandy bars, with sparse vegetation between channels, were probably subjected to intense eolian deflation. The Ukrainian loess, to the south of Kiev, probably originated in such a manner.

The European loess deposits occur as three main morphological types corresponding to the depositional environment and the presence of sedimentary traps.

1. The platform loess, or 'cover loess,' in western Europe, occurs as a mantle of relatively constant thickness. This loess is a homogeneous facies, characterized by considerable spatial continuity that corresponds to the coldest and driest phases of the upper Pleniglacial of the Weichselian (~30–15 ka BP).
2. Slope deposits – more localized, and of variable thickness – are preserved in sedimentary traps. This loess is deposited leeward of asymmetric valleys, features that occur frequently in Europe; loess deposition is influenced by a combination of valley orientation and wind direction. Dust accumulates on the leeward slope, where landform-induced turbulence allows dust to settle and where snow cover and local vegetation act as dust traps. In contrast, windward slopes are zones of deflation (nondeposition). Such phenomena also occur in a variety of local settings, including alluvial terraces and the acute slope angle between marine cliffs and fossil beaches, such as at Sangatte (northern France). The famous sequence at Red Hill near Brno shows the succession of several cycles of loess slope deposits linked to alluvial terraces. Kukla (1977) reviewed several such deposits in Europe.
3. A third dune-like morphology, known as loess Greda, is linked to platform loess. Loess Greda look like elongated dunes several kilometers long; they have been described in central Europe by Léger (1990), and have also been

observed on the right bank of the Rhine Valley near Heidelberg (Antoine et al., 2001). In the latter location, loess accumulation is mostly represented by upper Pleniglacial deposits (35–15 ka), which reach a thickness of 15–20 m; Greda are oriented NNW–ESE, with small, discrete valleys between them.

Paleosols and Their Stratigraphic Significance in the Loess of Europe

The different loess units show a fundamentally cyclic climatic origin (Kukla, 1977). This cyclicity is expressed as an alternating series of loess and paleosols that correspond to global glacial–interglacial climate cycles of 100-ka average duration for the most recent ones (Kukla and Cilek, 1996). Every cycle has a forest soil B-horizon, which is overlain by a steppe (chernozem) soil in central and eastern Europe, and a humiferous forest soil in western Europe. In slope deposits, a light-colored dust layer overlies the black humiferous horizon abruptly; it is overlain by a pellet sand layer, which is capped by loess deposits. In ‘platform’ settings, this sequence is not so apparent; the slope deposits, which have preserved a much more complete record of past environmental changes, are more informative than those in platform settings. Platform deposits contain only direct airfall loess trapped by the local vegetation. Six stages in the development of soil complexes in loess have been summarized by Kukla and Koci (1972) (Figure 3). The recognition of the different soils provides information on the paleoenvironmental conditions, and also provides a very useful tool for section-to-section correlation.

Many European loess studies have shown that abrupt changes in sedimentation are recorded in soil complexes as ‘markers.’ Markers are generally finer grained than normal loess but have no significant differences in composition (Hradilova and Stastny, 1994). An intriguing problem raised by markers is that of identification of the dust source which, given its fine grain size, may not have been close by (Rousseau et al., 1998a). Markers may represent long-distance eolian

transport because of their characteristic sharp contacts and much finer grain size. Thus, dust storms of continental magnitude seem to offer a possible explanation for the deposition of markers. Several studies report major dust events in historical time. Kukla (1977) reported that, on 5 April 1960, a storm deposited 3 cm of dust in Romania, carried from the Kalmyk steppe in Central Asia, for example, a distance of more than 2000 km. Several dust storms have also transported red dust from the Sahara to Europe during the past 20–30 years.

While loess sequences of the last climatic cycle are the best preserved in Europe, some loess–paleosol sequences show older cycles. In northwestern Europe, the St. Pierre-lès-Elbeuf sequence in Normandy shows four cycles (Lautridou, 1974), overlying a tufa with a mollusk fauna of probable MIS 11 (~400 ka) age (Rousseau et al., 1992). The Somme Valley shows overlapping loess and paleosol sequences that overlie the different river terraces (Antoine, 1994). The oldest (sandy) loess, located on top of the terrace dated at about 1 Ma, was deposited at the end of the Lower Pleistocene before the B–M magnetic reversal (Antoine et al., 2000, 2003a). The St. Vallier loess, near Lyon, is among the oldest in Europe, having been dated to 2.5–1.8 Ma by means of a tephra horizon of the Mont Dore volcanic system (Pastre et al., 1996). In the Rhine Valley, the Achenheim loess includes five loess–paleosol couplets, rich in terrestrial mollusks (Lautridou et al., 1986; Sommé et al., 1986). This sequence contains a yellow loess (the ‘canary loess’) that corresponds to MIS 12 (Rousseau, 1987), and indicates particularly cold conditions. Finally, in central Europe, the Krems and Red Hill sequences (Czech Republic) are famous for the long climatic history they preserve, stretching back to the Brunhes–Matuyama boundary at about 0.75 Ma (Kukla, 1977). The loess sequence at Starnzendorf (Austria) records the Gauss–Matuyama paleomagnetic boundary at about 2.5 Ma (Kukla et al., 1990).

Dating Loess in Europe: Geochronology

Most European last glacial loess chronology is based on luminescence dating methods, which include thermoluminescence (TL), optically stimulated luminescence (OSL), and infrared stimulated luminescence (IRSL) (Lang et al., 2003; Wintle et al., 1984; Zöller and Wagner, 1990). As the electron traps involved in these different solid-state physics processes appear to behave almost independently, several age determinations can be obtained from the same sample.

In Europe, the most common materials used for ^{14}C dating are charcoal and wood. These materials are uncommon in loess, and are rarely distributed in an order that provides a continuous, high-resolution chronology. A chronological framework was developed for the Nussloch loess sequence in Germany, based on AMS ^{14}C dating of loess organic matter (Hatté et al., 2001a). The protocol used in this study is adapted to the particular characteristics of the Nussloch sequence (low organic carbon content, high carbonate content, and iron under +2 oxidation state; Hatté et al., 2001b). The resulting radiocarbon chronology is in excellent agreement with OSL ages, although the two chronological methods do not date identical events. Indeed, since luminescence techniques measure the time elapsed since the last sunlight bleaching event,

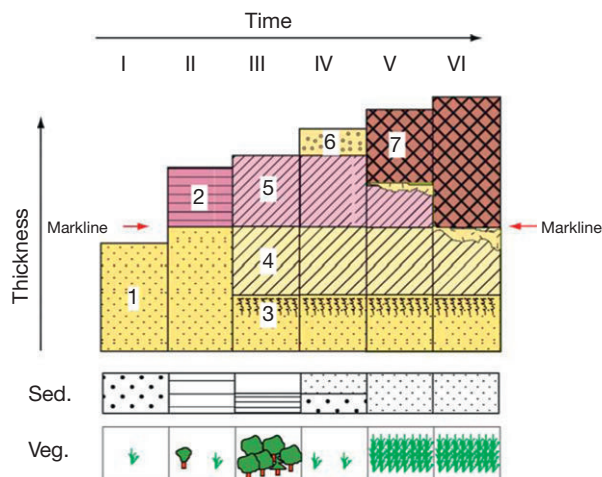


Figure 3 Schematic development of soil complex in loess areas. Reproduced from Kukla G and Koci A (1972) End of the Last Interglacial in the loess record. *Quaternary Research* 2: 374–383.

and thus characterize pulses of dust, ^{14}C on loess organic matter determines the time elapsed since the death of the plant that grew and died on a loess surface before being covered by a new dust pulse. There is no notion of pulse for ^{14}C chronology since vegetation is always present. The general difference between luminescence and ^{14}C chronologies can be summarized by saying that the first characterizes a temporal framework in steps whereas the second smooths and somewhat leads the first one (Figure 4).

Variability of Loess Sedimentation within a Single Glacial Period (Last Climatic Cycle) in Western Europe

Stratigraphic, paleopedologic, mollusk, and palynologic data, coupled with sedimentology, magnetic susceptibility, and TL/IRSL ages, provide a new picture of the last climatic cycle in northwestern Europe, and its connection with neighboring regions (Figure 5). The last interglacial period (including MIS 5e and part of MIS 5d; Kukla et al., 1997) in Europe is

called the Eemian. After the truncation of the Bt horizon that formed in the Eemian paleosol (the Rocourt soil), the early Weichselian is represented by a complex of humiferous paleosols, known as the St. Sauflieu soils (Antoine et al., 1994). This complex is characterized by the superimposition of a gray forest soil, locally doubled (Bettencourt-St. Ouen, Villiers Adam) and two or three steppe soils. The lower part, with its gray forest soils, correlates with the Brörup/Rederstall/Odderade succession (MIS 5d/5a; Figure 6). This sequence indicates a first continentalization of the environment, with development of gray forest soils in loess-derived colluvium, under a boreal forest of pine and beech (Munaut in Antoine et al., 1994, 1999). Above an erosional phase with evidence of deep seasonal frost (the end of MIS 5a), the upper part of the complex is characterized by soils that formed under some birch in a steppe environment with grass and aster family plants. This part of the pedocomplex probably represents the rapid climatic oscillations that prevailed during the transition between MIS 5 and 4 (interstadials 20 and 19 of the GRIP ice-core record; Dansgaard et al., 1993). The whole paleosol

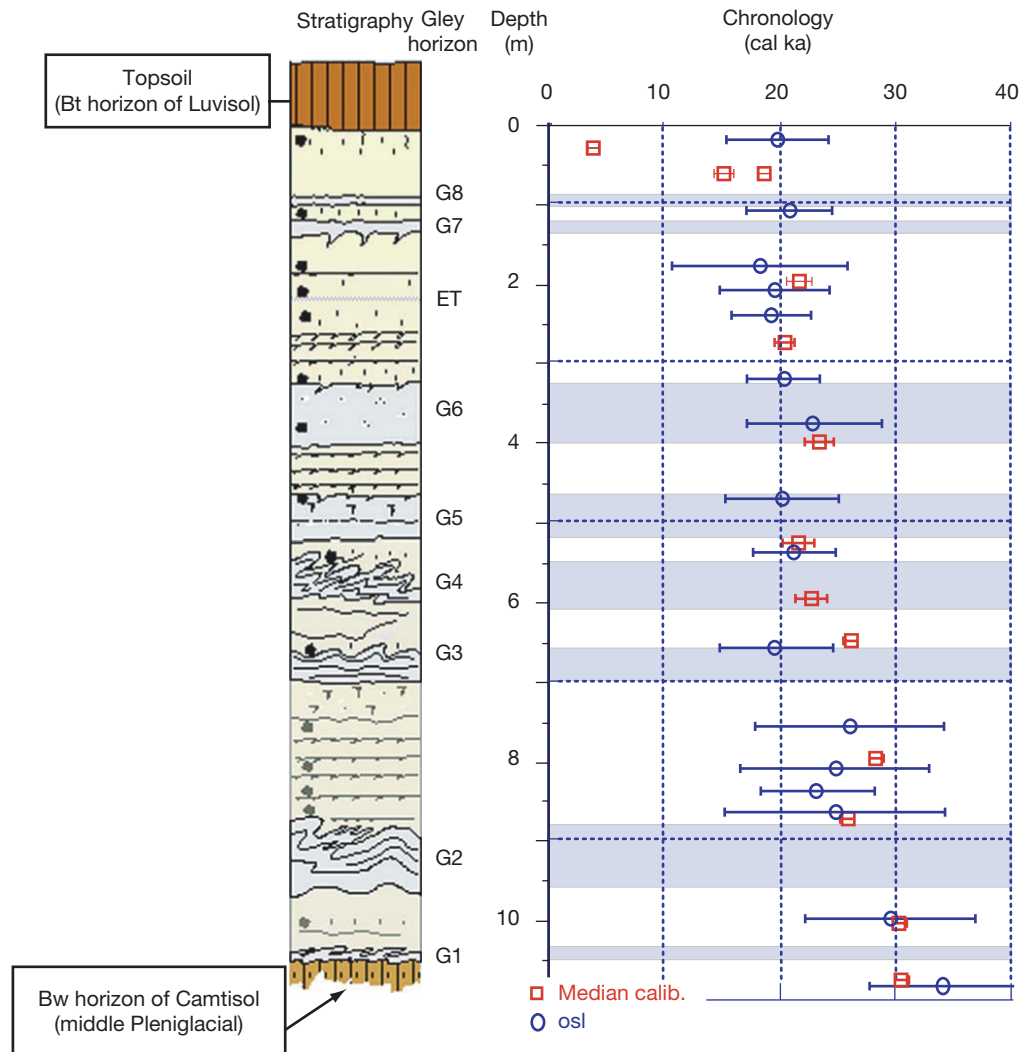


Figure 4 Comparison between the OSL and ^{14}C dates in the upper Pleniglacial sequence of Nussloch. Key to symbols: G, gley horizon; ET, Eltville Tuff. Data from Hatté et al. (2001a) and Lang et al. (2003).

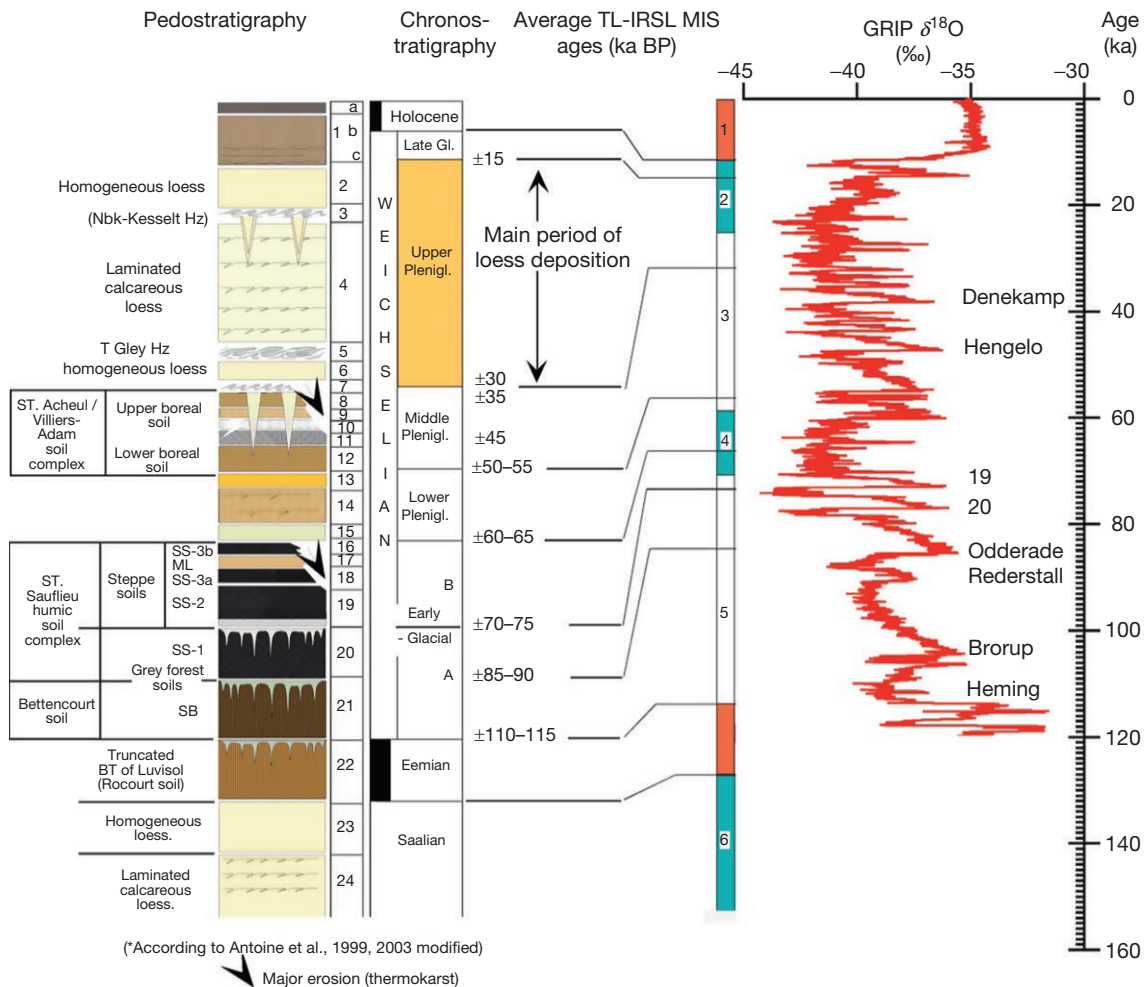


Figure 5 Pedostratigraphy of the upper Pleistocene in western Europe. Reproduced from Antoine P, Rousseau DD, Lautridou JP, and Hatté C (1999) Last interglacial–glacial climatic cycle in loess–paleosol successions of north-western France. *Boreas* 28: 551–563.

sequence shows an increasingly continental environment in two main phases, contemporaneous with sea-level lowering (Sommé et al., 1994). Thus, the paleogeographic change in the North Sea–Channel region contributed to the disappearance of the oceanic influence at the end of the last interglacial.

The upper boundary of the early glacial is defined by the erosive contact at the top of the last steppe soil (Antoine et al., 1994). After the deposition of the first loess (the lower Pleniglacial), an extensive but short erosive episode is marked by laminated colluvium with soil fragments, cryoturbation, and frost cracks, which indicate frost reworking of the underlying levels. This unit, up to 2 m thick in northern France, is a marker level in the oldest part of the Weichselian. A similar record can be traced all the way to the Rhine Valley (Antoine et al., 2001; Haesaerts et al., 2005).

The lower Pleniglacial loess is locally covered by younger loess, often heterogeneous and containing granules. Above this loess, a soil complex is found (Complex of St. Acheul–Villiers Adam) that corresponds to most of the middle Pleniglacial (MIS 3, 50–30 ka). During this period, loess sedimentation diminished and was interrupted by several phases of pedogenesis. These phases produced brown boreal soil to arctic brown soils.

In most of the profiles in the Somme and Normandy, this period is represented by a unique polygenetic horizon (St. Acheul).

Elsewhere, in the sequences found in Villiers Adam, the loess is thicker (4 m) and contains four paleosols: a leached boreal soil, a humiferous arctic meadow soil (Van Vliet-Lanoë, 1987), a tundra gley, and an Arctic brown soil.

The lowermost part of the upper Pleniglacial loess marks the end of pedogenesis; it contains frost wedges and evidence of thermokarst (Antoine et al., 2001). Following that, the main body of the upper Pleniglacial loess was deposited between ~25 and 15 ka. Typically, these deposits are about 4–5 m thick, but may locally reach 6–8 m in thickness. The upper Pleniglacial loess contains as many as three units, separated by periglacial paleosol horizons (Antoine, 1991). The most common unit is the Nagelbeek/‘Kesselt’ tongue horizon, dated to about 22 ka ^{14}C year (Haesaerts et al., 1981). The modern soil is developed in the uppermost upper Pleniglacial loess. A high-resolution investigation in some western European loess has documented climate variability through different indices (biological, sedimentological, geochemical, and geophysical). The results show that during the last climatic cycle, the main loess deposition interval started at ~70 ka



Figure 6 Paleosol complex of St. Sauflieu including the last interglacial paleosol and the humiferous soils of the early Pleniglacial. Photo by Pierre Antoine.

and ended at $\sim 16\text{--}15$ ka (Rousseau et al., 1998b). Two main phases of loess deposition, centered around 60 ± 5 and 23 ± 8 ka BP, are separated by a period with much lower sedimentation rates of between ± 55 and 35 ka (Antoine et al., 1999, 2001).

A similar sequence, with local variations due to the more continental conditions inferred from their geographical location, is available for central Europe (Kukla, 1977; Figure 7). A brown forest Bt soil at the base corresponds to the last interglacial. It is overlain by a biogenic steppe soil of chernozem type, interrupted by a Marker I, 2–10-cm thick. This is a sharply delimited band of light-colored dust. It separates the underlying chernozem from the overlying hillwash loam composed of sand-sized fragments known as pellet sands. Fine loess caps this first pedocomplex (PKIII). This first eolian deposit is succeeded by a second pedocomplex (PKII) with a pseudogley overlain by a chernozem interrupted in some rare places by a new Marker horizon (IIa), immediately followed by small pellet sands capped by another loess. This succession is repeated a second time with a thicker chernozem interrupted by Marker II, overlain, in turn, by thick pellet sands and a loess with an age placing it at the base of MIS 4. A third soil complex developed after this lower Pleniglacial loess; it consists of a brown decalcified soil overlain by a thin loess, then a humiferous chernozem. This is pedocomplex PKI. Finally, the upper Pleniglacial displays the thickest loess deposits, with intercalated gley horizons, as observed in western Europe. A similar pattern, with regional differentiation, has also been described in the Ukrainian loess deposits near Lubny (Rousseau et al.,

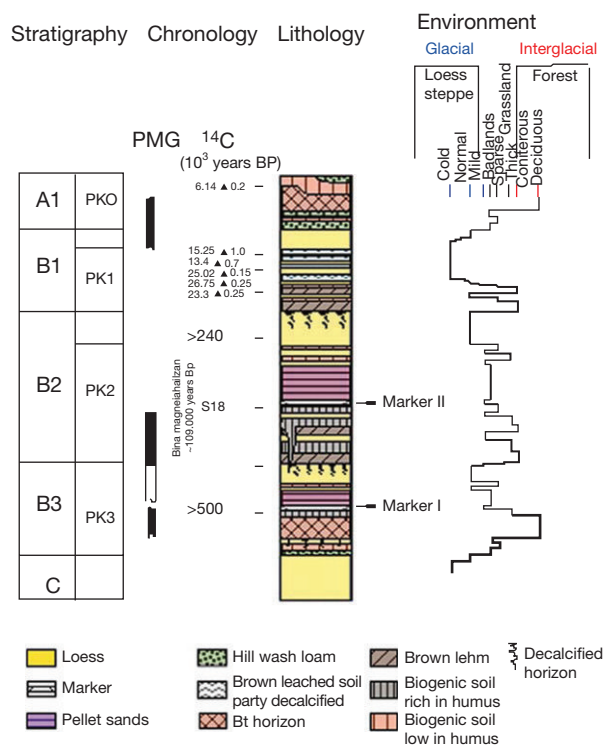


Figure 7 Pedomatigraphy of eastern Europe. Reproduced from Kukla G (1977) Pleistocene land-sea correlations. 1: Europe. *Earth-Science Reviews* 13: 307–374; Rousseau DD (2001) Loess biostratigraphy: New advances and approaches in mollusk studies. *Earth-Science Reviews* 54(1–3): 157–171.

2001). At Lubny, the loesses and paleosols correlate closely with those of central (Kukla, 1977) and western Europe. Thus, there is a remarkable consistency in the history of the last climatic cycle as recorded in the loess sequences of western, central, and eastern Europe over a distance of 2500 km, as shown by Haesaerts et al. (2003, 2005). The differences observed relate to local conditions arising from the proximity of the Fennoscandian ice sheet and Alpine glaciers, or the influence of the westerlies and climatic variations over the North Atlantic.

Paleoclimatic Proxies in European Loess

The past few decades have seen the development of new paleoclimatic proxies, allowing a more precise interpretation of the European loess sequences. These include loess and paleosol geochemistry, identification of periglacial features, molluscan paleozoogeography, magnetic susceptibility, and detailed sedimentology.

Geochemistry

Few organic geochemistry investigations are available for European loess, but several recent studies of loess have been completed in France (Hatté, 2000; Hatté et al., 1998) and Germany (Hatté and Guiot, 2005; Hatté et al., 1998, 1999, 2001a; Pustovoytov and Terhorst, 2004). Other investigations are in progress in eastern Europe. Organic geochemistry studies

are based on the 'fingerprint' of environmental conditions provided by plant $\delta^{13}\text{C}$, and on its undisturbed conservation during burial and subsequent sedimentation. During photosynthesis, plants discriminate against ^{13}C because of differences in chemical and physical properties due to its greater mass (O'Leary, 1981). Both major types of photosynthetic pathways have a characteristic isotopic signature. C4 plants living in rather severe climatic conditions, with high insolation and/or water stress, show a mean $\delta^{13}\text{C}$ of $-13 \pm 2\text{‰}$, whereas C3 plants, which prefer more temperate environments, present $\delta^{13}\text{C}$ values around $-26 \pm 4\text{‰}$ (O'Leary, 1988). Variability around the mean $\delta^{13}\text{C}$ values in leaves of terrestrial vegetation (foliar $\delta^{13}\text{C}$) results from environmental changes that influence stomatal conductance (e.g., Feng and Epstein, 1995). These results show that variation of the $\delta^{13}\text{C}$ in C3 plants within the range -30 to -22‰ is primarily influenced by the $\delta^{13}\text{C}$, the concentration of atmospheric CO_2 , and by precipitation, and, second, by soil type and texture and insolation. Temperature influence differs from one biome (association of plants) to another, but remains the most important parameter in the definition of the biome itself. On the other hand, variations of isotopic signature in C4 plants within the -15 to -11‰ interval must be almost exclusively linked to variations in the $\delta^{13}\text{C}$ in atmospheric CO_2 . All these metabolic responses to environmental changes indicate that carbon isotopic composition of plants reflects climatic variations.

In contrast to interglacial soils, typical loess is associated with sparse vegetation and a weak rhizosphere. The absence of well-established pedogenesis and the dry glacial environment favor the degradation of organic matter without distortion of the isotopic signal, making typical loess suitable for an organic geochemical study. When properly prepared (Hatté and Gauthier, 2006), the carbon isotopic composition of loess organic matter is a powerful paleoclimatic indicator, because it inherits the $\delta^{13}\text{C}$ of growing plants that trap dust at the time of deposition. As environmental conditions and vegetation types (C3 vs. C4 photosynthetic pathways) control the $\delta^{13}\text{C}$ levels in plants, the $\delta^{13}\text{C}$ values of organic matter in loess can be used to infer temporal variations in climate and vegetation. Thus, the isotopic signal cannot be interpreted solely in terms of change in the ratio of C3 to C4 plants. Indeed, considering only the C3 photosynthetic pathway, $\delta^{13}\text{C}$ variations can be linked to first order changes in atmospheric CO_2 ($\delta^{13}\text{C}$ and concentration) and precipitation and, at the second order, to temperature, soil type and texture, and insolation.

In the Rhine Valley (Achenheim, France, and Nussloch, Germany), Hatté et al. (1998) demonstrated, with values ranging from -23 to -26‰ , C3 origin of organic matter during the last glacial period (70–12 ka BP), whereas Pustovoytov and Terhorst (2004) exhibited some C4 carbon-enriched layers ($\delta^{13}\text{C}$ from -16 to -19‰) within the same period in Schattenhausen, <1 km to the east of Nussloch. This conflict of view remains to be resolved, but two interpretations can be proposed. One concerns complications arising from carbonates in the loess. Another possibility is that there existed a mosaic of vegetation types, such as a mixture of C4 grasses and C3 trees.

Considering the C3 photosynthetic pathway only, Hatté et al. (1998, 1999) interpreted the variations in the $\delta^{13}\text{C}$ of loess organic matter in Nussloch (Germany, Rhine Valley) and

Achenheim (France, Rhine Valley) as a response to changes in paleoprecipitation during the last glaciation. Using a linear relation between loess $\delta^{13}\text{C}$ and atmospheric CO_2 concentration and $\delta^{13}\text{C}$, on the one hand, and precipitation on the other, Hatté et al. (2001b) attempted a deconvolution of the loess $\delta^{13}\text{C}$ record to reconstruct paleoprecipitation. However, paleoclimatic inferences were limited because only parameters of the first order were taken into account. Nevertheless, use of a vegetation model (Biome4) provides the required greater complexity by considering first- and second-order parameters. Inverse modeling of loess $\delta^{13}\text{C}$ in Nussloch provided reconstructed paleoprecipitation values that varied between 240 and 400 mm year $^{-1}$ through the last glaciation (Hatté and Guiot, 2005). This clearly demonstrated atmospheric teleconnections with the Greenland ice sheet extension, by matching Dansgaard-Oeschger (D/O) events with a precipitation increase of ~ 100 –200 mm year $^{-1}$ (Hatté and Guiot, 2005; Figure 8).

Cryoturbation and Evidence of Ice Wedges

Cryoturbation features and ice-wedge casts occur at different stratigraphic levels in the northwestern European loess sequences, and represent marker horizons that allow correlation of sections (Lautridou and Sommé, 1974). Following the studies of Pissart (1987) and Van Vliet-Lanoë (1987), cryoturbation features are interpreted as the result of differential expansion of different surface materials in response to freezing. Indeed, in contrasting materials (e.g., loam vs. sand), the experiments performed in Caen indicate clearly the establishment of structures with a drop or pear shape when the pressure generated by this process is blocked at the surface by refreezing. In poorly drained environments with a surface water sheet, the cryogenic expansion, blocked by surface freezing, causes downward deformation. Conversely, in a well-drained environment, the cryogenic expansion can exert a force toward the surface, and so produce certain relief forms (tundra ostioles, soils with ice mounds).

Ice-wedge casts are produced from cracks caused by thermal contraction. They form progressively in a series of events, as follows: (1) cracking of the permafrost by thermal contraction in winter and (2) infilling of cracks with meltwater produced by thawing of snow in spring, followed by refreezing (French, 1996; Péwé, 1962). The development of these structures indicates the occurrence of permafrost, with a mean atmospheric temperature lower than -8°C , and lack of a thick snow cover. After the degradation of the ice wedge, the structure is often fossilized by loess sedimentation. The ice-wedge casts can be as large as 1 m wide and 2 m deep in the loess of northern France and Belgium (Figure 9). In plan view, ice-wedge casts make up a network of polygons ± 10 –12 m wide. Their presence in the loess indicates significantly colder temperatures during the last glacial period.

Terrestrial Mollusks and Paleozoogeography

Terrestrial mollusk (gastropod) assemblages are one of the most powerful paleoclimate proxies in carbonate loess sequences (Figure 10). Quaternary mollusk species are extant and do not show any changes in their ecological requirements.

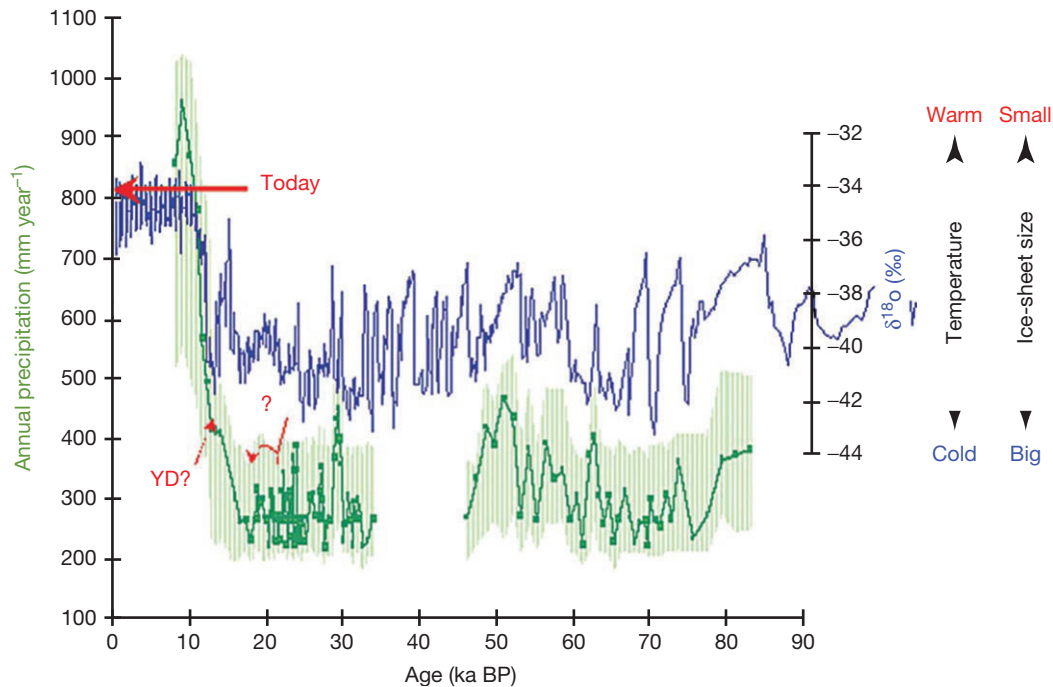


Figure 8 Reconstruction of annual precipitation in the Nussloch sequence. Reproduced from Hatté C and Guiot J (2005) Paleoprecipitation reconstruction by inverse modelling using the isotopic signal of loess organic matter: Application to the Nussloch loess sequence (Rhine Valley, Germany). *Climate Dynamics* 25: 315–327.

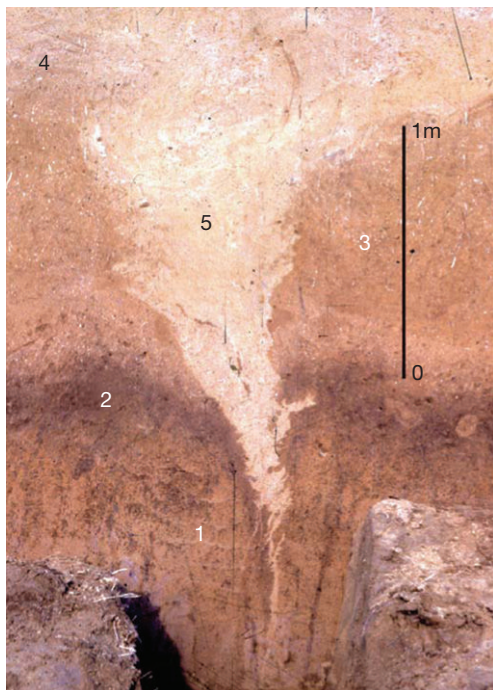


Figure 9 Ice wedge in the loess sequence at Sourdon. Photo by Pierre Antoine.

As the identification of the species is performed by considering the shell shape and ornamentation, it is then possible to use the present ecological requirements and zoogeography of living individuals to interpret the fossil assemblages. Variations

in the specific composition of gastropods can be used to characterize past environments, and to reconstruct ecological and climatic parameters. Mollusk communities, including forest species, mainly represent temperate assemblages regardless of the region under consideration. Interstadial assemblages include species indicating cool and open conditions, whereas two main assemblages represent cold environments (Lozek, 1990; Rousseau, 1987, 2001; Rousseau and Puisségur, 1990). One group best characterizes steppe environments and is dominated by four species of *Pupilla*. The coldest and wettest conditions of a tundra-like environment are represented by the *Columella* group, which includes more species than the *Pupilla* group. While climatic variations can be interpreted from multivariate analysis of the mollusk assemblages, the compilation of the assemblages over a large area also yields other important information. Indeed, the impoverishment of the mollusk assemblages in the western European Upper Pleistocene loess sequence has been interpreted as a response to the coldest and wettest condition that prevailed in this area. In contrast, conditions notably less maritime (Atlantic) characterized central Europe during the last glacial period, where the assemblages were both more abundant and more diverse (Rousseau, 2001; Rousseau et al., 1990). Transfer functions of terrestrial mollusks, following the modern analog principle, have been developed for European loess sequences, which show similar temperature reconstructions to those computed from pollen counts (Rousseau, 1991). Other methods have also been developed using the present distribution of the species, as well as the 'climate mutual range' method originally developed for beetles (Moine et al., 2002). Finally, studies of Hungarian mollusk assemblages have used the ecological ranges of the observed species, applying a

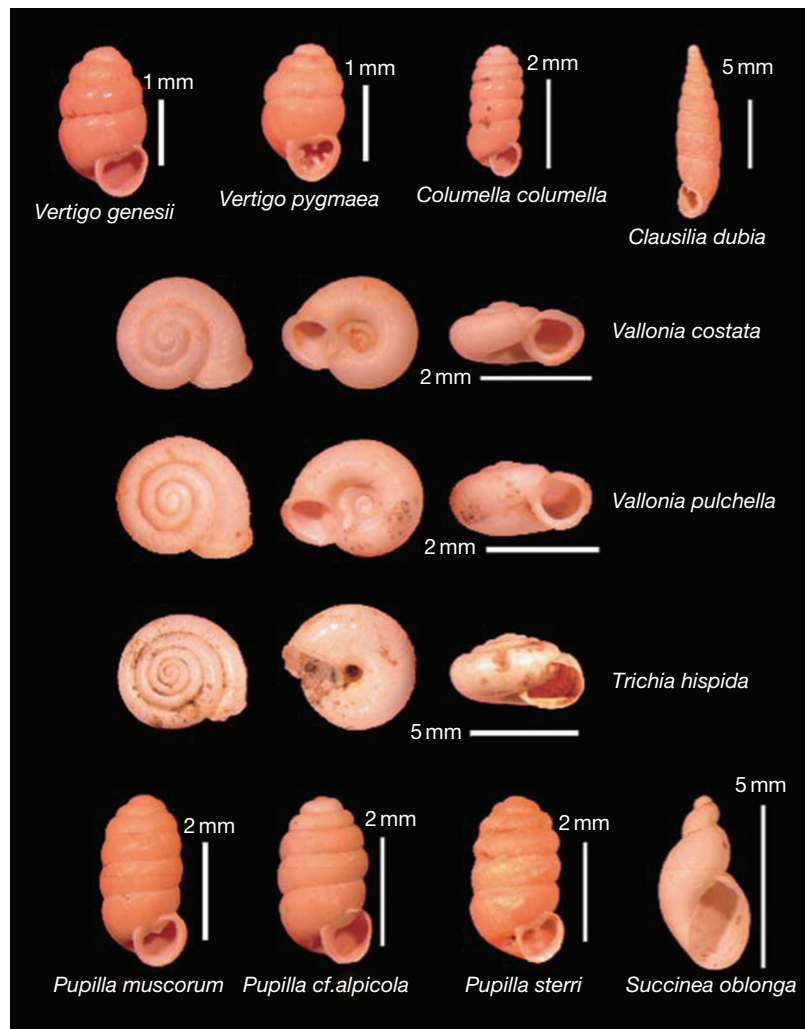


Figure 10 Main terrestrial mollusk species described from loess deposits. Photos by O. Moine.

method similar to the one developed for micromammal studies by Okhr (Sümegei and Hertelendi, 1998). Terrestrial mollusks have also been studied for their amino acid signatures, which show significant differences from one climatic cycle to another (Oches and McCoy, 1995).

Sedimentology and Grain Size

The particle size distribution of loess is often used to determine variations in the composition of the sediment. However, variation in the different grain size classes can also be used to determine the relative wind velocity at the time of particle transport. Comparison of the different grain size classes, from silt to coarse sands, is widely used in loess studies. Shi et al. (2003) interpreted an increase in the coarser grain size fractions at Dolni Vestonice (Czech Republic) as corresponding to the record of the major iceberg discharges in the North Atlantic, called Heinrich Events. However, another approach is to define a grain size index, which corresponds to a ratio of two main particle size classes. Higher values indicate stronger winds. Two main indices have been developed. Studying the loess sequence at Kesselt in Belgium, Vandenberghe et al.

(1998) defined the 'U ratio' as the ratio between the two size classes 16–44 and 5.5–16 μm . Similarly, in studying other European loess sequences, Antoine et al. (2002) developed the IGR index ratio defined as the ratio between coarse loam (20–50 μm) and fine loam and clay (<20 μm), which is similar to the 'U ratio.' The IGR index can be used not only to reconstruct the wind dynamics but also as a new tool for the detailed correlation of different sequences within a limited area or along a transect. Results of studies using this method indicate that similar patterns can be traced from west to east right across Europe from northern France to Ukraine.

European Loess as a Record of the Response of the Continental Environments to North Atlantic Climatic Variability

Loess sedimentation in Europe appears to be rhythmic. Compared to Chinese loess, the European loess sequences (and especially those in western Europe) show a more discontinuous and contrasting record. These characteristics are linked to the influence of the North Atlantic Ocean that gives rise to

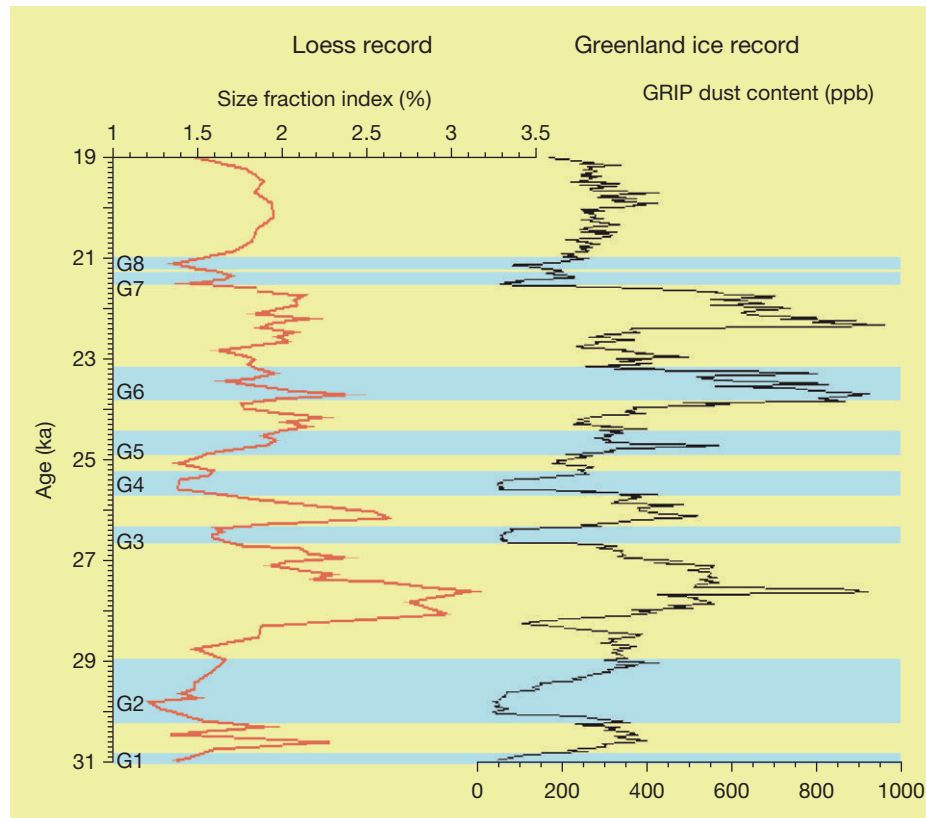


Figure 11 Comparison of the IGR grain size index at Nussloch and the GRIP dust record during the 35–15-ka interval. Reproduced from Rousseau DD, et al. (2002) Abrupt millennial climatic changes from Nussloch (Germany) Upper Weichselian eolian records during the Last Glaciation. *Quaternary Science Reviews* 21(14–15): 1577–1582.

more humid environments (increase in soil development, and periglacial structures) and to high-frequency (millennial) variations in loess depositional rates.

According to new studies, the loess of Europe records rapid climatic events similar to those described in the Greenland GISP2 and GRIP ice cores (D/O), and in marine cores from the North Atlantic (Bond cycles). The latest results (Hatté et al., 2001a; Lang et al., 2003) obtained in studies of the Nussloch loess sequence (Germany) indicate good correspondence between European loess and North Atlantic climate variability (Antoine et al., 2001, 2002; Rousseau et al., 2002; Figures 11 and 12).

The cornerstone hypothesis of the study of the Nussloch sequence in the Rhine Valley is that loess sedimentation, if it has global significance, should be coherent with the dust record preserved in Greenland ice. High-resolution analyses of the Nussloch loess sequence can best be compared to the Greenland dust record because both provide a record of Northern Hemisphere eolian dynamics. Both GRIP and GISP2 dust records show alternating phases of high or very low dust concentrations in the atmosphere, the latter corresponding to the D/O interstadial (IS) events (IS 2–24 of Dansgaard et al. (1993), being warm climatic intervals during which very little dust reached Greenland).

Measurements of magnetic susceptibility and pedostratigraphy show clearly that the succession of loess and paleosols at Nussloch matches the general stratigraphy of the last

climatic cycle in western Europe. This succession, dated by AMS radiocarbon and OSL, has been correlated with the GRIP dust concentration for the last climatic cycle (Figure 11). Numerous events in the Greenland record corresponding to D/O events also appear to be recorded in the loess sequence (Figure 10).

At Nussloch, the variations of the IGR index through time appear similar to the fluctuations in Greenland dust fluctuation during the 32–19 ka interval, where the temporal resolution is best (Figure 11). Furthermore, similar and synchronous variations have been determined in other western European sequences (Achenheim, Mainz-Weisenau) within the EOLE project (CNRS-ECLIPSE program), providing support for the global value of the IGR grain size index used. Thus, it appears that the western European loess sequences faithfully record D/O events, with the Nussloch section rightfully considered as a key reference locality (Antoine et al., 2002; Rousseau et al., 2002). However, this record of the D/O events is a function of the strength and duration of interstadial warming, as is also expressed in the $\delta^{18}\text{O}$ values in Greenland ice cores. Work within the EOLE project indicates that the warmest and longest interstadials are marked by well-developed paleosols (Bw horizon at the base of the thick sequence). In contrast, if the duration of the period of low dust concentration is short, then the interstadial will be marked in the stratigraphy by a gley paleosol, the signature of which will depend on the strength of the corresponding warming.

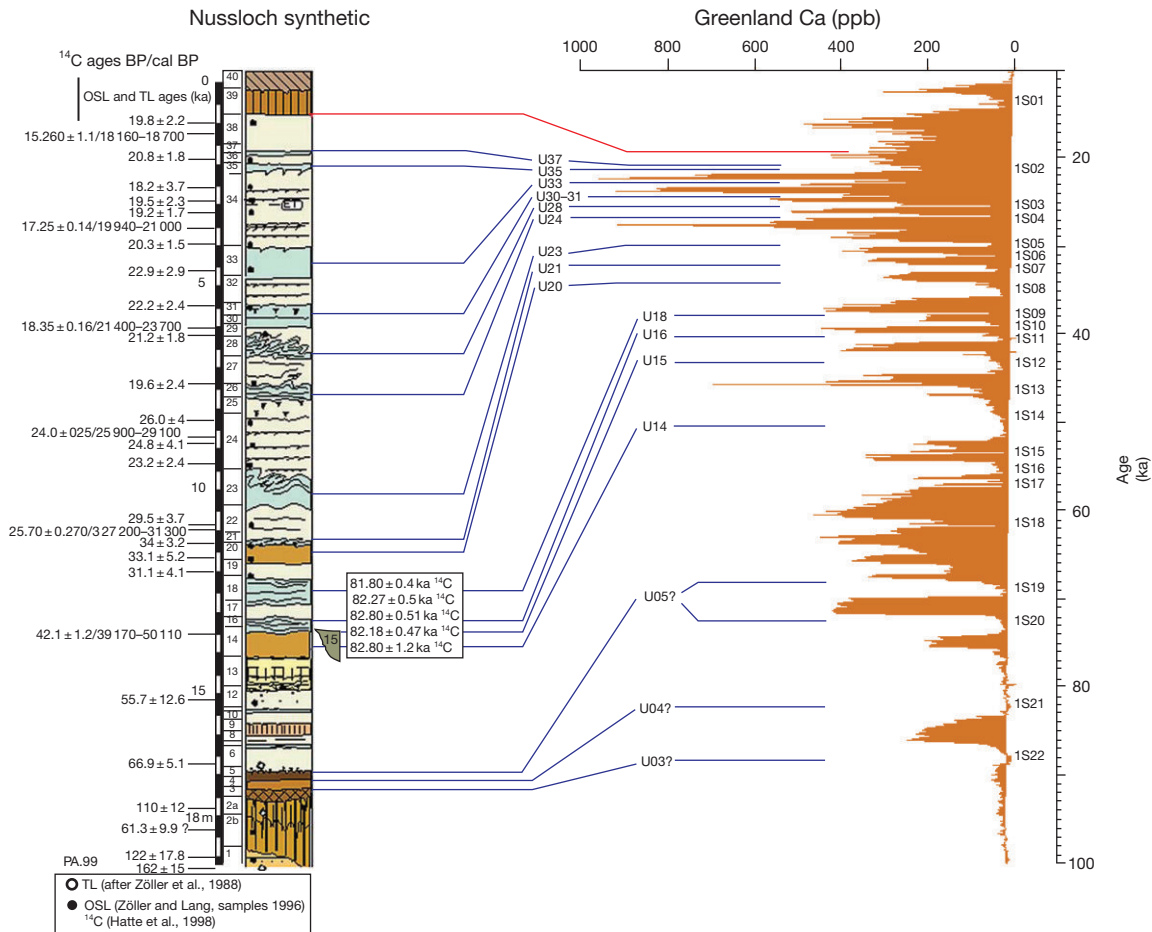


Figure 12 Pedostratigraphy of the Nussloch sequence compared with the Greenland dust record. Reproduced from Antoine P, et al. (2001) High-resolution record of the last interglacial–glacial cycle in the loess paleosol sequences of Nussloch (Rhine Valley–Germany). *Quaternary International* 76/77: 211–229; Rousseau DD, Antoine P, Hatté C, and Moine O (2003) Enregistrements des événements rapides de type Dansgaard–Oeschger dans les séquences loessiques européennes. *Lettre pigb-pmrc-France* 15: 21–24.

Variations in organic matter $\delta^{13}\text{C}$ support this interpretation, having been mainly linked to climatic fluctuations (availability of water and atmospheric CO_2 concentration) as recorded by the vegetation (Hatté et al., 1998, 2001b). Warming during D/O events is also marked by larger terrestrial mollusk populations, as indicated by a greater abundance of counted individuals. Although the loess record and its correlation with Greenland is well documented at Nussloch, similar patterns, especially in the stratigraphy and also in the grain size variations, have been described all along the loess belt from northern France eastward to the Czech Republic and Ukraine, demonstrating again that European loess has faithfully recorded the climatic variations that occurred in the North Atlantic.

Summary

The loess of Europe is mostly an eolian sediment, generally presenting elements of both local and global origin. It is indicative of periglacial environmental conditions, which made the fine material available to wind transport, originating mainly from sandurs or dried-out braided rivers, moraines, or

dried-out shelves. Considering their distribution, thickness, and complexity around the margins of the Quaternary ice sheets in the Northern Hemisphere, loess sequences can be considered as one of the best records of global environmental changes on the continents.

European loess sequences have been intensively studied for many decades, but increasingly higher stratigraphic resolution and availability of a growing range of climate proxy indicators have resulted in some notable advances in recent years. Climatic variability has been analyzed at high resolution based on different proxies. Sequences studied have revealed that the main loess deposition started at about 70 ka and ended around 16–15 ka. For example, the magnetic susceptibility record of the Achenheim loess sequence (France) has been correlated with the Upper Pleistocene (70–15 ka) Greenland dust content. Other results have shown that abrupt changes, named markers, are also recorded in the soil complexes. Markers are generally finer grained than most loess, but mineral content does not differ significantly. These markers correspond to long-distance wind-transport episodes, recording clearly visible events. The results from the study of the Nussloch sequence (Rhine Valley) show that the loess sedimentation, *in sensu*

stricto, is rhythmic, its fluctuations corresponding with rapid events of both marine and glacial type.

Analysis of particle size variation is a key method in loess research. Preliminary comparison of the grain size record from the Nussloch sequence, in the Rhine Valley, and the dust content from the GRIP ice core in Greenland shows high-frequency peaks that correlate with dust content in ~1.5-ka cycles, the main ones being associated with the North Atlantic Heinrich Events. This supports the hypothesis that European loess sequences contain a record of rapid climatic changes. Recently, work on $\delta^{13}\text{C}$ variations in organic matter from European loess sequences has shown that $\delta^{13}\text{C}$ parallels the GISP2 $\delta^{18}\text{O}$ variations, and is interpreted as recording the D/O events. Variation in this index was interpreted as recording the response of the local vegetation to climate changes. However, because there were no changes in the type of photosynthetic pathway, this index is also considered to be a proxy for local annual precipitation. In mid-latitudes, therefore, the dust intervals appear to correspond to periods when, although vegetation cover was reduced, it was adequate to provide sufficient organic matter from which to abstract a signal of biological activity (i.e., mollusks). The Greenland dust record also shows that isotope stages 2 and 4 were dustier than stage 3 and that important variations occurred in the dust content of the atmosphere during the same interval. Some of the oscillations were contemporaneous with the massive iceberg discharges named Heinrich Events.

See also: Loess Deposits: Origins and Properties. **Luminescence Dating:** Optical Dating; Thermoluminescence. **Paleoclimate Reconstruction:** Sub-Milankovitch (DO/Heinrich) Events. **Permafrost and Periglacial Features:** Cryoturbation Structures; Ice Wedges and Ice-Wedge Casts. **Quaternary Stratigraphy:** Pedostratigraphy.

References

- Antoine P (1991) Nouvelles données sur la stratigraphie du pléistocène supérieur de la France septentrionale, d'après les sondages effectués sur le tracé du TGV Nord. *Publications du CERP* 3: 9–20.
- Antoine P (1994) The Somme valley terrace system (northern France): A model of river response to Quaternary climatic variations since 800,000 BP. *Terra Nova* 6: 453–464.
- Antoine P, Catt J, Lautridou JP, and Sommé J (2003) The loess and coversands of northern France and southern England. *Journal of Quaternary Sciences* 18: 309–318.
- Antoine P, Lautridou JP, and Laurent M (2000) Long-term fluvial archives in NW France: Response of the Seine and Somme rivers to tectonic movements, climatic variations and sea level changes. *Geomorphology* 33: 183–207.
- Antoine P, Munaut AV, and Sommé J (1994) Réponse des environnements aux climats du début glaciaire weichselien: Données de la France du Nord-Ouest. *Quaternaire* 5: 151–156.
- Antoine P, Rousseau DD, Lautridou JP, and Hatté C (1999) Last interglacial–glacial climatic cycle in loess–paleosol successions of north-western France. *Boreas* 28: 551–563.
- Antoine P, et al. (2001) High-resolution record of the last interglacial–glacial cycle in the loess paleosol sequences of Nussloch (Rhine Valley–Germany). *Quaternary International* 76/77: 211–229.
- Antoine P, et al. (2002) Événements éoliens rapides dans les loess du Pléniglaciaire supérieur weichselien: l'exemple de la séquence de Nussloch (Vallée du Rhin–Allemagne). *Quaternaire* 13: 199–208.
- Antoine P, et al. (2003) The Pleistocene rivers of the English Channel region. *Journal of Quaternary Science* 18: 227–243.
- Auffret JP, Horn R, Larssonneur C, Curry D, and Smith AJ (1982) *La Manche orientale, carte des paléovallées et des bancs sableux*. Bureau de Recherches Géologiques et Minières edit.
- Dansgaard W, et al. (1993) Evidence for general instability of past climate from a 250-kyr ice-core record. *Nature* 364: 218–220.
- Derbyshire E, Billard A, van Vliet-Lanoë B, Lautridou J-P, and Cremaschi M (1988) Loess and palaeoenvironment: Some results of a European joint programme of research. *Journal of Quaternary Science* 3: 147–169.
- Derbyshire E and Mellors TW (1988) Geological and geotechnical characteristics of some loess and loessic soils from China and Britain: A comparison. *Engineering Geology* 25: 135–175.
- Feng X and Epstein S (1995) Carbon isotopes of trees from arid environments and implications for reconstructing atmospheric CO_2 concentration. *Geochimica et Cosmochimica Acta* 59: 2599–2608.
- Frechen M, Horváth E, and Gábris G (1997) Geochronology of Middle and Upper Pleistocene loess sections in Hungary. *Quaternary Research* 48: 291–312.
- Frechen M, Oches EA, and Kohfeld KE (2003) Loess in Europe – Mass accumulation rates during the Last Glacial Period. *Quaternary Science Reviews* 22(1819): 1835–1857.
- French HM (1996) *The Periglacial Environment*. London: Longman.
- Grabowska-Olszewska B (1988) Engineering – Geological problems of loess in Poland. *Engineering Geology* 25: 177–199.
- Grahmann R (1932) Der löss in Europa. *Mitteilungen der Gesellschaft für Erdkunde zu Leipzig* 51: 5–24.
- Haase G, Haase D, Ruske R, Jäger KD, and Altermann M (2006) Loess in Europe – Spatial distribution in a scale 1:2,500,000. *Quaternary Science Reviews* 26: 1301–1312.
- Haesaerts P, Borziak I, Chirica V, Damblon F, Koulakova L, and van der Plicht J (2003) The east Carpathian loess record: A reference for the Middle and Late Pleniglacial stratigraphy in central Europe. *Quaternaire* 14: 163–188.
- Haesaerts P, Juvigné E, Kuyl O, Mucher H, and Roebroeks W (1981) Compte rendu de l'excursion du 13 juin 1981, en Hesbaye et au Limbourg Néerlandais, consacrée à la chronostratigraphie des loess du Pléistocène supérieur. *Annales de la Société Géologique de Belgique* 104: 223–240.
- Haesaerts P, et al. (2005) The loess–paleosol succession of Kurtak (Yenisei Basin, Siberia): A reference record for the Karga stage (MIS 3). *Quaternaire* 16(1): 3–24.
- Hatté C (2000) *Les Isotopes du Carbone (^{14}C et ^{13}C) Dans la Matière Organique des Loess de l'Europe du Nord-Ouest: Applications Paléoclimatiques*, p. 175. Orsay: Université Paris-Sud. Thèse de Doctorat.
- Hatté C and Gauthier C (2006) *The loess war*. .
- Hatté C and Guiot J (2005) Paleoprecipitation reconstruction by inverse modelling using the isotopic signal of loess organic matter: Application to the Nussloch loess sequence (Rhine Valley, Germany). *Climate Dynamics* 25: 315–327.
- Hatté C, Pessenda LC, Lang A, and Paterne M (2001) Development of accurate and reliable ^{14}C chronologies for loess deposits: Application to the loess sequence of Nussloch (Rhine Valley, Germany). *Radiocarbon* 43(2B): 611–618.
- Hatté C, et al. (1998) $\delta^{13}\text{C}$ variations of loess organic matter as a record of the vegetation response to climatic changes during the Weichselian. *Geology* 26(7): 583–586.
- Hatté C, et al. (1999) New chronology and organic matter $\delta^{13}\text{C}$ paleoclimatic significance of Nussloch loess sequence (Rhine Valley, Germany). *Quaternary International* 62: 85–91.
- Hatté C, et al. (2001) $\delta^{13}\text{C}$ of loess organic matter as a potential proxy for precipitation. *Quaternary Research* 55(1): 33–38.
- Horváth E (2001) Marker horizons in the loesses of the Carpathian Basin. *Quaternary International* 76/77: 157–163.
- Hradilova J and Stastny M (1994) Changes in the clay fraction mineral composition in a loess profile of the last interglacial and early glacial in Praha-Sedlec. *Acta Universitatis Carolinae Geologica* 38: 229–238.
- Juvigné E (1985) The use of heavy mineral suites for loess stratigraphy. *Geologie en Mijnbouw* 64: 333–336.
- Juvigné EH and Wintle AG (1988) A new chronostratigraphy of the late Weichselian loess units in Middle Europe based on thermoluminescence dating. *Eiszeitalter und Gegenwart* 38: 94–105.
- Kukla G (1977) Pleistocene land-sea correlations. 1: Europe. *Earth-Science Reviews* 13: 307–374.
- Kukla G, An ZS, Melice JL, Gavin J, and Xiao JL (1990) Magnetic susceptibility record of Chinese Loess. *Transactions of the Royal Society Edinburgh: Earth Sciences* 81: 263–288.
- Kukla G and Cilek V (1996) Plio-Pleistocene megacycle: Record of climate and tectonics. *Palaeogeography, Palaeoclimatology, Palaeoecology* 120: 171–194.

- Kukla G and Koci A (1972) End of the last interglacial in the loess record. *Quaternary Research* 2: 374–383.
- Kukla G, McManus JF, Rousseau DD, and Chuine I (1997) How long and how stable was the last interglacial? *Quaternary Science Reviews* 16: 605–612.
- Lang A, et al. (2003) High-resolution chronologies for loess: Comparing AMS ^{14}C and optical dating results. *Quaternary Science Reviews* 22(10–13): 953–959.
- Lautridou JP (1974) La séquence loessique séquanienne du Würm à Saint-Pierres-Elbeuf. *Bulletin de l'Association Française pour l'Etude du Quaternaire* 40–41: 242–243.
- Lautridou JP (1985) *Le Cycle Périglaciaire Pléistocène en Europe du Nord-Ouest et Plus Particulièrement en Normandie*. Thèse Doctorat es Sciences Thesis, p. 908. Caen: Université Caen.
- Lautridou JP and Sommé J (1974) Les loess et les provinces climato-sédimentaires du Pléistocène supérieur dans le Nord-Ouest de la France. Essai de corrélation entre le Nord et la Normandie. *Bulletin de l'Association Française pour l'Etude du Quaternaire* 40–41: 237–241.
- Lautridou JP, et al. (1986) Corrélations entre sédiments quaternaires continentaux et marins (littoraux et profonds) dans le domaine France septentrionale-Manche. *Revue de Géologie Dynamique et Géographie Physique* 27(2): 105–112.
- Léger M (1990) Loess landforms. *Quaternary International* 7/8: 53–61.
- Lozek V (1990) Molluscs in loess, their paleoecological significance and role in geochronology – Principles and methods. *Quaternary International* 7/8: 71–79.
- Lyell C (1833) *The Principles in Geology*. London: John Murray.
- Maruszczak H and Wilgat M (1995) Stratigraphical and paleogeographical interpretation of the results of heavy mineral analyses in loesses of Vojvodina. In: Maruszczak H (ed.) *Problems of the Stratigraphy and Paleogeography of Loesses in Central Europe*, pp. 173–190. Lublin: Analecta Universitatis Mariae Curie-Skłodowska.
- Moine O, Rousseau DD, Jolly D, and Vianey-Liaud M (2002) Paleoclimatic reconstruction using mutual climatic range on terrestrial mollusks. *Quaternary Research* 57(1): 162–172.
- O'Leary MH (1981) Carbon isotope fractionation in plants. *Phytochemistry* 20(4): 553–567.
- O'Leary MH (1988) Carbon isotopes in photosynthesis. *Bioscience* 38(5): 328–336.
- Oches EA and McCoy W (1995) Amino acid geochronology applied to the correlation and dating of Central European loess deposits. *Quaternary Science Reviews* 14: 767–782.
- Pastre JF, Billard A, Debard E, Faure M, and Guerin C (1996) A tephritic horizon Mont-Dore provenance in the Plio-Pleistocene loessic sequence of Saint-Vallier (France). *Comptes Rendus de l'Académie des Sciences de Paris* 323(7): 607–614.
- Perederij VI (2001) Clay mineral composition and paleoclimatic interpretation of the Pleistocene deposits of Ukraine. *Quaternary International* 76/77: 113–121.
- Péwé TL (1962) Ice wedges in permafrost. Lower Yukon River Area, near Galena, Alaska. *Biuletyn Peryglacjalny* 11: 65–76.
- Pustovoytov K and Terhorst B (2004) An isotopic study of a late Quaternary loess–paleosol sequence in SW Germany. *Revista Mexicana de Ciencias Geológicas* 21(1): 88–93.
- Pye K (1984) Loess. *Progress in Physical Geography* 8: 176–217.
- Pye K (1995) The nature, origin and accumulation of loess. *Quaternary Science Reviews* 14: 653–657.
- Rousseau DD (1987) Paleoclimatology of the Achenheim series (middle and upper Pleistocene, Alsace, France) – A malacological analysis. *Palaeogeography, Palaeoclimatology, Palaeoecology* 59(4): 293–314.
- Rousseau DD (1991) Climatic transfer function from Quaternary molluscs in European loess deposits. *Quaternary Research* 36: 195–209.
- Rousseau DD (2001) Loess biostratigraphy: New advances and approaches in mollusk studies. *Earth-Science Reviews* 54(1–3): 157–171.
- Rousseau DD, Antoine P, Hatté C, and Moine O (2003) Enregistrements des événements rapides de type Dansgaard-Oeschger dans les séquences loessiques européennes. *Lettre pigb-pmrc-France* 15: 21–24.
- Rousseau DD, Gerasimenko N, Matviischina Z, and Kukla G (2001) Late Pleistocene environments of the Central Ukraine. *Quaternary Research* 56(3): 349–356.
- Rousseau DD, Kukla G, Zöller L, and Hradilova J (1998) Early Weichselian dust storm layer at Achenheim in Alsace, France. *Boreas* 27: 200–207.
- Rousseau DD and Puisségur JJ (1990) A 350,000 years climatic record from the loess sequence of Achenheim, Alsace, France. *Boreas* 19: 203–216.
- Rousseau DD, Puisségur JJ, and Lautridou JP (1990) Biogeography of the Pleistocene Pleniglacial malacofaunas in Europe. Stratigraphic and climatic implications. *Palaeogeography, Palaeoclimatology, Palaeoecology* 80: 7–23.
- Rousseau DD, Puisségur JJ, and Lecoille F (1992) West-European molluscs assemblages of stage 11: Climatic implications. *Palaeogeography, Palaeoclimatology, Palaeoecology* 92: 15–29.
- Rousseau DD, Zoller L, and Valet JP (1998) Late Pleistocene climatic variations at Achenheim, France, based on a magnetic susceptibility and TL chronology of loess. *Quaternary Research* 49(3): 255–263.
- Rousseau DD, et al. (2002) Abrupt millennial climatic changes from Nussloch (Germany) Upper Weichselian eolian records during the Last Glaciation. *Quaternary Science Reviews* 21(14–15): 1577–1582.
- Rozycki SZ (1991) *Loess and Loess-Like Deposits*, p. 187. Wrocław: Ossolineum-Polish Academy of Sciences.
- Semmel A (1967) Neue Fundstellen von vulkanischem Material in hessischen Lössen. *Notizblatt des Hessischen Landesamtes fuer Bodenforschung zu Wiesbaden* 95: 104–108.
- Shi C, et al. (2003) Climate variations since the last interglacial recorded in Czech loess. *Geophysical Research Letters* 30(11): 16. <http://dx.doi.org/10.1029/2003GL017251>.
- Smalley IJ (1966) The properties of glacial loess and the formation of loess deposits. *Journal of Sedimentary Petrology* 36: 669–676.
- Smalley IJ, Jefferson IF, Dijkstra TA, and Derbyshire E (2001) Some major events in the development of the scientific study of loess. *Earth-Science Reviews* 54(1–3): 5–18.
- Sommé J, et al. (1986) Le cycle climatique du Pléistocène Supérieur dans les loess d'Alsace à Achenheim. In: Morzadec-Kerfourn MT (ed.) *Oscillations climatiques entre 125 000 ans et le maximum glaciaire. Corrélations entre les domaines marin et continental*, *Bulletin de l'Association Française pour l'Etude du Quaternaire*, vol. 25–26, pp. 97–104.
- Sommé J, et al. (1994) The Watten boring – An Early Weichselian and Holocene climatic and paleoecological record from the French North Sea coastal plain. *Boreas* 23: 231–243.
- Van Vliet-Lanoë B (1987) *Le Rôle de la Glace de Ségrégation Dans les Formations Superficielles de l'Europe de l'Ouest*, p. 864. Doctorat es Sciences Thesis. Caen: Université Paris I.
- Van Vliet-Lanoë B and Coutard JP (1984) Structures caused by repeated freezing and thawing in various loamy sediments: A comparison of active, fossil and experimental data. *Earth Surface Process Landform* 9: 553–565.
- Vandenberghe J, Huijzer B, Múcher H, and Laan W (1998) Short climatic oscillations in a western European loess sequence (Kesselt, Belgium). *Journal of Quaternary Sciences* 13: 471–485.
- von Richtofen F (1882) On the mode of origin of the loess. *Geological Magazine* 9: 293–305.
- Wintle AG, Shackleton NJ, and Lautridou JP (1984) Thermoluminescence dating of periods of loess deposition and soil formation in Normandy. *Nature* 310: 491–493.
- Zöller L, Stremme HE, and Wagner GA (1988) Löss-Paläoboden-Sequenzen von Nieder-, Mittel- und Oberrhein. *Chemical Geology* 73: 39–62.
- Zöller L and Wagner GA (1990) Thermoluminescence dating of loess – Recent developments. *Quaternary International* 7/8: 119–128.

North America

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Introduction

The loess deposits of North America cover extensive areas of the midcontinent, northwestern United States, and Alaska and neighboring parts of the Yukon Territory of northwestern Canada. The loess of North America is derived from both glaciogenic and nonglaciogenic sources, and source areas vary both spatially and temporally. The resultant deposits of wind-blown (eolian) silt and intercalated paleosols preserve important records of past climate and environmental conditions during the Quaternary. The loess deposits may act as both a sink and a source of dust, and recently it has been suggested that loess not only reflects changing climatic conditions, but may also potentially cause climate change through radiative forcing, and also through biogeochemical changes in atmospheric, terrestrial, and marine environments (Harrison et al., 2001).

Distribution and Thickness of Loess Deposits

Loess is extensive over North America, particularly in the midcontinent region, Alaska, and the northwestern United States (Figure 1). The distribution of loess deposits in North America is governed primarily by the balance of supply of suitable sediment and the availability of conditions suitable for accumulation (Muhs et al., 2003); these conditions for accumulation include vegetation and topography (Mason et al., 1999; Pye, 1996); source proximity, wind strength, and direction also influencing the pattern of loess distribution (Muhs and Bettis, 2000). In terms of their spatial extent, the principal areas of North America where loess deposits are preserved are: the midcontinent (incorporating the Great Plains and Central Lowland physiographic provinces) plus the Lower Mississippi Valley and the Wabash and Ohio River Valleys; the Pacific northwestern United States (i.e., the Columbia Plateau physiographic province comprising two areas, the Columbia Plateau of eastern Washington and western Oregon, and the Snake River area of southern Idaho); plus Alaska and the neighboring Yukon Territory of northwestern Canada. The notable absence of areally extensive loess deposit in Canada is due to the fact that most of the area was covered by the Laurentide Ice Sheet (LIS) during the last glacial period. Loess deposits are even lacking in those areas of Canada that were deglaciated first. The reason for this is not well understood, but it is likely that the large proglacial lakes, such as glacial Lake Agassiz, served as the sinks for most glaciogenic silt (Flint, 1971).

Maximum loess thicknesses in the North American midcontinent can exceed 60 m; in the Lower Mississippi Valley, bluffs composed of loess 20–30 m high are common (Markewich et al., 1998), with maximum loess thicknesses

occurring to the east of the valley (Figure 2). In many locations in central North America, loess deposited during the last glacial period (Peoria Loess) accounts for more than 90% of the total loess thickness (Pye, 1987). The thickest known deposits of Late Glacial loess in the world occur in central Nebraska (48 m thickness at Bignell Hill), with considerable thicknesses of Peoria Loess also being found in Iowa (41 m at the Loveland Loess paratype) (Bettis et al., 2003). In the Lower Mississippi Valley, Peoria Loess is 10–20 m thick, but the thickness decreases with distance from the river bluffs, i.e. with distance from the likely source areas. The present thicknesses of older loesses here, however, seem to be more closely linked to geomorphic position and degree of dissection than proximity of source (Busacca et al., 2004).

The loess on the Columbia Plateau (Figure 3) ranges from approximately 0.1 to 75 m thickness, with the thickest deposits being found in an area straddling the eastern Washington/western Idaho border, called 'The Palouse' (Busacca and McDonald, 1994). Loess cover in the Columbia Plateau province beyond The Palouse is thinner and less continuous. Loess deposits also cover large areas to the north and south of the Snake River Plain in Idaho and into western Wyoming. Here, the loess is typically 2 m thick, although it can reach up to 12 m thickness in places (Lewis et al., 1975).

Loess is the most widespread surficial deposit in Alaska (Figure 4), covering many areas below 300–450 m altitude (Péwé, 1975). The thicknesses of the deposits range from a few millimeters in some areas to more than 60 m at Gold Hill, north of the Tanana River near Fairbanks (Péwé, 1955). Loess deposits along the north side of the Tanana valley, and in the lower Yukon River valley are 3–12 m in thickness, while on the Seward Peninsula, deposits are reported to be more than 1.5 m thick and cover 75% of the area (Péwé, 1975). Throughout Alaska, loess deposits are thickest near rivers, with thicknesses decreasing rapidly with distance away from the rivers (Péwé, 1975) and downwind of valley dust sources (Lindholm et al., 1959; Muhs et al., 2004). The transport and deposition of loess are processes that are still active today in Alaska (Figure 5), with most activity being focused along the braided glacial streams of the Delta River (Figure 6), the Knik and Matanuska Rivers, and near the junction of the Tanana and Yukon Rivers (Péwé, 1975).

Stratigraphic Framework and Chronology of Loess Deposits

In the central United States, five major loess units have been identified, representing middle-to-late Quaternary deposition. The oldest unit is termed 'pre-Loveland Loess,' although several



Figure 1 Map of North America showing physiographic provinces and the distribution of Quaternary loess. Loess distribution compiled by the authors from sources given in [Figures 2–4](#).

other names also have been proposed (e.g., Crowleys Ridge Loess, Mariana Loess, and Sicily Island Loess); however, while some extensive and well-preserved pre-Loveland loess is found in parts of the Great Plains ([Mason et al., 2004](#)), these deposits are typically poorly preserved in the Central Lowland and northeastern United States, where they are present at all. The more commonly observed lowermost loess unit present in the central United States is Loveland Loess ([Figure 7](#)), which probably correlates with the penultimate glacial period (Marine oxygen isotope stage 6 (MIS-6); [Forman et al., 1992](#); [Maat and Johnson, 1996](#)) and is usually no more than a few meters thick. Loveland Loess, in turn, is overlain by at least two younger loess units. The older of these two units is of mid-Wisconsin age (MIS-3) and is variably termed ‘Roxana Silt’ (for deposits found in the Central Lowland province), ‘Pisgah Formation’ (Iowa), or ‘Gilman Canyon Formation’ (Great Plains). This mid-Wisconsin age loess is overlain by the late-Wisconsin age (MIS-2) ‘Peoria Loess.’ A Holocene loess termed the ‘Bignell Loess,’ which is typically <2 m in thickness, is also present at some Great Plains locations ([Mason et al., 2003](#)). No Holocene loess unit has been identified east of the Missouri River valley; that is, in the Central Lowland ([Figure 2](#); [Bettis et al., 2003](#)).

Loveland Loess has a soil from the last interglacial period (MIS-5 and part of MIS-4) developed in the uppermost part of it, termed the ‘Sangamon Geosol’ ([Grimley et al., 2003](#); [Markewich et al., 1998](#); [Muhs and Bettis, 2000](#)). This paleosol is typically thick (3–5 m), and morphologically better expressed than the modern surface soil. The upper part of the

mid-Wisconsin loess has also been modified by the development of the ‘Farndale Soil’ in the Central Lowland province, and the equivalent Gilman Canyon Formation in the Great Plains province. Where these loesses are thin, the Farndale and Gilman Canyon Formation are welded to the Sangamon Geosol. Peoria Loess overlies the Farndale Soil (or the Gilman Canyon Formation in the Great Plains). The modern soil is developed in Peoria Loess in the Central Lowland. However, in the Great Plains, where Bignell Loess buries Peoria Loess the Brady Soil is formed in the upper part of the Peoria Loess and the modern soil is formed in Bignell Loess.

While the broad temporal framework of the various stratigraphic units identified in the loess sequences of the central United States as outlined above is reasonably well understood, the specific timing of deposition and cessation of accumulation varies spatially across the region. The timing of Peoria Loess deposition, for example, demonstrates important regional differences, varying from between approximately 30 000–20 000 carbon-14 (^{14}C) years before present (BP) for the onset of deposition, and approximately 14 000–10 000 ^{14}C years BP for the end of deposition, based respectively on maximum- and minimum-limiting radiocarbon dates of the bracketing paleosols ([Bettis et al., 2003](#); [Muhs and Zárte, 2001](#)). Luminescence and beryllium-10 (^{10}Be) ages obtained directly for the Peoria Loess also support these radiocarbon dates (e.g., [Bettis et al., 2003](#); [Roberts et al., 2003](#)). Those areas located close to major loess sources record accumulation of loess earlier than areas at more distal locations ([Bettis et al., 2003](#); [Muhs and Zárte, 2001](#); [Ruhe, 1983](#)). Furthermore, accumulation of loess

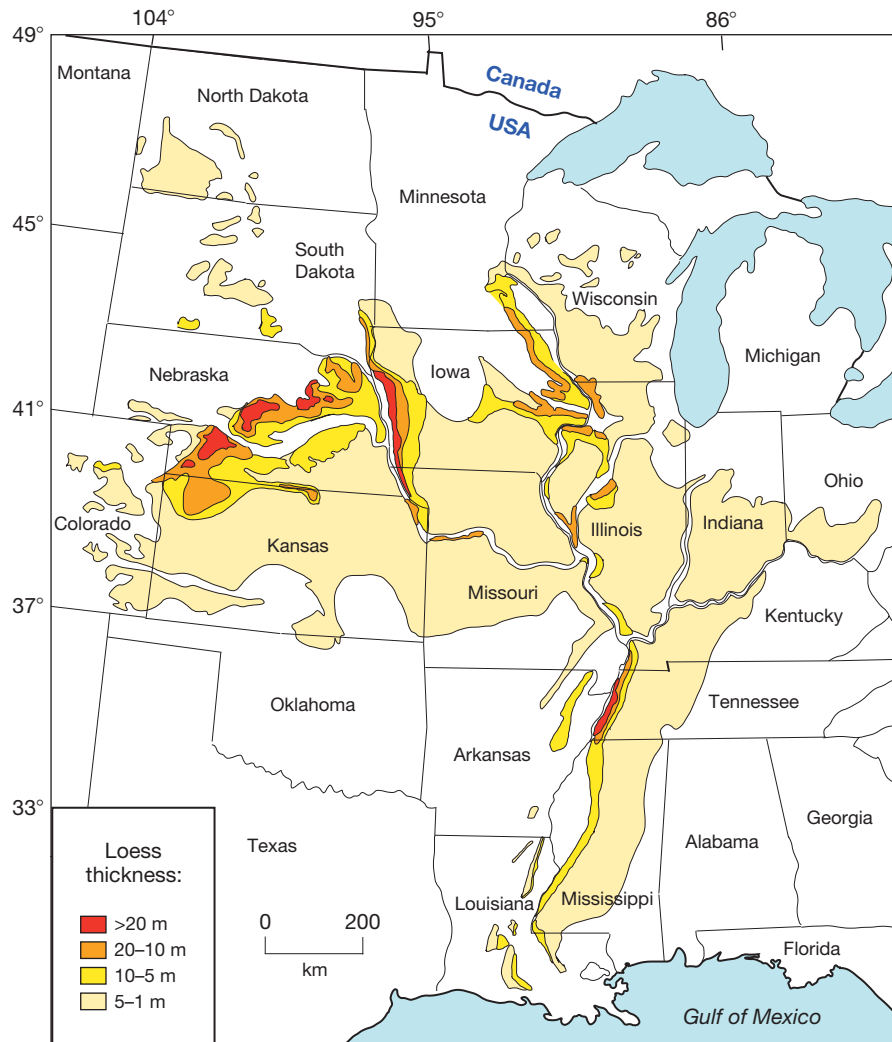


Figure 2 Map showing the distribution and thickness of last glacial (Peoria) loess in the midcontinent of North America (Central Lowland and Great Plains physiographic provinces). Redrawn from compilation in Figure 2 of Bettis et al. (2003) and sources therein.

also fluctuated over time, resulting in further local stratigraphic variations (e.g., the formation of weakly developed paleosols within Peoria Loess, indicating brief periods of no deposition) which have not been correlated on a regional scale (Busacca et al., 2004; Ruhe et al., 1971; Wang et al., 2000).

The loess deposits in the Columbia Plateau physiographic province preserve a very different record than those of the central United States, both in terms of duration and likely origins. In the Columbia Plateau area of eastern Washington (Figure 8), loess records may extend as far back as 1–2 Ma (Busacca, 1991) and they preserve many paleosols, reflecting periods of nondeposition and stability between times of more active loess deposition. McDonald and Busacca (1992) defined a regional soil-stratigraphic sequence comprising (from oldest to youngest) the Devil's Canyon Soil, the Old Maid Coulee Soil, the Washtucna Soil, and the Sand Hills Coulee Soil. Furthermore, Busacca and McDonald (1994) ascribed the terms 'L1' and 'L2' to the uppermost and the next oldest loess units in the region, respectively (Figure 9). The uppermost unit, L1, contains Holocene loess and the Mazama tephra (dated to ~7700 cal years BP) which, in turn, are located

above the Sand Hills Coulee Soil; the base of this unit is marked by the Mount St. Helens set S tephra (dated to ~15400 cal years BP). The next oldest unit, L2, contains the Washtucna Soil complex and the Old Maid Coulee Soil, and toward the base of the unit the Mount St. Helens set C tephra provides another useful correlative and chronologic marker. Unit L1 is of late Wisconsin age, while L2 is reported to be of early to middle Wisconsin age (Busacca and McDonald, 1994; Richardson et al., 1997). The oldest of the four soil-stratigraphic units identified by McDonald and Busacca (1992), the Devils Canyon Soil, is located at the top of the next oldest loess unit, L3.

In the Snake River Plain area of the Columbia Plateau physiographic province, at least four Quaternary loess units have been identified, separated by buried soils (McDole et al., 1973). The youngest two Quaternary loess units are separated by a well-developed paleosol, and were informally termed 'Loess A' (uppermost unit) and 'Loess B' by Pierce et al. (1982). Loess A is believed to have been deposited during the late Wisconsin (Forman et al., 1993; Kuntz et al., 1986; Pierce et al., 1982). There is, however, some disagreement regarding

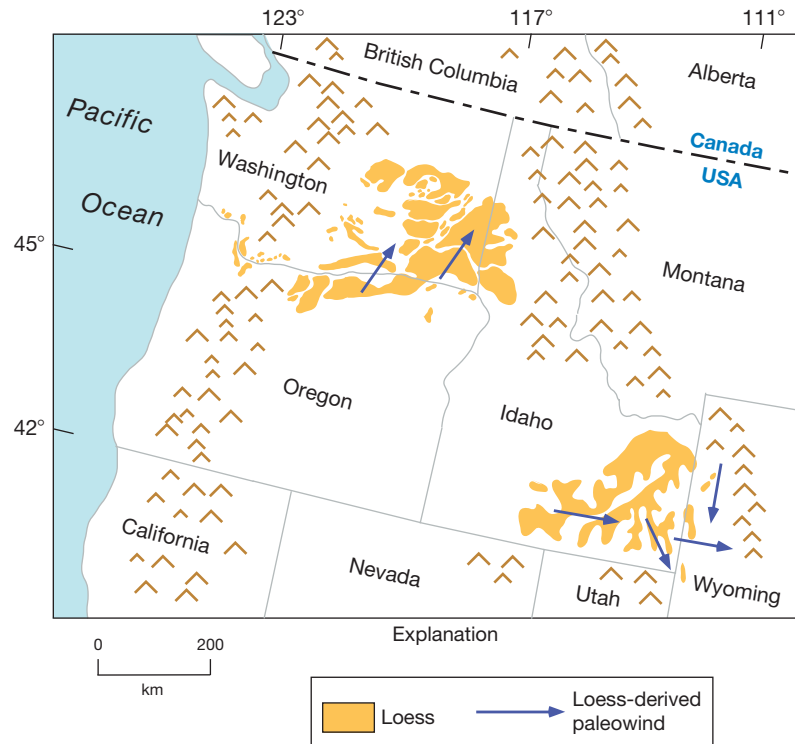


Figure 3 Map showing the distribution of loess and inferred paleowinds in the northwestern United States. Compiled from Lewis et al. (1975) and Busacca and McDonald (1994).

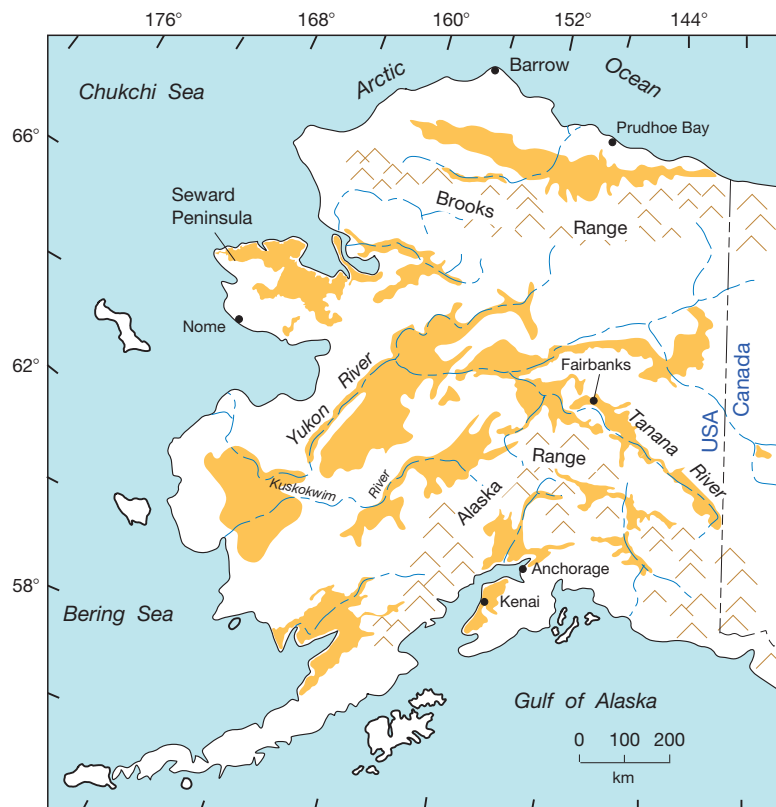


Figure 4 Map showing the distribution of loess in Alaska. Redrawn from Péwé (1975) and other sources given in Muhs et al. (2003).



Figure 5 Photograph of dust storm on the Richardson Highway, near Delta Junction, Alaska, derived from silts on the Delta River floodplain, July, 2004. Photograph by E.A. III Bettis.

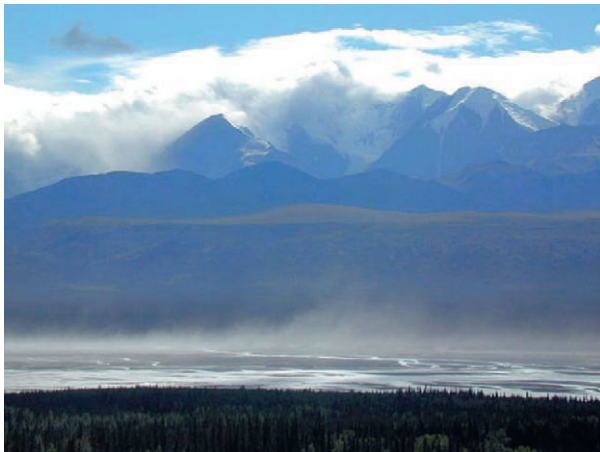


Figure 6 Photograph of dust storm (same as in [Figure 5](#)) on the Delta River floodplain, looking southwest, south of Delta Junction, Alaska, July 2004. Sources of silt in the Delta River are glaciers in the Alaska Range (background); Holocene loess is several meters thick on adjacent moraines and outwash terraces (foreground). Photograph by E.A. III Bettis.

the age of Loess B; an early Wisconsin age is proposed by [Forman et al. \(1993\)](#), on the basis of thermoluminescence ages, whereas [Pierce et al. \(1982\)](#) propose a penultimate glacial age for Loess B on the basis of the degree of soil development and stratigraphic relations with other well-dated deposits.

The loess deposits preserved in Alaska extend even further back in time than those of the Columbia Plateau province, spanning as many as 3 Ma from the end of the Tertiary and through the whole of the Quaternary Period for some localities in unglaciated central Alaska. This age range is based on fission-track dating of tephra preserved in the loess deposits, coupled with observations of paleomagnetic reversals ([Westgate, 1988; Westgate et al., 1990](#)). Numerous paleosols are preserved in the loess of Alaska ([Begét and Hawkins, 1989; Muhs et al., 2003; Figure 10](#)) and the neighboring Yukon Territory suggesting that deposition rates fluctuated over time. Deposits of loess in central Alaska accumulated both before

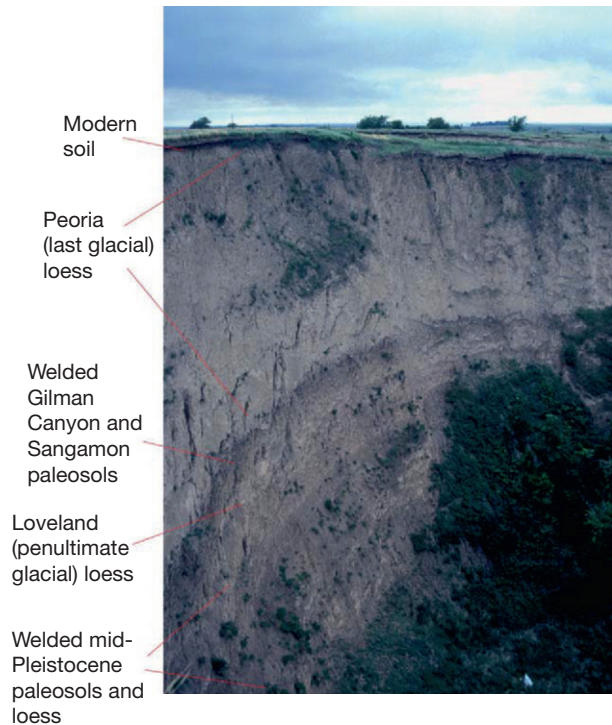


Figure 7 Photograph of modern soil, last glacial (Peoria) loess, mid-Wisconsin, last interglacial, and older paleosols near Eustis, Nebraska. See [Maat and Johnson \(1996\)](#) and [Roberts et al. \(2003\)](#) for details on stratigraphy and luminescence ages; paleoclimatic inferences of fossil snails in this section are given in [Rousseau and Kukla \(1994\)](#). Photograph by D.R. Muhs.

and during the last interglacial, the mid-Wisconsin stadials, during the last glacial, through the Holocene, and up to the present day ([Muhs et al., 2003; Péwé, 1975](#)).

Sources of Loess and Timing of Deposition

Loess is classically viewed as a glacially derived sediment, leading to the view that loess deposits denote times of increased glacial activity and the intercalated paleosols indicate warmer interglacial or interstadial periods. However, the loess of North America comes from both glaciogenic and nonglaciogenic sources. The response of the different loess regions of North America to glacial and interglacial conditions varies, with the timing of loess deposition depending on the interplay between various factors including climatic conditions, sediment supply, sediment source, and vegetation cover.

In the central United States, geochemical and isotopic provenance studies have shown that the loess deposits can be of both glaciogenic and nonglaciogenic origins, with changes in source possible over time. For example, Peoria Loess in western Iowa shows two different sources over this period; the oldest Peoria Loess is derived from the Missouri River valley (i.e., glacially derived material), while the younger Peoria Loess is derived from a mixture of local glaciogenic material and distal nonglaciogenic sources ([Muhs and Bettis, 2000](#)). Geochemical and isotopic studies of loess from the central United States have revealed that the composition of Peoria Loess varies



Figure 8 Photograph of typical loess-mantled landscape in the Palouse region of eastern Washington and roadcuts showing loess–paleosol stratigraphy. Photograph by H.M. Roberts.

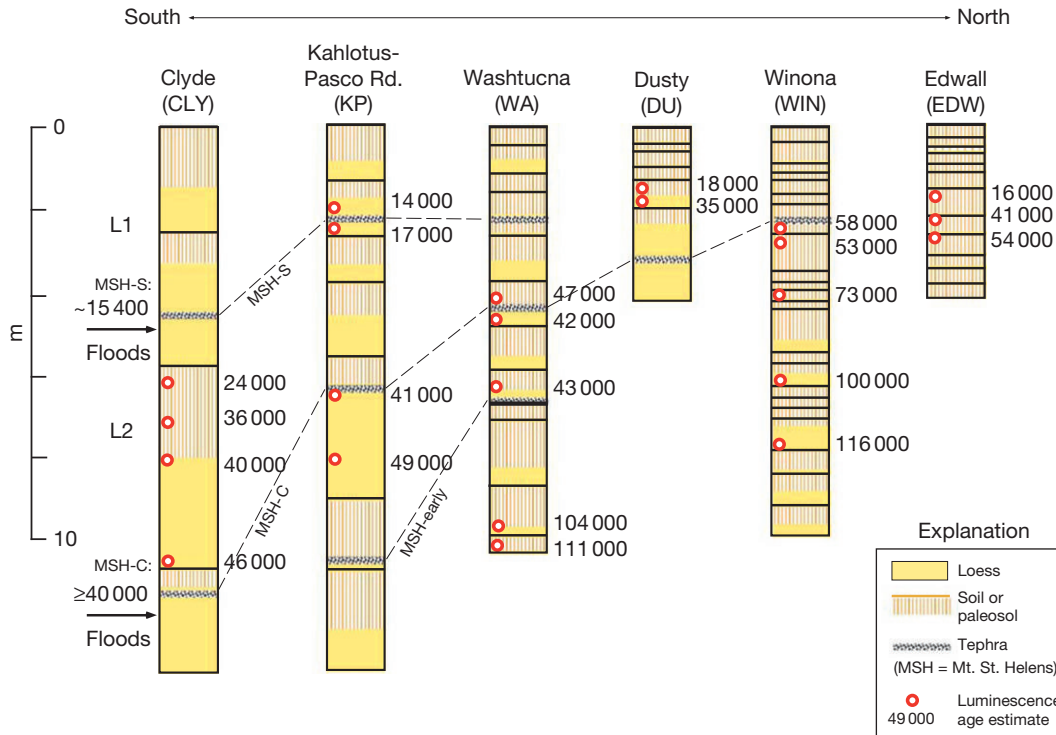


Figure 9 Loess stratigraphy and luminescence chronology at selected localities in the Palouse region of the Columbia Plateau region of North America. Redrawn from Richardson CA, McDonald EV, and Busacca AJ (1997) Luminescence dating of loess from the northwest United States. *Quaternary Science Reviews* 16: 403–415; Busacca AJ and McDonald EV (1994) Regional sedimentation of late Quaternary loess on the Columbia Plateau: Sediment source areas and loess distribution patterns, regional geology of Washington State. *Washington Division of Geology and Earth Resources Bulletin* 80: 181–190.

greatly across the region; this reflects the different sources of material even over relatively short distances. In the Great Plains province, for example, studies have shown that silts from Tertiary White River Group outcrops in southern South Dakota and northern Nebraska form a major constituent of Peoria Loess in Nebraska (Aleinikoff et al., 1998). In neighboring eastern Colorado loess source material is derived from both the White River Group and glaciogenic valley sources draining the Front Range of Colorado (Aleinikoff et al., 1999). In the Central Lowland province, loess deposits located east of the Missouri and Mississippi Rivers are composed dominantly of glaciogenic source material, but some nonglaciogenic sources provide significant local loess contributions (Bettis et al., 2003). In the loess deposits of the lower Illinois and central Mississippi River Valleys, between 10 and 40% of the sediment is estimated to be derived from nonglaciogenic sources, based on the comparison of magnetic susceptibility, and silt and clay mineralogy (Grimley, 2000).

Loess deposits accumulated during glacial periods or stadials in the central United States, while paleosols formed during interglacials or interstadials. The cumelic nature of the Farndale Soil of the Central Lowland province and the Gilman Canyon Formation of the Great Plains province, suggests that loess accumulation continued throughout the interstadial but was sufficiently slow to allow pedogenesis to occur (Muhs et al., 2003). Most of the loess deposits preserved in the central United States were deposited during the last glacial period, when silt production rates from the LIS and from midcontinent nonglacial sources were extremely high. Thus, even where nonglacial sources were important, such as the Great Plains, loess deposition took place during the last glacial period. Equally significant, at this time vegetation conditions were also suitable for accumulation of loess in the central United States, with boreal forest present in the Central Lowland province (Baker et al., 1986, 1989) and parkland in the Great Plains province (Wells and Stewart, 1987); both have a high

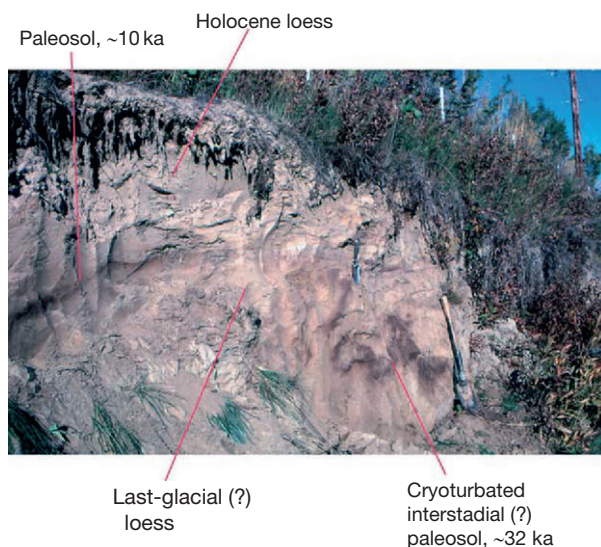


Figure 10 Photograph of Holocene loess and underlying paleosol, possible last glacial loess, and highly cryoturbated, probable mid-Wisconsin paleosol at the Chena Hot Springs Road loess exposure, near Fairbanks Alaska (see [Muhs et al., 2003](#), for details). Photograph by D.R. Muhs.

dust-trapping potential. The importance of both production rates and the potential for accumulation in the development of loess deposits is illustrated for the central United States by consideration of the Holocene loess deposits of this region. In the Central Lowland province, loess accumulation potential is high because of the vegetation cover, but the production rate is extremely low; hence there is little or no Holocene loess preserved in the deposits of this area, although there may be small amounts of loess incorporated into the modern soils ([Mason and Jacobs, 1998](#)). Holocene loess production rates were somewhat higher in the Great Plains province, however, the accumulation potential was typically low, giving rise to modest Holocene (Bignell) loess deposits in this region ([Mason et al., 2003](#); [Muhs et al., 2003](#)) accumulating only where more effective dust-trapping topography and/or vegetation cover exists ([Mason et al., 1999](#)).

The timing of the deposition of loess in some areas of the Columbia Plateau province is inferred to be different from that for other loess deposits of North America and, indeed, the world ([Busacca et al., 2004](#)). Loess deposition rates are greatest in the Palouse region of eastern Washington. Here, the principal source of the loess is believed to be the flood sediments of proglacial Lake Missoula; the production and deposition rates of these loess deposits are therefore believed to peak late in the glacial cycle. In contrast to other locations, therefore, in the Palouse area pedogenesis occurred during glacial periods and stadials, while loess accumulation began toward the end of these cold periods and continued during interglacial periods and interstadials when the ice sheets retreated and the glacial outburst flood sediment sources were available ([McDonald and Busacca, 1998](#)). Thus, unlike the Great Plains region, loess in the Palouse region has a glaciogenic source, but the timing of deposition is not during the glacial maximum.

Geochemical tracing of the late Pleistocene Palouse loess demonstrates that it is derived from the glacial outburst flood

‘slackwater’ deposits ([Sweeney et al., 2002](#)). Further evidence that the slackwater deposits are the source of the Palouse loess is provided by the thinning of loess deposits with increasing distance from the slackwater deposits of south-central Washington and north-central Oregon ([Busacca and McDonald, 1994](#)). The loess deposits of the Snake River Plain are believed to be derived from the alluvial deposits of the Snake River as the deposits thin with increasing distance to the east and southeast of the river ([Busacca et al., 2004](#)). In contrast with the Palouse loess, the loess of the Snake River Plain is probably associated with deposition during glacial times and the paleosols are associated with warmer periods, in the same way as observed for the central United States and for many other locations worldwide, although more chronological studies are needed for this area.

The mineralogy and the major and trace element geochemistry of central Alaskan loess is distinctive from that of all other North American loess deposits ([Muhs et al., 2003](#)); it has unusually high iron oxide (Fe_2O_3) and aluminum oxide (Al_2O_3) contents, in spite of having a low clay content, and it also contains extremely low concentrations of carbonates. Alaskan loess is dominantly glaciogenic, with silt production being directly linked to glacial activity ([Muhs et al., 2003](#)). [Hallet et al. \(1996\)](#) demonstrated an order of magnitude increase in suspended sediment yields for drainage basins with more than 30% glacier cover compared to unglaciated Alaskan drainage basins. Clearly, in the past, silt production would have been expected to vary as ice cover varied. However, Alaska demonstrates the situation where conflicting conditions occur between maximizing silt production and maximizing the accumulation of loess. Contrary to the situation observed elsewhere in North America, [Muhs et al. \(2003\)](#) report that only moderate accumulation of loess occurred during the last glacial period in Alaska. Furthermore, extensive deposits of Holocene loess are found in Alaska, which in many locations are as thick or thicker than any potential last glacial deposits ([Muhs and Zárate, 2001](#)), challenging the classical view that the periods of maximum loess accumulation in Alaska correspond to glacial times.

A model of loess stratigraphic development in North America outlined by [Muhs et al. \(2003\)](#) proposes that there is a balance between factors influencing production and those affecting accumulation of loess. During times of potential maximum loess production in Alaska, that is, when the ice sheets and glaciers were most extensive, conditions for loess accumulation are at their least favorable because the herb-tundra vegetation which persisted during these cold conditions was an ineffective loess trap ([Begét, 1988](#)). This offers a potential explanation for the rather modest last glacial loess deposits observed in Alaska, even though loess production rates were probably extremely high ([Muhs et al., 2003](#)). Holocene loess production rates were not as high as those during glacial times, but the Holocene vegetation has been dominantly boreal forest whose much higher loess trapping potential has fostered the abundant thick Holocene loess deposits observed in Alaska ([Muhs et al., 2003](#)).

Paleoclimatic Implications

Due to their eolian transport mechanism, loess deposits provide one of the few direct records of atmospheric circulation,

preserving information regarding paleowinds. They can also suggest the past moisture balance of a region at the time of formation, and the amount of past vegetation cover (Muhs and Zárte, 2001). They also preserve floral and faunal paleoenvironmental indicators, along with other paleoclimatic proxies. Furthermore, they provide a record of the dust flux over time which is of increasing interest for models of climate change.

In the central United States, trends in Peoria Loess thickness (Figure 2), particle size, and also geochemistry, all indicate that the dominant direction of paleowinds was from the west or northwest (Mason, 2001; Muhs and Bettis, 2000). This finding is in direct contradiction with paleowind simulations from some atmospheric general circulation models (AGCMs) which predict prevailing easterly or northeasterly paleowinds resulting from a strong glacial anticyclone generated over the LIS during the Last Glacial Maximum (LGM; COHMAP Members, 1988). Similarly, climate model predictions of paleowind directions for the Pacific northwestern United States and Alaska and the neighboring Yukon Territory are also in disagreement with the observations from loess deposits. In contrast, however, recent observations of decreased loess accumulation rates across the Columbia Plateau during the LGM, arising due to suppression of the prevailing southwesterly dust-transporting winds, have been proposed as evidence supporting the modeled easterly wind anomaly south of the Cordilleran Ice Sheet predicted by AGCMs (Sweeney et al., 2004). The reasons for the discrepancies noted between modeled output and the geomorphologic evidence from various regions are not presently understood, and further investigation is required to resolve the inconsistencies that remain.

Interestingly, model simulations do not identify the Great Plains as being a significant source of dust during the last glacial period (Kohfeld and Harrison, 2000; Mahowald et al., 1999), yet the thickest last glacial loess deposits in the world are found here. The presence of nonglaciogenic loess in the Great Plains (Aleinikoff et al., 1998, 1999) indicates that factors other than simply glacial silt production are important for explaining loess origins during glacial periods. Possible factors of paleoclimatic significance include a sparse vegetation cover (e.g., a cool steppe vegetation) reflecting drier glacial climates (leading to the subsequent erosion of nonglacially derived silts and clays, for example, the White River Group and the Pierre Shale), lower temperatures, and stronger winds (e.g., Muhs et al., 1999). The loess of the Palouse region of the Columbia Plateau demonstrates that the accumulation of loess is not necessarily associated primarily with peak glacial activity; here, loess accumulation reaches its peak during Late Glacial and Holocene times when the supply of material is at its greatest.

The loess deposits of North America often contain proxy data for paleoclimatic conditions. The study of land snail species found in loess deposits, for instance, can yield important information regarding paleotemperature and paleoprecipitation, in addition to the vegetation type. For example, the discovery of extralimital northern or Cordilleran land snails in the Peoria Loess deposits of the Great Plains and Central Lowland indicates that colder conditions prevailed here in Late Glacial times (Rousseau and Kukla, 1994). Similarly, success has also been achieved in obtaining important information regarding the occurrence of forest or cool grassland

(‘C3’ plant communities) or of warm grassland (‘C4’ communities) from the carbon and oxygen isotope compositions of loess and minimally developed intercalated paleosols (e.g., Wang et al., 2000).

Loess deposits also provide a direct record of atmospheric dust flux, which may be used to assess the role of dust in climate change. The application of luminescence techniques to date the sediments that make up the loess deposits (rather than using radiocarbon dating of the paleosols that bracket the loess) has revealed that the accumulation rate varies over time. The high rates of loess deposition found in midcontinental North America (Bettis et al., 2003; Roberts et al., 2003) may have important implications for far-field (i.e., distant downwind) areas in terms of radiative and biogeochemical impacts. In spite of the observation that loess deposition decreases rapidly with increasing distance from the source, the fine-grained component of the loess may travel much further; this fine-grained component is more effective than coarser material in inducing radiative backscatter (Harrison et al., 2001), resulting in potential modification of the radiation budget even at distal locations where more modest loess thicknesses are found. Furthermore, fine-grained, far-traveled dust can also alter the atmospheric chemistry and affect the major biogeochemical cycles by providing an important source of nutrients to marine ecosystems (Harrison et al., 2001). Significant contributions of fine-grained (and little weathered) material to surface soils downwind of loess source areas in North America have also been reported (e.g., Mason and Jacobs, 1998).

While North American loess is a less significant present-day dust source than many other areas in terms of its potential effect on global climate (Kohfeld and Harrison, 2001), recent work suggests that it may have influenced past climate (Roberts et al., 2003). The role of dust in climate change is not well understood (Harrison and Kohfeld, 2001) but further study may prove it to be a significant factor, both for past and future climate change.

See also: Dating Techniques; Loess Deposits: Origins and Properties. **Glacial Climates:** Effects of Atmospheric Dust. **Glaciations:** Late Quaternary in North America. **Introduction:** History of Dating Methods. **Paleoceanography, Physical and Chemical Proxies:** Oxygen-Isotope Stratigraphy of the Oceans. **Paleoclimate:** Introduction. **Paleoclimate Modeling:** Quaternary Environments; **Paleosols and Wind-Blown Sediments:** Mineral Magnetic Analysis; Overview. **Pollen Methods and Studies:** Use of Pollen as Climate Proxies.

References

- Aleinikoff JN, Muhs DR, and Fanning CM (1998) Isotopic evidence for the sources of late Wisconsin (Peoria) loess, Colorado and Nebraska: Implications for paleoclimate. In: Busacca AJ (ed.) *Dust Aerosols, Loess Soils and Global Change*, pp. 124–127. Pullman, WA: Washington State University College of Agriculture and Home Economics Miscellaneous Publication MISC0190.
- Aleinikoff JN, Muhs DR, Sauer R, and Fanning C (1999) Late Quaternary loess in northeastern Colorado, Part II – Pb isotopic evidence for the variability of loess sources. *Geological Society of America Bulletin* 111: 1876–1883.
- Baker RG, Rhodes RS II, Schwert DP, et al. (1986) A full-glacial biota from southeastern Iowa, USA. *Journal of Quaternary Science* 1: 91–107.

- Baker RG, Sullivan AE, Hallberg GR, and Horton DG (1989) Vegetational changes in western Illinois during the onset of late Wisconsinan glaciation. *Ecology* 70: 1363–1376.
- Begét JE (1988) Tephra and sedimentology of frozen loess. In: Senneset K (ed.) *Fifth International Permafrost Conference Proceedings*, vol. 1, pp. 672–677. Trondheim: Tapir.
- Begét JE and Hawkins DB (1989) Influence of orbital parameters on Pleistocene loess deposition in central Alaska. *Nature* 337: 151–153.
- Bettis EA III, Muhs DR, Roberts HM, and Wintle AG (2003) Last Glacial loess in the conterminous USA. *Quaternary Science Reviews* 22: 1907–1946.
- Busacca AJ (1991) Loess deposits and soils of the Palouse and vicinity. In: Morrison RB (ed.) *The Geology of North America, vol. K-2, Quaternary Nonglacial Geology*, pp. 216–228. Boulder, CO: Conterminous US Geological Society of America.
- Busacca AJ, Béget JE, Markewich HE, Muhs DR, Lancaster N, and Sweeney MR (2004) Eolian sediments. In: Gillespie AR, Porter SC, and Atwater BF (eds.) *The Quaternary Period in the United States*, pp. 275–309. Amsterdam: Elsevier.
- Busacca AJ and McDonald EV (1994) Regional sedimentation of late Quaternary loess on the Columbia Plateau: Sediment source areas and loess distribution patterns, regional geology of Washington State. *Washington Division of Geology and Earth Resources Bulletin* 80: 181–190.
- COHMAP Members (1988) Climatic changes of the last 18,000 years: Observations and model simulations. *Science* 241: 1043–1052.
- Flint RF (1971) *Glacial and Quaternary Geology*. New York: Wiley.
- Forman SL, Bettis EA, Kemmis TJ, and Miller BB (1992) Chronologic evidence for multiple periods of loess deposition during the late Pleistocene in the Missouri and Mississippi River valley, United States; implications for the activity of the Laurentide ice sheet. *Palaeogeography, Palaeoclimatology, Palaeoecology* 93: 71–83.
- Forman SL, Smith RP, Hackett WR, Tullis JA, and McDaniel PA (1993) Timing of late Quaternary glaciations in the western United States based on the age of loess on the eastern Snake River Plain, Idaho. *Quaternary Research* 40: 30–37.
- Grimley DA (2000) Glacial and nonglacial sediment contributions to Wisconsin Episode loess in the central United States. *GSA Bulletin* 112: 1475–1495.
- Grimley DA, Follmer LR, Hughes RE, and Solheid PA (2003) Modern, Sangamon and Yarmouth soil development in loess of unglaciated southwestern Illinois. *Quaternary Science Reviews* 22: 225–244.
- Hallet B, Hunter L, and Bogen J (1996) Rates of erosion and sediment evacuation by glaciers: A review of field data and their implications. *Global and Planetary Change* 12: 213–235.
- Harrison S, Kohfeld KE, Roelandt C, and Claquin T (2001) The role of dust in climate changes today, at the last glacial maximum and in the future. *Earth-Science Reviews* 54: 43–80.
- Kohfeld KE and Harrison SP (2000) How well can we simulate past climates? Evaluating the models using global palaeoenvironmental datasets. *Quaternary Science Reviews* 19: 321–346.
- Kuntz MA, Spiker EC, Rubin M, Champion DE, and Lefevre RH (1986) Radiocarbon studies of latest Pleistocene and Holocene lava flows of the Snake River Plain, Idaho: Data, lessons, interpretations. *Quaternary Research* 25: 163–176.
- Lewis GC, Fosberg MA, McDole RE, and Chugg JC (1975) Distribution and some properties of loess in south-central and south-eastern Idaho. *Soil Science Society of America Proceedings* 39: 1165–1168.
- Lindholm GF, Thomas LA, Davidson DT, Handy RL, and Roy CJ (1959) Silt near Big Delta and Fairbanks. In: Davidson DT, et al. (ed.) *The Geology and Engineering Characteristics of Some Alaskan Soils, Iowa State University Bulletin*, vol. 186, pp. 33–70.
- Maat PB and Johnson WC (1996) Thermoluminescence and new ^{14}C age estimates for late Quaternary loesses in southwestern Nebraska. *Geomorphology* 17: 115–128.
- Mahowald N, Kohfeld K, Hansson M, et al. (1999) Dust sources and deposition during the last glacial maximum and current climate: A comparison of model results with paleodata from ice cores and marine sediments. *Journal of Geophysical Research* 104: 15895–15916.
- Markewich HW, Wysocki DA, Pavich MJ, et al. (1998) Paleopedology plus TL, ^{10}Be , and ^{14}C dating as tools in stratigraphic and paleoclimatic investigations, Mississippi River Valley, USA. *Quaternary International* 51/52: 143–167.
- Mason JA (2001) Transport direction of Peoria loess in Nebraska and implications for loess sources on the central Great Plains. *Quaternary Research* 56: 79–86.
- Mason JA, Bettis EA III, Roberts HM, Muhs DR, and Joekel RM (2004) Last glacial loess sedimentary system of eastern Nebraska and western Iowa. In: Mandel RM (ed.) *Guidebook for Fieldtrips*. 18th Biennial Meeting of the American Quaternary Association, pp. 1–28. Lawrence, KS: Kansas Geological Survey. Open File Report 2004-33.
- Mason JA and Jacobs PW (1998) Chemical and particle size evidence for addition of fine dust to soils of the midwestern United States. *Geology* 26: 1135–1138.
- Mason JA, Jacobs PM, Hanson PR, Miao XD, and Goble RJ (2003) Sources and paleoclimatic significance of Holocene Bignell Loess, central Great Plains. *Quaternary Research* 60: 330–339.
- Mason JA, Nater EA, Zanner CW, and Bell JC (1999) A new model of topographic effects on the distribution of loess. *Geomorphology* 28: 223–236.
- McDole RE, Lewis GC, and Fosberg MA (1973) Identification of paleosols and the Fort Hall Geosol in southeastern Idaho loess deposits. *Soil Science Society of America Proceedings* 37: 611–616.
- McDonald EV and Busacca AJ (1992) Late Quaternary stratigraphy of loess in the Channeled Scabland and Palouse regions of Washington State. *Quaternary Research* 38: 141–156.
- McDonald EV and Busacca AJ (1998) Unusual timing of regional loess sedimentation triggered by glacial outburst flooding in the Pacific northwest US. In: Busacca AJ (ed.) *Dust Aerosols, Loess Soils, and Global Change*, pp. 163–166. Pullman: Washington State University College of Agriculture and Home Economics Miscellaneous Publication #MISC0190.
- Muhs DR, Ager TA, Bettis EA III, et al. (2003) Stratigraphy and palaeoclimatic significance of late Quaternary loess-paleosol sequences of the last interglacial-glacial cycle in central Alaska. *Quaternary Science Reviews* 22: 1947–1986.
- Muhs DR, Aleinikoff JN, Stafford TW Jr., et al. (1999) Late Quaternary loess in northeastern Colorado, Part I – Age and paleoclimatic significance. *Geological Society of America Bulletin* 111: 1861–1875.
- Muhs DR and Bettis EA III (2000) Geochemical variations in Peoria Loess of western Iowa indicate paleowinds of midcontinental North America during last glaciation. *Quaternary Research* 53: 49–61.
- Muhs DR, McGeehin JP, Beann J, and Fisher E (2004) Holocene loess deposition and soil formation as competing processes, Matanuska Valley, southern Alaska. *Quaternary Research* 61: 265–276.
- Muhs DR and Zárte M (2001) Late Quaternary eolian records of the Americas and their paleoclimatic significance. In: Markgraf V (ed.) *Interhemispheric Climate Linkages*, pp. 183–216. San Diego: Academic Press.
- Péwé TL (1955) Origin of the upland silt near Fairbanks, Alaska. *Geological Society of America Bulletin* 66: 699–724.
- Péwé TL (1975) *Quaternary Geology of Alaska*. Geological Survey Professional Paper, p. 835.
- Pierce KL, Fosberg MA, Scott WE, Lewis GC, and Colman SM (1982) Loess deposits of southeastern Idaho: Age and correlation of the upper two loess units. In: Bonnicksen B and Breckinridge RM (eds.) *Cenozoic Geology of Idaho, Idaho Bureau of Mines and Geology Bulletin*, vol. 26, pp. 712–725.
- Pye K (1987) *Aeolian Dust and Dust Deposits*. Orlando, FL: Academic Press.
- Pye K (1996) The nature, origin and accumulation of loess. *Quaternary Science Reviews* 14: 653–667.
- Richardson CA, McDonald EV, and Busacca AJ (1997) Luminescence dating of loess from the northwest United States. *Quaternary Science Reviews* 16: 403–415.
- Roberts HM, Muhs DR, Wintle AG, Duller GAT, and Bettis EA III (2003) Unprecedented last glacial mass accumulation rates determined by luminescence dating of loess from western Nebraska. *Quaternary Research* 59: 411–419.
- Rousseau DD and Kukla G (1994) Late Pleistocene climate record in the Eustis loess section, Nebraska, based on land snail assemblages and magnetic susceptibility. *Quaternary Research* 42: 176–187.
- Ruhe RV (1983) Depositional environment of late Wisconsin loess in the midcontinental United States. In: Wright HE Jr. and Porter SC (eds.) *Late-Quaternary Environments of the United States, the Late Pleistocene*, vol. 1, pp. 130–137. Minnesota: University of Minnesota Press.
- Ruhe RV, Miller GA, and Vreeken WJ (1971) Paleosols, loess sedimentation and soil stratigraphy. In: Yaalon DH (ed.) *Paleopedology – Origin, Nature and Dating of Paleosols*, pp. 41–59. Jerusalem: Israel Universities Press.
- Sweeney MR, Busacca AJ, Gaylord DA, and Zender C (2002) Provenance of Palouse loess and relation to late Pleistocene glacial outburst flooding, Washington State. *EOS Transactions American Geophysical Union* 83: 47 Abstract H22B-0899.
- Sweeney MR, Busacca AJ, Richardson CA, Blinnikov M, and McDonald EV (2004) Glacial anticyclone recorded in Palouse loess of northwestern United States. *Geology* 32: 705–708.
- Wang H, Follmer LR, and Liu JC (2000) Isotope evidence of paleo-El Niño-Southern Oscillation cycles in loess-paleosol record in the central United States. *Geology* 28: 771–774.
- Wells PV and Stewart JD (1987) Cordilleran-boreal taiga and fauna on the Central Great Plains of North America, 14,000–18,000 years ago. *American Midland Naturalist* 118: 94–106.
- Westgate J (1988) Isothermal plateau fission-track age of the Late Pleistocene Old Crow tephra, Alaska. *Geophysical Research Letters* 15: 376–379.
- Westgate JA, Stemper BA, and Péwé TL (1990) A 3 m.y. record of Pliocene-Pleistocene loess in interior Alaska. *Geology* 18: 858–861.

South America

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Introduction

The loess record of southern South America, the largest of the Southern Hemisphere, was originally reported in the nineteenth century under different names, among others by Charles Darwin in 1844. Its resemblance with the loess deposits of the Rhine valley was first noticed in 1866. Since then the term 'loess' has been loosely used in the literature to refer to the vast mantle of late Tertiary and Quaternary sediments of the Chaco-Pampean plain.

Loess-like deposits, loessoid deposits, reworked loess, loessic deposits, and secondary loess have been used to designate fine-grained, light-colored and massive sediments with the appearance of loess. Generally, loess-like and loessoid deposits are the terms most commonly found in the literature. Neotropical loess has been recently used to designate primary and reworked sediments of loessic character, distributed between latitudes 20° and 30° S. Also, the term tropical loess has been proposed to account for fine-grained surface deposits of massive appearance in northeastern Argentina, areas of southern Brazil, and northern Uruguay. These sediments, traditionally interpreted as *in situ* weathered byproducts from the underlying late Jurassic basaltic rocks, have been recently interpreted as a wide and extensive loess mantle that accumulated during the last glacial cycle. This hypothesis generated controversy and is presently under debate. Ongoing geochemical, mineralogical, and pedological analysis reinforce the *in situ* origin of the surface deposits.

During most of the twentieth century, research on South American loess was fragmentary and discontinuous. With few exceptions, a great number of contributions were focused on the stratigraphy and the fossil vertebrate content. A renewed interest began in the late 1980s specially driven by the necessity of testing global paleoclimatic models. Numerous studies were conducted particularly across the Pampean plain. The more detailed mapping, magnetostratigraphy, dating, paleopedological studies, and compositional analysis have substantially increased the understanding of the record. In recent years, detailed analysis substantially increased the knowledge of the loess–paleosol record of the mountain valley loess of Tucumán in northwestern Argentina. The Chaco loess is less known than the other two loess records, although regional aspects of its composition and geographical distribution have been reported.

Geographical Distribution

Loess distribution has always been associated with the extensive and heterogeneous Chaco-Pampean plain of southern South America, a region over 1 million km², covering most of northern central Argentina, western Uruguay, southern Brazil, western Paraguay, and southeastern Bolivia (Figure 1). This

generalization probably comes from the geographic distribution of loessoid deposits in Argentina and was later adopted in other studies, including not only loess but also sand dunes, sand mantles, and fine-grained deposits of noneolian origin. Subsequently, more detailed mapping across the Chaco-Pampean plain (Table 1) indicated a more restricted geographic distribution of loess; extensive areas of central Argentina are covered by eolian sands, while alluvial deposits are common in the northeast.

The Pampean Loess and Loess-Like Record

General Characteristics

The exposures of Pampean sediments are scarce and limited to quarries, road cuts, and river banks. The exceptions are the Chapadmalal sea cliffs near Mar del Plata and the Paraná River banks, 20–30 m thick and laterally continuous for several kilometers. Both of these unusually well-exposed sections have been the subject of numerous studies (Figure 2).

The Pampean sequences do not represent a continuous sedimentary record; stratigraphic discontinuities are indicated by several erosional unconformities while the rate and characteristics of the sediment supply varied through time. Loess is a secondary and minor component of the sedimentary sequences, being prevalent only during the last glacial cycle. Instead, loess-like deposits generated by the reworking of primary eolian sediments are dominant. Running water, mass-wasting processes on hill slopes, and deposition in swampy and lacustrine environments are thought to have been involved in the retransportation and resedimentation of primary loess. The reworking by aqueous transport agents resulted in several loess-like deposits including siltstones, clay siltstones, and sandy siltstones. They usually exhibit a general massive appearance or have poorly defined sedimentary structures frequently masked by pedogenesis and diagenesis.

Paleosols

The occurrence of pedogenic features is ubiquitous throughout the stratigraphic sequence of loess and loess-like deposits (Figure 3). The recognition and classification of paleosol soil horizons present several difficulties in field exposures. Paleosols were originally interpreted as truncated profiles with A horizons eroded by eolian and/or fluvial processes. Their identification has been mostly based on field morphological properties.

Detailed micromorphological analysis carried out at some stratigraphic intervals of the northern and southern Pampas show that pedogenesis was active throughout the sedimentation process resulting in accretionary soil profiles. Welding of pedological features on preexisting soil surfaces gives way to pedocomplexes. In addition, paleosols are not always discrete

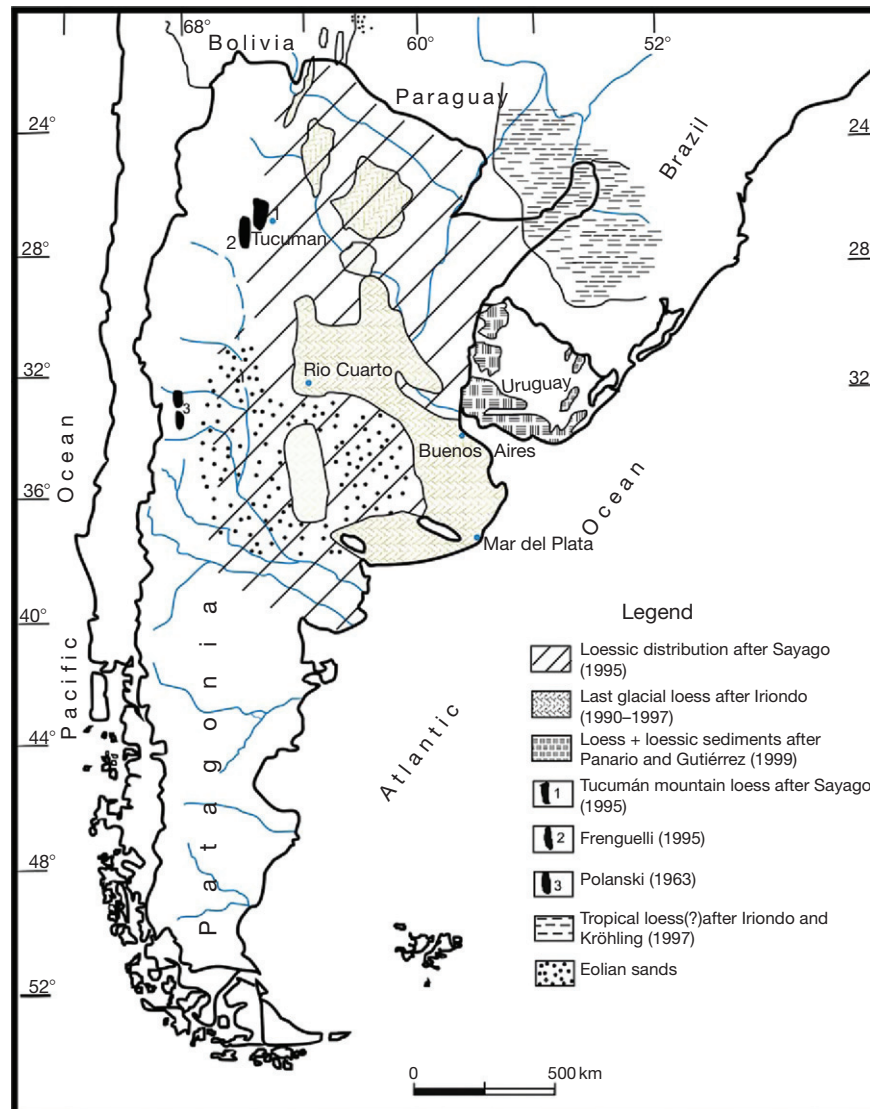


Figure 1 Loess and loessoid distribution following several authors. Adapted from Frenguelli J (1955) *Loess y Limos pampeanos*. Ministerio de Educación de la Nación. Serie Técnica y Didáctica. La Plata, No. 7; Iriondo M (1990) *Map of the South American plains – Its present state. Quaternary of South America and Antarctic Peninsula* 6: 297–308; Panario D and Gutiérrez O (1999) *The continental Uruguayan Cenozoic: An overview. Quaternary International* 62: 75–84; Polanski J (1963) *Estratigrafía, Neotectónica y Geomorfología del Pleistoceno pedemontano entre los ríos Diamante y Mendoza. Asociación Geológica Argentina Revista XVII*: 127–349; Zárate M (2003) *The loess record of Southern South America. Quaternary Science Reviews* 22: 1987–2006; Zárate M and Blasi A (1993) *late Pleistocene-Holocene eolian deposits of the southern Buenos Aires Province, Argentina: A preliminary model. Quaternary International* 17: 15–20.

entities, as their boundaries are sometimes obliterated by bioturbation, aqueous reworking, and grain size variation. Micromorphology also suggests that the balance between pedogenesis and sedimentation varied cyclically over time, with alternating phases of pedogenesis and intervals of much higher sedimentation rate. This pattern resulted in the identification of pedosedimentary cycles, thought to be ultimately driven by climatic changes.

Carbonate Accumulations

Carbonate accumulations, known locally as ‘tosca,’ constitute a very common feature of the loess and loess-like deposits of

the southern and northern Pampas. Their origin has been attributed to both pedogenesis and groundwater processes. Carbonate accumulations exhibit very diverse morphologies including carbonate-cemented sedimentary layers, nodules, and concretions ranging from millimeters to half a meter in size, as well as veins (Figure 4). Carbonate rhizoconcretions are also present, associated with paleosol horizons.

Calcrete crusts are another common and characteristic feature of the southern Pampas sequences. In addition, crusts are developed on the surfaces of late Miocene and Pleistocene loess-like deposits. They exhibit complex morphologies, resulting from multiple episodes of brecciation, dissolution, and reprecipitation. The thickness varies from a few centimeters

Table 1 Summarized characteristics of the loess record in the Pampas, Chaco and the mountain valleys of Tucumán

Area	General knowledge	Sedimentary record	Stratigraphic record
Pampean loess record	Numerous studies on stratigraphy, vertebrate fossil remains, paleopedology, magnetostratigraphy, and geochronology: several numerical dates by ^{14}C , TL, OSL, Ar/Ar	Discontinuous sedimentary record Loessoid deposits dominant component Loess prevalent during last glaciation forming a continuous apron	late Miocene (~10 Ma) to Holocene record
Chaco loess record	Not known in detail. Information available on areal distribution and partially on composition, very few dates (^{14}C)	Loess covering several areas of the plain	Last Glacial–Holocene loess
Northwest Argentina, Tucumán Mountain valley loess record	Studies performed since the 1980s; several recent papers on pedostratigraphy, micromorphology, ^{14}C , and OSL dates, geochemistry, magnetostratigraphy	Continuous stratigraphic loess–paleosol record. High resolution	Mid-Pleistocene to Holocene loess–paleosol record



Figure 2 Exposures of the Pampean loess and loess-like record. (a) Chapadmalal sea cliffs south of Mar del Plata. (b) late Miocene and Pliocene loess deposits at a river bank in Ventania range. (c) Gorina quarry in La Plata. (d) Outcrop along the left margin of the Paraná River in Baradero, northern Pampas.

to more than 2 m, depending on the relative age of the geomorphic surfaces where they are present. No numerical ages have been obtained.

Vertebrate Fossil Content

The abundant vertebrate fossil remains recovered from the late Cenozoic loess and loess-like deposits have been the subject

of numerous studies, dealing particularly with their taxonomy and biostratigraphic implications. Fossil assemblages consist of autochthonous endemic groups (including edentates, litopterns, notoungulates, and marsupials) and Holarctic immigrants (including carnivores, cervids, camelids, and horses) which gradually migrated to South America at different stages. A major faunal turnover, whose origin is still debated, is recorded in the late Pliocene loess-like deposits of Chapadmalal.



Figure 3 Field exposures of paleosols. (a) Truncated paleosol of the mid-Pleistocene developed on loess deposits, Gorina quarry, near La Plata. (b) late Pliocene loess–paleosol sequence at Chapadmalal, Bt horizons more eroded and with darker colors.

Fossil assemblages have been grouped into different land mammal ages or stages–ages based on their relative degree of evolution (Table 2). All the type localities of the late Miocene to Pleistocene stage–ages (except for the Huayquerian, situated in the Andean piedmont), are located in the southern and northern Pampas of Buenos Aires province.

In addition to fossil bones, loess and loess-like deposits exhibit striking mammal bioturbation features represented by burrows and caves of variable sizes (Figure 5). The smaller burrows, with diameters of less than 20 cm, were formed by fossil rodents. Large burrows, with diameters between 0.80 and 1.80 m, are also common. Their morphological characteristics and the occurrence of claw marks on the walls and roofs, along with biomechanical analysis of fossil limbs, suggest that the mylodontid sloths *Scelidotherium* and *Glossotherium* were the probable builders of these burrows.

Prior to the initiation of magnetostratigraphic studies, fossil assemblages were the only tools available to assign relative ages and correlate isolated and distant loess sections across the Pampean plain. This resulted in a regional chronostratigraphic scheme of the loess and loess-like record.

Impact Glasses

A remarkable feature of the loess and loess-like deposits of South America is the occurrence of vesicular and glassy fragments, ranging in size from a few millimeters up to clusters 1–2 m in diameter. These fragments, known as ‘escorias’ are found at localities of the southern Pampas and in Río Cuarto (northern Pampas; Figure 6). In stratigraphic sections, they form horizons characterized by a relatively high concentration of fragments of different sizes and can be traced laterally for several hundreds of meters or even kilometers at some localities.

Reworked escorias fragments from these main horizons are commonly embedded in younger fluvial deposits; some secondary reworking is due to vertebrate bioturbation. Their genesis, debated for decades, had been previously attributed to natural fires, volcanic and igneous activity, and even *in situ* diagenetic processes. However, recent studies of the geochemical and petrographic character of escorias show clear signatures of an impact origin. High-resolution argon-40/argon-39 ($^{40}\text{Ar}/^{39}\text{Ar}$) dating of the fresher-appearing glass indicates the occurrence of five impact events in the southern Pampas recorded at sections cropping out in Chapadmalal, Centinela del Mar (two events), Bahía Blanca, and Chasicó (Table 3). In Río Cuarto, impact glasses have been found scattered on exhumed surfaces with relative age, fission track, and $^{40}\text{Ar}/^{39}\text{Ar}$ dating indicating the presence of at least two additional Holocene and Pleistocene impact events.

Magnetostratigraphy

Magnetostratigraphic analysis has been performed on several sections of loess and loess-like deposits of the northern Pampas in the vicinity of Buenos Aires and the Chapadmalal sea cliffs in the southern Pampas (Table 4).

When no numerical ages are available for chronology the magnetostratigraphic interpretations, however, are constrained by the stratigraphic discontinuity and the common lateral facial changes. As a result, several alternative interpretations have been proposed at some sections. The magnetostratigraphy of Chapadmalal suggests an early Gauss-to-Gilbert magnetic age (>3.15 Ma) for the lowermost levels supported by later studies indicating ages close to 4 Ma. The sequence is completed by a discontinuous record of loess-like deposits accumulated during the Gauss–Matuyama and Brunhes magnetic ages.

In the northern Pampas, the paleomagnetic analysis suggests a Matuyama age younger than 2 Ma for the oldest levels cropping out at the bottom of several quarries. The Brunhes–Matuyama boundary is identified at several sections at a variable depth of 5–10 m from the surface. The onset of the Brunhes chron is thought to be characterized by a new pulse of loessic sedimentation in coincidence with an increase in the Andean volcanic activity and a change to drier and colder climatic conditions.

Magnetic Susceptibility

Near the city of Buenos Aires in the northern Pampas, low susceptibility is recorded in paleosols and higher susceptibility in loess layers. This contrasts with the loess–paleosol record of

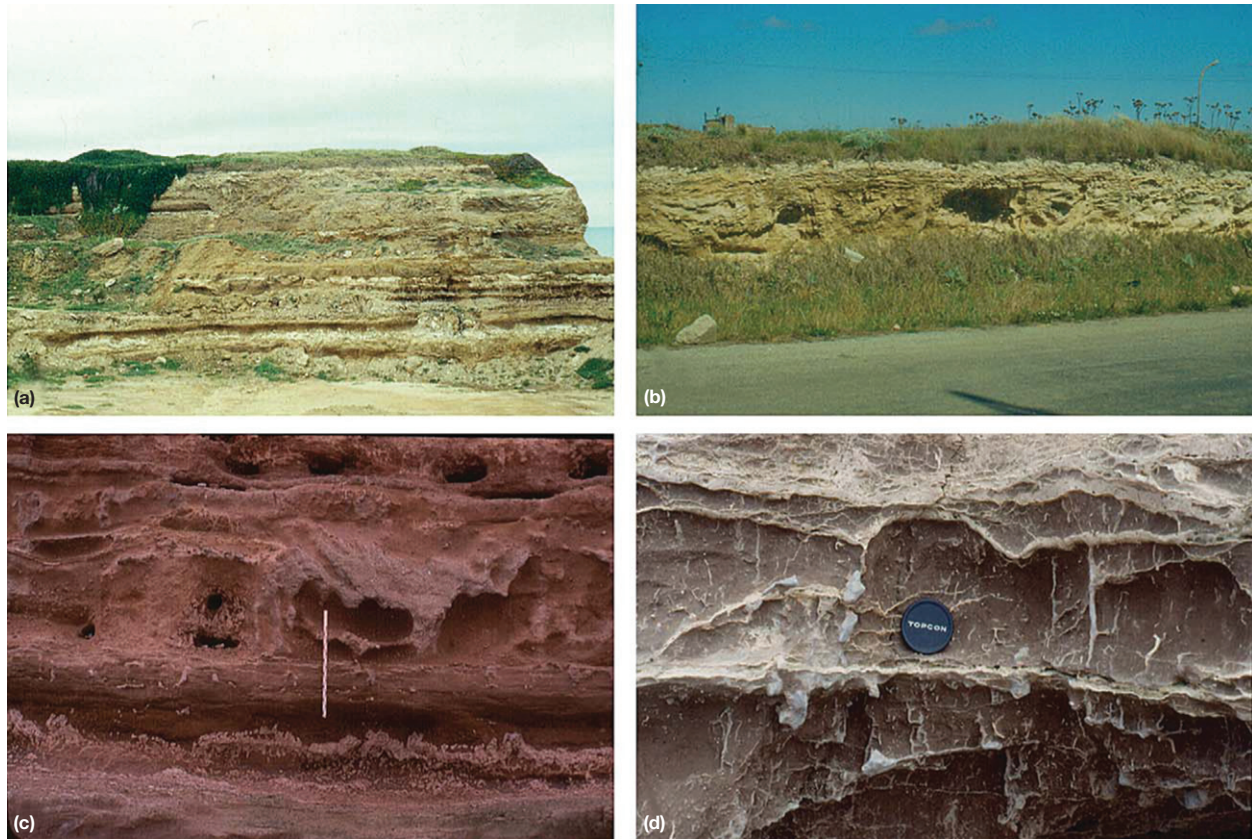


Figure 4 Morphologies of carbonate accumulations. (a) Calcrete crusts in late Pliocene and Pleistocene loess-like deposits at a Chapadmalal quarry. (b) Calcrete crust on top of Pliocene loess deposits in Tandilia. (c) Veins, nodular and massive calcium carbonate layers eroded by paleochannels, early Pleistocene loess-like deposits at Chapadmalal. (d) Rhizoconcretions and reticulate pattern of carbonate stringers associated with a paleosol level, Chapadmalal.

Table 2 Stages–ages (land mammal ages) and type localities in the Pampean region (adapted from Cione and Tonni (2001), and references therein)

Area	Type locality	Stages–ages (land mammal ages)	Sedimentary deposits
Northern Pampas of Buenos Aires province	Stratigraphic sections located in the surroundings of Buenos Aires city and La Plata	Platan	Lacustrine and fluvial facies – loess
		Lujanian	Fluvial deposits – loess
		Bonaerian	Loess–loessoid deposits
Southern Pampas of Buenos Aires province	Chapadmalal sea cliffs	Ensenadan	Loessoid deposits of paludal and fluvial environments
		Marplatan	Loessoid deposits of fluvial environments
	Chapadmalal sea cliffs	Chapadmalan	Loessoid deposits of hill slopes and of fluvial environments
	Monte Hermoso	Montehermosan	Loessoid deposits of fluvial environments
	Arroyo Chasicó in southern Buenos Aires	Chasicooan	Very fine sandstones and loessoid deposits siltstones of fluvial and paludal environments

China, but is similar to what has been reported for Alaska. Magnetic susceptibility changes have been interpreted as evidence of alternating periods of wet and dry conditions. Low susceptibility values are reported in calcrete layers and reduced sediments and high values in the B horizons. The magnetic

mineralogy is dominated by multidomain magnetite, behavior attributable to fine silt-sized particles most likely of detrital origin. Diagenetic processes and past climatic conditions have been considered as possible controlling factors of magnetic susceptibility. The magnetic signals of late Pleistocene and

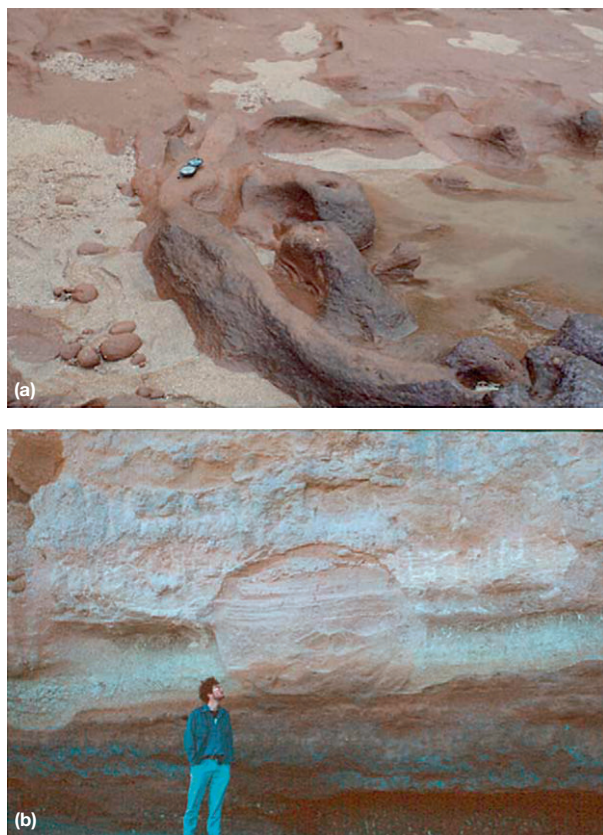


Figure 5 Bioturbation structures. (a) Rodent burrow in Pliocene loess at the Chapadmalal sea cliffs. (b) Cave excavated in loess-like deposits north of Mar del Plata.



Figure 6 Fragment of impact glass (escoria) found in late Pliocene loessoid deposits of the southern Pampas.

Holocene alluvial and eolian deposits at two Pampean fluvial environments of Buenos Aires suggest that the genesis of a high-coercive fraction in loessic sediments is tentatively associated with dissolution. Pedogenesis is considered the cause of magnetic signal variations in paleosols showing extensive dissolution of detrital ferrimagnetic minerals. High-coercivity minerals are thought to indicate climatic conditions characterized by a dry season.

Table 3 Numerical ages of impact glasses

Localities	Host sediments	Glass age
Rio Cuarto northern Pampas	Surface material	4–10 ka 2.3 ± 1.6 ka 6 ± 2 ka 570 ± 100 ka 114 ± 26 ka
Centinela del Mar southern Pampas	Loessoid deposits of fluvial environment	230 ± 30 ka 445 ± 21 ka
Chapadmalal sea cliffs (Mar del Plata) southern Pampas	Loess and loessoid deposits	3.27 ± 0.08 Ma
Bahia Blanca southern Pampas	Loessoid deposits	5.33 ± 0.05 Ma
Chasicó southern Pampas	Loessoid deposits and fine sandstones	9.23 ± 0.09 Ma

Source: Adapted from Schultz P, Zárate M, Hames B, et al. (2004) The Quaternary impact record from the Pampas, Argentina. *Earth and Planetary Science Letters* 219: 221–238.

The Mountain Valley Loess of Tucumán

Loess–paleosol sections over 40 m thick have been reported at various sites in a mountain valley of Sierras Pampeanas in Tucumán province (Figure 7). The sections, located at altitudes between 2300 and 2500 m above sea level, are exposed along gully and river walls of the Tafi del Valle intramountain basin, a closed basin that acted as a sediment trap to eolian dust throughout the Quaternary.

Twenty-eight discrete paleosols consisting of Bw, Bt, and C horizons were recognized at the La Mesada section on the basis of morphological properties (color, structure, texture, and clay coating). Detailed micromorphological analysis carried out at two stratigraphic intervals of La Mesada permitted the identification of a series of three main pedosedimentary stages (Table 5), interpretation probably applicable to the rest of the sequence. The pedosedimentary stages include an interval of accretionary pedogenesis, followed by a relatively stable surface characterized by the dominance of pedogenic process and a return to accretionary pedogenesis. Also, a cyclical pedosedimentary history is inferred at the 45-m-thick loess–paleosol sequence of El Lambadero on the basis of field properties, magnetic susceptibility, particle size, calcium carbonate content, and soil micromorphology.

Magnetostratigraphy, geochemistry, and micromorphology were carried out at the 50-m-thick Las Carreras loess–paleosol sequence. The succession comprises 32 paleosols classified as Luvisols and (Luvic) Kastanozems. Carbonate leaching and reprecipitation, soil aggregation, and clay translocation are, among others, the major pedogenic processes involved in paleosol development. It is considered a quasicontinuous and complete continental record of climatic changes and environmental conditions in subtropical South America. The paleomagnetic analysis indicates a Matuyama magnetic age for the lower 15 m with the Brunhes–Matuyama boundary found at a depth of 26.7 m. A minimum age of 1.15 Ma is assigned to the onset of loess sedimentation. The relative maxima of magnetic susceptibility correspond mainly with loess and relative

Table 4 Location of the main magnetostratigraphic sections studied in the Pampean loess sequences

Area	Localities	Magnetostratigraphy
Mountain valleys of Tucumán	Las Carreras section	Brunhes
Northern Pampas	Buenos Aires city: three magnetostratigraphic sections (building excavations)	Matuyama (<1.15 Ma)
		Brunhes
	Baradero: two magnetostratigraphic sections	Matuyama (upper part), Jaramillo subchron recorded. Exposed lower level >1.5 Ma
		Brunhes
Southern Pampas	La Plata: three main magnetostratigraphic sections in building excavations and quarries	Matuyama
		Brunhes
	Chapadmalal sea cliffs, Mar del Plata: four magnetostratigraphic sections	Matuyama (upper part), subchron Jaramillo recorded. Exposed lower levels >1.5 Ma
		Brunhes
		Matuyama
		Gauss
		Gilbert (lower levels exposed >4 Ma)

minima with paleosols following a pattern similar to the Pampean loess–paleosol sequences. High susceptibility values were found associated with the presence of organic matter in both fossil and the modern Ah soil horizons. The paleoenvironmental or sedimentological significance of magnetic susceptibility signals, however, remains unclear and more studies are needed to explore the relation of the magnetic properties and lithology.

The Last Glacial Loess Record

Geographic Distribution

The eolian sediments of the last glacial cycle form a very extensive and continuous apron that blankets the landscape of the Pampean plain as well as extensive areas of the Chaco plain. The deposits constitute the parent material of the present cultivated soils (Figure 8). Small patches of loess are also found along the proximal Andean piedmont and intramountain valleys of Sierras Pampeanas (Figure 1).

Grain Size Zonation

Across the Pampean plain eolian deposits exhibit a regional west-southwest to east-northeast grain size zonation.

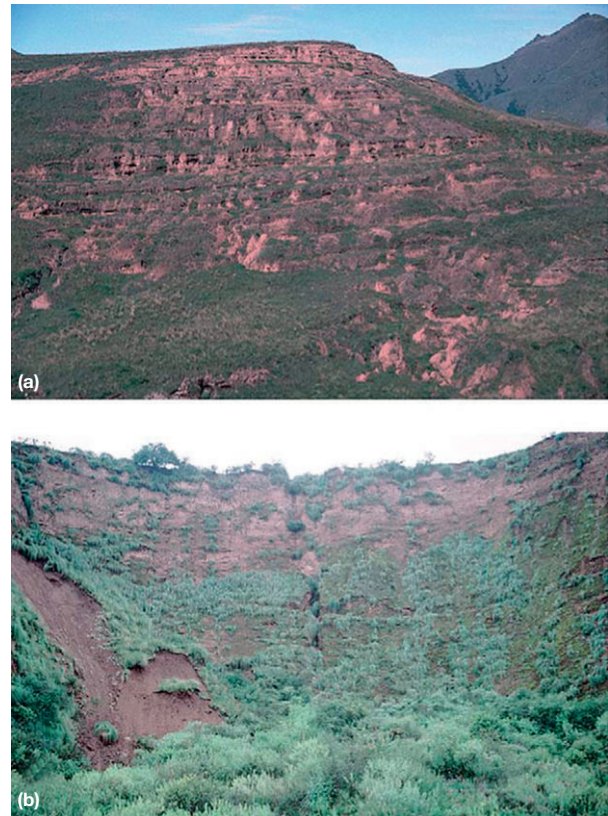


Figure 7 Mountain loess from Tafi del Valle, Tucumán. (a) Panoramic view of the upper part of the loess–paleosol sequence near La Mesada. (b) General view of El Lambadero section. Photos courtesy of Rob Kemp.

Loess is surrounded by a large sand dune field west and south-southwestward which extends along the distal eastern Andean piedmont, covering most of the wide valley of the Bermejo–Desaguadero–Salado fluvial system, as well as the floodplains and fan environments of the San Juan, Mendoza, Tunuyán, Diamante, and Atuel rivers. The sand dune fields and sand mantles extend eastward to around 150 km west of Buenos Aires, including large-scale dunes trending north-northeastward–south-southwestward. Southward, the sandy eolian facies continue along the Colorado and Negro fluvial valleys surrounding the southern Pampas where sandy deposits grade into loessial sands and sandy loess facies (Figure 9).

Local topographic factors promote the deposition of loess on the surface of structural plains of La Pampa province (Figure 9) and along the proximal eastern piedmont and intramountain basins of Sierras Pampeanas of Córdoba and San Luis.

In the Chaco plain, the loess cover is more discontinuous, alternating with alluvial deposits. Sand dunes composed of fine sands of dominantly quartz composition are reported in the Bolivia–Paraguay low plain.

Mineralogical Composition

In the southern Pampas, sandy loess is mostly volcanoclastic consisting of andesitic plagioclase, hornblende, pyroxene, volcanic glass shards, and quartz grains. Volcanic rock fragments

Table 5 Main characteristics of Tucumán mountain valley paleosols

Sections and thickness (m)	Number of paleosols	Paleosol characteristics
La Mesada (1, 2) (42 m)	28	(1) Bw and Bt horizons interbedded with loess (C horizons) (2) Micromorphology suggests three pedosedimentary stages: accretionary pedogenesis, relative stable surface, and return to accretionary pedogenesis
El Lamedero (2) (45 m)	One in the uppermost 6 m profile. Two in the lowermost 4.5 m profile	Bt, BC, and C horizons identified pedocomplex in the upper section; three pedosedimentary stages cyclical development with four pedosedimentary stages
Las Carreras (3) (50 m)	32	Bt, BC horizons, A horizons mostly present but difficult to identify Main pedogenic processes: carbonate leaching and reprecipitation, pedogenic formation and translocation of clay and organic matter

Source: Adapted from Kemp R, Toms P, Sayago JM, et al. (2003) Micromorphology and OSL dating of the basal part of the loess-paleosol sequence at La Mesada in Tucumán Province, Northwest Argentina. *Quaternary International* 106–107: 111–117; Kemp R, King M, Toms P, et al. (2004) Pedosedimentary development of part of a Late Quaternary loess-paleosol sequence in northwest Argentina. *Journal of Quaternary Science* 19: 1–10; Schellenberger A, Heller F, Veit H (2003) Magnetostratigraphy and magnetic susceptibility of the Las Carreras loess-paleosol sequence in Valle de Tafí, Tucumán, NW-Argentina. *Quaternary International* 106/107: 159–167; Zinck JA and Sayago JM (1999) Loess-paleosol sequence of La Mesada in Tucumán province, northwest Argentina. Characterization and paleoenvironmental interpretation. *Journal of South American Earth Sciences* 12: 293–310; Zinck JA and Sayago JM (2001) Climatic periodicity during the late Pleistocene from a loess-paleosol sequence in northwest Argentina. *Quaternary International* 78: 11–16.

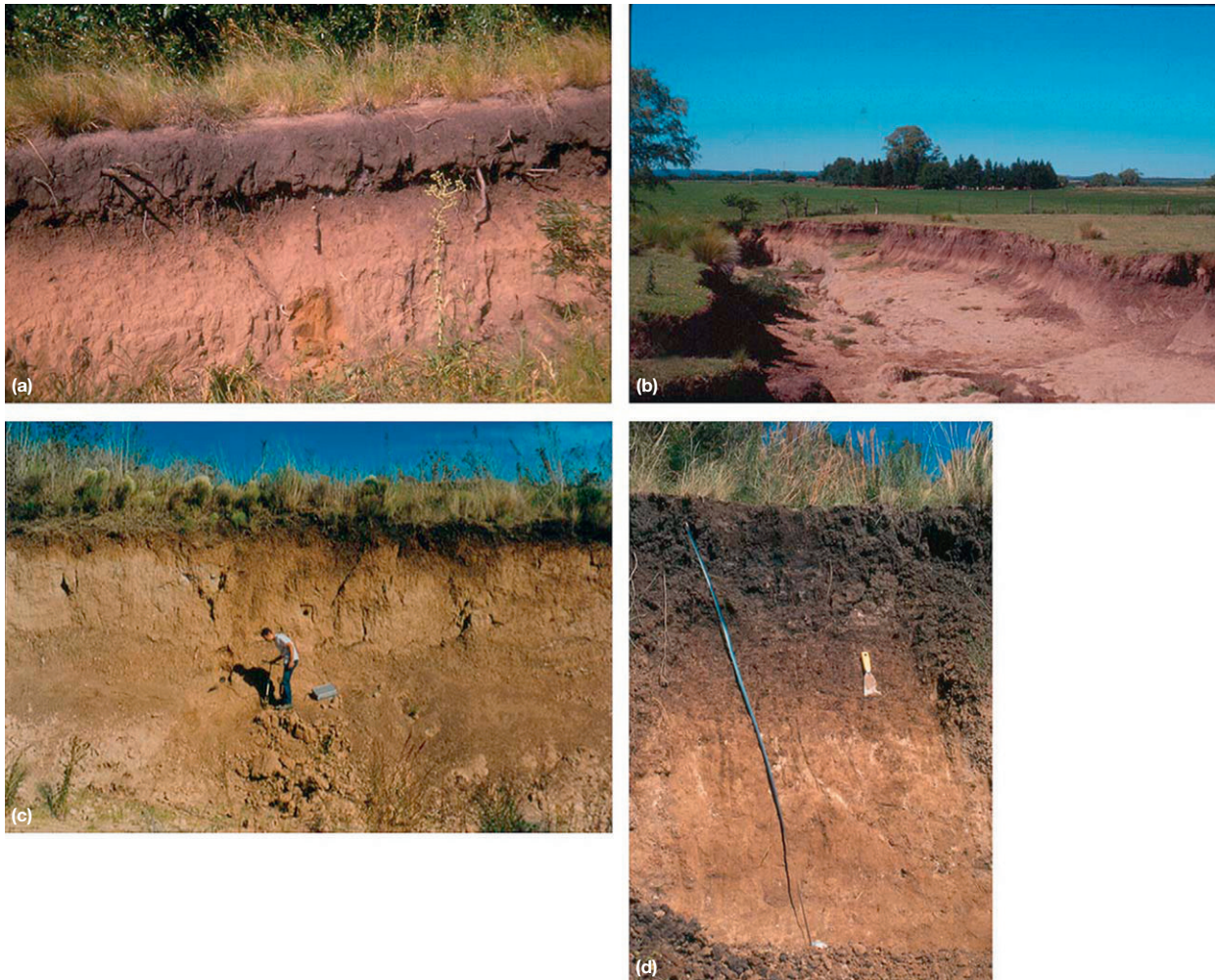


Figure 8 Last glacial loess deposits. (a) Late Glacial sandy loess with present soil profile (argiudoll) near Mar del Plata, southern Pampas. (b) Last glacial loess in Córdoba, northern Pampas, gully generated by agricultural practices. (c) Last glacial loess at the Gorina quarry, northern Pampas. (d) Detailed view of the present soil developed on last glacial loess at the Gorina quarry.

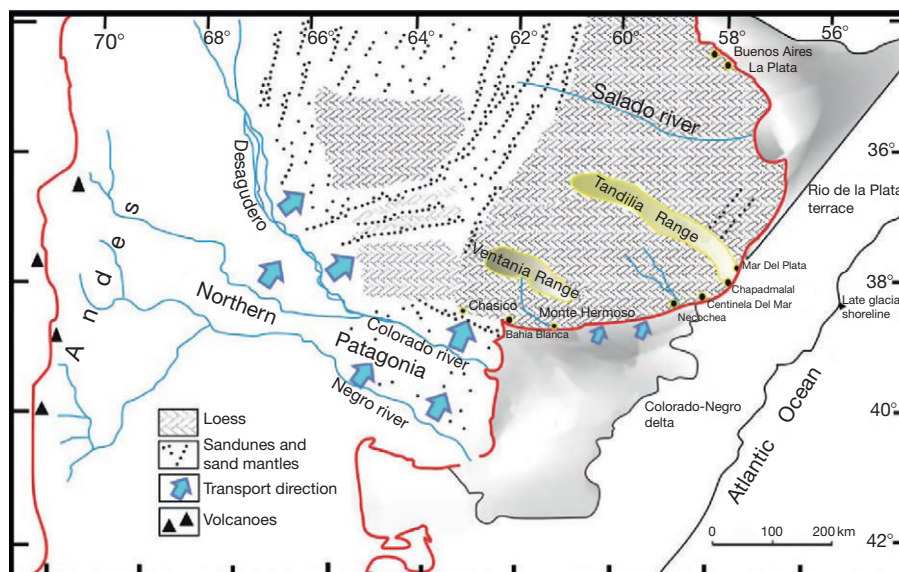


Figure 9 Loess and eolian facies of the southern Pampas. Source areas and inferred wind transport direction. Adapted from Zárate M and Blasi A (1993) late Pleistocene–Holocene eolian deposits of the southern Buenos Aires Province, Argentina: A preliminary model. *Quaternary International* 17: 15–20. Areal distribution of the Colorado and Negro rivers delta adapted from Parker G, Violante RA, Paterlini MC (1996) *Fisiografía de la plataforma continental*. In: Ramos VA and Turic MA (eds.) *XIII Congreso Geológico Argentino y III Congreso de exploración de hidrocarburos*, Buenos Aires. *Geología y Recursos Naturales de la Plataforma Continental Argentina*, Relatorio, vol. 1, pp. 1–16.

are very common. Particles are very fresh and usually well rounded (volcanic rock fragments, magnetite, pyroxenes) or subrounded to subangular (clinopyroxenes, orthoclase, plagioclase). Illite, poor in potassium, is the dominant clay mineral in Pleistocene–Holocene loess sequences, while smectite and kaolinite are secondary components (Table 6).

In the northern Pampas of Buenos Aires, the mineralogical composition of loess varies throughout the sections. It is predominantly composed of volcanoclastic material (glass shards and plagioclases); the sand fraction includes metamorphic rock fragments (gneisses, migmatites, etc.). Illite and smectite are the predominant clay minerals (Table 6). In southern Santa Fé, the last glacial loess is composed of volcanoclastic material as well as a secondary percentage of metamorphic and igneous rock. Quartz is a dominant component of Chaco loess.

Geochemistry and Isotopic Composition

The isotopic ratios (strontium-87/strontium-86 ($^{87}\text{Sr}/^{86}\text{Sr}$), neodymium-143/neodymium-144 ($^{143}\text{Nd}/^{144}\text{Nd}$)) of loess samples covering the time interval of ca. 10–150 ka from four localities of the Pampas (Hipodromo, Gorina, Baradero, and Lozada) and one locality from the mountain valleys of Tucumán (Lambadero) are characterized by a broad range of $^{87}\text{Sr}/^{86}\text{Sr}$ values and restricted $^{143}\text{Nd}/^{144}\text{Nd}$ ratios (0.5123–0.5126). A significant geographic variability exists and individual sites show less isotopic variability. Chondrite-normalized rare-Earth element (REE) patterns are generally consistent with loess from other regions and characterized by light REE enriched patterns, flat heavy REE trends, and negative europium anomalies.

Geochronology

Some radiocarbon ages together with a few thermoluminescence (TL) ages were first used to determine the chronology of

loess deposition in South America. A correlation with the marine isotopic stage stratigraphy was proposed for the northern Pampas of Santa Fé. New optically stimulated luminescence (OSL) analyses have provided numerical ages of some representative Pampean loess–paleosol sections and La Mesada and El Lambadero sections in the Tucumán mountain valleys. The intervals of dominant soil formation and periods of increasing depositional rates identified on the basis of detailed micromorphological analysis were chronologically bracketed and tentatively correlated with the marine isotope stratigraphy. Among other results, the OSL chronology suggests different deposition rates across the region with loess accumulating until the mid-Holocene in the western part of the northern Pampas. Discrepancies have arisen with OSL ages that suggest older ages than previously indicated by both radiocarbon and TL chronologies. At La Mesada, OSL ages ranging from 150 to 195 ka BP are much older than those indicated by the radiocarbon chronology (<30 ka BP). Consistent with this, the magnetostratigraphy in another nearby section (Las Carreras) suggests much older ages (Table 7).

Provenance of Loess Deposits

The mineralogy and geochemistry of loess in South America indicates that the Andes Cordillera, Sierras Pampeanas of Córdoba, and San Luis, and the Paraná River basin have been the most important source areas, with their relative inputs varying across the region (Figure 10). The isotopic data are consistent with multiple loess sources with a majority of particles being derived from the Andes. Discrete tephra layers, along high percentages of fresh volcanic shards within the loess suggest the direct input of Andean volcanic eruptions coming from the Central Volcanic Zone and the Southern Volcanic Zone. These two volcanic districts seem to have had a major

Table 6 Geographic distribution, mineralogical composition, source areas, and inferred wind systems of last glacial loess (adapted from Zárate, 2003 and references therein). Source areas of mountain valleys in Tucumán according to Smith et al. (2003) and Schellenberger and Veit (2006)

<i>Area</i>	<i>Eolian deposit</i>	<i>Mineralogical composition</i>	<i>Source area</i>	<i>Inferred transport-wind systems</i>
Western Chaco	Loess	Dominant quartz composition	Bolivian Andes	Northerly winds
Mountain valleys of Tucumán	Loess	Volcaniclastic composition dominant	Western Andean source (Chilean altiplano) Local sources. Tephra from CVZ	Northerly winds Westerly winds
Northern Pampas (Pampa Ondulada of Buenos Aires, southern Santa Fé, Córdoba)	Loess	Volcaniclastic composition dominant	Andes Cordillera, Sierras Pampeanas, Paraná basin, Uruguayan shield. Tephra from CVZ and the SVZ	South-southwesterly winds
	Clayey loess	Secondary amount of metamorphic and igneous rocks		(?)North/northeasterly winds (?)Tropospheric winds from the Puna/Altiplano
Northern Pampas (Pampa Deprimida)	Sandy loess (WSW) grading to typical loess (ENE)	Volcaniclastic composition dominant ENE area: not determined	Northern Patagonia (late Tertiary volcaniclastic sediments) Andes Cordillera (34–38° S)	South-southwesterly winds Holocene: westerly winds
Southern Pampas	Sand mantles	Volcaniclastic (<quartzitic and metamorphic rocks)	Sierras Pampeanas (?). Tephra from SVZ and CVZ (?) Northern Patagonia (late Tertiary volcaniclastic sediments)	South-southwesterly winds
	Loessial sands. Sandy loess. Clayey loess (<)		Andes (34–38° S). Tandilia and Ventania. Tephra from SVZ	Holocene: westerly winds

Table 7 Chronology proposed for the loess–paleosol sections studied in the mountain valleys of Tucumán according to Zinck and Sayago (1999), Zinck and Sayago (2001), Kemp et al. (2003, 2004), and Schellenberger et al. (2003)

<i>Sections</i>	<i>¹⁴C dates</i>	<i>OSL dates</i>	<i>Magnetostratigraphy</i>
La Mesada	Four dates between ~5–42 m of depth: age range from ca. 17.5 to 27.5 ka	Three dates between 35 and 42 m of depth: ages >152 ka	–
El Lamedero	–	Upper 6.1 m: 33–96 ka and lowermost 4.7 m: >165 ka	–
Las Carreras	–	–	Brunhes–Matuyama including Jaramillo subchron and Kamikatsura event

influence on the different compositions of loess in different regions of South America. The volcanoes of the Puna plateau probably were the main source of loess in the Chaco region and some of the northern Pampas tephras. The volcanic eruptions of the Southern Volcanic Zone are more significant as a loess source material in the southern Pampas.

The loess and sand deposits of the southern Pampas are mostly derived from Andean pyroclastic and volcanic rocks. Only a minor and localized contribution from the mountain ranges of southern Buenos Aires is present. In the northern Pampas, Andean volcaniclastic material is dominant in Santa Fé and northern Buenos Aires province. The Sierras Pampeanas of Córdoba and San Luis constitute a secondary source area, providing particles of metamorphic and igneous rocks and traces of kaolinite. In addition, the

Paraná River basin contributed as a secondary source. The metamorphic outcrops located on the Uruguayan margin of the Río de la Plata might be another potential secondary source area.

The isotopic signatures suggest that the Chilean Altiplano was the source area of the mountain valley loess of Tucumán. Presumably direct ash falls and local dust sources were secondary contributors.

Paleoclimatic Significance of the Loess Record

Wind Systems

Two regional-scale sedimentological models have been proposed to explain loess origins in South America. The ‘Chaco

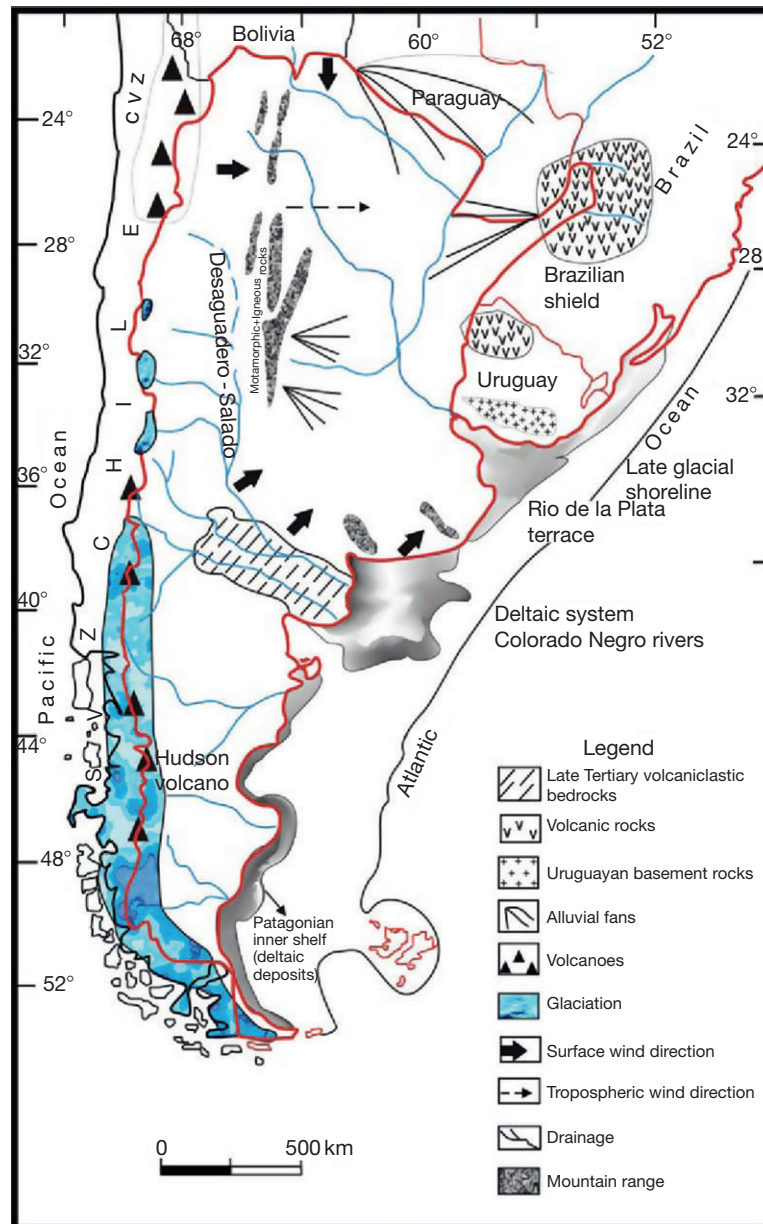


Figure 10 Source areas of loess and eolian sands. Adapted from Zárate M (2003) The Loess record of Southern South America. *Quaternary Science Reviews* 22: 1987–2006.

model' and the 'Pampean model' were formulated by Iriondo (1997) to illustrate source areas, transport pathways, and depositional areas for loess of the northern and southern regions, respectively. The Chaco model suggests that during last glacial period, northerly wind systems deflated the floodplains of the Parapetí, Pilcomayo, and Bermejo rivers, depositing the Chaco loess. The loess of the mountain valleys of Tucumán is now considered to be part of the Chaco model.

The Pampean model explains the loess record further south. Particles were deflated from the floodplains of the Colorado and Negro rivers, and the Desaguadero fluvial system, a tributary of the Colorado (Figure 9). The westerly and south-westerly transport directions inferred from geomorphological,

sedimentological, and geochemical evidences agree with westerly paleowind simulations derived from climate models.

South American Source of Antarctic Dust

A South American origin has been attributed to dust found in Antarctic ice cores. In general, the isotopic ratios of the eolian dust records from the Epica-Dome C and Vostok ice cores suggest that the Pampas, Patagonia, and possibly the exposed continental shelf of Argentina were the dominant sources of Antarctic dust. Detailed studies are needed to find out which of these regions were the main contributors of Antarctic dust.

Conclusions

Information is still lacking from several areas across the region, particularly the Chaco plain and the Andean mountain valleys north and south of Tucumán whereas recently proposed chronologies need to be tested in order to understand fully the paleoclimatic significance of the loess record. Many of the studies have been carried out in the Pampean plain. Here, the late Miocene to Holocene record is a complex stratigraphic succession mostly composed of loess-like deposits, numerous welded paleosols, and several erosional unconformities. The paleoenvironmental and paleoclimatic interpretations derived from the record should be examined carefully considering local and regional factors of controls. At a regional scale, loess deposits become a dominant component of the sequences during the last glacial period. Differences in sedimentation rates, sediment supplies, and soil development reflect heterogeneous environmental conditions through time across the vast Chaco-Pampean plain. To date, the apparently continuous and complete mountain valley loess–paleosol sequence of Tucumán, in north-western Argentina, dating back to around 1.15 Ma, is the most promising continental proxy record of paleoenvironmental and paleoclimatic conditions of southern South America.

See also: Loess Deposits: Origins and Properties. **Beetle Records:** Late Pleistocene of South America. **Glacial Climates:** Effects of Atmospheric Dust. **Glaciations:** Late Pleistocene in South America; Middle Pleistocene Glaciations in the Southern Hemisphere. **Ice Core Records:** South America. **Lake Level Studies:** Latin America. **Loess Records:** China; Europe; North America. **Luminescence Dating:** Optical Dating. **Paleosols and Wind-Blown Sediments:** Nature of Paleosols; Soil Micromorphology. **Pollen Records, Late Pleistocene:** South America. **Pollen Records, Postglacial:** South America. **Vertebrate Records:** Late Pleistocene of South America.

References

- Bidegain JC (1998) New evidence of the Brunhes/Matuyama polarity boundary in the Hernández-Gorina quarries, northwest of the city of La Plata, Buenos Aires Province, Argentina. *Quaternary of South America and Antarctic Peninsula* 11: 207–228.
- Cione AL and Tonni EP (2001) Correlation of Pliocene to Holocene southern South American and European vertebrate-bearing units. *Bollettino della Società Paleontologica Italiana* 40: 167–173.
- Clapperton CM (1993) *Quaternary Geology and Geomorphology of South America*, p. 779. Amsterdam: Elsevier.
- Delmonte B, Basile Doelsch I, Petit JR, et al. (2004) Comparing the Epica and Vostok dust records during the last 220,000 years: Stratigraphical correlation and provenance in glacial periods. *Earth Science Reviews* 66: 63–87.
- Frenguelli J (1955) *Loess y Limos pampeanos*. Ministerio de Educación de la Nación. Serie Técnica y Didáctica. La Plata, no. 7.
- Imbellone PA and Teruggi ME (1993) Paleosols in loess deposits of the Argentine Pampas. *Quaternary International* 17: 49–55.
- Iriondo M (1990) Map of the South American plains – Its present state. *Quaternary of South America and Antarctic Peninsula* 6: 297–308.
- Iriondo MH (1997) Models of deposition of loess and loessoids in the Upper Quaternary of South America. *Journal of South American Earth Sciences* 10: 71–79.
- Iriondo MH and Kröhling D (1997) The tropical loess. In: Zhisheng A and Weijian Z (eds.) *Quaternary Geology, Proceedings of the 30th International Geological Congress* 21: 61–77.
- Kemp R, King M, Toms P, et al. (2004) Pedosedimentary development of part of a Late Quaternary loess–paleosol sequence in northwest Argentina. *Journal of Quaternary Science* 19: 1–10.
- Kemp R, Toms P, Sayago JM, et al. (2003) Micromorphology and OSL dating of the basal part of the loess–paleosol sequence at La Mesada in Tucumán Province, Northwest Argentina. *Quaternary International* 106–107: 111–117.
- Kemp R, Zárate M, Toms M, et al. (2006) Late Quaternary paleosols, stratigraphy and landscape evolution in the Northern Pampas, Argentina. *Quaternary Research* 66: 119–132.
- Kröhling DM (1999) Upper Quaternary geology of the lower Carcarañá Basin, North Pampa, Argentina. *Quaternary International* 57/58: 135–148.
- Morrás H (1999) Geochemical differentiation of Quaternary sediments from the Pampean region based on soil phosphorus contents as detected in the early 20th century. *Quaternary International* 62: 57–67.
- Morrás H, Moretti L, Piccolo G, and Zech W (2005) New hypothesis and results on the origin of stonelines and subsurface structure in ferrallitic soils of Misiones, Argentina. *Geophysical Research Abstracts* 7: 05522.
- Muhs D and Zárate M (2001) Late Quaternary eolian records of the Americas and their paleoclimatic significance. In: Markgraf V (ed.) *Interhemispheric Climate Linkages*, pp. 183–216. San Diego: Academic Press.
- Nabel P, Cione A, and Tonni E (2000) Environmental changes in the Pampean area of Argentina at the Matuyama–Brunhes (C1r–C1n) Chrons boundary. *Palaeogeography, Palaeoclimatology, Palaeoecology* 162: 403–412.
- Nabel PE, Morrás HJM, Petersen N, and Zech W (1999) Correlation of magnetic and lithologic features of soils and Quaternary sediments from the Undulating Pampa, Argentina. *Journal of South American Earth Sciences* 12: 311–323.
- Orgeira MJ (1990) Paleomagnetism of late cenozoic fossiliferous sediments from Barranca de los Lobos (Buenos Aires province, Argentina). The magnetic age of the South American land-mammal ages. *Physics of the Earth and Planetary Interiors* 64: 121–132.
- Orgeira MJ, Walther AM, Tofalo RO, et al. (2003) Environmental magnetism in fluvial and loessic Holocene sediments and paleosols from the Chacopampean plain (Argentina). *Journal of South American Earth Sciences* 16: 259–264.
- Orgeira MJ, Walther AM, Vasquez CA, et al. (1998) Mineral magnetic record of paleoclimatic variation in loess and paleosol from the Buenos Aires formation (Buenos Aires, Argentina). *Journal of South American Earth Sciences* 11: 561–570.
- Panario D and Gutiérrez O (1999) The continental Uruguayan Cenozoic: An overview. *Quaternary International* 62: 75–84.
- Parker G, Violante RA, and Paterlini MC (1996) Fisiografía de la plataforma continental. In: Ramos VA and Turic MA (eds.) *XIII Congreso Geológico Argentino y III Congreso de exploración de hidrocarburos*, Buenos Aires. *Geología y Recursos Naturales de la Plataforma Continental Argentina*, Relatorio, vol. 1, pp. 1–16.
- Polanski J (1963) Estratigrafía, Neotectónica y Geomorfología del Pleistoceno pedemontano entre los ríos Diamante y Mendoza. *Asociación Geológica Argentina Revista XVII*: 127–349.
- Sayago JM (1995) The Argentinian neotropical loess: An overview. *Quaternary Science Reviews* 14: 755–766.
- Schellenberger A, Heller F, and Veit H (2003) Magnetostratigraphy and magnetic susceptibility of the Las Carreras loess–paleosol sequence in Valle de Tafí, Tucumán, NW-Argentina. *Quaternary International* 106/107: 159–167.
- Schellenberger A and Veit H (2006) Pedostratigraphic and pedological and geochemical characterization of Las Carreras loess–paleosols sequence, Valle de Tafí, NW-Argentina. *Quaternary Science Reviews* 25: 811–831.
- Schultz P, Zárate M, Hames B, et al. (2004) The Quaternary impact record from the Pampas, Argentina. *Earth and Planetary Science Letters* 219: 221–238.
- Smith J, Vance D, Kemp R, et al. (2003) Isotopic constraints on the source of Argentinian loess – With implications for atmospheric circulation and the provenance of Antarctic dust during recent glacial maxima. *Earth and Planetary Science Letters* 6682: 1–16.
- Teruggi ME (1957) The nature and origin of Argentine loess. *Journal of Sedimentary Petrology* 27: 322–332.
- Tonni EP, Nabel P, Cione AL, et al. (1999) The Ensenada and Buenos Aires formations (Pleistocene) in a quarry near La Plata, Argentina. *Journal of South American Earth Sciences* 12: 273–291.
- Vizcaino S, Fariña RA, Zárate MA, Bargo MS, and Schultz P (2004) Palaeoecological implications of the mid-Pliocene faunal turnover in the Pampean region (Argentina). *Palaeogeography, Palaeoclimatology, Palaeoecology* 213: 101–113.

- Vizcaino SF, Zárate MA, Bargo MS, and Dondas A (2001) Pleistocene burrows in the Mar del Plata area (Buenos Aires province, Argentina) and their probable builders. *Acta Paleontologica Polonica* 46: 289–301.
- Zárate M (2003) The loess record of southern South America. *Quaternary Science Reviews* 22: 1987–2006.
- Zárate M and Blasi A (1993) Late Pleistocene–Holocene eolian deposits of the southern Buenos Aires Province, Argentina: A preliminary model. *Quaternary International* 17: 15–20.
- Zinck JA and Sayago JM (1999) Loess–paleosol sequence of La Mesada in Tucumán province, northwest Argentina. Characterization and palaeoenvironmental interpretation. *Journal of South American Earth Sciences* 12: 293–310.
- Zinck JA and Sayago JM (2001) Climatic periodicity during the late Pleistocene from a loess–paleosol sequence in northwest Argentina. *Quaternary International* 78: 11–16.

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LUMINESCENCE DATING

Contents

Thermoluminescence

Optical Dating

Electron Spin Resonance Dating

Thermoluminescence

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Thermoluminescence (TL) dating is based on the fact that natural minerals can absorb and store ionizing energy. If a mineral is heated to a sufficiently high temperature, some of the stored energy is released in the form of light called TL; a subsequent exposure to heat will not result in more TL being emitted unless the mineral is first allowed to absorb more ionizing energy. Exposure to light (e.g., natural sunlight) will also cause some of this stored energy to be lost.

In the natural environment, the energy absorbed by a mineral mainly comes from radiation emitted from radioisotopes within the mineral grain, from its immediate surroundings, and from cosmic rays. The rate at which a dose of radiation energy is absorbed by a mineral (the dose rate) is therefore proportional to the concentrations of radioisotopes and to the intensity of cosmic rays at the sample site. Simply stated, for a given mineral sample, what is needed to determine a TL age is (1) the laboratory dose of radiation that produces the same intensity of TL as did the environmental radiation dose (called the equivalent dose) and (2) a measure of the environmental dose rate. The TL age is simply the equivalent dose divided by the dose rate. TL dating therefore gives the time elapsed since mineral grains were last exposed to sufficient heat or sunlight.

TL dating was developed in the early 1960s as a means of dating fired pottery (Aitken, 1985; Aitken et al., 1964, 1968; Roberts, 1997), but the technique was later modified so that it also could be used to date the last time that sediments were exposed to sunlight. TL dating is part of a family of chronological techniques collectively referred to as trapped charge dating. Related methods include electron spin resonance dating, optical dating, and radioluminescence dating (Krbetschek et al., 2000).

Following the seminal publication by Daniels et al. (1953) on the potential of using TL in geological and archaeological research, TL-dating techniques that could be used to date unheated sediments were developed independently in the mid- to late 1960s in the former Soviet Union and in North America. Historical reviews can be found in Dreimanis et al. (1978), Wintle and Huntley (1982), Aitken (1985), and Lian and

Roberts (2006). TL dating is now a well-established method with an established age range of approximately 500–500,000 years, but in special cases ages as old as 1 Ma have been calculated. Because Quaternary geoscientists are mainly concerned with natural sedimentary environments, only the TL-dating techniques relevant to this application are discussed in this article. Discussions of the methods relevant to archaeological materials (e.g., fired pottery, hearthstones, artificial glass, and slag) can be found in Fleming (1979), Aitken (1985, 1990), Roberts (1997), Feathers (1997), and Wagner (1998), although the differences between the methods are not major.

Mechanism Responsible for TL

TL refers to the light emitted by matter in response to heat. It is not to be confused with incandescence, which is the light emitted from red-hot coals or a light bulb filament. In the case of TL, further heating will not result in more light being emitted, unless the material is first allowed to absorb ionizing energy. The physical models used to explain the TL phenomena are complex and not fully understood; a comprehensive account of the physics behind the mechanism is provided by McKeever (1985), whereas concise, and varied, accounts relevant to dating can be found in Aitken (1985), Wintle (1997), McKeever and Chen (1997), and Wagner (1998).

Crystals, such as those comprising natural minerals, are not perfect and contain impurities and structural defects, some of which are attractive to electrons that have been freed from their normal sites in the valence bands of atoms. Free electrons are produced when atoms absorb ionizing radiation that occurs in the natural environment.

Free electrons may subsequently be trapped at impurity or structural defect sites (traps) for a time that depends on the strength, or thermal lifetime, of the particular trap. At environmental temperatures (~20 °C), traps with long thermal lifetimes (sometimes referred to as deep traps) may hold electrons for

millions of years or more, whereas traps with short thermal lifetimes (shallow traps) may hold electrons for only a few days or less. In any given mineral grain, there are many different kinds of defects and impurities; many of these can act as electron traps, and these, in turn, all have different thermal lifetimes.

When a mineral grain is exposed to an external source of energy (e.g., heat or light), an electron in a trap may acquire enough energy to escape from it. Once evicted, an electron will travel around the crystal lattice until it eventually becomes retrapped or comes across another kind of site attractive to electrons referred to as a recombination center; this process of eviction and recombination is nearly instantaneous. When an electron encounters a recombination center, it combines with the atom there and excess energy is given off either as heat (lattice vibrations or phonons) or as luminescence (e.g., TL); in the latter case, the process is referred to as radiative recombination. If irradiation continues, electron traps will continue to fill until a further dose of radiation results in no more TL being measured. This condition is referred to as saturation.

All electron traps can be emptied if a mineral grain is exposed to sufficient heat, but only some of these traps can be emptied by exposure to light. Of the light-sensitive traps responsible for TL, some can easily be emptied by a brief exposure to sunlight exposure, but there are others for which a much longer sunlight exposure is needed. During laboratory heating, both incandescence and TL are emitted, but these are separated using infrared and red absorbing optical filters so that it is mainly the violet or ultraviolet TL that is being recorded.

TL Dating

As mentioned previously, to calculate a TL age requires an estimate of the equivalent dose and measurement of the environmental dose rate. These are discussed in turn.

Estimating the Equivalent Dose

There are several TL-dating procedures available for estimating the equivalent dose. Detailed discussions with various emphases can be found in, for example, Wintle and Huntley (1982), Aitken (1985), Singhvi and Wagner (1986), Berger (1988a,b, 1995), Forman (1989), Lian and Huntley (2001), and Lian and Roberts (2006). Of the TL-dating procedures available, the best established are the additive dose, total bleach, partial bleach (sometimes called the *R*-gamma or *R*-beta method, depending on whether gamma or beta radiation is used), and the regeneration methods. All these methods require that 20–50 aliquots of a sample be prepared. These aliquots are then given various laboratory radiation doses, in some cases several are exposed to light (i.e., they are bleached for a specific duration under laboratory light or natural sunlight), and then their TL is measured using an apparatus such as that shown in Figure 1. These data are used to model the way a sample's TL responds to absorbed radiation dose. From this dose-response model, the radiation dose the sample had absorbed since it was last heated or exposed to sunlight is estimated. This absorbed dose estimate is commonly referred to as the equivalent dose because it is found using the dose-response constructed using either beta particles or gamma rays from laboratory radiation sources, whereas in the environment, the

dose that the sample absorbed has come from alpha, beta, gamma, and cosmic ray radiation. Equivalent dose is commonly denoted by ED, D_{eq} , or D_{eqr} , and it is sometimes referred to loosely (and incorrectly) as paleodose.

Of the techniques used to estimate the equivalent dose, the additive dose method is the simplest. It is used in cases in which mineral sediment has been exposed to enough heat to empty all the relevant electron traps. The dose-response is constructed by giving several aliquots of the prepared sample various radiation doses while leaving some of them undosed; these latter aliquots are referred to as naturals (commonly denoted by *N*). TL from all the aliquots is recorded as a function of temperature, and data sets consisting of TL intensity as a function of laboratory dose are determined for successively increasing temperatures (e.g., 200–500 °C). For each data set, a curve is fitted and extrapolated to where it intercepts the dose axis. When a temperature range associated with thermally stable traps has been reached, the dose axis-intercept values will be constant, forming what is commonly referred to as a TL plateau. The equivalent dose is usually taken to be the average value of the dose-intercept values that fall within the plateau. If the dose-intercept values do not form a plateau, then something is amiss and an equivalent dose cannot be determined from them. Achieving a plateau in the dose-response versus temperature data is therefore a necessary condition for providing a TL age. An example of the additive dose technique is shown in Figure 2.

In cases in which the event to be dated is the last exposure to extended sunlight, the total bleach method is often used (Singhvi et al., 1982; Wintle and Huntley, 1979). The concept is that the measured TL intensity, *I*, can be considered as being the sum of two components, I_0 and I_d . I_0 is the TL intensity that would have been measured had the sediment sample been collected immediately after burial, which results only from electrons in light-insensitive traps, whereas I_d is the TL intensity resulting from the environmental radiation dose absorbed since the sample was buried. A clear-cut division of the traps into two such types probably does not exist in reality, although in practice it appears to be a good approximation. The total bleach method involves extrapolating the dose-response curve to the TL intensity value of a natural aliquot that has received an extended exposure to sunlight of a duration and spectrum similar to that which it would have received in nature (Figure 3(a)).

As mentioned previously, the relationship between I_0 and I_d is an approximation. In reality, there are electron traps with a wide range of sensitivities to light, and therefore, it can be expected that there are cases for which some of the relevant traps remain full even after prolonged sunlight exposure. Moreover, there is concern that the extended sunlight exposure used in the total bleach method may result in TL sensitivity change and other unwanted effects. These problems are taken into account using the partial bleach method (Wintle and Huntley, 1980; Figure 3(b)). With the partial bleach method, the TL reduction caused by a relatively brief, and usually spectrally restricted, light exposure is measured as a function of laboratory radiation dose, with the goal that only the light-insensitive traps are sampled during TL measurement. These data are then extrapolated to zero reduction to find the dose axis intercept. The way this is done in practice is to construct a

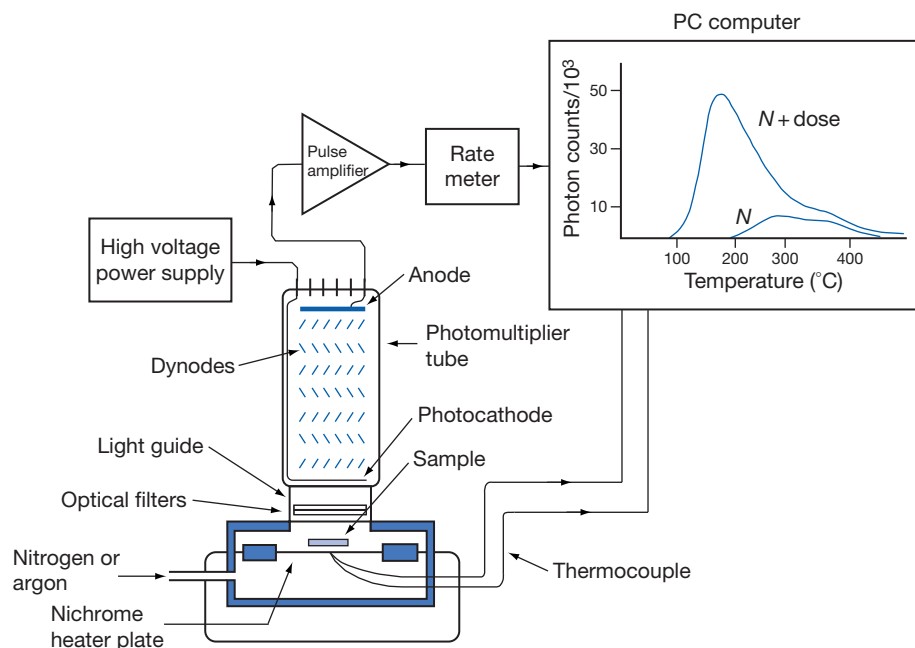


Figure 1 Schematic diagram of an apparatus (glow oven) designed to measure and record TL from sediments. The apparatus is a basic system designed to measure one sample aliquot at a time. More advanced systems, which are automated, can measure many aliquots in a run and can perform irradiations onboard. For the basic apparatus shown, a sediment sample is mounted on a ~1-cm-diameter aluminum or stainless-steel disk, which is in turn placed on the nichrome heater plate. The sample disk is commonly secured with a thermal compound paste so that there will be good thermal contact between it and the heater plate. The glow oven chamber is then evacuated, and the atmosphere is replaced with an inert gas, such as argon or nitrogen. The chamber needs to be evacuated to avoid the measurement of spurious TL, and an inert gas atmosphere is needed in the chamber to ensure good thermal contact between the disk, the sample grains, and the heater plate. The nichrome plate is heated at a constant rate by an electrical current, and TL photons are emitted from the sample as a consequence. The top of the chamber is transparent, allowing photons to pass from the sample through the light guide, which contains optical filters, to the photocathode of the photomultiplier tube (detector), where they result in the production of electrons (photoelectrons). The number of electrons that result from this process depends on the quantum efficiency of the photocathode, which typically is between 0% and ~25%, depending on the wavelength (energy) of the incident photons. Electrons produced at the photocathode are accelerated between several dynodes, each step resulting in the production of more electrons. For a typical photomultiplier tube, one electron at the photocathode will result in millions of electrons reaching the anode. The other equipment shown in the diagram is used to measure and control the heating rate (thermocouple) and to amplify and modify the pulses at the photomultiplier tube anode so that they can be recorded by the computer. The output example shown consists of two glow curves – plots of TL intensity (photon counts) versus temperature. The first glow curve is for a natural sample (*N*), and the second curve is for a natural sample that has been given a laboratory dose of radiation (*N*+dose). Note that the *N*+dose curve shows TL measured below approximately 200 °C, which has resulted from electrons in traps that are thermally unstable over geological time, with the electrons having been put there as a result of laboratory irradiation (see also [Figure 2](#)). Based on Figure 1.2 of [Aitken \(1985\)](#).

dose–response curve, as is done with the additive dose method discussed previously. In the case of the partial bleach method, however, some of the aliquots are given a short exposure to laboratory light (or natural sunlight) following laboratory dosing. Both data sets are fitted with suitable curves, and these are extrapolated to the point where they intersect above the dose axis. As with the additive dose and total bleach methods, dose axis-intercept values are determined at successively increasing temperatures until a plateau is achieved, and the equivalent dose is determined from it ([Figure 3\(b\)](#)). A modification of the partial bleach method is the selective bleach method ([Prescott and Mojarrabi, 1993](#)). This method is based on the fact that traps emptied at ~325 °C are significantly more sensitive to sunlight than others measured between 300 and 500 °C ([Franklin and Hornyak, 1990](#); [Prescott and Fox, 1990](#)). In practice, aliquots used for the partial bleach data set are selectively bleached with light of a wavelength of ~425 nm (green) or longer, and then the 380 nm TL associated with the 325 °C trap is measured for all aliquots.

If extrapolation of the dose–response curve(s) to the dose axis is large, the regeneration technique should be used. In its simplest form, it involves interpolation of the TL intensity of a natural sample aliquot onto its additive dose curve, and the equivalent dose is determined from that point's position on the curve above the dose axis. In its most complete form, the regeneration method involves constructing a regenerative dose–response data set in addition to the additive dose data set. The regenerative data set is produced by bleaching some of the sample aliquots for an extended time so as to empty all the light-sensitive traps and then giving them various laboratory radiation doses. The data points (laboratory dose values) that define the regenerative dose–response are chosen so as to largely overlap those that define the additive dose–response. The additive dose data set is then shifted along the dose axis to where it coincides with the regenerative dose data set, with the magnitude of the shift being equal to the equivalent dose after a correction for incomplete laboratory bleaching. A requirement of the technique is that the dose–responses of

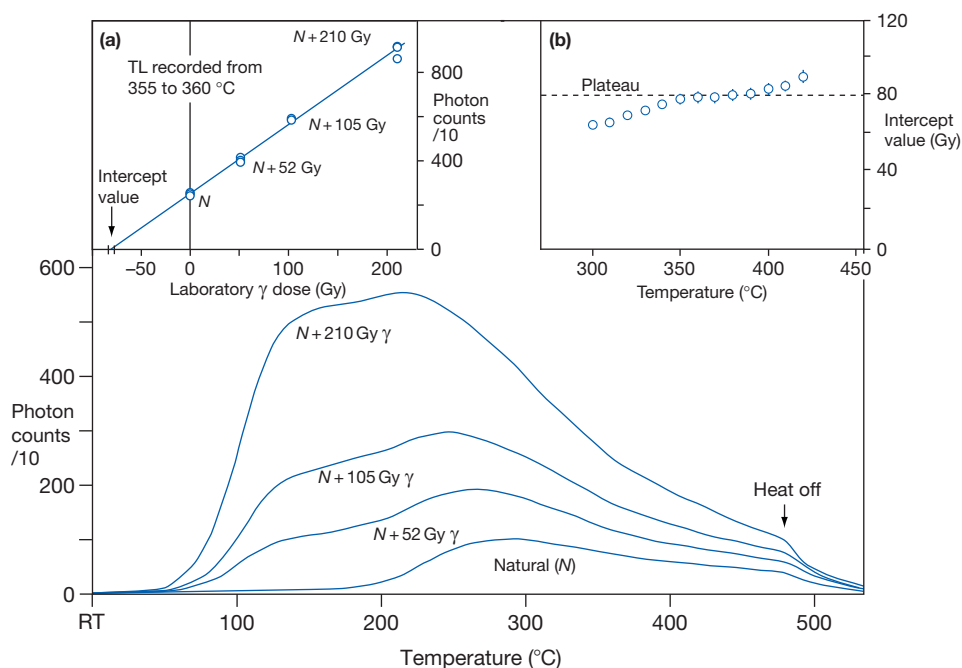


Figure 2 Example of the additive dose method used for finding the equivalent dose of a sediment sample that has been heated so that all the relevant electron traps have been emptied. The example is for fluvial silt that was heated by fire approximately 15 000 years ago. Aliquots of the sample are prepared; some are left as it is (the naturals, N), whereas others are given various doses of laboratory radiation ($N + \text{dose}$ data points), in this case from a ^{60}Co gamma (γ) source. Typically, several aliquots are used at each dose point. The TL from the sample aliquots is measured, resulting in glow curves that have peaks representing different electron traps; as the temperature increases, progressively deeper, or more thermally stable, traps are being emptied. Inset (a) shows the dose–response determined from the TL that was measured between 355 and 360 °C, and the dose axis intercept is found as shown. Inset (b) shows dose axis intercepts plotted as a function of glow curve temperature and the formation of a TL plateau at approximately 350 °C that corresponds to an equivalent dose of approximately 80 Gy. Modified from Figure 1 of Lian OB and Huntley DJ (2001) *Luminescence dating*. In: Last, WM and Smol JP (eds.) *Tracking Environmental Change Using Lake Sediments, vol. 1: Basin Analysis, Coring, and Chronological Techniques*, pp. 261–282. Dordrecht, The Netherlands: Kluwer.

both the additive dose and regenerative data sets are identical. This regeneration method is often called the Australian slide method (Prescott et al., 1993) due to its extensive use on the 1 million year-old stranded dune sequence in South Australia.

All of the protocols discussed previously involve constructing a sample's dose–response using either laboratory gamma or beta radiation. In the environment, a sample is also irradiated by alpha particles, which are less efficient in producing free electrons than are gamma or beta radiation. For sand-sized grains, the alpha-affected surfaces are removed by etching with acid, but this is impractical for silt-sized grains. Therefore, if silt-sized grains are to be dated, their alpha efficiency must be determined and included in the dose-rate calculation. This is done by constructing a separate alpha dose–response curve and comparing the equivalent dose found from it with that found using either gamma or beta radiation; see Aitken (1985) for a thorough discussion.

The wavelength (i.e., energy) of the TL emitted from a sample depends on the minerals present and the nature of the electron traps (i.e., the types of impurities and defects present) sampled during heating. The wavelength range that can be measured is largely dependent on the sensitivity of the detector (photomultiplier tube; Figure 1); most of these can only detect wavelengths below approximately 600 nm. This, together with the fact that light resulting from incandescence emitted at high temperatures must be avoided, means that blue or shorter

wavelengths typically are used for dating. In practice, optical filters are used to separate out the desired part of the spectrum (Figure 1). Blue pass filters are commonly used for quartz, and both blue and ultraviolet pass filters are used for feldspar dating. The choice of wavelength does have a bearing on the equivalent dose, but the relationship between the two is not clear (Berger et al., 1994). However, the ultraviolet TL from feldspar does bleach more quickly with sunlight exposure time than does the blue emission (see Aitken, 1985, Figure 8.8).

Determining the Dose Rate

The dose absorbed by the sediment grains of interest comes from alpha, beta, and gamma radiation emitted during the decay of ^{235}U , ^{238}U , ^{232}Th , ^{40}K , and ^{87}Rb and their daughter products, within both the mineral grains and their surroundings. Another component of the absorbed dose comes from cosmic rays, which consist of highly energetic particles originating from unknown sources in outer space. The intensity of cosmic rays diminishes with depth beneath the ground surface, and the rate at which it diminishes is related to the density of the material through which they are passing. The intensity of cosmic rays at the surface also varies with latitude and elevation and with fluctuations in Earth's magnetic field (Prescott and Hutton, 1994). The radiation dose absorbed by the minerals of interest depends not only on the concentration of the

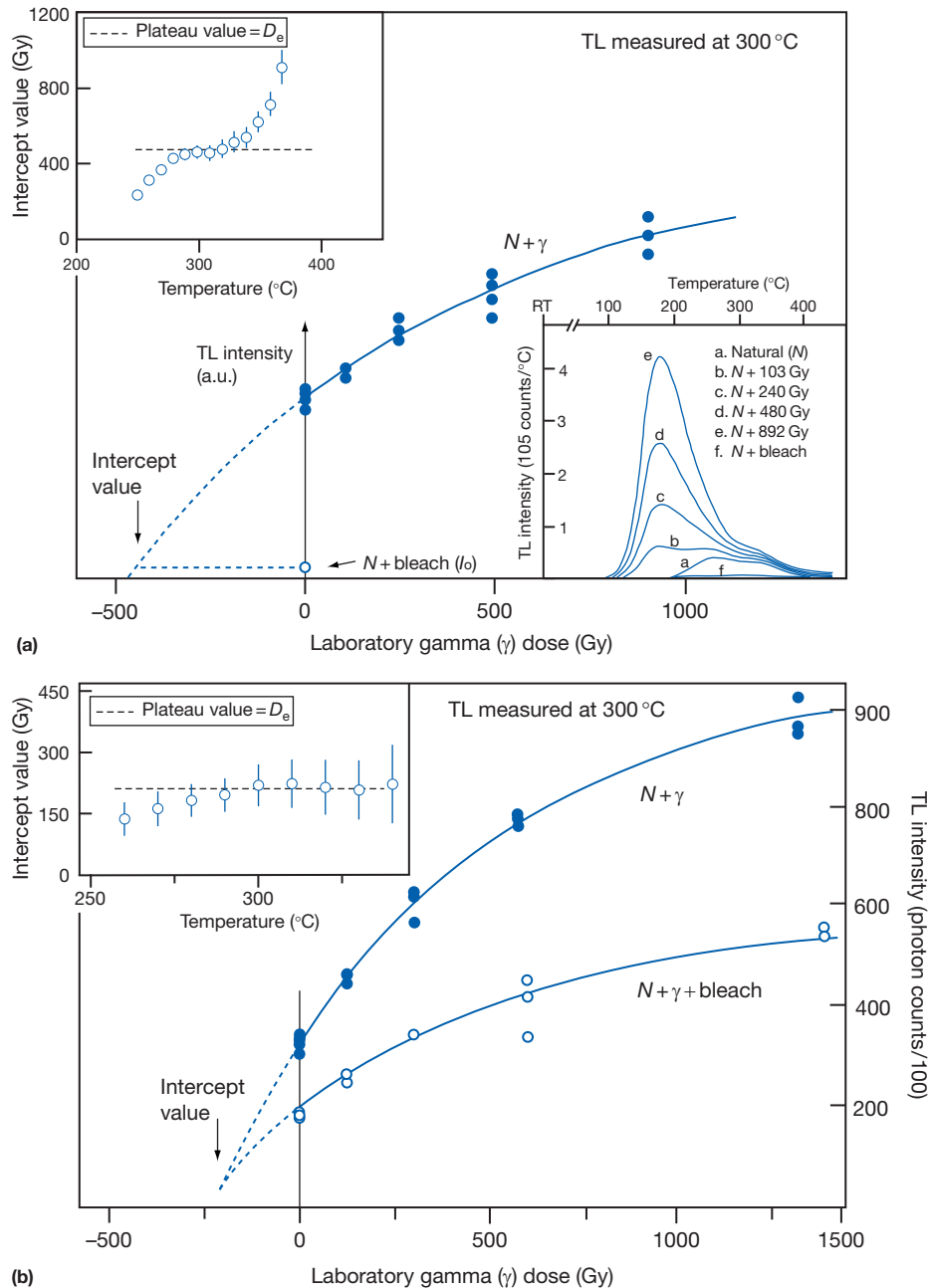


Figure 3 Examples of the total bleach and partial bleach methods for determining equivalent dose. (a) The total bleach method. Several sample aliquots are prepared; some are left as it is (naturals, N), whereas others are given various laboratory doses, in this case for a ^{60}Co gamma (γ) source, ($N + \gamma$ set). These data are fitted with a curve, which is extrapolated (dashed line) to where it intercepts the TL intensity measured from a natural aliquot that has been given a long exposure to light ($N + \text{bleach}$) so that only electrons in the most light-insensitive traps remain, which give rise to I_0 (see text). The inset at the bottom shows the TL glow curves for this sample. The inset at the top shows intercept values plotted as a function of progressively increasing temperatures; an intercept (TL) plateau is formed between ~ 300 and 330°C , and from it the equivalent dose (D_e) is estimated to be ~ 480 Gy. The data are for K-feldspar medium sand, and they gave a TL age between 240 and 290 ka. (b) The partial bleach method. The data are essentially the same as for the total bleach method shown in (a) except another (lower) curve is constructed from data measured from a set of aliquots that have been given various laboratory doses, followed by a restrictive bleach so that only the most light-sensitive traps are emptied ($N + \gamma + \text{bleach}$ set). Both curves are extrapolated (dashed lines) to where they intersect above the dose axis. As in (a), intercept values are determined as a function of temperature (inset graph), and the dose at which they become constant is taken to be the D_e , which is estimated to be ~ 225 Gy in this case. The TL, presumably from K-feldspar, was measured from fine silt polymineral grains extracted from an ~ 110 ka organic-rich (peaty) sediment. Optical filters in front of the photomultiplier tube included a Schott KG-1 (to absorb unwanted incandescence) and Schott BG-39 and Kopp 5-58 blue pass filters. The sample aliquots were preheated at 130°C for 5 days prior to TL measurement to minimize the effects of anomalous fading. From Lian OB and Roberts RG (2006) Dating the Quaternary: Progress in luminescence dating of sediments. *Quaternary Science Reviews* 25, Figure 2: the graphs in (a) are based on Figure 3 of Balescu and Lamothe (1993), and the graphs in (b) are modified from Figure 4(d) of Lian et al. (1995).

radioisotopes mentioned previously, and the intensity of cosmic rays, but also on the quantity of water and organic matter in the sediment matrix because these attenuate or absorb radiation differently from the mineral matter. All of these quantities have to be determined and included in the dose-rate calculations. General formulas used to calculate dose rate from concentrations of ambient U, Th, and K can be found in Aitken (1985, 1998), whereas the latest dose-rate conversion factors for these radioisotopes are those of Adamiec and Aitken (1998). Methods for estimating the contribution to the dose rate of beta particles emitted during the decay of ^{40}K and ^{87}Rb within K-feldspar grains are discussed by Huntley and Baril (1997) and Huntley and Hancock (2001), respectively, and the beta attenuation factors can be found in Mejdahl (1979). Those for ^{87}Rb have been updated by Readhead (2002a,b). Formulas modified to account for the occurrence of organic matter in the sediment matrix were developed by Divigalpitiya (1982) and can be found in Lian et al. (1995). The formula used to estimate the dose rate due to cosmic rays at a sample site is found in Prescott and Hutton (1994).

Radioisotope concentrations are determined by taking a representative subsample of the bulk (unprepared) fraction used for dating and employing a variety of laboratory analyses that give the concentration of the parent isotopes or of the various daughter products in the decay chains, from which the concentration of the parents can be determined. The assumption is that the isotopes of interest have remained in secular equilibrium throughout the lifetime of the sedimentary deposit that was sampled. In most instances this is the case, but there are situations in which the parent isotopes and/or the daughters have been removed from the sample site or have been introduced to it. For example, ^{222}Rn (a daughter product of ^{238}U) can leave by gaseous diffusion, and U, which is soluble in groundwater, can be leached away from a site or

be taken up by organic matter at the site. It is often difficult or impossible to determine if radioactive disequilibrium has occurred at a sample site in the past, but one can easily determine if it exists at the time of sample collection. This is done by measuring the activities of various daughter products in the decay chains. Some forms of disequilibrium are time dependent and, if significant, should be taken into account when the dose rate is calculated. Papers relevant to TL dating that discuss and assess radioactive disequilibrium for various sedimentary environments, and suggest ways to deal with it, are those of Prescott and Hutton (1995), Olley et al. (1996, 1997, 2004), Stokes et al. (2003), and Berger (2006).

Another variable that can complicate dosimetry at a sample site is variability in the sediment matrix. Beds or lenses of significantly different material (e.g., zones where heavy minerals are concentrated) that occur within 50 cm of the sample location can contribute significantly to the gamma dose delivered to the sample site. In such cases, the dose rate calculated from the material in the sample alone will be inappropriate, and *in situ* measurements are required. This can be done in the field using a portable gamma-ray spectrometer (Figure 4). Other options, albeit more laborious, are to calculate the dose rate from samples collected from the various strata or lenses surrounding the sample site and use these data to model the dose absorbed by the sediments to be dated (see Aitken, 1985, Appendix H) or to use *in situ* TL dosimeter capsules (Aitken, 1985, 1998).

Barring the existence of significant disequilibrium in the U decay chain, the largest uncertainty in the dose-rate calculation is commonly that associated with pore water content in the sediment matrix. This is especially so in cases in which water content has fluctuated significantly over time, such as would be expected in environments in which standing or flowing water once existed, such as former lake bottoms, bogs, glacier or ice



Figure 4 Example of a portable gamma-ray spectrometer (Exploranium model GR256). The instrument consists of a long cylindrical probe (shown inserted at the sample site) in which a sodium iodide crystal and a photomultiplier tube together serve as the detector; the probe is connected to electronics that record the gamma-ray spectra. Also shown at the lower left-hand side of the case is a handheld computer that is used to control the spectrometer and to upload data. Portable gamma-ray spectrometers are able to detect gamma rays from ^{40}K and from late daughters in the U and Th decay chains from which the concentrations of the parent isotopes can be calculated, assuming secular equilibrium.

sheet beds, and stream beds. In such cases, usually the best that can be done is to use an average water content value with an error term that will cover both extremes – dry and water saturated – at 2 standard deviations. In such extreme cases, the uncertainty associated with the past water content may contribute $\pm 10\%$ to the uncertainty in the calculated TL age. If one wishes to date sediments extracted from organic-rich environments, such as paleosols or fossil peat bogs, then the organic content of the bulk sample needs to be measured as well as water content, and as mentioned previously, the degree of disequilibrium must be assessed because organic matter absorbs uranium from groundwater (e.g., [Lian et al., 1995](#)).

Sample Collection and Preparation

Although sample collection for TL dating is relatively simple, it should be done by someone from the dating laboratory, together with a geoscientist familiar with the environmental history of the site and the region, so that the best sampling strategy can be established.

A sediment sample is usually collected by inserting a metal or opaque hard plastic open-ended cylindrical container into the cleaned section face (see [Wagner, 1998, Figure 94](#)). The container is then carefully excavated and the ends are promptly sealed to preserve the water content. Another technique suitable for cohesive deposits involves carving out a sediment block (see [Lian and Huntley, 2001, Figure 8\(b\)](#)) and wrapping it in several layers of aluminum foil for protection; a few millimeters of sediment is removed from the light-exposed surfaces in the laboratory before the block is sampled for dating. Sample collection is discussed in more detail by [Lian and Huntley \(2001\)](#) and [Wallinga \(2002\)](#).

The amount of sediment needed for TL dating depends on the desired grain size, the available amount of quartz and/or feldspar, and which dosimetry techniques will be employed, so it is useful to have some prior information about the grain size content of the sample. Typically, ~ 1 kg of material is sufficient, but there have been instances in which less than 1 g has been enough and other cases in which more than several kilograms have been needed. Obviously, for very large samples one has to be concerned about sampling across regions with very different ages.

Samples are prepared in a special laboratory under subdued orange or red light. Preparation usually involves two steps: (1) chemical treatments to remove carbonates, organic matter, and feldspars (if only quartz is to be dated) and to disperse (deflocculate) clay, and in some cases treatments are administered to remove TL-blocking oxidation coatings from the mineral grains; and (2) separation of the sample into the desired mineral types and grain sizes using sieves, heavy liquids, and settling in water if the fine silt fraction is required. The most common grain sizes used for TL dating are in the fine silt range – 2–8 or 4–11 μm diameter – or a range in the fine to medium sand fraction, such as 90–125 or 250–350 μm diameter. These size fractions are chosen because they are common in most depositional environments and because the microdosimetry has been well defined for them, but other grain sizes can be used as well. The laboratory techniques used to prepare a sample for dating are discussed in detail by [Wintle \(1997\)](#);

see also [Lian et al. \(1995\)](#) or [Lian and Shane \(2000\)](#) for samples that contain a significant organic component.

Environments Suitable for TL Dating

Any sedimentary deposit that contains an adequate amount of quartz and/or potassium feldspar grains in the desired size fraction, which has been exposed to sufficient sunlight or heat prior to final burial, has the potential for yielding an accurate TL age. Therefore, the most suitable environments are those in which the sediment has been transported over considerable distances in the atmosphere and/or have been reworked by other processes at the ground surface for an extensive amount of time before finally being buried. Much of TL dating of sediments has therefore been done on eolian deposits such as loess and beach dune sediment. Reviews of the application of TL dating to eolian sediments (especially loess) can be found in [Readhead \(1988\)](#), [Berger \(1988a,b, 1992a, 1994, 1995\)](#), [Wintle \(1990, 1993\)](#), [Zöller and Wagner \(1990\)](#), [Prescott and Robertson \(1997\)](#), and [Wagner \(1998\)](#). One of the most extensive and detailed TL dating studies was that done on the 0–800 ka stranded beach dune system in South Australia ([Huntley and Prescott, 2001](#); [Huntley et al., 1993, 1994](#)). These workers were able to demonstrate that their methods could produce TL ages from quartz sand between zero and at least 500 ka that were consistent with high sea-level stands that had been associated with marine oxygen isotope stages of known age ([Figure 5](#)); older ages were also obtained for dunes occurring further inland, but the accuracy of these ages is questionable ([Huntley and Prescott, 2001](#)). Other notable examples include those of [Berger et al. \(Berger, 1992a; Berger et al., 1994\)](#), who TL dated the fine silt feldspar fraction of loess of known age from Alaska and New Zealand and concluded that accurate ages up to 800 ka could be attained ([Figure 6](#)), although [Prescott and Robertson \(1997\)](#) questioned whether there is enough evidence to make this claim.

Sediments from many other environments have received attention as well, but with mixed results. These include those deposited in fluvial, colluvial, lacustrine, marine, peat bog, and pedogenic systems (see reviews by [Berger, 1995](#); [Prescott and Robertson, 1997](#); [Wagner, 1998](#); [Walker, 2005](#)). Heated sediments such as those that can be found directly underlying lava flows ([Forman et al., 1994](#); [Robertson et al., 1996](#)) can be dated by TL, and often tephra can be as well ([Berger, 1985, 1992b](#); [Berger and Basacca, 1995](#)).

Factors that Can Lead to an Inaccurate TL Age

The most significant factor that can lead to an inaccurate TL age is that the mineral grains comprising a sample collected from a sedimentary deposit have not all been exposed to sufficient heat or sunlight prior to burial. In such cases, the calculated TL age will lie somewhere between the true age of the deposit and the apparent age of its parent material(s). An inaccurate TL age can also result if an inappropriate protocol is chosen for determination of the equivalent dose; if an inappropriate curve is fitted to the dose–response data ([Wagner, 1998, Figure 81](#));

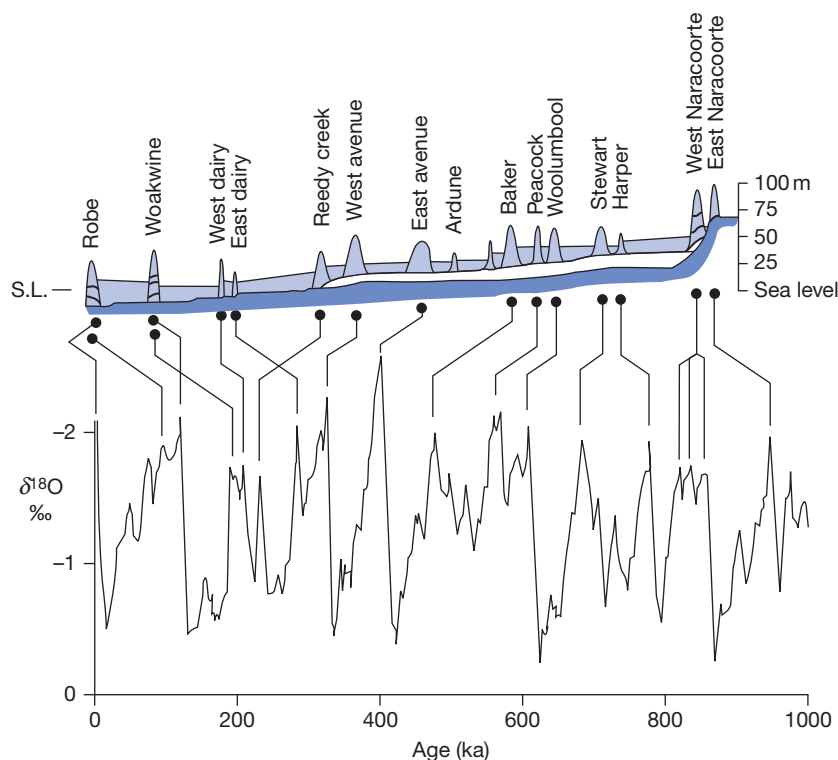


Figure 5 Cross-section through the stranded beach dunes at Robe, South Australia. The black dots show where samples were collected for dating. The TL ages for these are shown to correlate with peaks in the marine oxygen isotope curve, which are associated with sea-level-high stands. In many cases, the TL ages calculated were also consistent with other chronological information. In each case, the mineral dated was quartz, and a regeneration technique (Australian slide method) was used to find the equivalent dose. Two Kopp 7–59 optical filters were used to select the blue TL emission band. For many of the samples, a preheat procedure (166 °C for 33 h) was employed to isolate the rapidly bleached 325 °C TL peak from adjacent peaks. Reproduced from Huntley DJ and Prescott JR (2001) Improved methodology and new thermoluminescence ages for the dune sequence in southeast South Australia. *Quaternary Science Reviews* 20: 687–699, Figure 1. Reproduced with permission. See also [Huntley et al. \(1993, 1994\)](#).

if TL from part of the glow curve that represents electron traps that are difficult to bleach is used to construct the dose-response curve ([Prescott and Mojarabi, 1993](#); [Spooners, 1998](#)); if supralinearity in the low-dose part of the dose-response curve exists and it is not taken into account ([Aitken, 1985](#); [Wagner, 1998](#)); if so-called spurious (i.e., nonradiation-induced) TL is inadvertently measured ([Aitken, 1985](#)); or if there is, or has been, significant radioactive disequilibrium at the sample site. More complex phenomena that can cause problems are the presence of second-order kinetics, for which the position of the TL peaks is dependent on the radiation dose absorbed by a sample ([Aitken, 1985](#); [McKeever and Chen, 1997](#)), and, if feldspars are dated, anomalous fading ([Aitken, 1985](#); [Visocekas, 1985](#); [Wintle, 1973](#)), which refers to the loss of electrons at ambient temperatures, probably by quantum-mechanical tunneling, from traps that are expected to be thermally stable over geological time. The presence of all these malign effects, with perhaps the exception of anomalous fading, can be detected by the absence of a TL plateau. Some workers maintain that the effects of anomalous fading can be avoided by first heating (preheating) the sample at a low temperature for an extended period of time (e.g., 100 °C for 1 week; [Wintle, 1997](#)) before the TL is measured, by dating the fine silt fraction ([Berger, 1995](#)), or by measuring an appropriate TL emission wavelength. Other workers, however, maintain that anomalous fading cannot be circumvented, and if it is

significant in a sample, then the sample cannot be dated unless the fading can be corrected for.

Recent Advances in TL Dating

TL dating of unheated sediments has been largely replaced by optical dating, although TL dating still has an important role in dating heated materials. The main reasons that optical dating is now preferred over TL dating are that (1) the traps sampled for optical dating are much more sensitive to light than those measured for TL, and (2) optical dating protocols allow for the measurement of very small samples, often as small as a single grain of sand. Despite the establishment of optical dating, in some cases it cannot be applied to unheated sediments. These include instances in which anomalous fading cannot be avoided or for which it cannot be corrected (if feldspars are selected) or cases in which the mineral grains of interest emit optically stimulated luminescence that is too dim to be used effectively for dating (as is the case for some quartz samples). These problems have prompted some workers to revisit TL dating. One of most important advances has been the development of a dual-aliquot protocol for TL dating using the light-sensitive isothermally stimulated red TL emission that is prevalent in many quartz samples derived from volcanic rock ([Westaway and Roberts, 2006](#)). This technique allows for the

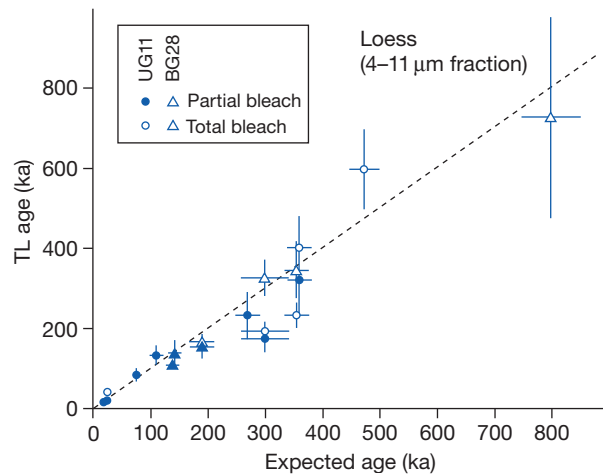


Figure 6 Graph showing TL age versus expected age (~18–800 ka) for various fine silt loess polymineral samples from Alaska and New Zealand; the measured TL was presumably from feldspar. In each case, sample aliquots were preheated before measurement (various temperatures and durations were used; see Berger et al., 1992, for details). Error bars are two standard deviations. Ages for most of the samples were calculated from equivalent dose values found using both the total bleach and partial bleach methods and either a UG11 (ultraviolet pass) or BG28 (blue pass) optical filter between the sample and the detector (photomultiplier tube). Berger et al. concluded that for all except two of their samples, the choice of output filter had no significant bearing on calculated TL age. They also concluded that for their samples, the partial bleach method could produce accurate TL ages up to approximately 350 ka, and that the total bleach method could yield accurate ages for samples in the range 100–800 ka. See also Berger et al. (1994) for a more detailed analysis and discussion of these samples; a critical discussion of this work can be found in Prescott and Robertson (1997). Modified from Berger GW, Pillans BJ, and Palmer AS (1992) Dating loess up to 800 ka by thermoluminescence. *Geology* 20: 403–406; Figure 3.

measurement of very small samples and helps eliminate some of the uncertainty inherent in the standard multiple-aliquot protocols, namely that associated with the assumption that the dose–response of every aliquot in a set is identical. This dual-aliquot method also allows information to be gained on equivalent dose distributions in samples consisting of grains with different age populations.

Another advance in TL dating includes the use of the red TL emitted from most feldspars, which has been shown not to suffer from anomalous fading, at least for the samples studied (Zink and Visocekas, 1997; see also the review by Fattahi and Stokes, 2003). One drawback of the aforementioned techniques is that a special photomultiplier tube is needed to detect the red TL. Finally, workers have also sought to employ the isothermal blue TL signal to extend the age range of quartz (Jain et al., 2005). It should be noted, however, that these new techniques are in developmental stages and therefore should not be used alone for routine dating until they have been thoroughly tested.

See also: **Luminescence Dating:** Electron Spin Resonance Dating; Optical Dating.

References

- Adamiec G and Aitken MJ (1998) Dose–rate conversion factors: Update. *Ancient TL* 16: 37–49.
- Aitken MJ (1985) *Thermoluminescence Dating*. London: Academic Press.
- Aitken MJ (1990) *Science-Based Dating in Archaeology*. London: Longman.
- Aitken MJ (1998) *An Introduction to Optical Dating*. Oxford: Oxford University Press.
- Aitken MJ, Tite MS, and Reid J (1964) Thermoluminescent dating of ancient ceramics. *Nature* 202: 1032–1033.
- Aitken MJ, Zimmerman DW, and Fleming SJ (1968) Thermoluminescent dating of ancient pottery. *Nature* 219: 442–444.
- Balescu S and Lamothe M (1993) Thermoluminescence dating of the Holsteinian marine formation of Herzele, northern France. *Journal of Quaternary Science* 8: 117–124.
- Berger GW (1985) Thermoluminescence dating of volcanic ash. *Journal of Volcanology* 25: 333–347.
- Berger GW (1988a) Dating Quaternary events by luminescence. In: Easterbrook DJ (ed.) *Dating Quaternary Sediments*, pp. 13–50. Boulder, CO: Geological Society of America.
- Berger GW (1988b) TL dating studies of tephra, loess, and lacustrine sediments. *Quaternary Science Reviews* 7: 295–303.
- Berger GW (1990) Effectiveness of natural zeroing of the thermoluminescence in sediments. *Journal of Geophysical Research* 95: 12375–12397.
- Berger GW (1992a) Dating loess up to 800 ka by thermoluminescence. *Geology* 20: 403–406.
- Berger GW (1992b) Dating volcanic ash by thermoluminescence. *Geology* 20: 11–14.
- Berger GW (1994) Thermoluminescence of sediments older than ~100 ka. *Quaternary Geochronology (Quaternary Science Reviews)* 13: 445–455.
- Berger GW (1995) Progress in luminescence dating methods for Quaternary sediments. In: Rutter NW and Catto NR (eds.) *Dating Methods for Quaternary Deposits*, vol. 2, pp. 81–104. St. John's, NL: Geological Association of Canada.
- Berger GW (2006) Trans-Arctic Ocean tests of fine silt luminescence sediment dating provide a basis for an additional geochronometer for this region. *Quaternary Science Reviews* 25: 2529–2551.
- Berger GW and Basacca AJ (1995) Thermoluminescence dating of Late Pleistocene loess and tephra from eastern Washington and southern Oregon and implications for the eruptive history of Mount St. Helens. *Journal of Geophysical Research* 100: 22361–22374.
- Berger GW, Pillans BJ, and Palmer AS (1992) Dating loess up to 800 ka by thermoluminescence. *Geology* 20: 403–406.
- Berger GW, Pillans BJ, and Palmer AS (1994) Test of thermoluminescence dating of loess from New Zealand and Alaska. *Quaternary Science Reviews* 13: 309–333.
- Daniels F, Boyd CA, and Saunders DF (1953) Thermoluminescence as a research tool. *Science* 117: 343–349.
- Divigalpitiya WMR (1982) *Thermoluminescence Dating of Sediments*. Unpublished master's thesis, Simon Fraser University: Burnaby.
- Dreimanis A, Hütt G, Raukas A, and Whippewy PW (1978) Dating methods of Pleistocene deposits: Thermoluminescence. *Geoscience Canada* 5: 55–60.
- Fattahi M and Stokes S (2000) Extending the time range of luminescence dating using red TL (RTL) from volcanic quartz. *Radiation Measurements* 32: 479–485.
- Fattahi M and Stokes S (2003) Dating volcanic and related sediments by luminescence methods: A review. *Earth-Science Reviews* 62: 229–264.
- Feathers JK (1997) The application of luminescence dating in American archaeology. *Journal of Archaeological Method and Theory* 4: 1–66.
- Fleming SJ (1979) *Thermoluminescence Techniques in Archaeology*. Oxford: Clarendon Press.
- Forman SL (1989) Applications and limitations of thermoluminescence to date Quaternary sediments. *Quaternary International* 1: 47–59.
- Forman SL, Pierson J, Smith RP, Hackett W, and Valentine G (1994) Assessing the accuracy of thermoluminescence for dating baked sediments beneath late Quaternary lava flows, Snake River Plain, Idaho. *Journal of Geophysical Research* 99: 15569–15576.
- Franklin AD and Hornyak WF (1990) Isolation of the rapidly bleaching peak in quartz TL glow curves. *Ancient TL* 8: 29–31.
- Huntley DJ and Baril MR (1997) The K content of the K-feldspars being measured in optical dating or thermoluminescence dating. *Ancient TL* 15: 11–13.
- Huntley DJ and Hancock RGV (2001) The Rb contents of K-feldspar grains being measured in optical dating. *Ancient TL* 19: 43–46.
- Huntley DJ, Hutton JT, and Prescott JR (1993) The stranded beach-dune sequence of southeast South Australia: A test of thermoluminescence dating, 0–800 ka. *Quaternary Science Reviews* 12: 1–20.
- Huntley DJ, Hutton JT, and Prescott JR (1994) Further thermoluminescence dates from the dune sequence in the southeast of South Australia. *Quaternary Science Reviews* 13: 201–207.

- Huntley DJ and Prescott JR (2001) Improved methodology and new thermoluminescence ages for the dune sequence in southeast South Australia. *Quaternary Science Reviews* 20: 687–699.
- Jain M, Bøtter-Jensen L, Murray AS, et al. (2005) Revisiting TL: Dose measurement beyond the OSL range using SAR. *Ancient TL* 23: 9–24.
- Krbetschek MR, Trautmann T, Dietrich A, and Stolz W (2000) Radioluminescence dating of sediments: Methodological aspects. *Radiation Measurements* 32: 493–498.
- Lian OB, Hu J, Huntley DJ, and Hicock SR (1995) Optical dating studies of Quaternary organic-rich sediments from southwestern British Columbia and northwestern Washington State. *Canadian Journal of Earth Sciences* 32: 1194–1207.
- Lian OB and Huntley DJ (2001) Luminescence dating. In: Last WM and Smol JP (eds.) *Basin Analysis, Coring, and Chronological Techniques: Tracking Environmental Change Using Lake Sediments*, vol. 1, pp. 261–282. Dordrecht, The Netherlands: Kluwer.
- Lian OB and Roberts RG (2006) Dating the Quaternary: Progress in luminescence dating of sediments. *Quaternary Science Reviews* 25: 2449–2468.
- Lian OB and Shane PA (2000) Optical dating of paleosols bracketing the widespread Rotoehu tephra, North Island, New Zealand. *Quaternary Science Reviews* 19: 1649–1662.
- McKeever SWS (1985) *Thermoluminescence of Solids*. Cambridge: Cambridge University Press.
- McKeever SWS and Chen R (1997) Luminescence models. *Radiation Measurements* 27: 625–661.
- Mejdahl V (1979) Thermoluminescence dating: Beta-dose attenuation in quartz grains. *Archaeometry* 21: 61–71.
- Olley JM, De Deckker P, Roberts RG, et al. (2004) Optical dating of deep-sea sediments using single grains of quartz: A comparison with radiocarbon. *Sedimentary Geology* 169: 175–189.
- Olley JM, Murray A, and Roberts RG (1996) The effects of disequilibria in the uranium and thorium decay chains on burial dose rates in fluvial sediments. *Quaternary Science Reviews* 15: 751–760.
- Olley JM, Roberts RG, and Murray AS (1997) Disequilibria in the uranium decay series in sedimentary deposits at Allen's Cave, Nullarbor Plain, Australia: Implications for dose rate determinations. *Radiation Measurements* 27: 433–443.
- Prescott JR and Fox PJ (1990) Dating quartz sediments using the 325 °C TL peak: New spectral data. *Ancient TL* 8: 32–34.
- Prescott JR, Huntley DJ, and Hutton JT (1993) Estimation of equivalent dose in thermoluminescence dating – The Australian slide method. *Ancient TL* 11: 1–5.
- Prescott JR and Hutton JT (1994) Cosmic ray contributions to dose rates for luminescence and ESR dating: Large depths and long-term time variations. *Radiation Measurements* 23: 497–500.
- Prescott JR and Hutton JT (1995) Environmental dose rates and radioactive disequilibrium from some Australian luminescence dating sites. *Quaternary Science Reviews* 14: 439–448.
- Prescott JR and Mojarrabi B (1993) Selective bleach: An improved partial bleach technique for finding equivalent doses for TL dating of quartz sediments. *Ancient TL* 11: 27–30.
- Prescott JR and Robertson GB (1997) Sediment dating by luminescence: A review. *Radiation Measurements* 27: 893–922.
- Readhead ML (1988) Thermoluminescence dating study of quartz in aeolian sediments from southeastern Australia. *Quaternary Science Reviews* 7: 257–264.
- Readhead ML (2002a) Absorbed dose fraction for ^{87}Rb β particles. *Ancient TL* 20: 25–28.
- Readhead ML (2002b) Addendum to "Absorbed dose fraction for ^{87}Rb β particles". *Ancient TL* 20: 47.
- Roberts RG (1997) Luminescence dating in archaeology: From origins to optical. *Radiation Measurements* 27: 819–892.
- Robertson GB, Prescott JR, and Hutton JT (1996) Thermoluminescence dating of volcanic activity at Mount Gambier, South Australia. *Transactions of the Royal Society of Southern Australia* 120: 7–12.
- Singhvi AK, Sharma YP, Agrawal DP, and Dhir RP (1982) Thermoluminescence dating of dune sands in Rajasthan, India. *Nature* 295: 313–315.
- Singhvi AK and Wagner GA (1986) Thermoluminescence dating and its application to young sedimentary deposits. In: Hurford AJ, Jäger E, and Ten Cate JAM (eds.) *Dating Young Sediments*, pp. 159–197. Bangkok: CCOP Technical Secretariat. CCOP Technical Publication No. 16.
- Spooner NA (1998) Human occupation at Jinmium, northern Australia: 116,000 years ago or much less? *Antiquity* 72: 173–178.
- Stokes S, Ingram S, Aitken MJ, et al. (2003) Alternative chronologies for late Quaternary (Last Interglacial–Holocene) deep sea sediments via optical dating of silt-size quartz. *Quaternary Science Reviews* 22: 925–941.
- Visocekas R (1985) Tunneling radiative recombination in labradorite; its association with anomalous fading of thermoluminescence. *Nuclear Tracks and Radiation Measurements* 10: 521–529.
- Wagner GA (1998) *Age Determination of Young Rocks and Artifacts*. New York: Springer-Verlag.
- Walker MJC (2005) *Quaternary Dating Methods*. New York: Wiley.
- Wallinga J (2002) Optically stimulated luminescence dating of fluvial deposits: A review. *Boreas* 31: 303–322.
- Westaway KE and Roberts RG (2006) A dual-aliquot regenerative-dose protocol (DAP) for thermoluminescence (TL) dating of quartz sediments using the light-sensitive and isothermally-stimulated red emissions. *Quaternary Science Reviews* 25: 2513–2528.
- Wintle AG (1973) Anomalous fading of thermoluminescence in mineral samples. *Nature* 245: 143–144.
- Wintle AG (1990) A review of the current research on TL dating of loess. *Quaternary Science Reviews* 9: 385–397.
- Wintle AG (1993) Luminescence dating of aeolian sands: An overview. In: Pye K (ed.) *The Dynamics and Environmental Context of Aeolian Sedimentary Systems*, pp. 49–58. Boulder, CO: Geological Society of America. Geological Society of America Special Publication No. 72.
- Wintle AG (1997) Luminescence dating: Laboratory procedures and protocols. *Radiation Measurements* 27: 769–817.
- Wintle AG and Huntley DJ (1979) Thermoluminescence dating of sediments. *PACT* 3: 374–380.
- Wintle AG and Huntley DJ (1980) Thermoluminescence dating of ocean sediments. *Canadian Journal of Earth Sciences* 17: 348–360.
- Wintle AG and Huntley DJ (1982) Thermoluminescence dating of sediments. *Quaternary Science Reviews* 1: 31–53.
- Zink AJC and Visocekas R (1997) Datability of sanidine feldspars using the near-infrared TL emission. *Radiation Measurements* 27: 251–261.
- Zöller L and Wagner GA (1990) Thermoluminescence dating of loess – Recent developments. *Quaternary International* 7(8): 119–128.

Optical Dating

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Introduction

Optical dating, sometimes referred to as optically stimulated luminescence (OSL) dating, is based on the fact that natural minerals can absorb and store ionizing energy. If a mineral is exposed to sufficient light (e.g., sunlight), or heated to a high temperature ($\sim 500^\circ\text{C}$), some or all of this stored energy will be lost.

In the environment, the energy absorbed by a mineral comes from radiation emitted from radioisotopes within the mineral grain, from its immediate surroundings, and from cosmic rays. The rate at which a dose of radiation energy is absorbed by a mineral (the dose rate) is therefore proportional to the concentrations of radioisotopes, and to the intensity of cosmic rays at the sample site. Simply put, for a given mineral sample, what is needed to determine an optical age is (1) the laboratory dose of radiation that produces the same intensity of OSL as did the environmental radiation dose (called the equivalent dose) and (2) a measure of the environmental dose rate. An optical age is simply the equivalent dose divided by the dose rate. Optical dating therefore gives the time elapsed since mineral grains were last exposed to sufficient sunlight or heat.

Optical dating developed from thermoluminescence (TL) dating in the mid-1980s (Huntley et al., 1985). Its main advantage over TL dating is that only the most light-sensitive OSL signal is sampled, which allows for much younger samples to be dated. Moreover, the optical dating experiments needed to produce an equivalent dose are generally simpler than those required for TL dating, which usually results in better precision. Optical dating has therefore all but replaced TL dating for unheated sediments. Optical and TL dating are part of a family of chronological techniques collectively referred to as trapped charge dating. Related methods include electron spin resonance (ESR) dating and radioluminescence dating (e.g., Krbetschek et al., 2000); discussion of ESR dating can be found elsewhere in this encyclopedia.

The Mechanism Responsible for OSL

Luminescence refers to the light emitted by matter when it is subjected to external stimulation. When this external stimulation is photons of visible light, the luminescence emitted is referred to as OSL; it is called infrared-stimulated luminescence (IRSL) if the stimulation is from far-infrared photons. In the discussion below, for simplicity, both phenomena are referred to as OSL unless additional detail is needed. The physical models used to explain the OSL phenomena are complex and not fully understood; general, and varied, accounts, most of which are relevant to dating, can be found in Aitken (1998), Wintle (1997), McKeever and Chen (1997), Bailey (2001), Bøtter-Jensen et al. (2003), Preusser et al. (2009), and McKeever and Yuhikara (2011).

Briefly, crystals, such as those comprising natural minerals, are not perfect, and contain impurities and structural defects, some of which are attractive to electrons that have been freed from their normal sites in the valence bands of atoms; these sites are referred to as electron traps. Free electrons are produced when atoms absorb ionizing radiation that occurs in the natural environment. Free electrons may subsequently be held in traps for a time that depends on the strength, or thermal lifetime, of the particular trap. For dating, only those traps that can hold electrons at environmental temperatures over geological time (thermally stable or 'deep' traps) are of interest. Further detail can be found in reference to TL dating.

Some of these traps can be emptied (bleached) by exposure to light. Of the light-sensitive traps responsible for OSL, some can easily be emptied by a brief exposure (e.g., a few seconds) to direct sunlight, but there are others for which a much longer sunlight exposure is needed. In any one mineral grain, therefore, there are many different types of electron trap that can contribute to OSL, and these have different sensitivities to light. In the laboratory, OSL is produced by exposing a sample to light (photons) of a specific wavelength, or wavelength range, thus freeing electrons from light-sensitive traps. Some of these freed electrons quickly recombine at other sites where excess energy is given off as luminescence (the OSL). The output is what is commonly referred to as a shine-down curve, which consists of OSL intensity plotted as a function of light stimulation (sometimes referred to as excitation) time (Figure 1(a)). For a given mineral grain, the intensity of the measured OSL is related to the number of filled electron traps. If irradiation continues, traps will fill until an additional dose of radiation will not result in more OSL being measured; this condition is referred to as saturation and it is one of the factors that sets the upper age limit for the method. The shape of the shine-down curve reflects the different electron trap populations present in the mineral, the initial OSL coming from the most light-sensitive traps, and subsequent OSL arising from traps that are not as light sensitive. A single shine-down curve can, therefore, be composed of the sum of several exponential decay curves, each associated with a different electron trap population with a specific sensitivity to light. Smith and Rhodes (1994) concluded that the shine-down curve for quartz consists of three components which they referred to as fast, medium, and slow, depending on how easily they could be bleached, while others have identified at least seven components for this mineral (Jain et al., 2003). It is not known exactly how many components exist in any one mineral grain, but they could theoretically form a continuum. It is becoming generally accepted that for quartz the OSL measured over the initial part of the shine-down curve, that recorded from when the stimulating beam is switched on until $\sim 5\%$ of the initial OSL signal remains (the fast component), is thermally stable and can be used to determine accurate age estimates, while the OSL recorded over subsequent time intervals (from the

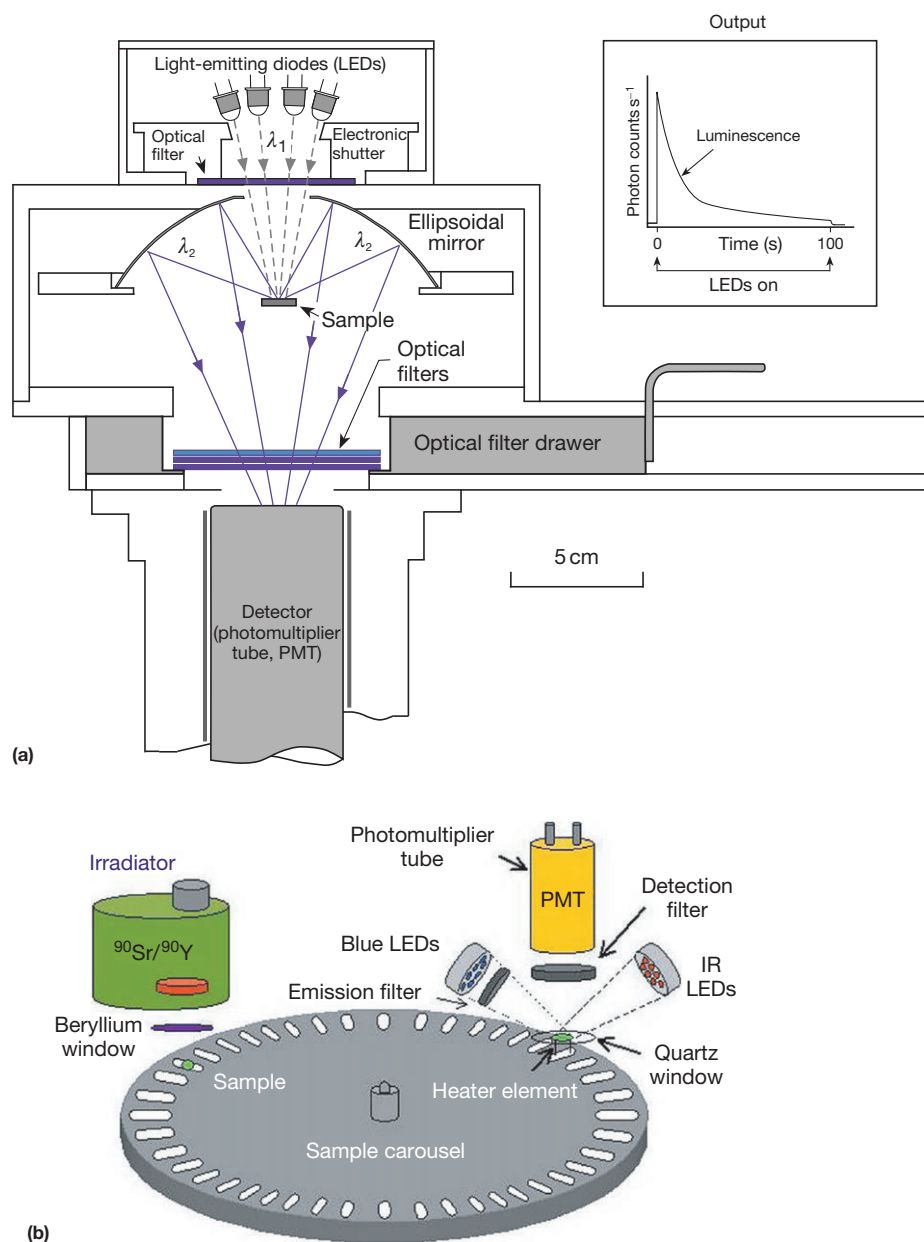


Figure 1 Instrumentation used to stimulate and record optically stimulated luminescence (OSL). (a) Schematic illustration of a single-sample apparatus. Light (photons) of wavelength λ_1 are emitted from high-power blue or infrared light-emitting diodes (LEDs) and are focused onto a sample aliquot, which is inserted manually into the apparatus by means of a drawer that extends perpendicular to the drawing. Photons of a shorter wavelength, λ_2 (the OSL), are emitted in response and are collected by an ellipsoidal mirror, which, in turn, directs them to the photomultiplier tube (PMT; the detector). A discussion of the principles behind the photomultiplier tube can be found elsewhere in this encyclopedia. The optical filter(s) immediately above the mirror are used to absorb photons from the LEDs that are of a shorter wavelength (higher energy) than λ_1 , while the optical filters immediately above the PMT are used to block scattered photons from the stimulation source (LEDs, in this case) and other unwanted photons (e.g., see Lian and Huntley, 2001). The output from the PMT is fed through electronics (not shown), and the result is a shine-down curve, such as the one depicted in the inset figure. The shine-down curve shows the OSL recorded as a function of stimulation time, and reflects the progressive depletion of trapped electrons. The greater the number of filled electron traps at the time of sample collection, the greater the intensity of measured OSL when the LEDs are switched on. Note that the shine-down curve does not show a simple exponential decay, but the sum of various decay curves, each representing a population of electron traps with a particular sensitivity to light. Based on Figure 3 of Lian OB and Huntley DJ (2001) *Luminescence dating*. In: Last WM and Smol JP (eds.) *Tracking Environmental Change Using Lake Sediments, Vol. 1: Basin Analysis, Coring, and Chronological Techniques*, pp. 261–282. Dordrecht: Kluwer. (b) Schematic representation of one of the automated luminescence readers that is available commercially: the Risø TL/OSL reader. In its basic form, the machine consists of a sample carousel which is used to move aliquots (up to 48 in a run) either to a position where they can be irradiated (by a shielded $^{90}\text{Sr}/^{90}\text{Y}$ beta source in this case), or to a position under a PMT where luminescence (OSL or IRSL) can be measured. In the latter position, the temperature of the aliquot can be raised, as is required for TL or single-aliquot OSL measurements. Optional attachments (not shown) include a system for measuring radioluminescence, an X-ray generator (as an alternative to the beta-source irradiator), and an apparatus for irradiating and measuring single grains. See Bøtter-Jensen et al. (2003) for details. Drawing reproduced from Bøtter-Jensen L, McKeever SWS, and Wintle AG (2003) *Optically Stimulated Luminescence Dosimetry*. Amsterdam: Elsevier.

medium and slow components, etc.) should be avoided (Bai-ley, 2010).

Generating and Measuring OSL for Optical Dating

The first instruments designed to measure OSL (e.g., as in Figure 1 of Huntley et al., 1985) were developed from those used to measure TL. Instruments used for optical dating are much simpler than those used for TL dating as they do not usually require that the sample chamber be evacuated, and a sample can usually be measured in air, whereas for TL dating an inert gas atmosphere is needed. These simplified measurement requirements usually lead to better precision. The first optical dating instruments could only hold one aliquot of sediment at a time, aliquots had to be changed manually, they required a separate irradiator facility, and gas lasers, such as the argon ion laser used by Huntley et al. (1985), were used for the stimulating light source. Expensive lasers were soon replaced by filtered quartz-halogen or xenon-arc lamps, and then eventually by more affordable high-power light-emitting diode (LED) arrays. The instruments also became automated with the irradiator unit mounted on board, and they could measure many aliquots in a run while keeping them at an elevated temperature, as is usually required for single-aliquot techniques (Figure 1).

As with the earlier TL dating apparatuses, current optical dating instruments require a photomultiplier tube (PMT) as a detector, and this typically is mounted above the sample (aliquot) position with space for optical filters between them (Figure 1(a) and 1(b)). Recent advances in commercial instrumentation include attachments to measure pulsed OSL, radioluminescence, focused semiconductor lasers as stimulating sources for single-grain dating, and X-ray generators as an alternative to shielded beta-source irradiators. Reviews and discussion of optical dating instrumentation can be found in Aitken (1998), Bortolot and Bluszcz (2003), Bøtter-Jensen et al. (2003), and Thomsen et al. (2008a).

Optical Dating

As mentioned earlier, what is required to calculate an optical age is to measure the environmental dose rate, and to make an estimate of the equivalent dose. These are discussed in turn below.

Determining the Environmental Dose Rate

The dose of radiation absorbed by the sediment grains of interest comes from alpha and beta particles, and from gamma rays emitted during the decay of ^{235}U , ^{238}U , ^{232}Th , ^{40}K , and ^{87}Rb , and their daughter products, both within the mineral grains and in their surroundings; another component of the absorbed dose comes from cosmic rays, which consist of highly energetic particles originating from unknown sources in the universe. All of these quantities have to be determined and included in the dose-rate calculations, as well as the pore-water content, and, in some cases, the organic content of the sediment matrix, as these attenuate or absorb radiation differently

than does mineral matter. The term dose refers to the amount of radiation energy absorbed per kilogram of matter. The SI unit for dose is the gray (Gy), where $1\text{ Gy} = 1\text{ J kg}^{-1}$. Dose rate commonly is denoted by $\dot{D}_T$ (where the subscript 'T' indicates the total dose rate, that due to alpha, beta, gamma, and cosmic-ray radiation) and it has units of Gy ka^{-1} or mGy a^{-1} .

General formulae used to calculate dose rate from the concentrations of radioisotopes in the environment, and from the intensity of cosmic rays, can be found in Aitken (1985, 1998), and in Prescott and Hutton (1994), respectively. Dose-rate formulas that take into account organic content can be found in Lian et al. (1995). Factors that can reduce the accuracy and precision of a calculated dose-rate value are (1) radioactive disequilibrium in the U and/or Th decay chains, (2) mineral inhomogeneity in the sediment matrix, (3) uncertainty about a sample's depth beneath the ground surface over the lifetime of the deposit (which adds uncertainty to the cosmic-ray dose rate), (4) uncertainty about the water content history at the sample site, and (5) analytical uncertainties associated with the procedure and instrumentation used to measure radioisotope concentrations. These topics are discussed in more detail elsewhere in this encyclopedia.

Estimating the Equivalent Dose

Equivalent dose, measured in grays (Gy, defined above), is found by recording how a sample's OSL intensity responds to increasing doses of laboratory radiation; this is referred to as a sample's dose response. The term equivalent dose is used because the laboratory radiation used to construct a sample's dose response consists only of gamma rays or beta particles (or sometimes X-rays), whereas in the natural environment the radiation absorbed by a sample comes from alpha and beta particles, gamma rays, and cosmic rays. Equivalent dose is represented by the notation ED , D_{eq} , or D_e , and is sometimes referred to loosely (and incorrectly) as paleodose.

One of the main differences between TL dating and optical dating is that for optical dating there is no way of selecting during measurement only those electron traps that are thermally stable over geological time. To overcome this problem, the sample aliquots are heated (preheated) prior to measurement. The objective of the preheat is to empty all of the thermally unstable electron traps, while leaving most of the thermally stable traps full; the former traps are filled during laboratory irradiation. The temperature and duration used for the preheat is, to a degree, specific to the sample, so in practice the severity of the preheat is increased until the equivalent dose determined becomes constant. Preheating a sample does, however, have some malign side-effects that have to be assessed, and, if severe, taken into account. One of these unwanted side-effects is that preheating changes the OSL sensitivity of the mineral grains, and another is that it causes electrons to be transferred from thermally stable light-insensitive traps to the thermally stable light-sensitive traps relevant to dating. The degree of this thermal transfer depends on the amount of electrons in thermally stable light-insensitive traps, and this, in turn, is related to the amount of light exposure that the sample received in the environment prior to burial, and on the magnitude of the laboratory radiation dose(s) given to the sample.

Several procedures are currently available that can be used to estimate equivalent dose. These protocols can be divided into those which require many aliquots of a sample, and those that need only one aliquot. Of the latter, a single aliquot can comprise hundreds or thousands of individual sediment grains, or it can consist merely of one grain of sand. Methods used to find equivalent dose are summarized below, while detailed discussions, with various emphases, can be found in, for example, Aitken (1998), Murray and Wintle (2000, 2003), Duller (2004), Wintle and Murray (2006), Lian and Roberts (2006), Jacobs and Roberts (2007), Duller (2008), and Wintle (2008).

Multiple-Aliquot Techniques

The multiple-aliquot procedures were the first protocols designed for optical dating and they evolved from those used for TL dating. They include the additive-dose method, the additive-dose with late-light subtraction technique, the additive-dose with thermal-transfer correction (ADTT) method, and the regeneration methods. All these protocols require that ~20–50 aliquots of a sample to be prepared. These aliquots are then given various laboratory radiation doses, in some cases several are exposed to light (i.e., they are bleached for a specific duration using a filtered laboratory light source, or natural sunlight) before or after irradiation, depending on the protocol, and then their OSL is measured using apparatuses like those described above. The OSL data are used together to deduce the sample's dose response, and from that an equivalent dose is estimated.

The additive-dose method is the simplest protocol, and it was the one used in the initial trials by Huntley et al. (1985). This method requires that several aliquots of a sample are left as is (the so-called naturals, commonly denoted by N), while several others are each given progressively increasing doses of laboratory radiation ($N + \text{dose}$). All the aliquots are then preheated and measured, usually after a delay. The luminescence intensity is then plotted as a function of laboratory radiation dose. A curve is fitted to these data, and extrapolated to where it intercepts the dose axis, the intercept point being the equivalent dose (e.g., as is shown in Figure 1A of Wintle, 2008). If a sample consists of grains that have all received sufficient sunlight exposure prior to burial, and if the shine-down curve consists of only one component, or of components that all represent light-sensitive thermally stable traps (which is commonly not the case for quartz – see, e.g., Bailey, 2000, but which might be the case for feldspar), then it should not matter which part of the shine-down curve is used to construct the dose response. That is, the intercept values will be constant with stimulation time; this is sometimes referred to as a shine plateau. On the other hand, intercept values will increase with stimulation time if the sample consists of grains that have not received sufficient sunlight prior to burial. Unfortunately, this indicative rise in intercept values with stimulation time will not occur if the sample consists of a population of grains that have been exposed to sufficient sunlight, mixed with one or more populations of grains that have not. A modification of the additive-dose technique is the late-light subtraction method (Aitken, 1998): aliquots are irradiated, preheated, and

measured in the manner described above, except that the OSL recorded during an interval of time at the end of the shine-down curve (the so-called late-light) is subtracted from the initial OSL used to construct the dose response. The idea is that this late-light consists mainly of scattered photons from the stimulation beam, background photon counts from the detector (PMT), and OSL arising from electrons remaining in hard to bleach traps. The time interval selected for the late-light clearly depends on the power of the stimulating beam at the sample.

The main limitation of the additive-dose and late-light subtraction methods is that if thermal transfer resulting from the preheat procedure is significant, a calculated optical age will be too old. This problem is, however, taken into account with the ADTT technique (Huntley et al., 1993). For the ADTT technique, a subset of the irradiated aliquots that are to be used to construct the dose–response curve are given a short exposure to natural sunlight or laboratory light ($N + \text{dose} + \text{bleach}$) so as to empty all of the electrons in light-sensitive traps, but leaving full the light-insensitive traps that are the source of the thermal transfer. The two sets of aliquots are then preheated together and measured. Curves are fitted to both data sets, and these are extrapolated to where they intersect above the dose axis, and the dose there is taken as the intercept value. The ADTT method not only corrects for thermal transfer, but accounts for background photon counts from the detector, scattered photons from the stimulation source, and Raman-scattered photons. As with the additive-dose method, dose-intercept values should be constant with stimulation time if only thermally stable components of the shine-down curve are sampled and all the light-sensitive traps in all the sample grains measured had been emptied in the environment before burial. An example of the ADTT method is shown in Figure 2. If extrapolation of the dose–response curve(s) to the dose axis is large, as is the case for most old samples, then the regeneration method should be used. In its most basic form, the regeneration method involves giving several sample aliquots an extended exposure to light so as to empty as many light-sensitive traps as possible, and then using these aliquots to construct a regenerative dose–response curve by giving them various radiation doses ($N + \text{bleach} + \text{dose}$). The OSL intensity from a natural aliquot is then interpolated onto the regenerative dose–response curve, the equivalent dose being the corresponding value of this point on the dose axis; an example is shown in Figure 1B of Wintle (2008). This is in essence the technique that is used in single-aliquot dating, and it is described in more detail in that context below. In its most complete form, however, the multiple-aliquot regeneration method involves constructing both a regenerative dose–response curve and an additive-dose curve (Figure 3). The regenerative-dose data set is constructed so that it largely overlaps the additive-dose data set. If the initial laboratory bleach has caused no sensitivity change in the regenerative set of aliquots, then the dose response deduced from it should be identical to that found from the additive-dose aliquots. If this is the case, then the additive-dose data set is shifted along the dose axis so that it coincides with the regenerative data set, and a single curve is fitted to all the data (Figure 3). The magnitude of the shift is equal to the equivalent dose after corrections for incomplete laboratory bleaching and for the slight decay in OSL incurred during normalization (discussed

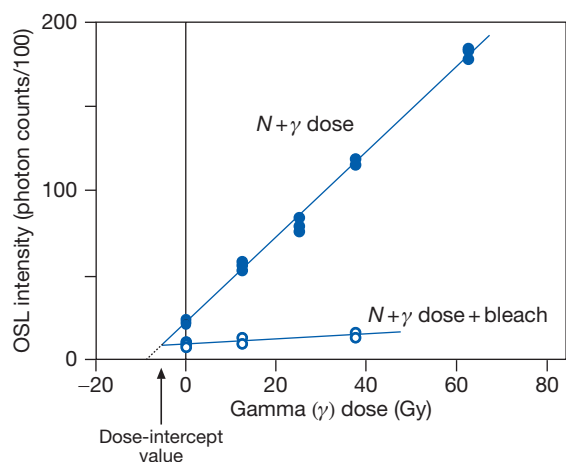


Figure 2 Example of the additive-dose with thermal-transfer correction (ADTT) method, as used for multiple-aliquot dating. The upper curve ($N + \gamma$ dose) is an additive-dose curve, while the lower curve ($N + \gamma$ dose + bleach) is used to correct for thermal transfer. Aliquots were irradiated with a ^{60}Co γ source in this case. The data shown is from the IRSL recorded during the first 5 s of stimulation. Both curves are extrapolated to where they intersect above the dose axis, and this point is taken as the dose-intercept value. Dose-intercept values are commonly determined as a function of stimulation time, and, if they are constant, an average value is usually taken as the equivalent dose, after small correction for the decay in luminescence (trapped electrons) that has resulted from the normalization procedure; see the text for additional detail. The data shown are from the violet (400 nm, 3.1 eV) luminescence emitted during the stimulation of the K-feldspar fraction in fine-silt polymineral aliquots with near-infrared (880 nm, 1.4 eV) light. IRSL was detected by an EMI 9365Q photomultiplier tube behind Schott BG-39 and Kopp 5–57 optical filters. The aliquots were preheated at 160 °C for 4 h before measurement. The sample is a 2.4 ka eolian deposit from south-central British Columbia, Canada (sample CCL4 of Lian and Huntley, 1999). Based on Figure 7 of Lian OB and Huntley DJ (1999) Optical dating studies of post-glacial aeolian deposits from the south-central interior of British Columbia, Canada. *Quaternary Science Reviews* 18: 1453–1466.

below). In practice, sensitivity change in the regenerative data set is identified by observing an intensity-scaling parameter in the curve-fitting algorithm. This regeneration method was introduced by Prescott et al. (1993) and is often referred to as the Australian slide method because of its extensive use in TL dating of a significant stranded beach dune sequence in southern Australia.

Normalization for Multiple-Aliquot Techniques

One of the drawbacks of the multiple-aliquot techniques is that the aliquots comprising a set can never be made to be identical. The reason for that is that a sample aliquot typically contains thousands of individual mineral grains, and these all have different OSL sensitivities. That is, for any given grain in a sample, the OSL emitted per dose of radiation is different, some grains emitting much more OSL than others. The OSL measured from an aliquot may, therefore, be dominated by that emitted from only a very small number of the grains. To compensate for this intrinsic difference between aliquots, all aliquots in a set are initially normalized. This is usually done

by recording the OSL intensity that results from a brief exposure to the stimulation beam (commonly referred to as a short shine) before the aliquots are given a laboratory dose. The severity of the short shine depends on the OSL intensity of the natural sample; if the affect of the short shine on the trapped electron population relevant to dating is significant, it can be accounted for when the equivalent dose is determined. Normalization factors are calculated for every aliquot by dividing the short-shine OSL intensity measured from them by the average short-shine OSL intensity of the entire set. These factors are used to reduce the scatter in the dose–response data.

Other normalization procedures include those for which the sample aliquots are given a short shine following a radiation dose and preheat that are administered after the final OSL measurements are made. These methods are used when the OSL intensities measured from the natural aliquots are too low, or if there is concern that the different laboratory doses (and the bleach treatment) administered to the various aliquots used to construct a sample's dose response have resulted in sensitivity change; a thorough discussion is given by Aitken (1998, p. 98). If the normalization factors are produced from the OSL measured from the natural aliquots, then the normalized dose–response data can also be used to give information on the degree of sunlight exposure a sample has experienced in the environment; this is discussed in detail by Huntley and Berger (1995) and Ollerhead et al. (2001).

Single-Aliquot Techniques

One of the advantages that optical dating has over TL dating is that only a small proportion of the relevant electron traps needs to be sampled to construct a dose–response curve. This was realized by Huntley et al. (1985) during the initial trials of optical dating and they suggested that a sample's dose response could be constructed from just a single aliquot of prepared sediment. This prompted other workers to develop techniques that required only one or two aliquots of a sample to estimate equivalent dose. Indeed, one can now calculate an optical age value from a single grain of sand. The main advantage of this is that it is possible to determine the desired optical age value of a sample (which usually is its burial age) which consists of grains that belong to different equivalent-dose populations. Moreover, since all the measurements are performed on a single aliquot, normalization is not required. All the single-aliquot methods described below therefore involve repeatedly preheating, stimulating (with light), and dosing the same sample aliquot.

The first single-aliquot protocols were developed in the early 1990s (e.g., Duller, 1991) and involved constructing an additive-dose curve for a sample, and extrapolating it to the dose axis. These single-aliquot additive-dose (SAAD) techniques were followed by the single-aliquot regeneration and added dose (SARA) method where the natural OSL is plotted on a regenerative-dose curve and the place where this point intersects the dose axis is taken to be the equivalent dose. In reality, these early single-aliquot protocols were not truly single aliquot as at least one additional aliquot is required to correct for the progressive loss of OSL signal and (or) to monitor sensitivity change. In the use of additional aliquots it is presumed, as with the multiple-aliquot techniques, that all the

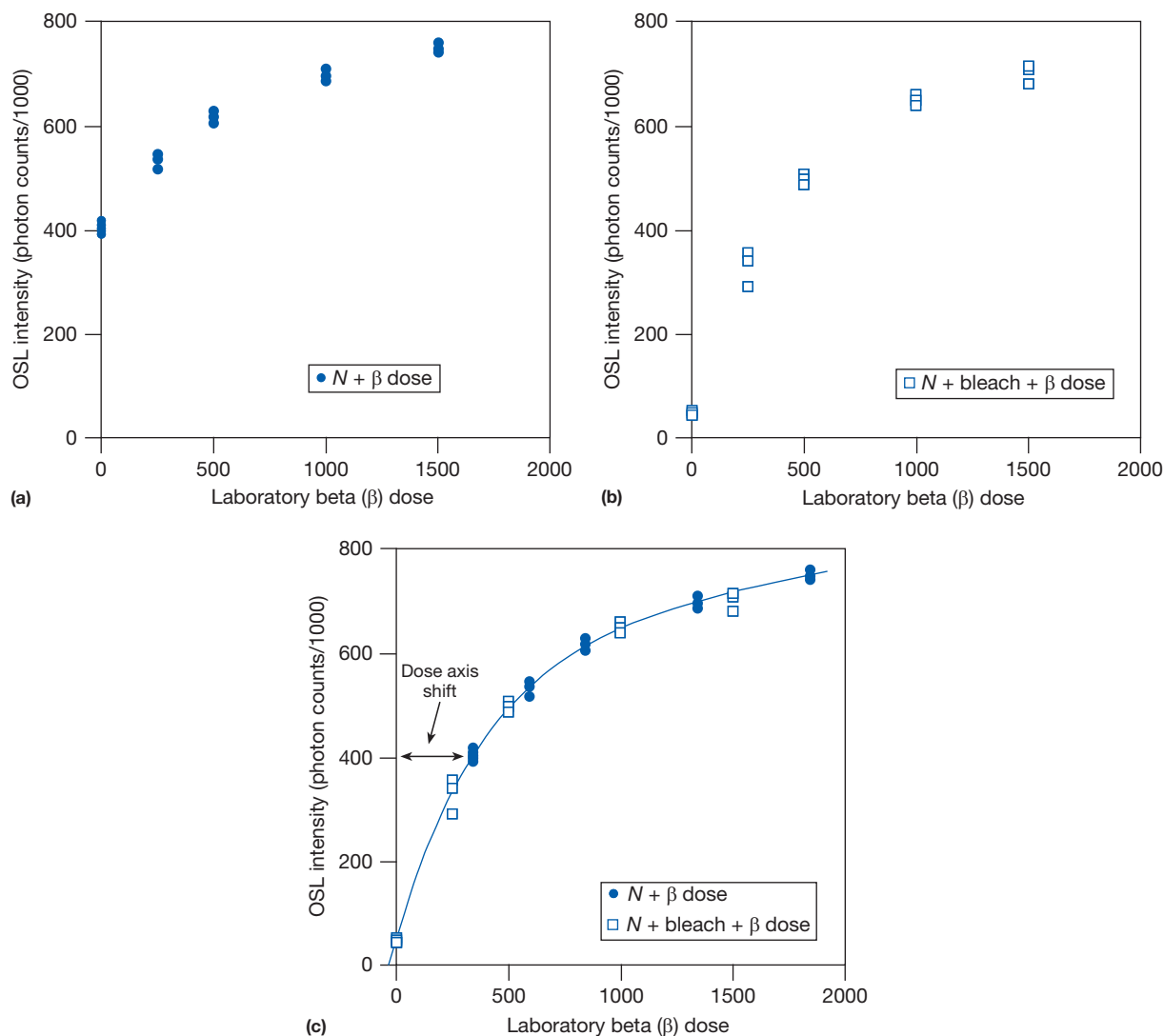


Figure 3 Illustration of the Australian slide method, a combined additive- and regenerative-dose method used for determining equivalent dose; it is especially useful in cases where extrapolation of the additive-dose curve to the dose axis is large. The data in (a) are the additive-dose data ($N + \beta$ dose), while the data shown in (b) are those from aliquots that have received an extended exposure to light before being given various laboratory doses of radiation – the regenerative-dose data set ($N + \text{bleach} + \beta$ dose). Irradiation was from a $^{90}\text{Sr}/^{90}\text{Y}$ β source in this case. The aliquots were preheated at 140 °C for 7 days before measurement. The additive-dose data are shifted along the dose axis until they coincide with the regenerative data, the magnitude of the shift being equal to the equivalent dose, after a correction for incomplete laboratory bleaching and for decay of the OSL signal (trapped electrons) incurred during normalization. If there has been no sensitivity change in the regenerative data, then both data sets should define the same curve. The data shown is from lacustrine sediment from Lake Poukawa, New Zealand, that gave an apparent optical age of 83 ± 7 ka (sample LPCS22 of [Shane et al., 2002](#)). The mineral type and size, and the measurement conditions, are the same as those described in [Figure 2](#), except in this case the photomultiplier tube was an EMI 9235QA, and a Kopp 5–58 filter was used instead of a Kopp 5–57 filter.

grains in a sample have had the same bleaching history and have the same OSL sensitivity. For that reason, the SARA technique has become limited to heated materials. Reviews of the early single-aliquot methods can be found in [Aitken \(1998\)](#).

The SAAD method was later modified so that it required only one aliquot, but the problem of sensitivity change due to repeated preheating and measurement remained. This led to the development of the single-aliquot regenerative-dose (SAR) technique, which incorporated a method for correcting for the sensitivity change. The first SAR protocols monitored sensitivity change by observing the response of the 110 °C TL glow-curve peak during preheating, but this practice was eventually

replaced by observing a sample's OSL response to a small radiation dose (a so-called test dose) that is administered at the end of each regenerative-dose cycle (e.g., [Murray and Wintle, 2000, 2003](#); [Roberts et al., 1998](#)). The SAR protocol described by [Murray and Wintle \(2000, 2003\)](#) and [Wintle and Murray \(2006\)](#), or some slight modification of it, currently is the method favored by most optical dating practitioners; it is illustrated, with additional detail, in [Figure 4](#). At present the protocol is used almost exclusively to date quartz as it has been found to be inadequate for correcting for sensitivity change in feldspar (e.g., [Blair et al., 2005](#)), but there have been several cases where it has been shown to be applicable

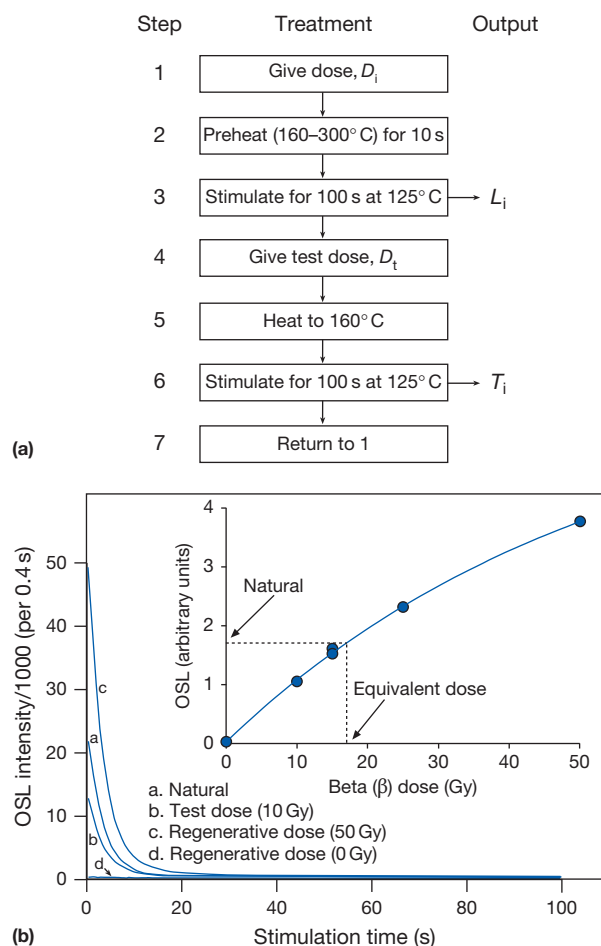


Figure 4 Example of the basic single-aliquot regenerative-dose (SAR) method of Murray and Wintle (2000). The procedure, based on that described in Table 1 of Murray and Wintle (2000), is shown in (a): For the natural sample, $i = 0$ and $D_0 = 0$ Gy. The OSL used for regenerative-dose data points (L_i), and the OSL measured from the test doses (T_i), typically are derived from that measured during the first second or less of stimulation thereby selecting the thermally stable fast component of the shine-down curve (although longer intervals are sometimes used), minus the OSL measured from the last part of the shine-down curve (the late-light). An alternative approach is to deduce L_i by subtracting from the initial luminescence signal the so-called early light instead of the late-light, the early light usually being that which is measured during an interval immediately following that used for the initial measurement. The intention of the early-light subtraction method is to minimize the influence of the OSL measured from the malign slow and medium components of the shine-down curve that are found in quartz. This method has shown success in some cases (e.g., Ballarini et al., 2007; Cunningham and Wallinga, 2010; Pawley et al., 2010), but it should still be considered novel and it therefore should be applied with due vigilance. An alternate approach to isolating the fast component is to apply curve stripping algorithms (e.g., Bailey et al., 1997), but mathematically deconvoluting overlapping exponential (shine-down) curves can be difficult (Bailey, 2010). Another method that is intended to isolate the thermally stable, and fast, OSL component in quartz from malign components involves stimulating the sample while at an elevated temperature with infrared (IR) light instead of blue light. For this method to be effective the sample has to be relatively bright and be free of feldspar contamination as quartz responds poorly to IR light; a thorough review of this novel approach is provided by Bailey (2010). In general, during standard quartz OSL measurement (i.e., with blue-light

also to this mineral (e.g., Banerjee et al., 2001; Davids et al., 2010; Lamothe, 2004; Steffen et al., 2009a). Single-aliquot techniques, like SAR, are not immune to thermal transfer resulting from preheating, but its effect on equivalent dose can be monitored by repeating the experiments using increasing preheat temperatures (this practice is sometimes called a preheat-plateau test). Moreover, its effect can also be monitored by observing the OSL intensity measured from a zero-dose point on the regeneration curve (the so-called recuperation test). Unfortunately, if thermal transfer is significant in a sample, and cannot be made tolerable by reducing the preheat temperature or by adding (for quartz) a high-temperature blue-light stimulation step at the end of each cycle (Figure 4), then the sample cannot be dated using SAR.

The success of the test-dose procedure for correcting sensitivity change is assessed by repeating at least one of the data points on the regenerative-dose curve (the so-called recycling test), and a dose recovery test is required. For the latter, fresh aliquots (naturals) are bleached so that essentially all of the relevant electron traps are emptied. These aliquots are then all given a laboratory radiation dose that is similar to the sample's equivalent dose, and then an attempt is made to recover the laboratory

stimulation and detection in the ultraviolet), the sample aliquots are held at 125 °C to prevent the re trapping of electrons in the 110 °C TL trap; the aliquots are allowed to cool to <60 °C before the next step in the SAR cycle. The preheat treatment that is given after the test dose is administered (step 5) is commonly less severe than that given after a regenerative dose so as not to induce additional sensitivity change: what is usually done for this test-dose preheat is that the temperature of the aliquot is increased to 160 °C and then the heat is abruptly turned off (a so-called cut-heat treatment). Other steps may be included in the SAR routine as well, such as a short stimulation with IR light to check for feldspar contamination in a quartz sample, or a longer stimulation with IR light to attempt to 'wash' away the unwanted feldspar signal (Olley et al., 2004), or the addition of a high-temperature preheat (e.g., 280 °C), while the aliquot is stimulated with blue light at the end of each cycle to reduce the effect of thermal transfer (Murray and Wintle, 2003). Whether or not these added treatments are successful depends to a large degree on the characteristics of the sample. (b) Example of SAR shine-down curves, and the corresponding regenerative-dose graph (inset), for a quartz sand sample. Modified from Figure 7 of Lian OB and Roberts RG (2006) *Dating the Quaternary: Progress in luminescence dating of sediments. Quaternary Science Reviews* 25: 2449–2468. The quartz sample in this example is from Lake Mungo, Australia (sample M3T 58.0 m of Bowler et al. (2003)), and it yielded an optical age of 42.2 ± 2.5 ka. For the regenerative-dose graph, the units on vertical scale are arbitrary (dimensionless) because they are derived from the ratio L_i/T_i , where T_i is used to correct for sensitivity change in L_i . Of special importance in this example is that the duplicate dose point of 15 Gy on the regenerative-dose graph has a so-called recycling ratio (1.06 ± 0.03) that is consistent with unity, indicating that the test-dose procedure intended to correct for sensitivity change has been successful. Also of importance is that the intensity of the zero-dose measurement is only 1.6% of the intensity of the OSL measured from the natural aliquot indicating that thermal transfer as the result of the preheat (220 °C for 10 s in this case) is tolerable. The aliquot was stimulated for 100 s with blue (470 nm, 2.6 eV) light from an LED array, with the diodes set at constant intensity. The ultraviolet (350 nm, 3.6 eV) OSL signal, recorded by an Electron Tubes Ltd. 9235QA photomultiplier tube mounted behind a Hoya U-340 optical filter, used for construction of the regenerative-dose curve consisted of that measured during the first 3 s of stimulation, minus that recorded during the final 30 s.

dose using the same SAR protocol that was used to determine the equivalent dose of the natural sample. That a sample pass these tests (recuperation, recycling, and dose recovery) is a necessary, but unfortunately not a sufficient, condition for a calculated age value to be accepted (e.g., see Steffen et al., 2009a).

Optical Dating Using Linear Modulation

In the vast majority of cases, it is desired to sample the most light-sensitive electron traps – in other words, those that are responsible for the OSL recorded at the beginning of a shine-down curve (i.e., that from the so-called fast component of a shine-down curve, which was discussed earlier). In practice, therefore, one typically constructs a sample's dose response from the OSL recorded in the first second (or less) of continuous wave (CW) stimulation. What is meant by CW is that OSL is recorded while the stimulating light source is held on a sample at constant power. An alternative way of recording OSL is to do so while ramping the power of the stimulating light source at the sample from zero to some maximum value. This measurement technique is referred to as linear modulation (LM; Bulur, 1996). Using LM one can separate and record the various decay components that make up a shine-down curve (Figure 5). For example, the fast or most slowly bleaching components can be separated, and dose-response curves can be constructed from each. This is significant as there are samples for which a fast component is missing or is too dim to be used for dating, and there are other situations where the measurement of a slow component might be desired over a fast one.

Optical Dating Using Pulsed Stimulation

When a mineral sample is exposed to the stimulation beam there is a small delay before the resulting OSL is emitted; this is referred to as the luminescence lifetime. Because the luminescence lifetime is different for quartz and feldspar, it is possible to use pulsed stimulation to discriminate between the OSL arising from these two minerals in unseparated samples. For feldspar most of the OSL is emitted in a few μs after the stimulating pulse is switched off, whereas for quartz it persists for much longer ($>100\ \mu\text{s}$); see Figure 4 in Denby et al. (2006). In practice, the OSL that is used to construct the dose response is measured in the time between the pulses of stimulating light. Technical details and tests of the method can be found in Denby et al. (2006), Thomsen et al. (2008b), and Niedermayer and Bayer (2010).

Analysis and Presentation of Equivalent-Dose Distributions Arising from Single-Aliquot Techniques

As mentioned above, the main advantage that the single-aliquot techniques have over the multiple-aliquot methods is that an equivalent dose value can be determined from a very small aliquot of sample, even from just a single grain of sand, and this has led to the possibility of deducing the true age of a sedimentary deposit even if it consists of grains with different bleaching histories (i.e., different optical age values). In practice, current instrumentation and analytical software allows for hundreds of equivalent-dose values to be efficiently determined from less than a gram of prepared sediment. The challenge,

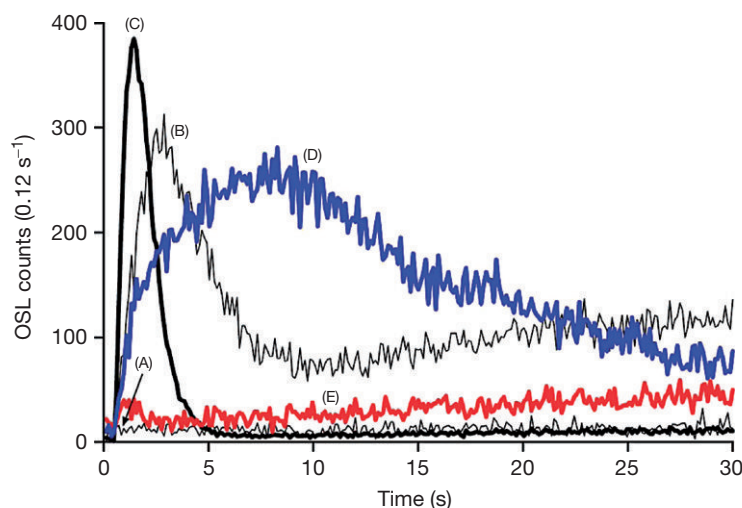


Figure 5 Example of linear modulation (LM) response curves for four individual sand-sized grains of quartz (naturals) extracted from the sample discussed in Figure 4(b). In each case, stimulation was from a focused green (532 nm, 2.3 eV) semiconductor laser, for which the intensity of the beam was steadily increased from zero to 90% of its maximum value (i.e., the laser power at the grain was ramped from 0 to 45 mW cm^{-2}) over a period of 30 s. Curve (A) consists of the background counts from the photomultiplier tube, while curve (B) is the LM-OSL from a grain that consists of more than one OSL component – in other words, the grain contains more than one population of electron trap, each population having a different sensitivity to light. The curve (C) is from a grain dominated by the so-called fast component, while curves (D) and (E) are LM-OSL data from grains dominated by the medium and slow components, respectively. Curve (C) has been multiplied by a factor 0.2 for the purpose of clarity of presentation. For this particular sample, the total LM-OSL output for a collection of grains over the first 5 s of stimulation is dominated by that from grains with an intense fast component, with 55% of the standardized estimate measured OSL coming from just 10% of the grains. The data are from Bowler et al. (2003). Modified from Figure 7 of Lian OB and Roberts RG (2006) Dating the Quaternary: Progress in luminescence dating of sediments. *Quaternary Science Reviews* 25: 2449–2468.

therefore, has been to analyze a distribution of equivalent-dose values and determine the desired equivalent dose and optical age from it. In reality, the spread in equivalent-dose values observed in a single sample is not only a result of its bleaching history, but also comes from differences in local dosimetry and variations in the intrinsic sensitivity of the grain(s) that make up an aliquot. Indeed, even samples that consist of grains that initially had all received extended sunlight exposure can show a spread in equivalent-dose values that is beyond that associated with analytical uncertainties. The cause of this so-called overdispersion is not fully understood.

Several methods have been devised to analyze equivalent-dose distributions, but the most useful arguably are those described by Galbraith *et al.* (1999). The models proposed by Galbraith are based on well-established statistical principles and are applicable to samples from a broad range of depositional environments: one of these statistical models is the central age model (CAM), which determines the equivalent dose from the weighted mean value, and takes into account overdispersion. Another is the minimum age model (MAM), which estimates the equivalent dose of a sample using the lowest equivalent-dose values; this approach has been validated in simulations on young fluvial sediments. The finite mixture model (FMM; Galbraith, 2005; Roberts *et al.*, 2000) can be applied in cases where a sample consists of grains belonging to two or more equivalent-dose populations. The FMM allows the number of equivalent-dose components to be estimated and ages can then be calculated from each one (e.g., Jacobs *et al.*, 2008). However, the FMM has been found to be applicable only in cases where each aliquot consists of a single grain of sand (Arnold and Roberts, 2009).

Statistical models should not be used blindly. Some prior information about the depositional environment that a sample came from is needed. For example, it is unlikely that the MAM should be used to date a loess deposit, or, for example, that the CAM should be applied to sediment extracted from poorly bleached alluvium. Clearly, information from a detailed sedimentological investigation is needed to help make a decision about what statistical model is most appropriate for the situation at hand. Indeed, Bailey and Arnold (2006) demonstrated through modeling that no one method of analysis is suited to all depositional environments, but that it is possible to select the most suitable method on the basis of the shape of the equivalent-dose distribution.

The method by which an equivalent-dose distribution is presented is also of importance. Basic histograms, as in Figure 6, were used initially but these do not show the variation in precision between aliquots that arise due to counting statistics – that is, equivalent-dose values calculated from highly luminescent (bright) aliquots being more precise than those determined from dim aliquots. The best graphical representation of equivalent-dose distributions probably is the radial plot (Galbraith, 2010; Galbraith *et al.*, 1999) in which the precision associated with each equivalent-dose value is clearly represented (Figure 6), and the degree of overdispersion can be evaluated. Weighted histograms and probability density plots may alleviate the problems associated with conventional histograms and probability density plots but they can be difficult to interpret. Berger (2010) has suggested that if probability

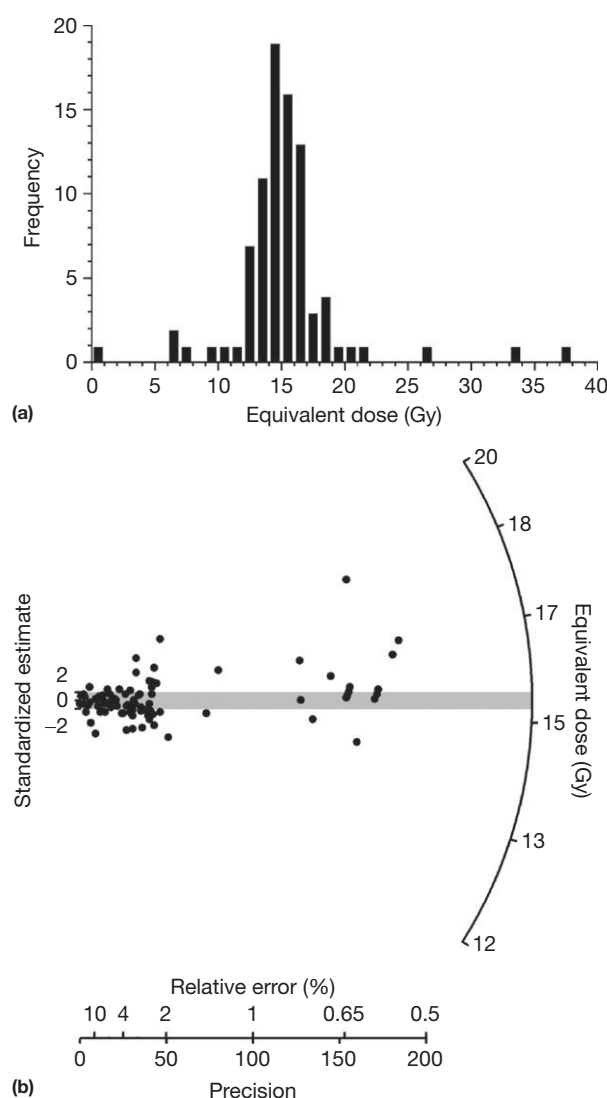


Figure 6 (a) Histogram showing the frequency distribution of the equivalent-dose values determined from 86 single quartz grains. The sample and the measurement conditions are the same as those discussed in Figures 4 and 5. (b) Radial plot of the same data shown in (a), where the horizontal shaded band represents the weighted-mean equivalent-dose value of all the data, which is 15.3 Gy. The histogram in (a) shows several equivalent-dose values >20 Gy, some which have poor precision, an aspect not represented in this kind of representation. The radial plot in (b), however, displays these data along with their precision, the more precise data plotting further away from the origin (center of the circle). For this sample, grains with relative errors of <1% (i.e., a precision value >100) have equivalent-dose values that are distributed over a range of 14–18 Gy. Also of note in this radial plot are the relatively high number of equivalent-dose values that fall outside the gray band. This gives a visual indication of the degree of overdispersion (see text) in this sample. Reproduced from Figure 8 of Lian OB and Roberts RG (2006) *Dating the Quaternary: Progress in luminescence dating of sediments. Quaternary Science Reviews* 25: 2449–2468.

density plots are desired, then the use of such plots with log-transformed equivalent-dose values and relative error estimates can further overcome the shortcomings of their conventional counterparts.

Sample Collection and Preparation, and Quartz or Feldspar?

Although sample collection for optical dating is relatively simple, it should be done by someone from the dating laboratory, together with a geoscientist familiar with the sedimentology (i.e., inferred depositional processes) of the sample site and the environmental history of the region, so that the best sampling strategy can be established. Sample collection and laboratory preparation are summarized elsewhere in this encyclopedia, while more detailed accounts can be found in Wintle (1997), Aitken (1985, 1998), and Lian and Huntley (2001); Lian et al. (1995) discuss cases where organic content is significant.

Whether quartz or feldspar is used for dating depends on the relative abundance of these minerals in a sample in the desired grain size, as well as the luminescence characteristics of the mineral which can be, to some degree, specific to the sample. The main advantages and disadvantages of using quartz and feldspar for optical dating are summarized in Figure 7 and are discussed in more detail by Wintle (2008).

Environments Suitable for Optical Dating

Any sedimentary deposit that contains an adequate amount of quartz and/or potassium feldspar grains in the desired size fraction, which have been exposed to sufficient sunlight or heat prior to final burial, has the potential for yielding an accurate optical age. As with TL dating, the most suitable environments are those in which the sediment has been transported over considerable distances in the atmosphere and/or have been reworked by other processes at the ground surface for an extensive amount of time before finally being buried and shielded from sunlight. Much of optical dating of sediments therefore has been done on eolian deposits such as loess and dune sediment. However, the greater sensitivity to sunlight of the electron traps responsible for OSL, compared to that of the traps responsible for TL, has resulted in the application of optical dating to a greater variety of depositional environments. Moreover, the development of single-aliquot techniques (especially SAR) has introduced the possibility of extracting the true burial age of a sediment from a deposit consisting of grains with different bleaching histories (i.e., with various age populations).

Examples that illustrate all of the applications of optical dating are too numerous and varied to be adequately discussed here, and the reader is directed to the published proceedings of the International Luminescence and Electron Spin Resonance Dating (LED) Conferences, where recent case studies and new developments are concisely presented. The two most current LED conference publications can be found in the journals *Quaternary Geochronology* (2(1–4), 2007; 5(2–3), 2010) and *Radiation Measurements* (41(7–8), 2006; 44(5–6), 2009). The journal *Ancient TL* (free online at: www.aber.ac.uk/ancient-tl/) publishes short articles on the experimental and theoretical aspects of optical dating; *Ancient TL* also includes papers related to TL and ESR dating. Illustrative examples can also be found in Prescott and Robertson (1997), Roberts (1997), Aitken (1998), Wagner (1998), Walker (2005), and Lian and

Roberts (2006) and in the special issue of the journal *Boreas*, 'Luminescence dating of Quaternary sediments' (35(4), 2008). Some examples that illustrate the diverse application, and the current age range, of optical dating are given below.

As mentioned above, one of the advantages that optical dating has over TL dating is that the electron traps responsible for the optical signal are much more sensitive to light, compared to those responsible for TL. This has allowed samples to be dated using OSL that are much younger than those assessable using TL. For example, Smith et al. (1990), using quartz, green laser light for stimulation, and the multiple-aliquot additive-dose method, were able to produce optical ages of 46 ± 30 , 95 ± 14 , and 350 ± 110 years, for eolian dune sand in Mali and France. More recently, Wolfe et al. (2002), using K-feldspar sand, infrared (IR) stimulation, and the multiple-aliquot ADTT technique, produced optical ages as young as a few decades with analytical uncertainties $<15\%$ (at two standard deviations) for eolian dune sands from the Canadian prairies. These optical ages agreed well with expected ages.

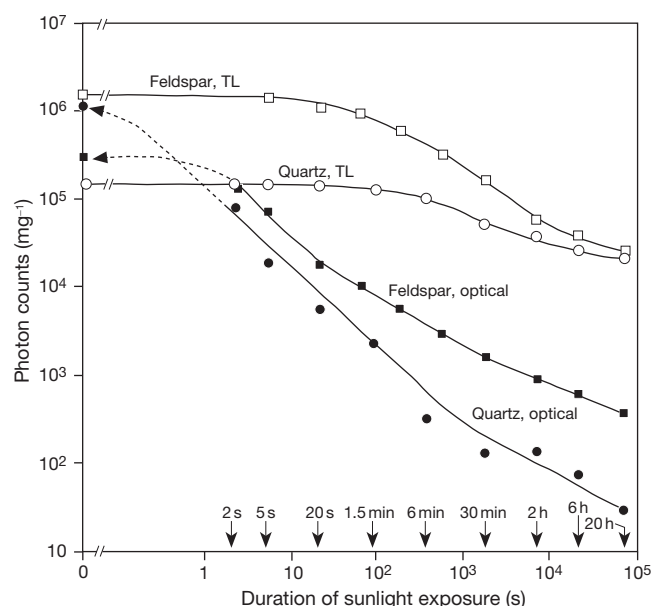
Huntley et al. (1993) demonstrated that IR stimulation could be used to preferentially stimulate K-feldspar inclusions in quartz grains. These workers used the multiple-aliquot ADTT method and quartz sand sampled from the stranded beach dune system in South Australia, and produced optical ages up to 400 ka that matched well with independent age information. Lian et al. (1995), on the other hand, used the multiple-aliquot ADTT protocol and IR stimulation to selectively date the K-feldspar grains in polymineral silt-sized samples extracted from peat deposits, and found good agreement between the optical ages and independent age information for samples ranging in age from 0 to ~ 120 ka once corrections for organic content was incorporated into the dose-rate calculations.

Examples of using OSL to date heated sediments include that of Lang and Wagner (1996), who successfully dated the fine-silt fraction of alluvium in southwestern Germany that was interpreted to have been heated by a campfire during the Roman period. The mineral of choice in this case was K-feldspar, which was stimulated by IR light. Equivalent doses, and optical ages, were determined using the multiple-aliquot additive-dose with late-light subtraction technique. Another example of dating fired sediment is provided by Lian and Brooks (2004) who sampled heat-altered alluvial mud from the base of small hearths in southern Manitoba (Canada) thought to be of anthropogenic origin. These workers also selected K-feldspar in the fine-silt polymineral fraction by selectively stimulating it with IR light. Optical ages were calculated from equivalent doses found using the multiple-aliquot ADTT method. These optical ages were consistent with accelerator mass spectrometry (AMS) radiocarbon ages of charcoal fragments found in the heated sediments, which dated the hearths to about 1 ka. A similar study was completed by Fu et al. (2010) who found consistency between quartz-based SAR ages of heated and light-exposed sediments (~ 5 ka) at an archaeological site at Qujialing, China.

One of the most novel applications of a single-aliquot protocol was provided by Roberts et al. (1997) who sampled quartz grains from relict mud wasp nests covering prehistoric paintings on cave walls in the Kimberly region of northern Australia. The premise was that wasps had previously collected

Quartz		Feldspar	
Advantages	Disadvantages	Advantages	Disadvantages
Highly resistant to weathering	Relatively low OSL intensity; some quartz samples do not emit measurable OSL; signal may consist of thermally unstable components	OSL saturates at higher radiation doses than does that from quartz	Weathers more readily from the environment than does quartz
OSL signal bleaches more rapidly in sunlight than that from feldspar	OSL saturates at lower radiation doses compared to that emitted from feldspar	OSL intensity may be orders of magnitude higher than that emitted from quartz	Suffers from anomalous fading
Does not appear to suffer from anomalous fading	Thermal transfer can be higher in quartz than in feldspar	IRSL can be stimulated preferentially in unseparated mineral mixtures (polyminerall samples)	Difficult or impossible to correct for sensitivity change in regenerative-dose data when using SAR

(a)



(b)

Figure 7 (a) Summary of the principal advantages and disadvantages of using quartz or K-feldspar for optical dating. The fact that the IRSL emitted from K-feldspar typically saturates at higher radiation doses than that emitted from quartz, and because the luminescence emitted from K-feldspar is usually much more intense than that emitted from quartz, and that the K-feldspar IRSL signal does not appear to include malign components as does that of some quartz samples, makes feldspar ideal for dating both very young and very old samples. Feldspar, however, suffers from anomalous fading where quartz appears not to. It is mainly for that reason that quartz is used more widely than feldspar when both minerals are present at a sample site. Anomalous fading was first observed in feldspar by Wintle (1973) and refers to the loss of electrons from traps that are expected to be thermally stable and ambient temperatures over geological time, to other defects and centers in the crystal lattice; it is thought that quantum-mechanical tunneling is the mechanism responsible for this phenomenon. Some studies have suggested that alkali feldspars (K-feldspars) exhibit less anomalous fading than does most plagioclase feldspars (but see below), and this is one of the reasons that the former feldspar mineral has been preferred over the latter for optical dating. Experiments using the dominate IR-stimulated violet emission (centered at 400 nm, 3.1 eV) from a variety of sand-sized K-feldspar natural sediment samples, and from small slices taken from whole K-feldspar crystals, collected from various parts of the world (80 samples in total), have strongly suggested that anomalous fading is ubiquitous in this mineral (Huntley and Lamothe, 2001; Huntley and Lian, 2006). It has also been observed that plagioclase feldspar with low Ca content show no tunneling (fading), and that such minerals may be used to extend the current upper age limit of optical dating (Huntley et al., 2007). For sufficiently bright feldspar samples, anomalous fading can be measured in the laboratory with the necessary precision, and if its magnitude is significant, then either it must be corrected for or the sample cannot be dated. Huntley and Lamothe (2001) devised a way to correct for anomalous fading by using a sample's measured fading rate, but their technique only can be applied to a sample that is in the linear part of its dose response. It has been observed, however, that the 400 nm IR-stimulated luminescence from fine-silt sized K-feldspar can often yield optical ages that are in good agreement with independent age information without correction for anomalous fading. This suggests that, for some samples, grain size, or perhaps the mineral's resistance to weathering, might be a factor affecting the degree of anomalous fading in a sample. Other studies have indicated that using various preheat protocols, or by sharply blocking the luminescence emitted at wavelengths below 400 nm, will reduce anomalous fading to a point where it can be tolerated (e.g., see discussion in Berger et al., 2004), and it has also been shown that the 400 nm emission from K-feldspar measured while the sample is held at a temperature of 220 °C shows reduced anomalous fading, if this measurement is preceded by an infrared wash at low temperature (~50 °C; Buylaert et al., 2009, 2012; Thomsen et al., 2008c).

(b) Comparison of the effectiveness of direct sunlight in emptying the traps in quartz and feldspar responsible for OSL and TL (reproduced from Godfrey-Smith DJ, Huntley DJ, and Chen W-H (1988) Optical dating studies of quartz and feldspar sediment extracts. *Quaternary Science Reviews* 7: 373–380). The TL data are from the 310–320 °C range, while the OSL data are that from the first 20 s of stimulation, using green (514 nm, 2.4 eV) photons from an argon ion laser. Note that the OSL signal from quartz bleaches faster than that from feldspar, and much faster than the TL signals from both minerals. The quartz was from a 500-ka eolian dune sand, while the feldspar was from a 70-ka fluvial sand.

mud (containing quartz sand grains) from well-bleached surfaces distant to the cave mouth, and that the luminescence clock had started after the grains had been deposited on the cave wall and covered by a subsequent nest layer(s). These workers, using a SAAD method, showed that at least some of the cave paintings must have been constructed before 17 ka. A somewhat similar study was undertaken by Baran et al. (2003), but in this case the features of interest were prehistoric stone structures in southern Sweden. Baran et al. sampled sediment that had been used to fill the spaces between the stones, extracted quartz sand-sized grains from it, and applied the SAR protocol (Murray and Wintle, 2000) to small aliquots and single quartz grains, and deduced that construction of the stone structures had probably commenced as early as 7 ka.

The advent of single-grain dating has opened up the possibility of analyzing the character of the dose responses associated with individual grains. Yoshida et al. (2000) looked at the dose responses of single grains of quartz from samples ranging in age from ~250 to ~950 ka collected from the well-studied stranded beach dune system in South Australia. They found that all but one of the nine samples studied contained a small number of grains (~15%) that were strongly luminescent, and whose luminescence signal saturated at unusually high laboratory radiation doses. Some of these grains yielded ages approaching 1 Ma that were consistent with the independent chronology of the dunes. A similar approach has been used by Rhodes et al. (2006) to date marine sands and eolian deposits, some approaching 1 Ma, associated with archaeological material in northwestern Africa. A relatively new approach that has the potential to extend the upper age limit of optical dating even further was introduced by Rink et al. (2007). This method involves measuring the light-sensitive defects associated with aluminum and titanium atoms on silicon sites in quartz using ESR. This new method is referred to as ESR optical dating, and it is thought to have the potential to date samples that were last exposed to an extended period of sunlight 5 Ma ago or more (Rink et al., 2007).

A good example of the use of the LM technique is that of Olley et al. (2006) who used this technique to show that grains incased in resin-impregnated sample blocks, that held grave-infill sediments associated with ancient human fossils, had been sufficiently protected from room light exposure while being archived in a laboratory. First, these workers applied the SAR protocol to single quartz grains, and used the fast component of the continuous-wave OSL shine-down curve, together with the maximum age model, to produce equivalent-dose values (and optical ages) which were consistent with the expected age for sediments. They then applied the LM technique to grains from the same sample and found that both the fast and slow components produced equivalent-dose values that were consistent with each other. This confirmed that the grains had remained sheltered from light during storage. A similar approach had previously been used by Yoshida to date archaeological samples, and by Olley to investigate deep-sea sediments (see Lian and Roberts, 2006).

Factors that can Lead to an Inaccurate Optical Age

The most significant factor that can lead to an inaccurate optical age is that the mineral grains comprising a sample

collected from a sedimentary deposit have not all been exposed to sufficient sunlight, or heat, prior to burial. If this is the case, and if multiple-aliquot techniques are used, the calculated optical age will either represent the true age of the deposit, or it will lie somewhere between the true age and the apparent age of its parent material(s), although some information about equivalent-dose (age) populations can be gained from studying the scatter in the normalized dose-response data (e.g., Ollerhead et al., 2001). If single-aliquot methods are used, it is much easier to identify equivalent-dose distributions, and in many cases the equivalent dose representing the event of interest (e.g., the time elapsed since burial) can be extracted from it.

An inaccurate or unacceptably imprecise optical age can also result (1) if an inappropriate laboratory protocol is chosen for determination of the equivalent dose, (2) if an inappropriate curve is fitted to the dose-response data, (3) if there is, or has been, significant radioactive disequilibrium at the sample site, and (4) if single grains are dated and variations in microdosimetry are significant.

More complex phenomena that can cause problems are (1) thermal transfer that cannot be corrected for, (2) sensitivity change that cannot be detected (as can occur when using multiple-aliquot techniques) or corrected for (when using single-aliquot methods), (3) the failure of a dose recovery test in single-aliquot dating, (4) overdispersion in equivalent-dose values found using single-aliquot techniques that is not recognized or is not accounted for, and (5) if OSL from a part of the shine-down curve that represents electron traps that are difficult to bleach, or are thermally unstable (the so-called medium and slow components), is used to construct the dose response (e.g., Steffen et al., 2009b).

If feldspars are dated, the severity of anomalous fading has to be assessed, and if it is significant, it has to be avoided or corrected for, as discussed in Figure 7. In light of all these potential pitfalls, it is highly advisable, when using any of the optical dating protocols, no matter how well established, to date at least one sample from the sedimentary sequence or landform of interest for which the age has previously been determined by an independent, and reliable, method.

See also: **Luminescence Dating: Electron Spin Resonance Dating; Thermoluminescence.**

References

- Aitken MJ (1985) *Thermoluminescence Dating*. London: Academic Press.
- Aitken MJ (1998) *An Introduction to Optical Dating*. Oxford: Oxford University Press.
- Arnold LJ and Roberts RG (2009) Stochastic modelling of multi-grain equivalent dose (D_e) distributions: Implications for OSL dating of sediment mixtures. *Quaternary Geochronology* 4: 204–230.
- Bailey RM (2000) The interpretation of quartz optically stimulated luminescence equivalent dose versus time plots. *Radiation Measurements* 32: 129–140.
- Bailey RM (2001) Towards a general kinetic model for optically and thermally stimulated luminescence of quartz. *Radiation Measurements* 33: 17–45.
- Bailey RM (2010) Direct measurement of the fast component of quartz optically stimulated luminescence and implications for the accuracy in optical dating. *Quaternary Geochronology* 5: 559–568.

- Bailey RM and Arnold LJ (2006) Statistical modelling of single-grain quartz D_e distributions and an assessment of procedures for estimating burial dose. *Quaternary Science Reviews* 25: 2475–2502.
- Bailey RM, Smith BW, and Rhodes EJ (1997) Partial bleaching and the decay form characteristics of quartz OSL. *Radiation Measurements* 27: 123–136.
- Ballarín M, Wallinga J, Wintle AG, and Bos AJJ (2007) A modified SAR protocol for optical dating of individual grains from young quartz samples. *Radiation Measurements* 42: 360–369.
- Banerjee D, Murray AS, Bøtter-Jensen L, and Lang A (2001) Equivalent dose estimation using a single aliquot of polycrystalline feldspar. *Radiation Measurements* 33: 73–94.
- Baran J, Murray AS, and Haggstrom L (2003) Estimating the age of stone structures using OSL: The potential of entrapped sediment. *Quaternary Science Reviews* 22: 1265–1271.
- Berger GW (2010) An alternative form of probability-distribution plot for D_e values. *Ancient TL* 28: 11–21.
- Berger GW, Melles M, Banerjee D, Murray AA, and Raab A (2004) Luminescence chronology of non-glacial sediments in Changeable Lake, Russian High Arctic, and implications for limited Eurasian ice-sheet extent during the LGM. *Journal of Quaternary Science* 19: 513–523.
- Blair MW, Yukihiro EG, and McKeever SWS (2005) Experiences with single-aliquot OSL procedures using coarse-grained feldspar. *Radiation Measurements* 39: 361–374.
- Bortolot VJ and Bluszcz A (2003) Strategies for flexibility in luminescence dating: Procedure-oriented measurement and hardware modularity. *Radiation Measurements* 37: 551–555.
- Bøtter-Jensen L, McKeever SWS, and Wintle AG (2003) *Optically Stimulated Luminescence Dosimetry*. Amsterdam: Elsevier.
- Bowler JM, Johnston H, Olley JM, et al. (2003) New ages for human occupation and climatic change at Lake Mungo, Australia. *Nature* 421: 837–840.
- Bulur E (1996) An alternative technique for optically stimulated luminescence (OSL) experiment. *Radiation Measurements* 26: 701–709.
- Buylaert J-P, Jain M, Murray AS, Thomsen KJ, Thiel C, and Sohbati R (2012) A robust feldspar luminescence dating method for Middle and Late Pleistocene sediments. *Boreas*, in press. <http://dx.doi.org/10.1111/j.1502-3885.2012.00248.x>.
- Buylaert JP, Murray AS, Thomsen KJ, and Jain M (2009) Testing the potential of an elevated temperature IRSL signal from K-feldspar. *Radiation Measurements* 44: 560–565.
- Cunningham AC and Wallinga J (2010) Selection of integration time intervals for quartz OSL decay curves. *Quaternary Geochronology* 5: 657–666.
- Davids F, Duller GAT, and Roberts RG (2010) Testing the use of feldspars for optical dating of hurricane overwash deposits. *Quaternary Geochronology* 5: 125–130.
- Denby PM, Bøtter-Jensen L, Murray AS, Thomsen KJ, and Moska P (2006) Application of pulsed OSL to the separation of the luminescence components from a mixed quartz/feldspar sample. *Radiation Measurements* 41: 774–779.
- Duller GAT (1991) Equivalent dose determination using single aliquots. *Nuclear Tracks and Radiation Measurements* 18: 371–378.
- Duller GAT (2004) Luminescence dating of Quaternary sediments: Recent advances. *Journal of Quaternary Science* 19: 183–192.
- Duller GAT (2008) Single-grain optical dating of Quaternary sediments: Why aliquot size matters in luminescence dating. *Boreas* 37: 589–612.
- Fu X, Zhang J-F, Mo D-W, Shi C-X, Lium H, and Zhou L-P (2010) Luminescence dating of baked earth and sediments from Qujialing archaeological site, China. *Quaternary Geochronology* 5: 353–359.
- Galbraith RF (2005) *Statistics for Fission Track Analysis*. London: Chapman & Hall.
- Galbraith RF (2010) On plotting OSL equivalent doses. *Ancient TL* 28: 1–9.
- Galbraith RF, Roberts RG, Laslett GM, Yoshida H, and Olley JM (1999) Optical dating of single and multiple grains of quartz from Jinmium rock shelter, northern Australia. Part I: Experimental design and statistical models. *Archaeometry* 41: 339–364.
- Huntley DJ, Baril MR, and Haidar S (2007) Tunnelling in plagioclase feldspar. *Journal of Physics D: Applied Physics* 40: 900–906.
- Huntley DJ and Berger GW (1995) Scatter in luminescence data for optical dating – Some models. *Ancient TL* 13: 5–9.
- Huntley DJ, Godfrey-Smith DI, and Thewalt MLW (1985) Optical dating of sediments. *Nature* 313: 105–107.
- Huntley DJ, Hutton JT, and Prescott JR (1993) Optical dating using inclusions within quartz grains. *Geology* 21: 1087–1090.
- Huntley DJ and Lamothe M (2001) Ubiquity of anomalous fading in K-feldspars, and the measurement and correction for it in optical dating. *Canadian Journal of Earth Sciences* 38: 1093–1106.
- Huntley DJ and Lian OB (2006) Some observations on the tunnelling of trapped electrons in feldspars and their implications for optical dating. *Quaternary Science Reviews* 25: 2503–2512.
- Jacobs Z and Roberts RG (2007) Advances in optically stimulated luminescence dating of individual grains of quartz from archaeological deposits. *Evolutionary Anthropology* 16: 210–223.
- Jacobs Z, Wintle AG, Duller GAT, Roberts RG, and Wadley L (2008) New ages for the post-Howiesons Poort, late and final Middle Stone Age at Sibudu, South Africa. *Journal of Archaeological Science* 35: 1790–1907.
- Jain M, Murray AS, and Bøtter-Jensen L (2003) Characterisation of blue light stimulated luminescence components in different quartz samples: Implications for dose measurement. *Radiation Measurements* 37: 441–449.
- Krbetschek MR, Trautmann T, Dietrich A, and Stolz W (2000) Radioluminescence dating of sediments: Methodological aspects. *Radiation Measurements* 32: 493–498.
- Lamothe M (2004) Optical dating of pottery, burnt stones, and sediments from selected Quebec archaeological sites. *Canadian Journal of Earth Sciences* 41: 659–667.
- Lang A and Wagner GA (1996) Infrared stimulated luminescence dating of archaeosediments. *Archaeometry* 28: 129–141.
- Lian OB and Brooks GR (2004) Optical dating studies of mud-dominated alluvium and buried hearth-like features from Red River Valley, southern Manitoba, Canada. *The Holocene* 14: 570–578.
- Lian OB, Hu J, Huntley DJ, and Hicock SR (1995) Optical dating studies of Quaternary organic-rich sediments from southwestern British Columbia and northwestern Washington State. *Canadian Journal of Earth Sciences* 32: 1194–1207.
- Lian OB and Huntley DJ (1999) Optical dating studies of post-glacial aeolian deposits from the south-central interior of British Columbia, Canada. *Quaternary Science Reviews* 18: 1453–1466.
- Lian OB and Huntley DJ (2001) Luminescence dating. In: Last WM and Smol JP (eds.) *Tracking Environmental Change Using Lake Sediments, Vol. 1: Basin Analysis, Coring, and Chronological Techniques*, pp. 261–282. Dordrecht: Kluwer.
- Lian OB and Roberts RG (2006) Dating the Quaternary: Progress in luminescence dating of sediments. *Quaternary Science Reviews* 25: 2449–2468.
- McKeever SWS and Chen R (1997) Luminescence models. *Radiation Measurements* 27: 625–661.
- McKeever SWS and Yukihiro EG (2011) *Optically Stimulated Luminescence: Fundamentals and Applications*. Chichester: Wiley.
- Murray AS and Wintle AG (2000) Luminescence dating of quartz using an improved regenerative-dose protocol. *Radiation Measurements* 32: 57–73.
- Murray AS and Wintle AG (2003) The single aliquot regenerative dose protocol: Potential for improvements in reliability. *Radiation Measurements* 37: 377–381.
- Niedermayer M and Bayer A (2010) Pulsed photon-stimulated luminescence (PPSL) as a tool in retrospective dosimetry. *Radiation Protection Dosimetry* 139: 494–504.
- Ollerhead J, Huntley DJ, Nelson AR, and Kelsey HM (2001) Optical dating of tsunami-laid sand from an Oregon coastal lake. *Quaternary Science Reviews* 20: 1915–1926.
- Olley JM, Pietsch T, and Roberts RG (2004) Optical dating of Holocene sediments from a variety of geomorphic settings using single grains of quartz. *Geomorphology* 60: 337–358.
- Olley JM, Roberts RG, Yoshida H, and Bowler JM (2006) Single-grain optical dating of grave-infill associated with human burials at Lake Mungo, Australia. *Quaternary Science Reviews* 25: 2469–2474.
- Pawley SM, Toms P, Armitage SJ, and Rose J (2010) Quartz luminescence dating of Anglian Stage (MIS 12) fluvial sediments: Comparison of SAR age estimates to the terrace chronology of the Middle Thames valley, UK. *Quaternary Geochronology* 5: 569–582.
- Prescott JR, Huntley DJ, and Hutton JT (1993) Estimation of equivalent dose in thermoluminescence dating – The Australian slide method. *Ancient TL* 11: 1–5.
- Prescott JR and Hutton JT (1994) Cosmic ray contributions to dose rates for luminescence and ESR dating: Large depths and long-term time variations. *Radiation Measurements* 23: 497–500.
- Prescott JR and Robertson GB (1997) Sediment dating by luminescence: A review. *Radiation Measurements* 27: 893–922.
- Preusser F, Chithambo M, Götze T, et al. (2009) Quartz as a natural dosimeter. *Earth-Science Reviews* 97: 184–214.
- Rhodes EJ, Singarayer JS, Raynal J-P, Westaway KE, and Sbihi-Alaoui FZ (2006) New age estimates for the Palaeolithic assemblages and Pleistocene succession of Casablanca, Morocco. *Quaternary Science Reviews* 25: 2569–2585.
- Rink WJ, Bartoll J, Schwarcz HP, Shane P, and Bar-Yosef O (2007) Testing the reliability of ESR dating of optically exposed buried quartz sediments. *Radiation Measurements* 42: 1618–1626.
- Roberts RG (1997) Luminescence dating in archaeology: From origins to optical. *Radiation Measurements* 27: 819–892.
- Roberts RG, Galbraith RF, Yoshida H, Laslett GM, and Olley JM (2000) Distinguishing dose populations in sediment mixtures: A test of single-grain optical dating

- procedures using mixtures of laboratory-dosed quartz. *Radiation Measurements* 32: 459–465.
- Roberts R, Walsh G, Murray A, et al. (1997) Luminescence dating of rock art and past environments using mud-wasp nests in northern Australia. *Nature* 387: 696–699.
- Roberts R, Yoshida H, Galbraith R, Laslett G, Jones R, and Smith M (1998) Single-aliquot and single-grain optical dating confirm thermoluminescence age estimates at Malakunanja II rock shelter in northern Australia. *Ancient TL* 16: 19–24.
- Shane PA, Lian OB, Augustinus P, Chisari R, and Heijnis H (2002) Tephrostratigraphy and geochronology of a 100 kyr terrestrial record at Lake Poukawa, North Island, New Zealand. *Global and Planetary Change* 33: 221–242.
- Smith BW and Rhodes EJ (1994) Charge movements in quartz and their relevance to optical dating. *Radiation Measurements* 23: 329–333.
- Smith BW, Rhodes EJ, Stokes S, Spooner NA, and Aitken MJ (1990) Optical dating of sediments: Initial quartz results from Oxford. *Archaeometry* 32: 19–31.
- Steffen D, Preusser F, and Schlunegger F (2009a) OSL quartz age underestimation due to unstable signal components. *Quaternary Geochronology* 4: 353–362.
- Steffen D, Schlunegger F, and Preusser F (2009b) Drainage basin response to climate change in Pisco valley, Peru. *Geology* 37: 491–494.
- Thomsen KJ, Bøtter-Jensen L, Jain M, Denby PM, and Murray AS (2008a) Recent instrumental developments for trapped electron dosimetry. *Radiation Measurements* 43: 414–421.
- Thomsen KJ, Jain M, Murray AS, Denby PM, Roy N, and Bøtter-Jensen L (2008b) Minimizing feldspar OSL contamination in quartz UV-OSL using pulsed blue stimulation. *Radiation Measurements* 43: 752–757.
- Thomsen KJ, Murray AS, Jain M, and Bøtter-Jensen L (2008c) Laboratory fading rates of various luminescence signals from feldspar-rich sediment extracts. *Radiation Measurements* 43: 1474–1486.
- Wagner GA (1998) *Age Determination of Young Rocks and Artifacts*, p. 466. New York: Springer.
- Walker MJC (2005) *Quaternary Dating Methods*, p. 286. New York: Wiley.
- Wintle AG (1973) Anomalous fading of thermoluminescence in mineral samples. *Nature* 245: 143–144.
- Wintle AG (1997) Luminescence dating: Laboratory procedures and protocols. *Radiation Measurements* 27: 769–817.
- Wintle AG (2008) Luminescence dating: Where it has been and where it is going. *Boreas* 37: 471–482.
- Wintle AG and Murray AS (2006) A review of quartz optically stimulated luminescence characteristics and their relevance in single-aliquot regeneration dating protocols. *Radiation Measurements* 41: 369–391.
- Wolfe SA, Huntley DJ, and Ollerhead J (2002) Optical dating of modern and Late Holocene dune sands in the Brandon Sand Hills, southwestern Manitoba. *Géographie Physique et Quaternaire* 56: 203–214.
- Yoshida H, Roberts RG, Olley JM, Laslett GM, and Galbraith RF (2000) Extending the age range of optical dating using single 'supergrains' of quartz. *Radiation Measurements* 32: 439–446.

Relevant Website

www.aber.ac.uk – The Official Website of the University of Wales, Aberystwyth.

Electron Spin Resonance Dating

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Introduction

Electron spin resonance (ESR), also called electron paramagnetic resonance, is a dating technique based on the accumulation of free electrons or holes (electron defects) in the crystal lattices of minerals. The process is due to exposure to radiation emitted from radioactive nuclides in the sample and its environment. Hence, it is a 'radiation-dose' or 'trapped-charge' method, similar to thermoluminescence (TL) and optically stimulated luminescence (OSL). These methods differ in the instrumentation used and the physical processes involved in the accumulated radiation dose.

In ESR, the mineral accumulates radiation damage and acts as a kind of dosimeter of its total exposure to radiation from the surrounding environment, and the sample itself. An ESR age is derived from the measurements of the dose that the sample has received over time. The most common applications of this method are to tooth enamel, corals, mollusk shells, and quartz.

Basic Principles

There are two energy levels at which electrons may exist: the ground state (valence band) and a higher energy state (conduction band). These are shown schematically in Figure 1. In insulators, such as quartz, feldspar, apatite, and calcite, the two energy bands are separated by energy levels that are forbidden transitions from the ground state. Naturally occurring radionuclides emit a variety of radiation, for example, α -particles (He nuclei), β -particles (electrons), and γ -radiation (photons). When these particles interact with matter, electrons in the atoms can be excited from the valence band to the conduction band, and electron defects ('holes') accumulate in the valence band. In general, these holes are rapidly filled by other electrons. However, electrons (and holes) can also become trapped in intermediate energy levels where they cannot decay back to the ground state. All minerals contain such defect sites where electrons and holes can become trapped (Figure 1). When the sample is initially crystallized, it should not contain any defects, and the ESR signal should be zero. After formation, the mineral is exposed to radiation, leading to the accumulation of the electronic defects. These trapped electrons and holes form paramagnetic centers, since there are unpaired electrons present. Paramagnetic centers show weak magnetic response in an applied magnetic field, which can be detected with an ESR spectrometer, giving rise to a characteristic resonance (Figure 2).

In addition to the accumulation of electronic defects, ESR signals can arise from the breaking of covalent bonds to form free radicals, such as CO_2^- (an ionized free radical) from

carbonate ions. This is the most common process for trapped-charge accumulation in secondary carbonates such as speleothems, shells, and corals, as well as tooth enamel. Grün et al. (2008) noted that CO_2^- can occur in two forms with different symmetries, leading to a small difference in the ESR signal and additional errors of <5%. Later, Joannes-Boyau and Grün (2011) showed that these differences could account for the underestimation of ages by up to 30%.

The measured ESR signal is called the 'natural intensity' and is dependent on the number of traps or radiation dose (D), the dose rate of the radioactivity (dD/dt), and the radiation time T , which should be the age of the sample. The product of dose rate and time is the equivalent dose D_e that the sample has received in the past. Hence,

$$D_e = \int_0^T D(t) dt \quad [1]$$

If dD/dt is a constant, this equation integrates to

$$T = D_e / (dD/dt) \quad [2]$$

The determination of the D_e value is the ESR part of the procedure. It is necessary to separately evaluate the radiation dose from analysis of radioactive elements (mainly Th, U, and K) from the sample and its surroundings. We will discuss this later.

Measurement of the Dose Value

D_e is the equivalent dose and is derived from the fact that laboratory sources use γ -rays whereas the accumulated dose is the sum of different energies of α -, β -, and γ -radiation. Thus, the experimentally determined dose value is the γ equivalent of the naturally received dose, which is frequently called the paleodose P . The SI unit for absorbed radiation dose is the gray (Gy), which is equivalent to the deposition of one joule of energy (J) in 1 kg of matter.

In order to produce reliable results, the measured ESR signals must have the following properties:

- The signal intensity increases in proportion to the received dose.
- The signals are thermally stable, to at least an order of magnitude longer than the age of the sample.
- The number of traps is consistent in the material. Recrystallization, crystal growth, or phase transformations should not have occurred.
- The signal does not show anomalous fading.
- The signals are not influenced by sample preparation, such as grinding, exposure to laboratory light, etc.

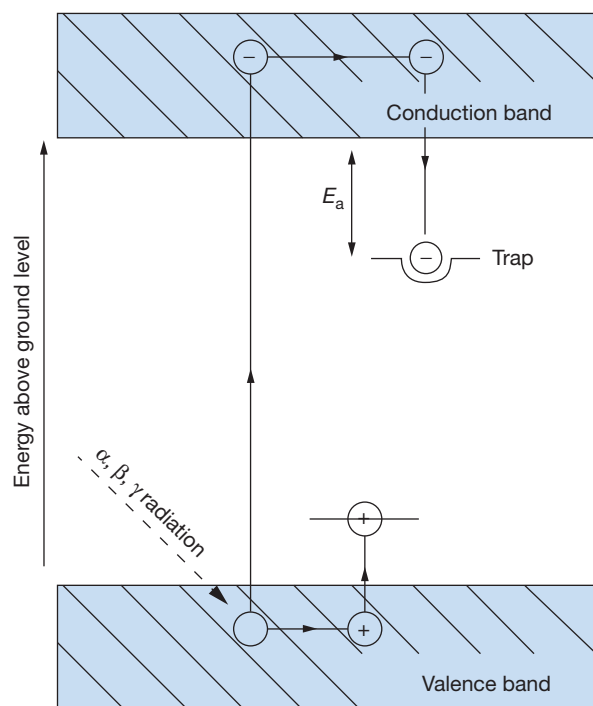


Figure 1 Trapping of electrons and holes, which is the basis for electron spin resonance (ESR) dating. An insulating mineral has two energy levels at which electrons may exist. The lower energy level (valence band) is separated from the higher energy level (conduction band) by a so-called forbidden zone. When a mineral is formed or reset, all electrons are in the ground state. Ionizing radiation emitted from radioactive elements (U, Th, and K) removes negatively charged electrons from atoms. The electrons are transferred to the conduction band, and positively charged holes are left behind near the valence band. After a short time of diffusion, most of the electrons recombine with the holes. Some electrons can be trapped by impurities (electron traps) in the crystal lattice. These unpaired (free) electrons can be directly measured by ESR spectroscopy. E_a is the activation energy or trap depth.

Two processes limit the applicability of ESR for dating:

Saturation: When all the available traps are filled, any further irradiation does not result in further formation of paramagnetic centers. In high-dose-rate environments, saturation may occur within a short period (a few thousand years), whereas in low-dose environments, this may occur in a few million years.

Thermal stability: Trapped electrons have only a limited probability of remaining in the traps. After a mean lifetime (τ), ~64% of the electrons will have left the trapped and returned to the ground state. The effects of 'fading' (loss of the ESR signal) can be neglected only if the lifetime τ is about ten times longer than the sample age. The thermal stability of different paramagnetic centers may be different.

It is not possible to give a general upper limit for ESR dating because of the dependence of both saturation and thermal stability on the specific nature of the paramagnetic centers and the environment of the samples. Each case needs to be evaluated individually. On the younger end of the dating scale, ESR measurements on tooth enamel have been used in evaluating small doses that would naturally accumulate in a few hundred years. Similarly, good ESR results have been obtained

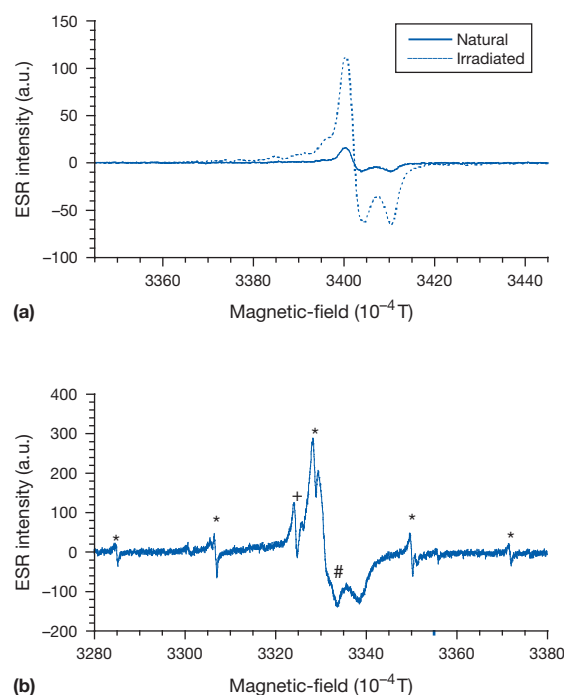


Figure 2 ESR spectra of tooth enamel. The spectrum is dominated by a CO_2^- radical (a). Some spectra show interferences by methyl (* in (b)), SO_2 (+), and a rotating CO_2^- (#).

on shell and coral samples of a few hundred years' age. At the other end of the timescale, EST analysis of tooth enamel has provided reasonable results over 2 million years (Curnoe et al., 2001; Schwarcz et al., 1994). Corals and shells produce systematic underestimates older than about 300 000 years as a result of the limited thermal stability of this ESR signal.

For measurement of the D_e value, the sample is irradiated in the laboratory with increasing doses from a calibrated γ -ray source. This irradiation produces more paramagnetic centers and a consequent increased ESR signal. A dose-response curve can be established by plotting the ESR signal intensity against the applied dose (Figure 3). The data produced can be extrapolated to zero intensity, and the D_e value results from the intercept of the plot with the x -axis. In contrast to luminescence studies, it is not usually possible to regenerate the ESR signal in the way that can be often done in TL and OSL, because heating often changes the ESR characteristics of the minerals. The dose-response curve is not necessarily linear because of the saturated effects of different paramagnetic centers.

ESR Measurement

ESR intensities can be measured with off-the-shelf instrumentation (manufactured by Bruker, JEOL, and Varian, for example). An ESR spectrometer has three basic components (Figure 4(a)): a strong electromagnet, a microwave generator (either a Gunn diode or klystron), and an electronic data-processing unit. The sample is located in a cavity between the poles of the magnet. The cavity is connected to the microwave generator by waveguides. The paramagnetic centers have

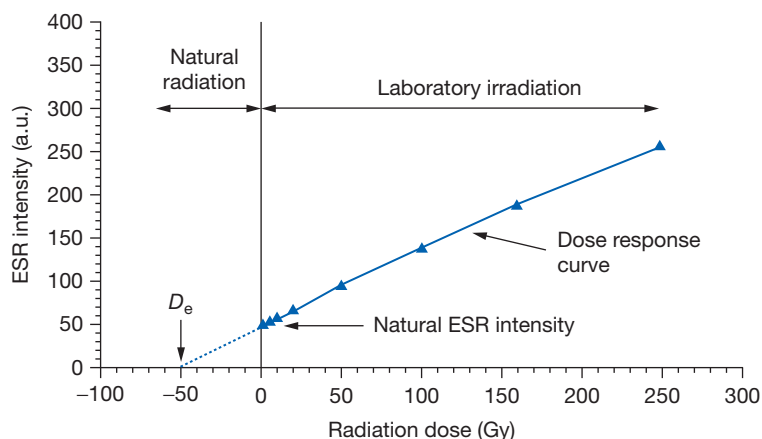


Figure 3 Dose (D_e) determination using the additive dose method. After measurement of the natural intensity, the sample is irradiated in the laboratory. This leads to the production of more paramagnetic centers (see [Figure 1](#)). Consequently, the signal intensity increases. The plot of ESR intensity versus laboratory dose is called the dose–response curve. The D_e value results from fitting the data points of the dose–response with an exponential function and extrapolation to zero ESR intensity.

permanent magnetic moments (μ_s) due to the spin of the unpaired electrons. In an external magnetic field, this magnetic moment has two energy states: the ground state, when the spin is oriented toward the magnetic field direction; and an excited state, when the spin is in the opposite direction ([Figure 4\(b\)](#)). The difference between these two energy levels, ΔE , can be expressed as

$$\Delta E = g\mu_B H \quad [3]$$

where g is the Landé factor, μ_B is the Bohr magneton ($9.274078 \times 10^{-24} \text{ J T}^{-1}$), and H is the magnetic field strength in tesla (T). A transfer from the lower to the higher state can occur at the resonant frequency (ν), where $\nu = \omega/2\pi$ and ω is the precession frequency of the magnetic moment, μ_s . Therefore, the resonant condition is

$$\nu h = g\mu_B H \quad [4]$$

where h is the Planck constant ($6.626176 \times 10^{-34} \text{ J s}^{-1}$). Hence, applying the resonant signal frequency ν will result in ‘flipping’ the electrons from the lower to the higher state. The amount of microwave energy absorbed is directly proportional to the number of paramagnetic centers, and therefore, in principle, the age of the sample.

After microwave absorption, the two electron populations should be equalized, and no further absorption observed. As can be seen from [Figure 4](#), the resonance condition can be observed either by varying the microwave frequency or the magnetic field intensity. However, because the physical dimensions of the waveguide and the microwave cavity are critically dependent on the microwave frequency, the magnetic field is varied linearly for the ESR measurement. To optimize the signal-to-noise ratio, ESR signals are recorded as derivatives using field modulation ([Figures 2](#) and [4\(a\)](#)). Note that ESR measurements do not require absolute determination of the number of paramagnetic centers; only the relative concentrations (through the ESR response) are needed to establish a dose–response curve. Most modern ESR spectrometers are very stable and do not require internal or external standardization.

Determination of the Dose Rate

The overall dose rate is calculated from the concentrations of the radioactive elements in the sample and their surroundings; only the U and Th decay chains and ^{40}K are of relevance, although there is a minor contribution from ^{87}Rb in sediments, plus a small component from cosmic radiation. There are three types of ionizing radiation from these sources:

- Gamma (γ) rays (photons), with a range in sediments of ~ 30 cm.
- Beta (β) rays (electrons), with a range of ~ 2 mm.
- Alpha (α) rays (He nuclei), with a range of $20\text{--}40$ μm . Alpha particles are less efficient in producing ESR intensity than β and γ radiation. An α -particle efficiency needs to be determined ([Aitken, 1985](#)). The α -efficiency is expressed as a fraction of the equivalent dose for β and γ required to produce the same ESR intensity. This is ~ 0.06 for coral and 0.13 for tooth enamel.

For conversion of the elemental concentrations into dose rates see [Aitken \(1985\)](#), and [Adamiec and Aitken \(1998\)](#). The concentrations within samples are usually very different from those from their surroundings, meaning that the internal and external dose rates need separate evaluation.

External Dose Rate

The calculation of the external dose rate depends on the size of the samples ([Figure 5](#)). It is possible to remove the outer ~ 50 μm of the sample to eliminate the effects of external α -radiation. If it is possible to remove 2 mm, then the β -radiation effects can also be eliminated, leaving only the γ - and cosmic-ray effects. If the grain size of the sample is < 2 mm, obviously it will have β -radiation effects. It is also important to separately calculate the dose for β and γ , since they originate from different volumes of the surrounding sediment, to a distance of 2 mm or 30 cm, respectively ([Figure 5](#)). The external β dose rate can be estimated from the composition

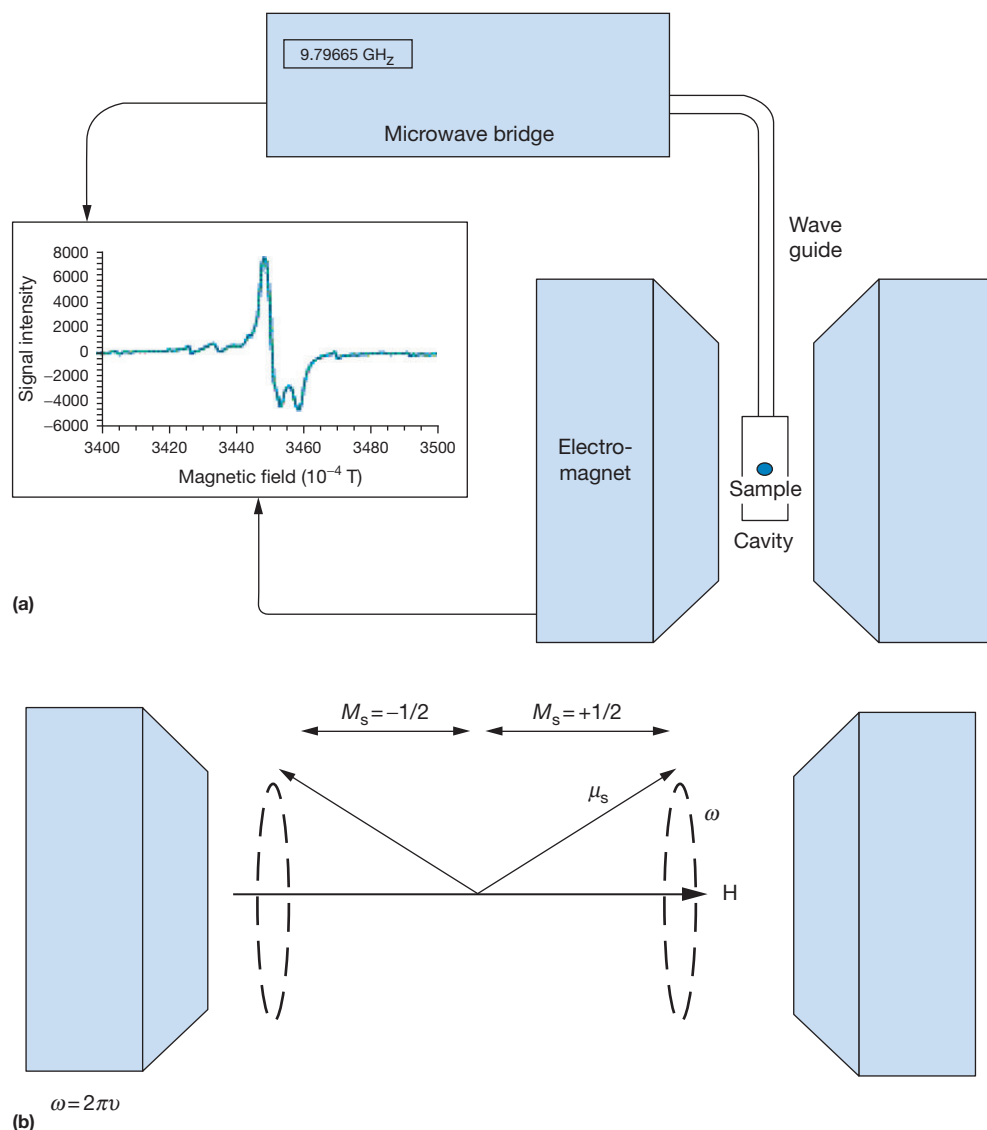


Figure 4 (a) ESR spectrometer. It consists of three main units: a strong electromagnet, a microwave generator, and an electronic processing unit. The sample is measured in the cavity, which is located in a strong magnetic field. The magnetic field can be linearly changed. In resonance, microwave energy is absorbed by the sample, giving rise to an ESR spectrum as shown in Figure 2. (b) Orientation of the magnetic moment of an electron in an external magnetic field. M_s , spin quantum number; μ_s , magnetic moment of spin; ω , precession frequency; ν , microwave frequency that is required for inducing a flip of the magnetic moment into the opposite direction.

of U, Th, and K in the surrounding sediment. In general, the γ dose must be determined using *in situ* measurements using a portable, calibrated γ spectrometer or TL dosimeters. *In situ* measurements have the advantage of reflecting the current-day water content of the sediment, since the presence of water affects the flux of β and γ radiation. It is also important to estimate the cosmic-ray dose rate, which is $\sim 300 \mu\text{Gy year}^{-1}$ at sea level but also is present at a reduced level below ground and varies with altitude as well as latitude.

Internal Dose Rate

The internal dose rate is mainly generated by α and β rays emitted from elements in the sample itself. In secondary carbonates and the constituents of teeth (enamel and

dentine), the U-decay chains are in disequilibrium. This affects the calculated dose, which needs to be calculated mathematically from the observed disequilibrium of ^{238}U . Generally, the α -efficiency is difficult to estimate because of the size requirements, but the best estimates are 0.13 ± 0.02 , 0.07 ± 0.01 , and 0.06 ± 0.02 for tooth enamel, mollusks, and coral, respectively. Dose-rate calculations can become complex if U-series disequilibrium or attenuation needs to be considered (Grün, 1989).

In tooth enamel and mollusks, dose-rate calculations can be further complicated by the accumulation of U in these materials over time. Modern samples are virtually free of U (1–10 ppb), whereas fossil samples may contain up to several hundred ppm U in bones and teeth. The history of the uranium uptake with time may have significant implications for the

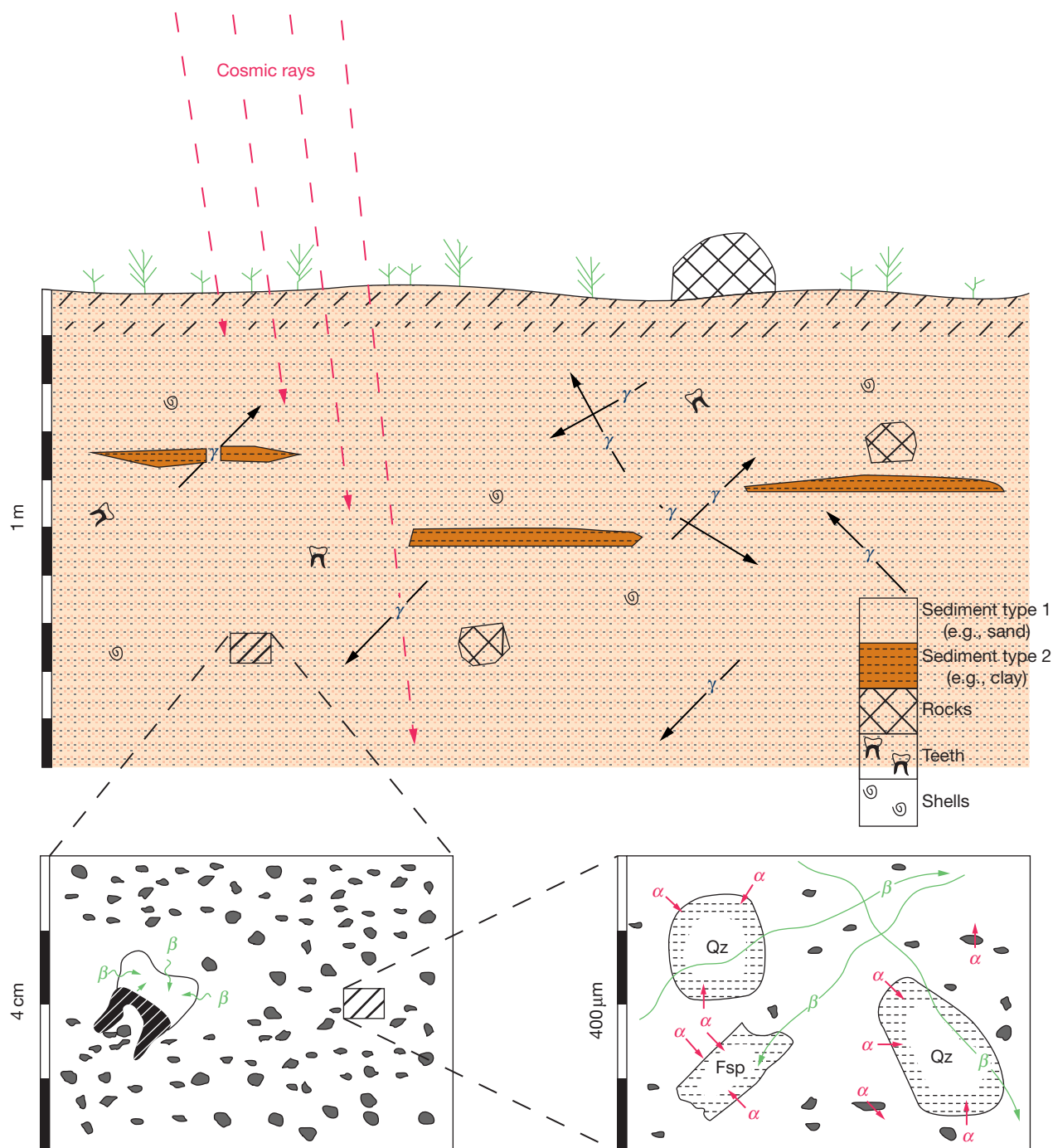


Figure 5 Schematic representation of the different external components of natural radiation relevant for dose-rate calculations (Qz, quartz; Fsp, feldspar). Cosmic rays are high-energy particles and are attenuated once they penetrate the sedimentary layers. For practical purposes, the cosmic-ray dose rate becomes negligible at a depth of approximately 20 m. Gamma rays have an average range of 30 cm. In homogeneous, fine-grained sediments, the gamma dose rate can be calculated from the chemical analysis of the bulk sediment. However, if the sediment contains boulders or intercalation of layers with different radioactivity (represented by the two sediment types), the gamma dose rate should be measured *in situ*. Beta rays have average ranges of a few millimeters. In smaller samples, such as teeth or shells, the volume that has received external beta dosage cannot be removed. For dose-rate calculations, the sediment in the immediate surroundings of the sample is analyzed for radioactive elements. Alpha rays have average ranges of a few tens of micrometers. In addition to the external dose-rate sources, samples receive a portion of the internally generated alpha, beta, and gamma rays. Adapted version of Figure 2.2 (p. 42) originally drawn by Steven Stokes from 'An Introduction to Optical Dating' by Aitken (1988), by permission of Oxford University Press.

average dose-rate calculation. Conventionally, two different U-uptake models have been considered: early U uptake (EU) and linear U uptake (LU). For the EU model, it is assumed that all the U present in the teeth today was accumulated shortly after burial, while the LU model assumes the U has been accumulated continuously in a linear fashion. The differences in these calculations are small if the U concentration is moderate. However, with increasing U concentration, the differences between the LU and EU age can be a factor of two. In general interpretations of ESR results, the correct age is considered to be bracketed by the LU and EU results. It has to be recognized that the U-uptake history might be outside the two limiting approaches of LU and EU for 10–20% of teeth samples. Examples of differences are U leaching or strongly delayed U uptake due to hydrologic changes. It should be emphasized that if teeth or shells contain considerable amounts of U, then the standard models may yield grossly discrepant results. There is no *a priori* way to prefer one model over the other.

To overcome the question of unknown U-uptake history, Grün et al. (1988) and Grün (2009) suggested analyzing U-series isotopes in the ESR samples. A one-parameter U-uptake model equation is used:

$$U(t) = U_m(t/T)^{p+1} \quad [5]$$

where $U(t)$ is the uranium concentration at time t , U_m is the measured present-day U concentration, T is the age of the sample (this must be estimated from the U-series result!), and p is an uptake parameter. Some examples of the resulting functions are shown in Figure 6(a). The measurement of U-series isotopic data is used to establish the relationship between apparent U-series age and p -value, starting from a closed system, which corresponds to $p = -1$ (Figure 6(b)). Different constituents of teeth tend to give somewhat different U-series isotopic results, indicating variations in their closed system behavior. The p versus age relationship can then be used, along with other dose information, to compute the dose values for any given age (Figure 6(c)). The projection of measured D_e onto this function gives a combined U-series/ESR age, and returning this calculated age allows one to estimate the corresponding p -values.

Measurement of Radioactive Elements and Isotopes

A large number of analytical techniques are available for estimating the concentrations of radioactive elements. In this field, external γ dose can be measured with TL dosimeters or a γ spectrometer. Other methods for elemental analysis include X-ray fluorescence, inductively coupled plasma mass spectrometry (ICPMS), and neutron activation analysis. In teeth and mollusk shells with U concentrations >0.5 ppm, uranium and thorium concentrations, as well as U-series isotopes, can be measured *in situ* by laser-ablation multicollector ICPMS (Eggins et al., 2005).

Errors

Although error calculations ought to be straightforward (Aitken, 1985, Appendix B), most publications mix random and systematic errors. Most ESR results are quoted to one standard deviation (1σ). The error in D_e is usually within 1–5%. However,

depending on the number of dose parameters and the analytical techniques used, typical precision can be in the range 3–7%. Sources of systematic error derive from calibration of radiation sources of field γ -ray spectrometers. Other systematic errors can arise because of assumptions, particularly if certain parameters cannot be directly measured. The water content of sediments may have changed over time, or some sediments might have had disequilibrium in the U-series system. Additionally, the α -efficiency is difficult to measure and based on a few analyses. These systematic effects can be evaluated only by comparison of ESR results with those of independent dating techniques.

Many archaeological and paleoanthropological sites have been excavated in the past, so determination of external dose rate is not possible. Sometimes, reconstructions have been attempted from museum sediment samples. Clearly, these estimates can have large errors ($>20\%$) since there are many effects that can arise: the samples may be inhomogeneous in U, Th, and K; the physical characteristics of the original sedimentary environment are not preserved; nor is the water content. If teeth contain significant U, errors can be as much as 50% because of the unknown uptake history of U. In the optimal situation where all parameters are known, 1σ errors for teeth and mollusk shells are about 3–9%, and those of corals are lower at 1–3%. Systematic errors in these contexts may be as high as 5%.

Applications of ESR Dating

ESR dating has been applied to a wide variety of materials. Unfortunately, results have sometimes turned out to be questionable, or other dating methods could be applied to a particular material more easily. For example, although it is possible to use ESR for speleothem dating, U-series is much more precise and requires less sample and effort. Particularly because of advances in ICPMS measurements of U-series isotopes, U-series dating of secondary carbonates is now much faster and cheaper than ESR. Some examples of exotic materials where ESR was used to estimate ages are by Okumura et al. (2010), who tried to use ESR to date barites from hydrothermal vent chimneys in the ocean – the ages resulting were considerably different from $^{210}\text{Pb}/^{226}\text{Ra}$ ages. Another example is by Wildberger et al. (2010), who unsuccessfully tried to use ESR in a speleothem study, compared to conventional U–Th approaches. The most successful applications of ESR have been to tooth-enamel dating.

Corals

ESR has played an important role in sorting out the chronology of raised coral reef tracts in Barbados (Schellmann and Radtke, 2004). Figure 7 shows a comparison of ESR age estimates with calibrated radiocarbon and thermal ionization mass spectrometry U-series dates. The ESR age estimates agree well with the ^{14}C results in the range of 0–3500 years, and both methods result in similar errors (Figure 7(a)). When comparing the ESR to U-series results in the range of 70–130 ka, ESR appears to give results somewhat younger than U-series, and errors are significantly larger. This may be due to an overestimate of the U-series age, as ESR is significantly less affected by diagenetic process than U-series (Eggins et al., 2005). Blackwell et al.

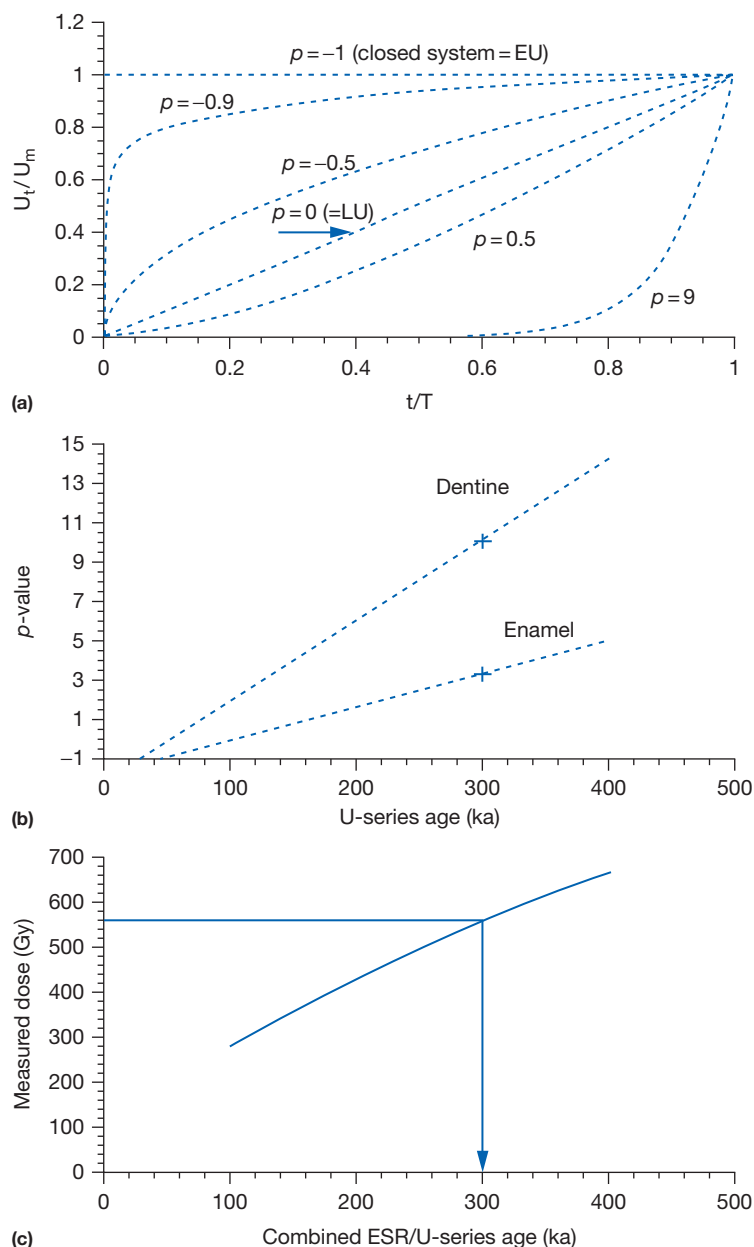


Figure 6 Combination of U-series and ESR dating. (a) The U-uptake history using eqn [5] for different p -values. A p -value of -1 corresponds to a closed system (= EU model), $p = 0$ to the linear-uptake model, and $p > 0$ to sublinear, initially delayed U uptake. (b) Apparent U-series ages are calculated from the measured isotopic ratios (i.e., $^{230}\text{Th}/^{234}\text{U}$ and $^{234}\text{U}/^{238}\text{U}$) and a p -value starting from $p = -1$. The crosses indicate the p -values for dentine and enamel that are obtained after the calculation of the U-series/ESR age (c). (c) Using the relationship between p -values and apparent U-series age estimates (from (b)) in combination with all other dose-rate parameters, doses are calculated for any given age. The projection of the measured D_e value onto this function yields the age of the sample, and the p -values for the different constituencies of the tooth can be read from (b). Reproduced from Grün R, Schwarcz HP, and Chadam JM (1988) ESR dating of tooth enamel: Coupled correction for U-uptake and U-series disequilibrium. *Nuclear Tracks and Radiation Measurements* 14: 237–241.

(2007) were able to compare ESR and U-series dates on corals from 6.4 to about 137 ka in age, although the ESR dates of the oldest corals deviated from the U–Th ages to the younger age, as already noted by Eggin et al. (2005).

Mollusk Shells

Mollusk shells are notoriously difficult to date since they tend to accumulate U and exchange CO_2 , causing problems both in

U-series and radiocarbon dating. ESR is also complicated by the fact that dose–response curves are complex, with inflection points, and cannot be easily described by a simple dose-rate function. Based on a large number of samples of known age, a ‘dose-plateau’ procedure was developed and applied to a number of Patagonian shoreline deposits (Figure 8; Schellmann and Radtke, 2003). For deposits correlated with marine isotope stages (MIS) 1–7, it appears possible to use ESR for general assignment of samples to one particular stage. For

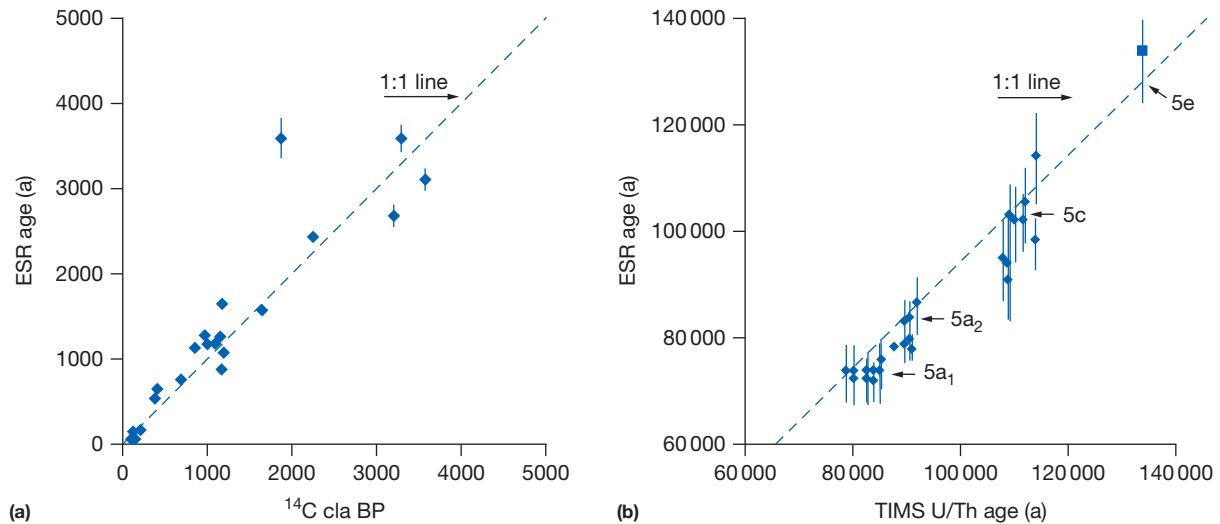


Figure 7 Comparison of ESR age estimates on corals with independent age estimates on the same samples. (a) ESR versus calibrated radiocarbon dates. The error bars of both methods are comparable; ESR seems generally to produce slightly older ages. (b) ESR versus U-series ages (screened for initial $^{234}\text{U}/^{238}\text{U}$ ratios close to seawater) on samples from Barbados. The ages of the different units are indicated by arrows (their extension to the 1:1 line). Generally, ESR has significantly larger errors than U-series and seems to slightly underestimate the expected ages, whereas U-series has a tendency to give slightly older age estimates. Adapted from Schellmann G and Radtke U (2004) The marine Quaternary of Barbados. *Kölner Geographische Arbeiten* 81: 1–137, with permission.

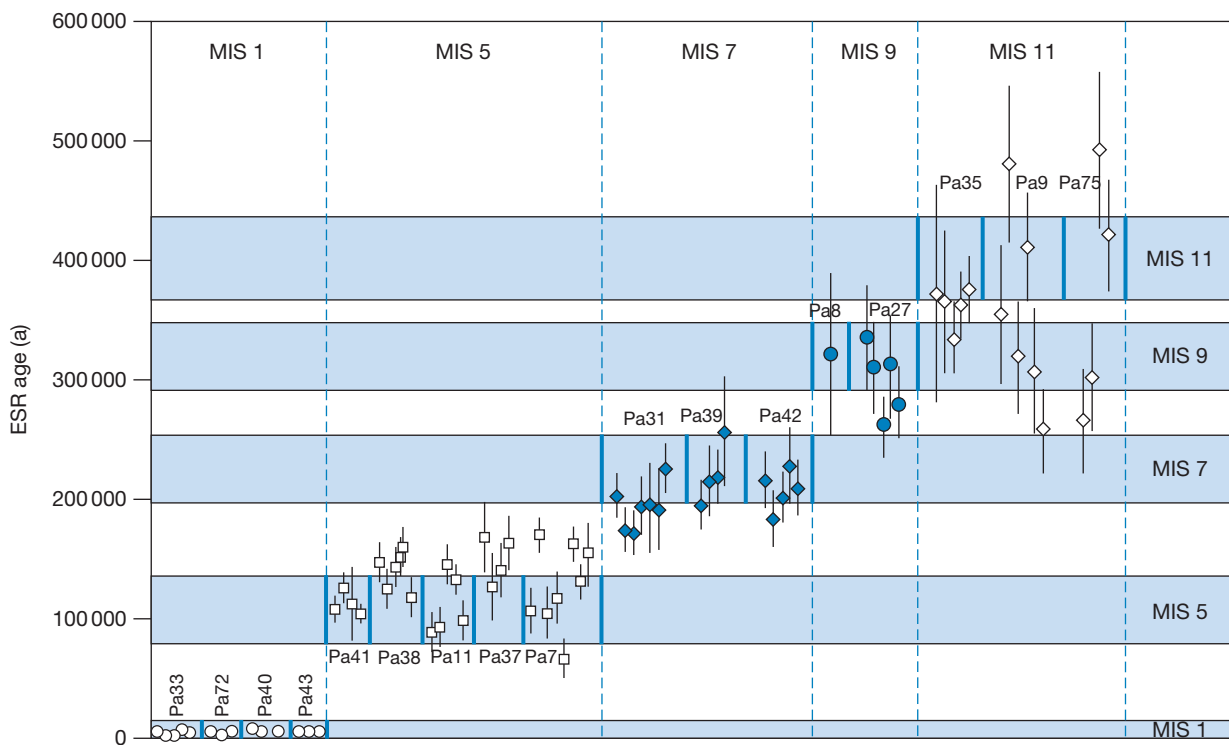


Figure 8 Comparison of ESR age estimates on mollusk shells from marine deposits (indicated by the Pa numbers) in Patagonia with independent geomorphological assessments. ESR can be used for the secure assignment of the remnants of sea-level high stands to marine isotope stages (MIS) 1, 5, and 7. However, the uncertainties in the ESR age estimates of older deposits do not allow correlations of geological formations to a specific MIS. Adapted from Schellmann G and Radtke U (2003) Die Datierung litoraler Ablagerungen (Korallenriffe, Stranwälle, Küstendünen) mit Hilfe der Elektronenspin-Resonanz-Methode (ESR). *Essener Geographische Arbeiten* 35: 95–113, with permission.

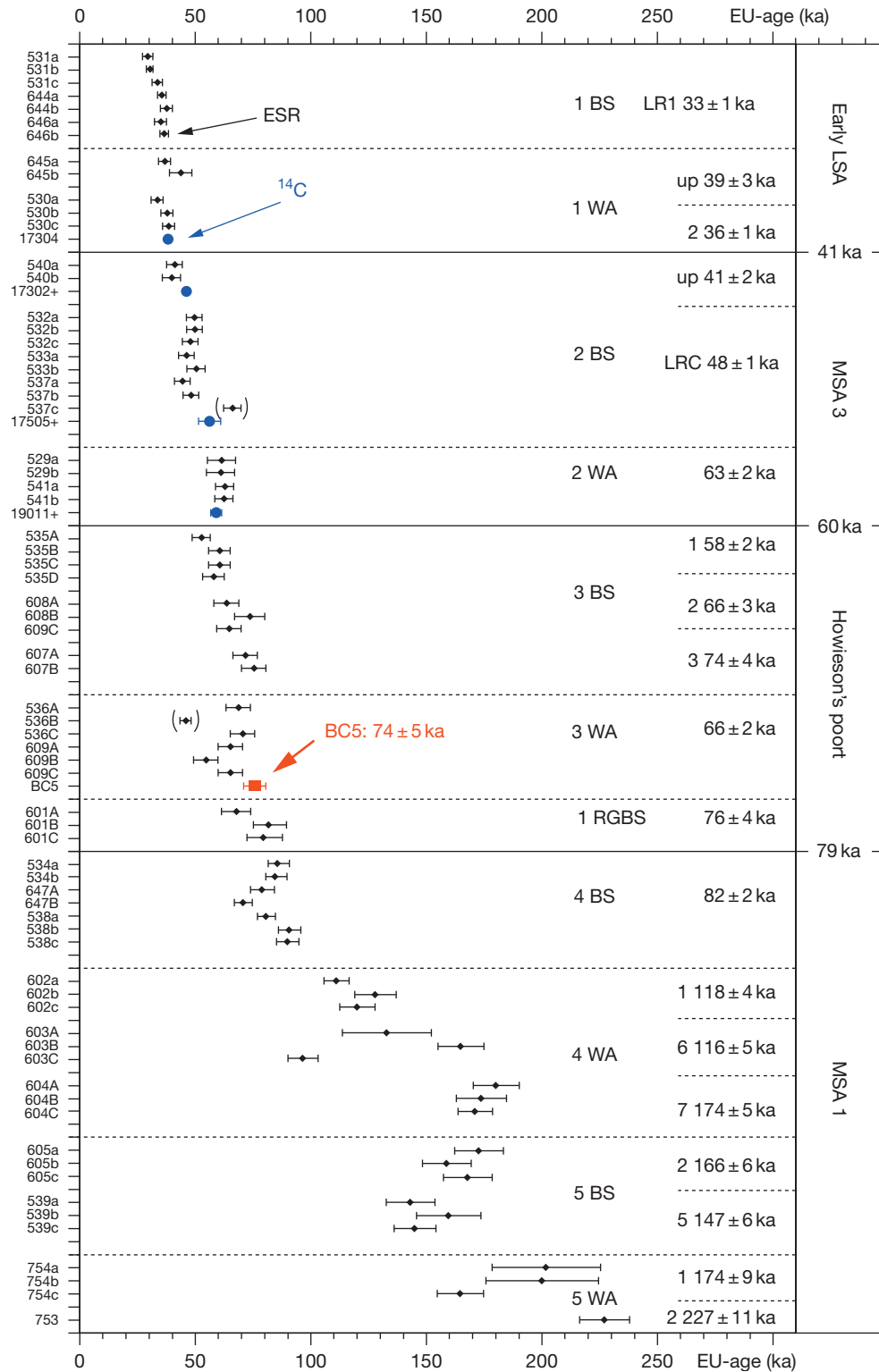


Figure 9 The age of BC5 in context with the ESR and ^{14}C chronology for Border Cave. Lowercase letters following the sample number denote subsamples of a single tooth, and uppercase letters denote separate enamel fragments. The two bracketed results were not used for the calculation of the average ages of the units. Adapted from Grün R, Beaumont P, Tobias PV, and Eggins S (2003) On the age of Border Cave 5 human mandible. *Journal of Human Evolution* 45: 155–167; Bird MI, Fifield LK, Santos GM, et al. (2003) Radiocarbon dating from 40–0 ka BP at Border Cave, South Africa. *Quaternary Science Reviews* 22: 943–947.

older units, errors become too large. However, ESR was very important in eradicating the myth of an MIS-3 sea-level high stand in South America, which was due entirely to radiocarbon dating of samples without proper cleaning to remove CO₂ contamination (Radtko, 1989). Titschack et al. (2009) were able to combine ESR with microstructure and stable-isotope studies of shells to suggest that ESR could be used to indicate times of diagenesis from aragonite to calcite in shells from Rhodes (Greece).

Tooth Enamel

ESR dating of tooth enamel has been applied in a number of archaeological and paleoanthropological sites (Grün, 1997, 2000; Grün and Stringer, 1991; Rink, 1997). For early modern human sites in the Klasies River Mouth Cave (South Africa), as well as Qafzeh and Skhul in Israel, ESR dating has demonstrated the occurrence of anatomically modern humans approximately 100 000 years ago.

Recently, Falguères et al. (2010) were able to demonstrate the usefulness of combined ESR and U-series methods to date several major Acheulian sites in Spain, Italy, and France to the period of 300–600 ka. Another interesting example is the study of Skinner et al. (2007b), who used ESR coupled with ²³⁰Th/²³⁴U to develop an LU model to establish a date of 57–87 ka on teeth from a Paleolithic site, Obi-Rakhmat, in Uzbekistan. ¹⁴C ages at this site suggest a later age, although this dating is still part of an ongoing discussion. Krivoshapkin et al. (2010) noted that different uptake models in Skinner et al. (2007b) give different results and argue for a lower age close to the limit of ¹⁴C dating. Liberda et al. (2010) have been able to use ESR in tooth enamel of the Mousterian layer (#20) from El Castillo, Spain, to obtain ages of 42.7 ± 6.4 ka, consistent with ¹⁴C dates on the same horizons. Yin et al. (2011) applied combined ESR and U-series to dating of the controversial Dali skull. U-series and ESR gave consistent results for one tooth, of about 281 ± 46 ka, although measurements on other teeth gave discrepancies between U-series and ESR dates.

The important strength of ESR is that it is virtually nondestructive and allows direct analysis of tooth-enamel fragments from human fossils. The archaeological and paleoanthropological site of Border Cave contains one of the best examples of a detailed ESR dating sequence (Figure 9). The dating results for the younger horizons agree well with radiocarbon ages on charcoal collected from the same layers. Of particular interest is the age of Border Cave 5, one of the few early anatomically modern hominid fossils found *in situ* at this location. A study based on the analysis of nitrogen contents and infrared splitting factors indicated that BC5 could be as young as the Iron Age (~1000 years old at this site; Sillen and Morris, 1996). ESR analysis of tooth fragments and determination of the U concentration by ICPMS agreed with the age of the older samples in the surrounding for this level (Figure 9). The best age estimate of 74 ± 5 ka implies that fully modern humans lived in southern Africa significantly earlier than in Europe.

In some types of samples, the ESR signal can be very durable; Skinner et al. (2007a) discuss the dating of crocodile teeth from Australia with ages of >1 My.

Bones

All attempts to obtain reliable ESR dates on bones have been unsuccessful, although this would be a most useful application. Unfortunately, bones absorb more U than teeth, and the mineral phase is only 40–60% of the bone. During fossilization, the mineralogical composition changes, resulting in new minerals and disintegration of others, as well as conversion of an amorphous phase into hydroxyapatite. Formation of new mineral phases with time will result in new electronic defects (traps), resulting in difficulty in estimating age due to these additional variables, no matter which U-uptake model is used.

Quartz

ESR study of quartz has been used for dating of river terraces (Voinchet et al., 2004). The main advantage of ESR over luminescence is that the Al and Ti centers in quartz give significantly higher ESR saturation levels than luminescence methods.

This means older samples should be routinely datable. The disadvantage of ESR is the very long bleaching times required to reset these paramagnetic centers (several hours for Ti and days for Al). This makes it difficult to know whether samples were completely bleached before burial. Comprehensive comparisons of ESR versus TL and OSL for quartz are still needed. In a pilot study, ESR was used to reconstruct the thermal history of the Eldzhurtinskiy granite, in the Caucasus region (Grün et al., 1999). The implication of this study was that ESR has a possible application in detailed landscape reconstruction in rapidly uplifting terrains. More recently, Tissoux et al. (2008) have attempted to apply ESR to quartz in fluvial deposits, who also undertook bleaching experiments on the Ti and Al centers. Similarly, Voinchet et al. (2007) looked at fluvial sediments in order to try to establish the optimal Al bleaching time, showing that the samples can be bleached in the first kilometer of transport, even though laboratory experiments suggest longer bleaching times.

See also: U-Series Dating. **Archaeological Records:** Neanderthal Demise. **Introduction:** History of Dating Methods. **Luminescence Dating:** Optical Dating; Thermoluminescence.

References

- Adamiec G and Aitken MJ (1998) Dose-rate conversion factors: Update. *Ancient TL* 16: 37–50.
- Aitken MJ (1985) *Thermoluminescence Dating*. New York: Academic Press.
- Aitken MJ (1988) *Optical Dating*. Oxford: Oxford University Press.
- Blackwell BAB, Teng SJT, Lundberg JA, Blickstein JIB, and Skinner AR (2007) Coupled ²³⁰Th/²³⁴U-ESR analyses for corals: A new method to assess sea level change. *Radiation Measurements* 42: 1250–1255.
- Curnoe D, Grün R, and Thackeray F (2001) Direct ESR dating of a Pliocene hominid from Swartkrans. *Journal of Human Evolution* 40: 379–391.
- Eggins S, Grün R, McCulloch M, et al. (2005) In situ U-series dating by laser-ablation multi-collector ICPMS: New prospects for Quaternary geochronology. *Quaternary Science Reviews* 24: 2523–2528.
- Falguères C, Bahain JJ, Duval M, et al. (2010) A 300–600 ka ESR/U-series chronology of Acheulian sites in Western Europe. *Quaternary International* 223–224: 293–298.

- Grün R (1989) Electron spin resonance (ESR) dating. *Quaternary International* 1: 65–109.
- Grün R (1997) Electron spin resonance dating. In: Taylor RE and Aitken MJ (eds.) *Chronometric Dating in Archaeology*, pp. 217–261. New York: Plenum.
- Grün R (2000) Electron spin resonance dating. In: Clilberto E and Spoto G (eds.) *Modern Analytical Methods in Art and Archaeology. Chemical Analysis Series* 155: 641–679. New York: Wiley.
- Grün R (2009) The relevance of parametric U-uptake models in ESR age calculations. *Radiation Measurements* 44: 472–476.
- Grün R, Joannes-Boyau R, and Stringer C (2008) Two types of CO_2^- radicals threaten the fundamentals of ESR dating of tooth enamel. *Quaternary Geochronology* 3: 150–172.
- Grün R, Schwarcz HP, and Chadam JM (1988) ESR dating of tooth enamel: Coupled correction for U-uptake and U-series disequilibrium. *Nuclear Tracks and Radiation Measurements* 14: 237–241.
- Grün R and Stringer CB (1991) ESR dating and the evolution of modern humans. *Archaeometry* 33: 153–199.
- Grün R, Tani A, Gurbanov A, Koshug D, Williams I, and Braun J (1999) A new method for the estimation of cooling and denudation rates using paramagnetic centers in quartz: A case study on the Eldzhurtinskiy Granite, Caucasus. *Journal of Geophysical Research* 104(B8): 17531–17549.
- Joannes-Boyau R and Grün R (2011) A comprehensive model for CO_2^- radicals in fossil teeth enamel: Implications for ESR dating. *Quaternary Geochronology* 6: 82–87.
- Krivoshapkin AI, Kuzmin YV, and Jull AJT (2010) Chronology of the Obi-Rakhmat grotto (Uzbekistan): First results on the dating and problems of the Paleolithic key site in central Asia. *Radiocarbon* 52: 549–554.
- Liberda JJ, Thompson JW, Rink WJ, et al. (2010) ESR dating of tooth enamel in Mousterian layer 20, El Castillo, Spain. *Geoarchaeology* 25: 467–474.
- Okumura T, Toyoda S, Sato F, Uchida A, Ishibashi J-I, and Nakai S (2010) ESR dating of marine barite in chimneys deposited from hydrothermal vents. *Geochronometria* 37: 57–61.
- Radtke U (1989) Marine Terrassen und Korallenriffe—das Problem der quartären Meeresspiegelschwankungen erläutert an Fallstudien aus Chile, Argentinien und Barbados. *Düsseldorfer Geographische Arbeiten* 27: 1–245.
- Rink WJ (1997) Electron spin resonance (ESR) dating and ESR applications in Quaternary Science and archaeometry. *Radiation Measurements* 27: 975–1025.
- Schellmann G and Radtke U (2003) Die Datierung litorals Ablagerungen (Korallenriffe, Stranwälle, Küstendünen) mit Hilfe der Elektronenspin-Resonanz-Methode (ESR). *Essener Geographische Arbeiten* 35: 95–113.
- Schellmann G and Radtke U (2004) The marine Quaternary of Barbados. *Kölner Geographische Arbeiten* 81: 1–137.
- Schwarcz HP, Grün R, and Tobias PV (1994) ESR dating studies of the australopithecine site of Sterkfontein, South Africa. *Journal of Human Evolution* 26: 175–181.
- Sillen A and Morris AG (1996) Diagenesis of bone from Border Cave: Implications for the age of the Border Cave hominids. *Journal of Human Evolution* 31: 499–506.
- Skinner AR, Blackwell BAB, Hasan MM, and Blickstein JIB (2007) Expanding the range of electron spin resonance dating. *Archaeological Chemistry. American Chemical Society Symposium Series*, vol. 968, pp. 1–14. Washington, DC: American Chemical Society.
- Skinner AR, Blackwell BAB, Mian A, et al. (2007) ESR analyses on tooth enamel from the Paleolithic layers at the Obi-Rakhmat hominid site, Uzbekistan: Tackling a dating controversy. *Radiation Measurements* 42: 1237–1242.
- Tissoux H, Toyoda S, Falguères C, et al. (2008) ESR dating of sedimentary quartz from two Pleistocene deposits using Al and Ti centers. *Geochronometria* 30: 23–31.
- Titschack J, Radtke U, and Freiwald A (2009) Dating and characterization of polymorphic transformation of aragonite to calcite in Pleistocene bivalves from Rhodes (Greece) by combined shell microstructure, stable isotope and electron spin resonance study. *Journal of Sedimentary Research* 79: 332–346.
- Voinchet P, Bahain JJ, Falguères C, et al. (2004) ESR dating of quartz extracted from Quaternary sediments: Application to fluvial terrace systems of northern France. *Quaternaire* 15: 135–141.
- Voinchet P, Falguères C, Tissoux H, Bahain JJ, Despriée J, and Pirouelle F (2007) ESR dating of fluvial quartz: Estimate of the minimal distance transport required for getting a maximum optical bleaching. *Quaternary Geochronology* 2: 363–366.
- Wildberger A, Geyh MA, Groner U, Häuselmann P, Heller F, and Poletze M (2010) Dating speleothems from the Silberer cave system and surrounding areas: Speleogenesis in the Muota Valley, central Switzerland. *Zeitschrift für Geomorphologie* 54(2): 307–328. Supplementary Issues.
- Yin G, Bahain JJ, Shen G, et al. (2011) ESR/U-series study of teeth recovered from the paleoanthropological stratum of the Dali man (Shaanxi Province, China). *Quaternary Geochronology* 6: 98–105.

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M

Magnetic Polarity Studies *see* Geomagnetic Excursions and Secular Variations

Mammalian Evolution *see* Vertebrate Overview; **Vertebrate Records:** Early Pleistocene; Mid-Pleistocene of Africa; Mid-Pleistocene of Southern Asia; **Vertebrate Studies:** Ancient DNA; Speciation and Evolutionary Trends in Quaternary Vertebrates

Marine Isotope Stages *see* **Paleoceanography, Physical and Chemical Proxies:** Oxygen-Isotope Stratigraphy of the Oceans

Megafauna, Pleistocene *see* Vertebrate Overview; **Vertebrate Records:** Mid-Pleistocene of Australia; Mid-Pleistocene of North America; **Vertebrate Records:** Early Pleistocene; Late Pleistocene of Africa; Late Pleistocene of North America; Late Pleistocene of South America; Late Pleistocene of Southeast Asia; Late Pleistocene Megafaunal Extinctions; Mid-Pleistocene of Africa; Mid-Pleistocene of Europe; Mid-Pleistocene of Southern Asia; **Vertebrate Studies:** Interactions with Hominids

Megafaunal Extinction *see* **Vertebrate Records:** Late Pleistocene Megafaunal Extinctions; **Vertebrate Studies:** Interactions with Hominids

Methane Studies, Ice Cores *see* **Ice Core Methods:** Methane Studies

Mg/Ca and Sr/Ca Studies *see* **Paleoceanography, Physical and Chemical Proxies:** Mg/Ca and Sr/Ca Paleothermometry from Calcareous Marine Fossils

Micromorphology of Sediments *see* **Glacial Landforms, Sediments:** Micromorphology of Glacial Sediments; **Paleosols and Wind-Blown Sediments:** Soil Micromorphology

Midges *see* **Chironomid Records:** Africa; Chironomid Overview; Late Pleistocene of Europe; Postglacial Europe; Postglacial Southern Hemisphere

Milankovitch Theory *see* **Glaciation, Causes:** Astronomical Theory of Paleoclimates

Mollusks, Marine *see* **Paleoceanography, Biological Proxies:** Corals, Sclerosponges and Mollusks

Moraines, Glacial *see* **Glacial Landforms:** Moraine Forms and Genesis; Evidence of Glacier Recession; Glacial Landsystems; Glacitectonic Structures and Landforms; **Glacial Landforms, Sediments:** Glacial Erratics and Till Dispersal Indicators

Morphostratigraphy *see* **Quaternary Stratigraphy:** Morphostratigraphy/Allostratigraphy

N

Neanderthal Demise *see* **Archaeological Records:** Neanderthal Demise

Neoglaciation *see* **Glaciations:** Neoglaciation in Europe; Neoglaciation in the American Cordilleras

O

Optically-Stimulated Luminescence Dating *see* **Luminescence Dating:** Optical Dating

Oribatid Mites

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Introduction

Oribatid mite studies are only now reaching the main stream of Quaternary investigation as researchers seek additional methods for recognizing climate change based on organismal proxies of habitat. In most aquatic and semiaquatic environments, these minute arthropods ([Figure 1](#)) are preserved well enough and in sufficient numbers to make them very useful for serious investigations of paleoclimate, paleoecology, stratigraphy, or paleolimnology. But they have been largely disregarded, perhaps because their study requires a patient fossil picker and a skilled microscopist with systematic knowledge of the group. Because rewards of information are potentially large, this situation has begun to change. At this point in the development of Paleoacarology, workers are trying to learn enough about the quality and subtlety of the oribatid fossil record to be able to trust interpretations of

that record without a need for secondary corroboration. Errors will be made on the way, and this is a time for conservative interpretation, but it appears that the oribatid fossil record is very robust and full of subtlety and will emerge from careful efforts by knowledgeable paleoacarologists.

Mites in postglacial sediments are often alluded to as being 'subfossil.' When an organism dies, or when it molts, the remains fall to the substrate. When they are interred in a natural manner, they become fossils. A vast array of postdepositional circumstances may befall these fossils; on rare occasions they will even be replaced by mineral matter, but more likely they will not, nor is replacement of any sort of validation required for them to be 'fossil.' Most will be lost to processes of decay and/or biogeochemistry. To be fossil, they need only survive these threats long enough to fall under the inspection of the curious scientist. Often the term 'subfossil' is employed as an

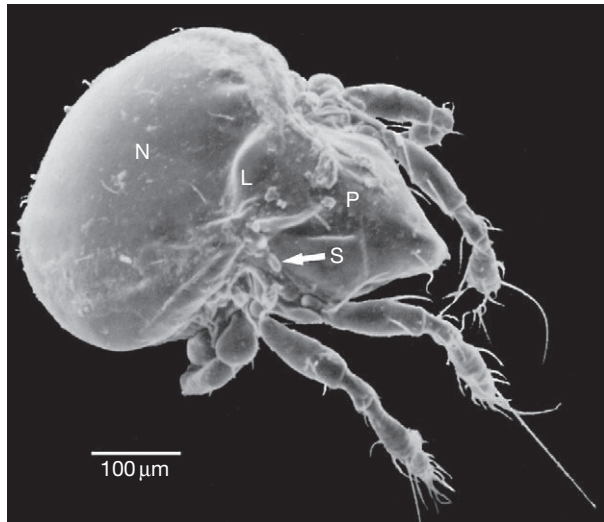


Figure 1 *Hydrozetes* sp. seen in dorsal oblique view illustrating the dorsal shield (N, notogaster) and prodorsal shield (P), the lenticulus (L), and the right sensory sensillus (S) at tip of arrow. Fossil *Hydrozetes* sp. often preserve proximal leg segments but distal segments seen on this recent specimen are usually absent on fossils. Modified from Erickson JM (1988) Fossil oribatid mites as tools for Quaternary paleoecologists: Preservation quality, quantities, and taphonomy. In: Laub RS, Miller NG, and Steadman DW (eds.) *late Pleistocene and early Holocene Paleocology and Archaeology of the Eastern Great Lakes Region*. *Bulletin of the Buffalo Society of Natural Sciences*, vol. 33, pp. 207–226. Buffalo, NY: Buffalo Society of Natural Sciences.

apology for expressing the geologic age of the specimen when none is needed. There is no gradation from cadaver to skeleton to fossilwannabe to subfossil to perifossil to fossil such as the term subfossil implies. We consider this term inappropriate and recommend it not be used. As soon as an organism is deposited as part of the sediment it is a fossil and workers must begin to think about it in terms of geologic processes and their implications. In particular, one of us (JME) believes that by considering late Pleistocene and Holocene mites, and most other geologically young fossils, as ‘subfossils’ they have not been properly treated by either neontologists or paleontologists.

The presence of fossil oribatids was recognized a century ago by Nordenskiöld (1901), and their potential was recognized by Frey (1964) who suggested that fossil mites would one day prove useful to Quaternary geoscientists in much the same ways that fossil pollen and diatoms were then proving to be. Coleoptera, Cladocera, and Chironomida were beginning to be recognized in the Quaternary fossil record (Coope, 1967; Frey, 1964) as the concept of paleoclimate and paleohabitat proxies emerged. Each of these taxa has its own distinctive information to provide to Quaternary paleoecologists. Our purpose here will be to clarify current, distinctive uses of fossil Oribatida. It will include an introduction to the two eras of fossil mite study, the descriptive era and the applied era, as these can be useful ways to approach the information now available to paleoacarologists. First, a brief introduction to the biological affinities and habits of the group is in order.

What are Oribatid Mites?

The arthropod subclass Acari represents an enormous diversity of minute, eight-legged animals identified as mites and ticks. Reduced segmentation and lack of an independent head structure readily differentiate them from arachnid relatives such as spiders and scorpions (Walter and Proctor, 1999). These globally distributed and ancient animals utilize a wide variety of trophic styles, while behaviorally interacting with other organisms and natural systems to fill countless niches. Oribatid mites, commonly referred to as ‘moss mites’ or ‘beetle mites,’ are a diverse group of mostly terrestrial acarids with densities reaching hundreds of thousands per square meter in rich organic soil horizons, particularly those that have well-developed fermentation layers (Norton, 1990). They are the dominant arthropod in temperate to subtropical forest habitats. In addition, they are well distributed in or on other substrates, including decaying woody and nonwoody vascular plants, macro fungi, mosses, algae, lichens and even forest canopies (Behan-Pelletier, 2000; Behan-Pelletier and Bissett, 1994; Norton, 1985). Figure 2 presents an array of extant oribatids derived primarily from semiaquatic and aquatic habitats. These are species that are commonly fossilized. Their small size, usually between 200 and 800 μm, allows them to dwell in the microclimate boundary layer of the aforementioned vegetation types.

These minute chelicerates play a significant role in decomposition and nutrient cycling processes as saprophagous and mycophagous organisms grazing on organic residues and microflora by means of direct (selective feeding) and indirect (comminution of particles) relationships with microflora. The average generation time from egg to adult is approximately 1–3 years in temperate climates, and up to 7 years in cold climates (Cannon and Block, 1988; Norton, 1990). Rather than being adaptive, long generation times and relatively low fecundity have been explained as constraints of low secondary production, associated with relatively low quality food compared to that of other arachnids (Norton, 1994). In addition to high densities, more than 10 000 described species in over 1100 genera representing approximately 173 families create a diverse fauna that frequently exceeds 80% of all microarthropods in the soil (Walter and Proctor, 1999). Only a small set of oribatid taxa, 15 genera in 11 families, is hydrophilic, or affiliated with aquatic and semiaquatic habitats (Behan-Pelletier and Eamer, 2007). They are widely distributed in freshwater habitats, but their species diversity is low compared to that of terrestrial communities, and little is known about the microhabitat requirements of particular species. Although most oribatid mites reproduce sexually, approximately 10% are obligatorily parthenogenic, and this mode is dominant in aquatic habitats (Norton and Palmer, 1991; Norton et al., 1993). Fossil *Hydrozetes* spp. are occasionally preserved with unborn juveniles in the body cavity.

Hydrophilic oribatids exhibit multiple morphological and physiological adaptations to the aquatic medium. Brachypylina are active, and have well-developed tracheal systems, usually aided by a ‘plastron’ for gas exchange. Another adaptation is controlled floatation or ‘levitation’ that entails the unexplained formation of an air bubble in the gut, best known in the genus *Hydrozetes* (Newell, 1945). Though not unique to hydrophilic oribatids, a primitive light receptor (lenticulus;

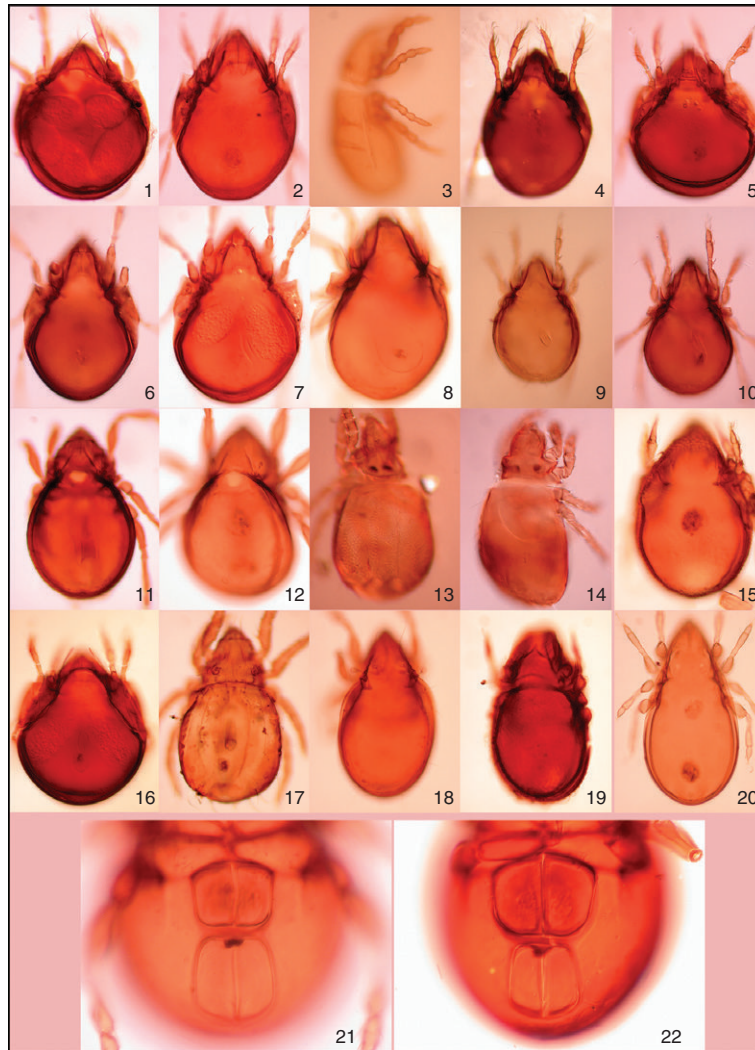


Figure 2 Recent, adult oribatid mites such as may be found in and around aquatic and semiaquatic habitats in northeastern North America. Dorsal views: 1, *Zetomimus francisi*; 2, *Chamobates* sp.; 3, *Eniochthonius minutissimus*; 4, *Heterozetes aquaticus*; 5, *Pelopsis bifurcata*; 6, *Naiazetes* (male); 7, *Naiazetes reevesi* (female); 8, *Limnozetes onondaga*; 9, *Limnozetes* n. sp.; 10, *Limnozetes amnicus*; 11, *Hydrozetes lemnae*; 12, *Hydrozetes indicator*; 13, *Trimalaconothrus maior*; 14, *Trhypochthoniellus crassus*; 15, *Rostrozetes ovulum*; 16, *Punctoribates armipes*; 17, *Platynothrus peltifer*; 18, *Hemileius* sp.; 19, *Carabodes granulatus*; 20, *Liebstadia humerata*. Ventral views genital-anal region: 21, *Tegeocranellus kethleyi*; 22, *Tegeocranellus muscorum*. Most specimens approximately 350 μ m in size, but 21 and 22 are greatly enlarged.

Figure 1) is well developed in *Hydrozetes* and *Tegeocranellus* (**Figures 2(12), 2(13), 2(21), 2(22)**). No oribatids are known to swim, and few show any leg modifications. One exception is *Mucronothrus nasalis*, which has exceptionally flexible legs and is particularly tenacious in moving currents (**Norton et al., 1988**).

How are Oribatids Fossilized?

Widespread distribution and large population numbers of the Oribatida make them excellent candidates for fossilization in deposits near their normal dwelling site. Most fossil mites belong to the Brachypylina because of the chitinous sclerotization and fused skeleton of the group. Ptychotine oribatids have skeletons sclerotized with calcium oxalate (**Norton and Behan-**

Pelletier, 1991) which seem readily destroyed geochemically after death. **Erickson et al. (1985)** and **Erickson (1986, 1988)** have noted the advantages mites offer the Quaternary scientist who seeks local paleoenvironmental information about a site. They do not fly, nor do they suffer much overland or stream transport before they disarticulate, so where found, they are demonstrably close to their life habitat. Hydrophilic taxa likely matured within the basin of deposition; they were autochthonous, whereas specimens arriving at the site of deposition from surrounding terrain are considered allogenic or allochthonous (**Drouk, 1997; Erickson, 1988**). Mites enter the fossil record by overland precipitation runoff, local stream inflow, swash zone wave action, in-fall of vegetation, air-fall from overhanging vegetation and local wind transport of sediment (**Figure 3**). Because they do not fly as beetles do, nor blow long distances

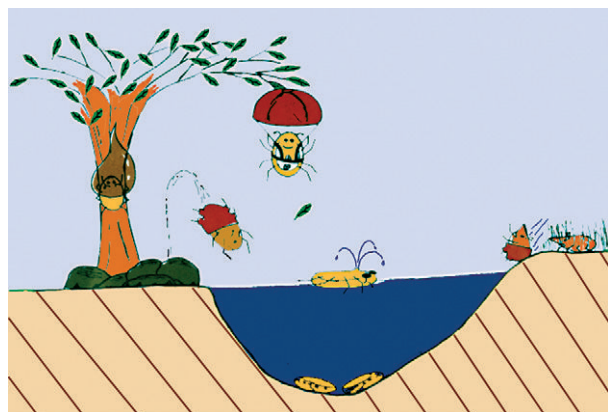


Figure 3 Cartoon demonstrating that oribatid mites reach the depositional environment by several means representative of their original microhabitat domiciles. Tree resins preserved oribatids in amber, a source for most pre-Quaternary oribatid specimens. Mites enter lakes and marshes by overland runoff and shoreline erosion, by dropping from overhanging vegetation, and by *in situ* death of aquatic taxa. Sediment-focusing processes may transport them toward the center of the basin.

on the wind as many pollen species do, they offer excellent characteristics as proxies of local habitat and climate, particularly when those are a reflection of moisture conditions. Because of this they are beginning to emerge as an important tool for Quaternary studies.

Development of Quaternary Oribatid Studies

Human knowledge of organisms is organized and framed around the identity of the creature being studied. That identification is the link between past and future studies making essential the ability to identify the subject of study. The description of oribatid genera and species has been essential to the science, and knowledge of its literature among key workers is a necessity for the use of mites in Quaternary studies.

The Descriptive Era

Quaternary mite neontological studies are based upon the work of acarologists who pioneered the field and devoted great effort toward discovering and describing new taxa. Much of the early work beginning around the beginning of the twentieth century was done by European and North American acarologists that included researchers such as Berlese, Michael, Banks, and Ewing and the expansive morphological works of Grandjean. The next wave of acarologists who sought to better know the global fauna included such workers as Hammer, J. Balogh, Mahunka, Aoki, Jacot, and Woolley. The two-volume work by Balogh and Balogh (1992) summarized the genera of the world in a most useful synthesis. Many more acarologists have contributed significantly to the field but our aim here is to provide entry to the literature of the group.

As much of the pioneer literature was being compiled, Strenzke, Knülle, Willmann, Karpinnen, and Sellnick explored

wider directions of physiology and ecology. Strenzke and Knülle were both interested in community structure and habitat partitioning that influenced present paleontological studies. Sellnick pioneered studies of fossil mites in amber. Amber has been the source of most geologically old fossil material bearing on the evolution of the Oribatida with new taxa continuing to be recognized (Arillo et al., 2008).

Contemporary experts who have made major contributions to oribatid study include Kethley, Walter, Reeves, Alberti, Marshall, and Bocard. Two North American acarologists who have greatly influenced Quaternary fossil mite applications are R. A. Norton and V. Behan-Pelletier. Norton has contributed to comparative morphology, systematics, and paleontology, as well as evolutionary aspects of mite reproductive biology, and ecological roles in decomposer food webs. Behan-Pelletier's expertise includes systematics, biogeography, and ecology of boreal brachypylina mites, especially members of the Ceratozetoidea and aquatic and peatland taxa. All of the aforementioned acarologists' names are intended to be valuable to researchers as keywords to the primary literature. Lastly, the bibliography of oribatid papers of the world compiled by Schatz (2004) and the emerging PDF database being compiled by Norton and Behan-Pelletier are invaluable reference tools.

The Applications Era

Applications for study of oribatid mites have been clear in the agricultural disciplines for a half century or more, but their study in the fossil record remained largely systematic or a cataloging of species occurrences until the rise of interest in climate-related Quaternary research became part of the basic research of Geology after the second half of the twentieth century. At that point, the ecological approaches to oribatid study by Strenzke (1952) and Knülle (1957) made clear their potential utility as proxies of microhabitat and, by extension, as indices of the paleoclimate. It remained only to raise awareness of researchers to those possibilities.

Two areas of application began to emerge. In Europe, Denford (1978) discussed the possible information that oribatid and mesostigmatid mites and related taxa, such as ticks and lice that accompany humans and occupy their dwelling sites, should be able to provide to archaeologists. Generally brachypylina oribatids are far better represented as fossils than their predatory mesostigmatid relatives which have more unsclerotized connective tissue between body segments as we discuss below.

After Denford (1978) called attention to their potential use in archaeology, European acarologists initiated a group of studies aimed at clarification of habitat conditions during human occupation of various archaeological sites, particularly in lowland mires where peat accumulation preserved oribatids (Schelvis, 1987, 1989, 1990). In 1989, Schelvis and Van Geel demonstrated that the record of fossil mite occurrences and abundances through a 2000-year-long section at Usselo, Netherlands, reflected changes similar to those of paleobotanical data from the same section. Schelvis' (1990) study improved on the use of fossil oribatids by integrating the concept of modern habitat groups, 'synusiae' of Strenzke (1952), and the seventeen 'isovalent' mite occurrence groups

of Knülle (1957), predicated on soil moisture, into a semi-quantitative framework based on numbers of recovered individuals of each species. Occurrence values based on counts of each mite species as indexed to its assigned group, ratioed to totals for each group, then provide the basis for calculation of percentages of occurrence of each group among the entire fauna. The resulting spectra of ecological groups, plotted as pie diagrams, formed the basis for making strong inferences of habitat types that prevailed in the past at particular archaeological sites. Schelvis' (1990, 1998) method potentially has great power, especially if it is employed through closely-sampled sections of otherwise homogeneous-appearing sediments, but it has not been exploited yet to full advantage by the archaeological community. In Europe, synusia and isovalent groups of mites are available as tools at the species level whereas these are yet to be established elsewhere.

A group of acarologists in Scandinavia and Russia combined a tradition of systematic and community structure study of modern bog and mire faunas with analysis of the historical development of those faunas based on fossils. These works are more numerous than can be cited here, but some workers who have been deeply involved include Karppinen, Koponen, Markkula, Krivolutsky, and Drouk. Drouk (1997) has pointed out both the applications and the pitfalls of acarological analysis that have emerged from these efforts by suggesting that present knowledge of oribatid distributions and their controlling factors are not yet well enough known, and further, it appears that preglacial mite community species compositions have not been precisely duplicated when glacial conditions have left an area thus making analysis by analogy less trustworthy. Nevertheless, analyses of species distribution, by habitat-

related assemblage structure and by following changes in composition of an ecological group (e.g., Schelvis, 1990) through succession have all been applied satisfactorily. Presently, oribatid community structure in carefully analyzed habitats (e.g., de la Riva-Caballero et al., 2010) is providing ranges of temperature, moisture content, pH and trophic conditions. Thus, oribatid species have demonstrated their utility as contributors to multiproxy reconstructions, having greater precision than that afforded by single proxy studies (Larsen et al., 2006). This is one future for oribatid applications.

In North America, beginnings of the application process focused on understanding relationships of fossil mites to sediments typically available from Quaternary sites. Waters (1979) examined fossil content of several lithologies of late Pleistocene and Holocene sediments cored from Glovers Pond, a 20 000-year-old spring-fed lake in New Jersey, USA. Clastics, including sand, silt and clay, were devoid of oribatids, but biogenic sediments from both littoral and profundal settings, including peat, marl and gyttja contain fossil oribatids in quantities large enough to allow quantitative comparative analyses that are replicable even at small sample sizes available from 5- or 6-cm-diameter cores. Waters' work established that sample sizes as small as 10 g (dry weight) of marl or gyttja contained enough fossil mites to allow comparisons within cores and between cores in the same lake basin. Erickson et al. (1985) noted that 1 g dry weight of lake sediment contained an average of 11 mites thus suggesting that samples >5–7 g wet samples could give internally consistent comparisons of mite populations within cores when data are standardized to 10 g dry weight. A single 30 g, wet-weight sample produced 620 fossil mites. Metzger (1985) conducted up-core

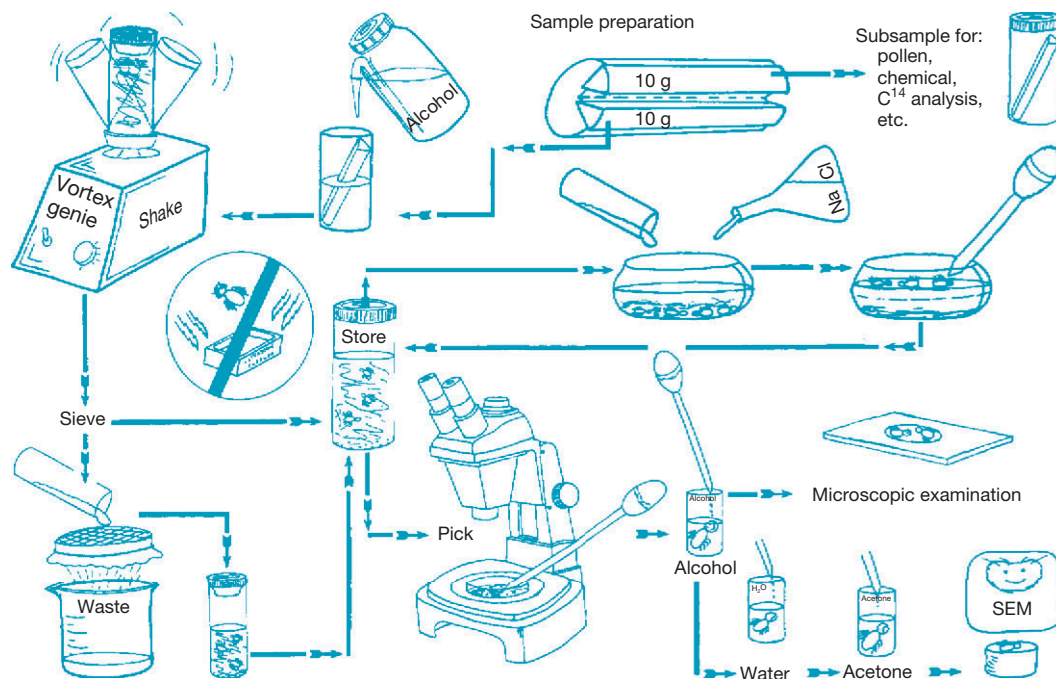


Figure 4 Suggested protocol for lab procedures useful when retrieving, separating, and identifying fossil oribatid mites. Modified from Erickson JM (1988) Fossil oribatid mites as tools for Quaternary paleoecologists: Preservation quality, quantities, and taphonomy. In: Laub RS, Miller NG, and Steadman DW (eds.) *late Pleistocene and early Holocene Paleoecology and Archaeology of the Eastern Great Lakes Region*. *Bulletin of the Buffalo Society of Natural Sciences*, vol. 33, pp. 207–226. Buffalo, NY: Buffalo Society of Natural Sciences.

analysis of mite faunas at Glovers Pond by determining numbers of aquatic, semiaquatic and terrestrial taxa as percentages of mite populations in each equalized, 10-g sample through 3 m of core. Erickson (1987, 1988) pointed out that changes in percentage of aquatic mites up-core document changes in lake level within the basin over time preparing the foundation for biostratigraphic studies. In 1997, Erickson suggested that oribatid mite studies should be applied to questions of global climate change. Since that time, reports based on interesting applications of oribatid data have appeared with some regularity. Thus the taphonomic or preservational aspects as well as habitat preferences of oribatids have been emphasized in much of this work. These will provide focus for the discussions that follows.

Techniques and Methods

Obviously, during micropaleontological studies the value of the final product is controlled dramatically by the care and planning of the initial sampling. Sampling techniques for live oribatids include those employed by soil scientists who investigate soil faunas of all types. Soil or vegetative litter samples are placed on a screen in Tullgren or Berlese funnels and then under light (Krantz, 1978) which desiccates the mineral and organic matrix prompting soil organisms to move downward under positive geotaxis until they drop through a screen into the funnel for capture. Fossil mites are more challenging to sample and a variety of techniques may be employed to extract and study them (Krivolutsky and Druk, 1986; Krivolutsky et al., 1990).

Sediment Sampling for Oribatid Mites

Field sampling techniques as would be used for many Quaternary micropaleontological studies such as pollen, mollusks, or Coleoptera are applicable to oribatid sampling, but it should be pointed out that archaeological sampling and geological sampling requirements are often distinctly different (Coetzee and Brink, 2003; Erickson, 1988; Schelvis, 1997, 1998; Solhøy, 2001). Slot and trench samples of exposed soil or quarry faces are readily obtained using clean, durable, self-sealing plastic bags which are easily labeled. Of primary import is that stratigraphic interval and sample integrity are maintained from the outset. We wish to reinforce the notion that oribatid mites, like pollen, are ubiquitous in the modern environment. Samples that may be set aside in the field, or even in the lab, although monitored for contamination by physical factors, may yet be contaminated by mobile oribatids unless they are sealed immediately (Drouk, 1997). If one encounters an oribatid that still possesses complete tibial and tarsal leg segmentation, it should be regarded with suspicion during faunal analysis because fossil mites rarely retain distal leg appendages. Field samples and cores may be refrigerated to prevent molding, and desiccation should be prevented at all times. Sample sizes ranging from 10 to 1000 g can produce useful data depending on the objectives of the study.

Cored sediments present more rigid sampling situations in which core diameter controls sample availability and sample thickness determine the amount of sediment for use by the

paleoacarologist. Most cores are intended for multiple types of analysis. Sampling protocols (Figure 4) that extract organic matter by wet-sieving may permit physical and chemical studies of sediments passed through the sieves if distilled water is the washing agent. Oribatids, and most entomological remains, are best preserved in ethanol (70% or greater) and sediment samples may be immediately disbursed in this when gathered if other analyses are not needed. One half of a 1 cm-thick slice through a 4-cm-diameter core will provide sufficient numbers of oribatid mites for studies of lake level and climate change in gyttja, marl or peat substrates.

A cautionary note is appropriate. Fossil oribatids seem not uniformly distributed throughout some deposits. An unpublished experience of the senior author involves research done with Dr. Paul Karrow on the Don Brickyard Beds of Ontario,

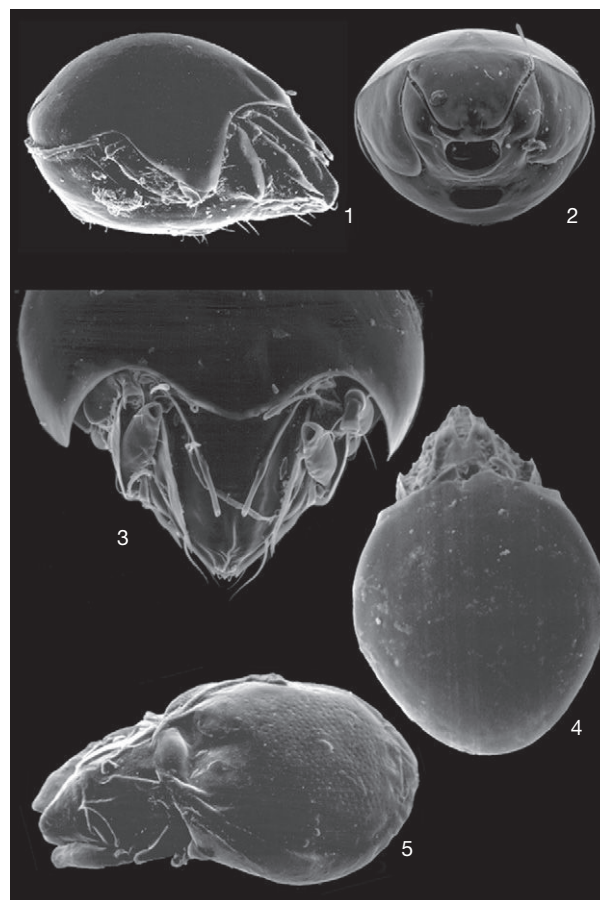


Figure 5 SEM images of typical fossil hydrophilic and aquatic mites from Erickson et al. (2003), illustrating the quality of preservation that is often present in lacustrine-authigenic sediment. 1, *Heterozetes aquaticus* lateral view; 2, anterior view of *Tegeribates* sp.; 3, dorsal view of anterior half of *H. aquaticus*; 4, *Naiazetes reevesi* seen in dorsal view; 5, oblique view of *Hydrozetes* sp. Reproduced from Erickson JM, Platt RB, Jr., and Jennings DH (2003) Holocene fossil oribatid mite biofacies as proxies of paleohabitat at the Hiscock Site, Byron, NY. In: Laub RS (ed.) *Proceedings of the Second Smith Symposium: The Hiscock Site: Late Pleistocene and Holocene Paleoeecology and Archaeology of Western New York State. Bulletin of the Buffalo Society of Natural Sciences*, vol. 37, pp. 176–189. Buffalo, NY: Buffalo Society of Natural Sciences.

Canada, that involved resampling of units which had produced a mite fauna some years earlier. The original work was interrupted and samples lost. Resampling of the same horizons was accomplished several years later by the same researcher who took the original samples. Processing of the resampled sediments by the same techniques produced much insect material but did not yield any oribatids. This lack of reproducibility is surprising because it is counter to a number of other experiences, yet it is appropriate cautionary information to provide herein.

Weight versus volume

Efforts to apply oribatid data to solution of geologic problems require that the data be internally comparable within a study. Hopefully, results will be comparable between studies as well. Cored sediments induce uncertainties that make data semi-quantitative at best, because of both deposition- and compaction-rate inequalities. We suggest that sampling by weight rather than by volume is appropriate for oribatid study. Furthermore, adjusting sample sizes to dry weight equivalents is

the most consistent means of assuring comparability. We have normalized samples to 10 g dry weight in several studies (Erickson, 1997; Pickard, 1993) either by sampling dried sediments (not recommended) or by determining desiccation factors of wet sediment and standardizing by using the wet/dry ratio.

Fossil mite extraction

Disbursed samples may be processed in several ways as suggested in Figure 4. From marl they may be separated by wet-sieving to retain material >0.05 mm after which they may be decanted with very weak HCl (maximum strength three parts per thousand) to remove the calcium carbonate. Alternatively, the samples may be stored in ethanol and carbonates removed from individual mites by immersing specimens of interest in a drop or two of the HCl on a depression slide. From gyttja they may be dispersed and sieved and the organics placed in ethanol. Peat should be mechanically disbursed in alcohol as thoroughly as possible. A shaking device may be used during the disbursal process (Figure 4) but this may further remove leg segments that have remained with the mite, nor are they

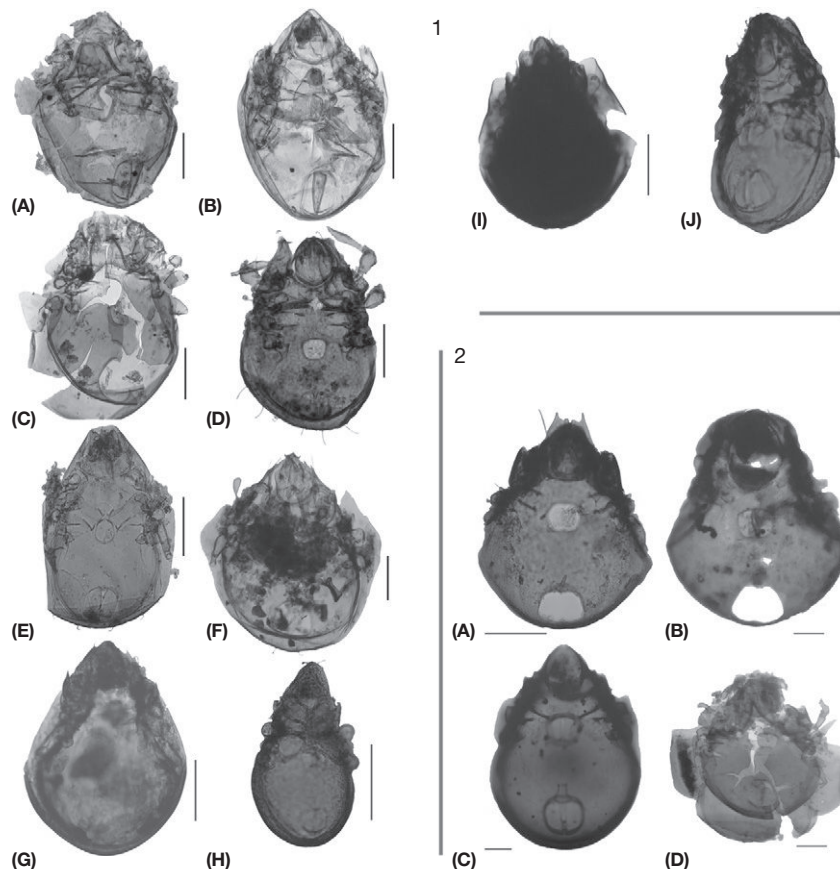


Figure 6 Fossil oribatids from the Hiscock Site, Byron, NY. 1, Ventral views (except 1G, I) of fossils from biofacies 1. (1A) *Limnozetes onondaga*, a *Sphagnum*-dweller; (1B) *Xylobates lophotrichus*; (1C) undetermined ceratozetoid; (1D) *Carabodes radiatus*; (1E) *Oripoda* sp.; (1F) *Naiazetes reevesi*; (1G) *Pelopsis* represented by *P. bifurcatus* specimen from biofacies 4; (1H) *Suctobelbella* sp.; (1I) *Peloptulus* sp.; and (1J) *Tectocephus velatus*, a cosmopolitan species found in most biofacies. 2, Oribatids occurring in biofacies 2. All except (2A) are ventral views: (2A) *Anachipteria signata*; (2B) undetermined *Galumna* sp.; (2C) *Podoribates pratensis*, a commonly occurring species in the Hiscock fauna; and (2D) *Achipteria* sp. crushed during mounting. Scale bars represent 100 µm. Modified from Erickson JM, Platt RB, Jr., and Jennings DH (2003) Holocene fossil oribatid mite biofacies as proxies of paleohabitat at the Hiscock Site, Byron, NY. In: Laub RS (ed.) *Proceedings of the Second Smith Symposium: The Hiscock Site: Late Pleistocene and Holocene Paleocology and Archaeology of Western New York State. Bulletin of the Buffalo Society of Natural Sciences*, vol. 37, pp. 176–189. Buffalo, NY: Buffalo Society of Natural Sciences.

recommended if taphonomic study (see section 'Taphonomic Analysis') is an objective. Under no circumstances should oribatids be cleaned by sonic cleaning methods because they are brittle and shatter easily with only brief exposure.

Separation of individual mites from the background sample is accomplished by hand-picking under magnification either with or without employing a floatation process (Figure 4). Kerosene (e.g., Coope, 1986; Kenward et al., 1980), heptane

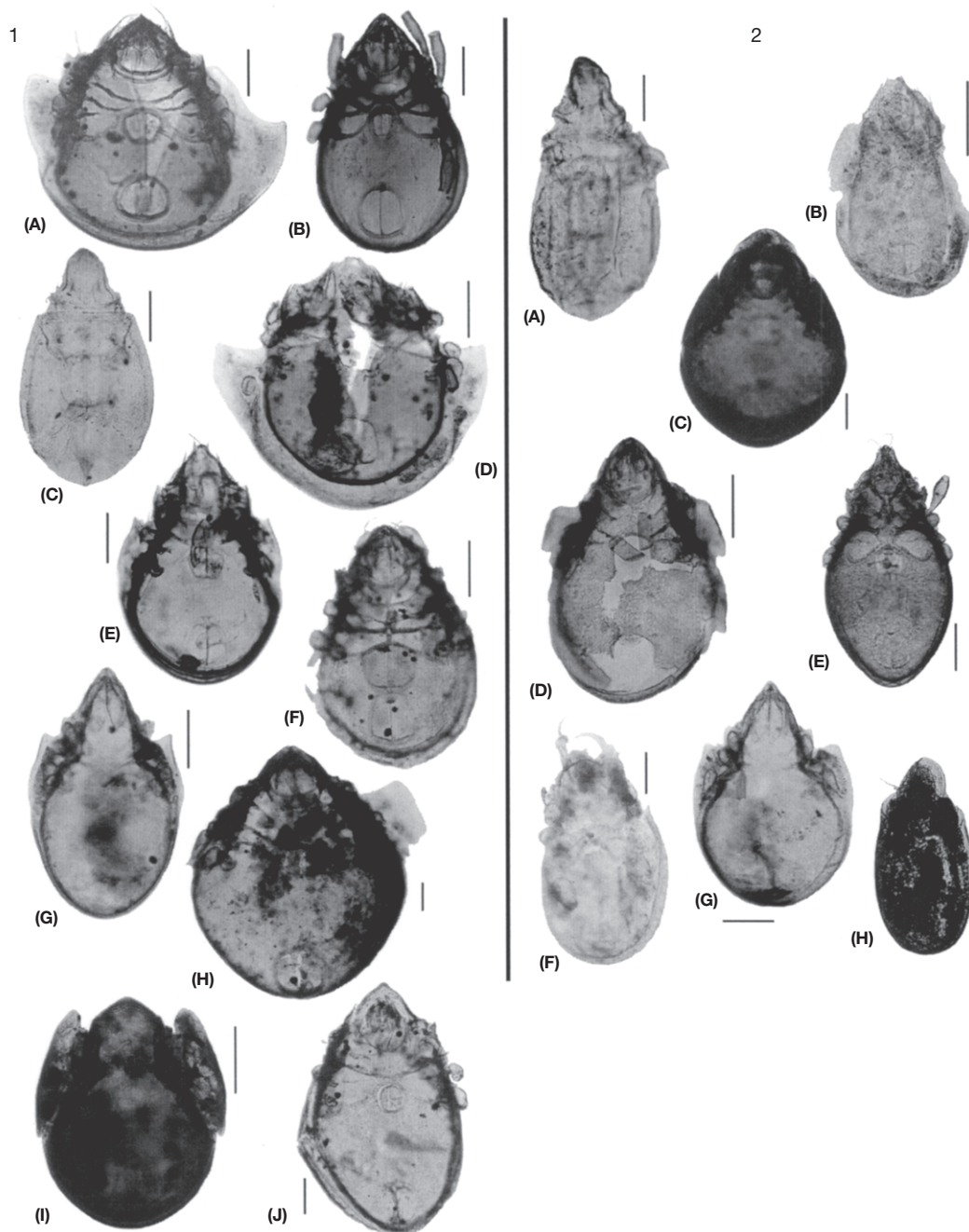


Figure 7 Hiscock Site biofacies 3 oribatid mites have a strong aquatic component. (1A) *Heterozetes aquaticus*; (1B) *Hydrozetes parisiensis*; (1C) *Trimalaconothrus maior*; (1D) *Zetomimus furcatus*; (1E) *Naiazetes reevesi*; (1F) *Tegeocranellus muscorum*, nonaquatic taxa; (1G) *Scheloribates* sp. A; (1H) *Podoribates pratensis*; (1I) *Pergalumna curva*; and (1J) *Phauloppia pilosa*. Ventral views are (1B, E, F, H, and J). Biofacies 4 includes an expansion of arboreal and epigeal elements of the fauna. (2A) *Trimalaconothrus foveolatus*, an aquatic mite; (2B) *Cultrobates quadricuspidatus*; (2C) *Tegoribates* sp.; (2D) *Rostrozetes ovulum*; (2E) *Allosuctobelba obtusa*; (2F) *Ommatocepheus clavatus*; (2G) *Scheloribates* sp. B; (2H) *Oripoda elongata*. Ventral views are (2B, D, E, and F). Scale bars represent 100 μm. Modified from Erickson JM, Platt RB, Jr., and Jennings DH (2003) Holocene fossil oribatid mite biofacies as proxies of paleohabitat at the Hiscock Site, Byron, NY. In: Laub RS (ed.) *Proceedings of the Second Smith Symposium: The Hiscock Site: Late Pleistocene and Holocene Paleoecology and Archaeology of Western New York State*. Bulletin of the Buffalo Society of Natural Sciences, vol. 37, pp. 176–189. Buffalo, NY: Buffalo Society of Natural Sciences.

(Walter et al., 1987), drying and rewetting (Drouk, 1997) saturated magnesium sulfate (Salt and Hollick, 1944) or saturated NaCl solution (Fain and Hart, 1986) may each be used to suspend oribatids above dispersed sediment by means of density differences, but each requires a distinct protocol of sample preparation not considered here. We prefer to use the saturated salt (NaCl) solution because it is effective and benign (Coetzee and Brink, 2003; Erickson, 1988). Samples stored in ethanol are spread thinly over the bottom of a petri dish and covered by a thin layer of ethanol. Using a dropper or pipette, introduce saturated salt solution to completely cover the sample. Stir sediment gently with a pick and place on microscope stage. Sediment-filled, alcohol-saturated specimens generally will float somewhat, others will reach the surface of the dense, salt solution. As oribatid fossils float to the surface, they may be removed by fine-tipped dropper to a storage vial of ethanol. We have verified the quality of NaCl separations by hand-picking the postflotation residues of random samples and have found that some few oribatids deeply entwined within

organic debris are not floated although they often reorient in the dish so that the mite is upward and visible for inspection. Final visual inspection of the sample is always advised. The drying-wetting method (Drouk, 1997) often introduces air bubbles into the body cavity that are difficult to remove. Alternatively, fossils may be hand-picked from dispersed sediment by dropper without first floating them. Reminders of salt-treated samples may be washed through a sieve (80 µm) and stored again in ethanol or discarded.

Identification

Utility of mite information is dependant upon the level to which taxa are identified. A phase contrast microscope with 1000× magnification is essential for species-level identification. Fossils may be mounted permanently on glass slides using Hoyers or de Faures medium (Krantz, 1978), cured for 24 h in a warm oven and sealed by ringing with glyptol after which they may be stored.

Table 1 Oribatid mite biofacies based on fossil occurrences in excavation pits G7SW, H5SW (=G5NW), H3SW, and I3NE at the Hiscock Site, Byron, NY. Large samples from the top of the Pleistocene Cobble and Fibrous Gravelly Clay Layers of H3SW yielded no mite fossils. Samples are arranged lithostratigraphically, and taxa are grouped by habitat categories as described in text. Biofacies, shaded horizontally for ease of recognition, suggest the following conditions: Basal zone corresponding to Pleistocene clastic sediments barren of mite fossils; biofacies 1, cool or mild, moist, fluctuating water table (=WT); 2, dry, hot? or cold?, very low WT; 3, mild, very wet, very high WT; 4, similar to present but more moist, with fluctuating WT and vernal pond conditions, becoming drier toward the top where an unsampled biofacies may exist. Position of the discontinuous, early Holocene, Gelatinous Woody Layer is also shaded to highlight its stratigraphic uniqueness

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	<i>Tectocepheus velutus</i>	<i>Opella cf. O. nova</i>	<i>Anachipteria signata</i>	Undetermined ceratocetids	<i>Galumoid</i> species	<i>Tegoribates</i> sp.	<i>Pergalumna curva</i>	<i>Podoribates pratensis</i>	<i>Rostozetes ovulum</i>	<i>Tegocarellus muscorum</i>	<i>Nalaezites reevesi</i>	<i>Limnozetes orondaga</i>	<i>Heterozetes aquaticus</i>	<i>Trimalacnothrus major</i>	<i>Trimalacnothrus foveolatus</i>	<i>Zetomimus furcatus</i>	<i>Zetomimus</i> sp.	<i>Hydrozetes parisiensis</i>		<i>Oripoda cf. O. elongata</i>	<i>Phialooppia pilosa</i>	<i>Onmatopopeus clavatus</i>	<i>enopodid</i> species	<i>Protoribates</i> sp.	<i>Alopirobella obtusa</i>	<i>Scheloribates</i> sp.	<i>Cultrobrates quadricuspidatus</i>	<i>scheleoribatid</i> species	<i>Achipteria</i> sp.	<i>Pelophilus</i> sp.	<i>Carabodes radiatus</i>	<i>Xylobates lophotrichus</i>	<i>Pelopsis bifurcatus</i>	<i>Pelopsis</i> sp.	<i>Suctobelbella</i> spp.																																																																																																																																																																																																																																																																																																																																																																																																																																																																																																																																																																																																																																																																																																																																																																																																																																																																																																																																																																																																																																																																																																																																																				
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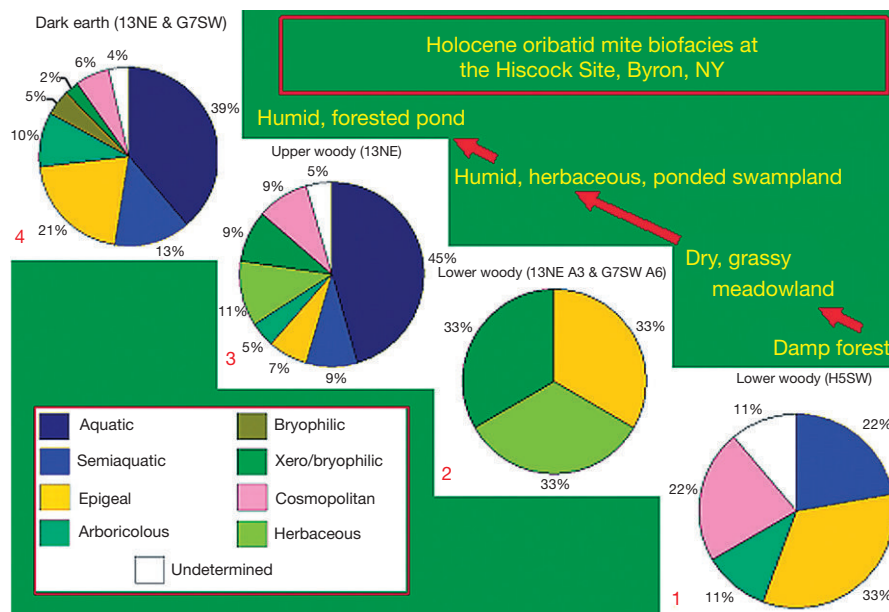


Figure 8 Habitat associations of fossil oribatids from biofacies 1 through 4 at the Hiscock Site, Byron, NY. Moisture and vegetation conditions are interpreted from mite associations arranged in ascending stratigraphic order.

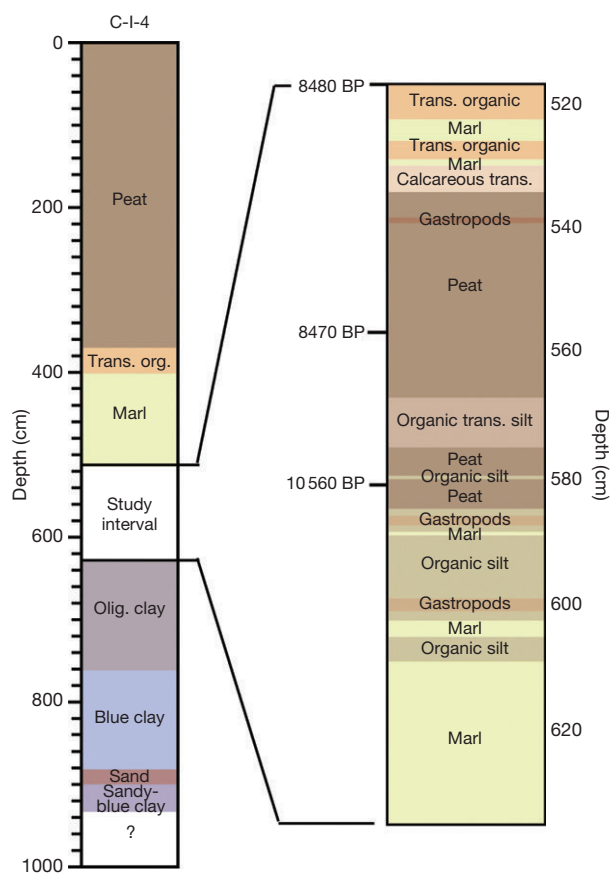


Figure 9 Lithologies and study interval from core C-I-4 at Glovers Pond, NJ. Interval from 630 to 520 cm depth was studied at 2 cm sample resolution for oribatid mite populations. ¹⁴C ages are uncalibrated years BP (Erickson and Solod, 2007).

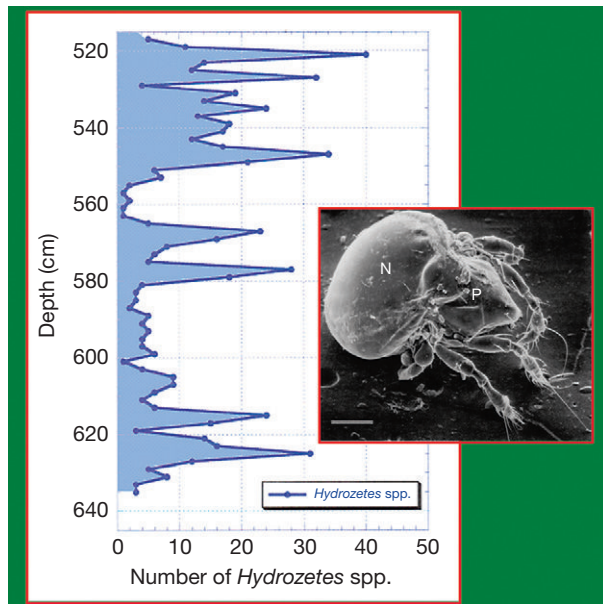


Figure 10 Fossil numbers of species of the aquatic oribatid genus *Hydrozetes* in core C-I-4 at Glovers Pond, NJ. Notogaster (N) and prodorsum (P) are indicated on *Hydrozetes* dorsum.

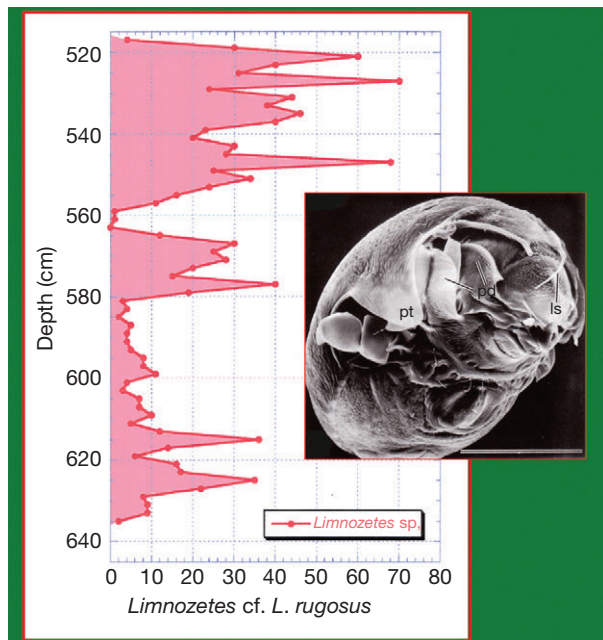


Figure 11 Fossil numbers of the *Sphagnum* mite *Limnozetes* cf. *L. rugosus* shown in inset. Letters designate pteromorph (pt), pedotectum (pd), and left lamellar seta (ls).

Many oribatid workers prefer to examine specimens using the Grandjean method by suspending them in lactic acid in a depression slide half covered by a moveable cover slip under which the specimen is maneuvered with a fine needle. The mite may be rotated for inspection of all necessary features after which it is returned to the storage vial of ethanol. This method works best for cleared specimens of live material. Fossils are

often not preserved well enough to justify manipulation and the attendant risk of damage in their brittle state. Practicing the technique with recent material is helpful before working with fossils. A thorough discussion of sorting and study techniques can be found in the summary by Solhøy (2001) or in other synoptic works (Drouk, 1997; Schelvis, 1997).

Scanning electron microscopy (SEM) can be an important tool for identification of fossil mites (Figure 5). It is most useful for satisfying questions raised during light microscopy. Sediment-filled specimens and those that are badly deformed may yet be identified using SEM. The only caveat is that SEM requires a permanent mount of the specimen and therefore archiving of the SEM stub along with other voucher materials is necessary.

Identification is aided by keys available at several taxonomic levels. The generic key by Norton (1990) or that by Balogh and Balogh (1992) permit initial identification. Workers should then refer to the extensive oribatid primary literature, summarized by Schatz (2004), containing keys to species. In Canada, many important systematic and ecological studies by Behan-Pelletier are focused on hydrophilic species, such as *Limnozetes*, the Family Zetomimidae (Behan-Pelletier, 1989; Behan-Pelletier and Eamer, 2003) and peatland oribatid faunas.

Sectors of Application

Applications for fossil oribatid mites generally parallel those for other microbiota in geological and archaeological research. The uniqueness of the group as a tool for Quaternary studies emerges from its unique life properties and the unique conditions of preservation which are a response to those life habits. Biostratigraphic, biofacies, faunal assemblage, paleoclimate and taphonomic studies, all of which potentially rest on the importance of initial taxonomic study – the skill being lost with the passing generation – each has import to the Quaternary scientist because when these organisms occur in the fossil record they are: (1) essentially of local derivation (see discussions in Erickson, 1988 and Drouk, 1997); (2) reflective of nearby soil and vegetation conditions; (3) preserved in numbers that often reflect population sizes; (4) preserved in taphonomic conditions that reflect local geologic setting; (5) are locally and regionally distributed according to climate factors of moisture and temperature; and (6) readily recognized after the initial identifications are established. Recent investigations of fossil oribatid mites serve as examples for mite applications to each of these sectors.

Biostratigraphic Applications

Many archaeological studies of oribatid fossils involve analysis of surface samples or of spot samples associated with human or animal fossils or artifacts (Foddai and Minelli, 1994; Schelvis, 1990, 1992, 1997). Studies are now demonstrating that close sampling for oribatids through longer stratigraphic intervals (Bennike, 2000; Erickson, 1987, 1988, 1997; Erickson and Solod, 2002; Luoto, 2009; Solhøy and Solhøy, 2000) can provide an unusual wealth of information of several sorts important to studies of postglacial history and climate change. We discuss some examples of these here.

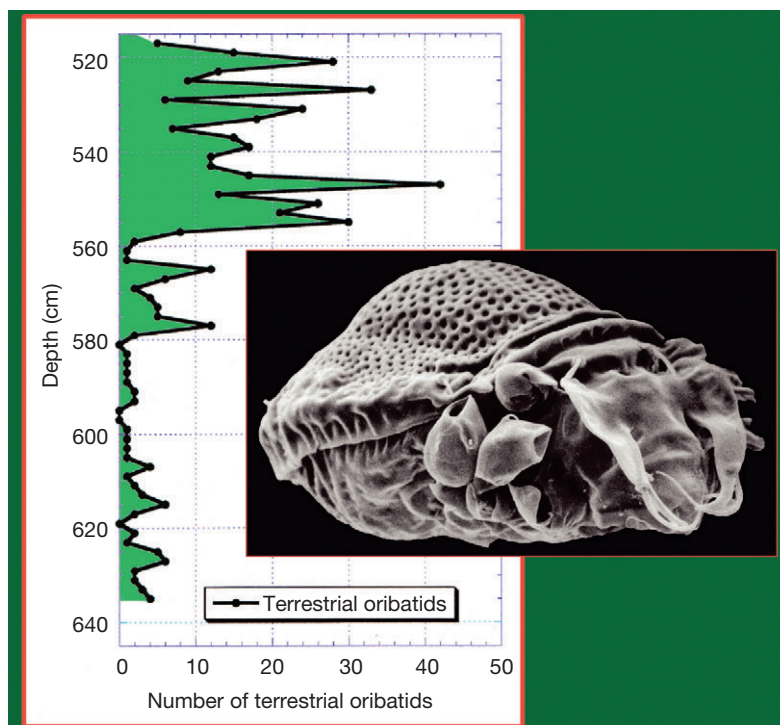


Figure 12 Numbers of terrestrial oribatid fossils in the study interval at Glovers Pond, NJ. Inset is *Ommatocepheus clavatus*, an inhabitant of lichen grown on rocks and tree bark.

Biozonation by species acme zones

In the study by Schelvis and Van Geel (1989) noted above, the first attempt to finely sample a stratigraphic section and to relate oribatid content to previously understood biostratigraphic parameters such as pollen and macrophyte zonation was made. It demonstrated that oribatid populations through a 2.5 m section of lake and bog sediment spanning roughly 2000 years responded to changing climate, and climate-influenced habitat alteration, in patterns that were very similar to those found in pollen spectra alone. Mite abundance data coincided remarkably well with pollen zone and subzone boundaries. A close examination of occurrence data would suggest that more information is available in this stratigraphic record than the newness of the approach allowed Schelvis and Van Geel to recognize. In their work, populations of the aquatic species *Hydrozetes lacustris* demonstrate two additional biostratigraphic subzones suggestive of those related to climate change in the Preboreal recorded by *Hydrozetes* species elsewhere (Erickson and Solod, 2002, 2007; Solhøy and Solhøy, 2000). In these studies, increases in populations of the aquatic mites reflect increases in water temperature, or depth, and availability of nutrients in the aquatic system. The acme of these populations defines the peak of such conditions in the sediment record. Records of very low, or no, occurrences of *Hydrozetes* spp. suggest cold climates with reduced nutrient cycling in lakes and bogs, or prolonged dry conditions (Erickson, 1987).

It is noteworthy that these same interpretations are recognized independently using different species in a 7000-year record from Signy Island in the Antarctic where two species

of oribatid mite are the largest land organisms on the island (Hodgson and Convey, 2005). *Alaskozetes antarcticus* and *Halozetes belgicae*, basically terrestrial taxa, have partitioned the lowland terrains from the tidal flats to 200 m elevation where they are both found in lake catchments in differing numbers. Hodgson and Convey (2005) report peak occurrences (acmes) of each species in similar positions in Heywood and Sombre Lakes approximately 1600–1700 cal years BP. This important study demonstrates utility of oribatid acme zones as a means of correlation in Quaternary sediments between local basins, but the climate change interpretation is of central interest.

Biofacies analysis

Biofacies analysis by oribatid distributions in five separate 1-m² quadrants exposing a 1.5-m section of peat at the Hiscock Site in New York, USA, was accomplished by taking horizontal slot samples 1.5 cm thick × 10 cm long × 5 cm deep in the wall of excavated quadrants with the goals of defining biostratigraphic zones and testing their relationships to the physical stratigraphy of the site (Erickson et al., 2003). Biofacies were based on the presence and absence of the species seen in Figures 6 and 7. These were next assigned to eight habitat-related assemblages including cosmopolitan, xerophilic/bryophilic, bryophilic, herbaceous, semiaquatic, aquatic, arboricolous, and epigeal groupings loosely based upon preferred habitat information related to the more than 35 species determined. Occurrence data are displayed stratigraphically in Table 1.

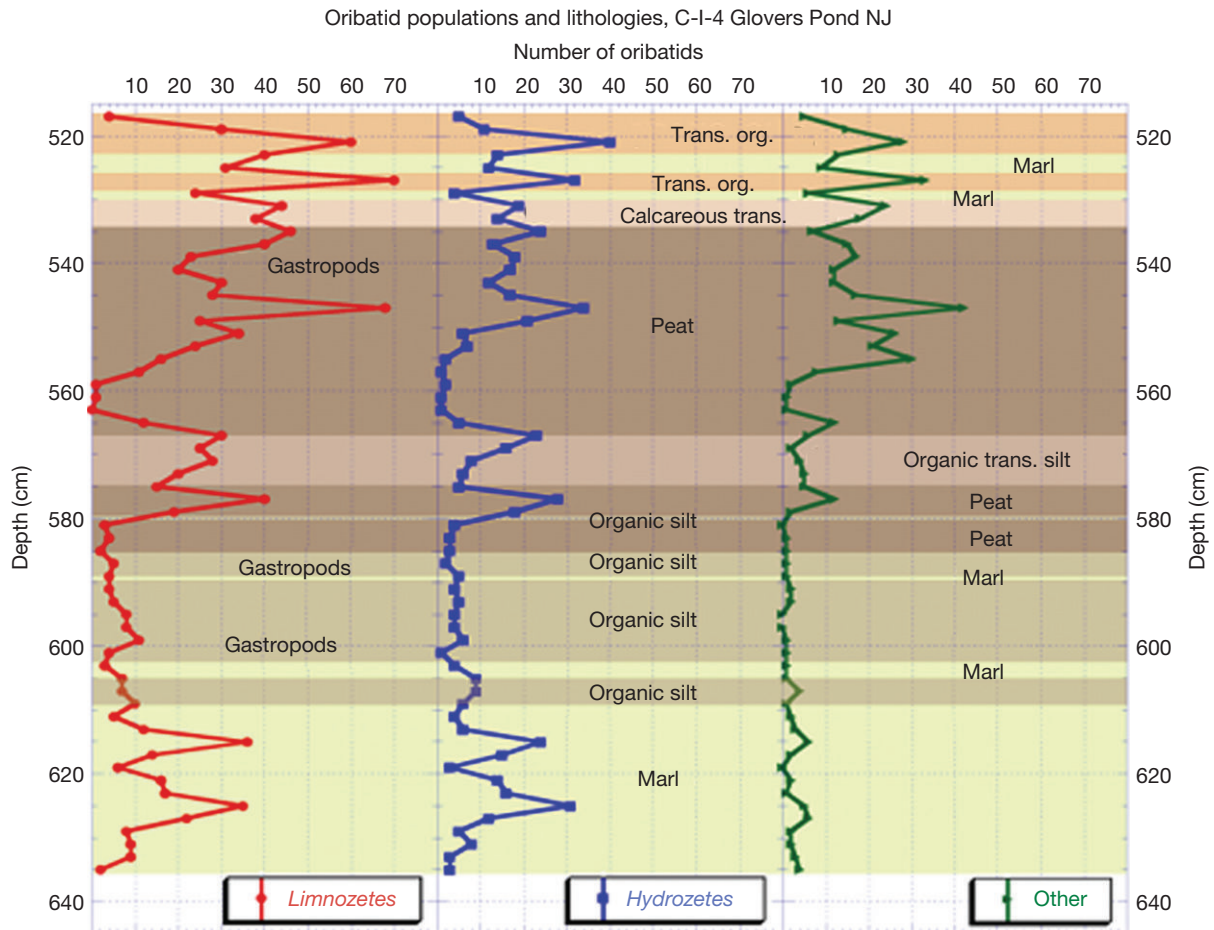


Figure 13 Fossil population data for *Limnozetes rugosus*, *Hydrozetes* spp. and terrestrial taxa (undifferentiated) displayed against stratigraphic and lithologic information from Glovers Pond core C-I-4. Data for Glovers Pond core C-I-4 with depth in centimeters after Solod (2001). Reproduced from Erickson JM and Solod AM (2002) Recognition of the Younger Dryas interval by quantitative biozonation of fossil oribatid mites at Glovers Pond, NJ. *XI International Congress of Acarology, Program and Abstract Book*, pp. 175–176; Erickson JM and Solod AM (2007) Recognition of postglacial cold intervals by quantitative biozonation of fossil oribatid mites. In: Morales-Malacara JB, Behan-Pelletier V, Ueckermann E, et al. (eds.) *Acarology XI: Proceedings of the International Congress*, pp. 9–16. Mexico: Instituto de Biología, UNAM; Facultad de Ciencias, UNAM; Sociedad Latinoamericana de Acarología, 2006.

Results of the biofacies analysis led to recognition of four mite-bearing biozones and one zone devoid of mites, and it was shown that biofacies boundaries were not coincidental with sediment contacts (Erickson et al., 2003). When the habitat groups of each biofacies are analyzed as percentages of the total mite count in each biofacies and plotted, they produce a visual means for recognizing dominant elements of habitat. The concept is similar to the isoalent groups of Knülle (1957) as employed by Schelvis (1990, 1998) although our group assignments are more subjective. Figure 8 relates the groupings sequentially and includes interpretations of habitat inferred from the mites.

The biofacies approach is useful when the sedimentary section is homogeneous, particularly if it is all peat, so there is little direct interpretation to be made from changes in sediment type. Oribatid assemblages that have been used stratigraphically to define biofacies demonstrate variations in depositional environment that are otherwise masked within

sedimentary sections (Erickson et al., 2003). Species assemblages subdivided by habitat preference yield further information about environmental change (Table 1; Figure 8) and thus paleoclimate.

Paleoclimate Proxies

Virtually all types of fossil oribatid studies offer information about paleoclimate. Many interpretations are second or third order inferences that fall out of assemblages of fossil species (Figure 8) and their habitat preference data rather than originating as direct proxy applications. Oribatid studies designed to provide direct first order proxies of temperature and moisture are becoming more common although the results they produce will continue to be semiquantitative or qualitative until physiologic limits for the modern fauna are better known (Behan-Pelletier and Bissett, 1994; Behan-Pelletier and Eamer, 2007). Multiproxy studies of Holocene deposits are providing

increasingly more refined paleoclimate and climate change information (Larsen et al., 2006; Sidorchuk, 2004). We will describe some examples below.

Proxies of temperature and moisture – lake and bog deposits

Close sampling of a 2-m section of core from Glovers Pond (Figure 9) known to contain the Younger Dryas cold interval was designed to use counts of the aquatic mites *Hydrozetes* spp. (Figure 10), the *Sphagnum* mite *Limnozetes* cf. *rugosus* (Figure 11), and terrestrial species (Figure 12) to define biozones that were climate-controlled (Erickson and Solod, 2007). Productivity of the lake and lake shore environments in postglacial Glovers Pond was temperature-driven and oribatid populations responded to warmer temperatures by increasing their numbers independent of sediment types as seen in Figure 13. Subsequent interpretation of biozonation was based upon synchronous changes in mite numbers as proxies of climate (Figure 14). C^{14} ages bracket the study

interval and temperature fluctuations were verified by O^{18} isotope analyses (Quevedo et al., 2002).

Some of the fossil oribatids that frequently occurred in Glovers Pond biozones include those in Figure 15. Declines in productivity included oribatid populations that are not aquatic, or semiaquatic, suggesting the data reflect climate changes large enough to influence all habitats, as opposed to short-duration drought or wet episodes. Oribatid biozones are interpreted to indicate established postglacial climate intervals from the Allerød through the Preboreal Oscillation into the Holocene warming (Figure 14). A sudden decline at 619 cm depth is believed to correlate with the Gurnensee/Killarney cold event recorded in only the best of stratigraphic records.

We believe that there is a great deal more information contained in these data (Figure 14) an example of which is our observation that within the cold interval of the Younger Dryas (Zone B) there is a cold, damp interval (612–581 cm

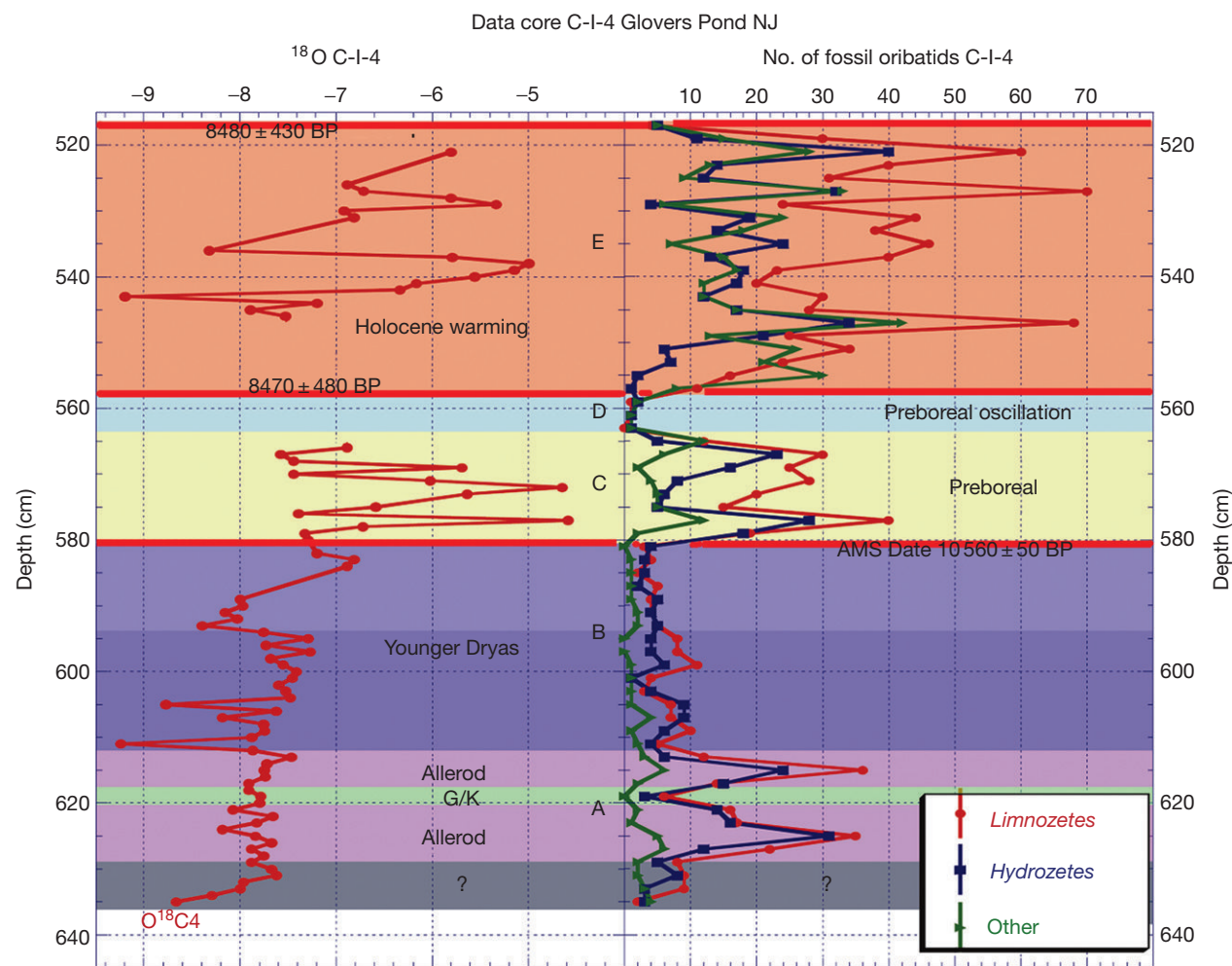


Figure 14 Interpretation of paleoclimate intervals based on distributions of oribatid mites zoned by population structure. Oxygen-isotope data from C-I-4 provide independent temperature proxy. The top of the Younger Dryas (Zone B) may be unconformable with the Preboreal. Modified from Erickson JM and Solod AM (2007). Recognition of postglacial cold intervals by quantitative biozonation of fossil oribatid mites. In: Morales-Malacara JB, Behan-Pelletier V, Ueckermann E, et al. (eds.) *Acarology XI: Proceedings of the International Congress*, pp. 9–16. Mexico: Instituto de Biología, UNAM; Facultad de Ciencias, UNAM; Sociedad Latinoamericana de Acarología, 2006.

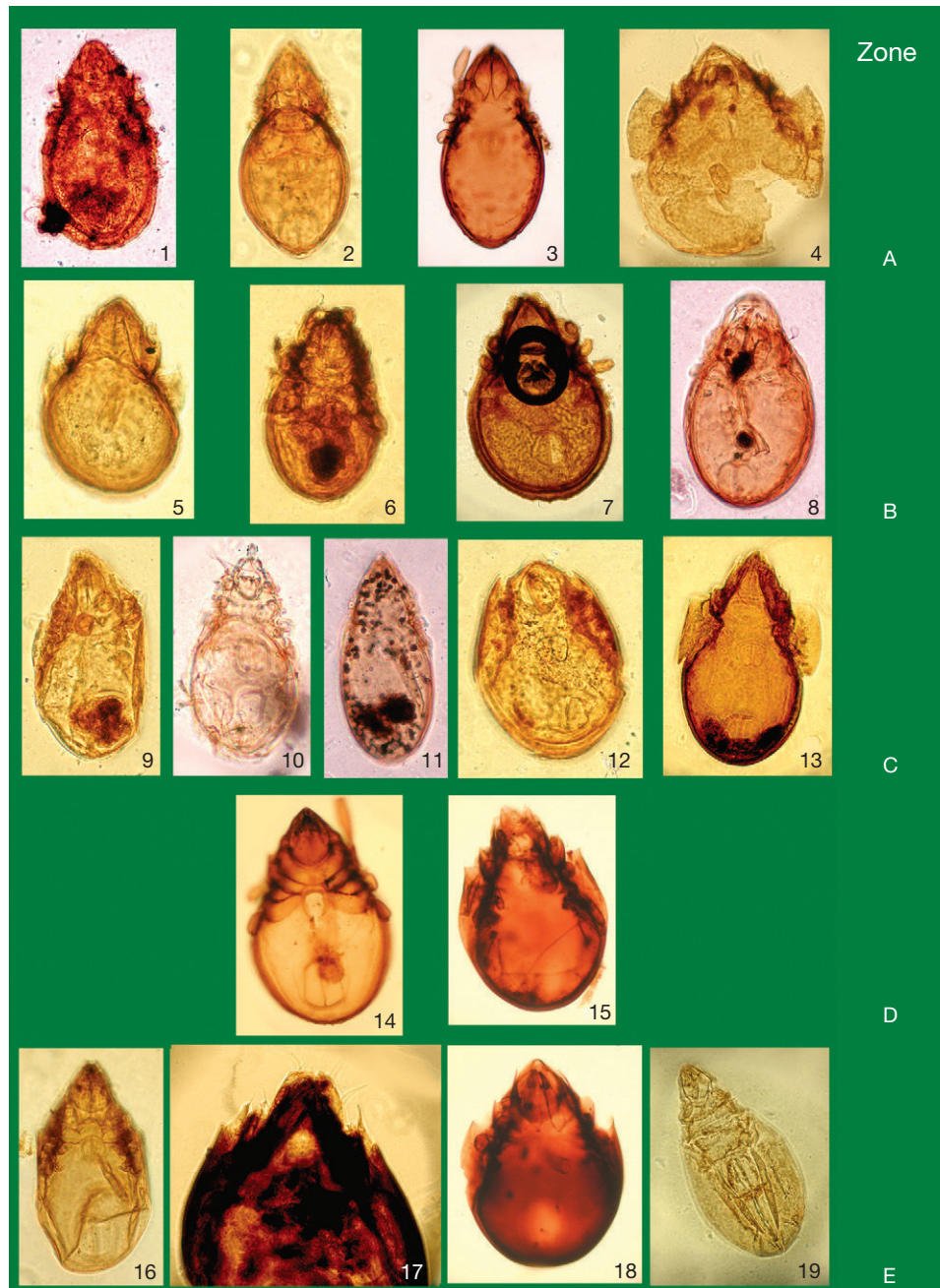


Figure 15 Fossil oribatid mites from biozones A (oldest) to E (youngest) at Glovers Pond, NJ. 1, *Scapheremaeus* sp.; 2, *Suctobelbella* sp.; 3, *Phauloppia* sp.; 4, *Heterozetes aquaticus*; 5, *Limnozetes* sp.; 6, *Tectocephus* sp.; 7, *Cepheus* sp.; 8, oripodid undet.; 9, *Pirnodus* sp.; 10, undet. belbid; 11, *Oripoda elongata*; 12, undet. galumnid; 13, *Cultrobates quadricuspispidatus*; 14, *Hydrozetes* sp.; 15, *Pelopsis* sp.; 16, *Xylobates* sp.; 17, *Achipteria* sp.; 18, *Heterozetes aquaticus*; and 19, *Trimalaconothrus maior*.

depth) of fluctuating temperatures followed by a stable, cold, dry interval (Erickson and Solod, 2007). The cold damp interval itself may be partitioned by a dry episode at approximately 601–603 cm depth. A second example is found at 557 cm when recovery from the Preboreal Oscillation cold event began, first by an increase in terrestrial taxa and then *Sphagnum*-dwelling *Limnozetes* populations, followed in 6 cm by an increase in populations of *Hydrozetes* spp. We interpret these

data to mean that the Preboreal Oscillation was dry and cold, and as climate warmed, moisture levels also increased, but slowly. Terrestrial productivity improved as groundwater levels rose and *Sphagnum* production returned in bogs, but lake level did not return to normal for a considerable time. This recovery contrasts with several earlier episodes in which aquatic and semiaquatic species recover in consort. Fossil specimens illustrating examples from each biozone are included in Figure 15.

Proxies of temperature and moisture – cave deposits

An unusual setting for oribatid study was described by Polyak et al. (2001). Oribatids were discovered within laminations of two stalagmites in Hidden Cave, a New Mexico cavern in the Guadalupe Mountains. To determine similarity with modern conditions a study of the living fauna of the cave environs was conducted. It was then compared with the 11 oribatid fossil taxa of the stalagmites which was a mismatch. Wetter and cooler suggested by presence of *Epidamaeus* (*Epidamaeus*) sp. implied that pine forests now existing at higher elevations were likely present at the cave elevation during formation of the stalagmite intervals examined, 3135–887 years BP.

Taphonomic Analysis

Study of the causes of death and fossilization of organisms, biostratinomy, are often focused on the events of transport and interment of fossils that leave a 'taphonomic imprint' on the fossil. When mites are considered fossil, the study of how they became so is legitimized as it were. Every fossil has a taphonomic history and each is potentially unique. Erickson (1988) examined hundreds of fossil mites during the course of picking them from core samples, and he recognized that preservation characteristics varied stratigraphically. He presented a description of the variation in the form of a Taphonomic Scale (Figure 16) of oribatid preservation condition (Erickson, 1988). He recognized that because the starting condition of the fossil is known, the endpoint it arrived at can be used to infer some of the geologic processes that acted on the corpse on its way to the deposition site. Drouk (1997) also discussed the concept.

Recognition of unconformities

Solod (2001) included study of taphonomic condition of mites through a study section of core from Glovers Pond. Mites were classed and ranked from 1 to 5 (Erickson, 1988) then indexed by taxon and interval using a simple formula. For example, the number of *Hydrozetes* spp. of each class multiplied by the class rank and added for all classes present in the sample is divided by the total number of fossil mites in the sample thus giving a value between 0 and 5 which is graphed by taxon and stratigraphic interval (Figure 16). The value serves as a Preservation Index that can be related to the scales of mite condition and used for stratigraphic comparison. Results are not fully interpreted, but Erickson has noted that there is tendency for fully-compressed or sediment-filled specimens to occur in restricted units near or at interfaces overlain by offlap-onlap depositional cycles. Furthermore, sediments filling the class 2 mites are often organic matter although the specimen may be originating from marl when exhumed. Class 3 mites are thought to have been compressed by compaction of sediments resulting from lowered water tables in basin-margin peatlands.

In one such interval, 569–557 cm, there is some evidence the oribatids have been reworked from an eroding basin-margin peat, or soil horizon, exposed by prolonged low water conditions and redeposited both in upper sediments of the declining-water phase and in basal sediments of the rising-water phase centered on the Preboreal Oscillation (Figures 14 and 16). The Preservation Index alerts us to the condition

of potential unconformity with as much sensitivity as do the numbers of mites to temperature and moisture-level changes (Figure 14). Application of this Taphonomic Scale and the Preservation Index to entire stratigraphic sections may make it possible, or easier, to assess unconformable relationships concealed within lacustrine cores. In many instances these are difficult to recognize by sedimentologic examination alone. Greater study of mite taphonomy is in order.

Archaeological Research Applications

Some of the earliest uses of fossil mite data were made by archaeologists in order to evaluate environmental conditions associated with prehistoric settlements that may have been influenced by climate change or environmental crises. As described above, Denford (1978), Schelvis (1987, 1989, 1990, 1992, 1997, 1998), and Schelvis and Van Geel (1989) pioneered application of oribatid and predatory mites for understanding habitat conditions related to archaeological sites. Fossil oribatids are underutilized in this discipline.

Socio-economic patterns

In 2007, a group led by Chepstow-Lusty reasoned that socio-economic conditions at a site in the Andes were linked to the size of domestic animal populations locally. Oribatid populations were suggested to be enlarged in response to larger amounts of dung present in grazing yards from whence it was also washed into a nearby lake. Mites from the yard were also washed into the lake and deposited. Lake waters likely suffered eutrophication at the same time as there were increasing populations of aquatic oribatids. Chepstow-Lusty et al. (2007) found three taxa, *Galumna* sp., *Trichoribates* sp., and *Hydrozetes* sp. present in sufficient numbers in cored lake sediments to allow quantitative evaluation of fossil populations. These data demonstrated three episodes of enhanced animal productivity interspersed with two episodes of reduced domestic productivity over the past 1200 years. The authors called for additional study of mite taphonomy and local community structure and composition.

Spring deposits

As awareness of fossil oribatids expands, so too will their applications. The well-studied Quaternary deposits at Florisbad, South Africa, include disturbed paleosols as old as the last interglacial, eolian silts and spring deposits as young as Middle Stone Age (MSA). They have been studied for numerous paleontological and archaeological factors. To this knowledge Coetzee and Brink (2003) have recently added data on fossil oribatid and mesostigmatid mites, some as old as 125 000 years BP. It was noted that most of the species found were able to reproduce parthenogenically which suggest that they were aquatic species. Thelytokous mites are currently being investigated by R. A. Norton in an effort to understand ecological advantages of this, and other, oribatid reproductive modes (Norton and Palmer, 1991). Coetzee and Brink's oribatid data permitted more accurate interpretation of organic sediments of the MSA horizon which contained *Papillacarus capucinus*, suggesting they are more likely related to organics produced in the spring than to eolian sedimentation (Coetzee and Brink, 2003).

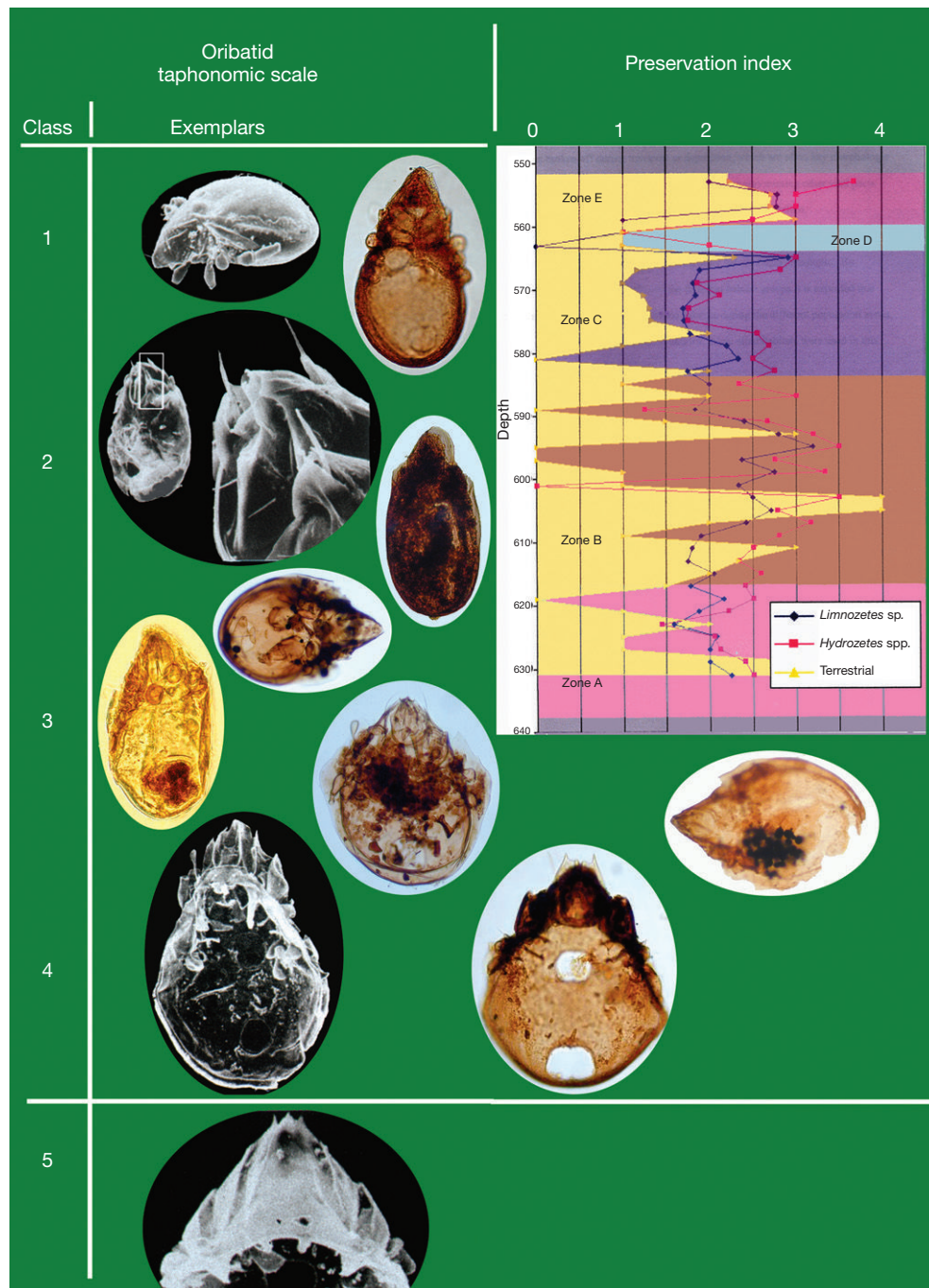


Figure 16 Oribatid taphonomic scale (Erickson, 1988) includes five recurring types of preservation; classes 1–4 are gradational to some extent with class 1 being inflated and pristine, class 2 somewhat inflated with setae preserved but missing some plates and mouthparts thus leading to filling body cavity with sediment, class 3 flattened, complete skeletons, class 4 complete prodorsal and ventral shields but notogaster, genital, and anal plates missing, class 5 prodorsal region only suggesting predation of hysterosoma. Inset is graphic display of Preservation Index data from Glovers Pond core C-I-4 after Solod (2001). Data for Glovers Pond core C-I-4 with depth in centimeters after Solod (2001).

Acknowledgments

This article draws heavily on the works of acarologists and paleoacarologists throughout the world, and we are most appreciative of their efforts although space limits prohibit more complete analysis of all contributions. Assistance over many

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See also: Lake Level Studies: North America. **Paleoclimate Reconstruction:** Historical Climatology. **Quaternary Stratigraphy:** Continental Biostratigraphy. **Radiocarbon Dating:** AMS Radiocarbon Dating.

References

- Arillo A, Subias LS, and Shtanchaeva U (2008) A new fossil oribatid mite, *Ommatocepheus nortoni* sp. nov. (Acariformes, Oribatida, Cepheidae), from a new outcrop of Lower Cretaceous Alava amber (northern Spain). *Systematic and Applied Acarology* 13: 252–255.
- Balogh J and Balogh P (1992) *The Oribatid Genera of the World, 2 vols.*, p. 263 and 375. Budapest: Hungarian Natural History Museum Press.
- Behan-Pelletier VM (1989) *Limnozetes* (Acari: Oribatida: Limnozetestidae) of Northeastern North America. *The Canadian Entomologist* 21: 453–506.
- Behan-Pelletier VM (2000) Ceratozetidae (Acari: Oribatida) of arboreal habitats. *The Canadian Entomologist* 132: 153–182.
- Behan-Pelletier VM and Bissett B (1994) Oribatida of Canadian peatlands. In: Finnmore AT and Marshall SA (eds.) *Terrestrial Arthropods of Peatlands, with Particular Reference to Canada. Memoirs of the Entomological Society of Canada*, pp. 73–88. Canada: Entomological Society of Canada.
- Behan-Pelletier VM and Eamer B (2003) Zetomimidae (Acari: Oribatida) of North America. In: Smith IM (ed.) *An Acarological Tribute to David R. Cook (From Yankee Springs to Wheeny Creek)*, p. 331. West Bloomfield: Indira Publishing House.
- Behan-Pelletier VM and Eamer B (2007) Aquatic Oribatida: Adaptations, constraints, distribution and ecology. In: Morales-Malacara JB, Behan-Pelletier V, and Ueckermann E, et al. (eds.) *Acarology XI: Proceedings of the International Congress*. México: Instituto de Biología, UNAM; Facultad de Ciencias, UNAM; Sociedad Latinoamericana de Acarología.
- Bennike O (2000) Palaeoecological studies of Holocene lake sediments from west Greenland. *Palaeogeography, Palaeoclimatology, Palaeoecology* 155: 285–304.
- Cannon RCS and Block W (1988) Cold tolerance of microarthropods. *Biological Reviews* 63: 23–77.
- Chepstow-Lusty AJ, Frogley MR, Bauer BS, et al. (2007) Evaluating socio-economic change in the Andes using oribatid mite abundances as indicators of domestic animal densities. *Journal of Archaeological Science* 34: 1178–1186.
- Coetzee L and Brink JS (2003) Fossil oribatid mites (Acari, Oribatida) from the Florisbad Quaternary deposits, South Africa. *Quaternary Research* 59: 246–254.
- Coope GR (1967) The value of Quaternary insect faunas in the interpretation of ancient ecology. In: Cushing EJ and Wright HE Jr. (eds.) *Quaternary Paleoeecology*, pp. 359–380. New Haven, CT: Yale University Press.
- Coope GR (1986) Coleoptera analysis. In: Berglund BE (ed.) *Handbook of Holocene Palaeoecology and Palaeohydrology*, pp. 703–713. Chichester: Wiley.
- de la Riva-Caballero AH, Birks AE, Bjune H Birks, and Solhøy T (2010) Oribatid mite assemblages across the tree-line in western Norway and their representation in lake sediments. *Journal of Paleolimnology* 44: 361–374.
- Denford S (1978) Mites and their potential use in archaeology. In: Brothwell DR, Thomas KD, and Clutton-Brock J (eds.) *Research Problems in Zooarchaeology*, vol. 3, pp. 77–83. London: Institute of Archaeology. Occasional Publication.
- Drouk AY (1997) Aracological analysis: Problems of paleoecological reconstructions. In: Edwards ME, Sher AV, and Guthrie RD (eds.) *Terrestrial Paleoenvironmental Studies on Beringia*, pp. 91–97. Fairbanks, Alaska: The Alaska Quaternary Center, University of Alaska.
- Erickson JM (1986) The promise for fossil oribatid mites in paleoecological analyses of Quaternary lake and bog deposits: Preservation quality and data extraction (Abstract). *Smith Symposium: Late Pleistocene and Early Holocene Paleoeecology and Archeology of the Eastern Great Lakes Region*. Buffalo, New York: Buffalo Museum of Science. Program and Abstracts, p. 18.
- Erickson JM (1987) Response of fossil aquatic mite populations (Arthropoda: Oribatida) to changing hydrogeologic conditions in Quaternary lacustrine sediments (abstract). *100th Annual Meeting*, vol. 19, no. 7, p. 655. Boulder, CO: Geological Society of America. Abstracts with Programs.
- Erickson JM (1988) Fossil oribatid mites as tools for Quaternary paleoecologists: Preservation quality, quantities, and taphonomy. In: Laub RS, Miller NG, and Steadman DW (eds.) *Late Pleistocene and Early Holocene Paleoeecology and Archeology of the Eastern Great Lakes Region. Bulletin of the Buffalo Society of Natural Sciences*, vol. 33, pp. 207–226. Buffalo, NY: Buffalo Society of Natural Sciences.
- Erickson JM (1997) Can paleoacarology contribute to global climate change research? In: Mitchell R, Horn DJ, Needham GR, and Welbourn WC (eds.) *Proceedings of the 9th International Congress of Acarology*, pp. 533–537. Ohio Biological Survey.
- Erickson JM, Platt RB Jr., and Jennings DH (2003) Holocene fossil oribatid mite biofacies as proxies of paleohabitat at the Hiscock Site, Byron, NY. In: Laub RS (ed.) *The Hiscock Site: Late Pleistocene and Holocene Paleoeecology and Archaeology of Western New York State. Proceedings of the Second Smith Symposium. Bulletin of the Buffalo Society of Natural Sciences*, vol. 37, pp. 176–189. Buffalo, NY: Buffalo Society of Natural Sciences.
- Erickson JM and Solod AM (2002) Recognition of the Younger Dryas Interval by quantitative biozonation of fossil oribatid mites at Glovers Pond, N.J. *XI International Congress of Acarology Program and Abstract Book*, pp. 175–176.
- Erickson JM and Solod AM (2007) Recognition of postglacial cold intervals by quantitative biozonation of fossil oribatid mites. In: Morales-Malacara JB, Behan-Pelletier V, and Ueckermann E, et al. (eds.) *Acarology XI: Proceedings of the International Congress*, pp. 9–16. México: Instituto de Biología, UNAM; Facultad de Ciencias, UNAM; Sociedad Latinoamericana de Acarología.
- Erickson JM, Waters DL, and Metzger RA (1985) Fossil soil mites (Arthropoda: Oribatida): A new tool for Quaternary bio- and ecostratigraphers (abstract). *Canadian Paleontology and Biostratigraphy Seminar, Quebec City, Programme with Abstracts*, pp. 9–10.
- Fain A and Hart BJ (1986) A new, simple technique for extraction of mites using the difference in density between ethanol and saturated NaCl (preliminary note). *Acarologia* 27: 125.
- Foddai D and Minelli A (1994) Fossil arthropods from a full-glacial site in northeastern Italy. *Quaternary Research* 41: 336–342.
- Frey DG (1964) Remains of animals in Quaternary lake and bog sediments and their interpretation. *Archiv für Hydrobiologie, Ergebnisse Der Limnologie* 2: 1–116.
- Hodgson DA and Convey P (2005) A 7000-year record of oribatid mite communities on a maritane-Antarctic island: Responses to climate change. *Arctic, Antarctic, and Alpine Research* 37: 239–245.
- Kenward HK, Hall AR, and Jones AKG (1980) A tested set of techniques for the extraction of plant and animal macro-fossils from waterlogged archaeological deposits. *Science and Archaeology* 22: 3–15.
- Knülle W (1957) Die Verteilung der Acari: Oribatei im Boden. *Zeitschrift Für Morphologie Und Ökologie der Tiere* 46: 397–432.
- Krantz GW (1978) *A Manual of Acarology*, 2nd edn., p. 509. Corvallis, OR: Oregon State University Book Stores.
- Krivolutsky DA and Druk AJ (1986) Fossil oribatid mites. *Annual Review of Entomology* 31: 533–545.
- Krivolutsky DA, Druk AJ, Eitminaviciute IS, Laskova LM, and Karppinen E (1990) *Fossil Oribatid Mites*, p. 110. Vilnius, Lithuania: Mokslas Publishers.
- Larsen JA, Bjune AE, and de la Riva-Caballero AH (2006) Holocene environmental and climate history of Trettetjorn, a low alpine lake in Western Norway, based on subfossil pollen, diatoms, oribatid mites and plant macrofossils. *Arctic, Antarctic, and Alpine Research* 38(4): 571–583.
- Luoto TP (2009) An assessment of lentic ceratopogonids, ephemeropterans, trichopterans, and oribatid mites as indicators of past environmental change in Finland. *Annales Zoologici Fennici* 46: 256–270.
- Metzger RA (1985) *Numerical Distribution of Mites Relative to Depositional Environment for Postglacial Sediments Collected from Glovers Pond in Northwestern New Jersey*, p. 123. BS Thesis. St. Lawrence University.
- Newell IM (1945) *Hydrozetes* Berlese (Acari, Oribatidea): The occurrence of the genus in North America and the phenomenon of levitation. *Transactions of the Connecticut Academy of Arts and Sciences* 36: 253–275.
- Nordenskiöld E (1901) Zur Kenntnis der Oribatidenfauna Finnlands. *Acta Societatis pro Fauna et Flora Fennica* 21: 1–34.
- Norton RA (1985) Aspects of the biology and systematics of soil arachnids, particularly saprophagous and mycophagous mites. *Quaestiones Entomologicae* 21: 523–541.
- Norton RA (1990) Acarina: Oribatida. In: Dindal DL (ed.) *Soil Biology Guide*, pp. 779–803. New York: Wiley.
- Norton RA (1994) Evolutionary aspects of oribatid mite life histories and consequences for the origin of the Astigmata. In: Houck M (ed.) *Mites: Ecological and Evolutionary Life-History Patterns*, pp. 99–135. New York: Chapman and Hall.
- Norton RA and Behan-Pelletier VM (1991) Calcium carbonate and calcium oxalate as cuticular hardening agents in oribatid mites (Acari: Oribatida). *Canadian Journal of Zoology* 69: 1504–1511.
- Norton RA, Kethley JB, Johnston DE, and O'Connor BM (1993) Phylogenetic perspectives on genetic systems and reproductive modes of mites. In: Wrensch DL and Ebbert MA (eds.) *Evolution and Diversity of Sex Ratio in Insects and Mites*, pp. 8–99. New York: Chapman and Hall.
- Norton RA and Palmer SC (1991) The distribution, mechanisms and evolutionary significance of parthenogenesis in oribatid mites. In: Schuster R and Murphy PW (eds.) *The Acari – Reproduction, Development and Life-History Strategies*, pp. 107–136. London: Chapman and Hall.

- Norton RA, Williams DD, Hogg ID, and Palmer SC (1988) Biology of the oribatid mite *Mucronothrus nasalis* (Acari: Oribatida: Trhypochthoniidae) from a small coldwater springbrook in Eastern Canada. *Canadian Journal of Zoology* 66: 622–629.
- Pickard RA (1993) *Systematic Paleontology of Fossil Mites from Lower Portion of Core C-I-3, Glovers Pond, New Jersey*, p. 112. Unpublished BS Thesis. St. Lawrence University.
- Polyak VJ, Cokendolpher JC, Norton RA, and Asmerom Y (2001) Wetter and cooler Late Holocene climate in the southwestern United States from mites preserved in stalagmites. *Geology* 29: 643–646.
- Quevedo HL, Carpenter SJ, and Erickson JM (2002) An exceptional high-resolution stable isotope record of the last deglaciation from NW New Jersey. *Geological Society of America Programs with Abstracts* 34(6): 352.
- Salt G and Hollick FSJ (1944) Studies of wireworm populations. I. A census of wireworms in pasture. *Annals of Applied Biology* 31: 52–64.
- Schatz H (2004) *World Literature of the Oribatida*. Innsbruck, Austria.
- Schelvis J (1987) Some aspects of research on mites (Acari) in archaeology samples. *Paleohistoria* 29: 211–218.
- Schelvis J (1989) Mites (Acari) from the Late Neolithic well at Kolhorn (The Netherlands). *Paleohistoria* 31: 165–171.
- Schelvis J (1990) The reconstruction of local environments on the basis of remains of oribatid mites (Acari; Oribatida). *Journal of Archaeological Science* 17: 559–571.
- Schelvis J (1992) Mites and mammoths. *Proceedings of the Section Experimental and Applied Entomology of the Netherlands Entomological Society* 3: 140–141.
- Schelvis J (1997) Mites in the background. Use and origin of remains of mites (Acari) in Quaternary deposits. *Studies in Quaternary Entomology – An Inordinate Fondness for Insects. Quaternary Proceedings*, vol. 5, pp. 233–236. Chichester: Wiley.
- Schelvis J (1998) Mites. In: Preece RC and Bridgland DR (eds.) *Late Quaternary Environmental Change in North-west Europe*, pp. 234–241. London: Chapman and Hall.
- Schelvis J and Van Geel B (1989) A palaeoecological study of the mites (Acari) from a Lateglacial deposit at Usselo (The Netherlands). *Boreas* 18: 237–243.
- Sidorchuk EA (2004) Subfossil oribatid mites as the bioindicators of profound environmental changes during the Holocene. *Doklady Biological Sciences* 396(1–6): 236–239.
- Solhøy T (2001) Oribatid mites. In: Smol JP, Birks HJB, and Last WM (eds.) *Tracking Environmental Change Using Lake Sediments*, vol. 4, *Zoological Indicators*, pp. 81–104. Dordrecht: Kluwer Academic.
- Solhøy IW and Solhøy T (2000) The fossil oribatid mite fauna (Acari: Oribatida) in late-glacial and early-Holocene sediments in Kråkenes Lake, western Norway. *Journal of Paleolimnology* 23: 35–47.
- Solod AM (2001) *Fossil Moss Mites (Arthropoda: Oribatida) as Indicators of the Younger Dryas Cooling Episode, Glovers Pond, Northwestern, New Jersey*, p. 61. BS Thesis. St. Lawrence University.
- Strenze K (1952) Untersuchungen über die Tiergemeinschaften des Bodens: Die Oribatiden und ihre Synusien in den Böden Norddeutschlands. *Zoologica* 104: 1–173.
- Walter DE, Kethley J, and Moore JC (1987) A heptane flotation method for recovering microarthropods from semiarid soils, with comparison to the Merchant–Crossley high-gradient extraction method and estimates of microarthropod biomass. *Pedobiologia* 30: 221–232.
- Walter DE and Proctor HC (1999) *Mites: Ecology, Evolution and Behaviour*. Wallingford: CABI Publishing.
- Waters DL (1979) *Oribatid Mites Collected from Postglacial Sediments in Northern New Jersey*, p. 59. Unpublished BS Thesis. St. Lawrence University.

P

Packrat Middens *see* **Plant Macrofossil Methods and Studies:** Rodent Middens

Paleoanthropology *see* **Archaeological Records:** Overview; 2.7 Myr–300 000 Years Ago in Africa; 2.7 Myr–300 000 Years Ago in Asia; 1.9 Myr–300 000 Years Ago in Europe; Global Expansion 300 000–8000 Years Ago, Africa; Global Expansion 300 000–8000 Years Ago, Asia; Global Expansion 300 000–8000 Years Ago, Australia; Global Expansion 300 000–8000 Years Ago, Americas; Human Evolution in the Quaternary; Neanderthal Demise; Postglacial Adaptations

PALEOBOTANY

Contents

Overview of Terrestrial Pollen Data

Ancient Plant DNA

Charred Particle Analyses

Paleophytogeography

Silicon Isotopes in Diatoms

Overview of Terrestrial Pollen Data

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Some History

Pollen analysis or palynology lies at the interface of geology and biology but has close contacts with other disciplines such as archaeology. Pollen analysis has tended to develop in different directions in different countries, largely dependent on the interests of the influential pioneers in the field. The study of Quaternary vegetation history began in Scandinavia, at the end of the nineteenth century, primarily through the work of Blytt

and Sernander, botanists who examined peat bog stratigraphy and plant macrofossil content to develop a scheme of postglacial climate change. Their ideas were further developed by the Swedish geologist [von Post \(1916\)](#) with the technique of statistical pollen analysis that is the most widely used method for reconstructing Quaternary vegetation history today. Von Post attracted other scientists to the field, including the Swedish botanist Gunnar Erdtman, who through his international outlook and subsequent work on pollen morphology and

preparation technique deserves to be widely remembered. Erdtman was the first to publish widely on von Post's methods outside Scandinavia and paid an influential visit to Scotland, Ireland, and the Isle of Man in the mid-1920s. However, he was long overshadowed by von Post and was unable to obtain a research position in Sweden until late in his career. Consequently, Quaternary palynology in Sweden has tended to be carried out within the Earth sciences (Traverse, 1988).

In Denmark, the pioneer palynologists (Iversen, Jessen, and Troels-Smith) were based at the Geological Survey, National Museum, and Botanical Institute and helped develop the interdisciplinary nature of Quaternary science. Jessen brought palynology to Ireland. The links to botany and forest ecology were strong, with the Danish Geological Survey being a major force for forest conservation. The establishment of an experimental field station at Draved Forest, Jutland, in 1949 was a landmark. At the Geological Survey, Iversen created a department with a strong international reputation that was a formative influence on the later development of the subject in the United States, as both Cushing (Cushing and Wright, 1967) and Davis (1976) studied in Copenhagen. Iversen linked with the Norwegian botanist Faegri to produce a major textbook of pollen analysis in 1950, now in a fourth but aging edition (Faegri and Iversen, 1989). Birks (2005) has reviewed the interesting development of pollen analysis in Scandinavia.

Erdtman inspired, among others, Godwin to work on the East Anglian fens and submerged forests of coastal peat beds. Godwin (1934) published a critical assessment of the potential of von Post's methods and dominated the British paleoecological scene between 1940 and 1960, becoming Professor of Botany at Cambridge and founding a great British centre of botanical paleoecology. *The History of the British Flora*, a factual basis for phytogeography (Godwin, 1975), first published in 1956, is still a useful paleoecological text summarizing early pollen work in Britain. He emphasized the close links between pollen and plant macrofossil studies. Pollen studies developed somewhat independently in Central and Eastern Europe with Rudolph, Bertsch, and Firbas being important names among the pioneers. Auer was influential in both Finland and Canada and von Neijstadt in the Soviet Union. The contributions of many of these early scientists are reviewed by Lang (1994). Sears, Potzger, and Wilson were important founders of North American Quaternary palynology (Canright, 1995).

The schools set up by Firbas (1949) in Central Europe and Godwin in England collected enough data to enable the reconstruction of vegetation and environmental change on a far larger scale than had been hitherto attempted. Godwin's initiatives were developed at Cambridge by West, who exemplified the geological aspects while Birks and Birks (1980) developed the biological links. In Poland, Szafer, director of the Botanic Garden in Cracow, began a series of studies of the Quaternary flora of Poland as early as 1912. A fortuitous meeting with botanist Tansley from Britain, on an excursion to Poland and Czechoslovakia in 1928, led to an invitation to participate in the International Botanical Congress in Cambridge in 1930, where he contributed to the symposium on the distributional ecology of beech. His interests were very wide-ranging and included both pollen and macrofossils, and he first demonstrated the value of isopoll maps in studying the Holocene history of beech and spruce in Central–Eastern Europe (Szafer, 1935).

Recent Developments

Von Post (1916) and Godwin (1934) speculated about the ways in which the study of vegetation history could contribute to the biological sciences. Many of these speculations have subsequently become reality, and the results are outlined in the articles that follow this introduction. After the initial enthusiasm and excitement from the pioneer days of palynology, a significant development was the routine application of independent chronologies to Late Quaternary sedimentary sequences through the development of radiocarbon dating (Taylor, 2000). This development took place during the 1960s, in research groups with access to sufficient funding, and contributed to a generally increased level of precision in the reconstruction of vegetation. With independent dating for the Holocene, it was possible to make meaningful comparisons within and between regions and eventually build databases that made pollen data accessible to a broad range of scientists. Quaternary pollen analysis developed from being merely a method of geological correlation into a useful tool for exploring vegetation dynamics.

Davis (1976) and Huntley and Birks (1983) published continental scale maps showing the Holocene spreading of tree taxa from their glacial refugia to their present distribution. Their work illustrated the dynamic nature of tree distributions that have undergone continuous change. The subsequent work of the COHMAP (Cooperative Holocene Mapping Project) group (Wright et al., 1993) demonstrated that vegetation was continuously changing in response to climate forcing, bringing into focus the quasi-equilibrium situation between vegetation and climate. This line of research, which originated in a strictly geological approach, has led to a paradigm shift in biology. The concepts of climax vegetation and traditional succession were shown to be largely untenable under typical disturbance regimes with a constantly changing climate. Rather, the research suggested that plant species acted in an individualistic manner, leading to temporary assemblages of plants in vegetation communities that are nevertheless recognizable in the pollen records. This development in the subject is a good example of results from one discipline leading to constructive development in another – the best possible outcome of interdisciplinary research.

Increased awareness of the role of climate change gave a new impetus to palynological research, beginning in the 1980s. Pollen data provided evidence for the impacts of constant climate change and are used as one proxy among many for the reconstruction of past climate. A complicating issue is the undoubted impact of human activities on vegetation during recent millennia that is well recorded by pollen data, and considerable progress has been made in placing archaeological sites in a broader environmental setting. Identifying and assessing the relative importance of the varied drivers of long-term vegetational change is proving to be one of the current directions of palynological research.

Pollen diagrams have tended to increase in length and detail over the years, with more types identified and increased precision in all aspects of the work that has transformed the nature of the subject. In a landmark paper, Davis (1963) made an important attempt to relate pollen spectra in a more quantitative manner to its source vegetation through the development of R-values. Her work laid the foundation for later developments in pollen representation theory that are now culminating in

practical tools for the quantitative reconstruction of vegetation at various scales (Gaillard et al., 2008; Prentice, 1985; Sugita, 1993). Increasing quantitative precision has stimulated the application and development of new statistical analytical tools to pollen data that were summarized by Birks and Gordon (1985) and are reviewed in Numerical Analysis Methods.

The increasing quantitative and spatial precision in vegetation reconstruction has led to exciting new insights being generated in the research areas of evolution, ecology, and conservation biology (Bennett, 1997). These topics included plant migration, disturbance biology, population ecology, and the long-term evolution of genetic structure. The development of the Long-Term Ecological Research program has brought into focus the value of longer timescales in the understanding of the current status and direction of change of biological systems, which has increased the interest of biologists in palynological research (Willis et al., 2010).

Variable Scales: A Special Feature of Pollen Data

A unique feature of pollen analysis is the flexible way in which the method generates data at different spatial and temporal scales (Figure 1). Traditional pollen analysis utilizes pollen preserved in lake basins and peatland sites to study the effects of climate change and anthropogenic influence on vegetation. Increasing the temporal resolution with annually laminated sediments allows short-term events to be resolved and permits time series analyses to be made with ease. Choosing sites with longer sedimentary sequences from unglaciated regions yields the long-term records that cover more of the Pleistocene and allows comparison with marine and ice core records. The same analytical techniques can be applied at fine spatial scales to describe vegetation

succession after a local fire or record the effects of death of a single tree in a mor humus profile. The pollen analyst has a unique set of controls at the fingertips with which to manipulate temporal and spatial scale and study a wide range of processes.

Pollen analysis has been joined over the years by a range of other analytical methods using the same sediments to generate more complete environmental reconstruction. Some of these other methods are covered in this section, including phytoliths, charcoal, and ancient DNA. Pollen, analyzed together with plant macrofossils, beetle remains, diatoms, chironomids, and other biological, physical, and chemical analyses, comprises multiproxy data that are used to make reconstructions of climate and the biological and physical environments that are as complete as possible.

Key Insights from Pollen Analysis

The articles that follow document a range of insights into the nature of vegetation dynamics obtained from pollen data. A few of these are highlighted here.

Nonanalog plant communities are unique assemblages of plants that no longer grow together. Examples are the ash–ironwood (*Fraxinus–Ostrya*) communities of the midwestern United States during the Late Glacial and the rich deciduous forests with pine found in the mid-Holocene in northwest Europe. The American example is thought to reflect the unique combinations of climatic parameters occurring at that time, while the European example has been lost due to almost complete anthropogenic modification of natural forest communities.

Individualistic dynamics of tree distributions during the Holocene have been of major importance to the development of plant

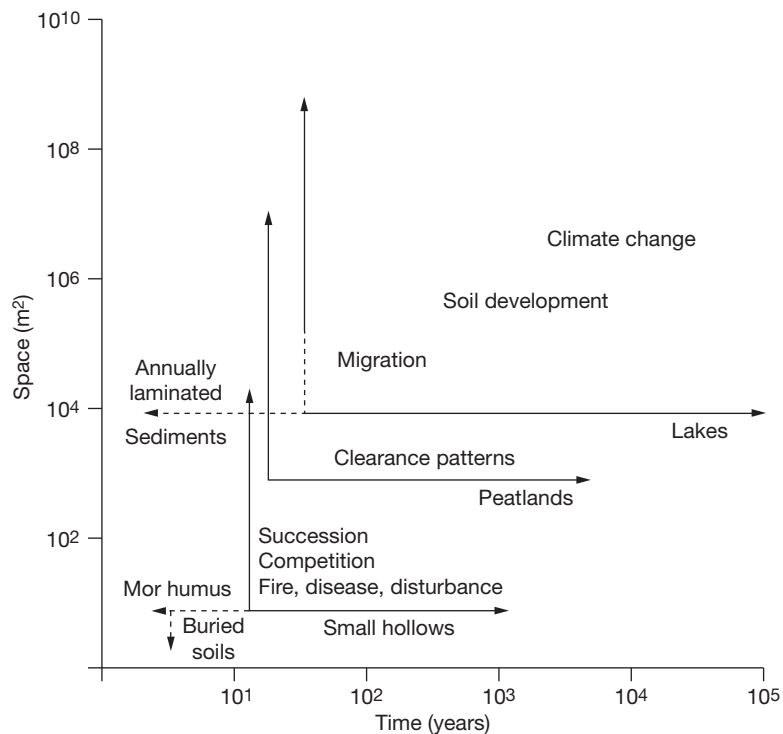


Figure 1 The spatial and temporal scales available to the pollen analyst.

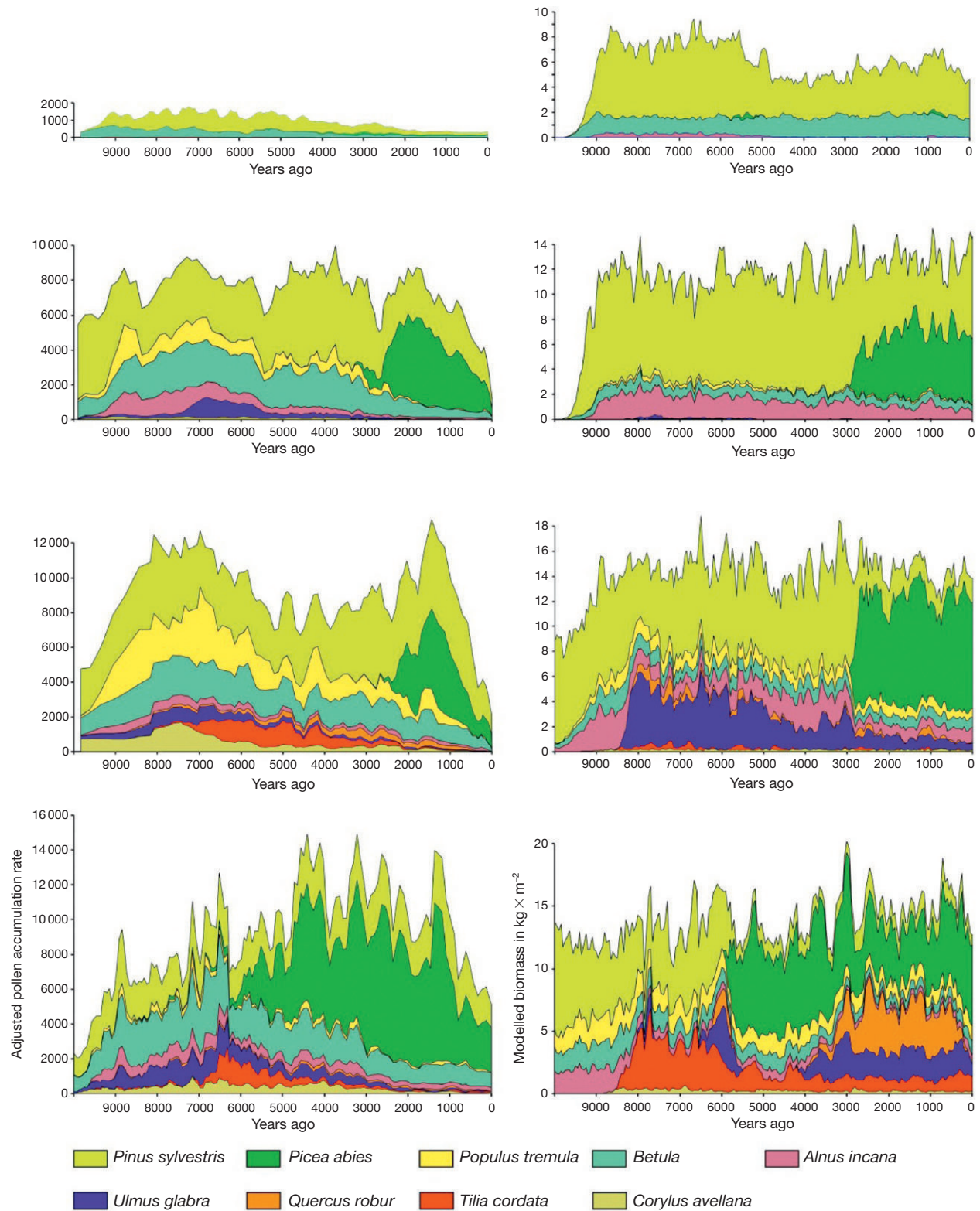


Figure 2 Converted pollen accumulation rates (left column) and modeled biomass (right column) for each species, at Tsuolbmajavri (top row, 68.69° N 22.08° E), Abborrtjärnen (second row, 63.88° N 14.45° E), Holtjärnen (third row, 60.65° N 14.92° E), Sweden, and Laihalampi, Finland (bottom row, 61.49° N 26.08° E; Miller et al., 2008).

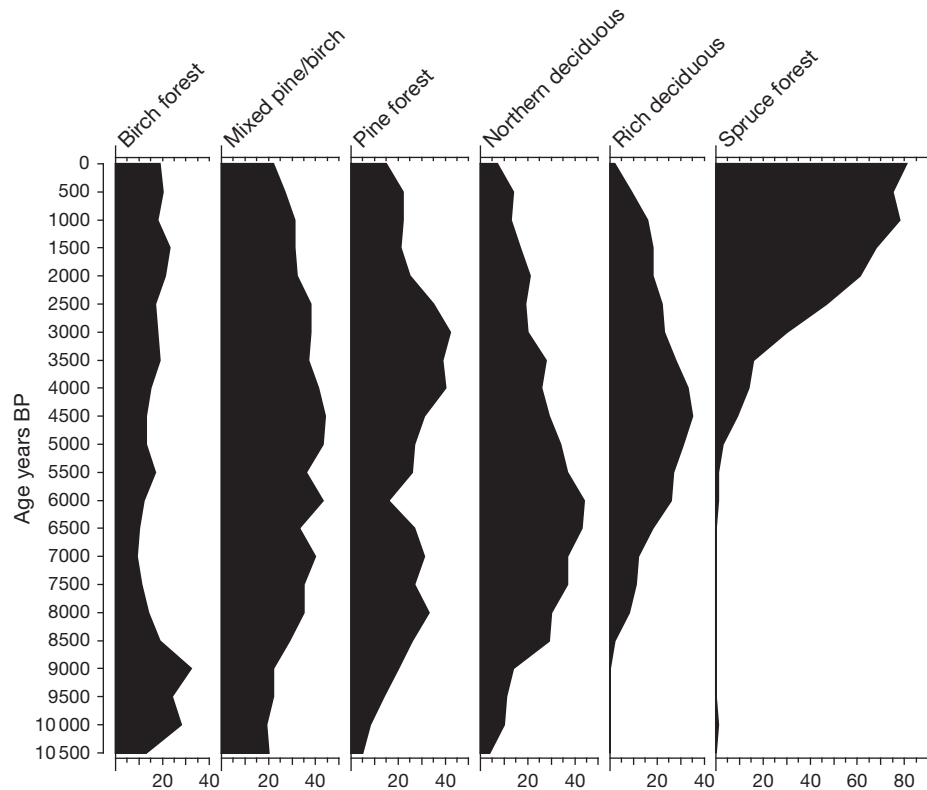


Figure 3 The changing area of Fennoscandian forest cover during the Holocene. Forest is divided into six categories. The diagram covers about 0.7% of global land area and is based on 308 individual pollen diagrams. The x-axis shows the area covered by each forest type, and one unit is approximately 7500 km².

ecological theory. Species have tended not to migrate in biomes or communities but rather in response to individualistic environmental forcing, although [Clark and McLachlan \(2003\)](#) show that intersite variance for Holocene tree abundances in North America tended to stabilize rapidly, showing that plant distributions cannot be explained by individualistic behavior alone but include a component of community dynamics.

Refugial theories have developed to help explain present patterns of genetic diversity and the regionally variable loss of species during periods of rapid climate change and climatic extremes. The links between paleoecology and evolutionary genetics are rapidly developing, with vigorous debate about the locations, nature, and abundance of these refugia. European refugia appear to have had different characteristics to those in North America with consequences for the long-term survival of species.

The quasi-equilibrium of vegetation with climate has been debated among palynologists for many years. Large international projects and syntheses have concluded that pollen data can in many situations be used as a proxy for climate data and that the concept of 'migrational lag' or delayed response of vegetation to climate change caused by dispersal limitation is either of short duration or not easily detected in the pollen record. European beech for example was slow to dominate in the European Holocene, but its dynamics are broadly consistent with its relationships to regional climate ([Bradshaw et al., 2010](#)).

The extent of human influence on vegetation is considerable but varies both temporally and spatially. In many parts of the world, human influence is now a more important driver

of vegetational change than climate, but this is most likely to be a relatively recent development. Restoration ecology is consequently an emerging topic that is influenced by the paleovegetational record ([Jackson and Hobbs, 2009](#)). Human influence does extend far back in time, primarily through its effect on fire regimes and populations of large browsing and grazing animals. Pollen analysis has always had a close link to environmental archaeology and has contributed to placing archaeological sites in a broader environmental context.

Invasions of new and exotic species are a major concern within nature conservation. Pollen analysis has documented the spread and invasion of species in the past and provides insight into the invasion process and population growth, and their long-term consequences. The spread of hemlock (*Tsuga canadensis*) into the midwestern United States showed an interesting balance between dynamics and stability at a fine spatial scale, while the spread of spruce (*Picea abies*) through Scandinavia was associated with competitive replacement of lime (*Tilia cordata*) and a recent uniform spreading rate of about 250 m year⁻¹ in Sweden ([Bradshaw and Lindbladh, 2005](#); [Seppä et al., 2009](#)).

Disturbance biology is another topic where pollen analysis has contributed to biological understanding and long-term ecological research. Periodic disturbance by fire is documented by paleoecological techniques, while chronic disturbance from animals can often be inferred. Fire has long been a major ecological factor in several ecosystems and has a complex relationship with human activity. There can be abundant charcoal remains in Quaternary sediment cores, and fire interacts with vegetation type to be a

major disturbance agency in the boreal region (Ohlson et al., 2011), although natural fire regimes are now rare in populated regions. The hemlock decline and elm decline are examples of the widespread impact of disease.

The interglacial–glacial cycle and its links to evolutionary biology is a topic where pollen analysis is altering the way scientists think (Bennett, 1997; Hu et al., 2009). The nature of the glacial–interglacial cycle and its biological consequences have been investigated for many years, and the relative importance of soil development, climate change and the biological properties of organisms have been intensively debated. Some interesting conclusions are emerging that have profound scientific implications. For example, it seems that most generation of genetic diversity predates the Quaternary, which is characterized by a general loss of diversity in, for example, woody plants (Willis and Niklas, 2004). Vegetation has usually tracked rapid climate change, but the response can be moderated by changing soil nutrients. History matters, and community structure is constrained and shaped by its past.

The Way Ahead

No one can be certain what the future will bring, but studies of the past dynamics of predictable systems do help us to build likely scenarios. One current development within pollen analysis is to use past reconstructions to test climate-driven simulations of vegetation. Models that can correctly simulate past vegetation are more likely to prove reliable in the future. There are a growing number of these validation exercises in progress, and they give pollen data an important role in forecasting possible consequences of future climate change. Data–model comparison has established that the northern distributional limits of temperate trees in Fennoscandia were controlled by millennial variations in temperatures during the Holocene (Figure 2; Miller et al., 2008).

The building of global databases of environmental data from the past and present is likely to continue. As pollen databases grow and improve in quality, there will be new analyses made on the basis of their content. Pollen data can be combined in innovative ways to represent vegetation cover from large areas. A composite forest type diagram has been drawn for the whole of Fennoscandia showing the area covered by six forest types throughout the Holocene (Figure 3) (Holmqvist et al., unpublished). Records of changing vegetation over large areas of the globe are useful tools in understanding global system dynamics and helping monitor the scale and consequences of human impact on terrestrial ecosystems.

Pollen analysis is moving from the era of primary data collection into the exciting phase of secondary data analysis. Pollen databases are building a new sense of community among palynologists and with pooled data of high quality, new insights into the nature of vegetation dynamics and its relationship to the dominant drivers of human activity and climate change await discovery.

See also: Phytoliths; Pollen Analysis, Principles; Pollen Records, Last Interglacial of Europe. **Paleobotany:** Ancient Plant DNA; Charred Particle Analyses. **Pollen Methods and Studies:** Archaeological Applications; Changing Plant Distributions and Abundances; Databases and Their Application; POLLSCAPE Model: Simulation Approach for Pollen Representation of Vegetation and Land Cover; Stand-Scale Palynology; Surface Samples and Trapping; Use of Pollen as Climate Proxies. **Pollen Records, Late Pleistocene:** Africa; Australasia; Northern Asia. **Pollen Records, Postglacial:**

Northeastern North America; Northern Europe; South America; Southeastern North America; Southern Europe; Southwestern North America. **Radiocarbon Dating:** Charcoal; Conventional Method.

References

- Bennett KD (1997) *Ecology and Evolution: The Pace of Life*. Cambridge: Cambridge University Press.
- Birks HJB (2005) Fifty years of Quaternary pollen analysis in Fennoscandia 1954–2004. *Grana* 44: 1–22.
- Birks HJB and Birks HH (1980) *Quaternary Palaeoecology*. London: Arnold.
- Birks HJB and Gordon AD (1985) *Numerical Methods in Quaternary Pollen Analysis*. London: Academic Press.
- Bradshaw RHW, Kito N, and Giesecke T (2010) Factors influencing the Holocene history of *Fagus*. *Forest Ecology and Management* 259: 2204–2212.
- Bradshaw RHW and Lindbladh M (2005) Regional spread and stand-scale establishment of *Fagus sylvatica* and *Picea abies* in Scandinavia. *Ecology* 86: 1679–1686.
- Canright JE (1995) A brief history of some major contributors to the development of palynology in the United States. *Palynos* 18: 1–7.
- Clark JS and McLachlan JS (2003) Stability of forest biodiversity. *Nature* 423: 635–638.
- Cushing EJ and Wright HE Jr. (1967) *Quaternary Paleoeecology*. New Haven, CT: Yale University Press.
- Davis MB (1963) On the theory of pollen analysis. *American Journal of Science* 261: 897–912.
- Davis MB (1976) Pleistocene biogeography of temperate deciduous forests. *Geoscience and Man* 13: 13–26.
- Fægri K and Iversen J (1989) In: Fægri K, Kaland PE, and Krzywinski K (eds.) *Textbook of Pollen Analysis*, 3rd edn. Chichester: Wiley.
- Firbas F (1949) *Spät- und nacheiszeitliche waldgeschichte Mitteleuropas nördlich der Alpen. Erster band: Allgemeine Waldgeschichte*. Jena: Fischer.
- Gaillard M-J, Sugita S, Bunting JM, et al. (2008) The use of modelling and simulation approach in reconstructing past landscapes from fossil pollen data: A review and results from the POLLANDCAL network. *Vegetation History and Archaeobotany* 17: 419–443.
- Godwin H (1934) Pollen analysis: An outline of the problems and potentialities of the method. *New Phytologist* 33: 278–305.
- Godwin H (1975) *History of the British Flora*. Cambridge: Cambridge University Press.
- Hu FS, Hampe A, and Petit RJ (2009) Paleoeecology meets genetics: Deciphering past vegetational dynamics. *Frontiers in Ecology and the Environment* 7: 371–379.
- Huntley B and Birks HJB (1983) *An Atlas of Past and Present Pollen Maps for Europe 0–13000 Years Ago*. Cambridge: Cambridge University Press.
- Jackson ST and Hobbs RJ (2009) Ecological restoration in the light of ecological history. *Science* 325: 567–569.
- Lang G (1994) *Quartäre vegetationsgeschichte Europas*. Jena: Gustav Fischer.
- Miller PA, Giesecke T, Hickler T, et al. (2008) Exploring climatic and biotic controls on Holocene vegetation change in Fennoscandia. *Journal of Ecology* 96: 247–259.
- Ohlson M, Brown KJ, Birks HJB, et al. (2011) Invasion of Norway spruce diversifies the fire regime in boreal European forests. *Journal of Ecology* 99: 395–403.
- Prentice IC (1985) Pollen representation, source area, and basin size – Toward a unified theory of pollen analysis. *Quaternary Research* 23: 76–86.
- Seppä H, Alenius T, Bradshaw RHW, Giesecke T, Heikkilä M, and Muukkonen P (2009) Invasion of Norway spruce (*Picea abies*) and the rise of the boreal ecosystem in Fennoscandia. *Journal of Ecology* 97: 620–640.
- Sugita S (1993) A model of pollen source area for an entire lake surface. *Quaternary Research* 39: 239–244.
- Szafer W (1935) The significance of isopollen lines for the investigation of the geographical distribution of trees in the post-glacial period. *Bulletin de l'Académie Polonaise des Sciences B* 1935: 235–239.
- Taylor RE (2000) Fifty years of radiocarbon dating. *American Scientist* 88: 60–67.
- Traverse A (1988) *Paleopalynology*. London: Unwin Hyman.
- von Post L (1916) Skogsträdpollen i sydsvenska torfmosselagerföljder. *Geologiska föreningens i Stockholm förhandlingar* 38: 384–394.
- Willis KJ, Bailey RM, Bhagwat SA, and Birks HJB (2010) Biodiversity baselines, thresholds and resilience: Testing predictions and assumptions using palaeoecological data. *Trends in Ecology and Evolution* 25: 583–591.
- Willis KJ and Niklas KJ (2004) The role of Quaternary environmental change in plant macroevolution: The exception or the rule? *Philosophical Transactions of the Royal Society of London Series B* 359: 159–172.
- Wright HE Jr., Kutzbach JE, Webb T III, Ruddiman WF, Street-Perrott FA, and Bartlein PJ (1993) *Global Climates Since the Last Glacial Maximum*. Minneapolis, MN: University of Minnesota Press.

Ancient Plant DNA

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Introduction

The study of DNA recovered from ancient plant material has had a short, but turbulent, history. Similar to other ancient DNA (aDNA) subfields, plant aDNA studies experienced early highs, with high-profile papers reporting the recovery of groundbreaking genetic material from samples that dated back millions of years. However, these reports from the early 1990s were quickly followed by disappointing lows, most notably the discovery that contamination problems plagued the results. Fortunately, by the late 1990s, the field staged a healthy recovery, and DNA from old plant material has now been used in a wide range of applications, from forensics to archaeology to ecology (Palmer et al., 2012). However, despite the interest, plant aDNA studies have lacked the focus that other areas of aDNA have achieved. For example, one review (Gugerli et al., 2005) notes that of a sample of almost 500 papers on aDNA published between 1984 and 2004, less than 60 (11%) dealt with plants. Furthermore, despite the power that aDNA studies of other types of biota have had to offer in the Quaternary context (cf. Willerslev and Cooper, 2005), plant aDNA has remained essentially unexploited. This is a pity, because conventional terrestrial paleoecology relies predominantly on plant remains (e.g., pollen and spores), thus fossil plant material is abundant and detailed paleoecological information is available for hypothesis testing using aDNA.

The Recovery and Analysis of Ancient Plant DNA

Ancient DNA

The term aDNA can be used to loosely describe DNA that is preserved in, or recovered from, any nonliving source that has not been specifically preserved at the time of death or sampled for genetic analyses. Characteristically, such DNA will have undergone some degree of degradation, resulting in both reductions in the overall yields and limitations on the size of analytical fragments. As such, there are limits to the analytical techniques that can be applied to aDNA. Many methods that are currently becoming popular in studies on modern DNA require better quality DNA than can be recovered from aDNA specimens. For most of the aDNA field's history, researchers developed their analyses to focus on the amplification of short <500 base pair (bp) fragments of DNA using the polymerase chain reaction (PCR). One of the key features of PCR is that particular genes or genetic areas can be targeted in order to answer research questions. Often, genetic regions can be chosen so as to be either unique to the species of interest or conversely shared among a variety of related taxa, enabling the exclusion, or inclusion, of different original sources of

DNA from the analysis. For example, studies that aim to identify to which species or taxonomic group a particular sample belongs must find a suitable genetic sequence that is known to be variable between potential source taxa, in such a way that if the sequence can be recovered from the questionable sample, and it matches a sequence obtained from the candidate sources, the identification can be made. Studies that are focused on the phylogenetic relationship between specimens or those that investigate population-specific factors also rely on markers that can discriminate individuals and taxa. However, these studies also face the requirement of containing enough informative data that can be combined with computational analyses to resolve the questions.

Genetic Regions Analyzed

In comparison to animals, plant genomes offer an increased number of distinct genetic targets for analyses. Plant genomes are comprised of three unique components: mitochondrial DNA (mtDNA), chloroplast DNA (cpDNA), and nuclear DNA (nuDNA). All three have been investigated in plant aDNA studies, although not in equal frequency due to various strengths and weaknesses. Most plant cells contain multiple copies of mtDNA and cpDNA, thereby increasing the chance of their recovery from ancient samples. However, plant mtDNA is not frequently used in aDNA research, because unlike animal mtDNA, it frequently recombines and has a very slow mutation rate. In contrast, cpDNA does not recombine and evolves more quickly, albeit still at a fourth of the rate of nuDNA. Therefore, cpDNA is particularly useful for aDNA in taxonomic identification and phylogenetic analyses, commonly using the *rbcl* gene and *trnL* intron. Taberlet et al. (2007) have recently proposed the chloroplast *trnL* (UAA) intron as a candidate for genetically 'barcoding' modern plants. In principle, a barcode is a genetic region with a unique sequence for every species. Like barcodes on commercial products, a quick analysis can yield the exact item, or in this case, taxon. This approach has become popular in many branches of biology, because it allows researchers to identify the taxonomic affiliation of unknown samples. The chloroplast *trnL* (UAA) intron may serve as such a DNA barcode for plants because it has sufficient variability to distinguish between many species and genera, while still having conserved flanking regions for PCR amplification in all plant families. Furthermore, the P6 loop of the intron is short enough (10–143 bp) to be appropriate for aDNA research.

Compared to mtDNA and cpDNA, plant cells contain very few copies of nuDNA. However, nuDNA codes for the phenotype of the plant and therefore has been used to understand various functional properties of ancient specimens (e.g., Jaenicke-Després et al., 2003). Microsatellites (short repeated sequences of variable length) in nuDNA have also been studied

to understand relationships between varieties of a species, such as grapes (Manen et al., 2003). Another important feature of the nuclear genome is polyploidy – the trait of having more than two pairs of homologous chromosomes per cell. Polyploidy often leads to important phenotypic changes, especially for agricultural crops. Naturally, it is an important concern for aDNA research. However, it is challenging to determine the number of copies of a chromosome given the degraded nature of ancient samples. Rather, researchers have investigated this topic through the use of markers unique to each polyploid type (e.g., Blatter et al., 2002).

Source Materials Analyzed for Ancient Plant DNA

Naturally Occurring Source Materials

Many different source materials are available for ancient plant DNA studies. All plant cellular tissues contain DNA, and thus, any that can be found preserved *postmortem* are of interest. Plant material represents the bulk of the global necromass (the weight of dead organisms), and as such, a large amount of naturally preserved material exists that can be exploited in aDNA analyses. Much of this material can be visually identified as plant matter (see [Plant Macrofossil Introduction](#)) such as leaves, woods (e.g., Dumolin-Lapegue et al., 1999; Liepelt et al., 2006), or pollen (e.g., Paffetti et al., 2007; Parducci et al., 2005). Other studies, however, have successfully recovered ancient plant DNA from samples where visual identification is impossible – including permafrost and sediment cores (Figure 1) dating back several hundred thousand years (e.g., Willerslev et al., 2003, 2007), human and animal coprolites, some which date back nearly 30 thousand years (e.g., Poinar et al., 1998; Figure 2), packrat middens (Küch et al., 2002), and even from within the colon of mummified specimens (Rollo et al., 2002). Although the above are limited to samples less than 800,000 years old, claims have also been made for the recovery of much older DNA – from samples that date back millions of years.

However, as discussed in more detail below, it is likely that the DNA recovered from such ancient remains was probably derived from modern contaminants, and as such, the oldest yet authentic DNA recovered dates only as far back as 450,000–800,000 years (Willerslev et al., 2007). Furthermore, other potentially useful sources of ancient plant DNA remain to be exploited. The successful recovery of DNA from sediment and permafrost cores suggests that authentic plant aDNA might be expected to survive in stable anaerobic environments, such as sediments found at the bottom of certain lakes (Gugerli et al., 2005) and other marine environments, as well as in semi-permanent ice, such as permafrost (Willerslev et al., 2003) and ice sheets (Willerslev et al., 2007).

Herbarium Collections

In addition to naturally preserved material, many other sources of DNA are available for use in ancient plant DNA studies. Globally, herbariums contain vast amounts of dried or otherwise preserved specimens that may often contain preserved DNA (cf. Drabkova et al., 2002; Saltonstall, 2002). It is worth noting that studies using herbarium specimens have not been limited to plant DNA. Several interesting studies have also

reported the recovery of historic plant pathogen DNA from herbarium specimens, including that of the potato blight strains (*Phytophthora infestans*) dating as far back as the Great Irish Famine of 1845, recovered from dried potato specimens (e.g., Ristaino et al., 2001), and of the bacterium *Xanthomonas axonopodis* pv. citri from historic citrus canker specimens, some of which were more than 90 years old (Li et al., 2006).

Archaeological and Historical Collections

Plant tissues and secondary products have played an important role in human society. A wide variety of preserved materials have been demonstrated suitable for DNA-based study. Examples include wood used for building construction, furniture, cultural and other historical artifacts (e.g., Burger et al., 2000; Deguilloux et al., 2002), and papyrus (Marota et al., 2002). Seeds of cultivated plants from numerous archaeological sites have been studied, including wheat (*Triticum* spp.; e.g., Allaby et al., 1999), maize (*Zea mays*) (Figure 3; e.g., Jaenicke-Després et al., 2003), and various grape (*Vitis vinifera*) varieties (Manen et al., 2003). Plant DNA has also been recovered from the interior surfaces of waterlogged amphoras, thereby revealing past contents (Hansson and Foley, 2008). It is even possible to isolate and identify genetic sequences directly from archaeological sediment (Hebsgaard et al., 2009). The source of the so-called ‘dirt’ DNA is for the most part unclear. However, Tanaka et al. (2010) claim to have recovered DNA from plant phytoliths isolated from archaeological sediment samples (but see Elbaum et al., 2009). (For a more detailed discussion of plant DNA as it pertains to archaeology, see Schlumbaum et al.’s (2008) thorough review.)

The Scope of Ancient Plant DNA

Plant Domestication

Due to their relevance in human history and their fundamental position in human society, a large portion (almost 50% of studies assessed in a recent comprehensive survey by Gugerli et al., 2005) of plant aDNA studies have focused on agricultural species. One key area of research is the history of plant domestication, using the large quantities of well-preserved and accurately dated material that is found in many archaeological sites (mostly no older than a few thousand years). The genetics of modern crops has been extensively studied, and as such, many molecular markers developed in those studies have been useful in the archaeogenetic context (Brown, 1999) to investigate such questions as the origin and spread of agriculture in species that include maize (*Zea mays*; Freitas et al., 2003; Lia et al., 2007), bread wheat (*Triticum aestivum*; Li et al., 2011), the bottle gourd (*Lagenaria siceraria*; Erickson et al., 2005), and *Vitis vinifera*, the wine grape (Manen et al., 2003). Furthermore, some studies have even investigated the selection undergone in the genome of such crops. For example, by using phylogenetic methods and DNA recovered from a variety of modern and ancient specimens, Allaby et al. (1999) were able to investigate the evolution of the high-molecular-weight glutenin gene family and, using their results, argue that Emmer wheat (*Triticum dicoccum*) may have been domesticated on two separate occasions. While that study provided insights into evolution



Figure 1 DNA from permafrost and soil. (a) Exposed permafrost section where DNA can be obtained through horizontal drilling into the cliff face, Kolyma Lowland, Yakutia, Siberian Far East (Photograph by David Gilichinsky). (b) Permafrost where DNA must be obtained by conventional vertical drilling techniques, Kolyma Lowland, Yakutia, Siberian Far East (Photograph by David Gilichinsky). (c) Hand drilling equipment for use in horizontal permafrost coring into exposures, Taimyr, North Eastern Siberia (Photograph by Clare Flemming). (d) Heavy drilling equipment for vertical permafrost coring, Kolyma Lowland, Yakutia, Siberian Far East (Photograph by David Gilichinsky). (e) Permafrost core obtained as described earlier. The drilling head is spiked with known control vector DNA to monitor for sample contamination when removing the outer surface of the cores during subsequent laboratory manipulation (Photograph by Clare Flemming). (f) Soil samples taken from Scladina Cave in Belgium for aDNA analyses using sterile conical centrifuge tubes (Photograph by Jonas Binladen and Eske Willerslev).

prior to the selection imposed by humans during domestication, other studies actually investigated evolution during the domestication process. For example, using archaeological samples spanning a range of dates (Figure 3), Jaenicke-Després et al. (2003) obtained DNA sequences from three genes (one involved in the control of plant architecture, one in storage protein synthesis, and one in starch production) that were hypothetically selected for (either deliberately or incidentally) during the domestication of maize and provide insights into the timing and sequence of selection on those characteristics. In a similar fashion, Palmer et al. (2009) have used aDNA to uncover the surprising origins of Egyptian two-rowed barley (*Hordeum vulgare* ssp. *vulgare*). They found that, despite having a more primitive morphology, the barley evolved from the derived six-row form and may have been perpetuated due to adaptations to local environmental conditions.

Population Paleogenetics

An additional area where plant aDNA has begun to play an important role is in population paleogenetics, that is, analyses that investigate how species or related taxa are genetically

changing through time. On a relatively recent scale, herbarium specimens of the common reed (*Phragmites australis*) have been used to reconstruct the timing and route of invasion into North America by a nonnative genotype during the twentieth century (Saltonstall, 2002). Specifically, by comparing two regions of cpDNA from non-North American samples collected prior to 1910 with those from modern North American specimens, Saltonstall (2002) demonstrated both that an introduction had occurred and that the introduced type (which is morphologically identical to the native) has displaced the native plant and even invaded previously uninhabited regions. On a deeper timescale, several studies have investigated nuDNA allelic diversity between modern and much older specimens, such as 4000-year-old seagrass, *Posidonia oceanica* (Raniello and Procaccini, 2002), to investigate changes in population diversity.

Bulked Specimen Analyses

Environmental reconstruction

In order to generate statistically meaningful results, studies such as those outlined above often require the analysis of DNA from multiple individual samples from different



Figure 2 Coprolite as a source of ancient plant DNA. Plant remains have been recovered from various coprolite sources, including 2000-year-old human (Poinar et al., 2001) (a) and 20 000-year-old ground sloth (Poinar et al., 1998) (b). (a) Reprinted from Poinar HN, Kuch M, Sobolik KD, et al. (2001) A molecular analysis of dietary diversity for three archaic Native Americans. *Proceedings of the National Academy of Sciences of the United States of America* 98: 4317–4322. Copyright 2001 National Academy of Sciences, USA. (b) Reprinted with permission from Macmillan Publishers Ltd: Nature Reviews Genetics: Hofreiter M, Serre D, Poinar HN, Kuch M, and Pääbo S (2001) Ancient DNA. *Nature Reviews Genetics* 2: 353–359, copyright 2001.



Figure 3 DNA from archaeological maize. Archaeological maize cob from the Ocampo Caves (Valenzuela Cave), in northeast Mexico, dated to approximately 4000 years before the present (length, 47 mm). DNA from this and similar cobs has been used for aDNA studies on the domestication of maize (Jaenicke-Després et al., 2003). Reprinted from Jaenicke-Després V, Buckler ES, Smith BD, et al. (2003) Early allelic selection in maize as revealed by ancient DNA. *Science* 302: 1206–1208. Reprinted with permission from AAAS.

locations and time periods. Furthermore, should the research questions relate to plant communities as a whole, and not simply individual species, these methods are less than ideal. Recent developments in ‘bulk-screening’ techniques provide the means to effectively reconstruct past environments as well as investigate vertebrate diets from ancient coprolites (Tringali and Rubin, 2005).

Environmental reconstructions have so far been conducted using a variety of source materials, including glacial silt (Gould et al., 2010), marine cores (Paffetti et al., 2007), terrestrial sediment samples (Haile et al., 2007), rodent middens (Küch et al., 2002), and coprolites from ancient humans (Rasmussen et al., 2009) and extinct mammals (Poinar et al., 1998). Studies that present information about a range of coexisting plants are particularly exciting. For example, using samples from permafrost sediment cores dated between 10 000 and 400 000 years (Figure 1(a)–1(e)), Willerslev et al. (2003) were able to recover DNA sequences from at least 19 different plant taxa, providing insights into how the taxonomic diversity and composition of the Beringian vegetation changed over that interval. Among other things, this approach has several useful advantages in that as long as the cores can be accurately dated, and DNA preservation is adequate, DNA can be recovered from specific target points in time. Obviously, due to the somewhat stochastic nature of the fossil and archaeological record, this cannot always be achieved in other situations. An important question in sediment-based DNA studies is whether aDNA may have moved across stratigraphic layers. For example, Haile et al. (2007) investigated DNA leaching in dry sediments from a New Zealand cave and found evidence for downward migration of DNA.

Diet

Genetic analysis of desiccated feces – also known as coprolites – can be used to identify which organism produced the feces and what plants and animals formed its diet. Despite the corrosive nature of mammalian digestive processes, a significant amount of amplifiable DNA can be recovered from both recent and ancient coprolites. For aDNA studies, bulk screening of plants in coprolites presents a fascinating opportunity to better understand the diets of past mammals. For example, van Geel et al. (2008) tested feces found in the digestive tract of a

22 500-year-old Siberian woolly mammoth for the *rbcl* gene and recovered ten DNA sequences which were consistent with the plant macrofossil and microfossil remains in the coprolite. Bulk screening of ancient human coprolites is also feasible; however, apart from a handful of studies (e.g., Rasmussen et al., 2009), this approach has been arguably underutilized by the archaeological community. Indeed, the analysis of plant DNA within a coprolite can be readily tested in conjunction with the DNA of the individual who produced the stool (cf. Gilbert et al., 2008; Rasmussen et al., 2009).

Before attempting to recover plant DNA from ancient coprolites, researchers should be aware of a number of noteworthy limitations and caveats. First, coprolites frequently contain various chemical compounds which impede DNA extraction and inhibit PCR amplification. Second, it has recently become recognized that plant DNA which is not part of an organism's diet can be incorporated in coprolites through inadvertent swallowing of airborne and waterborne pollen (cf. Reinhard et al., 2008). This occurrence may complicate interpretations of past diets but conversely may aid environmental reconstructions by providing a snapshot of animal's habitat shortly before its death. Another critical factor in bulk-specimen research is the database used for sequence queries. In order to determine the taxonomic origin of a genetic sequence, computational searches must be made against DNA databases, such as GenBank, in order to locate the most likely candidate. Thus, taxonomic identifications are fundamentally limited by the scope of the database. On the other hand, it is possible to revisit old data as online databases grow larger. For example, Reinhard et al. (2008) reanalyzed plant sequences recovered from prehistoric human coprolites from Texas first reported by Poinar et al. (2001). With the aid of an expanded database, they found that some previously unidentified sequences matched plants with medicinal and hallucinogenic properties. A final concern in coprolite DNA research is that it suffers from stochasticity in the finding of samples, thereby limiting choice of time periods and species that can be studied.

Forensic and Archaeological Uses

Not all plant aDNA work is focused on population or evolutionary questions. Plant aDNA studies have also played a role in simpler identification studies, such as the forensic or archaeological uses of the identification, and even tracing and certification of plant products. Recent studies have included analyses on plant-derived foods and drinks (e.g., Manen et al., 2003), fibers and textiles (e.g., Burger et al., 2000), woods (e.g., Deguilloux et al., 2004), and psychoactive drugs (e.g., Mukherjee et al., 2008).

Problems and Limitations of aDNA Studies

Extraction of Amplifiable Plant DNA

As with all aDNA studies, several problems challenge this field of study and must be appreciated in order to gauge the possibilities offered by plant aDNA. One of the simplest issues is something faced by all plant DNA researchers and involves the difficulties associated with purifying nucleic acids from plant tissues. Plant tissues are generally richer in primary or

secondary metabolites (such as carbohydrates or phenolic compounds) than animal tissues. Furthermore, many of the plant tissues that preserve best in the archaeological or paleontological record (such as wood or charred seeds) naturally contain these inhibitors (Gugerli et al., 2005). On top of this, many man-made products (such as treated wood or fibers) have been treated with other products (such as dyes) that inhibit PCR. However, due to the commonplace nature of these problems in modern plant studies, they are not insurmountable. Future research will determine which existing methods to purify nucleic acids away from inhibitors are compatible with ancient samples. Instead, two much bigger problems seriously challenge plant (indeed all) aDNA studies, those of *postmortem* DNA damage and contamination.

The Credibility of Early Plant aDNA Studies

Questions about the credibility of results from aDNA studies arose almost immediately following the first publications of plant aDNA. In 1990, 770-bp-length DNA sequences were published that were attributed to 17–20-million-year-old fossilized plant cpDNA (Golenberg et al., 1990). However, although the fossils showed remarkable preservation, including cell structures such as the chloroplasts, and at first glance seemed ideal specimens, an attempt to replicate the results in a second laboratory failed (Sidow et al., 1991). Sidow et al. (1991) reported only the presence of high concentrations of high-molecular-mass bacterial DNA derived from living bacteria and suggested that the most likely origin of such DNA was through recent contamination during the excavation process. Furthermore, Pääbo and Wilson (1991) also expressed skepticism that a 770 bp sequence could be recovered after 20 million years. Based on laboratory calculations of depurination rates (and subsequent fragmentation) under hydrated conditions, these authors claimed that no sequence over 800 bp should exist after 5000 years, and nothing amplifiable should exist after 4 million years. Despite some counter arguments, and a few other studies in which researchers claim to recover aDNA from Miocene samples (e.g., Kim et al., 2004; Panieri et al., 2010; Figure 4), an overwhelming body of evidence now exists that suggests that Golenberg et al. (1990) did not successfully retrieve authentic aDNA sequences (cf. Hebsgaard et al., 2005).

The Postmortem Degradation of DNA

As the field of study has matured, it has become apparent that the scope of aDNA is limited by the *postmortem* degradation of DNA. Although *in vivo* DNA is protected by an array of repair mechanisms, *postmortem*, these cease to be active. Therefore, the damage processes that target DNA confer irreparable damage. The end result of most of these processes, which include, among other things, various oxidative and hydrolytic reactions, as well as cross-linking of the DNA to other molecules (including other DNA strands), is that the DNA rapidly becomes unsuitable for PCR, within days in extreme situations, but more often on the scale of years. At first, these limitations will simply be reflected through reductions in the number of templates available for amplification and the potential size of amplifiable fragments. Several studies on plant DNA

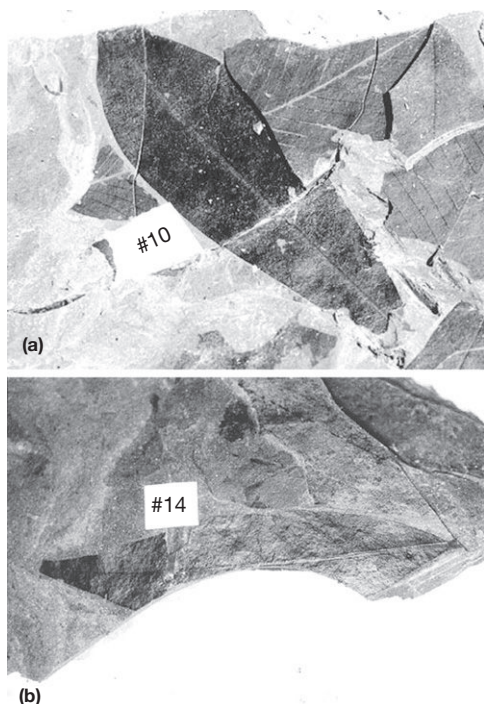


Figure 4 Examples of fossil plant material reported as a source of ancient DNA. Voucher photograph of fossil specimens used in an ancient DNA study claiming the recovery of Miocene *Magnolia* DNA sequences (17–20 Ma; Kim et al., 2004). (a) *Magnolia latahensis* (Berry) Brown. (b) *Persea pseudocarolinensis* Lesquereux. Many aDNA researchers believe that DNA is unlikely to survive for such a long time and would argue that the results are likely derived from contamination. Reprinted from Kim S, Soltis DE, Soltis PS, and Suh Y (2004) DNA sequences from Miocene fossils: An *ndhF* sequence of *Magnolia latahensis* (Magnoliaceae) and an *rbcL* sequence of *Persea pseudocarolinensis* (Lauraceae). *American Journal of Botany* 91: 615–620, with permission from Botanical Society of America, copyright 2004.

demonstrate this phenomenon. For example, Deguilloux et al. (2002) found an inverse relationship between amplification success rate and amplicon size in dry wood; smaller amplicons were present in larger numbers than longer ones due to degradation. The authors also demonstrated that, in the same samples, cpDNA, mtDNA, and nuclear ribosomal DNA (rDNA) genes could be amplified more often than single-copy nuDNA genes. This is consistent with DNA degradation, as individual cells contain multiple copies of cpDNA, mtDNA, and rDNA, and assuming DNA degradation affects all DNA sequences in a similar manner, these templates will out-survive those of the rarer single-copy nuDNA genes. Interestingly, the authors also found the same relationship when DNA content was investigated from young tissues from the outer part of tree trunks. That is, they are more likely to yield DNA, and yield larger fragments of DNA, than tissues from older heartwood samples (Deguilloux et al., 2002). Other researchers have approached the issue of *postmortem* degradation by searching for proxy measures of DNA preservation, thereby avoiding the testing of futile samples. For example, Elbaum et al. (2006) employed infrared spectroscopy to determine the amount of lignin in ancient olive pits and subsequently predict whether a given sample is likely to yield amplifiable DNA.

Impact of Specimen Age on DNA Survival

The implications of DNA degradation for any particular study depend largely on whether a sample is beyond the age at which DNA remains sufficiently intact and undamaged. Unfortunately, there is no simple answer to what this age is, because too little is known about *postmortem* DNA degradation (specifically which damage processes affect what specimens, in what combinations, and at what rates) to provide definitive answers. However, some estimates have been made, based on the knowledge that the rate of DNA degradation is inversely correlated with heat. Using calculated deamination and depurination kinetics for the four nucleotides, it has been estimated that under mammalian physiological salt conditions, neutral pH and an ambient temperature of 15 °C, 100 000 years is a likely estimate of the time beyond which DNA will be degraded to a level below the size of a useful amplifiable fragment (Lindahl, 1993). Under certain environmental conditions, however, such as encasement in ice or permafrost, this period may be extended by perhaps several times (Willerslev et al., 2003), while in warmer environments, the survival limit of amplifiable DNA is likely to be much shorter.

Problems Associated with DNA Degradation

DNA degradation presents several problems for the aDNA field (cf. Gilbert et al., 2005). First, researchers are limited as to the quantities of amplifiable DNA that they can recover. Second, due to DNA fragmentation and cross-linking, researchers are often limited in their ability to amplify templates of less than a few hundred bp in size (cf. Willerslev and Cooper, 2005). When this is combined with the insights about the implications of the relative abundance of different genetic targets, it becomes apparent that there are limitations on what can be achieved in this field of study. A further problem, and one that is most serious for the credibility of the discipline, is that of sample contamination. The low abundance of DNA extracted from a degraded specimen can easily be masked by modern DNA from external contaminants of the same or a similar DNA sequence. Without proper treatment to remove the environmental DNA, it is a very simple (and, unfortunately, common) matter to coamplify both sources of DNA. Depending on the relative concentrations of host-to-contaminant DNA in the PCR, the target DNA may be either completely swamped or sufficiently modified to result in erroneous data.

Contamination

Due to the multistep processes that are employed to successfully extract the low concentrations of DNA associated with ancient remains, contaminants can enter the system at many stages (cf. Roberts and Ingham, 2008). The most common sources of contaminants include the following:

Microorganisms. Most PCR reactions are based upon the hypothesis that primers will bind to specific loci not present in unrelated organisms, thus will selectively amplify only the DNA of interest. However, due to the potentially large numbers of genetically different, and as yet unsequenced, microorganisms that are estimated to exist, it is very difficult to

rule out the presence of short sequence homologues between the sample of interest and contaminant microorganisms in the environment.

Modern DNA introduced during sample collection. It is possible to inadvertently transfer modern DNA to ancient samples during field collection or later analyses. Therefore, it is recommended to collect and analyze plant aDNA samples with sterile equipment while wearing examination gloves. Although it has been shown to be advantageous to collect hominin fossils while wearing full body suits (Fortea et al., 2008), such a complicated procedure is not common for ancient plant collection and may be unnecessary because human contamination can be readily identified and ignored in ancient plant studies. In contrast, food particles and plant-based textiles are more problematical sources of contamination for plant aDNA research.

Sample conservation or other treatments. Many archaeological or historical collections may have been treated with glues and varnishes containing some plant-derived products and thus provide a contaminating DNA source.

Related specimens. It is possible for tissue or DNA extracted from one specimen to cross-contaminate others extracted in the same laboratory. In this case, both sources are coextracted; PCR primers may bind equally well to the authentic and contaminant sequence (or in the worst case, preferentially to the contaminant), and thus coamplify.

Previously amplified PCR fragments. The exponential nature of PCR results in huge numbers of amplified products (amplicons) concentrated into tiny volumes, which may spread rapidly and contaminate ancient samples with DNA, swamping the authentic aDNA. In addition, high levels of contamination can persist in a laboratory environment for long periods of time. Consequently, it is essential that all aDNA work be conducted in a dedicated clean laboratory far removed from amplified DNA. Furthermore, aDNA labs are required to be designed in ways to minimize outside contamination (Willerslev and Cooper, 2005).

Future Directions

Next-Generation Sequencing

Plant aDNA research is arguably on the cusp of being transformed through state-of-the-art technologies collectively known as next-generation sequencing (NGS; Palmer et al., 2012). NGS is quickly changing how DNA is studied in many fields, including aDNA. Most NGS platforms operate by isolating hundreds of thousands of fragments of DNA, and then sequencing each fragment in parallel. In this way, NGS platforms, such as Roche/454 FLX series, Life Technologies SOLiD series, and Illumina Genome Analyzer series, are capable of generating data in the magnitude of hundreds of millions to billions of nucleotides in a single run (Millar et al., 2008). In addition to yielding several orders of magnitude more data than PCR, NGS also retrieves shorter DNA sequences than is possible with conventional methods; PCR is normally used to amplify DNA fragments at least 60 bp in length, while NGS can recover DNA segments as short as 25 bp. Many other fields view this as a limitation, but it is a boon for aDNA research (Figure 5). As discussed previously, ancient samples generally yield many short DNA fragments and relatively few long

fragments; taken to the extreme, NGS can recover DNA from samples which are too degraded for PCR amplification. In addition, by only studying very short DNA fragments, one should theoretically encounter a lower rate of contaminating exogenous DNA (Figure 5). This is possible because modern DNA tends to have a higher molecular weight than aDNA and comprises a relatively smaller proportion of short DNA strands found in ancient samples. Nevertheless, NGS has drawbacks including expensive instruments and the challenge of analyzing gigabytes of data. Several high-profile publications in vertebrate aDNA have utilized NGS (e.g., Green et al., 2010; Miller et al., 2008), although once again, plant aDNA research lags slightly behind. Based on findings from other NGS studies, it is anticipated that plant aDNA will soon employ NGS for deep cloning, shotgun sequencing, and targeted capture of certain genetic regions.

Deep cloning

Through most of the history of aDNA research, bacterial cloning of PCR amplicons was used as a means to investigate the composition of DNA sources within complex sources of DNA, such as the plant components of coprolites. Unfortunately, however, while useful, the number of cloned PCR products that required sequencing in order to fully characterize a complex substrate was difficult to achieve. Thus, the method rarely was able to describe, for example, the rarest components in a sample. NGS provides an efficient – and reasonably affordable – alternative. Rasmussen et al. (2009) employed NGS deep cloning of plant cpDNA PCR products from ancient human coprolites and recovered 286 sequences. Thus, deep cloning using NGS is a viable alternative to transforming and sequencing hundreds of bacterial colonies. Furthermore, deep cloning can become relatively inexpensive when multiple samples are combined into a single NGS run; the origin of each DNA segment is easily determined using sample-specific DNA index tags which are attached to aDNA molecules (ligated) before samples are merged into a common pool.

Shotgun sequencing

Determining the DNA sequence of large sections of a genome is challenging for people studying aDNA due to the fragmentation of the template molecules. NGS offers a solution to this problem through shotgun sequencing of the fragmented templates. Although randomly chosen from the aDNA extract, the large number of sequences generated in parallel in each sequencing run (hundreds of thousands to tens of millions) means that large amounts of aDNA sequence can, in theory, be rapidly generated. Shotgun sequencing with NGS platforms has already been performed on ancient vertebrate samples, most notably for whole-genome sequencing (WGS) of Neanderthals (Green et al., 2010) and woolly mammoths (Miller et al., 2008). In spite of these advances, the shotgun sequencing of ancient samples is problematic because most recovered sequences are derived from bacteria and fungi – primarily saprotrophs which feed on organic matter *postmortem*. Therefore, it is necessary to shotgun sequence more DNA from ancient samples than modern ones. Nevertheless, plant aDNA researchers can apply NGS

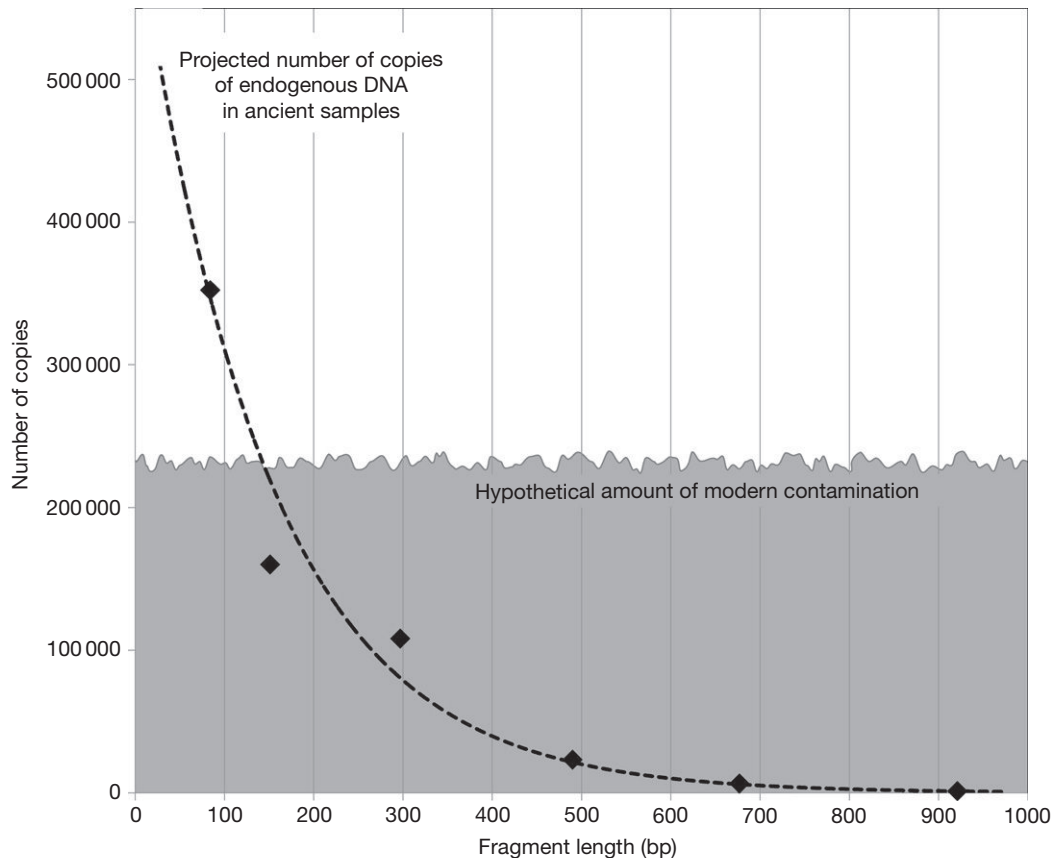


Figure 5 Relationship between DNA fragment length and copy number in ancient samples. Due to DNA degradation, ancient samples contain many short DNA strands but very few long segments. Contamination from modern sources, on the other hand, is expected to be of higher average molecular weight with a relatively random distribution of sequence length. Therefore, regardless of the extent of contamination, it is advantageous to analyze short DNA fragments from ancient samples. By using NGS to study DNA strands 25–35 bp in length, one can efficiently isolate aDNA sequences from the ‘background noise’ of modern contamination. Based on DNA sequences found in 28 000-year-old skeletal remains of a Siberian mammoth (Poinar et al., 2006, Figure S1).

shotgun sequencing for various purposes. For example, WGS of extinct species or agricultural cultivars will undoubtedly reveal lost genes and unknown genetic diversity. Bulk specimen samples, like coprolites and archaeological sediment, can also be shotgun sequenced on NGS platforms in order to recover and identify all taxa contributing to the sample. In other words, it is possible to forgo an amplification of a common genetic marker and directly obtain a picture of the entire biological community represented within a sample. Rates of contamination could be similarly quantified using the relative frequency of DNA sequences that would not be expected to be found in the sample, at least according to well-established paleontological and archaeological knowledge. For example, Rasmussen et al. (2009) recovered fragments of cattle (*Bos taurus*) DNA in a 14 000-year-old Native American coprolite but argue it is a laboratory reagent contaminant because cattle were not present in the Americas before their introduction by Europeans.

Targeted capture

As discussed, shotgun sequencing ancient samples frequently yields a disproportionately low amount of endogenous DNA. New technologies have been developed to target sequences of interest, including primer extension capture (PEC) and

array- and solution-based sequence capture (Knapp and Hofreiter, 2010). PEC uses biotinylated oligonucleotide primers to hybridize with sequences of interest, for example, around an mtDNA genome. Hybridized sequences are isolated through the use of streptavidin-coated magnetic beads, and all other DNA fragments are washed away. In this way, NGS can be efficiently applied to just the regions of interest. This approach has been used on ancient Neanderthal samples to recover the complete mtDNA genomes of five Neanderthals (Briggs et al., 2009). Like PEC, microarrays on glass slides target sequences of interest, but they have the advantage of capturing many more loci. Burbano et al. (2010) used microarrays to investigate ~14 000 single-nucleotide polymorphisms (SNPs) from Neanderthal samples; remarkably, this approach enriched targeted areas by nearly 200 000 times. The first targeted capture experiment on plant aDNA was published by Ávila-Arcos et al. (2011) on a collection of maize cobs and kernels up to 1000 years old. Using a set of RNA ‘baits’ in solution, the authors targeted 670 loci in the chloroplast and nuclear genomes and recovered up to 29 times more DNA of interest than shotgun (random) sequencing. The vast difference in enrichment values between Burbano et al.’s (2010) Neanderthal study and Ávila-Arcos et al.’s (2011)

maize study is partially due to optimization but also reflects different ways of calculating enrichment. Undoubtedly, plant aDNA research has only begun to capitalize on targeted capture methods to investigate genes of interest.

Analysis of Non-DNA Biomolecules

Ancient RNA

Although millennial-scale preservation of plant DNA has been studied for two decades, researchers have recently recognized that RNA can also be preserved at levels that enable informative sequence analysis. Paleogeneticists have so far ignored the transcriptome because RNA's single-stranded structure lacks the chemical stability of double-helical DNA, presumably leading to degradation of RNA shortly after the death of an organism. Contrary to this assumption, various observations have suggested that it is possible for RNA to survive for thousands of years, at least in seeds. Most notably, [Sallon et al. \(2008\)](#) report the successful germination and growth of a ~2000-year-old date (*Phoenix dactylifera*) seed excavated at Masada (Israel). For the seed to have germinated, RNA must have been present in a relatively undamaged state. Research into the preservation of ancient RNA is ongoing, and preliminary findings have been presented at conferences and reported by scientific media ([Callaway, 2010](#)). Future analyses of ancient transcriptomes will provide an important counterpart to aDNA research because RNA plays several special roles in living cells. For example, messenger RNA (mRNA) acts as the intermediary between DNA and protein synthesis. Unlike DNA, many copies of an mRNA molecule can be present in a cell at one time, as required by cellular demands and environmental conditions. Conversely, a gene may be inactive during certain developmental stages and no mRNA molecules are produced. Another form of RNA, microRNA (miRNA) acts as a gene regulator, turning on and off different parts of the genome. This is particularly interesting because it is believed that the bulk of evolutionary change occurs through gene regulation rather than novel mutation (see [Chen and Rajewsky, 2007](#)). Thus, by studying ancient mRNA and miRNA, it will be possible to better understand the evolutionary histories of plants. It should be noted that seeds are perhaps the only organic substrate that preserves RNA because they have evolved specifically to disperse and protect genetic sequences.

Ancient protein

As with ancient RNA, the analysis of proteins preserved in ancient samples can serve as a complement to aDNA research. Because proteins represent an end product of the genome in any specific tissue, they therefore represent an alternative way to investigate ancient organisms. Due to the redundancy of the genetic code – that is, multiple codons code for the same amino acid – closely related species may share identical proteins although their DNA sequences may differ. For this reason, proteins are rarely studied to infer evolutionary relationships. On the other hand, some proteins are better suited to long-term preservation than DNA or RNA. Thus, it may be possible to extract proteins from samples that are too degraded to yield nucleic acids. Analysis of 1400-year-old grape seeds by [Cappellini et al. \(2010\)](#) demonstrates the

advantages of combining the avenues of aDNA and ancient proteins.

Ancient protein resurrection

The reconstruction of extinct proteins may sound like science fiction but is now a reality in cutting-edge genetic research. Rather than extracting proteins from ancient samples, this approach starts with aDNA sequences and builds fresh proteins as specified by the genetic code. This allows scientists to investigate how ancient proteins compare to modern analogs. [Campbell et al. \(2010\)](#) were the first to 'resurrect' ancient proteins with the production of woolly mammoth hemoglobin. The researchers determined the sequence of the mammoth hemoglobin gene, expressed the protein using plasmids, and then discovered unknown adaptations which improve hemoglobin's effectiveness at low temperatures. This approach will be especially important in plant aDNA for understanding how domesticated crops have evolved as they were transported and grown around the world.

Conclusions

Developments over the last 20 years in molecular biology, sample retrieval, and analytical techniques have enabled the field of ancient plant DNA to mature into a position where it can be a valuable asset to Quaternary science. aDNA studies are, however, limited by DNA survival, and researchers must be aware of the serious consequences that this can have on the validity of aDNA results (cf. [Gilbert et al., 2005](#)). Thus, although a potentially powerful tool, future contributions are likely to be limited to samples from relatively young or cold environments, and as such, studies should be designed with these limitations of preservation specifically in mind. This will enable the maximization of potential gains and ensure that ancient plant DNA studies continue to provide unparalleled insights into the past.

See also: Dating Techniques; Plant Macrofossil Introduction. **Permafrost and Periglacial Features:** Permafrost. **Plant Macrofossil Records:** Arctic Eurasia. **Pollen Methods and Studies:** Databases and Their Application. **Vertebrate Studies:** Ancient DNA.

References

- Allaby RG, Banerjee M, and Brown TA (1999) Evolution of the high-molecular-weight glutenin loci of the A, B, D and G genomes of wheat. *Genome* 42: 296–307.
- Ávila-Arcos MC, Cappellini E, Romero-Navarro JA, et al. (2011) Application and comparison of large-scale solution-based DNA capture-enrichment methods on ancient DNA. *Scientific Reports* 1: 74.
- Blatter RHE, Jacomet S, and Schulmbaum A (2002) Spelt-specific alleles in HMW glutenin genes from modern and historical European spelt (*Triticum spelta* L.). *Theoretical and Applied Genetics* 104: 329–337.
- Briggs AW, Good JM, Green RE, et al. (2009) Targeted retrieval and analysis of five Neandertal mtDNA genomes. *Science* 325: 318–321.
- Brown TA (1999) How ancient DNA may help in understanding the origin and spread of agriculture. *Proceedings of the Royal Society B* 354: 89–98.
- Burbano HA, Hodges E, Green RE, et al. (2010) Targeted investigation of the Neandertal genome by array-based sequence capture. *Science* 328: 723–725.

- Burger J, Hummel S, and Herrmann B (2000) Palaeogenetics and cultural heritage. Species determination and STR-genotyping from ancient DNA in art and artefacts. *Thermochimica Acta* 365: 141–146.
- Callaway E (2010) Taking molecular snaps of ancient crops. *Nature News* <http://dx.doi.org/10.1038/news.2010.464> Published, online 13 September 2010.
- Campbell KL, Roberts JEE, Watson LN, et al. (2010) Substitutions in woolly mammoth hemoglobin confer biochemical properties adaptive for cold tolerance. *Nature Genetics* 42: 536–540.
- Cappellini E, Gilbert M, Geuna F, et al. (2010) A multidisciplinary study of archaeological grape seeds. *Naturwissenschaften* 97: 205–217.
- Chen K and Rajewsky N (2007) The evolution of gene regulation by transcription factors and microRNAs. *Nature Reviews Genetics* 8: 93–103.
- Deguilloux M-F, Pemonge M-H, and Petit RJ (2002) Novel perspectives in wood certification and forensics: Dry wood as a source of DNA. *Proceedings of the Royal Society B* 269: 1039–1046.
- Deguilloux M-F, Pemonge M-H, and Petit RJ (2004) DNA-based control of oak wood geographic origin in the context of the cooperage industry. *Annals of Forest Science* 61: 97–104.
- Drabkova L, Kirschner J, and Vlcek C (2002) Comparison of seven DNA extraction and amplification protocols in historical herbarium specimens of *Juncaceae*. *Plant Molecular Biology Reporter* 20: 161–175.
- Dumolin-Lapegue S, Pemonge H-M, Gielly L, Taberlet P, and Petit RJ (1999) Amplification of oak DNA from ancient and modern wood. *Molecular Ecology* 8: 2137–2140.
- Elbaum R, Melamed-Bessudo C, Boaretto E, et al. (2006) Ancient olive DNA in pits: Preservation, amplification and sequence analysis. *Journal of Archaeological Science* 33: 77–88.
- Elbaum R, Melamed-Bessudo C, Tuross N, Levy AA, and Weiner S (2009) New methods to isolate organic materials from silicified phytoliths reveal fragmented glycoproteins but no DNA. *Quaternary International* 193: 11–19.
- Erickson DL, Smith BD, Clarke AC, Sandweiss DH, and Tuross N (2005) An Asian origin for a 10,000-year-old domesticated plant in the Americas. *Proceedings of the National Academy of Sciences of the United States of America* 102: 18315–18320.
- Fortea J, de la Rasilla M, García-Tabernero A, Gigli E, Rosas A, and Lalueza-Fox C (2008) Excavation protocol of bone remains for Neandertal DNA analysis in El Sidrón Cave (Asturias, Spain). *Journal of Human Evolution* 55: 353–357.
- Freitas FO, Bendel G, Allaby RG, and Brown TA (2003) DNA from primitive maize landraces and archaeological remains, implications for the domestication of maize and its expansion into South America. *Journal of Archaeological Science* 30: 901–908.
- Gilbert MTP, Bandelt HJ, Hofreiter M, and Barnes I (2005) Assessing ancient DNA studies. *Trends in Ecology and Evolution* 20: 541–544.
- Gilbert MTP, Jenkins DL, Gotherstrom A, et al. (2008) DNA from pre-Clovis human coprolites in Oregon, North America. *Science* 320: 786–789.
- Golenberg EM, Giannasi DE, Clegg MT, et al. (1990) Chloroplast DNA sequence from a Miocene *Magnolia* species. *Nature* 344: 656–658.
- Gould BA, Leon B, Buffen AM, and Thompson LG (2010) Evidence of a high-Andean, mid-Holocene plant community: An ancient DNA analysis of glacially preserved remains. *American Journal of Botany* 97: 1579–1584.
- Green RE, Krause J, Briggs AW, et al. (2010) A draft sequence of the Neandertal genome. *Science* 328: 710–722.
- Gugerli F, Parducci L, and Petit RJ (2005) Ancient plant DNA, review and prospects. *New Phytologist* 166: 409–418.
- Haile J, Holdaway R, Oliver K, et al. (2007) Ancient DNA chronology within sediment deposits: Are paleobiological reconstructions possible and is DNA leaching a factor? *Molecular Biology and Evolution* 24: 982–989.
- Hansson MC and Foley BP (2008) Ancient DNA fragments inside Classical Greek amphoras reveal cargo of 2400-year-old shipwreck. *Journal of Archaeological Science* 35: 1169–1176.
- Hebsgaard MB, Gilbert MTP, Arneborg J, et al. (2009) 'The Farm Beneath the Sand' – An archaeological case study on ancient 'dirty' DNA. *Antiquity* 83: 430–444.
- Hebsgaard MB, Phillips MJ, and Willerslev E (2005) Geologically ancient DNA: Fact or artefact? *Trends in Microbiology* 13: 212–220.
- Hofreiter M, Serre D, Poinar HN, Kuch M, and Pääbo S (2001) Ancient DNA. *Nature Reviews Genetics* 2: 353–359.
- Jaenicke-Després V, Buckler ES, Smith BD, et al. (2003) Early allelic selection in maize as revealed by ancient DNA. *Science* 302: 1206–1208.
- Kim S, Soltis DE, Soltis PS, and Suh Y (2004) DNA sequences from Miocene fossils: an *ndhF* sequence of *Magnolia latahensis* (Magnoliaceae) and an *rbcL* sequence of *Persea pseudocarolinensis* (Lauraceae). *American Journal of Botany* 91: 615–620.
- Knapp M and Hofreiter M (2010) Next generation sequencing of ancient DNA: Requirements, strategies and perspectives. *Genes* 1: 227–243.
- Küch M, Rohland N, Betancourt JL, Latorre C, Stepan S, and Poinar HN (2002) Molecular analysis of an 11,700-year-old rodent midden from the Atacama Desert, Chile. *Molecular Ecology* 11: 913–924.
- Li W, Brlansky RH, and Hartung JS (2006) Amplification of DNA of *Xanthomonas axonopodis* pv. citri from historic citrus canker herbarium specimens. *Journal of Microbiological Methods* 65: 237–246.
- Li C, Lister DL, Li H, et al. (2011) Ancient DNA analysis of desiccated wheat grains excavated from a Bronze Age cemetery in Xinjiang. *Journal of Archaeological Science* 38: 115–119.
- Lia VV, Confalonieri VA, Ratto N, et al. (2007) Microsatellite typing of ancient maize: Insights into the history of agriculture in southern South America. *Proceedings of the Royal Society B* 274: 545–554.
- Liepert S, Sperisen C, Deguilloux M, et al. (2006) Authenticated DNA from ancient wood remains. *Annals of Botany* 98: 1107–1111.
- Lindahl T (1993) Instability and decay of the primary structure of DNA. *Nature* 362: 709–715.
- Manen J-F, Bouby L, Dalnoki O, Marival P, Turgay M, and Schlumbaum A (2003) Microsatellites from archaeological *Vitis vinifera* seeds allow a tentative assignment of the geographical origin of ancient cultivars. *Journal of Archaeological Science* 30: 721–729.
- Marota I, Basile C, Ubaldi M, and Rollo F (2002) DNA decay rate in papyri and human remains from Egyptian archaeological sites. *American Journal of Physical Anthropology* 117: 310–318.
- Millar CD, Huynen L, Subramanian S, Mohandesan E, and Lambert DM (2008) New developments in ancient genomics. *Trends in Ecology and Evolution* 23(7): 386–393.
- Miller W, Drautz DI, Ratan A, et al. (2008) Sequencing the nuclear genome of the extinct woolly mammoth. *Nature* 456: 387–390.
- Mukherjee A, Roy S, De Bera S, et al. (2008) Results of molecular analysis of an archaeological hemp (*Cannabis sativa* L.) DNA sample from North West China. *Genetic Resources and Crop Evolution* 55: 481–485.
- Pääbo S and Wilson AC (1991) Miocene DNA sequences – A dream come true? *Current Biology* 1: 45–46.
- Paffetti D, Vettori C, Caramelli D, et al. (2007) Unexpected presence of *Fagus orientalis* complex in Italy as inferred from 45,000-year-old DNA pollen samples from Venice lagoon. *BMC Evolutionary Biology* 7: S6.
- Palmer SA, Moore JD, Clapham AJ, Rose P, and Allaby RG (2009) Archaeogenetic evidence of ancient Nubian Barley evolution from six to two-row indicates local adaptation. *PLoS One* 4: e6301. <http://dx.doi.org/10.1371/journal.pone.0006301>.
- Palmer SA, Smith O, and Allaby RG (2012) The blossoming of plant archaeogenetics. *Annals of Anatomy* 194: 146–156.
- Panieri G, Lugli S, Manzi V, Roveri M, Schreiber BC, and Palinska KA (2010) Ribosomal RNA gene fragments from fossilized cyanobacteria identified in primary gypsum from the late Miocene, Italy. *Geobiology* 8: 101–111.
- Parducci L, Suyama Y, Lascoux M, and Bennett KD (2005) Ancient DNA from pollen, a genetic record of population history in Scots pine. *Molecular Ecology* 14: 2873–2882.
- Poinar HN, Hofreiter M, Spaulding WG, et al. (1998) Molecular coproscopy: Dung and diet of the extinct Ground Sloth *Nothrotheriops shastensis*. *Science* 281: 402–406.
- Poinar HN, Kuch M, Sobolik KD, et al. (2001) A molecular analysis of dietary diversity for three archaic Native Americans. *Proceedings of the National Academy of Sciences of the United States of America* 98: 4317–4322.
- Poinar HN, Schwarz C, Qi J, et al. (2006) Metagenomics to paleogenomics: Large-scale sequencing of Mammoth DNA. *Science* 311: 392–394.
- Raniello R and Procaccini G (2002) Ancient DNA in the seagrass *Posidonia oceanica*. *Marine Ecology Progress Series* 227: 269–273.
- Rasmussen M, Cummings LS, Gilbert MTP, et al. (2009) Response to comment by Goldberg et al. on 'DNA from pre-Clovis human coprolites in Oregon, North America'. *Science* 325: 148.
- Reinhard KJ, Chaves SM, Jones JG, and Iñiguez AM (2008) Evaluating chloroplast DNA in prehistoric Texas coprolites: Medicinal, dietary, or ambient ancient DNA? *Journal of Archaeological Science* 35: 1748–1755.
- Ristaino JB, Groves CT, and Parra G (2001) PCR amplification of the Irish potato famine pathogen from historic specimens. *Nature* 411: 695–697.
- Roberts C and Ingham S (2008) Using ancient DNA analysis in palaeopathology: A critical analysis of published papers, with recommendations for future work. *International Journal of Osteoarchaeology* 18: 600–613.
- Rollo F, Ubaldi M, Ermini L, and Marota I (2002) Ötzi's last meals, DNA analysis of the intestinal content of the Neolithic glacier mummy from the Alps. *Proceedings of the National Academy of Sciences of the United States of America* 99: 12594–12599.

- Sallon S, Solowey E, Cohen Y, et al. (2008) Germination, genetics, and growth of an ancient date seed. *Science* 320: 1464.
- Saltonstall K (2002) Cryptic invasion by a non-native genotype of the common reed, *Phragmites australis*, into North America. *Proceedings of the National Academy of Sciences of the United States of America* 99: 2445–2449.
- Schlumbaum A, Tensen M, and Jaenicke-Després V (2008) Ancient plant DNA in archaeobotany. *Vegetation History and Archaeobotany* 17: 233–244.
- Sidow A, Wilson AC, and Pääbo S (1991) Bacterial DNA in *Clarkia* fossils. *Philosophical Transactions of the Royal Society B* 333: 429–433.
- Taberlet P, Coissac E, Pompanon F, et al. (2007) Power and limitations of the chloroplast *trnL* (UAA) intron for plant DNA barcoding. *Nucleic Acids Research* 35: e14.
- Tanaka K, Honda T, and Ishikawa R (2010) Rice archaeological remains and the possibility of DNA archaeology: Examples from Yayoi and Heian periods of Northern Japan. *Archaeological and Anthropological Sciences* 2: 69–78.
- Tringe SG and Rubin EM (2005) Metagenomics: DNA sequencing of environmental samples. *Nature Reviews Genetics* 6: 805–814.
- van Geel B, Aptroot A, Baittinger C, et al. (2008) The ecological implications of a Yakutian mammoth's last meal. *Quaternary Research* 69: 361–376.
- Willerslev E, Cappellini E, Boomsma W, et al. (2007) Ancient biomolecules from deep ice cores reveal a forested southern Greenland. *Science* 317: 111–114.
- Willerslev E and Cooper A (2005) Ancient DNA. *Proceedings of the Royal Society B* 272: 3–16.
- Willerslev E, Hansen AJ, Binladen J, et al. (2003) Diverse plant and animal genetic records from Holocene and Pleistocene sediments. *Science* 300: 791–795.

Charred Particle Analyses

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Introduction

Burning, or combustion, of biomass is a complex process that rapidly oxidizes fuel, releasing energy in the form of heat, as well as greenhouse gases. The flames, often considered synonymous with fire, constitute the visible part of the reaction (Saito, 2001). The combustion equation for vegetation, estimated using cellulose, is $(C_6H_{10}O_5)_n + (6O_2)_n \rightarrow (6CO_2)_n + (5H_2O)_n + \text{energy}$ (Cofer et al., 1997). In addition to carbon dioxide and water vapor, small amounts of nitrogen, sulfur, halogens, and minerals may also be released during a fire event. In nature, however, the combustion reaction is rarely complete, with incomplete combustion resulting in the production of other products such as carbon monoxide, methane, hydrogen gas, hydrocarbons, soot, and charcoal (Figure 1). During a fire, organic matter may not be equally exposed to the combustion process, resulting in the additional production of charred particles. According to Jones et al. (1997), charcoal refers to any black-colored plant-derived material whose chemical composition and ultrastructure (i.e., the submicroscopic structure at the cellular level) have been significantly altered as a result of heating in a fire, though retaining the recognizable anatomic structure of the parent plant. In contrast, charred particles refer to plant material that has been partially blackened and chemically altered as a result of heating in a fire but without significant alteration of the ultrastructure. In this article, these particles will be collectively referred to as charcoal. All Quaternary studies dealing with charcoal, by quantifying particles or taxonomic identification, are called anthracology. Soil charcoal analysis (pedoanthracology) and the analysis based on archaeological material (archaeoanthracology) are important subdisciplines within anthracology.

From an Earth history perspective, stratigraphic deposits containing fossil charcoal reveal correspondence between the first appearance of terrestrial plants in the Silurian period approximately 420 million years ago and biomass burning (Figure 2; Bowman et al., 2009). Higher resolution charcoal records from the last 80 000 years suggest that the incidence of fire disturbance increases during warmer intervals and decreases during cooler intervals, commensurate with climate-driven changes in biomass (Daniau et al., 2010; Marlon et al., 2009). Other studies show that fire disturbance also responds to temporal changes in moisture and vegetation composition (Brown et al., 2005; Grimm et al., 2011; Hu et al., 2006; Ohlson et al., 2011; Power et al., 2008). Combined, these studies and others clearly reveal that fire is integral to the Earth system (Figure 2), both as part of the global carbon cycle (Ohlson et al., 2009) and as an ecosystem process that has influenced the development, dynamics, and configurations of disparate ecosystems worldwide (Bond et al., 2005; Flannigan et al., 2009; Power et al., 2008).

Fire has also played an important role in human development because it has been used for a variety of purposes, such as lighting, heating, and cooking, throughout all world cultures. More recently, since the end of the last Ice Age, fire activity is known to have changed through time in response to human action. For example, ancient peoples in Eurasia used fire to help clear virgin forests in order to create vast agricultural lands (Odgaard and Rasmussen, 2000; Ruddiman, 2003). Today, fire is heavily managed, with potentially troublesome wildfires actively suppressed when possible (Pyne, 1984). At the same time, prescribed fire is also widely used as a tool in forest management, where it is variously applied in hazard reduction, silviculture, habitat enhancement, range burning, insect and disease control, and conservation–restoration of natural systems (Weber and Taylor, 1992).

Study of Charred Particles

Naturalists in the nineteenth century recognized that charcoal was persistent in the natural environment after observing its occurrence in both soil and sediments. However, it was not until the early mid-twentieth century that charcoal was used as a paleoenvironmental proxy for fire disturbance (Iversen, 1941). Thereafter, the use of charcoal in describing and understanding the functioning of Quaternary environments remained limited until the 1970s. At that point, paleoecologists began developing methods to study different aspects of fire and vegetation history. Since the mid-1990s, charcoal records have been widely used as key indicators of natural fire disturbance (Figure 3). In addition, the records have also been used to examine the historical use of fire by people.

Following the pioneering work of Wright (1974), the measurement of charcoal in Quaternary sediment was mainly developed and applied in North America during the early 1970s. Wright contended that past fire could potentially be identified as a key ecological factor influencing both ecosystem changes, through vegetation and soil dynamics, as well as biogeochemical cycling. However, the main development of sedimentary charcoal studies occurred during the late 1980s and early 1990s under the leadership of Clark (1988a,b, 1990), who, as a former student of Wright, developed and advanced specific concepts and methodologies to reconstruct fire history, including continuous high-resolution charcoal studies.

Since then, the reconstruction approach has been further refined through the introduction of quantitative methodologies (Gavin et al., 2006; Higuera et al., 2010; Long et al., 1998), coupled with development of software specifically designed to analyze charcoal records, including both Charster and K1D developed by D.G. Gavin, University of Oregon, and

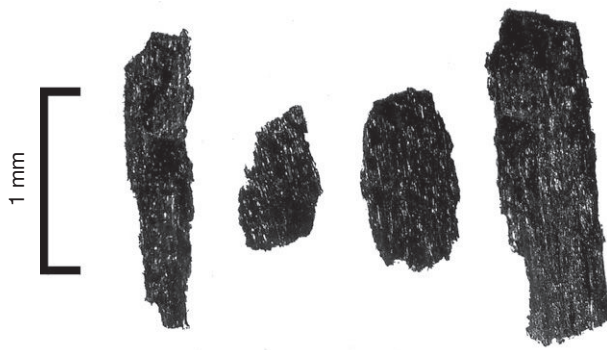


Figure 1 Charcoal fragments extracted from a sediment core that was collected from Scum Lake, British Columbia, Canada. In general, fragments are classified as charcoal if they are black, opaque with submetallic luster, and exhibit cellular structure (Sander and Gee, 1990). Photographs by K. Brown and N. Hebda.

CharAnalysis (Higuera et al., 2010). Another important advance relates to the development of the Global Charcoal Database (GCD; Power et al., 2008), which has been used to generate regional to global fire disturbance syntheses that have improved our collective understanding of the drivers of fire disturbance.

Longer Term Perspectives

In recent years, the public has become increasingly aware of forest fires, largely due to the fact that there has been an overall global increase in the frequency and magnitude of fire combined with the expansion of human settlements into fire-prone environments. Future changes in fire activity induced by climate (Flannigan et al., 2009) will likely further exacerbate public concern about increasing wildfire along the urban-wildland interface. The overall increase in global fire disturbance during the last two decades may be related to several processes, including increased fuel loads associated with decades of fire suppression as well as changing climate (Gavin et al., 2007; IPCC, 2007; Pitman et al., 2007). Indeed, the general interaction between climate, vegetation, and fire is known to be complex and multifaceted, although climatic factors are considered to be the predominant controls regulating the fire regime (Marlon et al., 2009; Power et al., 2008; Swetnam and Betancourt, 1998). However, vegetation is another important factor that regulates fire by controlling the configuration and flammability of fuels (Bond and Keeley, 2005; Brown et al., 2005; Hu et al., 2006) and by altering local microclimatic conditions (Ohlson et al., 2011). In turn, fire is capable of modifying both climate and the surrounding vegetation. The climate system can be influenced by fire through the emission of greenhouse gases and carbonaceous aerosols (Scholes and Andreae, 2000), whereas forest configuration may be mediated by altering species turnover rates or by selecting for fire-adapted species (Gavin et al., 2007; Overpeck et al., 1990).

Given that recent land management practices have dramatically altered natural fire regimes in many parts of the world,

long-term records provide a context for understanding modern fire regimes (Gavin et al., 2007). To understand the full importance of the factors and processes that regulate fire regimes, scientists, forest managers, policymakers, and other officials need greater insight into the long-term dynamics and variability of fire. There are several approaches through which longer term perspectives can be obtained, including assessment of historical records, fire-scarred tree rings, stand age distributions, and charcoal records. In general, the former approaches provide insight into fire disturbance at subannual to centennial scales, whereas charcoal records can provide subdecadal to millennial perspectives. Clearly, one advantage of charcoal records is that they do not attenuate through time as do tree ring records. Charcoal records may be obtained from lake sediment sequences, peat deposits, small hollows, alluvial deposits, and soils, as well as from archaeological sites where they provide insight into human use of fire.

Stratigraphic Charcoal Records

Through a variety of transport mechanisms and taphonomic processes including convective dispersal as well as postfire wind transport, overland flow, and sediment focusing (Clark, 1988a; Higuera et al., 2007), charcoal particles may enter nearby depositional environments such as lakes or wetlands, where they preserve (Figure 4). With time, charcoal from subsequent fires accumulates within these various environments, eventually forming long-term records of disturbance history. Often, small deep lakes with no inflowing streams or surrounding peatlands are selected as the best sites for Quaternary fire studies because they permit local fire history reconstruction and reduce the amount of ambiguity associated with charcoal source areas from larger basins or those with inflowing streams, which can deliver additional charcoal from the watershed. Records of past fire activity can be recovered using a variety of sampling approaches that are designed to collect continuous sediment sequences (Figure 4). Several different samplers are suitable for the recovery of sediment, with perhaps the Livingstone piston corer (Wright, 1967) being the most well known. The stratigraphic records are typically logged or photographed prior to additional sampling, which often involves extracting known volumes of sediment in contiguous samples, ideally in increments of 1 cm or less. Clark (1988a) and Higuera et al. (2007) indicate that the optimal sampling interval for detecting individual fires should be <0.2 and <0.12 times the mean fire return interval, respectively.

Historically, several techniques have been used to estimate charcoal quantity (Rhodes, 1998), including pollen slide, thin-section, and sieving estimates. In the pollen slide approach, a sediment sample is processed for pollen analysis, which involves chemical digestion (see Pollen Analysis, Principles). The resulting residue is mounted on a glass slide, protected with a cover slip, and examined using a compound light microscope at 400–1000 $\times$ magnification. In addition to pollen, small charcoal fragments will often survive the processing and be found on the slide, and these fragments can be tallied as particles or areas. In general, fragments on pollen slides can be classified as charcoal if they are black, opaque, and tabular or angular in shape (Rhodes, 1998). Wood anatomical structure is generally

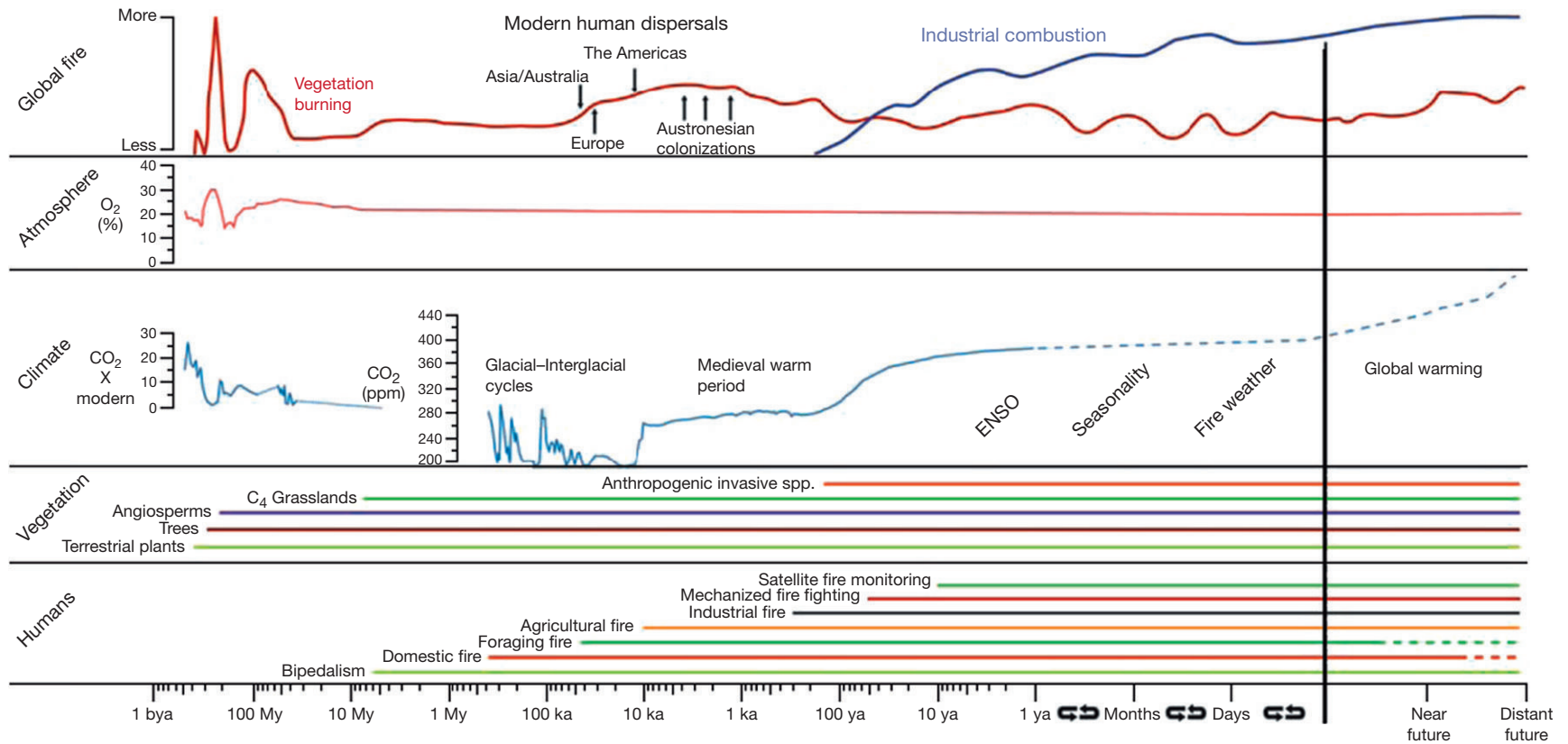


Figure 2 Temporal schematic of global fire activity derived using stratigraphic charcoal deposits relative to other aspects of earth history, including evolution of the atmosphere, climate system, and vegetation. The evolution of the genus *Homo* and subsequent human-related fire activities are likewise illustrated. Note that the time scale is logarithmic. Reproduced from Bowman DMJS, Balch JK, Artaxo P, et al. (2009) Fire in the Earth system. *Science* 324: 481–484.

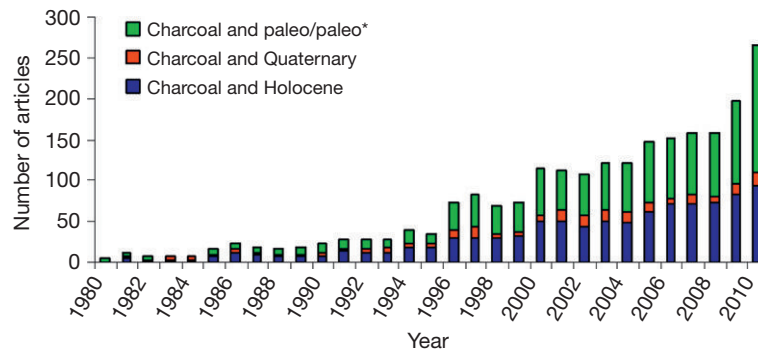


Figure 3 Evolution of the number of articles archived in the bibliographic database Scopus using the keywords charcoal and Holocene, charcoal and Quaternary, and charcoal and paleo/paleo*. The figure does not contain all Quaternary articles dealing with charcoal because the searches focused only on the title–abstract–keyword fields for the years 1980–2010.

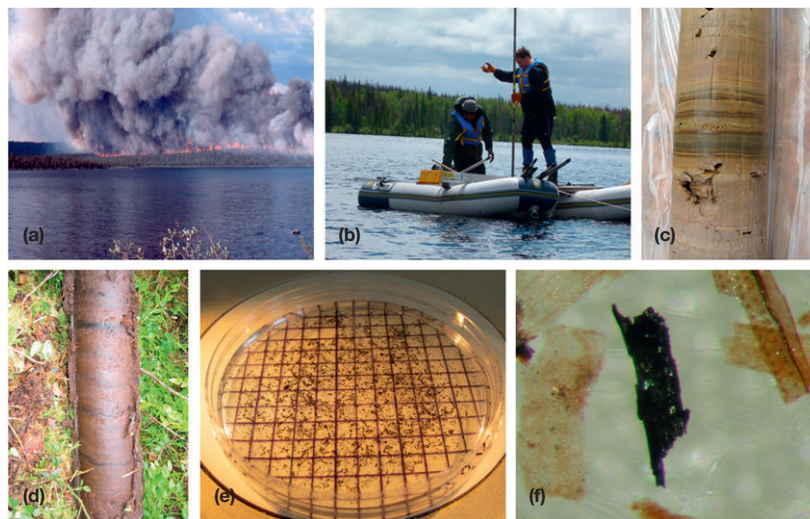


Figure 4 A photographic series showing the production, collection, and tabulation of charcoal. (a) Smoke plume from a forest fire, illustrating how charcoal can be transported to lake depositional environments. Photograph from Canadian Forest Service. (b) Recovering a sediment record using a Livingstone corer. Photograph by R. McKay. (c) Example of a lake sediment core showing laminations. Charcoal fragments are preserved in the sediment and will be extracted in the laboratory. (d) Peat core showing possible charcoal bands. Photograph by R. Bradshaw. (e) Sieved sediment, including charcoal fragments, in a gridded Petri dish. (f) Charcoal fragment, approximately 1 mm long.

difficult to discern in pollen slide charcoal since the light source is from below the sample. The pollen slide approach is attractive because it combines the tabulation of pollen with that of charcoal. A disadvantage of the approach is that charcoal fragments may be chemically or physically degraded during preparation. Another methodological challenge relates to charcoal identification, whereby the definitive classification of a fragment as charcoal can be difficult, given the general small size of the fragments coupled with the persistence of other dark organic material on the slides. In the petrographic thin-section approach, the recovered sedimentary archive is dehydrated under a vacuum, impregnated with epoxy resin, cured, and sectioned (Clark, 1988b). This technique, however, is generally considered both time-consuming and expensive and has not been widely adopted. In contrast, the sieving approach is both simple and relatively quick. Initially, sediment samples are soaked in a deflocculating agent such as 5–10% sodium metaphosphate solution, followed

by gentle sieving through a mesh. The diameter of the mesh openings may vary, although 125–180- μ m mesh diameters are commonly used. The residue caught on the sieve is suspended in water in a gridded Petri dish and examined using a stereomicroscope at 20–40 $\times$ magnification (Figure 4). Charcoal identification can be made using the same criteria as in the pollen slide samples, though in this case, anatomical structure is readily visible and the fragments are often lustrous. Both particle counts and area determinations can be made. Regarding the latter, areas can be estimated using an eyepiece graticule or measured more precisely using specialized imaging software (Lynch et al., 2002). Empirical studies have shown a strong correlation between count and area measures of charcoal abundance (Figure 5; Ali et al., 2009). In addition to providing an estimate of charcoal quantity, another advantage of the sieving approach is that other plant macrofossils present in the Petri dish can be simultaneously tabulated.

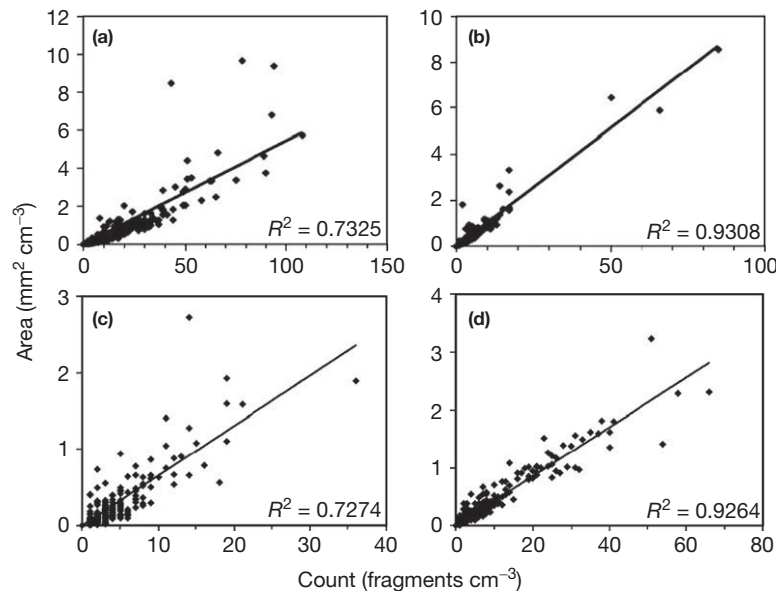


Figure 5 Correlation between charcoal fragment and area concentrations in lake sediments. (a) Data from Lac Pas-de-Fond, Quebec ($N=354$); (b) data from Lake Raigejeppe pond, Sweden ($N=245$); (c) data from Lake Klotärnen, Sweden ($N=322$); (d) data from Sarup Sø, Denmark ($N=605$). Although charcoal concentration, expressed in area, volume, or mass, is linked to burned biomass, fragment numbers are not because they undergo breakage during transportation and deposition in nature or during sieving at the laboratory. However, the two descriptors (number and area) are positively correlated. Panels (a) and (b) are reproduced from Carcaillet C (2007) Charred particle analysis. In: Elias SA (ed.) *Encyclopedia of Quaternary Science*, pp. 1582–1593. Amsterdam: Elsevier. The data used in panels (c) and (d) were provided by K. Brown.

From a source area perspective, charcoal recorded from pollen slides is considered a proxy for both local and regional burning. Given its small size (usually $<150\ \mu\text{m}$), pollen slide charcoal is often referred to as microcharcoal (Mooney and Tinner, 2011). Charcoal in this size range is capable of being convected long distances (1–100 km; Clark, 1988a; Peters and Higuera, 2007). In contrast, charcoal isolated with the sieving approach is often referred to as macrocharcoal. Because the fragments are larger (typically $>125\ \mu\text{m}$), the transport distances are generally shorter. Thus, macrocharcoal serves as a proxy for stand- to local-scale fire disturbance where local reflects fires within about 1 km of the basin (Gavin et al., 2003; Higuera et al., 2007). Initial measurements from prescribed burn experiments suggest that most macroscopic charcoal is deposited near the burn edge (Figure 6; Clark et al., 1998; Lynch et al., 2004; Ohlson and Tryterud, 2000). More recent observations from wildland fires (Pisaric, 2002; Tinner et al., 2006), coupled with the development of a particle distribution model (Higuera et al., 2007; Peters and Higuera, 2007), show that macrocharcoal fragments have a potential to be transported several to tens of kilometers from source (Figure 6). Critically, however, this increased transport distance does not preclude identification of local fires though peak charcoal detection in lake sediment records.

In addition to these two approaches, several other sediment-based fire reconstruction approaches are being developed. Though relatively nascent in paleofire reconstruction, chemical groups such as monosaccharide anhydrides, light carboxylic acids, polycyclic aromatic hydrocarbons, and lignin burning products have potential to serve as proxies for past fire. Within the monosaccharide anhydride group, Holocene records of biomass burning have been developed using the

molecular marker levoglucosan (Elias et al., 2001). Levoglucosan is a monosaccharide derivative formed from the thermal breakdown of cellulose during combustion, and its concentration can be measured in any residue of fuel that contained cellulose (Louchouart et al., 2009). It is produced in large quantities, transported in smoke plumes, and deposited by both wet and dry deposition mechanisms (Kerhwalid et al., 2010). Another interesting application in charcoal research uses reflected light microscopy to estimate the temperature of charcoal formation (McParland et al., 2007). This method has implications for both understanding combustion temperatures of natural wildfires and for understanding the human use of wood and charcoal as a fuel source.

Analysis of Sediment Charcoal Records

Given that the sieved charcoal method now constitutes the most widely used and recommended approach in paleofire reconstruction, the method will be examined in greater detail (Figure 7; Higuera et al., 2010). Following sieving and quantification, the estimates of charcoal are expressed as concentrations (Figure 8), either as counts (fragments cm^{-3}) or areas ($\text{mm}^2\text{cm}^{-3}$). The charcoal accumulation rate (particles $\text{cm}^{-2}\text{year}^{-1}$ or $\text{mm}^2\text{cm}^{-2}\text{year}^{-1}$; CHAR), alternatively referred to as charcoal flux, is determined by multiplying the charcoal concentration by vertical accretion rate (cm year^{-1}), where the latter rate is usually determined with the aid of an age-depth model that is derived using radiometric dating of selected samples and any other available chronological indicators. CHAR is often the preferred metric because it accounts for changes in accretion rate through time that can either

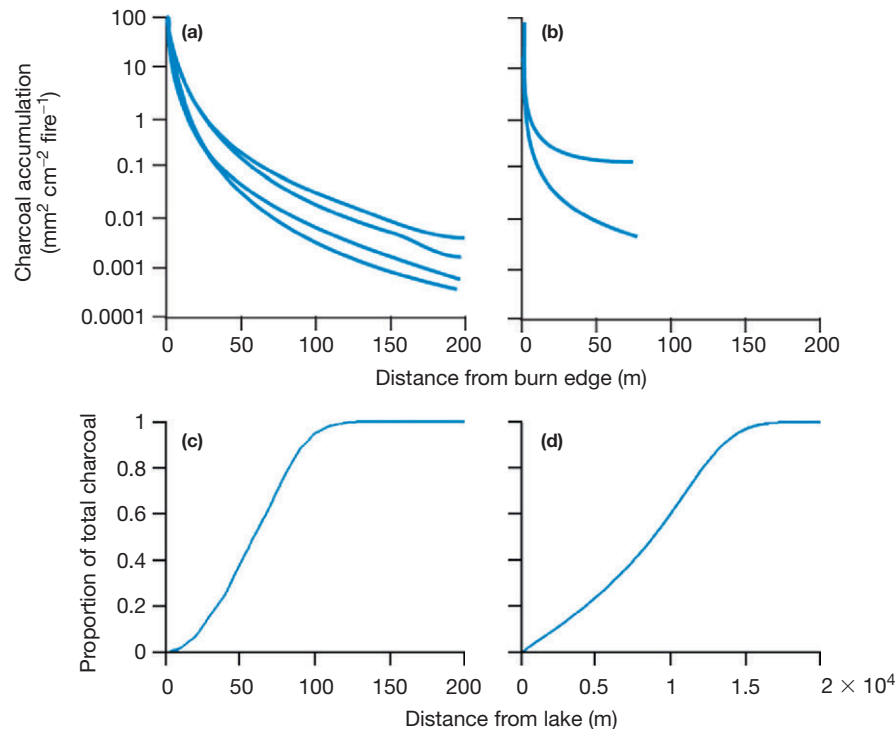


Figure 6 Pattern of charcoal accumulation rates with distance from burned forests. (a) The International Crown Fire Modeling Experiment was performed in *Picea mariana* (black spruce) forest, Northwest Territories, Canada, and (b) the Bor Island experiment fire in *Pinus sylvestris* (Scots pine) forest, Siberia. Charcoal accumulation declines rapidly with distance from the burn edge, with most charcoal deposited within 60 m of the burned edge (reproduced from Lynch JA, Clark JS, and Stocks BJ (2004) Charcoal production, dispersal, and deposition from the Fort Province experimental fire: Interpreting fire regimes from charcoal records in boreal forests. *Canadian Journal of Forest Research* 34: 1642–1656). Particle dispersal model output showing cumulative charcoal deposition in a lake with increasing distance for (c) 10 m and (d) 1000 m modal injection height scenarios. Reproduced from Peters ME and Higuera PE (2007) Quantifying the source area of macroscopic charcoal with a particle dispersal model. *Quaternary Research* 67: 304–310.

concentrate or dilute the quantity of charcoal particles, enabling sample-to-sample comparison. CHAR samples are typically interpolated to a constant temporal resolution to accommodate uneven sampling intervals caused by downcore variations in accretion rate. The interpolated CHAR record consists of discrete peaks that are superimposed on a slowly varying mean (Higuera et al., 2010), with the former reflecting the peaks component and the latter the background.

Interpolated CHAR records can then be decomposed into both background and peaks components (Figures 7 and 8). The background component reflects the low-frequency portion of the CHAR series that varies slowly through time in response to changes in charcoal production and transport, as well as sediment mixing and sampling. It can be modeled using a curve fitting algorithm (Briles et al., 2008; Higuera et al., 2010). Removal of the background component from the CHAR series, either through subtraction (residual) or division (ratio), results in the detection of the high-frequency peak component. The peak component is comprised of both fire-induced signal and non-fire-related noise, with the latter related to distant fires, analytical noise, and charcoal redeposition (Hallett and Anderson, 2010). Consequently, a threshold can be applied to the peak component in order to separate signal from noise. A Gaussian mixture model (Gavin et al., 2006) can be used to identify the noise, with a specified percentile value (i.e., the 99th percentile) of the peak component

distribution used as a separation threshold that distinguishes signal from noise. In rare instances when background charcoal does not vary through the record, a global threshold can be determined using the peak distribution from the entire record. Alternatively, local thresholds, robust to changes in background charcoal, peak variance, and the detrending model, can likewise be generated for each sample in a record by sampling the peak distribution within a moving window (Higuera et al., 2010). Peaks that are identified as being derived from local fires are referred to as fire episodes because they cannot unambiguously be related to a single event and may owe their origin to more than one fire that burned near the basin. Importantly, a minimum count test can be used to evaluate whether the peaks identified result from a statistically insignificant change in charcoal, and if so, then the peak can be rejected. The time difference between peaks reflects the fire return interval. Estimates of fire frequency can be obtained by summing the peak series over a specified window width and smoothing. In addition to this technique, other approaches can be used to estimate temporal trends in fire frequency. For example, Clark et al. (2002) and Brown et al. (2005) used Fourier power spectral analysis and wavelets to examine frequencies and periodicities in grassland charcoal records.

Critically, CHAR can vary greatly from location to location because each study site possesses a unique set of characteristics related to the size and type of basin, physiography of the

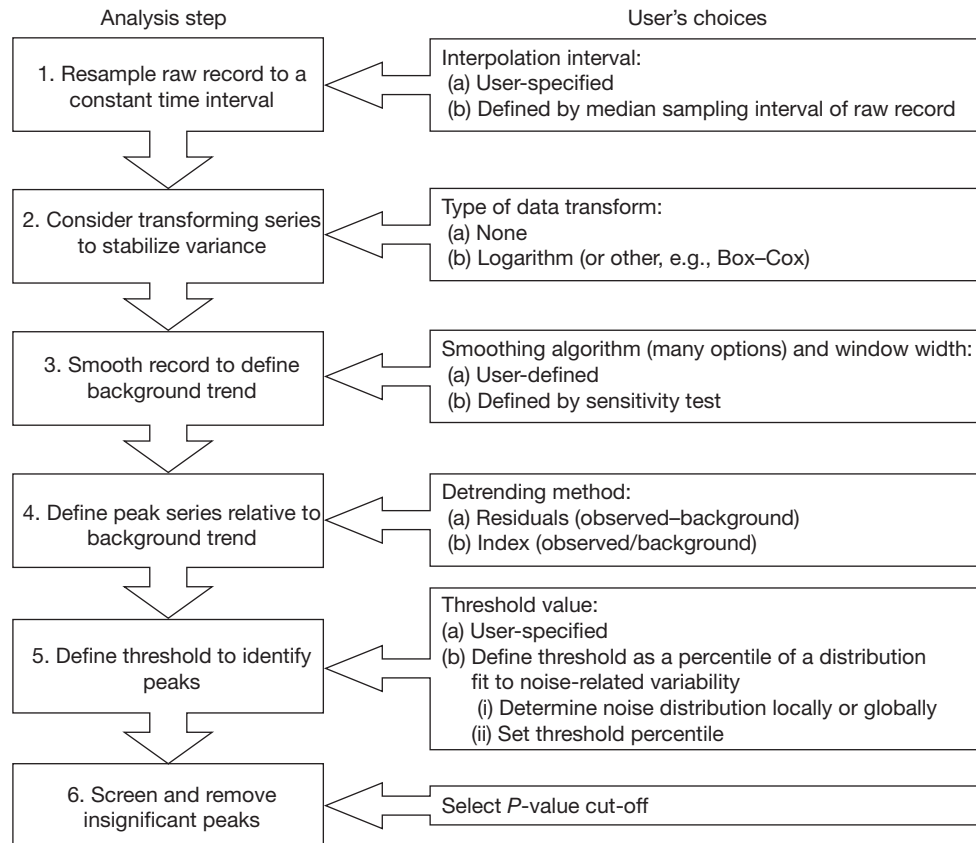


Figure 7 Flow diagram showing the steps and choices required during charcoal decomposition. Reproduced from Higuera PE, Gavin DG, Bartlein PJ, and Hallett DJ (2010) Peak detection in sediment-charcoal records: Impacts of alternative data analysis methods on fire history interpretations. *International Journal of Wildland Fire* 19: 996–1014. Copyright © International Association of Wildland Fire (USA) 2010.

catchment area, accretion rates, vegetation types, landscape structure, and fire regime. Because of this inherent spatial variability, single records yield site-specific reconstructions sensitive to watershed-scale processes, whereas collated records that combine multiple fire history reconstructions allow for larger spatial syntheses and insights into the regional climate controls of fire (Carcaillet and Richard, 2000; Power et al., 2008).

Soil Charcoal

In a forest fire, much of the charcoal produced falls within the burned area (Ohlson and Tryterud, 2000) and is gradually incorporated into the soil by bioturbation (Carcaillet, 2001). In addition to mixing by biological activities, climatic processes and tree uprooting by windthrow can mix soil, contributing to a general lack of stratification. Consequently, soil charcoal is of limited use for Quaternary reconstruction because of compromised stratigraphic integrity, though the age of single charcoal fragments can be measured by accelerator mass spectrometry (AMS) dating to establish time of fire, with consideration given to the inbuilt age of the wood (Gavin et al., 2003). From a methodological perspective, analysis of soil charcoal involves the collection and sampling of soil profiles, with charcoal

separated from other soil particles by flotation or sieving and examined using a stereomicroscope. If taxonomic identification is desired, individual charcoal fragments can be compared to descriptions from specialized literature (e.g., Neumann et al., 2001) or regionally explicit reference collections.

Soil charcoal archives have several unique applications. For example, they provide records from areas where depositional basins may be rare (Hopkins et al., 1993) and can be used to examine general changes in forest composition and fire disturbance at the landscape scale (Ohlson et al., 2011). In addition, they are also useful in determining the past position of treeline (Carneli et al., 2004). Soil charcoal records can also be used to examine the spatial variability of fire resulting from landscape structure and human activity (Gavin et al., 2003). For instance, it has been observed that on the west coast of Vancouver Island, Canada, topography is an important factor affecting the distribution of fire disturbance, with south-facing slopes exhibiting a greater incidence of fire compared to north-facing slopes. Gavin (2003) also used soil charcoal records to examine trends in soil disturbance, which are typically too long to measure using plot-based studies and thus are poorly understood. Finally, by measuring the carbon concentration in soil charcoal, it is possible to report on soil charcoal carbon pools (Ohlson et al., 2009), which are important for determining carbon budgets (Table 1).

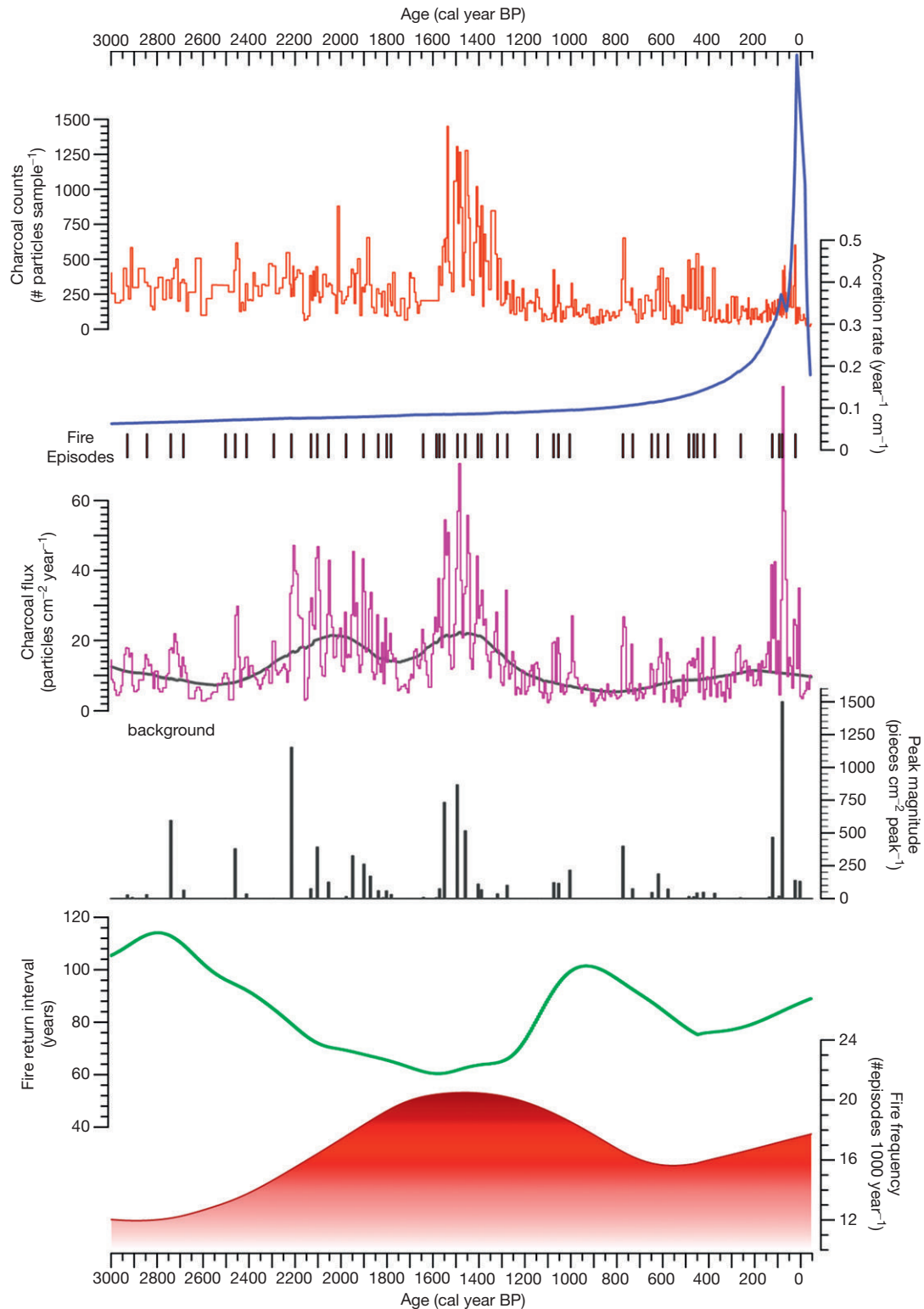


Figure 8 Decomposition of a 3000-year-long sedimentary charcoal record from Foy Lake, Montana, USA. Charcoal counts (red) were converted to charcoal accumulation rates ($\text{CHAR} = \text{particles cm}^{-2} \text{year}^{-1}$) (magenta) by multiplying the charcoal concentration by accretion rate (blue). CHAR values were interpolated to 5-year intervals and smoothed using a 500-year window width (background). Fire episodes are indicated by a vertical (red) symbol when CHAR exceeds background by calculating residuals and using a locally defined threshold and Gaussian mixture model. Peak magnitude illustrates the size of the charcoal peak above background, and the smoothed fire return interval (green) illustrates the years per fire. The bottom panel shows smoothed fire frequency using a 1000-year window to illustrate the change in peak frequency through time. The data used in this figure were provided by M. Power.

Table 1 Range of soil charcoal quantities in various vegetation types and locations

<i>Vegetation type and location</i>	<i>Quantity (kg charcoal ha⁻¹)</i>	<i>Reference</i>
Deciduous forest (Canada)	350–1300	Bélanger et al. (2004)
Boreal forest (Sweden)	1000–2100	Zackrisson et al. (1996)
Boreal forest (Norway and Sweden)	0–2200	Ohlson et al. (2009)
Alpine grassland (France)	0–200	Carcaillet and Talon (2001)
Sub-Alpine forest (France)	5–20 000	Carcaillet and Talon (2001)
Mediterranean forest (France)	8500–100 000	Carcaillet et al. (1997)
Savanna (Africa)	150	Carcaillet and Talon (2001)

Archaeological Charcoal

Charcoal from archaeological sites can be used to reconstruct past environments and ascertain how prehistoric populations may have used or exploited fire (Heinz and Thiébault, 1998; February, 2000; see [Archaeological Applications](#)). Charcoal sampled in archaeological sites largely results from the charring of wood used as fuel, building material, or material for the manufacture of domestic implements. Furthermore, the burning of dwellings can often be associated with site abandonment, yielding large amounts of other archaeological charred material such as seeds and bones. This material provides additional information about the precise use of wood by distinct human populations or societies. Consequently, archaeological charcoal can be used to examine past lifeways and the relationship that humans have had with their environment (Chabal, 1997). Laboratory methods for analyzing archaeological charcoal are similar to those used in pedoanthracology, although the size of identified fragments is generally larger in archaeology (>2mm).

As an example, analysis of charcoal composition and tree ring structure from Mesolithic (early Holocene) cooking pits in northern Sweden reveals that large pieces of pine (*Pinus*) wood were often selected as fuel, even though other types of vegetation were available (Carcaillet, 2007). Critically, this restrictive charcoal assemblage pattern suggests that Mesolithic hunters preferred pine as fuel, likely because it was abundant in the environment and well suited for combustion. In another European example, archaeological charcoal was used to infer the type of fuel selected for charcoal production in charcoal kiln soils (Ludemann and Nelle, 2002). In certain regions of Europe, charcoal kilns abound, demonstrating the importance of charcoal production in the recent past where it was used in industrial activities such as glass making. Charcoal is lighter than wood but with the same calorific benefit, so the production of charcoal saved both energy and effort because the transportation of wood from forest to town was avoided.

In North America, lacustrine and archaeological records have both been used to examine Aboriginal and European burning practices (Clark and Royall, 1996; Turner, 1991), which vary spatially. For example, in coastal western Canada, archaeological findings suggest that large sophisticated Aboriginal populations with permanent settlements developed during the last 2000 years (Brown and Hebda, 2002; Hebda and Mathewes, 1984) and used fire both domestically and as a landscape management tool. Thereafter, a switch from Aboriginal burning to European forest clearing and

utilization is shown by a marked change in charcoal deposition across the pre- to post-European settlement transition (McDadi and Hebda, 2008). In another example from Central America, archaeological and paleoecological research suggests significant areas of Neotropical forests were cleared and burned to facilitate agricultural activities before the arrival of Europeans (Dull et al., 2010). Recent synthesis has established that a major reduction in biomass burning occurred ~500–600 years ago, as a result of both climatic (Little Ice Age) and anthropogenic forcing (pandemics; Power et al., in press; Ruddiman, 2003; Ruddiman et al., 2011). As more archaeological and paleoclimate information become available for the region, the drivers of changing charcoal abundance in the Neotropics will become clearer.

Global Charcoal Database

The Global Charcoal Database (GCD) is an emerging tool for mapping and characterizing changing fire regimes on regional to global scales through time (Power et al., 2009). The database is unrestricted and publically accessible (see [‘Relevant Websites’](#)). In total, 405 sedimentary charcoal records, spanning over 200 000 years of Earth history are featured in the GCD (Figure 9). The primary function of the database is to serve as a tool for testing hypotheses about the changing drivers of fire regime on centennial to millennial timescales (see [Databases and Their Application](#)). Within the database, charcoal quantities, sample volumes, depths, and modeled ages are included, along with appropriate metadata for parsing individual sites based on unique characters such as vegetation type or size of watershed. The radiometric dating approaches (AMS or conventional dating), measured ages, and calibrated ages for all age models are likewise included. The creation of the database also represents the first step in providing the necessary empirical data to test the validity of vegetation and fire models that simulate past and future ecosystems under markedly different biological and physical conditions from present. Currently, the GCD is being used to produce regional and global syntheses that examine the linkages between Quaternary climate variability and biomass burning (Marlon et al., 2009; Mooney et al., 2011; Power et al., 2008; Vannière et al., 2011).

To facilitate syntheses using GCD data, a standardization approach needs to be used (Figure 10) that normalizes individual records and stabilizes the variance so that charcoal data can be compared from hundreds of sites that employ a wide range of sampling, processing, and quantification techniques

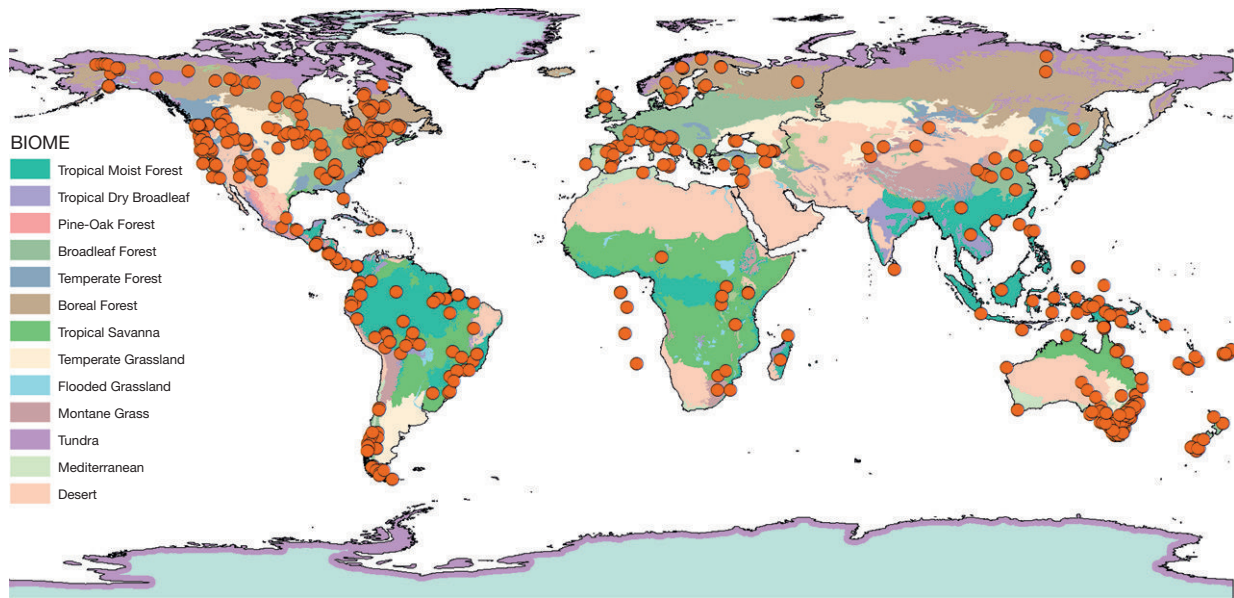


Figure 9 Locations of available records in the Global Charcoal Database (red dots) relative to the distribution of global biomes (World Wildlife Federation).

(Power et al., 2008, 2009). Composite records for regional analysis are created by smoothing standardized charcoal time series (Marlon et al., 2009). Individual records are first sampled and smoothed (Cleveland and Devlin, 1988). Once the individual records have been smoothed, a three-step standardization protocol is used to facilitate comparisons between sites and through time. The standardization procedure involves rescaling values using a minimax transformation, homogenizing the variance using a Box-Cox transformation (Figure 10) and rescaling values once more to Z-scores using an appropriate base period (e.g., the last 2000 years) to achieve a common mean and variance. Using this methodology, several papers have highlighted key findings in regional biomass burning that demonstrate strong linkages between fire and past climate (e.g., Marlon et al., 2009; Mooney et al., 2011; Power et al., in press; Vannière et al., 2011).

There are relatively few global charcoal records available for documenting the Last Glacial Maximum (LGM), but the number of sites doubles for the interval spanning the last 10 000 calendar years before present (cal BP). These data, when standardized and combined, document broad-scale changes in charcoal accumulation through time and are interpreted as reflecting regional changes in biomass burning. The dominant patterns of biomass burning have been discussed in Power et al. (2008) and Marlon et al. (2009) for broad geographic regions, within the context of millennial-to-centennial scale trends in fire and climate. These studies have concluded that climate forcing, both directly and indirectly, has had a major impact on global biomass burning since glacial times. The most important factors influencing regional climate and biomass burning since the LGM are (1) the seasonal and latitudinal distribution of insolation, (2) disappearance of the Northern Hemisphere ice sheets, (3) changes in paleogeography related to isostatic rebound and rising sea levels affecting the degree of continentality, (4) changes to ice caps in the

tropics and Southern Hemisphere, (5) changes in global sea surface temperatures, and (6) changes in atmospheric composition. These aspects of the global climate system have been directly and/or indirectly linked to changes in regional fire activity through regional-scale atmospheric circulation anomalies and concomitant changes to the incidence of fire-supporting climate regimes.

Reconstructions of global biomass burning since the LGM suggest that overall fire disturbance was generally low during the LGM and remained low during the glacial and late-glacial interval from 21 000 to ca. 14 000 cal BP (Power et al., 2008). At this time, global climate was colder and drier, atmospheric CO₂ levels lower, and terrestrial biomass significantly reduced relative to present. Given the cool, dry nature of the troposphere, both convection and lightning activity would have been limited. This factor, combined with the noted general reduction of fuel load, may have contributed to the prolonged period of low global fire activity. As Northern Hemisphere ice sheets waned and atmospheric CO₂ concentrations increased, terrestrial vegetation expanded across the high latitudes, and global fire activity generally increased over the last ~10 000 years, with an exception during the last two millennia. For example, in the early Holocene, sea surface and terrestrial temperatures increased in the Northern Hemisphere in response to changes in insolation. At the same time, vegetation cover expanded, and fire became an increasingly dominant process within the Earth system. However, global biomass burning has declined during the last two millennia. Embedded within that long-term trend are centennial-scale increases and decreases in fire associated with temperature variability such as the Medieval Climate Anomaly (Stine, 1994) and Little Ice Age, respectively. The most obvious human impacts on global fire regimes have occurred since the Industrial Revolution, with dramatic increases in fire activity after AD ~1750 because of land clearance and decreases after AD ~1850 as a result of forest management (Marlon et al., 2012).

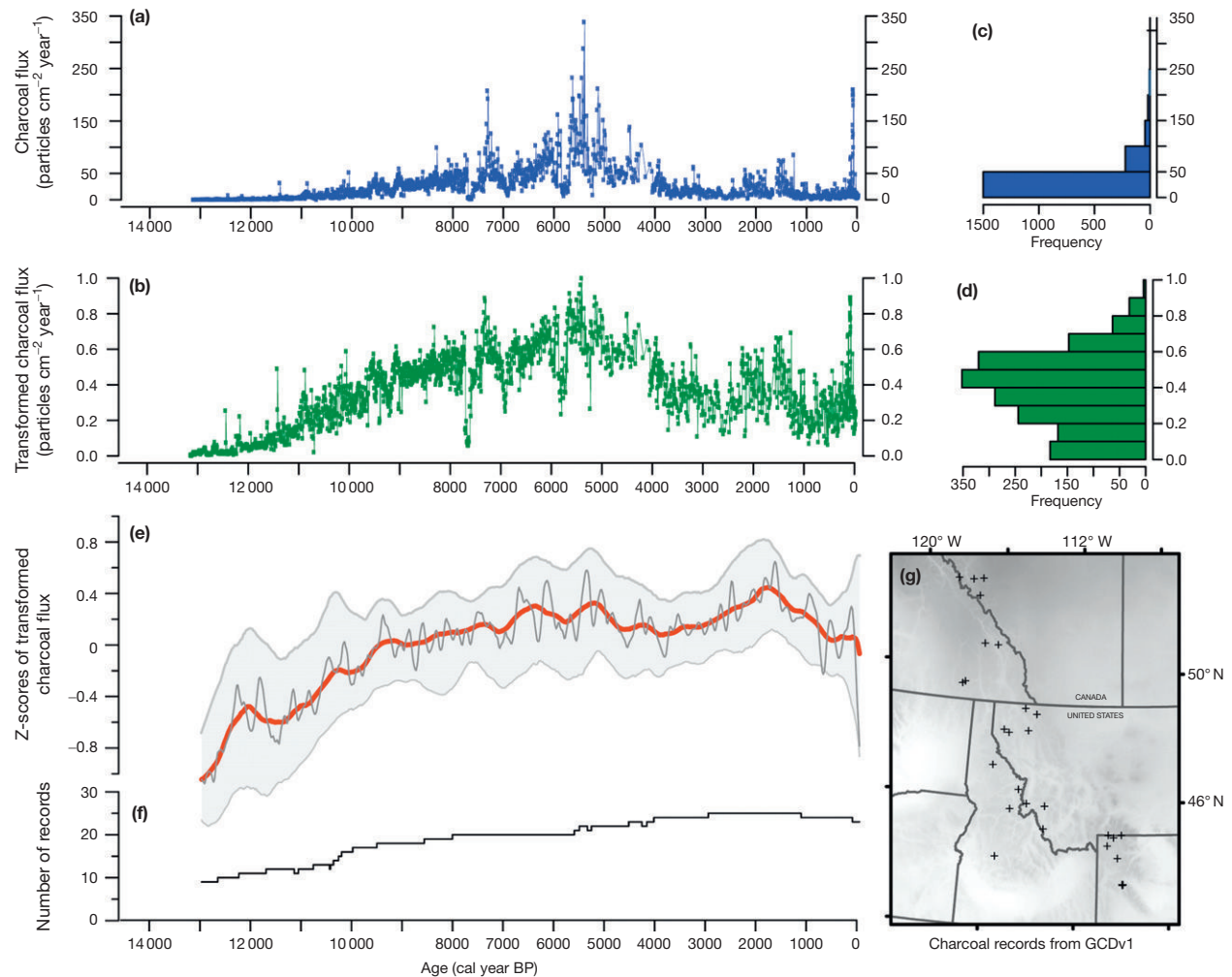


Figure 10 A comparison of (a) original and (b) transformed flux data (particles cm⁻² year⁻¹) from Foy Lake, Montana (Power et al., 2011), demonstrating the impact of the Box-Cox transformation. Comparing the distribution of the (c) original and (d) transformed data as frequency distributions illustrates how the optimal value of λ (e.g., 0.0311 in this example) normalized the distribution of the data. (e) A 13000-year regional composite that was created from 26 transformed charcoal records from the northern Rocky Mountains, expressed as Z-scores of charcoal anomalies. In this case, the entire records were used to calculate the base period for the Z-scores. All charcoal samples were first transformed and then smoothed using a 500-year window (red line). The dark gray line illustrates a 100-year smoothing window, and the upper and lower 95% confidence intervals are shown in light gray. (f) The numbers of records contributing to the 500-year smoothed time series. (g) Locations of the charcoal records used in the composite. The data used in this figure were provided by M. Power.

Future Directions

Although significant progress has been made in recent decades regarding use of charcoal in Quaternary studies, much work remains. From a spatial perspective, the GCD reveals significant gaps in the distribution of charcoal records (Figure 9). Increased sampling is required in regions that are currently underrepresented, including Russia, eastern Europe, middle Asia, and Africa. Furthermore, it would be desirable, though perhaps not practical, to resample and analyze some previously published records that are poorly resolved. From a methodological perspective, recovered sedimentary sequences should ideally be sampled contiguously and at high resolution. The recent advances in the decomposition of charcoal time series for fire return interval and frequency estimates have resulted in a surge of research (Figure 3). However, detailed

understanding of the peak magnitude component remains elusive and needs to be investigated further. This will require additional calibration studies that relate peak magnitudes to burn areas, fire intensities, and charcoal delivery. Furthermore, many of the recent methodological advances were developed using charcoal records from conifer-dominated ecosystems that often experience infrequent, stand-replacing fires. An evaluation of these advances is now required using data from different vegetation types to assess the universality of the methods across ecosystems. For example, in fires in Mediterranean ecosystems, the wood is often mineralized from high combustion temperature as a result of volatile compounds such as terpenes and low residual humidity in the vegetation. Consequently, wood charcoal production tends to be low, whereas ash production may be high relative to boreal and temperate regions. In addition, other methods of fire

reconstruction such as molecular and reflected light microscopy approaches should be further developed and evaluated. Finally, more data model comparisons are needed, with a view towards improving our understanding of the mechanisms that cause fire regime shifts (Brubaker et al., 2009).

Reconstructing and modeling carbon flux and storage at landscape and global scales should be further explored using charcoal databases. However, certain parameters need to be highlighted, such as the rate of carbon emission during fire, with consideration given to different vegetation types. Notably, sedimentary and soil charcoal records provide a means for evaluating the historical role of fire in sequestering refractory carbon and should be used to measure charcoal carbon pools. Given the long-term nature of the sedimentary records, they could also be used in studies that focus on the recalcitrant nature of charcoal.

Paleofire reconstructions using the GCD provide an additional opportunity for hypothesis testing and validation of paleofire simulations that attempt to capture the linkages among climate, vegetation, and fire. Accurately simulating past patterns and mechanisms of global and regional fire activity is a critical step toward assessing whether future changes in climate and fire can be realistically projected. Simulations that capture the spatial and temporal patterns of future fire regimes will help landscape managers prepare for the affects of climate change and help governmental agencies decide on how best to manage and allocate fire-related resources. The GCD is one of the many new tools available for fire scientists to deepen our understanding of fire in the Earth system, and it should be further developed through increased data contributions, leading to the production of additional informative analyses.

Conclusions

Quaternary investigations using charcoal as a proxy or measure continue to increase (Figure 3). These investigations are not discipline-specific but rather cross-disciplinary and include paleoecology, fire ecology, and archaeology, and also soil, carbon, and climate studies. Within the last decade, advances in the numerical treatment of charcoal time series and the development of a GCD have resulted in improved characterization of past fire regimes and the drivers of fire regime change. Future studies should focus on the continued improvement and testing of quantitative methods, advancement of new reconstruction approaches, data model comparisons, and issues related to human use of fire, carbon cycling, and sampling of under-represented regions.

See also: Plant Macrofossil Introduction. **Plant Macrofossil Methods and Studies:** Surface Samples, Taphonomy, Representation; Treeline Studies. **Radiocarbon Dating:** ^{14}C of Plant Macrofossils; Charcoal.

References

- Ali AA, Higuera PE, Bergeron Y, and Carcaillet C (2009) Comparing fire-history interpretations based on area, number and estimated volume of macroscopic charcoal in lake sediments. *Quaternary Research* 72: 462–468.
- Bélanger N, Côté B, Fyles JW, et al. (2004) Forest regrowth as the controlling factor of soil nutrient availability 75 years after fire in a deciduous forest of southern Quebec. *Plant and Soil* 262: 363–372.
- Bond WJ and Keeley JE (2005) Fire as a global 'herbivore': The ecology and evolution of flammable ecosystems. *Trends in Ecology and Evolution* 20: 387–394.
- Bond WJ, Woodward FI, and Midgley GF (2005) The global distribution of ecosystems in a world without fire. *New Phytologist* 165: 525–538.
- Bowman DMJS, Balch JK, Artaxo P, et al. (2009) Fire in the Earth system. *Science* 324: 481–484.
- Briles CE, Whitlock C, Bartlein PJ, and Higuera P (2008) Regional and local controls on postglacial vegetation and fire in the Siskiyou Mountains, northern California, USA. *Palaeogeography, Palaeoclimatology, Palaeoecology* 265: 159–169.
- Brown KJ, Clark JS, Grimm EC, et al. (2005) Fire cycles in North American interior grasslands and their relation to prairie drought. *Proceedings of the National Academy of Sciences of the United States of America* 102: 8865–8870.
- Brown KJ and Hebda RJ (2002) Ancient fires on southern Vancouver Island, British Columbia, Canada: A change in causal mechanisms at about 2,000 ybp. *Environmental Archaeology* 7: 1–13.
- Brubaker LB, Higuera PE, Rupp TS, et al. (2009) Linking sediment-charcoal records and ecological modeling to understand causes of fire-regime change in boreal forests. *Ecology* 90: 1788–1801.
- Carcaillet C (2001) Are Holocene wood-charcoal fragments stratified in alpine and subalpine soils? Evidence from the Alps based on AMS ^{14}C dates. *The Holocene* 11: 231–242.
- Carcaillet C (2007) Charred particle analysis. In: Elias SA (ed.) *Encyclopedia of Quaternary Science*, pp. 1582–1593. Amsterdam: Elsevier.
- Carcaillet C, Barakat HN, Panaiotis C, and Loisel R (1997) Fire and late Holocene expansion of *Quercus ilex* and *Pinus pinaster* in Corsica. *Journal of Vegetation Science* 8: 85–94.
- Carcaillet C and Richard PJH (2000) Holocene changes in seasonal precipitation highlighted by fire incidence in eastern Canada. *Climate Dynamics* 16: 549–559.
- Carcaillet C and Talon B (2001) Soil carbon sequestration by Holocene fires inferred from soil charcoal in the dry French Alps. *Arctic, Antarctic, and Alpine Research* 33: 282–288.
- Carnelli AL, Theurillat J-P, Thimon M, et al. (2004) Past uppermost tree limit in the central European Alps (Switzerland) based on soil and soil charcoal. *The Holocene* 14: 393–405.
- Chabal L (1997) *Forêts et sociétés en Languedoc (Néolithique final, Antiquité tardive): l'anthracologie, méthode et paléocologie*, p. 192. Paris: Edition de la Maison des Sciences de l'Homme.
- Clark JS (1988a) Particle motion and the theory of charcoal analysis: Source area, transport, deposition, and sampling. *Quaternary Research* 30: 67–80.
- Clark JS (1988b) Stratigraphic charcoal analysis on petrographic thin sections: Application to fire history in northwestern Minnesota. *Quaternary Research* 30: 81–91.
- Clark JS (1990) Fire and climate change during the last 750 yr in northwestern Minnesota. *Ecological Monographs* 60: 135–159.
- Clark JS, Grimm EC, Donovan JJ, et al. (2002) Drought cycles and landscape responses to past aridity on prairies of the Northern Great Plains, USA. *Ecology* 83: 595–601.
- Clark JS, Lynch JA, Stocks BJ, and Goldammer JG (1998) Relationships between charcoal particles in the air and sediments in west-central Siberia. *The Holocene* 8: 19–29.
- Clark JS and Royall PD (1996) Local and regional sediment charcoal evidence for fire regimes in presettlement north-eastern North America. *Journal of Ecology* 84: 365–382.
- Cleveland WS and Devlin SJ (1988) Locally weighted regression: An approach to regression analysis by local fitting. *Journal of the American Statistical Association* 83: 596–610.
- Cofer WR III, Koutzenogii KP, Kokorin A, and Ezcurra A (1997) Biomass burning emissions and the atmosphere. In: Clark JS, Cachier H, Goldammer JG, and Stocks B (eds.) *Sediment Records of Biomass Burning and Global Change*, pp. 189–206. Berlin, Heidelberg: Springer.
- Daniau A-L, Harrison SP, and Bartlein PJ (2010) Fire regimes during the last glacial. *Quaternary Science Reviews* 29: 2918–2930.
- Dull R, Nevle RJ, Woods WI, et al. (2010) The Columbian encounter and the little ice age: Abrupt land use change, fire, and greenhouse forcing. *Annals of the Association of American Geographers* 4: 755–771.
- Elias VO, Simoneit BRT, Cordeiro RC, and Turco B (2001) Evaluating levoglucosan as an indicator of biomass burning in Carajás, Amazônia: A comparison to the charcoal record. *Geochimica et Cosmochimica Acta* 65: 267–272.
- Enache MD and Cumming BF (2006) Tracking recorded fires using charcoal morphology from the sedimentary sequence of Prosser Lake, British Columbia (Canada). *Quaternary Research* 65: 282–292.

- February EC (2000) Archaeological charcoal and dendrochronology to reconstruct past environments of southern Africa. *South African Journal of Science* 96: 111–116.
- Flannigan MD, Krawchuk MA, de Groot WJ, et al. (2009) Implications of changing climate for global wildland fire. *International Journal of Wildland Fire* 18: 483–507.
- Gavin DG (2003) Forest soil disturbance intervals inferred from soil charcoal radiocarbon dates. *Canadian Journal of Forest Research* 33: 2514–2518.
- Gavin DG, Brubaker LB, and Lertzman KP (2003) Holocene fire history of a coastal temperate rain forest based on soil charcoal radiocarbon dates. *Ecology* 84: 186–201.
- Gavin DG, Hallett D, Hu FS, et al. (2007) Forest fire and climate change: Insights from sediment charcoal records. *Frontiers in Ecology and the Environment* 5: 499–506.
- Gavin DG, Hu FS, Lertzman KP, and Corbett P (2006) Weak climatic control of forest fire history during the late Holocene. *Ecology* 87: 1722–1732.
- Grimm EC, Donovan JJ, and Brown KJ (2011) A high-resolution record of climate variability and landscape response from Kettle Lake, northern Great Plains, North America. *Quaternary Science Reviews* 30: 2626–2650.
- Hallett DJ and Anderson RS (2010) Paleofire reconstruction for high-elevation forests in the Sierra Nevada, California, with implications for wildfire synchrony and climate variability in the late Holocene. *Quaternary Research* 73: 180–190.
- Hebda RJ and Mathewes RW (1984) Holocene history of cedar and Native Indian cultures of the North American Pacific coast. *Science* 255: 711–713.
- Heinz C and Thiébaud S (1998) Characterization and palaeoecological significance of archaeological charcoal assemblages during late and post-glacial phases in southern France. *Quaternary Research* 50: 56–68.
- Higuera PE, Gavin DG, Bartlein PJ, and Hallett DJ (2010) Peak detection in sediment-charcoal records: Impacts of alternative data analysis methods on fire-history interpretations. *International Journal of Wildland Fire* 19: 996–1014.
- Higuera PE, Peters ME, Brubaker LB, and Gavin DG (2007) Understanding the origin and analysis of sediment-charcoal records with a simulation model. *Quaternary Science Reviews* 26: 1790–1809.
- Hopkins MS, Ash J, Graham AW, et al. (1993) Charcoal evidence of the spatial extent of the Eucalyptus woodland expansions and rainforest contractions in North Queensland during the late Pleistocene. *Journal of Biogeography* 20: 357–372.
- Hu FS, Brubaker LB, Gavin DG, et al. (2006) How climate and vegetation influence the fire regime of the Alaskan boreal biome: The Holocene perspective. *Mitigation and Adaptation Strategies for Global Change* 11: 829–846.
- IPCC (2007) Climate Change 2007: The Physical Science Basis. In: Solomon S, Qin D, and Manning M, et al. (eds.) *Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change*. Cambridge; New York, NY: Cambridge University Press.
- Iversen J (1941) Land occupation in Denmark's stone age. *Danmarks Geologiske Undersøgelse. II. Raekke* 66: 1–68.
- Jones TP, Chaloner WG, and Kuhlbusch TAJ (1997) Proposed bio-geological and chemical based terminology for fire-altered plant matter. In: Clark JS, Cachier H, Goldammer JG, and Stocks B (eds.) *Sediment Records of Biomass Burning and Global Change*, pp. 9–22. Berlin, Heidelberg: Springer.
- Kerhwal N, Zangrando R, Gambaro A, et al. (2010) Specific molecular markers in ice cores provide large-scale patterns in biomass burning. *PAGES News* 8: 59–61.
- Long C, Whitlock C, Bartlein PJ, and Millsaugh SH (1998) A 9000-year fire history from the Oregon Coast range, based on a high-resolution charcoal study. *Canadian Journal of Forest Research* 28: 774–787.
- Louchouart P, Kuo L-J, Wade TL, and Schantz M (2009) Determination of levoglucosan and its isomers in size fractions of aerosol standard reference materials. *Atmospheric Environment* 43: 5630–5636.
- Ludemann T and Nelle O (2002) Die Wälder am Schainsland und ihre Nutzung durch Bergau und Kölerei. *Berichte Freiburger Forstliche Forschung* 15: 1–139.
- Lynch JA, Clark JS, Bigelow NH, et al. (2002) Geographic and temporal variations in fire history in boreal ecosystems of Alaska. *Journal of Geophysical Research* 107: 1–17.
- Lynch JA, Clark JS, and Stocks BJ (2004) Charcoal production, dispersal, and deposition from the Fort Providence experimental fire: Interpreting fire regimes from charcoal records in boreal forests. *Canadian Journal of Forest Research* 34: 1642–1656.
- Marlon JR, Bartlein PJ, Long C, et al. (2012) Natural versus human causes of fire in the western U.S. *Proceedings of the National Academy of Sciences of the United States of America* 109: 535–543.
- Marlon JR, Bartlein PJ, Walsh M, et al. (2009) Wildfire responses to abrupt climate change in North America. *Proceedings of the National Academy of Sciences of the United States of America* 106: 2519–2524.
- McDadi O and Hebda RJ (2008) Change in historic fire disturbance in a Garry oak (*Quercus garryana*) meadow and Douglas-fir (*Pseudotsuga menziesii*) mosaic, University of Victoria, British Columbia, Canada: A possible link with First Nations and Europeans. *Forest Ecology and Management* 256: 1704–1710.
- McParland LC, Collinson ME, Scott AC, et al. (2007) Ferns and fires: Experimental charring of ferns compared to wood and implications for paleobiology, paleoecology, coal petrology, and isotope geochemistry. *PALAIOS* 22: 528–538.
- Mooney SD, Harrison SP, Bartlein PJ, et al. (2011) Late Quaternary fire regimes of Australasia. *Quaternary Science Reviews* 20: 28–46.
- Mooney SD and Tinner W (2011) The analysis of charcoal in peat and organic sediments. *Mires and Peat* 7: 1–18.
- Neumann K, Schoch W, Detienne P, and Schweingruber FH (2001) *Woods of the Sahara and the Sahel – An Anatomical Atlas*. Wiesbaden: Ludwig Reichert.
- Odgaard BV and Rasmussen P (2000) Origin and temporal development of macro-scale vegetation patterns in the cultural landscape of Denmark. *Journal of Ecology* 88: 733–748.
- Ohlson M, Brown KJ, Birks HJB, et al. (2011) Forest composition as a dominant driver of late-Holocene wildfire in boreal Europe. *Journal of Ecology* 99: 395–403.
- Ohlson M, Dahlberg B, Økland T, et al. (2009) The charcoal carbon pool in boreal forest soils. *Nature Geoscience* 2: 692–695.
- Ohlson M and Tryterud E (2000) Interpretation of the charcoal record in forest soils: forest fires and their production and deposition of macroscopic charcoal. *The Holocene* 10: 519–525.
- Overpeck JT, Rind D, and Goldberg R (1990) Climate-induced changes in forest and vegetation. *Nature* 343: 51–53.
- Peters ME and Higuera PE (2007) Quantifying the source area of macroscopic charcoal with a particle dispersal model. *Quaternary Research* 67: 304–310.
- Pisaric MFJ (2002) Long-distance transport of terrestrial plant material by convection resulting from forest fires. *Journal of Paleolimnology* 28: 349–354.
- Pitman AJ, Narisma GT, and McAneney J (2007) The impact of climate change on the risk of forest and grassland fires in Australia. *Climatic Change* 84: 383–401.
- Power MJ, Marlon JR, Bartlein PJ, and Harrison S (2009) Fire history and the global charcoal database: A new tool for hypothesis testing and data exploration. *Palaeogeography, Palaeoclimatology, Palaeoecology* 291: 52–59.
- Power MJ, Marlon J, Ortiz N, et al. (2008) Changes in fire regime since the LGM: An assessment based on a global synthesis of charcoal data. *Climate Dynamics* 30: 897–907.
- Power MJ, Mayle FE, Bartlein PJ, et al. (in press) 16th Century burning decline in the Americas: Population collapse or climate change? *Holocene*.
- Power MJ, Whitlock C, and Bartlein PJ (2011) Postglacial fire, vegetation, and climate history across an elevational gradient in the Northern Rocky Mountains, USA. *Quaternary Science Reviews* 30: 2520–2533.
- Pyne SJ (1984) *Introduction to Wildland Fire*, p. 455. New York: Wiley.
- Rhodes AN (1998) A method for the preparation and quantification of microscopic charcoal from terrestrial and lacustrine sediment cores. *The Holocene* 8: 113–117.
- Ruddiman WF (2003) The anthropogenic greenhouse era began thousands of years ago. *Climatic Change* 61: 261–293.
- Ruddiman WF, Kutzbach JE, and Vavrus SJ (2011) Can natural or anthropogenic explanations of late-Holocene CO₂ and CH₄ increases be falsified? *The Holocene* 21: 865–879.
- Saito K (2001) Flames. In: Johnson EA and Miyanish K (eds.) *Forest Fires: Behavior and Ecological Effects*, pp. 12–54. San Diego, CA: Academic Press.
- Sander PM and Gee CT (1990) Fossil charcoal: Techniques and applications. *Review of Palaeobotany and Palynology* 63: 269–279.
- Scholes MJ and Andreae MO (2000) Biogenic and pyrogenic emissions from Africa and their impact on the global atmosphere. *Ambio* 29: 23–29.
- Stine S (1994) Extreme and persistent drought in California and Patagonia during Medieval time. *Nature* 369: 546–549.
- Swetnam TW and Betancourt JL (1998) Mesoscale disturbance and ecological response to decadal climatic variability in the American Southwest. *Journal of Climate* 11: 3128–3147.
- Tinner W, Hofstetter S, Zeuglin F, et al. (2006) Long-distance transport of macroscopic charcoal by an intensive crown fire in the Swiss Alps – Implications for fire history reconstruction. *The Holocene* 16: 287–292.
- Turner NJ (1991) Burning mountain sides for better crops: Aboriginal landscape burning in British Columbia. *Archaeology in Montana* 32: 57–73.
- Vannièr B, Power MJ, Roberts N, et al. (2011) Circum-Mediterranean fire activity and climate changes during the mid-Holocene environmental transition (8500–2500 cal. BP). *The Holocene* 21: 53–73.
- Weber MG and Taylor SW (1992) The use of prescribed fire in the management of Canada's forested lands. *The Forestry Chronicle* 68: 324–334.

- Wright HE Jr. (1967) A square-rod piston sampler for lake sediments. *Journal of Sedimentary Research* 37: 975–976.
- Wright HE Jr. (1974) Landscape development, forest fires, and wilderness management. *Science* 186: 487–495.
- Zackrisson O, Nilsson MC, and Wardel DK (1996) Key ecological function of charcoal from wild fire in the Boreal forest. *Oikos* 77: 10–19.

Relevant Websites

<http://www.gpwg.org/>
<http://www.ncdc.noaa.gov/paleo/impd/paleofire.html>
<http://www.neotomadb.org/>

Introduction

Phytogeography is the study of the distribution of plants or taxonomic groups of plants and its focus is to explain the ranges of plants in terms of their origin, dispersal, and evolution (Matthews et al., 2003). As the minimum time period needed to observe changes in abundance and distribution of plants most often is longer than the career or life span of a geographer, paleoecological data and their understanding are of great importance. The phytogeography of the fossil record is, according to Wnuk and Pfefferkorn (1996), just as important as understanding modern phytogeography. The role of climate change in changing the distribution of species is the core area of biogeographical research (MacDonald et al., 2008).

The need for observations on longer timescales is acute because most species distributions are not in equilibrium with present climate and changes are occurring at larger scales. Historical processes, such as dispersal and colonization since the last glacial stage, and time are important for explaining and understanding changes in the distribution of species, vegetation belts or biotas, and ranges and patterns of population responses to environmental change (e.g., Bhagwat and Willis, 2008; Birks, 1989; Cruzan and Templeton, 2000; Grytnes et al., 1999; Huntley and Webb, 1989). Pollen and plant macrofossil data from different sedimentary records are well suited, and the construction of databases such as the European Pollen Database (Fyfe et al., 2009) and the Neotoma database (see [Databases and Their Application](#)) has facilitated the compilation of data on larger temporal and spatial scales as well as testing hypotheses (Binney et al., 2009; Cruzan and Templeton, 2000; Gajewski, 2008). For large-scale studies or biome-scale studies, pollen data can be aggregated to a higher taxonomic level, thus avoiding the problem in pollen analysis with low-resolution taxonomy (Gajewski, 2008). Additionally, there are an increasing number of sites where pollen and macrofossil analyses have been done at high temporal resolution and where more details can be studied. These may show that equilibrium states are never reached, but vegetation is under continuous change in a dynamic world.

Climate (temperature, precipitation) is believed to be one of the main forcing factors for plant distributions, but there are also other possible contributing factors or types of environmental change that affect ecosystems, such as fires, in addition to human impact, competition, succession, and invasions (McLachlan and Clark, 2004). These factors might covary spatially, so it is sometimes difficult to separate the direct importance of one specific factor (Grytnes et al., 1999).

According to MacDonald et al. (2008), it is important to understand climate change and its potential impact on species, populations, and communities. Because future climate changes are expected to take place at a greater rate than those in the recent geological past (Overpeck, 1996), it is extremely

important to use insights from paleoecology and paleophytogeography to be able to fully understand present-day and possible future biotic responses to environmental change.

Impact of Ice Ages

Ice Ages have had significant impacts on the geography of global biotas (Willis et al., 2004). The changes between cold and warm climates have led to repeated restriction and isolation of many populations, which has driven the formation of new species. Evidence for this can be found in the fossil pollen and plant macrofossil records. Repeated glacial–interglacial cycles have occurred throughout Earth's history on Milankovich timescales (400, 100, 40, and 23–19 ka). The most recent and the best studied periods are the Quaternary Ice Ages covering the last 2.6 million years (My) (Hooghiemstra and Van der Hammen, 2004; Willis et al., 2004). According to Frogley et al. (1999), up to 80% of Quaternary time has been characterized by glacial conditions, while only 20% of the time had conditions similar to or warmer than the present.

During the Quaternary Ice Ages, dramatic environmental changes led to major shifts in the distribution and abundance of vegetation and vegetation belts (Bennett, 2004; Hooghiemstra and Van der Hammen, 2004). The main response of vegetation has been changes from forest to non-forested vegetation when climates have been colder (Tzedakis et al., 1997). Willis and Niklas (2004) found little evidence for increases in plant diversity during the Quaternary but concluded that stasis and extinction were the main responses to loss and fragmentation of habitats. More recent work has shown that the Quaternary Ice Ages have been essential for genetic diversification and speciation. According to Willis and Niklas (2004), three evolutionary scenarios can explain the fragmentation and re-expansion of species ranges during the Quaternary: (1) species persistence, (2) extinction, and (3) classical allopatric speciation due to geographical isolation. Some species were restricted to isolated glacial refugia, whereas other species persisted in large populations. There is no evidence of increased speciation during the Quaternary period, and local extinctions (extirpations) were more the case than total extinctions (Willis and Niklas, 2004). One of the few examples of plant species extinction is that of *Picea critchfieldii* in eastern North America in the Late Quaternary (Jackson and Weng, 1999). In the Colombian Andes, pollen records show that the treeless paramo vegetation was more widespread during Pleistocene glacial intervals, whereas during interglacial times, it was limited to the higher altitudes, as the upper tree limit advanced upslope (Hooghiemstra and Van der Hammen, 2004). Similar patterns have been noted in many other mountainous regions (e.g., the western United States).

The Quaternary Ice Ages also had an impact on the genetic diversity within species. [Petit et al. \(2003\)](#) examined the distribution of variation in neutral chloroplast DNA markers throughout the ranges of 22 common tree species in Europe. They found that genetic diversity was highly differentiated in southern populations, consistent with the expected effect of population isolation during the glacial intervals. The highest genetic diversity level was found at mid-latitudes where lineages mixed during the Holocene recolonization, giving the northern populations a small subset of genetic diversity generated in glacial refuges ([Petit et al., 2003](#)).

The Last Glacial Phase

A major problem arising from glaciation is the effect of glaciers and ice sheets overriding the land surface and removing much of the surficial sediment, making the included fossils as well as the number of suitable sites scarce. Reconstruction of Late Glacial Maximum (LGM) (ca. 21 ka BP) ranges of plants is important for assessing climate and biotic sensitivity to various forcings, because orbital parameters were similar to today, but the boundary conditions and forcing factors were different from today during the LGM ([Jackson et al., 2000](#)). In North America and Europe, LGM sites are few, many do not yield good dates, and the derived paleobotanical records lack modern analogs in both climate and vegetation ([Jackson et al., 2000](#); [McLachlan and Clark, 2004](#)). This has led paleoecologists to infer vegetation on broader scales using biomes and plant functional types.

In Europe, temperate trees and woodland species must have survived in smaller refugia, whereas in North America, the vegetation migrated to more restricted zones or larger areas in the south. According to [Stewart and Lister \(2001\)](#), treeless vegetation dominated the unglaciated parts of northern Europe, Asia, and northern America during the last glacial interval. It has for long been the only view based on pollen data that woody taxa in Europe survived in three refugia during the LGM: on the Iberian Peninsula, in Italy, and in the Balkans ([Bennett et al., 1991](#); [Huntley and Birks, 1983](#); [Willis, 1994](#)). The position of these has been debated in recent decades. Pollen alone is often not sufficient to indicate the presence of species, due to its long distance transport and poor taxonomic resolution ([Jackson et al., 2000](#)). Therefore, it has been difficult to detect these small, isolated populations by pollen only ([Birks, 1989](#)). It has been demonstrated by [Loehle \(2007\)](#) that pollen production is reduced in cold environments in combination with low concentrations of atmospheric carbon dioxide. One approach for tracing these small refugial populations is to examine other proxies, such as plant macrofossils ([Bhagwat and Willis, 2008](#); [Birks and Birks, 2003](#)) and genetic data ([McLachlan et al., 2005](#); [Qiu et al., 2011](#)).

New macrofossil data and genetic studies indicate that cold-adapted taxa survived further north during the full glacial ([Bhagwat and Willis, 2008](#); [Binney et al., 2009](#); [Birks and Willis, 2008](#); [Väliranta et al., 2011](#); [Willis and van Andel, 2004](#)). For Beringia, [Brubaker et al. \(2005\)](#) conclude that small populations of boreal trees and shrubs must have existed as small refugial populations during the LGM. In central and eastern Europe, [Willis and van Andel \(2004\)](#) argue, on the

basis of macrofossil charcoal from trees, that 20 different tree taxa were growing in the region during the last full-glacial interval. According to [Bhagwat and Willis \(2008\)](#), these trees survived in these northerly refugia due to specific biogeographical traits, such as short generation time, small seed size, and vegetative reproduction, and not by chance. In North America, it has been shown that broad-leaved trees were distributed more widely than first believed. Syntheses of pollen and macrofossil data show that boreal tree taxa and/or tundra were growing close to the Laurentide Ice margin during the LGM ([Jackson et al., 2000](#)). [Anderson et al. \(2006\)](#) have shown, from multiple lines of evidence including DNA and macrofossil analyses, that white spruce (*Picea glauca*) survived in glacial refugia in Alaska.

The Present Interglacial: The Holocene

In light of new macrofossil evidence, both for the LGM and the early Holocene, the postglacial vegetation migration rates of 500–1000 m year⁻¹ suggested by [Huntley and Birks \(1983\)](#) have been questioned by several authors (e.g., [Binney et al., 2009](#); [McLachlan et al., 2005](#)) because observations of mean dispersal distances of most trees and herbs are too small to account for these recolonization rates in northern latitudes, following retreat of the glaciers. This lack of consistency between migration rates derived from different biological data sets has been termed Reid's paradox ([Clark et al., 1998](#)). Resolving this issue has been a major task in recent paleoecological research.

With the new evidence from macrofossil data showing that many tree species survived in refugia further north, a more detailed picture of their postglacial migration routes and rates can be drawn ([Binney et al., 2009](#); [Väliranta et al., 2011](#)). The existence of these northern refugia can explain the rapid spread and expansion of trees in the early Holocene and thus also explain early finds of tree remains at high altitudes in Scandinavia ([Kullman, 1998](#)).

Past behavior of the treeline is also of interest because it is linked to large-scale climate features ([Binney et al., 2009](#)). Altitudinal and latitudinal Holocene treeline changes in relation to climate change have been studied extensively in many areas, based both on pollen and macrofossils (e.g., [Binney et al., 2009](#); [Bjune, 2005](#); [Hooghiemstra and Van der Hammen, 2004](#); [MacDonald et al., 2000](#); [Payette et al., 2002](#)). In addition to climate, the spatial and temporal variations in treeline are also related to tree dispersal rates and distances to glacial refugia. From LGM to mid-Holocene, latitudinal treelines in Europe migrated north more than 2000 km from areas close to the Alps to northern parts of Scandinavia ([Willis et al., 2000](#)). Similar trends are seen in the Colombian Andes ([Hooghiemstra and Van der Hammen, 2004](#); [Van der Hammen, 1974](#)). In addition to latitudinal changes in the treeline, large changes have been observed in the altitudinal treeline in Scandinavia (e.g., [Barnekow, 1999](#); [Bjune, 2005](#); [Kullman, 1998](#)) as well as in the Alps (e.g., [Tinner and Kaltenrieder, 2005](#)) in the early and middle part of the Holocene. The pattern of treeline movements in North America differed somewhat from that of Eurasia. Trees migrated north through Canada rather late, because these areas only became free of ice in

the mid-Holocene in some cases. Most of these forest trees spread toward northern areas from a large southern refugium (Payette et al., 2002).

One significant recent ecosystem change is the establishment of spruce (*Picea abies*) forest in southern Scandinavia in the later part of the Holocene. This led to a switch in keystone species in the forests, as spruce replaced the existing forests dominated by birch (*Betula*) and/or pine (*Pinus sylvestris*) in some areas and broad-leaved forests in others (e.g., Bjune et al., 2009; Giesecke, 2005; Giesecke and Bennett, 2004; see **Changing Plant Distributions and Abundances**). During the last 5000–6000 years, small spruce populations expanded rapidly northward from areas close to the ice sheet (Giesecke, 2005; Giesecke and Bennett, 2004). Also, in northwest Russia, there is evidence for spruce growth at and beyond the present-day treeline at the onset of the Holocene (Väliranta et al., 2011).

Future

It is evident from the fossil and genetic records that species have changed their distribution as a reaction to climate and this will happen in the future as well. Understanding how plant distribution and biodiversity have responded to past environmental factors, including climate change, natural disturbance, and human activities, is essential to predicting how plant distribution and diversity may change in response to global warming. The understanding of past responses gained through paleophytogeography and paleoecology is important for the prediction of future migration rates and distributional changes. This knowledge will also help determine regions with the highest genetic diversity, a factor that is important for the long-term persistence and conservation of plant species. Paleoecological data also allow comparisons between present-day ecological dynamics and dynamics before human impact. Closer integration of data sets and more high-resolution analyses will improve the quality and extend the applicability of paleoecological data.

To fully utilize data from the past, much can be gained by combining traditional paleoecological data with new techniques such as molecular methods and climate and population modeling. Genetic data provide an effective approach to detect cryptic refugia (Anderson et al., 2006), population sizes, and lineages of ancestry but are, relative to paleoecological data, less accurate in providing evidence of timing and location. Extracting DNA from the fossil records, such as fossil pollen, can answer questions on plant migration in space and time at a high taxonomic level (MacDonald et al., 2008). Modeling will be important to estimate potential climatic and ecological forcing mechanisms. Ecological models and simulations will provide tests of dynamic responses of vegetation to climate changes and are thus important for conservation and management.

See also: Plant Macrofossil Introduction; Pollen Analysis, Principles. **Introduction:** History of Quaternary Science. **Paleobotany:** Ancient Plant DNA; Overview of Terrestrial Pollen Data. **Pollen Methods and Studies:** Changing Plant Distributions and Abundances; Databases and Their Application.

References

- Anderson LL, Hu FS, Nelson DM, et al. (2006) Ice-age endurance: DNA evidence of a white spruce refugium in Alaska. *Proceedings of the National Academy of Sciences* 103: 12447–12450.
- Barnekow L (1999) Holocene tree-line dynamics and inferred climatic changes in the Abisko area, northern Sweden, based on macrofossil and pollen records. *The Holocene* 9: 253–265.
- Bennett KD (2004) Continuing the debate on the role of Quaternary environmental change for macroevolution. *Philosophical Transactions Royal Society London B* 359: 295–303.
- Bennett KD, Tzedakis PC, and Willis KJ (1991) Quaternary refugia of north European trees. *Journal of Biogeography* 18: 103–115.
- Bhagwat SA and Willis KJ (2008) Species persistence in northerly refugia of Europe: A matter of chance or biogeographical traits? *Journal of Biogeography* 35: 464–482.
- Binney HA, Willis KJ, Edwards MJ, et al. (2009) The distribution of late-Quaternary woody taxa in northern Eurasia: Evidence from a new macrofossil database. *Quaternary Science Reviews* 28: 2445–2464.
- Birks HJB (1989) Holocene isochrone maps and patterns of tree-spreading in the British Isles. *Journal of Biogeography* 16: 503–540.
- Birks HH and Birks HJB (2003) Reconstructing Holocene climates from pollen and plant macrofossils. In: Mackay A, Battarbee RW, Birks HJB, and Oldfield F (eds.) *Global Change in the Holocene*, pp. 341–357. London: Hodder Arnold.
- Birks HJB and Willis KJ (2008) Alpines, trees, and refugia in Europe. *Plant Ecology and Diversity* 1: 147–160.
- Bjune AE (2005) Holocene vegetation history and tree-line changes on a north–south transect crossing major climate gradients in southern Norway – Evidence from pollen and plant macrofossils in lake sediments. *Review of Palaeobotany and Palynology* 133: 249–275.
- Bjune AE, Ohlson M, Birks HJB, et al. (2009) The development and local stand-scale dynamics of a *Picea abies* forest in southeastern Norway. *The Holocene* 19: 1073–1082.
- Brubaker LB, Anderson PM, Edwards ME, et al. (2005) Beringia as a glacial refugium for boreal trees and shrubs: New perspectives from mapped pollen data. *Journal of Biogeography* 32: 833–848.
- Clark JS, Fastie C, Hurr G, et al. (1998) Reid's paradox of rapid plant migration. *Bioscience* 48: 13–24.
- Cruzan MB and Templeton AR (2000) Paleogeography and coalescence: Phylogeographic analysis of hypotheses from the fossil record. *Trends in Ecology & Evolution* 15: 491–496.
- Frogley MR, Tzedakis PC, and Heaton THE (1999) Climatic variability in northwest Greece during the last interglacial. *Science* 285: 1886–1889.
- Fyfe RM, de Beaulieu JL, Binney H, et al. (2009) The European pollen database: Past efforts and current activities. *Vegetation History and Archaeobotany* 18: 417–424.
- Gajewski K (2008) The global pollen database in biogeographical and palaeoclimatic studies. *Progress in Physical Geography* 32: 379–402.
- Giesecke T (2005) Moving front or population expansion: How did *Picea abies* (L.) Karst. become frequent in central Sweden? *Quaternary Science Reviews* 24: 2495–2509.
- Giesecke T and Bennett KD (2004) The Holocene spread of *Picea abies* (L.) Karst. in Fennoscandia and adjacent areas. *Journal of Biogeography* 31: 1523–1548.
- Grytnes JA, Birks HJB, and Peglar SM (1999) Plant species richness in Fennoscandia: Evaluating the relative importance of climate and history. *Nordic Journal of Botany* 19: 489–503.
- Hooghiemstra H and Van der Hammen T (2004) Quaternary Ice-Age dynamics in the Colombian Andes: Developing an understanding of our legacy. *Philosophical Transactions Royal Society London B* 359: 173–181.
- Huntley B and Birks HJB (1983) An Atlas of Past and Present Pollen Maps of Europe: 0–13,000 years ago, p. 667. Cambridge: Cambridge University Press.
- Huntley B and Webb T III (1989) Migration: Species' response to climatic variations caused by changes in the Earth's orbit. *Journal of Biogeography* 16: 5–19.
- Jackson ST, Webb RS, Anderson KH, et al. (2000) Vegetation and environment in eastern North America during the Last Glacial Maximum. *Quaternary Science Reviews* 19: 489–508.
- Jackson ST and Weng C (1999) Late Quaternary extinction of a tree species in eastern North America. *Proceedings of the National Academy of Sciences of the United States of America* 93: 13847–13852.
- Kullman L (1998) Non-analogous tree flora in the Scandes Mountains, Sweden, during the early Holocene – Macrofossil evidence of rapid geographic spread and response to palaeoclimate. *Boreas* 27: 153–161.

- Loehle C (2007) Predicting Pleistocene climate from vegetation in North America. *Climates of the Past* 3: 109–118.
- MacDonald GM, Bennett KD, Jackson ST, et al. (2008) Impacts of climate change on species, populations and communities: Palaeobiogeographical insights and frontiers. *Progress in Physical Geography* 32: 139–172.
- MacDonald GM, Velichko AA, Kremenetski CV, et al. (2000) Holocene treeline history and climate change across northern Eurasia. *Quaternary Research* 53: 302–311.
- Matthews JA, Bridges EM, Caseldine CJ, et al. (2003) *The Encyclopedic Dictionary of Environmental Change*, p. 690. London: Arnold.
- McLachlan JS and Clark JS (2004) Reconstructing historical ranges with fossil data at continental scales. *Forest Ecology and Management* 197: 139–147.
- McLachlan JS, Clark JS, and Manos PS (2005) Molecular indicators of tree migration capacity under rapid climate change. *Ecology* 86: 2088–2098.
- Overpeck JT (1996) Warm climate surprises. *Science* 271: 1820–1821.
- Payette S, Eronen M, and Jasinski JJP (2002) The circumboreal Tundra–Taiga interface: Late Pleistocene and Holocene changes. *Ambio, Special Report Number 12. Dynamics of the Tundra-Taiga Interface*, 15–22.
- Petit RJ, Aguinagalde I, de Beaulieu JL, et al. (2003) Glacial refugia: Hotspots but not melting pots of genetic diversity. *Science* 300: 1563–1565.
- Qiu YX, Fu CX, and Comes HP (2011) Plant molecular phylogeography in China and adjacent regions: Tracing the genetic imprints of Quaternary climate and environmental change in the world's most diverse temperate flora. *Molecular Phylogenetics and Evolution* 59: 225–244.
- Stewart JR and Lister AM (2001) Cryptic northern refugia and the origins of the modern biota. *Trends in Ecology & Evolution* 16: 608–613.
- Tinner W and Kaltenrieder P (2005) Rapid responses of high-mountain vegetation to early Holocene environmental changes in the Swiss Alps. *Journal of Ecology* 93: 936–947.
- Tzedakis PC, Andrieu V, de Beaulieu JL, et al. (1997) Comparison of terrestrial and marine records of changing climate of the last 500,000 years. *Earth and Planetary Science Letters* 150: 171–176.
- Väliranta M, Kaakinen A, Kuhry P, et al. (2011) Scattered late-glacial and early Holocene tree populations as dispersal nuclei for forest development in north-eastern European Russia. *Journal of Biogeography* 38: 922–932.
- Van der Hammen T (1974) The Pleistocene changes of vegetation and climate in tropical South America. *Journal of Biogeography* 1: 3–26.
- Willis KJ (1994) The vegetational history of the Balkans. *Quaternary Science Reviews* 13: 769–788.
- Willis KJ, Bennett KD, and Walker D (2004) Introduction. The evolutionary legacy of the ice ages. *Philosophical Transactions Royal Society London B* 359: 157–158.
- Willis KJ and Niklas KJ (2004) The role of Quaternary environmental change in plant macroevolution: The exception or the rule? *Philosophical Transactions Royal Society London B* 359: 159–172.
- Willis KJ, Rudner E, and Sumegi P (2000) The full-glacial forests of central and southeastern Europe. *Quaternary Research* 53: 203–213.
- Willis KJ and van Andel TH (2004) Trees or no trees? The environments of central and eastern Europe during the Last Glaciation. *Quaternary Science Reviews* 23: 2369–2387.
- Wnuk C and Pfefferkorn HW (1996) Foreword: The value of paleophytogeography. *Review of Palaeobotany and Palynology* 90: 1–3.

Relevant Websites

<http://www.europeanpollendatabase.net/> – The European Pollen Database.
<http://www.neotomadb.org/> – Neotoma Paleocology Database.

Silicon Isotopes in Diatoms

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Introduction

Silicon isotope analysis of diatom silica ($\delta^{30}\text{Si}_{\text{diatom}}$) is a rapidly evolving technique for tracking changes in silicon uptake by diatoms within both marine and freshwater environments. In combination with other isotopic proxies, $\delta^{30}\text{Si}_{\text{diatom}}$ can provide an alternative insight into biogeochemical cycles of the past. To date, $\delta^{30}\text{Si}_{\text{diatom}}$ measurements have played a key role in formulating and testing hypotheses concerning phytoplankton export of carbon to the deep ocean, with implications for understanding the long-term controls over the global carbon cycle (Beucher et al., 2007; Brzezinski et al., 2002; De La Rocha et al., 1998; Pichevin et al., 2009). $\delta^{30}\text{Si}_{\text{diatom}}$ also has potential for tracking environmental change using lake sediments, either as a proxy for diatom biogeochemistry, or as a tracer of past changes in the inorganic silicon cycle, related to past weathering processes in hydrological systems (Alleman et al., 2005; Street-Perrott et al., 2008; Swann et al., 2010).

Silicon is the second most abundant element on Earth, with three major stable isotopes, ^{28}Si , ^{29}Si , and ^{30}Si , with abundances of 92.22, 4.68, and 3.08%, respectively (Rosman and Taylor, 1998). Silicon isotope ratios are reported as either $\delta^{29}\text{Si}$ or $\delta^{30}\text{Si}$, whereby:

$$\delta^{30}\text{Si} = 1000 \left(\frac{R_{\text{Sample}} - R_{\text{Std}}}{R_{\text{Std}}} \right) \quad [1]$$

where R is the $^{30}\text{Si}/^{28}\text{Si}$ (or $^{29}\text{Si}/^{28}\text{Si}$ for $\delta^{29}\text{Si}$) ratio of either the sample or standard (NBS-28), as indicated by the subset label. Today, majority of techniques allow measurement of $\delta^{30}\text{Si}$; however, analytical limitations with some methods require instead the use of $\delta^{29}\text{Si}$ (e.g., Cardinal et al., 2003; De La Rocha, 2002). Because silicon isotope fractionation is mass dependent, $\delta^{30}\text{Si}$ can be estimated from $\delta^{29}\text{Si}$ as (Cardinal et al., 2003; De La Rocha, 2002):

$$\delta^{30}\text{Si} = \delta^{29}\text{Si} \times 1.93 \quad [2]$$

Diatoms assimilate and polymerize hydrous silica ($\text{Si}[\text{OSi}]_n[\text{OH}]_{4-n}$) from silicic acid ($\text{Si}[\text{OH}]_4$) to form the distinctive silica frustules which distinguish them from other microalgae and which accumulate in marine and lacustrine deposits. The fundamental origin of silicic acid is the dissolution of silicate minerals, transported to the lake or marine environment via airborne dust, rivers, or groundwater (Basile-Doelsch, 2006; De La Rocha and Bickle, 2005; De La Rocha et al., 2000). In many cases, a large percentage of the silicic acid available to diatoms originates from the dissolution of sedimenting diatom frustules (e.g., Alleman et al., 2005; Treguer et al., 1995). The pathway by which silicic acid reaches the environments in which diatoms grow can have a significant effect on the isotopic composition of the silicic acid and subsequent isotopic composition of diatom silica. Therefore, $\delta^{30}\text{Si}_{\text{diatom}}$ is a function of the silicon isotope composition of dissolved silica/silicic acid ($\delta^{30}\text{Si}_{\text{DSi}}$), the amount of silicic acid

uptake by diatoms, and the isotopic fractionation between the silicic acid and diatom silica (De La Rocha et al., 1997, 2000). Consequently, understanding the $\delta^{30}\text{Si}_{\text{diatom}}$ of sedimentary deposits requires an appreciation of both the processes that control the transport of silicic acid to diatom habitats, and those which affect the uptake of silicic acid by diatoms. Because silicon isotopes can be used as a proxy for past changes in diatom productivity, the biological effects over $\delta^{30}\text{Si}_{\text{diatom}}$ have attracted the most attention (Brzezinski et al., 2002; De La Rocha et al., 1997, 1998; Pichevin et al., 2009). However, there is also potential for using $\delta^{30}\text{Si}_{\text{diatom}}$ as a proxy for past changes in weathering and transport of nutrients to aquatic systems (Georg et al., 2009a; Street-Perrott et al., 2008; Swann et al., 2010). This article aims to detail the most important aspects of silicon isotope variability in the contemporary environment, while demonstrating the considerable potential for using $\delta^{30}\text{Si}_{\text{diatom}}$ in Quaternary research. Additional reviews on silicon isotopes, in diatoms, other organisms and inorganic systems are presented by De La Rocha (2006), Leng et al. (2009), and Basile-Doelsch (2006).

Analytical Procedures

As is the case with the oxygen isotope composition of diatoms ($\delta^{18}\text{O}_{\text{diatom}}$) $\delta^{30}\text{Si}_{\text{diatom}}$ analysis requires samples of pure diatom silica, free of any minerogenic or organic contamination. The procedures for purification of diatom silica and quantification of contaminants largely mirror those for $\delta^{18}\text{O}_{\text{diatom}}$, and are thus not covered here. It has been suggested that $\delta^{30}\text{Si}_{\text{diatom}}$ preparation requires the avoidance of glassware, which can be a source of contamination. In addition, some researchers involve extra trace metal cleaning steps prior to isotope analysis, which may be beneficial in stripping the diatoms of externally attached clay particles and other contaminating oxides (e.g., Hendry et al., 2010). While $\delta^{18}\text{O}_{\text{diatom}}$ is more sensitive to sample preparation temperature, reagent and heating time, the primary concern with $\delta^{30}\text{Si}_{\text{diatom}}$ is avoidance of sample dissolution, which can lead to isotopic bias (Demarest et al., 2009).

A number of methods exist for the measurement of $\delta^{30}\text{Si}$ from particulate diatom silica and dissolved silicic acid (Leng et al., 2009; Reynolds et al., 2007). These include: fluorine based conversion of silica into SiF_4 and isotope ratio mass spectrometry (IRMS) (De Freitas et al., 1991; De La Rocha et al., 1996; Ding et al., 1996; Leng and Sloane, 2008), multiple collector inductively coupled plasma mass spectrometry (MC-ICP-MS; Cardinal et al., 2003; De La Rocha, 2002; Reynolds et al., 2006b), acid decomposition and IRMS (Brzezinski et al., 2006), and secondary ion mass spectrometry (SIMS) (Basile-Doelsch et al., 2005). An interlaboratory comparison study found that each of the different methods generate identical values within analytical error

(Reynolds et al., 2007), with reproducibility typically between 0.05 and 0.1‰ for all methods except SIMS.

Fluorination

Fluorination methods involve the conversion of silica into SiF_4 using a fluorine based reagent, such as ClF_3 or BrF_5 (De La Rocha et al., 1996; Douthitt, 1982; Leng and Sloane, 2008; Tilles, 1961). The silica sample is reacted in three steps, and subsequently the Si is extracted. Stage one involves ‘outgassing’ (dehydration) to remove surficial and loosely bonded water in the reaction tubes at room temperature. Stage two involves a prefluorination step involving a stoichiometric deficiency of the reagent at low temperatures which removes hydroxyl and loosely bound water. The third stage is a full reaction at 450 °C for around 12 h with an excess of reagent to disassociate the silica into O_2 and SiF_4 . Isotopes are subsequently measured by IRMS. Advantages of using fluorination include the ability to measure oxygen and silicon isotopes from the same sample (Leng and Sloane, 2008), while recent developments using a laser microprobe offer the potential to analyze individual diatom frustules (Gao and Ding, 2009).

Multiple Collector Inductively Coupled Plasma Mass Spectrometry

MC-ICP-MS, operating in either wet or dry plasma mode, offers the option to avoid using fluorine reagents (Cardinal et al., 2003; De La Rocha, 2002; Georg et al., 2006b; Reynolds et al., 2006b). Reliable silicon isotope measurements can be obtained with analytical errors that are equivalent to or lower than those achieved using fluorination from much smaller samples. Further improvements in sensitivity and reproducibility have been made by dissolving the silica either in aqueous NaOH or by alkaline fusion followed by cation exchange chromatography to purify Si for analysis (Georg et al., 2006b; van den Boorn et al., 2006). When operating at normal resolution, MC-ICP-MS is limited by atomic interference over the Si mass range, predominantly from $^{14}\text{N}^{16}\text{O}$ which has a mass of 30. In such circumstances, only $\delta^{29}\text{Si}$ can be calculated (Cardinal et al., 2003; De La Rocha, 2002). High resolution MC-ICP-MS eliminates the effects of any interference and allows direct measurement of $\delta^{30}\text{Si}$ (Chmeleff et al., 2008; Reynolds et al., 2006b).

Acid Decomposition

The acid decomposition method enables part automation of sample preparation and analysis, using a modified Finnigan Kiel device and IRMS (Beucher et al., 2007, 2008; Brzezinski et al., 2006). Silica samples of high purity are converted to Cs_2SiF_6 by dissolution in HF and the addition of CsCl. Using the modified Kiel device, the Cs_2SiF_6 product is decomposed in sulfuric acid to generate SiF_4 gas online with a gas source isotope ratio mass spectrometer for analysis. Samples that contain significant levels of nonsiliceous impurity can be pre-processed by dissolution in HF followed by precipitation of

triethylamine silicomolybdate and combustion of the precipitate to generate SiO_2 , following De La Rocha et al (1996) (Brzezinski et al., 2006).

Secondary Ion Mass Spectrometry

SIMS is predominantly used for analysis of mineral grains, however, the ability to analyze particles larger than ~50 µm diameter offers the potential for analysis of individual diatom frustules (Alexander et al., 2002; Basile-Doelsch et al., 2005). SIMS requires polished thin sections and individual grains must be mounted in epoxy resin prior to analysis. In terms of analytical precision, SIMS measurements are less precise ($\pm 0.75\%$ (2σ)). However, an improvement in analytical precision would offer great potential in terms of examining the isotopic homogeneity of sediment samples, identifying species-specific isotope ratios, or for obtaining high resolution isotope records using extremely small samples.

Fractionation of Silicon Isotopes During Diatom Growth

Central to the use of $\delta^{30}\text{Si}_{\text{diatom}}$ as a biogeochemical tracer is the preferential assimilation by diatoms of the lighter ^{28}Si over ^{30}Si during synthesis of their silica frustules. As a result of this isotopic discrimination, uptake of silicic acid from the substrate leads to a progressive increase in the $\delta^{30}\text{Si}$ of residual silicic acid and a corresponding increase in the $\delta^{30}\text{Si}$ of subsequent diatom silica (Figure 1). Therefore, on a first order, there is a positive relationship between the amount of silica used by diatoms, the isotopic composition of dissolved silicon ($\delta^{30}\text{Si}_{\text{DSi}}$), and the isotopic composition of the diatoms ($\delta^{30}\text{Si}_{\text{diatom}}$; Figure 1). The discrimination by diatoms against ^{30}Si is expressed in terms of the fractionation factor (ϵ), which is approximately equal to the difference in δ between the dissolved and particulate phase:

$$\epsilon = 1000 \times \ln \left(\frac{R_{\text{DSi}}}{R_{\text{diatom}}} \right) \approx \delta^{30}\text{Si}_{\text{DSi}} - \delta^{30}\text{Si}_{\text{diatom}} \quad [3]$$

De La Rocha et al. (1997) estimated ϵ as -1.1% , independent of species and growth temperature, based on culture experiments using three species of marine phytoplankton: *Skeletonema costatum*, *Thalassiosira weissflogii*, and *Thalassiosira* sp. These findings are supported by an earlier study by Spadero (1983) who estimated a similar value of ϵ for *Phaeodactylum tricornutum* and by Milligan et al (2004), who cultured *T. weissflogii* across a range of dissolved CO_2 concentrations. To date, it is unclear when and where silicon isotopic fractionation occurs within the diatom cell during silicification. However, because changes in dissolved CO_2 concentration affect the amount of silica polymerization by *T. weissflogii*, but not $\delta^{30}\text{Si}_{\text{diatom}}$, it appears that the isotopic fractionation does not take place during the polymerization step (Milligan et al., 2004). Further studies such as this are required in order to develop a mechanistic perspective of silicon isotope fractionation by diatoms. For example, the four species of marine phytoplankton examined so far represent a narrow group with respect to the diversity of diatoms and the range of habitats in which they exist (Round et al., 1990). In particular, there are no

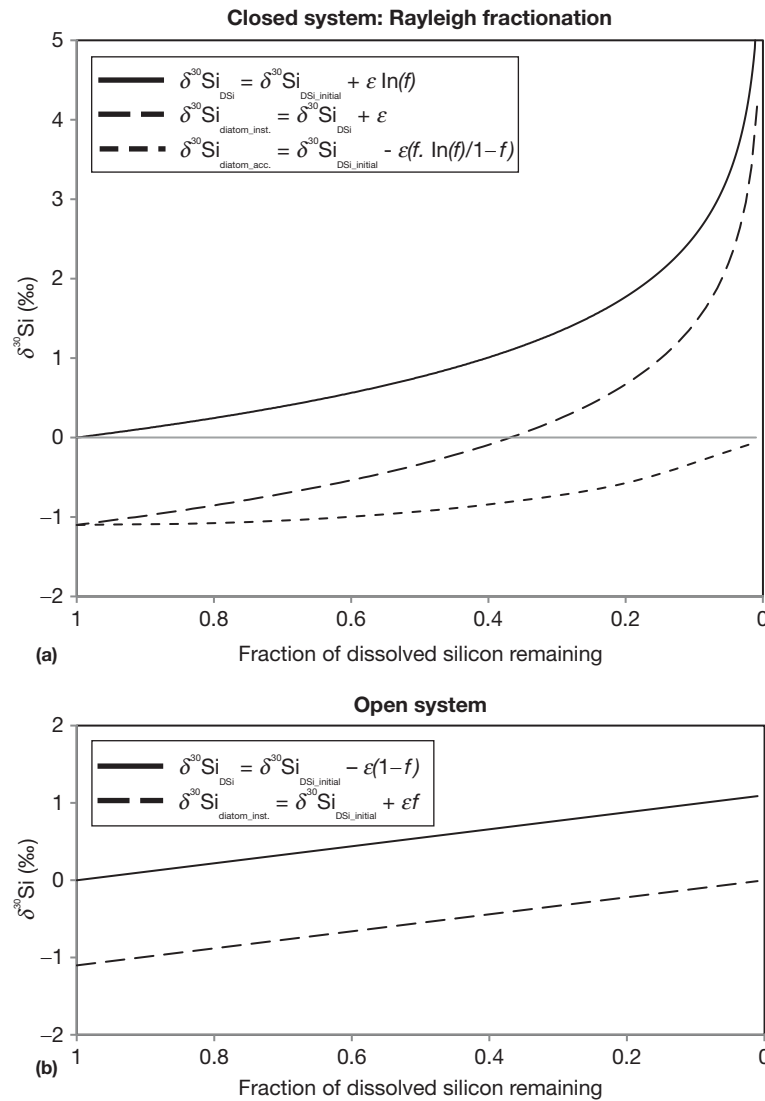


Figure 1 Conceptual models for isotope fractionation between silicic acid (dissolved silicon) and diatom silica. (a) Rayleigh distillation within a closed system, assuming a single initial silicic acid source with no replenishment. (b) Isotope fractionation in a steady state open system, assuming constant replenishment of the silicic acid substrate, balanced by constant removal of diatom silica. In equations, f = fraction of dissolved silica remaining; $\delta^{30}\text{Si}_{\text{DSi}}$ = silicon isotope composition of dissolved silicon; $\delta^{30}\text{Si}_{\text{DSi, initial}}$ = initial $\delta^{30}\text{Si}_{\text{DSi}}$ at $f=1$; $\delta^{30}\text{Si}_{\text{diatom, inst.}}$ = instantaneous $\delta^{30}\text{Si}$ of diatom silica; $\delta^{30}\text{Si}_{\text{diatom, acc.}}$ = cumulative $\delta^{30}\text{Si}$ of diatom silica.

controlled estimates for isotope fractionation at temperatures relevant to the Southern Ocean, which is an important region for diatom productivity. Nor have there been any culture based studies of freshwater diatoms, despite a marked difference in silicification between freshwater and marine taxa (Conley et al., 1989).

Beyond the laboratory, the validity of $\delta^{30}\text{Si}_{\text{diatom}}$ as a biogeochemical tracer has been investigated by monitoring the natural environment. At first glance, the three dimensional distribution of $\delta^{30}\text{Si}_{\text{DSi}}$ in the ocean appears to conform with culture observations, whereby high $\delta^{30}\text{Si}_{\text{DSi}}$ values are found in regions of low silicic acid concentration due to uptake by diatom growth (e.g., Figure 2; Reynolds et al., 2006a). Surface transects and depth profiles have been studied in the Southern Ocean (Cardinal et al., 2005, 2007; De La Rocha et al., 2000; Varela et al., 2004), Monterey Bay, California (De La Rocha et al., 2000), the north

Pacific (Reynolds et al., 2006a) and the eastern equatorial Pacific (Beucher et al., 2008). In general, these studies support a -1.1‰ fractionation factor, however, such estimates come with a high degree of uncertainty due to poor control over the source and isotopic composition of the silicic acid used by the diatoms.

In particular, elucidating isotope fractionation (ϵ) from field samples is hampered by a lack of control over the degree of mixing between water bodies of different Si concentration and isotopic composition, and this process alone can explain the distribution observed in Figure 2 (Reynolds et al., 2006a). Because the majority of Si entering the ocean is assimilated by diatoms, which subsequently dissolve during sedimentation, the $\delta^{30}\text{Si}$ of the deep ocean has a largely homogenous $\delta^{30}\text{Si}$ of $+0.9\text{‰}$ roughly equal to that of the sum of inputs to the ocean (De La Rocha and Bickle, 2005; De La Rocha et al., 2000). Upwelling of this deep ocean silicic acid provides

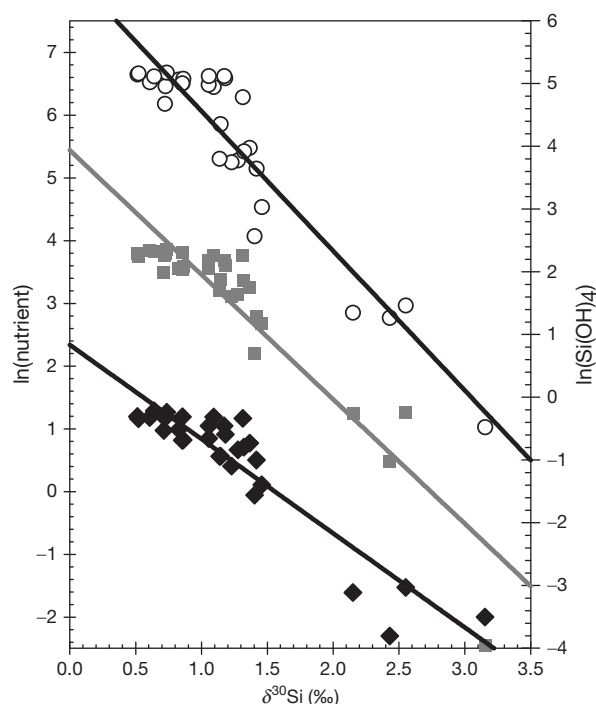


Figure 2 Variations in North Pacific seawater $\delta^{30}\text{Si}$ correlate with nutrient concentrations (in μM) due to mixing of nutrient depleted surface water with nutrient replete deep waters. Plotted as $\ln[\text{phosphate}]$ in black diamonds, $\ln[\text{nitrate}]$ in gray squares, and $\ln[\text{Si}(\text{OH})_4]$ in open circles with right axis (Reynolds et al., 2006a).

the raw material upon which diatom silicification depends, however in order to estimate ϵ from natural variations in $\delta^{30}\text{Si}$, an assumption is required either that the diatoms grew (a) in a closed system, whereby an initial pool of silicic acid is progressively assimilated with no further replenishment (as in batch culture), in which case isotope fractionation follows a Rayleigh-type distillation curve (Figure 1(a)), or (b) within an open system under steady state with continuous supply of Si and removal of diatoms (Figure 1(b); Cardinal et al., 2007; De La Rocha et al., 1997; 2000; Varela et al., 2004). For example, Varela et al. (2004) measured $\delta^{30}\text{Si}_{\text{DSi}}$ and $\delta^{30}\text{Si}_{\text{diatom}}$ from surface water samples in the Southern Ocean and generated estimates for ϵ using both open and closed system models which ranged between +1 and +2.1‰ depending on the parameters used (Varela et al., 2004). A similar comparison by Cardinal et al. (2007) suggests a correlation between ϵ and silicic acid concentration. However, uncertainties over the effect of mixing hampers attempts to rigorously validate the isotope fractionation factor using field comparisons, as well as raising uncertainty in interpreting paleo- $\delta^{30}\text{Si}_{\text{diatom}}$ changes.

Fractionation of Silicon Isotopes During Diatom Dissolution

Recycling of diatom silica is a key component of the global silicon cycle, with for example, 97% of diatoms living in the marine photic zone lost via dissolution during sinking (Ragueneau et al., 2000; Treguer et al., 1995). Dissolution

can also have a significant effect on diatom sedimentation and silicon cycling within lakes, particularly saline lakes and deep lakes such as Lake Baikal. Little is known about the effects of taphonomic processes, such as dissolution on the $\delta^{30}\text{Si}$ of sedimentary diatoms (Demarest et al., 2009). De La Rocha et al. (1998) reported an absence of any change in $\delta^{30}\text{Si}_{\text{diatom}}$ following a 26% dissolution of opal frustules. However, a more detailed experimental study reported that a systematic isotope effect accompanied the dissolution of diatom silica, with the accumulating dissolved silicon being consistently depleted in ^{30}Si by $\sim 0.55\%$ relative to the corresponding residual silica (Demarest et al., 2009; Figure 3). This fractionation, which essentially acts to reverse that which occurs during diatom growth, was unaffected by differences in temperature, sample composition or solubility (Demarest et al., 2009; Figure 3).

The implications of these findings are widespread. For example, dissolution could affect the spatial distribution of $\delta^{30}\text{Si}_{\text{DSi}}$ in the ocean, a process which has not been taken into account within existing models (e.g., Reynolds, 2009; Reynolds et al., 2006a; Wischmeyer et al., 2003). In addition, unknown dissolution effects will have compromised attempts to calibrate ϵ from field studies (Cardinal et al., 2005, 2007; De La Rocha et al., 2000; Reynolds et al., 2006a; Varela et al., 2004) and may explain apparent spatial variability in ϵ (Cardinal et al., 2007). Importantly, the effects of dissolution can also lead to an overestimation of past silicon utilization from sedimentary $\delta^{30}\text{Si}_{\text{diatom}}$ data (Demarest et al., 2009), particularly if the amount of dissolution changed with time. Demarest et al. (2009) suggest that given the analytical uncertainty of $\sim 0.1\%$ for $\delta^{30}\text{Si}$ measurements, dissolution differences of less than $\sim 20\%$ between samples are unlikely to impose measureable bias in their respective $\delta^{30}\text{Si}_{\text{diatom}}$. In addition, despite the large amount of silica lost to dissolution within the ocean (Treguer et al., 1995), it is possible that the diatoms which remain have undergone minimal dissolution (Demarest et al., 2009). Nevertheless, it is clear that attempts to quantify the amount of dissolution of diatom samples should be given a high priority in future research projects.

Factors Affecting Silicon Isotopes in Terrestrial Hydrological Systems

An understanding of the non-diatom related processes that affect $\delta^{30}\text{Si}_{\text{DSi}}$ in water is needed before sedimentary records of $\delta^{30}\text{Si}_{\text{diatom}}$ can be interpreted. An increasingly large volume of work has documented the wide natural variability of $\delta^{30}\text{Si}_{\text{DSi}}$ in terrestrial hydrological systems such as groundwaters, soil water, and rivers. Inorganic processes such as dissolution of crustal rock, precipitation of secondary minerals, adsorption of silicic acid onto secondary oxides of iron and aluminum, and subsequent dissolution of secondary minerals and oxides, present a complex array of factors which can alter the $\delta^{30}\text{Si}$ of silicic acid entering lakes and oceans (Basile-Doelsch, 2006; De La Rocha et al., 2000; Douthitt, 1982). Generally speaking, the $\delta^{30}\text{Si}_{\text{DSi}}$ of river water is usually higher than that of crustal rock (De La Rocha et al., 2000; Ding et al., 2004; Georg et al., 2006a). By contrast, secondary precipitates, adsorbed silicic acid and higher plant biogenic silica (phytoliths) have $\delta^{30}\text{Si}$

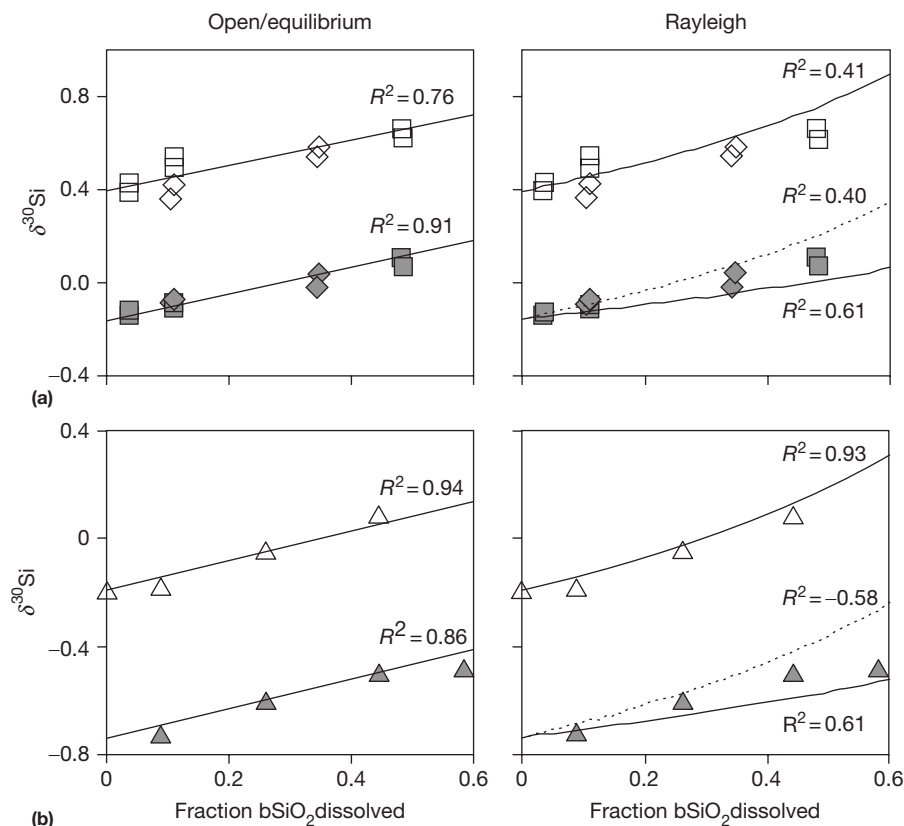


Figure 3 Effects of silica dissolution on $\delta^{30}\text{Si}_{\text{diatom}}$. Silicon isotope composition of residual diatom silica (open symbols) and accumulating $\text{Si}(\text{OH})_4(\text{aq})$ (filled symbols) as a function of the fraction of Southern Ocean surface water diatom silica dissolved during (a) single batch and (b) serial stage dissolution experiments in a closed system, as compared to expectations for an open kinetic/equilibrium (left panels) or closed kinetic Rayleigh (right panels) fractionation process. Squares, diamonds and triangles reflect different experiments and lines indicate the isotope trends expected for each open kinetic or equilibrium models (see Demarest et al., 2009 for details).

values lower than bedrock (Basile-Doelsch et al., 2005; Ding et al., 2008; Opfergelt et al., 2009). Subsequent dissolution of secondary minerals and biogenic silica therefore generates silicic acid with lower $\delta^{30}\text{Si}$ than that derived from primary weathering. Therefore, the $\delta^{30}\text{Si}_{\text{DSi}}$ of water entering lakes and oceans can largely be explained by the relative influence of these different silicon sources and pathways (Georg et al., 2006a, 2009a).

For example, in the Verzasca River, Switzerland, high $\delta^{30}\text{Si}_{\text{DSi}}$ was observed to correspond with periods of low discharge and high base cation flux (Georg et al., 2006a). This was interpreted to be due to isotopic fractionation during precipitation of secondary clays within shallow groundwaters and soils. By contrast, periods of high river discharge had low $\delta^{30}\text{Si}$ and low base cation flux characteristic of overland runoff, where the low $\delta^{30}\text{Si}_{\text{DSi}}$ values reflect the original weathering of crustal bedrock (Georg et al., 2006a). Groundwaters in the Bengal Basin, India (Georg et al., 2009a), and in the Navajo Sandstone aquifer, Arizona (Georg et al., 2009b) were found to have low $\delta^{30}\text{Si}_{\text{DSi}}$ (as low as -1.42‰ in Arizona, -0.2‰ in Bengal) relative to regional river waters; however, their silicic acid concentration was higher. These low $\delta^{30}\text{Si}_{\text{DSi}}$ values were interpreted to reflect dissolution of secondary clays or silcretes within the aquifer leading to a progressive decrease in $\delta^{30}\text{Si}_{\text{DSi}}$ along the flowpath (Georg et al., 2009b). Where biological

uptake plays an important role in the silicon cycle, interpreting the $\delta^{30}\text{Si}_{\text{DSi}}$ of freshwaters becomes increasingly complex and difficult to disentangle (Georg et al., 2006a). In heavily vegetated systems, assimilation of silica by higher plants can have a significant effect on the $\delta^{30}\text{Si}_{\text{DSi}}$ in river water (e.g., Ding et al., 2004).

Applications of $\delta^{30}\text{Si}_{\text{diatom}}$ to Quaternary Marine Sediments

Understanding temporal patterns in marine primary productivity and organic carbon drawdown is critical for understanding glacial–interglacial changes in atmospheric $p\text{CO}_2$. In this respect, studies of marine $\delta^{30}\text{Si}_{\text{diatom}}$ records have had a significant impact, despite uncertainties concerning the controls over silicon isotopes in the contemporary environment (Beucher et al., 2007; Brzezinski et al., 2002; De La Rocha et al., 1998; Hendry et al., 2010; Pichevin et al., 2009). For example, $\delta^{30}\text{Si}_{\text{diatom}}$ records from sediments collected south of the Antarctic polar front show declines of $\sim 1\text{‰}$ at the onset of the Last Glacial Maximum (LGM; De La Rocha et al., 1998) and several previous glacials (Brzezinski et al., 2002; Figure 4). One interpretation is that these data reflect a reduction in diatom productivity in the Southern Ocean during glacial maxima due

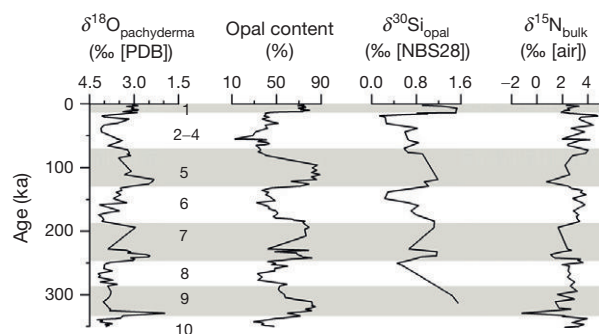


Figure 4 Southern Ocean sediments demonstrate enhanced silicic acid uptake during glacial periods, inversely correlated with nitrate utilization ($\delta^{15}\text{N}$; Brzezinski et al., 2002), $\delta^{18}\text{O}$ (*N. pachyderma*; Charles et al., 1991), opal content (Charles et al., 1991), $\delta^{30}\text{Si}$ of diatom opal and $\delta^{15}\text{N}$ of bulk sediment (Rau and Froelich, 1993) from RC13-259 (53°53' S; 4°56' W; 2677 m). Numbers indicate oxygen isotope stages 1–10 (Brzezinski et al., 2002). Gray shading indicates interglacial periods.

to expansion of sea ice cover (De La Rocha et al., 1998). However, the $\delta^{30}\text{Si}_{\text{diatom}}$ evidence contrasts with the nitrogen isotope ratio of organic matter ($\delta^{15}\text{N}_{\text{org}}$; Figure 4), which suggest an increase in nitrate utilization (increased productivity) during the glacials (Brzezinski et al., 2002). This inverse relationship between $\delta^{30}\text{Si}_{\text{diatom}}$ and $\delta^{15}\text{N}_{\text{org}}$ can be explained by a reduction of silicon uptake in diatoms relative to carbon and nitrogen assimilation due to increased iron availability during glacial periods (Hutchins and Bruland, 1998). Today, the Southern Ocean is iron limited and the ratio of Si:N uptake by phytoplankton is ~4:1 (Pondaven et al., 2000). However, under saturating light and nutrient conditions, and during Fe fertilization experiments, ratios of ~1:1 have been observed (Coale et al., 2004). It is proposed that desert expansion during glacial maxima increased the flux of iron-rich eolian dust which fertilized the ocean. However, although productivity (carbon and nitrogen uptake) increased, the higher iron concentration led to a reduction in diatom silicic acid uptake (Figure 4; Brzezinski et al., 2002; Matsumoto et al., 2002).

One potential outcome of this proposed decoupling between primary productivity and silicification during glacial maxima is an excess of unused silicic acid within the Southern Ocean surface waters. The silicic acid leakage hypothesis (Brzezinski et al., 2002; Matsumoto et al., 2002) proposes that this excess silicon was transported northwards, via the sub-Antarctic mode water, where it provided a competitive advantage to diatoms over coccolithophores in the mid- to low latitudes. In turn, this led to a global reduction in calcite biomineralization, raising ocean alkalinity and increasing atmospheric CO_2 drawdown (Brzezinski et al., 2002; Matsumoto et al., 2002). In support of this hypothesis, a sediment core from the sub-Antarctic sector of the Indian Ocean exhibits decreases in $\delta^{30}\text{Si}_{\text{diatom}}$ which suggest that excess silicic acid reached this site during the LGM (Beucher et al., 2007). However, similar changes in $\delta^{30}\text{Si}_{\text{diatom}}$ are not apparent in the subtropics, making it unclear whether the pool of silicic acid extended into the subtropical region during glacial periods (Beucher et al., 2007; Crosta et al., 2007).

Silicon leakage may have also occurred from the tropical Pacific Ocean during glacial times. Iron fertilization via

airborne dust is invoked to explain low $\delta^{30}\text{Si}_{\text{diatom}}$ values during the LGM in the Eastern Equatorial Pacific (Pichevin et al., 2009). In this instance, a reduction in biogenic silica sedimentation rate, alongside lower $\delta^{30}\text{Si}_{\text{diatom}}$ values, coincides with an increase in the deposition of organic carbon during the LGM, indicating enhanced carbon export during this period (Figure 5). Estimated silicic acid and nitrate uptake rates, based on silicon and nitrogen isotopes, suggest that neither of these nutrients limited productivity during the LGM (Pichevin et al., 2009). Instead, it is possible that phosphate may have played a limiting role in the tropical glacial ocean, which is a marked contrast to the situation in the contemporary ocean (Pichevin et al., 2009).

Although interpretations of $\delta^{30}\text{Si}_{\text{diatom}}$ such as these are compelling, alternative explanations can be offered to explain variability in $\delta^{30}\text{Si}_{\text{diatom}}$ over Quaternary timescales. One major limitation to using $\delta^{30}\text{Si}$ as a proxy for past diatom productivity is the potential variability in the concentration and $\delta^{30}\text{Si}$ of upwelling silicic acid. At present, the deep ocean is thought to be isotopically homogenous (Cardinal et al., 2005), however given that the oceanic residence time of Si is roughly 15000 years (Treguer et al., 1995) and a full glacial-interglacial cycle covers 100000 years, some portion of the observed variability in $\delta^{30}\text{Si}_{\text{diatom}}$ may be due to changes in the flux of silicon to the ocean via terrestrial weathering. The concentration of silicic acid in upwelling waters can be estimated using the $\delta^{30}\text{Si}$ of benthic sponge spicules (Hendry et al., 2010; Wille et al., 2010). Sponge spicule $\delta^{30}\text{Si}$ records from sediments beneath the Southern Ocean appear to demonstrate that the concentration and $\delta^{30}\text{Si}_{\text{DSi}}$ of upwelling water at the LGM exhibited no significant difference from contemporary conditions (Ellwood et al., 2010; Hendry et al., 2010), thus supporting the interpretation of $\delta^{30}\text{Si}_{\text{diatom}}$ as a proxy for silicic acid uptake in the photic zone (Ellwood et al., 2010; Hendry et al., 2010).

Additional uncertainties concern the effect of species change on $\delta^{30}\text{Si}_{\text{diatom}}$, since any major shift in the biogeochemistry of surface waters will undoubtedly be accompanied by a floristic shift (De La Rocha, 2006). Although culture experiments have demonstrated no significant difference between the isotopic fractionation by different taxa (De La Rocha et al., 1997) changes in diatom species, associated with different bloom forming patterns and seasonal preferences, could lead to seasonal bias within sedimentary $\delta^{30}\text{Si}_{\text{diatom}}$ records. Such issues could be addressed by species-specific analyses of diatom remains. Alternatively, where laminated sediments are available, microstratigraphic analysis on a subannual scale would enable an estimate of the seasonal weighting given by different species, relative to the bulk sample.

$\delta^{30}\text{Si}_{\text{diatom}}$ in Lakes and Lake Sediments

To date, very little research has been published on silicon isotopes in either lakes or lake sediments. Alleman et al. (2005) investigated the seasonal pattern of $\delta^{29}\text{Si}_{\text{DSi}}$ in Lake Tanganyika, in conjunction with analyses of tributary waters and five $\delta^{29}\text{Si}_{\text{diatom}}$ measurements from lake diatom phytoplankton. In addition, Street-Perrott et al. (2008) and Swann et al. (2010) generated late-glacial and Holocene

$\delta^{30}\text{Si}_{\text{diatom}}$ records from Lake Rutundu, Kenya and Lake El'gytgytyn, Siberia, respectively. In principle, the processes which control lacustrine $\delta^{30}\text{Si}_{\text{diatom}}$ are the same as those that operate in the ocean (Alleman et al., 2005); thus, there is potential to use silicon isotopes to trace aquatic biogeochemical change in freshwaters, with significance for understanding global and regional carbon cycling (Street-Perrott and Barker, 2008). However, while the deep ocean environment can be considered to be buffered against short-term variability in terrestrial inputs and processes (De La Rocha and Bickle, 2005), the small volume of lakes, and their short residence time for Si, leads to a much greater sensitivity to silicon isotope changes originating from the surrounding catchment (Street-Perrott et al., 2008). Lake $\delta^{30}\text{Si}_{\text{diatom}}$ records could therefore offer a valuable archive of changes in terrestrial weathering processes.

Large lakes are more likely to yield data relevant to our understanding of freshwater biogeochemical silicon cycling in the past, as their greater silicon and water residence times mean the system is more analogous to the marine environment. For example, the long residence time of silica within Lake Tanganyika means that the surface silica budget is largely a function of vertical mixing and silicon recycling, while Si loading from tributaries accounts for 2–3% at most (Alleman et al., 2005). Water depth $\delta^{29}\text{Si}$ profiles from Lake Tanganyika are similar to those analyzed in sea water, with low $\delta^{29}\text{Si}_{\text{DSi}}$ values in deep water (mean $\delta^{29}\text{Si}_{\text{DSi}} = +0.61 \pm 0.05\text{‰}$, calculated $\delta^{30}\text{Si}_{\text{DSi}} = 1.18\text{‰}$; eqn [2]) and high values in the epilimnion due to uptake of silicic acid by diatoms (Alleman et al., 2005). Biweekly monitoring of Si concentration and $\delta^{29}\text{Si}_{\text{DSi}}$ over 1 year demonstrated three periods of silicon transport to the lake, characterized by increases in Si concentration and low $\delta^{29}\text{Si}_{\text{DSi}}$ values, followed by a decrease in concentration and increase in $\delta^{29}\text{Si}_{\text{DSi}}$ related to diatom productivity. In addition, a comparison between $\delta^{29}\text{Si}_{\text{diatom}}$ from five phytoplankton samples and $\delta^{29}\text{Si}_{\text{DSi}}$ from surrounding water generated estimates of ϵ close to the -1.1‰ reported for marine diatoms (Alleman et al., 2005).

Street-Perrott et al. (2008) conducted the first analyses of silicon isotopes from lake sediments as part of a multiproxy approach studying the terrestrial biogeochemical cycle of silicon at Lake Rutundu, Kenya. Between 37 and 18 ka, $\delta^{30}\text{Si}_{\text{diatom}}$ varied between -1.3 and $+0.5\text{‰}$, with most samples yielding near-maximum values, but with two notable minima at ~ 33 and 23 ka (Street-Perrott et al., 2008). The highest values of $+0.5\text{‰}$ fall within the expected range of surface runoff in this region, which could indicate complete utilization of silicic acid within the lake. However, the low $\delta^{30}\text{Si}_{\text{diatom}}$ values fall beyond the expected range for Rayleigh fractionation and possibly reflect enhanced groundwater influences on lake water (e.g., Georg et al., 2009a,b) during periods of low lake level (Street-Perrott et al., 2008).

Changes in $\delta^{30}\text{Si}_{\text{diatom}}$ within the sediments of Lake El'gytgytyn, Siberia, are interpreted as reflecting changes in silicic acid availability with time, owing to the restricted hydrological and weathering activity in this arctic region, and due to the oligotrophic nature of the lake (Swann et al., 2010). During the late-glacial and Holocene (~ 23 ka to present), $\delta^{30}\text{Si}_{\text{diatom}}$ exhibits marked changes – in particular higher values coinciding with the LGM (~ 20 ka) and Greenland Stadial 1 (a.k.a. Younger

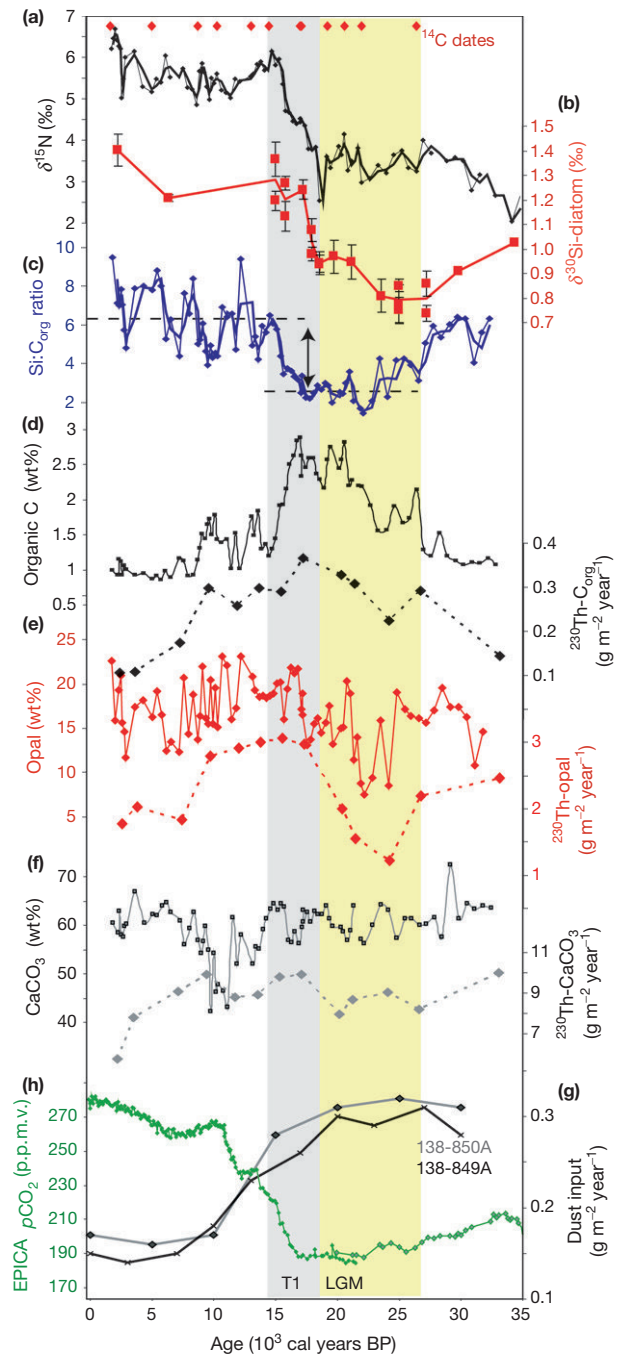


Figure 5 A record of decreasing productivity but increasing silicification in the Eastern Equatorial Pacific during the transition from the Last Glacial Maximum (LGM) to the Holocene (Pichevin et al., 2009). Data from Core 1240, taken in the eastern Pacific Ocean, west of Equator. (a) $\delta^{15}\text{N}$ signal of bulk material (bold line indicates 2-point average); (b) $\delta^{30}\text{Si}_{\text{diatom}}$ (error bars 1σ); (c) Si:C ratio (bold line indicates 2-point average); (d)–(f) elemental concentrations (solid lines) and ^{230}Th -normalized accumulation rates (dotted lines) of (d) organic carbon, (e) opal and (f) carbonate; (g) dust fluxes in Core ODP 138-849A and 138-850A (McGee et al., 2007); (h) atmospheric $p\text{CO}_2$ record from EPICA dome C ice core (Monnin et al., 2001) in parts per million by volume. T1 = termination 1.

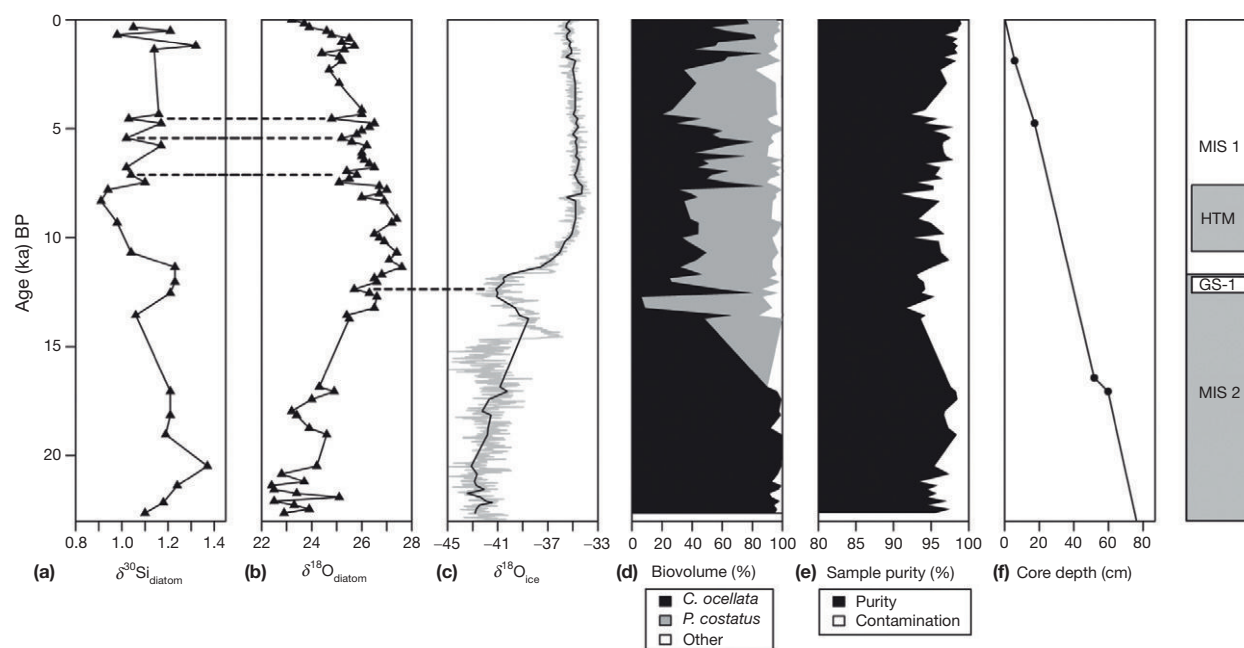


Figure 6 Climate and environmental change in north-eastern Siberia (Swann et al., 2010), as recorded by changes in (a) $\delta^{30}\text{Si}_{\text{diatom}}$ and (b) $\delta^{18}\text{O}_{\text{diatom}}$ in Lake El'gygytyn; (c) corresponding record from Greenland for comparison, NGRIP $\delta^{18}\text{O}_{\text{ice}}$ (NGRIP Project Members, 2004); (d) diatom species biovolumes; (e) diatom sample purity in the analyzed material and (f) ^{14}C age model for site Lz1029. Gray line for $\delta^{18}\text{O}_{\text{ice}}$ indicates the original NGRIP data, black line is a local polynomial regression (Loess) used to estimate comparable values of $\delta^{18}\text{O}_{\text{ice}}$ for the $\delta^{18}\text{O}_{\text{diatom}}$ data using a smoothing window to reflect the resolution of the sediment record. Age model uses a linear interpolation between calibrated ^{14}C dates (plotted) that incorporate a 1.3 ka reservoir effect correction. HTM on zonation indicates duration of 'Holocene thermal maximum' in Lake El'gygytyn. GS-1 indicates the short period of climatic cooling in Lake El'gygytyn that occurs during the GS-1 event in the Greenland ice core record.

Dryas; ~ 12.5 ka). Low $\delta^{30}\text{Si}_{\text{diatom}}$ values coincide with relatively warmer periods, namely the Holocene Thermal Maximum (10–9 ka BP; Swann et al., 2010). Swann et al. (2010) argue that during cold climates, deep permafrost, reduced weathering and limited moisture availability resulted in lower concentrations but a more complete consumption of silicic acid, which in turn led to higher $\delta^{30}\text{Si}_{\text{diatom}}$ values (Figure 6).

Analysis of $\delta^{30}\text{Si}_{\text{diatom}}$ from lake sediments offers considerable potential for obtaining extra information regarding the biogeochemical cycles of paleo- and contemporary lakes, in particular, when coupled with other isotopic and microfossil proxies. The ability to analyze $\delta^{30}\text{Si}_{\text{diatom}}$ alongside $\delta^{18}\text{O}_{\text{diatom}}$ using the same sample (Leng and Sloane, 2008) is likely to result in a marked increase in applications using lake sediments. However, continued research into $\delta^{30}\text{Si}$ patterns within contemporary lakes is required in order to better understand the relative controls over the isotope ratios in the sedimentary record.

Conclusion

Understanding changes in weathering, nutrient cycles and the cycling of carbon within aquatic systems is one of the key objectives of Quaternary research, due to the importance of these processes in driving global climate change. In this respect, silicon isotope analysis of diatom silica offers a valuable tool, particularly within a multiproxy framework. Where changes in the flux and isotopic composition of silicic acid remain constant with time, $\delta^{30}\text{Si}_{\text{diatom}}$ should reflect changes

in diatom silicic acid utilization (De La Rocha et al., 1997), dampened by the effects of diatom dissolution in some instances (Demarest et al., 2009). This interpretation of marine $\delta^{30}\text{Si}_{\text{diatom}}$ records has had considerable influence on the debate over the role of the oceanic carbon pump in driving long-term changes in atmospheric carbon dioxide concentration (e.g., Brzezinski et al., 2002). However, $\delta^{30}\text{Si}_{\text{diatom}}$ can also vary as a function of changes in the terrestrial silicon cycle (Basile-Doelsch, 2006), which potentially compromises the use of $\delta^{30}\text{Si}_{\text{diatom}}$ as a proxy for diatom silicic acid utilization. Such effects are particularly relevant where lake sediments are concerned, and in this respect, $\delta^{30}\text{Si}_{\text{diatom}}$ records may instead provide useful archives of past weathering processes and nutrient mobility (Street-Perrott et al., 2008). However, continued research into the controls over silicon isotopes in all facets of the contemporary environment is required in order to realize this potential.

See also: Diatom Introduction. Diatom Methods: $\delta^{18}\text{O}$ Records. Diatom Records: Diatom Fossil Records from Marine Laminated Sediments; Freshwater Laminated Sequences. Glaciations: Late Pleistocene Glacial Events in Beringia.

References

- Alexander CMO, Taylor S, Delaney JS, Ma PX, and Herzog GF (2002) Mass-dependent fractionation of Mg, Si, and Fe isotopes in five stony cosmic spherules. *Geochimica et Cosmochimica Acta* 66(1): 173–183.

- Alleman LY, Cardinal D, Cocquyt C, et al. (2005) Silicon isotopic fractionation in Lake Tanganyika and its main tributaries. *Journal of Great Lakes Research* 31(4): 509–519.
- Basile-Doelsch I (2006) Si stable isotopes in the Earth's surface: A review. *Journal of Geochemical Exploration* 88(1–3): 252–256.
- Basile-Doelsch I, Meunier JD, and Parron C (2005) Another continental pool in the terrestrial silicon cycle. *Nature* 433(7024): 399–402.
- Beucher CP, Brzezinski MA, and Crosta X (2007) Silicic acid dynamics in the glacial sub-Antarctic: Implications for the silicic acid leakage hypothesis. *Global Biogeochemical Cycles* 21(3): GB3015.
- Beucher CP, Brzezinski MA, and Jones JL (2008) Sources and biological fractionation of silicon isotopes in the Eastern Equatorial Pacific. *Geochimica et Cosmochimica Acta* 72(13): 3063–3073.
- Brzezinski MA, Pride CJ, Franck VM, et al. (2002) A switch from $\text{Si}(\text{OH})_4$ to NO_3^- depletion in the glacial Southern Ocean. *Geophysical Research Letters* 29(12): 1564.
- Brzezinski MA, Jones JL, Beucher CP, Demarest MS, and Berg HL (2006) Automated determination of silicon isotope natural abundance by the acid decomposition of cesium hexafluorosilicate. *Analytical Chemistry* 78(17): 6109–6114.
- Cardinal D, Alleman LY, de Jong J, Ziegler K, and Andre L (2003) Isotopic composition of silicon measured by multicollector plasma source mass spectrometry in dry plasma mode. *Journal of Analytical Atomic Spectrometry* 18(3): 213–218.
- Cardinal D, et al. (2005) Relevance of silicon isotopes to Si-nutrient utilization and Si-source assessment in Antarctic. *Global Biogeochemical Cycles* 19(2): GB2007.
- Cardinal D, et al. (2007) Silicon isotopes in spring Southern Ocean diatoms: Large zonal changes despite homogeneity among size fractions. *Marine Chemistry* 106(1–2): 46–62.
- Charles CD, Froelich PN, Zibello MA, Mortlock RA, and Morley JJ (1991) Biogenic opal in Southern Ocean sediments over the last 450,000 years: Implications for surface water chemistry and circulation. *Paleoceanography* 6: 697–728.
- Chmieleff J, Horn I, Steinhofel G, and von Blanckenburg F (2008) In situ determination of precise stable Si isotope ratios by UV-femtosecond laser ablation high-resolution multi-collector ICP-MS. *Chemical Geology* 249(1–2): 155–166.
- Coale KH, et al. (2004) Southern ocean iron enrichment experiment: Carbon cycling in high- and low-Si waters. *Science* 304(5669): 408–414.
- Conley DJ, Kilham SS, and Theriot E (1989) Differences in silica content between marine and fresh-water diatoms. *Limnology and Oceanography* 34(1): 205–213.
- Crosta X, Beucher C, Pahnke K, and Brzezinski MA (2007) Silicic acid leakage from the Southern Ocean: Opposing effects of nutrient uptake and oceanic circulation. *Geophysical Research Letters* 34(13): L13601.
- De Freitas ASW, McCulloch AW, and McInnes AG (1991) Recovery of silica from aqueous silicate solutions via trialkyl or tetraalkylammonium silicomolybdate. *Canadian Journal of Chemistry – Revue Canadienne de Chimie* 69(4): 611–614.
- De La Rocha CL (2002) Measurement of silicon stable isotope natural abundances via multicollector inductively coupled plasma mass spectrometry (MC-ICP-MS). *Geochemistry, Geophysics, Geosystems* 3: 1045–1052.
- De La Rocha CL (2006) Opal-based isotopic proxies of paleoenvironmental conditions. *Global Biogeochemical Cycles* 20: GB4S09.
- De La Rocha CL and Bickle MJ (2005) Sensitivity of silicon isotopes to whole-ocean changes in the silica cycle. *Marine Geology* 217(3–4): 267–282.
- De La Rocha CL, Brzezinski MA, and De Niro MJ (1996) Purification, recovery, and laser-driven fluorination of silicon from dissolved and particulate silica for the measurement of natural stable isotope abundances. *Analytical Chemistry* 68(21): 3746–3750.
- De La Rocha CL, Brzezinski MA, and De Niro MJ (1997) Fractionation of silicon isotopes by marine diatoms during biogenic silica formation. *Geochimica et Cosmochimica Acta* 61(23): 5051–5056.
- De La Rocha CL, Brzezinski MA, De Niro MJ, and Shemesh A (1998) Silicon-isotope composition of diatoms as an indicator of past oceanic change. *Nature* 395(6703): 680–683.
- De La Rocha CL, Brzezinski MA, and De Niro MJ (2000) A first look at the distribution of the stable isotopes of silicon in natural waters. *Geochimica et Cosmochimica Acta* 64(14): 2467–2477.
- Demarest MS, Brzezinski MA, and Beucher CP (2009) Fractionation of silicon isotopes during biogenic silica dissolution. *Geochimica et Cosmochimica Acta* 73(19): 5572–5583.
- Ding T, et al. (1996) *Silicon Isotope Geochemistry*. Beijing: Geological Publishing House.
- Ding T, Wan D, Wang C, and Zhang F (2004) Silicon isotope compositions of dissolved silicon and suspended matter in the Yangtze River, China. *Geochimica et Cosmochimica Acta* 68(2): 205–216.
- Ding TP, et al. (2008) Silicon isotope fractionation between rice plants and nutrient solution and its significance to the study of the silicon cycle. *Geochimica et Cosmochimica Acta* 72(23): 5600–5615.
- Douthitt CB (1982) The geochemistry of the stable isotopes of silicon. *Geochimica et Cosmochimica Acta* 46: 1449–1458.
- Ellwood MJ, Wille M, and Maher W (2010) Glacial silicic acid concentrations in the Southern Ocean. *Science* 330(6007): 1088–1091.
- Gao JF and Ding TP (2009) Laser microprobe silicon isotope analysis method and geology applications. *Geochimica et Cosmochimica Acta* 73(13): A413.
- Georg RB, Reynolds BC, Frank M, and Halliday AN (2006a) Mechanisms controlling the silicon isotopic compositions of river waters. *Earth and Planetary Science Letters* 249(3–4): 290–306.
- Georg RB, Reynolds BC, Frank M, and Halliday AN (2006b) New sample preparation techniques for the determination of Si isotopic compositions using MC-ICPMS. *Chemical Geology* 235(1–2): 95–104.
- Georg RB, West AJ, Basu AR, and Halliday AN (2009a) Silicon fluxes and isotope composition of direct groundwater discharge into the Bay of Bengal and the effect on the global ocean silicon isotope budget. *Earth and Planetary Science Letters* 283(1–4): 67–74.
- Georg RB, Zhu C, Reynolds BC, and Halliday AN (2009b) Stable silicon isotopes of groundwater, feldspars, and clay coatings in the Navajo Sandstone aquifer, Black Mesa, Arizona, USA. *Geochimica et Cosmochimica Acta* 73(8): 2229–2241.
- Hendry KR, Georg RB, Rickaby REM, Robinson LF, and Halliday AN (2010) Deep ocean nutrients during the Last Glacial Maximum deduced from sponge silicon isotopic compositions. *Earth and Planetary Science Letters* 292(3–4): 290–300.
- Hutchins DA and Bruland KW (1998) Iron-limited diatom growth and Si:N uptake ratios in a coastal upwelling regime. *Nature* 393(6685): 561–564.
- Leng MJ and Sloane HJ (2008) Combined oxygen and silicon isotope analysis of biogenic silica. *Journal of Quaternary Science* 23: 313–319.
- Leng MJ, et al. (2009) The potential use of silicon isotope composition of biogenic silica as a proxy for environmental change. *Silicon* 1: 65–77.
- Matsumoto KK, Sarmiento JL, and Brzezinski MA (2002) Silicic acid leakage from the Southern Ocean: A possible explanation for glacial atmospheric pCO_2 . *Global Biogeochemical Cycles* 16: 1031–1053.
- McGee D, Marcantonio F, and Lynch-Stieglitz J (2007) Deglacial changes in dust flux in the eastern equatorial Pacific. *Earth and Planetary Science Letters* 257(1–2): 215–230.
- Milligan AJ, Varela DE, Brzezinski MA, and Morel F (2004) Dynamics of silicon metabolism and silicon isotopic discrimination in a marine diatom as a function of pCO_2 . *Limnology and Oceanography* 49(2): 322–329.
- Monnin E, et al. (2001) Atmospheric CO_2 concentrations over the last glacial termination. *Science* 291(5501): 112–114.
- NGRIP Project Members (2004) High resolution record of Northern Hemisphere climate extending into the last interglacial period. *Nature* 431: 147–151.
- Opergelt S, et al. (2009) Impact of soil weathering degree on silicon isotopic fractionation during adsorption onto iron oxides in basaltic ash soils, Cameroon. *Geochimica et Cosmochimica Acta* 73(24): 7226–7240.
- Picchevin LE, et al. (2009) Enhanced carbon pump inferred from relaxation of nutrient limitation in the glacial ocean. *Nature* 459: 1114–1117.
- Pondaven P, Ruiz-Pino D, Fravallo C, Treguer P, and Jeandel C (2000) Interannual variability of Si and N cycles at the time-series station KERFIX between 1990 and 1995 – A 1-D modelling study. *Deep Sea Research Part I: Oceanographic Research Papers* 47(2): 223–257.
- Ragueneau O, et al. (2000) A review of the Si cycle in the modern ocean: Recent progress and missing gaps in the application of biogenic opal as a paleoproductivity proxy. *Global and Planetary Change* 26(4): 317–365.
- Rau GH and Froelich PN (1993) A 450 kyr record of bulk sediment $\text{d}^{13}\text{C}_{\text{org}}$ and d^{15}N at 54oS: Significant glacial–interglacial changes in Southern Ocean C and N supply/demand? *Transactions of the American Geophysical Union* 74: 998.
- Reynolds BC (2009) Modeling the modern marine delta Si-30 distribution. *Global Biogeochemical Cycles* 23: GB2015.
- Reynolds BC, Frank M, and Halliday AN (2006a) Silicon isotope fractionation during nutrient utilization in the North Pacific. *Earth and Planetary Science Letters* 244(1–2): 431–443.
- Reynolds BC, Georg RB, Oberli F, Wiechert U, and Halliday AN (2006b) Re-assessment of silicon isotope reference materials using high-resolution multi-collector ICP-MS. *Journal of Analytical Atomic Spectrometry* 21(3): 266–269 GB2015.
- Reynolds BC, et al. (2007) An inter-laboratory comparison of Si isotope reference materials. *Journal of Analytical Atomic Spectrometry* 22(5): 561–568.
- Rosman KJR and Taylor PDP (1998) Isotopic compositions of the elements 1997. *Pure and Applied Chemistry* 70(1): 217–235.
- Round FE, Crawford RM, and Mann DG (1990) *The Diatoms: Biology and Morphology of the Genera*. Cambridge: Cambridge University Press.

- Spadero PA (1983) *Silicon Isotope Fractionation by the Marine Diatom Phaeodactylum tricornutum*. Chicago: University of Chicago.
- Street-Perrott FA and Barker PA (2008) Biogenic silica: A neglected component of the coupled global continental biogeochemical cycles of carbon and silicon. *Earth Surface Processes and Landforms* 33(9): 1436–1457.
- Street-Perrott FA, et al. (2008) Towards an understanding of late Quaternary variations in the continental biogeochemical cycle of silicon: Multi-isotope and sediment-flux data for Lake Rutundu, Mt Kenya, East Africa, since 38 ka BP. *Journal of Quaternary Science* 23(4): 375–387.
- Swann GEA, et al. (2010) A combined oxygen and silicon diatom isotope record of Late Quaternary change in Lake El'gygytyn, North East Siberia. *Quaternary Science Reviews* 29(5–6): 774–786.
- Tilles D (1961) Variations of silicon isotope ratios in a zoned pegmatite. *Journal of Geophysical Research* 66(9): 3015–3020.
- Treguer P, et al. (1995) The silica balance in the world ocean – A reestimate. *Science* 268(5209): 375–379.
- van den Boorn S, et al. (2006) Determination of silicon isotope ratios in silicate materials by high-resolution MC-ICP-MS using a sodium hydroxide sample digestion method. *Journal of Analytical Atomic Spectrometry* 21(8): 734–742.
- Varela DE, Pride CJ, and Brzezinski MA (2004) Biological fractionation of silicon isotopes in Southern Ocean surface waters. *Global Biogeochemical Cycles* 18: GB1047.
- Wille M, et al. (2010) Silicon isotopic fractionation in marine sponges: A new model for understanding silicon isotopic variations in sponges. *Earth and Planetary Science Letters* 292(3–4): 281–289.
- Wischmeyer AG, De La Rocha CL, Maier-Reimer E, and Wolf-Gladrow DA (2003) Control mechanisms for the oceanic distribution of silicon isotopes. *Global Biogeochemical Cycles* 17: 1083–1095.

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PALEOCEANOGRAPHY

Paleoceanography An Overview

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Introduction

Paleoceanographic records derived from marine sediment cores provide an ideal means to reconstruct climate and environmental change during the Quaternary. Deep sea sediment records are usually continuous and contain a mixture of terrigenous sediment (mostly clay) and plankton fossil remains, each providing many different proxies of environmental change ([Figure 1](#)). Typical deep sea sediments accumulate at rates of 1–5 cm ka⁻¹; a 10-m piston core contains a record of the past several hundred thousand years. Longer records are obtained from ocean drilling by the Deep Sea Drilling Project and its successors, the Ocean Drilling Program and Integrated Ocean Drilling Program. While initial sediment coring and deep sea drilling focused on the deep sea where accumulation rates are low, exciting progress in the past two decades has been made coring continental margins, drift deposits, and basins where high sedimentation rates (1 m ka⁻¹) and/or low sediment mixing allow rapid changes in climate and ocean circulation to be resolved.

The effect of bioturbation (mixing by worms and other organisms at the seafloor) is an important consideration in higher resolution studies of climate change. Bioturbation can thoroughly mix 10 cm of sediment (equivalent to 1000–2000 years at typical sedimentation rates) before the sediment is buried below the reach of benthic organisms. Bioturbation is sometimes evident by mottled coloring of sediments, as seen in core photographs ([Figure 1](#)). The smoothing effect is greatest when the rate of change is fast, the change is large, and the sedimentation rate is low. Milankovitch cycles (100–23-ka period) are readily preserved at typical deep sea sedimentation rates, while millennial-scale cycles are often attenuated. To minimize the effects of bioturbation, studies have exploited high sedimentation rate cores or cores from anoxic marine basins where benthic activity is minimal. Under the highest sedimentation rates, in the absence of sediment mixing, finely sampled marine cores appear capable of resolving events occurring within a few years or even a single year.

Oxygen Isotope Stratigraphy

Beyond the limit of radiocarbon (approximately 40000 years BP), the stratigraphic framework for the study of the Quaternary in marine sediment cores has been derived by measuring the oxygen isotope ratio in marine carbonates (see [Oxygen-](#)

[Isotope Stratigraphy of the Oceans](#)). Two stable isotopes of oxygen exist, ¹⁸O and ¹⁶O, and their ratio in marine carbonates varies according to the ratio in seawater and the temperature at which the carbonate is secreted. The seawater ¹⁸O/¹⁶O ratio is sensitive to climate and particularly to the amount of ice stored in continental ice sheets. Oxygen isotope variations found in marine sediments are thus related to climate, and the variations through time are dominated by cycles that can be correlated with the orbital cycles in solar radiation. The orbital cycles in solar radiation can be calculated with high accuracy independently of the sediment record (see [Astronomical Theory of Paleoclimates](#)). The offset between the orbital cycles and the oxygen isotope record can be determined empirically from sediment cores by radiocarbon dating the last 30000 years and from U–Th dating of the coral terraces that indicate sea level maxima (and oxygen isotope minima) during marine isotope stages 3, 5, and 7 ([Imbrie et al., 1984](#)). The offset can alternately be determined from a model of ice sheet growth ([Lisiecki and Raymo, 2005](#)). Using this offset, a chronology for the oxygen isotope series (usually a composite (stacked) record of oxygen isotope variations) can be developed by aligning the oxygen isotope record to the orbital variations. In practice, chronologies for new cores are produced by aligning an oxygen isotope record to a stacked $\delta^{18}\text{O}$ time series, such as the one produced by [Lisiecki and Raymo \(2005\)](#). The orbital tuning approach using stacked oxygen isotope records has now been extended back more than 5.3 Ma, providing numbered isotope events and a chronology with uncertainty of 6000 years or less for the Quaternary.

Proxies

Over the past four decades, paleoceanographers have developed an extensive array of proxies of past climate and environment ([Table 1](#)), developed principally from proxies found in whole sediment and from the geochemical composition of plankton shells such as the calcium carbonate skeleton of the foraminifer *Globigerina bulloides* ([Figure 2](#)). The attributes of foraminifer shells make them ideally suited to record environmental information. Planktonic foraminifers inhabit and record variations in the sea surface environment (with possible species-specific seasonal and depth habitats; see [Planktic Foraminifera](#)), while benthic foraminifers record the environment at the seafloor or within the uppermost sediment layers (see [Benthic Foraminifera](#)). Foraminifer shells are composed

of calcium carbonate (CaCO_3), incorporating the isotopes of oxygen and carbon, and the element calcium, as well as trace metals that substitute for calcium or become incorporated into the calcite lattice. Almost all proxies are subject to multiple influences, and while the relationship between oxygen isotope ratios in modern specimens and modern ocean temperatures can be constrained, the possibility remains that other influences have affected the proxy in the past. Oxygen isotopes, influenced both by the isotopic composition of the water and by the temperature at which the calcite is precipitated, are a well-known example of a proxy that is subject to multiple influences. For this reason, paleoceanographic data are almost always presented in their original units (e.g., $\delta^{18}\text{O}$ -calcite), accompanied by an interpretation in terms of the dominant influence(s) (temperature or $\delta^{18}\text{O}$ -water in the case of $\delta^{18}\text{O}$ -calcite). Constraining the other influences using additional proxies or by measuring more than one proxy for the same

environmental variable remains a preferred strategy in interpreting paleoceanographic records. The recent application of Mg-/Ca-based paleothermometry measured on the same shells where oxygen isotopes have been measured is a notable success in constraining the influence of temperature on the oxygen isotopic composition and enables the reconstruction of the two most important variables in physical oceanography: temperature and salinity (see [Oxygen Isotope Composition of Seawater, Salinity Proxies \$\delta^{18}\text{O}\$](#)). The use of radioisotopes to constrain accumulation rates shows promise in the reconstruction of sediment fluxes (see [Radioisotope Proxies](#)). Paleoceanographic proxies and their interpretation are discussed in articles listed in [Table 1](#).

Evolution During the Quaternary

Proxy records, such as benthic and planktonic foraminiferal oxygen isotopes and ice-rafted debris from high latitudes, have expanded our understanding of the evolution of climate during the Quaternary. A distinctive feature is the onset of the Quaternary Ice Ages as well as the progressive cooling of the high latitude sea surface and the deep ocean ([Lisiecki and Raymo, 2005](#)). In contrast, records from low latitudes indicate that the sea surface temperature (SST) of the tropical warm pool regions remained stable during the last 5 Ma. The result is a dramatic increase in the zonal (west–east) and meridional (north–south) gradients in SST, accompanied by a progressive cooling of the water upwelled along the eastern margins of the Pacific (Peru, California) and South Atlantic (Africa).

Zonal and Meridional Gradients in SST

A notable feature of SST reconstructions is the emergence of the modern equatorial Pacific cold tongue and changes in upwelling during the mid-Pleistocene. A comparison of SST between the western and eastern equatorial Pacific shows that there is a

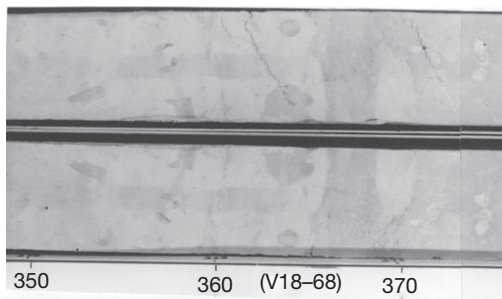


Figure 1 Piston core V18-68 from the Southeast Pacific (55°S , 78°E), split into two halves. This core is comprised of foraminifer ooze (a mixture of clay and plankton remains that include the calcium carbonate skeletons of foraminifers and coccoliths). The mottling and color variations in this section (350–370 cm) are the result of mixing of sediment by organisms that occurred at the seafloor prior to the accumulation of additional sediment (from the Index to Geological Samples, <http://www.ngdc.noaa.gov/mgg/>).

Table 1 Paleoceanographic proxies for climate and environmental change

Proxy	Climate/environmental significance	Article
Mg/Ca, Sr/Ca	Sea surface temperature	Mg/Ca and Sr/Ca Paleothermometry from Calcareous Marine Fossils
Alkenone saturation (U^k_{37})	Sea surface temperature	Alkenone Paleothermometry Based on the Haptophyte Algae; Biomarker Indicators of Past Climate
Fossil plankton census counts	Sea surface temperature	Temperature Proxies, Census Counts
$\delta^{13}\text{C}$, Cd/Ca, Ba, Zn/Ca, $\delta^{15}\text{N}$	Nutrients, ocean circulation, and ocean ventilation	Nutrient Proxies; Carbon Cycle Proxies ($\delta^{11}\text{B}$, $\delta^{13}\text{C}_{\text{calcite}}$, $\delta^{13}\text{C}_{\text{organic}}$, Shell Weights, B/Ca, U/Ca, Zn/Ca, Ba/Ca)
$\delta^{18}\text{O}$	Ice volume, ocean temperature	Oxygen-Isotope Stratigraphy of the Oceans; Oxygen Isotope Composition of Seawater; Salinity Proxies $\delta^{18}\text{O}$
CaCO_3	Plankton production, dissolution, carbon cycle	Dissolution of Deep-Sea Carbonates
Tex_{86}	Temperature	Biomarker Indicators of Past Climate
HBI (highly branched isoprenoid alkenes)	Temperature	Biomarker Indicators of Past Climate
Lipid δD	Salinity	Biomarker Indicators of Past Climate
$\delta^{11}\text{B}$, $\delta^{13}\text{C}$ -calcite, $\delta^{13}\text{C}$ -organic, shell weight, U/Ca, Li/Ca, Cd/Ca, Zn/Ca, Ba/Ca	Carbon cycle	Carbon Cycle Proxies ($\delta^{11}\text{B}$, $\delta^{13}\text{C}_{\text{calcite}}$, $\delta^{13}\text{C}_{\text{organic}}$, Shell Weights, B/Ca, U/Ca, Zn/Ca, Ba/Ca)
^{230}Th , ^{231}Pa	Sediment accumulation, ocean circulation	Radioisotope Proxies

notable increase (~ 3.5 to $\sim 5^\circ\text{C}$) in the equatorial Pacific zonal SST gradient during the interval between 1.2 and 0.8 million years (My) ago (Figure 3). Liu and Herbert (2004) reported decreasing alkenone-derived SSTs from a record from the eastern equatorial Pacific for the last 1.8 My. In contrast, SSTs in the western Pacific warm pool (WPWP) remained relatively stable throughout the Pleistocene epoch (Degardel-Thoron et al., 2005). Oxygen isotopes, alkenones, and Mg/Ca measurements from the western and eastern equatorial Pacific show that there appear to be permanent El-Niño-like conditions during the Pliocene warm period (Dekens et al., 2004; Wara et al., 2005). During the warm early Pliocene (4.5–3.0 Ma), the eastern Pacific thermocline was deep and the west-to-east

SST difference across the equatorial Pacific was relatively small ($1.5 \pm 0.9^\circ\text{C}$), much like a modern El Niño event. After 1.7 Ma, however, SST in the eastern equatorial Pacific began decreasing, while the SST in the western equatorial Pacific remained warm. Hence, the zonal SST difference increased up to $4\text{--}5^\circ\text{C}$ at 1.7 Ma. Meanwhile, SSTs cooled in middle-latitude eastern margin boundary upwelling regions (California, Peru, Africa), increasing the north–south temperature gradient (Dekens et al., 2004; Federov et al., 2006; Marlow et al., 2000). The data reveal that the meridional temperature gradient increased from $\sim 2^\circ\text{C}$ to $\sim 8^\circ\text{C}$, indicating a vastly poleward expansion of the tropical ocean warm pool at that time. Federov et al. and Dekens et al. hypothesize that both the zonal and meridional gradients were accompanied by steepening of the vertical temperature gradient in the ocean, a hypothesis supported by the analysis of surface and deep-dwelling Foraminifera (Wara et al., 2005). Both of these changes likely contributed to the increased sensitivity to orbital forcing that marked the mid-Pleistocene transition (Figure 4).

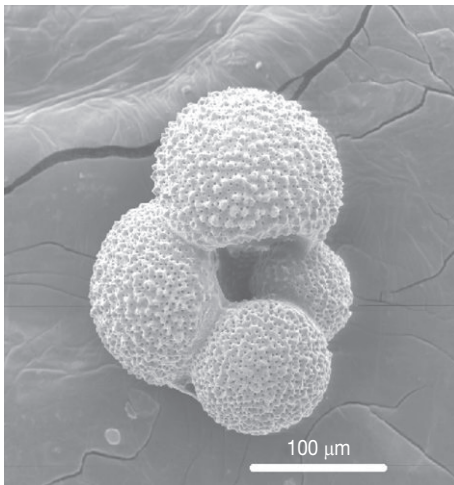


Figure 2 Calcium carbonate shell of *Globigerina bulloides* from a sediment trap sample collected in the Santa Barbara Basin in the eastern Pacific (photograph by Robert Fisher, University of Colorado).

Mid-Pleistocene Transition

There have been several hypotheses regarding the mid-Pleistocene climate change. Medina-Elizalde and Lea (2005) investigated the SST record from a site in the western equatorial Pacific warm pool based on planktonic foraminiferal magnesium/calcium ratios. They focused on the timing and nature of tropical Pacific SST changes over the mid-Pleistocene transition, and showed that SST changes in the western equatorial Pacific warm pool were synchronous with eastern Pacific cold-tongue SST, but preceded changes in continental ice volume. Based on the records, they considered changes in atmospheric carbon dioxide as the cause of the switch in climate periodicities at this time. Herbert et al. (2010) also emphasized the

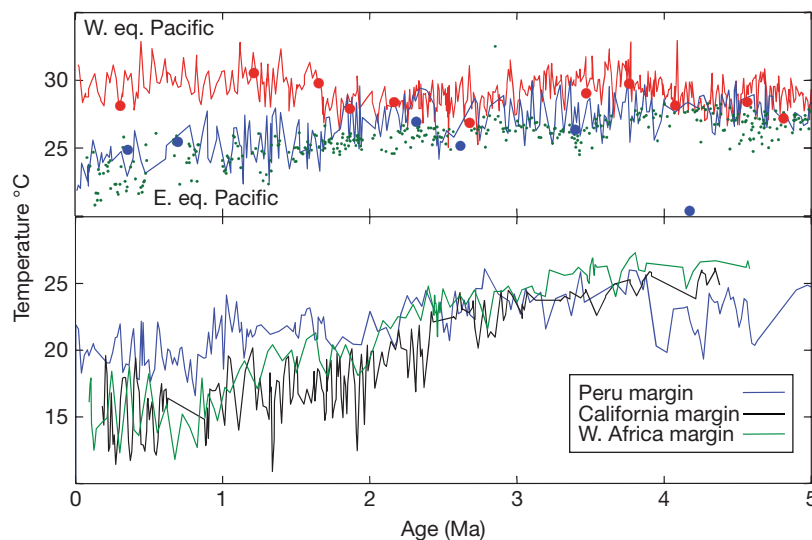


Figure 3 Top: Evolution of the Pacific West–East sea surface temperature gradient during the past 4 Ma based on Mg/Ca (red, from ODP 806 in the western equatorial Pacific) and Mg/Ca (blue) and alkenones (green dots) in the eastern equatorial Pacific (ODP 847). Larger circles are additional Mg/Ca data for ODP 806 (red) and 847 (blue). Bottom: Progressive cooling of coastal upwelling regions revealed by alkenone records from Peru (blue, ODP 1237), California Margin (black, ODP 1014), and the West African Margin (green, ODP 1084). Modified from Federov AV, Dekens PS, Mccarthy MD, et al. (2006) The Pliocene paradox (mechanisms for a permanent El Niño). *Science* 312: 1485–1489.

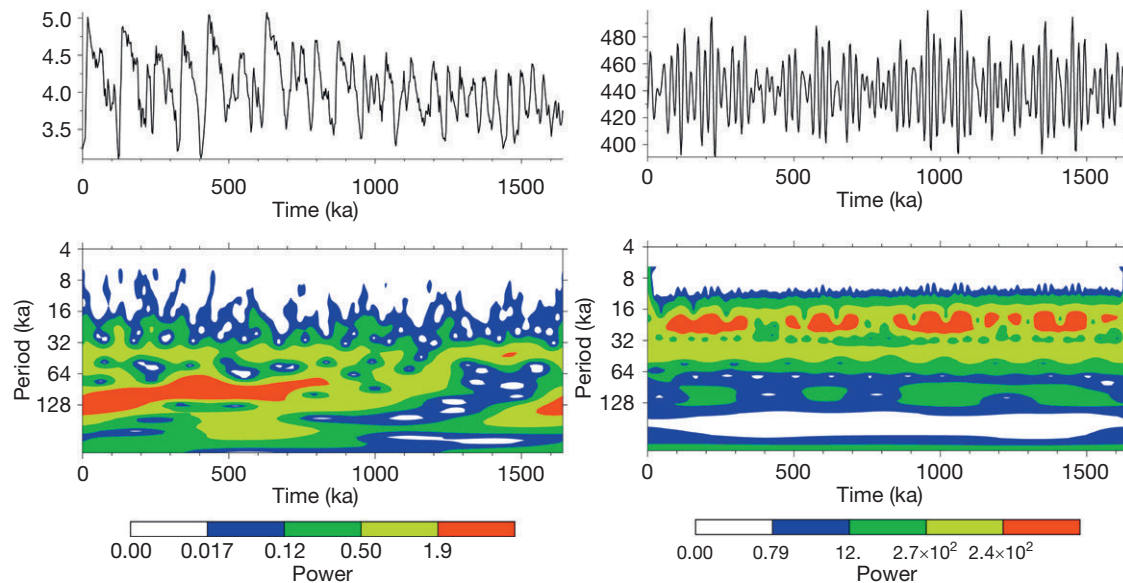


Figure 4 Time series and wavelet analysis of June solar radiation at 65° N (Berger and Loutre, 1991) (W m^{-2}) (right) and oxygen isotope variations (Lisiecki and Raymo, 2005) (per mil) for the past 1.6 million years (left). Increasing ice volume and/or cooling is upward in this figure. The wavelet spectrum (bottom panels) reveals the variance at the orbital frequencies of eccentricity (100 000 years), obliquity (41 000 years), and precession (23 000 years) and shows the increase in 100 000-year variance between 1 and 0.65 Ma that characterized the mid-Pleistocene transition.

importance of determining the timing and amplitude of tropical SST in solving the puzzle of the Plio–Pleistocene. According to them, alkenone-based tropical SST records from the major ocean basins show coherent Glacial–Interglacial temperature changes that align with global changes in ice volume and deep ocean temperature over the past 3.5 My. They pointed out that tropical temperatures became tightly coupled with benthic $\delta^{18}\text{O}$ and orbital forcing after 2.7 Ma, suggesting that the inception of a strong carbon-dioxide-greenhouse-gas feedback and amplification of orbital forcing at 2.7 Ma corresponded to the onset of Northern Hemisphere glaciation.

Others, however, challenge the hypothesis of a gradual decrease in atmospheric CO_2 concentration as a dominant trigger of the larger 100-ka glacial cycles since 850 000 years BP. According to Degardel-Thoron et al. (2005), the mid-Pleistocene cooling was restricted to upwelling regions, whereas surface temperatures remained at the same mean level before and after 0.8 Ma in the WPWP. They suggest an alternative hypothesis that nonlinear, coupled ocean-atmosphere dynamics regulate the equatorial Pacific climate. The increased temperature contrast between the eastern equatorial Pacific upwelling regions and the thermally stable WPWP in the mid-Pleistocene probably led to the intensification of the Walker circulation. This strengthening of the zonal Pacific atmospheric circulation is further supported by the surface freshening of the WPWP. These changes might have led to atmospheric processes and oceanic interactions which could have altered the meridional heat and moisture transfers to the Northern Hemisphere ice sheets during the mid-Pleistocene transition and thus contributed to Northern Hemisphere glaciations.

The meridional expansion of the tropical warm pool during the warm Pliocene period evidently had enormous impacts on the Pliocene climate, including a slowdown of the atmospheric Hadley circulation and El-Niño-like conditions in the equatorial region. On the other hand, the atmospheric response to the

reduced zonal SST gradient indicates relatively weak zonal Walker circulation. The mechanisms of ocean heat uptake, such as enhanced ocean vertical mixing, were also considered to play a role in sustaining a climate state with a weak tropical SST gradient. A positive feedback between hurricanes and the upper-ocean circulation in the tropical Pacific Ocean may have been essential for maintaining warm, El-Niño-like conditions during the early Pliocene (Fedorov et al., 2010). More frequent and/or stronger hurricanes in the central Pacific imply greater heating of water parcels, warmer temperatures in the eastern equatorial Pacific, warmer tropics, and in turn, even more hurricanes. The increase in meridional and zonal SST gradients in the Pacific during the mid-Pleistocene implies a change in the mean climate state from El-Niño-like state to more La-Niña-like condition, which implies an intensification of the Walker circulation during the mid-Pleistocene.

Glacial–Interglacial Cycles

Paleoceanographic records provide a valuable complement to terrestrial evidence of the Quaternary Glacial–Interglacial cycles. Continuous in time and widespread in geographic distribution, paleoceanographic records provide a way of understanding how the climate system changes during glacial cycles. The following discussion emphasizes the changes that occurred during the Last Glacial Maximum (LGM), 21 000 calendar years BP, recognizing that previous glacial cycles of the last 850 000 years share many of the same characteristics. Two aspects of the glacial world are particularly significant: (1) the large sensitivity of the North Atlantic region compared to other regions (glacial changes were larger here than anywhere else), and (2) the similarity between temperature and atmospheric carbon dioxide change (and changes in carbon in the ocean),

clearly implicating the role of carbon dioxide as an amplifying influence on the glacial–interglacial cycles.

Climate Sensitivity of the North Atlantic Compared to Other Regions

Since the pioneering CLIMAP (Climate: Long-Range Investigation Mapping and Prediction) studies (CLIMAP Project Members, 1981), many efforts have been made to map SST changes that occurred between the present and the LGM, 21 000 years BP. While the earliest widespread compilations were based on faunal assemblages, later studies have used oxygen isotopes, alkenones, Mg/Ca, and other proxies (Waelbroeck et al., 2009). All approaches are consistent in showing that large cooling occurred in the North Atlantic compared to other regions. Comparing alkenone-based estimates (Rosell-Mele et al., 2004) with the original CLIMAP estimates (the latter expressed as annual mean zonal averages; Shin et al., 2003) for the LGM, the North Atlantic (40–60° N) was 7–8 °C colder than present, compared to a cooling of up to 5 °C in the North Pacific and 4 °C in the southeastern Atlantic. Higher latitude regions (>60°) that are close to freezing today cooled less, forming sea ice instead. The tropical oceans were a few degrees colder (Table 2). The least amount of cooling occurred in the subtropical gyres. Based on these reconstructions, the North Atlantic appears to be about 2–3 times more sensitive than the low latitude oceans due to a combination of factors that include the growth of ice on the surrounding continents, changes in sea ice extent, and changes in ocean circulation. The original CLIMAP studies showed some regions within several of the subtropical gyres to be warmer than present at the LGM. Some subsequent studies have supported this result, while others have indicated that the subtropical gyres were no different or slightly cooler than today (Waelbroeck et al., 2009). All the studies indicate that the gyres are remarkably insensitive to change compared to other regions.

Changes in the meridional overturning circulation could contribute to the large climate sensitivity of the North Atlantic, provide a link between ice sheets and ocean circulation, and influence global climate as well. The decrease in deep water formation results in a much colder surface waters in the North Atlantic, with widespread effects across the Northern Hemisphere. The meridional overturning circulation has been interpreted from different tracers. The most extensive reconstruction uses $\delta^{13}\text{C}$ and Cd/Ca to produce a glacial section that can

be compared to the modern GEOSECS $\delta^{13}\text{C}$ and phosphate measurements (Curry and Oppo, 2005; Lynch-Stieglitz et al., 2007) (Figure 5). The glacial section reveals three different end-member water masses: (1) a shallow (1000 m) southern-source water mass (2), a mid-depth (1500 m) northern-source water mass (3), and a deep (>2000 m) southern-source water mass. Similar to previous compilations and studies based on other proxies, the section shows reduced southward penetration of the northern-source water mass corresponding to present-day NADW and the increased northward penetration of the deepest, southern-source water mass. The $\delta^{13}\text{C}$ of the intermediate depth is enriched relative to modern, consistent with a more shallow overturning circulation, forming a water mass that has been called Glacial North Atlantic Intermediate Water. Proxies for the rate of circulation (^{14}C , $^{231}\text{Pa}/^{230}\text{Th}$) indicate the glacial deep circulation was not much slower than today. A key unresolved aspect of the overturning circulation, relevant to the glacial state, millennial-scale variability, and future change, is whether different stable modes exist (Hofmann and Rahmstorf, 2009).

Role of the Ocean in Carbon Dioxide and Climate Change

One of the most significant findings with regard to future climate change is the observed correlation between changing temperature and carbon dioxide concentrations in the atmosphere. By comparing the magnitude of glacial to interglacial temperature change with the magnitude of carbon dioxide change, it is possible to independently confirm the climate sensitivity: the amount of change that might be expected in the future due to emissions of carbon dioxide from the burning of fossil fuels. While ice cores provide this correlation, paleoceanographic proxies can also be used to make this comparison, employing proxies for temperature and carbon in the surface ocean.

In addition to helping calibrate the sensitivity to carbon dioxide, paleoceanographic records are also useful in understanding the ocean's role in the carbon cycle. The ocean contains sixty times more carbon than the atmosphere due to its large volume and the ability of carbon dioxide to chemically react in seawater to form carbonic acid (H_2CO_3), carbonate (CO_3^{2-}), and bicarbonate (HCO_3^-) ions (see Dissolution of Deep-Sea Carbonates). Most of this carbon resides in the deep sea, kept out of the surface ocean by the continuous formation and sinking of fixed carbon by biological activity (termed the biological pump), and the slow return to the surface by ocean circulation. One of the first-developed proxies for the biological pump was the difference in the carbon isotope ratio $\delta^{13}\text{C}$ recorded in planktic and benthic foraminifers ($\Delta\delta^{13}\text{C}$) (Shackleton et al., 1983). The surface ocean is high in $\delta^{13}\text{C}$ because ^{12}C is preferentially removed as fixed carbon, and the deep ocean is low in $\delta^{13}\text{C}$ because the fixed carbon with low $\delta^{13}\text{C}$ is later added to the deep sea by respiration and regeneration (see Carbon Cycle Proxies ($\delta^{11}\text{B}$, $\delta^{13}\text{C}_{\text{calcite}}$, $\delta^{13}\text{C}_{\text{organic}}$, Shell Weights, B/Ca, U/Ca, Zn/Ca, Ba/Ca)). Comparison of SST derived from Mg/Ca in the western tropical Pacific (Degaridel-Thoron et al., 2005) with the record of $\Delta\delta^{13}\text{C}$ from a Pacific core reveals that the carbon isotope gradient is larger during glacial times when the SST is cooler, and the gradient is smaller during interglacial times when the temperature is warmer (Figure 6). More light carbon was in the deep sea, and less in the surface

Table 2 Change in sea surface temperature (°C) between the Last Glacial Maximum and the present based on the alkenone proxy and faunal changes (CLIMAP)

Region	Difference from modern (alkenones)	Difference from modern (CLIMAP annual zonal mean)
Northeast Atlantic (30–45° N)	Up to –7	–8
Northeast Pacific (40–60° N)	Up to –5	–2
Tropical Pacific	–2.0	–1
Tropical Atlantic	–2.7	–2
Tropical Indian	–1.7	–2
Southeastern Atlantic	Up to –4	–2 to –3

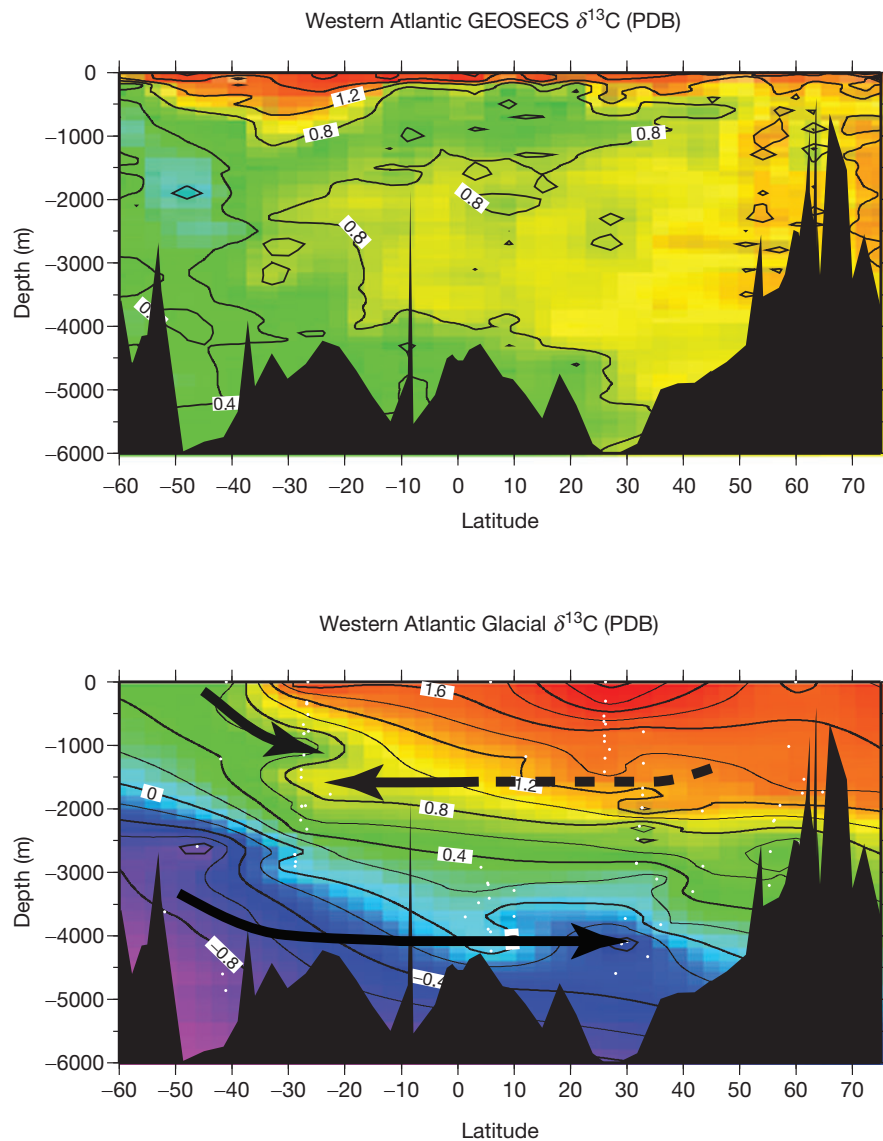


Figure 5 North Atlantic Deep Water production during the Last Glacial Maximum (LGM). Reproduced/modified from Curry WB and Oppo DW (2005) Glacial water mass geometry and the distribution of $\delta^{13}\text{C}$ of total CO_2 in the western Atlantic Ocean. *Paleoceanography* 20: PA1017, with permission of American Geophysical Union.

ocean, during glacial times. These changes in the carbon isotope gradient can be scaled to provide a record of atmospheric carbon dioxide. More recently developed evidence from boron isotopes confirms the reduction in dissolved carbon dioxide in the surface ocean during glacial times (Hoenisch et al., 2009) and also reveals the stability of carbon dioxide since at least 2.1 Ma.

Millennial-Scale Variability During Glacial Times

In addition to the Quaternary glacial cycles found in all marine sediment cores, some paleoceanographic records contain finer-scale variations that are largest during glacial times, such as the interval from 60 000 to 20 000 years BP (see [Sub-Milankovitch \(DO/Heinrich\) Events](#)). Two different types of millennial-scale variability are observed during the last

glacial interval. Dansgaard-Oeschger (D-O) events describe a series of warm (interstadial) and cool (stadial) events that are recorded in the oxygen isotopes of Greenland ice cores. Twenty-five events are numbered during the last glacial interval, each lasting from 600 to 2000 years and each characterized by progressive cooling followed by abrupt warming. Heinrich (H) events are other millennial-scale events that occurred during the last glacial. Heinrich events were initially defined on the basis of coarse sediments found in discrete layers in North Atlantic sediments and represent times when coarse ice-rafted material was transported farther south. Six events have been widely identified in North Atlantic sediments (H1, 14 300; H2, 21 000; H3, 27 000; H4, 35 500; H5, 52 000; and H6, 69 000 years BP), each occurring at the coldest part of a D-O cycle. The Younger Dryas shares some similarities with both D-O stadial events and H events, and can be considered the last H event

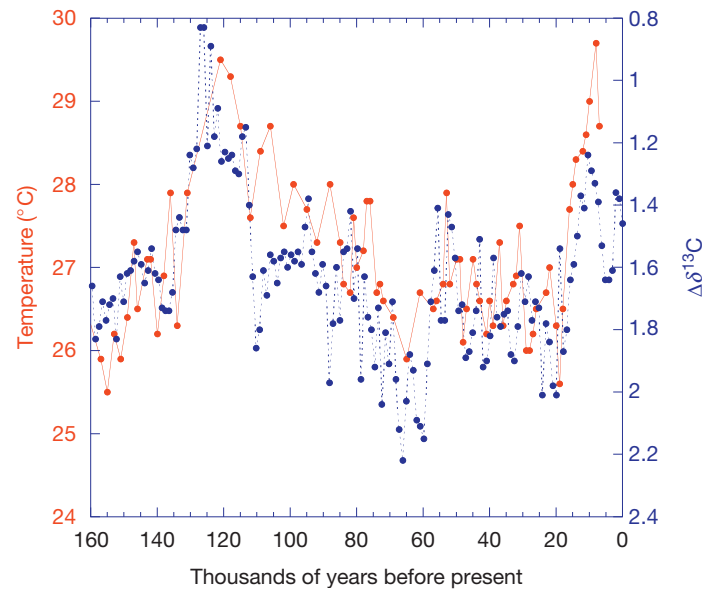


Figure 6 Paleoceanographic records reveal the covariation between carbon gradients (Shackleton et al., 1983) and sea surface temperature (Degaridel-Thoron et al., 2005) in the western Pacific warm pool. The vertical gradient in $\delta^{13}\text{C}$ was one of the first of many paleoceanographic proxies that record the partitioning of carbon between the surface and deep ocean, revealing the ocean's role in determining the Ice Age variations in atmospheric carbon dioxide.

(H0, 11 000 years BP) on the basis of cold, fresh conditions found in the North Atlantic. Subsequent studies have also shown that the H events are characterized by colder SSTs, in addition to extensive ice-rafting, and by a freshwater anomaly further north, as indicated by $\delta^{18}\text{O}$ records. One of the early records of millennial-scale cooling of the North Atlantic, sea surface is found in increasing percentages of *Neogloboquadrina pachyderma* left-coiling (Figure 7(b); Bond et al., 1992).

While the amplitude of SST changes and the episodic ice-rafting events are largest and best documented in the North Atlantic, paleoceanographic records show evidence of millennial-scale variability far from the North Atlantic (Voelker et al., 2002). The extent of millennial-scale variability outside the North Atlantic and the climate processes involved are the subject of intense research and remain limited by less certain dating beyond the range of radiocarbon dating (40 000 years BP) and also by bioturbation (see Introduction) that attenuates millennial-scale variability. Two regions where appropriate sediments exist are the Santa Barbara Basin off California (Site 893) and the Oman Margin in the Arabian Sea (RC2714). Both regions are characterized by high sedimentation rates ($>25 \text{ cm ka}^{-1}$) and low oxygen concentrations at the seafloor. Both sites record millennial-scale events during the past 50 000 years, indicative of increased ventilation and/or reduced productivity during cool periods in Santa Barbara Basin (Figure 7(c); Cannariato and Kennett, 1999), and indicative of decreasing denitrification and/or reduced productivity along the Oman Margin during cool periods (Figure 7(d); Altabet et al., 2002). Several different mechanisms have been proposed to explain the link between the North Atlantic and these remote sites, some involving atmospheric circulation (both sites are characterized by strong wind-driven upwelling today), and another group of

explanations tied to global changes in the ventilation of the upper intermediate ocean.

While the preceding discussion examines the millennial-scale variability during the last glacial interval, from 120 000 to 10 000 years BP, longer records show that millennial-scale variability was a persistent feature of at least the preceding four glacial cycles (the last 500 000 years), prevalent whenever benthic $\delta^{18}\text{O}$ exceeded 3.5 per mil (McManus et al., 1999). These oscillations and events are absent or much reduced during the postglacial interval. The presence of this variability only during glacial times, the abrupt, sawtooth character of the cycles, and the long time between events (thousands of years) have led to many explanations that involve the interaction between ice sheet growth and decay, the surface and deep circulation in the North Atlantic, and the bipolar seesaw (Barker et al., 2011).

The Last Deglaciation

Regional differences in the timing and the pattern of deglaciation provide important evidence regarding the mechanisms driving the Earth's glacial cycles. Unfortunately, age uncertainty in marine sediments still makes it difficult to evaluate the events and rapid changes that occurred during the deglaciation that differ by only several hundred years. While the measurement uncertainty of radiocarbon (one standard deviation) during this time interval is on the order of 50 years, the practical uncertainty due to changes in reservoir age, bioturbation, and other effects is on the order of several hundred years. This uncertainty makes it extremely difficult to determine the alignment of Younger Dryas cooling observed in ice cores with sea level determined from

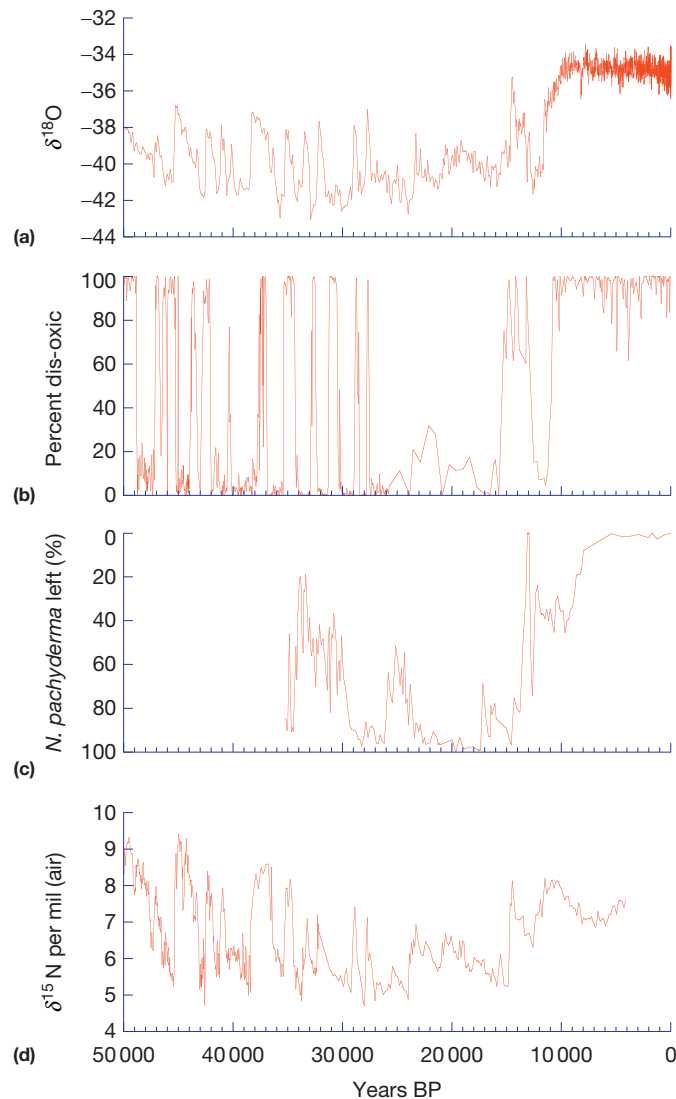


Figure 7 Paleoceanographic records of millennial-scale variability compared to the GISP 2 oxygen isotope record (a) (Grootes et al., 1993) includes *N. pachyderma* (left-coiling) abundance in the North Atlantic (greater abundance indicates cooler sea surface temperature) (c) (Bond et al., 1992), the percent of disoxic foraminifers in Santa Barbara Basin Site 893 (increasing percentages indicate greater ventilation or reduced productivity) (b) (Cannariato and Kennett, 1999), and the $\delta^{15}\text{N}$ ratio in core RC2714 from the Arabian Sea (higher $\delta^{15}\text{N}$ indicates higher nutrient utilization, a measure of increased productivity) (d) (Altabet et al., 2002).

corals, SSTs, and meltwater events observed in oxygen isotope records. Despite this uncertainty, two different broadscale patterns of the last deglaciation can be observed, upon which, regional differences are superimposed. One pattern is typical of the North Atlantic region and is characterized by a rapid change from full glacial conditions beginning 15 ka (Figure 8; Lea et al., 2003). The abrupt change toward interglacial conditions was interrupted by a brief return to almost full glacial conditions during the Younger Dryas interval (13 000–11 700 years BP). The Younger Dryas interval ended with an abrupt transition toward the present interglacial. This pattern typifies most North Atlantic records and is also found in regions far from the North Atlantic, such as the California Margin and the Asian Monsoon region (Figure 7). Records that show large

millennial-scale variability also tend to show the abrupt pattern during the deglaciation.

In contrast, sediments from western tropical Pacific and high southern latitudes are characterized by a different pattern that begins earlier (17 000 years BP) and warms more gradually (Figure 8; Stott et al., 2007; Visser et al., 2003). This pattern also contains a brief return toward colder conditions, but the reversal is not as large and the timing is different from the Younger Dryas, beginning around 15 000 years BP. The sequence of events during the last termination remains an active research topic. Hypotheses regarding mechanisms can be divided into those that emphasize changes in the tropics, high southern latitudes (Anderson et al., 2009) and possibly southern insolation, and those that emphasize initiation by increased northern

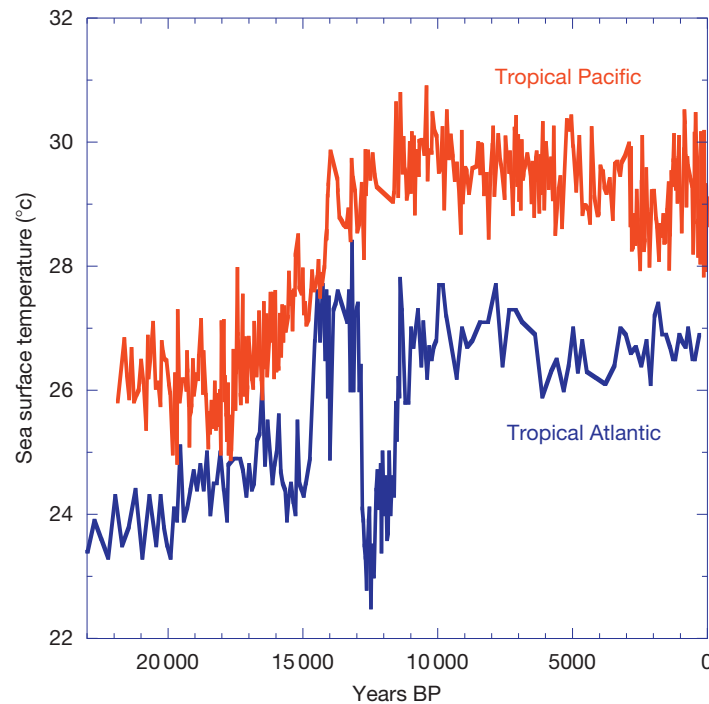


Figure 8 Temperature change during the last deglaciation revealed by alkenone-based reconstructions of sea surface temperature in the tropical Pacific (red) and tropical Atlantic (blue) (Stott et al., 2007; Visser et al., 2003).

seasonality (65° N) (Clark et al., 2009), with a possible role of the Asian monsoon and the intertropical convergence zone (ITCZ) (Cheng et al., 2010). More information on the last deglaciation, including its regional characteristics, is provided in the individual articles.

Postglacial

The largest external influence on Holocene climate was the 8% decrease in summer radiation (and increase in winter radiation) that occurred between 11,000 years BP and the present due to the precession of the Earth's axis of rotation. Summers were warmer and winters colder over the Northern Hemisphere 11,000 years ago, a contrast that is much reduced today. These changes are expected to have caused stronger monsoon circulation 11,000 years BP, and the strengthening of the monsoon in the early Holocene observed in both data and model simulations is one of the successes of Quaternary paleoclimatology (see [Postglacial Indian Ocean](#)).

New proxies have refined our understanding of other climate changes during the postglacial. In the tropics, the ITCZ appears to have been located northward of its present mean position during the early Holocene and appears to have moved south during the Holocene. In the African and Asian sectors, this would have contributed to weakening the monsoon circulation. In the Atlantic, the southward migration of the ITCZ has been documented by decreasing Ti and Fe concentrations in Cariaco Basin sediments, an indicator of decreasing rainfall over the adjacent continent that is best explained by southward movement of the

ITCZ during the Holocene (Haug et al., 2001) (see [Postglacial North Atlantic](#)). In the Pacific, the east–west SST gradient, indicative of the strength of the Walker circulation, was largest during the mid-Holocene, reduced today, and much reduced during the last glacial interval (Koutavis et al., 2002; see [Postglacial North Pacific](#), [Postglacial South Pacific](#)). These changes support the hypothesis of an El-Niño-like mean state in the Pacific during the last glacial interval, and more La-Niña-like mean conditions during the middle Holocene. Progressive freshening of the western Pacific sea surface from the beginning to the end of the Holocene is supported by the oxygen isotope record in the western Pacific, after removal of the temperature effect using Mg/Ca (Stott et al., 2004).

An intriguing aspect of paleoceanographic records of the postglacial interval is the evidence for small-amplitude century-scale cycles that correlate with records of solar variability derived from ^{10}Be and ^{14}C variations. First observed in records of ice-rafting in the North Atlantic (Bond et al., 2001), millennial-scale variability has also been observed in paleoceanographic records of the Asian summer monsoon (Fleitmann et al., 2003; Gupta et al., 2003). Unlike the Milankovitch changes in summer insolation that are large (8%), these changes in (total annual) radiation are small (less than 1%), requiring amplifying mechanisms not yet understood. The changes observed in paleoceanographic records are so small that the changes are close to the uncertainty in most proxies and perhaps only detectable in regions where the changes are unusually large. Some consider the Medieval Warm Period, the Little Ice Age, and part of the warming since the Little Ice Age (during the last century) to be the most recent

manifestations of the solar cycles. The societal relevance of this aspect of natural climate variability warrants further research, where paleoceanographic records have great potential.

See also: **Glaciation, Causes:** Astronomical Theory of Paleoclimates. **Ice Core Methods:** Overview. **Paleoceanography, Biological Proxies:** Benthic Foraminifera; Biomarker Indicators of Past Climate; Planktic Foraminifera; Temperature Proxies, Census Counts. **Paleoceanography, Physical and Chemical Proxies:** Carbon Cycle Proxies ($\delta^{11}\text{B}$, $\delta^{13}\text{C}_{\text{calcite}}$, $\delta^{13}\text{C}_{\text{organic}}$, Shell Weights, B/Ca, U/Ca, Zn/Ca, Ba/Ca); Dissolution of Deep-Sea Carbonates; Oxygen-Isotope Stratigraphy of the Oceans; Radioisotope Proxies; Salinity Proxies $\delta^{18}\text{O}$. **Paleoceanography, Records:** Late Pleistocene North Atlantic; Postglacial Indian Ocean; Postglacial North Atlantic; Postglacial North Pacific.

References

- Altabet MA, Higginson MJ, and Murray DW (2002) The effect of millennial-scale changes in Arabian Sea denitrification on atmospheric CO_2 . *Nature* 415: 159–162.
- Anderson RF, Ali S, Bradtmiller LI, et al. (2009) Wind-driven upwelling in the southern ocean and the deglacial rise in atmospheric CO_2 . *Science* 323: 1443–1448.
- Barker S, Knorr G, Edwards RL, et al. (2011) 800,000 years of abrupt climate variability. *Science* 334: 347.
- Berger A and Loutre MF (1991) Insolation values for the climate of the last 10 million years. *Quaternary Science Reviews* 10: 297–317.
- Bond GC, Heinrich H, Broecker W, et al. (1992) Evidence for massive discharges of icebergs into the North Atlantic Ocean during the last glacial. *Nature* 360: 245–249.
- Bond G, Kromer B, Beer J, et al. (2001) Persistent solar influence on North Atlantic climate during the Holocene. *Science* 294: 2130–2136.
- Cannariato K and Kennett PJ (1999) Climatically related millennial-scale fluctuations in the strength of California margin oxygen-minimum zone during the past 60 ky. *Geology* 27: 975–987.
- Cheng H, Edwards RL, Broecker WS, et al. (2010) Ice age terminations. *Science* 326: 248.
- Clark PU, Dyke AS, Shakun JD, et al. (2009) The last glacial maximum. *Science* 325: 710–714.
- Climap Project Members (1981) Seasonal reconstruction of the Earth's surface at the last glacial maximum. *Geological Society of America, Map and Chart Series* 36: 1–18.
- Curry WB and Oppo DW (2005) Glacial water mass geometry and the distribution of $\delta^{13}\text{C}$ of total CO_2 in the western Atlantic Ocean. *Paleoceanography* 20: PA1017.
- Degardel-Thoron T, Rosenthal Y, and Bassinot F (2005) Stable sea surface temperatures in the western Pacific warm pool over the past 1.75 million years. *Nature* 433: 294–298.
- Dekens PS, Ravelo AC, and Mccarthy MD (2004) Warm upwelling regions in the Pliocene warm interval. *Paleoceanography* 22: PA3211.
- Fedorov AV, Dekens PS, Mccarthy MD, et al. (2006) The Pliocene paradox (mechanisms for a permanent El Niño). *Science* 312: 1485–1489.
- Fedorov A, Brierley CM, and Emanuel K (2010) Tropical cyclones and permanent El Niño in the early Pliocene epoch. *Nature* 463: 1066–1070.
- Fleitmann D, Burns SJ, Mudelsee M, et al. (2003) Holocene forcing of the Indian monsoon recorded in a stalagmite from Southern Oman. *Science* 300: 1737–1739.
- Groote PM, Stuiver M, White JWC, Johnsen SJ, and Jouzel J (1993) Comparison of oxygen isotope records from the GISP2 and GRIP Greenland ice cores. *Nature* 366: 552–554.
- Gupta AK, Anderson DM, and Overpeck JT (2003) Abrupt changes in the Holocene Asian Southwest Monsoon and their links to the North Atlantic Ocean. *Nature* 421: 354–356.
- Haug GH, Hughen KA, Sigman DM, Peterson LC, and Rohl U (2001) Southward migration of the intertropical convergence zone through the Holocene. *Science* 293: 1304–1308.
- Herbert T, Peterson LC, Lawrence KT, and Liu Z (2010) Tropical ocean temperatures over the past 3.5 million years. *Science* 328: 1530–1534.
- Hoenisch B, Hemming NG, Archer D, Siddall M, and Mcmanus J (2009) Atmospheric carbon dioxide concentration across the mid-Pleistocene transition. *Science* 324: 1551–1554.
- Hofmann M and Rahmstorf S (2009) On the stability of the Atlantic meridional overturning circulation. *Proceedings of the National Academy of Sciences of the United States of America* 106: 20584–20589.
- Imbrie J, Hays JD, Martinson DG, et al. (1984) The orbital theory of Pleistocene climate: Support from a revised chronology of the marine 180 record. In: Berger AL (ed.) *Milankovitch and Climate*. Dordrecht: D. Reidel Publishing Company.
- Koutavis A, Lynch-Stieglitz J, Marchitto T, and Sachs JP (2002) El Niño-like pattern in Ice Age Tropical Pacific sea surface temperature. *Science* 297: 226–230.
- Lea DW, Pak DK, Peterson LC, and Hughen KA (2003) Synchronicity of tropical and high-latitude Atlantic temperatures over the last glacial termination. *Science* 301: 1361–1364.
- Lisiecki LE and Raymo ME (2005) A Pliocene–Pleistocene stack of 57 globally-distributed benthic $\delta^{18}\text{O}$ records. *Paleoceanography* 20: PA1003.
- Liu Z and Herbert T (2004) High-latitude influence on the eastern equatorial Pacific climate in the early Pleistocene epoch. *Nature* 427: 720–723.
- Lynch-Stieglitz J, Adkins JF, Curry WB, et al. (2007) Atlantic meridional overturning circulation during the last glacial maximum. *Science* 316: 66–69.
- Marlow FR, Lange CB, and Wefer G (2000) Upwelling intensification as part of the Pliocene–Pleistocene climate transition. *Science* 290: 2288–2291.
- Mcmanus JF, Oppo DW, and Cullen JL (1999) A 0.5 million year record of millennial-scale climate variability in the North Atlantic. *Science* 283: 971–975.
- Medina-Elizalde M and Lea D (2005) The mid-Pleistocene transition in the Tropical Pacific. *Science* 310: 1009–1012.
- Rosell-Mele A, Bard E, Emeis K, et al. (2004) Sea surface temperature anomalies in the oceans at the LGM estimated from the alkenone- $\text{U}^{K'}$ index: Comparison with GCMs. *Geophysical Research Letters* 31: L03208.
- Shackleton NJ, Hall MA, Line J, and Shuxi C (1983) Carbon isotope data in core V19-30 confirm reduced carbon dioxide concentration in the ice age atmosphere. *Nature* 306: 319–322.
- Shin S-I, Liu Z, Otto-Bliesner B, Brady E, Kutzbach JE, and Harrison SP (2003) A simulation of the last glacial maximum climate using the NCAR–CCSM. *Climate Dynamics* 20: 127–151.
- Stott L, Cannariato K, Thunell RC, Haug GH, Koutavis A, and Lund S (2004) Decline of surface temperature and salinity in the western tropical Pacific Ocean during the Holocene Epoch. *Nature* 431: 56–59.
- Stott L, Timmermann A, and Thunell R (2007) Southern hemisphere and deep-sea warming led deglacial atmospheric CO_2 rise and tropical warming. *Science* 318: 435–438.
- Visser K, Thunell RC, and Stott L (2003) Magnitude and timing of temperature change in the Indo-Pacific warm pool during deglaciation. *Nature* 421: 152–155.
- Voelker AHL, et al. (2002) Global distribution of centennial-scale records for marine isotope stage (MIS) 3: A database. *Quaternary Science Reviews* 21: 1185–1212.
- Waelbroeck C, Paul A, Kucera M, et al. (2009) Constraints on the magnitude and patterns of ocean cooling at the last glacial maximum. *Nature Geoscience* 2: 127–132.
- Wara MW, Ravelo AC, and Delaney ML (2005) Permanent El Niño-like conditions during the Pliocene warm period. *Science* 309: 758–761.

Relevant Websites

- <http://geology.er.usgs.gov/eesteam/prism/> – Pliocene Research, Interpretation, and Synoptic Mapping.
- <http://www.glacialoceanatlas.org/> – Glacial Ocean Atlas.
- <http://www.ncdc.noaa.gov/paleo/> – National Climatic Data Center.

PALEOCEANOGRAPHY, BIOLOGICAL PROXIES

Contents

Alkenone Paleothermometry Based on the Haptophyte Algae
Benthic Foraminifera
Biomarker Indicators of Past Climate
Coccolithophores
Corals, Sclerosponges and Mollusks
Dinoflagellates
Marine Diatoms
Planktic Foraminifera
Radiolarians and Silicoflagellates
Temperature Proxies, Census Counts

Alkenone Paleothermometry Based on the Haptophyte Algae

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Introduction

In paleoceanography, the term ‘alkenones’ refers to a series of C₃₇–C₃₉ straight-chain methyl and ethyl ketones with one to four unsaturations and all-*trans* configuration in the double bonds (Figure 1). Besides this group of compounds, there are homologues (compounds with similar general chemical formula due to the presence of the same functional group but differs in their molecular mass by a CH₂ group) of shorter or longer alkyl chains, but to date, these additional ketones are of very limited paleoceanographic interest. The alkenones have been found in recent marine sediments of widespread geographical locations from the tropics to the subpolar regions and in marine sediment as old as the Cretaceous (Brassell and Dumitrescu, 2004).

Since their first reported occurrence in the sediment from the Walvis Ridge (Boon et al., 1978) and subsequent identification (de Leeuw et al., 1980), the alkenones have been intensively studied, especially the C₃₇ di- and triunsaturated homologues. Over the years, these biomarkers have shown to be a remarkably robust organic molecular sea surface temperature (SST) proxy in the form of the alkenone unsaturation index (see review by Herbert et al., 2003) and have allowed the description of the SST variability on various timescales in the global ocean throughout the late Cenozoic, ranging from the Pliocene (e.g., Herbert et al., 2010; Lawrence et al., 2006) and the Pleistocene (e.g., Lamy et al., 2004; Martínez-García et al., 2009; Martrat et al., 2007) to shorter timescales such as centennial (Eglinton et al., 1992), decadal (Sicre et al., 2008), or even as short as interannual changes (Kennedy and

Brassell, 1992). It has also been shown to be able to register abrupt climatic changes, such as Dansgaard–Oeschger cycles (Cacho et al., 1999; Harada et al., 2006). To date, the alkenones are probably the best biomarker for SST reconstruction. They offer several advantages over the classical proxies based on carbonate microfossils. For instance, the alkenone unsaturation index is found to be unaffected by carbonate dissolution (Sikes et al., 1991), making it applicable for sediments found beneath the calcium compensation depth which are barren of carbonate tests. It also provides an opportunity for SST reconstruction in regions where the diversity of microfossil taxa is too low for a successful assemblage transfer-function-based SST reconstruction. Within the past two decades, the alkenone unsaturation index has become one of the most widely used SST proxies in the world. Interlaboratory results can easily be compared as the maximum difference between any two laboratories for the alkenone-derived SST on the same samples is less than 2.1 °C (with a probability of 95%) (Rosell-Melé et al., 2001).

Quantification and Analytical Challenges

One of the advantages of using the C₃₇ alkenones for paleoceanographic studies is that the laboratory protocols involved in obtaining and quantifying these biomarkers are relatively simple and straightforward. Work by Sikes et al. (1991) demonstrated that the alkenone unsaturation index is not affected by the condition of storage (at room temperature or frozen) for up to several years. This finding has great implications for the general applicability of this proxy, that is, it could be applied on marine sediments that were retrieved several decades ago

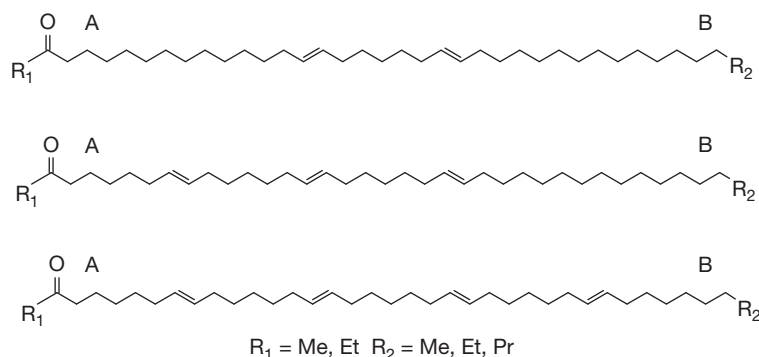


Figure 1 Chemical structures of the C_{37} alkenones, which are of main interest for paleoceanographic studies.

and might not have been kept frozen since their retrieval. Moreover, although there are cases where the plastic sample bags were found to contaminate the sediment by introducing a compound that coelutes with the $C_{37:3}$ alkenone (Sachs and Lehman, 1999), storing the sediment cores in plastic liners does not generally seem to hinder alkenone analysis.

Prior to analysis, the extractable organic matter (of which the alkenones are part) are obtained from bulk sediment samples, either by microwave-assisted extraction, by repeated sonication, by Soxhlet apparatus, or by an accelerated solvent extractor system. Additional sample clean-up procedures, such as saponification, silica gel chromatography, or urea adduction may be necessary if the sample has high organic matter content, or there are compounds that coelute with the alkenones during analysis.

Gas chromatography coupled with a flame ionization detector (GC-FID) is the most widely used method for alkenone analysis. The GC is equipped with a long nonpolar column coated with chemical film to enhance separation. Along with the implementation of a slow high-temperature program, this enables the alkenones to be separated from other compounds and subsequently quantified by the FID. The commonly used column length is between 30 and 60 m. The longer the column, the better the separation of alkenones in the chromatograms. Although FID is a sensitive detector, it does not provide information about the chemical structure of the compounds. In a GC-FID system, identification of alkenones is achieved principally by the comparison of their elution time with that of external standards. Therefore, it is strongly recommended to run duplicates of the samples on another system that could provide extra information about the chemical structure of measured compounds, such as a gas chromatography coupled with mass spectrometry (GC-MS), to confirm the alkenone identification and to ensure that the alkenone-derived SST estimates are not biased by coeluting compounds. Recently, a novel approach based on a fast GC/time-of-flight mass spectrometry was proposed by Hefter (2008) as an alternative for alkenone identification done with the conventional GC-FID system. This approach could reduce the analysis time by a factor of 10 and circumvent the bias introduced by coelution that may not be detected in a GC-FID, in addition to yielding a higher sensitivity compared to a conventional GC-MS.

Apart from the bias caused by coeluting compounds, the accuracy in the estimation of paleotemperatures might also be compromised at low alkenone abundances. Irreversible

adsorption of a portion of the alkenones, preferentially $C_{37:3}$ alkenone into the chromatographic column, poses one of the potential challenges (Grimalt et al., 2001; Villanueva and Grimalt, 1997). This problem could be amplified at the high-temperature end of the unsaturation index scale where the $C_{37:3}$ alkenone is in low abundance, resulting in spuriously warm SST estimates. One plausible way of circumventing this problem is via long-term monitoring of the GC system by running reference standards with a known alkenone unsaturation index.

Chemical Structures

The chemical structure of alkenones was first elucidated by de Leeuw et al. (1980). They found that alkenones have several unusual characteristics for biolipids, in terms of their chain length (C_{37-39}) and the spacing of the unsaturation (C7 instead of the more common C2 or C3). The C_{37} alkenones in marine sediments could be di-, tri-, and tetraunsaturated. The synthesis of these alkenones in both all-*cis* and all-*trans* configurations and coinjection with natural samples allows determination that the double bonds in alkenones have the rare all-*trans* configuration (Rechka and Maxwell, 1988) as opposed to the more common *cis* configuration, that is, (*E,E*)-15,22-heptatriacontadien-2-one ($\text{Me}_{37:2}$), (*E,E,E*)-8,15,22-heptatriacontatrien-2-one ($\text{Me}_{37:3}$), and (*E,E,E,E*)-8,15,22,29-heptatriacontatetraen-2-one ($\text{Me}_{37:4}$). These compounds are also widely known as $C_{37:2}$, $C_{37:3}$, and $C_{37:4}$ alkenones or ketones in the literature. In general, GC-MS is the technique of choice for structural characterization of the alkenones, since it has enough sensitivity, based on the fragmentation pattern of the molecules, to detect their presence in sediments, algae, or waters where they are often found in low concentrations.

The alkenone distributions so far identified could be grouped arbitrarily in two types (tentatively named type A and type B for convenience of discussion). Type A consists of C_{37} methylketones, C_{38} methyl and ethylketones, C_{39} methyl and ethylketones, and C_{40} ethylketones (Figure 2). Alkenones with this distribution are synthesized by several haptophyte algae such as coccolithophorids *Emiliania huxleyi* (Figure 3), *Gephyrocapsa oceanica*, and several strains of the species *Isochrysis* (Marlowe et al., 1990) found in marine waters and sediments and in a few nonmarine environments (Figure 4). On the other hand, the only known algal precursor of the less

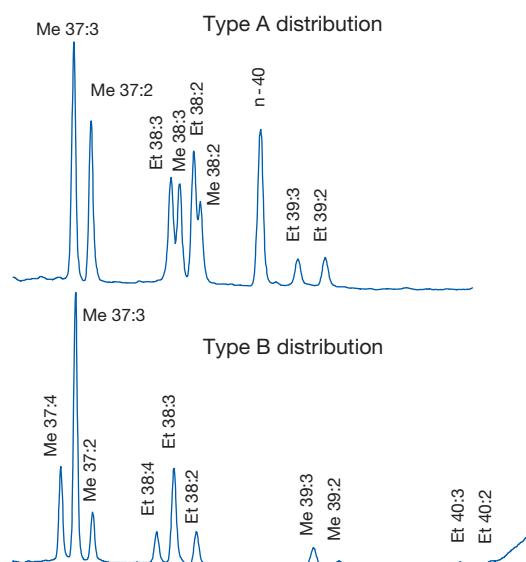


Figure 2 Gas chromatograms using GC-FID, showing examples of the two types of distributions of the C_{37} – C_{40} alkenones. Full chemical names of the compounds are given in [Table 1](#).

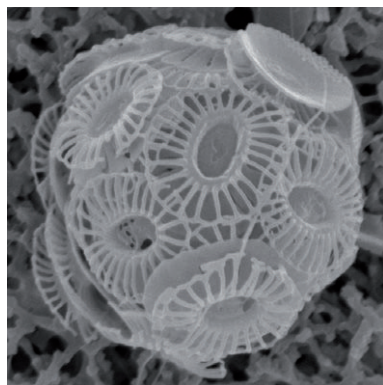


Figure 3 Coccoliths of *Emiliana huxleyi*, at present the main alkenone producer in the global ocean.

common type B distribution is *Chrysotila lamellosa* ([Rontani et al., 2004](#)), and it has been found in sediments from freshwater lakes and hypersaline ponds ([Figure 4](#)). Recently, novel pentaunsaturated C_{38} and C_{39} alkenones have been found in the surface sediments of a perennially ice-covered Lake Fryxell in the Antarctica, which are thought to be synthesized by *Isochrysis* and/or *C. lamellosa* based on the genetics study performed on the same materials ([Jaraula et al., 2010](#)).

Biological Origin

To date, alkenones are known to be synthesized by a small group of haptophyte algae, which includes some coccolithophorids from the family Gephyrocapsaceae, such as the living species *E. huxleyi*, *G. oceanica*, and the extinct genus *Reticulofenestra* ([Marlowe et al., 1990](#); [Volkman et al., 1995](#)). Some of these taxa are coccolith producers and others are naked.

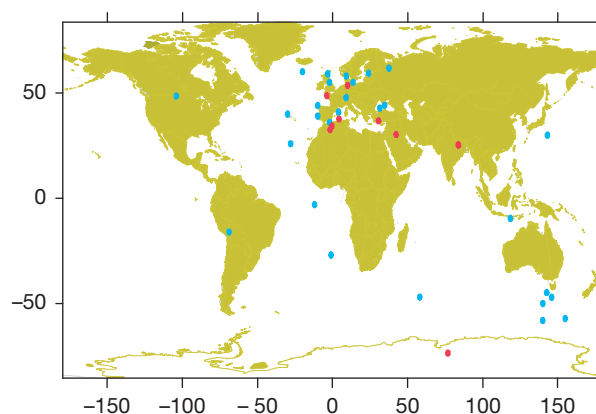


Figure 4 Examples of sites in which type A (blue) and type B (red) distributions of algae have been identified.

The low number of algal precursors is very important from the paleoclimatic or paleoceanographic standpoint, since the translation of the climatic event into a chemical signal depends on a limited number of organisms, thus providing a better constraint on the data interpretation. Conversely, in compounds synthesized by large groups of organisms, it cannot be excluded that changes in the sedimentary record could preferentially be related to variations in the precursor species rather than to climatic changes. In this respect, it has to be pointed out that alkenone analyses are based on extracts of sediments recovered at specific core sections, that is, there is no prior isolation of fossils from individual species of precursor organisms such as in Foraminifera-based geochemical proxies. Haptophyte algae are very small (on the order of $5\ \mu\text{m}$), and isolation of the carbonate tests from individual species is not feasible. In addition, there are alkenone-producing taxa that are naked, for example, *Isochrysis*. Thus, alkenone analyses of dry sediment always correspond to the composite contributions of all organisms producing these compounds present in the core section under study. Consequently, alkenone-derived SST estimates reflect a more integrated signal, with a high signal-to-noise ratio in the measurement, relative to those estimated via methods employing a small number of microfossils, such as the analysis of Mg/Ca from Foraminifera.

Despite the limited number of precursor species, these alkenones are widespread in the waters and sediments of all oceans ([Figure 4](#)), which shows that the precursor algae are widely distributed. This general distribution offers the potential to obtain paleoclimatic information from a wide diversity of marine environments that essentially cover the world's oceans, except the polar regions that are covered by sea ice ([Sikes et al., 1997](#)).

Calibration to SST

The variation in the degree of alkenone unsaturation (i.e., the number of the double bonds) obtained from the Quaternary sediments from around the world, in addition to the good correlation with planktonic foraminiferal oxygen isotope ratios during the past 550 000 years in a marine sediment core from the tropical Atlantic, led to the conclusion that the extent of unsaturation in alkenones is associated with SST ([Brassell et al.,](#)

1986). These authors thus proposed an index, known as the U'_{37} (eqn [1]), based on the relative abundance of di-, tri-, and tetraunsaturated alkenones as an approximation to quantify the degree of the unsaturation:

$$U'_{37} = (Me_{37:2} - Me_{37:4}) / (Me_{37:2} + Me_{37:3} + Me_{37:4}) \quad [1]$$

where $Me_{37:x}$ denotes the concentration of the C_{37} methyl alkenone with x carbon-carbon double bonds. To correlate the alkenone unsaturation index to temperature, laboratory cultures of *E. huxleyi* were grown under different temperatures, and the findings showed that the relationship between the two alkenone indices and growth temperature was linear in the range of 8–25 °C (Prahl and Wakeham, 1987; Prahl et al., 1988). However, the proportion of $Me_{37:4}$ alkenone is generally low in relation to $Me_{37:3}$ alkenone and $Me_{37:2}$ alkenone in marine sediments. Therefore, the original U'_{37} index is simplified to U'_{37} index (eqn [2] below) (Prahl and Wakeham, 1987):

$$U'_{37} = Me_{37:2} / (Me_{37:2} + Me_{37:3}) \quad [2]$$

Over the years, the U'_{37} index has become the more widely used index among the two, and most published calibrations (e.g., Conte et al., 2006; Müller et al., 1998) are based on this index. This is partly attributed to the fact that $C_{37:4}$ alkenones are rarely detected in the marine sediments. Consequently, in most cases, the U'_{37} values are equivalent to the U'_{37} values. The exception is probably in the polar regions, where some studies have found high relative abundance of $C_{37:4}$ alkenones in the sediments (e.g., Bendle et al., 2005). Some workers reported that in subpolar and polar settings where substantial amount of $C_{37:4}$ alkenones are found, U'_{37} yields more reasonable SST estimates that are comparable to other proxy data across the glacial-interglacial cycles (e.g., Bard, 2001; Calvo et al., 2002; Rosell-Melé, 1998). However, to date, there is still no consensus on the relationship of the relative abundance of $C_{37:4}$ alkenone with temperature. Some studies (e.g., Sikes and Sicre, 2002) indicate little or no relationship between $C_{37:4}$ alkenones and growth temperature, which suggests that in most cases, the U'_{37} ratio is probably a better proxy for paleotemperature estimate than U'_{37} . On the other hand, other studies (e.g., Harada et al., 2003; Rosell-Melé, 1998) found a negative correlation between the relative abundance of the $Me_{37:4}$ alkenone (the percentage of $Me_{37:4}$ alkenone, $\%C_{37:4}$, with respect to the sum of all C_{37} alkenones) and surface salinity in the high latitudes and suggested that $C_{37:4}\%$ could be used as a surface paleosalinity proxy. More recent work by Bendle et al. (2005) proposed that $\%C_{37:4}$ could serve as an indicator for polar water mass (low temperature and salinity) in the North Atlantic. Thus, further investigation is required to shed more light on the dependence of $C_{37:4}\%$ on temperature and/or salinity and its potential as a paleoclimatic proxy.

The laboratory culture study suggested a linear correlation between the alkenone unsaturation index and growth temperature, giving rise to the following equation: $U'_{37} = 0.034 \times T + 0.039$ ($r^2 = 0.994$, $n = 22$). Subsequently, many other calibration studies have also been performed, using other algal cultures, examining the relationship between water column particulates and *in situ* temperature measurements, or comparing core-top alkenone measurements with water column temperatures from oceanic databases. A decade

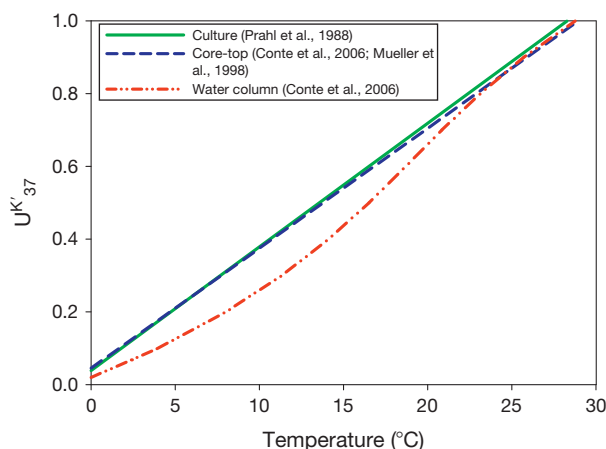


Figure 5 Comparison of the calibration equations of greatest general use for the estimation of SSTs from the unsaturation alkenone index (U'_{37}). The equation of Prahl et al. (1988) was obtained from cultures of *E. huxleyi* under controlled conditions. The core-top equation of Müller et al. (1998) and Conte et al. (2006), which were statistically identical, was obtained from measurements of core tops and comparison with sea surface temperatures from oceanic databases. The polynomial surface water calibration of Conte et al. (2006) was based on measurements of suspended particulate matter in the water column.

after the initial culture work (Prahl and Wakeham, 1987), surface sediments collected between 60°N and 60°S in the world oceans, spanning a temperature range from 1 to 29 °C and diverse biographical provinces, were analyzed for U'_{37} values (Müller et al., 1998). The resulting regression also demonstrated an exceptionally strong linear correlation with mean annual SST and a regression line ($U'_{37} = 0.033 \times T + 0.044$; $r^2 = 0.958$, $n = 370$) that is strikingly identical, within error limits, to that previously reported based on *E. huxleyi* cultures (Prahl et al., 1988; Figure 5). More recently, a more extensive global core-top compilation ($n = 592$) also found a statistically identical regression line (Conte et al., 2006). The strong agreement between the laboratory culture and core-top studies has led to most studies using one of these two equations for SST estimations. Furthermore, the results from some regional core-top calibration studies from the Atlantic Ocean (Rosell-Melé et al., 1995), the Indian Ocean (Sonzogni et al., 1997), and the Pacific Ocean (Prahl et al., 2006) further attest to the remarkable linear relationship between the alkenone unsaturation index and SST.

Even so, there is some debate on whether the relationship is linear over the whole range of the alkenone unsaturation index, or whether different equations should be used to account for deviations at the cold or warm ends ($U'_{37} \approx 0$ or 1, respectively). The linearity of the global core-top calibration over a large temperature range (~ 0 to 29 °C) seems to be confirmed by the core-top calibration studies (Conte et al., 2006; Müller et al., 1998), but contradicting conclusions could be drawn from previous regional studies that suggest a flattening of U'_{37} -SST relationship at high-temperature ends (e.g., Sonzogni et al., 1997) and those that argue against it (e.g., Pelejero and Calvo, 2003). Nevertheless, caution should be exercised in interpreting the U'_{37} -derived SST estimates at both temperature ends due to the low abundance of $C_{37:3}$ (warm end) and $C_{37:2}$ (cold end).

In addition, the fidelity of the alkenone paleothermometry could also be compromised by nonthermal physiological growth factors including nutrient and light availability, as evidenced by the findings of the laboratory culture studies (e.g., Epstein et al., 1998, 2001; Prahl et al., 2003; Versteegh et al., 2001), the suspended particulate matters in the water column (Popp et al., 2006), and the sediment (e.g., Prahl et al., 2006). The *E. huxleyi* culture study by Prahl et al. (2006) revealed that anomalously warm alkenone-derived SSTs are obtained under light-deprived conditions, while the opposite is true for nutrient-depleted conditions. Besides, the authors found that the alkenone/alkenoate composition also seems to respond to these physiological stresses, prompting the authors to propose that one could infer nonthermal physiological imprints on the alkenone-derived SST estimates by comparing the alkenone/alkenoate signature in sediments. However, it is not yet clear whether this index is representative of the global ocean.

In any case, it should be mentioned that the only alkenone distributions showing a direct and clear relationship with SST are those of type A, the most common distributions in the marine sediments. The distributions of type B alkenones in sedimentary samples give results that strongly deviate from the instrumental temperature values when their unsaturation indices are used for calculation with the above-mentioned equations, or other equivalent expressions available in the literature. The usefulness of type B alkenone distributions for temperature estimation remains to be demonstrated. In other words, the alkenones are good paleothermometers for marine systems, but up to now, they cannot be used for temperature estimation in brackish or freshwater environments.

Relationship Between Suspended Particulate Matters in Water Column and SST

Although the alkenone paleothermometer is one of the most robust proxies available to paleoceanographers, there are still some uncertainties that might constrain its accuracy in SST estimation. For instance, the global sediment core-top calibration differs substantially from the surface water calibration (see Figure 5). In the environment where SSTs are lower than

20 °C, the SST estimates based on core-top and culture calibration are consistently warmer than those derived from the surface water calibration, especially in the intermediate temperature range (10–15 °C). The underlying causes for this discrepancy are unclear. Based on the result of a simple model calculation, Conte et al. (2006) speculated that the combination of differential degradation of C_{37:3} over C_{37:2} alkenones, and the seasonality of alkenone production are the likely causative factors. Indeed, it is difficult to compare the C₃₇ alkenone composition in the water column and in underlying sediments due to the large differences in timescale the two signals represent. Water column particles are collected either by active filtration systems or by passive collection devices, namely, sediment traps. In the first case, sample collection may involve hours or 1 or 2 days at the most. In the second, samples may be collected for 1 or 2 years at weekly or monthly intervals in the most favorable cases. These samples are compared with sediment sections that may represent time-integrated periods of hundreds of years. Furthermore, sediments are not always collected with devices that could ensure collection of the true sediment–water interface section. It may happen that core-top sediments available for comparison are 1 or 2 ka old due to sediment mixing or to failure to recover the uppermost sediment during coring.

Comparison between alkenone analysis of filtered particles and *in situ* temperature measurements in the Gulf of Lions, under well-defined water column stratification conditions (Bentaleb et al., 1999), showed that the U^K₃₇ values of the suspended particles at 5-, 30-, and 1100-m depth corresponded to the water temperature where the maximum Haptophyceae abundance (30 m) occurred. Remarkably, the U^K₃₇ index at 5 and 1100 m was the same as that found at 30 m, indicating that the alkenones eventually incorporated into the sediment were likely to record the water temperature where the highest abundance of the source organism was found. These results are in general agreement with sediment trap data from 42° N in the eastern Pacific Ocean, where it was observed that alkenone production during periods of offshore stratification occurred at the subsurface chlorophyll maximum, and therefore, the U^K₃₇ index reflected the temperature measured at this depth

Table 1 Structures of the alkenones found in sedimentary environments

Alkenone	Structure	Molecular weight
Me37:4	(E,E,E,E)-8,15,22,29-heptatriacontatetraen-2-one	526
Me37:3	(E,E,E)-8,15,22-heptatriacontatrien-2-one	528
Me37:2	(E,E)-15,22-heptatriacontadien-2-one	530
Et38:4	(E,E,E,E)-9,16,23,30-octatriacontatetraen-3-one	540
Me38:4	(E,E,E,E)-8,15,22,29-octatriacontatetraen-2-one	540
Et38:3	(E,E,E)-9,16,23-octatriacontatrien-3-one	542
Me38:3	(E,E,E)-8,15,22-octatriacontatrien-2-one	542
Et38:2	(E,E)-16,23-octatriacontadien-3-one	544
Me38:2	(E,E)-15,22-octatriacontadien-2-one	544
Et39:4	(E,E,E,E)-9,16,23,30-nonatriacontatetraen-3-one	562
Me39:4	(E,E,E,E)-8,15,22,29-nonatriacontatetraen-2-one	554
Et39:3	(E,E,E)-9,16,23-nonatriacontatrien-3-one	556
Me39:3	(E,E,E)-8,15,22-nonatriacontatrien-2-one	556
Et39:2	(E,E)-16,23-nonatriacontadien-3-one	558
Me39:2	(E,E)-15,22-nonatriacontadien-2-one	558
Et40:3	(E,E,E)-9,16,23-tetracontatrien-3-one	570
Et40:2	(E,E)-16,23-tetracontadien-3-one	572

Table 2 Reported average $U^{K'}_{37}$ data on sediment trap particles and underlying sediments

Location	Deployment period	$U^{K'}_{37}$ Measurements		References
		Sediment traps	Underlying sediment	
Gulf of California 27°53' N, 111°40' W	14 January 1996–21 September 1997	0.87 (500) ^a	0.90 (665) ^a	Göni et al. (2001)
Northeast Pacific Ocean 41°55' N, 125°40' W	22 September 1987–16 September 1988			Prahl et al. (1993)
42°10' N, 127°30' W		0.38 (1000)	0.42 (2712)	
41°35' N, 131°45' W		0.38 (1000)	0.43 (3111)	
Northeast Atlantic 47°54' N, 21°30' E	April 1989–April 1990	0.33 (1000)	0.45 (3680)	Rosell-Melé et al. (2000)
Norwegian Basin 69°41.2' N, 0°27.8' E	August 1991–July 1992	0.40 (3700)	0.57 (4400)	Thomsen et al. (1998)
Barents Sea 75°11.8' N, 12°29.2' E	March–July 1991	0.53 (500)	0.53 (3290)	Thomsen et al. (1998)
North Pacific off Japan 34°10' N, 142° E	5 March 1991–2 March 1992	0.59 (3000)	0.58 (2050)	Sawada et al. (1998)
		0.18 (1840)	0.74 (9200)	
		0.76 (1950)		
		0.76 (1674)		
		0.74 (4180)		
		0.74 (5687)		
		0.72 (8688)		

^aWater column depth (m).

(i.e., temperatures colder than the SST) (Prahl et al., 1993). A summary of studies dealing with correspondences between $U^{K'}_{37}$ indices in water column-suspended particles and underlying sediments is reported in Table 2. Good agreements were found in the Gulf of California (Göni et al., 2001), the Norwegian Basin (Thomsen et al., 1998), and the North Pacific off Japan (Sawada et al., 1998). In the Norwegian Sea, the $U^{K'}_{37}$ index of the sediment perfectly matched with the alkenone composition of material collected in a trap at 500-m water depth. The sediment traps deployed in the North Pacific off Japan extend over the deepest water column ever investigated (9200 m). No significant change was observed between the average $U^{K'}_{37}$ of sediment traps deployed at 1674, 4180, 5687, and 8688 m depth (13 samples collected in successive time intervals) and the underlying sediment (Sawada et al., 1998).

Cases of lack of correspondence in the alkenone composition in the water column and underlying sediments have also been reported (see Table 2 for some examples). In the Barents Sea, sediment trap data at 1840- and 1950-m water depth and underlying sediment (2050 m) showed considerable scatter. Resuspension and lateral advection were proposed as an explanation of these discrepancies (Thomsen et al., 1998). A major contrast between annual average sediment trap measurements at 3700 m and underlying sediment (4400 m) was reported in the northeast Atlantic (Rosell-Melé et al., 2000). Allochthonous alkenone input from elsewhere, for example, the polar front, instead of overlying waters, was proposed as an explanation. A transect of three sediment traps and underlying sediments in the northeast Pacific Ocean extending from the coast to the open ocean (Prahl et al., 1993) revealed an increase in $U^{K'}_{37}$ index between the sediment traps and sediments with increasing distance from the shore (Table 2): 0.04 (nearshore), 0.05 (further offshore), and 0.12 (gyre). The authors suggested that the observed differences did not necessarily involve selective degradation of the more unsaturated alkenones. The discrepancies could arise from other mechanisms, such as differential

preservation of the settling particles arriving at the seafloor over the year, because the $U^{K'}_{37}$ composition of the sediment trap particles collected monthly or weekly was not uniform.

Overall, these results show a good correspondence between $U^{K'}_{37}$ indices in water column particles and underlying sediments, indicating that the processes involving quantitative changes in C_{37} alkenones during particle settling do not result, in general, in significant changes of $U^{K'}_{37}$. To this end, the results of Sawada et al. (1998) concerning a water column of 9200 m are particularly significant.

Validity of Alkenone Paleothermometry in the Geological Past

Another important aspect in relation to the validity of alkenone paleothermometry throughout the Quaternary concerns the issue of whether the record could be affected by changes in evolving species. The dominance of *E. huxleyi*, the best-known alkenone-producing organism in the modern ocean, is restricted only to the last 70–80 ka, and the first appearance of this coccolithophorid in the oceans dates to 280 ka BP (Thierstein et al., 1977). The presence of alkenones in sediments before the appearance of *E. huxleyi* has been related to other algal species from the family *Gephyrocapsaceae* (*Gephyrocapsa* and *Emiliania* species) (Marlowe et al., 1990). However, studies comparing downcore SST profiles with abundances of coccoliths have not observed abrupt discontinuities paralleling the changes in predominant species. For example, the reversal in the dominant coccolithophorid species observed 74 ka BP at the Walvis Ridge did not significantly influence the alkenone unsaturation index–SST relationship (Müller et al., 1997). Similarly, in a study of a core situated in the central north Atlantic, significant changes in the relative composition of the C_{37} alkenones were not related to the reversals of the populations of *Gephyrocapsa* and *Emiliania* species over the last 290 ka (Villanueva et al., 2002). On a longer timescale, a comparison of the coccoliths and alkenone unsaturation index in the southeast Atlantic between 1500 and 500 ka BP

also demonstrates that evolutionary events and changes in species dominance (among *Pseudoemiliania lacunosa*, medium *Reticulofenestra*, *G. caribbeanica*, *G. oceanica*, small *Gephyrocapsa*, and others) did not affect the alkenone unsaturation index (McClymont et al., 2005). These results suggest that the alkenone paleothermometry can be applied to older sediments, prior to the first occurrence of *E. huxleyi* and *G. oceanica*. In recent years, alkenone paleothermometry has been successfully used to generate long-term SST records (e.g., Lawrence et al., 2006), showing the ability to provide reasonable temperature reconstruction for at least the last 5 My. In addition, the alkenone unsaturation index has recently been applied to older sediments of the Paleogene, where alkenone-based SSTs are similar to those inferred from other temperature proxies (Bijl et al., 2010), suggesting that even in sediments of this age, the index may still provide plausible SST estimates.

In fact, the influence of population genetics on the relationship between the alkenone unsaturation index and SST in older sediments could be examined using the relative abundance of the C_{37} and the C_{38} alkenones (C_{37}/C_{38}). This ratio shows a range of values in modern sediment and water column particulates between 0.7 and 1.4 (e.g., Conte et al., 1998, 2001) and around 1.8 in a culture study (Prahl et al., 1988). Therefore, if the alkenones found in sediments prior to the appearance of *E. huxleyi* exhibit C_{37}/C_{38} values within this range, one could assume that the fossil alkenone source organisms had a similar temperature adaptation mechanism in the lipids, validating the application of the calibration based on global surface sediments and culture studies. This approach has been attempted in several studies of samples from the late Pleistocene (e.g., Müller et al., 1997), across the mid-Pleistocene transition (e.g., McClymont et al., 2005) and the early Pliocene (e.g., Beltran et al., 2011). In addition, this ratio has been used to distinguish between marine and lacustrine alkenone producers during the Eocene (Weller and Stein, 2008), via the comparison of their sedimentary C_{37}/C_{38} data with a suite of published C_{37}/C_{38} values of several culture studies on different species of organism, that is, *G. oceanica* (Volkman et al., 1995), *Isochrysis galbana*, and *C. lamellosa* (Marlowe et al., 1984). Nevertheless, one should keep in mind that this ratio is empirical and must thus be interpreted with caution, since it has been found to vary as a function of growth temperature in the *E. huxleyi* cultures (Prahl et al., 1988), and some differences could be observed between cultured and natural alkenone producers (e.g., Conte et al., 1998; Yamamoto et al., 2000).

Alkenone Production: Depth and Seasonality

Due to the autotrophic lifestyle of the source organism of alkenones, that is, haptophyte algae, it is generally assumed that the alkenones reflect the algal growth temperature in the photic zone where light is available for them to carry out photosynthesis. Empirical evidence provided by the global core-top studies suggest that the best correlation exists for the water surface (0 m), suggesting that in most cases, the alkenones incorporated into the sediment are exported from the upper photic zone. Several water column studies suggest that the alkenone synthesizers could also thrive at the subsurface depths, such as the mixed layer (Bentaleb et al., 1999; Cortés et al., 2001) and thermocline waters (Ohkouchi et al., 1999).

Especially, in the latter case, where exported alkenones originated from the thermocline which could be as deep as 200 m and receive little heat from the sun, one might find that the alkenone-derived SST estimates are spuriously cold relative to the temperature of the overlaying surface water.

In spite of the best correlation found between the alkenone unsaturation index and mean annual SST (relatively to other seasons) exhibited by the global core-top compilation (Müller et al., 1998), care should be exercised in interpreting alkenone-derived SSTs as annual mean temperature estimates, especially in the high latitudes. This is because the seasonality of alkenone production in the ocean varies progressively with the latitudes. At high latitudes where temperatures are low and light is not available year-round, alkenone production is most likely limited to summer, thus, it is plausible that the alkenone-derived SSTs in this region might reflect summer SST instead of the annual mean. This has been observed in the core-top studies in the high latitudes in the Pacific Ocean (Prahl et al., 2010) and the Southern Ocean (Sikes et al., 1997). In contrast, the peak production of alkenones in the low and mid-latitudes occurs in winter or spring (e.g., Prahl et al., 1993), with some contribution during other seasons as well. Consequently, the sedimentary alkenones from these oceanic settings reflect a more integrated signal over the whole year.

Diagenetic Alteration of Alkenone Unsaturation Index

C_{37} – C_{39} alkenones are among the most refractory lipids known (Volkman et al., 1980). The causes of this resistance to degradation may be related to low solubility, protective package within the algal cells, and their unusual Z (*trans*) double bond configuration (Table 1), because many bacteria do not have enzymatic systems for the transformation of this type of double bond. The principal structural difference between the alkenones included in the U^K_{37} indices relates to the degrees of unsaturation, for example, $C_{37:2}$, $C_{37:3}$, and $C_{37:4}$. In general, the greater the degree of unsaturation, the more easily a lipid is degraded. According to the ratios previously defined (eqns [1] and [2]), SST estimation will be biased if selective degradation of $C_{37:2}$, $C_{37:3}$, or $C_{37:4}$ alkenones occurs during transport through the water column or in sediments. On the other hand, no bias in the SST estimates will be produced if all C_{37} alkenones are degraded in the same proportion, even in cases of extensive alteration.

Reported microbial degradation experiments of $C_{37:3}$ and $C_{37:2}$ alkenone mixtures with $U^K_{37} = 0.78$ and 0.6 under oxic, sulfate-reducing, and methanogenic conditions showed that the maximal percentages of consumption were 83, 43, and 80%, respectively, but that after this extent of degradation, the differences in U^K_{37} index were small, with a 0.4 °C bias in the worst case (Teece et al., 1998). Meanwhile, feeding experiments of zooplankton on alkenone producers indicated that the alkenone unsaturation index is not significantly altered by grazing, in which the C_{37} alkenones were found to be absent from the grazer bodies but abundant in the fecal pellets (Grice et al., 1998). The maximal U^K_{37} difference between dietary algae and released pellets was 0.03, equivalent to 0.9 °C. Nevertheless, recent studies focusing on the effect of the bacterial degradation on alkenones (e.g., Rontani et al., 2008) demonstrate that some aerobic bacteria have the

propensity to preferentially degrade the more unsaturated alkenones, introducing a 'warm' bias in alkenone-derived SST estimates.

After sedimentation, the transformation of organic matter, including lipids, strongly depends on the activity of microbiota and benthic organisms, at least during the first stages of diagenesis. Obviously, the degradation rates depend on sedimentary conditions. The concentrations of C_{37} alkenones decrease faster in oxic than in anoxic environments. Postdepositional oxidation episodes also involve strong decreases in C_{37} alkenone concentrations. However, the relevant point for paleoceanographic studies is whether the U'_{37} ratio is altered as a consequence of these diagenetic processes. In a study of core tops from the Peru margin, no significant changes in U'_{37} were observed, despite a decrease of total alkenone concentration of 30% (McCaffrey et al., 1990). In contrast, other studies (Gong and Hollander, 1999; Hoefs et al., 1998; Huguet et al., 2009) reported the selective degradation of the $C_{37:3}$ alkenone when comparing sediment records in oxic and anoxic sediments, resulting in spuriously warm SST estimates. Recently, epoxyketones were discovered in the sediments from the coastal region of southeast Alaska (Prah et al., 2010), which are produced as a consequence of selective alkenone degradation by an aerobic bacterial process (Rontani et al., 2008). By performing a semiquantitative calculation, it was proposed that the preferential degradation of alkenones could account for a considerable warming effect in the SST estimates. Since it is still in the early stage of development, more work is needed before the epoxyketones can be used to quantify the warm bias in SST estimates caused by the preferential degradation of alkenones in the sediments.

Stratigraphic Offsets Between the Alkenones and Other Sedimentary Proxies

In some depositional environments, vigorous currents or eddies are able to transport alkenones laterally over great distances from their source, as seen in the southwestern Atlantic (Benthien and Müller, 2000) and sediment drifts in the Cape Basin (Sachs and Anderson, 2003). Depending on the source of these advected alkenones, one could observe cold or warm biased SSTs in the sediment, or a discrepancy between different paleoceanographic proxies at the same site. However, lateral advection at sediment drifts does not necessarily result in erratic downcore SST variation patterns as demonstrated by Sachs and Anderson (2003) using a multiproxy approach and thorium-derived focusing factors.

Recent advancements in the radiocarbon technique provide additional ways of constraining the source origin of alkenones in marine sediments. For instance, at the Bermuda Rise (Ohkouchi et al., 2002) and the Benguela upwelling system off Namibia (Mollenhauer et al., 2003), studies using accelerator mass spectrometry (AMS) ^{14}C revealed discrepancies between the fine organic matter ($<63 \mu m$), including the algal C_{37} alkenones, and planktonic Foraminifera in the marine sediments, which implied that the alkenones were several hundreds or even thousands of years older than the Foraminifera that are taken as reference materials deposited at the same depth. This sedimentary asynchrony is explained by the strong differences in size between coccoliths (on the order of $5 \mu m$)

and the foraminiferal tests ($>150 \mu m$). In cases of resuspension due to intense bottom currents, the finer alkenone-bearing particles are susceptible to easier mobilization than the larger foraminiferal remains (Ohkouchi et al., 2002). The result is that older, previously deposited material can be remobilized and deposited in a new environment, leaving the fine fraction signal old with respect to the adjacent coarse material. However, the differences in particle size are not a priori very relevant for sedimentation through the water column because most algal remains arrive to the bottom as aggregates, for example, in the form of zooplankton fecal pellets ($\sim 50 \mu m$).

The ages of the marine sedimentary sections are currently determined from the composition of Foraminifera, either by interpretation of the $\delta^{18}O$ curve or by measurement of AMS ^{14}C in their carbonate skeleton. The potential temporal offset between alkenone SST and foraminiferal records may thus jeopardize the possibility of using alkenones for high-resolution studies in sedimentary environments because the signal that is stored in their composition may be older than the age determined for the sedimentary sections in which they are found.

Close comparison of high-resolution alkenone-derived SST and $\delta^{18}O$ profiles afford a practical way to detect whether this problem is significant in the sedimentary core sections under study. When both proxies exhibit synchronous changes, possible temporal offsets due to dissimilar sedimentation processes between organic matter and Foraminifera are negligible. This approach has been followed in some studies (e.g., Martrat et al., 2004), and in most cases, no sedimentation asynchrony was observed.

Concluding Remarks

Alkenone unsaturation indices are reliable tools for past SST estimations. The results from core-top analyses are coincident with laboratory culture experiments and confirm that the calibration equations obtained from cultures of *E. huxleyi* provide reliable estimates of mean annual SST in all oceans, except perhaps in the high latitudes where alkenone production is substantially biased toward the warm seasons. There is, however, a need for further investigation on the reliability of this equation in the cold- and warm-temperature ends. In this respect, sufficient quantitative information should be provided together with the alkenone studies, for possible SST reassessment in the future.

There is no consistent evidence of selective biogenic degradation of the $C_{37:2}$ and $C_{37:3}$ alkenones by organisms that consume them. Studies in the literature report little or no changes of U'_{37} as a consequence of C_{37} alkenone degradation in sediments or in the water column. These studies outweigh those in which significant changes were observed. More studies encompassing the whole range of U'_{37} values, different water column environments, and sedimentary conditions would help to clarify this issue.

In any case, the alkenone paleothermometry should be used with careful attention to the possible analytical biases due to low concentrations and irreversible adsorption in the chromatographic column. In addition, caution is warranted in the interpretation of the alkenone-inferred SSTs, when offsets are detected between the alkenone- and Foraminifera-derived

proxy records, as these offsets could reflect problems of asynchronous sedimentation.

In summary, the alkenone paleothermometry is a very useful tool for the understanding of the climatic evolution of the oceans, but its predictive power is strengthened by placing it in the context of the other paleoceanographic proxies.

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See also: [Paleoceanography: Paleoceanography An Overview.](#)
[Paleoceanography, Biological Proxies: Coccolithophores.](#)

References

- Bard E (2001) Comparison of alkenone estimates with other paleotemperature proxies. *Geochemistry, Geophysics, Geosystems* 2: 1002.
- Beltran C, Flores JA, Sicre MA, Baudin F, Renard M, and de Rafélis M (2011) Long chain alkenones in the Early Pliocene Sicilian sediments (Trubi Formation – Punta di Maïata section): Implications for the alkenone paleothermometry. *Palaeogeography, Palaeoclimatology, Palaeoecology* 308: 253–263.
- Bendle J, Rosell-Melé A, and Ziveri P (2005) Variability of unusual distributions of alkenones in the surface waters of the Nordic seas. *Paleoceanography* 20: PA2001.
- Bentaleb I, Grimalt JO, Vidussi F, et al. (1999) The C-37 alkenone record of seawater temperature during seasonal thermocline stratification. *Marine Chemistry* 64: 301–313.
- Benthien A and Müller PJ (2000) Anomalous low alkenone temperatures caused by lateral particle and sediment transport in the Malvinas Current region, western Argentine Basin. *Deep Sea Research Part I: Oceanographic Research Papers* 47: 2369–2393.
- Bijl PK, Houben AJP, Schouten S, et al. (2010) Transient middle Eocene atmospheric CO₂ and temperature variations. *Science* 330: 819–821.
- Boon JJ, van der Meer FW, Schuyl JW, De Leeuw J, and Schenck PA (1978) Organic geochemical analyses of core samples from Site 362, Walvis Ridge. *DSDP Leg 40, Initial Reports*, Volume 38–41(supplement): 627–637.
- Brassell SC and Dumitrescu M (2004) Recognition of alkenones in a lower Aptian porcellanite from the west-central Pacific. *Organic Geochemistry* 35: 181–188.
- Brassell SC, Eglinton G, Marlowe IT, Pflaumann U, and Sarnthein M (1986) Molecular stratigraphy – A new tool for climatic assessment. *Nature* 320: 129–133.
- Cacho I, Grimalt JO, Pelejero C, et al. (1999) Dansgaard-Oeschger and Heinrich event imprints in Alboran Sea paleotemperatures. *Paleoceanography* 14: 698–705.
- Calvo E, Grimalt JO, and Jansen E (2002) High resolution U^K₃₇ sea surface temperature reconstruction in the Norwegian Sea during the Holocene. *Quaternary Science Reviews* 21: 1385–1394.
- Conte MH, Sicre MA, Ruhlmann C, et al. (2006) Global temperature calibration of the alkenone unsaturation index U^K₃₇ in surface waters and comparison with surface sediments. *Geochemistry, Geophysics, Geosystems* 7: Q02005.
- Conte MH, Thompson A, Lesley D, and Harris RP (1998) Genetic and physiological influences on the alkenone/alkenoate versus growth temperature relationship in *Emiliania huxleyi* and *Gephyrocapsa oceanica*. *Geochimica Et Cosmochimica Acta* 62: 51–68.
- Conte MH, Weber JC, King LL, and Wakeham SG (2001) The alkenone temperature signal in western North Atlantic surface waters. *Geochimica Et Cosmochimica Acta* 65: 4275–4287.
- Cortés MY, Bollmann J, and Thierstein HR (2001) Coccolithophore ecology at the HOT station ALOHA, Hawaii. *Deep Sea Research Part II: Topical Studies in Oceanography* 48: 1957–1981.
- de Leeuw JW, van der Meer FW, Rijpstra WIC, and Schenck PA (1980) On the occurrence and structural identification of long chain unsaturated ketones and hydrocarbons in sediments. *Physics and Chemistry of the Earth* 12: 211–217.
- Eglinton G, Bradshaw SA, Rosell A, Sarnthein M, Pflaumann U, and Tiedemann R (1992) Molecular record of secular sea-surface temperature-changes on 100-year timescales for glacial termination-I, termination-II and termination-IV. *Nature* 356: 423–426.
- Epstein BL, D'Hondt S, and Hargraves PE (2001) The possible metabolic role of C₃₇ alkenones in *Emiliania huxleyi*. *Organic Geochemistry* 32: 867–875.
- Epstein BL, D'Hondt S, Quinn JG, Zhang JP, and Hargraves PE (1998) An effect of dissolved nutrient concentrations on alkenone-based temperature estimates. *Paleoceanography* 13: 122–126.
- Gong C and Hollander DJ (1999) Evidence for differential degradation of alkenones under contrasting bottom water oxygen conditions: Implication for paleotemperature reconstruction. *Geochimica Et Cosmochimica Acta* 63: 405–411.
- Göni MA, Hartz DM, Thunell RC, and Tappa E (2001) Oceanographic considerations for the application of the alkenone-based paleotemperature U^K₃₇ index in the Gulf of California. *Geochimica Et Cosmochimica Acta* 65: 545–557.
- Grice K, Klein Breteler WCM, Schouten S, Grossi V, de Leeuw JW, and Damsté JSS (1998) Effects of zooplankton herbivory on biomarker proxy records. *Paleoceanography* 13: 686–693.
- Grimalt JO, Calvo E, and Pelejero C (2001) Sea surface paleotemperature errors in U^K₃₇ estimation due to alkenone measurements near the limit of detection. *Paleoceanography* 16: 226–232.
- Harada N, Sato M, Shiraishi A, and Honda MC (2006) Characteristics of alkenone distributions in suspended and sinking particles in the northwestern North Pacific. *Geochimica Et Cosmochimica Acta* 70: 2045–2062.
- Harada N, Shin KH, Murata A, Uchida M, and Nakatani T (2003) Characteristics of alkenones synthesized by a bloom of *Emiliania huxleyi* in the Bering Sea. *Geochimica Et Cosmochimica Acta* 67: 1507–1519.
- Hefter J (2008) Analysis of alkenone unsaturation indices with fast gas chromatography/time-of-flight mass spectrometry. *Analytical Chemistry* 80: 2161–2170.
- Herbert TD, Heinrich DH, and Karl KT (2003) Alkenone paleotemperature determinations. In: *Treatise on Geochemistry*. Oxford: Pergamon.
- Herbert TD, Peterson LC, Lawrence KT, and Liu Z (2010) Tropical ocean temperatures over the past 3.5 million years. *Science* 328: 1530–1534.
- Hoefs MJL, Versteegh GJM, Rijpstra WIC, de Leeuw JW, and Damsté JSS (1998) Postdepositional oxic degradation of alkenones: Implications for the measurement of palaeo sea surface temperatures. *Paleoceanography* 13: 42–49.
- Huguet C, Kim JH, de Lange GJ, Damsté JSS, and Schouten S (2009) Effects of long term oxic degradation on the U^K₃₇, TEX₈₆ and BIT organic proxies. *Organic Geochemistry* 40: 1188–1194.
- Jaraula CMB, Brassell SC, Morgan-Kiss RM, Doran PT, and Kenig F (2010) Origin and tentative identification of tri to pentaunsaturated ketones in sediments from Lake Fryxell, East Antarctica. *Organic Geochemistry* 41: 386–397.
- Kennedy JA and Brassell SC (1992) Molecular records of 20th-century El-Niño events in Laminated sediments from the Santa-Barbara basin. *Nature* 357: 62–64.
- Lamy F, Kaiser JRM, Ninnemann U, Hebbeln D, Arz HW, and Stoner J (2004) Antarctic timing of surface water changes off Chile and Patagonian ice sheet response. *Science* 304: 1959–1962.
- Lawrence KT, Liu ZH, and Herbert TD (2006) Evolution of the eastern tropical Pacific through Plio-Pleistocene glaciation. *Science* 312: 79–83.
- Marlowe IT, Brassell SC, Eglinton G, and Green JC (1990) Long-chain alkenones and alkyl alkenoates and the fossil coccolith record of marine sediments. *Chemical Geology* 88: 349–375.
- Marlowe IT, Green JC, Neal AC, Brassell SC, Eglinton G, and Course PA (1984) Long-chain (N-C₃₇-C₃₉) alkenones in the Prymnesiophyceae – Distribution of alkenones and other lipids and their taxonomic significance. *British Phycological Journal* 19: 203–216.
- Martínez-García A, Rosell-Melé A, Geibert W, et al. (2009) Links between iron supply, marine productivity, sea surface temperature, and CO₂ over the last 1.1 Ma. *Paleoceanography* 24: PA1207.
- Martrat B, Grimalt JO, Lopez-Martínez C, et al. (2004) Abrupt temperature changes in the Western Mediterranean over the past 250,000 years. *Science* 306: 1762–1765.
- Martrat B, Grimalt JO, Shackleton NJ, de Abreu L, Hutterli MA, and Stocker TF (2007) Four climate cycles of recurring deep and surface water destabilizations on the Iberian margin. *Science* 317: 502–507.
- Mccaffrey MA, Farrington JW, and Repeta DJ (1990) The organic geochemistry of Peru margin surface sediments: 1. A comparison of the C₃₇ alkenone and historical El Niño records. *Geochimica Et Cosmochimica Acta* 54: 1671–1682.

- McClymont EL, Rosell-Melé A, Giraudeau J, Pierre C, and Lloyd JM (2005) Alkenone and coccolith records of the mid-Pleistocene in the south-east Atlantic: Implications for the U^{K}_{37} index and South African climate. *Quaternary Science Reviews* 24: 1559–1572.
- Mollenhauer G, Eglinton TI, Ohkouchi N, et al. (2003) Asynchronous alkenone and foraminifera records from the Benguela Upwelling System. *Geochimica Et Cosmochimica Acta* 67: 2157–2171.
- Müller PJ, Cepek M, Ruhland G, and Schneider RR (1997) Alkenone and coccolithophorid species changes in late Quaternary sediments from the Walvis Ridge: Implications for the alkenone paleotemperature method. *Palaeogeography, Palaeoclimatology, Palaeoecology* 135: 71–96.
- Müller PJ, Kirst G, Ruhland G, Keigwin LD, and Rosell-Melé A (1998) Calibration of the alkenone paleotemperature index U^{K}_{37} based on core-tops from the eastern South Atlantic and the global ocean (60 degrees N–60 degrees S). *Geochimica Et Cosmochimica Acta* 62: 1757–1772.
- Ohkouchi N, Eglinton TI, Keigwin LD, and Hayes JM (2002) Spatial and temporal offsets between proxy records in a sediment drift. *Science* 298: 1224–1227.
- Ohkouchi N, Kawamura K, Kawahata H, and Okada H (1999) Depth ranges of alkenone production in the central Pacific Ocean. *Global Biogeochemical Cycles* 13: 695–704.
- Pelejero C and Calvo E (2003) The upper end of the U^{K}_{37} temperature calibration revisited. *Geochemistry, Geophysics, Geosystems* 4: 1014.
- Popp BN, Bidigare RR, Deschenes B, et al. (2006) A new method for estimating growth rates of alkenone-producing haptophytes. *Limnology and Oceanography-Methods* 4: 114–129.
- Prahl FG, Collier RB, Dymond J, Lyle M, and Sparrow MA (1993) A biomarker perspective on prymnesiophyte productivity in the northeast Pacific-Ocean. *Deep Sea Research Part I: Oceanographic Research Papers* 40: 2061–2076.
- Prahl FG, Mix AC, and Sparrow MA (2006) Alkenone paleothermometry: Biological lessons from marine sediment records off western South America. *Geochimica Et Cosmochimica Acta* 70: 101–117.
- Prahl FG, Muehlhausen LA, and Zahnle DL (1988) Further evaluation of long-chain alkenones as indicators of paleoceanographic conditions. *Geochimica Et Cosmochimica Acta* 52: 2303–2310.
- Prahl FG, Rontani JF, Zabeti N, Walinsky SE, and Sparrow MA (2010) Systematic pattern in U^{K}_{37} – Temperature residuals for surface sediments from high latitude and other oceanographic settings. *Geochimica Et Cosmochimica Acta* 74: 131–143.
- Prahl FG and Wakeham SG (1987) Calibration of unsaturation patterns in long-chain ketone compositions for paleotemperature assessment. *Nature* 330: 367–369.
- Prahl FG, Wolfe GV, and Sparrow MA (2003) Physiological impacts on alkenone paleothermometry. *Paleoceanography* 18(2): 1025.
- Rechka JA and Maxwell JR (1988) Unusual long chain ketones of algal origin. *Tetrahedron Letters* 29: 2599–2600.
- Rontani JF, Beker B, and Volkman JK (2004) Long-chain alkenones and related compounds in the benthic haptophyte *Chrysotila lamellosa* Anand HAP 17. *Phytochemistry* 65: 117–126.
- Rontani JF, Harji R, Guasco S, et al. (2008) Degradation of alkenones by aerobic heterotrophic bacteria: Selective or not? *Organic Geochemistry* 39: 34–51.
- Rosell-Melé A (1998) Interhemispheric appraisal of the value of alkenone indices as temperature and salinity proxies in high-latitude locations. *Paleoceanography* 13: 694–703.
- Rosell-Melé A, Bard E, Emeis KC, et al. (2001) Precision of the current methods to measure the alkenone proxy U^{K}_{37} and absolute alkenone abundance in sediments: Results of an interlaboratory comparison study. *Geochemistry, Geophysics, Geosystems* 2: 1046.
- Rosell-Melé A, Combes P, Muller PJ, and Ziveri P (2000) Alkenone fluxes and anomalous U^{K}_{37} values during 1989–1990 in the Northeast Atlantic (48 degrees N 21 degrees W). *Marine Chemistry* 71: 251–264.
- Rosell-Melé A, Eglinton G, Pflaumann U, and Sarinthein M (1995) Atlantic core-top calibration of the U^{K}_{37} index as a sea-surface palaeotemperature indicator. *Geochimica Et Cosmochimica Acta* 59: 3099–3107.
- Sachs JP and Anderson RF (2003) Fidelity of alkenone paleotemperatures in southern Cape Basin sediment drifts. *Paleoceanography* 18: 1082.
- Sachs JP and Lehman SJ (1999) Subtropical north Atlantic temperatures 60,000 to 30,000 years ago. *Science* 286: 756–759.
- Sawada M, Handa N, and Nakatsuka T (1998) Production and transport of long-chain alkenones and alkyl alkenoates in a sea water column in the northwestern Pacific off central Japan. *Marine Chemistry* 59: 219–234.
- Sicre MA, Yiou P, Eiriksson J, et al. (2008) A 4500-year reconstruction of sea surface temperature variability at decadal time-scales off North Iceland. *Quaternary Science Reviews* 27: 2041–2047.
- Sikes EL, Farrington JW, and Keigwin LD (1991) Use of the alkenone unsaturation ratio U^{K}_{37} to determine past sea-surface temperatures – Core-top SST calibrations and methodology considerations. *Earth and Planetary Science Letters* 104: 36–47.
- Sikes EL and Sicre MA (2002) Relationship of the tetra-unsaturated C_{37} alkenone to salinity and temperature: Implications for paleoproxy applications. *Geochemistry, Geophysics, Geosystems* 3(11): 1063.
- Sikes EL, Volkman JK, Robertson LG, and Pichon JJ (1997) Alkenones and alkenes in surface waters and sediments of the Southern Ocean: Implications for paleotemperature estimation in polar regions. *Geochimica Et Cosmochimica Acta* 61: 1495–1505.
- Sonzogni C, Bard E, Rostek F, Lafont R, Rosell-Melé A, and Eglinton G (1997) Core-top calibration of the alkenone index vs sea surface temperature in the Indian Ocean. *Deep Sea Research Part II: Topical Studies in Oceanography* 44: 1445–1460.
- Teece MA, Getliff JM, Lettley JW, Parkes RJ, and Maxwell JR (1998) Microbial degradation of the marine prymnesiophyte *Emiliania huxleyi* under oxic and anoxic conditions as a model for early diagenesis: Long chain alkadienes, alkenones and alkyl alkenoates. *Organic Geochemistry* 29: 863–880.
- Thierstein HR, Geitzenauer KR, and Molino B (1977) Global synchronicity of late Quaternary coccolith datum levels – Validation by oxygen isotopes. *Geology* 5: 400–404.
- Thomsen C, Schulz-Bull DE, Petrick G, and Duinker JC (1998) Seasonal variability of the long-chain alkenone flux and the effect on the U^{K}_{37} -index in the Norwegian Sea. *Organic Geochemistry* 28: 311–323.
- Versteegh GJM, Riegman R, de Leeuw JW, and Jansen JHF (2001) U^{K}_{37} values for *Isochrysis galbana* as a function of culture temperature, light intensity and nutrient concentrations. *Organic Geochemistry* 32: 785–794.
- Villanueva J, Flores JA, and Grimalt JO (2002) A detailed comparison of the U^{K}_{37} and coccolith records over the past 290 years: implications to the alkenone paleotemperature method. *Organic Geochemistry* 33: 897–905.
- Villanueva J and Grimalt JO (1997) Gas chromatographic tuning of the U^{K}_{37} paleothermometer. *Analytical Chemistry* 69: 3329–3332.
- Volkman JK, Barrett SM, Blackburn SI, and Sikes EL (1995) Alkenones in *Gephyrocapsa oceanica* – Implications for studies of paleoclimate. *Geochimica Et Cosmochimica Acta* 59: 513–520.
- Volkman JK, Eglinton G, Corner EDS, and Sargent JR (1980) *Novel Unsaturated Straight-Chain C_{37} – C_{39} Methyl and Ethyl Ketones in Marine Sediments and a Coccolithophore *Emiliania huxleyi**. Oxford: Pergamon.
- Weller P and Stein R (2008) Paleogene biomarker records from the central Arctic Ocean (Integrated Ocean Drilling Program Expedition 302): Organic carbon sources, anoxia, and sea surface temperature. *Paleoceanography* 23: PA1S17.
- Yamamoto M, Shiraiwa Y, and Inouye I (2000) Physiological responses of lipids in *Emiliania huxleyi* and *Gephyrocapsa oceanica* (Haptophyceae) to growth status and their implications for alkenone paleothermometry. *Organic Geochemistry* 31: 799–811.

Benthic Foraminifera

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Biology

Benthic Foraminifera are single-celled, eukaryotic, preferentially marine microorganisms with an external protective shell called a test. Benthic foraminiferal shells can be of either calcium carbonate secreted by the cell (calcareous) or of clay, silt, and sand size sediments cemented together (agglutinated). A few of the freshwater benthic Foraminifera have no shell (allogromids). Tests begin with an initial chamber (proloculus) secreted by the zooid, and grow through the progressive addition of further chambers, all of which are connected. Chambers are separated by a wall (septum) and joined by the foramen (aperture). The shells vary in size from being as large as over ~5 cm or even more (e.g., *nummulites*) to several microns. The shell shapes vary from being single chambered (unilocular) to chambers arranged in one, two, or more rows (uniserial, biserial, and multiserial, respectively) as well as with chambers arranged in spiral fashion (planispiral and trochospiral forms; [Figure 1](#)). The external surface of the shell is often ornamented with pits, ribs, spines, plates, etc. The benthic foraminifers communicate with the surrounding environment through an opening in the shell called the aperture. A web of protoplasm (pseudopodia) is extended through the aperture and other secondary openings called pores, to collect food material as well as for locomotion ([Figure 2](#)). The position of the aperture varies from species to species and this, combined with chamber shape, form of the test margin, structures on the test, and morphology, serves as the primary basis for species identification ([Boersma, 1980](#); [Loeblich and Tappan, 1987](#)).

At higher taxonomic levels, Foraminifera are subdivided on the basis of test-wall-forming architecture ([Loeblich and Tappan, 1987](#)). This has largely to do with the organization of calcite crystals in relation to the organic membrane that serves as a template for test secretion. A review of classification is presented in [Sen Gupta \(1999\)](#). There are three significant groups: (1) test made of organic material (Allogromida), (2) test particles attached to a proteinaceous or mineralized matrix (agglutinated, 3 orders), and (3) test wall made of secreted carbonate (calcareous, 8 orders). At the generic and species levels, Foraminifera are recognized by test morphology.

Benthic Foraminifera reproduce by both sexual and asexual modes with alternation of generations reported in several species (see [Goldstein, 1999](#)). They are abundant in nearshore coastal waters and their abundance usually decreases with increasing water depth. Benthic Foraminifera evolved during the Cambrian and at present, they are very diverse, with over 10000 species reported from different parts of the world's oceans.

Ecology

Benthic Foraminifera live in the upper 5–10 cm of the sediments in all marine environments. Their distribution is

controlled by the physical (temperature, salinity, depth, turbulence, grain size, etc.), chemical (pH, dissolved oxygen, nutrients, etc.), and biological (organic matter, predators, etc.) characteristics of the sediments and the overlying seawater. Out of all these parameters, shallow water coastal benthic Foraminifera are able to tolerate a large range of seawater temperature and salinity ([Saraswat et al., 2011](#)). However, organic flux and dissolved oxygen are limiting factors for both the surface and subsurface distribution of the majority of benthic Foraminifera. Benthic foraminiferal distribution (including abundance, diversity, species richness, etc.) varies with sediment depth ([Goody, 2003](#); [Loubere and Fariduddin, 1999a](#); [Mackensen and Douglas, 1989](#); [Sun et al., 2006](#)). The depth distribution of benthic Foraminifera in an oligotrophic setting is controlled by food availability, while in eutrophic waters, the depth to which benthic Foraminifera can survive depends on oxygen levels, as explained by the TROX model (trophic-oxygen model) ([Figure 3](#); see [Jorissen et al., 2007](#), for review). The depth-wise species zonation is better developed in the deep oceans than in shallow waters because of differing turbulence conditions. However, in restricted environments such as silled basins and within the sediments, depletion of oxygen does not seem to be a limiting factor for a suite of benthic species. Living Foraminifera are routinely found in subsurface or basinal settings where oxygen levels are very low, and some have special adaptations to these conditions ([Bernhard, 1996](#)). Many are active searchers within their environment and may have associations with larger organisms. These associations may involve adaptation to the use of larger organism burrows as a way of having access to aerated water deeper within the sediments and so overcoming the low oxygen barrier ([Loubere et al., 1995](#)).

The calcareous (both calcite and aragonite) forms are present in all types of marine environments above the carbonate compensation depth (CCD). The agglutinated benthic Foraminifera are more abundant in brackish waters, and they also survive below the CCD. In low salinity and brackish environments, calcareous foraminiferal diversity is severely reduced to only one or two tolerant species, mainly because of low pH. In restricted or biologically challenging environments, agglutinated forms become dominant. Some Foraminifera in reef settings maintain symbionts in a manner similar to coral coelenterates ([Hallock, 1999](#)). Foraminifera may live attached to rock or plant surfaces as epiphytes; they may be epibenthic on sediment surfaces or shallow to deep infaunal in soft substrate ([Jorissen, 1999](#)). Foraminiferal test morphology appears related to habitat preference ([Corliss, 1985](#)).

On the larger scale, Foraminifera show two patterns in distribution and abundance of taxa. Firstly, they are zoned according to water depth. Zonation, and the taxa involved, varies from one region to the next, but the underlying factors are probably the same. In the shallow marine setting, on the

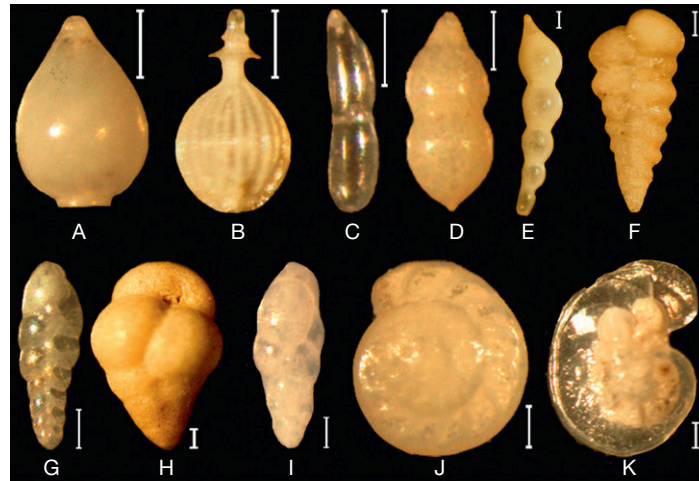


Figure 1 Different shapes of benthic foraminiferal shells. A–B, unilocular; C–D, bilocular; E, uniserial; F–G, biserial; H–I, multiserial; J–K, spiral.

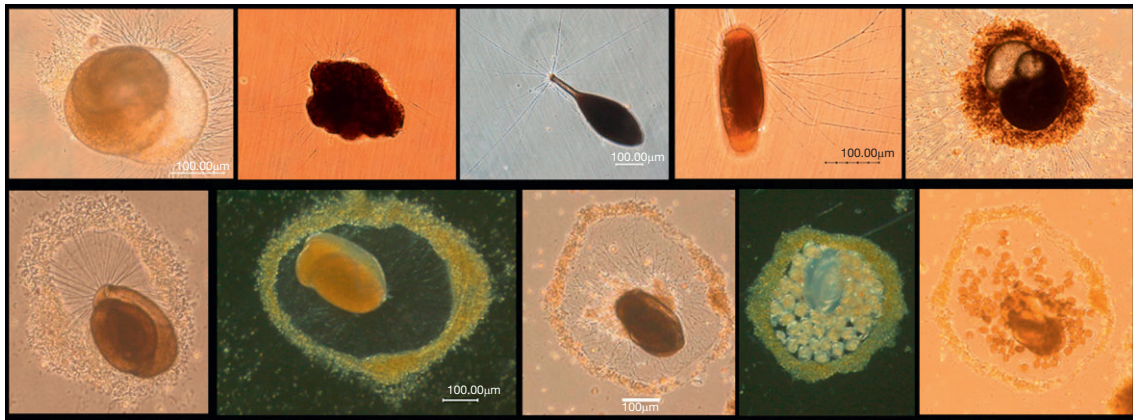


Figure 2 Live benthic Foraminifera with extended web of pseudopodia (upper row) along with growth stages including cyst formation in miliolids prior to reproduction (lower row).

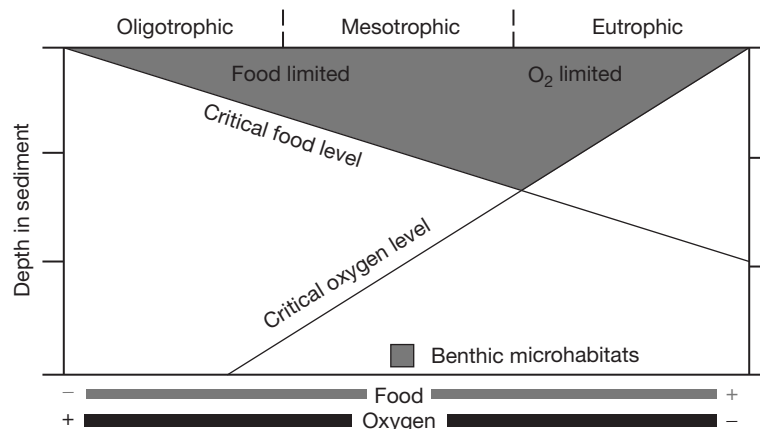


Figure 3 TROX model explaining factors affecting benthic foraminiferal distribution, especially in deep-sea environments. Modified from Jorissen FJ, Fontanier C, and Thomas E (2007) Paleooceanographical proxies based on deep-sea benthic foraminiferal assemblage characteristics. In: Hillaire-Marcel C and de Vernal A (eds.) *Proxies in Late Cenozoic Paleooceanography: Pt. 2: Biological Tracers and Biomarkers*, p. 263–326. Amsterdam: Elsevier.

inner continental shelf, substrate and physical conditions are highly variable, resulting in complex and patchy distributions. As with other organisms, a fundamental boundary exists between the sunlit ocean where plants are abundant and outer shelf and deeper habitats which are in the dark. Zonation of foraminiferal taxa is likely driven by the main ocean temperature gradient and the exponentially decreasing supply of useable organic matter with water depth. Beyond the continental shelf, the substrate becomes generally more uniform, allowing for broader distribution of biofacies (Culver, 1988). The second pattern of benthic foraminiferal distribution is biogeographic and strongly associated with latitude (Culver and Buzas, 1999). This pattern is most prominent with neritic, shallow ocean taxa and is clearly related to ocean thermal gradients.

From Living to Fossil

Foraminifera do not die a natural death, as either the whole cell divides into several small nuclei with little protoplasm, each of which grows into an individual, or the protoplasm of two individuals combines together, divides into spores, two of which fuse together to give rise to an individual. Thus, the original cell material keeps on multiplying and giving rise to offspring. Therefore, foraminiferal populations generate tests which accumulate in the sediments as a result of predation or reproduction. In reproduction, foraminiferal cytoplasm subdivides into zooids which abandon the parent's test. Test production is thus a function not only of standing stock of a taxon but also of generation time. Thus, the foraminiferal 'death' assemblage need not bear close numerical similarity to the living community.

Beyond the initial 'death' assemblage formation, there are a number of processes that influence what is preserved in the longer term (Murray, 2006). Test production occurs in a range of microhabitats from the epiphytic to deep infaunal, and this production is then mixed into accumulating sediments and through the geochemically active diagenetic zone within the upper 10–50 cm of the seabed. In this zone, sediments are bioturbated, well mixed, and passed from generally oxic conditions at the surface to anoxic, even sulfate reducing, conditions at depth. Geochemical gradients are generated in response to oxidation of organic matter within the sediments. Acids produced from this, and probably also associated biological activity, may degrade carbonate shells. The dissolution may lead to partial decalcification of calcareous foraminiferal tests and even to an increase in abundance of agglutinated tests in the death assemblages (Murray and Alve, 1999). The taphonomic alteration of calcareous benthic species depends on surface texture. Contrary to that, the agglutinated tests are relatively less prone to diagenetic changes and degrade in a more arbitrary manner, indicating that their degradation not only depends on test architecture but also the physical/mechanical processes (Berkeley et al., 2009). There is a net, and species selective, loss of tests within the sediment column. The extent of this depends on the intensity of the geochemical gradients (supply of easily oxidized organic carbon), the species involved, and the microhabitat in which the species lived. Deeper infaunal taxa have generally better preservation potential than any more lightly calcified epibenthic species. Thus, the living foraminiferal community represents a moment in time

and transient conditions, while the fossil assemblage represents the integration of production under a number of different conditions over long time period.

Benthic Foraminifera as Recorders of the Quaternary

The variation in abundance, morphology, and composition of the foraminiferal tests is an adaptive response of Foraminifera to changes in the environments they inhabit. Thus, the climatic response of benthic Foraminifera is used to infer past environmental conditions. Various characteristics of the benthic foraminiferal tests that are used to infer past climatic variations include total foraminiferal number, species abundance, assemblages, test size/diameter, coiling direction, average number of chambers, proloculus size, test abnormalities, stable isotopic composition of the tests, trace metal composition of the test, etc. (Boltovskoy and Wright, 1976; Murray, 2006; Sen Gupta, 1999).

Use of presence-absence and indicator taxa is the simplest approach and has been developed extensively for industrial (petroleum exploration) applications. However, taxa respond to a spectrum of environmental factors, so that it is often difficult to confidently determine the nature of environmental changes responsible for observed events in the fossil record. Laboratory culture studies can help elucidate specific responses to particular parameters, which can then be applied to fossils (see Linshy et al., 2007 for review; Nigam et al., 2008; Saraswat et al., 2011). Assemblage analysis is more time consuming but can be more revealing of unique biofacies. With careful organization of a modern data set, it is possible to make quantitative calibrations of seabed fossil assemblages to important environmental variables (see Loubere and Qian, 1997, for theory). Uniformitarian assumptions then allow for the development of 'transfer functions' with which paleoenvironmental parameters may be at least semiquantitatively estimated from fossil assemblage composition. Abnormalities in the tests in response to anthropogenic pollution have been reported, both from the field as well as in controlled laboratory culture studies (see Nigam et al., 2006 for review; Nigam et al., 2009a; Frontalini and Coccioni, 2011). The stable isotope record of benthic Foraminifera is one of the primary correlations and age dating tools in the Quaternary, particularly oxygen isotopes (Lisiecki and Raymo, 2005). The primary driver of oxygen isotope shifts in calcareous tests is changing ice volume at the poles, with a secondary imprint from changes in local temperature, precipitation, evaporation, and runoff. Carbon isotope signals are more complex, incorporating shifts in surrounding water mass, global transfers of carbon isotopes between ocean and continents, and microhabitat selection for infaunal taxa (Schumacher et al., 2010). Trace elements in benthic Foraminifera serve as recorders of global ocean circulation and chemical state of the ocean (Bryan and Marchitto, 2008; Lear et al., 2008; Toyofuku et al., 2011). They also have the potential for recording local environmental geochemical conditions.

Applications

Intermediate Water Oxygenation

Benthic Foraminifera live both on the surface of sediments (epifaunal) as well as several centimeters below (infaunal).

The infaunal forms, in general, can tolerate low dissolved oxygen conditions. This species-specific response of benthic Foraminifera to variation in dissolved oxygen concentration in the ambient seawater (Bernhard and Sen Gupta, 1999; Gooday et al., 2010) has been used to reconstruct past variation in shelf water oxygenation (den Dulk et al., 2000; Nigam et al., 2009b). The high relative abundance of miliolids has been suggested as an indicator of periods of increased ventilation in the northern Arabian Sea, whereas deep-infaunal species indicate oxygen-deficient waters. The Quaternary changes in the intensity and extent of the oxygen minimum zone in the northern Arabian Sea, inferred from benthic foraminiferal assemblages, have been correlated with a change in organic matter flux in response to monsoon-controlled productivity changes (den Dulk et al., 2000). The development of a seasonal shallow water hypoxic zone off the west coast of India has been traced to preanthropogenic time, based on increased abundance of rectilinear benthic Foraminifera in low oxygen waters (Nigam et al., 2009b).

Surface Productivity

Benthic foraminiferal assemblages preserved in sediments are strongly affected by the supply of labile organic matter to the environment (reviewed in Gooday, 2003). This is particularly true for the deep ocean, where other environmental factors are more uniform. Above the depth at which thermodynamic dissolution (due to bottom water calcite undersaturation) begins to strongly affect foraminiferal preservation, it can be argued that organic carbon flux, and its seasonal variation, is the primary determinant on assemblages (Corliss et al., 2009; Soma and Gupta, 2010). Organic carbon flux to the seabed in the open ocean is a function of new biological production in the upper ocean. This new production depends on upper ocean nutrient distribution and on patterns of wind-induced upwelling. Thus, reconstructing biogenic flux to the seafloor and upper ocean productivity provides windows to past oceanic nutrient distributions, circulation, and wind fields. These are important variables in the control of atmospheric CO₂ content and in the modeling of global climate change mechanisms.

Analysis of a large suite of surface sediment fossil assemblages in a data set organized to minimize organic carbon flux correlation with other potentially significant ecological variables has made it possible to construct a transfer function for upper ocean productivity (Loubere and Fariduddin, 1999b). In higher productivity settings, infaunal taxa are abundant, whereas in the oligotrophic domain, the assemblages are dominated by epifaunal and opportunistic forms. Down-core assemblages can thus be used to estimate past oceanic productivity.

The benthic foraminiferal productivity transfer function has been used extensively in the eastern equatorial Pacific since this region is globally important to the marine new production budget and also to the exchange of CO₂ between ocean and atmosphere. Application of the transfer function in a regional network of cores, and its comparison to other paleoceanic tracers, has allowed reconstruction of upper ocean nutrient concentration gradients and patterns of biological production as they were during the late Pleistocene glacial intervals (Loubere et al., 2003). This work has demonstrated important changes in nutrients and productivity over glacial–interglacial

cycles that would have affected oceanic transfer of CO₂ to the atmosphere, and hence, the atmospheric CO₂ cycles recorded in ice cores.

High primary productivity off South-West Africa during Heinrich 2 (25.4–24.3 ka BP) and 1 (16.8–15.8 ka BP) was inferred from increased abundance of infaunal and deep-infaunal benthic foraminiferal specimens, along with high benthic Foraminifera accumulation rates. At present, this region is marked by highly seasonal primary productivity (Zarriess and Mackensen, 2010). A distinct increase in the abundance of opportunistic species feeding on phytodetritus (*Epistominella exigua*) or adapted to elevated but intermittent organic carbon fluxes (*Bulimina aculeata* and *B. marginata*) has been inferred to indicate episodes of recurrent phytoplankton bloom at the end of marine isotopic stage (MIS) 13 and during the early and late parts of MIS 11 in the Blake Outer Ridge, whereas increased abundance of infaunal taxa (*Uvigerina* spp. and *B. alazanensis*) during glacial MIS 12 and 10 indicates sustained organic carbon delivery and diminished pore water oxygenation (Poli et al., 2010).

Monsoon

High turbulence near the mouths of rivers, in response to freshwater influx from the rivers, favors rounded symmetrical benthic Foraminifera (Nigam et al., 1992). Additionally, the morphology of selected benthic Foraminifera in the shallow water regions close to river mouth varies with changes in freshwater influx from the rivers (Nigam and Khare, 1995). These characteristics have been used to infer cyclic changes in the monsoon during the late Quaternary (Nigam et al., 1995). Large salinity variations in shallow water regions also favor characteristic benthic foraminiferal assemblages. *Ammonia beccarii*, along with a number of other calcareous taxa (*Haynesina* sp., *Elphidium* spp., *Quinqueloculina* spp., and *Rotalinoides annectens*) and agglutinated taxa (*Haplophragmoides* sp. and *Ammobaculites formosensis*), dominate a large proportion of the Pearl River estuary, southeast China Sea, and a gradual increase in their relative abundance has been used to infer declining fluvial influx since the mid-Holocene (Garrett, 2010). The establishment of the modern alternating northeast and southwest monsoon system in the Indian subcontinent at 2.8 Ma and its prevalence throughout the Quaternary was inferred on the basis of increased abundance of benthic foraminiferal taxa indicating strongly seasonal pulsed organic matter flux (Gupta and Thomas, 2003).

High Latitude Glaciations

In the middle to high latitudes, the pattern of repeating glacial–interglacial cycles through the Quaternary has left a clear signal in the marine sedimentary archive. The sensitivity of restricted exchange environments, such as fjords, to climatic and environmental change is increasingly recognized. These glacially overdeepened and often sheltered coastal environments preserve exceptionally well-resolved sedimentary records close to the land–ocean interface. The associated benthic foraminiferal assemblage changes preserved in these sediments provide an excellent record of paleoenvironments, from the deep ocean to the shallow subtidal zone. In the high latitude coastal and shelf

sea environments, where salinities are often reduced and planktonic Foraminifera are absent, benthic Foraminifera represent an extremely valuable microfossil group. In recent years, a number of very detailed paleoclimatic records, based on benthic Foraminifera, have emerged from these environments, many highlighting the nature and instability of Holocene paleoclimates (e.g., Husum and Hald, 2004).

Feyling-Hanssen (1964) pioneered the use of benthic Foraminifera in the reconstruction of late Quaternary paleoenvironments in deposits from the Oslofjord area of southern Norway, employing what was termed an *ecostratigraphic* approach. The presence of distinct benthic foraminiferal biofacies, including that of the ice edge, has also been inferred (Ishman and Szymeczek, 2003). Recently, Majewski (2010) summarized benthic foraminiferal distributions from the west Antarctic fjord environments and their paleoecological implications. Increased relative abundance of the benthic foraminiferal species *Cassidulinina neoteretis* from the Bølling/Allerød through the Younger Dryas suggests that Atlantic Intermediate Water entered the Kangerlussuaq Trough during the Allerød and the Younger Dryas via the Irminger Current and further that advection of Atlantic Intermediate Water into the Kangerlussuaq Trough, south of the Denmark Strait, promoted melting and retreat of

the marine portions of the Greenland ice sheet (Jennings et al., 2006). A high-resolution record of glacial retreat and Holocene climate change in the Weddell Sea side of the Antarctic Peninsula has been generated on the basis of benthic foraminiferal assemblages. It is inferred that the most glacier-proximal conditions lasted until ~8800 years BP, as indicated by *Criboelphidium* cf. *webbi*. A marked increase in *Paratrochammina bartrami* and *Paratrochammina antarctica*, the major constituents of the *P. bartrami*–*P. antarctica* assemblage after ~3500 years BP, indicates the onset of the neoglaciation interval (Majewski and Anderson, 2009).

Sea Level

The environmental gradients of intertidal settings are often extreme and include large changes in salinity, subaerial exposure, and marked changes in pH. For this reason, intertidal environments often exhibit a marked intertidal zonation pattern and benthic Foraminifera. In particular, the cosmopolitan agglutinated species *Trochammina inflata*, *Miliammina fusca*, and *Jadammina macrescens* appear to be universal in their occurrence and vertical zonation (Figure 4).

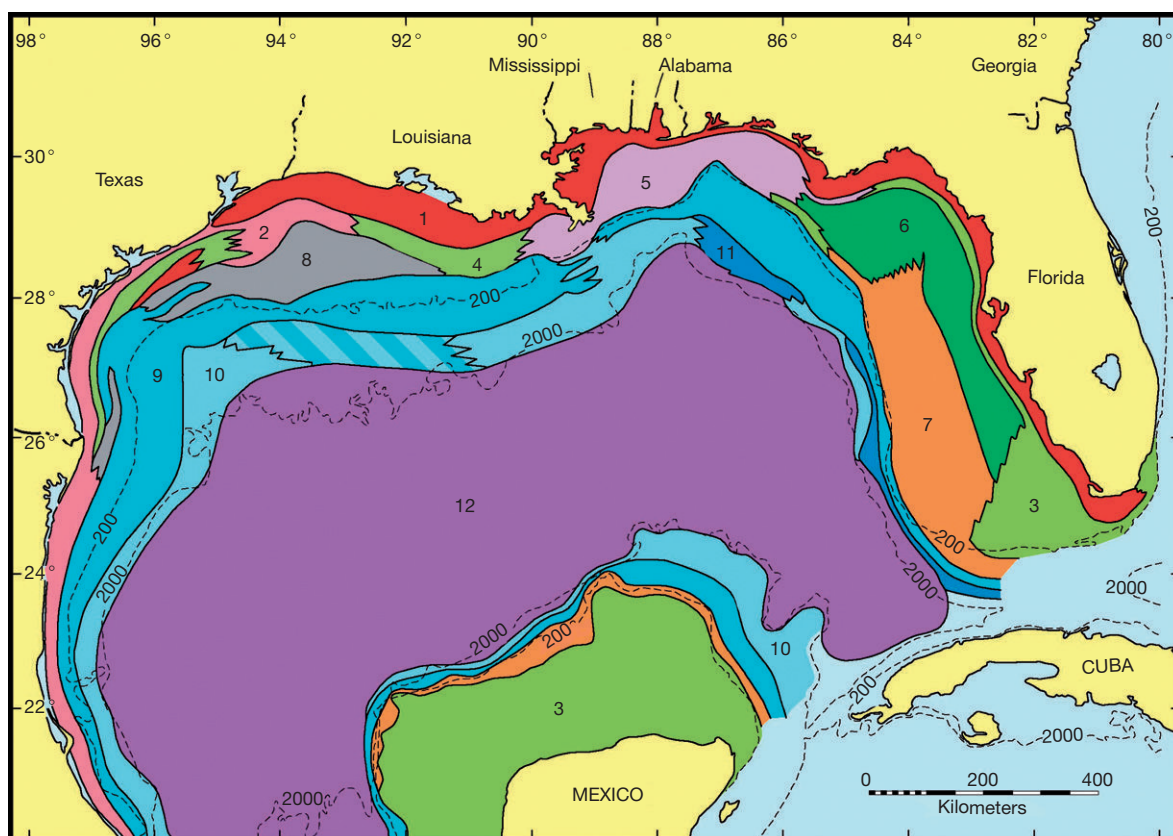


Figure 4 Zonation of benthic foraminiferal taxa down the continental slope of the Gulf of Mexico. These regions can be distinguished by the assemblage of Foraminifera found in the sediments. Notice how the distributional pattern is clearly influenced by water depth and regional substrate (e.g., Mississippi delta and Florida/Yucatan carbonate platforms). Zonation named taxa: 1 = *Ammonia*, 2 = *Elphidium*, 3 = *Miliolid*, 4 = *Saccammina*–*Ammobaculites*, 5 = *Rosalina*–*Hanzawaia*–*Islandiella*, 6 = *Asterigerina*–*Soritid*–*Archaias*, 7 = *Planulina*, 8 = *Elphidium*–*Hanzawaia*–*Bigenerina*, 9 = *Brizalina*, 10 = *Bulimina*, 11 = *Islandiella*, 12 = *Eponides*–*Nuttallides*. Redrawn from Poag W (1981) *Ecologic Atlas of Benthic Foraminifera of the Gulf of Mexico*, 174 p. Woods Hole, MA: Marine Science International, with permission.

Scott and Medioli (1978) provided an elegant illustration of the application of benthic foraminiferal intertidal zonation to the reconstruction of past sea level in a study of salt-marsh deposits in Nova Scotia, Canada. They argued that past sea-level positions could be reconstructed from sediment cores with a precision of ± 0.05 m. In particular, the use of *Trochammina macrescens*, which is the highest occurring salt-marsh Foraminifera living on the salt-marsh surfaces of New England and the Canadian Maritimes, is arguably the most accurate sea-level indicator in the world (Figure 5; Gehrels, 2002). In sediment records, particularly those overlying hard, difficult to compact substrates, the presence of species such as *T. macrescens*, which occurs in a narrow zone near the highest astronomical tide level, provides a very useful indicator of former sea level. When dated, these salt-marsh deposits provide so-called sea-level index points, and these may be combined to provide a record of changing regional sea levels (see Gehrels, 2002).

In formerly glaciated regions, where isostatic uplift of the Earth's crust is active, salt-marsh deposits are rare, and isolation basins often provide the best available evidence of former sea levels. In such settings, basins with well-defined rock sills can become isolated from the sea during uplift, so that they preserve a sedimentary record of the marine to freshwater transition. This transition, once dated, provides a sea-level index point. Numerous basins within a region, with rock sills at differing altitudes, provide a series of sea-level index points. A detailed study by Lloyd (2000) of Loch nan Corr, an isolation basin in north west Scotland (a region heavily glaciated during the Last Glacial Maximum (LGM)), reveals a total of eight foraminiferal biozones, with the lower zones rich in calcareous estuarine and shelf species, dominated by *Haynesina germanica*. As the Loch nan Corr basin went through the isolation process, the foraminiferal assemblages recorded a steady decrease in the number of calcareous taxa and a gradual increase in agglutinated taxa. Initially, *J. macrescens* dominated, giving way to *M. fusca*, until freshwater conditions are finally indicated by the sudden appearance of testate amoebae. Sea level also plays an important control in many shelf sea processes, exerting a particularly strong influence on the position of shelf sea fronts that separate the seasonally stratified from well-mixed water columns of the middle and high latitudes. Typically, as sea levels rise following deglaciation, continental margins are flooded and the locations of fronts migrate landward across the shelf. Austin and Scourse (1997) used benthic foraminiferal assemblage and stable isotope data from a core in the central Celtic Sea to test the temporal evolution from a mixed to stratified water column during the Holocene transgression of the European shelf.

Benthic foraminiferal assemblages show marked faunal changes as 'high-energy' epifaunal species, such as *Cibicides lobatulus*, give way to 'low-energy' infaunal species, such as *B. marginata*, during sea-level rise. Since the mixed water column situation will translate summer heating throughout the depth of the water column, summer bottom water temperatures, when Foraminifera are growing and calcifying their tests, will be relatively high. However, once sea level has risen sufficiently to enable the water column to stratify, summer heating of the surface ocean is no longer mixed throughout the depth of the water column, and summer bottom water temperatures remain low. As well as the benthic assemblage changes, Austin and Scourse (1997) used stable isotopes recorded in two benthic species, *Ammonia batavus*

and *Quinqueloculina seminulum*, to infer a mid-Holocene cooling of 4–5 °C in bottom water temperatures as stratification of the central Celtic Sea was established, following sea-level rise.

Thermohaline Circulation

The continuous inter-basin exchange of seawater due to the difference in its temperature and salinity in different oceans is termed as thermohaline circulation. It includes formation of deepwater in the northern Atlantic and around Antarctica, distribution of this deepwater to all major oceans as deepwater, bottom water or intermediate water, and a subsequent movement of warm water from the Pacific and the Indian Ocean, back to northern Atlantic as surface currents.

Changes in deepwater production in the northern Atlantic and a subsequent change in ventilation of deep, intermediate, and bottom waters are a part of thermohaline circulation changes during glacial–interglacial time periods as well as long-term adjustments in thermohaline circulation. Such periodic adjustments of thermohaline circulation alter physicochemical characteristics of intermediate, deep, and bottom water, which in turn affect benthic Foraminifera. Therefore, benthic Foraminifera are ideal to track changes in thermohaline circulation. Extinction and decreased abundance of selected benthic foraminiferal species are often used to track changes in thermohaline circulation. In addition, stable oxygen and carbon isotopic ratios of selected benthic Foraminifera have also been used to infer past changes in thermohaline circulation. A decrease in abundance and disappearance of elongate, bathyal–upper abyssal benthic Foraminifera (Extinction Group = Stilosomellidae, Pleurostomellidae, some Nodosariidae) in the northern Indian and southwest Pacific Ocean at the time of onset of the mid-Pleistocene transition was attributed to the increased production of cold, well-ventilated Glacial North Atlantic Intermediate Water (Hayward, 2002; Kawagata et al., 2006). The dominance of epifaunal species within the pre-S5 and post-S5 sediments, combined with the low abundance and dominance of infaunal taxa at the base and top of S5, has been ascribed to changes in state of anoxia resulting from variation in ventilation of the deep waters (Nolet and Corliss, 1990). Differences in the abundance of oxygen-sensitive and dissolution-prone benthic foraminiferal species between the LGM and the Holocene in the abyssal waters of the southwestern Gulf of Mexico were used to infer changes in the production of glacial North Atlantic deep water (NADW) during LGM and NADW during the Holocene (Machain-Castillo et al., 2010).

Storm/Tsunami

Long-term past records of storms/tsunami events can help us understand recurrence intervals of these events. The storms/tsunamis leave behind evidence in several forms, especially the material transported from its natural place of occurrence to an unlikely location. Storms/tsunamis can erode sediments from deeper waters, transport them to shallow regions against the normal gradient, and redeposit them in shallower regions. Even though it is possible to determine anomalous origin of the sediments, based on the sediment texture and composition, it is difficult to ascertain their origin from the landward or seaward side of their occurrence. Benthic Foraminifera which have characteristic depth zonations of different species can

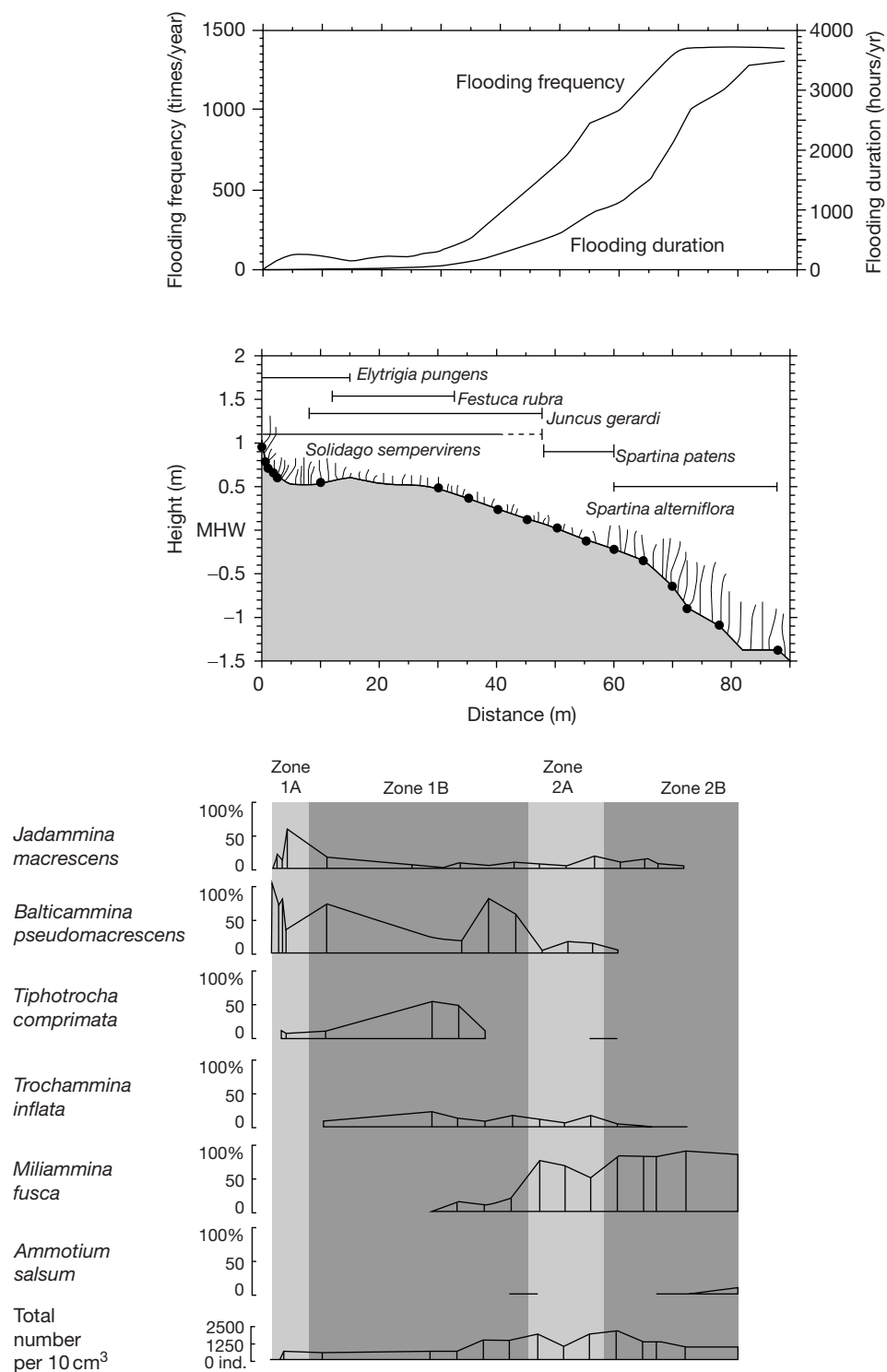


Figure 5 The foraminiferal zonation at Sanborn Cove marsh, Machiasport, Maine, in relation to the zonation of plants and the frequency and duration of tidal flooding. The foraminiferal zones (1A, 1B, 2A, 2B) correspond to the zones from Nova Scotian marshes described by Scott and Medioli (1978), except for the fact that *Jadammina macrescens* and *Balticammina pseudomacrescens* are grouped as *Trochammina macrescens* in Nova Scotia. MHW – mean high water (1.97 m above sea level). Reproduced from Gehrels WR (2002) Intertidal foraminifera as palaeoenvironmental indicators. In: Haslett SK (ed.) *Quaternary Environmental Micropalaeontology*, pp. 91–114. London: Arnold.

help determine the direction of transport of sediments. Benthic Foraminifera, being very small and light in weight, also get transported along with the sediments during such events. Benthic Foraminifera show a marked zonation in shallow waters, with coastal assemblages being significantly different than deep-water assemblages. Thus, the transport of sediments against gravity by extreme events like storms or tsunamis can be inferred on the basis of benthic Foraminifera (Mamo et al., 2009).

Making use of modern storm traces (offshore benthic Foraminifera in marsh sediments), intense storm activities (six storms) were inferred between 0 and 1000 years BP interval along the South Carolina coast. Three additional storm events were identified during the 1000–2000 yr BP interval and two between the 2000 and 3000 yr BP interval (Scott et al., 2003). The presence of ca. 12 000–10 000-year-BP-old sediments above ca. 8000-year-BP-old sediments in cores collected from the eastern Arabian Sea was attributed to ancient storm–tsunami events, based on characteristic benthic foraminiferal assemblages (Nigam and Chaturvedi, 2006). Here, inverted sediments were found sandwiched between typical marine sedimentary sequences. The top of the inverted section was older than the top of the typical sequence just below it. This inverted sequence contained benthic Foraminifera that are found in comparatively deeper waters than their present water depth, which ruled out the possibility of transport of this inverted sequence from the landward side of its present location. The older age and deeper benthic foraminiferal assemblage in the inverted sequence were inferred to have been caused by transport of these sediments from a deeper location by a storm or tsunami. A second similar storm–tsunami-deposited a foram assemblage that was identified from ca. 10 000-year-old sediments over a normally deposited, ca. 7000-year-old sequence in the same region. Schafer and Medioli (2009) inferred erosion and transport of sediments from the adjacent inner shelf by an onshore-directed storm/hurricane event based on the presence of benthic and planktic foraminiferal tests in a core collected from the landward side of a baymouth barrier in Chezzetcook, Nova Scotia, Canada.

Marine Archaeology

Calcareous benthic Foraminifera are exclusively marine. Therefore, the presence of autochthonous fossil calcareous benthic foraminiferal shells can be considered proof of marine environments. This has been used to settle a long-standing controversy regarding the usage of a large rectangular tank of Harappan (Indus Valley Civilization) times, excavated at a location far away from the present coast. Large triangular stones with a hole were recovered from the tank. It was long debated whether the stones were used as anchors, suggesting the structure was part of a dockyard for small ships, or whether the stones were used as deadweight to draw water out of the freshwater tank. The presence of fossil benthic foraminiferal shells, characteristic of shallow water, in the large rectangular structure confirmed it to be the dockyard for small ships (Nigam, 1988). A distinct difference in benthic foraminiferal assemblage characteristic of salt marsh and lagoon was used to reconstruct the evolution and subsequent fifth-century human settlement of San Francesco del Deserto Island in the Lagoon of Venice (Barbero et al., 1997).

Pollution Monitoring

Benthic Foraminifera live in all marine environments, from the intertidal regions to the deepest trenches. In shallow water coastal habitats, industrial waste effluents and domestic sewage outfall affect benthic Foraminifera. The pollutants affect the normal life cycle of benthic Foraminifera, resulting in highly deformed and stunted shells. The shells from polluted regions often also lack ornamentation (Bhalla and Nigam, 1986; Geslin et al., 2002; Scott et al., 2001; see Nigam et al., 2006 for review). Anthropogenic pollutants not only affect the morphology of benthic Foraminifera but also their diversity and abundance (Arieli et al., 2011; Frontalini and Coccioni, 2008, 2011). This distinct benthic foraminiferal response to anthropogenic pollution has often been used to reconstruct historical changes in pollution levels (Scott et al., 2005; see Nigam et al., 2006 for review). A gradual increase in pollutants over a period of time will affect the benthic foraminiferal population in the vicinity with increasing severity. Thus, temporal changes in the number of deformed or stunted specimens can be used to infer changes in the concentrations of pollutants. The change in benthic foraminiferal abundance has also been used to delineate shifts in the level of mining activity in monsoon-dominated tropical regions, as huge precipitation and surface runoff during monsoon season carries mining waste to the estuaries, leading to increased total suspended matter and thus turbidity, which affects benthic Foraminifera (Panchang et al., 2005).

See also: **Glacial Climates:** Thermohaline Circulation. **Glacial Landforms:** Evidence of Glacier Recession. **Introduction:** Understanding Quaternary Climatic Change. **Paleoceanography, Physical and Chemical Proxies:** Carbon Cycle Proxies ($\delta^{11}\text{B}$, $\delta^{13}\text{C}_{\text{calcite}}$, $\delta^{13}\text{C}_{\text{organic}}$, Shell Weights, B/Ca, U/Ca, Zn/Ca, Ba/Ca); Dissolution of Deep-Sea Carbonates; Nutrient Proxies; Oxygen Isotope Composition of Seawater; Oxygen-Isotope Stratigraphy of the Oceans. **Paleoclimate:** Paleoclimate History of the Arctic. **Paleoclimate Modeling:** Quaternary Environments. **Quaternary Stratigraphy:** Continental Biostratigraphy. **Sea Level Studies:** Microfossil-Based Reconstructions of Holocene Relative Sea-Level Change.

References

- Arieli RN, Almogi-Labin A, Abramovich S, and Herut B (2011) The effect of thermal pollution on benthic foraminiferal assemblages in the Mediterranean shoreface adjacent to Hadera power plant (Israel). *Marine Pollution Bulletin* 62: 1002–1012.
- Austin WEN and Scourse JD (1997) Evolution of seasonal stratification in the Celtic Sea during the Holocene. *Journal of the Geological Society* 154: 249–256.
- Barbero RS, Albani AD, and Zecchetto S (1997) Palaeoenvironmental significance of a benthic foraminiferal fauna from an archaeological excavation in the Lagoon of Venice, Italy. *Palaeogeography, Palaeoclimatology, Palaeoecology* 136: 41–52.
- Berkeley A, Perry CT, and Smithers SG (2009) Taphonomic signatures and patterns of test degradation on tropical, intertidal benthic foraminifera. *Marine Micropaleontology* 73: 148–163.
- Bernhard J (1996) Microaerophilic and facultative anaerobic benthic foraminifera: A review of experimental and ultrastructural evidence. *Revue de Paleobiologie* 15: 261–275.
- Bernhard JM and Sen Gupta BK (1999) Foraminifera of oxygen depleted environments. In: Sen Gupta BK (ed.) *Modern Foraminifera*, pp. 201–216. Great Britain: Kluwer.

- Bhalla SN and Nigam R (1986) Recent foraminifera from polluted marine environment of Velsao beach, South Goa, India. *Revue de Paléobiologie* 5: 43–46.
- Boersma A (1980) Foraminifera. In: Haq B and Boersma A (eds.) *Introduction to Marine Micropaleontology*, pp. 19–78. New York: Elsevier North-Holland.
- Boltovskoy E and Wright R (1976) *Recent Foraminifera*. The Hague: W. Junk.
- Bryan SP and Marchitto TM (2008) Mg/Ca – Temperature proxy in benthic foraminifera: New calibrations from the Florida Straits and a hypothesis regarding Mg/Li. *Paleoceanography* 23: PA2220. <http://dx.doi.org/10.1029/2007PA001553>.
- Corliss B (1985) Microhabitats of benthic foraminifera within deep sea sediments. *Nature* 314: 435–438.
- Corliss BH, Brown CW, Sun X, and Showers WJ (2009) Deep-sea benthic diversity linked to seasonality of pelagic productivity. *Deep Sea Research Part I: Oceanographic Research Papers* 56: 835–841.
- Culver S (1988) New foraminiferal depth zonation of the northwestern Gulf of Mexico. *Palaios* 3: 69–85.
- Culver S and Buzas M (1999) Biogeography of neritic benthic foraminifera. In: Sen Gupta B (ed.) *Modern Foraminifera*, pp. 93–102. Dordrecht: Kluwer.
- Den Dulk M, Reichart GJ, Van Heyst S, Zachariasse WJ, and Van der Zwaan GJ (2000) Benthic foraminifera as proxies of organic matter flux and bottom water oxygenation? A case history from the northern Arabian Sea. *Palaeogeography, Palaeoclimatology, Palaeoecology* 161: 337–359.
- Feyling-Hanssen RW (1964) Foraminifera in Late Quaternary deposits from the Oslofjord area. *Norges Geologiske Undersøkelse* 225: 383.
- Frontalini F and Coccioni R (2008) Benthic foraminifera for heavy metal pollution monitoring: A case study from the central Adriatic Sea coast of Italy. *Estuarine, Coastal and Shelf Science* 76: 404–417.
- Frontalini F and Coccioni R (2011) Benthic foraminifera as bioindicators of pollution: A review of Italian research over the last three decades. *Revue de Micropaleontology* 54: 115–127.
- Garrett E (2010) The contemporary distribution of benthic foraminifera in the Pearl River estuary, southeast China, and their use in reconstructing mid to late Holocene fluvial flux. Masters Thesis, Durham University.
- Gehrels WR (2002) Intertidal foraminifera as palaeoenvironmental indicators. In: Haslett SK (ed.) *Quaternary Environmental Micropaleontology*, pp. 91–114. London: Arnold.
- Geslin E, Debenay JP, Duleba W, and Bonetti C (2002) Morphological abnormalities of foraminiferal tests in Brazilian environments: Comparison between polluted and non-polluted areas. *Marine Micropaleontology* 45: 151–168.
- Goldstein S (1999) Foraminifera: A biological overview. In: Sen Gupta B (ed.) *Modern Foraminifera*, pp. 37–56. Dordrecht: Kluwer.
- Gooday AJ (2003) Benthic foraminifera (Protista) as tools in deep-water paleoceanography: Environmental influences on faunal characteristics. *Advances in Marine Biology* 46: 1–90.
- Gooday AJ, Bett BJ, Escobar E, et al. (2010) Habitat heterogeneity and its influence on benthic biodiversity in oxygen minimum zones. *Marine Ecology: An Evolutionary Perspective* 31: 125–147.
- Gupta AK and Thomas E (2003) Initiation of Northern Hemisphere glaciation and strengthening of the northeast Indian monsoon: Ocean Drilling Program Site 758, eastern equatorial Indian Ocean. *Geology* 31: 47–50.
- Hallock P (1999) Symbiont-bearing foraminifera. In: Sen Gupta B (ed.) *Modern Foraminifera*, pp. 123–140. Dordrecht: Kluwer.
- Hayward BW (2002) Late Pliocene to middle Pleistocene extinctions of deep-sea benthic foraminifera (*Stilostomella* Extinction) in the Southwest Pacific. *Journal of Foraminiferal Research* 32: 274–307.
- Husum K and Hald M (2004) A continuous marine record 8000–1600 cal. yr BP from the Malangenfjord, north Norway: Foraminiferal and isotopic evidence. *The Holocene* 14: 877–887.
- Ishman SE and Szymek P (2003) Foraminiferal distributions in the former Larsen-A Ice Shelf and Prince Gustav Channel region, eastern Antarctic Peninsula margin: A baseline for Holocene paleoenvironmental interpretation. *Antarctic Research Series* 79: 239–260.
- Jennings AE, Hald M, Smith M, and Andrews JT (2006) Freshwater forcing from the Greenland Ice Sheet during the Younger Dryas: Evidence from southeastern Greenland shelf cores. *Quaternary Science Reviews* 25: 282–298.
- Jorissen F (1999) Benthic foraminiferal microhabitats below the sediment–water interface. In: Sen Gupta B (ed.) *Modern Foraminifera*, pp. 161–180. Dordrecht: Kluwer.
- Jorissen FJ, Fontanier C, and Thomas E (2007) Paleoceanographical proxies based on deep-sea benthic foraminiferal assemblage characteristics. In: Hillaire-Marcel C and de Vernal A (eds.) *Proxies in Late Cenozoic Paleoceanography. Biological Tracers and Biomarkers* Part 2: pp. 263–326. Amsterdam: Elsevier.
- Kawagata S, Hayward BW, and Gupta AK (2006) Benthic foraminiferal extinctions linked to late Pliocene–Pleistocene deep-sea circulation changes in the northern Indian Ocean (ODP Sites 722 and 758). *Marine Micropaleontology* 58: 219–242.
- Lear CH, Bailey TR, Pearson PN, Coxall HK, and Rosenthal Y (2008) Cooling and ice growth across the Eocene–Oligocene transition. *Geology* 36: 251–254.
- Linshy VN, Rana SS, Kurtarkar SR, Saraswat R, and Nigam R (2007) Appraisal of laboratory culture experiments on benthic foraminifera to assess/develop paleoceanographic proxies. *Indian Journal of Marine Sciences* 36: 301–321.
- Lisiecki LE and Raymo ME (2005) A Pliocene–Pleistocene stack of 57 globally distributed benthic $\delta^{18}\text{O}$ records. *Paleoceanography* 20: PA1003. <http://dx.doi.org/10.1029/2004PA001071>.
- Lloyd JM (2000) Combined foraminiferal and thecamoebian environmental reconstruction from an isolation basin in NW Scotland: Implications for sea-level studies. *Journal of Foraminiferal Research* 30: 294–305.
- Loeblich A and Tappan H (1987) *Foraminiferal Genera and their Classification*, vol. 1–2. New York: Van Nostrand Reinhold.
- Loubere P and Fariduddin M (1999a) Benthic foraminifera and the flux of organic carbon to the seabed. In: Sen Gupta B (ed.) *Modern Foraminifera*, pp. 181–200. Dordrecht: Kluwer.
- Loubere P and Fariduddin M (1999b) Quantitative estimation of global patterns of surface ocean biological productivity and its seasonal variation on time scales from centuries to millennia. *Global Biogeochemical Cycles* 13: 115–133.
- Loubere P, Fariduddin M, and Murray R (2003) Patterns of export production in the eastern equatorial Pacific over the past 130,000 years. *Paleoceanography* 18: 1028. <http://dx.doi.org/10.1029/2001PA000658> (p. 6–1 to 6–21).
- Loubere P, Meyers P, and Gary A (1995) Benthic foraminiferal microhabitat selection, carbon isotope values, and association with larger animals: A test with *Uvigerina peregrina*. *Journal of Foraminiferal Research* 25: 83–95.
- Loubere P and Qian H (1997) Reconstructing paleoecology and paleoenvironmental variables using factor analysis and regression: Some limitations. *Marine Micropaleontology* 31: 205–217.
- Machain-Castillo ML, Argáez FRG, Castillo LBC, Herrera JAA, and Sen Gupta BK (2010) Last glacial maximum deep water masses in southwestern Gulf of Mexico: Clues from benthic foraminifera. *Boletín de la Sociedad Geológica Mexicana* 62: 453–467.
- Mackensen A and Douglas RG (1989) Down-core distribution of live and dead deep-water benthic foraminifera in box cores from the Weddell Sea and the California continental borderland. *Deep-Sea Research I* 36: 879–900.
- Majewski W (2010) Benthic foraminifera from West Antarctic fiord environments: An overview. *Polish Polar Research* 31: 61–82.
- Majewski W and Anderson JB (2009) Holocene foraminiferal assemblages from Firth of Tay, Antarctic Peninsula: Paleoclimate implications. *Marine Micropaleontology* 73: 135–147.
- Mamo B, Strotz L, and Dominey-Howes D (2009) Tsunami sediments and their foraminiferal assemblages. *Earth-Science Reviews* 96: 263–278.
- Murray J (2006) *Ecology and Applications of Benthic Foraminifera*, p. 426. Cambridge: Cambridge University Press.
- Murray JW and Alve E (1999) Natural dissolution of modern shallow water benthic foraminifera: Taphonomic effects on the paleoecological record. *Palaeogeography, Palaeoclimatology, Palaeoecology* 146: 195–209.
- Nigam R (1988) Was the large rectangular structure at Lothal (Harappan settlement) a 'Dockyard' or an 'Irrigation Tank'? In: Rao SR (ed.) *Proceedings of the Marine Archaeology of Indian Ocean Countries*, pp. 20–21. India: NIO.
- Nigam R and Chaturvedi SK (2006) Do inverted depositional sequences and allochthonous foraminifera in sediments along the Coast of Kachchh, NW India, indicate palaeostorm and/or tsunami effects? *Geo-Marine Letters* 26: 42–50.
- Nigam R and Khare N (1995) Significance of correspondence between river discharge and proloculus size of benthic foraminifera in paleomonsoonal studies. *Geo-Marine Letters* 15: 45–50.
- Nigam R, Khare N, and Borole DV (1992) Can benthic foraminiferal morpho-groups be used as indicators of paleomonsoonal precipitation? *Estuarine, Coastal and Shelf Science* 34: 533–542.
- Nigam R, Khare N, and Nair RR (1995) Foraminiferal evidences for 77-year cycles of droughts in India and its possible modulation by the Gleissberg solar cycle. *Journal of Coastal Research* 11: 1099–1107.
- Nigam R, Kurtarkar SR, Saraswat R, Linshy VN, and Rana SS (2008) Response of benthic foraminifera *Rosalina leei* to different temperature and salinity, under laboratory culture experiment. *Marine Biological Association of the United Kingdom* 88: 699–704.
- Nigam R, Linshy VN, Kurtarkar SR, and Saraswat R (2009a) Effects of sudden stress due to heavy metal mercury on benthic foraminifera *Rosalina leei*: Laboratory culture experiment. *Marine Pollution Bulletin* 59: 362–368.
- Nigam R, Prasad V, Mazumder A, Garg R, Saraswat R, and Henriques PJ (2009b) Late Holocene changes in hypoxia off the west coast of India: Micropaleontological evidences. *Current Science* 96: 708–713.

- Nigam R, Saraswat R, and Panchang R (2006) Application of foraminifers in ecotoxicology: Retrospect, prospect and prospect. *Environmental International* 32: 273–283.
- Nolet GJ and Corliss BH (1990) Benthic foraminiferal evidence for reduced deep-water circulation during sapropel deposition in the eastern Mediterranean. *Marine Geology* 94: 109–130.
- Panchang R, Nigam R, Baig N, and Nayak GN (2005) A foraminiferal testimony for the reduced adverse effects of mining in Zuari Estuary, Goa. *International Journal of Environmental Studies* 62: 579–591.
- Poag W (1981) *Ecologic Atlas of Benthic Foraminifera of the Gulf of Mexico*, 174 p. Woods Hole, MA: Marine Science International.
- Poli MS, Meyers PA, and Thunell RC (2010) The western North Atlantic record of MIS 13 to 10: Changes in primary productivity, organic carbon accumulation and benthic foraminiferal assemblages in sediments from the Blake Outer Ridge (ODP Site 1058). *Palaeogeography, Palaeoclimatology, Palaeoecology* 295: 89–101.
- Saraswat R, Nigam R, and Pachkhane S (2011) Difference in optimum temperature for growth and reproduction in benthic foraminifer *Rosalina globularis*: Implications for paleoclimatic studies. *Journal of Experimental Marine Biology and Ecology* 405: 105–110.
- Schafer CT and Medioli FS (2009) Pilot study of fossil evidence of onshore-directed storm events in estuarine sediments: Chezzetcook Inlet, Nova Scotia. *Canadian Journal of Earth Sciences* 46: 193–205.
- Schumacher S, Jorissen FJ, Mackensen A, Gooday AJ, and Pays O (2010) Ontogenetic effects on stable carbon and oxygen isotopes in tests of live (Rose Bengal stained) benthic foraminifera from the Pakistan continental margin. *Marine Micropaleontology* 76: 92–103.
- Scott DB, Collins E, Gayes P, and Wright E (2003) Records of prehistoric hurricanes on the South Carolina coast based on micropaleontological and sedimentological evidence, with comparison to other Atlantic coast records. *Geological Society of America Bulletin* 115: 1027–1039.
- Scott DB and Medioli FS (1978) Vertical zonation of marsh foraminifera as accurate indicators of former sea-levels. *Nature* 272: 538–541.
- Scott DB, Medioli FS, and Schafer CT (2001) *Monitoring of Coastal Environments using Foraminifera and Thecamoebian Indicators*, p. 176. Cambridge: Cambridge University Press.
- Scott DB, Tobin R, Williamson M, et al. (2005) Pollution monitoring in two North American estuaries: Historical reconstructions using benthic foraminifera. *Journal of Foraminiferal Research* 35: 65–82.
- Sen Gupta B (1999) *Modern Foraminifera*, 371 p. Dordrecht: Kluwer.
- De Soma and Gupta AK (2010) Deep-sea faunal provinces and their inferred environments in the Indian Ocean based on distribution of recent benthic foraminifera. *Palaeogeography, Palaeoclimatology, Palaeoecology* 291: 429–442.
- Sun X, Corliss BH, Brown CW, and Showers WJ (2006) The effect of primary productivity and seasonality on the distribution of deep-sea benthic foraminifera in the North Atlantic. *Deep Sea Research Part I: Oceanographic Research Papers* 53: 28–47.
- Toyofuku T, Suzuki M, Suga H, et al. (2011) Mg/Ca and $\delta^{18}\text{O}$ in the brackish shallow-water benthic foraminifer *Ammonia 'beccarii'*. *Marine Micropaleontology* 78: 113–120.
- Zarriess M and Mackensen A (2010) The tropical rainbelt and productivity changes off northwest Africa: A 31,000-year high-resolution record. *Marine Micropaleontology* 76: 76–91.

Biomarker Indicators of Past Climate

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Introduction

Molecular fossils, 'biomarkers,' add to the growing arsenal of climate indicators scientists are using to reconstruct climate history and understand what causes climate to change. Advances in analytical chemistry and instrumentation over the last two decades have expanded the analytical window in which organic geochemists can routinely work. Precise and highly sensitive mass spectrometers that can determine molecular structures and isotope ratios are now coupled to gas and liquid chromatographs that can separate the complex mixtures of organic constituents found in geologic samples into pure components.

As more geoscientists become familiar with these new techniques, there are likely to be tremendous advances in organic geochemical climate proxies. This article reviews some of the more promising and novel biomarker proxies now being used in paleoclimate research. The article is organized around the climate parameters of most interest to paleoceanographers and paleoclimatologists: temperature, salinity, and sea ice.

Biomarker Proxies of Temperature

The reconstruction of ocean and lake surface temperatures is central to understanding past climatic changes. Impressive advances have been made in organic geochemical approaches to temperature reconstructions. This article focuses on the two most promising and widely applied organic geochemical water temperature proxies: U^k_{37} and TEX_{86} . A relatively new organic geochemical indicator of air temperature known as cyclization ratio of branched tetraethers/methylation index of branched tetraethers (CBT/MBT) is also briefly discussed.

Alkenone Paleothermometry

The discovery of temperature-controlled unsaturation patterns in long-chain ketones and alkenones (nC_{37} – nC_{39} methyl and ethyl ketones) in the mid-1980s (Brassell et al., 1986), and their identification as biomolecules unique to prymnesiophyte algae, principally the coccolithophorid *Emiliania huxleyi* (Marlowe et al., 1984; Volkman et al., 1980), was a watershed event in paleoclimate research. For the first time paleoceanographers had a tool for reconstructing sea surface temperatures (SST) that was unambiguously derived from organisms living in the uppermost part of the ocean where sunlight can penetrate, that was not influenced by the chemical composition of seawater (as, for instance, oxygen isotope ratios in planktonic Foraminifera are), and that was not corrupted by the chemical composition of ocean bottom water. Thus, a sizeable advantage of the

alkenone paleotemperature technique over other temperature proxies, such as Mg/Ca ratios in planktonic Foraminifera or faunal (Foraminifera, radiolarians, diatoms) assemblage changes, is the fact that alkenones are preserved in the sediments even after the dissolution of the calcareous remains of their producers. The technique can, therefore, be applied in regions where other techniques would fail (e.g., the Southern Ocean).

The original definition of the unsaturation ratio (U^k_{37} index) included the tetraunsaturated C_{37} alkenone (Brassell et al., 1986), which was later omitted due to its insignificant contribution ($U^k_{37} = C_{37:2}/(C_{37:2} + C_{37:3})$). Prahl et al. (1988) were the first to publish a quantitative calibration of the U^k_{37} index to growth temperature ($U^k_{37} = 0.034 T + 0.039$, $r^2 = 0.994$) based on cultured *E. huxleyi*, and initial field measurements showed that this calibration reproduced SSTs in the northeast Pacific Ocean (Prahl and Wakeham, 1987). This calibration has since been widely used for the reconstruction of past SSTs in most parts of the world's oceans (e.g., Bard et al., 1997; Herbert et al., 2001; Koutavas and Sachs, 2008; Pahnke and Sachs, 2006; Sachs and Lehman, 1999). Extensive alkenone analyses on a global set of core-top samples and correlation of the U^k_{37} index with overlying mean annual SSTs ($U^k_{37} = 0.033 T + 0.069$, $r^2 = 0.981$) have further confirmed the culture-derived relation of Prahl et al. (1988) (Müller et al., 1998) and demonstrated its applicability to most oceanic areas (Figure 1). Though slight differences exist in the U^k_{37} -temperature dependence of different coccolithophorid species and genetic variations (Conte et al., 1998), past changes in species composition do not appear to compromise paleo-SST reconstructions (Müller et al., 1997).

Most studies, including the core-top U^k_{37} -SST calibration of Müller et al. (1998), assume that U^k_{37} values reflect temperatures at the sea surface. However, alkenone concentrations in sediment traps indicate that the depth of maximum production varies regionally and can even change between different seasons. Alkenone-producing haptophytes show highest abundance in the surface mixed layer in the North Atlantic (Conte et al., 2001) and in the northwest Pacific (Hamanaka et al., 2000). Similarly, Ohkouchi et al. (1999) concluded from a sediment transect in the Pacific that alkenones are produced in the surface mixed layer at high and low latitudes, while thermocline production prevails at Mid-Latitudes. In the gyre region of the subtropical North Pacific, haptophyte growth is linked to the depth of the chlorophyll maximum, which can vary in depth throughout the year (Prahl et al., 2005). Alkenone production shows a strong seasonal cycle in most areas (Conte et al., 2001; Leduc et al., 2010; Prahl et al., 2005), but the unsaturation ratio preserved in the sediments predominantly reflects mean annual temperatures (Müller and Fischer, 2001; Prahl et al., 1993, 2005). In contrast, a strong summer

to fall bias in U^k_{37} -SST estimates is observed in cold, high-latitude regions of the ocean (Prah et al., 2010; Sikes and Volkman, 1993; Sikes et al., 2005; Ternois et al., 1998), suggesting that in some areas, regional U^k_{37} -SST calibrations better reproduce SSTs.

Alkenones have been found in sediments as old as the early Cretaceous (Brassell et al., 2004) and they have been used to reconstruct SSTs on time scales ranging from 10^3 – 10^6 years.

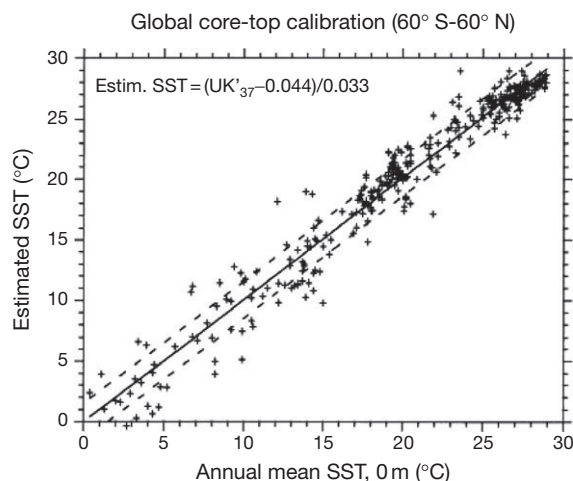


Figure 1 Global core-top calibration of U^k_{37} -derived SST versus overlying mean annual SST from Müller et al. (1998) ($SST = U^k_{37} - 0.044 / 0.033$, $r^2 = 0.958$).

Despite substantial alkenone loss in both the water column and the sediment after deposition, several studies indicate stability of the U^k_{37} -index over geologic time and, therefore, no effect on paleo-SST reconstructions (Conte et al., 1992; Müller and Fischer, 2001; Prah et al., 1989, 1993). However, contrasting studies indicate differential removal of $C_{37:3}$ (Gong and Hollander, 1999; Rontani et al., 2005, 2008), which can have implications for alkenone paleothermometry, depending on the conditions under which degradation occurred (e.g., anoxic, oxic, denitrifying). Other environmental factors potentially affecting the unsaturation ratio of alkenones include light limitation (causing a bias towards warmer SSTs) and nutrient limitation (bias towards lower SSTs; Prah et al., 2003).

The TEX_{86} and MBT/CBT Paleotemperature Proxies

Over the last decade, two new paleotemperature proxies were developed based upon the relative abundances of various glycerol dialkyl glycerol tetraethers (GDGTs) (Figure 2). Isoprenoidal GDGTs with *sn*-2,3 stereochemistry are produced by at least two of three main groups of archaea, the *Crenarchaeota* and the *Euryarchaeota* (De Rosa and Gambacorta, 1988; Koga and Morii, 2005), that occur ubiquitously in the environment, including in the open ocean (e.g., Karner et al., 2001), lakes (e.g., Powers et al., 2010), and soils (Leininger et al., 2006). *Crenarchaeota* biosynthesize different types of GDGTs containing 0–3 cyclopentane moieties (I–IV; Figure 2) and crenarchaeol (V), which, in addition to four cyclopentane moieties, has a cyclohexane ring (Schouten et al., 2000; Sinninghe Damsté

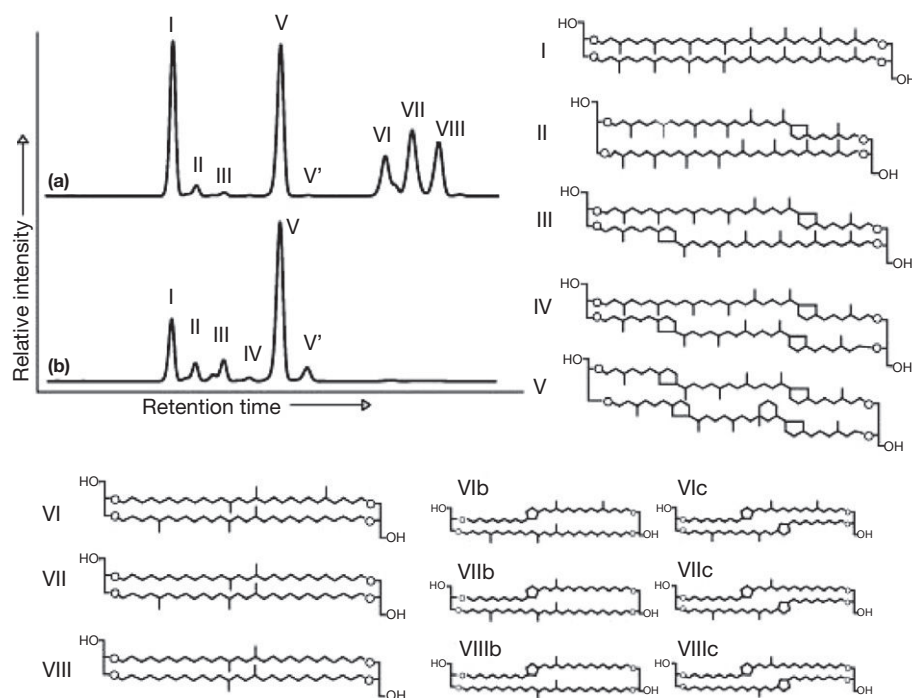


Figure 2 Base peak HPLC-MS chromatograms of GDGTs with the various structures indicated. (a) Sample from the Norwegian Drammensfjord, with a TEX_{86} value of 0.42 and a BIT index value of around 0.6. (b) Sample from the Arabian Sea, with a TEX_{86} value of 0.68 and a BIT index of <0.05. Reproduced from Schouten S, et al. (2009) An interlaboratory study of TEX_{86} and BIT analysis using high-performance liquid chromatography-mass spectrometry. *Geochemistry, Geophysics, Geosystems* 10: Q03012.

et al., 2002). They also synthesize small quantities of a regioisomer of crenarchaeol (V').

Other GDGTs have an *sn*-1,2 stereochemistry and *n*-alkyl chain architecture, and contain varying amounts of methyl branches and cyclopentane moieties (VI–VIII, Figure 2). The exact biological source of these so-called branched GDGTs remains uncertain. Initially attributed to soil bacteria (Herfort et al., 2006; Hopmans et al., 2004; Sinninghe Damsté et al., 2000; Weijers et al., 2006), they can also be produced, as suggested by recent evidence, in sediments or in the water column (Peterse et al., 2009a,b; Sinninghe Damsté et al., 2009; Tierney and Russell, 2009).

Culture studies have shown that the relative number of cyclopentane moieties in the GDGTs from hyperthermophilic archaea increases with growth temperature (Gliozzi et al., 1983) (Uda et al., 2001), which has been interpreted as a means by which archaea adjust their cell membrane rigidity. This dependency on growth temperature was also demonstrated for the nonhyperthermophilic archaea (Wuchter et al., 2004). Thus, by measuring the relative amounts of GDGTs present in marine sediments, the temperature at which *Crenarchaeota* were living when they produced their membranes can be determined. The TEX₈₆ proxy was proposed as a means to quantify the relative abundance of GDGTs (Schouten et al., 2002):

$$\text{TEX}_{86} = \frac{[\text{III}] + [\text{IV}] + [\text{V}]}{[\text{II}] + [\text{III}] + [\text{IV}] + [\text{V}]} \quad [1]$$

The roman numerals refer to structures in Figure 2. The original 44 sediment core-tops from 15 different locations used to establish this new paleotemperature proxy were expanded in the years that followed. Kim et al. (2008) analyzed 287 core-top sediments distributed over the world oceans from different depths, and after exclusion of samples from the Red Sea and the polar oceans, as well as another fifteen outliers, they arrived at the following global temperature calibration:

$$\text{SST} = -10.78 + 56.2 \times \text{TEX}_{86} \quad (r^2 = 0.935, n = 223) \quad [2]$$

Two years later, Kim et al. (2010) expanded this dataset again to 426 core-tops, and proposed a new calibration:

$$\text{SST} = 67.5 \times \text{TEX}_{86}^L + 46.9 \quad (r^2 = 0.86, n = 396, p < 0.0001) \quad [3]$$

with

$$\text{TEX}_{86}^L = \log \left(\frac{[\text{III}]}{[\text{II}] + [\text{III}] + [\text{IV}]} \right) \quad [4]$$

Besides being logarithmic, this new calibration excludes the regioisomer of crenarchaeol (structure V'). It appears better suited to colder temperatures (<15 °C) including those of the polar oceans. For temperatures >15 °C, Kim et al. (2010) proposed a logarithmic calibration based on the original TEX₈₆ proxy:

$$\text{SST} = 68.4 \times \log \text{TEX}_{86} + 38.6 \quad (r^2 = 0.87, n = 255, p < 0.0001) \quad [5]$$

The TEX₈₆ temperature proxy has also been shown to work in large lakes characterized by high *Crenarchaeota* production (Powers et al., 2010). They proposed the following modified TEX₈₆ calibration specifically for lakes:

$$\text{LST} = -14.0 \times \text{TEX}_{86} + 55.2 \quad [6]$$

This calibration varies only slightly from the 'original' marine calibration (eqn [2]), suggesting that salinity does not have a large effect on the TEX₈₆ proxy, a conclusion also reached by Wuchter et al. (2004) based on mesocosm experiments.

The TEX₈₆ SST proxy has been primarily applied in open ocean studies, where the confounding influence of terrestrially sourced GDGTs is minimized. Examples are late Quaternary water temperature reconstructions from the Arabian Sea (Huguet et al., 2006b), the African lakes (Tierney et al., 2008), the Mediterranean Sea (Castañeda et al., 2010), the Eocene Arctic Ocean (Brinkhuis et al., 2006; Sluijs et al., 2006), and the Cretaceous proto-Atlantic (Schouten et al., 2003).

Several limitations and potential confounding influences must be considered when applying and interpreting TEX₈₆-based temperature reconstructions. Several studies (Wakeham et al., 2003; Wuchter et al., 2005, 2006) have shown that, even though archaea live throughout the water column and hence produce GDGTs at a variety of depths (Herndl et al., 2005; Ingalls et al., 2006; Karner et al., 2001), sedimentary TEX₈₆ values typically reflect the SST of water layers <100 m deep. This is likely due to the efficient export of organic matter from the surface ocean in the form of aggregates and zooplankton fecal pellets (Huguet et al., 2006a). This is, however, not always the case, and sedimentary GDGTs can reflect the average temperature of the entire water column (Menzel et al., 2006). Another potential source of uncertainty is the seasonally varying production of GDGTs (e.g. Leider et al., 2010; Sinninghe Damsté et al., 2009; Wuchter et al., 2005) though sediment core-top TEX₈₆ values do appear to reflect mean annual temperatures (Kim et al., 2008). This is likely the result of seasonally invariant archaeal growth and export. Oxidation of GDGTs under prolonged oxic conditions, such as can occur in turbidites deposited in the deep ocean, can also lead to changes in TEX₈₆ values equivalent to 2–6 °C (Huguet et al., 2009). The relative abundance of *Crenarchaeota* versus *Euryarchaeota* or even different *Crenarchaeota* populations that may produce different relative abundances of GDGTs can also influence TEX₈₆-based SSTs (Schouten et al., 2008; Turich et al., 2007, 2008). This is particularly likely in coastal, lacustrine, or isolated basins (Trommer et al., 2009) as well as in upwelling areas (Lee et al., 2008), where certain strains of *Crenarchaeota* and *Euryarchaeota* might be favored. In anoxic settings, such as highly organic-rich sediments and anoxic basins, methanotrophic archaea may also be a significant source of GDGTs to sediments (Blaga et al., 2009; Sinninghe Damsté et al., 2009; Wakeham et al., 2004).

Besides the branched GDGTs, soil organic matter also typically contains some of the isoprenoidal GDGTs I, II, and III and some crenarchaeol. A significant contribution of these GDGTs prohibits the use of the TEX₈₆ proxy; this can be evaluated using the branched versus isoprenoid tetraether (BIT) index:

$$\text{BIT} = \frac{[\text{VI}] + [\text{VII}] + [\text{VIII}]}{[\text{V}] + [\text{VI}] + [\text{VII}] + [\text{VIII}]} \quad [7]$$

Values above 0.3 may be used as a general limit, although evaluation of the terrestrial and marine end-members is recommended when possible (Weijers et al., 2006). Because of this

complication, the TEX₈₆ cannot be used in many coastal settings and in most lakes (Blaga et al., 2009; Powers et al., 2004, 2010).

The distribution patterns of branched GDGTs have also been explored for their utility as paleoenvironmental proxies. The relative amount of cyclopentane moieties (VI–VIIIb and c, Figure 2), expressed as the CBT index, was found to be related to the pH of the soil, while the relative amount of methyl branches, expressed as the MBT, was positively correlated with the annual mean air temperature and, to a lesser extent, to soil pH (Peterse et al., 2009b; Weijers et al., 2007).

$$\text{CBT} = -\log \left(\frac{[\text{VIIb}] + [\text{VIIIb}]}{[\text{VII}] + [\text{VIII}]} \right) \quad [8]$$

MBT=

$$\frac{[\text{VIII}] + [\text{VIIIa}] + [\text{VIIIb}]}{[\text{VI}] + [\text{VII}] + [\text{VIII}] + [\text{VIa}] + [\text{VIIa}] + [\text{VIIIa}] + [\text{VIb}] + [\text{VIIb}] + [\text{VIIIb}]} \quad [9]$$

$$\text{CBT} = 3.33 - 0.38 \times \text{pH} (r^2 = 0.70, n = 134) \quad [10]$$

$$\text{MBT} = 0.867 - 0.096 \times \text{pH} + 0.021 \times \text{MAT} (r^2 = 0.82, n = 134) \quad [11]$$

The latter two equations can be combined:

$$\text{MBT} = 0.122 - 0.187 \times \text{CBT} + 0.020 \times \text{MAT} (r^2 = 0.77) \quad [12]$$

One example of the use of this proxy is the reconstruction of East Asian continental temperatures during the last deglaciation period (Peterse et al., 2010).

Biomarker Proxies of Salinity

Along with temperature, salinity is the most important oceanographic parameter. Its reconstruction can shed light on changes in ocean stratification, water mass circulation, and precipitation, which are tightly linked to climate. Recent developments in organic geochemistry have produced two promising paleosalinity proxies: the relative abundance of tetraunsaturated C₃₇ methyl ketones (alkenones) and the hydrogen isotopic composition (D/H) of organic compounds produced by phytoplankton and plants.

The %C_{37:4} Paleosalinity Proxy

The C_{37:4} alkenone is produced primarily in cold surface waters of the ocean (temperatures <10 °C) (Rosell-Melé et al., 1994) and is not well preserved in sediments compared to the di- and triunsaturated alkenones. It is, therefore, most often excluded from alkenone paleotemperature reconstructions (see the

section ‘Alkenone Paleothermometry’). In polar regions, however, the inclusion of C_{37:4} into the unsaturation ratio (U_k³⁷) can, in some cases, increase its correlation with SST (Bendle and Rosell-Melé, 2004).

At concentrations of %C_{37:4} (relative abundance of C_{37:4} to the total abundance of C_{37:2}, C_{37:3}, and C_{37:4}, as percentage) >5%, the temperature relation breaks down (Rosell-Melé, 1998; Sicre et al., 2002; Sikes et al., 1997), however, and a correlation with salinity has, instead, been suggested in the Nordic Seas and the North Atlantic (Rosell-Melé, 1998; Rosell-Melé et al., 2002; Sicre et al., 2002; Table 1). A %C_{37:4}-salinity dependence was also observed in the low-salinity waters of the Sea of Okhotsk (Seki, 2005), the Bering Strait (Harada et al., 2003), and the Baltic Sea (Blanz et al., 2005; Schulz et al., 2000; Table 1), leading to the suggestion that low salinities may impose stress on the alkenone-producing haptophytes, causing them to increase production of C_{37:4} (Harada et al., 2003).

The established %C_{37:4}-salinity relations, however, vary significantly between oceanic regions (Table 1), and no correlation with salinity was found in particulate and surface sediment samples from the Nordic Seas (Bendle et al., 2005) or the Atlantic, Pacific, and Indian sectors of the Southern Ocean (Sikes and Sicre, 2002). The applicability of %C_{37:4} as a salinity proxy may thus be restricted to certain cold, low-salinity marine environments. Further studies will need to establish whether this is due to a possible contribution to C_{37:4} production from other environmental factors. Nevertheless, the importance of establishing past salinity variations justifies this further work.

The Lipid δD Paleosalinity Proxy

The hydrogen isotopic composition ($\delta D = [(D/H)_{\text{sample}} / (D/H)_{\text{standard}} - 1] \times 1000$), where the standard is Vienna Standard Mean Ocean Water of lipids from plants and algae reflect the δD value of environmental water (Englebrecht and Sachs, 2005; Sachse et al., 2004; Sauer et al., 2001; Sessions et al., 1999; Zhang and Sachs, 2007). Precipitation δD values, in turn, reflect air temperature, air mass trajectory, intensity of precipitation, and the isotopic composition of source water. Temperature tends to dominate at high latitudes, while precipitation intensity tends to dominate in the tropics, such that lower precipitation δD values occur at lower temperatures and higher precipitation intensities (Dansgaard, 1964). Since most hydrogen in lipids is bound to carbon and, therefore, nonexchangeable on time scales of millions of years, the δD value of sedimentary lipids contains important information on past hydrological conditions and climate.

δD values in alkenones have been shown in culture and field studies to closely record the isotopic composition of the

Table 1 %C_{37:4}-Salinity relationships in different ocean basins

Ocean basin	Linear regression	R ²	Reference
North Atlantic and Nordic Seas	%C _{37:4} = 146.9(±10.6) – 4.1S (±0.3)	0.69	Rosell-Melé et al. (2002), p. 1301
North Atlantic	%C _{37:4} = 48.1S + 1691	0.78	Sicre et al. (2002), p. 1304
Bering Strait	%C _{37:4} = 11.7S + 397.6	0.76	Harada et al. (2003), p. 1310

water in which they were produced (Englebrecht and Sachs, 2005). Alkenone δD values are depleted relative to that of the water by some -193‰ (Englebrecht and Sachs, 2005), reflecting the initial deuterium depletion of the primary photosynthate of -171‰ (Yakir and DeNiro, 1990) and further fractionation during alkenone biosynthesis. A greater D depletion of $C_{37:3}$ relative to $C_{37:2}$ requires chromatographic separation of the individual alkenones prior to isotope analysis (D'Andrea et al., 2007; Schwab and Sachs, 2009). Moreover, different hydrogen isotope fractionation by the two most abundant producers of alkenones in the modern ocean, *E. huxleyi* and *Gephyrocapsa oceanica* (δD offset $\sim 30\text{‰}$; Schouten et al., 2006), requires knowledge of the species distribution and their temporal changes for robust paleoceanographic reconstructions based upon alkenone δD values.

D/H fractionation during the biosynthesis of alkenones and a wide variety of other phytoplanktonic and cyanobacterial lipids has been recently shown to occur in response to salinity changes (Schouten et al., 2006; Sachse and Sachs, 2008; Sachs and Schwab, 2011). D/H fractionation decreased by an average of $0.9 \pm 0.2\text{‰}$ per unit ($\sim \text{ppt}$) increase in salinity in a variety of lipids from hypersaline ponds on Christmas Island in the tropical Pacific Ocean (Sachse and Sachs, 2008) and in the dinoflagellate sterol dinosterol from the Chesapeake Bay estuary in the eastern United States (Sachs and Schwab, 2011) (Figure 3). This response to salinity is most often additive to the effect of the δD value of the environmental water, in which arid conditions result in higher evaporation relative to precipitation, and, therefore, higher water δD values and vice versa. This increases the sensitivity of biomarker δD as a salinity proxy for the quantitative reconstruction of past changes in the hydrological cycle.

The lipid δD paleosalinity proxy has been applied to reconstruct salinity variations in late Quaternary ocean sediments (Pahnke et al., 2007; van der Meer et al., 2007) and from late

Holocene sediments in the Black Sea (van der Meer et al., 2008) and from saline lakes on tropical Pacific islands (Sachs et al., 2009; Smittenberg et al., 2011).

There is still much we do not understand about the environmental influences on D/H fractionation during lipid synthesis in phytoplankton and cyanobacteria. In addition to salinity, recent work indicates that growth rate, growth phase, and growth temperature, all seem to influence D/H fractionation in phytoplankton (Wolhowe et al., 2009; Zhang et al., 2009). But these effects may be small in comparison to the water δD and salinity effect in many settings.

The Highly Branched Isoprenoid Sea Ice Proxy (IP₂₅)

C_{25} and C_{30} highly branched isoprenoid (HBI) alkenes are ubiquitous components in recent sediments that appear to originate from four genera of marine diatoms (Class Bacillariophyceae), *Rhizosolenia*, *Haslea*, *Navicula*, and *Pleurosigma* (Grossi et al., 2004). Several C_{25} HBI alkenes ($C_{25:1}$ to $C_{25:6}$) have been characterized in cultures of *Haslea* and *Pleurosigma* (Wraige et al., 1997).

The reported covariance of numbers of unsaturations in HBI alkenes in diatoms with temperature once represented a hopeful new paleotemperature proxy (Rowland et al., 2001). However, this proxy has not been successfully applied to paleotemperature reconstructions.

Recently, a unique C_{25} monounsaturated HBI has been detected in sea ice samples. Some *Haslea* species have been reported to inhabit Arctic sea ice (Booth and Horner, 1998). The recent sediments from the Arctic with this C_{25} monounsaturated HBI underlie either seasonal or permanent ice cover, and it has yet to be reported in sediments from more temperate environments or from northern high-latitude open-water or ice-free phytoplankton (Belt et al., 2007). As a result, the

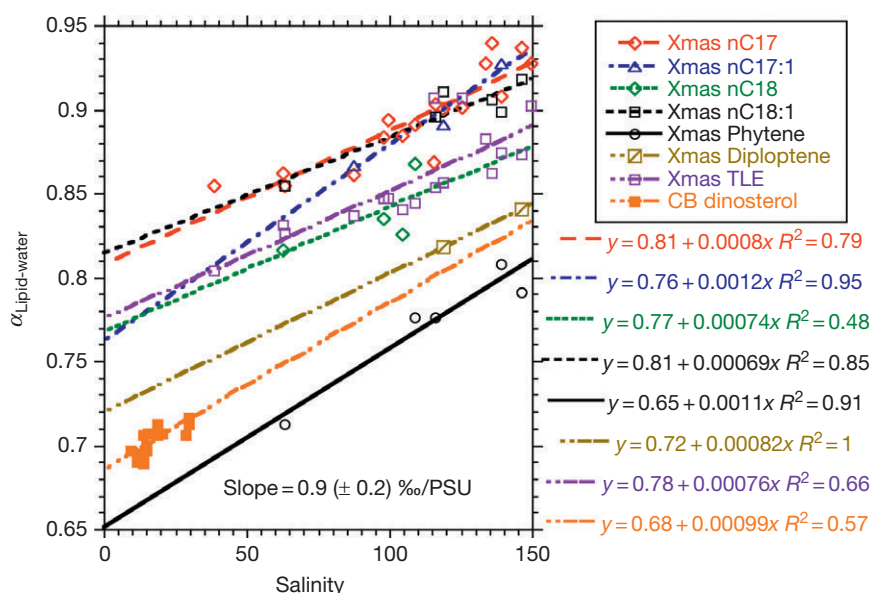


Figure 3 Fractionation factors for a variety of lipids from hypersaline ponds on Christmas Island (Sachse and Sachs, 2008) and the Chesapeake Bay estuary (Sachs and Schwab, 2011). The average slope of all regressions is 0.0009 ± 0.0002 , implying a relatively constant decrease in D/H fractionation of 0.9‰ per unit increase in salinity.

monounsaturated isomer has been attributed to sea ice diatoms, and its existence in sediments is an indicator of sea ice. The proxy has been named IP25 (ice proxy with 25 carbons) (Belt et al., 2007).

The IP25 records from sediments have proven to be remarkably consistent with historical sea ice records and with other paleoclimatic parameters (Gregory et al., 2010; Massé et al., 2008). The $\delta^{13}\text{C}$ values of IP25 in sea ice and in sediment trap material collected under sea ice are also distinguishably heavier than that of organic matter from planktonic or terrigenous sources, supporting the use of IP25 as an Arctic sea ice proxy (Belt et al., 2008).

IP25 has been applied for the reconstruction of sea ice over the last several centuries (Vare et al., 2010) and during the last 30 000 years (Müller et al., 2009).

Summary

Tremendous progress has been made in the last decade developing and applying organic geochemical indicators of past climate. A small set of the most significant of these advances has been reviewed here. Advances in analytical chemistry in the coming years ought to make all of the temperature and hydrologic proxies described here available to a broad range of geochemists. The different biological, physical, and chemical underpinnings of these proxies, as compared to the inorganic paleoclimate proxies widely used today, such as the oxygen and carbon isotopes and trace metal ratios in foraminiferal shells, make them complementary. In some climatic or depositional environments, one or another of the temperature or hydrologic proxies may work best, while in other settings it ought to be possible to apply two or more indicators of the same climate parameter within a single sediment core. As a result, increasingly robust reconstructions of climate are certain to emerge.

See also: Diatom Introduction; Structures and Applications of Biomarkers from Arctic Sea Ice Diatoms. **Carbonate Stable Isotopes:** Lake Sediments. **Diatom Methods:** $\delta^{18}\text{O}$ Records; Diatomites: Their Formation, Distribution, and Uses; Salinity and Climate Reconstructions from Continental Lakes. **Paleobotany:** Ancient Plant DNA; Overview of Terrestrial Pollen Data. **Paleoceanography, Biological Proxies:** Coccolithophores; Dinoflagellates; Marine Diatoms; Planktic Foraminifera; Radiolarians and Silicoflagellates; Temperature Proxies, Census Counts. **Paleoceanography, Physical and Chemical Proxies:** Carbon Cycle Proxies ($\delta^{11}\text{B}$, $\delta^{13}\text{C}_{\text{calcite}}$, $\delta^{13}\text{C}_{\text{organic}}$; Shell Weights, B/Ca, U/Ca, Zn/Ca, Ba/Ca); Mg/Ca and Sr/Ca Paleothermometry from Calcareous Marine Fossils; Nutrient Proxies. **Paleoclimate Reconstruction:** Approaches. **Paleolimnology:** Overview of Paleolimnology; Pigment Studies.

References

- Bard E, Rostek F, and Sonzogni C (1997) Interhemispheric synchrony of the last deglaciation inferred from alkenone palaeothermometry. *Nature* 385: 707–710.
- Belt ST, Massé G, Rowland SJ, Poulin M, Michel C, and LeBlanc B (2007) A novel chemical fossil of palaeo sea ice: IP25. *Organic Geochemistry* 38: 16–27.
- Belt ST, Massé G, Vare LL, et al. (2008) Distinctive ^{13}C isotopic signature distinguishes a novel sea ice biomarker in Arctic sediments and sediment traps. *Marine Chemistry* 112: 158–167.
- Bendle J and Rosell-Melé A (2004) Distributions of UK37 and $\text{U}^{K_{37}}$ in the surface waters and sediments of the Nordic Seas: Implications for paleoceanography. *Geochimica et Cosmochimica Acta* 5(11): <http://dx.doi.org/10.1029/2004GC000741>.
- Bendle J, Rosell-Melé A, and Ziveri P (2005) Variability of unusual distributions of alkenones in the surface waters of the Nordic seas. *Paleoceanography* 20: <http://dx.doi.org/10.1029/2004PA001025>.
- Blaga CI, Reichart GJ, Heiri O, and Sinninghe Damsté JS (2009) Tetraether membrane lipid distributions in water-column particulate matter and sediments: A study of 47 European lakes along a north–south transect. *Journal of Paleolimnology* 41: 523–540.
- Blanz T, Ernis KC, and Siegel H (2005) Controls on alkenone unsaturation ratios along the salinity gradient between the open ocean and the Baltic Sea. *Geochimica et Cosmochimica Acta* 69(14): 3589–3600.
- Booth BC and Horner RA (1998) Microalgae on the Arctic Ocean Section, 1994: Species abundance and biomass. *Deep-Sea Research II* 44: 1607–1622.
- Brassell SC, Dumitrescu M, and Party OLSS (2004) Recognition of alkenones in a lower Aptian porcellanite from the west-central Pacific. *Organic Geochemistry* 35: 181–188.
- Brassell SC, Eglinton G, Marlowe IT, Pflaumann U, and Sarnthein M (1986) Molecular stratigraphy: A new tool for climatic assessment. *Nature* 320: 129–133.
- Brinkhuis H, Schouten S, Collinson ME, et al. (2006) Episodic fresh surface waters in the Eocene Arctic Ocean. *Nature* 441: 606–609.
- Castañeda IS, Schefuss E, Pätzold J, Sinninghe Damsté JS, Weldeab S, and Schouten S (2010) Millennial-scale sea surface temperature changes in the eastern Mediterranean (Nile River Delta region) over the last 27,000 years. *Paleoceanography* 25: <http://dx.doi.org/10.1029/2009PA001740>.
- Conte MH, Eglinton G, and Madureira LAS (1992) Long-chain alkenones and alkyl alkenoates as paleotemperature indicators: Their production, flux and early sedimentary diagenesis in the eastern North Atlantic. *Organic Geochemistry* 19(1–3): 287–298.
- Conte MH, Thompson A, Lesley D, and Harris RP (1998) Genetic and physiological influences on the alkenone/alkenoate versus growth temperature relationship in *Emiliania huxleyi* and *Gephyrocapsa oceanica*. *Geochimica et Cosmochimica Acta* 62(1): 51–64.
- Conte MH, Weber JC, King LL, and Wakeham SG (2001) The alkenone temperature signal in western North Atlantic surface waters. *Geochimica et Cosmochimica Acta* 65(23): 4275–4287.
- D'Andrea WJ, Liu Z, Da Rosa Alexandre M, Wattley S, Herbert TD, and Huang Y (2007) An efficient method for isolating individual long-chain alkenones for compound-specific hydrogen isotope analysis. *Analytical Chemistry* 79: 3430–3435.
- Dansgaard W (1964) Stable isotopes in precipitation. *Tellus* 16: 436–468.
- De Rosa M and Gambacorta A (1988) The lipids of Archaeobacteria. *Progress in Lipid Research* 27: 153–177.
- Englebrecht AC and Sachs JP (2005) Determination of sediment provenance at drift sites using hydrogen isotopes and unsaturation ratios in alkenones. *Geochimica et Cosmochimica Acta* 69(17): 4253–4265.
- Gliozzi A, Paoli G, De Rosa M, and Gambacorta A (1983) Effect of isoprenoid cyclization on the transition temperature of lipids in thermophilic archaeobacteria. *Biochimica et Biophysica Acta – Protein Structure and Molecular Enzymology* 735: 234–242.
- Gong C and Hollander DJ (1999) Evidence for differential degradation of alkenones under contrasting bottom water oxygen conditions: Implication for paleotemperature reconstruction. *Geochimica et Cosmochimica Acta* 63(3/4): 405–411.
- Gregory TR, Smart CW, Hart MB, Massé G, Vare LL, and Belt ST (2010) Holocene palaeoceanographic changes in Barrow Strait, Canadian Arctic: Foraminiferal evidence. *Journal of Quaternary Science* 25: 903–910.
- Grossi V, Beker B, Geenevasen JAJ, et al. (2004) C25 highly branched isoprenoid alkenes from the marine benthic diatom *Pleurosigma strigosum*. *Phytochemistry* 65: 3049–3055.
- Hamanaka J, Sawada K, and Tanoue E (2000) Production rates of C37 alkenones determined by ^{13}C -labeling technique in the euphotic zone of Sagami Bay, Japan. *Organic Geochemistry* 31: 1095–1102.
- Harada N, Shin KH, Murata A, Uchida M, and Nakatani T (2003) Characteristics of alkenones synthesized by a bloom of *Emiliania huxleyi* in the Bering Sea. *Geochimica et Cosmochimica Acta* 67(8): 1507–1519.
- Herbert TD, Schuffert JD, Andreassen D, et al. (2001) Collapse of the California Current during glacial maxima linked to climate change on land. *Science* 293: 71–76.
- Herfort L, Schouten S, Boon JP, et al. (2006) Characterization of transport and deposition of terrestrial organic matter in the southern North Sea using the BIT index. *Limnology and Oceanography* 51: 2196–2205.

- Herndl GJ, Reinthaler T, Teira E, et al. (2005) Contribution of Archaea to total prokaryotic production in the deep Atlantic Ocean. *Applied and Environmental Microbiology* 71: 2303–2309.
- Hopmans EC, Weijers JWH, Schefu[s] E, Herfort L, Sinninghe Damsté JS, and Schouten S (2004) A novel proxy for terrestrial organic matter in sediments based on branched and isoprenoid tetraether lipids. *Earth and Planetary Science Letters* 224: 107–116.
- Huguet C, Cartes JE, Sinninghe Damsté JS, and Schouten S (2006a) Marine crenarchaeotal membrane lipids in decapods: Implications for the TEX₈₆ paleothermometer. *Geochemistry, Geophysics, Geosystems* 7: Q11010. <http://dx.doi.org/10.1029/2006GC001305>.
- Huguet C, Kim J-H, de Lange GJ, Sinninghe Damsté JS, and Schouten S (2009) Effects of long-term oxic degradation of the U^K₃₇, TEX₈₆ and BIT organic proxies. *Organic Geochemistry* 40: 1188–1194.
- Huguet C, Kim J-H, Sinninghe Damsté JS, and Schouten S (2006b) Reconstruction of sea surface temperature variations in the Arabian Sea over the last 23 kyr using organic proxies (TEX₈₆ and U^K₃₇). *Paleoceanography* 21: PA3099. <http://dx.doi.org/10.1029/2006PA001351>.
- Ingalls AE, Shah SR, Hansman RL, et al. (2006) Quantifying archaeal community autotrophy in the mesopelagic ocean using natural radiocarbon. *Proc Natl Acad Sci USA* 103: 6442–6447.
- Jenkyns HC, Forster A, Schouten S, and Sinninghe Damsté JS (2004) High temperatures in the Late Cretaceous Arctic Ocean. *Nature* 432: 888–892.
- Karner MB, DeLong EF, and Karl DM (2001) Archaeal dominance in the mesopelagic zone of the Pacific Ocean. *Nature* 409: 507–510.
- Kim J-H, Schouten S, Hopmans EC, Donner B, and Sinninghe Damsté JS (2008) Global sediment core-top calibration of the TEX₈₆ paleothermometer in the ocean. *Geochimica et Cosmochimica Acta* 72: 1154–1173.
- Kim J-H, Van der Meer J, Schouten S, et al. (2010) New indices and calibrations derived from the distribution of crenarchaeal isoprenoid tetraether lipids: Implications for past sea surface temperature reconstructions. *Geochimica et Cosmochimica Acta* 74: 4639–4654.
- Koga Y and Morii H (2005) Recent advances in structural research on ether lipids from Archaea including comparative and physiological aspects. *Bioscience, Biotechnology, and Biochemistry* 69: 2019–2034.
- Koutavas A and Sachs JP (2008) Northern timing of deglaciation in the eastern equatorial Pacific from alkenone paleothermometry. *Paleoceanography* 23: <http://dx.doi.org/10.1029/2008PA001593>.
- Leduc G, Schneider R, Kim JH, and Lohmann G (2010) Holocene and Eemian sea surface temperature trends as revealed by alkenone and Mg/Ca paleothermometry. *Quaternary Science Reviews* 29(7–8): 989–1004.
- Lee KE, Kim JH, Wilke I, Helmke P, and Schouten S (2008) A study of the alkenone, TEX₈₆, and planktonic foraminifera in the Benguela upwelling system: Implications for past sea surface temperature estimates. *Geochemistry, Geophysics, Geosystems* 9: Q10019. <http://dx.doi.org/10.1029/2008GC002056>.
- Leider A, Hinrich K-U, Mollenhauer G, and Versteegh G (2010) Core-top calibration of the lipid-based U^K₃₇ and TEX₈₆ temperature proxies on the southern Italian shelf (SW Adriatic Sea, Gulf of Taranto). *Earth and Planetary Science Letters* 300: 112–124.
- Leininger S, Ulrich T, Schlöter M, et al. (2006) Archaea predominate among ammonia-oxidizing prokaryotes in soils. *Nature* 442: 806–809.
- Marlowe IT, et al. (1984) Long chain (n-C37–C39) alkenones in the Prymnesiophyceae. Distribution of alkenones and other lipids and their taxonomic significance. *British Phycological Journal* 19: 203–216.
- Massé G, Rowland SJ, Sicre M-A, Jacob J, Jansen E, and Belt ST (2008) Abrupt climate changes for Iceland during the last millennium: Evidence from high resolution sea ice reconstructions. *Earth and Planetary Science Letters* 269: 565–569.
- Menzel D, Hopmans EC, Schouten S, and Sinninghe Damsté JS (2006) Membrane tetraether lipids of planktonic Crenarchaeota in Pliocene sapropels of the eastern Mediterranean Sea. *Paleogeogr Paleoclimatol Paleoecol* 239: 1–15.
- Müller PJ, Cepek M, Ruhland G, and Schneider RR (1997) Alkenone and coccolithophorid species changes in Late Quaternary sediments from the Walvis Ridge: Implications for the alkenone paleotemperature method. *Paleogeography, Paleoclimatology, Paleocology* 135: 71–96.
- Müller PA and Fischer G (2001) A 4-year sediment trap record of alkenones from the filamentous upwelling region off Cape Blanc, NW Africa and a comparison with distributions in underlying sediments. *Deep Sea Research Part I: Oceanographic Research Papers* 48: 1877–1903.
- Müller PJ, Kirst G, Ruhland G, von Storch I, and Rosell-Mele A (1998) Calibration of the alkenone paleotemperature index U^K₃₇ based on core-tops from the eastern South Atlantic and the global ocean (60°N–60°S). *Geochimica et Cosmochimica Acta* 62(10): 1757–1772.
- Müller J, Massé G, Stein R, and Belt ST (2009) Variability of sea-ice conditions in the Fram Strait over the past 30,000 years. *Nature Geoscience* 2: 772–776.
- Ohkouchi N, Kawamura K, Kawahata H, and Okada H (1999) Depth ranges of alkenone production in the central Pacific Ocean. *Global Biogeochemical Cycles* 13(2): 695–704.
- Pahnke K and Sachs JP (2006) Sea surface temperatures of southern midlatitudes 0–160 kyr B.P. *Paleoceanography* 21: PA2003. <http://dx.doi.org/10.1029/2005PA001191>.
- Pahnke K, Sachs JP, Keigwin LD, Timmermann A, and Xie S-P (2007) Eastern tropical Pacific hydrologic changes during the past 27,000 years from D/H ratios in alkenones. *Paleoceanography* 22: PA4214. <http://dx.doi.org/10.1029/2007PA001468>.
- Peterse F, Kim JH, Schouten S, Kristensen DK, Koc N, and Sinninghe Damsté JS (2009a) Constraints on the application of the MBT/CBT paleothermometer at high latitude environments (Svalbard, Norway). *Organic Geochemistry* 40: 692–699.
- Peterse F, Prins MA, Beets CJ, et al. (2010) Decoupled warming and monsoon precipitation in East Asia over the last deglaciation. *Earth and Planetary Science Letters*. <http://dx.doi.org/10.1016/j.epsl.2010.11.010>.
- Peterse F, Schouten S, van der Meer J, van der Meer MTJ, and Sinninghe Damsté JS (2009b) Distribution of branched tetraether lipids in geothermally heated soils: Implications for the MBT/CBT temperature proxy. *Organic Geochemistry* 40: 201–205.
- Powers LA, Werne JP, Johnson TC, Hopmans EC, Sinninghe Damsté JS, and Schouten S (2004) Crenarchaeotal membrane lipids in lake sediments: A new paleotemperature proxy for continental paleoclimate reconstruction? *Geology* 32: 613–616.
- Powers LA, Werne JP, Vanderwoude AJ, Sinninghe Damsté JS, Hopmans EC, and Schouten S (2010) Applicability and calibration of the TEX₈₆ paleothermometer in lakes. *Organic Geochemistry* 41: 404–413.
- Prahl FG, Collier RB, Dymond J, Lyle M, and Sparrow MA (1993) A biomarker perspective on pyromnesiophyte productivity in the northeast Pacific Ocean. *Deep-Sea Research I* 40: 2061–2076.
- Prahl FG, de Lange GJ, Lyle M, and Sparrow MA (1989) Post-depositional stability of long-chain alkenones under contrasting redox conditions. *Nature* 341: 434–437.
- Prahl FG, Muehlhausen LA, and Zahnle DL (1988) Further evaluation of long-chain alkenones as indicators of paleoceanographic conditions. *Geochimica et Cosmochimica Acta* 52: 2303–2310.
- Prahl FG, Popp BN, Karl DM, and Sparrow MA (2005) Ecology and biogeochemistry of alkenone production at Station Aloha. *Deep Sea Research Part I: Oceanographic Research Papers* 52(5): 699–719.
- Prahl FG, Rontani JF, Zabeti N, Walinsky SE, and Sparrow MA (2010) Systematic pattern in – Temperature residuals for surface sediments from high latitude and other oceanographic settings. *Geochimica et Cosmochimica Acta* 74(1): 131–143.
- Prahl FG and Wakeham SG (1987) Calibration of unsaturation patterns in long-chain ketone compositions for paleotemperature assessment. *Nature* 330: 367–369.
- Prahl FG, Wolfe GV, and Sparrow MA (2003) Physiological impacts on alkenone paleothermometry. *Paleoceanography* 18(2): 1025. <http://dx.doi.org/10.1029/2002PA000803>.
- Rontani J-F, Bonin P, Jameson I, and Volkman JK (2005) Degradation of alkenones and related compounds during oxic and anoxic incubation of the marine haptophyte *Emiliania huxleyi* with bacterial consortia isolated from microbial mats from the Camargue, France. *Organic Geochemistry* 36: 603–618.
- Rontani J-F, Harji R, Guasco S, et al. (2008) Degradation of alkenones by aerobic heterotrophic bacteria: Selective or not. *Organic Geochemistry* 39: 34–51.
- Rosell-Mele A (1998) Interhemispheric appraisal of the value of alkenone indices as temperature and salinity proxies in high-latitude locations. *Paleoceanography* 13(6): 694–703.
- Rosell-Melé A, Carter J, and Eglinton G (1994) Distributions of long-chain alkenones and alkyl alkenoates in marine surface sediments from the North East Atlantic. *Organic Geochemistry* 22: 501–509.
- Rosell-Melé A, Jansen E, and Weinelt M (2002) Appraisal of a molecular approach to infer variations in surface ocean freshwater inputs into the North Atlantic during the last glacial. *Global and Planetary Change* 34(3–4): 143–152.
- Rowland SJ, Belt ST, Wraige EJ, Massé G, Roussakis C, and Robert J-M (2001) Effects of temperature on polyunsaturation in cytosolic lipids of *Haslea ostrearia*. *Phytochemistry* 56: 597–602.
- Sachs JP and Lehman SJ (1999) Subtropical North Atlantic temperatures 60,000 to 30,000 years ago. *Science* 286: 756–759.
- Sachs JP, Sachse D, Smittenberg RH, Zhang Z, Battisti DS, and Golubic S (2009) Southward movement of the Pacific intertropical convergence zone AD 1400–1850. *Nature Geoscience* 2: 519–525.
- Sachs JP and Schwab VF (2011) Hydrogen isotopes in dinosterol from the Chesapeake Bay estuary. *Geochimica et Cosmochimica Acta* 75(2): 444–459.
- Sachse D, Radke J, and Gleixner G (2004) Hydrogen isotope ratios of recent lacustrine sedimentary *n*-alkanes record modern climate variability. *Geochimica et Cosmochimica Acta* 68(23): 4877–4889.

- Sachs D and Sachs JP (2008) Inverse relationship between D/H fractionation in cyanobacterial lipids and salinity in Christmas Island saline ponds. *Geochimica et Cosmochimica Acta* 72(3): 793–806.
- Sauer PE, Eglinton TI, Hayes JM, Schimmelmann A, and Sessions AL (2001) Compound-specific D/H ratios of lipid biomarkers from sediments as a proxy for environmental and climatic conditions. *Geochimica et Cosmochimica Acta* 65(2): 213–222.
- Schouten S, Hopmans EC, Forster A, van Breugel Y, Kuypers MMM, and Damsté JSS (2003) Extremely high sea-surface temperatures at low latitudes during the middle Cretaceous as revealed by archaeal membrane lipids. *Geology* 31: 1069–1072.
- Schouten S, Hopmans EC, Pancost RD, and Damsté JSS (2000) Widespread occurrence of structurally diverse tetraether membrane lipids: Evidence for the ubiquitous presence of low-temperature relatives of hyperthermophiles. *PNAS* 97: 14421–14426.
- Schouten S, Hopmans EC, Schefuß E, and Sinninghe Damsté JS (2002) Distributional variations in marine crenarchaeotal membrane lipids: A new tool for reconstructing ancient sea water temperatures? *Earth and Planetary Science Letters* 204: 265–274.
- Schouten S, Hopmans EC, and Sinninghe Damsté JS (2004) The effect of maturity and depositional redox conditions on archaeal tetraether lipid palaeothermometry. *Organic Geochemistry* 35: 567–571.
- Schouten S, Ossebaer J, Kienhuis M, and Bijma J (2005) Stable hydrogen isotopic fractionation patterns of long chain alkenones. *Organic Geochemistry: Challenges for the 21st Century* 1: 603–604.
- Schouten S, Ossebaer J, Schreiber K, et al. (2006) The effect of temperature, salinity and growth rate on the stable hydrogen isotopic composition of long chain alkenones produced by *Emiliania huxleyi* and *Gephyrocapsa oceanica*. *Biogeosciences* 3: 113–119.
- Schouten S, van der Meer MTJ, Hopmans EC, and Sinninghe Damsté JS (2008) Comment on 'Lipids of marine Archaea: Patterns and provenance in the water column and sediments' by Turich et al. (2007). *Geochimica et Cosmochimica Acta* 72: 5342–5346.
- Schouten S, et al. (2009) An interlaboratory study of TEX₈₆ and BIT analysis using high-performance liquid chromatography-mass spectrometry. *Geochemistry, Geophysics, Geosystems* 10: Q03012.
- Schulz H-M, Anne S, and Kay-Christian E (2000) Long-chain alkenone patterns in the Baltic sea – An ocean-freshwater transition. *Geochimica et Cosmochimica Acta* 64(3): 469–477.
- Schwab VF and Sachs JP (2009) The measurement of D/H ratio in alkenones and their isotopic heterogeneity. *Organic Geochemistry* 40(1): 111–118.
- Seki O (2005) Decreased surface salinity in the Sea of Okhotsk during the last glacial period estimated from alkenones. *Geophysical Research Letters* 32(8): <http://dx.doi.org/10.1029/2004GL022177>.
- Sessions AL, Burgoyne TW, Schimmelmann A, and Hayes JM (1999) Fractionation of hydrogen isotopes in lipid biosynthesis. *Organic Geochemistry* 30(9): 1193–1200.
- Sicre M-A, Bard E, Ezat U, and Rostek F (2002) Alkenone distributions in the North Atlantic and Nordic sea surface waters. *Geochemistry, Geophysics, Geosystems* 3: <http://dx.doi.org/10.1029/2001GC000159>.
- Sikes EL, O'Leary T, Nodder SD, and Volkman JK (2005) Alkenone temperature records and biomarker flux at the subtropical point on the Chatham Rise, SW Pacific Ocean. *Deep Sea Research Part I: Oceanographic Research Papers* 52(5): 721–748.
- Sikes EL and Sicre MA (2002) Relationship of the tetra-unsaturated C-37 alkenone to salinity and temperature: Implications for paleoproxy applications. *Geochemistry, Geophysics, Geosystems* 3: art. no.-1063.
- Sikes EL and Volkman JK (1993) Calibration of alkenone unsaturation ratios (U37k') for paleotemperature estimation in cold polar waters. *Geochimica et Cosmochimica Acta* 57(8): 1883–1889.
- Sikes EL, Volkman JK, Robertson LG, and Pichon J-J (1997) Alkenones and alkenes in surface waters and sediments of the Southern Ocean: Implications for paleotemperature estimation in polar regions. *Geochimica et Cosmochimica Acta* 61(7): 1495–1505.
- Sinninghe Damsté JS, Hopmans EC, Pancost RD, Schouten S, and Geenevasen JAJ (2000) Newly discovered non-isoprenoid glycerol dialkyl glycerol tetraether lipids in sediments. *Chemical Communications* 1683–1684.
- Sinninghe Damsté JS, Ossebaer J, Abbas B, Schouten S, and Verschuren D (2009) Fluxes and distribution of tetraether lipids in an equatorial African lake: Constraints on the application of the TEX₈₆ palaeothermometer and BIT index in lacustrine settings. *Geochimica et Cosmochimica Acta* 73: 4232–4249.
- Sinninghe Damsté JS, Schouten S, Hopmans EC, van Duin ACT, and Geenevasen JAJ (2002) Crenarchaeol: The characteristic core glycerol dibiphytanyl glycerol tetraether membrane lipid of cosmopolitan pelagic crenarchaeota. *Journal of Lipid Research* 43: 1641–1651.
- Sluijs A, Schouten S, Pagani M, et al. (2006) Subtropical Arctic Ocean temperatures during the Palaeocene/Eocene thermal maximum. *Nature* 441: 610–613.
- Smittenberg RH, Saenger C, Dawson MN, and Sachs JP (2011) Compound-specific D/H ratios of the marine lakes of Palau as proxies for West Pacific Warm Pool hydrologic variability. *Quaternary Science Reviews* 30: 921–933.
- Ternois Y, Sicre M-A, Boireau A, Beaufort L, Miquel J-C, and Jeandel C (1998) Hydrocarbons, sterols and alkenones in sinking particles in the Indian Ocean sector of the Southern Ocean. *Organic Geochemistry* 28(7–8): 489–501.
- Tierney JE and Russell JM (2009) Distributions of branched GDGTs in a tropical lake system: Implications for lacustrine application of the MBT/CBT paleoproxy. *Organic Geochemistry* 40: 1032–1036.
- Tierney JE, Russell JM, Huang Y, Sinninghe Damsté JS, Hopmans EC, and Cohen AS (2008) Northern hemisphere controls on tropical southeast African climate during the past 60,000 years. *Science* 322: 252–255.
- Trommer G, Siccha M, van der Meer MTJ, et al. (2009) Distribution of Crenarchaeota tetraether membrane lipids in surface sediments from the Red Sea. *Organic Geochemistry* 40: 724–731.
- Turich C, Freeman KH, Bruns MA, Conte M, Jones AD, and Wakeham SG (2007) Lipids of marine Archaea: Patterns and provenance in the water-column and sediments. *Geochimica et Cosmochimica Acta* 71: 3272–3291.
- Turich C, Freeman KH, Jones AD, Bruns MA, Conte M, and Wakeham SG (2008) Reply to the Comment by S. Schouten, M. van der Meer, E. Hopmans and J.S. Sinninghe Damsté on 'Lipids of marine Archaea: Patterns and provenance in the water column'. *Geochimica et Cosmochimica Acta* 72: 5347–5349.
- Uda I, Sugai A, Itoh Y, and Itoh T (2001) Variation in molecular species of polar lipids from *Thermoplasma acidophilum* depends on growth temperature. *Lipids* 36: 103–105.
- van der Meer MTJ, Marianne Baas W, Rijpstra IC, et al. (2007) Hydrogen isotopic compositions of long-chain alkenones record freshwater flooding of the Eastern Mediterranean at the onset of sapropel deposition. *Earth and Planetary Science Letters* 262(3–4): 594–600.
- van der Meer MTJ, Sangiorgi F, Baas M, Brinkhuis H, Sinninghe Damsté JS, and Stefan S (2008) Molecular isotopic and dinoflagellate evidence for Late Holocene freshening of the Black Sea. *Earth and Planetary Science Letters* 267(3–4): 426–434.
- Vare LL, Massé G, and Belt ST (2010) A biomarker-based reconstruction of sea ice conditions for the Barents Sea in recent centuries. *The Holocene* 20: 637–643.
- Volkman JK, Eglinton G, Corner EDS, and Sargent JR (1980) Novel unsaturated straight-chain C37–C39 methyl and ethyl ketones in marine sediments and a coccolithophore *Emiliania huxleyi*. In: Douglas AG and Maxwell JR (eds.) *Advances in Organic Geochemistry, 1979. Physics and Chemistry of the Earth*, pp. 219–227. Oxford: Pergamon Press.
- Wakeham SG, Hopmans EC, Schouten S, and Sinninghe Damsté JS (2004) Archaeal lipids and anaerobic oxidation of methane in euxinic water columns: A comparative study of the Black Sea and Cariaco Basin. *Chemical Geology* 205: 427–442.
- Wakeham SG, Lewis CM, Hopmans EC, Schouten S, and Sinninghe Damsté JS (2003) Archaea mediate anaerobic oxidation of methane in deep euxinic waters of the Black Sea. *Geochimica et Cosmochimica Acta* 67: 1359–1374.
- Weijers JWH, Schouten S, Hopmans EC, et al. (2006) Membrane lipids of mesophilic anaerobic bacteria thriving in peats have typical archaeal traits. *Environmental Microbiology* 8: 648–657.
- Weijers JWH, Schouten S, van den Donker JC, Hopmans EC, and Sinninghe Damsté JS (2007) Environmental controls on bacterial tetraether membrane lipid distribution in soils. *Geochimica et Cosmochimica Acta* 71: 703–713.
- Wolhowe MD, Prah FG, Probert I, and Maldonado M (2009) Growth phase dependent hydrogen isotopic fractionation in alkenone-producing haptophytes. *Biogeosciences* 6: 1681–1694.
- Wraige EJ, Belt ST, Lewis CA, et al. (1997) Variations in structures and distributions of C25 highly branched isoprenoid (HBI) alkenes in cultures of the diatom, *Haslea ostrearia* (Simonsen). *Organic Geochemistry* 27: 497–505.
- Wuchter C, Schouten S, Coolen MJL, and Sinninghe Damsté JS (2004) Temperature-dependent variation in the distribution of tetraether membrane lipids of marine Crenarchaeota: Implications for TEX₈₆ paleothermometry. *Paleoceanography* 19: PA4028.
- Wuchter C, Schouten S, Wakeham SG, and Sinninghe Damsté JS (2005) Temporal and spatial variation in tetraether membrane lipids of marine Crenarchaeota in particulate organic matter: Implications for TEX₈₆ paleothermometry. *Paleoceanography* 20: <http://dx.doi.org/10.1029/2004PA001110>.
- Wuchter C, Schouten S, Wakeham SG, and Sinninghe Damsté JS (2006) Archaeal tetraether membrane lipid fluxes in the northeastern Pacific and the Arabian Sea: Implications for TEX₈₆ paleothermometry. *Paleoceanography* 21(4): <http://dx.doi.org/10.1029/2006PA001279>.
- Yakir D and DeNiro MJ (1990) Oxygen and hydrogen isotope fractionation during cellulose metabolism in *Lemna gibba* L. *Plant Physiology* 93: 325–332.
- Zhang Z and Sachs JP (2007) Hydrogen isotope fractionation in freshwater algae: I. Variations among lipids and species. *Organic Geochemistry* 38: 582–608.
- Zhang Z, Sachs JP, and Marchetti A (2009) Hydrogen isotope fractionation in freshwater and marine algae: II Temperature and nitrogen limited growth rate effects. *Organic Geochemistry* 40: 428–439.

Coccolithophores

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Introduction

Coccolithophores are unicellular, autotrophic organisms belonging to the marine phytoplankton that during certain phases of their cycle of life are able to produce calcium carbonate plates called coccoliths. In consequence, they are included in the more general term calcareous nannoplankton. Cell size in general is lower than 30 μm , and their reproduction is mainly by binary division (mitotic), although in several cases, more complex processes are observed. Once the cell has died, the coccosphere descends to the sea floor, constituting one of the most important contributors to the oceanic sediments since the late Triassic (Figure 1).

Their habitat is restricted to the upper part of the water column with a wide latitudinal range extending from the polar regions to the tropics. Owing to their great abundance and the existing relationship between photosynthesis and calcification, coccolithophores are a key element in the biochemical cycles of the Earth. Blooms of these algae are observed today, covering large regions of the ocean that can be monitored by satellite imagery (Figure 2).

The use of coccolithophores as paleoclimatic and paleoceanographic indicators needs precise knowledge of their biology and ecology. Culture experiments are completed with the recovery of living assemblages in different oceans, as well as annual and seasonal records in sediment traps, providing valuable information about their behavior. Additionally, the relationship between the coccolithophore assemblages and environmental parameters, such as surface water temperature or nutrient contents, makes it possible to use them as paleoceanographic proxies in the sedimentary material recovered by coring, drilling, or in continental outcrops.

The abundance and regularity of these organisms in the oceanic sediment record, as well as their relatively high rate of evolutionary divergence, confers important biostratigraphic value to this group.

Biology

Included in the Division Haptophyta (Table 1), coccolithophores are characterized by having a golden-brown chloroplast and, in most observed living species (with a motile phase), a flagella system. A third element, called haptonema, is sometimes observed between the long regular flagella (Figure 3). The secretion of scales takes place in the vicinity of the Golgi body, and these calcify inside the cell to produce a coccolith (in most cases calcitic) that finally occupies the cell periphery. The function of these plates is not well understood, but probably they serve to protect the cell against grazers and viruses or as a tool to control the light absorbed by the cell. The morphology of coccoliths and the structure and organization of their crystalline units are the

criteria used for their classification (Figure 4). The orientation of the crystal units produces an interference figure when observed under the cross-polarized light of a petrographic microscope, and their individual patterns of interference are diagnostic at the species level (Green and Leadbeater, 1994; Haq, 1983; Haq and Boersma, 1978; Tappan, 1981; Thierstein and Young, 2004; Winter and Siesser, 1994; Young et al., 1997).

Three major types of coccoliths can be distinguished (Figure 4; Young et al., 2003):

Heterococcoliths: These are characterized by calcitic units of varying shapes and sizes and ordered in cycles. Their generation is intracellular and seems to occur in most cases during the diploid phase (Figures 4 and 5).

Holococcoliths: These are characterized by small crystal units organized in circular or oval aggregates. The calcification seems to be extracellular during the haploid phase (Figures 3 and 4).

Nannoliths: Calcitic plates different from the hetero- and holococcoliths are grouped under this classification. In general, they correspond to simple crystal structures that originally occupied internal or external positions in the cell, formed under different biomineralization processes, and may include, in some cases, forms not linked to coccolithophores (Figures 5 and 6).

In most cultivated species, the dominant mechanism of reproduction is asexual by binary division (mitotic); each new cell shares the coccoliths previously generated by its parents. However, in other cases, including some of the most abundant taxa (e.g., *Emiliania huxleyi*, *Helicosphaera carteri*, and *Coccolithus pelagicus*), a more complex cycle is observed. The succession of diploid ($2n$) and haploid (n) phases generates heterococcolith and holococcoliths (plus organic scales), respectively. The origin of nannoliths is diverse and is not well known at present (Figure 6).

The most common and diverse types are heterococcoliths. Three main morphological types have been distinguished, although these types are not taxonomically significant (Figure 6; Young et al., 2003):

Placolith: A disc-shaped coccolith with two shields connected by a tube

Murolith: A bowl-shape coccolith with a dilated central area bounded by a wall of variable height

Planolith: A disc-shaped coccolith with a single shield

Modifications of these generic types at the periphery (circular, elliptical, or polygonal), or the presence of bars, spines, or stems are the criteria used for taxonomy.

Both the reproductive mechanisms and the morphology of some taxa seem to be linked to environmental features, opening new possibilities to explore their use as tracers.

Ecology

As autotrophic (photosynthesizing) organisms that live in the uppermost part of the water column (photic zone), light intensity is the main limiting factor for coccolithophore proliferation (Figure 7). Other factors, such as temperature, nutrient contents or salinity, as well as competition with other organisms, also control their distribution. It is difficult to consider these factors separately because they interact continuously with each other and with the physicochemical properties of water masses, such as upwelling, stratification, and turbidity.

Light and Temperature

Luminosity directly affects the development and characteristics of coccolithophore assemblages. This parameter, however, must be considered from different points of view: first, latitudinal

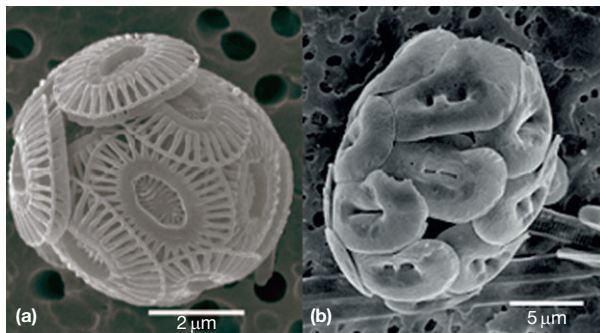


Figure 1 Coccospheres of *Emiliana huxleyi* (a) and *Helicosphaera carteri* (b). Photo courtesy of Lluisa Cros.

control, due to differences in the intensity of solar radiance, as well as in the seasonal (annual and daily) cycles; second, water-depth control, related to the differential absorption of wavelengths. These factors are closely related to sea-surface temperature, which also has a latitudinal and water-depth gradient (Thierstein and Young, 2004; Winter and Siesser, 1994).

Light penetration and seasonality

The progressive attenuation of light in the water column, as a consequence of the absorption of different wavelengths, affects coccolithophore assemblages. Most species live in the upper photic zone (UPZ; 100–10% of light), corresponding to the first 50 m (Figure 7). A reduced number of species are specialized to use low wavelength light (blue and violet) and grow in the middle and lower photic zone (MPZ and LPZ; less than 1% of light), which may reach 200 m water depth. This is the case of *Oolithotus* spp. and *Florisphaera profunda*, inhabitants of the MPZ and LPZ, respectively (low light intensity at shallower depths may be caused by an increase in turbidity due to high terrigenous sedimentation, the presence of volcanic ash, or excess biological production, causing an increase in organisms normally living in the LPZ in the UPZ). As with light intensity, temperature also decreases with depth and, as a result, LPZ inhabitants grow in a colder environment in any given latitude (Figure 7).

Latitude

In general terms, coccolithophores are widespread in the ocean, extending from the tropics with temperatures over 28 °C, as in the Gulf of Mexico, to the cold waters in the Pacific sector of the Southern Ocean, with temperatures around 2 °C. Day length, combined with low temperature, is the factor

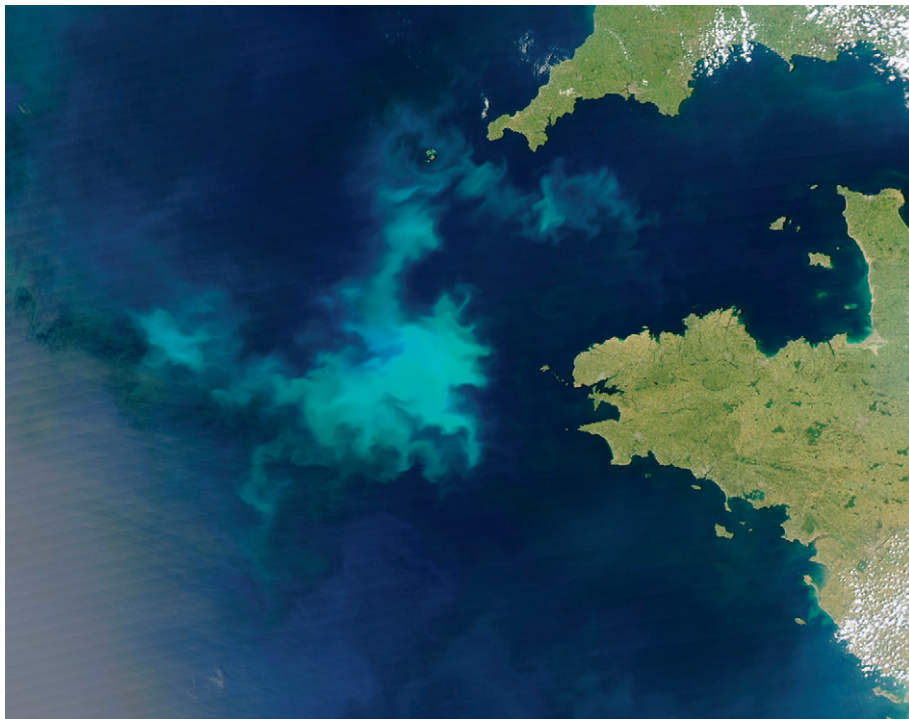


Figure 2 Coccolithophore bloom (light green) in the North Atlantic, close to France and United Kingdom. Provided by the SeaWiFS Project, NASA/Goddard Space Flight Center, and ORBIMAGE (<http://oceancolor.gsfc.nasa.gov/>).

Table 1 Extant coccolithophore classification

Kingdom Chromista	
Division Haptophyta	
Class	
Prymnesiophyceae	
Order Isochrysidales	
Family Isochrysidaceae	
Noelaerhabdaceae	<i>Emiliania</i> , <i>Reticulofenestra</i> , <i>Gephyrocapsa</i>
Coccosphaerales	
Occolithaceae	<i>Coccolithus</i>
Calcidiscaceae	<i>Calcidiscus</i> , <i>Umbilicosphaera</i> , <i>Oolithotus</i> , <i>Hayaster</i>
Pleurochrysidaceae	
Hymenomonadaceae	
Zygodiscals	
Helicosphaeraceae	<i>Helicosphaera</i>
Pontosphaeraceae	<i>Pontosphaera</i> , <i>Scyphosphaera</i>
Syracosphaerales	
Syracosphaeraceae	<i>Syracosphaera</i> , <i>Coronosphaera</i>
Calciosoleniaceae	<i>Calciosolenia</i>
Rhabdosphaeraceae	<i>Rhabdosphaera</i> , <i>Discosphaera</i> , <i>Acanthoica</i> , <i>Algirosphaera</i>
Heterococcoliths	
<i>incertae sedis</i>	
Alisphaeraceae	
Umbellosphaeraceae	<i>Umbellosphaera</i>
Papposphaeraceae	<i>Papposphaera</i>
Nannoliths	
Braarudosphaeraceae	<i>Braarudosphaera</i>
Ceratolithaceae	<i>Ceratolithus</i> ^a , <i>Neosphaera</i> ^a
<i>Incertae sedis</i>	<i>Florisphaera</i> , <i>Gladiolithus</i>

^aDifferent life-cycle phases of same taxon.

Source: Based on Young J, Geisen M, Cros L, et al. (2003) Special issue: A guide to extant Coccolithophore taxonomy. *Journal of Nannoplankton Research* 1:1–121.

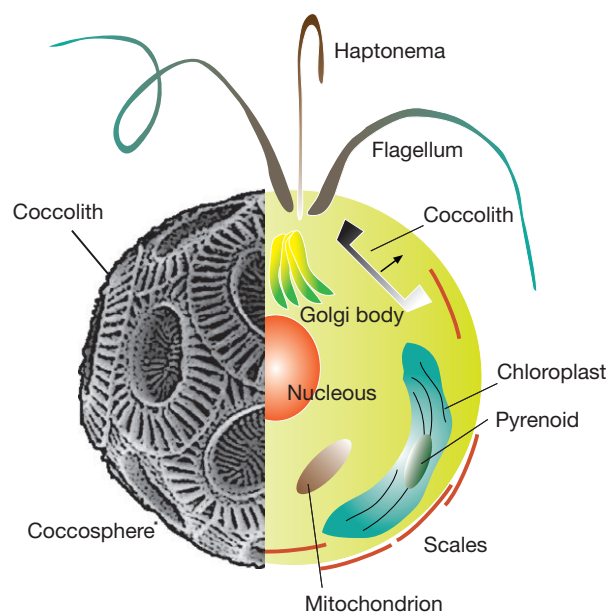


Figure 3 Schematic representation of a generic coccolithophore cell, as well as position of scales and/or coccoliths (nonmotile (left) and motile (right) phases). Type of coccolith, cytological characteristics, and definition of organelles vary in different taxa.

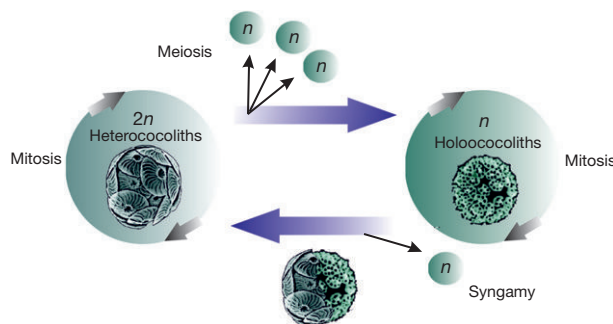


Figure 4 Generalized life cycle of a coccolithophore. Haploid: cells that contain only one copy of each chromosome (haploid number = n). Diploid: cells that have two sets of chromosomes, one from each parent (diploid number = $2n$). Holococcoliths: coccoliths constituted by small crystal units organized in circular or oval aggregates. Heterococcoliths: coccoliths constituted by calcitic units of varying shapes and sizes and ordered in cycles. Modified from Geisen M, Billard C, Broerse ATC, Cros L, Probert I, and Young JR (2002) Life-cycle associations variation or cryptic speciation. *European Journal of Phycology* 37: 531–550; Young JR, Geisen M, Cros L, et al. (2003) Special Issue: A guide to Extant Coccolithophore Taxonomy. *Journal of Nannoplankton Research* 1: 1–125.

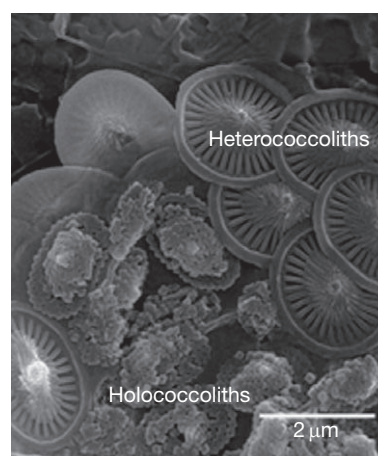


Figure 5 Holococcoliths and heterococcoliths of *Acanthoica quattropsina*. Photo courtesy of Lluïsa Cros.

affecting coccolithophore assemblages in high latitudes, where the assemblages include only a few euryhaline and eurythermal (cosmopolitan) species. Toward lower latitudes, assemblages become more diverse, with more specialized populations.

Traditionally, different biogeographic regions have been recognized from the Arctic to the Antarctic oceans, related to the most extensive surface water masses: subantarctic, temperate (or transitional), subtropical (central), tropical (equatorial), and subantarctic (Figure 8). Each biogeographic zone is characterized by a coccolithophore assemblage, where cosmopolitan species live together with other geographically restricted forms (McIntyre and Bè, 1967).

This is, however, a simplistic picture that only provides a general idea about coccolithophore distribution; particular features, such as eddies, upwelling regions, or closed basins, are not considered, and coccolithophore assemblages in these environments may be significantly different. On the other hand, there is

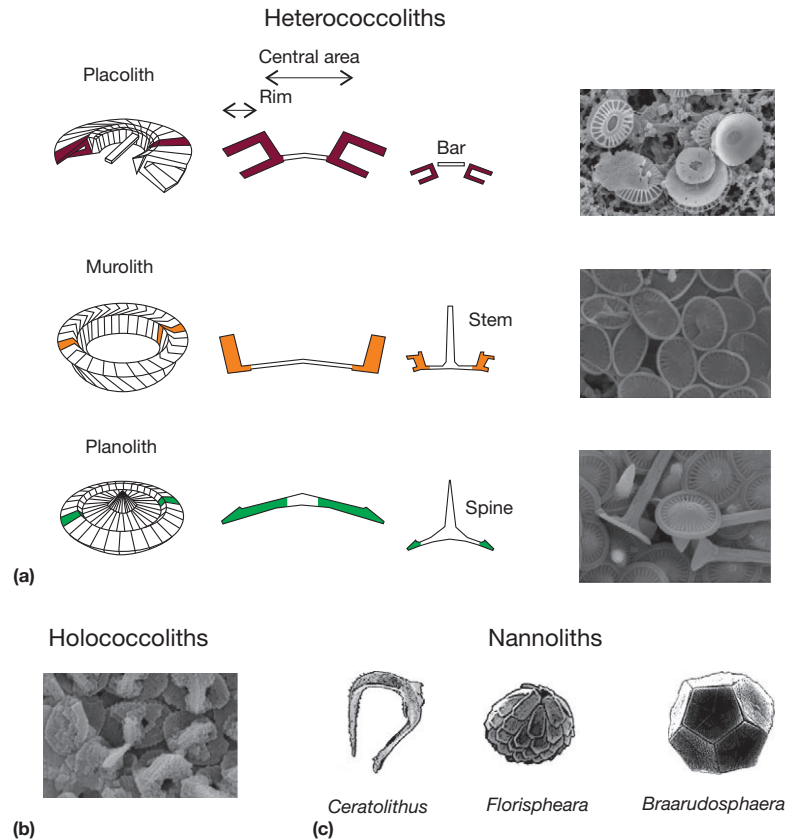


Figure 6 Structure and morphology of (a) heterococcoliths, (b) holococcoliths, and (c) nannoliths. Modified from Young JR, Geisen M, Cros L, et al. (2003) Special Issue: A guide to extant coccolithophore taxonomy. *Journal of Nannoplankton Research* 1: 1–125, Photo courtesy of José Gravalosa.

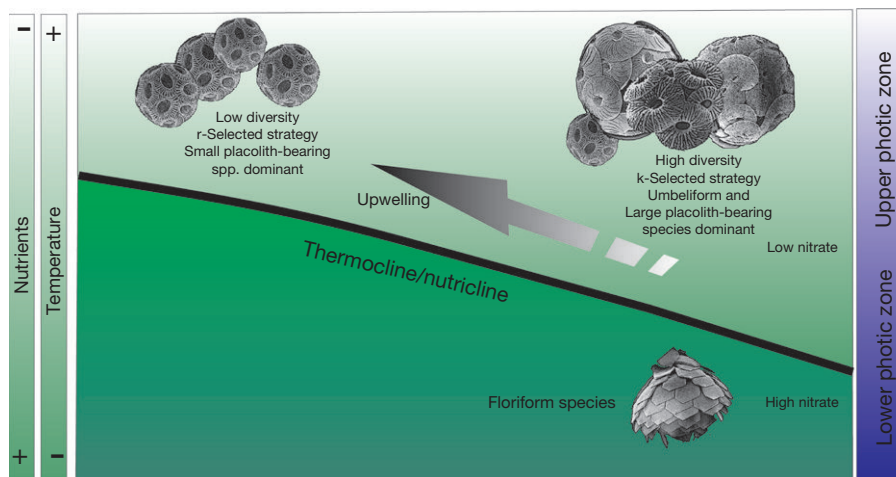


Figure 7 Distribution of coccolithophores in the ocean water column and their relationship with oceanographic and physical properties.

a significant degree of hemispheric and ocean provincialism, which is especially patent in the distribution of some species, and this can introduce noteworthy modifications.

Nutrients

As part of the marine phytoplankton, the availability of bio-limiting nutrients, such as phosphorus and nitrogen (in the form of phosphates and nitrates), is fundamental for

coccolithophore development and diversification. Both nutrients act directly on cellular constitution and calcification, but in environments with high nutrient concentrations, as in the case of upwellings, the proliferation of other organisms (e.g., diatoms that are better adapted to proliferating in these scenarios) competes with that of coccolithophores, limiting their growth. The phytoplankton community responds to increasing levels of nutrient availability following a biological succession (Margalef, 1978) that has been defined as follows

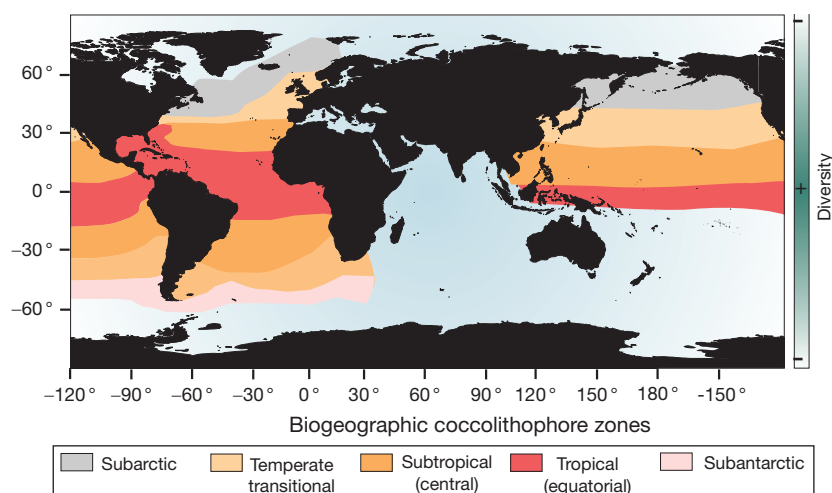


Figure 8 Biogeographic zonation for the Atlantic and Pacific oceans. Modified from McIntyre A and Bé AWH (1967) Modern Coccolithophoridae of the Atlantic Ocean I: Placoliths and Cyrtoliths. *Deep-Sea Research* 14: 561–597.

(Figure 8): in regimes of high turbulence and high nutrient contents (eutrophic), diatoms are the dominant organisms; coccolithophores are also abundant, developing an r-selected ecological strategy (opportunistic; growth rate maximized). In contrast, under conditions of low-turbulence and low-nutrient contents (oligotrophic), dinoflagellates are proportionally more abundant. In between these conditions (mesotrophic), coccolithophores are, in relative terms, the most abundant organisms, developing a k-selected ecological strategy (specialist; efficiency maximized). Biological succession is a generic concept that explains a general behavior, but particular aspects, such as physiological characteristics of some organisms or local variations, have not yet been considered in the issue. Furthermore, it is important to consider that, in the absence of strong competition, environments with high nutrient contents favor the production of all organisms, and the succession occurs when some of the nutrients become depleted, or are present in low proportions (e.g., Si, for diatoms), but nitrates are abundant (Figure 9).

In this succession, two particular situations are interesting: one is the high nutrient concentration scenario in low-turbulence regimes that elicits red tides (by dinoflagellates). The second situation occurs in low-turbulence regimes, with moderate-to-low nutrient contents; to this, a third factor should be added: long day length, causing coccolithophore blooms. This phenomenon is well known in *Emiliana huxleyi*, which produces spectacular blooms that can even be observed from space (Figure 2). Similar mechanisms persisting over time can be invoked to explain the high concentration of species morphologically similar to *Emiliana huxleyi* in the middle and late Pleistocene, such as those observed for the 'small' *Gephyrocapsa* (<3 μm) and *Gephyrocapsa caribbeana* (Figure 10).

In conclusion, the abundance and characteristics of coccolithophore assemblages are directly linked to the stability of the mixed layer (turbulence), the nutrient content, and light penetration. The thickness of the mixed layer in the ocean

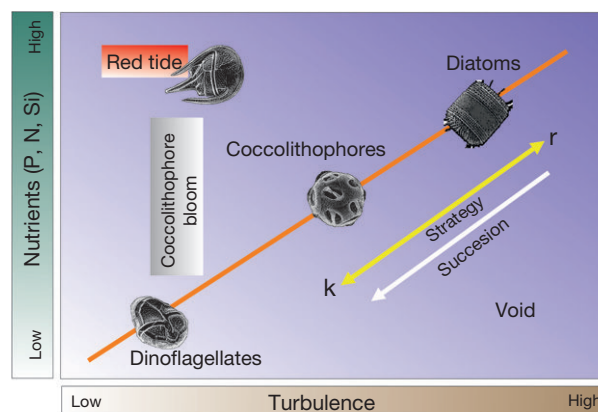


Figure 9 Relationship between organisms, nutrients, and turbulence: biological succession. The behavior of coccolithophores is considered generically; some species can respond in different ways. Coccolithophore blooms are referred to *Emiliana huxleyi* (Balch, 2004). Modified from Margalef R (1978) Life-forms of phytoplankton as survival alternatives in an unstable environment. *Oceanologica Acta* 1: 493–509.

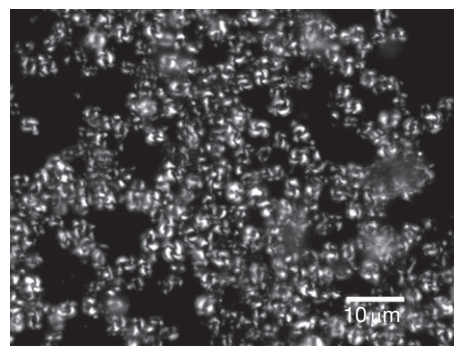


Figure 10 Petrographic microscope photography of *Gephyrocapsa caribbeana* (ODP Site 1089) Atlantic Ocean, during MIS 11.

determines the position of the thermocline (pycnocline), which is also linked to the nutricline (depth of maximum rate of change in the concentration of nutrients). The position of the nutricline in the water column controls the coccolithophore assemblage: a deep nutricline situated in the LPZ, in accordance with a stratified and stable oceanic environment, is characterized by the dominance of LPZ coccolithophore species. Alternatively, a shallow thermocline (and nutricline) is consistent with the development of a UPZ coccolithophore assemblage. The quantification of these nanoflora can be used to reconstruct past variations in the position of those parameters, which in turn have been controlled by atmospheric dynamics (Molfinio and McIntyre, 1990).

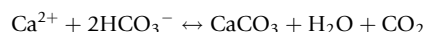
Three main groups of coccolithophores can be distinguished, based on the stability of oceanic environments (Figure 7; Okada and Honjo, 1973; Winter and Siesser, 1994):

1. The placolith-bearing assemblage dominates eu- and mesotrophic environments. Small cells (e.g., *Emiliania huxleyi*) are relatively more abundant in upwelling regimes, whereas large cells (e.g., *Calcidiscus*, *Umbilicosphaera*...) are more frequent in the open-ocean, in mid-latitudes.
2. The umbelliform assemblage is characteristic of oligotrophic regimes, especially in subtropical gyres. This assemblage occurs with large placolith-bearing and abundant discoliths and muroliths (e.g., *Pontosphaera*, *Rhabdosphaera*, or *Syracosphaera*).
3. The floriform assemblage, abundant in low and mid-latitudes, occupies the LPZ (*Florisphaera profunda*) and MPZ (*Oolithotus fragilis*, not properly a floriform species).

Although some of these assemblages have latitudinal restrictions, an estimation of their ratios can be used to monitor past oceanographic changes.

Coccolith Production and Accumulation Mechanisms

Around 60% of the total oceanic calcium carbonate comes from coccoliths. The production of these plates (see the section 'Biology') is linked to photosynthesis, which provides the energy required to transport inorganic carbon and calcium inside the cell according to the expressions:



During this process, CO_2 is released. However, the generation of organic matter involves the absorption of CO_2 , according to:

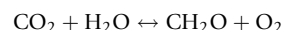


Figure 11 summarizes the existing relationship between production of organic matter and calcium carbonate precipitated by coccolithophores. The uptake of Ca for shell construction and of CO_2 for the generation of cellular and skeletal matter controls the coccolithophore $\text{CaCO}_3/\text{C}_{\text{org}}$ rain ratio.

As shown by the equations (Figure 11), coccolith production affects the concentration of CO_2 . The production of organic matter takes up CO_2 (the organic carbon pump), while the production of calcium carbonate releases CO_2 (carbonate pump). The sinking coccoliths remove organic carbon and

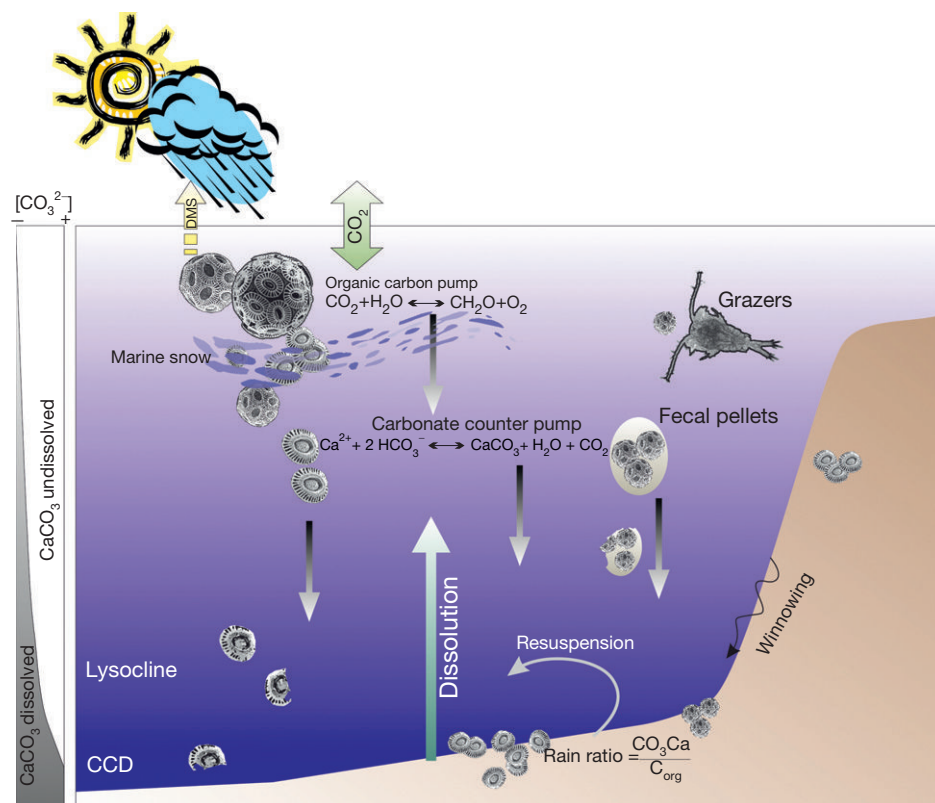


Figure 11 Relationship between processes of production and dissolution of calcium carbonate and coccolithophore action.

carbonate carbon from the surface ocean in a proportion of approximately 5:1 (Corg/CaCO₃), termed the rain ratio.

Once cells have died, disaggregated coccospheres descend through the water column. In theory, in a quiet environment without lateral water movements, the sinking rate of coccolith-sized particles of about 3 µm is estimated at 14 cm per day, implying hundreds of years to reach the sea floor. However, coccoliths are affected by two processes that considerably increase their sedimentation rate (Figure 11). First, grazers ingest high amounts of coccolithophores, which, after digestion, are expelled as fecal pellets, reaching a critical mass that increases their sedimentation rate; in addition, some of these pellets form an organic capsule that protects coccoliths from corrosive deep waters (Honjo, 1976). Second, 'marine snow' (the continuous shower of mostly organic detritus that falls through the water column) agglutinates organic and inorganic particles to form high-density aggregates that reach the sea floor at rates close to 200 m per day (Thierstein and Young, 2004).

Although carbonate dissolution occurs at all depths, it is especially significant after the first thousand meters. The carbonate ion concentration ([CO₃²⁻]) is larger in the sea surface and decreases with water depth. This pattern, combined with the temperature and pressure dependence of the saturation concentration, results in the upper ocean being everywhere supersaturated, and the deep ocean being progressively undersaturated with increasing water depth. Calcium carbonate dissolves when waters are subsaturated in calcite or aragonite. The water depth at which calcite subsaturation starts is called the calcite lysocline, while aragonite subsaturation usually occurs at much shallower depths. Below the calcite lysocline, carbonate shells and fragments are still present in the sediments, but calcite dissolution is complete at the calcite compensation depth (CCD). During sedimentation, coccoliths are affected by dissolution, especially if they sink in subsaturated waters. The effect of these waters depends on the length of time the coccoliths remain in contact with the corrosive waters, the effectiveness of protection systems (e.g., pellets), and the coccolith structure. Rapid sedimentation and burial prevent (or at least diminish) dissolution, whereas coccoliths constituted by large crystals are more resistant than those formed by small structures.

Holococcoliths are easily dissolved forms, while heterococcoliths (with large elements) and monocrystalline nannoliths are the most resistant. The study of coccolith dissolution is a potential tool, not much used, to monitor past variations in the CCD. Taking into account that calcium carbonate dissolution is linked to atmospheric, oceanic, and biosphere dynamics, monitoring dissolution can help to understand global processes such as CO₂ variability in the global system (Thierstein and Young, 2004).

Large regions of the sea floor are covered by nannofossil oozes, where the coccolith (or nannofossils, coccolithophore-related fossil organisms) concentration constitutes over 90% of the sediment. In other lithologies, coccoliths are diluted by terrigenous or other biogenic components. Their absence or low abundance in deep-ocean sediments (for mid- and low latitudes) is usually related to dissolution or winnowing processes. The persistence of bottom currents over biogenic sediments can produce the loss of tiny forms (coccoliths) and the enrichment of coarse microfossils (e.g., planktonic foraminifers). The winnowing of the small size forms favors the

remobilization and ultimate creation of an allochthonous assemblage, causing changes in the original coccolithophore biocenosis (biological community). Consequently, a preliminary evaluation of these factors is crucial to understanding the sedimentary record. Additionally, paleoceanographic studies are based on direct comparisons between present-day data recovered from living assemblages or from sediment traps affected by seasonality, and on the coccolithophores from surface sediments. Most authors agree that in the absence of dissolution, reworking or mixing, the surface sediment records mean annual values.

In addition to the particular significance of certain species and their potential for monitoring variations in productivity, a direct correlation between the coccolithophore accumulation rate and productivity proxies can be observed. In oceanic settings not affected by dissolution, the coccolithophore accumulation rate is a good proxy for paleoproductivity (Flores et al., 2003).

Coccolithophores as Paleoceanographic Indicators

The study of coccolithophores in sediment cores provides useful information about the location of surface water masses. Modification of the assemblages and the geochemical composition of the shell by environmental features such as SST, distribution, and nutrient abundance, and the relationship of these with oceanic and atmospheric processes, is the basis for using coccolithophores as paleoceanographic proxies. Even though coccoliths are abundant in the sedimentary record, this group has been little used for paleoceanographic reconstructions. The cosmopolitanism of the most abundant species, as well as their rapid evolutionary trends, prevents the application of actualistic methods. Curiously, however, this explains their success in biostratigraphy (Bown, 1998).

Reconstruction of Latitudinal Variations and Temperatures

In a general sense, global climatic changes are manifested in the ocean as variations in the position of front-limiting surface water masses, which, as discussed above, have their characteristic coccolithophore assemblages. The identification of these assemblages in the sedimentary sequence allows past spatial changes in these fronts to be monitored. One of the first attempts was made in the North Atlantic, in an attempt to trace the northward movement of the polar front over the last 20 000 years (McIntyre, 1967): Subarctic assemblages were found in sediment cores at latitudes between 42° and 45° N, marking the southernmost penetration of the north polar front during glacial periods. This simple technique permitted an orbital-related signal to be identified that was consistent with glacial–interglacial pulses of different intensities. However, the availability of high-resolution sedimentary records has shown that coccolithophores can help detect abrupt and short-term rapid changes in SST. Figure 12 shows the results obtained in a sediment core from the western Mediterranean, where Heinrich and Dansgaard–Oeschger stadial cold events were detected using the cold water morphotype of *Emiliania huxleyi* (>4 µm; Colmenero-Hidalgo et al., 2004; Sierro et al., 2005).

With the improvement of biogeographic referential patterns, as well as the advent of new data about the behavior of

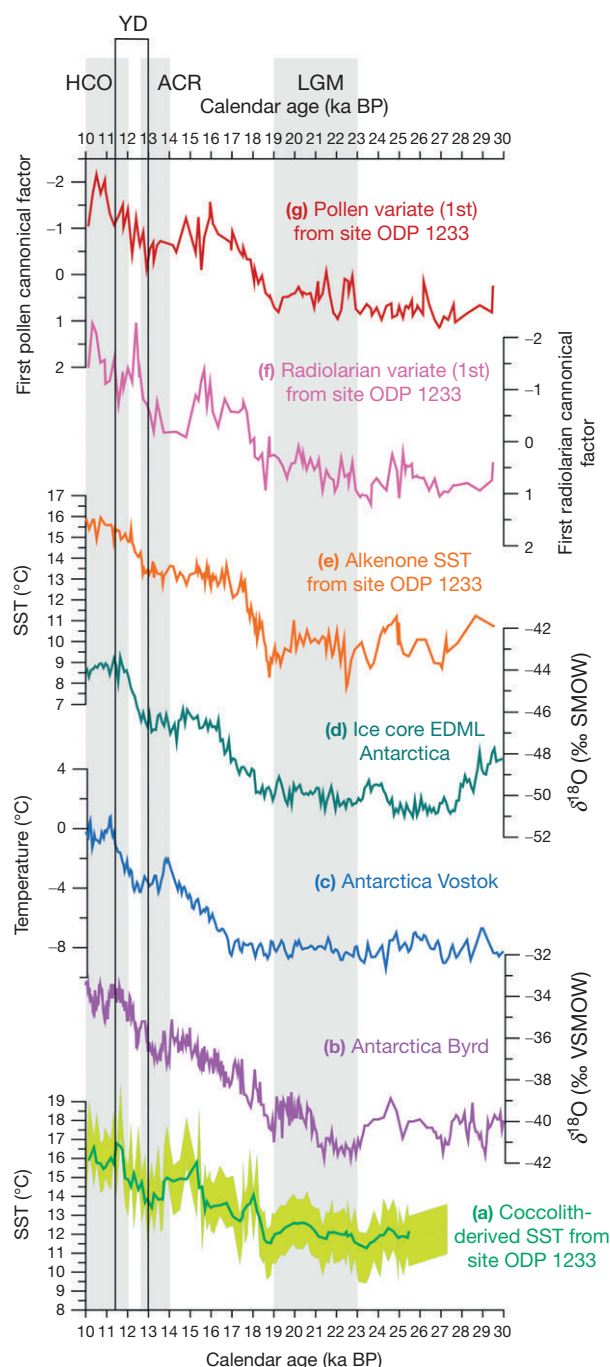


Figure 13 Down-core paleotemperature obtained in ODP Site 1233 situated in the NE coast of South America by a coccolithophore transfer function, and its comparison with results obtained with biogeochemical and micropaleontological techniques. Reproduced from Saavedra-Pellitero M, Flores JA, Lamy F, Sierro FJ, Cortina A (2010) Coccolithophore estimates of paleotemperature and paleoproductivity changes in the Southeast Pacific over the past ~27 ka. *Paleoceanography*, <http://dx.doi.org/10.1029/2009PA001824>.

as a paleoproductivity proxy, useful in the reconstruction of Pleistocene sequences (Stoll et al., 2002).

Oceanic productivity is related to the characteristics of the water masses and the ability of the ocean in this region to distribute nutrients in the photic layer. A stable stratified

system, such as in subtropical gyres, is coherent with a deep thermocline/nutricline, favoring the development of inhabitants of the LPZ. The opposite scenario occurs in upwelling regions, with a narrow mixed layer and a shallow nutricline, where the UPZ inhabitants are the dominant forms. According to this pattern, the ratio between LPZ and UPZ offers the possibility of monitoring fluctuations in the position of the thermocline (Molfino and McIntyre, 1990). Additionally, comparisons with present-day data allow the development of transfer functions to estimate paleoproductivity and the position of the thermocline, which in turn is related to wind intensity and direction. Application of this technique in the equatorial Indian and Pacific oceans has led to the identification of precessional-controlled variations in the position of the thermocline, linked to ENSO oscillations for the last climatic cycle. A deep nutricline is associated with frequent El Niño-like scenarios during interglacials, whereas intervals of high productivity correspond to a shallow (and/or a lower gradient) nutricline during glacials. It has been demonstrated that this oscillation is directly controlled by insolation and is independent of global ice volume variations (Beaufort et al., 2001).

Carbon isotope values may be used for fertility reconstructions in the ocean. Preliminary $\delta^{13}\text{C}$ values measured in coccoliths seem to reflect the seawater composition, and consequently surface water productivity, although additional studies are needed and should be applied in sediment series. Recently, other geochemical proxies, such as the Sr/Ca ratio, have been carried out using coccoliths. This ratio is considered a good indicator of surface water fertility when applied to monospecific samples of *Emiliania huxleyi* (Stoll et al., 2002).

How Do Coccolithophores Affect the Climate System?

As primary producers, coccolithophores are important contributors to the biological pump that removes CO_2 from surface ocean waters; however, they also remove significant amounts of calcium carbonate. Calcite burial and dissolution in deep-sea sediments have important impacts on the CO_2 cycle. The balance between organic matter and calcium carbonate production determines the $\text{CaCO}_3/\text{C}_{\text{org}}$ rain ratio, a parameter that can be used to estimate the effectiveness of the process (Fischer and Wefer, 1999). A high accumulation of calcium carbonate in sediments contributes to increase the relative proportion of CO_2 that, as stated above, can be balanced by a high production of organic matter in surface waters (CO_2 consuming). This process also depends on the contribution of other planktonic organisms (e.g., diatoms and foraminifers).

During the Pleistocene, episodes of high coccolith production have been recognized, such as one reconstructed for the warm interglacial episode of MIS 11. At that time, a single species, such as the small but robust placolith *Gephyrocapsa caribbeanica*, was dominant in the ocean (Figures 10 and 15) (Bollmann et al., 1998; Flores et al., 2003). These episodes of inferred high coccolith production are coincident with intense dissolution at the deepest regions of the sea floor and a rising CCD.

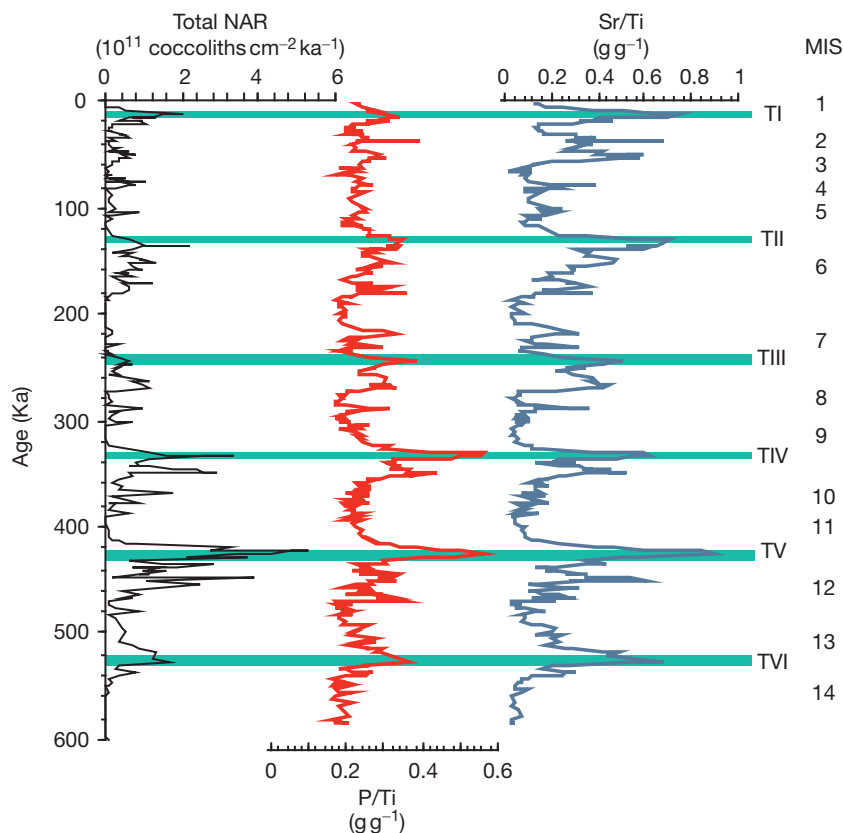


Figure 14 Relationship between total accumulation rate of coccolithophores (NAR = nannofossils accumulation rate) and some paleoproductivity proxies (P/Ti and Sr/Ti) in ODP Site 1089 (Atlantic Southern Ocean), for the last 500 kys. T represents terminations, and MIS marine isotope stages (Flores and Filippelli, unpublished).

Additionally, coccolithophores, like other phytoplanktonic organisms, are producers of dimethyl sulfide (DMS). When released to the air, this compound is the primary source of the sulfur particles that stimulate the formation of clouds. An increase in cloud formation by enhanced DMS emission can cause an increase in the albedo (the Earth's reflectivity) which can affect the climate system (Thierstein and Young, 2004). Further studies on this topic, both in present-day and past environments, are needed.

Pleistocene Biostratigraphy and Biochronology

Age control is the first step in any paleoceanographic study. Geochronological, geomagnetic, and geochemical techniques currently provide the main framework for obtaining time-reference points. However, sometimes, these techniques are time consuming and are not always easy to apply (e.g., in routine studies in oceanographic vessels), and alternative techniques such as biochronology (identification of time-calibrated biological events) are needed. Coccolithophores (and biologically related calcareous nannofossils) are an abundant

and widespread oceanic micropaleontological group, with a high evolutionary rate that generates identifiable changes in taxa and in assemblages, which are easy to identify using petrographic and electron microscopy techniques. These characteristics have played a key role in the success of the group in biostratigraphy (Figure 15).

Standard biostratigraphic patterns are restricted, offering low-resolution solutions. Over the last decade, exhaustive studies have allowed the identification of different events, calibrated by comparison with geomagnetic and oxygen isotope (orbital) time scales. The parallel correlation of biostratigraphic, paleomagnetic, and isotope events allows us to evaluate whether or not they occurred synchronously in different regions, together with their relationship with environmental parameters (Wei, 1993; Raffi, 2002; Colmenero-Hidalgo, 2004; Flores et al., 2003; Raffi et al., 2006).

As already discussed, coccolithophore assemblages are latitudinally controlled, and hence, this may limit the usefulness of some events for global correlations. In addition, some species, morphotypes, or assemblage characteristics may be different in particular regions or oceanic settings, such as in upwelling or restricted basins, and a general pattern is difficult to apply.

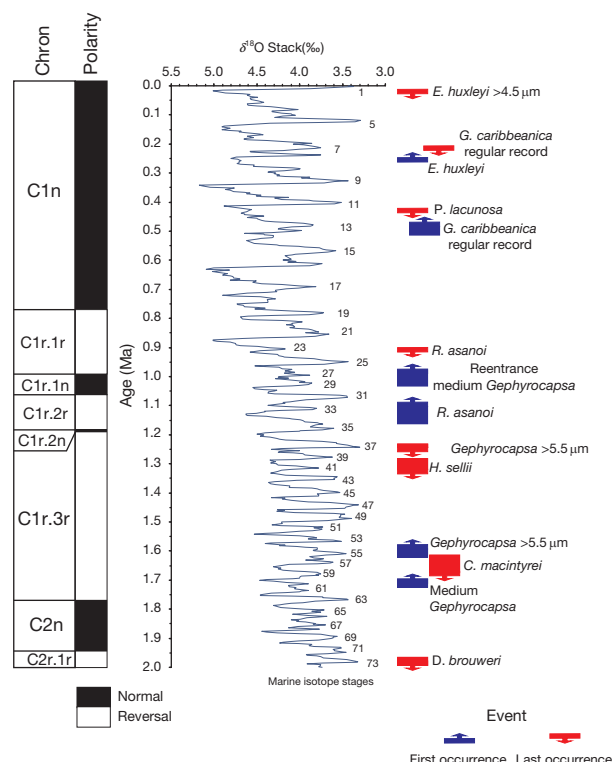


Figure 15 Compilation of the most significant biostratigraphic coccolithophore events and their geochronological calibration. The thickness of the colored bars represents the interval of definition of each biostratigraphic event (diachronism). Reproduced from Wei W (1993) calibration of Upper Pliocene–Lower Pleistocene nannofossil events with oxygen isotope stratigraphy. *Paleoceanography* 8: 85–99; Raffi I (2002) Revision of the early–middle pleistocene calcareous nannofossil biochronology (1.75–0.85 Ma). *Marine Micropaleontology* 45: 25–55; Flores JA, Marino M., Sierro FJ, Hodel DA, and Charles CD (2003) Calcareous plankton dissolution pattern and coccolithophore assemblages during the last 600 ka at ODP Site 1089 (Cape Basin, South Atlantic): Paleoceanographic implications. *Palaeogeography, Palaeoclimatology, Palaeoecology* 196: 409–426; Raffi I, Backman J, Fornaciari E, Pälike H, Rio D, Lourens L, and Hilgen F (2006) Calcareous nannofossil astrochronology encompassing the past 25 million years. *Quaternary Science Reviews* 25: 3113–3137; Flores JA, Colmenero-Hidalgo E, Mejía-Molina A, Baumann KH, Henderiks J, Larsson K, et al. (2010) Distribution of large *Emiliania huxleyi* in the Central and Northeast Atlantic as a tracer of surface ocean dynamics during the last 25 000 years. *Marine Micropaleontology* 76: 53–66.

See also: Diatom Introduction; Varved Marine Sediments. **Carbonate Stable Isotopes:** Overview. **Glacial Climates:** Biosphere Feedbacks; **Paleobotany:** Overview of Terrestrial Pollen Data. **Paleoceanography:** Paleoceanography An Overview. **Paleoceanography, Biological Proxies:** Biomarker Indicators of Past Climate; Dinoflagellates; Marine Diatoms; Planktic Foraminifera; Radiolarians and Silicoflagellates; Temperature Proxies, Census Counts. **Paleoceanography, Physical and Chemical Proxies:** Carbon Cycle Proxies ($\delta^{11}\text{B}$, $\delta^{13}\text{C}_{\text{calcite}}$, $\delta^{13}\text{C}_{\text{organic}}$, Shell Weights, B/Ca, U/Ca, Zn/Ca, Ba/Ca); Mg/Ca and Sr/Ca Paleothermometry from

Calcareous Marine Fossils; Nutrient Proxies; Oxygen-Isotope Stratigraphy of the Oceans. **Paleoclimate:** Modern Analog Approaches in Paleoclimatology. **Paleoclimate Modeling:** Paleo-ENSO; Quaternary Environments. **Quaternary Stratigraphy:** Chronostratigraphy; Climastratigraphy.

References

- Balch WM (2004) Re-evaluation of the physiological ecology of coccolithophore. In: Thierstein H and Young JR (eds.) *Coccolithophores. From Molecular Processes to Global Impact*, pp. 165–190. Berlin: Springer.
- Beaufort L, Garidel-Thoron T, Mix AC, and Pisias NG (2001) ENSO-like forcing on oceanic primary production during the late Pleistocene. *Science* 293: 2440–2444.
- Bollmann J, Baumann KH, and Thierstein H (1998) Global dominance of *Gephyrocapsa* coccoliths in the late Pleistocene: Selective dissolution, evolution, or global environmental change? *Paleoceanography* 13: 517–529.
- Bown PR (ed.) (1998) *Calcareous Nannofossil Biostratigraphy*. Cambridge: Chapman and Hall.
- Colmenero-Hidalgo E, Flores J-A, Sierro FJ, et al. (2004) Ocean surface water response to short-term climate changes revealed by coccolithophores from the Gulf of Cadiz (NE Atlantic) and Alboran Sea (W Mediterranean). *Palaeogeography, Palaeoclimatology, Palaeoecology* 205(3–4): 317–336.
- Fischer G and Wefer G (eds.) (1999) *Use of Proxies in Paleoceanography, Examples from the South Atlantic*. Berlin, Heidelberg: Springer-Verlag.
- Flores J-A, Marino M, Sierro FJ, Hodel DA, and Charles CD (2003) Calcareous plankton dissolution pattern and coccolithophore assemblages during the last 600 kyr at ODP Site 1089 (Cape Basin, South Atlantic): Paleoceanographic implications. *Palaeogeography, Palaeoclimatology, Palaeoecology* 196: 409–426.
- Geisen M, Billard C, Broerse ATC, Cros L, Probert I, and Young JR (2002) Life-cycle associations involving pairs of holococcolithophorid species: Intraspecific variation or cryptic speciation? *European Journal of Phycology* 37: 531–550.
- Green JC and Leadbeater BSC (1994) *The Haptophyte Algae*. Oxford: Clarendon Press.
- Haq BU (1983) *Calcareous Nannoplankton*. Stroudsburg, PA: Hutchinson Ross Publishing Co.
- Haq BU and Boersma A (eds.) (1978) *Introduction to Marine Micropaleontology*. New York: Elsevier.
- Honjo S (1976) Coccoliths: Production, transportation and sedimentation. *Marine Micropaleontology* 1(1): 65–79.
- Margalef R (1978) Life-forms of phytoplankton as survival alternatives in an unstable environment. *Oceanologica Acta* 1: 493–509.
- McIntyre A (1967) Coccoliths as paleoclimatic indicators of Pleistocene glaciation. *Science* 158: 1314–1317.
- McIntyre A and Bé AWH (1967) Modern Coccolithophoridae of the Atlantic Ocean I. Placoliths and Cyrtoliths. *Deep-Sea Research* 14: 561–597.
- Molffino B and McIntyre A (1990) Nutricline variations in the Equatorial Atlantic coincident with the Younger-Dryas. *Paleoceanography* 5(6): 997–1008.
- Okada H and Honjo S (1973) The distribution of oceanic coccolithophores in the Pacific. *Deep Sea Research* 20: 355–374.
- Raffi I (2002) Revision of the early-middle pleistocene calcareous nannofossil biochronology (1.75–0.85 Ma). *Marine Micropaleontology* 45: 25–55.
- Raffi I, Backman J, Fornaciari E, et al. (2006) Calcareous nannofossil astrochronology encompassing the past 25 million years. *Quaternary Science Reviews* 25: 3113–3137.
- Saavedra-Pellitero M, Flores JA, Lamy F, Sierro FJ, and Cortina A (2010) Coccolithophore estimates of paleotemperature and paleoproductivity changes in the Southeast Pacific over the past 27 kyr. *Paleoceanography* <http://dx.doi.org/10.1029/2009PA001824>.
- Sierro FJ, Hodel DA, Curtis JH, et al. (2005) Impact of iceberg melting on Mediterranean thermohaline circulation during Heinrich events. *Paleoceanography* 20: 1029–1051.
- Stoll H, Rosenthal Y, and Falkowski P (2002) Climate proxies from Sr/Ca of coccolith calcite: Calibrations from continuous culture *Emiliania huxleyi*. *Geochemistry and Cosmochimica Acta* 66(6): 927–936.

- Tappan H (1981) *The Paleobiology of Plant Protists*. New York: W.H. Freeman and Co.
- Thierstein H and Young J (eds.) (2004) *Coccolithophores. From Molecular Processes to Global Impact*. Berlin: Springer.
- Wei W (1993) Calibration of upper Pliocene-lower Pleistocene nannofossil events with oxygen isotope stratigraphy. *Paleoceanography* 8: 85–99.
- Winter A and Siesser W (1994) *Coccolithophores*. Cambridge: Cambridge University Press.
- Young JR, Bergen JA, Bown PR, et al. (1997) Guidelines for Coccolith and calcareous nannofossil terminology. *Palaeontology* 40(4): 875–912.
- Young J, Geisen M, Cros L, Kleijne A, Sprengel C, Probert I, et al. (2003) A guide to extant coccolithophore taxonomy. *Journal of Nanoplankton Research* (Sp. Issue 1): 1–121.

Relevant Websites

- <http://iodp.tamu.edu/curation/mrc.html> – Integrated Ocean Drilling Program. Micropaleontological Reference Centers (MRCs).
- <http://nannotax.org/> – Nannotax.
- http://www.nhm.ac.uk/hosted_sites/ina/CODENET/index.html – Coccolithophorid evolutionary biodiversity and ecology network – CODENET.
- http://www.nhm.ac.uk/hosted_sites/ina/terminology/index.htm – International Nanoplankton Association.
- <http://oceancolor.gsfc.nasa.gov/> – SeaWiFS Project, NASA/Goddard Space Flight Center, and ORBIMAGE.

Corals, Sclerosponges and Mollusks

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Glossary

δ : Standard reporting notation for light stable isotopes is delta (δ) notation. Delta values are calculated using the equation $(R_{\text{sample}}/R_{\text{standard}} - 1) \times 1000$, where R is the ratio of the heavy to light isotope (e.g., $^{18}\text{O}/^{16}\text{O}$).

‰: Standard unit of delta (δ) notation is per mil (‰; per thousand).

$\Delta^{14}\text{C}$: Standard reporting notation for radiocarbon (^{14}C) content, which is expressed as the per mil deviation from the $^{14}\text{C}/^{12}\text{C}$ ratio in 19th-century wood. Units are ‰.

$^{14}\text{C}_{\text{AMS}}$: Radiocarbon age determinations made using accelerator mass spectrometric techniques.

Introduction

The early identification and understanding of the temporal significance of growth increments in many calcareous skeletons, but especially those in mollusks and corals, facilitated their use as a means to study environmental variability in the Quaternary. Armed with this understanding, geochemical studies of the skeletons of mollusks and corals, and later of sclerosponges, began in earnest. These early investigations established many of the fundamental principles that now form the foundation of nearly all studies using the skeletons of corals, sclerosponges, and mollusks to study environmental and climate change. In this article, we provide a general overview of the use of corals, sclerosponges, and mollusks as archives of variability in the Quaternary Earth system.

Corals

Scleractinian or stony star corals belong to the subclass Zoantharia (Hexacorallia), in the class Anthozoa of the phylum Cnidaria. Scleractinia are polyp animals found exclusively in marine environments, which first appeared in the middle Triassic (~240 Ma). Scleractinians include solitary and colonial forms, many of which secrete a hard calcareous (aragonite) exoskeleton that provides support and protection for the polyps. Scleractinian corals form two ecological groups: reef builders (hermatypic) and non-reef builders (ahermatypic). Hermatypic corals, the best-known scleractinia, contain symbiotic dinoflagellates (autotrophic zooxanthellae) and inhabit shallow waters of the tropical oceans where they secrete massive, calcium carbonate skeletons that, along with calcareous algae, form the physical structure of coral reefs. Ahermatypic corals lack symbiotic algae (azooxanthellae) and hence are not restricted to shallow waters; they inhabit all regions of the ocean.

The corallite is the central macroarchitectural component of the aragonite skeleton of corals, as each corallite houses a polyp animal. A corallite may be thought of as a calcareous tube with vertical walls (theca), radiating vertical partitions

(septa), and horizontal layers (dissepiments) at the base of the coral polyp. As the surface of the corallite extends, new dissepiments are formed higher in the theca and old dissepiments are abandoned as the polyp raises itself up its theca. The living tissue, which is the site of calcification, occupies only a small fraction of the coral exoskeleton. Coral growth includes elongation (lengthening of the skeleton over time by extension) and thickening (mass addition of the skeleton over time). The interplay between extension and thickening results in variations in skeletal density and produces annual density bands in most hermatypic corals.

Quaternary environmental reconstructions using ahermatypic corals are in their nascent phase and hold great promise; however, the bulk of the literature is based on studies of massive hermatypic corals and this article will focus on these shallow-water corals. Optimal growth conditions for hermatypic corals include warm, shallow, and protected water (20–26 °C; 5–15 m), low sediment load, and normal ocean salinity. Massive corals are exceptional archives of late Quaternary environmental change in the tropics because: (1) they can be precisely dated (U-series; radiocarbon annual growth increments), (2) they are long-lived (tens to hundreds of years), (3) the geochemistry of their aragonite skeletons provides a record of temperature and salinity variations, and (4) the ecological distribution of some species permit robust reconstruction of sea level elevation. Corals that have proven particularly useful in Quaternary environmental reconstructions include massive *Porites* sp. corals from the Indo-Pacific, massive *Montastraea* sp. from the Atlantic, and to a somewhat lesser extent *Diploria* sp. from the Atlantic.

Patterns of annual density banding reflect variations in a number of environmental variables (e.g., temperature, water quality, light conditions, nutrient availability). These variations may be used to retrospectively monitor growth characteristics of massive corals and infer past environmental changes. The most commonly reported growth variable is linear extension rate, which can vary considerably within and between genera and individual coral colonies from the same reef. As an example, annual average extension rates vary from ~13 mm yr⁻¹ in massive *Porites* corals from the Great Barrier

Reef (GBR), to $\sim 8 \text{ mm yr}^{-1}$ in massive *Montastraea* corals from the Florida Keys, to $\sim 4 \text{ mm yr}^{-1}$ in massive *Diploria* corals from Bermuda. Early and subsequent work on growth variations in Bermuda corals demonstrated a link between skeletal extension and temperature and air pressure. Researchers in Australia have generated multicentury length skeletal extension records from a multitude of *Porites* heads from the GBR, Australia, and they have demonstrated that linear extension and calcification rates were significantly and directly correlated with latitudinal variations in average sea surface temperature. Ongoing work on growth variations in corals from the Caribbean and Gulf of Mexico holds significant promise.

A diverse record of environmental variability is also encoded into the geochemistry of the aragonite skeleton of massive, hermatypic corals. The vast majority of geochemical records generated over the years from corals have been stable oxygen ($\delta^{18}\text{O}$) and carbon ($\delta^{13}\text{C}$) isotopes. Early, seminal work demonstrated kinetic and metabolic fractionation results in the precipitation of coralline aragonite that is not in isotopic equilibrium with ambient seawater. The degree of isotopic disequilibrium differs amongst the different coral genera, with corals having lower $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values than inorganic aragonite precipitated from the same fluid as coralline aragonite. Despite this isotopic disequilibrium, which likely reflects biological mediation of the fluid involved in precipitation, a multitude of studies over the years have demonstrated that robust records of environmental variability can be constructed using stable isotopic variations in coralline aragonite, provided that the skeleton is properly sampled.

The $\delta^{18}\text{O}$ of coralline aragonite is primarily influenced by the sea surface temperature (SST) and $\delta^{18}\text{O}$ composition of seawater during skeletal precipitation. There is an inverse relation between coral $\delta^{18}\text{O}$ and temperature, such that when temperature increases, coral $\delta^{18}\text{O}$ values become more negative. The slope of the $\delta^{18}\text{O}_{\text{coral}}$ -SST relationship has been estimated by various workers to range from 0.18‰ per $^{\circ}\text{C}$ to 0.22‰ per $^{\circ}\text{C}^{-1}$, which is close to that determined in inorganic aragonite. Variations in $\delta^{18}\text{O}_{\text{seawater}}$ reflect changes in the volume of water stored in continental ice sheets and changes in the hydrologic balance (evaporation and precipitation). Increases (decreases) in continental ice volume result in increases (decreases) in $\delta^{18}\text{O}_{\text{seawater}}$. Precipitation (evaporation) results in decreases (increases) in $\delta^{18}\text{O}_{\text{seawater}}$. Time series of coral $\delta^{18}\text{O}$ variations, many of which have subannual resolution, have played a critical role in extending the instrumental record of tropical climate variability on timescales of seasons, years, decades, and centuries. These records have been used to constrain past behavior of coupled ocean-atmosphere systems including the Asian monsoon, the El Niño-Southern Oscillation (ENSO), Pacific Decadal Oscillation (PDO), and North Atlantic Oscillation (NAO). Coral $\delta^{18}\text{O}$ records have also been instrumental in understanding the behavior of the tropical climate system during times of changes in background climate states (e.g., glacial states).

Coral $\delta^{13}\text{C}$ variations have proven more difficult to interpret in terms of environmental variability. Skeletal $\delta^{13}\text{C}$ values reflect the two main carbon sources for corals: dissolved inorganic carbon (DIC) in seawater and zooplankton. Variations in skeletal $\delta^{13}\text{C}$ values reflect some combination of metabolic activities within the tissue layer (respiration and photosynthesis) and

heterotrophy (the consumption of other organisms). Environmental factors postulated to drive variations in skeletal $\delta^{13}\text{C}$ include light levels (as modulated by water depth, cloud cover, and/or seasonality), dietary changes (heterotrophy vs. autotrophy), and changes in seawater DIC.

The radiocarbon ($\Delta^{14}\text{C}$) in coral skeletons closely matches that of surrounding seawater DIC and coral $\Delta^{14}\text{C}$ records reflect changes in ocean circulation patterns and processes on all timescales, which are often driven by changes in climate (e.g., ENSO). The classical application of coral radiocarbon is the study of ocean circulation, which takes advantage of differences between surface and deep-water radiocarbon values to study vertical mixing and differences in the radiocarbon values of different source regions to study surface advection.

Variations in element-to-calcium ratios (e.g., Sr/Ca, Mg/Ca, U/Ca, Ba/Ca, Cd/Ca) in coral skeletons have also been used to reconstruct environmental variability, with various degrees of success. Coral Ba/Ca variations have been demonstrated to reflect changes in riverine input and upwelling in reef sequences of the GBR and the Galapagos, respectively. Coral Pb/Ca variations in a Bermuda coral record industrial pollution from the east coast of North America through Pb/Ca variations. A great deal of effort has been put forth in attempts to establish elemental ratio-based coral thermometer; U/Ca, Mg/Ca and Sr/Ca have been proposed. Most workers in the field agree that factors other than temperature affect coral U/Ca and Mg/Ca variations, which confounds their use as coral thermometers. Coral Sr/Ca is a most widely applied coral thermometer, although its use as a thermometer is not without critics. Coral Sr/Ca varies in response to changes in seawater temperature and seawater Sr/Ca. Many workers have argued for the constancy of seawater Sr/Ca on millennial timescales, based on residence time considerations; other workers have argued for variations in seawater Sr/Ca in the modern ocean based on select measurements. Some workers have argued for a strong biological control on coral Sr/Ca; whereas others have produced reproducible records of Sr/Ca from multiple coral heads, variations which have strong coherency with known climate phenomena such as ENSO and PDO.

Coral Sr/Ca determinations, when paired with coral $\delta^{18}\text{O}$ determinations made from the same sample, offer promise as a means to isolate the thermal and hydrologic components of the signal. The temperature component of the coral $\delta^{18}\text{O}$ is resolved using coral Sr/Ca, leaving a residual $\delta^{18}\text{O}$ record, which should reflect changes in seawater $\delta^{18}\text{O}$. Several of these coral-based, residual or calculated seawater $\delta^{18}\text{O}$ records have been produced. Some extend back several centuries and suggest that significant and sometimes abrupt changes in the hydrologic regime of the tropics was far more prevalent than previously thought.

Sclerosponges

Sclerosponges, also referred to as coralline sponges, are marine organisms with a dense skeleton that is primarily calcareous, with minor amount of siliceous spicules. Sclerosponges are calcified demosponges (Phylum Porifera) with a long geologic history, extending from the late Cambrian to the present. Fifteen species of sclerosponges, some of which are extinct,

have been identified, which belong to 2 classes, 4 subclasses, 8 families, and 11 genera. Sclerosponges have proven particularly useful in environmental reconstructions in the Quaternary: *Ceratoporella nicholsoni*, the most massive Caribbean sclerosponge with an aragonite skeleton, and *Astrosclera willleyana* (aragonite) and *Acanthochaetes wellsi* (calcite), which are widely distributed and abundant sponges in the Indo-Pacific. Sclerosponges tend to inhabit cryptic (i.e., sheltered) environments. *Ceratoporella nicholsoni* are common in deeper sections of fore reef as well as light-starved submarine caves, and *A. willleyana* and *A. wellsi* are common in well-flushed, sublit-toral caves in reef walls.

Sclerosponges are thought to build their skeleton in a manner analogous to scleractinian corals, in which the porous upper portion of the skeleton houses the living organism and the dense lower skeleton is devoid of living sponge tissue. Calcification occurs within the living tissue layer, and is characterized by extension in the uppermost region of the tissue layer and thickening in the lower regions. The two-stage calcification process results in time averaging of the environmental signal, the magnitude of which is a function of tissue layer thickness, growth rate, and density difference between the porous upper skeleton and dense basal skeleton. The basal skeleton of sclerosponges is very dense, approaching that of solid aragonite (2.93 g cm^{-3}), and denser than the aragonite skeletons of reef-building corals. Hence, sclerosponges are thought to be less susceptible to postdepositional alteration than corals.

Sclerosponges are slow-growing, long-lived organisms; average growth rates range from 0.1 to 0.3 mm yr^{-1} as estimated using staining (Alizarin Red-S, Calcein) and/or radiometric techniques (^{14}C , ^{210}Pb , U-Th). Recent work suggests growth rates can reach 1.2 mm yr^{-1} in *A. wellsi*. The internal chronology of sclerosponges has been constrained using ratios of Sr/Ca (*C. nicholsoni*) and $\delta^{13}\text{C}$ (*A. wellsi*), which have been shown to respond to seasonal temperature changes. Growth rates within a single sponge can deviate from the average growth rate by a factor of 2 using such high-resolution techniques; however, the mean growth rate is similar to constant growth rate estimates. Such methods have been proposed to provide high-resolution age models for cross-correlation in time of sclerosponges from different locations for paleoclimate studies, much the same way that coral sclerochronology is often made more precise using geochemical signals. In *C. nicholsoni* the overall slow growth rate indicates that even small sponges may live for hundreds to thousands of years. Sclerosponges do not have annual density bands, unlike scleractinian corals, but the skeletons of some species do contain concentric bands of darker material. The chronologic significance of such color variations is still being investigated. Sclerosponge skeletons contain relatively high uranium and low thorium concentrations, making them amenable to absolute dating with high precision via mass spectrometric U-series techniques. Sclerosponges skeletons are also amenable to radiocarbon age determinations.

The combination of slow growth rates, long life spans, depth distribution in the upper water column, and amenability to absolute dating make sclerosponges particularly attractive as an archive of Quaternary climate variations. Pioneering early work on sclerosponges concluded that these sponges precipitate their aragonite skeleton at (near?) stable isotopic

equilibrium with ambient seawater, a determination that has been confirmed by subsequent studies. Such equilibrium precipitation is an attribute that distinguishes them from reef-building corals.

Carbon isotopic records in sclerosponges have played a seminal role in providing independent evidence for the influence of deforestation and fossil-fuel burning on the $\delta^{13}\text{C}$ of DIC in surface seawater (i.e., Suess effect). These records exhibit significant and reproducible depletions in $\delta^{13}\text{C}$, as first documented in records from the Caribbean and then in those from the Indo-Pacific. The highly reproducible nature of the twentieth century trend toward more depleted $\delta^{13}\text{C}$ values in sclerosponge $\delta^{13}\text{C}$ records provides additional chronological constraints in proxy records using sclerosponges. In detail, sclerosponge $\delta^{13}\text{C}$ records vary from region to region, reflecting the local surface seawater DIC, the variability of which is a function of oceanic carbon cycling.

Oxygen isotopic values determined from skeletons of sclerosponges agree with the expected isotopic equilibrium values for ambient seawater temperature and $\delta^{18}\text{O}$ composition, although detailed calibration studies of the sclerosponge $\delta^{18}\text{O}$ -temperature relationship, like those performed for scleractinian corals, are in their infancy. Initial relationships between $\delta^{18}\text{O}$ and temperature have been based on bulk oxygen isotope values and annual average temperatures from seven sclerosponges, and these results match the biogenic aragonite line over a small temperature range. This similarity facilitates direct comparison between $\delta^{18}\text{O}$ records among different sedimentary archives. Records of $\delta^{18}\text{O}$ from sclerosponges having a depth distribution have also been combined with $\delta^{18}\text{O}$ records from surface-dwelling corals to document variations in thermocline depth.

Variations in elemental concentrations (e.g., Pb) in sclerosponges have shown to reflect environmental variability. For example, in Caribbean and Bahamas sclerosponges, there are sharp increases in Pb concentrations from the post-1850 industrial period, which coincide with increases in Pb concentrations documented in Greenland snow and a scleractinian coral from the Florida Keys. Differences in the shape of the Pb peak have been demonstrated at different depths from the same location and between different locations, most likely a result of differences in particulate settling. Nonetheless, the Pb peak in sclerosponges is very pronounced and can be used as another age constraint.

The concentrations of boron, magnesium, and barium are lower in sclerosponges relative to those in scleractinian corals, whereas concentrations of strontium and uranium are significantly higher in sclerosponges. Such differences likely reflect differences in calcification mechanisms between the non-zooxanthellate sclerosponges and the zooxanthellate corals. Element-to-calcium ratios (e.g., Sr/Ca, U/Ca, B/Ca, Mg/Ca) in sclerosponges have recently been evaluated in terms of their utility as proxies of environmental change. The emerging consensus is that sclerosponge Sr/Ca holds the most promise as a proxy for changes in seawater temperature. The temperature sensitivity of Sr/Ca ratios in sclerosponges appears to be greater than in both inorganic aragonite and scleractinian coral skeletons. Such heightened temperature sensitivity is significant because of the small amplitude of the seawater temperature signal below the surface mixed layer.

Mollusks

During >500 Myr of evolution, mollusks (Phylum Mollusca) have occupied an extraordinarily broad variety of different ecological niches ranging from deep-sea to shallow-water habitats, from the poles to the equator, from lakes, rivers, and intertidal zones to dry land. Whereas all seven extant mollusk classes (Aplacophora, Monoplacophora, Polyplacophora, Scaphopoda, Bivalvia, Gastropoda, and Cephalopoda) occur in the marine realm, bivalves and gastropods also inhabit brackish and freshwater environments. Some snails even live in habitats on dry land. This enormous biogeographic distribution is uncontested in the animal kingdom.

Most mollusks form calcareous hard tissues that serve a multitude of different tasks (e.g., providing shape, stability, and protection). In addition, mollusk shells, particularly those of bivalves, fulfill the prerequisites of Quaternary climate archives because they function as precise calendars and record environmental information, in chronological order, during their entire life.

Mollusk shells grow by periodic accretion of skeletal material (mineral phase plus soluble and insoluble organic matrices) along the ventral margin. Precipitation of calcium carbonate from the pallial fluid is mediated by proteins. This process is periodically retarded or interrupted and results in the formation of distinct growth lines: zones enriched in organic material and characterized by different crystallization patterns (smaller crystallites, different growth direction, etc.). Growth lines segment the growth pattern into time slices of equal duration, so-called growth increments. Numerous studies have identified annual, fortnightly, daily, and ultradian (more than once a day) growth increments in mollusk shells reflecting cycles in temperature and food availability or other environmental variables, biological clocks (circadian/diurnal and circalunidian/tidal rhythms), or chemical cycles. Regular growth patterns can be used for a variety of different purposes, such as precise calendar dating of each shell portion, to estimate the time interval represented by samples taken from the shell for geochemical studies, to reconstruct life history traits and past environmental and climate changes or to determine the ontogenetic age of each specimen. The study of growth pattern analysis of molluscan shells is commonly known as mollusk sclerochronology.

In contrast to many ephemeral mollusk species that live only for a few years, some bivalves can live for many centuries, thereby adding less than $10\text{--}100\ \mu\text{m}$ of $\text{CaCO}_3\ \text{yr}^{-1}$ to their shells during gerontic stages. In fact, bivalves lead the list of the longest-lived animals that produce distinct incremental growth patterns in their hard parts. Examples are geoducks, *Panopea abrupta* (Conrad) from the Northwest Pacific (>160 yr old; ca. 15 cm shell width) or the freshwater pearl mussel, *Margaritifera margaritifera* (Linnaeus) from Scandinavian rivers (>300 yr; ca. 10 cm shell width). The ocean quahog, *Arctica islandica* (Linnaeus) from the North Atlantic can live for more than 375 yr and attains a shell height of less than 8 cm. Sclerochronological age estimates were substantiated by mark-and-recovery experiments (Tetracycline, Calcein, or Alizarin Red-S staining), absolute dating techniques ($^{14}\text{C}_{\text{AMS}}$, amino acid racemization), and cyclic variations of geochemical properties

in annual growth increments reflecting the seasonal temperature or primary productivity cycle.

Mollusk shells reliably record changes of the ambient environment as variable growth rates and variable geochemical properties. Usually, mollusk shells grow at faster rates (broader increments) at optimal growth conditions, for example, higher temperature, food availability, and food quality. Extraction of environmental signals from growth increment chronologies, however, requires removal of growth trends that are related to physiological aging. With increasing ontogenetic age, the absolute growth rate and the year-to-year variance (heteroscedasticity) decrease. These 'age' trends can be estimated and removed from the growth increment time series using detrending and standardizing techniques developed by dendrochronologists. The resulting chronologies are dimensionless measures of shell growth expressed as standardized growth indices (SGI). Several studies demonstrated that contemporaneous specimens alive in the same habitat exhibit similar SGI chronologies, indicating that shell growth is actually controlled by external factors. Based on similar growth patterns, SGI time series of many individuals with overlapping life spans can thus be strung together to form long, uninterrupted chronologies, so-called master or composite chronologies. Master chronologies cover many mollusk generations and provide a record of environmental variation that affected all specimens alive at the same time and in the same habitat. For example, growth patterns of geoduck shells from the NW Pacific were used to reconstruct decadal-scale variability of ocean currents (PDO) during the last 150 yr. Freshwater pearl mussels from Scandinavia demonstrated that the number of extreme cold summers has significantly decreased since the early twentieth century. Ocean quahogs from around Iceland indicated that heat advection (which itself varies in strength on quasi-decadal periods of 12–14 yr) to northern latitudes was strongly reduced during the Little Ice Age.

Geochemical properties of mollusk shells provide another possibility for reconstructing past climates and environments. Shell oxygen and stable carbon isotopes have frequently been used to confirm the annual formation of shell material, to infer subseasonal and inter-annual changes of water temperature, primary productivity, and to provide further evidence for an anthropogenic increase of CO_2 in the ocean. For example, shells of *Phacosoma japonicum* (Reeve) from shell middens revealed that Japan was exposed to large swings of monsoon-related precipitation during the Holocene. Several studies have confirmed that mollusks form their shell in oxygen isotopic equilibrium with the ambient water. If the $\delta^{18}\text{O}$ value of the water is known, the temperature during formation of the shell can be estimated. For example, the inter-annual temperature variability of the Labrador Current during the late nineteenth and twentieth century was reconstructed through the analyses of shell $\delta^{18}\text{O}$ values of *Arctica islandica*.

However, other stable isotope ratios (e.g., $\delta^{13}\text{C}$), trace and minor elements or element-to-calcium ratios (Sr/Ca, Mg/Ca, Ba/Ca) of mollusk shells seem to exhibit significant "vital effects" (i.e., trends related to growth rate, metabolism and/or ontogenetic age). Some bivalves may also be fairly selective against certain trace and minor elements, which may pose

limits on using shells as environmental biomonitors (tracking pollution sources etc.). Several potential proxies of environmental variables are currently being evaluated, including $\delta^{44}\text{Ca}$ (temperature proxy), $\delta^{11}\text{B}$ (pH proxy), and molecules containing more than one rare isotope (e.g., $^{13}\text{C}^{18}\text{O}^{16}\text{O}$ in $\text{CaCO}_3 = \delta^{18}\text{O}_{\text{water}}$ -independent temperature proxy).

The combination of widespread occurrence in time and space, multiproxy archiving and temporal control (calendar alignment) make mollusks ideal archives of Quaternary climate and environmental variability. Fast-growing young shell portions provide unique paleoweather data and information about seasonal environmental extremes. Long-lived species, however, can open century-long windows in the climatic past and provide detailed information on decadal-scale climate variability. An integrated approach that combines sclerochronological and geochemical techniques provides the most powerful tool to separate environmental signals from ontogenetic age- and growth rate-related noise.

Summary

In this short article, we have presented a survey of decades worth of research in the establishment and application of using the skeletons of corals, sclerosponges, and mollusks as sedimentary archives of environmental variability in the Quaternary. The environmental and climate record encoded in the skeletons of these organisms provide a background in which to assess future climate change. Those interested in learning additional information on the topic of this chapter are encouraged to use the articles listed in the References section as a starting point and to then delve into the abundant and additional primary literature.

See also: **Paleoceanography: Paleoceanography An Overview.**
Paleoceanography, Biological Proxies: Temperature Proxies,

Census Counts. Paleoceanography, Physical and Chemical Proxies: Carbon Cycle Proxies ($\delta^{11}\text{B}$, $\delta^{13}\text{C}_{\text{calcite}}$, $\delta^{13}\text{C}_{\text{organic}}$, Shell Weights, B/Ca, U/Ca, Zn/Ca, Ba/Ca); Nutrient Proxies; Radioisotope Proxies; Salinity Proxies $\delta^{18}\text{O}$.

References

- Böhm F, Joachimski MM, Dullo W-C, et al. (2000) Oxygen isotope fractionation in marine aragonite of coralline sponges. *Geochimica et Cosmochimica Acta* 64: 1695–1703.
- Clark GR II (2005) Daily growth lines in some living Pectens (Mollusca: Bivalvia), and some applications in a fossil relative: Time and tide will tell. *Palaeogeography, Palaeoclimatology, Paleoecology* 228: 26–42.
- Druffel ERM and Benavides LM (1986) Input of excess CO_2 to the surface ocean based on $^{13}\text{C}/^{12}\text{C}$ ratios in a banded Jamaican sclerosponge. *Nature* 321: 58.
- Druffel ERM (1997) Geochemistry of corals: Proxies of past ocean chemistry, ocean circulation, and climate. *Proceedings of the National Academy of Sciences* 94: 8354–8361.
- Felis T and Pätzold J (2004) Corals as Climate Archive. In: Fischer H, Kumke T, and Lohmann G, et al. (eds.) *The Climate in Historical Times: Towards a Synthesis of Holocene Proxy Data and Climate Models*, pp. 91–108. Berlin: Springer.
- Gagan MK, Ayliffe LK, Beck JW, et al. (2000) New views of tropical paleoclimate from corals. *Quaternary Science Reviews* 19: 167–182.
- Jones DS (1983) Sclerochronology: Reading the record of the molluscan shell. *American Scientist* 71: 384–391.
- Lorrain A, Gillikan DP, Paulet Y-M, et al. (2005) Strong kinetic effects on Sr/Ca ratios in the calcitic bivalve *Pecten maximus*. *Geology* 33: 965–968.
- Rosenheim BE, Swart PK, Thorrold SR, Willenz P, Berry L, and Latkoczy C (2004) High-resolution Sr/Ca records in sclerosponges calibrated to temperature *in situ*. *Geology* 32(2): 145–148.
- Schöne BR, Rodland DL, Fiebig J, et al. (2006) Reliability of multi-taxon, multi-proxy reconstructions of environmental conditions from accretionary biogenic skeletons. *Journal of Geology* 114: 267–285.
- Swart PK, Thorrold SR, Rosenheim BE, et al. (2002) Intra-annual variation in the stable oxygen and carbon and trace element composition of sclerosponges. *Paleoceanography* 17: 1045. <http://dx.doi.org/10.1029/2000PA000622>.
- Tudhope AW, Chilcott CP, McCulloch MT, et al. (2001) Variability in the El Niño–Southern oscillation through a glacial–interglacial cycle. *Science* 291: 1511–1517.

Dinoflagellates

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Introduction

Dinoflagellates are microscopic unicellular organisms occupying most aquatic environments, from freshwater bodies to the open ocean. Most dinoflagellates are planktonic and use their two flagella to swim in spiral-like motion, from which their name originates (from the Greek word *dinos* meaning whirling). In biological sciences, dinoflagellates receive special attention since they represent an important part of the net primary production in aquatic environments and because they are responsible for harmful algal blooms or red tides. In paleontology and Earth sciences, dinoflagellates also deserve our particular interest since they yield microfossils, which constitute good biostratigraphic markers of the Mesozoic and Cenozoic. They have been shown to be useful paleoecological indicators of changes in sea-surface waters.

The motile stage of dinoflagellates does not produce remains that preserve as fossils, but about 10–20% of species experience a cyst stage that do yield microfossils. These microfossils mostly consist of highly resistant organic-walled cysts, also called dinocysts, which can be routinely observed under optical microscope in palynological slides. Therefore, the fossil remains of dinoflagellates do not correspond to the vegetative stage of the cells, and the morphology of living and fossil forms may differ considerably (Figures 1 and 2). Living dinoflagellates were described by naturalists in the eighteenth century, but the affinities of microfossils related to dinoflagellates have only been known since the second half of the twentieth century. Most fossil dinoflagellate cysts described from the beginning of nineteenth century until the end of the twentieth century were associated with ‘hystrichospheres’ and believed to be the remineralized skeletons of algae. Important progress in the understanding of the status of dinoflagellate cysts in the living world was made during the 1960s. The in-depth morphological observations of [Evvitt \(1963\)](#) permitted the identification of structures relating the fossil organic-walled cysts with dinoflagellates. Moreover, the *in vitro* experiments by [Wall and Dale \(e.g., 1966, 1968\)](#) demonstrated the biological affinities of dinocysts and contributed to the documentation of the stages of the life cycle of many taxa (Table 1).

Since the 1970s, many studies have been undertaken to document the distribution of dinocysts on the sea floor. About one hundred Quaternary dinocyst species have been described to date. The dinocyst assemblages in surface sediment samples of the marine environment show a biogeographic distribution with distinct latitudinal gradients from the Arctic seas to the circum-Antarctic Ocean, demonstrating their dependence on temperature and sea-ice conditions (e.g., [de Vernal et al., 2008](#); [Rochon et al., 1999](#)). Dinocyst distribution patterns also reflect inshore to offshore gradients

that have been linked to salinity, productivity, and eutrophication (e.g., [Dale, 1996](#); [Marret and Zonneveld, 2003](#); [Radi and de Vernal, 2008](#)). Based on these observations, it is possible to discern relationships between dinocyst assemblages and sea-surface conditions, including temperature, salinity, sea-ice cover, productivity and trophic characteristics of the waters, and therefore to use dinocyst assemblages for qualitative or quantitative reconstruction of paleoceanographic conditions (e.g., [de Vernal et al., 2005, 2008](#); [Radi and de Vernal, 2008](#)).

In addition to organic-walled cysts or dinocysts, some dinoflagellate taxa yield calcareous microfossils generally associated with the cyst stage. Calcareous dinoflagellates appear to be abundant in oceanic sediments at middle to low latitudes, where they can be used as tracers of hydrographic conditions in the upper water column (e.g., [Vink, 2004](#)).

Here, we principally summarize information on dinoflagellates and their organic-walled cysts, which are studied within the field of marine palynology and currently used for paleoenvironmental investigation in Quaternary sciences.

Ecology of Dinoflagellates

Dinoflagellates live in various types of aquatic environments, including lakes, estuaries, epicontinental seas and oceans, from equatorial to polar latitudes (for a synthesis of the ecology of dinoflagellates, see [Taylor and Pollinger, 1987](#) or [Matthiessen et al., 2005](#)). However, most dinoflagellates are marine: a few thousand species are known from marine waters, whereas only a few hundred species are known to live in freshwater.

In the oceans, dinoflagellates seem to be particularly well adapted to neritic (nearshore) environments, including the estuaries, epicontinental seas, and continental shelves. This is probably due to their broad tolerance to low salinity conditions, in addition to nutrient availability and stratification of water masses. In general, the diversity of dinoflagellates in surface waters increases toward the equator and decreases toward the poles.

The feeding strategy of dinoflagellates is variable. Many are autotrophic (i.e., photosynthetic) and form an important part of the planktonic primary production in lakes and oceans. About half of dinoflagellate species are heterotrophic or mixotrophic (i.e., feeding on other organisms or on dissolved organic substances), and others are symbionts of marine invertebrates such as corals in which they are known as zooxanthellae. The autotrophic dinoflagellates depend upon light and nutrients (notably nitrogen and phosphorus). They live in the photic zone at relatively shallow depths, usually within the upper 50–100 m. The heterotrophic dinoflagellates depend on the overall biological productivity of their environment

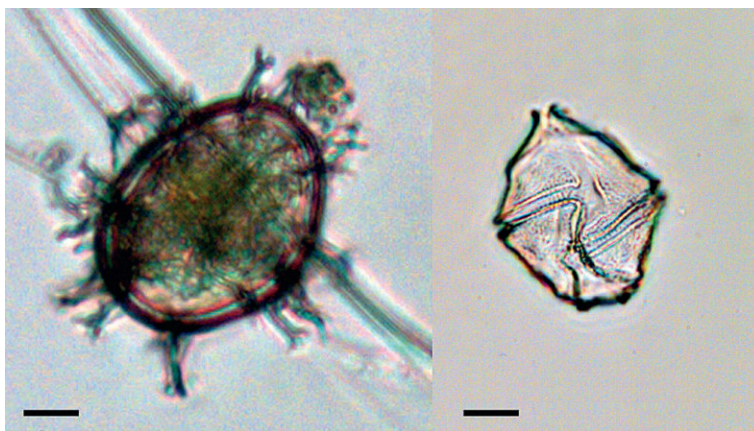


Figure 1 Photograph of *Gonyaulax* cyst and theca in optical microscopy (scale bar = 10 μm).

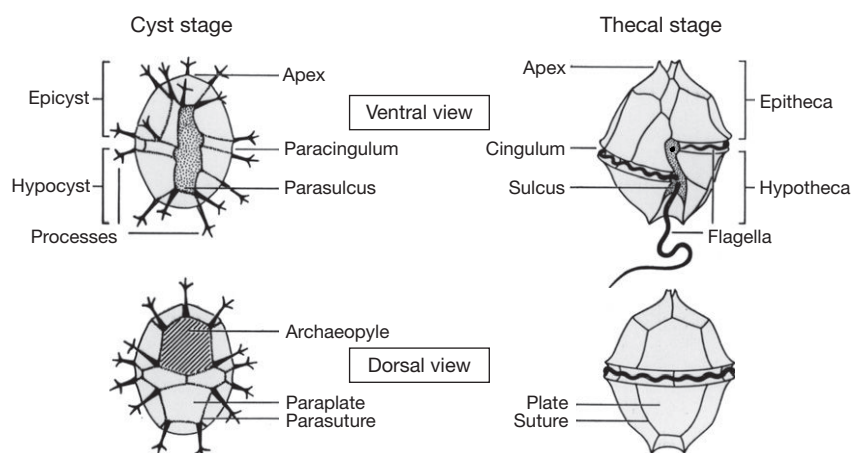


Figure 2 Scheme of dinoflagellate cyst and theca (*Gonyaulax*-type).

and generally live in surface waters where their prey can be found. Studies on feeding behaviors of dinoflagellates have also shown that several autotrophic species are actually mixotrophic, using either phagotrophy (engulfing particulates) or osmotrophy (assimilation of dissolved organic or inorganic compounds), possibly as a source of vitamins or to supplement their nutrition (e.g., Stoecker, 1999).

Dinoflagellates are mobile in the water column. They have two flagella, one around the cingulum and the other longitudinal, which permit swimming with a spiral-like 'whirling' motion with a speed ranging from a few centimeters to a few meters per hour. They use their flagella together with physiologic adjustment of buoyancy to migrate vertically in the water column, in order to optimize their metabolic and feeding activities. During the day, dinoflagellates migrate to the surface for photosynthesis, and during the night, they migrate down, away from the nutrient-depleted surface waters. Despite their ability to move vertically, dinoflagellates generally inhabit a relatively thin and shallow surface water layer, especially in stratified marine environments, because they cannot migrate across the pycnocline, an important physical barrier.

The reproduction of dinoflagellates is most commonly asexual by mitosis. In the blooming period, vegetative cell

divisions occur at a rate of about one per day. Sexual reproduction is also observed for many species. When blooming, dinoflagellates can be responsible for 'red tides,' so called because the large density of cells in the surface water induces a color change (green, brown, or red). Many dinoflagellates are bioluminescent and cause a glow at the sea surface by night. A few dinoflagellate species produce neurotoxins that may be bioconcentrated by filtering organisms, notably shellfish, which then become poisonous and dangerous for the health of animals (or people) that eat them.

In marine environments, dinoflagellates constitute one of the main primary producers, together with diatoms and coccolithophorids. Typically, dinoflagellates experience their blooms after diatoms.

The Life Cycle of Dinoflagellates and the Cyst Stage

Most dinoflagellates have a complex life cycle involving several stages: asexual and sexual, motile, and nonmotile (cf. Pfister and Anderson, 1987 for a synthesis, and Figure 3 for a schematic illustration). In general, during the course of sexual reproduction, the haploid vegetative cells ($1 - n$ chromosomes) become

Table 1 List of marine late Quaternary dinocyst taxa

Species – Dinocyst name	Motile affinity	Order
<i>Ataxiodinium choane</i>	<i>Gonyaulax</i> sp.	Gonyaulacales
<i>Bitectatodinium spongium</i>	Unknown	Gonyaulacales
<i>Bitectatodinium tepikiense</i>	<i>Gonyaulax digitale</i>	Gonyaulacales
<i>Brigantedinium cariacense</i>	<i>Protoperidinium avellanum</i>	Peridinales
<i>Brigantedinium simplex</i>	<i>Protoperidinium conicoides</i>	Peridinales
<i>Dalella chathamensis</i>	<i>Gonyaulax</i> sp.	Gonyaulacales
<i>Dubridinium caperatum</i>	<i>Preperidinium meunieri</i>	Peridinales
<i>Echinidinium aculeatum</i>	Unknown	Peridinales
<i>Echinidinium bispiniformum</i>	Unknown	Peridinales
<i>Echinidinium delicatum</i>	Unknown	Peridinales
<i>Echinidinium granulatum</i>	Unknown	Peridinales
<i>Echinidinium karaense</i>	Unknown	Peridinales
<i>Echinidinium transparentum</i>	Unknown	Peridinales
<i>Echinidinium zonneveldii</i>	Unknown	Peridinales
Cysts of <i>Gymnodinium catenatum</i>	<i>Gymnodinium catenatum</i>	Gymnodinales
<i>Impagidinium aculeatum</i>	<i>Gonyaulax</i> sp.	Gonyaulacales
<i>Impagidinium japonicum</i>	<i>Gonyaulax</i> sp.	Gonyaulacales
<i>Impagidinium pallidum</i>	<i>Gonyaulax</i> sp.	Gonyaulacales
<i>Impagidinium paradoxum</i>	<i>Gonyaulax</i> sp.	Gonyaulacales
<i>Impagidinium patulum</i>	<i>Gonyaulax</i> sp.	Gonyaulacales
<i>Impagidinium plicatum</i>	<i>Gonyaulax</i> sp.	Gonyaulacales
<i>Impagidinium sphaericum</i>	<i>Gonyaulax</i> sp.	Gonyaulacales
<i>Impagidinium striatum</i>	<i>Gonyaulax</i> sp.	Gonyaulacales
<i>Impagidinium variaseptum</i>	<i>Gonyaulax</i> sp.	Gonyaulacales
<i>Impagidinium velorum</i>	<i>Gonyaulax</i> sp.	Gonyaulacales
<i>Islandinium brevispinosum</i>	<i>Protoperidinium</i> sp.	Peridinales
<i>Islandinium? cezare</i>	<i>Protoperidinium</i> sp.	Peridinales
<i>Islandinium minutum</i>	<i>Protoperidinium</i> sp.	Peridinales
<i>Leipokatium invisitatum</i>	Unknown	Peridinales
<i>Lejeunecysta oliva</i>	Unknown	Peridinales
<i>Lejeunecysta sabrina</i>	? <i>Protoperidinium leone</i>	Peridinales
<i>Lingulodinium machaerophorum</i>	<i>Lingulodinium polyedrum</i>	Gonyaulacales
<i>Nematosphaeropsis labyrinthus</i>	<i>Gonyaulax spinifera</i>	Gonyaulacales
<i>Nematosphaeropsis rigida</i>	<i>Gonyaulax</i> sp.	Gonyaulacales
<i>Operculodinium aguinaense</i>	? <i>Protoceratium reticulatum</i>	Gonyaulacales
<i>Operculodinium centrocarpum</i>	<i>Protoceratium reticulatum</i>	Gonyaulacales
<i>Operculodinium short processes</i>	? <i>Protoceratium reticulatum</i>	Gonyaulacales
<i>Operculodinium israelianum</i>	? <i>Protoceratium</i> sp.	Gonyaulacales
<i>Operculodinium janduchenei</i>	Unknown	Gonyaulacales
<i>Operculodinium longispinigerum</i>	Unknown	Gonyaulacales
Cysts of <i>Pentaparsodinium dalei</i>	<i>Pentaparsodinium dalei</i>	Peridinales
Cysts of <i>Pheopolykrikos hartmannii</i>	<i>Pheopolykrikos hartmannii</i>	Gymnodinales
Cyst of <i>Polykrikos</i> sp.-Arctic morphotype	<i>Polykrikos</i> sp.	Gymnodinales
Cysts of <i>Polykrikos kofoidii</i>	<i>Polykrikos kofoidii</i>	Gymnodinales

(Continued)

Table 1 (Continued)

Species – Dinocyst name	Motile affinity	Order
Cysts of <i>Polykrikos schwartzii</i>	<i>Polykrikos schwartzii</i>	
<i>Polysphaeridium zoharyi</i>	<i>Pyrodinium bahamense</i>	Gonyaulacales
Cysts of <i>Protoperidinium americanum</i>	<i>Protoperidinium americanum</i>	Peridinales
Cysts of <i>Protoperidinium nudum</i>	<i>Protoperidinium nudum</i>	Peridinales
Cysts of <i>Protoperidinium stellatum</i>	<i>Protoperidinium stellatum</i>	Peridinales
<i>Pyxidinospis psilata</i>	Unknown	Gonyaulacales
<i>Pyxidinospis reticulata</i>	Unknown	Gonyaulacales
<i>Quinquecusps concreta</i>	? <i>Protoperidinium leone</i>	Peridinales
Cysts of cf. <i>Scrippsiella trifida</i>	<i>Scrippsiella trifida</i>	Gonyaulacales
<i>Selenopemphix antarctica</i>	Unknown	Peridinales
<i>Selenopemphix nephroides</i>	<i>Protoperidinium subinermis</i>	Peridinales
<i>Selenopemphix quanta</i>	<i>Protoperidinium conicum</i>	Peridinales
<i>Spiniferites alaskensis</i>	<i>Gonyaulax</i> sp.	Gonyaulacales
<i>Spiniferites belerius</i>	<i>Gonyaulax scrippsae</i>	Gonyaulacales
<i>Spiniferites bentori</i>	<i>Gonyaulax digitale</i>	Gonyaulacales
<i>Spiniferites bulloideus</i>	<i>Gonyaulax scrippsae</i>	Gonyaulacales
<i>Spiniferites cruciformis</i>	<i>Gonyaulax</i> sp.	Gonyaulacales
<i>Spiniferites delicatus</i>	<i>Gonyaulax</i> sp.	Gonyaulacales
<i>Spiniferites elongatus</i>	<i>Gonyaulax elongata</i>	Gonyaulacales
<i>Spiniferites frigidus</i>	<i>Gonyaulax</i> sp.	Gonyaulacales
<i>Spiniferites hyperacanthus</i>	<i>Gonyaulax</i> sp.	Gonyaulacales
<i>Spiniferites lazus</i>	<i>Gonyaulax</i> sp.	Gonyaulacales
<i>Spiniferites membranaceus</i>	<i>Gonyaulax membranacea</i>	Gonyaulacales
<i>Spiniferites mirabilis</i>	<i>Gonyaulax spinifera</i>	Gonyaulacales
<i>Spiniferites pachydermus</i>	<i>Gonyaulax</i> sp.	Gonyaulacales
<i>Spiniferites ramosus</i>	<i>Gonyaulax</i> sp.	Gonyaulacales
<i>Stelladinium reidii</i>	Unknown	Peridinales
<i>Stelladinium robustum</i>	Unknown	Peridinales
<i>Tectatodinium pellitum</i>	<i>Gonyaulax spinifera</i>	Gonyaulacales
<i>Trinovantedinium applanatum</i>	<i>Protoperidinium pentagonum</i>	Peridinales
<i>Trinovantedinium variabile</i>	Unknown	Peridinales
<i>Tuberculodinium vancampoe</i>	<i>Pyrophacus steinii</i>	Gonyaulacales
<i>Votadinium calvum</i>	<i>Protoperidinium oblongum</i>	Peridinales
<i>Votadinium spinosum</i>	<i>Protoperidinium claudicans</i>	Peridinales
<i>Xandarodinium xanthum</i>	<i>Protoperidinium divaricatum</i>	Peridinales

Gray zones: heterotrophic or mixotrophic taxa.

White zones: autotrophic taxa.

or produce haploid gametes that fuse to develop a zygote. The zygote rapidly enters into a dormancy stage called 'hypnozygote,' during which the diploid cell (with $2 - n$ chromosomes) is protected by forming a cyst. After a dormancy period of variable length, the protoplasm excysts and meiosis occurs. The excystment of the cell occurs through an opening in the cyst called the

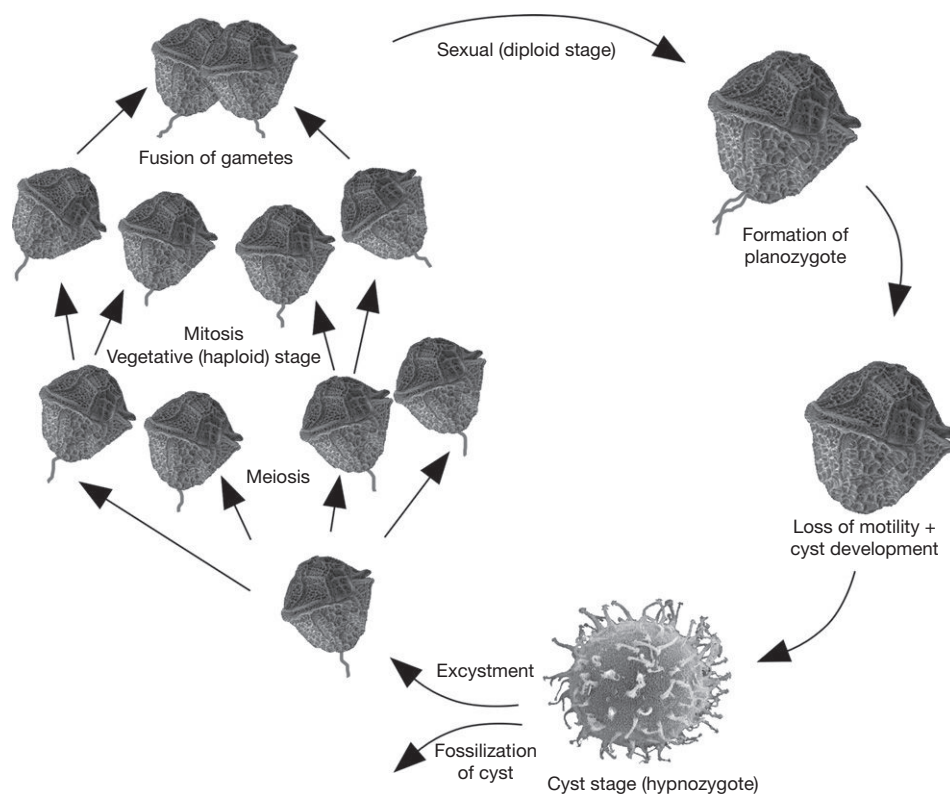


Figure 3 Schematic illustration of idealized dinoflagellate life cycle. Adapted from Evitt WR (1985) *Sporopollenin Dinoflagellate Cysts – Their Morphology and Interpretation*, 333 pp. Dallas, TX: American Association of Stratigraphic Palynologists Foundation.

archaeopyle, which has geometry determined by the morphology of the original theca (see Section ‘[The Taxonomy and General Morphology of Dinoflagellates and Dinocysts](#)’ below).

In the life cycle of dinoflagellates, the cysts have an extremely important function since they form a protective structure that fosters survival of the cell over long periods of time (several years to decades) and permits propagation and dispersal (cf. [Dale, 1983](#)).

The cyst wall of many species is made of highly resistant organic matter called dinosporin. This is similar in composition to the sporopollenin of pollen grains. These organic-walled cysts called ‘dinocysts’ are typically 15–100 μm in diameter. They are observed in palynological slides prepared following standard laboratory procedures used for the study of pollen and spores, which involve treatments with hydrofluoric and hydrochloric acids. The study of dinocysts is thus considered a subdiscipline of palynology.

Some dinoflagellates also produce cysts made of calcium carbonate, which fossilize in sediment. These are analyzed together with the calcareous nannoplankton. In addition to hypnozygotes, some dinoflagellates produce temporary cysts to survive adverse conditions. This type of cyst does not preserve as fossils.

There are variations in the life cycle of dinoflagellates. Sexual reproduction has been observed only in a few taxa, and more *in vitro* and *in vivo* studies are necessary for a comprehensive understanding of their life cycle. Only 10–20% of dinoflagellate taxa are known to produce cysts that preserve as fossils. The correspondence between the organic-walled or calcareous dinoflagellate cysts recovered

in sediments and their motile stage in the water column is known for several species but still needs to be established for many others.

The Taxonomy and General Morphology of Dinoflagellates and Dinocysts

Dinoflagellates are unicellular eukaryotic organisms that belong to the Protista, based on the classification system of five kingdoms. Because they include both autotrophic and heterotrophic species, dinoflagellates can be classified both by zoologists and botanists. Within the International Code of Zoological Nomenclature, they have been grouped under the phylum of Mastigophora. Dinoflagellates and their cysts, however, are more commonly treated as algae and are thus classified following the rules of the International Code of Botanical Nomenclature. For a long time, they have been associated to the division of Pyrrophyta named after the bioluminescent forms (from the Greek word *pyrrhos* meaning fire). Dinoflagellates are now grouped within the division of Dinoflagellata, which is distinguished on the bases of several criteria. A main one is the cell membrane that is characterized by a complex outer layer (the Amphiesma) containing a single layer of flattened vesicles (the amphiesmal vesicles). In the cases of many dinoflagellates, the amphiesmal vesicles contain thecal plates composed of cellulose. These dinoflagellates are classified as thecate, whereas those lacking such cellulosic plates are classified as naked or unarmored. Dinoflagellates also bear a unique

type of nucleus, the dinokaryon, which is characterized by chromosomes that remain condensed between cell divisions and which lack histones. Another peculiarity of dinoflagellates is their asymmetric morphology, and their possession of two dissimilar flagella.

The classification of dinoflagellates and their cysts is principally based on the structure of the Amphiesma and the organization of the thecal plates, referred to as tabulation. Among dinoflagellates producing cysts that have been recovered in Quaternary geological samples, three principal groups are represented: the Gymnodiniales that are unarmored and the Gonyaulales and Peridiniales that are thecate and show distinct types of tabulation (Figure 4).

In the Peridiniales and Gonyaulales, the tabulation pattern is described according to the organization of the theca from the apex to the cingulum and antapex, and according to the ventral or dorsal orientation. On the ventral (or sulcal) side, are inserted the flagella, whereas the archaeopyle is often located on the dorsal side of the cysts. The tabulation includes apical, precingular, cingular, postcingular, and antapical plates, which are labeled according to various systems (see Edwards, 1993; Evitt, 1985). The most commonly used is the Kofoid's system, based on the number of plates in each level of the theca. For the cyst stage, the prefix 'para' is used when referring to the outline of the plates on the cyst wall, and certain plates are called 'homologues' of their thecal counterpart. The tabulation system of Evitt and Taylor, which is based on the interconnection between plates, is also often used for the description and classification of Gonyaulales. The main morphological characteristic of Peridiniales is the presence of intercalary plates between the apical and precingular plate series on the dorsal side, which are lacking in Gonyaulales (Figure 5).

Independent classifications of dinoflagellates have been erected by biologists and paleontologists, and most of the literature of the nineteenth and twentieth centuries refers to distinct classification schemes for living and fossil dinoflagellates. Nevertheless, Fensome and colleagues (1993) made an attempt to reconcile the classification schemes by biologists and paleontologists, using various criteria (nucleus type, structure of the Amphiesma, tabulation) and proposed a common suprageneric classification for cyst and motile stages of dinoflagellates. At the generic and species levels, the motile stage name could be retained for both living and fossil forms for which the biological affinities of the cysts are known. However, these affinities are still uncertain for many Quaternary dinoflagellate cysts and for most pre-Quaternary ones. Therefore, the genera and species names given to dinocysts often differ from the ones of their living counterpart (cf. Head, 1996).

In the classification of living and fossil dinoflagellates, tabulation constitutes an important diagnostic character. The tabulation pattern of the dinoflagellates can be reflected in various ways in the cysts. Some dinocysts may show sutures or septa defining the original pattern of thecal plates. Others possess extensions called processes whose position may be determined by the original tabulation. Some do not show any evidence of tabular pattern, with the exception of the archaeopyle that has a given position. Beyond the tabulation patterns, as reflected by sutures or processes and from the archaeopyle, there are many morphological criteria used to distinguish and identify dinocyst species. They include the structure of the layers of the wall (cavate or proximate), the overall shape of the cyst (elongate, round, or flattened), the presence of apical or antapical horns, the presence of processes and sutures, their respective positions (following

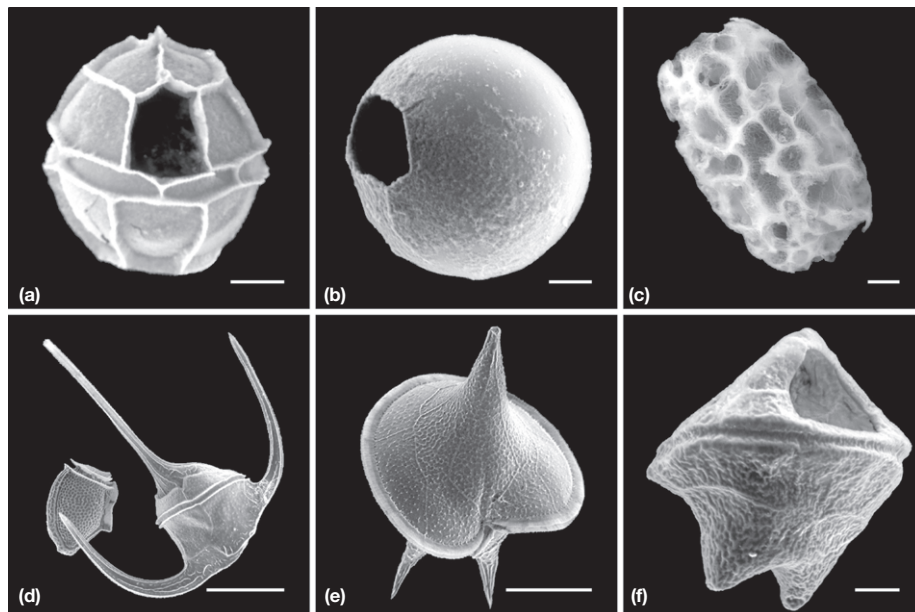
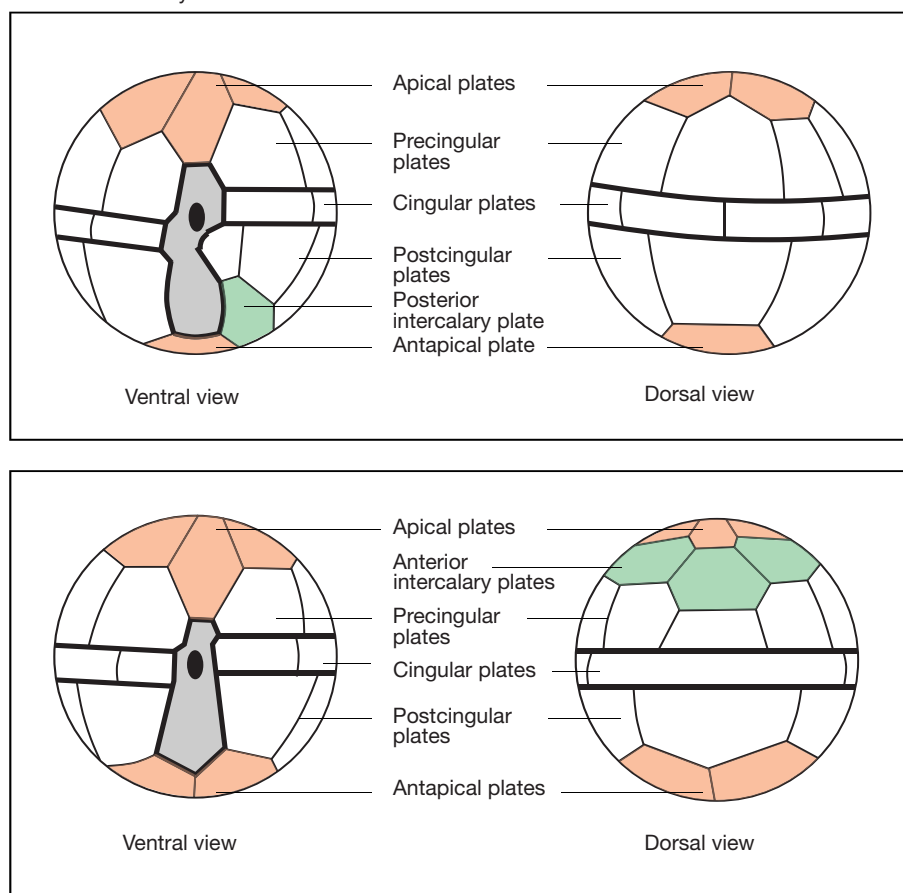


Figure 4 Scanning electron micrographs of dinoflagellates (d–e) and dinocysts. (a–c, f). Scale bar = 10 μ m; (d–e) scale bar = 50 μ m. (a) *Impagidinium patulum* (cyst of *Gonyaulax* sp., order Gonyaulales) North Atlantic Ocean; (b) *Brigantedinium cariacense* (cyst of *Protoperidinium avellanum*, order Peridiniales), Bedford Basin, NS, Canada; (c) Cyst of *Polykrikos schwartzii* (order Gymnodiniales), Lawrence Town, NS, Canada; (d) *Ceratium tripos* and *Dinophysis acuta* (theca; orders Gonyaulales and Dinophysiales, respectively), Bedford Basin, NS, Canada; (e) *Protoperidinium* cf. *depressum* (theca; order Peridiniales) Bedford Basin, NS, Canada; and (f) *Quinquecuspis concreta* (cyst of *Protoperidinium leone*, order Peridiniales), Saanich Inlet, BC, Canada.

Tabulation of Gonyaulacales



Tabulation of Peridinales

Figure 5 Simplified scheme showing the tabulation pattern of Peridinales and Gonyaulacales. The ventral view corresponds to the surface of the sulcus (in gray) where the flagellar pore is. The dorsal view corresponds to the surface where the archaepyle is often located (often made of precingular plate or plates for Gonyaulacales and of intercalary plate or plates for Peridinales). Adapted from Fensome RA, Taylor FJR, Norris G, Sarjeant WAS, Wharton DI, and Williams GL (1993) *A Classification of Living and Fossil Dinoflagellates*. Micropaleontology Special Publication Number 7, 351 pp, New York: American Museum of Natural History.

tabulation or not), the ornamentation of the cyst wall, the size, the color (semitransparent or brown), etc. Explicit descriptions of the morphological nomenclature of dinocysts have notably been published by the [Groupe de travail 'Dinoflagellés' \(Dinoflagellate working group, 1986\)](#), [Head et al. \(1989\)](#), [Edwards \(1993\)](#).

The Organic-Walled Dinoflagellate Cysts or Dinocysts in the Marine Realm

Biostratigraphy of Dinocysts

Dinocysts are the most common organic-walled microfossils or palynomorphs in marine sediments. In the field of marine palynology, they have been intensively studied for biostratigraphic purposes, notably within the frame of petroleum exploration programs. The most ancient dinocysts known thus far date from the Silurian, but they are more abundant from the Triassic to the modern, with maximum diversity of species recorded during the Cretaceous and Paleogene (e.g., [Powell, 1992](#); for an exhaustive index of fossil dinoflagellate genera and species with their respective biostratigraphic ranges, see [Fensome and](#)

[Williams, 2004](#)). The diversity of species in the Quaternary sediments is relatively low, with close to hundred species formally described to date. Although it still needs investigation, the Pliocene and Quaternary biostratigraphic record of dinocysts shows extinctions of a few species and very rare first appearances of species (for North Atlantic biostratigraphic scheme see [de Vernal et al., 1992](#); [De Schepper and Head, 2008](#)).

Biogeographic Distribution of Dinocysts

Since the early work of [Wall et al. \(1977\)](#) documenting the modern distribution of dinocysts in sediments, many palynological studies have permitted the description of the distribution patterns of dinocysts on the sea floor. There are now regional data sets for the North Atlantic and the Arctic Ocean, the circum-Antarctic Ocean, the low latitudes of the Atlantic Ocean, and the eastern and western Pacific margins (see, e.g., the synthesis by [Rochon et al., 1999](#), the compilation by [Marret and Zonneveld, 2003](#), and the overview of [de Vernal and Marret, 2007](#)).

Studies providing indications of dinocyst abundance in sediments reveal high concentrations (up to 10^6 cysts cm^{-3}) along continental margins (shelves, inland seas, estuaries) and decreasing concentrations offshore, suggesting higher dinocyst fluxes in neritic (near shore) areas than in open oceans (Figure 6).

The species assemblages show large variations from the equator to the poles, with maximum species diversity at low latitudes. Many species appear ubiquitous or cosmopolitan and occur in wide latitudinal ranges of both hemispheres, but several appear to be restricted to equatorial–tropical environments. Only a few taxa seem to occur exclusively at polar and subpolar latitudes. Some species have been reported only

from the Northern Hemisphere, whereas others only occur in the Southern Ocean. Moreover, the species composition of assemblages apparently differs in the Atlantic and Pacific oceans. Thus, there is some regionalization in dinocyst distributions (Figures 7–13).

In general, the relative abundance of taxa in dinocyst assemblages follows distribution patterns closely related to the latitudinal and to nearshore-to-open-ocean gradients. Multivariate analyses of assemblages clearly demonstrate relationships with sea-surface temperatures, whatever the spatial scale being considered (e.g., Marret and Zonneveld, 2003; Radi and de Vernal, 2008). The analyses of assemblage distribution patterns at

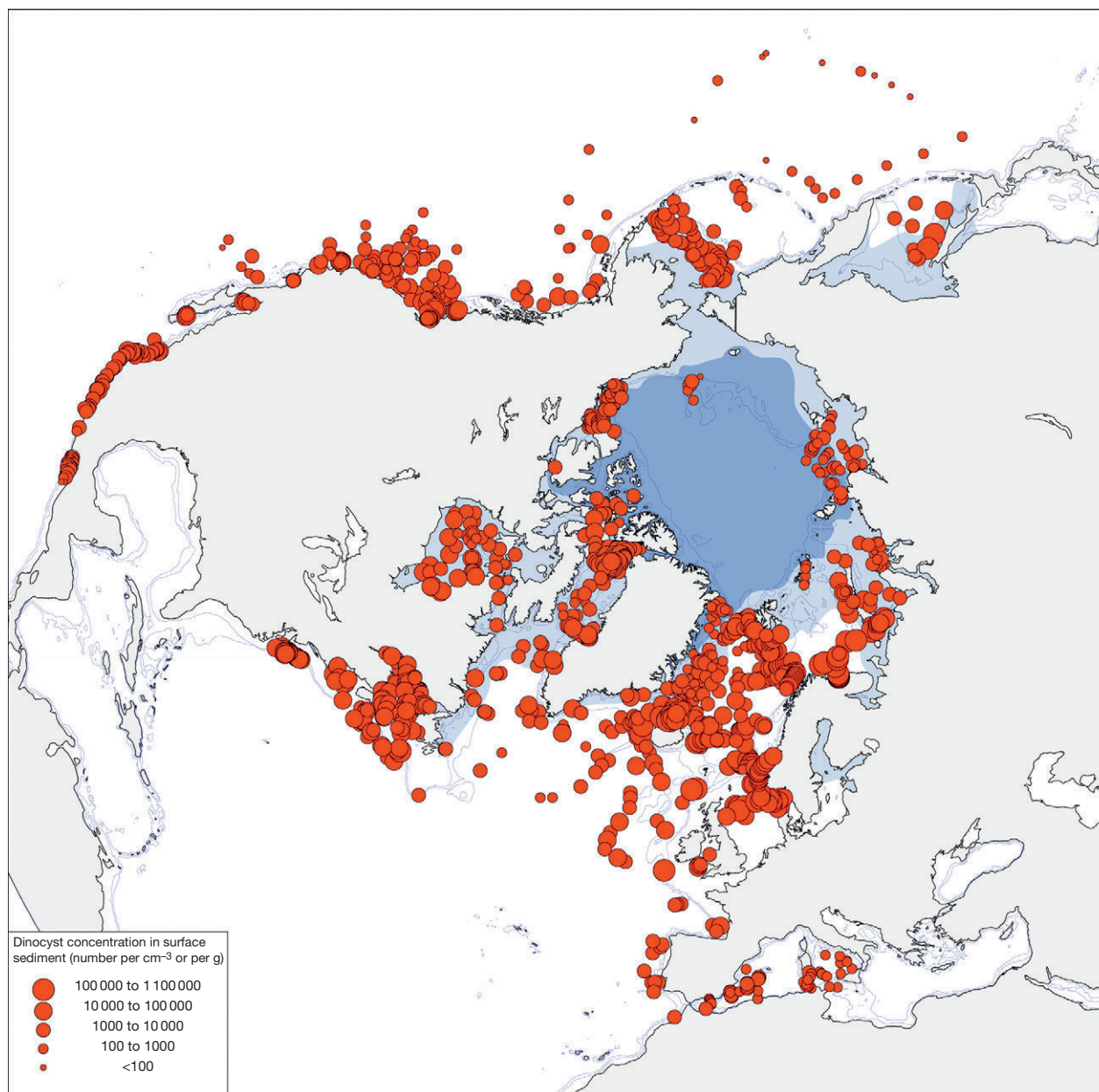


Figure 6 Dinocyst concentration in surface sediment of the Northern Hemisphere (data updated from de Vernal et al., 2005 and Radi and de Vernal, 2008; see dinocyst databases at <http://www.geotop.ca>). The shaded light and dark blue areas correspond, respectively, to the median (1979–2000) of the minimum and maximum sea-ice extent (sea-ice data from the National Snow and Ice Data Center at <http://nsidc.org>). The isobaths correspond to water depths of 200 and 1000 m.

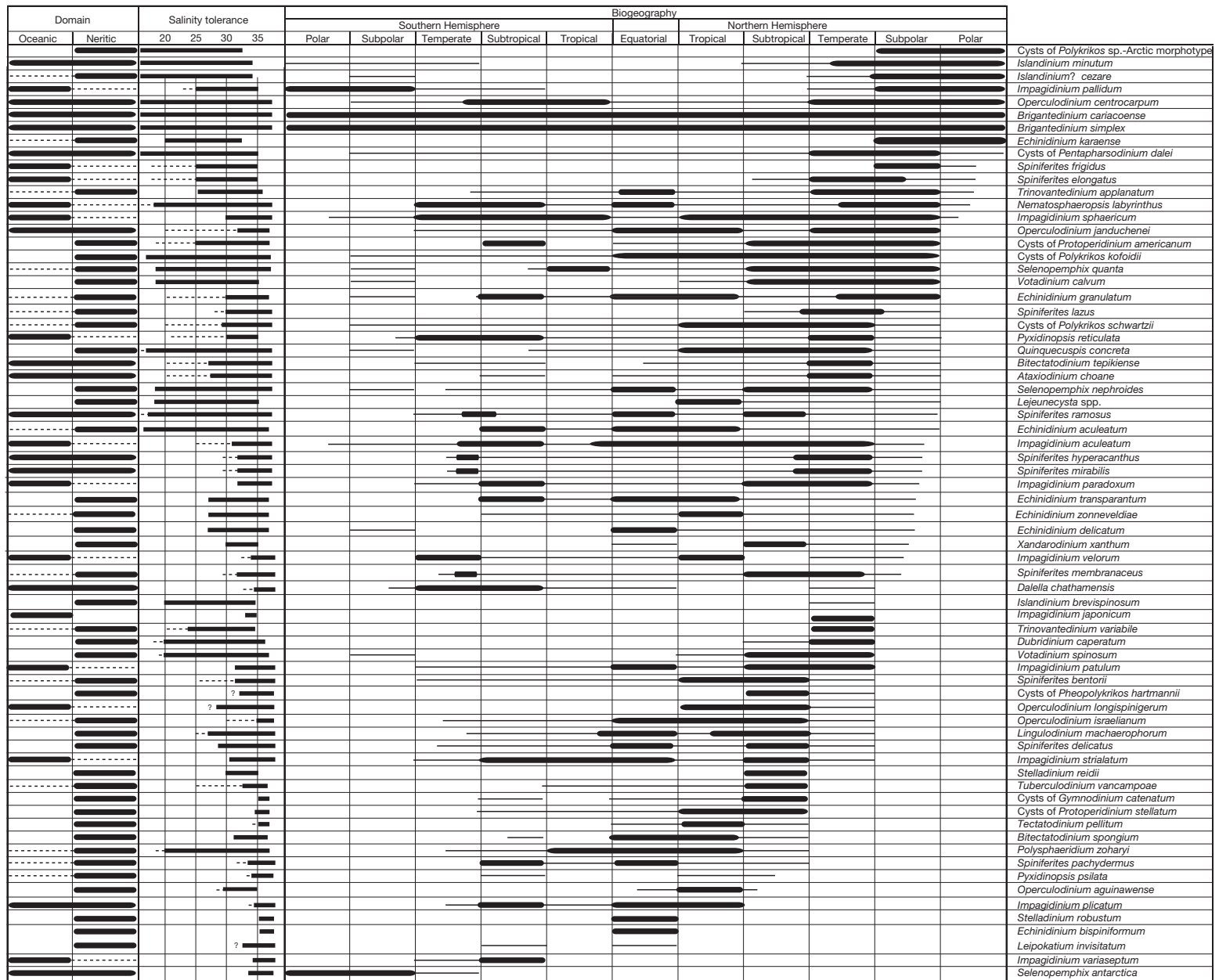


Figure 7 Latitudinal distribution of dinocyst taxa in surface sediments, salinity tolerance and neritic to oceanic occurrences (updated from Marret and Zonneveld, 2003; de Vernal et al., 2005; Radi and de Vernal, 2008).

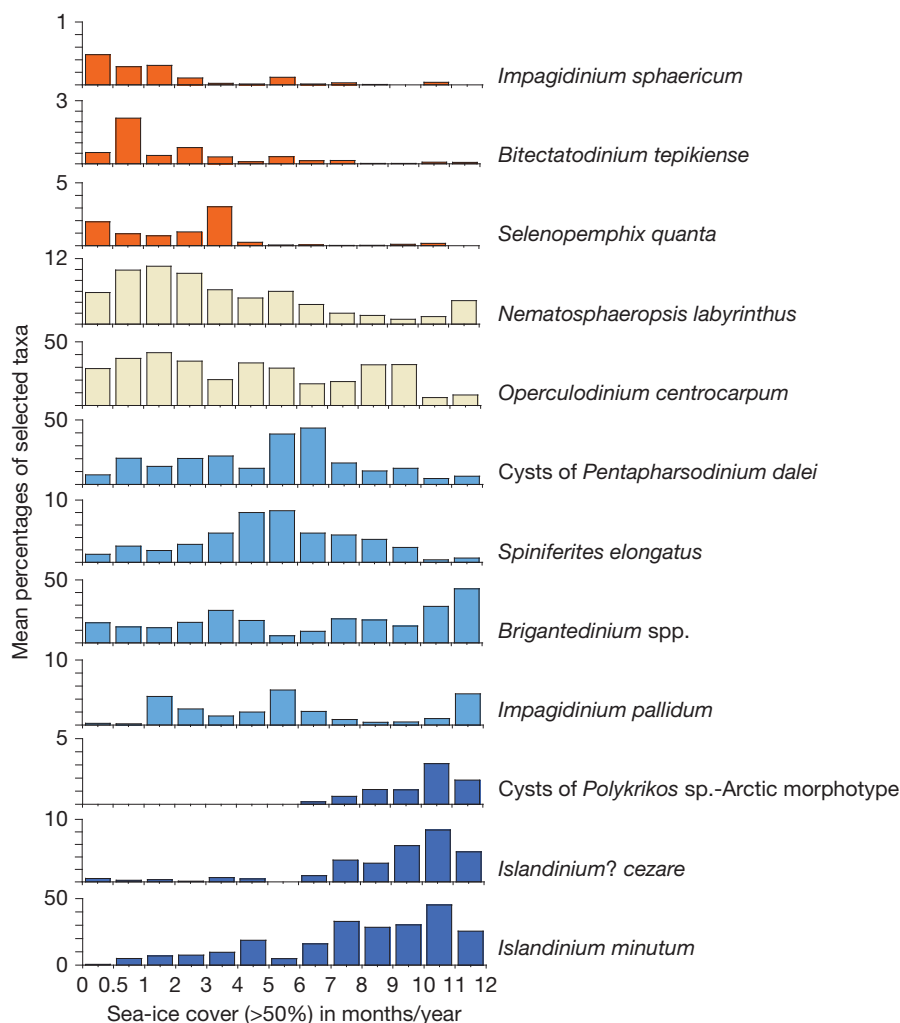


Figure 8 Mean percentage of selected dinocyst taxa versus sea-ice extent in months per year with more than 50% of ice coverage. Note that most other taxa do not occur in areas characterized by seasonal sea ice (data updated from de Vernal et al., 2005 and Radi and de Vernal, 2008; see dinocyst databases at <http://www.geotop.ca>).

regional scales (subhemispheric ocean basins) also show relationships between the cyst assemblages and the sea-surface salinity, annual amplitude of sea-surface temperatures, sea-ice cover, primary productivity, and upwelling intensity (e.g., de Vernal et al., 2005, 2008; Radi and de Vernal, 2004, 2008; Rochon et al., 1999). At more local scales (estuarine systems), relationships with eutrophication can also be evidenced (e.g., Dale and Dale, 2002). However, because the composition of assemblages may vary significantly from a region to another, the quantitative relationships between dinocyst assemblages and hydrographic or oceanic conditions at regional or local scales cannot be extrapolated to a global scale (Figures 9–13).

Ecological–Paleoecological Significance of Dinocysts in Sediment

The dinocysts represent only a fragmentary picture of the original dinoflagellate populations (e.g., Dale, 1976; Head, 1996; Matthiessen et al., 2005). The ecological affinities of dinocyst taxa are known only by default, more from

distribution of the assemblages and species on the sea floor than from the ecology of their living counterpart. There are only few studies documenting the ecology of nontoxic dinoflagellates and the works coupling information on planktonic dinoflagellates and fluxes of dinocysts to the sea floor are extremely rare. Nevertheless, it seems that the biogeographic distributions of cyst-forming dinoflagellates in surface waters and that of dinocysts in sediments are generally consistent with each other (e.g., Dodge, 1994). There is a general consistency with respect to onshore–offshore patterns and latitudinal gradients, which closely relate to salinity and temperature controlled by current patterns. However, the correspondence between observations of motile and cyst assemblages is not perfect, notably due to the fact that the motile dinoflagellates in the plankton assemblages correspond to an instantaneous time interval, whereas the cysts in upper first centimeter of surface sediments may represent several years, decades or centuries of sediment deposition.

In paleoceanographic studies, dinocysts are usually associated with productivity in the upper part of the water column at

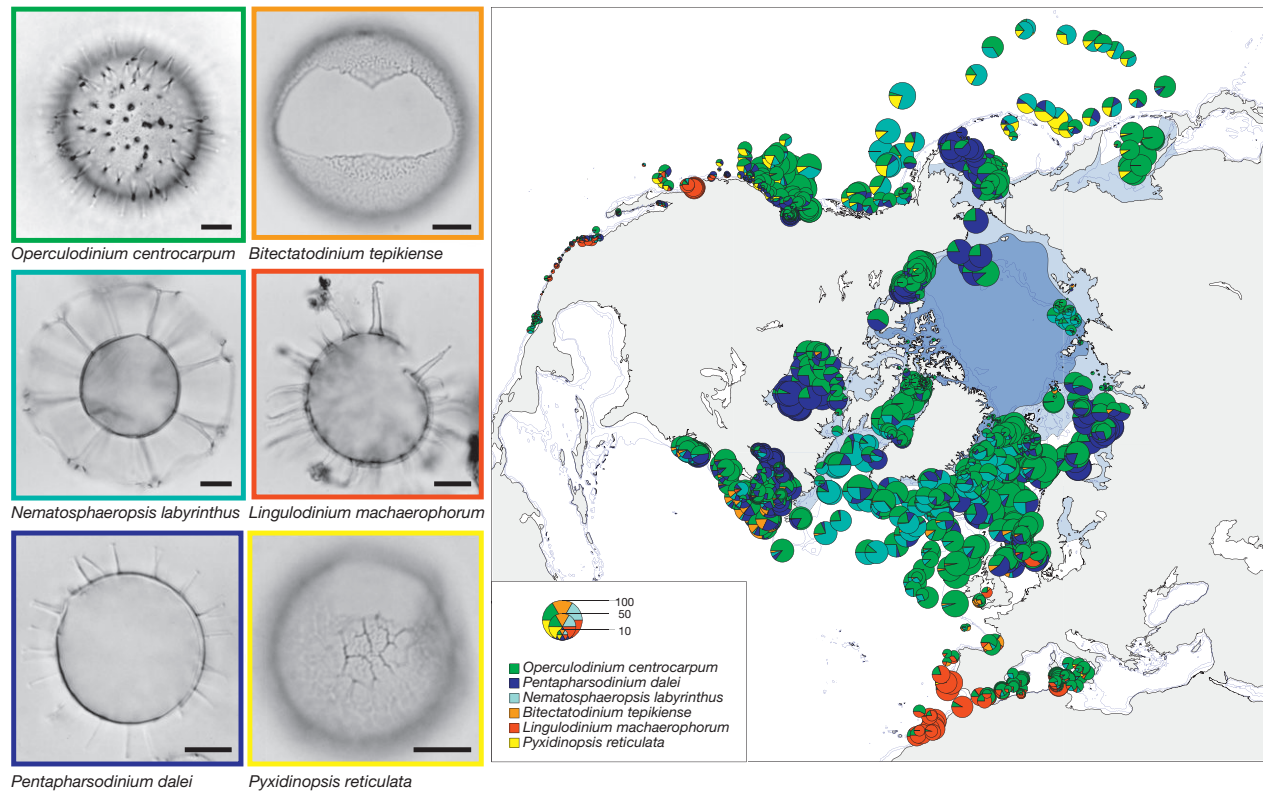


Figure 9 Photographs and distribution of selected dinocyst species belonging mostly to Gonyaulacales in surface sediment samples from the Northern Hemisphere (updated from de Vernal et al., 2005; Radi and de Vernal, 2008). The scale bars on the photographs correspond to 10 μm . The circle diameter illustrates the cumulative percentage of the selected species in the dinocyst assemblages.

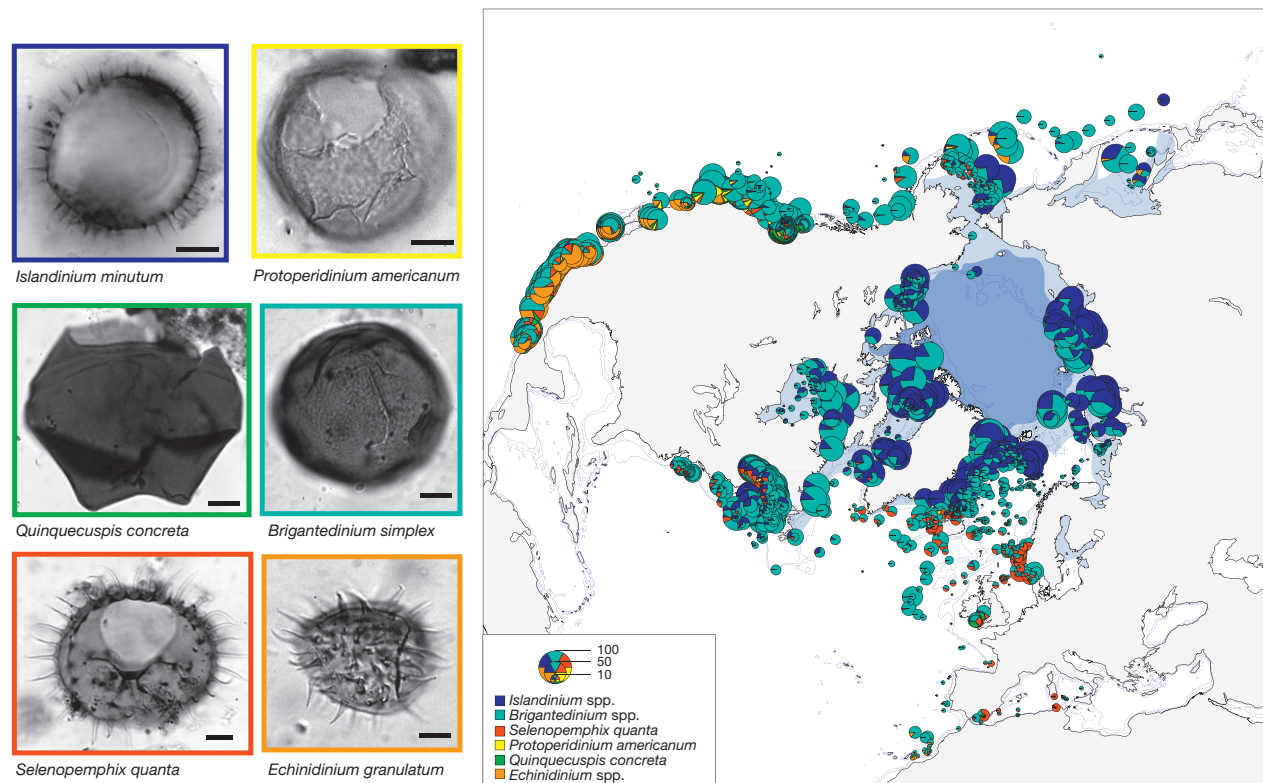


Figure 10 Photographs and distribution of selected dinocyst taxa belonging to Peridiniales in surface sediment samples from the Northern Hemisphere (updated from de Vernal et al., 2005; Radi and de Vernal, 2008). The scale bars on the photographs correspond to 10 μm . The circle diameter illustrates the cumulative percentage of the selected species in the dinocyst assemblages.

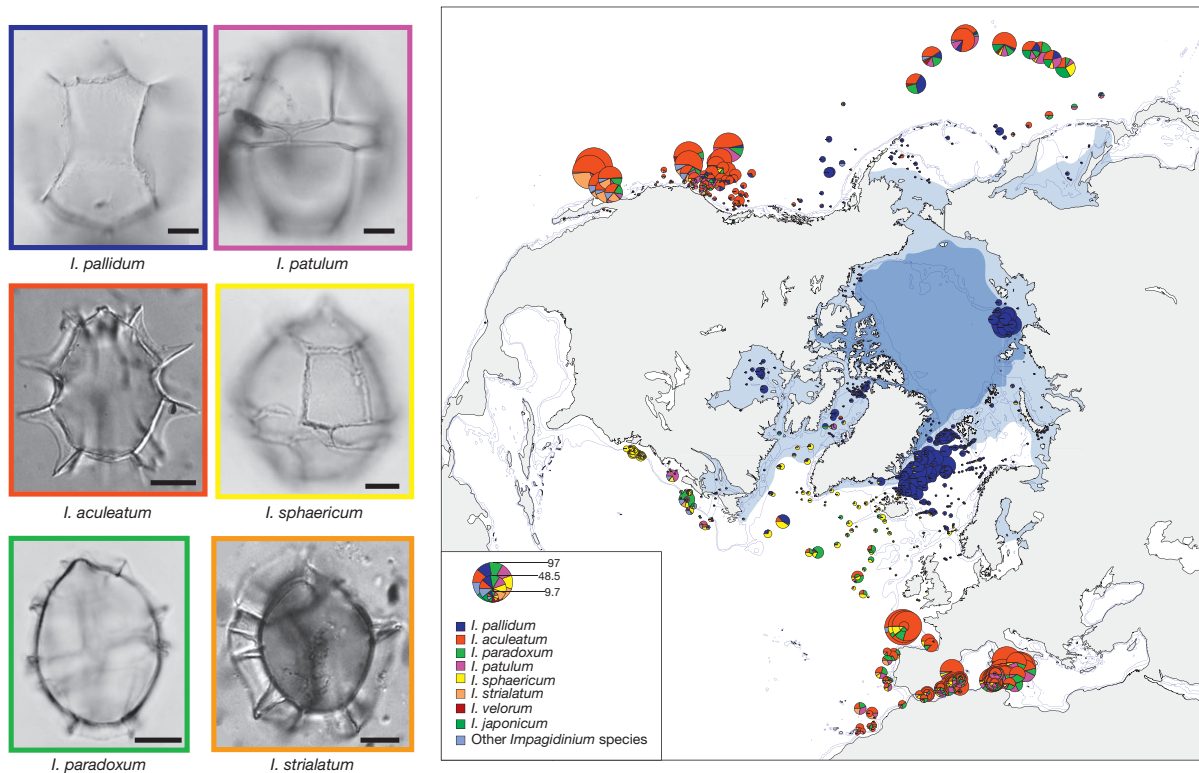


Figure 11 Photographs and distribution of selected dinocyst species of the genus *Impagidinium* (Order Gonyaulacales) in surface sediment samples from the Northern Hemisphere (updated from de Vernal et al., 2005; Radi and de Vernal, 2008). The scale bars on the photographs correspond to 10 μm . The circle diameter illustrates the cumulative percentage of the selected species in the dinocyst assemblages.

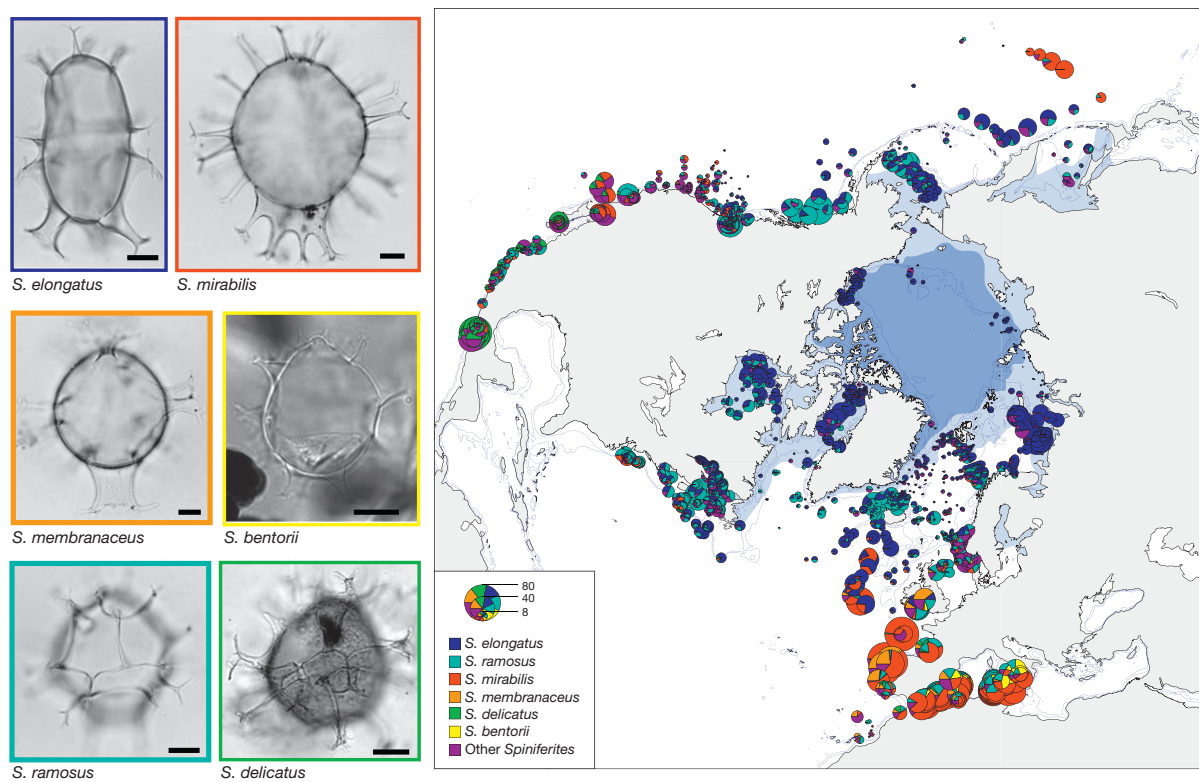


Figure 12 Photographs and distribution of selected dinocyst species of the genus *Spiniferites* (Order Gonyaulacales) in surface sediment samples from the Northern Hemisphere (updated from de Vernal et al., 2005; Radi and de Vernal, 2008). The scale bars on the photographs correspond to 10 μm . The circle diameter illustrates the cumulative percentage of the selected species in the dinocyst assemblages.

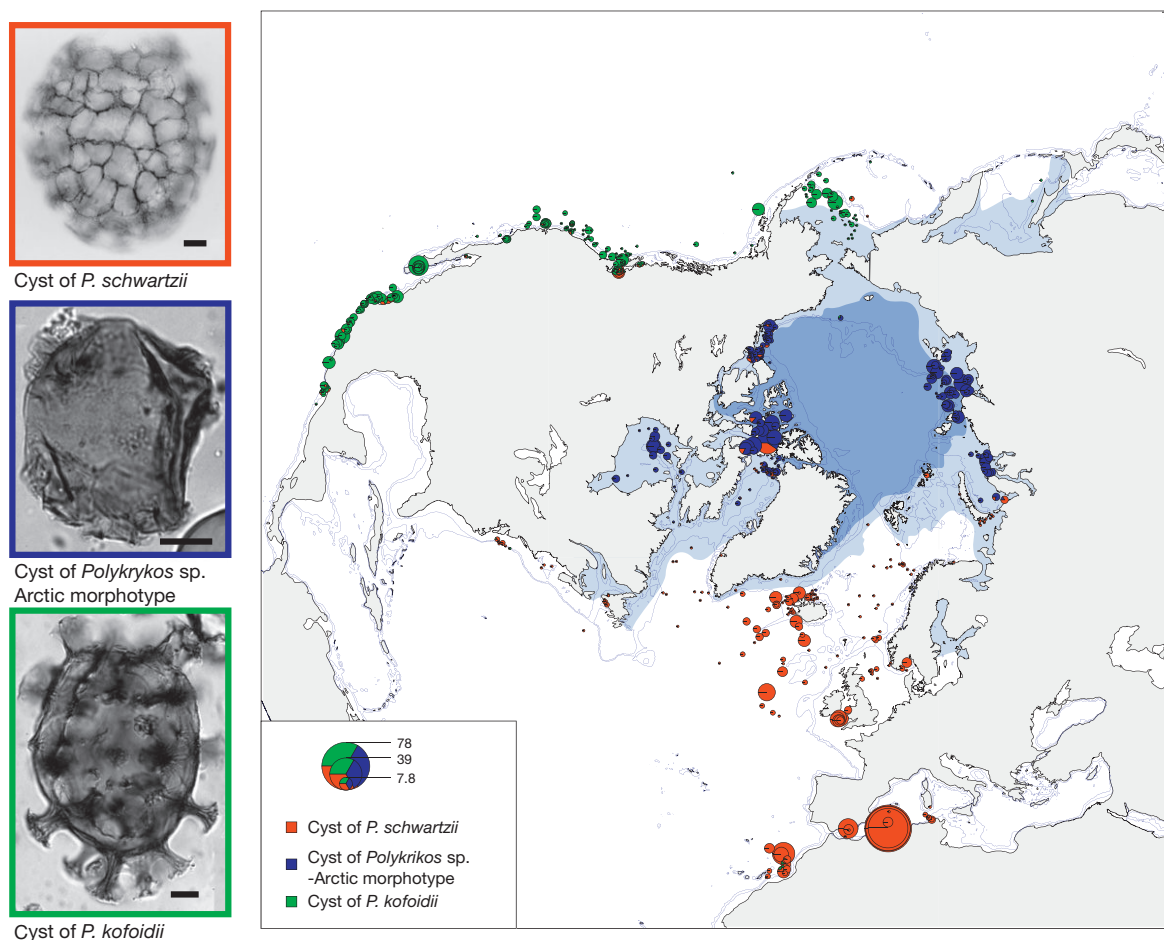


Figure 13 Photographs and distribution of selected dinocyst species of the genus *Polykrikos* (Order Gymnodiniales) in surface sediment samples from the Northern Hemisphere (updated from de Vernal et al., 2005; Radi and de Vernal, 2008). The scale bars on the photographs correspond to 10 μm . The circle diameter illustrates the cumulative percentage of the selected species in the dinocyst assemblages.

the site location. However, the sinking rate of individual cells or microscopic remains such as dinocysts is extremely slow (on the order of meters per day). Thus, incorporation within fecal pellets or marine snow (coherent aggregates of microscopic organic material having a density high enough to sink in the water column) constitutes mechanisms which likely explain the vertical fluxes of pelagic particles over hundreds or thousands of meters (e.g., Turner, 2002). Although dinocyst assemblages on the sea floor likely relate to sedimentation within marine snow and fecal pellets, some lateral transport with surface, intermediate or bottom currents, cannot be ruled out.

Dinocyst are made of refractory organic matter and usually preserve very well in marine sediments, unlike diatoms or foraminifers that are subject to dissolution of opal silica and calcium carbonate, respectively. From this point of view, dinocysts constitute an extremely useful proxy of ocean changes in many neritic environments and other regions of the world's oceans where calcium carbonate dissolution occurs because of a shallow lysocline and/or oxidation of organic rich sediments that fosters high pCO_2 . However, the cyst wall of some dinocyst taxa seems to be more susceptible to selective degradation when using strong oxidation techniques (e.g., acetolysis, KOH). Such techniques are occasionally used in palynology

but are avoided in the study of dinocysts. Good preservation of dinocyst taxa is generally assumed, although it is possible that selective degradation occurs in sediment or at the water–sediment interface under oxygen-rich conditions (cf. Zonneveld et al., 2001). The dinocyst taxa that are the most susceptible to degradation from oxidation are the cysts of *Protoperidinium* spp. and *Echinidinium* spp. (i.e., Peridinales taxa that are associated with heterotrophic productivity). These taxa are typically characterized by thin and brownish cyst walls.

Applications of Marine Dinocysts in Quaternary Sciences

Despite the caveats mentioned above, dinocysts constitute excellent tracers of sea-surface conditions. They are usually well preserved in sediments and the assemblages are characterized by relatively high species diversity, even in nearshore environments and in the Arctic and subarctic domains. From these points of view, dinocysts are complementary to calcareous microfossils such as planktonic foraminifers and coccoliths that are stenohaline (have narrow limits of salt concentration in water) and are used primarily in paleoceanography in the

low to middle latitudes. During recent decades, there have been increasing numbers of studies based on dinocysts in the field of Quaternary geology.

Hydrographic Reconstruction in Epicontinental Environments and Along the Continental Margin

In estuaries and marginal marine environments, dinocysts are particularly good proxies to document changes in sea-surface conditions, since they occur in a wide range of salinity and record both variations in salinity and temperature. Among dinocyst records from inland seas, the data from the Marmara and Black seas are particularly spectacular (e.g., [Mudie et al., 2002](#)) because the assemblages reveal extremely large salinity changes and because some taxa show morphological variations that reflect endemism that is likely due to isolation and low salinity ([Figure 14](#)).

In estuaries or deltaic environments, the various proportions of salinity-tolerant and stenohaline species can be used to estimate changes in salinity and freshwater discharges, for example off eastern coast of Canada, in the northwestern North Atlantic, or off Africa in the Equatorial Atlantic ([Bouimetarhan et al., 2009](#); [de Vernal and Marret, 2007](#)).

Paleoceanographic Reconstructions

The development of reference dinocyst databases from the studies of surface sediment samples has allowed the development of quantitative techniques for the reconstruction of sea-surface conditions such as temperature and salinity. Various attempts have been made to adapt regression and multiregression techniques (i.e., the conventional transfer functions), the best analog method or neural network approaches. Based on various trials, it seems that extrapolation approaches (neural networks and regression) that rely on calibration between the

assemblages and given oceanographical parameters strongly depend on the definition of the database ([Bonnet et al., 2010](#); [Guiot and de Vernal, 2007](#)). As a consequence, although extrapolation techniques may yield very good validation results, the reconstructions can differ significantly, depending on the calibration data set. Therefore, most paleoceanographic reconstructions from dinocyst data have been made on conservative grounds, using the modern analog technique that does not involve any extrapolation, but simply uses the similarity between the fossil spectrum and the modern assemblages to infer past sea-surface conditions (for a discussion, cf. [de Vernal et al., 2001, 2005, 2008](#)).

Quantitative reconstructions of sea-surface temperature, salinity and sea-ice cover have been made from many Quaternary sequences in the North Atlantic and data compilations are available for the Last Glacial Maximum (LGM), 21 000 years ago ([de Vernal et al., 2005](#)). In addition to significant changes in temperature, the dinocyst-based reconstructions point toward variations in the annual amplitude of temperature, and change in salinity and sea-ice cover ([Figure 15](#)).

Eutrophication, Productivity, and Upwelling

Dinocysts include fossil remains from both microzooplanktonic and phytoplanktonic communities. Thus, the dinocyst assemblages, together with the overall fluxes, depend on the primary productivity and the structure of the planktonic population. Studies from the fjords of northwest Europe, and along the coasts of Japan and America, have shown increased fluxes of dinocysts with higher percentages of heterotrophic taxa (cysts of *Protoperidinium* sp., notably). These changes in assemblages appear related to recent eutrophication, notably due to human activities. Some attempts have been made to distinguish the effects of eutrophication from those of pollution in dinocyst assemblages, but the results are equivocal (e.g., see [Dale and Dale, 2002](#)).

In ocean environments where upwelling results in high primary productivity and where the phytoplankton is dominated by diatoms, dinocyst fluxes are high and are characterized by the abundance of heterotrophic taxa over autotrophic ones (e.g., [Pospelova et al., 2008](#); [Radi and de Vernal, 2004](#)). Notably, downcore analyses performed off the African margin have shown variations in the abundance of Peridinales over the last hundred thousand years, which indicate changes in productivity and upwelling strength in relation to orbital forcing ([Figure 16](#); [Höll et al., 2000](#)).

Conclusion

Fossil dinocysts are mainly known from marine sediments and appear to be particularly abundant along continental margins (estuaries, continental shelves and slopes, epicontinental seas). Dinocysts have been widely used in biostratigraphy and paleoecology of the Mesozoic and Tertiary. In the field of Quaternary paleoceanography, paleoclimatology, and paleoecology, the study of dinocysts is of growing interest. Because they are very resistant to decomposition, dinocysts are generally well preserved in sediments, unlike the dissolution that

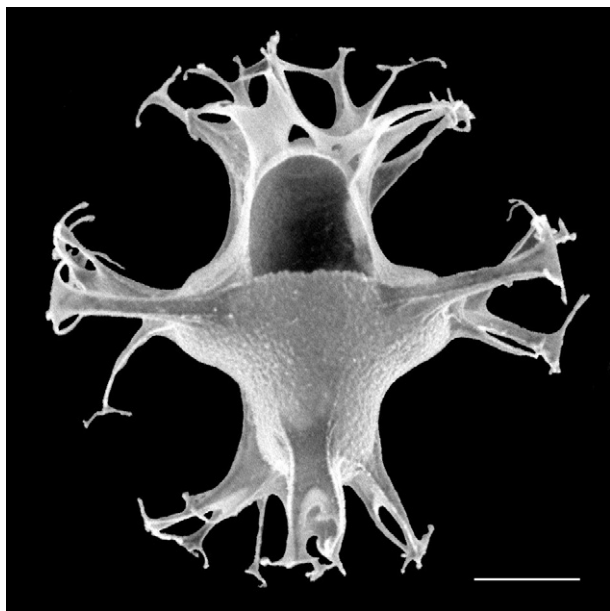


Figure 14 Photograph of *Spiniferites cruciformis*. Specimen from sediment collected in the Black Sea. Scale bar = 20 μm .

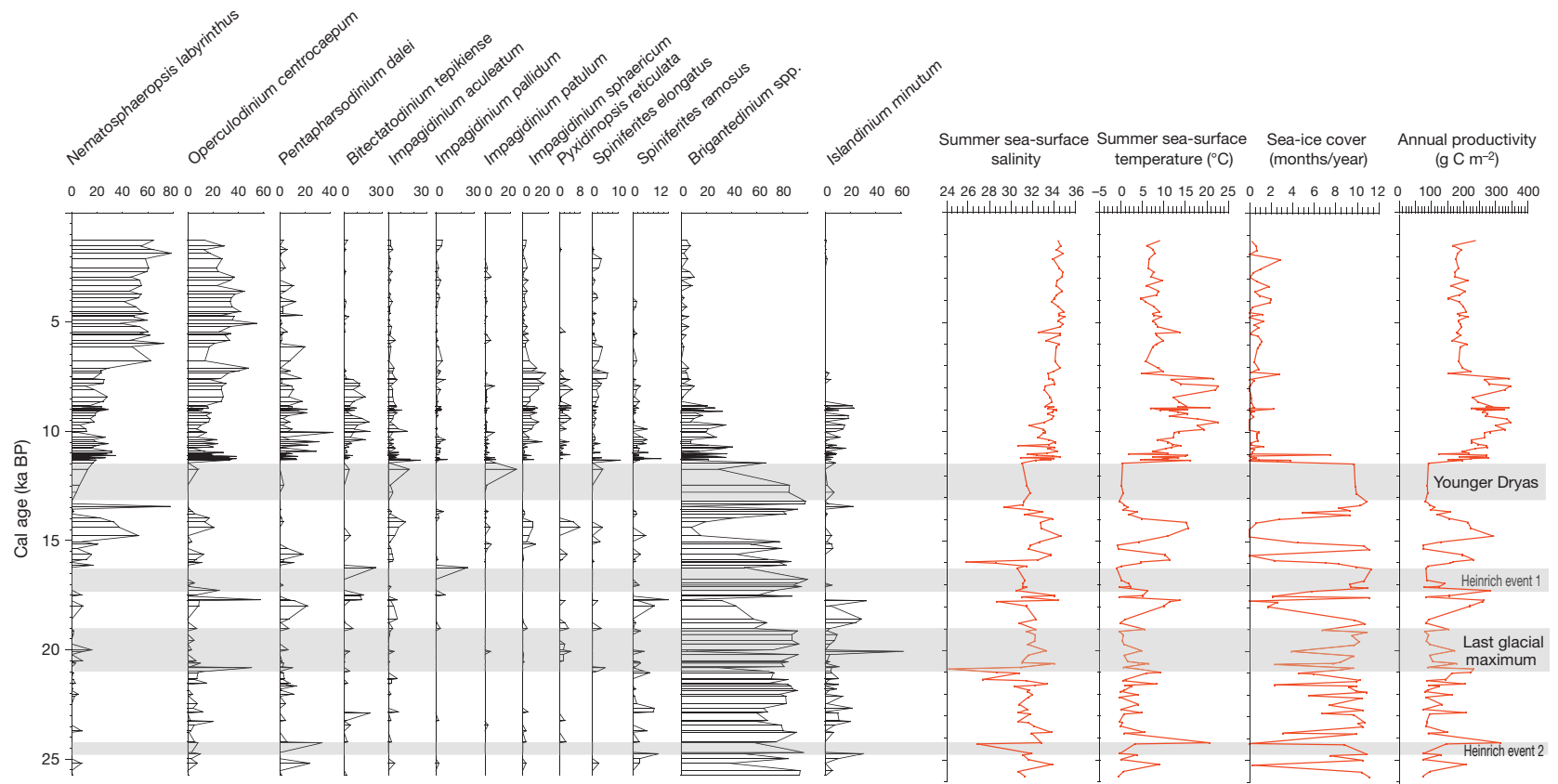


Figure 15 Example of late Quaternary dinocyst assemblages in the Labrador Sea, northwest North Atlantic, and corresponding reconstruction of sea-surface salinity, temperature, sea-ice cover, and mean annual productivity based on the modern analog technique. The figure shows the record from core HU91-045-094 (45°41.14 W 50°12.26 N; water depth = 3448 m). For more detail on the record, see for example, [Radi and de Vernal \(2008\)](#).

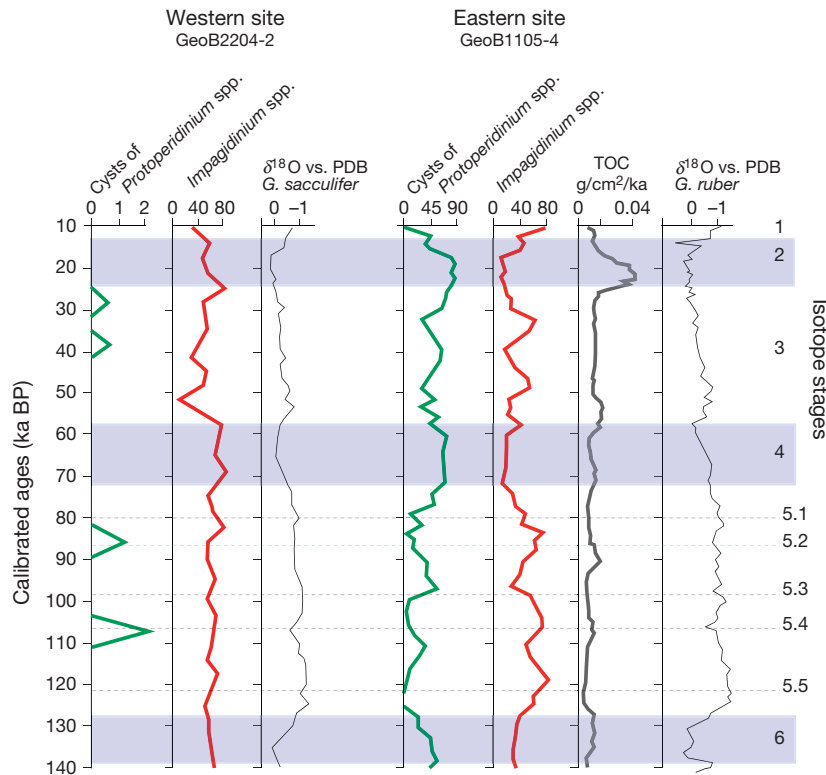


Figure 16 Example of dinocyst records representative of productivity and upwelling changes during the middle and late Pleistocene in the tropical Atlantic. The western site (core GeoB2204-2, 8°32'S, 34°01'W; water depth = 2072 m) is from a low productivity area and the eastern site (core GeoB1105-4: 01°40'S, 12°26'W; water depth = 3225 m) was collected in a productive upwelling region. In the sediments, the cysts of *Protoperidinium* spp. are associated with high productivity in surface waters whereas *Impagidinium* spp. are associated with oligotrophic conditions. The blue bands correspond to glacial episodes, which were marked by relatively high productivity in the eastern tropical Atlantic. Data redrafted from Höll et al. (2000).

may affect calcareous or siliceous biological remains. Moreover, the development of reference databases from surface sediment samples (e.g., Dale, 1996; de Vernal et al., 2001, 2005, 2008; Radi and de Vernal, 2008; Rochon et al., 1999; Marret and Zonneveld, 2003) has led to the documentation of strong relationships between the distribution of dinocyst assemblages and sea-surface parameters, including productivity and hydrographic conditions. Quantitative approaches, such as the best analog methods, have permitted the quantitative reconstruction of temperature, salinity and sea-ice cover extent. For example, hydrographic maps of the northern North Atlantic during the LGM have been established using dinocyst data, and many regional reconstructions are currently being developed. Other applications of dinocysts in Quaternary sciences include the reconstruction of hydrological changes from the study of cores collected in epicontinental seas or deltaic environments. Dinocyst assemblages may also provide insights into variation of the trophic character of upper water masses, leading to the identification of eutrophication in nearshore environments, and to estimations of changes in productivity in upwelling areas.

See also: **Paleoceanography:** Paleoceanography An Overview.
Paleoceanography, Biological Proxies: Coccolithophores;
 Marine Diatoms. **Quaternary Stratigraphy:** Continental
 Biostratigraphy.

References

- Bonnet S, de Vernal A, Hillaire-Marcel C, Radi T, and Husum K (2010) Variability of sea-surface temperature and sea-ice cover in the Fram Strait over the last two millennia. *Marine Micropaleontology* 74: 59–74.
- Bouimmetarhan I, Marret F, Dupont L, and Zonneveld K (2009) Dinoflagellate cyst distribution in marine surface sediments off West Africa (17–6° N) in relation to sea-surface conditions, freshwater input and seasonal coastal upwelling. *Marine Micropaleontology* 71: 113–130.
- Dale B (1976) Cyst formation, sedimentation, and preservation: Factors affecting dinoflagellate assemblages in recent sediments from Trondheimsfjord, Norway. *Review of Palaeobotany and Palynology* 22: 39–60.
- Dale B (1983) Dinoflagellate resting cysts: 'Benthic plankton'. In: Fryxell GA (ed.) *Survival Strategies of Algae*, pp. 69–136. London: Cambridge University Press.
- Dale B (1996) Dinoflagellate cyst ecology: Modelling and geological applications. In: Jansonius J and McGregor DC (eds.) *Palynology: Principles and Applications*. Dallas, TX: American Association of Stratigraphic Palynologists Foundation.
- Dale B and Dale A (2002) Environmental application of dinoflagellate cysts and acritarchs. In: Haslett SK (ed.) *Quaternary Environmental Micropaleontology*, pp. 207–240. London: Arnold.
- De Schepper S and Head MJ (2008) Age calibration of dinoflagellate cyst and acritarch events in the Pliocene–Pleistocene of the eastern North Atlantic (DSDP Hole 610A). *Stratigraphy* 5: 131–167.
- de Vernal A, Eynaud F, Henry M, et al. (2005) Reconstruction of sea-surface conditions at middle to high latitudes of the Northern Hemisphere during the Last Glacial Maximum (LGM) based on dinoflagellate cyst assemblages. *Quaternary Science Reviews* 24: 897–924.
- de Vernal A, Henry M, Matthiessen J, et al. (2001) Dinoflagellate cyst assemblages as tracers of sea-surface conditions in the northern North Atlantic, Arctic and sub-Arctic seas: The new 'n = 677' database and application for quantitative paleoceanographical reconstruction. *Journal of Quaternary Science* 16: 681–699.

- de Vernal A, Hillaire-Marcel C, Solignac S, Radi T, and Rochon A (2008) Reconstructing sea-ice conditions in the Arctic and subarctic prior to human observations. In: Weaver E (ed.) *Arctic Sea Ice Decline: Observations, Projections, Mechanisms, and Implications*. AGU Monograph Series 180: pp. 27–45. Washington, DC: AGU.
- de Vernal A, Londeix L, Harland R, et al. (1992) The Quaternary organic walled dinoflagellate cyst of the North Atlantic Ocean and adjacent seas: Ecostratigraphic and biostratigraphic records. In: Head MJ and Wrenn JH (eds.) *Neogene and Quaternary Dinoflagellate Cysts and Acritarchs*. Dallas, TX: American Association of Stratigraphic Palynologists Foundation.
- de Vernal A and Marret F (2007) Organic-walled dinoflagellates: Tracers of sea-surface conditions. In: Hillaire-Marcel C and de Vernal A (eds.) *Proxies in Late Cenozoic Paleoceanography*, pp. 371–408. Amsterdam: Elsevier.
- Dodge JD (1994) Biogeography of marine armoured dinoflagellates and dinocysts in the NE Atlantic and North Sea. *Review of Palaeobotany and Palynology* 84: 169–180.
- Edwards L (1993) Dinoflagellates. In: Lipps JH (ed.) *Fossil Prokaryotes and Protists*, pp. 105–129. Boston, MA: Blackwell.
- Evitt W (1963) A discussion and proposals concerning fossil dinoflagellates, hystrichospheres and acritarchs. *Proceedings of the National Academy of Sciences* 49: 158–164.
- Evitt WR (1985) *Sporopollenin Dinoflagellate Cysts – Their Morphology and Interpretation*. Austin, TX: American Association of Stratigraphic Palynologists Foundation 333 pp.
- Fensome RA, Taylor FJR, Norris G, Sarjeant WAS, Wharton DI, and Williams GL (1993) *A Classification of Living and Fossil Dinoflagellates*. Micropaleontology Special Publication Number 7, 351 pp, New York: American Museum of Natural History.
- Fensome RA and Williams GL (2004) *The Lentin and Williams Index of Fossil Dinoflagellate*, Contribution Series 42, 909 pp. Dallas, TX: American Association of Stratigraphic Palynologists Foundation.
- Groupe de travail 'Dinoflagellés' (1986) Guide pour la détermination de kystes de dinoflagellés fossiles, Le complexe. Gonyaulacysta, Mémoire Elf Aquitaine, 12, 478 pp.
- Guiot J and de Vernal A (2007) Transfer functions: Methods for quantitative paleoceanography based on microfossils. In: Hillaire-Marcel C and de Vernal A (eds.) *Proxies in Late Cenozoic Paleoceanography*, pp. 523–563. Amsterdam: Elsevier.
- Head MJ (1996) Modern dinoflagellate cysts and their biological affinities. In: Jansonius J and McGregor DC (eds.) *Palynology: Principles and Applications*, pp. 1197–1248. Dallas, TX: American Association of Stratigraphic Palynologists Foundation.
- Head MJ, Edwards LE, and Steidinger KA (1989) *Short Course on Neogene-Recent Dinoflagellates, Course Manual*. Dallas, TX: American Association of Stratigraphic Palynologists Foundation.
- Höfl C, Zonneveld KAF, and Willems H (2000) Organic-walled dinoflagellate cyst assemblages in the tropical Atlantic Ocean and oceanographical changes over the last 140 ka. *Palaeogeography, Palaeoclimatology, Palaeoecology* 160: 69–90.
- Marret F and Zonneveld K (2003) Atlas of modern global organic-walled dinoflagellate cyst distribution. *Review of Palaeobotany and Palynology* 125: 1–200.
- Matthiessen J, de Vernal A, Head M, Okolodkov Y, Zonneveld K, and Harland R (2005) Modern organic-walled dinoflagellate cysts in Arctic marine environments and their (paleo-) environmental significance. *Paläontologische Zeitschrift* 79(1): 3–51.
- Mudie PJ, Rochon A, Aksu AE, and Gillespie H (2002) Dinoflagellate cysts, freshwater algae and fungal spores as salinity indicators in Late Quaternary cores from Marmara and Black seas. *Marine Geology* 190: 203–231.
- Pfiester LA and Anderson DM (1987) Dinoflagellate reproduction. In: Taylor FJR (ed.) *The Biology of Dinoflagellates*. Oxford: Blackwell.
- Pospelova V, de Vernal A, and Pedersen TF (2008) Distribution of dinoflagellate cysts in surface sediments from the northeastern Pacific Ocean (43–25°N) in relation to sea-surface temperature, productivity and coastal upwelling. *Marine Micropaleontology* 68: 21–48.
- Powell AJ (ed.) (1992) *A Stratigraphic Index of Dinoflagellate Cysts*. In: *British Micropaleontological Society Publication Series*. London: Chapman & Hall 290 pp.
- Radi T and de Vernal A (2004) Dinocyst distribution in surface sediments from the northeastern Pacific margin (40–60°N) in relation to hydrographic conditions, productivity and upwelling. *Review of Palaeobotany and Palynology* 128: 163–193.
- Radi T and de Vernal A (2008) Dinocysts as proxy of primary productivity in mid-high latitudes of the Northern Hemisphere. *Marine Micropaleontology* 68: 84–114.
- Rochon A, de Vernal A, Turon J-L, Matthiessen J, and Head MJ (1999) Distribution of dinoflagellate cysts in surface sediments from the North Atlantic Ocean and adjacent seas in relation to sea-surface parameters. *Special Contribution Series of the American Association of Stratigraphic Palynologists Foundation* 35: 1–152.
- Stoecker DK (1999) Mixotrophy among dinoflagellates. *Journal of Eukaryotic Microbiology* 46: 397–401.
- Taylor FJR and Pollinger U (1987) The ecology of dinoflagellates. In: Taylor FJR (ed.) *The Biology of Dinoflagellates*. Oxford: Blackwell.
- Turner JT (2002) Zooplankton fecal pellets, marine snow and sinking phytoplankton blooms. *Aquatic Microbial Ecology* 27: 57–102.
- Vink A (2004) Calcareous dinoflagellate cysts in South and equatorial Atlantic surface sediments: Diversity, distribution, ecology and potential for palaeoenvironmental reconstruction. *Marine Micropaleontology* 50: 43–88.
- Wall D and Dale B (1966) 'Living fossils' in western Atlantic plankton. *Nature* 211: 1025–1026.
- Wall D and Dale B (1968) Modern dinoflagellate cysts and evolution of the Peridinales. *Micropaleontology* 14: 265–304.
- Wall D, Dale B, Lohmann GP, and Smith WK (1977) The environmental and climatic distribution of dinoflagellate cysts in the North and South Atlantic Oceans and adjacent seas. *Marine Micropaleontology* 2: 121–200.
- Zonneveld KAF, Versteegh GJM, and De Lange GJ (2001) Palaeoproductivity and post-depositional aerobic organic matter decay reflected by dinoflagellate cyst assemblages of the Eastern Mediterranean S1 sapropel. *Marine Geology* 172: 181–195.

Marine Diatoms

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Introduction

The first genomic blueprint of a diatom confirms that diatoms evolved when a heterotroph, a single-cell microbe, engulfed a red alga. The two became one organism (endosymbiosis) and swapped some genetic material to create a new hybrid genome (Armbrust et al., 2004). Their origin, according to genetic and sedimentary evidence, must have occurred earlier than their first fossil record in the Jurassic. One possibility is that their origin may be related to the end-Permian mass extinction, after which many marine niches were opened.

Although individual cells are mostly in the size range of 10–200 μm , many planktonic diatoms form multicellular chains that can be several millimeters in length, occasionally exceeding a centimeter (Figure 1). Both benthic and planktonic forms exist, but because they are autotrophic (i.e., they photosynthesize carbohydrates in the presence of sunlight), they are restricted to the photic zone and are generally considered phytoplankton (photosynthetic organisms that live by floating in seawater).

Most species reproduce by mitosis, a process that repeats itself until a size threshold is attained. Then, they regain their size by auxospore formation. Cell division can occur every 4–8 h, leading to a billion cells in only 10 days.

Diatoms are formally classified as belonging to the division Chrysophyta, class Bacillariophyceae, and are divided in two orders: Centrales, radially symmetric, and Pennales, bilaterally symmetric (Figure 2). Their taxonomy is based on the form and ornamentation of their frustules, an advantage to paleontology since the same features can be used to define extant and extinct species (Hustedt, 1962). It is estimated that there are more than 200 genera and approximately 100 000 extant (living) species.

When nutrients and light conditions in the upper mixed layer are favorable, planktonic diatoms quickly dominate phytoplankton communities (bloom). As such, they are often ecologically classified as opportunistic r-strategists. In temperate continental shelves, a spring bloom occurs when sufficient light and nutrients are available. It is, however, in coastal, equatorial, and high-latitude upwelling regions that sudden population explosions occur. Upon depletion of nutrients, diatom cells typically increase in sinking rate or form resting spores (RS) and fall to the sediment (bust; Raymont, 1980; Figure 3).

Nitrogen has long been considered a limiting nutrient. However, genome work has revealed that some diatoms have a nitrogen waste pathway common to animal species that has never been seen before in photosynthetic organisms. This reduces their dependence on nitrogen from the surrounding waters. But what typically causes the end of diatom blooms is a lack of silicon (Si). Si is a major requirement for diatoms, and

it is not as efficiently regenerated as nitrogen or phosphorus. This is the reason why they play a key role in the regulation of the silicon cycle in the modern ocean.

Many species have special requirements, such as vitamins and micronutrients, which may explain the occurrence of high-nitrate low-chlorophyll (HNLC) areas in the ocean. Iron (Fe) enrichment experiments, both in culture and *in situ*, show a diatom growth response in a matter of days and reveal that Si drawdown by phytoplankton greatly exceeds that of nitrate, suggesting that Si is the ultimate limiting nutrient for diatoms under Fe-rich conditions in HNLC areas of the ocean (Peterson and Whitney, 2003). Fe limitation was also shown to increase the Si/C ratio and half Si/N and Si/P uptake by diatoms in HNLC regions, as well as increase the abundance of large diatoms that grow thicker shells. Coastal regions, always believed not to be limited by Fe, were also proved to have phytoplankton growth constrained by iron limitation (Hutchings et al., 2002).

The study of diatoms preserved/buried on the seafloor is of both economic and scientific interest. The economic interest comes from the exploration of diatomites, as well as from the fact that much petroleum is of diatom origin. Nevertheless, the most consequential activity of marine diatoms is their possible influence on climate. Oceans contribute about half of the global annual primary production (PP), a process by which carbon dioxide (CO_2) is converted into organic carbon and which is mostly carried out by phytoplankton (Falkowski et al., 2000). In modern oceans, diatoms are dominant in the high latitudes, and in equatorial and coastal upwelling regimes, the latter of which concentrate 80–90% of the ocean PP. This makes them the most important group of phytoplankton, responsible for almost half of marine PP. Due to their bloom-and-bust behavior, diatoms are believed to play a disproportionately important role in the export of carbon from oceanic surface waters (the biological pump). Moreover, Joint Global Ocean Flux Study (JGOFS) process studies support the hypothesis that highly productive areas are potential major centers of carbon sequestration. Furthermore, tests of marine diatoms growing under different CO_2 concentrations suggest potentially important biotic feedbacks as the level of global CO_2 increases and foresee a vital role for diatoms as mediators of global warming.

Paleoecology/Paleoceanography

Diatoms' small size and the fact that a high proportion of fossil taxa are still extant make it possible to use quantitative and statistical methods to attempt paleoenvironmental reconstructions from small samples.

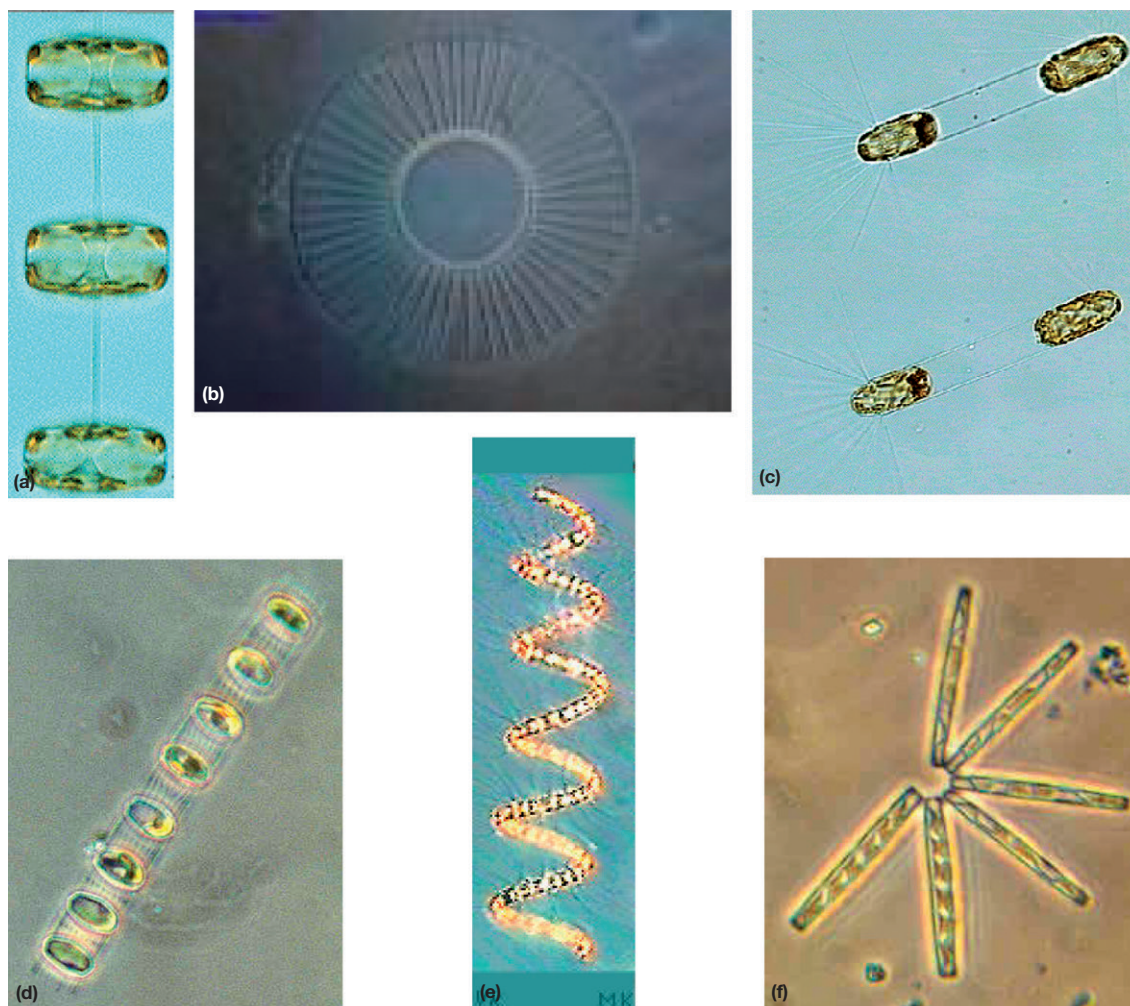


Figure 1 Diatoms may be solitary free-living cells or attached to other cells to form many different types of colonies. (a) Chain colonies of *Thalassiosira nordenskiöldii* Cleve. Cells connected by mucilaginous threads; (b) single cell of *Planktoniella sol* (Wallich) Schütt; (c) single cells of *Corethron criophilum* Castracane; and (d) chain colonies of *Skeletonema costatum* (Greville) Cleve. Cells connected by marginal spines (<http://www.dnr.state.md.us/bay/cblife/algae/diatom>); (e) helical colony of *Chaetoceros debilis* Cleve; and (f) zigzag chain of cells of *Thalassionema nitzschoides* Grunow.

Methods

Quantitative methods

The first step is to start with a sample of known volume and weight so that diatom abundance ($\# \text{valves g}^{-1}$) and diatom accumulation rate (DAR: number of valves per cm^{-2} per ka) can be determined. The second step is the identification, to the most specific taxonomic level possible, of at least 300 individuals per sample. When diversity is high, the count number should be increased, as needed, to the level where no additional species are encountered.

The major problem faced by diatom paleoecologists is opal solubility. Silicon is undersaturated all over and all along the water column. After death and disintegration of the protective organic skin, opal frustules suffer dissolution during descent through the water column, on the sediment surface and during diagenesis. In the water column, regeneration rates depend on diatom type and bacterial assemblage, suggesting that regeneration is regulated by bacterial activity, which in turn depends on the initial physiological condition of the diatom

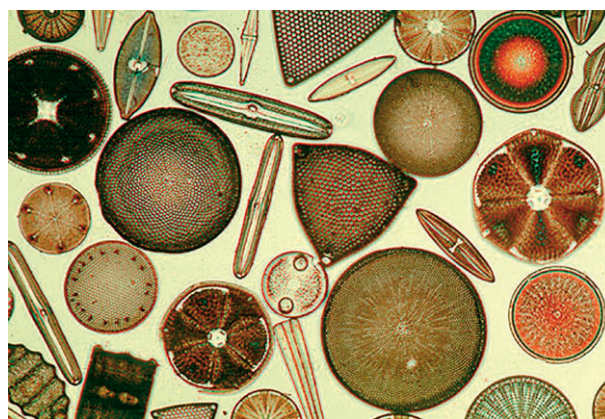


Figure 2 Example of arranged diatom slides made by naturalists in the late 1800s. These slides were traded worldwide and are still valued by collectors today. Note the radial symmetry in the Centrales diatoms and the bilateral symmetry in the Pennales. Slide from the collection of the Academy of Natural Sciences of Philadelphia. Photographs by Jane Rines. <http://thalassa.gso.uri.edu/flora/index.html>.



Figure 3 When conditions turn unfavorable, diatom cells typically increase in sinking rate and fall from the upper mixed layer. In the case of the genus *Chaetoceros*, this sinking is induced by the production of heavy resting spores (RS). (a) *Chaetoceros constrictus* Gran RS; (b) *Chaetoceros diadema* (Ehrenberg) Gran RS; (c) RS record of a bloom of *Chaetoceros* (www.marbot.gu.se/sss/ssshome.htm; photographs by Mats Kuylenskierna); and (d) a *Skeletonema costatum* record of a monospecific bloom. (c and d) are BSEI photos of Saanich Inlet laminated sediments. Reproduced from Dean JM, Kemp AES, and Pearce RB (2001) Palaeo-flux records from electron microscope studies of late Holocene laminated sediments, Saanich Inlet, British Columbia. In Bornhold BD and Kemp AES (eds.) Late Quaternary Sedimentation in Saanich Inlet, British Columbia, Canada-Ocean Drilling Program Leg 169S, *Marine Geology* 174: 139–158.

assemblage. On the seafloor, dissolution is mainly a function of the degree of Si undersaturation in the sediment pore waters and the rate of detrital accumulation, with small concentrations of Al inhibiting opal dissolution. Incorporation into fecal pellets or aggregates and the formation of heavily silicified RS tend to counteract dissolution. Trap studies have shown that in marine sediments, only 1–5% of the living assemblage in surface plankton is represented in the sediments. This selective dissolution results in the moderate enrichment of the sediments in resistant and robust taxa that are not important constituents of the water column or trap records. However, in coastal upwelling environments, most of the diatom taxa found in the water column and traps also occur in the sediment surface, as the dominant taxa are regularly maintained. Possibly because plankton taxa produced during blooms are preserved with greater efficiency than those produced during non bloom periods, and the dominant genus *Chaetoceros* Ehrenberg produces RS. Besides, relatively little lateral transport of diatoms during settling in the water column has been observed, and similar distribution patterns of the main species in the

plankton and sediment indicate that the diatom sedimentary record of productive areas can effectively be used for paleoecological inferences.

The lack of an effective and reliable method to quantify dissolution in the sedimentary record is still a problem. The use of a fragmentation index combined with the abundance of diatoms helps to identify samples marked by dissolution, but the best solution is the application of molecular marker techniques to identify the former presence of diatoms.

Diatom diversity, calculated on the basis of the Shannon–Wiener method, has also been used as an indicator of surface productivity, with low values indicating a near monospecific assemblage, as found in highly productive environments.

Sea surface temperature (SST) and PP are the two most likely parameters to reconstruct from this phytoplankton group: SST because diatoms record the true surface conditions and PP because diatoms are the most important primary producer in highly productive areas.

A first approach to calculate past SST was a temperature index developed through diatom modern biogeography

(Kanaya and Koizumi, 1966), using the temperature preferences of specific diatoms. The diatom-derived SST's qualitative index is still being used in Pacific Ocean studies and can indicate important trends.

A more quantitative approach to reconstruct past SST was pioneered by Imbrie and Kipp (1971) who generated a Foraminifera-based transfer function using factor analysis, followed by a multistep regression of factor loadings against SST measured *in situ*. In contrast to oceanic level reconstructions based on Foraminifera, better results are likely to be obtained from diatoms at the regional level and from regions/areas with great variability in the environmental parameter being reconstructed.

Diatom transfer functions (TFs) to estimate SST have been generated for the eastern equatorial Pacific, the North Atlantic, Antarctica, and coastal upwelling areas. While in the early years the Imbrie and Kipp approach was followed, more recent studies have included either linear or unimodal models, looking for which model best describes the ecological signal of the diatoms. The exploration of the relationship between diatom species and environmental variables using canonical correspondence analysis (for unimodal approaches) allows the development of TFs without having a predetermined environmental variable and taking into consideration the environmental variables' intercorrelations. This method puts considerable emphasis on the calibration and validation of the TF and allows past PP, salinity, and nutrient reconstructions, besides SST, for annual and several seasonal averages (Lopes et al., 2010).

To develop TF, it is imperative that the fossil assemblage is consistent with the assemblage on which the inference model is based. In order to minimize the effect of dissolution before burial, one attempt was made through the analysis of the fossil assemblages to which a set of surface samples partially dissolved in the laboratory was added.

Valve size

The diameter of *Azpeitia nodulifera* (A. Schmidt) Fryxell and Sims (formerly *Coscinodiscus nodulifer*) ranges from 10 to >100 μm . Diameter measurements in equatorial Pacific cores showed good correlation between maximum diameter and glacial periods. Since wind strength during glacial intervals intensifies oceanic circulation and generates stronger equatorial upwelling, larger forms of *A. nodulifera* may trace this increased upwelling. Another species that shows changes in valve diameter is *Paralia sulcata* (Ehrenberg) Cleve. In coastal environments, valve diameter appears negatively correlated to water column mixing, nutrient input, and PP rates. Temperature has also a positive effect on *P. sulcata* valve diameter, but in terms of salinity, conflicting reports have been published for different environments, reinforcing the idea that variations observed in the water column and/or the sediments should only be used as analogs on a regional/local level. More recently, a study by Crosta (2009) of size and abundance of two species, *Fragilariopsis kerguelensis* (O'Meara) Hustedt and *Fragilariopsis curta* (Van Heurck) Hustedt from East Antarctica Holocene sediments, supports the hypothesis that larger species populations reflect enhanced sexual reproduction.

Isotopic approaches

Isotopic and biomarker techniques have been developed that take into consideration the high dissolution potential of diatoms, as well as the possible winnowing and focusing of sediments, both of which may bias the quantitative and the assemblage data:

- $\delta^{18}\text{O}$ – Oxygen isotopes in marine diatoms should reflect seawater isotopic composition and temperature. Data obtained in eight deep-sea cores from the Antarctic–Atlantic sector revealed core-to-core consistency and similarity to the sequence of events recorded for glacial–interglacial cycles in standard Foraminifera-derived isotopic records, proving the utility of diatoms to generate much-needed $\delta^{18}\text{O}$ stratigraphies for the high latitudes (Shemesh et al., 1995). Generation of new $\delta^{18}\text{O}$ records has been limited by difficulties with the application of the technique, in particular with the obtaining of a clean and pure diatom sample, since even the cleaning methods can add oxygen from other components in the sample (Swann and Leng, 2009).
- $\delta^{13}\text{C}$ – The isotopic composition of carbon from organic matter enclosed within diatom frustules was proposed as a proxy for paleoproductivity and paleo-dissolved carbon concentrations in ocean surface waters.
- $\delta^{15}\text{N}$ – Nitrogen plays a role in the construction of the diatom silica frustules and represents an N archive related to the primary photosynthetic signal. Isotopic measurements of diatom $\delta^{15}\text{N}$ from the Southern Ocean (SO) surface nitrate pool and surface sediments reflect the proportion of nitrate supply from the subsurface that is consumed in the euphotic zone. Several downcore profiles of diatom-bound $^{15}\text{N}/^{14}\text{N}$ from the Antarctic show a consistent trend in the Indian sector but a marked contrast in Atlantic records (Robinson et al., 2004). The differences obtained with different measuring methodologies make it clear that more work needs to be done before the proxy can be broadly used.
- $\delta^{29}\text{Si}$ – Variations in the ratio of ^{30}Si – ^{28}Si in sedimentary opal have been proposed as a reliable proxy for past silicon cycle and diatom productivity. Seasonal variations detected in $\delta^{29}\text{Si}$ in the surface waters linked to upwelling events may also be of great use for paleo-upwelling reconstructions in the future.
- $^{231}\text{Pa}/^{230}\text{Th}$ – The preferential scavenging of opal by ^{231}Pa made the $^{231}\text{Pa}/^{230}\text{Th}$ ratio a possible proxy for opal paleoproductivity, at least in the Indian sector of the SO (Dezileau et al., 2003). Nevertheless, it is necessary to check if the measured opal is actually derived from diatoms.

Molecular Marker Techniques

Chlorine sterol esters (CSE) can be traced back to their biological precursors and are specific for each phytoplankton group. CSE sterols 3 and 4 are both derived by diatoms and have been proposed as a good marker for past diatom productivity (Dahl et al., 2004). Furthermore, long-chain diols and mid-chain hydroxy methyl alkanolates of diatomaceous origin were reported as indicators of high-nutrient conditions.

Proboscia laboratory cultures showed a strong relationship between long-chain diol and 12-hydroxy methyl alkanolate

composition with growth temperature; however, these results did not hold in nature (Rampen et al., 2009).

Finally, IP₂₅ (C₂₅ monounsaturated hydrocarbon), an organic compound believed to be produced mainly by diatoms living in sea ice, may prove to be a good paleoindicator of sea ice (Belt et al., 2007) and is now in the phase of global testing.

Trace metals

Diatoms incorporate trace elements in their opaline frustules. A Lake Geneva study indicates that the incorporation of different trace elements into the diatom frustules has substantially changed in consonance with climate state and the lake's ecological conditions (Jaccard et al., 2009), though the exact processes behind this correlation are yet unknown. In the ocean, several ratios using trace metals have been used as proxies for different parameters traced by the diatoms: in Antarctic sediments, the (Zn/Si)_{opal} ratio appears to be a proxy for the salinity of the mixed layer (Hendry and Rickaby, 2008). The Ge/Si ratio is used to study the Si cycle and climate change, as Ge mimics the behavior of Si in ocean: changes in the ratio mainly indicate changes in continental weathering, but the biological control (via diatoms) also appears to be important (Sutton et al., 2010).

Backscattered electron imagery

Backscattered electron imagery (BSEI) is used to study diatom-rich laminae and mats. The observation of polished thin sections produced from an undisturbed sample slab embedded in Spurr epoxy resin allows the visualization of diatom successions within individual laminae.

Diatoms and Oceanic Processes

Productivity

Glacial–interglacial variability

To verify the hypothesized diatom role in lowering atmospheric $p\text{CO}_2$ during glacial intervals, it is vital to develop a good understanding of past productivity conditions.

Diatom abundance is much used as an indication of past surface productivity, but DAR represents a more reliable measure to assess productivity estimations. In terms of diatom assemblages, coastal upwelling systems have similar assemblages, not only in the water column, but also in the sediment record, at least in respect to the dominant species. The high contribution to the assemblage of the chain-forming *Chaetoceros* reflects bloom conditions. *Chaetoceros* is an opportunist genus known to respond very fast to nutrient pulses and water turbulence. Its capacity to produce RS results in a good sedimentary record of coastal upwelling conditions. But the genus is also abundant in the North Atlantic where their spring blooming reflects nutrient availability and ideal light conditions.

Equatorial upwelling regions have a totally different diatom assemblage from the coastal zones, but there is not much difference between Pacific and Atlantic diatom assemblages. The genus *Thalassionema* Grunow is the dominant taxon found in sediment assemblages, with contributions of the upwelling-related fragile forms of *Pseudo-nitzschia* Peragallo and *Nitzschia* Hassall.

In the high northern latitudes of the open ocean, diatom assemblages contain largely endemic species. In the north Pacific, *Neodenticula* Akiba and Yanagisawa, together with several species of *Rhizosolenia* Ehrenberg, *Thalassiosira* Cleve, and *Coscinodiscus* Ehrenberg, dominates the assemblages. In the North Atlantic, *Neodenticula* is not present, and *Thalassiosira gravida* Cleve and *Thalassionema nitzschioides* (Grunow) Mereschkowsky are the dominant.

In the SO, at least five species are unique to Antarctica: *F. kerguelensis* (O'Meara) Hustedt, *Eucampia balaustium* Castracane, *Thalassiosira lentiginosa* (Janish) Fryxell, *C. margaritae* Freguelli, and *Thalassiosira gracilis* (Karsten) Hustedt.

Time series of DAR and assemblage composition from the Atlantic major coastal upwelling loci of the eastern boundary currents and eastern equator indicate glacial periods as the most productive times in both hemispheres of the eastern Atlantic. An increase in PP was also found in the Angola Basin and attributed to the northern penetration of a stronger Benguela current. A more vigorous atmospheric circulation, intensifying the trade winds and the wind-driven equatorial divergence, could explain this general increase in PP in the glacial Atlantic (Figure 4).

In the SE Pacific, productivity does not vary in consonance with glacial–interglacial cycles. At the Peru margin, variability in paleoproductivity occurs across oxygen isotope boundaries; off the coast of Chile, productivity is higher during glacials.

In the NW Pacific and the NE Pacific upwelling margin, diatom records point to a decrease in productivity during glacials and increase in interglacials.

Studies have also been performed on sediments underlying HNLC regions: the SO and the equatorial Pacific. An excess of Si in the glacial Antarctic, as a result of dust Fe input during glacials, was hypothesized in the 1990s. The northward transport of that Si to low- and mid-latitudes by the sub-Antarctic mode waters would generate a floristic change from coccolithophores to diatoms, resulting in the lowering of atmospheric $p\text{CO}_2$.

Changes in productivity in phase with inputs of Fe from the continent, reported off the coast of Chile, may provide the link between past changes of iron availability and biological productivity.

In the HNLC equatorial Pacific, which has also been shown to be an Si-limited ecosystem, downcore studies of benthic Foraminifera indicate higher productivity in the glacials, while diatoms increase in the interglacials.

An apparent contradiction between productivity proxies is found in the SO, where most of the biogenic opal burial occurs at present under the Antarctic Circumpolar Current (ACC). Assuming the observed HNLC conditions are a consequence of Fe limitation and that the Fe supply results in thinner diatom valves, the supply of Fe to the oceans can increase diatom carbon export to the seafloor without enhancing biogenic Si export. Furthermore, an increase in diatoms in sedimentary sequences from the open south Atlantic, located 6° N of the present Polar front, as well as a suite of 50 cores from the Atlantic and western Indian sector, suggests a northward expansion of the sub-Antarctic front by 5–6° in latitude. However, the contradiction reappears when diatom-bonded $\delta^{13}\text{C}$ and $\delta^{29}\text{Si}$ data point to no increase in productivity, nor in the percentage of Si utilization by diatoms during the last

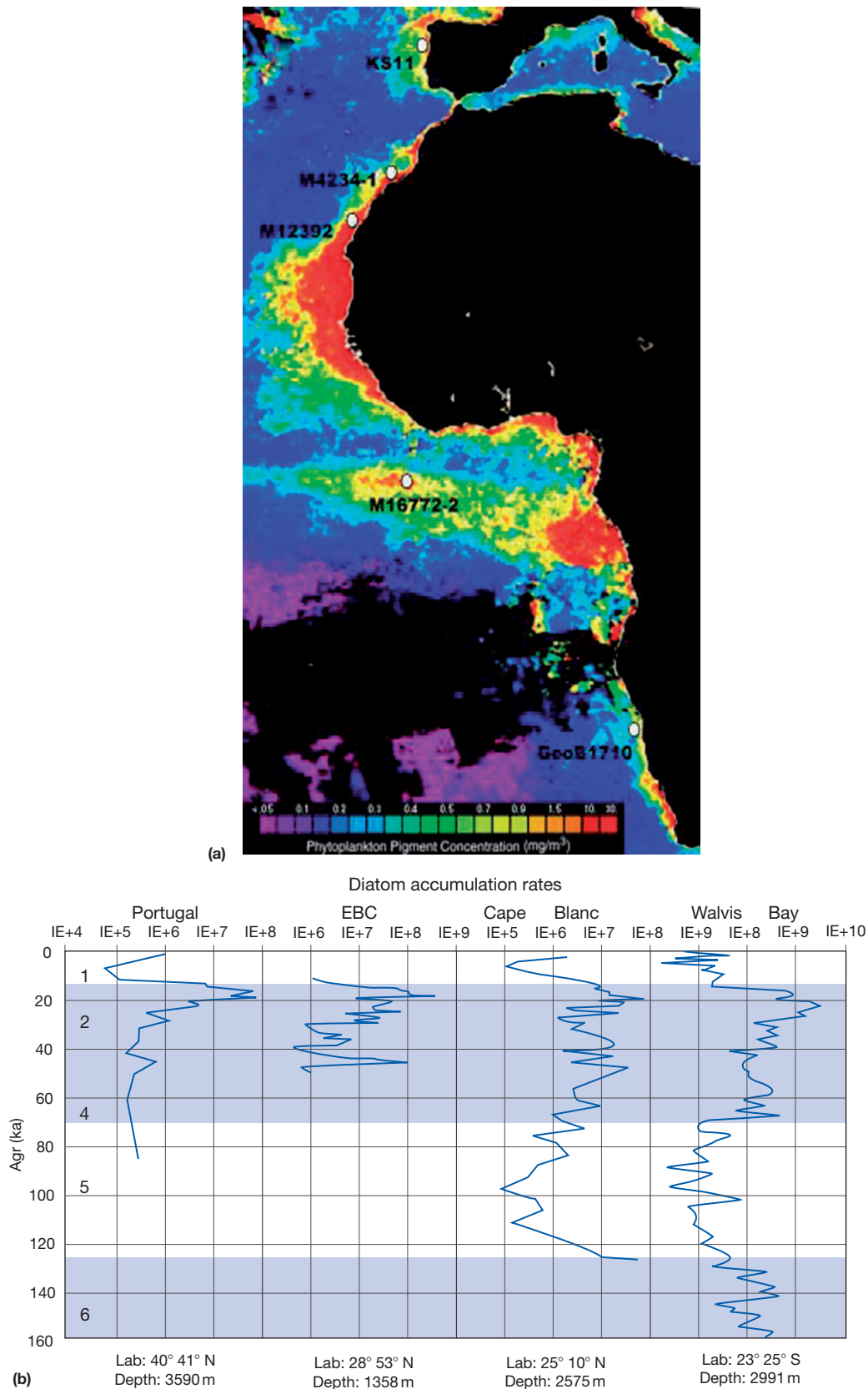


Figure 4 (a) Cores' approximate location on a 3-month composite coastal zone color scanner estimates of phytoplankton pigment concentration (June–July 1979 to NSF/NASA 1989). (b) DARs (number of valves $\text{cm}^{-2} \text{ka}^{-1}$) versus age (^{14}C ka) along the cores retrieved from the major coastal upwelling areas of the eastern Atlantic (KS11 – Portugal; GeoB M1410 – Canary Islands; M13229 – Cape Blanc; GeoB M1710 – Walvis Bay.). Reproduced from Abrantes F (2000) 200 000 yr diatom records from Atlantic upwelling sites reveal maximum productivity during LGM and a shift in phytoplankton community structure at 185 000 yr. *Earth Planetary Science Letters* 176: 7–16.

glaciation, relatively to the Holocene. According to Dugdale et al. (2004), the teleconnection for nutrients between the equatorial Pacific and Antarctica suggests an oscillatory relationship that may influence or control the Si glacial cycles. This hypothesis connects to the silica leakage hypothesis (SALH) that Matsumoto et al. (2002) have invoked to explain the drawdown of atmospheric CO₂ during glacial intervals. However, this has been challenged by Bradt Miller et al. (2009) who found convincing evidence for SALH in the equatorial Atlantic but not in the Pacific. However, large increases in Si in Pacific sediments in some terminations and other periods suggest that the SO might not be the only source for silica.

Millennial variability

Several workers have shown the dramatic climatic variability, known as Dansgaard–Oeschger (D–O) cycles, in Greenland ice cores and in many marine sequences. The largest D–O oscillations, known as Heinrich events (HE), are believed to represent major iceberg discharges in the North Atlantic and are associated with a southern invasion of polar waters, disruption of the North Atlantic current circulation, and decrease in deep-water ventilation. The relation between oceanic productivity and these major climatic variations has been assessed by diatom quantification for HE1 and HE4 through the open North Atlantic, from north to south of the iceberg discharge belt (the Ruddiman belt). During HE4, diatom productivity decreased drastically at all latitudes, but throughout HE1, productivity increased in the Sargasso Sea (Gil et al., 2009) and apparently was not affected in northern latitudes, possibly due to the associated increase in summer insolation not observed during H4 (Nave et al., 2007).

Productivity variations have also been assessed for the anoxic Cariaco Basin laminated sediments, from the analysis of CSE for the Bølling/Allerød–Younger Dryas (YD) interval. Diatom populations increased during the YD in the Cariaco Basin, reflecting a higher productivity attributed to a regional intensification in trade winds and upwelling.

Centennial to decadal variability

Centennial-scale variability in productivity, temperature, and trajectory of water masses, associated with the most recent Holocene climatic shifts: the Little Ice Age and the Medieval Warm Period, are also revealed by diatom abundance and assemblage composition in high-sedimentation rate cores from the NW and NE Atlantic margins. At present, the North Atlantic oscillation (NAO) is believed to dictate climate variability across most of the Northern Hemisphere, especially during wintertime. Those diatom records are, at all sites, coherent with the action of a NAO-like mechanism or the NAO itself.

One promising proxy for the reconstruction of El Niño events through time is *A. nodulifera*, an open ocean tropical form that reaches the Peruvian and California coasts and has a major flux to the bottom during such events. Besides, its spatial distribution in the Peruvian coastal sediments marks the area where unusually warm waters are registered in El Niño years. In a 15-ka record from the Guayama Basin, this species abundance reveals that modern oceanographic conditions with intensified ENSO cycles was established between 2.8 and 2.4 ka

and 500–200-year cycles of more frequent and/or stronger El Niño-like events (Barron et al., 2004).

Monospecific mats

In addition to the spring bloom and upwelling productivity, large diatoms, typical of oligotrophic areas, are capable of generating substantial production that may contribute significantly to the biological pump.

A long line of diatom-rich waters was spotted in the equatorial Pacific by the space shuttle crew, and subsequent *in situ* sampling revealed a rhizosolenid mat along a thermal front that corresponded to a tropical wave instability. These mats can remain in the surface waters until changes in wind circulation or loss of buoyancy due to physiological stress causes them to sink to the ocean bottom where they form a thick and unbreakable net that prevents sediment bioturbation.

Quaternary mat deposits have been recognized beneath major frontal regions in the North Atlantic and SO (Figure 5). Other examples come from the Benguela current upwelling system, where a mat deposited from 2.6 to 2 Ma may reflect

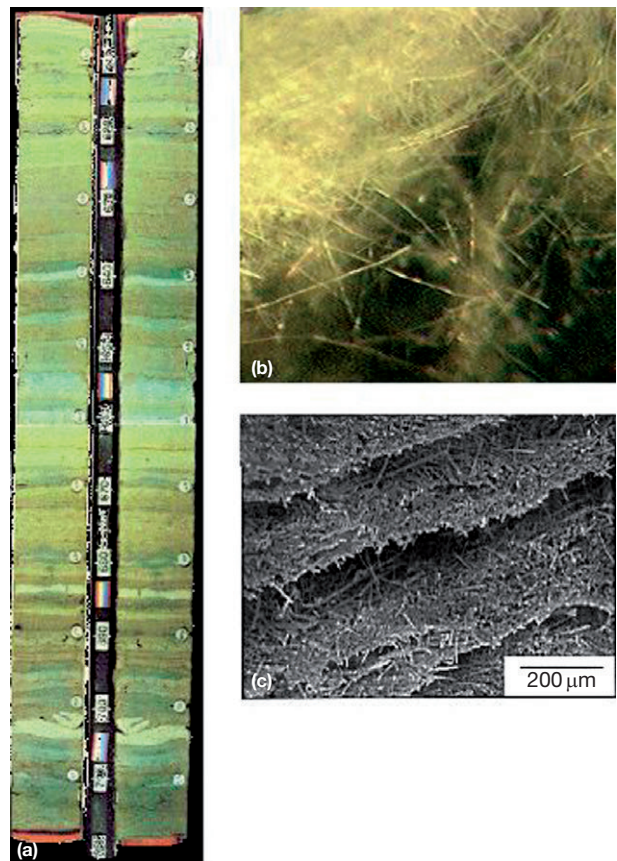


Figure 5 A 3-m diatom mat composed of billions of layers of the hairlike diatom *Thalassiothrix longissima* found in the North Atlantic. (a) Section of the split core showing the layers (or laminae); (b) fibrous diatoms; and (c) scanning electron microscope photomicrograph showing the sheet of paper like individual mats of diatoms http://www.log.furg.br/WEBens/ideo/CORE_REPOSITORY/RHP4Boden.html. Reproduced from Bodén P and Backman J (1996) A laminated sediment sequence from the northern north Atlantic Ocean and its climatic record. *Geology*, 24, 507–510.

an increased influence of SO water on the SE Atlantic. In the inner continental shelf basins of the Western Antarctica Peninsula, a *Chaetoceros*-dominated ooze was found in laminated sediments deposited during the last deglaciation, representing increased PP. Rhizosolenid-laminated sediments were also described in the eastern Mediterranean and related to density variations in the water column resulting from variations in the Nile river input. Analogous oozes of the giant diatom *Ethmodiscus rex* Castracane were described in the equatorial Atlantic and at 9° S in the Indian Ocean.

General circulation

Many water column diatom assemblages show close affinity to water masses and help to trace past positions and interactions between water masses and current systems. Downcore assemblage studies defined on a regional level were successfully used as paleoceanographic indicators in many cases: in the Norwegian Sea, revealing variations in the influx of temperate Atlantic surface waters during the Holocene, and in the north Icelandic shelf, revealing interactions between the cold East Greenland Current and the warm Irminger Current during the latest Holocene. In the Gulf of St. Lawrence, deglaciation and Holocene diatom data support the suggestion that the YD cooling event could not have been caused by meltwater input from the Laurentide ice sheet. In the ACC (at a site that lies immediately downstream from where North Atlantic Deep Water (NADW) is entrained into the ACC), deglacial submillennial-scale variations in the numbers of *Chaetoceros* RS are related to shifts in the path of the ACC and in the strength of NADW overflow in the Falkland Plateau and document important interhemispheric links between the North Atlantic and the SO (Allen et al., 2005).

Sea ice dynamics

The presence in the sediments of *Eucampia antarctica* (Castracane) Mangin, a species that lives at the ice margin in Antarctica, has been used as a tracer for Antarctic sea ice. Another group of diatoms live in brine pockets at the base of sea ice in both the Arctic and the Antarctic (Figure 6). These diatoms bloom twice a year, within the ice in spring and in the ocean in summer, which leads to different species in each bloom and generates a transition in species composition between the ice margin and the open ocean. Tracing these two assemblages in the surface sediments allowed the



Figure 6 View of a pack of sea ice at Terra Nova Bay in January of 1990, where a brownish-colored band due to the presence of high numbers of diatoms is clearly visible. Photograph by Caterina Nuccio.

estimation of a more reduced summer sea ice field during the Last Glacial Maximum.

Furthermore, the presence of Quaternary diatoms in glacial sediment samples recovered from beneath the West Antarctic ice sheet (grounded ice) and their correlation with beryllium-10 dates imply one or more partial (Ross Sector) or complete marine ice sheet collapse events in interglacials of the past 750 ka.

Lateral transport/displacement

Displaced diatoms can provide important paleoecological information. The presence of benthic marine diatoms in deep sediments (>200 m) indicate lateral displacement of shallower sediments. Lateral advection can be traced by the planktonic neritic taxa normally present in highly productive coastal regions but found offshore, when live specimens are laterally advected by upwelling filaments. The distribution of Antarctic endemic diatoms in the sediments of the Pacific and the southwest Indian Ocean reveals the northward path of Antarctic bottom water, as far as the equatorial regions.

Sea-level variability and other coastal processes

Marine littoral diatoms can be used to reconstruct sea-level changes in the recent past.

Many of these species, which are similar throughout the world, show tidal zonation; as such, changes in diatom community indicate past sea level. Diatom TF have been developed and successfully applied to reconstruct sea-level changes in arctic (Woodroffe and Long, 2010) and tropical environments. These data help address a variety of problems in coastal geology (paleogeography, natural hazard assessment).

Diatoms as Tracers of Continental Climate

Freshwater diatoms can also be found in deep-sea cores, and their presence in atmospheric dust off NW Africa allows their use as tracers of the path and intensity of past storms. Off Africa, the Sahel and sub-Saharan intermittent lakes are the dust source. In the equatorial Atlantic, freshwater diatom records have made it possible to determine the time of the first influx of wind-blown continental detritus (1.2 Ma). Over the last 900 ka, peaks in the abundance of the freshwater diatom *Aulacoseira granulata* were found in equatorial Atlantic cores between 3° N and 1° S and 21° to 25° further N, with a periodicity that corresponded to the Earth's precession cycle and related to African monsoon cycles.

In Greenland ice cores, layers of freshwater diatoms parallel those of smaller dust particles and are related to seasonal winds originating from dry areas of North America. In central Antarctic ice cores, freshwater and marine diatoms occur in smaller abundances than in Greenland, but their downcore pattern follows the dust data with higher concentrations during glacial periods (Kellogg and Kellogg, 1996). Their origin was traced to Patagonia.

Regarding a more quantitative approach to the freshwater diatom signal in marine sediments, Lopes et al. (2010) developed a linear equation calibrated with modern conditions that allowed the identification and dating of undisturbed marine material, linked with the Glacial Missoula flood signal.

Biostratigraphy

Biostratigraphy is only possible because species evolve and become extinct at specific times, and the evolution of one species is not repeatable. Diatoms are particularly advantageous for biostratigraphic studies of high-latitude sediments where calcareous microfossils are often poorly preserved or of low diversity.

With the advent of the Deep Sea Drilling Project, many kilometers of sediment cores have been recovered, and it has been possible to determine that the evolutionary history of diatoms has been punctuated by several floristic turnovers. But, recognition of diatoms as biostratigraphic tools by the geologic community came only when Burckle (1972) correlated late Cenozoic diatoms from the equatorial Pacific to the same site magnetic polarity zonations.

The bimodal size–frequency distribution of *A. nodulifera* in the equatorial Pacific allows size changes to be applied biostratigraphically in late Quaternary fossil assemblages from this region.

The first appearance of *Fragilariopsis doliolus* Husted, a species mainly from equatorial regions but present up to 40° from the equator, marks the beginning of the Quaternary in marine records. By the early Quaternary, the modern diatom flora was developed in most marine environments.

See also: **Paleoceanography:** Paleoceanography An Overview. **Diatom Records:** Antarctic Waters; Diatom Fossil Records from Marine Laminated Sediments; North Atlantic and Arctic; Pacific. **Paleoclimate:** The Younger Dryas Climate Event.

References

- Allen CS, Pike J, Pudsey CJ, and Leventer A (2005) Submillennial variations in ocean conditions during deglaciation based on diatom assemblages from the southwest Atlantic. *Paleoceanography* 20: PA2012. <http://dx.doi.org/10.1029/2004PA001055>.
- Armbrust E, Berges J, Bowler C, et al. (2004) The genome of the diatom *Thalassiosira pseudonana*: Ecology, evolution, and metabolism. *Science* 306: 79–86.
- Barron J, Bukry D, and Bischoff L (2004) High resolution paleoceanography of the Guaymas basin, Gulf of California, during the Past 15 000 years. *Marine Micropaleontology* 50: 185–207.
- Belt ST, Massé G, Rowland SJ, Poulin M, Michel C, and LeBlanc B (2007) A novel chemical fossil of palaeo sea ice: IP₂₅. *Organic Geochemistry* 38: 16–27.
- Bradt Miller LI, Anderson RF, Fleisher MQ, and Burckle LH (2009) Comparing glacial and Holocene opal fluxes in the Pacific sector of the Southern Ocean. *Paleoceanography* 24: PA2214. <http://dx.doi.org/10.1029/2008PA001693>.
- Burckle L (1972) Late Cenozoic planktonic diatom zones from the Eastern Equatorial Pacific. *Nova Hedwigia* 39: 217–246.
- Crosta X (2009) Holocene size variations in two diatoms species, East Antarctica: Productivity vs environmental conditions. *Deep-Sea Research I* 56: 1983–1993.
- Dahl KA, Repeta D, and Goericke R (2004) Reconstruction of phytoplankton community of the Cariaco basin during the Younger Dryas cold event using chlorinesteryl esters. *Paleoceanography* 19: PA1006. <http://dx.doi.org/10.1029/2003PA000907>.
- Dezileau L, Reyss JL, and Lemoine F (2003) Late Quaternary changes in biogenic opal fluxes in the Southern Indian Ocean. *Marine Geology* 3398: 1–16.
- Dugdale RC, Lyle M, Wilkerson FP, Chai F, Barber R, and Peng T-H (2004) Influence of equatorial diatom processes on Si deposition and atmospheric CO₂ cycles at glacial/interglacial timescales. *Paleoceanography* 19: PA3011. <http://dx.doi.org/10.1029/2003PA000929>.
- Falkowski P, Scholes RJ, Boyle E, et al. (2000) The global carbon cycle: A test of our knowledge of Earth as a system. *Science* 290: 291–296.
- Gil IM, Keigwin LD, and Abrantes FG (2009) Deglacial diatom productivity and surface ocean properties over the Bermuda Rise, northeast Sargasso Sea. *Paleoceanography* 24: PA4101. <http://dx.doi.org/10.1029/2008PA001729>.
- Hendry KR and Rickaby REM (2008) Opal (Zn/Si) ratios as a nearshore geochemical proxy in coastal Antarctica. *Paleoceanography* 23: PA2218. <http://dx.doi.org/10.1029/2007PA001576>.
- Hustedt F (1962) *Die Kieselalgen. Dr. L. Rabenhorsts Kryptogamen-Flora*, pp. 920. Weinheim: Verlag Von J. Cramer.
- Hutchings DA, Hare CE, Weaver RS, et al. (2002) Phytoplankton iron limitation in the Humboldt Current and Peru Upwelling. *Limnology and Oceanography* 47(4): 997–1011.
- Imbrie J and Kipp N (1971) A new micropaleontological method for quantitative paleoclimatology. In: Turekian K (ed.) *Late Cenozoic Glacial Ages*, pp. 71–182. New Haven, CT: Yale University Press.
- Jaccard T, Ariztegui D, and Wilkinson KJ (2009) Incorporation of zinc into the frustule of the freshwater diatom *Stephanodiscus hantzschii*. *Chemical Geology* 265(3–4): 381–386.
- Kanaya T and Koizumi I (1966) Interpretation of diatom thanatocoenoses from the North Pacific applied to a study of core V20–130 (studies of a deep-sea core V20–130, part IV). *The Science Reports of the Tohoku University. Second Series, Geology* 37(2): 89–130.
- Kellogg D and Kellogg T (1996) Diatoms in South Pole ice: Implications for eolian contamination of Sirius group deposits. *Geology* 24: 115–118.
- Lopes C, Mix AC, and Abrantes F (2010) Environmental controls of diatom species in Northeast Pacific sediments. *Palaeogeography, Palaeoclimatology, Palaeoecology* 297: 188–200.
- Matsumoto K, Sarmiento JL, and Brzezinski MA (2002) Silicic acid leakage from the Southern Ocean: A possible explanation for glacial atmospheric pCO₂. *Global Biogeochemical Cycles* 16(3): 1031. <http://dx.doi.org/10.1029/2001GB001442>.
- Nave S, Labeyrie L, Gherardi J, et al. (2007) Primary productivity response to Heinrich events in the North Atlantic Ocean and Norwegian Sea. *Paleoceanography* 22: PA3216. <http://dx.doi.org/10.1029/2006PA001335>.
- Peterson T and Whitney F (2003) The role of Fe:Si supply ratio in regulating export production in HNLC waters of the Northeast Subarctic Pacific Ocean. In: Kemp A and Dugdale R (eds.) *AGU Chapman Conference on the Role of Diatom Production and Si Flux and Burial in the Regulation of Global Cycles*, p. 31. Paros, Greece, 22–26 September, AGU.
- Rampen SW, Schouten S, Schefuss E, and Sinninghe Damsté JS (2009) Impact of temperature on long chain diol and mid-chain hydroxyl methyl alkanolate composition in *Proboscia* diatoms: Results from culture and field studies. *Organic Geochemistry* 40: 1124–1131.
- Raymont J (1980) *Plankton and Productivity in the Oceans 1*. Oxford: Pergamon Press.
- Robinson R, Brunelle B, and Sigman DM (2004) Revisiting nutrient utilization in the glacial Antarctic: Evidence from a new method for diatom-bound N isotopic analysis. *Paleoceanography* 19: PA3001. <http://dx.doi.org/10.1029/2003PA000996>.
- Shemesh A, Burckle L, and Hays J (1995) Late Pleistocene oxygen isotope records of biogenic silica from the Atlantic sector of the Southern Ocean. *Paleoceanography* 10: 176–196.
- Sutton J, Ellwood MJ, Maher WA, and Croot P (2010) Oceanic distribution of inorganic germanium relative to silicon: Germanium discrimination by diatoms. *Global Biogeochemical Cycles* 24: GB2017. <http://dx.doi.org/10.1029/2009GB003689>.
- Swann GE and Leng MJ (2009) A review of diatom $\delta^{18}\text{O}$ in palaeoceanography. *Quaternary Science Reviews* 28(5–6): 384–398.
- Woodroffe SA and Long AJ (2010) Reconstructing recent relative sea-level changes in West Greenland: Local diatom-based transfer functions are superior to regional models. *Quaternary International* 221(1–2): 91–103.

Planktic Foraminifera

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Description

Planktic Foraminifera are single-celled eukaryotic organisms that live in the photic zone of the marine environment and exhibit passive floating lifestyles (Figure 1(a)). They range in size from 100 to 1 mm in length and gain nutrition by capturing food particles and digesting them within their protoplasm. Many species also carry symbiotic algae within their protoplasm that photosynthesize during the day. Planktic Foraminifera (often referred to as planktic forams) secrete shells (tests) composed of calcium carbonate (CaCO_3) that are usually characterized by a number of chambers added during growth. Individuals are thought to reproduce about every 28 days with the lunar cycle and may live up to a few months, after which their tests sink to the ocean floor. Approximately 40 species of planktic Foraminifera live in the ocean today. This comprises about 1% of the extant species of Foraminifera (99% are benthic).

Historical Background

The geologic record of Foraminifera begins in the earliest Cambrian. Foraminifers probably existed as cells without tests long before that. At ~200 Ma, during the Jurassic, some foraminifers adapted to a planktic mode of life. The rapid evolution of planktic Foraminifera during the Cretaceous and throughout the Cenozoic makes them ideal biostratigraphic markers, and their sensitivity to physical and biological environmental changes makes them useful indicators of paleoceanographic and paleoclimatic history.

Prior to the advent of the microscope, seventeenth century naturalists observed foraminifers with hand lenses and often classified them as gastropods or cephalopods because many possess a coiled-chamber arrangement. As microscopy advanced, more detail in the structure of tests was observed, and Foraminifera were recognized as single-celled organisms. The first detailed descriptions of planktic Foraminifera were published by d'Orbigny (1826). Foraminiferal studies advanced with the sailing of the *Challenger* expedition of 1872–76. Brady (1884) illustrated a number of planktic foraminifers from this first oceanographic cruise, and our understanding of planktic foraminiferal biogeography began to take shape. The next major advancement was the development of scanning electron microscopy, which enabled detailed analysis of test wall ultrastructure for the first time. The Deep Sea Drilling Project (DSDP) and its successors, the Ocean Drilling Program (ODP), and the Integrated Ocean Drilling Program (IODP), have provided a wealth of quantitative data on both the temporal, geographic, and paleoecologic distribution of planktic foraminifers.

Classification

The taxonomic classification of Foraminifera has proved problematic, with little consensus across the scientific community. For simplicity, the authors ignore the higher classification categories of the Linnaean system and adopt the well-known foraminifer classification scheme of Loeblich and Tappan (1988) which places planktic foraminifers into order Foraminiferida and suborder Globigerinida.

Traditionally, planktic foraminifers are classified primarily by the morphology and ultrastructure of their secreted calcium carbonate (CaCO_3) tests. The number, shape, and arrangement of chambers, the position and shape of apertures, and test ornamentation are important criteria (Figure 1(b)), reflecting a long history of study and utilization by micropaleontologists using optical microscopes. Because classification based solely on test morphology ignores phylogenetic relationships among species, most modern classification schemes incorporate the present understanding of ancestor–descendant relationships and the conservative morphologic features of lineages such as surface ultrastructure. Recent studies of foraminiferal rDNA sequencing have led to advances in molecular phylogeny, at times upsetting existing classification schemes.

Geographic Distribution of Living Planktic Foraminifera

The geographic distribution of living planktic foraminifers at any instant in time depends upon a number of controlling conditions. Temperature and salinity are considered first-order controls, but nutrient concentrations at various depths in the water column, water density, CO_2 , O_2 , distribution of symbiotic organisms, predation, and food supply are also significant factors.

Planktic foraminifers live in the upper 200 m of the ocean at all temperatures, from the freezing point of seawater (approximately -1.8°C) to the warmest sea surface conditions near 31°C (Be, 1971). Extensive monitoring through plankton tows and sediment trap studies shows that different taxa live at different depths and that the vertical distribution of planktic foraminifers in the water column changes daily and seasonally. In general, species living very close to the sea surface (0–50 m) include *Globigerinoides ruber*, *Globigerinoides sacculifer*, *Globigerinoides conglobatus*, *Turborotalita quinqueloba*, and *Globigerina rubescens*. Deeper-dwelling planktic forams (>100 m) include *Globorotalia crassaformis*, *Globorotalia inflata*, *Globorotalia tumida*, *Globorotalia menardii*, *Globorotalia truncatulinoides*, *Neoglobobulimina pachyderma*, and *Sphaeroidinella dehiscens*.

Present-day planktic foraminifers are arrayed in generally zonal assemblages (polar, subpolar, temperate, subtropical, and tropical) that are globally distributed and bipolar in nature

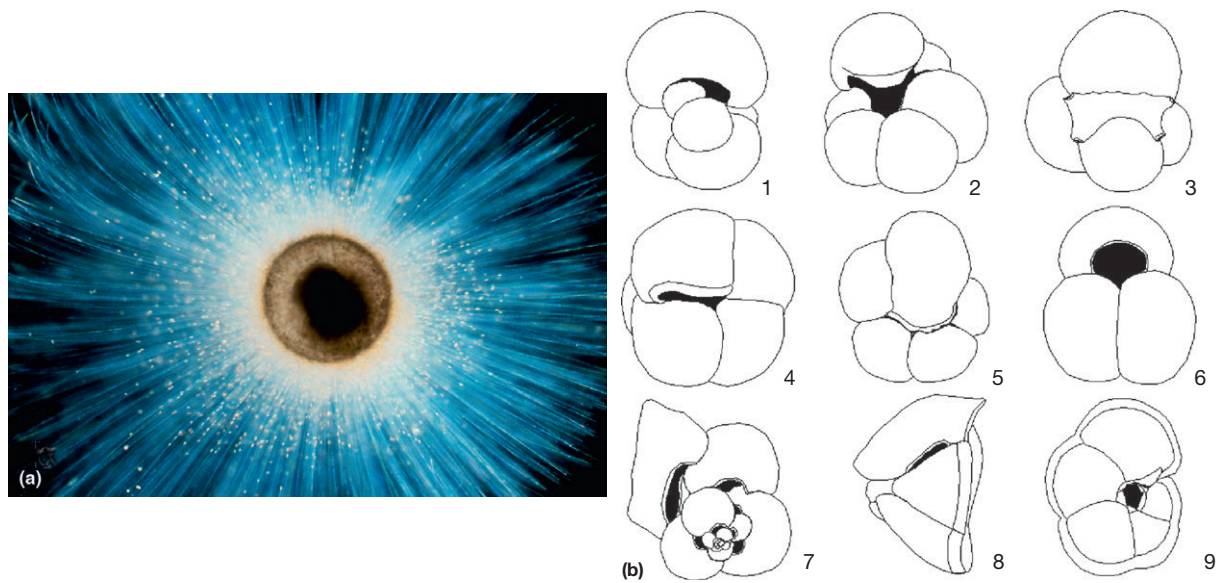


Figure 1 (a) Light micrograph of the living Foraminifera, *Orbulina universa*, with its final spherical chamber surrounding an internal multichambered shell (dark cytoplasm filled region within the sphere). Numerous symbiotic dinoflagellates form a spherical region of photosynthetic activity around the shell (golden color). The calcite spines typically extend 2.0–2.5 mm beyond the surface of the shell that is 0.5 mm in diameter. Courtesy of Dr. Howard Spero, Department of Geology, University of California at Davis. (b) Line sketches of common species showing a variety of chamber arrangements. (1) *Globigerinella aequilateralis*; (2) *Neogloboquadrina dutertrei*; (3) *Globigerinita glutinata*; (4) *Neogloboquadrina pachyderma* (dextral); (5) *Turborotalita quinqueloba*; (6) *Globigerinoides ruber*; (7) *Globigerinoides sacculifer*; (8) *Globorotalia truncatulinoides*; and (9) *Globorotalia menardii*.

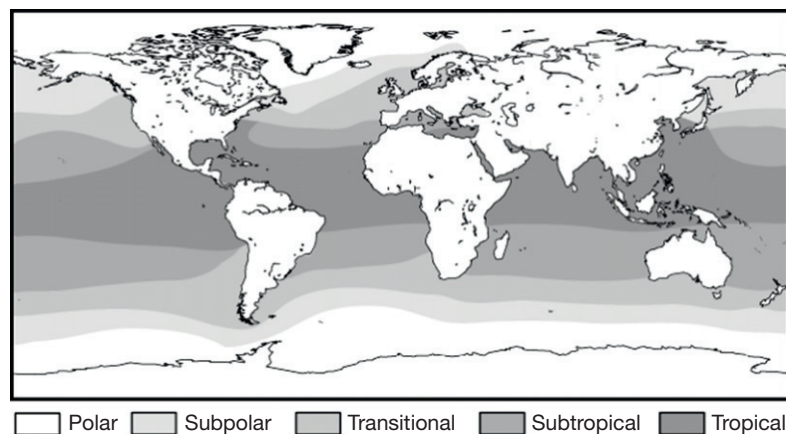


Figure 2 Geographic distribution of living planktic Foraminifera grouped by biogeographic assemblages.

(Figure 2). The polar regions are characterized by nearly monospecific assemblages composed of *N. pachyderma*. These polar assemblages sometimes contain minor concentrations of *T. quinqueloba*, *Globigerina bulloides*, and *Globigerinita glutinata*. Coiling direction of some taxa is significant, and the coldest regions are occupied almost exclusively by sinistrally (left) coiling forms of *N. pachyderma*.

Subpolar waters are characterized by *Globigerina bulloides*, dextrally coiling *N. pachyderma*, *T. quinqueloba*, *Globorotalia inflata*, *Globorotalia truncatulinoides*, *Globorotalia scitula*, and *Globigerinita glutinata*.

Species diversity in general increases toward the tropics where a typical open ocean assemblage comprises 25–30 different species of planktic Foraminifera including

Globigerinoides sacculifer, *Globigerinoides ruber*, *Globigerinoides conglobatus*, *Globorotalia menardii*, *Globorotalia tumida*, *Globorotalia crassaformis*, *Globorotalia truncatulinoides*, *Globigerina rubescens*, *Globigerina bulloides*, *N. dutertrei*, *Globigerinella aequilateralis*, *Globigerinita glutinata*, *Orbulina universa*, *Candeina nitida*, *S. dehiscens*, and *Pulleniatina obliquiloculata*. These low-latitude assemblages are dominated by *Globigerinoides ruber* and *Globigerinoides sacculifer*. Both taxa have maximum abundance in the tropics, but *G. sacculifer* is euryhaline and an excellent indicator of high-salinity conditions.

The subtropical assemblage contains *Globigerinoides ruber*, *Globigerinoides sacculifer*, *Globigerinoides conglobatus*, *Globorotalia menardii*, *Globorotalia inflata*, *Globorotalia truncatulinoides*, *Globorotalia hirsuta*, *Globigerina falconensis*, *Globigerina bulloides*,

N. dutertrei, *Globigerinella aequilateralis*, *Globigerinita glutinata*, *Globorotalia crassaformis*, and *O. universa*.

The temperate assemblage exists at the gradient between the subtropical warm waters and subpolar cool waters. It is characterized by a mixed assemblage containing both cool and warm water taxa.

Seafloor Distribution of Planktic Foraminifera Tests

The seafloor thanatocoenosis (death assemblage) is the result of complex interactions between foram tests and the biological, chemical, and physical environment. Despite minor differences between plankton tow or sediment trap observations and core top assemblages, there exists a robust and overwhelming surface temperature signal in the quantitative distribution of planktic foraminifera on the seafloor. This fact has been utilized by paleoceanographers to help unlock the Quaternary history of the world's oceans. For example, during glacial periods, the range of polar and subpolar assemblages expanded, relative to today, and shifted in their geographic distribution toward the equator (CLIMAP, 1981; MARGO, 2009).

Stratigraphic Distribution and Biostratigraphy

The combined characteristics of high abundance, small size, facies independence, ease of identification, and high preservation potential make planktic foraminifera ideal for biochronological analyses. For any pre-Quaternary study, foraminiferal biochronology remains one of the leading age-dating tools in terms of accuracy and utility. Planktic foraminifera zonation have been developed by a number of workers, and individual

events (biotic first appearance datums (FAD) and last appearance datums (LAD)) have been dated by calibration to magnetostratigraphy (e.g., Berggren et al., 1985, 1995a; Dowsett, 1989; Wade et al., 2010) and oxygen isotope stratigraphy (Berggren et al., 1995b). Within the Quaternary, there are relatively few evolutionary events that can be used for foram biostratigraphy.

The most widely used tropical to subtropical planktic foraminifera zonation (Berggren et al., 1995a modified by Wade et al., 2011) characterizes the Quaternary by one low-latitude zone, Pt1 or the *Globorotalia truncatulinoides* partial range zone. Pt1 is defined as the interval containing *Globorotalia truncatulinoides* between the highest occurrence of *Globigerinoides fistulosus* and the recent. The last appearance of *Globigerinoides fistulosus* is approximately synchronous (~1.88 Ma) in both Atlantic and Pacific tropical sites. The first appearance of *Globorotalia truncatulinoides* is generally synchronous outside the South-West Pacific region at ~1.8 Ma (Figure 3).

Pt1 can be subdivided into a lower Pt1a or *Globorotalia tosaensis* highest occurrence subzone (Wade et al., 2011). Pt1a is defined as the interval between the highest occurrence of *Globigerinoides fistulosus* and the highest occurrence of *Globorotalia tosaensis*. This zone has an estimated age of ~1.77–0.65 Ma. Pt1b or the *Globorotalia truncatulinoides* Partial range subzone is defined as the biostratigraphic interval characterized by the partial range of the nominate taxon following the highest occurrence of *Globorotalia tosaensis* and extending to the recent (Wade et al., 2011).

Stratigraphic information (summarized from Berggren et al. (1995b) and Wade et al. (2011)) shows the following planktic foraminiferal events to be useful for developing Quaternary biochronology: LAD *Globorotalia flexuosa* (0.068 Ma; $\delta^{18}\text{O}$ stage 4), LAD *Globigerinoides ruber* (pink) (0.12 Ma;

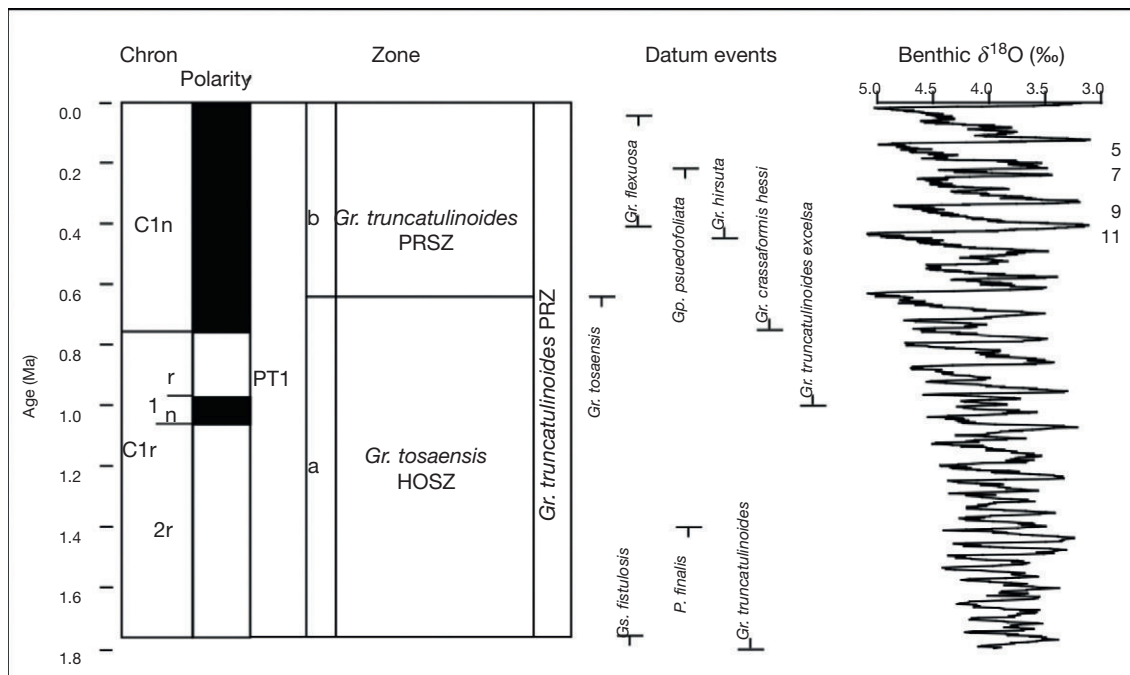


Figure 3 Quaternary (sub)tropical planktic foraminifer biostratigraphy and magnetobiochronology. After Berggren et al. (1995a,b) and Wade et al. (2010) and Benthic $\delta^{18}\text{O}$ from Lisiecki and Raymo (2005).

Indo-Pacific only), LAD *Globoquadrina pseudofoliata* (0.22 Ma; $\delta^{18}\text{O}$ stage 7), FAD *Globorotalia flexuosa* (0.401 Ma; $\delta^{18}\text{O}$ stage 11), FAD *Globorotalia hirsuta* (0.45 Ma; $\delta^{18}\text{O}$ stage 12), LAD *Globorotalia tosaensis* (0.65 Ma), FAD *Globorotalia crassaformis hessi* (0.75 Ma), FAD *Globorotalia truncatulinoides excelsa* (1.00 Ma), and LAD *Pulleniatina finalis* (1.40 Ma). A good overview of other planktic foraminifer zonal schemes and stratigraphic distribution of various taxa within the Quaternary can be found in Bolli et al. (1985).

One last biochronological application of planktic Foraminifera is based upon coiling direction changes. Some taxa (e.g., *Globorotalia truncatulinoides* and *N. pachyderma*), as indicated in the section on living Foraminifera, show a relationship between sea surface temperature (SST) and the ratio of sinistrally (left) to dextrally (right) coiled tests. In high-latitude cold regions, *N. pachyderma* exhibits sinistral coiling almost exclusively. In subpolar regions, more dextral-coiling specimens are observed. In local to regional settings, stratigraphic correlations can be based upon regular shifts in coiling patterns.

Applications to Paleoclimatology and Oceanic Paleoclimatology

With the recognition that the distribution of planktic Foraminifera on the seafloor closely matches overlying oceanic conditions, utilization of the ecological tolerances and preferences of extant taxa (taxonomic uniformitarianism) have been used both qualitatively and quantitatively to reconstruct past conditions.

The occurrence of warm-water taxa like *Globorotalia menardii* at latitudes far North or South of their present-day distribution is commonly used as evidence for warmer conditions. Likewise, the equatorward retreat of other taxa can signal cooling, relative to today. Quantitative analysis of assemblages of planktic foraminifera has been widely used to reconstruct past sea surface temperature using transfer function or nearest modern analog approaches.

Transfer functions range from simple (Ericson and Wollin, 1956) to the complex factor analytic approaches pioneered by Imbrie and Kipp (1971). This latter approach formed the basis of the CLIMAP Last Glacial Maximum reconstruction (CLIMAP, 1981) and has been modified and used to reconstruct paleoclimate conditions during other Quaternary intervals and even back to the mid-Pliocene (Dowsett et al., 2005, 2010).

Modern analog techniques were first introduced by Hutson (1979). Today there are many variations on the basic technique, but all are based upon statistical matching of a fossil sample (census) to a modern assemblage and assigning the physical characteristics (e.g., temperature) of the closest modern analog to the fossil sample.

Calcium carbonate tests are thought to be secreted in equilibrium with seawater at the time of formation. For this reason, planktic foraminifera have been the most widely used organisms for shell chemistry (e.g., Mg:Ca) and stable isotope geochemistry ($\delta^{18}\text{O}$, $\delta^{13}\text{C}$, etc.). In addition, planktic foraminifera are generally found in sufficient abundance for AMS ^{14}C dating. Their small size, abundance, and distribution make them ideal paleoenvironmental tools. With the knowledge that different species of planktic foraminifer live at different depths

and reach maximum abundance at different times of the year, geochemical analyses of planktic foraminifera tests can provide a wealth of information on the composition and structure of the water column in the past. With benthic foraminifera providing information on bottom waters, elaborate paleoenvironmental and paleoclimatic information can be mined from samples containing foraminifera.

In summary, planktic foraminifera are the original multi-tool for Quaternary marine studies. The quantitative distribution of foraminifera in both temporal and spatial dimensions provides information on the age of the sample and many aspects of the paleoclimatic environment. When combined with geochemical and isotopic analyses of planktic foraminifera tests, this fossil group surpasses all others in its contribution to the field of Quaternary paleoclimatology.

See also: [Paleoclimatology: Paleoclimatology An Overview.](#)
[Paleoclimatology, Biological Proxies: Benthic Foraminifera.](#)
[Paleoclimatology Reconstruction: Pliocene Environments.](#)
[Quaternary Stratigraphy: Continental Biostratigraphy.](#)
[Radiocarbon Dating: AMS Radiocarbon Dating.](#)

References

- Be AWH (1971) An ecological, zoogeographic and taxonomic review of recent planktonic foraminifera. In: Ramsay ATS (ed.) *Oceanic Micropaleontology*, vol. 1, p. 808. London: Academic Press.
- Berggren WA, Hilgen FJ, Langeris CG, et al. (1995a) Late neogene chronology: New perspectives in high-resolution stratigraphy. *Geological Society of America Bulletin* 107(11): 1272–1287.
- Berggren WA, Kent DV, Swisher CC, and Aubry M-P (1995b) A revised Cenozoic geochronology and chronostratigraphy. In: Berggren WA, Kent DV, Aubry M-P, and Hardenbol J (eds.) *Geochronology, Time Scales and Global Stratigraphic Correlation*, pp. 129–212. Tulsa: Society for Sedimentary Geology. Special Publication 54.
- Berggren WA, Kent DV, and Van Couvering JA (1985) Neogene geochronology and chronostratigraphy. In: Snelling NJ (ed.) *The Chronology of the Geological Record*, pp. 211–260. London: Geological society of London Memoir 10.
- Bolli HM, Saunders JB, and Perch-Nielsen K (eds.) (1985) *Plankton Stratigraphy*, p. 1032. Cambridge: Cambridge University Press.
- Brady H.B. (1884). Report on the foraminifera dredged by H.M.S. *Challenger* during the years 1873–1876. *Report on the Scientific Results of the Voyage of H.M.S. Challenger, Zoology* 9, p. 814.
- CLIMAP (1981) Seasonal reconstruction of the Earth's surface at the Last Glacial Maximum. *Geological Society of America Map and Chart Series* MC-36. pp. 1–18.
- d'Orbigny AD (1826) Tableau methodique de la classe des Cephalopodes. *Annales des Sciences Naturelles* 1(7): 96–314.
- Dowsett HJ (1989) Application of the graphic correlation method to Pliocene marine sequences. *Marine Micropaleontology* 35: 279–292.
- Dowsett HJ, Chandler MA, Cronin TM, and Dwyer GS (2005) Middle Pliocene sea surface temperature variability. *Paleoclimatology* 20: PA2014. <http://dx.doi.org/10.1029/2005PA001133>.
- Dowsett HJ, Robinson MM, Stoll DK, and Foley K (2010) Mid-Piacenzian mean annual sea surface temperature analysis for data-model comparisons. *Stratigraphy* 7(2–3): 189–198.
- Ericson DB and Wollin G (1956) Correlation of six cores from the equatorial Atlantic and the Caribbean. *Deep Sea Research* 3: 104–125.
- Hutson WH (1979) The Agulhas current during the late Pleistocene: Analysis of modern faunal analogs. *Science* 207: 64–66.
- Imbrie J and Kipp NG (1971) A new micropaleontological method for quantitative paleoclimatology: Applications to a late Pleistocene Caribbean core. In: Turekian KK (ed.) *The Late Cenozoic Glacial Ages*, pp. 71–181. New Haven: Yale University Press.

- Lisiecki LE and Raymo ME (2005) A Plio-Pleistocene stack of 57 globally distributed benthic $\delta^{18}\text{O}$ records. *Paleoceanography* 20: PA1003. <http://dx.doi.org/10.1029/2004PA001071>.
- Loeblich AR and Tappan H (1988) *Foraminiferal Genera and their Classification* p. 970. New York: Van Nostrand Reinhold.
- MARGO (2009) Constraints on the magnitude and patterns of ocean cooling at the Last Glacial Maximum. *Nature Geoscience* 2: 127–132. <http://dx.doi.org/10.1038/NGEO411>.
- Srinivasan MS and Kennett JP (1981) Neogene planktonic foraminiferal biostratigraphy and evolution. Equatorial to Subantarctic South Pacific. *Marine Micropaleontology* 6: 499–533.
- Wade BS, Pearson PN, Berggren WA, and Pälike H (2011) Review and revision of Cenozoic tropical planktonic foraminiferal biostratigraphy and calibration to the geomagnetic polarity and astronomical time scale. *Earth-Science Reviews* 104: 111–142.

Radiolarians and Silicoflagellates

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Introduction

This article discusses the application of radiolarians and silicoflagellates to Quaternary studies. For the most part, the article discusses radiolarians, marine protist microzooplankton, whose siliceous shells are found in sediments as old as the Cambrian. Radiolarians are widely used in Quaternary stratigraphy and paleoceanographic research. Silicoflagellates are marine protist phytoplankton. Their shells, also of opaline silica, are uncommon in sediments but occur sporadically from early Cretaceous to the Recent, and have occasionally been used for stratigraphy and paleoceanographic purposes.

General books on radiolarians include Anderson (1983) on the group's biology and De Wever et al. (2001) on fossil forms. Boltovskoy et al. (2010) provide a synthesis of biogeographic and ecologic information for living species, while a reasonably comprehensive list of named living species is maintained online at WoRMS. A short general review of radiolarian research is provided by Lazarus (2005). Monographic length literature on silicoflagellates is essentially lacking, but short chapters by Perch-Nielsen (1985) (emphasizing biostratigraphy), Bukry (1995) and McCartney (1995) (emphasizing biology) provide a reasonable introduction to the group and its literature.

Radiolarians

Modern Ecology and Preservation

The term radiolarian refers to several closely related groups including Phaeodarians and Acantharians but only the polycystine radiolarians (with two primary divisions, Spumellaria and Nassellaria) form robust siliceous shells that fossilize well, and all paleontologic uses to date are of polycystines. Radiolarians are small (~40–400 µm) grazers of phytoplankton, bacteria, and exported organic detritus in the surface to deep water layers of all modern oceans. Radiolarian shells (Figure 1) show an amazing range of shapes and geometries, and can be quite elaborate. Ever since Haeckel first published his illustrations of radiolarians, they have attracted the interest of artists and designers. For scientists, the complex forms of radiolarian shells provide numerous morphologic characteristics for distinguishing species. There are approximately 400 known living species of polycystine radiolarians (WoRMS, 2010), most of which are also found as fossils in Quaternary sediments. Most species are restricted to specific habitats. Some are found only in polar regions, others only in tropical waters, while yet others are found only in deep water masses (Boltovskoy et al., 2010; Figure 2). Radiolarians, as is generally true of siliceous skeletons, are preserved in sediments only when the flux of opaline material into the surface sediment is relatively high and dilution within the sediment by other components is not

too strong. They generally are most common in ocean sediments where carbonate preservation is poor: polar regions, deep Indo-Pacific sediments along the equator, and ocean margin upwelling zones. In the modern ocean, radiolarian preservation patterns closely mirror that of the more abundant producers of opal – diatoms and both distributions are similar to that of surface ocean productivity (Figure 3). The presence of radiolarians is thus sometimes taken as an indicator of productivity, although this is more a secondary coincidence due to the linkage of preservation patterns to diatom productivity in surface waters.

Preparation Methods

Quaternary radiolarians (and silicoflagellates) are extracted from sediments using simple chemicals such as HCl and H₂O₂ to dissolve/disaggregate the sedimentary matrix, and sieving to obtain clean specimens. Smaller radiolarian species and larger silicoflagellates require 38 µm sieves, although this often causes significant dilution by diatom frustules in diatomaceous sediments. Larger mesh sieves, particularly 63 µm, are often used for biostratigraphic studies for this reason, but may not adequately recover many small forms important for ecologic analysis. Transmitted light microscopes with brightfield illumination and 100–400× total magnification are fully adequate for observation and counting purposes. SEM imaging rarely yields significant additional morphologic information and is primarily used for older (e.g., Mesozoic) faunas where recrystallization of opal to other minerals makes transmitted light observation difficult.

Applications

Water mass boundaries

The geographic distribution of most marine plankton closely reflects that of the water masses they live in and is also true of radiolarians, which show a strong latitudinal diversity gradient as well (Figure 4). Highly diverse faunas of radiolarians (up to 200 species) are found in the warm surface waters of the tropics and subtropics. Mid-latitude surface waters contain a lower diversity assemblage, partially of cosmopolitan forms, and partially of species characteristic of temperate regions. Distinctly different species are found in the lower diversity assemblages (50–100 species) of polar regions. Some of these polar species appear to be high-latitude cosmopolites, but others are found only in one or two of the three modern polar regions (North Pacific; Norwegian-Greenland Sea, and Southern Ocean) (Figure 2). Other species are primarily associated with subpolar fronts, or exclusively in low- to mid-latitude upwelling zones (Figure 2). This remarkably detailed faunal adaptation to surface water mass distributions is largely preserved in the fossil faunas found in underlying surface

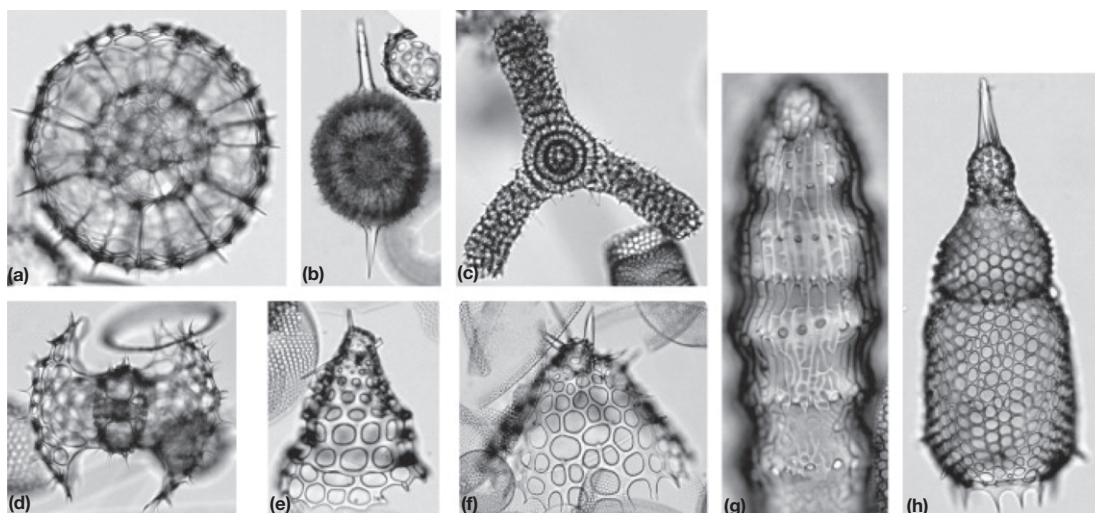


Figure 1 Examples of radiolarian shells (a–d, Spumellaria; e–g, Nassellaria). (a) *Actinomma antarcticum*, (b) *Stylatractus universus*, (c) *Euchitonia* sp., (d) *Tetrapyle* sp., (e) *Cycladophora davisiana*, (f) *Ceratocyrtis histicosa*, (g) *Siphocampe arachnea*, and (h) *Theocorythium trachelium*.

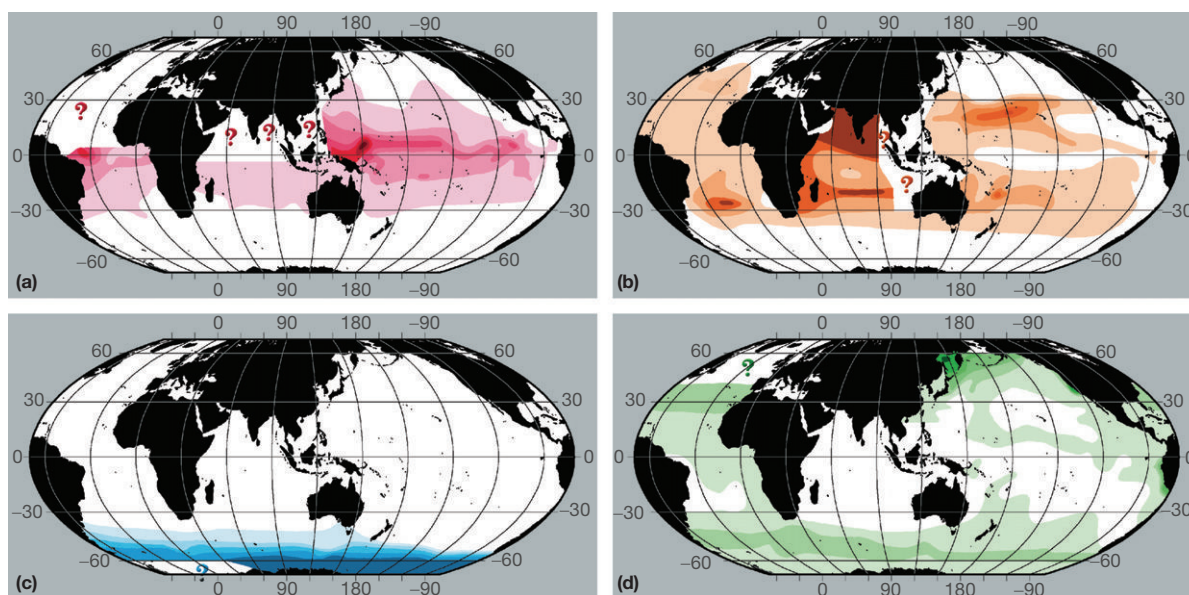


Figure 2 Radiolarian biogeographic patterns. Darker shades indicate higher relative abundance within the whole radiolarian assemblages in surface sediments. The distributions of four species are shown. Most living radiolarian species can be classified as having one of these distribution patterns: tropical taxa (*Spongaster tetras*, a), subtropical (*Acrosphaera lappacea*, b), polar (*Antarctissa denticulata*, c), and frontal/gyre margin/upwelling (*Cycladophora davisiana*, d). Redrawn from Lombardi G and Boden G (1985) Modern radiolarian global distributions. Cushman Foundation for Foraminiferal Research, Special Publication, 16A.

sediments, despite changes in individual species' relative abundances due to preservation artifacts. Thus, the biogeography of fossil radiolarian faunas can be used to map out the history of ocean water mass geography, and, particularly in high latitudes, to a much finer resolution than is possible with calcareous microfossils, as these latter have very low diversities (generally only a few rather cosmopolitan taxa) in cool temperate to polar regions. A good Quaternary example of this method using radiolarians is [Morley \(1989\)](#).

Biostratigraphy

Although biostratigraphic applications of radiolarians are primarily in older sediments (indeed, most radiolarian workers are engaged in biostratigraphy of Mesozoic and Paleozoic rocks), radiolarians are also occasionally important as biostratigraphic markers in Quaternary sediments. Two important examples include the last occurrence of the spumellarian species *Stylatractus universus* and abundance fluctuations in the nassellarian species *Cycladophora davisiana*.

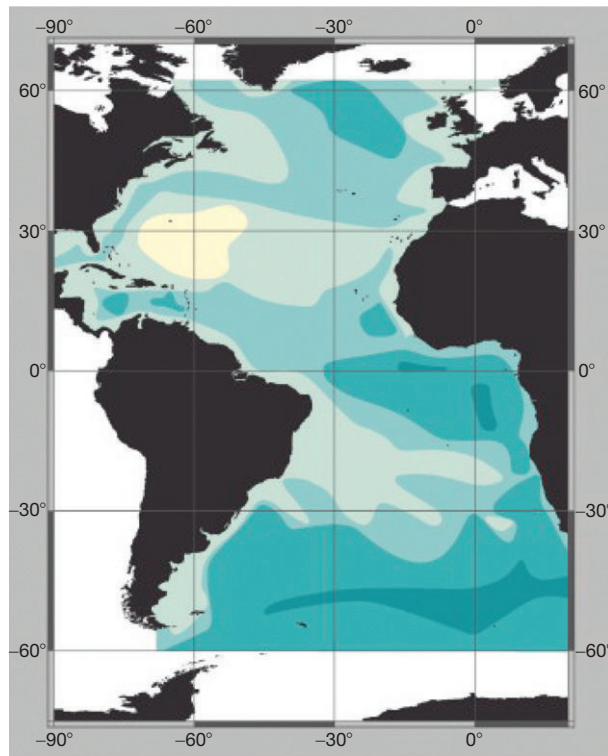


Figure 3 Radiolarian abundance in Atlantic Ocean surface sediments. Darker shades indicate higher absolute abundance of radiolarians (shells/gram sediment). The distribution primarily reflects opal preservation in sediments, due to the saturation of silica in sediment pore waters in opal-rich sediments. Opal abundance in sediments is primarily caused by high surface water productivity and the dominance in such waters of siliceous diatom phytoplankton. In regions without abundant opal production (e.g., oligotrophic central gyres) virtually all radiolarian skeletons eventually dissolve as modern sea water is very undersaturated in silica. Redrawn from Goll RM and Bjørklund KR (1971) Radiolaria in surface sediments of the North Atlantic. *Micropaleontology* 17(4): 434–454; Goll RM and Bjørklund KR (1974) Radiolaria in surface sediments of the South Atlantic. *Micropaleontology* 20(1): 38–75.

The last occurrence datum (LAD) of *Stylatractus universus* Hays 1970 (Figure 1) was first noted from the late Pleistocene in Antarctic sediments and subsequently noted in several other ocean basins, all in the late Pleistocene (see review in Hays and Shackleton, 1976). Hays and Shackleton (1976) calibrated the age of this event to the same interval in glacial–interglacial oxygen isotope curves in several widely distributed piston cores from the Indo-Pacific (stage 11/12 boundary; currently ca. 430 ka (Moore et al., 1995)). Morley and Shackleton (1978) subsequently extended this calibration with additional cores from the Atlantic Ocean. This datum is thus global (Figures 5 and 6) and has proved to be very useful in interpreting and cross-correlating oxygen isotope curves during the CLIMAP and SPECMAP projects, as the event conveniently falls approximately in the middle of the Brunhes magnetochron, thus providing a globally recognizable calibration point above the base of the Brunhes.

Of all radiolarian taxa used in Quaternary research, *Cycladophora davisiana* Ehrenberg 1862 (Figure 1) is perhaps the most important, being valuable both for biostratigraphic and paleoecologic studies. In addition to using this species as an

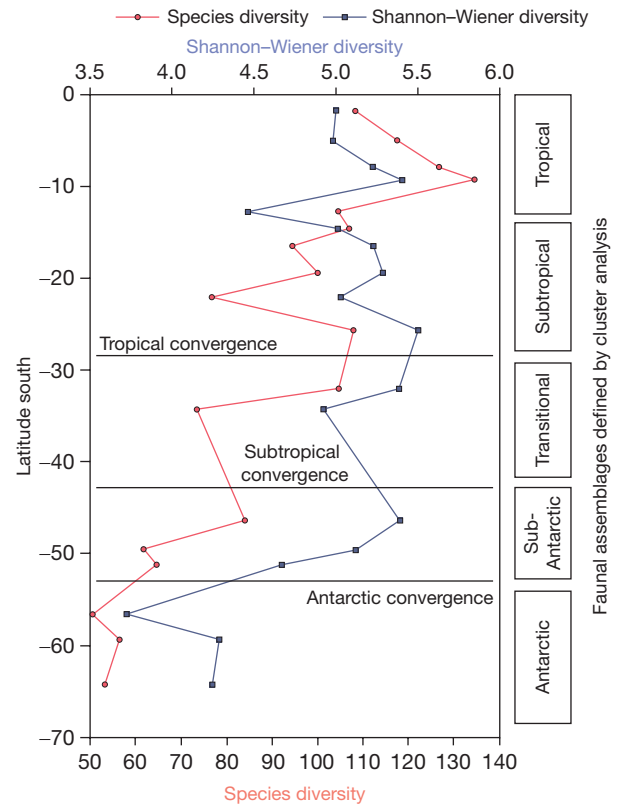


Figure 4 Radiolarian diversity versus latitude, showing faunal groups in relation to oceanographic fronts. The red curve shows the number of species found in each of a series of surface sediment samples taken in a transect from the western equatorial Pacific to the Southern Ocean southeast of New Zealand. The blue curve shows the Shannon–Wiener relative abundance weighted diversity metric calculated from the same samples. Patterns of covariation in abundance in the entire sample set were analyzed using a clustering algorithm to classify the samples into a small set of similar faunal groups. The boundaries between these groups, shown on the right side of the figure, closely match physical oceanographic boundaries that are crossed by the transect (labeled lines). Redrawn and replotted from Boltovskoy D (1987) Sedimentary record of radiolarian biogeography in the equatorial to Antarctic western Pacific Ocean. *Micropaleontology* 33(3): 267–281.

important indicator of glacial conditions, Hays et al. (1976a) in their seminal ‘pacemaker’ paper (which established the dominance of Milankovitch frequencies in Quaternary climate change) also noted the importance of *C. davisiana* to Quaternary stratigraphy. This was more explicitly studied by Morley and Hays (1979) who noted that this species increased dramatically in relative abundance in synchrony with oxygen isotope glacial–interglacial stages in North Atlantic cores. In a series of papers this was also established for most other ocean regions (see review in Brathauer et al., 2001). In modern oceans, this species is common only in the Sea of Okhotsk and in the centers of low to mid-latitude upwelling regions (Figure 2; Lombardi and Boden, 1985). Its increased abundance in glacial intervals (Figure 7) is thought to reflect changes in midwater ecology, as this species primarily inhabits the intermediate (subthermocline) layers in the Sea of Okhotsk today (Nimmergut and Abelman 2002).

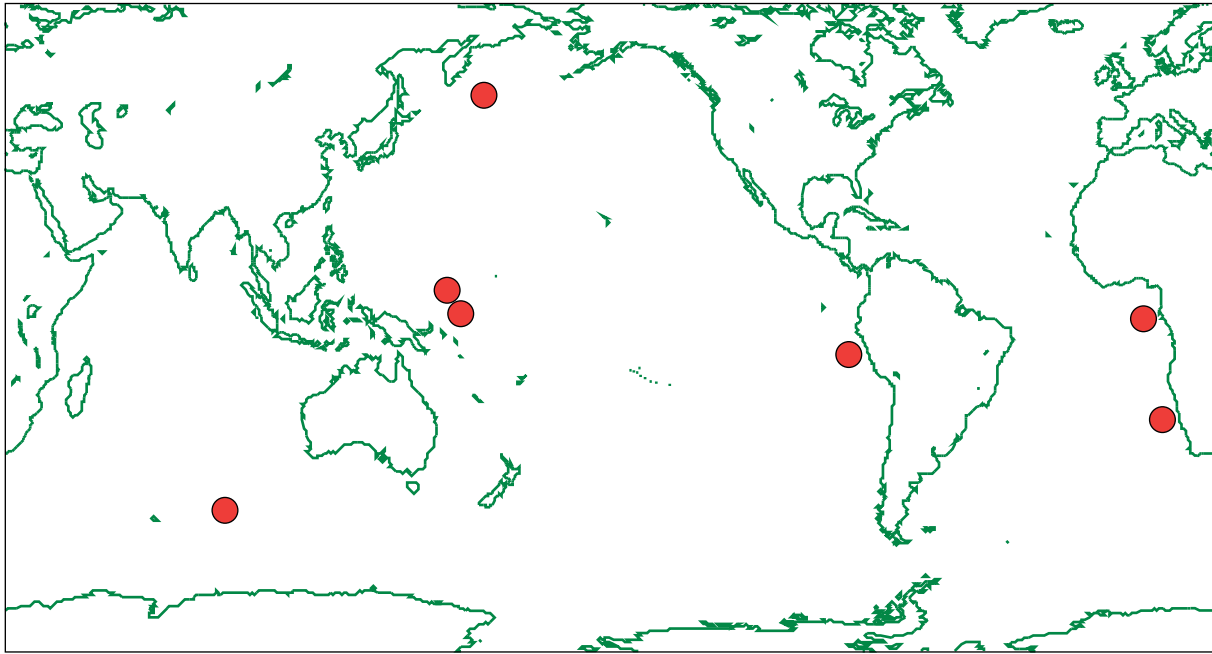


Figure 5 *Stylatractus universus* LAD calibration core sites location map (based on data in Hays and Shackleton, 1976; Morley and Shackleton, 1978). The last occurrence of this species has been correlated to the same stage of the globally synchronous time scale of the Pleistocene oxygen isotope curve.

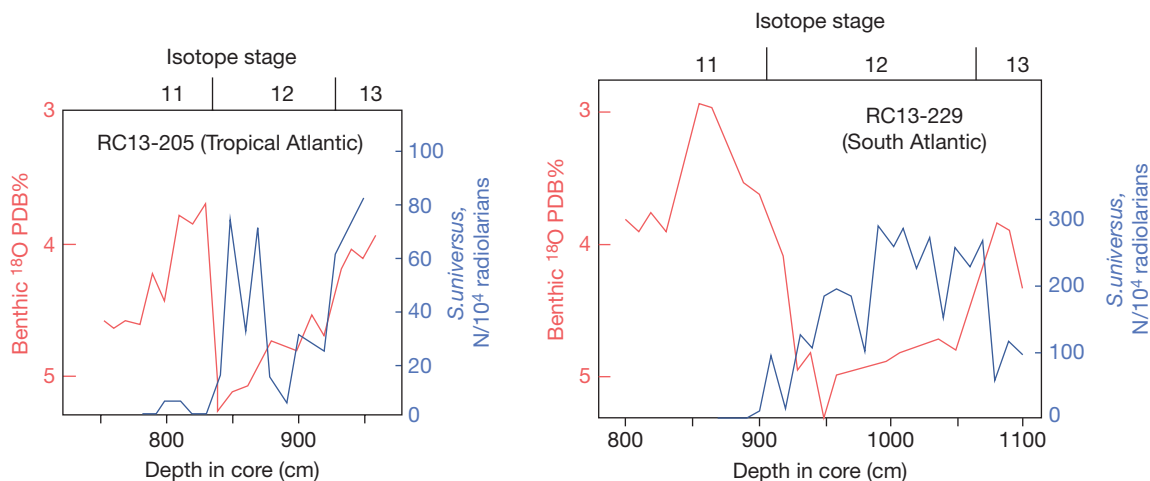


Figure 6 Isochroneity of the LAD *Stylatractus universus* as shown by comparison to oxygen isotope stratigraphy in two representative cores. Blue lines show abundance of *S. universus*; red lines the oxygen isotope composition of benthic foraminiferal shells in down-core sediment samples. The isotope curves are part of a longer time series of data from the cores which permit each cycle of the curve to be uniquely identified to the global reference standard isotope stages (top of each figure), and thus to absolute time. *S. universus* abundance declines to very low levels during the transition between isotope stages 12 and 11, and the species disappears completely within the early part of isotope stage 11. Redrawn from Morley JJ and Shackleton NL (1978) Extension of the radiolarian *Stylatractus universus* as a biostratigraphic datum to the Atlantic Ocean. *Geology* 6(5): 309–311.

Many other radiolarian biostratigraphic events are found in the Quaternary and are of at least regional importance. Summaries of biostratigraphic zonations for radiolarians in different geographic regions that include Quaternary events can be found in Sanfilippo et al. (1985).

It should be noted that most radiolarian biostratigraphic events are either only found within one biogeographic region, or if present in more than one region, may have slightly

different ages, or even show regular gradational variation in event age with geography (Moore et al., 1995). This is due to the very close adaptation of most radiolarian species to specific water masses and to environmental gradients within them. The high precision and global validity of the LAD of *S. universus* is exceptional and is most likely due to a preference for globally distributed deep water masses, although this can only be inferred for this now extinct form.

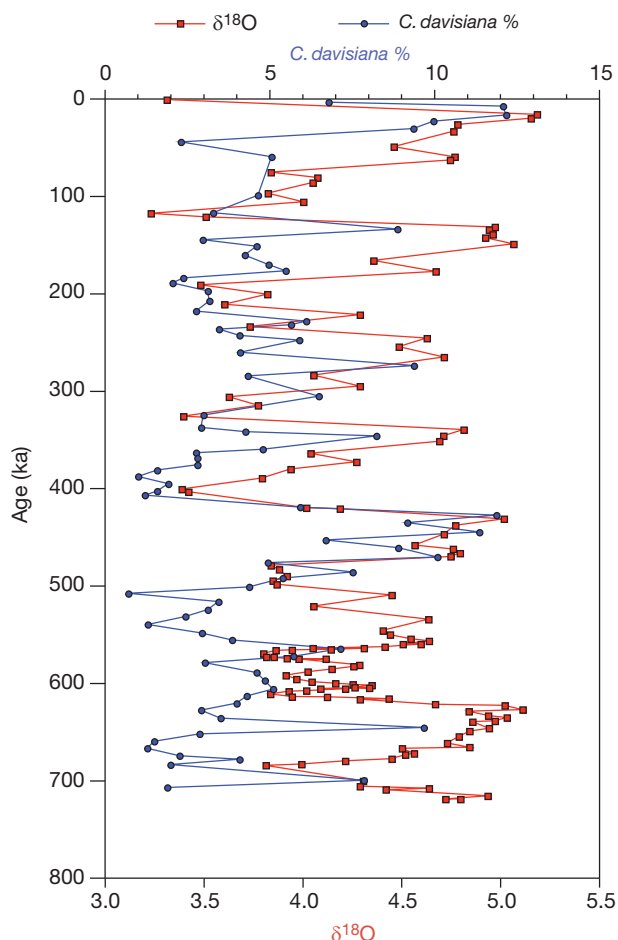


Figure 7 Covariation of *Cycladophora davisiana* and glacial-interglacial oxygen isotopes in down-core samples from a sub-Antarctic deep-sea sediment core. Blue line shows variation in relative abundance of *C. davisiana*, red line shows oxygen isotope composition of benthic foraminiferal shells (glacial intervals are indicated by more positive oxygen isotope values). *C. davisiana* is generally much more abundant during glacial intervals. *C. davisiana* abundance curves can be used in sediment cores lacking stable isotope data to identify glacial-interglacial time intervals. Reproduced from Brathauer U, Abelmann A, Gersonde R, Niebler H-S, and Fütterer DK (2001) Calibration of *Cycladophora davisiana* events versus oxygen isotope stratigraphy in the subantarctic Atlantic Ocean – A stratigraphic tool for carbonate-poor Quaternary sediments. *Marine Geology* 175: 167–181.

Paleotemperature estimation

One of the main contributions marine microfossils make to Quaternary studies is the estimation of past ocean water temperatures. Radiolarians have been one of the most important groups of fossils for such studies. The most important early paleotemperature estimates using radiolarians were those of CLIMAP, both in down-core studies such as Hays et al. (1976a) and in a series of regional studies mapping out the temperatures of the surface ocean during the Last Glacial Maximum, as summarized in Hays et al. (1976b) and Moore (1978, 1980). Radiolarians, with their relatively high diversity and insensitivity to carbonate dissolution, provided much of the temperature reconstructions for CLIMAP in high-latitude regions, as well as in deeper basins of the Pacific (CLIMAP, 1976).

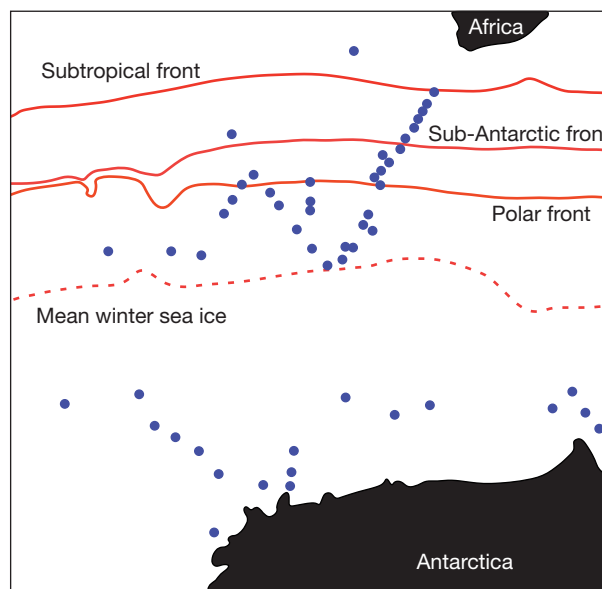


Figure 8 Location of surface sediment samples used to create a radiolarian temperature transfer function for the sub-Antarctic to Antarctic South Atlantic. The relative abundances of species from these samples were analyzed using factor analysis to define four main groups of species that vary in geographic abundance in a uniform way. Reproduced from Lazarus DB (2005) A brief review of radiolarian research. *Paläontologische Zeitschrift* 79(1): 183–200; redrawn from Abelmann A, Brathauer U, Gersonde R, Sieger R, and Zielinski U (1999) Radiolarian-based transfer function for the estimation of sea surface temperatures in the Southern Ocean (Atlantic sector). *Paleoceanography* 14(3): 410–421.

Several more recent studies using similar methods have been published, including those of Heusser and Morley (1985) in the North Pacific and Dolven et al. (2002) in the Norwegian–Greenland Sea. These more recent studies have sometimes introduced methodological improvements over those of CLIMAP. Pisias et al. (1997), in a review of radiolarian paleotemperature equations for the Pacific, noted the problem of covariation in summer and winter temperatures, and instead created regression equations for mean annual water temperature and seasonality in water temperature, which are much less highly correlated to each other. Abelmann et al. (1999; Figures 8–10) in the Antarctic recognized differences in water depth habitat of different radiolarian species and those species known to live only in subsurface layers were excluded from the set of taxa used to estimate surface water temperatures. Cortese et al. (2005) used a neural network program instead of a multivariate statistical method to estimate paleotemperatures in Norwegian–Greenland sea sediments. However, despite these methodological improvements, the magnitude of error in paleotemperature estimation has improved only slightly, for example, to 1–2 °C, which is similar to that achieved for other microfossil groups. It is likely that this level of error represents limits due to other methodologic and ecologic problems, such as taxonomic error, and more importantly, limits on the extent to which radiolarian assemblage composition is actually controlled by water temperature.

Paleoproductivity estimation

Organisms in the real world are rarely adapted to and primarily responsive to only one major environmental parameter such

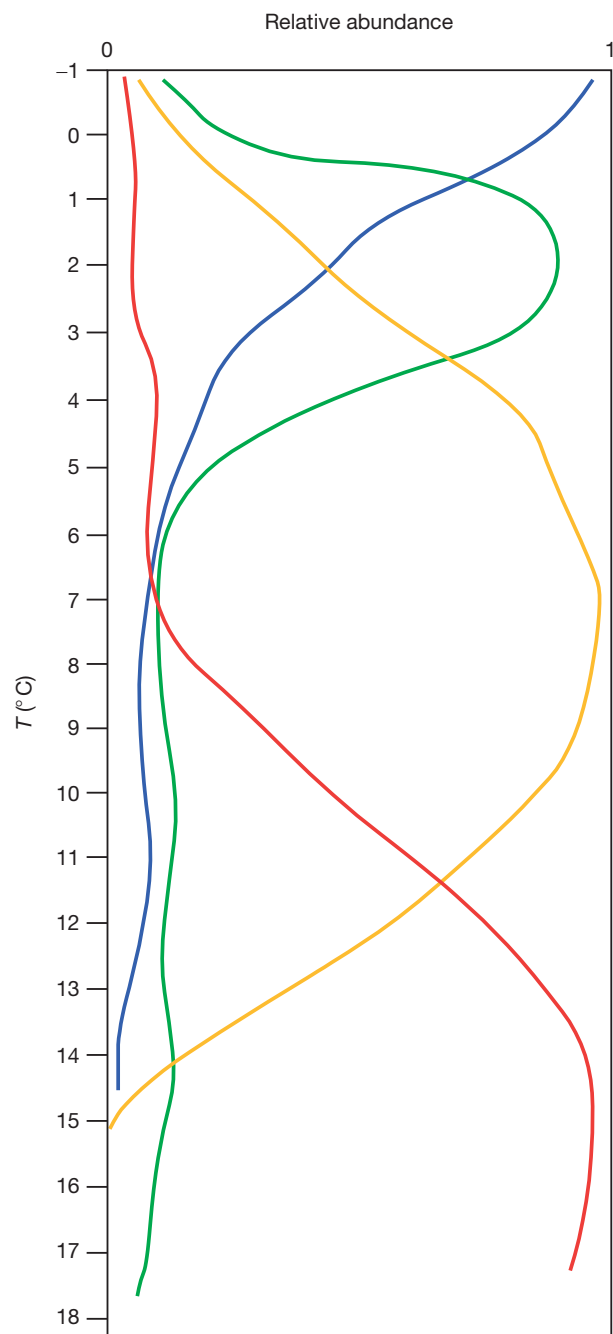


Figure 9 South Atlantic surface sediment radiolarian assemblage factor distributions (relative abundance in sediment samples; see [Figure 8](#)) versus modern surface ocean temperature. Each differently colored curve represents a distinct geographic radiolarian species assemblage (red, tropical/subtropical; yellow, subpolar, green, Polar Front; blue, Southern Ocean. Reproduced from Lazarus DB (2005) A brief review of radiolarian research. *Paläontologische Zeitschrift* 79(1): 183–200; redrawn from Abelmann A, Brathauer U, Gersonde R, Sieger R, and Zielinski U (1999) Radiolarian-based transfer function for the estimation of sea surface temperatures in the Southern Ocean (Atlantic sector). *Paleoceanography* 14(3): 410–421.

as temperature. This is certainly also true of radiolarians, although to what degree is generally unknown. This is, however, substantial indirect evidence that the distributions of radiolarians and other planktonic organisms are controlled at least

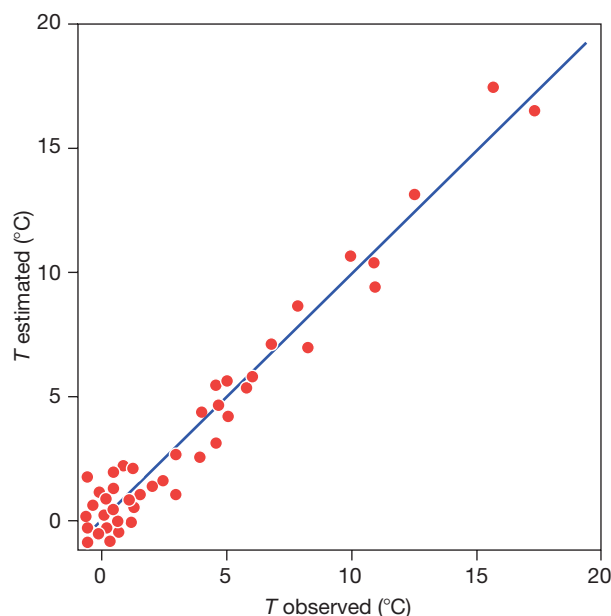


Figure 10 Correlation between actual ocean water temperature and water temperature estimated from radiolarian transfer function in South Atlantic. Estimated temperatures are from multiple regression of radiolarian factors defined in earlier steps ([Figures 8 and 9](#)). Reproduced from Lazarus DB (2005) A brief review of radiolarian research. *Paläontologische Zeitschrift* 79(1): 183–200; redrawn from Abelmann A, Brathauer U, Gersonde R, Sieger R, and Zielinski U (1999) Radiolarian-based transfer function for the estimation of sea surface temperatures in the Southern Ocean (Atlantic sector). *Paleoceanography* 14(3): 410–421.

as much by food availability as by any other factor. This was already apparent in the global distributional data collected for radiolarians from surface sediments by the CLIMAP project (summarized in [Lombardi and Boden, 1985](#); [Figure 2](#)), as well as in analyses of species distribution in the water column by [Casey et al. \(1982\)](#). Many species distributions do not match temperature distributions but instead are more closely aligned with frontal systems or follow large-scale patterns of primary productivity in the ocean. [Molina-Cruz \(1977, 1984\)](#) first explicitly attempted to apply the distribution of radiolarians to map ocean productivity. This effort was limited by CLIMAP's focus on temperature reconstruction, incomplete knowledge of radiolarian habitats in the water column, and by the limited data available at that time on ocean productivity distribution.

A second approach to this problem was developed by [Nigrini and Caulet \(1992\)](#), based on their observation of a recurrent pattern of unusual species distributions in association with low- to mid-latitude upwelling zones. They noted that certain radiolarian species are found, either exclusively or in unusual abundance, in widely separated upwelling regions, including the Somali system in the Indian Ocean and the Peru-Chile upwelling system in the Pacific. The relative summed abundance of these 'upwelling' species in the total radiolarian assemblage was used to define an upwelling radiolarian index, or URI. In subsequent papers Caulet and coworkers generated URI time series for several cores in late Pleistocene sediments, mostly from the Somali upwelling system, and compared to other inferred proxies of paleoproductivity such as planktonic foraminiferal faunas and $\delta^{13}\text{C}$. In general, URI curves show moderate to good correlation with other paleoproductivity

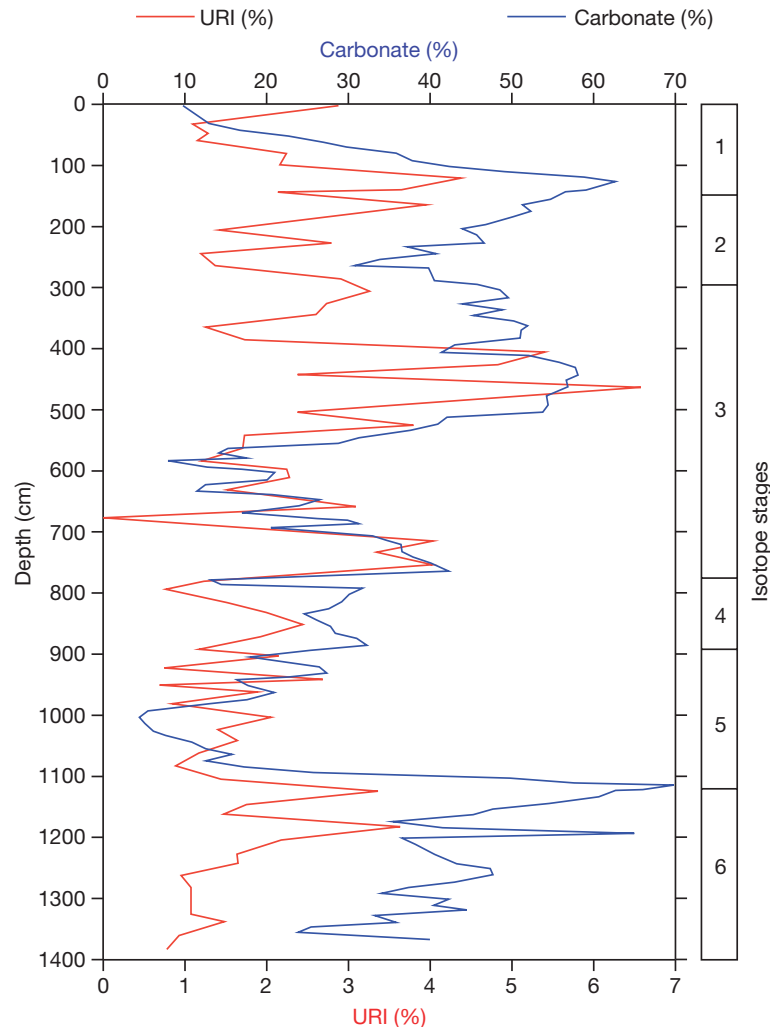


Figure 11 Variation over the last ca. 140 ka in the upwelling radiolarian index (URI) versus accumulation rate of carbonate in a tropical Indian Ocean core located above the calcium carbonate compensation depth (CCD). Because the core is situated above the CCD and has only minor nonbiogenic material, the bulk accumulation rate is used as a proxy for productivity. Replotted from data shown in [Caulet et al. \(1992\)](#).

proxies, although substantial differences are also often seen in certain time intervals ([Figure 11](#)). URI curves also occasionally show patterns that are difficult to interpret, or are at variance to other paleoceanographic proxies of productivity ([Jacot des Combes and Abelmann, 2007](#)). Use of the URI as a paleoproductivity proxy is, however, limited by many factors. These include a lack of knowledge of the underlying biologic controls on species distributions (what specifically causes these enhanced relative abundances within the upwelling zone), by the limited stratigraphic range of many of the species, several of which have so far only been reported from the Pleistocene, and by latitude. In higher latitudes, several of the species used are relatively common even outside of upwelling regions and thus not suitable as markers of upwelling activity.

[Jacot des Combes et al. \(1999\)](#) began the development of a third approach to estimating paleoproductivity, based on an early proposal by [Casey et al. \(1982\)](#), who had suggested that the relative abundance of deep-water dwelling species should be enhanced in areas of upwelling and/or enhanced export productivity. This would be due to both the cooling and

eutrophication of surface waters, which would depress the abundance of warm, oligotrophic adapted surface water taxa, and by the enhanced abundance of deeper water taxa, that, directly or indirectly, would be able to feed off the higher levels of exported organic matter to intermediate and deep water layers. Methods based on this ecologic principle were christened WADE methods (for WAter Depth Ecology) by [Lazarus \(2005\)](#). In [Jacot des Combes](#)' approach, the ratio in relative abundance between a small number of surface water and deep water taxa was employed as a thermocline–surface radiolarian index (TSRI), with higher relative abundances of deep-water taxa yielding higher TSRI values and an inferred higher level of export productivity. The TSRI index often correlates reasonably well with other productivity indicators, although down-core patterns of variation can be complex and shift over time ([Jacot des Combes et al., 2005](#)). The TSRI can also be applied in open ocean regions outside of upwelling regions, as demonstrated by [Jacot des Combes et al. \(1999\)](#). The TSRI, however, is limited by the small number of species used. Due to complex adaptation to multiple environmental constraints, individual species of radiolarians (and generally most organisms) are not

always common in regions or individual samples within their preferred biogeographic province, or even at 'optimum' conditions for any one environmental parameter. This variation in abundance can strongly bias indices, such as the TSRI, that are based on just a few taxa. Weinheimer (2001) and Lazarus et al. (2006, 2008) developed an extended version of the WADE

method by employing a much larger number of taxa (>50) to document both long-term trends and glacial–interglacial scale variation in radiolarian faunas within the Benguela upwelling system (Figure 12(a) and 12(b)).

WADE methods, due to their lack of dependence on specific taxa, can in principle be applied even to older sediments and

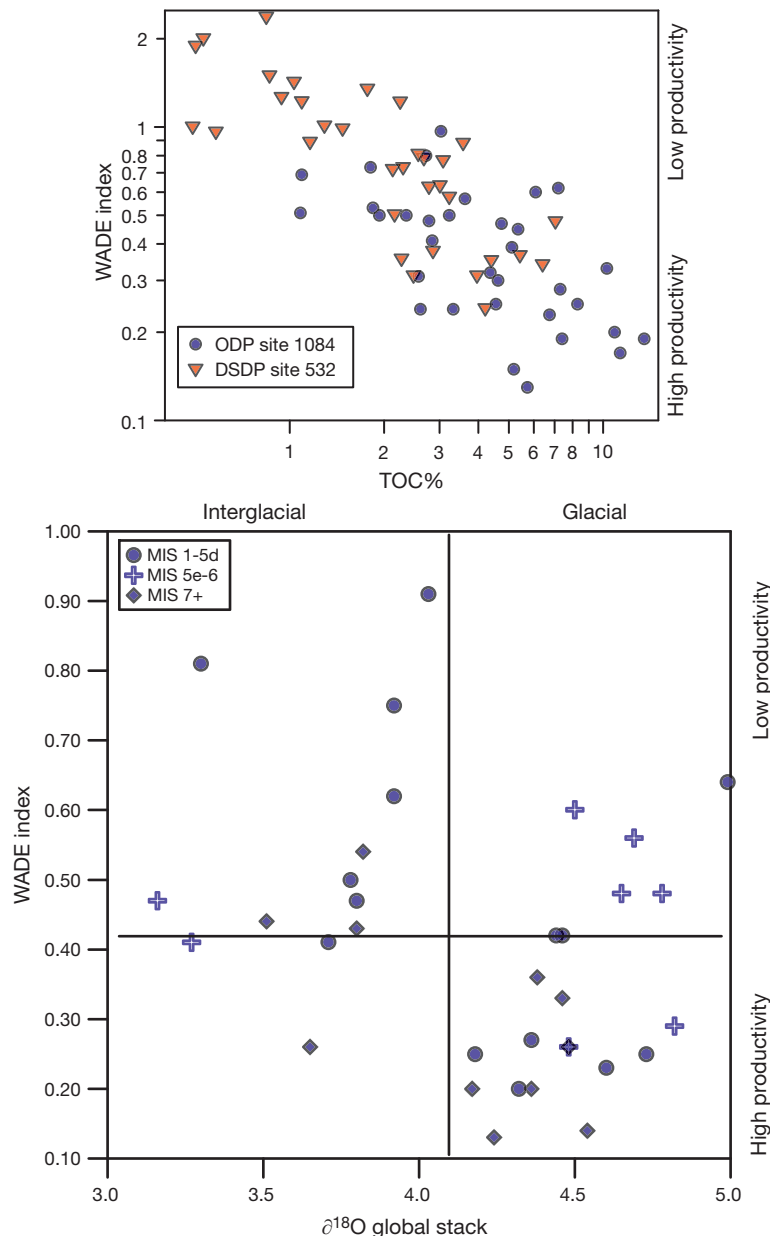


Figure 12 Radiolarian Water Depth Ecology (WADE) indices based on large numbers (>50) of surface and deep-water dwelling species, and their correlation to environmental change indicators in sediments from the Benguela Upwelling system off of Africa in the western South Atlantic. (a) Correlation of WADE index to total organic carbon content of sediments (TOC) over the last 6 million years in two deep-sea drilling sites: Site 1084 near the center of upwelling, and Site 532 near the edge. (b) WADE productivity index in samples covering the last few glacial–interglacial cycles from upwelling Site 1084. Samples are assigned estimated global benthic $\delta^{18}\text{O}$ values according to sample age as a proxy for degree of glacial–interglacial environment. Glacial samples have generally much higher productivity, as indicated by lower values of the WADE index. Reproduced from Lazarus D, Bittniok B, Diester-Haass L, Meyers P, and Billups K (2006) Comparison of radiolarian and sedimentologic paleoproductivity proxies in the latest Miocene–Recent Benguela Upwelling System. *Marine Micropaleontology* 60: 269–294; Lazarus D, Bittniok B, Diester-Haass L, et al. (2008) Radiolarian and sedimentologic paleoproductivity proxies in late Pleistocene sediments of the Benguela Upwelling System, ODP Site 1084. *Marine Micropaleontology* 68: 223–235.

with extinct taxa, so long as the water depth preferences of these forms can be independently established. WADE methods are also based on a clear ecologic model and thus the data should be relatively easy to interpret. However, the choice of which taxa to use is still rather arbitrary and the number of studies using these methods is still very small. Furthermore, neither URI nor WADE indices have yet been calibrated in surface sediment assemblages to independent measurements of ocean productivity. It is thus too early to make any general conclusions as to their robustness.

Outlook for the Use of Radiolarians in Quaternary Studies

Radiolarians, due to their wide-ranging ecology, robust opaline shell chemistry, and high diversity, provide an important record of Quaternary marine environments that complements that provided by other microfossil groups such as diatoms and planktonic Foraminifera. They provide paleotemperature estimates and estimates of paleoproductivity and provide useful biochronologic information as well. Importantly, radiolarians provide useful information from sediments lacking well preserved carbonate. Such sediments are common in environmentally critical areas of the ocean, for example, high latitudes and low-latitude upwelling zones. The limitations to the use of radiolarians are primarily our own. For example, the first reasonably comprehensive lists of living species names, and maps of living species distributions, both appeared only quite recently (Boltovskoy et al., 2010; WoRMS, 2010) and many of the entries in these syntheses are still quite sketchy. We still know far too little about living radiolarian ecology, particularly for individual species, and even the descriptive taxonomy of Quaternary radiolarians is incomplete. As our taxonomic and ecologic knowledge of the group improves, the importance of radiolarians in Quaternary research is likely to increase.

Silicoflagellates

Silicoflagellate skeletons are generally intermediate in size between diatoms and radiolarians (mostly 20–100 μm), and often co-occur on microscope slides prepared for examination of these other groups (Figure 13). They thus are occasionally studied by diatom and radiolarian workers, as well as by silicoflagellate specialists. Their living diversity is very low – just a few species – and moderately low diversity has been characteristic of the group since its origin in the Cretaceous. The silicoflagellate skeleton is composed of tubular elements connected into simple ring-like structures, external radial spines, plus a central apical structure in some taxa. The taxonomy of silicoflagellates is based on the number of sides of the geometric basal ring and the complexity of the apical structure. The small number of resulting morphologic characters may contribute to the low described diversity of taxa. However, recent work suggests that paleoceanographically significant subspecies level variation may be present within traditional species definitions, which can be exploited in future research (Malinverno, 2010). Silicoflagellates are normally found as only a minor component of biosiliceous sediments, and then only when silica is relatively well

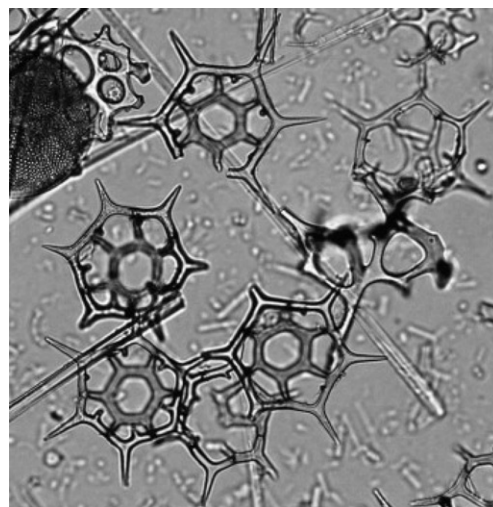


Figure 13 Silicoflagellate skeletons (*Distephanus* sp.). Skeleton size, with spines, c.40 μm .

preserved. Their occurrence can thus be used as a rough indicator of overall good opal preservation state, but the preservation characteristics of silicoflagellates have never been studied in detail.

Silicoflagellates have been studied in the plankton for decades and numerous studies have reported both seasonal and geographic variations in relative taxa abundances, and in overall levels of absolute and relative flux, in comparison to other groups (see e.g., Onodera and Takahashi, 2005). Thus, in principle it should be possible to use fossil silicoflagellate occurrence patterns to study past environments, as is done for radiolarians. However, uncertainties in species-level taxonomy, low overall diversity, and sporadic occurrence in surface sediments have limited the development of normal paleoenvironmental tools such as transfer functions and paleoenvironmental analysis has remained primarily qualitative. However, the ratio between the two current living genera, *Dictyocha* and *Distephanus*, has sometimes been useful as a relative temperature indicator in paleoceanographic research (Barron et al., 2010; Figure 14).

There are one or two Quaternary biostratigraphic events reported for silicoflagellates (McCartney, 1995; Perch-Nielsen, 1985) but these have not been used much in biostratigraphic research.

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See also: **Paleoceanography:** Paleoceanography An Overview.
Paleoceanography, Biological Proxies: Planktic Foraminifera.
Quaternary Stratigraphy: Continental Biostratigraphy.

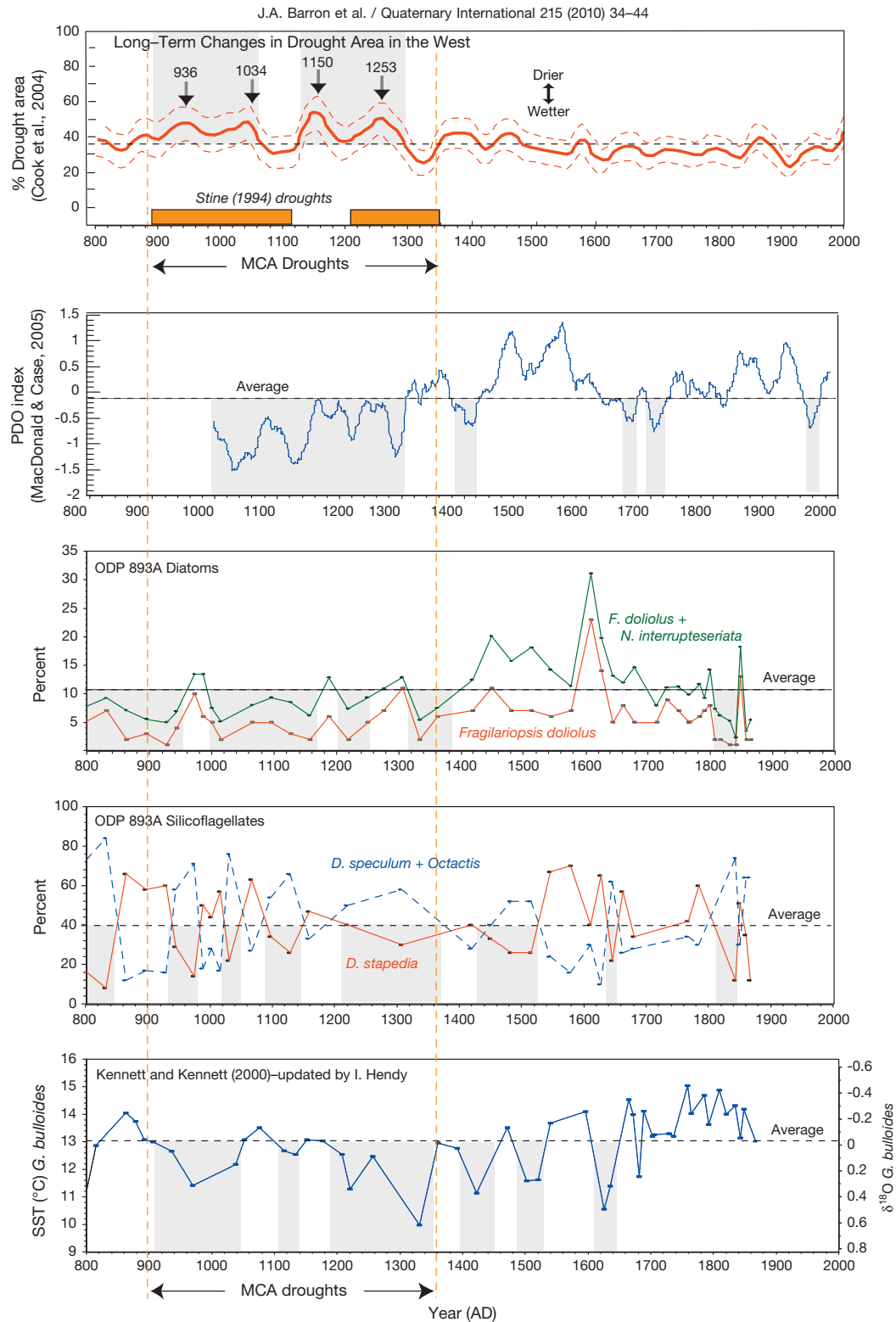


Figure 14 Variation in silicoflagellate generic ratios over the last 1100 years in ODP Hole 893A sediments from the Santa Barbara Basin (middle panel) compared to sea surface temperature estimates (SST) from planktonic Foraminifera oxygen isotopes (lower panel) and changes in key diatom species (upper panel). Despite low sample density, a good correlation between silicoflagellate ratios and estimated SST can be seen. Reproduced from Barron JA, Bukry D, and Field D (2010) Santa Barbara Basin diatom and silicoflagellate response to global climate anomalies during the past 2200 years. *Quaternary International* 215: 34–44.

References

- Abelmann A, Brathauer U, Gersonde R, Sieger R, and Zielinski U (1999) Radiolarian-based transfer function for the estimation of sea surface temperatures in the Southern Ocean (Atlantic sector). *Paleoceanography* 14(3): 410–421.
- Anderson OR (1983) *Radiolaria*. New York: Springer-Verlag 365 pp.
- Barron JA, Bukry D, and Field D (2010) Santa Barbara Basin diatom and silicoflagellate response to global climate anomalies during the past 2200 years. *Quaternary International* 215: 34–44.
- Boltovskoy D (1987) Sedimentary record of radiolarian biogeography in the equatorial to Antarctic western Pacific Ocean. *Micropaleontology* 33(3): 267–281.
- Boltovskoy D, Kling SA, Takahashi K, and Björklund K (2010) World atlas of distribution of living radiolaria. *Palaeontologia Electronica* 13(3): (web). c.230 pp.
- Brathauer U, Abelmann A, Gersonde R, Niebler H-S, and Fütterer DK (2001) Calibration of *Cycladophora davisiana* events versus oxygen isotope stratigraphy in the subantarctic Atlantic Ocean – A stratigraphic tool for carbonate-poor Quaternary sediments. *Marine Geology* 175: 167–181.
- Bukry D (1995) Silicoflagellates and their geologic applications. *Open-File Report 95–260*, pp. 1–23. Washington, DC: US Geological Survey.
- Casey RE, Spaw JM, and Kunze FR (1982) Polycystine radiolarian distributions and enhancements related to oceanographic conditions in a hypothetical ocean. *Transactions of the Gulf Coast Association of Geological Societies* 32: 319–332.
- Caulet JP, Venec-Peyre M-T, Vergnaud-Grazzini C, and Nigrini C (1992) Variation of South Somali upwelling during the last 160 ka: Radiolarian and foraminifera records in core MD 85674. In: Summerhayes CP, Prell WL, and Emeis KC (eds.) *Upwelling Systems: Evolution Since the Early Miocene*, pp. 379–389. London: Geological Society. Special Publication.
- CLIMAP (1976) The surface of the Ice-Age earth. *Science* 191(4232): 1131–1137.
- Cortese G, Dolven JK, Björklund K, and Malmgren BA (2005) Late Pleistocene–Holocene radiolarian paleotemperatures in the Norwegian Sea based on artificial neural networks. *Palaeogeography, Palaeoclimatology, Palaeoecology* 224: 311–332.
- De Wever P, Dumitrica P, Caulet JP, Nigrini C, and Caridroit M (2001) *Radiolarians in the Sedimentary Record*. Amsterdam: Gordon and Breach 533 pp.
- Dolven JK, Cortese G, and Björklund KR (2002) A high-resolution radiolarian-derived paleotemperature record for the late Pleistocene–Holocene in the Norwegian Sea. *Paleoceanography* 17(4): 1–13.
- Goll RM and Björklund KR (1972) Radiolaria in surface sediments of the North Atlantic Ocean. *Micropaleontology* 17(4): 434–454.
- Goll RM and Björklund KR (1974) Radiolaria in surface sediments of the South Atlantic. *Micropaleontology* 20(1): 38–75.
- Hays JD (1970) Stratigraphy and evolutionary trends of radiolaria in North Pacific deep-sea sediments. *Geological Society of America Memoirs* 126: 185–218.
- Hays JD, Imbrie J, and Shackleton NJ (1976) Variations in the earth's orbit: Pacemaker of the ice ages. *Science* 194(4270): 1121–1132.
- Hays JD, Lozano JA, Shackleton N, and Irving G (1976) Reconstruction of the Atlantic and western Indian Ocean sectors of the 18,000 B.P. Antarctic Ocean. In: Cline RM and Hays JD (eds.) *Investigation of Late Quaternary Paleoceanography and Paleoclimatology*, pp. 337–372. Boulder, CO: Geological Society of America Memoir.
- Hays JD and Shackleton NJ (1976) Globally synchronous extinction of the radiolarian *Stylatractus universus*. *Geology* 4(11): 649–652.
- Heusser LE and Morley JJ (1985) Pollen and radiolarian records from deep-sea core RC14-103: Climatic reconstructions of northeast Japan and northwest Pacific for the last 90,000 years. *Quaternary Research* 24: 60–72.
- Jacot des Combes H and Abelmann A (2007) A 350-ky radiolarian record off Luederitz, Namibia – Evidence for changes in the upwelling regime. *Marine Micropaleontology* 62: 194–210.
- Jacot des Combes H, Caulet JP, and Tribouillard NP (1999) Pelagic productivity changes in the equatorial area of the northwest Indian Ocean during the last 400,000 years. *Marine Geology* 158(1–4): 27–55.
- Jacot des Combes H, Caulet JP, and Tribouillard NP (2005) Monitoring the variations of the Socotra upwelling system during the last 250 kyr: A biogenic and geochemical approach. *Palaeogeography, Palaeoclimatology, Palaeoecology* 223: 243–259.
- Lazarus DB (2005) A brief review of radiolarian research. *Paläontologische Zeitschrift* 79(1): 183–200.
- Lazarus D, Bittniok B, Diester-Haass L, Meyers P, and Billups K (2006) Comparison of radiolarian and sedimentologic paleoproductivity proxies in the latest Miocene–Recent Benguela Upwelling System. *Marine Micropaleontology* 60: 269–294.
- Lazarus D, Bittniok B, Diester-Haass L, et al. (2008) Radiolarian and sedimentologic paleoproductivity proxies in late Pleistocene sediments of the Benguela Upwelling System, ODP Site 1084. *Marine Micropaleontology* 68: 223–235.
- Lombardi G and Boden G (1985) Modern radiolarian global distributions. Cushman Foundation for Foraminiferal Research, Special Publication, 16A.
- Malinverno E (2010) Extant morphotypes of *Distephanus speculum* (Silicoflagellata) from the Australian sector of the Southern Ocean: Morphology, morphometry and biogeography. *Marine Micropaleontology* 77: 154–174.
- McCartney K (1995) Silicoflagellates. In: Blome CD, Whalen PM, and Reed KM (eds.) *Siliceous Microfossils: Short Courses in Paleontology* Number 8: pp. 159–176. Knoxville: University of Tennessee.
- Molina-Cruz A (1977) The relation of the southern trade winds to upwelling processes during the last 75,000 years. *Quaternary Research* 8(3): 324–338.
- Molina-Cruz A (1984) Radiolaria as indicators of upwelling processes: The Peruvian connection. *Marine Micropaleontology* 9(1): 53–75.
- Moore TC Jr. (1978) The distribution of radiolarian assemblages in the modern and ice-age Pacific. *Marine Micropaleontology* 3(3): 229–266.
- Moore TC, Burckle LH, Geitzenauer K, et al. (1980) The reconstruction of sea surface temperatures in the Pacific Ocean of 18,000 BP (on the basis of all microfossils). *Marine Micropaleontology* 5(3): 215–247.
- Moore TC Jr., Shackleton NJ, and Pisias NG (1995) Paleoceanography and the diachrony of radiolarian events in the eastern Equatorial Pacific. In: Pisias NG, Meyer LA, Janecek TR, Palmer-Judson A, and van Andel TH (eds.) *Proceedings of the Ocean Drilling Program, Scientific Results*, pp. 911–930. College Station, TX: Ocean Drilling Program.
- Morley J (1989) Variations in high-latitude oceanographic fronts in the southern Indian Ocean: An estimation based on faunal changes. *Paleoceanography* 4(5): 547–554.
- Morley JJ and Hays JD (1979) *Cycladophora davisiana*: A stratigraphic tool for Pleistocene North Atlantic and interhemispheric correlation. *Earth and Planetary Science Letters* 44(3): 383–389.
- Morley JJ and Shackleton NL (1978) Extension of the radiolarian *Stylatractus universus* as a biostratigraphic datum to the Atlantic Ocean. *Geology* 6(5): 309–311.
- Nigrini C and Caulet JP (1992) Late Neogene radiolarian assemblages characteristic of Indo-Pacific areas of upwelling. *Micropaleontology* 38(2): 139–164.
- Nimmergut A and Abelmann A (2002) Spatial and seasonal changes of radiolarian standing stocks in the Sea of Okhotsk. *Deep Sea Research Part I: Oceanographic Research Papers* 49(3): 463–493.
- Onodera J and Takahashi K (2005) Silicoflagellate fluxes and environmental variations in the northwestern North Pacific during December 1997–May 2000. *Deep Sea Research Part I: Oceanographic Research Papers* 52: 371–388.
- Perch-Nielsen K (1985) Silicoflagellates. In: Bolli HM, Saunders JB, and Perch-Nielsen K (eds.) *Plankton Stratigraphy*, pp. 811–847. Cambridge: Cambridge University Press.
- Pisias NG, Roelofs A, and Weber M (1997) Radiolarian-based transfer functions for estimating mean surface ocean temperatures and seasonal range. *Paleoceanography* 12(3): 365–379.
- Sanfilippo A, Westberg-Smith MJ, and Riedel WR (1985) Cenozoic radiolaria. In: Bolli HM, Saunders JB, and Perch-Nielsen K (eds.) *Plankton Stratigraphy*, pp. 631–712. Cambridge, UK: Cambridge University Press.
- Weinheimer A (2001) Radiolarians from the Northern Cape Basin, Site 1082. In: Wefer G, Berger WH, and Richter C (eds.) *Proceedings of the Ocean Drilling Program, Science Results*, vol. 175, pp. 1–16. College Station, TX: Ocean Drilling Program.
- WoRMS (2010) Polycystinea. Accessed through: World Register of Marine Species at <http://www.marinespecies.org/aphia.php?p=taxdetails&id=235740> on 2010–11–23.

Relevant Website

www.radiolaria.org – Website for radiolarian specialists.

Temperature Proxies, Census Counts

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Introduction

Because the historical sea-surface temperature (SST) record extends back only to the late 1800s, paleoceanographers must rely on indirect measures of temperature – referred to as temperature proxies – to reconstruct thermal fluctuations during most of the Quaternary and earlier geologic periods. An ideal temperature proxy would instantaneously record unbiased temperature changes in a way that was insensitive to postdepositional change. While all temperature proxies fall short of this ideal, a number of approaches are sufficiently robust to provide quantitative results. Here we discuss quantitative methods that assume that the biogeography of marine microorganisms is controlled either by temperature or a correlated environmental variable such as thermocline structure or nutrient content. The statistical methods described below employ census counts of microfossils of marine fauna and flora and have been used to reconstruct conditions at the surface or in the deep sea. Planktonic Foraminifera are widely distributed in the world's oceans. Temperature reconstructions using the shells of these organisms capture the global range of temperature variability and biogeographic provinces within the ocean (e.g., cool eastern margins, warm gyres, cold polar regions, etc.). In modern usage, four types of multivariate methods are used: coiling ratios of *Neoglobquadrina pachyderma*, transfer functions employing Q-mode factor models, modern analog methods, and neural networks. Coiling ratios of *N. pachyderma* are applied only in the high latitudes where taxonomic diversity of the planktonic Foraminifera tends toward a single morphotype or taxon. The latter methods, which can be applied throughout the world's ocean, are relatively unbiased, and capable of precision that ranges from ± 0.8 to ± 1.5 °C.

Ecological Basis

Theory

Their simple physiology, great abundance, relatively short lifespan, and planktonic habit make microorganisms one of the best choices among marine organisms as the basis for quantitative reconstruction of marine properties such as surface temperature and productivity. While marine microfossils are extremely useful due to their widespread distribution, they cannot provide high resolution geochemical time series with subannual resolution such as those that can be extracted from corals, sclerosponges, or mollusks (see [Oxygen-Isotope Stratigraphy of the Oceans](#)). The distribution of a marine microorganism is controlled by its ecological niche, the sum total of the environmental factors that affect the organism during its lifespan. Organisms reach maximum abundance within their optimal niche zone and decrease in abundance away from this optimum. A common misconception regarding microfossil-based paleotemperature proxy methods is that the fundamental assumption upon which they are based that the distribution

of the organisms studied is controlled solely by temperature. While it is certainly true that temperature plays a very important role in the distribution of marine microorganisms, it is not the sole factor determining their distribution. Rather, factors such as food availability, thermocline structure, nutrient levels, competition, and predation help to shape the ecological success, and thus biogeographic distribution of these organisms. These competing factors place a theoretical lower limit on the precision that can be achieved using these methods.

This truth was not lost on [Imbrie and Kipp \(1971\)](#) who proposed a regression-based approach to reconstructing paleoceanographic temperature change using statistically transformed planktonic foraminiferal assemblages. This transfer function approach has come to be known as the Imbrie–Kipp method (IKM). They recognized that the biogeography of simple microorganisms is controlled by a suite of correlated environmental variables. Because surface temperature in the modern ocean is highly correlated with other temperatures within the euphotic zone and with most other near-surface environmental fields, they reasoned that the reconstruction of sea surface temperatures is well justified. The fundamental assumption behind these statistical methods is thus best stated as follows: the biogeography of simple microorganisms is controlled either by temperature or a correlated environmental variable. As a corollary, the relationship between the organism and its environment must not change during the interval of paleoceanographic reconstruction. The task at hand thus becomes determining a means by which to scale the observed variability in the foraminiferal data to yield robust estimates of paleoceanographic changes in temperature or some other environmental variable. There is, of course, a limit to the number of environmental variables that can conceivably be reconstructed in this way. While temperature is the most commonly reconstructed parameter, paleoproductivity has been reconstructed with success (e.g., [Mix, 1989a,b](#)) because on global scales, temperature and production are uncorrelated. Traditionally, warm and cold season temperatures have been reconstructed (e.g., [CLIMAP, 1976, 1981](#)), with the rationale that different organisms live during different times of the year and thus impart independent information on the fossil record. This is a largely pragmatic decision, as seasonal reconstruction is desirable. This particular approach has been criticized because monthly and seasonal temperatures are so very highly correlated due to the strength of the seasonal cycle. [Mix \(1989a,b\)](#), proposed a novel solution to this problem by reconstructing mean annual temperature and annual temperature range, two environmental variables that are essentially uncorrelated.

Validity of Assumptions

The validity of the fundamental assumptions has been tested in a number of ways. The most common approach is to argue that the

assumptions are valid because a statistically significant relationship can be generated between a single environmental field such as temperature and core-top foraminiferal faunal data. This is not an entirely satisfying argument because the least squares methods which are often applied in these approaches must yield positive, definite results and thus any additional variables added to a statistical model must increase the amount of variance explained. More direct approaches to testing the hypothesis include comparison of statistical methods using different families of microorganisms (e.g., Foraminifera vs. radiolarian) or against geochemical paleotemperature proxy methods. In many cases, the results may disagree primarily because the proxies may be responding to different variables, seasons, or depth habitats. Because of this, perhaps the most stringent test of the fundamental assumptions arises from experimental culture work on planktonic Foraminifera (e.g., [Bijma et al., 1990](#)), and field observations of planktonic foraminifer using plankton tows (e.g., [Bé and Hutson, 1977](#)) or sediment traps (e.g., [Mohiuddin et al., 2005](#); [Thunell and Reynolds, 1984](#)), or studies that combined plankton tow, sediment trap, and core top data ([Wilke et al., 2009](#)). [Ortiz and Mix \(1997\)](#), for example, compared the effectiveness of the IKM against the traditional modern analog method, using both sediment trap and core top calibration data sets. Their work indicated that the modern analog technique (MAT) produced global temperature estimates with smaller residuals and that the sediment trap faunas produced reliable temperature estimates when the core top methods were applied to them.

As an alternative, [Morey et al. \(2005\)](#), using canonical correspondence analysis (CCA), found support for these assumptions by demonstrating that core top foraminiferal assemblages from the Atlantic and Pacific Oceans were correlated with two CCA axes. The primary axis was controlled by surface temperature, while the secondary axis was inversely correlated with temperature and positively correlated with a suite of environmental variables related to productivity. They also found no statistically significant overprint of dissolution recorded in the foraminiferal faunas, suggesting that post-depositional changes may not be as severe a problem as has often been assumed. [Žarić et al. \(2005\)](#) came to similar conclusions regarding the relative importance of temperature and production on the flux and relative abundance of planktonic foraminiferal populations collected from 42 sediment traps distributed through the world's oceans. This generalization may not hold on all regional scales; however, [Siccha et al. \(2009\)](#) found that in the warm waters of the Red Sea, productivity seemed to be the dominant factor controlling glacial-interglacial variations of foraminiferal faunas potentially because the low latitude position of this environment minimizes the range of observed SST variation.

Other studies have explored the impact of seasonality on paleotemperature estimation. [Salguero et al. \(2008\)](#), employing core top faunas from the Iberian Margin, found that the variety of paleotemperature methods discussed here tended to underestimate SST in southern regions with intermittent upwelling and overestimate SST in the north where colder waters are present year round. [Fraile et al. \(2009\)](#) modeled fluxes of foraminiferal species to document how seasonally shifting temperatures can alter the seasonal flux patterns and thus influence foraminiferal faunal paleotemperature estimates.

Approaches

Statistically Significant Census Counts and Percent Abundance Data

A number of approaches have been proposed to estimate paleotemperature using microfossil data. The microfossil shells or tests of organisms that once lived in the water column settle to the sea floor and are incorporated into the sedimentary record in great abundance. Although the distribution of these organisms varies as a function of time, water depth, and location, there may be thousands to tens of thousands of foraminiferal tests per cubic gram of sediment. These microfossils are extracted from the sediment and sorted into taxonomic groups. The number of recent morphotypic foraminiferal taxa recognized by various authors range from around 26 to as many as 45. The discrepancy arises from variations in regional faunal abundance and differences in opinion regarding taxonomic distinctions.

The great abundance of these organisms in the sediment means that only a subsample of the total population is usually analyzed. A basic question that must be addressed then is the number of individual shells that should be analyzed to produce statistically meaningful results. Traditionally, 300–350 individuals from all taxa are counted ([Imbrie and Kipp, 1971](#)), a value which is selected as a balance between accuracy and effort. [Dryden \(1931\)](#), working with mineral grains, presented a probability relationship and figures through which this 'rule' can be understood:

$$p_e(\text{in shells}) = 0.6745 \times \sqrt{(np(1-p))}$$

where p_e is the probable error, n =total number of shells counted, and p =fractional abundance of the taxa in question. While Dryden's work was conducted with mineral grains, it applies equally well to microfossil data. Because the error function follows a square root, p_e must fall off asymptotically as n increases. The turning point in the function lies between 50 and 100 shells. Thus the cost, in terms of the number of shells that must be identified to minimize a given error, grows disproportionately relative to the resulting improvement in accuracy. Using calculations from [Dryden \(1931\)](#), when $p=0.5$, to obtain a 1% accuracy would require counting 4549 total shells, while a 5% accuracy would require only 200! Because the error increases with decreasing p (i.e., as taxa become rare), a value of $n \approx 300$ was selected as a reasonable compromise between accuracy and analytical efficiency. The errors associated with counting this number of total shells range from around 17% error for a taxon with 5% abundance down to about 2.5% error for a taxon with 80% abundance. Increasing the count to around 750 total shells decreases this range only to 11% error for the rare species, and about 1.5% for the abundant species. Therefore, counting as many as 1000 shells per sample represents significant effort for only modest returns.

Morphotypes Versus Genotypes

The relative percentages of the organisms in the fossil assemblage record environmental information that can be extracted through a variety of quantitative means, once they have been sorted into morphotypes or taxa defined based on phenotypic

traits observable in the foraminiferal shell. In general, these morphotypes have specific geographic distributions and are often referred to in the literature as 'species.' Recently, however, this assumption has been tested through the application of genetic techniques to assess the level of genetic variability within the Foraminifera. The first important result to arise from this work is that rates of genetic drift within the Foraminifera are consistent with the molecular clock hypothesis, implying that rates of speciation within the Foraminifera should be sufficiently slow to certify their reliability as climate recorders on evolutionarily short timescales such as the Quaternary. Perhaps more importantly, these studies have enabled the development of genotypically defined foraminiferal species for the first time. The results of this work have been striking. To date, 33 cryptic genetic types have been found in 9 of the 22 sequenced morphotypes that have been studied (Kucera and Darling, 2002). These genetic types have specific geographic ranges. When an estimate of the modern geographic distribution of these cryptic genotypes was incorporated into core top

temperature reconstructions using the range of multivariate methods discussed below, the results improved the predicted temperature estimates by 5–34% (Kucera and Darling, 2002). Unfortunately, we currently cannot apply this approach to the fossil record, as it is as yet not possible to differentiate these genetic morphotypes on the basis of observed phenotypic variation (Kucera and Darling, 2002).

Classes of Microfossil-Based Paleotemperature Proxy Methods

In general, the worker seeks to determine a quantitative relationship between a foraminiferal response and environmental forcing (Figure 1) using a calibration data set based on faunas generally extracted from core tops (e.g., Mix et al., 1999; Prell et al., 1999) and environmental data from an oceanographic atlas (e.g., WOA, 1998). The relationship developed using this training or calibration data set is then applied to downcore data to infer past environmental conditions.

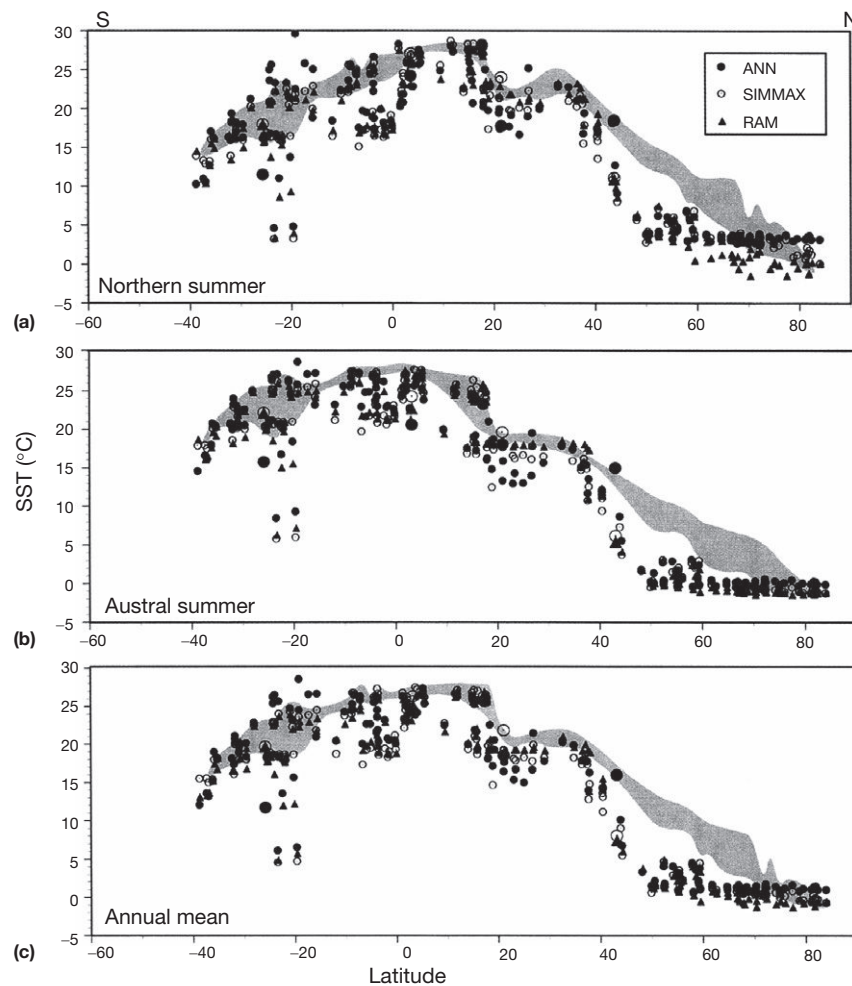


Figure 1 Latitudinal transects of SST estimates in the Atlantic comparing SST reconstructions by artificial neural network (ANN), similarity maximum method (SIMMAX) and revised analog method (RAM). Gray shaded area envelops modern SST values (WOA, 1998) at the same stations, (a) Northern summer; (b) austral summer; (c) annual average. Enlarged symbols mark severe no-analog samples excluded from triangulation. Error estimates for each method average on the order of 1 °C. Reprinted from Kucera M, Weinelt M, Kiefer T, et al. (2005b) Reconstruction of sea-surface temperatures from assemblages of planktonic foraminifera: Multi-technique approach based on geographically constrained calibration data sets and its application to glacial Atlantic and Pacific Oceans. *Quaternary Science Reviews* 24: 951–998.

Table 1 Overview of the MARGO planktonic foraminifer data sets with prediction errors for individual regions

Calibration data set				Root mean square error of prediction (°C)					LGM data set	
Region	Source	Samples	Remarks	SST	ANN	RAM	SIMMAX	MAT	Cores	Samples
North Atlantic	1	862		Summer	1.14	1.09	0.91	1.42	102	550
				Winter	0.96	0.79	0.86	1.32		
				Annual	0.96	0.85	0.81	1.26		
South Atlantic	1	321	83 tropical samples in common with North Atlantic	Summer	0.81	0.76	0.98	1.36	69	198
				Winter	0.95	0.96	0.98	1.47		
				Annual	0.83	0.62	0.93	1.36		
Pacific	1	1111	Some core tops common to IndoPacific data set 13 LGM cores common to IndoPacific data set	Summer	1.19	0.78	1.00	1.63	82	265
				Winter	1.64	1.17	1.27	1.88		
				Annual	1.35	0.93	1.10	1.67		
Mediterranean	2	274	Includes 129 samples from North Atlantic	Summer	1.14	0.74			37	273
				Winter	0.79	0.47				
				Annual	0.91	0.54				
Indian Ocean and Australia (IndoPacific total)	3	1344	Some core tops common to Pacific data set 13 LGM cores common to Pacific data set	Summer	0.93	0.91		0.89	170	301
				Winter	1.01	1.98		1.98		
				Annual	0.86	0.86		0.84		
									460	1574

The error rates have been determined by cross validation of the calibration data sets using ten independent partitions (artificial neural network (ANN) and modern analog technique (MAT) and revised analog method (RAM) for Indian Ocean and Australia) and the leaving-one-out method (similarity maximum method (SIMMAX), RAM, MAT). MAT values for North and South Atlantic and Pacific data sets refer to error rates for SIMMAX without distance weighting. Note that the leaving-one-out procedure in RAM02 software, used for all regions except the Indian Ocean and Australia underestimates the error rate as the samples are 'left out' after the two-dimensional interpolation procedure.

Sources for tabulated results: Kucera et al., (2005b); Hayes et al., (2005); Barrows and Juggins (2005). Reprinted from Kucera M, Weinelt M, Kiefer T, et al. (2005b) Reconstruction of sea-surface temperatures from assemblages of planktonic foraminifera: Multi-technique approach based on geographically constrained calibration data sets and its application to glacial Atlantic and Pacific Oceans. *Quaternary Science Reviews* 24: 951–998.

One approach to this problem is to generate a multiple linear regression between surface temperature and the raw percent abundance of the taxa found in the sediments (e.g., Vincent and Berger, 1981). This approach proved to be less than satisfactory due to instability of the regression coefficients as a result of intercorrelation of the input foraminiferal data. This problem, sometimes referred to as 'multicollinearity' results in an overfitted regression model that may produce plausible regression statistics during the initial modern calibration phase, but which yields unrealistic downcore proxy temperature estimates. That is to say that multiple taxa of Foraminifera may respond to correlated environmental information in one way in the modern environment and in a different way in the past. Therefore, when the interaction between the environmental forcing changes downcore, the regression will no longer yields valid results.

In modern usage, four types of methods are employed: coiling ratios of *N. pachyderma*, transfer functions employing Q-mode factor models (the IKM), modern analog methods, and neural networks. Coiling ratios of *N. pachyderma* are applied only in the high latitudes where taxonomic diversity of the planktonic Foraminifera tends toward a single morphotype or taxon. The latter methods, which can be applied throughout the world's oceans, are relatively unbiased, and capable of precision that ranges from ± 0.8 to ± 1.5 °C (Figure 1, Table 1). At high latitudes, the multivariate methods perform poorly. This is evident in Figure 1 as a systematic bias in

paleotemperature estimates in the North Atlantic. As pointed out in Kucera et al. (2005b), each of these multivariate methods has difficulty reproducing temperatures in cold high-latitude regions where faunal diversity decreases rapidly toward assemblages dominated by a single morphotype, *N. pachyderma* (s.).

Imbrie–Kipp Transfer Function Method

The IKM was proposed as the means of solving the multicollinearity problem (Imbrie and Kipp, 1971). The method employs Q-mode factor analysis to simplify a large matrix of variables measured on many samples. Q-mode factor analysis is often used when many variables are measured at multiple spatial or temporal points. The Q-mode method seeks to preserve the 'information' within the samples of the data set, rather than the variance within the variables. We use the term 'information' rather than variance because each sample represents a set of measurements on many variables that may differ greatly in their variance. Fundamentally, a data matrix ($\mathbf{U}_{ct}$) is decomposed into a factor score matrix ($\mathbf{B}_{ct}$), a factor loading matrix ($\mathbf{F}_{ct}$), and an error matrix ($\mathbf{E}_{ct}$):

$$\mathbf{U}_{ct} = \mathbf{B}_{ct}\mathbf{F}_{ct} + \mathbf{E}_{ct}$$

The factor score matrix represents linear combinations of the original foraminiferal census count data (factors

assemblages) that are combined in such a way that they are independent of each other, or orthogonal. The rows in the factor loading matrix indicate the relative contribution of each factor assemblage to each sample in the data set. This approach effectively deals with the intercorrelation between the input variables, allowing stable regression parameters to be determined from stepwise multiple linear regression of the factor loading matrix against the environmental data. This was the basis of the CLIMAP temperature reconstructions and became a standard procedure for estimation of environmental information from a variety of microfossil groups. [Ortiz and Mix \(1997\)](#) report that this method can produce errors on the order of $\pm 1.9^\circ\text{C}$ when employed with a global calibration data set.

Modern Analog Methods: MAT, SIMMAX, RAM

Until recently, the principle alternative to the IKM was the modern analog method ([Prell, 1985](#)). This approach remains in many ways closer to the original microfossil census data than the IKM. The basic assumptions behind the modern analog methods are that similar foraminiferal faunas are produced by similar environmental conditions, and as in the IKM, that temperature either controls or is correlated with the environmental factors that give rise to the foraminiferal assemblages. The task of the investigator is thus to employ some index with which to compare multidimensional foraminiferal assemblage data to determine which are most similar (or least distinct). A variety of metrics or dissimilarity indices have been proposed and these form the basis for differentiating between the three primary variants of the modern analog method that have developed.

The traditional MAT as applied by [Prell \(1985\)](#) made use of the squared chord distance, defined as:

$$d_{ij} = \sum \left[p_{ik}^{1/2} - p_{jk}^{1/2} \right]^2$$

where the sum is calculated over all taxa recognized, and p_{ik} and p_{jk} represent the percentages of taxa k , in two samples i and j . Samples with unknown environmental conditions are compared against a calibration data set to find the best analogs. Environmental conditions at the locations of the best analogs are then averaged as specified by the user to estimate the paleoclimate conditions under which the assemblage from the unknown sample were likely generated. Advantages of this approach are that in general analysis of the environmental data is independent of the selection of the best analogs. The quality of the analogs obtained from the calibration data set can be assessed using the mean and standard deviation of the dissimilarity index and by the standard deviation of the environmental data obtained from the locations of the best analogs. This final consideration is important as it can be used to assess the uniqueness of the relationship between a particular fauna and the environment. However, a primary consideration in the application of the method is that the results are strongly dependent on the quality of the calibration data set. Larger and more environmentally comprehensive calibration data sets yield more reliable results. [Ortiz and Mix \(1997\)](#) reported that this method can produce errors to the order of $\pm 1.5^\circ\text{C}$ when employed with a global calibration data set consisting of

1121 samples. Two variants of the modern analog method have been developed. These are referred to as the similarity maximum method (SIMMAX) and revised analog method (RAM) methods. Both of these methods exhibit precision to the order of ± 0.8 to 1.0°C , but these improved error estimates may underestimate the true error in the methods for reasons described below ([Table 1](#)).

The SIMMAX method (e.g., [Pflaumann et al., 1996](#)) differs from the traditional MAT method by using the scalar product of the normalized faunal assemblages as the similarity index, and then weighting the contribution of the best analogs obtained by the inverse of their distance from the location where the sample for paleotemperature reconstruction was obtained. This latter condition places a strong constraint on the resulting paleotemperature reconstruction, effectively nudging the solution back toward the modern environmental field. [Kucera et al. \(2005a,b\)](#) report an average SIMMAX error of $\pm 0.98^\circ\text{C}$ and an average MAT error of $\pm 1.3^\circ\text{C}$, but the latter was determined using the SIMMAX method without its distance-weighting function, a standard way of applying the method ([Table 1](#)). [Kucera et al. \(2005a,b\)](#) state that this is identical to the classical MAT method but is not strictly correct since SIMMAX and the traditional MAT of [Prell \(1985\)](#) do not use the same dissimilarity index.

The RAM merges the traditional MAT with the response surface method, which is a standard procedure in palynology ([Waelbroeck et al., 1998](#)). This method seeks to describe the faunal or floral distribution as a simultaneous function of two or more environmental variables (i.e., to map out an environmental response surface). These variables could be warm and cold season temperature or annual temperature and productivity, for example. The first step in the RAM process is to interpolate the foraminiferal and environmental fields to a fixed grid specified by the user. The gridding procedure enables the faunal composition to be estimated for all combinations of the input environmental fields in the data set. This, of course, requires that the data sets be sufficiently well distributed to yield reliable results. Following this step, the traditional MAT method of [Prell \(1985\)](#) is applied to the interpolated fields with the modification that the number of best analogs retained is determined based on the derivative of the calculated dissimilarity coefficients using a user-determined threshold ([Kucera et al., 2005b](#)). The RAM method is capable of errors to the order of $\pm 0.8^\circ\text{C}$, but this error is perhaps an underestimate, since the errors are calculated on the basis of the interpolated fields and not the ungridded data ([Table 1](#)).

Artificial Neural Networks

The implementation of artificial neural networks (ANNs) as a tool to infer paleotemperature using foraminiferal census data ([Malmgren and Nordlund, 1997](#)) represents a significant departure from the statistical approaches employed, and has the potential to be the most significant advance in the field in recent decades. The ANN method differs significantly from previous methods in that sets of interconnected processing units (neurons) explore the relationship between the input census data and the environmental field(s) believed to force the foraminiferal response. The network then uses an

optimization procedure to minimize the errors in prediction of the environmental variable in relation to the observed foraminiferal faunas. The learning process is completed when additional information no longer minimizes the prediction error. A subset of samples is reserved for validation of the resulting network. The advantages of this approach are that properly trained ANNs are less dependent on the size of the calibration data set than other methods such as the MAT, and the method is capable of extrapolating to conditions outside of the initial environmental data set (Kucera et al., 2005b). The primary disadvantages of the approach are that the ANN solution obtained from a pairing of foraminiferal census data and environmental data is not unique, the method is computationally expensive, and it can be difficult to determine the underlying relationship between the foraminiferal census data and the environmental forcing. The lack of a unique solution is addressed by averaging the results from multiple ANN solutions. Considerable caution must be exercised when establishing the initial conditions under which the method is applied (e.g., number of neurons employed, fraction of data held in reserve, and number of networks developed for comparison). Despite these potential difficulties, Kucera et al., 2005b demonstrate that the method is capable of producing results with an average precision of $\pm 1.0^\circ\text{C}$ (Table 1).

Selected Case Studies

Planktonic foraminiferal faunal temperature estimates were initially developed for use with late Quaternary faunas. Over the span of several thousand years, it was argued that the evolution of these taxa – and thus potential changes in their relationship to the environment due to genetic drift – were minimal. Qualitative methods, such as changes in the relative proportion of tropical versus subtropical taxa, or quantitative application of transfer function methods are now being applied over a much broader range of geologic time. At one extreme, workers are applying these methods in environments with extremely high sedimentation rates to study variation on multidecadal to centennial timescales. In contrast, others are employing these approaches to infer changes in paleotemperature millions of years ago. Staines-Urías et al. (2009) used a combination of foraminiferal faunas and geochemical proxies to study multidecadal and centennial variations in SST and thermocline structure in the Gulf of California. Kandiano and Bauch (2007) argued that marine isotope stage (MIS) 11 may not serve as a good analog for future Holocene climate because faunal, geochemical, and lithologic data suggested that the early evolution of water masses during MIS 11 differed significantly from that of the early Holocene. Schaefer et al. (2005) studied the evolution of the Subtropical Front east of New Zealand over the past 1 Ma. Sato et al. (2008) compared variations in planktonic foraminiferal faunas with oxygen isotopes and Mg/Ca derived from planktonic Foraminifera to determine that the modern Western Pacific Warm Pool was established during the Pliocene by 3.6 Ma, in response to the constriction of the Indonesian seaway and the closure of the Central American seaway. Perhaps the most ambitious effort to expand the use of planktonic foraminiferal faunal SST estimates is the Pliocene Research, Interpretation and Synoptic Mapping

Project, which is combining planktonic foraminiferal faunal estimates with alkenone and Mg/Ca temperature estimates to construct boundary conditions to force coupled ocean atmosphere general circulation models to simulate ocean circulation during the Pliocene, a potential analog for future warming (Dowsett and Robinson, 2009; Dowsett et al., 2009). Their results suggest that the pronounced warmth of the Pliocene was driven by a combination of enhanced greenhouse forcing and meridional heat transport (Dowsett et al., 2010).

Controversies

Microfossil-based paleotemperature proxy methods have been employed in numerous studies since their inception, largely as part of the CLIMAP project. The results of these studies have significantly advanced our understanding of climate change on a variety of timescales, but not without controversy. Perhaps the two most significant problems associated with this class of methods are the generic no-analog problem, and the specific warm Last Glacial Maximum (LGM) tropics problem.

All of the methods described above are sensitive – to some degree – to the no-analog problem. This is effectively a violation of the assumption that the response of the foraminiferal faunas to climate is invariant between the calibration faunas and the downcore faunas of interest. In extreme cases, no-analog conditions can result from excessive environmental stress (e.g., mass extinction events), which can modify foraminiferal assemblages and morphology (Keller and Abramovich, 2009). The no-analog problem is generally manifest through poor goodness-of-fit statistics, such as communality with respect to the IKM, or large dissimilarity functions with respect to the modern analog methods. Because it postulates a specific statistical relationship between the foraminiferal faunas and the environmental response, the IKM is more sensitive to no-analog problems than the various modern analog approaches, provided that they employ sufficiently large calibration data sets.

Perhaps the greatest controversy associated with microfossil-based paleotemperature proxy methods is the warm glacial tropics problem. A surprising result of the CLIMAP reconstruction was the prediction that the tropics either did not cool significantly or may have warmed to the order of 1°C in some regions. These results did not agree with theoretical estimates, modeling studies, or geochemical estimates that predicted changes as great as $5\text{--}6^\circ\text{C}$ in some places. This seemingly implausible result has spawned numerous studies using a variety of techniques. Revised geochemical work now infers tropical glacial cooling to the order of 3°C with considerable spatial heterogeneity.

Mix et al. (1999) presented another interesting approach toward dealing with the no-analog problem in the Atlantic basin. Rather than using a core top calibration data set, they generated downcore factors, then projected these factors onto core top faunas, which formed the basis for an Imbrie–Kipp paleotemperature calibration. When the paleotemperature calibration was used to scale the observed downcore factor pattern, significant cooling, to the order of $2\text{--}3^\circ\text{C}$ was observed, particularly in eastern boundary current regions. Similarly, Barrows and Juggins (2005) using the MAT, RAM, and ANN methods observed glacial cooling to the order of 4°C in the

Indian Ocean, while Kucera et al. (2005b) observed glacial cooling of $>3^{\circ}\text{C}$ in the eastern equatorial Pacific and eastern boundary currents using the ANN, RAM, and SIMMAX with and without distance scaling. Chen et al. (2005) employed the IKM, MAT, SIMMAX, RAM, and ANN methods in a study of glacial temperatures in the western Pacific. Their results indicate that while the MAT, SIMMAX, RAM, and ANN methods all predicted temperatures $\sim 1^{\circ}\text{C}$ cooler, the IKM method (which performed poorly in the calibration process) inferred no glacial temperature change relative to modern. Steinke et al. (2008) compared five different faunal transfer function methods and found nearly unchanged or unusually high paleotemperatures in the South China Sea during the Last Glacial Maximum. These results were in contrast to the $2\text{--}5^{\circ}\text{C}$ cooling observed with U^{k}_{37} and Mg/Ca geochemical methods. They concluded that the anomalously warm paleotemperatures derived from the foraminiferal faunal methods were the result of a no-analog fauna in the glacial interval, which exhibited abundances of *N. pachyderma* (dex.) and *P. obiquiloculata* greater than their abundance in the core top calibration data set. Li et al. (2009) proposed use of a morphometric paleotemperature index based on the test size of *O. universa* in the South China Sea as a means of addressing this particular no-analog faunal problem.

While sampling issues may have contributed to the warm glacial tropics problem, the majority of the thermal discrepancy appears to have arisen from a no-analog problem and the use of the traditional IKM, which is most sensitive to this problem. One approach toward addressing this problem is to base results on the comparison of a variety of methods when generating microfossil-based paleotemperature estimates (e.g., Chen et al., 2005; Hayes et al., 2005; Kucera et al., 2005a,b; Nyberg et al., 2002). This approach is consistent with the ‘no free lunch’ theorem from the field of machine learning, which states that no single statistical method can provide optimal classification or prediction results with all data types, although some methods may perform better with specific data sets than others. A promising study along these lines was provided by Assareh et al. (2010). They compared a series of machine-learning algorithms to evaluate their relative potential in paleotemperature estimation. They employed a Monte Carlo method consisting of a genetic algorithm (GA) that compared the performance of 1000 different classifier teams in four classes (IKM, MAT, regression trees, and ANNs) over 500 randomized iterations. The GA randomized the input parameters of the predictors in the classifier team, retaining the best classifiers and discarding the poorly performing classifiers at each iteration step. In this way, the classifier team ‘evolves’ over the course of the Monte Carlo iterations. The resulting paleotemperature classifier team produced errors to the order of $\pm 0.88^{\circ}\text{C}$ and was dominated by the ANN team members.

See also: Paleoclimate Relevance to Global Warming.
Paleoceanography, Biological Proxies: Biomarker Indicators of Past Climate; Coccolithophores; Dinoflagellates; Marine Diatoms; Planktic Foraminifera; Radiolarians and Silicoflagellates.
Paleoceanography, Physical and Chemical Proxies: Mg/Ca and Sr/Ca Paleothermometry from Calcareous Marine Fossils.
Paleoclimate Reconstruction: Pliocene Environments.

References

- Assareh A, Volkert LG, and Ortiz JD (2010) Evolutionary selection of regression predictors to enhance the performance of microfossil-based paleotemperature proxies. *Ninth International Conference on Machine Learning and Applications (ICMLA, 2010)*, pp. 379–385.
- Barrows TT and Juggins S (2005) Sea-surface temperatures around the Australian margin and Indian Ocean during the Last Glacial Maximum. *Quaternary Science Reviews* 24: 1017–1047.
- Bé A and Hutson WH (1977) Ecology of planktonic foraminifera and biogeographic patterns of life and fossil assemblages in the Indian Ocean. *Micropaleontology* 23: 369–414.
- Bijma J, Faber WW Jr., and Hemleben C (1990) Temperature and salinity limits for growth and survival of planktonic foraminifera in laboratory cultures. *Journal of Foraminiferal Research* 20: 128–148.
- Chen MT, Huang C-C, Pflaumann U, Waelbroeck C, and Kucera M (2005) Estimating glacial western Pacific sea-surface temperature: Methodological overview and data compilations of surface sediment planktonic foraminifer faunas. *Quaternary Science Reviews* 24: 1049–1062.
- CLIMAP Project Members (1981) Seasonal reconstructions of the Earth surface at the last glacial maximum. *Map Chart Series* No. 36: pp. 1–18. Boulder, CO: Geological Society of America.
- Climate: Long Range Investigations, Prediction and Mapping. (CLIMAP) Project Members (1976) The surface of the ice age Earth. *Science* 191: 1131–1137.
- Dowsett H and Robinson M (2009) Mid-Pliocene equatorial Pacific sea surface temperature reconstruction: A multiproxy perspective. *Philosophical Transactions of the Royal Society A* 367: 109–125.
- Dowsett H, Robinson M, and Foley K (2009) Pliocene three-dimensional global ocean temperature reconstruction. *Climate of the Past* 5: 769–783.
- Dowsett H, Robinson M, Stoll D, and Foley K (2010) Mid-Piacenzian mean annual sea surface temperature analysis for data-model comparisons. *Stratigraphy* 7: 189–198.
- Dryden AL (1931) Accuracy in percentage representation of heavy mineral frequencies. *Proceedings of the National Academy of Sciences of the United States of America* 17: 233–238.
- Fraile I, Multiza S, and Schultz M (2009) Modeling planktonic foraminiferal seasonality: Implications for sea-surface temperature reconstructions. *Marine Micropaleontology* 72: 1–9.
- Hayes A, Kucera M, Kallel N, Saffi L, and Rohling EJ (2005) Glacial Mediterranean sea surface temperatures based on planktonic foraminiferal assemblages. *Quaternary Science Reviews* 24: 999–1016.
- Imbrie J and Kipp N (1971) A new micropaleontological method for quantitative paleoclimatology: Application to a Late Pleistocene Caribbean core. In: Turekian K (ed.) *The Late Cenozoic Glacial Ages*, pp. 71–181. New Haven, CT: Yale University Press.
- Kandiano E and Bauch H (2007) Phase relationship and surface water mass change in the Northeast Atlantic during Marine Isotope stage 11 (MIS 11). *Quaternary Research* 68: 445–455.
- Keller G and Abramovich S (2009) Lilliput effect in late Maastrichtian planktonic foraminifera: Response to environmental stress. *Palaeogeography, Palaeoclimatology, Palaeoecology* 284: 47–62.
- Kucera M and Darling KF (2002) Cryptic species of planktonic foraminifera: Their effect on paleoceanographic reconstruction. *Philosophical Transactions of the Royal Society of London A* 360: 695–718.
- Kucera M, Roselle-Mele A, Schneider R, Waelbroeck C, and Weinelt M (2005a) Multiproxy approach for reconstruction of the Glacial Surface Ocean (MARGO). *Quaternary Science Reviews* 24: 813–819.
- Kucera M, Weinelt M, Kiefer T, et al. (2005b) Reconstruction of sea-surface temperatures from assemblages of planktonic foraminifera: Multi-technique approach based on geographically constrained calibration data sets and its application to glacial Atlantic and Pacific Oceans. *Quaternary Science Reviews* 24: 951–998.
- Li B-H, Wang X-Y, Jian Z-M, and Wang P-X (2009) Sea surface environment inferred from planktonic foraminifera in the southern South China Sea since the last glacial period. *Paleoworld* 18: 23–33.
- Luciani V, Giusberti L, Claudia Agnini, Beckman J, Fornaciari E, and Rio D (2007) Paleocene–Eocene thermal maximum as recorded by Tethyan planktonic foraminifera in the Forada section (northern Italy). *Marine Micropaleontology* 64: 189–214.
- Malmgren B and Nordlund U (1997) Application of artificial neural networks to paleoceanographic data. *Palaeogeography, Palaeoclimatology, Palaeoecology* 136: 359–373.
- Mix AC (1989a) Pleistocene paleoproductivity: Evidence from organic carbon and foraminiferal species. In: Berger WH, Smetacek VS, and Wever G (eds.) *Productivity of the Oceans: Past and Present*, pp. 313–340. New York: Wiley.

- Mix AC (1989b) Influence of productivity variations on long-term atmospheric CO₂. *Nature* 337: 541–544.
- Mix AC, Morey AE, Pisias NG, and Hostetler S (1999) Foraminiferal faunal estimates of paleotemperatures: Circumventing the no-analog problems yields cool ice-age tropics. *Paleoceanography* 14: 350–359.
- Mohiuddin M, Nishimura A, and Tanaka Y (2005) Seasonal succession, vertical distribution, and dissolution of planktonic foraminifera along the subarctic front: Implications for paleoceanographic reconstruction in the northwestern Pacific. *Marine Micropaleontology* 55: 129–156.
- Morey A, Mix A, and Pisias N (2005) Planktonic foraminiferal assemblages preserved in surface sediments correspond to multiple environmental variables. *Quaternary Science Reviews* 24: 925–950.
- Nyberg J, Malmgren BA, Kuijpers A, and Winter A (2002) A centennial-scale variability of tropical North Atlantic surface hydrography during the late Holocene. *Palaeogeography, Palaeoclimatology, Palaeoecology* 183: 25–41.
- Ortiz JD and Mix AL (1997) Comparison of Imbrie–Kipp transfer function and modern analog temperature estimates using sediment trap and core top foraminiferal faunas. *Paleoceanography* 12(2): 175–190.
- Pflaumann U, Duprat J, Pujol C, and Labeyrie LD (1996) SIMMAX: A modern analog technique to deduce Atlantic sea surface temperatures from planktonic foraminifera in deep-sea sediments. *Paleoceanography* 11: 15–35.
- Pflaumann U and Jian Z (1999) Modern distribution patterns of planktonic foraminifera in the South China Sea and the western Pacific: A new transfer technique to estimate regional sea-surface temperatures. *Marine Geology* 156: 41–83.
- Prell WL (1985) The stability of low-latitude sea-surface temperatures: An evaluation of the CLIMAP reconstruction with emphasis on the positive SST anomalies. *Report TRO25*, United States Department of Energy, Washington, DC, 60 p.
- Prell WL, Martin A, Cullen J, and Trend M (1999) The Brown University Foraminiferal Data Base. IGBP PAGES/World Data Center-A for Paleoclimatology Data Contribution Series # 1999-027. Boulder, CO: NOAA/NGDC Paleoclimatology Program.
- Salguero E, Voelker A, Abrantes F, et al. (2008) Planktonic foraminifera from modern sediments reflect upwelling patterns off Iberia: Insights from a regional transfer function. *Marine Micropaleontology* 66: 135–164.
- Sato K, Oda M, Chiyonobu S, Kimoto K, and Domitsu H (2008) Establishment of the western Pacific warm pool during the Pliocene: Evidence from planktonic foraminifera, oxygen isotopes, and Mg/Ca ratios. *Palaeogeography, Palaeoclimatology, Palaeoecology* 265: 140–147.
- Schaefer G, Rodger J, Hayward B, Kennet J, Sabaa A, and Scott G (2005) Planktonic foraminiferal and sea surface temperature record during the last 1 My across the subtropical front, Southwest Pacific. *Marine Micropaleontology* 54: 191–212.
- Siccha M, Trommer G, Schulz H, Hemleben C, and Kucera M (2009) Factors controlling the distribution of planktonic foraminifera in the Red Sea and implications for the development of transfer functions. *Marine Micropaleontology* 72: 146–156.
- Staines-Urías F, Douglas R, and Gorsline D (2009) Oceanographic variability in the southern Gulf of California over the past 400 years: Evidence from faunal and isotopic records from planktonic foraminifera. *Palaeogeography, Palaeoclimatology, Palaeoecology* 284: 337–354.
- Steinke S, Yu P-S, Kucera M, and Chen MT (2008) No-analog planktonic foraminiferal faunas in the glacial southern South China Sea: Implications for the magnitude of the glacial cooling in the western Pacific warm pool. *Mar. Micro* 66: 71–90.
- Thunell RC and Reynolds LA (1984) Sedimentation of planktonic foraminifera: Seasonal changes in species flux in the Panama Basin. *Micropaleontology* 30: 243–262.
- Vincent E and Berger W (1981) Planktonic foraminifera and their use in paleoceanography. In: Emiliani C (ed.) *The Oceanic Lithosphere*, vol. 7. New York: Wiley.
- Waelbroeck C, Labeyrie L, Duplessy J-C, et al. (1998) Improving past sea surface temperature estimates based on planktonic fossil faunas. *Paleoceanography* 13: 272–283.
- Wilke I, Meggers H, and Bickert T (2009) Depth habitats and seasonal distribution of recent planktonic foraminifera in the Canary Islands region (20° N) based on oxygen isotopes. *Deep-Sea Research I* 56: 89–106.
- World Ocean Atlas (1998) Version 2, <http://www.nodc.noaa.gov/oc5/woa98.html>, Technical Report, National Oceanographic Data Center, Silver Spring, MD.
- Žarić S, Donner B, Fisher G, Mulitza S, and Wefer G (2005) Sensitivity of planktonic foraminifera to sea surface temperature and export production as derived from sediment trap data. *Marine Micropaleontology* 55: 75–105.

PALEOCEANOGRAPHY, PHYSICAL AND CHEMICAL PROXIES

Contents

Carbon Cycle Proxies ($\delta^{11}\text{B}$, $\delta^{13}\text{C}_{\text{calcite}}$, $\delta^{13}\text{C}_{\text{organic}}$, Shell Weights, B/Ca, U/Ca, Zn/Ca, Ba/Ca)

Dissolution of Deep-Sea Carbonates

Mg/Ca and Sr/Ca Paleothermometry from Calcareous Marine Fossils

Magnetic Proxies and Susceptibility

Nutrient Proxies

Oxygen-Isotope Stratigraphy of the Oceans

Oxygen Isotope Composition of Seawater

Radioisotope Proxies

Salinity Proxies $\delta^{18}\text{O}$

Terrigenous Sediments

Carbon Cycle Proxies ($\delta^{11}\text{B}$, $\delta^{13}\text{C}_{\text{calcite}}$, $\delta^{13}\text{C}_{\text{organic}}$, Shell Weights, B/Ca, U/Ca, Zn/Ca, Ba/Ca)

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Introduction

Observations from Antarctic ice cores have revealed a tight correlation between temperature and atmospheric pCO_2 over the past 800 ka, with glacial periods displaying $\sim 30\%$ lower pCO_2 compared to interglacials. Although the temperature– pCO_2 covariation strongly suggests CO_2 may be an important amplifier of climate change, ice cores do not reveal the causal mechanisms of the glacial–interglacial CO_2 variations.

The ocean is the only reservoir capable of absorbing the amount of carbon needed to explain the lowering of atmospheric pCO_2 from interglacial to glacial levels. Due to the larger land ice extent during glacial periods, the terrestrial biosphere must have been smaller and thus acted as a source of CO_2 to the atmosphere, rather than a sink. Because CO_2 does not only physically dissolve in seawater but chemically reacts with it to form carbonic acid and bicarbonate and carbonate ions, the ocean contains ~ 50 times more carbon than the atmosphere. The 30% decrease in glacial atmospheric pCO_2 and terrestrial biosphere reduction therefore translate into a mere $1\text{--}2\%$ increase in the marine carbon inventory, demonstrating how a small change in the ocean inventory can cause dramatic changes in the atmospheric inventory. Knowledge of the origin and amplitude of past fluxes between carbon reservoirs is therefore important for evaluating the role of CO_2 in climate forcing.

Marine carbonate chemistry can be described by six parameters: aqueous PCO_2 , HCO_3^- , and CO_3^{2-} concentrations, the sum of all dissolved carbon species in the ocean

($\text{DIC} = [\text{CO}_2] + [\text{HCO}_3^-] + [\text{CO}_3^{2-}]$), alkalinity, and seawater pH. Because these parameters are governed by equilibrium reactions, knowledge of any two of these parameters, in addition to temperature, salinity, and pressure, allows for calculation of the entire marine carbonate system. In contrast to inclusions of ancient air in glacier ice, an archive for pristine seawater does not exist for the past and carbonate parameters cannot be measured directly. Paleocceanographers therefore use proxies, that is, measurable quantities that hold a close relationship to the desired parameter but which are preserved in the geological archive. These are often chemical and isotopic compositions of the fossil remains of organisms, as well as physical descriptors such as grain sizes, shell weights, and microstructures. Water pressure can be approximated by the preferred depth habitat of the organism in consideration; we focus here on proxies describing specific parameters of the marine carbonate system. Tools available to date are the boron isotopic composition ($\delta^{11}\text{B}$) of marine carbonates as a proxy for seawater pH, trace metal to calcium ratios of B, U, Cd, Zn, and Ba, as well as foraminiferal shell weights and stable carbon isotopes ($\delta^{13}\text{C}$) for reconstructing the amount of dissolved carbon, and $\delta^{13}\text{C}$ of alkenones as a proxy for aqueous PCO_2 .

Carbon Cycle Proxies

The Boron Isotope Proxy for Seawater pH

The basis for the boron isotope–pH proxy has been described in detail by [Hemming and Hanson \(1992\)](#). Briefly, there are

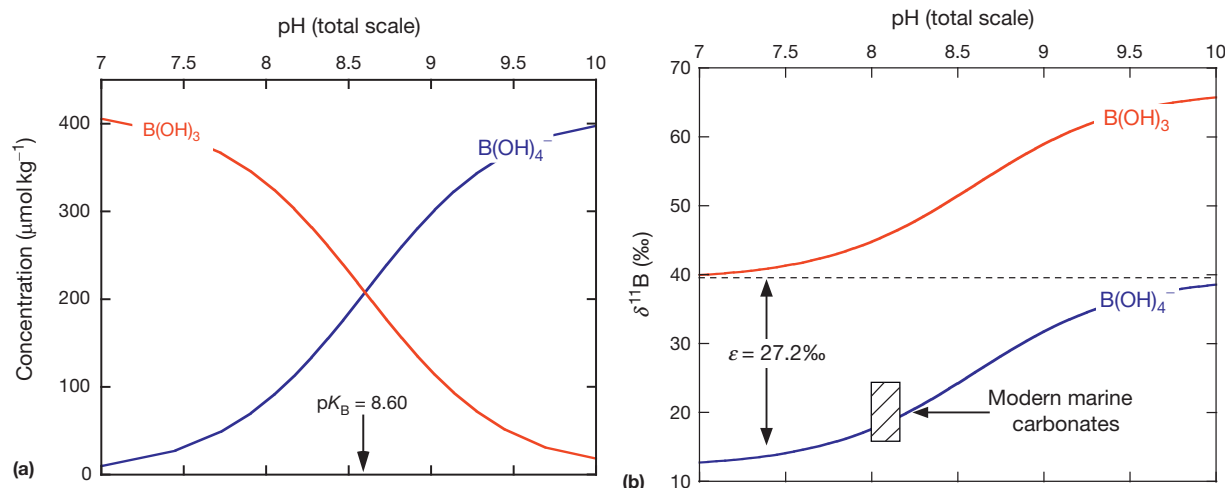


Figure 1 (a) Distribution of the two dominant boron species dissolved in seawater, boric acid ($B(OH)_3$) and borate ($B(OH)_4^-$), versus pH. The equilibrium constant, pK_B , was determined by Dickson (1990). (b) Boron isotopic composition of aqueous boron species versus seawater pH, calculated from the fractionation factor of Klochko et al. (2006). The dashed line indicates $\delta^{11}B$ of seawater, which is constant at $\sim 39.6\text{‰}$. Modern marine carbonates record an isotopic composition that falls close to the isotopic composition of dissolved borate in seawater (Hemming and Hanson, 1992).

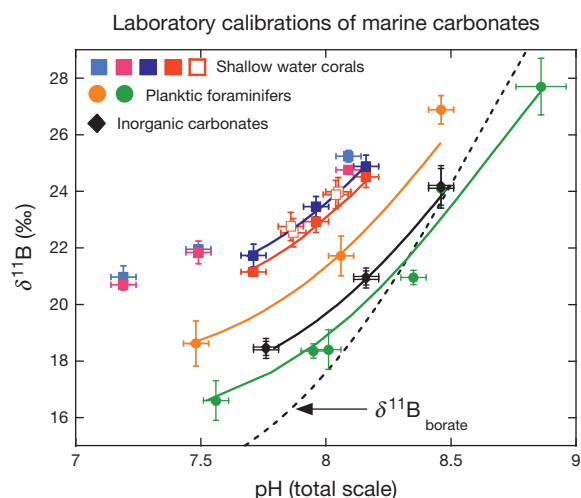


Figure 2 Empirical relationships between $\delta^{11}B$ and experimental seawater pH measured in five species of corals, two species of planktic foraminifers, and inorganically precipitated calcite. Although all carbonate calibration curves fall close to the isotopic composition of dissolved borate in seawater (dashed black line), there are significant offsets between them and they consistently describe a shallower slope than suggested by the aqueous fractionation. Paleoreconstructions therefore need to be restricted to monospecific carbonates that are calibrated for the $\delta^{11}B$ /pH relationship.

two dominant species of dissolved boron in the ocean, boric acid [$B(OH)_3$] and borate [$B(OH)_4^-$], and their relative concentration changes with pH. At low pH (<7) boron exists in the form of $B(OH)_3$, whereas at high pH (>10) boron occurs predominately in the form of $B(OH)_4^-$ (Figure 1(a)). There are also two isotopes of boron, ^{10}B and ^{11}B , and because of differences in vibrational energy and molecular geometry, $B(OH)_3$ is enriched in ^{11}B relative to $B(OH)_4^-$, with a measured isotope fractionation of $\sim 27\text{‰}$ between the two boron species

(Klochko et al., 2006). Consequently, as the relative concentration of the dissolved species changes with pH, so does their isotopic composition (Figure 1(b)).

$\delta^{11}B$ of planktic foraminifers, corals, and inorganically precipitated $CaCO_3$ fall close to the boron isotopic composition of $B(OH)_4^-$ over a wide range of seawater pH (Figure 2). However, all empirical calibration curves describe a $\delta^{11}B$ -pH relationship with a shallower slope than the one measured for aqueous boron isotope fractionation. To reflect the difference between the larger aqueous fractionation of $\epsilon \sim 27\text{‰}$ and the smaller value that best describes the carbonates, we distinguish the use of an empirical value for the boron isotope fractionation into marine carbonates as $\epsilon^* \sim 20\text{‰}$. To translate the $\delta^{11}B$ of fossil carbonates into paleo-pH values, the following equation can be used:

$$pH = pK_B - \log \left(\frac{-(\delta^{11}B_{\text{seawater}} - \delta^{11}B_{\text{carbonate}} - a)}{(\delta^{11}B_{\text{seawater}} - \alpha^* \times (\delta^{11}B_{\text{carbonate}} + a) - \epsilon^*)} \right)$$

where pK_B is the equilibrium constant for the boric acid/borate system, $\delta^{11}B_{\text{seawater}}$ in the modern ocean is 39.6‰ , fractionation $\epsilon^* = (\text{carbonate fractionation factor } \alpha^* - 1) \times 1000$, and a is the constant offset between the apparent aqueous borate and empirical $\delta^{11}B$ -pH-calibration curve for a specific calcium carbonate. Because offsets between different calcifiers are significant (Figure 2), it is important to restrict paleoreconstructions to single species that are calibrated for the $\delta^{11}B$ /pH relationship.

It has also been shown that physiological processes such as respiration, calcification, and symbiont photosynthesis change pH in the microenvironment of a calcifying organism. Culture experiments with living, symbiont-bearing foraminifers have revealed a higher $\delta^{11}B$ of specimens grown in the light relative to individuals grown in the dark (Hönisch et al., 2003). Because the empirical carbonate calibrations shown in Figure 2 are largely parallel to each other despite the different nature of calibrated organisms and their respective calcification schemes,

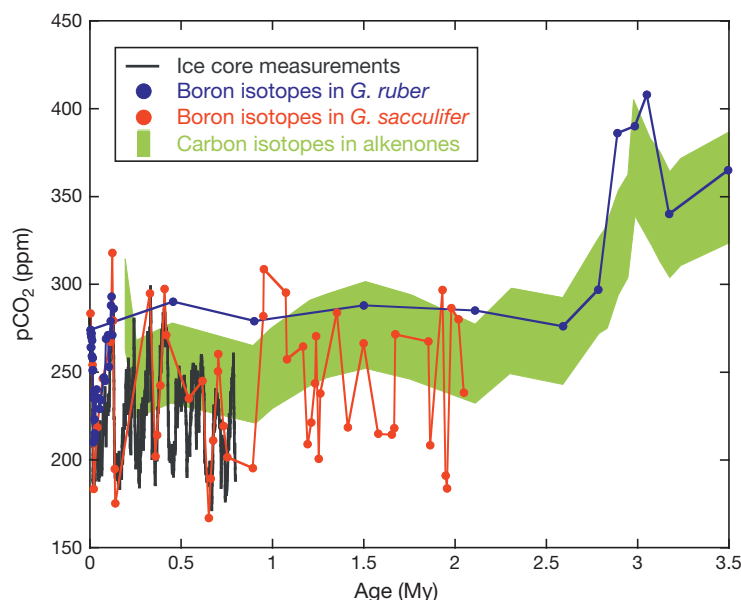


Figure 3 Comparison of atmospheric $p\text{CO}_2$ measured in ice cores (gray line) and surface ocean PCO_2 reconstructed from boron isotopes in shells of the planktic foraminifers *G. ruber* (Seki et al., 2010) and *G. sacculifer* (Hönisch et al., 2009) and ϵ_p in alkenones (Seki et al., 2010). The width of the PCO_2 band from the alkenone reconstruction indicates the PCO_2 -uncertainty based on assumed high (lower boundary) and very high (upper boundary) phosphate availability to alkenone-producing coccolithophores. The average PCO_2 uncertainty of boron isotope reconstructions for the Pleistocene is $\sim 20 \mu\text{atm}$. PCO_2 values from oligotrophic ocean regions, where modern aqueous PCO_2 is in equilibrium with atmospheric $p\text{CO}_2$, generally match atmospheric $p\text{CO}_2$ within the analytical uncertainty of the reconstruction. However, alkenone PCO_2 estimates heavily rely on nutrient estimates, and reconstructions using the same global calibration do not always match atmospheric $p\text{CO}_2$ (cf. Figure 9). Sample selection for boron isotope and alkenone reconstruction by Seki et al. (2010) is believed to be biased toward interglacial periods.

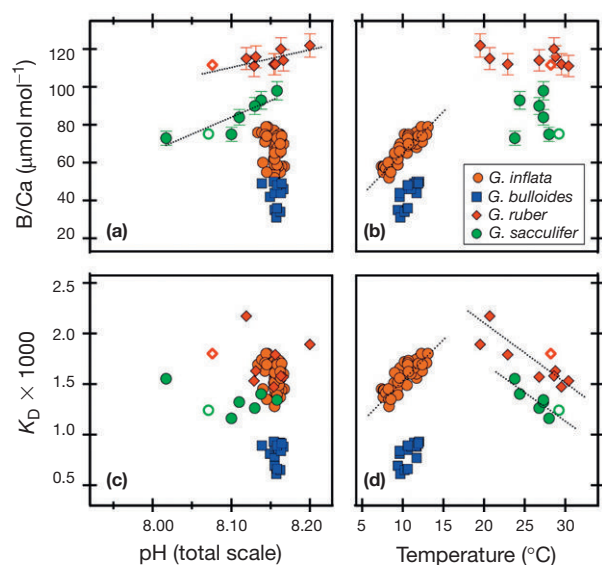


Figure 4 In some species of planktic foraminifers, B/Ca exhibits relationships with pH and/or temperature (a and b). The empirical partition coefficient, K_D , also varies with temperature (d), though the relationship is not consistent across species. The K_D -temperature correlation is positive for *G. inflata* and *G. bulloides*, and negative for *G. ruber* and *G. sacculifer*. It is not yet clear whether this indicates a species-specific effect or whether different relationships are observed because of the influence of an unidentified environmental control. All data shown are derived from modern marine sediments. B/Ca from cultured *O. universa* increases with pH (a), but does not exhibit a strong relationship with temperature (b).

it appears that vital effects due to physiological processes are constant over a wide range of seawater pH and thus do not generally compromise the use of the boron isotope proxy. However, analyses of different shell size classes of the symbiont-bearing foraminifer *Globigerinoides sacculifer* have revealed an increase of $\delta^{11}\text{B}$ with shell size (Hönisch and Hemming, 2004; Ni et al., 2007). This pattern could be explained by larger individuals spending relatively more time closer to the sea surface, where light levels are higher and the effect of symbiont photosynthesis on $\delta^{11}\text{B}$ is larger (Hönisch and Hemming, 2004). Because $\delta^{11}\text{B}$ is also affected by dissolution and the dissolution effect is smallest in the largest shells, it is preferable to use large shells in a narrow size range for paleoreconstructions (Hönisch and Hemming, 2004).

Despite these potential complications, Hönisch and Hemming (2005) and Foster (2008) have demonstrated that quantitative reconstruction of ancient seawater pH and atmospheric $p\text{CO}_2$ is feasible if sample selection and data analysis are done carefully (Figure 3). However, beyond the past 3–5 million years, paleoreconstructions are hampered by uncertainties regarding $\delta^{11}\text{B}_{\text{seawater}}$. Until paleoceanographers find a way to better constrain the value of $\delta^{11}\text{B}_{\text{seawater}}$ through time, reconstructions of the distant past are restricted to relative changes across a climate event rather than absolute pH values.

The B/Ca Proxy for Surface Seawater pH and Bottom Water Carbonate Saturation

The ratio of boron to calcium in foraminiferal calcite (B/Ca) is influenced by seawater carbonate chemistry, with a theoretical

basis similar to that of boron isotopes. Borate ($\text{B}(\text{OH})_4^-$) is more abundant in seawater at higher pH, and if $\text{B}(\text{OH})_4^-$ is the species predominantly incorporated into calcite (as boron isotopes suggest), then B/Ca should increase with pH. Planktic foraminifers from coretop sediments (Foster, 2008; Yu et al., 2007) do indeed exhibit B/Ca relationships with environmental parameters such as pH and temperature (Figure 4(a) and 4(b)). However, because many parameters tend to covary in surface seawater (e.g., temperature, salinity, alkalinity), it has been difficult to distinguish specific control(s) on B/Ca from coretop samples.

To isolate the influence of single variables on B/Ca in planktic foraminifers, culture experiments have been performed with *Orbulina universa* under tightly controlled laboratory conditions (Allen et al., 2011). In these experiments, B/Ca increases with pH, salinity, and seawater boron concentration, but not with temperature. Because raising pH of seawater concurrently affects $[\text{CO}_3^{2-}]$, $[\text{HCO}_3^-]$, and alkalinity, the *O. universa* data cannot be used to discern the

direct influence of these different parameters on B/Ca. Efforts to tease apart and quantify their respective influences on B/Ca are currently in progress. The effect of salinity ($2.6 \mu\text{mol mol}^{-1}$ per salinity unit in *O. universa*) does not pose a large complication for glacial–interglacial timescale reconstructions when salinity changes are small, but will need to be considered for larger changes such as occurred earlier in Earth history.

To constrain the carbonate system in past surface seawater, some researchers have applied calibrations with an empirical boron partition coefficient (K_D , defined as $[\text{B/Ca}]_{\text{carbonate}}/[\text{B}(\text{OH})_4^-/\text{HCO}_3^-]_{\text{seawater}}$). Determination of K_D in this manner is based on the assumption that $\text{B}(\text{OH})_4^-$ and HCO_3^- are the dominant aqueous species incorporated into calcite, and that $\text{B}(\text{OH})_4^-$ substitutes into the CO_3^{2-} lattice site. Modern K_D calibrations have been established by combining coretop B/Ca with modern surface-ocean hydrographic data (Figure 4(d); Foster, 2008; Yu et al., 2007). However, paleoreconstructions utilizing K_D calibrations must be treated with caution until the carbonate system controls on B/Ca are better understood and quantified. We do not know whether borate and bicarbonate are truly the only species relevant to B incorporation, so K_D may not accurately represent B partitioning into calcite. In addition, temperature effects are not consistent between studies and are not observed in culture experiments (Figures 3 and 4(c)), so temperature-dependent K_D calibrations may need to be reconsidered.

In benthic foraminifers, B/Ca does not appear to be directly controlled by seawater pH, but instead increases with carbonate saturation or ΔCO_3^{2-} . Both the absolute B/Ca value and the slope of the B/Ca– ΔCO_3^{2-} relationship vary by species

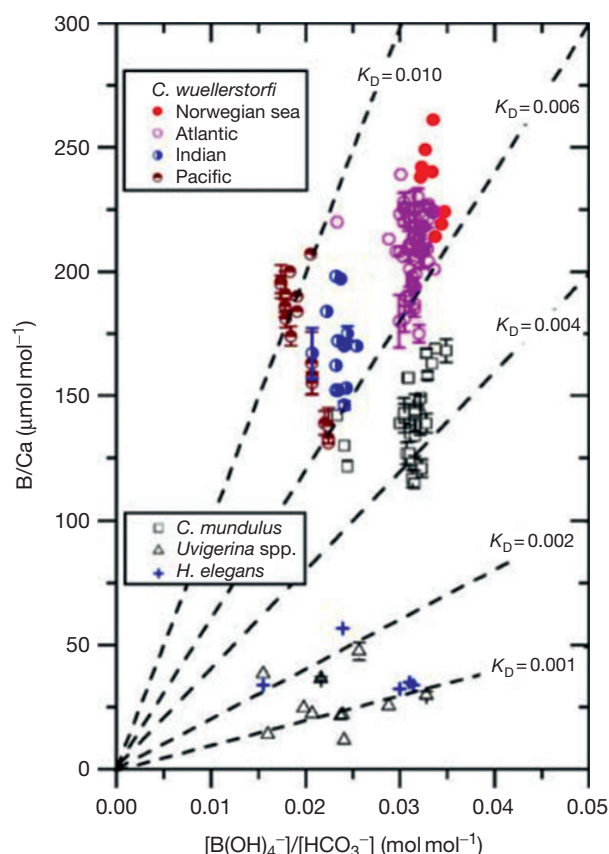


Figure 5 In benthic foraminifers, B/Ca increases with bottom water ΔCO_3^{2-} . Different species are characterized by different B/Ca concentrations, ranging across an order of magnitude from $\sim 25 \mu\text{mol mol}^{-1}$ (*Uvigerina* spp.) to $\sim 250 \mu\text{mol mol}^{-1}$ (*C. wuellerstorfi*). The large B/Ca differences between species and their variable B/Ca – ΔCO_3^{2-} slopes suggest at least a partial biological control on B incorporation into benthic foraminiferal calcite. Reprinted from Yu J and Elderfield H (2007) Benthic foraminiferal B/Ca ratios reflect deep water carbonate saturation state. *Earth and Planetary Science Letters* 258: 73–86.

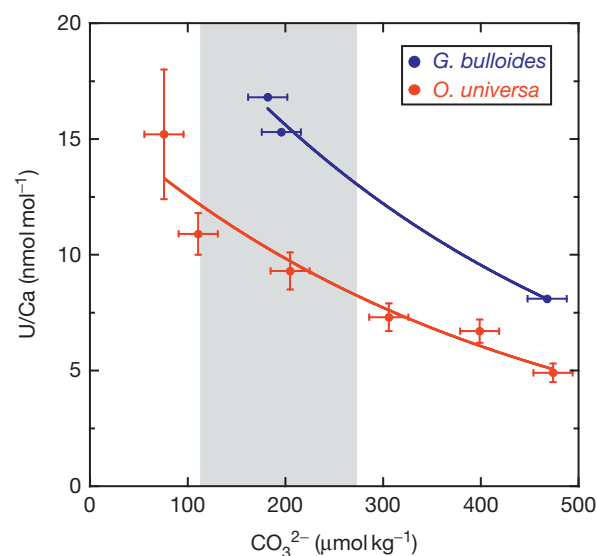


Figure 6 Empirical relationships between U/Ca in shells of *O. universa* and *G. bulloides* and seawater $[\text{CO}_3^{2-}]$. Foraminifers were grown under controlled conditions in the laboratory. The shaded area indicates the range of $[\text{CO}_3^{2-}]$ in modern surface seawater. Reprinted from Russell AD, Hönisch B, Spero HJ, and Lea DW (2004) Effects of changes in seawater carbonate ion concentration and temperature on shell U/Ca, Mg/Ca, and Sr/Ca of planktonic foraminifera. *Geochimica et Cosmochimica Acta* 68(21): 4347–4361.

(Figure 5), which indicates at least a partial biological control on B incorporation. The mechanism for the observed B/Ca– ΔCO_3^{2-} linkage is unclear, but application to glacial North Atlantic sediments gives water-column ΔCO_3^{2-} values that are consistent with $\delta^{13}\text{C}$ (Yu and Elderfield, 2007). The consistency of benthic B/Ca with water mass tracers and with other carbonate system proxies such as $\delta^{11}\text{B}$ suggest that B/Ca may serve as a valuable constraint on deep water chemistry in future studies.

U/Ca to Infer $[\text{CO}_3^{2-}]$

An early study by Russell et al. (1994) investigated the potential of foraminiferal U/Ca for reconstructing the seawater uranium inventory, which is a function of river inflow and removal into sediments overlain by anoxic bottom waters. Using culture experiments with living planktic and benthic foraminifers, the authors demonstrated that U/Ca ratio is incorporated proportional to the seawater concentration and thus may be a useful proxy. However, studying U/Ca from planktic foraminifers through geological time, Russell et al. (1996) found glacial U/Ca ratios approximately 25% lower than Holocene values. Such a large change could not be explained by changes in the seawater uranium concentration and Russell et al. (1996) therefore suggested the observed glacial–interglacial change may be predominantly controlled by seawater $[\text{CO}_3^{2-}]$ or temperature. This hypothesis has been confirmed in culture experiments with the living planktic foraminifers *O. universa* and *Globigerina bulloides*, which showed that while temperature does not exert a significant influence on uranium incorporation, foraminiferal U/Ca is inversely related to $[\text{CO}_3^{2-}]$ (Figure 6; Russell et al., 2004):

$$\text{U/Ca}_{O.universa} = 15.5 \exp(-0.0025[\text{CO}_3^{2-}])$$

$$\text{U/Ca}_{G.bulloides} = 27.9 \exp(-0.0025[\text{CO}_3^{2-}])$$

From a mechanistic point of view, the marine chemistry of uranium is linked to $[\text{CO}_3^{2-}]$ in two ways: (1) uranium occurs in seawater predominantly in the form of carbonate complexes and the concentration of the prevalent triscarbonate complex ($\text{UO}_2(\text{CO}_3)_3^{4-}$) decreases with pH, and (2) uranium adsorption onto shell surfaces is inhibited by $[\text{CO}_3^{2-}]$ (Russell et al., 2004, and references therein). Both mechanisms are consistent with the observation of decreasing U/Ca in foraminifer shells secreted under higher pH (or $[\text{CO}_3^{2-}]$).

Application of the experimental U/Ca relationship to equatorial Atlantic and Caribbean cores suggests glacial surface ocean $[\text{CO}_3^{2-}]$ was $80\text{--}110 \pm 60 \mu\text{mol kg}^{-1}$ higher than during the Holocene (Russell et al., 2004), an estimate that is similar to reconstructions using $\delta^{11}\text{B}$ or the deconvolution of the carbonate ion effect (CIE) on planktic foraminiferal carbon isotopes.

Although first applications demonstrate the proxy is generally promising, the study of several sediment cores revealed that contamination by authigenic carbonate formation may place a significant diagenetic overprint on the incorporated U/Ca. This may limit the general applicability of U/Ca to sediments above the redox front. Another diagenetic complication often observed in chemical proxies is the susceptibility

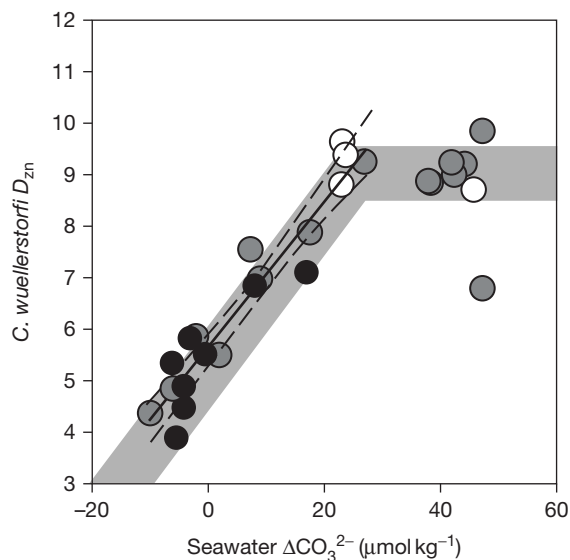


Figure 7 Zn partition coefficients (D_{Zn}) in *Cibicides wuellerstorfi* versus seawater ΔCO_3^{2-} . Measurements are from Holocene core tops from the North Atlantic (white), South Atlantic (dark gray), and Indo-Pacific (black). Line shows relationship used for converting D_{Zn} into ΔCO_3^{2-} (below $30 \mu\text{mol kg}^{-1}$) with 95% confidence intervals (dashed). Also shown is the original relationship proposed by Marchitto et al. (2000) (gray band). Reprinted from Marchitto TM, Lynch-Stieglitz J, and Hemming SR (2005) Deep Pacific CaCO_3 compensation and glacial–interglacial atmospheric CO_2 . *Earth and Planetary Science Letters* 231: 317–336.

of U/Ca to dissolution, which appears to decrease the original ratio (Russell et al., 1994).

Zn/Ca in Benthic Foraminifers

Marchitto et al. (2000) have shown that the partitioning of Zn (D_{Zn}) in tests of the benthic foraminifer species *Cibicides wuellerstorfi* and *Uvigerina* spp. reflects the concentration of dissolved zinc in bottom waters according to the following equation:

$$D_{\text{Zn}} = (\text{Zn/Ca})_{\text{foram}} / (\text{Zn/Ca})_{\text{seawater}}$$

However, Zn/Ca is also strongly affected by the saturation state of bottom waters during calcite precipitation such that D_{Zn} decreases with decreasing seawater calcite saturation state (ΔCO_3^{2-}). Although it has been suggested that these empirical data could represent reduced incorporation of trace metals in undersaturated waters during foraminiferal growth and/or preferential loss of Zn during postdepositional dissolution, investigations using Rose Bengal stained (i.e., recently living) foraminifer tests have shown that Zn/Ca incorporation is dominated by the effects of calcification rather than dissolution (Marchitto et al., 2000). As such, if ancient seawater zinc concentrations can be constrained, then foraminiferal Zn/Ca can be used to infer ΔCO_3^{2-} at the time of calcification.

In contrast to U/Ca and B/Ca, which are sensitive to seawater carbonate chemistry over a wide range of conditions, D_{Zn} is only sensitive to ΔCO_3^{2-} at $\Delta\text{CO}_3^{2-} < 30 \mu\text{mol kg}^{-1}$, that is, close to the saturation horizon in the deep ocean. Assuming a

linear relationship below $\sim 30 \mu\text{mol kg}^{-1} \Delta\text{CO}_3^{2-}$, the association between D_{Zn} and changes in carbonate ion saturation can be described by the following equation (Figure 7, Marchitto et al., 2005):

$$D_{\text{Zn}} = 0.143 \times \Delta\text{CO}_3^{2-} + 5.63$$

Application of this relationship to benthic foraminifers from sediments in the deep tropical Pacific allowed Marchitto et al. (2005) to infer deep ocean paleo- ΔCO_3^{2-} from the last glacial to the modern. Their data show glacial ΔCO_3^{2-} slightly lower than in the modern ocean as well as a brief positive excursion of $\sim 25\text{--}30 \mu\text{mol kg}^{-1}$ during the deglaciation. These data are broadly consistent with the hypothesis that the glacial addition of atmospheric CO_2 to the deep ocean caused dissolution of sedimentary CaCO_3 . During the deglaciation, CO_2 returned to the atmosphere, leaving the deep ocean temporarily oversaturated with respect to CaCO_3 , until deposition increased and a new equilibrium was subsequently established during the Holocene.

This study is a promising first attempt at using benthic foraminiferal Zn/Ca, but it has not yet been complemented by newer studies. Of particular concern for deeper time paleoreconstructions is the relatively short residence of Zn in the ocean, which is on the order of $\sim 10\text{--}30$ ka (Marchitto et al., 2005). In addition, Zn behaves like a nutrient and its deep ocean concentration varies as a function of organic matter degradation, making direct comparisons between different locations difficult in the absence of additional constraints. Further studies are needed to verify these initial results and in particular the Zn inventory of the paleocean should be better constrained.

Foraminiferal Ba/Ca

The distribution of dissolved barium in the modern ocean varies with alkalinity and silica in that all three are partially depleted in the surface ocean and more concentrated with increasing water depth. The removal of alkalinity, silica, and barium from surface waters is attributed to biological processes in which planktic calcifiers, diatoms, and radiolaria secrete CaCO_3 and opal skeletons, and barite (BaSO_4) is precipitated in association with decaying organic matter. Lea and Boyle (1989) pointed out that Ba and alkalinity are ocean-wide more closely related than Ba and silica, and although stoichiometric analysis of sinking particles and their regeneration in the deep sea could not fully resolve the mechanistic linkage between alkalinity and Ba (Lea, 1993), Lea (1993) suggested that changes in thermohaline circulation should still redistribute barium and alkalinity similarly, thereby allowing reconstruction of past alkalinity distributions from benthic foraminiferal Ba/Ca (e.g., Hall and Chan, 2004; Lea, 1993; Lea and Boyle, 1989).

In order for this proxy to faithfully record seawater alkalinity, the incorporation of barium in foraminiferal tests must be in proportion to the seawater Ba concentration, which has been verified in laboratory culture with planktic foraminifers. In addition, Ba incorporation furthermore does not appear to vary with temperature, salinity, or seawater pH (Hönisch et al., 2011, and references therein). However, benthic flux measurements in the modern Pacific and Atlantic Oceans found that

the Ba/alkalinity flux ratio was only $\sim 1/3$ of that expected from the modern deep-water trend, suggesting that the locations of precipitation and regeneration are not linked closely (Rubin et al., 2003). This may be an indication that the oceanic Ba-alkalinity correlation is not coupled on longer time scales and because the residence time of barium is relatively short (~ 10 ka), the accuracy of benthic foraminiferal Ba/Ca as an indicator of paleo-alkalinity remains in doubt.

Deconvolution of the CIE to Infer $[\text{CO}_3^{2-}]$

The stable isotopic composition of dissolved inorganic carbon in seawater ($\delta^{13}\text{C}_{\text{DIC}}$) is affected by a number of different processes, including global carbon cycling, ocean circulation, primary productivity, and organic carbon degradation. The carbon isotopic composition of marine carbonates ($\delta^{13}\text{C}_c$) records changes in $\delta^{13}\text{C}_{\text{DIC}}$, although carbon incorporation into the skeleton is also affected by additional parameters such as physiological life processes, shell size of planktic foraminifers and growth temperature. In addition, culture experiments with living planktic foraminifers have revealed that $\delta^{13}\text{C}_c$ and $\delta^{18}\text{O}_c$ decrease with increasing carbonate ion concentration and pH of the culture water (Spero et al., 1999, and references therein).

Mechanistic explanations for this CIE point to isotope fractionation between the dissolved carbon species in seawater. With regard to $\delta^{18}\text{O}_c$, the observed effect was attributed to the isotope partitioning between $[\text{HCO}_3^-]$ and $[\text{CO}_3^{2-}]$ in seawater and the relatively enhanced availability of isotopically lighter CO_3^{2-} at high pH (Zeebe, 1999). The effect on $\delta^{13}\text{C}_c$ is less well understood but may be related to physiological life processes, particularly respiration. Because at high pH the amount of respired CO_2 in the vicinity of a foraminifer is extremely high relative to ambient seawater concentration, Zeebe et al. (1999) suggested that conversion of this isotopically light CO_2 to HCO_3^- and CO_3^{2-} leads to an isotopic depletion of these carbon species relative to bulk seawater and thus to low shell $\delta^{13}\text{C}_c$.

Although these influences on foraminiferal $\delta^{13}\text{C}_c$ and $\delta^{18}\text{O}_c$ are relatively small, there is good evidence that marine carbonate chemistry measurably affects paleoceanographic records. For instance, the pH effect on foraminiferal $\delta^{18}\text{O}_c$ may be responsible for an underestimation of Cretaceous tropical sea surface temperatures (Zeebe, 2001), and instead of reporting the transfer of organic carbon from the terrestrial to the oceanic biosphere, the glacial-interglacial shift in $\delta^{13}\text{C}_c$ may be explained by higher glacial $[\text{CO}_3^{2-}]$ (Spero et al., 1997). Correcting $\delta^{13}\text{C}_c$ and $\delta^{18}\text{O}_c$ records for the CIE is therefore important for accurate paleoceanographic interpretations.

In addition, species-specific differences in the CIE also present a new way of reconstructing $[\text{CO}_3^{2-}]$ in the paleocean. Among the tested planktic foraminifers, *G. sacculifer* and *G. ruber* share approximately the same depth habitat, but the $\delta^{13}\text{C}_c/[\text{CO}_3^{2-}]$ slope (m) is twice as large in *G. ruber* as in *G. sacculifer* ($m_{G. \text{ ruber}} = -0.0089\text{‰} \mu\text{mol}^{-1} \text{ kg}$ and $m_{G. \text{ sacculifer}} = -0.0047\text{‰} \mu\text{mol}^{-1} \text{ kg}$). Assuming specific vital effects and depth habitat of these species did not change through time, the species-specific difference in the slopes can be used to distinguish between the effect of $[\text{CO}_3^{2-}]$ and a simultaneous change in $\delta^{13}\text{C}_{\text{DIC}}$ (for a detailed explanation see Spero et al., 1999). Application to the sediment record yields

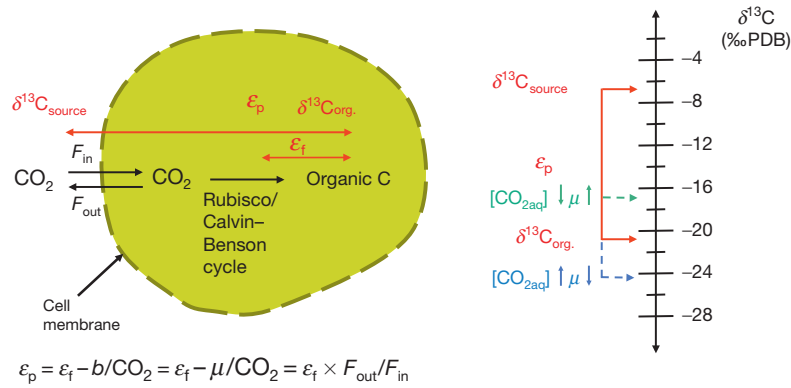


Figure 8 Conceptual model of the $\delta^{13}\text{C}_{\text{alkenones}}$ proxy for seawater PCO_2 . The left panel shows a schematic phytoplankton cell with a membrane that is permeable for CO_2 to diffuse in and out of the cell. ϵ_p denotes the total photosynthetic carbon isotope fractionation between $\delta^{13}\text{C}$ of the carbon source in seawater and the organic product. ϵ_f reflects carbon fractionation associated with the enzymatic carbon fixation. μ is the growth rate and b is an empirical factor combining the effects of all physiological influences. The right panel visualizes the respective effects of CO_2 and μ on ϵ_p , where lower CO_2 and higher growth rate (shown in green) decrease ϵ_p , whereas higher CO_2 and lower growth rate (shown in blue) increase ϵ_p . Figures provided by courtesy of A. Benthien.

an estimate of $+55 \pm 63 \mu\text{mol kg}^{-1}$ in $[\text{CO}_3^{2-}]$ for the equatorial Indian Ocean during the last glacial period (Spero et al., 1999). Unfortunately, this method is restricted to tropical surface waters, where *G. sacculifer* and *G. ruber* both occur.

Carbon Isotope Fractionation in Alkenones to Infer Aqueous PCO_2

Alkenones are specific biomolecules that are exclusively synthesized by a group of calcite-secreting phytoplankton, coccolithophores. Alkenones are well preserved in marine sediments and their carbon isotope composition ($\delta^{13}\text{C}_{\text{alkenones}}$) has been used to reconstruct the aqueous concentration of CO_2 , $[\text{CO}_{2\text{aq}}]$, in the surface ocean. Reconstructing $[\text{CO}_{2\text{aq}}]$ from $\delta^{13}\text{C}_{\text{alkenones}}$ is based on the idea that marine phytoplankton assimilate CO_2 by passive diffusion from seawater into the cell, and the carbon isotope fractionation associated with the carbon fluxes in (F_{in}) and out (F_{out}) of the cell thus would be mostly controlled by $[\text{CO}_{2\text{aq}}]$. However, laboratory and field experiments have shown that photosynthetic carbon isotope fractionation (ϵ_p) can be influenced by a number of parameters, including nutrient concentration, irradiance, growth rate, and the carbon source, that is, CO_2 and/or HCO_3^- . Changing ideas of how photosynthetic carbon fractionation actually takes place is expressed by the different models used to describe the processes involved (Figure 8):

$$\epsilon_p = \epsilon_f - b/\text{CO}_2 \quad (1)$$

$$\epsilon_p = \epsilon_f - \mu/\text{CO}_2 \quad (2)$$

$$\epsilon_p = \epsilon_f \times F_{\text{out}}/F_{\text{in}} \quad (3)$$

where ϵ_f denotes the fractionation associated with enzymatic carbon fixation (e.g., 29‰ for the enzyme Rubisco), b is an empirical factor combining the effects of physiological influences, μ is the growth rate, and the term $F_{\text{out}}/F_{\text{in}}$ denotes CO_2 leakage, that is, the ratio of CO_2 efflux over influx. ϵ_p is thus small when growth rate is high and increases under elevated

$[\text{CO}_{2\text{aq}}]$, that is, when CO_2 leakage out of the cell increases (Figure 8). Traditionally, measured values of ϵ_p are shown versus $1/[\text{CO}_{2\text{aq}}]$ and if the equations described above are correct and ϵ_p was predominantly controlled by $[\text{CO}_{2\text{aq}}]$, alkenones should show a negative correlation between ϵ_p and $1/[\text{CO}_{2\text{aq}}]$. However, alkenones taken from core-top sediments in the South Atlantic display a positive correlation between ϵ_p and $1/[\text{CO}_{2\text{aq}}]$, indicating the dominant importance of nitrate and phosphate availability for the carbon isotope fractionation (Benthien et al., 2002). In addition, mesocosm culture experiments show that the isotopic fractionation varies by only 3–4‰ over a PCO_2 range from 180 to 700 ppmv (Benthien et al., 2007), demonstrating that the CO_2 sensitivity of $\epsilon_{\text{alkenones}}$ is low in the modern ocean, where coccolithophores are adapted to low $[\text{CO}_{2\text{aq}}]$. Despite these apparent complications, recent reconstructions of the Pliocene pCO_2 history from alkenones (Figure 9; Pagani et al., 2010) and boron isotopes and alkenones (Figure 3; Seki et al., 2010) have given promisingly similar estimates of Pliocene pCO_2 that are ~ 100 ppm higher compared to preindustrial times. In addition, early Cenozoic alkenone estimates are high and consistent with independent evidence for a warmer climate at that time (Pagani et al., 2005). It is clear from these studies that reasonable assumptions regarding nutrient concentrations have to be made to enable meaningful CO_2 estimates from alkenones. However, the comparison of reconstructions from six different sediment cores from the Atlantic and Pacific Ocean with atmospheric pCO_2 measured in ice cores (Figure 9; Pagani et al., 2010) demonstrates that a global $\delta^{13}\text{C}_{\text{alkenones}}$ calibration is insufficient to accurately describe different oceanographic settings (Figure 9).

Size-Normalized Shell Weights

The average shell weight of planktic foraminifers is primarily determined by their size, but there are measurable secondary relationships of shell mass to (1) ambient $[\text{CO}_3^{2-}]$ during shell precipitation (Barker and Elderfield, 2002), and (2) bottom water carbonate saturation, which drives shell dissolution

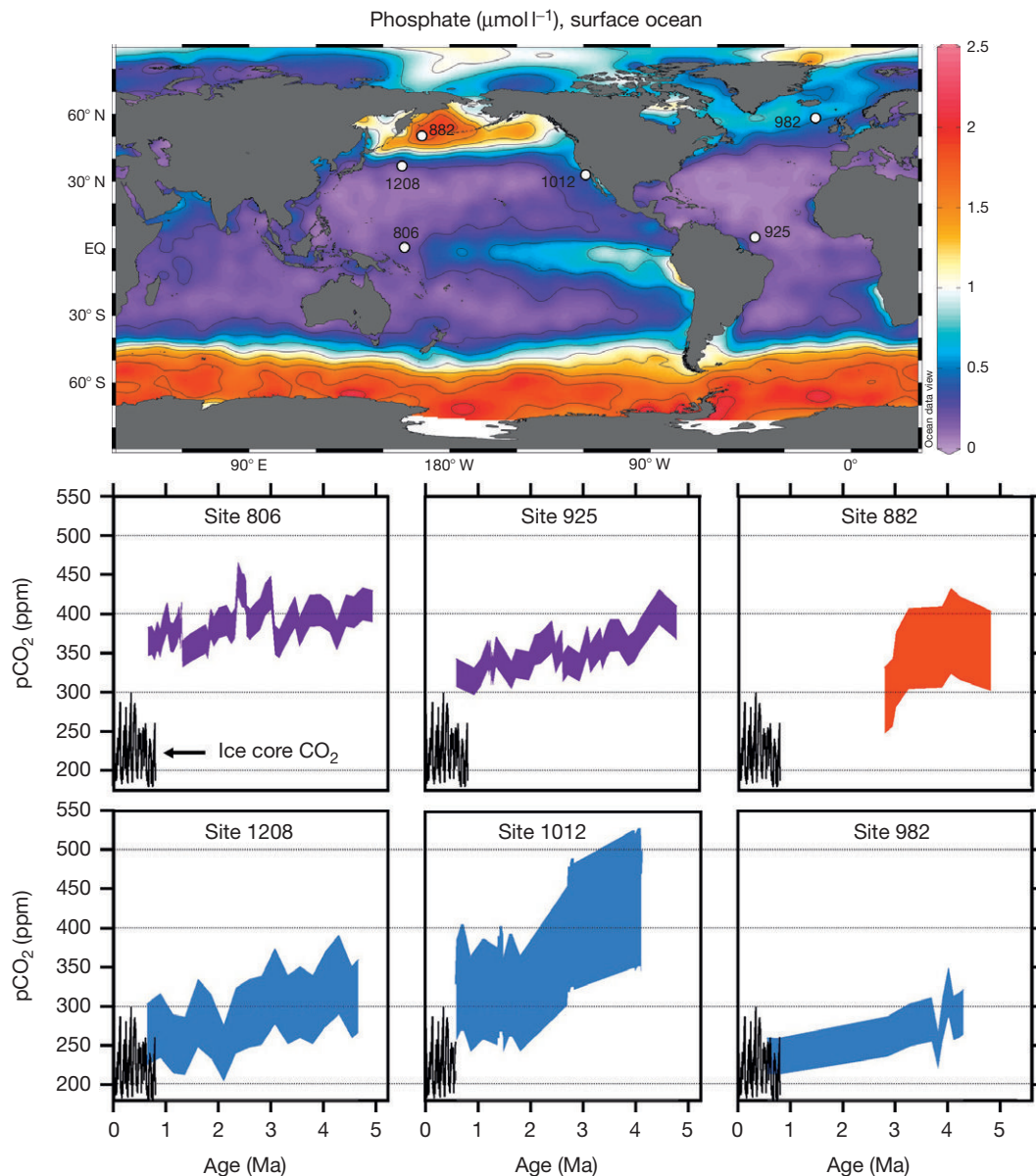


Figure 9 Plio-Pleistocene surface-ocean PCO_2 estimated from six different sediment cores in the Atlantic and Pacific Ocean. The map shows the core sites with their modern annual average phosphate concentrations ($\mu\text{mol l}^{-1}$), plotted from the World Ocean Atlas 2005 data set in Ocean Data View v. 4.0.1). The width of the data bands reflects maximum and minimum PCO_2 estimates as a function of nutrient concentrations at different alkenone production depths. Ice core pCO_2 is shown for comparison with each alkenone estimate and demonstrates that few alkenone estimates match atmospheric pCO_2 in coeval intervals. In contrast to the expectation that surface-ocean PCO_2 should be in equilibrium with atmospheric pCO_2 in oligotrophic ocean areas (sites 806 and 925, in purple), alkenone estimates best match ice core measurements in the mesotrophic–eutrophic North Atlantic (site 982, in blue). It is promising that all sites consistently show higher PCO_2 during the Pliocene but it is also obvious that single-site reconstructions may not yield reliable atmospheric pCO_2 estimates. Adapted from Pagani M, Liu Z, LaRiviere J, and Ravelo AC (2010) High Earth-system climate sensitivity determined from Pliocene carbon dioxide concentrations. *Nature Geoscience* 3: 27–30.

(Lohmann, 1995). To better constrain the small weight changes associated with these processes and to correct for size variability, paleoceanographers restrict their examinations to defined, narrow size ranges of single foraminifer species and collect a minimum of 50 shells per sample. Data from such samples are referred to as size-normalized shell weights (SNW), which can then be compared to similar samples of different age or location.

Because $[\text{CO}_3^{2-}]$ in both surface and deep waters changes through geological time, one needs to differentiate between variations in surface ocean calcification and deep ocean dissolution on a given sample, Barker et al. (2004) introduced a new approach where proximal sediment cores from depth profiles are compared. The shallowest cores are supposedly unaffected by dissolution and thus reflect shell weight changes through differences in $[\text{CO}_3^{2-}]$ during calcification in the surface ocean

only. This information can then be used to extract the effect of bottom water carbonate saturation on SNW changes in deeper cores.

Estimating changes in seawater $[\text{CO}_3^{2-}]$ from SNW of planktic foraminifers requires empirically calibrated relationships. Due to partial dissolution, the shell weight of nearly all species is lower in deeper water than it is in shallow water, and the decrease is continuous over a wide range of carbonate saturation states, even well above the calcite lysocline. Based on shell weights of three different planktic foraminifer species, Broecker and Clark (2001) determined an average weight-loss slope of $0.3 \pm 0.05 \mu\text{g } \mu\text{mol}^{-1} \text{kg}^{-1}$ decrease in pressure-corrected deep-sea carbonate ion concentration. This slope is provided as an example and should not be considered universal. Because susceptibility to dissolution differs between foraminifer species, species-specific relationships have to be applied to enable meaningful paleoceanographic reconstructions. Similarly, the SNW/ $[\text{CO}_3^{2-}]$ relationship in the surface ocean differs between species and also requires specific calibrations (Barker et al., 2004). In addition to specific calibrations, first studies indicate that additional factors such as growth habitat, food supply, salinity etc. need to be considered as potential secondary influences on the initial shell weight variability (Barker et al., 2004; Broecker and Clark, 2001). Although the relationship thus seems to be more complex than originally thought, application to the glacial north Atlantic resulted in realistic surface ocean PCO_2 estimates and $[\text{CO}_3^{2-}] \sim 60 \mu\text{mol kg}^{-1}$ higher than Holocene values (Barker and Elderfield, 2002).

To use SNW as a paleocarbonate ion proxy in ocean bottom water, this method requires the offset between pore and bottom water saturation to be constant. However, this may not be a valid assumption because calcite dissolution also occurs above the saturation horizon, where it is attributed to pore-water acidification caused by degradation of organic matter. The amount of organic matter reaching the seafloor varies between sites and also depends on depth. Assuming increased productivity during glacial intervals, the magnitude of this effect might have been even stronger. Application of this proxy should therefore be restricted to locations where strong changes in paleoproductivity are not expected.

Conclusions

Quantifying past changes in the amount and distribution of dissolved carbon in the ocean can clarify the role of CO_2 in climate change, and this may be accomplished by careful application of a variety of proxies for the marine carbonate system. The proxies described above are based on a wide range of independent organic and inorganic controls and are characterized by different strengths and weaknesses. The record of a target parameter may be compromised or limited by secondary influences of competing physical or chemical parameters, vital effects, diagenesis, limited availability of carbonates in the sedimentary archive, and temporal limitations via the residence time of an element in seawater. To improve confidence in paleoceanographic reconstructions of marine carbonate chemistry, comparison of various proxy data sets is therefore desirable. The general advantage of preservational proxies such

as shell weights is their non-destructive character, as weight analysis leaves foraminiferal samples untouched and allows further chemical investigation. Apart from providing additional information on carbonate saturation, knowledge of the foraminiferal preservation state can also elucidate the possible influence of dissolution on chemical proxies like $\delta^{11}\text{B}$ and U/Ca. Due to extensive culture studies with living planktic foraminifers, physical and chemical proxies are often better understood in planktic than in benthic foraminifers. This will probably change in the near future because recent progress in culturing benthic foraminifers promises the establishment of empirical proxy relationships under controlled laboratory conditions and better understanding of benthic proxies.

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See also: **Paleoceanography, Biological Proxies:** Alkenone Paleothermometry Based on the Haptophyte Algae; Biomarker Indicators of Past Climate; Temperature Proxies, Census Counts. **Paleoceanography, Physical and Chemical Proxies:** Salinity Proxies $\delta^{18}\text{O}$.

References

- Allen KA, Hönisch B, Eggins SM, Yu JM, Spero HJ, and Elderfield H (2011) Controls on boron incorporation in cultured tests of the planktic foraminifer *Orbulina universa*. *Earth and Planetary Science Letters* 309: 291–301.
- Barker S and Elderfield H (2002) Foraminiferal calcification response to glacial–interglacial changes in atmospheric CO_2 . *Science* 297: 833–836.
- Barker S, Kiefer T, and Elderfield H (2004) Temporal changes in North Atlantic circulation constrained by planktonic foraminiferal shell weights. *Paleoceanography* 19: 1–15.
- Benthien A, Andersen N, Müller PJ, Schneider RR, and Wefer G (2002) Carbon isotopic composition of C37:2 alkenones in core-top sediments of the South Atlantic Ocean: Effects of CO_2 and nutrient concentrations. *Global Biogeochemical Cycles* 16: 1012. <http://dx.doi.org/10.1029/2001GB001433>.
- Benthien A, Zondervan I, Engel A, Heffer J, Terbruggen A, and Riebesell U (2007) Carbon isotopic fractionation during a mesocosm bloom experiment dominated by *Emiliania huxleyi*: Effects of CO_2 concentration and primary production. *Geochimica et Cosmochimica Acta* 71: 1528–1541.
- Broecker W and Clark E (2001) An evaluation of Lohmann's foraminifera weight dissolution index. *Paleoceanography* 16: 531–534.
- Dickson AG (1990) Thermodynamics of the dissociation of boric acid in synthetic seawater from 273.15 to 318.15 K. *Deep Sea Research* 37: 755–766.
- Foster GL (2008) Seawater pH, pCO_2 and $[\text{CO}_3^{2-}]$ variations in the Caribbean Sea over the last 130 kyr: A boron isotope and B/Ca study of planktic foraminifera. *Earth and Planetary Science Letters* 271: 254–266.
- Hall JM and Chan LH (2004) Ba/Ca in benthic foraminifera: Thermocline and middepth circulation in the North Atlantic during the last glaciation. *Paleoceanography* 19: 13.
- Hemming NG and Hanson GN (1992) Boron isotopic composition and concentration in modern marine carbonates. *Geochimica et Cosmochimica Acta* 56: 537–543.
- Hönisch B, Allen KA, Russell AD, et al. (2011) Planktic foraminifers as recorders of seawater Ba/Ca. *Marine Micropaleontology* 79: 52–57.
- Hönisch B, Bijma J, Russell AD, et al. (2003) The influence of symbiont photosynthesis on the boron isotopic composition of foraminifera shells. *Marine Micropaleontology* 49: 87–96.

- Hönisch B and Hemming NG (2004) Ground-truthing the boron isotope paleo-pH proxy in planktonic foraminifera shells: Partial dissolution and shell size effects. *Paleoceanography* 19: 1–13.
- Hönisch B and Hemming NG (2005) Surface ocean pH response to variations in pCO₂ through two full glacial cycles. *Earth and Planetary Science Letters* 236: 305–314.
- Hönisch B, Hemming NG, Archer D, Siddall M, and McManus JF (2009) Atmospheric carbon dioxide concentration across the mid-Pleistocene transition. *Science* 324: 1551–1554.
- Klochko K, Kaufman AJ, Yao W, Byrne RH, and Tossell JA (2006) Experimental measurement of boron isotope fractionation in seawater. *Earth and Planetary Science Letters* 248: 261–270.
- Lea DW (1993) Constraints on the alkalinity and circulation of glacial circumpolar deep water from benthic foraminiferal barium. *Global Biogeochemical Cycles* 7: 695–710.
- Lea DW and Boyle EA (1989) Barium content of benthic foraminifera controlled by bottom-water composition. *Nature* 338: 751–753.
- Lohmann GP (1995) A model for variation in the chemistry of planktonic foraminifera due to secondary calcification and selective dissolution. *Paleoceanography* 10: 445–457.
- Marchitto TM Jr., Curry WB, and Oppo DW (2000) Zinc concentrations in benthic foraminifera reflect seawater carbonate chemistry. *Paleoceanography* 15: 299–306.
- Marchitto TM, Lynch-Stieglitz J, and Hemming SR (2005) Deep Pacific CaCO₃ compensation and glacial–interglacial atmospheric CO₂. *Earth and Planetary Science Letters* 231: 317–336.
- Ni Y, Foster GL, Bailey T, et al. (2007) A core top assessment of proxies for the ocean carbonate system in surface-dwelling foraminifers. *Paleoceanography* 22: 1–14.
- Pagani M, Liu Z, LaRiviere J, and Ravelo AC (2010) High Earth-system climate sensitivity determined from Pliocene carbon dioxide concentrations. *Nature Geoscience* 3: 27–30.
- Pagani M, Zachos JC, Freeman KH, Tipple B, and Bohaty S (2005) Marked decline in atmospheric carbon dioxide concentrations during the Paleogene. *Science* 309: 600–603.
- Rubin SI, King SL, Jahnke RA, and Froelich PN (2003) Benthic barium and alkalinity fluxes: Is Ba an oceanic paleo-alkalinity proxy for glacial atmospheric CO₂? *Geophysical Research Letters* 30: 1885.
- Russell AD, Emerson S, Mix AC, and Peterson LC (1996) The use of foraminiferal uranium/calcium ratios as an indicator of changes in seawater uranium content. *Paleoceanography* 11: 649–663.
- Russell AD, Emerson S, Nelson BK, Erez J, and Lea DW (1994) Uranium in foraminiferal calcite as a recorder of seawater uranium concentrations. *Geochimica et Cosmochimica Acta* 58: 671–681.
- Russell AD, Hönisch B, Spero HJ, and Lea DW (2004) Effects of changes in seawater carbonate ion concentration and temperature on shell U/Ca, Mg/Ca, and Sr/Ca of planktonic foraminifera. *Geochimica et Cosmochimica Acta* 68: 4347–4361.
- Seki O, Foster GL, Schmidt DN, Mackensen A, Kawamura K, and Pancost RD (2010) Alkenone and boron-based Pliocene pCO₂ records. *Earth and Planetary Science Letters* 292: 201–211.
- Spero HJ, Bijma J, Lea DW, and Bemis BE (1997) Effect of seawater carbonate concentration on foraminiferal carbon and oxygen isotopes. *Nature* 390: 497–500.
- Spero HJ, Bijma J, Lea DW, and Russell AD (1999) Deconvolving glacial ocean carbonate chemistry from the planktonic foraminifera carbon isotope record. In: Abrantes F and Mix A (eds.) *Reconstructing Ocean History: A Window into the Future*, pp. 329–342. New York: Kluwer Academic/Plenum.
- Yu J and Elderfield H (2007) Benthic foraminiferal B/Ca ratios reflect deep water carbonate saturation state. *Earth and Planetary Science Letters* 258: 73–86.
- Yu J, Elderfield H, and Hönisch B (2007) B/Ca in planktonic foraminifera as a proxy for surface seawater pH. *Paleoceanography* 22: PA2202.
- Zeebe RE (1999) An explanation of the effect of seawater carbonate concentration on foraminiferal oxygen isotopes. *Geochimica et Cosmochimica Acta* 63: 2001–2007.
- Zeebe RE (2001) Seawater pH and isotopic paleotemperatures of Cretaceous oceans. *Palaeo* 170: 49–57.
- Zeebe RE, Bijma J, and Wolf-Gladrow D (1999) A diffusion–reaction model of carbon isotope fractionation in foraminifera. *Marine Chemistry* 64: 199–227.

Dissolution of Deep-Sea Carbonates

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Introduction

Marine carbonates represent the major component of deep-sea sediments (Figure 1; Table 1). The distribution of carbonates throughout the world's oceans reflects the interplay between production in the surface ocean and dissolution, either in the water column, or at the seafloor before final burial. Since the modern rate of carbonate production is several times greater than the global burial rate, dissolution is an important term. The dissolution of carbonates in the deep sea is ultimately controlled by the saturation state of mean ocean water, which itself is a function of the balance between the rate of supply (through weathering) and burial of carbonate at the sea floor. In this way, the system can be thought of as self-regulating. The importance of carbonate dissolution within the global carbon cycle can be appreciated by its relation to mean ocean saturation and atmospheric CO₂. As a rule of thumb, increased dissolution will result from increased levels of CO₂ and a corresponding decrease in ocean saturation state. Past changes in atmospheric CO₂ have been linked to changes in carbonate dissolution intensity. Similarly, much of the CO₂ currently emitted by fossil fuel burning will ultimately be sequestered by an increase in deep-sea carbonate dissolution.

Global Distribution of Deep-Sea Carbonates

Surface Production

Almost all the carbonate in deep-sea sediments is manufactured by marine organisms living in the top few hundred meters of the water column. Various species of phytoplankton and zoo plankton produce calcium carbonate (CaCO₃) shells or platelets, that fall to the seafloor after death. Some organisms build tests made of calcite (e.g., coccolithophores and Foraminifera) while others (e.g., pteropods) form aragonite shells. Differences in the carbonate mineral and form result in different distributions with water depth due to their varying solubilities (see the Section 'Mechanisms of Carbonate Dissolution').

The primary control on the distribution of deep-sea carbonates is therefore the pattern of biological production in the surface ocean (Figure 1). Carbonate-rich sediments (oozes) are to be found beneath highly productive regions such as coastal or equatorial upwelling zones (a notable exception being the Southern Ocean), whereas oligotrophic areas are associated with sediments lower in carbonate content.

Dissolution in the Deep Sea

One of the most striking features of ocean sedimentation is the dramatic decrease in the carbonate content of seafloor sediments with increasing water depth. Topographic highs such as mid-ocean ridges tend to be draped in carbonate-rich oozes,

whereas abyssal depth sediments are almost devoid of carbonates (Figure 2). The transition between these two realms is a zone, several hundred meters wide, over which the CaCO₃ content drops from perhaps 85–95% to almost zero. The top of this zone (where CaCO₃ content starts to decrease rapidly) is generally termed the sedimentary lysocline (this term has also been applied to the entire transitional zone). The depth below which %CaCO₃ is effectively zero (where the flux of carbonate to the seafloor is matched by the rate of dissolution) is termed the carbonate compensation depth (the calcite compensation depth, CCD, is deeper than the aragonite compensation depth, ACD, due to the greater solubility of aragonite; see the Section 'Mechanisms of Carbonate Dissolution'). Understanding why this transition occurs and the overall role of carbonate dissolution within the global carbon cycle requires an appreciation of certain thermodynamic and kinetic considerations.

Thermodynamics of the Carbonate System in Seawater

Dissolved Inorganic Carbon, DIC

The dissolution of carbonate in seawater is intimately related to the marine carbon dioxide (CO₂) system. CO₂ dissolved in seawater exists in three inorganic forms: CO₂ (aq.) (aqueous CO₂), HCO₃[−] (bicarbonate ion), and CO₃^{2−} (carbonate ion). A fourth form, H₂CO₃ (carbonic acid), is present only in very low concentrations and is generally considered in combination with CO₂ (aq.) and denoted as CO₂:

$$[\text{CO}_2] = [\text{CO}_2(\text{aq.})] + [\text{H}_2\text{CO}_3] \quad [1]$$

Here square brackets denote the stoichiometric concentrations (molarity or mol kg^{−1}). The total dissolved inorganic carbon (DIC or ΣCO₂) in seawater is therefore defined as

$$\text{DIC} \equiv \Sigma\text{CO}_2 = [\text{CO}_2] + [\text{HCO}_3^-] + [\text{CO}_3^{2-}] \quad [2]$$

Under equilibrium conditions (no net exchange of CO₂ between air and sea), the concentration of CO₂ in the surface ocean is related to the fugacity (or partial pressure) of CO₂ in the atmosphere according to Henry's law (the fugacity of a gas is similar to its partial pressure but allows for its nonideality):

$$\text{CO}_2(\text{g}) \stackrel{K_0}{=} \text{CO}_2(\text{aq.}) \quad [3]$$

Table 1 Global distribution of the major sedimentary types

Component	Atlantic	Pacific	Indian	World ocean
Calcareous ooze	68	36.2	54.3	47.7
Siliceous ooze	6.7	14.7	20.4	14.2
Red clay	25.3	49.1	25.3	38.1

Data from Thurman HV and Trujillo AP (2002) *Essentials of Oceanography*, 7th ed. Upper Saddle River: Prentice-Hall.

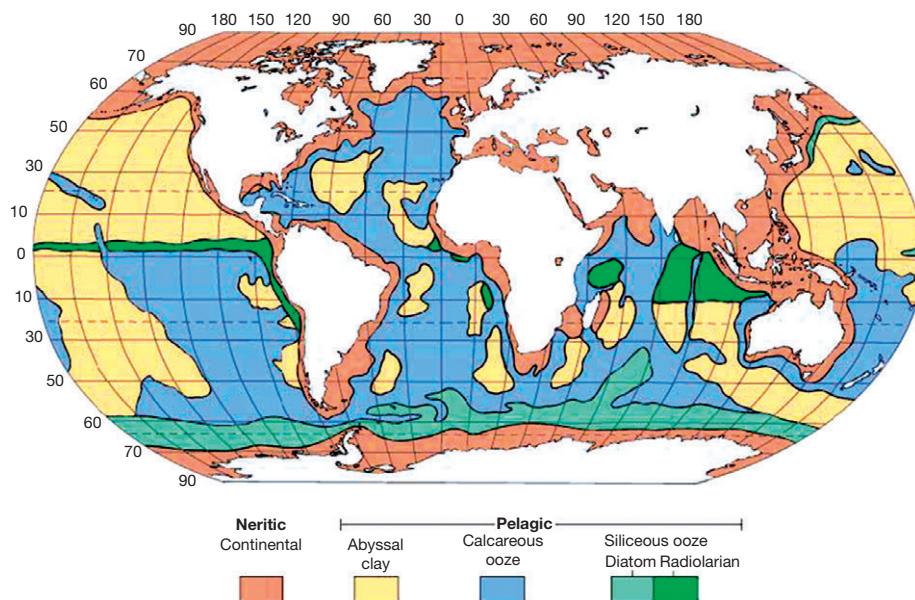
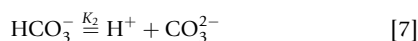
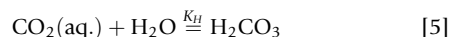


Figure 1 Global distribution of dominant sedimentary types. Reprinted from Thurman HV and Trujillo AP (2002) *Essentials of Oceanography*, 7th ed. Upper Saddle River: Prentice-Hall.

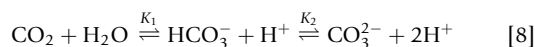
where K_0 is the solubility coefficient of CO_2 in seawater:

$$K_0 = [\text{CO}_2]/f \text{CO}_2 \quad [4]$$

The dissolved carbonate species are related by the following equilibria:



where K_1 and K_2 are the first and second dissociation constants of carbonic acid (which do not differentiate between $\text{CO}_2(\text{aq.})$ and H_2CO_3). Thus, the system may be simplified to



It is most common to use the stoichiometric equilibrium constants, K_1^* and K_2^* , which are defined as

$$K_1^* = [\text{H}^+][\text{HCO}_3^-]/[\text{CO}_2] \quad [9]$$

$$K_2^* = [\text{H}^+][\text{CO}_3^{2-}]/[\text{HCO}_3^-] \quad [10]$$

K_1^* and K_2^* are dependent on temperature (T), salinity (S), and pressure (P) (Table 2, Figure 3). Expressions for $K_{0,1,2}$ have been determined repeatedly and the reader is referred to DOE (1994) for current recommended usage and references.

The partitioning of DIC between CO_2 , HCO_3^- , and CO_3^{2-} can be usefully illustrated with a Bjerrum plot (Figure 3). It can be seen that, for typical ocean water, HCO_3^- dominates ($\sim 90\%$), followed by CO_3^{2-} . CO_2 represents only a few percent of the total dissolved inorganic carbon in seawater. This biasing away from CO_2 explains why the oceanic carbon reservoir is so much (some 50–60 times) larger than the atmosphere and

leads to the thought that the oceans must play an important role in controlling atmospheric CO_2 levels.

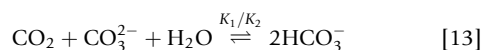
It can also be seen from Figure 3 that $[\text{HCO}_3^-]$ is relatively constant for average oceanic pH values. This leads to a valuable rule of thumb; the concentration of carbonate ion, $[\text{CO}_3^{2-}]$, is inversely related to $[\text{CO}_2]$. This generalized statement can be derived numerically, by combining Eqns [9] and [10]:

$$K_1^*/K_2^* = [\text{HCO}_3^-]^2/[\text{CO}_2]/[\text{CO}_3^{2-}] \quad [11]$$

Since K_1^* , K_2^* , and $[\text{HCO}_3^-]$ are effectively constant for a given T , S , and P , Eqn [11] reduces to

$$[\text{CO}_2]/[\text{CO}_3^{2-}] \approx \text{constant} \quad [12]$$

From Eqn [11] we may also derive the pH reaction



Alkalinity, TA

The total alkalinity (TA) of seawater may be defined as the charge difference between the major conservative (concentration unaffected by changes in pH, pressure, or temperature) cations and anions:

$$\text{TA} = \Sigma \text{conservative cations} - \Sigma \text{conservative anions}$$

$$\text{TA} \approx ([\text{Na}^+] + 2[\text{Mg}^{2+}] + 2[\text{Ca}^{2+}] + [\text{K}^+]) - ([\text{Cl}^-] + 2[\text{SO}_4^{2-}]) \quad [14]$$

The small excess positive charge (~ 2 as compared with $\sim 600 \text{ mmol kg}^{-1}$ total positive charge concentration) is balanced by contributions from carbonate, borate, and water alkalinities (Figure 4):

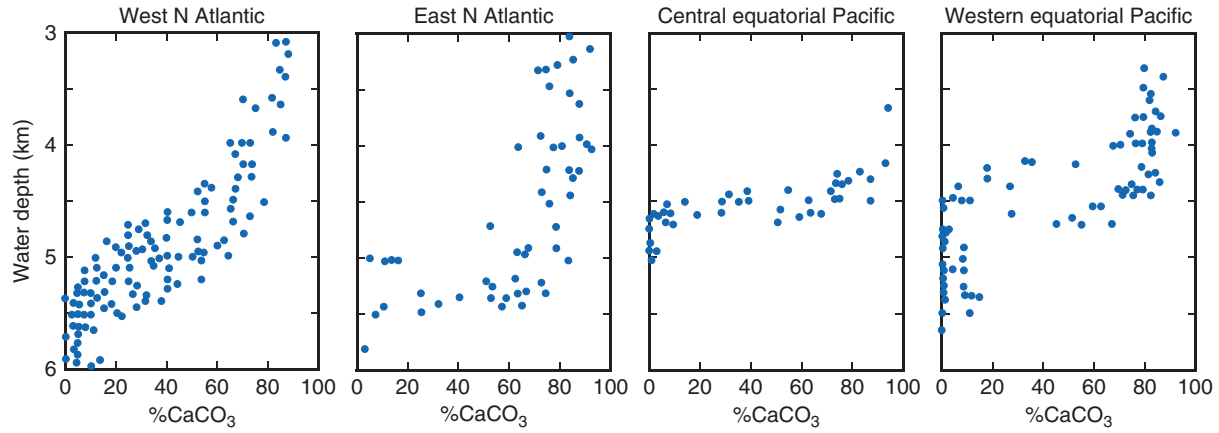


Figure 2 %CaCO₃ in core-top sediments from the Atlantic and Pacific Oceans. Carbonate is preserved to significantly deeper depths in the Atlantic. Reprinted from Broecker WS and Peng T-H (1982) *Tracers in the sea*. Palisades: Eldigio Press.

Table 2 Influence of salinity, temperature, and pressure on the dissociation constants of carbonic acid and the solubility products of calcite and aragonite. 1 dbar $\approx$ 1 m

T (C)	S	P (dbar)	pK_1^*	pK_2^*	pK_{sp}^* (calcite)	pK_{sp}^* (aragonite)
25	35	0	5.8563 ^a	8.9249 ^a	6.3692 ^b	6.1880 ^b
2	35	0	6.0860 ^a	9.3433 ^a	6.3667 ^b	6.1656 ^b
25	34	0	5.8594 ^a	8.9342 ^a	6.3854 ^b	6.2034 ^b
2	35	3000	5.9397 ^c	9.2536 ^c	6.1043 ^c	5.9189 ^c

^aDOE (1994).

^bMucci (1983).

^cPressure correction from Millero FJ (1995) Thermodynamics of the carbon dioxide system in the oceans. *Geochimica et Cosmochimica Acta*, 59: 661–677.

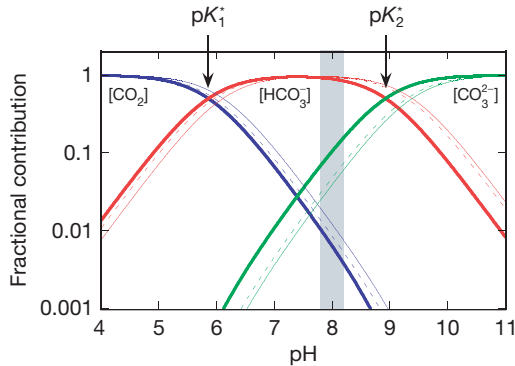


Figure 3 Bjerrum plot for the carbonate system in seawater for various P , T , S . Blue, red, and green represent [CO₂], [HCO₃⁻], and [CO₃²⁻], respectively. Heavy curves are for $S=35$, $T=25$ C, $P=0$ dbar. Narrow curves are for $S=35$, $T=0$ C, $P=0$ dbar, dashed curves are $S=35$, $T=0$ C, $P=3000$ dbar. Note that the pK^* values ($= -\log_{10} (K^*)$) for K_1^* and K_2^* are equal to pH when [HCO₃⁻] = [CO₂] and [CO₃²⁻] = [HCO₃⁻], respectively. Shaded area represents typical ocean pH values.

$$\text{TA} \approx \text{PA} = [\text{HCO}_3^-] + 2[\text{CO}_3^{2-}] + [\text{B}(\text{OH})_4^-] + [\text{OH}^-] - [\text{H}^+] \quad [15]$$

where PA is the practical alkalinity (alkalinity for most practical purposes). In fact, carbonate alkalinity (CA) is by far the most

important contributor to TA, which leads to a useful approximation:

$$\text{TA} \approx \text{CA} = [\text{HCO}_3^-] + 2[\text{CO}_3^{2-}] \quad [16]$$

Since [CO₂] is only a minor component of ΣCO_2 ,

$$\Sigma\text{CO}_2 \approx [\text{HCO}_3^-] + [\text{CO}_3^{2-}] \quad [17]$$

which leads to

$$[\text{CO}_3^{2-}] \approx \text{TA} - \Sigma\text{CO}_2 \quad [18]$$

Balance between DIC and TA

Alkalinity is delivered to the oceans by continental weathering and hydrothermal exchange and is lost mainly through sedimentary CaCO₃ burial and reef formation. Weathering or burial of CaCO₃ adds or removes TA and DIC in the ratio 2:1. Therefore, an increase in one or other of these terms will cause a change in [CO₃²⁻] (Eqn [18]) and therefore $p\text{CO}_2$ (see the Section ‘Changes in Carbonate Dissolution through Time’). Similarly, the manufacture or dissolution of CaCO₃ within the marine system causes a decrease or increase in TA and DIC in a 2:1 ratio. This means that the formation of CaCO₃ actually causes an increase in $p\text{CO}_2$ (Figure 5).

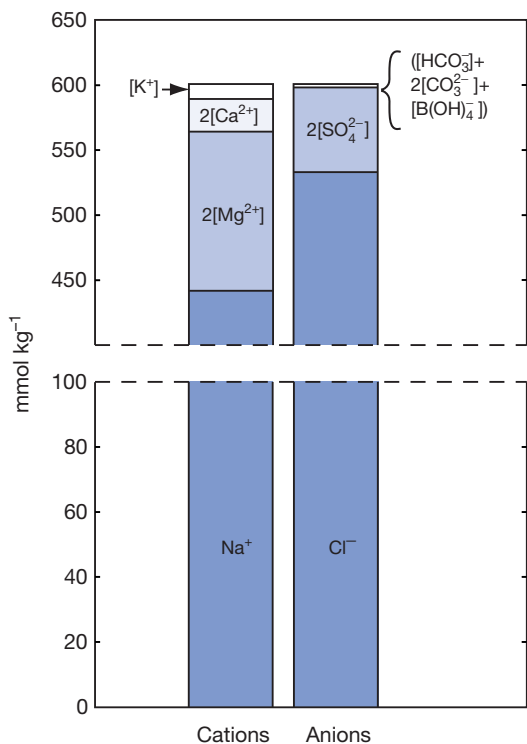


Figure 4 Charge balance for the major ions in seawater. The small excess positive charge of the conservative cations over conservative anions is balanced (mainly) by $[\text{HCO}_3^-]$, $[\text{CO}_3^{2-}]$ and $[\text{B}(\text{OH})_4^-]$ with minor contributions from $[\text{OH}^-]$ and $[\text{H}^+]$. Reprinted from Zeebe RE and Wolf-Gladrow D (2001) *CO₂ in Seawater: Equilibrium, Kinetics, Isotopes*. Amsterdam: Elsevier.

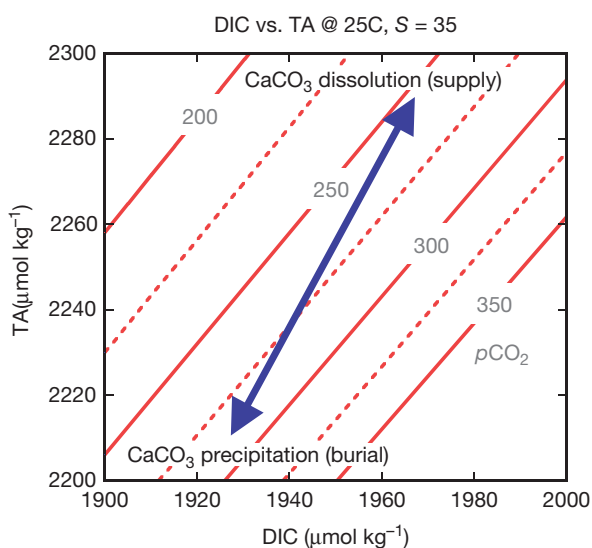


Figure 5 DIC versus TA for the formation and dissolution of CaCO_3 in seawater. Contours are constant $p\text{CO}_2$. Formation of CaCO_3 from its dissolved constituents causes a decrease in TA and DIC in the ratio 2 : 1 and a corresponding increase in $p\text{CO}_2$. The supply of CaCO_3 to the oceans from weathering (or its burial as sediments) also increases (or decreases) TA and DIC in the ratio 2 : 1.

Determining Carbonate Parameters

It is not possible to measure the individual species of the carbonate system directly. However, the four parameters, DIC, TA, $f\text{CO}_2$, and pH, can be determined. From a knowledge of any two of these parameters, along with temperature, salinity, and pressure, it is possible to calculate all six (ΣCO_2 , $[\text{CO}_3^{2-}]$, $[\text{HCO}_3^-]$, $[\text{CO}_2]$, pH, TA).

Mechanisms of Carbonate Dissolution

Thermodynamics

Efforts to understand the mechanisms of carbonate dissolution have involved many laboratory and field-based studies on both synthetic and biogenic carbonates. However, there remains considerable uncertainty regarding the specific parameters involved.

Equilibrium solution of calcium carbonate in seawater may be expressed as



The stoichiometric solubility product, K_{sp}^* is then defined as

$$K_{\text{sp}}^* = [\text{Ca}^{2+}]_{\text{sat}} [\text{CO}_3^{2-}]_{\text{sat}} \quad [20]$$

where the subscript 'sat' refers to the concentration at saturation. The tendency for CaCO_3 to dissolve in seawater depends on the degree of saturation (Ω) of the surrounding seawater with respect to the specific mineral phase in question (calcite or aragonite):

$$\Omega = [\text{Ca}^{2+}][\text{CO}_3^{2-}]/K_{\text{sp}}^* \quad [21]$$

where $[\text{Ca}^{2+}]$ and $[\text{CO}_3^{2-}]$ are the concentrations in solution. Since the concentration of Ca^{2+} in seawater is relatively constant, we can make the approximation

$$\Omega = [\text{CO}_3^{2-}]/[\text{CO}_3^{2-}]_{\text{sat}} \quad [22]$$

When $\Omega=1$, the system is at equilibrium and should be stable. $\Omega<1$ indicates that the solution is undersaturated with respect to carbonate and we should expect dissolution to occur. When $\Omega>1$, the solution is said to be supersaturated and, thermodynamically, we may expect precipitation of carbonate. However, due to complexities such as the composition of seawater, spontaneous precipitation is rare (see Morse and He, 1993). A widely used alternative for expressing the degree of saturation is the $\Delta[\text{CO}_3^{2-}]$ notation (Figure 8), where

$$\Delta[\text{CO}_3^{2-}] = [\text{CO}_3^{2-}] - [\text{CO}_3^{2-}]_{\text{sat}} \quad [23]$$

The value of K_{sp}^* depends on the particular mineral phase of interest and varies with temperature, salinity, and pressure such that solubility increases with decreasing temperature and increasing pressure (Figure 6; Table 2). For our concerns, pressure has the most important influence given the great depth of the ocean. The pressure dependence of carbonate solubility relates to the difference in volume occupied by the solid phase of CaCO_3 compared with the dissolved ions. The change in partial molar volume, ΔV , between solid and dissolved phases, where

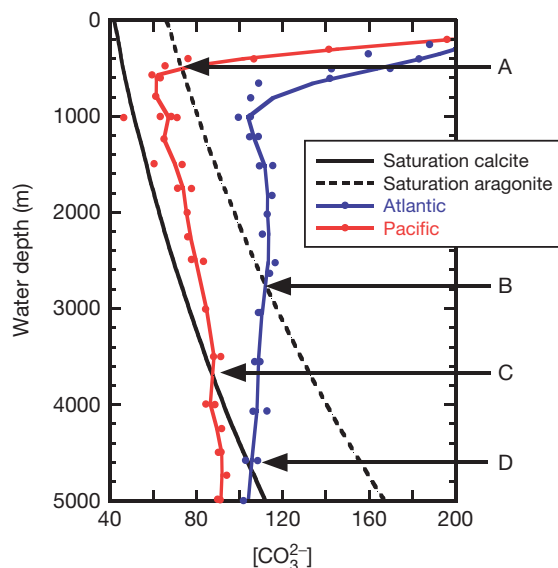


Figure 6 Water-column profiles of *in situ* $[\text{CO}_3^{2-}]$ from the Atlantic and Pacific Oceans plotted with the saturation $[\text{CO}_3^{2-}]$ profiles for aragonite and calcite. Saturation horizons for aragonite and calcite in the Pacific and Atlantic are labeled A–D, respectively. *In situ* data are from WOCE (sections A05 and P15N, available from <http://cdiac.ornl.gov>). Solubility coefficients are from Mucci (1983). Pressure correction from Millero (1995).

$$\Delta V = V_{\text{Ca}} + V_{\text{CO}_3} - V_{\text{CaCO}_3} \quad [24]$$

is negative for calcite and aragonite, that is, the volume occupied by Ca^{2+} and CO_3^{2-} is smaller than when they are combined to form CaCO_3 . This explains why solubility increases with increasing water depth. Aragonite is more soluble than calcite for any given T , S , and P . It is for this reason that the saturation horizon for aragonite is shallower than for calcite (Figure 6, Table 2).

Experimental determinations of K_{SP}^* at various T and S and 1 atm pressure suggest an uncertainty of $\sim 5\%$ for both calcite and aragonite (Mucci, 1983). Difficulties in determining accurate values for ΔV (a range of $\sim 35\text{--}45 \text{ cm}^3 \text{ mol}^{-1}$ for calcite, e.g., Ingle (1975)) lead to a total uncertainty in K_{SP}^* of $\sim 11\%$ at 4 km water depth. Accordingly, our ability to define the vertical profiles of $[\text{CO}_3^{2-}]_{\text{sat}}$ (Figure 6) are similarly uncertain. Including the error associated with determining *in situ* $[\text{CO}_3^{2-}]$, our uncertainty in calculating Ω becomes $\pm 12\%$. In all, this means that we cannot define the depth of the saturation horizon in the oceans to better than $\pm 0.90 \text{ km}$ (see Emerson and Hedges (2003) for a fuller description).

From purely thermodynamic considerations, we should expect the CaCO_3 content of marine sediments to drop from a maximum value to zero precisely where the water column becomes undersaturated. But as stated earlier, the actual transition from carbonate-rich to carbonate-poor sediments occurs over a zone spanning several hundred meters with undissolved carbonate being found considerably deeper than the water-column saturation horizon. It is therefore necessary to consider the rate of CaCO_3 dissolution.

Kinetics

If solid CaCO_3 is in contact with undersaturated water, it will tend to dissolve. The rate of dissolution ultimately depends on the degree of undersaturation and therefore both $[\text{CO}_3^{2-}]$ and K_{SP}^* need to be known accurately. The most commonly employed form of kinetic rate equation for the dissolution of CaCO_3 (Morse and Berner, 1972) is given by

$$R = k(1 - \Omega)^n \quad [25]$$

where R is the rate (usually $\% \text{ day}^{-1}$), k is the rate constant (same units as R), and n is the reaction order. Determination of k and n has proved to be extremely nontrivial due to the difficulties in extrapolating between laboratory results and the real ocean and the fact that CaCO_3 dissolution in seawater is the sum of several different chemical reaction pathways and involves the effects of impurities and surface coatings as well as multiple carbonate phases. A notable uncertainty in our understanding of carbonate dissolution kinetics is highlighted by the disparity between dissolution rates obtained in the laboratory with those predicted from field observations (which are smaller by 2–3 orders of magnitude). The most extensive laboratory-based measurements of carbonate dissolution were made by Keir (1980) using a variety of biogenic and synthetic carbonates at near-equilibrium conditions. Keir obtained a reaction order of $n=4.5$ for calcite and a very wide range of rate constants depending on the form of calcite tested, in many cases k being inversely correlated with grain size. Later work by Hales and Emerson (1997) suggested that uncertainties in Keir's calculation of Ω had led to a particularly high rate order. They reinterpreted his results in terms of first-order kinetics. Using a value of $n \sim 1\text{--}2$ reduced the associated range of rate constants and reduced discrepancies between various field observations. However, differences in rate constants determined in the laboratory compared with *in situ* field experiments remain unexplained.

Shape of the Sedimentary Lysocline

The transition from carbonate-rich to carbonate-poor sediments as illustrated in Figure 2 is the ultimate expression of carbonate dissolution in the deep sea. The shape and position of the transition zone is sensitive to deep-water carbonate chemistry, the flux of CaCO_3 and refractory material, and also the flux of organic carbon and its oxidation to CO_2 (see also Archer (1991)). Interbasin differences in the depth of the lysocline are mainly due to deep-water chemistry variations. For example, dissolution occurs at much shallower depths in the Pacific than in the Atlantic Ocean (Figures 2 and 6). This is a result of mixing water masses with differing $[\text{CO}_3^{2-}]$ (North Atlantic Deep Water, NADW, has relatively high $[\text{CO}_3^{2-}]$ compared with deep water produced in the Southern Ocean) as well as the constant addition of CO_2 (and corresponding decrease in $[\text{CO}_3^{2-}]$) to deep waters by the oxidation of raining organic carbon as they progress along the ocean conveyor.

The fact that carbonate may be preserved below the depth at which the water column becomes undersaturated (i.e., where $\Omega < 1$) implies that dissolution is sufficiently slow (relative to the flux arriving at the sea floor) to allow burial. Also, diffusion of ocean water through the sediments must be slow enough to

allow pore waters to reach equilibrium saturation (by addition of CO_3^{2-} through dissolution). Higher fluxes of CaCO_3 to the sediment will tend to deepen the lysocline relative to the saturation horizon. This is apparent under certain upwelling areas (Figure 1). In regions where productivity is low, the CCD will tend to approach the saturation horizon.

The proportion of noncarbonate material (e.g., clays) will also affect the shape of the transition zone. For example, the generally thicker lysocline of the Atlantic relative to the Pacific Ocean is thought to be a consequence of a higher flux of terrigenous material to Atlantic sediments (Archer, 1996). The effect is related to the sensitivity of $\%\text{CaCO}_3$ to dissolution (see the Section 'Quantifying the Extent of Dissolution'); a lower initial CaCO_3 content will mean greater sensitivity of $\%\text{CaCO}_3$ to a given degree of dissolution (Archer, 1991).

Supralysoclineal Dissolution

Besides carbonate preservation below the saturation horizon, significant dissolution may occur above this depth. This phenomenon is the result of pore waters becoming undersaturated with respect to CaCO_3 through the addition of metabolic CO_2 by oxidation of organic carbon within the sediment (cf. eqn [13]). An early modeling study by Emerson and Bender (1981) suggested that release of respiratory CO_2 within pore waters could cause up to 50% dissolution of CaCO_3 even at the saturation horizon although the actual number is probably somewhat lower, perhaps 20–40%. *In situ* microelectrode measurements of O_2 and pH confirm that dissolution of CaCO_3 within pore waters must be occurring in response to the release of CO_2 (e.g., Hales and Emerson, (1997); Figure 7). Further studies, using *in situ* benthic flux measurements have generally confirmed this conclusion although supersaturated settings

with a high initial $\%\text{CaCO}_3$ have proved to be equivocal. R. A. Jahnke and D. B. Jahnke (2004) have suggested that the low alkalinity fluxes observed at these sites may be the result of inorganic CaCO_3 precipitation in the supersaturated sediment surface, followed by CO_2 -driven redissolution deeper in the sediments. Hales (2003) argues that the observations can be reconciled using revised estimates for the rates of respiration and dissolution.

A consequence of pore-water dissolution is that loss of CaCO_3 will tend to be more intense under highly productive regions (such as continental margins) for a given bottom water saturation. This is somewhat opposed to the notion that increased CaCO_3 production will tend to deepen the lysocline. However, Archer (1991) has shown that it is the ratio of C_{org} to CaCO_3 reaching the sediments (the benthic rain ratio) which makes the most difference to the proportion of CaCO_3 dissolved. Increasing the benthic rain ratio causes a shoaling of the sedimentary lysocline, whereas increasing the absolute flux of CaCO_3 will generally cause a deepening of the lysocline (Figure 8).

Quantifying the Extent of Dissolution

Dissolution Proxies

The importance of deep-sea carbonate dissolution within the global carbon cycle is reflected by the enumerable attempts to reconstruct the history of dissolution on various timescales. Furthermore, because dissolution can affect other paleoceanographic proxies employed to reconstruct climatic variations, it is evermore important to be able to quantitatively assess the extent of dissolution at any given location at any given time. At this time there is no perfect dissolution proxy; each suffers from

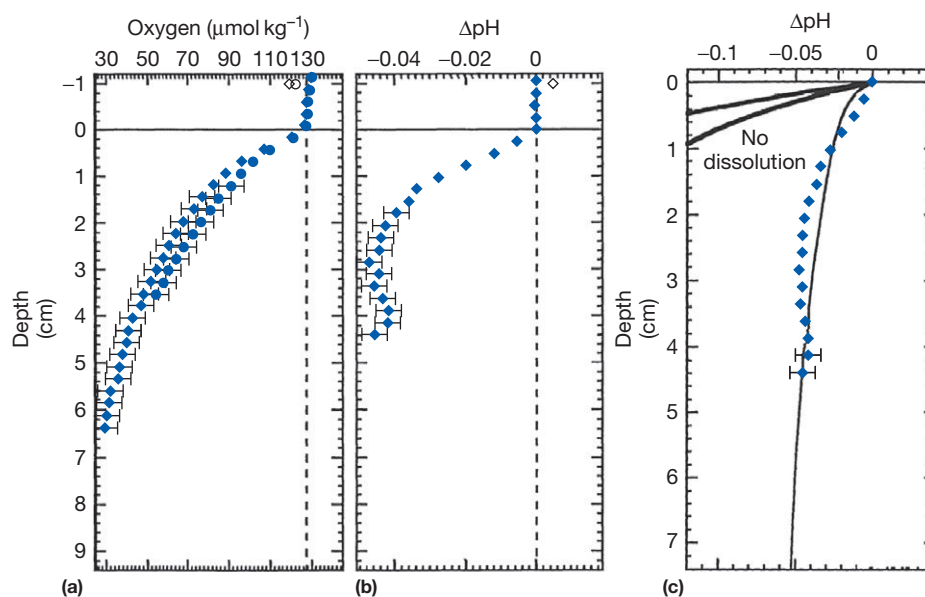


Figure 7 Pore-water profiles of (a) oxygen; (b) ΔpH ; (c) modeled ΔpH from 2,235 m water depth in the west equatorial Pacific (Ontong Java Plateau) from Hales and Emerson (1996). Measurements were made *in situ* using microelectrodes. Decreasing O_2 reflects the oxidation of organic matter and the production of respiration CO_2 . Heavy curves in (c) were computed assuming that respiration CO_2 does not cause dissolution of carbonate. Light curve is a best fit to the data. The observed pH values demand that dissolution occurs. Reprinted from Hales B and Emerson S (1996) Calcite dissolution in sediments of the Ontong-Java Plateau: *In situ* measurements of pore water O_2 and pH. *Global Biogeochemical Cycle*, 10: 527–541.

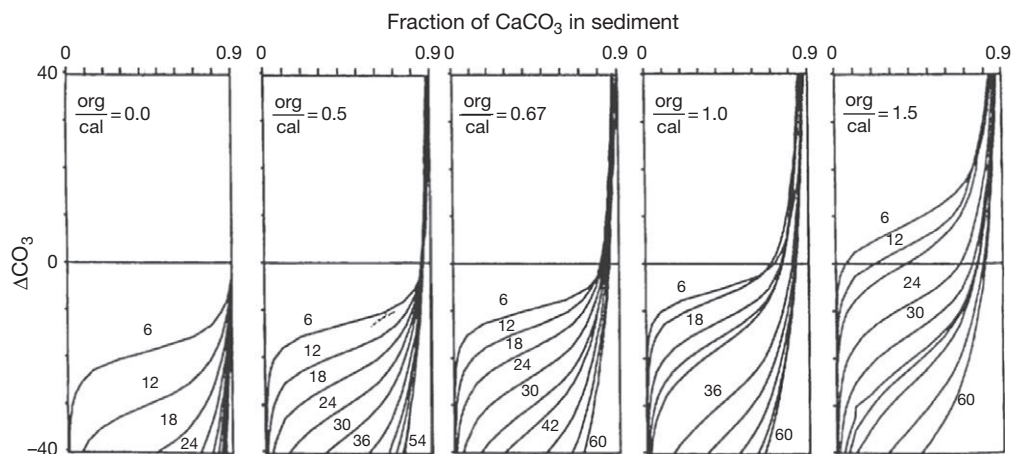


Figure 8 Results from Archer's (1991) sedimentary lysocline model. Curve numbers represent various CaCO_3 fluxes ($\text{micro-mol.cm}^{-2}.\text{yr}^{-1}$). Increasing the benthic rain ratio ($C_{\text{org}}/\text{CaCO}_3$) causes a shoaling of the lysocline, whereas increasing the absolute flux of CaCO_3 generally causes a deepening (these results are from Archer's oxic/anoxic model. Note results are plotted versus $\Delta[\text{CO}_3^{2-}]$ (cf. eqn [23]). Reprinted from Archer D (1991) Modeling the calcite lysocline. *Journal of Geophysical Research-Oceans* 96: 17037–17050.

particular difficulties, most commonly a sensitivity to changes in environmental conditions at the sea surface. It is wise therefore to consider more than one proxy at any given location. A further complication involves the offset in $[\text{CO}_3^{2-}]$ between bottom waters and pore waters. For example, in sediments above the water-column saturation horizon, dissolution can only occur under the influence of undersaturated pore waters. To infer changes in bottom-water $[\text{CO}_3^{2-}]$ would therefore require, at the very least, a knowledge of the contemporary benthic flux ratio of $C_{\text{org}}/\text{CaCO}_3$. (see **Carbon Cycle Proxies** ($\delta^{11}\text{B}$, $\delta^{13}\text{C}_{\text{calcite}}$, $\delta^{13}\text{C}_{\text{organic}}$, **Shell Weights**, B/Ca , U/Ca , Zn/Ca , Ba/Ca)).

Carbonate content (% CaCO_3)

The most apparent effect of dissolution in deep-sea sediments is the decrease in carbonate content with increasing water depth (Figure 2). However, the relationship between % CaCO_3 and the extent of dissolution is not simple and as a result % CaCO_3 is generally not considered a reliable indicator of dissolution. As described above, the carbonate content of a particular sediment is a function of the production and dissolution of CaCO_3 and dilution by noncarbonate material. Additionally, % CaCO_3 is fairly insensitive to the degree of dissolution until this becomes significant. For example, dissolving 50% of the carbonate from a sediment originally containing 90% CaCO_3 would only decrease this value by 8% (Figure 9).

Changes in species composition

Dissolution causes the thinning and breakup of Foraminifera tests and coccoliths. Different species are more or less susceptible to the effects of dissolution depending on the initial thickness of their tests as well as more cryptic differences such as crystal habit or chemical composition. These differences lead to changes in the species composition (faunal or floral assemblage) as dissolution proceeds and can be correlated with the extent of dissolution. Many variations of this type of proxy have been developed. As well as Berger's (1970) classic solubility ranking of planktonic Foraminifera species (Table 3),

other workers have considered the ratio of benthic to planktonic Foraminifera (benthic or bottom-dwelling species tending to be more resistant to dissolution) or pteropods to Foraminifera as well as coccolith assemblages. A major complication to this approach is the environmental control on initial species composition. Planktonic foraminifera species assemblages are very sensitive to the prevailing environmental conditions at the sea surface and are in fact widely used to reconstruct changes in, for example, sea-surface temperatures (SSTs) through time. This complication makes interpretation in terms of dissolution subject to significant uncertainty.

Fragmentation indices

The extent of breakup or fragmentation of Foraminifera tests during dissolution can be used to assess the overall preservation state of the sediment. This approach has been applied extensively to reconstruct dissolution intensity in a wide variety of settings. A potential problem with fragmentation indices is their nonlinearity; since a single shell can break into a number of fragments, a simple ratio of fragments to total entities tends to be oversensitive to changes at the onset of dissolution and much less sensitive as dissolution proceeds (Figure 10). This matter was addressed by Le and Shackleton (1992), who suggested taking the average number of fragments per whole shell into account.

$$\% \text{Fragment} = 100 \times (\# \text{fragments}/x) / \{(\# \text{fragments}/x) + \# \text{whole shells}\} \quad [26]$$

where x is the average number of fragments that a typical Foraminifera shell will break into (Figure 10). Another variant is to compare the number of fragments to whole shells of a single species (e.g., Mekik *et al.* (2002)).

Grain-size indices

Grain-size indices are based on the fact that progressive dissolution causes a decrease in the average grain size of a given sediment packet as entities such as Foraminifera tests break up. A typical example of this type of indicator is the proportion of coarse fraction to total sediment (e.g., % >150 μm). Grain-size

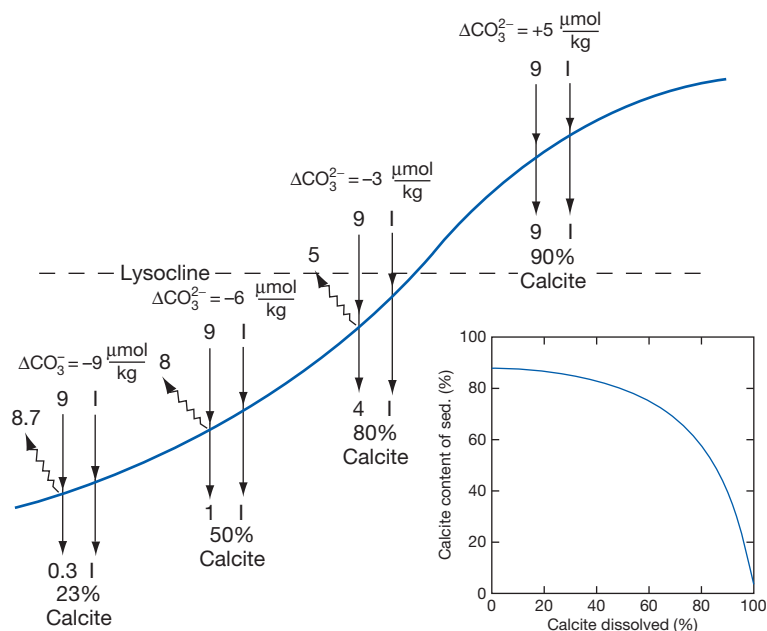


Figure 9 Illustration of the insensitivity of %CaCO₃ to dissolution until the fraction dissolved exceeds ~50%. Note in this case initial carbonate content is 90%. A lower initial value would result in increased sensitivity to dissolution. Straight arrows represent fluxes of carbonate and noncarbonate to the sediment and burial after dissolution (indicated by wavy arrows). Reprinted from Broecker WS and Peng T-H (1982) *Tracers in the Sea*. Palisades: Eldigio Press.

Table 3 Berger's (1970) solubility index of planktonic Foraminifera (rank 1 is most soluble)

Low resistance	High resistance
1. <i>Globigerinoides ruber</i>	12. <i>Globorotalia hirsuta</i>
2. <i>Orbulina universa</i>	13. <i>Globorotalia truncatulinoides</i>
3. <i>Globigerinella siphonifera</i>	14. <i>Globorotalia inflata</i>
4. <i>Globigerina rubescens</i>	15. <i>Globorotalia cultrata</i>
5. <i>Globigerinoides sacculifer</i>	16. <i>Globorotalia dutertrei</i>
6. <i>Globigerinoides tenellus</i>	17. <i>Globigerina pachyderma</i> , s.l.
7. <i>Globigerinoides conglobatus</i>	18. <i>Pulleniatina obliquiloculata</i>
8. <i>Globigerina bulloides</i>	19. <i>Globorotalia crassiformis</i>
9. <i>Globigerina quinqueloba</i>	20. <i>Sphaeroidinella dehiscens</i>
10. <i>Globigerinita glutinata</i>	21. <i>Globorotalia tumida</i>
11. <i>Candela nitida</i>	22. <i>Turborotalia humulis</i>

From Berger WH (1970) Planktonic foraminifera: Selective solution and the lysocline. *Marine Geology* 8: 111–138.

indices are sensitive to the initial particle size distribution, for example, coccoliths versus Foraminifera, and are therefore subject to environmental change. They may also suffer from a lack of sensitivity at the onset of dissolution.

Foraminiferal shell weights

The individual weights of Foraminifera tests within a narrow size range have been correlated to bottom water [CO₃²⁻] and used to infer changes in this parameter (related to CaCO₃ dissolution, cf. Eqn [21]) through time (e.g., Broecker and Clark (2001)). Again, this proxy is subject to variations in initial shell weight, which may result from changing conditions during growth (e.g., Barker and Elderfield (2002)).

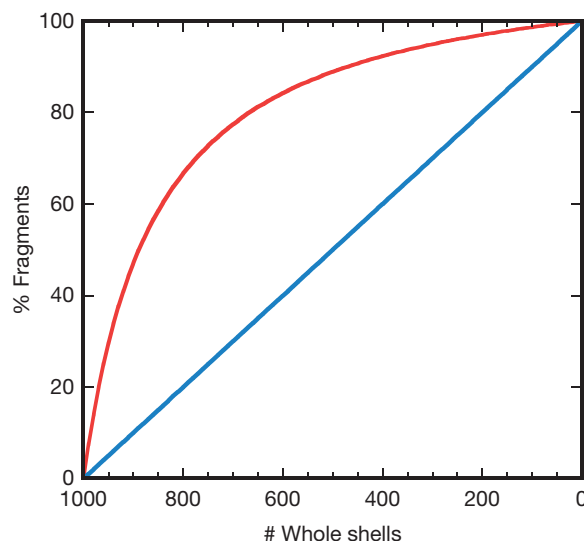


Figure 10 Dissolution causes progressive fragmentation of foraminiferal shells. Each shell will produce several fragments, leading to nonlinearity of a straightforward measure of %fragmentation (red curve). Le and Shackleton (1992) accounted for this by including a term for average number of fragments per whole shell into their fragmentation index (cf. Eqn [26], blue curve).

Observational indices

Changes in the ultrastructure of Foraminifera tests, caused by progressive dissolution, can be observed using scanning electron microscopy (SEM) (Figure 11). Various studies have employed this technique in an attempt to reconstruct the history of dissolution in different locations although difficulties

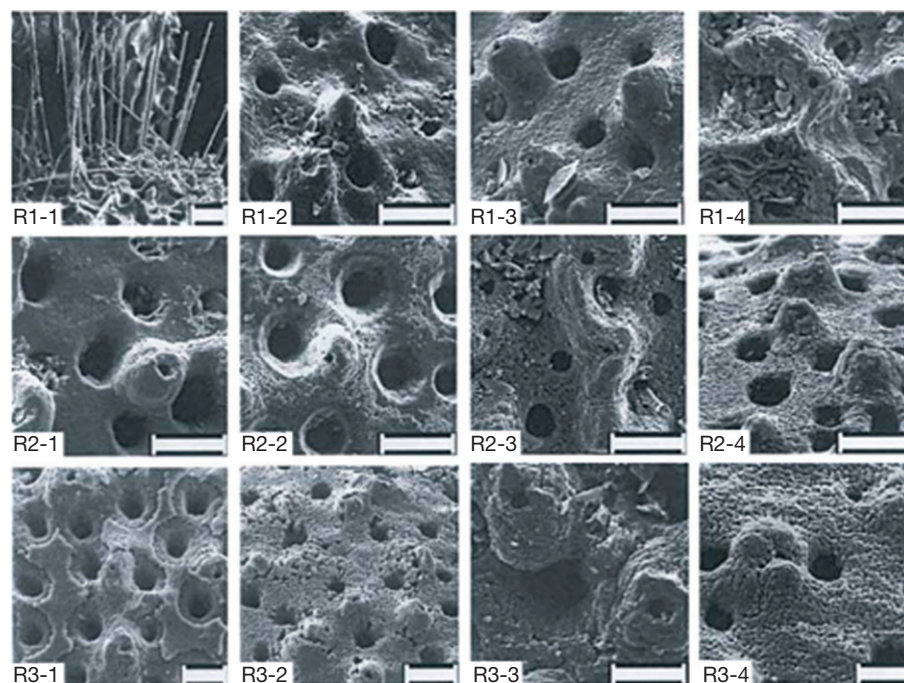


Figure 11 Scanning electron microscopy (SEM) can be used to assess the preservation state of carbonate sediments. In this case, numbers 1–4 show progressive dissolution of tests of *Globigerina bulloides*. Reprinted from Dittert et al. (2000).

may arise from changes in initial shell-wall structure as well as the inherent subjectivity of the method (this is also a problem with many of the other proxies discussed here).

Crystallinity

A relatively unexplored indicator of carbonate dissolution is the so-called crystallinity of foraminiferal calcite. Crystallinity, defined as the peak width (at half maximum height) of the (104) calcite X-ray diffraction peak, is essentially a measure of how perfect the calcite crystal lattice is. As dissolution proceeds, more poorly crystallized calcite is thought to be removed, causing a narrowing of the (104) diffraction peak. Crystallinity has been applied to core-top sediments and may, in the future, prove to be a reliable dissolution indicator, providing environmental conditions do not play a major role (Bassinot et al., 2004).

Composite indices

In order to circumvent the various difficulties associated with many of the individual proxies for carbonate dissolution, several workers have adopted so-called composite dissolution indices (CDIs). Peterson and Prell (1985) used a combination of six different measurements: %CaCO₃, a size index (>63 µm), three fragmentation indices (% whole Foraminifera, ratio of benthic to planktonic Foraminifera, and % whole *Globorotalia menardii*) and % radiolarians (ratio of radiolarians to radiolarians + whole planktonic Foraminifera).

Chemical indices

As well as indicators based on the physical effects of dissolution, various chemical proxies have been developed with the aim of directly reconstructing deep-sea carbonate system

parameters. For example, elemental ratios, such as Zn/Ca and Cd/Ca measured on benthic Foraminifera, have been applied to assess changes in deep ocean [CO₃²⁻] (e.g., Marchitto et al. (2005)) whereas the boron isotopic composition (δ¹¹B) of the same organisms has the potential for quantifying changes in surface and deep-water pH (e.g., Sanyal et al. (1995)).

Effect of Dissolution on Other Paleocceanographic Proxies

Many of the proxies employed by paleoceanographers utilize foraminiferal calcite. It has been widely shown that many of these proxies are actually susceptible to the effects of dissolution. For example, planktonic foraminiferal Mg/Ca ratios (used as a proxy for temperature) are known to decrease with progressive dissolution of the test. The actual mechanism is poorly understood but probably reflects the chemical heterogeneity of foraminiferal calcite. The effects of dissolution on Mg/Ca-derived temperatures can be significant and as such it is important to assess the state of sediment preservation before interpretation of results. Other foram-based proxies that may be affected by dissolution include δ¹¹B, δ¹⁸O, and faunal assemblages (see above).

Changes in Carbonate Dissolution through Time

From the relations described in Sections ‘Thermodynamics of the carbonate system in seawater’ and ‘Mechanisms of Carbonate Dissolution,’ it is possible to understand the importance of carbonate dissolution within the broader context of the global carbon cycle, particularly with respect to atmospheric CO₂. Our understanding of atmospheric CO₂ changes

in the past as well as the consequences of future changes depends on our understanding of their interaction with marine carbonate chemistry and CaCO_3 dissolution.

Carbonate Budgets and Carbonate Compensation

The marine carbonate cycle involves the balance between the production of CaCO_3 (mainly through biogenic processes) from its dissolved constituents, the delivery of those dissolved ingredients to the ocean from continental weathering or hydrothermal exchange, and the burial of solid CaCO_3 as sediment or reefs. The actual rates of carbonate production, burial, and the influx from weathering are sources of significant uncertainty, as discussed by Milliman and Droxler (1996) (Table 4). However, it is clear that the global production rate of carbonate is significantly greater than the rate of supply from weathering (by a factor of ~ 4 or more). This overproduction of carbonate is (approximately) balanced by dissolution; the depth of the lysocline is such that it intersects and dissolves approximately the correct proportion of sinking CaCO_3 to maintain the balance between supply and burial. As mentioned in the Section ‘Thermodynamics of the Carbonate System in Seawater,’ CaCO_3 weathering (or burial) adds (or removes) alkalinity and DIC in a ratio of 2 : 1. From Eqn [18], we see, for example, that an increase in CaCO_3 weathering will cause an increase in $[\text{CO}_3^{2-}]$ and deepening of the lysocline such that a smaller proportion of raining CaCO_3 will be dissolved, thereby regaining the balance between supply and burial. This process is known as carbonate compensation and has a timescale of adjustment of the order 4–5 ka (Broecker and Peng, 1982). Thus, changes in CaCO_3 dissolution through time may be linked with changes in weathering, the rate of carbonate production, or changes in the locus of CaCO_3 deposition (see below). (The present discussion concerns timescales of thousands of years. On longer (Myr) timescales, the control of weathering on CO_2 is rather different, being dependent on the rate of silicate weathering alone and not on the imbalance between weathering and burial rates (see Berner (2004) for a full description).) Importantly, these processes are also linked to atmospheric CO_2 by the inverse relation between CO_2 and $[\text{CO}_3^{2-}]$ (Eqn [12]).

Table 4 Global estimates for the rates of supply, production, and burial of CaCO_3 (units are 10^{12} mol CaCO_3 per year). Uncertainties are probably less than a factor of 1 for most values

Inputs	Production	Outputs (burial)
Rivers (~ 13)	Coral reefs, shelves (~ 25)	Neritic (~ 15)
Hydrothermal (~ 3 – 12)	Pelagic (inc. slopes) (~ 65)	Deep-sea (~ 17)
Groundwater (≤ 5)		
Total ~ 21 – 30	~ 90	~ 32

Data compiled from Milliman JD and Droxler AW (1996) Neritic and pelagic carbonate sedimentation in the marine environment: Ignorance is not bliss. *Geologische Rundschau* 85: 496–504.

Glacial–Interglacial Changes in Atmospheric CO_2

Ice-core records tell us that atmospheric CO_2 was approximately 30% lower during the Last Glacial Maximum (LGM) than the preindustrial Holocene (see CO_2 Studies). They also tell us that CO_2 has varied in tandem with the late Pleistocene glacial cycles, each cycle comprising a stepwise decline from an interglacial high of about 280 ppmv to a glacial low of about 190 ppmv followed by a more rapid increase to interglacial values during glacial terminations (Petit *et al.*, 1999). The reason for lowered atmospheric CO_2 during glacial times has been debated for more than quarter of a century and is still the subject of enquiry. However, since the oceans contain some 50–60 times more carbon than the atmosphere (see the Section ‘Thermodynamics of the Carbonate System in Seawater’), it is assumed that they must play a major role in the drawdown of CO_2 during glacial times. Currently favored mechanisms involve variations in vertical stratification and air–sea exchange in the Southern Ocean, enhanced biological activity under the influence of Fe fertilization, and changes in the dominant plankton type (see Sigman and Boyle (2000) for a review and additional references). However, any mechanism will ultimately involve interaction with the marine carbonate cycle. A few examples of mechanisms involving changing carbonate dissolution patterns are described below.

Records of % CaCO_3 from the deep Pacific Ocean show cycles of increased glacial CaCO_3 content and decreased % CaCO_3 during interglacials (Figure 12). There has been debate over whether the Pacific CaCO_3 cycles represent changes in dissolution intensity or surface ocean CaCO_3 production although dissolution is the generally favored explanation. The observation of a deeper calcite lysocline during glacial times (Figure 12) suggests that deep ocean $[\text{CO}_3^{2-}]$ was higher and could help to explain the decrease in CO_2 .

In an attempt to explain the glacial decrease in CO_2 , Berger (1982) suggested that rising sea levels associated with glacial terminations would increase the accommodation space for coral reef communities. Since coral reefs and shallow carbonate deposition account for almost half of all modern CaCO_3 burial (Table 4; Milliman and Droxler, 1996), an increase in this mode of deposition at the end of glacial times would cause global CaCO_3 burial to exceed supply, leading to a decrease in whole ocean $[\text{CO}_3^{2-}]$ and an increase in atmospheric CO_2 . This would in turn lead to a shoaling of the lysocline through compensation and a shift in the locus of carbonate deposition from the deep sea to more shallow waters. However, the mean change in lysocline depth for the glacial period is not sufficient to explain the 30% reduction in CO_2 via this mechanism. CaCO_3 dissolution actually increased in the deep Atlantic as a consequence of low $[\text{CO}_3^{2-}]$ Antarctic Bottom Water (AABW) invading the glacial deep Atlantic (Barker *et al.*, 2004). In fact, an estimate of global deep-sea CaCO_3 accumulation during the last glacial period suggests that it was probably very similar to the modern day (Catubig *et al.*, 1998). Furthermore, since the deglacial rise in CO_2 preceded the rise in sea level, it cannot have been driven mainly by changes in shallow-water accommodation space.

Another suggestion was made by Archer and Maier-Reimer (1994). This involved the influence of pore-water CaCO_3 dissolution and the benthic rain ratio of C_{org} to CaCO_3 (see

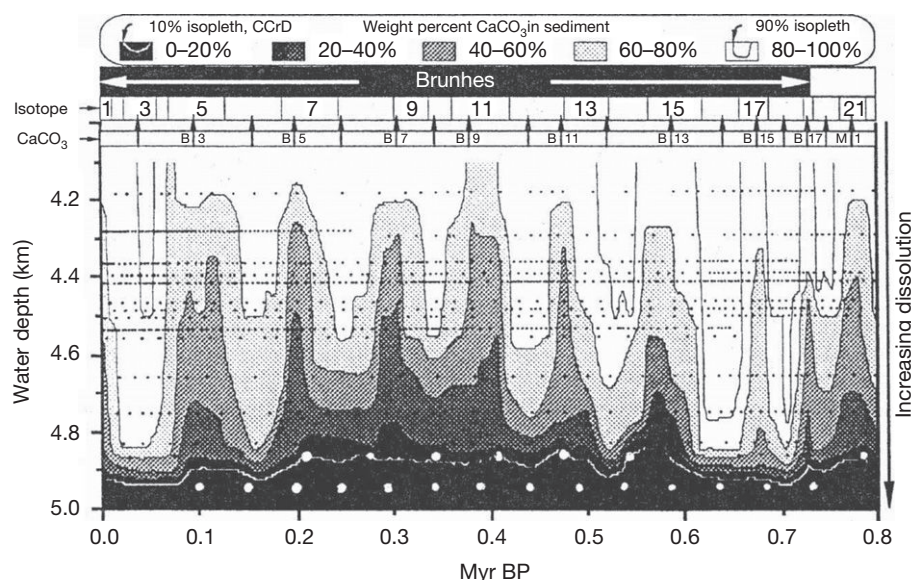


Figure 12 Reconstructed sedimentary lysocline in the equatorial Pacific made by combining several records of %CaCO₃ from different water depths (Farrell and Prell, 1989). Glacial periods are characterized by a deepened and narrower transition zone suggesting better carbonate preservation. Reprinted from Farrell BF and Prell W (1989) Climatic change and CaCO₃ preservation; an 800,000 year bathymetric reconstruction from the Central Equatorial Pacific Ocean. *Paleoceanography* 4: 447–466.

the Section ‘Shape of the Sedimentary Lysocline’. By increasing the production ratio of organic carbon to inorganic carbonate (e.g., by increasing the ratio of diatom to coccolithophore production), Archer and Maier-Reimer argued that increased supralysoclineal dissolution would cause carbonate supply to exceed burial, thereby raising [CO₃²⁻] and lowering CO₂. In this scenario, the depth of the sedimentary lysocline would not need to change by very much since it would shoal relative to a deepened saturation horizon. Although the ‘rain-ratio’ hypothesis escapes some of the problems associated with the coral reef hypothesis, it has not yet passed the rigors of validation (e.g., there are difficulties associated with the width of the predicted transition zone (Sigman *et al.*, 1998)). Thus the search for a viable explanation for the glacial lowering of atmospheric CO₂ continues. In all likelihood, it will comprise a combination of several mechanisms rather than a single process.

Carbonate Dissolution in the Future: Sequestration of Anthropogenic CO₂

Atmospheric CO₂ has increased by around 90 ppmv (~30%) since the start of the industrial revolution. However, this increase reflects only about half of the total CO₂ added; most of the rest has been taken up by the oceans. The oceans provide a buffer to increasing CO₂ by reaction with CO₃²⁻, thereby lowering the saturation state. This has consequences for marine life in the surface ocean but will also result in deep-sea CaCO₃ dissolution. In fact, much of the CO₂ released by fossil fuel burning will ultimately be neutralized by CaCO₃ dissolution in the deep sea. Estimates of the inventory of deep-sea CaCO₃ available for CO₂ neutralization are of the order 2000–5000 Gt C (Archer *et al.*, 1997). These can be compared to a global

inventory of fossil fuel reserves of ~5000 Gt C (Sundquist, 1985). Modeling studies (Archer *et al.*, 1997, 1998) suggest that on timescales of hundreds of years, 70–80% of anthropogenic CO₂ will dissolve into the oceans. Neutralization by CaCO₃ dissolution will occur on timescales of thousands of years and will eventually account for 60–70% of total CO₂ emissions.

See also: Ice Core Methods: CO₂ Studies; Overview.
Paleoceanography: Paleoceanography An Overview.
Paleoceanography, Biological Proxies: Benthic Foraminifera; Coccolithophores; Planktic Foraminifera. **Paleoceanography, Physical and Chemical Proxies:** Carbon Cycle Proxies ($\delta^{13}\text{C}$, $\delta^{13}\text{C}_{\text{calcite}}$, $\delta^{13}\text{C}_{\text{organic}}$, Shell Weights, B/Ca, U/Ca, Zn/Ca, Ba/Ca).

References

- Archer D (1991) Modeling the calcite lysocline. *Journal of Geophysical Research-Oceans* 96: 17037–17050.
- Archer D (1996) A data-driven model of the global calcite lysocline. *Global Biogeochemical Cycle* 10: 511–526.
- Archer D, Kheshgi H, and Maier-Reimer E (1997) Multiple timescales for neutralization of fossil fuel CO₂. *Geophysical Research Letters* 24: 405–408.
- Archer D, Kheshgi H, and Maier-Reimer E (1998) Dynamics of fossil fuel CO₂ neutralization by marine CaCO₃. *Global Biogeochemical Cycle* 12: 259–276.
- Archer D and Maier-Reimer E (1994) Effect of deep-sea sedimentary calcite preservation on atmospheric CO₂ concentration. *Nature* 367: 260–263.
- Barker S and Elderfield H (2002) Foraminiferal calcification response to glacial–interglacial changes in atmospheric CO₂. *Science* 297: 833–836.
- Barker S, Kiefer T, and Elderfield H (2004) Temporal changes in North Atlantic circulation constrained by planktonic foraminiferal shell weights. *Paleoceanography* 19: <http://dx.doi.org/10.1029/2004PA001004>.
- Bassinot F, Melieres F, Gehlen M, Levi C, and Labeyrie L (2004) Crystallinity of foraminifer shells: A proxy to reconstruct past bottom water CO₃ changes? *Geochemistry Geophysics Geosystems* 5: <http://dx.doi.org/10.1029/2003GC000668>.

- Berger WH (1970) Planktonic foraminifera: Selective solution and the lysocline. *Marine Geology* 8: 111–138.
- Berger WH (1982) Increase of carbon-dioxide in the atmosphere during Deglaciation – the coral-reef hypothesis. *Naturwissenschaften* 69: 87–88.
- Berner A (2004) *The Phanerozoic carbon cycle: CO₂ and O₂*, p. 158. New York: Oxford University Press.
- Broecker WS and Clark E (2001) Glacial to Holocene redistribution of carbonate ion in the deep sea. *Science* 294: 2152–2155.
- Broecker WS and Peng T-H (1982) *Tracers in the Sea*. Palisades: Eldigio Press.
- Catubig NR, Archer DE, Francois R, deMenocal P, Howard W, and Yu EF (1998) Global deep-sea burial rate of calcium carbonate during the last glacial maximum. *Paleoceanography* 13: 298–310.
- Dittert N and Henrich R (2000) Carbonate dissolution in the South Atlantic Ocean: Evidence from ultrastructure breakdown in *Globigerina bulloides*. *Deep-Sea Research Part I* 47: 603–620.
- DOE (1994) Dickson AG and Goyet C (eds.) In: *Handbook of Methods for the Analysis of the Various Parameters of the Carbon dioxide System in Sea Water, version 2*, (ORNL/CDIAC-74).
- Emerson S and Bender M (1981) Carbon fluxes at the sediment-water interface of the deep sea: Calcium carbonate preservation. *Journal Marine Research* 39: 139–162.
- Emerson S and Hedges JI (2003) Sediment diagenesis and benthic flux. In: Elderfield H (ed.) *Treatise on Geochemistry: The Oceans and Marine Geochemistry*, vol. 6, pp. 293–319. Amsterdam: Elsevier.
- Farrell BF and Prell W (1989) Climatic change and CaCO₃ preservation; an 800,000 year bathymetric reconstruction from the Central Equatorial Pacific Ocean. *Paleoceanography* 4: 447–466.
- Hales B (2003) Respiration, dissolution, and the lysocline. *Paleoceanography* 18.
- Hales B and Emerson S (1996) Calcite dissolution in sediments of the Ontong-Java Plateau: *In situ* measurements of pore water O₂ and pH. *Global Biogeochemical Cycle* 10: 527–541.
- Hales B and Emerson S (1997) Evidence in support of first-order dissolution kinetics of calcite in seawater. *Earth and Planetary Science Letter* 148: 317–327.
- Ingle SE (1975) Solubility of calcite in the ocean. *Marine Chemistry* 3: 301–319.
- Jahnke RA and Jahnke DB (2004) Calcium carbonate dissolution in deep sea sediments: Reconciling microelectrode, pore water and benthic flux chamber results. *Geochimica et Cosmochimica Acta* 68: 47–59.
- Keir RS (1980) Dissolution kinetics of biogenic calcium carbonates in seawater. *Geochimica et Cosmochimica Acta* 44: 241–252.
- Le J and Shackleton NJ (1992) Carbonate dissolution fluctuations in the western Equatorial Pacific during the late Quaternary. *Paleoceanography* 7: 21–42.
- Marchitto TM, Lynch-Stieglitz J, and Hemming SR (2005) Deep Pacific CaCO₃ compensation and glacial–interglacial atmospheric CO₂. *Earth and Planetary Science Letter* 231: 317–336.
- Mekik FA, Loubere PW, and Archer DE (2002) Organic carbon flux and organic carbon to calcite flux ratio recorded in deep-sea carbonates: Demonstration and a new proxy. *Global Biogeochemical Cycle* 16: (art. no.-1052).
- Millero FJ (1995) Thermodynamics of the carbon dioxide system in the oceans. *Geochimica et Cosmochimica Acta* 59: 661–677.
- Milliman JD and Droxler AW (1996) Neritic and pelagic carbonate sedimentation in the marine environment: Ignorance is not bliss. *Geologische Rundschau* 85: 496–504.
- Morse JW and Berner RA (1972) Dissolution kinetics of calcium-carbonate in sea-water 2: Kinetic origin for Lysocline. *American Journal Science* 272: 840.
- Morse JW and He SL (1993) Influences of T, S and P(CO₂) on the pseudo-homogeneous precipitation of CaCO₃ from seawater-implications for Whiting formation. *Marine Chemistry* 41: 291–297.
- Mucci A (1983) The solubility of Calcite and Aragonite in seawater at various salinities, temperatures, and one atmosphere total pressure. *American Journal of Science* 283: 780–799.
- Peterson LC and Prell WL (1985) Carbonate preservation and rates of climatic change: An 800 kyr record from the Indian Ocean. In: Sundquist E and Broecker WS (eds.) *The Carbon Cycle and Atmospheric CO₂: Natural Variations Archean to Present*. *Geophysical Monograph Series*, vol. 32, pp. 251–269. Washington, DC: AGU.
- Petit JR, Jouzel J, Raynaud D, et al. (1999) Climate and atmospheric history of the past 420,000 years from the Vostok ice core Antarctica. *Nature* 399: 429–436.
- Sanyal A, Hemming NG, Hanson GN, and Broecker WS (1995) Evidence for a higher pH in the Glacial ocean from boron isotopes in foraminifera. *Nature* 373: 234–236.
- Sigman DM and Boyle EA (2000) Glacial/interglacial variations in atmospheric carbon dioxide. *Nature* 407: 859–869.
- Sigman DM, McCorkle DC, and Martin WR (1998) The calcite lysocline as a constraint on glacial/interglacial low-latitude production changes. *Global Biogeochemical Cycle* 12: 409–427.
- Sundquist E (1985) Geological perspectives on carbon dioxide and the carbon cycle. In: Sundquist E and Broecker WS (eds.) *The Carbon Cycle and Atmospheric CO₂: Natural Variations, Archean to Present*, pp. 5–59. Washington, D. C: AGU.
- Thurman HV and Trujillo AP (2002) *Essentials of Oceanography*, 7th ed. Upper Saddle River: Prentice-Hall.
- Zeebe RE and Wolf-Gladrow D (2001) *CO₂ in Seawater: Equilibrium, Kinetics, Isotopes*. Amsterdam: Elsevier.

Relevant Website

<http://cdiac.ornl.gov> – Carbon Dioxide Information Analysis Center.

Mg/Ca and Sr/Ca Paleothermometry from Calcareous Marine Fossils

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Introduction

In the ocean, the distribution of sea surface and bottom water temperature (BWT) and salinity is perhaps the best representation of the state of the climate system, as the ocean plays a fundamental role in the evolution of Earth's climate. Therefore, determining the past temperature evolution of the ocean is key for understanding Earth's climate history. Among the most often used elemental proxies of paleotemperature are Mg/Ca in planktonic Foraminifera and Sr/Ca in corals as recorders of sea surface temperature (SST) and Mg/Ca in benthic Foraminifera and ostracodes as proxies of BWT. Mg, Sr, and Ca have relatively long oceanic residence times, about 13 My for Mg, 5 My for Sr, and 1 My for Ca, implying nearly constant Mg-, and Sr- to- Ca ratios in seawater on timescales of less than a million years. A clear advantage of these carbonate-based thermometers is that coupling $\delta^{18}\text{O}$ and Mg/Ca measurements in Foraminifera, or Sr/Ca in corals, potentially provides a novel way to adjust for the temperature-dependency of $\delta^{18}\text{O}$ in CaCO_3 minerals and isolate the record of seawater $\delta^{18}\text{O}$ ($\delta^{18}\text{O}_{\text{seawater}}$; Mashiotto et al., 1999), which can then be used to reconstruct local changes in evaporation–precipitation (and by inference salinity; e.g., Lea et al., 2000), advection of water with different seawater oxygen isotopic composition ($\delta^{18}\text{O}_{\text{seawater}}$) and provide valuable information about changes in continental ice volume (e.g., Lear et al., 2000).

Thermodynamic Effects on Mg and Sr Coprecipitation in Carbonates

Among the alkaline earth elements, Mg^{2+} and Sr^{2+} form the most important solid solutions with carbonates. Because of the difference in the ionic radii between Sr^{2+} (large) and Mg^{2+} (small), SrCO_3 is isostructural with the aragonite crystal cell (orthorhombic), whereas MgCO_3 is isostructural with calcite (rhombohedral). The Mg/Ca and Sr/Ca ratios in CaCO_3 minerals depend on two factors: the Mg/Ca and Sr/Ca activity ratios of ocean water, and the distribution coefficients of Mg/Ca and Sr/Ca between calcite, aragonite, and seawater, respectively. These relationships are expressed as $D_{\text{El}} = (\text{El}/\text{Ca})_{\text{mineral}}/(\text{El}/\text{Ca})_{\text{seawater}}$, where D_{El} is the empirical homogeneous distribution coefficient calculated based on the molar concentration ratios of Mg/Ca and Sr/Ca in calcite, aragonite, and seawater, respectively. At equilibrium, the partitioning constant between the two pure mineral phases depends on temperature, as the substitution of Mg into calcite and Sr into aragonite, are both associated with a change in enthalpy or heat of the reaction, which is sensitive to temperature. As the substitution of Mg into calcite is an endothermic reaction, the Mg/Ca ratio of calcite is expected to increase with increasing temperature (Rosenthal et al., 1997). In contrast, the

substitution of Sr into aragonite is an exothermic reaction, so the Sr/Ca ratio of aragonite should be inversely correlated with temperature.

Based on thermodynamic considerations, Rosenthal et al. (1997) and Lea et al. (1999) proposed an exponential temperature dependence of Mg uptake into calcite of about 3% per K. This prediction is consistent with experiments of Mg incorporation into inorganically precipitated calcites; inorganic coprecipitation experiments of Sr with aragonite show a weak inverse linear dependence on temperature. However, as instructive as these data may be, biologically mediated processes are rarely at thermodynamic equilibrium. For example, Foraminifera contain 1–2 orders of magnitude lower Mg than found in marine inorganic calcites. Such offsets suggest that biological processes exert a major influence on the coprecipitation of metals in biogenic carbonates, thus highlighting the need for species-specific empirical calibrations.

Foraminiferal Mg/Ca Paleothermometry

Temperature Calibrations of Mg/Ca in Planktonic Foraminifera

The temperature dependence of Mg uptake into planktonic foraminiferal tests has been determined using three different approaches: (i) controlled culture experiments in which temperature is fixed independently of other environmental parameters (e.g., light, salinity, and pH); (ii) measurements of planktonic Foraminifera from sediment trap time series in sites characterized by significant seasonal variability in SST; (iii) analyses of fossil (dead) Foraminifera tests obtained from surface sediments, typically from core tops. A summary of published multiple-, and single-species calibrations is given in Table 1. At present, Mg/Ca calibrations are mostly expressed as an exponential dependence of temperature in the form: $\text{Mg}/\text{Ca} \text{ (mmol mol}^{-1}\text{)} = B e^{AT}$ where A and B are the exponential and pre-exponential constants, respectively, and T is temperature in $^{\circ}\text{C}$. Justification for the choice of exponential fit to the empirical data comes primarily from the thermodynamic prediction of exponential response. In this relationship, the exponential constant reflects the Mg/Ca response to a given temperature change (in $\text{mmol mol}^{-1}^{\circ}\text{C}^{-1}$), implying increased sensitivity with temperature. The pre-exponential constant determines the absolute temperature.

Culture-based calibrations are available for several species including *Globigerinoides sacculifer* (Nürnberg et al., 1996), *Globigerinoides ruber* (white; Kisakurek et al., 2008), *Orbulina universa* (Lea et al., 1999; Russell et al., 2004), *Globigerina bulloides* (Lea et al., 1999; Mashiotto et al., 1999), and *Neoglobobulimina papyderma* (dextral; Langen et al., 2005) (Figure 1(a)). Combined, the available experiments suggest a temperature sensitivity of $9.4 \pm 0.9\%$ (1 S.D.) change in Mg/Ca per $^{\circ}\text{C}$, thus providing the strongest direct evidence for temperature control

Table 1 Summary of published Mg/Ca–temperature calibrations for multiple and single species of planktonic Foraminifera (Mg/Ca in mmol mol⁻¹ and *T* in °C). Except when noted, all calibrations are based on exponential equations $Mg/Ca = A e^{BT}$

Species	A	B	Material	Region	Temperature range (°C)	References
Multiple-species equations						
Two species	0.47	0.082	Core tops	Atlantic and Pacific	0–30	Nürnberg et al. (1996)
Eight species	0.52	0.100	Core tops	North Atlantic	8–22	Elderfield and Ganssen (2000)
Eight species	0.78	0.095	Core tops	North Atlantic	8–22	Rosenthal and Lohmann (2002) based on data from Elderfield and Ganssen (2000)
Ten species	0.38	0.090	Sediment traps	North Atlantic	15–28	Anand et al. (2003)
Two shallow and thermocline dwellers	0.22	0.113	Core tops	North Atlantic	19–28	Regenberg et al. (2009)
Two deep dwellers	0.84	0.083	Core tops	North Atlantic	8–15	Regenberg et al. (2009)
Single-species equations						
<i>G. ruber</i> (white)	0.50	0.080	Culture		17–32	Kisakurek et al. (2008)
<i>G. sacculifer</i>	0.39	0.089	Culture		19–30	Nürnberg et al. (1996)
<i>O. universa</i>	1.36	0.085	Culture		16–25	Lea et al. (1999)
<i>O. universa</i>	0.85	0.096	Culture		15–25	Russell et al. (2004)
<i>G. bulloides</i>	0.53	0.100	Culture		16–25	Lea et al. (1999)
<i>G. bulloides</i>	0.47	0.107	Culture/core tops		10–25	Mashiotta et al. (1999)
<i>N. pachyderma</i> (d)	0.51	0.104	Culture		9–19	Langen et al. (2005)
<i>G. ruber</i> (white; 250–350)	0.34	0.102	Sediment traps	North Atlantic	22–28	Anand et al. (2003)
<i>G. ruber</i> (white; 212–355)	0.69	0.068	Sediment traps	Gulf of California	20–33	McConnell and Thunell (2005)
<i>G. ruber</i> (white; 250–350)	0.5	0.084	Sediment traps	Java, Indonesia	25–30	Mohtadi et al. (2009)
<i>G. sacculifer</i> with sac (350–500)	0.67	0.069	Sediment traps	North Atlantic	22–28	Anand et al. (2003)
<i>G. sacculifer</i> w/out sac (350–500)	1.06	0.048	Sediment traps	North Atlantic	22–28	Anand et al. (2003)
<i>G. bulloides</i> (212–355)	1.20	0.057	Sediment traps	Gulf of California	16–31	McConnell and Thunell (2005)
<i>G. ruber</i> (white; 250–350)	0.30	0.089	Core tops	Equatorial Pacific	24–29	Lea et al. (2000)
<i>G. ruber</i> (white; 250–350)	0.38	0.090	Core tops	Atlantic and Pacific	21–29	Dekens et al. (2002)
<i>G. ruber</i> (white; 250–315)	0.76	0.070	Core tops	Atlantic	17–28	Cleroux et al. (2008)
<i>O. universa</i>	0.95	0.086	Core tops	North Atlantic	8–22	Hathorne et al. (2003)
<i>G. bulloides</i>	0.56	0.100	Core tops	North Atlantic	8–22	Elderfield and Ganssen (2000)
<i>G. bulloides</i> (250–315)	0.78	0.082	Core tops	Atlantic	10–18	Cleroux et al. (2008)
<i>N. pachyderma</i> (s)	0.55	0.099	Core tops	Norwegian Sea	1–12	Nürnberg (1995)
<i>N. pachyderma</i> (s)	0.41	0.083	Core tops	South Atlantic	0–15	Nürnberg (1995)
<i>N. pachyderma</i> (s) (linear equation $Mg/Ca = AT + B$)	0.13	0.35	Core tops	Norwegian Sea	0–15	Kozdon et al. (2009)
<i>G. inflata</i> (355–400)	0.71	0.056	Core tops	Atlantic	11–18	Cleroux et al. (2008)
<i>G. inflata</i> (315–400)	0.72	0.076	Core tops	South Atlantic	3.1–16.5	Groeneveld and Chiessi (2011)
<i>P. obliquiloculata</i> (355–400)	1.02	0.039	Core tops	Atlantic	20–25	Cleroux et al. (2008)
<i>G. trancatulinioides</i> (r) (355–400)	0.62	0.074	Core tops	Atlantic	11–17	Cleroux et al. (2008)
<i>G. trancatulinioides</i> (r)	0.36	0.098	Core tops	Indian Ocean	8–25	McKenna and Prell (2005)
<i>G. trancatulinioides</i> (s) (355–400)	0.88	0.045	Core tops	Atlantic	12–17	Cleroux et al. (2008)

on planktonic foraminiferal Mg/Ca, much in the same way that inorganic experiments supported the thermodynamic expectations. Sediment trap calibrations provide additional support to the dominant temperature control. The sediment-trap based multispecies calibration of Anand et al. (2003) from the Sargasso Sea (Figure 1(b)) is in good agreement with culture-based calibrations, suggesting a multispecies temperature sensitivity of $9 \pm 0.3\%$ per °C. Subsequent calibrations of *G. ruber* (white) from sediment traps in the Indian Ocean, off Java Island (Indonesia), and the South China Sea yield consistent results (Huang et al., 2008; Mohtadi et al., 2009). Consistent with these results, the multispecies core-top calibration of Elderfield and Ganssen (2000), reevaluated by Rosenthal and Lohmann (2002) suggests temperature sensitivity of $9.5 \pm 0.5\%$ per °C (Figure 1(c)).

A main advantage of multispecies calibration is the large temperature range it provides, which may offer a broader base

for statistical analysis. However, offsets among species can bias the calibration, and therefore, should be considered when choosing the appropriate equation, be it a single- or a multi-species calibration (Regenberg et al., 2009). Indeed, such offsets in Mg/Ca among individual species stress the need for single-species calibrations. Although there is significant variability among the different equations, the median exponential value of the 27 calibrations given in Table 1 is $8.5 \pm 1.6\%$ (1 S.D.) per °C (average 8.2%), in accordance with the multispecies calibrations. Collectively, a sensitivity of $9 \pm 1\%$ per °C seems to be a robust feature of planktonic Foraminifera; differences among calibration may reflect additional controls. It is also noteworthy that for single-species calibrations based on a narrow temperature range, a linear regression might provide a better fit to the data. The pre-exponential constant (*B*) differs significantly among calibrations, for both the same and

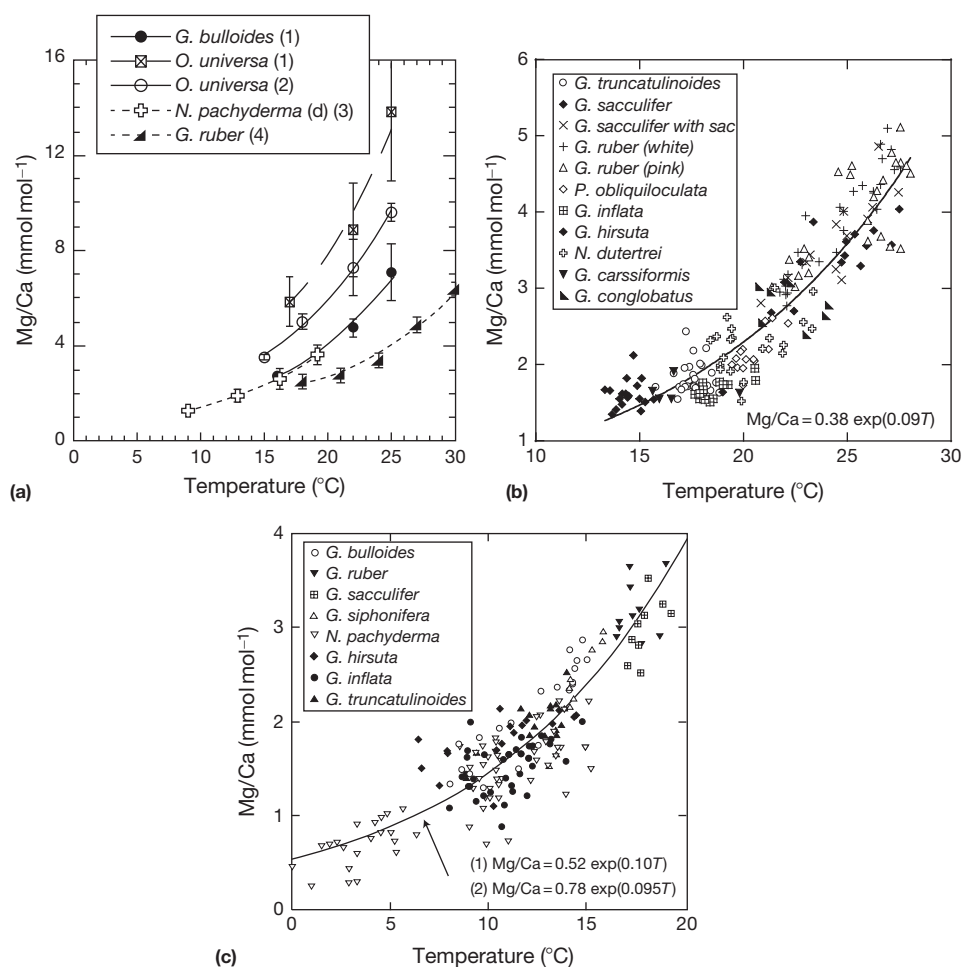


Figure 1 Mg/Ca–temperature calibrations based on: (a) culture experiments (data from (1) Lea et al. (1999), (2) Russell et al. (2004), (3) (Langen et al. (2005), (4) Kisakurek et al. (2008)); (b) sediment traps from the Sargasso Sea (Anand et al., 2003); (c) core tops (Elderfield and Ganssen, 2000) (regression line 1 from Elderfield and Ganssen (2000), line 2 from Rosenthal and Lohmann (2002) based on the same data set).

different species. This high variability reflects a combination of several factors including intra- and interspecies variability (Anand et al., 2003; Elderfield and Ganssen, 2000), diagenetic overprints including postdepositional dissolution (Rosenthal and Lohmann, 2002), contamination from sedimentary contaminants (e.g., Lea et al., 2005; Pena et al., 2005, 2008). It is also important to note that methodological biases among laboratories should be considered and accounted for when comparing data sets cleaned with and without the reductive step (Rosenthal et al., 2004). It follows, that the accuracy of Mg/Ca thermometry is often better in estimating relative changes in seawater temperatures (i.e., anomalies) than absolute temperatures.

The good agreement between the core-top, sediment trap and culture-based calibrations also provides strong evidence that the temperature signal imprinted during the test formation is reliably transferred into the sediment although, as discussed below, secondary effects of salinity and pH on the calcification process and postdepositional alteration may complicate the interpretation of Mg/Ca records in terms of variability due strictly to temperature.

Inter- and Intraspecies Variability

Interspecies variability in Mg/Ca is generally correlated with their calcification depth. Accordingly, shallow mix-layer dwellers (e.g., *G. ruber* and *G. sacculifer*) have high Mg/Ca, whereas deep dwellers (e.g., *Pulleniatina obliquiloculata*, *Globigerinoides inflata*, *Globigerinoides truncatulinoides*), are characterized by relatively low Mg/Ca (Rosenthal and Boyle, 1993). By assuming specific calcification depth to each foraminiferal species, and using the appropriate calibration (Anand et al., 2003; Cleroux et al., 2008; Regenberg et al., 2009; Skinner and Elderfield, 2005) one can reconstruct the temperature structure of the upper thermocline (e.g., Groeneweld and Chiessi, 2011; Xu et al., 2008). However, although it is common to assign planktonic Foraminifera specific calcification depths, it is also well known that many species continue to calcify while migrating vertically through the water column and some species, such as *G. sacculifer*, also add an outer calcite crust prior to reproduction (aka ‘gemetogenic crust’) at depths significantly deeper (and colder) than their principal habitat depth. These processes are likely to cause significant variability in the Mg

distribution within and among Foraminifera tests, which should be linked to changes in calcification temperature. It follows, that the 'whole test' Mg/Ca composition most commonly used for temperature reconstructions represents a weighted average of calcite layers formed at different depths/temperatures. This is less of an issue for the shallow, mixed-layer species of *G. ruber* and *G. sacculifer* and more critical for deep species such as *P. obliquiloculata*, *Neogloboquadrina dutertrei*, *Globigerinoides tumida* and *G. truncatulinoides*.

There is also significant size-dependent variability in Mg/Ca ratios among tests of the same species. Mg/Ca increases with increasing test size in most species, possibly because small individuals calcify faster than large individuals (Elderfield et al., 2002) or because larger individuals spend relatively more time near the sea surface, where the water is warmer and the higher light intensity increases symbiont photosynthetic activity, which in turn allows growth to larger sizes (Honisch and Hemming, 2004; Spero and Lea, 1993). In addition, Mg/Ca can vary among different morphotypes of the same species, reflecting differences in growth season (e.g., the white vs. pink variety of *G. ruber*; Anand et al., 2003) or depth habitat (e.g., *G. ruber* s.s. vs. *G. ruber* s.l.; Steinke et al., 2005). The intraspecific variability implies that Mg/Ca analyses and calibration should be performed on the same morphotypes and within a narrow size range. Also, when possible, the same size fraction should be used for both Mg/Ca and isotopic analyses.

Studies using microanalytical techniques also reveal large variability in Mg/Ca (and other trace elements) within and among individual tests that is apparently too large to be explained by the temperature change due to the vertical migration experienced by the different species during their adult life cycle (Hathorne et al., 2009; Kunioka et al., 2006; Sadekov et al., 2008). The pattern of Mg/Ca variation is significantly different between symbiont-bearing and symbiont-free species (Eggins et al., 2003, 2004). Tests of symbiont-bearing Foraminifera show cyclic variations between high and low Mg/Ca layers within individual chambers, which are largely absent from symbiont-free species (Sadekov et al., 2005). There are also indications that high Mg/Ca ratios are associated with organic-matrix rich bands (Kunioka et al., 2006). Nonetheless, despite the large, hitherto unexplained, intratest variability, it is widely accepted now that whole test Mg/Ca measurements produce estimates that closely match measured temperatures along the water column at the assumed individual species calcification depth and over a large geographical extent (Eggins et al., 2003; Hathorne et al., 2009; Sadekov et al., 2008, 2009, 2010).

Secondary Nontemperature Effects

Salinity and pH effect

Culture studies have provided evidence for a positive relationship between planktonic foraminiferal Mg/Ca and seawater salinity (Lea et al., 1999; Nürnberg et al., 1996) and arguably an inverse dependence on pH (Lea et al., 1999; Russell et al., 2004). These experiments have documented a relatively small salinity effect, about $6 \pm 4\%$ change in Mg/Ca per salinity unit (psu) over a salinity range of 20–45 for *G. ruber*, *G. sacculifer* and *O. universa* (Dueñas-Bohórquez et al., 2009; Kisakurek et al., 2008; Lea et al., 1999). In contrast, core-top calibrations suggest substantially higher and quite variable salinity

dependence. For *G. ruber*, these calibrations suggest salinity sensitivity of about 14%, 27%, and up to 40% per psu. for sediments from the equatorial Indian-Pacific Oceans (Mathien-Blard and Bassinot, 2009), Atlantic Ocean (Arbuszewski et al., 2010), and Mediterranean Sea (Ferguson et al., 2008), respectively (although note that the very high estimates might reflect contamination by secondary high-Mg overgrowths; Hoogakker et al., 2009). Interestingly, the core-top calibrations suggest that the dependence on salinity is significantly higher than suggested from culture experiments and apparently increases with salinity. Indeed, a recent paper suggests little or no Mg/Ca dependence at low salinities (32–34) and a high dependence at high salinities (>35) (Arbuszewski et al., 2010).

The pH effect on planktonic Foraminifera is still uncertain. In early culture experiments, Mg/Ca in *O. universa* and *G. bulloides* decreased by about $7 \pm 6\%$ per 0.1 pH unit (Lea et al., 1999; Russell et al., 2004). Subsequent culture experiments of *G. ruber* and *G. sacculifer* suggest a negligible effect at modern seawater pH or $[\text{CO}_3]$ concentration and an inverse relationship below that (Dueñas-Bohórquez et al., 2009; Kisakurek et al., 2008). Indeed, the $[\text{CO}_3]$ effect may differ among species and also depend on other environmental conditions. For example, a sediment trap study in the Antarctic Ocean shows significant positive relationship between Mg/Ca of *N. pachyderma* (sinistral) and surface water $[\text{CO}_3]$ (Hendry et al., 2009). This and complementary studies in the Nordic Seas (Kozdon et al., 2009; Meland et al., 2006) suggest that temperature may exert the dominant control on Mg/Ca in *N. pachyderma* (s) above 3 °C, whereas below that there is an additional saturation effect that might mask the temperature dependence. It could also be that long-term adaptation in the field induces a different response than observed under short-term culture experiments.

Whereas more studies are required to accurately determine the sensitivity of Mg/Ca to different environmental conditions, a new generation of equations may be required in order to more accurately describe the dependence of planktonic foraminiferal Mg/Ca on temperature, salinity, and carbonate concentrations. These multivariate equations should be used in future studies and for reappraising published records (e.g., Arbuszewski et al., 2010; Mathien-Blard and Bassinot, 2009).

Dissolution Effects

Core-top studies along bathymetric transects indicate a systematic decrease in Mg/Ca ratios of planktonic Foraminifera with increasing depth, independent of the overlying SSTs, thereby suggesting that foraminiferal Mg/Ca is altered by post-depositional dissolution on the seafloor, driven by the depth-dependent decrease in carbonate saturation levels (Dekens et al., 2002; Rosenthal et al., 2000). The dissolution effect is significantly stronger below the lysocline, although the exact response depends on the species (Regenberg et al., 2006). It has been argued that the systematic decrease in Mg/Ca ratios of planktonic Foraminifera is likely caused by preferential dissolution of the Mg-rich and more dissolution-susceptible calcite (Brown and Elderfield, 1996), which results in a decrease in the whole test Mg/Ca ratio that shifts the temperature estimates toward apparently colder values (Rosenthal et al., 2000). The decrease in Mg/Ca with depth is higher in

nonspinose, thermocline dwelling species (e.g., *G. tumida* and *N. dutertrei*) than in spinose, shallow dwelling species (*G. ruber* and *G. sacculifer*), in accordance with the larger range of temperatures over which the deep dwellers calcify (Brown and Elderfield, 1996; Dekens et al., 2002; Huang et al., 2008).

Various approaches have been offered to account for dissolution effects on Mg/Ca. Several studies propose that combining core-top measurements, where the magnitude of the depth-dependent decrease of Mg/Ca has been determined with independent estimates for past shifts in lysocline depth, could be used to estimate the bias in Mg/Ca–temperatures due to dissolution effects (Dekens et al., 2002; Lea et al., 2000; Regenberg et al., 2006). Applying this approach to two Pacific records, these studies suggest a potential error of 0.5 °C in the estimates of the tropical Pacific glacial–interglacial temperature amplitude. This approach, however, does not quantify the dissolution effects on specific samples, which might respond more to changes in pore-water chemistry than to shifts in the hydrographic lysocline depth. Taking a different approach, Rosenthal and Lohmann (2002) suggested that the relationship between size-normalized test weight, and the dissolution-driven decrease in Mg/Ca, could be used for correcting down core Mg-based temperature estimates for dissolution effects. While this method is very attractive, as it directly addresses

dissolution effects experienced by a particular sample, further studies have demonstrated that the test's size–weight relationship is not strictly a function of dissolution, as initially thought, but also depends on the [CO₃] content of seawater in which they grow (Barker and Elderfield, 2002). It is possible that using multiple species at different water depths (i.e., above and below the lysocline) may provide a way to distinguish primary from postdepositional effects.

Temperature calibrations of Mg/Ca in benthic Foraminifera

Initial calibrations, largely based on the *Cibicidoides pachyderma* from a shallow bathymetric transect in Little Bahama Banks (spanning a temperature range of 5–18 °C), suggested that the temperature sensitivity of *Cibicidoides* spp. Mg/Ca is essentially identical with the ~9% increase in Mg/Ca per °C observed in planktonic Foraminifera (Lear et al., 2002; Martin et al., 2002; Rosenthal et al., 1997). Similar temperature sensitivities were observed in other benthic species including *Planulina* spp., *Oridorslis umbonatus*, and *Melonis* spp. (Lear et al., 2002), however, as is the case for planktonic Foraminifera, it was also noted that interspecies offsets in benthic Foraminifera exist for which corrections are needed (e.g., Lear et al., 2000). Subsequent studies have raised questions, however, about the accuracy of these initial calibrations (Table 2).

Table 2 Summary of published Mg/Ca–temperature calibrations for benthic Foraminifera (Mg/Ca in mmol mol^{−1} and temperature in °C)

Species	A	B	C	Mineral	Region	Temperature range (°C)	References
Exponential equations $Mg/Ca = A e^{BT}$							
<i>C. pachyderma</i>	1.36	0.100		Low-Mg calcite	Little Bahama Banks	4.5–18	Rosenthal et al. (1997)
<i>C. pachyderma</i> , <i>C. wuellerstorfi</i>	0.85	0.11		Low-Mg calcite	Atlantic and Pacific	0–18	Martin et al. (2002)
<i>C. pachyderma</i> , <i>C. wuellerstorfi</i> , <i>C. compressus</i>	0.87	0.11		Low-Mg calcite	Atlantic and Pacific	1–18	Lear et al. (2002)
<i>Planulina</i> spp.	0.79	0.120		Low-Mg calcite	Pacific	2–12	Lear et al. (2002)
<i>O. umbonatus</i>	1.01	0.114		Low-Mg calcite	Atlantic and Pacific	1–10	Lear et al. (2002)
<i>O. umbonatus</i>	1.53	0.09		Low-Mg calcite	Atlantic	3–10	Rathmann et al. (2004)
<i>Melonis</i> spp.	0.98	0.101		Low-Mg calcite	Atlantic and Pacific	1–18	Lear et al. (2002)
<i>Melonis barleeanus</i>	0.658	0.137		Low-Mg calcite	Iceland	0–7	Kristjánsdóttir et al. (2007)
<i>P. ariminensis</i>	0.91	0.062		Low-Mg calcite	Pacific	3–15	Lear et al. (2002)
<i>Cassidulina neoreretis</i>	0.864	0.082		Low-Mg calcite	Iceland	8–14	Kristjánsdóttir et al. (2007)
<i>Gyroidinoides</i> spp.	1.71	0.109		Low-Mg calcite	Mediterranean Sea	0–7	Cacho et al. (2006)
<i>Uvigerina</i> spp.	0.92	0.061		Low-Mg calcite	Atlantic and Pacific	2–18	Lear et al. (2002)
<i>Uvigerina</i> spp.	0.98	0.043		Low-Mg calcite	Florida Straits	5–18	Marchitto et al. (2007)
<i>Uvigerina</i> spp.	0.98	0.045		Low-Mg calcite	South Pacific	1–20	Elderfield et al. (2009)
Linear equations $Mg/Ca = A + BT$							
<i>C. pachyderma</i>	1.12	0.12		Low-Mg calcite	Florida Straits	5–18	Marchitto et al. (2007)
<i>C. pachyderma</i>	0.89	0.13		Low-Mg calcite	Little Bahama Banks	5–18	Curry and Marchitto (2008)
<i>Uvigerina</i> spp.	0.77	0.079		Low-Mg calcite	Florida Straits	5–18	Marchitto et al. (2007)
<i>Uvigerina</i> spp.	0.98	0.065		Low-Mg calcite	South Pacific	1–20	Elderfield et al. (2009)
<i>Planoglabratella opercularis</i>	89.7	2.22		High-Mg calcite	culture	10–25	Toyofuku et al. (2000)
<i>Quinqueloculina yabei</i>	66.0	2.9		High-Mg calcite	culture	10–25	Toyofuku et al. (2000)
<i>Planoglabratella opercularis</i>	81.5	1.6		High-Mg calcite	culture	10–25	Toyofuku and Kitazato (2005)
<i>Hyalinea balthica</i>	0	0.488		Low-Mg calcite	Global	5–14	Rosenthal et al. (2011)
<i>Hoeglundona elegans</i>	0.96	0.034		Aragonite	Global	>4	Rosenthal et al. (2006)
Linear equations $Mg/Ca = A + BT + C\Delta[CO_3]$							
<i>C. wuellerstorfi</i>	0.82	0.056	0.0087	Low-Mg calcite	Global	<3	Elderfield et al. (2006)

First, it has been argued that the shallow Little Bahama Banks samples are compromised by diagenetic calcite overgrowths enriched in Mg, thus yielding an apparently high exponential dependence on temperature. New calibration studies of *C. pachyderma* from depth transects in the Florida Straits and Great Bahama Banks, in areas where the problem of diagenetic calcite overgrowths is apparently minimal, suggest a linear relationship between Mg/Ca and BWT with a sensitivity of $\sim 0.12 \text{ mmol mol}^{-1} \text{ }^{\circ}\text{C}^{-1}$ (Table 2) for the temperature range from 5.8 to 18.6 $^{\circ}\text{C}$ (Marchitto et al., 2007). This calibration has been replicated in Little Bahama Banks by Curry and Marchitto (2008) who used a microanalytical technique to avoid contamination by diagenetic overgrowths.

Secondly, in deep waters the degree of CaCO_3 saturation starts to exert a significant control on the incorporation of Mg into benthic foraminiferal tests, thus altering the Mg/Ca-temperature relationships (Elderfield et al., 2006; Martin et al., 2002; Rosenthal et al., 2006). New calibrations based on global datasets suggest a $\Delta[\text{CO}_3]$ -Mg/Ca sensitivity of $\sim 0.009 \text{ mmol mol}^{-1} \mu\text{mol}^{-1} \text{ kg}^{-1}$ (where $\Delta[\text{CO}_3]$ is the degree of calcite bottom-water saturation) in calcitic benthic tests that, in the modern ocean, becomes increasingly important below $\sim 3^{\circ}\text{C}$ (Elderfield et al., 2006; Healey et al., 2008). There are also indications for a significant $[\text{CO}_3]$ ion effect at warmer, oversaturated waters (Bryan and Marchitto, 2008). After adjusting for the additional $\Delta[\text{CO}_3]$ effect, a sensitivity of about $0.12\text{--}0.15 \text{ mmol mol}^{-1} \text{ }^{\circ}\text{C}^{-1}$ is found in other epifaunal species including *Cibicidoides wuellerstorfi*, *Cibicidoides mundulus* and *Oridorsalis umbonatus* (Healey et al., 2008; Raitzsch et al., 2008; Sosdian and Rosenthal, 2009; Yu and Elderfield, 2008), consistent with the *C. pachyderma* calibration (Marchitto et al., 2007).

It is now evident that benthic foraminiferal Mg/Ca thermometry is not as straightforward as initially thought, and the additional $\Delta[\text{CO}_3]$ effect needs to be constrained in any down core reconstructions of deep-water temperatures. One possible approach is using paired elemental ratios with different sensitivities for temperature and $\Delta[\text{CO}_3]$ (e.g., Mg/Ca coupled with Li/Ca or B/Ca) to correct for, or deconvolve the contribution of different environmental parameters (Bryan and Marchitto, 2008; Lear et al., 2010; Yu and Elderfield, 2008) or through using independent estimates of the carbonate ion effect (Sosdian and Rosenthal, 2009). However, the low sensitivity suggested for the above-mentioned Foraminifera and the additional $\Delta[\text{CO}_3]$ effect can also lead to large errors associated with estimating deep ocean temperatures (e.g., $\pm 2.4^{\circ}\text{C}$ for *C. pachyderma*: Marchitto et al., 2007). Alternatively, it has been argued that infaunal taxa such as *Uvigerina* sp. may be less affected by the $\Delta[\text{CO}_3]$ effect as it calcifies from saturated pore waters and thus is less affected by bottom water chemistry (Elderfield et al., 2009). The exact temperature sensitivity of *Uvigerina* species is, however, still uncertain. On the other hand, core-top calibrations suggest sensitivity of about 6–8% per $^{\circ}\text{C}$ (Elderfield et al., 2009; Lear et al., 2002), so a significantly higher sensitivity is required to explain the relatively large down core variability (Elderfield et al., 2009; Martin et al., 2002). If indeed some shallow infaunal benthic Foraminifera are free of saturation effects and characterized by high temperature sensitivity, they may offer a new tool for accurate and

precise reconstruction of past deep ocean temperatures. One promising example is the use of the shallow infaunal calcitic species, *Hyalinea balthica*, which exhibits high temperature sensitivity ($\sim 0.5 \text{ mmol mol}^{-1} \text{ }^{\circ}\text{C}^{-1}$) and no discernible $\Delta[\text{CO}_3]$ effect (Bamberg et al., 2010; Rosenthal et al., 2011).

Inferring Seawater $\delta^{18}\text{O}$ and Salinity

The $\delta^{18}\text{O}$ composition of calcareous marine fossils reflects, at a minimum, both the temperature and the isotopic composition of seawater in which the shell or skeleton was formed. The $\delta^{18}\text{O}$ of modern surface water reflects, in part, the hydrological balance between evaporation and precipitation (E/P), that is, the surface salinity. Advection can also alter $\delta^{18}\text{O}_{\text{sw}}$ by mixing waters of different salinities (or $\delta^{18}\text{O}_{\text{sw}}$). Over very long time scales, seawater $\delta^{18}\text{O}$ is also a function of the extent of continental ice sheets. Paired $\delta^{18}\text{O}$ and Mg/Ca measurements in Foraminifera allow for the adjustment of temperature-dependency of carbonate $\delta^{18}\text{O}$ and isolation of $\delta^{18}\text{O}_{\text{sw}}$ (Lea et al., 2000; Mashiota et al., 1999), which can then be used to reconstruct local changes in E/P and by inference salinity (e.g., Oppo et al., 2009; Schmidt et al., 2004; Stott et al., 2004), and provide valuable information about changes in continental ice volume (e.g., Lear et al., 2000). A major advantage of measuring both proxies on the same samples is that it allows for the determination of temporal leads and lags between variations in seawater temperature and the build-up or erosion of continental ice sheets, which is fundamental for understanding climate change.

In the past decade, the application of this method has already led to important findings. New Mg-derived SST records have contributed to our understanding of the role of the equatorial Pacific in global climate change on millennial (e.g., Koutavas et al., 2006; Linsley et al., 2010; Stott et al., 2002), orbital (e.g., Lea et al., 2000; Rosenthal et al., 2003), and secular time scales (e.g., de Garidel-Thoron et al., 2005; Wara et al., 2005). Among the most significant findings that come from these studies are: (1) the equatorial Pacific was characterized by a 'super El Nino' state during the mid-Pliocene warm period (Wara et al., 2005); (2) except for glacial-interglacial variability, Western Pacific warm pool (WPWP) SSTs were stable throughout the Pleistocene, despite Northern Hemisphere glaciations and global cooling (de Garidel-Thoron et al., 2005); (3) SST in the WPWP was 2–3 $^{\circ}\text{C}$ cooler during glacial periods than at present (e.g., de Garidel-Thoron et al., 2007; Lea et al., 2000; Linsley et al., 2010); and (4) during glacial terminations, changes in equatorial SST arguably led the Northern Hemisphere deglacial record by ~ 3000 years (Lea et al., 2000; Visser et al., 2003).

Other examples of a breakthrough offered by this new approach are benthic foraminiferal Mg/Ca records documenting the global cooling trend throughout the Cenozoic (Billups, 2003; Billups and Schrag, 2002; Lear et al., 2000, 2004; Sosdian and Rosenthal, 2009). The low-resolution seawater oxygen isotope record ($\delta^{18}\text{O}_{\text{seawater}}$) of Lear et al. (2000) suggests that the Cenozoic expansion of continental ice sheets occurred primarily in three steps: the Antarctic glaciation during the Eocene-Oligocene (E/O) transition and the middle Miocene, and the Northern Hemisphere glaciation during the Pliocene-Pleistocene. A high-resolution benthic Mg/Ca record from the

North Atlantic (Figure 2) shows two distinct cooling events: the first associated with the late Pliocene (LPT, 2.5–3 Ma) and the second preceding the transition from a dominant 41–100 ka climate variability during mid-Pleistocene (the Mid-Pleistocene Transition, 1.2–0.85 Ma; Sosdian and Rosenthal, 2009). The reconstructed $\delta^{18}\text{O}_{\text{seawater}}$ record suggests that sea level was apparently 25 ± 20 m higher than at present during the mid-Pliocene, a period considered to be a good analog for future climate. While this estimate needs to be validated, given the uncertainties in the method, it offers a new way to assess climate variability and impacts under different boundary conditions. On time scales longer than ~ 1 My, secular variations in seawater Mg/Ca need to be considered before interpreting foraminiferal Mg/Ca records strictly in terms of temperature. First, the paleotemperature equation should include a term accounting for the difference between the seawater Mg/Ca ratio in the past and in the modern ocean. Secondly, studies of various calcareous organisms suggest that the partitioning of Mg between the solution and biogenic CaCO_3 minerals is nonlinearly dependent on the Mg/Ca ratio of the seawater (Hasiuk and Lohmann, 2010; Ries, 2004; Segev and Erez, 2006). Therefore, these parameters need to be considered in order to obtain accurate reconstructions of the secular variations of ocean temperature (e.g., Cramer et al., 2011 and references therein).

Marine Ostracode Mg/Ca Paleothermometry

Studies over the past decade have provided strong evidence for the temperature dependence of Mg/Ca in marine ostracodes (Dwyer et al., 2002, and references therein). Indeed records of Mg/Ca in marine ostracodes have been used successfully for reconstructing the history of BWTs through the Pliocene and Pleistocene (e.g., Cronin et al., 2005; Dwyer et al., 1995). Ostracodes are bivalved crustaceans that grow via a process of molting. Ostracodes excrete an exoskeleton known as a carapace that in some taxa consists of 80–90% calcite and the rest of chitin and proteins. Tests of adult specimens, used in paleoceanography, are flat, typically 0.5–1 mm in length and weigh between 20 and 200 μg , thus providing a good target for geochemical analyses. Temperature calibrations are currently available for two marine, epifaunal benthic ostracodes, the deep-sea genus *Krithe* and shallow marine/estuarine genus *Loxococoncha* (Dwyer et al., 1995, 2002), both long-lived genera known since the Cretaceous. Mg/Ca ratios in these genera are significantly higher than in deep-sea benthic Foraminifera. Based on core-top calibrations, Dwyer et al. (2002) proposed a linear relationship between *Krithe* Mg/Ca and temperature, with a slope of $0.95 \pm 0.15 \text{ mmol mol}^{-1} \text{ } ^\circ\text{C}^{-1}$ or expressed as an exponential dependence of 6% per $^\circ\text{C}$, which is within the range of sensitivities observed in Foraminifera (Figure 3).

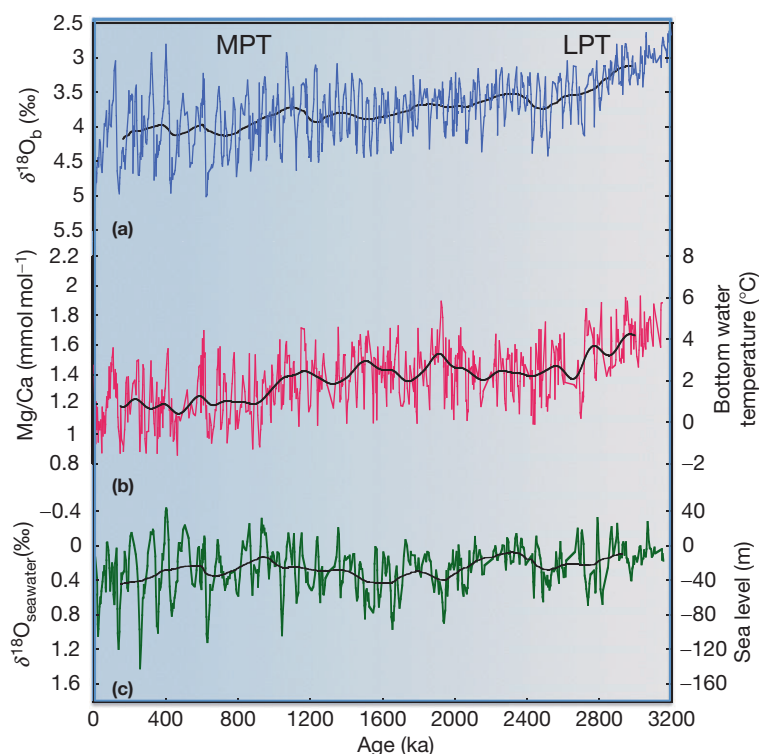


Figure 2 Plio-Pleistocene benthic foraminiferal records from Deep Sea Drilling Project (DSDP) site 607 (41° N, 32° W, 3427-m water depth) in the North Atlantic: (a) benthic oxygen isotope record ($\delta^{18}\text{O}_b$ in ‰PDB); (b) benthic foraminiferal Mg/Ca record, derived from *C. wuellerstorfi* and *O. umbonatus*, converted to bottom water temperature (BWT) using the equation: $\text{Mg/Ca} = 0.15 \text{ BWT} + 1.16$; (c) seawater ($\delta^{18}\text{O}_{\text{sw}}$ in ‰SMOW) and sea level estimates (green line represents three-point smoothed curve) derived from the Mg/Ca-BWT and $\delta^{18}\text{O}_b$ records. MPT and LPT refer to mid-Pleistocene and late Pliocene transitions. Black lines represent average signals for each variable determined by applying a Gaussian filter with a cut-off frequency of 400 ka. Modified from Sosdian S and Rosenthal Y (2009) Deep sea temperature and ice volume changes across the Pliocene–Pleistocene climate transitions. *Science*, 325: 306–310.

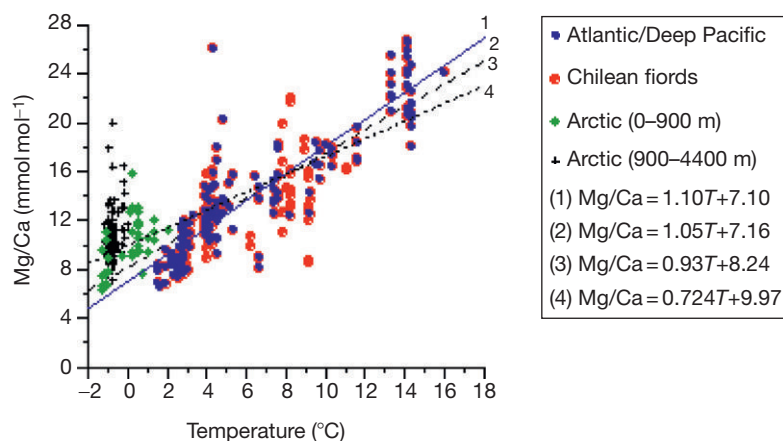


Figure 3 Various Mg/Ca–temperature calibrations for adult shells of genus *Krithe*. Regression lines are based on samples from (1) Atlantic and central Pacific core tops; (2) data from regression 1 and core tops from Chilean Fjords; (3) same as 2 with Arctic core tops between 0 and 900 m; (4) same as 3 with Arctic core tops deeper than 900 m. Modified from Dwyer GS, Cronin TM, and Baker PA (2002) Trace elements in marine ostracodes. *The Ostracoda: Applications in Quaternary Research*, pp. 205–224. American Geophysical Union.

Significant differences between the two ostracode genera and ontogenic effects within each genus can be minimized by only studying tests of the adult stage for a particular genus (Dwyer et al., 2002). The offsets among the different calibrations suggest that secondary, nontemperature parameters might influence the Mg/Ca record of marine ostracodes. Culture studies show no discernible salinity effects, and dissolution experiments suggest that the effects of postdepositional dissolution on the test composition are minimal (Dwyer et al., 2002). A $\Delta[\text{CO}_3]$ effect on Mg/Ca in ostracodes, which has not yet been fully investigated, might explain the divergence among the calibrations.

Coralline Sr/Ca Paleothermometry

Temperature Calibrations of Sr/Ca in Corals

The variability of skeletal Sr/Ca in hermatypic coral skeletons has been shown in many studies to be highly correlated with monthly or seasonal variations in water temperature. Early studies indicated that the Sr/Ca in coral skeletal aragonite is primarily regulated by water temperature (Beck et al., 1992) with more Sr incorporated in coral skeleton at lower temperatures, as discussed above. The relationship obtained in these studies is of the form:

$$(\text{Sr/Ca})_{\text{coral}} (\text{mmol mol}^{-1}) = B + A(\text{SST})$$

where SST is given in $^{\circ}\text{C}$, and the slope $A = -0.062 \text{ mmol mol}^{-1} ^{\circ}\text{C}^{-1}$ (Beck et al., 1992). These initial observations are supported by more recent studies showing strong correlations between near-monthly coral Sr/Ca and surface water temperature (e.g., Alibert and McCulloch, 1997; DeLong et al., 2007; Goodkin et al., 2007; Linsley et al., 2000, 2004; Marshall and McCulloch, 2002; Nurhati et al., 2009; Wu, 2010) (Figure 4). Generally, the slope of the Sr/Ca–SST relationship is similar for near-monthly coral Sr/Ca and monthly SST (Table 3). Despite the success of the coral Sr/Ca paleo-thermometer at many sites, these and other studies have also revealed potential complicating factors that can limit

the use of this tracer. Some of these factors—uncertainties are summarized below.

Secondary Nontemperature Effects

Disequilibrium offset

In *Porites* corals, although the slope of the Sr/Ca–temperature relationships derived at different sites with different individual corals is generally the same, the intercepts of the linear relationship can be significantly different among sites, even for coral analyzed on the same type of instrument. Marshall and Marshall and McCulloch (2002), working with *Porites* corals from the Great Barrier Reef, attribute this effect to be due in part to the disequilibrium Sr/Ca offset in corals. Similar to the $^{18}\text{O}/^{16}\text{O}$ in corals, the skeletal Sr/Ca is out of equilibrium with seawater Sr/Ca.

Using three intracolony and three intercolony *Porites* coral records from the reefs offshore of Amédée Island, New Caledonia, DeLong et al. (2007) found the coral Sr/Ca signal over the twentieth century to be highly reproducible. They report that the average absolute offset between coeval monthly Sr/Ca determinations between any two coral time series is $0.035 \pm 0.026 \text{ mmol mol}^{-1}$ ($\sim 0.65 ^{\circ}\text{C}$), which is less than twice the analytical precision of the coral Sr/Ca measurements.

At Clipperton Atoll (10°N , 109°W) in the eastern Pacific, Wu (2010) and (Wu et al. (in review) observed an offset in mean Sr/Ca value (mmol mol^{-1}) between three *Porites lobata* colonies as previously observed in other coral skeletal Sr/Ca records (Bagnato et al., 2004; De Villiers et al., 1995; DeLong et al., 2007; Marshall and McCulloch, 2002). The offsets between the cores ranged from a low of $0.04 \text{ mmol mol}^{-1}$ to a high of $0.09 \text{ mmol mol}^{-1}$, which would equate to a 0.5 – $1.3 ^{\circ}\text{C}$ difference in temperature reconstruction.

In Savusavu Bay on Fiji's north island of Vanua Levu ($16^{\circ}49'\text{S}$, $179^{\circ}10'\text{E}$), Wu (2010) observed mean Sr/Ca offset of $0.37 \text{ mmol mol}^{-1}$ between the two adjacent Fiji *Porites* colonies in the same water depth. The colonies are thought to be the same species, based on morphology and skeletal architecture. It is worth noting that the two coral cores were analyzed

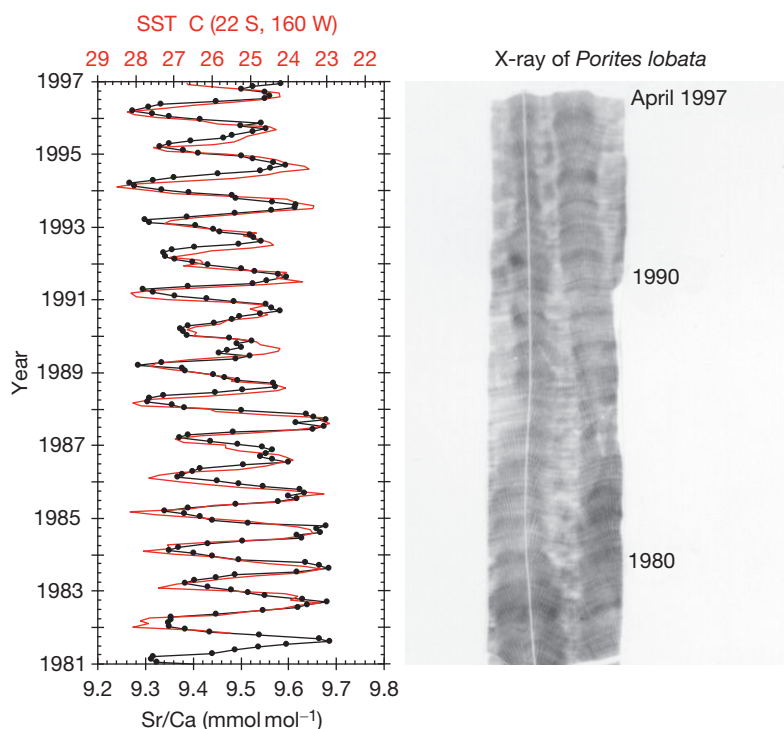


Figure 4 Comparison of monthly sea surface temperature (red) for the latitude–longitude grid including Rarotonga in the South Pacific (1° by 1° ; centered at 22° S, 160° W) and Sr/Ca measurements (black) analyzed at 1 mm intervals in a *Porites lutea* coral core from Rarotonga. The x-ray positive of the coral core shows the sampling track and annual growth density bands.

Table 3 Several published monthly Sr/Ca–temperature calibrations for *Porites* corals from the Pacific Ocean (Sr/Ca in mmol mol^{-1} and T in $^\circ\text{C}$)

Site	Species	Sr/Ca = B + AT		References	Method
		B	A		
New Caledonia	<i>Porites lobata</i>	10.479	−0.062	Beck et al. (1992)	TIMS
New Caledonia	<i>Porites lobata</i>	10.451	−0.054	DeLong et al. (2007)	ICP-OES
Hawaii	<i>Porites lobata</i>	10.956	−0.079	De Villiers et al. (1994)	TIMS
Palmyra	<i>Porites</i> sp.	11.43	−0.088	Nurhati et al. (2009)	ICP-OES
Fanning	<i>Porites</i> sp.	10.78	−0.065	Nurhati et al. (2009)	ICP-OES
Christmas (Pacific)	<i>Porites</i> sp.	11.18	−0.079	Nurhati et al. (2009)	ICP-OES
Taiwan	<i>Porites lobata</i>	10.286	−0.051	Shen et al. (1996)	TIMS
Great Barrier Reef	<i>Porites</i> sp.	10.38	−0.055	Alibert and McCulloch (1997)	TIMS
Great Barrier Reef	<i>Porites</i> sp.	10.4	−0.059	Marshall and McCulloch (2002)	TIMS
Fiji	<i>Porites lutea</i> , <i>Porites lutea</i>	10.63, 9.78	−0.052, −0.034	Linsley et al. (2004) and Wu (2010, PhD diss.)	ICP-OES
Rarotonga	<i>Porites lutea</i>	11.12	−0.065	Linsley et al. (2000)	ICP-OES
Tonga	<i>Porites lutea</i>	10.927	−0.071	Wu (2010, PhD diss.)	ICP-OES
Clipperton	<i>Porites lobata</i>	11.104	−0.068	Wu (2010, PhD diss.)	ICP-OES
Fiji	<i>Diploastrea (Septa)</i>	10.08	−0.034	Bagnato et al. (2004)	ICP-OES
Bermuda	<i>Diploria labyrinthiformis</i>	10.4	−0.048	Goodkin et al. (2007)	ICP-OES
Bermuda	<i>Diploria labyrinthiformis</i>	11.2	−0.084 ^a	Goodkin et al. (2007)	ICP-OES
Bermuda	<i>Diploria labyrinthiformis</i>	10.3	−0.045	Goodkin et al. (2007)	ICP-OES

TIMS, thermal ion mass spectrometry; ICP-OES, inductively coupled plasma optical emission spectrometer.

^aBefore growth rate correction.

on exactly the same ICP-OES instrument in two different laboratories, following interlaboratory calibration (Harvard and Lamont Doherty). This large offset would equate to an unrealistic temperature reconstruction difference of 10°C based on the Sr/Ca–SST calibration derived in this study, or $\sim 6^\circ\text{C}$

based on the commonly accepted Sr/Ca–SST calibration of $-0.06 \text{ mmol mol}^{-1} ^\circ\text{C}^{-1}$.

The different intercepts for coral Sr/Ca–SST relationships shown in Table 3 reflect the fact that the disequilibrium offset can vary between corals of the same species, even in close

proximity. For the genus *Porites*, the similarity of the slopes of the Sr/Ca–SST relationships (Table 3) indicates that Sr/Ca can be used to accurately reconstruct paleo-water temperature, as long as adequate site- and coral-specific temperature calibrations are carried out. For fossil corals lacking a site-specific calibration, the consistent Sr/Ca–SST slope, but potentially different intercept implies that only relative changes in SST over time can accurately be reconstructed ($\pm 0.5^\circ\text{C}$). In all cases, however, for an individual coral Sr/Ca record to be an absolute temperature proxy, the degree of disequilibrium offset must be constant in that coral. This is generally the case for corals sampled along the maximum growth axis (e.g., DeLong et al., 2007; Wu, 2010).

Many other studies report no or insignificant Sr/Ca offsets from corals at the same or closely adjacent sites. For example, in the Line Islands, Nurhati et al. (2009) found insignificant Sr/Ca offsets between *Porites* corals from Christmas, Fanning, and Palmyra Atolls when measured in the same laboratory. They report correlation coefficients (*r*-values) between SST and monthly Sr/Ca of -0.80 , -0.79 , and -0.71 for Christmas, Fanning, and Palmyra, respectively.

Extension rate

Some studies of slow growing corals have reported growth rate effects on skeletal Sr/Ca. De Villiers et al. (1994) reported significantly higher Sr/Ca ratios associated with slower skeletal extension rates in *Pavona clavus*. De Villiers et al. (1994) argued that the difference between skeleton accretion at 12 and 6 mm year⁻¹ is equivalent to $1\text{--}2^\circ\text{C}$ based on near-monthly calibration with monthly SST. However, Marshall and McCulloch (2002) noted that because De Villiers et al. (1994) did not sample exclusively along the coral's maximum growth axis, this invalidates their calcification rate argument, since it is widely known that sampling in off-axis sections of hermatypic coral skeletons generally yields higher Sr/Ca ratios, apparently due to smaller polyp size and lower skeletal density. In other slow growing corals, Bagnato et al. (2004) and Goodkin et al. (2007) both found strong correlations between Sr/Ca and SST in the slow growing corals *Diploastrea* and *Diploria*, respectively. Goodkin et al. (2007) also observed a skeletal extension rate correlation and developed a correction method for adjusted time series Sr/Ca for extension rate effects.

In faster growing *Porites* corals, Shen et al. (1996) and Alibert and McCulloch (1997) find little growth rate effect on Sr/Ca in *Porites lutea* skeletons accreted in the maximum growth axis and Shen et al. (1996) found no perceptible growth rate effect in *Porites* at accretion rates between 18 and 23 mm year⁻¹. DeLong et al. (2007) analyzed Sr/Ca along several *Porites* cores and slabs from New Caledonia *Porites* and found excellent replication and no apparent skeletal extension rate effects. Wu (2010) measured 5-year average extension and calcification rates on *Porites* cores from Fiji, Tonga, and Clipperton. The results revealed no clear and consistent growth trends and no obvious growth effects on the geochemical proxies $\delta^{18}\text{O}$, $\delta^{13}\text{C}$, and Sr/Ca.

Monthly Sr/Ca oscillations in coral aragonite

Using high spatial resolution ion microprobe analyses, Meibom et al. (2003) have found distinct near-monthly oscillations of *Porites lutea* skeletal Sr/Ca with an amplitude greater

than 10%. They propose that these 'monthly' Sr/Ca oscillations result from metabolic changes that are synchronous with the lunar cycle. As has been previously observed in *Porites* (Cohen et al., 2001), their study show that temporally equivalent skeletal elements in *Porites* can have different Sr/Ca ratios. Some of the skeletal Sr/Ca heterogeneity is apparently due to the presence of night-time and day-time deposited skeletal elements (Cohen et al., 2001). During the night, small micron-sized crystals are deposited at the axial spines of the corallites (centers of calcification). Larger needle shaped crystals are formed during the day along all surfaces of the precipitating skeleton. Microprobe analyses of Sr/Ca show that the night-time and day-time skeletons have significantly different Sr/Ca ratios (e.g., Cohen et al., 2001) and that Sr incorporation into the skeleton may be inversely related to calcification rate (Ferrier-Pages et al., 2003).

Replication of decadal and long-term secular variability in coral Sr/Ca

One means of potentially separating climate-related Sr/Ca variability from variability induced by biologic or diagenetic processes is to develop replicated coral Sr/Ca series from a specific site or in a specific region. Only limited replication of century-scale and decadal Sr/Ca variability has been completed to date, and more replicated multicentury Sr/Ca corals records are needed to assess potential biologic or diagenetic effects on Sr/Ca.

From the South Pacific, Linsley et al. (2006) present partially replicated annually averaged multicentury long *Porites* Sr/Ca series from Fiji and Rarotonga. In this study, Linsley et al. (2006) reported that application of the slope of Sr/Ca–SST relationship derived from near-monthly coral Sr/Ca and monthly SST over a calibration interval from 1970 to the core tops to the entire annually averaged multicentury Sr/Ca series at both sites results in unrealistically large amplitude long-term secular trends in SST. To get the Sr/Ca secular trend and interdecadal amplitude at Fiji and Rarotonga to fit the amplitude of instrumental SST and night time marine air temperature variability from 1870 to 1997 required a scaling of $\sim 0.2 \text{ mmol} \cdot \text{mol}^{-1} \cdot ^\circ\text{C}^{-1}$ whereas the monthly calibrations are 0.053 and 0.065 $\text{mmol} \cdot \text{mol}^{-1} \cdot ^\circ\text{C}^{-1}$ at Fiji and Rarotonga, respectively. This 0.2 $\text{mmol} \cdot \text{mol}^{-1} \cdot ^\circ\text{C}^{-1}$ scaling also resulted in a more realistic amplitude of estimated SST changes before 1870. At Fiji, Sr/Ca results for the 20-year period centered on 1660 would indicate a $4\text{--}5^\circ\text{C}$ cooling relative to the late twentieth century using the calibration derived from comparison to monthly instrumental SST (Linsley et al., 2004). The apparent amplification of inferred SST changes from coral Sr/Ca prior to the calibration period was previously discussed with respect to interdecadal variability in Sr/Ca from both Fiji and Rarotonga by Linsley et al. (2004) and may suggest that at these sites the Sr/Ca–SST relationship has not been constant over the last 380 years.

Replicated monthly resolution Sr/Ca time series from three *P. lobata* coral colonies at Clipperton Atoll ($10^\circ 18' \text{N}$, $109^\circ 13' \text{W}$) in the eastern Pacific are used to assess the fidelity of coral Sr/Ca as a SST proxy in this region of relatively stable SST (Wu, 2010; Wu et al., in review). Significant correlations were found between individual cores and gridded instrumental SST for the calibration period (1981–94) with the three-core average increasing the correlation to SST. Down-core, the

intercolony monthly Sr/Ca records showed good coherence with the exception of a 12-year section of growth in one colony. Although, Wu (2010) could find no evidence of diagenesis or other significant skeletal alterations in this section, the presence of this 'anomalous' interval serves to support the multicoral replication approach. Since AD ~1900, there is an observed secular warming trend of ~1 °C recorded in the coral Sr/Ca records which is 0.5–0.75 °C greater than the trend in instrumental SST. The decadal Sr/Ca variability is also greater than that observed in instrumental SST. The magnitude of these coral Sr/Ca–SST trends are in line with other coral Sr/Ca reconstructions and further supports the conclusion that decadal and secular trends in SST at shallow reef sites are not fully reflected in gridded SST products.

A similar amplification also appears in the results of a digenetically altered coral from Ningaloo reef in Western Australia reported by Müller et al. (2001). However, in New Caledonia *Porites*, it is noteworthy that a network of coral Sr/Ca records shows no apparent amplification of the long-term secular SST signal (DeLong et al., 2007).

Coral Sr/Ca Summary

Coral skeletal Sr/Ca seasonal variability has proven to be highly correlated with monthly instrumental water temperature at many sites and appears to be a potentially useful tracer in paleoceanographic studies. As noted above, the results of several studies have raised questions/concerns about the accuracy of the coral Sr/Ca thermometer in some modern corals that are important since these issues may directly affect interpretation of Sr/Ca data from fossil corals and reports of significant tropical Pacific and Caribbean cooling prior to 10000 years BP (e.g., Beck et al., 1997; Guilderson et al., 1994). Although this tracer clearly has the potential to greatly improve our understanding of oceanic conditions, there remains important work to be done to more fully understand the significance of all modes of Sr/Ca variability in corals.

See also: **Paleoceanography: Paleoceanography An Overview.**
Paleoceanography, Biological Proxies: Benthic Foraminifera; Corals, Sclerosponges and Mollusks; Planktic Foraminifera.
Paleoceanography, Physical and Chemical Proxies: Nutrient Proxies.

References

- Alibert C and McCulloch MT (1997) Strontium/calcium ratios in modern *Porites* corals from the Great Barrier Reef as a proxy for sea surface temperature: Calibration of the thermometer and monitoring of ENSO. *Paleoceanography* 12: 345–363.
- Anand P, Elderfield H, and Conte MH (2003) Calibration of Mg/Ca thermometry in planktonic foraminifera from a sediment trap time-series. *Paleoceanography* 18(2): 1050. <http://dx.doi.org/10.1029/2002PA000846>.
- Arbuszewski J, deMenocal P, Kaplan A, and Farmer CE (2010) On the fidelity of shell-derived $\delta^{18}\text{O}_{\text{seawater}}$ estimates. *Earth and Planetary Science Letters* 300: 185–196.
- Bagnato S, Linsley BK, Howe SS, Wellington GM, and Salinger J (2004) Evaluating the use of the massive coral *Diploastrea heliopora* for paleoclimate reconstruction. *Paleoceanography* 19(PA 1032): 1–12.
- Bamberg A, Rosenthal Y, Paul A, et al. (2010) Reduced North Atlantic Central Water formation in response to Holocene ice sheet melting. *Geophysical Research Letters* 37: L17705. <http://dx.doi.org/10.1029/2010GL043878>.
- Barker S and Elderfield H (2002) Foraminiferal calcification response to glacial–interglacial changes in atmospheric CO₂. *Science* 297: 833–836.
- Beck JW, Recy J, Taylor F, Edwards RL, and Cabloch G (1997) Abrupt changes in early Holocene tropical sea surface temperature derived from coral records. *Nature* 385: 705–707.
- Beck WJ, Edwards RL, Ito E, et al. (1992) Sea-surface temperature from coral skeletal strontium/calcium ratios. *Science* 257: 644–647.
- Billups K (2003) Application of benthic foraminiferal Mg/Ca ratios to questions of early Cenozoic climate change. *Earth and Planetary Science Letters* 209: 181–195.
- Billups K and Schrag DP (2002) Paleotemperatures and ice volume of the past 27 Myr revisited with paired Mg/Ca and $^{18}\text{O}/^{16}\text{O}$ measurements on benthic foraminifera. *Paleoceanography* 17: 1–11.
- Brown SJ and Elderfield H (1996) Variations in Mg/Ca and Sr/Ca ratios of planktonic foraminifera caused by postdepositional dissolution: Evidence of shallow Mg-dependent dissolution. *Paleoceanography* 11: 543–551.
- Bryan SP and Marchitto TM (2008) Mg/Ca-temperature proxy in benthic foraminifera: New calibrations from the Florida Straits and a hypothesis regarding Mg/Li. *Paleoceanography* 23: PA2220. <http://dx.doi.org/10.1029/2007PA001553>.
- Cacho I, Shackleton N, Elderfield H, Sierrac FJ, and Grimald JO (2006) Glacial rapid variability in deep-water temperature and $\delta^{18}\text{O}$ from the Western Mediterranean Sea. *Quaternary Science Reviews* 25: 3294–3311.
- Cléroux C, Cortijo E, Anand P, et al. (2008) Mg/Ca and Sr/Ca ratios in planktonic foraminifera: Proxies for upper water column temperature reconstruction. *Paleoceanography* 23: PA3214. <http://dx.doi.org/10.1029/2007PA001505>.
- Cohen AL, Lane GD, Hart SR, and Lobel PS (2001) Kinetic control of skeletal Sr/Ca in a symbiotic coral: Implications for the paleoceanographic proxy. *Paleoceanography* 16: 20–26.
- Cramer BS, Miller KG, Barrett PJ, and Wright JD (2011) Late Cretaceous–Neogene trends in deep ocean temperature and continental ice volume: Reconciling records of benthic foraminiferal geochemistry ($\delta^{18}\text{O}$ and Mg/Ca) with sea level history. *Journal of Geophysical Research* 116(C12): C12023. <http://dx.doi.org/10.1029/2011JC007255>.
- Cronin TM, Dowsett HJ, Dwyer GS, Baker PA, and Chandler MA (2005) Mid-Pliocene deep-sea bottom-water temperatures based on ostracode Mg/Ca ratios. *Marine Micropaleontology* 54: 249–261.
- Curry WB and Marchitto TM (2008) A secondary ionization mass spectrometry calibration of *Cibicides* *pachyderma* Mg/Ca with temperature. *Geochemistry, Geophysics, Geosystems* 9(4): Q04009.
- de Garidel-Thoron T, Rosenthal Y, Bassinot F, and Beaufort L (2005) Stable sea surface temperatures in the western Pacific warm pool over the past 1.75 million years. *Nature* 433: 294–298.
- de Garidel-Thoron T, Rosenthal Y, Beaufort L, Bard E, and Mix A (2007) A multiproxy assessment of the equatorial Pacific hydrography during the last deglaciation. *Paleoceanography* 22: PA3204. <http://dx.doi.org/10.1029/2006PA001269>.
- De Villiers S, Nelson BK, and Chivas AR (1995) Biological controls on coral Sr/Ca and $\delta^{18}\text{O}$ reconstructions of sea surface temperatures. *Science* 269: 1247–1249.
- De Villiers S, Shen GT, and Nelson BK (1994) The Sr/Ca-temperature relationship in coralline aragonite: Influence of variability in (Sr/Ca)_{seawater} and skeletal growth parameters. *Geochimica et Cosmochimica Acta* 58: 197–208.
- Dekens PS, Lea DW, Pak DK, and Spero HJ (2002) Core top calibration of Mg/Ca in tropical foraminifera: Refining paleo-temperature estimation. *Geochemistry, Geophysics, Geosystems* 3(4): 1022.
- DeLong KL, Quinn TM, and Taylor FW (2007) Reconstructing twentieth-century sea surface temperature variability in the southwest Pacific: A replication study using multiple coral Sr/Ca records from New Caledonia. *Paleoceanography* 22: PA4212.
- Dueñas-Bohórquez A, da Rocha RE, Kuroyanagi A, Bijma J, and Reichert GJ (2009) Effect of salinity and seawater calcite saturation state on Mg and Sr incorporation in cultured planktonic foraminifera. *Marine Micropaleontology* 73: 178–189.
- Dwyer GS, Cronin TM, and Baker PA (2002) Trace elements in marine ostracodes. *The Ostracoda: Applications in Quaternary Research*, pp. 205–224. Washington, DC: American Geophysical Union.
- Dwyer GS, Cronin TM, Baker PA, Raymo ME, Buzas JS, and Corrège T (1995) North Atlantic deepwater temperature change during late Pliocene and late Quaternary climatic cycles. *Science* 270: 1347–1351.
- Eggins S, De Deckker P, and Marshall J (2003) Mg/Ca variation in planktonic foraminifera tests: Implications for reconstructing palaeo-seawater temperature and habitat migration. *Earth and Planetary Science Letters* 212: 291–306.
- Eggins S, Sadekov A, and De Deckker P (2004) Modulation and daily banding of Mg/Ca in *Orbulina universa* tests by symbiont photosynthesis and respiration: A complication for seawater thermometry? *Earth and Planetary Science Letters* 225: 411–419.

- Elderfield H and Ganssen G (2000) Past temperature and $\delta^{18}\text{O}$ of surface ocean waters inferred from foraminiferal Mg/Ca ratios. *Nature* 405: 442–445.
- Elderfield H, Greaves M, Barker S, et al. (2009) A record of bottom water temperature and seawater $\delta^{18}\text{O}$ for the Southern Ocean over the past 440 kyr based on Mg/Ca of benthic foraminiferal *Uvigerina* spp. *Quaternary Science Reviews* 29: 160–170.
- Elderfield H, Vautravers M, and Cooper M (2002) The relationship between shell size and Mg/Ca, Sr/Ca, $\delta^{18}\text{O}$, and $\delta^{13}\text{C}$ of species of planktonic foraminifera. *Geochemistry, Geophysics, Geosystems* 3: <http://dx.doi.org/10.1029/2001GC000194>.
- Elderfield H, Yu J, Anand P, Kiefer T, and Nyland B (2006) Calibrations for benthic foraminiferal Mg/Ca paleothermometry and the carbonate ion hypothesis. *Earth and Planetary Science Letters* 250: 633–649.
- Ferguson JE, Henderson GM, Kucera M, and Rickaby REM (2008) Systematic change of foraminiferal Mg/Ca ratios across a strong salinity gradient. *Earth and Planetary Science Letters* 265: 153–166.
- Ferrier-Pages C, Boisson F, Allemand D, and Tambutte E (2003) Kinetics of strontium uptake in the scleractinian coral *Stylophora pistillata*. *Marine Ecology Progress Series* 245: 93–100.
- Goodkin NF, Hughen KA, and Cohen AL (2007) A multicoral calibration method to approximate a universal equation relating Sr/Ca and growth rate to sea surface temperature. *Paleoceanography* 22.
- Groeneveld J and Chiessi CM (2011) Mg/Ca ratios of *Globorotalia inflata* as a recorder of permanent thermocline temperatures in the South Atlantic. *Paleoceanography* 26: PA2203.
- Guilderson TP, Fairbanks RG, and Rubenstone JL (1994) Tropical temperature variations since 20,000 years ago: Modulating interhemispheric climate change. *Science* 263: 663–665.
- Hasiuk FJ and Lohmann KC (2010) Application of calcite Mg partitioning functions to the reconstruction of paleocean Mg/Ca. *Geochimica et Cosmochimica Acta* 74(23): 6751–6763. <http://dx.doi.org/10.1016/j.gca.2010.07.030>.
- Hathorne EC, Alard O, James RH, and Rogers NW (2003) Determination of intratest variability of trace elements in foraminifera by laser ablation inductively coupled plasma-mass spectrometry. *Geochemistry, Geophysics, Geosystems* 4: <http://dx.doi.org/10.1029/2003GC000539>.
- Hathorne EC, James RH, and Lampitt RS (2009) Environmental versus biomineralization controls on the intratest variation in the trace element composition of the planktonic foraminifera *G. inflata* and *G. scitula*. *Paleoceanography* 24: <http://dx.doi.org/10.1029/2009PA001742>.
- Healey SL, Thunell RC, and Corliss BH (2008) The Mg/Ca-temperature relationship of benthic foraminiferal calcite: New core-top calibrations in the $<4^\circ\text{C}$ temperature range. *Earth and Planetary Science Letters* 272: 523–530.
- Hendry KR, Rickaby REM, Meredith MP, and Elderfield H (2009) Controls on stable isotope and trace metal uptake in *Neogloboquadrina pachyderma* (sinistral) from an Antarctic sea-ice environment. *Earth and Planetary Science Letters* 278: 67–77.
- Honisch B and Hemming NG (2004) Ground-truthing the boron isotope-paleo-pH proxy in planktonic foraminifera shells: Partial dissolution and shell size effects. *Paleoceanography* 19: <http://dx.doi.org/10.1029/2004PA001026>.
- Hoogakker BAA, Klinkhammer GP, Elderfield H, Rohling EJ, and Hayward C (2009) Mg/Ca paleothermometry in high salinity environments. *Earth and Planetary Science Letters* 284: 583–589.
- Huang KF, You CF, Lin HL, and Shieh YT (2008) In situ calibration of Mg/Ca ratio in planktonic foraminiferal shell using time series sediment trap: A case study of intense dissolution artifact in the South China Sea. *Geochemistry, Geophysics, Geosystems* 9(4): <http://dx.doi.org/10.1029/2007GC001660>.
- Kisakurek B, Eisenhauer A, Böhm F, Garbe-Schonberg D, and Erez J (2008) Controls on shell Mg/Ca and Sr/Ca in cultured planktonic foraminifera, *Globigerinoides ruber* (white). *Earth and Planetary Science Letters* 273: 260–269.
- Koutavas A, DeMenocal P, Olive GC, and Lynch-Stieglitz J (2006) Mid-Holocene El Niño–Southern Oscillation (ENSO) attenuation revealed by individual foraminifera in eastern tropical Pacific sediments. *Geology* 34: 993–996. <http://dx.doi.org/10.1130/G22810A.1>.
- Kozdon R, Eisenhauer A, Weinelt M, Meland MY, and Nurnberg D (2009) Reassessing Mg/Ca temperature calibrations of *Neogloboquadrina pachyderma* (sinistral) using paired $\delta^{44}\text{Ca}$ and Mg/Ca measurements. *Geochemistry, Geophysics, Geosystems* 10(3): <http://dx.doi.org/10.1029/2008GC002169>.
- Kristjansdottir GB, Lea DW, Jennings AE, Pak DK, and Belanger C (2007) New spatial Mg/Ca-temperature calibrations for three Arctic, benthic foraminifera and reconstruction of north Iceland shelf temperature for the past 4000 years. *Geochemistry, Geophysics, Geosystems* 8(3): Q03P21. <http://dx.doi.org/10.1029/2006GC001425>.
- Kunioka D, Shirai S, Takahata N, and Sano Y (2006) Microdistribution of Mg/Ca, Sr/Ca, and Ba/Ca ratios in *Pulleniatina obliquiloculata* test by using a NanoSIMS: Implication for the vital effect mechanism. *Geophysics Geochemistry Geosystems* 7(12): Q12P20. <http://dx.doi.org/10.1029/2006GC001280>.
- Langen PJV, Pak DK, Spero HJ, and Lea DW (2005) Effects of temperature on Mg/Ca in neogloboquadrinid shells determined by live culturing. *Geochemistry, Geophysics, Geosystems* 6: <http://dx.doi.org/10.1029/2005GC000989>.
- Lea DW, Mashiotta TA, and Spero HJ (1999) Controls on magnesium and strontium uptake in planktonic foraminifera determined by live culturing. *Geochimica et Cosmochimica Acta* 63: 2369–2379.
- Lea DW, Pak DK, and Paradis G (2005) Influence of volcanic shards on foraminiferal Mg/Ca in a core from the Galápagos region. *Geochemistry, Geophysics, Geosystems* 6: <http://dx.doi.org/10.1029/2005GC000970>.
- Lea DW, Pak DK, and Spero HJ (2000) Climate impact of late Quaternary equatorial Pacific sea surface temperature variations. *Science* 289: 1719–1724.
- Lear CH, Elderfield H, and Wilson PA (2000) Cenozoic deep-sea temperatures and global ice volumes from Mg/Ca in benthic foraminiferal calcite. *Science* 287: 269–272.
- Lear CH, Mawbey EM, and Rosenthal Y (2010) Ocean temperatures and saturation state across Cenozoic glacial transitions. *Paleoceanography* 25: <http://dx.doi.org/10.1029/2009PA001880>.
- Lear CH, Rosenthal Y, Coxall HK, and Wilson PA (2004) Late Eocene to early Miocene ice sheet dynamics and the global carbon cycle. *Paleoceanography* 19: PA4015. <http://dx.doi.org/10.1029/2004PA001039>.
- Lear CH, Rosenthal Y, and Slowey N (2002) Benthic foraminiferal Mg/Ca-paleothermometry: A revised core-top calibration. *Geochimica et Cosmochimica Acta* 66: 3375–3387.
- Linsley BK, Rosenthal Y, and Oppo DW (2010) Evolution of the Indonesian throughflow and western Pacific warm pool during the Holocene. *Nature Geoscience* 3: 578–583. <http://dx.doi.org/10.1038/ngeo920>.
- Linsley BK, Wellington GM, and Schrag DP (2000) Decadal sea surface temperature variability in the sub-tropical South Pacific from 1726 to 1997 AD. *Science* 290: 1145–1148.
- Linsley BK, Wellington GM, Schrag DP, Ren L, Salinger MJ, and Tudhope AW (2004) Geochemical evidence from corals for changes in the amplitude and spatial pattern of South Pacific interdecadal climate variability over the last 300 years. *Climate Dynamics* 22(22): 1–11. <http://dx.doi.org/10.1007/s00382-003-0364-y>.
- Linsley BK, Kaplan A, Gouriou Y, et al. (2006) Tracking the extent of the South Pacific Convergence Zone since the early 1600s. *Geochemistry, Geophysics, Geosystems* 7(4): <http://dx.doi.org/10.1029/2005GC001115>.
- Marchitto TM, Bryan SP, Curry WB, and McCorkle DC (2007) Mg/Ca temperature calibration for the benthic foraminifer *Cibicides* pachyderma. *Paleoceanography* 22: 1–9. <http://dx.doi.org/10.1029/2006PA001287>.
- Marshall JF and McCulloch MT (2002) An assessment of the Sr/Ca ratio in shallow water hermatypic corals as a proxy for sea surface temperature. *Geochimica et Cosmochimica Acta* 66(18): 3263–3280.
- Martin PA, Lea DW, Rosenthal Y, Papenfuss TP, and Sarnthein M (2002) Late Quaternary deep-sea temperatures inferred from benthic foraminiferal magnesium. *Earth and Planetary Science Letters* 198: 193–209.
- Mashiotta TA, Lea DW, and Spero HJ (1999) Glacial–interglacial changes in subantarctic sea surface temperature and $\delta^{18}\text{O}$ -water using foraminiferal Mg. *Earth and Planetary Science Letters* 170: 417–432.
- Mathien-Blard E and Bassinot F (2009) Salinity bias on the foraminifera Mg/Ca thermometry: Correction procedure and implications for past ocean hydrographic reconstructions. *Geochemistry, Geophysics, Geosystems* 10: <http://dx.doi.org/10.1029/2008GC002353>.
- McConnell MC and Thunell RC (2005) Calibration of the planktonic foraminiferal Mg/Ca paleothermometer: Sediment trap results from the Guaymas Basin Gulf of California. *Paleoceanography* 20: <http://dx.doi.org/10.1029/2004PA001077>.
- McKenna VM and Prell WL (2005) Calibration of the Mg/Ca of *Globorotalia truncatulinoides* (r) for the reconstruction of marine temperature gradients. *Paleoceanography* 19: <http://dx.doi.org/10.1029/2000PA000604>.
- Meibom A, Stage M, Wooden J, et al. (2003) Monthly strontium/calcium oscillations in symbiotic coral argonite: Biological effects limiting the precision of the paleotemperature proxy. *Geophysical Research Letters* 30: <http://dx.doi.org/10.1029/2002GL016864>.
- Meland MYE, Jansen E, Elderfield H, Dokken TM, Olsen A, and Bellerby RGJ (2006) Mg/Ca ratios in the planktonic foraminifer *Neogloboquadrina pachyderma* (sinistral) in the northern North Atlantic/Nordic Seas. *Geochemistry, Geophysics, Geosystems* 7(6): <http://dx.doi.org/10.1029/2005GC001078>.
- Mohitadi M, Steinke S, Groeneveld J, et al. (2009) Low-latitude control on seasonal and interannual changes in planktonic foraminiferal flux and shell geochemistry off south Java: A sediment trap study. *Paleoceanography* 24: <http://dx.doi.org/10.1029/2008PA001636>.
- Müller A, Gagan MK, and McCulloch MT (2001) Early marine diagenesis in corals and geochemical consequences for paleoceanographic reconstructions. *Geophysical Research Letters* 28: 4471–4474.
- Nurhati IS, Cobb KM, Charles CD, and Dunbar RB (2009) Late 20th century warming and freshening in the central Tropical Pacific. *Geophysical Research Letters* 36.

- Nürnberg D (1995) Magnesium in tests of *Neoglobobulimina pachyderma* sinistral from high northern and southern latitudes. *Journal of Foraminiferal Research* 25: 350–368.
- Nürnberg D, Bijma J, and Hemleben C (1996) Assessing the reliability of magnesium in foraminiferal calcite as a proxy for water mass temperature. *Geochimica et Cosmochimica Acta* 60: 803–814.
- Oppo DW, Rosenthal Y, and Linsley BK (2009) 2,000-year-long temperature and hydrology reconstructions from the western Pacific warm pool. *Nature* 460: 1113–1116.
- Pena LD, Calvo E, Cacho I, Eggins S, and Pelejero C (2005) Identification and removal of Mn–Mg-rich contaminant phases on foraminiferal tests: Implications for Mg/Ca past temperature reconstructions. *Geochemistry, Geophysics, Geosystems* 6: <http://dx.doi.org/10.1029/2005GC000930>.
- Pena LD, Cacho I, Calvo E, Pelejero C, Eggins S, and Sadekov A (2008) Characterization of contaminant phases in foraminifera carbonates by electron microprobe mapping. *Geochemistry, Geophysics, Geosystems* 9: <http://dx.doi.org/10.1029/2008GC000218>.
- Raitzsch M, Kuhnert H, Groeneveld J, and Bickert T (2008) Benthic foraminifer Mg/Ca anomalies in South Atlantic core top sediments and their implications for paleothermometry. *Geochemistry, Geophysics, Geosystems* 9: <http://dx.doi.org/10.1029/2007GC001788>.
- Rathmann S, Hess S, Kuhnert H, and Mulitz S (2004) Mg/Ca ratios of the benthic foraminifera *Oridorsalis umbonatus* obtained by laser ablation from core top sediments: Relationship to bottom water temperature. *Geochemistry, Geophysics, Geosystems* 5: <http://dx.doi.org/10.1029/2004GC000808>.
- Regenberg M, Nürnberg N, Steph S, Garbe-Schönberg D, and Tiedemann R (2006) Assessing the effect of dissolution on planktonic foraminiferal Mg/Ca ratios: Evidence from Caribbean core tops. *Geochemistry, Geophysics, Geosystems* 7: <http://dx.doi.org/10.1029/2005GC001019>.
- Regenberg M, Steph S, Nürnberg N, Tiedemann R, and Garbe-Schönberg D (2009) Calibrating Mg/Ca ratios of multiple planktonic foraminiferal species with $\delta^{18}\text{O}$ -calcification temperatures: Paleothermometry for the upper water column. *Earth and Planetary Science Letters* 278: 324–336.
- Ries JB (2004) Effect of ambient Mg/Ca ratio on Mg fractionation in calcareous marine invertebrates: A record of the oceanic Mg/Ca ratio over the Phanerozoic. *Geology* 32(11): 981–984. <http://dx.doi.org/10.1130/G20851.1>.
- Rosenthal Y and Boyle EA (1993) Factors controlling the fluoride content of planktonic foraminifera: An evaluation of its paleoceanographic applicability. *Geochimica et Cosmochimica Acta* 57: 335–346.
- Rosenthal Y, Boyle EA, and Slowey N (1997) Environmental controls on the incorporation of Mg, Sr, and Cd into benthic foraminiferal shells from Little Bahama Bank: Prospects for thermocline paleoceanography. *Geochimica et Cosmochimica Acta* 61: 3633–3643.
- Rosenthal Y, Lear CH, Oppo DW, and Linsley BK (2006) Temperature and carbonate ion effects on Mg/Ca and Sr/Ca ratios in benthic foraminifera: The aragonitic species *Hoeglundina elegans*. *Paleoceanography* 21: <http://dx.doi.org/10.1029/2005PA001158>.
- Rosenthal Y and Lohmann GP (2002) Accurate estimation of sea surface temperatures using dissolution-corrected calibrations for Mg/Ca paleothermometry. *Paleoceanography* 17: <http://dx.doi.org/10.1029/2001PA000749>.
- Rosenthal Y, Lohmann GP, Lohmann KC, and Sherrell RM (2000) Incorporation and preservation of Mg in *G. sacculifer*: Implications for reconstructing sea surface temperatures and the oxygen isotopic composition of seawater. *Paleoceanography* 15: 135–145.
- Rosenthal Y, Oppo DW, and Linsley BK (2003) The amplitude and phasing of climate change during the last deglaciation in the Sulu Sea, western equatorial Pacific. *Geophysical Research Letters* 30: <http://dx.doi.org/10.1029/2002GL016612>.
- Rosenthal Y, Perron-Cashman S, Lear CH, et al. (2004) Interlaboratory comparison study of Mg/Ca and Sr/Ca measurements in planktonic foraminifera for paleoceanographic research. *Geochemistry, Geophysics, Geosystems* 5: <http://dx.doi.org/10.1029/2003GC000650>.
- Rosenthal Y, Morley A, Barras C, et al. (2011) Temperature calibration of Mg/Ca ratios in the intermediate water benthic foraminifer *Hyalina balthica*. *Geochemistry, Geophysics, Geosystems* 12(4): Q04003. <http://dx.doi.org/10.1029/2010GC003333>.
- Russell AD, Hoenisch B, Spero HJ, and Lea DW (2004) Effects of seawater carbonate ion concentration and temperature on shell U, Mg, and Sr in cultured planktonic foraminifera. *Geochimica et Cosmochimica Acta* 68: 4347–4361.
- Sadekov A, Eggins SM, De Deckker P, and Kroon D (2008) Uncertainties in seawater thermometry deriving from intratest and intertest Mg/Ca variability in *Globigerinoides ruber*. *Paleoceanography* 23: <http://dx.doi.org/10.1029/2007PA001452>.
- Sadekov A, Eggins SM, De Deckker P, Ninnemann U, Kuhnt W, and Bassinot F (2009) Surface and subsurface seawater temperature reconstruction using Mg/Ca microanalysis of planktonic foraminifera *Globigerinoides ruber*, *Globigerinoides sacculifer*, and *Pulleniatina obliquiloculata*. *Paleoceanography* 24: <http://dx.doi.org/10.1029/2008PA001664>.
- Sadekov AU, Eggins S, Klinkhammer GP, and Rosenthal Y (2010) Effects of seafloor and laboratory dissolution on the Mg/Ca composition of *Globigerinoides sacculifer* and *Orbulina universa* tests – A laser ablation ICPMS microanalysis perspective. *Earth and Planetary Science Letters* 292: 312–324.
- Sadekov AY, Eggins SM, and De Deckker P (2005) Characterization of Mg/Ca distributions in planktonic foraminifera species by electron microprobe mapping. *Geochemistry, Geophysics, Geosystems* 6: <http://dx.doi.org/10.1029/2005GC000973>.
- Schmidt MW, Spero HJ, and Lea DW (2004) Links between salinity variation in the Caribbean and North Atlantic thermohaline circulation. *Nature* 428: 160–163.
- Segev E and Erez J (2006) Effect of Mg/Ca ratio in seawater on shell composition in shallow benthic foraminifera. *Geochemistry, Geophysics, Geosystems* 7: Q02P09. <http://dx.doi.org/10.1029/2005GC000969>.
- Shen C-C, Lee T, Chen C-Y, Wang C-H, Dai C-F, and Li L-A (1996) The calibration of D [Sr/Ca] versus sea surface temperature relationship for *Porites* corals. *Geochimica et Cosmochimica Acta* 60: 3849–3858.
- Skinner LC and Elderfield H (2005) Constraining ecological and biological bias in planktonic foraminiferal Mg/Ca and $\delta^{18}\text{O}$: A multispecies approach to proxy calibration testing. *Paleoceanography* 20(8): <http://dx.doi.org/10.1029/2004PA00105>.
- Sosdian S and Rosenthal Y (2009) Deep sea temperature and ice volume changes across the Pliocene–Pleistocene climate transitions. *Science* 325: 306–310.
- Spero HJ and Lea DW (1993) Intraspecific stable isotope variability in the planktic foraminifera *Globigerinoides sacculifer*: Results from laboratory experiments. *Marine Micropaleontology* 22: 221–234.
- Steinke S, Chiu H-Y, Yu P-S, et al. (2005) Mg/Ca ratios of two *Globigerinoides ruber* (white) morphotypes: Implications for reconstructing past tropical/subtropical surface water conditions. *Geochemistry, Geophysics, Geosystems* 6: <http://dx.doi.org/10.1029/2005GC000926>.
- Stott L, Poulsen C, Lund S, and Thunell R (2002) Super ENSO and global climate oscillations at millennial time scales. *Science* 297: 222–226.
- Stott L, Cannariato K, Thunell R, Haug GH, Koutavas A, and Lund S (2004) Decline of surface temperature and salinity in the western tropical Pacific Ocean in the Holocene epoch. *Nature* 431: 56–59.
- Toyofuku T and Kitazato H (2005) Micromapping of Mg/Ca values in cultured specimens of the high-magnesium benthic foraminifera. *Geochemistry, Geophysics, Geosystems* 6: <http://dx.doi.org/10.1029/2005GC0009619>.
- Toyofuku T, Kitazato H, Kawahata H, Tsuchiya M, and Nohara M (2000) Evaluation of Mg/Ca thermometry in foraminifera: Comparison of experimental results and measurements in nature. *Paleoceanography* 15: 456–464.
- Visser K, Thunell RC, and Stott L (2003) Magnitude and timing of temperature change in the Indo-Pacific warm pool during deglaciation. *Nature* 421: 152–155.
- Wara MW, Ravelo AC, and Delaney ML (2005) Permanent El Niño-like conditions during the Pliocene warm period. *Science* 309: 758–761.
- Wu H (2010) Assessing replicated coral trace element (Sr/Ca and Mg/Ca) variability and skeletal growth records from the tropical Pacific, University at Albany–State University of New York, 211 pp.
- Wu H, Linsley BK, and Schrag DP (in review) Replicated *Porites lobata* coral Sr/Ca records in the eastern equatorial Pacific (Clipperton Atoll) document increasing sea surface temperature with implications for salinity variability over the last century. *Paleogeography, Palaeoclimatology, Palaeoecology*.
- Xu J, Holbourn A, Kuhnt W, Jian Z, and Kawamura H (2008) Changes in the thermocline structure of the Indonesian outflow during Terminations I and II. *Earth and Planetary Science Letters* 273: 152–162.
- Yu J and Elderfield H (2008) Mg/Ca in the benthic foraminifera *Cibicides wuellerstorfi* and *Cibicides mundulus*: Temperature versus carbonate ion saturation. *Earth and Planetary Science Letters* 276: 129–139.

Magnetic Proxies and Susceptibility

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Introduction: Magnetic Susceptibility and Environmental Magnetism in Paleoceanography

Magnetic susceptibility (MS) is one of the most commonly employed proxies in paleoresearch and is routinely measured on materials collected by the Integrated Ocean Drilling Program (IODP, www.iodp.org) and international piston coring cruises. MS measurements are not restricted to the marine realm and are commonly employed in terrestrial drilling programs like the Antarctic Geological Drilling Program (Andrill, www.andrill.org) and the International Continental Drilling Program (ICDP, www.icdp-online.org), as well as many lake coring, terrestrial stratigraphic, Quaternary geological, and archaeological projects. The widespread use of MS reflects its sensitivity as an indicator of lithologic variation, coupled with the practicality and versatility of measurement. MS measurements can be made on many types of samples, from sediment cores and boreholes to discrete samples of any shape or even directly on an outcrop. Setup costs are low and running costs are negligible, data acquisition is fast, sample preparation is minimal, and as it is both nondestructive and does not interfere with other measurements, it is often the first measurement made in many studies. These factors make MS uniquely applicable to a host of problems, yet MS is still poorly understood by many in the Quaternary and geological communities. Even for those who specialize in magnetism, determining what exactly drives MS variations within any particular record is a nontrivial undertaking. MS is just the simplest of many measurements that make up the tool kit known as environmental magnetism (Thompson and Oldfield, 1986) and often many other magnetic measurements are needed to truly understand what drives MS changes (e.g., Table 1). Here we review and assess both the potential and the opportunities MS and environmental magnetism have for deciphering the Quaternary paleoceanographic record.

What Is MS?

All substances respond to a magnetic field, the question is by how much and in what direction. MS is the relationship between an induced magnetization (M) to an applied magnetic field (H), or simply how magnetizable a material becomes when placed in an external magnetic field.

Low-field (~ 0.1 mT) volumetric MS (κ) is the type typically measured in most paleostudies:

$$\kappa = M/H$$

where M is the magnetic moment induced per unit volume (amperes per meter; Am^{-1}) and H is the applied field (Am^{-1}). As both M and H have the same système international d'unités (SI) units, κ is dimensionless. Centimeter–gram–second (cgs) units are also dimensionless, where 4π SI = 1 cgs. Unfortunately, it is often not explicitly stated which system is used in

many MS studies, making comparisons difficult. To exclude density-related effects and permit comparison between diverse samples, volumetric susceptibility can be density-corrected to produce mass-normalized MS values (χ).

$$\chi = \kappa/\rho$$

where ρ is the sample density. SI units of χ are $\text{m}^3 \text{kg}^{-1}$; cgs units are $\text{emu Oe}^{-1} \text{g}^{-1}$.

Influences on MS Values

Different materials, their concentration, and their magnetic grain size and shape, can all affect the MS values of any sample. Resulting from the orbital and spin motions of electrons and their interactions, all materials produce a magnetic response when exposed to a magnetic field, though the nature and strength of this response varies between materials. In the transition metals (Fe, Co, Ni), and their compounds, the magnetization can be particularly strong with parallel (but unequal) coupling of unpaired electrons within a crystal lattice producing a strong magnetization and ferromagnetic or ferrimagnetic behavior (Figure 1; Table 2). True ferromagnetism only exists in the elemental state of these transition metals. As these elements are commonly oxidized (or reduced) into various ferrimagnetic forms in the environment, ferromagnetic behavior is extremely rare in natural samples. Common ferrimagnetic minerals include magnetite, maghemite, and greigite (Table 2). These possess strong positive magnetic susceptibilities and, when present, these minerals will often dominate the MS signal (Figure 1). When atomic arrangements are equal, but opposite, there is no (theoretical) net magnetization as parallel pairings cancel out (antiferromagnetism). However, for a variety of reasons, alignment may not be perfectly parallel in antiferromagnetic minerals and a residual magnetization may result, inducing canted antiferromagnetic behavior (Figure 1). Hematite and goethite are common canted antiferromagnetic minerals possessing weak positive susceptibility (Figure 1; Table 2). Paramagnetic and diamagnetic behavior (Figure 1; Table 2) is induced in samples that have some or no unpaired electrons, respectively. Unlike ferro/ferrimagnetic and canted antiferromagnetic behavior, any induced magnetization in a paramagnetic or diamagnetic material is nonpermanent and lost upon removal of the field. Common examples of paramagnetic magnetic materials are ilmenite, biotite, lepidocrocite, siderite, and pyrite, all exhibiting weakly positive magnetic susceptibilities. Common diamagnetic materials include quartz, calcite, and water with weakly negative magnetic susceptibilities (Figure 1; Table 2). In the absence of ferrimagnetic minerals, paramagnetic or diamagnetic minerals may drive the MS signal.

In many environments, the concentration and type of magnetic minerals are often the dominant controls on MS.

Table 1 Some common magnetic parameters and the discrimination they afford

Parameter	Notes
κ, χ	Magnetic susceptibility (MS): the ratio of magnetization induced to the intensity of the magnetizing field. Measured within a small (~ 0.1 mT) reversible magnetic field either volumetrically (κ) or on a mass-normalized basis (χ). Roughly proportional to the concentration of ferrimagnetic minerals within a sample Units: κ (dimensionless), χ ($\text{m}^3 \text{kg}^{-1}$)
χ_{fd}	Frequency-dependent susceptibility: variation of MS with frequency. Measured at 0.46 and 4.6 kHz on a Bartington MS2B. Indicative of superparamagnetic (SP) grains around the SP/single domain (SD) boundary ($\sim 0.03 \mu\text{m}$). At higher frequencies (χ_{hf}), grains at the SP/SD boundary respond as SD grains; thus, high frequency MS is proportionally lower than low frequency (χ_{lf}) MS Units: expressed as a % of χ_{lf}
ARM	Anhyseretic remanent magnetization: acquired when a sample is placed in a decreasing alternating magnetic field (e.g., 100 mT) in the presence of a small steady biasing field (e.g., 0.1 mT). Sensitive to both ferrimagnetic concentration and ferrimagnetic grain size Units: $\text{Am}^2 \text{kg}^{-1}$
(S)IRM	(Saturation) isothermal remanent magnetization (SIRM): magnetic remanence induced in a sample upon removal of the field. Usually attained in increasing field strengths (e.g., IRM 20 mT, IRM 100 mT) up to the saturating field (SIRM), which is usually an arbitrary term for largest field generated in the laboratory (~ 1000 mT). SIRM is an indicator for the concentration of magnetic minerals but also responds to grain size variations Units: $\text{Am}^2 \text{kg}^{-1}$
SIRM/ χ	This ratio can be diagnostic of either mineralogy, or if mineral type and concentrations are similar, magnetic grain size Units: Am^{-1}
ARM/ χ	This ratio is dominantly used to determine magnetic grain size variations, but it is also somewhat sensitive to concentration Units: Am^{-1}
IRM $\times$ mT/SIRM	Ratio of remanence acquired in a smaller field than SIRM to the SIRM. Different field sizes can reveal different aspects of the magnetic signal independent of magnetic concentration. For example, low fields such as 20 mT are sensitive to coarse-grained magnetite, whereas acquisition at higher fields (e.g., 300 mT) can indicate hematite contributions Units: dimensionless
(Bo)cr SIRM- $\times$ mT/ SIRM S-ratio	Demagnetization parameters: obtained by applying reversed magnetic fields to a previously saturated sample. The coercivity of remanence (Bo)cr is the reverse field (mT) at which the SIRM returns to zero. Loss in other fields SIRM- $\times$ mT/SIRM can be expressed as a ratio. Loss at a field of 100 mT (which discriminates between ferrimagnetic and canted antiferromagnetic minerals) has been termed the S-ratio Units: (Bo)cr (mT), SIRM- $\times$ mT/SIRM, and the S-ratio are dimensionless

However, magnetic grain size and shape may also influence the MS signal. Ferrimagnetic grains organize their magnetization into different regions, or domains, to achieve minimum energy states; this results in a grain size dependence of MS values. For example, in magnetite ultrafine particles known as superparamagnetic (SP) grains ($<0.03 \mu\text{m}$) and larger particles known as multidomain (MD) grains ($>\sim 10 \mu\text{m}$) possess higher MS values than smaller single domain (SD) and pseudo SD (PSD) grains in the 0.03 to $\sim 10 \mu\text{m}$ size range (Figure 2). Anisotropy also affects MS values as not all grains are spherical. Magnetic energies favor ordering along the easiest direction, the long axis of elongate grains, with grain shape having as much as a factor of 2 influence on MS. For more information on the factors affecting MS measurement, see Thompson and Oldfield (1986), Dearing (1999), Evans and Heller (2003), and Tauxe (2010).

MS Measurement

MS is the ratio of the induced magnetic moment to the applied field. Changes to the strength and frequency of the applied field and the temperature at which it is measured induce non-linear magnetic responses in a sample. These differences can provide information about the composition of the materials carrying the magnetization, which is important for their interpretation.

Most common measurements of MS are made in low fields (~ 0.1 mT), which do not permanently magnetize the sample, allowing them to return to their original magnetic state on removal of the field. When ferrimagnetic minerals dominate the mineral assemblage, low-field MS is often assumed to be proportional to their concentration. In greater field strengths, a variety of different minerals and magnetic grain sizes can contribute to the MS signal. However, above a certain critical field, the induced magnetization becomes nonreversible and a magnetization, or remanence, is retained by the sample. Measurement of the magnetization in different field sizes permits construction of remanence parameters and ratios (when measured in zero field) (i.e., isothermal remanent magnetization (IRM; Table 1)) and construction of hysteresis loops (when measured in field) that can be diagnostic of magnetic mineralogy and/or magnetic grain size. MS is dependent on the frequency of the applied field, which has some magnetic grain size dependence. Many systems allow measurements to be made at variable frequencies providing information on the proportion of ultrafine SP grains. MS also displays strong temperature dependence, which is a function of intrinsic magnetic mineralogical properties. Regular measurements made through warming or cooling cycles can produce characteristic curves indicative of different magnetic mineral or grain size populations. In summary, MS is an aggregated function of magnetic concentration, mineralogy, and grain size and

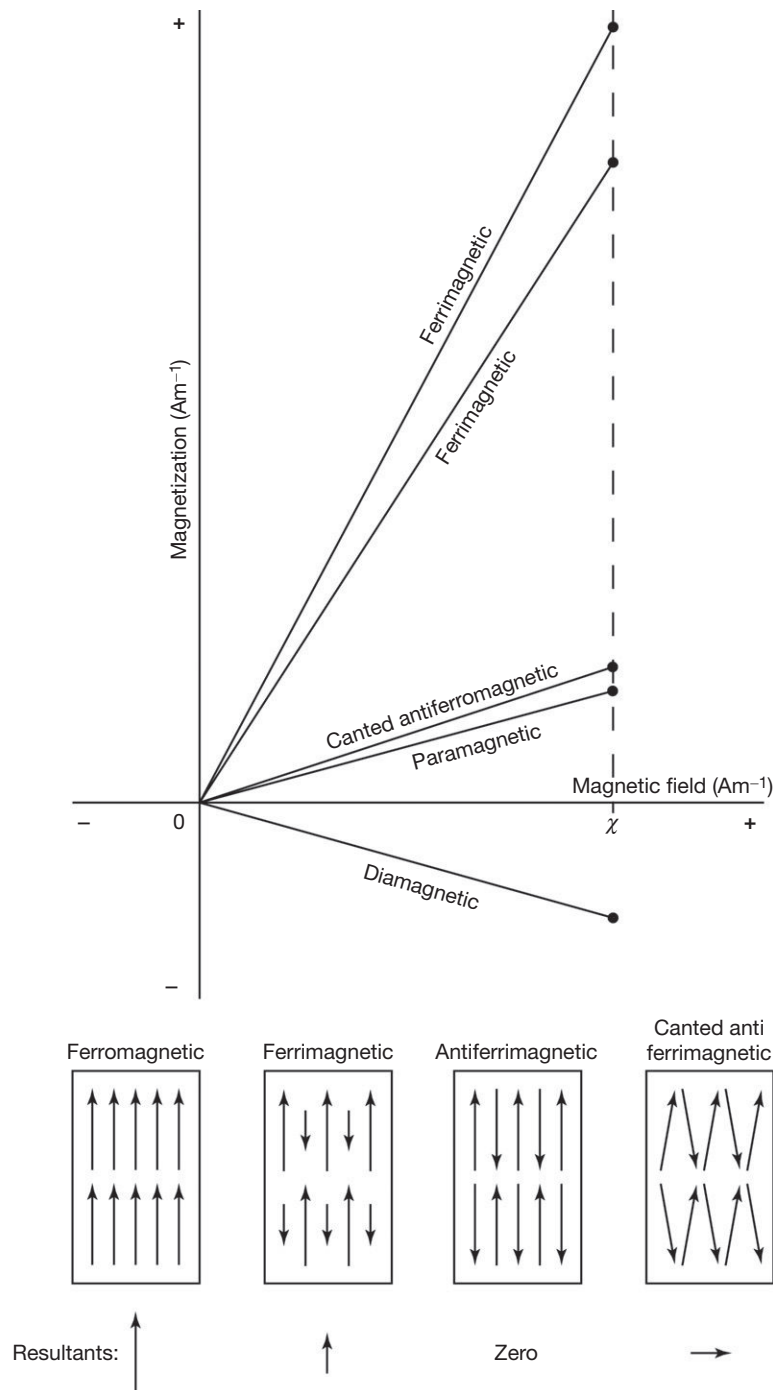


Figure 1 Schematic diagram illustrating the relative strengths and organizations for a range of magnetic behaviors. Adapted from Dearing J (1999) Magnetic susceptibility. In: Walden J, Oldfield F, and Smith J (eds.) *Environmental Magnetism: A Practical Guide*, Technical Guide No. 6, pp. 35–62. London: Quaternary Research Association.

shape that is dependent upon the applied field, frequency, and temperature of measurement.

MS Measurement Systems

MS (κ or χ) can be measured on a variety of field and laboratory equipment providing MS measurements of discrete

samples, whole cores, split cores, u-channels, or *in situ* field samples, and can be made in varying field strengths, frequencies, and at different temperatures. For all systems, the magnetic susceptibility produces a magnetic field that induces and measures an infield magnetic moment in response to these three factors.

Due to cost, ease of use, and versatility, the Bartington MS2 is often the most widely available and used MS measurement

Table 2 Magnetic behavior, crystal arrangement and example minerals and mass normalized susceptibilities. Adapted from Dearing J (1999) Magnetic susceptibility. In: Walden J, Oldfield F, and Smith J (eds.) *Environmental Magnetism: A Practical Guide*, Technical Guide No. 6, pp. 35–62. London: Quaternary Research Association.

Magnetic behavior	Crystal arrangement	Minerals	MS
'True' ferromagnetism	Parallel coupling of all unpaired electrons to the applied field	Fe – iron Co – cobalt Ni – nickel	276 000 204 000 68 850
Antiferromagnetism	In oxides of Fe, Co, and Ni, electron coupling occurs via an intermediate oxygen atom with alternate crystal layers aligning opposite to each other resulting in no net magnetic moment		
Ferrimagnetism	If the magnetic moments of the alternating iron compounds are unequal, then a magnetic moment is created	Fe ₃ O ₄ – magnetite γFe ₂ O ₃ – maghemite Fe ₃ S ₄ – greigite	660 440 170
Canted antiferromagnetism	For various reasons, sublattices in otherwise antiferromagnetic arrangements may not be perfectly aligned and a small residual magnetization exists	αFe ₂ O ₃ – hematite αFeOOH – goethite	~0.7 ~0.4
Paramagnetism	Some unpaired electrons align to an applied field; this magnetization is lost on removal of the field	FeTiO ₃ – ilmenite Mg,Fe,Al silicate – biotite γFeOOH – lepidocrocite FeCO ₃ – siderite FeS ₂ – pyrite	1.7 ~0.5 0.7 1.0 0.3
Diamagnetism	Compounds with no unpaired electrons act to repel the applied field; this magnetization is lost on removal of the field	SiO ₂ – quartz CaCO ₃ – calcite H ₂ O – water	–0.0058 –0.0048 –0.009

Units of mass-normalized magnetic susceptibility (MS) – $10^{-6} \text{ m}^3 \text{ kg}^{-1}$.

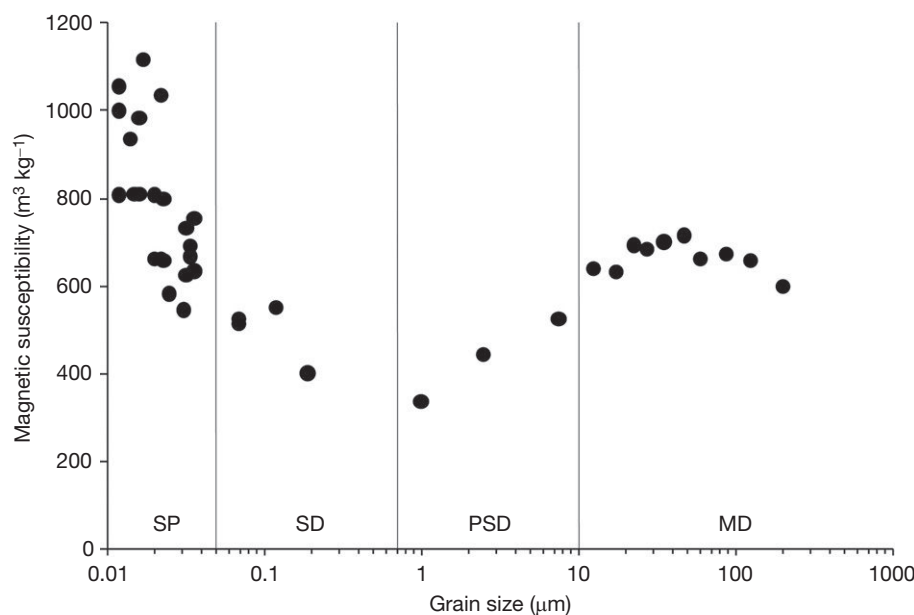


Figure 2 Domain-size boundaries and variation in MS as a function of magnetic grain size for magnetite. Domain abbreviations: SP = superparamagnetic, SD = single domain, PSD = pseudo-single domain, MD = multi domain. Data from Maher (1988), Ozdemir and Banerjee (1982) and Dankers (1978).

system. Capable of parts per million (ppm) detection of magnetite, it can be coupled with a variety of sensors to allow initial characterization of the magnetic properties of a sample. These measurements are most commonly made at low field (80 Am^{-1}), low frequency (0.47 kHz), and room temperature.

For example, MS2C loop sensors coupled with measurements of γ -ray porosity, p-wave velocity, and electrical resistivity on multisensor tracks are a standard shipboard measurement on

both coring and drilling operations alike. Standard 1.5-m-long core sections can be characterized in as rapidly as 10 min in some configurations, providing real-time data to aid core correlation, site selection, and for initial sediment interpretation. Such measurements should account for the background MS before and after measurement to help correction for local environment and drift complication. This can be easily corrected for when measuring discrete samples as the background can be recorded after

every measurement. Whole-core- or other long-core-logging measurements at relatively high (e.g., centimeter scale) resolution require many MS measurements between background air measurements. Sensor drift is usually assumed to be linear; however, this may not always be a valid assumption. More accurate measurements of drift can be obtained by measuring background MS mid-logging, though this increases measurement time and introduces the potential for positional errors. Core-logging sensors also integrate data over a stratigraphic interval (e.g., the MS2C loop smoothes over twice its sensor diameter; Dearing, 1999); thus, data are smoothed and edge effects at core ends are introduced. It is theoretically possible to reintroduce resolution and lost edge effect data through deconvolution, though in practice this is almost never employed. However, because these measurements are nondestructive, sections can be relogged, subsampled, and remeasured discretely to provide greater precision, accuracy, and resolution where needed.

Different susceptometers have demonstrated precision; however, little work has been done to calibrate systems to examine data accuracy. Sagnotti et al. (2003) noted that different MS2 systems can underestimate reference samples by up to 15% and that differences between MS2 and Kappabridge systems can range from 2% to 13%. MS2 systems can be susceptible to sample volume and positioning differences between measurements with 1.5% differences between 1 and 7 cc samples of the same material. These factors coupled with simple variances in other factors, for example, changes in room temperature, can introduce measurement error. With these types of measurements becoming increasingly popular, more attention is being focused on these problems and standards and protocols are currently being developed to help address some of these issues, which will permit greater accuracy in comparisons of records across the globe.

Environmental Records of Susceptibility

Compounds of iron comprise roughly 5% of the Earth's crust and are ubiquitous in many environments in a variety of inorganic and bioavailable forms (Table 2); for example, iron oxides can constitute 2–6% of basalts (Thompson and Oldfield, 1986). Iron-bearing minerals are often eroded, transported, and deposited coevally with other minerals, nutrients, or pollutants. Thus, magnetic properties of environmental materials often correlate well with other independently determined properties. In many sedimentary environments, magnetite is the most common ferrimagnetic mineral. Given its strong response to a magnetic field (Table 2), low-field MS is often considered proportional to magnetite concentration (Dearing, 1999). However, although mineral concentration is often considered the principal factor driving MS, this is often not the case and these assumptions are not always valid. Natural samples are rarely unimodally composed and instead are aggregates of varying amounts of different minerals and grain sizes, complicating such a simple interpretation. As magnetic properties are not uniform across magnetic mineralogies or grain sizes, the interpretation of what drives MS variation is often not simply a reflection of ferrimagnetic concentration. Similar values can potentially be driven by different

compositions, for example, the presence of a small amount of magnetite or larger concentrations of hematite. MS can be diluted by diamagnetic minerals (e.g., biogenic carbonates) that are often significant components of ocean sediments. The nature of the magnetic carriers also can affect the MS. For example, authigenic bacterial magnetosomes are composed of SP to SD magnetite, while terrigenous inputs to marine systems are often PSD to MD grain sizes. The interplay of all of these factors can sometimes produce different and unique MS values characteristic of a specific environment or process. However, it is rarely possible to determine the factors driving the MS signal from MS measurements alone; a greater suite of magnetic measurements are normally necessary (e.g., Table 1). Such an evaluation is almost always required, not only to identify the absolute drivers of change, but also to recognize the causes of differences between measurements and to help with the understanding and accounting for the potential processes involved in driving these changes.

However, given the abundance of Fe oxides in the environment, their ability to harmonize with natural processes, and the ability of magnetic measurements to detect subtle changes in concentration, mineralogy, and grain size, links are often drawn between measurements of MS and environmental change in many diverse settings. Lacustrine MS records often show strong linkages to other independent climate proxies, and compared to the deep-sea environment, lake records possess relatively high sedimentation rates, enabling high-resolution reconstructions of climate. However, many of the most climatically sensitive lakes are found in glacially proximal environments and are thus temporally restricted due to their postglacial formation. Similarly, loess records have shown to possess remarkable MS archives of climate changes; however, their formation is quasi-continuous and restricted to certain areas of the globe. In contrast, relatively slow but continuous sedimentation rates in marine records permit the potential for long climate change reconstructions from large areas of the globe.

MS as a Paleocceanographic Proxy

Like all other environmental records of susceptibility, the MS records of marine sediments are an aggregated sum of the interplay of magnetic concentration, magnetic mineralogy, and magnetic grain size. Variations in these properties can be forced by, and often be characteristic of, several different depositional processes and postdepositional (diagenetic) transformations. Depositional processes generally reflect the balance between terrigenous flux, modulated by some combination of surface and bottom currents, eolian transport, ice-rafted and other glacial marine processes, volcanogenic inputs, tectonic and downslope transport, and biogenic preservation. Postdepositional alterations may result in the creation, destruction, and/or transformation of magnetic minerals. All these factors can influence an MS signal and their relative influence can vary both spatially and temporally. Unraveling these often codependent influences is a nontrivial matter, but with careful selection of both site location and time period, coupled with analysis and understanding of the factors driving the MS signal, MS records of ocean sediments can potentially reveal much

about the paleohistory of both oceans and climate. Here we will report some of the important ways MS has been used in the study of paleoceanographic and paleoclimatic records of ocean cores and some of the important factors to consider in their interpretation.

MS as a Proxy for Terrigenous Flux

The delivery and distribution of terrigenous sediments to the world's oceans is subject to a number of complex processes. One of the major factors governing terrigenous sediment concentration in the ocean is the availability and transport of detrital grains. Transport and delivery-related processes can be, and have been, climatically mediated over a range of timescales in response to changes in climatic regimes. In environments with minimal diagenetic transformations, the MS records of ocean cores can simply reflect (and are often assumed to reflect) the composition of the terrigenous material delivered to a site. Given that the production of biogenic materials (e.g., carbonate and silica) can also be climatically dependent, the MS records of many marine sediments often reflect the balance between terrigenous flux and biogenic dilution. Common examples of this type of behavior are found in Pacific, Indian, and Atlantic Oceans.

The MS of ODP sites 721 and 722 in the Arabian Sea strongly correlate with the percentage of terrigenous materials ($r=0.98$; Bloemendal and deMenocal, 1989). Delivery of this terrigenous fraction is linked to eolian transport and monsoon intensity. Spectral analysis of the Arabian Sea MS records and ODP site 661 from the eastern equatorial Atlantic showed strong influence of the 23 and 19 ka precession-related orbital periodicities prior to 2.4 My with a shift to 41 ka periodicities after 2.4 My, possibly reflecting ice sheet effects on monsoon climate or dust source regions (Bloemendal and deMenocal, 1989; Figure 3). Similarly for ODP site 882 in the North Pacific, Maslin et al. (1996) used MS as a proxy for ice-rafted debris (IRD), and showed synchronous changes between IRD, the $\delta^{18}\text{O}$ record, and the biogenic opal record (Figure 4). Despite the different processes responsible for delivering sediment to these two sites, both MS records identify a major climate reorganization 2.4–2.7 Ma, resulting in global changes in terrestrial flux from the continents. It is thought that these shifts are associated with the initiation of northern hemisphere glaciations and the start of the Quaternary. These long records linked to climatically mediated terrestrial fluxes helped establish the links between orbital and climatic forcings of MS records.

Greater sedimentation rates in the North Atlantic, coupled with greater carbonate preservation, have permitted higher resolution studies of individual events, with their chronologies constrained by more accurate dating. Surrounded by a terrestrial regime dominated by the growth and decay of continental-scale ice sheets, bedrock geology composed of both highly magnetic basalts and crystalline basement rocks and weakly magnetic carbonates, and a sediment transport regime dominated by dynamic surface and subsurface circulations, the North Atlantic has proven to be an exceptional archive for MS studies of paleoclimate and paleoceanography.

In the mid-latitude IRD belt (40° and 55° N) of the glacial North Atlantic (Ruddiman, 1977), MS has been widely used to

correlate IRD events between cores, illustrating the basin-wide coherence of the IRD pattern. Through glacial and interglacial cycles, Robinson (1986) showed that MS peaks were driven by fluxes of IRD in the mid-latitude North Atlantic, linked to increased glacial calving. This signal is also mediated by productivity (Figure 5). During glacial periods, increased delivery of IRD is coupled with a reduction in the production of biogenic carbonates resulting in high MS values. During interglacials, decreased delivery of IRD is coupled with increased biogenic production resulting in increased diamagnetic carbonates resulting in low MS values. This flux/dilution relationship results in strongly contrasting MS values on glacial/interglacial timescales that are replicable across large spatial scales. The highest glacial susceptibility peaks are associated with sand layers (>150 μm) deposited during Heinrich events (Bond et al., 1992), providing a diagnostic signature of these canonical Laurentide Ice Sheet discharge events that prominently characterize the glacial North Atlantic (e.g., Bond et al., 1992). MS records south of the glacial polar front in the North Atlantic differ greatly from those within the IRD belt, reflecting the strong melting gradient and IRD-forced nature of MS. Mapping of these spatial MS signatures also permitted reconstruction of iceberg trajectory paths, paleosurface currents, and the migration of the glacial polar front.

From the Southern Ocean, Pugh et al. (2009) show MS records from three cores from the Scotia Sea to be almost identical to the European Project for Ice Coring in Antarctica (EPICA) dust record (Figure 6). Again, glacial periods are characterized by high susceptibility and interglacials with susceptibility minima. However, these records are apparently not driven by IRD, but by dust export from continental landmasses. The processes involved in driving ice rafting in the North Atlantic and dust flux in the Southern Ocean are very different; however, MS records show similar trends. This similarity is forced by shifts in dominant global climatic patterns over glacial/interglacial timescales, which in turn mediates the terrestrial flux and the attendant MS record.

While global climate changes can have a large impact on global sediment flux and the MS record, it is essential to interpret any MS record within the context of its location and processes acting in its formation. For example, Stoner et al. (1995) and Carlson et al. (2008) examined the magnetic and geochemical record of cores from the Eirik Drift in the NE Labrador Sea, southwest of Greenland. Unlike the majority of records from further south in the North Atlantic Basin, glacial terminations and interglacial periods are characterized by higher MS values than glacial periods. During the glacial–interglacial transitions of marine isotope stage (MIS) 2/1 and MIS 6/5, enhanced MS is accompanied by a coarsening of magnetic grain sizes and an increase in Ti/Fe concentrations. Here, minima in IRD (represented by the >125 μm fraction) and increases in the silt fraction are synchronous with increases in MS and magnetic grain sizes and are thought to be associated with sediment-laden melt water pulses associated with increased melting of the Greenland Ice Sheet during deglaciation (Carlson et al., 2008; Stoner et al., 1995). Thus, while these Greenland proximal records, like the North Atlantic IRD and Southern and Indian Ocean eolian records are driven by flux of terrigenous fragments responding to changing climate drivers, their MS signals produce very different temporal signatures.

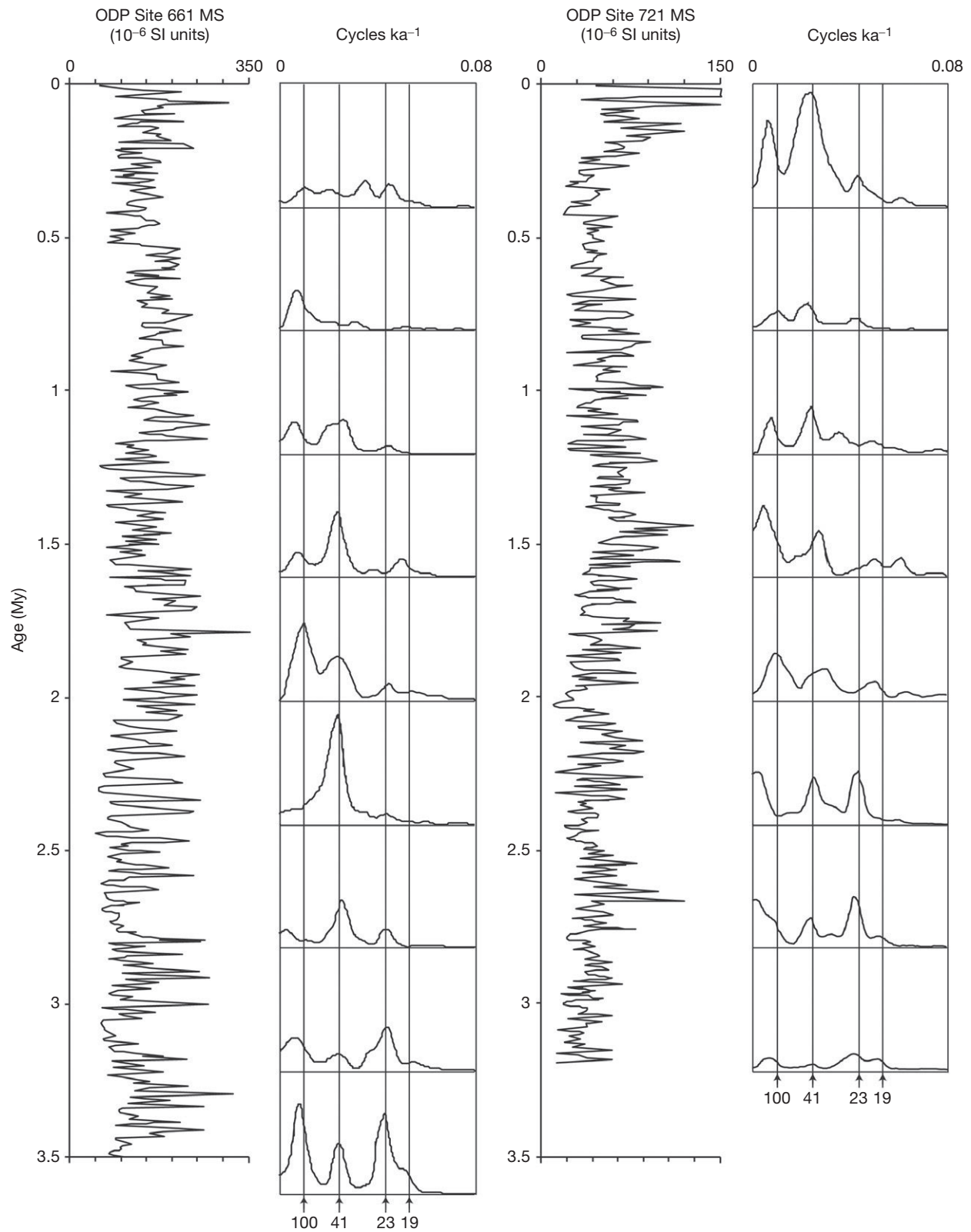


Figure 3 Magnetic susceptibility time series for ODP Site 661 and ODP site 721, and the variance spectra (inset figures) for time intervals between 3.5 Ma and the present. Adapted by permission from Macmillan Publishers Ltd: [Nature] (Bloemendal J and deMenocal P (1989) Evidence for a change in the periodicity of tropical climate cycles at 2.4 Myr from whole-core magnetic susceptibility measurements. *Nature* 342: 897–900), copyright 1989.

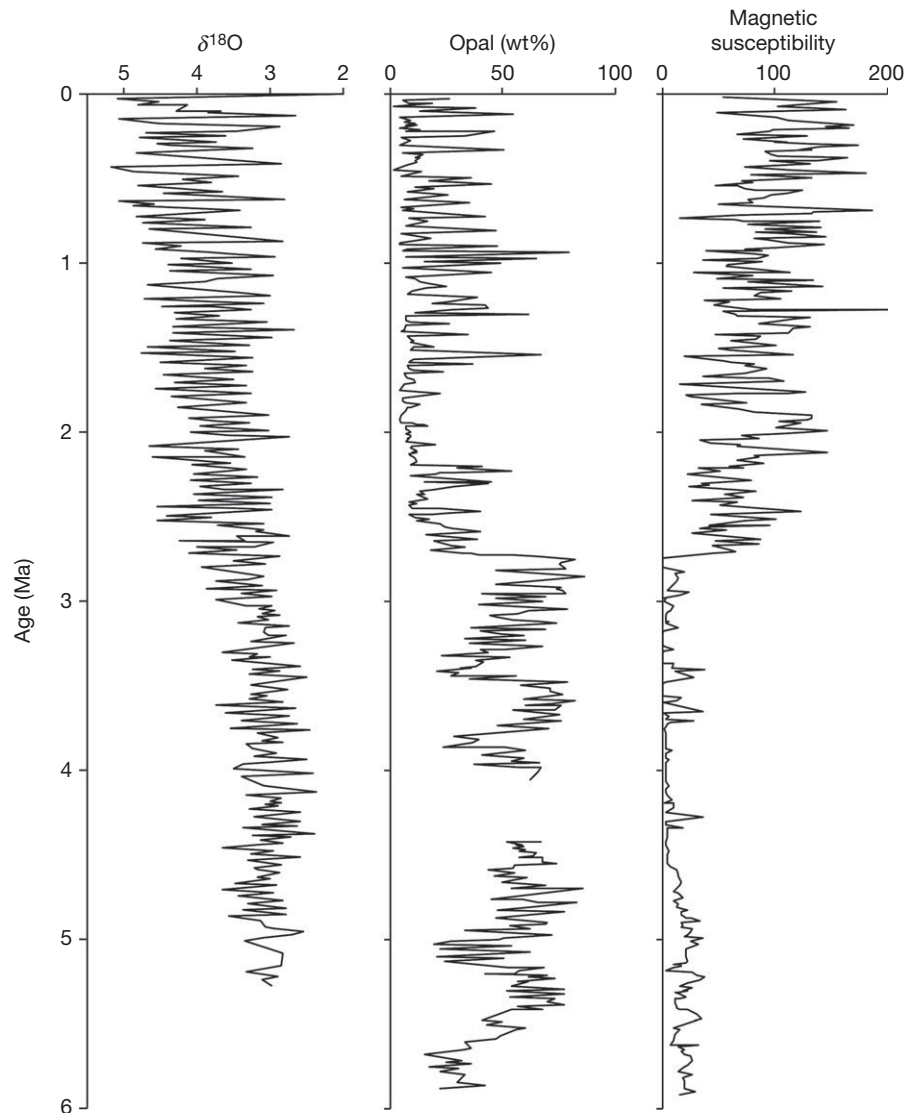


Figure 4 Oxygen isotope, opal, and MS records from ODP site 882. Adapted with kind permission from Springer Science+Business Media: Geologische Rundschau, *The progressive intensification of northern hemisphere glaciation as seen from the North Pacific*, 85, 1996, 452–465, Maslin MA, Haug GH, Sarnthein M and Tiedemann R Copyright (1996).

MS records driven by the flux of terrestrial fragments can yield high-resolution information about changes in climate and its effect on the terrestrial landscape. The glacial/interglacial, flux/dilution relationship in the ocean remains a major driver behind many MS records and permits comparisons over local, regional, and to some degree even global scales. However, these records must first be interpreted within the context of the processes driving these records and other factors which may potentially modify this relationship must also be assessed.

MS as a Proxy for Sediment Redistribution

In the deep ocean, MS records often correlate well with IRD percentage and reflect surface currents and iceberg melt routes. However, [Rasmussen et al. \(1997\)](#) showed a remarkable similarity between the MS record of ENAM93-21 from the Faeroe-

Shetland Channel in the North Atlantic and the $\delta^{18}\text{O}$ of the Greenland Ice Core Project (GRIP) record ([Figure 7](#)). This MS record captured all 15 Dansgaard-Oeschger (D-O) cycles within the MIS 3 and MIS 2 intervals. In contrast to the North Atlantic IRD pattern, abrupt warming is characterized by sharp increases in MS, with gradual cooling associated with decreasing MS values, showing little relation to IRD inputs ([Figure 7\(c\)](#)) either in the model proposed by [Robinson \(1986\)](#) or regionally. MS lows are associated with colder stadial periods and Heinrich event intervals. Subsequent studies have found similar patterns along the path of dense overflow waters in the subarctic basins of the North Atlantic that contribute to North Atlantic Deep Water (NADW). Because these overflow waters have been implicated as drivers of the D-O cycles observed in the Greenland ice cores, these MS observations have implications for examination of the processes involved. MS patterns were shown to be strongly controlled by variations

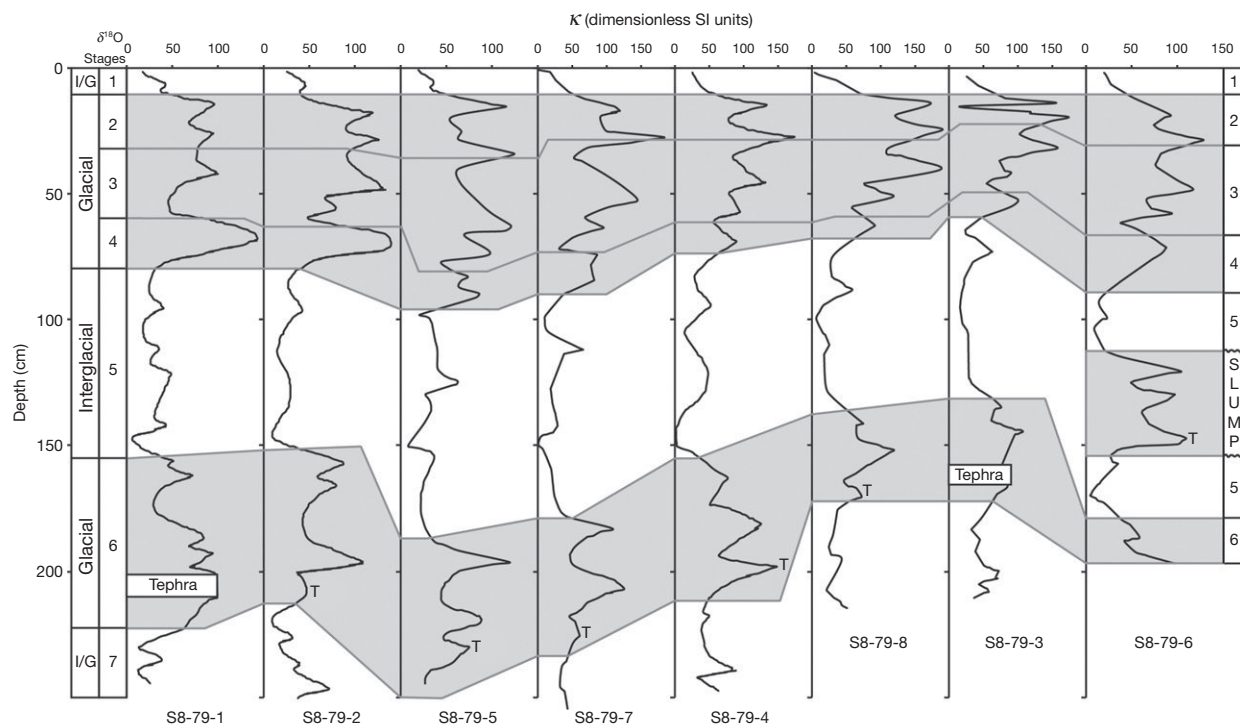


Figure 5 Correlation of cores using MS through seven glacial/interglacial stages. Reprinted from *Physics of the Earth and Planetary Interiors*, 42, Robinson SG, The late Pleistocene paleoclimatic record of North Atlantic deep-sea sediments revealed by mineral-magnetic measurements, 22–47, Copyright (1986), with permission from Elsevier.

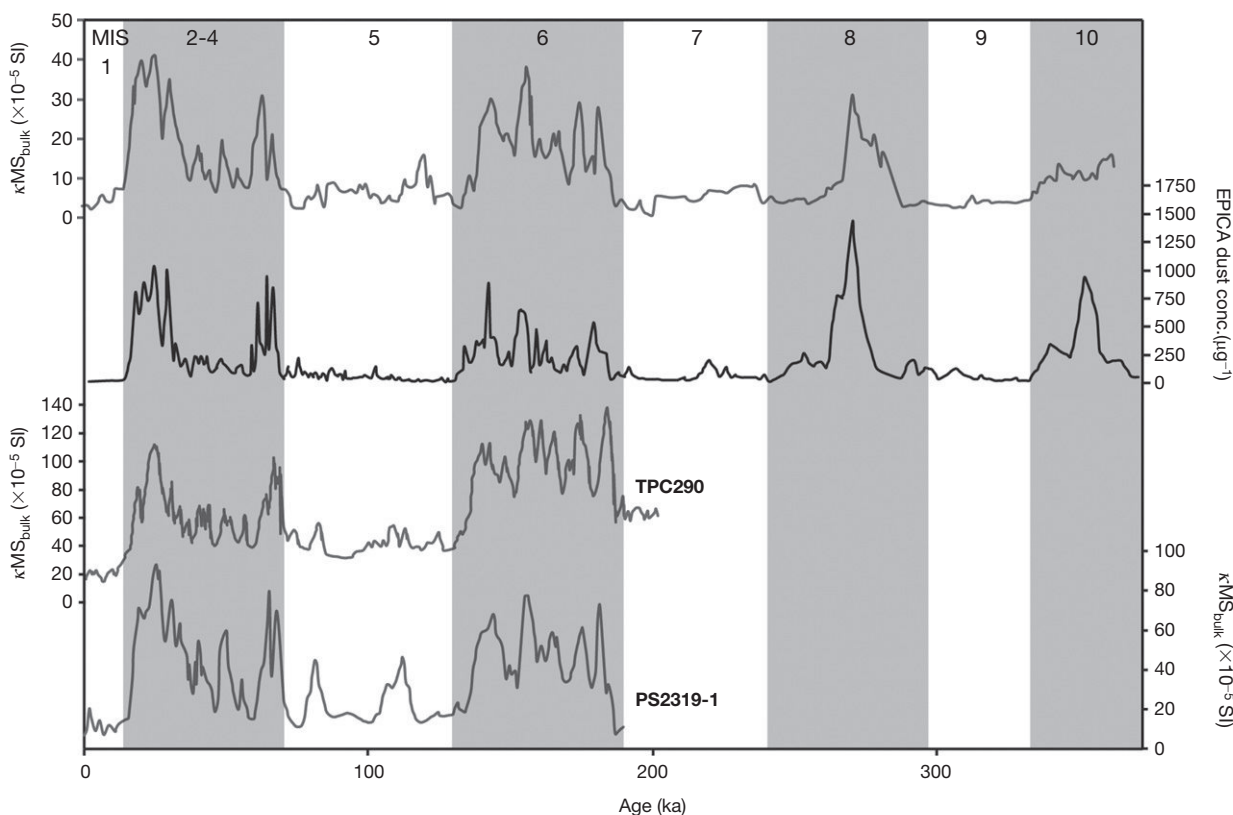


Figure 6 Correlation between EPICA dust concentration and three Scotia Sea cores. Note the high susceptibility in glacial phases and low MS during interglacials, similar to the IRD record. Reprinted from *Earth and Planetary Science Letters*, 284, Pugh R, McCave I, Hillenbrand C, and Kuhn G, Circum-Antarctic age modelling of Quaternary marine cores under the Antarctic Circumpolar Current: Ice-core dust-magnetic correlation, 113–123, Copyright 2009, with permission from Elsevier.

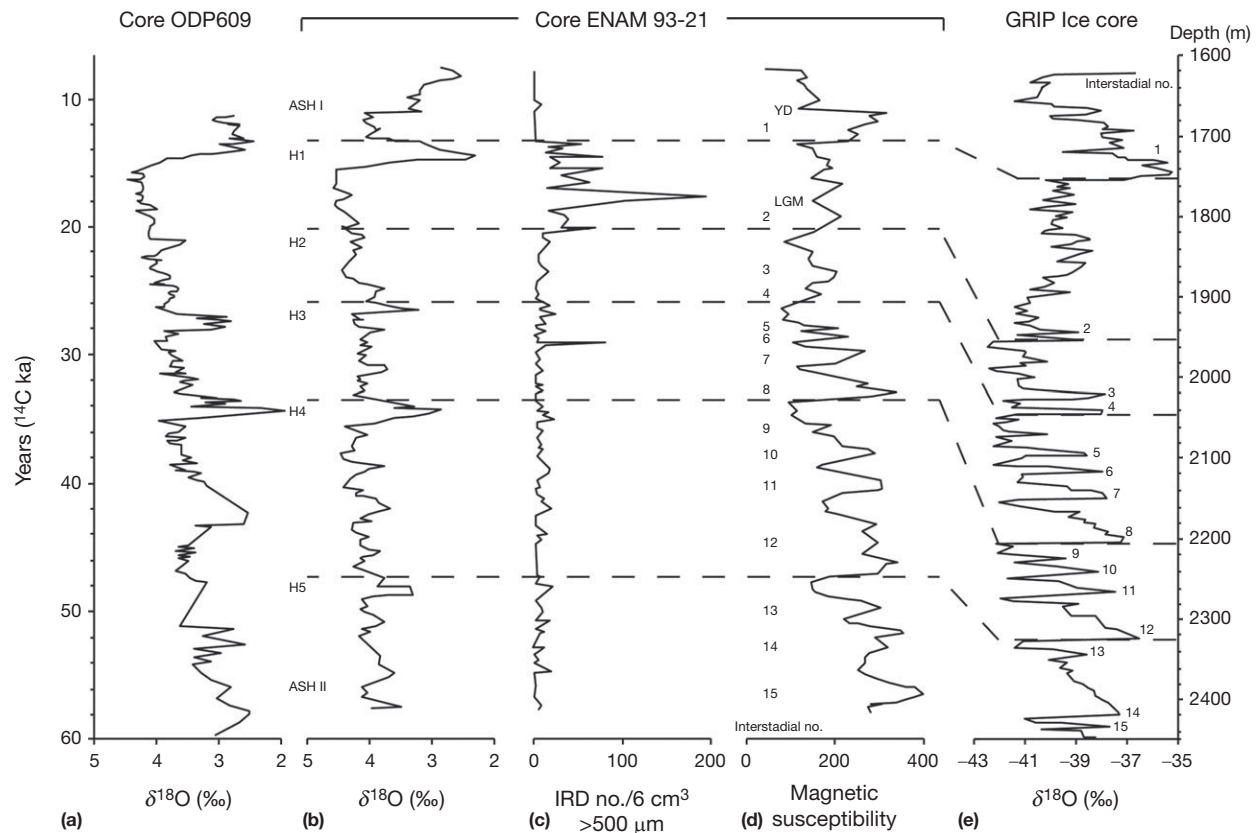


Figure 7 Comparison of (a) $\delta^{18}\text{O}$ from ODP site 609, (b) $\delta^{18}\text{O}$, (c) IRD grains $>500\ \mu\text{m}$, (d) MS from ENAM 93-21, and (e) $\delta^{18}\text{O}$ from the GRIP ice core record. Reprinted from *Quaternary Science Reviews*, 16, Rasmussen TL, van Weering TCE, and Labeyrie L, Climatic instability, ice sheets and ocean dynamics at high northern latitudes during the last glacial period (58–10 ka BP), 71–80, Copyright (1997), with permission from Elsevier.

in titanomagnetite concentration (Kissel et al., 1997; Snowball and Moros, 2003) and similar patterns were shown to be spatially replicable over a transect of several cores (e.g., Kissel, et al., 2009). It is suggested that during colder phases when NADW formation was sluggish, transport and delivery of ferromagnetic magnetite to drift sites was restricted resulting in lower MS values. Conversely, during warmer phases, NADW strength increased, as did basaltic fragment transport, increasing the MS values at these locations.

Peaks and minima of MS in oceanic sediments can often be interpreted in a paleoclimatic context. However, resuspension during downslope turbidite flows and subsequent density-related settling of relatively heavy magnetic minerals within an otherwise magnetically weak matrix can also produce characteristic signatures. These MS peaks can be important for identifying turbidite layers and for distinguishing them from paleosignals, as these events can have severe implications for the paleointerpretation of sedimentary records (e.g. core S8-79-6 in Figure 5). MS peaks generated by these events can often be process and event diagnostic. Goldfinger et al. (2007, 2008) used MS to study paleoseismology, correlating MS peaks between earthquake-event-triggered turbidite flows over several hundred kilometers along the Cascadia margin. Earthquake-generated turbidite deposits have not only been recorded by magnetic measurements on the Pacific margins, but also in Quebec and off the Portuguese Coast, illustrating the potential of MS for identifying these deposits.

MS and Source Variation

IRD, bottom current, and slope process records of MS often assume that magnetic concentration linked to flux/dilution relationships are the principal drivers behind the MS record. However, Pirrung et al. (2002a,b) showed that MS varied over three orders of magnitude in the surface sediments of both the Nordic Seas (Pirrung et al., 2002a) and the Atlantic sector of the Southern Ocean (Pirrung et al., 2002b), linked to the source of minerals from different provinces (Figure 8). Highest values were found around Scoresby Sund and along the Iceland–Faeroe–Shetland Ridge, linked to basaltic sources in the Nordic Seas; similarly, in the Southern Ocean, highest values are proximal to mafic sources. It is often assumed that all ferrimagnetic minerals are terrigenously derived, yet all terrigenous materials are not ferrimagnetic. In the absence of a strong ferrimagnetic source component, MS values may be influenced by weaker magnetic components; such areas are evident, for example, in the Norwegian and western Barents Sea (Figure 8). Variations are evident in other regions of the North Atlantic, with Andrew and Hardardottir (2009) magnetically discriminating between Greenlandic and Icelandic sources on either side of the Denmark Strait. Stoner and Andrews (1999) also showed that sediments sourced from the Hudson Bay ice streams were carbonate-rich and this was reflected in the sediments of the Labrador Sea that possess relatively low MS compared to the rest of the North Atlantic.

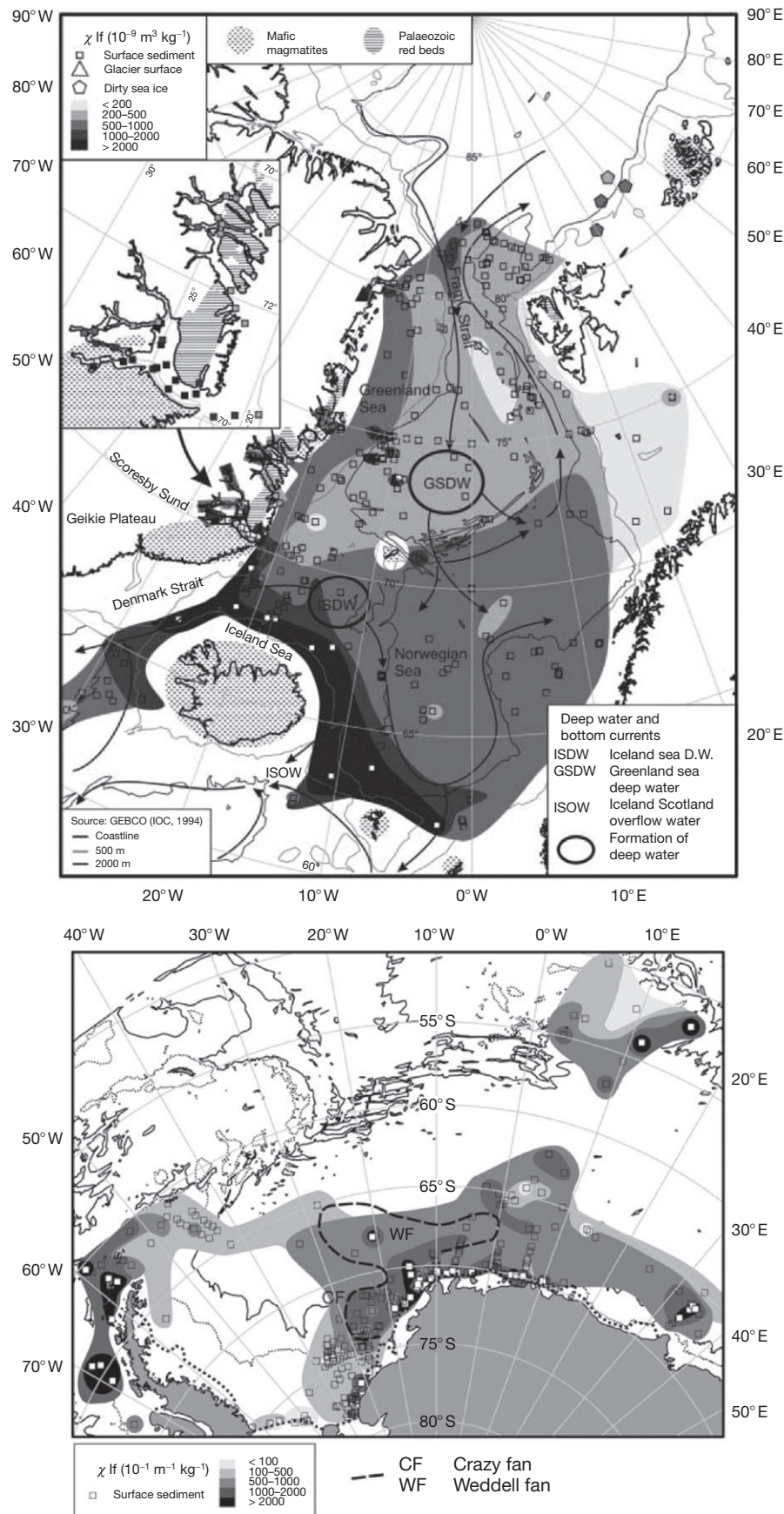


Figure 8 MS variation in surface sediments from the Nordic Seas and the Atlantic sector of the Southern Ocean. Reproduced with kind permission from Springer Science+Business Media: *Geo-Marine Letters*, Magnetic susceptibility and ice-rafted debris in surface sediments of the Nordic Seas: Implications for Isotope Stage 3 oscillations, 22, 2002a, 1–11, Pirrung M, Fütterer D, Grobe H, Matthiessen J, and Niessen F, Figure 1; *Geo-Marine Letters*, Magnetic susceptibility and ice-rafted debris in surface sediments of the Atlantic sector of the Southern Ocean, 22, 2002b, 170–180, Pirrung M, Hillenbrand C, Diekmann B, Fütterer D, Grobe H, and Kuhn G, Figure 2.

Heinrich layer MS values are also low in this region, illustrating MS dilution due to diamagnetic carbonate background terrigenous concentrations. Instantaneous tephra deposits can also possess distinct magnetic signatures and can be used as stratigraphic markers (e.g., Peters et al., 2010).

Watkins and Maher (2003) and Watkins et al. (2007) extended the work of Pirrung et al. (2002a) to include both potential sources and core sediments surrounding the whole North Atlantic Basin (Figure 9), both at the present day and during the Last Glacial Maximum (LGM). Their results neatly illustrated that even as sediment from the IRD belt (poleward of 40° N) during the LGM possessed higher susceptibility than present (Watkins et al., 2007), there exists significant regional variation in MS, both for the present day and at the LGM. For example, IRD originating from Icelandic basalts and the North American Shield possess characteristically high susceptibilities, and are reflected in the sedimentary properties of the Western North Atlantic. In contrast, the lower susceptibility areas identified by Pirrung et al. (2002a) are proximal to the gray and red sandstones from Spitsbergen which outcrop and surround the Norwegian and Barents Seas, and eolian dust sources are responsible for the relatively low susceptibility of the central Atlantic. With this variation defined, MS can potentially be used to map the timing and intensity of IRD events,

and under optimal conditions can provide information on their sources through the last glacial–interglacial transition (cf. Watkins and Maher, 2003; Watkins et al., 2007). However, isolating the relative influences of sediment flux and sediment source is an almost impossible process using MS alone. Instead, a wider range of magnetic measurements are often needed (see Table 1), with combinations of these parameters permitting assessment independent of magnetic concentration. Source variation and the roles of IRD and bottom water currents also rarely act in isolation of each other or indeed several other processes which may deliver or restrict transport of terrigenous fractions. Site selection and magnetic measurement which consider these processes, and an interpretation that accounts for them, are often key to successfully relate the MS signal to a certain aspect of the climate system.

Postdepositional Records of MS

Magnetic minerals accumulating in sedimentary environments are often assumed by the paleocommunity to be conservative, maintaining properties that reflect their original source materials and mechanisms of deposition. In areas with high organic matter accumulation and/or in low oxygen environments, organic material may be buried before fully oxidizing, promoting

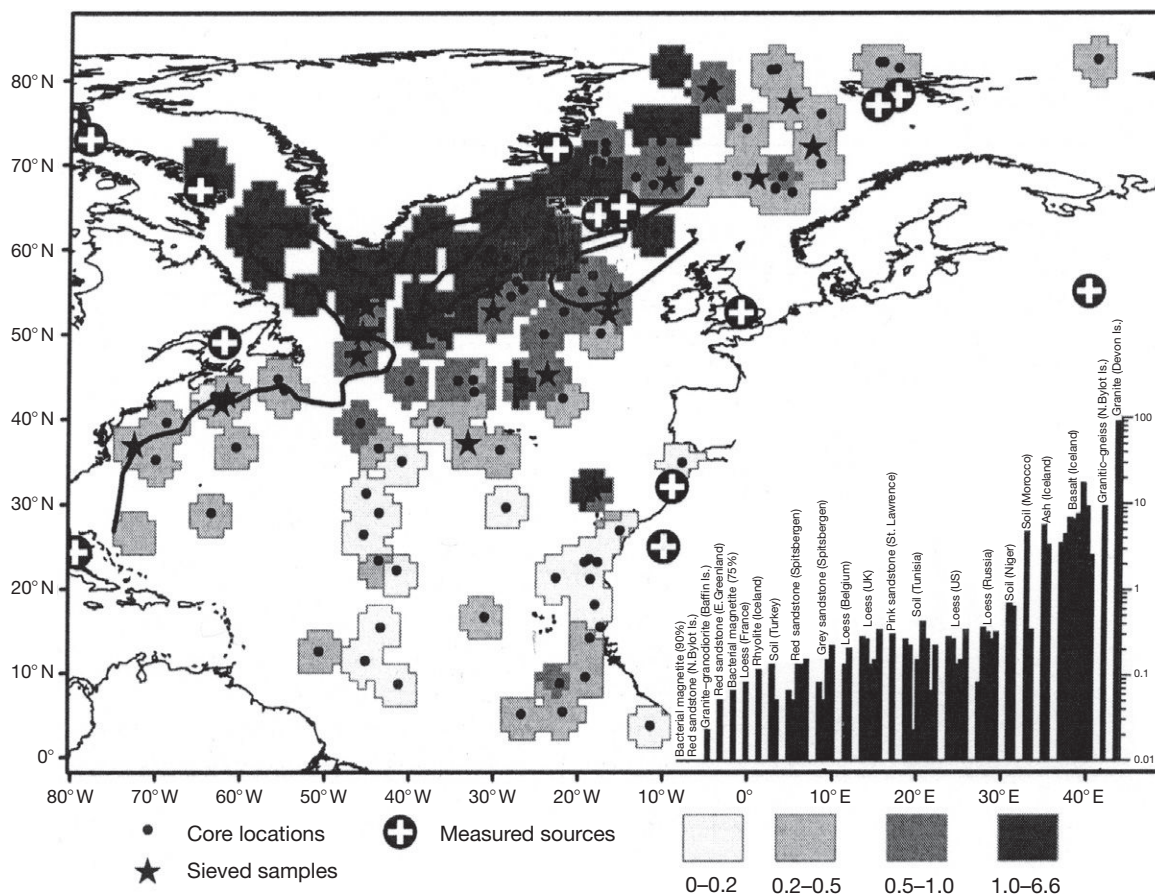


Figure 9 MS values ($10^{-6} \text{ m}^3 \text{ kg}^{-1}$) of potential sources (bar chart inset) and LGM values of MS for cores covering the North Atlantic (shaded ranges). Reprinted from *Paleoceanography*, 22, Watkins SJ, Maher BA and Bigg GR, Ocean circulation at the Last Glacial Maximum: A combined modeling and magnetic proxy-based study, 1–20, Copyright (2007), with permission from Elsevier.

bacterially mediated reduction diagenesis (Froelich et al., 1979). Fine ferrimagnetic grains (either of biogenic or terrestrial origin) are most susceptible to reduction in a profile undergoing diagenetic changes. This can result in partial loss of the deposited magnetic signal and is therefore a significant issue for interpretation of MS records. Deeper in these records, soluble iron can be reprecipitated as other magnetic phases, including pyrite and greigite, and these minerals have shown to be significant components of oceanic records. These minerals can nucleate around larger surviving ferrimagnetic grains protecting their core (and magnetic signal) from further reduction.

Oceanic areas producing strong diagenetic change are generally restricted to continental margins and/or where sedimentation and organic accumulation rates are high and in areas of upwelling and in partially landlocked basins (e.g., the Mediterranean and the Sea of Japan). Such areas are generally well known and while assessment of diagenetic change cannot be conclusive using MS alone, simple additional diagnostic magnetic measurements can help document evidence for diagenetic change. Distinctive signatures include steep decreases in concentration indicators (e.g., MS) with a parallel coarsening of parameters sensitive to magnetic grain size pointing to the selective loss of finer ferrimagnetic grains. As hematite and

goethite are less affected by suboxic diagenesis, an increase in coercivity may also accompany selective loss of finer grains.

Diagenetic processes usually result in degradation of the depositional signal and the paleoclimatic information it contains. However, reduction-driven diagenetic transformations may not always result from steady-state conditions, but from changing boundary conditions providing additional opportunities for the generation of paleoinformation. For example, in an analysis of four sediment records from the Sea of Japan, light/dark rhythms in the sediment record suggested cyclic switching between times of oxic and anoxic conditions, related to glacio-eustatic sea level changes (Vigliotti, 1997). Here MS correlates with the $\delta^{18}\text{O}$ record, showing sea level-lowered glacial periods are more favorable for production of anoxic basin conditions resulting in an increase in sulfuric compounds and a reduction in magnetic concentration and coarsening of the magnetic grain size. Interglacial periods contain deeper waters, greater basin oxygenation, and greater retention of the fine ferrimagnetic fraction (Figure 10).

Postdepositional transformations act not only to remove ferrimagnetic minerals from a profile, but *in situ* production of fine-grained authigenically derived bacterial magnetite can also potentially significantly alter the depositional record. Bacterial

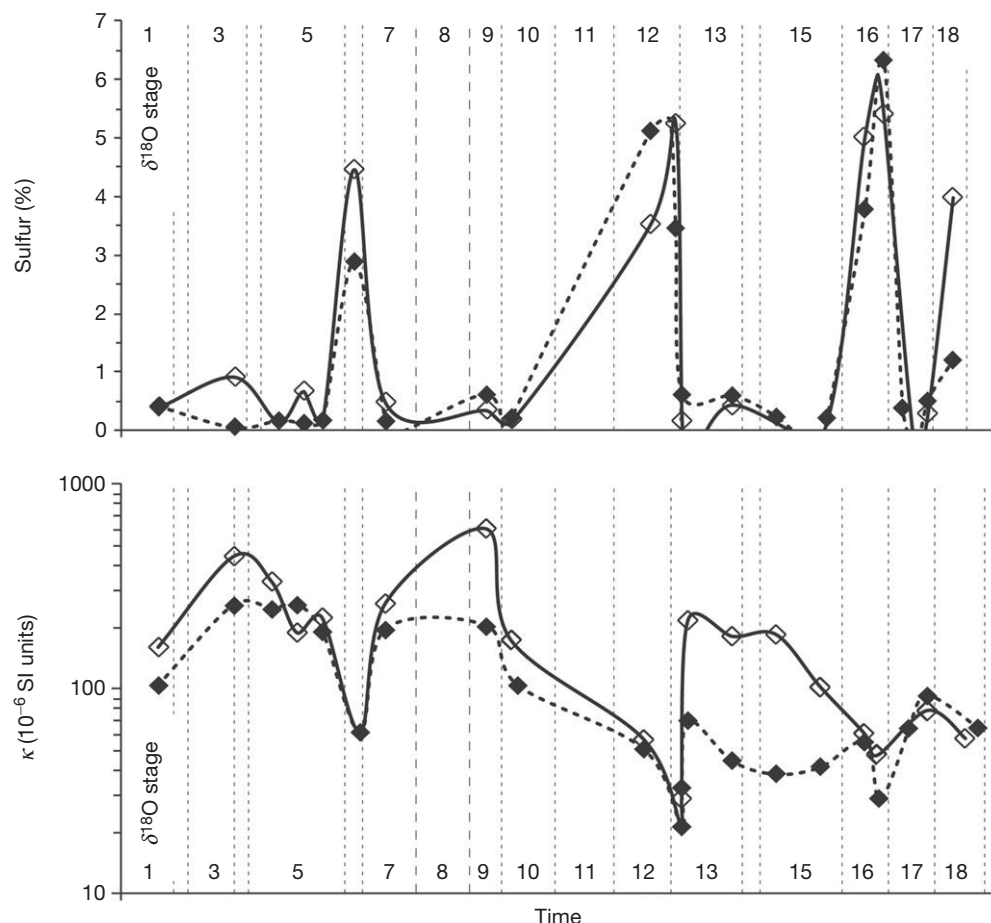


Figure 10 Sulfur (%) and MS for ODP sites 794 (open symbols) and 795 (full symbols) from the Sea of Japan through 18 marine isotope stages. Reprinted from *Quaternary Science Reviews*, 16, Vigliotti L, Magnetic properties of light and dark sediment layers from the Japan sea: Diagenetic and paleoclimatic implications, 1093–1114, Copyright (1997), with permission from Elsevier.

magnetite chains have been identified as significant remanence carriers in both oxic and anoxic oceanic records (e.g., Kirschvink, 1982; Petersen et al., 1986; Stolz et al., 1990). Mechanisms controlling their formation can also possess certain boundary conditions. Lean and McCave (1998) examined a deep-sea core east of New Zealand, identifying bacterial magnetite phases and proposing climatic mediation of conditions favoring bacterial magnetite formation. However, this record is probably not reflective of all authigenic contributions, especially in areas susceptible to diagenesis. As with fine-grained ferrimagnetic constituents within terrigenous sediment, these phases are rapidly lost as they pass through the upwardly migrating Fe redox boundary by iron-reducing bacteria (Karlin et al., 1987; Robinson et al., 2000). Thus, in general, bacterial magnetosomes typically are only found in the upper few meters of sediment. These organisms produce extracellular SP magnetite, which can be readily distinguished from coarse-grained IRD or high-coercivity eolian dusts by additional magnetic measurements. However, if they go undetected in a sediment record, they have the potential to overprint the depositional magnetic signal which may result in an erroneous interpretation.

Conclusions

The MS record at any one site is the sum of the influences of many factors including flux and subsequent redistribution (by numerous processes, the source and mineralogy of the terrigenous material, and dilution by diamagnetic materials). This signal can then be a modified transformation of the magnetic fraction postdepositionally.

Correlations between MS and other climate proxies result from the ubiquitous nature of iron oxides in marine sedimentary environments and their sensitivity to climatically driven processes. The key is to derive which aspect of the climate system they are recording. Similar records can be produced by very different processes, which can lead to erroneous interpretations. Therefore, independent constraints are required when interpreting MS record beyond the local scale. As similar delivery processes can often produce different MS records or as postdepositional processes can affect records to different degrees, it is often necessary to couple MS measurements with other simple magnetic analyses to understand the drivers of the MS changes before attempts are made to extract any paleoinformation.

In summary, MS measurements have proven to be a highly effective tool for the interpretation of paleoenvironmental change, possessing sensitivity to aspects often unresolvable by other methods. With good understanding of the reasons driving any MS changes, MS records have become part of a two-way iterative feedback process for the understanding of paleoclimate. We have to understand the climatic drivers before we can interpret the MS signal, but the MS record can provide excellent new insights into paleoclimate that may not be available using other proxies. Ongoing work to address issues of inter-lab calibration and recognition of the potential for increased resolution through particle-sized specific measurement will only increase the robustness and understanding of these techniques.

See also: [Paleoceanography, Physical and Chemical Proxies: Terrigenous Sediments](#), [Paleosols and Wind-Blown Sediments: Mineral Magnetic Analysis](#).

References

- Andrew JT and Hardardottir J (2009) A comparison of Holocene sediment- and paleo-magnetic characteristics from the margins of Iceland and East Greenland. *Jokull* 59: 51–66.
- Bloemendal J and deMenocal P (1989) Evidence for a change in the periodicity of tropical climate cycles at 2.4 Myr from whole-core magnetic susceptibility measurements. *Nature* 342: 897–900.
- Bond G, Heinrich H, Broecker W, et al. (1992) Evidence for massive discharges of icebergs into the North Atlantic Ocean during the last glacial period. *Nature* 360: 245–249.
- Carlson AE, Stoner JS, Donnelly JP, and Hillaire-Marcel C (2008) Response of the southern Greenland Ice Sheet during the last two deglaciations. *Geology* 36: 359–362.
- Dankers PH (1978) *Magnetic Properties of Dispersed Natural Iron Oxides of Known Grain Size*. PhD Thesis, University of Utrecht.
- Dearing J (1999) Magnetic susceptibility. In: Walden J, Oldfield F, and Smith J (eds.) *Environmental Magnetism: A Practical Guide*, pp. 35–62. London: Quaternary Research Association Technical Guide No. 6.
- Evans ME and Heller F (2003) *Environmental Magnetism: Principles and Applications of Enviromagnetics*. San Diego, CA: Academic Press.
- Froelich P, Klinkhammer G, Bender M, et al. (1979) Early oxidation of organic matter in pelagic sediments of the eastern equatorial Atlantic: Suboxic diagenesis. *Geochimica et Cosmochimica Acta* 43: 1075–1090.
- Goldfinger C, Grijalva K, Burgmann R, et al. (2008) Late Holocene rupture of the northern San Andreas Fault and possible stress linkage to the Cascadia subduction zone. *Bulletin of the Seismological Society of America* 98: 861–889.
- Goldfinger C, Morey AE, Nelson CH, et al. (2007) Rupture lengths and temporal history of significant earthquakes on the offshore and north coast segments of the Northern San Andreas Fault based on turbidite stratigraphy. *Earth and Planetary Science Letters* 254: 9–27.
- Hassold N, Rea D, van der Pluijm B, Parés J, Gleason J, and Ravelo A (2006) Late Miocene to Pleistocene paleoceanographic records from the Feni and Gardar Drifts: Pliocene reduction in abyssal flow. *Palaeogeography, Palaeoclimatology, Palaeoecology* 236: 290–301.
- Karlin R, Lyle M, and Heath GR (1987) Authigenic magnetite formation in suboxic marine sediments. *Nature* 326: 490–493.
- Kirschvink JL (1982) Paleomagnetic evidence for fossil biogenic magnetite in western Crete. *Earth and Planetary Science Letters* 59: 388–392.
- Kissel C, Laj C, Lehman B, Labyrie L, and Bout-Roumaizilles V (1997) Changes in the strength of the Iceland-Scotland Overflow Water in the last 200,000 years: Evidence from magnetic anisotropy analysis of core SU90-33. *Earth and Planetary Science Letters* 152: 25–36.
- Kissel C, Laj C, Mulder T, Wandres C, and Cremer M (2005) The magnetic fraction: A tracer of deep water circulation in the North Atlantic. *Earth and Planetary Science Letters* 288: 444–454.
- Lean C and McCave I (1998) Glacial to interglacial mineral magnetic and palaeoceanographic changes at Chatham Rise, SW Pacific Ocean. *Earth and Planetary Science Letters* 163: 247–260.
- Maher BA (1988) Magnetic properties of synthetic magnetites. *Geophysical Journal* 94: 83–96.
- Maslin MA, Haug GH, Sarnthein M, and Tiedemann R (1996) The progressive intensification of northern hemisphere glaciation as seen from the North Pacific. *Geologische Rundschau* 85: 452–465.
- McCave IN (2007) Deep-sea sediment deposits and properties controlled by currents. In: Hillaire-Marcel C and De Vernal A (eds.) *Developments in Marine Geology*, vol. 1, pp. 19–62. New York: Elsevier.
- Ozdemir O and Banerjee SK (1982) A preliminary magnetic study of soil samples from west-central Minnesota. *Earth and Planetary Science Letters* 59: 393–403.
- Peters C, Austin WEN, Walden J, and Hibbert FD (2010) Magnetic characterization and correlation of a Younger Dryas tephra in North Atlantic marine sediments. *Journal of Quaternary Science* 25: 339–347.
- Petersen N, von Dobeneck T, and Vali H (1986) Fossil bacterial magnetite in deep-sea sediments from the South Atlantic Ocean. *Nature* 320: 611–615.

- Pirrung M, Fütterer D, Grobe H, Matthiessen J, and Niessen F (2002a) Magnetic susceptibility and ice-rafted debris in surface sediments of the Nordic Seas: Implications for Isotope Stage 3 oscillations. *Geo-Marine Letters* 22: 1–11.
- Pirrung M, Hillenbrand C, Diekmann B, Fütterer D, Grobe H, and Kuhn G (2002b) Magnetic susceptibility and ice-rafted debris in surface sediments of the Atlantic sector of the Southern Ocean. *Geo-Marine Letters* 22: 170–180.
- Pugh R, McCave I, Hillenbrand C, and Kuhn G (2009) Circum-Antarctic age modelling of Quaternary marine cores under the Antarctic Circumpolar Current: Ice-core dust–magnetic correlation. *Earth and Planetary Science Letters* 284: 113–123.
- Rasmussen TL, van Weering TCE, and Labeyrie L (1997) Climatic instability, ice sheets and ocean dynamics at high northern latitudes during the last glacial period (58–10 ka BP). *Quaternary Science Reviews* 16: 71–80.
- Robinson SG (1986) The late Pleistocene palaeoclimatic record of North Atlantic deep-sea sediments revealed by mineral-magnetic measurements. *Physics of the Earth and Planetary Interiors* 42: 22–47.
- Robinson SG, Sahota JTS, and Oldfield F (2000) Early diagenesis in North Atlantic abyssal plain sediments characterized by rock-magnetic and geochemical indices. *Marine Geology* 163: 77–107.
- Ruddiman WF (1977) North Atlantic ice-rafting: A major change at 75,000 years before the present. *Science* 196: 1208–1211.
- Sagnotti L, Rochette P, Jackson M, Vadeboin F, Dinarès-Turell J, and Winkler A (2003) Inter-laboratory calibration of low-field magnetic and anhysteretic susceptibility measurements. *Physics of the Earth and Planetary Interiors* 138: 25–38.
- Snowball I and Moros M (2003) Saw-tooth pattern of North Atlantic current speed during Dansgaard–Oeschger cycles revealed by the magnetic grain size of Reykjanes Ridge sediments at 59°N. *Paleoceanography* 18: 1026–1037.
- Stolz JF, Lovley DR, and Haggerty SE (1990) Biogenic magnetite and the magnetization of sediments. *Journal of Geophysical Research* 95: 4355–4361.
- Stoner JS and Andrews JT (1999) The North Atlantic as a Quaternary magnetic archive. In: Maher BA and Thompson R (eds.) *Quaternary Climates, Environments and Magnetism*, pp. 49–80. Cambridge: Cambridge University Press.
- Stoner JS, Channell JET, and Andresen C (1995) Magnetic properties of deep-sea sediments off southwest Greenland: Evidence for major differences between the last two deglaciations. *Geology* 23: 241–244.
- Tauxe L (2010) *Essentials of Paleomagnetism*. Berkeley, CA: University of California Press.
- Thompson R and Oldfield F (1986) *Environmental Magnetism*. London: Allen & Unwin.
- Vigliotti L (1997) Magnetic properties of light and dark sediment layers from the Japan sea: Diagenetic and paleoclimatic implications. *Quaternary Science Reviews* 16: 1093–1114.
- Watkins S and Maher BA (2003) Magnetic characterization of present-day deep-sea sediments and sources in the North Atlantic. *Earth and Planetary Science Letters* 214: 379–394.
- Watkins SJ, Maher BA, and Bigg GR (2007) Ocean circulation at the Last Glacial Maximum: A combined modeling and magnetic proxy-based study. *Paleoceanography* 22: 1–20.

Relevant Websites

- http://www.gmw.com/magnetic_properties/pdf/Om0409%20J_Dearing_Handbook_iss7.pdf – Environmental Magnetic Susceptibility: Using the Bartington MS2 system.
- www.icdp-online.org – International Continental Drilling Program (ICDP).
- www.iodp.org – International Ocean Drilling Program (IODP).
- <http://magician.ucsd.edu/Essentials/index.html> – Essentials of Paleomagnetism.

Nutrient Proxies

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Introduction

The distribution of marine nutrients in the past is of interest to paleoceanographers for two main reasons. First, nutrients control oceanic primary productivity, which is believed to be an important control on atmospheric CO₂ on glacial–interglacial timescales. Second, nutrients are useful tracers of deep-water masses, and changes in the modes and locations of deep-water formation are believed to have fundamentally influenced Quaternary climates. Reconstruction of paleonutrients relies mainly on five proxies that are recorded in marine sediments: foraminiferal $\delta^{13}\text{C}$, Cd/Ca, Ba/Ca, and Zn/Ca; and organic matter $\delta^{15}\text{N}$.

Carbon-13

Systematics

Dissolved inorganic carbon (DIC), also known as ΣCO_2 , is composed of three dominant species in seawater: dissolved carbon dioxide (CO₂(aq)), bicarbonate (HCO₃[−]), and carbonate (CO₃^{2−}), having relative concentrations on the order of 1%, 90%, and 10%, respectively. Furthermore, carbon has two stable isotopes: ¹²C (98.9%) and ¹³C (1.1%). Fractionation between these two isotopes is expressed in delta notation:

$$\delta^{13}\text{C} = \left[\left(^{13}\text{C}/^{12}\text{C} \right)_{\text{sample}} / \left(^{13}\text{C}/^{12}\text{C} \right)_{\text{standard}} - 1 \right] \times 1000 \quad [1]$$

where the standard is a calcium carbonate, usually referenced to the Pee Dee Belemnite standard. On this scale, the $\delta^{13}\text{C}$ of DIC ranges between about −1‰ and 2.5‰ in the world's oceans (Kroopnick, 1985). Photosynthetic fixation of carbon preferentially utilizes ¹²C such that primary marine organic matter $\delta^{13}\text{C}$ is typically between −20‰ and −30‰. This leaves the remaining surface ocean DIC pool slightly enriched in $\delta^{13}\text{C}$. As organic matter sinks through the water column and decays, its ¹²C-rich carbon is regenerated along with other nutrients such as phosphate and nitrate, resulting in a water column $\delta^{13}\text{C}$ profile that is inversely correlated with these nutrients (Figure 1). If biogeochemical cycling were the only process acting on the DIC pool, $\delta^{13}\text{C}$ would decrease by about 1.1‰ for each 1 μmol kg^{−1} increase in dissolved phosphate, and an oceanic range of about 3.3‰ would be possible (Lynch-Stieglitz et al., 1995). Regional differences in the $\delta^{13}\text{C}$ of marine organic matter may alter the slope of this relationship.

In addition to biogeochemical cycling, the $\delta^{13}\text{C}$ of surface ocean DIC is significantly affected by air–sea exchange of CO₂. If DIC were at isotopic equilibrium with atmospheric CO₂, it would be enriched by about 8‰ relative to the atmosphere at 20 °C. This value results from the dominance of HCO₃[−], which is enriched by 8.5‰ relative to the atmosphere, with smaller contributions from CO₃^{2−} (6‰) and CO₂(aq) (−1‰) (Lynch-Stieglitz et al., 1995). The enrichment for HCO₃[−] (and

therefore DIC) increases by about 0.1‰ per degree of cooling. However, isotopic equilibration between the surface ocean mixed layer and the atmosphere is never reached because the time required (on the order of a decade) is longer than surface ocean mixing times. The extent of equilibration may be increased through longer air–sea contact time or high winds, which would increase DIC $\delta^{13}\text{C}$. A final air–sea effect occurs in areas where there is net movement of CO₂ into or out of the surface ocean. Both CO₂(aq) and atmospheric CO₂ are isotopically light compared to DIC. Regions that absorb CO₂ from the atmosphere, such as the low-*p*CO₂ North Atlantic, therefore experience $\delta^{13}\text{C}$ depletion. Areas that emit CO₂ to the atmosphere, such as the high-*p*CO₂ eastern equatorial Pacific, experience $\delta^{13}\text{C}$ enrichment. The net effect of the various air–sea exchange processes spans a surface ocean range of about 2‰ (Lynch-Stieglitz et al., 1995). In addition, the anthropogenic evolution of atmospheric CO₂ toward lower $\delta^{13}\text{C}$ values, known as the Suess effect, results in a progressive lowering of surface water $\delta^{13}\text{C}$ that has already propagated into the deep North Atlantic (Olsen and Ninnemann, 2010).

Paleoceanographic Reconstruction

Reconstruction of past ocean $\delta^{13}\text{C}$ relies mainly on the calcitic (CaCO₃) tests of protozoa called Foraminifera. Foraminiferal calcite carbon is presumed to be derived from dissolved HCO₃[−], and may therefore be expected to record the $\delta^{13}\text{C}$ of DIC, in the absence of complicating biological effects. Core-top calibrations show that various taxa of benthic Foraminifera faithfully reflect the $\delta^{13}\text{C}$ of bottom water DIC (Duplessy et al., 1984). Epifaunal species are preferred for reconstructions because infaunal taxa record pore water values, which are typically lower than those of bottom waters due to organic matter remineralization (Zahn et al., 1986). However, the $\delta^{13}\text{C}$ of even epifaunal species may be lower than that of bottom waters in regions where organic matter rain rates are very high (Mackensen et al., 1993).

Benthic foraminiferal $\delta^{13}\text{C}$ provides a picture of the past distribution of deep-water masses. Biogeochemical cycling and air–sea exchange cause most surface waters, particularly those stripped of nutrients, to be more enriched in $\delta^{13}\text{C}$ compared to the deep ocean. North Atlantic Deep Water (NADW) is today formed from such low-nutrient waters, and therefore carries a high- $\delta^{13}\text{C}$ signature (Figure 2). In contrast, Antarctic Bottom Water (AABW) and Antarctic Intermediate Water (AAIW) are formed from poorly ventilated surface waters and therefore carry low- $\delta^{13}\text{C}$ signatures. Lowest $\delta^{13}\text{C}$ values are found today in the deep North Pacific, far from the well-ventilated NADW source.

Today NADW transports a large amount of heat into the North Atlantic, and past changes in its formation are believed to have strongly impacted regional climates. The most extensively studied period in this regard is the Last Glacial Maximum

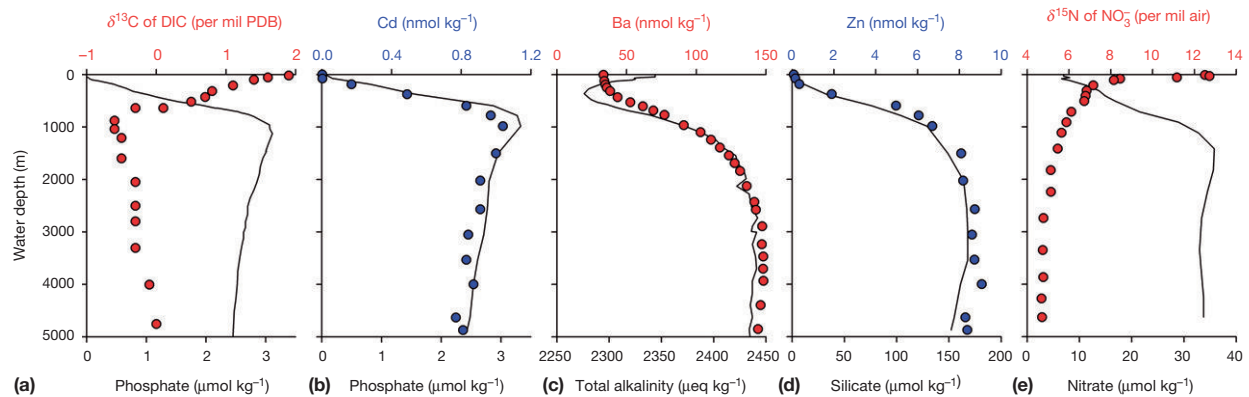


Figure 1 Modern seawater profiles of the main nutrient proxies. (a) $\delta^{13}\text{C}$ of DIC (Kroopnick, 1985) compared to dissolved phosphate (Broecker et al., 1982) in the central North Pacific; (b) Dissolved Cd compared to dissolved phosphate in the eastern North Pacific (Bruland, 1980); (c) Dissolved Ba (Ostlund et al., 1987) compared to total alkalinity (Broecker et al., 1982) in the central North Pacific; (d) Dissolved Zn compared to dissolved silica in the eastern North Pacific (Bruland, 1980); (e) $\delta^{15}\text{N}$ of dissolved nitrate compared to dissolved nitrate in the Indian sector of the Southern Ocean (Sigman et al., 1999).

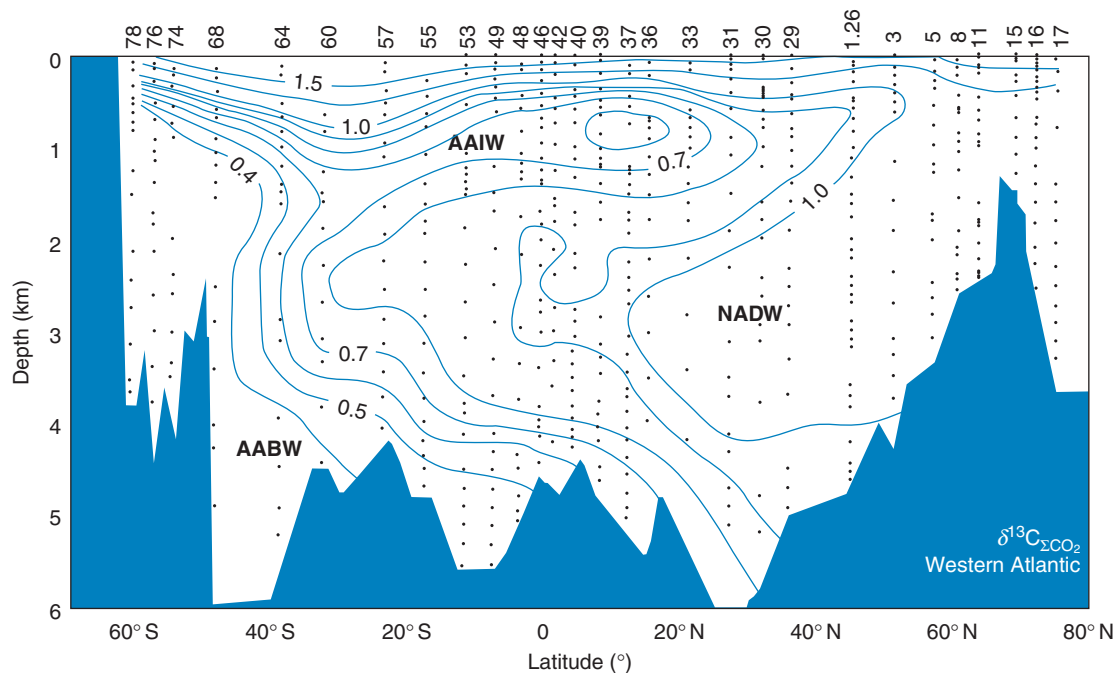


Figure 2 Meridional section of the $\delta^{13}\text{C}$ of DIC in the western Atlantic Ocean during the 1970s (reproduced from Kroopnick PM (1985) The distribution of ^{13}C of ΣCO_2 in the world oceans. *Deep-Sea Research* 32: 57–84). Major water masses are indicated, and the numbers at the top refer to oceanographic cruise stations. Preindustrial $\delta^{13}\text{C}$ values were slightly higher at the sea surface and in the North Atlantic, shallower than ~ 2 km (Olsen and Ninnemann, 2010).

(LGM, ~ 20 ka BP), for which meridional Atlantic $\delta^{13}\text{C}$ sections have been created. Such reconstructions suggest that low- $\delta^{13}\text{C}$ AABW expanded into the North Atlantic and that its boundary with high- $\delta^{13}\text{C}$ NADW shoaled from a water depth of 4 to ~ 2 –3 km (Figure 3; Curry and Oppo, 2005; Duplessy et al., 1988). This reorganization was long assumed to be due to a reduction in the formation rate of NADW, but other more rate-sensitive proxies have since underscored the fact that paleonutrients constrain only the spatial extent of water masses and not their fluxes. The glacial form of NADW, dubbed Glacial North

Atlantic Intermediate Water (GNAIW), likely formed significantly southward of its modern deep convection regions. This migration may therefore have cooled high latitudes even if formation rates were not reduced. On longer Quaternary time-scales, the $\delta^{13}\text{C}$ gradient between the North Atlantic and Pacific has been used to monitor NADW extent. As during the LGM, NADW volume was apparently reduced during all Northern Hemisphere glaciations.

One interesting aspect of the glacial $\delta^{13}\text{C}$ distribution is that the lowest values occurred in the deep Southern Ocean, rather

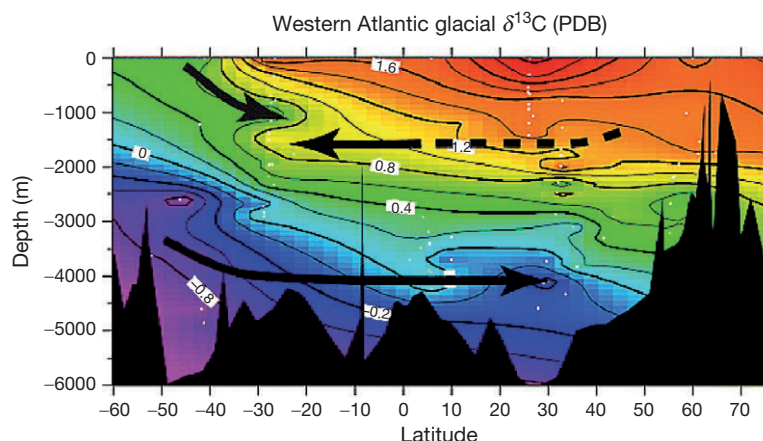


Figure 3 Meridional section of benthic foraminiferal $\delta^{13}\text{C}$ for the LGM western Atlantic (Curry and Oppo, 2005). Arrows indicate presumed flow directions of AAIW, GNAIW, and AABW, and small white dots show locations of sediment cores used in the reconstruction.

than in the North Pacific as they do today. Glacial production of a North Pacific Deep Water could explain the pattern, but such a water mass is difficult to document, in part because of the poor preservation of calcite in the North Pacific. Some have attributed low Southern Ocean $\delta^{13}\text{C}$ values to an organic matter microhabitat effect (Mackensen et al., 1993), noting that Cd/Ca (see Section ‘Cadmium’) does not support a very high-nutrient glacial AABW (Boyle, 1992). However, the relative uniformity of glacial $\delta^{13}\text{C}$ in the Atlantic sector of the Southern Ocean suggests that the values are probably reliable (Ninnemann and Charles, 2002). Independent estimates of deep sea salinity and temperature require glacial AABW to have been particularly dense, so perhaps GNAIW avoided extensive mixing in the Southern Ocean en route to the North Pacific, leaving AABW anomalously low in $\delta^{13}\text{C}$. Reduced vertical mixing between these two water masses is supported by an enhanced oxygen isotope vertical gradient in the deep Atlantic (Lund et al., 2011).

The $\delta^{13}\text{C}$ pattern in the glacial Atlantic implies that there was a net shift of DIC from the upper ocean into the deep ocean. A comparable $\delta^{13}\text{C}$ shift has been documented in the glacial Indian Ocean, and although data coverage in the Pacific is comparatively poor, an intermediate- to mid-depth $\delta^{13}\text{C}$ gradient is generally supported (Herguera et al., 2010). Glacial DIC deepening may have acted to lower atmospheric CO_2 , through both the direct effect of decreased surface ocean DIC and the compounding effect of increased oceanic alkalinity due to greater dissolution of seafloor CaCO_3 (Sigman et al., 2010).

Reconstruction of surface water $\delta^{13}\text{C}$ appears to be more complicated than deep-water work because of confounding influences on planktonic foraminiferal $\delta^{13}\text{C}$ (Keigwin and Boyle, 1989; Spero, 1998). For example, culture work shows that photosymbionts, respiration, and carbonate chemistry affect the $\delta^{13}\text{C}$ of various taxa. Nevertheless, planktonic-benthic comparisons may record past shifts of DIC partitioning between the upper and deep oceans, with the implications for atmospheric CO_2 mentioned earlier (Shackleton et al., 1983). A widespread planktonic $\delta^{13}\text{C}$ minimum during the last deglaciation may represent the release of DIC that was

stored in the isotopically light and isolated AABW during glacial times (Spero and Lea, 2002).

On glacial-interglacial timescales, the partitioning of carbon between land and ocean must also be considered. It is estimated from benthic Foraminifera that the average $\delta^{13}\text{C}$ value of the world’s oceans was 0.3‰ lower during the LGM (Duplessy et al., 1988). This estimate is subject to considerable error because of the scarcity of data in some regions, particularly the North Pacific. Assuming an average terrestrial carbon isotopic composition of -25‰ , the glacial lowering could be explained by a transfer of 400–500 Gt (10^{15} g) of carbon from land into the ocean, or about 20% of the terrestrial biomass (Crowley, 1995). Independent reconstructions of the LGM terrestrial biosphere, based on paleoecological data, argue for a greater reduction, by up to two or three times that inferred from mean ocean $\delta^{13}\text{C}$. Factors that might reconcile these disparate estimates include glacial carbon storage on exposed continental shelves or in isotopically light methane hydrates.

Trace Metals: Cadmium, Barium, and Zinc

Numerous dissolved trace elements in the ocean exhibit vertical profiles that resemble those of nutrients. That is, they are at low concentrations in surface waters and at greater concentrations at depth. In some cases, this behavior is linked to the element’s importance as a micronutrient. For example, iron is essential for the synthesis of chlorophyll and various algal proteins and is the limiting nutrient in some regions of the ocean. In other cases, nutrient-like behavior may result simply from adsorption onto particulate organic matter in the photic zone and co-remineralization in the deep sea. Regardless of biogeochemical mechanisms, such elements may be useful as nutrient and water mass tracers. Paleoreconstruction relies on the fact that various elements are incorporated into foraminiferal calcite during precipitation. In particular, divalent cations are believed to substitute for calcium in the calcite crystal matrix. Three divalent trace metals have been developed as paleonutrient tracers: cadmium (Cd), barium (Ba), and zinc (Zn).

Cadmium

Dissolved Cd has an oceanic distribution very similar to that of the major nutrient phosphate (Figure 1). Both are nearly completely removed from most surface waters and regenerated at depth, with an intermediate-depth concentration maximum near 1 km. Cd therefore behaves like a labile (easily remineralized) nutrient, and its concentration increases about fivefold between the deep North Atlantic and North Pacific. Although Cd has been linked to at least one important algal metalloenzyme, it is not clear if this use is sufficient to explain its ocean-wide nutrient-like behavior. The global correlation between Cd and phosphate follows a slight curve that may be explained by preferential uptake of Cd over phosphate by particulate organic matter (Elderfield and Rickaby, 2000).

Benthic foraminiferal Cd/Ca ratios reflect seawater Cd concentrations (Boyle, 1992). The relationship between foraminiferal Cd/Ca and seawater Cd is expressed in terms of the partition coefficient:

$$D_{Cd} = (Cd/Ca)_{\text{foram}} / (Cd/Ca)_{\text{seawater}} \quad [2]$$

Calcareous benthic foraminiferal D_{Cd} varies with water depth, from ~ 1.3 in the upper ocean to 2.9 below 3 km. The aragonitic benthic foraminifer *Hoeglundina elegans* has a D_{Cd} of 1.0, invariant with depth. Calcitic partition coefficients also appear to be reduced in waters that are very undersaturated with respect to calcite, such as in the deep Pacific (McCorkle et al., 1995).

Cd/Ca generally supports $\delta^{13}C$ observations indicating that low-nutrient NADW was less extensive in the glacial Atlantic Ocean and was present as the shallower GNAIW (Figure 4; Boyle and Keigwin, 1987; Marchitto and Broecker, 2006). A notable difference between the two proxies is the already mentioned deep Southern Ocean discrepancy, where glacial $\delta^{13}C$ shows very low values in contrast to Cd/Ca, which is similar to the current value. An undersaturation effect on Cd/Ca and/or a microhabitat effect on $\delta^{13}C$ might explain some of the difference. Alternatively, large-scale changes in deep-water ventilation might have decoupled the two tracers such that AABW was depleted in $\delta^{13}C$ without being significantly enriched in Cd.

The Southern Ocean problem highlights the fact that because $\delta^{13}C$ and Cd behave somewhat differently in the modern ocean, unique information may be derived from paired

measurements that is not available from either tracer alone. Specifically, air–sea exchange affects $\delta^{13}C$ but not Cd, so combining the two can potentially reveal past air–sea processes (Lynch-Stieglitz and Fairbanks, 1994). Paired measurements from various regions of the glacial ocean suggest that different deep-water masses had distinct air–sea signatures, allowing for the separation of biogeochemical aging and water mass mixing. LGM Atlantic observations may be largely explained by mixing between well-ventilated GNAIW and poorly ventilated AABW (Marchitto and Broecker, 2006).

Planktonic Foraminifera incorporate Cd with a similar range of partition coefficients as benthics (Delaney, 1989), but less paleoceanographic work has been done with planktonics. Cd/Ca from a polar species suggests that LGM nutrient levels in the high-latitude North and South Atlantic were not much different from today, arguing against changes in NADW and AABW end member properties as explaining the deep sea record (Keigwin and Boyle, 1989). Temperature appears to affect the incorporation of Cd in at least one planktonic species (Rickaby and Elderfield, 1999), and correction for this influence implies that glacial Southern Ocean surface nutrient levels were elevated south of the modern Polar Front, but relatively unchanged north of it (Elderfield and Rickaby, 2000), seemingly in contradiction with $\delta^{15}N$ data (see Section ‘Nitrate Utilization’).

Barium

Dissolved Ba is moderately depleted in surface waters and reaches maximum concentrations below ~ 2 km (Figure 1). Its distribution resembles that of alkalinity, but the association is coincidental (Lea and Boyle, 1989). Ba is removed from shallow waters mainly by barite formation in decaying organic matter, while alkalinity is removed mainly by $CaCO_3$ formation. Both are regenerated at depth as their carrier phases dissolve, and deep-water masses have characteristic Ba and alkalinity values, with Ba increasing about threefold between the deep North Atlantic and North Pacific. Due to its refractory (less easily remineralized) behavior, Ba may offer information that is distinct from the more labile Cd. Ba is incorporated into several taxa of calcitic benthic Foraminifera with a partition coefficient of ~ 0.4 (Lea and Boyle, 1989), and there is

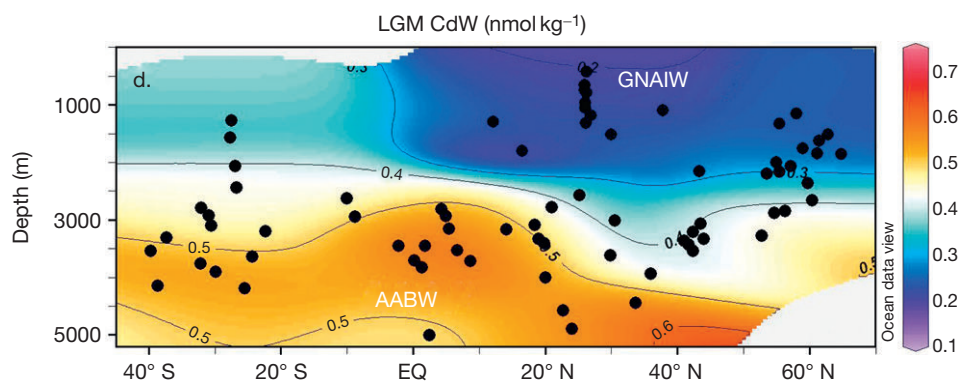


Figure 4 Meridional section of seawater Cd concentration in the LGM Atlantic, reconstructed from benthic foraminiferal Cd/Ca (Marchitto and Broecker, 2006). Black dots show locations of sediment cores used in the reconstruction.

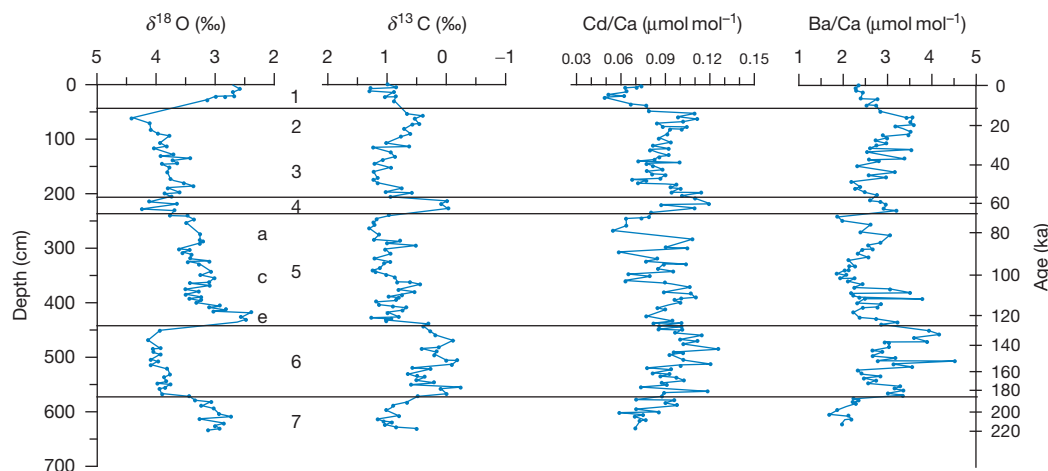


Figure 5 Records of benthic foraminiferal $\delta^{18}\text{O}$, $\delta^{13}\text{C}$, Cd/Ca, and Ba/Ca in a sediment core from the deep North Atlantic spanning the past 210 ka (Lea and Boyle, 1990a). Low $\delta^{13}\text{C}$ and high Cd/Ca and Ba/Ca generally occur during glacial stages (indicated by high $\delta^{18}\text{O}$ and labeled as marine isotope stages 2, 4, and 6), indicative of high nutrients and the presence of AABW. Note the reversed scales for $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$.

evidence that this value is reduced in strongly undersaturated waters (McCorkle et al., 1995).

For the LGM Atlantic, benthic Ba/Ca supports the view that low-nutrient NADW was replaced by the shallower GNAIW (Figure 5; Lea and Boyle, 1990b). Low intermediate-depth Ba/Ca also appears to rule out the Mediterranean Sea as an important contributor to GNAIW, since water from that basin carries high Ba concentrations. In contrast to $\delta^{13}\text{C}$ and Cd/Ca, however, there was no apparent glacial Ba gradient between the deep Atlantic and the Pacific. Deep ocean Ba may have become decoupled from the other nutrient tracers because of an increase in barite regeneration at the seafloor, possibly associated with increased productivity (Lea and Boyle, 1990b).

Zinc

Dissolved Zn has an oceanic distribution very similar to that of the nutrient silica (Figure 1). Both are nearly completely removed from most surface waters, but unlike the labile Cd and phosphate, they lack intermediate-depth concentration maxima. Zn therefore behaves like a refractory nutrient, with maximum concentrations below $\sim 1\text{--}2$ km water depth. Zn is an essential micronutrient for many marine organisms, second only to iron among the biologically important trace metals. In particular, its use by diatoms (algal protists with siliceous shells) in the enzyme carbonic anhydrase may explain its oceanic association with silica. Zn concentrations increase more than tenfold between the deep North Atlantic and North Pacific, and there is a sevenfold meridional increase within the deep Atlantic alone. Zn/Ca may therefore be a very sensitive tracer of past interactions between NADW and AABW (Marchitto et al., 2002).

At least two species of calcitic benthic Foraminifera incorporate Zn with a partition coefficient that depends strongly on the saturation state. In sufficiently supersaturated waters D_{Zn} is ~ 9 , but it may be as low as ~ 4 in the corrosive deep Pacific. The glacial increase in deep North Atlantic Zn due to the increased presence of AABW was therefore partially muted in Zn/Ca because of AABW's lower saturation state (Marchitto et al., 2002). Nevertheless, paired Zn/Ca and Cd/Ca measurements provide

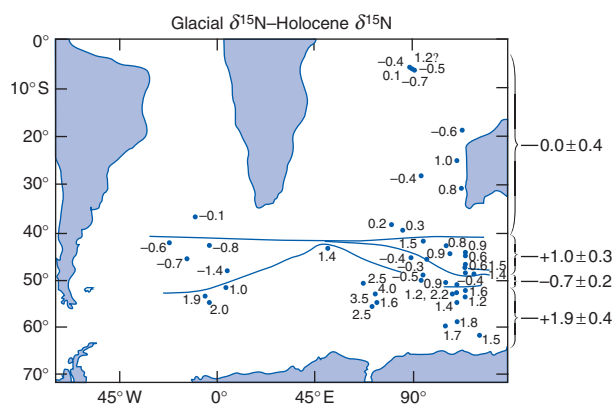


Figure 6 Difference between LGM and Holocene bulk sediment $\delta^{15}\text{N}$ in the Southern Ocean and Indian Ocean (Francois et al., 1997). Numbers at right show mean values for four regions. There was an increase in $\delta^{15}\text{N}$ south of the modern Polar Front ($\sim 50^\circ\text{S}$) during the LGM, suggesting increased NO_3^- utilization due to reduced nutrient upwelling.

strong evidence for glacial AABW expansion, and not some independent change in NADW nutrient content.

Nitrogen-15

Nitrate Utilization

Along with carbon and phosphate, nitrate (NO_3^-) is an essential major nutrient for marine primary production, and is often the limiting nutrient in the open ocean. Like carbon, nitrogen has two stable isotopes: ^{14}N (99.6%) and ^{15}N (0.4%). Fractionation is again expressed in delta notation:

$$\delta^{15}\text{N} = \left[\left(\frac{^{15}\text{N}}{^{14}\text{N}} \right)_{\text{sample}} / \left(\frac{^{15}\text{N}}{^{14}\text{N}} \right)_{\text{standard}} - 1 \right] \times 1000 \quad [3]$$

where the standard is atmospheric N_2 . Phytoplankton preferentially incorporate the lighter isotope, leading to $\delta^{15}\text{N}$ enrichment of the remaining dissolved NO_3^- pool. In regions where

the NO_3^- supply is heavily utilized, the surface ocean isotopic enrichment can be relatively large (order of 10‰; **Figure 1**), and organic matter $\delta^{15}\text{N}$ increases in parallel. To the first order, the $\delta^{15}\text{N}$ of sedimentary organic matter may therefore be used to reconstruct past surface ocean NO_3^- $\delta^{15}\text{N}$, and thus the degree of NO_3^- utilization (**Altabet and Francois, 1994**). There are indications that sedimentary bulk organic matter $\delta^{15}\text{N}$ may be subject to diagenetic alteration in some environments, which might be avoidable by measuring diatom-bound $\delta^{15}\text{N}$ (**Robinson and Sigman, 2008**).

Past NO_3^- utilization is of particular interest in the Southern Ocean, the largest area of incomplete consumption of NO_3^- and phosphate by phytoplankton in the modern oceans. Here, deep waters charged with nutrients and DIC upwell to the surface, but they are subducted again before their nutrient

load can be efficiently extracted, resulting in a missed opportunity for the ocean's 'biological pump' to sequester CO_2 (**Sigman et al., 2010**). Hypothetically, the resulting leak of CO_2 into the atmosphere could have been stemmed during glacial times either by increased biological productivity or by reduced vertical mixing. Both processes would be recorded as increases in $\delta^{15}\text{N}$ due to a more complete NO_3^- utilization. South of the modern Polar Front, $\delta^{15}\text{N}$ data indeed suggest that NO_3^- utilization was enhanced during the last glaciation (**Figure 6**), and proxy evidence for reduced export production confirms that this change was related to reduced upwelling and/or stronger surface stratification (**Francois et al., 1997; Robinson and Sigman, 2008**). North of the modern Polar Front, export production increased, but NO_3^- utilization apparently did not rise as much as it did south of the front, implying that the

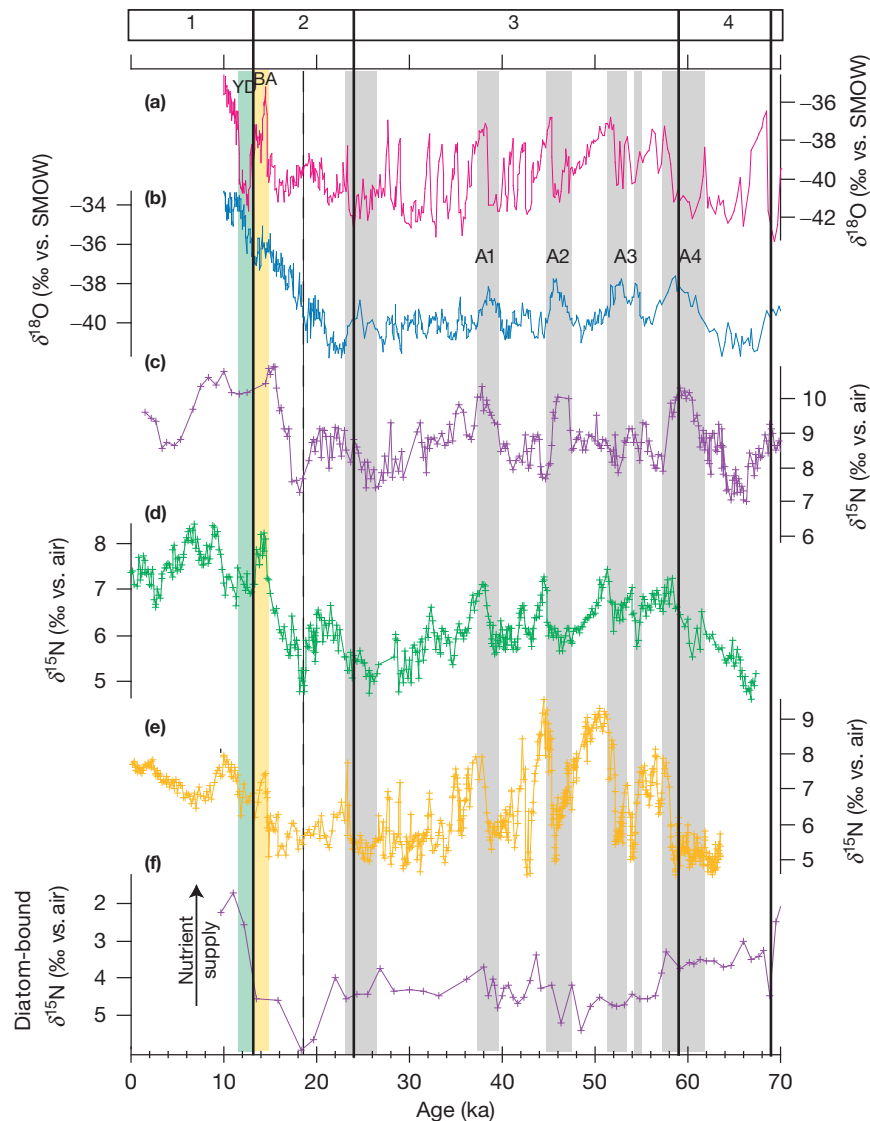


Figure 7 Sedimentary $\delta^{15}\text{N}$ records from off southern Chile (c) (**Robinson et al., 2007**), California (d) (**Hendy et al., 2004**), and Oman (e) (**Altabet et al., 2002**), compared to $\delta^{18}\text{O}$ of ice (a proxy for air temperature) in Greenland (a) and Antarctica (b). Numbers at the top indicate marine isotope stages, and gray bars denote millennial-scale warm events in Antarctica. Also shown (on a reversed scale) is $\delta^{15}\text{N}$ in the subantarctic zone of the Southern Ocean (f). Reproduced from **Robinson RS, Mix A, and Martinez P (2007) Southern Ocean control on the extent of denitrification in the southeast Pacific over the last 70 ka. *Quaternary Science Reviews* 26: 201–212.**

enhanced productivity could have resulted from increased upwelling. It has been suggested that the changes south of the Polar Front had a greater impact on the atmosphere, contributing to glacial CO₂ lowering.

Denitrification

A second important influence on the $\delta^{15}\text{N}$ of dissolved NO₃[−] is water column denitrification, which represents an important loss of fixed nitrogen from the oceans. Under suboxic conditions, NO₃[−] may be used by certain bacteria as an electron acceptor during organic matter degradation, producing N₂O and N₂. Like NO₃[−] utilization, this process preferentially uses ¹⁴N and results in $\delta^{15}\text{N}$ enrichment of the remaining NO₃[−] pool. While well-oxygenated deep waters are typically ~5–6‰ in $\delta^{15}\text{N}$, denitrification may drive the values above 18‰. Today, denitrification occurs primarily in the intermediate-depth oxygen minimum zones (OMZs) of the eastern tropical North and South Pacific, and the Arabian Sea.

Along the western Mexico continental margin, bulk sediment $\delta^{15}\text{N}$ was about 2–3‰ lower during late Quaternary glacial intervals, indicative of reduced denitrification in this region (Ganeshram et al., 2002). Other sediment cores from the eastern tropical North Pacific exhibit reduced organic carbon content and increased bioturbation during glacial times, consistent with erosion of the regional OMZ. Arabian Sea sedimentary $\delta^{15}\text{N}$ was similarly reduced (Altabet et al., 2002), suggesting that global rates of denitrification were significantly lower during glacial periods. This may have resulted in an increased oceanic inventory of NO₃[−], possibly stimulating primary production in oligotrophic (subtropical gyre) regions and contributing to the glacial drawdown of atmospheric CO₂. However, phosphate limitation likely became important, reducing overall NO₃[−] fixation and limiting the biological impact of reduced denitrification (Ganeshram et al., 2002).

OMZ weakening during glacial periods may have resulted from decreased upwelling-driven productivity in those regions, and/or increased ventilation by high-oxygen intermediate waters. In the Arabian Sea, sediment $\delta^{15}\text{N}$ also decreased during millennial-scale intervals corresponding to the Northern Hemisphere cold stages of the last Ice Age, the so-called Dansgaard-Oeschger stadials (Figure 7(a) and (e); Altabet et al., 2002). Reduced denitrification at these times was likely due to reduced summer upwelling and productivity in response to a weakened southwest Indian monsoon. $\delta^{15}\text{N}$ off southern Chile also displayed millennial-scale variability during the last Ice Age, but with a timing that more closely mimics the temperature history of Antarctica (Figure 7(b) and (c); Robinson et al., 2007). $\delta^{15}\text{N}$ records from the eastern North Pacific OMZ appear to transition from Antarctic-like timing near the equator to Greenland-like timing further north (Figure 7(d); Hendy et al., 2004; Robinson et al., 2007; Pichevin et al., 2010), pointing to some combination of southern and northern forcing along this margin. Enhanced ventilation of Subantarctic Mode Water during Antarctic cold phases may have delivered more oxygen to the eastern tropical and South Pacific OMZs, resulting in less denitrification. Furthermore, if enhanced NO₃[−] utilization occurred in the Subantarctic Southern Ocean during cold phases, as suggested

by higher diatom-bound $\delta^{15}\text{N}$ (Figure 7(f); Robinson et al., 2007), then nutrient delivery to the low-latitude world ocean would have been reduced, leading to lower productivity and less demand for oxygen in the regions of the modern OMZs. In contrast to the Southern Ocean, changes in upwelling and productivity above the low-latitude OMZs are believed to exert little leverage on atmospheric CO₂ because nutrients upwelled there are completely utilized eventually (Sigman et al., 2010).

See also: **Diatom Records:** Antarctic Waters. **Glacial Climates:** Thermohaline Circulation. **Paleoceanography:** Paleoceanography An Overview. **Paleoceanography, Biological Proxies:** Benthic Foraminifera; Planktic Foraminifera. **Paleoceanography, Physical and Chemical Proxies:** Carbon Cycle Proxies ($\delta^{11}\text{B}$, $\delta^{13}\text{C}_{\text{calcite}}$, $\delta^{13}\text{C}_{\text{organic}}$, Shell Weights, B/Ca, U/Ca, Zn/Ca, Ba/Ca); Dissolution of Deep-Sea Carbonates; Mg/Ca and Sr/Ca Paleothermometry from Calcareous Marine Fossils. **Paleoclimate Reconstruction:** Sub-Milankovitch (DO/Heinrich) Events.

References

- Altabet MA and Francois R (1994) Sedimentary nitrogen isotopic ratio as a recorder for surface ocean nitrate utilization. *Global Biogeochemical Cycles* 8: 103–116.
- Altabet MA, Higginson MJ, and Murray DW (2002) The effect of millennial-scale changes in Arabian Sea denitrification on atmospheric CO₂. *Nature* 415: 159–162.
- Boyle EA (1992) Cadmium and $\delta^{13}\text{C}$ paleochemical ocean distributions during the stage 2 glacial maximum. *Annual Review of Earth and Planetary Science* 20: 245–287.
- Boyle EA and Keigwin LD (1987) North Atlantic thermohaline circulation during the past 20,000 years linked to high-latitude surface temperature. *Nature* 330: 35–40.
- Broecker WS, Spencer DW, and Craig H (1982) *GEOSCS Pacific Expedition*. Washington, DC: US Government Printing Office.
- Bruland KW (1980) Oceanographic distributions of cadmium, zinc, nickel, and copper in the North Pacific. *Earth and Planetary Science Letters* 47: 176–198.
- Crowley T (1995) Ice age terrestrial carbon changes revisited. *Global Biogeochemical Cycles* 9: 377–390.
- Curry WB and Oppo DW (2005) Glacial water mass geometry and the distribution of $\delta^{13}\text{C}$ of ΣCO_2 in the western Atlantic Ocean. *Paleoceanography* 20: PA1017. <http://dx.doi.org/10.1029/2004PA001021>.
- Delaney ML (1989) Uptake of cadmium into calcite shells by planktonic foraminifera. *Chemical Geology* 78: 159–165.
- Duplessy JC, Shackleton NJ, Fairbanks RG, Labeyrie L, Oppo D, and Kallel N (1988) Deepwater source variations during the last climatic cycle and their impact on global deepwater circulation. *Paleoceanography* 3: 343–360.
- Duplessy JC, Shackleton NJ, Matthews RK, et al. (1984) $\delta^{13}\text{C}$ record of benthic foraminifera in the Last Interglacial Ocean: Implications for the carbon cycle and the global deep water circulation. *Quaternary Research* 21: 225–243.
- Elderfield H and Rickaby REM (2000) Oceanic Cd/P ratio and nutrient utilization in the glacial Southern Ocean. *Nature* 405: 305–310.
- Francois R, Altabet MA, Yu E-F, et al. (1997) Contribution of Southern Ocean surface-water stratification to low atmospheric CO₂ concentrations during the last glacial period. *Nature* 389: 929–935.
- Ganeshram RS, Pedersen TF, Calvert SE, and Francois R (2002) Reduced nitrogen fixation in the glacial ocean inferred from changes in marine nitrogen and phosphorous inventories. *Nature* 415: 156–159.
- Hendy IL, Pedersen TF, Kennett JP, and Tada R (2004) Intermittent existence of a southern Californian upwelling cell during submillennial climate change of the last 60 kyr. *Paleoceanography* 19: PA3007. <http://dx.doi.org/10.1029/2003PA000965>.
- Herguera JC, Herbert T, Kashgarian M, and Charles C (2010) Intermediate and deep water mass distribution in the Pacific during the Last Glacial Maximum inferred from oxygen and carbon stable isotopes. *Quaternary Science Reviews* 29: 1228–1245.
- Keigwin LD and Boyle EA (1989) Late quaternary paleochemistry of high-latitude surface waters. *Palaeogeography, Palaeoclimatology, Palaeoecology* 73: 85–106.
- Kroopnick PM (1985) The distribution of ^{13}C of ΣCO_2 in the world oceans. *Deep Sea Research* 32: 57–84.

- Lea DW and Boyle EA (1989) Barium content of benthic foraminifera controlled by bottom-water composition. *Nature* 338: 751–753.
- Lea DW and Boyle EA (1990a) A 210,000-year record of barium variability in the deep northwest Atlantic Ocean. *Nature* 347: 269–272.
- Lea DW and Boyle EA (1990b) Foraminiferal reconstruction of barium distributions in water masses of the glacial oceans. *Paleoceanography* 5: 719–742.
- Lund DC, Adkins JF, and Ferrari R (2011) Abyssal Atlantic circulation during the Last Glacial Maximum: Constraining the ratio between transport and vertical mixing. *Paleoceanography* 26: PA1213. <http://dx.doi.org/10.1029/2010PA001938>.
- Lynch-Stieglitz J and Fairbanks RG (1994) A conservative tracer for glacial ocean circulation from carbon isotope and palaeo-nutrient measurements in benthic foraminifera. *Nature* 369: 41–43.
- Lynch-Stieglitz J, Stocker TF, Broecker WS, and Fairbanks RG (1995) The influence of air-sea exchange on the isotopic composition of oceanic carbon: Observations and modeling. *Global Biogeochemical Cycles* 9: 653–665.
- Mackensen A, Hubberten H-W, Bickert T, Fischer G, and Fütterer DK (1993) $\delta^{13}\text{C}$ in benthic foraminiferal tests of *Fontbotia wuellerstorfi* (Schwager) relative to $\delta^{13}\text{C}$ of dissolved inorganic carbon in Southern Ocean deep water: Implications for Glacial ocean circulation models. *Paleoceanography* 8: 587–610.
- Marchitto TM and Broecker WS (2006) Deep water mass geometry in the glacial Atlantic Ocean: A review of constraints from the paleonutrient proxy Cd/Ca. *Geochemistry, Geophysics, Geosystems* 7: Q12003. <http://dx.doi.org/10.1029/2006GC001323>.
- Marchitto TM, Oppo DW, and Curry WB (2002) Paired benthic foraminiferal Cd/Ca and Zn/Ca evidence for a greatly increased presence of Southern Ocean Water in the glacial North Atlantic. *Paleoceanography* 17: 1038. <http://dx.doi.org/10.1029/2000PA000598>.
- McCorkle DC, Martin PA, Lea DW, and Klinkhammer GP (1995) Evidence of a dissolution effect on benthic foraminiferal shell chemistry: $\delta^{13}\text{C}$, Cd/Ca, Ba/Ca, and Sr/Ca results from the Ontong Java Plateau. *Paleoceanography* 10: 699–714.
- Ninnemann US and Charles CD (2002) Changes in the mode of Southern Ocean circulation over the last glacial cycle revealed by foraminiferal stable isotopic variability. *Earth and Planetary Science Letters* 201: 383–396.
- Olsen A and Ninnemann US (2010) Large $\delta^{13}\text{C}$ gradients in the preindustrial North Atlantic revealed. *Science* 330: 658–659.
- Ostlund HG, Craig H, Broecker WS, and Spencer D (1987) GEOSECS Atlantic, Pacific and Indian Ocean expeditions. *Shorebased Data and Graphics*, vol. 7. Washington, DC: US Government Printing Office.
- Pichevin LE, Ganeshram RS, Francavilla S, Arellano-Torres E, Pedersen TF, and Beaufort L (2010) Interhemispheric leakage of isotopically heavy nitrate in the eastern tropical Pacific during the last glacial period. *Paleoceanography* 25: PA1204. <http://dx.doi.org/10.1029/2009PA001754>.
- Rickaby REM and Elderfield H (1999) Planktonic foraminiferal Cd/Ca: Paleonutrients or paleotemperature? *Paleoceanography* 14: 293–303.
- Robinson RS, Mix A, and Martinez P (2007) Southern Ocean control on the extent of denitrification in the southeast Pacific over the last 70 ka. *Quaternary Science Reviews* 26: 201–212.
- Robinson RS and Sigman DM (2008) Nitrogen isotopic evidence for a poleward decrease in surface nitrate within the ice age Antarctic. *Quaternary Science Reviews* 27: 1076–1090.
- Shackleton NJ, Hall MA, Line J, and Shuxi C (1983) Carbon isotope data in core V19-30 confirm reduced carbon dioxide concentration in the ice age atmosphere. *Nature* 306: 319–322.
- Sigman DM, Altabet MA, McCorkle DC, Francois R, and Fischer G (1999) The $\delta^{15}\text{N}$ of nitrate in the Southern Ocean: Consumption of nitrate in surface waters. *Global Biogeochemical Cycles* 13: 1149–1166.
- Sigman DM, Hain MP, and Haug GH (2010) The polar ocean and glacial cycles in atmospheric CO_2 concentration. *Nature* 466: 47–55.
- Spero HJ (1998) Life history and stable isotope geochemistry of planktonic foraminifera. In: Norris RD and Corfield RM (eds.) *Isotope Paleobiology and Paleoecology*, pp. 7–36. Paleontological Society.
- Spero HJ and Lea DW (2002) The cause of carbon isotope minimum events on glacial terminations. *Science* 296: 522–525.
- Zahn R, Winn K, and Sarnthein M (1986) Benthic foraminiferal $\delta^{13}\text{C}$ and accumulation rates of organic carbon: *Uvigerina peregrina* group and *Cibicides wuellerstorfi*. *Paleoceanography* 1: 27–42.

Oxygen-Isotope Stratigraphy of the Oceans

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Introduction

The Pleistocene, also known as the 'Ice Ages,' starts 2.6 Ma. At the beginning of the nineteenth century, the first geological evidence of major Quaternary glacial episodes came from observations by Louis Agassiz of sediment deposits (moraines) and erosion features attributed to ancient glaciers in the Jura mountains of Western Europe. Up until about 1960, however, most of the Ice Age puzzle remained unexplained. One theory was very promising in explaining the Ice Ages: the astronomical theory of climate developed by a Serbian professor of mathematics, Milutin Milankovitch. Based on his calculations of seasonal and latitudinal distribution of solar energy at the surface of the Earth, [Milankovitch \(1930\)](#) proposed an orbital theory of Ice Ages in which waxing and waning of continental ice sheets are under the control of long-term, orbitally driven changes in summer insolation over the high latitudes of the Northern Hemisphere. One of the most interesting aspects of this theory was that it made it possible to test predictions about climatic records such as how many ice-age deposits geologists would find, and when these deposits had been formed during the past 650 ka. During a few decades, however, the astronomical theory of climate was largely disputed because discussions were based on fragmentary geological records obtained on land and supported by incomplete, and frequently incorrect, radiometric age data. The key to unravel the mystery of the Ice Ages would come from the oceanic seafloor covered by continuous sedimentary sections that had recorded hundreds of thousands of years of the Earth's climatic and oceanographic evolution.

A new branch of Earth's Sciences emerged in the 1950s – paleoceanography. One of its major tools for reconstructing the climate evolution of the Earth is stable oxygen isotopes measured from the remains of marine calcareous organisms. This tool provides clues about past changes in seawater temperature and seawater isotopic composition, which is controlled by changes in waxing and waning of large continental ice sheets and is, therefore, directly linked to glacial–interglacial history of the Earth. This article presents the major steps in our understanding of oxygen-isotope fluctuations in the ocean and their use for stratigraphic purposes.

Oxygen Stable Isotopes in Marine Carbonate to Remains: A Brief Overview

Oxygen Stable Isotopes and the δ Notation

There are three stable isotopes of oxygen in nature: ^{16}O , ^{17}O , and ^{18}O , with relative natural abundances of 99.76, 0.04, and 0.20%, respectively. Because of the higher abundances and the greater mass difference between ^{16}O and ^{18}O , research on oxygen isotopic ratios deals normally with the $^{18}\text{O}/^{16}\text{O}$ ratio. The ratio of stable isotopes of oxygen in carbonates is analyzed by gas mass spectrometric determination of the mass ratios of

carbon dioxide (CO_2) released during reaction of the sample with a strong acid. This ratio is expressed as a difference with reference to a standard carbon dioxide of known composition. This difference in isotope ratios, known as δ -value and given in percentage, is calculated as follows:

$$\delta^{18}\text{O} = \frac{(^{18}\text{O}/^{16}\text{O} \text{ sample} - ^{18}\text{O}/^{16}\text{O} \text{ standard})}{^{18}\text{O}/^{16}\text{O} \text{ standard}} \times 1000 \quad [1]$$

A sample enriched in ^{18}O relative to the standard will show a positive δ -value (with a corresponding negative value for a sample enriched in ^{16}O relative to the standard). The standard commonly used in carbonates is referred to as Pee Dee belemnite (PDB; a cretaceous belemnite from the Pee Dee formation in North Carolina, United States). This standard is not available any longer; however, various international standards have been run against PDB for comparative purposes. Two standards are commonly used and distributed by the National Institute of Standards and Technology in the United States, and the International Atomic Energy Agency in Vienna. They are NBS-18 (carbonate) and NBS-19 (limestone; [Coplen, 1988, 1994](#)).

Carbonate Fractionation: The Pioneering Work of H. Urey and S. Epstein

Harold Urey, a Nobel laureate at the University of Chicago, pioneered the analysis of stable isotopes of oxygen in carbonates in the late 1940s ([Urey, 1947](#)). Urey and his colleagues had found that the carbonate shells of marine organisms from cold water contained a higher proportion of the heavier ^{18}O isotope than did organisms living in warmer water ([Urey et al., 1948](#)). In the early 1950s, pushing this promising approach a step forward, Samuel Epstein and colleagues looked for empirical calibrations of modern mollusk shell oxygen-isotope ratios relative to seawater temperatures. This calibration exercise made it possible to derive equations that could be used to estimate past temperatures from fossilized biologic carbonate remains ([Epstein et al., 1951, 1953](#)).

$$T = 16.9 - 4.2 \times (\delta^{18}\text{O} \text{ sample} - ^{18}\text{O} \text{ water}) + 0.13 \times (\delta^{18}\text{O} \text{ sample} - ^{18}\text{O} \text{ water})^2 \quad [2]$$

As is readily seen from the above formula, the oxygen-isotope composition of marine carbonate samples ($\delta^{18}\text{O}$ sample) is not only related to the temperature of precipitation (T , in $^{\circ}\text{C}$), but also to the isotopic composition of seawater ($\delta^{18}\text{O}$ water). In 1954, Willi Dansgaard had pointed out that evaporation/precipitation processes might act like distilleries and concentrate particular oxygen isotopes during the water cycle ([Figure 1](#)). More particularly, the isotopic fractionation process related to evaporation withdraws from the oceans more of the lighter ^{16}O isotope, leaving the ocean water enriched in the heavier ^{18}O isotope. As a consequence, it was hypothesized that during glaciation periods, when the glaciers

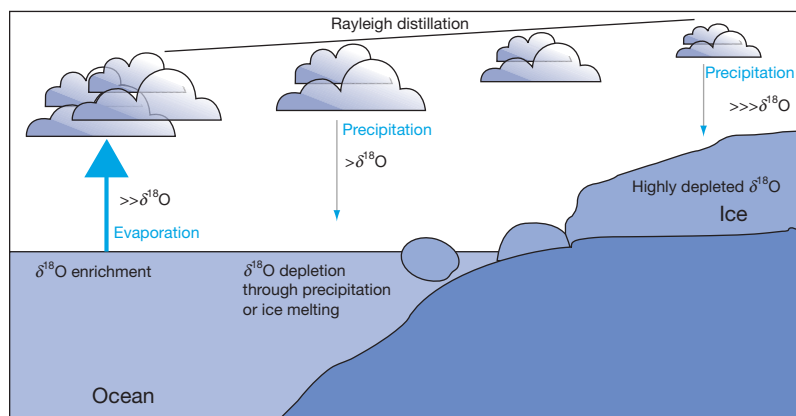


Figure 1 The hydrological cycle and its influences on the oxygen-isotope ratios. Molecules of water containing the lighter ^{16}O isotope have a higher vapor pressure and these molecules are preferentially enriched in the vapor phase, showing a much smaller $\delta^{18}\text{O}$ ($\gg\delta^{18}\text{O}$) than the ocean waters from which they derive. During precipitation processes, fractionation acts in the opposite way than during evaporation, leaving the remaining water vapor even more depleted in ^{18}O in the clouds. This Rayleigh fractionation process in the clouds explains why the water vapor that ultimately precipitates at low temperature to form the ice caps is extremely depleted in ^{18}O ($\gg\gg\delta^{18}\text{O}$) relative to the ocean water.

expanded, light ^{16}O atoms of oxygen were preferentially extracted from the sea and stored in the ice sheets, leaving the seawater enriched in the heavier ^{18}O isotope. When the glaciers melted, the stored isotopes flooded back into the ocean, returning it to its interglacial composition.

Pleistocene Oxygen-Isotope Stratigraphy: Early Works

Cesare Emiliani, the Father of Marine Isotopic Stratigraphy

The early work of Urey and his colleagues had involved studies of the relation between oxygen isotopes and temperature in recent mollusks. In the 1950s, the Italian–American geologist Cesare Emiliani started the measurements of oxygen isotopes from calcite shells of Foraminifera. The Foraminifera are unicellular organisms floating in the water column (planktonic species) or living on the seafloor (benthic species) whose calcite tests accumulate in oceanic sediments after their death. Analyzing tests of planktonic Foraminifera sampled down the length of several marine sediment cores, Emiliani found a periodic variation in the ratio of $^{18}\text{O}/^{16}\text{O}$, and interpreted these oscillations as related to glacial–interglacial changes (Emiliani, 1955; Figure 2(a)). Before Emiliani’s work, it was thought that there had been only four major glaciations during the Pleistocene. Working on sediments sampled with long piston cores, Emiliani showed that there had been many more, with at least eight glaciations during the late Pleistocene (Emiliani, 1966; Figure 2(b)).

Emiliani proposed a way of numbering the marine oxygen isotopic stages (MISs) observed in his deep-sea records. This numbering scheme is still used today. Starting from the most recent interglacial (warm) period – the Holocene (MIS 1) – Emiliani attributed an odd number to light isotopic stages, whereas the glacial (cold) periods, characterized by heavier $\delta^{18}\text{O}$ values, were given an even number (Figure 2). Later on, more subtle isotopic variations, such as the episodes of the isotopic stage 5, were broken into a–e substages (Shackleton, 1969).

Continental Ice-Sheet Waxing/Waning and the Global $\delta^{18}\text{O}$ Signal

As seen above (eqn [2]), it was known that the $\delta^{18}\text{O}$ measured from biologic marine carbonates reflected two major effects: the temperature of the seawater in which the organisms developed and the volume of continental ice sheets (which would change the isotopic composition of seawater). Cooler temperatures and greater ice volumes both result in more positive $^{18}\text{O}/^{16}\text{O}$ ratios (higher $\delta^{18}\text{O}$). Emiliani estimated that 60% of the variations he observed in deep-sea records were due to the temperature effect, 40% were due to the ice effect. Based on this estimation, he concluded that equatorial and tropical ocean surface temperatures had been several degrees cooler during times of glaciation. However, this conclusion was soon to be challenged.

Sir Nicolas Shackleton, from Cambridge University (United Kingdom), compared the $\delta^{18}\text{O}$ measured in the tests of surface-dwelling Foraminifera to oxygen isotopic ratios obtained, in the same core, from benthic Foraminifera that had lived on the seafloor. If, as Cesare Emiliani had estimated, the ratio of oxygen isotopes in marine fossils is chiefly governed by sea temperature at the glacial–interglacial timescale, the benthic Foraminifera would have shown much smaller deviations than the planktonic Foraminifera since the temperature of the water at the bottom of the ocean would have likely remained unchanged, close to freezing. But Shackleton’s results showed that the two isotopic curves were nearly identical, indicating that the temperature was not the main effect at play and that most of the glacial–interglacial isotopic signal was related to changes of the continental ice-sheet volume, which in turn controlled the $^{18}\text{O}/^{16}\text{O}$ ratio in the ocean. Instead of having recorded major temperature changes, Emiliani’s $\delta^{18}\text{O}$ curve had been mostly an isotopic message from the ancient ice sheets.

The most accurate constraints to estimate the change of seawater $\delta^{18}\text{O}$ between the Last Glacial Maximum (LGM) and the Holocene are provided by the measurements of pore water $\delta^{18}\text{O}$ and by high-resolution records of benthic foraminifer

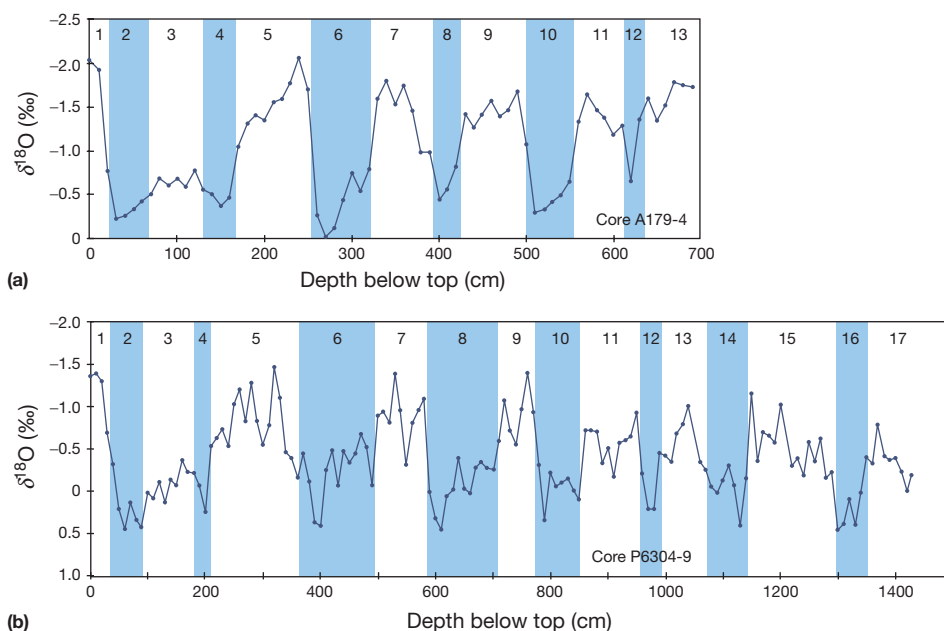


Figure 2 (a) One of the first Upper Pleistocene, marine oxygen-isotope records obtained from planktonic Foraminifera (*Globigerinoides ruber*) picked along Caribbean core 179–4 (Emiliani, 1955). Emiliani converted the $\delta^{18}\text{O}$ values to sea-surface temperatures (SSTs) assuming that 60% of the oxygen-isotope variance was related to temperature changes and only 40% to glacial–interglacial changes in ice volume. (b) A longer planktonic isotopic record obtained on the Foraminifera species *G. sacculifer* picked from Caribbean core P6304–9. This record shows the eight major glaciation stages from the Upper Pleistocene (Emiliani, 1966). Isotopic stages as defined by Emiliani.

$\delta^{18}\text{O}$ in the high latitudes. They show that the $\delta^{18}\text{O}$ of seawater in the deep ocean during the LGM was $1.05 \pm 0.2\%$ heavier than today (see Duplessy et al. (2002) for a review).

Currents and global thermohaline circulation mix the ocean water in such way that any chemical change in one part of the ocean should be reflected everywhere within a thousand years. Thus, rapidly, it became obvious that isotopic variations could be used as a powerful stratigraphic correlation tool. Further efforts were, therefore, devoted to test the global nature of the isotopic signal and to date the major oscillations first observed by Emiliani.

Developing a Timescale for the Oxygen-Isotope Stratigraphy

Absolute Dating (^{14}C and U/Th on Coral Terraces)

Since massive corals can grow only within a shallow water depth, past coral terraces provide an accurate record of former sea-level changes. A geochemist at Lamont Doherty Observatory (New York, United States), Wallace S. Broecker, dated fossil coral reefs in the Florida Keys and the Bahamas, and showed that the sea reached a high level 120 and 80 ka ago, presumably during periods of warm climate when melted ice sheets had released large amounts of water to the sea. These periods of high sea levels closely correspond to warm periods predicted by the astronomical theory of climates of M. Milankovitch.

Broecker and Van Donk (1970), working on a $\delta^{18}\text{O}$ record from a Caribbean core, concluded that the major cycle in the marine oxygen isotopic records was 100 ka in duration. Moreover, they emphasized on the fact that the isotopic records showed a characteristic sawtooth pattern in which periods of

glacial expansion averaging about 100 ka in length were abruptly terminated by rapid deglaciations. They labeled these ‘terminations’ (episodes of rapid deglaciation) using roman numbers (I, II, III, ...), starting with the most recent termination (I) corresponding to the deglaciation located between the LGM (at about 20 ka) and the Holocene.

Magnetostratigraphy and the ‘Rosetta Stone’ for Quaternary Ice Ages

The dating of long marine sedimentary series was greatly improved by using magnetic stratigraphy, a technique first established in the early 1960s. This approach is based on the fact that the Earth’s magnetic field periodically reversed in the past. By looking at the magnetic signal imprinted into rocks or sediments, the geologist can pinpoint reversals that had been previously accurately dated (i.e., K/Ar or Ar/Ar dating applied to volcanic rock samples). Neil Opdyke applied this magnetostratigraphy approach to cores from the Lamont’s core repository. Core V28–238, retrieved in the eastern equatorial Pacific, was particularly promising, with the first distinct magnetic reversal of the core corresponding to the Brunhes/Matuyama (B/M) reversal that had been previously dated on land at 730 ka by K/Ar method.

The ‘Brunhes–Matuyama’ reversal was found to occur just before MIS 19 (Shackleton and Opdyke, 1973; Figure 3). Assuming a constant age–depth model for the core, it was now possible to assign ages to each isotope stage by interpolation. Stage 5e, the penultimate interglacial period equivalent to the Holocene, came out at an age of 123.5 ka. A timescale was now available for the entire period for which Milankovitch had made his calculations.

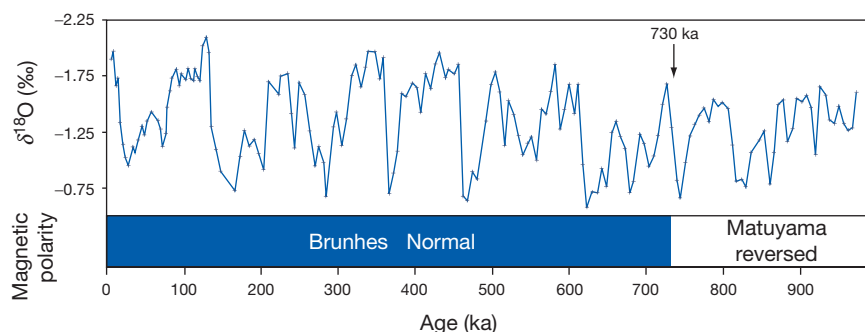


Figure 3 The ‘Rosetta Stone’ of the Ice Ages, oxygen-isotope record from core V28-238 tied into the magnetostratigraphic timescale. This work provided the first accurate chronology of late Pleistocene climate. Modified from Shackleton NJ and Opdyke ND (1973) Oxygen isotope and paleomagnetic stratigraphy of equatorial Pacific core V28-238: Oxygen isotope temperatures and ice volumes on a 105 year and 106 year scale. *Quaternary Research* 3: 39–55.

During the decades between the 1930s and the 1960s, the astronomical theory of climate had been largely disputed, with discussion based on fragmentary geological records supported by incomplete and frequently incorrect radiometric age data. From the 1960s onward, the access to continuous and longer oceanic records revealed many more than the initial four glacial periods recognized from the terrestrial records in Europe, and made it possible to date relatively well these periods. These marine records provided a strong support to the astronomical theory of climate. In 1976, Hays and coauthors published a landmark paper in which they convincingly demonstrated that all the orbital frequencies predicted by the astronomical theory of climate could be found through spectral analysis of climate-sensitive indicators from deep-sea records. One important implication of the astronomical ‘pacemaker’ discovery is that the calculation of past insolation cycle variations could be used as the basis for an astronomically derived timescale.

Astronomical Tuning: The SPECMAP Approach

With the convincing validation of the astronomical theory of climate by Hays et al. (1976), several paleoceanographers proposed the development of a high-precision timescale from deep-sea sedimentary records by using orbitally-related oscillations of paleoclimatic proxies as internal ‘pacemakers’ (i.e., Morley and Hays, 1981). The so-called astronomical (or orbital) ‘tuning’ approach for developing timescales rests on the phase locking of orbitally-related oscillations in sedimentary records to primary orbital oscillations calculated by astronomers, namely, the precession of equinoxes (with periods of 19–23 ka) and the obliquity of the Earth’s axis (with a main period of 41 ka).

To get a global, marine stratigraphic framework on which to develop an Upper Pleistocene astronomical timescale, Imbrie et al. (1984) selected five planktonic Foraminifera oxygen isotopic records from low- and mid-latitudes in which globally persistent isotopic excursions could be graphically correlated. As seen above, over the Ice Age glacial–interglacial cycles, global seawater oxygen isotopic excursions recorded in the Foraminifera calcite chiefly reflect the waxing and waning of continental ice sheets, enriched in the light ^{16}O isotope relative to seawater.

An initial, radiometrically controlled age model was developed for each of the five isotopic records, using few stratigraphic levels: core tops (with a zero age); two isotopic events in stage 2 with ^{14}C ages of 18 and 21 ka; the MIS6–MIS5 isotopic stage boundary (termination II), taken at 127 ka; and the Brunhes–Matuyama magnetic reversal (at the base of MIS19), taken at 730 ka. The astronomical tuning procedure applied by Imbrie et al. (1984) iteratively modifies the initial depth–age model to increase the coherency and phase lock between filtered 19–23 ka and 41 ka oscillations of the oxygen isotopic records, on the one hand, and the precession and obliquity components of Earth’s orbital variations, on the other hand. Such an orbital-tuning approach assumes constant phase lags at the primary Milankovitch frequencies between the marine oxygen isotopic variations and Earth’s orbital changes. Those phase lags had been determined in an earlier paper (Imbrie and Imbrie, 1980) by adjusting the parametrization of a nonlinear (exponential) climatic model predicting ice-volume variations so that the model output would match an 130-ka-long isotopic record, radiometrically dated.

The five astronomically dated isotopic records were finally normalized and stacked, and the resulting curve smoothed with a nine-point Gaussian filter to produce a reference stratigraphic record (Figure 4(a)). The rationale for the stacking approach is that global isotopic changes related to ice-sheet waxing/waning would be enhanced in the resulting composite curve, whereas local or regional isotopic variations would tend to cancel out. Stacking is also useful because it results in a set of well-defined isotopic oscillations that can, in principle, be found in all sediment cores. Figure 5 shows the comparison of filtered precession- and obliquity-related oscillations from the stacked, planktonic isotopic record with obliquity (upper panel) and precession (lower panel) orbital oscillations.

Imbrie and coworkers estimated their geological timescale for the Upper Pleistocene to be accurate to within 5 ka (Imbrie et al., 1984). Since its publication, this orbitally aged isotopic stacked record has been used in hundreds of paleoceanographic studies as a reference curve for developing age models and tying marine sedimentary records into a common stratigraphic framework.

Three years after the landmark paper by Imbrie et al. (1984), Martinson et al. developed an orbital timescale for the last 300 ka (Martinson et al., 1987) that they transferred

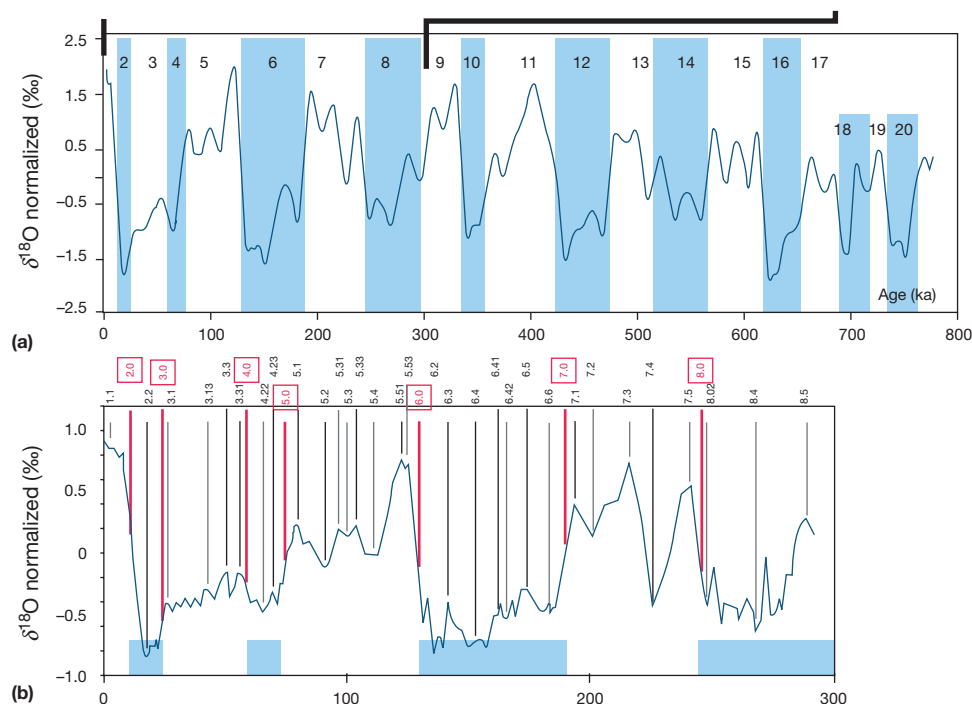


Figure 4 (a) The reference oxygen-isotope record established by the SPECMAP group (Spectral Mapping) by stacking five planktonic isotopic records. The age model was developed by phase-locking oxygen-isotope oscillations to orbital-forcing functions (Imbrie et al., 1984). (b) The detailed oxygen-isotope record from the last 300 ka obtained by stacking high-resolution benthic records (Pisias et al., 1984) and tied to orbital-forcing functions (Martinson et al., 1987). Isotopic stages are labeled according to the Pisias et al. (1984) modification of Emiliani's scheme.

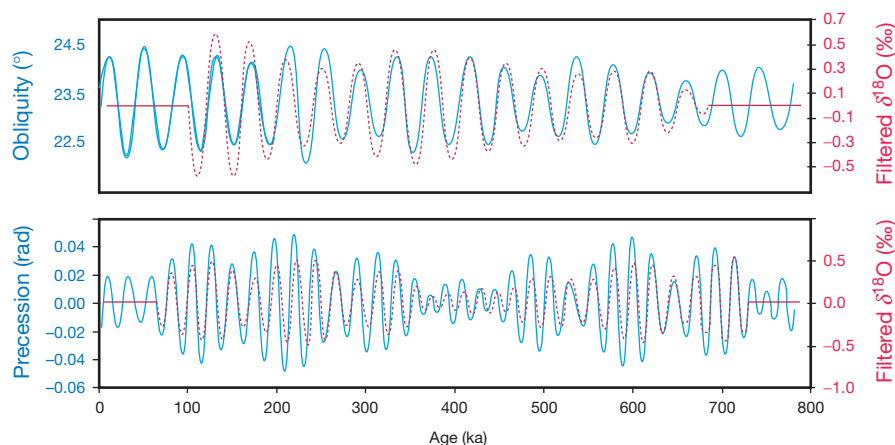


Figure 5 In solid lines, variations in obliquity (upper panel) and precession (lower panel), and the corresponding frequency components extracted by band filtering the SPECMAP δ¹⁸O stack record (dashed line) over the past 780 ka. Reproduced from Imbrie J, Hays JD, Martinson DG, et al. (1984). The orbital theory of Pleistocene climate: Support from a revised chronology of the marine delta ¹⁸O record. In: Berger A et al. (eds.) *Milankovitch and Climate*, pp. 269–305. Dordrecht: Plenum Reidel.

to the stacked, benthic oxygen-isotope stratigraphy from Pisias et al. (1984; Figure 4(b)). Martinson et al. followed four different tuning strategies. This made it possible to test the robustness and accuracy of astronomical tuning for developing an Upper Pleistocene timescale. The error measured by the standard deviation about the mean of their four chronologies has an average magnitude of 2.5 ka. Transferring the final chronology to the stacked isotopic record of Pisias et al. (1984) leads to additional errors. The authors concluded this

final chronology has an average error of ±5 ka. Comparison of the younger portion of the Imbrie et al.'s chronology and Martinson et al.'s 300-ka chronology shows that ages of isotopic events agree to within the error bars (Figure 4).

The 300,000-year-long stacked δ¹⁸O record produced by Martinson et al. (1987) has been the preferred target record for upper Pleistocene benthic correlations since its publication. Having a higher temporal resolution, it also contains additional, short events of global significance that were not seen

in the SPECMAP planktonic $\delta^{18}\text{O}$ record. Beyond the last 300 000 ka, the longer, stacked benthic $\delta^{18}\text{O}$ record produced by Lisiecki and Raymo (2005) is now a widely used target for isotopic stratigraphy (see later).

Oxygen-Isotope Stratigraphy and the Imprint of the Primary Orbital Oscillations

The Deep Sea Drilling Project and the subsequent Ocean Drilling Program provided long, deep-sea sedimentary records that made it possible to look at climatic oscillations over the entire Pleistocene and beyond. These long records made it immediately obvious that the 100-ka oscillations only dominate the climate of the Earth over the last 600 ka, whereas the early Pleistocene variability is dominated by 41 ka, obliquity-related oscillations (i.e., Ruddiman et al., 1986). The exact timing and mechanisms of this shift between dominant periodicities of the climate system are still a matter of debate. Some studies have concluded that the transition from 41 to 100 ka dominant oscillations was gradual, taking place approximately in the interval 1.0–1.2 to 0.6 Ma (Berger et al., 1994; Imbrie et al., 1993; Ruddiman et al., 1989). However, other studies favored a rapid transition at about 0.8–0.9 Ma (Bassinot et al., 1997; Pisias and Moore, 1981; Shackleton and Opdyke, 1976). Statistical detection of the mid-Pleistocene transition in mean and variance of Pleistocene $\delta^{18}\text{O}$ records indicates that Earth's climate system may have experienced an abrupt change around 0.9 Ma (the so-called mid-Pleistocene revolution) that was characterized by an increase in ice mass and a global cooling (Maasch, 1988; Mudelsee and Stattegger, 1997).

The dominance of the 100-ka oscillation in the Upper Pleistocene is a major puzzle of the astronomical theory of climates since eccentricity variations of the Earth's orbit lead to only minor changes in the insolation budget (see review of this problem in Imbrie et al., 1993). Numerous authors have proposed that this 100-ka oscillation may result from non-linear response of the climate system that would transfer power from the envelope of the precession oscillations (i.e., Imbrie et al., 1993). Numerical modeling has suggested that the existence of large Northern Hemisphere ice sheets is an essential condition for developing the sawtooth 100-ka oscillations of the late Pleistocene (i.e., Imbrie et al., 1993; Le Treut and Ghil, 1983; Saltzman, 1987).

State of the Art and Limits of the Marine Oxygen-Isotope Stratigraphy

Improvement in the Marine Oxygen-Isotope Stratigraphy

Orbital tuning provides a global reference standard for dating long, continuous sediment records. However, the approach is not without its problems. Subsequent revisions of the 0–780 ka SPECMAP timescale showed, for instance, that several oscillations predicted by the astronomical theory of climate (in isotopic stages 17 and 18) were apparently missing in the stacked record (Bassinot et al., 1994; Shackleton et al., 1990). Their absence is related to (1) the incompleteness of some of the isotopic records that reached those stages and to (2) an erroneous Brunhes–Matuyama magnetic reversal age assignment in the age model developed at the first step of the

Imbrie et al. (1984) tuning approach. $^{40}\text{Ar}/^{39}\text{Ar}$ radiometric dating of the Brunhes–Matuyama magnetic boundary gives an age of about 780 ka (Baksi et al., 1992), significantly older than the K/Ar age (730 ka) available in the early 1980s, and used by Imbrie et al. (1984) to constrain their initial age model. Although the time constraint at the B/M boundary was removed midway in Imbrie et al.'s orbital-tuning approach, it is quite clear that the final age model suffered from this erroneous tie point.

Applying astronomical tuning strategies with a higher degree of freedom (i.e., no initial age assignment to the B/M boundary) on carefully selected, high-resolution isotopic records, Shackleton et al. (1990) and Bassinot et al. (1994) obtained astronomical ages for the Brunhes–Matuyama magnetic reversal that are in good agreement with $^{40}\text{Ar}/^{39}\text{Ar}$ radiometric ages (Figure 6). The Shackleton et al. (1990) record extends back to 2.5 Ma, and all the magnetic reversal age assignments are in good accordance with $^{40}\text{Ar}/^{39}\text{Ar}$ radiometric ages. These examples illustrate the fact that theoretical precision of astronomical timescales (usually a few thousand years) may actually mask potential, larger uncertainties related to the tuning strategy and/or to incompleteness of paleoclimatic records used to provide the stratigraphic framework on which the astronomical tuning is conducted. This implies that great care should be taken in selecting isotopic records used as references.

Developing a Global, Reference Curve

In order to build a globally meaningful stratigraphic reference curve, a time-consuming and robust approach is to use as many good and coherent oxygen-isotope records as possible; this is the approach followed by Lisiecki and Raymo (2005). These authors recently developed a 5.3 My stack (the 'LR04' stack) of benthic $\delta^{18}\text{O}$ records from 57 globally distributed sites aligned by an automated graphic correlation algorithm (Figure 7). They developed an age model for the Pliocene–Pleistocene, obtained from tuning the $\delta^{18}\text{O}$ stack to a simple ice model based on 21 June insolation at 65°N (Imbrie and Imbrie, 1980). Including all sources of errors (i.e., uncertainty in the orbital solution, number of records covering the oldest interval analyzed, assumed response time of the ice sheets), the authors estimated the uncertainty in the LR04 age model to be ± 6 ka from 3 to 1 Ma (Lower Pleistocene) and ± 4 ka from 1 to 0 Ma (Upper Pleistocene). The LR04 age model, designed to minimize changes in global sedimentation rates, generally supports the SPECMAP timescale back to 625 ka. As to date, the LR04 record is one of the most widely used reference for benthic oxygen isotopic stratigraphy of the Pliocene–Pleistocene.

Limits of Marine Oxygen-Isotope Stratigraphy

The accuracy of oxygen-isotope stratigraphy and its use in paleoceanographic studies rest on the basic assumption that global $\delta^{18}\text{O}$ variations recorded in the remains of marine organisms are mainly related to ice-volume changes and are, therefore, synchronous to within the mixing rate of the ocean (~ 1000 years). However, for detailed and accurate correlations at major climatic terminations, this set of assumption has its

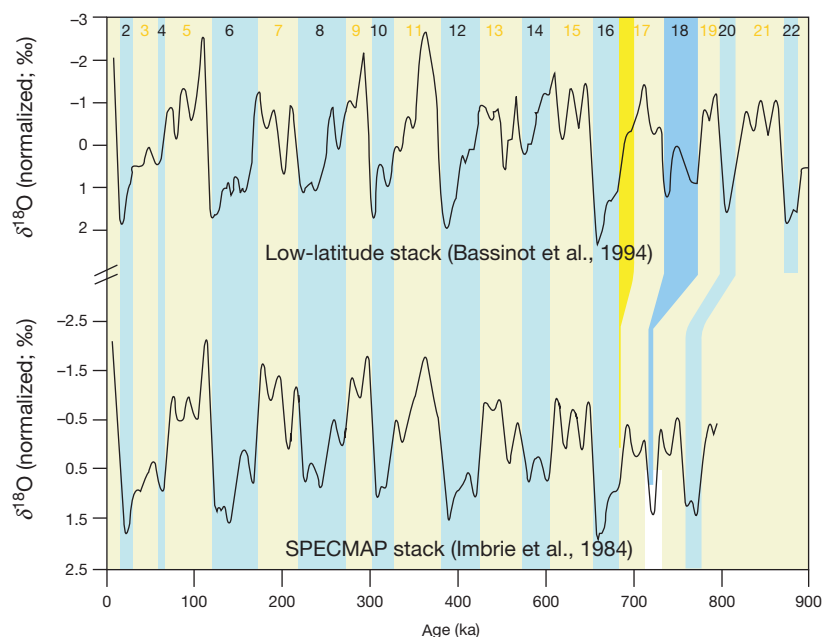


Figure 6 Lower panel, the SPECMAP isotopic record (Imbrie et al., 1984). Upper panel, the low-latitude stack (Bassinot et al., 1994). The yellow and dark blue connections indicate the missing cycles from the SPECMAP stack that lead to a revision of the orbital solution in the astronomical timescale of the low-latitude stack. This new tuning solution led to a revised age of the Brunhes–Matuyama reversal (found at the base of isotopic stage 19), in good agreement with Ar/Ar dates.

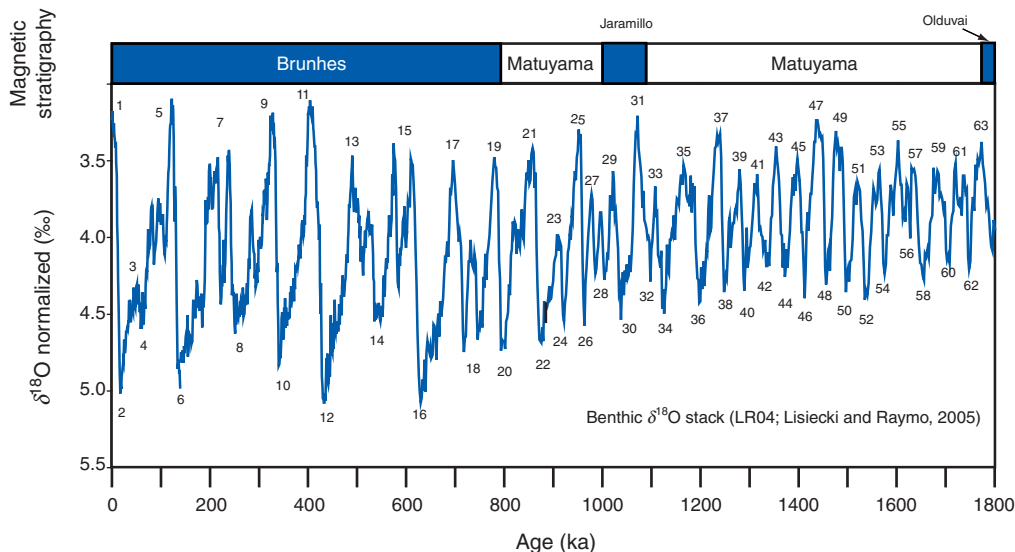


Figure 7 The isotopic curve LR04 obtained by stacking 57 global benthic isotopic records. Age model was obtained through orbital tuning (Lisiecki and Raymo, 2005). To this date, this record is one of the most widely used for isotopic stratigraphy of the Plio–Pleistocene.

limits as shown recently by the radiocarbon dating of two high-resolution records, one from the Iberian margin (Atlantic) and the other from the eastern equatorial Pacific. This study revealed that the benthic $\delta^{18}\text{O}$ signal at the Pacific site lagged the Atlantic site by 3.9 ka across the last termination, a difference which is not explained by ocean mixing alone (Skinner and Shackleton, 2005). This lag has been chiefly attributed to the temperature-related part of the benthic Foraminifera $\delta^{18}\text{O}$ records, with the temperature increase in the deep Pacific being

significantly delayed relative to the ice-related $\delta^{18}\text{O}$ change of seawater across the last deglaciation.

Beyond the time interval that can be precisely radiochronologically dated, over the last six terminations, the analysis of sedimentation rates derived from the alignment of benthic $\delta^{18}\text{O}$ confirmed that there is a statistically significant Atlantic lead over the Pacific benthic $\delta^{18}\text{O}$ change that exceeds the delay expected from ocean mixing rates (Lisiecki and Raimo, 2009). To circumvent part of this problem and provide

paleoceanographers with more accurate stratigraphic targets across terminations, Lisiecki and Raimo (2009) produced two benthic stacked records, one for the Atlantic Ocean and the other for the Pacific Ocean. But diachronisms are not only restricted to interbasin correlations. Studies based on well-dated Atlantic, Indian, and Pacific sediments across the last termination suggest that, although some $\delta^{18}\text{O}$ deglacial changes are rapidly transferred to intermediate water depths, they can take an additional ~ 1500 years to reach the deep waters of the same basin (Labeyrie et al., 2005; Waelbroeck et al., 2006).

All the above results remind us that, while most of the $\delta^{18}\text{O}$ signal recorded in benthic carbonates is ice volume related and globally distributed to within the mixing rate of the ocean, complex circulation schemes and hydrological changes such as temperature modifications can lead to inter- and intra-basin diachronisms of up to a few thousand years across terminations. This should be taken into account for detailed stratigraphic correlations, and should be also carefully kept in mind when using benthic $\delta^{18}\text{O}$ records as a proxy for the timing of ice volume changes. This includes the studies that compare the phases of different proxies to $\delta^{18}\text{O}$ changes in a single sediment core.

See also: K/Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ Dating. **Glaciation, Causes: Astronomical Theory of Paleoclimates. Quaternary Stratigraphy: Overview.**

References

- Baksi AK, Hsu V, McWilliams MO, et al. (1992) $^{40}\text{Ar}/^{39}\text{Ar}$ dating of the Brunhes–Matuyama geomagnetic field reversal. *Science* 256: 356–357.
- Bassinot FC, Beaufort L, Vincent E, et al. (1997) Changes in the dynamics of western equatorial Atlantic surface currents and biogenic productivity at the ‘mid-Pleistocene revolution’ (930 ka). *Proceedings of the Ocean Drilling Program, Scientific Results* 154: 269–284.
- Bassinot FC, Labeyrie LD, Vincent E, et al. (1994) The astronomical theory of climate and the age of the Brunhes–Matuyama magnetic reversal. *Earth and Planetary Science Letters* 126: 91–108.
- Berger WH, Yasuda M, Bickert T, et al. (1994) Quaternary time scale for the Ontong Java Plateau: Milankovitch template for Ocean Drilling Program Site 806. *Geology* 180: 105–116.
- Broecker WS and Van Donk J (1970) Insolation changes, ice volumes, and the ^{18}O in deep-sea cores. *Reviews of Geophysics and Space Physics* 8: 169–198.
- Coplen TB (1988) Normalization of oxygen and hydrogen isotope data. *Chemical Geology: Isotope Geoscience Section* 72: 293–297.
- Coplen TB (1994) Reporting of stable hydrogen, carbon, oxygen isotopic abundances. *Pure and Applied Chemistry* 66: 273–276.
- Dansgaard W (1954) The O^{18} -abundance in fresh water. *Geochimica et Cosmochimica Acta* 6: 241–260.
- Duplessy JC, Labeyrie L, and Waelbroeck C (2002) Constraints on the ocean oxygen isotopic enrichment between the Last Glacial Maximum and the Holocene: Paleocceanographic implications. *Quaternary Science Reviews* 21: 315–330.
- Emiliani C (1955) Pleistocene temperatures. *Journal of Geology* 63: 538–578.
- Emiliani C (1966) Paleotemperature analysis of Caribbean cores P6304–8 and P6304–9 and a generalized temperature curve for the past 425,000 years. *Journal of Geology* 74: 109–126.
- Epstein S, Buchsbaum R, Lowenstam H, and Urey HC (1951) Carbonate–water temperature scale. *Bulletin of the Geological Society of America* 62: 417–426.
- Epstein S, Buchsbaum R, Lowenstam HA, and Urey HC (1953) Revised carbonate–water isotopic temperature scale. *Bulletin of the Geological Society of America* 64: 1315–1326.
- Hays JD, Imbrie J, and Shackleton NJ (1976) Variations in the Earth’s orbit: Pacemaker of the Ice Ages. *Science* 194: 1121–1132.
- Imbrie J, Berger A, Boyle EA, et al. (1993) On the structure and origin of major glaciation cycles 2: The 100,000-year cycle. *Paleoceanography* 8: 699–735.
- Imbrie J, Hays JD, Martinson DG, et al. (1984) The orbital theory of Pleistocene climate: Support from a revised chronology of the marine delta ^{18}O record. In: Berger A, et al. (ed.) *Milankovitch and Climate*, pp. 269–305. Dordrecht: Plenum Reidel.
- Imbrie J and Imbrie Z (1980) Modeling the climatic response to orbital variations. *Science* 207: 943–953.
- Labeyrie L, Waelbroeck C, Cortijo E, Michel E, and Duplessy J-C (2005) Changes in deep water hydrology during the Last Deglaciation. *Comptes Rendus Geoscience* 337: 919–927.
- Le Treut H and Ghil M (1983) Orbital forcing, climatic interactions, and glaciation cycles. *Journal of Geophysical Research* 88: 5167–5190.
- Lisiecki LE and Raimo ME (2009) Diachronous benthic $\delta^{18}\text{O}$ responses during late Pleistocene terminations. *Paleoceanography* 24: PA3210. <http://dx.doi.org/10.1029/2009PA001732>.
- Lisiecki LE and Raymo ME (2005) A Pliocene–Pleistocene stack of 57 globally distributed benthic $\delta^{18}\text{O}$ records. *Paleoceanography* 20: PA1003. <http://dx.doi.org/10.1029/2004PA001071>.
- Maasch KA (1988) Statistical detection of the mid-Pleistocene transition. *Climate Dynamics* 2: 133–143.
- Martinson DG, Pisias NG, Hays JD, Imbrie J, Moore TC Jr., and Shackleton NJ (1987) Age dating and the orbital theory of the Ice Ages: Development of a high-resolution 0–300,000 year chronostratigraphy. *Quaternary Research* 27: 1–29.
- Milankovitch M (1930) *Mathematische klimalehre und astronomische theorie der klimaschwankungen*, p. 176. Berlin: Gebruder Borntraeger.
- Morley JJ and Hays JD (1981) Towards a high-resolution, global, deep-sea chronology for the last 750,000 years. *Earth and Planetary Science Letters* 53: 279–295.
- Mudelsee M and Stattegger K (1997) Exploring the structure of the mid-Pleistocene revolution with advance methods of time-series analysis. *Geologische Rundschau* 86: 499–511.
- Pisias NG, Martinson DG, Moore TC, et al. (1984) High resolution stratigraphic correlation of benthic oxygen isotopic records spanning the last 300,000 years. *Marine Geology* 56: 119–136.
- Pisias NG and Moore TC (1981) The evolution of Pleistocene climate: A time series approach. *Earth and Planetary Science Letters* 52: 450–456.
- Ruddiman WF, Raymo ME, Martinson DG, Clement BM, and Backman J (1989) Pleistocene evolution: Northern hemisphere ice sheets and North Atlantic Ocean. *Paleoceanography* 4(4): 353–412.
- Ruddiman WF, Raymo M, and McIntyre A (1986) Matuyama 41,000-year cycles: North Atlantic Ocean and northern hemisphere ice sheets. *Earth and Planetary Science Letters* 80: 117–129.
- Saltzman B (1987) Carbon dioxide and the $\delta^{18}\text{O}$ record of late Quaternary climatic change: A global model. *Climate Dynamics* 1: 77–85.
- Shackleton NJ (1967) Oxygen isotope analyses and Pleistocene temperatures re-assessed. *Nature* 215: 15–17.
- Shackleton NJ (1969) The last interglacial in the marine and terrestrial records. *Proceedings of the Royal Society B* 174: 135–154.
- Shackleton NJ, Berger A, and Peltier WA (1990) An alternative astronomical calibration of the lower Pleistocene timescale based on ODP Site 677. *Transactions of the Royal Society of Edinburgh: Earth Sciences* 81: 251–261.
- Shackleton NJ and Opdyke ND (1973) Oxygen isotope and paleomagnetic stratigraphy of equatorial Pacific core V28–238: Oxygen isotope temperatures and ice volumes on a 105 year and 106 year scale. *Quaternary Research* 3: 39–55.
- Shackleton NJ and Opdyke ND (1976) Oxygen-isotope and paleomagnetic stratigraphy of Pacific core V28–239: Late Pliocene to latest Pleistocene. In: Cline RM and Hays JD (eds.) *Investigations of Late Quaternary Paleoceanography and Paleoclimatology. Memoir – Geological Society of America*, vol. 145, pp. 449–464. London: Geological Society of America.
- Skinner LC and Shackleton NJ (2005) An Atlantic lead over Pacific deep-water change across termination I: Implications for the application of the marine isotope stage stratigraphy. *Quaternary Science Reviews* 24: 571–580.
- Urey HC (1947) The thermodynamic properties of isotopic substances. *Journal of the Chemical Society (London)* 562–581.
- Urey HC, Epstein S, McKinney C, and McCrea J (1948) Method for measurement of paleotemperatures. *Bulletin of the Geological Society of America* 59: 1359–1360.
- Waelbroeck C, Levi C, Duplessy J-C, et al. (2006) Distant origin of circulation changes in the Indian Ocean during the last deglaciation. *Earth and Planetary Science Letters* 243: 244–251.

Oxygen Isotope Composition of Seawater

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Introduction

This article deals with processes that affect the ratio of the two most common stable isotopes of oxygen in seawater and explores ways in which this has changed during the Quaternary. Both text and figures draw heavily on the water-based oxygen isotope part of a previously published essay on stable oxygen and carbon isotopes in Foraminifera (Rohling and Cooke, 1999). In contrast to the current text, which offers only the fundamental key references, the essay of Rohling and Cooke (1999) contains a very extensive set of references, and would therefore be a useful first step for further and more specialized reading.

Natural Abundance, Measurement, and Physicochemical Behavior

Oxygen exists in nature in the form of three stable isotopes: ^{16}O , ^{17}O , and ^{18}O . Their relative natural abundances are 99.76, 0.04, and 0.20%, respectively. Because of the higher abundances and the greater mass difference between ^{16}O and ^{18}O , research on oxygen isotope ratios normally concerns $^{18}\text{O}/^{16}\text{O}$ ratios.

The word isotope (Greek, meaning 'equal places') implies that the various isotopes occupy the same position in the Periodic Table. The difference between the atomic masses of the isotopic species for each element consists of a different number of neutrons in the nucleus, but isotopes have the same number of protons. The oxygen isotope $^{16}_8\text{O}$ contains 8 protons and 8 neutrons, giving it an atomic mass of 16. The isotope $^{18}_8\text{O}$ contains 8 protons and 10 neutrons, giving it an atomic mass of 18.

While the absolute abundances of minor isotopes (such as ^{18}O) cannot be determined accurately, it is still possible to get quantitative results by comparing the result given for a known external standard with that for the unknown sample. These differences in isotope ratios, known as δ values, are defined as:

$$\delta_{\text{sam}} = \frac{R_{\text{sam}} - R_{\text{std}}}{R_{\text{std}}} \times 1000$$

Here, sam is the sample value and std is the standard or reference value. These variations in composition are given in delta (δ) notation and are reported in parts per thousand (per mille, ‰). R stands for the heavy/light ratio between the abundances of any two isotopes (e.g., $^{18}\text{O}/^{16}\text{O}$). A positive δ value indicates enrichment in the heavy isotope, relative to the standard, and conversely, depletion is shown by a negative δ value. The standard used today in the analysis of $\delta^{18}\text{O}$ in water is the Vienna Standard Mean Ocean Water (VSMOW).

All isotopes of a given element contain the same number and arrangement of electrons, and so broadly display similarity in their chemical behavior. However, the mass differences of

different isotopes impose subtle differences in their physicochemical properties. The mass differences are particularly important in light elements (low numbers in the Periodic Table). Molecules vibrate with a fundamental frequency which depends on the mass of the isotopes from which they are composed. The resultant differences in dissociation energy of the light and heavy isotopes imply that bonds formed by light isotopes are weaker than those formed by heavy isotopes. Hence, as a rule of thumb, molecules comprised of the light isotopes react somewhat more easily than those comprised of the heavy isotopes.

The partitioning of isotopes between substances with different isotopic compositions is called 'fractionation.' The fractionation factor (α), which quantifies isotopic fractionation between two substances A and B, is defined as $\alpha = R_A/R_B$. Here, R_A and R_B are the heavy/light ratios between the abundances of any two isotopes (e.g., $^{18}\text{O}/^{16}\text{O}$) in the exchanging chemical compounds A and B, respectively. Fractionation mainly results from: (1) isotope exchange reactions and (2) kinetic effects. Isotope exchange reactions concern partitioning of isotopes between phases that are in equilibrium, and are therefore also known as 'equilibrium isotope fractionation' processes. Processes of equilibrium fractionation are essentially temperature dependent. Kinetic effects cause deviations from the simple equilibrium processes due to different rates of reaction for the various isotopic species (due directly to vibration differences, or indirectly through differences in bonding energies). Important kinetic effects are associated with diffusion.

Processes Controlling Oxygen Isotope Ratios in Seawater

The oxygen isotope ratio of seawater is intimately linked with fractionation processes within the hydrological cycle. Schematically, this cycle is comprised of evaporation, atmospheric vapor transport, precipitation, and subsequent return of freshwater to the ocean (directly via precipitation and via runoff or iceberg melting). Long-term storage of freshwater in aquifers and especially ice sheets is also important for seawater isotope ratios (Figure 1). Formation and melting of seasonal sea ice imposes strong local variability. Finally, the spatial distribution of oxygen isotopes in the world's oceans depends on processes of advection and mixing of water masses from different source regions with different isotopic signatures. The various influences are discussed below, but it should be noted that this forms only a basic introduction, and specialist texts and reviews should be consulted before advanced applications. Key texts and overviews can be found in Craig and Gordon (1965), Dansgaard (1964), Garlick (1974), Gonfiantini (1986), Hoefs (1997), Hoffmann and Heimann (1997), Joussame and Jouzel (1993), Jouzel et al. (1975), Majoube (1971), Merlivat

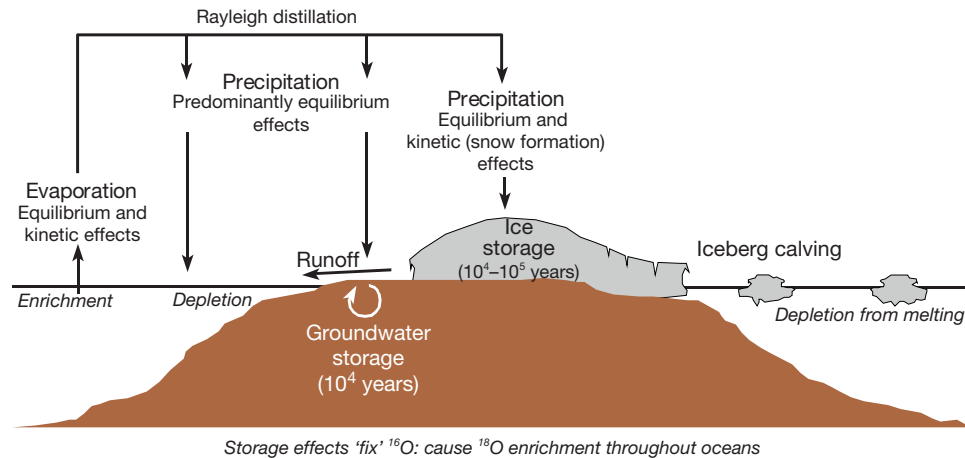


Figure 1 Schematic presentation of the hydrological cycle influences on oxygen isotope ratios. Effects on seawater are described in italics (reproduced from Rohling EJ and Cooke S (1999) Stable oxygen and carbon isotope ratios in foraminiferal carbonate. In: Sen Gupta BK (ed.) *Modern Foraminifera*, pp. 239–258. Dordrecht, The Netherlands: Kluwer Academic). The ‘fix’ comment refers to the storage of preferentially ^{16}O -enriched precipitation in ice sheets and groundwater, which constitutes a preferential removal of ^{16}O from the oceans and thus a relative ^{18}O -enrichment in the oceans.

and Jouzel (1979), Rohling and Cooke (1999), Rozanski et al. (1982, 1993), and Stewart (1975).

Evaporation

The isotopic exchange at the sea–air interface is given by:



Molecules composed of lighter isotopes have higher vapor pressures and the lighter molecular species are therefore preferentially enriched in the vapor phase. The fractionation factor for the equilibrium exchange is $\alpha_{l-v} = [^{18}\text{O}/^{16}\text{O}]_l / [^{18}\text{O}/^{16}\text{O}]_v$. The most commonly used relationship between α_{l-v} and temperature during evaporation is that given by Majoube (1971):

$$\alpha_{l-v} = \exp \{ (1.137 \text{ T}^{-2}) \times 10^3 - (0.4156 \text{ T}^{-1}) - 2.0667 \times 10^{-3} \}$$

where T is in Kelvin. This relationship illustrates a decrease in fractionation with increasing temperature. The fractionation causes a difference between $\delta^{18}\text{O}$ of seawater and $\delta^{18}\text{O}$ of vapor evaporated from that seawater equal to $\delta^{18}\text{O}_l - \delta^{18}\text{O}_v = 10^3 \ln(\alpha) \text{‰}$. At a temperature of 20 °C this difference amounts to about 9.8‰. The equilibrium enrichment factor ε is defined as $\varepsilon = \alpha - 1$, and is often reported as a ‰ value. For $\alpha = 1.010$, $\varepsilon = 10\text{‰}$.

Craig and Gordon (1965) found that, in addition to equilibrium fractionation, further fractionation should have occurred to explain their observations of stronger depletions in vapor than expected from equilibrium fractionation only. This further depletion is thought to be due to kinetic effects during molecular diffusion within the boundary layer between the water–air interface and the fully turbulent region (where no further fractionation occurs). Within that boundary layer, diffusion predominates because of slow atmospheric transport. The kinetic enrichment factor ($\Delta\varepsilon$) for oxygen isotopes depends on the relative air humidity (h) in the turbulent region outside the boundary layer, giving a relationship for most natural circumstances of $\Delta\varepsilon = 14.2 (1 - h) \text{‰}$, where h is presented as a fraction (i.e., 0.7 means 70% relative humidity).

Since diffusion rates depend on property gradients, the final isotopic composition of newly evaporated water also depends on the isotopic composition ($\delta^{18}\text{O}_{\text{atm}}$) of vapor already present in the turbulent region of the atmosphere, and on $\delta^{18}\text{O}$ of the surface water (Gonfiantini, 1986):

$$\delta^{18}\text{O}_E = \frac{1}{1 - h + \Delta\varepsilon} \left(\frac{\delta^{18}\text{O}_{\text{Sea surface}} - \varepsilon}{\alpha} - h\delta^{18}\text{O}_{\text{atm}} - \Delta\varepsilon \right)$$

Here, the $\delta^{18}\text{O}$, $\Delta\varepsilon$, and ε values, normally reported in ‰, are to be used in true form, that is, value $\times 10^{-3}$. It will be obvious that the preferential uptake of the lighter isotope during evaporation causes a shift to heavier $\delta^{18}\text{O}$ values in the remaining surface waters.

Precipitation and Atmospheric Vapor Transport

Fractionation processes during the formation of droplets are basically the same as during evaporation, but work in the opposite direction. Droplets are normally near equilibrium with atmospheric vapor, due to an absence of significant kinetic fractionation during condensation when relative humidity is 100%. The equilibrium fractionation factor for condensation is the same as that for evaporation and may therefore be determined with Majoube’s (1971) equation for the appropriate temperature, and then used in $\delta^{18}\text{O}_l - \delta^{18}\text{O}_v = 10^3 \ln(\alpha) \text{‰}$ to determine the isotopic composition of the droplets for any given atmospheric vapor composition.

Since condensation preferentially removes the heavier isotope ^{18}O , the $\delta^{18}\text{O}$ of the remaining atmospheric vapor becomes progressively more depleted. Relative to an original atmospheric vapor composition that is given by the composition of evaporated water ($\delta^{18}\text{O}_E$), the isotopic composition of the atmospheric vapor (AV) is approximated by $\delta^{18}\text{O}_{\text{AV}} = \delta^{18}\text{O}_E + 10^3(\alpha - 1) \ln f$ (Dansgaard, 1964), where α is the fractionation factor at condensation temperature and f is the fraction of atmospheric vapor remaining after rain-out (e.g., 0.7 if 70% of the evaporated water remains as atmospheric vapor due to 30% rain-out). The isotopic composition

of precipitation formed in equilibrium with the atmospheric vapor is found according to $\delta^{18}\text{O}_\text{P} - \delta^{18}\text{O}_\text{AV} = 10^3 \ln \alpha$. From these relationships, which represent a basic Rayleigh distillation process, it is immediately obvious that the first precipitation will be of a similar isotopic composition as the original seawater, and that a longer pathway from the source region (and consequently more rain-out) causes atmospheric vapor to become more and more depleted. New precipitation from this vapor reflects this depletion (Figure 2). As a result, precipitation is significantly more depleted at high latitudes than in the tropics (reaching values as low as -57‰ in Antarctica).

The magnitude of rain-out is strongly temperature dependent. As atmospheric vapor is transported toward colder regions, it experiences successive condensations. This effect – rather than temperature dependence of the fractionation factor – determines a basic quasilinear relationship of $0.69\text{‰}^\circ\text{C}^{-1}$ between $\delta^{18}\text{O}$ of precipitation and temperature, as has been observed for temperatures between -40 and $+15^\circ\text{C}$. Above 15°C , this relationship with temperature breaks down. There, mainly in areas with strong convection in the atmosphere, the so-called amount effect dominates. This effect approximately determines 1.5‰ depletion in the $\delta^{18}\text{O}$ of precipitation for every 100 mm increase in rainfall.

The changes in oxygen-isotope composition of atmospheric vapor and precipitation affect surface waters in the world's oceans through addition of freshwater via precipitation directly onto the sea surface or via runoff. Arid, evaporative areas demonstrate the evaporative surface water $\delta^{18}\text{O}$ enrichment. Regions in reasonable proximity to a river mouth will be affected by the volumetrically weighted average isotopic composition of precipitation over the river's catchment area. A

high-latitude river system imports freshwater with generally lower $\delta^{18}\text{O}$ values than those of a low latitude river – for example, the average (pre-Aswan dam) Nile river $\delta^{18}\text{O}$ composition was near -2‰ , whereas the Arctic McKenzie river discharges waters with a composition around -20‰ .

Two crucial 'delayed' responses need to be taken into account. Both result from long-term storage of precipitation. Besides runoff from rivers, icebergs calving from continental ice sheets also provide a source of 'continental' freshwater. Icebergs import into the ocean a 'fossil' isotopic signature depending on the age of the calving ice. Since ice cores in the Greenland and Antarctic Ice Sheets penetrate ice aged well over 100 000 years, and (in Antarctica) even reach 900 000 years, the isotopic signatures of bergs shed by those ice sheets should not be considered as a result of the present-day freshwater cycle. Similarly, but much less importantly, fossil waters may accumulate in aquifers and at a much later stage contribute to river discharge into the oceans, thus causing deviations from the isotopic compositions of rivers expected from precipitation in the catchment basin. Fossil ground waters may be as old as 35 000 years.

Long-Term 'Storage': Glacial Ice Volume

Apart from delayed return of 'fossil' isotopic signals back into the ocean, long-term storage systems, such as glacial ice sheets and – to a much lesser extent – major aquifers, may also affect the global $\delta^{18}\text{O}$ balance. Because the timescales of storage (on the order of 10^4 – 10^5 years) exceed those of ocean ventilation (on the order of 10^3 years), the modification of ocean water $\delta^{18}\text{O}$ due to these storage effects will be mixed throughout the ocean and, therefore, will be manifest both at the surface and at depth.

The main influence is related to the volume of glacial ice sheets. These are built up by high latitude precipitation (snow) at very low temperatures toward the end of the Rayleigh distillation process (Figure 1), and so record extremely light $\delta^{18}\text{O}$ values (Figure 2). This preferential sequestration of ^{16}O in ice sheets leaves the oceans relatively enriched in ^{18}O . At the same time, build-up of ice volume lowers global sea level. Research on $\delta^{18}\text{O}$ changes in fossil carbonate, complemented by accurate constraints on sea level changes from fossil coral reef studies, suggests that the relationship between sea level lowering and mean oceanic $\delta^{18}\text{O}$ enrichment approximates $0.012 \pm 0.001\text{‰m}^{-1}$. Work on $\delta^{18}\text{O}$ values of actual glacial waters, trapped as pore water in marine sediments, has revised this relationship to about $0.010 \pm 0.001\text{‰m}^{-1}$ (see the Section 'Processes Controlling Oxygen Isotope Ratios in Seawater').

The above relationships provide a sound working model for a time with a well-matured ice sheet, but it should be noted that the processes behind it may invoke a more nonlinear relationship for growing or recently matured ice sheets. Ice sheets go through a cycle of growth (young), equilibrium (mature), and decay. Mature ice sheets are mass balanced: ice accumulates at the surface, then flows from the surface to the bottom and out to the edges, where it is shed through iceberg calving and/or melting. Precipitation reaching the summit of a recently 'matured' ice sheet may therefore be significantly more depleted in $\delta^{18}\text{O}$ than that lost from the edges at the same time,

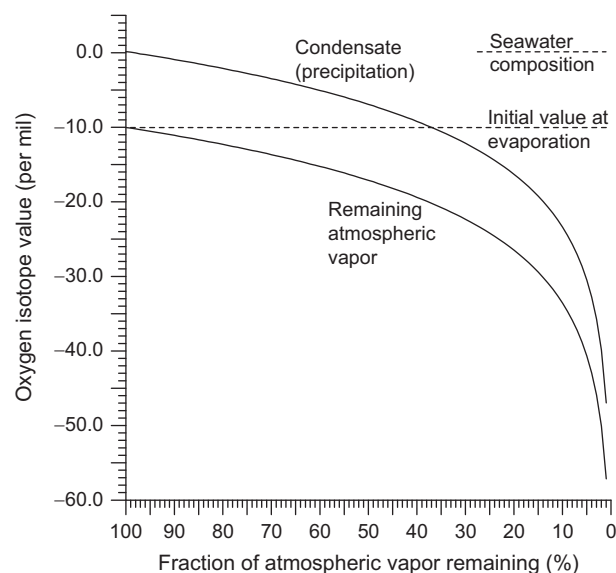


Figure 2 The relationships between isotopic composition of atmospheric vapor and precipitation relative to evaporation with an original composition of -10‰ and fractionation at a constant temperature of 15°C . Reproduced from Rohling EJ and Cooke S (1999) Stable oxygen and carbon isotope ratios in foraminiferal carbonate. In: Sen Gupta BK (ed.) *Modern Foraminifera*, pp. 239–258. Dordrecht, The Netherlands: Kluwer Academic.

which concerns ice deposited when the ice sheet was much 'younger' (smaller/lower with higher temperatures than over the mature ice sheet). Therefore, an ice sheet may be mass balanced, but not yet isotopically balanced. After a significant lag ($>10^4$ years), the ice reaching the edges will be that built up on a mature ice sheet, and the isotopic composition of accumulating and calving ice will be more or less the same; only then will the ice sheet have reached isotopic balance.

Sea Ice Freezing and Melting

The $\delta^{18}\text{O}$ of newly formed sea ice is $2.57 \pm 0.10\text{‰}$ enriched relative to that of seawater. Although this is a small offset compared with the massive salinity difference between seawater (S usually >30 p.s.u.) and sea ice ($S < 20$ p.s.u.), the isotopic difference still imposes a major seasonal fluctuation associated with sea ice formation and melting. It is tempting to assume that such seasonal influences would cancel out in the long term, but increases in surface water salinity due to salt (brine) rejection when sea ice is formed may lead to convection and transport of existing surface waters into the ocean interior. The removed surface waters would be replaced by surface waters that were not (as much) affected by freezing processes, and when the sea ice subsequently melts the isotopic effect would not cancel out. The 'brine rejection' influence of sea ice formation can thus lead to significant regional vertical and horizontal $\delta^{18}\text{O}$ anomalies.

Advection and Diffusion

Advection and mixing of water masses from different source areas strongly affect the basic $\delta^{18}\text{O}$ composition observed at any site. Each 'source area' concerns a basin or region – which may be very remote from the study site – where surface waters are 'imprinted' with a certain $\delta^{18}\text{O}$ composition, through action of the freshwater cycle, freezing/melting of sea ice, etc. This 'preset' isotopic composition may be considered as a virtually conservative property for the newly formed water

mass, as long as this water mass does not come into contact with other sinks or sources of $\delta^{18}\text{O}$, notably the freshwater cycle. In practice, therefore, $\delta^{18}\text{O}$ may be used as a conservative tracer for transport (advection) and mixing of water masses in the subsurface ocean.

A mixing end product receives a $\delta^{18}\text{O}$ composition resulting from the volumetrically weighted averaged $\delta^{18}\text{O}$ compositions of its components: $\delta^{18}\text{O}_{\text{end product}} = (A \delta^{18}\text{O}_A + B \delta^{18}\text{O}_B + C \delta^{18}\text{O}_C) / (A + B + C)$ where A , B , and C are the volumes of three mixing components and $\delta^{18}\text{O}_A$, $\delta^{18}\text{O}_B$, and $\delta^{18}\text{O}_C$ are their isotopic compositions, respectively. Any change in relative proportions or the individual 'preset' isotopic compositions of the various mixing components would, therefore, affect the isotopic composition of the mixing end product. Any change in isotopic composition must therefore be viewed within the broader context of water mass formation and mixing on ocean-wide scales, and should not be purely ascribed to local changes in surface forcing (e.g., freshwater budget). It has been found that these hitherto 'ignored' processes may be particularly important for the global $\delta^{18}\text{O}$ distribution during times with comprehensively rearranged ocean circulation (e.g., glacial periods).

Global Seawater $\delta^{18}\text{O}$ Distribution

The combined processes listed above result in a global seawater $\delta^{18}\text{O}$ distribution. A current collection exists of over 20000 seawater $\delta^{18}\text{O}$ measurements made since about 1950 (Schmidt et al., 1999). The date of that reference refers to the original appearance date, based on compilations made for two publications (Bigg and Rohling, 2000; Schmidt et al., 1999), but the database is current up to the present. From that database, Figure 3 shows the integrated surface ocean (top 50 m) $\delta^{18}\text{O}$ field interpolated onto a $4^\circ \times 5^\circ$ grid. Data are registered from all depths and are accompanied by a variety of other oceanographic measurements, most notably salinity. A diverse series of search modes allows extraction of data from the database according to specific research needs.

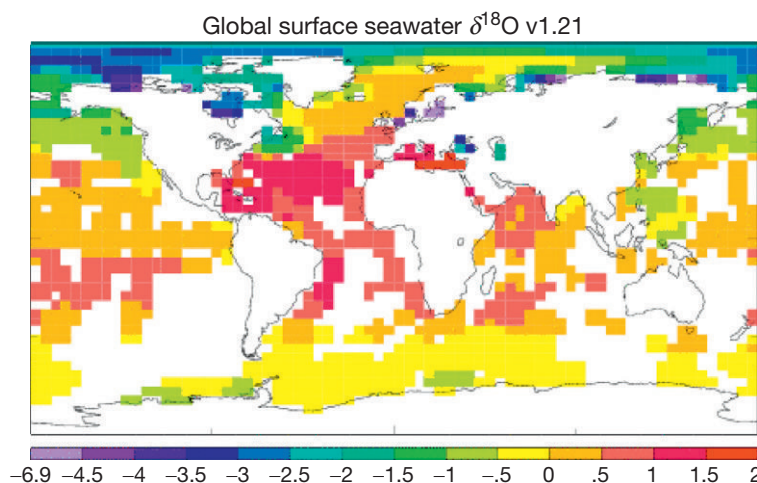


Figure 3 Example of output from the Schmidt et al. (1999) global sea water $\delta^{18}\text{O}$ database (extracted 13 January 2012). The image shows surface water $\delta^{18}\text{O}$ integrated over the top 50 m on a $4^\circ \times 5^\circ$ grid.

Oxygen Isotope Ratios in Seawaters of the Quaternary

Oxygen isotope ratios in seawater have changed considerably in the geological past. Apart from regional and short-term temporal variability, there have been fluctuations on global scales and of long ($\geq 10^4$ years) duration. These global-scale and long-duration fluctuations were mostly associated with changes in continental ice volume. In this section, it is discussed how we know of such past $\delta^{18}\text{O}$ variations.

Seawater $\delta^{18}\text{O}$ Inferred from Measurements on Fossil Carbonates

By far the most common manner to evaluate past changes in seawater $\delta^{18}\text{O}$ is via an indirect method, which starts with the analysis of $\delta^{18}\text{O}$ in fossil carbonates – mostly the shells of unicellular zooplankton (Foraminifera). The stable isotopes of oxygen (and carbon) in carbonates are analyzed by mass spectrometric determination of the mass ratios of carbon dioxide (CO_2) obtained from the sample, with reference to a standard carbon dioxide of known composition. The CO_2 is produced by reaction of the carbonate with phosphoric acid: $\text{CaCO}_3 + \text{H}_3\text{PO}_4 \rightleftharpoons \text{CaHPO}_4 + \text{CO}_2 + \text{H}_2\text{O}$. The $\delta^{18}\text{O}$ of carbonates is not measured against the VSMOW standard but against the Vienna Pee Dee Belemnite (VPDB) standard. The VPDB and VSMOW scales relate to one another according to $\delta^{18}\text{O}_{\text{VSMOW}} = 1.03092 \delta^{18}\text{O}_{\text{VPDB}} + 30.92$.

There are several processes that determine the oxygen isotope composition of newly formed foraminiferal carbonate. Firstly, equilibrium fractionations between water and the various carbonate species (CaCO_3 , H_2CO_3 , HCO_3^- , CO_3^{2-}) determine an important temperature influence on the $\delta^{18}\text{O}$ of foraminiferal carbonate. Secondly, there are considerable kinetic effects, notably as the shell growth-rate changes through the organism's life cycle. To minimize the influence of kinetic effects, often grouped under the name 'vital effects,' analytical strategies have been designed that focus on time series of single species (to avoid vital effect differences between species), and that use very narrow size windows from which specimens are picked for analysis (to minimize changes in the vital effects with growth). There are other, currently less well-constrained influences, such as the concentration of the carbonate ion (CO_3^{2-}). Stable isotope ratios in carbonate are discussed in a separate chapter, and the present discussion is restricted to equilibrium fractionation between water and carbonate, which holds important clues regarding the way we can determine past seawater $\delta^{18}\text{O}$ changes.

The overall reaction involved in the precipitation of carbonate is: $\text{Ca}^{2+} + 2\text{HCO}_3^- \rightleftharpoons \text{CaCO}_3 + \text{CO}_2 + \text{H}_2\text{O}$. Where oxygen is concerned, the various carbonate species in seawater, including the predominant bicarbonate (HCO_3^-) ion, are isotopically equilibrated with the water (H_2O) molecules. The overall equilibrium fractionation between calcite and water is a function of temperature. For the range of 0–500 °C, O'Neil et al. (1969) determined that the equilibrium fractionation factor α_{c-w} between calcite and water (α_{c-w}) changes with temperature according to: $\alpha_{c-w} = \exp\{(2.78 T^{-2}) \times 10^3 - 3.39 \times 10^{-3}\}$ where T is in Kelvin. Other workers have reported slightly different equations, for example, with the last coefficient at

2.89×10^{-3} rather than 3.39×10^{-3} . The summary effect of equilibrium fractionation is a roughly 0.2‰ depletion in carbonate $\delta^{18}\text{O}$ for every 1 °C temperature increase, although Kim and O'Neil (1997) presented a more detailed relationship where the $\delta^{18}\text{O}$ change with temperature is more pronounced at low temperatures (up to $0.25\text{‰} \text{ } ^\circ\text{C}^{-1}$) than at higher temperatures (around $0.2\text{‰} \text{ } ^\circ\text{C}^{-1}$).

A first observation to be made regarding planktonic Foraminifera is that several species may inhabit different depths at different stages of their life cycle. Since temperature decreases with increasing depth in the surface oceans, vertical migrations influence equilibrium fractionation. This again is a problem that can be circumvented by analytical strategy; time series of data are considered for single species in very narrow size windows, so the habitat-depth effect should be similar for all and thus will not affect the inferred variability sea water $\delta^{18}\text{O}$ over time.

Simplified, therefore, a time series of $\delta^{18}\text{O}$ analyzed from fossil planktonic Foraminifera in a sediment core is dominated by two main variables: temperature variability and seawater $\delta^{18}\text{O}$ variability. The temperature changes can be 'corrected for' using other methods for estimating past seawater temperatures (e.g., Mg/Ca ratios in the foraminiferal carbonate, organic geochemical biomarker ratios, or statistical transforms of microfossil abundance variations). This then allows the estimation of past changes in seawater $\delta^{18}\text{O}$.

Time series of foraminiferal $\delta^{18}\text{O}$ and – by implication – seawater $\delta^{18}\text{O}$ illustrate that glacial ice volume has varied strongly over the last 2 million years, a period known on the geological timescale to fall within the Quaternary. During the last 400 000 years, the main variability can be seen as a rough sawtooth pattern with a 100 000-year rhythm of glacial–interglacial cycles, which is mainly related to global ice volume fluctuations (Figure 4). Ice sheets grew spasmodically over periods of roughly 90 000 years, reaching a maximum extent that was followed by a very abrupt ice volume reduction heralding the onset of an interglacial maximum with a typical mean duration of some 10 000 years. Between 0.4 and 1 million years ago, there was a less well-defined state, with a complex interplay between main periods of 100 000 and 41 000 years for the glacial cycles. Prior to 1 million years ago (especially in the interval between 2.5 and 1 million years ago), the glacial cycles followed a well-developed 41 000-year period, and in that period the glacial cycles did not have a conspicuously sawtoothed temporal structure (Figure 4). Following calibration with dated fossil coral reefs, the past $\delta^{18}\text{O}$ fluctuations were found to indicate that global ice volume has waxed and waned so much that global sea level variations occurred between 140 m below and 5–10 m above the present-day level. For the last glacial cycle, such studies suggested a relationship between sea level lowering and mean oceanic $\delta^{18}\text{O}$ enrichment in the region of $0.012 \pm 0.001\text{‰} \text{ m}^{-1}$.

Other Uses of Foraminifer-Based $\delta^{18}\text{O}$ in Paleoceanography

There have been recent forays into alternative uses of foraminiferal carbonate-based $\delta^{18}\text{O}$ that avoid the complications involved in deconvolving the carbonate $\delta^{18}\text{O}$ data in terms of past temperature and seawater $\delta^{18}\text{O}$. These studies have focused instead on the fact that, in specific settings, carbonate

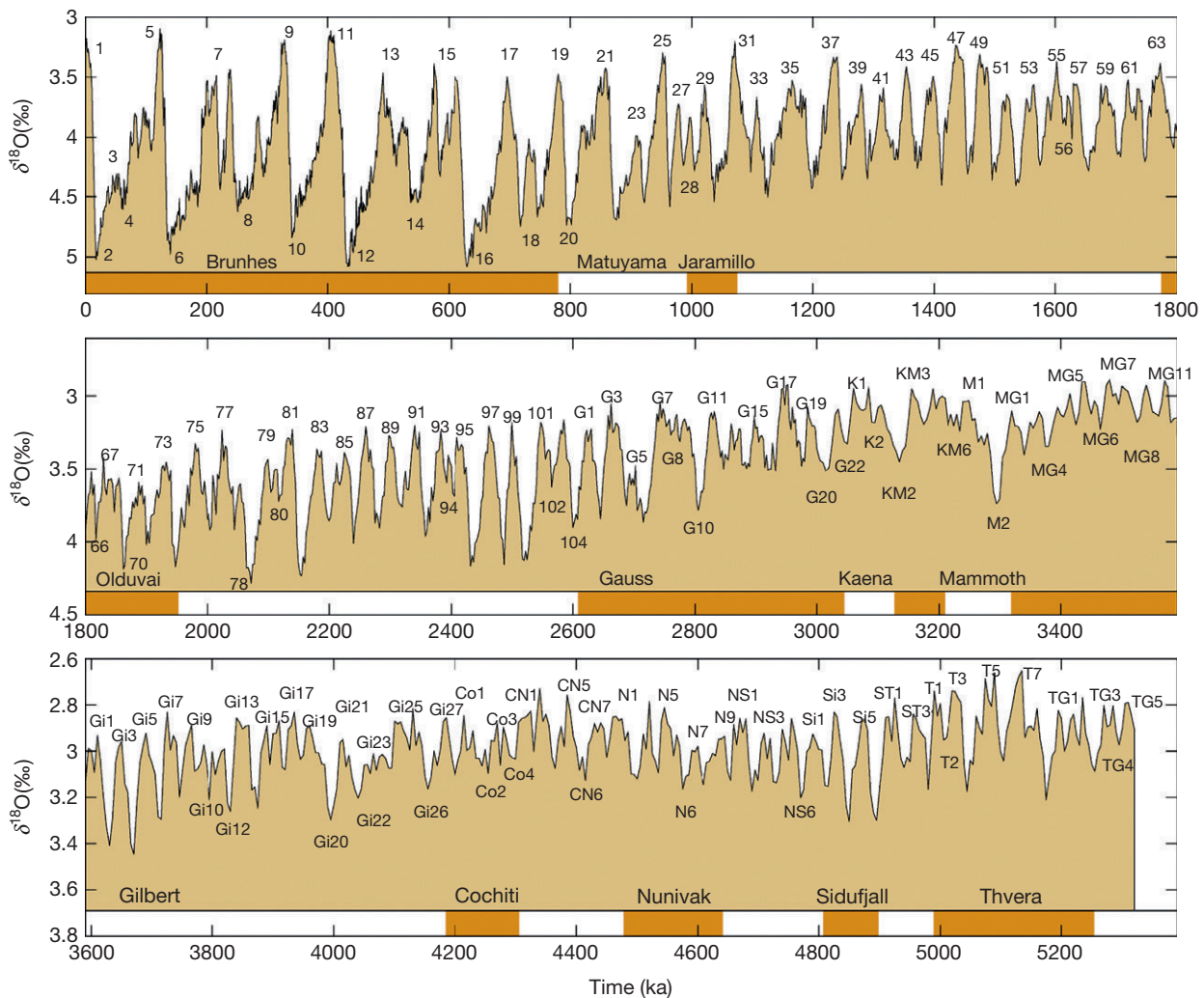


Figure 4 A stacked record of stable oxygen isotope variations for the last 5.3 million years, constructed from 57 globally distributed deep-sea (benthic foraminiferal) records. Numbers indicate glacial (down = heavy $\delta^{18}\text{O}$ values) and interglacial (up = light $\delta^{18}\text{O}$ values) intervals. The dark/light bar at the bottom of each panel indicates (named) intervals of normal (dark) and reversed (light) geomagnetic polarity. Time is indicated in units of 1000 years (ka). Reproduced from Lisiecki LE and Rayme ME (2005) A Pliocene-Pleistocene stack of 57 globally distributed benthic $\delta^{18}\text{O}$ records. *Paleoceanography* 20: PA1003, <http://dx.doi.org/10.1029/2004PA001071>.

$\delta^{18}\text{O}$ offers a very useful approximation of past seawater density, because it is a function of temperature and the hydrological budget (as is density). Thus the carbonate $\delta^{18}\text{O}$ data can be used as a dynamic proxy for the reconstruction of density gradients with depth or in a spatial sense. Lynch-Stieglitz et al. (1999) used depth transects of benthic foraminiferal $\delta^{18}\text{O}$ data on both sides of Florida Straits in geostrophic (where the horizontal pressure gradient is balanced by the Coriolis force) current intensity reconstructions of the Florida Current during glacial times. Billups and Schrag (2000) argued that, in tropical and subtropical regions, the $\delta^{18}\text{O}$ data obtained from various planktonic foraminiferal species may be used to reconstruct past changes in spatial surface-water density gradients, on a global scale.

Another line of work has focused on the sea-level-dependent concentration effect on $\delta^{18}\text{O}$ in evaporative marginal seas with limited connection to the open ocean. A key study along these lines concerns the Red Sea, a highly evaporative basin with a very

shallow (137 m) connection to the open ocean. As sea level varied on glacial-interglacial timescales between about 130 m below the present and about 5 m above the present, the broadly triangular strait profile imposed very serious limitations on the amount of water exchange between the Red Sea and the open ocean. With lowering sea level, the exchange is progressively inhibited, which increases the residence time of waters inside the highly evaporative Red Sea. As a consequence of this sea level effect, the glacial-interglacial $\delta^{18}\text{O}$ contrast in the Red Sea is amplified about five times relative to the open ocean. By quantifying the relationship between sea level and Red Sea $\delta^{18}\text{O}$ change, Siddall et al. (2003) developed a new technique for reconstruction of global sea level variability, using foraminiferal $\delta^{18}\text{O}$ records from central Red Sea sediment cores. Records employing this method have been further developed in detail through the last 520,000 years (Rohling et al., 2009, 2010; Figure 5). This technique relies on a physically defined

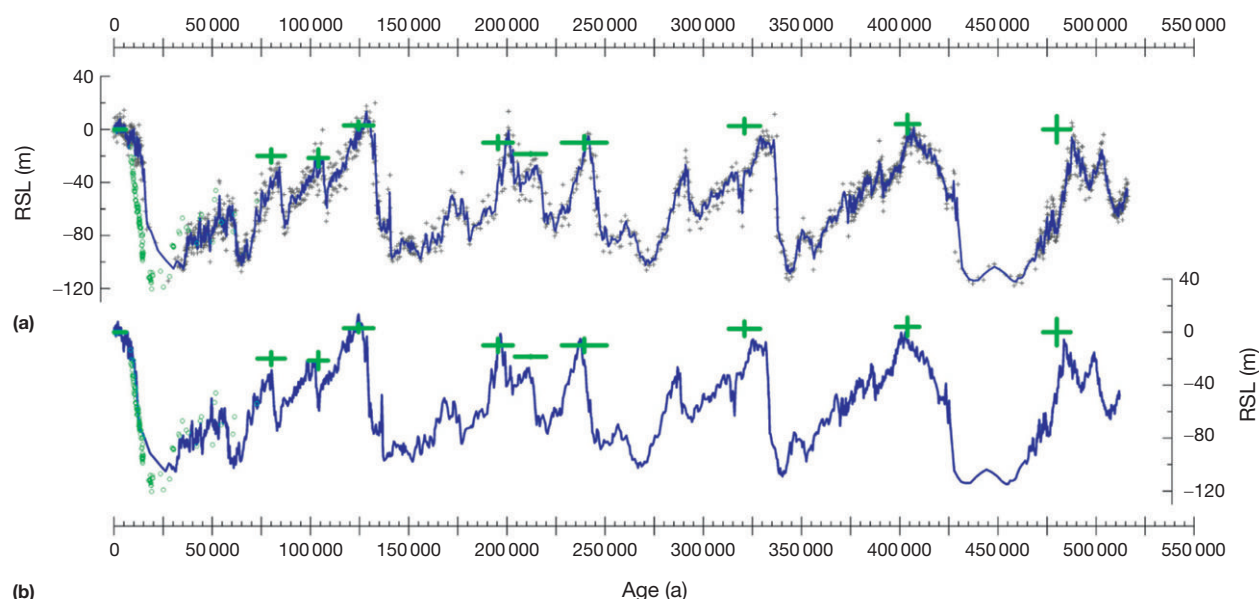


Figure 5 Sea level reconstruction from the Red Sea method (Rohling et al., 2009). Gray dots are individual values; blue line is a smoothing function so that individual data are spread around it a normal distribution with $1\sigma = 6$ m, which is the method's realistic uncertainty (Siddall et al., 2003). Green symbols denote benchmark data from U–Th dated corals and submerged speleothems. Panel (a) shows the record on its initial chronology, based on graphical correlation with the Antarctic EPICA Dome C temperature anomaly record (Jouzel et al., 2007). Panel (b) shows the record after a systematic adjustment of -4 ka, based on the offset seen in (a) between the Red Sea sea level record and the radiometrically dated points during the last deglaciation (19–8 ka) (Rohling et al., 2010). This brought midpoint ages for interglacials in the Red Sea record within 1.5% of the midpoint ages from radiometric datings (Rohling et al., 2010). Although a good match was obtained on glacial–interglacial timescales, further improvements to the chronology remain needed on shorter timescales. Data available from the DATA link with Rohling et al. (2010) at <http://www.highstand.org/erohling/ejrhome.htm>.

relationship between Red Sea $\delta^{18}\text{O}$ and sea level in the strait and is therefore fundamentally different in approach and sensitivity to the global deep-sea $\delta^{18}\text{O}$ records shown in Figure 4.

'Direct' Measurement of Past Seawater

Recent work has introduced significant new insight into the history of seawater $\delta^{18}\text{O}$. Adkins et al. (2002) measured $\delta^{18}\text{O}$ profiles for pore waters extracted from marine sediment cores. The pore water profile reflects the past bottom water $\delta^{18}\text{O}$ conditions, subject to advective and diffusive processes within the sediment, and the Last Glacial Maximum ($\sim 20,000$ years ago) stands out as the interval with highest $\delta^{18}\text{O}$ values (i.e., maximum ice volume). Hence the actual Last Glacial Maximum bottom water $\delta^{18}\text{O}$ was determined from sediment cores for a variety of sites around the world, from a range of different water depths. The sea level difference between the Last Glacial Maximum and the present was taken from other analyses, and overall, this study suggests that the actual relationship between sea level lowering and mean oceanic $\delta^{18}\text{O}$ enrichment is closer to $0.010 \pm 0.001\text{‰ m}^{-1}$. This indicates that the steeper relationship previously derived from carbonate-based $\delta^{18}\text{O}$ changes likely contained some residual temperature effects (i.e., glacial deep-water temperatures were colder than was assumed).

See also: K/Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ Dating. **Paleoceanography, Physical and Chemical Proxies: Oxygen-Isotope Stratigraphy of the Oceans. Quaternary Stratigraphy: Overview.**

References

- Adkins JF, McIntyre K, and Schrag DP (2002) The salinity, temperature and $\delta^{18}\text{O}$ of the glacial deep ocean. *Science* 298: 1769–1773.
- Bigg GR and Rohling EJ (2000) An oxygen isotope data set for marine water. *Journal of Geophysical Research* 105: 8527–8535.
- Billups K and Schrag DP (2000) Surface ocean density gradients during the last glacial maximum. *Paleoceanography* 15: 110–123.
- Craig H and Gordon LI (1965) Isotope oceanography: Deuterium and oxygen 18 variations in the ocean and the marine atmosphere. *Symposium on Marine Geochemistry*, pp. 277–374. University of Rhode Island Occasional Publications 3.
- Dansgaard W (1964) Stable isotopes in precipitation. *Tellus* 16: 436–468.
- Garlick GD (1974) The stable isotopes of oxygen, carbon, hydrogen in the marine environment. In: Goldberg ED (ed.) *The Sea*, vol. 5, pp. 393–425. New York: Wiley.
- Gonfiantini R (1986) Environmental isotopes in lake studies. In: Fritz P and Fontes JC (eds.) *Handbook of Environmental Isotope Geochemistry*, vol. 2, pp. 113–168. Amsterdam: Elsevier.
- Hoefs J (1997) *Stable Isotope Geochemistry*, 4th edn. Berlin: Springer-Verlag.
- Hoffmann G and Heimann M (1997) Water isotope modeling in the Asian monsoon region. *Quaternary International* 37: 115–128.
- Joussaume S and Jouzel J (1993) Paleoclimatic tracers: An investigation using an atmospheric general circulation model under ice age conditions, 2. Water isotopes. *Journal of Geophysical Research* 98(D2): 2807–2830.
- Jouzel J, Masson-Delmotte V, Cattani O, et al. (2007) Orbital and millennial Antarctic climate variability over the past 800,000 years. *Science* 317: 793–796.
- Jouzel J, Merlivat L, and Roth E (1975) Isotopic study of hail. *Journal of Geophysical Research* 80: 5015–5030.
- Kim ST and O'Neil JR (1997) Equilibrium and nonequilibrium oxygen isotope effects in synthetic calcites. *Geochimica et Cosmochimica Acta* 61: 3461–3475.
- Lisiecki LE and Raymo ME (2005) A Pliocene–Pleistocene stack of 57 globally distributed benthic $\delta^{18}\text{O}$ records. *Paleoceanography* 20: PA1003. <http://dx.doi.org/10.1029/2004PA001071>.
- Lynch-Stieglitz J, Curry BC, and Slowey N (1999) A geostrophic transport estimate for the Florida Current from the oxygen isotope composition of benthic foraminifera. *Paleoceanography* 14: 360–373.

- Majoube M (1971) Fractionnement en oxygène 18 et en deutérium entre l'eau et sa vapeur. *Journal de Chimie Physique* 10: 1423–1436.
- Merlivat L and Jouzel J (1979) Global climatic interpretation of the deuterium-oxygen 18 relationship for precipitation. *Journal of Geophysical Research* 84(C8): 5029–5033.
- O'Neil JR, Clayton RN, and Mayeda TK (1969) Oxygen isotope fractionation on divalent metal carbonates. *Journal of Chemical Physics* 51: 5547–5558.
- Rohling EJ, Braun K, Grant K, et al. (2010) Comparison between Holocene and Marine Isotope Stage-11 sea-level histories. *Earth and Planetary Science Letters* 291: 97–105.
- Rohling EJ and Cooke S (1999) Stable oxygen and carbon isotope ratios in foraminiferal carbonate. In: Sen Gupta BK (ed.) *Modern Foraminifera*, pp. 239–258. Dordrecht, The Netherlands: Kluwer Academic.
- Rohling EJ, Grant K, Bolshaw M, et al. (2009) Antarctic temperature and global sea level closely coupled over the past five glacial cycles. *Nature Geoscience* 2: 500–504.
- Rozanski K, Araguas-Araguas L, and Gonfiantini R (1993) Isotopic patterns in modern global precipitation. In: Swart PK, Lohmann KC, McKenzie J, and Savin S (eds.) *Climate Change in Continental Isotopic Records. Geophysical Monograph*, vol. 78, pp. 1–36. Washington, DC: American Geophysical Union.
- Rozanski K, Sonntag C, and Münnich KO (1982) Factors controlling stable isotope composition of European precipitation. *Tellus* 34: 142–150.
- Schmidt GA (1999) Forward modeling of carbonate proxy data from planktonic foraminifera using oxygen isotope tracers in a global ocean model. *Paleoceanography* 14: 482–497.
- Schmidt GA, Bigg GR, and Rohling EJ (1999) *Global Seawater Oxygen-18 Database*. <http://data.giss.nasa.gov/o18data/>.
- Siddall M, Rohling EJ, Almogi-Labin A, et al. (2003) Sea-level fluctuations during the last glacial cycle. *Nature* 423: 853–858.
- Stewart MK (1975) Stable isotope fractionation due to evaporation and isotopic exchange of falling waterdrops: Applications to atmospheric processes and evaporation of lakes. *Journal of Geophysical Research* 80(9): 1133–1146.

Radioisotope Proxies

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Introduction

What are Radioisotopes?

Atoms consist of a cloud of negatively charged electrons circling around a positively charged nucleus. The nucleus accounts for most of the atomic mass and is composed of positively charged protons and uncharged neutrons. Electrons and protons have electrical charges of equal magnitude but are opposite in polarity. They occur in equal numbers, so that single atoms have no net charges. The number of protons is referred to as the atomic number (Z). Each element of the periodic table is characterized by its atomic number (i.e., its number of protons), which dictates the number of electrons and the configuration of valence electrons establishing the chemical properties of the atom.

Unlike the number of protons and electrons, the number of neutrons in the nucleus (N) can vary without changing the chemical properties of the atom. This is because changing the number of uncharged neutrons in the nucleus does not require changing the number of electrons and the configuration of the valence electrons. A given element can thus consist of atoms with different numbers of neutrons. These atoms have similar chemical properties, but different atomic masses or mass number ($A = Z + N = \text{number of protons} + \text{number of neutrons}$). Atoms of the same element (i.e., with identical Z) but with different masses (i.e., with different A) are called isotopes. By convention, isotopes are identified by the chemical symbol of the element and the atomic mass number (A) in superscript (e.g., ^{238}U , ^{235}U).

Depending on the relative proportion of protons and neutrons, the nucleus of an atom can be stable (resulting in a stable isotope) or unstable (resulting in a radioisotope). Radioisotopes 'decay' spontaneously by changing the proton/neutron ratios of their nucleus until they reach a stable nuclear configuration. They do so by emitting energetic particles (alpha or beta) from their nuclei. During alpha decay, the nucleus emits an alpha particle consisting of two neutrons and two protons. Consequently, the daughter (i.e., the product of the decay) has lost four atomic mass units ($A_{\text{daughter}} = A_{\text{parent}} - 4$) and two protons ($Z_{\text{daughter}} = Z_{\text{parent}} - 2$). The atom that undergoes decay thus becomes a different element with different chemical properties (e.g., $^{238}\text{Th} \rightarrow ^{234}\text{Ra} + \alpha$). During beta decay, the nucleus ejects an electron, which can be charged negatively (negatron) or positively (positron). When the nucleus emits a negatron, one of the neutrons in the nucleus becomes a proton. Consequently, the daughter keeps the same atomic mass as the parent atom ($A_{\text{daughter}} = A_{\text{parent}}$) but gains one proton ($Z_{\text{daughter}} = Z_{\text{parent}} + 1$), thereby also changing its chemical identity (e.g., $^{234}\text{Th} \rightarrow ^{234}\text{Pa} + \beta^-$). When the nucleus emits a positron, one of the protons in the nucleus becomes a neutron.

General Applications of Radioisotopes

The rate at which radioisotopes decay ($-dN/dt$; i.e., the rate at which atoms of the parent are transformed into daughters) is proportional to the number of atoms of the radioisotope present (N):

$$-dN/dt = \lambda N \quad [1]$$

where λ is the decay constant, which is a fixed property for a particular radioisotope and is independent of environmental factors (temperature, pressure, etc.). λN is also called the 'activity' of the radioisotope N and is often expressed in 'disintegration per minutes' (dpm).

Integrating this equation results in a simple exponential law that describes the change in the number of parent isotopes as a function of time:

$$N(t) = N_{(0)} e^{-\lambda t} \quad [2]$$

where $N(t)$ is the number of parent isotopes remaining at time t and $N_{(0)}$ is the initial number of parent isotopes at $t=0$.

If we know $N_{(0)}$, measuring $N(t)$ provides a means to calculating the period of time over which decay occurred ($t = -1/\lambda \ln(N(t)/N_{(0)})$). Estimating $N_{(0)}$ is often the tricky part of this approach. Nonetheless, applying this simple equation provides 'clocks' that have been widely used to date environmental samples on timescales ranging from billions of years to a few months (see the section on chronology).

Besides their wide applications in chronology, radioisotopes also provide quantitative information on a wide range of environmental processes. This is particularly true for the radioactive isotopes produced in the three naturally occurring radioactive decay series (Table 1). These decay series are initiated by the very slow radioactive disintegration of three long-lived isotopes: ^{238}U (half-life = 4.47×10^9 years), ^{235}U (half-life = 7.04×10^8 years), and ^{232}Th (half-life = 1.41×10^{10} years). The half-life is the amount of time it takes for half of the radioactive atoms in a sample to decay. The atoms of ^{238}U , ^{235}U , and ^{232}Th that now decay were produced by nucleosynthesis in supernovas billions of years ago, but because of their very long half-lives, they still persist to this day. When left undisturbed, all the daughters in the series have the same decay rate (or the same 'activity'). The decay series is then said to be in 'secular equilibrium.' In this situation, each radioactive isotope of the series decays at a rate that is equal to the rate at which it is produced (which is the rate at which its parent is decaying). Ultimately, the rate of decay of all the daughters is dictated by the rate of decay of the initial parent (^{238}U , ^{235}U , or ^{232}Th), which very slowly decreases with time.

Table 1 Natural decay series initiated by the ^{238}U , ^{235}U , and ^{232}Th

	U-238 series						Th-232 series						U-235 series					
Np																		
U	U-238 4.47 × 10 ⁹ yr		U-234 2.48 × 10 ⁵ yr										U-235 7.04 × 10 ⁸ yr					
Pa	↓	Po-234 1.18 m	↓										↓	Po-231 3.2 × 10 ⁴ yr				
Th	Th-234 24.1 d		Th-230 7.52 × 10 ⁴ yr				Th-232 1.41 × 10 ¹⁰ yr		Th-228 1.90 yr				Th-231 25.6 h			Th-227 18.6 d		
Ac			↓				↓	Ac-228 6.13 h	↓					↓	Ac-227 22.0 yr			
Ra			↓				↓	↓	↓							↓		
Fr																		
Rn			Rn-222 3.825 d															
At			↓															
Po			Po-218 3.05 m		Po-214 1.6 × 10 ⁻⁷ s		Po-210 138.4 d		Po-216 0.158 s	65%	Po-212 30 × 10 ⁻⁷ s					Po-215 1.63 × 10 ⁻³ s		
Bi			↓	Bi-214 19.7 m	↓	Bi-210 5.0 d	↓		↓	Bi-212 60.5 m	↓					↓	Bi-211 2.16 m	
Pb			Pb-214 26.8 m		Pb-210 223 yr		Pb-206		Pb-212 10.6 h	35%	Pb-208					Pb-211 36.1 m		Pb-207
Tl											Tl-208 3.1 m						Tl-207 4.79 m	

The usefulness of these isotopes in environmental sciences stems from the fact that processes on the surface of the Earth often separate daughters from parents, thereby perturbing secular equilibrium (Boudron et al., 2003). This arises because the successive daughters of the decay series consist of different elements with very different chemical properties (Table 1). For instance, while uranium and radium are relatively soluble in water, thorium, protactinium, polonium, and lead have very low solubility and are readily removed from aqueous solutions by adsorption on surfaces of particles or sediments. On the other hand, radon is a gas and tends to diffuse into the atmosphere from the rocks or water where it is produced from the decay of radium. When an environmental process separates a daughter from its parent (for instance, when settling particles remove Th isotopes from seawater into the underlying sediment, thereby separating them from their more soluble parent uranium isotopes), the activity of daughter isotopes (i.e., the rate at which they decay) decreases and becomes lower than the activity of parent isotopes. Assuming steady state, we can write a simple mass balance equation stating that the rate at which the daughter is produced (i.e., the rate at which the parent decays or the activity of the parent; $\lambda_{\text{parent}}N_{\text{parent}}$) must be equal to the rate at which the daughter decays (i.e., the activity of the daughter; $\lambda_{\text{daughter}}N_{\text{daughter}}$) plus the rate at

which the daughter is removed by the environmental process (R_{daughter}). Thus

$$R_{\text{daughter}} (\text{atoms min}^{-1}) = \lambda_{\text{parent}}N_{\text{parent}} - \lambda_{\text{daughter}}N_{\text{daughter}} \quad [3]$$

Measuring the difference in activity between parents and daughters in environmental samples (such as seawater, sediments, corals) thus provides means of quantifying the rate of removal of the daughter from the sample. In turn, determining the rate of removal of the daughter by a given process under different environmental conditions provides quantitative information on the relative intensity of this process under these environmental conditions and adds insight into the mechanisms involved. For instance, the parent isotope ^{238}U dissolved in seawater decays to its daughter ^{234}Th . Because uranium is very soluble, ^{238}U remains in seawater on average for 400 ka before being removed (mainly by incorporation into biogenic carbonates and basaltic rocks at mid-ocean ridges or by diffusion into anoxic sediments). ^{234}Th , on the other hand, being very insoluble, is quickly removed from seawater by adsorption on the surface of settling particles (a process called 'scavenging'). As a result, we find a deficit in the activity of ^{234}Th in seawater ($\lambda_{\text{Th-234}}N_{\text{Th-234}}$) compared to the activity of ^{238}U ($\lambda_{\text{U-238}}N_{\text{U-238}}$). Such a deficit is only found in the upper

water column because it is where particle concentrations and fluxes are sufficiently high to compete with the rapid rate of decay of ^{234}Th , whose half-life is only 24.1 days (Cochran and Masqué, 2003). In deeper water, particle concentrations and fluxes quickly decrease as a result of their remineralization, and the scavenging intensity resulting from this smaller particle flux is too small to remove ^{234}Th at a rate that would be comparable to its rapid rate of decay. Consequently, secular equilibrium between ^{238}U and ^{234}Th is essentially maintained below surface waters ($\lambda_{\text{U-238}}N_{\text{U-238}} = \lambda_{\text{Th-234}}N_{\text{Th-234}}$). As per eqn [3], the deficit in the upper water column is a quantitative estimate of the removal of ^{234}Th by scavenging from surface waters, and we find that it is larger in regions where biological productivity is higher. This observation clearly points to the export of biogenic particles from surface waters as the main variable regulating the scavenging of ^{234}Th and other insoluble elements from seawater. By measuring the ratio of the concentration of any constituent of the settling particles ($[X]$) to the activity of ^{234}Th in the settling material ($A_{\text{Th-234}}$), we can also calculate the export flux of this constituent (F_x):

$$F_x(\text{g min}^{-1}) = R_{\text{Th-234}}(\text{atoms min}^{-1}) \times [X]/A_{\text{Th-234}}(\text{g atoms}^{-1}) \quad [4]$$

This method has been widely used to estimate the export flux of carbon from surface to deep ocean water and evaluate the role of the marine biota in controlling the concentration of carbon dioxide in the atmosphere (Cochran and Masqué, 2003).

Applications of Radioisotope Tracers in Quaternary Paleooceanography

For Quaternary paleoceanographic applications, we are restricted to isotopes that are incorporated into paleoceanographic archives (mainly sediments, corals, or ferromanganese crusts) and that have half-lives long enough to persist on Quaternary timescales. The decay series initiated by ^{232}Th does not provide useful paleoceanographic tracers since the longest-lived radioisotope it generates (^{228}Ra) has a half-life of 5.75 years (Table 1). On the other hand, the ^{238}U and ^{235}U decay series both have one long-lived isotope with important paleoceanographic applications other than chronology (Francois, 2007; Henderson and Anderson, 2003). They are ^{230}Th (half-life = 75.38 ka) from the ^{238}U decay series and ^{231}Pa (half-life = 32.76 ka) from the ^{235}U decay series (Table 1). Information recorded in the sediment by these radioisotopes remains for a period of time equivalent to four or five times their half-life, beyond which their activity in sediment decreases to values that cannot be measured accurately. Paleooceanographic applications of these isotopes are thus largely limited to the last glacial cycle (i.e., the past 150 ka).

Paleoceanographic Applications Based on the Activity of ^{230}Th in Marine Sediments

^{230}Th is produced in seawater from the radioactive decay of ^{234}U (Figure 1). Since uranium resides in the ocean for

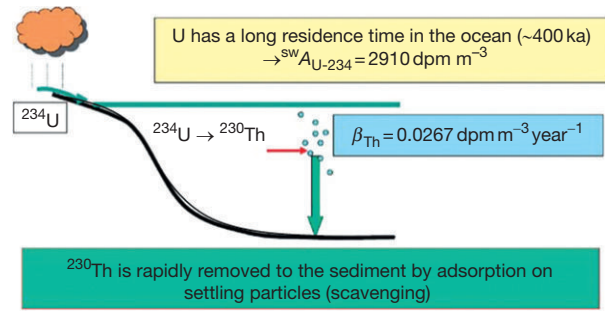


Figure 1 ^{234}U is leached from continental rocks by chemical weathering and added to the ocean by rivers and groundwater. ^{234}U decays in the ocean to produce ^{230}Th which is quickly adsorbed on settling particles and removed into marine sediments, a process called 'particle scavenging.'

about 400 ka, it is well mixed and its seawater concentration or activity is uniform ($^{sw}A_{\text{U-234}} = 2910 \text{ dpm m}^{-3}$), resulting in a constant rate of formation of ^{230}Th in seawater (rate of formation of ^{230}Th ($\text{atoms m}^{-3} \text{ min}^{-1}$) = ^{234}U decay rate = $2910 \text{ atoms m}^{-3} \text{ min}^{-1}$). The rate of production of ^{230}Th in seawater ($\beta_{\text{Th-230}}$) is often expressed in $\text{dpm m}^{-3} \text{ year}^{-1}$:

$$\beta_{\text{Th-230}} = \lambda_{230} \times \text{Rate of formation of } ^{230}\text{Th} (\text{atoms m}^{-3} \text{ min}^{-1}) \quad [5]$$

$$\beta_{\text{Th-230}} = \lambda_{230} \times ^{sw}A_{\text{U-234}} \quad [6]$$

$$\beta_{\text{Th-230}} = 9.16 \times 10^{-6} \text{ year}^{-1} \times 2910 \text{ dpm m}^{-3} \quad [7]$$

$$\beta_{\text{Th-230}} = 0.0267 \text{ dpm m}^{-3} \text{ year}^{-1} \quad [8]$$

As indicated above, Th is highly insoluble in seawater, and ^{230}Th produced from the decay of dissolved ^{234}U is promptly removed from the water column by scavenging (i.e., adsorption on settling particles; Figure 1).

Removal by scavenging can follow two distinct pathways (Anderson et al., 1990): a vertical pathway (proximal scavenging) due to rapid adsorption on particles settling locally and a lateral pathway (boundary scavenging) resulting from intensified scavenging at the ocean margins and other regions of high particle flux, following lateral transport (Figure 2). Partitioning between these two modes of removal is dictated by the residence time (or particle reactivity) of the scavenged element. Elements with lesser affinity for particles have longer residence time in the water column and can be transported laterally over longer distances to be preferentially removed by boundary scavenging in high particle flux regions. On the other hand, elements with very high affinity for particles have very short residence times in the water column, which limit the extent to which they can be laterally transported before removal. As a result, they tend to be removed locally by proximal scavenging (Bacon, 1988).

Thorium is among the most particle-reactive elements and is thus rapidly removed from the water column, primarily by proximal scavenging. The activity of ^{230}Th in seawater increases with depth from $<0.1 \text{ dpm m}^{-3}$ in surface water to $\approx 1 \text{ dpm m}^{-3}$ at the bottom of the ocean (Francois, 2007). This is nearly four orders of magnitude lower than the activity of its

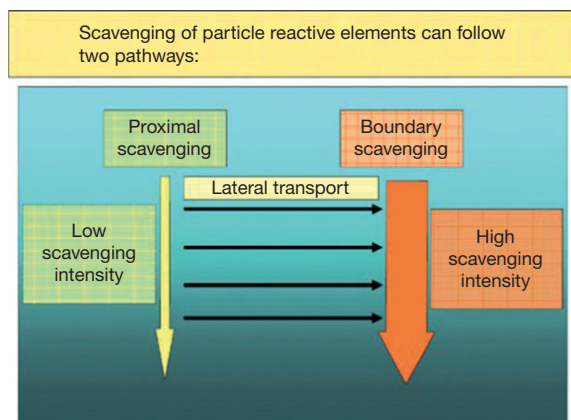


Figure 2 Proximal versus boundary scavenging. Elements with high affinity for particles are quickly removed from seawater by proximal scavenging. Elements with lesser affinity and longer residence times in seawater travel laterally over long distances to be preferentially removed in areas of high particle flux. Lateral transport occurs as a result of deep water circulation or eddy diffusion (see text for further explanations).

parent ^{234}U . The activity of ^{230}Th drops to much lower levels than the activity of ^{234}Th discussed above because the half-life of ^{230}Th is much longer (75.38 ka). With such a slow decay rate, ^{230}Th is overwhelmingly removed by scavenging rather than decay, even in deep waters where scavenging intensity is relatively weak.

The residence time of ^{230}Th in the water column (τ_{Th}) can be estimated by comparing its activity in seawater ($^{\text{sw}}A_{\text{Th-230}}$) to that of its parent ^{234}U ($^{\text{sw}}A_{\text{U-234}}$). Equation [3] indicates that the rate of removal of ^{230}Th (atoms min^{-1}) is equal to the activity of ^{234}U minus that of ^{230}Th . The residence time of ^{230}Th in seawater (τ_{Th}) with respect to scavenging is equal to the number of atoms of ^{230}Th present in seawater ($^{\text{sw}}A_{\text{Th-230}}/\lambda_{230}$) divided by their rate of removal by scavenging ($R_{\text{Th-230}}$). Therefore,

$$\tau_{\text{Th}} = ^{\text{sw}}A_{\text{Th-230}}/(\lambda_{230} \times R_{\text{Th-230}}) \quad [9]$$

$$\tau_{\text{Th}} = ^{\text{sw}}A_{\text{Th-230}}/(\lambda_{230} \times (^{\text{sw}}A_{\text{Th-234}} - ^{\text{sw}}A_{\text{Th-230}})) \quad [10]$$

$$\text{Since } ^{\text{sw}}A_{\text{Th-230}} \ll ^{\text{sw}}A_{\text{U-234}}$$

$$\tau_{\text{Th}} \cong ^{\text{sw}}A_{\text{Th-230}}/(\lambda_{230} \times ^{\text{sw}}A_{\text{U-234}}) \quad [11]$$

With a ^{234}U activity of 2910 dpm m^{-3} and a range from <0.1 to 1 dpm m^{-3} for ^{230}Th , the residence time of the latter in seawater thus ranges from <4 years in the upper water column to ≈ 40 years in deep water. Since the time required for lateral transport in a typical ocean basin is ≈ 100 years (Anderson et al., 1990), such short residence times indicate that removal of ^{230}Th by boundary scavenging is very limited, and thus ^{230}Th removal from the water column occurs primarily via the proximal route. This has been confirmed in modeling studies (Henderson et al., 1999; Siddall et al., 2008) and by direct measurements of the settling flux of ^{230}Th in the ocean (Yu et al., 2001).

An important paleoceanographic application has emerged from the rapid removal of ^{230}Th by proximal scavenging (François et al., 2004). To better understand past oceanic processes, paleoceanographers need to know the vertical rain rate

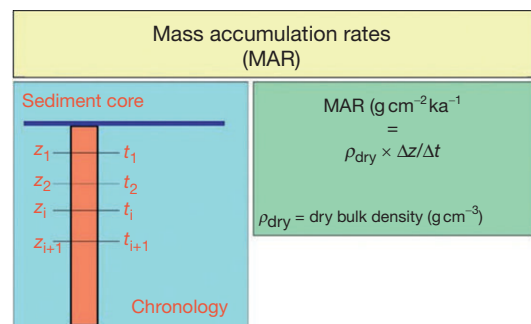


Figure 3 Estimating mass accumulation rates of sediments on the seafloor (see text for explanations).

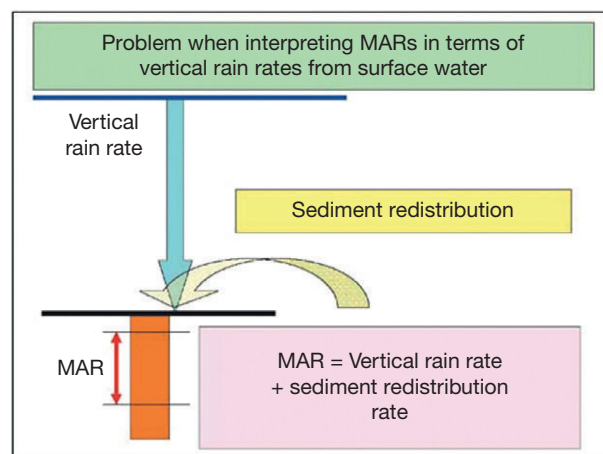


Figure 4 Sediment redistribution on the seafloor after initial deposition results in mass accumulation rates (MAR) that are different from the vertical rain rate of particles originating from the overlying surface waters.

of particles originating from surface waters, which can be interpreted in terms of export production of biogenic material or rates of supply of lithogenic particles from continents. Traditionally, this information is obtained by calculating mass accumulation rates (MARs) of sedimentary constituents on the seafloor (Figure 3). Sediment cores are taken from the seafloor and sediment subsamples are dated (mostly using ^{14}C or oxygen isotope stratigraphy; see Chapter [Oxygen Isotope Composition of Seawater](#)). The dates and depths of the subsamples within the sediment column provide sedimentation rates (cm year^{-1}) between dated horizons ($= (z_{i+1} - z_i) / (t_{i+1} - t_i)$; Figure 3). Total MARs ($\text{g cm}^{-2} \text{ year}^{-1}$) are then obtained by multiplying sedimentation rates by the dry bulk density (g cm^{-3}) of the sediment (i.e., the dry weight of 1 cm^3 of wet sediment) and the accumulation rates of any sedimentary constituent can be calculated by multiplying these MARs by the weight fraction of this sedimentary constituent.

A major problem with this approach, however, is that once settling particles reach the seafloor they are often redistributed (Figure 4) either by bottom currents that erode them from their initial sites of deposition and transport them to more quiescent areas of the seafloor, or by downslope transport, if they are initially deposited on sloping surfaces (François et al., 2004). As a result, MAR on the seafloor often bears little

resemblance to the initial rain rates from the overlying surface water, which is the information that is needed to address key paleoceanographic questions such as the changes in export fluxes of biogenic materials from surface to deep water or changes in supply rates of continental dust carried by winds to the ocean surface.

^{230}Th provides a solution to this problem (Bacon, 1984). Since the rate of production of ^{230}Th per m^3 of seawater is constant and well known (eqn [8]), we can easily calculate the rate of ^{230}Th production over the entire water column overlying the seafloor per unit surface area by simply multiplying $\beta_{\text{Th-230}}$ by the water-column depth (z):

$$P_{\text{Th-230}} = \beta_{\text{Th-230}} \times z \text{ (dpm m}^{-2}\text{year}^{-1}) \quad [12]$$

We have seen that because of its very short residence time in seawater, practically all the ^{230}Th that is produced in seawater by ^{234}U decay is removed vertically by proximal scavenging into the underlying sediment. Therefore, the vertical flux of scavenged ^{230}Th reaching the seafloor ($^vF_{\text{Th-230}}$) is always nearly equal to its known rate of production in the overlying water ($P_{\text{Th-230}}$). A small fraction of ^{230}Th is removed laterally by boundary scavenging, but this can generally be neglected (Yu et al., 2001). As a result, there is a simple inverse relationship between vertical flux of settling particles reaching the seafloor (vF_p) and the ^{230}Th concentration or activity in this settling material ($^pA_{\text{Th-230}}$; Figure 5):

$$^vF_p = ^vF_{\text{Th-230}} / ^pA_{\text{Th-230}} \approx P_{\text{Th-230}} / ^pA_{\text{Th-230}} \quad [13]$$

Downcore variations in the ^{230}Th activity measured in sediment (decay corrected to the time of deposition using ^{14}C or oxygen isotopes core chronology to take into account ^{230}Th decay during burial) thus provide estimates of past changes in the vertical flux of particles that reached the seafloor and accurate information on past changes in biological productivity in the overlying surface water or supply rates of mineral dust. In instances of sediment redistribution after initial deposition, since particles and adsorbed ^{230}Th are redistributed together, the ^{230}Th activity in sediments is not affected, keeping the information on vertical rain rate intact.

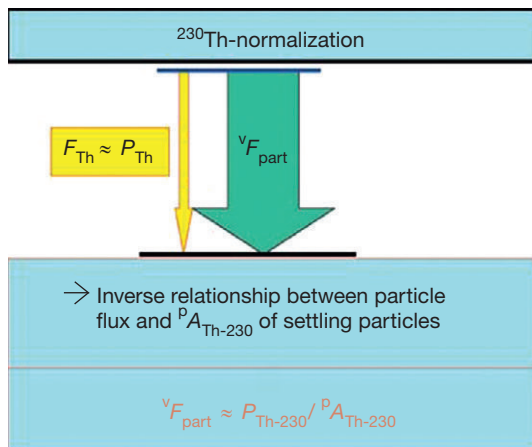


Figure 5 Principle of ^{230}Th -normalization to reconstruct past changes in vertical rain rates of particles to the seafloor (see text for explanations).

Application of this simple approach has revealed the widespread occurrence of sediment redistribution on the seafloor and the often erroneous estimates of particle rain rates based on sediment mass accumulation (François et al., 2004).

Paleoceanographic Applications Based on ^{231}Pa Activities in Marine Sediments

^{231}Pa is also produced uniformly in seawater but from the decay of ^{235}U (Table 1):

$$\beta_{\text{Pa-231}} = \lambda_{231} \times ^{\text{sw}}A_{\text{U-235}} \quad [14]$$

$$\beta_{\text{Pa-231}} = 2.17 \times 10^{-5} \text{ year}^{-1} \times 117 \text{ dpm m}^{-3} \quad [15]$$

$$\beta_{\text{Pa-231}} = 0.00247 \text{ dpm m}^{-3} \text{ year}^{-1} \quad [16]$$

Combining eqns [8] and [16], ^{231}Pa and ^{230}Th are produced at a constant ratio of

$$\beta_{\text{Pa-231}} / \beta_{\text{Th-230}} = 0.093 \quad [17]$$

Like thorium, protactinium is insoluble and is rapidly adsorbed on settling particles. It is, however, less particle reactive, resulting in a residence time in seawater that is about five times longer than that of thorium (Bacon, 1988). As a result, ^{231}Pa can be laterally transported over longer distances to be preferentially removed by boundary scavenging in regions of high particle scavenging where sediments are imparted a $^{231}\text{Pa}/^{230}\text{Th}$ ratio higher than 0.093 (Figure 6). This fractionation between the two radioisotopes during removal by scavenging and the resulting variations in sediment $^{231}\text{Pa}/^{230}\text{Th}$ ratios have seen distinct paleoceanographic applications in the Pacific and Atlantic Ocean. Differences in applications in these two oceans arise from different modes of lateral transport of ^{231}Pa .

$^{231}\text{Pa}/^{230}\text{Th}$ in Atlantic Sediments

The rate of the Atlantic meridional overturning circulation (MOC) (see Chapter [Late Pleistocene North Atlantic](#)) is the main factor controlling sediment $^{231}\text{Pa}/^{230}\text{Th}$ in the sediment deposited in the Atlantic (Yu et al., 1996; Luo et al., 2010). The Gulf Stream carries warm saline water from the subtropical regions to the northern Atlantic. As this water cools by transferring heat to the atmosphere, its density increases and it eventually sinks to produce a deep-water mass known as the North Atlantic Deep Water (NADW), which transits through the entire Atlantic basin to emerge in the Southern Ocean, from where it fills the deep basins of the Indian and Pacific Ocean (Broecker, 1991). The heat brought to high northern latitudes ameliorates the climate of northern continental masses and is implicated in the large and abrupt climate changes that characterize the Quaternary period (see Chapter [Late Pleistocene North Atlantic](#)). $^{231}\text{Pa}/^{230}\text{Th}$ in Atlantic sediments thus provides a tool to reconstruct past changes in the rate of MOC in relation to climate changes.

In the modern ocean, the residence time of ^{231}Pa in seawater is equivalent to the transit time of NADW through the Atlantic (Figure 7). As a result, nearly 50% of the ^{231}Pa produced in the NADW during its transit is exported to the

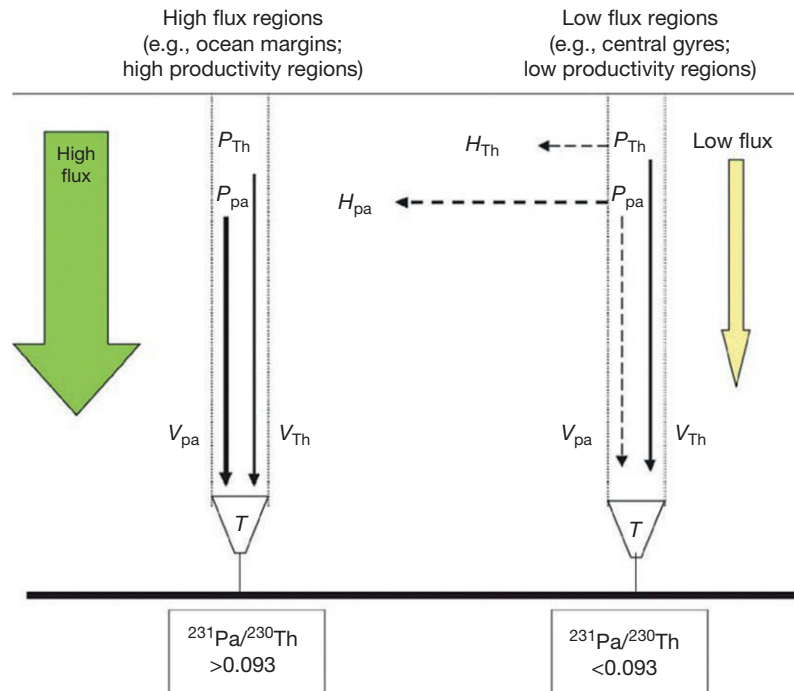


Figure 6 Fractionation between ^{230}Th and ^{231}Pa during their removal from seawater by particle scavenging. ^{230}Th and ^{231}Pa are produced at a constant rate (P_{Th} and P_{Pa}) and uniform rate ratio (0.093) in seawater. ^{230}Th has a shorter residence time in seawater limiting horizontal transport (H_{Th}). ^{230}Th is thus mainly removed by proximal scavenging and its vertical flux (V_{Th}) is always nearly equal to its production rate (P_{Th}). ^{231}Pa has a longer residence time resulting in lateral transport of ^{231}Pa (H_{Pa}) from regions with low scavenging intensity to regions of high scavenging intensity. As a result, the vertical flux of ^{231}Pa (V_{Pa}) is higher than its production rate in the overlying water column (P_{Pa}) in high flux regions, while it is lower in low flux region. Settling particles and sediments have thus a $^{231}\text{Pa}/^{230}\text{Th}$ activity ratio that is higher than the production rate ratio (0.093) in high flux regions and lower in low flux regions.

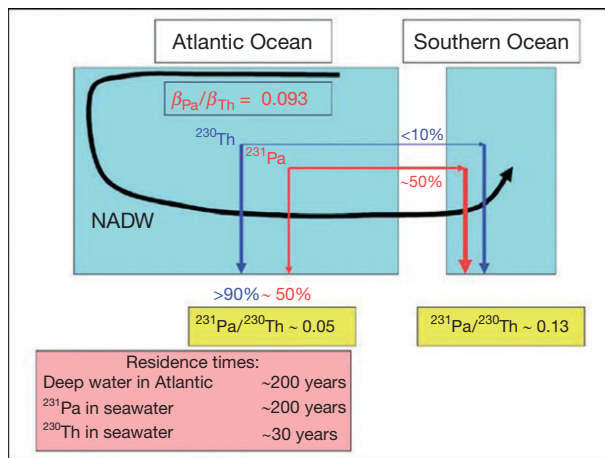


Figure 7 Principle for estimating the rate of the Atlantic meridional overturning circulation (MOC) from $^{231}\text{Pa}/^{230}\text{Th}$ in Atlantic sediments. ^{231}Pa and ^{230}Th are produced uniformly in seawater at a production rate ratio ($\beta_{\text{Pa}}/\beta_{\text{Th}}$) of 0.093. Because of its shorter residence time in the water column, nearly all the ^{230}Th produced in the North Atlantic Deep Water (NADW) during transit through the Atlantic is removed in the underlying sediment. On the other hand, since the residence time of ^{231}Pa in seawater is similar to today's residence time of deep water in the Atlantic, approximately half of the ^{231}Pa produced in the Atlantic is exported into the Southern Ocean with NADW resulting in a deficit of ^{231}Pa in Atlantic sediment. Surface sediment $^{231}\text{Pa}/^{230}\text{Th}$ increases with decreasing rate of NADW formation and export to the Southern Ocean.

Southern Ocean. On the other hand, because of its much shorter residence time, most of the ^{230}Th is removed in the Atlantic sediment. Reflecting this differential export, the mean $^{231}\text{Pa}/^{230}\text{Th}$ of Atlantic sediments (mean activity ratio ≈ 0.05) is much lower than the ratio of the rate of production of the two radioisotopes in the water column (0.093). Should the rate of NADW formation and transit have decreased in the past, the extent of ^{231}Pa export to the Southern Ocean would have also diminished, resulting in higher $^{231}\text{Pa}/^{230}\text{Th}$ in Atlantic sediments (Marchal et al., 2000; Siddall et al., 2007). The interpretation of this proxy is complicated by changes in particle flux and composition, which must also be monitored when interpreting sediment $^{231}\text{Pa}/^{230}\text{Th}$.

A first high-resolution $^{231}\text{Pa}/^{230}\text{Th}$ profile was obtained from a core retrieved from the North Atlantic. It revealed large changes in sediment $^{231}\text{Pa}/^{230}\text{Th}$ coincident with climatic variability during the last deglaciation (McManus et al., 2004; Figure 8). This record confirmed the steady formation of NADW over the entire warm Holocene period at a rate similar to present ($^{231}\text{Pa}/^{230}\text{Th} \sim 0.055$), and also documented a rapid increase in $^{231}\text{Pa}/^{230}\text{Th}$ to values approaching the production rate ratio (0.093) between 17 500 and 15 000 years BP. This was taken to indicate that MOC had nearly ceased during that period (particle fluxes decreased during that time interval and cannot account for the observation), which coincides with a massive iceberg discharge known as Heinrich Event H1 (see section 24). Deep water formation in the North Atlantic resumed abruptly at 15 000 year BP, coinciding with a rapid

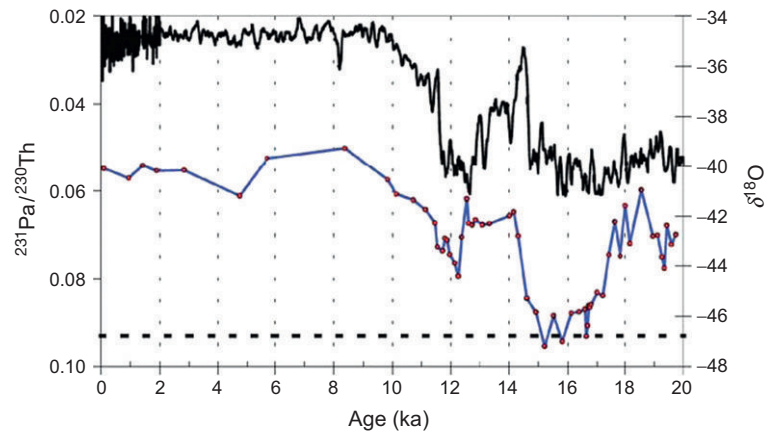


Figure 8 Changes in sediment $^{231}\text{Pa}/^{230}\text{Th}$ in the North Atlantic during the last 20,000 years (McManus et al., 2004). Top black line shows the oxygen isotopic composition of ice accumulating on Greenland and reflecting air temperature (higher $\delta^{18}\text{O}$ = warmer temperature). Bottom blue line shows $^{231}\text{Pa}/^{230}\text{Th}$ in sediment. Dashed line is the production rate ratio of ^{231}Pa and ^{230}Th in seawater (see text for discussion).

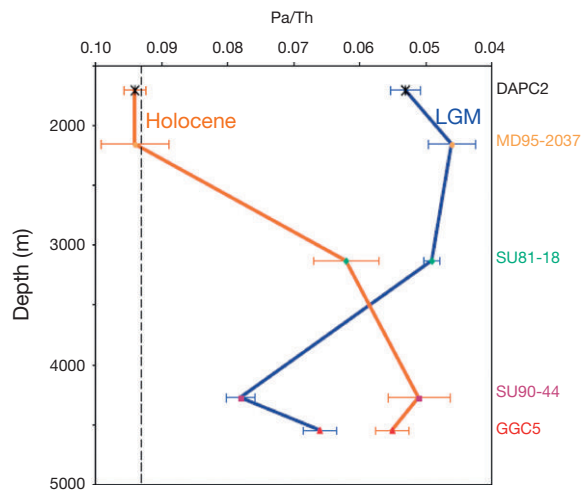


Figure 9 $^{231}\text{Pa}/^{230}\text{Th}$ measured in the Holocene and the Last Glacial Maximum section of 5 cores taken at different depths in the North Atlantic (Gherardi et al., 2009). The Holocene sections have high $^{231}\text{Pa}/^{230}\text{Th}$ at 2000 m depth and much lower ratios below, consistent with the rapid formation of deep water in the modern Atlantic Ocean. In contrast, $^{231}\text{Pa}/^{230}\text{Th}$ in the glacial sections are low at depth shallower than 3000 m and higher below, indicating that the glacial MOC was vigorous but restricted to the upper 3000 m of the water column.

warming (Bølling) and rise in sea level. The analysis of additional cores (Gherardi et al., 2009) confirmed that general evolution of the Atlantic deep water circulation during the last deglaciation. These additional data also indicated, that the MOC did not completely cease during H1. Cores taken at shallower depth indicate the presence of relatively weak but significant rate of overturning within the upper 2000m of the water column. Results from this study also showed a clear contrast in the geometry of the glacial and Holocene MOC (Fig. 9). While these data show the potential of the approach, recent modeling studies, however, also indicate that their quantitative interpretation will require a much larger database (Siddall et al., 2007; Luo et al., 2010).

$^{231}\text{Pa}/^{230}\text{Th}$ in Pacific Sediments

In the Pacific Ocean, ^{231}Pa is effectively removed by boundary scavenging, as it is transported laterally by eddy diffusion (Figure 2) from low particle flux regions in the central gyres of the ocean to high particle flux regions near continental margins and at the equator, where biological productivity is higher (Bacon, 1988). As a result, the distribution of surface sediment $^{231}\text{Pa}/^{230}\text{Th}$ mirrors the ocean color measured from satellite, which reflects the amount of chlorophyll present and the biological productivity of surface waters (Figures 10). Pacific sediment $^{231}\text{Pa}/^{230}\text{Th}$ therefore offers the possibility of mapping past changes in ocean productivity, an important factor that affects the level of atmospheric CO_2 and greenhouse warming. Downcore profiles suggest a significant decrease in biological productivity in regions of the equatorial Pacific during colder glacial periods (Pichat et al., 2004; Figure 11). Complications arise, however, in the quantitative interpretation of such profiles, as $^{231}\text{Pa}/^{230}\text{Th}$ at a given site is affected by particle flux in neighboring areas and by the chemical composition of the settling material (Chase et al., 2002). Recent studies have actually used sediment $^{231}\text{Pa}/^{230}\text{Th}$ in the sediment of the equatorial regions to assess past changes in the settling flux of opal, which has a particularly high affinity for ^{231}Pa (Bradt Miller et al., 2006).

Why is the Interpretation of Sediment $^{231}\text{Pa}/^{230}\text{Th}$ Different in the Atlantic and the Pacific?

This contrast stems from the different mechanisms that transport ^{231}Pa laterally in the two oceans (Yu et al., 2001). In the Atlantic Ocean, the lateral transport of ^{231}Pa is mostly controlled by the meridional overturning circulation (Figure 7), which rapidly 'flushes' a large fraction of the ^{231}Pa produced in the Atlantic into the Southern Ocean. In the Pacific Ocean, there is no deep-water formation at high northern latitudes. As a result, the ^{231}Pa produced in Pacific waters remains in this ocean basin and is removed in Pacific sediments. The contrast

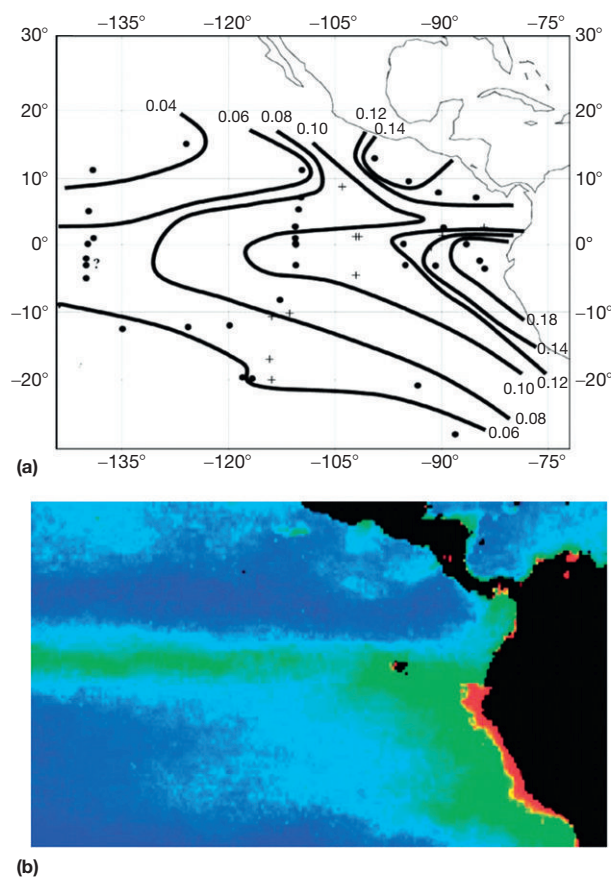


Figure 10 Correspondence between surface sediment in the Eastern Equatorial Pacific and ocean color measured by satellites.

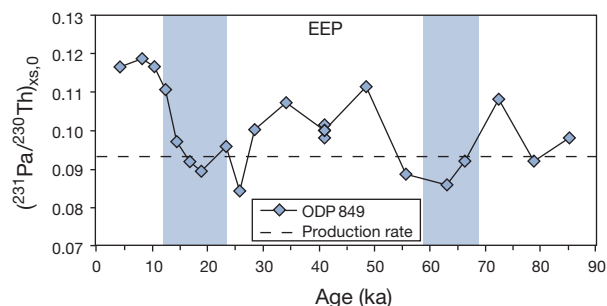


Figure 11 Sediment $^{231}\text{Pa}/^{230}\text{Th}$ in a core from the eastern equatorial Pacific (ODP 849) covering the last 85,000 years and showing lower values during the colder glacial periods highlighted in gray (Pichat et al., 2004).

in scavenging intensity between regions of low particle flux (central gyres) and high particle flux (ocean margins and equatorial upwelling) creates horizontal gradients in seawater ^{231}Pa concentration that drives a lateral flux of ^{231}Pa from low to high particle flux regions by eddy diffusion, a process known as boundary scavenging (Figure 6). Consequently, downcore changes in $^{231}\text{Pa}/^{230}\text{Th}$ recorded in Pacific sediments mostly reflect changes in the contrast in particle flux between low and high productivity regions of the Pacific. On the other hand, the effect of boundary scavenging in the Atlantic Ocean is limited.

This is because the transit time of deep water through this basin is too short for the establishment of the lateral concentration gradients needed to transport ^{231}Pa by eddy diffusion from the central gyres toward the Atlantic margins. Instead, downcore variations in $^{231}\text{Pa}/^{230}\text{Th}$ recorded in Atlantic sediments reflect mostly the rate at which ^{231}Pa is exported to the Southern Ocean as a result of the meridional overturning.

Conclusions

Radioisotopes provide important quantitative tools to address key paleoceanographic questions during the last Quaternary climatic cycle (Henderson and Anderson, 2003). In particular, ^{230}Th provides a means of accurately quantifying past changes in the rain rate of particles from surface waters, which in turn provides insights into the evolution of oceanic productivity (François et al., 2004), while $^{231}\text{Pa}/^{230}\text{Th}$ brings important constraints on past changes in ocean circulation and its impact on climate (McManus et al., 2004). Our understanding of the role of the ocean on global climate that results from the application of these versatile tools enables us to better predict the future evolution of climate in response to anthropogenic forcing.

See also: [Paleoceanography, Records: Late Pleistocene North Atlantic. Paleoclimate Reconstruction: Sub-Milankovitch \(DO/Heinrich\) Events.](#)

References

- Anderson RF, Lao Y, Broecker WS, Trumbore S, Hofmann HJ, and Wolfli W (1990) Boundary scavenging in the Pacific Ocean: A comparison of Be-10 and Pa-231. *Earth and Planetary Science Letters* 96: 287–304.
- Bacon MP (1984) Glacial to interglacial changes in carbonate and clay sedimentation in the Atlantic ocean estimated from ^{230}Th measurements. *Isotope Geoscience* 2: 97–111.
- Bacon MP (1988) Tracers of chemical scavenging in the ocean: Boundary effects and large-scale chemical fractionation. *Philosophical Transactions of the Royal Society of London A* 325: 147–160.
- Boudron B, Turner S, Henderson GM, and Lundstrom CC, Eds. (2003). Introduction to U-series geochemistry. In: *Reviews in Mineralogy and Geochemistry* vol. 52, Uranium-series Geochemistry, p. 1.
- Bradtmiller LI, Anderson RF, Fleisher MQ, and Burckle LH (2006) Diatom productivity in the equatorial Pacific Ocean from the last glacial period to the present: A test of the silicic acid leakage hypothesis. *Paleoceanography* 21: PA4201. <http://dx.doi.org/10.1029/2006PA001282>.
- Broecker WS (1991) The great ocean conveyor. *Oceanography* 4: 79–89.
- Chase Z, Anderson RF, Fleisher MQ, and Kubik PW (2002) The influence of particle composition and particle flux on scavenging of Th, Pa and Be in the ocean. *Earth and Planetary Science Letters* 204: 215–229.
- Cochran JK and Masqué P (2003) Short-lived U/Th series radionuclides in the ocean: Tracers for scavenging rates, export fluxes and particle dynamics. In: Boudron B, Turner S, Henderson GM, and Lundstrom CC (eds.) *Uranium-series Geochemistry, Reviews in Mineralogy and Geochemistry*, vol. 52, p. 461.
- François R (2007) Paleoflux and paleocirculation from sediment ^{230}Th and $^{231}\text{Pa}/^{230}\text{Th}$. In: Hillaire-Marcel C and De Vernal A (eds.) *Proxies in Late Cenozoic Paleoceanography*, ch. 16. Amsterdam: Elsevier.
- François R, Frank M, Rutgers van der Loeff MM, and Bacon MP (2004) ^{230}Th -normalization: An essential tool for interpreting sedimentary fluxes during the late Quaternary. *Paleoceanography* 19(1): PA1018. <http://dx.doi.org/10.1029/2003PA000939>.
- Gherardi J-M, Labeyrie L, Nave S, François R, McManus JF, and Cortijo E (2009) Glacial–interglacial circulation changes inferred from $^{231}\text{Pa}/^{230}\text{Th}$ sedimentary record in the North Atlantic region. *Paleoceanography* 24: PA2204. <http://dx.doi.org/10.1029/2008PA001696>.

- Henderson GM and Anderson RF (2003) U-series toolbox for paleoceanography. In: Boudron B, Turner S, Henderson GM, and Lundstrom CC (eds.) *Uranium-series Geochemistry Reviews in Mineralogy and Geochemistry*, vol. 52, p. 493.
- Henderson GM, Heinze C, Anderson RF, and Winguth AME (1999) Global distribution of the ^{230}Th flux to ocean sediments constrained by GCM modeling. *Deep-Sea Research* 46: 1861–1893.
- Luo Y, Francois R, and Allen SE (2010). Sediment $^{231}\text{Pa}/^{230}\text{Th}$ as a recorder of the rate of the Atlantic meridional overturning circulation: Insights from a 2-D model. *Ocean Sci.* 6: 381–400.
- Marchal O, Francois R, Stocker TF, and Joos F (2000) Ocean thermohaline circulation and sedimentary $^{231}\text{Pa}/^{230}\text{Th}$ ratio. *Paleoceanography* 15: 625–641.
- McManus JF, Francois R, Gherardi J, Keigwin LD, and Brown-Leger S (2004) Collapse and rapid resumption of Atlantic meridional circulation linked to deglacial climate changes. *Nature* 428: 834–837.
- Pichat S, Sims KW, Francois R, McManus JF, Brown Leger S, and Albarede F (2004) Lower biological productivity during glacial periods in the equatorial Pacific as derived from $(^{231}\text{Pa}/^{230}\text{Th})_{\text{xs},0}$ measurements in deep-sea sediments. *Paleoceanography* 19: PA4023. <http://dx.doi.org/10.1029/2003PA000994>.
- Siddall M, Anderson RF, Winckler G, et al. (2008) *Paleoceanography* 23: PA2208. <http://dx.doi.org/10.1029/2007PA001556>.
- Siddall M, Stocker TF, Henderson GM, et al. (2007) Modelling the relationship between $^{231}\text{Pa}/^{230}\text{Th}$ distribution in North Atlantic sediment and Atlantic meridional overturning circulation. *Paleoceanography* 22: PA2214. <http://dx.doi.org/10.1029/2006PA001358>.
- Yu E-F, Francois R, and Bacon MP (1996) Similar rates of modern and last-glacial ocean thermohaline circulation inferred from radiochemical data. *Nature* 379: 689–694.
- Yu E-F, Francois R, Bacon MP, and Fleer AP (2001) Fluxes of ^{230}Th and ^{231}Pa to the deep sea: Implications for the interpretation of excess ^{230}Th and $^{231}\text{Pa}/^{230}\text{Th}$ profiles in sediments. *Earth and Planetary Science Letters* 191: 219–230.

Salinity Proxies $\delta^{18}\text{O}$

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Introduction

Temperature and salinity are two of the key properties of ocean water masses. The distribution of these two independent but related characteristics reflects the interplay of incoming solar radiation (insolation) and the uneven distribution of heat loss and gain by the ocean, with that of precipitation, evaporation, and the freezing and melting of ice. Temperature and salinity, to a large extent, determine the density of a parcel of water. Small differences in temperature and salinity can increase or decrease the density of a water parcel, which can lead to convection. Once removed from the surface of the ocean where 'local' changes in temperature and salinity can occur, the water parcel retains its distinct relationship between (potential) temperature and salinity. One can take advantage of this 'conservative' behavior where changes only occur as a result of mixing processes, to track the movement of water in the deep ocean (Figure 1). The distribution of density in the ocean is directly related to horizontal pressure gradients and thus geostrophic ocean currents (currents resulting from the balance between gravitational forces and the Coriolis effect). During the Quaternary when we have had systematic growth and decay of large land-based ice sheets, salinity has had to change. A quick scaling argument following that of Broecker and Peng (1982) is as follows: the modern ocean has a mean salinity of 34.7 psu (practical salinity units) and is on average 3500 m deep. During glacial maxima, sea level was on the order of ~ 120 m lower than present. Simple scaling of the loss of freshwater (3–4%) requires an average increase in salinity of a similar percentage or to ~ 35.9 psu. Because much of the deep ocean is of similar temperature, small changes in salinity have a large impact on density, yielding a potentially different distribution of water masses and control of the density-driven (thermohaline) ocean circulation. It is partly for this reason that reconstructions of past salinity are of interest to paleoceanographers.

What is salinity? At the simplest of all levels, salinity is the total amount of dissolved materials in seawater. It is not practical to evaporate seawater samples to dryness to determine the total dissolved materials, and some materials such as chlorides are lost during drying. A more complete definition was created by the International Council for the Exploration of the Sea, whereby salinity was defined as the 'total amount of solid materials in grams dissolved in 1 kg of seawater when all the carbonate has been converted to oxide, the bromine and iodine replaced by chlorine and all organic matter completely oxidized.' Again, this is a difficult measurement to make accurately. Taking advantage of the near-constant proportionality of the major salts in seawater, salinity was 'redefined' using chlorinity and a simple linear relation $S = 1.80655 \times \text{Cl}$, where chlorinity is 'the amount of silver required to precipitate completely the halogens in 0.3285234 kg of seawater' (Wooster et al., 1969). At about the same time, a relationship between salinity and conductivity was

also proposed. Conductivity measurements are relatively easy compared to the chemistry-based techniques and are accurate. The relationship between salinity and conductivity has been revised slightly, such that the practical salinity scale of 1978 is now the official definition, and knowing the conductivity, pressure, and *in situ* temperature allows salinity to be calculated (JPOTS, 1980; Millero, 1996). Surface salinities reflect the precipitation–evaporation freshwater balance and span the full range of potential values from 0 psu in freshwater rivers and coastal environments to 42 psu in regions of high evaporation such as the Red Sea. Interior ocean salinities do not vary as much: ~ 33.8 psu (Antarctic intermediate water) to ~ 36.5 psu (North Atlantic central water), with abyssal waters in general differing by less than 1 psu (Figure 2).

Reconstructing Past Salinity Variations (with an Emphasis on the Quaternary)

As paleoceanographers, there are several different approaches that can be utilized to reconstruct past variations in salinity. These can be grossly classified as 'biological' and 'chemical.' In the purely biological approach, one relates the distribution and relative abundance of certain species to salinity and makes an assumption that this empirical relationship has been the same in the past. The salinity gradients in the deep ocean are very small, so this approach has been mainly used to attempt to reconstruct surface salinity variations in the open ocean and changes in estuarine or coastal environments. A classic example of this approach is the work of Climate: Long-Range Investigation, Mapping, and Prediction researchers (e.g., Kipp, 1976) who related the distribution of planktonic Foraminifera in core-top sediments to seasonal and annual salinity. In this example, the relative abundances of the (26) species of planktonic Foraminifera are deconstructed using Q-mode factor analysis (e.g., Imbrie and Kipp, 1971) into major modes or factors which are then related to surface salinity observations using a multiple linear regression scheme. The empirically derived relationship between the major factors and salinity is then used as a model to estimate past surface salinity from planktonic Foraminifera census counts in sediment samples taken downcore and back through time (Figure 3).

How might one otherwise reconstruct past salinity variations and take advantage of other 'conventional' biogeochemical proxies? Are there relationships between elements or isotopes in the skeletons of, for example, carbonate fossils and salinity? Can one directly measure past salinity variations?

'Current' Methods of Reconstructing Past Salinity

Current methods of reconstructing past ocean salinity focus on using either direct measurements of past $\delta^{18}\text{O}$ water

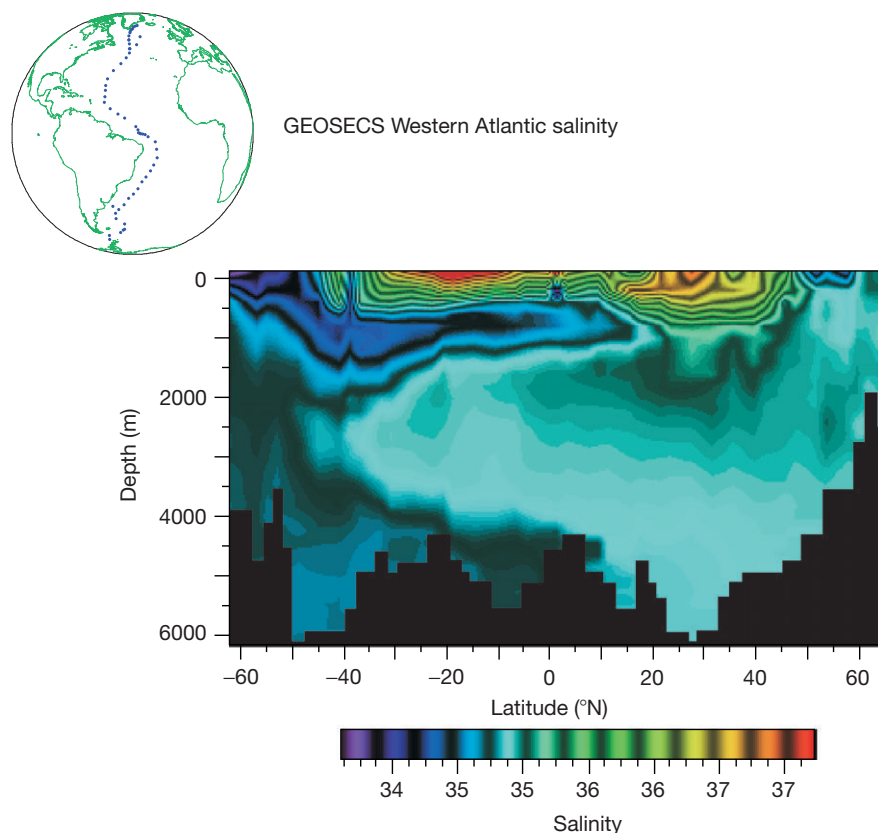


Figure 1 Western Atlantic salinity (psu) profile as a function of latitude and depth (m) observed during the GEOSECS expedition.

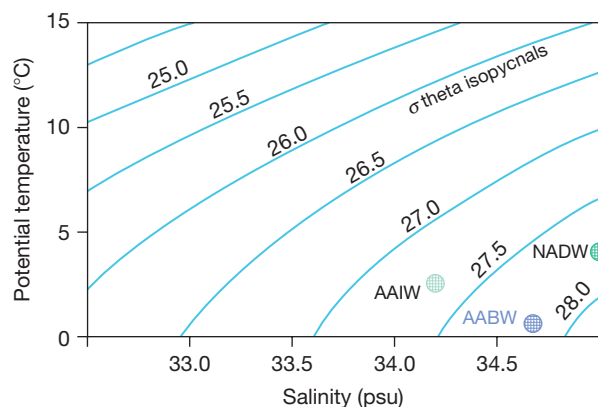


Figure 2 Relationship between potential temperature ($^{\circ}\text{C}$), salinity (psu), and seawater potential density (σ_{θ} : kg m^{-3}). Representative values for the three main interior water masses in the Atlantic Basin are shown: Antarctic intermediate water (AAIW), Antarctic bottom water (AABW), and North Atlantic deep water (NADW).

(e.g., glacial pore-water $\delta^{18}\text{O}$) or independent paleotemperature proxies (e.g., alkenones, foraminiferal Mg/Ca , coral Sr/Ca , and faunal assemblages) to extract the past $\delta^{18}\text{O}$ water signal from the $\delta^{18}\text{O}$ carbonate record (e.g., Malaize and Caley, 2009). Pore-water chlorinity measurements can be used to directly estimate glacial salinity. The indirect methods using biogeochemical proxies have the advantage over the direct pore-water estimates by facilitating time series and having the

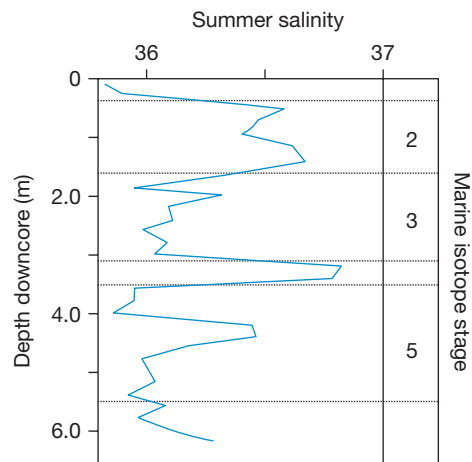


Figure 3 Transfer function-based surface salinity estimate from the Columbian Basin through marine isotope stage 6. Left vertical axis is depth downcore (meters) and corresponding carbonate (isotope) stage designations are provided. Reproduced from [Prell WL and Hays JD \(1976\)](#) late Pleistocene faunal and temperature patterns of the Colombia Basin, Caribbean Sea. *Geological Society of America Memoir 145*, pp. 201–220.

potential to document so-called abrupt climate change events at the submillennial to seasonal timescale. The reconstructed $\delta^{18}\text{O}$ water record can then be deconvolved into its ice volume and salinity components. For the sake of this article, we focus

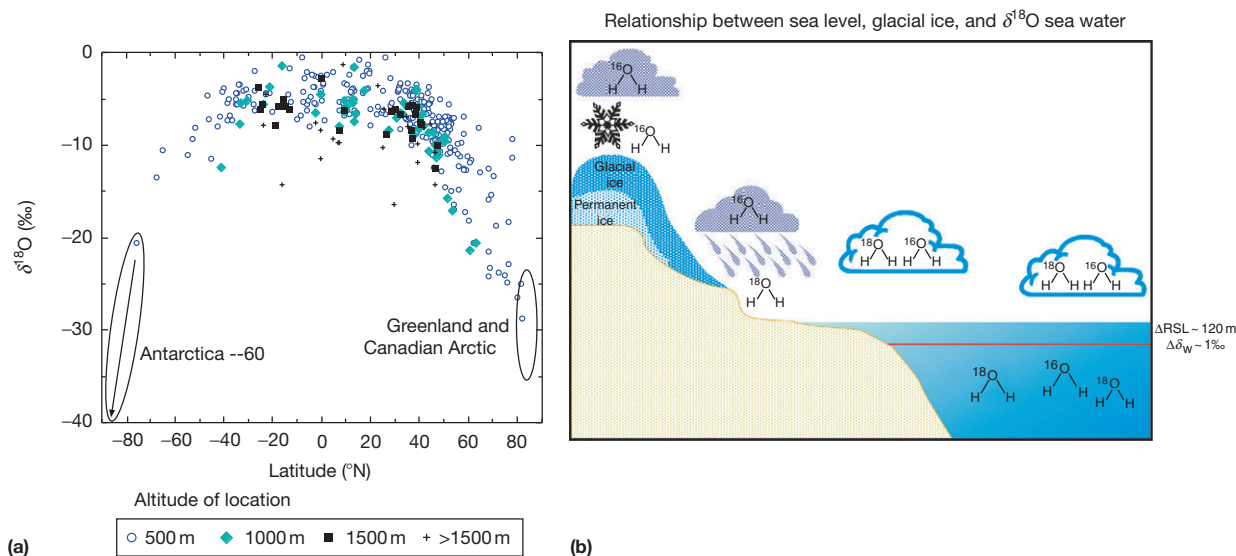


Figure 4 (a) Mean annual $\delta^{18}\text{O}$ of precipitation as a function of latitude (°N) and altitude, an update using the IAEA ISOHIS database to figures originally reported in, for example, Rozanski et al. (1993) among others. (b) Cartoon schematic of the effect of a growing ice sheet on mean ocean $\delta^{18}\text{O}$.

on open ocean salinity changes, although additional methods of reconstructing estuarine and brackish water salinities are used. For example, the large salinity changes associated with brackish environments allow the use of faunal assemblage changes as direct salinity proxies, whereas in open ocean environments salinity changes are generally too subtle to cause large faunal changes. We focus particularly on combining Mg/Ca and $\delta^{18}\text{O}$ measurements in marine carbonates as a means to reconstruct open ocean glacial–interglacial $\delta^{18}\text{O}$ water (salinity) estimates. An analogous technique applied to reef-building hermatypic corals is the combination of Sr/Ca and $\delta^{18}\text{O}$ (cf. McCulloch et al., 1994; see corals, sclerosponges, and mollusks).

One begins with the consideration of the workhorse of paleoceanography: $\delta^{18}\text{O}$ thermometry. This is done because the direct estimates of salinity during the last glaciation provide a constraint on one of the primary factors in the $\delta^{18}\text{O}$ of marine carbonate, and thus, there is a need to have a common basis of understanding. The ratio of $^{18}\text{O}:^{16}\text{O}$ (expressed as $\delta^{18}\text{O}$) in biogenic carbonates has been exploited as a geochemical paleothermometer, since the pioneering work of Urey (1947) and Epstein et al. (1953 and references therein). Weber and Woodhead (1972) documented that the oxygen isotope temperature sensitivity of reef-building hermatypic corals is similar to that of the oxygen paleotemperature scale developed for mollusks (e.g., Epstein et al., 1953 and references therein), $\sim 0.21\text{‰}$ per °C. Similar comparisons have been made using Foraminifera (e.g., Bemis et al., 1998; Shackleton, 1974, among many). Thus, biogenic carbonate $\delta^{18}\text{O}$ is a function of temperature and variations in the oxygen isotopic composition of water (δw). To be complete, carbonate ion content and pH also exert a small effect (Spero et al., 1997). For this discussion, it is assumed that this is a minor effect. Most marine biogenic carbonate materials have a temperature sensitivity similar to that of inorganic carbonate.

So, what controls the oxygen isotopic composition of seawater, and does this relate back to salinity? On long geologic

timescales, water–rock interactions control the mean isotopic composition. On glacial–interglacial timescales, ^{18}O -depleted water is locked up in ice sheets leaving the ocean more enriched in ^{18}O . In a gross oversimplification, ^{18}O is heavier than its more abundant ^{16}O cousin, and as moisture travels from the evaporating sources of the ocean to the growing ice sheet, the ^{18}O rains out in precipitation, leaving the remaining vapor depleted or more negative (cf. Craig and Gordon, 1965; Rozanski et al., 1993; Figure 4); this is in addition to the lower vapor pressure of ^{18}O which yields a more negative (fewer H_2^{18}O than H_2^{16}O compared to the water source) vapor phase to begin with. Thus, during glacial periods when this high-latitude precipitation is locked up in ice sheets, the ocean has more ^{18}O in it than during interglacials or, if preferred, less ^{16}O , which leaves the ocean relatively ‘enriched’ in ^{18}O with a corresponding more positive $\delta^{18}\text{O}$ water. Even without growing ice sheets, high-latitude (and altitude) precipitation has a more negative $\delta^{18}\text{O}$ value, and as this water enters the ocean as freshwater with zero salinity, it yields a relationship between salinity and $\delta^{18}\text{O}$ water. Craig and Gordon (1965) proffered the first oceanwide or global surface salinity– $\delta^{18}\text{O}$ water relationship: $\sim 0.6\text{‰}$ for every salinity unit (Figure 5(a)). In actuality, the whole ocean salinity– $\delta^{18}\text{O}$ water relationship is the result of a series of regional relationships (Figure 5(b) and 5(c)) that reflect the local precipitation–evaporation balance, the input of river or glacial meltwater, and the ocean dynamics (e.g., Fairbanks et al., 1992; Schmidt et al., 1999). Local/regional surface water $\delta^{18}\text{O}$ water–salinity relations range by an order of magnitude from $\sim 0.2\text{‰}$ to $\sim 1.5\text{‰}$ per psu (Figure 5(c)). The deep ocean $\delta^{18}\text{O}$ water–salinity relationship is on the order of $\sim 1.5\text{‰}$ per psu (e.g., Craig and Gordon, 1965).

If one can independently determine or estimate temperature and the eustatic (global) ice volume component, these can be converted into $\delta^{18}\text{O}$ space and subtracted from the measured (biogenic) $\delta^{18}\text{O}$ carbonate. The residual $\delta^{18}\text{O}$ is an estimate of regional $\delta^{18}\text{O}$ water (salinity) variations. Implicitly, it

sediment pore waters and determined that the mean ocean was $\sim 1\text{‰}$ heavier during the Last Glacial Maximum. Over time via diffusive processes, pore-water $\delta^{18}\text{O}$ equilibrates with that of the overlying deep ocean. Thus, the change from full glacial conditions ~ 20 ka to the present interglacial is 'remembered' in the sediments. Combining measurements with a pore-water diffusion model allows the glacial condition to be reconstructed from the current pore-water $\delta^{18}\text{O}$ water profile. To do so, they compressed or squeezed the pore waters out of sediment and analyzed their oxygen isotopic composition. These results provide a direct measure of removing the whole ocean $\delta^{18}\text{O}$ water from marine carbonate $\delta^{18}\text{O}$ analyses. This change can then be partitioned as a function of sea level (time) over the last glacial cycle in a fashion similar to that proposed by Fairbanks and Matthews (1978) and Fairbanks (1989). It is generally assumed that this relation holds for previous Pleistocene glacial cycles and is insensitive to the oxygen isotope composition of Northern versus Southern Hemisphere ice sheets. Adkins et al. (2002) performed a more difficult suite of pore-water chlorinity measurements, thus providing for the first time a direct measure of salinity of the deep ocean during the Last Glacial Maximum. Adkins et al. confirmed that the deep ocean was significantly saltier 20 ka and that the deep ocean was significantly colder than present (Figure 6). Astonishingly, Adkins et al. measured waters from the deep Southern and Pacific oceans that are significantly more salty than our whole ocean glacial estimate of 35.9 psu. They determined that, unlike today when the saltiest deep water is formed in the North Atlantic, the Southern Ocean was saltier by a large margin. Adkins et al. correctly infer that brine rejection during sea ice formation played a much more dominant role in deep-water production. A corollary to this observation is that the pole-to-pole freshwater budget must

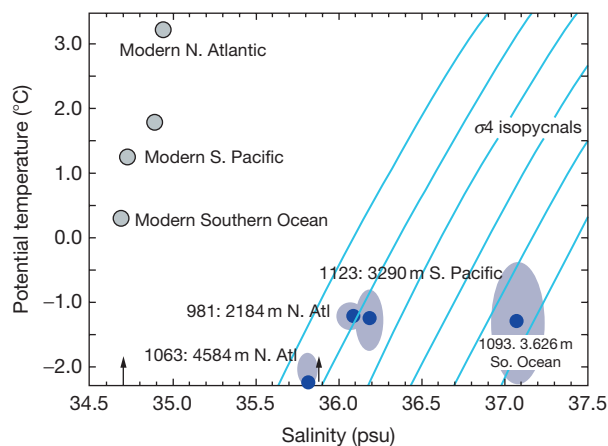


Figure 6 Last Glacial Maximum salinity (psu) and potential temperature ($^{\circ}\text{C}$) as estimated from pore-water chlorinity, $\delta^{18}\text{O}$ pore waters, and benthic foraminiferal $\delta^{18}\text{O}$ at four ODP core sites. The stippled regions represent propagated error ranges, and curved lines denote lines of potential density referenced to 4000 dbar (σ_4) during the Last Glacial Maximum. The arrows demark mean ocean salinity for present (~ 34.7) and during the glaciation (~ 35.9). Present-day values/estimates for the core locations are also shown (filled circles). Error ellipses for glacial values are demarked as stippled pattern. Redrawn from Adkins JF, McIntyre K, and Schrag DP (2002) The salinity, temperature, and ^{18}O of the glacial deep ocean. *Science* 298: 1789–1773.

have been very different than today. A very strong constraint on the mechanisms of formation and redistribution of deep water during the last glaciation would be a systematic basin-wide pore-water study similar to Adkins et al. to accurately constrain the geometry of glacial deep waters. This direct reconstruction of paleosalinity (through chlorinity) removes assumptions and complications due to nutrient chemistry implicit in other biogeochemical proxies such as $\delta^{13}\text{C}$ (Curry and Oppo, 2005 and references therein).

Combined Mg/Ca and $\delta^{18}\text{O}$

At present, the combination of Mg/Ca and $\delta^{18}\text{O}$ of Foraminifera (and ostracods) has the greatest potential for reconstruction of Quaternary salinity changes, and this combined method is beginning to be widely applied. Foraminiferal Mg/Ca provides an independent temperature estimate that can be used to extract the $\delta^{18}\text{O}$ water signal from the $\delta^{18}\text{O}$ carbonate record. Although there are a number of commonly used paleotemperature proxies, including alkenones, foraminiferal faunal assemblages, Mg/Ca of Foraminifera, and Sr/Ca of corals, all of which have their own advantages and disadvantages, as a salinity proxy, the combination of Mg/Ca and $\delta^{18}\text{O}$ has the advantage that both analyses are conducted on the same material (foraminiferal calcite), thus alleviating issues related to differences in seasonality, habitat, and ecology.

Mg/Ca Paleothermometry

Although the relationship between Mg and temperature in calcite has been recognized for many years, Mg paleothermometry did not come widely into use until the 1990s. Several early studies documented that magnesium was higher in Foraminifera shells precipitated in warmer waters (Bender et al., 1975; Savin and Douglas, 1973), although initial concerns about secondary effects precluded wide acceptance of the Mg/Ca proxy. However, studies of inorganic calcite precipitation demonstrated that magnesium incorporation is largely controlled by thermodynamics and indicated that Mg substitution into the calcite lattice takes an exponential form that can be described by the equation

$$\text{Mg/Ca} = b e^{mT}$$

where b is the preexponential constant, m is the exponential constant, and T is the temperature (Katz, 1973; Oomori et al., 1987).

Mg/Ca Temperature Calibration Equations

Since these early studies, work has focused on developing Mg temperature calibration equations for biogenic calcites (benthic and planktonic Foraminifera, ostracods, and mollusks) using culturing, sediment trap, and core-top material. Calibration equations derived from live culturing experiments of *Globigerinoides sacculifer*, *Orbulina universa*, *Globigerina bulloides*, and *Neoglobobulimina pachyderma* take an exponential form and indicate that these Foraminifera species all incorporate Mg at a rate of $\sim 10\%$ per $^{\circ}\text{C}$ (exponential constants vary between 0.085 and 0.102 (e.g., Lea et al., 1999; Mashiotto et al., 1999; Nürnberg

et al., 1996); and also varies in few planktonic foraminifer species (Eggins et al., 2003)). The pre-exponential constant, which determines the absolute temperature, ranges from 0.3 to 1.36, although all cultured species except *O. universa* fall in the range of 0.3–0.5. The reason for the large difference between *O. universa* and the other planktonic species is unknown, although *O. universa* is morphologically unique in its formation of a final spherical chamber. Mg paleothermometry errors are on the order of 1.0–1.4 °C. Core-top and sediment trap studies are in good agreement with the culturing equations, and these studies have extended the calibration equations to additional planktonic species (Anand et al., 2003; Elderfield and Ganssen, 2000). Although both tropical and temperate species have similar exponential constants, tropical species have lower preexponential constants than temperate and subpolar species (Figure 7).

The development of benthic foraminiferal calibration equations has lagged behind that of planktonic Foraminifera, in part due to the difficulty in culturing these species, and the acquisition of live (e.g., rose bengal stained) benthic Foraminifera from multicores or box cores with good sediment–water interfaces. Rosenthal et al. (1997) defined a calibration equation for the

epifaunal genus *Cibicidoides* between 5 °C and 18 °C, which was extended to bottom water temperatures (<3 °C) by Martin et al. (2002), yielding an exponential equation of

$$\text{Mg/Ca} = 0.85.11\text{BWT}$$

The standard error derived from the data set is 1.14 °C. A study of Mg/Ca in other species of benthic Foraminifera indicated that although several common species yielded results similar to that of *Cibicidoides*, the paleoceanographically important taxon *Uvigerina* apparently has a lower temperature sensitivity (Lear et al., 2002; Martin et al., 2002).

Mg/Ca paleothermometry has also been successfully applied to marine ostracods (Cronin et al., 2000; Dwyer et al., 1995). These small, benthic crustaceans secrete a calcite shell that incorporates Mg/Ca very similarly to Foraminifera (~9% per °C), although linear calibration equations have been applied (Dwyer et al., 1995).

Secondary Effects

Because paleoceanographers are interested in isolating small temperature differences, it is important to accurately identify temperature changes. In addition to the temperature effect, the most prominent secondary effect on Mg/Ca in biogenic calcites is dissolution. Mg/Ca in shell calcite decreases with increasing dissolution on the seafloor. The cause of Mg loss is as yet uncertain but may be due to uneven distribution of Mg in shells and higher solubility of Mg-rich shell components. For instance, studies have shown that species with gametogenic calcite crusts or certain types of shell ornamentation, such as *N. dutertrei* or *G. tumida*, respectively, are more susceptible to Mg loss (Brown and Elderfield, 1996; Rosenthal et al., 2000). In order to have confidence in the Mg/Ca proxy, it is necessary to either analyze only pristine shells, or quantify the amount of dissolution and its effect on the Mg/Ca ratio. Dekens et al. (2002) analyzed Mg/Ca in three different planktonic species along Pacific and Atlantic depth transects in order to quantify the percentage of Mg loss with depth increase, and developed dissolution-corrected calibration equations. For instance, for Pacific *G. ruber*, the calibration equation is adjusted by 5% per kilometer depth. An alternative approach is to estimate the amount of postdepositional dissolution by measuring average shell masses and normalizing Mg/Ca ratios accordingly (Rosenthal and Lohmann, 2002).

Foraminiferal culturing experiments have been performed to quantify the effects of salinity and pH on Mg/Ca. These studies indicate that although temperature is the dominant control, both salinity and pH exert a minor influence on shell Mg/Ca. For *G. bulloides* and *O. universa*, shell Mg/Ca decreases 6% per 0.1 pH unit. Salinity exerts a smaller influence of ~4‰ per psu on *O. universa* (Lea et al., 1999 and references therein).

What Are the Errors for Past Salinity Estimates Using Mg/Ca and $\delta^{18}\text{O}$?

It is important to recall that $\delta^{18}\text{O}$ carbonate has a temperature sensitivity on the order of ~0.21‰ per °C and that Mg/Ca temperature estimates have an error of at least 1.1–1.4 °C, that is, the Mg/Ca temperature error alone contributes at

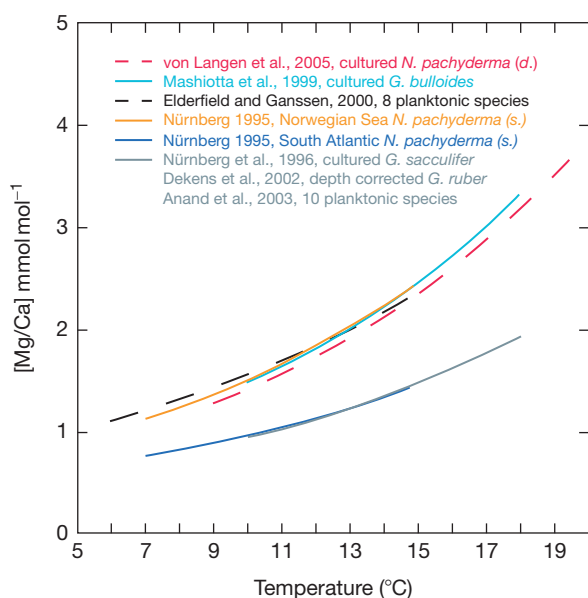


Figure 7 A composite of planktonic Foraminifera temperature (°C) versus Mg/Ca (mmol mol⁻¹) 'calibrations.' Data from von Langen et al. (2005), cultured *N. pachyderma* (d.); Mashiotta TA, Lea DW, and Spero HJ (1999) Glacial–interglacial changes in Subantarctic sea surface temperature and $\delta^{18}\text{O}$ water using foraminiferal Mg. *Earth and Planetary Science Letters* 170: 417–432; Elderfield H and Ganssen G (2000) Past temperature and $\delta^{18}\text{O}$ of surface ocean waters inferred from foraminiferal Mg/Ca ratios. *Nature* 400: 442–445; Nürnberg (1995), Norwegian Sea *N. pachyderma* (s.); Nürnberg (1995), South Atlantic *N. pachyderma* (s.); Nürnberg D, et al. (1996) Assessing the reliability of magnesium in foraminiferal calcite as a proxy for water mass temperatures. *Geochimica et Cosmochimica Acta* 60: 803–814; Dekens P, Lea DW, Pak DK, and Spero HJ (2002) A core-top calibration of Mg/Ca in tropical foraminifera: Refining paleo-temperature estimation. *Geochimica Geophysica Geosystems* 8; Anand P, Elderfield H, and Conte MH (2003) Calibration of Mg/Ca thermometry in planktonic foraminifera from a sediment trap time series. *Paleoceanography* 18(2): 1050. <http://dx.doi.org/10.1029/2002PA000846>, 002003.

least $\pm 0.2\text{‰}$ error to our estimate of past $\delta^{18}\text{O}$ water. There is the implicit assumption that the observed salinity– $\delta^{18}\text{O}$ water relations are constant through time. Regions with steep salinity– $\delta^{18}\text{O}$ water relationships have correspondingly smaller errors in their salinity estimate. In regions that have a well-constrained salinity– $\delta^{18}\text{O}$ water relation, there is the potential to reconstruct salinity to on the order of ± 0.3 – 0.4 psu. Assumptions on whole ocean $\delta^{18}\text{O}$ water changes during the Quaternary add at least another 0.1‰ $\delta^{18}\text{O}$ water error. Thus, errors of past $\delta^{18}\text{O}$ water–salinity estimates vary in space and time (e.g., Rohling, 2007). In spite of limitations in estimating salinity from $\delta^{18}\text{O}$ water, it would be possible to reconstruct the monsoon variability based on the $\delta^{18}\text{O}$ water and salinity by using paired measurements of $\delta^{18}\text{O}$ and Mg/Ca in planktonic foraminifer species from the Arabian Sea (Govil and Naidu, 2010) and Atlantic Ocean (Weldeab et al., 2007).

Quaternary $\delta^{18}\text{O}$ Water Variations: An Example from the Tropical Pacific

As an example of the combined Mg/Ca and $\delta^{18}\text{O}$ technique, the work of Lea et al. (2002) from the equatorial Pacific is highlighted. The focus is first on TR163-19 (2°N , 91°W) near the Galapagos in the eastern Pacific (Figure 8). Oxygen isotope measurements of the surface-dwelling planktonic

Foraminifera *G. ruber* exhibit the characteristic ‘sawtooth’ pattern observed in most if not all Quaternary $\delta^{18}\text{O}$ profiles. The glacial–interglacial difference in $\delta^{18}\text{O}$ carbonate is $\sim 1.8\text{‰}$. Mg/Ca concentrations on the same species of planktonic Foraminifera are interpreted to reflect a glacial–interglacial SST difference on the order of $\sim 2.6^\circ\text{C}$. If we strictly interpret the Mg/Ca results as SST and convert the temperature difference into $\delta^{18}\text{O}$ space, we can then subtract the ‘SST’ signature from the observed $\delta^{18}\text{O}$ carbonate. This yields a predicted $\delta^{18}\text{O}$ water time series that includes glacial (eustatic) and local/regional $\delta^{18}\text{O}$ water differences. The difference during the Last Glacial Maximum is $\sim 1.2\text{‰}$ which, if the whole ocean were heavier by $\sim 1\text{‰}$, leaves $\sim 0.2\text{‰}$. If the glacial salinity– $\delta^{18}\text{O}$ water relationship were similar to today’s, this difference implies an ~ 1 psu increase (i.e., more salty), either by reduced rainfall or by a stronger influence of subtropical surface waters brought to the equator in the north equatorial current. Given the location of the core relative to the modern-day intertropical convergence zone (ITCZ), a southward displacement of the ITCZ would allow for more subtropical water to gain influence over the site. Clearly apparent in the time series is the fact that tropical SST leads $\delta^{18}\text{O}$ water by ~ 3 ka, particularly on transitions to interglacial periods.

Because of questions regarding the exact history of eustatic (glacial) $\delta^{18}\text{O}$ water and the lack of a standardized curve (where at least every use would be working with the same record), Lea et al. (2002) like many others do not explicitly

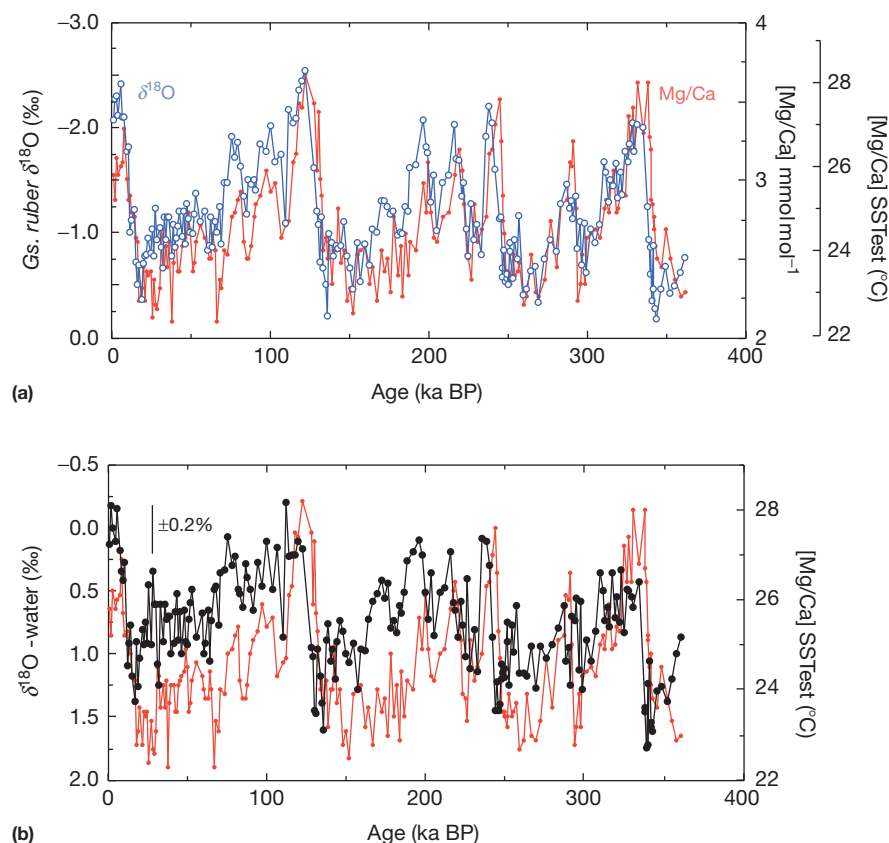


Figure 8 (a) TR163-19, eastern equatorial Pacific downcore *G. ruber* $\delta^{18}\text{O}$ (‰) – blue and Mg/Ca (mmol mol⁻¹) and SST (°C) estimate – red. (b) Deconvolved $\delta^{18}\text{O}$ water (‰) – blue and Mg/Ca-SST (°C) – red using the data of (a). Data from Lea et al. (2002).

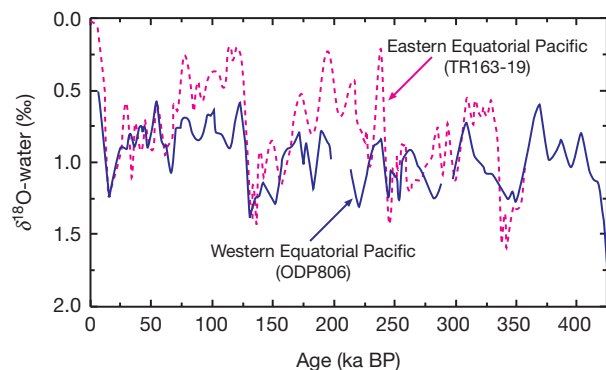


Figure 9 A comparison of Quaternary (0–400 ka BP) planktonic Foraminifera surface $\delta^{18}\text{O}$ water (‰) reconstructions from the western (ODP 806; solid blue) and eastern (TR163-19; dotted red) equatorial Pacific. To minimize noise, the individual data were passed through a triangle filter, and data have not had a sea level $\delta^{18}\text{O}$ correction applied. Core-top $\delta^{18}\text{O}$ water estimates are consistent with modern-day observations. Redrawn from Lea DW, Martin PA, Pak DK, and Spero HJ (2002) Reconstructing a 350 ka history of sea level using planktonic Mg/Ca and oxygen isotope records from a Cocos Ridge core. *Quaternary Science Review* 21: 283–293.

subtract out an ice volume component. Instead, Lea et al. (2002) use a 'site-to-site' comparison with ODP 806, a core from the western equatorial (0° , 159°E) Pacific warm pool (Figure 9). The site-to-site comparison assumes that the ice volume signal is common in each record, and therefore, differences between the two $\delta^{18}\text{O}$ water records are a reflection of changes in local salinity ($\delta^{18}\text{O}$ water). In the western Pacific the $\delta^{18}\text{O}$ water glacial–interglacial amplitude is only $\sim 0.75\text{‰}$ and less than that expected due to ice volume. This observation requires that the surface waters overlying ODP 806 had a more negative $\delta^{18}\text{O}$ water than today and were likely fresher or less saline. In combination with the results from TR163-169, the ODP 806 results indicate a different surface water $\delta^{18}\text{O}$ water gradient in the tropical Pacific during glacials with the east being more positive (saltier) and the west being more negative (fresher). An additional feature in both records, and especially in that of ODP 806, is what appears to be a long-term secular trend in $\delta^{18}\text{O}$ water on the order of $\sim 0.1\text{‰}$ per ka. If not an artifact, such a trend could indicate a larger glacial sea level drop earlier in the record (i.e., marine isotope stage 12) and/or a change in the mean $\delta^{18}\text{O}$ of glacial ice.

Where do $\delta^{18}\text{O}$ water–salinity reconstructions stand? The signals that one would like to directly interpret are small, and there have been inconsistencies when one study applies a calibration from one lab to another set of results or species measured by a different technique. The observation of interlaboratory offsets has necessitated standardization of Mg/Ca techniques (cf. Rosenthal and Lohmann, 2002; Lea et al., 2002). Consistently applied error propagation is necessary so that one can gauge the validity of any single downcore result: ideally, we want to be able to interpret submillennial variability when the sedimentary-depositional environment allows us to. To remove the ice volume component from the estimated $\delta^{18}\text{O}$ water requires, at the very least, a common internationally agreed upon standard Quaternary sea level record and ice volume ($\delta^{18}\text{O}$) scaling.

See also: **Paleoceanography, Biological Proxies:** Benthic Foraminifera; Planktic Foraminifera; Temperature Proxies, Census Counts.

References

- Adkins JF, McIntyre K, and Schrag DP (2002) The salinity, temperature, and ^{18}O of the glacial deep ocean. *Science* 298: 1769–1783.
- Anand P, Elderfield H, and Conte MH (2003) Calibration of Mg/Ca thermometry in planktonic foraminifera from a sediment trap time series. *Paleoceanography* 18(2): 1050. <http://dx.doi.org/10.1029/2002PA000846>, 002003.
- Bemis BE, Spero HJ, Bijma J, and Lea DW (1998) Reevaluation of the oxygen isotopic composition of planktonic foraminifera: Experimental results and revised paleotemperature equations. *Paleoceanography* 13(2): 150–160.
- Bender ML, Lores RB, and Williams DF (1975) Sodium, magnesium, and strontium in the tests of planktonic foraminifera. *Micropaleontology* 21: 448–459.
- Broecker WS and Peng T-H (1982) *Tracers in the Sea*. Palisades, NY: Eldigio Press.
- Brown S and Elderfield H (1996) Variations in Mg/Ca and Sr/Ca ratios of planktonic foraminifera caused by postdepositional dissolution – Evidence of shallow Mg-dependent dissolution. *Paleoceanography* 11: 543–551.
- Craig H and Gordon LI (1965) Deuterium and oxygen-18 variations in the ocean and the marine atmosphere. In: Tongiorgi E (ed.) *Stable Isotopes in Oceanographic Studies and Paleotemperatures, Soletto, 26–27 July 1965*, pp. 1–22. Pisa: Consiglio Nazionale delle Ricerche, Laboratorio di Geologia Nucleare.
- Cronin TM, Dwyer GS, Baker PA, Rodriguez-Lazaro J, and DeMartino DM (2000) Orbital and suborbital variability in North Atlantic bottom water temperature obtained from deep-sea ostracod Mg/Ca ratios. *Palaeogeography, Palaeoclimatology, Palaeoecology* 162: 45–57.
- Curry WB and Oppo DW (2005) Glacial water mass geometry and the distribution of $\delta^{13}\text{C}$ of ΣCO_2 in the western Atlantic Ocean. *Paleoceanography* 20: PA1017. <http://dx.doi.org/10.1029/2004PA001021>.
- Dekens PS, Lea DW, Pak DK, and Spero HJ (2002) Core top calibration of Mg/Ca in tropical foraminifera: Refining paleotemperature estimation. *Geochemistry, Geophysics, Geosystems* 3: 29pp. <http://dx.doi.org/10.1029/2001GC000200>.
- Dwyer GS, Baker PA, and Cronin TM (1995) North Atlantic deepwater temperature change during late Pliocene and late Quaternary climatic cycles. *Science* 270: 1347–1351.
- Eggins S, Deckker PD, and Marshall J (2003) Mg/Ca variation in planktonic foraminifera tests: Implications for reconstructing palaeo-seawater temperature and habitat migration. *Earth and Planetary Science Letters* 212: 291–306.
- Elderfield H and Ganssen G (2000) Past temperature and $\delta^{18}\text{O}$ of surface ocean waters inferred from foraminiferal Mg/Ca ratios. *Nature* 4005: 442–445.
- Epstein S, Buchsbaum R, Lowenstanu HA, and Urey H (1953) The revised carbonate-water oxygen isotope paleotemperature scale. *Geological Society of America Bulletin* 64: 1315–1326.
- Fairbanks RG (1989) A 17,000 year glacio-eustatic sea level record: Influence of glacial melting rates on the Younger Dryas event and deep-ocean circulation. *Nature* 342: 637–642.
- Fairbanks RG, Charles CD, and Wright JD (1992) Origin of global meltwater pulses. In: Taylor RE, et al. (ed.) *Radiocarbon After Four Decades*, pp. 473–500. New York: Springer.
- Fairbanks RG and Matthews RK (1978) The marine oxygen isotope record in Pleistocene coral, Barbados, West Indies. *Quaternary Research* 10: 181–196.
- Govil P and Naidu PD (2010) Evaporation–precipitation changes in the eastern Arabian Sea for the last 68 ka: Implications on monsoon variability. *Paleoceanography* 25: P1210. <http://dx.doi.org/10.1029/2008PA001687>.
- Imbrie JI and Kipp NG (1971) A new micropaleontological method for quantitative paleoclimatology: Application to a late Pleistocene Caribbean core. In: Turekian KK (ed.) *The Late Cenozoic Glacial Ages*, pp. 71–181. New Haven, CT: Yale University Press.
- JPOTS (1980) *The Practical Salinity Scale 1978 and the International Equation of State of Seawater*, p. 25. Paris: UNESCO.
- Katz A (1973) The interaction of magnesium with calcite during crystal growth at $25\text{--}90^\circ\text{C}$ and one atmosphere. *Geochimica et Cosmochimica Acta* 37: 1563–1586.
- Kipp NG (1976) New transfer function for estimating past sea-surface conditions from sea-bed distribution of planktonic foraminiferal assemblages in the North Atlantic. *Geological Society of America Memoir* 145, pp. 3–41.
- Lea DW, Mashiotta TA, and Spero HJ (1999) Controls on magnesium and strontium uptake in planktonic foraminifera determined by live culturing. *Geochimica et Cosmochimica Acta* 63: 2369–2379.

- Lea DW, Pak DK, and Spero HJ (2000) Climate impact of late quaternary equatorial Pacific sea surface temperature variations. *Science* 289: 1719–1724.
- Lear C, Rosenthal Y, and Slowey NC (2002) Benthic foraminiferal Mg/Ca-paleothermometry: A revised core-top calibration. *Geochimica et Cosmochimica Acta* 66: 3375–3387.
- Malaize B and Caley T (2009) Sea surface salinity reconstruction as seen with foraminifera shells: Methods and case studies. *European Physical Journal Conference* 1: 177–188.
- Martin PA, Lea DW, Rosenthal Y, Shackleton NJ, Sarinthein M, and Papenfuss T (2002) Quaternary deep sea temperature histories derived from benthic foraminiferal Mg/Ca. *Earth and Planetary Science Letters* 198: 193–209.
- Mashiotto TA, Lea DW, and Spero HJ (1999) Glacial–interglacial changes in Subantarctic sea surface temperature and $\delta^{18}\text{O}$ water using foraminiferal Mg. *Earth and Planetary Science Letters* 170: 417–432.
- Maslin MA and Burns SJ (2000) Reconstruction of the Amazon Basin effective moisture availability over the past 14,000 years. *Science* 290: 2285–2287.
- McCulloch MT, Gagan MK, Mortimer GE, Chivas AR, and Isdale PJ (1994) A high-resolution Sr/Ca and $\delta^{18}\text{O}$ coral record from the Great Barrier Reef, Australia, and the 1982–1983 El Niño. *Geochimica et Cosmochimica Acta* 58: 2747–2754.
- Millero FJ (1996) *Chemical Oceanography*, 2nd edn. New York: CRC Press.
- Nürnberg D (1995) Magnesium in tests of *Neogloboquadrina pachyderma* sinistral from high Northern and Southern latitudes. *Journal of Foraminiferal Research* 25(4): 350–368.
- Nürnberg D, et al. (1996) Assessing the reliability of magnesium in foraminiferal calcite as a proxy for water mass temperatures. *Geochimica et Cosmochimica Acta* 60: 803–814.
- Oomori T, et al. (1987) Distribution coefficient of Mg^{2+} ions between calcite and solution at 10–50 °C. *Marine Chemistry* 20: 327–336.
- Prell WL and Hays JD (1976) Late Pleistocene faunal and temperature patterns of the Colombia Basin, Caribbean Sea. *Geological Society of America Memoir* 145, pp. 201–220.
- Rohling EJ (2007) Progress in paleosalinity: Overview and presentation of a new approach. *Paleoceanography* 22: PA3215. <http://dx.doi.org/10.1029/2007PA001437>.
- Rosenthal Y, Boyle E, and Slowey N (1997) Temperature control on the incorporation of Mg, Sr, F and Cd into benthic foraminiferal shells from Little Bahama Bank: Prospects for thermocline paleoceanography. *Geochimica et Cosmochimica Acta* 61(17): 3633–3643.
- Rosenthal Y and Lohmann GP (2002) Accurate estimation of sea surface temperatures using dissolution-corrected calibrations for Mg/Capaleothermometry. *Paleoceanography* 17: <http://dx.doi.org/10.1029/2001PA000749>.
- Rosenthal Y, Lohmann GP, Lohmann KC, and Sherrell RM (2000) Incorporation and preservation of Mg in *Globigerinoides sacculifer*: Implications for reconstructing the temperature and $^{18}\text{O}/^{16}\text{O}$ of seawater. *Paleoceanography* 15: 135–145.
- Rozanski K, Araguas-Araguas L, and Gonfiantini R (1993) Isotopic patterns in modern global precipitation. In: Swart PK, Lohman KC, McKenzie J, and Savin S (eds.) *Climate Change in Continental Isotopic Records. Geophysical Monograph Series* 78: pp. 1–36. Washington, DC: American Geophysical Union.
- Savin SM and Douglas RG (1973) Stable isotope and magnesium geochemistry of recent planktonic foraminifera from the south Pacific. *Geological Society of America Bulletin* 84: 2327–2342.
- Schmidt GA, Bigg GR, and Rohling EJ (1999) Global seawater oxygen-18 database. <http://data.giss.gov/data/>.
- Schrag D, Hampt G, and Murray DW (1996) Pore fluid constraints on the temperature and oxygen isotopic composition of the glacial ocean. *Science* 272: 1930–1932.
- Shackleton NJ (1974) Attainment of isotopic equilibrium between ocean waters and the benthonic foraminifera genus *Uvigerina*: Isotopic changes in the ocean during the last glacial. *Colloques Internationaux du Centre National de la Recherche Scientifique* 219: 203–209.
- Spero HJ, Bijma J, Lea DW, and Bemis BE (1997) Effect of seawater carbonate concentration on planktonic foraminiferal carbon and oxygen isotopes. *Nature* 390: 497–500.
- Urey H (1947) The thermodynamic properties of isotopic substances. *Journal of the Chemical Society* 1: 562–581.
- von Langen P, Pak D, Spero H, and Lea D (2005) Effects of temperature on Mg/Ca in neogloboquadrinid shells determined by live culturing. *Geochemistry, Geophysics, Geosystems* 6(10). <http://dx.doi.org/10.1029/2005GC000989>.
- Weber JN and Woodhead PMJ (1972) Temperature dependence of oxygen-18 concentration in reef coral carbonates. *Journal of Geophysical Research* 77(3): 463–473.
- Weldeab S, Lea DW, Schneider RR, and Anderson N (2007) 155,000 years of west African monsoon and ocean thermal evolution. *Science* 316: 1303–1307.
- Wooster WS, Lee AJ, and Dietrich G (1969) Redefinition of salinity. *Deep Sea Research* 16: 321–322.

Terrigenous Sediments

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Introduction

Terrigenous sediments are the weathering products of rocks exposed at the Earth's surface. They are brought to the ocean by rivers, winds, and ice. Although substantially more sediment is transported to the ocean by rivers than winds, most is retained near margins. Most of the clays deposited in areas far away from the continents are delivered by winds (Kennett, 1982; Rea and Hovan, 1995), see also review in Kohfeld and Harrison (2001), particularly in the Pacific Ocean. In polar and subpolar regions, melting icebergs may also contribute a significant amount of ice-rafted detritus (IRD) (see [Glaciomarine Sediments and Ice-Rafted Debris](#)). Once shed from the continent, terrigenous sediments are subject almost exclusively to physical processes, most importantly, transport by ocean currents. Turbidity currents and other gravity flows bring sediments from the continental shelves and river deltas to the deep ocean, where they can be picked up and transported by bottom currents. Strong surface and intermediate level ocean currents are also important for sediment transport and redistribution in the oceans (e.g., Franzese et al., 2006; Kennett, 1982; Latimer et al., 2006; Petschick et al., 1996).

The ability of an ocean current to transport particles depends on the speed of the current relative to the settling velocity of the particles, which is proportional to grain size (Figure 1). Changes in flow speed (either spatial or temporal) will alter the transport capacity of a current and the grain size distribution of the underlying sediment. Grain size is, therefore, a useful proxy for flow speed and an important measure of bottom currents' relationship to climate variability (e.g., McCave and Hall, 2006). Grain size alone, however, cannot give us information about changes in flow direction or about the upper-level ocean circulation.

Tying a deep-sea sediment sample to its continental source region (provenance) can yield important information about the processes responsible for its transport and deposition, including important paleoclimate variables such as wind speed and direction, aridity, glacial activity, and both surface and deep ocean currents. The composition of a terrigenous sediment sample depends on many variables related to the lithology and age of the source rock, as well as the weathering and transport history of the resulting sediment. An important problem in sediment provenance studies is distinguishing between provenance and process controls on the final, observed sedimentary composition. This article gives an overview that outlines the potential of terrigenous sediments in paleoceanographic applications, specifically concerning their provenance (sources), weathering history, sorting (grain size distribution), and flux (see [Paleoceanography An Overview](#)).

Provenance (Lithologic Source)

As previously mentioned, the geochemical and mineralogical composition of a terrigenous sediment sample depends on the

lithology and age of the source rock, as well as the weathering and transport history of the resulting sediment. The primary signature, however, is determined by the initial bulk composition of the source rock (i.e., felsic vs. mafic), its age, and geologic history.

The Earth's crust is formed by partial melting and subsequent fractional crystallization of the upper mantle. During this process, elements are fractionated based on their compatibility within the mantle rocks, and minerals are formed according to their crystallization temperatures (Figure 2; Bowen, 1928). During partial melting, elements that are more incompatible will be enriched in the melts and depleted in the residual solids, while more compatible elements are enriched in the residual solids and depleted in the partial melts. During fractional crystallization, olivine or ultramafic minerals are the first to crystallize, followed by mafic minerals (pyroxene, amphibole, biotite, and plagioclase), and finally the felsic minerals (potassium feldspar, muscovite, and quartz) most common on the Earth's surface. As these minerals form, they leave the residual magma depleted in the elements which are incorporated in those minerals and enriched in incompatible elements. For example, olivine tends to concentrate Ni, pyroxenes concentrate Sc and Cr, and plagioclase feldspars concentrate Ca, Sr, and Eu (e.g., Albarede, 2003).

The continental crust is formed by fractional crystallization of enriched magmas, giving them felsic compositions with high $\text{SiO}_2/\text{Al}_2\text{O}_3$ and $\text{K}_2\text{O}/\text{Na}_2\text{O}$, enrichments of incompatible elements, such as the alkali and alkaline earth elements, especially the large-ion-lithophile elements (K, Rb, Cs, and Ba), over compatible trace elements, and depletions in Eu compared to its neighboring rare earth elements (REE). Oceanic crust is thought to be formed from the residual depleted upper mantle and, therefore, has a mafic mineral composition with low $\text{SiO}_2/\text{Al}_2\text{O}_3$ and $\text{K}_2\text{O}/\text{Na}_2\text{O}$, relatively high concentration of refractory elements such as Fe and Mg, enrichments in compatible trace elements (e.g., Ni, Cr, and Sc) over incompatible ones, and no significant Eu depletions (e.g., Albarede, 2003; Taylor and McLennan, 1985).

The differences in bulk mineralogical and major and trace element concentrations also translate into different isotopic signatures for more mafic, depleted sources versus more felsic, enriched sources. This is due to the chemical properties of the radioactive parent elements compared to the radiogenic daughter elements. For the Rb–Sr system, the parent isotope, rubidium (Rb) is more incompatible than the daughter Strontium (Sr), leading to a higher Rb/Sr in magma melts compared to residual solids. This translates into higher Rb/Sr in the igneous rocks that make up the bulk of the continental crust and lower Rb/Sr in the mantle and the mafic igneous rocks of the ocean crust. Over time, ^{87}Rb decays to ^{87}Sr ($t_{1/2} = 48.8 \text{ Ga}$), increasing the concentration of ^{87}Sr relative to the stable, non-radiogenic isotope ^{86}Sr . For a given age, felsic rocks will have

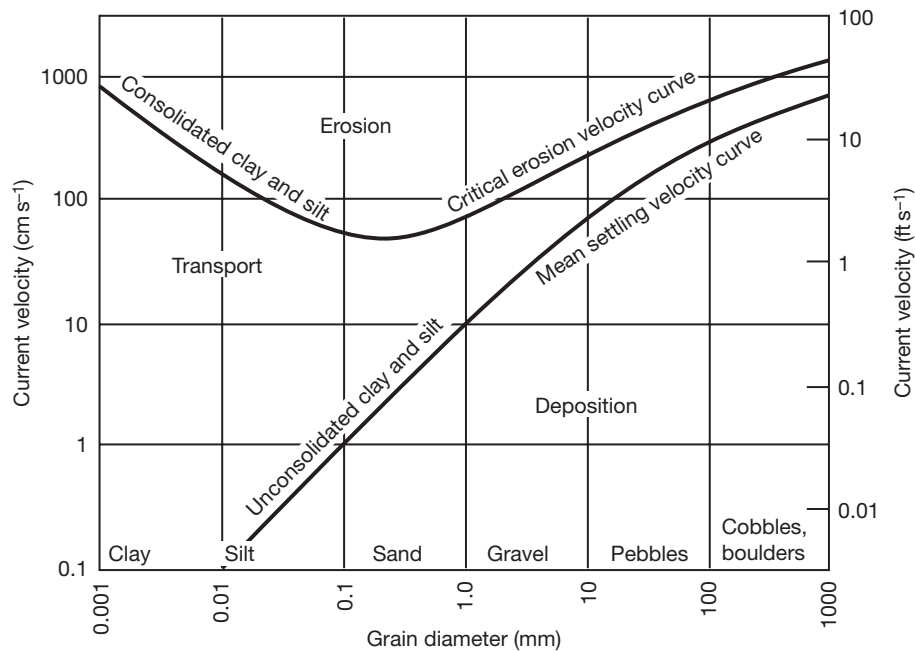


Figure 1 Hjulstrom's diagram, showing the critical current velocity required to move quartz grains on a flat, uniform bed at a water depth of 1 m as well as the minimum velocity required to sustain transport for sediments of varying sizes. Electrostatic cohesion increases the critical erosion velocities of small particles (clay and fine silt).

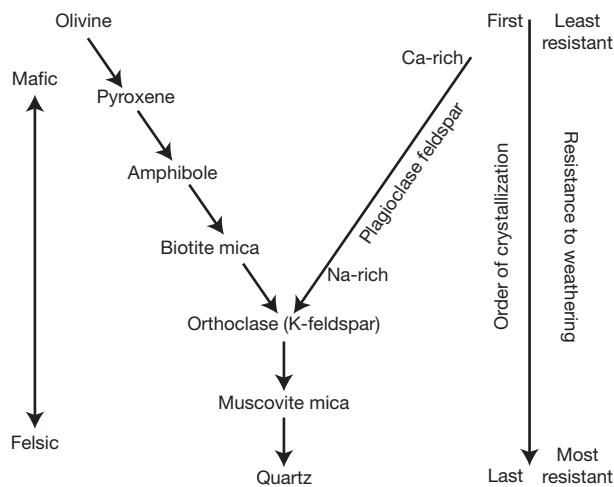


Figure 2 Bowen's reaction series; the relative order of crystallization of the common silicate minerals from a cooling magma (Bowen, 1928).

higher $^{87}\text{Sr}/^{86}\text{Sr}$ than mafic rocks, and for a given composition, $^{87}\text{Sr}/^{86}\text{Sr}$ increases with age. The Rb–Sr system is, therefore, extremely useful for discriminating among potential source areas with very different bulk chemistries (i.e., felsic vs. mafic) or very different ages. Conversely, for the Sm–Nd system, the parent isotope Samarium (Sm) is slightly less incompatible than the daughter neodymium (Nd), leading to a lower Sm/Nd in magma melts compared to residual solids. This translates into higher Sm/Nd in the upper continental crust and lower Sm/Nd in the mantle and ocean crust. Over time, ^{147}Sm decays to ^{143}Nd ($t_{1/2} = 106$ Ga), increasing the $^{143}\text{Nd}/^{144}\text{Nd}$. In this case, for a given age, felsic rocks will

have lower $^{143}\text{Nd}/^{144}\text{Nd}$ than mafic rocks. This leads to a gross negative correlation between Nd and Sr isotopes in terrigenous sediments (Figure 3). The Sm–Nd and Rb–Sr systems are most widely used for constraining the provenance of terrigenous marine sediments (e.g., Dasch, 1969; Franzese et al., 2006; Goldstein et al., 1984; Grousset and Biscaye, 2005; Grousset et al., 1988; Jones et al., 1994). Although less widely used to date, the U–Th–Pb, Lu–Hf, and K–Ar systems have also been shown to provide powerful provenance constraints. The commonality in these is that they are the radioactive decay products of parents with long half lives. They have different sensitivities depending on the different geochemical behaviors of the parents and daughters and on the rock types and geological history of the sources. Thus, they provide provenance constraints that are related to both the age and time-averaged geochemical characteristics of sedimentary sources.

Geologic age will most obviously affect a continental source rock in terms of its radiogenic isotopic composition. Geologic age can also be expressed in the mineralogical, major and trace element composition, as older rocks are more likely to have been involved in processes such as metamorphism, weathering, and crustal recycling (discussed further below). For example, metamorphism affects the composition when soluble elements are mobilized and, subsequently, lost from the host rock. Weathering also tends to release soluble elements, so old continental rocks tend to have high ratios of insoluble/soluble elements (e.g., Rb/Sr and Th/U Dasch, 1969; Goldstein and Jacobsen, 1988; McLennan and Taylor, 1980). These effects are magnified if they occur over many rock cycles, so sediments derived from old upper continental crust tend to have extended histories of metamorphism and weathering and a substantial component of recycled sedimentary rocks in their provenance.

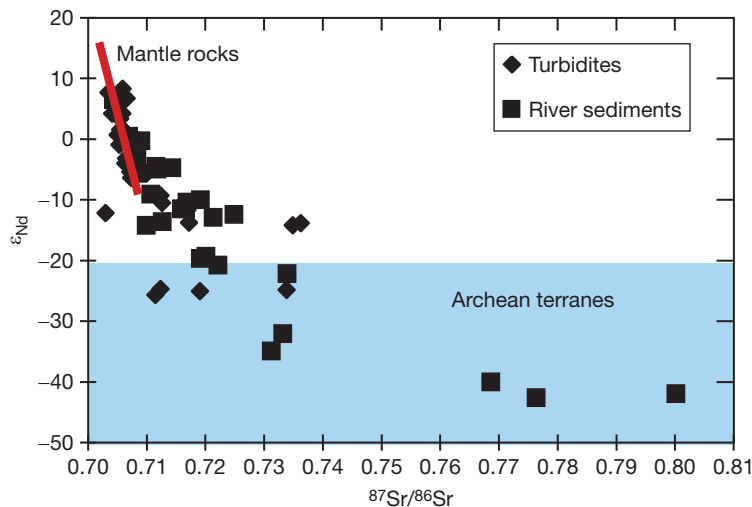


Figure 3 Nd (reported as ϵ_{Nd} , which is the fractional deviation of $^{143}\text{Nd}/^{144}\text{Nd}$ from the bulk Earth value in parts per 10 000) versus Sr isotope variations in river sediments (Goldstein and Jacobsen, 1988) and deep-sea turbidites (Hemming and McLennan, 2001; McLennan et al., 1990). Although there is generally a negative correlation between these two isotopes, the range in $^{87}\text{Sr}/^{86}\text{Sr}$ for a given provenance is great, particularly for sediments of ancient (low ϵ_{Nd}) provenance.

Sedimentary Processes

In addition to provenance variability and mixing, the processes of weathering and transport can have significant impact on terrigenous sediment composition. During these processes, elements are fractionated based on their chemical behavior (soluble vs. insoluble or refractory) and the relative stability of their mineralogical host(s). Sediments are the byproduct of weathering; some of them are minerals altered from the original rock, some simply reduced in size, and some are newly formed minerals.

Physical weathering tends to break down softer minerals, leaving behind the hardest, most resistant minerals. Chemical weathering involves the breakdown of the primary minerals in the rock to new secondary minerals that are more stable in the surface environment or to material that may be carried away in solution. The stability of minerals can be predicted using Bowen's reaction series (Figure 2; Bowen, 1928). In general, mafic minerals, the first ones to crystallize from a magma, are the most unstable at the Earth's surface and quickest to weather because the surface is most different from the conditions under which they formed at equilibrium. Felsic minerals are much more stable because the conditions at the surface are much more similar to the conditions under which they formed. Quartz is the most stable of the common rock-forming minerals, while feldspars are readily converted to clay minerals. For this reason, the quartz/feldspar ratio of sediment is often used as an indicator of the extent of weathering, and/or the importance of an old, recycled provenance (e.g., Diekmann and Kuhn, 2002; Diekmann et al., 2000; Kuhn and Diekmann, 2002).

As minerals are broken down and/or converted to clay minerals, the relative behavior of the elements becomes important. Insoluble, or refractory, elements (e.g., Al, Th, and REE) are retained in the rock regolith or newly formed minerals, while soluble elements (e.g., alkali and alkaline earth

elements and U) are more likely to be removed from the system. Postweathering, sediment transport tends to separate the weathering products from the host rock, fractionating the minerals and elements accordingly. In addition, particles are sorted by grain size, with the larger grains from a given source rock containing the most resistant minerals.

Formation and Distribution of Clay Minerals

Away from continental slopes, clay minerals are the most widely distributed terrigenous components in pelagic marine sediments, and their occurrence and distribution are well reviewed by (Kennett, 1982) (see *Eolian Records, Deep-Sea Sediments*). Some clays are formed as weathering or alteration products of primary silicate minerals, and others are recycled from sedimentary sources. All clay minerals are very fine-grained due to their crystal structure. Four types of clays are dominantly found in marine sediments: chlorite, illite, kaolinite, and smectite. In most cases, chlorite and illite are primary minerals from the source, while kaolinite and smectite are newly formed alteration products. Mineral composition of the clays is related to a number of factors, including source rock type and weathering. Weathering in turn is controlled by relief, temperature, precipitation, and vegetation, and consequently, the different kinds of clay reflect the major soil types in the source areas (Gibbs, 1967). Most clays are carried to the ocean and deposited without being changed, and thus marine sediments contain clays that reflect the weathering process that produced them (e.g., Biscaye, 1965). Primary minerals such as chlorite and illite are relatively more abundant in colder climates, while kaolinite is characteristic of tropical zones (Figure 4). Smectite tends to be associated with distribution of volcanic sources. The fact that much illite is brought to the ocean as a primary mineral is supported by the survey of K/Ar ages by Hurley et al., 1963.

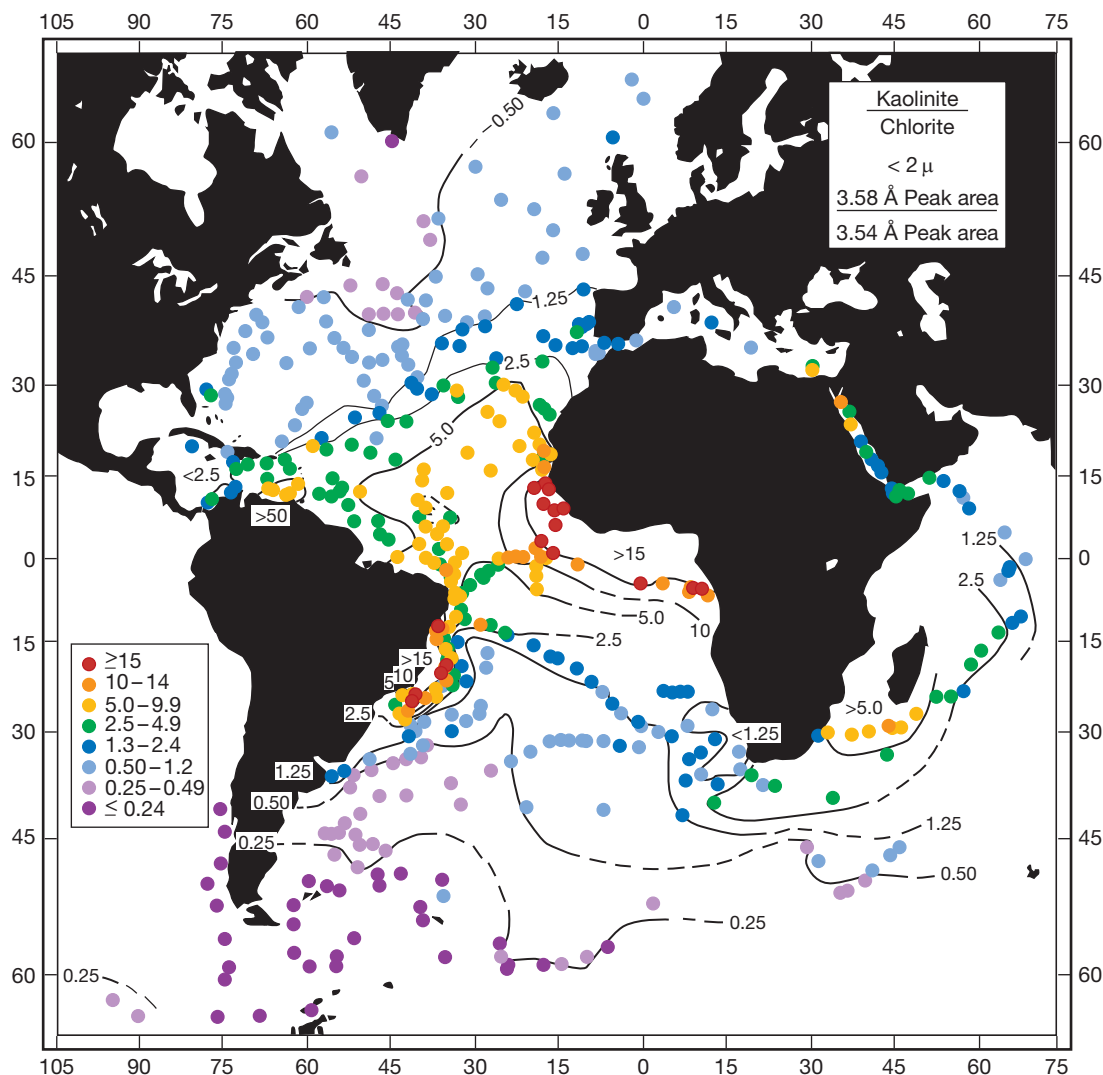


Figure 4 Map of kaolinite/chlorite distribution of the $<2\ \mu\text{m}$ terrigenous fraction of sediments from the Atlantic Ocean. Modified from Biscaye PE (1965) Mineralogy and sedimentation of recent deep-sea clay in Atlantic Ocean and adjacent seas and oceans. *Geological Society of America Bulletin* 76(7): 803–832.

Geochemical Signatures of Weathering

In addition to mineralogical changes, predictable bulk chemical changes occur during weathering, and the chemical index of alteration (CIA) has been established as a general guide to the degree of weathering (Nesbitt and Young, 1982). Using molecular proportions (and only considering CaO residing in silicate minerals rather than in phosphate and carbonate):

$$\text{CIA} = 100 \times \frac{\text{Al}_2\text{O}_3}{[\text{Al}_2\text{O}_3 + \text{K}_2\text{O} + \text{Na}_2\text{O} + \text{CaO}]}$$

For marine sediments, it is also necessary to make a correction on Na_2O for seawater salt (e.g., McLennan et al., 1990). On the CIA scale, unweathered igneous rocks have values of 50 or slightly below, while the most extreme residual clays have values of almost 100. Typical shales average about 70–75 (Figure 5). The bulk-rock chemistry is dominated by the conversion of feldspars or glass to clay minerals during weathering.

The contents of some trace elements in sediments are different from those of their sources due to sedimentary processes of weathering and sorting, and measurement of these may provide clues about the processes and pathways of sediment distribution. For example, during weathering, there is a substantial increase in the Rb/Sr ratio of most solid residues (e.g., Dasch, 1969). Rb tends to follow K and is enriched in K-feldspar and micas, whereas Sr tends to follow Ca and is enriched in plagioclase and pyroxene. Plagioclase and pyroxene are generally susceptible to chemical attack during weathering, and Sr is removed to the aqueous system. Rubidium's hosts have greater resistance to chemical weathering, and additionally, Rb released during weathering tends to follow Al and thus be retained in the solid system. This is because Rb^+ , a large alkali trace element, is more readily retained on exchange sites of clays than the smaller alkaline earth element, Sr^{2+} , which is more easily lost to aqueous solution. Thus, terrigenous sediments that are derived from highly weathered terrains

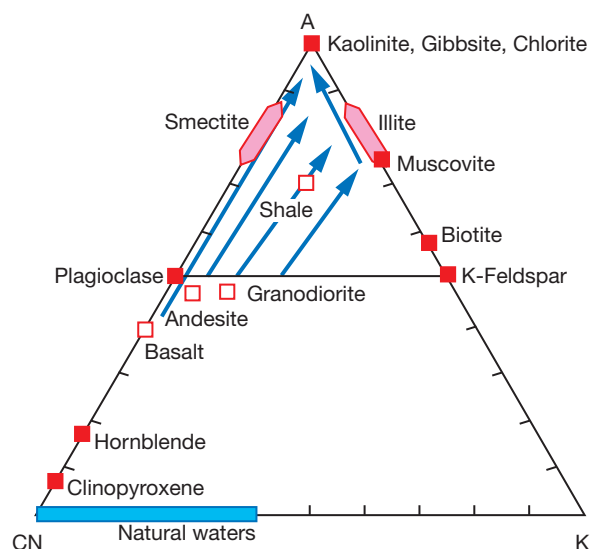


Figure 5 Ternary plot of Al_2O_3 (A), $\text{CaO} + \text{Na}_2\text{O}$ (CN), and K_2O (K) of the terrigenous silicate fraction of sediments. Chemical index of alteration (CIA) is a measure of the degree of chemical weathering. The terminology comes from (Nesbitt and Young, 1982) and is reviewed in (McLennan et al., 1990, 1993).

and/or sedimentary sources will tend to have high Rb/Sr ratios. Over time, this evolves into higher $^{87}\text{Sr}/^{86}\text{Sr}$. Recent enrichments in Rb relative to Sr will result in lower $^{87}\text{Sr}/^{86}\text{Sr}$ and thus younger Rb–Sr model ages, so in favorable cases, it is possible to distinguish recent weathering from old sedimentary sources in terrigenous marine sediments.

The $\text{K}_2\text{O}/\text{Na}_2\text{O}$ and $\text{K}_2\text{O}/\text{CaO}$ of solid residues of weathering are also increased. Additionally, the K–Ar system may be almost entirely reset during extreme weathering. This is because K^+ is retained on exchange sites of clays, analogous to Rb^+ , and Ar is a noble gas that is released to the atmosphere during the breakdown of feldspar. Accordingly, the CAI, Rb–Sr, and K–Ar systems are complementary approaches to study time-integrated weathering effects on terrigenous sediments. Because K is much more soluble than Al or Rb, the ratios $\text{Al}_2\text{O}_3/\text{K}_2\text{O}$ and Rb/K should also increase in the sediments with enhanced chemical weathering and the conversion of silicate minerals to clay. Similarly, U can be oxidized to U^{6+} and forms the highly stable species $(\text{UO}_2)^{2+}$, which is much more soluble than Th, so weathering tends to increase Th/U in the sediments. This phenomenon is most easily observed with old, recycled sedimentary sources, as U would be continually lost during successive cycles of weathering (e.g., McLennan and Taylor, 1980).

Sorting

The processes of sediment transport and deposition tend to sort sediments by grain size. Grain size distributions can provide compelling information about the transporting agent (wind, water, and ice) as well as the velocity of wind and water. For example, wind tends to carry fine-grained sediments, with average grain size decreasing with distance from source (e.g., Rea, 1994). Bottom currents act to move sediments their

paths along shaped by bathymetric boundaries. Drift deposits accumulate in association with the movement of these deep water boundary currents. Bottom currents tend to enrich drift deposits in silt ($2\text{--}63\ \mu\text{m}$) and clay ($<2\ \mu\text{m}$) sized sediment. Sortable silt (SS) is defined as the $10\text{--}63\ \mu\text{m}$ silt fraction, because below about $10\ \mu\text{m}$, particles tend to behave cohesively and may not be sorted by primary particle size (e.g., McCave and Hall, 2006). This is due primarily to the dominance of clay minerals in the finer sediment fraction but also to van der Waals forces becoming significant. In general, the mean grain size of SS mean is proportional to bottom current speed, as is the percentage of SS% in the total $<63\ \mu\text{m}$ terrigenous fraction (reviewed in McCave and Hall, 2006), and the ratio of silt/clay (e.g., Diekmann et al., 2003; Hall and McCave, 2000; Kuhn and Diekmann, 2002).

Particles of different grain sizes often have different mineral and geochemical compositions, so sedimentary sorting can alter the composition of sediments deposited in the deep sea. Grain size tends to correlate negatively with elements that are associated with clay minerals, such as Al, Rb, and Cs. Clay fractions will also generally contain higher Rb/Sr than the coarse fractions. This can be conceived simply by considering the main host of Rb and Sr in the clay fraction to be illite and in the silt and sand fractions to be feldspar. In the first order, there is a separation of these elements according to mineralogy such that Rb is mostly found in mica and Sr is mostly found in feldspar. This is the reason there tends to be higher $^{87}\text{Sr}/^{86}\text{Sr}$ in clays compared to sands.

Sorting also tends to separate denser minerals from the bulk sediment. Zr, Hf, and heavy REE are enriched in the heavy mineral fraction of sediments, particularly in zircon (ZrSiO_4), which is among the most robust minerals, surviving multiple sedimentary cycles, as well as high-grade metamorphism and even melting, and tend to be more abundant in coarser sediments (e.g., Albarede, 2003; Taylor and McLennan, 1985). The ratio Zr/Rb has also been used as proxy for grain size or degree of sorting, since Zr is heavily concentrated in the coarse sediment, while Rb tends to be concentrated in clay (Liu et al., 2006).

In summary, sorting during sediment transport is a process of unmixing, whereby grains of different sizes are physically separated from one another. Because the different grain size fractions tend to have different compositions, differential sorting can create spatial or temporal variability in sediment composition with no change in average provenance, only in the average grain size of particles being deposited.

Application of Terrigenous Sediments in Paleooceanography

Distinguishing Between Provenance and Process

Sediments are inherently mixtures of particles from various sources. It is rare to find deep-sea sediment with a single point source of terrigenous sediment (outside of a delta). The composition of terrigenous sediment, therefore, represents a mixture of the multiple end-member source compositions. Variations in terrigenous sediment composition can be caused by variations in the composition or flux of one or more of the

end-member sediment sources, transport mechanism, and/or by the sedimentary processes described above.

As mentioned previously, it can be difficult to distinguish recent weathering from an old, crustal provenance. A change in an index of chemical weathering measured on terrigenous seafloor sediments could theoretically be interpreted as: (1) a change in the intensity of continental weathering just prior to the time of deposition, (2) a change in average provenance between more juvenile and more chemically mature parent materials, or (3) a combination of both. As an example, a relative depletion in Sr could be the result of surficial Sr loss or of a provenance change between sources with differing lithologies. Strontium is a moderately incompatible element that is highly mobile during surficial weathering. Comparing Sr concentrations with those of other elements that are refractory in surficial processes but have different compatibilities in igneous systems should permit discrimination between lithologic variability and weathering. Nd, Th, and Y are all relatively insoluble; Nd has a similar compatibility to Sr; Th is more incompatible than Sr; and Y is much more compatible. Therefore, with provenance variability, lower Sr/Th should be accompanied by lower Nd/Th, and lower Sr/Y should be accompanied by lower Nd/Y. With chemical weathering, on the other hand, Sr/Th and Sr/Y should decrease with no change in Nd/Th or Nd/Y.

It is also possible to use trace element ratios to isolate source compositional variability from sedimentary sorting. For example, Th and Zr are both incompatible elements, while Sc is moderately compatible during melting and crystallization of magmas. In the absence of any sedimentary processes, the ratios Th/Sc and Zr/Sc should correlate (Figure 6). Because sorting does not appreciably separate Th from Sc, but sorting of zircons makes large differences in the Zr/Sc, any increase in Zr/Sc for relatively constant Th/Sc provides a clear indication of cases where substantial enrichment of zircon has occurred as a result of sorting (McLennan et al., 1990, 1993).

As described in previous sections, the Rb–Sr isotope system may be disturbed by many geological processes, some of which are part of the continental provenance, and others that may be related to recent weathering and transport processes. The Sm–Nd isotope system, on the other hand, provides an average age of crust formation of the sediment's sources (Goldstein et al., 1984). As reviewed by Taylor and McLennan (1985), this is because Sm and Nd are REE with similar radii and, thus, are generally not separated by most sedimentary processes (although there are exceptions). The Rb–Sr system thus provides a wide range of compositions for a given Sm–Nd model age or $^{143}\text{Nd}/^{144}\text{Nd}$ (Dasch, 1969; Goldstein and Jacobsen, 1988). Comparing variations in the two isotope systems is therefore a useful way to confirm whether Rb–Sr system changes are related to provenance or sedimentary process (e.g., Franzese et al., 2006; Revel et al., 1996a,b; Walter et al., 2000).

Grain size distributions can also provide valuable information on provenance and dispersal of sediments, especially when combined with geochemical data and inverse end-member modeling (e.g., Weltje and Prins, 2003). Such a combined approach of quantitative grain size and compositional variability can often discriminate between process-related changes (e.g., wind speed or current speed) and changes in sediment source or delivery mechanism (e.g., wind vs. ocean).

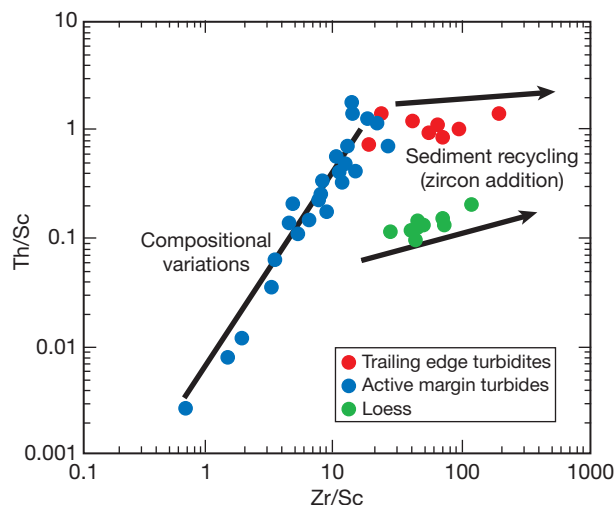


Figure 6 Th/Sc versus Zr/Sc. Th and Zr are both incompatible in igneous processes, while Sc is compatible. Accordingly, increasing ratios in source rocks indicate more evolved compositions. Because sorting does not appreciably separate Th from Sc, but sorting of zircons makes large differences in the Zr/Sc, increase in Zr/Sc for relatively constant Th/Sc provides a clear indication of cases where substantial enrichment of zircon has occurred as a result of sorting. Reproduced from McLennan SM, Taylor SR, McCulloch MT, and Maynard JB (1990) Geochemical and Nd–Sr isotopic composition of deep-sea turbidites – Crustal evolution and plate tectonic associations. *Geochimica et Cosmochimica Acta* 54(7): 2015–2050; McLennan SM, Hemming S, McDaniel DK, and Hanson GN (1993) Geochemical approaches to sedimentation, provenance, and tectonics. *Geological Society of America* 284: 21–40 (Special Paper).

Terrane Types or Components

Any compositional variability (in time or space) that cannot be explained by recent sedimentary processes must be due to provenance variations. In some cases, potential source areas can be identified and directly compared with the deep-sea sediment. In many cases, however, there are no unique ways of identifying or ‘fingerprinting’ sediment sources, and it is useful to group sediment provenance characteristic into terrane types. Five such groupings have been described extensively (McLennan et al., 1990, 1993); they are (1) old upper continental crust, (2) recycled sedimentary rocks, (3) young undifferentiated arc, (4) young differentiated arc, and (5) exotic components. A sixth component that may be a significant portion of the terrigenous marine sediments in the pelagic realm, and away from continental inputs (e.g., in the Antarctic Circumpolar current, downstream from oceanic plateaus and ridges) is (6) oceanic basalt. Continental crust is distinguished from undifferentiated sources (arc and oceanic basalt) by evolved major element compositions (e.g., high $\text{SiO}_2/\text{Al}_2\text{O}_3$ and $\text{K}_2\text{O}/\text{Na}_2\text{O}$), negative Eu anomalies, and enrichments of normally incompatible over compatible trace elements. For example, continental crust yields sediments with light REE enrichment, high Th/Sc, and high La/Sc.

Sediments derived from old upper continental crust tend to have a substantial component of recycled sedimentary rocks in their provenance. In addition to the characteristics generally noted for continental crust, sediments derived from old

continental crust will tend to have high Rb/Sr and high Th/U reflecting weathering and sedimentary recycling. The presence of old continental sources is most simply identified by radiogenic isotope analyses.

Recycled sedimentary rocks are a common and important contributor to terrigenous sediments, and it is difficult to distinguish ambient weathering from recycled sedimentary sources. Measurements such as K/Ar ages, Rb/Sr model ages, and clay mineralogy are probably the best ways to distinguish these in young terrigenous sediments.

Young undifferentiated arc sources are recently derived from the mantle without significant intracrustal differentiation. This provenance component has many attributes in common with oceanic basalt sources but may have different implications for sedimentary (and thus paleoclimate/paleoceanography) applications. In Quaternary applications, the distribution of possible sources is largely still available for study, and the geographic pattern of variation in provenance helps to distinguish these contributions. Compared to continental sources, they provide sediments with less evolved major element compositions (e.g., low but variable $\text{SiO}_2/\text{Al}_2\text{O}_3$ and $\text{K}_2\text{O}/\text{Na}_2\text{O}$), absence of negative Eu anomalies, and little enrichments of normally incompatible over compatible trace elements, so they have low Th/Sc and La/Sc smaller recycled sedimentary component, so their CIA and Rb/Sr are low.

Young differentiated arc sources have very similar characteristics to continental crust and can be thought of as young continental crust. In most cases, these sources also tend to be less weathered than old continental sources, with a smaller recycled sedimentary component, so their CIA and Rb/Sr are low.

In specific cases, identification of even minuscule amounts of certain exotic components can have very important application. McLennan et al. (1993) cite published examples of their utility for constraining the timing of approach of allochthonous terranes as well as identifying ocean crust components in modern forearc turbidites. If identified, exotic components can provide near-unique information for interpreting provenance of terrigenous sediments in Quaternary settings.

Unmixing and Interpreting Changes in Mixing Proportions (Fluxes)

Spatial or temporal variations in the provenance of terrigenous sediment can often be linked to paleoceanographic and/or climatic changes. Such variations, even when clearly identified as independent of recent sedimentary processes, should be interpreted with caution. Seafloor sediments are often mixtures of various end-members, each with their own provenance. Any change in the composition of the mixture could, therefore, be the result of (1) a change in the composition of one or more end-member sources due to recent weathering and/or sorting; (2) the addition or removal of a sedimentary source; or (3) a change in the flux of one or more end-members to the site. The two previous sections describe some ways in which the first two options can be evaluated. The third option can be evaluated by first identifying the various end-members and then applying a set of end-member mixing equations to the data. Depending on the number of end-members and the number of variables, these can be as simple as single algebraic equation or

a complex matrix of algorithms such as those used in end-member mixing algorithm, reviewed in Weltje and Prins (2003). In general, the mixing equations take the following form:

$$R_{\text{mix}}C_{\text{mix}} = \sum (R_n C_n f_n) \text{ and } C_{\text{mix}} = \sum (C_n f_n)$$

where R stands for the isotope ratio, C for the concentration, and f for the mixing fraction of each of n end-members.

An important caveat is that the mixing fractions must always add up to 1 (or 100%), so the mixing equations can only solve for the relative proportions of the end-members. For example, in a simple situation with two end-members, A and B, one cannot distinguish an increased contribution of A from a decreased contribution of B with compositional data alone. By multiplying the mixing fraction by the total mass accumulation rates (MAR) of terrigenous sediment, one can calculate end-member MARs and identify which end-member (or end-members) actually changed. Going one step further, the MAR may not always be directly related to changes in local sediment flux. Lateral advection of sediment at the sea floor, whether syndepositional or postdepositional, can produce major changes in terrigenous MAR (reviewed in Francois et al., 2004). When evaluating provenance changes in the context of transport pathways ocean circulation, this is an important parameter to consider.

Constant flux proxies provide important constraints on sedimentary redistribution in the ocean. Examples are $^{230}\text{Th}_{\text{xs}}$ (derived from radioactive decay of ^{234}U in the water column, e.g., Bacon, 1984; Henderson and Anderson, 2003; Francois et al., 2004) and ^3He (derived from interplanetary dust particles, e.g., Ozima et al., 1984). ^{230}Th is a constant flux proxy because thorium is highly particle reactive, and thus, it is quickly removed from the water column after forming by radioactive decay of ^{234}U which uniformly produces ^{230}Th in the water column at a known and constant rate. The ^3He content of marine sediments is a constant flux proxy because interplanetary dust particles rain to the Earth at approximately constant rates, and they have extremely high ^3He contents. Thus, they act like an isotope spike. The enhancement of sedimentation rate by advection is well-known on drift deposits, for example, Bermuda Rise. In fact, they are favored sites for paleoceanography research due to the high accumulation rates that provide great temporal resolution. This process is termed 'focusing,' and the 'focusing factor' is the ratio of accumulation rate from chronology to accumulation rate that would hold if only sediment settling vertically through the water column above the deposition site were deposited. However, geochemical proxies for constant flux have identified additional sites where the sediment morphology does not clearly indicate drift deposits, yet the vertical accumulation of sediments based on chronology of the sediments is greater than that estimated from the concentrations of constant flux proxies in the sediment.

In a recent example, geochemical data (isotope ratios and trace element concentrations) was used to identify three major end-member sources for the sediments deposited south of Africa and to calculate the fraction of each end-member that contributes to each sediment sample. ^{230}Th -normalized fluxes were then used to determine that the provenance changes were

due in part to a lower supply of sediment derived from the Agulhas Current during the Last Glacial Maximum about 20 ka (Franzese et al., 2006). This was the first study to combine provenance tracers with ^{230}Th -normalization, which together have great potential for tracing sediment transport and source history.

Tools

Petrography

The simplest tool used to study terrigenous sediment is the petrographic microscope. Petrographic studies allow for visual identification of major lithological components in the sand ($>63\ \mu\text{m}$) fraction, as well as potentially diagnostic minerals and/or rock types. This approach has been particularly useful for identifying sources of IRD in the ocean. For example, it was the first line of evidence used to infer the importance of Hudson Strait sources to the Heinrich layers, layers rich in IRD in the North Atlantic associated with major abrupt climate change (Bond et al., 1992, reviewed in Hemming, 2004). It was also used by Bond and Lotti (1995), as well as others afterwards, to examine the relative timing of different iceberg contributors in the North Atlantic. Petrographic studies of IRD can sometimes be used to identify IRD from specific ice sheets and thus be used to make inferences about the behavior of specific ice sheets during times of climate change and can also be used to trace the average path of the main surface currents in regions where icebergs and sea ice are produced.

IRD is composed of a large spectrum of grain sizes, ranging from clay to sand and gravel, but its presence is identified in marine sediments by the coarse fraction, because rafting is the only means of delivering coarse sediment to these regions (e.g., Ruddiman, 1977). The recognition of IRD allows paleoceanographers to study ice-sheet dynamics in the past, and its distribution indicates the flow of surface currents that carried the icebergs (Figure 7).

Mineralogy by X-Ray Diffraction

X-ray diffraction (XRD) is the primary, nondestructive tool for identifying the mineralogy of fine-grained terrigenous sediments. Every mineral has a characteristic XRD pattern whose 'fingerprint' can be matched against a database of known minerals. XRD data can be analyzed to determine the proportion (weight percent) of the different minerals present.

Clay mineralogy by XRD provides important information on weathering history of marine sediments (e.g., Biscaye, 1965). For example, the kaolinite/chlorite of the $<2\ \mu\text{m}$ fraction of surface sediments from the Atlantic Ocean gives an important first-order view of the sources and processes responsible for distributing sediments in the ocean (Figure 4). Mineralogy by XRD has become a widely used tool in paleoceanography to constrain weathering intensity as well as sources and transportation pathways.

Geographic variations in clay mineralogy can help to constrain sediment transportation processes. For example, Petschick et al. (1996) made a more detailed map of the clay mineral distribution in the South Atlantic Ocean than the one from Biscaye (1965). The summary map from Petschick et al. (1996) is shown in Figure 8(a), along with surface

contributions and deep current systems. An example of inherited 'tropical' signal is seen in the kaolinite-rich zone near the Filchner-Ronne Ice shelf. Although deep currents represent an important means of transporting sediment, for example, the tongue of chlorite-rich sediment in the western South Atlantic along the South American margin, comparison of the clay province boundaries with the distribution of surface currents in the South Atlantic (Figure 8(b)), the bathymetry of the South Atlantic (Figure 8(c)), and the July and January average winds (Figure 8(d) and 8(e)), it becomes clear that surface currents (and/or the winds that drive them) are also responsible for the first-order distribution of terrigenous sediments in the South Atlantic.

In combination with clay mineralogy, the relative proportions of quartz and feldspar measured by XRD have been used to distinguish old, weathered, or recycled terranes with high Qz/Fsp ratios from younger, fresher, or undifferentiated terranes (e.g., Diekmann and Kuhn, 2002; Diekmann et al., 2000; Kuhn and Diekmann, 2002).

Geochemistry

The combined provenance and transport history is expressed in the isotopic and major and trace element composition of the terrigenous sediment. Geochemical approaches, including isotopes, therefore, complement information inferred from mineralogical investigations. Although the composition of organic matter from terrigenous sources is acknowledged as a powerful provenance tracer, only the silicate portion of these sediments is reviewed here. Geochemical approaches to the silicate portion of terrigenous sediments allow us to constrain provenance age and geochemical history of the source as well as the weathering history. As described by McLennan et al. (1993), among the important geochemical characteristics that define fundamental provenance types are radiogenic isotope compositions (reflecting average provenance age), large-ion lithophile element enrichments, and other trace element characteristics such as Eu anomalies (reflecting provenance compositions), alkali and alkaline earth element depletions, and aluminum enrichment (reflecting weathering and alteration), Zr and Hf enrichments (reflecting increased concentration of zircon by sedimentary sorting), and high Cr abundances (indicating ultramafic sources).

Major and Trace Element Concentrations

Solution techniques

Paleoceanographic studies of terrigenous sediments use many of the same geochemical techniques commonly used to study rocks. In the case of deep-sea sediments, however, sequential leaching techniques are usually needed to remove biogenic and authigenic sediments prior to sediment digestion (e.g., Bayon et al., 2002). Major and trace element concentrations can then be analyzed in solution by inductively coupled plasma mass spectrometry (ICP-MS), optical emission spectroscopy (ICP-OES or ICP-AES), or atomic-absorption spectroscopy.

X-ray fluorescence

X-ray fluorescence (XRF) is a nondestructive technique used for chemical analysis of materials. An X-ray source is used to

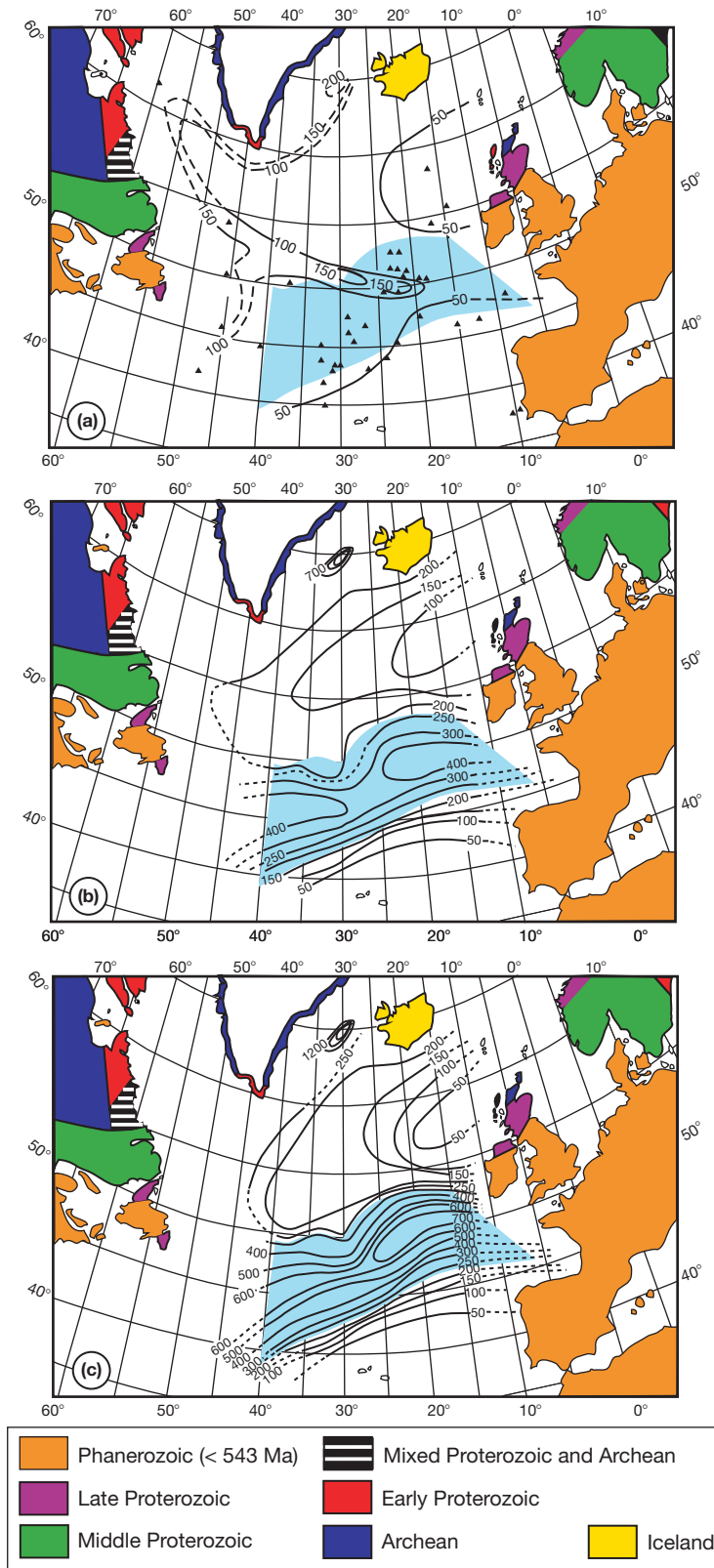


Figure 7 Maps of IRD flux in the North Atlantic. Units are $\text{mg cm}^{-2} \text{ ka}^{-1}$ of terrigenous silicates in the $>63 \mu\text{m}$ fractions. (a) 125–115 ka, (b) 40–25 ka, (c) 25–13 ka (Last Glacial Maximum). The zone with Last Glacial Maximum flux of $>250 \text{ mg cm}^{-2} \text{ ka}^{-1}$ is shown on each panel for comparison purposes. Reproduced from Ruddiman WF (1977) Late quaternary deposition of ice-rafted sand in subpolar North-Atlantic (Lat. 40° to 65°N). *Geological Society of America Bulletin* 88(12): 1813–1827; Hemming SR (2004) Heinrich events: Massive late Pleistocene detritus layers of the North Atlantic and their global climate imprint. *Reviews of Geophysics* 42(1).

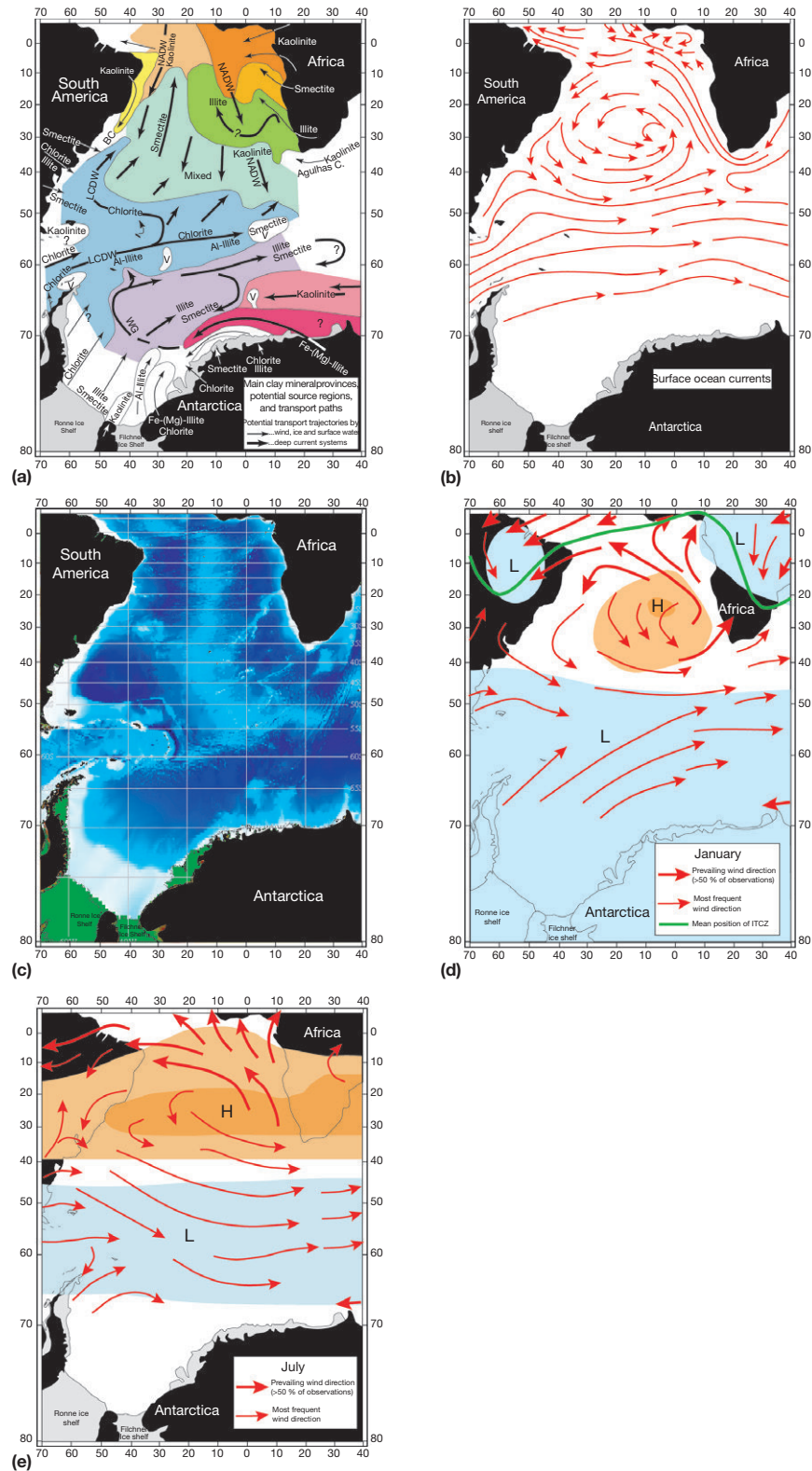


Figure 8 Maps of clay mineral distribution in the South Atlantic, along with regional deep currents and local, near-continent surface currents (a) (reproduced from [Petschick R, Kuhn G, and Franz G \(1996\)](#) Clay mineral distribution in surface sediments of the South Atlantic: Sources, transport, and relation to oceanography. *Marine Geology* 130: 203–229), surface currents in the South Atlantic (b) (reproduced from Open University Oceanography Course Team (1989) *Ocean Circulation*. Oxford/New York: Pergamon Press; Milton Keynes), bathymetry of the South Atlantic (c), January winds (d), and July winds (e) (reproduced from Open University Oceanography Course Team (1989) *Ocean Circulation*. Oxford/New York: Pergamon Press; Milton Keynes). The distribution of clay provinces in the South Atlantic appears to be controlled by a combination of surface and deep currents and possibly surface winds.

irradiate the specimen and to cause the elements in the specimen to emit (or fluoresce) their characteristic X-rays. XRF is particularly well-suited for investigations that involve bulk chemical analyses of major elements (Si, Ti, Al, Fe, Mn, Mg, Ca, Na, K, and P) and trace elements (in abundances >1 ppm; Ba, Ce, Co, Cr, Cu, Ga, La, Nb, Ni, Rb, Sc, Sr, Rh, U, V, Y, Zr, and Zn) in sediment. In most cases, the sample is ground to a fine powder, mixed with a chemical flux, and melted in a furnace to create a homogenous glass (often known as a fused bead) that can be analyzed and the abundances of the elements calculated. This type of analysis is limited to relatively large discrete samples, typically >1 g.

XRF scanners are part of emerging technology that allows rapid semiquantitative determination of elemental concentrations in marine sediments at resolutions down to 100 μm . This has important implications for high-resolution surveying of marine sediment cores. An important example is the recognition of greatly increased flux of terrigenous sediments based on the Fe/Ca ratio of sediments from the Brazil margin, at the approximate times of Heinrich Events (massive iceberg events in the North Atlantic), that correlated to rapid climate variability in many parts of the world (Arz et al., 1998; Hemming, 2004). Another fundamental example is the correlation of Ti concentrations, as a proxy for rainfall in the Cariaco Basin, with collapse of the Mayan civilization (Haug et al., 2003) used terrigenous sediment compositions to evaluate variations in rainfall in the source areas to the Cariaco Basin, South America. In this case, the bulk titanium content of undisturbed sediment reflects variations in riverine input and the hydrological cycle over northern tropical South America and ties drought to the collapse of the Mayan civilization. With recent efforts to better calibrate the data from XRF scanners to elemental concentrations (e.g., Böning et al., 2007; Weltje and Tjallingii, 2008) we should expect to see more widespread use of this technique in paleoceanography in the future.

Isotope Ratios

Isotope ratios are measured by MS, though the exact technique depends on the isotope system of interest. For provenance work, most methods involve sequential leaching, sediment digestion, and separation chemistry prior to analysis by MS. Common isotope systems such as Rb–Sr and Sm–Nd can be measured by multicollector ICP–MS or thermal ionization MS, while Ar–Ar requires a gas source MS. Constant flux proxies must be measured on bulk sediment samples; $^{230}\text{Th}_{\text{xs}}$ measurements are easily made by ICP–MS, while ^3He measurements require a special noble gas MS.

Grain Size Distributions

Grain size measurements can be made by separating a sample into different grain size fractions and recording the weight of each. Sand sized particles (>63 μm) can be sieved dry, while coarse silt (20–63 μm) must be wet sieved. Below 20 μm , fine silt can be further separated based on settling velocity according to Stokes' Law. Various devices are available for making more detailed measurements of grain size distributions, including SS mean and SS%, and these have been reviewed in detail (Bianchi et al., 1999; McCave and Hall, 2006; McCave et al.,

2006). The instruments of choice for the study of deep-sea sediments as proxies for current intensity are the Sedigraph, which uses X-ray scanning to measure the gravity-induced settling velocities of different size particles in a liquid with known properties and the Coulter counter, passes particles suspended in a weak electrolyte solution through small apertures, and electrically senses the volume of electrolyte displaced (Bianchi et al., 1999; McCave and Hall, 2006; McCave et al., 2006).

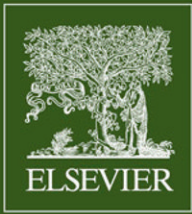
See also: K/Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ Dating; **Glacial Landforms, Sediments:** Glaciomarine Sediments and Ice-Rafted Debris. **Paleoceanography:** Paleoceanography An Overview. **Permafrost and Periglacial Features:** Rock Weathering.

References

- Albarede F (2003) *Geochemistry: An Introduction*. Cambridge: Cambridge University Press.
- Arz HW, Pätzold J, and Wefer G (1998) Correlated millennial-scale changes in surface hydrography and terrigenous sediment yield inferred from last-glacial marine deposits off Northeastern Brazil. *Quaternary Research* 50(2): 157–166.
- Bacon MP (1984) Glacial to interglacial changes in carbonate and clay sedimentation in the Atlantic–Ocean estimated from Th-230 measurements. *Isotope Geoscience* 2(2): 97–111.
- Bayon G, German CR, Boella RM, Milton JA, Taylor RN, and Nesbitt RW (2002) An improved method for extracting marine sediment fractions and its application to Sr and Nd isotopic analysis. *Chemical Geology* 187(3–4): 179–199.
- Bianchi GG, Hall IR, McCave IN, and Joseph L (1999) Measurement of the sortable silt current speed proxy using the Sedigraph 5100 and Coulter Multisizer II: Precision and accuracy. *Sedimentology* 46(6): 1001–1014.
- Biscaye PE (1965) Mineralogy and sedimentation of recent deep-sea clay in Atlantic Ocean and adjacent seas and oceans. *Geological Society of America Bulletin* 76(7): 803–832.
- Bond G, Heinrich H, Broecker W, et al. (1992) Evidence for massive discharges of icebergs into the North Atlantic ocean during the last glacial period. *Nature* 360: 245–249.
- Bond GC and Lotti R (1995) Iceberg discharges into the North Atlantic on millennial time scales during the last deglaciation. *Science* 267: 1005–1010.
- Böning P, Bard E, and Rose J (2007) Toward direct, micron-scale XRF elemental maps and quantitative profiles of wet marine sediments. *Geochemistry, Geophysics, Geosystems* 8(5): Q05004.
- Bowen NL (1928) *The Evolution of the Igneous Rocks*. Princeton, NJ: Princeton University Press.
- Dasch EJ (1969) Strontium isotopes in weathering profiles, deep-sea sediments, and sedimentary rocks. *Geochimica et Cosmochimica Acta* 33: 1521–1552.
- Diekmann B, Falker M, and Kuhn G (2003) Environmental history of the south-eastern South Atlantic since the Middle Miocene: Evidence from the sedimentological records of ODP Sites 1088 and 1092. *Sedimentology* 50: 511–529.
- Diekmann B and Kuhn G (2002) Sedimentary record of the mid-Pleistocene climate transition in the southeastern South Atlantic (ODP Site 1090). *Paleogeography, Paleoclimatology, Paleoeocology* 182: 241–258.
- Diekmann B, Kuhn G, Rachold V, et al. (2000) Terrigenous sediment supply in the Scotia Sea (Southern Ocean): Response to Late Quaternary ice dynamics in Patagonia and on the Antarctic Peninsula. *Paleogeography, Paleoclimatology, Paleoeocology* 162(3–4): 357–387.
- Francois R, Frank M, van der Loeff MMR, and Bacon MP (2004) Th-230 normalization: An essential tool for interpreting sedimentary fluxes during the late Quaternary. *Paleoceanography* 19(1).
- Franzese AM, Hemming SR, Goldstein SL, and Anderson RF (2006) Reduced Agulhas leakage during the last glacial maximum inferred from an integrated provenance and flux study. *Earth and Planetary Science Letters* 250(1–2): 72–88.
- Gibbs RJ (1967) Geochemistry of Amazon River system. I. Factors that control salinity and composition and concentration of suspended solids. *Geological Society of America Bulletin* 78(10): 1203–1232.
- Goldstein SJ and Jacobsen SB (1988) Nd and Sr isotopic systematics of river water suspended material: Implications for crustal evolution. *Earth and Planetary Science Letters* 87: 249–265.

- Goldstein SL, O'Nions RK, and Hamilton PJ (1984) A Sm–Nd isotopic study of atmospheric dusts and particulates from major river systems. *Earth and Planetary Science Letters* 70: 221–236.
- Grousset FE and Biscaye PE (2005) Tracing dust sources and transport patterns using Sr, Nd and Pb isotopes. *Chemical Geology* 222(3–4): 149–167.
- Grousset FE, Biscaye PE, Zindler A, Prosper J, and Chester R (1988) Neodymium isotopes as tracers in marine sediments and aerosols: North Atlantic. *Earth and Planetary Science Letters* 87: 367–378.
- Hall IR and McCave NI (2000) Paleocurrent reconstruction, sediment and thorium focussing on the Iberian margin over the last 140 ka. *Earth and Planetary Science Letters* 178: 151–164.
- Haug GH, Gunther D, Peterson LC, Sigman DM, Hughen KA, and Aeschlimann B (2003) Climate and the collapse of Maya civilization. *Science* 299(5613): 1731–1735.
- Hemming SR (2004) Heinrich events: Massive late Pleistocene detritus layers of the North Atlantic and their global climate imprint. *Reviews of Geophysics* 42: RG1005, 43 pp.
- Hemming SR and McLennan SM (2001) Pb isotope compositions of modern deep sea turbidites. *Earth and Planetary Science Letters* 184(2): 489–503.
- Henderson GM and Anderson RF (2003) The U-series toolbox for paleoceanography. *Uranium-Series Geochemistry* 52: 493–531.
- Hurley PM, Heezen BC, Pinson WH, and Fairbairn HW (1963) K–Ar age values in pelagic sediments of the North-Atlantic. *Geochimica et Cosmochimica Acta* 27: 393–399.
- Jones CE, Halliday AN, Rea DK, and Owen RM (1994) Neodymium isotopic variations in North Pacific modern silicate sediment and the insignificance of detrital REE contributions to seawater. *Earth and Planetary Science Letters* 127: 55–66.
- Kennett JP (1982) *Marine Geology*. Englewood Cliffs, NJ: Prentice Hall.
- Kohfeld KE and Harrison SP (2001) DIRTMAP: The geological record of dust. *Earth-Science Reviews* 54(1–3): 81–114.
- Kuhn G and Diekmann B (2002) Late Quaternary variability of ocean circulation in the southeastern South Atlantic inferred from the terrigenous sediment record of a drift deposit in the southern Cape Basin (ODP Site 1089). *Palaeogeography, Palaeoclimatology, Palaeoecology* 182: 287–303.
- Latimer JC, Filippelli GM, Hendy IL, Gleason JD, and Blum JD (2006) Glacial–interglacial terrigenous provenance in the southeastern Atlantic Ocean: The importance of deep-water sources and surface currents. *Geology* 34(7): 545–548.
- Liu LW, Chen J, Ji JF, Chen Y, and Balsam W (2006) Variation of Zr/Rb ratios in the Chinese loess deposits during the past 18 Myr and its implication for the change of East Asian monsoon intensity. *Geochemistry, Geophysics, Geosystems* 7.
- McCave IN and Hall IR (2006) Size sorting in marine muds: Processes, pitfalls, and prospects for paleoflow-speed proxies. *Geochemistry, Geophysics, Geosystems* 7.
- McCave IN, Hall IR, and Bianchi GG (2006) Laser vs. settling velocity differences in silt grain size measurements: Estimation of palaeocurrent vigour. *Sedimentology* 53(4): 919–928.
- McLennan SM, Hemming S, McDaniel DK, and Hanson GN (1993) Geochemical approaches to sedimentation, provenance, and tectonics. *Geological Society of America* 284: 21–40 (Special Paper).
- McLennan SM and Taylor SR (1980) Th and U in sedimentary rocks: Crustal evolution and sedimentary recycling. *Nature* 285: 621–624.
- McLennan SM, Taylor SR, McCulloch MT, and Maynard JB (1990) Geochemical and Nd–Sr isotopic composition of deep-sea turbidites – Crustal evolution and plate tectonic associations. *Geochimica et Cosmochimica Acta* 54(7): 2015–2050.
- Nesbitt HW and Young GM (1982) Early Proterozoic climates and plate motions inferred from major element chemistry of lutites. *Nature* 299(5885): 715–717.
- Open University Oceanography Course Team (1989) *Ocean Circulation*. Oxford: Pergamon Press New York: Milton Keynes.
- Ozima M, Takayanagi M, Zashu S, and Amari S (1984) High He-3–He-4 ratio in ocean sediments. *Nature* 311(5985): 448–450.
- Petschick R, Kuhn G, and Franz G (1996) Clay mineral distribution in surface sediments of the South Atlantic: Sources, transport, and relation to oceanography. *Marine Geology* 130: 203–229.
- Rea DK (1994) The paleoclimatic record provided by eolian deposition in the deep-sea – The geologic history of wind. *Reviews of Geophysics* 32(2): 159–195.
- Rea DK and Hovan SA (1995) Grain-size distribution and depositional processes of the mineral component of abyssal sediments – Lessons from the North Pacific. *Paleoceanography* 10(2): 251–258.
- Revel M, Grousset FE, and Labeyrie L (1996a) Grain-size and Sr–Nd isotopes as a tracer of paleo-bottom current strength, Northeast Atlantic Ocean. *Marine Geology* 131(3–4): 233–249.
- Revel M, Sinko JA, Grousset FE, and Biscaye PE (1996b) Sr and Nd isotopes as tracers of North Atlantic lithic particles: Paleoclimatic implications. *Paleoceanography* 11(1): 95–112.
- Ruddiman WF (1977) Late quaternary deposition of ice-rafted sand in subpolar North-Atlantic (Lat. 40° to 65°N). *Geological Society of America Bulletin* 88(12): 1813–1827.
- Taylor SR and McLennan SM (1985) *The Continental Crust: Its Composition and Evolution*. Oxford: Blackwell Scientific Publications Ltd..
- Walter HJ, Hegner E, Diekmann B, Kuhn G, and van der Loeff MMR (2000) Provenance and transport of terrigenous sediment in the South Atlantic Ocean and their relations to glacial and interglacial cycles: Nd and Sr isotopic evidence. *Geochimica et Cosmochimica Acta* 64(22): 3813–3827.
- Weltje GJ and Prins MA (2003) Muddled or mixed? Inferring palaeoclimate from size distributions of deep-sea clastics. *Sedimentary Geology* 162(1–2): 39–62.
- Weltje GJ and Tjallingii R (2008) Calibration of XRF core scanners for quantitative geochemical logging of sediment cores: Theory and application. *Earth and Planetary Science Letters* 274(3–4): 423–438.





VOLUME THREE

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2ND EDITION

SCOTT A. ELIAS EDITOR-IN-CHIEF
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SECOND EDITION

Volume 3

PAL–POL

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DEDICATIONS

The second edition of the Encyclopedia of Quaternary Science is dedicated to the memory of three outstanding scientists who have recently passed away: Russell Coope, Bill Watts, and Aleksis Dreimanis.

Russell Coope

G. Russell Coope (1930–2011) was essentially the founding father of Quaternary Entomology, having initiated the modern era of Quaternary insect fossil research in the late 1950s. Prior to that time, and going back to the nineteenth century, there had been a number of false starts and stops in this field. Previous researchers had assumed that any ice-age insect fossils they found would necessarily represent extinct taxa. This, as it turns out, was a completely false assumption. Beginning with the Late Pleistocene Upton Warren and Chelford sites, Russell was able to determine that fossil insects, especially beetles (Coleoptera) from Pleistocene deposits, represent extant species, albeit species that may live in dramatically different regions today than those where they lived in the past. For many years at the start of this endeavor, Russell was ‘a voice in the wilderness,’ having to convince colleagues in both the Quaternary sciences and modern Entomology that it is possible to identify ‘beetle bits’ (individual sclerites of beetle exoskeleton) to the species level, and that fossil beetle assemblages can yield reliable reconstructions of past environments. One of his greatest battles was to convince skeptical palynologists that the interstadial interval in what is now known as MIS 3, and another one toward the end of the last glaciation, were almost as warm as today, in spite of the fact that the vegetation cover represented in pollen records failed to register this. Likewise, Russell championed the idea that climate oscillations, particularly in the late glacial interval and early Holocene, were extremely rapid and dramatic: switching from full glacial to full interglacial conditions in a matter of decades, at most. Russell’s paleoclimate reconstructions were finally given full credence by the scientific community at large when the oxygen isotope records from Greenland ice cores began to be published, showing nearly the exact same amplitude and timing of climate change he had been publishing since the 1960s.

Russell’s research has had important implications for the field of zoogeography. On the basis of his work, we now know that insects respond to climate change chiefly by shifting their distributional ranges, so that they



Figure 1 Russell Coope at the Chelford quarry fossil site, August 1977. Photograph by Alan V. Morgan, used with permission.

track these changes through space. Through his tenacity and hard work, Russell was able to track down the modern ranges of beetle species he found in Pleistocene deposits to such remote locations as northeastern Siberia and the Tibetan Plateau (cold-adapted species), as well as North Africa and the Mediterranean islands (warm-adapted species).

Russell brought a level of curiosity and enthusiasm for natural history and all things scientific that was infectious. He was a brilliant lecturer and storyteller, holding his audiences practically spellbound. He trained many of the paleoentomologists who took the field forward in recent decades. His seemingly boundless energy kept him working and publishing papers, right up to the end of his life. He published more than 200 papers in his long career. He collaborated with dozens of other Quaternary scientists and was one of the first to appreciate and promote multidisciplinary studies of Pleistocene fossil sites. There is no doubt that his research and publishing legacies will continue to inspire future generations of paleoentomologists.

Bill Watts

William (Bill) A. Watts (1930–2010) was a towering figure in Quaternary paleoecology. He was a skilled botanist and ecologist and this was a secret to his success in unraveling the interglacial, late-glacial, and Holocene vegetation history not only in his native Ireland but also in the midwestern, the southeastern, and eastern USA, and in Italy. A major contribution was his reintroduction and development of quantitative plant macrofossil analysis. Macrofossils can often be identified to lower taxonomic levels than pollen, and Bill took full advantage of the complementarity of both methods. Thus, he was able to determine important features of vegetation history by identifying trees to species level and to illuminate the history of aquatic plants and local vegetational developments.



Figure 2 Photograph of Bill Watts, taken at Laguna de la Roya, Spain. Photograph by Fraser Mitchell, used with permission.

Bill always enjoyed fieldwork and microscope work, carrying out the latter in spare time left over from his job as Provost of Trinity College Dublin and his participation on committees dealing with hospitals and conservation in Ireland. Bill's quiet but intense enthusiasm and his great sense of humor added much to discussions, macrofossil identification sessions, and excursions with him. He and his work are inspirations to those who follow him.



Figure 3 Aleksis Dreimanis, photograph by Stephen Hicock.

Aleksis Dreimanis

Aleksis Dreimanis (1914–2011) was widely considered the world's foremost Quaternary glacial geologist in the 1970s and 80s. Aleksis was born in Latvia. While still an undergraduate at the University of Latvia, he published his first paper on the structural geology of deformed glacial deposits along the banks of the River Daugava – a study that he would return to and revise 60 years later. In 1944, Aleksis was conscripted as a military geologist by the Germans who occupied Latvia. For the next two years, he was separated from his family and forced to do strategic geological mapping in eastern Germany and northern Italy until the end of WWII, followed by internment in prisoner of war and displaced persons camps. In 1948, he escaped to London, England, where he met with the President of the University of Western Ontario who gave him a job in London, Canada (his family was able to join him the following year).

Over the next six decades at the University of Western Ontario, Aleksis ascended to the top of his discipline while gaining international recognition through landmark publications and inspired leadership on committees. His 1970s synthesis and definition of the Wisconsin Stage with Paul Karrow, Jan Terasmae, Dick Goldthwait, and Jane Forsyth became a world standard to which other records were compared. In 1989, he expanded this synthesis and introduced the conceptual till tetrahedron that showed the till continuum with end members occupying the corners of an inverted pyramid. He was among the first geologists to use radiocarbon dating, and he provided samples for inter-lab calibration. He was among the first to quantitatively study the relationship between till lithology and till formation, and to investigate how glacial transport and dynamics determine till texture, deformation, and rheologic states. Aleksis produced over 200 publications spanning eight decades and over a wide range of topics. Largely because of his influence, the study of till genesis and its relation to till properties and past glacial behavior has become a major focus of glacial geologic research in recent decades. The contribution and legacy that forms the body of work of Dreimanis will continue as a major benchmark within the diverse fields of scientific enquiry that are sometimes loosely referred to as 'glacial geology.'

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FOREWORD

As with the publication of the first edition in 2007, the publication of the second edition of the *Encyclopedia of Quaternary Science* represents a landmark in the history of publishing in the field of Quaternary Science. Quaternary Science is a multidisciplinary endeavor which seeks to establish as detailed a picture as possible of the manifold environmental changes that have occurred during the most recent geological period, the Quaternary – an interval of time that spans the past 2.59 million years or the past 0.056% of geological time. It is a period of significant climate and environmental change and witnessed the widespread dispersal of our species, *Homo sapiens*, across the planet.

Since Louis Agassiz and Reverend William Buckland traipsed over parts of the Scottish landscape in 1840 in search of evidence of glaciation, a huge literature has emerged on the science of long-term climate and environmental change. Rapid technological advances in the late twentieth century and the proliferation of scientific journals, particularly in an era of electronic publishing, have resulted in an exponential growth and documentation of knowledge on the climate and environmental changes that have occurred during the Quaternary period. More recently, there has been an increasing public interest in applied Quaternary research as a framework for understanding the basis for recent climate changes and for understanding the nature and frequency of geological hazards and vexing issues such as soil erosion and land degradation, and the adverse effects of ocean water temperature increases and acidification on coral reef environments. In a similar manner, the likely magnitude of future sea-level rise and the associated impacts on coastal landscapes in the twenty-first century have attracted wide public interest. In this sense, Quaternary Science is very much on the political agenda and is a critically important subject to address issues of public concern.

The *Encyclopedia of Quaternary Science* edited by Scott Elias of Royal Holloway, University of London, UK, accordingly represents a particularly welcome addition to the literature. The encyclopedia presents an up-to-date and authoritative overview of Quaternary Science. The encyclopedia should enjoy a wide readership as the entries are presented in a very clear and easily readable style. The text of the articles is written at a level that allows undergraduate students to understand the material, while providing active researchers with a ready reference resource for information in the field. Each entry of up to 4000 words covers the salient points of each topic with very clear illustrations. A central theme that pervades the work is the importance of Quaternary Science in providing an historical context for assessing present environmental changes and as basis for modeling potential future changes.

The encyclopedia consists of four volumes in print form and is available electronically. All the entries have been updated and the text totals around 3,500,000 words. As a major reference work, the encyclopedia has a very wide coverage of topics within the Quaternary sciences reflecting the complex and interdisciplinary nature of the science. Each of the major sections begins with a general overview of the topic prepared by a leading expert in the field. The major sections, for example, examine the analytical methods commonly used in paleoenvironmental reconstructions to unravel in a forensic-like manner the nature of former environments and the tempo of environmental change. Accordingly, a great range of topics such as the former extent of ice cover and nature of Quaternary glaciations, the biological responses and resultant fossil records to fluctuating climate, the expansion and contraction of desert environments, and global and local changes in relative sea levels are examined. Other topics covered include dating techniques, Quaternary stratigraphy, fluvial environments, lake level studies, paleosols, paleobotany, ancient DNA, paleolimnology, vertebrate studies, insect fossil studies, paleoceanography, stable isotope studies, ice core records, and human evolution in the Quaternary. One of the new sections in the encyclopedia examines the application of Quaternary proxy evidence in forensic science. All sections provide a clear summary of the latest advances in the fields of research.

This is an outstanding work and the editors and the publisher are to be congratulated for producing an encyclopedia that cogently summarizes the current state of the science.

A handwritten signature in black ink, appearing to read 'C. Murray-Wallace', with a stylized flourish at the end.

Colin V. Murray-Wallace
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INTRODUCTION

It has been my privilege to work with more than 420 leading Quaternary scientists in developing the second edition of the *Encyclopedia of Quaternary Science*. This team of writers and editors represents 28 countries in Europe, Asia, Africa, the Americas, Australia, and New Zealand. Starting with the first edition in 2006, I have had my finger on the pulse of Quaternary science, and this branch of science is truly pulsating! Information now comes from an incredible variety of disciplines: geochemistry, numerical modeling, history, vulcanology, paleobiology, nuclear physics, stratigraphy, sedimentology, climatology, anthropology, archaeology, glacial geology, soil science, ice-stream modeling – the list is staggering. This highly disparate group of people are bound together by one common thread: the desire to know the history of the planet during the last 2.6 million years – the time of the ice ages. For Quaternary scientists, this is a pressing need, not an idle curiosity. Any doubts about this statement can be easily dispelled by a consideration of the lengths to which many of them go to gather the necessary data. Some of them have worked for months in sub-zero temperatures on top of very high mountains, or near the center of polar ice sheets, collecting ice cores. Others have spent many weeks on some of the roughest seas in the world, drilling deep-sea sediment cores. Often the work is more mundane. An oxygen isotope curve for a lengthy marine sediment core represents thousands of hours of patiently picking tiny fossils from layer after layer of sediments, in order to obtain sufficient numbers of calcium carbonate shells to yield samples for isotopic analysis. A map showing proposed ice limits from the last glaciation represents thousands of hours of field mapping of glacial features by dozens of people. Why do all of these people devote their lives to this pursuit of knowledge? Does it really matter so much? The answer becomes clear when you step back and examine the topic of Quaternary science in its proper context.

The world we inhabit has largely been shaped by the events of the Quaternary. All the biological communities that exist today are the end product of a long series of species associations that came together in the past, largely driven by climatic change during the Pleistocene. We cannot properly understand the functioning of modern ecosystems without a solid knowledge of their history, any more than we can understand the plot of a long novel by reading just the final page. We are also living in a time of alarming climate changes. Even though the pace and intensity of some of these changes have not been seen in historical times, there were many rapid, large-scale climatic shifts in the Pleistocene. The best way to predict the effects of global warming on the planet's climate and ecosystems is to look at the history of similarly intense, rapid changes in the prehistoric past. The interval that is most relevant today is the most recent geologic period: the Quaternary. As human populations rise exponentially, increasing numbers of people are exposed to geologic hazards, such as earthquakes (and attendant tsunamis), and volcanic eruptions. These are short-lived events that take place only rarely in any one region. The interval between major events, such as volcanic eruptions, may be centuries or millennia. How do we come to grips with predicting the future likelihood of such erratic phenomena? Again, the answers come from piecing together the ancient history of such events, over many thousands of years.

The Quaternary has been the time when our own species came of age. The beginning of the Quaternary, roughly 2.6 million years ago, was about the time when the earliest member of our genus (*Homo*) first appeared in Africa. Pleistocene environments shaped the course of human evolution, culminating in anatomically modern *Homo sapiens* spreading from Africa throughout most of the world during the last glaciation. Even though human beings largely shape their own environments today, for the vast majority of our species' history, it has been the environment that has shaped us. Our direct ancestors' adaptations to environmental change are deeply ingrained in our genes. Thus, an understanding of the environmental conditions that shaped our species is critical to our understanding of humanity.

Quaternary science is a rapidly changing field, and the articles that appear in this encyclopedia reflect this. New dating techniques, such as cosmogenic nuclide dating, are revolutionizing our understanding of many earth surface processes. The ability to analyze increasingly smaller samples for radiocarbon and stable isotopes of oxygen and hydrogen means that we are gaining a level of precision in the reconstruction of past events that was unheard of just a few years ago. Stable isotope studies of air bubbles trapped in ice cores from Greenland and Antarctica have given Quaternary scientists an entirely new perspective on the rapidity and intensity of climatic change during the last glacial cycle and beyond. Likewise, the discovery of long sequences of annually laminated sediments in both marine and freshwater environments has provided a great leap forward in our ability to resolve the timing of environmental changes in nonpolar regions. The ability to extract and analyze ancient DNA sequences from Pleistocene fossils (both plants and animals) is revolutionizing the field of paleobiology. We are beginning to be able to trace the genetic lineages of a number of different organisms, from beetles to bison. In short, these are very exciting times to be a Quaternary scientist! While it is virtually impossible for any Quaternary researcher or student to keep abreast of all the new discoveries in this multifaceted science, this encyclopedia can be of great help. The articles contained here represent the state of the art in a huge variety of topics, and they offer the opportunity to dig deeper into their respective subjects by providing full citations of the most pertinent literature available. I invite you to come and explore the Quaternary Period in the pages that follow. It is a fascinating story.

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HOW TO USE THE ENCYCLOPEDIA

The material in this encyclopedia is organized into two broad sections. At the start of Volume 1 are four *Introductory Articles* that serve as a broad introduction to the field. It is intended that these articles should serve as the first port of call for readers, to enable them to embark on any of the main entries without the need to consult other material beforehand.

Following the Introductory Articles, the main body of The Encyclopedia is arranged as a series of entries in alphabetical order. These entries fill the remainder of the volumes.

To help realize the full potential of the material in the Encyclopedia we have provided four features to help you find the topic of your choice: a contents list by subject; an alphabetical contents list; cross-references to other articles; and a full subject index.

1. Contents list by subject

Your first point of reference will probably be the contents list by subject. This list appears at the front of each volume, and groups the entries under subject headings describing the broad themes of Quaternary science. This will enable the reader to make quick connections between entries and to locate the entry of interest. Under each main section heading, you will find several subject areas and under each subject area is a list of those entries that covers aspects of that subject, together with the volume and page numbers on which these entries may be found.

2. Alphabetical contents list

The alphabetical contents list, which also appears at the front of each volume, lists the entries in the order in which they appear in the Encyclopedia. This list provides both the volume number and the page number of each entry.

You will find ‘dummy’ entries where obvious synonyms exist for entries or where we have grouped together related topics. Dummy entries appear in both the contents list and the body of the text.

Example

If you were attempting to locate material on *peat studies* via the alphabetical contents list:

PEAT STUDIES *see* PLANT MACROFOSSIL METHODS AND STUDIES: Mire and Peat Macros

The dummy entry directs you to the entry in which *peat* is covered.

If you were trying to locate the material by browsing through the text and you had looked up *Peat Studies*, then the following information would be provided in the dummy entry:

Peat Studies *see* PLANT MACROFOSSIL METHODS AND STUDIES: Mire and Peat Macros

3. Cross-references

All of the entries in the Encyclopedia have been extensively cross-referenced. The cross-references, which appear at the end of the entry, serve three different functions:

- i. To indicate if a topic is discussed in greater detail elsewhere.
- ii. To draw the reader’s attention to parallel discussions in other entries.
- iii. To indicate material that broadens the discussion.

Example

The following list of cross-references appear at the end of the entry GLACIAL LANDFORMS, ICE SHEETS | Growth and Decay

See also: **Introduction:** History of Quaternary Science; **Paleoclimate Reconstruction:** Sub-Milankovitch (DO/Heinrich) Events; **Glacial Climates:** Thermohaline Circulation; **Glaciation, Causes:** Astronomical Theory of Paleoclimates; **Paleoclimate Modeling:** Last Glacial Maximum GCMs; **Glaciations:** Overview; **Sea Level Studies:** Overview; Paleocceanography; **Ice Core Records:** Correlations Between Greenland and Antarctica.

Here you will find examples of all three functions of the cross-reference list: a topic discussed in greater detail elsewhere (e.g., **Paleoclimate Modeling:** Last Glacial Maximum GCMs), parallel discussion in other entries (e.g., **Glacial Climates:** Thermohaline Circulation), and reference to entries that broaden the discussion (e.g., **Glaciations:** Overview).

4. Index

The index provides you with the page number where the material is located. The index entries differentiate between material that is a whole entry, is part of an entry, or is data presented in a figure or a table. Detailed notes are provided on the opening page of the index.

5. Contributors

A full list of contributors is listed at the beginning of each volume.

CONTENTS LIST BY SUBJECT

Location references refer to the volume number and page number (separated by a colon)

Introduction and History of Science

INTRODUCTION | History of Quaternary Science, 1:1
INTRODUCTION | History of Dating Methods, 1:9
INTRODUCTION | Societal Relevance of Quaternary Research, 1:17
INTRODUCTION | Understanding Quaternary Climatic Change, 1:26

Paleoclimatology

Overview and Approaches

PALEOCLIMATE | Introduction, 3:87
PALEOCLIMATE RECONSTRUCTION | Approaches, 3:179
PALEOCLIMATE MODELING | Data-Model Comparisons, 3:135
PALEOCLIMATE | Timescales of Climate Change, 3:93
PALEOCLIMATE | Modern Analog Approaches in Paleoclimatology, 3:102
PALEOCLIMATE | Paleoclimate History of the Arctic, 3:113
PALEOCLIMATE | The Younger Dryas Climate Event, 3:126

Paleoclimate Reconstructions

PALEOCLIMATE RECONSTRUCTION | Pliocene Environments, 3:185
GLACIATION, CAUSES | Tectonic Uplift and Continental Configurations, 2:127
GLACIATION, CAUSES | Astronomical Theory of Paleoclimates, 2:136
PALEOCLIMATE RECONSTRUCTION | Paleodroughts and Society, 3:194
PALEOCLIMATE RECONSTRUCTION | Paleotempestology, 3:209

PALEOCLIMATE RECONSTRUCTION | Sub-Milankovitch (DO/Heinrich) Events, 3:200
PALEOCLIMATE RECONSTRUCTION | Younger Dryas Oscillation, Global Evidence, 3:222
PALEOCLIMATE RECONSTRUCTION | Historical Climatology, 3:237
PALEOCLIMATE RECONSTRUCTION | The Last Millennium, 3:229
PALEOCLIMATE RELEVANCE TO GLOBAL WARMING, 3:244

Climate Forcing Mechanisms

GLACIAL CLIMATES | Thermohaline Circulation, 1:737
GLACIAL CLIMATES | Biosphere Feedbacks, 1:721
GLACIAL CLIMATES | Volcanic and Solar Forcing, 1:748
GLACIAL CLIMATES | Effects of Atmospheric Dust, 1:729

Climate Modelling

PALEOCLIMATE MODELING | Quaternary Environments, 3:147
PALEOCLIMATE MODELING | The Last Interglacial, 3:155
PALEOCLIMATE MODELING | Last Glacial Maximum GCMs, 3:165
PALEOCLIMATE MODELING | Paleo-ENSO, 3:171

Dating Techniques

Radiometric Methods

DATING TECHNIQUES, 1:447
RADIOCARBON DATING | Conventional Method, 4:305
RADIOCARBON DATING | AMS Radiocarbon Dating, 4:316

RADIOCARBON DATING | Sources of Error, 4:324
 RADIOCARBON DATING | Variations in Atmospheric ^{14}C , 4:329
 RADIOCARBON DATING | Causes of Temporal ^{14}C Variations, 4:336
 RADIOCARBON DATING | Calibration of the ^{14}C Record, 4:345
 RADIOCARBON DATING | Charcoal, 4:353
 RADIOCARBON DATING | ^{14}C of Plant Macrofossils, 4:361
 K/Ar AND $^{40}\text{Ar}/^{39}\text{Ar}$ DATING, 2:477
 U-SERIES DATING, 4:567
 FISSION-TRACK DATING, 1:643
 LUMINESCENCE DATING | Thermoluminescence, 2:643
 LUMINESCENCE DATING | Optical Dating, 2:653
 LUMINESCENCE DATING | Electron Spin Resonance Dating, 2:667
 COSMOGENIC NUCLIDE DATING | Overview, 1:407
 COSMOGENIC NUCLIDE DATING | Cosmic Ray Interactions in Minerals, 1:418
 COSMOGENIC NUCLIDE DATING | Methods, 1:410
 COSMOGENIC NUCLIDE DATING | Exposure Geochronology, 1:432
 COSMOGENIC NUCLIDE DATING | Landscape Evolution, 1:440

Geomagnetic Excursions and Secular Variations in the Magnetic Field

Geomagnetic Excursions and Secular Variations, 1:705

Biological Methods

LICHENOMETRY, 2:565

Annual Layer Methods

VARVED MARINE SEDIMENTS, 4:582
 VARVED LAKE SEDIMENTS, 4:573

Chemical Methods

AMINO-ACID DATING, 1:37

Quaternary Stratigraphy

QUATERNARY STRATIGRAPHY | Tephrochronology, 4:277
 QUATERNARY STRATIGRAPHY | Overview, 4:189
 QUATERNARY STRATIGRAPHY | Lithostratigraphy, 4:227

QUATERNARY STRATIGRAPHY | Continental Biostratigraphy, 4:206
 QUATERNARY STRATIGRAPHY | Pedostratigraphy, 4:250
 QUATERNARY STRATIGRAPHY | Morphostratigraphy/Allostratigraphy, 4:243
 QUATERNARY STRATIGRAPHY | Climatostratigraphy, 4:222
 QUATERNARY STRATIGRAPHY | Sequence Stratigraphy, 4:260
 QUATERNARY STRATIGRAPHY | Chronostratigraphy, 4:215

Quaternary Glaciation

Glacial Landforms

GLACIAL LANDFORMS | Introduction, 1:755
 GLACIAL LANDFORMS, ICE SHEETS | Growth and Decay, 1:877
 GLACIAL LANDFORMS, ICE SHEETS | Evidence of Glacier and Ice Sheet Extent, 1:884
 GLACIAL LANDFORMS | Evidence of Glacier Recession, 1:803
 GLACIAL LANDFORMS, ICE SHEETS | PaleoeLAS, 1:909
 GLACIAL LANDFORMS | Glacitectonic Structures and Landforms, 1:839
 GLACIAL LANDFORMS, SEDIMENTS | Glaciogenic Lithofacies, 2:18
 GLACIAL LANDFORMS | Moraine Forms and Genesis, 1:769
 GLACIAL LANDFORMS, SEDIMENTS | Glaciofluvial Landforms of Deposition, 2:6
 GLACIAL LANDFORMS | Glaciofluvial Landforms of Erosion, 1:825
 GLACIAL LANDFORMS, SEDIMENTS | Glacial Erratics and Till Dispersal Indicators, 2:81
 GLACIAL LANDFORMS, SEDIMENTS | Glaciomarine Sediments and Ice-Rafted Debris, 2:30
 GLACIAL LANDFORMS, SEDIMENTS | Tills, 2:62
 GLACIAL LANDFORMS, SEDIMENTS | Till Fabric Analysis, 2:76
 GLACIAL LANDFORMS | Glacial Land Systems, 1:813
 GLACIAL LANDFORMS, EROSIONAL FEATURES | Micro- to Macroscale Forms, 1:865
 GLACIAL LANDFORMS, EROSIONAL FEATURES | Major Scale Forms, 1:847
 GLACIAL LANDFORMS, ICE SHEETS | Trimlines and Palaeonunataks, 1:918
 GLACIAL LANDFORMS | Quaternary Vulcanism: Subglacial Landforms, 1:780
 GLACIAL LANDFORMS, SEDIMENTS | Micromorphology of Glacial Sediments, 2:52

GLACIAL LANDFORMS, SEDIMENTS | Clast Form Analysis, 2:1
 GLACIAL LANDFORMS, ICE SHEETS | Evidence of Glacier Flow Directions, 1:895
 GLACIAL LANDFORMS, SEDIMENTS | Glaciolacustrine, 2:43
 GLACIAL LANDFORMS, SEDIMENTS | Glacial Sequence Stratigraphy, 2:85

Permafrost and Periglacial Features

PERMAFROST AND PERIGLACIAL FEATURES | Permafrost, 3:464
 PERMAFROST AND PERIGLACIAL FEATURES | Active Layer Processes, 3:421
 PERMAFROST AND PERIGLACIAL FEATURES | Paraglacial Geomorphology, 3:553
 PERMAFROST AND PERIGLACIAL FEATURES | Slope Deposits and Forms, 3:481
 PERMAFROST AND PERIGLACIAL FEATURES | Frost Mounds: Active and Relict Forms, 3:472
 PERMAFROST AND PERIGLACIAL FEATURES | Patterned Ground, 3:452
 PERMAFROST AND PERIGLACIAL FEATURES | Talus Slopes, 3:566
 PERMAFROST AND PERIGLACIAL FEATURES | Thermokarst Topography, 3:574
 PERMAFROST AND PERIGLACIAL FEATURES | Ice Wedges and Ice-Wedge Casts, 3:436
 PERMAFROST AND PERIGLACIAL FEATURES | Cryoturbation Structures, 3:430
 PERMAFROST AND PERIGLACIAL FEATURES | Blockfields (Felsenmeer), 3:523
 PERMAFROST AND PERIGLACIAL FEATURES | Block/Rock Streams, 3:514
 PERMAFROST AND PERIGLACIAL FEATURES | Rock Weathering, 3:500
 PERMAFROST AND PERIGLACIAL FEATURES | Rock Glaciers and Protalus Forms, 3:535
 PERMAFROST AND PERIGLACIAL FEATURES | Periglacial Fluvial Sediments and Forms, 3:490
 PERMAFROST AND PERIGLACIAL FEATURES | Permafrost and Glacier Interactions, 3:507
 PERMAFROST AND PERIGLACIAL FEATURES | Yedoma: Late Pleistocene Ice-Rich Syngenetic Permafrost of Beringia, 3:542

Tree-ring section

DENDROCHRONOLOGY, 1:453
 DENDROCLIMATOLOGY, 1:459
 GLACIAL LANDFORMS, TREE RINGS | Dendrogeomorphology, 2:91
 GLACIAL LANDFORMS, TREE RINGS | Dendroglaciology, 2:104

Fluvial Environments

FLUVIAL ENVIRONMENTS | Sediments, 1:663
 FLUVIAL ENVIRONMENTS | Terrace Sequences, 1:684
 Glacial–Interglacial Scale Fluvial Responses, 2:112
 FLUVIAL ENVIRONMENTS | Responses to Rapid Environmental Change, 1:676
 FLUVIAL ENVIRONMENTS | Deltaic Environments, 1:693
 PALEOHYDROLOGY, 3:253

History of Quaternary Glaciations

GLACIATIONS | Overview, 2:143
 GLACIATIONS | Transition from Late Neogene to Early Quaternary Environments, 2:151
 GLACIATIONS | Early Quaternary (Pleistocene) and Precursors, 2:167
 GLACIATIONS | Middle Pleistocene in Eurasia, 2:172
 GLACIATIONS | Mid-Quaternary in North America, 2:180
 GLACIATIONS | Middle Pleistocene Glaciations in the Southern Hemisphere, 2:187
 GLACIATIONS | Late Quaternary of Antarctica, 2:216
 GLACIATIONS | Late Quaternary in North America, 2:245
 GLACIATIONS | Late Quaternary in Highland Asia, 2:236
 GLACIATIONS | Late Quaternary of the Southwest Pacific Region, 2:202
 GLACIATIONS | Late Pleistocene in Eurasia, 2:224
 GLACIATIONS | Late Pleistocene in South America, 2:250
 GLACIATIONS | Late Pleistocene Glacial Events in Beringia, 2:191
 GLACIATIONS | Neoglaciation in Europe, 2:257
 GLACIATIONS | Neoglaciation in the American Cordilleras, 2:269

Sea Level Studies

SEA LEVEL STUDIES | Overview, 4:369
 SEA LEVEL STUDIES | Geomorphological Indicators, 4:377
 SEA LEVEL STUDIES | Coral Records of Relative Sea-Level Changes, 4:409
 SEA LEVEL STUDIES | Use of Cave Data in Sea-Level Reconstructions, 4:460
 SEA LEVEL STUDIES | Eustatic Sea-Level Changes – Glacial–Interglacial Cycles, 4:429
 SEA LEVEL STUDIES | Eustatic Sea-Level Changes Since the Last Glacial Maximum, 4:439

SEA LEVEL STUDIES | Isostasy: Glaciation-Induced Sea-Level Change, 4:452
SEA LEVEL STUDIES | Sedimentary Indicators of Relative Sea-Level Changes – High Energy, 4:385
SEA LEVEL STUDIES | Sedimentary Indicators of Relative Sea-Level Changes – Low Energy, 4:396
SEA LEVEL STUDIES | Microfossil-Based Reconstructions of Holocene Relative Sea-Level Change, 4:419
SEA-LEVELS, LATE QUATERNARY | Late Quaternary Relative Sea-Level Changes in High Latitudes, 4:467
SEA-LEVELS, LATE QUATERNARY | Late Quaternary Sea-Level Changes in Greenland, 4:481
SEA-LEVELS, LATE QUATERNARY | Late Quaternary Relative Sea-Level Changes at Mid-Latitudes, 4:489
SEA-LEVELS, LATE QUATERNARY | Late Quaternary Relative Sea-Level Changes in the Tropics, 4:495
SEA-LEVELS, LATE QUATERNARY | Tectonics and Relative Sea-Level Change, 4:503

Paleosols and Wind-Blown Sediments

Paleosol Research

PALEOSOLS AND WIND-BLOWN SEDIMENTS | Overview, 3:357
PALEOSOLS AND WIND-BLOWN SEDIMENTS | Nature of Paleosols, 3:367
LOESS DEPOSITS: ORIGINS AND PROPERTIES, 2:573
PALEOSOLS AND WIND-BLOWN SEDIMENTS | Weathering Profiles, 3:392
PALEOSOLS AND WIND-BLOWN SEDIMENTS | Soil Micromorphology, 3:381
PALEOSOLS AND WIND-BLOWN SEDIMENTS | Soil Morphology in Quaternary Studies, 3:402
PALEOSOLS AND WIND-BLOWN SEDIMENTS | Mineral Magnetic Analysis, 3:375
PALEOSOLS AND WIND-BLOWN SEDIMENTS | Biogeochemical Role of Dust in Quaternary Climate Cycles, 3:412

Dune Studies

DUNE FIELDS | High Latitudes, 1:597
DUNE FIELDS | Mid-Latitudes, 1:606
DUNE FIELDS | Low Latitudes, 1:623

Loess Studies

LOESS RECORDS | Central Asia, 2:585
LOESS RECORDS | China, 2:595
LOESS RECORDS | Europe, 2:606
LOESS RECORDS | North America, 2:620
LOESS RECORDS | South America, 2:629
EOLIAN RECORDS, DEEP-SEA SEDIMENTS, 1:637

Lake Level Studies

LAKE LEVEL STUDIES | Overview, 2:483
LAKE LEVEL STUDIES | Modeling, 2:558
LAKE LEVEL STUDIES | Africa during the Late Quaternary, 2:499
LAKE LEVEL STUDIES | Asia, 2:506
LAKE LEVEL STUDIES | Australia, 2:524
LAKE LEVEL STUDIES | Latin America, 2:531
LAKE LEVEL STUDIES | North America, 2:537
LAKE LEVEL STUDIES | West-Central-Europe, 2:549

Paleobotany

Overview

PALEOBOTANY | Overview of Terrestrial Pollen Data, 2:699

Pollen Studies

POLLEN ANALYSIS, PRINCIPLES, 3:794
POLLEN METHODS AND STUDIES | Use of Pollen as Climate Proxies, 3:805
POLLEN METHODS AND STUDIES | Surface Samples and Trapping, 3:839
POLLEN METHODS AND STUDIES | The Biome Approach to Reconstructing Past Vegetation, 3:861
POLLEN METHODS AND STUDIES | POLLSCAPE Model, 3:871
POLLEN METHODS AND STUDIES | Reconstructing Past Biodiversity Development, 3:816
POLLEN METHODS AND STUDIES | Numerical Analysis Methods, 3:821
POLLEN METHODS AND STUDIES | Changing Plant Distributions and Abundances, 3:854
POLLEN METHODS AND STUDIES | Stand-Scale Palynology, 3:846
PALEOBOTANY | Paleophytogeography, 2:730
POLLEN METHODS AND STUDIES | Databases and Their Application, 3:831

Charred Particle Analysis

PALEOBOTANY | Charred Particle Analyses, 2:716

Quaternary Pollen Records

POLLEN RECORDS, LAST INTERGLACIAL OF EUROPE, 4:1
POLLEN RECORDS, LATE PLEISTOCENE | Africa, 4:9
POLLEN RECORDS, LATE PLEISTOCENE | Australasia, 4:18
POLLEN RECORDS, LATE PLEISTOCENE | Northern Asia, 4:27

POLLEN RECORDS, LATE PLEISTOCENE | Northern North America, 4:39
 POLLEN RECORDS, LATE PLEISTOCENE | Western North America, 4:72
 POLLEN RECORDS, LATE PLEISTOCENE | South America, 4:52
 POLLEN RECORDS, LATE PLEISTOCENE | Middle and Late Pleistocene in Southern Europe, 4:63
 POLLEN RECORDS, POSTGLACIAL | Africa, 4:85
 POLLEN RECORDS, POSTGLACIAL | Australia and New Zealand, 4:104
 POLLEN RECORDS, POSTGLACIAL | Northeastern North America, 4:115
 POLLEN RECORDS, POSTGLACIAL | Northwestern North America, 4:124
 POLLEN RECORDS, POSTGLACIAL | Southeastern North America, 4:133
 POLLEN RECORDS, POSTGLACIAL | Southwestern North America, 4:142
 POLLEN RECORDS, POSTGLACIAL | South America, 4:156
 POLLEN RECORDS, POSTGLACIAL | Northern Asia, 4:164
 POLLEN RECORDS, POSTGLACIAL | Northern Europe, 4:173
 POLLEN RECORDS, POSTGLACIAL | Southern Europe, 4:179
 POLLEN METHODS AND STUDIES | Archaeological Applications, 3:880

Phytoliths

PHYTOLITHS, 3:582

Ancient Plant DNA

PALEOBOTANY | Ancient Plant DNA, 2:705

Plant Macrofossil Studies

PLANT MACROFOSSIL INTRODUCTION, 3:593
 PLANT MACROFOSSIL METHODS AND STUDIES | Surface Samples, Taphonomy, Representation, 3:684
 PLANT MACROFOSSIL METHODS AND STUDIES | Treeline Studies, 3:690
 PLANT MACROFOSSIL METHODS AND STUDIES | Megafossils, 3:621
 PLANT MACROFOSSIL METHODS AND STUDIES | Paleolimnological Applications, 3:657
 PLANT MACROFOSSIL METHODS AND STUDIES | CO₂ Reconstruction from Fossil Leaves, 3:613
 PLANT MACROFOSSIL METHODS AND STUDIES | Rodent Middens, 3:674

PLANT MACROFOSSIL METHODS AND STUDIES | Mire and Peat Macros, 3:637
 PLANT MACROFOSSIL METHODS AND STUDIES | Use in Environmental Archaeology, 3:699
 PLANT MACROFOSSIL METHODS AND STUDIES | Dendroarchaeology, 3:630
 PLANT MACROFOSSIL METHODS AND STUDIES | Validation of Pollen Studies, 3:725

Plant Macrofossil Records

PLANT MACROFOSSIL RECORDS | Arctic Eurasia, 3:733
 PLANT MACROFOSSIL RECORDS | Greenland, 3:760
 PLANT MACROFOSSIL RECORDS | Arctic North America, 3:746
 PLANT MACROFOSSIL RECORDS | Holocene North America, 3:768
 PLANT MACROFOSSIL RECORDS | Late Glacial Multidisciplinary Studies, 3:785

Diatom Studies

DIATOM INTRODUCTION, 1:471
 DIATOM METHODS | Data Interpretation, 1:489
 DIATOM RECORDS | North Atlantic and Arctic, 1:562
 DIATOM RECORDS | Pacific, 1:571
 DIATOM RECORDS | Antarctic Waters, 1:527
 DIATOM RECORDS | Diatom Fossil Records from Marine Laminated Sediments, 1:554
 DIATOM RECORDS | Freshwater Laminated Sequences, 1:540
 DIATOM RECORDS | Large Lakes, 1:546
 DIATOM METHODS | Salinity and Climate Reconstructions from Continental Lakes, 1:507
 DIATOM METHODS | $\delta^{18}\text{O}$ Records, 1:481
 PALEOBOTANY | Silicon Isotopes in Diatoms, 2:734
 DIATOM METHODS | Use in Archaeology, 1:516
 DIATOM METHODS | Diatomites: Their Formation, Distribution, and Uses, 1:501
 Structures and Applications of Biomarkers from Arctic Sea Ice Diatoms, 1:588

Paleolimnology

PALEOLIMNOLOGY | Overview of Paleolimnology, 3:259
 PALEOLIMNOLOGY | Lake Chemistry, 3:292
 PALEOLIMNOLOGY | Physical Properties of Lake Sediments, 3:300
 PALEOLIMNOLOGY | Cladocera, 3:271
 PALEOLIMNOLOGY | Freshwater Mollusks, 3:281

PALEOLIMNOLOGY | Contributions of
Paleolimnological Research to
Biogeography, 3:313
PALEOLIMNOLOGY | Pigment Studies, 3:326
PALEOLIMNOLOGY | Visible and Infrared
Spectroscopical Applications, 3:349
PALEOLIMNOLOGY | Multiproxy Approaches, 3:339

Vertebrate Studies

VERTEBRATE OVERVIEW, 4:590
VERTEBRATE RECORDS | Early Pleistocene, 4:599
VERTEBRATE RECORDS | Early and Middle
Pleistocene of Northern Eurasia, 4:605
VERTEBRATE RECORDS | Mid-Pleistocene of
Southern Asia, 4:651
VERTEBRATE RECORDS | Mid-Pleistocene of
Africa, 4:615
VERTEBRATE RECORDS | Mid-Pleistocene of
Australia, 4:621
VERTEBRATE RECORDS | Mid-Pleistocene of
Europe, 4:639
VERTEBRATE RECORDS | Mid-Pleistocene of
North America, 4:646
VERTEBRATE RECORDS | Late Pleistocene of
Africa, 4:664
VERTEBRATE RECORDS | Late Pleistocene of North
America, 4:673
VERTEBRATE RECORDS | Late Pleistocene of South
America, 4:680
VERTEBRATE RECORDS | Late Pleistocene of
Southeast Asia, 4:693
VERTEBRATE RECORDS | Late Pleistocene
Megafaunal Extinctions, 4:700
VERTEBRATE RECORDS | Late Pleistocene
Mummified Mammals, 4:713
VERTEBRATE STUDIES | Ancient DNA, 4:719
VERTEBRATE STUDIES | Speciation and
Evolutionary Trends in Quaternary
Vertebrates, 4:723
VERTEBRATE STUDIES | Dwarfing and Gigantism in
Quaternary Vertebrates, 4:733
VERTEBRATE STUDIES | Interactions with
Hominids, 4:748

Insect Fossil Studies

Beetle Studies

BEETLE RECORDS | Overview, 1:161
BEETLE RECORDS | Late Tertiary and Early
Quaternary Records, 1:173
BEETLE RECORDS | Middle Pleistocene of Western
Europe, 1:184

BEETLE RECORDS | Late Pleistocene of
Europe, 1:200
BEETLE RECORDS | Late Pleistocene of North
America, 1:221
BEETLE RECORDS | Late Pleistocene of South
America, 1:235
BEETLE RECORDS | Late Pleistocene of
Australia, 1:191
BEETLE RECORDS | Late Pleistocene of Northern
Asia, 1:255
BEETLE RECORDS | Late Pleistocene of Japan, 1:207
BEETLE RECORDS | Late Pleistocene of New
Zealand, 1:244
BEETLE RECORDS | Postglacial Europe, 1:274
BEETLE RECORDS | Postglacial North America, 1:282

Chironomid Studies

CHIRONOMID RECORDS | Chironomid
Overview, 1:355
CHIRONOMID RECORDS | Late Pleistocene of
Europe, 1:373
CHIRONOMID RECORDS | Africa, 1:361
CHIRONOMID RECORDS | Late Pleistocene of
Europe, 1:373
CHIRONOMID RECORDS | Postglacial
Europe, 1:386
CHIRONOMID RECORDS | Postglacial Southern
Hemisphere, 1:398

Oribatid Mite Studies

ORIBATID MITES, 2:680

Paleoceanography

Paleoceanographic Studies

PALEOCEANOGRAPHY | Paleoceanography An
Overview, 2:745
PALEOCEANOGRAPHY, BIOLOGICAL PROXIES |
Corals, Sclerosponges and Mollusks, 2:795
PALEOCEANOGRAPHY, BIOLOGICAL PROXIES |
Benthic Foraminifera, 2:765
PALEOCEANOGRAPHY, BIOLOGICAL PROXIES |
Planktic Foraminifera, 2:825
PALEOCEANOGRAPHY, BIOLOGICAL PROXIES |
Radiolarians and Silicoflagellates, 2:830
PALEOCEANOGRAPHY, BIOLOGICAL PROXIES |
Coccolithophores, 2:783
PALEOCEANOGRAPHY, BIOLOGICAL PROXIES |
Alkenone Paleothermometry Based on the
Haptophyte Algae, 2:755
PALEOCEANOGRAPHY, BIOLOGICAL PROXIES |
Marine Diatoms, 2:816

PALEOCEANOGRAPHY, BIOLOGICAL PROXIES |
Dinoflagellates, 2:800

PALEOCEANOGRAPHY, PHYSICAL AND
CHEMICAL PROXIES | Terrigenous
Sediments, 2:941

PALEOCEANOGRAPHY, PHYSICAL AND
CHEMICAL PROXIES | Oxygen Isotopic
Composition of Seawater, 2:915

PALEOCEANOGRAPHY, PHYSICAL AND
CHEMICAL PROXIES | Oxygen-Isotope
Stratigraphy of the Oceans, 2:907

PALEOCEANOGRAPHY, PHYSICAL AND
CHEMICAL PROXIES | Dissolution of Deep-Sea
Carbonates, 2:859

PALEOCEANOGRAPHY, PHYSICAL AND
CHEMICAL PROXIES | Mg/Ca and Sr/Ca
Paleothermometry from Calcareous Marine
Fossils, 2:871

PALEOCEANOGRAPHY, BIOLOGICAL PROXIES |
Temperature Proxies Census Counts, 2:841

PALEOCEANOGRAPHY, PHYSICAL AND
CHEMICAL PROXIES | Salinity Proxies
 $\delta^{18}\text{O}$, 2:932

PALEOCEANOGRAPHY, PHYSICAL AND
CHEMICAL PROXIES | Nutrient Proxies, 2:899

PALEOCEANOGRAPHY, PHYSICAL AND
CHEMICAL PROXIES | Carbon Cycle
Proxies, 2:849

PALEOCEANOGRAPHY, PHYSICAL AND
CHEMICAL PROXIES | Radioisotope
Proxies, 2:923

PALEOCEANOGRAPHY, BIOLOGICAL PROXIES |
Biomarkers Indicators of Past Climate, 2:775

PALEOCEANOGRAPHY, PHYSICAL AND
CHEMICAL PROXIES | Magnetic Proxies and
Susceptibility, 2:884

Paleoceanographic Records

PALEOCEANOGRAPHY, RECORDS | Early
Pleistocene, 3:1

PALEOCEANOGRAPHY, RECORDS | Late
Pleistocene North Atlantic, 3:9

PALEOCEANOGRAPHY, RECORDS | Late
Pleistocene South Atlantic, 3:18

PALEOCEANOGRAPHY, RECORDS | Late
Pleistocene North Pacific, 3:33

PALEOCEANOGRAPHY, RECORDS | Postglacial
North Atlantic, 3:55

PALEOCEANOGRAPHY, RECORDS | Postglacial
North Pacific, 3:62

PALEOCEANOGRAPHY, RECORDS | Postglacial
South Pacific, 3:73

PALEOCEANOGRAPHY, RECORDS | Postglacial
Indian Ocean, 3:46

Ice Core Studies

Ice Core Studies

ICE CORE METHODS | Overview, 2:278

ICE CORES | History of Research, Greenland and
Antarctica, 2:435

ICE CORE METHODS | Chronologies, 2:303

ICE CORE METHODS | Stable Isotopes, 2:347

ICE CORE METHODS | Conductivity Studies, 2:319

ICE CORE METHODS | Glaciochemistry, 2:326

ICE CORE METHODS | Biological Material, 2:288

ICE CORE METHODS | Microparticle and Trace
Element Studies, 2:342

ICE CORE METHODS | CO₂ Studies, 2:311

ICE CORE METHODS | Methane Studies, 2:334

ICE CORE METHODS | ¹⁰Be and Cosmogenic
Radionuclides in Ice Cores, 2:353

ICE CORE METHODS | Studies of Firn Air, 2:361

Ice Core Records

ICE CORES | History of Carbon Monoxide and
Ultra-Trace Gases from Ice Cores, 2:463

ICE CORES | History of Nitrous Oxide from Ice
Cores, 2:471

ICE CORES | Dynamics of the Greenland Ice
Sheet, 2:439

ICE CORES | Dynamics of the East Antarctic Ice
Sheet, 2:456

ICE CORES | Dynamics of the West Antarctic Ice
Sheet, 2:448

ICE CORE METHODS | Borehole Temperature
Records, 2:298

ICE CORE RECORDS | Greenland Stable
Isotopes, 2:403

ICE CORE RECORDS | Antarctic Stable Isotopes, 2:395

ICE CORE RECORDS | Ice Margin Sites, 2:416

ICE CORE RECORDS | Correlations Between
Greenland and Antarctica, 2:410

ICE CORE RECORDS | Africa, 2:373

ICE CORE RECORDS | Chinese, Tibetan
Mountains, 2:379

ICE CORE RECORDS | South America, 2:387

ICE CORE RECORDS | Thermal Diffusion
Paleotemperature Records, 2:431

Stable Isotope Research – Carbonates

CARBONATE STABLE ISOTOPES | Overview, 1:291

CARBONATE STABLE ISOTOPES |
Speleothems, 1:294

CARBONATE STABLE ISOTOPES | Terrestrial Teeth
and Bones, 1:304

CARBONATE STABLE ISOTOPES | Terrestrial
Organic Materials, 1:315
CARBONATE STABLE ISOTOPES | Non-Lacustrine
Terrestrial Studies, 1:322
CARBONATE STABLE ISOTOPES | Lake
Sediments, 1:333
CARBONATE STABLE ISOTOPES | Nonmarine
Biogenic Carbonates, 1:341

Humans in the Quaternary

ARCHAEOLOGICAL RECORDS | Overview, 1:49
ARCHAEOLOGICAL RECORDS | 2.7 Myr–300 000
Years Ago in Africa, 1:59
ARCHAEOLOGICAL RECORDS | 2.7 Myr–300 000
Years Ago in Asia, 1:67
ARCHAEOLOGICAL RECORDS | 1.9 Myr–300 000
Years Ago in Europe, 1:83
ARCHAEOLOGICAL RECORDS | Global Expansion
300 000–8000 Years Ago, Africa, 1:91
ARCHAEOLOGICAL RECORDS | Global Expansion
300 000–8000 Years Ago, Asia, 1:98
ARCHAEOLOGICAL RECORDS | Global
Expansion 300 000–8000 Years Ago,
Australia, 1:108

ARCHAEOLOGICAL RECORDS | Global Expansion
300 000–8000 Years Ago, Americas, 1:119
ARCHAEOLOGICAL RECORDS | Neanderthal
Demise, 1:146
ARCHAEOLOGICAL RECORDS | Human Evolution
in the Quaternary, 1:135
ARCHAEOLOGICAL RECORDS | Postglacial
Adaptations, 1:154

Use of Quaternary proxies in forensic science

USE OF QUATERNARY PROXIES IN FORENSIC
SCIENCE | Use of Geology in Forensic Science:
Search to Locate Burials, 4:522
USE OF QUATERNARY PROXIES IN FORENSIC
SCIENCE | Analysis Techniques in Forensic
Palynology, 4:556
USE OF QUATERNARY PROXIES IN FORENSIC
SCIENCE | Soils and Sediment, 4:535
USE OF QUATERNARY PROXIES IN FORENSIC
SCIENCE | The Use of Macroscopic Plant Remains
in Forensic Science, 4:542
USE OF QUATERNARY PROXIES IN FORENSIC
SCIENCE | Insects, 4:548
Diatoms, 1:522

CONTENTS

<i>Dedications</i>	<i>v</i>
<i>Foreword</i>	<i>ix</i>
<i>Introduction</i>	<i>xi</i>
<i>Editorial Advisory Board</i>	<i>xiii</i>
<i>Contributors</i>	<i>xv</i>
<i>How to Use the Encyclopedia</i>	<i>xxvii</i>
<i>Contents List by Subject</i>	<i>xxix</i>

VOLUME 1

INTRODUCTORY ARTICLES	1
History of Quaternary Science <i>S A Elias</i>	1
History of Dating Methods <i>A G Wintle</i>	9
Societal Relevance of Quaternary Research <i>S A Elias</i>	17
Understanding Quaternary Climatic Change <i>J J Lowe, M J C Walker, and S C Porter</i>	26
A	
ALKENONE STUDIES <i>see</i> PALEOCEANOGRAPHY, BIOLOGICAL PROXIES: Alkenone Paleothermometry Based on the Haptophyte Algae	
ALLOSTRATIGRAPHY <i>see</i> QUATERNARY STRATIGRAPHY: Morphostratigraphy/Allostratigraphy	
Amino Acid Dating <i>G H Miller, D S Kaufman, and S J Clarke</i>	37
ANATOMICALLY MODERN HUMANS <i>see</i> ARCHAEOLOGICAL RECORDS: Global Expansion 300 000–8000 Years Ago, Africa; Global Expansion 300 000–8000 Years Ago, Asia; Global Expansion 300 000–8000 Years Ago, Australia; Global Expansion 300 000–8000 Years Ago, Americas	
ARCHAEOLOGICAL RECORDS	49
Overview <i>C Gamble</i>	49

2.7 Myr–300 000 Years Ago in Africa <i>J W K Harris, D R Braun, and M Pante</i>	59
2.7 Myr–300 000 Years Ago in Asia <i>R Dennell</i>	67
1.9 Myr–300 000 Years Ago in Europe <i>J McNabb</i>	83
Global Expansion 300 000–8000 Years Ago, Africa <i>A E Close and T Minichillo</i>	91
Global Expansion 300 000–8000 Years Ago, Asia <i>M D Petraglia and R Dennell</i>	98
Global Expansion 300 000–8000 Years Ago, Australia <i>R Cosgrove</i>	108
Global Expansion 300 000–8000 Years Ago, Americas <i>T Goebel</i>	119
Human Evolution in the Quaternary <i>L Cashmore</i>	135
Neanderthal Demise <i>W Davies</i>	146
Postglacial Adaptations <i>G Bailey</i>	154

B

BEETLE RECORDS	161
Overview <i>S A Elias</i>	161
Late Tertiary and Early Quaternary Records <i>S A Elias and S Kuzmina</i>	173
Middle Pleistocene of Western Europe <i>G R Coope</i>	184
Late Pleistocene of Australia <i>N Porch</i>	191
Late Pleistocene of Europe <i>G Lemdahl and G R Coope</i>	200
Late Pleistocene of Japan <i>M Hayashi</i>	207
Late Pleistocene of North America <i>S A Elias</i>	221
Late Pleistocene of South America <i>A C Ashworth</i>	235
Late Pleistocene of New Zealand <i>M Marra</i>	244
Late Pleistocene of Northern Asia <i>A Sher and S Kuzmina</i>	255

Postglacial Europe <i>P Ponel</i>	274
Postglacial North America <i>S A Elias</i>	282

BERINGIA *see* ARCHAEOLOGICAL RECORDS: Global Expansion 300 000–8000 Years Ago, Americas;
 BEETLE RECORDS: Late Pleistocene of North America; DUNE FIELDS: High Latitudes; GLACIATIONS:
 Late Pleistocene Glacial Events in Beringia; PALEOCEANOGRAPHY, RECORDS: Postglacial North
 Pacific; PLANT MACROFOSSIL RECORDS: Arctic North America; POLLEN RECORDS, LATE
 PLEISTOCENE: Northern North America; VERTEBRATE RECORDS: Late Pleistocene of North America

BIOGENIC CARBONATE STUDIES *see* CARBONATE STABLE ISOTOPES: Nonmarine Biogenic Carbonates

BOND CYCLES *see* PALEOCLIMATE RECONSTRUCTION: Sub-Milankovitch (DO/Heinrich) Events

C

CARBONATE STABLE ISOTOPES 291

Overview <i>H Schwarcz</i>	291
Speleothems <i>H Schwarcz</i>	294
Terrestrial Teeth and Bones <i>H Bocherens and D G Drucker</i>	304
Terrestrial Organic Materials <i>D McCarroll and N Loader</i>	315
Non-Lacustrine Terrestrial Studies <i>J Quade and T Cerling</i>	322
Lake Sediments <i>S M Bernasconi and J A McKenzie</i>	333
Nonmarine Biogenic Carbonates <i>S J Carpenter</i>	341

CARBON DIOXIDE, ATMOSPHERIC CONCENTRATIONS *see* CARBONATE STABLE ISOTOPES:
 Overview; ICE CORE METHODS: CO₂ Studies; PALEOCEANOGRAPHY, PHYSICAL AND CHEMICAL
 PROXIES: Carbon Cycle Proxies ($\delta^{11}\text{B}$, $\delta^{13}\text{C}_{\text{calcite}}$, $\delta^{13}\text{C}_{\text{organic}}$, Shell Weights, B/Ca, U/Ca, Zn/Ca, Ba/Ca);
 PLANT MACROFOSSIL METHODS AND STUDIES: CO₂ Reconstruction from Fossil Leaves

CAVE ART *see* VERTEBRATE STUDIES: Interactions with Hominids

CHARCOAL STUDIES *see* PALEOBOTANY: Charred Particle Analyses

CHIRONOMID RECORDS 355

Chironomid Overview <i>I R Walker</i>	355
Africa <i>H Eggermont and D Verschuren</i>	361
Late Pleistocene of Europe <i>S J Brooks</i>	373
Postglacial Europe <i>G Velle and O Heiri</i>	386

Postglacial Southern Hemisphere <i>J Massferro and M Vandergoes</i>	398
--	-----

CLADOCERA STUDIES *see* PALEOLIMNOLOGY: Cladocera

CLIMATE CHANGE *see* PALEOCLIMATE: Introduction; Timescales of Climate Change; PALEOCLIMATE MODELING: Data–Model Comparisons; Quaternary Environments; The Last Interglacial; Last Glacial Maximum GCMs; Paleo-ENSO; PALEOCLIMATE RECONSTRUCTION: Approaches; Pliocene Environments; Paleodroughts and Society; Sub-Milankovitch (DO/Heinrich) Events; Paleotempestology; Younger Dryas Oscillation, Global Evidence; The Last Millennium; Historical Climatology; Paleoclimate Relevance to Global Warming

CLIMATE MODELING, QUATERNARY *see* PALEOCLIMATE MODELING: Quaternary Environments

COCCOLITHOPHORE STUDIES *see* PALEOCEANOGRAPHY, BIOLOGICAL PROXIES: Alkenone Paleothermometry Based on the Haptophyte Algae; Coccolithophores

COLEOPTERA FOSSIL RECORDS *see* BEETLE RECORDS: Overview; Late Tertiary and Early Quaternary Records; Middle Pleistocene of Western Europe; Late Pleistocene of Australia; Late Pleistocene of Europe; Late Pleistocene of Japan; Late Pleistocene of North America; Late Pleistocene of South America; Late Pleistocene of New Zealand; Late Pleistocene of Northern Asia; Postglacial Europe; Postglacial North America

CORAL STUDIES *see* PALEOCEANOGRAPHY, BIOLOGICAL PROXIES: Corals, Sclerosponges and Mollusks; SEA LEVEL STUDIES: Coral Records of Relative Sea-Level Changes

COSMOGENIC NUCLIDE DATING 407

Overview <i>J C Gosse</i>	407
------------------------------	-----

Methods <i>J M Schaefer and N Lifton</i>	410
---	-----

Cosmic Ray Interactions in Minerals <i>D Lal</i>	418
---	-----

Exposure Geochronology <i>S Ivy-Ochs and F Kober</i>	432
---	-----

Landscape Evolution <i>D E Granger</i>	440
---	-----

CUT-MARKED BONE *see* VERTEBRATE STUDIES: Interactions with Hominids

D

DANSGAARD-OESCHGER EVENTS *see* PALEOCLIMATE RECONSTRUCTION: Paleodroughts and Society

Dating Techniques <i>A J T Jull</i>	447
--	-----

DENDROARCHAEOLOGY *see* PLANT MACROFOSSIL METHODS AND STUDIES: Dendroarchaeology

Dendrochronology <i>B L Coulthard and D J Smith</i>	453
--	-----

Dendroclimatology <i>B H Luckman</i>	459
---	-----

Diatom Introduction	471
<i>V J Jones</i>	
DIATOM METHODS	481
$\delta^{18}\text{O}$ Records	481
<i>M J Leng, P A Barker, G E A Swann, and A M Snelling</i>	
Data Interpretation	489
<i>A Korhola</i>	
Diatomites: Their Formation, Distribution, and Uses	501
<i>R J Flower</i>	
Salinity and Climate Reconstructions from Continental Lakes	507
<i>S C Fritz</i>	
Use in Archaeology	516
<i>N G Cameron</i>	
Diatoms	522
<i>N G Cameron</i>	
DIATOM RECORDS	527
Antarctic Waters	527
<i>C E Stickley, J Pike, and V J Jones</i>	
Freshwater Laminated Sequences	540
<i>H Simola</i>	
Large Lakes	546
<i>A W Mackay</i>	
Diatom Fossil Records from Marine Laminated Sediments	554
<i>J Pike and C E Stickley</i>	
North Atlantic and Arctic	562
<i>N Koç, A Miettinen, and C E Stickley</i>	
Pacific	571
<i>I Koizumi</i>	
Structures and Applications of Biomarkers from Arctic Sea Ice Diatoms	588
<i>S T Belt, G Massé, and M Poulin</i>	
DIATOMS, MARINE <i>see</i> PALEOCEANOGRAPHY, BIOLOGICAL PROXIES: Marine Diatoms	
DINOFLAGELLATES <i>see</i> PALEOCEANOGRAPHY, BIOLOGICAL PROXIES: Dinoflagellates	
DNA, FOSSIL ANIMAL <i>see</i> VERTEBRATE STUDIES: Ancient DNA	
DNA, FOSSIL PLANTS <i>see</i> PALEOBOTANY: Ancient Plant DNA	
DROUGHT HISTORY RECONSTRUCTION <i>see</i> PALEOCLIMATE RECONSTRUCTION: Paleodroughts and Society	
DUNE FIELDS	597
High Latitudes	597
<i>S A Wolfe</i>	
Mid-Latitudes	606
<i>J Sun and D R Muhs</i>	

Low Latitudes <i>N Lancaster</i>	623
-------------------------------------	-----

E

EL NIÑO SOUTHERN OSCILLATION <i>see</i> PALEOCEANOGRAPHY, RECORDS: Postglacial South Pacific; PALEOCLIMATE MODELING: Paleo-ENSO	
ELECTRON SPIN RESONANCE DATING <i>see</i> LUMINESCENCE DATING: Electron Spin Resonance Dating	
EOLIAN SEDIMENTS <i>see</i> PALEOSOLS AND WIND-BLOWN SEDIMENTS: Overview; Nature of Paleosols; Mineral Magnetic Analysis; Soil Micromorphology; Weathering Profiles; Soil Morphology in Quaternary Studies; Biogeochemical Role of Dust in Quaternary Climate Cycles	
Eolian Records, Deep-Sea Sediments <i>D K Rea</i>	637
EQUILIBRIUM LINE ALTITUDE (ELA) RECONSTRUCTION <i>see</i> GLACIAL LANDFORMS, ICE SHEETS: Paleo-ELAs	
ERRATICS, GLACIAL <i>see</i> GLACIAL LANDFORMS, SEDIMENTS: Glacial Erratics and Till Dispersal Indicators	
EXTINCTIONS, QUATERNARY VERTEBRATES <i>see</i> VERTEBRATE RECORDS: Late Pleistocene Megafaunal Extinctions	

F

FELSENMEER (BLOCKFIELDS) <i>see</i> PERMAFROST AND PERIGLACIAL FEATURES: Block/Rock Streams	
Fission-Track Dating <i>J A Westgate, N D Naeser, and B V Alloway</i>	643

FLUVIAL ENVIRONMENTS 663

Sediments <i>A Aslan</i>	663
Responses to Rapid Environmental Change <i>T E Törnqvist</i>	676
Terrace Sequences <i>D J Merritts</i>	684
Deltaic Environments <i>L Giosan and S L Goodbred</i>	693

FORAMINIFERA STUDIES *see* PALEOCEANOGRAPHY: Paleocceanography An Overview; PALEOCEANOGRAPHY, BIOLOGICAL PROXIES: Temperature Proxies, Census Counts; Benthic Foraminifera; Planktic Foraminifera; PALEOCEANOGRAPHY, PHYSICAL AND CHEMICAL PROXIES: Oxygen-Isotope Stratigraphy of the Oceans

FOSSIL MITES *see* Oribatid Mites

G

Geomagnetic Excursions and Secular Variations <i>A P Roberts and G M Turner</i>	705
--	-----

GLACIAL CLIMATES	721
Biosphere Feedbacks <i>V Brovkin</i>	721
Effects of Atmospheric Dust <i>I Tegen</i>	729
Thermohaline Circulation <i>S Rahmstorf</i>	737
Volcanic and Solar Forcing <i>D T Shindell</i>	748
GLACIAL LANDFORMS	755
Introduction <i>D J A Evans and D I Benn</i>	755
Moraine Forms and Genesis <i>D J A Evans</i>	769
Quaternary Vulcanism: Subglacial Landforms <i>J L Smellie</i>	780
Evidence of Glacier Recession <i>P M Colgan</i>	803
Glacial Landsystems <i>D J A Evans</i>	813
Glaciofluvial Landforms of Erosion <i>A E Kehew, M L Lord, and A L Kozlowski</i>	825
Glacitectonic Structures and Landforms <i>D J A Evans</i>	839
GLACIAL LANDFORMS, EROSIONAL FEATURES	847
Major Scale Forms <i>I S Evans</i>	847
Micro- to Macroscale Forms <i>B R Rea</i>	865
GLACIAL LANDFORMS, ICE SHEETS	877
Growth and Decay <i>M J Siegert</i>	877
Evidence of Glacier and Ice Sheet Extent <i>D M Mickelson and C Winguth</i>	884
Evidence of Glacier Flow Directions <i>C R Stokes</i>	895
Paleo-ELAs <i>A Nesje</i>	909
Trimlines and Paleonunataks <i>C K Ballantyne</i>	918

VOLUME 2**GLACIAL LANDFORMS, SEDIMENTS 1**

Clast Form Analysis 1
D I Benn

Glaciofluvial Landforms of Deposition 6
J L Carrivick and A J Russell

Glaciogenic Lithofacies 18
N Eyles and M Lazorek

Glaciomarine Sediments and Ice-Rafted Debris 30
C Ó Cofaigh

Glaciolacustrine 43
D J A Evans

Micromorphology of Glacial Sediments 52
J F Hiemstra

Tills 62
D J A Evans

Till Fabric Analysis 76
D I Benn

Glacial Erratics and Till Dispersal Indicators 81
D J A Evans

Glacial Sequence Stratigraphy 85
D J A Evans

GLACIAL LANDFORMS, TREE RINGS 91

Dendrogeomorphology 91
H Gärtner and I Heinrich

Dendroglaciology 104
B L Coulthard and D J Smith

Glacial–Interglacial Scale Fluvial Responses 112
M D Blum

GLACIATION, CAUSES 127

Tectonic Uplift and Continental Configurations 127
L A Owen

Astronomical Theory of Paleoclimates 136
A Berger, M-F Loutre, and Q Z Yin

GLACIATIONS 143

Overview 143
J Ehlers and P L Gibbard

Transition from Late Neogene to Early Quaternary Environments 151
M Sarnthein

Early Quaternary (Pleistocene) and Precursors 167
J Ehlers, V Astakhov, P L Gibbard, Ó Ingólfsson, J Mangerud, and J I Svendsen

Middle Pleistocene in Eurasia <i>J Ehlers, V Astakhov, P L Gibbard, J Mangerud, and J I Svendsen</i>	172
Mid-Quaternary in North America <i>C E Jennings, J S Aber, G Balco, R Barendregt, P R Bierman, C W Rovey II, M Roy, L H Thorleifson, and J A Mason</i>	180
Middle Pleistocene Glaciations in the Southern Hemisphere <i>A Coronato and J Rabassa</i>	187
Late Pleistocene Glacial Events in Beringia <i>S A Elias and J Brigham-Grette</i>	191
Late Quaternary of the Southwest Pacific Region <i>D J A Barrell</i>	202
Late Quaternary of Antarctica <i>Ó Ingólfsson</i>	216
Late Pleistocene in Eurasia <i>J Ehlers, V Astakhov, P L Gibbard, J Mangerud, and J I Svendsen</i>	224
Late Quaternary in Highland Asia <i>L A Owen</i>	236
Late Quaternary in North America <i>J T Andrews and A S Dyke</i>	245
Late Pleistocene in South America <i>A Coronato and J Rabassa</i>	250
Neoglaciation in Europe <i>J A Matthews</i>	257
Neoglaciation in the American Cordilleras <i>S C Porter</i>	269

GLOBAL WARMING *see* Paleoclimate Relevance to Global Warming

H

HEINRICH EVENTS *see* PALEOCLIMATE RECONSTRUCTION: Paleodroughts and Society

HISTORICAL CLIMATE RECORDS *see* PALEOCLIMATE RECONSTRUCTION: Historical Climatology; The Last Millennium

HOLOCENE ENVIRONMENTS *see* Dendroclimatology; ARCHAEOLOGICAL RECORDS: Postglacial Adaptations; BEETLE RECORDS: Postglacial Europe; Postglacial North America; CHIRONOMID RECORDS: Postglacial Europe; Postglacial Southern Hemisphere; DIATOM METHODS: Use in Archaeology; GLACIATIONS: Neoglaciation in Europe; Neoglaciation in the American Cordilleras; ICE CORES: Dynamics of the Greenland Ice Sheet; Dynamics of the West Antarctic Ice Sheet; Dynamics of the East Antarctic Ice Sheet; PALEOCEANOGRAPHY, RECORDS: Postglacial Indian Ocean; Postglacial North Atlantic; Postglacial North Pacific; Postglacial South Pacific; PALEOCLIMATE: Timescales of Climate Change; PALEOCLIMATE MODELING: Paleo-ENSO; PALEOCLIMATE RECONSTRUCTION: Historical Climatology; The Last Millennium; PLANT MACROFOSSIL METHODS AND STUDIES: Treeline Studies; PLANT MACROFOSSIL RECORDS: Holocene North America; POLLEN RECORDS, POSTGLACIAL: Africa; Australia and New Zealand; Northeastern North America; Northwestern North America; Southeastern North America; Southwestern North America; South America; Northern Asia; Northern Europe; Southern Europe

HUMAN EVOLUTION *see* ARCHAEOLOGICAL RECORDS: Overview; 2.7 Myr–300 000 Years Ago in Africa; 2.7 Myr–300 000 Years Ago in Asia; 1.9 Myr–300 000 Years Ago in Europe; Human Evolution in the Quaternary

I

ICE CORE METHODS 277

Overview 278
E J Brook

Biological Material 288
J C Priscu, B C Christner, C M Foreman, and G Royston-Bishop

Borehole Temperature Records 298
K M Cuffey

Chronologies 303
J Schwander

CO₂ Studies 311
T Blunier and T M Jenk

Conductivity Studies 319
R Mulvaney

Glaciochemistry 326
K J Kreutz and B G Koffman

Methane Studies 334
J Chappellaz

Microparticle and Trace Element Studies 342
J R McConnell

Stable Isotopes 347
E J Brook

¹⁰Be and Cosmogenic Radionuclides in Ice Cores 353
R Muscheler

Studies of Firn Air 361
C Buizert

ICE CORE RECORDS 373

Africa 373
L G Thompson and M E Davis

Chinese, Tibetan Mountains 379
C P Wake

South America 387
L G Thompson and M E Davis

Antarctic Stable Isotopes 395
E J Brook

Greenland Stable Isotopes 403
B M Vinther and S J Johnsen

Correlations Between Greenland and Antarctica 410
E J Brook

Ice Margin Sites <i>V V Petrenko</i>	416
Thermal Diffusion Paleotemperature Records <i>A M Grachev</i>	431
ICE CORES	435
History of Research, Greenland and Antarctica <i>M Aydin</i>	435
Dynamics of the Greenland Ice Sheet <i>C S Hvidberg, A Svensson, and S L Buchardt</i>	439
Dynamics of the West Antarctic Ice Sheet <i>R Bindshadler</i>	448
Dynamics of the East Antarctic Ice Sheet <i>E D Waddington and C S Lingle</i>	456
History of Carbon Monoxide and Ultra-Trace Gases from Ice Cores <i>M Aydin</i>	463
History of Nitrous Oxide from Ice Cores <i>A Schilt</i>	471
ICE SHEETS, PLEISTOCENE <i>see</i> GLACIAL LANDFORMS, ICE SHEETS: Growth and Decay; Evidence of Glacier and Ice Sheet Extent; Trimlines and Paleonunataks; GLACIATIONS: Early Quaternary (Pleistocene) and Precursors; Middle Pleistocene in Eurasia; Mid-Quaternary in North America; Middle Pleistocene Glaciations in the Southern Hemisphere; Late Pleistocene Glacial Events in Beringia; Late Quaternary of the Southwest Pacific Region; Late Quaternary of Antarctica; Late Pleistocene in Eurasia; Late Quaternary in Highland Asia; Late Quaternary in North America; Late Pleistocene in South America; ICE CORES: Dynamics of the Greenland Ice Sheet; Dynamics of the West Antarctic Ice Sheet; Dynamics of the East Antarctic Ice Sheet	
ICE WEDGES, ICE WEDGE CASTS <i>see</i> PERMAFROST AND PERIGLACIAL FEATURES: Ice Wedges and Ice-Wedge Casts	
K	
K/Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ Dating <i>J R Wijbrans and K F Kuiper</i>	477
L	
LAKE CHEMISTRY RECONSTRUCTION <i>see</i> PALEOLIMNOLOGY: Lake Chemistry	
LAKE LEVEL STUDIES	483
Overview <i>R T Jones and J T Jordan</i>	483
Africa during the Late Quaternary <i>M E Edwards</i>	499
Asia <i>G Yu, B Xue, and Y Li</i>	506
Australia <i>J Magee</i>	524

Latin America <i>S E Metcalfe</i>	531
North America <i>J R Stone and S C Fritz</i>	537
West-Central Europe <i>M Magny</i>	549
Modeling <i>J Vassiljev</i>	558
Lichenometry <i>D P McCarthy</i>	565

LITHOSTRATIGRAPHY *see* QUATERNARY STRATIGRAPHY: Lithostratigraphy

Loess Deposits: Origins and Properties <i>D R Muhs</i>	573
---	-----

LOESS RECORDS 585

Central Asia <i>A E Dodonov</i>	585
China <i>S C Porter</i>	595
Europe <i>D-D Rousseau, E Derbyshire, P Antoine, and C Hatté</i>	606
North America <i>H M Roberts, D R Muhs, and E A Bettis III</i>	620
South America <i>M A Zárate</i>	629

LUMINESCENCE DATING 643

Thermoluminescence <i>O B Lian</i>	643
Optical Dating <i>O B Lian</i>	653
Electron Spin Resonance Dating <i>A J T Jull</i>	667

M

MAGNETIC POLARITY STUDIES *see* Geomagnetic Excursions and Secular Variations

MAMMALIAN EVOLUTION *see* Vertebrate Overview; VERTEBRATE RECORDS: Early Pleistocene; Mid-Pleistocene of Africa; Mid-Pleistocene of Southern Asia; VERTEBRATE STUDIES: Ancient DNA; Speciation and Evolutionary Trends in Quaternary Vertebrates

MARINE ISOTOPE STAGES *see* PALEOCEANOGRAPHY, PHYSICAL AND CHEMICAL PROXIES: Oxygen-Isotope Stratigraphy of the Oceans

MEGAFAUNA, PLEISTOCENE *see* Vertebrate Overview; VERTEBRATE RECORDS: Mid-Pleistocene of Australia; Mid-Pleistocene of North America; Early Pleistocene; Late Pleistocene of Africa;

Late Pleistocene of North America; Late Pleistocene of South America; Late Pleistocene of Southeast Asia; Late Pleistocene Megafaunal Extinctions; Mid-Pleistocene of Africa; Mid-Pleistocene of Europe; Mid-Pleistocene of Southern Asia; VERTEBRATE STUDIES: Interactions with Hominids

MEGAFAUNAL EXTINCTION *see* VERTEBRATE RECORDS: Late Pleistocene Megafaunal Extinctions; VERTEBRATE STUDIES: Interactions with Hominids

METHANE STUDIES, ICE CORES *see* ICE CORE METHODS: Methane Studies

Mg/Ca AND Sr/Ca STUDIES *see* PALEOCEANOGRAPHY, PHYSICAL AND CHEMICAL PROXIES: Mg/Ca and Sr/Ca Paleothermometry from Calcareous Marine Fossils

MICROMORPHOLOGY OF SEDIMENTS *see* GLACIAL LANDFORMS, SEDIMENTS: Micromorphology of Glacial Sediments; PALEOSOLS AND WIND-BLOWN SEDIMENTS: Soil Micromorphology

MIDGES *see* CHIRONOMID RECORDS: Africa; Chironomid Overview; Late Pleistocene of Europe; Postglacial Europe; Postglacial Southern Hemisphere

MILANKOVITCH THEORY *see* GLACIATION, CAUSES: Astronomical Theory of Paleoclimates

MOLLUSKS, MARINE *see* PALEOCEANOGRAPHY, BIOLOGICAL PROXIES: Corals, Sclerosponges and Mollusks

MORAINES, GLACIAL *see* GLACIAL LANDFORMS: Moraine Forms and Genesis; Evidence of Glacier Recession; Glacial Landsystems; Glacitectonic Structures and Landforms; GLACIAL LANDFORMS, SEDIMENTS: Glacial Erratics and Till Dispersal Indicators

MORPHOSTRATIGRAPHY *see* QUATERNARY STRATIGRAPHY: Morphostratigraphy/Allostratigraphy

N

NEANDERTHAL DEMISE *see* ARCHAEOLOGICAL RECORDS: Neanderthal Demise

NEOGLACIATION *see* GLACIATIONS: Neoglaciation in Europe; Neoglaciation in the American Cordilleras

O

OPTICALLY-STIMULATED LUMINESCENCE DATING *see* LUMINESCENCE DATING: Optical Dating

Oribatid Mites

J M Erickson and R B Platt Jr.

680

OXYGEN ISOTOPE STAGES *see* PALEOCEANOGRAPHY, PHYSICAL AND CHEMICAL PROXIES: Oxygen-Isotope Stratigraphy of the Oceans

P

PACKRAT MIDDENS *see* PLANT MACROFOSSIL METHODS AND STUDIES: Rodent Middens

PALEOANTHROPOLOGY *see* ARCHAEOLOGICAL RECORDS: Overview; 2.7 Myr–300 000 Years Ago in Africa; 2.7 Myr–300 000 Years Ago in Asia; 1.9 Myr–300 000 Years Ago in Europe; Global Expansion 300 000–8000 Years Ago, Africa; Global Expansion 300 000–8000 Years Ago, Asia; Global Expansion 300 000–8000 Years Ago, Australia; Global Expansion 300 000–8000 Years Ago, Americas; Human Evolution in the Quaternary; Neanderthal Demise; Postglacial Adaptations

PALEOBOTANY

699

Overview of Terrestrial Pollen Data

699

R H W Bradshaw

Ancient Plant DNA

705

N Wales, R Allaby, E Willerslev, and M T P Gilbert

Charred Particle Analyses <i>K J Brown and M J Power</i>	716
Paleophytogeography <i>A E Bjune</i>	730
Silicon Isotopes in Diatoms <i>J J Tyler</i>	734
PALEOCEANOGRAPHY	745
Paleoceanography An Overview <i>D M Anderson and K E Lee</i>	745
PALEOCEANOGRAPHY, BIOLOGICAL PROXIES	755
Alkenone Paleothermometry Based on the Haptophyte Algae <i>S L Ho, B D A Naafs, and F Lamy</i>	755
Benthic Foraminifera <i>R Saraswat and R Nigam</i>	765
Biomarker Indicators of Past Climate <i>J P Sachs, K Pahnke, R Smittenberg, and Z Zhang</i>	775
Coccolithophores <i>J-A Flores and F J Sierro</i>	783
Corals, Sclerosponges and Mollusks <i>T M Quinn and B R Schöne</i>	795
Dinoflagellates <i>A de Vernal, A Rochon, and T Radi</i>	800
Marine Diatoms <i>F Abrantes and I M Gil</i>	816
Planktic Foraminifera <i>H J Dowsett and M M Robinson</i>	825
Radiolarians and Silicoflagellates <i>D Lazarus</i>	830
Temperature Proxies, Census Counts <i>J D Ortiz</i>	841
PALEOCEANOGRAPHY, PHYSICAL AND CHEMICAL PROXIES	849
Carbon Cycle Proxies ($\delta^{11}\text{B}$, $\delta^{13}\text{C}_{\text{calcite}}$, $\delta^{13}\text{C}_{\text{organic}}$, Shell Weights, B/Ca, U/Ca, Zn/Ca, Ba/Ca) <i>B Hönisch and K A Allen</i>	849
Dissolution of Deep-Sea Carbonates <i>S Barker</i>	859
Mg/Ca and Sr/Ca Paleothermometry from Calcareous Marine Fossils <i>Y Rosenthal and B Linsley</i>	871
Magnetic Proxies and Susceptibility <i>R G Hatfield and J S Stoner</i>	884
Nutrient Proxies <i>T M Marchitto</i>	899

Oxygen-Isotope Stratigraphy of the Oceans <i>F C Bassinot</i>	907
Oxygen Isotope Composition of Seawater <i>E J Rohling</i>	915
Radioisotope Proxies <i>R Francois</i>	923
Salinity Proxies $\delta^{18}\text{O}$ <i>P D Naidu</i>	932
Terrigenous Sediments <i>A M Franzese and S R Hemming</i>	941

VOLUME 3

PALEOCEANOGRAPHY, RECORDS	1
Early Pleistocene <i>T de Garidel-Thoron</i>	1
Late Pleistocene North Atlantic <i>D M Anderson</i>	9
Late Pleistocene South Atlantic <i>S Mulitza, A Paul, and G Wefer</i>	18
Late Pleistocene North Pacific <i>T Kiefer</i>	33
Postglacial Indian Ocean <i>P D Naidu</i>	46
Postglacial North Atlantic <i>M W Kerwin and K A Huguen</i>	55
Postglacial North Pacific <i>J I Martínez</i>	62
Postglacial South Pacific <i>F Lamy and R de Pol-Holz</i>	73
PALEOCLIMATE	87
Introduction <i>C J Mock</i>	87
Timescales of Climate Change <i>P J Bartlein</i>	93
Modern Analog Approaches in Paleoclimatology <i>C J Mock and J J Shinker</i>	102
Paleoclimate History of the Arctic <i>G H Miller, J Brigham-Grette, R B Alley, L Anderson, H A Bauch, M S V Douglas, M E Edwards, S A Elias, B P Finney, J J Fitzpatrick, S V Funder, A Geirsdóttir, T D Herbert, L D Hinzman, D S Kaufman, G M MacDonald, L Polyak, A Robock, M C Serreze, J P Smol, R Spielhagen, J W C White, A P Wolfe, and E W Wolff</i>	113
The Younger Dryas Climate Event <i>A E Carlson</i>	126

PALEOCLIMATE MODELING	135
Data–Model Comparisons <i>S P Harrison</i>	135
Quaternary Environments <i>C S Jackson</i>	147
The Last Interglacial <i>M Montoya</i>	155
Last Glacial Maximum GCMs <i>A J Broccoli</i>	165
Paleo-ENSO <i>C J Mock</i>	171
 PALEOCLIMATE RECONSTRUCTION	 179
Approaches <i>B N Shuman</i>	179
Pliocene Environments <i>R Z Poore</i>	185
Paleodroughts and Society <i>C J Mock</i>	194
Sub-Milankovitch (DO/Heinrich) Events <i>L Labeyrie, L Skinner, and E Cortijo</i>	200
Paleotempestology <i>K-b Liu</i>	209
Younger Dryas Oscillation, Global Evidence <i>S Björck</i>	222
The Last Millennium <i>M E Mann</i>	229
Historical Climatology <i>M Chenoweth</i>	237
 Paleoclimate Relevance to Global Warming <i>S H Schneider and M D Mastrandrea</i>	 244
Paleohydrology <i>V R Thorndycraft</i>	253
 PALEOLIMNOLOGY	 259
Overview of Paleolimnology <i>M S V Douglas</i>	259
Cladocera <i>M Rautio and L Nevalainen</i>	271
Freshwater Mollusks <i>C G De Francesco</i>	281
Lake Chemistry <i>D Antoniadis</i>	292

Physical Properties of Lake Sediments <i>K R Hodder and R Gilbert</i>	300
Contributions of Paleolimnological Research to Biogeography <i>K A Moser</i>	313
Pigment Studies <i>S McGowan</i>	326
Multiproxy Approaches <i>N Michelutti and J P Smol</i>	339
Visible and Infrared Spectroscopical Applications <i>P Rosén and H Vogel</i>	349
PALEOLITHIC <i>see</i> ARCHAEOLOGICAL RECORDS: Overview; 2.7 Myr–300 000 Years Ago in Africa; 2.7 Myr–300 000 Years Ago in Asia; 1.9 Myr–300 000 Years Ago in Europe; Global Expansion 300 000–8000 Years Ago, Africa; Global Expansion 300 000–8000 Years Ago, Asia; Global Expansion 300 000–8000 Years Ago, Australia; Global Expansion 300 000–8000 Years Ago, Americas; Neanderthal Demise	
PALEOSOLS AND WIND-BLOWN SEDIMENTS	357
Overview <i>D R Muhs</i>	357
Nature of Paleosols <i>J A Mason and P M Jacobs</i>	367
Mineral Magnetic Analysis <i>M J Singer and K L Verosub</i>	375
Soil Micromorphology <i>R A Kemp</i>	381
Weathering Profiles <i>E A Bettis III</i>	392
Soil Morphology in Quaternary Studies <i>L McFadden</i>	402
Biogeochemical Role of Dust in Quaternary Climate Cycles <i>K E Kohfeld</i>	412
PATTERNED GROUND <i>see</i> PERMAFROST AND PERIGLACIAL FEATURES: Patterned Ground	
PEAT STUDIES <i>see</i> PLANT MACROFOSSIL METHODS AND STUDIES: Mire and Peat Macros	
PEDOSTRATIGRAPHY <i>see</i> QUATERNARY STRATIGRAPHY: Pedostratigraphy	
PERMAFROST AND PERIGLACIAL FEATURES	421
Active Layer Processes <i>N I Shiklomanov and F E Nelson</i>	421
Cryoturbation Structures <i>J Vandenberghe</i>	430
Ice Wedges and Ice-Wedge Casts <i>J Murton</i>	436
Patterned Ground <i>C K Ballantyne</i>	452

Permafrost <i>C R Burn</i>	464
Frost Mounds: Active and Relict Forms <i>N Ross</i>	472
Slope Deposits and Forms <i>C Harris</i>	481
Periglacial Fluvial Sediments and Forms <i>J van Huissteden, J Vandenberghe, P L Gibbard, and J Lewin</i>	490
Rock Weathering <i>J Murton</i>	500
Permafrost and Glacier Interactions <i>R I Waller</i>	507
Block/Rock Streams <i>P Wilson</i>	514
Blockfields (Felsenmeer) <i>B R Rea</i>	523
Rock Glaciers and Protalus Forms <i>A Kääb</i>	535
Yedoma: Late Pleistocene Ice-Rich Syngenetic Permafrost of Beringia <i>L Schirrmeister, D Froese, V Tumskey, G Grosse, and S Wetterich</i>	542
Paraglacial Geomorphology <i>C K Ballantyne</i>	553
Talus Slopes <i>B H Luckman</i>	566
Thermokarst Topography <i>C R Burn</i>	574
PERMAFROST HISTORY <i>see</i> PERMAFROST AND PERIGLACIAL FEATURES: Active Layer Processes; Cryoturbation Structures; Patterned Ground; Permafrost	
Phytoliths <i>M S Blinnikov</i>	582
PIGMENTS, FOSSIL <i>see</i> PALEOLIMNOLOGY: Pigment Studies	
PINGOS <i>see</i> PERMAFROST AND PERIGLACIAL FEATURES: Frost Mounds: Active and Relict Forms	
Plant Macrofossil Introduction <i>H H Birks</i>	593
PLANT MACROFOSSIL METHODS AND STUDIES	613
CO ₂ Reconstruction from Fossil Leaves <i>M Rundgren</i>	613
Megafossils <i>G M MacDonald</i>	621
Dendroarchaeology <i>R H Towner</i>	630

Mire and Peat Macros <i>D Mauquoy and B van Geel</i>	637
Paleolimnological Applications <i>M-J Gaillard and H H Birks</i>	657
Rodent Middens <i>S A Elias</i>	674
Surface Samples, Taphonomy, Representation <i>A C Dieffenbacher-Krall</i>	684
Treeline Studies <i>W Tinner</i>	690
Use in Environmental Archaeology <i>S Jacomet</i>	699
Validation of Pollen Studies <i>S T Jackson and R K Booth</i>	725
PLANT MACROFOSSIL RECORDS	733
Arctic Eurasia <i>F Kienast</i>	733
Arctic North America <i>N H Bigelow, G D Zazula, and D E Atkinson</i>	746
Greenland <i>O Bennike</i>	760
Holocene North America <i>R G Baker</i>	768
Late Glacial Multidisciplinary Studies <i>B Ammann, H H Birks, A Walanus, and K Wasylkowa</i>	785
PLATE TECTONICS <i>see</i> GLACIAL LANDFORMS: Glacitectonic Structures and Landforms; GLACIATION, CAUSES: Tectonic Uplift and Continental Configurations; GLACIATIONS: Late Quaternary in North America	
PLIOCENE ENVIRONMENTS <i>see</i> PALEOCLIMATE RECONSTRUCTION: Pliocene Environments	
Pollen Analysis, Principles <i>H Seppä</i>	794
POLLEN METHODS AND STUDIES	805
Use of Pollen as Climate Proxies <i>S Brewer, J Guiot, and D Barboni</i>	805
Reconstructing Past Biodiversity Development <i>B V Odgaard</i>	816
Numerical Analysis Methods <i>H J B Birks</i>	821
Databases and Their Application <i>E C Grimm, R H W Bradshaw, S Brewer, S Flantua, T Giesecke, A-M Lézine, H Takahara, and J W Williams</i>	831
Surface Samples and Trapping <i>A Poska</i>	839

Stand-Scale Palynology <i>R H W Bradshaw</i>	846
Changing Plant Distributions and Abundances <i>T Giesecke</i>	854
The Biome Approach to Reconstructing Past Vegetation <i>M E Edwards</i>	861
POLLSCAPE Model: Simulation Approach for Pollen Representation of Vegetation and Land Cover <i>S Sugita</i>	871
Archaeological Applications <i>M-J Gaillard</i>	880

VOLUME 4

Pollen Records, Last Interglacial of Europe <i>C Tzedakis</i>	1
--	---

POLLEN RECORDS, LATE PLEISTOCENE 9

Africa <i>M E Meadows and B M Chase</i>	9
Australasia <i>P Kershaw and S van der Kaars</i>	18
Northern Asia <i>A V Lozhkin and P M Anderson</i>	27
Northern North America <i>N H Bigelow</i>	39
South America <i>H Hooghiemstra and J C Berrio</i>	52
Middle and Late Pleistocene in Southern Europe <i>J-L de Beaulieu, P C Tzedakis, V Andrieu-Ponel, and F Guiter</i>	63
Western North America <i>R S Thompson</i>	72

POLLEN RECORDS, POSTGLACIAL 85

Africa <i>A-M Lézine</i>	85
Australia and New Zealand <i>J R Dodson</i>	104
Northeastern North America <i>J W Williams and B N Shuman</i>	115
Northwestern North America <i>D G Gavin and F S Hu</i>	124
Southeastern North America <i>D A Willard</i>	133
Southwestern North America <i>P E Wigand</i>	142

South America <i>H Behling</i>	156
Northern Asia <i>A A Andreev and P E Tarasov</i>	164
Northern Europe <i>M J Bunting</i>	173
Southern Europe <i>L Sadori</i>	179

POTASSIUM-ARGON DATING *see* K/Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ Dating

Q

QUATERNARY STRATIGRAPHY 189

Overview <i>B Pillans</i>	189
Continental Biostratigraphy <i>T van Kolfschoten</i>	206
Chronostratigraphy <i>B Pillans</i>	215
Climatostratigraphy <i>P L Gibbard</i>	222
Lithostratigraphy <i>W E Westerhoff and H J T Weerts</i>	227
Morphostratigraphy/Allostratigraphy <i>P D Hughes</i>	243
Pedostratigraphy <i>A Palmer</i>	250
Sequence Stratigraphy <i>T R Naish, S T Abbott, and R M Carter</i>	260
Tephrochronology <i>B V Alloway, D J Lowe, G Larsen, P A R Shane, and J A Westgate</i>	277

R

RADIOLARIAN STUDIES *see* PALEOCEANOGRAPHY, BIOLOGICAL PROXIES: Radiolarians and Silicoflagellates

RADIOCARBON DATING 305

Conventional Method <i>G T Cook and J van der Plicht</i>	305
AMS Radiocarbon Dating <i>A J T Jull</i>	316
Sources of Error <i>E M Scott</i>	324

Variations in Atmospheric ^{14}C <i>J van der Plicht</i>	329
Causes of Temporal ^{14}C Variations <i>G S Burr</i>	336
Calibration of the ^{14}C Record <i>P J Reimer, R W Reimer, and M Blaauw</i>	345
Charcoal <i>M I Bird</i>	353
^{14}C of Plant Macrofossils <i>C Hatté and A J T Jull</i>	361

ROCK GLACIERS *see* PERMAFROST AND PERIGLACIAL FEATURES: Rock Glaciers and Protalus Forms

RODENT MIDDENS *see* PLANT MACROFOSSIL METHODS AND STUDIES: Rodent Middens

S

SEA LEVEL STUDIES 369

Overview <i>I Shennan</i>	369
Geomorphological Indicators <i>P A Pirazzoli</i>	377
Sedimentary Indicators of Relative Sea-Level Changes – High Energy <i>J A G Cooper</i>	385
Sedimentary Indicators of Relative Sea-Level Changes – Low Energy <i>R J Edwards</i>	396
Coral Records of Relative Sea-Level Changes <i>C D Woodroffe</i>	409
Microfossil-Based Reconstructions of Holocene Relative Sea-Level Change <i>W R Gehrels</i>	419
Eustatic Sea-Level Changes – Glacial–Interglacial Cycles <i>C V Murray-Wallace</i>	429
Eustatic Sea-Level Changes Since the Last Glacial Maximum <i>P L Whitehouse and S L Bradley</i>	439
Isostasy: Glaciation-Induced Sea-Level Change <i>G Milne and I Shennan</i>	452
Use of Cave Data in Sea-Level Reconstructions <i>A Dutton</i>	460

SEA-LEVELS, LATE QUATERNARY 467

Late Quaternary Relative Sea-Level Changes in High Latitudes <i>C Ó Cofaigh and M J Bentley</i>	467
Late Quaternary Sea-Level Changes in Greenland <i>S A Woodroffe and A J Long</i>	481
Late Quaternary Relative Sea-Level Changes at Mid-Latitudes <i>A C Kemp, B P Horton, and S E Engelhart</i>	489

Late Quaternary Relative Sea-Level Changes in the Tropics <i>Y Zong</i>	495
Tectonics and Relative Sea-Level Change <i>A R Nelson</i>	503

SEA SURFACE TEMPERATURE RECONSTRUCTION *see* PALEOCEANOGRAPHY: Paleoceanography An Overview; PALEOCEANOGRAPHY, BIOLOGICAL PROXIES: Alkenone Paleothermometry Based on the Haptophyte Algae; Biomarker Indicators of Past Climate; Coccolithophores; Dinoflagellates; Marine Diatoms; Planktic Foraminifera; Temperature Proxies, Census Counts; PALEOCEANOGRAPHY, PHYSICAL AND CHEMICAL PROXIES: Carbon Cycle Proxies ($\delta^{11}\text{B}$, $\delta^{13}\text{C}_{\text{calcite}}$, $\delta^{13}\text{C}_{\text{organic}}$, Shell Weights, B/Ca, U/Ca, Zn/Ca, Ba/Ca); Mg/Ca and Sr/Ca Paleothermometry from Calcareous Marine Fossils; Oxygen-Isotope Stratigraphy of the Oceans; PALEOCEANOGRAPHY, RECORDS: Late Pleistocene North Atlantic; Late Pleistocene North Pacific; Late Pleistocene South Atlantic; Postglacial Indian Ocean; Postglacial North Atlantic; Postglacial North Pacific; Postglacial South Pacific

SILICOFLAGELLATES *see* PALEOCEANOGRAPHY, BIOLOGICAL PROXIES: Radiolarians and Silicoflagellates

SPELEOTHEMS *see* CARBONATE STABLE ISOTOPES: Speleothems

STABLE ISOTOPE STUDIES, CARBONATES *see* CARBONATE STABLE ISOTOPES: Overview; Speleothems; Terrestrial Teeth and Bones; Nonmarine Biogenic Carbonates; Terrestrial Organic Materials; Non-Lacustrine Terrestrial Studies; Lake Sediments

STABLE ISOTOPE STUDIES, DEEP SEA RECORDS *see* PALEOCEANOGRAPHY, PHYSICAL AND CHEMICAL PROXIES: Oxygen-Isotope Stratigraphy of the Oceans; Oxygen Isotope Composition of Seawater; Salinity Proxies $\delta^{18}\text{O}$; PALEOCEANOGRAPHY, RECORDS: Early Pleistocene; Late Pleistocene North Atlantic; Late Pleistocene South Atlantic; Postglacial Indian Ocean; Postglacial North Atlantic; Postglacial North Pacific; Postglacial South Pacific; ICE CORE METHODS: Stable Isotopes; ICE CORE RECORDS: Antarctic Stable Isotopes; Greenland Stable Isotopes; Correlations Between Greenland and Antarctica

STABLE ISOTOPE STUDIES, ICE CORES *see* ICE CORE METHODS: Stable Isotopes; ICE CORE RECORDS: Antarctic Stable Isotopes; Greenland Stable Isotopes; Correlations Between Greenland and Antarctica

SUB-MILANKOVITCH EVENTS *see* PALEOCLIMATE RECONSTRUCTION: Sub-Milankovitch (DO/Heinrich) Events

T

TALUS SLOPES *see* PERMAFROST AND PERIGLACIAL FEATURES: Talus Slopes

TEETH AND BONES, STABLE ISOTOPE STUDIES *see* CARBONATE STABLE ISOTOPES: Terrestrial Teeth and Bones

TEPHROCHRONOLOGY *see* QUATERNARY STRATIGRAPHY: Tephrochronology

THERMOHALINE CIRCULATION OF THE OCEANS *see* GLACIAL CLIMATES: Thermohaline Circulation

THERMOLUMINESCENCE DATING *see* LUMINESCENCE DATING: Thermoluminescence

TILL, GLACIAL *see* GLACIAL LANDFORMS, SEDIMENTS: Till Fabric Analysis; Tills; Glacial Erratics and Till Dispersal Indicators

TREE RINGS *see* Dendrochronology; Dendroclimatology; GLACIAL LANDFORMS, TREE RINGS: Dendrogeomorphology; Dendroglaciology; PLANT MACROFOSSIL METHODS AND STUDIES: Dendroarchaeology

TREELINE RECONSTRUCTION *see* POLLEN METHODS AND STUDIES: Archaeological Applications

U**USE OF QUATERNARY PROXIES IN FORENSIC SCIENCE 522**

Use of Geology in Forensic Science: Search to Locate Burials 522
L Donnelly

Soils and Sediment 535
R C Murray

The Use of Macroscopic Plant Remains in Forensic Science 542
J H Bock

Insects 548
D E Gennard

Analytical Techniques in Forensic Palynology 556
V M Bryant

U-Series Dating 567
W G Thompson

V

Varved Lake Sediments 573
B Zolitschka

Varved Marine Sediments 582
K A Hughen and B Zolitschka

Vertebrate Overview 590
D C Schreve

VERTEBRATE RECORDS 599

Early Pleistocene 599
L Rook, M Delfino, M P Ferretti, and L Abbazzi

Early and Middle Pleistocene of Northern Eurasia 605
I Vislobokova and A Tesakov

Mid-Pleistocene of Africa 615
L C Bishop and A Turner

Mid-Pleistocene of Australia 621
G J Prideaux

Mid-Pleistocene of Europe 639
R Sardella

Mid-Pleistocene of North America 646
C J Bell

Mid-Pleistocene of Southern Asia 651
J de Vos

Late Pleistocene of Africa 664
T E Steele

Late Pleistocene of North America 673
J I Mead

Late Pleistocene of South America <i>M Ubilla</i>	680
Late Pleistocene of Southeast Asia <i>Y Chaimanee</i>	693
Late Pleistocene Megafaunal Extinctions <i>S A Elias and D C Schreve</i>	700
Late Pleistocene Mummified Mammals <i>C R Harington</i>	713
VERTEBRATE STUDIES	719
Ancient DNA <i>I Barnes, R Barnett, and B Shapiro</i>	719
Speciation and Evolutionary Trends in Quaternary Vertebrates <i>A M Lister</i>	723
Dwarfing and Gigantism in Quaternary Vertebrates <i>M R Palombo and R Rozzi</i>	733
Interactions with Hominids <i>K V Boyle</i>	748
VOLCANIC ASH <i>see</i> QUATERNARY STRATIGRAPHY: Tephrochronology	
VULCANISM <i>see</i> GLACIAL CLIMATES: Volcanic and Solar Forcing; GLACIAL LANDFORMS: Quaternary Vulcanism: Subglacial Landforms; QUATERNARY STRATIGRAPHY: Tephrochronology	
W	
WIND-BLOWN SEDIMENTS (LOESS, SAND DUNES) <i>see</i> Eolian Records, Deep-Sea Sediments; Loess Deposits: Origins and Properties; DUNE FIELDS: High Latitudes; Mid-Latitudes; Low Latitudes; LOESS RECORDS: Central Asia; China; Europe; North America; South America	
Y	
YOUNGER DRYAS OSCILLATION <i>see</i> PALEOCLIMATE RECONSTRUCTION: Younger Dryas Oscillation, Global Evidence	
<i>Index</i>	757

PALEOCEANOGRAPHY, RECORDS

Contents

Early Pleistocene

Late Pleistocene North Atlantic

Late Pleistocene South Atlantic

Late Pleistocene North Pacific

Postglacial Indian Ocean

Postglacial North Atlantic

Postglacial North Pacific

Postglacial South Pacific

Early Pleistocene

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Introduction

From about 3 My ago to now, the climate of the Earth has been marked by the succession of large glacial–interglacial changes which shaped the relief of the Earth and eventually led toward modern biodiversity. Oceanic sedimentary sequences provide the most continuous records of these dramatic variations. Among the oceanic indicators of past climatic changes, the oxygen isotope composition of Foraminifera in deep-sea cores, as a proxy of past changes in global ice cap volume, has been one of the most insightful tool to reveal the history of the continental ice sheets. The most striking result of paleoceanographic studies using $\delta^{18}\text{O}$ was the discovery of periodic glacial–interglacial successions, which were paced by orbital forcing as theorized by Milankovitch's theory. While the late Pleistocene spectrum is mainly marked by the ~ 100 ka period, the succession of glacial and interglacial periods during the early Pleistocene period – from 2.6 to about 0.8 Ma – was marked by a very regular pacing of 41 ka. Moreover, oxygen isotope records showed that within the same interval, the buildup of ice caps in the high latitudes during glaciations was considerably smaller during the early Pleistocene than later. Finally, the configuration of the continents during the early Pleistocene was not significantly different than during the late Pleistocene. The 41 ka 'world' – the early Pleistocene – is thus a key period of the Earth's history when global climate changed toward modern climate without any strong external forcing changes. This period offers unique opportunities to infer processes at play in glacial to interglacial transitions.

Stratigraphy of the Early Pleistocene

In the oceanic realm, the combination of orbital cyclostratigraphy with paleomagnetism, biostratigraphy, and absolute

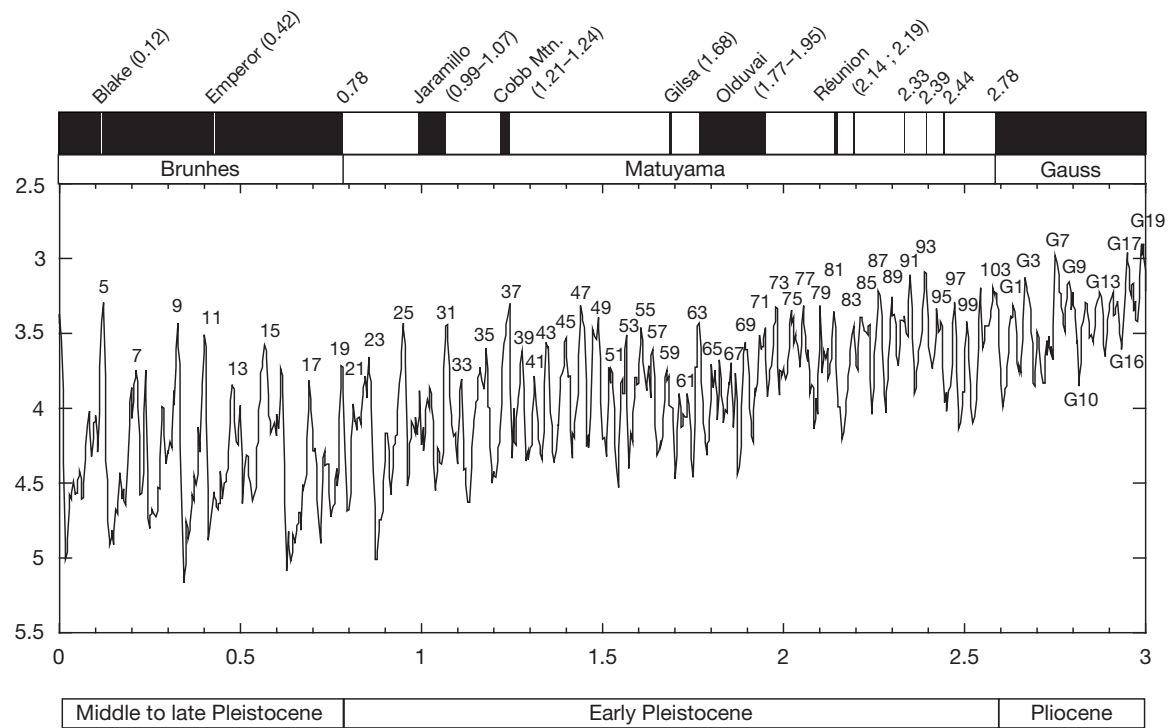
dating made possible the development of an extremely accurate and precise stratigraphy of early Pleistocene paleoceanographic records ([Figure 1](#)). The early Pleistocene encompasses the Gelasian, Calabrian, and Tyrhenian subdivisions. Recent changes in the stratigraphy have put the Pliocene–Pleistocene boundary at about 2.6 Ma, during marine isotope stage (MIS) 103 ([Rio et al., 1998](#)), also corresponding with the lower boundary of the Quaternary.

Pliocene–Pleistocene Boundary

The Global Stratotype Section and Point (GSSP) for the Pliocene–Pleistocene boundary was proposed in 1998 for the Gelasian stage, and placed close to the Gauss–Matuyama boundary (2.588 Ma; [Rio et al., 1998](#)). The GSSP is at Monte San Nicola, Sicily. In oceanic stable oxygen isotope records, the Pliocene–Pleistocene boundary is associated with MIS 103, corresponding to the onset of Northern Hemisphere glaciations as reflected in proxy records which are discussed later (ice-rafted debris (IRD), sea-surface temperatures (SSTs), loess records).

Early/Late Pleistocene Boundary and the Mid-Pleistocene Transition

While the beginning of the early Pleistocene is clearly defined, the end of the early Pleistocene period is informal. Conventionally, the Brunhes–Matuyama reversal (0.78 Ma) defines the early to mid-Pleistocene boundary ([Gradstein et al., 2005](#)). In paleoceanographic records, the period that leads to the dominance of the large 100 ka glaciations is called the mid-Pleistocene transition or mid-Pleistocene revolution. Because no formal definition was ever voted, and because it corresponds to a change in both time and frequency domain that depends upon the paleoceanographic record used, the age of this



boundary can vary from about 1 to about 0.65 Ma. Advanced time series statistical analyses using benthic $\delta^{18}\text{O}$ records as a proxy of past global changes in ice volume indicate that the midpoint of the transition in mean ice volume was 922 ± 12 ka, whereas the maximum amplitude in the 100 ka band was attained at 641 ± 9 ka (Mudelsee and Schulz, 1997).

Ice-Volume and Sea-Level Changes

During the early Pleistocene, a secular trend led to the onset of the large glaciations of the late Pleistocene. In the oceanic realm, this secular trend has been clearly recorded by deep-sea benthic Foraminifera $\delta^{18}\text{O}$ reconstructions, which are influenced by global ice-volume changes and by a minor contribution of deep-sea temperature changes, recently deconvolved using Mg/Ca paleothermometry. A number of records have been documented throughout the different oceanic basins that display early Pleistocene glacial–interglacial changes (Figure 2). Shackleton et al. (1976), using the historic V28-239 piston core from the equatorial Pacific ocean, were the first to reconstruct a complete record of $\delta^{18}\text{O}$ variations at a sampling rate that allowed the detection of glacial–interglacial cycles throughout the Pleistocene. Later, high-resolution records from the equatorial Atlantic (site DSDP 502) and from the Northern Atlantic (site DSDP 607) confirmed that the glacial–interglacial amplitude of $\delta^{18}\text{O}$ changes was reduced during the early Pleistocene (Prell et al., 1982; Ruddiman et al., 1989). More recently, at the same site, Sosdian and Rosenthal reconstructed the first long-term record of glacial–interglacial deep-sea temperature and ice-volume changes (Sosdian and Rosenthal, 2009). Three salient features characterize the early Pleistocene: (1) the global ice-volume difference between glacial and interglacial stages was reduced during the early Pleistocene; (2) the glacial–interglacial cycles were paced by a 41 ka rhythm during the early Pleistocene, whereas the late Pleistocene is marked by the onset of ~ 100 ka glacial–interglacial cycles; (3) global ice-volume buildup during glacial inceptions and ice-volume melting during terminations (deglaciations) are fairly symmetric during the early Pleistocene.

Amplitude of Ice-Volume Changes

The early Pleistocene was characterized by smaller glacial ice caps during glacial intervals, but also by ‘somewhat’ more ice during the interglacials (Prell et al., 1982). Accurate estimates of glacial and interglacial ice volumes during the early Pleistocene are still poorly constrained because of uncertainties in proxy-based methods. Two approaches have been followed to unravel deep-sea temperature effects from ice-volume changes on the benthic isotopic record. The first approach uses benthic paleothermometry (Sosdian and Rosenthal, 2009), and the second uses ice-sheet modeling to constrain the ice-volume component (Bintanja et al., 2005). Overall, using the former approach, temperature uncertainties and $\delta^{18}\text{O}$ -associated uncertainties induce a ± 32 m uncertainty in reconstructed sea levels during glacial–interglacial transitions (Figure 3). The amplitude of glacial–interglacial sea-level changes is reduced, compared to late Pleistocene oscillations, by approximately 20 m.

Pacing of Ice-Volume Changes

The main cyclicity of glacial–interglacial successions during the early Pleistocene has been shown to be close to obliquity (41 ka), in contrast to precession during the late Pliocene. This period is clearly detected in deep-sea benthic $\delta^{18}\text{O}$ records from all around the world (Lisiecki and Raymo, 2005), but also in planktonic records from the major ocean basins. Moreover, this 41 ka rhythm is a robust feature, because it can be detected in records whose timescales are based on independent stratigraphy (i.e., not based on Milankovitch cycles), as demonstrated in the landmark DSDP 607 record from the North Atlantic (Raymo and Nisancioglu, 2003; Ruddiman et al., 1989) and from the world’s oceans (Huybers and Wunsch, 2004). Lastly, this North Atlantic record clearly showed a co-variation of the $\delta^{18}\text{O}$ signal with IRD records, confirming that the melting of the large Northern ice caps was synchronous with the isotopic record.

Symmetry of Glacial and Interglacial Transitions

The last feature of ice-volume changes during the early Pleistocene is that the curve representing the succession of glacial and interglacial stages recorded in benthic $\delta^{18}\text{O}$ records following the 41 ka cycle described earlier is marked by a distinctive sinusoidal shape. This pattern clearly contrasts with the 100 ka sawtooth shape in the curve representing late Pleistocene ice-volume changes (Imbrie et al., 1993). The symmetry of early Pleistocene 41 ka glacial–interglacial changes implies a linear response of the climate system to orbital obliquity forcing, in contrast with late Pleistocene Ice Ages that suggest a strong nonlinear response of the climate system to orbital forcing. As discussed, the importance of total summer insolation, forced mainly by obliquity changes, may explain the nonskewed shape of glacial–interglacial changes.

Paleotemperature Changes

Since the early 2000s, the quantification of the oceanic temperature changes across glacial–interglacial cycles during the early Pleistocene has considerably improved. This progress is mainly due to the advance of geochemical proxies, such as alkenone paleothermometry and Mg/Ca in planktonic Foraminifera shells. Indeed, classical micropaleontological markers are affected by increased uncertainties with regard to the relevance of analogs as one gets further back into the past. Geochemical paleothermometers are less sensitive to this issue, though two issues may be critical for estimates of past temperature changes. First, preservation issues are critical for Mg/Ca paleothermometry, and floristic changes throughout the Pleistocene may change the specific phytoplankton biosynthesizing alkenones. A number of records have now been developed throughout the world’s oceans that shed light on both the long-term trends and glacial–interglacial changes in ocean-surface temperature during the early Pleistocene (Figure 3).

Long-Term Trend of Paleotemperature Changes

During the early Pleistocene and toward the mid-Pleistocene transition, a long-term cooling has been shown to affect both

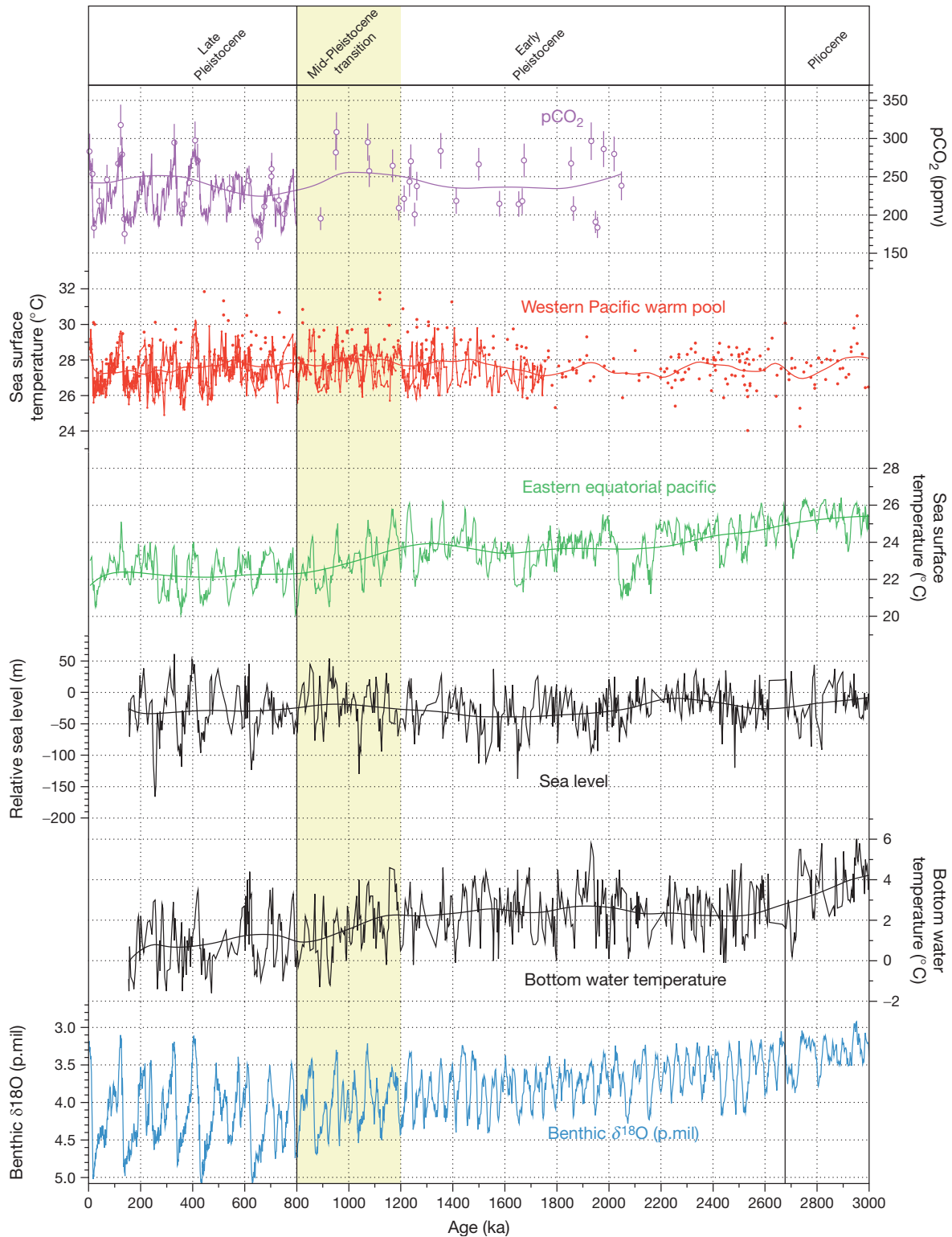


Figure 3 Comparison of Pleistocene atmospheric pCO_2 concentrations measured in ice-core records (purple line; Lüthi et al., 2008) and inferred from boron isotope measurements in Foraminifera (Hönisch et al., 2009); western equatorial Pacific compilation of Mg/Ca-derived sea-surface temperatures (SSTs) (de Garidel-Thoron et al., 2005; Medina-Elizalde and Lea, 2005; Wara et al., 2005); eastern equatorial Pacific SSTs (Liu et al., 2008); sea-level changes inferred from benthic Foraminifera paleotemperature, deep-sea temperature changes (Sodrian and Rosenthal, 2009), and a benthic $\delta^{18}O$ stack (Lisiecki and Raymo, 2005).

the high latitudes and deep-sea waters. This cooling is a continuation of the long-term Pliocene trend that finally led to Pleistocene glaciations. The long-term cooling of deep-sea waters, first inferred from $\delta^{18}\text{O}$ in benthic Foraminifera, was confirmed by Mg/Ca paleothermometry on the same organisms at low resolution and has been further resolved more recently. This cooling also affected polar SSTs, as demonstrated by North Atlantic ocean $\delta^{18}\text{O}$ reconstructions of surface-dwelling Foraminifera. The high-latitude cooling is indirectly confirmed by low-latitude SST records from upwelling areas. Indeed, records of SSTs from the Benguela current and eastern equatorial Pacific showed that the long-term cooling observed in the high latitudes was also present at low latitudes. However, this pattern of cooling was restricted to the upwelling area (Marlow et al., 2000). Indeed, most of the subtropical surface ocean did not show any long-term cooling or warming throughout the Pleistocene (de Garidel-Thoron et al., 2005). Such a stability of Pleistocene subtropical temperatures has been observed in the western equatorial Pacific, in the western equatorial Atlantic (Sepulcre et al., 2011), and in the eastern equatorial Atlantic (de Garidel-Thoron et al., 2005; McClymont et al., 2005; Schefuß et al., 2004). The long-term cooling of high latitudes and upwelling SSTs, compared to the stability of subtropical SSTs, suggests an increase of meridional and zonal thermal gradients across the different oceans. This finding confirms the robustness of the alkenone and Mg/Ca methods, as their independent estimates of past SSTs converge toward the same results. In summary, while the subtropical ocean remained warm throughout the early Pleistocene, high latitudes and deep waters experienced a significant cooling during the early Pleistocene.

Glacial–Interglacial Changes

While there is a clear decoupling of subtropical SSTs with high latitudes and upwelling SSTs on the long-term trend, such a difference does not exist at the orbital timescale. During the 41 ka glacial–interglacials, both low- and high-latitude SSTs seem to have responded in phase to the obliquity forcing. During a maximum in obliquity, ice volumes were reduced, and SSTs at different latitudes around the world reached their maximum. Maximum SSTs at low latitudes in records from the eastern and western equatorial Pacific are thus out of phase with local insolation forcing (de Garidel-Thoron et al., 2005; Liu and Herbert, 2004). Moreover, the strong 41 ka signal in low-latitude SST records implies a remote forcing in the low latitudes by some high-latitude processes. The weakness of a strong precessional signal in low-latitude SST records further underlines the role of high-latitude insolation forcing on low-latitude climate response. Among the processes suggested to be at play for the phasing of low-latitude SST signals, low- to high-latitude insolation gradients have been suggested as a potential driver of early Pleistocene climate (Raymo and Nisancioglu, 2003). It was recently proposed that total insolation over the duration of summer in both Northern and Southern Hemispheres was the most likely driver of 41 ka glacial oscillations, rather than insolation changes in the Northern Hemisphere or latitudinal insolation gradients (Huybers, 2006).

Thermohaline Circulation Changes

Early Pleistocene global topography and bathymetry were not significantly different from those of the modern Earth. No major tectonic change can be invoked to explain the main Pleistocene paleoceanographic changes. The Panama Strait was closed at least 1 Ma prior to the Pliocene–Pleistocene boundary, and the configuration of straits in the Indonesian archipelago was likewise close to the modern one.

Thermohaline circulation has been clearly shown to dramatically change over the last climatic cycle (late Pleistocene). During an interglacial period such as the Holocene, the ‘conveyor’ is stronger than during glacial periods when thermohaline circulation is reduced. At the millennial timescale, there is a large body of evidence demonstrating that thermohaline circulation was weaker during stadials of the last glacial stage and during millennial cold snaps, at least in the North Atlantic region. Early Pleistocene thermohaline circulation changes have been extensively investigated in order to test whether similar changes in oceanic circulation occurred in the distant past.

The main evidence used to reconstruct past changes in thermohaline circulation is provided by the $\delta^{13}\text{C}$ composition of benthic Foraminifera from North Atlantic deep-sea cores at different depths over the water column. Past $\delta^{13}\text{C}$ gradients are traditionally used as a tracer of the different water masses, and give insights on past changes in North Atlantic Deep Water production. However, $\delta^{13}\text{C}$ in deep waters from the North Atlantic is also influenced by the production rate of deep water in the Southern Ocean. During the early Pleistocene, few changes were observed in the modes of variability of North Atlantic deep waters. However, at the end of the early Pleistocene, deep sites from the North Atlantic record a mid-Pleistocene reduction in glacial $\delta^{13}\text{C}$ (Raymo et al., 1997). Observations on the relative influence of northern and southern ocean deep waters led to the formulation of two different hypotheses: (1) this observation was first interpreted in the early 1990s to be indicative of enhanced glacial suppression of the North Atlantic Deep Water at the end of the early Pleistocene (Raymo et al., 1990); (2) however, recent studies point toward a more critical role of the production rate of deep waters around Antarctica on the isotopic composition of deep-sea waters in deepest sites from the Northern Atlantic (Raymo et al., 2004). This increased role of the Southern ocean in deep-water masses would imply slighter variations in thermohaline circulation during the mid-Pleistocene. In summary, though thermohaline circulation likely changed over the course of the early Pleistocene, there is as yet no clear evidence of the amplitude of these changes and of their role on global climate.

From 41 to 100 ka: The Mid-Pleistocene Transition

The period that marks the onset of the largest glaciations of the Cenozoic era, at the end of the early Pleistocene, is called mid-Pleistocene transition or mid-Pleistocene revolution. During this transition, the climate system bifurcated from a linear response to obliquity forcing to a nonlinear response to orbital forcing, which is characterized by a strong 100 ka period, presumably correlated with orbital eccentricity. Note that a

recent study indicates that the statistical significance of late Pleistocene 100 ka cycles is weak, and that deglaciations (terminations) occurred each two or three obliquity cycles (41 ka; Huybers and Wunsch, 2005). In any case, the period preceding the mid-Pleistocene revolution led to the largest Pleistocene ice caps, and a shift from a linear (early Pleistocene; Imbrie et al., 1992) to a nonlinear climate response (late Pleistocene; Imbrie et al., 1993). The Mid-Pleistocene transition is not associated with any abrupt changes in orbital forcing. Many hypotheses have been formulated to account for this. All these hypotheses suggest a mechanism that preserves Northern Hemisphere ice sheets throughout one or two obliquity cycles, therefore allowing the buildup of large late Pleistocene ice sheets.

Thermohaline Switch

In this hypothesis, marine feedback triggered a marine-induced instability of continental ice volume at the mid-Pleistocene transition. This feedback was attributed to the reorganization of the 'global conveyor belt' of thermohaline circulation (DeBlonde and Peltier, 1991). It was also suggested that the critical increase in Northern ice volume modified Arctic oceanic circulation, and therefore altered global patterns of ocean circulation (Berger et al., 1994). As discussed earlier (see section 'Thermohaline Circulation Changes'), changes in early Pleistocene thermohaline circulation are still poorly resolved and cannot be used to assess this hypothesis.

Atmospheric CO₂ Hypothesis

The long-term cooling trend that affected the high latitudes might have been caused by a secular decrease of atmospheric CO₂ concentration. In this hypothesis, during the early Pleistocene, higher concentrations in atmospheric CO₂ prevented the conservation of Northern Hemisphere ice sheets over two precession cycles. This hypothesis has been successfully simulated (Berger et al., 1999), but direct evidence for the secular decrease in atmospheric CO₂, likely from tectonic processes, has not been noted in ice-core records (Lüthi et al., 2008) or in recent boron isotope reconstructions of atmospheric CO₂ (Figure 3; Hönlisch et al., 2009).

Sea-Ice Switch Hypothesis

This hypothesis is based on deep-water cooling during the Pleistocene, inferred from benthic Foraminifera records (see section 'Ice-Volume and Sea-Level Changes'). In this hypothesis, the long-term cooling of the deep ocean resulted in an extensive sea-ice cover of the polar oceans. Such an extensive sea-ice cover activated the 'sea-ice switch' that changed the ablation and accumulation rates over the continental ice sheets (Tziperman and Gildor, 2003). This mechanism reproduces a change from a linear to a nonlinear response of the climate system. However, this hypothesis has not yet been validated by any proxy data of sea-ice cover.

Regolith Hypothesis

In this hypothesis, the weathered soil (regolith) lying beneath the Laurentide ice sheet was progressively eroded during the

Pliocene and early Pleistocene glacial cycles. This soft bed could sustain only thin ice sheets which responded linearly to orbital forcing (precession and obliquity). By the mid-Pleistocene, mineralogical and geochemical data indicate that the regolith was largely eroded, allowing the ice sheet to rest directly on the crystalline bedrock and enabling the ice sheets to reach greater heights (Clark and Pollard, 1998).

Chaotic Response to Obliquity Forcing

This hypothesis posits that the switch from the 41 to 100 ka results from the chaotic nature of glacial Pleistocene cycles, and that the change in frequencies might occur spontaneously (Huybers, 2009). In this hypothesis, the late Pleistocene 100 ka cycles are, in fact, close to multiples of obliquity cycles (mostly 80 ka), and no large change in external forcing occurred during this time interval. This hypothesis was tested using a modified version of the Imbrie and Imbrie ice-volume model forced by obliquity.

Walker Circulation Hypothesis

Preceding the mid-Pleistocene transition (starting about 1 Ma ago), the difference between western and eastern equatorial Pacific SSTs increased, enhancing the zonal meridional Walker circulation (see section 'Paleotemperature Changes'), following the long trend initiated in the late Pliocene (Wara et al., 2005). The increase in the Walker circulation was synchronous with a long-term freshening of the western equatorial Pacific. Such a pattern, which nowadays occurs during the La Niña phase of the El Niño Southern Oscillation, is correlated with enhanced freshwater fluxes toward the high latitudes, and also to a likely decrease of the poleward heat flux, based on the global energy balance (Philander and Fedorov, 2003). In this hypothesis, increased accumulation of continental ice was thus induced by low-latitude climate changes which altered poleward freshwater and heat fluxes (de Garidel-Thoron et al., 2005; McClymont and Rosell-Melé, 2005). This hypothesis posits that permanent El Niño conditions, such as in the late Pliocene–early Pleistocene, would inhibit the growth of large ice sheets over North America (review by Chiang, 2009). In line with this hypothesis, Huybers and Molnar (2007) inferred that the effect of warmer eastern Pacific surface waters would exceed the effect of obliquity forcing on the ablation rate of North Hemisphere ice sheets, based on modern observations. Therefore, the establishment of the zonal temperature gradient in the equatorial Pacific during the mid-Pleistocene might have favored the onset of large late Pleistocene glaciations.

Millennial-Scale Dynamics in the Early Pleistocene?

Abrupt climatic changes in the late Pleistocene have been extensively described from global oceanic sedimentary records. In the millennial band, Dansgaard–Oeschger events and Heinrich events paced MIS 3. Though these events were first characterized from sites in the North Atlantic, these abrupt changes have been shown to have remote effects on climate. These events, characterized by massive iceberg discharges in the North Atlantic ocean, correlated with significant sea-surface

cooling at relatively low latitudes, and were eventually linked to changes in global thermohaline circulation. In the early Pleistocene, very few sedimentary records have been studied at millennial resolution (Mc Intyre et al., 2001; Raymo et al., 1998). Among the few existing studies, Raymo et al. (1998) have shown that during the early Pleistocene (MIS 40 and 44), IRD, indicative of iceberg discharges, were present in deep-sea sediments from the North Atlantic and were found to exhibit a clear millennial signal (site ODP983). Though surface Foraminifera fauna did not show a similar pattern of millennial changes, these IRD indicate that 'Dansgaard-Oeschger-like' events occurred during the early Pleistocene glacials when the ice sheets were reduced, compared to the last climatic cycle. Deep-sea sediment records cannot yet characterize the patterns and phasing of global climate changes associated with these millennial events, but there is robust evidence for millennial-scale climate change during early Pleistocene Ice Ages. The origin of millennial-scale oscillations during the early Pleistocene cannot yet be ascribed to any particular process, though it is likely that the same processes at play during the late Pleistocene should also have been critical to abrupt changes in the 41 ka world.

Conclusions

The early Pleistocene offers a unique perspective on how the global climate system evolved from a warmer mode relative to the preindustrial mean state. Paleoclimatographic records from all around the world clearly indicate that the climate system responded linearly to orbital obliquity (41 ka) and precessional (21 ka) forcing during the early Pleistocene, and changed during the mid-Pleistocene to become dominated by 100 ka cycles. However, the exact processes leading to the transmission of the obliquity signal to the global climate (and in particular to the low latitudes) and the evolution of the dominant periodicities remain uncertain. The role of thermohaline circulation in early Pleistocene glacial-interglacial cycles does not seem to be preponderant, though the available data cannot precisely constrain its past changes. Finally, the processes leading to the onset of the largest glaciations of the Cenozoic, after the mid-Pleistocene transition, took place after a long-term cooling of the high latitudes, contrasting with a low-latitude stability in SSTs, and of an increase in the meridional Walker circulation.

See also: **Glacial Climates:** Thermohaline Circulation.

Paleoceanography: Paleoceanography An Overview.

Paleoceanography, Biological Proxies: Alkenone

Paleothermometry Based on the Haptophyte Algae. **Quaternary**

Stratigraphy: Continental Biostratigraphy.

References

- Berger WH, Jansen E, Johannessen O, Muench R, and Overland J (1994) Mid-Pleistocene climate shift: The Nansen connection. *The Polar Oceans and their Role in Shaping the Global Environment: The Nansen Centennial Volume. Geophysical Monograph* 84: pp. 295–311. Washington, DC: AGU.
- Berger A, Li X, and Loutre M-F (1999) Modelling northern hemisphere ice volume over the last 3 Ma. *Quaternary Science Reviews* 18: 1–11.
- Bintanja R, van de Wal RSW, and Oerlemans J (2005) Modelled atmospheric temperatures and global sea levels over the past million years. *Nature* 437: 125–128.
- Chiang JC (2009) The tropics in Paleoclimate. *Annual Review of Earth and Planetary Sciences* 37: 263–297.
- Clark P and Pollard D (1998) Origin of the middle Pleistocene transition by ice sheet erosion of regolith. *Paleoceanography* 13: 1–9.
- de Garidel-Thoron T, Rosenthal Y, Bassinot F, and Beaufort L (2005) Stable sea surface temperatures in the western Pacific warm pool over the past 1.75 million years. *Nature* 433: 294–298.
- DeBlonde G and Peltier W (1991) A one-dimensional model of continental ice volume fluctuations through the Pleistocene: Implications for the origin of the mid-Pleistocene climate transition. *Journal of Climate* 4: 318–344.
- Gradstein F, Ogg J, Smith A (2005) A Geologic Time Scale 2004. Cambridge University Press, Cambridge, 589 p.
- Hönisch B, Hemming N, Archer D, Siddall M, and McManus J (2009) Atmospheric carbon dioxide concentration across the mid-Pleistocene transition. *Science* 324: 1551–1554.
- Huybers P (2006) Early Pleistocene glacial cycles and the integrated summer insolation forcing. *Science* 313: 508–511.
- Huybers P (2009) Pleistocene glacial variability as a chaotic response to obliquity forcing. *Climate of the Past* 5: 481–488.
- Huybers P and Molnar P (2007) Tropical cooling and the onset of North American glaciation. *Climate of the Past* 3: 549–557.
- Huybers P and Wunsch C (2004) A depth-derived Pleistocene age model: Uncertainty estimates, sedimentation variability, and nonlinear climate change. *Paleoceanography* 19: PA1028.
- Huybers P and Wunsch C (2005) Obliquity pacing of the late Pleistocene glacial terminations. *Nature* 434: 491–494.
- Imbrie J, Berger A, Boyle E, et al. (1993) On the structure and origin of major glaciation cycles: 2. The 100,000-year cycle. *Paleoceanography* 8: 699–735.
- Imbrie J, Boyle E, Clemens SC, et al. (1992) On the structure and origin of major glaciation cycles: 1. Linear responses to Milankovitch forcing. *Paleoceanography* 7: 701–738.
- Lisiecki LE and Raymo M (2005) A Pliocene–Pleistocene stack of 57 globally distributed benthic $\delta^{18}\text{O}$ records. *Paleoceanography* 20: 17.
- Liu Z, Cleaveland LC, and Herbert TD (2008) Early onset and origin of 100-kyr cycles in Pleistocene tropical SST records. *Earth and Planetary Science Letters* 265: 703–715.
- Liu Z and Herbert T (2004) High-latitude influence on the eastern equatorial Pacific climate in the early Pleistocene epoch. *Nature* 427: 720–723.
- Lüthi D, Le Floch M, Bereiter B, et al. (2008) High-resolution carbon dioxide concentration record 650,000–800,000 years before present. *Nature* 453: 379–382.
- Marlow J, Lange C, Wefer G, and Rosell-Mele A (2000) Upwelling intensification as part of the Pliocene–Pleistocene climate transition. *Science* 290: 2288–2291.
- Mc Intyre K, Delaney M, and Ravelo A (2001) Millennial-scale climate change and oceanic processes in the late Pliocene and early Pleistocene. *Paleoceanography* 16: 535–543.
- McClymont EL and Rosell-Melé A (2005) Links between the onset of modern Walker circulation and the mid-Pleistocene climate transition. *Geology* 33: 389–392.
- McClymont E, Rosellmele A, Giraudeau J, Pierre C, and Lloyd J (2005) Alkenone and coccolith records of the mid-Pleistocene in the south-east Atlantic: Implications for the index and South African climate. *Quaternary Science Reviews* 24: 1559–1572.
- Medina-Elizalde M and Lea D (2005) The mid-Pleistocene transition in the tropical Pacific. *Science* 310: 1009–1012. <http://dx.doi.org/10.1126/science.1115933>.
- Mudelsee M and Schulz M (1997) The mid-Pleistocene climate transition: Onset of 100 ka cycle lags ice volume build-up by 280 ka. *Earth and Planetary Science Letters* 151: 117–123.
- Philander SG and Fedorov AV (2003) Role of tropics in changing the response to Milankovitch forcing some three million years ago. *Paleoceanography* 18: 1–12.
- Prell W, Gardner J, et al. (1982) Oxygen and carbon isotope stratigraphy for the quaternary of hole 502B: Evidence of two modes of variability. *Initial Reports DSDP*.
- Raymo M, Ganley K, Carter S, Oppo D, and McManus J (1998) High latitude climate instability in the early Pleistocene. *Nature* 392: 699–702.
- Raymo M and Nisancioglu K (2003) The 41 kyr world: Milankovitch's other unsolved mystery. *Paleoceanography* 18: 1011–1015.
- Raymo M, Oppo D, and Curry W (1997) The mid-Pleistocene climate transition: A deep sea carbon isotopic perspective. *Paleoceanography* 12: 546–559.
- Raymo M, Oppo D, Flower B, et al. (2004) Stability of North Atlantic water masses in face of pronounced climate variability during the Pleistocene. *Paleoceanography* 19: PA2008.

- Raymo M, Ruddiman W, Shackleton N, and Oppo D (1990) Evolution of Atlantic-Pacific C13 gradients over the last 2.5 my. *Earth and Planetary Science Letters* 97: 353–368.
- Rio D, Sprovieri R, Castradori D, and Di Stefano E (1998) The Gelasian Stage (Upper Pliocene): A new unit of the global standard chronostratigraphic scale. *Episodes* 21: 82–87.
- Ruddiman W, Raymo M, Martinson D, Clement B, and Backman J (1989) Pleistocene evolution: Northern Hemisphere ice sheets and North Atlantic ocean. *Paleoceanography* 4: 353–412.
- Schefuß E, Sinninghe Damsté J, and Jansen J (2004) Forcing of tropical Atlantic sea surface temperatures during the mid-Pleistocene transition. *Paleoceanography* 19: PA4029.
- Sepulcre S, Vidal L, Tachikawa K, and Rostek F (2011) Sea-surface salinity variations in the northern Caribbean Sea across the mid-Pleistocene transition. *Climate of the Past* 7: 75–90.
- Shackleton N, Opdyke N, Cline R, and Hays J (1976) Oxygen isotope and paleomagnetic stratigraphy of Pacific core V28-239, late Pliocene to latest Pleistocene. In: Cline RM and Hays JD (eds.) *Investigation of Late Quaternary Paleoceanography, and Paleoclimatology*, Geological Society of America Memoir 145: 449–464.
- Sosdian S and Rosenthal Y (2009) Deep-sea temperature and ice volume changes across the Pliocene–Pleistocene climate transitions. *Science* 325: 306–310.
- Tziperman E and Gildor H (2003) On the mid-Pleistocene transition to 100-kyr glacial cycles and the asymmetry between glaciation and deglaciation times. *Paleoceanography* 18: 1001–1008.
- Wara MW, Ravelo A, and Delaney ML (2005) Permanent El Niño-like conditions during the Pliocene warm period. *Science* 309: 758–761.

Relevant Website

<http://www.quaternary.stratigraphy.org.uk/> – Subcommission on Quaternary Stratigraphy (SQS).

Late Pleistocene North Atlantic

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Modern Surface and Deep Circulation of the North Atlantic

The modern circulation of the North Atlantic can be divided into the wind-driven circulation in the upper few 100 m and the meridional overturning circulation that moves water masses formed at the surface through the abyss. Meridional overturning is driven by a combination of buoyancy and dynamic forcing, including processes operating in the Southern Ocean, and exciting research today in oceanography and paleoceanography concerns the processes that control the overturning circulation and thus influence climate on a broad scale.

Surface Circulation

The westerly winds blowing across the North Atlantic at mid-latitudes and the easterly trade winds at lower latitudes drive a clockwise gyre circulation in the North Atlantic (**Figure 1**). The Gulf Stream flows rapidly northward along the eastern coast of North America and northeastward across the North Atlantic, the Canary Current flows slowly southward along the west coast of Africa, and an equatorial current flows westward along the equator. One branch of the surface circulation continues northward in the Labrador, Greenland, and Iceland Seas, and even farther north into the Arctic. Cold Arctic water flows southward along the east coast of Greenland, around the southern tip of Greenland, and forms a small gyre in the Labrador Sea (not shown).

Deep Circulation and the Meridional Overturning Circulation

Several processes, including heat loss to the atmosphere and mixing with cold Arctic water, increase the density of surface waters in the region of the Greenland, Iceland, and Norwegian Seas (the Nordic Seas) and in the Labrador Sea, so that the surface waters convect (sink), forming the North Atlantic Deepwater (NADW) that flows southward along the western boundary between 2000 and 4000 m in the North Atlantic (**Figure 1**, blue arrows). The deep branch of the meridional overturning circulation is much slower than the surface circulation, making a circuit over centuries rather than decades, and the current speeds are not easily directly measured. The deep circulation can be traced from conservative physical properties such as temperature and salinity that are acquired at the surface and subsequently changed only by mixing with other water masses (**Figure 2**). NADW forms in the Nordic and Labrador Seas and flows southward into the South Atlantic, observed as a maximum in salinity (**Figure 2**). In the South Atlantic, the Antarctic Intermediate Water, identified by low salinity, flows northward above NADW. Antarctic Bottom Water originates near Antarctica and flows northward across the equator in the deep Atlantic below NADW. The combination of intermediate salinity combined with very low temperature makes Antarctic Bottom Water the most dense water mass in this simplified view of the deep circulation.

The conveyor belt is another term given to the combined surface and deep circulation that carries surface water northward in the Gulf Stream, where it sinks in the Nordic Seas and flows out of the North Atlantic and into the other ocean basins. The surface return flow to the North Atlantic is principally from the Pacific via the Indian Ocean. The conveyor analogy describes the ideas that the circulation forms a complete loop, and that the motion transports heat, salt, and nutrients throughout the ocean basins. Model-based studies indicate that the strength of the overturning circulation depends partly upon surface water in the Nordic Seas being dense enough to convect. This requirement adds a nonlinear, threshold process to the Atlantic circulation. If ice sheet melting or precipitation or runoff were to freshen the Nordic Seas so that surface waters became too buoyant, the overturning circulation might be reduced or stop completely, abruptly when the threshold is crossed. Conversely, when the surface waters become dense enough to sink, the overturning circulation might resume abruptly. The late Pleistocene paleoceanographic records discussed later provide concrete evidence for on/off behavior of the thermohaline circulation during the late Pleistocene.

Wind-driven and deep circulations are not independent. The Gulf Stream can be considered the upper limb of the overturning circulation and is dynamically driven by the surface wind field. The atmospheric circulation influences not only the winds that drive the surface circulation but also the deep ocean circulation by creating salty, dense waters through excess evaporation in the subtropical region, by producing cold, freshwater in the Arctic region through excess precipitation and runoff, and heat loss, and by creating a region of high seasonality and deep winter mixing beneath the westerly winds in the North Atlantic.

Evolution During The Pleistocene

The stacked benthic $\delta^{18}\text{O}$ composited from 57 sites reflects a combination of ice volume change and deep ocean temperature change, and reveals a gradual evolution throughout the Pleistocene (Lisiecki and Raymo, 2005; **Figure 3**). The average temperature of the Earth cooled, likely dominated by cooling at high latitudes with little change at low latitudes. The volume of the ice sheets increased. A steplike increase in the concentration of ice-rafted debris (IRD) signaled the expansion of large-scale Northern Hemisphere ice sheets 2.74 Ma (**Figure 3**; Jansen et al., 2000). Expansion of the Northern Hemisphere ice sheets around 2.7 Ma has also been interpreted from an increase in the concentration of long-chain n-alkanes in North Atlantic sediments (Naafs et al., 2012). The development of glacial outwash plains would be expected to increase the supply of aeolian sediment to the North Atlantic, including n-alkanes derived from higher plant waxes (Naafs et al., 2012). Sea surface temperature (SST) in the North Atlantic (Site 607, 41°N) cooled throughout the Pleistocene, with two steps

occurring in the late Pliocene (3.1–2.4 Ma) and the mid-Pleistocene transition (1.5–0.8 Ma; Lawrence et al., 2010). The most remarkable aspect of this evolution was that it occurred in the absence of a change in orbital forcing, implicating a fundamental change internal to the climate system (Clark et al., 2006). Many hypotheses for the origin of this evolution focus on the mid-Pleistocene transition, when the variance in the benthic oxygen isotope record shifted from a dominantly 41 000-year period to a dominantly 100 000-year period. Why ice volume varies at the 100 000-year period when

there is so little variation in insolation at this periodicity has been termed the 100-ka cycle problem, and the unexpectedly small variance in the precessional band (23 000 years) prior to the transition has been attributed to a shift in the relative proportions of ice volume in the Northern and Southern Hemispheres (Raymo et al., 2006).

Mid-Pleistocene Transition

Many aspects of the climate changed across the mid-Pleistocene transition between 1.25 and 0.7 Ma (Clark et al., 2006). The average $\delta^{18}\text{O}$ increased, indicating deep ocean cooling and greater ice volume, the variance increased (Figure 3), and the dominant period of variation shifted from the 41 000-year obliquity cycle to the 100 000-year period. The increase in variance is attributed to a combination of deep ocean cooling and increase in ice volume; the North Atlantic sea surface also cooled (Figure 3). Upwelling increased in tropical ocean regions and African and Asian monsoon intensity and aridity increased, summarized by Clark et al., 2006. Changes in the meridional overturning are difficult to reconstruct. Vertical and between-ocean gradients in benthic foraminifer $\delta^{13}\text{C}$ from an array of sites suggest that the coupling between overturning and glaciation has changed through time and across the mid-Pleistocene transition; however, the multiple influences on $\delta^{13}\text{C}$ complicate this interpretation.

Glacial–Interglacial Cycles

The glacial–interglacial cycles of the late Pleistocene North Atlantic are one of the best-studied aspects of the Pleistocene. SSTs have been reconstructed from many different proxies, and surface and deep circulations inferred from tracers such as $\delta^{13}\text{C}$. The following discussion focuses on the changes that occurred during the Last Glacial Maximum (LGM), recognizing that previous glacial stages share many of the same characteristics at least for the last five glacial cycles.

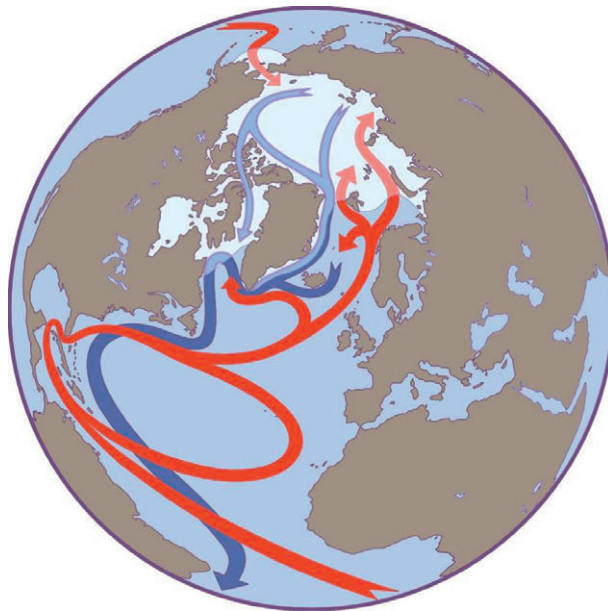


Figure 1 Simplified view of the North Atlantic circulation shows warm water flowing clockwise around the North Atlantic gyre (red loop), with a branch flowing northward to the Labrador and Greenland/Iceland Seas (red arrows), where it cools, sinks, and flows southward along the western boundary of the deep Atlantic (blue arrows). Reproduced from Holloway G and Proshutinsky A (2007) Role of tides in arctic ocean/ice climate. *Journal of Geophysical Research* 112: C04S06.

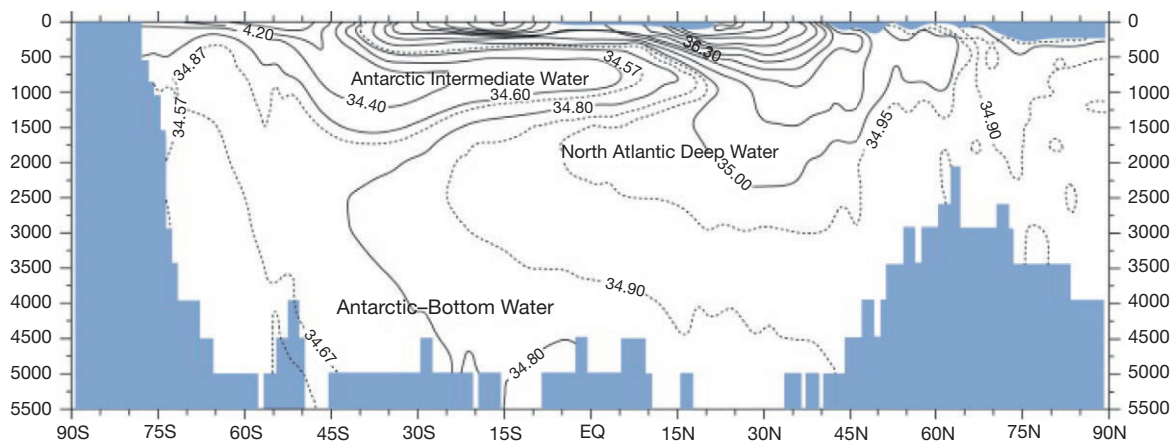


Figure 2 Salinity (psu) as a function of depth and latitude in the Atlantic. The principal water masses can be identified and traced back to their origin at the high-latitude sea surface by their characteristic salinity and temperature characteristics. Reproduced from Conkright ME, Locarnini RA, Garcia HE, et al. (2002) *World Ocean Atlas 2001: Objective Analyses, Data Statistics, and Figures, CD-ROM Documentation*. Silver Spring, MD: National Oceanographic Data Center; from the World Ocean Atlas.

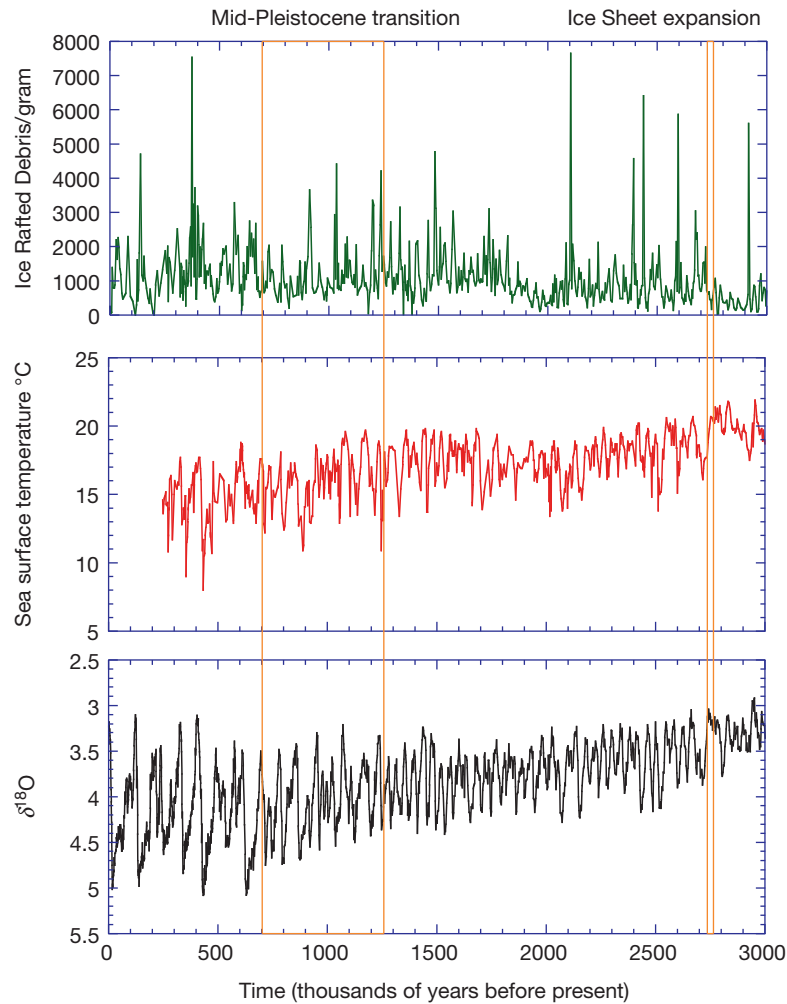


Figure 3 North Atlantic ice-rafted debris (IRD) (Jansen et al., 2000), sea-surface temperature (SST) (Site 607; Lawrence et al., 2010), and the global benthic oxygen isotope stacked record (Lisiecki and Raymo, 2005). Ice sheet expansion refers to expansion of large-scale Northern Hemisphere ice sheets identified on the basis of IRD, and the mid-Pleistocene transition refers to the period of increasing variance in $\delta^{18}\text{O}$, culminating in the dominance of the 100 000-year cycle after 700 000 years BP.

Climate Sensitivity of the North Atlantic

The most dramatic aspect of the glacial Atlantic is the large difference in SST with respect to the present. Whether one examines the pioneering maps created by Climate: Long-Range Investigation, Mapping, and Prediction (CLIMAP; CLIMAP Project Members, 1981) or the Glacial Atlantic Ocean Mapping (GLAMAP; Pflaumann et al., 2003), or Multiproxy Approach for the Reconstruction of the Glacial Ocean surface (MARGO) compilation (Waelbroeck et al., 2009), one observes that the middle-to-high latitudes cooled more than lower latitudes, and also that the North Atlantic sea surface cooled much more than other oceans at similar latitudes (Figure 4). According to the CLIMAP and GLAMAP reconstructions, the region between 45° and 60° N cooled by more than 10 °C (Figure 4). North of 60° N, the difference from today is not as large because seawater cannot cool below about −1.5 °C before sea ice forms. Here sea ice extended farther south. In contrast, mid-latitudes in other ocean basins cooled by only a few degrees. In the subtropical gyres in all oceans, the sea

surface cooling was smallest, and in the CLIMAP reconstruction, several of the gyres were warmer than today. Along the equator in the Atlantic, the cooling was small but larger than in the subtropics. Cooling along the equator was most pronounced at the eastern margin, and has been attributed to stronger coastal and equatorial upwelling driven by stronger trade winds (Waelbroeck et al., 2009).

Today, the sea ice margin migrates seasonally in the high North Atlantic, extending in winter across the Nordic Seas north of Fram Strait (80° N) and southward along the east coast of Greenland, and retreating northward each summer. During glacial times as the sea surface cooled, the southern sea ice margin moved toward the equator. The southern extent of sea ice and its seasonal cycle has always been poorly constrained, its boundary often inferred only from reduced sediment accumulation. Early CLIMAP reconstructions positioned the summer and winter sea ice margin far south of its present latitude, with perennial sea ice cover across the Nordic Seas. Later reconstructions (Pflaumann et al., 2003) based on foraminifer faunal temperature reconstructions have indicated that

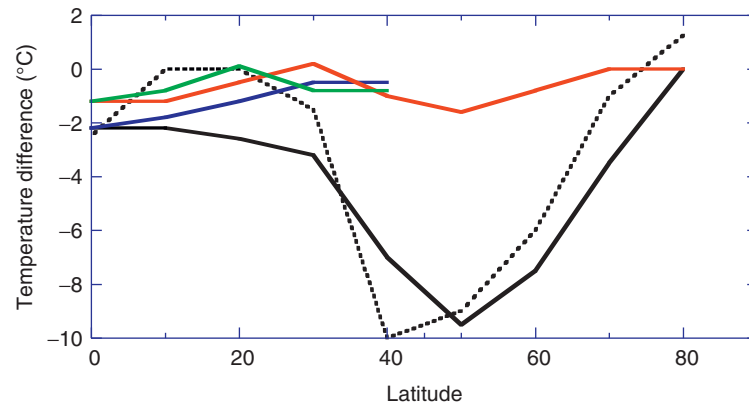


Figure 4 Change in zonal average sea surface temperature between glacial and present (negative indicates glacial was cooler than today). North Atlantic from CLIMAP (black line), North Atlantic from GLAMAP (black dashed), North Indian-Pacific from CLIMAP (red), South Atlantic (blue), South Indian-Pacific (green) (Pflaumann, 1985; CLIMAP Project Members, 1981; Pflaumann et al., 2003).

the winter sea ice margin was not so extensive, with an ice-free channel extending in winter as far north as 50° or 60°, and that the summer sea ice margin was much less extensive than originally determined, extending across the Arctic and Fram Strait. The revised summer sea ice margin is not much different than today, and leaves the Nordic Seas ice-free during glacial summers. Until unambiguous indicators of sea ice for each season are found, reconstructing the sea ice margin will remain an active area of research, highly significant in understanding possible sites of deepwater formation during glacial times, the albedo of the northern North Atlantic, and the receipt of solar radiation and the exchange of heat between ocean and atmosphere.

Meridional Overturning

Changes in the production of NADW, and the vertical gradients in water mass properties have been inferred from many proxies, including $\delta^{13}\text{C}$, Cd/Ca , $^{231}\text{Pa}/^{230}\text{Th}$, and $^{143}\text{Nd}/^{144}\text{Nd}$. NADW is depleted in nutrients (high $\delta^{13}\text{C}$ and low Cd/Ca) and can be distinguished from Antarctic bottom water (low $\delta^{13}\text{C}$ and high Cd/Ca). One of the most well-documented discoveries in the North Atlantic is the change in $\delta^{13}\text{C}$ between modern and glacial times (Figure 5), inferred to represent a decrease in NADW production during glacial times, and its replacement by a water mass that formed farther south and descended to a shallower depth (Glacial North Atlantic Intermediate Water; Lynch-Stieglitz et al., 2007). Results based on Cd/Ca are consistent with the section developed from $\delta^{13}\text{C}$ (Figure 5, bottom panel). One additional approach, based on U-series decay products (^{231}Pa and ^{230}Th), provides the only measure of the rate of circulation. The export of ^{231}Pa relative to ^{230}Th indicates no significant difference at the LGM relative to the present (Yu et al., 1996); however, abrupt changes during the deglaciation are indicated at some sites (McManus et al., 2004; see Section 'The Last Deglaciation').

One of the critical questions has been the extent to which NADW production drives, or responds to, glacial-interglacial climate change in the North Atlantic region. Studying different proxies found in South Atlantic cores, Piotrowski et al. (2005)

show that changes in benthic foraminifer $\delta^{13}\text{C}$ and changes in ϵ_{Nd} , both reflecting NADW formation, followed the ice growth (inferred from benthic $\delta^{18}\text{O}$) during the transition from the last interglacial interval (marine isotope stage (MIS) 5a) to the next glacial interval (MIS 4), decreasing NADW production, and later followed the initial warming (inferred from Greenland and Antarctic ice cores) at the end of the last glaciation (MIS 2), increasing NADW production). The results are consistent with the hypothesis that NADW production changes are an amplifier, but not an initial trigger, of the late Pleistocene Ice Ages.

Millennial-Scale Variability

Millennial-scale variability is discussed in detail in a separate article, and its inclusion here seeks to highlight North Atlantic paleoceanographic records and their significance. Any discussion of marine records of millennial-scale variability must include the caveat that millennial-scale events are not often preserved in marine sediments. Deep-sea sedimentation rates are typically 1–5 cm ka^{-1} , and bioturbation (mixing of sediment by benthic organisms at the seafloor) can homogenize sediment across 10 cm, attenuating the amplitude, and sometimes completely erasing events of 1000-year (1–3 cm) duration. The discovery of millennial-scale events in ice cores motivated a vigorous effort to find and analyze higher sedimentation rate (>10 or >20 cm ka^{-1}), or minimally bioturbated cores. Since then, millennial-scale variability has been reported in cores from the North Atlantic, and sites as far away as the Cariaco Basin, the California Margin, the Arabian Sea, and the South China Sea. The magnitude of SST change is greatest in the North Atlantic, consistent with the amplitude of glacial-interglacial change. While the following discussion examines the millennial-scale variability during the last glacial, from 120,000 to 10,000 years BP (Figure 6), longer records show that millennial-scale variability was a persistent feature of at least the preceding four glacial intervals (the last 500,000 years), prevalent whenever benthic $\delta^{18}\text{O}$ exceeded 3.5 per mil (McManus et al., 1999). These oscillations and events are

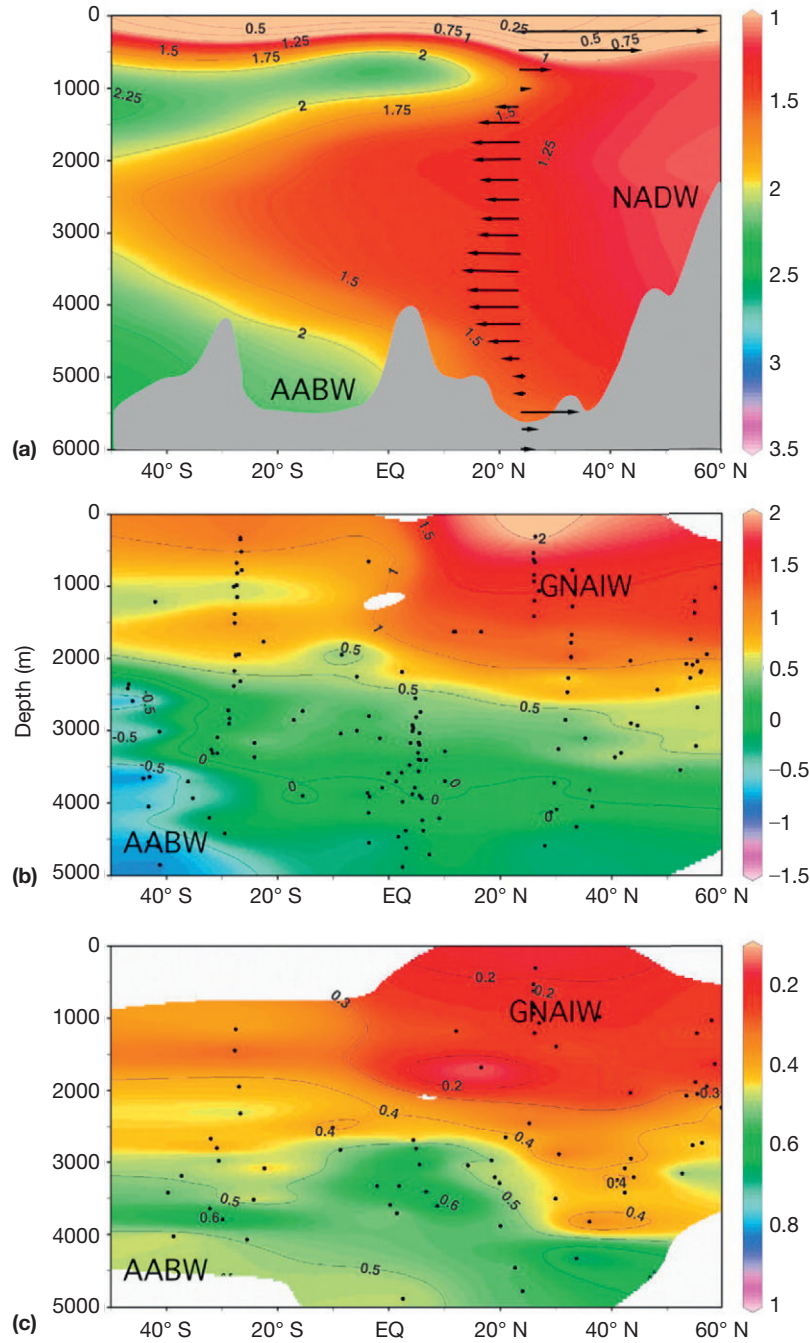


Figure 5 (a) The modern distribution of dissolved phosphate ($\mu\text{mol l}^{-1}$) – a biological nutrient – in the western Atlantic (Conkright et al., 2002). Also indicated is the southward flow of NADW, which is compensated by the northward flow of warmer waters above 1 km, and the Antarctic Bottom Water (AABW) below. (b) The distribution of the carbon isotopic composition ($^{13}\text{C}/^{12}\text{C}$, expressed as $\delta^{13}\text{C}$, Vienna Pee Dee belemnite standard) of the shells of benthic Foraminifera in the western and central Atlantic during the LGM (Bickert and Mackensen, 2004; Curry and Oppo, 2005). Data from different longitudes are collapsed in the same meridional plane. (c) Estimates of the Cd (nmol kg^{-1}) concentration for LGM from the ratio of Cd/Ca in the shells of benthic Foraminifera (Marchitto and Broecker, 2006). Today, the isotopic composition of dissolved inorganic carbon and the concentration of dissolved Cd in seawater both show “nutrient”-type distributions similar to that of PO_4 .

absent or much reduced during the postglacial period. The absence of this variability during interglacial times, the abrupt, sawtooth character of the cycles, and the long time between events (thousands of years) have led to many explanations that involve the interaction between ice sheet growth and decay, sea

ice extent, and the surface and deep circulations in the North Atlantic.

Two different types of millennial-scale variability are observed during the last glacial cycle in the North Atlantic. Dansgaard–Oeschger events describe a series of warm (interstadial)

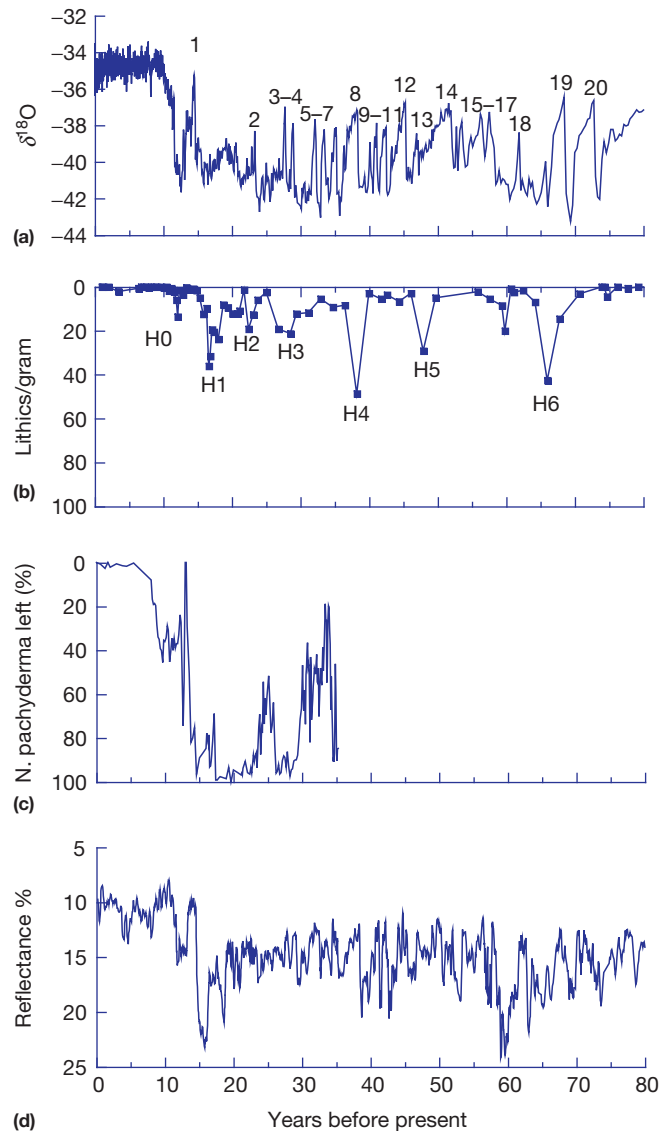


Figure 6 Millennial-scale variability in the North Atlantic during the late Pleistocene. (a) GISP2 oxygen isotope record with interstadial (warm) Dansgaard–Oeschger events labeled; (b) concentration of ice-rafted lithic fragments at Site 980 (H0–H6 are Heinrich layers); (c) percentages of *Neogloboquadrina pachyderma* left coiling in North Atlantic core; sediment reflectance in Cariaco Basin Site 1008. Data obtained from the World Data Center for Paleoclimatology <http://www.ncdc.noaa.gov/paleo/> from; Bond et al. (1993); Grootes and Stuiver (1997); McManus et al. (1999); Peterson et al. (2000).

and cool (stadial) events that are recorded in the oxygen isotopes of Greenland ice cores. Twenty-five events are numbered during the last glacial period, each lasting from 600 to 2000 years, and each is characterized by progressive cooling followed by abrupt warming (Figure 6(a)). At Site 980 in the North Atlantic, D–O events are evident as shifts in foraminifer abundance (Figure 6(c)), high ratios of *Neogloboquadrina pachyderma* left indicating cooler SSTs (Bond et al., 1993), and also in alkenone-based SST reconstructions of the Bermuda Rise at 35°N in the North Atlantic (Sachs and Lehman, 1999). Farther south in the Cariaco Basin, millennial-scale events appear in time series of gray scale in the Cariaco Basin (a proxy for plankton productivity and basin anoxia), indicating that at least the basin-scale surface circulation is involved.

Compared to the ice cores, the cycles found in the marine sediment records are not as pronounced; however, bioturbation complicates the effort to map the geographic extent of D–O variability throughout the Atlantic.

Heinrich events are other millennial-scale events that occurred during the last glacial period. Heinrich events were initially defined on the basis of coarse sediments found in discrete layers in North Atlantic sediments, and represent times when coarse ice-rafted material was transported farther south during massive iceberg discharge events from the Laurentide and other smaller ice sheets. Six events have been widely identified in North Atlantic sediments (H1, 14 300; H2, 21 000; H3, 27 000; H4, 35 500; H5, 52 000; and H6, 69 000 years BP), each occurring at the coldest part of a D–O cycle

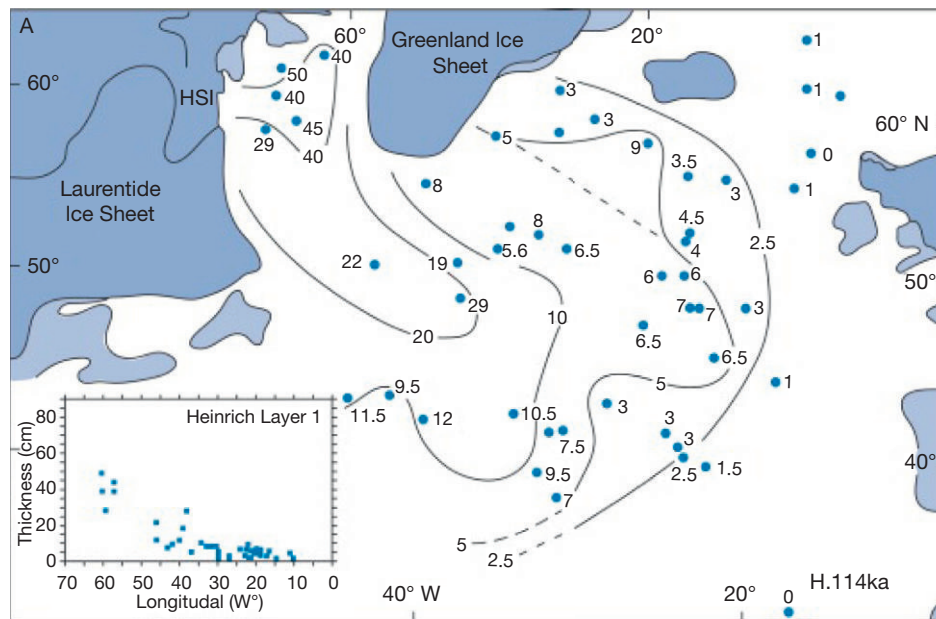


Figure 7 Thickness (cm) of Heinrich layer 1 based on whole-core magnetic susceptibility; inset shows the thickness as a function of longitude. Reproduced from Dowdeswell JA, Maslin M, Andrews JA, and McCave IN (1995) Iceberg production, debris rafting, and the extent and thickness of Heinrich layers (H-1, H-2) in North Atlantic sediments. *Geology* 23: 301–304. <http://geology.gsapubs.org/>. © Geological Society of America.

(Figure 6). Most of the melting of icebergs occurs along the Polar Front, and the southern edge of the IRD belt therefore reveals that the Polar Front extended farther south during H events (Figure 7). Subsequent studies have also shown that the H events are characterized by colder SST in addition to extensive ice rafting, and by a freshwater anomaly farther north indicated by $\delta^{18}\text{O}$.

The H events are also characterized by changes in NADW production, in the sense that the cold stadial events were accompanied by reduced NADW production, possibly reduced production of Glacial North Atlantic Intermediate Water, and a relative increase in the proportion of Antarctic Bottom Water and Antarctic Intermediate Water at Atlantic sites. NADW production appears most reduced during H events. One sequence of events that emphasizes the instability of ice sheets in driving millennial-scale variability considers a gradual accumulation of ice over thousands of years, accompanied by decreased NADW production that further cools the North Atlantic toward the stadial state. Instability in the ice sheet eventually leads to abrupt, catastrophic release of ice into the North Atlantic, which freshens the North Atlantic upon melting and stops the production of NADW. Following ice sheet collapse, freshwater input to the North Atlantic is reduced, salinity increases again, and NADW production resumes, rapidly warming the North Atlantic. Problems exist even with this simplified sequence of events, and alternate models have been proposed that attempt to fit all the observations.

The Last Deglaciation

The sequence of events that occurred in the North Atlantic during the last deglaciation provides important clues regarding cause and effect, and the interaction between ice sheets,

atmosphere, and ocean circulation. Uncertainties in radiocarbon dating (± 100 years for marine sediments) continue to make correlation among records difficult, and even the strategy to analyze different proxies from a single core to determine leads and lags among proxies often has uncertainty (± 100 or more years). The timing and changing routes of meltwater discharge from the Arctic, and from the Laurentide and Eurasian Ice Sheets remain poorly constrained. Therefore, while the general sequence is becoming increasingly well known, significant uncertainty surrounds leads and lags and the timing of abrupt climate changes that occurred during the last deglaciation.

The oxygen isotope record from the GISP2, reflecting primarily the surface air temperature over Greenland, provides a useful framework to identify many of the events that occurred during the deglaciation (Figure 8). Readily apparent is the cooling that occurred during the LGM with respect to the Holocene, the initial warming during the Bølling/Allerød (B/A), the return to colder conditions during the Younger Dryas (YD), and the final warming that led to the postglacial North Atlantic climate (Holocene). The record of SST from the subpolar North Atlantic (core SU8118 off Portugal) follows a very similar history with respect to both the timing and amplitude of events (Figure 8(b)). The planktonic oxygen isotope record (Figure 8(c)) from the subtropical North Atlantic (GGC5, Bermuda Rise) reflects both the melting of the ice sheets, about 1.1 per mil, and a component related to warming of the sea surface (also about 1 per mil, equal to about 4 °C). Based on these records, during the deglaciation the subpolar North Atlantic warmed by 12 °C, and the subtropical North Atlantic warmed by about 4 °C, both increases interrupted by a cooling of a few degrees during the YD. The sedimentary $^{231}\text{Pa}/^{230}\text{Th}$ measured on the same core (GGC5) shows that NADW production was reduced compared to the Holocene at

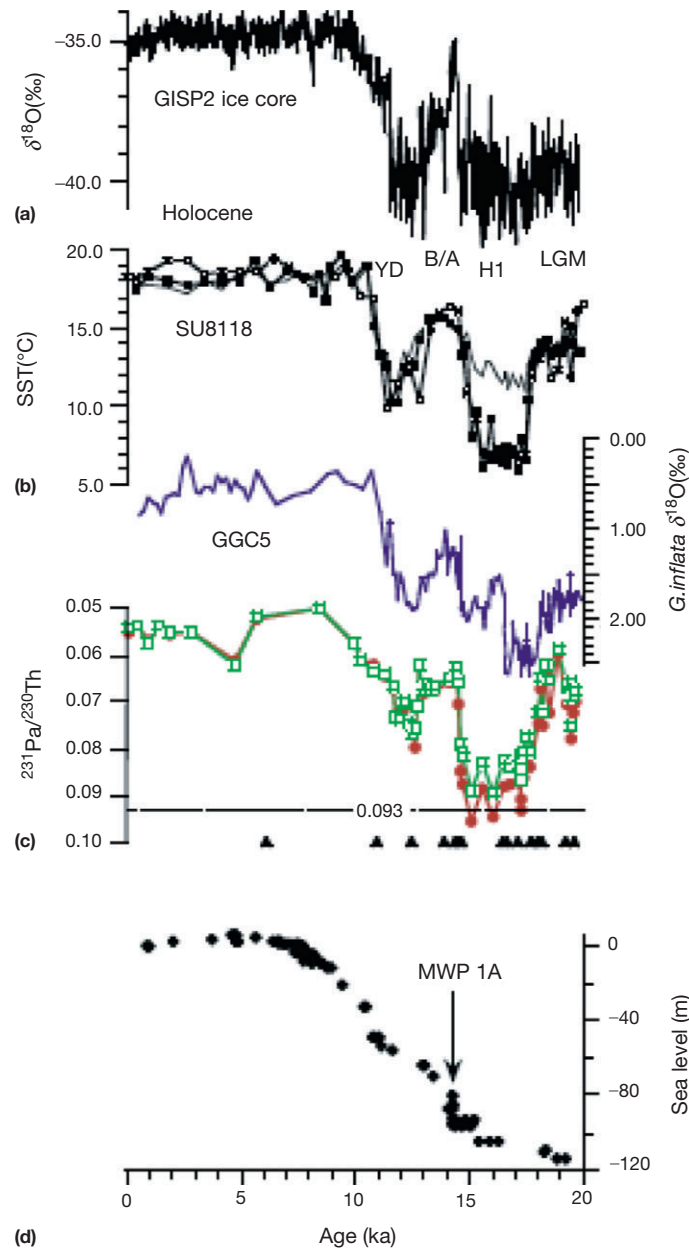


Figure 8 Sequence of events during the last deglaciation in the North Atlantic. Reproduced from [McManus JF, Francois R, Gherardi J-M, Keigwin LD, and Brown-Leger S \(2004\)](#) Collapse and rapid resumption of Atlantic meridional circulation linked to deglacial climate changes. *Nature* 428: 834–837. Events labeled include the Last Glacial Maximum (LGM), Heinrich event 1 (H1), the Bølling/Allerød (B/A), the Younger Dryas (YD), and meltwater pulse 1A (MWP 1A). (a) GISP2 oxygen isotope record, (b) sea-surface temperature (SST) estimates from the subpolar North Atlantic based on planktonic foraminifer assemblages in core SU8118, (c) three-point running mean of planktonic $\delta^{18}\text{O}$ and sedimentary $^{231}\text{Pa}/^{230}\text{Th}$ from core GGC5 (decreases in this ratio indicate greater NADW production, and increases in the ratio indicated reduced NADW production; the ratio 0.093 indicates no differential removal of the isotopes by the deep circulation), and (d) sea level rise determined from Barbados corals.

the LGM, ceased toward the end of the LGM during H1 (reaching the production ratio of 0.093), increased during the B/A, and decreased during the YD before increasing toward Holocene values ([Figure 8\(c\)](#)). A peak in the concentration of IRD occurred in cores throughout the region during H1 ([Figure 7](#)), and other evidence has been compiled indicating that the sea surface cooled and freshened, changes that could have led to reduced NADW production as a result of ice melting ([Sarthein et al., 1992](#)). Reconstructing the melt history of individual ice

sheets remains difficult. The record of sea level determined from Barbados corals, integrating melting from all the ice sheets ([Figure 8\(d\)](#)) shows that sea level was lowest 18 000 years ago, and increased most rapidly during two intervals (meltwater pulse 1A (MWP-1A), [Figure 8\(d\)](#); and MWP-1B, not labeled). These meltwater pulses have been suggested as the cause of thermohaline circulation change during the deglaciation; however, this interpretation remains controversial, and difficult to evaluate because of dating uncertainty.

Based on the comparison between summer insolation at 65° N and the records of early ice melting in the Nordic Seas, a series of papers examined the role of the North Atlantic freshening in driving the deglaciation through changing NADW production. This hypothesis has been challenged by data from the tropics. By comparing the $\delta^{18}\text{O}$ signal in planktic foraminifers with the record of tropical SST based on Mg/Ca from the same samples, it is apparent that the increase in tropical SST leads the melting of the ice sheets (Visser et al., 2003). When the ice core record of atmospheric carbon dioxide is correlated to these records, tropical (and Antarctic) warming and atmospheric carbon dioxide increase are shown to be in phase, both occurring before the $\delta^{18}\text{O}$ depletion during the last deglaciation (Visser et al., 2003). Comparing proxies for NADW production based on $\delta^{13}\text{C}$ and ϵ_{Nd} , Piotrowski et al. (2005) show that the resumption of NADW during the deglaciation followed the warming observed in ice core records. Thus, while ice melt and NADW production may be linked, they appear to follow the warming in the tropics and the increase in carbon dioxide in the atmosphere. In this interpretation, ice sheet melting in the North Atlantic does not initiate the deglacial sequence; however, changes in NADW, sea ice, and land and ocean albedo may be powerful amplifiers that lead to the high sensitivity of climate in the North Atlantic.

See also: **Paleoceanography, Records:** Late Pleistocene North Pacific; Late Pleistocene South Atlantic.

References

- Bickert T and Mackensen A (2004) Late glacial to Holocene changes in South Atlantic deep water circulation. In: Wefer G, Mulitza S, and Ratmeyer V (eds.) *The South Atlantic in the Late Quaternary: Reconstruction of Material Budget and Current Systems*. Berlin: Springer.
- Bond G, Broecker W, Johnsen S, et al. (1993) Correlations between climate records from North Atlantic sediments and Greenland ice. *Nature* 365: 143–147.
- Clark PU, Archer D, Pollard D, et al. (2006) The middle Pleistocene transition: Characteristics, mechanisms, and implications for long-term changes in atmospheric pCO_2 . *Quaternary Science Reviews* 25: 3150–3184.
- CLIMAP Project Members (1981) Seasonal reconstruction of the Earth's surface at the last glacial maximum. *Geological Society of America Map and Chart Series* 36: 1–18.
- Conkright ME, Locarnini RA, Garcia HE, et al. (2002) *World Ocean Atlas 2001: Objective Analyses, Data Statistics, and Figures, CD-ROM Documentation*. Silver Spring, MD: National Oceanographic Data Center.
- Curry WB and Oppo DW (2005) Glacial water mass geometry and the distribution of $\delta^{13}\text{C}$ of ΣCO_2 in the western Atlantic Ocean. *Paleoceanography* 20: PA1017.
- Dowdeswell JA, Maslin M, Andrews JA, and McCave IN (1995) Iceberg production, debris rafting, and the extent and thickness of Heinrich layers (H-1, H-2) in North Atlantic sediments. *Geology* 23: 301–304.
- Groote PM and Stuiver M (1997) Oxygen 18/16 variability in Greenland snow and ice with 10^3 to 10^5 -year time resolution. *Journal of Geophysical Research* 102: 26455–26470.
- Holloway G and Proshutinsky A (2007) Role of tides in Arctic ocean/ice climate. *Journal of Geophysical Research* 112: C04S06.
- Jansen E, Fronval T, Rack F, and Channell JET (2000) Pliocene–Pleistocene ice rafting history and cyclicity in the Nordic Seas during the last 3.5 Myr. *Paleoceanography* 16: 709–721.
- Lawrence KT, Sosdian S, White HE, and Rosenthal Y (2010) North Atlantic climate evolution through the Plio–Pleistocene climate transitions. *Earth and Planetary Science Letters* 300: 329–342.
- Lisiecki LE and Raymo ME (2005) A Pliocene–Pleistocene stack of 57 globally distributed Benthic $\delta^{18}\text{O}$ records. *Paleoceanography* 20: PA1003.
- Lynch-Stieglitz J, Adkins JF, Curry WB, et al. (2007) Atlantic meridional overturning circulation during the last glacial maximum. *Science* 316: 66–69.
- Marchitto TM and Broecker WS (2006) Deep water mass geometry in the glacial Atlantic Ocean: A review of constraints from the paleonutrient proxy Cd/Ca. *Geochemistry, Geophysics, Geosystems* 7: Q12003.
- McManus JF, Francois R, Gherardi J-M, Keigwin LD, and Brown-Leger S (2004) Collapse and rapid resumption of Atlantic meridional circulation linked to deglacial climate changes. *Nature* 428: 834–837.
- McManus JF, Oppo DW, and Cullen JL (1999) A 0.5 million year record of millennial-scale climate variability in the North Atlantic. *Science* 283: 971–975.
- Naafs BDA, Heffer J, Acton G, et al. (2012) Strengthening of North American dust sources during the Late Pliocene (2.7 Ma). *Earth and Planetary Science Letters* 317–318: 8–19.
- Peterson LC, Haug GH, Hughen KA, and Rohl U (2000) Rapid changes in the hydrologic cycle of the tropical Atlantic during the last glacial. *Science* 290: 1947–1951.
- Pflaumann U (1985) Transfer-function '134/6' – A new approach to estimate sea-surface temperatures and salinities of the eastern North Atlantic from the planktonic foraminifers in the sediment. *Meteor Forschungsergeb. Reihe C* 39: 37–71.
- Pflaumann U, Sarinthein M, Chapman M, et al. (2003) Glacial North Atlantic: Surface conditions reconstructed by GLAMAP 2000. *Paleoceanography* 18: 1065.
- Piotrowski AM, Goldstein SL, Hemming SR, and Fairbanks RG (2005) Temporal relationships of carbon cycling and ocean circulation at glacial boundaries. *Science* 307: 1933–1938.
- Raymo ME, Lisiecki LE, and Nisancioglu KH (2006) Plio–Pleistocene ice volume, antarctic climate, and the global $\delta^{18}\text{O}$ record. *Science* 313: 492–495.
- Sachs JP and Lehman SJ (1999) Subtropical North Atlantic temperatures 60,000 to 30,000 years ago. *Science* 286: 756–759.
- Sarinthein M, Jansen E, Arnold M, et al. (1992) $\delta^{18}\text{O}$ time-slice reconstruction of meltwater anomalies at termination I in the North Atlantic between 50 and 80°N. In: Bard E and Broecker WS (eds.) *The Last Deglaciation: Absolute and Radiocarbon Chronologies*. Berlin: Springer.
- Visser K, Thunell RC, and Stott L (2003) Magnitude and timing of temperature change in the Indo-Pacific warm pool during deglaciation. *Nature* 421: 152–155.
- Waelbroeck C, Paul A, Kucera M, et al. (2009) Constraints on the magnitude and patterns of ocean cooling at the last glacial maximum. *Nature Geoscience* 2: 127–132.
- Yu E-F, Bacon MP, and Francois R (1996) Similar rates of modern and last glacial ocean thermohaline circulation inferred from radiochemical data. *Nature* 379: 689–694.

Late Pleistocene South Atlantic

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Introduction

The South Atlantic Ocean receives inflows from the Pacific, the Indian and the Southern Ocean and plays a significant role in the coupling of oceanic processes between these basins and the North Atlantic. An important feature of the present-day circulation in the South Atlantic is the cross-equatorial flow of warm upper layer waters into the North Atlantic, which are imported to compensate for the southward flow of North Atlantic Deep Water (NADW). The associated heat transport into the North Atlantic has a considerable effect on the modern climate (e.g., Wefer et al. (2004)). Processes in the South Atlantic can affect this heat transport and thus can have consequences for climates elsewhere.

The current systems in the South Atlantic are also critical for the nutrient budget of the ocean. The frontal region between the South Atlantic and the Southern Ocean is the ventilation area for South Atlantic central and intermediate water masses that spread northward into the Atlantic. Since these water masses are derived from nutrient-rich Antarctic surface waters, they are the main source of nutrients to sustain the productivity in the eastern boundary upwelling regions of the tropical and subtropical Atlantic.

The knowledge of past variations of the Atlantic circulation and how these variations led to changes in sea-surface temperature (SST), productivity, and atmospheric circulation are critical to our understanding of the role of the South Atlantic in long-term and abrupt climate change. In recent decades, these issues have been addressed by investigating late Pleistocene sediments in the South Atlantic.

Present-Day Circulation of the South Atlantic

The dominant circulation system of the South Atlantic Ocean is the anticyclonic Subtropical Gyre (Figure 1; see Stramma and England (1999) and Wefer et al. (1996) and references therein). The gyre is driven by the general atmospheric circulation, which is dominated by SE trade winds. The eastern part of the Subtropical Gyre, the Benguela Current, has two distinct branches: the Benguela Oceanic Current (BOC) and the Benguela Coastal Current (BCC). The BOC includes thermocline as well as intermediate waters. At the surface, it is replenished from two sources. There is a supply of Indian Ocean water which originates in the Agulhas Current region and slips into the South Atlantic Ocean around the Cape of Good Hope. Because Indian Ocean water is relatively warm, this source is called the 'warm water route.' The remaining water is subantarctic surface water added at the Subtropical Front, or intermediate water that enters the South Atlantic Ocean through Drake Passage, upwells into the thermocline, warms up and flows eastward with the South Atlantic Current.

This second source largely corresponds to the 'cold water route.' The northward flowing surface waters become warmer and, because of the high evaporation in the subtropics, they become saltier. In the western equatorial South Atlantic a fraction of the warm water flows southward with the Brazil Current to close the subtropical gyre. The remaining fraction crosses the equator and feeds the Gulf Stream and the North Atlantic Drift to reach the subpolar North Atlantic Ocean. Here these waters release their heat to the atmosphere and cool. Because of their high salinity, they become dense enough to sink to great depth in the Labrador and Nordic Seas and thus form NADW that eventually flows back into the South Atlantic. In the deep South Atlantic, NADW interacts with the northward flowing Circumpolar Deep Water (CDW). At about 45° S, the NADW divides the CDW into an upper Circumpolar Deep Water (UCDW) and lower Circumpolar Deep Water (LCDW) branch. The relatively warm and saline NADW occupies the depth interval between 1500 and 4000 m, while below 4000 m LCDW is encountered. The mixing zone between NADW and LCDW is marked by gradients in temperature, salinity, nutrient concentrations, and in the corrosiveness of the water with respect to calcium carbonate.

Late Quaternary Surface-Water Records

Sea-Surface Temperature

The distribution of SST in the South Atlantic is an important constraint for validating the results of coupled global climate models and serves as a boundary condition for simulating past ocean states with ocean general circulation models (e.g., Paul and Schäfer-Neth (2004)). SST reconstructions in the South Atlantic are based on transfer functions derived from assemblages of radiolaria, diatoms, or Foraminifera, alkenones, oxygen isotopes, Mg/Ca ratios, and calcium isotopes of planktic Foraminifera (for further information on paleoceanographic proxies, see Fischer and Wefer (1999)). Generally, SSTs in the tropical and South Atlantic were several degrees colder than today for much of the late Pleistocene. During the last interglacial and the preceding deglaciation (Termination II), however, temperatures were about 1.5 °C warmer than today in the subtropical South Atlantic and up to 3 °C warmer than today in the Southern Ocean (see Gersonde et al. (2004) and references therein). Most of the available time series indicate that SST in the South Atlantic and the Southern Ocean leads North Atlantic SST and ice volume at all orbital frequencies by several thousands of years. The early response of the Southern Ocean may be due to an early retreat of Antarctic sea ice in response to changes in summer insolation (e.g., Shemesh et al. (2002)). This early warming might then be transferred to the entire South Atlantic via advection.

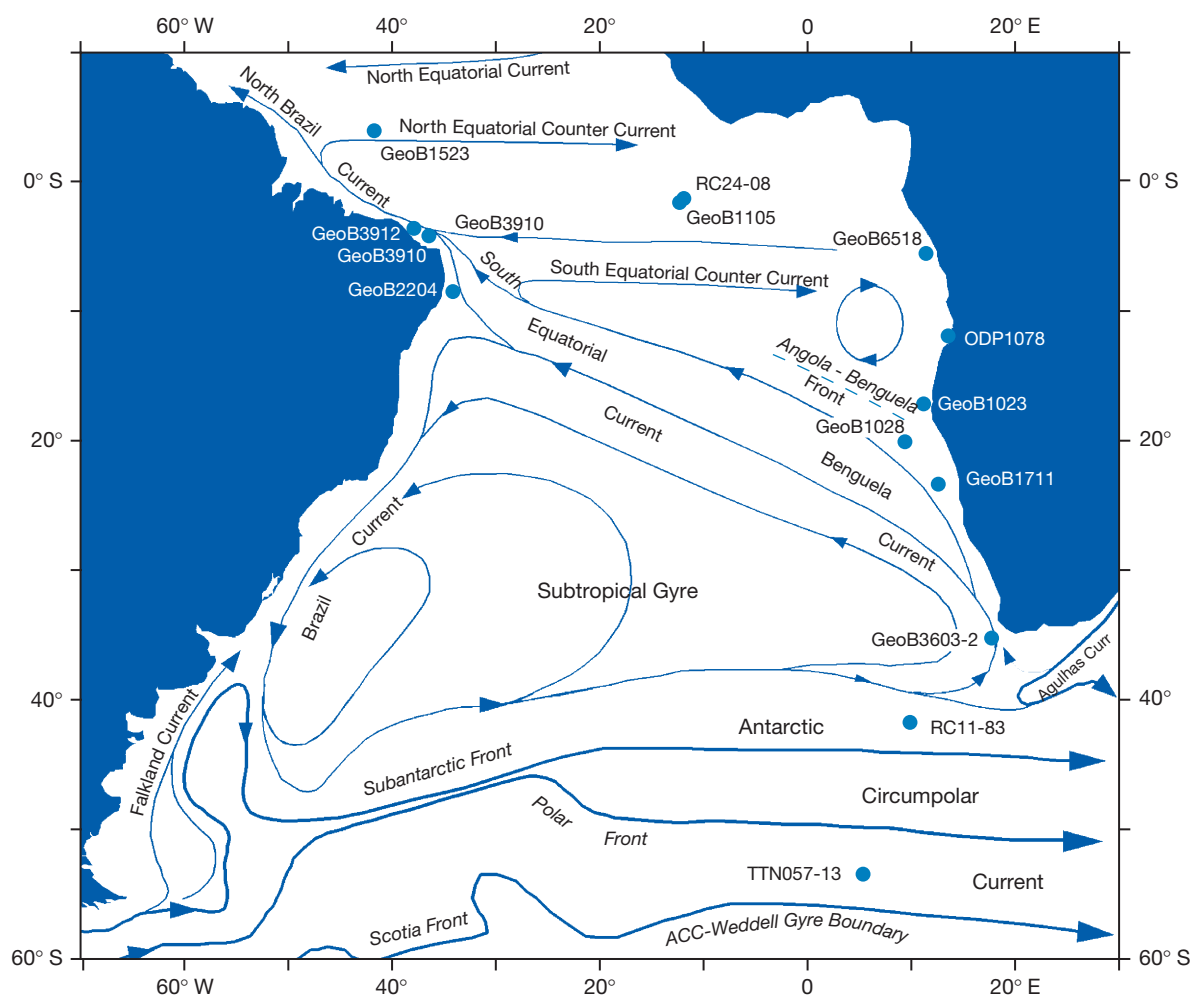


Figure 1 Schematic representation of the large-scale, upper-level circulation in the South Atlantic Ocean (see also contributions in Wefer et al. (1996)) and locations of some important late Pleistocene paleoceanographic records cited in the text. Redrawn from Stramma L and England M (1999) On the water masses and mean circulation of the South Atlantic Ocean. *Journal of Geophysical Research* 104(C9): 20, 863–20, 883.

Another important effect on SST at millennial timescales is the so-called bipolar seesaw: caused by the massive decrease in the Atlantic northward heat transport, the surface northern North Atlantic cooled by several degrees Celsius, while the tropical Atlantic and South Atlantic warmed. The seesaw effect is most pronounced during deglaciations, when a large amount of freshwater is shed into the North Atlantic, disrupting the thermohaline circulation. Evidence that the seesaw effect is actually working on shorter timescales comes from high-resolution cores from the tropical Atlantic (Mix et al., 1986). Here, the anti-phased warming shows a clear two-step structure as expected from two meltwater pulses into the North Atlantic.

The best spatial coverage of SST reconstructions is available for the Last Glacial Maximum (LGM). In the South Atlantic, the most important features of the glacial distribution of SSTs are (1) a distinct cooling in the Southern Ocean, (2) a strong cooling in the upwelling region off Namibia, (3) a moderate cooling of 2–3 °C in the western tropical Atlantic (Figure 2; e.g., Gersonde et al. (2004)). The glacial cooling in the tropical and South Atlantic Ocean is generally explained as the result of

several combined effects: lower glacial CO₂ values, stronger upwelling and offshore displacement of upwelling zones, northward displacement of the frontal zones in the Southern Ocean (see discussion on Agulhas leakage, below), and advection of cold water as well as northward expansion of sea ice in the Southern Ocean.

Upper Ocean Stratification and Thermocline Depth

In the tropical Atlantic, mixed-layer thickness and thermocline depth are strongly related to wind strength and direction (see references in Wolff et al. (1999)). For this reason, reconstructions of mixed-layer and thermocline depth provide important constraints on the Pleistocene evolution of the equatorial wind system. On orbital timescales, thermocline depth in the eastern equatorial Atlantic is characterized by variations in the 23-ka cycle, whereas the western equatorial Atlantic thermocline shows a dominant periodicity in the 41-ka band (Figure 3; Wolff et al., 1999). The phase relationships with respect to

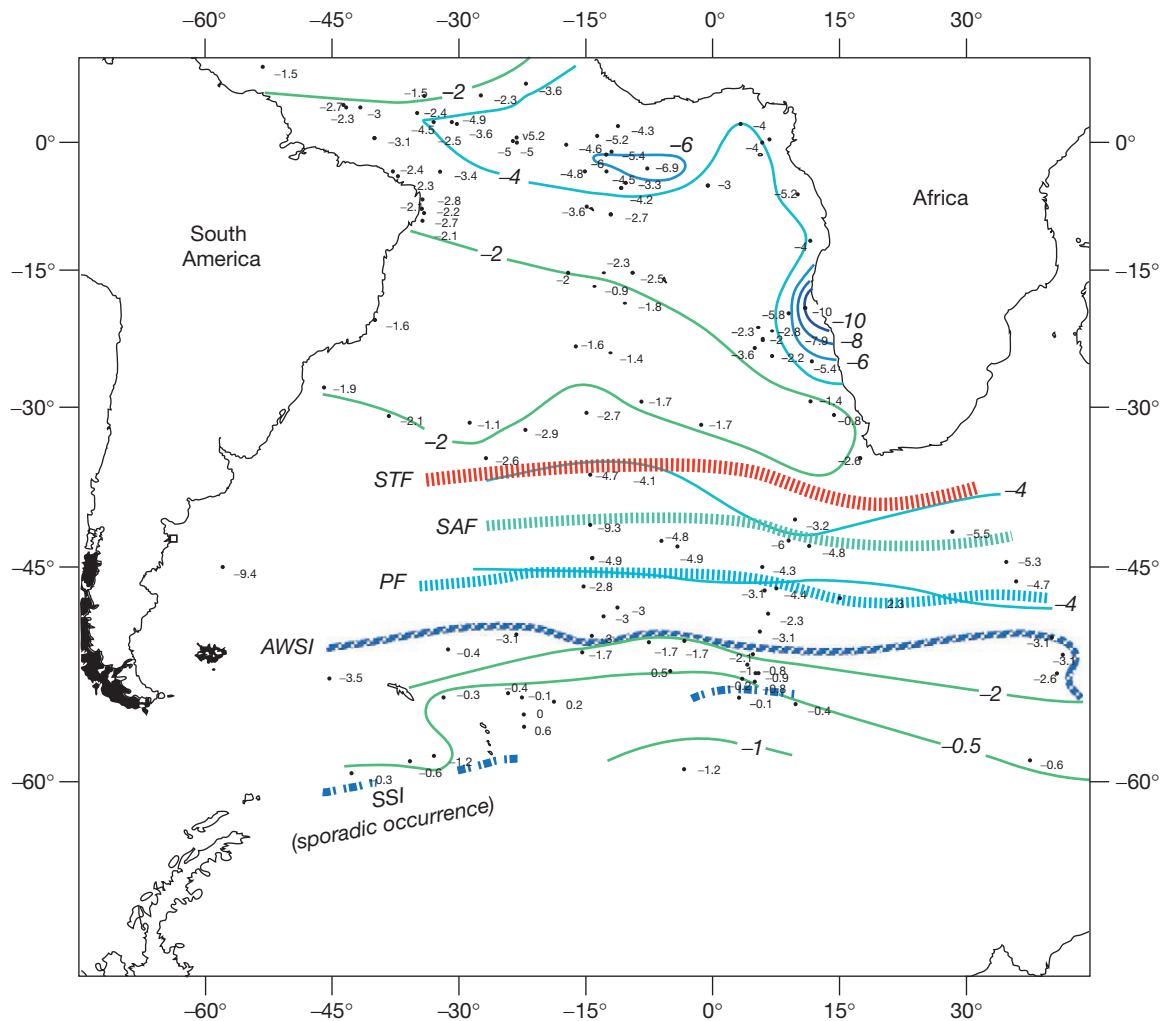


Figure 2 Austral summer sea-surface temperature anomaly (LGM – modern). Also indicated is the location of the estimated position of the Subtropical Front (STF), Sub-Antarctic Front (SAF), and Polar Front (PF), as well as the boundary of average winter sea-ice (AWSI) extent during the LGM. Arrows indicate sporadic expansion of summer sea ice (SSI) into the northern Weddell Basin. Compiled from various sources by Gersonde R, Abelmann A, Cortese G, et al. (2004) The late Pleistocene South Atlantic and Southern Ocean surface – A summary of time-series studies. In: Wefer G, Mulitza S, and Ratmeyer V (eds.) *The South Atlantic in the Late Quaternary: Reconstruction of Material Budgets and Current Systems*, p. 499. Springer.

insolation and ice volume suggest that thermocline fluctuations in the east are primarily driven by the intensity of the monsoon due to variations in low-latitude insolation. In contrast, changes in the thermocline depth in the western tropical Atlantic, which lead global ice volume at the 41-ka band by 3–4 ka, seem to depend on the strength of the SE trade winds in response to the meridional temperature gradients in the South Atlantic.

On shorter timescales, McIntyre and Molino (1996) proposed a link between tropical trade wind zonality and ice-raftering events (Heinrich Events) in the North Atlantic, based on the presence of the coccolithophore *Florisphaera profunda*, which is mainly controlled by thermocline depth. They report that maxima in mixed-layer depth in the eastern tropical Atlantic due to low easterlies correspond to Heinrich Events 1–5 in the North Atlantic (Figure 4). Warm water is released to the North Atlantic from the Caribbean warm water reservoir when tropical easterlies diminish, producing the rapid melting

of ice and hence Heinrich Events. A possible forcing of these events can be the precession cycle of insolation. Since the sun passes overhead twice a year in the tropics, two maxima in insolation exist during the course of a year. If climate is responding to the largest maximum of insolation, independent of the date it occurs during the year, it seems possible that precession is stimulating climate variability on half-precessional periods.

Evidence for higher seasonal stratification in the Southern Ocean, south of the modern Polar Front, comes from the radiolarian species *Cycladophora davisiana* (see references in Francois et al. (1997)). The only place where the modern abundances of this species are comparable to the glacial Southern Ocean is the Sea of Okhotsk, a marginal sea in the north-west Pacific. Here, the upper water column is characterized by a low sea-surface salinity layer with a strong temperature minimum at its base. Taking the Sea of Okhotsk as a modern analog, it has been proposed that the density stratification in the Southern Ocean was stronger than today during the LGM.

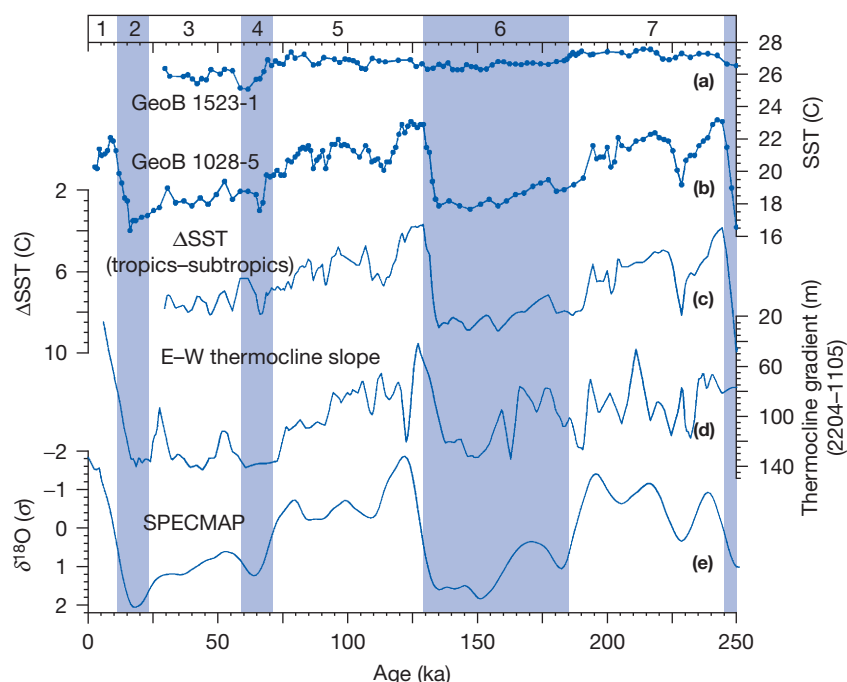


Figure 3 Comparison of the east-west thermocline slope with the tropical-subtropical SST evolution: (a) and (b) time series of alkenone-derived SSTs at core locations GeoB1523-1 and GeoB1028-5; (c) SST differences between cores GeoB1523-1 and GeoB1028-5; (d) thermocline differences (east-west slope) that were estimated by subtracting the thermocline depths of cores GeoB1105-4 and GeoB2204-2; and (e) SPECMAP stack indicative for global ice volume. From Wolff T, Mulitza S, Rühlemann C, and Wefer G (1999) Response of the tropical Atlantic thermocline to late Quaternary trade wind changes. *Paleoceanography* 14: 374–383, and references therein. Reproduced/modified by permission of American Geophysical Union.

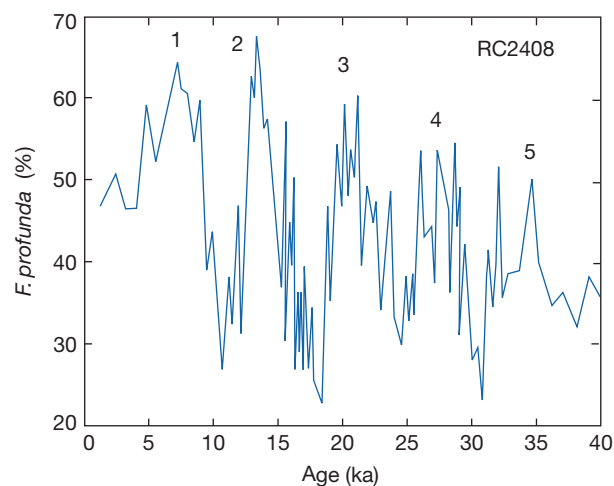


Figure 4 Percentage of the coccolithophoridae species *Florisphaera profunda* in core RC2408 from the eastern equatorial Atlantic vs. ^{14}C age. Maxima in abundance are assumed to reflect times of a deep mixed layer in the eastern equatorial Atlantic. Modified from McIntyre A and Molino B (1996) Forcing of Atlantic equatorial and subpolar millennial cycles by precession. *Science* 274: 1867–1870.

However, the common view that the formation of southern source deep water was enhanced during glacials (see section on ‘intermediate and deep-water masses’) requires convection rather than stratification, at least during winter time.

Coastal Upwelling

Upwelling areas are of great importance for regional and global climate through their contribution to biological productivity and their interaction with winds, cloud formation, and precipitation. In the South Atlantic, the largest upwelling regions are located off Namibia, in the Benguela System. Today, these upwelling areas are fed by South Atlantic Central Water from about 200 m water depth. The planktonic foraminifer *Neogloboquadrina pachyderma* attains its highest abundance in modern coastal upwelling areas and thus has been used as an indicator for coastal upwelling (e.g., West et al. (2004)). In sedimentary records from the Benguela region, cyclic variations in the abundance of *N. pachyderma* indicate increased upwelling intensity in glacial maxima and precession maxima, and correlate well with the wind-strength record reconstructed from grain-size analysis (see references in West et al. (2004)). During glacial maxima, steep temperature gradients over the Southern Hemisphere led to strong SE trade winds and strong upwelling. Precession maxima caused a weak monsoonal circulation, more zonal SE trade winds, and strong coastal upwelling.

Off Namibia, the abundance of *N. pachyderma* also shows variability at sub-Milankovitch frequencies during the last 140 ka (Figure 5; Little et al. (1997)). Periods of high abundance in left coiling *N. pachyderma* (so-called ‘PS events’) correlate with North Atlantic Heinrich Events and Dansgaard–Oeschger cycles from Greenland Ice Cores. In accordance with the eastern tropical Atlantic thermocline variability, this suggests large-scale

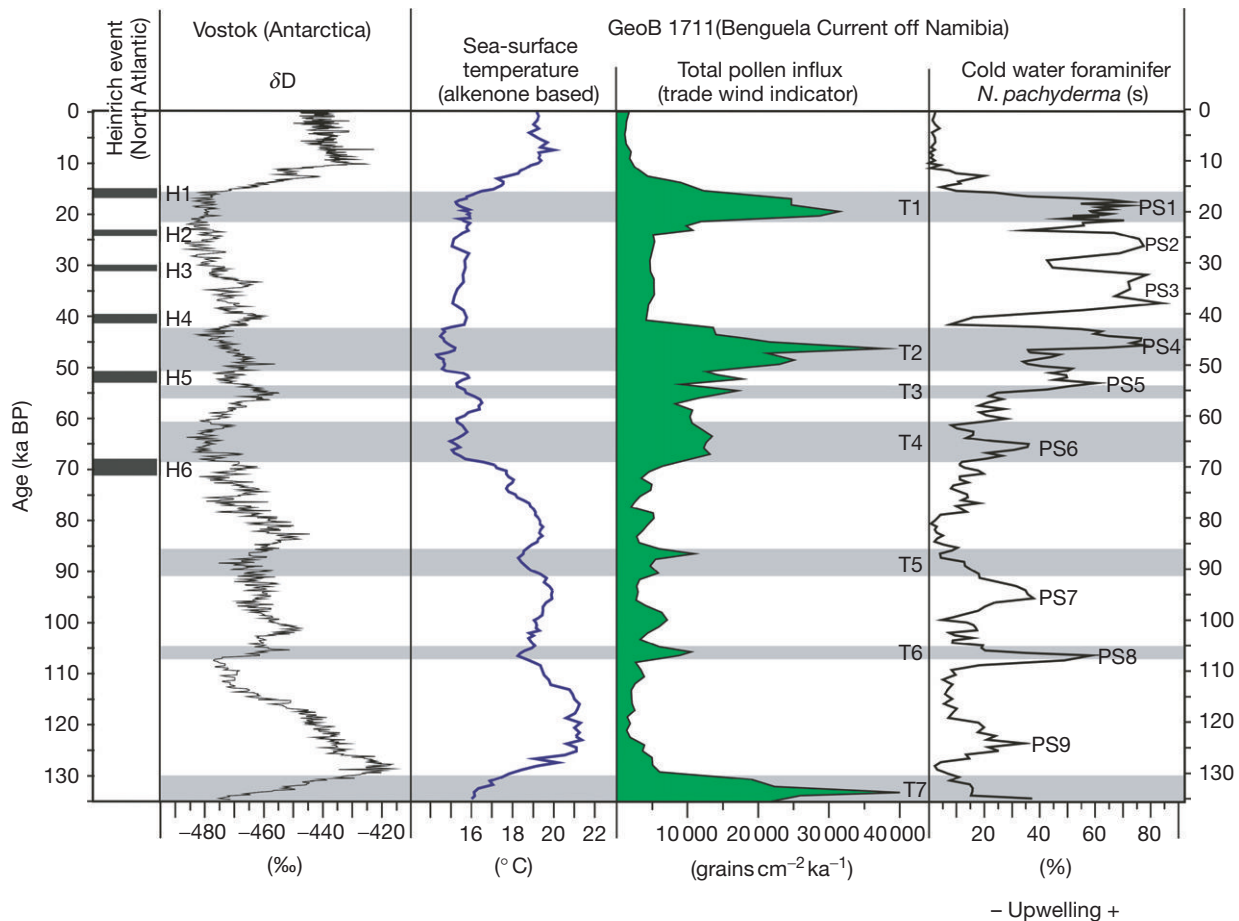


Figure 5 Comparison between alkenone-derived sea-surface temperature, total pollen influx, abundance of the planktic foraminifer *Neogloboquadrina pachyderma* in core GeoB1711-4 and the deuterium record of the Vostok ice core. Redrawn from Shi N, Schneider R, Beug HJ, and Dupont LM (2001) Southeast trade wind variations during the last 135 ka: Evidence from pollen spectra in eastern South Atlantic sediments. *Earth and Planetary Science Letters* 187: 311–321, Elsevier.

global oceanographic or climatic teleconnections between the South and North Atlantic via the trade wind system.

A problem concerning most upwelling indicators is to distinguish between variations in local upwelling strength and changes in the 'preformed' characteristics of the source water mass of the upwelled water. For example, inferred changes in upwelling strength might simply reflect changes in the temperature or nutrient content of the ascending central waters that are related to processes in higher latitudes rather than variations in the local wind system.

Agulhas Leakage

Today, the Cape of Good Hope is an important site of warm water and salt import into the Atlantic Ocean (see section on 'present-day circulation of the South Atlantic' and Paul and Schäfer-Neth (2004)). This salt and heat import from the Indian and Atlantic Oceans (the so-called Agulhas leakage) is considered to be an important process to maintain the global thermohaline circulation. Changes in the amount of Agulhas leakage have been reconstructed by means of foraminiferal assemblages (Peeters et al., 2004). Foraminiferal records

suggest that the Agulhas leakage was enhanced during interglacials when increased strength of the Indian monsoon system promoted the transport of warm tropical waters feeding the Agulhas current (Figure 6). The highest Agulhas leakage took place during early deglaciations, suggesting a crucial role for Agulhas leakage in glacial terminations (Peeters et al., 2004). A major concern is whether foraminiferal assemblages in the Agulhas region really reflect Agulhas leakage or are just a manifestation of the general peak warming of the Southern Ocean during deglaciations (see above). Moreover, studies of coccolithophore assemblages do not suggest a significant change in Agulhas flow (Winter and Martin, 1990).

Freshwater Input

River runoff and precipitation in the tropical Atlantic

The Amazon and Congo rivers have the biggest flow in the world ($\sim 0.2 \cdot 10^6 \text{ m}^3 \text{ s}^{-1}$), and are responsible for a large part of the negative sea-surface salinity anomaly observed in the present-day tropical Atlantic Ocean. Information about freshwater run-off has been reconstructed either directly via the oxygen isotope composition of planktic Foraminifera or via

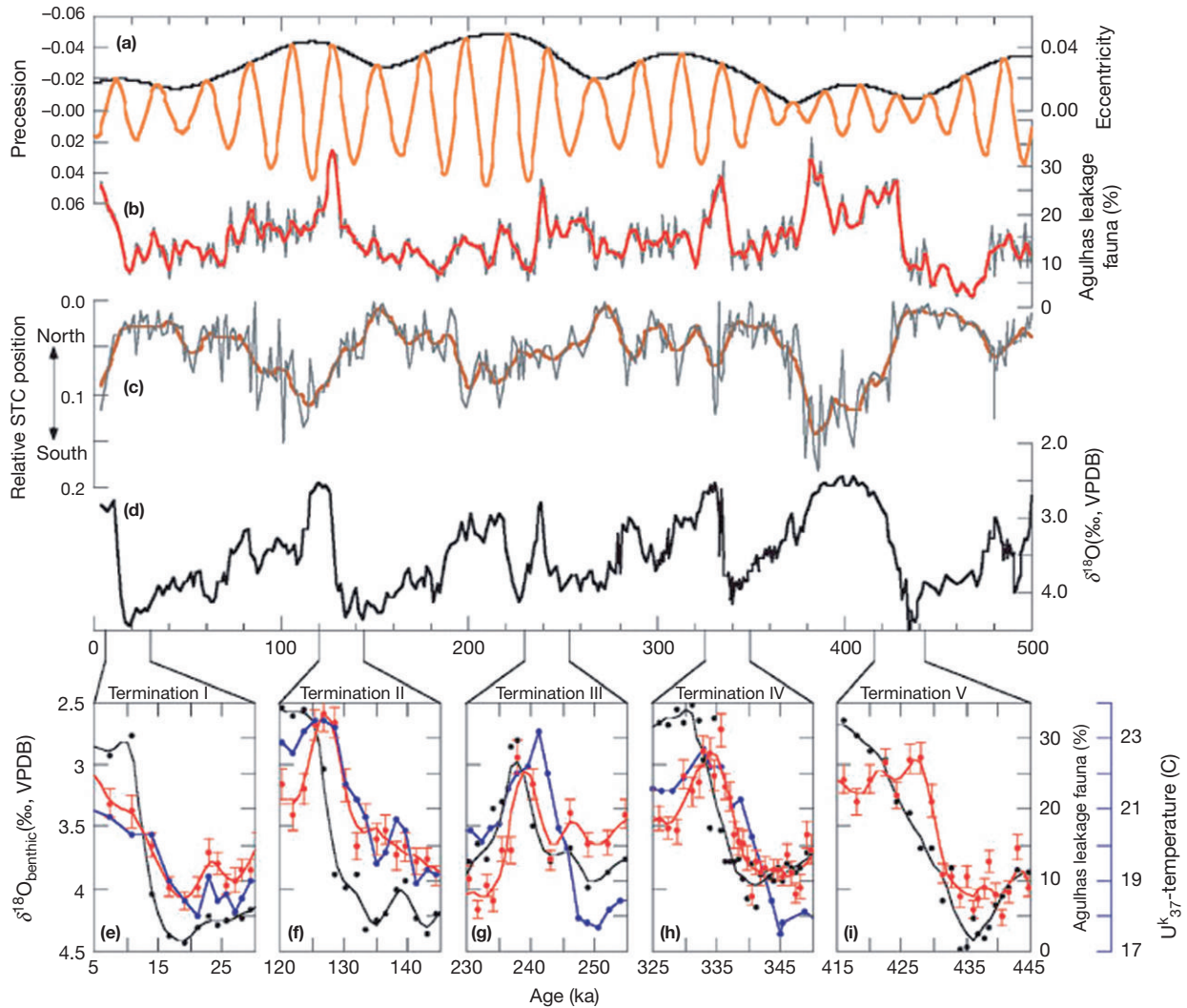


Figure 6 (a) Orbital eccentricity (black line) and precession (note inverted axis). (b) The Agulhas leakage fauna. Periods of maximum Agulhas leakage appear to be related to minima in precession (maxima in northern summer insolation). (c) The ratio of *G. truncatulinoides*/*N. pachyderma* dex. + *G. inflata* + *G. truncatulinoides* as a proxy for the distance of the STC relative to the position of the Cape basin record. High values indicate a southward position for the STC (note inverted axis). (d) The $\delta^{18}\text{O}$ of *Cibicides wuellerstorfi* (e–i). The Cape basin proxy records of Agulhas leakage (red line), global ice volume (black line), and SST (blue line) during the major late Pleistocene glacial Terminations I–V. From [Peeters FJC, Acheson R, Brummer G-JA, et al. \(2004\)](#) Vigorous exchange between the Indian and Atlantic oceans at the end of the past five glacial periods. *Nature* 430: 661–665.

indirect indicators, for example, the input of terrigenous components or freshwater diatoms. On orbital timescales monsoonal precipitation seems to follow the precessional cycle (see references in [Mulitza and Rühlemann \(2000\)](#)). Low Northern Hemisphere summer radiation during precession maxima suppresses the monsoonal circulation and leads to less precipitation over Africa and the tropical Atlantic. Actually, clay mineral proxies reveal that Congo River discharge and associated sediment input fluctuated in tune with precessional cycles of African monsoon intensity ([Gingele et al., 1998](#)). However, apart from the more gradual changes in monsoonal intensity on orbital time-scales, there is evidence for abrupt changes in freshwater supply during the last glacial period and during deglaciation. Measurements of elemental ratios in a high-resolution core from the continental slope off tropical Brazil show abrupt terrigenous sedimentation pulses that

suggest humid conditions and high surface runoff in north-east Brazil ([Figure 7; Arz et al. \(1998\)](#)). These pulses occur during times of ice rafting, reduced meridional overturning circulation, and low SSTs in the North Atlantic, the so-called Heinrich Events. Since SSTs in the tropical Atlantic are at a maximum during times of reduced overturning circulation (see section on ‘sea-surface temperature’) an increase of surface runoff off Brazil might be explained by a southward shift of the thermal equator and the associated Intertropical Convergence Zone ([Mulitza and Rühlemann \(2000\)](#) and references therein).

An alternative explanation might be an increased moisture supply with the SE trade winds. Today SE trade winds deliver intense precipitation to SE Brazil during Southern Hemisphere Winter. Increased tropical–subtropical temperature gradients in the South Atlantic during Heinrich Events might have led to

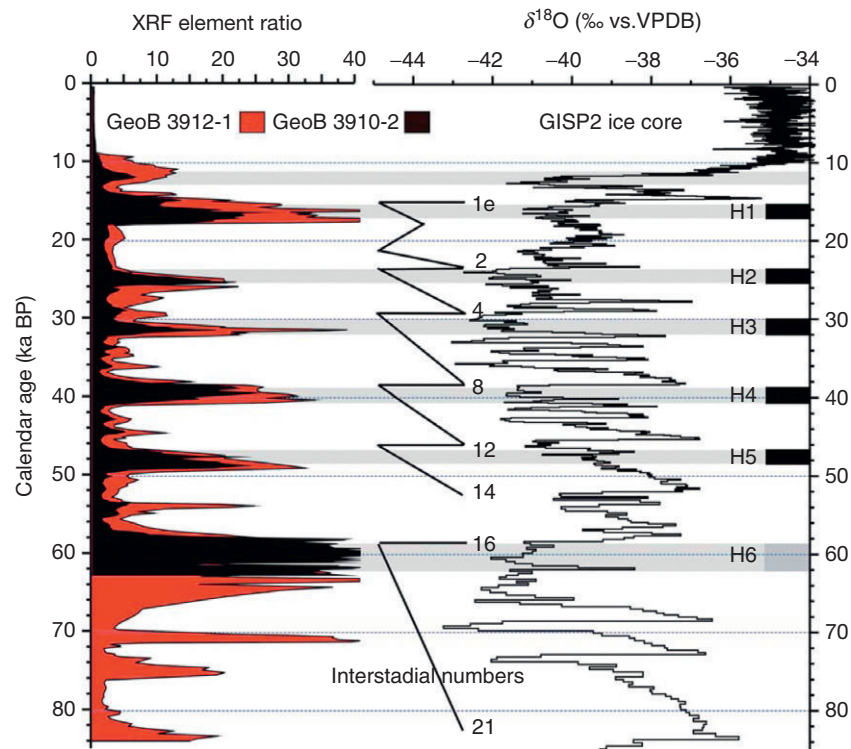


Figure 7 Comparison of the Ti/Ca and Fe/Ca ratio (Ti–Ca record scaled up fourfold) of cores GeoB 3910-2/3912-1 with the $\delta^{18}\text{O}$ record of the GISP2 ice core, the Heinrich Events, and the schematic longer-term cooling (Bond) cycles (see references in [Arz et al. \(1998\)](#)) (courtesy of Helge Arz).

stronger trade winds and more precipitation in the Brazilian hinterland. Since this increased supply of moisture should act at the expense of moisture supply to tropical Africa, we should expect that conditions are drier in tropical Africa during Heinrich Events. The isotopic composition of terrigenous plant lipids, extracted from a marine sediment core close to the Congo River mouth, indeed indicates that central African precipitation was reduced during Heinrich Event 1 and the Younger Dryas, contemporaneous with high differences in SSTs between the tropics and subtropics of the South Atlantic ([Figure 8](#); [Scheffuß et al. \(2005\)](#)). It must be noted, however, that organic components can have a significantly higher age than the surrounding sediment and may not be an optimal signal carrier for paleoclimatic proxies.

Meltwater and ice-rafted detritus (IRD)

The Quaternary history of meltwater and IRD input into the Southern Ocean has important implications for ice-sheet dynamics and the formation of intermediate and deep waters. The oxygen isotope composition of polar planktic Foraminifera together with records of IRD have been used to show that the region of the polar front has been covered by a meltwater lid containing significant contributions from melting icebergs in the period between 35 and 17 ka BP ([Labeyrie et al., 1986](#)). Cores from the southeast Atlantic Ocean show millennial-scale pulses of IRD delivery between 20 and 74 ka ([Figure 9](#); [Kanfoush et al. \(2000\)](#)). In this region, IRD events coincided with strong increases in NADW production and inferred warming (interstadials) in the high-latitude North Atlantic. Sea-level

rise or increased NADW production associated with strong interstadials may have resulted in destabilization of grounded ice shelves and surging in the Weddell Sea region. A recent compilation of IRD records by [Diekmann et al. \(2004\)](#), however, suggest that the temporal patterns of IRD distribution are not correlated over the entire Atlantic Sector of the Southern Ocean ([Figure 10](#)). In contrast to the southeast Atlantic Ocean, higher IRD fluxes are related to interglacial stages in the Weddell Sea. Marked IRD fluxes in the Bellingshausen Sea off the Antarctic Peninsula and in the southern and the northern Scotia Sea occur at glacial terminations and show small short-term spikes during MIS 3 and MIS 2. At present, it seems to be unclear whether the IRD signal really represents the dynamics of the Antarctic Ice Sheet, or is controlled by the temperature effect on the survival of icebergs or other sources of IRD from offshore Antarctica ([Diekmann et al., 2004](#)).

Sea Ice

Sea-ice variability plays a major role in global climate change through its effect on surface albedo and heat and salt balances in polar oceans. Recent sea-ice reconstructions in the Southern Ocean have mainly been done using the distribution of sea-ice diatoms (see [Gersonde et al. \(2004\)](#) and references therein). During the LGM, winter sea ice increased significantly by 60–70% with respect to the present ([Figure 2](#); [Gersonde et al. \(2004\)](#)). The summer sea-ice edge probably reached as far north as 55° S in the eastern Weddell Sea sector. In the southeast Atlantic Ocean, sea ice started to decline early

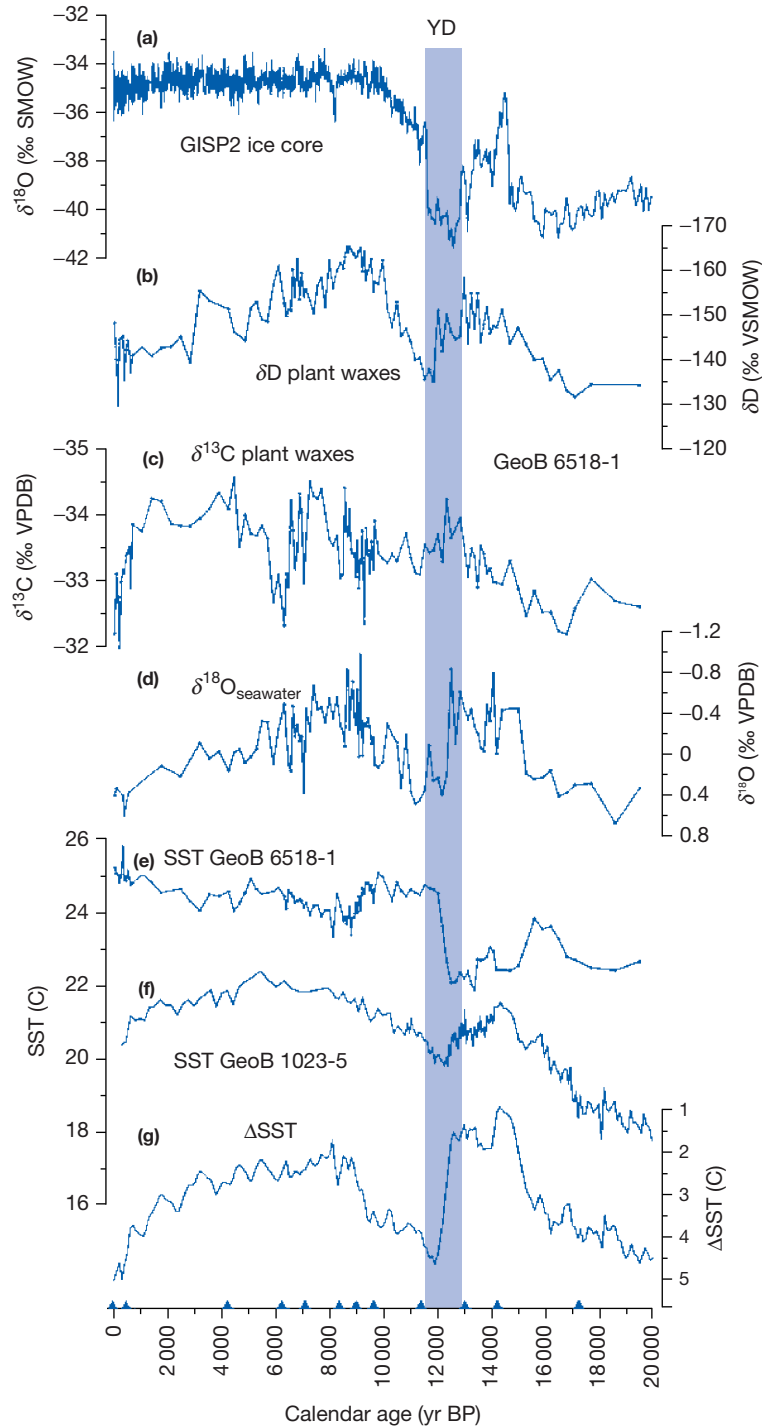


Figure 8 Deglacial-Holocene records of northern high-latitude and central African climate and SST variations in the tropical and subtropical Atlantic. Shown is (a) the GISP2 $\delta^{18}\text{O}$ record; (b) the plant-wax δD record from GeoB 6518-1 reflecting large-scale central African precipitation changes; (c) the plant-wax $\delta^{13}\text{C}$ record reflecting relative contributions from C3 and C4 plants; (d) the $\delta^{18}\text{O}$ record corrected for global ice volume and local temperature changes; (e) alkenone-derived sea-surface temperatures of core GeoB 6518-1; and (f) core GeoB 1023-5 from the subtropical South Atlantic and (g) the temperature difference between both cores. Triangles on lower axis indicate ^{14}C -AMS dates. The grey bar indicates the Younger Dryas period (YD; 12.9–11.6 ka BP). From Schefuß E, Schouten S, and Schneider RR (2005) Climatic controls on central African hydrology during the past 20,000 years. *Nature* 437: 1003–1006, and references therein.

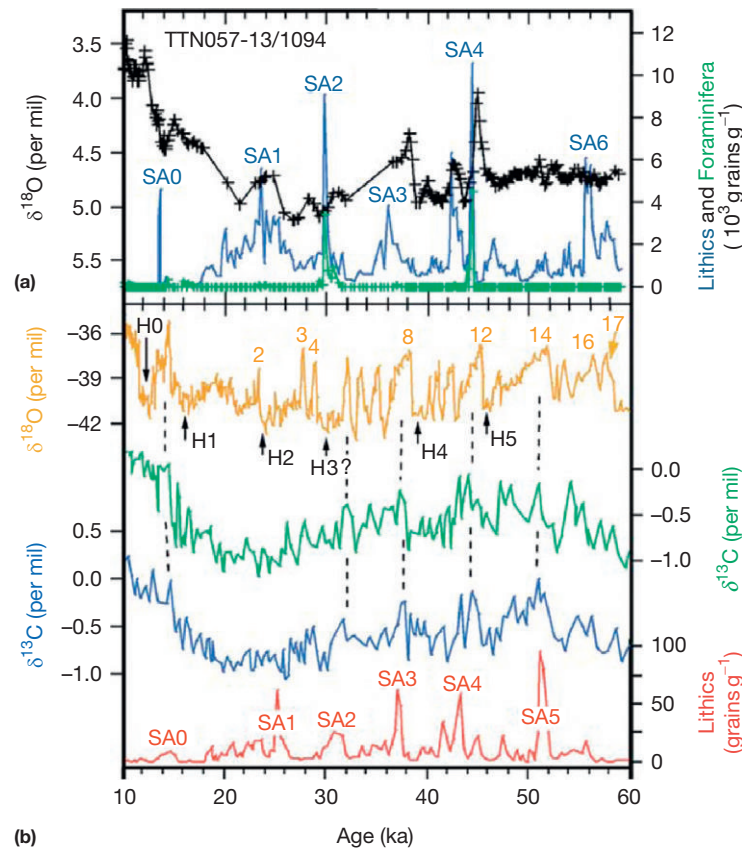


Figure 9 (a) Ice-rafting (SA) events in the Atlantic sector of the Southern Ocean. Concentration of lithic components (blue), $\delta^{18}\text{O}$ of *N. pachyderma* (black), and abundance of Foraminifera (green) in core TTN057-13 and ODP Site 1094 at 53°S . Event SA0 occurs during the Antarctic Cold Reversal, as indicated by an increase in planktic $\delta^{18}\text{O}$ values during Termination I. Events SA1, SA3, and SA4 are associated with low planktic $\delta^{18}\text{O}$ values, which indicate either the influence of meltwater or warm SST. Events SA2 and SA4 are associated with large increases in foraminiferal abundance. (b) Similarities exist between the $\delta^{18}\text{O}$ record of the Greenland Ice Sheet Project 2 (GISP2) ice core (gold) and the benthic $\delta^{13}\text{C}$ (*Cibicidoides*) records of sediment cores RC11-83 (green) and TTN057-21 (blue). Prominent interstadials as defined in the GISP2 record are numbered, and the North Atlantic Heinrich (H) Events are indicated in relation to climatic events in the Greenland ice cores. South Atlantic IRD events SA0 through SA5 are expressed as peaks in the total lithic concentration in core TTN057-21 at 41°S (red). Events SA2 through SA5 are each associated with high benthic $\delta^{13}\text{C}$ values, indicating strong NADW production and inferred warming in the North Atlantic. From Kanfoush SL, Hodell DA, Charles CD, Guilderson TP, Mortyn G, and Ninnemann US (2000) Millennial-scale instability of the Antarctic ice sheet during the last glaciation. *Science* 288: 1815–1818.

during deglaciation at about 19 ka BP (Figure 11; Shemesh et al. (2002)). The early retreat of sea ice probably also affected the wind and precipitation regimes over the eastern South Atlantic. The patterns of terrigenous sedimentation on the continental margin off southwest Africa reveals coherent changes between Antarctic sea-ice extent and the southwest African winter rain since 45 ka BP, with enhanced winter rainfall and trade wind strength during periods of increased sea-ice presence (Stuut et al., 2004).

Productivity

Primary production is an important factor in the climate system because it influences the partitioning of carbon dioxide between ocean and atmosphere (see references in Rühlemann et al. (1999)). During the late Quaternary, high and variable productivity in the eastern equatorial Atlantic contrasts with relatively constant and low productivity in the West. During

cold substages, paleoproductivity was at a maximum in the east, whereas minimum values were attained in the West. This countercyclicity is probably due to the equatorial thermocline tilt with upwelling of nutrient-rich waters in the east and a contemporaneous nutricline deepening in the West due to the piling up of warm and nutrient-poor surface waters. Furthermore, there is evidence that the glacial central waters were depleted in nutrients relative to today. Since thermocline waters are a primary source of nutrients in oligotrophic areas, a lowering of their nutrient concentrations might also have contributed to lower paleoproductivity in the glacial western tropical Atlantic.

In the Southern Ocean, glacial productivity was lower south of the present-day Polar Front and increased to the North relative to today (Francois et al., 1997). Increased surface-water stratification south of the Polar Front may have contributed to the lowering of atmospheric CO_2 concentration during the Last Glacial Maximum.

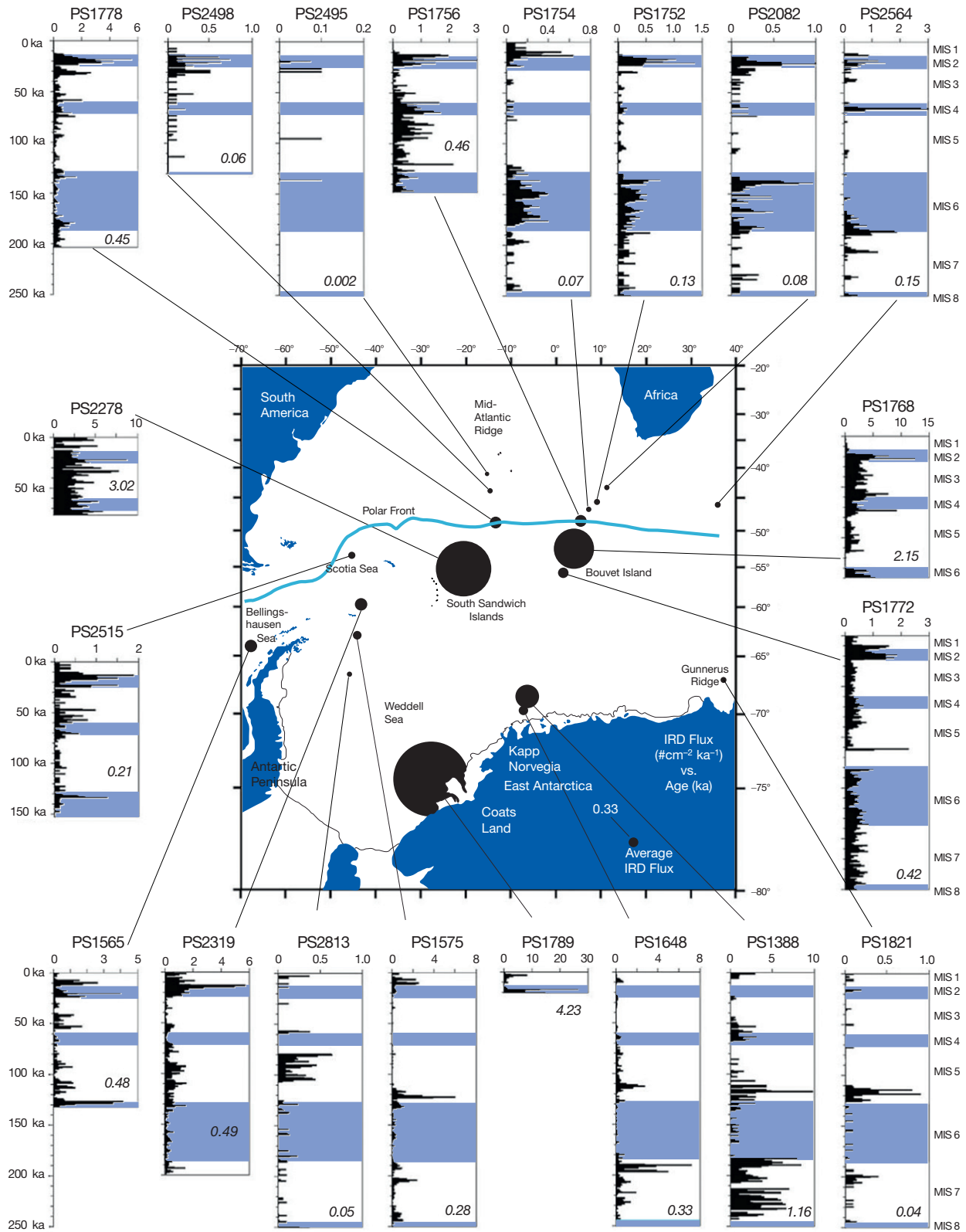


Figure 10 Temporal and spatial variability of flux rates of ice-rafted detritus (IRD) recorded in sediment cores of the study area. Diagrams display IRD fluxes in $\# \text{cm}^{-2} \text{ka}^{-1}$ vs. age in ka (last 250 ka) with indication of average IRD value (italic number) and marine isotope stages (glacial stages in gray). Black circles on the map are proportional to average IRD values in the respective sediment cores. From [Diekmann B, Fütterer DK, Grobe H, et al. \(2004\)](#) Terrestrial sediment supply in the polar to temperate South Atlantic: Land-ocean links of environmental changes during the late Quaternary. In: Wefer G, Mulitza S, and Ratmeyer V (eds.) *The South Atlantic in the Late Quaternary: Reconstruction of Material Budgets and Current Systems*, p. 375. Springer.

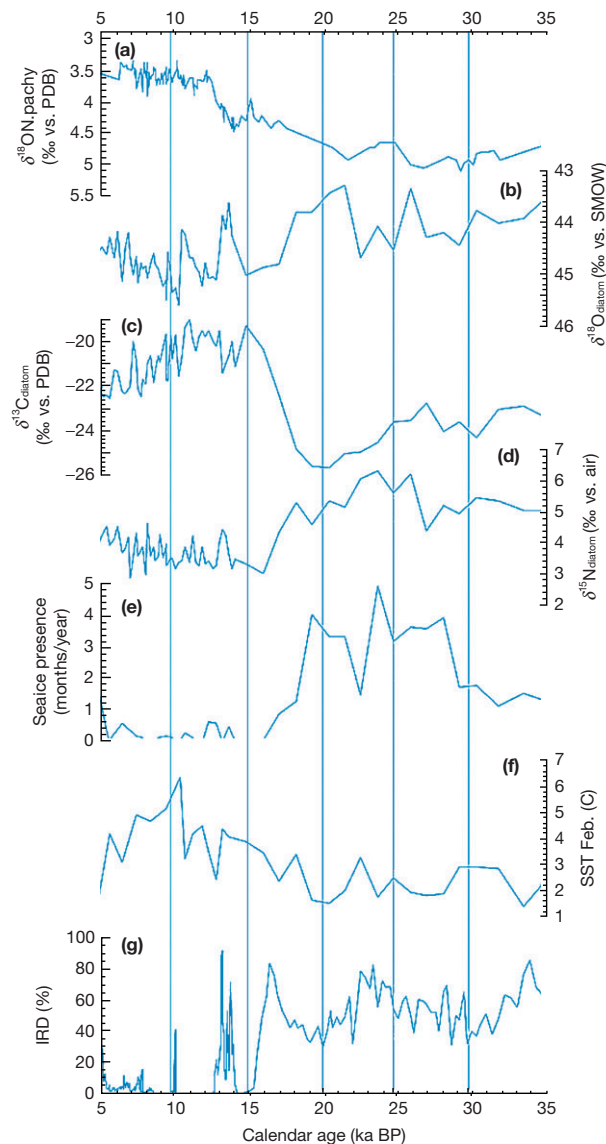


Figure 11 Downcore records of (a) $\delta^{18}\text{O}$ of the planktic foraminifer *N. pachyderma*; (b) $\delta^{18}\text{O}$ of biogenic silica; (c) diatom $\delta^{13}\text{C}$; (d) diatom $\delta^{15}\text{N}$; (e) sea-ice presence; (f) February sea-surface temperature; and (g) ice-rafted debris in core TTN057-13 (see Figure 9). From Shemesh A, Hodell D, Crosta X, Kanfoush S, Charles C, and Guilderson T (2002) Sequence of events during the last deglaciation in Southern Ocean sediments and Antarctic ice cores. *Paleoceanography* 17: <http://dx.doi.org/10.1029/2000PA000599>. Reproduced/modified by permission of American Geophysical Union.

Late Quaternary Intermediate- and Deep-Water Records

Thermocline and Central Water Masses

The water masses of the permanent thermocline are known as central water masses. They originate at the subtropical convergences of both hemispheres and spread toward the equatorial regions. The circulation within the permanent thermocline is held to be responsible for transmitting the glacial cooling from high latitudes to the tropics through a stronger circulation

within a cooler and shallower thermocline (see references in Paul and Schäfer-Neth (2004)). For this reason, changes of the circulation within the thermocline can have important effects on tropical/subtropical productivity. Carbon isotope ratios of deep-dwelling Foraminifera indicate that nutrients were depleted in the thermocline during cold substages of the late Quaternary. Covariance between the $\delta^{13}\text{C}$ records of the deep-dwelling foraminifer *Globorotalia truncatulinoides* from the western equatorial Atlantic and the benthic foraminifer *Cibicides wuellerstorfi* from Caribbean intermediate water suggests that the upper part of the western equatorial Atlantic water column was largely composed of nutrient-poor central and intermediate waters of northern origin (Mulitza et al., 1998). This pattern might have been the result of a circulation mode in which sub-Antarctic surface waters formed nutrient-rich deep waters, rather than intermediate waters.

Benthic foraminiferal oxygen isotope ratios of a shallow sediment core recovered at 426 m water depth in the eastern tropical Atlantic indicate a rapid and intense warming of central waters during Heinrich Event H1 and the Younger Dryas (Figure 12; Rühlemann et al. (2004)). This warming is consistent with a reduced ventilation of cold intermediate and deep waters in conjunction with downward mixing of heat from the thermocline as predicted by the 'seesaw effect.'

Intermediate- and Deep-Water Masses

The deep South Atlantic is an important site to monitor changes in deep-water circulation, because it contains the mixing zone between the NADW and the southern source water masses. Most reconstructions of deep-water circulation rely on the carbon isotopic composition of benthic Foraminifera, which is inversely related to the nutrient concentrations and hence can be used to trace deep-water masses. Generally, the deep basins of the South Atlantic were occupied by cold, ^{13}C -depleted water masses during glaciations, whereas intermediate levels were dominated by ^{13}C -enriched water masses of northern origin (Duplessy et al., 1988). This pattern can be explained by a shallower overturning and a greater input of deep water from the Southern Ocean. Meanwhile, a wealth of cores provides a very detailed picture of the distribution of deep- and intermediate-water masses during the LGM (see compilation by Curry and Oppo 2005, Figures 13 and 14). These data indicate the presence of three distinct water masses in the glacial Atlantic Ocean: a shallow (~ 1000 m), southern source water mass with a $\delta^{13}\text{C}$ value of about 0.3–0.5‰, a mid-depth (~ 1500 m), a northern source water mass with an end-member value of about 1.5‰, and a deep (>2000 m), southern source water mass with $\delta^{13}\text{C}$ values of less than -0.2 ‰.

Deep-water properties in the South Atlantic are also influenced by millennial-scale climate signals (Figure 15). In a core from the deep Cape Basin, Charles et al. (1996) found abrupt changes in the carbon isotopic composition of benthic Foraminifera that lag Southern Hemisphere climate proxies by about 1.5 ka. This lag either reflects a very early response of Southern Hemisphere surface waters or is due to the seesaw effect that predicts warming of the South Atlantic during thermohaline circulation slowdown.

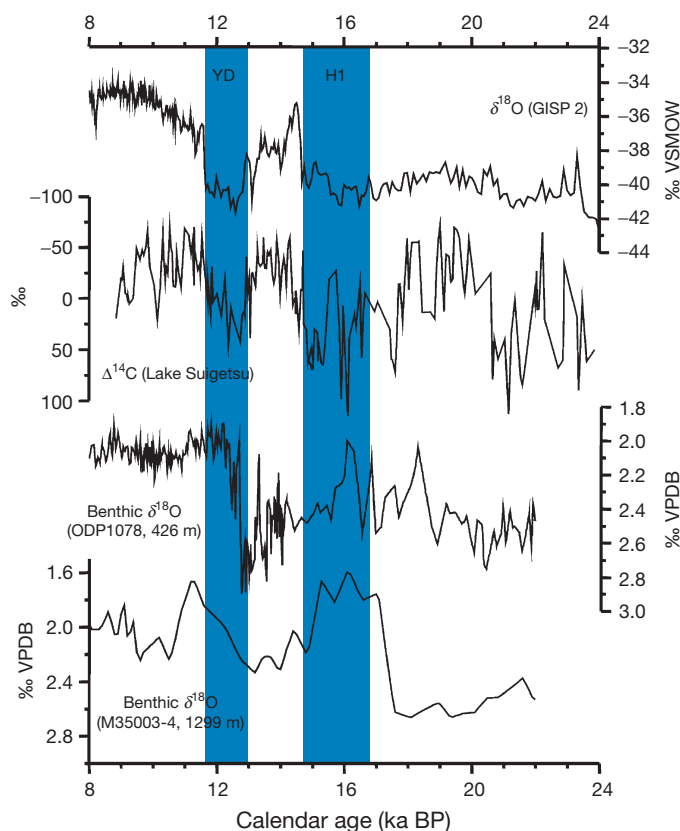


Figure 12 Evolution of benthic $\delta^{18}\text{O}$ in the cores M35003-4 ($12^{\circ}50' \text{ N}$, $61^{\circ}15' \text{ W}$; 1299-m water depth), and ODP1078 ($11^{\circ}55' \text{ S}$, $13^{\circ}24' \text{ E}$; 426-m water depth) during the last deglaciation and comparison to oxygen isotopes from the GISP2 ice core and atmospheric radiocarbon from sediments of Lake Suigetsu, Japan. Data from [Ruhlemann C, Mulitza S, Lohmann G, Paul A, Prange M, and Wefer G \(2004\)](#) Abrupt warming of the intermediate-depth tropical Atlantic caused by thermohaline circulation weakening: Evidence from paleoclimate data and model simulations. *Paleoceanography* 19: PA1025, (<http://dx.doi.org/10.1029/2003PA000948>). Reproduced/modified by permission of American Geophysical Union.

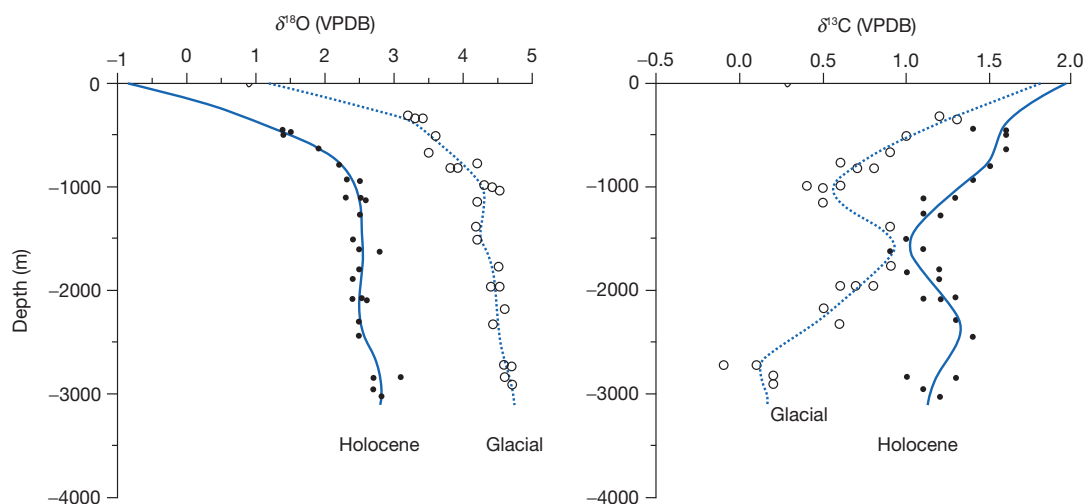


Figure 13 Holocene (solid line, solid symbols) and glacial (dashed line, open symbols) profiles of the stable isotopic compositions of benthic foraminifers *Cibicidoides* and *Planulina* from a vertical core transect on the Brazilian continental slope. Surface values represent stable isotope values of the planktonic foraminifer *G. ruber*. Several features of the Holocene and glacial water masses are depicted in the figures. The steep $\delta^{18}\text{O}$ gradient in the upper water column is controlled by the thermocline. The $\delta^{13}\text{C}$ profiles document changes in water-mass position and characteristics. The high $\delta^{13}\text{C}$ values associated with northern source waters shoaled during the glacial period to about 1500 m. A very steep bathymetric gradient in $\delta^{13}\text{C}$ below 1500 m was caused by gradual mixing between the northern source and southern source water masses at this location. [Curry WB and Oppo DW \(2005\)](#) Glacial water mass geometry and the distribution of $\delta^{13}\text{C}$ of ΣCO_2 in the western Atlantic Ocean. *Paleoceanography* 20: PA1017, (<http://dx.doi.org/10.1029/2004PA001021>). Reproduced/modified by permission of American Geophysical Union.

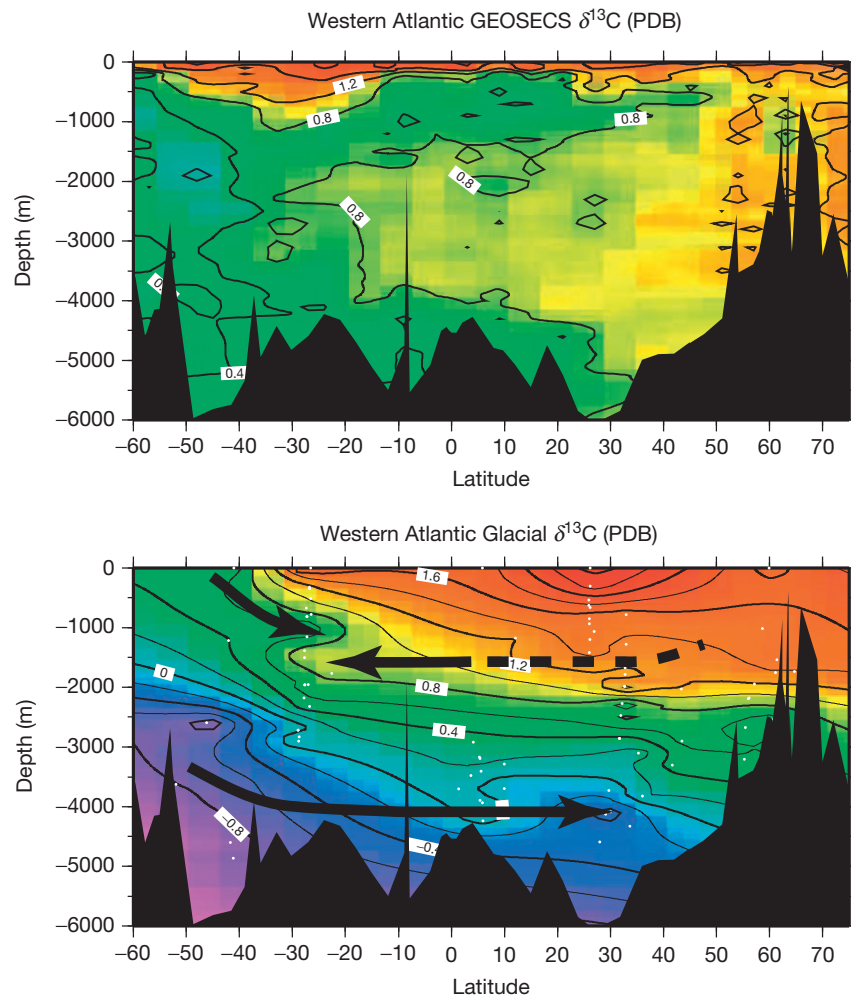


Figure 14 The distribution of $\delta^{13}\text{C}$ of dissolved CO_2 in the modern western Atlantic (top panel); and for the Last Glacial Maximum (bottom panel). A three-component model is proposed to explain the water-mass distribution and gradients. The patterns in the South Atlantic imply that the northern source water at 1,500 m penetrated as far as 30° S latitude. The penetration of the deep Southern Ocean water mass can be observed as far as 60° N. From Curry WB and Oppo DW (2005) Glacial water mass geometry and the distribution of $\delta^{13}\text{C}$ of ΣCO_2 in the western Atlantic Ocean. *Paleoceanography* 20: PA1017 (doi:10.1029/2004PA001021). Reproduced/modified by permission of American Geophysical Union.

Conclusions

Many of the processes that are at work in the present-day South Atlantic and Southern Ocean were affected by millennial- and orbital-scale climate changes during the late Pleistocene. The South Atlantic is not only reacting but also actively contributing to global climate change. For example, stratification and productivity of the Southern Ocean influenced the concentration of atmospheric CO_2 , and variations in sea-ice extent modified planetary albedo. The heat transport in the South Atlantic may be influenced by the changes in the input of warm water at the Cape of Good Hope. Furthermore, millennial-scale changes in wind-driven heat transport, potentially forced by changes in

orbital insolation, have been suggested as a potential trigger of ice-rafter events and deglaciations in the Northern Hemisphere. Future work must explore whether the tropical Atlantic can actually be a driver of abrupt climate change.

See also: **Diatom Records:** Antarctic Waters. **Paleoceanography, Biological Proxies:** Alkenone Paleothermometry Based on the Haptophyte Algae; Benthic Foraminifera. **Paleoceanography, Records:** Late Pleistocene North Atlantic. **Paleoclimate Modeling:** Quaternary Environments. **Radiocarbon Dating:** AMS Radiocarbon Dating.

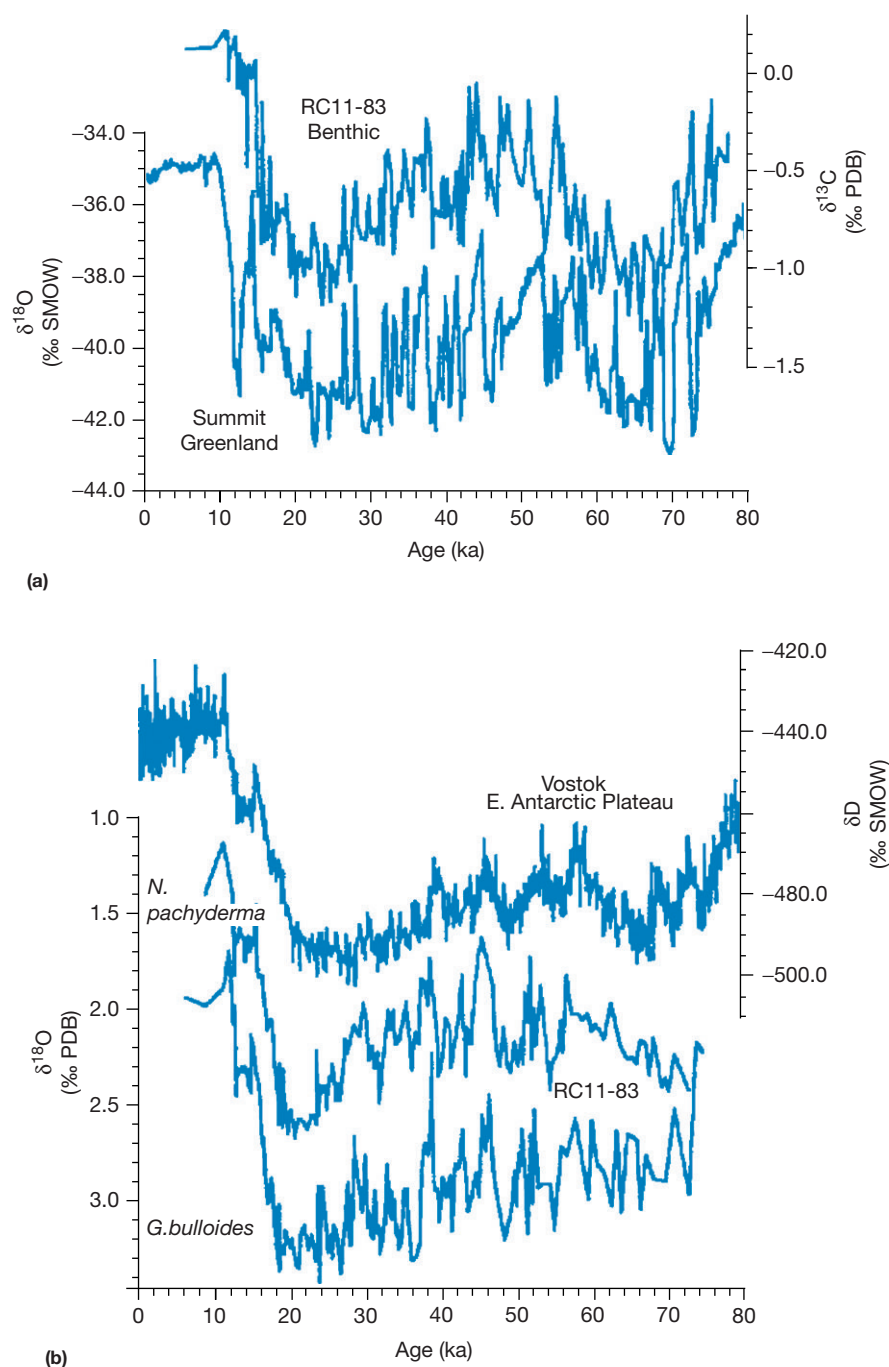


Figure 15 (a) Comparison of time series of benthic foraminiferal $\delta^{13}\text{C}$ in core RC11-83 with the Summit ice core $\delta^{18}\text{O}$ from Greenland. (b) Time series of $\delta^{18}\text{O}$ from two planktonic foraminiferal species with the deuterium (δD) record of Vostok. From Charles CD, Lynch-Stieglitz J, Ninnemann US, and Fairbanks RG (1996) Climate connections between the hemispheres revealed by deep sea sediment core/ice core correlations. *Earth and Planetary Science Letters* 142: 19–27.

References

- Arz HW, Pätzold J, and Wefer G (1998) Correlated millennial-scale changes in surface hydrography and terrigenous sediment yield inferred from last-glacial marine deposits off northeastern Brazil. *Quaternary Research* 50: 157–166.
- Charles CD, Lynch-Stieglitz J, Ninnemann US, and Fairbanks RG (1996) Climate connections between the hemispheres revealed by deep sea sediment core/ice core correlations. *Earth and Planetary Science Letters* 142: 19–27.
- Curry WB and Oppo DW (2005) Glacial water mass geometry and the distribution of $\delta^{13}\text{C}$ of ΣCO_2 in the western Atlantic Ocean. *Paleoceanography* 20: PA1017, <http://dx.doi.org/10.1029/2004PA001021>.
- Diekmann B, Fütterer DK, Grobe H, et al. (2004) Terrigenous sediment supply in the polar to temperate South Atlantic: Land-ocean links of environmental changes during the late Quaternary. In: Wefer G, Mulitza S, and Ratmeyer V (eds.) *The South Atlantic in the Late Quaternary: Reconstruction of Material Budgets and Current Systems*, p. 375. Berlin Heidelberg New York: Springer-Verlag.

- Duplessy JC, Shackleton NJ, Fairbanks RG, Labeyrie L, Oppo D, and Kallel N (1988) Deepwater source variations during the last climatic cycle and their impact on the global deepwater circulation. *Paleoceanography* 3: 343–360.
- Fischer G and Wefer G (1999) *Use of Proxies in Paleoceanography: Examples from the South Atlantic*. Berlin Heidelberg New York: Springer-Verlag.
- Francois R, Altabet MA, Yu E-F, et al. (1997) Contribution of Southern Ocean surface-water stratification to low atmospheric CO₂ concentrations during the last glacial period. *Nature* 389: 929–935.
- Gersonde R, Abelmann A, Cortese G, et al. (2004) The Late Pleistocene South Atlantic and Southern Ocean surface – A summary of time-series studies. In: Wefer G, Mulitza S, and Ratmeyer V (eds.) *The South Atlantic in the Late Quaternary: Reconstruction of Material Budgets and Current Systems*, p. 499. Berlin Heidelberg New York: Springer-Verlag.
- Gingele FX, Müller PJ, and Schneider RR (1998) Orbital forcing of freshwater input in the Zaire Fan area – Clay mineral evidence from the last 200 kyr. *Palaeogeography, Palaeoclimatology, Palaeoecology* 138: 17–26.
- Kanfoush SL, Hodell DA, Charles CD, Guilderson TP, Mortyn G, and Ninnemann US (2000) Millennial-scale instability of the Antarctic ice sheet during the last glaciation. *Science* 288: 1815–1818.
- Labeyrie LD, Pichon JJ, Labracherie M, Ippolito P, Duprat J, and Duplessy JC (1986) Melting history of Antarctica during the past 60,000 years. *Nature* 322: 701–706.
- Little MG, Schneider RR, Kroon D, Price B, Summerhayes CP, and Segl M (1997) Trade wind forcing of upwelling, seasonality, and Heinrich Events as a response to Sub-Milankovitch climate variability. *Paleoceanography* 12: 568–576.
- McIntyre A and Molfino B (1996) Forcing of Atlantic equatorial and subpolar millennial cycles by precession. *Science* 274: 1867–1870.
- Mix AC, Ruddiman WF, and McIntyre A (1986) Late Quaternary paleoceanography of the tropical Atlantic. 1: Spatial variability of annual mean sea-surface temperatures, 0–20,000 years B.P. *Paleoceanography* 1: 43–66.
- Mulitza S and Rühlemann C (2000) African monsoonal precipitation modulated by interhemispheric temperature gradients. *Quaternary Research* 53: 270–274.
- Mulitza S, Rühlemann C, Bickert T, Hale W, Pätzold J, and Wefer G (1998) Late Quaternary $\delta^{13}\text{C}$ gradients and carbonate accumulation in the western equatorial Atlantic. *Earth and Planetary Science Letters* 155: 237–249.
- Paul A and Schäfer-Neth C (2004) The Atlantic Ocean at the Last Glacial Maximum. 2: Reconstructing the current systems with a global ocean model. In: Wefer G, Mulitza S, and Ratmeyer V (eds.) *The South Atlantic in the Late Quaternary: Reconstruction of Material Budgets and Current Systems*, p. 549. Berlin Heidelberg New York: Springer-Verlag.
- Peeters FJC, Acheson R, Brummer G-JA, et al. (2004) Vigorous exchange between the Indian and Atlantic oceans at the end of the past five glacial periods. *Nature* 430: 661–665.
- Rühlemann C, Müller PJ, and Schneider RR (1999) Organic carbon and carbonate as paleoproductivity proxies: Examples from high and low productivity areas of the tropical Atlantic. In: Fischer G and Wefer G (eds.) *Use of Proxies in Paleoceanography: Examples from the South Atlantic*, p. 315. Berlin Heidelberg New York: Springer-Verlag.
- Rühlemann C, Mulitza S, Lohmann G, Paul A, Prange M, and Wefer G (2004) Abrupt warming of the intermediate-depth tropical Atlantic caused by thermohaline circulation weakening: Evidence from paleoclimate data and model simulations. *Paleoceanography* 19: PA1025. <http://dx.doi.org/10.1029/2003PA000948>.
- Schefufl E, Schouten S, and Schneider RR (2005) Climatic controls on central African hydrology during the past 20,000 years. *Nature* 437: 1003–1006.
- Shemesh A, Hodell D, Crosta X, Kanfoush S, Charles C, and Guilderson T (2002) Sequence of events during the last deglaciation in Southern Ocean sediments and Antarctic ice cores. *Paleoceanography* 17: <http://dx.doi.org/10.1029/2000PA000599>.
- Shi N, Schneider R, Beug HJ, and Dupont LM (2001) Southeast trade wind variations during the last 135 kyr: Evidence from pollen spectra in eastern South Atlantic sediments. *Earth and Planetary Science Letters* 187: 311–321.
- Stramma L and England M (1999) On the water masses and mean circulation of the South Atlantic Ocean. *Journal of Geophysical Research* 104(C9): 20,863–20,883.
- Stuut JWB, Crosta X, van der Borg K, and Schneider R (2004) Relationship between Antarctic sea ice and southwest African climate during the late Quaternary. *Geology* 32: 909–912.
- Wefer G, Berger WH, Siedler G, and Webb DJ (1996) *The South Atlantic: Present and Past Circulation*. Berlin Heidelberg New York: Springer-Verlag.
- Wefer G, Mulitza S, and Ratmeyer V (2004) *The South Atlantic in the Late Quaternary: Reconstruction of Material Budgets and Current Systems*. Berlin Heidelberg New York: Springer-Verlag.
- West S, Jansen JHF, and Stuut J-B (2004) Surface water conditions in the Northern Benguela Region (SE Atlantic) during the last 450 ky reconstructed from assemblages of planktonic foraminifera. *Marine Micropaleontology* 51: 321–344.
- Winter A and Martin K (1990) Late Quaternary history of the Agulhas current. *Paleoceanography* 5: 479–486.
- Wolff T, Mulitza S, Rühlemann C, and Wefer G (1999) Response of the tropical Atlantic thermocline to late Quaternary trade wind changes. *Paleoceanography* 14: 374–383.

Late Pleistocene North Pacific

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Introduction

The paleoceanography of the North Pacific is less well-known than that from other regions of the world's oceans, especially its Northern Hemisphere counterpart, the North Atlantic (see [Late Pleistocene North Atlantic](#)). The scarcity of records from the North Pacific has to do with the paucity or complete absence of carbonate in the deep-sea sediments in large parts of the Pacific (see [Dissolution of Deep-Sea Carbonates](#)). Most areas of the North Pacific lie below the carbonate compensation depth (CCD) which means that Foraminifera are absent from the sediments. In the modern North Pacific the CCD is as deep as 5 km in the central equatorial region and shoals to 3 km in the northern sub-Arctic Pacific. Even above those depths, in the lysocline (see [Dissolution of Deep-Sea Carbonates](#)) partial dissolution of carbonate shells is prevalent, which can lead to a bias of proxy records. However, in contrast to the Atlantic Ocean, the CCD was deeper by a few hundred meters during the last glaciation, resulting in somewhat better preservation of Foraminifera in glacial sediments ([Broecker, 1982](#)) and hence in improved conditions for paleoceanographic reconstructions. Unlike the North Atlantic, the North Pacific appeared as a passive ocean with little dynamics in ocean circulation and hydrography. This perception had already manifested itself when the ocean received its Latin name *mare pacificum*, 'peaceful sea,' by the explorer Ferdinand Magellan after his pioneering crossing of the Pacific Ocean. Oceanographically, the Pacific Ocean also appears relatively calm with its lack of deep-water formation, absence of large Arctic ice sheets throughout the late Pleistocene (see [Late Pleistocene Glacial Events in Beringia](#)) and the generally zonally aligned hydrographic pattern. However, the paleoceanographic records retrieved from the North Pacific reveal that the region was subject to a dynamic climate and oceanography during the late Pleistocene and before. Moreover, since the recognition of the far-field effects of El Niño, it has also become clear that oceanographic changes in the Pacific have the potential to substantially influence climate and environment on a global scale and most likely have done so in the past. For instance, the North Pacific provides moisture to the monsoon regions in Southeast Asia, to precipitation in North America, and to the growth of the Northern Hemisphere ice sheets. Furthermore, the North Pacific, due to its position at the far end of the global ocean circulation pathway, stores large quantities of nutrients and carbon in its subsurface waters. The importance of the Pacific Ocean as a whole is amplified by its sheer size and volume. With $180 \times 10^6 \text{ km}^2$ and volume of about $700 \times 10^6 \text{ km}^3$, the Pacific accounts for approximately half of the world's ocean volume, hence providing a huge storage capacity for all kinds of environmental substances, most importantly CO_2 .

Oceanographic Setting

Upper Ocean Circulation

The surface ocean circulation of the North Pacific is dominantly driven by the westerly and trade winds. The wind system generates the anticyclonic subtropical gyre and the cyclonic sub-Arctic gyre, which are separated from each other by the eastward flowing North Pacific Current at 30° – 42° N , the southern part of which is known as the Kuroshio Current Extension. The exact location of the North Pacific Current, an associated sub-Arctic–subtropical frontal system, moves around the mid-latitude area today and was almost certainly not stationary on paleoceanographic timescales. The same is therefore true for the location of the bifurcation of the North Pacific Current into the northward flowing eastern limb of the subpolar gyre and the southward flowing California Current.

Subsurface Ocean Circulation

No deep water is formed in the North Pacific today due to low salinity of the surface waters in the sub-Arctic gyre, which maintain a density contrast relative to the underlying waters, even if cooled close to freezing temperature. Pacific Deepwater below 1000 m therefore originates from southern source Circumpolar Deepwater that enters the northern basin along the western Pacific margin via the Deep Western Boundary Current. Gain of buoyancy through mixing with overlying less dense waters and, to a lesser extent, through geothermal heating ([Adcroft et al., 2001](#)) result in some rising of bottom waters to shallower deep-water levels and recirculation out of the North Pacific partly through the Indonesian Passage into the Indian Ocean and partly into the South Pacific, and back into the Antarctic Circumpolar Current, and from there through the Drake Passage into the Atlantic and Southern Ocean. The only subsurface water-mass forming in the North Pacific is the North Pacific intermediate water (NPIW). NPIW forms as a result of winter sea ice formation in the Sea of Okhotsk and mixing and subduction processes at the convergence of the subtropical Kuroshio and sub-Arctic Oyashio currents in the western North Pacific off Japan. The water mass spreads widely eastwards below the subtropical gyre at water depths between approximately 200 and 800 m. An important question of North Pacific paleoceanography is whether in past times, such as the Last Glacial Maximum (LGM), NPIW formation was increased, maybe even to the extent that it generated a deep-water mass of North Pacific origin. Deep-water formation in the North Pacific would alter inventories of oxygen and nutrients, and hence reduce the capacity of North Pacific waters to store carbon.

Occurrence of Paleoceanographic Records

The sedimentological characteristics described in the 'Introduction' determine the distribution of paleoceanographic records in the North Pacific. Most records are derived from the eastern and western continental margins and marginal seas. These locations are sufficiently shallow to preserve carbonate. Furthermore, sediments at these locations were often deposited at sedimentation rates of at least several centimeters per 1 ka, high enough to make them useful for paleoceanographic studies. North Pacific records from open ocean sites are rare. They are associated with seamounts that reach above the carbonate compensation depth, for example, on the Emperor–Hawaii Seamount chain, or singular seamounts under the Alaskan gyre in the sub-Arctic northeastern Pacific, or under the subtropical gyre in the western Pacific. It is evident that the Pleistocene paleoceanography of large regions in the central sub-Arctic and central subtropical North Pacific will never be investigated by means of carbonate-based geochemistry or micropaleontology.

The Paleoceanographic Record

Deep and Intermediate Water Records

Glacial–interglacial changes

Studies of deep and intermediate water paleoceanography in the North Pacific focus on changes in ocean circulation and hydrography. The Pacific has a significant capacity to store heat and various dissolved species of gases and salts. However, this capacity can vary with changing ocean circulation. Such variations may also have played a substantial role in global change during the late Pleistocene. Most studies of intermediate and deep water circulation in the North Pacific use the carbon isotopic composition of epibenthic Foraminifera shells ($\delta^{13}\text{C}_{\text{bent}}$), a proxy for the ventilation (oxygen content) of water masses (see **Nutrient Proxies, Carbon Cycle Proxies** ($\delta^{11}\text{B}$, $\delta^{13}\text{C}_{\text{calcite}}$, $\delta^{13}\text{C}_{\text{organic}}$, **Shell Weights, B/Ca, U/Ca, Zn/Ca, Ba/Ca**)). A number of other studies rely on the radiocarbon age difference ($\Delta^{14}\text{C}$) between benthic and planktonic Foraminifera to approximate the ages of subsurface water masses since their isolation from the atmosphere by sinking from the surface ocean into the ocean interior. Fewer studies are based on trace metal distribution in sediments and foraminiferal shells (see **Mg/Ca and Sr/Ca Paleothermometry from Calcareous Marine Fossils**) and on micropaleontological and sedimentological evidence. The overwhelming bulk of studies address the difference between the status during the LGM (ca. 18–24 ka) and the last deglaciation (ca. 18–11 ka) (see **Postglacial North Pacific**), while only singular studies reconstruct the development across the late Pleistocene (see Section '**Millennial-scale ventilation changes**').

$\delta^{13}\text{C}_{\text{bent}}$. Depth transects of benthic carbon isotope records (Duplessy et al., 1988; Keigwin, 1998; Stott et al., 2000; Matsumoto et al., 2002; Ohkushi et al., 2003) draw an overall consistent picture of glacial water column ventilation. Above about 2 km water depth $\delta^{13}\text{C}_{\text{bent}}$ values are as much as ca. 0.8‰ higher during the LGM than during the late Holocene, after addition of a global 'secular' value that adjusts for 0.3‰ lower average oceanic glacial $\delta^{13}\text{C}$ values. Below 2 km water

depth, the (corrected) glacial $\delta^{13}\text{C}_{\text{bent}}$ values were the same as or slightly lower (no more than 0.2‰) than today. Accordingly, $\delta^{13}\text{C}$ data suggest that the glacial North Pacific was better ventilated above 2 km, and – if anything – slightly less well ventilated in deep waters below 2 km. The finding of improved glacial ventilation above 2 km appears to be robust and of almost North Pacific-wide relevance. The result is supported by $\delta^{13}\text{C}$ transects from the far northwestern Pacific and Okhotsk Sea (Keigwin, 1998) as well as from a depth transect of seven cores from the northeastern Pacific California Continental Borderland (Stott et al., 2000). Matsumoto et al. (2002) compiled $\delta^{13}\text{C}$ data from the Pacific that were available at the time (exclusive of those from the California Borderland) and mapped the ventilation state of the Pacific Ocean in a zonally averaged profile. The most striking feature is a persistent zone of minimum ventilation in the Pacific, north of 20° N. The ventilation minimum was centered between 1 and 2 km water depth during the Holocene but more than 1 km deeper during the LGM (Figure 1). The relatively shallow depth of better glacial ventilation suggests a scenario in which intensified glacial NPIW production allowed waters to sink to about 1 km deeper than today and spread more widely through the North Pacific Ocean.

Cd/Ca. While $\delta^{13}\text{C}$ is a widely used proxy for the ventilation (oxygenation) state of water masses, it has its limitations in the clarity of interpretation. For example, in ocean regions or during times of high phytoplankton production, the $\delta^{13}\text{C}$ ventilation signal can be masked by locally high flux of fresh organic matter to the seafloor (the 'Mackensen effect'; see **Carbon Cycle Proxies** ($\delta^{11}\text{B}$, $\delta^{13}\text{C}_{\text{calcite}}$, $\delta^{13}\text{C}_{\text{organic}}$, **Shell Weights, B/Ca, U/Ca, Zn/Ca, Ba/Ca**), **Nutrient Proxies**). Complementary to the $\delta^{13}\text{C}$ records there exist a few trace metal-based reconstructions of the nutrient status of past bottom waters. Records of Cd/Ca ratios from benthic Foraminifera (see **Nutrient Proxies**) from water depths below 2700 m indicate that glacial nutrient concentrations were up to 50% lower than during the Holocene (Ohkouchi et al., 1994). These data seem to be in conflict with $\delta^{13}\text{C}_{\text{bent}}$ evidence for a slightly reduced ventilation of glacial deep waters. However, the Holocene Cd/Ca reference data in the records compiled by Ohkouchi et al. (1994) are based on only a few ambiguous values. Records with a better resolution, especially in the Holocene section, will be required to underpin the apparent disagreement between ventilation and nutrient proxy records.

$\Delta^{14}\text{C}$. As another independent way to reconstruct past circulation changes, some studies used the radiocarbon age difference between the surface and the bottom water, as determined from planktonic and benthic Foraminifera. The age difference provides a measure for the ventilation age of the subsurface water mass in which the benthic Foraminifera grew. The $\Delta^{14}\text{C}$ record of the North Pacific is patchy and not always unambiguous. With 2000 ± 450 yr being the average ventilation age measured for the glacial deep Pacific, the ventilation ages of glacial deepwater (3200 m water depth) were approximately 500 years older than today (Shackleton et al., 1988). On the contrary, glacial ventilation ages from water depths above 2000 m are slightly younger (Broecker et al., 2004a). The inferred scenario of better ventilated intermediate waters and less well-ventilated deep waters qualitatively match the

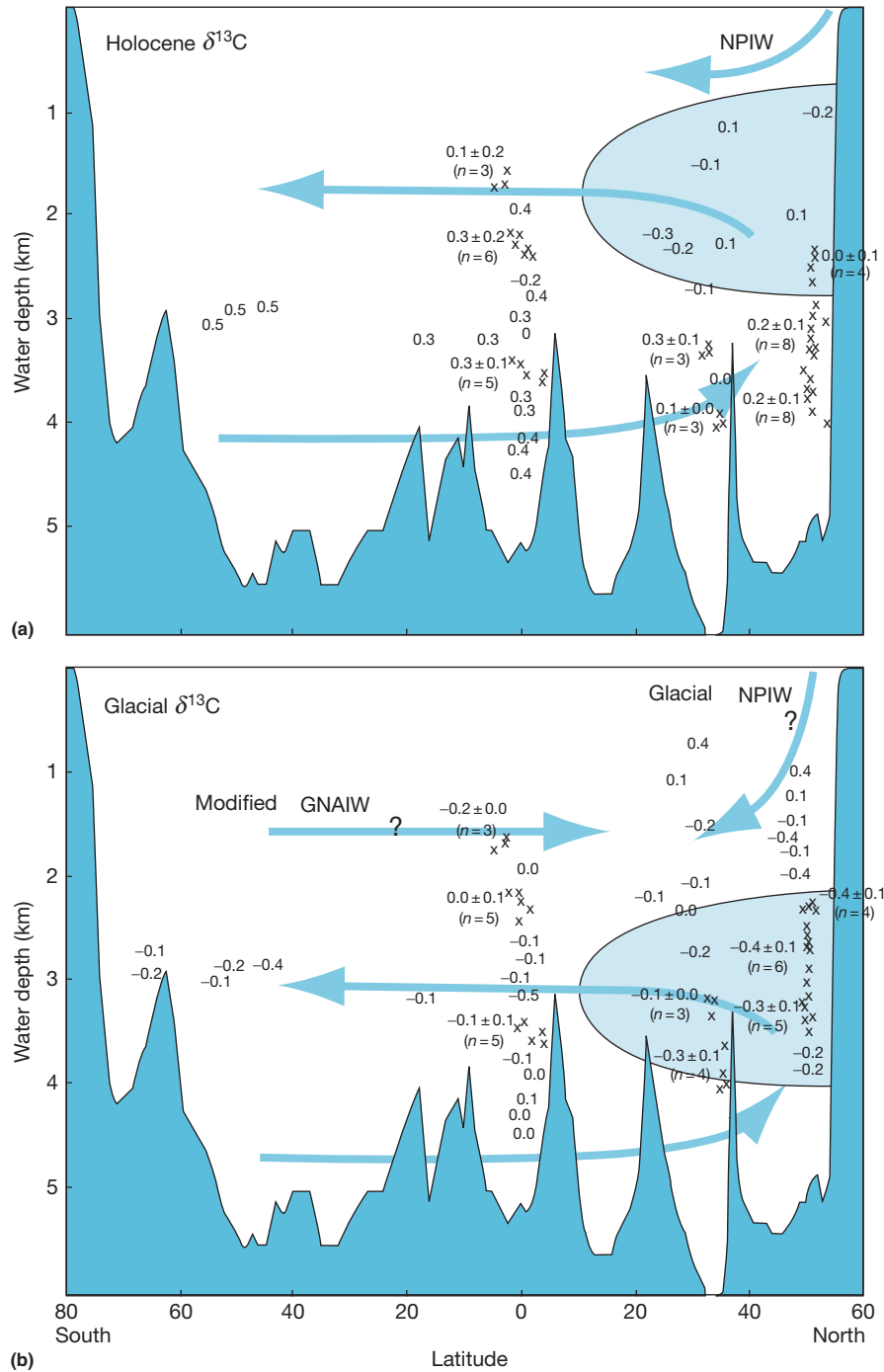


Figure 1 Meridional distribution of benthic foraminiferal $\delta^{13}\text{C}$ in the western Pacific Ocean. The lowest $\delta^{13}\text{C}$ indicated by shading is found in the deep North Pacific centered around 2000 m water depth during the Holocene and around 3000 m during the glaciation. Arrows indicate suggested circulation. The higher $\delta^{13}\text{C}$ in the upper 2000 m in the North Pacific reflect a local source of newly ventilated glacial NPIW, influx of modified Glacial North Atlantic Intermediate Water, or both. From Matsumoto K, Oba T, Lynch-Stieglitz J, and Yamamoto H (2002) Interior hydrography and circulation of the glacial Pacific Ocean. *Quaternary Science Reviews* 21: 1693–1704.

reconstructions from benthic $\delta^{13}\text{C}$. However, in an unsorted compilation of benthic-planktic $\Delta^{14}\text{C}$ data from the Pacific Ocean, Broecker et al. (2004b) demonstrate that the range of values span over 1 ka (Figure 2) for reasons of spatial heterogeneity, contamination with secondary calcite, bioturbational mixing of the sediment, and probably other causes. Future

studies using benthic-planktic $\Delta^{14}\text{C}$ as a ventilation proxy will therefore need to carefully choose suitable samples and exclude biasing factors.

Shell thickness of Foraminifera. A novel proxy for the estimation of water-mass chemistry is the shell thickness of planktonic Foraminifera, measured as the weight of shells of a

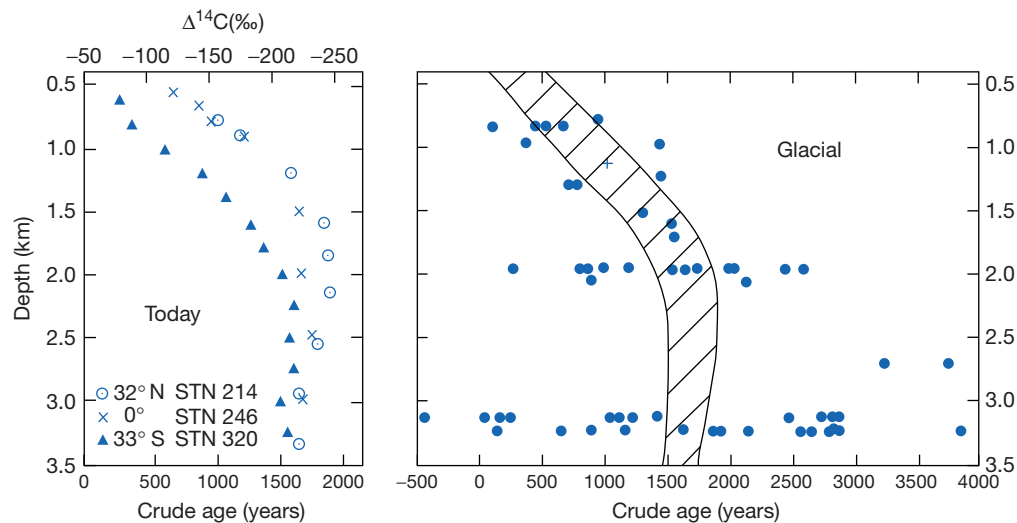


Figure 2 In the left-hand panel are the measurements of the $\Delta^{14}\text{C}$ (upper scale) for dissolved inorganic carbon (ΣCO_2) at three stations occupied during the GEOSECS survey. Also shown are the equivalent crude ages (lower scale) calculated assuming that the $\Delta^{14}\text{C}$ for surface mixed layer ΣCO_2 is 50‰. In the right-hand panel are crude age differences for coexisting glacial age benthic-planktic pairs (solid circles). The hatched region represents the present-day range of ages in the Pacific Ocean as a function of water depth. From Broecker WS, Clark E, Hajdas I, and Bonani G (2004b) Glacial ventilation rates for the deep Pacific Ocean. *Paleoceanography* 19: PA2002.

defined grain size. Assuming that the shell thickness decreases with gradual carbonate dissolution, one can approximate the corrosiveness, hence carbonate ion CO_3^{2-} concentration of bottom waters. A depth transect from the Ontong Java Plateau suggests higher-than-today CO_3^{2-} ion concentrations in the glacial western equatorial Pacific below 3 km water depth (Broecker and Clark, 2001). Higher CO_3^{2-} concentration in otherwise unchanged conditions would imply an improved ventilation of Pacific deep water. On a glacial-interglacial timescale, however, the lower concentration of CO_2 in the atmosphere may have biased the CO_3^{2-} reconstructions. On a global scale, it resulted in a shift of ocean chemistry toward higher CO_3^{2-} concentrations. Furthermore, it has recently been found that the initial thickness of Foraminifera shells is not constant as assumed, but varies according to surface water CO_3^{2-} saturation, which is linked to atmospheric CO_2 and other parameters. The shell weight record of the Pacific will therefore need re-examination in order to extract an unambiguous paleoceanographic signal.

Micropaleontological evidence. The species composition of the benthic Foraminifera community was investigated at sites of intermediate water depths at 1083 m and 2048 m off Japan (Ohkushi et al., 2003). Glacial assemblages are characteristic of well-oxygenated bottom waters compared with the low-oxygen assemblages found in Holocene sections. These results support geochemical evidence of better ventilation of glacial NPIWs to a depth of at least 2000 m. Moreover, the distribution of fossil abundances of the radiolarian species *Cycladophoria davisiana* provides information on potential regions of glacial deep convection and hence intermediate water formation (Ohkushi et al., 2003). Today, high abundances of this species are found mainly in the Okhotsk Sea, where deep convection forms part of the NPIW. In contrast to this, in last glacial sediments abundance maxima were found in the Bering Sea (Figure 3), suggesting that the Bering Sea contributed

substantially to the increased ventilation of intermediate depth waters in the glacial North Pacific.

Millennial-scale ventilation changes

The few existing records of changes in North Pacific deep ocean circulation before the LGM are restricted to the northeastern Pacific margin. Variations in the North Pacific are usually described in the context of rapid changes elsewhere, mostly of the Dansgaard-Oeschger and Heinrich events as recorded in Greenland and the North Atlantic (see Sub-Milankovitch (DO/Heinrich) Events). However, it should be kept in mind that these interbasin comparisons are in most cases limited by poor chronostratigraphic control, the improvement of which is presently a major challenge to high-resolution paleoceanography. A sediment core from 2712 m water depth off Oregon provides a benthic foraminiferal $\delta^{13}\text{C}$ record of deep-water ventilation over the last 60 ka (Lund and Mix, 1998). Increases in $\delta^{13}\text{C}$ on the order of 0.2‰ display repeated millennial-scale events of improved ventilation during stadial episodes (cold events in Greenland), in particular, during the pronounced stadials marked by Heinrich melt-water events in the North Atlantic (Figure 4). The millennial-scale pattern of deep ocean ventilation in the Pacific was therefore suggested as being opposite to the pattern in the Atlantic. This could mean that during cold stadials, deep water formed in the North Pacific down to at least 2700 m. Such deep-water formation could have been enabled by higher surface-water salinity in the North Pacific in response to reduced freshwater import due to cooler Atlantic surface waters and decreased intensity of the Southeast Asian summer monsoon. Alternatively, faster turnover of Pacific deep waters, driven by Southern Ocean winds, may have compensated for suppressed North Atlantic deep water production during stadial intervals.

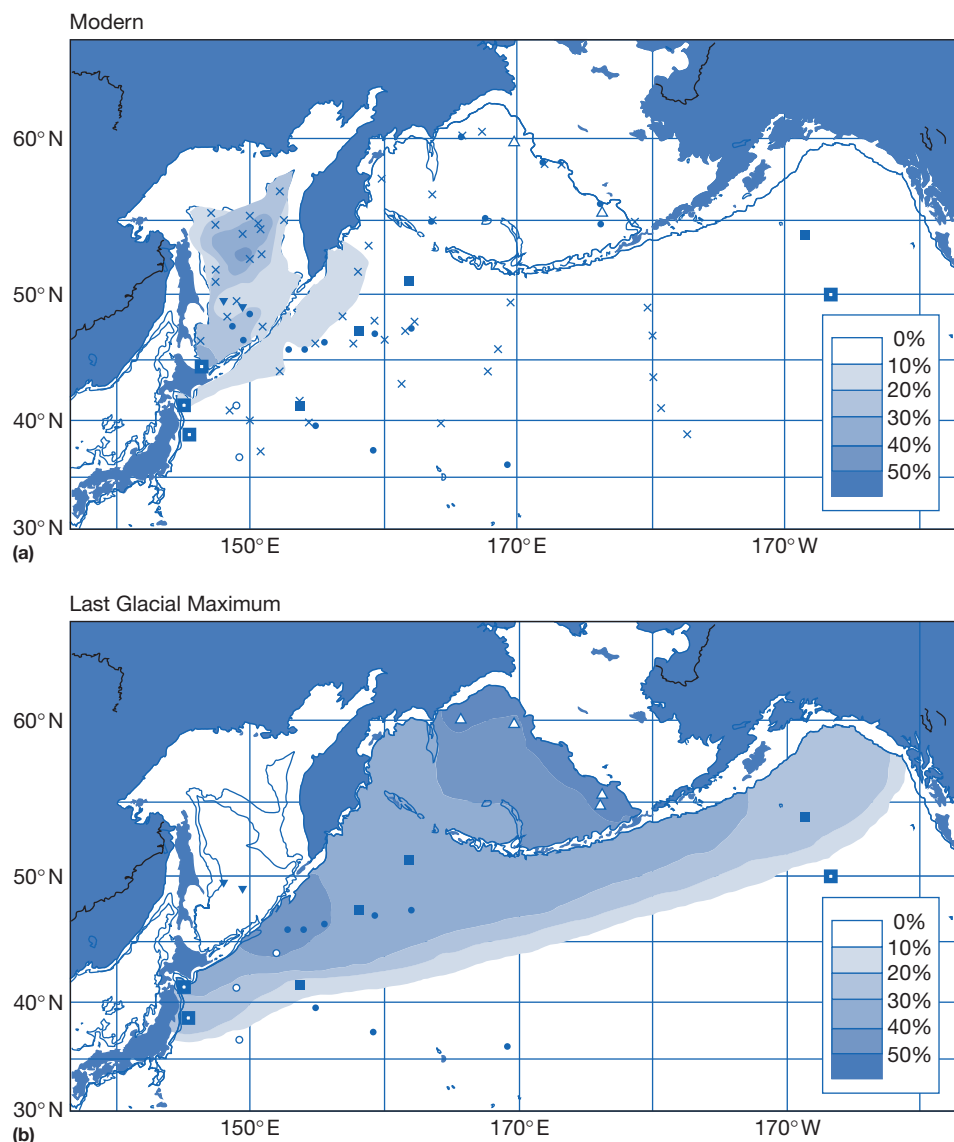


Figure 3 Comparison between (a) present and (b) last glacial distribution patterns for *Cycladophora davisiana* in the North Pacific deep-sea sediments indicating favorable conditions for deep convection. From Ohkushi K, Itaki T, and Nemoto N (2003) Last Glacial–Holocene change in intermediate-water ventilation in the Northwestern Pacific. *Quaternary Science Reviews* 22: 1477–1484.

Records of intermediate water ventilation come from the California Margin, most famously from the Santa Barbara Basin, where sediments are partially laminated. Laminations prevail in sections representing the Holocene and the interstadials, while sediment sections coeval with the cold stadials in Greenland and the North Atlantic are bioturbated (Behl and Kennett, 1996). These sedimentary structures, alongside a suite of other proxy evidence including assemblage changes of benthic Foraminifera and of redox-sensitive elements, suggest that suboxic to anoxic conditions during (the Holocene) interstadials alternated with high oxygen conditions in the Santa Barbara Basin during times of cold stadials in Greenland and the North Atlantic (Cannariato and Kennett, 1999). The combined records from the Oregon and California margins suggest

that deep and intermediate waters in the North Pacific had a similar history with respect to millennial-scale ventilation changes during the last glacial cycle.

The Surface-Water Hydrographic Record

As in the deep-water record, reconstructions of surface-water hydrography have focused on the LGM and even allowed spatial mapping of the sea-surface temperature (SST) field for that time slice (Climate Long-Range Investigation, Mapping, and Prediction (CLIMAP), Multiproxy Approach for the Reconstruction of the Glacial Ocean Surface (MARGO)). For other time periods the coverage of paleoceanographic reconstruction is fragmentary and relies on individual records.

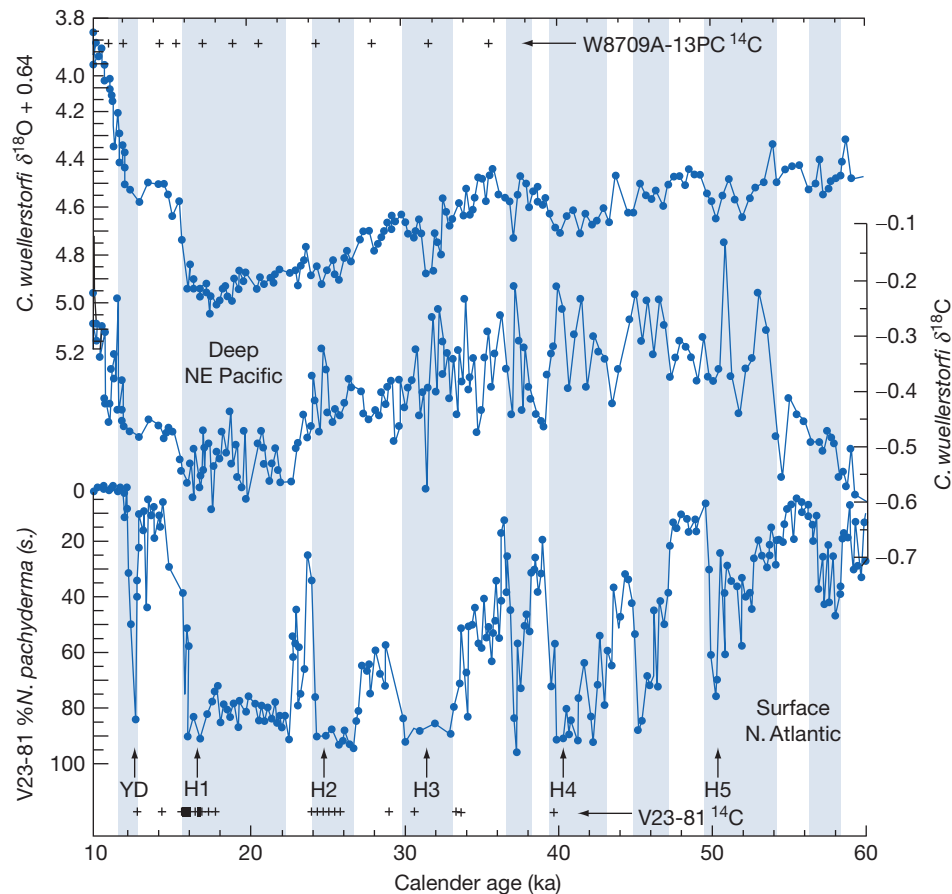


Figure 4 Benthic foraminiferal $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ in core W8709A-13A compared to the percentage of a cool-water planktonic Foraminifera species in a North Atlantic core. Shaded areas represent cold (stadial) intervals in the North Atlantic. Millennial-scale $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ maxima in the North Pacific generally correspond to stadial conditions in the Atlantic. From Lund DC and Mix AC (1998) Millennial-scale deep-water oscillations: Reflections of the North Atlantic in the deep Pacific from 10 to 60 ka. *Paleoceano-graphy* 13: 10–19.

The Last Glacial Maximum

The CLIMAP project (CLIMAP Projekt Members, 1981) produced a map of SST during the LGM (Figure 5). The reconstructions were based on faunal (mostly planktonic foraminiferal) assemblages that were processed with a statistical algorithm, a so-called transfer function. One intriguing result was that only parts of the glacial North Pacific were reconstructed as significantly colder than today, such as the sub-Arctic North Pacific, the western Pacific margin, and the equatorial Pacific, while the subtropical North Pacific was reconstructed as up to 2 °C warmer than today.

With the availability of more and better-dated data, improved statistical techniques, and additional proxies, the task was re-addressed by a joint effort called MARGO. Their Foraminifera-based reconstruction refined some of the earlier CLIMAP results (Kucera et al., 2005), such as a more southerly position of the sub-Arctic front, the contraction and maybe even cooling of the Western Pacific Warm Pool, and a substantial (>3 °C) cooling of the eastern equatorial Pacific and the southward flowing California Current off the US coast. Contrary to CLIMAP, the MARGO reconstruction does not provide conclusive evidence for LGM warming anywhere in the Pacific (Figure 6). This Foraminifera-based reconstruction is generally

consistent with alkenone-derived LGM SST (Rosell-Mele et al., 2004; Figure 7).

The shallow lysocline in the North Pacific results in an imperfect preservation of foraminifera shells in most regions and in large areas where carbonate-based reconstructions are not possible, such as for the North Pacific subtropical gyre. However, a suite of sediment cores from shallow water depths well above the lysocline from off Hawaii provides invaluable paleotemperature data (Lee et al., 2001). SST estimates from foraminiferal assemblages and from alkenones both confirm that the subtropical gyre was indeed colder by 1–3 °C during the LGM (Figure 8).

Marine Isotope Stage 5

The Earth's orbital configuration during the peak of the last interglacial (marine isotope stage 5e; Eemian, around 130–120 ka) was different from the Holocene interglacial, in particular, with summer insolation higher in the Northern Hemisphere (see *Astronomical Theory of Paleoclimates*). Accordingly, the CLIMAP reconstruction of the Eemian surface ocean found that SST in the North Pacific were predominantly higher by 1–3 °C than during the Holocene. Since then, other records with complementary methods lend further support

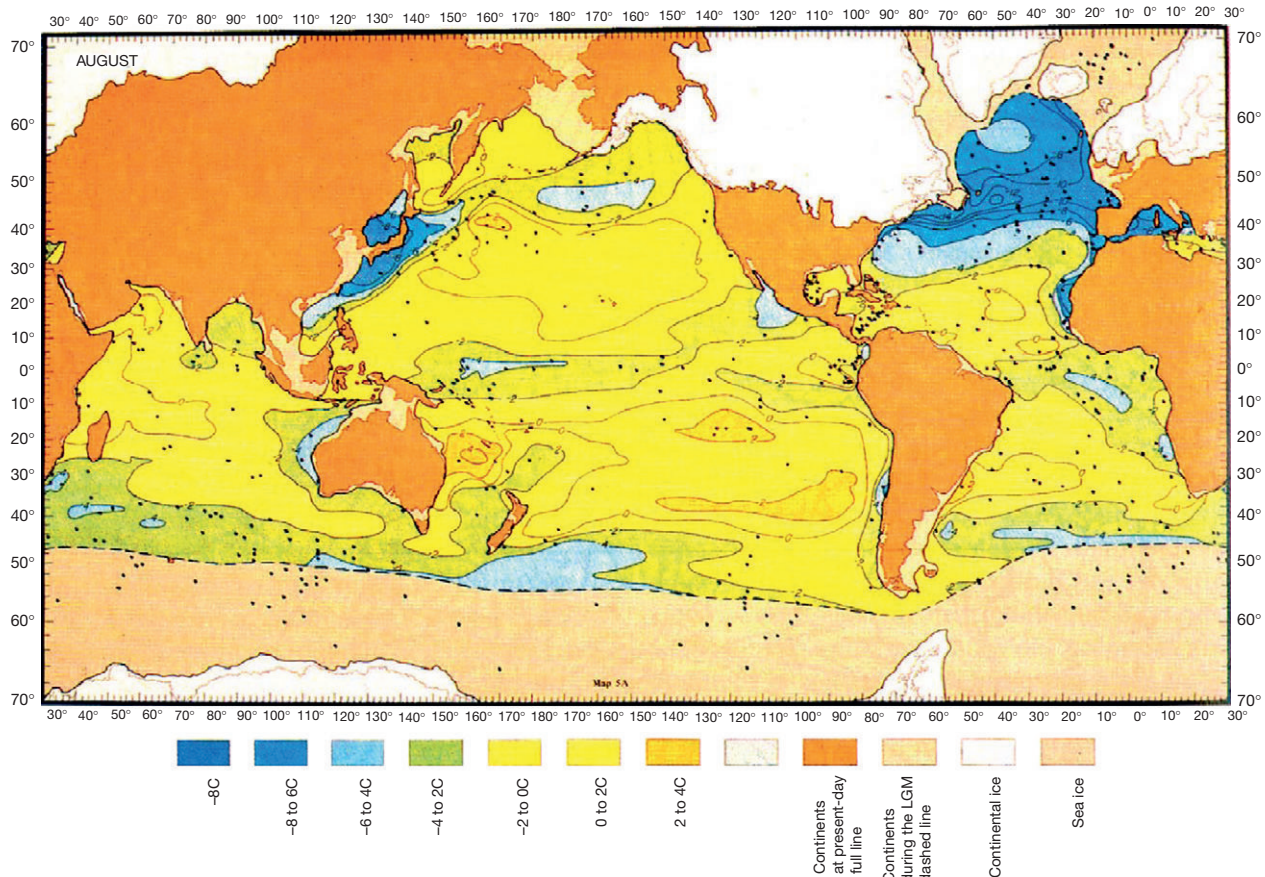


Figure 5 Map of summer SST differences between reconstructions for the Last Glacial Maximum (LGM) and for the late Holocene. Figure taken from Kucera et al. (2005). From CLIMAP Project Members (1981) Seasonal reconstructions of the Earth's surface at the Last Glacial Maximum. *Geological Society of America Map and Chart Series*: 1–18.

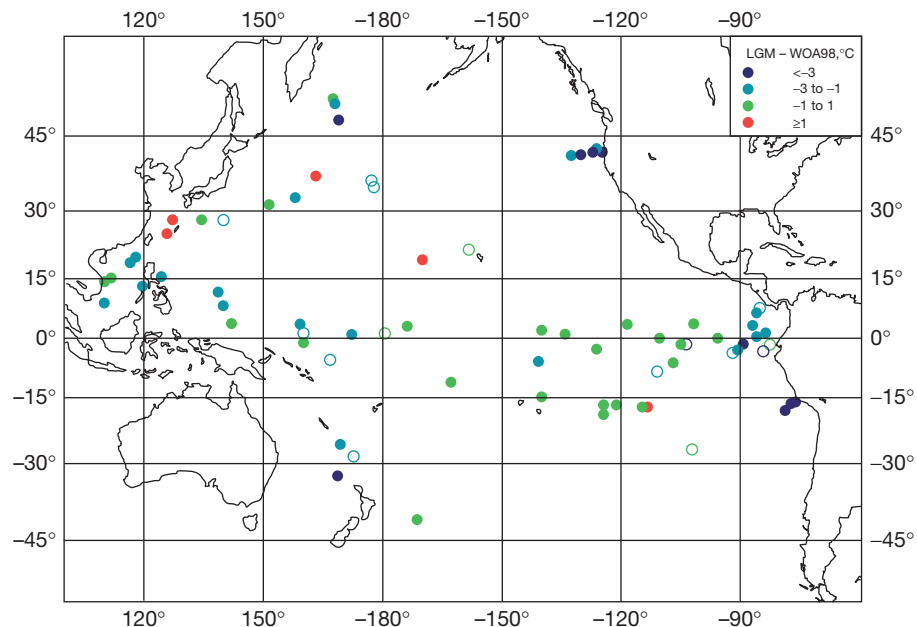


Figure 6 Differences between LGM annual SST in the Pacific estimated from foraminifer assemblages and modern measured SST (World Ocean Atlas, 1998). The SST values represent averages of the estimates from three different methods (ANN, SIMMAX, and RAM). Samples where the maximum difference in SST estimates among the four techniques is $< 2^{\circ}\text{C}$ are indicated by solid circles (highest reliability), between $2\text{--}4^{\circ}\text{C}$ by solid diamonds (medium reliability), and $> 4^{\circ}\text{C}$ by open circles (lowest reliability). Figure taken from Kucera et al. (2005).

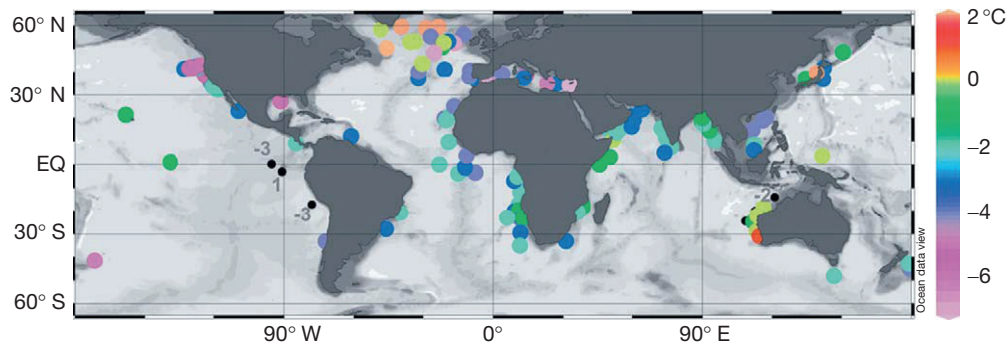


Figure 7 Change in annual mean sea-surface temperature between the Last Glacial Maximum and the present derived from alkenones using the U_{37}^K index. From Rosell-Melé A, Bard E, Emeis K-C, et al. (2004) Sea surface temperature anomalies in the oceans at the LGM estimated from the alkenone U_{37}^K index: Comparison with GCMs. *Geophysical Research Letters* 31: L03208.

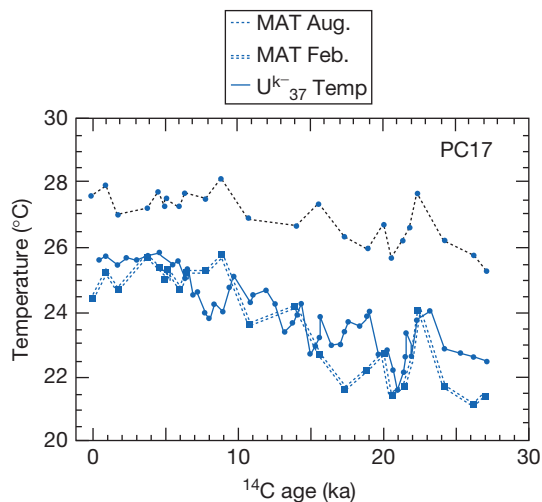


Figure 8 Comparison of sea-surface temperature (SST) variations over the last 30 ka reconstructed from sediment core PC17 off Hawaii. SST were derived from planktonic Foraminifera assemblages using the Modern Analog Technique and from alkenones using the U_{37}^K index.

that the Eemian was warmer than today. Examples come from the western equatorial Pacific including the South China Sea with at least 1 °C, from the California Current with 2–3 °C (Kreitz et al., 2000; Figure 9), and off central Japan with 3 °C higher SST. The long record from the California Current furthermore suggests that, at least regionally, the Eemian interglacial period was among the warmest periods in the North Pacific since at least the mid-Pleistocene.

Millennial-scale variability

Millennial-scale hydrographic variability during the last glacial is documented along the eastern and western North Pacific margins and in the western sub-Arctic gyre, and usually compared to concurrent changes in Greenland and the North Atlantic (Heinrich and Dansgaard-Oeschger events; see Sub-Milankovitch (DO/Heinrich) Events). In the South China Sea (Wang et al., 1999), the Sulu Sea (Dannenmann et al., 2002), and south of the Philippines (Stott et al., 2002) in the tropical–subtropical western Pacific, hydrographic changes

mostly concern sea surface salinity (SSS) as indicated by $\delta^{18}O$ of planktonic foraminiferal calcite. Western Pacific surface waters were more saline (saltier) during cold (stadial) times in the North Atlantic and fresher during times of relative Atlantic warmth (Figure 10). Various records, especially speleothems from caves in eastern China, have compellingly demonstrated that southeast Asian summer monsoon precipitation goes along with warmth in the North Atlantic. Hence, monsoonal variability is a prime candidate to explain the paleosalinity oscillations in the western Pacific. More relevant to the open ocean tropical Pacific are changes in local precipitation caused by past changes in the Walker circulation associated with ENSO variability. The paleosalinity record would imply a weakened Walker circulation associated with more frequent or stronger El Niño conditions during stadials. Validation of this scenario will require more well-dated millennial-scale records from the western and eastern equatorial Pacific, of parameters sensitive to ENSO, for example, thermocline depth and phytoplankton productivity.

Different from the paleosalinity record, millennial-scale variations of SST observed at western Pacific low latitudes were small (Dannenmann et al., 2002; Figure 10). Only one planktonic foraminiferal record from the Okinawa Trough in the East China Sea points to systematic cooling of ca. 2 °C during the time of Heinrich melt-water events in the North Atlantic (Li et al., 2001). However, in the northeastern Pacific, several paleohydrographic records off the US west coast also show millennial-scale fluctuations of remarkable resemblance with records from Greenland and the North Atlantic. In particular, the cores from the Santa Barbara Basin off California display such detailed similarity between planktonic foraminiferal isotopic and assemblage changes with the temperature changes in Greenland that quasi-synchronicity can be inferred (Hendy et al., 1999). This very important finding has promoted the awareness of paleoceanographic coupling between ocean basins and the concept of global-scale teleconnections in paleoclimatology in general. Temperature variations in the Santa Barbara Basin deduced from the planktonic $\delta^{18}O$ record reached amplitudes of up to 5 °C, or even up to 8 °C including brief temperature overshoots at the beginning of warm phases. Alkenone-derived SST variations from a nearby site with a more open-ocean setting off Point Conception were on the

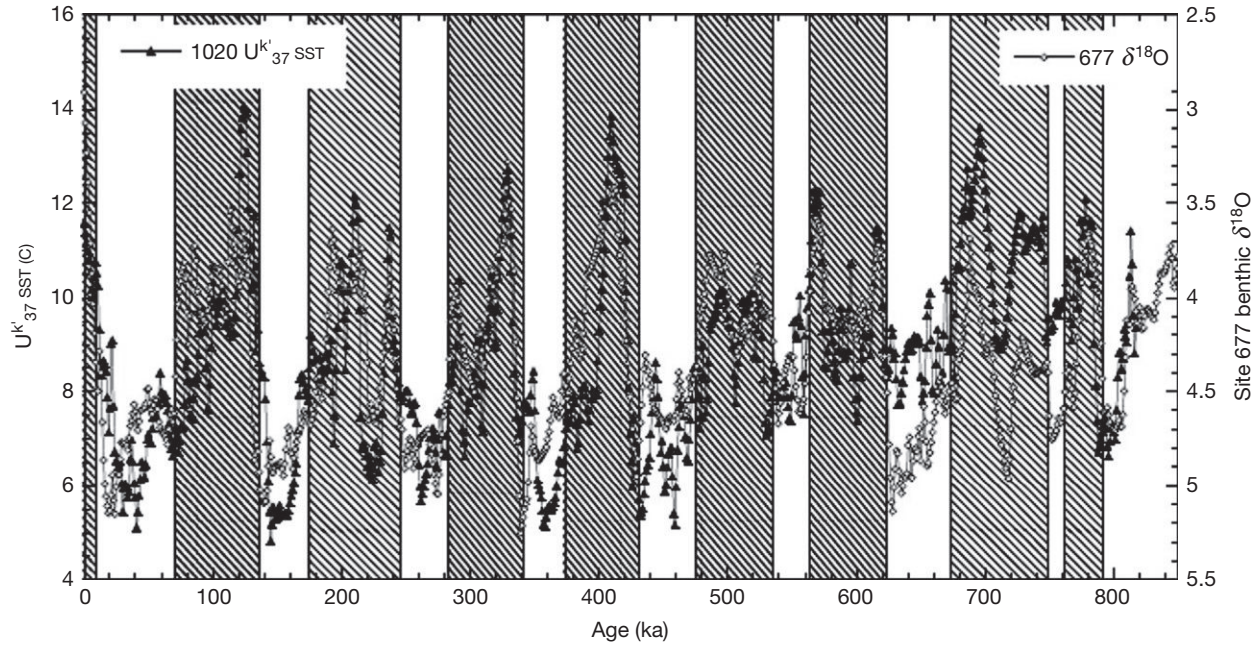


Figure 9 Alkenone SST from off California (DP Site 1020) SST and $\delta^{18}\text{O}$ from ODP site 677 for reference. Age model from tuning site 1020 to site 677 $\delta^{18}\text{O}$. Interglacial periods are shaded. From [Kreitz SF, Herbert TD, and Schuffert JD \(2000\)](#) Alkenone paleothermometry and orbital-scale changes in sea-surface temperature at site 1020, northern California margin. *Proceedings of the Ocean Drilling Program, Scientific Results* 167: 153–161.

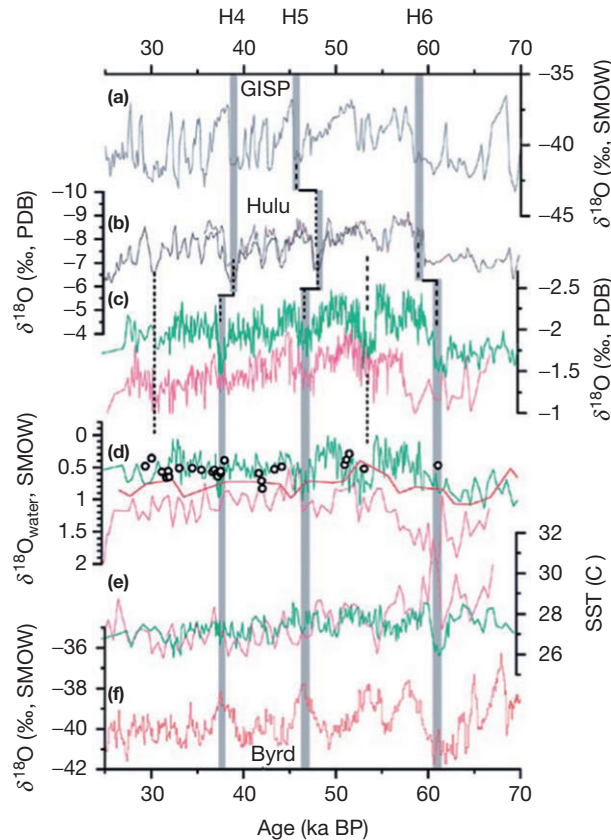


Figure 10 Comparison of the paleoclimatic records of the last glacial period from (a) Greenland and F, Antarctica (as in [Blunier and Brook, 2001](#)), and (b) the Chinese Hulu Cave ([Wang et al., 2001](#)) to (c–e) paleoceanographic records from the Sulu Sea (green; core MD97-2141; Dannenmann et al. (2004)), from Mindanao (purple; core MD97-2181; Stott et al. (2002)), and from the western equatorial Pacific (red; core ODP 806B; [Lea et al., 2000](#)). The paleoceanographic records are planktonic $\delta^{18}\text{O}$ (C), SST estimates from Mg/Ca ratios in (e) foraminiferal calcite, and seawater $\delta^{18}\text{O}$ deduced from planktonic $\delta^{18}\text{O}$ and SST estimates (d). Superimposed are sea-level data (black open circles; Chappell et al. (1996) and Yokoyama et al. (2001)). Grey bars indicate occurrences of Heinrich events H4, H5, and H6. Figure from [Dannenmann S, Linsley BK, Oppo DW, Rosenthal Y, and Beaufort L \(2003\)](#) East Asian monsoon forcing of suborbital variability in the Sulu Sea during marine isotope stage 3: Link to northern hemisphere climate. *Geochemistry, Geophysics, Geosystems* 4: 1001.

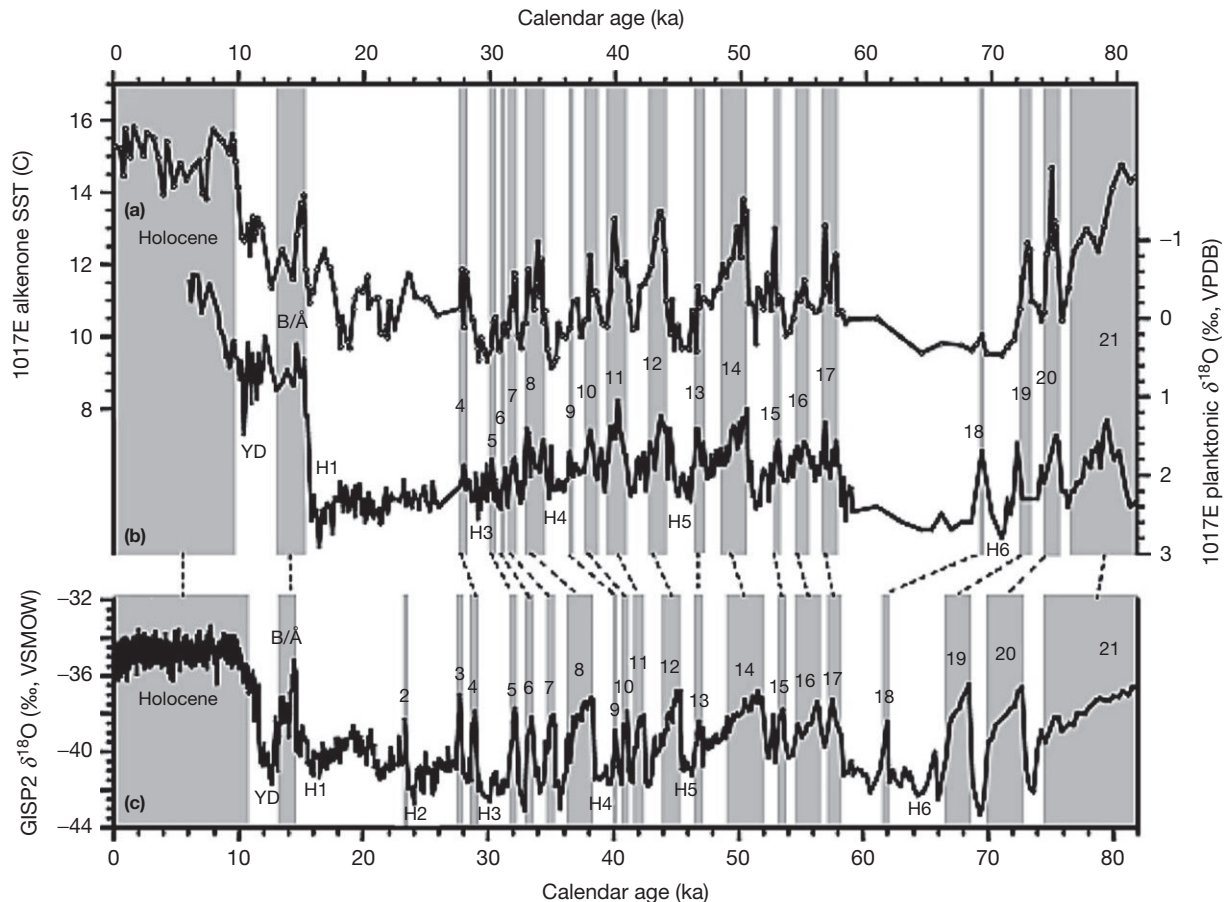


Figure 11 Paleoclimatological records of Alkenone SSTs (a) and $\delta^{18}\text{O}$ of planktonic Foraminifera (b) for Hole 1017E (ODP Leg 167) on the California continental margin, and $\delta^{18}\text{O}$ record of Greenland Ice (c). Interstadials (shading areas) are numbered according to the GISP2 scheme. Heinrich events are marked by H symbols with their numbers. YD = Younger Dryas. B/Å = Bölling-Allerød. Figure from Seki O, Ishiwatari R, Matsumoto K (2002) Millennial climate oscillations in NE Pacific surface waters over the last 82 ka: New evidence from alkenones. *Geophysical Research Letters* 29: 59-1-4.

order of 3–4 °C (Seki et al., 2002; Figure 11). Further north, from the California Current off Oregon, SST records based on radiolarian assemblages also display millennial-scale variations of 2–3 °C that can be correlated to the Greenland temperature pattern (Pisias et al., 2001). The ongoing discussion about mechanisms for the Atlantic–Pacific signal transfer include mid-latitude atmospheric circulation changes due to an oscillating morphology of the Laurentide ice sheet, and oceanic circulation changes in response to Atlantic–Pacific water vapor transport or far-field effects of global oceanic overturning. One single record from the sub-Arctic North Pacific – from the western gyre – also reflects millennial-scale hydrographic changes during the last glacial cycle (Kiefer et al., 2001). A multi-proxy SST record based on faunal composition, $\delta^{18}\text{O}$ and Mg/Ca ratios of planktonic Foraminifera suggests that SST varied by at least 3 °C during marine isotope stage 3 (Figure 12). In contrast to the more southerly sites, however, SST was apparently out of phase with the Greenland/North Atlantic temperature pattern. The stratigraphy of this core relies on the Pacific–Atlantic correlation of characteristics of the geomagnetic record that are assumed to be globally synchronous (see [Geomagnetic Excursions and Secular Variations](#)). An out-of-phase temperature pattern in the subarctic Pacific

suggests a dominant oceanic control on SST, such as a shallowing of the mixed layer in response to basin-wide changes in ocean circulation or changes in the local hydrography.

The Surface-Water Biogeochemical Record

Glacial–interglacial changes

In order to understand the role of the North Pacific in the sequestration or emission of atmospheric CO_2 , it is important to develop our knowledge of past changes in nutrient status and phytoplankton productivity. The modern sub-Arctic North Pacific and the central eastern equatorial Pacific are both ocean regions of open-ocean upwelling where supply of nutrients is not readily utilized for phytoplankton production, making those regions net sources for CO_2 to the atmosphere. It is important to reconstruct past CO_2 budgets of these regions and learn whether these regions were past sinks or sources for atmospheric CO_2 . Late Pleistocene records of phytoplankton productivity changes reveal a large-scale pattern of glacial–interglacial changes with opposing patterns between open ocean regions of the subarctic and the tropical–subtropical North Pacific. In the eastern and central tropical Pacific, Barite accumulation records over the last 3–4 glacial cycles (350 and

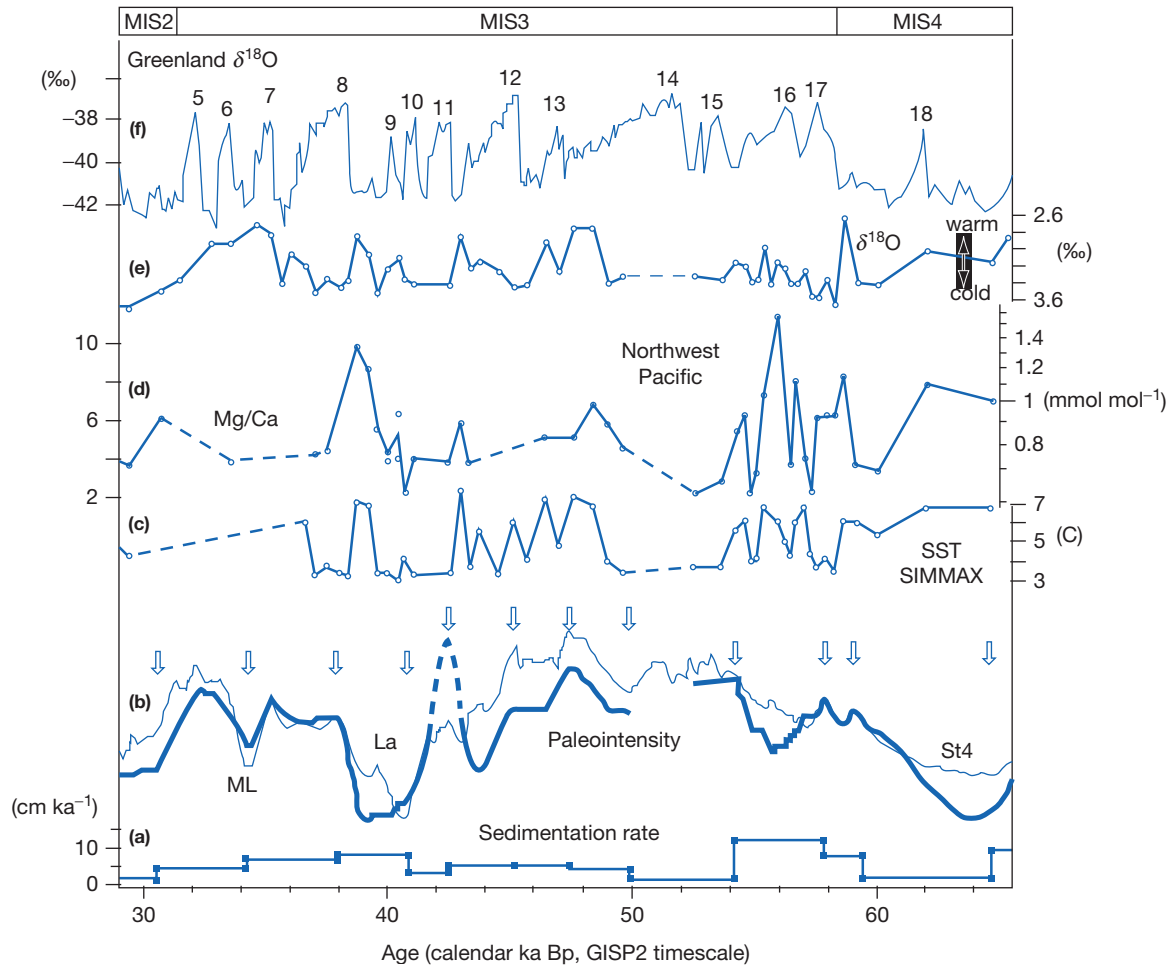


Figure 12 (c–e) Sea-surface temperature (SST), (a) sedimentation rates, and (b) geomagnetic paleointensity records at sub-Arctic northwest Pacific ODP site 883 between 29 and 65 cal ka BP compared to (f) the GISP2 $\delta^{18}\text{O}$ temperature record. From Kiefer et al. (2002).

450 ka) show that the export of fresh organic matter (export productivity) was about twice as high during glacials than during interglacials (Paytan et al., 1996). Most other proxies support this observation (Paytan et al. (2004), and references therein). This implies equatorial upwelling of nutrient-rich subsurface waters driven by stronger glacial trade winds or increased nutrient utilization and changes in ecosystem structure. The increased upwelling may have been a significant glacial source of CO_2 to the atmosphere, although the balance between export production and upwelling is what determines the new CO_2 flux. Higher glacial productivity is also repeatedly reported from sites below the Kuroshio Extension in the sub-Arctic–subtropical transition zone and often attributed to increased glacial dust fertilization, while a southward shift of the transition zone and associated southward expansion of nutrient-rich sub-Arctic surface waters provides an additional explanation. In the western sub-Arctic Pacific and its marginal seas, the Okhotsk Sea and Bering Sea, export productivity was decreased during glacial periods. A record of biogenic Barium concentration strongly suggests that export production fluctuated in parallel with temperatures over Antarctica and the Southern Ocean (Jaccard et al., 2005; Figure 13). Records

available from the sub-Arctic eastern Pacific do not reveal a clear pattern. Better stratification of the upper ocean inhibiting nutrient supply from below, extensive sea ice, and light limitation were all suggested as possible causes for low glacial productivity and for counteracting any possible iron fertilization in these domains (see review of Kienast et al. (2004)). If the glacial North Pacific was indeed better stratified, this would imply that the outgassing of CO_2 during the process of subsurface-water upwelling was reduced or prevented, hence contributing to the lower glacial concentrations of CO_2 in the atmosphere.

Millennial-scale variability

Evidence for millennial-scale productivity changes in concert with the Dansgaard–Oeschger temperature oscillations over Greenland (see Sub-Milankovitch (DO/Heinrich) Events) comes from the California margin cores (see Section ‘Millennial-scale variability’), where increased export productivity during interstadials was proposed to have contributed to or even to account for the low bottom-water oxygen content in the Santa Barbara Basin at intermediate water depths. Alternation between predominant La Niña conditions with a

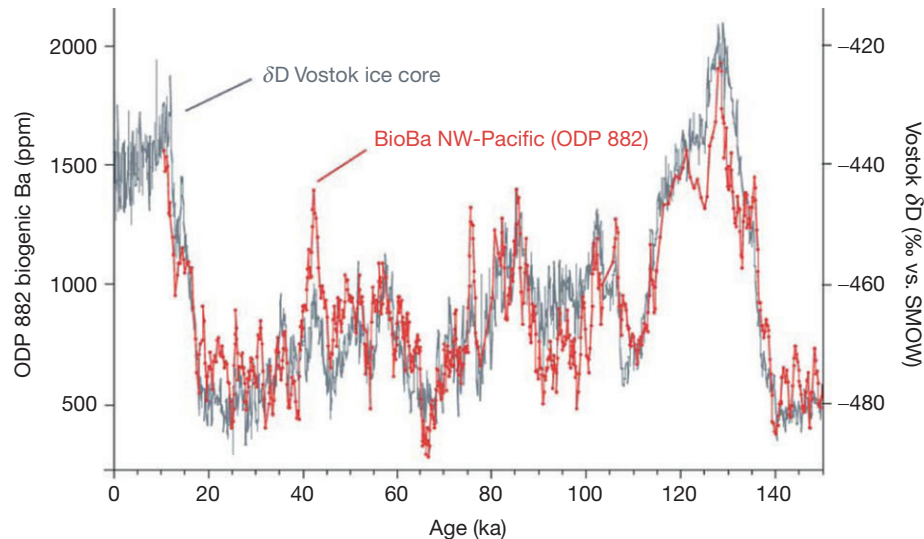


Figure 13 Biogenic barium concentration at North Pacific ODP site 882 (red) and the Antarctic ice core δD record at Vostok station (gray) over the past 150 ka, illustrating the strong correlation between productivity in the North Pacific and temperatures over Antarctica. From Jaccard SL, Haug GH, Sigman DM, Pedersen TF, Thierstein HR, and Rohl U (2005) Glacial-interglacial changes in subarctic North Pacific stratification. *Science* 308: 1003–1006.

shallow nutricline during warm intervals on the one hand, and predominant El Niño conditions with a deep nutricline hampering algal productivity during cold intervals on the other hand, was discussed as a plausible explanation. In contrast to this, the record of biogenic Barium from the western sub-Arctic Pacific correlates surprisingly well with temperatures over Antarctica (Jaccard et al., 2005). The proposed control mechanism for productivity involves increased polar stratification inhibiting the supply of nutrients from below into the sunlit zone during global cold times. However, at present it is still not clear how the temperature signal was transferred from the Southern Hemisphere to the Northern Hemisphere high latitude Pacific.

See also: Geomagnetic Excursions and Secular Variations.
Glaciation, Causes: Astronomical Theory of Paleoclimates.
Glaciations: Late Pleistocene Glacial Events in Beringia.
Paleoceanography, Physical and Chemical Proxies: Carbon Cycle Proxies ($\delta^{11}\text{B}$, $\delta^{13}\text{C}_{\text{calcite}}$, $\delta^{13}\text{C}_{\text{organic}}$; Shell Weights, B/Ca, U/Ca, Zn/Ca, Ba/Ca); Dissolution of Deep-Sea Carbonates; Mg/Ca and Sr/Ca Paleothermometry from Calcareous Marine Fossils; Nutrient Proxies.
Paleoceanography, Records: Late Pleistocene North Atlantic; Postglacial North Pacific. **Paleoclimate Reconstruction:** Sub-Milankovitch (DO/Heinrich) Events.

References

- Adcroft A, Scott JR, and Marotzke J (2001) Impact of geothermal heating on the global ocean circulation. *Geophysical Research Letters* 28: 1735–1738.
- Behl RJ and Kennett JP (1996) Brief interstadial events in the Santa Barbara basin, NE Pacific, during the past 60 kyr. *Nature* 379: 243–246.
- Blunier T and Brook EJ (2001) Timing of millennial-scale climate change in Antarctica and Greenland during the last glacial period. *Science* 291: 109–112.
- Broecker WS (1982) Ocean chemistry during glacial time. *Geochimica et Cosmochimica Acta* 46: 1689–1705.
- Broecker WS and Clark E (2001) Glacial-to-Holocene Redistribution of Carbonate Ion in the Deep Sea. *Science* 294: 2152–2155.
- Broecker WS, Barker S, Clark E, Hajdas I, Bonani G, and Stott L (2004a) Ventilation of the glacial deep Pacific Ocean. *Science* 306: 1169–1172.
- Broecker WS, Clark E, Hajdas I, and Bonani G (2004b) Glacial ventilation rates for the deep Pacific Ocean. *Paleoceanography* 19: PA2002.
- Cannariato KG and Kennett JP (1999) Climatically related millennial-scale fluctuations in strength of California margin oxygen-minimum zone during the past 60 k.y. *Geology* 27: 975–978.
- CLIMAP Projekt Members (1981) Seasonal reconstructions of the Earth's surface at the last glacial maximum. *Geological Society of America Map and Chart Series* MC-36, 1–18.
- Dannenmann S, Linsley BK, Oppo DW, Rosenthal Y, and Beaufort L (2003) East Asian monsoon forcing of suborbital variability in the Sulu Sea during marine isotope stage 3: Link to northern hemisphere climate. *Geochemistry, Geophysics, Geosystems* 4: 1001.
- Duplessy J-C, Shackleton NJ, Fairbanks RG, Labeyrie L, Oppo D, and Kallel N (1988) Deepwater source variations during the last climatic cycle and their impact on the global deepwater circulation. *Paleoceanography* 3: 343–360.
- Hendy IL and Kennett JP (1999) Latest Quaternary North Pacific surface-water responses imply atmosphere-driven climate instability. *Geology* 27: 291–294.
- Herbert TD, Schuffert JD, Andreasen D, et al. (2001) Collapse of the California Current during glacial maxima linked to climate change on land. *Science* 293: 71–76.
- Jaccard SL, Haug GH, Sigman DM, Pedersen TF, Thierstein HR, and Rohl U (2005) Glacial/interglacial changes in subarctic North Pacific stratification. *Science* 308: 1003–1006.
- Keigwin LD (1998) Glacial-age hydrography of the far northwest Pacific Ocean. *Paleoceanography* 13: 323–339.
- Kiefer T, Lorenz S, Schulz M, Lohmann G, Sarnthein M, and Elderfield H (2002) Response of precipitation over Greenland and the adjacent ocean to North Pacific warm spells during Dansgaard-Oeschger stadials. *Terra Nova* 14: 295–300.
- Kiefer T, Sarnthein M, Erlenkeuser H, Grootes P, and Roberts A (2001) North Pacific response to millennial-scale changes in ocean circulation over the last 60 ky. *Paleoceanography* 16: 179–189.
- Kienast SS, Hendy IL, Crusius J, Pedersen TF, and Calvert SE (2004) Export production in the subarctic North Pacific over the last 800 kyrs: No evidence for iron fertilization? *Journal of Oceanography* 60: 189–203.
- Kreitz SF, Herbert TD, and Schuffert JD (2000) Alkenone paleothermometry and orbital-scale changes in sea-surface temperature at site 1020, northern California margin. *Proceedings of the Ocean Drilling Program, Scientific Results* 167: 153–161.
- Kucera M, Weinelt M, Kiefer T, et al. (2005) Reconstruction of sea-surface temperatures from assemblages of planktonic foraminifera: Multi-technique

- approach based on geographically constrained calibration data sets and its application to glacial Atlantic and Pacific Oceans. *Quaternary Science Reviews* 24: 951–998.
- Lea DW, Pak DK, and Spero HJ (2000) Climate impact of late Quaternary equatorial Pacific sea surface temperature variations. *Science* 289: 1719–1724.
- Lee KE, Slowey NC, and Herbert TD (2001) Glacial SSTs in the subtropical North Pacific: A comparison of U k1/37 delta 18O and foraminiferal assemblage temperature estimates. *Paleoceanography* 16: 268–279.
- Li T, Liu Z, Hall MA, et al. (2001) Heinrich event imprints in the Okinawa Trough: Evidence from oxygen isotope and planktonic foraminifera. *Palaeogeography, Palaeoclimatology, Palaeoecology* 176: 133–146.
- Lund DC and Mix AC (1998) Millennial-scale deep water oscillations: Reflections of the North Atlantic in the deep Pacific from 10 to 60 ka. *Paleoceanography* 13: 10–19.
- Matsumoto K, Oba T, Lynch-Stieglitz J, and Yamamoto H (2002) Interior hydrography and circulation of the glacial Pacific Ocean. *Quaternary Science Reviews* 21: 1693–1704.
- Ohkouchi N, Kawahata H, Murayama M, Okada M, Nakamura T, and Taira A (1994) Was deep water formed in the North Pacific during the late Quaternary? Cadmium evidence from the northwest Pacific. *Earth and Planetary Science Letters* 124: 185–194.
- Ohkushi K, Itaki T, and Nemoto N (2003) Last Glacial–Holocene change in intermediate-water ventilation in the Northwestern Pacific. *Quaternary Science Reviews* 22: 1477–1484.
- Paytan A, Kastner M, and Chavez FP (1996) Glacial to interglacial fluctuations in productivity in the equatorial Pacific as indicated by marine barite. *Science* 274: 1355–1357.
- Paytan A, Lyle M, Mix A, and Chase Z (2004) Climatically driven changes in oceanic processes throughout the equatorial Pacific. *Paleoceanography* 19: PA4017.
- Pisias NG, Mix AC, and Heusser L (2001) Millennial scale climate variability of the northeast Pacific Ocean and northwest North America based on radiolaria and pollen. *Quaternary Science Reviews* 20: 1561–1576.
- Rosell-Melé A, Bard E, Emeis K-C, et al. (2004) Sea surface temperature anomalies in the oceans at the LGM estimated from the alkenone-U[37]vK' index: Comparison with GCMs. *Geophysical Research Letters* 31: L03208.
- Seki O, Ishiwatari R, and Matsumoto K (2002) Millennial climate oscillations in NE Pacific surface waters over the last 82 kyr: New evidence from alkenones. *Geophysical Research Letters* 29(59): 2144.
- Shackleton NJ, Duplessy J-C, Arnold M, Maurice P, Hall MA, and Cartlidge J (1988) Radiocarbon age of last glacial Pacific deep water. *Nature* 335: 708–711.
- Stoot L, Poulsen C, Lund S, and Thunell R (2002) Super ENSO and global climate oscillations at millennial time scales. *Science* 297: 222–226.
- Stott LD, Neumann M, and Hammond D (2000) Intermediate water ventilation on the northeastern Pacific margin during the late Pleistocene inferred from benthic foraminiferal $\delta^{13}\text{C}$. *Paleoceanography* 15: 161.
- Wang YJ, Cheng H, Edwards RL, et al. (2001) A high-resolution absolute-dated late Pleistocene monsoon record from Hulu Cave, China. *Science* 294: 2345–2348.
- Wang L, Sarnthein M, Erlenkeuser H, et al. (1999) East Asian monsoon climate during the Late Pleistocene: High-resolution sediment records from the South China Sea. *Marine Geology* 156: 245–284.

Postglacial Indian Ocean

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Introduction

The summer monsoon is the dominant climatic feature of the Indian Ocean tropics and the adjacent continent. Boreal summer is characterized by high solar radiation that causes intense heating over northern India and the Tibet Plateau. This pattern of heating causes ascending air flow and the development of an intense low-pressure cell that is centered over Asia around 30° N (Figure 1). The atmospheric pressure gradient between the Asian Continent and the cooler southern Indian Ocean induces large-scale meridional overturning with the lower circulation limb being the strong low-level southwesterly summer monsoon winds of the western Indian Ocean. The convergence of these air masses and their uplift due to heating and orographic steering cause seasonal monsoon rains (Figure 2). By contrast, the winter season of the Asian sector is characterized by low solar radiation, cold temperature, and northeasterly winds, which flow from the cold Asian continent toward the Arabian Sea. These continental winter monsoon winds carry little moisture and have relatively low velocity. Monsoon rainfall in India is highly variable. The northeastern region, the western slopes of the Western Ghats, and parts of the Himalayas have very high rainfall of around 2000 mm annually. The eastern part of the peninsula extending up to the northern plain receive around 1000–2000 mm rainfall, while the area from western Deccan up to the Plains of the Punjab gets around 100–500 mm rainfall. Thus the unique monsoon circulation in the Indian Ocean and associated rainfall over Asia have two fundamental regional effects: (1) socioeconomic and agricultural development in the densely populated Asian countries and (2) the biogeochemistry of the Indian Ocean sediments.

Pre- and Postglacial Changes

The last deglaciation was not a smooth transition from one climate state to another but rather occurred in a series of steps. Recent evidences suggest that the deglacial warming was initiated ~19 ka BP in the Indian Ocean and alike $\delta^{18}\text{O}$ trend prevailed in the Antarctica and Indian Ocean record (Figure 3; Naidu and Govil, 2010). The $\delta^{18}\text{O}$ records of the Indian Ocean exhibit a two-step deglaciation model that included two phases of rapid ice melting separated by a brief period of climate stability (Figure 4; Peeters et al., 1999); however, sea surface temperature (SST) show a gradual warming during deglaciation. Earlier it was believed that only in the circum-Atlantic region deglaciation was interrupted by a brief return to colder conditions, the so-called Younger Dryas event centered at about 12 ka BP. A Younger Dryas event is also documented in the $\delta^{18}\text{O}$ records of the Indian Ocean sediment records, but conspicuous cooling was not noticed during Younger Dryas in the Indian Ocean.

The carbon and oxygen isotope records of benthic Foraminifera show that the water-column structure of the Indian

Ocean during the Last Glacial Maximum (LGM) was marked by the presence of a deep front separating intermediate and deepwater masses with very different characteristics (Kallel et al., 1988). The intermediate water temperature and carbon isotope values were similar to those of today. By contrast, the deepwater was cooler than now by at least 1.5 °C, more depleted in ^{13}C , and poorly oxygenated. The vertical salinity/ $\delta^{18}\text{O}$ gradient in the northern Indian Ocean during the LGM was similar to the modern one. The sharp deep front between intermediate and deepwater is thus best explained by the decrease in deepwater temperature in the Indian Ocean and development of a deep thermocline. The hydrological boundary between deep and intermediate waters became suppressed after deglaciation.

Salinity Contrast

Oxygen isotopic ratios ($\delta^{18}\text{O}$) of *Globigerinoides ruber* in the northern Indian Ocean reveal a pattern similar to that of salinity conditions. In the Bay of Bengal, the $\delta^{18}\text{O}$ values decrease progressively toward the mouth of the Ganges, Brahmaputra, Irrawaddy River system, clearly reflecting the salinity pattern. The general north–south $\delta^{18}\text{O}$ increase in the Arabian Sea also fits the salinity pattern and reflects the strong evaporation excess in this basin (Figure 5(a)). The $\delta^{18}\text{O}$ difference between glacial and Holocene indicates a sharp salinity contrast from the glacial to Holocene in the Bay of Bengal (Figure 5(b)). During the Holocene from south to north, *G. ruber* $\delta^{18}\text{O}$ values decrease by 1.1‰ in the Bay of Bengal and increase by 0.8‰ in the Arabian Sea, whereas mean annual sea surface salinity decreases by 3 PSU in the Bay of Bengal and increases by 1.5 PSU in the Arabian Sea (Duplessy, 1982). During the LGM, reduction in the $\delta^{18}\text{O}$ gradient observed in the Bay of Bengal and the Andaman Sea is associated with a decrease in river discharge due to weaker southwest monsoon.

A marked increase in *G. ruber* $\delta^{18}\text{O}$ gradient is observed in the Arabian Sea. This strong glacial isotope gradient, which is larger than 1‰, is not related to the geographical distribution of the SST which is more uniform than today but indicates a strong salinity increase from south to north due to more evaporation. The difference between the oxygen isotope ratios of planktic Foraminifera of LGM and Holocene (Figure 5(b)) shows that the southwest monsoon was weaker than today but that the northeast monsoon was stronger. However, it would be difficult to interpret the $\delta^{18}\text{O}$ of planktic foraminifer as a proxy of monsoon precipitation in the Indian Ocean, as $\delta^{18}\text{O}$ of planktonic Foraminifera is influenced by temperature and the $\delta^{18}\text{O}$ of seawater where the latter covaries with salinity and, on millennial time scales, global ice volume. Subsequently, these problems are circumvented and $\delta^{18}\text{O}_{\text{sw}}$ were reconstructed by removing the effect of temperature (temperature estimates by Mg/Ca ratios in foraminifer) and global ice volume effect from the $\delta^{18}\text{O}$ of planktic Foraminifera. Such studies from

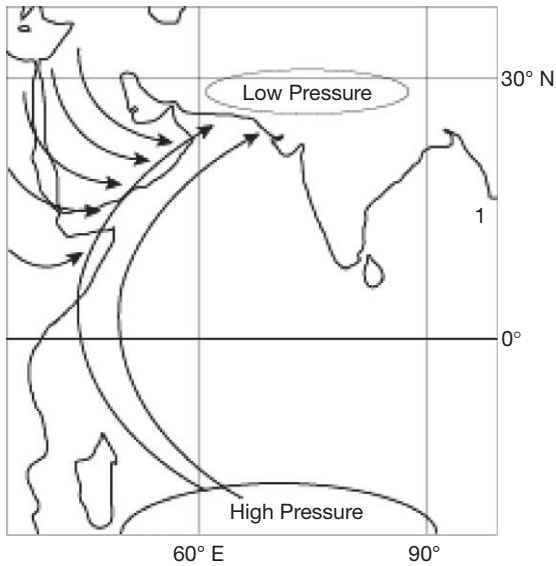


Figure 1 Atmospheric circulation during the summer months, regional wind patterns, north–south pressure gradient and cyclonic circulation around the Asian low-pressure center. Modified from Clemens S, et al. (1991) Forcing mechanisms of the Indian Ocean monsoon. *Nature* 353: 720–725.



Figure 2 Pattern of rainfall during peak monsoon along the Western Ghats of India.

the Arabian Sea show heavy $\delta^{18}\text{O}_{\text{sw}}$ values during the last glacial period than the present day reflecting more evaporation and less precipitation (Anand et al., 2008; Dahl and Oppo, 2006). Similarly, $\delta^{18}\text{O}_{\text{sw}}$ variability in the Bay of Bengal (Govil and Naidu, 2011) and Andaman Sea (Rashid et al., 2007) also points out that Indian Ocean monsoon intensity was stronger during early Holocene and weaker during last glaciation.

SST Changes

SST estimates based on the faunal composition, alkenones, and magnesium/calcium ratios show about 3–4 °C regional cooling in the Indian Ocean during the glacial maximum as compared to the Holocene. Seasonal SST reconstructions show

that summer and winter SST were 1.2 and 2.6 °C cooler, respectively, during the LGM than in the Holocene. Earlier SST estimates based on the alkenones in the Arabian Sea document a rather stable SST changes in the range of 27.1–27.7 °C during the Holocene (Rostek et al., 1998). Recent reconstructions of SST variations based on the Mg/Ca ratios in planktic foraminifer document ~2 °C fluctuations during the Holocene in the Bay of Bengal and Arabian Sea (Naidu and Govil, 2010). Similarly, the monsoon upwelling regions of the western Arabian Sea also shows a temperature change of 2 °C during the Holocene (Naidu and Malmgren, 2005; Saher et al., 2007). In general, both summer and winter SST estimates show greater magnitude of fluctuations during the last glaciation compared to the Holocene (Figure 6). However, within the Holocene, a 2 °C change in summer SST has occurred mainly due to the variation of upwelling intensity. In the tropical Indian Ocean, last deglacial warming was initiated between 18 and 19 ka. Before the onset of deglacial warming, 20 ka BP is marked with a rise of 1–2 °C temperature (Figure 7). Overall, a 4 °C rise in SST from 19 to 14 ka BP documented in the Indian Ocean and significant cooling was not noticed during Younger Dryas.

Recently Ganssen et al. (2011) have analyzed $\delta^{18}\text{O}$ analyses on individual specimens of planktonic foraminifer species (*G. ruber* and *Globigerina bulloides*) from a sediment core off the Somalia, which reveals that seasonal SST has fluctuated between 13 and 11 °C, with greater seasonal SST change in the Holocene than in last glacial period.

Outflows into the Indian Ocean During the Deglaciation

Western Pacific, Red Sea, and Persian Gulf waters flow into the Indian Ocean as surface water outflow at the sill depths ranging from 100 to 300 m. These three outflow waters have significant influence on the hydrography at the thermocline depth in the Indian Ocean. The outflow has varied through time depending on the sea-level changes. The outflow of western Pacific, Red Sea, and Persian Gulf waters was suppressed during the LGM due to lower sea levels. A vigorous flow of Red Sea outflow into the Arabian Sea was reconstructed for 2 ka BP before a major monsoon intensification caused the cessation of deepwater formation from 15.5 to 7.3 ka BP (Figure 8; Naqvi and Fairbanks, 1996). Similarly, during glacial periods, the Bab el Mandab Strait was reduced to <20 m in depth or even completely closed off so that advection of Red Sea waters to the Arabian Sea was drastically reduced, if not completely cut off.

Monsoon Variability

Various biogeochemical proxies have been used to reconstruct the summer monsoon variability over geological time scales. This section discusses the variability of monsoon during the postglacial period (mainly the Holocene). After a weak summer monsoon activity during the LGM, the initiation of strengthening monsoon activity began around 15 ka BP, which coincides with the Northern Hemisphere deglaciation. Monsoon activity further increased from 12 ka BP and reached maximum intensity between 9 and 6 ka BP. Thereafter, it decreased from 5 to 1.2 ka BP (Figure 9; Naidu and Malmgren, 1996). As a consequence of strong monsoon rainfall, the

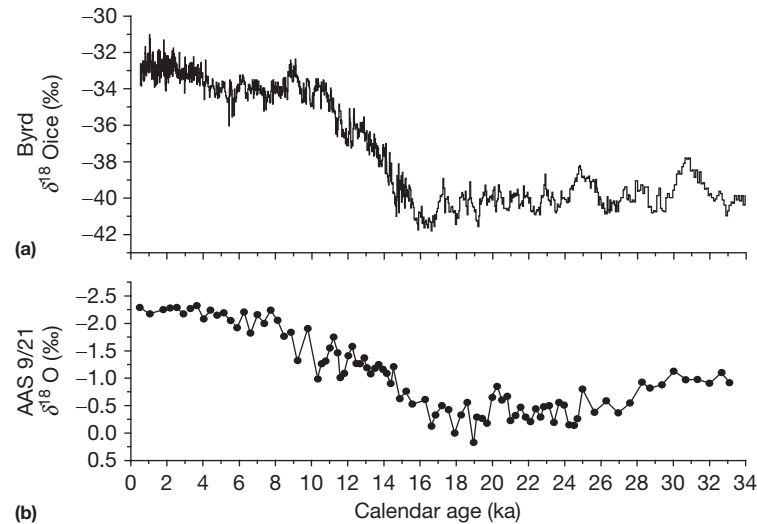


Figure 3 Similarities between (a) $\delta^{18}\text{O}$ of ice from Antarctica Ice Core Record and (b) $\delta^{18}\text{O}$ of planktic foraminifer from the northern Indian Ocean during the deglaciation. Modified from [Naidu PD and Govil P \(2010\)](#) New evidence on the sequence of deglacial warming in the tropical Indian Ocean. *Journal of Quaternary Science* 25: 1138–1143.

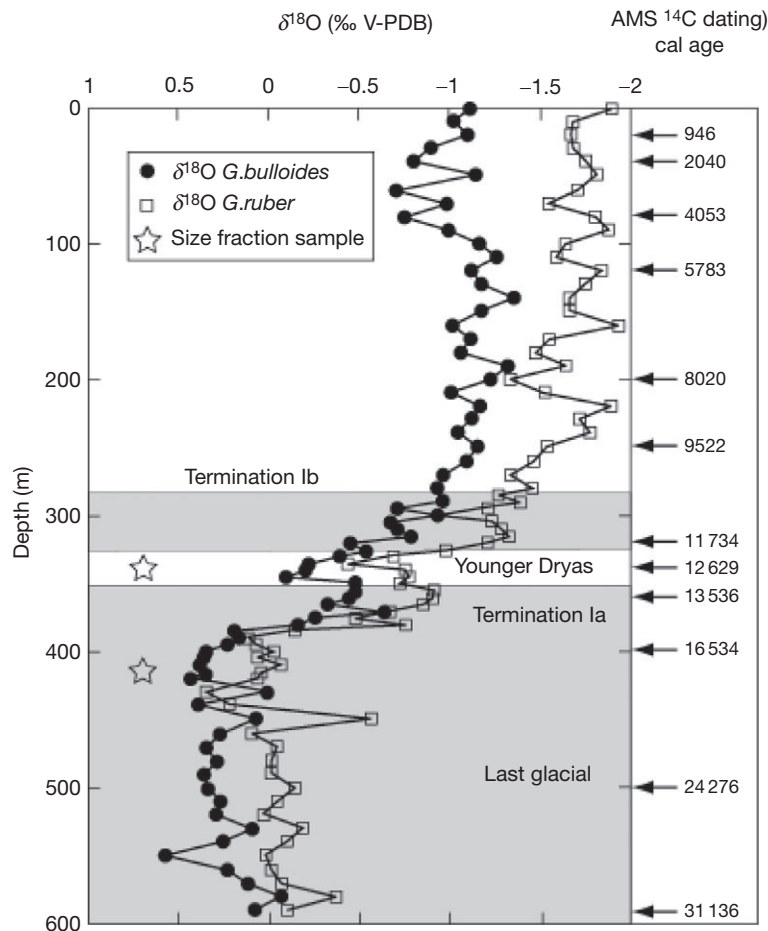


Figure 4 Oxygen isotopic curve of a core from the western Arabian Sea showing two-step deglaciation. Reproduced from [Peeters F, et al. \(1999\)](#) A size analysis of planktic Foraminifera from the Arabian Sea. *Marine Micropaleontology* 36: 31–63.

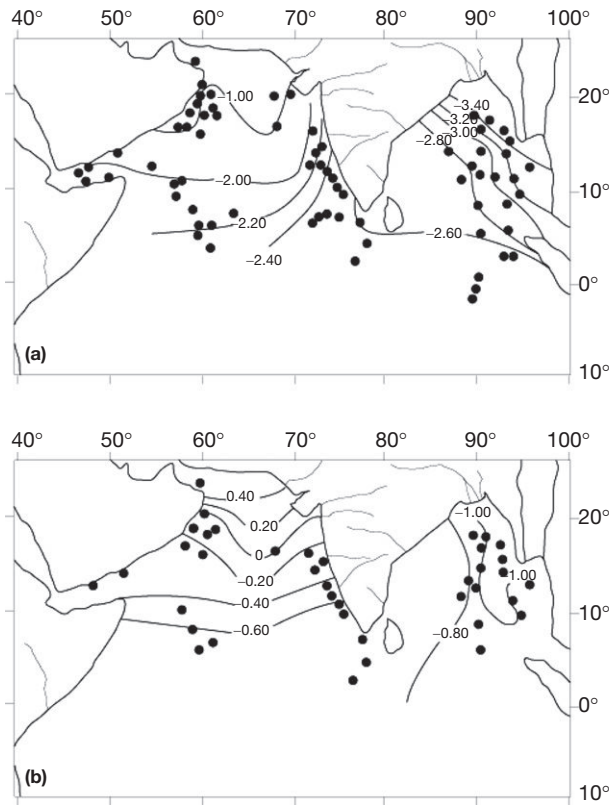


Figure 5 (a) Variation of oxygen isotope ratios during the Holocene. (b) Variations in oxygen isotopic differences between Holocene and Last Glacial Maximum in the northern Indian Ocean. Modified from Duplessy JC (1982) Glacial to interglacial contrasts in the northern Indian Ocean. *Nature* 295: 494–498.

outflow of the rivers draining the Himalayan Mountains increased at the end of the deglaciation, and it was larger than today during early and mid-Holocene. About $5 \times 10^{12} \text{ m}^3$ of sediment was deposited in the Bengal Basin from 11 to 7 ka BP when the southwest monsoon was strong (Goodbred and Kuehl, 2000). Furthermore, the development of Ganges–Brahmaputra River Delta began 11 ka BP, when rising sea level flooded the Bengal Basin, thereby trapping most of the river discharge on the inner margin. Recent studies based on $\delta^{18}\text{O}_{\text{sw}}$ (which is more reliable proxy for monsoon rainfall) reconstructions from the Bay of Bengal and the Arabian Sea also supports that initiation of SW monsoon around 15 ka and reduction of monsoon rainfall during Younger Dryas (Anand et al., 2008; Rashid et al., 2007).

Ultra high-resolution record of monsoon variability during the time span of 1 ka reveals that the monsoon wind strength increased during the past four centuries as the Northern Hemisphere warmed (Anderson et al., 2002). This suggests that the warming trend may directly affect the monsoon, and the observed link between Eurasian snow cover and the southwest monsoon persists on a centennial scale. However, comparison of all India rainfall data with the Northern Hemisphere air temperature anomalies over last 100 years do not support that global warming effect on the Indian rainfall at decadal time scale.

Coherent Occurrence of Abrupt Climate Shifts in the North Atlantic and Monsoon Changes

During the last Ice Age, the Indian Ocean summer monsoon exhibited abrupt changes that were closely correlated with millennial-scale climate events in the North Atlantic region, suggesting a causal link. The amplitude of abrupt changes in the North Atlantic region was smaller during the Holocene than in the glacial period; hence, the Holocene epoch is considered as a stable phase. A high-resolution study from the Oman Margin documented seven distinct intervals of weak summer monsoon during the Holocene (Figure 10). These weaker summer monsoon phases are correlated with the millennial-scale cooling in the North Atlantic. On the other hand, monsoon maxima have occurred in several pulses during which the North Atlantic was warmest, including the most recent climate changes from the Medieval Warm Period to the Little Ice Age (Gupta et al., 2003). This makes a strong case linking the monsoon events of the Indian Ocean with that of North Atlantic climate shifts. Thus, the link between North Atlantic climate and the Asian monsoon is a long-term aspect of global climate.

Similar coherent changes between monsoon variability and high-latitude climate events have also been reconstructed prior to deglaciation. For example, sediment cores from the northeastern Arabian Sea show laminated, organic-carbon-rich bands, reflecting strong monsoon-induced biological productivity that correlate with mild interstadial climate events in the northern North Atlantic region. In contrast, periods of lowered southwest monsoonal intensity, indicated by bioturbated, organic-carbon-poor bands, are associated with intervals of high-latitude atmospheric cooling and the injection of melt water into the North Atlantic basin. Arabian Sea monsoon records suggest that Dansgaard–Oeschger and Heinrich events are strongly expressed in low-latitude monsoon climate variability. Sediment records from the Pakistan Margin suggest that the rapid Dansgaard–Oeschger and Heinrich-style events known from Greenland and the North Atlantic are also a significant component also of low-latitude climate variability during the past 110 ka due to millennial to centennial fluctuations in the intensity of southwest monsoonal circulation (Schulz et al., 1998). Thus, several lines of evidence suggest a synchronicity between monsoon variability of the Indian Ocean and high-latitude climate change during pre- and post-deglaciation times.

Monsoon Forcing Mechanisms After Deglaciation

Linking the climatic events of the high latitudes to the tropical climatic events in general and Asian monsoon events in particular gained much importance during last decade. Synchronicity has been demonstrated between changes of monsoon intensity and variation of temperatures in Greenland Ice Cores and North Atlantic. Are these mere coincidences or is there a physical mechanism to explain a link between high-latitude climate changes and monsoons? None of these studies, however, has shown a connection between SST changes in the North Atlantic and the tropical latitudes and Indian monsoon system in quantitative terms. Two propositions have been made to address the forcing mechanism of the monsoon at millennial and centennial time scales after deglaciation.

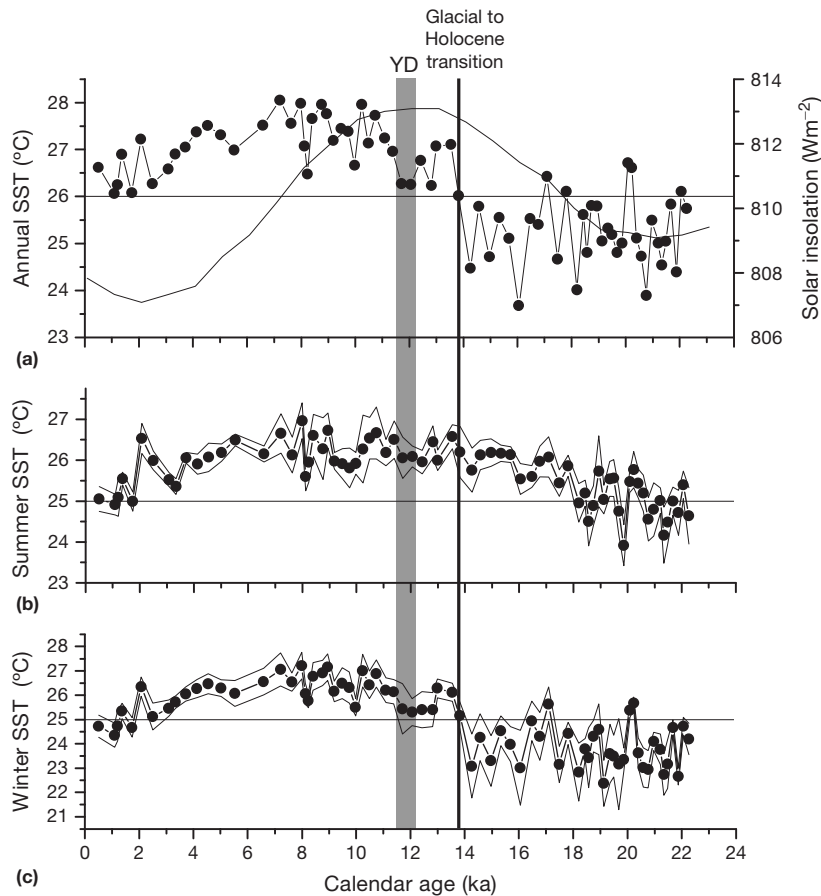


Figure 6 Annual, summer, and winter sea surface temperatures variations at the ODP Site 723A from the western Arabian Sea. These temperatures were estimated using artificial neural networks based on planktic foraminiferal census data. Modern sea surface temperature for annual, summer, and winter is shown by a straight gray line. Annual solar insolation at 20°N curve in the panel (a) is shown by a gray line. Thin gray lines in the panel (b) and (c) are minimum and maximum error bars of temperature estimations. Reproduced from Naidu PD and Malmgren BA (2005) Seasonal sea surface temperature contrast between the Holocene and last glacial period in the western Arabian Sea (Ocean Drilling Program Site 723A): Modulated by monsoon upwelling. *Paleoceanography* 20: PA1004, <http://dx.doi.org/10.1029/2004PA001078>.

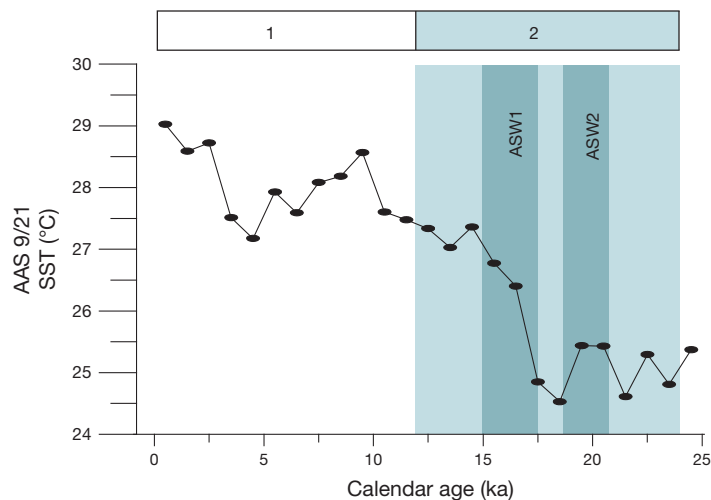


Figure 7 Sea surface temperature variations in the eastern Arabian Sea derived by the Mg/Ca ratios in planktic foraminifer species *Globigerinoides ruber*. Reproduced from Govil and Naidu (2010) Evaporation–precipitation changes in the eastern Arabian Sea for the last 68 ka: Implications on monsoon variability. *Paleoceanography* 25: PA1210, 11 pp.

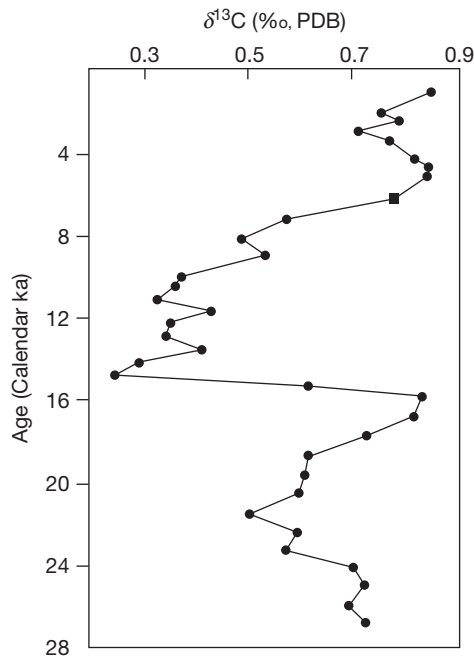


Figure 8 Changes of $\delta^{13}\text{C}$ in benthic Foraminifera of intermediate water depth (650 m) from the NW Arabian Sea. The $\delta^{13}\text{C}$ shifts between Holocene and last glacial was much smaller than the global mean. Thus the observed changes of $\delta^{13}\text{C}$ in a sediment core from the Gulf of Aden reflect changes of Red Sea outflow. The increase in $\delta^{13}\text{C}$ represent greater outflow of Red Sea water and vice versa. Modified from [Naqvi WA and Fairbanks RG \(1996\)](#) A 27 000 year record of Red Sea outflow: Implications for timing of post-glacial monsoon intensification. *Geophysical Research Letters* 23: 1501–1504.

El Niño teleconnection

The interdecadal variation of the Indian monsoon rainfall is strongly correlated with the interdecadal variations of various indices of El Niño-Southern Oscillation (ENSO; [Krishnamurthy and Goswami, 2000](#)). The decrease in the Indian monsoon rainfall associated with the warm phases of ENSO is due to an anomalous regional Hadley circulation with descending motion over the Indian continent and ascending motion near the equator sustained by the ascending phase of the anomalous Walker circulation in the equatorial Indian Ocean. However, inverse relationship between the ENSO and the Indian summer monsoon has broken down in recent decades ([Kumar et al., 1999](#)). A southeastward shift in the Walker circulation anomalies associated with ENSO events may lead to a reduced subsidence over the Indian region, thus favoring normal monsoon conditions. Additionally, increased surface temperatures over Eurasia in winter and spring, which are a part of the mid-latitude continental warming trend, may favor the enhanced land-ocean thermal gradient conducive to a strong monsoon. Therefore, Eurasian warming in recent decades helps to sustain the monsoon rainfall at a normal level despite strong ENSO events.

The Last deglaciation monsoon records of the Arabian Sea also show synchronicity between monsoon changes and high-latitude climate. It has been proposed that the Pacific El Niño is capable of effecting remote parts of northern and Southern Hemisphere and that Asian monsoon and Pacific El Niño are coupled systems ([Sirocko, 1996](#)). Such low and high-latitude teleconnections could be mediated through the Jet Stream of the upper troposphere and lower stratosphere.

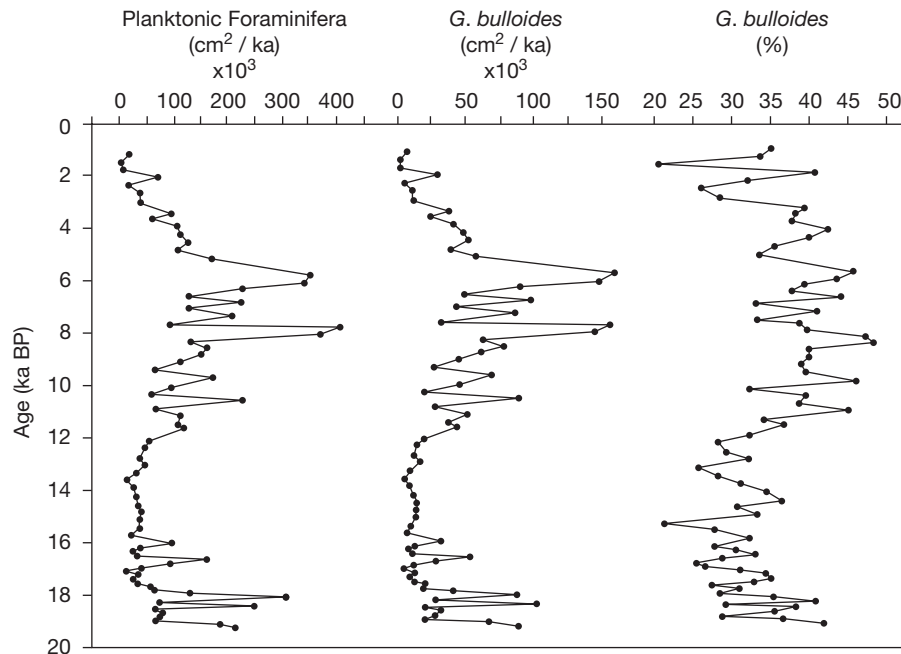


Figure 9 Variation in mass accumulation rates of planktonic Foraminifera *G. bulloides* through the last 19000 years BP in ODP Site 723A from the western Arabian Sea. These microfossil indices are sensitive to upwelling intensity and may, therefore, be employed to monitor the SW monsoon variability. Reproduced from [Naidu PD and Malmgren BA \(1996\)](#) A high resolution record of late Quaternary upwelling along the Oman Margin, Arabian Sea based on planktonic Foraminifera. *Paleoceanography* 11: 129–140.

Solar activity

There is sound evidence that the intensity of the summer monsoon has fluctuated dramatically in the past, varying with precessional changes in the distribution of solar radiation reaching the earth. The Holocene epoch has spanned half of precessional cycle. During this time the Northern Hemisphere has changed from having maximum seasonality approximately 11 ka BP, when the summer solstice was aligned with perihelion, to a minimum today. The tropical seasonal monsoons are highly influenced by precessional variations in insolation, resulting in a wetter tropical climate as monsoon intensity increases. Furthermore, variation of solar output has been suggested as the driver of both the North Atlantic cycles (Bond et al., 2001) and the variation of monsoon precipitation in Oman (Neff et al., 2001). Intensity of the Indian monsoon is found to have decreased during periods of solar minima during the last millennium suggesting solar control on the Indian monsoon on multidecadal time scale (Agnihotri et al., 2002).

Dipole

A dipole mode in the Indian Ocean has been noticed in late 1990s: a pattern of internal variability with anomalously low SST off Sumatra and high SSTs in the western Indian Ocean, with accompanying wind and precipitation anomalies (Saji et al., 1999). This air-sea interaction process is unique and inherent in the Indian Ocean and is shown to be independent of the ENSO. The dipole mode accounts for about 12% of the SST variability in the Indian Ocean. The $\delta^{18}\text{O}$ and Sr/Ca composition of modern corals from the Mentawai Island clearly show distinct cooling signals in coral Sr/Ca which represent Indian Ocean Dipole events from 1961 to 1994.

The Holocene coral records preserve similar signals and provide the first evidence that the Indian Ocean Dipole has operated since at least the mid-Holocene (Abram, 2001). These Dipole events also occurred during times when mean ocean SST and rainfall conditions were different from today, showing that the Indian Ocean Dipole is a robust climatic feature that persists in a range of mean climatic states. The fossil coral records also provide evidence that the Indian Ocean Dipole system can operate independently of the ENSO, with dipole events continuing in the mid-Holocene when ENSO is thought to have been substantially weaker than today.

Productivity Variations

The biological productivity of the Indian Ocean is highly seasonal, being highest during the summer monsoon, followed by the winter monsoon and the inter monsoon season. Flux of carbonate, organic carbon, and foraminiferal tests is tightly coupled to monsoon intensity in the western, central, and eastern Arabian Sea, as shown by sediment trap studies. However, in the past, the relationship between monsoon and productivity was not straight forward. All paleoproductivity records from the western Arabian Sea show that productivity was higher after deglaciation (Holocene) driven

by strong southwest monsoons. In the eastern Arabian Sea, the productivity proxies such as organic carbon and alkenones show higher values during the LGM as compared to the Holocene; therefore, it was suggested that in the eastern Arabian Sea paleoproductivity is predominantly linked to variation in the NE monsoon winds (Rostek et al., 1998). Recent studies from the eastern Arabian Sea also support that overall high productivity during last glacial period than in Holocene, within the last glacial period a drop in productivity corresponds Heinrich Events, which again suggests a teleconnection between the productivity in the Arabian Sea and North Atlantic climate (Singh et al., 2011). However, it is emphasized here that there is no consistent agreement between various productivity proxies from different regions of the eastern Arabian Sea. Therefore, the differences between productivity proxies such as calcium carbonate and organic carbon concentrations in the eastern Arabian Sea between glacial and interglacials are attributed to regional differences in sedimentation rates, dilution, and preservation, which modify the signal of carbonate and carbon production (Guptha et al., 2005).

Oxygen Minimum Zone Variability

The northern Arabian Sea is one of the few regions in the open ocean where thermocline water is severely depleted in oxygen. This intense oxygen minimum zone (OMZ) results from high rates of oxygen consumption in the water column due to high surface water productivity values and organic particle fluxes combined with a moderate rate of thermocline ventilation by Central Indian Ocean Water. The intensity of the OMZ has varied through orbital and suborbital time scales, and this variability is thought to be driven by changes in surface water productivity and the depth of winter mixing. Periods with a pronounced OMZ (such as the present times) are characterized by shallow winter mixing, while annual surface water productivity is high. Periods with a weak OMZ, on the other hand, are marked by low annual surface water productivity conditions and winter mixing that is deeper than today. Similar to OMZ, the denitrification process in the Arabian Sea also varied through time during glacial and interglacial periods. Isotopic fractionation of nitrogen associated with denitrification enriches $^{15}\text{NO}_3^-$ as $^{14}\text{NO}_3^-$ and is preferentially converted to gaseous products. As denitrification proceeds, the residual nitrate in the oxygen-deficit subsurface waters become progressively enriched in $^{15}\text{NO}_3^-$. A consistent pattern of heavy $\delta^{15}\text{N}$ and high accumulation rates of biogenic components during the Holocene and lighter $\delta^{15}\text{N}$ and low accumulation rates of biogenic components during glacials have been documented in the Arabian Sea (Figure 11) (Ganeshram et al., 2000). Reconstruction of denitrification at centennial time scale from the coastal eastern Arabian Sea document a substantially weakened denitrification during Little Ice Age (AD 1500 to 1750), as a result weak SW monsoon (Agnihotri et al., 2008). This suggests that the denitrification process of the Arabian Sea is tightly coupled with the SW monsoon strength in both millennial and centennial time scale. The influence of anthropogenic inducing of atmospheric CO_2 and subsequent absorption by ocean over last two centuries has resulted in ocean acidification in the Arabian Sea as evident from the presence

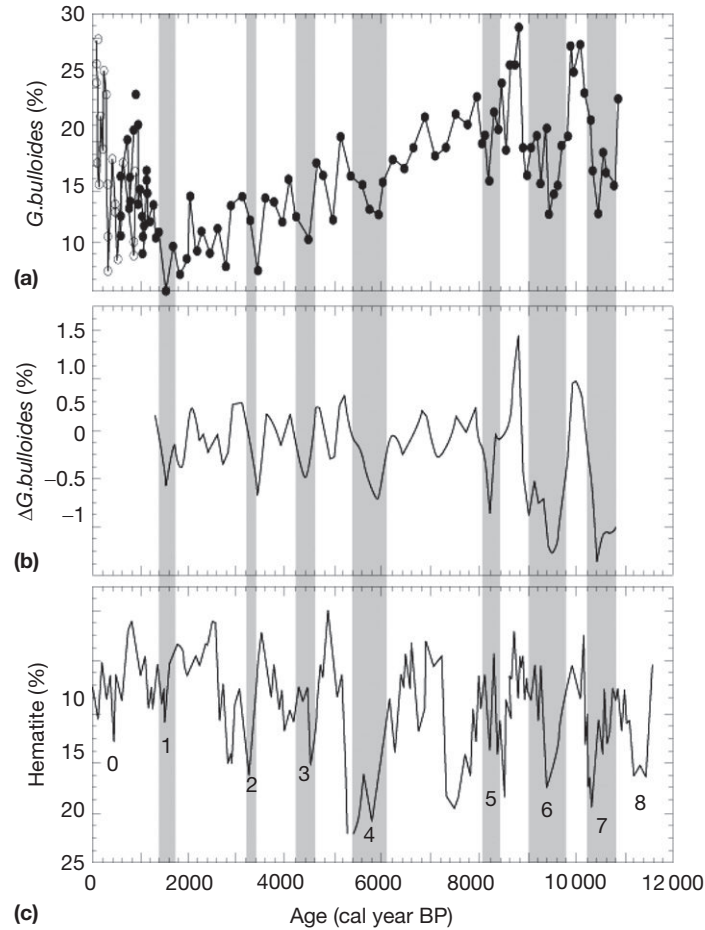


Figure 10 The abundance of *G. bulloides* in the sediment is relatively proportional to the monsoon activity in the Arabian Sea and percentage of hematite from the North Atlantic sediment reflects the temperature changes in the North Atlantic. Therefore, here the *G. bulloides* is used to infer the monsoon intensity and hematite percentage as proxy for temperature. The vertical gray bars indicate intervals of weak monsoon that coincides with the cold temperature in the North Atlantic. Modified from Gupta AK, et al. (2003) Abrupt changes in the Asian southwest monsoon during the Holocene and their links to the North Atlantic Ocean. *Nature* 421: 354–356.

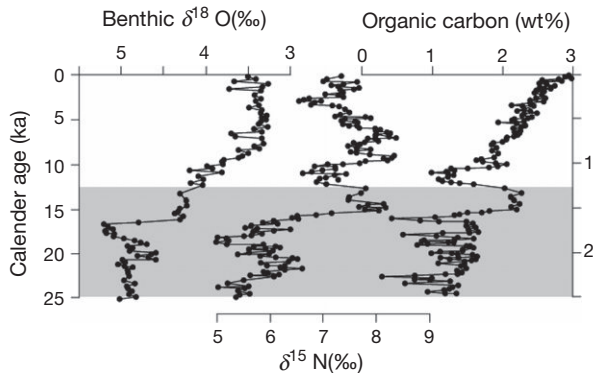


Figure 11 Benthic Foraminifera $\delta^{18}\text{O}$, $\delta^{15}\text{N}$ and organic-carbon content from a sediment core along western continental margin of India (MD76-131). The $\delta^{15}\text{N}$ and organic-carbon profiles indicate higher productivity and denitrification during the Holocene as compared to the last glacial period. Modified from Ganeshram RS, et al. (2000) Glacial–interglacial variability in denitrification in the world’s oceans: Causes and consequences. *Paleoceanography* 15: 361–376.

of thin-shelled Foraminifera in the sediment trap samples as compared with late Holocene (Moel et al., 2009).

The organic-carbon record from the Indian margin suggests lower productivity during glacial periods. This phenomenon has also been reconstructed for the Oman margin, based on a glacial age decline in biobarium and organic-carbon accumulation recorded in regional sediments. It has also been found that the lowest oxygen levels in the oxygen minimum correlates with productivity maxima in sedimentary records from the Pakistan margin. Thus, glacial decreases in productivity and denitrification in the Arabian Sea are attributed to the weakening of the summer monsoonal winds that drive seasonal upwelling and dominate mean annual export production. This substantiates that lower denitrification during glacial periods in the Arabian Sea could have resulted from a decrease in upwelling and the settling flux of organic material through the water column.

Therefore, the $\delta^{15}\text{N}$ record and associated data indicate that the Arabian Sea has exhibited large, climate-driven variations in denitrification, probably forced directly by changes in

the intensity of the summer monsoon and associated upwelling (Altabet et al., 2002). The denitrification in the Arabian Sea has been reported as being between 12 and 30 Tg years⁻¹ and thus could account for up to 10–25% of total marine denitrification.

Conclusions

The initial onset of climatic change in the Indian Ocean at the end of the glacial is marked around 15 ka BP. SST reconstructions of the Indian Ocean show 3–4 °C regional cooling during the LGM as compared to the Holocene. A striking contrast between LGM and Holocene concerns monsoon strength, productivity, denitrification, outflows of the red Sea, and Indonesia through flow. A prominent synchronicity between the strength of monsoon and high-latitude climate has been documented for the intervals before and after deglaciation. These teleconnections between high and low latitudes may be driven by the ENSO. The variability of biological productivity, monsoon strength, and denitrification are strongly coupled in the Indian Ocean. Strong monsoon during the Holocene results in higher productivity, which in turn triggers the denitrification in the Arabian Sea. It appears that monsoon, productivity, and denitrification are greatly influenced by high-latitude climate changes.

See also: **Paleoceanography:** Paleoceanography An Overview.
Paleoclimate Reconstruction: Younger Dryas Oscillation, Global Evidence.

References

- Abram NJ (2001) Holocene records of the Indian Ocean dipole. *American Geophysical Union Fall Meeting 2001, Abstract #PP11A-0443*.
- Agnihotri R, et al. (2002) Evidence for solar forcing on the Indian Monsoon during the last millennium. *Earth and Planetary Science Letters* 198: 521–527.
- Agnihotri R, et al. (2008) Variability of subsurface denitrification and surface productivity in the coastal eastern Arabian Sea over the past seven centuries. *The Holocene* 18: 755–764.
- Altabet MA, et al. (2002) The effect of millennial-scale changes in Arabian Sea denitrification on atmospheric CO₂. *Nature* 415: 159–162.
- Anand P, Kroon D, Singh AD, Ganeshram RS, Ganssen G, and Elderfield H (2008) Coupled sea surface temperature–sea water $\delta^{18}\text{O}$ reconstruction in the Arabian Sea at the millennial scale for the last 35 ka. *Paleoceanography* 23: PA4207. <http://dx.doi.org/10.1029/2007PA001564>.
- Anderson DM, et al. (2002) Increase in the Asian SW monsoon during the past four centuries. *Science* 297: 596–599.
- Bond G, et al. (2001) Persistent solar influence on North Atlantic climate during the Holocene. *Science* 294: 2130–2136.
- Clemens S, et al. (1991) Forcing mechanisms of the Indian Ocean monsoon. *Nature* 353: 720–725.
- Dahl KA and Oppo DW (2006) Sea surface pattern reconstructions in the Arabian Sea. *Paleoceanography* 21: PA1014. <http://dx.doi.org/10.1029/2005PA001162>.
- Duplessy JC (1982) Glacial to interglacial contrasts in the northern Indian Ocean. *Nature* 295: 494–498.
- Ganeshram RS, et al. (2000) Glacial–interglacial variability in denitrification in the world's oceans: Causes and consequences. *Paleoceanography* 15: 361–376.
- Ganssen GM, et al. (2011) Quantifying sea surface temperature ranges of the Arabian Sea for the past 20,000 years. *Climate of the Past* 7: 1337–1349.
- Goodbred SL and Kuehl SA (2000) Enormous Ganges–Brahmaputra sediment discharge during strengthened early Holocene monsoon. *Geology* 28: 1083–1086.
- Govil P and Naidu PD (2011) Variations of Indian monsoon precipitation during the last 32 kyr reflected in the surface hydrography of the western Bay of Bengal. *Quaternary Science Reviews* 30: 3871–3879.
- Gupta AK, et al. (2003) Abrupt changes in the Asian southwest monsoon during the Holocene and their links to the North Atlantic Ocean. *Nature* 421: 354–356.
- Guptha MVS, et al. (2005) Carbonate and carbon fluctuations in the eastern Arabian Sea over 140 ka: Implications on productivity changes. *Deep Sea Research II* 52: 1981–1993.
- Kallel N, et al. (1988) A deep hydrological front between intermediate and deep-water masses in the glacial Indian Ocean. *Nature* 333: 651–655.
- Krishnamurthy V and Goswami BN (2000) Indian monsoon – ENSO relationship on inter decadal time scale. *Journal of Climate* 13: 579–595.
- Kumar KK, Rajagopal R, and Cane MA (1999) On the weakening relationship between the Indian Monsoon and ENSO. *Science* 284: 2156–2159.
- Moel HD, et al. (2009) Planktonic foraminiferal shell thinning in the Arabian Sea due to anthropogenic ocean acidification? *Biogeosciences* 6: 1917–1925.
- Naidu PD and Govil P (2010) New evidence on the sequence of deglacial warming in the tropical Indian Ocean. *Journal of Quaternary Science* 25: 1138–1143.
- Naidu PD and Malmgren BA (1996) A high resolution record of late Quaternary upwelling along the Oman Margin, Arabian Sea based on planktonic foraminifera. *Paleoceanography* 11: 129–140.
- Naidu PD and Malmgren BA (2005) Seasonal sea surface temperature contrast between the Holocene and last glacial period in the western Arabian Sea (Ocean Drilling Program Site 723A): Modulated by monsoon upwelling. *Paleoceanography* 20: PA1004. <http://dx.doi.org/10.1029/2004PA001078>.
- Naqvi WA and Fairbanks RG (1996) A 27,000 year record of Red Sea outflow: Implications for timing of post-glacial monsoon intensification. *Geophysical Research Letters* 23: 1501–1504.
- Neff U, et al. (2001) Strong coherence between solar variability and the monsoon in Oman between 9 and 6 kyr ago. *Nature* 411: 290–293.
- Peeters F, et al. (1999) A size analysis of planktic foraminifera from the Arabian Sea. *Marine Micropaleontology* 36: 31–63.
- Rashid H, et al. (2007) A 25 ka Indian Ocean monsoon variability record from the Andaman Sea. *Quaternary Science Reviews* 26: 2586–2597.
- Rostek F, et al. (1998) Sea surface temperature and productivity records for the past 240 kyr in the Arabian Sea. *Deep Sea Research Part II* 44: 1461–1480.
- Saher M, et al. (2007) Western Arabian Sea SST during the last deglaciation. *Paleoceanography* 22: PA2208. <http://dx.doi.org/10.1029/2006PA001292>.
- Saji NH, et al. (1999) A dipole mode in the tropical Indian Ocean. *Nature* 401: 360–363.
- Schulz H, et al. (1998) Correlation between Arabian Sea and Greenland climate oscillation of the past 110,000 years. *Nature* 393: 54–57.
- Singh AD, et al. (2011) Productivity collapses in the Arabian Sea during glacial cold phases. *Paleoceanography* 26: PA3210. <http://dx.doi.org/10.1029/2009PA001923>.
- Sirocko F (1996) Teleconnections between the subtropical monsoons and high-latitude climates during the last deglaciation. *Science* 272: 526–529.

Postglacial North Atlantic

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Introduction

The most recent warm interglacial period, the Holocene, can be characterized as 10 ka of relative climatic stability compared to the previous 100 ka of Earth's history, the last glaciation ([Figure 1](#)). This notion of stability disappears however when examining the magnitude and duration of Holocene climate changes on their own merit ([Figure 2](#)). Numerous high-resolution paleoclimate reconstructions using tree-ring, lake-sediment, and ice-core data document that the Holocene was beset by global changes in temperature and hydrologic variability that persisted for up to several centuries. Most notably, the North Atlantic region experienced a sustained cold event approximately 8.2 ka that lasted for up to 400–600 yr in certain regions ([Figure 3](#)). In the Norwegian Sea, for example, local cooling has been detected as early as 8.6 ka ([Rohling and Palike, 2005](#)). This gradual cooling was suddenly intensified 8.2 ka when two North American glacial lakes flooded into the North Atlantic, slowing the formation of North Atlantic Deep Water and its associated northward heat transport for several centuries. The resulting multicentury cold event has now been recognized throughout the Northern Hemisphere as the '8.2-ka event'.

Less abrupt centennial-scale climate shifts have also punctuated the Holocene, resulting in three subdivisions based on climatic changes. These include (1) the pre-Boreal/Boreal climate transformation (from 10 to 8.5 ka), (2) the relatively warm Hypsithermal interval from 8.5 to 5 ka, and (3) the most recent Neoglacial cooling from 5 ka to today. Whereas these names have largely been abandoned, it is widely agreed that changes in the Earth's precession, obliquity, and eccentricity, in addition to changes in the global thermohaline circulation, have contributed to North Atlantic climatic variability during the Holocene.

Identifying subtle Holocene climatic variations in marine records can be challenging for several reasons. First, the record of Holocene oceanographic conditions is often preserved in only a few centimeters of poorly compacted marine silt and clay. In order to decipher SST and water-mass changes that may have persisted for a century or less, one must identify regions where sediments have accumulated rapidly over the past 10 ka, and use precision coring techniques that can preserve the delicate sediments that rest on the sea floor. Second, past oceanographic conditions must be interpreted from a combination of temperature, salinity, and density proxy data at a variety of depths in the water column. In addition, changes in these water properties could result from a number of climatic forcings, including changes in solar insolation, thermohaline circulation, and surface winds.

Today, warm and relatively fresh surface waters are transported from the equatorial Atlantic to the NE North Atlantic

along the Gulf Stream, North Atlantic Drift, and Norwegian Atlantic Current ([Figure 4](#)). Along this northward journey, evaporation and cooling increase the density of surface waters until they become unstable and sink at several key locations in the NE Atlantic ([Figure 5](#)). The conversion of surface water to deepwater releases heat from the ocean to the atmosphere, bathing nearby locations in anomalous warmth that buffers the entire region from nearby cold Arctic air masses.

Variations in solar insolation and the amount and location of deep-water production have led to regional climate oscillations during the Holocene. In this article, we review detailed records of major Holocene (Postglacial) variability in SST and ocean circulation within a select group of high sediment-accumulation-rate basins in the North Atlantic.

SST Variations

Holocene SST reconstructions are more spatially complete in the North Atlantic compared to other oceans, although still primarily limited to high sediment-accumulation-rate, near-shore basins. Pioneering research seeking to identify a long-term trend in upper ocean temperatures focused on reconstructing SSTs for 9 and 6 ka, using planktonic foraminiferal assemblages ([Ruddiman and Mix, 1993](#)). Calibration studies in the 1950s and 1960s had demonstrated that certain groupings of planktonic Foraminifera are highly correlated to SST, in addition to salinity and water-column nutrients ([Morey et al., 2005](#)). [Ruddiman and Mix \(1993\)](#) found no discernible pattern of past SST variability in the North Atlantic and questioned both the methodology used for quantitative reconstructions and the age control associated with variable sediment-accumulation rates.

A subsequent study of reconstructed summer SST for 6 ka, using a variety of methods (diatom assemblages, marine mollusks, and dinoflagellate cysts), found that warmer conditions (+1–4 °C relative to today) inundated the Norwegian Sea, Hudson Bay, Baffin Bay, Nares Strait, Davis Strait, Labrador Sea, and Greenland Sea during a time when summertime insolation was approximately 7% higher than today ([Kerwin et al. \(1999\)](#), and references therein). This trend of warmer-than-present SSTs during the early to middle Holocene is perhaps best illustrated in core MD952011 ([Calvo et al., 2002](#)) from the Norwegian Sea where maximum SST warming (+2 °C relative to today) accompanied the Holocene Thermal Maximum between 8.6 and 5.5 ka, followed by a gradual cooling to today's temperatures.

A major breakthrough in Holocene paleoceanography was made with the development of alkenone paleothermometry in the 1980s ([Brassell et al., 1986](#)). Alkenones are long-chained

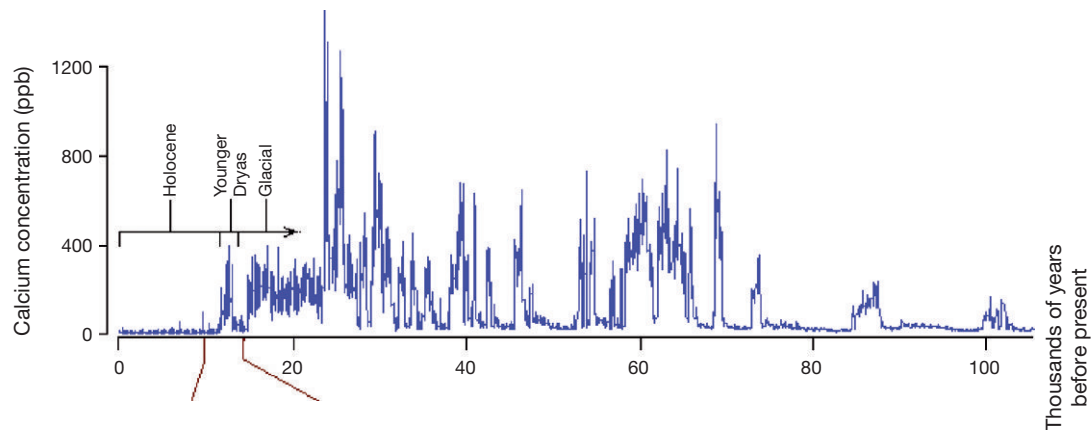


Figure 1 Time series of calcium concentration in Greenland over the past 110 ka. Calcium concentration is a proxy for the amount of dust in the atmosphere over Greenland and illustrates the relative stability of the Holocene compared to the last glacial period. From <http://www.gisp2.sr.unh.edu>; modified from Mayewski PA, Meeker LD, Whitlow SI, et al. (1994) Changes in atmospheric circulation and ocean ice cover over the North Atlantic during the last 41 000 years. *Science* 263: 1747–1751.

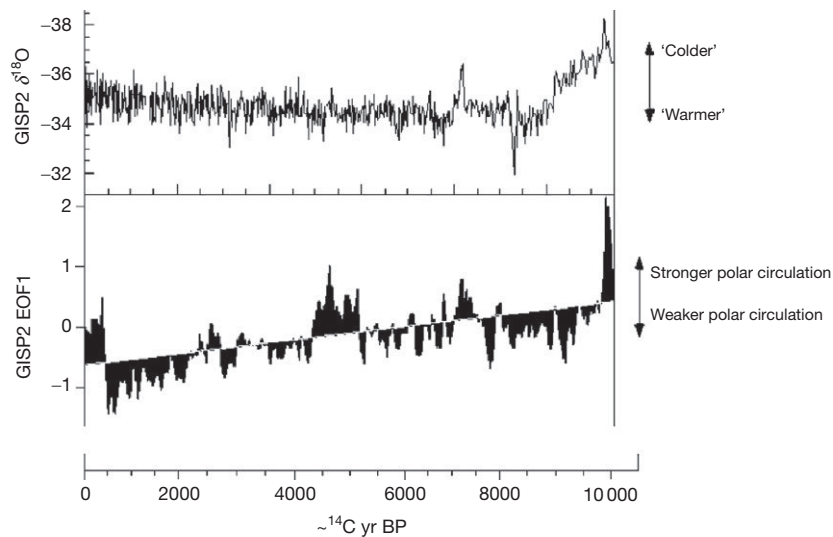


Figure 2 Holocene climate variability illustrated in the Greenland Ice Sheet Project II ice core. The top panel shows the $\delta^{18}\text{O}$ record (proxy for temperature) from Greenland over the past 10 ka (Grootes et al., 1993). The bottom panel shows variations in EOF1 which is a composite measure of major chemistry representing atmospheric circulation (O'Brien et al., 1995).

ketones synthesized as membrane lipids by certain kinds of algae that are well preserved in marine sediments (see [Alkenone Paleothermometry Based on the Haptophyte Algae](#)). The alkenone unsaturation index U_{37}^K is considered to be a robust proxy for SST because the algal organisms synthesizing alkenones must live in the photic zone, where their growth temperature is closely tied to SST. By measuring the alkenone unsaturation ratio preserved in a wide array of marine-sediment cores, a detailed picture of past surface ocean conditions may be possible (Bard, 2001).

Alkenone-based SST reconstructions from seven high-accumulation sediment cores in the North Atlantic (north of 35°N) and the western Mediterranean Sea show a cooling trend of $0.15\text{--}0.27^\circ\text{C}$ per thousand years over the past 10 ka (Marchal et al., 2002). These records are significant in that they

span a large area of the NE North Atlantic from 36° to 74°N latitude (Figure 6). This surface ocean cooling trend is also apparent in three non-alkenone reconstructions from the Labrador Sea and Iceland Basin (Figure 7). Isotopic analyses ($\delta^{18}\text{O}$) on two foraminiferal species, in combination with dinocyst assemblage data, suggest that SSTs have been cooling steadily since the early Holocene (Solignac et al., 2004). These $\delta^{18}\text{O}$ records each show more high-frequency variability than the alkenone-derived records, likely because of the sensitivity of isotopes to changes in sea-surface salinity as well as surface temperature (Solignac et al., 2004).

In contrast to the records from the high-latitude North Atlantic, alkenone data from sediment cores in the low-latitude North Atlantic (south of 30°N), eastern Mediterranean, and northern Red Sea (Figure 8) suggest that SSTs gradually warmed

during the Holocene (Rimbu et al., 2003). This apparent contradictory pattern of SST variability within the North Atlantic can be explained by opposing trends in the mean annual insolation forcing, which decreases throughout the Holocene in the high latitude North Atlantic but increases at low latitudes (Figure 9). Opposite SST trends in the eastern and western Mediterranean may reflect the continuous weakening of the North Atlantic Oscillation (NAO) from the early to the late Holocene (Duplessy et al., 2005; Rimbu et al., 2003).

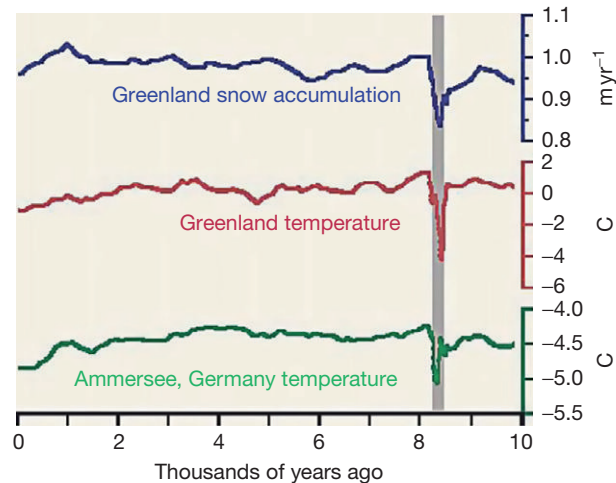


Figure 3 The 8.2-ka event recorded in the snow accumulation and temperature ($\delta^{18}\text{O}$) records from the GISP2 ice core and the oxygen-isotope record of fossil shells from Lake Ammer in Germany. Source: Von Grafenstein U; Erlenkeuser H, Müller J, Jouzel J, and Johnsen S (1998) The cold event 8 200 years ago documented in oxygen isotope records of precipitation in Europe and Greenland. *Climate Dynamics* 14: 73–81.

Ocean Circulation Variations

During the last glaciation, many of the large global millennial-scale climate oscillations coincided with alterations in thermohaline circulation in the North Atlantic. Understanding variability in thermohaline circulation during the Holocene has more recently become an objective of paleoclimatologists, leading to the refinement of a variety of proxies. In the Labrador Sea, for example, $\delta^{18}\text{O}$ measurements on planktonic Foraminifera, combined with analog reconstructions using dinocyst assemblages, have been used to document past thermohaline properties of the upper water column (Solignac et al., 2004). Similarly, grain size variations in deep ocean sediments have been correlated to changes in the speed of bottom currents south of Iceland (Bianchi and McCave, 1999). The concentration of

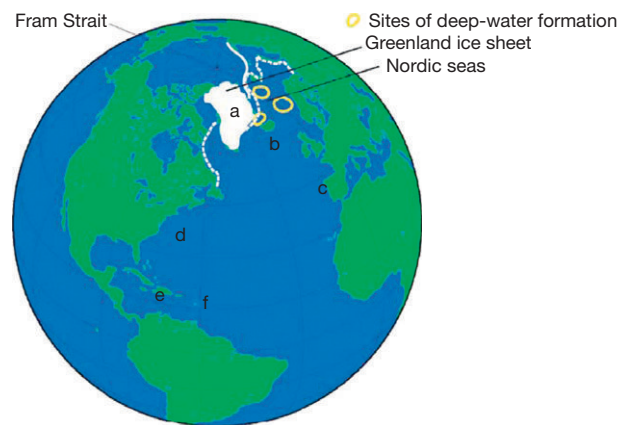


Figure 5 Sites of deep-water formation in the NE North Atlantic.

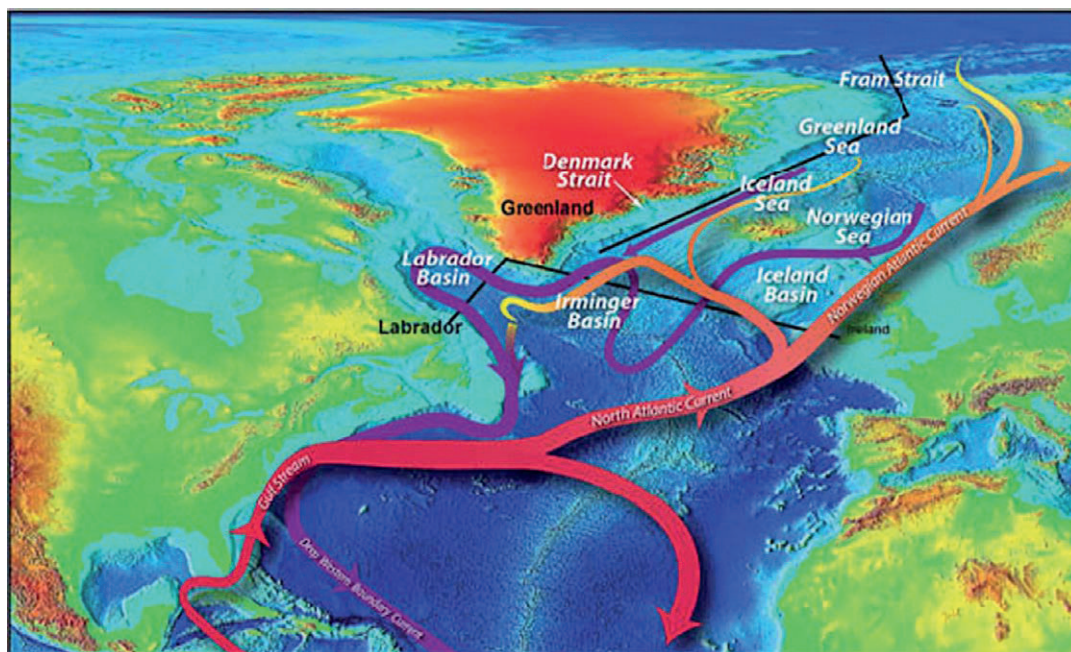


Figure 4 Map of North Atlantic region showing the Gulf Stream, North Atlantic Drift, and Norwegian Atlantic Current. From the Woods Hole Oceanographic Institution's Ocean and Climate Change Institute.

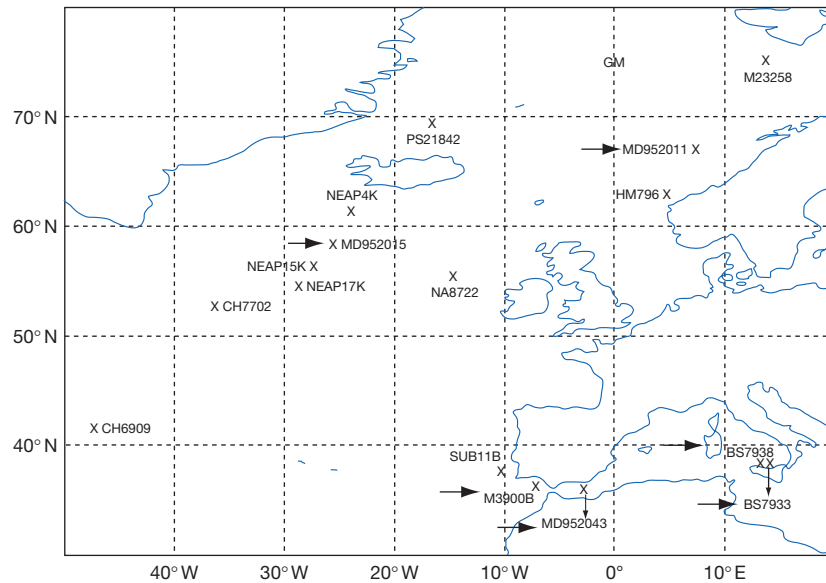


Figure 6 Location map of seven sediment cores from the North Atlantic and western Mediterranean Sea. These alkenone records all suggest that SSTs decreased through time during the Holocene. From Marchal O, et al. (2002) Apparent long-term cooling of the sea surface in the northeast Atlantic and Mediterranean during the Holocene. *Quaternary Science Reviews* 21: 455–583.

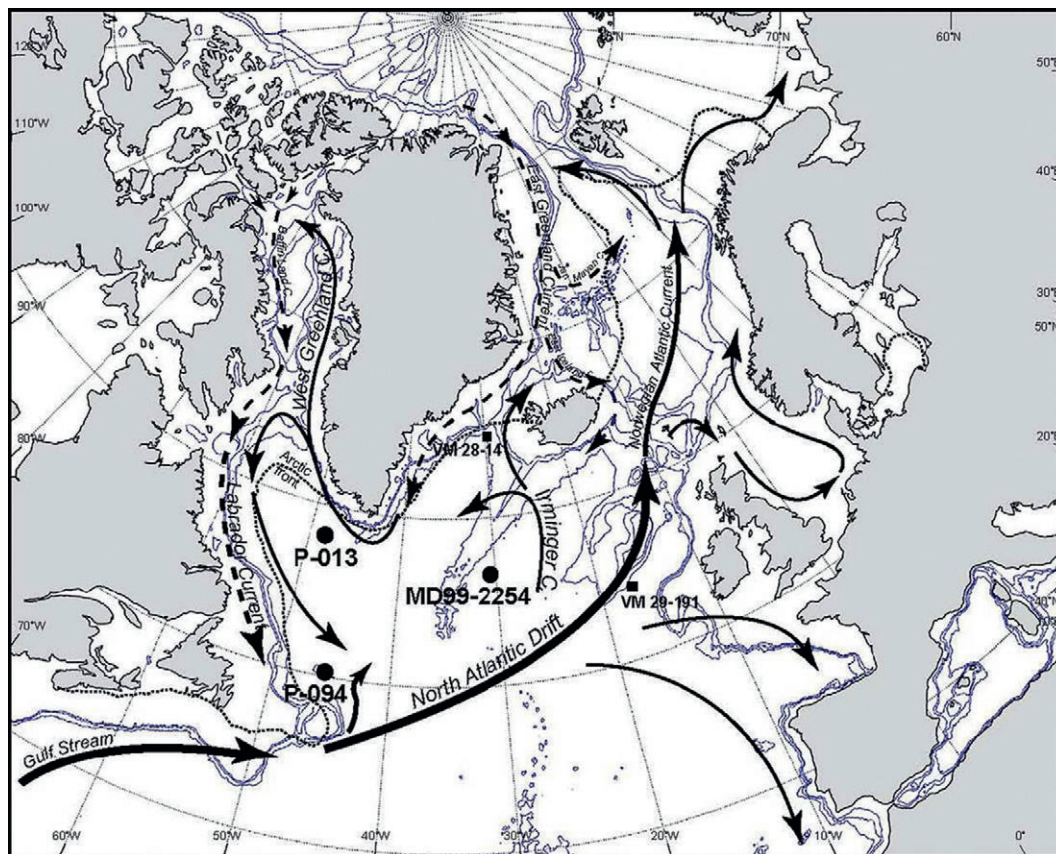


Figure 7 The location of sediment cores P-094, P-013, and MD99-2254 from the Labrador Sea and Iceland Basin. From Solignac S, de Vernal A, and Hillaire-Marcel C (2004) Holocene sea-surface conditions in the North Atlantic – contrasted trends and regimes in the western and eastern sectors (Labrador Sea vs. Iceland Basin). *Quaternary Science Reviews* 23: 319–334.

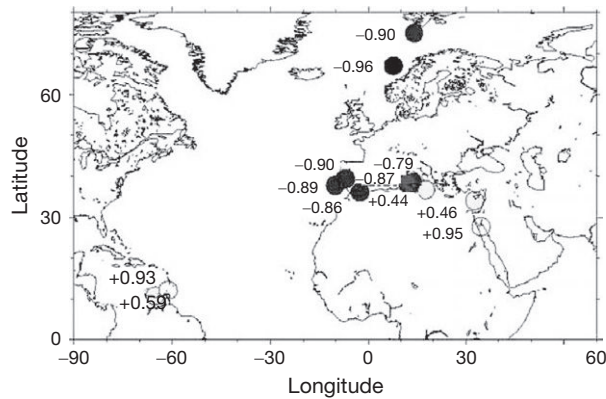


Figure 8 The location of five sediment cores from the low-latitude North Atlantic, eastern Mediterranean, and northern Red Sea, which suggest that SSTs have been increasing through time during the Holocene (Rimbu et al., 2003).

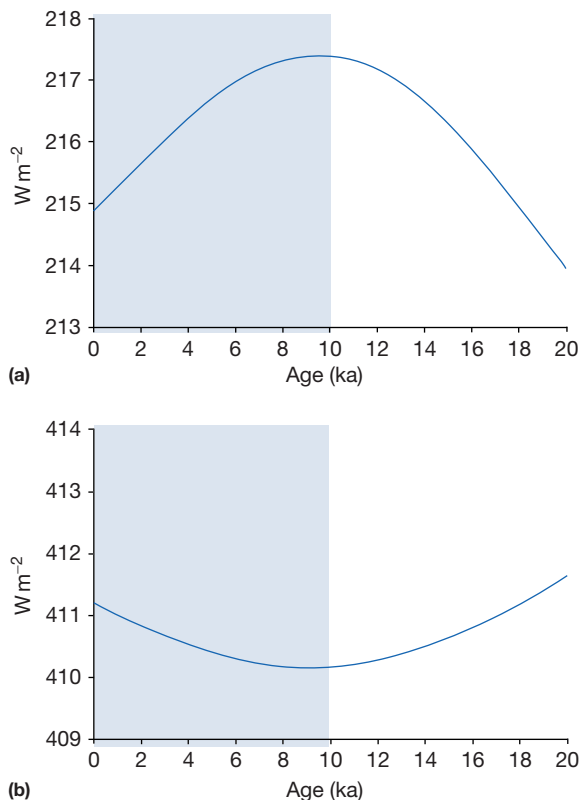


Figure 9 Mean annual insolation over the past 20 ka at A, 65° N latitude and B, 10° N latitude. The grey zone highlights the Holocene. From Duplessy JC, Cortijo E, and Kallel N (2005) Marine records of Holocene climatic variations. *Comptes Rendus Geoscience* 337: 87–95.

carbonate from the Bermuda Rise (Keigwin and Jones, 1989) and changes in the export flux of coccoliths to the south of Iceland (Giraudeau et al., 2000) have been shown to reflect instabilities in surface circulation. Carbon isotopic variations in benthic Foraminifera from the Feni Drift (Oppo et al., 2003) suggest that relatively high $\delta^{13}\text{C}$ values indicate a higher

proportion of North Atlantic Deep Water relative to Southern Ocean water. Th/Pa ratios from sinking particles that accumulated in Bermuda Rise sediments reflect the export of water from the North Atlantic basin and show a brief episode of significantly reduced deep-water formation during the mid-Holocene (McManus et al., 2004).

Paleoceanographic reconstructions using these techniques (and others) suggest that millennial-scale and shorter-term fluctuations in thermohaline circulation were a prevalent feature of the Holocene. Whereas changes in solar activity have been linked to a 1.5-ka cycle of North Atlantic hydrologic variability (Bond et al., 2001) and possibly century-scale solar variability, other, shorter-term variations result from changes in global thermohaline circulation, meltwater fluxes, and the general atmospheric circulation of the North Atlantic region. Although many more well-dated proxy records are needed to establish a complete picture of Holocene ocean circulation variability, the well-dated records that exist reveal several consistent patterns. In particular, during the early Holocene, deep-water formation increased in the northeastern corner of the North Atlantic, coincident with a period of reduced deep-sea ventilation in the Labrador Sea in the western North Atlantic.

In the eastern North Atlantic, where the Barents Sea borders the Arctic Ocean (Figure 10), oxygen isotopes from planktonic and benthic Foraminifera provide insight into Holocene thermohaline circulation changes. Today, cold and relatively fresh Arctic water enters the Barents Sea from the north and tends to remain at the surface. In contrast, relatively warm ($>1^\circ\text{C}$) and saline (>35 per mil) Atlantic water arrives from the south via thermohaline circulation. This northward-flowing Atlantic water sinks near the Barents Sea, but only to an intermediate depth, because the bottom waters are occupied by extremely cold and dense water that originates from brine rejection during sea-ice formation (Duplessy et al., 2005).

As a result, the hydrologic profile of the Barents Sea is strongly dependent on the amount of Atlantic water that reaches the area and settles in as intermediate water, due to thermohaline circulation.

Isotopic variations in sediment core ASV 880 from the Franz Victoria Trough (northern Barents Sea) provide a high-resolution record of deep-sea ventilation in the NE North Atlantic (Duplessy et al., 2005, and references therein). This record shows that thermohaline circulation and deep-water production peaked from 7.8 to 6.9 ka. During this time, Atlantic water occupied the entire water column as very little Arctic water sank along the Barents Sea Shelf. This was also the time of maximum Holocene SSTs, due in part to the strong northward flow of warm Atlantic water and the enhanced summer insolation forcing during the Holocene Thermal Maximum. After 6.9 ka, the flux of Atlantic water into the Arctic decreased and surface temperatures cooled. This reduction in thermohaline circulation created thicker and more extensive sea ice in winter. Subsequent brine rejection created a cold, dense water mass which promptly began to sink to the bottom of the Barents Sea, progressively replacing the Atlantic water that had occupied the entire water column. More recent hydrologic variability in the Barents Sea seems to correlate to long-term changes in the NAO, rather than the thermohaline circulation (Blindheim et al., 2000).

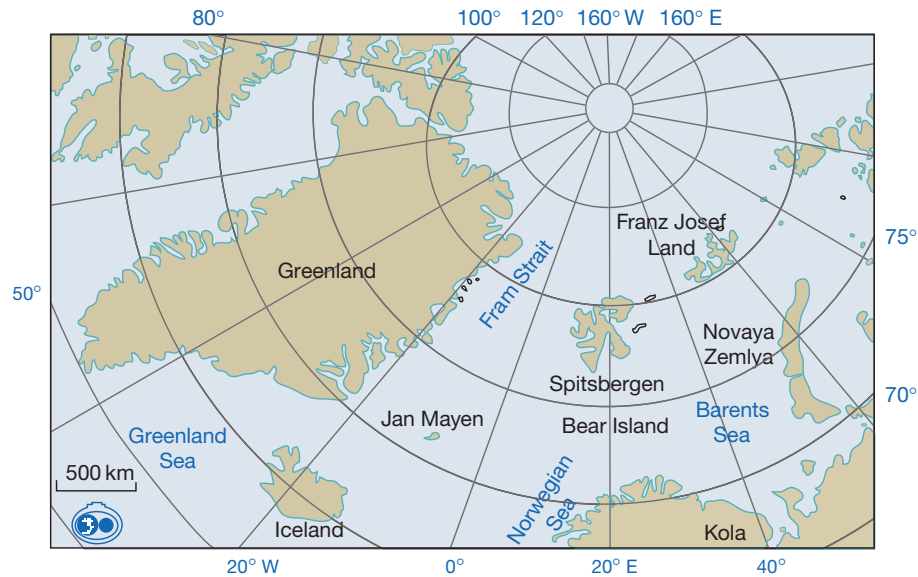


Figure 10 Location map of the Barents Sea.

During the early to middle Holocene, equally dynamic ocean circulation changes were occurring in the western North Atlantic. Here the analog method has been used to match fossil dinocyst assemblages from Holocene sediments to modern dinocyst assemblages from nearly 700 sites in the mid- to high-latitude North Atlantic and Arctic oceans (Solignac et al., 2004). This approach allows quantitative reconstructions of SST, sea-surface salinity, and sea ice cover. Two sediment cores from the western North Atlantic show that the surface of the Labrador Sea was relatively warm and fresh from 10 to 7.5 ka, when the final remnants of the Laurentide Ice Sheet melted away (Solignac et al., 2004). After 7.5 ka, surface temperatures decreased and the salinity increased as the meltwater flux from Laurentide Ice Sheet was eliminated. Surface temperatures and salinity then remained relatively stable throughout the remainder of the Holocene, with minor fluctuations representing changes in the relative strength and position of nearby minor ocean currents. One of the more significant fluctuations occurred about 7 ka, when the surface of the Labrador Sea became dense enough that intermediate water formation intensified. This period of enhanced deep-sea ventilation in the western Atlantic came at a time when warm Atlantic waters entering the Barents Sea diminished, resulting in a slowing down of the thermohaline circulation along the NE North Atlantic route. This pattern illustrates the antiphase nature of thermohaline circulation changes in the western and eastern North Atlantic.

Other records from high sediment-accumulation regions between the Labrador Sea and Barents Sea reinforce the dynamic nature of Holocene ocean circulation variations. A thick accumulation of Holocene sediments at the Rockall Plateau provides a well-dated record of oceanographic variability along the North Atlantic Drift. Core NA87-22 (Duplessy et al., 1992; Samthein et al., 1994) shows that changes in sea-surface salinity and density accompanied shifts in the amount of meltwater that reached the North Atlantic and the pattern of atmospheric circulation. Like the records from the Labrador Sea,

surface-water salinity near the Rockall Plateau was relatively low from 9 to 7.5 ka because of large meltwater contributions from the Laurentide Ice Sheet and the discharge of glacial lakes Agassiz and Ojibway during the 8.2-ka event. During the middle Holocene (7–5 ka), salinity increased after the final disappearance of the Laurentide Ice Sheet. Another major decline in surface water salinity began about 4.5 ka and lasted until 2.8 ka. This more recent Holocene salinity minimum is not associated with a meltwater peak because the Laurentide Ice Sheet had completely melted by this time. Rather, it appears to result from large-scale changes in atmospheric circulation that increased the amount of precipitation relative to evaporation over the North Atlantic (Duplessy et al., 2005).

Core MD99-2254 from the Greenland Rise shows several short intervals of cold, low-salinity surface water during the past 8 ka (Solignac et al., 2004). These pulses have been interpreted as increased fluxes of Arctic water from the north. They are consistent with periodic penetrations of cold surface waters and increased drift ice in the Nordic and Labrador Seas that were first described by Bond et al. (1997) and subsequently observed in numerous other studies throughout the Atlantic basin. Relatively fresh Arctic water in the Greenland Rise would have formed a buoyant surface-water layer that strengthens the stratification of the upper water column. This strong stratification would have limited the diffusion of heat through the entire water column and led to more warming of the surface layer in the summer and more cooling in the winter. Early work suggested that this persistent millennial-scale cycle was tied into fluctuations in the energy output of the Sun, and Bond et al. (2001) suggest that solar fluctuations, despite being small as an overall forcing, can affect climate in the North Atlantic over the entire Holocene.

It is worth noting that spatial variations in surface-water conditions are not solely driven by changes in insolation and thermohaline circulation. Recent research has documented that variations in the NAO are a persistent feature of the Holocene climate in the North Atlantic. When the NAO

index is high, strong winds from the west accompany a vibrant northward-flowing Norwegian current (Blindheim et al., 2000). The prevailing winds push Atlantic water toward the Scandinavian coasts, causing Atlantic water to penetrate deep into the Barents Sea. In contrast, when the NAO index is low, Atlantic water does not penetrate far into the Barents Sea.

Conclusions

Holocene oceanographic changes in the North Atlantic Ocean have been smaller than those of the last glacial period because of the absence of the large terrestrial ice sheets surrounding the basin. Temporal and spatial variations in insolation forcing and deep-water production during the Holocene have led to regional climate oscillations in the North Atlantic region. SST changes of the order of 1–2 °C have been observed throughout the North Atlantic and can be explained by a linear response to mean annual insolation variations (Figure 9). Changes in the thermohaline circulation are becoming better recognized in the high-latitude regions, but remain ambiguous at middle and low latitudes. Deep-water formation and thermohaline circulation changes in the NE North Atlantic appear to be out of phase with similar changes in the western North Atlantic, particularly the Labrador Sea. For example, a reduction in thermohaline circulation in the Barents Sea about 7 ka coincides with an increase in the formation of intermediate water to the west in the Labrador Sea.

In addition to the direct effects of thermohaline circulation and solar variability, inherent variations in the coupled ocean–atmosphere system such as the NAO can cause patterns of North Atlantic change that persist for decades and possibly centuries. Similarly, changes in solar variability may alter the surface wind and ocean dynamics in the North Atlantic, causing repeated occurrences of cold, iceberg-rich Arctic water to penetrate deep into the Nordic and Labrador seas. These oscillations underscore the complex dynamics that is the North Atlantic climate system.

See also: **Paleoceanography, Records:** Late Pleistocene North Atlantic.

References

- Bianchi GG and McCave IN (1999) Holocene periodicity in North Atlantic climate and deep-ocean flow south of Iceland. *Nature* 397: 515–517.
- Blindheim J, Borokov V, Hansen B, Malmberg SA, Turell WR, and Osterhus S (2000) Upper layer cooling and freshening in the Norwegian Sea in relation to atmospheric forcing. *Deep Sea Research* 147: 655–680.

- Bond G, Kromer B, Beer J, et al. (2001) Persistent solar influence on North Atlantic climate during the Holocene. *Science* 294: 2130–2135.
- Bond G, Showers W, Cheseby M, et al. (1997) A pervasive millennial-scale cycle in the North Atlantic Holocene and glacial climates. *Science* 278: 1257–1266.
- Brassell SC, Eglinton G, Marlowe IT, Pflaumann U, and Sarnthein M (1986) Molecular stratigraphy: A new tool for climatic assessment. *Nature* 320: 129–133.
- Calvo E, Grimalt J, and Jansen E (2002) High resolution UK37 sea surface temperature reconstructions in the Norwegian Sea during the Holocene. *Quaternary Sciences Reviews* 21: 1385–1394.
- Duplessy JC, Cortijo E, and Kallel N (2005) Marine records of Holocene climatic variations. *Comptes Rendus Geoscience* 337: 87–95.
- Duplessy JC, Labeyrie L, Arnold M, Paterne M, Duprat J, and van Weering T (1992) Changes of surface salinity of the North Atlantic ocean during the last deglaciation. *Nature* 358: 485–488.
- Giraudeau J, Cremer M, Manthe S, Labeyrie L, and Bond G (2000) Coccolith evidence for instabilities in surface circulation south of Iceland during Holocene times. *Earth and Planetary Science Letters* 179: 257–268.
- Groote PM, Stuiver M, White JWC, Johnsen SJ, and Jouzel J (1993) Comparison of oxygen isotope records from the GISP2 and GRIP Greenland ice cores. *Nature* 366: 552–554.
- Keigwin LD and Jones GA (1989) Glacial-Holocene stratigraphy, chronology, and paleoceanographic observations on some North Atlantic sediment drifts. *Deep-Sea Research* 36: 845–867.
- Kerwin MW, Overpeck JT, Webb RS, De Vernal A, Rind DH, and Healy RJ (1999) The role of oceanic forcing in mid-Holocene Northern Hemisphere climatic change. *Paleoceanography* 14: 200–210.
- Koc N, Jansen E, and Hafliðason H (1993) Paleoceanographic reconstructions of surface ocean conditions in the Greenland, Iceland, and Norwegian seas through the last 14 kyr based on diatoms. *Quaternary Science Reviews* 12: 115–140.
- Marchal O, et al. (2002) Apparent long-term cooling of the sea surface in the northeast Atlantic and Mediterranean during the Holocene. *Quaternary Science Reviews* 21: 455–583.
- Mayewski PA, Meeker LD, Whitlow SI, et al. (1994) Changes in atmospheric circulation and ocean ice cover over the North Atlantic during the last 41,000 years. *Science* 263: 1747–1751.
- McManus JF, Francois R, Gherardi J-M, Keigwin LD, and Brown-Leger S (2004) Collapse and rapid resumption of Atlantic meridional circulation linked to deglacial climate changes. *Nature* 428: 834–837.
- Morey AM, Mix AC, and Pisias NG (2005) Planktonic foraminiferal assemblages preserved in surface sediments correspond to multiple environment variables. *Quaternary Science Review* 24: 925–950.
- Oppo DW, McManus JF, and Cullen JL (2003) Deepwater variability in the Holocene epoch. *Nature* 422: 277–278.
- Rimbu N, Lohmann G, Kim JH, Arz HW, and Schneider R (2003) Arctic/North Atlantic Oscillation signature in Holocene sea surface temperature trends as observed from alkenone data.
- Rohling E and Palike H (2005) Centennial-scale cooling with a sudden cold event around 8200 years ago. *Nature* 434: 975–979.
- Ruddiman WF and Mix AC (1993) The North and Equatorial Atlantic at 9000 and 6000 yr B.P. In: Wright HE, Kutzbach JE, Webb T, Ruddiman WF, Street-Perrott FA, and Bartlein PJ (eds.) *Global Climates since the Last Glacial Maximum*, pp. 94–124. Minneapolis: University of Minnesota Press.
- Sarnthein M, Winn K, Jung S, et al. (1994) Changes in east Atlantic deepwater circulation over the last 30,000 years: Eight time slice reconstructions. *Paleoceanography* 9: 209–267.
- Solignac S, de Vernal A, and Hillaire-Marcel C (2004) Holocene sea-surface conditions in the North Atlantic – contrasted trends and regimes in the western and eastern sectors (Labrador Sea vs. Iceland Basin). *Quaternary Science Reviews* 23: 319–334.
- Von Grafenstein U, Erlenkeuser H, Müller J, Jouzel J, and Johnsen S (1998) The cold event 8200 years ago documented in oxygen isotope records of precipitation in Europe and Greenland. *Climate Dynamics* 14: 73–81.

Postglacial North Pacific

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Introduction

The North Pacific Ocean hosts the largest warm pool, the lowest surface salinities of a major ocean basin, and the oldest deep waters in the world's oceans (Tomczak and Godfrey, 1994). Unlike the Atlantic Ocean, where a vigorous thermohaline circulation exists, the low surface salinity of the North Pacific inhibits deepwater formation at high latitudes, and limits poleward heat transport. In the absence of major rivers draining to the North Pacific, the source of the low surface salinity veneer can be traced in part to the heavy rainfall of the Intertropical Convergence Zone (ITCZ). Today, the ITCZ is located north of the equator most of the year (Waliser and Gautier, 1993), though the evidence that the ITCZ might have occurred in a southerly position during the last glacial period is not strong (McGee et al., 2007). As a consequence of the present northerly location of the ITCZ, and heavy precipitation in the Panama Basin (e.g., Poveda et al., 2006) and the Indonesian Archipelago (e.g., Dai, 2006), a belt of low surface salinity forms in the tropical North Pacific at $\sim 10^\circ\text{N}$, which is advected west via the North equatorial current, and then north via the Kuroshio Current, thus contributing to surface stratification at high latitudes, which prevents deepwater formation and ventilation of the deep ocean (e.g., Tomczak and Godfrey, 1994). The ultimate source of precipitation in the tropical North Pacific resides in the export of moisture from the tropical Atlantic by the northeast trade winds over the isthmus of Central America, which controls the global thermohaline circulation of the ocean (e.g., Zucker and Broecker, 1992).

Therefore, when reconstructing the salinity structure and its modes of variability in the North Pacific, the dynamics of the ITCZ and its sensitivity to latitudinal shifts through time should be considered. For example, a more symmetric distribution of rainfall around the equator should tend to increase North Pacific surface salinity and the production of North Pacific Intermediate Water (NPIW) in the northwest Pacific (NWP) (Koutavas, 2004; Tomczak and Godfrey, 1994).

Increasing sea surface temperature (SST) trends on the order of $0.34\text{--}1.8^\circ\text{C}$, which are opposite to those of the North Atlantic, have been documented in the North Pacific for the past 7 ka. A seesaw pattern possibly responds to inter-oceanic teleconnections driven by the heat budget in the tropics (Kim et al., 2004). By contrast, a decreasing trend in SST has been documented for the same time interval in the Indo-Pacific warm pool (IPWP; e.g., Stott et al., 2004).

Even though the ITCZ is closely linked to the dynamics of the El Niño/Southern Oscillation (ENSO) phenomenon, which exerts a major influence on the North Pacific and global climate, the former has an important paleoceanographic effect due to its strong seasonal dynamics (e.g., Leduc et al., 2009). The zonal SST gradient along the equatorial Pacific Ocean is responsible for the state of the ENSO system, that is, a reduced

zonal SST gradient between the IPWP and the eastern equatorial Pacific (EEP) is typical of the El Niño phase as compared to a strong SST gradient typical of the La Niña phase. Therefore, when these zonal SST gradients persist over centuries to millennia, ENSO-like states result. In addition to this simplistic model, however, other modes of operation of the ENSO system should be considered. This is the case of the Modoki or Central Pacific El Niño (Ashok et al., 2007; Yu and Kim, 2010), which might explain some 'odd' paleoceanographic patterns. The yearly evolution of these Central Pacific events can result in either warming or cooling of the northeastern or eastern Pacific, depending on the thermocline depth at the onset of each event (Ashok et al., 2007).

The ENSO system appears to have experienced significant variability during the Holocene, with evidence suggesting its progressive strengthening of variance from the early-middle to the late Holocene (e.g., Cane, 2005; Clement et al., 2000; Conroy et al., 2008; Gagan et al., 2004; Moy et al., 2002; Rodbell et al., 1999; Sandweiss et al., 2001). Therefore, given the close coupling between ENSO and the ITCZ, and their sensitivity to equatorial winds, the reconstruction of their dynamics during the Holocene needs to be examined together (Koutavas, 2004).

In this article, which updates Koutavas (2004), postglacial and in particular Holocene paleoceanographic records are synthesized, looking for evidence of the relationship between tropical and extratropical processes linking ENSO and the Pacific decadal oscillation (PDO), the ITCZ, and the NPIW, and their influence on North Pacific climate variability. A detailed discussion of the deglaciation interval can be found in Kiefer and Kienast (2005). The present revision follows the North Pacific gyre (Figure 1), beginning in the EEP, the core of the atmospheric moisture transfer between the Caribbean and the Pacific Ocean. Then it follows the path of the cold tongue to the IPWP, the NWP, that is, the Kuroshio Current, and the Subarctic gyre, to return to the tropical Pacific via the northeastern Pacific (NEP) along the California Current.

Paleoceanographic Records from the Equatorial Pacific Ocean: The EEP and the Cold Tongue

Moisture transfer from the Caribbean to the EEP is enhanced during the boreal summer when the ITCZ is in the north (e.g., Leduc et al., 2007; Liu and Tang, 2005; Xie et al., 2008). Therefore, insights can be gained into the dynamics of this major climate feedback mechanism by reconstructing the former position of the ITCZ in the EEP. On the other hand, by comparing Pacific zonal SST gradients, it is possible to reconstruct the dynamics of the ENSO system. This is the case of the comparison between Holocene Mg/Ca SST reconstructions in planktonic Foraminifera from the EEP (Koutavas et al., 2002)

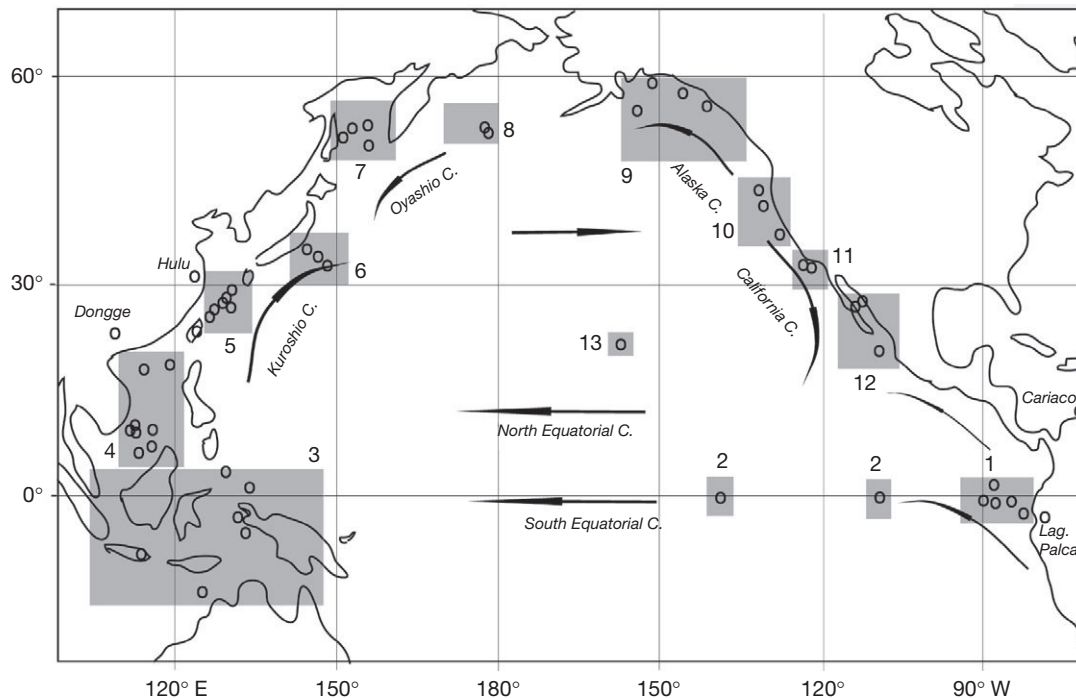


Figure 1 Map of the North Pacific, indicating major sea surface circulation patterns, and the location of records discussed in the text. Broad localities are indicated by gray rectangles, whereas sedimentary sites are indicated by dots: (1) the eastern equatorial Pacific (EEP; e.g., Koutavas et al., 2006; Leduc et al., 2009), (2) the central Pacific (Anderson et al., 2006; McGee et al., 2007), (3) the Indo-Pacific warm pool (IPWP, e.g., Griffiths et al., 2009; Beaufort et al., 2010), (4) the South China Sea (e.g., Huang et al., 2011; Tian et al., 2010), (5) the Okinawa Trough (e.g., Morimoto et al., 2007; Sun et al., 2005), (6) the Kuroshio, Central Japan (e.g., Sawada and Handa, 1998), (7) the Sea of Okhotsk (e.g., Seki et al., 2004), (8) the Bering Sea (Gorbarenko et al., 2010), (9) the Gulf of Alaska (e.g., Barron et al., 2009), (10) northern California (e.g., Barron et al., 2003), (11) Santa Barbara Basin (e.g., Friddell et al., 2003; Roark et al., 2003), (12) southern California (e.g., Barron et al., 2005; Marchitto et al., 2010), and (13) Hawaii (Uchikawa et al., 2010). Laguna Pallcacocha (Moy et al., 2002), the Dongge and Hulu Caves (China, Yuan et al., 2004; Wang et al., 2005) and the Cariaco Basin (southern Caribbean, Haug et al., 2001) are also indicated.

and the IPWP (Figure 2(a); Stott et al., 2004), which suggests that the two regions followed distinct SST trends (Koutavas, 2004; Koutavas et al., 2006). In the early and middle Holocene (10–4 ka), the warm pool and the cold tongue were warmer and cooler by $\sim 0.5^\circ\text{C}$, respectively. As a result, the zonal SST gradient was higher by about 1°C , suggesting a large-scale circulation pattern akin to a La Niña-like state, and a northward shift of the ITCZ (Koutavas et al., 2006). A northward ITCZ location has also been suggested for the Atlantic, based on evidence from the Cariaco Basin, where elevated concentrations of terrigenous titanium in Ocean Drilling Program (ODP) site 1002 occur (Figure 2(b); Haug et al., 2001).

It has been suggested that enhanced insolation in the mid-Holocene (e.g., Berger and Loutre, 1991; Liu et al., 2003) would have made boreal and equatorial summers warmer than present, which in turn would have resulted in a stronger zonal SST gradient across the tropical Pacific, leading to stronger easterly winds and a cooler EEP (Conroy et al., 2008), that is, a stronger Walker circulation and enhanced Bjerknes feedback (Cane, 2005). Other model responses also suggest drastic diminution of interannual ENSO variability (e.g., Clement et al., 2000), which is supported by evidence from coral oxygen isotopes ($\delta^{18}\text{O}$) and laminated sediments from Laguna Pallcacocha, Ecuador (Moy et al., 2002) and El Junco Crater Lake in the Galapagos (Conroy et al., 2008). These lacustrine records

offer unique and continuous records of ENSO variability for the last ~ 12 ka (Figure 2(c); Conroy et al., 2008; Moy et al., 2002). However, there are some conspicuous differences between the Pallcacocha and El Junco records for the early-middle Holocene, which suggests a more complex behavior of the La Niña-like state for the EEP region, perhaps related to unexplored dynamics of the Central Pacific El Niño phenomenon in the past.

The interpreted increase in variance of the ENSO system in the Pallcacocha record during the late Holocene (Moy et al., 2002) follows the general trend of the southward migration of the ITCZ in the Cariaco record (Figures 2(a) and 2(b); Haug et al., 2001). This trend has been interpreted to be driven by declining northern summer insolation, which in turn introduced enhanced air–sea interactions and westerly wind anomalies that triggered an El Niño-like period characterized by more intense/frequent episodes (Koutavas, 2004).

A complementary line of evidence for ENSO dynamics is provided by paleoproductivity records in the EEP. This is the case of enhanced opal burial during deglaciation which was apparently due to ‘silicic acid leakage’ advected from the Southern Ocean at a time of Fe limitation, thus favoring diatom blooms (Dubois et al., 2010). This silica excess increased export production, which in turn would have led to higher oxygen demand and denitrification within the oxygen

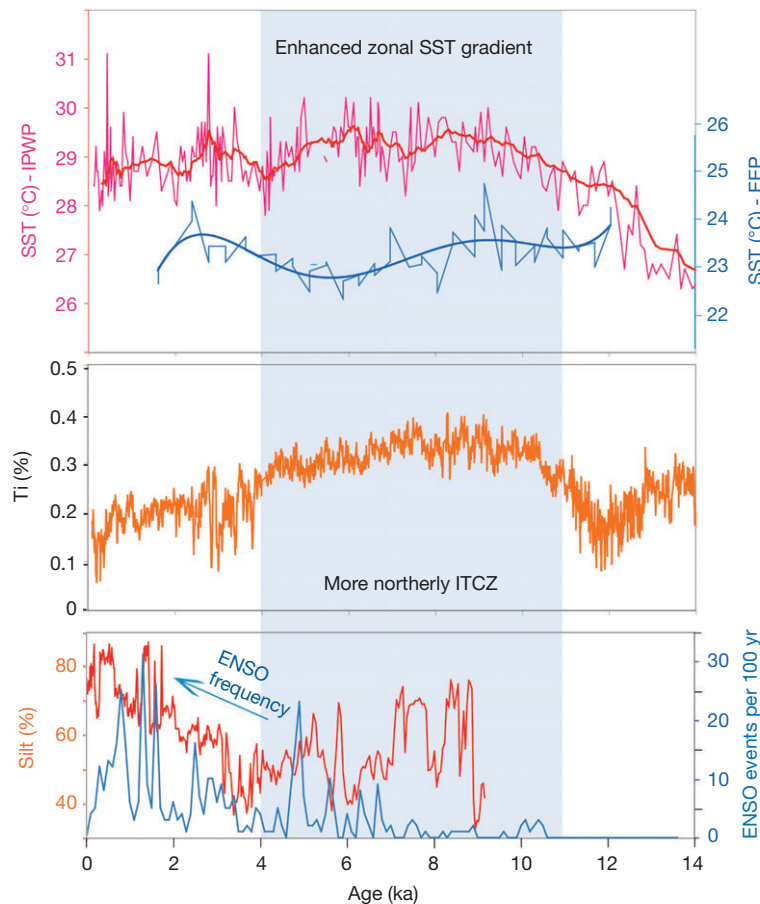


Figure 2 Comparisons between postglacial paleoclimate records from the tropical Pacific Ocean: (a) Mg/Ca SSTs from western equatorial Pacific sites MD98-2176 (red; [Stott et al., 2004](#)) and from eastern equatorial Pacific sites V21-30 (blue; [Koutavas et al., 2006](#)); (b) titanium (%) in ODP site 1002 from the Cariaco Basin (southern Caribbean, [Haug et al., 2001](#)); and (c) ENSO frequency inferred from color analysis of sediments from the Equadorian Laguna Pallcacocha (blue; [Moy et al., 2002](#)) and silt (%) from the Galapagos Is., Laguna El Junco (red; [Conroy et al., 2008](#)).

minimum zone in the EEP and the NEP ([Martínez and Robinson, 2010](#); [Robinson et al., 2009](#)). The supply of nutrients was advected from the sub-Antarctic mode water, via the equatorial undercurrent ([Hayes et al., 2011](#); [Peña et al., 2010](#)). Therefore, the increase in productivity is suggestive of enhanced upwelling ([Kienast et al., 2006](#)) under a La Niña-like scenario ([Peña et al., 2008](#)). Other proxy evidence from various records from the EEP support this increase in productivity during deglaciation and the early Holocene (e.g., [Kienast et al., 2007](#); [Martínez et al., 2006](#); [Warnock et al., 2007](#)).

By applying the ^{230}Th normalization technique, [Anderson et al. \(2006\)](#) derived new records of dust flux through the equator along 140°W and found a positive correlation with global ice volume for the past 300 ka. Minimum fluxes of continental mineral aerosol during glacial intervals exceeded those of interglacial maxima by about a factor of 2, similar to changes found elsewhere at low and mid-latitudes. [McGee et al. \(2007\)](#) studied a north–south transect through the equator along 110°W and suggested that during the Holocene the ITCZ has acted as a more effective barrier against interhemispheric dust transport, and that dust fluxes are 25–100% lower than during the last glacial interval. Whether or not this is due to weaker convection or the mean location of the ITCZ remains to be tested.

Paleoceanographic Records from the IPWP, the Kuroshio Current, and the Subarctic Gyre

Accepting a La Niña-like scenario in the early–middle Holocene, presumably sea surface salinities (SSS) in the northern tropical Pacific must have been lower than at present due to persistent rainfall north of the equator ([Koutavas, 2004](#)). This is supported by SSS reconstructions from the IPWP which suggest that SSS might have been somewhat greater during the early Holocene (e.g., [Griffiths et al., 2009](#); [Stott et al., 2004](#)). This scenario seems at odds with $\delta^{15}\text{N}$ data from a set of cores off Mindanao (6.4°N ; 126.4°E) which suggest that the thermocline was shallow during the early Holocene ([Kienast et al., 2008](#)).

In addition to these paleoceanographic inferences, the reconstruction of monsoon dynamics provides some insights into rainfall patterns, leading to a better understanding of salinity patterns in the region. Monsoon dynamics have been reconstructed from paleoceanographic records, and $\delta^{18}\text{O}$ measurements in stalagmites and coral calcite. Being part of the same system, two monsoon regimes are recognized in the western Pacific. The Australian–Indonesian monsoon (AIM) and the East Asian monsoon (EAM) are synchronous in their

opposing hemispheres, that is, while there are dry (wet) conditions at AIM during Boreal winter (summer), there are wet (dry) conditions at the EAM. The ITCZ is located in a southerly location during Boreal winter, and a northerly location in summer.

Griffiths et al. (2009) studied the Liang Luar stalagmite $\delta^{18}\text{O}$ record (Flores Island, Indonesia, Figure 3(a)) and argued in favor of a steady increase in Australian–Indonesian summer monsoon (AISM) during the early Holocene. They argue that precipitation, which increases during austral summer, does not appear to show a direct relation to insolation, but was apparently controlled by sea-level rise. In contrast, the coccolith and pollen record of core MD98-2175 (5°S–133.45°E; eastern Banda Sea) suggest that the AISM rainfall follows the pattern of glacial–interglacial dynamics, whereas the AIWM proxies (primary production and length of the dry season) are strongly influenced by precessional cycles, that is, low-latitude insolation during the austral winter (Beaufort et al., 2010). Furthermore, the synchrony of the AIM with other regional monsoon systems and ENSO at precession timescales suggests a common climatic forcing and highlights the role of the tropics on global climate (Beaufort et al., 2010).

As for the latest Holocene, geochemical data from a sediment core in the shallow silled and intermittently dysoxic Kau

Bay in Halmahera (Indonesia, 1°N; 127.5°E) were used to reconstruct century-scale variability in climate and basin ventilation within the IPWP (Langton et al., 2008). Downcore variations in bulk sedimentary $\delta^{15}\text{N}$ were attributed to changes in oceanographic conditions related to fluctuations in the ENSO system, that is, a century-long increase in El Niño activity beginning ca. 1.7 ka, with peaks in El Niño activity ca. 1.5, 1.15, and 0.7 ka. Kau Bay results suggest that there was diminished ENSO amplitude, or a departure from El Niño-like conditions during the Medieval Warm Period, and a distinctive, but steady decrease of El Niño activity during and after the Little Ice Age (Langton et al., 2008). These are interesting results that appear to be the opposite of those for the entire Holocene at the IPWP, that is, El Niño during warm periods, and vice versa. More data will lead to a better understanding of these high-frequency fluctuations through the Holocene.

Unlike the Liang Luar stalagmite record, the $\delta^{18}\text{O}$ record of planktonic Foraminifera in core G5-2-056P (3.58°S; 132.18°E, Seram Trough, eastern Indonesia, Brijker et al., 2007) suggests a steady decline in SST and SSS from the early to the mid-Holocene (10.6–6 ka). Climate conditions were ‘stable’ during the early Holocene compared to the mid–late Holocene when El Niño-like conditions apparently arose; the transition occurred at 6–4.5 ka (Brijker et al., 2007), in agreement with

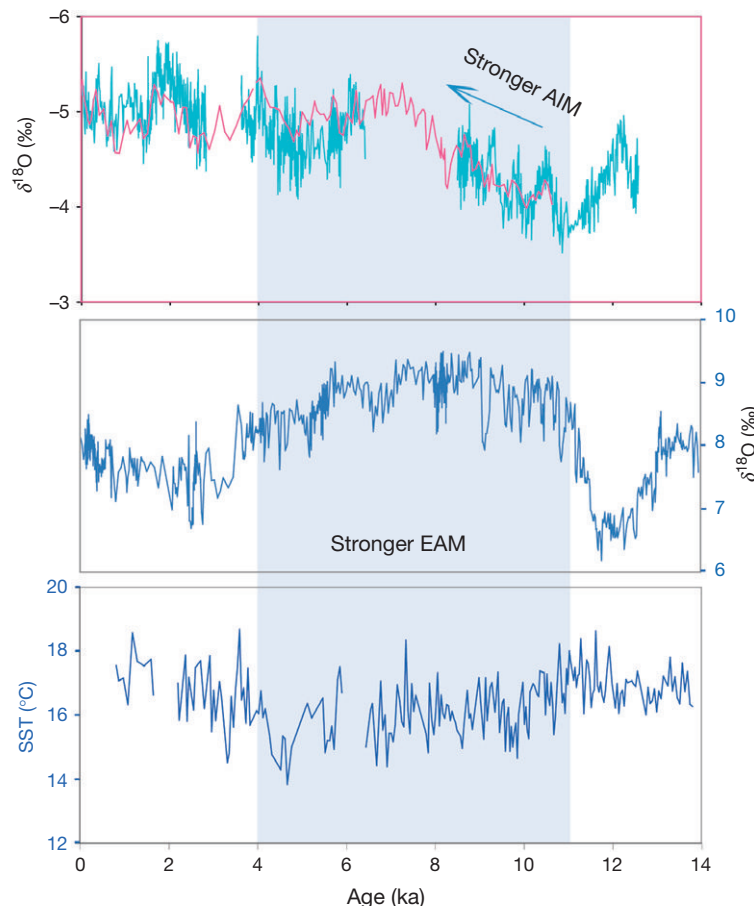


Figure 3 Comparisons between postglacial paleoclimate records from the northern Pacific. (a) $\delta^{18}\text{O}$ of the Indonesian Liang Luar stalagmite (Griffiths et al., 2009), (b) $\delta^{18}\text{O}$ of the Chinese Dongge Cave stalagmite (Yuan et al., 2004), and (c) Mg/Ca SST record of the Baja California Sur, Soledad Basin (Marchitto et al., 2010).

the EEP, to end in a steady increase from 4 ka onward. A similar pattern was found in core MD01-2378 (13.08° S, 121.79° E, Timor Sea, eastern Indian Ocean; Xu et al., 2008). In both cases, it is argued that orbital configuration might have favored the northward shift of the ITCZ, resulting in intensification of the EAM, weakening of the AISM, and strengthening of southeastern trade winds during the austral summer. The timing and pattern difference between the stalagmite and the sedimentary records accounts for the regional variability, which is typical of the IPWP.

A high-resolution planktonic foraminifer Mg/Ca SST and $\delta^{18}\text{O}$ record of core MD052896 (southern South China Sea, SCS) provides a history of EAM dynamics over the past 23 ka, as indicated by the latitudinal temperature gradient (ΔSST ; Tian et al., 2010). The EAM was stronger during cold periods, including Heinrich 1 (H1, 17.5–14.7 ka) and the Younger Dryas (YD, 12.8–11.7 ka) events. The ΔSST record also indicates that after ~ 8.5 ka, the EAM strengthened relative to deglaciation, reaching similar levels to those of the last glacial period. By comparing the SCS ΔSST with stalagmite $\delta^{18}\text{O}$ records, it is suggested that both the East Asian summer monsoon (EASM) and East Asian winter monsoon (EAWM) strengthened from the last glacial to the Holocene, and that the migration of the ITCZ most probably has not been the only cause of EAM variability (Tian et al., 2010).

Reconstructed alkenone unsaturation index (U^{k}_{37}) SSTs from the SCS suggest warming trends through the Holocene (e.g., Kiefer and Kienast, 2005; Kim et al., 2004; Wang et al., 1999) in comparison with cooling trends, reconstructed with the foraminiferal Mg/Ca proxy, in the Philippine–Indonesian Seas region (Stott et al., 2004). The difference might be real or due to biases imposed by the various proxies used (e.g., due to seasonal, ecological, or diagenetic effects; Koutavas, 2004), or due to effects of the EAM and AIM dynamics. The divergence has also been observed in SSS trends, as shown by a high-resolution record from site 17940 (20° N; northern SCS) that suggests a progressive increase in salinity through the Holocene, consistent with the decreasing strength of the EASM (Wang et al., 1999). Moreover, power spectrum analysis of this salinity record reveals frequencies consistent with a possible solar forcing of monsoonal rainfall. Solar forcing of the EAM has also been proposed for the Dongge, China $\delta^{18}\text{O}$ stalagmite record on the basis of high-frequency precipitation changes (Wang et al., 2005). Salinity reconstructions from the Indonesian Seas, unlike the SCS, suggest freshening through the Holocene (Stott et al., 2004).

The connection between EAM dynamics and the cold tongue, based on U^{k}_{37} SSTs estimates, suggests that the zonal SST gradient in the southern SCS reflects the temporal dynamics of the winter ‘cold tongue’ off southern Vietnam and variations in the EAM intensity, and that winter ‘cold tongue’ SSTs were lower at H1 and the YD events, resulting in an increased zonal SST gradient in the southern SCS due to a strengthened EAM (Huang et al., 2011). Within dating uncertainties, an intensified EAM during cold stadials was apparently coeval with the shutdown or a reduction in strength of the Atlantic meridional overturning circulation (AMOC). The zonal SCS SST gradient also indicates an enhanced EAM during the early Holocene, which may be induced by postglacial ice-sheet dynamics and a strong seasonal contrast in solar insolation (Huang et al., 2011). In accord to Tian et al. (2010), results

suggest that the EAM was probably modulated by a complex interplay between the AMOC, solar insolation and ice-sheet dynamics on sub-orbital time scales.

Sedimentary geochemical records from subalpine Retreat Lake (24.5° N; 121.44° E; Taiwan) appear to indicate an unstable EAM climate over the past ~ 10.3 ka, with a weak EAM between ~ 10.3 and 8.6 ka, more intense peaks between 8.6 and 7.7 ka, and a gradual decrease in response to summer insolation, heat, and moisture transport (Selvaraj et al., 2007). The total organic carbon record apparently reveals several weak monsoon intervals that correlate to low SSTs in the western tropical Pacific and cold events in the North Atlantic. These weak EAM events at 8.2, 5.4, and 4.5–2.1 ka also appear to correspond to low values of atmospheric methane and periods of reduced AMOC (Selvaraj et al., 2007).

Further north in the NWP, the reconstruction of the EAM strength from the Hulu and Dongge (China) stalagmite $\delta^{18}\text{O}$ records reveals significant variability on millennial and sub-millennial timescales for the last glacial period and the Holocene (e.g., Dykoski et al., 2005; Wang et al., 2005; Yuan et al., 2004; Figure 3(b)). According to the stalagmite records, postglacial EAM strength peaked in the early to mid-Holocene (11.5 to ~ 3.5 ka) and gradually declined through the late Holocene, with a slight increase after 2 ka (Figure 3(b)). As expected, this pattern is opposite to the AISM (Figures 3(a) and 3(b)).

In addition to the apparent consistency of the EAM trend with forcing by Northern Hemisphere summer insolation, abrupt shifts in the $\delta^{18}\text{O}$ signal suggest that threshold effects may be important (Dykoski et al., 2005). An important observation of Dykoski et al. (2005) is that the Holocene EAM centennial- and multidecadal-scale monsoon variability in the Dongge record is significant, although not as large as glacial millennial-scale variability, that is, it is less than half the amplitude. Additionally, they show how this high-frequency variability correlates with the Ti record of the Cariaco Basin (Figure 2(b)), and that there is a strong connection with northern South American hydrological changes related to the dynamics of the ITCZ. Furthermore, Dykoski et al. (2005) found peaks at solar periodicities of 208 and 86 years in a spectral analysis of the $\delta^{18}\text{O}$ record, together with peaks at ENSO frequencies, suggesting that both solar forcing (cf. Emile-Geay et al., 2006) and changes in oceanic and atmospheric circulation patterns might have been important in controlling Holocene monsoon climate.

The Mg/Ca and $\delta^{18}\text{O}$ signals of *Globigerinoides ruber* SST and SSS reconstructions in core A7 (28° N) from the Okinawa Trough were used to explore the link between EAM strength and subtropical SSTs for the last 18 ka (Sun et al., 2005). Results show that even though there are common features in variability between monsoon strength and SST during the deglaciation interval, Holocene SSTs at A7 site appear to be more stable than the strength of the EAM and apparently SSS only increased slightly in the early–middle Holocene, which contrasts with the notion of a stronger EAM (Koutavas, 2004; Sun et al., 2005). Sun et al.’s (2005) interpretation is that there was a decoupling between sea surface conditions and the EAM in the Okinawa Trough, in response to an inland shift of monsoonal rainfall, resulting in a net precipitation decrease in the ocean during the early–middle Holocene. This

interpretation seems to be supported by (1) a correlation between an increased freshening and high productivity with high monsoon intensity in the East China Sea during interstadials including the Holocene (Chang et al., 2009), and (2) an SST and SSS reconstruction from two fossil corals from Kikai Island (28.33°N; 130.03°E), which show that at ~6.2 ka, SSTs were similar to today in summer and winter, whereas SSSs were higher both in summer (+0.5‰, +1.1 psu) and in winter (+0.2‰, +0.6 psu) than modern values (Morimoto et al., 2007).

The Kuroshio Current transports large amounts of heat from the IPWP to high latitudes. Therefore, the reconstruction of the Kuroshio strength, pathway, and volume are important paleoceanographic aspects. The Kuroshio Current was probably confined east of its present position during the last glacial interval (Ujiie et al., 2003), then it reentered the Okinawa Trough at ~7.3 ka and subsequently experienced substantial millennial-scale fluctuations, possibly related to Northern Hemisphere climate shifts. This is suggested by the abundance of *Pulleniatina obliquiloculata*, a proxy of the Kuroshio Current, and foraminiferal SST reconstructions (Jian et al., 2000). The timing of this reentry in the Okinawa Trough is supported by a reconstruction of the north–south SST gradient from a set of cores off central Japan; results suggest that the Kuroshio Current reached maximum strength at ~7 ka and progressively weakened toward the present day (Sawada and Handa, 1998). Jian et al. (2000) attributed the mid-Holocene peak in transport to an intensification of the subtropical gyre and the North equatorial current, which was driven by an insolation maximum in summer. The ENSO state may also have influenced the variability of the Kuroshio Current during the Holocene, that is, through cold tongue dynamics. Because the Kuroshio Current weakens today during El Niño events, by analogy, the proposed increase in ENSO frequency in the late Holocene (Moy et al., 2002) may have contributed to its reduction (Koutavas, 2004). Furthermore, the conspicuous decrease in abundance of *P. obliquiloculata* in cores from the Kuroshio source region and the SCS between ~4.5 and 2.7 ka, which suggests a marked decrease in the influence of the Kuroshio and associated climatic cooling, also suggests a link to the EAM (Jian et al., 2000). This interpretation appears consistent with a significant decrease in EAM strength at about the same time, as revealed in the stalagmite $\delta^{18}\text{O}$ records from Dongge Cave in China (Figure 3(b)). By contrast, Ujiie et al. (2003) suggest that the *P. obliquiloculata* event might not reflect sea surface cooling, but (a) a regional change of the subsurface structure of the Kuroshio, or (b) reduced primary productivity in the Kuroshio source region due to enhanced activity of the ENSO system.

Further support for a reduced Kuroshio Current after the middle Holocene is provided by Sagawa et al. (2011), who reconstruct variations in SST and vertical thermal structure in the western subtropical North Pacific, based on planktonic $\delta^{18}\text{O}$ and Mg/Ca records from core ASM-5PC (28.38°N; 132.75°E) for the last 30 ka, and conclude that when the EAWM strengthens, water convection increases and intermediate depth temperatures decrease. Therefore, by the mid-Holocene, when intermediate temperatures were warmer, the EAWM strength was weak. This reconstruction agrees with the Hulu and Dongge stalagmite $\delta^{18}\text{O}$ records.

Core PC6 (40°N, 143°E, off Sanriku, Japan), retrieved under the present-day Oyashio front in the southern Subarctic gyre, provided a record of biological production and U^{37}_{37} SST for the last 27 ka (Minoshima et al., 2007). Biogenic opal content was low until the early Holocene, and rapidly increased at 6 ka. Low biological production in the mixed layer resulted from nutrient depletion, especially silica, during the deglaciation (15–9.5 ka) in comparison to the late Holocene (6 ka – present) when biological productivity increased. Nutrients were probably supplied from deeper levels by eddies rather than by the Oyashio Current. Despite limitations in chronology, these results seem to reveal an opposite trend to the Kuroshio region, that is, a weakening of the Kuroshio Current was accompanied by a strengthening of the Oyashio Current after the middle Holocene (Minoshima et al., 2007).

From a set of cores (51°N, 152°E; Sea of Okhotsk), productivity in the source region of the NPIW was found to change from a coccolithophorid- to a diatom-dominated system in the mid-Holocene, which might have resulted from a salinity-induced breakdown of surface stratification, allowing the upwelling of silica-rich subsurface water (Seki et al., 2004). The hypothesis here is that the increase in salinity in the late Holocene may involve a decrease in East Asian continental rainfall and a broad reorganization of the EAM (Seki et al., 2004).

In a more oceanic setting, core MD01-2416 (51°N, 168°E) provides a centennial-scale multiproxy record of subarctic North Pacific SST, SSS, and productivity for the Holocene (Sarnthein et al., 2004). Results reveal a gradual decrease in SST by 3–5°C from 11.1 to 4.2 ka and a stepwise long-term decrease in SSS by 2–3 psu. early Holocene SSSs were as high as in the modern subtropical Pacific. The steep halocline and stratification that characterizes present-day subarctic North Pacific waters is the product of mid-Holocene environmental change. High SSSs matched a productivity maximum of biogenic opal from the Bølling interstadial interval to the early Holocene (Sarnthein et al., 2004).

Finally, at the northern edge of the NWP, the $\delta^{18}\text{O}$ benthic Foraminifera of cores GC-11 (53.5°N, 178.85°E) and G-13 (53.68°N, 178.72°E, Bering Sea) provide a record of the last glacial to the mid-Holocene changes in global deepwater $\delta^{18}\text{O}$ and a rise in deepwater temperature from 1.2°C to 1.6°C between 8.5 and 6.5 ka. The subsequent environmental changes in the late Holocene led to modern-day conditions, characterized by a strong stratification and reduced deepwater ventilation (Gorbarenko et al., 2010).

Paleoceanographic Records from the NEP

High-resolution records of diatoms, silicoflagellates, and geochemistry (U and Mo) from the Gulf of Alaska show a decrease in productivity during the YD, which was followed by an increase during the early Holocene, a pattern that is analogous to several locations on the California margin. The increase in abundance of upwelling-related diatoms and Mo suggest that low oxygen bottom conditions were more intense between ~6.5 and 2.8 ka (Barron et al., 2009). In a similar approach, diatoms, silicoflagellates, and biogenic silica were analyzed from two cores from the Effingham Inlet, British

Columbia, where conditions of relatively high productivity were found between 4.85 and 4 ka, followed by a transition to the modern, and variable, ocean–climate regime since 2.8 ka, except for a brief period of warmer and drier conditions and possibly increased coastal upwelling offshore ca. 1.45–1.05 ka (Hay et al., 2007). Even though there are some common features between the Alaskan and British Columbia records, such as the 2.8 ka event, other differences possibly reveal regional variability. By contrast, the U^{k}_{37} SST reconstruction from core JT96-06 PC (49° N, Canadian margin) exhibits a strong YD cooling, followed by a warm peak in the early Holocene, cooling in the middle Holocene and a slight return to warmer SSTs in the late Holocene (Kienast and McKay, 2001). This SST pattern is similar to ODP 1019 from northern California (Barron et al., 2003), therefore suggesting a broad pattern of climate variability for the NEP.

The California Margin, and in particular the Santa Barbara basin (SBB), has been the focus of high-resolution paleoceanographic studies for the late Quaternary due to the influence of the California Current and the restricted bathymetry of the basins, which make sedimentary records sensitive to surface (upwelling) and bottom (oxygen) water properties, which in turn respond to millennial- and submillennial-scale climate variations (e.g., Behl and Kennett, 1996; Hendy and Kennett, 1999; Kennett and Ingram, 1995).

The study of diatoms, pollen census data, and U^{k}_{37} in ODP site 1019 (42° N, off northern California) revealed a strong cooling during the YD, warming in the early Holocene (11.5–8.5 ka), cooling by 1–2 °C in the middle Holocene (8.5–3.5 ka), and then returning to present-day values (Barron et al., 2003). The mid-Holocene cool interval is associated with opposite variations in the abundance of diatoms, that is, the decrease of the gyre species *Pseudoemotia doliolus* at the expense of the increase of the upwelling species *Thalassionema nitzschioides*, suggesting that intensification in upwelling was at least partly the cause of cooler temperatures (Barron et al., 2003). This interpretation is in accord with the interpretations of Gardner et al. (1997) and Barron and Burky (2007), who concluded that increases in biogenic mass accumulation rates along the California margin resulted from enhanced productivity due to stronger upwelling during the mid-Holocene. This would mean a stronger subtropical gyre in a La Niña-like scenario.

Pollen, geochemical (organic carbon, carbonate, and biogenic silica), and U^{k}_{37} analyses of ODP site 1018 (36.98° N, 123.28° W) provide evidence of wet conditions associated with a rapid rise in SST during deglaciation (Lyle et al., 2010). While SST gradients along the California margin suggest the weakening of the California Current and coastal upwelling, they proposed that wet periods in central California are connected to the tropical Pacific by the northward movement of temperate Pacific water into the Alaska gyre. The wettest period in the SBB peaked at 16 ka, preceding the wet peak in central and northern California by 4 ka (Lyle et al., 2010).

Bottom-water oxygenation levels at ODP site 893 (500 m sill depth; 34° N; California margin) were reconstructed using a bioturbation index, where laminated sediments are indicative of low oxygen conditions and warm climate intervals (Behl and Kennett, 1996). As for the Holocene period, coccolith assemblages and biogenic burial rates at ODP site 893 show that bioturbation varied with oxygen changes at the sea floor,

compared to the lamination that varied with lithogenic particles, thus favoring the interpretation that deepwater ventilation (Nederbragt et al., 2008) and the strength of the NPIW must have been important (Koutavas, 2004). In fact, the northern origin of intermediate water masses on postglacial times is supported by carbon isotope ($\delta^{13}C$) measurements on *Cibicides wuellerstorfi* in the EEP core MD02-2529 (8.2° N; 84.12° W; Leduc et al., 2010). Furthermore, in support of the NPIW hypothesis, Roark et al. (2003) measured the radiocarbon age to document the difference between benthic and planktonic Foraminifera in the SBB, and found that there is some degree of consistency between it and the Holocene bioturbation index, a connection which is probably related to the strength or source of NPIW and the AMOC. Of interest is the significant increase in the benthic–planktonic age difference at ~2 ka BP, which has persisted to the present day. This change is close to the one reported for Effingham Inlet (British Columbia, Hay et al., 2007) and other sites in the EEP (e.g., El Junco, Galapagos Is., Conroy et al., 2008). This change implies a shift to weaker NPIW ventilation in the latest Holocene, which coincides with the presence of increased laminations in the sediments (Roark et al., 2003).

Measurements of $\delta^{18}O$ in the planktonic Foraminifera *Globigerina bulloides* (a surface dweller) and *Neogloboquadrina pachyderma* (a thermocline dweller) in core AII-125 JPC-76 (34° N, SBB), and their isotopic difference ($\Delta\delta^{18}O$), were used to interpret the degree of surface stratification in relation to ENSO system dynamics, that is, positive ENSO anomalies led to increased stratification in the SBB, and vice versa (Friddell et al., 2003). Their results show a somewhat reduced isotopic difference (reduced stratification) between 9 and 5.5 ka, implying a decline in ENSO variability, that is, a La Niña-like early–middle Holocene state, documented in the tropical Pacific (Figure 2(a)). Conversely, a larger isotopic gradient (increased stratification) and enhanced decadal-scale surface variability between 5.5 and 3.5 ka may imply the onset of a strong ENSO system. In addition to the influence of the ENSO system on the SBB, Friddell et al. (2003) conclude that the variability of the PDO might significantly help explain the apparent weakening of the ENSO system by ~3.5 ka. By contrast, Barron and Burky (2007) point to enhanced tropical variability, which is not restricted to the Pacific, but has been documented in the Cariaco basin (Haug et al., 2001).

Multiscale dynamic interactions affecting the California Current were explored by Diffenbaugh and Ashfaq (2007), who used a Regional Climate Model (RegCM3) and suggested that mid-Holocene insolation might have caused systematic changes in wind forcing of the current, suppressing and enhancing wind stress in spring, and summer–autumn, respectively. Their climate model simulations for the mid-Holocene are supported by proxy-inferred cooler-than-present SSTs as well as by high $\delta^{15}N$ values in the northern and southern California Current system, respectively (Diffenbaugh and Ashfaq, 2007).

Diatom and silicoflagellate data from GGC55, JPC56, and DSDP 480 cores in the Guaymas Basin (Gulf of California) show a pattern of reduced productivity, which suggests cool SSTs in the early–middle Holocene until ~6 ka when productivity increased (Barron et al., 2005). This pattern apparently

parallels the evolution of the ENSO system through the Holocene (Barron et al., 2005).

Further south, and analogous to the SST pattern of the EEP and numerical model simulations of orbital forcing, the Mg/Ca SST record of the Soledad Basin (25.2° N; 112.7° W; Baja California, Mexico) shows cool values during the early–middle Holocene (Figure 3(c); Marchitto et al., 2010). In the early Holocene, millennial-scale fluctuations in SST correlate with cosmogenic nuclide (proxies of solar variability), with inferred solar minima corresponding to El Niño-like conditions, in apparent agreement with the theoretical ‘ocean dynamical thermostat’ response of ENSO to exogenous radiative forcing. The results of Marchitto et al. (2010) are in agreement with a previous interpretation of the $\delta^{18}\text{O}$ record of a stalagmite from the Pink Panther Cave (New Mexico) which was used to reconstruct the dynamics of the North American monsoon (Asmerom et al., 2007).

Through the study of core MV-99, GC-31 (23.47° N, 111.60° W; Magdalena margin, Baja California, Mexico), Ortiz et al. (2004) linked changes in bottom-water oxygen content to millennial-scale climate changes, as recorded by the $\delta^{18}\text{O}$ composition of Greenland ice, and suggested that this teleconnection predominantly resulted from changes in marine productivity, rather than changes in ventilation of the North Pacific. In accord with other studies, the interpretation is that the deepening of the thermocline and weakening of the nutricline during El Niño events reduce productivity (e.g., Barron et al., 2005).

The postglacial paleoceanography of the NEP is synthesized by Barron and Anderson (2011), who indicate that there were (1) La Niña-like or more negative PDO-like conditions, during the middle Holocene (8–4 ka), and (2) a climate transition between ~4.2 and 3 ka, leading to modern ENSO variability. More detailed interpretations, in relation to modern-day conditions, suggest that during the middle Holocene, the Aleutian Low was weaker during the winter and/or located to the west, while the North Pacific High was stronger during the summer and located to the north. Coastal upwelling off California was more vigorous during the summer and fall but suppressed during the spring, whereas Oregon and California SSTs were cooler compared to the SBB. late Holocene records indicate a more variable, El Niño-like and more positive, PDO-like system, with the Aleutian Low more intense during the winter and/or located to the east, and the North Pacific High weaker and/or displaced to the south. Coastal upwelling off California intensified during the spring but decreased during the fall. Oregon and California SSTs became warmer, recording the shoreward migration of the subtropical gyre during the fall, while spring upwelling increased in the SBB. The high-resolution proxy records indicate enhanced ENSO and PDO variability after 4 ka off southern California, 3.4 ka off northern California, and by 2 ka in southwestern Yukon. A progressively northward migration of the ENSO teleconnection during the late Holocene is proposed. This 2-ka time difference contrasts with the 4-ka differential reported for deglaciation by Lyle et al. (2010), as indicated before.

For the center of the North Pacific gyre, the carbon isotope record from Ordy Pond (21.3° N, 158.05° W, Oahu Is. Hawaii) provides evidence for an increasing C_4 trend with respect to C_3 plants along the Holocene, thus suggesting that the central

subtropical Pacific became drier, possibly due to a winter warming of the cold tongue in the EEP which was orbitally forced in an analogous way to the modern ENSO phenomenon (Uchikawa et al., 2010). Except for two major peaks at 5 and 0.5 ka, the trend seems coherent with SST reconstructions, and cold tongue weakening in the EEP (e.g., Kienast et al., 2006; Koutavas and Sachs, 2008; Leduc et al., 2007).

The ENSO System at Millennial Timescales

Besides the proxy recognition of ENSO-like states during the Holocene, there has been a search for their forcing mechanisms. This is the case of the ~1 or ~1.5 ka periodicity in SST variability at millennial timescales in the North Pacific during the Holocene, which is attributed to solar and orbital forcing that is still a matter of debate (e.g., Ditlevsen et al., 2007; Isono et al., 2009; Marchitto et al., 2010). One proposition is that tropical convection, controlled by the meridional location of the ITCZ, controls the emissions of the greenhouse gases CH_4 and N_2O and acts as a global climate amplifier (Ivanochko et al., 2005). In accord with proxy evidence, model results based on radiocarbon production suggest that there was an increase in ENSO variability and that it may have acted as a mediator between the Sun and the Earth’s climate (Emile-Geay et al., 2006). The increase in ENSO variability, reduced during the mid-Holocene, is evident from the spectral analysis of 124 proxy time series (Wirtz et al., 2010).

To mention one model, Lorenz et al. (2006) used a coupled atmosphere–ocean circulation model to explore insolation forcing on U^{k}_{37} global SST trends over the last 7 ka. They found that the extratropics cooled while the tropics warmed during the middle to late Holocene, in response to seasonally opposite insolation changes, and that climate mode changes similar to the Arctic/North Atlantic Oscillation are superimposed on the prevalent pattern. Furthermore, nonlinear changes in the seasonal cycle of insolation appear to be a key factor in the evolution of Holocene SSTs (Lorenz et al., 2006). An important observation of this experiment is that the timing of phytoplankton production is controlled not only by summer insolation in high latitudes, but also by the shift in the maximum insolation in low latitudes. This might explain the timing differences along the California margin described earlier. The difficulties reproducing accurately the mid-Holocene La Niña-like state are summarized and discussed by Chiang (2009).

Summary

From this compilation, it is apparent that the influence of the ENSO system extends all over the North Pacific during the Holocene, and that for centuries to thousands of years, El Niño-like and La Niña-like conditions can become the dominant state. Apart from a general pattern where the Kuroshio Current is weakened and the California Current is strengthened during El Niño- and La Niña-like states, respectively, there are significant regional differences. These are imposed by the interaction between the EAM and AIM in the IPWP and the northwest Pacific, and the PDO and the NPIW in the

northeast Pacific. Of major interest is the recognition that the AISM is not only controlled by solar insolation, but also by sea-level change. Similarly, more robust evidence is building for teleconnections between the EAM, the California Current, and the AMOC, modulated by insolation and ice-sheet dynamics on suborbital time scales. The general paleoceanographic picture for the North Pacific in the Holocene is one of a La Niña-like (negative PDO) state during the early–middle Holocene to one of El Niño-like (positive PDO) state during the late Holocene. Therefore, the ITCZ changed from a northern to a southern location from the middle Holocene to the late Holocene. This might have reduced the strength of the Kuroshio and California Currents.

See also: [Glacial Climates: Volcanic and Solar Forcing.](#)
Introduction: [Understanding Quaternary Climatic Change.](#)
Paleoceanography, Records: [Late Pleistocene North Pacific.](#)
Paleoclimate Modeling: [Paleo-ENSO.](#)

References

- Anderson RF, Fleisher MQ, and Lao Y (2006) Glacial–interglacial variability in the delivery of dust to the central equatorial Pacific Ocean. *Earth and Planetary Science Letters* 242: 406–414.
- Ashok K, Behera SK, Rao SA, Weng H, and Yamagata T (2007) El Niño Modoki and its possible teleconnection. *Journal of Geophysical Research* 112: C11007. <http://dx.doi.org/10.1029/2006JC003798> 27 pp.
- Asmerom Y, Polyak V, Burns S, and Rasmussen J (2007) Solar forcing of Holocene climate: New insights from a speleothem record, southwestern United States. *Geology* 35: 1–4.
- Barron JA and Anderson L (2011) Enhanced Late Holocene ENSO/PDO expression along the margins of the eastern North Pacific. *Quaternary International* 235: 3–12.
- Barron JA and Burky D (2007) Development of the California Current during the past 12,000 yr based on diatoms and silicoflagellates. *Paleogeography, Paleoclimatology, Paleoecology* 248: 313–338.
- Barron JA, Burky D, and Dean WE (2005) Paleoclimatographic history of the Guaymas Basin, Gulf of California, during the past 15,000 years based on diatoms, silicoflagellates, and biogenic sediments. *Marine Micropaleontology* 56: 81–102.
- Barron JA, Burky D, Dean WE, and Addison JA (2009) Paleoclimatography of the Gulf of Alaska during the past 15,000 years: Results from diatoms, silicoflagellates, and geochemistry. *Marine Micropaleontology* 72: 176–195.
- Barron JA, Heusser L, Herbert T, and Lyle M (2003) High-resolution climatic evolution of coastal northern California during the past 16,000 years. *Paleoceanography* 18. <http://dx.doi.org/10.1029/2002PA000768>.
- Beaufort L, van der Kaars S, Bassinot FC, and Moron V (2010) Past dynamics of the Australian monsoon: Precession, phase and links to the global monsoon concept. *Climate of the Past* 6: 695–706. <http://dx.doi.org/10.5194/cp-6-695-2010>. www.clim-past.net/6/695/2010/.
- Behl RJ and Kennett JP (1996) Brief interstadial events in the Santa Barbara basin, NE Pacific, during the past 60 kyr. *Nature* 379: 243–246.
- Berger A and Loutre MF (1991) Insolation values for the climate of the last 10 million years. *Quaternary Science Reviews* 10: 297–317.
- Briker JM, Jung SJA, Ganzen GM, Bickert T, and Kroon D (2007) ENSO related decadal scale climate variability from the Indo-Pacific warm pool. *Earth and Planetary Science Letters* 253: 67–82.
- Cane MA (2005) The evolution of El Niño, past and future. *Earth and Planetary Science Letters* 230: 227–240.
- Chang Y-P, Chen M-T, Yokoyama Y, et al. (2009) Monsoon hydrography and productivity changes in the East China Sea during the past 100,000 years: Okinawa Trough evidence (MD012404). *Paleoceanography* 24: PA3208. <http://dx.doi.org/10.1029/2007PA001577>.
- Chiang JCH (2009) The tropics in paleoclimate. *Annual Review of Earth and Planetary Sciences* 37: 263–297.
- Clement AC, Seager R, and Cane MA (2000) Suppression of El Niño during the mid-Holocene by changes in the Earth's orbit. *Paleoceanography* 15: 731–737.
- Conroy JL, Overpeck JT, Cole JE, Sanan TM, and Steinitz-Kannan M (2008) Holocene changes in eastern tropical Pacific climate inferred from a Galápagos lake sediment record. *Quaternary Science Reviews* 27: 1166–1180.
- Dai A (2006) Precipitation characteristics in eighteen coupled climate models. *Journal of Climate* 19: 4605–4630.
- Diffenbaugh NS and Asfaq M (2007) Response of California Current forcing to mid-Holocene insolation and sea surface temperatures. *Paleoceanography* 22: PA3101. <http://dx.doi.org/10.1029/2006PA001382>.
- Ditlevsen PD, Andersen KK, and Svensson A (2007) The DO-climate events are probably noise induced: Statistical investigation of the claimed 1470 years cycle. *Climate of the Past* 3: 129–134.
- Dubois N, Kienast M, Kienast S, Calvert SE, François R, and Anderson RF (2010) Sedimentary opal records in the eastern equatorial Pacific: It is not all about leakage. *Global Biogeochemical Cycles* 24: GB4020. <http://dx.doi.org/10.1029/2010GB003821>.
- Dykoski CA, Edwards RL, Cheng H, et al. (2005) A high-resolution, absolute-dated Holocene and deglacial Asian monsoon record from Dongge Cave, China. *Earth and Planetary Science Letters* 233: 71–86.
- Emile-Geay J, Cane M, Seager R, Kaplan A, and Almasi P (2006) ENSO as a mediator of the solar influence on climate. *Paleoceanography* 21: PA1207. <http://dx.doi.org/10.1029/2007JEG123456>.
- Fridell JE, Thunell RC, Guilderson TP, and Kashgarian M (2003) Increased northeast Pacific climatic variability during the warm middle Holocene. *Geophysical Research Letters* 30. <http://dx.doi.org/10.1029/2002GL016834>.
- Gagan MK, Hendy EJ, Haberle SG, and Hantoro WS (2004) Post-glacial evolution of the Indo-Pacific warm pool and El Niño–Southern Oscillation. *Quaternary International* 2004: 127–143.
- Gardner JV, Dean WE, and Dartnell P (1997) Biogenic sedimentation beneath the California Current system for the past 30 kyr and its paleoceanographic significance. *Paleoceanography* 12: 207–225.
- Gorbarenko SA, Wang P, Wang W, and Cheng X (2010) Orbital and suborbital environmental changes in the southern Bering Sea during the last 50 kyr. *Paleogeography, Paleoclimatology, Paleoecology* 286: 97–106.
- Griffiths ML, Drysdale RN, Gagan MK, et al. (2009) Increasing Australian–Indonesian monsoon rainfall linked to early Holocene sea-level rise. *Nature Geoscience* 2: 636–639.
- Haug GH, Hughen KA, Sigman DM, Peterson LC, and Röhl U (2001) Southward migration of the intertropical convergence zone through the Holocene. *Science* 293: 1304–1308.
- Hay MB, Dallimore A, Thomson RE, Calvert SE, and Pienitz R (2007) Siliceous microfossil record of late Holocene oceanography and climate along the west coast of Vancouver Island, British Columbia (Canada). *Quaternary Research* 67: 33–49.
- Hayes CT, Anderson RF, and Fleisher MQ (2011) Opal accumulation rates in the equatorial Pacific and mechanisms of deglaciation. *Paleoceanography* 26: PA1207. <http://dx.doi.org/10.1029/2010PA002008>.
- Hendy IL and Kennett JP (1999) Latest Quaternary North Pacific surface-water responses imply atmosphere-driven climate instability. *Geology* 27: 291–294.
- Huang E, Tian J, and Steinke S (2011) Millennial-scale dynamics of the winter cold tongue in the southern South China Sea over the past 26 ka and the East Asian winter monsoon. *Quaternary Research* 75: 196–204.
- Isono D, Yamamoto M, Irino T, et al. (2009) The 1500-year climate oscillation in the midlatitude North Pacific during the Holocene. *Geology* 37(7): 591–594.
- Ivanochko TS, Ganeshram RS, Brummer G-JA, et al. (2005) Variations in tropical convection as an amplifier of global climate change at the millennial scale. *Earth and Planetary Science Letters* 235: 302–314.
- Jian Z, Wang P, Saito Y, et al. (2000) Holocene variability of the Kuroshio Current in the Okinawa Trough, northwestern Pacific Ocean. *Earth and Planetary Science Letters* 184: 305–319.
- Kennett JP and Ingram BL (1995) A 20,000-year record of ocean circulation and climate change from the Santa Barbara basin. *Nature* 377: 510–514.
- Kiefer T and Kienast M (2005) Patterns of deglacial warming in the Pacific Ocean: A review with emphasis on the time interval of Heinrich event 1. *Quaternary Science Reviews* 24: 1063–1081.
- Kienast M, Kienast SS, Calvert SE, et al. (2006) Eastern Pacific cooling and Atlantic overturning circulation during the last deglaciation. *Nature* 443: 846–849.
- Kienast SS, Kienast M, Mix AC, Calvert SE, and François R (2007) Thorium-230 normalized particle flux and sediment focusing in the Panama Basin region during the last 30,000 years. *Paleoceanography* 22: PA2213. <http://dx.doi.org/10.1029/2006PA001357>.
- Kienast M, Lehmann MF, Timmermann A, et al. (2008) A mid-Holocene transition in the nitrogen dynamics of the western equatorial Pacific: Evidence of a deepening thermocline? *Geophysical Research Letters* 35: L23610. <http://dx.doi.org/10.1029/2008GL035464>.

- Kienast SS and McKay JL (2001) Sea surface temperatures in the subarctic Northeast Pacific reflect millennial-scale climate oscillations during the last 16 kyr. *Geophysical Research Letters* 28: 1563–1566.
- Kim J-H, Rimbu N, Lorenz SJ, et al. (2004) North Pacific and North Atlantic sea-surface temperature variability during the Holocene. *Quaternary Science Reviews* 23: 2141–2154.
- Koutavas A (2004) Postglacial paleoceanographic records of the North Pacific Ocean. *Encyclopedia of Quaternary Sciences*. Amsterdam: Elsevier.
- Koutavas A, deMenocal PB, Olive GC, and Lynch-Stieglitz J (2006) Mid-Holocene El Niño–Southern Oscillation (ENSO) attenuation revealed by individual foraminifera in eastern tropical Pacific sediments. *Geology* 34(12): 993–996.
- Koutavas A, Lynch-Stieglitz J, Marchitto TM, and Sachs JP (2002) El Niño-like pattern in ice age tropical Pacific sea surface temperature. *Science* 297: 226–230.
- Koutavas A and Sachs JP (2008) Northern timing of deglaciation in the eastern equatorial Pacific from alkenone paleothermometry. *Paleoceanography* 23: PA4205. <http://dx.doi.org/10.1029/2008PA001593>.
- Langton SJ, Linsley BK, Robinson RS, et al. (2008) 3500 yr record of centennial-scale climate variability from the Western Pacific warm pool. *Geology* 36(10): 795–798. <http://dx.doi.org/10.1130/G24926A>.
- Leduc G, Vidal L, Tachikawa K, and Bard E (2009) ITCZ rather than ENSO signature for abrupt climate changes across the tropical Pacific? *Quaternary Research* 72: 123–131.
- Leduc G, Vidal L, Tachikawa K, and Bard E (2010) Changes in Eastern Pacific ocean ventilation at intermediate depth over the last 150 kyr BP. *Earth and Planetary Science Letters* 298: 217–228.
- Leduc G, Vidal L, Tachikawa K, et al. (2007) Moisture transport across Central America as a positive feedback on abrupt climatic changes. *Nature* 445: 908–911.
- Liu Z, Brady E, and Lynch-Stieglitz J (2003) Global ocean response to orbital forcing in the Holocene. *Paleoceanography* 18: <http://dx.doi.org/10.1029/2002PA000819>.
- Liu WT and Tang W (2005) Estimating moisture transport over oceans using space-based observations. *Journal of Geophysical Research* 110: D10101. <http://dx.doi.org/10.1029/2004JD005300>.
- Lorenz SJ, Kim J-H, Rimbu N, Schneider RR, and Lohmann G (2006) Orbitally driven insolation forcing on Holocene climate trends: Evidence from alkenone data and climate modeling. *Paleoceanography* 21: PA1002. <http://dx.doi.org/10.1029/2005PA001152>.
- Lyle M, Heusser L, Ravelo C, Andreasen D, Olivarez Lyle A, and Diffenbaugh N (2010) Pleistocene water cycle and eastern boundary current processes along the California continental margin. *Paleoceanography* 25: PA4211. <http://dx.doi.org/10.1029/2009PA001836>.
- Marchitto TM, Mischler R, Ortiz JD, Carriquiry JD, and van Geen A (2010) Dynamical response of the tropical Pacific Ocean to solar forcing during the early Holocene. *Science* 330: 1378–1381.
- Martínez JL, Rincon D, Yokoyama Y, and Barrows T (2006) Foraminifera and coccolithophorid assemblage changes in the Panama Basin during the last deglaciation: Response to sea-surface productivity induced by a transient climate change. *Paleogeography, Palaeoclimatology, Palaeoecology* 234: 114–126.
- Martínez P and Robinson RS (2010) Increase in water column denitrification during the last deglaciation: The influence of oxygen demand in the eastern equatorial Pacific. *Biogeosciences* 7: 1–9.
- McGee D, Marcantonio F, and Lynch-Stieglitz J (2007) Deglacial changes in dust flux in the eastern equatorial Pacific. *Earth and Planetary Science Letters* 257: 215–230.
- Minoshima K, Kawahata H, and Ikehara K (2007) Changes in biological production in the mixed water region (MWR) of the northwestern North Pacific during the last 27 kyr. *Paleogeography, Palaeoclimatology, Palaeoecology* 254: 430–447.
- Morimoto M, Kayanne H, Abe O, and McCulloch MT (2007) Intensified mid-Holocene Asian monsoon recorded in corals from Kikai Island, subtropical northwestern Pacific. *Quaternary Research* 67: 204–214.
- Moy CM, Seltzer GO, Rodbell DT, and Anderson DM (2002) Variability of El Niño/Southern Oscillation activity at millennial time scales during the Holocene epoch. *Nature* 420: 162–165.
- Nederbragt AJ, Thurnow JW, and Bown PR (2008) Paleoproductivity, ventilation, and organic carbon burial in the Santa Barbara Basin (ODP Site 893, off California) since the last glacial. *Paleoceanography* 23: PA1211. <http://dx.doi.org/10.1029/2007PA001501> 15 pp.
- Ortiz JD, O'Connell SB, DelViscio J, et al. (2004) Enhanced marine productivity off western North America during warm climate intervals of the past 52 ky. *Geology* 32(6): 521–524. <http://dx.doi.org/10.1130/G20234.1>.
- Peña LD, Cacho I, Ferretti P, and Hall MA (2008) El Niño–Southern Oscillation-like variability during glacial terminations and interlatitudinal teleconnections. *Paleoceanography* 23: PA3101. <http://dx.doi.org/10.1029/2008PA001620>.
- Peña LD, Jones KM, Goldstein SL, et al. (2010) Advection of Southern Ocean intermediate waters to the tropics during the last deglaciation from Nd isotopes in foraminifera. In: *10th International Conference on Paleoceanography, Scripps Institution of Oceanography, University of California, San Diego, La Jolla, CA, 29 August–3 September*.
- Poveda G, Waylen PR, and Pulwarty R (2006) Modern climate variability in northern South America and southern Mesoamerica. *Paleogeography, Palaeoclimatology, Palaeoecology* 234: 3–27.
- Roark EB, Ingram BL, Southon J, and Kennett JP (2003) Holocene foraminiferal radiocarbon record of paleocirculation in the Santa Barbara Basin. *Geology* 31: 379–382.
- Robinson RS, Martínez P, Peña LD, and Cacho I (2009) Nitrogen isotopic evidence for deglacial changes in nutrient supply in the eastern equatorial Pacific. *Paleoceanography* 24: PA4213. <http://dx.doi.org/10.1029/2008PA001702>.
- Rodbell DT, Seltzer GO, Anderson DM, Abbott MA, Enfield DB, and Newman JH (1999) An 15,000-year record of El Niño-driven alluviation in Southwestern Ecuador. *Science* 283: 516–520.
- Sagawa T, Yokoyama Y, Ikehara M, and Kuwae M (2011) Vertical thermal structure history in the western subtropical North Pacific since the Last Glacial Maximum. *Geophysical Research Letters* 38: L00F02. <http://dx.doi.org/10.1029/2010GL045827>.
- Sandweiss DH, Maasch KA, Burger RL, Richardson JB, Rollins HB, and Clement A (2001) Variation in Holocene El Niño frequencies: Climate records and cultural consequences in ancient Peru. *Geology* 29: 603–606.
- Sarnthein M, Gebhardt H, Kiefer T, Lucera M, Cook M, and Erlenkeuser H (2004) Mid-Holocene origin of the sea-surface salinity low in the subarctic North Pacific. *Quaternary Science Reviews* 23: 2089–2099.
- Sawada K and Handa N (1998) Variability of the Kuroshio ocean current over the past 25,000 years. *Nature* 392: 592–595.
- Seki O, Ikehara M, Kawamura K, et al. (2004) Reconstruction of paleoproductivity in the Sea of Okhotsk over the last 30 kyr. *Paleoceanography* 19: <http://dx.doi.org/10.1029/2002PA000808>.
- Selvaraj K, Chen CTA, and Lou J-Y (2007) Holocene East Asian monsoon variability: Links to solar and tropical Pacific forcing. *Geophysical Research Letters* 34: L01703. <http://dx.doi.org/10.1029/2006GL028155>.
- Stott L, Cannariato K, Thunell R, Haug G, Koutavas A, and Lund S (2004) Decline of surface temperature and salinity in the western tropical Pacific Ocean in the Holocene epoch. *Nature* 431: 56–59.
- Sun Y, Oppo DW, Xiang R, Liu W, and Gao S (2005) Last deglaciation in the Okinawa Trough: Subtropical northwest Pacific link to Northern Hemisphere and tropical climate. *Paleoceanography* 20: <http://dx.doi.org/10.1029/2004PA001061>.
- Tian J, Huang E, and Pak DK (2010) East Asian winter monsoon variability over the last glacial cycle: Insights from a latitudinal sea-surface temperature gradient across the South China Sea. *Paleogeography, Palaeoclimatology, Palaeoecology* 292: 319–324.
- Tomczak M and Godfrey JS (1994) *Regional Oceanography: An Introduction*. Oxford: Pergamon Press.
- Uchikawa J, Popp BN, Schoonmaker JE, Timmermann A, and Lorenz SJ (2010) Geochemical and climate modeling evidence for Holocene aridification in Hawaii: Dynamic response to a weakening equatorial cold tongue. *Quaternary Science Reviews* 29: 3057–3066.
- Ujiié Y, Ujiié H, Taira A, Nakamura T, and Oguri K (2003) Spatial and temporal variability of surface water in the Kuroshio source region, Pacific Ocean, over the past 21,000 years: Evidence from planktonic foraminifera. *Marine Micropaleontology* 49: 335–364.
- Waliser DE and Gautier C (1993) A satellite-derived climatology of the ITCZ. *Journal of Climate* 6: 2162–2174.
- Wang Y, Cheng H, Edwards RL, et al. (2005) The Holocene Asian monsoon: Links to solar changes and North Atlantic climate. *Science* 308: 854–857.
- Wang L, Sarnthein M, Erlenkeuser H, et al. (1999) East Asian monsoon climate during the Late Pleistocene: High-resolution sediment records from the South China Sea. *Marine Geology* 156: 245–284.
- Wanner H, Beer J, Bütikofer J, et al. (2008) Mid- to late Holocene climate change: An overview. *Quaternary Science Reviews* 27: 1791–1828.
- Warnock J, Scherer R, and Loubere P (2007) A quantitative assessment of diatom dissolution and late quaternary primary productivity in the eastern equatorial Pacific. *Deep-Sea Research Part II* 54: 772–783.
- Wirtz KW, Lohmann G, Bernhardt K, and Lemmen C (2010) Mid-Holocene regional reorganization of climate variability: Analyses of proxy data in the frequency domain. *Paleogeography, Palaeoclimatology, Palaeoecology* 298(2010): 189–200.
- Xie S-P, Okumura Y, Miyama T, and Timmermann A (2008) Influences of Atlantic climate change on the tropical Pacific via the Central American isthmus. *Journal of Climate* 21: 3914–3928.

- Xu J, Holbourn A, Kuhnt W, Jian Z, and Kawamura H (2008) Changes in the thermocline structure of the Indonesian outflow during Terminations I and II. *Earth and Planetary Science Letters* 273: 152–162.
- Yu J-Y and Kim ST (2010) Three evolution patterns of Central Pacific El Niño. *Geophysical Research Letters* 37: L08706. <http://dx.doi.org/10.1029/2010GL042810>.
- Yuan D, Cheng H, Edwards RL, et al. (2004) Timing, duration, and transitions of the last interglacial Asian monsoon. *Science* 304: 575–578.
- Zaucker F and Broecker WS (1992) The influence of atmospheric moisture transport on the fresh water balance of the Atlantic drainage basin: General circulation model simulations and observations. *Journal of Geophysical Research* 97: 2765–2773.

Postglacial South Pacific

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Introduction

The South Pacific Ocean covers a major part of the Southern Hemisphere (SH) and plays an important role in the global climate system. Nevertheless, postglacial (Holocene) paleoceanographic records from this ocean are scarce because large parts are covered by deep sea basins, often too deep for carbonate preservation and thus not suitable to obtain records with sufficient time control and resolution (Figure 1). Therefore, most available Holocene paleoceanographic data originate from the margins of the surrounding continents where higher terrigenous sediment supply, combined with enhanced biological productivity and preservation, provides a much better potential for studying Holocene paleoenvironmental variability. After briefly discussing the modern oceanographic and climatic background, the main focus of this article concerns postglacial records from the continental margins off South America (Peru and Chile), New Zealand/Australia, and the Antarctic Peninsula (Palmer Deep). The available paleoceanographic data primarily focus on surface ocean records, such as sea surface temperatures (SSTs), paleosalinity, and paleoproductivity estimates. These data are complemented with information on terrestrial paleoclimates, as derived from the marine records. In addition, major forcing mechanisms involved in Holocene paleoenvironmental variability in the South Pacific realm are summarized, and critical issues for future research are identified.

Oceanographic and Climatic Setting

General Ocean Circulation

The South Pacific Ocean is dominated by deep basins, including the Peru and Chile basins in the eastern part, and the large southwest Pacific basin in the west (Figure 1; Tomczak and Godfrey, 2003). These basins are separated from each other by oceanic ridge systems, the most important being the Nazca Ridge and the East Pacific Rise. In the westernmost part of the South Pacific, New Zealand and the Melanesian Islands build the boundary for smaller adjacent basins including the Tasman Sea, the South Fiji Basin, and the Coral Sea. Deep basins are also characteristic of the Pacific sector of the Southern Ocean, south of the Pacific Antarctic Rise and eastwards reaching the Drake Passage and the Antarctic Peninsula (Figure 1).

The most prominent surface circulation feature of the South Pacific is the subtropical gyre (Tomczak and Godfrey, 2003; Figure 2). Its eastern boundary current is the Peru–Chile Current (PCC) system that largely exchanges heat and nutrients between high and low latitudes in the eastern Pacific. The main surface current originates from the northern margin of the Antarctic Circumpolar Current (ACC) that splits at around $\sim 43^\circ\text{S}$ off the coast of western South America into the PCC

flowing northward and the Cape Horn Current turning toward the south (Figure 3; Strub et al., 1998). Northward, the PCC is divided into coastal and oceanic branches through the poleward-flowing Peru–Chile countercurrent (PCCC). The PCC is the most extensive Eastern Boundary Current system in the world, driven by south-easterly winds along the Pacific coast of South America (Strub et al., 1998). The resulting offshore Ekman flow drives perennial upwelling of cool, nutrient-rich waters that support one of the most biologically productive regions in the oceans. At about 5°S , the PCC is deflected offshore, feeding the South Equatorial Current and flowing westward as the equatorial cold tongue between 10°S and 4°N . The western boundary currents of the South Pacific are less well developed (Tomczak and Godfrey, 2003) and consist of the East Australia Current between 18°S and Tasmania and another southward current along the east coast of New Zealand (the East Cape Current; Figure 2) that interrupts the more usual western boundary current formation seen in other oceans. The ACC system south of the subtropical gyre is one of the most important oceanic current systems, as it is the principle route for interocean transport. The ACC separates subtropical and polar water masses through three principal oceanic fronts: the subtropical front, the subantarctic front, and the Antarctic polar front where water mass characteristics change rather abruptly (Tomczak and Godfrey, 2003).

Wind Pattern

Corresponding to the general surface ocean circulation, the large-scale atmospheric circulation of the South Pacific region is comparatively simple (Tomczak and Godfrey, 2003). The subtropics are largely controlled by the semipermanent anticyclone over the southeast Pacific, resulting in moderately strong but very persistent southeast trade winds. South of the high pressure cell, the mid- to high-latitude South Pacific is dominated by extremely intense and prevalent westerly air flow. These strong so-called Southern Westerlies are annually centered around $50\text{--}55^\circ\text{S}$, corresponding to the steepest SST gradients within the ACC. Seasonally, the westerlies are latitudinally more confined and stronger in austral summer. During austral winter, zonal wind strengths are reduced in the center of the westerlies, but the whole wind belt extends further northward where zonal winds are thus enhanced. The westerlies result from the strong thermal gradient and atmospheric pressure difference between cold air masses over Antarctica and the warmer air and water masses in the subtropical regions (Cerveny, 1998).

El Niño Southern Oscillation

The South Pacific is the classical region for studying the impacts and the atmospheric and oceanographic processes of the

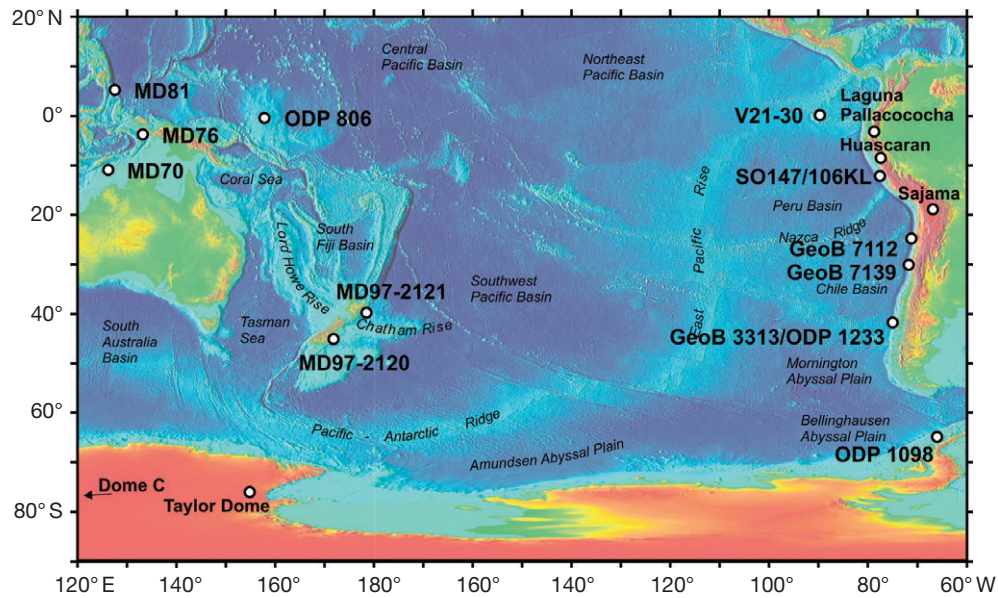


Figure 1 Map of the South Pacific showing major bathymetric features and the location of marine sediment cores and other paleoclimate archives discussed in the text.

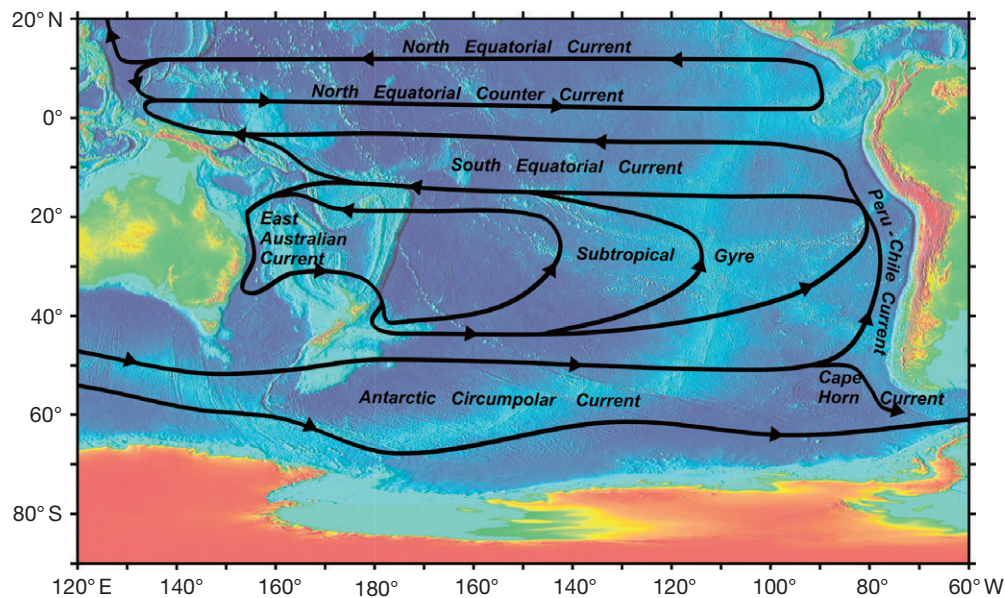


Figure 2 Map of the South Pacific showing major surface currents (after Tomczak and Godfrey, 2003).

El Niño Southern Oscillation (ENSO) as a major source for global climate variability on interannual timescales. During the positive phase of ENSO (El Niño events) that occurs roughly every 3–7 years, SSTs substantially increase in the eastern equatorial Pacific and the northern part of the PCC system, with important effects on continental rainfall in the subtropics and marine productivity of the Peruvian upwelling area (Tomczak and Godfrey, 2003). At the same time, the southeast Pacific anticyclone weakens and the storm tracks of the Southern Westerlies expand toward the north, resulting in above-average rainfall in mid-latitude Chile (Cerveny, 1998),

showing that the effects of ENSO are not restricted to the tropics/subtropics.

Continental Margin Records

South American Margin off Central and Southern Chile

Compared to other regions of the SH, marine sediments along the continental margin off Chile have been comparatively thoroughly studied after a number of international research cruises during the past 15 years. However, the first paleoenvironmental

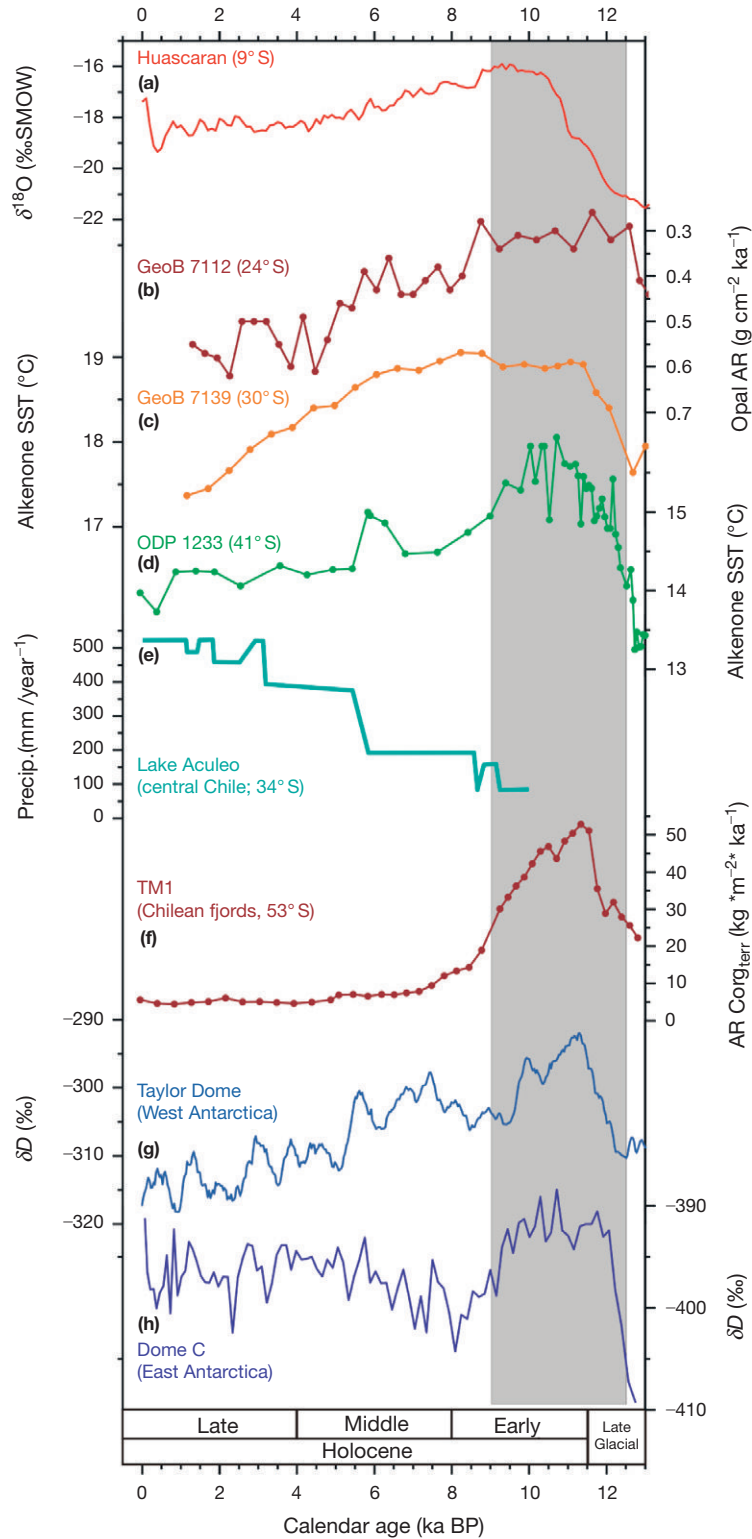


Figure 3 Sea surface temperature (SST) records off Chile compared to Antarctic and South American ice core data. Gray bar marks early Holocene maximum warm phase as defined at ODP Site 1233 (Kaiser et al., 2005). (a) Huascarán $\delta^{18}\text{O}$ record as a proxy for temperature changes in tropical South America (Thompson et al., 1995). (b) Opal accumulation rate (AR) record from core GeoB 7112-2 (Mohtadi et al., 2004) as a proxy for marine productivity. Higher values imply enhanced productivity and cooler SSTs. (c) SST record from core GeoB 7139-2 off north-central Chile (Kaiser et al., 2008). (d) SST record from ODP Site 1233 (Kaiser et al., 2005; Lamy et al., 2007). (e) Precipitation reconstruction for central Chile based on hydrological modeling of lake-level data (Jenny et al., 2003). (f) Terrestrial organic carbon AR record of core TM1 as a proxy for precipitation in southernmost Chile (Lamy et al., 2010). (g) δD record from the Taylor Dome ice core as a proxy for western Antarctic temperature (Steig et al., 1998). (h) δD record from the Dome C ice core as a proxy for eastern Antarctic temperature (EPICA Community Members, 2004).

records from the remote southern Chilean Fjord Region have only recently become available (Kilian et al., 2007; Lamy et al., 2010; Sepulveda et al., 2009). Pronounced gradients in oceanographic and climatic features make the Chilean margin particularly suitable for the study of postglacial paleoenvironmental variability at the northern margin of the ACC and the Southern Westerlies.

In terms of postglacial 'long-term' trends, SST records from the Chilean margin suggest an early Holocene maximum warming most clearly depicted at Ocean Drilling Program (ODP) Site 1233 (41° S), where SSTs peak between ~12.5 and 9 ka BP (Kaiser et al., 2005; Lamy et al., 2007; Figure 3(d)). This early Holocene optimum is likewise visible in a low-resolution alkenone SST record off southernmost Chile (MD07-3128; ~53° S; Caniupan et al., 2011) and coincides with warm conditions recorded both in the Eastern Antarctic Dome C (EPICA Community Members, 2004), and the Western Antarctic Taylor Dome ice cores (Steig et al., 1998), where the maximum warming appears to be restricted to a shorter interval, however (Figure 3(g) and 3(h)). An SST record off northern mid-latitude Chile (30° S; Figure 3(c); Kaiser et al., 2008) displays a slightly delayed optimum warming that lasted longer, that is, into the middle Holocene, a pattern that likewise appears in the tropical alpine South American ice core record recovered at the Nevado de Huascarán (Figure 3(a); Thompson et al., 1995). Except for the Dome C record, both Antarctic ice core and the paleoceanographic records display a cooling trend toward the late Holocene (Figure 3). The general temperature pattern during the Holocene is paralleled by opposing trends in continental rainfall between central and southern Chile (Lamy et al., 2010). In central Chile, located at the northern margin of the Southern Westerlies, regional terrestrial records and the terrigenous components in marine records generally suggest drier conditions during the early and middle Holocene and increased precipitation during the late Holocene (Figure 3(e); Jenny et al., 2003; Lamy et al., 1999; Latorre et al., 2007). In southern Chile, currently located within the core of the westerlies, new sedimentological and palynological records from fjords, peat bogs, and lakes point to more humid conditions during the early Holocene and decreasing rainfall thereafter (Figure 3(f)). These opposing rainfall trends across the northern margin and core of the westerlies point to antiphased intensity changes of the wind belt that resemble changes over the present seasonal cycle and do not necessarily require a latitudinal shift of the wind belt as a whole (Lamy et al., 2010).

The early Holocene optimum was not documented in the previously published SST record based on a shorter core (GeoB 3313-1) drilled at the same location as Site 1233. The shorter core only covers the last ~8 ka (Lamy et al., 2001, 2002). However, this shorter core provided very high-resolution records of SSTs, paleosalinity, paleoproductivity, and continental rainfall changes during the middle and late Holocene (Figures 4 and 5). The long-term trends in the records suggest a close coupling of marine and continental changes. Iron content decreases from the middle to the late Holocene, implying more humid conditions (Figure 4(a)). Similarly, SSTs as well as paleosalinities decrease and suggest increasing advection of subpolar water masses through the ACC and enhanced freshwater input in the Chilean Fjord region (Figure 4(b) and 4(c)). Finally, paleoproductivity generally increases throughout the record, consistent with

enhanced advection of high-nutrient/low-chlorophyll by the ACC combined with higher input of terrestrial micronutrients (Figure 4(d) and 4(e)). Similar continental rainfall and marine productivity changes have been found further north at 33° S (Hebbeln et al., 2002; Lamy et al., 1999).

The pattern of shorter term variability during the Holocene is more complex and many of the multicentennial- to millennial-scale variations in paleotemperatures and paleosalinities at site GeoB 3313 do not exactly match the reconstructed shifts of the westerlies in particular after ~5 ka BP. However, they do follow the millennial- to centennial-scale temperature changes recorded in Antarctica (Figure 5(a) and 5(b); Lamy et al., 2002). It has been speculated that this shift may be related to the onset of the modern state of ENSO (see later). The paleoproductivity records (Figure 5(d) and 5(e)) exhibit two century-scale excursions to higher values between 5.5 and 5 ka BP and shortly after 4 ka BP which do not exactly coincide with SST minima. Instead, they occur during cooling phases shortly after the two most significant increases in rainfall in the continental hinterland (Figure 5(c)). The iron record (Figure 4(a)) indicates that these rainfall augmentations terminated the less humid phase of the middle Holocene in southern Chile, pointing to a significant contribution of continent-derived micronutrients which resulted in an increase of marine productivity off southern Chile for several centuries (Lamy et al., 2002).

Studies on even shorter, that is, decadal- to centennial-scale timescales, are still rare. One such study has been performed on the latest Holocene sequence at ODP Site 1233 (Euler and Ninnemann, 2010). Planktonic and benthic foraminiferal $\delta^{18}\text{O}$ data suggest large decadal- and centennial-scale variations in surface and intermediate water physical properties that have been associated with shifts in tropical temperatures. Partly abrupt centennial-scale shifts in planktonic foraminiferal $\delta^{18}\text{O}$ data have been also observed further south at site MD07-3088 (~46° S) and interpreted in terms of significant SST and/or salinity variability (Siani et al., 2010). Within the fjords, there are indications of centennial-scale coolings during the past ~2000 years that partly coincide with the Northern Hemisphere Little Ice Age (Sepulveda et al., 2009).

South American Margin off Northern Chile and Peru

North of mid-latitude Chile, there are very few sedimentary records along the South American margin that cover the Holocene with a reasonable resolution because terrestrial sediment input is minimal off the hyperarid Atacama Desert. One of the few records comes from core GeoB 7112, located on the continental slope at ~24° S (Mohtadi et al., 2004). Though there is no direct SST proxy record available, several proxy records for marine productivity changes suggest that upwelling intensity was substantially reduced and the advection of warmer subtropical surface waters through a stronger PCCC was enhanced during the early and middle Holocene (Figure 3(b)). After ~5 ka BP, the pattern reversed, that is, upwelling became slightly stronger and surface waters were colder, consistent with the SST record at ~30° S off mid-latitude Chile (Figure 3(b) and 3(c)). As observed in most of the Chilean records further south, the variability in the records appears to be higher during the late Holocene, a pattern that may relate to more frequent and stronger El Niño events.

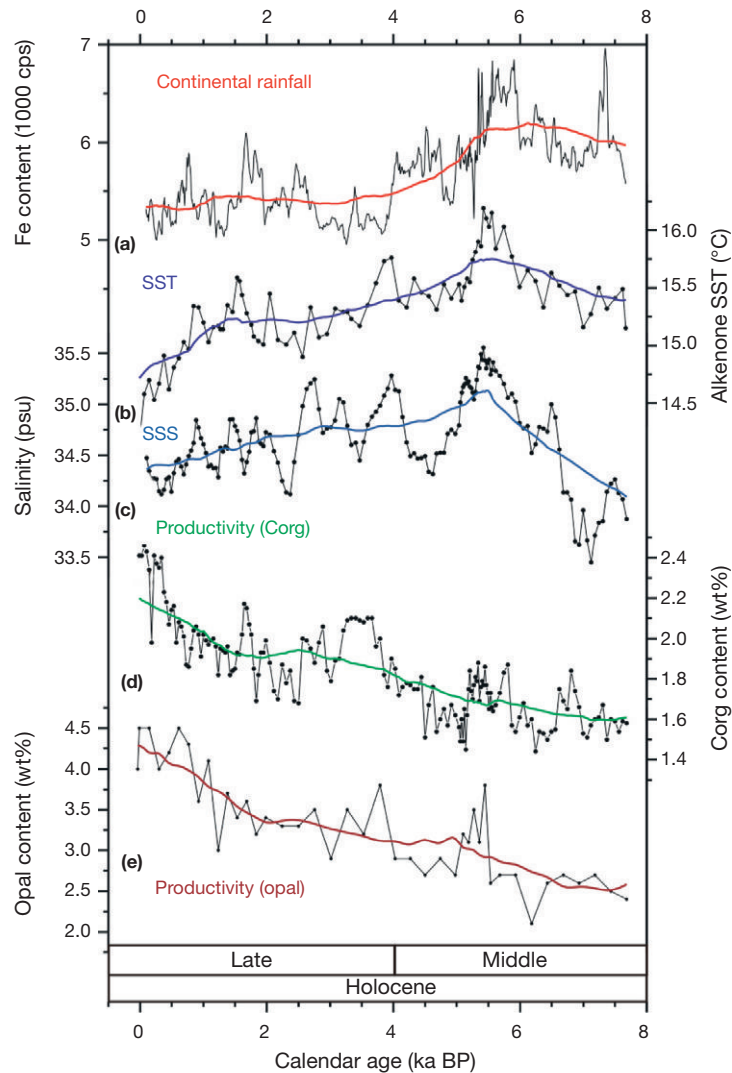


Figure 4 Paleoclimatic and paleoceanographic records from core GeoB 3313-1 off southern Chile covering the middle and late Holocene. Bold lines are ~2000-year moving averages (data from Lamy et al., 2002). (a) Iron contents interpreted as a proxy for rainfall and the position of the Southern Westerlies (high iron contents imply increased contribution of iron-rich volcanic Andean source rocks and decreased supply of iron-poor Coastal Range source rocks indicating decreased rainfall (Lamy et al., 2001)). (b) SST record tracing the advection of subpolar water by the ACC. (c) Sea surface salinity (SSS) reconstruction (5-point moving average) as a proxy for freshwater input in the Chilean Fjord region. (d) Corg contents and (e) biogenic opal contents are taken as proxies for paleoproductivity.

ENSO is also a major issue in the discussion of records from the Peruvian margin and adjacent tropical South America. An alkenone-based SST record from marine sediment core SO147-106KL (Figure 6(f); Rein et al., 2005), located on the continental shelf off Lima/Peru, shows a comparatively warm early Holocene, a cooling during the middle Holocene, and a significant late Holocene warming, culminating around 1.5 ka BP. The cooler SSTs would imply reduced El Niño events and mean La Niña-like conditions during the middle Holocene, consistent with SST records from the Eastern Pacific Cold Tongue (Figure 6(e); Koutavas et al., 2002; see section ‘Open Ocean Records’). These data are corroborated by terrigenous sediment flux rates, likewise based on core SO147-106KL, that suggest less El Niño-related rainfall events between 8 and 5.6 ka BP (Figure 6(g); Rein et al., 2005). In addition, lake records from

Equador (Moy et al., 2002) suggest reduced El Niño activity during the early and middle Holocene and significant enhancement of El Niño thereafter (particularly after ~5 ka BP; Figure 6(h)). In contrast to this broadly consistent picture, the presence of warm water mollusc shells that are presently not found along the coast of Peru, and $\delta^{18}\text{O}$ records from otoliths of sea catfish, imply warmer-than-present SSTs during the middle Holocene (Andrus et al., 2002; Figure 6(d)). Such warmer SSTs have been interpreted as reduced upwelling off Peru and thus a mean El Niño-like state (Andrus et al., 2002).

New Zealand/Australia

New alkenone SST records have recently become available from the southwest Pacific off New Zealand. A high-resolution

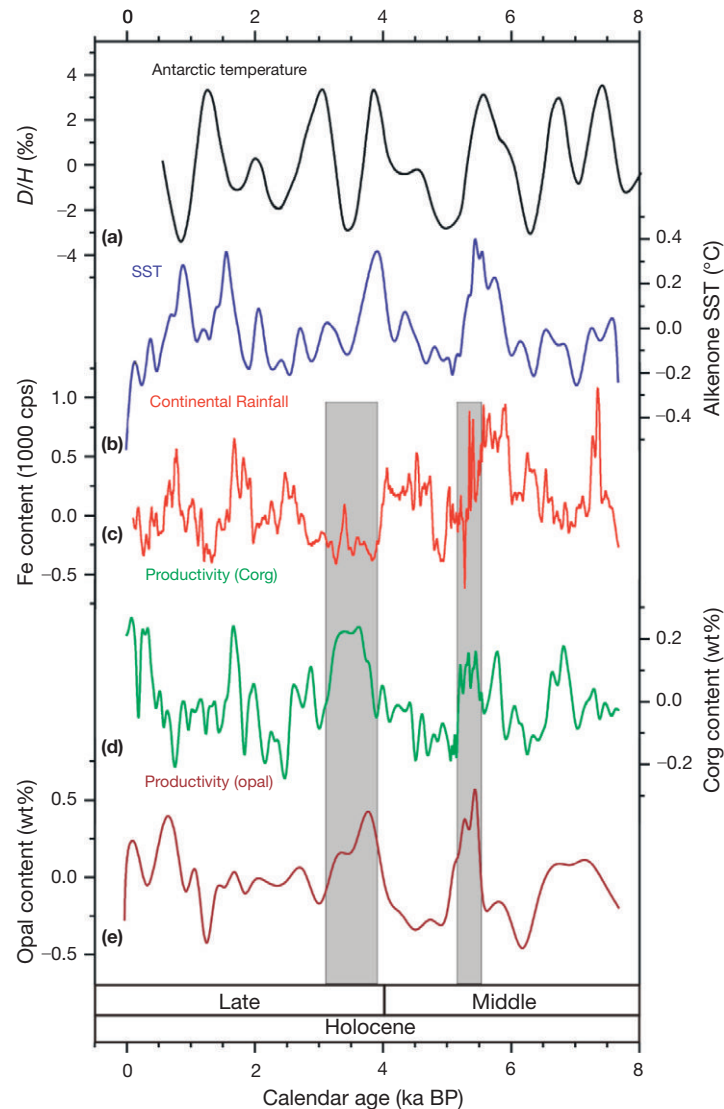


Figure 5 Paleoclimatic and paleoceanographic records from core GeoB 3313-1 off southern Chile focusing on the millennial- to centennial-scale variations compared to Antarctic ice core records. Data sets have been detrended by subtracting the long-term trends shown in Figure 4 (data from Lamy et al., 2002). (a) Short-term variability in Antarctic temperature shown as the first component of an empirical orthogonal function (EOF) analysis calculated from isotopic records from the Ross sea sector (Byrd and Taylor Dome sites; Masson et al., 2000). (b) SST record off Chile showing a significant positive correlation ($r=0.57$, $n=57$; significant on the 99.9% level after applying a 100-year time-resolution interpolation) to Antarctic temperatures until ca. 2000 BP (c) Iron contents of core GeoB 3313-1 indicating rainfall and the position of the Southern Westerlies. Paleoproductivity proxies (d) Corg content and (e) biogenic opal content. Vertical gray bars mark time intervals of century-scale increases in paleoproductivity as discussed in the text.

alkenone SST record based on core MD97-2120 ($\sim 46^\circ\text{S}$), located south of the Subtropical Convergence (Pahnke and Sachs, 2006) east of New Zealand, reveals a pronounced early Holocene temperature maximum lasting from ~ 12 to 8.5 ka BP with SSTs about $2\text{--}3^\circ\text{C}$ warmer than in the middle and late Holocene (Figure 7(b)). At the western continental margin of New Zealand, the alkenone SST record of core SO136-011GC ($\sim 44^\circ\text{S}$; Barrows et al., 2007) shows a somewhat less pronounced early Holocene temperature maximum and decreasing temperatures already beginning by ~ 9.5 ka BP and an additional cooling step in the late Holocene at ~ 2 ka BP (Figure 7(d)). A third alkenone SST record

based on core MD07-2121 ($\sim 40^\circ\text{S}$; Pahnke and Sachs, 2006), located east of New Zealand, north of the Subtropical Convergence, reveals maximum Holocene temperatures from ~ 9 to 6.5 ka BP which are, however, only $\sim 1\text{--}1.5^\circ\text{C}$ above late Holocene values (Figure 7(c)). Taken together, the southwest Pacific records reveal a Holocene pattern that is broadly similar to the southeast Pacific SST records (Figure 7(a)) and Antarctic ice core data sets (Figure 8(e) and 8(f)). Thus, the South Pacific mid-latitudes appear to be characterized by in-phase SST changes, at least regarding the general trends, similar to the pattern derived for the last glacial period.

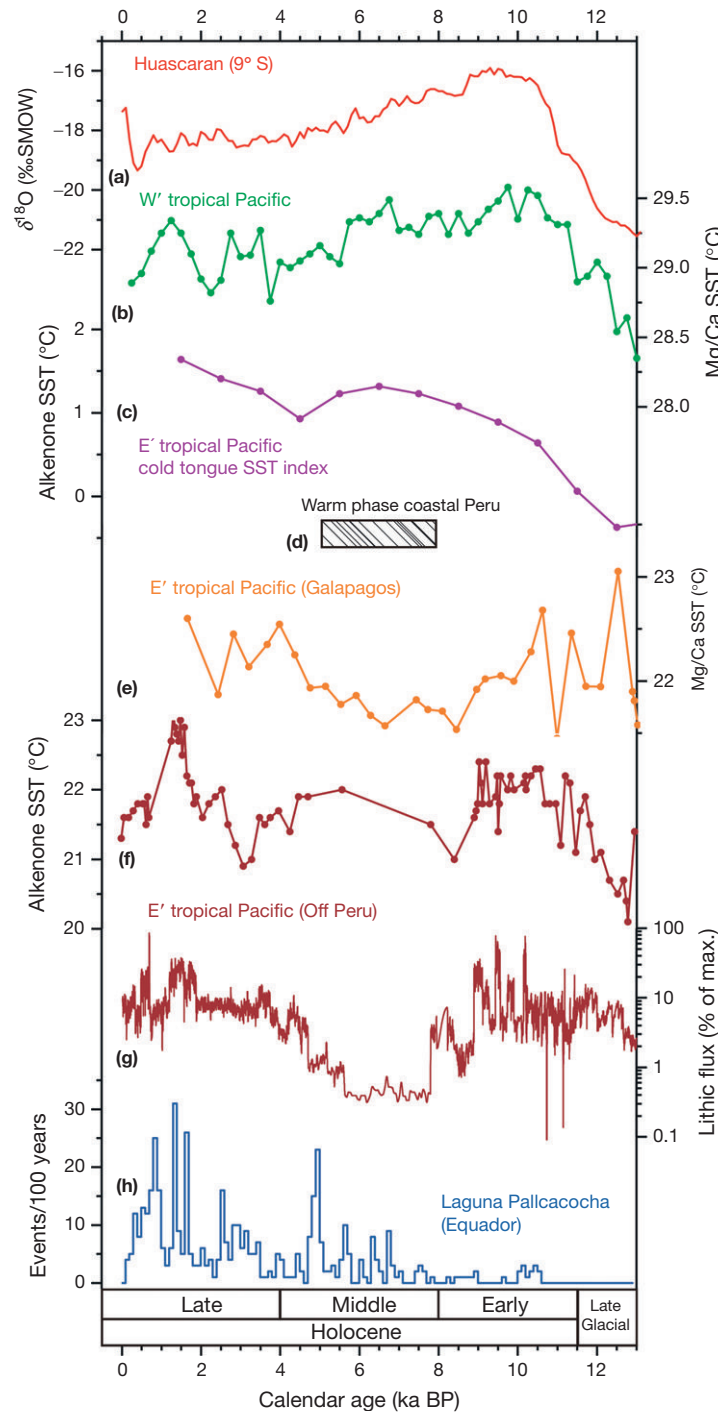


Figure 6 Paleoclimatic data from the subtropical and tropical Pacific and other terrestrial records relevant for the discussion of long-term ENSO changes. (a) Huascarán $\delta^{18}\text{O}$ record as a proxy for temperature changes in tropical South America (Thompson et al., 1995). (b) SST record from the western Pacific Warm Pool (stacked data from several cores as in Stott et al. (2004)). (c) Composite alkenone SST index based on five SST records from the eastern Pacific equatorial cold tongue that have been averaged for nonoverlapping 1000-year time windows (Koutavas and Sachs, 2008). (d) Warm phase recorded in coastal Peru (Andrus et al., 2002). (e) Mg/Ca-based SST record from the eastern Pacific cold tongue (Koutavas et al., 2002). (f) Alkenone-derived SST record from the Peruvian shelf (Rein et al., 2005). (g) Lithic flux record from the Peruvian shelf as a proxy for El Niño-related torrential rainfall events in the region (Rein et al., 2005). (h) Frequency of El Niño events reconstructed from turbidity frequency recorded in a lake in Ecuador (Moy et al., 2002).

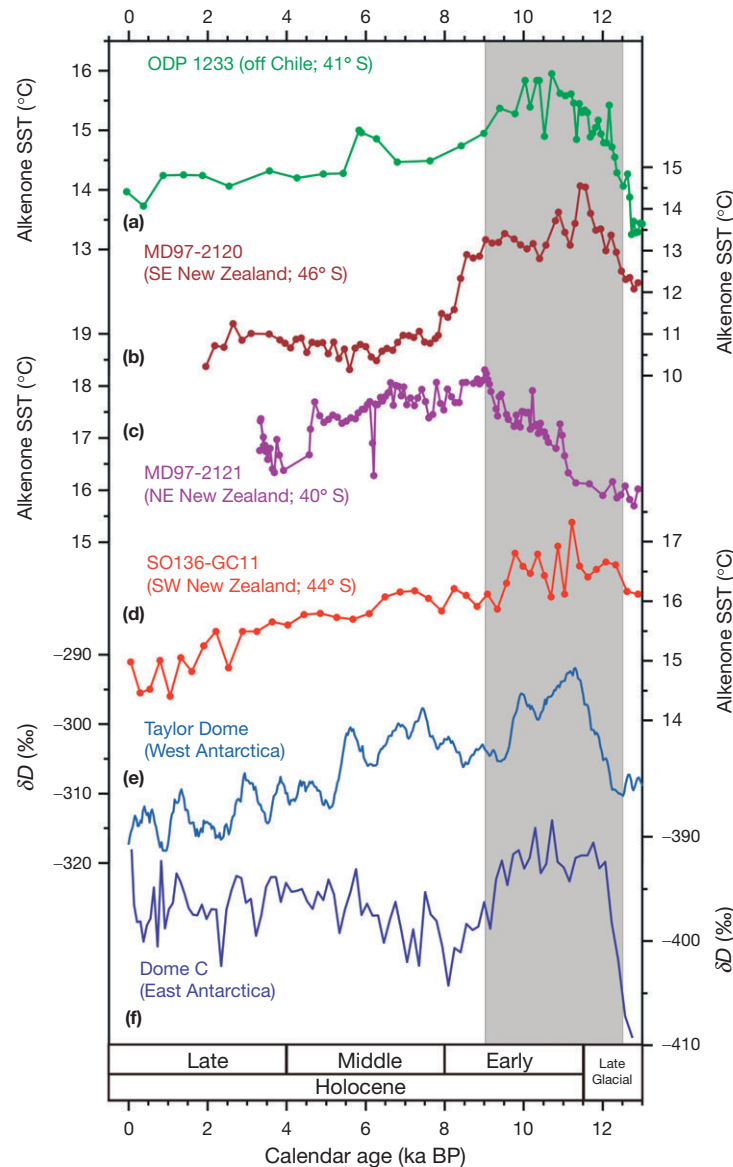


Figure 7 Alkenone-based SST records from the New Zealand continental margins compared to the SST record from ODP Site 1233 off southern Chile and Antarctic ice core records. Gray bar marks early Holocene maximum warm phase as defined at ODP Site 1233 (Kaiser et al., 2005). (a) SST record from ODP Site 1233 (Kaiser et al., 2005; Lamy et al., 2007). (b) SST record from core MD91-2120 (Pahnke and Sachs, 2006). (c) SST record from core MD91-2121 (Pahnke and Sachs, 2006). (d) SST record from core SO136-GC11 (Barrows et al., 2007). (e) δD record from the Taylor Dome ice core as a proxy for western Antarctic temperature (Steig et al., 1998). (f) δD record from the Dome C ice core as a proxy for eastern Antarctic temperature (EPICA Community Members, 2004).

Additionally, core MD07-2121 provided more indirect information on postglacial circulation changes off eastern New Zealand through changes in terrigenous sediment flux (Gomez et al., 2004). This record suggests a strengthened East Cape current from ~ 9 to 3.5 ka BP and reduced inflow of subantarctic water masses. A major sediment texture shift in this core at ~ 4 ka BP has been correlated to a similar shift in terrestrial floodplain and marine shelf records. This shift has been correlated with the onset of increased storminess connected to the establishment of modern intensified atmospheric circulation, strongly influenced by ENSO. Enhanced ENSO

activity during the late Holocene is consistent with the records from the South American margin of the Pacific.

Antarctic Peninsula (Palmer Deep)

Along the Pacific margin of the Antarctic Peninsula, high-resolution and well-dated sedimentary records have been recovered from the Palmer Deep, an inner shelf basin (Figure 1). The records from ODP Site 1098 cover the last ~ 13 ka, often in centennial or better resolution, and can be subdivided into four different intervals (Figure 8(b); Domack et al., 2001).

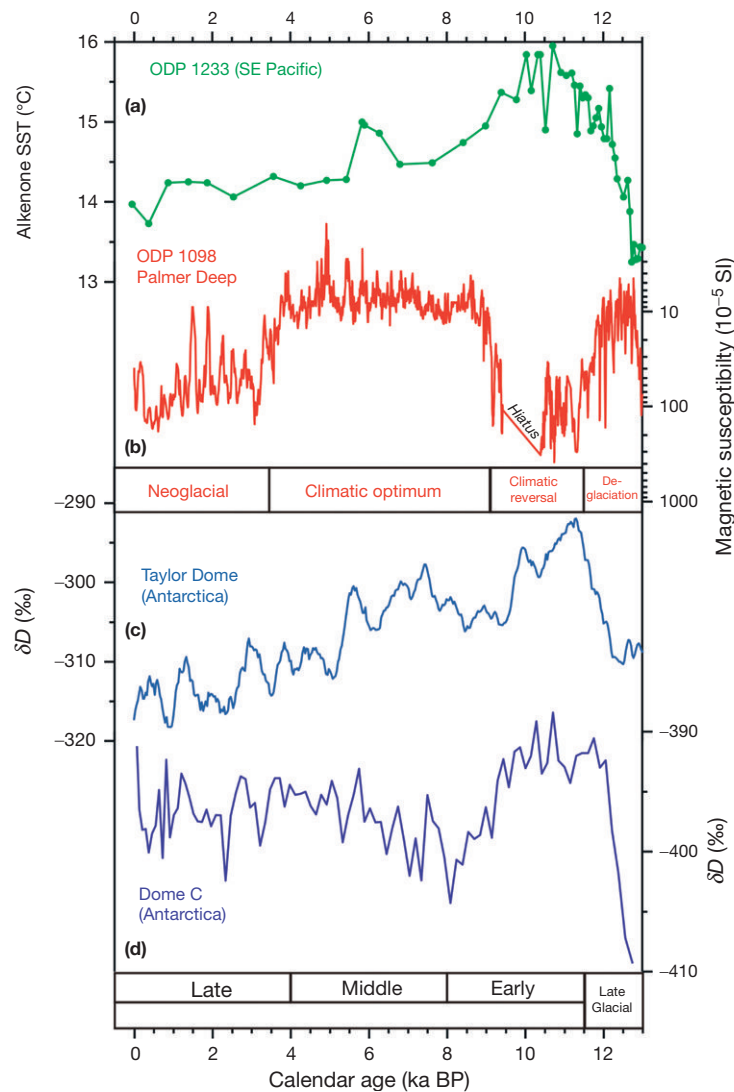


Figure 8 Paleoenvironmental record from Palmer Deep (Antarctic Peninsula) compared to the SST record from ODP Site 1233 off southern Chile and Antarctic ice core records. (a) SST record from ODP Site 1233 (Kaiser et al., 2005; Lamy et al., 2007). (b) Magnetic susceptibility record from ODP Site 1098 at Palmer Deep (Domack et al., 2001). High values are interpreted to reflect stronger terrigenous sediment input and reduced marine productivity coinciding with enhanced input of ice-rafted detritus (IRD) (Domack et al., 2001). Shown are also the major Holocene intervals that have been defined for Palmer Deep (reproduced from Domack E, Leventer A, Dunbar R, Taylor F, Brachfeld S, and Sjunneskog C (2001) Chronology of the Palmer Deep site, Antarctic Peninsula: A Holocene paleoenvironmental reference for the circum-Antarctic. *Holocene* 11: 1–9). (c) δD record from the Taylor Dome ice core as a proxy for western Antarctic temperature (Steig et al., 1998). (d) δD record from the Dome C ice core as a proxy for eastern Antarctic temperature (EPICA Community Members, 2004).

Deglaciation lasted until ~ 11.5 ka BP and was followed by a climatic reversal ending abruptly at ~ 9 ka BP. Thereafter, a middle Holocene ‘climatic optimum’ is recorded, characterized by enhanced marine productivity and decreased input of ice-rafted detritus (IRD). After ~ 4.7 ka BP, productivity decreased and IRD input increased, leading to a ‘neoglacial’ phase since ~ 3.4 ka BP. This is consistent with enhanced IRD content in South Atlantic sediment cores. Coeval changes were recorded in proxy records for deepwater conditions at the ~ 1000 m deep drilling site. During the middle Holocene ‘optimum,’ bottom waters were poorly ventilated, suggesting a reduction of circumpolar deepwater production. This, in turn, is possibly related to reduced North Atlantic Deepwater production (see (Thermohaline Circulation), providing a link to Northern

Hemisphere climate (Ishman and Sperling, 2002). Bottom water conditions changed rapidly during the late Holocene with variable inflow of circumpolar deepwater, possibly related to short-term changes in the position of the Southern Ocean sea-ice margin and the position of the ACC providing a link to the Chile margin records.

Bentley et al. (2009) recently reviewed the Holocene paleoenvironmental record of the Antarctic Peninsula region and their study revealed significant differences between terrestrial and marine records. This is particularly manifested for a pronounced warming period from ~ 4.5 to 2.8 ka BP seen in the terrestrial records that overlaps with the beginning of the ‘neoglacial’ phase in the marine records of Palmer Deep.

Open Ocean Records

Away from the continental margins, high-resolution Holocene records are rare in the mid- and high-latitude Pacific, where even the glacial–interglacial patterns are not well documented. A radiolarian-based SST record from the southwest Pacific recovered from the northeastern flanks of the Bounty Plateau (~900 km east of the New Zealand's South Island) reveals a 2–3 °C warmer early Holocene warm period between ~10 and 5 ka BP (Luer et al., 2009), roughly similar in timing and amplitude to the New Zealand and southeast Pacific margin records. In the Pacific Sector of the Southern Ocean, Bradtmiller et al. (2009) compared glacial and Holocene opal flux rates based on 31 cores and found that Holocene opal fluxes were lower north of the Polar Front and higher south of it. However, the resolution of their records does not allow for discussing Holocene temporal trends in these data sets.

In the tropics, Holocene paleoceanographic records have been derived from sediment records of the Western Pacific Warm Pool and the Eastern Pacific Cold Tongue (Figure 1; Koutavas et al., 2002; Stott et al., 2004). Today, both these are regions that are particularly sensitive to SST changes connected to ENSO. The western Pacific records display a slight (~0.5 °C) cooling over the past 10 ka, while the Mg/Ca-based SST record from the Cold Tongue core V21-30 reveals a broad middle Holocene cooling on the order of 1 °C (Figure 6(b) and 6(e)). These data suggest enhanced zonal SST gradients during the middle Holocene that decrease after ~5 ka, a pattern that would suggest a mean La Niña-like state of ENSO during the middle Holocene and a larger impact of El Niño events during the later part, broadly consistent with the records based on core SO147-106KL on the Peruvian shelf (Figure 6(f) and 6(g)). The above-mentioned SST records are both Mg/Ca-based. More recently, several alkenone-derived SST records from the Cold Tongue have been published that are all characterized by a consistent warming in the range of ~0.5–2 °C over the course of the Holocene (Figure 6(c); Koutavas and Sachs, 2008; Leduc et al., 2010). This warming trend has been explained by changes in low-latitude boreal winter insolation, assuming that the alkenone-producing haptophytes preferentially bloom during this season (Leduc et al., 2010). Moreover, Leduc et al. (2010) showed that the Mg/Ca-derived SST records from the eastern tropical Pacific are much more heterogeneous, due to complex combinations of different ecological preferences of planktonic Foraminifera and the impact of carbonate dissolution. Therefore, Mg/Ca SST records are more complicated to interpret, in terms of the overall temperature trends. However, the general pattern of enhanced zonal SST gradients during the early to middle Holocene, compared to the late Holocene, remains valid, based on comparisons with the alkenone-based SST reconstructions (Figure 6(b) and 6(c)).

Forcing Mechanisms

Orbital Timescales

The major 'long-term' trends within the postglacial records from the South Pacific are most likely related to orbital forcing. During the early Holocene, summer insolation was higher in

the Northern Hemisphere and reduced in the south, resulting in distinct seasonality differences between the two hemispheres. In the course of the Holocene, austral summer (December–February) insolation and seasonality increase toward the present. These changes are induced by precessional forcing, whereas changes in the obliquity of the Earth's axis produce hemispherically symmetric patterns in annual insolation (Figure 9). Throughout the Holocene, annual insolation decreased in high latitudes and increased in the tropics.

Most extratropical SST records from the South Pacific do indeed show a trend toward colder temperatures from the early to late Holocene, a pattern that is also found in some (but not all) Antarctic ice core records. In addition, atmospheric and oceanographic changes appear to be closely coupled, suggesting, for example, simultaneous latitudinal shifts of the northern margins of the Southern Westerlies and the ACC system on orbital timescales (see earlier). A recent model of global surface ocean changes during the Holocene shows early Holocene SST about 0.5 °C warmer in the Southern Ocean compared to ~1 °C warming in the Arctic Ocean (Liu et al., 2003). However, based on the paleorecords, the observed cooling in the course of the Holocene was substantially larger, for example, on the order of 1.5–2 °C in the Chilean margin records (Figure 3). Additional SH warming during the early Holocene might be attributable to changes in the global ocean circulation which involve a bipolar seesaw-like surface temperature pattern. For the early Holocene, such a mechanism has been invoked to explain warm temperatures in the Southern Ocean and Antarctica (Masson et al., 2000) and is consistent with a reconstruction of the Atlantic meridional overturning circulation strength in the North Atlantic (McManus et al., 2004; Figure 9(f)) (see [Thermohaline Circulation](#)).

Precessional changes in seasonal insolation might have affected the records in the subtropics and tropics. Reduced early and middle Holocene summer insolation in the SH could have, for example, strengthened the Hadley Cell circulation in the eastern Pacific as the South American summer monsoon was largely reduced (Figure 9). This would indirectly affect the position of the Southern Westerlies and the position of the northern margin of the ACC, contributing to the observed early Holocene warming in the southeast Pacific and dry conditions in central Chile. At the same time, however, the core of the Southern Westerlies over southernmost Chile was enhanced (see earlier). During the late Holocene, warming in the eastern tropical Pacific and cooling further south possibly enhanced the mid- to low-latitude SST gradient resulting in stronger westerlies at their northern margin (Lamy et al., 2010). In addition, seasonal insolation changes largely affect ENSO dynamics and the position of the Intertropical Convergence Zone (ITCZ). Modeling studies (Cane, 2005) suggest that the ITCZ was located north of its present position and the frequency and strength of El Niño events were strongly reduced during the early and middle Holocene, consistent with reconstructions based on paleodata from the eastern tropical Pacific and adjacent subtropical South America (Figure 6).

Millennial–Centennial Timescales

Millennial- to centennial-scale climate patterns are only documented in a few records in the South Pacific realm.

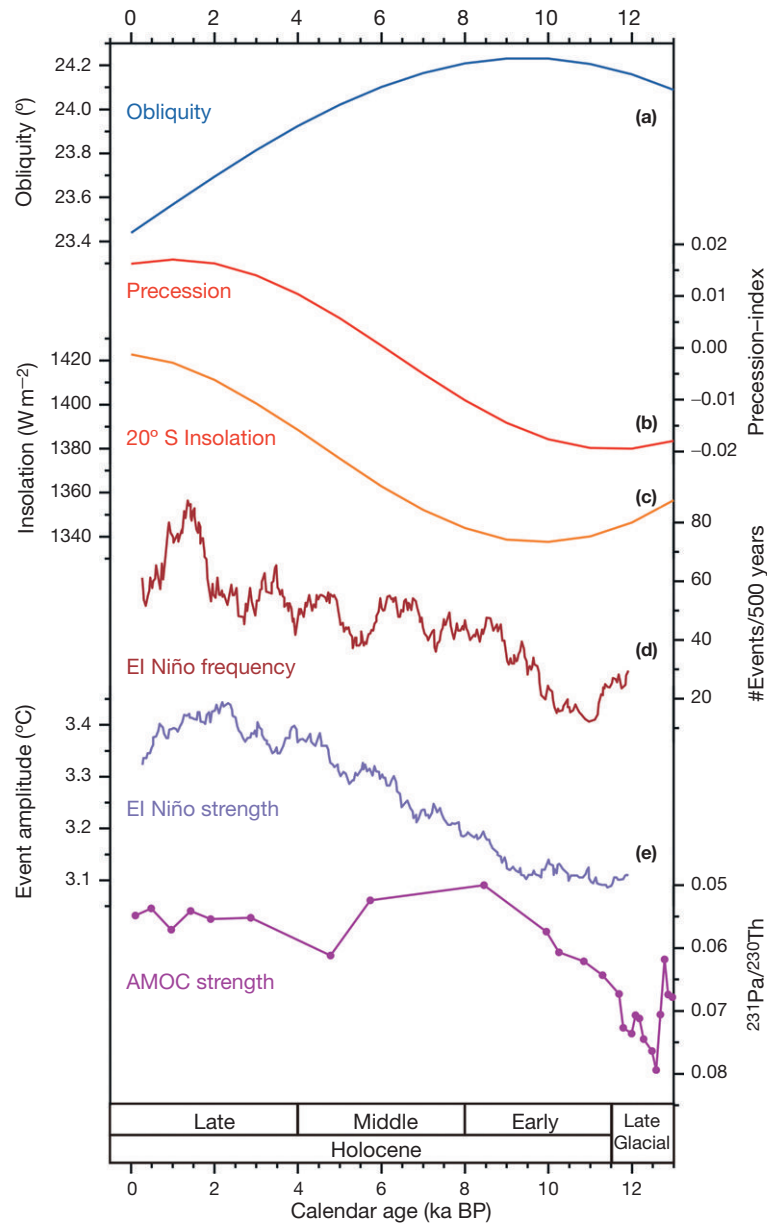


Figure 9 Orbital forcing parameters and modeled ENSO variability during the Holocene. (a) Obliquity (data from Berger and Loutre, 1991). (b) Precession index (data from Berger and Loutre, 1991). (c) Austral summer insolation (December–February; reproduced from Berger and Loutre, 1991). (d) Number of El Niño events per 500 years (years with $>3^{\circ}\text{C}$ NINO3 SST anomaly; data from Clement et al., 2000). (e) El Niño strength shown as the mean amplitude of the $>3^{\circ}\text{C}$ NINO3 SST anomaly in a 500-year window (data from Clement et al., 2000). (f) $^{231}\text{Pa}/^{230}\text{Th}$ record from a subtropical North Atlantic sediment core taken as a proxy for the strength of the Atlantic meridional overturning circulation (AMOC; data from McManus et al., 2004).

The high-resolution record of rainfall variability from the southern Chilean margin (41°S) suggests two dominant bands of variability with cycles of ~ 900 and ~ 1500 years (Lamy et al., 2001). Comparisons of band pass filters with records from Antarctica and tropical South America (Sajama ice core) show a close connection to the tropics, particularly regarding the ~ 900 -year cycle (Figure 10) most likely involving changes in the strength of the Hadley Cell over the southeast Pacific and adjacent South America (Lamy et al., 2001). On the other hand, the correlation to climate conditions in Antarctica on millennial and multicentennial

timescales reveals a major phase shift at ~ 4 ka BP (Figure 10) that roughly coincides with the onset of modern ENSO. A shift at the transition from the middle to the late Holocene is also evident in the comparison of the short-term rainfall fluctuations to surface ocean changes. This suggests a mismatch between millennial- and multicentennial-scale variability in atmospheric (i.e., Southern Westerlies) and oceanographic (i.e., southern PCC and ACC) circulation patterns during the late Holocene (Figure 6). The proposed explanation is that temperatures in the southern PCC during the late Holocene have probably not been affected by changes in ENSO (Lamy

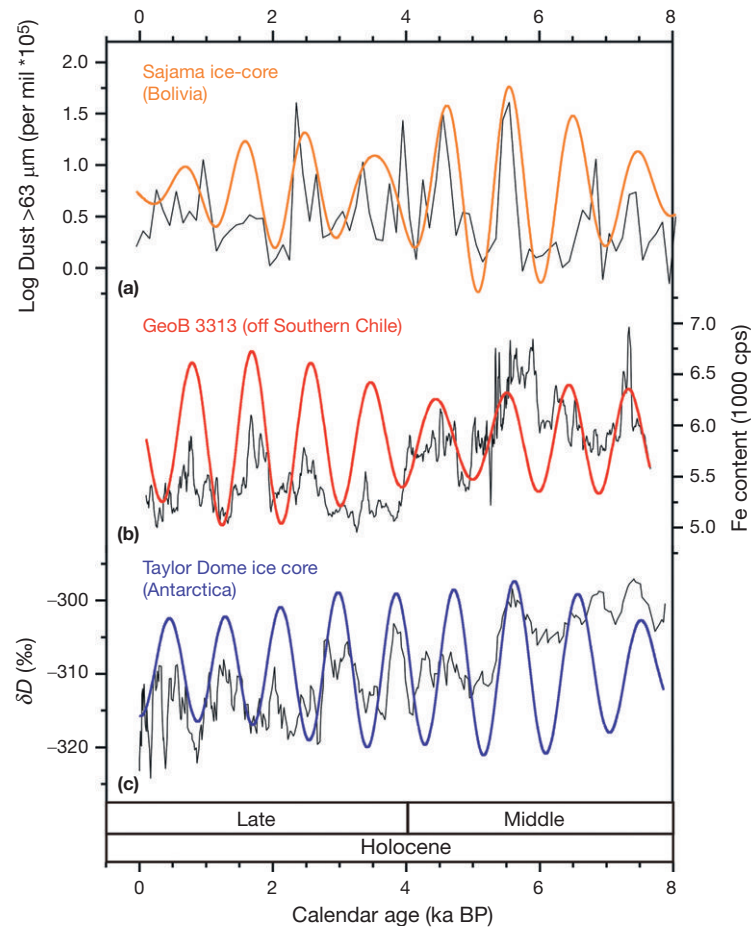


Figure 10 Comparisons of 900-year band pass filters of Chilean rainfall record with records from Antarctica and tropical South America. (a) Dust content of the Sajama ice core (thin line; data from Thompson et al., 1998) interpreted as a proxy for the aridity of the Bolivian Altiplano and the Hadley cell intensity over the southeast Pacific and tropical South America. (b) Iron content from core GeoB 3313-1 interpreted as a proxy for rainfall and the position of the Southern Westerlies (thin line = 5-point moving average; data from Lamy et al., 2001). (c) δD record of the Taylor Dome ice core (thin line = 5-point moving average) interpreted as a proxy for atmospheric temperature in coastal Antarctica and the extension of Antarctic sea ice (data from Steig et al., 1998). Thick lines represent Gaussian band pass filter centered at 900 years ($0.00111 \pm 0.0002 \text{ year}^{-1}$). The bandwidth chosen covers the dominant periodicities around 900 years in the records as indicated by spectral analyses (data from Lamy et al., 2001). High iron contents in the Chilean record correlate to high dust contents in the Sajama ice core in the 900-year band. Correlation of iron contents to δD record of the Taylor Dome ice core shifts from positive in the middle Holocene (high iron contents correlate to high δD values) to negative in the late Holocene (high iron contents correlate to low δD values).

et al., 2002). Modern ENSO-related SST anomalies only reach south to $33\text{--}36^\circ\text{S}$ (Strub et al., 1998), whereas meteorological studies suggest a strong effect of ENSO on rainfall anomalies in both central and southern Chile (Montecinos and Aceituno, 2003). The possible global origins of the ~ 900 and ~ 1500 cycles that are found in a number of Holocene records on a global scale are discussed by Lamy et al. (2001). The ~ 900 -year cycle could be related either to subtle changes in orbital parameters and/or to solar activity, as it is also present in the $\Delta^{14}\text{C}$ record from tree rings. Interestingly, long-term grain-size variability at Palmer Deep of the Antarctic Peninsula is characterized by ~ 1800 -year cycles that have been related to lunar tidal cycles (Warner and Domack, 2002) or they may simply represent multiples of the ~ 900 -year cycle. The high-resolution Palmer Deep records also suggest substantial variability of oceanic and atmospheric circulation close to Antarctica on shorter, centennial

timescales (e.g., ~ 200 and ~ 400 years, interpreted as solar variability) and even on multidecadal scales (Domack et al., 2001).

References

- Andrus CFT, Crowe DE, Sandweiss DH, Reitz EJ, and Romanek CS (2002) Otolith $\delta^{18}\text{O}$ record of mid-holocene sea surface temperatures in Peru. *Science* 295: 1508–1511.
- Barrows TT, Lehman SJ, Fifield LK, and De Deckker P (2007) Absence of cooling in New Zealand and the adjacent ocean during the Younger Dryas chronozone. *Science* 318: 86–89.
- Bentley MJ, Hodgson DA, Smith JA, et al. (2009) Mechanisms of Holocene palaeoenvironmental change in the Antarctic Peninsula region. *The Holocene* 19: 51–69.
- Berger A and Loutre MF (1991) Insolation values for the climate of the last 10 million years. *Quaternary Science Review* 10: 297–317.
- Bradt Miller LI, Anderson RF, Fleisher MQ, and Burckle LH (2009) Comparing glacial and Holocene opal fluxes in the Pacific sector of the Southern Ocean. *Paleoceanography* 24: PA2214.

- Cane MA (2005) The evolution of El Niño, past and future. *Earth and Planetary Science Letters* 230: 227–240.
- Caniupan AM, Lamy F, Lange CB, et al. (2011) Millennial-scale surface water changes in the Southeast Pacific off southernmost Chile 1 (53°S) over the past ~60 kyr. *Paleoceanography* 26: PA3221.
- Cerveny R (1998) Present climates of South America. In: Hobbs JE (ed.) *Climates of the Southern Continents: Present, Past and Future*, pp. 107–135. West Sussex: Wiley.
- Clement A, Saeger E, and Cane M (2000) Suppression of El Niño during the mid-Holocene by changes in the Earth's orbit. *Paleoceanography* 15: 731–737.
- Domack E, Leventer A, Dunbar R, Taylor F, Brachfeld S, and Sjunneskog C (2001) Chronology of the Palmer Deep site, Antarctic Peninsula: A Holocene palaeoenvironmental reference for the circum-Antarctic. *The Holocene* 11: 1–9.
- EPICA Community Members (2004) Eight glacial cycles from an Antarctic ice core. *Nature* 429: 623–628.
- Euler C and Ninnemann US (2010) Climate and Antarctic intermediate water coupling during the late Holocene. *Geology* 38: 647–650.
- Gomez B, Carter L, Trustrum NA, Palmer AS, and Roberts AP (2004) El Niño–Southern Oscillation signal associated with middle Holocene climate change in intercorrelated terrestrial and marine sediment cores, North Island, New Zealand. *Geology* 32: 653–656.
- Hebbeln D, Marchant M, and Wefer G (2002) Paleoproductivity in the southern Peru–Chile Current through the last 33,000 years. *Marine Geology* 186: 487–504.
- Ishman SE and Sperling MR (2002) Benthic foraminiferal record of Holocene deep-water evolution in the Palmer Deep, western Antarctic Peninsula. *Geology* 30: 435–438.
- Jenny B, Wilhelm D, and Valero-Garcés BL (2003) The Southern Westerlies in central Chile: Holocene precipitation estimates based on a water balance model for Laguna Aculeo (33°–50°S). *Climate Dynamics* 20: 269–280.
- Kaiser J, Lamy F, and Hebbeln D (2005) A 70-kyr sea surface temperature record off southern Chile (ODP Site 1233). *Paleoceanography* 20: PA4009. <http://dx.doi.org/10.1029/2005PA001146>.
- Kaiser J, Schefuss E, Lamy F, Mohtadi M, and Hebbeln D (2008) Glacial to Holocene changes in sea surface temperature and coastal vegetation in north central Chile: High versus low latitude forcing. *Quaternary Science Reviews* 27: 2064–2075.
- Kilian R, Baeza O, Steinke T, Arevalo M, Rios C, and Schneider C (2007) Late Pleistocene to Holocene marine transgression and thermohaline control on sediment transport in the western Magellanes fjord system of Chile (53°S). *Quaternary International* 161: 90–107.
- Koutavas A, Lynch-Stieglitz J, Marchitto TM Jr., and Sachs JP (2002) El Niño-like pattern in ice age tropical Pacific sea surface temperature. *Science* 297: 226–230.
- Koutavas A and Sachs JP (2008) Northern timing of deglaciation in the eastern equatorial Pacific from alkenone paleothermometry. *Paleoceanography* 23: 10 pp.
- Lamy F, Hebbeln D, Rohl U, and Wefer G (2001) Holocene rainfall variability in southern Chile: A marine record of latitudinal shifts of the Southern Westerlies. *Earth and Planetary Science Letters* 185: 369–382.
- Lamy F, Hebbeln D, and Wefer G (1999) High-resolution marine record of climatic change in mid-latitude Chile during the last 28,000 years based on terrigenous sediment parameters. *Quaternary Research* 51: 83–93.
- Lamy F, Kaiser J, Arz HW, et al. (2007) Modulation of the bipolar seesaw in the southeast Pacific during Termination 1. *Earth and Planetary Science Letters* 259: 400–413.
- Lamy F, Kilian R, Arz HW, et al. (2010) Holocene changes in the position and intensity of the southern westerly wind belt. *Nature Geoscience* 3: 695–699.
- Lamy F, Rühlemann C, Hebbeln D, and Wefer G (2002) High- and low-latitude climate control on the position of the southern Peru–Chile Current during the Holocene. *Paleoceanography* 17(1028) Article No. 1028.
- Latorre C, Moreno PI, Vargas G, et al. (2007) Late Quaternary environments and paleoclimate. In: Moreno T and Gibbons W (eds.) *The Geology of Chile*, pp. 309–328. London: Geological Society.
- Leduc G, Schneider R, Kim JH, and Lohmann G (2010) Holocene and Eemian sea surface temperature trends as revealed by alkenone and Mg/Ca paleothermometry. *Quaternary Science Reviews* 29: 989–1004.
- Liu Z, Brady E, and Lynch Stieglitz J (2003) Global ocean response to orbital forcing in the Holocene. *Paleoceanography* 18: 19–19–20.
- Luer V, Cortese G, Neil HL, Hollis CJ, and Willems H (2009) Radiolarian-based sea surface temperatures and paleoceanographic changes during the Late Pleistocene–Holocene in the subantarctic southwest Pacific. *Marine Micropaleontology* 70: 151–165.
- Masson V, Vimeux F, Jouzel J, et al. (2000) Holocene climate variability in Antarctica based on 11 ice-core isotopic records. *Quaternary Research* 54: 348–358.
- McManus J, Francois R, Gherardi J-M, Kelgin LD, and Brown-Leger S (2004) Collapse and rapid resumption of Atlantic meridional circulation linked to deglacial climate changes. *Nature* 428: 834–837.
- Mohtadi M, Romero O, and Hebbeln D (2004) Changing marine productivity off northern Chile during the past 19 000 years: A multivariable approach. *Journal of Quaternary Sciences* 19: 347–360.
- Montecinos A and Aceituno P (2003) Seasonality of the ENSO-related rainfall variability in central Chile and associated circulation anomalies. *Journal of Climate* 16: 281–296.
- Moy CM, Seltzer GO, Rodbell DT, and Anderson DM (2002) Variability of El Niño/Southern Oscillation activity at millennial timescales during the Holocene epoch. *Nature* 420: 162–165.
- Pahnke K and Sachs JP (2006) Sea surface temperatures of southern midlatitudes 0–160 kyr BP. *Paleoceanography* 21: PA2003. <http://dx.doi.org/10.1029/2005PA001191>.
- Rein B, Lückge A, Reinhardt L, Sirocko F, Wolf A, and Dullo W-C (2005) El Niño variability off Peru during the last 20,000 years. *Paleoceanography* 20: PA4003. <http://dx.doi.org/10.1029/2004Pa001099>.
- Sepúlveda J, Pantoja S, Hughen KA, et al. (2009) Late Holocene sea-surface temperature and precipitation variability in northern Patagonia, Chile (Jacaf Fjord, 44°S). *Quaternary Research* 72: 400–409.
- Siani G, Colin C, Michel E, et al. (2010) Late Glacial to Holocene terrigenous sediment record in the Northern Patagonian margin: Paleoclimate implications. *Paleogeography, Palaeoclimatology, Palaeoecology* 297: 26–36.
- Steig EJ, Hart CH, White JWC, Cunningham WL, Davis MD, and Saltzman ES (1998) Changes in climate, ocean and ice-sheet conditions in the Ross embayment, Antarctica, at 6 ka. *Annals of Glaciology* 27: 305–310.
- Stott L, Cannariato K, Thunell R, Haug GH, Koutavas A, and Lund S (2004) Decline of surface temperature and salinity in the western tropical Pacific Ocean in the Holocene epoch. *Nature* 430: 56–59.
- Strub PT, Mesias JM, Montecino V, Rutllant J, and Salinas S (1998) Coastal ocean circulation off Western South America. In: Robinson AR and Brink KH (eds.) *The Global Coastal Ocean. Regional Studies and Syntheses*, pp. 273–315. New York: Wiley.
- Thompson LG, Davis ME, Mosley-Thompson E, et al. (1998) A 25,000 year tropical climate history from the Bolivian ice cores. *Science* 282: 1858–1864.
- Thompson LG, Mosley-Thompson E, Davis ME, et al. (1995) Late Glacial stage and Holocene tropical ice core records from Huascaran, Peru. *Science* 269: 46–50.
- Tomczak M and Godfrey JS (2003) *Regional Oceanography: An Introduction*. Delhi: Daya Pub. House.
- Warner NR and Domack EW (2002) Millennial- to decadal-scale paleoenvironmental change during the Holocene in the Palmer Deep, Antarctica, as recorded by particle size analysis. *Paleoceanography* 17: 5–1–5–14.

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PALEOCLIMATE

Contents

Introduction

Timescales of Climate Change

Modern Analog Approaches in Paleoclimatology

Paleoclimate History of the Arctic

The Younger Dryas Climate Event

Introduction

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Introduction, Why Study Paleoclimatology?

The Earth's climate constantly changes, exhibiting variations both temporally and spatially. In a shorter timeframe of every other year or so, climatic changes such as those resulting from El Niño events, prominently affect regions throughout the globe. In much longer timeframes, such as tens of thousands of years ago, dramatic climatic changes have occurred, associated with environmental changes such as the equatorward expansion of large ice sheets and characteristics of vegetation assemblages much different than those seen today. A thorough understanding of the causes of climatic variations is important for society because of the effects of climate on human activities such as agriculture, water resources, energy, biological productivity, and environmental quality. These variations include not only present-day natural variability, such as periodically recurring droughts and floods, but also include future climatic changes. In particular, during the last few decades, scholars have shown increasing concern about human-induced climatic changes such as a doubling of carbon dioxide concentration, which may cause global warming (IPCC, 2001). Some future global warming scenarios from computer models predict potentially widespread catastrophic consequences for society, such as widespread drought at mid-latitudes and unprecedented rapid warming in the Arctic.

Paleoclimatology is defined as the study of climates during any part of the geological past (Bradley, 1999), and specifically deals with climates prior to the period of modern instrumental records, using proxy data (natural phenomena that are climate-dependent) to reconstruct past climates, and using computer models (hereafter referred to as general circulation models (GCMs)) to explain the causes of past climatic change (Kohfeld and Harrison, 2000; Figure 1). The Quaternary period, defined as encompassing the last 2.6 My, is unique in the field of paleoclimatology. This is due to the availability of copious

amounts of data by which to thoroughly understand the climate system at critical times of the past. The study of past climates can greatly help our understanding of climatic change by:

1. Providing a longer-term perspective than the modern instrumental record on understanding the controls, magnitudes, and spatial as well as temporal aspects of climatic change. For example, failure of the Asian monsoon may have occurred more frequently in the past, ranging from several centuries to many thousands of years ago, and the examination of many such cases may provide clues on how particular controls (e.g., El Niño events, mountain uplift) govern monsoonal activity (An et al., 2001; Morrill et al., 2003). Paleoclimatology also provides a longer record of greenhouse gas concentrations and other phenomena linked to climatic forcing mechanisms, and assessment of these longer records, such as carbon dioxide concentrations, may provide different interpretations on the timing of global warming (Ruddiman, 2005).
2. Testing the accuracy of GCMs. Some paleoclimates have boundary conditions and forcing mechanisms that were much different than today and what is expected for the future (Figure 2), especially at longer time scales, such as the peak of the last glaciation around 21 000 years ago (Webb, 1998) and just before the onset of the Quaternary when tectonic uplift played a prominent role in climatic change (Berger, 1992; Ruddiman et al., 1997). However, the variability of forcing mechanisms also exhibits different characteristics within shorter time scales, such as pre-industrial carbon dioxide concentrations in the last 500 years (Robertson et al., 2001). Since the forecasting of future climate relies on the use of GCMs, if model simulations of paleoclimate correspond with those reconstructed from fossil data, we could place more confidence in using these GCMs for simulating and understanding future climatic changes, such as global warming.

3. Providing a longer climate record of natural variability. We need a clear understanding of natural climatic variability, in order to delineate natural from anthropological variability. As a result of increased anthropogenic effects on climate, such as increasing population and urbanization, it is difficult to disentangle these two factors. The general rise in global mean temperature since the mid-1800s provides a clear example (Mann et al., 1998), but other prominent warm periods in the past provide valuable information on how the climate systems operate to further assess future climates (Webb et al., 1993). Scholars know that increased urban heat island effects and carbon dioxide concentrations may play important roles, but could the temperature rise also be a natural response from a colder period known as the Little Ice Age in the seventeenth to nineteenth centuries when carbon dioxide concentrations were lower than today?

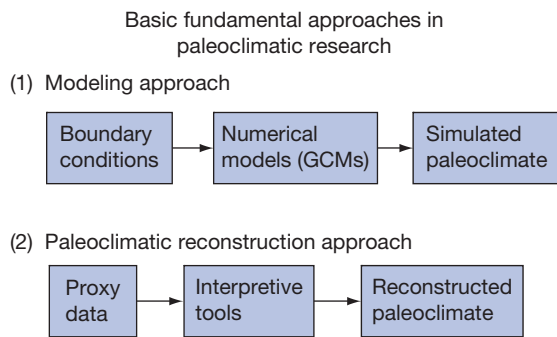


Figure 1 The two basic fundamental approaches in paleoclimate research: (1) the modelling approach and (2) the paleoclimate reconstruction ('data') approach. A common theme that has emerged in paleoclimatology in the last three decades is the synthesis and analysis of 'data/model' comparisons.

The first three reasons for studying paleoclimate help to understand the controls of climatic change and how the climate system operates. However, paleoclimatology is multidisciplinary, benefiting from several fields to understand and reconstruct climates of the past, thus offering important information for paleoecologists. This leads to a fourth important reason for studying paleoclimatology.

4. To provide detailed paleoclimatic records to understand the role of climatic change on ecosystems and society. This importance is best illustrated through some examples. Paleoecologists interested in vegetation histories of a species need to incorporate the role of climatic change over time scales of thousands to millions of years (Overpeck et al., 2003). Archaeologists studying the changing lifestyles of past societies need detailed paleoclimatic records to assess the role of climatic changes (Weiss and Bradley, 2001). The latter example also proves useful for scholars studying climate impact assessment and reconstructing natural hazards, such as the effects of widespread severe drought (Figure 3; The North American Drought Atlas, 2004). Some paleoclimatic proxies possess subseasonal resolution that can sometimes detect specific meteorological events, such as hurricanes and severe flooding, that affected society (Figure 4; Mock et al., 2006). Although most societies functioned differently in the past than they do today, we can learn much from climate impact studies of the past and apply them to understand potential future impacts (Jones et al., 2001).

Some Recent Developments in Paleoclimatology

Prior to 1970, much paleoclimatological research focused primarily on climatic reconstructions that described what happened, with studies involving a variety of different proxy data

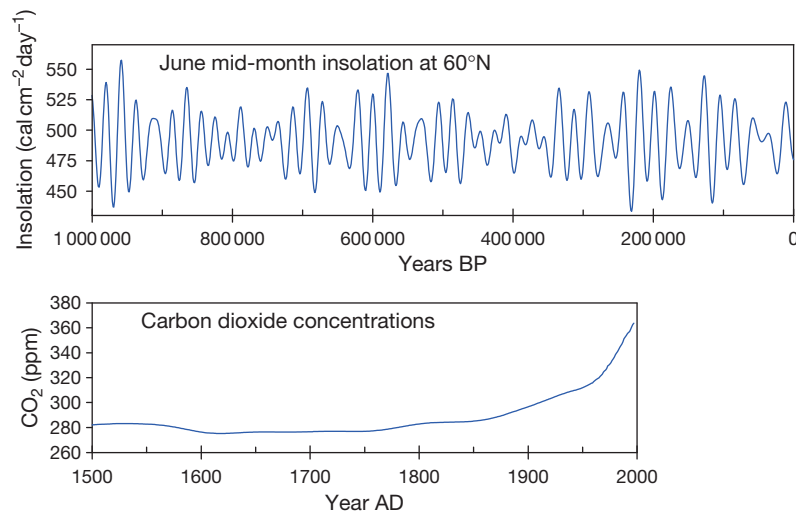


Figure 2 Some examples of boundary conditions of climate over time. Top: June mid-month insolation at 60°N in the last million years; note the cyclical behavior of insolation that represents variations in the Milankovitch orbital parameters. Reproduced from Berger A (1992) *Orbital Variations and Insolation Database*. IGBP PAGES/World Data Center-A for Paleoclimatology Data Contribution Series # 92-007. NOAA/NGDC Paleoclimatology Program, Boulder, CO. Bottom: Carbon dioxide concentrations since AD 1500. Reproduced from Robertson AD, Overpeck JT, Rind D, et al. (2001) Hypothesized climate forcing time series for the last 500 years. *Journal of Geophysical Research: Atmospheres* 106: 14783–14803. Paleoclimatic reconstruction data from the NOAA Paleoclimatology website at <http://www.ncdc.noaa.gov/paleo/paleo.html>.

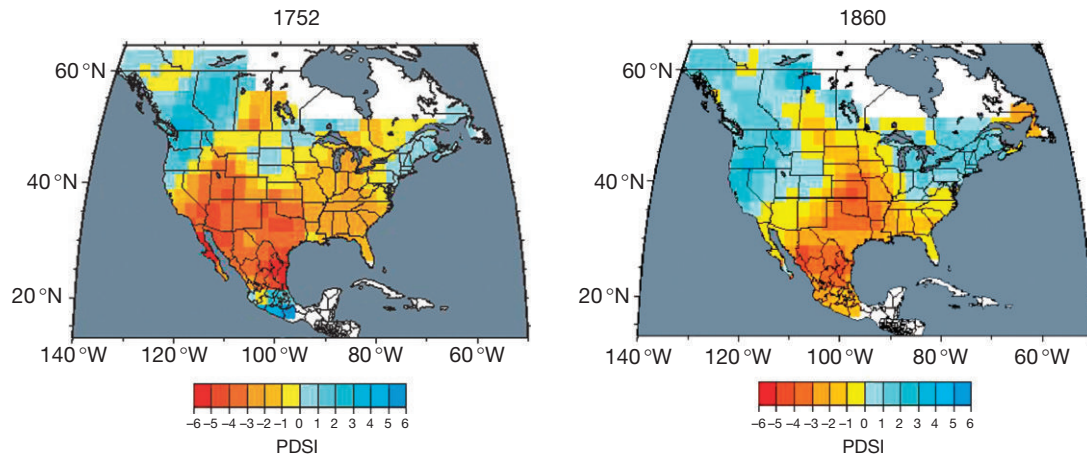


Figure 3 Geographical extent of Palmer Drought Severity Indices (PDSI) for two extreme drought years for the southern Great Plains, Mexico, and the Southeastern USA: 1752 and 1860. Negative indices represent the magnitude of severe drought. The drought maps were created from a network of 835 tree-ring sites. Reproduced from Cook ER and Krusic PJ (2004) North American Summer PDSI Reconstructions. IGBP PAGES/World Data Center for Paleoclimatology Data Contribution Series # 2004-045. NOAA/NGDC Paleoclimatology Program, Boulder, CO.

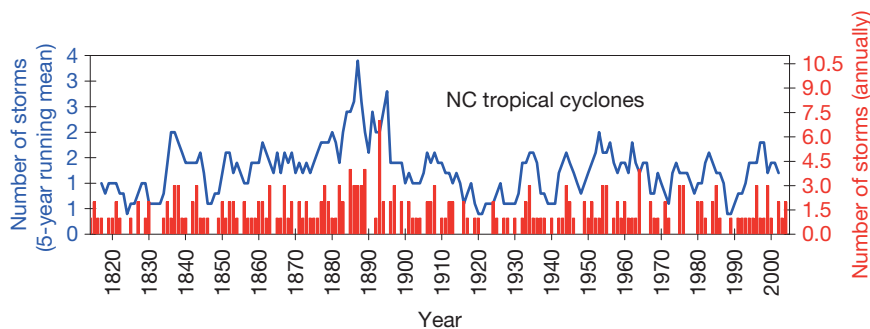


Figure 4 Annual time series of the number of tropical cyclones affecting the coast of North Carolina, USA. General trends are indicated by a 5-year running mean. Tropical cyclone data were taken from the HURDAT database and archival collections owned and compiled by Mock and colleagues at South Carolina.

types (Wendland, 1991). Radiometric dating techniques, such as radiocarbon and potassium-argon dating, provided a quantitative means to date past climatic change. Paleoclimatic research was propelled by the establishment of numerous research centers that specialized in particular proxy data and dating methods. For example, dendroclimatology, the study of tree-rings, accelerated after the establishment of the Laboratory of Tree Ring Research at the University of Arizona, USA in 1937. Similar dendroclimatic laboratories were established later at Columbia University in New York, the University of Arkansas, the Climatic Research Unit, University of East Anglia, United Kingdom, and the Swiss Federal Institute for Forest, Snow, and Landscape Research, Switzerland, among many. Research centers focusing on other proxies such as palynology also emerged, including the University of Minnesota in the United States, University of Cambridge in England, Lund University in Sweden, the University of Bern in Switzerland, and the Russian Academy of Sciences, Institute of Geography in Moscow. Similarly, prominent research centers in Quaternary paleoceanography emerged as well, with some notable centers being based at Cambridge University, Brown University, and Columbia University. As time progressed, improvements and new techniques in data analysis were developed at

prominent research centers such as the Quaternary Research Center in the United States, and the Xian Laboratory of Loess and Quaternary Geology at the Chinese Academy of Sciences. Newly trained academics that graduated from these research centers started research centers of their own, building up paleoclimatic databases.

Beginning in the early 1970s, the development of high-speed computers fostered a new type of paleoclimatology that specializes in analyzing large paleoclimatic datasets (Wright and Bartlein, 1993). Some interpretive tools in paleoclimatic analyses are qualitative in nature, which continue to this day and can involve analyses from local to hemispheric scales (Figure 1). Earlier quantitative studies applied basic transfer functions to convert proxy variables into climate variables, which also involved the calibration of modern climatic data with modern environmental data. The modern relationships were applied to fossil environmental data to quantitatively reconstruct past climate (Webb and Bryson, 1972). As the datasets grew, so did the sophistication of quantitative interpretive tools for analyzing large-scale paleoclimatic datasets (Mann et al., 1998; Prentice et al., 1991). For example, the recently compiled North American Drought Atlas, which provides geographical maps of drought severity by year, is based

on a geographical network of 835 tree-ring sites (Figure 3; Cook and Krusic, 2004). Many paleoclimatic data networks are now available through the World Data Center-A for Paleoclimatology (WDC-A) at Boulder, CO (Webb et al., 1994), and other mirror sites around the world at Johannesburg in South Africa, Lanzhou in China, Mendoza in Argentina, Nairobi in Kenya, and Pune in India (Eakin et al., 2003). These data networks range from the regional to global scale, with examples being the International Tree Ring Database (Grissino-Mayer and Fritts, 1997) and the Global Pollen Database.

The computer revolution also created a paleoclimatic perspective for dealing with GCMs. These models are similar to those used in daily weather forecasting but instead the principles are applied to simulate large-scale climate patterns of the past. Earlier attempts focused mostly on the atmosphere, but paleoclimatic modeling has evolved to link atmospheric models with detailed feedbacks as they relate to processes in the biosphere, lithosphere, and hydrosphere (Kohfeld and Harrison, 2000; Kutzbach et al., 1998). Considerable attention has been paid to ocean-atmospheric feedbacks. GCMs have been used to simulate paleoclimates ranging from a few hundred to millions of years ago (Kutzbach, 1992), as well as selected timeframes and phenomena of interest in the past (LeGrande et al., 2006; Seager et al., 2005). As opposed to paleoclimatic proxy data, which independently reconstruct 'what happened' (Figure 1), GCMs explain 'why things happened' and thus are an extremely useful tool for paleoclimatologists to test hypotheses concerning the causes of climatic change by comparing simulation results with those derived from proxy data (Harrison and Prentice, 2003; Mahowald et al., 1999). Within the last two decades, many different paleoclimatic modeling groups have emerged and remain active; some of these include modeling activities at the National Center for Atmospheric Research in Boulder, CO, the Hadley Centre in the United Kingdom, the Canadian Centre for Climate Modeling and Analysis, the Max-Planck Institute for Meteorology at Bremen University, the Laboratoire de Météorologie Dynamique in France, and the Goddard Institute of Space Studies in Maryland, USA.

Increased resolution in dating techniques and the growing body of paleoclimatic evidence, particularly from ice cores and marine sediments from the North Atlantic, indicate that abrupt decadal-centennial scale changes in climate occurred in the distant past, that are much different in magnitude and character than those observed in the modern instrumental record (Clark et al., 1999; Labeyrie et al., 2003; Overpeck, 1996). These changes are significant for society, for we now know that such abrupt climatic changes may occur within a single human lifetime. Paleoclimatic records thus offer the only means to test whether our predictive models can simulate such future changes. Modeling attempts have been made to simulate the causes and nature of these abrupt changes. These model runs are enabling scientists to conduct detailed data-model comparisons at the hemispheric and global scales (Clark et al., 2002). However, to date, most modeling studies on these events are still focusing on providing sensitivity tests to assess potential forcing mechanisms. We are only beginning to document and understand the controls and causes of these abrupt changes, and such questions will continue to be important for the paleoclimatology community for years to come.

Organization of the Articles

Numerous articles in this encyclopedia describe how paleoclimatology can contribute toward a further understanding of the climate system of the past. These articles represent the most up-to-date information from the paleoclimate research community. Bartlein provides an overview of the concepts on how climate varies at different time scales, ranging from inter-annual to pre-Quaternary, and how we can interpret some reconstructions from proxy data in relation to climate model simulations. Shuman further discusses the details of how Quaternary scientists use different data, methods, and tools for reconstructing past climates, including data/model comparisons as well as various geographical, quantitative, and qualitative approaches. Harrison describes the current status of data/model comparisons in paleoclimatology.

Several articles provide more details on important forcing mechanisms that affect feedbacks in the climate system, and these are important parts of climate model simulations. The importance of the thermohaline circulation on past climates is well-known, and Rahmstorf describes how changes of various fluxes in the oceanic environment may relate to glacial fluctuations and other rapid climatic changes. Brozkin illustrates that although climate is a prominent control on geographical aspects of the terrestrial biosphere, various other factors in the terrestrial biosphere may also alter feedbacks that affect the climate system. Tegen describes the importance of atmospheric dust on climate at different time scales, including attempts to quantify its magnitude. Shindell illustrates how solar and volcanic forcing are related to climatic variability at different time scales, also approaching this from a modeling perspective.

Another group of articles describe specifics on the nature of climate variability at different time scales from throughout the Quaternary and back into the Pliocene. Poore provides an important review of pre-Quaternary climate and forcing mechanisms that helps assess the onset of paleoclimatic variations in relation to multiple glacial and interglacial cycles in the Quaternary. Owen also contributes valuable information on pre-Quaternary paleoclimate, focusing attention on tectonics and continental configurations, but also addressing several different aspects of tectonics. Berger describes an overview of the astronomical theory of Quaternary paleoclimate, discussing the prominent orbital cycles as well as the role of carbon dioxide in climate change.

The late Pleistocene/Holocene timeframe of approximately the last 130 000 years has been the focus of many studies; some continuous, high-resolution proxies (e.g., ice cores) and climate modeling researchers have expressed a high degree of confidence in the prominent climate forcing mechanisms. Montoya illustrates the details of modeling simulations of the paleoclimate of the Last Interglacial (around 125 000 years ago), which is a time when orbital factors were conducive to widespread warmth in many parts of the world. Broccoli discusses the nature of research on model simulations of the Late Glacial Maximum (LGM) of 21 000 years ago, and demonstrates the importance of nonorbital factors such as ice-sheet size, dust, and greenhouse gases. Labeyrie illustrates that there are also more rapid, millennial-scale changes that occur within the orbital-induced changes and prominent glacial-interglacial

cycles. These events, commonly referred to as Dansgaard/Oeschger and Heinrich events, are prominent in many high-resolution paleoclimatic records but more work is needed on modeling efforts. Barber describes a regional example, illustrating how peatland records in northwest Europe can aid in multiproxy efforts of reconstructing past regional climate patterns through the Holocene.

Another group of articles describes the nature of paleoclimatic records at time scales within the last 2000 years, as well as addressing their importance to contemporary climatic issues and societal impacts. Smith provides an overview of dendroclimatology, which is a prominent paleoclimate proxy with annual temporal resolution for many terrestrial regions. Chenoweth provides an overview of research on historical climate reconstruction from archival materials, some of which include data resolved at the daily level. These materials are mostly limited to the last several centuries, depending on the geographical area of interest. Mann elegantly describes how different high-resolution proxies, including tree-rings and historical evidence, can be combined to quantitatively reconstruct temporal and spatial patterns of high-resolution climate over the last millennium. These reconstructions capture many important, dominant climatic modes that are seen in the modern instrumental record, but they also reveal changes in the importance of these modes over time. Liu provides an overview on paleotempestology, the reconstruction of past hurricanes and typhoons that encompass time scales from the last few hundred years to most of the Holocene, based on several different proxies. He also provides discussion of some of the forcing mechanisms that govern past hurricane activity. C Mock provides two articles on paleo-ENSO (El Niño-Southern Oscillation, Paleo-ENSO) and paleodrought, respectively. As with paleotempestology, both of these phenomena are climatic extremes of high interest to society that can be reconstructed from paleoclimatic proxies, also extending through the Holocene. Last, Schneider and Mastrandrea illustrate how longer records from paleoclimatology aid society in further understanding anthropogenic global climate change, particularly global warming, that may occur in the future.

See also: Dendroclimatology; Paleoclimate Relevance to Global Warming. **Glacial Climates:** Biosphere Feedbacks; Effects of Atmospheric Dust; Thermohaline Circulation; Volcanic and Solar Forcing. **Glaciation, Causes:** Astronomical Theory of Paleoclimates; Tectonic Uplift and Continental Configurations. **Paleoclimate:** Timescales of Climate Change. **Paleoclimate Modeling:** Data–Model Comparisons; Last Glacial Maximum GCMs; Paleo-ENSO; The Last Interglacial. **Paleoclimate Reconstruction:** Approaches; Historical Climatology; Paleodroughts and Society; Paleotempestology; Pliocene Environments; Sub-Milankovitch (DO/Heinrich) Events; The Last Millennium. **Plant Macrofossil Methods and Studies:** Mire and Peat Macros.

References

- An ZS, Kutzbach JE, Prell WL, and Porter SC (2001) Evolution of Asian monsoons and phased uplift of the Himalaya-Tibetan plateau since Late Miocene times. *Nature* 411: 62–66.
- Berger A (1992) *Orbital Variations and Insolation Database*. IGBP PAGES/World Data Center-A for Paleoclimatology Data Contribution Series # 92-007. NOAA/NGDC Paleoclimatology Program, Boulder, CO.
- Bradley RS (1999) *Paleoclimatology, Reconstructing Climates of the Quaternary*. San Diego, CA: Academic Press.
- Clark PU, Pisas NG, Stocker TF, and Weaver AJ (2002) The role of the thermohaline circulation in abrupt climate change. *Nature* 415: 863–869.
- Clark PU, Webb RS, and Keigwin LD (eds.) (1999) *Mechanisms of Millennial-Scale Global Climate Change*. In: *American Geophysical Monograph*, vol. 112. Washington, DC: American Geophysical Union.
- Cook ER and Krusic PJ (2004) North American Summer PDSI Reconstructions. IGBP PAGES/World Data Center for Paleoclimatology Data Contribution Series # 2004-045. NOAA/NGDC Paleoclimatology Program, Boulder, CO.
- Eakin CM, Diepenbroek, and Hoepffner M (2003) Appendix B – The PAGES data system. In: Alverson K, Bradley R, and Pedersen T (eds.) *Paleoclimate, Global Change and the Future*, pp. 176–179. Berlin: Springer.
- Grissino-Mayer HD and Fritts HC (1997) The International Tree-Ring Data Bank: An enhanced global database serving the global scientific community. *The Holocene* 7: 235–238.
- Harrison SP and Prentice IC (2003) Climate and CO₂ controls on global vegetation distribution at the last glacial maximum: Analysis based on palaeovegetation data, biome modelling and palaeoclimate simulations. *Global Change Biology* 9: 983–1004.
- IPCC (2001) Houghton JT, Ding Y, Griggs DJ, Noguer M, van der Linden PJ, Dai X, Maskell K, and Johnson CA (eds.) *Climate Change 2001: The Scientific Basis. Contribution of Working Group I to the Third Assessment Report of the Intergovernmental Panel on Climate Change*. Cambridge: Cambridge University Press.
- Jones PD, Ogilvie AEJ, Davies TD, and Briffa KB (2001) Unlocking the doors to the past: Recent developments in climate and climate-impact research. In: Jones PD, Ogilvie AEJ, Davies TD, and Briffa KR (eds.) *History and Climate: Memories of the Future?*, pp. 1–8. Dordrecht, The Netherlands: Kluwer Academic/Plenum.
- Kohfeld KE and Harrison SE (2000) How well can we simulate past climates? Evaluating earth system models using global palaeoenvironmental datasets. *Quaternary Science Reviews* 19: 321–346.
- Kutzbach JE (1992) Modeling large climatic changes of the past. In: Trenberth K (ed.) *Climate System Modeling*, pp. 669–688. Cambridge: Cambridge University Press.
- Kutzbach JE, Gallimore R, Harrison SP, Behling P, Selin R, and Laarif F (1998) Climate and biome simulations for the past 21 000 years. *Quaternary Science Reviews* 17: 473–506.
- Labeyrie L, Cole J, Alverson K, and Stocker T (2003) The history of climate dynamics in the late Quaternary. In: Alverson K, Bradley R, and Pedersen T (eds.) *Paleoclimate, Global Change and the Future*, pp. 33–61. Berlin: Springer.
- LeGrande AN, Schmidt GA, Shindell DT, et al. (2006) Consistent simulations of multiple proxy responses to an abrupt climate change event. *Proceedings of the National Academy of Sciences* 103: 837–842.
- Mahowald N, Kohfeld K, Hansson M, et al. (1999) Dust sources and deposition during the last glacial maximum and current climate: A comparison of model results with paleodata from ice cores and marine sediments. *Journal of Geophysical Research* 104: 15895–15916.
- Mann ME, Bradley RS, and Hughes MK (1998) Global-scale temperature patterns and climate forcing over the past six centuries. *Nature* 392: 779–787.
- Mock CJ, Holmberg SO, Glenn DA, Dodds SF, and Tanis JR (2006) Historical hurricane records of the Southeastern Atlantic Coast. Abstract. In: *102nd Annual Meeting of the Association of American Geographers*, Chicago, IL. The North Atlantic Hurricane Database (HURDAT) <http://www.nhc.noaa.gov/pastall.shtml>.
- Morrill C, Overpeck JT, and Cole JE (2003) A synthesis of abrupt changes in the Asian summer monsoon since the last deglaciation. *The Holocene* 13: 465–476.
- Overpeck JT (1996) Warm climate surprises. *Science* 271: 1820–1821.
- Overpeck JT, Whitlock C, and Huntley B (2003) Terrestrial biosphere dynamics in the climate system: Past and future. In: Alverson K, Bradley R, and Pedersen T (eds.) *Paleoclimate, Global Change and the Future*, pp. 81–111. Berlin: Springer.
- Prentice IC, Bartlein PJ, and Webb T III (1991) Vegetation and climate change in eastern North America since the last glacial maximum. *Ecology* 72: 2038–2056.
- Robertson AD, Overpeck JT, Rind D, et al. (2001) Hypothesized climate forcing time series for the last 500 years. *Journal of Geophysical Research: Atmospheres* 106: 14783–14803.
- Ruddiman WF (2005) *Plows, Plagues, and Petroleum: How Humans Took Control of Climate*. Princeton, NJ: Princeton University Press.
- Ruddiman WF, Raymo ME, Prell W, and Kutzbach JE (1997) The uplift-climate connection: A synthesis. In: Ruddiman WF and Prell W (eds.) *Global Tectonics and Climate Change*, pp. 471–515. New York: Plenum.

- Seager R, Kushnir Y, Herweijer C, Naik N, and Velez J (2005) Modeling of tropical forcing of persistent droughts and pluvials over western North America: 1856–2000. *Journal of Climate* 18: 4068–4091.
- The North American Drought Atlas (2004) <http://iridl.ldeo.columbia.edu/SOURCES/LDEO/TRL/NADA2004/pdsi-atlas.html>.
- Webb T III (ed.) (1998) Late Quaternary climates: Data syntheses and model experiments. *Quaternary Science Reviews* 17(6–7): 465–688.
- Webb RS, Anderson DM, and Overpeck JT (1994) Editorial: Archiving data at the World Data Center-A for paleoclimatology. *Paleoceanography* 9: 391–393.
- Webb T III and Bryson RA (1972) Late- and postglacial climatic change in the northern Midwest USA: Quantitative estimates derived from fossil pollen spectra by multivariate statistical analysis. *Quaternary Research* 2: 70–115.
- Webb T III, Crowley TJ, and Frenzel B (1993) Group Report: Use of paleoclimatic data as analogs for understanding future global changes. In: Eddy JA and Oeschger H (eds.) *Global Climate Changes in the Perspective of the Past*, pp. 51–70. Chichester: Wiley.
- Weiss H and Bradley RS (2001) What drives societal collapse? *Science* 291: 609–610.
- Wendland WM (1991) Emergence and evolution of paleoclimatology. *Physical Geography* 12: 287–299.
- Wright HE Jr. and Bartlein PJ (1993) Reflections on COHMAP. *The Holocene* 3: 89–92.

Timescales of Climate Change

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Introduction

Climate varies. Although often defined in terms of ‘average weather,’ or the expected values of statistical parameters over some interval, when looking at climatic time series plotted over any interval (from years to millions of years) for any location (from point to globe), one sees the tremendous variability of the record. Typical values or central tendencies for a climate variable can indeed be observed, as can general ranges of values within which the variable remains. However, most of the time a climate variable (and that part of the climate system it represents) might be described as continuously heading from one level to another, never remaining at a particular value for long. Intervals with low variability are generally the exceptions rather than the rule in climatic time series. Here, ‘variance’ refers to the statistical quantity (the mean of the squared deviations about the long-term mean); ‘variability’ refers to the tendency of records to vary; and ‘variation’ refers to a particular portion or feature of a record or to the time-course of a variable.

Variability of climate occurs across a range of timescales (the focus of this article) and does so in a fashion that exhibits considerable structure. Rather than displaying a simple pattern of variability that increases as a function of time span or record length, as with a system that obeyed simple scaling laws (Kantz and Schreiber, 2004), the climate system has some preferred temporal scales of variability that reflect the nature of the external forcing of the climate system or the internal time constants of the components of the climate system itself (Saltzman, 2002). Although some of these external controls are intrinsically periodic, such as the variations of the Earth’s orbital elements (Imbrie et al., 1984), most are not. Similarly, the internally generated variations of climate, including some that might be described as ‘cyclical,’ are best described as only quasiperiodic.

One approach for characterizing or describing the variability of climate is the variance spectrum, which displays a statistical property, variance, as a function of the period or frequency of the variation. ‘Peaks’ in such a display point to particular frequencies or timescales at which variability is concentrated. A schematic version of a variance spectrum for the climate system for timescales from a year to a billion years is illustrated later. In addition, evolutionary or ‘moving-window’ spectral analyses can be used to reveal the changing importance over time of variability on different timescales, and two examples are also described.

Perhaps owing to its heritage in signal processing (Jenkins and Watts, 1968) and to its wide application in developing the astronomical (Milankovitch) theory of climate change, the application of the variance spectrum to illustrate timescales of climate variability naturally directs a user toward the periodic or oscillatory features in a time series. Hence, explanations of those features usually invoke some kind of regular cyclical mechanisms, either external (Sun, Moon, planets) or internal (oscillatory ‘climate-modes’).

Several factors may predispose researchers to see cycles in time series (in addition to the tendency to see order in apparent chaos). These include the tendency for climate variables to vary between some general limits, and the consequences of applying some common data-analytical methods, such as the variance spectrum or moving-average filters. A few examples discussed later will show how spurious periodicities can arise when characterizing the timescales of climate variability.

An alternative approach for describing the nature of climate variability is to describe the characteristic features or kinds of variations that recur in many climatic time series. There is a relatively small number of kinds or patterns of variation that climatic time series display. These occur across a range of different timescales and can be distinguished by the extent of their influence throughout different levels in the climate-system hierarchy, their predictability, and the extent to which they represent progressive trends, or regular or irregular variations.

Climate Variations

The Hierarchy of Climate System Controls and Responses

Climatic variations occur within a hierarchy of controls and responses, which begin at the highest level with the external controls of climate, proceed through global, hemispheric, continental, and regional scales, and end with the variations of individual climate variables at specific locations at the lowest level (Bartlein, 1997). Responses at any one level of the hierarchy become the controls of variations of the components at lower levels. For example, on timescales of 10^4 – 10^6 years, ice sheets are dependent variables in the climate system, governed by orbitally controlled variations in insolation. At shorter timescales (10^4 – 10^3 years), ice sheets act as independent variables that have an important influence on global temperatures and atmospheric circulation. With the exception of the annual cycle, there is a general tendency for the variations of components at higher levels in the hierarchy to exhibit more long-term variability, while those at lower levels experience more short-term variability.

The existence of this hierarchy also has implications for attempts to explain the variations at a particular location. For example, although climatic variations at a location are ultimately governed by global-scale controls, a specific paleoclimatic record is not representative of the general state of the global system. This situation arises because the intermediate controls and responses have the potential of reinforcing, canceling, or even reversing the longer-term global trends. Gradual changes in large-scale controls may sometimes produce abrupt local changes when atmospheric circulation is reorganized. Conversely, abrupt changes in the large-scale circulation may produce warming in some regions and cooling or no change in others, as can be seen in the spatial anomaly patterns of year-to-year variations of climate. Consequently, while it may be

difficult or even impossible to ascribe a particular climate variation at a location to a specific configuration of higher-level controls, shorter-term variations at lower levels are strongly conditioned by the particular state of system at higher levels. Therefore, any discussion of the timescales of climatic variability should explicitly acknowledge the spatial scale or extent of the system being discussed.

Some Representative Paleoclimatic Time Series

To illustrate the spectrum and characteristic features of climatic variations across a range of timescales, it will be useful to examine some representative time series. **Figure 1(a)** is a composite global oxygen-isotope record that provides a general description of the global cooling during the Cenozoic that culminated in Quaternary glacial–interglacial variations (Zachos et al., 2001). This record was constructed from a number of individual records, which can be plotted as individual points, but not as time series. In addition to the general trend, individual steps toward colder conditions are evident. **Figure 1(b)** shows another composite record, this time constructed by superimposing or ‘stacking’ individual records to form a single series (Lisiecki and Raymo, 2005). This shows the periodic variations in global ice volume associated with the Earth’s orbital elements, as well as the changes in the relative importance of the variations at different timescales.

Oxygen-isotopic variations from the GISP 2 Greenland ice-core record are shown in **Figure 1(c)** (Stuiver et al., 1997). This series shows the large-amplitude millennial timescale variations known as the Dansgaard–Oeschger cycles that occur throughout most of the record, the Younger Dryas climate reversal (12 800–11 600 years BP, Alley et al., 1993), and the much lower amplitude variability in this record during the Holocene. **Figure 1(d)** shows a speleothem oxygen-isotope record that exhibits the long-term decline in the strength of the East Asian monsoon, along with decadal-to-century variations throughout the Holocene, a pattern typical of many Holocene records (Mayewski et al., 2004). **Figure 1(e)** illustrates a reconstruction of Northern Hemisphere temperature anomalies over the past two millennia (Moberg et al., 2005), along with the observational record of the past century (Hansen et al., 2006). Like other similar reconstructions (Mann and Jones, 2003), this record shows interannual-to-century-scale variations, and the unprecedented increase in temperature over recent decades.

These series do not illustrate the entire range of variations experienced by the climate system over the Quaternary. A network of time series that span the different timescales and levels in the hierarchy would be required, and such a network does not exist. However, this set of series does illustrate the variety of characteristic features in climatic time series that will be discussed further later.

The Spectrum of Climate Variability

There are several ways of schematically illustrating the nature of climate variability (in addition to simply plotting representative series). These include ‘powers-of-ten’ presentations (Webb, 1989), wherein climatic time series of different duration (e.g., the past 100 000, the past 10 000, the past 1000 years) are

plotted and examined; portrayal of the variance spectrum of climatic time series, which reveals the relative importance of variations at different timescales (Mitchell, 1976; Shackleton and Imbrie, 1990); or ‘scale diagrams’ that indicate the characteristic temporal and spatial scales on which the particular components of climate vary (McDowell et al., 1991). Of these, the variance spectrum has often been used to illustrate the relative importance of climatic variations on different timescales.

The Variance Spectrum

The variance spectrum displays the relative or absolute magnitude of variance in a time series at a particular period or frequency, as a function of that period or frequency (**Figure 2**), and provides a ‘frequency domain’ summarization of climate variability. Typically, variance spectra are plotted on log scales to accommodate the large range of timescales discussed, and the general tendency is for the total variance of longer-term variations of the climate system (e.g., glacial–interglacial variations) to be orders-of-magnitude larger than shorter-term variations (e.g., interannual variations). Mitchell (1976) drew a schematic variance spectrum for climate-system variations that covered timescales ranging from the age of Earth (geological timescales) down to hours (meteorological timescales). Based on analyses of series that represented parts of this time range (Hays et al., 1976; Kutzbach and Bryson, 1974), as well as considerable intuition, Mitchell’s spectrum was innovative in its attempt at presenting a general conceptual model that organized a large body of insight into how the climate system operates. Subsequent workers have progressively refined Mitchell’s schematic spectrum (Saltzman, 2002), or developed data-derived spectra for parts of the record (Shackleton and Imbrie, 1990; see also Wunsch, 2003).

Figure 2 is based on Mitchell’s spectrum, supplemented by material from the descendent papers. It differs somewhat from Mitchell’s in that the spectrum here was drawn to display an overall slope of -2 , that is, the ‘power-law’ spectrum of a simple random walk, in order to depict the long-run ‘nonstationarity’ of the climate system (Wunsch, 2003). Superimposed on this general background variability, which dominates the spectrum, are ‘peaks’ that represent the concentrations of variability associated with external drivers or internal autovariations. Some of these, like those associated with the annual cycle, ENSO, and the orbital elements, have underlying physical mechanisms that are truly cyclical (and hence are drawn with sharper peaks), while others, like the Holocene or tectonic ‘cycles’ do not. These are commonly referred to as ‘cycles’ but at best are only quasi-, or apparently, periodic. They are drawn here with broader peaks or ‘shoulders’ in the spectrum.

The first-order pattern on the schematic spectrum is the great increase in variance that occurs with increasing period, with second-order peaks of lower importance superimposed on this overall pattern. Comparisons can be made between the peaks and overall background variance at particular timescales. For example, the variations in global ice volume at a period of 100 ka that characterize the past million years (**Figure 1(b)**) have similar variance as the variations in the general mean state of climate that take place on 5–10 My timescales (**Figure 1(a)**). The tectonic ‘cycle’ peaks represent climate-system variability on timescales of 400 My (‘Wilson cycles’ of continental

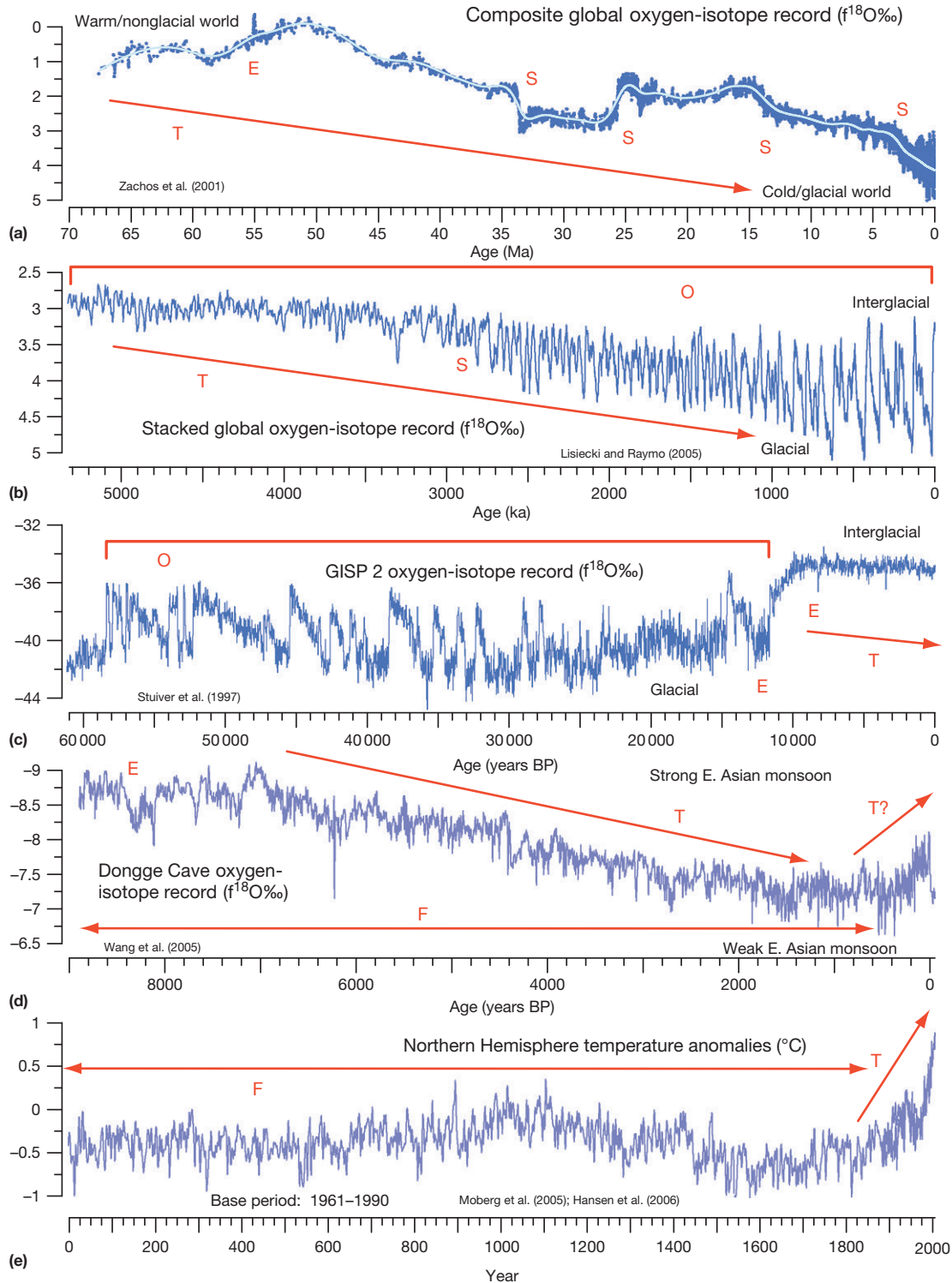


Figure 1 Some representative paleoclimatic time series and examples of typical kinds of climatic variations: (a) composite global oxygen-isotope record for the Cenozoic (Zachos et al., 2001); (b) stacked global oxygen-isotope record (Lisiecki and Raymo, 2005); (c) oxygen-isotopic variations from the GISP 2 Greenland ice-core record (Stuiver et al., 1997); (d) Dongge Cave speleothem oxygen-isotope record (Wang et al., 2005); (e) reconstruction of Northern Hemisphere temperature anomalies over the past two millennia (Moberg et al., 2005), along with the observational record of the past century (Hansen et al., 2006). Typical kinds of climatic variations (Table 1) are illustrated as follows: T, trend; S, step; O, oscillation; F, fluctuation; and E, event.

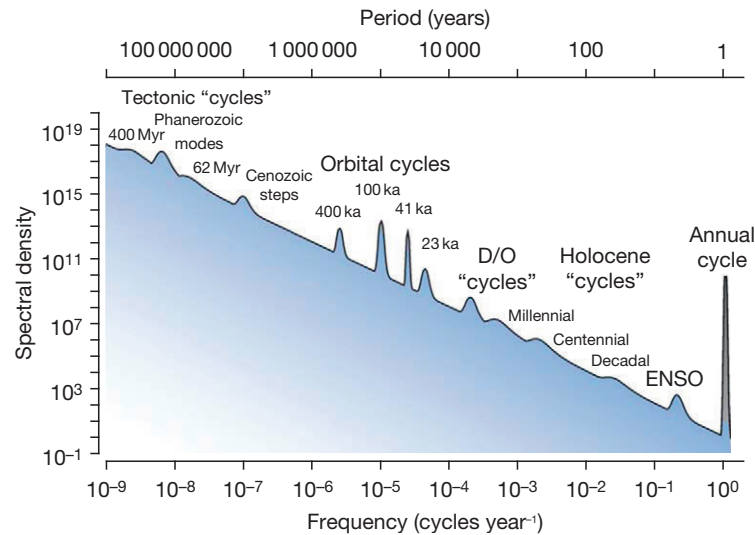


Figure 2 Schematic variance spectrum of climatic variations.

assembly and reassembly; Wilson, 1966), 200 My (variations among Phanerozoic climate modes, Frakes et al., 1992), 62 My (fossil diversity variations; Rhode and Muller, 2005), and steps in the Cenozoic oxygen-isotope record that are approximately 10 My apart (Figure 1(a); Zachos et al., 2001). Orbital cycles include the 400, 100, 41, and 23 (plus 19) ka climate (ice-volume and monsoon-strength) variations related to periodic insolation variations. A broad spectral peak is shown for Dansgaard-Oeschger cycles (Figure 1(c)), although these variations are not strictly periodic (Alley et al., 2001). Holocene timescale variations are shown schematically as broad peaks corresponding to millennial, centennial, and decadal-scale variability. Although often described as ‘cycles’ (e.g., ‘1500-year Bond cycles,’ see Bond et al., 2001), the periodicity of the variations is not sufficiently regular for them to be considered truly cyclical. Similarly, ENSO-related variations are only quasi-periodic and contribute a broad peak to the variance spectrum, while the truly periodic annual cycle contributes a very narrow peak. (The width of the annual-cycle peak on Figure 2 is exaggerated for visibility.)

The overall impression from either the schematic spectra, or from the time series themselves, is that (with the exception of the annual cycle) the variations of the climate system are almost completely stochastic. Even the variations arising from external controls that are nearly deterministic (such as the variations of the orbital elements that generate ice-volume variations) or from internal variations with physical generating mechanisms that are oscillatory in nature (such as ENSO) have considerable unpredictable ‘noise’ components. Variability of climate could thus be described as occurring over all timescales, with variability greatly increasing as scale increases, but with concentrations of variance at some preferred timescales.

Changing Variability

A variance spectrum, such as that in Figure 2, displays the ‘global’ level of variance at specific periods, and consequently masks the ‘local’ (over time) changes in the importance of

variability at particular periods, which is evident in many climatic time series (Figure 1(b) and (c)). One device for illustrating such changes in variability or in the relative importance of variations on different timescales is the evolutionary spectrum, also known as a moving-window spectrum. Evolutionary spectra constructed using the wavelet procedure (Torrence and Compo, 1998) for the stacked marine oxygen-isotope record (Lisiecki and Raymo, 2005) and the bidecadal GISP 2 oxygen-isotope record (Stuiver et al., 1997) are shown in Figure 3. The plots show ‘scaled power’ or variance plotted as a function of period of variation and time.

The evolutionary spectrum for the stacked global oxygen-isotope record clearly shows the increase in importance in 41-ka variations beginning around 2.65 Ma and in 100-ka variations beginning around 1 Ma, but it also shows that variations at these timescales were important during previous times. The 41-ka variations are evident throughout the record and variations at the other orbital periods can be seen to vary in their relative importance. The evolutionary spectrum for the GISP 2 oxygen-isotope data (Figure 3) clearly shows the importance of millennial-scale Dansgaard-Oeschger variations prior to the Holocene and also illustrates the changing period of those variations. From the beginning of the record plotted here until around 42 000 years BP, variance is concentrated at periods around 5000 years, and shifts to progressively shorter periods from then until 30 000 years BP. During MIS 2 and during the Holocene, there is little concentration of variance at any period shorter than 5000 years. The high variance at the longest timescales (e.g., 10 000 years and longer), although not demonstrably significant due to record-length constraints, reflects the important underlying signature of orbital-timescale variations in this record.

Changes in the relative importance of variations on different timescales are the rule in climatic time series, appearing in a variety of differing contexts (Overpeck and Webb, 2000). Explanation of the changes in variability of climatic time series presents a potentially greater challenge than explanation of changes in the mean, but must be accepted owing to the great impact climate variability has on human activity.

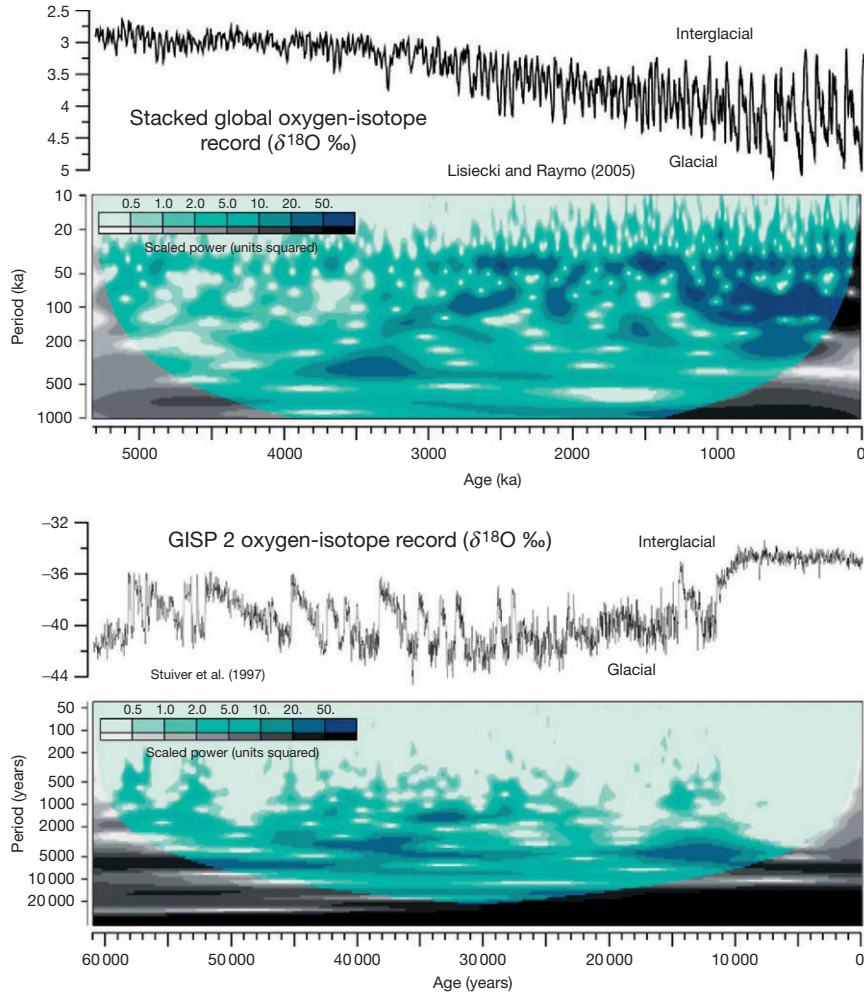


Figure 3 Evolutionary spectra calculated using the wavelet method (Torrence and Compo, 1998; <http://atoc.colorado.edu/research/wavelets/>) for the stacked global oxygen-isotope record (Lisiecki and Raymo, 2005) and oxygen-isotopic variations from the GISP 2 Greenland ice-core record (Stuiver et al., 1997). Gray areas indicate the regions outside of the 'cone of influence' where record-length considerations make the estimates less robust.

Spurious Periodicity

As mentioned earlier, the use of the variance spectrum to characterize climate variability, when coupled with the general tendency of climatic time series to wander back and forth, naturally directs attention toward explanations of climatic variability that invoke controls or mechanisms that are cyclic in nature. However, it is likely that the nature of the variability plus some commonly applied data-analytical tools yield series that merely appear periodic when they intrinsically are not so.

Figure 4 shows three examples of time series that feature quasiperiodic variations, each generated by filtering or transforming a series of normally distributed random numbers in an intrinsically aperiodic way. In each figure, the random numbers are plotted in gray in the background on an arbitrary scale, and the generated series in black.

Figure 4(a) shows a time series generated by the second-order autoregressive model:

$$z_t = \phi_1 z_{t-1} + \phi_2 z_{t-2} + a_t$$

where $\phi_1 = 1.32$ and $\phi_2 = -0.66$, and a_t is a normally distributed or Gaussian white-noise time series. This is the 'classical' second-order autoregressive or AR(2) model that describes Wolfer's sunspot series (Box and Jenkins, 1976), and autoregressive models have generally found wide application in representing time series with both short- and long-run memory. The resulting series clearly shows a rough 10 time-step oscillation, and further displays a change in amplitude and regularity of this oscillation mid-way through the series, which would probably provoke comment if this were a real paleoclimatic time series.

Figure 4(b) shows a linearly detrended time series, $\tilde{x}_t$, where

$$\tilde{x} = x_t - b_0 - b_1 t$$

and x_t is an integrated autoregressive process

$$x_t = x_{t-1} + y_t$$

$$y_t = \phi_1 y_{t-1} + a_t$$

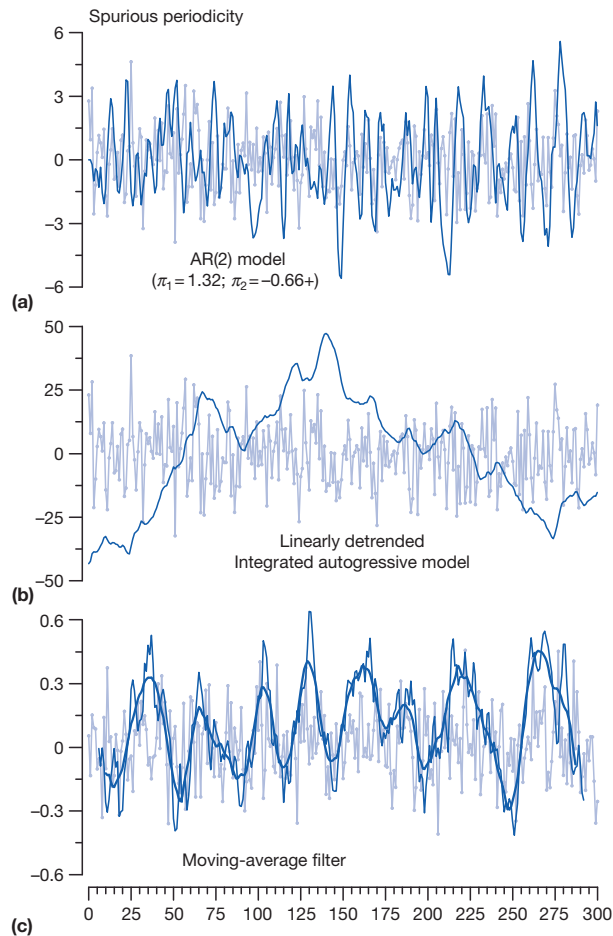


Figure 4 Some illustrations of spurious periodicity. Series plotted in pale blue are the white-noise time series used to generate the simulations in dark blue: (a) a second-order autoregressive process (AR(2)); (b) an improperly detrended integrated autoregressive model; and (c) running mean-filtered white noise (thick blue line represents the output from a 15-term running mean, thin blue line from a 10-term running mean of the data represented by the thick blue line).

with $\phi_1 = 0.8$ and a_t a normally distributed white-noise time series. The series x_t is an example of a ‘difference-stationary’ time series, or one in which the first differences of the series, $y_t = x_t - x_{t-1}$, are stationary or vary in a stable fashion over time. If treated instead as a ‘trend stationary’ time series, spurious periodicities can appear in the residual series, $\tilde{x}$, when x_t is linearly detrended (Nelson and Kang, 1981). The example series shows a broad cycle with a wavelength close to the record length and also shows lower-amplitude, higher-frequency variation, with a period around 20 time steps. Again, although generated by an aperiodic process, the resulting series could be viewed as being cyclic in nature. Paleoclimatic time series are frequently detrended as part of preliminary steps in data analysis.

Probably the most frequently applied method for generating apparently periodic variations in a time series when none really exist is illustrated in Figure 4(c), which demonstrates the Slutsky–Yule effect (von Storch and Zwiers, 2001). The dashed

line in Figure 4(c) is the result of applying a 15-term running mean to a white-noise time series:

$$y^t = \frac{1}{n} \sum_{i=1}^n w_i a_{t-i}$$

where $n = 15$, $w_1 = w_2 = \dots = w_n = 1$, and a_t is the white-noise time series. The solid line is simply the y_t ’s further smoothed by a 10-term running mean. The resulting series are clearly periodic and would likely be described or explained as ‘cyclic’ if they were real. Slutsky’s theorem (Jenkins and Watts, 1968) shows that by repeated application of summing (as in the running mean) or differencing filters, white noise can be reduced to a sine wave. In practice, smoothing like that done by the running mean can occur naturally in paleoclimatic records that integrate environmental conditions over time, but usually the smoothing occurs during data analysis.

All three series appear periodic yet are generated by relatively simple processes, without invoking any mechanism that could be intrinsically cyclical or periodic in nature. Because there are multiple ways of generating series with apparent periodicity, some of which are related directly to commonly applied data-analysis procedures, a relatively high standard should exist for declaring a particular series as cyclical and searching for explanations within the class of oscillatory or cyclical physical mechanisms.

Characteristic Kinds of Climate Variability

A Taxonomy of Typical Features in Climatic Time Series

An alternative ‘time domain’ approach for describing the nature of climate variability, other than that provided by the variance spectrum, is to describe the kinds of typical features or characteristic variations that recur in many climate climatic time series across a range of different timescales. There are a relatively small number of different kinds of variations that climatic time series display.

Table 1 gives examples of such typical variations, the levels in the climate system hierarchy that contain the controls of the variation, and those that are involved in the response. The table also indicates the predictability and reversibility of the individual kinds of variations. Climate variations are (a) predictable (P), if the timing and amplitude of the variations can be specified; (b) expectable (E), if either the timing or amplitude can be specified, but not both; or (c) unpredictable (U), if neither the timing nor amplitude of the variation can be specified. Variations are reversible (R), if the climate returns to a previous general mean state following the occurrence of the variation, or not reversible (N), if the climate does not return. The predictability and reversibility has implications for the biotic response to climate variations: Migration is the principal response to predictable and reversible climatic variations, while evolution (and extinction) is that for unpredictable and irreversible variations (Bartlein and Prentice, 1989; and Bennett, 1997). The table also shows the relative importance of trend, systematic (or periodic), and irregular components.

Table 1 Characteristic features or kinds of climatic variations

Kinds of variations	Examples	Levels in hierarchy ^A		Predictability/ reversibility ^B		Components of temporal variations ^C		
		Control	Response	Predict.	Reverse	Trend	System.	Irreg.
Trends								
Era-long	Cenozoic global cooling and glacierization	E	G–L	E	N	H	L	L
Glacial–interglacial	Last deglaciation	E,G	H–L	P	R	H	L	L
Multi-millennial	Late Holocene summer cooling in northern mid-latitudes	G–C	R–L	E	R	L	L	H
Century-long	Global warming	G–C	R–L	E	R	L	L	H
Steps								
Inter-epoch	Onset of N. Hemisphere glaciation (~2.65 Ma)	E	G–L	U	N	H	L	L
Millennial	5.5 ka African Monsoon collapse	C	R–L	E	R	H	L	L
Oscillations								
Periodic (orbital)	Ice sheet growth and decay, monsoon strengthening/weakening	E	G–L	P	R	L	H	L
Changing-periodic (orbital)	Strengthening of 100-ka cycle in past million years	E	G–L	U	N	L	H	L
Quasiperiodic (suborbital)	Dansgaard–Oeschger ‘cycles’ (75–10 ka)	E,G	H–L	E	R	L	M	H
Interannual (10 ⁰ –10 ¹ yrs)	ENSO variations, decadal-scale climate anomalies	H,C	R–L	E	R	L	M	M
Fluctuations								
Holocene (10 ³ –10 ⁴ years)	Holocene temperature and precipitation variations	G,H	C–L	E	R	M	L	H
Submillennial (10 ³ –10 ⁴ years)	Medieval warm period/Little Ice Age variations	G–C	R–L	E	R	L	L	H
Interannual (10 ⁰ –10 ¹ years)	Aperiodic interannual climate variations	H,C	R–L	E	R	L	L	H
Events								
Excursions (10 ⁵ –10 ⁶ years)	Late Paleocene Thermal Maximum	G	C–L	U	R	H	L	L
Climate reversals (10 ² –10 ⁴ years)	Younger Dryas climate reversal (~12.8–11.6 ka)	G–C	R–L	E	R	H	L	L
“Abrupt changes” (10 ¹ –10 ² years)	8.2 ka event	G–C	R–L	E	R	L	L	H
Interannual (10 ⁰ –10 ¹ years)	Pinatubo eruption global cooling	G	R–L	E	R	L	L	H

^aLevels in climate system hierarchy (E, external; G, global; H, hemispheric; C, continental; R, regional; L, local) controlling (Control) characteristic features of climatic variation or responding (Response) to them.

^bPredictability (Predict.) and reversibility (Reverse) of characteristic features of climatic variations: P, feature is predictable at the temporal or spatial scale indicated; E, feature is expectable, but not specifically predictable; U, feature is unpredictable.

^cRelative amplitude (H, high; M, medium; L, low) of components of temporal variations (Trend, Systematic (periodic or otherwise predictable), Irregular).

Trends

Trends are progressive increases or decreases in the levels of a particular climate variable; they appear in paleoclimatic variations on all timescales, but are particularly evident at certain scales. At the longest timescales displayed in [Figure 1\(a\)](#), cooling during the Cenozoic is evident in the general change in oxygen-isotopic ratios toward heavier values (i.e., toward cooler oceans, more ice). This trend is probably driven by the external controls of the climate system, with all the lower levels in the hierarchy responding. The general trend is broken, of course, by local increases in global temperature (as in the Eocene, ~50 Ma), and by locally more rapid decreases, but the overall impression when viewing this series is of a progressive

movement toward cooler conditions. The cooling trend is therefore expected once underway (but its particular amplitude is not predictable), nonreversing overall, and dominated by the long-term changes ([Table 1](#)). [Figure 1\(b\)](#), oxygen-isotope data for the past 5 My, also shows a generalized trend, which is the same climatic change as the more rapid decrease in global temperatures that is evident in the last part of Series A. On a shorter timescale ([Figure 2\(c\)](#)), the general transition between the Last Glacial Maximum (about 21000 years ago) and the present can be viewed as a trend (predictable and reversible), although broken by the Younger Dryas climate reversal (between 12.8 and 11.6 ka). During the Holocene ([Figure 1\(d\)](#)), many locations in the northern mid-latitudes experienced a cooling trend

in summer, or in the northern tropics and subtropics, a reduction in the strength of the monsoon. Finally, during the last century (Figure 1(d)), the global mean temperature, as well as that at individual stations, has generally increased.

Steps

Steps are the abrupt (relative to the timescale of variations under consideration) transitions from one level to another, and appear on many different timescales. Notable examples of steps in the representative series include the unpredictable and nonreversing interepoch decreases in global temperature during the Cenozoic (Figure 1(a)). Similarly, abrupt steps occur at the beginning and end of the Younger Dryas climate reversal and at the terminations of the 'Dansgaard-Oeschger' cycles during the interval between 60 and 20 ka (Figure 1(c)). These steps are expected, but not predictable, and have little irregular variation superimposed on the transition from one level to another.

Oscillations

Oscillations are either periodic or quasiperiodic variations about a stationary or slowly changing level and are one of the more prominent features of paleoclimatic time series. Oscillations dominate the glacial-interglacial variations in global temperature, and the variations in the continent-ocean temperature contrast that govern the monsoons, each generated by the periodic variations of the Earth's orbital elements (Figure 1(b)) (Imbrie et al., 1984; Prell and Kutzbach, 1992). Consequently, these variations are highly predictable and systematic. One important characteristic of the oxygen-isotope variations apparent in Figure 1(b) is the change in the relative importance of the different periodic components. Prior to 1 Ma, the 41-ka cycle is relatively more important, whereas thereafter, the 100-ka cycle becomes more prominent. These changes in periodicity seem inherently unpredictable and nonreversing. Quasiperiodic variations (i.e., variations that are less regular than the strictly periodic ones of the stacked global record) appear at 'suborbital' timescales, as in the aforementioned Dansgaard-Oeschger cycles. Interannual climate variations such as those constituting the El Niño/Southern Oscillation (ENSO) variations (Diaz and Markgraf, 2000) have also been described as quasiperiodic. Although systematic variation is apparent in these series, they are only expected owing to the absence of strict periodicity.

Fluctuations

Fluctuations are aperiodic variations of climate appearing at all timescales, but they tend to be more evident at shorter timescales. Variations in Northern Hemisphere temperature are dominated by interannual fluctuations, with decadal- and centennial-scale variations of less importance, with the exception of the past century. Fluctuations are expected, but not specifically predictable features in individual time series, and are distinguished from steps by their inherent reversibility.

Events

Events are similar in nature to steps, but are variations that return rapidly (relative to the timescale of variations under consideration) to a previous state, and hence are always reversible and distinct from steps. Examples of events are the late Paleocene Thermal Maximum (the distinct oxygen-isotope

peak at 55 Ma visible in Figure 1(a)), the Younger Dryas climate reversal, and 8.2-ka 'event' evident in the GISP 2 record (Figure 1(c)). What constitutes an event, as opposed to one of many fluctuations in a record is subjective, but generally events are both large and isolated, relative to nearby or background fluctuations.

At any given time or place, the prevailing climate is the product of the superimposition of all these features as they occur across all the different timescales. The efficacy of a particular feature in producing a response in some biotic or abiotic environment that is encoded in the paleoenvironmental record probably depends on the amplitude of the associated climatic variation and its duration, the number of features superimposed, the tendency for a particular variation to be predictable or reversible, and also on the existence of intrinsic thresholds.

Summary

The mean state of climate (or of the climate system) varies considerably, and these variations are expressed across a range of timescales. Moreover, the variability of the mean-state variations also varies over time. Despite this situation, there is considerable structure to the variability of climate. When viewed in the frequency domain, most of the variability of climate is related to the general increase in variance with increasing timescale, but there are several preferred timescales of variability related to the influence of external controls of climate and to internal mechanisms that generate quasiperiodic variations. The preferred timescales or periods related to external controls are those of 10^6 years and longer, which are related to tectonic controls, 10^4 – 10^6 years related to variations of the Earth's orbit, and the annual cycle, while those related to interval variations occur on interannual-to-millennial scales.

When viewed in the time domain, a small number of characteristic kinds of variation can be seen that recur across the full range of timescales. These typical variations include trends, steps, oscillations, fluctuations and events, and the superimposition of these generate climatic time series that show little stasis, or intervals of low variability over time.

See also: **Glaciation, Causes:** Astronomical Theory of Paleoclimates. **Paleoclimate:** Introduction. **Paleoclimate Modeling:** Quaternary Environments. **Paleoclimate Reconstruction:** Approaches.

References

- Alley RB, Anandakrishnan S, and Jung P (2001) Stochastic resonance in the North Atlantic. *Paleoceanography* 16: 190–198.
- Alley RB, Meese DA, Shuman CA, et al. (1993) Abrupt increase in Greenland snow accumulation at the end of the Younger Dryas event. *Nature* 362: 527–529.
- Bartlein PJ (1997) Past environmental changes: Characteristic features of Quaternary climate variations. In: Huntley B, Cramer W, Morgan A, Prentice HC, and Allen JRM (eds.) *Past and Future Rapid Environmental Change*, pp. 11–29. Berlin: Springer.
- Bartlein PJ and Prentice IC (1989) Orbital variations, climate and paleoecology. *Trends in Ecology & Evolution* 4: 195–199.

- Bennett KD (1997) *Evolution and Ecology: The Pace of Life*. Cambridge: Cambridge University Press.
- Bond G, Evans MN, Showers W, et al. (2001) Persistent solar influence on North Atlantic climate during the Holocene. *Science* 294: 2130–2136.
- Box GEP and Jenkins GM (1976) *Time Series Analysis: Forecasting and Control*. San Francisco, CA: Holden-Day.
- Diaz HF and Markgraf V (2000) *El Niño and the Southern Oscillation: Multiscale Variability and Global and Regional Impacts*. Cambridge: Cambridge University Press.
- Frakes LA, Francis JE, and Syktus JI (1992) *Climate Modes of the Phanerozoic: The History of the Earth's Climate over the Past 600 Million Years*. Cambridge: Cambridge University Press.
- Hansen J, Ruedy R, Sato M, and Lo k (2006) *GISS Surface Temperature Analysis, Global Temperature Trends: 2005 Summation*. Goddard Institute for Space Sciences, National Aeronautics and Space Administration 2006. <http://data.giss.nasa.gov/gistemp/2005/> (cited 20 February 2006).
- Hays JD, Imbrie J, and Shackleton N (1976) Variations in the earth's orbit: Pacemaker of the ice age. *Science* 194: 1121–1132.
- Imbrie J, Hays JD, Martinson DG, et al. (1984) The orbital theory of Pleistocene climate: Support from a revised chronology of the marine $\delta^{18}\text{O}$ record. In: Berger A, Imbrie J, Hays J, Kukla G, and Saltzman B (eds.) *Milankovitch and Climate*, pp. 269–305. Dordrecht: Reidel.
- Jenkins GM and Watts DG (1968) *Spectral Analysis and its Applications*, p. 297. San Francisco, CA: Holden-Day.
- Kantz H and Schreiber T (2004) *Nonlinear Time Series Analysis*. Cambridge: Cambridge University Press.
- Kutzbach JE and Bryson RA (1974) Variance spectrum of Holocene climatic fluctuations in the North Atlantic sector. *Journal of the Atmospheric Sciences* 31: 1958–1963.
- Lisiecki LE and Raymo ME (2005) A Pliocene-Pleistocene stack of 57 globally distributed benthic $\delta^{18}\text{O}$ records. *Paleoceanography* 20: PA1003.
- Mann ME and Jones PD (2003) Global surface temperatures over the past two millennia. *Geophysical Research Letters* 30: 1820–1824.
- Mayewski PA, Karlén W, Maasch KA, et al. (2004) Holocene climate variability. *Quaternary Research* 62: 243–255.
- McDowell PF, Webb T III, and Bartlein PJ (1991) Long-term environmental change. In: Turner BL II, Clark WC, Kates RW, Richards JF, Mathews JT, and Mayer WB (eds.) *The Earth as Transformed by Human Action: Global and Regional Changes in the Biosphere over the Past 300 Years*, pp. 143–162. Cambridge: Cambridge University Press.
- Mitchell JM Jr. (1976) An overview of climatic variability and its causal mechanisms. *Quaternary Research* 6: 481–493.
- Moberg A, Sonechkin DM, Holmgren K, Datsenko NM, and Karlen W (2005) Highly variable Northern Hemisphere temperatures reconstructed from low- and high-resolution proxy data. *Nature* 433: 613–617.
- Nelson CR and Kang H (1981) Spurious periodicity in inappropriately detrended time series. *Econometrica* 49: 741–751.
- Overpeck J and Webb R (2000) Nonglacial rapid climate events: Past and future. *PNAS* 97: 1335–1338.
- Prell WL and Kutzbach JE (1992) Sensitivity of the Indian monsoon to forcing parameters and implications for its evolution. *Nature* 360: 647–652.
- Rhode RA and Muller RA (2005) Cycles in fossil diversity. *Nature* 434: 208–210.
- Saltzman B (2002) *Dynamical Paleoclimatology: Generalized Theory of Global Climate Change*. San Diego, CA: Academic Press.
- Shackleton NJ and Imbrie J (1990) The $\delta^{18}\text{O}$ spectrum of oceanic deep water over a five-decade band. *Climatic Change* 16: 217–230.
- Stuiver M, Braziunas TF, Grootes PM, and Zielinski GA (1997) Is there evidence for solar forcing of climate in the GISP2 oxygen isotope record? *Quaternary Research* 48: 259–266.
- Torrence C and Compo GP (1998) A practical guide to wavelet analysis. *Bulletin of the American Meteorological Society* 79: 61–78.
- von Storch H and Zwiers FW (2001) *Statistical Analysis in Climate Research*. Cambridge: Cambridge University Press.
- Wang YJ, Cheng H, Edwards RL, et al. (2005) The Holocene Asian monsoon: Links to solar changes and North Atlantic climate. *Science* 308: 854–857.
- Webb T III (1989) The Spectrum of Temporal Climatic Variability: Current Estimates and the Need for Global and Regional Time Series. In: Bradley RS (ed.) *Global Changes of the Past*, pp. 61–81. Boulder, CO: UCAR/Office for Interdisciplinary Earth Studies.
- Wilson JT (1966) Did the Atlantic close and then re-open? *Nature* 211: 676–681.
- Wunsch C (2003) The spectral description of climate change including the 100 ky energy. *Climate Dynamics* 20: 353–363.
- Zachos J, Pagani H, Sloan L, Thomas E, and Billups K (2001) Trends, rhythms, and aberrations in global climate 65 Ma to present. *Science* 292: 686–693.

Modern Analog Approaches in Paleoclimatology

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Introduction

Quaternary paleoclimatic reconstructions of the last 21 000 years provide a long and detailed perspective on climatic variability, which in turn, enables a better understanding of the causes and mechanisms of climatic change (Bartlein et al., 1998; COHMAP Members, 1988). Synoptic climatological analyses provide useful information for understanding these causes and mechanisms as they relate to paleoclimatic evidence within a particular region (Barry and Perry, 1973; Barry, 1981; Hirschboeck et al., 1996; Serreze and Barry, 2005). Prior to the mid-1970s, most analyses of synoptic climatology of the past relied on the modern synoptic analog approach of past climates because this was the primary approach for reconstructing past atmospheric circulation. The modern analog approach is based on the uniformitarianism principle (Barry and Perry, 1973). It assumes that the range of modern conditions contains extreme events that may have been more frequent in the past and that the processes work in the same way during both times. However, a primary drawback to the modern analog approach is that no perfect analog for past climates prior to instrumental records exists, especially for longer Quaternary timescales, because external climatic controls such as ice sheet size, incoming insolation, and carbon dioxide concentrations differed dramatically when compared with today's conditions (e.g., Mitchell, 1990). The advent of GCM (general circulation model) simulations of past climates during the 1970s became a much more popular method of examining synoptic paleoclimatic patterns, as they provide a theoretical basis of how the climate system may have operated in the past.

However, GCMs have limitations in spatial resolution (usually about 1.5° longitude $\times$ 1.5° latitude of resolution) and in how they portray topography (Kohfeld and Harrison, 2000; McGuffie and Henderson-Sellers, 2005). RCMs (regional climate models) provide better resolution and more accurate topography but still greatly amplify it in some areas (e.g., Diffenbaugh and Sloan, 2004; Hostetler et al., 1994, 2000). Furthermore, GCMs have made some advances in simulating details on decadal- and century-scale climatic changes during the last 1000 years and matching results with proxy data (e.g., Seager et al., 2009), but this has not been the case for RCMs in the last 600 years. Modern synoptic analogs can provide greater spatial resolution as a result of copious station networks, and they can be applied much more readily to specific paleoclimate problems than previous applications of climate modeling.

Unlike the goals of most studies prior to the 1970s, the goal of modern synoptic analogs today is not simply to reconstruct past atmospheric circulation, but instead to provide explanations on how processes in the atmospheric portion of the climate system work in order to explain paleoclimatic variations of the past. Recent development of reanalysis datasets

such as the NCEP-NCAR Global Reanalysis (Kalnay et al., 1995; Kistler et al., 2001), North American Regional Reanalysis (Mesinger et al., 2006), and ECMWF-European Reanalysis (Uppala et al., 2005) provide new data resources for modern synoptic analog approaches. The reanalyzed data represent output from three-dimensional forecasting models initialized with observational data and interpolated onto a grid of similar global or regional resolution as GCMs and RCMs. Though simulated, reanalysis data are constrained (approximately every 6 h) by actual observations and therefore are regarded as a physically consistent depiction of past meteorological conditions. Output from such reanalyzed datasets often includes a broad suite of synoptic and dynamical meteorological fields (e.g., energy and water balance, atmospheric circulation), which can be used for modern synoptic analogs to understand climatic mechanisms associated with paleoclimate reconstructions (Shinker et al., 2006).

Modern synoptic analogs are obviously not perfect analogs but serve as partial analogs, or 'component' or 'process analogs' (Crowley, 1993; Webb et al., 1993). This article reviews the literature of modern synoptic analogs for paleoclimatic interpretation, discussing studies that involve longer (late Pleistocene to mid-Holocene) and shorter (AD 1000 to the present) timescales. It also discusses the drawbacks and assumptions of the analog approach and the changing nature of methodologies, and offers some suggestions for further research. Discussion is divided into two main parts: environment-to-circulation and circulation-to-environment approaches. These two approaches have been chosen because they represent the primary approaches in conventional synoptic climatological research (Barry and Carleton, 2001; Yarnal, 1993; Yarnal et al., 2001).

Environment-to-Circulation Approaches

The environment-to-circulation approach deals first with specifying certain conditions of the surface climate of interest in a particular area (e.g., extreme heavy rainfall), based on variations of modern instrumental data that are believed to resemble a particular climate in the past. The specification of these conditions is often selected according to a pattern of climatic anomalies as represented by a network of paleoclimatic proxy data. Based on the similar specified surface climate in the modern instrumental record, which represents a particular climatic extreme that may cover a month or a period of years, the corresponding atmospheric circulation is then examined in order to find the circulation patterns that may have occurred in the past.

Studies at Regional to Hemispheric Scales

One of the first systematic studies utilizing this approach for examining past climates was done by Simpson (1934), who constructed a hemispheric map of sea-level pressure for the late

Pleistocene, based on the distribution of abnormally extensive snow and ice. This approach was expanded by [Lamb \(1964\)](#), who constructed maps of winter and summer sea-level pressure for the late Pleistocene but used a more extensive dataset of snow and ice covering the period from 1920 to 1940. [Mather \(1954\)](#) also conducted a similar approach, using anomalies of sea-level pressure for the period 1931–40 as representative of full glacial climate because of generally lower mean annual temperatures in the Northern Hemisphere during this decade. As boundary conditions during the Last Glacial Maximum were dramatically different from what they are today, involving a large ice sheet at higher latitudes, lower sea-surface temperatures, lower concentrations of carbon dioxide, and different forcing conditions that involve the thermohaline circulation and sea ice ([COHMAP Members, 1988](#); [Otto-Bliesner et al., 2007](#)), the assumption that the present is a key to the past is invalid at the hemispheric as well as at the continental scale. For example, Mather's analysis implied that a northward and eastward shift of the Aleutian low would create drier conditions relative to today throughout the western United States, and this is inconsistent with implications from both proxy data and GCM simulations ([Otto-Bliesner et al., 2006](#); [Thompson et al., 1993](#)). Furthermore, the dating control as well as spatial coverage of proxy data used by Mather at the hemispheric scale is inadequate for conducting a detailed analog analysis.

[Barry \(1983\)](#), however, suggested that modern synoptic analogs may provide useful information regarding anomaly patterns of atmospheric circulation of past climates below the continental scale, given that sufficient proxy data are available for providing detailed surface conditions. [Nicholson \(1978, 1994\)](#) illustrated that this holds true at the spatial scale of central and northern Africa. She examined anomalies of precipitation from the nineteenth and twentieth century, comparing modern and historical composite anomalies with anomalies of the periods 20–15 and 10 ka reconstructed from proxy data. Patterns of the widespread reduced rainfall of the 1970s correspond to the climate anomalies of 20–15 ka, and the generally wet pattern for 1870–95 fits fairly well with that at 10 ka. Nicholson did not examine circulation data, but suggested that rainfall changes appear to be large scale in nature and that the characteristics of circulation associated with monsoonal flow in the present are similar to those in the past. The paleoclimatic and some twentieth century anomaly maps also show, interestingly, some opposite anomaly signs, particularly in East Africa and the western African tropics. Nicholson suggested that Saharan depressions may cause some variations that exhibit heterogeneous patterns.

[Edwards et al. \(2001\)](#) used a modern analog approach to examine how specific atmospheric configurations could have produced regional patterns of surface temperature and effective moisture during 14 000 cal year BP in interior Eastern Alaska seen in paleoclimate records but not captured by GCM simulations. They identified circulation anomalies during July 1964 as modern analogs to the cold and dry conditions that likely occurred in eastern Beringia until 14 000 cal year BP. They also constructed composites based on conditions in the instrumental record indicating a strong ridge leading to dry easterly flow and warm conditions as analogs to the early Holocene climate. Results were found to be consistent with proxy data networks (e.g., [Anderson et al., 2005](#)). [Edwards et al. \(2001\)](#) also

proposed a cold/wet pattern featuring enhancement of westerly flow into Alaska as a partial analog for the cold periods of the Little Ice Age. [Kokorowski et al. \(2008\)](#) described several synoptic controls associated with the position and strength of the Aleutian low and Siberian high leading to heterogeneous responses of Beringian climate during the Younger Dryas chronozone that help explain the contrasts between western Beringia and eastern Beringia, as well as topoclimatic variations within the region. However, their synoptic inferences were not based on any modern analog, as they believed that in this situation, an ideal close one could not be found.

Several scholars have utilized synoptic classification techniques commonly used in synoptic climatology in order to conduct their analog analyses. [Brinkmann and Barry \(1972\)](#) constructed 700-mb composite maps along with their associated precipitation, and temperature patterns, based on six extreme snowy winters and six extreme cool summers to determine the circulation patterns that explain the response of some proxy data in Keewatin and Labrador-Ungava, Canada. They assumed that the modern synoptic analogs may represent conditions conducive to glacial inception; with such modern conditions perhaps representative of conditions existing at least through the Holocene (cf. [Kaufman et al., 2004](#)). The 700-mb composite map for snowy winters shows a weaker-than-normal and westward migration of a trough in central Canada. Temperature anomalies for both Keewatin and Labrador-Ungava are positive. The 700-mb composite map for cool summers indicates a trough centered over eastern Hudson Bay, with dry conditions in Keewatin and wet conditions in Labrador-Ungava. The modern synoptic analogs suggest that vegetation changes at both locations would covary in response to summer temperature, but vegetation at both locations may respond differently as compared with one another in terms of precipitation.

[Antonsson et al. \(2008\)](#) used a synoptic classification scheme to identify groups of daily synoptic patterns to explain dry and warm mid-Holocene summer conditions in central Sweden. Their classification was based on daily mean sea-level pressure data from 1850 to 2003 using the Lamb–Jenkinson object classification for atmospheric circulation (also see [Lamb, 1950](#); [Jenkinson and Collison, 1977](#); [Chen and Li, 2004](#)). The dry and warm mid-Holocene summer climate in central Scandinavia likely occurred as a result of dominant summer anticyclonic circulation causing persistent blocking highs. Comparison of the reconstructed climate with simulations of modeled climate indicated that the modeled climate (predominantly driven by large-scale insolation patterns) underestimated the dry and warm period seen in the reconstruction; thus by assessing smaller-scale circulation analogs, [Antonsson et al. \(2008\)](#) were able to provide an alternative explanation for the regional summer climate patterns during the mid-Holocene.

[Diaz and Andrews \(1982\)](#) searched for a circulation regime for the warm mid-Holocene in northern North America by examining modern July temperature patterns derived from a principal component analysis and comparing the major modes of variability with patterns derived from a principal component analysis of fossil pollen data. [Diaz and Andrews \(1982\)](#) implied that the 700-mb circulation patterns for July 1961 and July 1972 are similar to the most important pattern portrayed by the pollen data, with a distinct ridge over mid-Canada, and

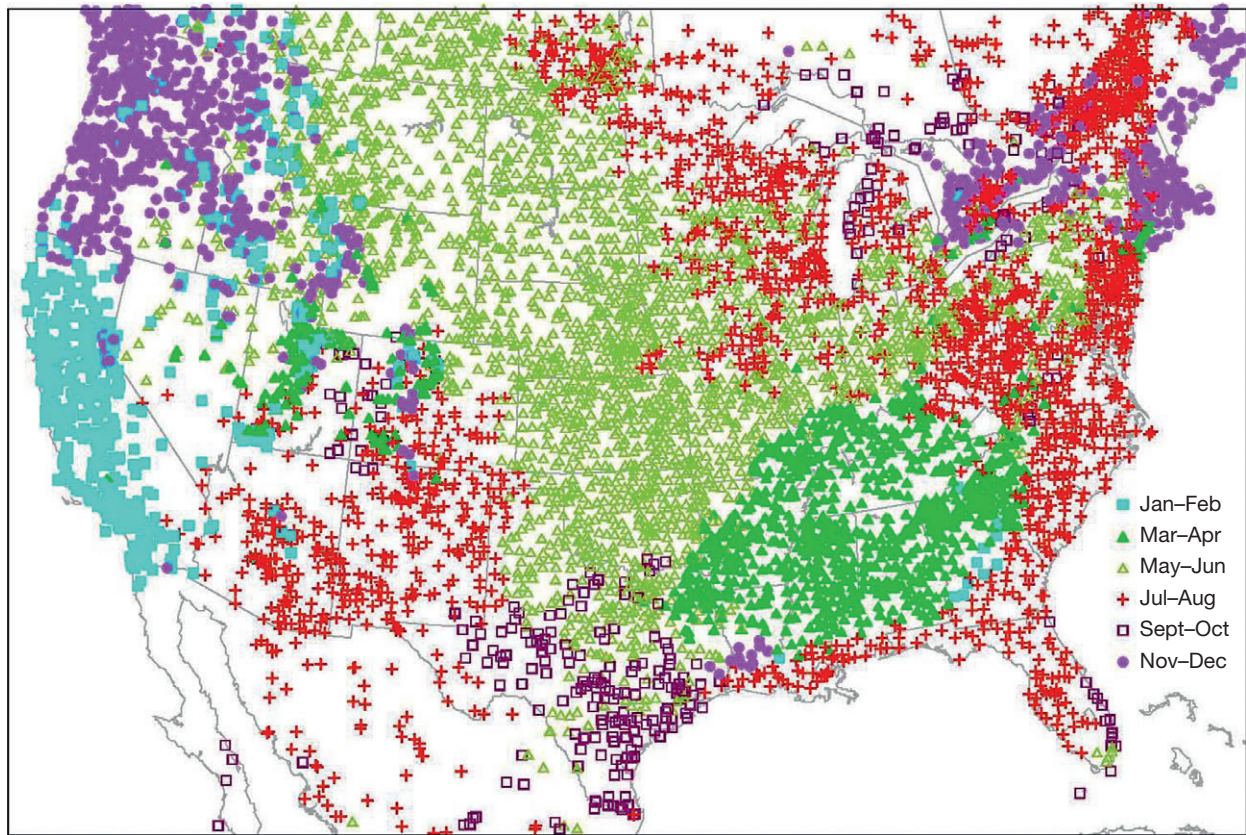


Figure 1 Distribution of the seasonal precipitation maximum. Refer to [Mock \(1996\)](#) regarding the timeframe of climatic normals, which differ slightly with respect to different observational networks.

that such patterns may prove useful for comparing with GCM simulations. The use of the first few principal component analysis and the somewhat spatially sparse pollen network, however, prevent the analysis of possible patterns of spatial heterogeneity of climate at smaller spatial scales.

Only one study, by [Sanchez and Kutzbach \(1974\)](#), was found to deal with the application of an environment-to-circulation approach at the regional-to-continental scale for shorter timescales. The authors examined differences of annual precipitation and temperature for the period 1961–70 covering parts of the southern United States, Central America, and part of South America in order to study spatial climatic variations that occurred during the Little Ice Age. They chose 1961–70 because it was characterized with widespread negative temperature anomalies throughout most of the American tropics. Widespread decreases of precipitation were also evident, and accompanying circulation changes indicated a southward displacement of the general circulation. Proxy data for comparison, unfortunately, is extremely sparse, and no clear consensus currently exists to conclude that the American tropics were clearly characterized by widespread colder temperatures during the Little Ice Age.

Studies at Local to Regional Scales

Surface climate responds to a hierarchy of climatic controls, involving larger-scale controls such as the polar jet stream and

smaller-scale controls such as topography ([Barry, 1970, 1982; Mock, 1996; Shinker, 2010](#)). As one analyzes climatic controls below the regional scale, more smaller-scale controls become superimposed over larger ones, creating a multitude of unique climatic regimes that are consistent with a particular set of larger-scale controls. Within some regions at smaller spatial scales, climates may be largely topographically controlled, thus the modern climate shares some characteristics of climatic controls that may have operated in the past. Spatial heterogeneity of precipitation seasonality is demonstrated by their spatial patterns over the United States and adjacent areas ([Figures 1 and 2](#)). Homogeneous precipitation maxima are apparent, such as the progression of winter precipitation maxima over the far Western United States by mid-latitude cyclones from the North Pacific, the extent of the summer precipitation from monsoonal processes in the Southwest, and the spring precipitation maxima over the Southeast United States from spring cyclogenesis ([Mojzisek and Mock, 2009](#)) ([Figure 1](#)). Heterogeneous seasonality patterns are most apparent in regions of diverse terrain such as the Rocky Mountains ([Mock, 1996; Shinker, 2010; Shinker and Bartlein, 2010](#)) and some places along water bodies that include the Great Lakes. Interestingly, seasonal precipitation minima over the United States generally exhibit more homogeneous patterns ([Figure 2](#)), which need to be assessed differently than precipitation maxima when also analyzing paleoclimate patterns. Illustrating several examples, winter precipitation minima are evident

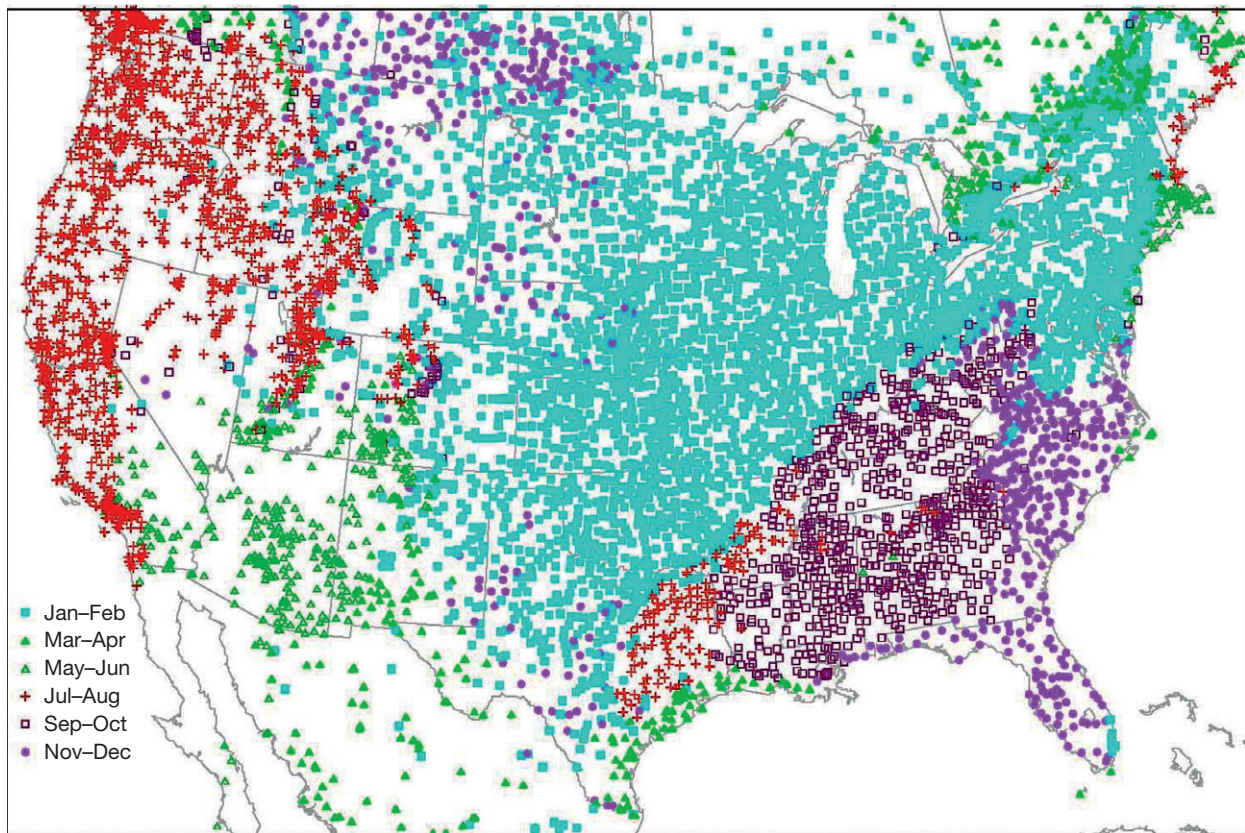


Figure 2 Distribution of the seasonal precipitation minimum. Refer to [Mock \(1996\)](#) regarding the timeframe of climatic normals, which differ slightly with respect to different observational networks.

over a large part of the eastern United States due to the absence of convection, a May–June minima over the Southwest, indicative of the circulation regime prior to monsoon onset, and summer minima over the West due to dominance of the Pacific subtropical high.

Modern synoptic analogs can provide valuable information regarding heterogeneous surface climatic responses to specific regional synoptic patterns in areas characterized by complex terrain. For example, [Mock and Brunelle-Daines \(1999\)](#) used a modern analog approach selecting August 1955 as a summer climate analog to describe the likely dominant circulation control leading to widespread drought over parts of the West and mid-continent, and enhanced monsoonal flow in the southwestern United States interpreted from proxy records during 6 ka. Modern synoptic analogs can also be conducted for localized subregions as well as for individual locations, but caution must be exercised in the selection of modern extremes in order to compare them with past climates. For example, abnormally wet conditions in the Great Basin of the western United States can arise from three different synoptic patterns: winter mid-latitude cyclones from the Pacific, cyclogenesis in the northeastern Great Basin, and monsoonal moisture from the Gulf of California ([Kay, 1982](#)). Thus, the investigator must know, specifically, what types of synoptic processes are of interest when formulating an environment-to-circulation analog approach.

[Streten \(1974\)](#) attempted to find a synoptic pattern that explained the dry and warm summers in interior Alaska that was believed to be representative of the last glaciation. From temperature and precipitation data, he examined 700-mb circulation during such modern extremes and noted that they correspond with a strong upper ridge over the region, with a deep low pressure system to the west. [Bowling \(1979\)](#) conducted a similar approach, but based the surface conditions from biological and geomorphological evidence, which showed wetter-than-normal conditions along the southern Alaska coast, and drier- and warmer-than-normal conditions inland. She implied that the winter circulation of 1976–77, characterized by increased southwesterly flow from the Gulf of Alaska, would be a synoptic pattern that explains the contrast of climatic anomalies along the coast and inland. It is similar to the pattern as implied by [Streten \(1974\)](#), thus such a pattern may have been more persistent during the Last Glacial Maximum during the course of the annual cycle.

Some analog studies specified a different type of surface environment other than temperature and precipitation to directly study synoptic climatic relationships with particular surface processes that may have operated more frequently in the past. [Enzel et al. \(1989, 1990\)](#) examined modern winter sea-level circulation patterns that may explain the existence of ephemeral lakes in the Mojave Desert, California. They constructed a sea-level composite-anomaly map on the basis of

8 months with lake-building flood events. Their map shows anomalous southwesterly flow with enhanced advection of warm, moist Pacific air into the region, associated with a deep cutoff low offshore and a blocking high in the Gulf of Alaska. Enzel et al. (1989, 1990) acknowledged that anomalously warm sea-surface temperatures may have affected circulation anomalies in the composite map, but implied that this general pattern was dominant prior to 9 ka and perhaps as far back as 18 ka. Ely et al. (1993) and Ely (1997) also examined modern flood analogs in Arizona and Utah. These studies classified three main types of flood analogs: the first caused by winter North Pacific frontal storms, the second caused by late-summer and fall storms associated with tropical storms and cutoff lows, and the third caused by localized summer convective thunderstorms. Composite maps of 700-mb heights were constructed for each analog type and compared to a 5000-year chronology of paleoflood events on the basis of flood deposits. Results from these studies suggest that long-term flood variability is associated with changes in large-scale circulation and cool, moist climate, including conditions involved in ENSO events, with flooding activity being the highest around 4800–3600 BP, around 1000 BP, and after 500 BP.

Van Heuklon (1977) used a similar analog approach to explain loess transport in Illinois during the last major Ice Age. He searched for daily events that were characterized by excessive dust transport for the region and then examined upper-level weather maps. Using weather maps from 13–15 October 1976, he implied that Alberta lows, coming from a northwesterly direction, may account for some fine-grained material to be transported in the region.

Several researchers have used an environment-to-circulation approach for studying atmospheric circulation to explain surface climate within the annual-to-historical (Little Ice Age) timescales. These analog approaches looked for a few ideal modern extreme events that best fit past climate. Bryson et al. (1974) examined precipitation patterns and associated circulation of the eastern Mediterranean for January 1955 and suggested that the occurrence of similar patterns may explain drought conditions in Mycenaean Greece around AD 1200. They also noted that the climate anomalies for January 1955 outside of southern Greece agree with climatic implications from pollen and historical evidence from the surrounding region. Weiss (1982) derived similar conclusions using a more dense station network and PDSI (Palmer Drought Severity Index) data, suggesting that January 1972 is also an accurate analog.

Newell (1992) examined modern synoptic analogs that correspond to summer 1816 temperature patterns in the Labrador Sea. He implied that months characterized with deep-low-pressure centers in the Labrador Sea, particularly June 1978 and June 1986, cause increased northwesterly winds and colder temperatures over the nearby land areas. Temperature anomalies for June 1978 and June 1986 were not as large as those for the summer of 1816, but Newell argued that the synoptic pattern for 1816 was most likely more intense. Stine (1994) implied that winter circulation of several dry years in the 1930s and for the winter of 1976–77 over California was more predominant during the Medieval Warm Period. He suggested that a contracted circumpolar vortex occurred in the past, but the winter synoptic analogs do not offer conclusive

proof, and all paleoclimatic evidence in mid-latitudes do not indicate widespread dryness (Hughes and Diaz, 1994). Mock (1991) specified patterns of summer freezing temperatures and heterogeneous but generally dry precipitation anomalies in the eastern Great Basin and adjacent Rocky Mountains during climatic extremes, and he used the associated circulation patterns from these extremes to study circulation patterns that may have prevailed during the summer of 1849. The analysis of the summer weather of 1849 was further studied by Blair and Rannie (1994), who searched for a larger-scale analog that fits the cold and dry summer as suggested by Mock (1991) and for an abnormally wet northern Great Plains. Both Mock (1991) and Blair and Rannie (1994) suggested that stronger westerlies and troughing over the western United States, with increased southwesterly flow at upper levels through the northern Plains, explain surface weather in the summer of 1849.

While most of the aforementioned studies focus on analogs of atmospheric pressure patterns to describe associated temperature and precipitation features of the past, Shinker et al. (2006) used atmospheric circulation variables (pressure and vertical motions) and energy- and water-balance variables from the NCEP-NCAR global reanalysis dataset (Kalnay et al., 1995; Kistler et al., 2001) to provide modern analogs for lake-level and moisture-balance trends from 11 000 to 8000 cal year BP. Their analysis focuses on the synoptic and dynamic controls of precipitation anomalies with an emphasis on the roles that large-scale vertical motions (omega) and surface feedbacks play in prolonged drought. One unique aspect of their modern synoptic analog approach was the interpretation of vertical motions at the 500 mb level via the variable omega, which is a secondary circulation feature embedded within ridges and troughs that provides information on sinking (within ridges) and rising (within troughs) motions that inevitably are the mechanisms that control the suppression or enhancement of precipitation, respectively. Their conceptual model explains the influence of atmospheric circulation on such precipitation mechanisms and regional moisture conditions in the present, in order to develop a more explicit understanding of the processes that can explain historic and prehistoric moisture anomalies. By illustrating the relationships between surface conditions, moisture availability, vertical motions, (omega) and surface water- and energy-balance components, they were able to demonstrate that dry conditions were not a result of anomalously strong zonal flow. Their results of modern analogs for mid-Holocene moisture patterns indicate that regional moisture flux and vertical motions can override the influence of broad circulation patterns.

This article presents two additional examples of modern synoptic analogs of past climates, using the NCEP-NCAR global reanalysis datasets, updating previous studies. Figure 3 illustrates August 1955, representing an analog for 6 ka as conducted by Mock and Brunelle-Daines (1999). The 500 mb geopotential height anomalies feature a strong upper-level anticyclone over the continent, likely creating widespread warmer-than-normal temperatures, and also being related to high specific humidity and increased monsoonal activity over the American Southwest (Figure 3). Two other centers of positive 500 mb height anomalies are indicative of a shifting long-wave pattern at the hemispheric scale. Heterogeneous patterns of omega at 500 mb are apparent, with strong sinking motions

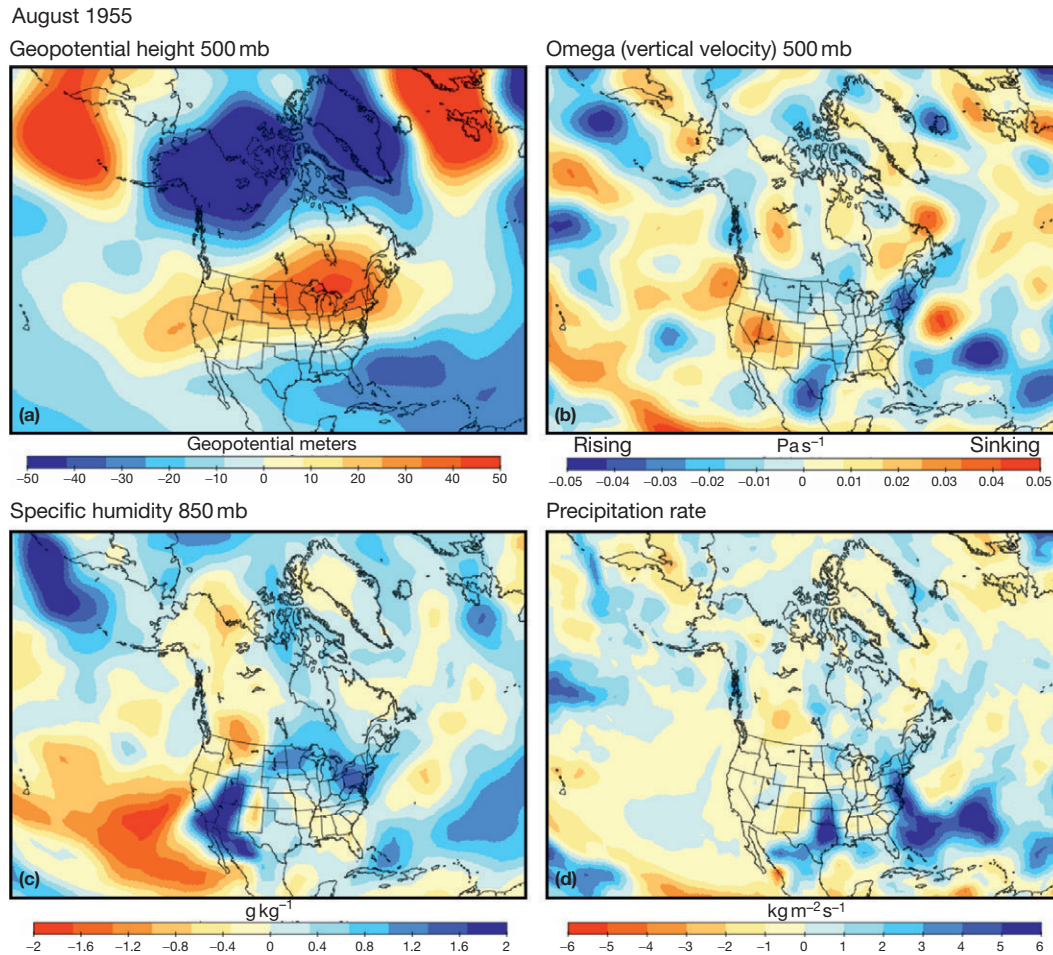


Figure 3 Composite-anomaly maps for August 1955, a modern analog for 6 ka. Geopotential heights at the 500 mb level (a), omega (vertical velocity) at the 500 mb level (b), specific humidity at the 850 mb level (c), and precipitation rate at the surface (d). Composite anomalies are based on the 1981–2010 base period.

over the Great Basin likely related to widespread dryness which is reflected from proxy data (Mock and Brunelle-Daines, 1999; also see Harrison et al., 2003). January 1957 was demonstrated by Mock and Bartlein (1995) to be a close modern analog of western US paleoclimate for the Last Glacial Maximum, but this analog should not be over generalized for adjacent regions. Mean sea-level pressure anomalies and omega patterns of sinking motion are indicative of a strong high off the Northwest coast, enabling anomalous dry easterly flow into the Pacific Northwest (Figure 4). Conversely, negative pressure anomalies, rising motion, and high relative humidity are indicative of wetter conditions in the American Southwest as a result of increased troughing of the polar jet stream.

Circulation-to-Environment Approaches

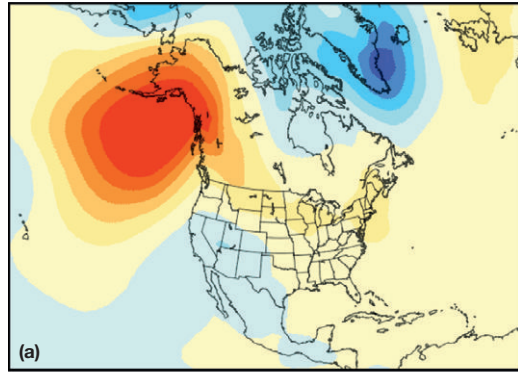
The circulation-to-environment approach is generally opposite of that of the environment-to-circulation approach. It is based on hypotheses of the prevailing atmospheric circulation that occurred during the period and in the study area of interest. Modern climatic extremes that most resemble these hypotheses

are selected as the modern synoptic analogs, the atmospheric circulation of these extremes are next examined, and then the corresponding surface climatic anomalies are analyzed as well as compared with patterns as suggested by paleoclimatic proxy data. The relationships between patterns of surface anomalies in the modern analogs and those from paleoclimatic data test the reliability of the modern analogs. The circulation-to-environment approach eliminates the problem of having to distinguish multiple circulation patterns that may explain surface paleoclimate, but it needs to have a robust hypothesis on the prevailing atmospheric circulation that may have predominated during the past. Such hypothesis can be taken from the output of GCMs or other conceptual models.

Most studies utilizing the circulation-to-environment approach deal with the examination of paleoclimates at longer timescales. Willett (1950) was one of the first to conduct such an approach. He examined circulation patterns with low (meridional) Rossby wave indices to construct a map of late Pleistocene Northern Hemisphere general circulation, assuming that low indices represent periods conducive for glacial advance. This approach was updated by Dzerdzevskii (1969), using circulation indices covering a longer timeframe and suggested

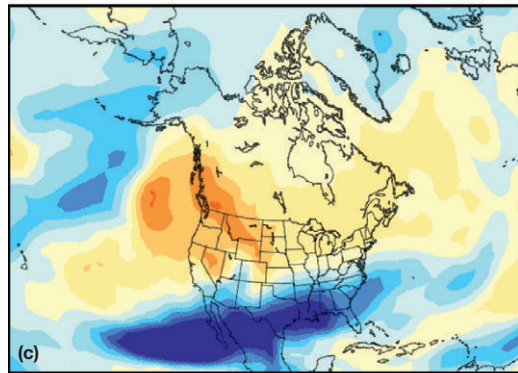
January 1957

Mean sea-level pressure

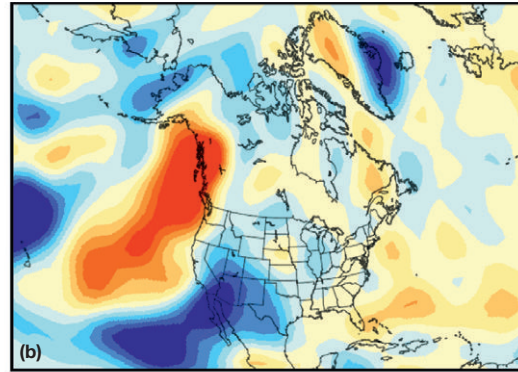


Geopotential meters

Specific humidity 850 mb

g kg⁻¹

Omega (vertical velocity) 500 mb

Rising Pa s⁻¹ Sinking

Precipitation rate

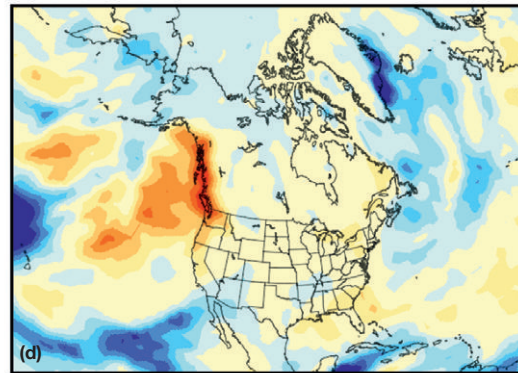
kg m⁻² s⁻¹

Figure 4 Composite-anomaly maps for January 1957, a modern analog for 18 ka. Mean sea-level pressure (a), omega (vertical velocity) at the 500 mb level (b), specific humidity at the 850 mb level (c), and precipitation rate at the surface (d). Composite anomalies are based on the 1981–2010 base period.

associated surface climatic anomalies during glacial and interglacial periods. As for the hemispheric studies discussed earlier, these reconstructions are not representative of late Pleistocene paleoclimate because of the external and internal boundary conditions being dramatically different from what they are today.

Some scholars have incorporated a similar approach in using hemispheric indices, but they have also examined relationships between hemispheric circulation and regional-scale surface climate. Butzer (1957) examined high (zonal) and low (meridional) indices of hemispheric circulation from 1920 to 1950 for the Mediterranean, European, and North Atlantic regions. Next, he compared regional sea-level pressure patterns and associated temperature and precipitation patterns in the region with paleoclimatic proxy data from the Mediterranean region. He suggested that periods with high indices represent paleoclimate from glacial periods and that low (meridional) indices represent the paleoclimate from interglacial periods. This suggestion is opposite of that from Willett (1950), as Butzer implied that a specific meridional circulation type may apply to the glacial paleoclimate for the region. Barry (1960, 1966) examined synoptic patterns of vapor-flux-

convergence and snowfall from the winters of 1956–57 and 1957–58 that might relate and explain some processes responsible for glacial inception for Labrador-Ungava, Canada. He found no single circulation type solely responsible, but suggested generally that low hemispheric indices lead to increased snowfall and that high indices lead to low precipitation. Mean total transport of streamlines was found to be primarily from the west, but the mean eddy component was found to be from the south. Barry's (1960, 1966) and Butzer's (1957) studies on hemispheric/regional scale relationships may suggest that modern synoptic analogs may not be reliable at the continental or larger spatial scales.

Several studies based the selection of modern circulation extremes at smaller spatial scales, assuming that the analogies of past circulation at the smaller spatial scale still hold true during the modern record. Derbyshire (1971) classified seven primary circulation patterns that most commonly occur over Australia. Associated temperature and precipitation anomalies with these circulation types were next examined. Derbyshire (1971) then selected two winters of a 'disturbed' westerly synoptic type that most closely corresponded to the spatial distribution of cirques, implying that this synoptic pattern prevailed

more frequently during the Last Glacial Maximum. [Alt \(1983\)](#) identified several main 500 mb patterns that explain glacial mass balance for the Queen Elizabeth Islands. She extrapolated these modern synoptic relationships to explain circulation characteristics during the last 5000 years. She implied that the year-to-year variability of circulation is larger than the variability in the long-term trends, but that this observation is valid only for the Queen Elizabeth Islands. Results suggest that the atmospheric circulation for 5 ka was characterized by a dominant anticyclonic circulation, which is conducive for warmer conditions.

Circulation-to-environment approaches have not typically dealt with the examination of a hierarchy of climatic controls at different spatial scales. [Cockcroft et al. \(1987\)](#) used an analog model from [Tyson \(1986\)](#), which explains present-day synoptic relationships between circulation and dry and wet spells in South Africa, in order to explain paleoclimatic changes in the region in the last 21 000 years. The model from [Tyson \(1986\)](#) used regional-scale zonal and meridional indices, and extremes of these indices were found to relate to dry and wet spells. [Cockcroft et al. \(1987\)](#) selected several individual climatic extremes and noted that winter and summer rainfall regimes tend to respond inversely with one another as a result of topographic influences on larger-scale circulation changes. These patterns of spatial heterogeneity corresponded with patterns of paleoclimatic proxy data for most of the late Pleistocene. [Mock and Bartlein \(1995\)](#) used a somewhat different approach in their investigation of synoptic patterns that may have caused patterns of spatial heterogeneity for the western United States during 18 ka and 9 ka. They based their hypotheses of atmospheric circulation controls from the output of GCMs and selected the modern synoptic analogs from regional circulation indices. Next, they examined composite maps of temperature and precipitation responses at the surface for the climatic extremes and compared these patterns with effective moisture maps as suggested by paleoclimatic proxy data. [Mock and Bartlein's \(1995\)](#) results clearly indicate that modern synoptic analogs are extremely useful for showing the additional role of smaller-scale climatic controls, such as low-level topographic gaps, with larger-scale controls in order to explain patterns of spatial heterogeneity that are clearly evident from spatial patterns of paleoclimatic proxy data.

[Mock et al. \(1998\)](#) used a synoptic classification scheme to identify modern atmospheric circulation patterns in Beringia for January and July sea-level pressure, and July 500 mb height anomalies and their associated temperature and precipitation patterns. Although these synoptic patterns are not specifically paleoclimate analogs, they note that spatial heterogeneity of responses likely occurred in the past as a result of variations in the position of trough and ridge axes, in addition to the role of topography, indicating the value of synoptic typing for understanding spatially variable regional temperature and/or precipitation patterns.

Only a few studies have utilized the circulation-to-environment analog approach at shorter (historical) timescales. [Baerreis and Bryson \(1968\)](#) chose several modern climatic extremes over the United States on the basis of the strength of the westerlies and constructed a composite-difference map of precipitation. They implied that a pattern of strong westerlies corresponds to a climate that prevailed

around AD 1200, with the summer of 1974 being a prime example ([Bryson, 1981](#)). Such implications may hold true for some areas such as Iowa where stronger westerlies generally lead to drought, which was recorded by a variety of paleoclimatic evidence ([Bryson and Murray, 1977](#)), but accurately dated paleoclimatic proxy data are not spatially plentiful enough for arriving at a conclusion that such a pattern is valid for the entire United States. The composite-difference map by [Baerreis and Bryson \(1968\)](#) show some smaller-scale climatic variations, such as generally drier conditions in the Snake River Plain, and such variations are not yet recorded by proxy data. [Wishman \(1994\)](#) conducted a circulation-to-environment approach of the Little Ice Age and Medieval Warm Period temperature patterns for western Norway. He based his hypotheses of atmospheric circulation from [Lamb \(1987\)](#), stating that the Little Ice Age was caused by a southward displacement of the Icelandic low and that the Medieval Warm Period was characterized by a northward displacement of the Azores high. [Wishman \(1994\)](#) used August 1979 and August 1976 circulation as representative of the Little Ice Age and Medieval Warm Periods, respectively. The surface temperature responses showed much greater decreases in the Little Ice Age at higher elevations than the lowland valleys, perhaps related to the abandonment of summer farms.

Discussion and Conclusion

It is clearly evident from the review of literature discussed earlier that the primary goal of modern synoptic analogs of past climates was once to reconstruct synoptic patterns of atmospheric circulation. Research on synoptic analogs of the past was clearly more active prior to the advent of GCMs. As GCMs now provide a theoretical means of doing the reconstruction over longer timescales, this is no longer the primary goal of modern synoptic analogs, although some implications of possible circulation anomaly patterns from the analogs at and below the regional level are still valuable. However, the reconstruction of atmospheric circulation from modern climate analogs is still an important goal at the shorter (Little Ice Age) timescales, for areas of very diverse terrain, and they may be useful as analogs of past extreme events ([Barriopedro et al., 2011](#)). For both longer and shorter timescales, the primary goal of modern synoptic analogs has shifted toward a process/dynamic approach of climatic forcing mechanisms that may have predominated in the past. These component or process analogs provide additional information that complement both GCMs and RCMs to understand relationships between synoptic-scale circulation patterns with climatic responses at the surface. The examination of climatic responses is also not limited to conventional temperature and precipitation, but can also include geomorphic and biological processes.

The following issues that concern the application of synoptic modern analogs for paleoclimatic assessment are clearly evident throughout the literature review; these are issues of 1) spatial scale of the circulation and surface environments, 2) spatial coverage of modern as well as of paleoclimatic data, and 3) the research design of the analog approach. The first issue, spatial scale of circulation and the surface environment, deals with the size of the study area of interest for searching a

modern synoptic analog. In terms of longer (Quaternary) timescales, continental-to-hemispheric scale analogs do not exist because of dramatic changes in surface and external boundary conditions in the past. Large ice sheets, greater magnitudes of oceanic currents, and much greater summer insolation, for example, cannot be found in any present-day climatic extreme. As the spatial scale decreases, more synoptic analogs generally become available. Nevertheless, one must be cautious of depicting too small of a surface environment in an analog approach because of the difficulties encountered in clearly depicting an environment-to-circulation approach that corresponds to the paleoclimate of interest. For example, one cannot choose high annual precipitation years at a few sites in parts of the western United States as analogs because of the multiplicity of circulation regimes that cause the same anomalies (Kay, 1982). A larger spatial scale would enable one to more clearly define the surface climate environment that may correspond to that in the past.

Adequate spatial coverage of both modern instrumental and paleoclimatic proxy data is crucial for conducting a synoptic analog approach. As a result of the hierarchy of climatic controls at different spatial scales, modern instrumental and paleoclimatic proxy data must be spatially copious to depict areas of spatial heterogeneity. Problems concerning some paleoclimatic proxy data also arise because of possible lags and other uncertain responses to climatic change (Mock and Bartlein, 1995), although new multiproxy methods are becoming more readily available for syntheses (e.g., Neukom et al., 2010). Thus, without a proper interpretation of paleoclimatic data, misleading patterns of spatial heterogeneity may result. Modern synoptic analogs conducted in a circulation-to-environment approach do not provide a means to reconstruct spatial patterns of paleoclimate; they only provide an understanding of processes that may have occurred in the past, and this requires a spatially detailed network of paleoclimatic proxy data. Researchers must also use the proper size grids of circulation data that represent all the synoptic features of interest (e.g., Pacific subtropical high; Siberian high) in their studies.

Neither the environment-to-circulation nor the circulation-to-environment approach shows any clear advantage; the choice depends on the timescale, region, and availability of data. The environment-to-circulation approach is largely inductive, being based heavily on how the surface environment is defined, and misinterpretations of surface data and poor theory can lead to erroneous implications of atmospheric circulation. The circulation-to-environment approach does not have the problem of defining a surface environment, but it is based on hypotheses of the prevailing circulation of the past, thus a clear framework of the climatic controls and the hypotheses must be formulated. The literature review has stated that generally the environment-to-circulation approach is more common in conducting synoptic analog studies. This is particularly true in terms of earlier studies dealing with longer timescales, because the primary goal was to reconstruct past atmospheric circulation. Studies dealing with shorter timescales still commonly employ such an approach because scholars are searching for circulation patterns that explain spatial variations of surface climatic proxy data responses. Longer timescale studies can use GCM output to conduct a circulation-to-environment approach; few similar studies

have been applied for shorter timescales. Yet, opportunities exist to conduct circulation-to-environment approaches for shorter timescales as better and more GCM simulations become available, as networks of proxy data improve spatially, and as a better understanding of hypotheses of climatic controls and synoptic patterns in the past for individual regions is achieved (Saenger et al., 2009; Palastanga et al., 2010; Luterbacher et al., 2010). Additionally, recent development of data sources (via reanalysis products) that have global and/or regional coverage representing surface and atmospheric variables are able to capture synoptic and dynamic processes and thus provide new opportunities for modern analog interpretations that focus on specific climatic mechanisms for understanding paleoenvironmental conditions (Shinker et al., 2006). Overall, both analog approaches, when exercised properly, make valuable contributions in interpreting past climates and will continue to do so as long as they have advantages over other methods in analyzing the hierarchy of climatic controls.

See also: [Introduction: History of Dating Methods.](#) **Paleoclimate:** [Introduction.](#) **Paleoclimate Modeling:** [Data–Model Comparisons;](#) [Last Glacial Maximum GCMs.](#) **Paleoclimate Reconstruction:** [The Last Millennium.](#)

References

- Alt BT (1983) Synoptic analogs: A technique for studying climatic change in the Canadian High Arctic. *Syllogeus* 49: 70–107.
- Anderson L, Abbott MB, Finney BP, and Burns SJ (2005) Regional atmospheric circulation changes in the North Pacific during the Holocene inferred from lacustrine carbonate oxygen isotopes, Yukon Territory, Canada. *Quaternary Research* 64: 21–35.
- Antonsson K, Chen D, and Seppä H (2008) Anticyclonic atmospheric circulation as an analog for the warm and dry mid-Holocene summer climate in central Scandinavia. *Climate of the Past* 4: 215–224.
- Baerreis DA and Bryson RA (1968) Climatic change and the Mill Creek culture of Iowa, Part I. *Journal of the Iowa Archaeological Society* 15: 1–34.
- Barriopedro D, Fischer EM, Luterbacher J, Trigo RM, and Garcia-Herrera R (2011) The Hot Summer of 2010: Redrawing the Temperature Record Map of Europe. *Science* 332: 220–224.
- Barry RG (1960) The application of synoptic studies in palaeoclimatology: A case study for Labrador-Ungava. *Geografiska Annaler* 42: 36–44.
- Barry RG (1966) Meteorological aspects of the glacial history of Labrador-Ungava with special reference to atmospheric vapor transport. *Geographical Bulletin (Canada)* 8: 319–340.
- Barry RG (1970) A framework for climatological research with particular reference to scale concepts. *Transactions of the Institute of British Geographers* 49: 61–70.
- Barry RG (1981) Atmospheric circulation and climatic change. I. Approaches to paleoclimatic reconstruction. In: Berger A (ed.) *Climatic Variations and Variability: Facts and Theories*, pp. 333–345. Dordrecht and Boston: D. Reidel Publishing Company.
- Barry RG (1982) Approaches to reconstructing the climate of the steppe-tundra biome. In: Hopkins DM, Matthews JV Jr., Schweger CE, and Young SB (eds.) *Paleoecology of Beringia*, pp. 195–204. New York: Academic Press.
- Barry RG (1983) Late Pleistocene climatology. In: Porter SC (ed.) *Late Quaternary Environments of the United States*, pp. 390–403. Minneapolis: University of Minnesota Press.
- Barry RG and Carleton AM (2001) *Synoptic and dynamic climatology*. London: Routledge.
- Barry RG and Perry AH (1973) *Synoptic Climatology*. London: Methuen.
- Bartlein PJ, Anderson KH, Anderson PM, et al. (1998) Paleoclimate simulations for North America over the past 21,000 years: Features of the simulated climate and comparison with paleoenvironmental data. *Quaternary Science Reviews* 17: 549–585.

- Blair D and Rannie WF (1994) 'Wading to Pembina': 1849 spring and summer weather in the valley of the Red River of the North and some climatic implications *Great Plains Research* 4: 3–26.
- Bowling SA (1979) Meteorological interpretation of ice age climate in Alaska. *Bulletin of the American Meteorological Society* 60: 845.
- Brinkmann WAR and Barry RG (1972) Palaeoclimatological aspects of the synoptic climatology of Keewatin, Northwest Territories. *Palaeogeography, Palaeoclimatology, Palaeoecology* 11: 77–91.
- Bryson RA (1981) Chinook climates and Plains peoples. *Great Plains Quarterly* 1: 5–15.
- Bryson RA, Lamb HH, and Donley DL (1974) Drought and the decline of Mycenae. *Antiquity* 48: 46–50.
- Bryson RA and Murray TJ (1977) *Climates of Hunger*. Madison: University of Wisconsin Press.
- Butzer KW (1957) The recent climatic fluctuation in lower latitudes and the general circulation of the Pleistocene. *Geografiska Annaler* 39: 105–113.
- Chen D and Li X (2004) Scale dependent relationship between maximum ice extent in the Baltic Sea and atmospheric circulation. *Global and Planetary Change* 41: 275–283.
- Cockcroft MJ, Wilkinson MJ, and Tyson PD (1987) The application of a present-day climatic model to the late Quaternary in Southern Africa. *Climatic Change* 10: 161–181.
- COHMAP Members (1988) Climatic changes of the last 18,000 years: Observations and model simulations. *Science* 241: 1043–1052.
- Crowley TJ (1993) Use and misuse of the geologic "analogs" concept. In: Eddy JA and Deschger H (eds.) *Global Changes in the Perspective of the Past*, pp. 17–27. New York: Wiley.
- Derbyshire E (1971) A synoptic approach to the atmospheric circulation of the last glacial maximum in southeastern Australia. *Palaeogeography, Palaeoclimatology, Palaeoecology* 10: 103–124.
- Diaz HF and Andrews JT (1982) Analysis of the spatial pattern of July temperature departures (1943–1972) over Canada and estimates of the 700 mb midsummer circulation during middle and late Holocene. *Journal of Climatology* 2: 251–265.
- Diffenbaugh NS and Sloan LC (2004) Mid-Holocene orbital forcing of regional-scale climate: A case study of western North America using a high-resolution RCM. *Journal of Climate* 17: 2927–2937.
- Dzerdzhevskii BL (1969) Climatic epochs in the twentieth century and some comments on the analysis of past climates. In: *Quaternary Geology and Climate*, pp. 49–60. Washington, DC: National Academy of Sciences.
- Edwards ME, Mock CJ, Finney BP, Barber VA, and Bartlein PJ (2001) Potential analogs for paleoclimatic variations in eastern interior Alaska during the past 14,000 yr: Atmospheric-circulation controls of regional temperature and moisture responses. *Quaternary Science Reviews* 20: 189–202.
- Ely LL (1997) Response of extreme floods in the southwestern United States to climatic variations in the late Holocene. *Geomorphology* 19: 175–201.
- Ely LL, Enzel Y, Baker VR, and Cayan DR (1993) A 5000-year record of extreme floods and climate change in the southwestern United States. *Science* 262: 410–412.
- Enzel Y, Cayan DR, Anderson RY, and Wells SG (1989) Atmospheric circulation during Holocene lake stands in the Mojave Desert: Evidence of regional climate change. *Nature* 341: 44–47.
- Enzel Y, Anderson RY, Brown WJ, Cayan DR, and Wells SG (1990) Tropical and subtropical moisture and southerly displaced North Pacific storm track: Factors in the growth of late Quaternary lakes in the Mojave Desert. In: Betancourt JL and MacKay AM (eds.) *Proceedings of the Sixth Annual Pacific Climate (PACLIM) Workshop*, pp. 135–139. Sacramento: California Department of Water Resources Interagency Ecological Studies Program Technical Report 23.
- Harrison SP, Kutzbach JE, Liu Z, et al. (2003) Mid-Holocene climates of the Americas: A dynamical response to changed seasonality. *Climate Dynamics* 20: 663–688.
- Hirschboeck KK, Ni F, Wood ML, and Woodhouse CA (1996) Synoptic dendroclimatology: Overview and prospectus. In: Dean JS, Meko DM, and Swetnam TW (eds.) *Tree Rings, Environment, and Humanity: Processes and Relationships* 1994: Tucson, Ariz., International Conference on Tree-Rings, Radiocarbon, pp. 205–223.
- Hostetler SW, Bartlein PJ, Clark PU, Small EE, and Solomon AM (2000) Stimulated influences of Lake Agassiz on the climate of central North America 11,000 years ago. *Nature* 405: 334–337.
- Hostetler SW, Giorgi F, Bates GT, and Bartlein PJ (1994) Lake-atmosphere feedbacks associated with paleolakes Bonneville and Lahontan. *Science* 263: 665–668.
- Hughes MK and Diaz HF (1994) *The Medieval Warm Period*. Dordrecht/Boston: Kluwer Academic.
- Jenkinson AF and Collison FP (1977) An initial climatology of gales over the North Sea, *Synoptic Climatology Branch Memorandum*, 62. Bracknell: Meteorological Office.
- Kalnay E, Kanamitsu M, Kistler R, et al. (1995) The NCEP/NCAR 40-year reanalysis project. *Bulletin of the American Meteorological Society* 77: 437–471.
- Kaufman DS, Ager TA, Anderson NJ, et al. (2004) Holocene thermal maximum in the western Arctic (0–180°W). *Quaternary Science Reviews* 23: 529–560.
- Kay PA (1982) A perspective on Great Basin paleoclimates. In: Madsen DB and O'Connell JF (eds.) *Man and Environment in the Great Basin*, pp. 76–81. Salt Lake City: Society for American Archaeology.
- Kistler R, Kalnay E, Collins W, et al. (2001) The NCEP-NCAR 50-year reanalysis: Monthly means CD-ROM and documentation. *Bulletin of the American Meteorological Society* 82: 247–267.
- Kohfeld KE and Harrison SP (2000) How well can we simulate past climates? Evaluating the models using global paleoenvironmental datasets. *Quaternary Science Reviews* 19: 321–346.
- Kokorowski HD, Anderson PM, Mock CJ, and Lozhkin AV (2008) A re-evaluation and spatial analysis of evidence for a Younger Dryas climatic reversal in Beringia. *Quaternary Science Reviews* 27: 1710–1722.
- Lamb HH (1950) Types and spells of weather around the year in the British Isles. *Quarterly Journal of the Royal Meteorological Society* 76: 393–438.
- Lamb HH (1964) Fundamentals of climate. In: Nairn AEM (ed.) *Problems in Palaeoclimatology*. New York: Interscience.
- Lamb HH (1987) What can historical records tell us about the breakdown of the Medieval warm climate in Europe in the fourteenth and fifteenth centuries – An experiment. *Beiträge zur Physik der Atmosphäre* 60: 131–144.
- Luterbacher J, Koenig SJ, Franke J, et al. (2010) Circulation dynamics and its influence on European and Mediterranean January–April climate over the past half millennium: Results and insights from instrumental data, documentary evidence and coupled climate models. *Climatic Change* 101: 201–234.
- Mather JR (1954) The present climatic fluctuation and its bearing on a reconstruction of the Pleistocene. *Tellus* 6: 287–301.
- McGuffie K and Henderson-Sellers A (2005) *A Climate Modelling Primer*. New York: Wiley.
- Mesinger F, DiMego G, Kalnay E, et al. (2006) North American regional reanalysis. *Bulletin of the American Meteorological Society* 87: 343–360.
- Mitchell JFB (1990) Greenhouse warming: Is the mid-Holocene a good analogue? *Journal of Climate* 3: 1177–1192.
- Mock CJ (1991) Historical evidence of a cold, dry summer in the northeastern Great Basin and adjacent Rocky Mountains during the summer of 1849. *Climatic Change* 18: 37–66.
- Mock CJ (1996) Climatic controls and spatial variations of precipitation in the Western United States. *Journal of Climate* 9: 1111–1124.
- Mock CJ and Bartlein PJ (1995) Spatial variability of late Quaternary paleoclimates in the western United States. *Quaternary Research* 44: 425–433.
- Mock CJ, Bartlein PJ, and Anderson PM (1998) Atmospheric circulation patterns and special climatic variations in Beringia. *International Journal of Climatology* 10: 1085–1104.
- Mock CJ and Brunelle-Daines AR (1999) A modern analogue of western United States summer palaeoclimate at 6000 years before present. *The Holocene* 9: 541–545.
- Mojzisek J and Mock CJ (2009) Historical changes to the Tennessee Precipitation Regime. In: Dupigny-Giroux LA and Mock CJ (eds.) *Historical Climate Variability and Impacts in North America*, pp. 23–34. New York/Dordrecht: Springer.
- Neukom R, Luterbacher J, Villalba R, et al. (2010) Multi-centennial summer and winter precipitation variability in southern South America. *Geophysical Research Letters* 37: L14708. <http://dx.doi.org/10.1029/2010GL043680>.
- Newell JP (1992) The climate of the Labrador Sea in the spring and summer of 1816, and comparisons with modern analogues. In: Harrington CR (ed.) *The Year Without a Summer*, pp. 245–254. Ottawa: Canadian Museum of Nature.
- Nicholson SE (1978) Climatic variations in the Sahel and other African regions during the past five centuries. *Journal of Arid Environments* 1: 3–24.
- Nicholson SE (1994) Recent rainfall fluctuations in Africa and their relationship to past conditions over the continent. *The Holocene* 4: 121–131.
- Otto-Bliesner BL, Brady EC, Clauzet G, Tomas R, Levis S, and Kothavala Z (2006) Last glacial maximum and Holocene climate in CCSM3. *Journal of Climate* 19: 2526–2544.
- Otto-Bliesner BL, Hewitt CD, Marchitto TM, et al. (2007) Last Glacial Maximum ocean thermohaline circulation: PMIP2model intercomparisons and data constraints. *Geophysical Research Letters* 34: L12706. <http://dx.doi.org/10.1029/2007GL029475>.
- Palastanga V, van der Schrier G, Weber SL, Kleinen T, Briffa KR, and Osborn TJ (2010) Atmosphere and ocean dynamics: Contributors to the European Little Ice Age? *Climate Dynamics* <http://dx.doi.org/10.1007/s00382-010-0751-0>.
- Saenger C, Chang P, Link J, Oppo DW, and Cohen AL (2009) Tropical Atlantic climate response to low-latitude and extratropical sea-surface temperature: A Little

- Ice Age perspective. *Geophysical Research Letters* 36: L11703. <http://dx.doi.org/10.1029/2009GL038677>.
- Sanchez WA and Kutzbach JE (1974) Climate of the American Tropics and Subtropics in the 1960s and possible comparisons with climatic variations of the last millennium. *Quaternary Research* 4: 128–135.
- Seager R, Ting M, Davis M, et al. (2009) Mexican drought: An observational, modeling and tree ring study of variability and climate change. *Atmosfera* 22: 1–31.
- Serreze MC and Barry RG (2005) *The Arctic Climate System*. Cambridge: Cambridge University Press.
- Shinker JJ (2010) Visualizing spatial heterogeneity of Western U.S. Climate Variability. *Earth Interactions* 14: 1–15. <http://dx.doi.org/10.1175/2010EI323.1>.
- Shinker JJ and Bartlein PJ (2010) Spatial variations of effective moisture in the western United States. *Geophysical Research Letters* 37: L02701. <http://dx.doi.org/10.1029/2009GL041387>.
- Shinker JJ, Bartlein PJ, and Shuman BN (2006) Synoptic and dynamic controls of North American mid-continental aridity. *Quaternary Science Reviews* 25: 1401–1417.
- Simpson GC (1934) World climate during the Quaternary Period. *Quarterly Journal of the Royal Meteorological Society* 60: 425–478.
- Stine S (1994) Extreme and persistent drought in California and Patagonia during medieval time. *Nature* 369: 546–549.
- Streten NA (1974) Some features of the summer climate of interior Alaska. *Arctic* 27: 273–286.
- Thompson RS, Whitlock C, Bartlein PJ, Harrison SP, and Spaulding WG (1993) Climatic changes in the western United States since 18,000 yr B.P. In: Wright HE Jr., Kutzbach JE, Webb T III, Ruddiman WF, Street-Perrott FA, and Bartlein PJ (eds.) *Global Climates since the Last Glacial Maximum*, pp. 468–513. Minneapolis: University of Minnesota Press.
- Tyson PD (1986) *Climatic Change and Variability in Southern Africa*. Cape Town: Oxford University Press.
- Uppala SM, Kallberg PW, Simmons AJ, et al. (2005) The ERA-40 re-analysis. *Quarterly Journal of the Royal Meteorological Society* 131: 2961–3012.
- Van Heuklon TK (1977) Distant source of 1976 dustfall in Illinois and Pleistocene weather models. *Geology* 5: 693–695.
- Webb T III, Crowley TJ, Frenzel B, et al. (1993) Group Report: Use of paleoclimatic data as analogs for understanding future global changes. In: Eddy JA and Oeschger H (eds.) *Global Changes in the Perspective of the Past*, pp. 51–71. New York: John Wiley & Sons.
- Weiss B (1982) The decline of Late Bronze Age Civilization as a possible response to climatic change. *Climatic Change* 4: 173–198.
- Willet HC (1950) The general circulation at the last (Würm) glacial maximum. *Geografiska Annaler* 32: 179–187.
- Wishman E (1994) A possible modern analogy to atmospheric circulation and climate in western Norway during the “Medieval Warm Epoch” and the “Little Ice Age Cold Epoch”. In: Frenzel B, Pfister C, and Gläser B (eds.) *Climate in Europe 1675–1715*, pp. 377–387. Stuttgart: Fischer Special Issue: ESF Project European Palaeoclimate and Man.
- Yarnal B (1993) *Synoptic Climatology in Environmental Analysis*. London: Belhaven Press.
- Yarnal B, Comrie AC, Frakes B, and Brown DP (2001) Developments and prospects in synoptic climatology. *International Journal of Climatology* 21: 1923–1950.

Paleoclimate History of the Arctic

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Early Cenozoic Warm Times

At the start of the Cenozoic, the planet was ice-free; there was no sea ice in the Arctic Ocean and there was no Greenland or Antarctic ice sheet. Benthic Foraminifera $\delta^{18}\text{O}$ document a long-term cooling of the global deep ocean during the Cenozoic (Zachos et al., 2001). Because oceanic bottom waters originate in the polar oceans, Arctic climate history is broadly consistent with the global deep-ocean record. Although general cooling and an increase in terrestrial ice volume characterize the Cenozoic, this first-order was punctuated by short- and longer-lived climate reversals, by variations in cooling rate, and by additional features related to growth and shrinkage of ice once ice sheets were well established. A detailed Arctic Ocean record that is equivalent to the global record is not yet available. Because the Arctic Ocean is connected to the global ocean only by a single narrow deep channel and some intermittently exposed shallow shelves, the Arctic Ocean $\delta^{18}\text{O}$ record often differs from that of the global ocean, hindering correlations. Emerging paleoclimate reconstructions from the Arctic Ocean derived from sediments recovered from the Lomonosov Ridge (Polyak et al., 2010) shed new light on the Cenozoic evolution of the Arctic Basin, but the data have yet to be fully integrated with evidence from high-latitude terrestrial records or with the other records from elsewhere in the Arctic Ocean. The presence of trace amounts of coarse clastic debris may indicate early onset of Arctic Ocean sea ice or calving land ice, but the evidence remains circumstantial and alternative explanations are possible.

General cooling through the Cenozoic is attributed mainly to a slow drawdown of greenhouse gases in the atmosphere through the weathering of silicic rocks that exceeded the release of stored carbon through volcanism and reprocessing. Many lines of proxy evidence show that atmospheric CO_2 was higher in the Cretaceous than it has been recently, and that it subsequently fell in parallel with the cooling. Changes in continental positions, sea level, and oceanic circulation as well as biological evolution also may have contributed, but are lesser factors.

Pliocene Warmth and the Transition into Quaternary Ice-Age Cycles

During the middle Pliocene (~3.5 Ma), forests occupied large regions near the Arctic Ocean that are currently polar deserts. Fossils of the marine bivalve *Arctica islandica*, which does not live where there is seasonal sea ice, found in marine deposits as young as 3.2 Ma old on Meighen Island at 80°N, likely record the peak Pliocene warmth of the ocean (Fyles et al., 1991). Multiple temperature proxies from terrestrial peat in the high Canadian Arctic indicate mean annual Pliocene temperatures 19 °C higher than present (Ballantyne et al., 2010); other sites have produced large temperature differences for the Pliocene of the North American Arctic, although slightly lower magnitudes (14 °C, Ellesmere Is., Ballantyne et al., 2006; beetles and plants indicative of summers as much as 10 °C warmer than recently measured temperatures and even larger wintertime warming to

a maximum of 15 °C or more, [Elias and Matthews, 2002](#)). Although continental configurations were similar to modern forests extended to the Arctic shoreline, nearly eliminating the Arctic tundra biome, and sea level stood ~25 m higher than present, implying much smaller Greenland (possibly limited to small mountain ice caps and glaciers) and Antarctic ice sheets (possibly without a West Antarctic Ice Sheet).

Explaining the warmth of the Arctic during the Pliocene remains enigmatic. The duration of warmth exceeds cycle-lengths related to orbital irregularities ($>10^5$ years), and continental configurations and gateways were similar to the present day. It is likely that the ocean transported more heat into the Arctic from the Atlantic during peak Pliocene warmth ([Haywood et al., 2009](#)), although the status of Arctic Ocean sea ice remains uncertain, as does the amount of ice on Greenland. Best estimates of atmospheric CO₂ loadings are similar to present (AD 2011) levels ([Pagani et al., 2010](#)), although most climate models driven by changes in greenhouse gases (mostly CO₂) tend to underestimate Arctic warmth (e.g., [Sloan and Barron, 1992](#)). A consensus view is that greater atmospheric and/or oceanic meridional heat transport, coupled with some combination of stronger greenhouse gas forcing and stronger long-term positive feedbacks than those found in many climate models, are required to explain the reconstructed temperatures for the Arctic during the warmest intervals of the Pliocene.

The Early Quaternary

The development of large continental ice sheets in the North American and Eurasian Arctic marks the onset of the Quaternary. The first secure evidence for the growth of these ice sheets is the appearance of ice-rafted debris (IRD) in the central North Atlantic between 3.0 and 2.5 Ma, far from shores where seasonal sea ice may transport sediment. Recently, the International Union of Geological Sciences has formalized a revision of the geological timescale that places the base of the Quaternary Period at 2.58 Ma, coincident with the Gauss–Matuyama paleomagnetic reversal ([Gibbard et al., 2010](#)), and closely aligned to the start of the ice-age cycles. The full Quaternary is best represented in deep-sea sediment, where the changes in $\delta^{18}\text{O}$ of benthic Foraminifera are used to subdivide the Quaternary into 104 numbered marine isotopic stages (MIS), counting consecutively back from the present ([Lisiecki and Raymo, 2005](#)), with odd-numbered stages representing global warm times/minimum ice volumes, and even-numbered stages representing peak ice volumes.

The appearance of large, quasistable continental ice sheets coincides approximately with the start of the Quaternary, but general cooling through the Cenozoic is reflected by the appearance of smaller glaciers and ice caps much earlier, during the Pliocene and possibly earlier. Although the pattern of glacial onset over Greenland remains uncertain, small amounts of IRD are found in marine sediment recovered from the continental shelf around Greenland as old as 7 Ma ([St. John and Krissek, 2002](#)). IRD in proximal marine sediment, but not in the central Atlantic, likely indicates that climate cooled sufficiently during the Pliocene and possibly even as early as the late Miocene, to allow small ice caps and glaciers to form at high elevations on Greenland. Outlet glaciers delivering small

icebergs to the sea during the coldest intervals likely explains the observed IRD, but the total ice volume on Greenland may have been very small. Indeed, the first significant appearance of likely Greenland-derived IRD in the open ocean is observed at ~3.2 Ma ([Kleiven et al., 2002](#)), which may represent the inception of a Greenland Ice Sheet, although the geographic distribution suggests Scandinavia and North America are also potential sources. Widespread IRD in the North Atlantic begins about 2.7 Ma ([Kleiven et al., 2002](#)), at about the onset of the Quaternary.

The best terrestrial record of the early phases of late Cenozoic ice growth comes from Iceland, where unusually high volcanic productivity, caused by the superposition of mantle plume and mid-ocean ridge magmatism, resulted in an environment where constructive geologic processes exceed destructive processes. This unusual state results in the unique preservation of glacial deposits in Iceland extending well back into the Pliocene. Because lava flows can only occur if the landscape is ice-free, the general consensus is that such flows represent interglacials, interstadials, or other ice-free periods, whereas hyaloclastites (subglacially formed volcanics) and till represent glaciations. Consequently, a record of past glaciations (often including ice-flow features) and interglaciations is preserved within stacked volcanic sequences. Mapping and dating of remnants of glacial deposits found imbedded within the lava flows of Iceland have revealed at least 22 glacial–interglacial cycles during the last ~3 Ma ([Geirsdóttir et al., 2011](#)). Four phases of glaciation have been identified: (1) The inception phase, between >4 and 3 Ma, when ice caps were concentrated over the mountainous regions in the south-east of Iceland. (2) A transitional growth phase (3.0–2.5 Ma), when the ice sheet(s) extended their margins to the north and west, reaching the sea in the north by Tjörnes by 2.5 Ma. (3) The presence of glacial deposits across all of the highlands by 2.5–2.4 Ma, suggests that progressive cooling resulted in continued incremental north and westward growth of the Quaternary ice sheet from its nucleus in southeast Iceland. By this time, the ice sheet covered more than half of the island. (4) Shortly afterwards, between 2.5 and 2.2 Ma, a single ice sheet covered the whole of Iceland for the first time, reaching the modern shoreline, and full-scale glacial–interglacial cyclicity was established ([Geirsdóttir, 2011](#)).

For the first 1.5–2.0 Ma of the Quaternary, cycles of waxing and waning continental glaciation appeared at 41 ka intervals, as the climate oscillated between glacial and interglacial states. Among the prominent but interglaciations during this period is one at about 2.4 Ma that is especially well represented by the Kap København Formation, a 100-m-thick sequence of estuarine sediments that covered an extensive lowland area near the northern tip of Greenland ([Funder et al., 2001](#)). The rich and well-preserved fossil fauna and flora in the Kap København Formation record warming from cold conditions as interglaciation, followed by cooling during the subsequent 10–20 ka. During the peak warmth, forest trees reached the Arctic Ocean coast, 1000 km north of the northernmost trees today. Based on this warmth, [Funder et al. \(2001\)](#) suggested that the Greenland Ice Sheet must have been reduced to local ice caps in mountain areas. Although finely resolved time records are not available throughout the Arctic Ocean of that time, by analogy with present faunas along the Russian coast,

the coastal zone would have been ice-free for 2–3 months in summer. Today, this coast of Greenland experiences year-round sea ice, and models of diminishing sea ice in a warming world generally indicate long-term persistence of summertime sea ice off these shores (e.g., [Holland et al., 2006](#)). Thus, the reduced sea ice off northern Greenland during deposition of the Kap København Formation suggests a widespread interglaciation during which Arctic sea ice was greatly diminished, and possibly absent in summer.

During the warm interval described above, precipitation was greater and temperatures were higher than at the peak of the current interglaciation, about 8 ka, and the temperature difference was larger during winter than during summer. Higher temperatures were not caused by notably greater solar insolation, owing to the relative repeatability of the Milankovitch variations over millions of years (e.g., [Berger et al., 1992](#)). Uncertainties in atmospheric CO₂ concentration, ocean heat transport, and perhaps other factors at the time of the Kap København Formation are sufficiently large to preclude strong conclusions about the causes of the unusual warmth.

Potentially correlative records of warm Early Quaternary interglacial conditions are found in deposits along coastal plains in northern and western Alaska. Interglacial marine transgressions that repeatedly flooded the Bering Strait modified the configuration of adjacent coastlines, altered regional continentality, and reinvigorated the exchange of water masses between the North Pacific, Arctic, and North Atlantic oceans. Since the first submergence of the Bering Strait about 5.5–5.0 Ma ([Marincovich and Gladenkov, 2001](#)), this marine gateway allowed relatively warm Pacific water to reach as far north as the Beaufort Sea ([Brigham-Grette and Carter, 1992](#)). The Gubik Formation of northern Alaska records at least three warm high sea stands in the early Quaternary. During the Colvillian transgression, about 2.7 Ma, the Alaskan Coastal Plain supported open boreal forest or spruce–birch woodland with scattered pine and rare fir and hemlock ([Nelson and Carter, 1991](#)). Warm marine conditions are confirmed by the general character of the ostracode fauna, which includes *Pterygocythereis vannieuwenhuisei* ([Brouwers, 1987](#)), an extinct species of a genus whose modern northern limit is the Norwegian Sea and which, in the northwestern Atlantic Ocean, is not found north of the southern cold-temperate zone ([Brouwers, 1987](#)). Despite the high sea level and relative warmth indicated by the Colvillian transgression, erratic boulder in Colvillian deposits southwest of Barrow, Alaska, indicate that glaciers then terminated in the Arctic Ocean and produced icebergs large enough to reach northwest Alaska.

The subsequent Bigbendian transgression (about 2.5 Ma), characterized by rich molluscan faunas including the gastropod *Littorina squalida* and the bivalve *Clinocardium californiense* ([Carter et al., 1986](#)), indicates renewed warmth along northern Alaska. The modern northern limit of both of these mollusk species is well to the south (Norton Sound, Alaska). The presence of sea otter bones suggests that the limit of seasonal ice on the Beaufort Sea was restricted during the Bigbendian interval to positions north of the Colville river and thus well north of typical twentieth century positions ([Carter et al., 1986](#)); modern sea otters cannot tolerate severe seasonal sea-ice conditions ([Schneider and Faro, 1975](#)).

The Fishcreekian transgression (about 2.4–2.1 Ma) has been correlated with the Kap København Formation on Greenland ([Brigham-Grette and Carter, 1992](#)). However, age control is imprecise, and [Brigham \(1985\)](#) and [Goodfriend et al. \(1996\)](#) suggested that the Fishcreekian could be as young as 1.4 Ma. This deposit contains several mollusk species that currently are found only south of the winter sea ice margin. Moreover, sea otter remains and the intertidal gastropod *L. squalida* at Fish Creek suggest that perennial sea ice was absent or severely restricted during the Fishcreekian transgression ([Carter et al., 1986](#)). Correlative deposits rich in mollusk species that currently live only well to the south are reported from the coastal plain at Nome, Alaska ([Kaufman and Brigham-Grette, 1993](#)).

The available data clearly indicate episodes of relatively warm conditions that correlate with high sea levels and reduced sea ice in the early Quaternary. The high sea levels suggest the melting of land ice. Thus, the correlation of warmth with diminished ice on land and at sea is analogous to features of contemporary change, as indicated by recent instrumental observations, model results, and data from other time intervals.

The Mid-Pleistocene Transition: 41 and 100 ka Worlds

Throughout the Quaternary the cyclical waxing and waning of continental ice sheets has dominated global climate variability in response to variations in solar insolation caused by features of Earth's orbit. After the onset of glaciation in North America, ice growth and decay occurred on the same frequency as the Earth's obliquity (tilt), a 41 ka cycle. But between 1.2 and 0.7 Ma, the variations in ice volume became larger and the periods much longer, averaging approximately 100 ka between interglaciations by 700 ka. Although Earth's eccentricity varies with an approximately 100 ka period, this variation results in only small changes in solar insolation in the key regions of ice growth, far less than precession and obliquity cycles, so eccentricity is unlikely to be the cause of this change; the reason for the dominant 100 ka period in ice volume remains debated.

The MPT is of particular interest because it was not caused by any major change in Earth's orbital behavior, and so the transition likely reflects a fundamental threshold within the climate system. Models for the 100 ka variability commonly assign a major role to the ice sheets themselves and especially to the Laurentide Ice Sheet, which dominated the total global change in ice volume. [Roe and Allen \(1999\)](#) assessed six different explanations of this behavior and found that all fit the data rather well; no one model has superiority over another.

A key observation that must be satisfied in any explanation for the mid-Pleistocene transition (MPT) is that the ice sheets of the last 700 ka were larger in volume but smaller in area than were the largest early Quaternary ice sheets ([Balco et al., 2005a,b](#)). [Clark and Pollard \(1998\)](#) used this observation to argue that the early Laurentide Ice Sheets must have had a substantially lower surface elevation than in the late Pleistocene, possibly by as much as 1 km. They argued that at the onset of the Quaternary, ice sheets advanced over easily deformed water-saturated weathered terrain. The low basal shear strength allowed fast ice flow, producing aerial extensive, but relatively thin ice sheets. Successive glacial cycles gradually

eroded the regolith, eventually exposing irregular bedrock in the central region. With increased resistance to basal sliding, the ice sheets eventually became thicker with a steeper surface profile, but not quite as extensive aerially as the early Quaternary ice sheets, despite their greater ice volume.

Other hypotheses emphasize the gradual global cooling that began in the early Cenozoic and continued through much of the Quaternary (Raymo et al., 2006; Ruddiman, 2003). For example, cooling may have passed a threshold whereby ice sheets attained sufficient thickness in their growth phase that the duration of the 'warm' portion of the 41 ka tilt cycle was too short to melt enough ice to cause deglaciation. Then the 100-ka cycle may result from the cumulative build up of the Laurentide Ice Sheet until it was sufficiently thick that it finally trapped enough of Earth's internal heat to thaw the ice-sheet bed, and the consequent faster ice flow lead to deglaciation. Marshall and Clark (2002) modeled the growth and decay of the Laurentide Ice Sheet and found that during growth the ice was frozen to its bed, decreasing flow velocities. After many tens of thousands of years, the modeled ice had thickened sufficiently to trap enough geothermal heat to thaw the bed, which allowed faster flow. Faster flow of the ice sheet lowered the surface gradient, which allowed warming and melting. Behavior such as that described could cause the main variations of ice volume to be slower than the orbitally driven variations in insolation. Alternative hypotheses require interactions in the Southern Ocean between the ocean and sea ice and between the ocean and the atmosphere (Gildor et al., 2002). For example, Toggweiler (2008) suggested that because of the close connection between the southern westerly winds and meridional overturning circulation in the Southern Ocean, shifts in wind fields could control the exchange of CO₂ between the ocean and the atmosphere. Carbon models support the notion that weathering and the burial of carbonate can be perturbed in ways that alter deep ocean carbon storage and result in 100 ka CO₂ cycles. Others have suggested that 100 ka cycles and CO₂ might be controlled by variability in obliquity cycles (i.e., two or three 41 ka cycles that produce a mixture of 80 and 120 ka cycles after 700 ka; Huybers, 2006, 2007, or by variable precession cycles altering the 19 and 23 ka cycles; Raymo, 1997). Ruddiman (2006) furthered these ideas but suggested that since 900 ka, CO₂-amplified ice growth continued at the 41 ka intervals but that polar cooling dampened ice ablation. His CO₂-feedback hypothesis suggests a mechanism that combines the control of 100 ka cycles with precession cycles (19 and 23 ka) and with tilt cycles (41 ka).

Other hypotheses also exist for these changes. A complete explanation of the onset of extensive glaciation on North America and Eurasia at the beginning of the Quaternary, and of the MPT from 41 to 100 ka ice-age cycles, remain the object of ongoing investigations.

A Link Between Ice Volume, Air Temperature, and Greenhouse Gases

The globally averaged temperature decrease during the 100-ka ice-age cycles since the MPT was about 5–6 °C (Jansen et al., 2007). Larger decreases are calculated for the Arctic and close

to the ice sheets, with a decrease of ~22 °C atop the Greenland Ice Sheet at the Last Glacial Maximum (LGM; Cuffey et al., 1995). The total change in solar radiation reaching the planet during these cycles was near zero; orbital features served primarily to change insolation seasonally and geographically.

Many factors probably contributed to this large temperature change despite the small global change in total insolation. Cooling produced growth of reflective ice that increased the planetary albedo, while the great height of the ice sheets on its own reduced Arctic temperatures as well. But changes in atmospheric greenhouse gases are necessary to explain the magnitude of the global cooling.

Antarctic ice cores now provide us with accurate estimates of atmospheric trace gas concentrations for the past 800 ka (Lüthi et al., 2008, and references therein). Complex changes, especially in the ocean, reduced atmospheric carbon dioxide concentrations during glaciations, and both oceanic and terrestrial changes reduced atmospheric water vapor, methane, and nitrous oxide, all of which are greenhouse gases. The changes in water vapor and carbon dioxide were most important in the planetary energy balance. Increasing aridity in some areas and glacial outwash in others produced additional dust that reduced the flux of insolation reaching the planet's surface (e.g., Mahowald et al., 2006). Cooling caused regions formerly forested to give way to grasslands, tundra or dune fields, further increasing the planetary albedo. While Earth's orbital features paced the ice-age cycles, strong positive feedbacks are required to provide quantitatively accurate explanations of the observed glacial and interglacial changes.

The relation between climate and carbon dioxide has been strongly correlated for at least the past 800 ka (Lüthi et al., 2008), and the growth and shrinkage of ice, cooling and warming of the globe, and other changes have repeated along similar, although not identical paths.

Marine Isotopic Stage 11 – A Long Interglaciation

Following the MPT, the growth and decay of ice sheets followed a 100 ka cycle: brief, warm interglaciations lasted 10 to about 40 ka, after which ice progressively increased in volume to a maximum, before ice volumes decreased rapidly and the planet transitioned into the next interglaciation. Although this 100 ka cycle is unlikely to be linked to changes in the 100 ka eccentricity cycle because it produces so little a change in solar energy reaching the Earth, there is an additional ~400 ka orbital cycle; Earth's orbit goes from almost round to more eccentric to almost round in about 100 ka, but the maximum eccentricity reached in these 100-ka cycles increases and decreases within a 400-ka cycle (Berger and Loutre, 1991). When the orbit is almost round, there is little effect from Earth's precession. But, about 400 ka, during MIS 11, the 400-ka cycle caused a nearly round orbit to persist, and the interglaciation as recorded in both ice cores and marine sediment lasted longer than previous or subsequent warm times (Jouzel et al., 2007). When Earth's orbit is nearly round, there is little change in insolation during a precession cycle. Consequently, summer insolation at high northern latitudes may not have become low enough at the end of the first 11 ka precession hemisphere to allow ice growth in the Arctic.

Mid- and low-latitude paleoclimate records show that MIS 11 lasted ~30 ka, rather than the typical 10 ka duration, but sediments containing MIS 11 are rare in the Arctic and their temperature signals are inconclusive (see [Helmke and Bauch, 2003](#), and references therein). Sea level during MIS 11 was higher than at any time since (cf. [Olson and Hearty, 2009](#)), and data from Greenland are consistent with notable shrinkage or loss of the ice sheet accompanying the long warmth, although the dating is poorly constrained ([Willerslev et al., 2007](#)).

Marine Isotopic Stage 5e: The Last Interglaciation

The warmest millennia of at least the past 250 000 years occurred during MIS 5, and especially during the warmest part of that interglaciation, MIS 5e ([Bauch et al., 1999](#); [Fronval and Jansen, 1997](#); [Kukla, 2000](#); [McManus et al., 1994](#)). During MIS 5e Earth's orbital parameters aligned to produce a strong positive anomaly in solar radiation during summer throughout the Northern Hemisphere ([Berger and Loutre, 1991](#)). Between 130 and 127 ka, the average solar radiation during the key summer months (May, June, and July) was about 11% greater than solar radiation at present throughout the Northern Hemisphere, and a slightly greater anomaly, 13%, existed over the Arctic. Greater solar energy in summer lead to melting of the Laurentide and Fennoscandinavian ice sheets earlier, during the precession cycle than during the last deglaciation. Sea level was higher than present 129 ka, coincident with the Northern Hemisphere insolation peak, whereas sea level only reached present in the Holocene about 6 ka, 5 ka after the peak in Northern Hemisphere insolation ([CAPE Last Interglacial Project Members, 2006](#)). The large insolation anomaly and intensification of the North Atlantic Drift combined to reduce Arctic Ocean sea ice, allow expansion of boreal forest to the Arctic Ocean shore throughout large regions, reduce permafrost, and melt almost all glaciers in the Northern Hemisphere ([CAPE Last Interglacial Project Members, 2006](#)).

High solar radiation in summer during MIS 5e, amplified by key boundary-condition feedbacks (especially sea ice, seasonal snow cover, and atmospheric water vapor), collectively produced summer temperature anomalies 4–5 °C above present over most Arctic lands ([CAPE Last Interglacial Project Members, 2006](#)), substantially above the average Northern Hemisphere summer temperature anomaly (1 ± 1 °C above present). MIS 5e demonstrates the strength of positive feedbacks on Arctic warming ([Miller et al., 2010b](#)).

Terrestrial MIS 5e Records

Summers throughout the Arctic were warmer during MIS 5e than at present, but the magnitude of warming differed spatially. Positive summer temperature anomalies were largest around the Atlantic sector, where summer warming was typically 4–6 °C higher than present. This anomaly extended into Siberia, but it decreased from Siberia westward to the European sector (0–2 °C), and eastward toward Beringia (2–4 °C). The Arctic coast of Alaska had sea-surface temperatures 3 °C above recent values and considerably less summer sea ice than recently, but much of interior Alaska had smaller anomalies

(0–2 °C) that probably extended into western Canada. In contrast, northeastern Canada and parts of Greenland had summer temperature anomalies of about 5 °C, and perhaps more. A stratified lacustrine sequence from the eastern Canadian Arctic that captures MIS 7, MIS 5e, and the Holocene, suggests that for at least this portion of the Arctic, only full interglacials were warm enough for lakes to be ice-free in summers long enough for the preservation of biotic-bearing sediment; the record also suggested that twentieth century warming represents a no-analog situation for the lake ([Axford et al., 2009](#)).

Precipitation and winter temperatures are more difficult to reconstruct for MIS 5e than are summer temperatures. In northeastern Europe, the latter part of MIS 5e was characterized by a marked increase in winter temperatures. A large positive winter temperature anomaly also occurred in Russia and western Siberia, although the timing is not as well constrained ([Funder et al., 2002](#); [Gudina et al., 1983](#); [Troitsky, 1964](#)). Qualitative precipitation estimates for most other sectors indicate wetter conditions than in the Holocene.

Marine MIS 5e Records

Low sedimentation rates in the central Arctic Ocean and the rare preservation of carbonate fossils limit the number of sites at which MIS 5e can be reliably identified in sediment cores. Peak concentrations of a foraminifer species that usually dwells in subpolar waters were found in MIS 5e zones and interpreted to indicate relatively warm interglacial conditions and much reduced sea-ice cover in the interior Arctic Ocean ([Adler et al., 2009](#); [Nørgaard-Pedersen et al., 2007a,b](#)). Interpretation of these and other proxies is complicated by the strong vertical stratification in the Arctic Ocean; today, warm Atlantic water (>1 °C) is in most areas isolated from the atmosphere by a relatively thin layer of cold (<–1 °C) low-salinity surface water, limiting the transfer of heat to the atmosphere. It is not always possible to determine whether subpolar Foraminifera found in Arctic Ocean sediment cores lived in warm waters that remained isolated from the atmosphere below the cold surface layer, or whether the warm Atlantic water had displaced the cold surface layer and was interacting with the atmosphere and affecting its energy balance.

Landforms and fossils from the western Arctic and Bering Strait indicate substantially reduced sea ice during MIS 5. The winter sea-ice limit is estimated to have been as much as 800 km farther north than its average twentieth century position, and summer sea ice was likely to have been much reduced relative to present ([Brigham-Grette and Hopkins, 1995](#)). These reconstructions are consistent with the northward migration of treeline by hundreds of kilometers throughout much of Alaska and the expansion of treeline to the Arctic Ocean coast in the Far East of Russia ([Lozhkin and Anderson, 1995](#)).

Marine Isotopic Stage 3 (70–30 ka): Rapid Climate Oscillations

The Arctic climate history through MIS 3 is difficult to reconstruct because of the paucity of continuous records and the

difficulty in providing a secure time frame. Continental ice sheets were reduced, relative to ice volumes during MIS 2 and MIS 4, and sea level was slightly higher, but the coastline was still well offshore in most places. The $\delta^{18}\text{O}$ record of temperature over the Greenland Ice Sheet and other ice-core data show that the North Atlantic region experienced repeated abrupt climate oscillations between fully glacial and interstadial conditions throughout MIS 3. These oscillations, known as Dansgaard-Oeschger (DO) events, each lasted several hundred to a few thousand years, with temperatures changing by as much as 15°C (Alley, 2007, and references therein). Similar climate excursions are also recorded in cave sediments from China and Yemen (Burns et al., 2003; Dykoski, et al., 2005; Wang et al., 2001), in high-resolution marine records off California (Behl and Kennett, 1996), and sediment cores from the Cariaco Basin (Hughen et al., 1996), the Arabian Sea (Schulz et al., 1998) and the Sea of Okhotsk (Nürnberg and Tiedemann, 2004), among many other sites. Rapid climate changes within MIS 3 were thus hemispheric to global in nature and are considered a sign of large-scale coupling between the cryosphere, ocean, and atmosphere. Although the cause of these events is still debated, they were likely the result of abrupt reorganizations of the ocean's thermohaline circulation, probably related to inherent ice sheet instabilities that introduced large quantities of icebergs and fresh water into the North Atlantic (MacAyeal, 1993).

The terrestrial Arctic vegetation record reflects increased continentality that very likely contributed to relatively high summer temperatures that presumably were offset by colder winters. Variable paleoenvironmental conditions were typical of the MIS 3 (Karaginskii) period throughout Arctic Russia, although regional correlations are difficult. However, evidence for repeated abrupt temperature oscillations is weak, possibly because of the difficulty of obtaining precise dates in this time range. On the New Siberian Islands, Andreev et al. (2001) document the existence of graminoid-rich tundra covering wide areas of the emergent shelf during MIS 3, with summer temperatures possibly 2°C higher than during the twentieth century. At Elikchan 4 Lake in the upper Kolyma drainage, the sediment record contains at least three MIS 3 intervals when summer temperatures and treeline reached late Holocene conditions (Anderson and Lozhkin, 2001). MIS 3 insect faunas in the lower Kolyma are thought to have thrived in summers that were $1\text{--}4.5^\circ\text{C}$ higher than recently (Alfimov et al., 2003).

The warmest widespread MIS 3 interval in Beringia occurred 44–35 ka (Anderson and Lozhkin, 2001); it is well represented in proxies from interior sites, although there is little or no vegetation response in areas closest to Bering Strait. Summer warmth appears to have been strongest in eastern Beringia where temperatures were only $1\text{--}2^\circ\text{C}$ lower than at present between 45 and 33 ka (Elias, 2007). The warmest interval in interior Alaska (Fox Thermal Event), about 40–35 ka, was marked by spruce forest tundra, although in the Yukon, forests were most dense a little earlier, about 43–39 ka (Anderson and Lozhkin, 2001).

The transition from MIS 3 to MIS 2 is marked by a shift from warm-moist to cold-dry conditions in western Beringia, but is absent or subtle in all but a few records in Alaska (Anderson and Lozhkin, 2001).

Marine Isotopic Stage 2 (30–15 ka): The LGM

The LGM was cold globally, but particularly cold in the Arctic. Planetary temperatures were $5\text{--}6^\circ\text{C}$ lower than at present (Jansen et al., 2007), whereas mean annual temperatures in central Greenland were depressed more than 20°C (Cuffey et al., 1995), with a similar summer temperature reduction estimated for Beringia (Elias et al., 1996). Much of the Arctic lands lay beneath continental ice sheets, and the Arctic Ocean was mantled by a continuous cover of sea ice and entrapped icebergs (Bradley and England, 2008). The lack of ocean-atmosphere transfer of heat during winter months would result in exceptionally cold winters, not only over the Arctic Ocean proper, but also across adjacent lands. These factors together limit the direct records of LGM climate across most of the Arctic. The available evidence for terrestrial ecosystems suggests that low- and high-shrub tundra were extremely restricted during the LGM, and largely replaced by polar desert or graminoid and forb tundra and an intergradation of steppe and tundra in the north Eurasian interior (Bigelow et al., 2003), ecosystem changes consistent with cold, dry climates with little snow cover.

Global ice volumes peaked about 21 ka. Rising solar insolation across the Northern Hemisphere in summer and increases in greenhouse gases caused Northern Hemisphere ice sheets to begin to recede from their maxima shortly after 20 ka, with the rate of recession increasing noticeably after about 16 ka (see, e.g., Clark et al., 2004; Dyke et al., 2003). Most coastlines became ice-free before 12 ka, and ice continued to melt rapidly as Northern Hemisphere summer insolation reached a peak ($\sim 9\%$ above modern) about 11 ka.

A major climate reversal during the deglacial cycle, the Younger Dryas, was initially recognized in pollen records from northwest Europe, and is now formally defined by the large isotopic excursion in the Greenland Ice Sheet cores, between 12.9 and 11.7 ka (Rasmussen et al., 2006). It can be considered that last of the large DO events, with its strongest expression centered on the northern North Atlantic, and downstream into Eurasia, but also effecting to a lesser degree the rest of the Northern Hemisphere.

Marine Isotopic Stage 1, The Holocene

The transition from MIS 2 to MIS 1, which marks the start of the Holocene, is commonly placed at the abrupt termination of the Younger Dryas cold event, about 11.7 ka (Rasmussen et al., 2006). By 8 ka the main unstable Northern Hemisphere ice sheets had been reduced in size that there impacts on Arctic climate was minimal, and by 6 ka, sea level and ice volumes were close to present, and greenhouse gases in the atmosphere differed little from preindustrial conditions (e.g., Jansen et al., 2007). The Holocene is identified as a separate epoch, not because the climate of the present interglaciation is particularly noteworthy, but because it is the period when humans became significant agents of geological change.

Holocene Thermal Maximum

A wide variety of evidence from terrestrial and marine archives indicates that peak Holocene Arctic summertime warmth was

achieved during the early Holocene, when most regions of the Arctic experienced sustained summer temperatures that exceeded the observed twentieth century values. This period of peak warmth, is geographically variable in its timing (Kaufman et al., 2004). The ultimate driver of the warming was orbital forcing, which produced increased summer solar radiation across the Northern Hemisphere. At 70° N June insolation 11 ka was about 45 W m⁻² larger than present, for a total change of about 9% (Berger and Loutre, 1991). January insolation did not change because there is no insolation that far north in January. Consequently, the net annual insolation change for the Arctic during a precession cycle exceeds that of lower latitudes.

High-resolution (decades to centuries) archives containing multiple climate proxies are available for most of the Holocene throughout the Arctic. Consequently, the Holocene record allows evaluation of the range of natural climate variability and of the magnitude of climate change in response to relatively small changes in forcings. Most Arctic paleoenvironmental records for the Holocene thermal maximum (HTM) record primarily summer temperatures. Many different proxies have been exploited to derive these reconstructions, including biological proxies (pollen, diatoms, chironomids, dinoflagellate cysts, and other microfossils), elemental and isotopic geochemical indices from lacustrine and marine sediments and ice cores, borehole temperatures, age distributions of radiocarbon-dated tree stumps north of (or above) current treeline, and the extralimital distributions of thermophilous marine invertebrates and whales.

A synthesis of 140 Arctic paleoclimatic and paleoenvironmental records extending from Beringia westward to Iceland (Kaufman et al., 2004) outlines the nature of the HTM in the western Arctic. The HTM has an average duration of 2.1 ka in Beringia to 3.5 ka in Greenland. The interval from 10 to 4 ka contains the greatest number of sites recording HTM conditions and the greatest spatial extent of those conditions in the western Arctic. The HTM began earliest in Beringia, where warmer-than-present summers became established 14–13 ka. Intermediate ages for HTM initiation (10–8 ka) are apparent in the Canadian Arctic Archipelago and in central Greenland. The HTM on Iceland occurred a bit later, 8–6 ka, whereas the onset on Svalbard was 10.8 ka. The latest general HTM onset (7–4 ka) was in central Canada. Most regions registered the onset of summer cooling by 6–5 ka. The average summer temperature anomaly during the HTM across the western Arctic was 1.6 °C, although it ranged from 0.5 to 3.0 °C, with the largest changes in the North Atlantic sector and the smallest over North Pacific sector.

The spatial pattern of HTM onset can be explained in part by regional variations in the proximity to continental ice sheets, which depressed temperatures nearby until the ice melted back, and to changes in freshwater delivery to the North Atlantic, which influenced the strength of the Atlantic meridional circulation.

Sea-ice conditions in the Arctic Ocean and adjacent channels have been developed by from a wide array of key proxies including the remains of whales and walrus, warm-water marine mollusks, changes in microfaunal assemblages, especially diatoms and dinoflagellate cysts, and more recently using biomarkers diagnostic of sea-ice or sea-ice-free organisms

(Polyak et al., 2010). These reconstructions parallel the terrestrial record for the most part. They demonstrate that an increased mass of warm Atlantic water moved into the Arctic Ocean at about the beginning of the Holocene. An obligate Atlantic-water mollusk appeared on northern Svalbard by 10.6 ka, with peak penetration of Atlantic water between 9.3 and 8 ka (Salvigsen, 2002), with most thermophilous mollusks disappearing from Svalbard by ~5 ka. Dates on the flooding of Bering Strait range from ~12 ka to slightly before 13 ka, although precise dating remains dependent on knowledge of the marine reservoir age at that time. The influx of Pacific water to the Arctic Ocean peaked between 9 and 5 ka, then diminished through the late Holocene. The extra ocean heat from both Atlantic and Pacific sources, coupled with increased summer insolation, decreased the area of perennial sea-ice cover during the early Holocene. Decreased sea-ice cover in the western Arctic also may be indicated by changes in concentrations of sea-salt sodium in the Penny Ice Cap, eastern Canadian Arctic, and the Greenland Ice Sheet.

As summer temperatures increased through the early Holocene, North American treeline expanded northward into regions formerly mantled by tundra, although the northward extent appears to have been limited to perhaps a few tens of kilometers beyond its recent position. In contrast, some tree-line species expanded well beyond their current limits in the Eurasian Arctic. Dated macrofossils indicate that individuals of the tree genera *Betula* (birch) and *Larix* (larch) lived far north of the modern treeline across most of northern Eurasia between 11 and 10 ka, and *Larix* extended beyond its modern limits as early as 13–12 ka in eastern Siberia (Binney et al., 2009, and references therein). In the Taimyr Peninsula of Siberia and nearby regions the most northerly limit reached by trees during the Holocene was more than 200 km north of the current treeline. Spruce (*Picea*) advanced slightly later than the other two genera. Interestingly, pine (*Pinus*), which now forms the conifer treeline in Fennoscandia and the Kola Peninsula of Russia, does not appear to have established appreciable forest cover at or beyond the present treeline in the Kola Peninsula until around 7 ka. However, quantitative reconstructions of temperature suggest that summer temperatures were above modern by 9 ka, and the development of extensive pine cover at and north of the present treeline appears to have been delayed relative to this warming. Treeline began to retreat southward across northern Eurasia 3–4 ka (Binney et al., 2009).

The timing of the HTM in the Eurasian Arctic overlaps the widest expression of the HTM in the western Arctic, but the timing of onset and termination in Eurasia shows less variability than in North America, and the magnitude of the treeline expansion and retreat is greater in the Eurasian Arctic.

Permafrost was degraded across much of the Russian Arctic during the HTM. Melting permafrost was widespread in the European north, but across Siberia permafrost partially thawed locally (reviewed by Astakhov, 1995). Areas south of the Arctic Circle appear to have experienced deep thawing (100–200 m depth) through the HTM, followed by renewed permafrost expansion after 3 ka.

Quantitative estimates of HTM summer temperature anomalies along the northern margins of Eurasia and adjacent islands range from 1 to 3 °C. Sea-surface temperature anomalies

during the HTM were as much as 4–5 °C for the eastern North Atlantic sector and adjacent Arctic Ocean.

Neoglaciation

Many climate proxies are available to characterize the overall pattern of late Holocene climate change. Although the timing of the HTM around the Arctic is at least partially modulated by the waning continental ice sheets, the last substantial ice-sheet-dominated climate perturbation was the brief cold excursion about 8.2 ka. By 8 ka the rapidly receding continental ice sheets had become small enough that they no longer exerted a strong influence on regional climates, and most proxy summer temperature records from the Arctic indicate an overall cooling trend through the late Holocene, as orbital influences steadily reduced summer insolation across the Northern Hemisphere.

The least ambiguous proxies that reflect the overall decrease in Northern Hemisphere summer temperatures through the Holocene are mountain glaciers and ice caps. Widespread evidence for the expansion or regrowth of glaciers after the HTM was summarized by [Porter and Denton \(1967\)](#), who introduced the term Neoglaciation to describe this reversal of the long deglaciation following the LGM. Although the timing of the onset of Neoglaciation across the Arctic is poorly constrained, most regions show evidence of initial glacier expansion or regrowth between 6 and 5 ka. Throughout the Holocene glacier mass balance has been dominantly controlled by summer temperature ([Koerner, 2005](#)), and the expansion of glaciers throughout the Arctic is interpreted to reflect a decrease in summer temperatures.

Neoglacial summer cooling is supported by several other lines of evidence: a reduction in melt layers in the Agassiz Ice Cap ([Koerner and Fisher, 1990](#)), and the decrease in $\delta^{18}\text{O}$ values in ice cores across Greenland and Arctic Canada ([Vinther et al., 2009](#)), clear patterns of change in borehole thermometry through the Greenland Ice Sheet; the retreat of large marine mammals and warm-water-dependent mollusks from the Canadian Arctic; the southward migration of the northern treeline across central Canada, Eurasia, and Scandinavia ([Barnekow and Sandgren, 2001](#)); the expansion of sea-ice cover along the shores of the Arctic Ocean on Ellesmere Island, in Baffin Bay, and in the Bering Sea; and the shift in vegetation communities inferred from plant macrofossils and pollen around the Arctic, including Wrangel Island ([Lozhkin et al., 2001](#)). The assemblage of microfossils and the stable isotope ratios of Foraminifera in marine sediment indicate a shift toward colder, lower salinity sea-surface conditions about 5 ka along the East Greenland Shelf ([Jennings et al., 2002](#)) and the western Nordic seas ([Koç and Jansen, 1994](#)), suggesting an increased flux of sea ice from the Arctic Ocean. Where quantitative estimates of temperature change are available, they generally indicate that summer temperature decreased by 1–2 °C during this initial phase of cooling.

Climate of the Past 2 ka

Climate reconstructions of the past 2 ka have gained increased attention because they provide a background reference for the range of natural variability against which changes that might be tied to anthropogenic activities can be compared. For most

of the Arctic, the pattern of overall summer cooling following an early HTM forms a multimillennial trend that culminated in the cold Little Ice Age (LIA) of the thirteenth through nineteenth centuries. Most Arctic glaciers and ice caps reached their maximum dimensions of the past 8 ka during the LIA ([Miller et al., 2010a](#)). High-resolution Arctic climate reconstructions for the first millennium AD are relative rare. The pan-Arctic (to 60° N latitude) compilation of [Kaufman et al. \(2009\)](#) included 23 proxy temperature records from lake sediments, tree rings, and glacier ice that extend back at least 1000 years are resolved at annual to subdecadal scale. The composite reconstruction is consistent with other evidence for millennial-scale trend of overall summer cooling since the HTM.

Warmth during Medieval times was initially recognized from evidence in Western Europe ([Lamb, 1977](#)), but the term is commonly applied to other regions over a wide temporal range. We restrict our consideration of Medieval warmth to the period between AD 950 and 1250. The strongest evidence for Medieval warmth comes from the northern North Atlantic region, where there is growing evidence for an episode of relatively cold summers between ~AD 600 and 950, followed by three centuries of relatively warmer summers during Medieval times. Evidence for warmth during this interval is based on reductions in glacier and ice cap dimensions, faunal changes in marine sediment, speleothems, ice cores, borehole temperatures, tree rings, and archaeology. Western Greenland ([Crowley and Lowery, 2000](#)), the Greenland Ice Sheet ([Dahl-Jensen et al., 1998](#)), Swedish Lapland ([Grudd et al., 2002](#)), northern Siberia ([Naurzbaev et al., 2002](#)), Arctic Canada ([Anderson et al., 2008](#)), and Iceland ([Geirsdóttir et al., 2009](#); [Larsen et al., 2011](#)) were all relatively warm around AD 1000. During Medieval time, Inuit populations moved out of Alaska into the eastern Canadian Arctic and hunted whale from skin boats in regions perennially ice-covered in the twentieth century ([McGhee, 2004](#)).

Medieval warmth is generally ascribed to a reduction in sulfate aerosols derived from explosive volcanism ([Jansen et al., 2007](#)), with the extra warmth around the North Atlantic and adjacent regions possibly linked to changes in oceanic circulation as well ([Broecker, 2001](#)).

The Arctic-wide reconstruction of [Kaufman et al. \(2009\)](#) indicates that average summer temperatures were similar to or higher than the warmest interval of the middle ages during much of the first half of the first millennium AD. Following the sixth century and prior to the twentieth, the average warmest interval occurred between AD 940 and 970. Only some sites registered peak preindustrial temperatures during this interval, however. The high average temperature in the composite record during Medieval time is dominated by sites in the northwestern North Atlantic sector (Greenland and Canadian High Arctic), especially isotope records from glacier ice. Indications of Medieval warmth outside the North Atlantic sector include the retreat of glaciers in southeastern Alaska ([Reyes et al., 2006](#); [Wiles et al., 2008](#)), and the wider tree rings in some high-latitude tree-ring records from Asia and North America ([D'Arrigo et al., 2006](#)). However, [D'Arrigo et al. \(2006\)](#) emphasized the uncertainties involved in estimating Medieval warmth relative to that of the twentieth century, owing in part to the sparse geographic distribution of proxy data as

well as to the less coherent variability of tree-growth temperature estimates. Hughes and Diaz (1994) argued that the Arctic as a whole was not anomalously warm throughout Medieval time.

Given the importance of understanding climate of the most recent past and the richness of the available evidence, intensive scientific effort has resulted in numerous Northern Hemisphere average temperature reconstructions for the past millennium (Mann et al., 2008, and references therein). These reconstructions are based on annually resolved proxy records, primarily from tree rings, with additional records from Greenland ice cores and a few annually laminated lake sediment records. Decadally resolved proxy temperature records from the Arctic (poleward of 60° N), primarily from lake sediments, have been explored for comparison (Kaufman et al., 2009, and references therein). In general, the Northern Hemisphere and the Arctic average temperature records are broadly similar: they show modest summer warmth during Medieval times, a variable, but colder climate from about AD 1250 to 1850, followed by twentieth century warming as shown by both paleoclimate proxies and the instrumental record. Less is known about changes in precipitation, which is spatially and temporally more variable than temperature.

The trend toward colder summers after ~AD 1250 coincides with the onset of the LIA, which persisted until ~AD 1850, although the timing and magnitude of specific cold intervals exhibit regional variability. The climate history of the LIA has been extensively studied in natural and historical archives, and it is well documented in Europe and North America (Grove, 1988). Historical evidence from the Arctic is relatively sparse, but it generally agrees with historical records from northwest Europe. Arctic proxy climate records from glacial and nonglacial sources show that the coldest interval of the Holocene occurred between ~AD 1450 and ~AD 1850, and during this interval most glaciers reached their Neoglacial maximum. Recent evidence from the Canadian Arctic indicates that, following substantial ice recession in Medieval times, glaciers and ice caps began to expand abruptly in the second half of the thirteenth century, and that ice expansion was further amplified ~AD 1450, after which ice caps receded to their pre-LIA margins only in recent decades (Miller et al., 2011).

The average summer temperature of the Northern Hemisphere during the LIA was no more than 1 °C lower than the twentieth century average (Mann et al., 2008), but was substantially greater in some regions. LIA cooling appears to have been stronger in the Atlantic sector of the Arctic than in the Pacific (Kaufman et al., 2004), perhaps because explosive volcanism promoted the development or transport of sea ice into the North Atlantic, leading to a reduction in the Atlantic meridional overturning circulation, which further amplified LIA cooling there (Miller et al., 2011). Although the initiation of the LIA and the structure of climate fluctuations during this multicentennial interval vary around the Arctic, most records show warming beginning in the late nineteenth century (Kaufman et al., 2009; Overpeck et al., 1997). The end of the LIA was apparently more uniform both spatially and temporally than its initiation.

Summer cooling that marks the transition into the LIA was in part a consequence of the Holocene-long decrease in Northern Hemisphere summer insolation due to the orbital changes

described earlier, and triggered by increased explosive volcanism in the late thirteenth and fifteenth centuries. This transition was likely aided by decreased solar irradiance at these times, and amplified by strong positive feedbacks, especially expanded Arctic Ocean sea ice and terrestrial snow cover, possibly enhanced by a weakening of the meridional circulation in the northern North Atlantic.

Placing Twentieth Century Arctic Warming in a Millennial Perspective

Much scientific effort has been devoted to learning how twentieth and early twenty-first century warmth compares with warmth during earlier times (e.g., Jansen et al., 2007). A data-model comparison indicates that the long-term cooling trend that dominated in the Arctic during the past nineteen centuries was linked to decreasing summer insolation from orbital factors, and that this trend was reversed during the twentieth century, despite continued reduction of summer insolation across the Arctic (Kaufman et al., 2009). Owing to the orbital changes affecting midsummer insolation (a drop in June insolation of about 1 W m⁻² at 75° N and 2 W m⁻² at 90° N during the last millennium; Berger and Loutre, 1991), additional forcing is required to explain why summer temperatures in recent decades have been similar to, and in most regions warmer than, summer temperatures achieved in Medieval times.

Thin, cold ice caps in the eastern Canadian Arctic preserve rooted vegetation beneath them that was killed by the expanding ice cover. A recent compilation of 94 ¹⁴C-dated plant samples recently emerged from beneath receding ice caps on northern Baffin Island shows that many of the ice caps that formed >1700 years ago persisted through Medieval times before melting early in the twenty-first century (Miller et al., 2011).

Records of surface melting from ice caps offer another view by which twentieth century warmth can be placed in a millennial perspective. The longest record comes from the Agassiz Ice Cap in the Canadian High Arctic, for which the percentage of summer melting of each season's snowfall is reconstructed for the past 10 ka. The percent of melt follows the general trend of decreasing summer insolation from orbital changes, but some brief departures are substantial. Of particular note is the significant increase in melt percentage during the past century; current percentages are higher than any other melt intensity since at least 4000 years ago (Fisher et al., 2011).

Changes in lake sediments and their biota record help place the recent century in a longer perspective. Extensive biotic changes, especially in the post AD 1850 interval, are interpreted to reflect recent warming above the Medieval warmth on Ellesmere Island and probably in other regions, although alternative explanations have been debated and explored (Smol and Douglas, 2007; Smol et al., 2005). D'Arrigo et al. (2006) show tree-ring evidence from a few North American and Eurasian sites that imply that summers were cooler in the Medieval Warm Period than in the late twentieth century, although the statistical confidence is weak. Tree-ring and tree-line studies in western Siberia (Esper and Schweingruber, 2004) and Alaska (Jacoby and D'Arrigo, 1995) suggest that

warming since 1970 has enhanced tree growth and follows a circumpolar trend. [Hantemirov and Shiyatov \(2002\)](#) present records from the Yamal Peninsula, Russia, well north of the Arctic Circle, showing summer temperatures of recent decades are the most favorable for tree growth within the past 4 ka.

The [National Research Council \(2006, p. 3\)](#) evaluated the available published data on globally and hemispherically averaged temperatures for the last millennium, and found that “Presently available proxy evidence indicates that temperatures at many, but not all, individual locations were higher during the past 25 years than during any period of comparable length since A.D. 900.” Greater uncertainties for hemispheric or global reconstructions were identified in assessing older comparisons. Climate models are unable to reproduce the observed warmth of the late twentieth century unless recent increases in greenhouse gases due to anthropogenic fossil fuel combustion are included, lending credibility to the argument that late twentieth century warmth is dominantly an artifact of human activities.

Conclusions

As the planet cooled from peak warmth in the early Cenozoic, extensive Northern Hemisphere ice sheets developed by 2.6 Ma, leading to changes in the circulation of both the atmosphere and oceans. From ~2.6 to ~1.0 Ma, ice sheets came and went about every 41 ka, in pace with cycles in the tilt of Earth’s axis, but for the past 700 ka, glacial cycles have been longer, lasting ~100 ka, separated by brief, warm interglaciations, when sea level and ice volumes were close to present. The cause of the shift from 41 to 100 ka glacial cycles is still debated. During the penultimate interglaciation, ~130 to ~120 ka, solar energy in summer in the Arctic was greater than at any time subsequently. As a consequence, Arctic summers were ~5 °C warmer than at present, and almost all Arctic glaciers melted completely except for the Greenland Ice Sheet, and even it was reduced in size substantially from its present extent. With the loss of land ice, sea level was ~5 m higher than present, with the extra melt coming from both Greenland and Antarctica, as well as small glaciers and ice caps. The LGM peaked ~21 ka, when mean annual temperatures over parts of the Arctic were as much as 20 °C lower than at present. Ice recession was well underway 16 ka; by 8 ka even the Laurentide Ice Sheet was too small to have a strong influence on the climate system, and most of the Northern Hemisphere ice sheets had melted by 6 ka. Insolation reached a summer maximum (9% higher than at present) ~11 ka and has been decreasing since then, primarily in response to the precession of the equinoxes. The extra energy elevated early Holocene summer temperatures throughout the Arctic 1–3 °C above twentieth century averages, enough to completely melt many small glaciers throughout the Arctic, although the Greenland Ice Sheet was only slightly smaller than at present. Early Holocene summer sea-ice limits were substantially smaller than their twentieth century average, and the flow of Atlantic water into the Arctic Ocean was substantially greater. As summer solar energy decreased in the second half of the Holocene, glaciers became re-established or advanced, sea ice expanded, and the flow of warm Atlantic water into the Arctic Ocean

diminished. Late Holocene cooling reached its nadir during the LIA (~AD 1250 to ~AD 1850), when sun-blocking aerosols from volcanic eruptions and perhaps other causes added to the orbital cooling, allowing most Arctic glaciers to reach their maximum Holocene extent. During the warming of the past century, glaciers have receded throughout the Arctic, terrestrial ecosystems have advanced northward, and perennial Arctic Ocean sea ice has diminished, with the most pronounced pan-Arctic changes occurring since AD 1970.

The histories of temperature and precipitation across Arctic lands during the Cenozoic exhibit clearly defined broad trends despite the limitations caused by scattered, and incomplete records. Arctic climate proxies for the most part reflect summer temperature anomalies, although ice-core isotopes reflect influences from changes in mean annual temperatures, moisture source area, and seasonality of precipitation. Precipitation changes in the past are more difficult to reconstruct than are paleotemperatures, in part due to the greater spatial variability of precipitation, but also due to the weak correlation between most proxies and precipitation amount. Nevertheless, some broad patterns clearly emerge.

On all time scales through the Cenozoic, temperature changes have been greater in the Arctic than for the Northern Hemisphere as a whole, for both summer and annual mean changes. This greater sensitivity reflects the powerful positive feedbacks acting across the Arctic that amplify changes due to external forcings. The strongest feedback on short timescales (decadal to millennial) is the expansion and contraction of sea ice, with additional positive feedbacks from changing seasonal snowcover, changing distribution of forest and tundra ecosystem boundaries, growth and decay of permafrost, and changes in the strength of the ocean’s meridional overturning circulation. On longer timescales (multimillennial), changes in the location and rate of deep convection in the North Atlantic, and the growth and decay of continental ice sheets are particularly important feedbacks, through changes in energy transfer, CO₂ and other greenhouse gas budgets, albedo, and ice-sheet height, and to a lesser extent, a negative feedback resulting from isostatic compensation to the increased ice load that lowers the ice surface. Accompanying changes in sea-surface temperatures control the amount of water vapor released to the atmosphere, influencing the magnitude of the greenhouse effect.

The great temperature reduction across the Arctic during glaciations increased the pole-to-equator thermal gradient, increasing planetary wind speeds, and the Arctic cold was amplified globally by reduced atmospheric water vapor and CO₂, the two most important greenhouse gases. Reduced atmospheric water vapor resulted in globally drier and dustier conditions on average, and ice sheet instabilities had dramatic impacts on the oceanic meridional overturning circulation, both with global consequences.

The primary forcing that explains the first-order climate trends through the Cenozoic has been the generally steady decrease in greenhouse gases in the atmosphere (with some notable perturbations) that produced a first-order temperature decline through most of the Cenozoic. Although this temperature decline continued through the Quaternary, high frequency temperature variations across the Arctic were superimposed on the gradual cooling trend, tied to the growth

and decay of the large continental ice sheets. Throughout the Quaternary, changes in the seasonal and geographic distribution of insolation caused by well-known oscillations in Earth's orbital parameters, primarily axial tilt (obliquity) and precession of the equinoxes, have paced the expansion and retreat of Northern Hemisphere ice sheets, with global climate consequences. Arctic Quaternary climates have also responded to variations in greenhouse gas forcings on glacial-interglacial timescales. During the Holocene, the general decrease in summer temperatures across the Arctic have been driven by precession, with perturbations to the millennial-scale cooling trend resulting from changes in the frequency of sulfur-rich explosive volcanic eruptions, and to small changes in solar luminosity. The strong warming trend over the past century across the Arctic, and since ~1970 in particular, stands in stark contrast to the first-order middle and late Holocene cooling trend, and is very likely a result of increased greenhouse gases that are a direct consequence of anthropogenic activities.

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See also: Paleoclimate Relevance to Global Warming. **Paleoclimate:** Introduction; Modern Analog Approaches in Paleoclimatology; The Younger Dryas Climate Event.

References

- Adler RE, Polyak L, Ortiz JD, et al. (2009) Sediment record from the western Arctic Ocean with an improved Late Quaternary age resolution: HOTRAX core HLY0503-8JPC, Mendeleev Ridge. *Global and Planetary Change* 68: 18–29.
- Alfimov AV, Berman DI, and Sher AV (2003) Tundra-steppe insect assemblages and the reconstruction of the Late Pleistocene climate in the lower reaches of the Kolyma River. *Zoologicheskii Zhurnal* 82: 281–300 (in Russian).
- Alley RB (2007) Wally was right: Predictive ability of the North Atlantic 'conveyor belt' hypothesis for abrupt climate change. *Annual Review of Earth and Planetary Sciences* 35: 241–272.
- Anderson PM and Lozhkin AV (2001) The stage 3 interstadial complex (Karginskii/middle Wisconsinan interval) of Beringia: Variations in paleoenvironments and implications for paleoclimatic interpretations. *Quaternary Science Reviews* 20: 93–125.
- Anderson RK, Miller GH, Briner JP, Lifton NA, and DeVogel SB (2008) A millennial perspective on Arctic warming from ^{14}C in quartz and plants emerging from beneath ice caps. *Geophysical Research Letters* 35: L01502. <http://dx.doi.org/10.1029/2007GL032057>.
- Andreev AA, Petet DM, Tarasov PE, Romanenko FA, Filimonova LV, and Sulerzhitsky LD (2001) Late pleistocene interstadial environment on Faddeyevskiy Island, East-Siberian Sea, Russia. *Arctic, Antarctic, and Alpine Research* 33: 28–35.
- Astakhov VI (1995) The mode of degradation of Pleistocene permafrost in West Siberia. *Quaternary International* 28: 119–121.
- Axford Y, Briner JP, Cooke CA, et al. (2009) Recent changes in a remote Arctic lake are unique within the past 200,000 years. *Proceedings of the National Academy of Sciences* 106(44): 18429–18430. <http://dx.doi.org/10.1073/pnas.0907094106>.
- Balco G, Rovey CW, and Stone OH (2005a) The first glacial maximum in North America. *Science* 307: 222.
- Balco G, Stone OH, and Jennings C (2005b) Dating Plio–Pleistocene glacial sediments using the cosmic-ray-produced radionuclides ^{10}Be and ^{26}Al . *American Journal of Science* 305: 1–41.
- Ballantyne AP, Greenwood DR, Sinninghe Damsté JS, Csank AZ, Eberle JJ, and Rybczynski N (2010) Significantly warmer Arctic surface temperatures during the Pliocene indicated by multiple independent proxies. *Geology* 38: 603–606. <http://dx.doi.org/10.1130/G30815.1>.
- Ballantyne AP, Rybczynski NL, Baker PA, Harington CR, and White D (2006) Pliocene Arctic temperature constraints from the growth rings and isotopic composition of fossil larch. *Palaeogeography, Palaeoclimatology, Palaeoecology* 242: 188–200.
- Barnekow L and Sandgren P (2001) Palaeoclimate and tree-line changes during the Holocene based on pollen and plant macrofossil records from six lakes at different altitudes in northern Sweden. *Review of Palaeobotany and Palynology* 117: 109–118.
- Bauch HA, Erlenkeuser H, Fahl K, et al. (1999) Evidence for a steeper Eemian than Holocene sea surface temperature gradient between Arctic and sub-Arctic regions. *Palaeogeography, Palaeoclimatology, Palaeoecology* 145: 95–117.
- Behl RJ and Kennett JP (1996) Brief interstadial events in the Santa Barbara Basin, NE Pacific, during the past 60 kyr. *Nature* 379: 243–246.
- Berger A and Loutre MF (1991) Insolation values for the climate of the last 10 million years. *Quaternary Science Reviews* 10: 297–317.
- Berger A, Loutre MF, and Laskar J (1992) Stability of the astronomical frequencies over the Earth's history for paleoclimate studies. *Science* 255(5044): 560–566.
- Bigelow N, Brubaker LB, Edwards ME, et al. (2003) Climate change and Arctic ecosystems – 1. Vegetation changes north of 55°N between the last glacial maximum, mid-Holocene, and present. *Journal of Geophysical Research* 108(D19): 8170. <http://dx.doi.org/10.1029/2002JD002558>.
- Binney HA, Willis KJ, Edwards ME, et al. (2009) The distribution of late-Quaternary woody taxa in northern Eurasia: Evidence from a new macrofossil database. *Quaternary Science Reviews* 28: 2445–2464.
- Bradley RS and England JH (2008) The Younger Dryas and the Sea of Ancient Ice. *Quaternary Research* 70: 1–10.
- Brigham JK (1985) *Marine Stratigraphy and Amino Acid Geochronology of the Gubik Formation, Western Arctic Coastal Plain, Alaska*, 21 pp. Doctoral dissertation, University of Colorado, Boulder. U.S. Geological Survey Open-File Report 85-381.
- Brigham-Grette J and Carter LD (1992) Pliocene marine transgressions of northern Alaska – Circumarctic correlations and paleoclimate. *Arctic* 43(4): 74–89.
- Brigham-Grette J and Hopkins DM (1995) Emergent-marine record and paleoclimate of the last interglaciation along the northwest Alaskan coast. *Quaternary Research* 43: 154–173.
- Broecker WS (2001) Was the medieval warm period global? *Science* 291(5508): 1497–1499. <http://dx.doi.org/10.1126/science.291.5508.1497>.
- Brouwers EM (1987) On *Prerogocythereis vunnieuwenhuisi* Brouwers sp. nov. In: Bate RH, Home DJ, Neale JW, and Siveter DJ (eds.) *A Stereo-Atlas of Ostracode Shells* 14: pp. 17–20. London: British Micropaleontological Society Part 1.
- Burns SJ, Fleitmann D, Matter D, Kramers J, and Al-Subbary AA (2003) Indian Ocean climate and an absolute chronology over Dansgaard/Oeschger Events 9 to 13. *Science* 301: 1365–1367.
- Carter LD, Brigham-Grette J, Marincovich L Jr., Pease VL, and Hillhouse US (1986) Late Cenozoic Arctic Ocean sea ice and terrestrial paleoclimate. *Geology* 14: 675–678.

- CircumArctic PaleoEnvironments (CAPE) – Last Interglacial Project Members (2006) Last Interglacial Arctic warmth confirms polar amplification of climate change. *Quaternary Science Reviews* 25: 1383–1400.
- Clark PU, McCabe AM, Mix AC, and Weaver AJ (2004) Rapid rise of sea level 19,000 years ago and its global implications. *Science* 304: 1141–1144.
- Clark PU and Pollard D (1998) Origin of the Middle Pleistocene transition by ice sheet erosion of regolith. *Paleoceanography* 13: 1–9.
- Crowley TJ and Lowery T (2000) How warm was the Medieval warm period? *Ambio* 29: 51–54.
- Cuffey KM, Clow GD, Alley RB, Stuiver M, Waddington ED, and Saltus RW (1995) Large arctic temperature change at the Wisconsin–Holocene glacial transition. *Science* 270(5235): 455–458.
- D'Arrigo R, Wilson R, and Jacoby G (2006) On the long-term context for late twentieth century warming. *Journal of Geophysical Research* 111: D03103. <http://dx.doi.org/10.1029/2005JD006352>.
- Dahl-Jensen D, Mosegaard K, Gundestrup N, et al. (1998) Past temperatures directly from the Greenland Ice Sheet. *Science* 282: 268–271.
- Dyke AS, Moore A, and Robertson L (2003) *Deglaciation of North America*. Geological Survey of Canada, Open File 1574 (published on CD-ROM).
- Dykoski CA, Edwards RL, Cheng H, et al. (2005) A high-resolution, absolute-dated Holocene and deglacial Asian monsoon record from Dongge Cave, China. *Earth and Planetary Science Letters* 233: 71–86.
- Elias SA (2007) Beetle records – Late Pleistocene North America. In: Elias SA (ed.) *Encyclopedia of Quaternary Science*, pp. 222–236. Amsterdam: Elsevier.
- Elias SA, Anderson K, and Andrews JT (1996) Late Wisconsin climate in the northeastern United States and southeastern Canada, reconstructed from fossil beetle assemblages. *Journal of Quaternary Science* 11: 417–421.
- Elias SA and Matthews JV Jr. (2002) Arctic North American seasonal temperatures in the Pliocene and Early Pleistocene, based on mutual climatic range analysis of fossil beetle assemblages. *Canadian Journal of Earth Sciences* 39: 911–920.
- Esper J and Schweingruber FH (2004) Large-scale treeline changes recorded in Siberia. *Geophysical Research Letters* 31: L06202. <http://dx.doi.org/10.1029/2003GL019178>.
- Fisher D, Zheng J, Burgess D, et al. (2011) Recent melt rates of Canadian Arctic ice caps are the highest in four millennia. *Global and Planetary Change* 84–85: 3–7.
- Fronval T and Jansen E (1997) Eemian and early Weichselian (140–60 ka) paleoceanography and paleoclimate in the Nordic seas with comparisons to Holocene conditions. *Paleoceanography* 12: 443–462.
- Funder S, Bennike O, Böcher J, Israelson C, Petersen KS, and Simonarson LA (2001) Late Pliocene Greenland – The Kap København formation in North Greenland. *Bulletin of the Geological Society of Denmark* 48: 117–134.
- Funder S, Demidov I, and Yelovicheva Y (2002) Hydrography and mollusc faunas of the Baltic and the White Sea–North Sea seaway in the Eemian. *Palaeogeography, Palaeoclimatology, Palaeoecology* 184: 275–304.
- Fyles JG, Marinovich L Jr., Matthews JV Jr., and Barendregt R (1991) Unique mollusc find in the Beaufort Formation (Pliocene) Meighen Island, Arctic Canada. *Current Research, Part B, Geological Survey of Canada, Paper 91-1B*, pp. 461–468.
- Geirsdóttir Á (2011) Pliocene and Pleistocene glaciations of Iceland: A brief overview of the glacial history. In: Ehlers J and Gibbard P (eds.) *Quaternary Glaciations – Extent and Chronology, Part IV – A Closer Look*, 2nd edn. Amsterdam: Elsevier INQUA Commission on Glaciation.
- Geirsdóttir Á, Miller GH, Thordarson T, and Olafsdóttir K (2009) A 2000 year record of climate variations reconstructed from Haukadalssvatn, West Iceland. *Journal of Paleolimnology* 41: 95–115. <http://dx.doi.org/10.1007/s10933-008-9253-z>.
- Geirsdóttir, Á (2011) Pliocene and Pleistocene glaciations of Iceland: A brief overview of the glacial history. In: Ehlers J, Gibbard PL, and Hughes PD (eds.) *Developments in Quaternary Science*, vol. 15, pp. 199–210. Amsterdam, The Netherlands. ISBN: 978-0-444-53447-7 © 2011 Elsevier B.V.
- Gibbard PL, Head MJ, and Walker MJC (2010) The Subcommission on Quaternary Stratigraphy. Formal ratification of the Quaternary System/Period and the Pleistocene Series/Epoch with a base at 2.58 Ma. *Journal of Quaternary Science* 25: 96–102.
- Gildor H, Tziperman E, and Toggweiler JR (2002) Sea ice switch mechanism and glacial–interglacial CO₂ variations. *Global Biogeochemical Cycles* 16: <http://dx.doi.org/10.1029/2001GB001446>.
- Goodfriend GA, Brigham-Grette J, and Miller GH (1996) Enhanced age resolution of the marine quaternary record in the Arctic using aspartic acid racemization dating of bivalve shells. *Quaternary Research* 45: 176–187.
- Grove JM (1988) *The Little Ice Age*, p. 498. London: Methuen.
- Grudd H, Briffa KR, Karlén W, Bartholin TS, Jones PD, and Kromer B (2002) A 7400-year tree-ring chronology in northern Swedish Lapland – Natural climatic variability expressed on annual to millennial timescales. *The Holocene* 12: 657–666.
- Gudina VI, Kryukov VD, Levchuk LK, and Sudakov LA (1983) Upper-Pleistocene sediments in north-eastern Taimyr. *Bulletin of Commission on Quaternary Researches* 52: 90–97 (in Russian).
- Hantemirov RM and Shiyatov SG (2002) A continuous multi-millennial ring-width chronology in Yamal, northwestern Siberia. *The Holocene* 12(6): 717–726.
- Haywood AM, et al. (2009) Comparison of mid-Pliocene climate predictions produced by the HadAM3 and CGAM3 General Circulation Models. *Global and Planetary Change* 66: 208–224.
- Helmke JP and Bauch HA (2003) Comparison of glacial and interglacial conditions between the polar and subpolar North Atlantic region over the last five climatic cycles. *Paleoceanography* 18(2): 1036. <http://dx.doi.org/10.1029/2002PA000794>.
- Holland MH, Bitz CM, and Tremblay B (2006) Future abrupt reductions in the summer Arctic sea ice. *Geophysical Research Letters* 33: L23503. <http://dx.doi.org/10.1029/2006GL028024>.
- Hughen K, Overpeck J, Anderson RF, and Williams KM (1996) The potential for paleoclimate records from varved Arctic lake sediments: Baffin Island, Eastern Canadian Arctic. In: Kemp AES (ed.) *Lacustrine Environments. Palaeoclimatology and Palaeoceanography from Laminated Sediments*, pp. 57–71. London: Geological Society. Special Publications 116.
- Hughes MK and Diaz HF (1994) Was there a 'Medieval Warm Period' and if so, where and when? *Journal of Climatic Change* 265: 109–142.
- Huybers P (2006) Early Pleistocene glacial cycles and the integrated summer insolation forcing. *Science* 313: 508–511.
- Huybers P (2007) Glacial variability over the last 2Ma: An extended depth-derived agemodel, continuous obliquity pacing, and the Pleistocene progression. *Quaternary Science Reviews* 26: 37–55.
- Jacoby GC and D'Arrigo RD (1995) Tree ring width and density evidence of climatic and potential forest change in Alaska. *Global Biogeochemical Cycles* 9(2): 227–234.
- Jansen E, Overpeck J, Briffa KR, et al. (2007) Palaeoclimate. In: Solomon S, Qin D, Manning M, Chen Z, Marquis M, Averyt KB, Tignor M, and Miller HL (eds.) *Climate Change 2007 – The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change*, pp. 434–497. Cambridge, UK and New York: Cambridge University Press.
- Jennings AE, Knudsen KL, Hald M, Hansen CV, and Andrews JT (2002) A mid-Holocene shift in Arctic sea ice variability on the East Greenland shelf. *The Holocene* 12: 49–58.
- Jouzel J, Masson-Delmotte V, Cattani O, et al. (2007) Orbital and millennial Antarctic climate variability over the past 800,000 years. *Science* 317: 793–796. <http://dx.doi.org/10.1126/science.1141038>.
- Kaufman DS, Ager TA, Anderson NJ, et al. (2004) Holocene thermal maximum in the western Arctic (0–180°W). *Quaternary Science Reviews* 23: 529–560.
- Kaufman DS and Brigham-Grette J (1993) Aminostratigraphic correlations and paleotemperature implications, Pliocene–Pleistocene high sea level deposits, northwestern Alaska. *Quaternary Science Reviews* 12: 21–33.
- Kaufman DS, Schneider DP, McKay NP, et al. (2009) Recent warming reverses long-term Arctic cooling. *Science* 235: 1236–1239.
- Kleiven HF, Jansen E, Fronval T, and Smith TM (2002) Intensification of Northern Hemisphere glaciations in the circum Atlantic region (3.5–2.4 Ma) – Ice-rafted detritus evidence. *Palaeogeography, Palaeoclimatology, Palaeoecology* 184: 213–223.
- Koç N and Jansen E (1994) Response of the high-latitude Northern Hemisphere to orbital climate forcing – Evidence from the Nordic Seas. *Geology* 22: 523–526.
- Koerner RM (2005) Mass Balance of glaciers in the Queen Elizabeth Islands, Nunavut, Canada. *Annals of Glaciology* 42(1): 417–423.
- Koerner RM and Fisher DA (1990) A record of Holocene summer climate from a Canadian high-Arctic ice core. *Nature* 343: 630–631.
- Kukla GJ (2000) The last interglacial. *Science* 287: 987–988.
- Lamb HH (1977) *Climate – Past, Present and Future. Volume 2: Climate History and the Future*, p. 835. London: Methuen.
- Larsen DJ, Miller GH, Geirsdóttir Á, and Thordarson T (2011) A 3000-year annually resolved glacier and climate record from the proglacial lake Hvítárvatn Iceland. *Quaternary Science Reviews* 30(19–20): 2715–2731.
- Lisiecki LE and Raymo ME (2005) A Pliocene–Pleistocene stack of 57 globally distributed benthic $\delta^{18}\text{O}$ records. *Paleoceanography* 20: PA1003. <http://dx.doi.org/10.1029/2004PA001071>.
- Lozhkin AV and Anderson PM (1995) The last interglaciation of northeast Siberia. *Quaternary Research* 43: 147–158.
- Lozhkin AV, Anderson PM, Vartanyan SL, Brown TA, Belaya BV, and Kotov AN (2001) Late Quaternary paleoenvironments and modern pollen data from Wrangel Island (Northern Chukotka). *Quaternary Science Reviews* 20: 217–233.
- Lüthi D, Le Floch M, Bereiter B, et al. (2008) High-resolution carbon dioxide concentration record 650,000–800,000 years before present. *Nature* 453: 379–382.

- MacAyeal DR (1993) Binge/purge oscillations of the Laurentide ice sheet as a cause of the North Atlantic's Heinrich events. *Paleoceanography* 8: 775–784.
- Mahowald NM, Muhs DR, Levis S, et al. (2006) Change in atmospheric mineral aerosols in response to climate: Last glacial period, preindustrial, modern, and doubled carbon dioxide climates. *Journal of Geophysical Research* 111: D10202. <http://dx.doi.org/10.1029/2005JD006653>.
- Mann ME, Zhang A, Hughes MK, et al. (2008) Proxy-based reconstructions of hemispheric and global surface temperature variations over the past two millennia. *Proceedings of the National Academy of Sciences* 105: 13252–13257.
- Marincovich L Jr. and Gladenkov AY (2001) New evidence for the age of Bering Strait. *Quaternary Science Reviews* 20(1–3): 329–335.
- Marshall SJ and Clark PU (2002) Basal temperature evolution of North American ice sheets and implications for the 100-kyr cycle. *Geophysical Research Letters* 29(24): 2214.
- McGhee R (2004) *The Last Imaginary Place: A Human History of the Arctic World*, p. 296. Ontario: Key Porter.
- McManus JF, Bond GC, Broecker WS, Johnsen S, Labeyrie L, and Higgins S (1994) High-resolution climate records from the North Atlantic during the last interglacial. *Nature* 371: 326–327.
- Miller GH, Alley RB, Brigham-Grette J, et al. (2010a) Arctic amplification: Can the past constrain the future? *Quaternary Science Reviews* 29: 1779–1790. <http://dx.doi.org/10.1016/j.quascirev.2010.02.008>.
- Miller GH, Brigham-Grette J, Anderson L, et al. (2010b) Temperature and precipitation history of the Arctic. *Quaternary Science Reviews* 29: 1679–1715. <http://dx.doi.org/10.1016/j.quascirev.2010.03.001>.
- Miller GH, Geirsdóttir Á, Zhong Y, et al. (2012) Abrupt onset of Little Ice Age cold triggered by volcanism and sustained by sea-ice/ocean feedbacks. *Geophysical Research Letters* 39: L02708. <http://dx.doi.org/10.1029/2011GL050168>.
- National Research Council (2006) *Surface temperature reconstructions for the last 2,000 years*, p. 160. Washington, DC: National Academies Press.
- Naurzbaev MM, Vaganov EA, Sidorova OV, and Schweingruber FH (2002) Summer temperatures in eastern Taimyr inferred from a 2427-year late-Holocene tree-ring chronology and earlier floating series. *The Holocene* 12: 727–736.
- Nelson RE and Carter LD (1991) Preliminary interpretation of vegetation and paleoclimate in northern Alaska during the late Pliocene Colvillean marine transgression. In: Bradley DC and Ford AB (eds.) *Geologic Studies in Alaska, U.S. Geological Survey Bulletin* 1999, pp. 219–222. Denver, CO: USGS.
- Nørgaard-Pedersen N, Mikkelsen N, and Kristoffersen Y (2007a) Arctic Ocean record of last two glacial–interglacial cycles off North Greenland/Ellesmere Island – Implications for glacial history. *Marine Geology* 244: 93–108.
- Nørgaard-Pedersen N, Mikkelsen N, Lassen SJ, Kristoffersen Y, and Sheldon E (2007b) Reduced sea ice concentrations in the Arctic Ocean during the last interglacial period revealed by sediment cores off northern Greenland. *Paleoceanography* 22: PA1218. <http://dx.doi.org/10.1029/2006PA001283>.
- Nürnberg D and Tiedemann R (2004) Environmental changes in the Sea of Okhotsk during the last 1.1 million years. *Paleoceanography* 19: PA4011.
- Olson SL and Hearty PJ (2009) A sustained +21 m sea-level highstand during MIS 11 (400 ka): Direct fossil and sedimentary evidence from Bermuda. *Quaternary Science Reviews* 28: 271–285.
- Overpeck J, Hughen K, Hardy D, et al. (1997) Arctic environmental changes of the last four centuries. *Science* 278: 1251–1256.
- Pagani M, Liu Z, LaRiviere J, and Ravelo AC (2010) High Earth-system climate sensitivity determined from Pliocene carbon dioxide concentrations. *Nature Geoscience* 3: 27–30. <http://dx.doi.org/10.1038/ngeo724>.
- Polyak L, Alley RB, Andrews JT, et al. (2010) History of Sea Ice in the Arctic. *Quaternary Science Reviews* 29: 1757–1778. <http://dx.doi.org/10.1016/j.quascirev.2010.02.010>.
- Porter SC and Denton GH (1967) Chronology of neoglaciation. *American Journal of Science* 165: 177–210.
- Rasmussen SO, Andersen KK, Svensson AM, et al. (2006) A new Greenland ice core chronology for the last glacial termination. *Journal of Geophysical Research* 111: D6. <http://dx.doi.org/10.1029/2005JD006079>.
- Raymo ME (1997) The timing of major climate terminations. *Paleoceanography* 12: 577–585.
- Raymo ME, Lisiecki LE, and Nisancioglu KH (2006) Plio–Pleistocene ice volume, Antarctic climate, and the global $\delta^{18}\text{O}$ record. *Science* 313: 492–495.
- Reyes AV, Wiles GC, Smith DJ, et al. (2006) Expansion of alpine glaciers in Pacific North America in the first millennium A.D. *Geology* 34: 57–60.
- Roe GH and Allen MR (1999) A comparison of competing explanations for the 100,000-yr ice age cycle. *Geophysical Research Letters* 26(15): 2259–2262.
- Ruddiman WF (2003) Insolation, ice sheets and greenhouse gases. *Quaternary Science Reviews* 22: 1597.
- Ruddiman WF (2006) Ice-driven CO_2 feedback on ice volume. *Climate of the Past* 2: 43–55.
- Salvigsen O (2002) Radiocarbon-dated *Mytilus edulis* and *Modiolus modiolus* from northern Svalbard: Climatic implications. *Norsk Geografisk Tidsskrift* 56: 56–61.
- Schneider KB and Faro B (1975) Effects of sea ice on sea otters (*Enhydra lutris*). *Journal of Mammalogy* 56: 91–101.
- Schulz H, von Rad U, and Erlenkeuser H (1998) Correlation between Arabian Sea and Greenland climate oscillations of the past 110,000 years. *Nature* 393: 54–57.
- Sloan LC and Barron EJ (1992) A comparison of Eocene climate model results to quantified paleoclimatic interpretations. *Palaeogeography, Palaeoclimatology, Palaeoecology* 93(3–4): 183–202.
- Smol JP and Douglas MSV (2007) Crossing the final ecological threshold in high Arctic ponds. *Proceedings of the National Academy of Sciences* 104: 12395–12397.
- Smol JP, Wolfe AP, Birks HJB, et al. (2005) Climate-driven regime shifts in the biological communities of arctic lakes. *Proceedings of the National Academy of Sciences* 102: 4397–4402.
- St. John KE and Krissek LA (2002) The late Miocene to Pleistocene ice-rafting history of southeast Greenland. *Boreas* 31: 28–35.
- Toggweiler JR (2008) Origin of the 100,000-year timescale in Antarctic temperatures and atmospheric CO_2 . *Paleoceanography* 23: PA2211. <http://dx.doi.org/10.1029/2006PA001405>.
- Troitsky SL (1964) Osnoviye zakonmernosti izmeneniya sostava fauny po razrezam morskikh meshmorenykh sloev ust-eniseyskoy vpadiny i nishne-pechorskoy depressii. Akademia NAUK SSSR. *Trudy instituta geologii i geofiziki* 9: 48–65 (in Russian).
- Vinther BM, Buchardt SL, Clausen HB, et al. (2009) Holocene thinning of the Greenland ice sheet. *Nature* 461: 385–388.
- Wang YJ, Cheng H, Edwards RL, et al. (2001) A high-resolution absolute-dated late Pleistocene monsoon record from Hulu Cave, China. *Science* 294: 2345–2348.
- Wiles GC, Barclay DJ, Calkin PE, and Lowell TV (2008) Century to millennial-scale temperature variations for the last two thousand years indicated from glacial geologic records of Southern Alaska. *Global and Planetary Change* 60: 15–125.
- Willerslev E, Cappellini E, Boomsma W, et al. (2007) Ancient biomolecules from deep ice cores reveal a forested Southern Greenland. *Science* 317: 111–114.
- Zachos J, Pagani M, Sloan L, Thomas E, and Billups K (2001) Trends, rhythms, and aberrations in global climate 65 Ma to present. *Science* 292(5517): 686–693.

The Younger Dryas Climate Event

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Introduction

The Younger Dryas has long been viewed as the canonical abrupt climate change event. The name is derived from the late 1800s Swedish and Danish pollen records that indicated a return of the cold-tolerant plant *Dryas octopetala* following a warm interval (the Allerød warm period) (Andersson, 1896; Hartz and Milthers, 1901). The advent of radiocarbon dating techniques constrained this European cooling event to ~11–10 ka (ka: 1000 years ago) in radiocarbon years (Mangerud et al., 1974) or ~12.9–11.7 ka in calendar years (Rasmussen et al., 2006). At this time, boreal summer insolation (the driving force behind deglaciation) neared its peak (Figure 1(a)) and the sea level was approximately halfway through its deglacial rise of 120–130 m (Figure 1(e)). Subsequent research identified climate changes correlative to the European Younger Dryas in the North Atlantic and North Pacific Oceans, across Asia and North America, and in the tropics (e.g., Clark et al., 2002; Shakun and Carlson, 2010). The footprint of Younger Dryas climate change suggests that it was caused by a reduction in Atlantic meridional overturning circulation (AMOC) (Stouffer et al., 2006), for which there has been growing evidence since the mid-1980s (Boyle and Keigwin, 1987). This article reviews the Younger Dryas climate event, starting with the geographic pattern of climate change, followed by a discussion of ocean circulation changes, and ending with its ultimate forcing.

Extent of the Younger Dryas

The Northern Hemisphere Younger Dryas

Greenland ice cores

In the Northern Hemisphere, Greenland ice cores provide the best expression of the Younger Dryas as a cold event. Ice core $\delta^{18}\text{O}$ (δ signifies a change in isotopic ratio relative to a standard multiplied by 1000) consistently decreases by ~3‰ with an ~3.5‰ increase at the end of the event, reflecting up to 9 °C of cooling and 11 °C of warming (Figure 1(b)) (Alley, 2000). Isotopic ratios of nitrogen and argon gases from the GISP2 (Greenland Ice Sheet Project 2) ice core suggest 10 ± 4 °C of warming at the end of the Younger Dryas (Grachev and Severinghaus, 2005). The $\delta^{18}\text{O}$ records could also have an imprint of precipitation source change because Greenland Ice Sheet accumulation decreased by ~40% (0.11–0.07 m year⁻¹) (Figure 1(c)) and dust records suggest atmospheric reorganization in only several decades (Alley, 2000). Greenland deuterium excess, a proxy of precipitation source, shifted in 1–3 years at the start of the Younger Dryas, suggesting that 2–4 °C of source cooling was imprinted on the isotopic records (Steffensen et al., 2008). The trapped nitrogen isotopes also indicate that high-altitude Greenland was ~15 °C colder than present, but this estimate should not be extended to the ice

margins as it may reflect a strengthening of the near-surface atmospheric inversion that is common over ice sheets (Severinghaus et al., 1998). Indeed, east Greenland valley glacier records suggest that summers were only 6–7 °C cooler than present (Kelly et al., 2008). Greenland runoff to the ocean was also enhanced (Figure 1(d)) (Carlson et al., 2008b) with mild terrestrial and marine summer conditions (Björck et al., 2002; Williams, 1993) during the Younger Dryas.

Europe

European pollen, chironomid, lake $\delta^{18}\text{O}$, and speleothem records indicate that a cooler (2–6 °C) and/or drier climate characterized the Younger Dryas, which potentially changed in decades (Figure 2(a) and 2(b)); e.g., Brauer et al., 1999; Genty et al., 2006; Heiri et al., 2007; von Grafenstein et al., 1999). Mountain glaciers readvanced in the Swiss Alps implying 3–4 °C of cooling (e.g., Ivy-Ochs et al., 2009). The Scandinavian Ice Sheet also readvanced, depositing an extensive moraine system across southern Norway, Sweden, and Finland (Andersen et al., 1995). Records from glacially influenced lakes in western Norway show an increase in glacier activity (Nesje, 2009), and the Iceland Ice Cap halted retreat during the Younger Dryas (Licciardi et al., 2007).

North Atlantic records

In the eastern North Atlantic, Iberian Margin and Mediterranean sea surface temperature (SST) records show 1–3 °C of cooling (Figure 2(d)) (e.g., Bard et al., 2000; Cacho et al., 2001). Northward up to the Norwegian Sea, SST records indicate 1–7 °C of cooling (Figure 2(c)) (e.g., Benway et al., 2010; Chapman and Maslin, 1999; Dolven et al., 2002; Ebbesen and Hald, 2004; Karpuz and Jansen, 1992). Westward, SSTs cooled by 1–2 °C in the subtropical gyre (Carlson et al., 2008a; Chapman and Maslin, 1999) and by 5–10 °C near maritime Canada (de Vernal et al., 1996; Keigwin and Jones, 1995). In contrast, SST increased by ~1.5 °C near southeastern United States (Figure 2(e)), which was related to the slowing of AMOC and decreased northward heat transport (Carlson et al., 2008a). The Gulf of Mexico also cooled slightly near the end of the Younger Dryas (Figure 2(f)) (Flower et al., 2004).

North America

Pollen, lake $\delta^{18}\text{O}$, and soil $\delta^{13}\text{C}$ records from maritime Canada to north-central United States suggest a cooler and drier Younger Dryas (e.g., Dorale et al., 2010; Shuman et al., 2005; Yu and Eicher, 1998). In the southeastern United States, however, the Younger Dryas is characterized by a warmer and/or wetter climate, reflecting the trapping of heat in the western subtropical gyre due to reduced AMOC (Grimm et al., 2006). The Pacific Coast of the United States and Canada experienced 2–3 °C of Younger Dryas cooling (Figure 3(a) and 3(b)) and a wetter climate (e.g., Barron et al., 2003; Kaufman et al., 2010; Kienast and McKay, 2001; MacDonald et al., 2008; Mathewes

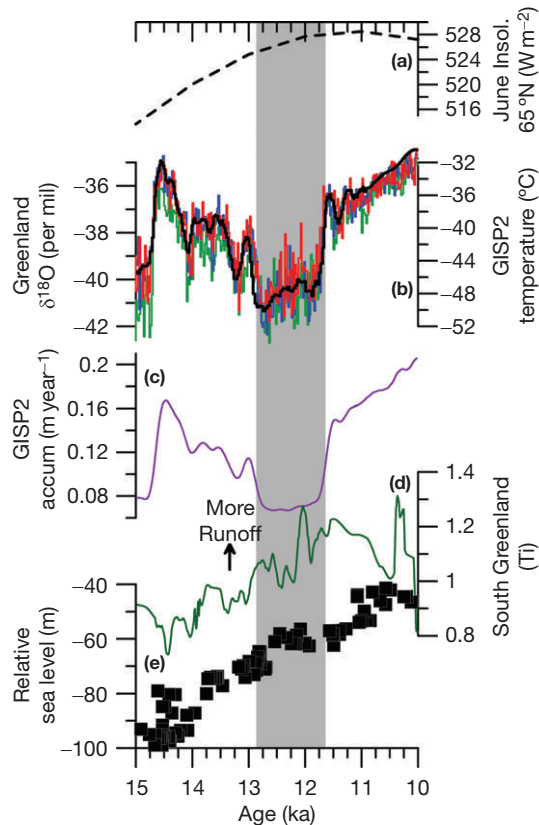


Figure 1 Greenland. (a) June insolation at 65°N (Berger and Loutre, 1991). (b) Greenland ice core $\delta^{18}\text{O}$ (red = GISP2 (Greenland Ice Sheet Project 2), blue = GRIP (Greenland Ice Core Project), green = NGRIP (North Greenland Ice Core Project); left) (Rasmussen et al., 2006) and GISP2 temperature (black; right) (Alley, 2000). (c) GISP2 ice core accumulation rate (Alley, 2000). (d) Ti concentration record off southern Greenland (Carlson et al., 2008b). (e) Relative sea-level data (Clark et al., 2009). Gray bar on this and other figures indicates timing of the Younger Dryas.

et al., 1993; Sea and Whitlock, 1995; Vacco et al., 2005). In the North American southwest, packrat midden $\delta^{13}\text{C}$ temperature reconstructions imply $>3^\circ\text{C}$ of cooling (Cole and Arundel, 2005), and lake, speleothem, and water table records suggest an increase in net precipitation during the Younger Dryas (Figure 3(c) and 3(d)) (Asmerom et al., 2010; Benson et al., 1997; Pigati et al., 2009; Polyack et al., 2004; Wagner et al., 2010). Valley glaciers readvanced or reoccupied cirques in many of the mountain ranges of the North American Cordillera (e.g., Davis et al., 2009). The southern margin of the Laurentide Ice Sheet also readvanced, depositing a moraine from western Lake Superior to southeastern Quebec (Lowell et al., 1999), with its northwest margin in Hudson Strait increasing iceberg discharge during the Younger Dryas (Andrews and Tedesco, 1992).

North Africa and the Middle East

In North Africa, climate cooled and net precipitation decreased during the Younger Dryas. A record of terrigenous input off the coast of northwest Africa indicates an increase in dust during the Younger Dryas, implying drying of the Sahara Desert (Figure 4(a)) (deMenocal et al., 2000), which is consistent

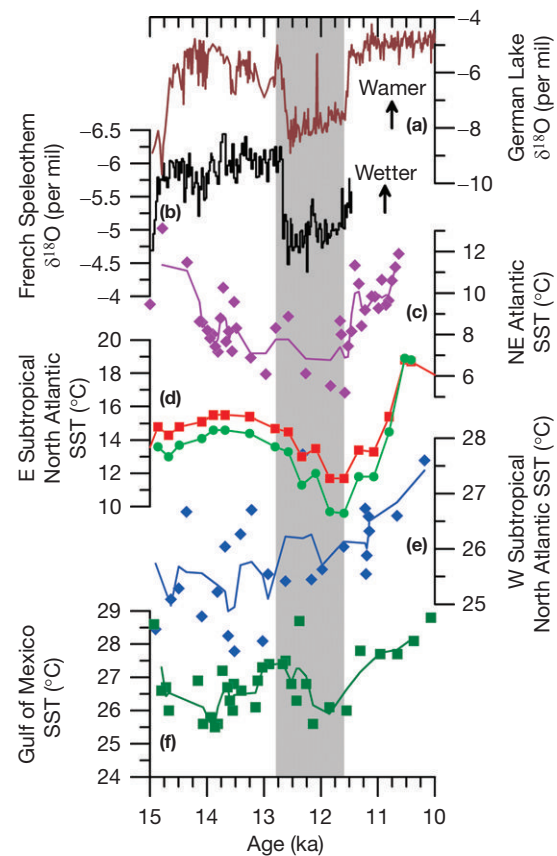


Figure 2 Europe and North Atlantic. (a) German lake $\delta^{18}\text{O}$ (von Grafenstein et al., 1999). (b) French speleothem $\delta^{18}\text{O}$ (Genty et al., 2006). (c) Fenian Drift Mg/Ca-SST (Benway et al., 2010). (d) Iberian Margin alkenone-SST (two different alkenone-SST relationships shown) (Bard et al., 2000). (e) Blake Outer Ridge Mg/Ca-SST (Carlson et al., 2008a). (f) Gulf of Mexico Mg/Ca-SST (Flower et al., 2004). Symbols are individual measurements; line is 3-point running average.

with speleothems from North Africa and the Middle East (Figure 4(b)) (e.g., Genty et al., 2006; Shakun et al., 2007). This precipitation decrease is attributable to the southward shift in the Intertropical Convergence Zone (ITCZ) in response to reduced AMOC and North Atlantic cooling (Stouffer et al., 2006). Reduced upwelling in the Arabian Sea because of decreased winds is also consistent with southward ITCZ migration (Altabet et al., 2002; Schulz et al., 1998). North African SSTs cooled $0.5\text{--}1^\circ\text{C}$ (deMenocal et al., 2000; Zhao et al., 1995), but Arabian SSTs were relatively constant during the Younger Dryas (Saher et al., 2007).

Asia

Large portions of central to northern Asia currently lack paleoclimate records spanning the Younger Dryas interval. Nevertheless, a decrease in $\delta^{18}\text{O}$ from Lake Baikal in southern Russia implies shifts in moisture sources to the lake and potential cooling during the Younger Dryas (Morley et al., 2005). Further east, a pollen record from central Japan suggests $\leq 3^\circ\text{C}$ of cooling during the Younger Dryas (Nakagawa et al., 2003), whereas SST records from the China Sea indicate $0.5\text{--}1^\circ\text{C}$ of cooling (Kubota et al., 2010; Sun et al., 2005). Multiple

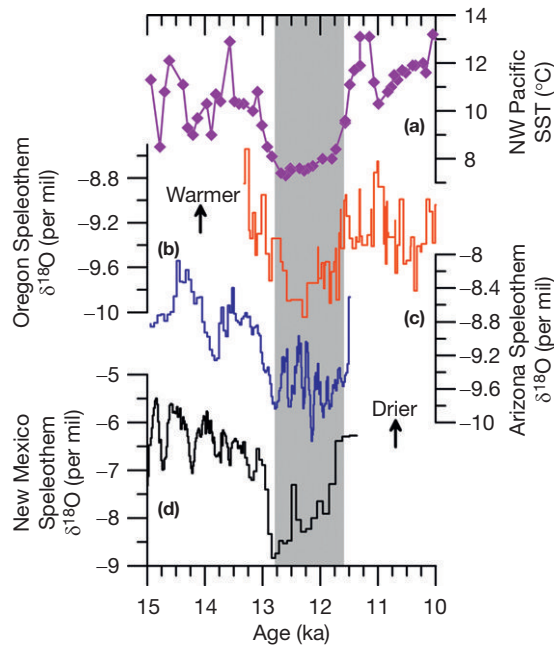


Figure 3 Western North America. (a) Northwest Pacific alkenone-SST (Barron et al., 2003). (b) Oregon speleothem $\delta^{18}\text{O}$ (Vacco et al., 2005). (c) Arizona speleothem $\delta^{18}\text{O}$ (Wagner et al., 2010). (d) New Mexico speleothem $\delta^{18}\text{O}$ (Asmerom et al., 2010).

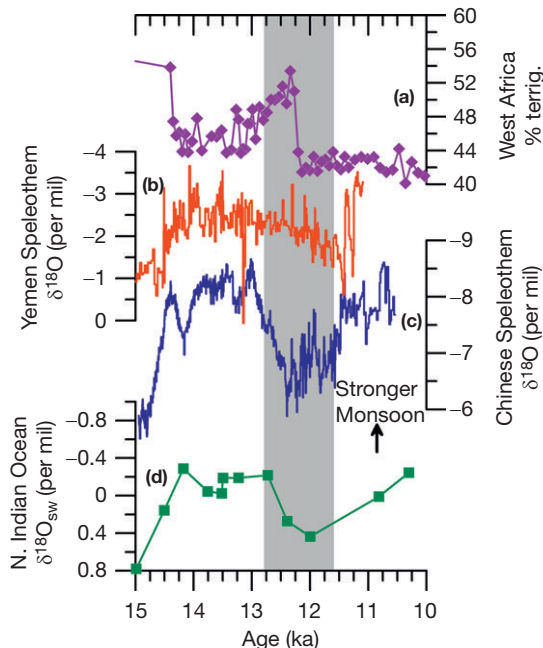


Figure 4 North Africa and Asia. (a) West Africa dustiness (deMenocal et al., 2000). (b) Yemen speleothem $\delta^{18}\text{O}$ (Shakun et al., 2007). (c) Chinese speleothem $\delta^{18}\text{O}$ (Wang et al., 2001). (d) North Indian Ocean $\delta^{18}\text{O}$ of seawater (sw) (Rashid et al., 2007).

speleothem $\delta^{18}\text{O}$ records from China and India and Indian Ocean marine salinity proxies imply a weakening of the summer Indo-Asian monsoon (Figure 4(c) and 4(d)) (Dykoski et al., 2005; Rashid et al., 2007; Sinha et al., 2005; Wang

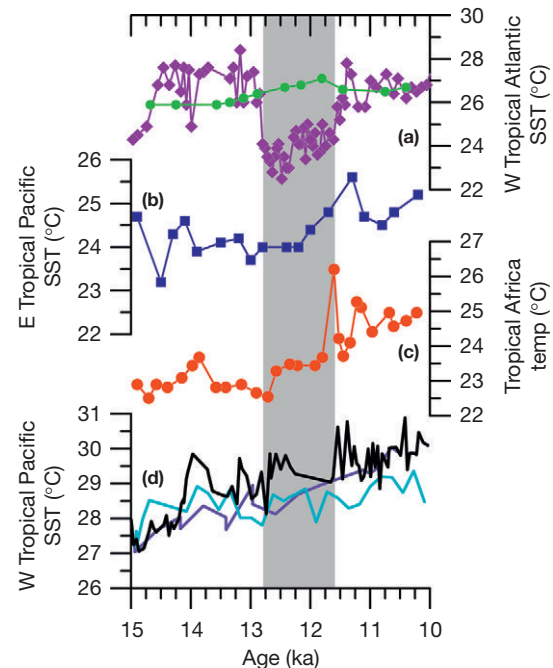


Figure 5 Tropics. (a) Mg/Ca-SST from Cariaco Basin (purple) (Lea et al., 2003) and alkenone-SST from Caribbean Sea (green) (Rühleman et al., 1999). (b) Mg/Ca-SST from eastern tropical Pacific (Lea et al., 2006). (c) Lake Tanganyika, Africa temperature (Tierney et al., 2008). (d) Mg/Ca-SST from western tropical Pacific (three records) (Stott et al., 2007).

et al., 2001), with a south China lake record suggesting an increase in winter Asian monsoon strength during the Younger Dryas (Yancheva et al., 2007). Overall, these paleomonsoon proxies are consistent with a southward shift in the ITCZ.

Tropical Climate and the Younger Dryas

East Atlantic and Central America

The temperature pattern observed in the tropics during the Younger Dryas is complex (Shakun and Carlson, 2010). SSTs show a warming of 0.25–1.2 °C in the Caribbean Sea and off northeast Brazil (Jaeschke et al., 2007; Rühleman et al., 1999; Schmidt et al., 2004; Weldeab et al., 2006), but a cooling of ~3.3 °C north of Venezuela (Figure 5(a)) (Lea et al., 2003), possibly from a southward shift in the ITCZ that increased trade wind strength and upwelling (Hughen et al., 1996). Indeed, precipitation proxies suggest that precipitation decreased north of the equator and increased south of the equator (Haug et al., 2001; Jaeschke et al., 2007; Wang et al., 2007). Ice cores from Peru and Bolivia show an ~2‰ decrease in $\delta^{18}\text{O}$ (Thompson et al., 1998), concurrent with higher lake levels in nearby Lake Titicaca and consistent with southward ITCZ migration (Baker et al., 2001). Despite wetter conditions, a warmer climate would explain glacier recession in Peru during the Younger Dryas (Rodbell and Seltzer, 2000).

West Atlantic and Africa

Off tropical West Africa, SST records indicate ~0.2 °C of warming during the Younger Dryas (Shefuß et al., 2005; Weldeab

et al., 2007) in agreement with air temperature proxies that suggest $\sim 0.8^\circ\text{C}$ of warming (Weijers et al., 2007). Another Younger Dryas SST record from southwestern tropical Africa, however, suggests a slight cooling from increased winds and an enhanced local upwelling of colder waters (Kim et al., 2002), which may extend to $\sim 25^\circ\text{S}$ (Farmer et al., 2005). Eastern tropical Africa temperature records indicate a complex pattern with warming at Lake Tanganyika (Figure 5(c)) and potential cooling at Lake Malawi and surrounding regions (Gasse et al., 2008; Tierney et al., 2008; Weldeab et al., 2007). Off the southeast coast of Africa near Madagascar, SSTs warmed during the Younger Dryas (Levi et al., 2007). Interestingly, despite evidence for a southward shift in the ITCZ in tropical Africa (e.g., Gasse et al., 2008; Johnson et al., 2002; Tierney and Russell, 2007), leaf wax $\delta^{13}\text{C}$ from Lakes Malawi and Tanganyika indicate arid conditions, highlighting a complex tropical African climate response to reduced AMOC (Castaneda et al., 2007; Tierney et al., 2008).

Tropical Pacific

In the eastern tropical Pacific, most SST records indicate little temperature change or slight warming during the Younger Dryas (Figure 5(b)) (e.g., Benway et al., 2006; Lea et al., 2006), with one record showing $\sim 0.4^\circ\text{C}$ of cooling (Kienast et al., 2006). Paleosalinity reconstructions suggest a net decrease in moisture transport from the Atlantic to the Pacific because of a southward displacement of the ITCZ (Benway et al., 2006). Further west, the Mauna Kea Ice Cap of the Hawaiian Islands may have readvanced during the Younger Dryas from a southward shift in the ITCZ, with enhanced extratropical storms and northward local winds that transported more moisture to Hawaii (Anslow et al., 2010). In the tropical west Pacific, SSTs generally warmed but with regions of cooling potentially reflecting proxy biases (Figure 5(d)) (Kienast et al., 2001; Levi et al., 2007; Linsley et al., 2010; Rosenthal et al., 2003; Stott et al., 2007).

Southern Hemisphere Climate During the Younger Dryas

During the Younger Dryas, the extratropical Southern Hemisphere generally warmed (Shakun and Carlson, 2010). Antarctic ice cores record increasing $\delta^{18}\text{O}$ and δD (Figure 6(e) and 6(f)) (e.g., Blunier and Brook, 2001; EPICA Community Members, 2006; Jouzel et al., 2007). SST records show a warming of $0.3\text{--}1.9^\circ\text{C}$ from the southeast Atlantic to New Zealand (Figure 6(a)–6(c)) (Barker et al., 2009; Barrows et al., 2007; Carlson et al., 2008b; Lamy et al., 2004; Pahnke and Sachs, 2006). Speleothem and pollen records from New Zealand and pollen records from South America confirm that the Younger Dryas was generally a period of warming (e.g., Hajdas et al., 2003; Moreno et al., 2009; Newnham and Lowe, 2000; Turney et al., 2003; Williams et al., 2005), consistent with New Zealand and Patagonian glacier retreat (Kaplan et al., 2008; 2010; Moreno et al., 2009; Putnam et al., 2010).

Geographic Response Summary

Globally, the Younger Dryas was a period of climate change. Plotting Younger Dryas temperature and climate anomalies against latitude shows that climate anomalies increased in magnitude toward the poles with opposite signs in the

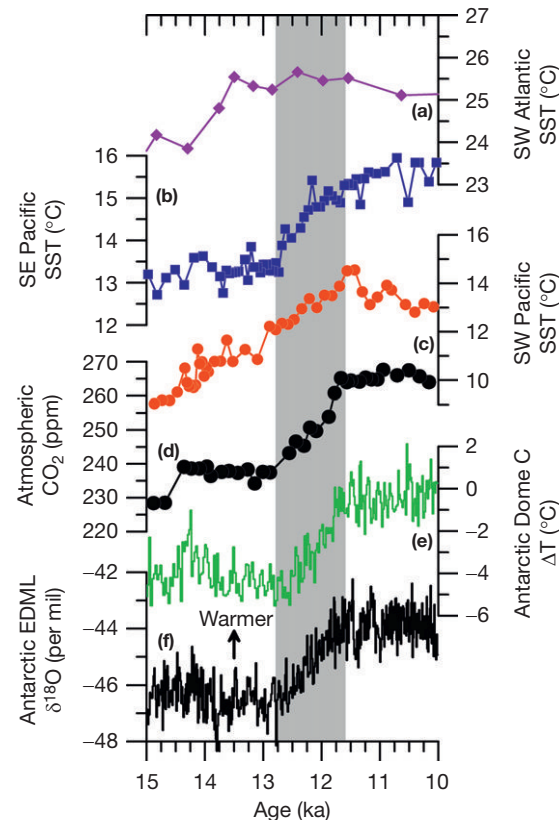


Figure 6 Southern Hemisphere. (a) Mg/Ca-SST near southeast Brazil (Carlson et al., 2008a). (b) Alkenone-SST near southern Chile (Lamy et al., 2004). (c) Alkenone-SST near New Zealand (Pahnke and Sachs, 2006). (d) Atmospheric CO_2 (Monnin et al., 2001). (e) Antarctic Dome C change in temperature (Jouzel et al., 2007). (f) Antarctic EDML ice core $\delta^{18}\text{O}$ (EPICA Community Members, 2006).

Northern and Southern Hemispheres (Figure 7) (Shakun and Carlson, 2010), reflecting the bipolar seesaw response (Blunier and Brook, 2001). Nevertheless, greater cooling at high northern latitudes than warming at high southern latitudes results in a net global cooling of $\sim 0.6^\circ\text{C}$ likely caused by more extensive snow and sea-ice cover increasing Northern Hemisphere albedo during the Younger Dryas (Shakun and Carlson, 2010).

Ocean Circulation During the Younger Dryas

The first evidence for a change in AMOC during the Younger Dryas came from North Atlantic proxies of water mass nutrients (benthic $\delta^{13}\text{C}$ and Cd/Ca) that showed that Southern Ocean Deep-water volume increased in the North Atlantic at the expense of North Atlantic Deep-water (Figure 8(b) and 8(c)) (Boyle and Keigwin, 1987). More recently, a proxy of water export ($^{231}\text{Pa}/^{230}\text{Th}$ in marine sediments) from the North Atlantic suggests a $\sim 30\%$ reduction in deep AMOC strength during the Younger Dryas (Figure 8(a)) (McManus et al., 2004), consistent with a reduction in the North Atlantic bottom water currents (Figure 8(d)) (Praetorius et al., 2008). Other tracers of water mass source and age confirm increased Southern Ocean Deep-water and Antarctic Intermediate-water and reduced North Atlantic Deep-water volume

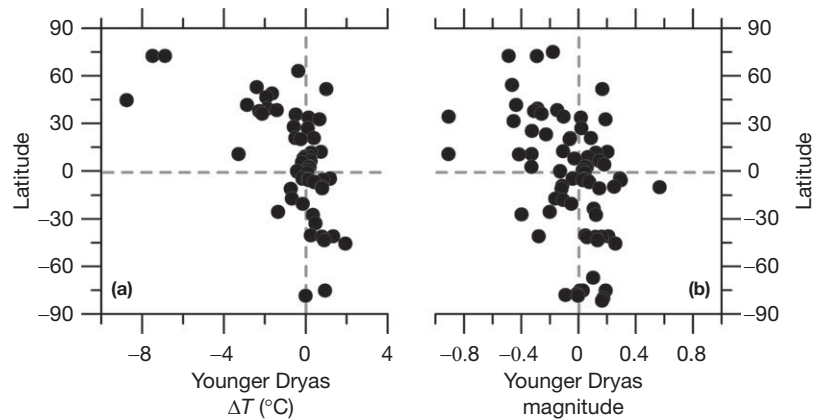


Figure 7 Latitudinal plots of the change in temperature between the Younger Dryas and Allerød (a) and the magnitude of Younger Dryas climate change relative to the glacial-interglacial change in a proxy record (b). Data from Shakun and Carlson (2010).

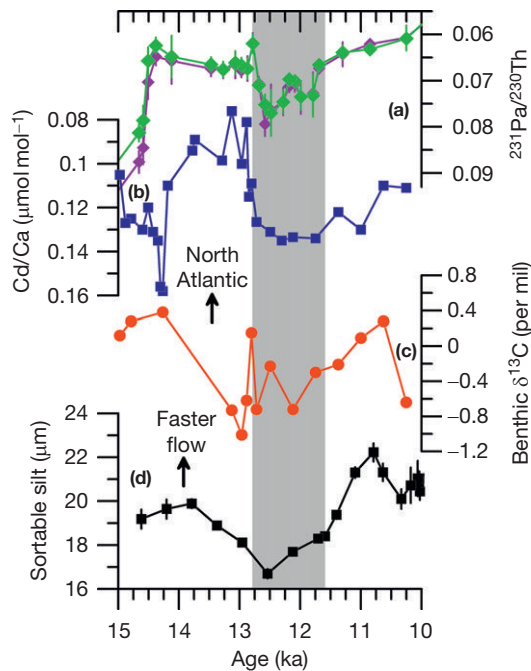


Figure 8 North Atlantic overturning circulation. (a) $^{231}\text{Pa}/^{230}\text{Th}$ for the North Atlantic with up indicating more water export past Bermuda Rise (McManus et al., 2004). (b) Benthic Cd/Ca from Bermuda Rise (Boyle and Keigwin, 1987). (c) Benthic $\delta^{13}\text{C}$ from Bermuda Rise (Boyle and Keigwin, 1987). (d) Sortable silt from the Iceland-Scotland Overflow (Praetorius et al., 2008). Gray bar indicates timing of Younger Dryas.

(Pahnke et al., 2008; Roberts et al., 2010; Robinson et al., 2005). The rise in atmospheric CO_2 during the Younger Dryas (Figure 6(d)) implies increased upwelling and degassing of carbon-rich water, evidence for which exists in the Southern Ocean and along Baja California (Anderson et al., 2009; Marchitto et al., 2007). The reduction in AMOC strength during the Younger Dryas provides the causal mechanism for Northern Hemisphere cooling and Southern Hemisphere warming, as well as the southward displacement of the ITCZ. Coupled

atmosphere-ocean climate models consistently simulate a reduction in AMOC and these attendant climate impacts in response to increased freshwater discharge to the North Atlantic (Liu et al., 2009; Meissner and Clark, 2006; Otto-Bliesner and Brady, 2009; Peltier et al., 2006; Stouffer et al., 2006).

The Cause of the Younger Dryas

The forcing of the Younger Dryas was originally inferred to be northward retreat of the southern Laurentide Ice Sheet margin out of the Great Lakes, thereby routing western Canadian Plains freshwater from the Mississippi River to the St. Lawrence River, with the attendant increase in freshwater discharge to the North Atlantic slowing AMOC (Johnson and McClure, 1976; Rooth, 1982). This hypothesis has more recently been questioned based on marine, terrestrial, and ice-sheet model results.

In the St. Lawrence Estuary, planktic $\delta^{18}\text{O}$ records show a 0.5–0.8‰ decrease at the start of the Younger Dryas (de Vernal et al., 1996; Keigwin and Jones, 1995; Keigwin et al., 2005) when $\delta^{18}\text{O}$ increases in the Gulf of Mexico (Figure 9(b)) (e.g., Flower et al., 2004). Concurrent Younger Dryas cooling of 5–10 °C in the St. Lawrence Estuary (de Vernal et al., 1996; Keigwin and Jones, 1995) may have masked a much larger $\delta^{18}\text{O}$ signal associated with eastward freshwater routing (1.8–2.7‰) (Carlson et al., 2007). Four independent geochemical freshwater source and amount proxies and mollusk $\delta^{18}\text{O}$ from the St. Lawrence River and Estuary support the eastward routing of western Canadian Plains freshwater at the start of the Younger Dryas (Figure 9(c) and 9(d)) (Brand and McCarthy, 2005; Carlson et al., 2007; Cronin et al., 2008). Modeling of the geochemical records indicates a base-flow freshwater discharge increase of ~ 0.12 Sv (Sv = Sverdrup, $10^6 \text{ m}^3 \text{ s}^{-1}$) (Carlson et al., 2007), which is sufficient to slow AMOC (Liu et al., 2009; Meissner and Clark, 2006; Otto-Bliesner and Brady, 2009; Peltier et al., 2006; Stouffer et al., 2006). Although dinoflagellate cyst reconstructed sea surface salinity does not decrease, which led de Vernal et al. (1996) to argue that St. Lawrence River freshwater discharge did not

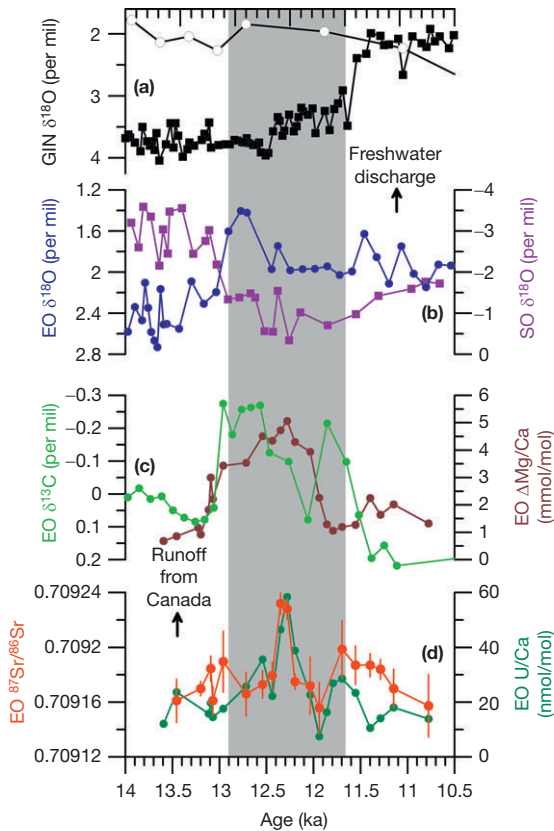


Figure 9 North American runoff records. (a) Two planktic $\delta^{18}\text{O}$ records from the Greenland-Iceland-Norwegian (GIN) Sea (Fram Strait, open symbols; Norwegian Sea, diamonds) (Dokken and Jansen, 1999; Nørgaard-Pedersen et al., 2003). (b) Planktic $\delta^{18}\text{O}$ records from the Gulf of Mexico (Southern Outlet: SO) (purple; Flower et al., 2004) and St. Lawrence Estuary (Eastern Outlet: EO) (blue; Keigwin et al., 2005). (c) Planktic $\delta^{13}\text{C}$ (light green; de Vernal et al., 1996) and $\Delta\text{Mg}/\text{Ca}$ (brown; Carlson et al., 2007). (d) Planktic $^{87}\text{Sr}/^{86}\text{Sr}$ (red) and U/Ca (dark green) from the St. Lawrence Estuary (Carlson et al., 2007).

increase during the Younger Dryas, subsequent research has shown that dinoflagellate cysts are insensitive to salinity changes above 12 practical salinity units (Telford, 2006). Similarly, ^{14}C dates from the eastern outlet of the western Canadian Plains have been interpreted as implying that the eastern outlet was not open until after the start of the Younger Dryas (Lowell et al., 2009). The dates, however, are only minimum limiting and indicate that the eastern outlet was open prior to ~ 12.6 ka, in agreement with the Mississippi and St. Lawrence runoff records (Carlson et al., 2009).

Arguments for a northern routing of western Canadian Plains freshwater and an abrupt discharge of Laurentide meltwater to the north have also been proposed as the forcing of the Younger Dryas. Murton et al. (2010) used optically stimulated luminescence (OSL) ages from sand layers bracketing an erosional surface on the Mackenzie River Delta to argue that the erosional event (e.g., flood) occurred shortly after ~ 13 ka based on the average of the stratigraphically lower dates. However, the OSL ages overlying the erosional surface of ~ 11.9 ka would be a closer age constraint on the erosion event, because erosion removed some unknown amount of sand and the

overlying ages are conformable with the end of the erosion event as flow waned, depositing the sand. Therefore, the erosion event is not related to the onset of the Younger Dryas. Tarasov and Peltier (2005) suggested that a 0.09 Sv discharge of Laurentide meltwater into the Arctic Ocean for 0.3 ka could have forced the Younger Dryas. Climate models do not, however, produce a >1 ka long reduction in AMOC from a 0.3 ka discharge of freshwater (Liu et al., 2009; Meissner and Clark, 2006; Otto-Bliesner and Brady, 2009; Peltier et al., 2006; Stouffer et al., 2006). Although such a melting event could be longer, sea-level records constrain the global meltwater discharge to <5 m of sea-level rise (Figure 1(f)) (Bard et al., 2010) or ~ 0.05 Sv over 1.2 ka. Furthermore, there is no existing paleoceanographic evidence for an increase in freshwater discharge to the Arctic Ocean at the start of the Younger Dryas (Figure 9(a)) (Carlson and Clark, 2008).

Another recent hypothesis for the source of the freshwater that forced the Younger Dryas is a comet that impacted on or near the Laurentide Ice Sheet (Firestone et al., 2007), but evidence for Younger Dryas-like events during earlier deglaciations precludes a unique comet forcing (Carlson, 2008). These earlier events occur at approximately the mid-point of the deglacial sea-level rise, similar to the sea level during the Younger Dryas, suggesting that the AMOC reduction is related to ice-sheet volume/extent (Carlson, 2008). Only continental routing from the Laurentide Ice Sheet retreat pertains to ice-sheet extent/size and the existing records all suggest that this freshwater routing was from the southern outlet to the eastern outlet, consistent with the originally hypothesized forcing of the Younger Dryas.

See also: Glacial Climates: Thermohaline Circulation. Ice Core Records: Correlations Between Greenland and Antarctica. Paleoceanography: Paleoceanography An Overview. Paleoclimate Reconstruction: Sub-Milankovitch (DO/Heinrich) Events; Younger Dryas Oscillation, Global Evidence.

References

- Alley RB (2000) The Younger Dryas cold interval as viewed from central Greenland. *Quaternary Science Reviews* 19: 213–226.
- Altabet MA, Higginson MJ, and Murray DW (2002) The effect of millennial-scale changes in Arabian Sea denitrification on atmospheric CO_2 . *Nature* 415: 159–162.
- Andersen BG, Lundqvist J, and Saarnisto M (1995) The Younger Dryas margin of the Scandinavian Ice Sheet – An introduction. *Quaternary International* 28: 145–146.
- Anderson RF, Ali S, Bradtmiller LI, et al. (2009) Wind-driven upwelling in the Southern Ocean and the deglacial rise in atmospheric CO_2 . *Science* 323: 1443–1448.
- Andersson G (1896) *Svenska växtvärldens historia*. Stockholm: P.A. Norstedt & Söner.
- Andrews JT and Tedesco K (1992) Detrital carbonate-rich sediments, northwestern Labrador Sea implications for ice-sheet dynamics and iceberg rafting (Heinrich) events in the North Atlantic. *Geology* 20: 1087–1090.
- Anslow FS, Clark PU, Kurz MD, and Hostettler SW (2010) Geochronology and paleoclimatic implications of the last deglaciation of the Mauna Kea Ice Cap, Hawaii. *Earth and Planetary Science Letters* 297: 234–248.
- Asmerom Y, Polyak VJ, and Burns J (2010) Variable winter moisture in the southwestern United States linked to rapid glacial climate shifts. *Nature Geoscience* 3: 114–117.
- Baker PA, Rigby CA, Seltzer GO, et al. (2001) Tropical climate changes at millennial and orbital timescales on the Bolivian Altiplano. *Nature* 409: 698–701.

- Bard E, Rostek F, Turon J-L, and Gendreau S (2000) Hydrological impact of Heinrich events in the subtropical Northeast Atlantic. *Science* 289: 1321–1324.
- Bard E, Hamelin B, and Delanghe-Sabatier D (2010) Deglacial meltwater pulse 1B and Younger Dryas sea levels revisited with boreholes at Tahiti. *Science* 327: 1235–1237.
- Barker S, et al. (2009) Interhemispheric Atlantic seesaw response during the last deglaciation. *Nature* 457: 1097–1102.
- Barron JA, Heusser L, Herbert T, and Lyle M (2003) High-resolution climatic evolution of coastal northern California during the past 16,000 years. *Paleoceanography* 18: <http://dx.doi.org/10.1029/2002PA000768>.
- Barrows TT, Lehman SJ, Fifield LK, and De Dekker P (2007) Absence of cooling in New Zealand and the adjacent ocean during the Younger Dryas chronozone. *Science* 318: 86–89.
- Benson L, Burdett J, Lund S, Kashgarian M, and Mensing S (1997) Nearly synchronous climate change in the Northern Hemisphere during the last glacial termination. *Nature* 388: 263–265.
- Benway HM, Mix AC, Haley BA, and Klinkhammer GP (2006) Eastern Pacific warm pool paleosalinity and climate variability: 0–30 kyr. *Paleoceanography* 21: <http://dx.doi.org/10.1029/2005PA001208>.
- Benway HM, McManus JF, Oppo DW, and Cullen JL (2010) Hydrographic changes in the eastern subpolar North Atlantic during the last deglaciation. *Quaternary Science Reviews* 29: 3336–3345.
- Berger A and Loutre MF (1991) Insolation values for the climate of the last 10 million years. *Quaternary Science Reviews* 10: 297–317.
- Björck S, et al. (2002) Anomalously mild Younger Dryas summer conditions in southern Greenland. *Geology* 30: 427–430.
- Blunier T and Brook EJ (2001) Timing of millennial-scale climate change in Antarctica and Greenland during the last glacial period. *Science* 291: 109–112.
- Boyle EA and Keigwin L (1987) North Atlantic thermohaline circulation during the past 20,000 years linked to high-latitude surface temperature. *Nature* 330: 35–40.
- Brand U and McCarthy FMG (2005) The Allerød–Younger Dryas–Holocene sequence in the west-central Champlain Sea, eastern Ontario: A record of glacial, oceanographic and climatic changes. *Quaternary Science Reviews* 24: 1463–1478.
- Brauer A, Endres C, Günter C, Litt T, Stebich M, and Negendank JFW (1999) High resolution sediment and vegetation responses to Younger Dryas climate change in varved lake sediments from Meerfeld Maar, Germany. *Quaternary Science Reviews* 18: 321–329.
- Cacho I, et al. (2001) Variability of the western Mediterranean Sea surface temperature during the last 25,000 years and its connection with the Northern Hemisphere climatic changes. *Paleoceanography* 16: 40–52.
- Carlson AE (2008) Why there was not a Younger Dryas-like event during the Penultimate Deglaciation. *Quaternary Science Reviews* 27: 882–887.
- Carlson AE and Clark PU (2008) Rapid climate change and Arctic Ocean freshening: Comment. *Geology* 36: e177.
- Carlson AE, et al. (2007) Geochemical proxies of North American freshwater routing during the Younger Dryas cold event. *Proceedings of the National Academy of Sciences of the United States of America* 104: 6556–6561.
- Carlson AE, Oppo DW, Came RE, LeGrande AN, Keigwin LD, and Curry WB (2008a) Subtropical Atlantic salinity variability and Atlantic meridional circulation during the last deglaciation. *Geology* 36: 991–994.
- Carlson AE, Stoner JS, Donnelly JP, and Hillaire-Marcel C (2008b) Response of the southern Greenland Ice Sheet during the last two deglaciations. *Geology* 36: 359–362.
- Carlson AE, Clark PU, and Hostetler SW (2009) Comment: Radiocarbon deglaciation chronology of the Thunder Bay, Ontario area and implications for ice sheet retreat patterns. *Quaternary Science Reviews* 28: 2546–2547.
- Castaneda IS, Werne JP, and Johnson TC (2007) Wet and arid phases in the southeast African tropics since the Last Glacial Maximum. *Geology* 35: 823–826.
- Chapman MR and Maslin MA (1999) Low-latitude forcing of meridional temperature and salinity gradients in the subpolar North Atlantic and the growth of glacial ice sheets. *Geology* 27: 875–878.
- Clark PU, Pisias NG, Stocker TF, and Weaver AJ (2002) The role of thermohaline circulation in abrupt climate change. *Nature* 415: 863–869.
- Clark PU, et al. (2009) The Last Glacial Maximum. *Science* 325: 710–714.
- Cole KL and Arundel ST (2005) Carbon isotopes from fossil packrat pellets and elevational movements of Utah agave plants reveal the Younger Dryas cold period in Grand Canyon, Arizona. *Geology* 33: 713–716.
- Cronin TM, et al. (2008) Impacts of post-glacial lake drainage events and revised chronology of the Champlain Sea episode 13–9 ka. *Paleogeography, Paleoclimatology, Paleocology* 262: 46–60.
- Davis PT, Menounos B, and Osborn G (2009) Holocene and latest Pleistocene alpine glacier fluctuations: A global perspective. *Quaternary Science Reviews* 28: 2021–2033.
- de Vernal A, Hillaire-Marcel C, and Bilodeau G (1996) Reduced meltwater outflow from the Laurentide ice margin during the Younger Dryas. *Nature* 381: 774–777.
- deMenocal P, Ortiz J, Guilderson T, and Sarnthein M (2000) Coherent high- and low-latitude climate variability during the Holocene warm period. *Science* 288: 2198–2202.
- Dokken TM and Jansen E (1999) Rapid changes in the mechanism of ocean convection during the last glacial period. *Nature* 401: 458–461.
- Dolven JK, Cortese G, and Björklund KR (2002) A high-resolution radiolarian-derived paleotemperature record for the late Pleistocene–Holocene in the Norwegian Sea. *Paleoceanography* 17: <http://dx.doi.org/10.1029/2002PA000780>.
- Dorale JA, et al. (2010) Isotopic evidence for Younger Dryas aridity in the North American midcontinent. *Geology* 38: 519–522.
- Dykoski CA, et al. (2005) A high-resolution, absolute-dated Holocene and deglacial Asian monsoon record from Dongge Cave, China. *Earth and Planetary Science Letters* 233: 71–86.
- Ebbesen H and Hald M (2004) Unstable Younger Dryas climate in the northeast North Atlantic. *Geology* 32: 673–676.
- EPICA Community Members (2006) One-to-one coupling of glacial climate variability in Greenland and Antarctica. *Nature* 444: 195–198.
- Farmer EC, deMenocal PB, and Marchitto TM (2005) Holocene and deglacial ocean temperature variability in the Benguela upwelling region: Implications for low-latitude atmospheric circulation. *Paleoceanography* 20: <http://dx.doi.org/10.1029/2004PA001049>.
- Firestone RB, et al. (2007) Evidence for an extraterrestrial impact 12,900 years ago that contributed to the megafaunal extinctions and the Younger Dryas cooling. *Proceedings of the National Academy of Sciences of the United States of America* 104: 16016–16021.
- Flower BP, Hastings DW, Hill HW, and Quinn TM (2004) Phasing of deglacial warming and Laurentide Ice Sheet meltwater in the Gulf of Mexico. *Geology* 32: 597–600.
- Gasse F, Chalif F, Vincens A, Williams MAJ, and Williamson D (2008) Climatic patterns in equatorial and southern Africa from 30,000 to 10,000 years ago reconstructed from terrestrial and near-shore proxy data. *Quaternary Science Reviews* 27: 2316–2340.
- Genty D, et al. (2006) Timing and dynamics of the last deglaciation from European and North African $\delta^{13}\text{C}$ stalagmite profiles—comparison with Chinese and Southern Hemisphere stalagmites. *Quaternary Science Reviews* 25: 2118–2142.
- Grachev AM and Severinghaus JP (2005) A revised $+10 \pm 4^\circ\text{C}$ magnitude of the abrupt change in Greenland temperature at the Younger Dryas termination using published GISP2 gas isotope data and air thermal diffusion constants. *Quaternary Science Reviews* 24: 513–519.
- Grimm EC, Watts WA, Jacobson GL, Hansen BCS, Almquist HR, and Dieffenbacher-Krall AC (2006) Evidence for warm wet Heinrich events in Florida. *Quaternary Science Reviews* 25: 2197–2211.
- Hajdas I, Bonani G, Moreno PI, and Ariztegui D (2003) Precise radiocarbon dating of late-glacial cooling in mid-latitude South America. *Quaternary Research* 59: 70–78.
- Hartz I and Milthers V (1901) Det senglacie ler i Allerød tegelværksgrav. *Meddelelser Danks Geologisk Foreningen* 8: 31–60.
- Haug GH, Hughen KA, Sigman DM, Peterson LC, and Röhl U (2001) Southward migration of the intertropical convergence zone through the Holocene. *Science* 293: 1304–1308.
- Heiri O, Cremer H, Engels S, Hoek WZ, Peeters W, and Lotter AF (2007) Lateglacial summer temperatures in the northwest European lowlands: A chironomid record from Hijkermeer, the Netherlands. *Quaternary Science Reviews* 26: 2420–2437.
- Hughen KA, Overpeck JT, Peterson LC, and Trumbore S (1996) Rapid climate changes in the tropical Atlantic region during the last deglaciation. *Nature* 380: 51–54.
- Ivy-Ochs S, Kerschner H, Maisch M, Christl M, Kubik PW, and Schlüchter C (2009) Latest Pleistocene and Holocene glacier variations in the European Alps. *Quaternary Science Reviews* 28: 2137–2149.
- Jaeschke A, Rühlemann C, Arz H, Heil G, and Lohmann G (2007) Coupling of millennial-scale changes in sea surface temperature and precipitation off northeastern Brazil with high-latitude climate shifts during the last glacial period. *Paleoceanography* 22: <http://dx.doi.org/10.1029/2006PA001391>.
- Johnson RG and McClure BT (1976) A model for Northern Hemisphere continental ice sheet variation. *Quaternary Research* 6: 973–985.
- Johnson TC, Brown ET, McManus J, Barry S, Barker P, and Gasse F (2002) A high-resolution paleoclimate record spanning the past 25,000 years in southern east Africa. *Science* 296: 113–116.
- Jouzel J, et al. (2007) Orbital and millennial Antarctic climate variability over the past 800,000 years. *Science* 317: 793–796.
- Kaplan MR, Moreno PI, and Rojas M (2008) Glacial dynamics in southernmost South America during Marine Isotope Stage 5e to the Younger Dryas chron: A brief review

- with a focus on cosmogenic nuclide measurements. *Journal of Quaternary Science* 23: 649–658.
- Kaplan MR, et al. (2010) Glacier retreat in New Zealand during the Younger Dryas stadial. *Nature* 467: 194–197.
- Karpuz NC and Jansen E (1992) A high-resolution diatom record of the last deglaciation from the SE Norwegian Sea: Documentation of rapid climatic changes. *Paleoceanography* 7: 499–520.
- Kaufman DS, Anderson RS, Hu FS, Berg E, and Werner A (2010) Evidence for a variable and wet Younger Dryas in southern Alaska. *Quaternary Science Reviews* 29: 1445–1452.
- Keigwin LD and Jones GA (1995) The marine record of deglaciation from the continental margin off Nova Scotia. *Paleoceanography* 10: 973–985.
- Keigwin LD, Sachs JP, Rosenthal Y, and Boyle EA (2005) The 8200 year B.P. event in the slope water system, western subpolar North Atlantic. *Paleoceanography* 20: <http://dx.doi.org/10.1029/2004PA001074>.
- Kelly MA, et al. (2008) A ^{10}Be chronology of lateglacial and Holocene mountain glaciation in the Scoresby Sund region, east Greenland: Implications for seasonality during lateglacial time. *Quaternary Science Reviews* 27: 2273–2282.
- Kienast SS and McKay JL (2001) Sea surface temperatures in the subarctic Northeast Pacific reflect millennial-scale climate oscillations during the last 16 kyr. *Geophysical Research Letters* 28: 1563–1566.
- Kienast M, Steinke S, Stettgen K, and Calvert SE (2001) Synchronous tropical South China Sea SST change and Greenland warming during deglaciation. *Science* 291: 2132–2134.
- Kienast M, et al. (2006) Eastern Pacific cooling and Atlantic overturning circulation during the last deglaciation. *Nature* 443: 846–849.
- Kim J-H, Schneider RR, Müller PJ, and Wefer G (2002) Interhemispheric comparison of deglacial sea-surface temperature patterns in Atlantic eastern boundary currents. *Earth and Planetary Science Letters* 194: 383–393.
- Kubota Y, Kimoto K, Tada R, Oda H, Yokoyama Y, and Matsuzaki H (2010) Variations of East Asian summer monsoon since the last deglaciation based on Mg/Ca and oxygen isotope of planktic foraminifera in the northern East China Sea. *Paleoceanography* 25: <http://dx.doi.org/10.1029/2009PA001891>.
- Lamy F, Kaiser J, Jinnemann U, Hebbeln D, Arz HW, and Stoner J (2004) Antarctic timing of surface water changes off Chile and Patagonian Ice Sheet response. *Science* 304: 1959–1962.
- Lea DW, Pak DK, Peterson LC, and Hughen KA (2003) Synchronicity of tropical and high-latitude Atlantic temperatures over the last glacial termination. *Science* 301: 1361–1364.
- Lea DW, Pak DK, Belanger CL, Spero HJ, Hall MA, and Shackleton NJ (2006) Paleoclimate history of Galápagos surface waters over the last 135,000 yr. *Quaternary Science Reviews* 25: 1152–1167.
- Levi C, et al. (2007) Low-latitude hydrological cycle and rapid climate changes during the last deglaciation. *Geochemistry, Geophysics, Geosystems* 8: <http://dx.doi.org/10.1029/2006GC001514>.
- Licciardi JM, Kurz MD, and Curtice JM (2007) Glacial and volcanic history of Icelandic table mountains from cosmogenic ^3He exposure ages. *Quaternary Science Reviews* 26: 1529–1546.
- Linsley BK, Rosenthal Y, and Oppo DW (2010) Holocene evolution of the Indonesian throughflow and the western Pacific warm pool. *Nature Geoscience* 3: 578–583.
- Liu Z, et al. (2009) Transient simulation of last deglaciation with a new mechanism for Bølling–Allerød warming. *Science* 325: 310–314.
- Lowell TV, Larson GJ, Hughes JD, and Denton GH (1999) Age verification of the Lake Gribben forest bed and the Younger Dryas advance of the Laurentide Ice Sheet. *Canadian Journal of Earth Sciences* 36: 383–393.
- Lowell TV, Fisher TG, Hajdas I, Glover K, Loope H, and Henry T (2009) Radiocarbon deglaciation chronology of the Thunder Bay, Ontario area and implications for ice sheet retreat patterns. *Quaternary Science Reviews* 28: 1597–1607.
- MacDonald GM, et al. (2008) Evidence of temperature depression and hydrological variations in the eastern Sierra Nevada during the Younger Dryas stage. *Quaternary Research* 70: 131–140.
- Mangerud J, Andersen ST, Berglund BE, and Donner JJ (1974) Quaternary stratigraphy of Norden, a proposal for terminology and classification. *Boreas* 3: 109–128.
- Marchitto TM, Lehman SJ, Ortiz JD, Flückiger J, and van Geen A (2007) Marine radiocarbon evidence for the mechanism of deglacial atmospheric CO_2 rise. *Science* 316: 1456–1459.
- Mathewes RW, Heusser LE, and Patterson RT (1993) Evidence for a Younger Dryas-like cooling event on the British Columbia coast. *Geology* 21: 101–104.
- McManus JF, Francois R, Gherardi J-M, Keigwin LD, and Brown-Leger S (2004) Collapse and rapid resumption of Atlantic meridional circulation linked to deglacial climate changes. *Nature* 428: 834–837.
- Meissner KJ and Clark PU (2006) Impact of floods versus routing events on the thermohaline circulation. *Geophysical Research Letters* 33: <http://dx.doi.org/10.1029/2006GL026705>.
- Monnin E, et al. (2001) Atmospheric CO_2 concentrations of the last glacial termination. *Science* 291: 112–114.
- Moreno PI, et al. (2009) Renewed glacial activity during the Antarctic cold reversal and persistence of cold conditions until 11.5 ka in southwestern Patagonia. *Geology* 37: 375–378.
- Morley DW, Leng MJ, Mackay AW, and Sloane HJ (2005) Late glacial and Holocene environmental change in the Lake Baikal region documented by oxygen isotopes from diatom silica. *Global and Planetary Change* 46: 221–233.
- Murton JB, Bateman MD, Dallimore SR, Teller JT, and Yang Z (2010) Identification of Younger Dryas outburst flood path from Lake Agassiz to the Arctic Ocean. *Nature* 464: 740–743.
- Nakagawa T, et al. (2003) Asynchronous climate changes in the North Atlantic and Japan during the last termination. *Science* 299: 688–691.
- Nesje A (2009) Latest Pleistocene and Holocene alpine glacier fluctuation in Scandinavia. *Quaternary Science Reviews* 28: 2119–2136.
- Newnham RM and Lowe DJ (2000) Fine-resolution pollen record of late-glacial climate reversal from New Zealand. *Geology* 28: 759–762.
- Nørgaard-Pedersen N, Spielhagen RF, Erlenkeuser H, Grootes PM, Heinemeier J, and Knies J (2003) Arctic Ocean during the Last Glacial Maximum: Atlantic and polar domains of surface water mass distribution and ice cover. *Paleoceanography* 18: <http://dx.doi.org/10.1029/2002PA000781>.
- Otto-Bliessner BL and Brady EC (2009) The sensitivity of the climate response to the magnitude and location of freshwater forcing: Last glacial maximum experiments. *Quaternary Science Reviews* 29: 56–73.
- Pahnke K and Sachs JP (2006) Sea surface temperatures of southern midlatitudes 0–160 kyr B.P. *Paleoceanography* 21: <http://dx.doi.org/10.1029/2005PA001191>.
- Pahnke K, Goldstein SL, and Hemming SR (2008) Abrupt changes in Antarctic Intermediate Water circulation over the past 25,000 years. *Nature Geoscience* 1: 870–874.
- Peltier WR, Vettoretti G, and Stastna M (2006) Atlantic meridional overturning and climate response to Arctic Ocean freshening. *Geophysical Research Letters* 33: <http://dx.doi.org/10.1029/2005GL025251>.
- Pigati JS, Bright JE, Shanahan TM, and Mahan SA (2009) Late Pleistocene paleohydrology near the boundary of the Sonoran and Chihuahuan Deserts, southeastern Arizona, USA. *Quaternary Science Reviews* 28: 286–300.
- Polyack VJ, Rasmussen JBT, and Asmerom Y (2004) Prolonged wet period in the southwestern United States through the Younger Dryas. *Geology* 32: 5–8.
- Praetorius SK, McManus JF, Oppo DW, and Curry WB (2008) Episodic reductions in bottom-water currents since the last ice age. *Nature Geoscience* 1: 449–452.
- Putnam AE, et al. (2010) Glacier advance in southern middle-latitudes during the Antarctic Cold Reversal. *Nature Geoscience* 3: 700–704.
- Rashid H, Flower BP, Poore RZ, and Quinn TM (2007) A ~25 ka Indian Ocean monsoon variability record from the Andaman Sea. *Quaternary Science Reviews* 26: 2586–2597.
- Rasmussen SO, et al. (2006) A new Greenland ice core chronology for the last glacial termination. *Journal of Geophysical Research* 111: <http://dx.doi.org/10.1029/2005JD006079>.
- Roberts NL, Piotrowski AM, McManus JF, and Keigwin LD (2010) Synchronous deglacial overturning and water mass source changes. *Science* 327: 75–78.
- Robinson LF, et al. (2005) Radiocarbon variability in the western North Atlantic during the last deglaciation. *Science* 310: 1469–1473.
- Rodbell DT and Seltzer GO (2000) Rapid ice margin fluctuations during the Younger Dryas in the tropical Andes. *Quaternary Research* 54: 328–338.
- Rooth C (1982) Hydrology and ocean circulation. *Progress in Oceanography* 11: 131–149.
- Rosenthal Y, Oppo DW, and Linsley BK (2003) The amplitude and phasing of climate change during the last deglaciation in the Sulu Sea, western equatorial Pacific. *Geophysical Research Letters* 30: <http://dx.doi.org/10.1029/2002GL016612>.
- Rühlemann C, Mulitz S, Müller PJ, Wefer G, and Zahn R (1999) Warming of the tropical Atlantic Ocean and slowdown of thermohaline circulation during the last deglaciation. *Nature* 402: 511–514.
- Saher MH, Jung SJA, Elderfield H, Greaves MJ, and Kroon D (2007) Sea surface temperatures of the western Arabian Sea during the last deglaciation. *Paleoceanography* 22: <http://dx.doi.org/10.1029/2006PA001292>.
- Schmidt MW, Spero HJ, and Lea DW (2004) Links between salinity variation in the Caribbean and North Atlantic thermohaline circulation. *Nature* 428: 160–163.
- Schulz H, von Rad U, and Erlenkeuser H (1998) Correlation between Arabian Sea and Greenland climate oscillations of the past 110,000 years. *Nature* 393: 54–57.
- Sea DS and Whitlock C (1995) Postglacial vegetation and climate of the Cascade Range, central Oregon. *Quaternary Research* 43: 370–381.
- Severinghaus JP, Sowers T, Brook EJ, Alley RB, and Bender BL (1998) Timing of abrupt climate change at the end of the Younger Dryas interval from thermally fractionated gases in polar ice. *Nature* 391: 141–146.

- Shakun JD and Carlson AE (2010) A global perspective on Last Glacial Maximum to Holocene climate change. *Quaternary Science Reviews* 29: 1801–1816.
- Shakun JD, Burns SJ, Fleitmann D, Kramers J, Matter A, and Al-Subary A (2007) A high-resolution, absolute-dated deglacial speleothem record of Indian Ocean climate from Socotra Island, Yemen. *Earth and Planetary Science Letters* 259: 442–456.
- Schefuß E, Schouten S, and Schneider RR (2005) Climatic controls on central African hydrology during the past 20,000 years. *Nature* 437: 1003–1006.
- Shuman B, Bartlein PJ, and Webb T (2005) The magnitudes of millennial- and orbital-scale climatic change in eastern North America during the late Quaternary. *Quaternary Science Reviews* 24: 2194–2206.
- Sinha A, et al. (2005) Variability of southwest Indian summer monsoon precipitation during the Bølling–Allerød. *Geology* 33: 813–816.
- Steffensen J, et al. (2008) High-resolution Greenland ice core data show abrupt climate change happens in a few years. *Science* 321: 680–684.
- Stott L, Timmermann A, and Thunell R (2007) Southern Hemisphere and deep-sea warming led deglacial atmospheric CO₂ rise and tropical warming. *Science* 318: 435–438.
- Stouffer RJ, et al. (2006) Investigating the causes of the response of the thermohaline circulation to past and future climate changes. *Journal of Climate* 19: 1365–1387.
- Sun Y, Oppo DW, Xiang R, Liu W, and Gao S (2005) Last deglaciation in the Okinawa Trough: Subtropical northwest Pacific link to Northern Hemisphere and tropical climate. *Paleoceanography* 20: <http://dx.doi.org/10.1029/2004PA001061>.
- Tarasov L and Peltier WR (2005) Arctic freshwater forcing of the Younger Dryas cold reversal. *Nature* 435: 662–665.
- Telford RJ (2006) Limitations of dinoflagellate cyst transfer functions. *Quaternary Science Reviews* 25: 1375–1382.
- Thompson LG, et al. (1998) A 25,000-year tropical climate history from Bolivian ice cores. *Science* 282: 1858–1864.
- Tierney JE and Russell JM (2007) Abrupt climate change in southeast tropical Africa influenced by monsoon variability and ITCZ migration. *Geophysical Research Letters* 34: <http://dx.doi.org/10.1029/2007GL029508>.
- Tierney JE, Russell JM, Huang Y, Sinninghe Damste JS, Hopmans EC, and Cohen AS (2008) Northern Hemisphere controls on tropical southeast African climate during the past 60,000 years. *Science* 322: 252–255.
- Turney CSM, McGlone MS, and Wilmshurst JM (2003) Asynchronous climate change between New Zealand and the North Atlantic during the last deglaciation. *Geology* 31: 223–226.
- Vacco DA, Clark PU, Mix AC, Chang H, and Edwards RL (2005) A speleothem record of Younger Dryas cooling, Klamath Mountains, Oregon, USA. *Quaternary Research* 64: 249–256.
- von Grafenstein U, Erlenkeuser H, Brauer A, Jouzel J, and Johnsen SJ (1999) A mid-European decadal isotope-climate record from 15,500 to 5000 years B.P. *Science* 284: 1654–1657.
- Wagner JDM, Cole JE, Beck JW, Patchett PJ, Henderson GM, and Barnett HR (2010) Moisture variability in the southwestern United States linked to abrupt glacial climate change. *Nature Geoscience* 3: 110–113.
- Wang YJ, et al. (2001) A high-resolution absolute-dated late Pleistocene monsoon record from Hulu Cave, China. *Science* 294: 2345–2348.
- Wang X, et al. (2007) Millennial-scale precipitation changes in southern Brazil over the past 90,000 years. *Geophysical Research Letters* 34: <http://dx.doi.org/10.1029/2007GL031149>.
- Weijers JWH, Schefuß E, Schouten S, and Sinninghe Damste JS (2007) Coupled thermal and hydrological evolution of tropical Africa over the last deglaciation. *Science* 315: 1701–1704.
- Weldeab S, Schneider RR, and Kölling M (2006) Deglacial sea surface temperature and salinity increase in the western tropical Atlantic in synchrony with high latitude climate instabilities. *Earth and Planetary Science Letters* 241: 699–706.
- Weldeab S, Lea DW, Schneider RR, and Andersen N (2007) 155,000 years of west African monsoon and ocean thermal evolution. *Science* 316: 1303–1307.
- Williams KM (1993) Ice sheet and ocean interactions, margin of the east Greenland Ice Sheet (14 ka to present): Diatom evidence. *Paleoceanography* 8: 69–83.
- Williams PW, King DNT, Zhao J-X, and Collerson KD (2005) Late Pleistocene to Holocene composite speleothem ¹⁸O and ¹³C chronologies from South Island, New Zealand – Did a global Younger Dryas really exist? *Earth and Planetary Science Letters* 230: 301–317.
- Yancheva G, et al. (2007) Influence of the intertropical convergence zone on the east Asian monsoon. *Nature* 445: 74–77.
- Yu Z and Eicher U (1998) Abrupt climate oscillations during the last deglaciation in central North America. *Science* 282: 2235–2238.
- Zhao M, Beveridge NAS, Shackleton NJ, Sarnthein M, and Eglinton G (1995) Molecular stratigraphy of cores off northwest Africa: Sea surface temperature history over the last 80 ka. *Paleoceanography* 10: 661–675.

PALEOCLIMATE MODELING

Contents

Data–Model Comparisons

Quaternary Environments

The Last Interglacial

Last Glacial Maximum GCMs

Paleo-ENSO

Data–Model Comparisons

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Introduction

Data–model comparison is the shorthand term used to describe the process of comparing output from climate or Earth-system model simulations of past time periods with paleoenvironmental observations. Such comparisons have two distinct purposes: at a fundamental level, the data–model comparison process leads to an understanding of the mechanisms of past climate and environmental changes; at a practical level, data–model comparisons can be used to evaluate how well state-of-the-art models work.

Paleodata record how environmental conditions have changed in response to changes in climate. However, these data cannot provide an unambiguous explanation of which aspects of the climate have changed or why they have changed. This is because the systems that record past changes are influenced by multiple aspects of the climate and, furthermore, generally display nonlinear responses to changes in specific climate parameters. In addition, the local climate regime itself is governed by multiple controls and a change in regime can be brought about in many different ways. Consequently, there are very many different pathways that could lead to the same response in a paleoclimatic or paleoenvironmental indicator.

To give a concrete example ([Figure 1](#)), pollen records and plant macrofossil data show that the northern treeline (i.e., the limit between tundra and taiga vegetation) was further north than today in Europe and Siberia during the mid-Holocene (ca. 6 ka). The location of the northern treeline is primarily controlled by the accumulated warmth of the growing season (usually expressed in terms of growing degree days, or the number of days with temperature higher than a given threshold multiplied by the temperature of each day). Increased growing season warmth can result from a warmer summer or from higher temperatures in spring and/or autumn resulting in

a longer growing season. Changes in the seasonal distribution of precipitation can also influence the length of the growing season: for example, a reduction in winter and spring snowfall will lead to a more rapid warming in spring and, hence, an earlier onset of the growing season. However, the growth of trees is also controlled by water availability. An increase in precipitation can allow trees to colonize areas from which they were previously excluded. Thus, an observed change in the position of the tundra–taiga boundary could have been caused by changes in many different aspects of the local climate regime. Similarly, the change in the local climate regime could have been brought about in many ways. For example, increased temperature during the growing season could be a response to a change in the seasonal cycle of incoming solar radiation (insolation), or it could result from an increased advection of warm air from adjacent oceans.

Models based on physical principles, or well-documented empirical representations of these physical principles, can provide mechanistic explanations of past climate and environmental changes ([Bartlein and Hostetler, 2004](#)). In order to provide such an explanation, the model must be shown to work under a wide range of different conditions, and it must explicitly include all the relevant components of the system implicated in a particular climate change. Furthermore, the model experiment must be appropriately designed to incorporate the key mechanisms involved. Thus, it would be possible to explore the causes of the northward shift in the tundra–taiga boundary during the mid-Holocene, for example, by using a fully coupled ocean–atmosphere–vegetation model, provided that the model was able to reproduce the position of the northern treeline today and during the mid-Holocene, and provided that the experimental design included factors such as the extent of ice sheets that were not explicitly simulated by the model.

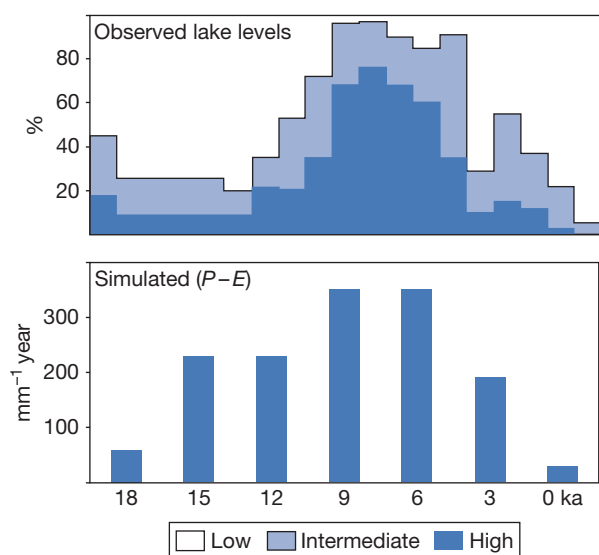


Figure 2 Observed lake levels in northern Africa during the past 18 ka (on the radiocarbon timescale), derived from the Global Lake Status Database, compared to change in runoff (precipitation minus evaporation, $P - E$) simulated by an atmospheric general circulation model (AGCM) driven by changes in orbital forcing and glacial boundary conditions. These simulations and the comparison with observed lake levels were made as part of the COHMAP project and show that observed climate changes in northern Africa during the last glacial–interglacial cycle were caused by orbitally driven changes in the African monsoon. Adapted from Kutzbach JE and Street-Perrott FA (1985) Milankovitch forcing of fluctuations in the level of tropical lakes from 18 to 0 ka BP. *Nature* 317: 130–134.

climate changes could be achieved only through the exercise of global climate models in conjunction with large-scale (preferably global) syntheses of paleoenvironmental data. COHMAP largely focused on pollen and plant macrofossil data of past vegetation changes, and geomorphic and biostratigraphic evidence of past changes in the level or extent of lakes, to provide a picture of regional changes in environmental conditions primarily caused by changes in regional climates.

The most famous example of the COHMAP approach is the now universally accepted explanation for formerly wetter conditions in areas of northern Africa that are desert today (Kutzbach and Street-Perrott, 1985). A wealth of archaeological, geomorphological, pollen, and lake-level data show that northern Africa was considerably wetter than today during the early and mid-Holocene (ca. 11–5.5 ka). Experiments with an atmospheric general circulation model (AGCM) showed that both the broad-scale spatial patterns and the temporal evolution of these patterns could be explained to first order as a consequence of orbital (Milankovitch) forcing: gradual changes in the seasonal and latitudinal distribution of insolation through the Holocene caused a transition from a state with greater land–sea temperature contrast and an expanded and intensified African monsoon toward the aridity seen today (Figure 2).

COHMAP focused on using data–model comparisons for an improved understanding of the mechanisms of climate change, and this idea was subsequently taken up by many other modeling groups. However, it was rapidly realized that

simulations of the same time period made by different modeling groups and using different models differed from one another. It was difficult to diagnose the cause of these differences because the simulations did not always use the same external forcing or internal boundary conditions. The need to understand the causes of intermodel differences was the underlying motivation for the establishment of the Paleoclimate Modeling Intercomparison Project (PMIP) (Joussaume et al., 1999) in the early 1990s. The aim of this international project, which was sponsored by the World Climate Research Programme (WCRP) and the International Geosphere–Biosphere Programme (IGBP), was to analyze the causes of the differences in the responses of different models when run with identical forcing and boundary conditions. PMIP initially focused on two time periods: the Last Glacial Maximum (LGM) (ca. 21 ka) and the mid-Holocene (ca. 6 ka). These two intervals were chosen because they represent extremes of climate forcing (Table 1). As a result of changes in the configuration of Earth’s orbit, specifically in the phasing of the precession and obliquity cycles, Northern Hemisphere insolation was substantially greater than today during boreal summer (and on annual average) during the first part of the Holocene. At the LGM, the Earth’s orbital configuration was similar to that today but greenhouse gas concentrations were low, Northern Hemisphere ice sheets were greatly expanded and sea level was lower, the ocean surface was colder, and there was more sea ice in winter.

Through systematic model–model comparisons, PMIP was able to show that some features of the simulated climate changes were robust and independent of the model used, and to diagnose the causes of differences between the climates simulated by individual models. However, PMIP aimed to go further than this: to determine whether climate models are able to simulate past climates correctly through extensive evaluation using paleoclimate and paleoenvironmental data sets. The existence of adequate benchmarks for model evaluation is a ‘given’ for the evaluation of climate models under present-day conditions. However, the accurate simulation of modern climate is not a sufficient guarantee that a model will correctly simulate climatic conditions different from those today, such as those expected as a consequence of the continuing increase in greenhouse gas emissions during the twenty-first century. By evaluating the LGM and mid-Holocene simulations against standard paleoclimate benchmarks, PMIP was able to show how a comprehensive set of models responds to large changes in forcing. These evaluations were used in the third assessment report of the Intergovernmental Panel on Climate Change (IPCC) (Figure 3; McAvaney et al., 2001). PMIP is continuing the systematic benchmarking of the models used to predict future climate changes.

Requirements for Data Syntheses for Data–Model Comparison

Large-scale data syntheses are required to document the diagnostic spatial patterns (or ‘fingerprint’) of specific climate-change mechanisms (Anderson, 1995; Kohfeld and Harrison, 2000). Maps presenting a view of the world at some key time in the past on the basis of the interpretation of selected data in the

Table 1 Boundary conditions used for mid-Holocene (6 ka BP) and Last Glacial Maximum (21 ka BP) experiments by the Paleoclimate Modeling Intercomparison Project

Boundary conditions	Phase 1, AGCM			Phase 1, AGCM + mixed-layer ocean		Phase 2, OAGCM			Phase 2, OAVGCM		
	Modern	6 ka BP	21 ka BP	Modern	21 ka BP	Modern	6 ka BP	21 ka BP	Modern	6 ka BP	21 ka BP
Sea-surface temperatures	Existing control	As modern simulation	CLIMAP (1981)	Existing control	Computed	Computed	Computed	Computed	Computed	Computed	Computed
Sea-ice cover	Existing control	As modern simulation	CLIMAP (1981)	Existing control	Computed	Computed	Computed	Computed	Computed	Computed	Computed
Ice-free land albedo	Existing control	As modern simulation	No change	Existing control	As modern simulation	Modern	As modern simulation	As modern simulation	Computed	Computed	Computed
Ice sheet extent and height	Existing control	As modern simulation	ICE 4G (Peltier, 1994)	Existing control	ICE 4G (Peltier, 1994)	Modern	No change	ICE 5G (Peltier, 2005)	Modern	As modern simulation	ICE 5G (Peltier, 2005)
Topography and coastline	Existing control	As modern simulation	−105 m	Existing control	−105 m	Modern	No change	ICE 5G (Peltier, 2005)	Modern	As modern simulation	ICE 5G (Peltier, 2005)
CO ₂	Existing control or 345 ppm	280 ppm or 280/345 existing control value	200 ppm or 200/345 existing control value	Existing control or 345 ppm	200 ppm or 200/345 existing control value	280 ppm	280 ppm	185 ppm	280 ppm	280 ppm	185 ppm
CH ₄	Not considered					760 ppb	650 ppb	350 ppb	760 ppb	650 ppb	350 ppb
N ₂ O	Not considered					270 ppb	270 ppb	200 ppb	270 ppb	270 ppb	200 ppb
CFC	Not considered					0	0	0	0	0	0
O ₃	Not considered					Modern-10DU	As modern simulation	As modern simulation	Modern-10DU	As modern simulation	As modern simulation
Solar constant	Existing control or 1365 W m ^{−2}	As modern simulation	As modern simulation	Existing control or 1365 W m ^{−2}	As modern simulation	1365 W m ^{−2}	1365 W m ^{−2}	1365 W m ^{−2}	1365 W m ^{−2}	1365 W m ^{−2}	1365 W m ^{−2}
Eccentricity	0.016724	0.01682	0.018994	0.016724	0.018994	0.016724	0.01682	0.018994	0.016724	0.01682	0.018994
Axial tilt	23.446	24.105	22.949	23.446	22.949	23.446	24.105	22.949	23.446	24.105	22.949
Angular Precession	102.04	0.87	114.42	102.04	114.42	102.04	0.87	114.42	102.04	0.87	114.42

<http://www-1sce.cea.fr>.

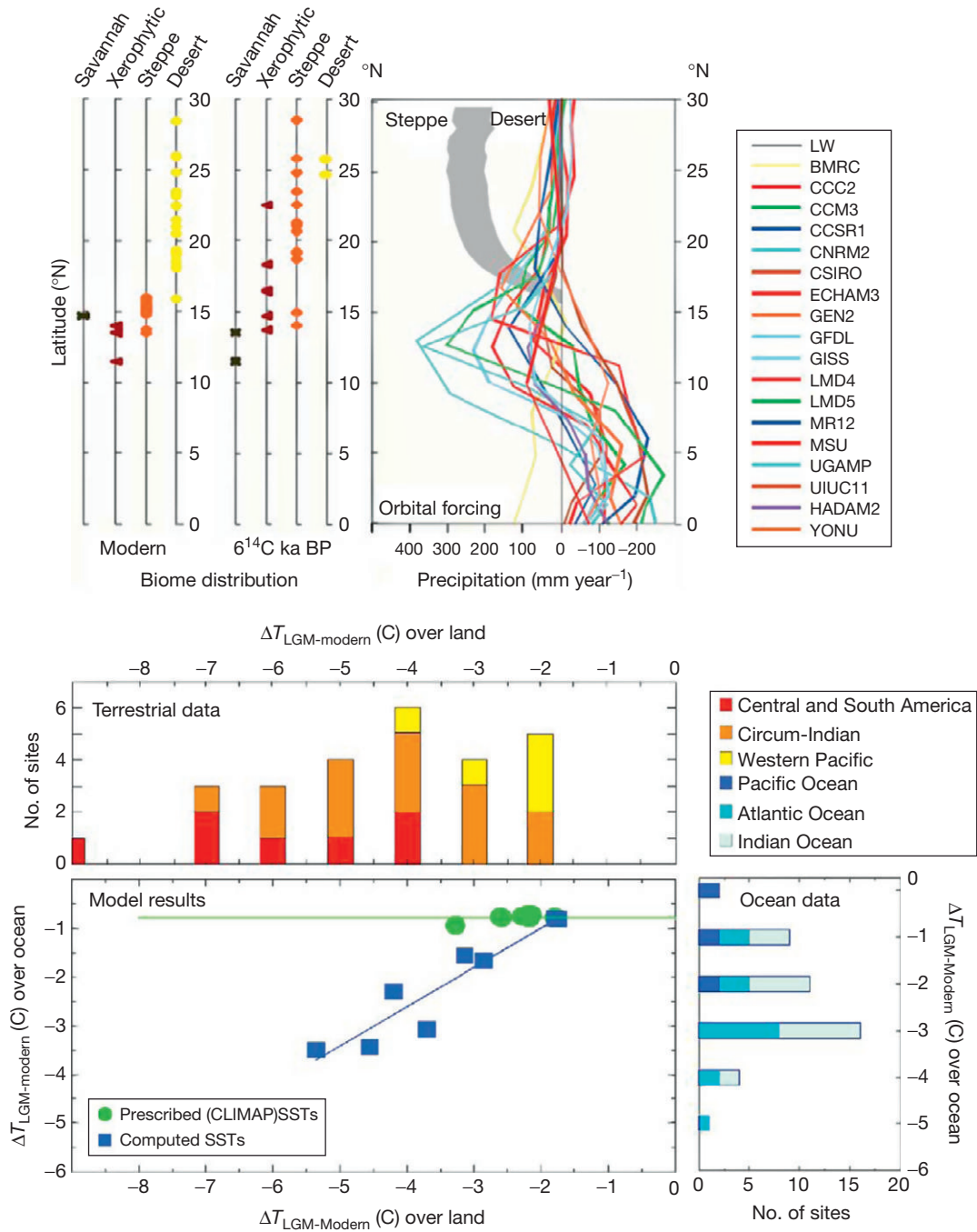


Figure 3 Examples of model evaluation using paleoclimate data for the mid-Holocene and Last Glacial Maximum (LGM) from the IPCC Third Assessment (McAvaney et al., 2001). The upper panel shows the latitudinal distribution of biomes across northern Africa today and at 6 ka BP (data derived from the BIOME 6000 dataset); the gray shaded area on the middle panel shows the increase in precipitation required at each latitude to cause a shift in vegetation from desert to grassland (steppe) and compares this to the changes in precipitation at each latitude simulated by the suite of AGCMs taking part in PMIP. The lower panel shows reconstructions of the magnitude of tropical cooling over land (data derived from the 21 ka TROPICS data set) and over the ocean (data derived from alkenone records) at the LGM and compares these reconstructions with the simulated change shown by multiple AGCMs and AGCMs coupled to a simple mixed-layer ocean model.

light of existing paradigms have long been part of the toolkit of Quaternary science. However, these maps are not suitable for systematic data–model comparisons or for model benchmarking. The creation of archival databases, such as the Global Pollen Database (GPD) (www.ncdc.noaa.gov), in the 1990s

represented a major step forward by ensuring access to large amounts of primary data and establishing full traceability for paleoreconstructions. Nevertheless, there is still a considerable amount of work required to produce data sets for model evaluation from these archive resources.

Data sets for model evaluation should have the following characteristics:

1. They should have a large enough spatial extent to distinguish regional patterns or ‘fingerprints.’ In practice, this means that the data set should be continental or ocean-basin-wide at a minimum, and preferably should be global in extent.
2. They should be transparent with respect to the primary data. Specifically, the data should be registered at individual sites and not spatially generalized in any way. This approach avoids the dual dangers of interpolating a signal on the basis of an inadequate conceptual model and of hiding the paucity of information for specific regions.
3. They should contain data that can be precisely located in time, and the time window should be limited in extent. In practice, the width of the time window that is considered acceptable depends on the magnitude and likely persistence of the climate signal being evaluated: given that the glacial state persisted for thousands of years and the difference between this state and the state today is large, it is generally considered acceptable to amalgamate data for a window of ± 1 ka around the glacial maximum for evaluation of glacial-type simulations. This approach would not be acceptable, however, if the aim of the model experiments was to evaluate the causes of climate oscillations within the glacial interval.
4. The data should be in a form that is directly comparable with model output. Initially, data sets for model comparison were focused on meteorological variables (e.g., July temperature, mean annual precipitation, and wind direction). The existence of offline models of components of the climate system, such as models of the marine and terrestrial biosphere or the dust cycle, means that a wide range of paleoenvironmental variables (e.g., vegetation type and dust-deposition rates) can now be used for model comparisons. The trend of using non-traditional variables for data–model comparison is likely to increase, as general circulation models evolve into full Earth-system models. However, the use of such variables places an additional constraint on the data syntheses, namely, that the controlling processes that give rise to particular kinds of records are well understood and made explicit in the data–model comparison process.
5. The data should be accompanied by adequate metadata (i.e., data describing the site, the record, and the age model) to allow users to select data for specific model comparisons.

Table 2 Large-scale and global data syntheses for model evaluation

<i>Data set</i>	<i>Variable</i>	<i>Region covered</i>	<i>Periods covered</i>	<i>Reference</i>	<i>Source</i>
GLSDB	Lake-level changes (runoff)	Global	Continuous from ca. 30 ka BP to present	Kohfeld and Harrison (2000)	Global dataset: BRIDGE website; Maps and data for selected time periods: PMIP2 website
BIOME 6000	Major vegetation types based on pollen and plant macrofossil data	Global	Modern, 6 and 21 ka BP	Prentice et al. (2000) and Pickett et al. (2004)	Maps and data for selected time periods: PMIP2 website
PAIN	Major vegetation types, based on pollen and plant macrofossil data	Northern high latitudes (N of 55° N)	Modern, 6 and 21 ka BP	Bigelow et al. (2003)	Maps and data for selected time periods: BRIDGE website
DIRTMAP	Mineral dust-deposition rates	Global, but predominately from ocean and ice-core sites	Past 150 ka, by marine isotope stages	Kohfeld and Harrison (2001)	Maps and data for selected time periods: BRIDGE website
SNOWLINE	Equilibrium line altitude of valley glaciers	Tropical (33° N–33° S)	Last Glacial Maximum, plus later glacial advance phases	Mark et al. (2005)	Maps and data for selected time periods: BRIDGE website
21ka TROPICS	Mean temperature of the coldest month, mean annual temperature	Tropical (33° N–33° S)	21 $\pm$ 1 ka BP	Farrera et al. (1999)	Global dataset: BRIDGE website; Maps and data: PMIP2 website
American monsoon	Multiproxy estimate of changes in effective moisture	North America and northern South America	Mid-Holocene (ca. 6 ka BP)	Harrison et al. (2003)	Maps and data: BRIDGE website
MARGO, LGM SSTs	Multiproxy estimates of mean annual SST	Global	21 $\pm$ 1 ka BP	Kucera et al. (2005)	Data sets: PANGAEA website
Holocene SSTs	Coccolith-based reconstructions of SST	Global	Holocene	Bollmann et al. (2002)	Data sets: NOAA NGD website
Marine productivity	Multiproxy estimates of marine productivity	Global	Marine isotope stages 5 through present	Kohfeld et al. (2005)	Maps and data: BRIDGE website

Data sources cited in the table are: BRIDGE website address: <http://www.bridge.bris.ac.uk>; PMIP2 website address: <http://www-1sce.cea.fr>; PANGAEA website address: <http://www.pangaea.de>; NOAA NGDC website address: <http://www.ncdc.noaa.gov>.

6. The data set should be adequately documented and publicly accessible in order to ensure full traceability of the data–model comparisons.

Requirements for Data–Model Comparisons

Early attempts to compare paleoenvironmental observations and model simulations were largely nonsystematic and subjective. This situation reflected the fact that the observations were reported on a site-by-site basis, and relatively few modelers had a command over the widely dispersed literature describing these records. However, the development of community-sponsored syntheses of different kinds of data means that it is now possible for comparisons to be made in a less *ad hoc* fashion. It is now widely accepted that data–model comparisons should fulfill certain basic requirements to be meaningful.

1. Data–model comparisons should be systematic and involve comparison with all of the available data from a region for a given time period. This avoids the situation of comparing model output with results from a few sites which might not be representative of the regional pattern of climate change.
2. Data–model comparisons should concern explicit model variables that can be matched to the controls on the paleo-observations. Thus, it is important to compare, for example, simulated July temperature with reconstructions based on proxies that respond explicitly to July temperature

rather than to some generalized requirement for warmth. In a similar way, comparisons of simulated changes in short-term variability in the El Niño/La Niña–Southern Oscillation (ENSO) signal cannot be evaluated by consideration of records that reflect changes in mean precipitation in ENSO-dominated regions.

3. They should be expressed in a way that makes it possible to evaluate matches/mismatches in an objective and quantitative way.
4. They should take into account the inherent variability of both the observed and simulated quantities being compared. In effect, this means taking into account the uncertainties in the paleoenvironmental reconstruction and the interannual variability in the simulated mean climate change.

Requirements for Model Benchmarking

The concept of ‘benchmarking’ is firmly entrenched in the procedure for evaluating performance in many areas of modeling, although the use of paleoclimatic and/or paleoenvironmental benchmarks for evaluating climate or Earth-system models is comparatively recent. The process of benchmarking simulations is based on the existence of a robust reconstruction of past climate or environmental changes, which can be used to measure the match to observations as models evolve. Thus, the requirements for benchmarks are somewhat different

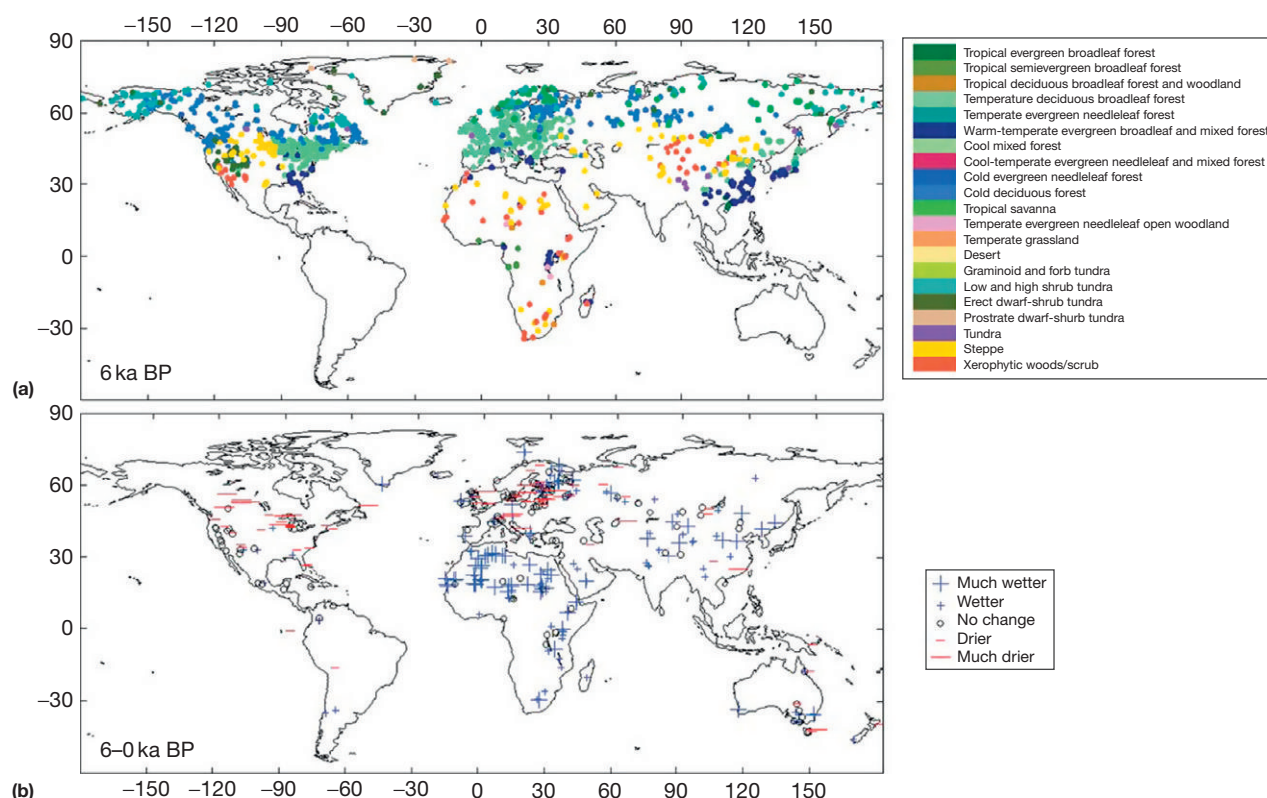


Figure 4 The world in the mid-Holocene (6 ka). (a) Reconstructed vegetation patterns as shown by data from the BIOME 6000 data set (Prentice et al., 2000), and (b) changes in lake status, a measure of the regional change in runoff, as shown by data from the Global Lake Status Database (Kohfeld and Harrison, 2000).

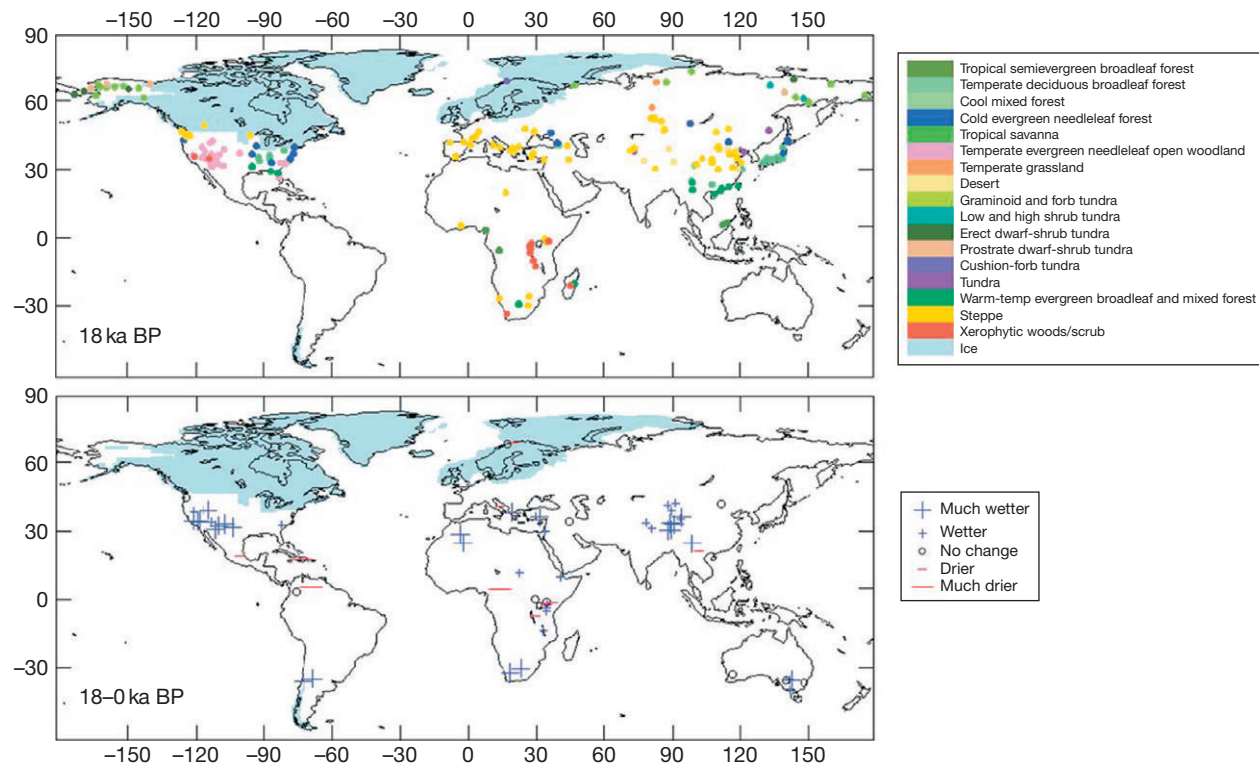


Figure 5 The world at the LGM (21 ka, ca. 18 ka on the radiocarbon timescale). Reconstructed vegetation patterns as shown by data from the BIOME 6000 data set (Prentice et al., 2000) and changes in lake status, a measure of the regional change in runoff, as shown by data from the Global Lake Status Database (Kohfeld and Harrison, 2000). The location of the Northern Hemisphere ice sheets is based on the ICE 5G data set (Peltier, 2004).

from the requirements for data syntheses for data–model comparison. Broadly speaking, benchmarks must

1. reflect large-scale patterns corresponding to first-order changes in the climate system;
2. be robust with respect to multiple data sources and reconstruction techniques;
3. be adequately supported by data, and not be based on single records;
4. be expressed in a way that makes objective diagnosis possible.

The observed change in the African monsoon during the mid-Holocene has provided one such benchmark, which has been widely used for model evaluation (see below). PMIP has also developed a set of other regional benchmarks, which are now being used systematically for model evaluation.

Existing Paleoenvironmental Data sets for Data–Model Comparison

There have been a number of community-wide exercises to create paleoclimate or paleoenvironmental data sets for model evaluation (see Table 2). Many of these data sets are of regional scope. The most widely used global data sets are those provided by the Paleovegetation Mapping Project (BIOME 6000) (Prentice et al., 2000) and its daughter projects (e.g., Pan-Arctic Initiative (PAIN): Bigelow et al., 2003), the Global Lake Status Database (GLSDB) (Kohfeld and Harrison,

2000), the DIRTMAP database (Kohfeld and Harrison, 2001), the LGM Tropical Terrestrial Data Synthesis (LGM TROPICS) data set (Farrera et al., 1999), the Snowline Database (SNOWLINE) (Mark et al., 2005), and the Multiproxy approach for the Reconstruction of the Glacial Ocean (MARGO) project (Kucera et al., 2005). These various data sets present a picture of a world radically different from what it is today.

Paleoenvironmental data for the mid-Holocene indicate that the most pronounced changes in climate occurred in the region dominated by the Afro-Asian monsoon and in the high northern latitudes. Data from the GLSDB, for example, show that northern Africa between 15° and 30° N was occupied by large lakes and extensive wetlands (Figure 4(a)). The BIOME 6000 data set (Figure 4(b)) shows that the transition from grassland to desert, which marks the Sahel/Sahara boundary, was at least 5° further north than today. True desert vegetation seems to have been confined to a relatively small area in the northeastern part of the modern Sahara. Although the changes in Asia were more muted, lake, vegetation, and eolian data from this region all indicate an increase in the area affected by monsoon rainfall. In the high northern latitudes, vegetation reconstructions show that the tundra–taiga boundary was further north than it is today in Fennoscandia and central Siberia, although there was little change in other regions, and the tree-line was south of its present position in eastern Canada.

At the LGM, paleoenvironmental data show that regional climates were colder and generally drier than today (Figure 5). Forests were absent (or at least highly restricted) north of 55° N and the landscape was dominated by treeless vegetation, with

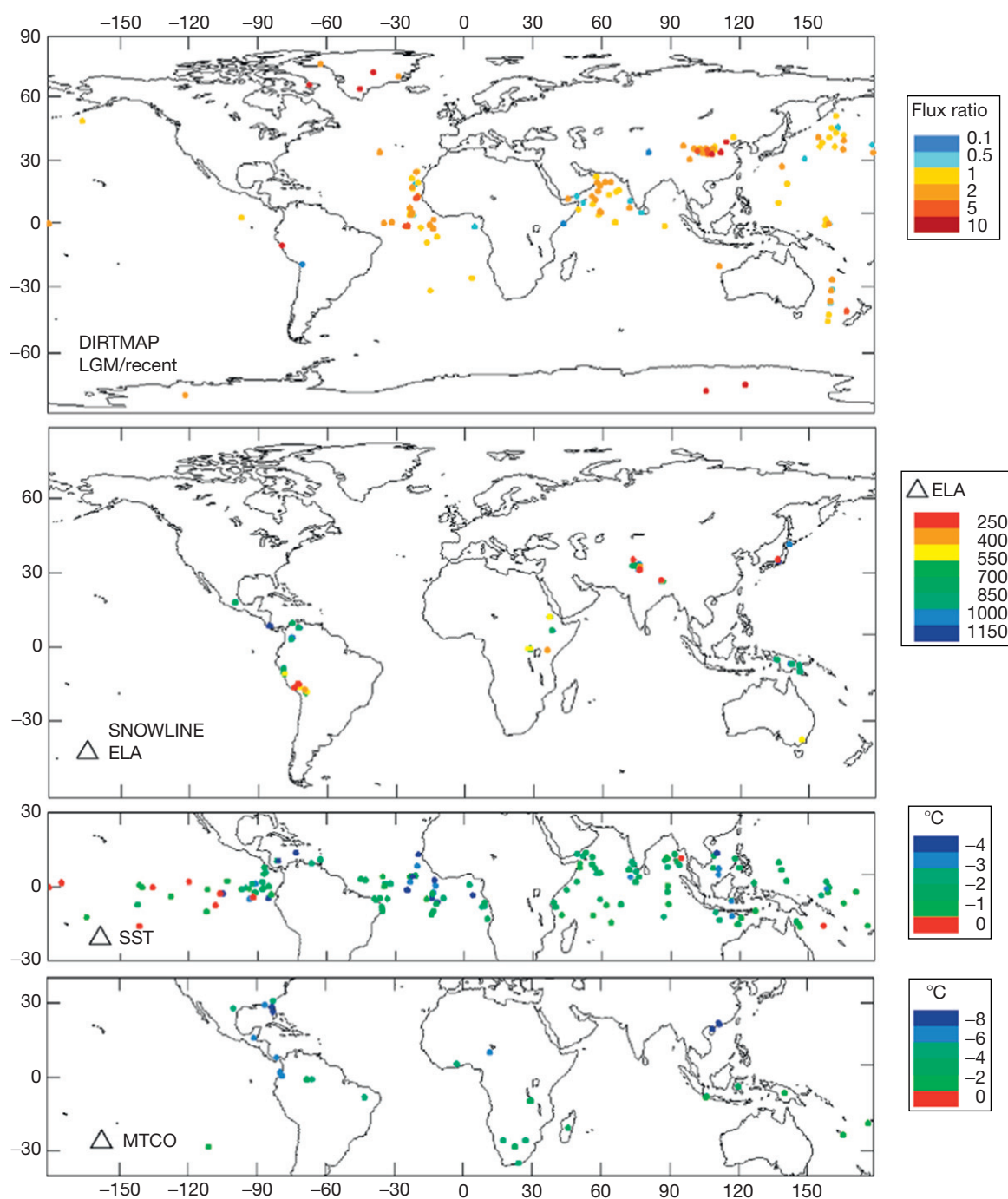


Figure 6 The tropics at the LGM (21 ka, ca. 18 ka on the radiocarbon timescale). The uppermost panel shows changes in dust deposition, as recorded in marine and ice-core records and by loess deposits on the continents, from the DIRTMAP database (Kohfeld and Harrison, 2001). The second panel shows changes in the equilibrium line altitude (Δ ELA) of tropical glaciers from the SNOWLINE database (Mark et al., 2005). The lower panels show reconstructed changes in mean annual sea-surface temperature from the MARGO project (Kucera et al., 2005) and reconstructed changes in mean temperature of the coldest month (MTCO) from the 21 ka TROPICS database (Farrera et al., 1999).

drier conditions giving rise to a significant expansion of graminoid and forb tundra, grassland, and xerophytic shrubland in northern and central Eurasia. Both lake and vegetation data indicate wetter conditions in the American Southwest, circum-Mediterranean region, and western China, most likely in response to the southward displacement of the westerlies caused

by the expanded northern ice sheets. The tropics were also colder and drier than today (Figure 6). Land-based temperature estimates from the LGM TROPICS data set indicate that the tropical lowlands were on average 2.5–3 °C cooler than today. These records show that the magnitude of the cooling differed regionally, with changes of only 1–2 °C in the

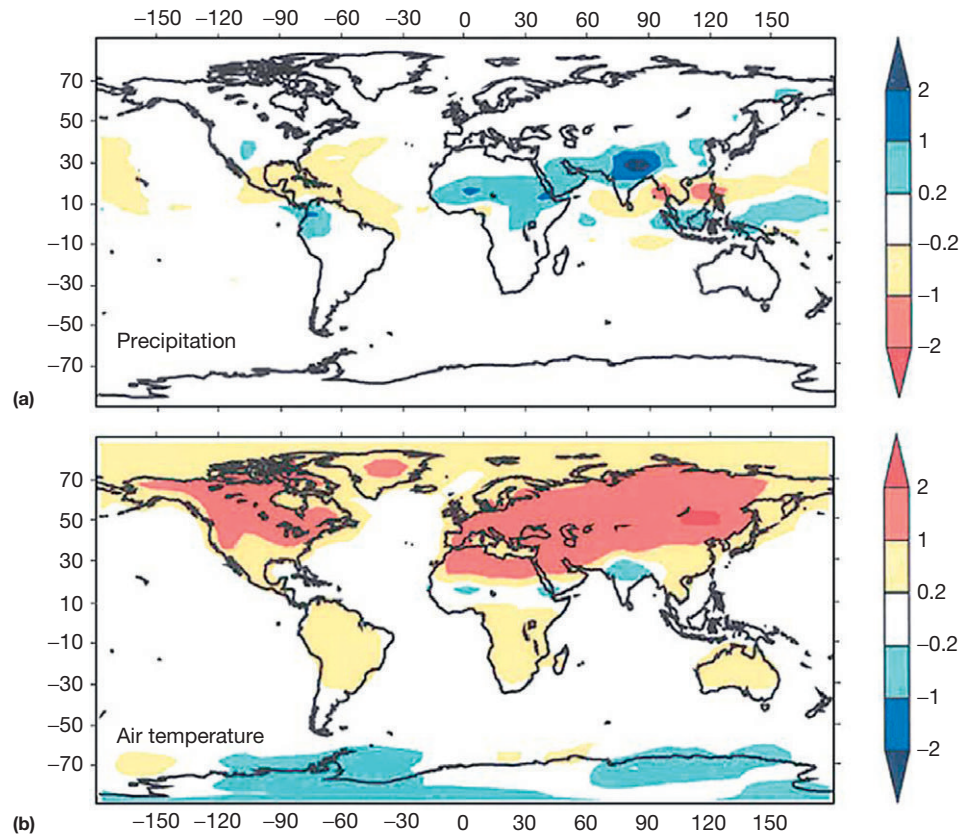


Figure 7 Changes (a) precipitation and (b) air temperature during the summer as a consequence of orbitally induced changes in incoming solar radiation (insolation) at 6 ka BP, based on a composite of the results of 17 AGCM simulations made within the first phase of the PMIP project.

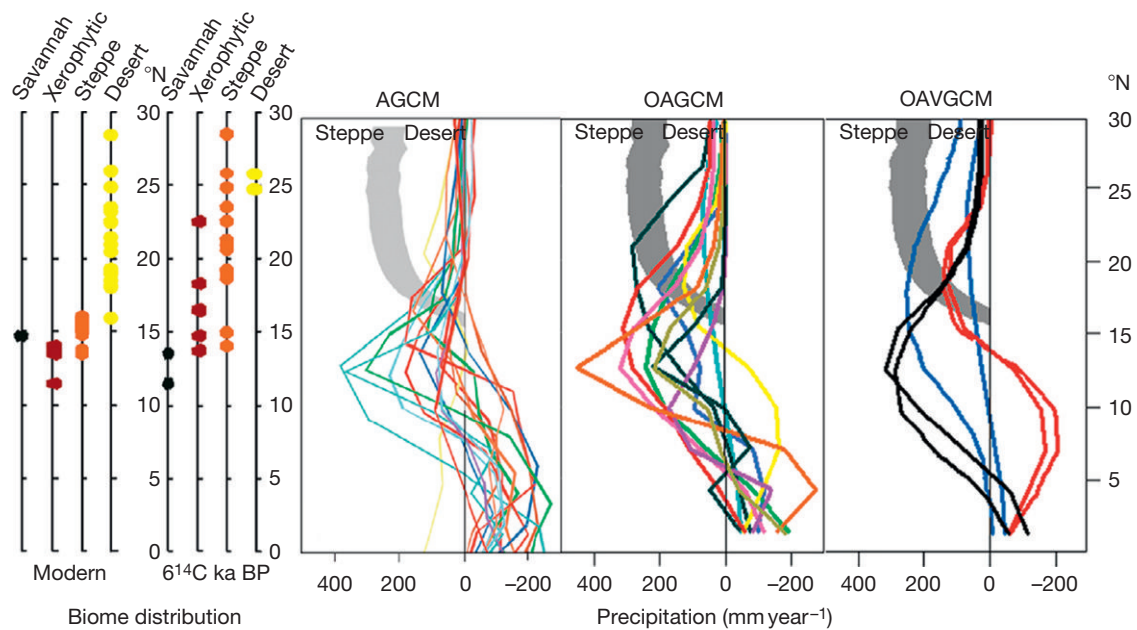


Figure 8 Benchmarking model improvement in the simulation of the mid-Holocene expansion of the northern African monsoon as a consequence of including ocean and land-surface feedbacks. The first panel shows the results of simulations made with an atmosphere-only general circulation model (AGCM) during the first phase of PMIP; results from coupled ocean–atmosphere general circulation models (OAGCMs) and coupled ocean–atmosphere–vegetation models (OAVGCMs) run in the second phase of PMIP are shown in the middle and right-hand panels. This diagram shows that the incorporation of ocean- and land-surface feedbacks results in a more realistic amplification of the African monsoon, although there are considerable differences in the responses of individual models.

circum-Pacific lowlands and changes of 5–6 °C in Central and South America. At higher elevations, snowline data also indicate considerable regional differentiation with only muted changes in climate in the Himalayas, southern Andes, and East Africa, but much more pronounced lowering of snowline in the northern Andes, Central America, and Papua New Guinea. The MARGO data sets also show considerable spatial patterning in the magnitude of the cooling of the tropical oceans. Reconstructed sea-surface temperatures (SSTs) in the Caribbean, the South China Sea, and in the tropical Atlantic off the western coast of northern Africa were >5 °C lower than present, consistent with terrestrial reconstructions from the Caribbean and South China. The cooling in the equatorial belt was more muted: in the western Pacific Ocean, reconstructed changes in mean annual temperature are <2 °C while changes in the Indian Ocean and in the southern part of the tropical Atlantic are 2–4 °C. This patterning is also consistent with the terrestrial reconstructions from adjacent regions.

The Mid-Holocene African Monsoon: An Example of Data–Model Comparisons and Benchmarking

The expansion of the area influenced by the African monsoon is one of the most striking features of the mid-Holocene record (Figure 4) and has therefore figured prominently in data–model comparisons. It also provides an iconic example of the use of paleodata for benchmarking. The data show that at 6 ka most of the modern-day Sahara was covered by large lakes, extensive wetlands, and grassland. The precipitation required to generate the observed latitudinal grassland distribution has been estimated using both statistical methods and a simple vegetation–soil water-balance model. The observed shift from desert to grassland at 20° N would require an increase of 200–300 mm in mean annual precipitation (Joussaume et al., 1999). Independent calculations of the change in runoff required to maintain the observed lakes yield a similar estimate

of the change in mean annual precipitation. Thus, the mid-Holocene expansion of the African monsoon meets the suitability criteria for benchmarking in that the change in spatial pattern shown by the paleoenvironmental data reflects a major climate-change signal; it is based on a large number of individual site observations, it is robust with respect to two independent sources of evidence, and it can be expressed quantitatively in terms of a variable (i.e., mean annual precipitation) that is directly available from model simulations. Furthermore, as the paleoenvironmental records are drawn from global data sets (BIOME 6000 and the GLSDB, respectively) that conform to the requirements for model evaluation, this benchmark is well documented and fully traceable.

In the first phase of PMIP, comparisons with the observed expansion of the African monsoon were used to evaluate the response of a suite of 17 AGCMs to mid-Holocene changes in orbital forcing. All the models showed an increase and a northward shift in monsoon precipitation (Figure 7(a)) in response to the substantial heating of the northern continents (Figure 7(b)), thus increasing the land–sea temperature contrast. Initial data–model comparisons used mapped patterns of lake-level changes from the GLSDB to demonstrate that the simulated runoff ($P-E$) did not increase sufficiently in the northern Sahara. Subsequently, output from each of the climate models was used to drive an offline vegetation model to determine whether the simulated change in climate could reproduce the changes in vegetation patterns shown by data from the BIOME 6000 database. Finally, a systematic comparison of the model simulations against the monsoon benchmark was made. This comparison (Figure 8) showed that although all these models showed an increase and a northward shift in monsoon precipitation, they all underestimated the precipitation required to maintain grassland between 20° and 23° N by at least 150 mm. As a result of complex feedbacks associated with orbitally induced changes in SST (Kutzbach et al., 2001), coupled ocean–atmosphere general circulation models produce a more

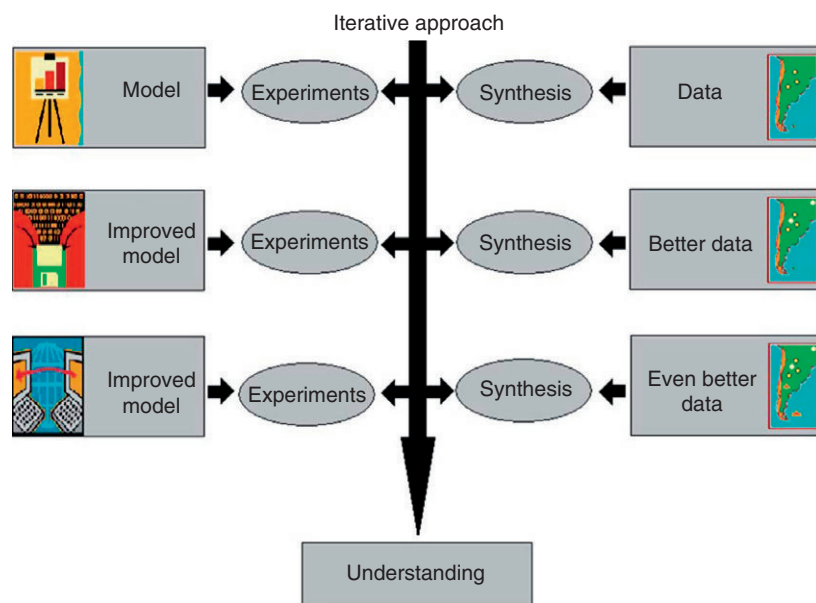


Figure 9 Illustration of the iterative process of data–model comparisons, showing how the data–model comparison approach leads to a constant improvement in the quality and coverage of data syntheses and in Earth-system modeling capacity.

substantial enhancement of the African monsoon than atmosphere-only models. Nevertheless, only a few of the models produce sufficient rainfall to maintain grassland at 20° N, and all the models underestimate precipitation further north by at least 100 mm. Feedbacks associated with changes in land-surface characteristics as a result of changes in vegetation cover can also result in a substantial enhancement of the African monsoon. Preliminary benchmarking of the relatively few coupled ocean–atmosphere–vegetation simulations that have been run for the mid-Holocene confirm this enhancement.

Data–Model Comparisons: An Iterative Process

The confrontation of models and data is an iterative process (Figure 9). Apparent disagreements between model simulations and paleoenvironmental observations can arise from many causes including inadequate and/or unrepresentative data coverage, imperfect understanding of the climatic significance of the observations, incorrect specification of the climate forcing, or inadequacies in the climate model itself. Attempts to deal with the first two of these problems have led to the progressive improvement of the data sets for data–model comparisons, the development of objective techniques for reconstructing climate variables from the data, and the increasing use of models to simulate the response of environmental sensors to climate change. Each of these improvements gives rise to a new phase of data–model comparison. Persistent disagreement between model simulations and well-documented phenomena in the paleorecord has led to the discovery of new climatic mechanisms and to model improvement. A classic example once again involves the African monsoon. Once it was recognized that AGCMs systematically underestimate the quantitative increase in precipitation in the present-day Sahara in mid-Holocene times, scientists noted that the observed presence of large lakes and moisture-demanding vegetation in this region would have led to significant changes in land-surface albedo and surface roughness. Experiments made by prescribing altered mid-Holocene land-surface conditions within a climate model showed that land-surface feedbacks produce a substantial amplification of the orbitally driven changes in the African monsoon. The recognition that changes in land-surface conditions were potentially important feedbacks on the African monsoon has paved the way for the recent trend toward the inclusion of vegetation as an interactive component in climate models.

See also: **Glacial Climates:** Biosphere Feedbacks. **Glaciation, Causes:** Astronomical Theory of Paleoclimates. **Paleoclimate Modeling:** Last Glacial Maximum GCMs.

References

- Alverson KD, Bradley RS, and Pedersen TF (2003) *Paleoclimate, Global Change and the Future*, p. 220. Berlin: Springer-Verlag.
- Alverson KD, Oldfield F, and Bradley RS (2000) Past global changes and their significance for the future. *Quaternary Sciences Reviews* 19: 1–479 (special issue).
- Anderson DM (1995) *Global Paleoenvironmental Data. PAGES Workshop Report Series 95-2*, p. 271. Bern: PAGES IPO.
- Bartlein PJ and Hostetler SW (2004) Modeling paleoclimates. In: Gillespie AR, Porter SC, and Atwater BF (eds.) *The Quaternary Period in the United States: Developments in Quaternary Science*, pp. 563–582. Amsterdam: Elsevier.
- Bigelow NH, Brubaker LB, Edwards ME, et al. (2003) Climatic change and Arctic ecosystems. I. Vegetation changes north of 55° N between the Last Glacial Maximum, mid-Holocene, and present. *Journal of Geophysical Research-Atmosphere* 108(D19): 8170. <http://dx.doi.org/10.1029/2002JD002558>.
- Bollmann J, Henderiks J, and Brabec B (2002) Global calibration of Gephyrocapsa coccolith abundance in Holocene sediments for paleotemperature assessment. *Paleoceanography* 17(3): 1035. <http://dx.doi.org/10.1029/2001PA000742>.
- COHMAP Members (1988) Climatic changes of the last 18,000 years: Observations and model simulations. *Science* 241: 1043–1052.
- Farrera I, Harrison SP, Prentice IC, et al. (1999) Tropical climates at the Last Glacial Maximum: A new synthesis of terrestrial palaeoclimate data. I. Vegetation, lake-levels and geochemistry. *Climate Dynamics* 15: 823–856.
- Harrison SP, Kutzbach JE, Liu Z, et al. (2003) Mid-Holocene climates of the Americas: A dynamical response to changed seasonality. *Climate Dynamics* 20: 663–688. <http://dx.doi.org/10.1007/s00382-002-0300-6>.
- Joussaume S, Taylor KE, Braconnot P, et al. (1999) Monsoon changes for 6000 years ago: Results of 18 simulations from the Palaeoclimate Modeling Intercomparison Project (PMIP). *Geophysical Research Letters* 26: 859–862.
- Kohfeld KE and Harrison SP (2000) How well can we simulate past climates? Evaluating the models using global paleoenvironmental data sets. *Quaternary Science Reviews* 19: 321–346.
- Kohfeld KE and Harrison SP (2001) DIRTMAP: The geological record of dust. *Earth-Science Reviews* 54: 81–114.
- Kohfeld KE, Le Quéré C, Harrison SP, and Anderson RF (2005) Role of marine biology in glacial–interglacial CO₂ cycles. *Science* 308: 74–78.
- Kucera M, Rosell-Melé A, Schneider R, Waelbroeck C, and Weinelt M (2005) Multiproxy approach for the reconstruction of the glacial ocean surface (MARGO). *Quaternary Science Reviews* 24: 813–819.
- Kutzbach JE, Harrison SP, and Coe MT (2001) Land–ocean–atmosphere interactions and monsoon climate change: A paleo-perspective. In: Schulze ED, Heimann M, and Harrison SP, et al. (eds.) *Global Biogeochemical Cycles in the Climate System*, pp. 73–86. San Diego, CA: Academic Press.
- Kutzbach JE and Street-Perrott FA (1985) Milankovitch forcing of fluctuations in the level of tropical lakes from 18 to 0 ka BP. *Nature* 317: 130–134.
- Mark B, Harrison SP, Spessa A, New A, Evans D, and Helmens K (2005) Snowline changes at the LGM: A global synthesis and evaluation. *Quaternary International* 138–139: 168–201.
- McAveney BJ, Covey C, Joussaume S, et al. (2001) Model evaluation. In: Houghton JT, Ding Y, and Griggs DJ, et al. (eds.) *Climate Change 2001: The Scientific Basis. Contribution of Working Group I to the Third Assessment Report of the Intergovernmental Panel on Climate Change*, p. 881. Cambridge: Cambridge University Press.
- McGuffie K and Henderson-Sellers A (2005) *A Climate Modelling Primer*, 3rd edn. London: Wiley.
- Peltier WR (1994) Ice age paleotopography. *Science* 265: 195–201.
- Peltier WR (2004) Global glacial isostasy and the surface of the ice-age Earth: The ICE-5G (VM2) model and GRACE. *Annual Review of Earth and Planetary Sciences* 32: 111–149.
- Peltier WR (2005) On the hemispheric origins of meltwater pulse 1a. *Quaternary Science Reviews* 24: 1655–1671.
- Pickett EJ, Harrison SP, Hope G, et al. (2004) Pollen-based reconstructions of biome distributions for Australia, Southeast Asia and the Pacific (SEAPAC region) at 0, 6000 and 18,000 14C yr B.P. *Journal of Biogeography* 31: 1381–1444 ISSN: 0305-0270 (paper), 1365-2699 (on-line).
- Prentice IC and Jolly D (2000) BIOME 6000 Participants: Mid-Holocene and glacial-maximum vegetation geography of the northern continents and Africa. *Journal of Biogeography* 27: 507–519.
- Washington WM and Parkinson CL (2005) *Introduction to Three-Dimensional Climate Modelling*, 2nd edn. Herndon, VA: University Science Books.
- Wright HE Jr., Kutzbach JE, Webb T III, Ruddiman WF, Street-Perrott FA, and Bartlein PJ (eds.) (1993) *Global Climates since the Last Glacial Maximum*. Minneapolis: University of Minnesota Press.

Relevant Websites

- <http://www.bridge.bris.ac.uk> – Bristol Research Initiative for Dynamic Global Environments.
- <http://www-lsce.cea.fr/pmip/2> – Palaeoclimate Modelling Intercomparison Project.
- <http://www.pangaea.de> – Publishing Network for Geoscientific & Environmental Data.
- <http://www.ncdc.noaa.gov/paleo/paleo.html> – World Data Centre A: Paleoclimatology.

Quaternary Environments

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Introduction

The modeling of Quaternary climate provides a unique opportunity to test our understanding of the balance of processes that maintain climate and its variability. Despite all that can be gleaned from the decades of instrumental data that exist on recent climate history, these data do not reveal the significant interactions between the atmosphere, ocean, cryosphere, and biosphere as are suggested by Quaternary climate proxy records. One of the most pressing questions about climate today is how these components of the climate system will respond as a whole to the projected increases in atmospheric carbon dioxide concentrations. The observation and modeling of Quaternary climate has the potential to provide the necessary insights, yet there still is significant progress that is needed before this becomes a reality.

There exists a vast array of models of varying complexity that have been applied to the study of Quaternary climate. Computational limitations provide a strong constraint on the level of detail that may be considered, such as the temporal and spatial resolution or the number of interacting components included in any particular climate-modeling framework. The timescales relating to many of the leading questions about Quaternary climate change, such as glacial–interglacial cycles of ice sheets and the global carbon cycle, are too long to be considered by the same models that are being used to model future climate. To add to the confusion, many of these different modeling approaches do not give consistent answers.

This article discusses the strength and limitations of the many approaches to the modeling of Quaternary climates. A central theme is the struggle to identify and model the balance of processes that are relevant to any given question. A more detailed treatment of climate modeling and other topics relevant to modeling Quaternary climate are provided by McGuffie and Henderson-Sellers (2005), Washington and Parkinson (2005), Bradley (1999), Crowley and North (1996), and Trenberth (1992).

Assessing Model Applicability

Complexity

Climate models are becoming increasingly complex and are providing new opportunities to consider how and why changes in Quaternary climate come about based on first principles. However, understanding physical relationships within a complex model can be as challenging as interpreting the real world, and the computational costs may preclude doing all the experiments that one would wish, in order to extract those relationships. Much of the modeling of Quaternary climate that has taken place up to this point makes use of ‘reduced-order’ climate models. Reduced-order climate models are developed by exploiting some physical understanding of a

system, in order to reduce the number of spatial dimensions or temporal resolution considered, by parameterizing the missing physics or degrees of freedom in terms of quantities that are being resolved. Carefully conceived simple models representing relevant physics and balances are better suited to yielding insights and physical intuition than more complex models, which have the capacity to give realistic results but for the wrong reasons. Thus, the ability to explicitly control balances is one of the principle advantages of reduced-order models.

One of the simplest models that has been applied to the study of paleoclimate is an energy balance model (EBM). The radiative balance between incoming short-wave radiation F_{sw} and outgoing long wave radiation F_{lw} is one of the strongest constraints on climate prediction. A few simple equations can be used to capture the radiative and convective processes that maintain this balance to get an excellent approximation of, for example:

1. the size of the greenhouse effect from the doubling of atmospheric carbon dioxide levels;
2. the strength of the water vapor feedback assuming a constant relative humidity for a given change in saturation vapor pressure; and
3. an answer to why more complex models of future climate are extremely sensitive to uncertainties in the distribution of clouds.

Following a model presented by Goody (1995), the atmospheric lapse rate $\Gamma = -6.5 \text{ K km}^{-1}$ controls the linkage between the surface air temperature T_g , the altitude z_e , and temperature T_e of the layer within the mid-troposphere, called the emission level. This is emitting long wave radiation to space in balance with incoming solar radiation (Figure 1):

$$T_g = T_e + \Gamma z_e \quad [1]$$

The height of the emission level may be related to a unitless measure of the absorptivity of the atmosphere, known as its optical thickness τ ,

$$z_e = H \ln \frac{3\tau_1}{2} \quad [2]$$

where τ_1 is the optical thickness of the whole atmosphere and $H \approx 2 \text{ km}$ provides an approximate e-folding length scale of the decrease in long wave absorptive gas density (principally from water vapor) with height of the atmosphere. Given the fact that atmospheric CO_2 makes up about 12.5% of τ , with most of the remaining portion coming from water vapor for an Earth-like atmosphere where $\tau_1 = 8$, it can be estimated that the direct radiative effect of doubling CO_2 should produce $\sim 1.1^\circ \text{C}$ warming with an additional 4.1°C warming arising from changes in water vapor concentrations that would be predicted assuming that relative humidity remains constant. These results are quantitatively and physically consistent with the response of general circulation models (GCMs). Following

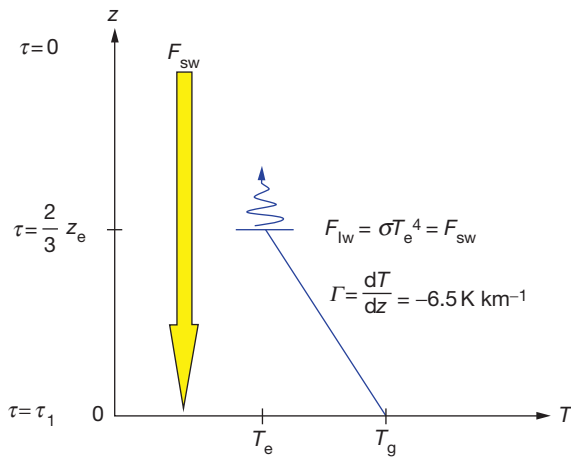


Figure 1 Conceptual greenhouse model of the relationship between surface air temperature T_g and the optical thickness of the atmosphere τ . Temperature is measured in units of degrees kelvin which is equivalent to degrees Celsius + 273.15. Adapted from Figure 5.9 of Goody R (1995) *Principles of Atmospheric Physics and Chemistry*. New York, NY: Oxford University Press, p. 324, with permission.

the example presented in Goody (1995), eqns [1] and [2] can be used along with the knowledge of the radiative properties of clouds, which act to either raise the altitude of the emission level z_e through increases in high clouds or decrease z_e by reflecting more short-wave radiation to space through increases in low clouds. Accordingly, the presence of observed concentrations of high clouds produces a net surface warming of 24 °C, whereas the presence of observed concentrations of low clouds produces a net surface cooling of 63 °C. Even this simple model of the greenhouse effect on climate, which does not take into account the complex physics of cloud formation, illustrates why clouds and water vapor are one of the primary sources of uncertainty in any model prediction of climate, whether relating to the past or the future.

Uncertainties

One of the more curious facts about model predictions of future climate is the surprising level of disagreement that exists among GCMs in their equilibrium surface air temperature sensitivity to a doubling of atmospheric CO_2 concentrations. This is despite their apparent success in reproducing observed surface air temperatures and surface air temperature trends over the last 100 years. The implication is that models are achieving agreement with the instrumental record of climate for different reasons. Differences in the treatment of clouds and radiation are likely to be important. Figure 2 shows observations and a range of model predictions for surface air temperature and percent total cloud cover, as obtained from the second Coupled Model Intercomparison Project (CMIP2). It is remarkable that the 12 models represented in the project achieved such close agreement in their predictions of surface air temperature, despite the extremely large range in predictions of total cloud amount. One of the reasons for the close agreement between the experimental predictions of surface air temperature is that nine of the experiments represented include

so-called ‘heat flux adjustments,’ which are extra sources or sinks of energy within the upper ocean that help models reproduce the observed sea-surface temperatures. Most modeling groups have moved away from the need to use heat flux adjustments, but as there has not been a parallel advancement in the understanding and elimination of sources of model uncertainty, it is assumed that the balances that maintain model climate are resulting from a different cancellation of errors.

Fortunately, there is broad agreement among climate models and climate proxies as to the large-scale qualitative response of the climate system to changes in the Earth’s orbital geometry, atmospheric $p\text{CO}_2$, and continental ice volume – for the particular late Quaternary time slices that have received most of the attention, including features of the last interglacial around 115 ka, Last Glacial Maximum (LGM) around 21 ka, and the mid-Holocene around 6 ka. However, this generalization does not apply in detail or with fields that are particularly sensitive to highly parameterized physics, such as clouds and precipitation.

One of the best ways to make an assessment of the uncertainty that exists for predicting changes in a given field of interest is to consider the range of predictions that exist among multiple climate models. At present, there are more than a few dozen model intercomparison projects that could be used in this way, with the results in Figure 2 being one example. This exercise is particularly effective for the kind of climate change experiments usually done for paleoclimate reconstructions, in which agreement among the models is less likely to be related to artifacts of the model tuning process and more likely to be related to physical processes or sets of processes that are well represented by the models. The series of volumes related to the IPCC assessments of global climate change contains excellent summaries of the current state of knowledge concerning global climate modeling, particularly with GCMs (IPCC, 2001).

Designing Models and Experiments

The climate models that exist for addressing questions of Quaternary climate change include compromises to meet the challenges of the range of processes of potential interest and the longer timescales involved. These compromises have the potential to interfere with reproducing the kind of balance of processes, and therefore sensitivities to external forcing, that exist in nature. A few examples are given here of choices that have been made to deal with the shortcomings of models, so that they may still be useful in addressing Quaternary climate science questions. An important point is that it has never been clear whether a given compromise is well justified, given the importance of the things known. Rather, the science of modeling Quaternary climate progresses as a sequence of tentative steps toward the objective of identifying and quantifying the dominant processes and assessing how these processes respond to a given forcing.

Heat Flux Adjustments

Major efforts in the modeling of Quaternary climate as a system of coupled component models have only recently begun.

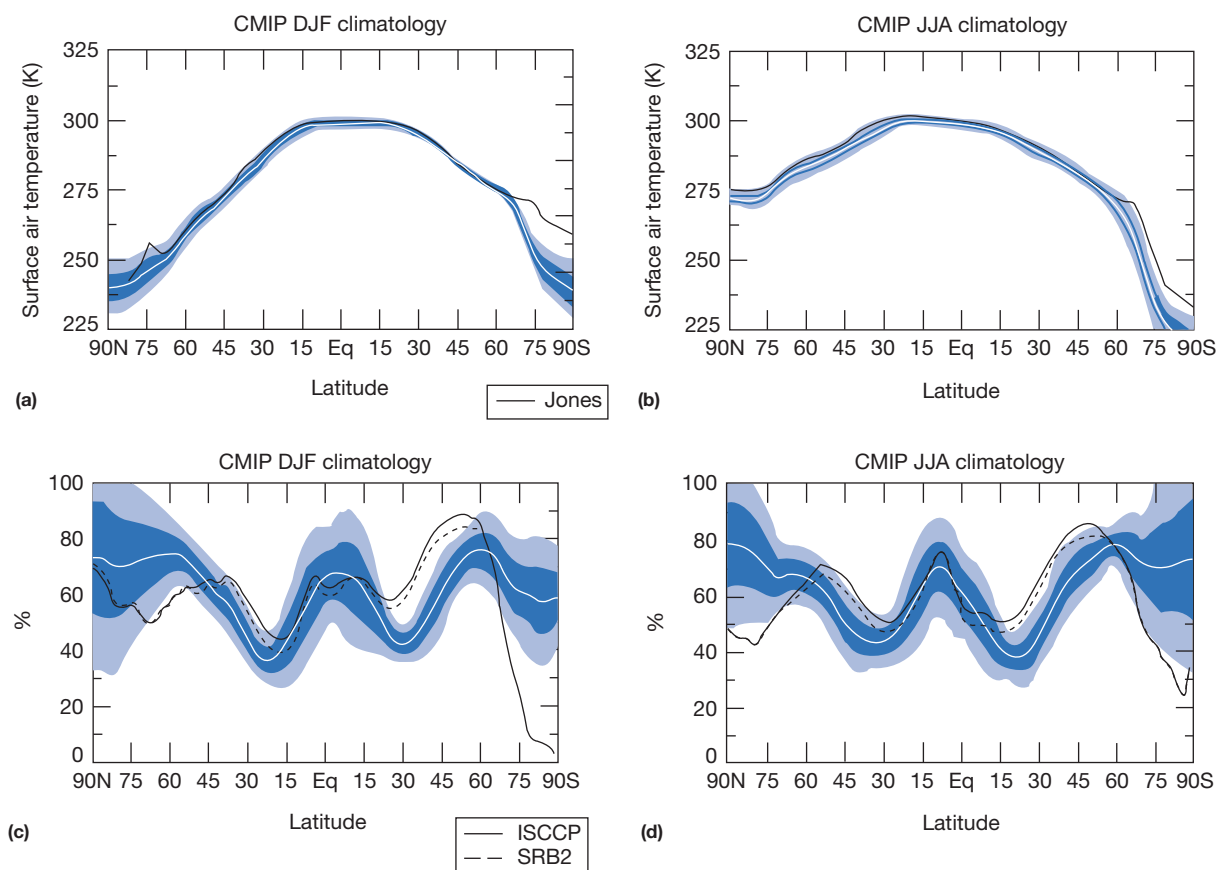


Figure 2 DJF and JJA comparison between observations (solid and dashed lines), and the 1 (dark gray) and 2 (light gray) intermodel standard deviation among 12 coupled atmosphere–ocean general circulation climate models and their predictions of surface air temperature (in degrees kelvin, equal to degrees Celsius + 273.15; (a) and (b)) and percent total cloud cover ((c) and (d)) (AchutaRao et al., 2004).

Until recently (approximately year 2000), the coupling between a model atmosphere and model ocean required additional fluxes of heat energy at the ocean surface (the flux adjustments mentioned above), which are designed to keep coupled models from drifting into an unrealistic state. Most researchers considered it unwise to attempt flux-adjusted coupled atmosphere–ocean model experiments of paleoclimate, because the errors that were so delicately handled for present climate would contribute a significant amount of uncertainty to model predictions of past climates. Models still contain significant biases which, when integrated for a sufficiently long enough time, could undermine a model's usefulness. However, more care is now being given to ensure that a model's global and annual mean energy balances are satisfied. This helps considerably in reducing the chance that an error in model predictions would accumulate unchecked. While the potentially problematic issues of climatic drift are being successfully addressed, many models still may contain significant biases and compensating errors.

Boundary Conditions

Another common practice in experimental design is to specify the boundary conditions of a climate model that one does not have the capacity or interest in modeling. Examples of boundary conditions include sea-surface temperatures, sea-ice extent,

ocean-heat transports, wind stresses, meridional transports of water vapor, vegetation, carbon dioxide concentrations, continental ice sheets, and river runoff. Many of these boundary conditions come from climate proxy data and the effort to provide global coverage for quantities like surface vegetation or the three-dimensional distribution of continental ice sheets is challenging and fraught with uncertainty. The idea is that all of the active components of the climate model would respond in the same way to the fixed boundary conditions as they would if the boundary conditions were predicted by a model. The extent to which this is true is not obvious.

One particular example that has received considerable attention is the reconstruction of sea-surface temperatures and sea-ice extent during the LGM (~21 ka BP), obtained through the CLIMAP project (Climate/Long-Range Investigation, Mapping and Prediction; CLIMAP Project Members, 1976) and updated by MARGO (Multiproxy Approach for the Reconstruction of the Glacial Ocean surface; Kucera et al., 2005). The purpose of modeling experiments that make use of these data as boundary conditions is to test the understanding and interpretation of land-based climate proxy records in terms of atmospheric processes. The lack of agreement between model predictions based on CLIMAP boundary conditions and many climate proxy records in the tropics raised questions about the accuracy of the reconstructed sea-surface temperatures, which

indicate little change from present across vast portions of the subtropics. From a modeling and experimental design point of view, the discrepancy raises a central point in the discussion about the balances that are being represented in models. Specifying sea-surface temperatures influences the sources or sinks of energy that are available to the atmosphere. If a large part of the ocean is kept unnaturally warm with respect to energy fluxes being provided by the atmosphere, the imbalance will shape a portion of the total solution that could affect model sensitivities important to particular scientific questions. Modeling groups differ widely in their willingness to design an experiment with this potential source of uncertainty.

An alternate experimental design that has helped in this regard is an atmospheric GCM that is coupled to a slab ocean of specified thickness. The slab ocean provides a representation of the thermal memory of the upper ocean but does not explicitly consider how upper-ocean temperatures may be affected by changes in ocean circulation or mixing. Ocean-heat transports are represented through heat-flux adjustments that only vary with the seasonal cycle. They are usually designed to provide the exact amount of heat energy required to support the observed climatological sea-surface temperature distributions. Maps of these flux adjustments show largely what would be expected from the effects of ocean currents on the transport of heat from the tropics and its release along the western boundary currents and the high northern latitudes where deep-water formation is taking place (Figure 3). These maps also include fluxes that compensate for biases that may exist in the model atmosphere. The predictions of LGM climate using an atmospheric GCM coupled to a slab-ocean model show a 1–2 °C cooling in the tropics and compare favorably with the 2–3 °C average cooling in experiments that include a dynamic model of the ocean (Figure 4). Currently, there is considerable uncertainty in atmosphere–ocean GCMs as to the rate of deep-water formation in the polar regions. In this case, it is helpful

to know how much of the LGM climate may be reproduced if ocean currents are unchanged. The experimental design of an atmosphere GCM coupled with a slab ocean can provide a helpful complement to the experiments that estimate the aspects of modeled climate change that exist because of atmospheric rather than oceanic processes.

Equilibration

Another source of uncertainty that has arisen with the advent of coupled atmosphere–ocean GCM experiments is the question of model equilibration. Predictions of paleo-ENSO (El Niño/Southern Oscillation) may be particularly sensitive to equilibration, as ENSO dynamics are very sensitive to processes that affect the structure of the thermocline. Stouffer (2004) found that the tropical Pacific thermocline required 200–300 years to attain 70% of its final equilibrium solution when forced by a cooling perturbation (Figure 5). The deep ocean requires the longest equilibration timescales of approximately 2000 years.

Considering many large-scale, and at times rapid, changes in climate throughout the Quaternary, it may have been unusual for the ocean to have ever been in equilibrium with changes in external forcing. This may add considerably to the uncertainties of modeling Quaternary climate. More needs to be done to quantify the potential problems this may create for model-data comparisons.

Examples

Stability of the meridional overturning circulation

The question of the ocean's response to high-latitude freshening is prominent in Quaternary climate-modeling studies because of the likelihood of sudden changes in continental ice sheet mass balance, particularly in connection with episodes of

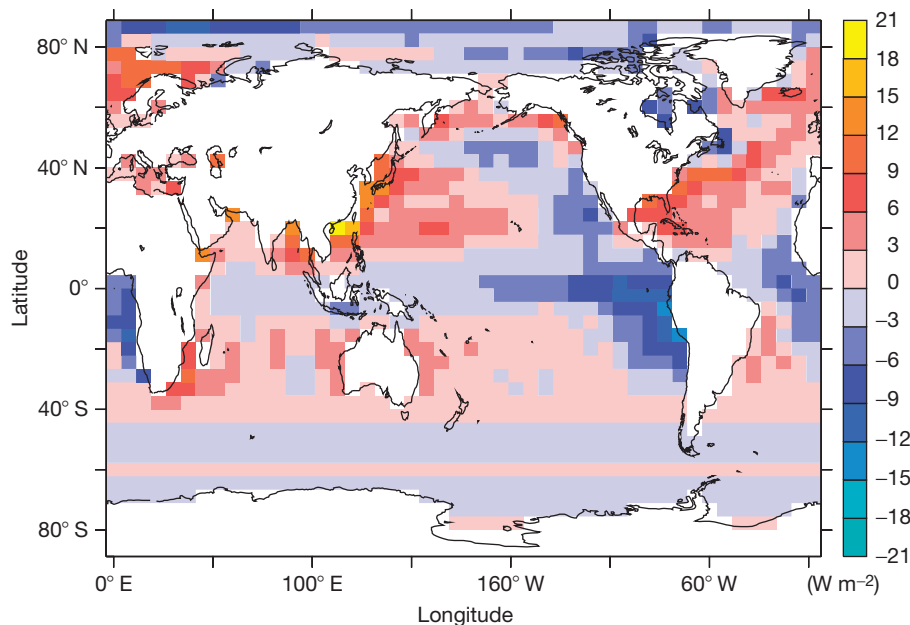


Figure 3 Example of annual mean heat flux adjustments applied to an atmospheric general circulation model (GCM) coupled to a 50-m slab ocean (Jackson and Broccoli, 2003).

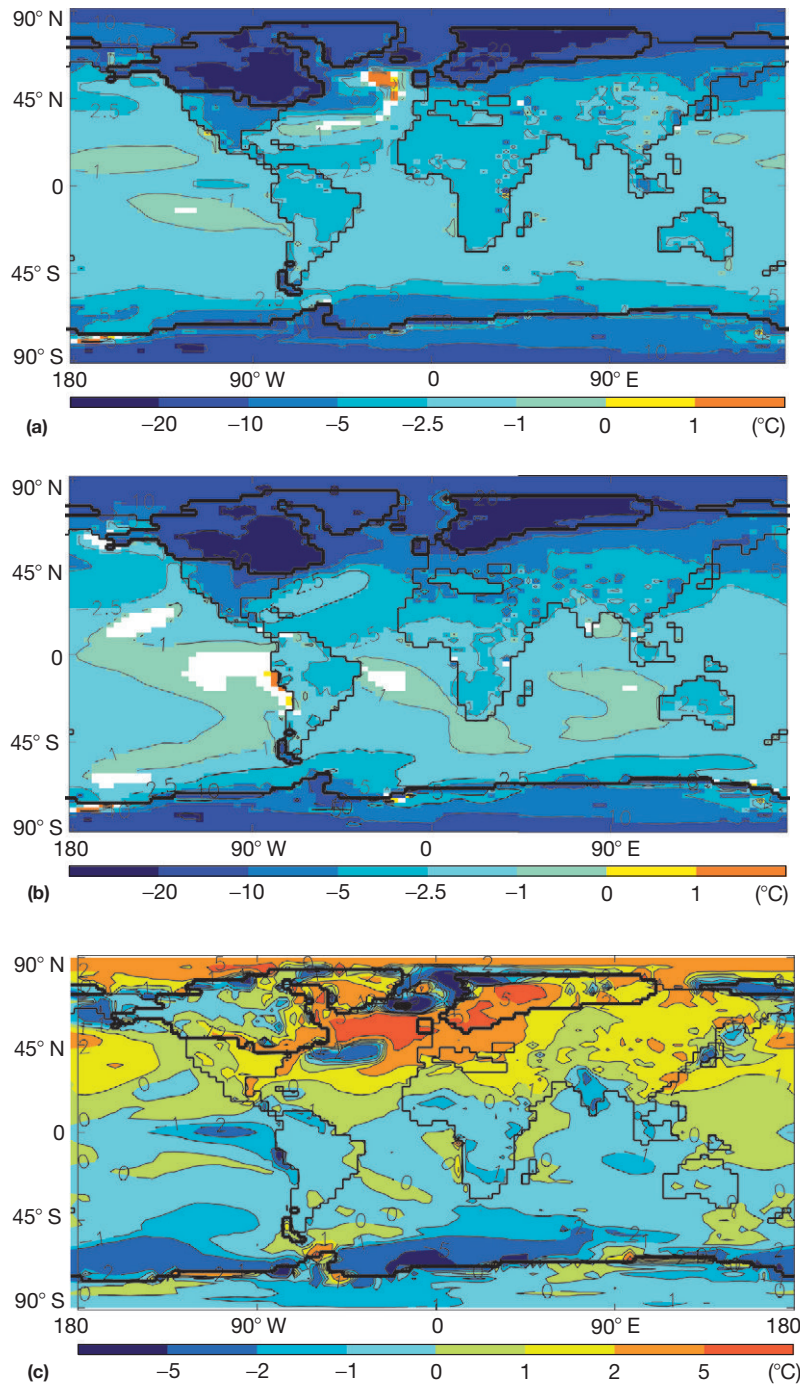


Figure 4 Dynamic versus slab-ocean versions of the Hadley Center climate model predictions of annual mean surface air temperature of Last Glacial Maximum climate change from present: (a) dynamic ocean experiment using HadCM3, (b) slab-ocean experiment using HadSM3, (c) HadCM3–HadSM3. Adapted from Hewitt CD, Stouffer RJ, Broccoli AJJ, Mitchell FB, and Valdes PJ (2003) The effect of ocean dynamics in a coupled GCM simulation of the Last Glacial Maximum. *Climate Dynamics* 20(2/3): 203–218. Crown Copyright 2006.

abrupt climate change. There exists a complex interplay as to the relative importance of thermal versus salinity forcing of the Atlantic Ocean's meridional overturning circulation and the role of the different parameterizations of horizontal and vertical mixing in the maintenance and stability of this circulation. A recent model intercomparison project considered the similarities and differences across a range of model classes,

including models of intermediate complexity (reduced-order models) as well as atmosphere–ocean GCMs. Of interest was the similar amplitude of the weakening that was predicted to occur across the different model designs, as well as the expression of these changes on surface climate across the globe for both weak and strong forcing experiments (Figure 6). One of the fundamental differences between the experiments

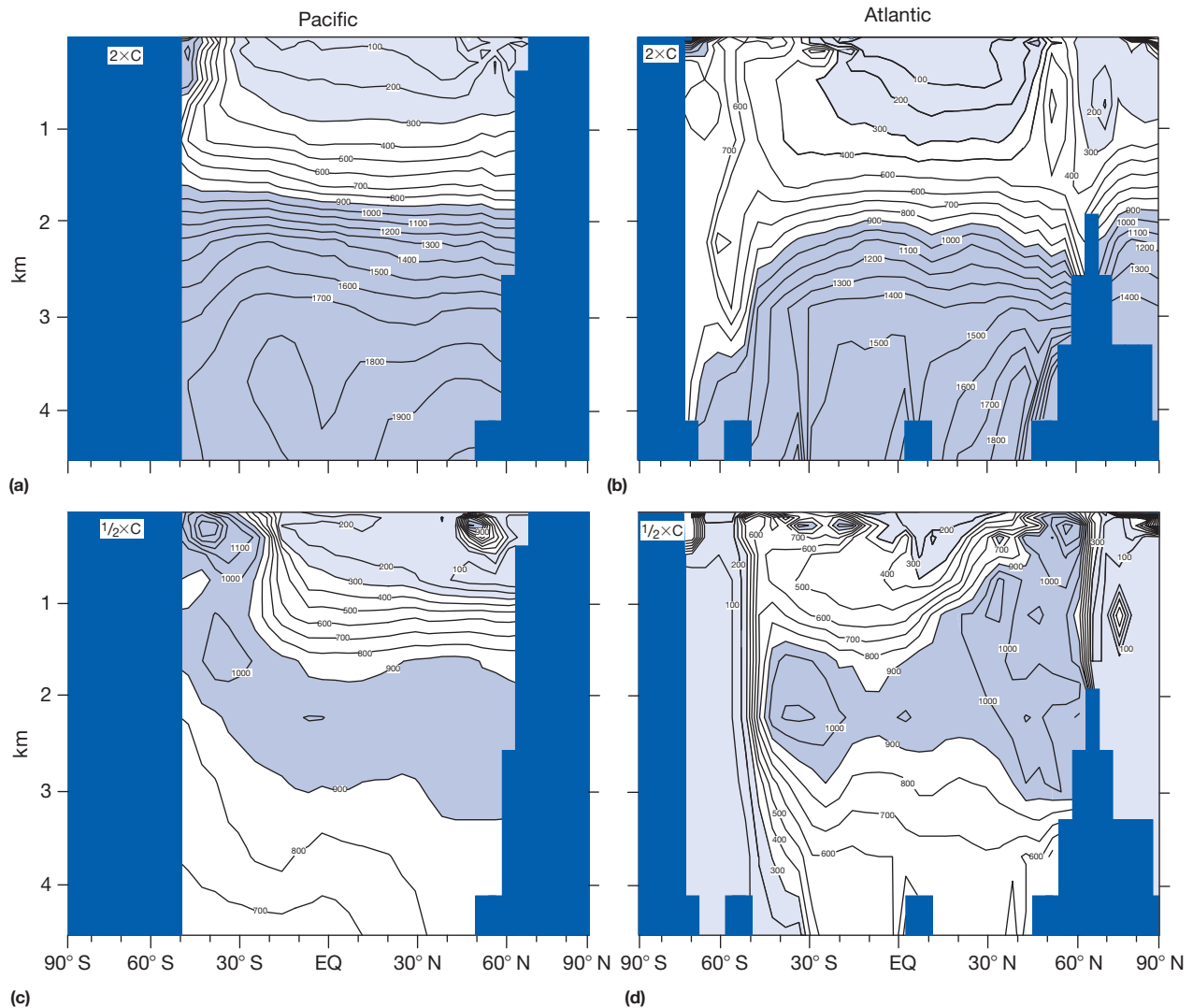


Figure 5 Time in years to reach 70% of the equilibrated response of the Pacific and Atlantic Ocean basins for a doubling ((a) and (b)) and halving ((c) and (d)) of atmospheric $p\text{CO}_2$ (Stouffer, 2004).

was the potential for the reduced-order ocean models to become locked in a state with a weak overturning circulation. It would appear that the mixing of ocean properties in ocean GCMs is a key part of their eventual recovery after forcing perturbations have ceased. This has potential implications for the interpretation of abrupt climate change, which conceptually has depended on the multiple stable equilibria observed in reduced-order models, in order to explain abrupt climate changes during the last-glacial cycle.

Glacial–interglacial changes in the carbon cycle

A range of model complexities have been formulated to consider hypotheses concerning glacial–interglacial changes in atmospheric CO_2 concentrations. These models consider processes that potentially affect the different pools of carbon that exist in the ocean and their preservation in ocean sediments. Many of these pools of carbon are either chemically or physically inaccessible for gas exchange with the atmosphere.

The relative sizes and exchange of carbon between these pools are controlled by numerous processes including ocean biology, availability of nutrients through ocean currents and supply through the atmosphere to the ocean surface, ocean temperatures, mixing between different water masses, ocean pH and alkalinity, and burial in sediments. The record of the preservation of calcium carbonate and opal in ocean sediments and proxies sensitive to ocean geochemistry and currents provides some constraints on the different hypotheses that have been proposed. However, there exists a curious dichotomy in the results of modeling studies that, depending on the modeling approach taken, suggests that the hypotheses that have been considered are either all insufficient (Archer et al., 2000a) or all sufficient (Legrand and Alverson, 2001) to explain the changes in the carbon cycle from LGM to present.

One of the focal points in the discussion of why these differences exist concerns the role of the low versus high

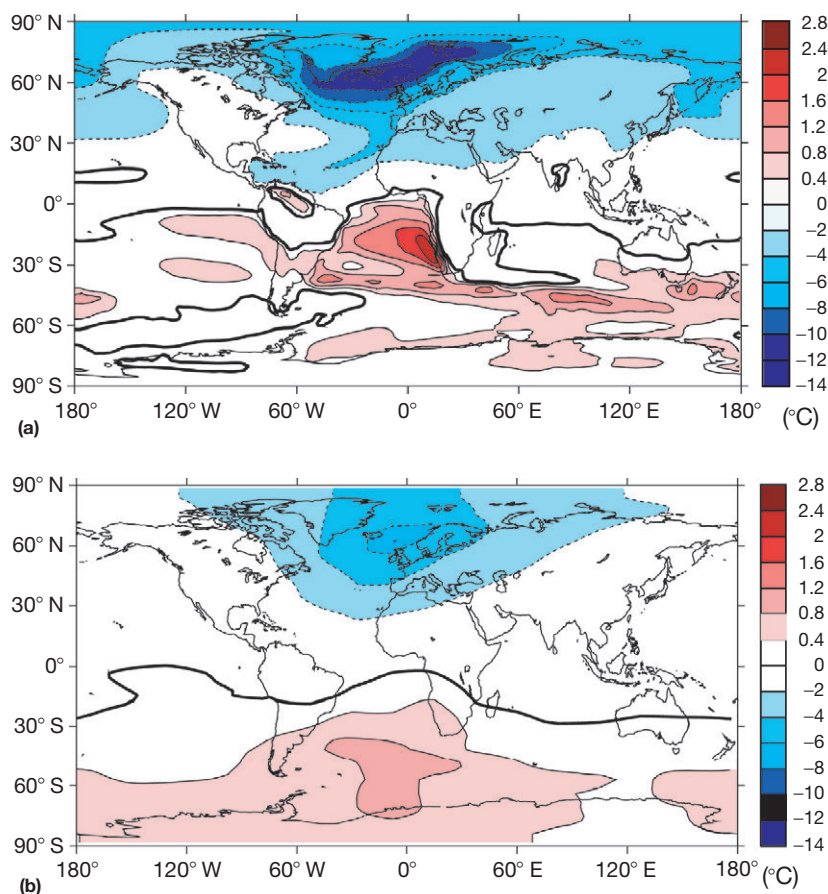


Figure 6 Surface air temperature response within an ensemble of (a) atmosphere–ocean GCMs and (b) models of intermediate complexity after 100 years of applying a 1 Sv ($10^6 \text{ m}^3 \text{ s}^{-1}$) (equivalent to raising sea level by 9 m) forcing to the North Atlantic (Stouffer et al., 2006).

latitudes in controlling the ocean's equilibration with the atmosphere (Archer et al., 2000b; Toggweiler et al., 2003). The warm water temperatures in the equatorial regions make carbon much less soluble in these waters than in the cooler polar waters. Because the pool of warmer tropical waters exchanges gases less effectively with the deep ocean than the surface waters of the higher latitudes, box models show that the higher latitudes would be much more important in these exchanges. The larger amount of vertical mixing in ocean GCMs appears to diminish the strength of this effect. However, because there still exists a significant amount of uncertainty concerning both the processes governing and the overall amplitude of vertical mixing in the world's oceans, the correct balance is not clear. Also being debated is the rate of horizontal mixing in the ocean that could be important for regulating the area of the ocean surface that is equilibrating with the atmosphere, and its effective exchange with the deep-ocean reservoir of carbon (Archer et al., 2003; Toggweiler et al., 2003). Box models assume that all water within a given box is well mixed, and so, these models tend to be overly efficient in their gas equilibration with the atmosphere, relative to how these processes are controlled in an ocean GCM. Therefore, box models show the plausibility of a number of hypotheses. The details of these hypotheses could potentially be worked out within ocean GCMs, but more

work is needed to establish whether ocean GCMs are modeling natural processes accurately.

Conclusions

The primary scientific opportunity of modeling Quaternary climate is to understand how the climate system operates as a whole. The precise balance of processes that maintain a given climate state is not readily apparent in instrumental data, nor is it obvious in comprehensive models of climate. The choices that are made to represent the climate system simply potentially affect how a model treats these balances, leading to misleading sensitivities. As was illustrated by the efforts to model glacial–interglacial changes of the carbon cycle, simplified models often are more sensitive than atmosphere–ocean GCMs. While there may be fewer explicit assumptions with GCMs, it is harder to know how to control the balance of processes that are being played out within them. Comparing two climate change experiments among several models could potentially provide some guidance as to which model predictions are more robust than others. Finally, it should not matter whether a model is made up of a few boxes with parameterized physics or millions of boxes with physical processes being resolved, if one is correctly capturing the relevant balances.

Getting to know what those balances should be is what makes modeling Quaternary climate interesting.

See also: Paleoclimate Relevance to Global Warming. **Glacial Climates:** Biosphere Feedbacks; Effects of Atmospheric Dust; Thermohaline Circulation. **Glacial Landforms, Ice Sheets:** Growth and Decay. **Glaciation, Causes:** Astronomical Theory of Paleoclimates. **Ice Core Methods:** CO₂ Studies. **Introduction:** Societal Relevance of Quaternary Research. **Lake Level Studies:** Modeling. **Paleoclimate:** Timescales of Climate Change. **Paleoclimate Modeling:** Data–Model Comparisons; Last Glacial Maximum GCMs; Paleo-ENSO; The Last Interglacial. **Paleoclimate Reconstruction:** Sub-Milankovitch (DO/Heinrich) Events. **Pollen Methods and Studies:** The Biome Approach to Reconstructing Past Vegetation.

References

- AchutaRao K, Covey C, Doutriaux C, et al. (2004) In: Bader D (ed.) *An Appraisal of Coupled Climate Model Simulations*. Livermore, CA: Lawrence Livermore National Laboratory UCRL-TR-202550.
- Archer DE, Eshel G, Winguth A, et al. (2000a) Atmospheric pCO₂ sensitivity to the biological pump in the ocean. *Global Biogeochemical Cycles* 14(4): 1219.
- Archer DE, Martin PA, Milovich J, Brovkin V, Plattner G-K, and Ashendel C (2003) Model sensitivity in the effect of Antarctic sea ice and stratification on atmospheric pCO₂. *Paleoceanography* 18(1): 1012.
- Archer D, Winguth A, Lea D, and Mahowald N (2000b) What caused the glacial/interglacial atmospheric pCO₂ cycles? *Reviews of Geophysics* 38: 159–189.
- Bradley RS (1999) *Paleoclimatology: Reconstructing Climates of the Quaternary*, 2nd edn., p. 613. New York: Academic Press.
- CLIMAP Project Members (1976) The surface of the ice-age earth. *Science* 191: 1131–1137.
- Crowley TJ and North GR (1996) *Paleoclimatology*, p. 360. New York, NY: Oxford University Press.
- Goody R (1995) *Principles of Atmospheric Physics and Chemistry*, p. 324. New York, NY: Oxford University Press.
- Hewitt CD, Stouffer RJ, Broccoli AJ, Mitchell FB, and Valdes PJ (2003) The effect of ocean dynamics in a coupled GCM simulation of the Last Glacial Maximum. *Climate Dynamics* 20(2/3): 203–218.
- IPCC (2001) Houghton JT, Ding Y, and Griggs DJ, et al. (eds.) *Climate Change 2001: The Scientific Basis. Contribution of Working Group I to the Third Assessment Report of the Intergovernmental Panel on Climate Change*, p. 881. Cambridge: Cambridge University Press.
- Jackson C and Broccoli AJ (2003) Orbital forcing of arctic climate: Mechanisms of climate response and implications for continental glaciation. *Climate Dynamics* 21: 539–557.
- Kucera M, Schneider R, and Weinelt M (eds.) (2005) *MARGO – Multiproxy Approach for the Reconstruction of the Glacial Ocean Surface*, p. 300. Amsterdam: Elsevier.
- McGuffie K and Henderson-Sellers A (2005) *A Climate Modelling Primer*, 3rd edn., p. 296. New York: John Wiley and Sons.
- Stouffer RJ (2004) Time scales of climate response. *Journal of Climate* 17(1): 209–217.
- Stouffer RJ, Yin J, Gregory JM, et al. (2006) Investigating the causes of the response of the thermohaline circulation to past and future climate changes. *Journal of Climate* 19: 1365–1387.
- Toggweiler JR, Gnanadesikan A, Carson S, Murnane R, and Sarmiento JL (2003) Representation of the carbon cycle in box models and GCMs: 1. Solubility pump. *Global Biogeochemical Cycles* 17(1): 10.
- Trenberth KE (ed.) (1992) *Climate System Modeling*, p. 818. Cambridge: Cambridge University Press.
- Washington WM and Parkinson CL (2005) *An Introduction to Three-Dimensional Climate Modeling*, 2nd edn., p. 368. Sausalito, CA: University Science Books.

The Last Interglacial

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Introduction

The last interglacial maximum (LIG; ca. 125 ka BP), also referred to as Marine Isotopic Stage (MIS) 5e or Eemian because of the warm biostratigraphic assemblages found for this time period in the Eem Valley (the Netherlands), is of interest for both modelers and geologists. In addition to being considered the warmest time period of the last 200 ka, its end represents the best documented transition from an interglacial to a glacial state (Kukla et al., 2002) and is relevant for our understanding of future anthropogenic global warming: although the forcing factors are very different, an adequate simulation of past warm periods enhances our confidence in climate models and thus in future climate projections.

Evidence for warmth during the LIG relies mainly on terrestrial and marine records from mid- and high northern latitudes (e.g., Aalsbersberg and Litt, 1998; Klotz et al., 2003; LIGA Members, 1991 and references therein; Zagwijn, 1996; see Figure 1). In agreement with isotopically light $\delta^{18}\text{O}$ values from the deep-sea record (e.g., CLIMAP, 1984 and references therein), global sea level may have been up to 6–7 m higher than the present level (Crowley and North, 1991 and references therein), implying a significant reduction in global ice volume, corresponding to the total of the Greenland or the West Antarctic ice sheets. Vostok ice-core data indicate carbon dioxide (CO_2) and methane (CH_4) levels comparable to pre-industrial ones and local temperatures about 2 °C higher than present (e.g., Petit et al., 1999). In contrast, sea-surface temperature (SST) reconstructions (CLIMAP, 1984) suggest that values were not significantly higher than that today (Figure 2).

Climate simulations of the LIG have been carried out with atmosphere-only general circulation models (AGCMs; e.g., De Noblet et al., 1996; Khodri et al., 2003; Kutzbach et al., 1991; Prell and Kutzbach, 1987; Royer et al., 1984), AGCMs coupled with static mixed-layer ocean models (e.g., Hall et al., 2005; Harrison et al., 1995), coupled atmosphere–ocean general circulation models (AOGCMs; e.g., Felis et al., 2005; Groll et al., 2005; Kaspar and Cubasch, 2005; Kaspar et al., 2005; Montoya et al., 1998, 2000), Earth system models of intermediate complexity (EMICs; e.g., Kubatzki et al., 2000), and energy balance models (EBMs; e.g., Crowley and Kim, 1994). Initially, the focus was mainly on the direct thermodynamic response to the insolation changes, such as that at high latitudes and in the monsoons (e.g., Harrison et al., 1995; Kutzbach et al., 1991; Prell and Kutzbach, 1987). More recently, modeling studies have focused on changes in dynamics forced by changes in insolation (Felis et al., 2005; Groll et al., 2005; Hall et al., 2005). EMICs have also been used to analyze the role of specific climatic components or forcing factors and to investigate the reasons for discrepancies between EMIC and AOGCM results (Kubatzki et al., 2000).

In this article, the latter studies are reviewed, focusing on equilibrium climate simulations of the LIG. After describing

the typical model setup, the seasonal and annual mean responses are analyzed, focusing on the changes in the surface-air temperature (SAT), the southwestern monsoons, atmospheric dynamics, and ocean circulation, and the role of vegetation feedbacks. Finally, the main conclusions are summarized, aiming at establishing robust model responses, the main disagreements among models, and the remaining open questions.

Boundary Conditions

Climate simulations of the LIG based on GCMs, which are computationally expensive, have mainly consisted in so-called equilibrium or time-slice simulations in which the relevant boundary conditions or external forcing factors are changed and the model is integrated until the simulated climate is in equilibrium with the specified forcing. Owing to the long timescales of some of its components, climate is really seldom in equilibrium with its boundary conditions, but in sufficiently long and relatively stable time periods, and as a first step, this constitutes a reasonable approach.

Since the establishment of Milankovitch's astronomical theory of ice-ages, variations in the insolation received by the Earth at the top of the atmosphere following variations in the Earth's orbital parameters are recognized as the key factors driving climate on orbital timescales (about 10^4 – 10^6 years), that is, glacial–interglacial cycles. Time-slice experiments are thus carried out by prescribing the Earth's orbital parameters at a given time period (Berger, 1978), which determines the insolation at the top of the atmosphere at that time. Table 1 shows the orbital parameters for 125 ka BP, suitable for a time-slice climate simulation of the LIG, together with those for the present. At 125 ka BP, the eccentricity of the Earth's orbit was greater, obliquity was slightly greater, and perihelion took place in northern summer instead of in northern winter as it does today (Figure 3). As a consequence, northern summer insolation was greater at 125 ka BP than it is today at all latitudes, with a peak value of 60 W m^{-2} at high northern latitudes (Figure 4). On average, insolation increased by 50 W m^{-2} (12%). Northern winter insolation decreased at all latitudes with an average value of 26 W m^{-2} (11%). At high latitudes, the decreasing effect of the timing of perihelion on insolation was partly compensated by the greater obliquity. Hence, maximum northern winter insolation decreases took place in the southern subtropics. To summarize, the seasonal cycle of insolation was amplified in the Northern Hemisphere and attenuated in the Southern Hemisphere. Mean annual insolation increased at high latitudes, where enhanced summer insolation is not offset in winter, and decreased at low latitudes, but the departure was below 2 W m^{-2} at all latitudes (not shown), and its globally averaged change was negligible (0.23 W m^{-2}).

Variations in greenhouse gases (GHGs) are also taken into account (Table 1). The CO_2 concentration is prescribed to be

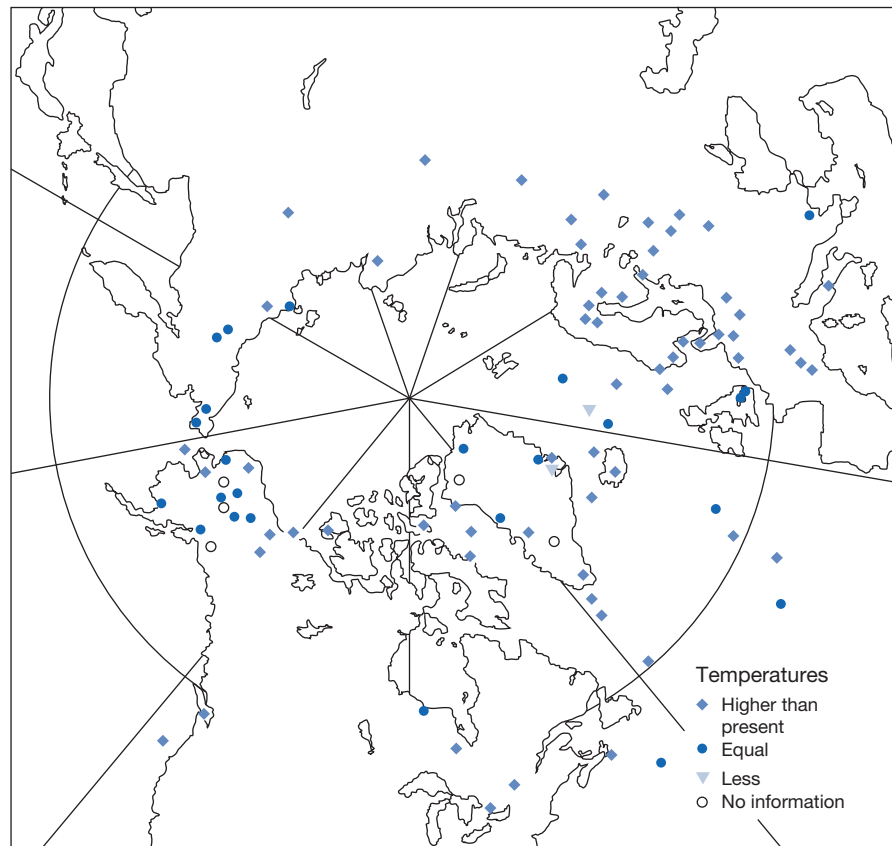


Figure 1 Temperature conditions at selected sites during the last interglacial (LIG) maximum as compared to modern values. Reprinted from LIGA Members (1991) Report of 1st discussion group: The last interglacial in high latitudes of the Northern Hemisphere: Terrestrial and marine evidence. *Quaternary International* 10–12: 9–28, © 1991, with permission from Elsevier.

about 270 ppmv, consistent with mean levels for the LIG estimated from ice-core measurements. This is similar to preindustrial levels (Petit et al., 1999). Although in the past control runs have been carried out with higher CO₂ levels (e.g., Khodri et al., 2003; Montoya et al., 2000), it is now recognized that the present climate is not in equilibrium with such high CO₂ levels, and that it is more reasonable to use a preindustrial level. In some simulations, other GHGs such as CH₄ and nitrous oxide (N₂O) have additionally been taken into account (Groll et al., 2005; Kaspar and Cubasch, 2005; Kaspar et al., 2005). Simpler models such as EMICs include interactive biosphere components (Kubatzki et al., 2000). Finally, since the estimated 6-m sea-level rise at the LIG relative to the present leads to negligible changes in the model land–sea mask, the latter is kept as in the control run.

Seasonal Responses

Owing to the wealth of proxy data for the LIG showing evidence for warmth at mid–high northern latitudes (e.g., LIGA Members, 1991) and for stronger southwestern monsoons in South Asia and Africa (e.g., Prell and Kutzbach, 1987), from the modeling perspective, attention has focused mainly on the latter regions. However, recently the role of changes in dynamics, especially in northern winter, has been highlighted.

Northern Summer Response: Direct Response to Insolation

The June–August (JJA) SAT responses simulated by all models, from EBMs to coupled AOGCMs (e.g., Crowley and Kim, 1994; De Noblet et al., 1996; Felis et al., 2005; Groll et al., 2005; Harrison et al., 1995; Kaspar and Cubasch, 2005; Kaspar et al., 2005; Kubatzki et al., 2000; Montoya et al., 2000; Prell and Kutzbach, 1987), agree well with one another and with the evidence for enhanced warming at the LIG from paleodata (e.g., Aalsbersberg and Litt, 1998; Klotz et al., 2003; LIGA Members, 1991; Zagwijn, 1996). As an example, Figure 5 shows the 2-m-temperature departures with respect to the control run of one of the most recent climate simulations of the LIG with a state-of-the-art coupled AOGCM (Groll et al., 2005; Kaspar and Cubasch, 2005; Kaspar et al., 2005). The JJA anomalies reflect the direct thermodynamic response to the insolation changes: as a consequence of the increase of insolation (Figure 4), models simulate overall enhanced warming over land. The largest departures are found in the northern continents, where the increase in both radiative forcing and the land-mass distribution, with low heat capacity compared to that of the ocean, are greatest. Owing to the large thermal inertia of the ocean, the SAT response is generally stronger in atmosphere-only models, some of which show maximum departures above 10 °C over the Eurasian continent (Khodri et al., 2003; Royer et al., 1984), compared to coupled atmosphere–ocean models,

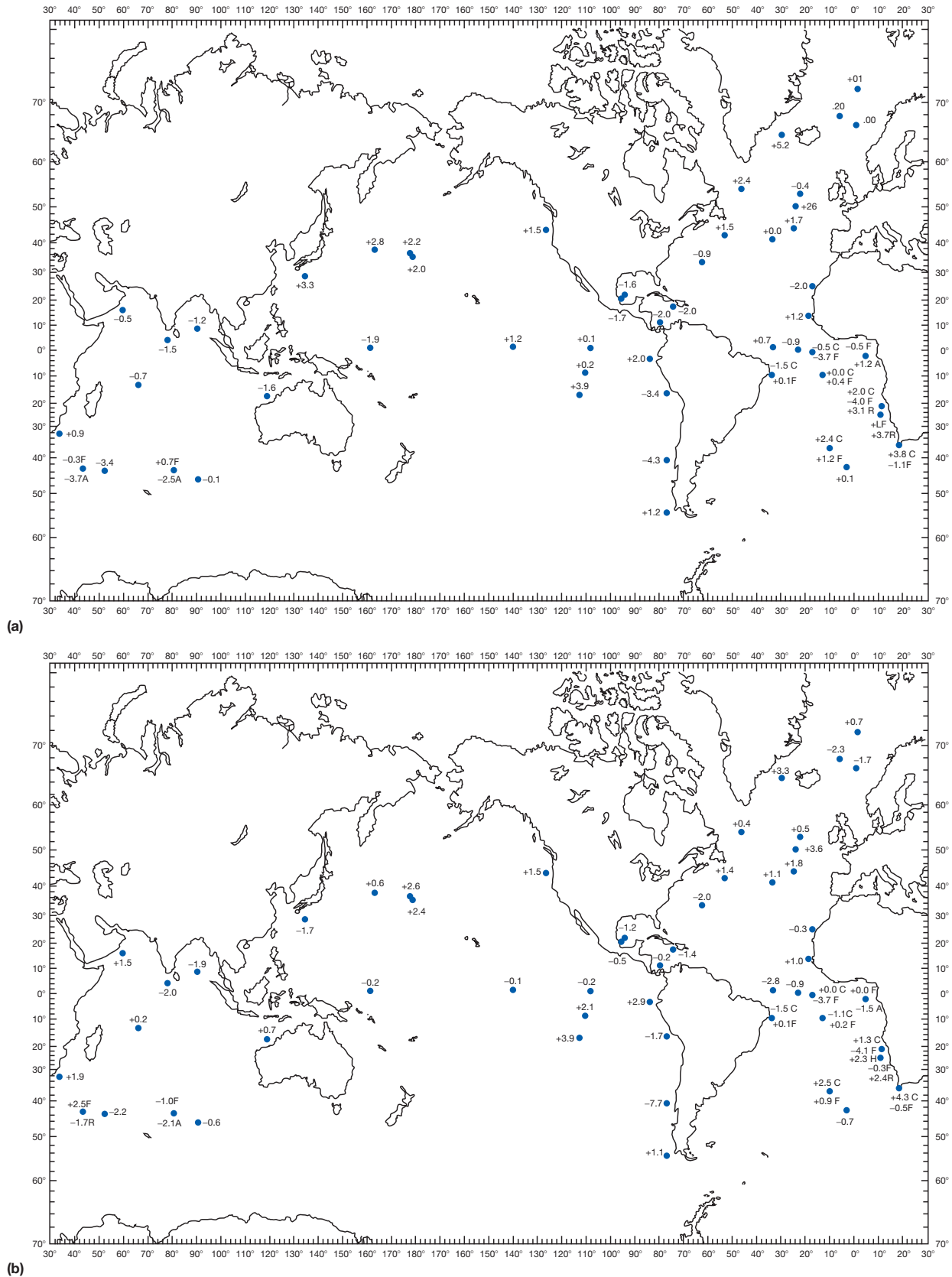
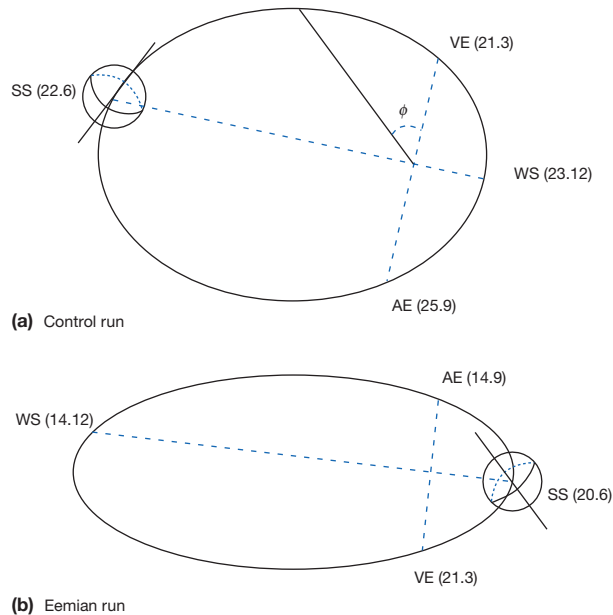


Figure 2 Difference in estimated sea-surface temperature (SST) between the last interglacial (LIG) and today (value at MIS-5e minus value taken for modern atlas data) for (a) February and (b) August, in degrees Celsius. Reprinted from CLIMAP Project Members (1984) The last interglacial ocean. *Quaternary Research* 21: 123–224, © 1984, with permission from Elsevier.

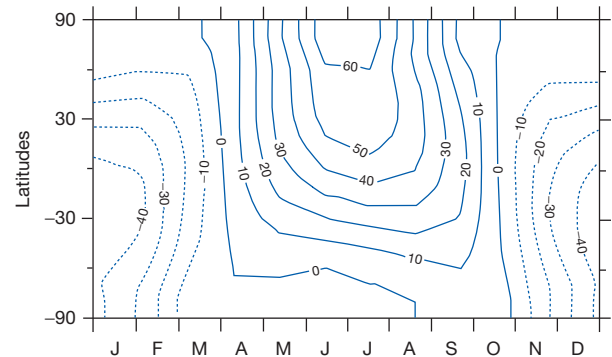
Table 1 Boundary conditions typically employed in time-slice climate simulations for the preindustrial climate and the LIG (125 ka BP)

	<i>Eccentricity</i> ^{a,b}	<i>Obliquity</i> ^{a,b} (°)	<i>Angle of perihelion</i> ^{a,b} (°)	<i>CO₂</i> ^{a,b} (ppmv)	<i>CH₄</i> ^b (ppmv)	<i>N₂O</i> ^b (ppbv)
Control run	0.017	23.45	282.16	280	630	160
125 ka BP	0.040	23.79	127.27	270–280	700	265

^aMontoya et al. (2000).^bGroll et al. (2005); Kaspar and Cubasch (2005); Kaspar et al. (2005); BP, before present; CH₄, methane; CO₂, carbon dioxide; ka, thousand years; N₂O, nitrous oxide; ppbv, parts per billion by volume; ppmv, parts per million by volume.**Figure 3** Position of the solstices and equinoxes for (a) the control run (the present), and (b) the Eemian run (125 ka BP). Dates of solstices and equinoxes are given in parentheses: VE, vernal equinox; AE, autumnal equinox; SS, summer solstice; WS, winter solstice. Reprinted from Montoya M, von Storch H, and Crowley TJ (2000) Climate simulation for 125 ka BP with a coupled ocean–atmosphere general circulation model. *Journal of Climate* 13: 1057–1072, © 2000, with permission from the American Meteorological Society.

in which the maximum SAT increases are $\sim 4\text{--}6^\circ\text{C}$ (Felis et al., 2005; Kubatzki et al., 2000; Montoya et al., 2000). For the same reason, the SAT response over the ocean is generally below 1°C (e.g., Groll et al., 2005; Harrison et al., 1995; Kaspar and Cubasch, 2005; Kubatzki et al., 2000; Montoya et al., 2000; see also Figure 5). The exceptions are the high northern latitudes, where insolation anomalies are largest (Figure 4) and SAT departures are amplified through the positive sea-ice albedo–SST feedback, that is, the sea-ice reduction following the increased JJA warming leads to enhanced absorption of surface solar radiation which further enhances the warming. This effect is not limited to JJA, but takes place all-year round (see below).

A common feature of many models including AGCMs (e.g., De Noblet et al., 1995; Felis et al., 2005; Groll et al., 2005; Khodri et al., 2003; Montoya et al., 2001; Royer et al., 1984) is a belt of reduced SATs over Central Africa and South Asia, related to the strengthened monsoons and enhanced cloud cover (not shown) in these areas. This is consistent with paleo-reconstructions of monsoon-related climatic indices suggesting

**Figure 4** Monthly mean insolation at the top of the atmosphere in watts per square meter (W m^{-2}) as a function of latitude for 126 ka BP minus present with 126 ka BP based on angular months. Reproduced from Joussaume S and Braconnot P (1997) Sensitivity of paleoclimate results to season definitions, *Journal of Geophysical Research* 102: 1943–1956, © 1997 American Geophysical Union, with permission of American Geophysical Union.

intensified monsoons during interglacial periods (e.g., Prell and Kutzbach, 1987). This result can be understood through the simulated response of the models, summarized in Figure 6: the JJA sea level pressure (SLP) change reflects the direct thermodynamic response to the JJA continental warming, decreasing over the warmer landmasses, especially over Eurasia, North Africa, and North America, and increasing, or decreasing less, over the cooler surrounding oceans (Groll et al., 2005; Khodri et al., 2003; Kubatzki et al., 2000; Montoya et al., 2000). As a consequence, the land–ocean SLP gradient is enhanced. In the Tropics, this leads to an intensification and northward migration of the on-shore low-level convergent flow, and thus of moisture transport further inland into equatorial Africa and South Asia, and a strengthening of the tropical easterly jet at high levels. This response reflects an intensification of the African and southwest Asian monsoons (De Noblet et al., 1995; Groll et al., 2005; Kubatzki et al., 2000; Montoya et al., 2000; Prell and Kutzbach, 1987). AGCMs suggest that the total precipitation is not affected significantly (De Noblet et al., 1995). However, EMICs suggest that when an interactive biosphere component is included, precipitation departures, both in JJA and in the annual mean, virtually double because of the existence of vegetation–precipitation feedbacks: the increase in precipitation at the LIG leads to an expansion of vegetation; subsequent changes in the radiation balance lead to a surplus in available energy used for evaporation. The increase in atmospheric total energy (sensible plus latent heat) results in enhanced convection and, together with changes in

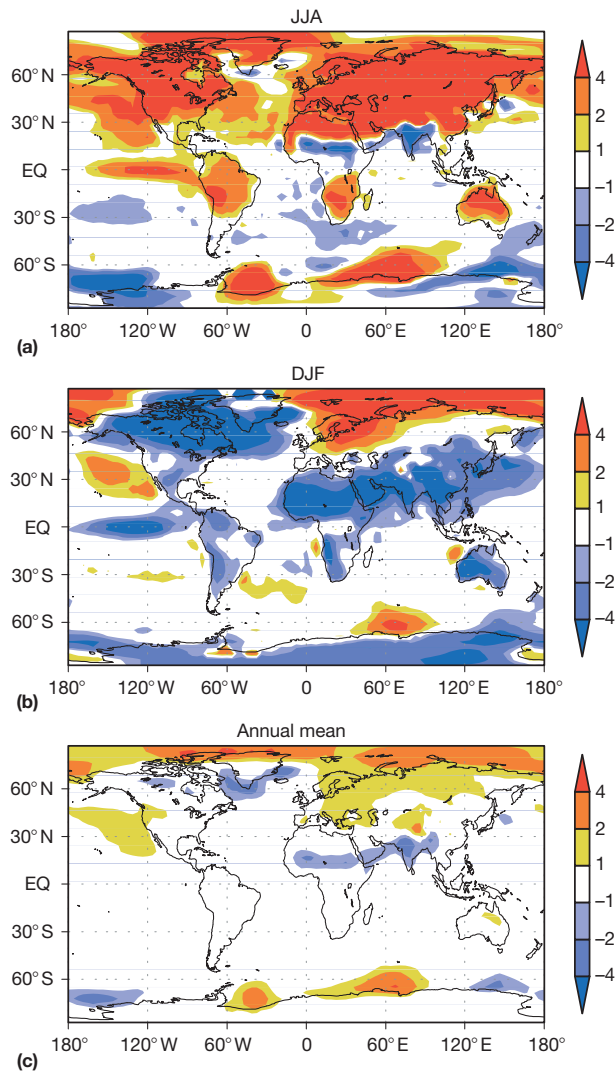


Figure 5 Departure in 2-m-temperature (125 ka BP minus control) in degrees Celsius ($^{\circ}\text{C}$) for (a) JJA, (b) DJF, and (c) annual mean, as simulated by the ECHO-G AGCM. Courtesy of F. Kaspar.

the moisture advection, in additional precipitation increase (Kubatzki et al., 2000 and references therein). As a consequence, 97% of North Africa is covered by vegetation, mainly grasses (Figure 7).

Northern Winter Response: The Role of Atmospheric Dynamics

Consistent with the decrease in insolation at almost all latitudes in December–February (DJF), models show considerable cooling over the continents (Figure 5). Although the maximum insolation decreases take place in the southern subtropics (Figure 2), maximum cooling takes place in the northern subtropical continents because of the larger fraction of surface covered by the ocean in the southern subtropics. As a consequence of the enhanced JJA insolation at high northern latitudes and the positive sea-ice albedo–SST feedback, in JJA, the heat storage by the ocean increases. This results in a delay in the onset of Arctic sea-ice growth in autumn, accelerated melting in

spring, and an all-year round warming (Figure 5, DJF) and sea-ice thickness and extent decrease (not shown). In models including oceanic components, warming up to 4°C , centered in the Arctic, takes place in DJF (e.g., Groll et al., 2005; Kubatzki et al., 2000; Montoya et al., 2000; see also Figure 5). In these aspects that reflect the direct thermodynamic response to insolation changes, the simpler models, such as EMICs (e.g., Kubatzki et al., 2000), are in good agreement with more complex models (e.g., Felis et al., 2005; Groll et al., 2005; Montoya et al., 2000).

However, models including AGCMs, even with SSTs prescribed to their present-day values, simulate considerable warming in DJF at mid-northern latitudes of Eurasia and North America (e.g., Groll et al., 2005; Harrison et al., 1991; Kaspar and Cubasch, 2005; Kaspar et al., 2005; Montoya et al., 1998). The warming pattern (Figure 5) agrees well with January SAT reconstructions based on pollen and plant macrofossil assemblages (Kaspar et al., 2005). Harrison et al. (1991) already showed such a warming as a consequence of stronger westerly winds, together with a northward shift and strengthening of the wintertime Atlantic jet, bringing warm and moist air into central and northeastern Europe. Kaspar and Cubasch (2005) have shown that the latter changes are a consequence of the SLP response over the North Atlantic, which is a dipole showing enhanced SLP over the subtropics and reduced SLP at high latitudes, comparable to a high-index state of the Arctic Oscillation (AO), which is the main mode of DJF northern extratropical climate variability.

An intensification of the AO index during the LIG was first suggested by Felis et al. (2005), who analyzed their AOGCM LIG climate simulation in comparison with coral records of the Red Sea, which indicated increased temperature seasonality during the LIG compared to the present. The AOGCM simulations suggest that this result is a consequence of the higher AO index leading to colder winters in the Middle East during the LIG compared to the present. This result is supported by the close correlation found between the present-day SST anomalies of the Red Sea and the AO index in the observations and simulations, for both the present day and the LIG. Furthermore, in an ensemble of transient simulations for the LIG with this model, the simulated AO index throughout most of the period is significantly higher than that for the modern climate.

A similar conclusion was reached by Hall et al. (2005), who studied the response of their AGCM coupled to a mixed-layer ocean model to insolation variations throughout the last 165 ka. To this end, they decomposed the model's circulation response into a linear response to obliquity and precession (orbital response) and a residual, mainly due to internal variability. In JJA, the orbital response is a local thermodynamic one to insolation changes, as described above. However, in DJF, the orbital response includes a dynamic response which is very similar to the AO (Figure 8). This result is consistent with models showing a northward shift of the Atlantic jet (e.g., Harrison et al., 1991; Kaspar and Cubasch, 2005; Kaspar et al., 2005; Montoya et al., 2000). Furthermore, European SATs are shown to be out of phase with DJF insolation changes but almost perfectly in phase with the orbital AO index, both reaching their maximum values at ca. 120 ka BP, when as a consequence of enhanced advection of warm, moist air into Europe, the model generates a $1\text{--}2^{\circ}\text{C}$ warm SAT anomaly over

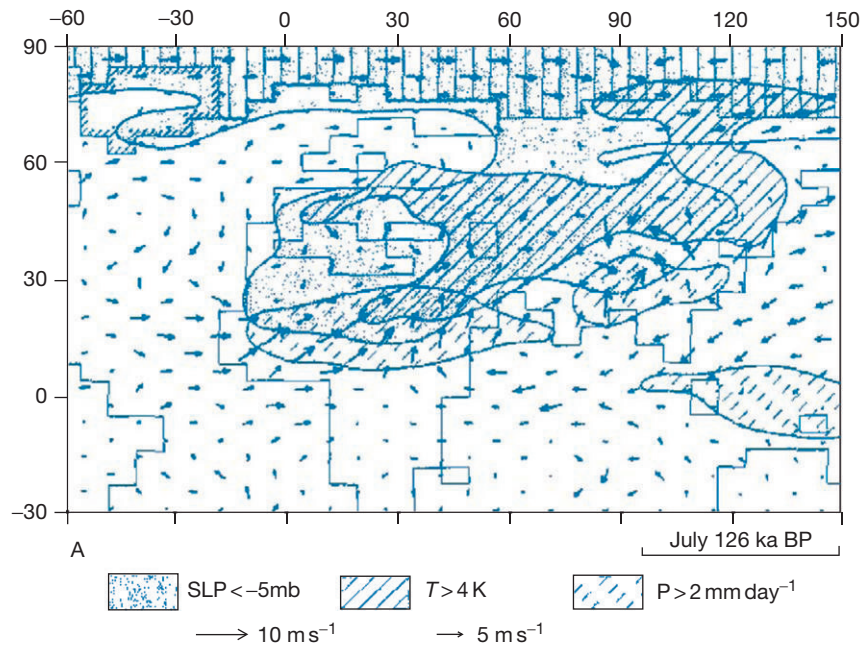


Figure 6 Departures (LIG minus control) of surface temperature (T), SLP, winds (arrows), and precipitation (P) for the Indian Ocean sector. Sea ice is indicated by vertical stripes. Reproduced from Prell WL and Kutzbach JE (1987) Monsoon variability over the past 150,000 years. *Journal of Geophysical Research* 92: 8411–8425, © 1987 American Geophysical Union, with permission.

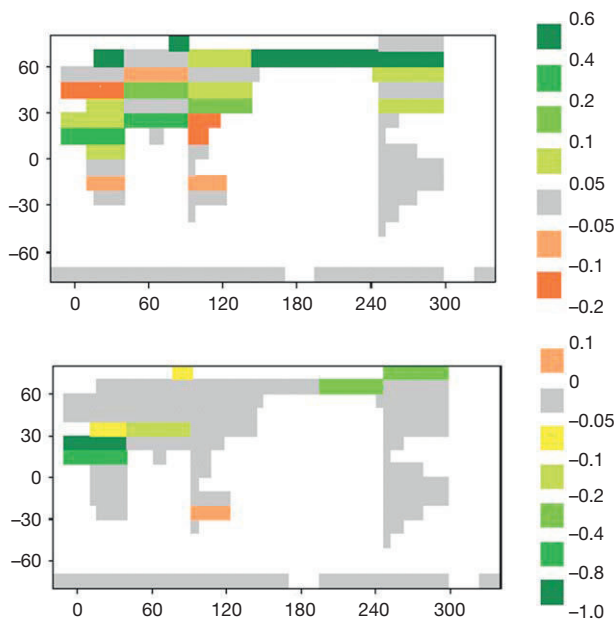


Figure 7 Departure (LIG minus present) in forest fraction (top) and desert fraction (bottom) simulated by the CLIMBER-2 EMIC in the model version including interactive atmospheric, oceanic, and biosphere components.

Europe (Figure 5). This result is consistent with temperature reconstructions indicating warmer than present LIG temperatures (Aalsbersberg and Litt, 1998; Klotz et al., 2003; LIGA Members, 1991; Zagwijn, 1996) and provides an explanation of the warm winter anomaly.

Annual Mean SAT Response and Globally Averaged Temperatures

Owing to the widespread evidence for warmth during the LIG, it has often been discussed whether globally averaged temperatures were significantly higher than the present temperature (e.g., Kubatzki et al., 2000; Montoya et al., 1998). Proxy data indeed indicate enhanced warming at high northern latitudes (e.g., LIGA Members, 1991), but SSTs were not significantly higher than the present value (CLIMAP Project Members, 1984). Coupled atmosphere–ocean climate models generally do not suggest significantly higher global temperatures either. The reason is that despite the large warming simulated in northern summer, most models show significant cooling in winter (e.g., Groll et al., 2005; Kubatzki et al., 2000; Montoya et al., 2000), with the net result that, except certain specific regions such as the Arctic and some regions in Eurasia, the mean annual SAT and SST changes are below 1 °C almost everywhere (e.g., Groll et al., 2005; Montoya et al., 2000; see also Figure 5). Montoya et al. (1998) compared the SST response simulated by their AOGCM with the CLIMAP Project Members (1984) SST reconstruction on a point-by-point basis (Figure 9). This high level of agreement suggests that the model results are consistent with the CLIMAP Project Members (1984) SST reconstruction for the LIG and supports the inference of similar or slightly higher mean annual and global temperatures.

Vegetation Feedbacks

However, one mechanism by which warming at the LIG might have been amplified all-year round, especially at mid- and high northern latitudes, is through vegetation feedbacks, as shown

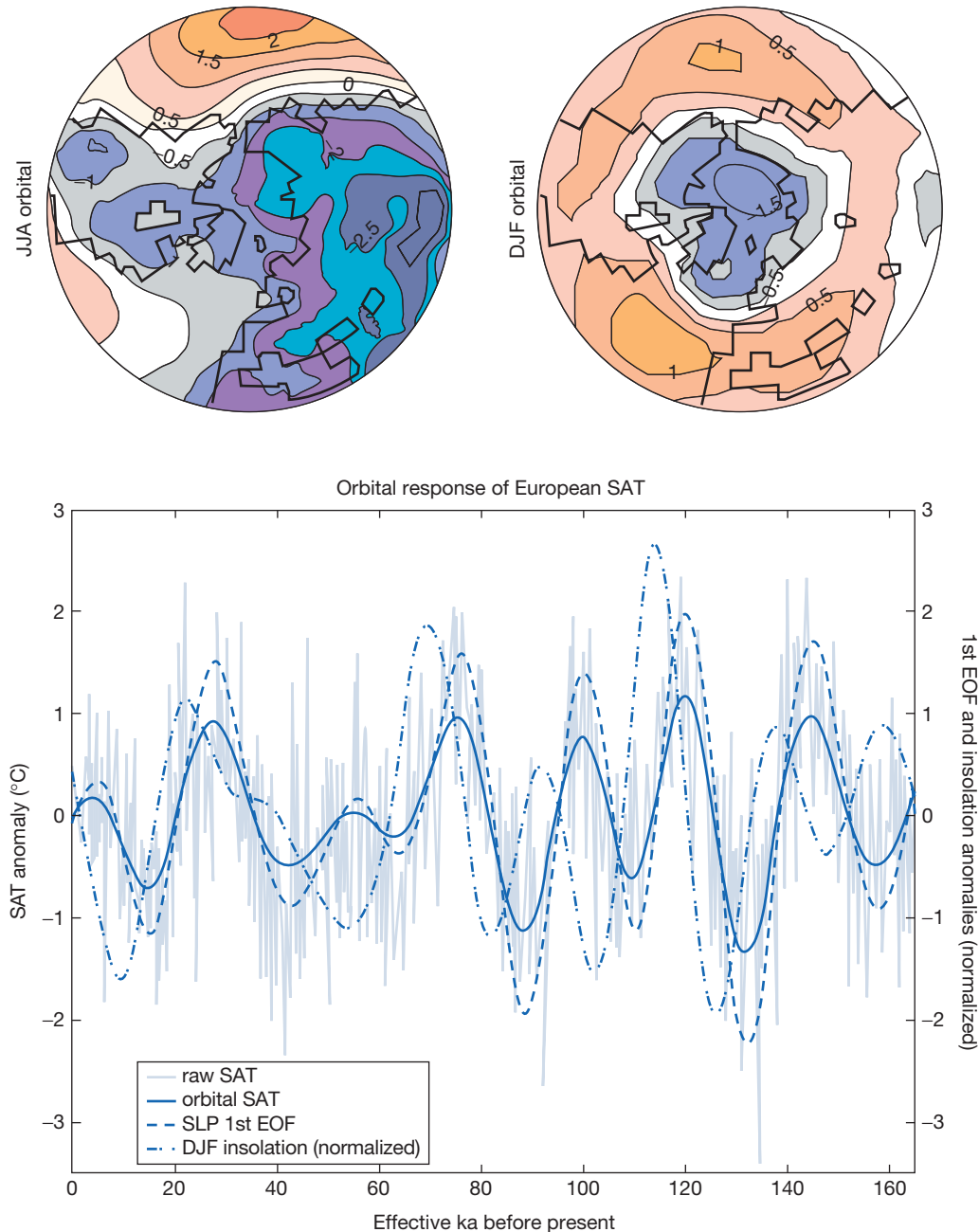


Figure 8 Top: First empirical orthogonal function (EOF) for orbital SLP variability north of 26°N , for JJA (left) and DJF (right). The patterns account for 96% and 80% of the total variance. Bottom: time series of 30-year-mean DJF SAT anomaly ($^{\circ}\text{C}$) averaged over the model's European continent (raw SAT, light gray), orbital DJF SAT ($^{\circ}\text{C}$) averaged over the model's European continent (solid), primary mode of orbital SLP variability (dashed), and DJF incoming insolation due to Milankovitch forcing averaged over the same latitudes, normalized by its standard deviation (dashed-dotted). Reprinted from Hall A, Clement A, Thompson DWJ, Broccoli A, and Jackson C (2003) The importance of atmospheric dynamics in the Northern Hemisphere wintertime climate response to changes in the Earth's orbit. *Journal of Climate* 18: 1315–1325, © 2003, with permission from the American Meteorological Society.

by Kubatzki et al. (2000) by including in their EMIC an interactive terrestrial vegetation module. This amplification is caused by the synergism between the sea-ice albedo–SST feedback and the vegetation–snow albedo feedback: because of warmer summers and long growing seasons during the LIG, boreal forests expanded northward. Since the snow coverage of high vegetation (trees) is less efficient than that for low

vegetation (grass or polar desert), the surface albedo thus decreases. Together with the earlier snow melting, this results in additional warming. As a result, the sea-ice area is further reduced and the area covered by boreal forests increases considerably (as an example, it doubles at 65°N), at the expense of tundra and polar desert (Figure 7) Maximum SAT differences over Eurasia increase up to 7°C in JJA, while in DJF, the Arctic

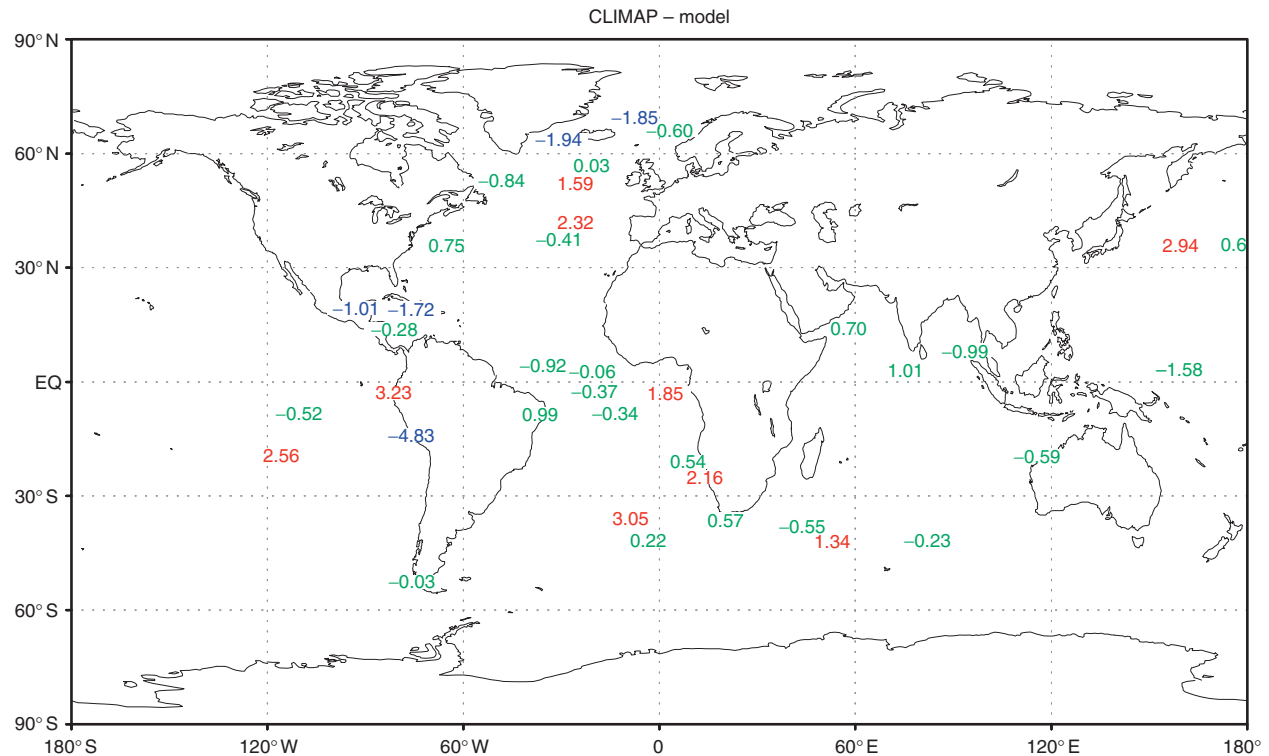


Figure 9 CLIMAP Project Members minus model-simulated annual mean SST anomalies (estimated as the summer and winter average SSTs) in degrees Celsius at CLIMAP core locations (blue indicates that the CLIMAP anomaly is colder than that predicted by the model, red indicates that it is warmer, and green indicates that there is no significant different at $1 - \sigma$ level). Reproduced from Montoya M, Crowley TJ, and von Storch H (1998) Temperatures at the last interglacial simulated by a coupled ocean-atmosphere model. *Paleoceanography* 13: 170–177, © 1998 American Geophysical Union, with permission.

warming reaches 8°C (Figure 10). In the annual mean, the Arctic Ocean and the northern continents are about 4 and 1°C warmer, respectively (not shown). Thus, the change in globally averaged temperature increases from 0.4 to 1.2°C when interactive vegetation is taken into account (see above). Because the effect is higher over land as well as at high latitudes, the good agreement with CLIMAP Project Members (1984) SSTs is not compromised. Thus, vegetation feedbacks might result in globally warmer conditions which cannot otherwise be accounted for. Such changes in vegetation are indeed consistent with vegetation reconstructions, indicating a northward expansion of boreal forests (e.g., LIGA Members, 1991).

Ocean Circulation Changes

Recent density reconstructions for the LIG indicate the presence of a fresh surface layer in the Labrador Sea which inhibited winter convection and, thus, Labrador Sea Water (LSW) formation (Hillaire-Marcel et al., 2001). This state lasted throughout the whole glacial cycle, and LSW formation resumed only during the Holocene, at about 7 ka BP. However, Denmark Strait Overflow Water (DSOW) production rates were still high. This means that, instead of the current LSW/North Atlantic deepwater (NADW) stratification, below a certain depth a single water mass occupied the North Atlantic basin.

Cottet-Puinel et al. (2005) have used the UVic coupled climate model, an EBM coupled to an OGCM, to analyze

LSW formation at the LIG. For the preindustrial climate, their model shows deepwater formation in the Labrador Sea as well as in the Irminger Sea instead of the GIN Seas, which is a common feature of coarse resolution models. However, for the LIG, deepwater formation occurs only in the Irminger Sea, in agreement with Hillaire-Marcel et al. (2001). As a result, the amplitude of the Atlantic meridional overturning circulation (AMOC), of 21 Sv ($1 \text{ Sv} = 10^6 \text{ m}^3 \text{ s}^{-1}$) in the control run, is reduced by about 6 Sv and is shallower, and Antarctic bottom water (AABW) penetrates slightly farther north, consistent with reconstructions of Duplessy et al. (1988). The model reproduces the fresh lid in the Labrador Sea found in the reconstructions. Hillaire-Marcel et al. (2001) argue that this is a consequence of enhanced export of fresh water out of the Arctic, possibly associated with an intensification of the hydrological cycle in the warmer LIG climate. However, Cottet-Puinel et al. (2004) discuss that in their model it is rather the consequence of enhanced sea-ice coverage due to the insolation decrease. As a consequence of the LSW formation halt, the model shows a decrease in SAT of up to 1.5°C centered over the North Atlantic, which is not consistent with the year-round warming at the LIG inferred from paleodata. Still, this could be due to excessive diffusion due to its relatively simple atmospheric component.

Other than that, modeling studies have not focused much on differences in the ocean circulation at the LIG relative to the present. The simulated changes are generally of small amplitude. Montoya et al. (2000) found a slight intensification of

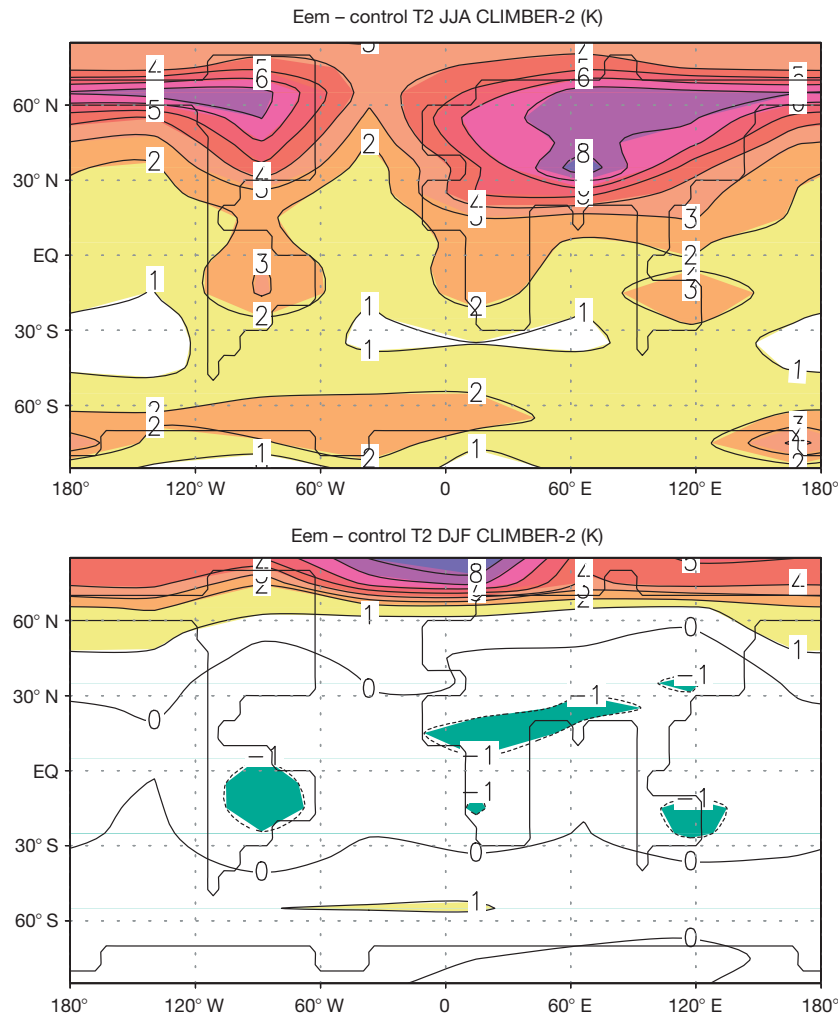


Figure 10 Departure in SAT (125 ka BP minus control) in kelvin for JJA (top) and DJF (bottom), as simulated by the CLIMBER-2 EMIC in the model version including interactive atmospheric, oceanic, and biosphere components.

the meridional overturning circulation in their AOGCM, in both the Atlantic and the Pacific Oceans, linked to enhanced evaporation at high northern latitudes. In contrast to this, the model simulation by Kubatzki et al. (2000) obtained a decrease, due to enhanced warming during the LIG. Crowley and Kim (1994) showed enhanced inflow of AABW into the Pacific, consistent with Montoya et al. (2000). Groll et al. (2005) indicate only minor changes of NADW formation rates during the LIG. These results are consistent with $\delta^{13}\text{C}$ data suggesting NADW formation rates for the LIG similar to modern ones (e.g., Adkins et al., 1997).

However, one possibility that has not been considered up to now is whether Greenland ice sheet meltwater might have contributed to the LSW halt. The extent to which Greenland ice sheet volume might have been reduced during the LIG is relevant to obtaining an estimate of its sensitivity, in particular with respect to potential future global warming. Global sea level during the LIG may have been up to 6–7 m higher than the present level, indicating a significant reduction in global ice volume corresponding to the total of the Greenland or the West Antarctic ice sheets. Cuffey and Marshall (2000) have

recently concluded that probably most of the global sea level change (4.5–5.0 m) was due to a strong reduction in the size of the Greenland ice sheet. An estimate of 5 m over 5 ka would represent a freshwater flux anomaly of ~ 0.01 Sv. This represents a very modest perturbation. Yet, the melting was strongest in southwest Greenland, closest to the Labrador Sea, and thus could potentially have affected LSW formation.

Conclusions and Discussion

Paleoclimatic reconstructions suggest widespread evidence for warmth during the LIG, especially at mid–high northern latitudes. Climate models provide plausible explanations for this warmth. Models of different complexities show a number of common features, such as enhanced JJA SAT warming over the continents leading to strengthened monsoons and high northern latitude warming throughout the year. Focusing on the Northern Hemisphere, where most of the paleodata are located and the response of the models is largest, the best agreement is found in JJA. This is consistent with the fact that

summer temperatures reflect the direct thermodynamic response to insolation changes. The DJF response, however, involves changes in dynamics. This leads to deviations in the SAT response from the direct thermodynamic response and, thus, to discrepancies between models of differing degrees of complexity and with different representations of the dynamics: while the simpler models reflect mainly the direct thermodynamic response to insolation changes, the more complex models suggest changes in the atmospheric dynamics reflecting an intensification of the AO. Through enhanced westerlies over the North Atlantic, this results in enhanced moisture and heat advection into Eurasia, which together with the sea-ice albedo–SST feedback could provide an explanation for warmth during the LIG, relative to the present, found especially at the mid- and high northern latitudes in northern winter. Yet, another possibility suggested by simpler models including interactive vegetation modules, and especially relevant at mid–high northern latitudes, is that feedbacks related to vegetation might also have played an important role. This hypothesis is supported by the evidence for boreal forests expanding further north during the LIG. Thus, the exact interplay of dynamics and biosphere during the LIG needs further clarification. The advent of comprehensive models including dynamic biosphere components will hopefully contribute to this purpose.

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See also: Pollen Records, Last Interglacial of Europe. **Glacial Climates:** Biosphere Feedbacks. **Glacial Landforms, Ice Sheets:** Evidence of Glacier and Ice Sheet Extent. **Glaciation, Causes:** Astronomical Theory of Paleoclimates. **Lake Level Studies:** Africa during the Late Quaternary. **Paleoclimate Modeling:** Data–Model Comparisons. **Paleoclimate Reconstruction:** Approaches. **Plant Macrofossil Records:** Arctic Eurasia; Arctic North America. **Sea Level Studies:** Overview.

References

- Aalsbersberg G and Litt T (1998) Multiproxy climate reconstructions for the last interglacial and early Weichselian. *Journal of Quaternary Science* 13: 367–390.
- Adkins JF, Boyle EA, Keigwin L, and Cortijo E (1997) Variability of the North Atlantic thermohaline circulation during the last interglacial period. *Nature* 390: 154–156.
- Berger A (1978) Long-term variations of daily insolation and Quaternary climatic changes. *Journal of Atmospheric Sciences* 35: 2362–2367.
- CLIMAP Project Members (1984) The last interglacial ocean. *Quaternary Research* 21: 123–224.
- Cottet-Puinel M, Weaver AJ, Hillaire-Marcel C, de Vernal A, Clark PU, and Eby M (2004) Variation of Labrador Sea water formation over the last glacial cycle in a climate model of intermediate complexity. *Quaternary Science Reviews* 23: 449–465.
- Crowley TJ and Kim KY (1994) Milankovitch forcing of the last interglacial sea level. *Science* 265: 1566–1568.
- Crowley TJ and North GR (1991) *Paleoclimatology*. New York: Oxford University Press.
- Cuffey KM and Marshall SJ (2000) Sea level rise from Greenland ice sheet retreat in the last interglacial period. *Nature* 404: 591–594.
- De Noblet N, Braconnot P, Joussaume S, and Masson V (1996) Sensitivity of simulated Asian and African summer monsoons to orbitally induced variations in insolation 126, 115 and 6 kBP. *Climate Dynamics* 12: 589–603.
- Duplessy JC, Shackleton NJ, Fairbanks RG, Labeyrie L, and Oppo D (1988) Deep water source variations during the last climatic cycle and their impact on the global deep water circulation. *Paleoceanography* 3: 343–360.
- Felis T, Lohmann G, Kuhnert H, et al. (2004) Increased seasonality in Middle East temperatures during the last interglacial period. *Nature* 429: 164–168.
- Groll N, Widmann M, Jones J, Kaspar F, and Lorenz S (2005) Simulated relationships between regional temperatures and large-scale circulation during the early last interglacial interglacial (125 kyr BP) and the pre-industrial period. *Journal of Climate* 18: 4035–4048.
- Hall A, Clement A, Thompson DWJ, Broccoli A, and Jackson C (2003) The importance of atmospheric dynamics in the Northern Hemisphere wintertime climate response to changes in the Earth's orbit. *Journal of Climate* 18: 1315–1325.
- Harrison SP, Kutzbach JE, and Behling P (1991) General circulation models, palaeoclimatic data and last interglacial climates. *Quaternary International* 10–12: 231–242.
- Harrison SP, Kutzbach JE, Prentice IC, Behling PJ, and Sykes MT (1995) The response of Northern Hemisphere extratropical climate and vegetation to orbitally induced changes in insolation during the last interglaciation. *Quaternary Research* 43: 174–184.
- Hillaire-Marcel C, de Vernal A, Bilodeau G, and Weaver AJ (2001) Absence of deep water formation in the Labrador Sea during the last interglacial period. *Nature* 410: 1073–1077.
- Joussaume S and Braconnot P (1997) Sensitivity of paleoclimate results to season definitions. *Journal of Geophysical Research* 102: 1943–1956.
- Kaspar F and Cubasch U (2005) Simulations of the Eemian interglacial and the subsequent glacial inception with a coupled ocean–atmosphere general circulation model. In: Sirocko F, Litt T, Claussen M, and Sánchez-Goni MF (eds.) *The Climate of Past Interglacials*. Amsterdam: Elsevier.
- Kaspar F, Kuhl N, Cubasch U, and Litt T (2005) A model-data comparison of European temperatures in the last interglacial interglacial. *Geophysical Research Letters* 32: L11703.
- Khodri M, Ramstein G, de Noblet-Ducoudré N, and Kageyama M (2003) Sensitivity of the northern extratropics hydrological cycle to the changing insolation forcing at 126 and 115 kyr BP. *Climate Dynamics* 21: 273–287.
- Klotz S, Guiot J, and Mosbrugger V (2003) Continental European last interglacial and early Würmian climate evolution: Comparing signals using different quantitative reconstruction approaches based on pollen. *Global and Planetary Change* 36: 277–294.
- Kubatzki C, Montoya M, Rahmstorf S, Ganopolski A, and Claussen M (2000) Comparison of a coupled global model of intermediate complexity and an AOGCM for the last interglacial. *Climate Dynamics* 16: 799–814.
- Kukla G, Bender ML, de Beaulieu JL, et al. (2002) Last interglacial climates. *Quaternary Research* 58: 2–13.
- Kutzbach JE, Gallimore RG, and Guetter PJ (1991) Sensitivity experiments on the effect of orbitally-caused insolation changes on the interglacial climate of high northern latitudes. *Quaternary International* 10–12: 223–229.
- LIGA Members (1991) Report of the 1st discussion group: The last interglacial in high latitudes of the northern hemisphere: Terrestrial and marine evidence. *Quaternary International* 10–12: 9–28.
- Montoya M, Crowley TJ, and von Storch H (1998) Temperatures at the last interglacial simulated by a coupled ocean–atmosphere model. *Paleoceanography* 13: 170–177.
- Montoya M, von Storch H, and Crowley TJ (2000) Climate simulation for 125,000 years ago with a coupled ocean–atmosphere general circulation model. *Journal of Climate* 13: 1057–1072.
- Petit JR, Jouzel J, Raynaud D, et al. (1999) Climate and atmospheric history of the past 420,000 years from the Vostok ice core, Antarctica. *Nature* 399: 429–436.
- Prell WL and Kutzbach JE (1987) Monsoon variability over the past 150,000 years. *Journal of Geophysical Research* 92: 8411–8425.
- Royer JF, Deque M, and Pestiaux P (1984) A sensitivity experiment to astronomical forcing with a spectral GCM: Simulation of the annual cycle at 125000 BP and 115000 BP. In: Berger A, Imbrie J, Hays J, Kukla G, and Saltzman B (eds.) *Milankovitch and Climate*. Dordrecht: Reidel.
- Zagwijn WH (1996) An analysis of last interglacial climate in western and central Europe. *Quaternary Science Reviews* 15: 451–469.

Last Glacial Maximum GCMs

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Introduction

General circulation models (GCMs) are valuable tools for developing a quantitative understanding of climate dynamics and climate change, and studies conducted using GCMs have provided important insights into the climate of the Last Glacial Maximum (LGM). The term ‘GCM’ is typically used to describe a numerical model in which the temporal evolution of the three-dimensional circulation of the atmosphere, ocean, or both is determined by solving the set of coupled, partial differential equations governing fluid motion. The equations are sufficiently complex that analytic solutions do not exist, and thus the equations must be solved by numerical methods. The atmosphere and ocean are divided into grid boxes, and the equations are solved for each of the boxes. Because the typical dimensions of a grid box are 100–400 km in the horizontal and 10–200 m (ocean) or 50–2000 m (atmosphere) in the vertical, a global GCM will have 10^5 – 10^6 grid boxes. The numerical solution of the equations governing fluid motion for this large number of grid boxes requires substantial computing power. Washington and Parkinson (2005) provided a detailed description of GCMs and their construction.

GCMs are among the most comprehensive members of the more general category of climate models. Simpler climate models reduce the complexity of the climate system, most often by averaging over one or more spatial dimensions. Although such simpler climate models have been useful in understanding Quaternary climate change, they are not the focus of this article. Instead, the focus is limited to results that have been obtained with the following types of climate models, listed in order of increasing complexity: atmospheric GCMs using prescribed sea-surface temperatures (SSTs) as a lower boundary condition, atmospheric GCMs coupled to highly simplified models of the ocean-mixed layer, and atmospheric GCMs coupled to oceanic GCMs. The decision to narrow the focus is made in order to keep the scope of this article manageable and to recognize the historical importance of these models.

Types of GCMs

Atmospheric GCMs with prescribed SSTs were used in the earliest attempts to model glacial climate, and were made possible by the pioneering work of the CLIMAP Project (1981) in reconstructing the global SST distribution for the LGM. Such models greatly simplify the interaction between the atmosphere and ocean by eliminating the influence of the atmosphere on the ocean, but can be useful in determining how the atmospheric circulation and land-surface climate might respond to a given set of surface boundary conditions. An important weakness of atmospheric GCMs with prescribed SSTs is that they do not conserve energy. The ocean can serve as

an infinite source or sink of energy, with a compensating energy flux at the top of the atmosphere. In a specific example, LGM simulations with atmospheric GCMs using SSTs prescribed according to Climate: Long-range Investigation, Mapping, and Prediction (CLIMAP) typically are in radiation deficit, with emitted radiation at the top of the atmosphere exceeding incoming radiation. If the atmospheric component of the model is assumed to be realistic, the radiation deficit would imply that the SSTs are too warm in an aggregate sense, as cooling would ensue if the SSTs were no longer constrained. Alternatively, the SSTs could be correct and the radiation deficit could result from deficiencies in the atmospheric model.

Atmospheric GCMs coupled to highly simplified models of the ocean-mixed layer alleviate the need for the global distribution of SST as a prescribed lower boundary condition. In these models, sometimes called atmosphere-mixed layer ocean models, the upper ocean is represented by a motionless slab of water of uniform temperature. (Such models are also known as atmosphere-slab ocean models in recognition of the simplicity with which the upper ocean is represented.) Atmosphere-slab ocean models, which typically include a sea-ice model in which ice can form or melt in response to thermodynamic processes, improve upon models with prescribed SSTs by allowing a two-way thermodynamic interaction between the atmosphere and the ocean. SST is an output variable in this type of climate model, which facilitates comparisons with reconstructions of glacial climate. The simplified ocean does not represent the effects of ocean circulation or thermal communication between the mixed layer and the deep ocean. In most atmosphere-slab ocean models, an artificial heat flux (often called a ‘q-flux’) is prescribed at each ocean grid point in order to represent the convergence or divergence of heat that would result from ocean heat transport in the real climate system. The inability to simulate changes in ocean heat transport or the dynamical interactions between the atmosphere and ocean (e.g., wind-driven ocean circulation or changes in ocean circulation resulting from density changes due to buoyancy fluxes) is a significant limitation of atmosphere-slab ocean models.

The coupling of an atmospheric GCM with an oceanic GCM yields a more complete model of the climate system with the capability of simulating changes in circulation in the ocean as well as the atmosphere. Because such coupled atmosphere–ocean GCMs (AOGCMs) can simulate two-way dynamic and thermodynamic interactions between the atmosphere and ocean, they can simulate coupled modes of climate variability such as the El Niño–Southern Oscillation (ENSO) phenomenon, a quasiperiodic fluctuation in the temperature of the central and eastern equatorial Pacific with far-ranging implications for climate. They can also simulate changes in the intensity of coastal upwelling in regions induced by changes in surface winds, or changes in ocean overturning (i.e., the so-called ocean conveyor; Broecker, 1991), that result from changes in the fluxes

of heat and fresh water. The additional realism of AOGCMs comes with a cost, however. Because the equations governing fluid motion must be solved for both the atmosphere and the ocean, and the long timescales ($\sim 10^3$ years) of the deep ocean circulation require AOGCMs to be integrated longer before they fully adjust to the forcing, AOGCMs require substantially greater amounts of computer time than atmosphere-only or atmosphere-slab ocean GCMs.

Design of GCM Experiments

For past epochs in which the climate was substantially different from today, such as the LGM, there are differences in atmospheric composition, Earth–Sun geometry, continental configuration, solar irradiance, or surface characteristics that are capable of altering the radiation balance, either locally or globally. Climate model simulations are conducted by altering one or more of these characteristics and comparing the resulting climate to a simulation of modern climate. Some of these changes, such as Earth–Sun geometry, continental configuration, and solar irradiance, are clearly external to the climate system. Others, such as atmospheric composition or the spatial distribution of continental ice sheets, are influenced by processes internal to the climate system.

A useful paradigm for thinking about the design of climate model experiments is the distinction between external forcing and internal feedbacks. Internal feedbacks are climate system processes that are explicitly represented within the model, whereas external forcing involves processes that are not represented in the model. A change that occurs outside the climate system, such as an increase in the brightness of the Sun, is considered to be an external forcing in this paradigm. The growth and decay of large continental ice sheets would also be an external forcing in most GCMs, because they do not include glacial dynamics. In a more comprehensive model that includes glacial dynamics, ice-sheet growth and decay would be an internal feedback. An unavoidable consequence of this paradigm that some may find counterintuitive is that whether a process is considered to be an external forcing or an internal feedback depends upon the construction of the model.

As an example, atmosphere-only GCMs require that the global distribution of SST and sea ice be specified. When simulating the LGM climate with such a model, the changes in specified SST and sea ice should be regarded as an external forcing. Thus, atmosphere-only GCMs cannot be used to understand the processes that led to changes in SST and sea ice during the LGM, but they can be used to simulate the atmospheric circulation and the spatial distribution of precipitation

that would accompany the specified changes in SST, sea ice, and other external forcings. Such models have also been useful for simulating the transport of isotopic tracers in water and the location and intensity of storm tracks during the LGM.

Atmosphere-slab ocean GCMs and AOGCMs do not require the specification of SST and sea ice, allowing them to be used when the objective is to determine if a model can simulate realistic changes in temperature in response to LGM external forcings (e.g., Manabe and Broccoli, 1985a). Such simulations allow the realism of internal feedbacks within the model to be evaluated, at least to the extent to which LGM temperature changes can be constrained by paleodata. Similarly, models that determine changes in temperature internally can be useful in understanding the mechanisms responsible for spatial variations in LGM cooling.

The external forcings that are relevant to most LGM simulations with atmosphere-slab ocean GCMs and AOGCMs are changes in atmospheric composition (including greenhouse gases and mineral dust), Earth–Sun geometry (i.e., Earth’s orbital parameters), continental ice-sheet distribution, sea level, and vegetation (Table 1). (Recall that changes in atmospheric composition, ice-sheet distribution, and vegetation are considered external forcings because most GCMs do not include biogeochemical cycling, glacial dynamics, or interactive vegetation.) Concentrations of greenhouse gases at the LGM are known from ice-core evidence, and orbital parameters for that time can also be calculated very accurately, making both of these forcings very well constrained. Sea-level changes are also relatively well known from multiple lines of evidence. Glacial geology provides reliable information about continental ice margins at the LGM, although there are larger uncertainties regarding LGM ice sheets that terminate on the now-inundated continental shelves. Ice-sheet thicknesses can be inferred from glacial geology and inverse modeling of postglacial isostatic adjustment, but with considerable uncertainties. The largest uncertainties involve vegetation changes and mineral dust concentrations. Insufficient data exist to develop a global vegetation map for the LGM, although there is ample evidence of vegetation changes on local and regional scales. Data from ice and sediment cores indicate increased dust deposition during the LGM, but these data are insufficient to construct a reliable three-dimensional distribution of atmospheric dust concentrations, although some estimates are available from model simulations.

The importance of various external forcings in contributing to the LGM climate can be quantified in terms of the radiative forcing (Table 1). Radiative forcing is the hypothetical change in the global mean radiative balance that would result from changing a particular forcing instantaneously without allowing

Table 1 Climate forcing mechanisms for the Last Glacial Maximum (LGM)

<i>Climate forcing mechanism</i>	<i>Radiative forcing</i>	<i>Confidence</i>	<i>References</i>
Greenhouse gases	-2 to -3 W m^{-2}	Very high	Hansen et al. (1993), Hewitt and Mitchell (1997), and Broccoli (2000)
Continental ice sheets	-2 to -3.3 W m^{-2}	Moderate	Hansen et al. (1993), Hewitt and Mitchell (1997), and Broccoli (2000)
Orbital parameters	$\pm 0.2 \text{ W m}^{-2}$	Very high	Hewitt and Mitchell (1997) and Broccoli (2000)
Sea level	-0.3 to -0.5 W m^{-2}	Moderate	Hewitt and Mitchell (1997) and Broccoli (2000)
Vegetation	0 to -1 W m^{-2}	Low	Crucifix and Hewitt (2005)
Atmospheric dust content	-1 to -2.3 W m^{-2}	Low	Hansen et al. (1993) and Claquin et al. (2003)

the climate system to respond. (For reasons that are beyond the scope of this article, radiative forcing is most commonly evaluated at the tropopause, which is the top of the 10–15 km deep atmospheric layer that undergoes efficient thermal communication with the surface.)

Some of the LGM radiative forcings can be quantified with relatively modest uncertainties. The two largest sources of radiative forcing of the Ice Age climate are the reduction in atmospheric greenhouse gas concentrations and the expansion of continental ice sheets, both of which are negative. Of the former, most of the forcing comes from the glacial reduction in carbon dioxide, with additional contributions from methane and nitrous oxide. Radiative forcing from the expanded LGM ice sheets is confined to the regions covered by continental ice, and is particularly intense near the southern margins of the Laurentide and Eurasian ice sheets where solar radiation (and hence reflection) is especially strong. The reduction in sea level at the LGM induces a modest negative radiative forcing, as the additional land areas have a higher surface albedo than the ocean surface. The radiative forcing due to the changes in orbital parameters is very small, despite the importance of orbital forcing for ice-sheet growth and decay, because of the similarity between the modern and LGM orbital configurations.

The radiative forcings contributed by changes in vegetation and atmospheric dust content are much more difficult to quantify. The global distribution of LGM vegetation is not known with any certainty, although there have been attempts to map it (Adams and Faure, 1997). One might expect that the magnitude of the radiative forcing from vegetation changes would be limited by some compensation between regions of increased and decreased surface albedo, but constraints are relatively weak. The uncertainties associated with reconstructing the global three-dimensional distribution of atmospheric dust concentration preclude a well-constrained value for dust forcing, and additional uncertainties in the optical properties of mineral dust add to the problem. Nonetheless, it has been estimated that LGM dust may contribute a substantial negative forcing, perhaps only exceeded by the greenhouse gas and ice-sheet forcings.

Results from GCM Experiments

GCM simulations of glacial climate have been conducted in substantial numbers, making it impractical to describe the findings from all of them in detail. There are, however, certain climate processes that have received special attention, and they are discussed in this section. A brief summary of these results is provided in Table 2.

Interhemispheric Asymmetry of Cooling

Virtually all LGM climate simulations are characterized by greater cooling in the Northern Hemisphere, particularly in the immediate vicinity of the ice sheets, but also extending over the subpolar and mid-latitude North Atlantic in many cases. In contrast, simulated cooling tends to be smaller in the tropics and at comparable latitudes of the Southern Hemisphere. The large cooling over the ice sheets has multiple causes. The high albedo of ice-covered surfaces results in

Table 2 Summary of changes in climate from GCM simulations of the LGM

<i>Spatial distribution of cooling</i>	<i>Largest cooling over Laurentide and Eurasian ice sheets; Northern Hemisphere cools more than Southern Hemisphere</i>
Atmospheric circulation	Polar jet stream splits around Laurentide ice sheet; enhanced ridge of high pressure in upper troposphere over western North America
Monsoons	Weakening of the South Asian monsoon; weakening of cross-equatorial flow in the Indian Ocean; decrease in intensity of south westerly winds over the Arabian Sea
Tropical cooling	Models simulate overall tropical cooling; magnitude varies among different models
Meridional overturning circulation	Weakens in two models and strengthens slightly in another; North Atlantic Deep Water does not sink as deep
Tropical Pacific climate and air–sea interactions	Amplitude of ENSO variability is increased in one model; mean climate becomes more ‘La Niña-like’ in two others

decreased absorption of solar radiation. Ice-sheet surfaces at the LGM were elevated in some locations by more than 2 km, with attendant cooling due to the decrease of temperature with height in the troposphere. Ice-covered surfaces also limit summer warming, as the latent heat required by surface melting acts as an energy sink.

The advection of cold air that originates over the ice sheets contributes to cooling in proximity to the ice sheets, allowing the cooling effects of the ice sheets to spread beyond their boundaries. When the atmospheric circulation carries ice-cooled air across the ocean surface, as in the North Atlantic, it can promote sea-ice growth, which further cools the atmosphere by reflecting solar radiation and insulating the atmosphere from the relatively warm ocean waters below (Manabe and Broccoli, 1985b). Experiments in which only the ice-sheet forcing was employed indicate that the thermal effects of continental ice sheets are primarily confined to the Northern Hemisphere, and thus constitute the primary cause of interhemispheric asymmetry in the magnitude of simulated LGM cooling. Some recent studies suggest that this asymmetry is capable of altering the atmospheric circulation in the tropics, leading to changes in the pattern of tropical precipitation (Chiang et al., 2003).

Alteration of the Northern Hemisphere Jet Stream

Some of the earliest simulations of LGM climate found that the expanded ice sheets of the Northern Hemisphere were capable of substantially altering the atmospheric circulation. Because the ice sheets can extend as much as 2–3 km above sea level, they act as topographic barriers to the circulation. The location of the ice sheets in the path of the polar jet stream creates the potential for the jet to be altered by their presence.

In simulations of the LGM climate, [Kutzbach and Guetter \(1984\)](#) and [Manabe and Broccoli \(1984\)](#) found that the polar jet stream was split into two branches by the Laurentide ice sheet. One branch skirted the southern edge of the ice sheet, and the other curved anticyclonically along its northern periphery.

The presence of the Laurentide ice sheet also amplified the upper tropospheric ridge of high pressure that is found over western North America in winter, enhancing southerly flow into what is now Alaska and producing modest LGM winter warming. The combination of the enhanced upper tropospheric ridge and the northern branch of the split polar jet stream facilitated the export of cold air to the North Atlantic, as discussed in Section ‘[Interhemispheric Asymmetry of Cooling](#).’ The southern branch of the jet enhanced winter precipitation in southwestern North America, leading to a wetter climate there.

More recent simulations of LGM climate have used a reconstruction of glacial topography in which the surface elevations of the ice sheets are somewhat lower than in the earlier simulations. This reduction in surface elevation reduces the barrier effect of the Laurentide ice sheet, with an attendant reduction in the degree to which the polar jet stream splits, although the resulting circulation is qualitatively similar to the results from the earlier simulations.

Monsoons

The nonuniform cooling in LGM simulations has an impact on monsoon circulations, particularly those in the Northern Hemisphere. [Prell and Kutzbach \(1987\)](#) identified a weakening of the South Asian monsoon system in their simulation of LGM climate with an atmosphere-only GCM. The weakening of the South Asian monsoon, which was attributed to a reduction of land temperatures during boreal summer, was accompanied by weaker cross-equatorial flow in the Indian Ocean and a decrease in the intensity of monsoon-related southwesterly winds over the Arabian Sea. More recent model simulations with atmosphere-slab ocean GCMs and AOGCMs produce a similar response, although the details of the circulation changes can be somewhat different because the SST pattern simulated by these models does not exactly correspond to the CLIMAP pattern used by [Prell and Kutzbach \(1987\)](#).

The glacial monsoon response is consistent, in terms of its physical mechanism, with other climate model simulations that indicate a strong response of the Asian monsoon to precessional forcing. The monsoon is strong when the phase of the precession cycle is such that perihelion occurs in boreal summer, which increases continental temperatures and the land–sea temperature contrast. The monsoon is weaker when perihelion occurs in winter, since this orbital configuration results in cooler summer temperatures, analogous to the glacial case.

Tropical Cooling

The magnitude of tropical cooling during the LGM has been a source of controversy for quite some time. The original reconstruction of glacial SSTs by CLIMAP indicated little or no cooling over large areas of the tropics and subtropics. Apparent

inconsistencies between relatively warm tropical SST and indicators of larger cooling from the tropical continents (snow-line depression, noble gases in aquifers, and pollen) have prompted questions about the actual magnitude of LGM tropical cooling (e.g., [Broecker, 1995](#); [Rind and Peteet, 1985](#)). More recent proxies for SST such as alkenone unsaturation and Mg/Ca ratio indicate that LGM tropical cooling was larger than the CLIMAP reconstruction.

Virtually, all LGM simulations with atmosphere-slab ocean GCMs or AOGCMs have simulated reduced temperatures throughout the tropics, although magnitudes vary among model simulations. Using a number of simulations performed within the framework of the Paleoclimate Modeling Intercomparison Project (PMIP), [Pinot et al. \(1999\)](#) found that all atmosphere-slab ocean GCMs examined produced cooling over the Indian and Pacific Oceans that contrasted with the very limited cooling indicated by CLIMAP. For roughly half of these simulations, the simulated terrestrial cooling in the tropics was in good agreement with paleodata. Focusing on the results of one particular PMIP simulation, [Broccoli \(2000\)](#) found that the simulated cooling was consistent with some tropical temperature proxies, but smaller than the cooling inferred from other proxies. He also noted that the differences in tropical cooling estimates based on different proxies were relatively large, precluding the use of tropical paleotemperatures as a conclusive test of climate models.

Recent LGM simulations with AOGCMs indicate that dynamical processes such as the ventilation of thermocline and intermediate waters from higher latitudes ([Liu et al., 2002](#)) can also influence the magnitude to tropical cooling, at least locally or regionally. Changes in wind stress and concomitant equatorial upwelling are thought to be responsible for enhanced cooling in the eastern equatorial Pacific in [Hewitt et al. \(2001\)](#).

Meridional Overturning Circulation in the Ocean

The advent of AOGCM simulations of the LGM has opened new areas of investigation for climate modelers. A particular focus of AOGCM simulation studies has been the response of the meridional overturning circulation (MOC) in the ocean to glacial forcing. The MOC, which is also known as the thermohaline circulation because it is thought to be driven by the effects of temperature and salinity on the density of sea water, is a global circulation system in which cold, salty surface waters sink in the subpolar North Atlantic, forming North Atlantic Deep Water (NADW), which spreads throughout the global ocean at depth before gradually upwelling into the thermocline.

Paleoceanographic evidence from the LGM indicates that NADW did not sink as deep as it does today, and there is some evidence that the overall volume of water involved in the MOC may have been less, although the latter interpretation remains somewhat controversial. Thus, there is an understandable interest in developing further insights into this question based on the results from glacial AOGCM simulations.

The behavior of the MOC in AOGCM simulations of the LGM varies. In an experiment with the Hadley Centre model HadCM3 ([Hewitt et al., 2001](#)), the MOC intensified and was accompanied by an increase in northward heat transport in the North Atlantic, producing relatively warm SSTs in parts of that ocean. Studies with the National Center for Atmospheric

Research model CSM (Shin et al., 2003) and the Canadian Centre for Climate Modeling and Analysis (CCCma) model (Kim et al., 2003) reported a weakening of the MOC and a reduction in the downward penetration of NADW. The weakening of NADW formation and the MOC in CCSM was attributed to changes in Antarctic Bottom Water formation that result from the expansion of sea ice in the Southern Ocean.

At present there is no definitive explanation for the difference between the HadCM3 response and that of the other models, although it is possible to speculate that HadCM3 had not completely adjusted to LGM forcing. It is worth noting that both HadCM3 and the CCCma model (Kim et al., 2002) experienced a transient increase in MOC intensity immediately after the LGM forcing was initiated, but the CCCma model eventually settled in a state with a weaker MOC.

Tropical Pacific Climate and Air–Sea Interactions

Air–sea interactions in the tropical Pacific Ocean give rise to the ENSO phenomenon. The warm phase of ENSO, known as El Niño, is characterized by an anomalous warming of the central and eastern equatorial Pacific, and SST anomalies have the opposite sign during the cold, or La Niña phase. In many parts of the world, primarily in the tropics, ENSO is the largest source of interannual climate variability.

The importance of ENSO in the modern climate has led many to wonder if ENSO variability was different in past climates. Recent AOGCM simulations provide some insights regarding the LGM. Otto-Bliesner et al. (2003) found that the amplitude of ENSO variability was enhanced during the LGM. Weakening of the zonal SST gradient, wind stresses, and upwelling in the tropical Pacific region and an intensification of the tropical thermocline were identified as causes of this enhanced variability. Other investigators have noted changes in the mean climate of the tropical Pacific, with the SST distribution being more La Niña-like (i.e., greater cooling in the eastern equatorial Pacific than in locations farther west) in the simulations by Hewitt et al. (2003) and Kim et al. (2003). The implications of these changes in mean climate on the amplitude of ENSO variability in these models are unknown.

Future Directions

Increases in computing power and experience with running AOGCMs are allowing these models to become more widely applied in studies of glacial climate. A number of challenges continue to exist, such as identifying the best way to initialize glacial AOGCM simulations and determining when a model was settled into a representative quasiequilibrium climate state. Models of intermediate complexity, in which both atmosphere and ocean are represented by simplified GCMs, can run much faster and should prove useful in guiding the use of full AOGCMs.

An important trend in climate modeling is toward the development of more comprehensive models of the climate system that include atmospheric and ocean chemistry and interactions with the biosphere. This trend is providing new directions to glacial climate simulations. For example, many modeling groups are developing the capability to include

isotopic tracers in AOGCMs, building upon past work that was done using atmosphere-only GCMs (Joussaume et al., 1984). Models that include interactive vegetation have become more common, and there is evidence that vegetation feedbacks can have important effects on LGM climate change on local and regional scales (e.g., Crucifix and Hewitt, 2005).

For simulating the climates of the Quaternary, the ultimate goal may be a model of the earth system that has an AOGCM at its core, but also includes the necessary elements to predict the growth and decay of continental ice sheets, the concentration of greenhouse gases and aerosols (including dust) in the atmosphere, the spatial distribution of vegetation, and the three-dimensional distribution of isotopic tracers in the atmosphere and ocean. A comprehensive model of this kind would allow simulations in which the only forcing comes from the variations in Earth's orbital parameters. The ability to conduct such experiments would represent an important step toward a thorough understanding of the causes of the large temporal variations in Quaternary climate.

See also: **Glacial Climates:** Biosphere Feedbacks; Effects of Atmospheric Dust; Thermohaline Circulation; Volcanic and Solar Forcing. **Glaciation, Causes:** Astronomical Theory of Paleoclimates. **Paleoclimate Modeling:** Paleo-ENSO; Quaternary Environments.

References

- Adams JM and Faure H (1997) Paleovegetation maps of the Earth during the Last Glacial Maximum and the early and mid-Holocene: An aid to archaeological research. *Journal of Archaeological Science* 24: 623–647.
- Broccoli AJ (2000) Tropical cooling at the Last Glacial Maximum: An atmosphere-mixed layer ocean model simulation. *Journal of Climate* 13: 951–976.
- Broecker WS (1991) The great ocean conveyor. *Oceanography* 4: 79–89.
- Broecker WS (1995) *The Glacial World According to Wally*. Palisades, NY: Eldigio Press.
- Chiang JCH, Biasutti M, and Battisti DJ (2003) Sensitivity of the Atlantic intertropical convergence zone to Last Glacial Maximum boundary conditions. *Paleoceanography* 18: 1094. <http://dx.doi.org/10.1029/2003PA000916>.
- Claquin T, Roelandt C, Kohfeld KE, et al. (2003) Radiative forcing of climate by ice-age atmospheric dust. *Climate Dynamics* 20: 193–202.
- CLIMAP Project Members (1981) Seasonal reconstructions of the Earth's surface at the last glacial maximum. In: *Geological Society of America Map and Chart Series MC-36*, 18. Available online at <http://geosociety.org>.
- Crucifix M and Hewitt CD (2005) Impact of vegetation changes on the dynamics of the atmosphere at the Last Glacial Maximum. *Climate Dynamics* 25: 447–459.
- Hansen J, Lacis A, Ruedy R, Sato M, and Wilson H (1993) How sensitive is the world's climate? *National Geographic Society Research and Exploration* 9: 142–158.
- Hewitt CD, Broccoli AJ, Mitchell JFB, and Stouffer RJ (2001) A coupled model study of the Last Glacial Maximum: Was part of the North Atlantic relatively warm? *Geophysical Research Letters* 28: 1571–1574.
- Hewitt CD and Mitchell JFB (1997) Radiative forcing and response of a GCM to ice age boundary conditions: Cloud feedback and climate sensitivity. *Climate Dynamics* 13: 821–834.
- Hewitt CD, Stouffer RJ, Broccoli AJ, Mitchell JFB, and Valdes PJ (2003) The effect of ocean dynamics in a coupled GCM simulation of the Last Glacial Maximum. *Climate Dynamics* 20: 203–218.
- Joussaume S, Jouzel J, and Sadourny R (1984) A general circulation model of water isotopes cycles in the atmosphere. *Nature* 311: 24–29.
- Kim SJ, Flato GM, and Boer GJ (2003) A coupled climate model simulation of the Last Glacial Maximum. Part 2: Approach to equilibrium. *Climate Dynamics* 20: 635–661.
- Kim SJ, Flato GM, Boer GJ, and McFarlane NA (2002) A coupled climate model simulation of the Last Glacial Maximum. Part 1: Transient multi-decadal response. *Climate Dynamics* 19: 515–537.

- Kutzbach JE and Guetter PJ (1984) Sensitivity of late-glacial and Holocene climates to the combined effects of orbital parameter changes and lower boundary condition changes: "Snapshot" simulations with a general circulation model for 18, 9 and 6 ka BP. *Annals of Glaciology* 5: 85–87.
- Liu ZY, Shin SI, Otto-Bliesner B, Kutzbach JE, Brady EC, and Lee DE (2002) Tropical cooling at the Last Glacial Maximum and extratropical ocean ventilation. *Geophysical Research Letters* 29: 1409. <http://dx.doi.org/10.1029/2001GL013938>.
- Manabe S and Broccoli AJ (1984) Ice-age climate and continental ice sheets: Some experiments with a general circulation model. *Annals of Glaciology* 5: 100–105.
- Manabe S and Broccoli AJ (1985a) A comparison of climate model sensitivity with data from the Last Glacial Maximum. *Journal of the Atmospheric Sciences* 42: 2643–2651.
- Manabe S and Broccoli AJ (1985b) The influence of continental ice sheets on the climate of an ice age. *Journal of Geophysical Research* 90: 2167–2190.
- Otto-Bliesner B, Brady EC, Shin SI, Liu ZY, and Shields C (2003) Modeling El Niño and its tropical teleconnections during the last glacial–interglacial cycle. *Geophysical Research Letters* 30: 2198. <http://dx.doi.org/10.1029/2003GL018553>.
- Pinot S, Ramstein G, Harrison SP, et al. (1999) Tropical paleoclimates at the Last Glacial Maximum: Comparison of Paleoclimate Modeling Intercomparison Project (PMIP) simulations and paleodata. *Climate Dynamics* 15: 857–874.
- Prell WL and Kutzbach JE (1987) Monsoon variability over the past 150,000 years. *Journal of Geophysical Research* 92: 8411–8425.
- Rind D and Peteet D (1985) Terrestrial conditions at the Last Glacial Maximum and CLIMAP sea surface temperature estimates: Are they consistent? *Quaternary Research* 24: 1–22.
- Shin SI, Liu Z, Otto-Bliesner B, Brady EC, Kutzbach JE, and Harrison SP (2003) A simulation of the Last Glacial Maximum climate using the NCAR-CCSM. *Climate Dynamics* 20: 127–151.
- Washington W and Parkinson CL (2005) *An Introduction to Three-Dimensional Climate Modeling*. Sausalito, CA: University Science Books.

Introduction

Teleconnections, such as El Niño, are climatic relationships over long distances that provide predictability of the climate system at interannual to decadal timeframes. They are often expressed as a single number to represent contrasting states of the climate system, often dealing with changes in prominent modes of atmospheric circulation. Teleconnections and their associated climatic changes at interannual and decadal time scales occur at frequencies every 1–100 years (Martinson et al., 1998). An understanding of these teleconnections at these time scales is extremely vital to society because they provide the framework in planning schemes for water resources and agriculture. For example, the monsoons of Asia and Africa affect hundreds of millions of people, with variations of occurrence at shorter time scales approximately every 3–7 years, and severe drought occasionally causing widespread famine.

Climatologists have studied interannual–decadal climatic changes in modern instrumental records, but records normally do not extend farther back than a century for most areas. Also during the twentieth century, anthropogenic climatic change became more apparent, causing some difficulties in disentangling anthropogenic from natural climatic effects such as involving fossil fuel-induced global warming. Therefore, a longer perspective on climatic changes at these short time scales contributes significantly to a further understanding on natural climatic variability and its causes (Overpeck and Webb, 2000). Furthermore, a longer historical perspective of climatic changes at interannual and decadal time scales can aid in analyzing the societal responses to climatic changes. This article describes an overview on what scholars know about the synoptic climatology of teleconnections such as the El Niño phenomena, the current status on how paleoclimatologists are utilizing proxy data to reconstruct these important teleconnections, and some potential forcing mechanisms of past El Niño/Southern Oscillation (ENSO) events.

Teleconnections

ENSO

Clearly, the most important source of interannual climatic variability at the global scale is the ENSO phenomenon. El Niño is defined as an anomalous warming of sea-surface temperatures of approximately 2 °C or more in the central and eastern Pacific Ocean of the Southern Hemisphere (Glantz, 1996). It is named after the ‘Christ Child’ since it usually initially appears around Christmas, off the western coast of South America. Normally, the atmospheric circulation in the South Pacific is characterized by high pressure off western South America and low pressure over Australia. However, during an El Niño year, lower pressure and increased convection are prevalent in the central and eastern South Pacific,

and higher pressure occurs over Australia, thus defining the ‘Southern Oscillation.’ Sea-level pressure differences between Tahiti, Indonesia, and Darwin, Australia define the Southern Oscillation Index (SOI), with abnormally negative SOI values indicating warm events and abnormally positive values indicating cold events. El Niño events, also termed as ‘warm events,’ occur approximately every 3–7 years although some events may occur in back-to-back years or may be lacking for up to a decade. Cold events or La Niña, also generally occur at the same time frequency and are regarded as part of ENSO. La Niña is the opposite condition of El Niño, associated with abnormal cooling of sea-surface temperatures in the central and eastern South Pacific, higher pressure off western South America, and lower pressure over eastern Australia.

During ENSO events, energy exchanges occur between the oceans and atmosphere, causing changes in the Southern Oscillation that eventually affect atmospheric circulation, temperature, and precipitation anomalies elsewhere over the globe. Climatologists have concluded that tropical forcing causes similar teleconnections to reoccur during individual warm and cold events (Kiladis and Diaz, 1989). Composite anomaly maps of 500 mb heights (upper air flow) reveal some generalized synoptic patterns that occur during El Niño (SOI negative) and La Niña (SOI positive; Figure 1). Note that the contrasting SOI extremes do not necessarily reveal completely opposite characteristics for every region. However, some generalizations are clearly apparent for some ENSO-sensitive regions. For example, a trough-ridge-trough pattern is evident over the North Pacific Ocean and adjacent North America during El Niño, while a ridge-trough-ridge pattern appears during La Niña. The increased troughing into the southeastern United States in the El Niño pattern would enable more colder air to penetrate into the region, while the pattern for La Niña indicates ridging over the southeastern United States that would create warmer temperatures. Precipitation responses during El Niño and La Niña also affect some regions with contrasting anomalies (Figure 2). For example, northern Mexico and the southwestern United States tend to be wetter during El Niño but drier during La Niña. Precipitation anomalies over northeastern Brazil indicate drier conditions during El Niño but wetter conditions during La Niña.

Note that some individual ENSO events may cause some climate anomaly patterns that differ from what is normally expected if such an event is of extremely high magnitude. For example, the 1982–83 and 1997–98 El Niños were very strong warm events documented in modern instrumental records, with their global teleconnections differing from many others that are typical of warm events. Furthermore, the relationships between tropical sea-surface temperatures and mid-latitude climate may change through time. We must therefore look at the paleoclimate record to gain a longer perspective of ENSO variability.

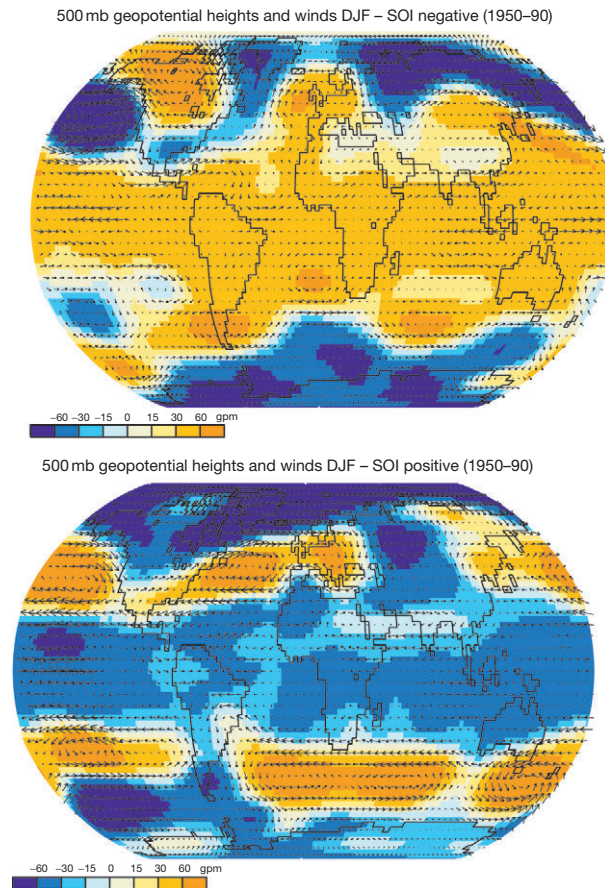


Figure 1 Composite anomaly maps of winter 500 mb heights and winds during SOI negative and SOI positive events. Negative anomalies indicate increased cyclonic flow, and positive anomalies indicate increased anticyclonic flow. Data are from the NCEP-Reanalysis Project (Kalnay et al., 1996).

Other Teleconnection Patterns

Two other important teleconnections affecting mid-latitude climate in the Northern Hemisphere at interannual timescales are the North Atlantic Oscillation (NAO) and Pacific North American (PNA) patterns. The NAO shows variation in the strength of the westerlies in the North Atlantic due to variations in the Icelandic low and Azores high (Rogers, 1984). The 500 mb height anomalies of the positive mode of the NAO indicate an anomalously stronger Icelandic low and a weak Azores high (Figure 3). Temperature responses to a positive NAO mode indicate cooler temperatures over Greenland and warmer temperatures over much of central and northern Europe (Figure 4). A weaker Icelandic low and stronger Azores high generally correspond with conditions of the negative mode, with warmer temperatures over Greenland and cooler temperatures over most of Europe (Figures 3 and 4). Variations in the NAO result from changes of North Atlantic sea-surface temperatures and ocean circulation, and perhaps sea ice at higher latitudes. The PNA teleconnection represents variations in mid-tropospheric flow that mostly affects North American climate (Leathers et al., 1991). A positive PNA pattern involves a trough in the east-central North Pacific, a ridge

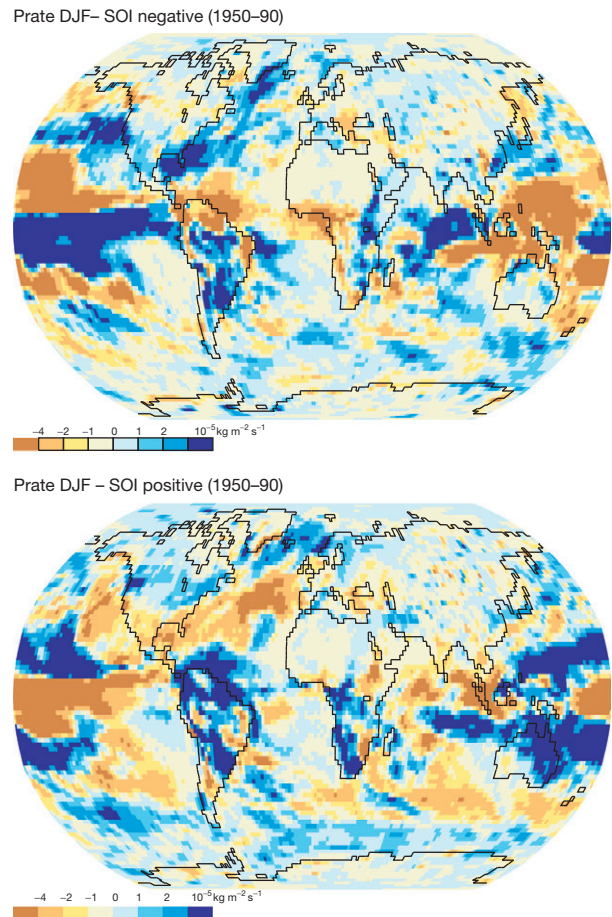
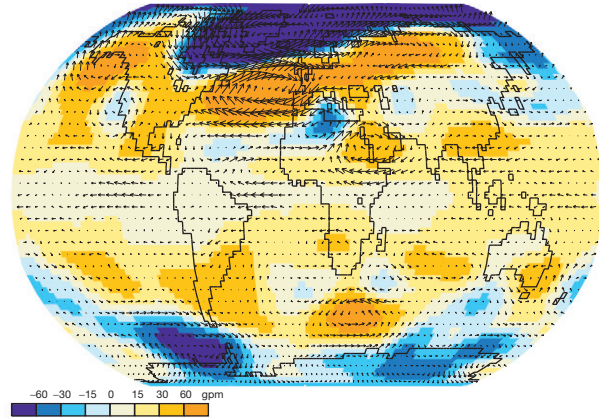


Figure 2 Composite anomaly maps of winter precipitation rates during SOI negative and SOI positive events. Negative anomalies indicate increased dryness, and positive anomalies indicate increased wetness. Data are from the NCEP-Reanalysis Project (Kalnay et al., 1996).

over western North America, and a trough over eastern North America, and a negative PNA pattern indicates opposite mid-tropospheric flow. Variations in the PNA pattern probably occur due to variations in sea-surface temperatures in the western and equatorial Pacific, and perhaps due to several climatic variables in East Asia.

The Pacific Decadal Oscillation (PDO) is an ‘ENSO-like’ phenomena that exhibits more persistent (decadal) variability as opposed to interannual (Zhang et al., 1997). Its index has been calculated based on Pacific Ocean sea-surface temperatures (Mantua et al., 1997). Positive PDO indices are indicative of colder SSTs in the north central Pacific Ocean and warmer SSTs off the west coast of North America – and negative PDO indices indicate the opposite condition. Its signal is a bit more prominent in the Mid-Latitudes and is probably driven more by sea-surface temperatures in the Mid-Latitudes, though tropical forcing cannot be completely excluded (Latif and Barnett, 1996). The Arctic Oscillation (AO) also generally exhibits somewhat more decadal-type behavior (Thompson and Wallace, 1998). The 500 mb winter composite anomaly maps of both AO modes reveal contrasting atmospheric circulation patterns for the polar regions of the AO, as compared to those at Mid-Latitudes (Figure 5). Winter temperatures during

500 mb geopotential heights and winds DJF – NAO positive (1950–90)



500 mb geopotential heights and winds DJF – NAO negative (1950–90)

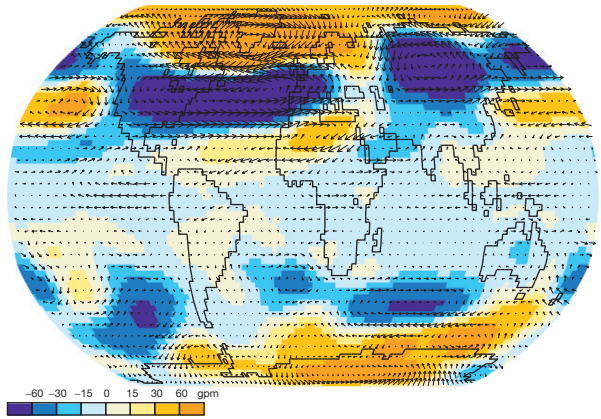
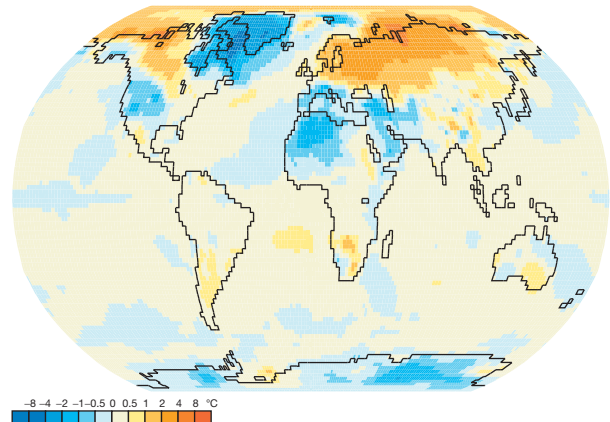


Figure 3 Composite anomaly maps of winter 500 mb heights and winds during NAO positive and NAO negative events. Negative anomalies indicate increased cyclonic flow, and positive anomalies indicate increased anticyclonic flow. Data are from the NCEP-Reanalysis Project (Kalnay et al., 1996).

tmp2m DJF – NAO positive (1950–90)



tmp2m DJF – NAO negative (1950–90)

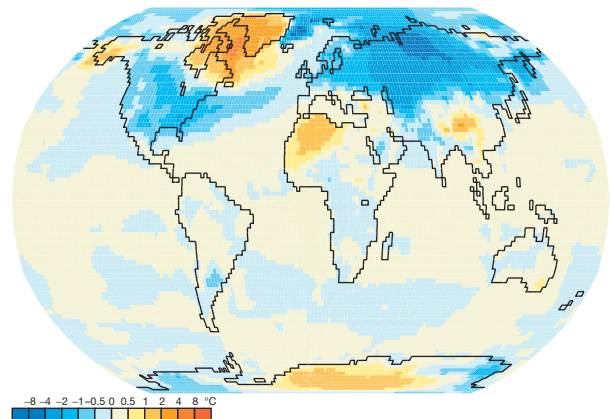
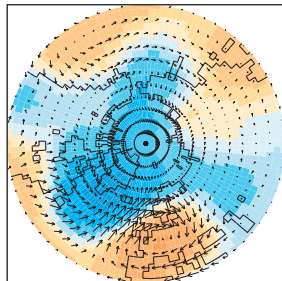


Figure 4 Composite anomaly maps of winter temperatures during NAO positive and NAO negative events. Negative anomalies indicate decreased temperatures, and positive anomalies indicate increased temperatures. Data are from the NCEP-Reanalysis Project (Kalnay et al., 1996).

500 mb heights and winds
AO positive


AO negative

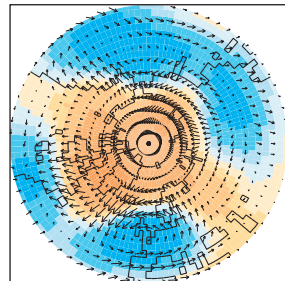
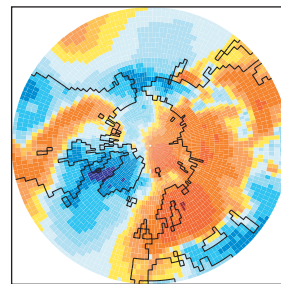


Figure 5 Composite anomaly maps of winter 500 mb heights and winds during AO positive and AO negative events. Negative anomalies indicate increased cyclonic flow, and positive anomalies indicate increased anticyclonic flow. Data are from the NCEP-Reanalysis Project (Kalnay et al., 1996).

Temperature at 2 m
AO positive


AO negative

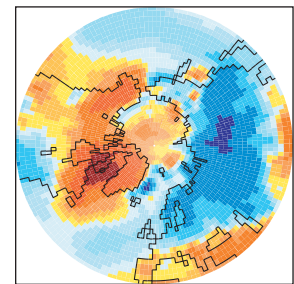


Figure 6 Composite anomaly maps of winter temperatures during AO positive and AO negative events. Negative anomalies indicate decreased temperatures, and positive anomalies indicate increased temperatures. Data are from the NCEP-Reanalysis Project (Kalnay et al., 1996).

the positive mode of the AO indicate colder temperatures in Greenland, warmer temperatures over much of Canada, and colder temperatures in Alaska (Figure 6). Temperatures during the negative mode of the AO show warmer conditions over

Greenland and Alaska, but the colder conditions over Canada is not well pronounced. Although these temperature responses of the AO are strongest seasonally in winter, a weaker signal is also evident in summer, which may show up in some paleoclimatic reconstructions (D'Arrigo et al., 2003).

Reconstruction of ENSO

Evidence from historical documents, tree-rings, and corals provides most of the ENSO reconstructions with annual resolution. Reconstructions of premodern ENSO events based on historical evidence have mostly been done in western South America, through the reconstruction of warm events (Ortlieb, 2000). Also, Quinn (1992) constructed a large-scale El Niño chronology by incorporating Nile River flood height data as well as records of droughts and floods in the Middle East and India. These historical reconstructions were each also assigned a relative strength (weak to very strong) and a confidence rating (1–5 with 5 being highest) relating to the reliability of historical evidence. The Quinn reconstruction extends back to AD 622, but the record prior to 1525 is based mostly on the Nile flood data. A comparison between the historical and modern instrumental record since 1877 shows approximately 75% of the El Niño years as being identical. Some anomalies exist, such as the occurrence of heavy rainfall in Peru during a non El Niño year, and some reconstructions may be suspect due to the validity and quality of historical sources (Ortlieb, 2000). Yet, a time series of the Quinn record since AD 622 shows the prevalence of a 2–6-year El Niño cycle that is similar to today.

Some coral records lie within the core area of ENSO activity, with continuous annual records extending back several hundred years (Tudhope et al., 2001). Most coral studies to date primarily analyze $\delta^{18}\text{O}$ isotopes in the skeletal carbonate as sea-surface temperature indicators (Cole, 2003), as $\delta^{18}\text{O}$ can represent seasonal curves within the annual cycle. However, $\delta^{18}\text{O}$ /climate relationships from corals are not simple, as for some areas, increased evaporation can enrich $\delta^{18}\text{O}$; conversely, increased precipitation can decrease $\delta^{18}\text{O}$. In addition, the $\Delta^{14}\text{C}$ in corals has been examined as an indicator of oceanic and atmospheric circulation changes, representative of ENSO events at interannual time scales. The trace elements cadmium and barium are also valuable in coral records, as they represent deep water upwelling conditions that are prevalent, for example, during cool periods of ENSO.

Several reconstructions of the winter SOI have been conducted from tree-rings in western North America and the Southern Hemisphere, by selecting tree-ring sites that are in ENSO-sensitive areas today. Perhaps the strongest and most coherent winter SOI reconstruction from tree rings comes from ENSO-sensitive trees in southwestern North America (Cook, 2000; Stahle et al., 1998). Some of the warm events from the reconstruction closely match the El Niño years from the historical record, but there are numerous disagreements due to non-ENSO climatic responses and problems with some of the historical records. A winter reconstruction of the SOI (Niño 3 region) from a network of tree rings in southwestern North America indicates that interannual variability is still prominent over the last four centuries, but less variability is evident for most of the time before the twentieth century (Figure 7).

The time period between the mid-1700s and early 1800s also appears to have somewhat longer periodicities between ENSO events. Luterbacher et al. (2002) provide a winter reconstruction of the NAO from old instrumental and historical records (Figure 7). The reconstruction reveals that much of the seventeenth to nineteenth centuries were characterized by somewhat less variability than that of the twentieth century, and the recent rise of the last 30–40 years is not evident in magnitude from the earlier record. Extreme values, both positive and negative, also exhibited times of increased frequency and absence, such as the active period of positive NAOs in the mid-nineteenth century. D'Arrigo et al. (2001) provide a reconstruction of the annual PDO (Figure 7). Results indicate somewhat increased decadal variability and generally lower indices prior to 1860. A reconstruction of the summer AO by D'Arrigo et al. (2003) from tree-ring sites in the Arctic Basin reveal mostly lower values prior to 1920, particularly in the late nineteenth century and late sixteenth century (Figure 7). Overall, paleoclimate reconstructions of the NAO, PDO, and AO seem to indicate that the twentieth century is rather anomalous when compared to most earlier records. This is an interesting result, and one that requires further investigation.

A long and well-dated ice-core record with annual resolution from Quelccaya, Peru, documents some past ENSO events (Thompson, 2000). However, the site lies a short distance south of the core region of ENSO activity, and it also may be influenced by Atlantic moisture sources; therefore it may not detect all ENSO events. ENSO reconstructions can also benefit from a multiproxy approach. For example, statistical analyses on several ENSO-sensitive historical, ice-core, and tree-ring records reveals 3–8 year periodicities from all of the records for the last 1000 years, though some longer periodicities are evident. Overall, the ENSO climatic expression did not differ dramatically, compared with variations in the modern record (Diaz and Pulwarty, 1994). Another approach to studying ENSO variability involves assessing the prominent patterns as reflected by empirical orthogonal functions (EOFs) of a multiproxy network (Mann et al., 2000). In their methods, Mann et al. demonstrate close calibration of the EOF that resembles ENSO activity with instrumental data from the modern record. Other evidence, such as from faunas (Sandweiss et al., 2001), flood deposits (Magilligan and Goldstein, 2001), paleoecological evidence from fossil pollen (McGlone et al., 1993), Foraminifera records in the tropical Pacific (Stott et al., 2002), and marine varves in the northeastern Pacific (Anderson et al., 1993), can reconstruct long decadal–century time scale variations of ENSO back through the Holocene and at times even into the late Pleistocene.

Regional Paleoclimatic Responses of Past El Niños and Climatic Forcing Mechanisms

Little work has been done on analyzing the consistency of regional and spatial variations of high resolution paleoclimatic responses to ENSO events prior to the modern record, especially for cold events. Yet, such opportunities exist, as demonstrated by Cole and Cook (1998) that ENSO/drought relationships over the United States have not always behaved consistently over the past 130 years. An interesting approach in

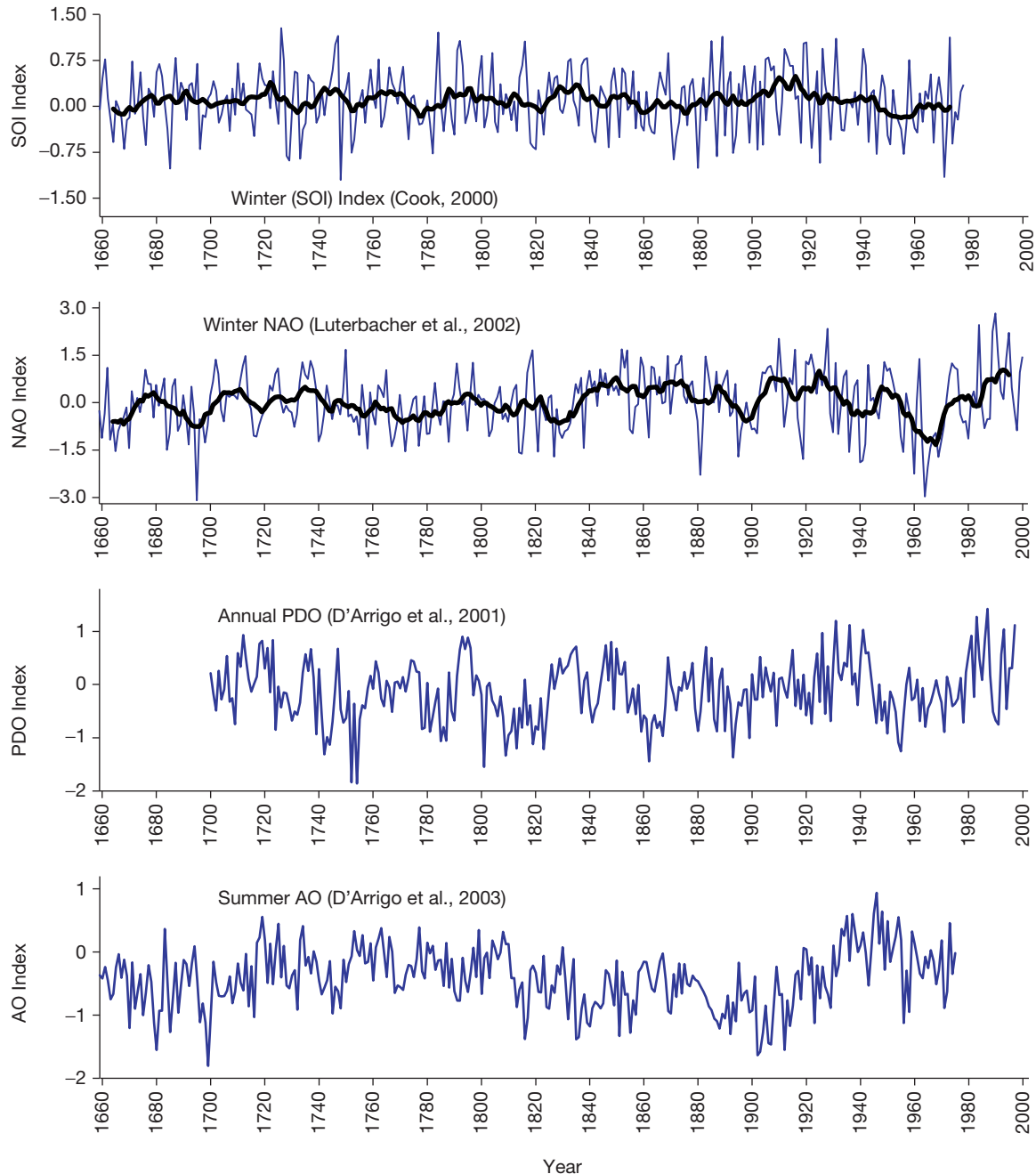


Figure 7 Annual time series of selected teleconnection indices as reconstructed by paleoclimatic data. Also shown are 10-year moving averages (bold line) for the SOI and NAO reconstructions. Data of the paleoclimatic reconstructions used in this paper were taken from the NOAA Paleoclimatology website at <http://www.ncdc.noaa.gov/paleo/paleo.html>.

assessing regional responses to paleo-ENSO events was conducted by Whetton and Rutherford (1994), later updated by Whetton et al. (1996). These studies analyzed statistical relationships between El Niño years from the Quinn record and proxy precipitation records in the Eastern Hemisphere. Statistically significant relationships between low precipitation and El Niño are evident for the Nile River and India since 1770, and it weakens for the Nile River prior to 1770, perhaps due to variations in the quality of historical data. Interestingly, correlations between El Niño and Northern China precipitation

records are not significant during the last century, but are statistically significant from 1525 to 1879. A record from Java suggests wetter conditions during El Niño years, but a statistically significant reversal in this relationships occurred during the 1700s. Such reversals in relationships that involve ENSO may be evident elsewhere around the world.

Numerous studies have analyzed the climatic forcing mechanisms that may drive varying periodicities of ENSO in the past at different time scales. This includes the use of general circulation models to simulate the variability of ENSO in model

years (e.g., for a 1000-year run) to further understand the mechanisms that drive the temporal variations in their phases, magnitude, and frequency. Mann et al. (2005) discussed how modeling efforts reveal the tendency for El Niño conditions to occur in response to volcanic forcing, with combined orbital and volcanic forcing explaining natural interannual variability of ENSO over the last 1000 years. They also describe some aspects of more La Niña type conditions that may have occurred during the Medieval Warm Period. The mid-Holocene has been of interest in determining the controls of ENSO activity, as numerous proxies suggest the likelihood of a prolonged weak ENSO period (Cole, 2001). Modeling efforts have recently been successful in simulating the natural variability of ENSO, and efforts in Holocene simulations that incorporate orbital forcing suggest the potential importance of cooler SSTs (Otto-Bliesner, 1999). At longer millennial time scales up to about 70 000 years ago, variations in greenhouse gases may be important controls of ENSO variability (Stott et al., 2002). The last interglacial, around 125 000 years ago, is also an interesting timeframe for examining paleo-ENSO, as the number of warm events may have been about twice as active as is normal today (Kukla et al., 2002).

Conclusions

Research on the paleoclimate and reconstruction of teleconnections, such as ENSO, the PNA, and NAO, is still in its early stages. Available data for calibrating annual proxies with the instrumental record are mostly restricted to the last 40 years. An understanding of teleconnections at time scales extending back into the Holocene and late Pleistocene will require more proxy data sites and improvements in modeling. However, the available reconstructions serve to illustrate the importance of a longer perspective of understanding paleo-ENSO events and other teleconnections that exhibit different aspects of variability and magnitude not apparent in the modern instrumental record.

See also: Dendroclimatology; Paleoclimate Relevance to Global Warming. **Paleoclimate:** Timescales of Climate Change. **Paleoclimate Modeling:** The Last Interglacial. **Paleoclimate Reconstruction:** Historical Climatology; Approaches; The Last Millennium; Paleodroughts and Society.

References

- Anderson RY, Soutar A, and Johnson TC (1993) Long-term changes in El Niño/Southern Oscillation: Evidence from marine and lacustrine sediments. In: Diaz HF and Markgraf V (eds.) *El Niño: Historical and Paleoclimatic Aspects of the Southern Oscillation*, pp. 419–433. Cambridge: Cambridge University Press.
- Cole JE (2001) A slow dance for El Niño. *Science* 291: 1496–1497.
- Cole JE (2003) Holocene coral records: Windows on tropical climate variability. In: McKay A, Battarbee R, Birks J, and Oldfield F (eds.) *Global Change in the Holocene*, pp. 168–184. London: Arnold.
- Cole JE and Cook ER (1998) The changing relationship between ENSO variability and moisture balance in the continental United States. *Geophysical Research Letters* 25: 4529–4532.
- Cook ER (2000) Niño 3 Index Reconstruction. International Tree-Ring Data Bank. IGBP PAGES/World Data Center-A for Paleoclimatology Data Contribution Series #2000-052. Boulder, CO, USA: NOAA/NGDC Paleoclimatology Program.
- D'Arrigo RD, Cook ER, Mann ME, and Jacoby GC (2003) Tree-ring reconstructions of temperature and sea-level pressure variability associated with the warm-season Arctic Oscillation since AD 1650. *Geophysical Research Letters* 30(11): 1549.
- D'Arrigo R, Villalba R, and Wiles G (2001) Tree-ring estimates of Pacific decadal climate variability. *Climate Dynamics* 18: 219–224.
- Diaz HF and Pulwarty RS (1994) An analysis of the time scales of variability in centuries-long ENSO-sensitive records in the last 1000 years. *Climate Change* 26: 317–342.
- Glantz MH (1996) *Currents of Change: El Niño's Impact on Climate and Society*. Cambridge: Cambridge University Press.
- Kalnay E, Kanamitsu M, Kistler R, et al. (1996) The NCEP/NCAR 40-year reanalysis project. *Bulletin of the American Meteorological Society* 77: 437–471.
- Kiladis GN and Diaz HF (1989) Global climate anomalies associated with extremes in the Southern Oscillation. *Journal of Climate* 2: 1069–1090.
- Kukla G, Clement AC, Cane MA, Gavin JE, and Zebiak SE (2002) Last interglacial and early glacial ENSO. *Quaternary Research* 58: 27–31.
- Latif M and Barnett TP (1996) Decadal climate variability over the North Pacific and North America: Dynamics and predictability. *Journal of Climate* 9: 2407–2432.
- Leathers DJ, Yarnal BM, and Palecki MA (1991) The Pacific/North American teleconnection pattern and United States climate. Part I: Regional temperature and precipitation associations. *Journal of Climate* 4: 517–528.
- Luterbacher J, Xoplaki E, Dietrich D, et al. (2002) Extending North Atlantic Oscillation reconstructions back to 1500. *Atmospheric Science Letters* 2: 114–124.
- Magilligan FJ and Goldstein PS (2001) El Niño floods and culture change: A late Holocene flood history for the Rio Moquegua, southern Peru. *Geology* 29: 431–434.
- Mann ME, Cane MA, Zebiak SE, and Clement A (2005) Volcanic and solar forcing of the tropical Pacific over the past 1000 years. *Journal of Climate* 18: 447–456.
- Mann ME, Gille E, Bradley RS, et al. (2000) Global temperature patterns in past centuries: An interactive presentation. *Earth Interactions* 4(4): 1–29.
- Mantua NJ, Hare SR, Zhang Y, Wallace JM, and Francis RC (1997) A Pacific interdecadal climate oscillation with impacts on salmon production. *Bulletin of the American Meteorological Society* 78: 1069–1079.
- Martinson DG, Battisti DS, Bradley RS, et al. (1998) *Decade-to-Century Scale Climate Variability and Change: A Science Strategy*. Washington, DC: National Academy Press.
- McGlone MS, Kershaw AP, and Markgraf V (1993) El Niño/Southern Oscillation climatic variability in Australasian and South American paleoenvironmental records. In: Diaz HF and Markgraf V (eds.) *El Niño: Historical and Paleoclimatic Aspects of the Southern Oscillation*, pp. 435–462. Cambridge: Cambridge University Press.
- Ortlieb L (2000) The documentary historical record of El Niño events in Peru: An update of the Quinn record sixteenth through nineteenth centuries. In: Diaz HF and Markgraf V (eds.) *El Niño and the Southern Oscillation: Multiscale Variability and its Impacts on Natural Ecosystems and Society*, pp. 207–295. Cambridge: Cambridge University Press.
- Otto-Bliesner BL (1999) El Niño/La Niña and Sahel precipitation during the middle Holocene. *Geophysical Research Letters* 26: 87–90.
- Overpeck JT and Webb RS (2000) Nonglacial rapid climate events: Past and future. *Proceedings of the National Academy of Sciences of the United States of America* 97: 1335–1338.
- Quinn WF (1992) A study of southern oscillation-related climatic activity for AD 622–1900 incorporating Nile River flood data. In: Diaz HF and Markgraf V (eds.) *El Niño: Historical and Paleoclimatic Aspects of the Southern Oscillation*, pp. 119–150. Cambridge: Cambridge University Press.
- Rogers JC (1984) The association between the North Atlantic Oscillation and the Southern Oscillation in the Northern Hemisphere. *Monthly Weather Review* 112: 1999–2015.
- Sandweiss DH, Maasch KA, Burger RL, Richardson JB III, Rollins HB, and Clement A (2001) Variation in Holocene El Niño frequencies: Climate records and cultural consequences in ancient Peru. *Geology* 29: 603–606.
- Stahle DW, Cleaveland M, Therrell MD, et al. (1998) Experimental dendroclimatic reconstruction of the Southern Oscillation. *Bulletin of the American Meteorological Society* 79: 2137–2152.
- Stott L, Poulson C, Lund S, and Thunell R (2002) Super ENSO and global climate oscillations at millennial time scales. *Science* 297: 222–226.

- Thompson LG (2000) Ice core evidence for climate change in the tropics: Implications for our future. *Quaternary Science Reviews* 19: 19–34.
- Thompson DWJ and Wallace JM (1998) The Arctic Oscillation signature in the wintertime geopotential height and temperature fields. *Geophysical Research Letters* 25: 1297–1300.
- Tudhope AW, Chilcott CP, McCulloch MT, et al. (2001) Variability in the El Niño–Southern Oscillation through a glacial–interglacial cycle. *Science* 291: 1511–1517.
- Whetton P, Allan R, and Rutherford I (1996) Historical ENSO teleconnections in the Eastern Hemisphere: Comparisons with latest El Niño series of Quinn. *Climatic Change* 32: 103–109.
- Whetton P and Rutherford I (1994) Historical ENSO teleconnections in the Eastern Hemisphere. *Climatic Change* 28: 221–253.
- Zhang Y, Wallace JM, and Battisti DS (1997) ENSO-like interdecadal variability: 1900–93. *Journal of Climate* 10: 1004–1020.

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PALEOCLIMATE RECONSTRUCTION

Contents

Approaches

Pliocene Environments

Paleodroughts and Society

Sub-Milankovitch (DO/Heinrich) Events

Paleotempestology

Younger Dryas Oscillation, Global Evidence

The Last Millennium

Historical Climatology

Approaches

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Introduction

A global network of meteorological stations and satellite observations documents the world's modern climates. There are far fewer observations of past conditions, and the need to search out – and to understand – archives of past climates means that despite more than a century of work, the knowledge of the history of climate remains sparse. The numbers of paleoclimate observations, however, are growing and filling in the maps needed to document the climate patterns of the past (e.g., [CLIMAP Project Members, 1981](#); [COHMAP Members, 1988](#); [Cook et al., 1999](#); [Mann et al., 1998](#)). In parallel, paleoclimate modeling of ever-increasing sophistication has provided key conceptual advances toward understanding the climate patterns of the past (e.g., [COHMAP Members, 1988](#); [Joussaume et al., 1999](#); [Rind et al., 1986](#)). Through such approaches, the field of paleoclimatology aims to understand the workings of the climate system.

Understanding past climate dynamics first requires reconstructing climate history. Reconstruction is, therefore, the primary work of paleoclimatologists and a two-part exercise. First, evidence for past conditions is accumulated, and second, the evidence is interpreted. The evidence includes biological, chemical, and geological traces of past events that are influenced by climate, such as tree growth and regional vegetation changes; the accumulation of certain elements and isotopes within glaciers, sediments, and the fossil remains of organisms; or the deposition of sediment in glacial moraines, lakes, and oceans. The interpretations range from simple qualitative inference to elegant quantitative estimates including computer-based process simulations. All approaches to paleoclimatic reconstruction rely on understanding the relationship between climate and the source of evidence; otherwise, the evidence

that exists equals only an unknown weather gauge: does it measure rainfall, temperature, or wind?

Paleoclimate reconstructions have been used as inputs to run computer climate model simulations of past conditions (e.g., sea-surface temperatures reconstructed by [CLIMAP Project Members, 1981](#), have been used by many studies) and have been used to test the accuracy of model output. Comparisons of empirically based reconstructions with model output also reveal the ways that atmospheric and ocean circulation, surface energy and moisture conditions, and other aspects of the climate system shaped past climates (e.g., [COHMAP Members, 1988](#)).

This article presents an overview of the approaches used to reconstruct paleoclimates. The first section introduces the wide range of paleoclimate data sources available. The second and third sections address qualitative and quantitative paleoclimate reconstructions, respectively, along with their utility and problems. The fourth section covers the important step of synthesizing and mapping paleoclimate data, as well as comparisons with climate model output, in order to better understand the nature of changing climatic gradients. Reconstructions alone cannot explain the observed changes. Models are required to understand the processes involved. Only paleoclimate reconstruction, however, can provide a long-term baseline for understanding modern and future conditions and can validate the ability of climate models to simulate conditions unlike those that are experienced today.

Proxies: Where Does Our Data Come From?

Evidence of past climate conditions come from many different sources. These sources can generally be divided into classes

based on the types of phenomena involved (i.e., biotic, geomorphic, geochemical, or geophysical) and according to the temporal resolution of the data (i.e., annual, decadal, centennial, millennial, or multimillennial) (for more details and examples, see [Bradley, 1999](#)). The sources derive from Earth processes that are controlled by climate and the evidence left by these processes acts as a stand-in, or proxy, for actual measurements of past conditions. In paleoclimatology, a proxy record tracks a phenomenon, such as changes in terrestrial vegetation composition, thought to have been closely shaped by climate and, therefore, providing a substitute for a direct measurement of climatic conditions. Proxies range from landscape features (e.g., glacial moraines, dry lake beds and abandoned shorelines, and inactive and vegetated dunes) to carefully calibrated measurements of biotic, geochemical, and geophysical phenomena (e.g., tree-ring widths, glacial ice isotopic values, and borehole temperatures).

Proxy data represent the world in different ways depending on the climatic factors that control the formation of each individual proxy. Multiple lines of evidence, therefore, produce the most complete paleoclimatic reconstructions including both temperature and moisture levels. The sections below briefly address the history of different paleoclimate studies, and consider some of specific proxies that are widely used today.

Qualitative Approaches

Beginnings

Paleoclimate reconstruction began, like many other fields, in a qualitative mode. Louis [Agassiz \(1840\)](#) recognized that hills and ridges of unconsolidated boulders, cobbles, and fine sediment, striations on smoothly worn bedrock and erratic boulders, among other features, represented the deposits left by once-expansive glaciers. Agassiz interpreted these features as glacial moraines left at the margins of flowing glacial ice, the patterns of glacial erosion of bedrock, and boulders transported by ice (rather than Biblical flood waters as others had contended) because he found strong similarities between these landscape features and the features of places that were currently glaciated in the Alps. Thus, Agassiz pioneered an important aspect of paleoclimate reconstruction using modern conditions or places as analogs for past conditions or the character of ancient landscapes. He proposed past 'Ice Ages' with large continental ice sheets based on extensive deposits in Europe and North America that he interpreted through analogy to be glacially derived.

Others followed Agassiz's approach. Another important example of early qualitative inferences about past conditions was the discovery of periods of wetter-than-modern climates in desert regions. [Gilbert \(1890\)](#) described extensive shorelines and river deltas that stood dry at high elevations on mountain sides above the Great Salt Lake in Utah (USA). The best explanation was that the lake had once been much higher than it is today and that the climate of Utah once had been wet. Similar studies recognized large ancient (now dry) lakes throughout the western United States, Australia, and Africa ([Street and Grove, 1979](#)). Conversely, vegetated and inactive dunes, as well as soils and paleosols, in the Great Plains and other regions (e.g., [Forman et al., 2001](#)) have long been interpreted as

signs of different-than-modern conditions in the past. Studies of loess sequences and river terraces have also long been studied as important qualitative signs of past climates that were different than today (e.g., [An et al., 1990](#)).

Continued Utility of Qualitative Approaches

Qualitative approaches to paleoclimate reconstruction remain important today, despite the advances of more quantitative approaches, and have been improved greatly by the development of radiocarbon and other dating methods. For example, the ages of many moraines are now known from the last glaciation because associated organic matter (e.g., logs and peat) have been radiocarbon dated. Maps of glacial extent over time provide vital data to the understanding of how conditions have changed (e.g., [Dyke and Prest, 1987](#)). For example, detailed maps of glacial extent help to inform studies which seek to explain the causes of abrupt climate change, including numerous century- and millennial-scale changes in temperature during the Pleistocene (e.g., [Rind et al., 1986](#)). Thus, long-used qualitative approaches to reconstructing past conditions merge well with late-twentieth and early-twenty-first century advances in paleoclimatology.

Qualitative data have also been important for appropriately parameterizing and testing computer model simulations of climate. Glacial coverage strongly influences the amount of energy coming into the Earth system from the sun because bright white snow has a high albedo, and thus, maps of glacial extent have become important inputs, or 'boundary conditions,' for computer simulations of 'Ice Age' conditions (e.g., [COHMAP Members, 1988](#)). Likewise, maps of lake status based on qualitative reconstructions of past lake shoreline levels, not unlike those initially conducted by [Gilbert \(1890\)](#), provide an important check on whether or not climate models accurately simulate past patterns of moisture balance (e.g., [COHMAP Members, 1988](#)). Similarly, qualitative reconstructions of temperature have been made from the past elevations and latitudes of treeline in mountains and polar regions (e.g., [MacDonald et al., 2000](#)) and have provided important benchmarks for evaluating climate-model simulations representing conditions as recent as the mid-Holocene (e.g., [Joussaume et al., 1999](#)) and as ancient as the Eocene when much of the Arctic was forested (e.g., [Sloan and Barron, 1990](#)).

Current Qualitative Techniques

As noted earlier, many qualitative approaches remain widely used and often focus on the recognition and dating of important landscape features. However, qualitative paleoclimate reconstruction has expanded to include the interpretation of many quantitative paleoenvironmental measurements.

Many climate proxies are precisely measured values of aspects of lake sediment, ocean sediment, or ice cores collected from key locations around the world. The interpretation of these proxies, however, is often qualitative. Examples include inferences about:

- past temperatures from lake sediment core characteristics, such as the amount of organic matter or biogenic silica produced by diatoms, thought to represent changes in

aquatic productivity governed by temperature (e.g., [Levesque et al., 1993](#));

- past temperatures from fossil pollen records of vegetation change, based on assumptions about the controls on plant distributions (e.g., shifts in tree cover during the Younger Dryas Chronozone, ~12 900–11 600 calendar years before present (cal yr BP; [Rind et al., 1986](#));
- past temperature and moisture levels from isotopic values of ice, lake sediment, and marine fossils thought to be closely governed by temperature- or evaporation-dependent effects on isotopic fractionation (e.g., the oxygen isotopic records from the Greenland Summit, beginning with [Dansgaard et al. \(1971\)](#), have been widely used as an uncalibrated record of past temperature);
- past moisture levels from aquatic plant macrofossil assemblages, sediment grain size, and unconformities in lake sediment cores thought to represent changes in lake water level (e.g., [Digerfeldt, 1986](#));
- past moisture levels from oceanic sediment characteristics thought to represent changes in river discharge (e.g., [Peterson et al., 2000](#));
- past moisture levels from sequences of soil development in loess deposits, such as those brought on by periods of intense monsoon rainfall ([An et al., 1990](#));
- past wind speeds and directions, as well as atmospheric pressure gradients, based on changes in lake and ocean sediment composition thought to have been influenced by terrestrial dust inputs, oceanic upwelling, and other wind-driven phenomena (e.g., [Clemens et al., 1991](#));
- past wind speeds and directions, as well as atmospheric pressure gradients, based on changes in the elemental and isotopic chemistry of ice cores thought to have been controlled by the deposition of wind-born dust and sea salt (e.g., [O'Brien et al., 1995](#));
- past frequencies of storm events based on coarse sediment layers deposited in lakes, coastal lagoons, salt marshes, and mangrove swamps thought to represent sediment deposited during floods (e.g., [Donnelly et al., 2001](#)).

Limitations

Qualitative paleoclimate reconstructions can be limited in their utility simply because they do not record the absolute magnitude of past temperature or precipitation changes. This caveat is more complex, however, than the simple lack of quantitative values. Misinterpretations can emerge from studies that do not explicitly calibrate the response of proxies to their governing factors. Many proxies behave in complex ways and can have multiple controls. Attempting to quantify an interpretation has the benefit of establishing the relationship between the proxies and the factors that control it.

Quantitative Approaches

General Concepts

Four types of quantitative paleoclimate reconstructions have received a large amount of attention. First, tree-ring measurements have offered a widely useful tool for quantifying past conditions because tree-ring growth is generally governed

by climate. Local relationships between growth and climate conditions can be calibrated and verified using the overlap between recent growth patterns and actual climate measurements for the same region ([Fritts, 1976](#); [Hughes et al., 1982](#)). Second, paleoclimate conditions have been inferred from assemblages of micro- and macrofossil remains of organisms preserved in ocean, lake, mire, and soil sediments by quantifying modern spatial relationships between the biota and climate and then assuming the same relationships governed past biotic changes (e.g., [Imbrie and Kipp, 1971](#); [Webb and Bryson, 1972](#)). Third, geochemical records have been used to provide quantitative records of climate by also applying calibration functions based on modern conditions, and/or the current understanding of the chemistry, to the proxy records (e.g., [Brassell et al., 1986](#); [Emiliani, 1966](#)). Fourth, physical measurements of temperature from boreholes in rock and ice have yielded quantitative estimates of surface temperatures based on the physics of heat flow (e.g., [Dahl-Jensen et al., 1998](#)).

Quantitative methods initially developed with a focus on fossil and subfossil remains of organisms and were explored in parallel for multiple settings and proxies (e.g., tree-rings foraminifera in ocean sediment cores and pollen in lake sediment cores). Tree-ring records extend back through the entire Holocene, but their application has been especially useful for studies of the past few centuries (e.g., [Mann et al., 1998](#)). Regressions between tree growth and historic climate values stand out as a relatively intuitive approach to tracking past conditions, but complications arise from the multiple local (soil, aspect) to regional (macroclimate) factors involved and from standardization of tree-ring widths to account for changes in tree volume over time ([Fritts, 1976](#)). Tree-ring-based methods incorporate a wide range of statistical methods to deconvolve the climatic signal in ensembles of tree-ring records (e.g., [Hughes et al., 1982](#)) and to map spatial patterns of climate using limited tree-ring data (e.g., [Cook et al., 1999](#)).

An important advance for many other types of biotic remains was the multivariate approach presented by [Imbrie and Kipp \(1971\)](#). The Imbrie and Kipp technique enabled investigators to reconstruct sea-surface temperatures from compositional analyses of foraminifera assemblages in ocean sediment cores based on the development of 'ecological' equations that relate assemblages to their living conditions today and then apply the equations to fossil assemblages. The method was then adapted for use with fossil pollen assemblages (e.g., [Webb and Bryson, 1972](#)) and many advances and additional applications to other taxa followed. Today, multivariate statistical methods are used to infer paleoclimate conditions from the assemblages of (1) many different types of marine, lacustrine, and mire organisms (e.g., [Charman, 2001](#); [Koç Karpuz and Schrader, 1990](#); [Smith, 1993](#)), (2) pollen from terrestrial vegetation (e.g., [Bartlién and Whitlock, 1993](#); [Guiot, 1990](#)), and (3) beetles and other terrestrial organisms (e.g., [Atkinson et al., 1986](#)).

Geochemical techniques also arose from a long history of investigation into the relationships between climate and geochemical systems. Early advances showed the influence of temperature on the relative abundance of light versus heavy isotopes of oxygen and hydrogen in many settings, including minerals and precipitation ([Dansgaard, 1964](#); [Urey, 1947](#)), and measurements of oxygen isotope ratios for paleoclimate

reconstruction followed (Dansgaard et al., 1971; Emiliani, 1966; Stuiver, 1968). The list of modern geochemical techniques includes multiple methods for reconstructing past water temperatures or salinity levels using elemental ratios (e.g., magnesium/calcium (Mg/Ca)) in sedimentary minerals, and using organic geochemistry. An important example of an organic geochemical technique used to measure temperature relies on the ratio of two plankton-derived compounds (alkenones) in marine sediments (Brassell et al., 1986). Ocean temperatures influence alkenone biosynthesis in a systematic manner, and thus the ratios of these compounds in sediments preserve past temperature trends (Müller et al., 1998; Prah and Wakeham, 1987).

Like qualitative paleoclimate data, quantitative estimates of past temperature and moisture levels have been important for computer-based studies of climate. Ocean temperature patterns inferred from fossil foraminifera assemblages (CLIMAP Project Members, 1981) were long used to provide the baseline ocean temperature data for computer simulations of Pleistocene conditions (e.g., COHMAP Members, 1988; Rind et al., 1986). Pollen-based inferences of terrestrial temperature and precipitation patterns have been used to test the accuracy of climate model output (e.g., Webb et al., 1993); syntheses of multiple quantitative approaches have been used to reconstruct temperature gradients during key intervals such as the Last Glacial Maximum (21 000 cal yr BP) and have also been the target of model experiments (e.g., COHMAP Members, 1988).

Present is the Key to the Past: Calibrating Proxies

When Agassiz (1840) postulated that ice ages had been part of Earth's history, he based his interpretation on the similarity between many geomorphic features found throughout Europe and the features found today near glaciers in the Alps. Likewise, many paleoclimatologists have used modern 'analogs' to inform their interpretation of proxy data. Tree-ring data have been interpreted through the establishment of calibration functions (regressions between historic conditions and historic growth patterns; Fritts, 1976), and similar calibrations have been done for many biotic and geochemical proxies (e.g., Müller et al., 1998; Smith, 1993). However, unlike tree-ring data, many calibrations rely on relationships (regressions) that link spatial (not temporal) patterns of climate with spatial patterns in biotic or geochemical data. The relationships then are applied to proxy data spanning centuries, millennia, or even the whole of the Quaternary, and thus, use information in space to gain information in time (e.g., Guiot, 1990).

Many quantitative techniques use equations, derived using modern climate-biotic assemblage datasets, to 'transfer' climate information onto biotic information (e.g., Bartlien and Whitlock, 1993; Imbrie and Kipp, 1971). Other techniques, however, explicitly rely on such an approach by statistically matching the composition of an assemblage of fossils to the composition of a modern biotic assemblage, thus inferring that the fossil assemblage represents the same set of conditions as the modern assemblage (e.g., Guiot, 1990). This 'modern analog technique' has often been applied to fossil assemblages of foraminifera in deep-ocean sediments (e.g., CLIMAP Project Members, 1981) and pollen in lake sediment (e.g., Bartlien and Whitlock, 1993; Guiot, 1990). Some modifications of the

approach include the generation of smoothed grids of possible analogs that span a wide range of climatic conditions by interpolation among modern samples (e.g., 'response surfaces' in Bartlien and Whitlock, 1993; Webb et al., 1993) and the use of neural networks, rather than explicit measurements of similarity, to find the best analogs (e.g., Peyron et al., 1998). More simply, growing-season and winter temperature inferences from beetle remains use the overlapping modern temperature preferences of the taxa found in a given sample (Atkinson et al., 1986).

Widely used geochemical proxies, such as isotope and alkenone records, also use relationships based on modern analogs to guide their interpretation (e.g., Müller et al., 1998), but many of these techniques rely on more fundamental relationships – established either from first principles or empirically (e.g., Prah and Wakeham, 1987; Urey, 1947). Similarly, temperature measurements from boreholes are quantitative in nature, but an understanding of physical processes must be applied to either establish the age of the measured temperature or the magnitude of signal attenuation over time. In this way, these geophysical measurements do not rely on 'analogs' but process-based understanding of the physics involved (Dahl-Jensen et al., 1998). Borehole temperatures have also been used to calibrate more detailed and chronologically controlled geochemical measurements of glacial ice (Cuffey and Clow, 1997). Thus, modern 'analogs' and their associated drawbacks can sometimes be avoided.

Quantitative Problems

A major problem for quantitative paleoclimate reconstruction arises from periods of the past that have conditions with no equivalent anywhere on Earth today. Modern analog methods fail if no analog exists. For example, fossil pollen records from the mid-continent of North America contain numerous samples from the late Pleistocene and early Holocene, which are compositionally unlike any modern sample of pollen; the assemblages appear to be the result of climatic conditions that had a more extreme difference in seasonal conditions (colder winters and warmer summers) than found anywhere today (Williams et al., 2001). Similarly, many biotic proxies may not find adequate analogs for Pleistocene conditions because atmospheric carbon dioxide levels were lower than those present during the Last Glacial Maximum and much of the Pleistocene. Higher plants and algae depend upon carbon dioxide for photosynthesis, and thus, ecosystems that developed under low carbon dioxide levels in the Pleistocene likely behaved differently to ecosystems today and conditions inferred from matching fossil and modern biotic assemblages may not be accurate (Cowling and Sykes, 1999).

Equation-based reconstructions can overcome the lack of modern analogs by extrapolation. Such equations may produce values beyond the range of modern data; however, the functions are not constrained under such circumstances and may be erroneous. Another approach to solving the problem involves modeling the processes involved in generating a proxy record, and thus, creating a paleoclimate reconstruction. The advantages of such an approach stem from the ability to understand the proxy mechanistically and account for complications which may not be tractable using statistical

reconstruction methods. For example, pollen-climate transfer equations can provide values of temperature and precipitation from Pleistocene pollen assemblages that lack modern analogs (e.g., Bartlién and Whitlock, 1993), but such reconstructions may be erroneous due to the influence of atmospheric carbon dioxide levels on vegetation dynamics (Cowling and Sykes, 1999). However, process-modeling of vegetation dynamics, including the influences of both climate and carbon dioxide, can help to constrain and improve the reconstructions (Guiot et al., 2000). An important caveat with such modeling approaches is the limited nature of models: processes may not all be incorporated, understood, or appropriately parameterized.

Processes Are the Key to Understanding: Process-Modeling as a Way for Reconstruction

Model-based reconstructions have played an important role in paleoclimate reconstruction. In particular, forward, process-based models have been used to make quantitative climate inferences from data that lack a method for statistical climate-proxy calibrations. Most prominently, estimates of past lake levels have been modeled to infer past precipitation rates (e.g., Kutzbach, 1980). In these cases, the hydrologic model is run in an attempt to simulate the observed changes in lake levels, and the range of conditions used as inputs into the model that produce satisfactory simulations can be inferred to represent the past conditions. Similarly, borehole temperature records rely on models of conduction and diffusion of heat over time to accurately interpret the raw down-hole data (e.g., Dahl-Jensen et al., 1998).

Synthesis

Paleoclimate reconstructions are most often made based on data from a single location. Interpretation of the climatic changes at a given study location, however, depends on knowledge about other places – perhaps even the rest of the world. Maps of paleoclimate data therefore are the ultimate reconstructions. They show empirically based representations of past climate gradients, which can be used to infer circulation patterns and other important dynamics (e.g., CLIMAP Project Members, 1981; COHMAP Members, 1988; Mann et al., 1998). Such inferences are limited if made from single proxy records alone, and like all paleoclimate reconstruction, can be made using qualitative or quantitative approaches. Qualitative approaches include either (1) inferring synoptic dynamics from climatic patterns (e.g., Shuman et al., 2002) or (2) finding a modern analog amid historic climate data that shows a pattern similar to the inferred past climate pattern (e.g., Mock and Brunelle-Daines, 1999). Quantitative approaches involve either (1) using statistical linkages between climatic values and synoptic climate dynamics (e.g., Webb and Bryson, 1972) or (2) comparing the data with climate model output (e.g., COHMAP Members, 1988).

Like the real-climate system, climate models produce fields of data that represent the spatial gradients of temperature, precipitation, pressure, and wind, as well as other variables (Bartlién and Hostetler, 2004). Comparisons of model output with maps of paleoclimate data thus provide important

insights, which benefit both the quest to improve climate models and to better understand past-climate dynamics. The comparisons reveal the ability of the model to accurately simulate conditions unlike those found today and show the types of dynamics that may explain past climate patterns (COHMAP Members, 1988). Therefore, in combination with climate model output, the paleoclimate record offers a rich source of insight into conditions unlike those experienced today.

See also: Dendroclimatology. Carbonate Stable Isotopes: Nonmarine Biogenic Carbonates. Paleoceanography, Biological Proxies: Benthic Foraminifera; Planktic Foraminifera. Paleoclimate: Introduction.

References

- Agassiz L (1840) *Etudes sur les Glaciers*. Neuchâtel: Jent et Gassmann. Published privately.
- An Z, Liu T, Lu Y, et al. (1990) The long term paleomonsoon variation recorded by loess paleosol sequence in central China. *Quaternary International* 7(8): 91–95.
- Atkinson TC, Briffa KR, Coope GR, Joachim M, and Perry DW (1986) Climatic calibration of coleopteran data. In: Berglund B (ed.) *Handbook of Holocene Palaeoecology and Palaeohydrology*, pp. 851–858. New York: Wiley.
- Bartlién PJ and Hostetler SW (2004) Modeling paleoclimates. In: Gillespie AR, Porter SC, and Atwater BF (eds.) *The Quaternary Period in the United States*, pp. 565–584. Elsevier: Amsterdam.
- Bartlién PJ and Whitlock C (1993) Paleoclimatic interpretation of the Elk Lake Pollen Record. In: Bradbury JP and Dean WE (eds.) *Elk Lake, Minnesota: Evidence for Rapid Climate Change in the North-Central United States*, pp. 275–293. Boulder, CO: Geological Society of America.
- Bradley RS (1999) *Paleoclimatology: Reconstructing Climates of the Quaternary*, 2nd edn. New York: Harcourt Academic Press.
- Brassell SC, Eglinton G, Marlowe IT, Pflaumann U, and Sarinthein M (1986) Molecular stratigraphy: A new tool for climatic assessment. *Nature* 320: 129–133.
- Charman DJ (2001) Biostratigraphic and palaeoenvironmental applications of testate amoebae. *Quaternary Science Reviews* 20: 1753–1764.
- Clemens S, Prell W, Murray D, Shimmield G, and Weedon G (1991) Forcing mechanisms of the Indian Ocean monsoon. *Nature* 353: 720–725.
- CLIMAP Project Members (1981) *Seasonal Reconstructions of the Earth's Surface at the Last Glacial Maximum*. Boulder, CO: Geological Society of America. Map and Chart Series MC-36.
- COHMAP Members (1988) Climatic changes of the last 18,000 years: Observations and model simulations. *Science* 241: 1043–1052.
- Cook ER, Meko DM, Stahle DW, and Cleaveland MK (1999) Drought reconstructions for the continental United States. *Journal of Climate* 12: 1145–1162.
- Cowling SA and Sykes MT (1999) Physiological significance of low atmospheric CO₂ for plant-climate interactions. *Quaternary Research* 52: 237–242.
- Cuffey KM and Clow GD (1997) Temperature, accumulation, and ice sheet elevation in central Greenland through the last deglacial transition. *Journal of Geophysical Research* 102: 26,383–26,396.
- Dahl-Jensen D, Mosegaard K, Gundestrup N, et al. (1998) Past temperatures directly from the Greenland ice sheet. *Science* 282: 268–271.
- Dansgaard WS (1964) Stable isotopes in precipitation. *Tellus* 16: 436–468.
- Dansgaard WS, Johnsen SJ, Clausen HB, and Langway CC Jr. (1971) Climatic record revealed by the Camp Century ice core. In: Turekian K (ed.) *The Late Cenozoic Glacial Ages*, pp. 37–56. New Haven, CT: Yale University Press.
- Digerfeldt G (1986) Studies on past lake-level fluctuations. In: Berglund BE (ed.) *Handbook of Holocene Palaeoecology and Palaeohydrology*, pp. 127–142. Chichester, UK: Wiley.
- Donnelly JP, Bryant SS, Butler J, et al. (2001) A 700-year sedimentary record of intense hurricane landfalls in southern New England. *Geological Society of America Bulletin* 113: 714–727.
- Dyke AS and Prest VK (1987) Late-Wisconsinan and Holocene history of the Laurentide Ice Sheet. *Geographie Physique et Quaternaire* 41: 237–263.
- Emiliani C (1966) Isotopic paleotemperatures. *Science* 154: 851–857.

- Forman SL, Oglesby R, and Webb RS (2001) Temporal and spatial patterns of Holocene dune activity on the Great Plains of North America: Megadroughts and climate links. *Global and Planetary Change* 29: 1–29.
- Fritts HC (1976) *Tree Rings and Climate*. London: Academic Press.
- Gilbert GK (1890) Lake Bonneville. *U.S. Geological Survey Monograph* 1: 1–275.
- Guiot J (1990) Methodology of the last climatic cycle reconstruction in France from pollen data. *Palaeogeography, Palaeoclimatology, Palaeoecology* 80: 49–69.
- Guiot J, Torre F, Jolly D, Peyron O, Boreaux JJ, and Cheddadi R (2000) Inverse vegetation modeling by Monte Carlo sampling to reconstruct palaeoclimates under changed precipitation seasonality and CO₂ conditions: Application to glacial climate in Mediterranean region. *Ecological Modelling* 127: 119–140.
- Hughes MK, Kelley PM, Pilcher JR, and LaMarche VC (1982) *Climate from Tree Rings*. Cambridge: Cambridge University Press.
- Imbrie J and Kipp NG (1971) A new micropaleontological method for quantitative paleoclimatology: Application to a Late Pleistocene Caribbean core. In: Turekian K (ed.) *The Late Cenozoic Glacial Ages*, pp. 71–181. New Haven, CT: Yale University Press.
- Joussaume S, Taylor KE, Braconnot P, et al. (1999) Monsoon changes for 6000 years ago: Results of 18 simulations from the Paleoclimate Modeling Intercomparison Project (PMIP). *Geophysical Research Letters* 26: 859–862.
- Koç Karpuz N and Schrader H (1990) Surface sediment diatom distribution and Holocene paleotemperature variations in the Greenland, Iceland and Norwegian Seas. *Paleoceanography* 5: 557–580.
- Kutzbach JE (1980) Estimates of past climate at Paleolake Chad, North Africa, based on a hydrological and energy-balance model. *Quaternary Research* 14: 210.
- Levesque AJ, Mayle FE, Walker IR, and Cwynar LC (1993) A previously unrecognized late-glacial cold event in eastern North America. *Nature* 361: 623–626.
- MacDonald GM, Velichko AA, Kremenetski CV, et al. (2000) Holocene treeline history and climate change across northern Eurasia. *Quaternary Research* 53: 302–311.
- Mann ME, Bradley RS, and Hughes MK (1998) Global-scale temperature patterns and climate forcing over the past six centuries. *Nature* 392: 779–787.
- Mock CJ and Brunelle-Daines AR (1999) A modern analog of western United States summer palaeoclimate at 6000 years before present. *The Holocene* 9: 541–545.
- Müller PJ, Kirst G, Ruhland G, von Storch I, and Rosell-Mele A (1998) Calibration of the alkenone paleotemperature index UK'37 based on core tops from the eastern South Atlantic and the global ocean (60°N–60°S). *Geochimica et Cosmochimica Acta* 62: 1757–1772.
- O'Brien SR, Mayewski PA, Meeker LD, Meese DA, Twickler MS, and Whitlow SI (1995) Complexity of Holocene climate as reconstructed from a Greenland ice core. *Science* 270: 1962–1964.
- Peterson LC, Haug GH, Hughen KA, and Röhl U (2000) Rapid changes in the Hydrologic cycle of the tropical Atlantic during the last glacial. *Science* 290: 1947–1951.
- Peyron O, Guiot J, Cheddadi R, et al. (1998) Climatic reconstruction in Europe for 18,000 yr BP from pollen data. *Quaternary Research* 49: 183–196.
- Prahl FG and Wakeham SG (1987) Calibration of unsaturation patterns in long-chain ketone compositions for paleotemperature assessment. *Nature* 330: 367–369.
- Rind D, Peteet D, Broecker W, McIntyre A, and Ruddiman W (1986) The impact of cold North Atlantic sea surface temperatures on climate: Implications for the Younger Dryas cooling (11–10 k). *Climate Dynamics* 1: 3–33.
- Shuman B, Bartlien P, Logar N, Newby P, and Webb T III (2002) Parallel climate and vegetation responses to the early Holocene collapse of the Laurentide Ice Sheet. *Quaternary Science Reviews* 21: 1793–1805.
- Sloan LC and Barron EJ (1990) "Equable" climates during Earth history? *Geology* 18: 489–492.
- Smith AJ (1993) Lacustrine ostracodes as hydrochemical indicators in lakes of the north-central United States. *Journal of Paleolimnology* 8: 135–147.
- Street FA and Grove AT (1979) Global maps of lake-level fluctuations since 30,000 yr B.P. *Quaternary Research* 12: 83–118.
- Stuiver M (1968) Oxygen-18 content of atmospheric precipitation during the last 11,000 years in the Great Lakes region. *Science* 162: 994–997.
- Urey HC (1947) The thermodynamic properties of isotopic substances. *Journal of the Chemical Society* 562–581.
- Webb T III, Bartlien PJ, Harrison SP, and Anderson KH (1993) Vegetation, lake level, and climate change in eastern North America. In: Wright HE Jr., Kutzbach JE, Webb T III, Ruddiman WF, Street-Perrott FA, and Bartlien PJ (eds.) *Global Climates since the Last Glacial Maximum*, pp. 415–467. Minneapolis, MN: University of Minnesota Press.
- Webb T III and Bryson RA (1972) Late- and postglacial climatic change in the northern Midwest, USA: Quantitative estimates derived from fossil pollen spectra by multivariate statistical analysis. *Quaternary Research* 2: 70–115.
- Williams JW, Shuman BN, and Webb T III (2001) Dissimilarity analyses of late-Quaternary vegetation and climate in eastern North America. *Ecology* 82: 3346–3362.

Pliocene Environments

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Introduction

The Pliocene spans the interval of Earth history from ca. 5.3 to 1.8 Ma (including the Gelasian Stage of the Upper Pliocene, which is now considered the base of the Quaternary subera, beginning at 2.6 Ma). During the Pliocene, the Earth completed a transition from relatively warm climatic conditions, with little or no continental ice in the Northern Hemisphere, to more variable climatic conditions, with significant continental ice sheets in the Northern Hemisphere, and alternating glacial–interglacial cycles. The early Pliocene, from about 5.3 to 3.0 Ma, represents the last time in Earth's history when climates were generally warmer and more equable than modern climate. At the start of the late Pliocene (after 3.0 Ma), significant Northern Hemisphere ice sheets began to build-up, mid- and high-latitude areas cooled, and climate variability, as measured by $\delta^{18}\text{O}$ in deep-sea carbonate microfossils, increased. Distinct glacial–interglacial cycles developed between 2.9 and 2.7 Ma.

The high-latitude cooling and onset of Northern Hemisphere glaciation (NHG) that occurred in the late Pliocene represent the culmination of a trend that started 50 Ma in the early Eocene. [Figure 1](#) shows a composite 65 Ma $\delta^{18}\text{O}$ record measured in calcareous benthic foraminifers in deep-sea sediments. The $\delta^{18}\text{O}$ is a measure of the ratio of ^{16}O : ^{18}O in the shells of the Foraminifera, which is controlled primarily by the temperature of sea water and the $\delta^{18}\text{O}$ of sea water when the shell is formed. Lower temperatures and/or more global continental ice volume at the time of shell formation reduces the ^{16}O in the shell and results in a more positive $\delta^{18}\text{O}$. The global deep-sea $\delta^{18}\text{O}$ record for the last 50 Ma shows an overall increase continuing to the present day. From about 40 Ma to the present, much of the change in the $\delta^{18}\text{O}$ record reflects the buildup of polar ice volume and the related cooling of high latitudes ([Zachos et al., 2001](#)). Prior to 3 Ma, most of the ice buildup and variability was in Antarctica. The increase in $\delta^{18}\text{O}$ after 3 Ma is primarily related to the development and variability of Northern Hemisphere ice sheets. Most workers concur that the long-term trend seen in [Figure 1](#) is related to the overall decrease in atmosphere carbon dioxide (CO_2) concentrations ([Figure 2](#)) that led to an overall cooling of Earth.

Changes in the Earth's orbit alter the distribution of incoming solar radiation. The change in solar radiation causes long-term 10^4 – 10^6 oscillations in the Earth's climate. Orbital cycles include 19–23 ka cycles related to the precession of the Earth's rotational axis, a 41-ka cycle related to the tilt of the Earth's rotational axis (obliquity), and a 100-ka cycle related to the changes in the shape of the Earth's orbit around the Sun (eccentricity). The orbital forcing and the Earth's climate response to the forcing vary through time.

[Figure 3](#) shows a typical composite deep-sea bottom-water $\delta^{18}\text{O}$ record for the last 5 My. Most $\delta^{18}\text{O}$ values from 5 to 3 Ma

vary within a narrow range between 3.5 and 3.0‰. Isotopic values are generally similar or smaller than late Pleistocene interglacial values (shown by the dashed vertical line in [Figure 3](#)), which indicates that mid- and high-latitude conditions during the early Pliocene were usually as warm or warmer than late Pleistocene interglacial intervals. The amplitude of variability in the isotopic record is relatively small compared to the late Pliocene and Pleistocene, which indicates that variability in early Pliocene climate was subdued compared to later intervals. The isotope record contains 19–23-ka precession cycles. Early Pliocene climate was variable but in general warm intervals of the early Pliocene were as warm as or warmer than Pleistocene interglacial intervals whereas cool intervals were much warmer than Pleistocene glacial intervals.

After 3.0 Ma, excursions to increasingly more positive $\delta^{18}\text{O}$ are evident and distinct oscillations between smaller (interglacial) and larger (glacial) $\delta^{18}\text{O}$ become well defined. The increasingly positive $\delta^{18}\text{O}$ and the resulting higher amplitude cycles reflect the progressive intensification of NHG and development of glacial–interglacial oscillations that are typical of the late Pleistocene. Intervals when $\delta^{18}\text{O}$ were as small as late Pleistocene interglacial values became increasingly brief. Cooler and more glacial conditions became more prevalent toward the present. At about the same time that NHG intensified, 41 ka obliquity-related cycles became prominent in the isotope record. The change from precession-related variability to obliquity-related variability at about 3.0 Ma is seen in a variety of geologic records. The obliquity-related cycles continued to be prominent into the Pleistocene until about 1.0 Ma when 100-ka eccentricity cycles became prominent.

[Figure 3](#) also shows estimates of sea-surface temperature (SST) in the South Atlantic Ocean off the coast of South Africa as measured by the alkenone technique. From 4.8 to about 3.2 Ma, SST was warm and relatively stable. However, after 3.0 Ma the SST record shows a marked cooling trend and variability increased. The cooling trend in the SST record at this locality is a measure of the intensification of wind-driven upwelling associated with the Benguela Current ([deMenocal, 2004](#)). The SST record in [Figure 3](#) is typical of many climate records from the Pliocene in that most records from mid- and high latitudes indicate overall warmer conditions during the early Pliocene compared to modern and then show evidence of cooling and higher amplitude variability after about 3.0 Ma.

Warm Climates of the Early Pliocene

A survey of Pliocene climates, mostly covering the interval from about 4 to 2 Ma, has been done by the Pliocene Research, Interpretation, and Synoptic Mapping (PRISM) project ([Dowsett et al., 1999](#); [Poore and Sloan, 1996](#)). Evidence for warmer conditions during the early Pliocene was found in

many records from a variety of areas. For example, early Pliocene pollen records show that trees grew on the margin of the Arctic Ocean and on the continent of Antarctica (Thompson and Fleming, 1996). Warm water planktic microfossils were found in high latitudes of the North Atlantic and North Pacific

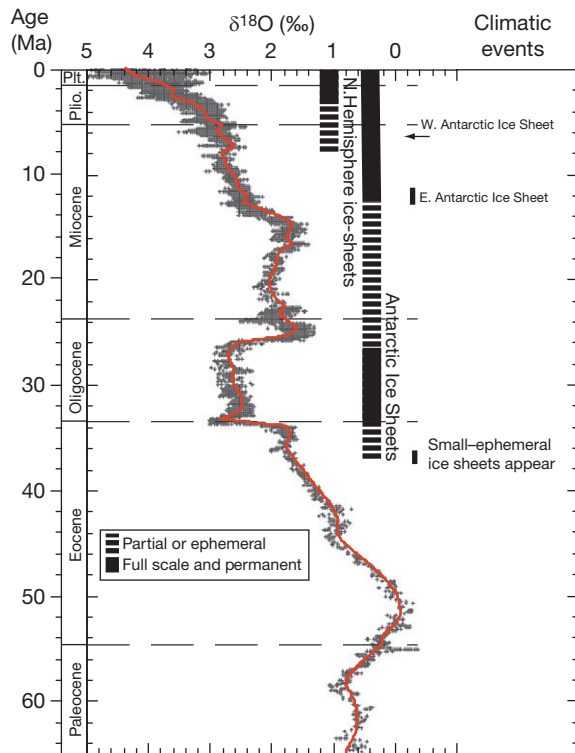


Figure 1 Composite global oxygen isotope record compiled from measurements of $\delta^{18}\text{O}$ in shells of deep-sea benthic foraminifers. The measurements were adjusted for vital effects. The red line represents an adjusted five-point running mean. Highly generalized occurrence of polar ice sheets is indicated by black bars. Black vertical bars show the occurrence of east and west Antarctic Ice Sheets and Northern Hemisphere ice sheets. Small and ephemeral ice sheets may have developed in the Northern Hemisphere as early as the late Miocene but permanent ice sheets did not develop until the middle of the Pliocene. Modified from Zachos J, Pagani M, Sloan L, Thomas E, and Billups K (2001) Trends, rhythms, and aberrations in global climate 65 Ma to present. *Science* 292: 686–693, reproduced with permission.

Oceans (Dowsett et al., 1996). The PRISM project integrated environmental records from 74 terrestrial localities and 77 marine localities into a synoptic compilation of representative conditions during a typical warm interval of a ‘time-slab’ from 3.15 to 2.85 Ma. Figure 4 shows a comparison of modern and the PRISM Pliocene global compilation of SST, sea ice, continental ice, and vegetation cover developed for climate modeling experiments. The compilation for the Pliocene is highly generalized and summarized on a 2° latitude by 2° longitude grid. The number of control points is especially limited in the Southern Hemisphere, and the details of the magnitude and timing of inferred Pliocene conditions can be questioned. However, the major differences indicated between the present day and early Pliocene are supported by multiple studies (Dowsett et al., 1999). SST in high and mid-latitudes were several degrees Celsius above modern. Evergreen vegetation was much more extensive in the Northern Hemisphere high latitudes during the early Pliocene, largely at the expense of more arid and cooler tundra and polar desert environments. Note the reduction in continental ice sheets (glaciers) compared to the present day and the absence of summer sea ice in the Arctic Ocean. The inferred reduction in continental ice sheets, as well as the elevation of geomorphic features such as Pliocene shorelines on continental margins, indicates that early Pliocene sea level was 20–30 m above modern sea level.

Mechanisms for Pliocene Warmth

Although climate records from a variety of areas indicate that climate during the early Pliocene was often warmer than modern climate, the explanation for the warmer climate is still a matter of debate. Mechanisms commonly mentioned as the forcing for warm early Pliocene climate include higher atmospheric concentrations of CO_2 , increased oceanic heat transport, and a permanent El Niño SST patterns in the equatorial Pacific Ocean.

Estimates based on carbon isotope records in marine deposits (Raymo et al., 1996) indicate that maximum global atmospheric CO_2 concentrations in the early Pliocene were around 425 ppm which is above the preindustrial (1850) atmospheric CO_2 level of 280 ppm but only slightly higher than the 2000 level of ~370 ppm. Estimates of Pliocene atmospheric CO_2 concentrations based on leaf stomatal frequency and size variation (Kurschner et al., 1996) suggest that

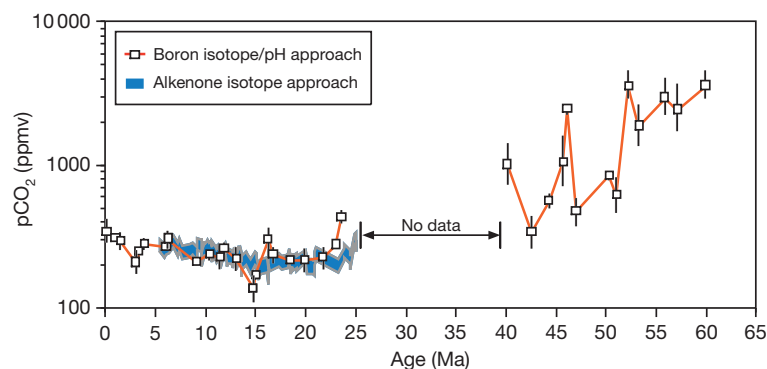


Figure 2 Estimates of atmospheric pCO_2 for the last 60 Ma. Reproduced from Zachos J, Pagani M, Sloan L, Thomas E, and Billups K (2001) Trends, rhythms, and aberrations in global climate 65 Ma to present. *Science* 292: 686–693, reproduced with permission. © 2001 AAAS.

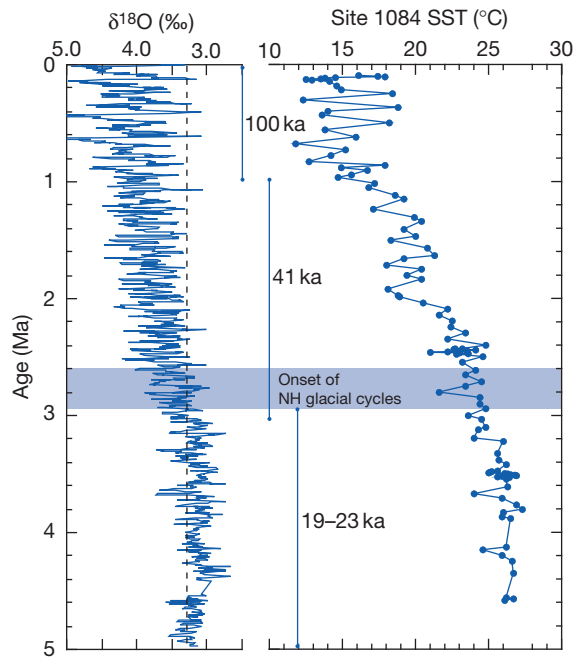


Figure 3 Composite benthic foraminifer oxygen isotope record and estimated sea-surface temperature (SST) record derived from alkenone measurements in ODP Site 1048 off the coast of Namibia in the South Atlantic Ocean. The SST record shows that warm and relatively stable temperatures in the early Pliocene became progressively cooler and more variable after about 3.2 Ma. The cooling and higher variability reflect intensification of the wind-driven Benguela Current. The composite benthic foraminifer isotope record shows progressive cooling and build-up of global ice volume and during the late Pliocene into the Pleistocene. The dashed vertical line with the composite isotope record represents the expected value for Pleistocene interglacial intervals. Note that conditions as warm and with ice volume similar to Pleistocene interglacial times become increasingly brief from the late Pliocene to the Pleistocene. See text for discussion. Adapted from deMenocal PB (2004) African climate change and faunal evolution during the Pliocene–Pleistocene. *Earth and Planetary Science Letters* 220: 3–24. © 2004, reproduced with permission from Elsevier.

values varied between 280 and 370 ppm. The difference in atmospheric CO_2 levels between the Pliocene and today seems too small to account for the climate differences inferred from the geologic record (Crowley, 1996). In addition, warming due solely to increased atmospheric CO_2 should result in warming at all locations. In contrast, the available Pliocene observations indicate temperatures in most tropical areas were very similar to modern.

Increased oceanic heat transport is another mechanism that has been proposed to explain Pliocene warmth. Mapping of Pliocene SST estimates in the North Atlantic Ocean (Dowsett et al., 1992) and estimates of deep-ocean circulation based on carbon isotope gradients in deep-ocean basins indicate that thermohaline circulation (THC) was intensified during much of the early Pliocene. Increased THC would result in increased transport of warm low-latitude waters via the Gulf Stream into the North Atlantic Ocean and explain the increased SST inferred from microfossil assemblages preserved in deep-sea sediments (Figure 5). Increased transport of heat into the North Atlantic should result in cooler equatorial SST. However,

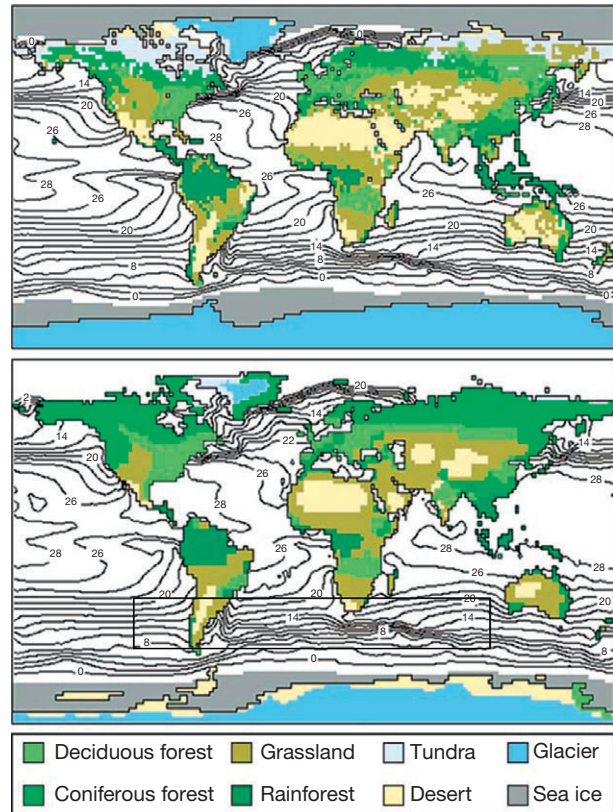


Figure 4 Comparison of generalized reconstruction of global environmental conditions for modern (top panel) and early Pliocene (bottom panel) warm intervals. Reconstruction is designed for use in modeling experiments and is done on a 2° latitude by 2° longitude grid. Top panel shows modern summer conditions. Bottom panel shows early Pliocene warm interval summer conditions. Land cover including glaciers and sea ice extent are color coded. SSTs are shown by isotherms in degrees centigrade. Open ocean areas are white. See text for discussion. Adapted from Dowsett HJ, Barron JA, Poore RZ, et al. (1999) Middle Pliocene paleoenvironmental reconstruction; PRISM2. *U.S. Geological Survey Open-File Report* 99-535, reproduced with permission.

available SST estimates from equatorial areas of the Atlantic indicate SST was similar to modern. It is possible that early Pliocene warming was due to some combination of slightly higher atmospheric CO_2 and increased THC.

Another proposed explanation for Pliocene warmth is that a permanent El Niño-like state existed in the early Pliocene. In the modern ocean, easterly winds in the equatorial Pacific essentially 'pile up' warm water in the western Pacific and cause upwelling of deeper cool waters to the surface in the eastern equatorial Pacific. The result is a west-to-east thermal gradient of about $3\text{--}7^\circ\text{C}$. During El Niño events, the easterly winds relax and the warm water in the western equatorial Pacific flows back to the east resulting in higher SST in the east and a reduction in the west-to-east thermal gradient. The change in equatorial surface water conditions alters atmospheric circulation patterns and affects climate conditions in a number of areas of the world through atmospheric teleconnections. The permanent El Niño theory for Pliocene warming argues that the modern temperature gradient across the equatorial Pacific Ocean was absent due to warmer SST in the

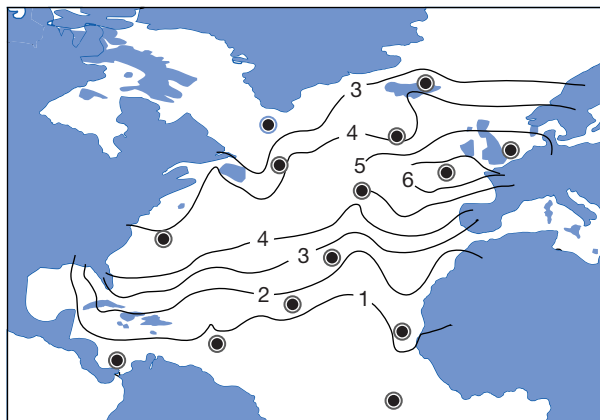


Figure 5 Map of North Atlantic Ocean showing difference in summer SST between early Pliocene warm interval and modern. Circled dots represent localities at which early Pliocene SST estimates were made. Contours represent estimated differences in SST in degrees Celsius. Early Pliocene SST in northeastern Atlantic was 5–6 °C higher in early Pliocene compared to today, suggesting increased heat transport by the Gulf Stream. Adapted from Dowsett HJ, Cronin TM, Poore RZ, Thompson RS, Whitley RC, and Wood AM (1992) Micropaleontological evidence for increased meridional heat transport in the North Atlantic Ocean during the Pliocene. *Science* 258: 1133–1135, reproduced with permission. © 2001 AAAS.

eastern equatorial Pacific, possibly because the Indonesian Seaway was more open, which would prevent pooling of water in the western equatorial Pacific. The more uniform temperature conditions from west to east would then foster increased meridional atmospheric heat transport (enhanced Hadley cell circulation) which provided the mechanism for increased warming in high and mid-latitudes (Molnar and Cane, 2002). Comparison of early Pliocene reconstructed climate estimates with climate anomalies experienced during El Niño events (Figure 6) show some similarities and provide support for the interpretation that modern El Niño teleconnections could explain early Pliocene conditions. The comparison must be evaluated with caution because the estimates of Pliocene conditions are derived from different times in the Pliocene and integrate different intervals of time. A recent SST reconstruction for the eastern equatorial Pacific (Wara et al., 2005) indicates that the temperature gradient across the equatorial Pacific was substantially reduced during the Pliocene compared to today, which is consistent with a permanent El Niño-like state (Figure 7). However, the reduced temperature gradient persisted until ~1.8 Ma, well after high-latitude cooling and build-up of continental ice sheets began in the late Pliocene. Thus, there is conflicting evidence for the correspondence of a reduced west-to-east thermal gradient in the equatorial Pacific and mid-to-high-latitude warming.

A number of modeling studies using PRISM boundary conditions have attempted to test the different mechanisms for explaining the warm Pliocene climates. A recent experiment using a fully coupled ocean–atmosphere model with PRISM boundary conditions and an atmospheric CO₂ level of 400 ppm indicated that lower albedo related to reduction of continental and sea ice may be the primary reason for increased warmth in mid- and high latitudes. The model

results also suggest that poleward heat transport was not an important factor in Pliocene warmth (Figure 8; Haywood and Valdes, 2004).

Intensification of NHG

The late Pliocene is marked by the development of significant Northern Hemisphere continental ice sheets, associated cooling of high latitudes, and the development of distinct glacial–interglacial cycles that are characteristic of the Pleistocene. The history of NHG is difficult to decipher from continental records because recurring glaciations often override and obscure the evidence for prior glacial advances. Examination of ice-rafted grains in deep-sea sediments adjacent to continents is often used to trace the evolution of continental ice sheets. When continental ice sheets reach coastlines, icebergs are calved and float out into the ocean. The subsequent melting of the icebergs releases rock fragments and other debris trapped in the ice and the debris drops to the sea floor. The amount, composition, and areal distribution of the ice-rafted debris (IRD) in marine sediments can then be used to infer the history of ice sheets and glaciers on the continents. Examination of sediment cores from the Iceland and Voring Plateaus reveals sporadic occurrences of IRD as far back as the Miocene. However, the ice-rafted material is minor and it is not clear if it is derived from sea ice or icebergs. Figure 9 shows the abundance of IRD in a deep-sea core recovered from Ocean Drilling Site 907 in the Norwegian–Greenland Sea adjacent to southwestern Greenland. The occurrence of IRD in the core at about 3.3 Ma provides evidence that enough continental ice had built up at that time in southern Greenland to release icebergs out into Norwegian–Greenland Sea. After 3.0 Ma, the consistent presence of IRD and recurring peaks in abundance represent the beginning of permanent ice caps that periodically varied in size (Jansen et al., 2000).

The character and age of rocks on continents is variable. Analyses of the types of ice-rafted rock fragments and mineral grains contained in sediment cores adjacent to continental areas can be used to identify the source area of grains and thus provide information on the time of formation and geographic extent of past glaciers. A summary of the build-up of continental ice sheets in the Northern Hemisphere is shown in Figure 10. The formation of small ice sheets began during the early Pliocene in southern Greenland and then spread to northern Greenland by about 3.0 Ma. Between 2.8 and 2.5 Ma ice sheets developed in Eurasia, Alaska, and North America. As noted in Figure 10, the development of Northern Hemisphere continental ice sheets was a phased process. As much as 200 ka may have elapsed between the establishment of major ice sheets in Alaska and Eurasia and the establishment of ice sheets in North America. Glacial–interglacial oscillations become distinct in the deep-sea oxygen isotope records between 2.9 and 2.7 Ma coincident with the continued build-up of the Northern Hemisphere continental ice sheets. At the same time that continental ice sheets were building up in the Northern Hemisphere, the Antarctic Ice Sheet was also expanding. Details of the late Pliocene history of the Antarctic Ice Sheet are not well known.

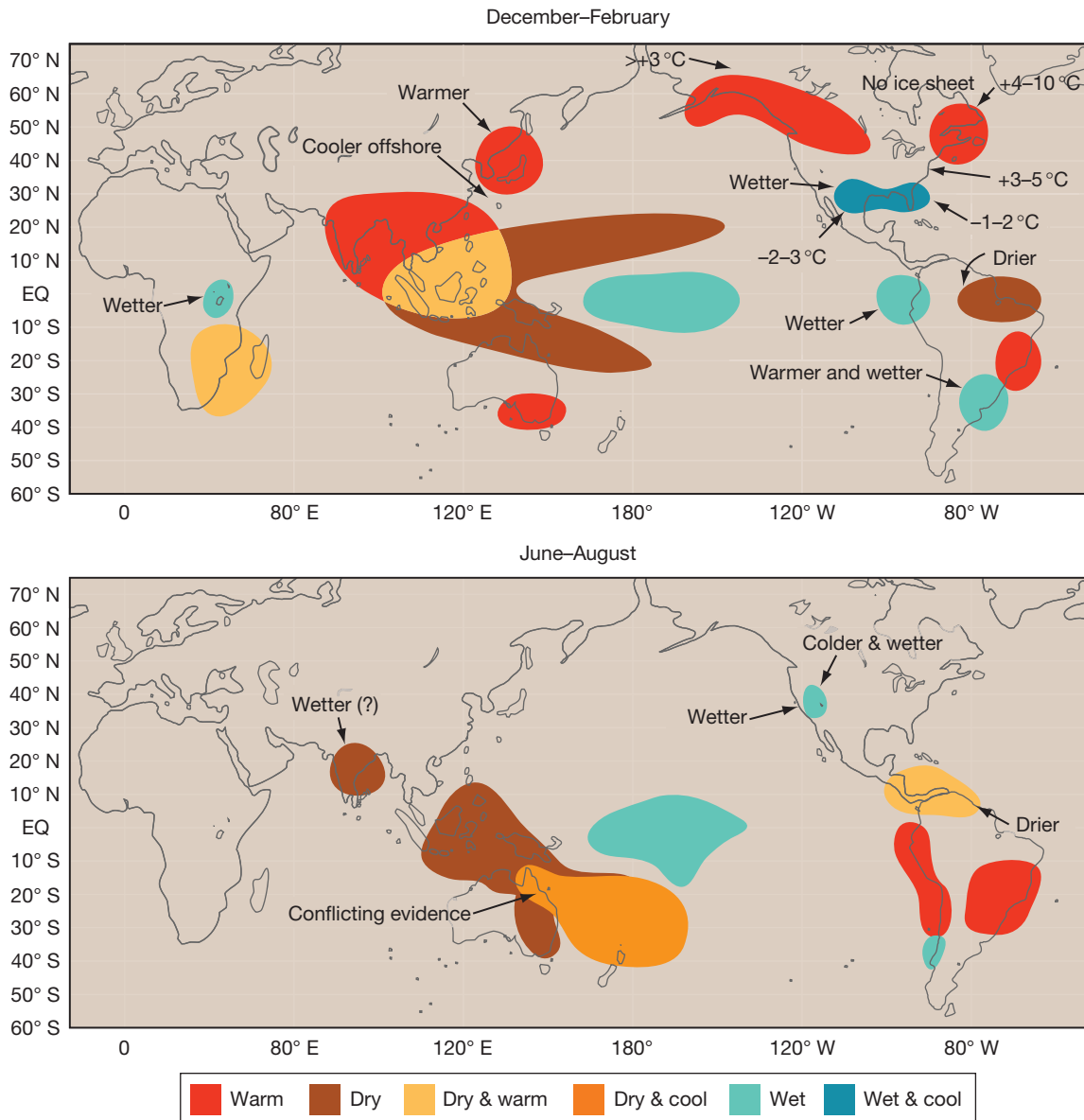


Figure 6 Map comparing typical El Niño precipitation and temperature anomalies in winter (December–February) and summer (June–August) with selected early Pliocene climate estimates. Colored shapes designate areas of El Niño anomalies. Pliocene climate estimates are noted in black lettering. Modern El Niño anomalies are from the National Weather Service Climate Prediction Center. Pliocene climate estimates are from various sources. See discussion in [Molnar and Cane \(2002\)](#). Adapted from Molnar R and Cane M (2002) El Niño's tropical climate and teleconnections as a blueprint for pre-Ice Age climates. *Paleoceanography* 17: 11-1–11-12, reproduced with permission. © 2002 American Geophysical Union.

Cause of NHG

Several tectonic events that altered oceanic or perhaps atmospheric circulation have been proposed to explain the intensification of NHG in the late Pliocene. Some of the possible events include the closing of the Isthmus of Panama, the closure of the Indonesian seaway, the uplift of the Tibetan Plateau, the deepening of the Bering Straits, and the deepening of the Greenland–Scotland Ridge ([Cane and Molnar, 2001](#); [Maslin et al., 1998](#)). When examined closely, none of these events appears to provide a complete explanation for the intensification of NHG. For example, the closure of the Isthmus

of Panama caused an increase in salinity of the Caribbean and presumably increased the strength of the Gulf Stream. It has been suggested that increased transport of warm, salty water to the North Atlantic by an enhanced Gulf Stream would have provided a source of moisture to grow ice sheets on the adjacent continents and also increase formation of North Atlantic deep water (NADW). However, the closure of the Isthmus of Panama was not a discrete event; it occurred gradually and over a long period, from about 4.5 to 2.0 Ma. In addition, carbon isotope records from North Atlantic deep-sea cores indicate NADW formation decreased from 3.5 to 2.0 Ma.

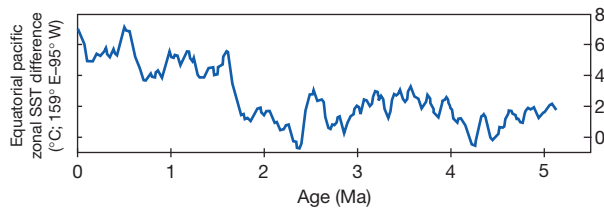


Figure 7 Estimated temperature gradient between western and eastern equatorial Pacific Ocean for last 5 Ma. Note that the gradient was greatly reduced from 1.8 to 5.0 Ma compared to gradient of 4–7 °C after 1.8 Ma. See text for discussion. Adapted from Wara M, Ravelo A, and Delaney M (2005) Permanent El Niño-like conditions during the Pliocene warm period. *Science* 309: 758–761, reproduced with permission. © 2005 AAAS.

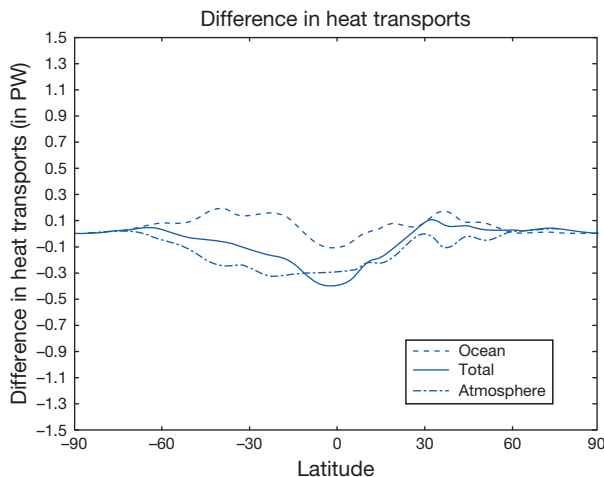


Figure 8 Model predicted difference in poleward heat transport (PW) between Pliocene coupled and present-day coupled model experiments using the HadCM3 GCM. The plot indicates that poleward heat transport in the Pliocene and modern simulations are very similar. Reprinted from Haywood AM and Valdes PJ (2004) Modelling Pliocene warmth: Contributions of atmosphere, oceans and cryosphere. *Earth and Planetary Science Letters* 218: 363–377. © 2004, reproduced with permission from Elsevier.

A reduction in atmospheric CO₂ concentration has been proposed as the cause for NHG. However, there is no clear evidence for a significant lowering of atmospheric CO₂ levels coincident with the intensification of NHG. In addition, a cooling related to lowered atmospheric CO₂ should be recorded in low latitudes but there is no evidence that tropical climates cooled at the beginning of the late Pliocene.

Variations in solar insolation reaching Earth's surface due to changes in Earth's orbit (orbital forcing) have also been proposed as an important cause for the intensification of NHG. In theory, ice sheets should form when summer temperatures are low enough to allow winter frozen precipitation (snow and ice) to survive the summer and thus begin a positive cooling feedback through increased surface albedo. The initial build-up of ice sheets in Alaska, Eurasia, and northeastern North America between 2.75 and 2.55 Ma (**Figure 10**) occurred when calculations show Northern Hemisphere summer insolation at 65°N reached minimum values (Maslin et al., 1998; **Figure 6**). However, similar or even lower summer

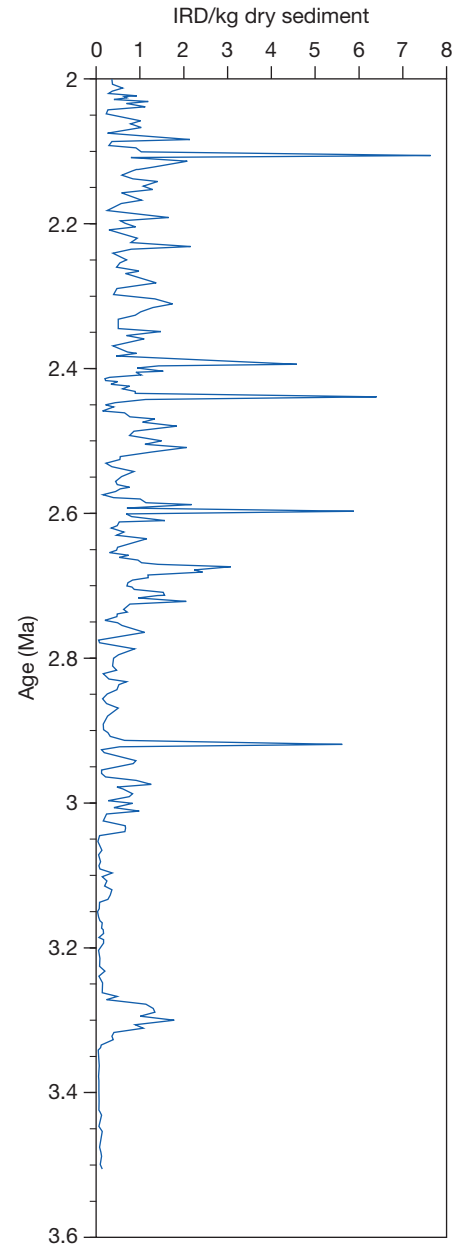


Figure 9 Abundance of ice-rafted debris (IRD) in sediments from ODP Site 907 just east of southern Greenland. Note consistent occurrence of significant amounts of IRD after 3.0 Ma. Adapted from Jansen E, Fronval T, Rack R, and Channell J (2000) Pliocene–Pleistocene ice rafting history and cyclicity in the Nordic Seas during the last 3.5 Myr. *Paleoceanography* 15: 709–721, reproduced with permission. © 2000 American Geophysical Union.

insolation values occur periodically earlier in the Pliocene, so summer insolation variability is not the sole cause of NHG.

Stratification and the resulting late summer warming of North Pacific surface waters at about 2.7 Ma have also been linked with the intensification of NHG. In addition to cooling, a source of moisture must be available to permit sufficient snow to grow ice sheets. In the modern ocean there is a large seasonal variation in SST in the sub-Arctic Northwest Pacific due to the formation of a seasonal thermocline. During

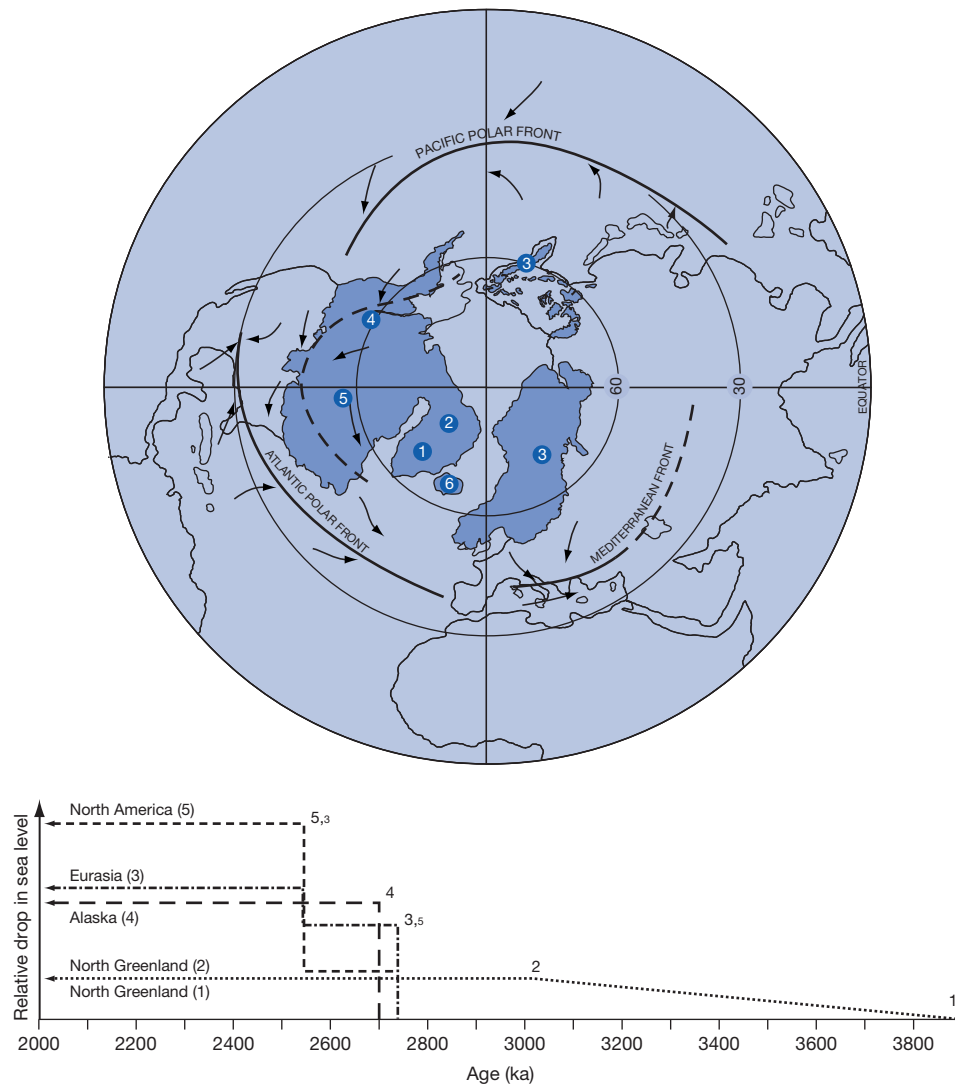


Figure 10 Map and graph showing location and timing of build-up of Northern Hemisphere continental ice sheets large enough to release icebergs into the ocean. Repeated glacial build-up occurred in Iceland during the Pliocene but the timing of the first sustained ice sheet is not well established. Open circles with three digit numbers represent Ocean Drilling Program cores used to establish ice sheet history. Arrows and fronts outline possible atmospheric circulation in preglacial Pliocene. Adapted from Maslin MA, Li XS, Loutre MF, and Berger A (1998) The contribution of orbital forcing to the progressive intensification of Northern Hemisphere glaciation. *Quaternary Science Reviews* 17: 411–426. © 1998, reproduced with permission from Elsevier.

winter, when no thermocline is present, surface waters are well mixed with subsurface water. SST is about 1 °C (February) and surface waters are rich in nutrients mixed from below. In summer, when a thermocline is present, surface waters are isolated from cool subsurface waters and late summer SST (September) is 12 °C. Estimates of North Pacific SST derived from analyses of alkenones, which likely reflect late summer SST (Haug et al., 2005) indicated SST increased by about 7 °C near 2.7 Ma. The increase in temperature is considered to reflect the development of a seasonal thermocline in the North Pacific, which would allow the large seasonal range in SST observed in the modern ocean. The presence of warm surface waters in the North Pacific in fall would enhance transport of moisture onto the adjacent cooler continents and increase snowfall and thus promote build-up of continental ice sheets. Haug et al. (2005) proposed that prior to the development of a seasonal thermocline at

2.7 Ma, surface waters would have remained cool during the seasonal cycle due to mixing with cooler subsurface waters. One problem with this theory is that alkenone SST estimates indicate that sub-Arctic North Pacific SSTs were nearly as warm at 2.9 and 3.0 Ma as they were at 2.7 Ma. In addition, the assumption that seasonal signals can be extracted from Pliocene proxy records needs additional testing.

Comparison of detailed climate proxy records from tropical to subpolar areas in the Northern Hemisphere indicates no clear pattern of cooling in the late Pliocene across latitudes or regions. Ravelo et al. (2004) concluded that regionally specific processes caused climatic cooling at different times and that the termination of the early to mid-Pliocene warm interval was not forced by a single event whose effects were propagated globally. Thus, as of this writing, no clear single cause has been identified as the ‘triggering event’ leading to the NHG.

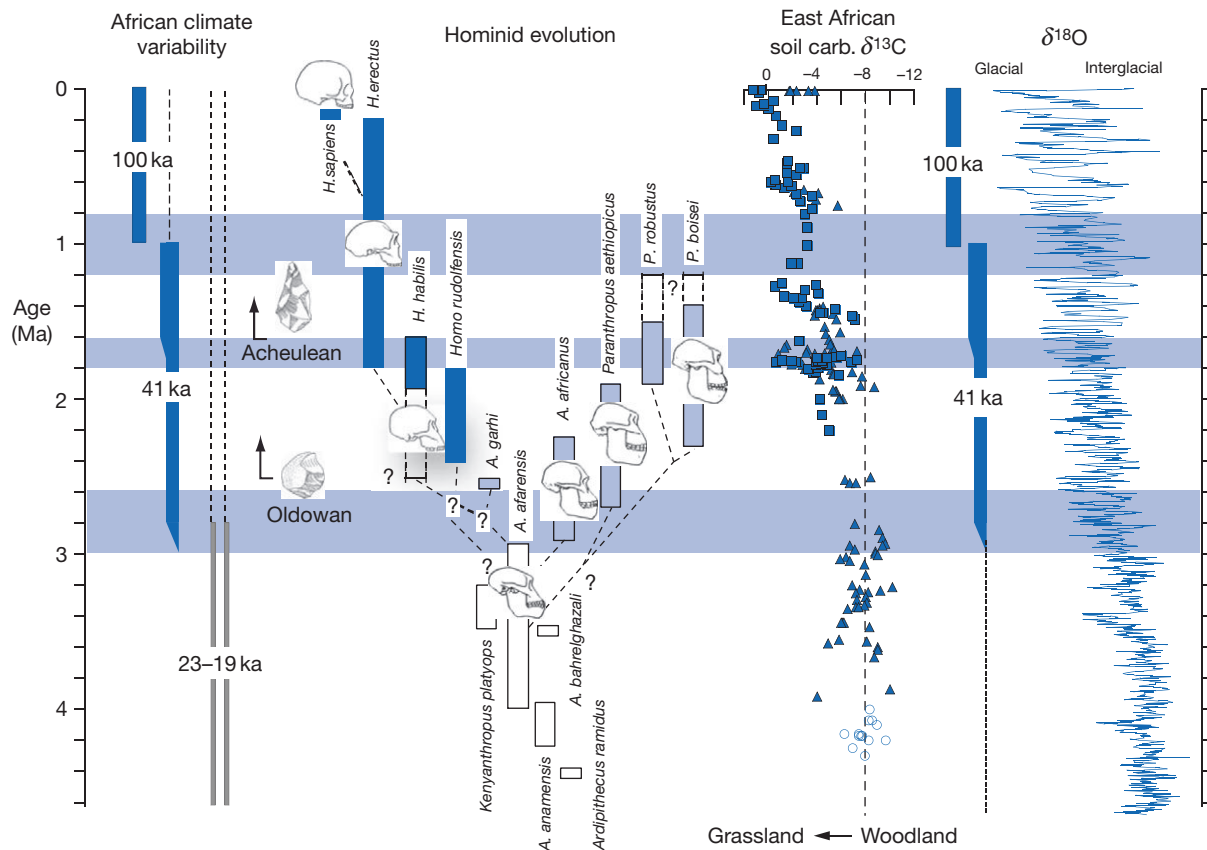


Figure 11 Summary of Hominid Evolution in Africa compared with evolution of Pliocene and Pleistocene climate as measured by soil carbon isotope values in East Africa and composite benthic foraminifer deep-sea isotope record. Note increase in Hominid speciation coincides with expansion of grasslands, initiation of NHG, and change to 41 ka cycles near 3.0 Ma. Adapted from deMenocal PB (2004) African climate change and faunal evolution during the Pliocene–Pleistocene. *Earth and Planetary Science Letters* 220: 3–24. © 2004, reproduced with permission from Elsevier.

Links Between NHG and Human Evolution

Records of African climate and climate variability derived from studies of deep-sea cores and continental records generally show a change from wetter and warmer conditions in the early Pliocene to cooler and drier conditions with higher variability in the late Pliocene. Pollen records from a variety of sites indicate that woodland areas were replaced by more open savannah grasslands (deMenocal, 2004; Thompson and Fleming, 1996). After about 3.0 Ma, variability in African wet-dry cycles as monitored by the abundance of African-source dust in deep-sea cores indicates that African climate cycles correspond closely with variation in high-latitude continental ice volume cycles. In general, glacial climates result in cooler and more arid conditions. Changes in the African mammal fauna including hominids broadly coincides with the change in climate and landscape that marks the development and intensification of NHG (Figure 11). Analyses of bovids (antelope group) and micromammal assemblages (e.g., rodents) indicate that arid-adapted taxa were more prevalent after 3.0 Ma. The diversification of hominids coincides with the change in pace of climate variability and the expansion of grasslands. It is not clear if the change in timing and magnitude of climate variability or the change to more arid conditions or some combination of both was the primary cause of the change in mammal faunas

(deMenocal, 2004). Although the details are still being debated, the available evidence indicates that the evolution of hominids was linked to the change in climate associated with the onset and intensification of NHG during the late Pliocene.

See also: **Glacial Landforms, Sediments:** Glaciomarine Sediments and Ice-Rafted Debris. **Glaciation, Causes:** Astronomical Theory of Paleoclimates. **Paleoceanography, Biological Proxies:** Alkenone Paleothermometry Based on the Haptophyte Algae. **Paleoceanography, Physical and Chemical Proxies:** Oxygen-Isotope Stratigraphy of the Oceans. **Paleoclimate Reconstruction:** Paleodroughts and Society. **Plant Macrofossil Methods and Studies:** CO₂ Reconstruction from Fossil Leaves.

References

- Cane M and Molnar P (2001) Closing of the Indonesian seaway as a precursor to east African aridification around 3–4 million years ago. *Nature* 411: 157–162.
- Crowley TJ (1996) Pliocene climates: The nature of the problem. *Marine Micropaleontology* 27: 3–12.
- deMenocal PB (2004) African climate change and faunal evolution during the Pliocene–Pleistocene. *Earth and Planetary Science Letters* 220: 3–24.
- Dowsett HJ, Barron JA, and Poore RZ (1996) Middle Pliocene sea surface temperatures: A global reconstruction. *Marine Micropaleontology* 27: 13–25.

- Dowsett HJ, Cronin TM, Poore RZ, Thompson RS, Whatley RC, and Wood AM (1992) Micropaleontological evidence for increased meridional heat transport in the North Atlantic Ocean during the Pliocene. *Science* 258: 1133–1135.
- Dowsett HJ, Barron JA, Poore RZ, et al. (1999) Middle Pliocene paleoenvironmental reconstruction; PRISM2. U.S. Geological Survey Open-File Report 99-535.
- Haug GH, Ganopolski A, Sigman DM, et al. (2005) North Pacific seasonality and the glaciation of North America 2.7 million years ago. *Nature* 433: 821–825.
- Haywood AM and Valdes PJ (2004) Modelling Pliocene warmth: Contributions of atmosphere, oceans and cryosphere. *Earth and Planetary Science Letters* 218: 363–377.
- Jansen E, Fronval T, Rack R, and Channell J (2000) Pliocene–Pleistocene ice rafting history and cyclicity in the Nordic Seas during the last 3.5 Myr. *Paleoceanography* 15: 709–721.
- Kurschner WM, van der Burgh J, Visscher H, and Dilcher D (1996) Oak leaves as biosensors of Late Neogene and Early Pleistocene paleoatmospheric CO₂ concentrations. *Marine Micropaleontology* 27: 299–312.
- Maslin MA, Li XS, Loutre MF, and Berger A (1998) The contribution of orbital forcing to the progressive intensification of Northern Hemisphere glaciation. *Quaternary Science Reviews* 17: 411–426.
- Molnar R and Cane M (2002) El Niño's tropical climate and teleconnections as a blueprint for pre-Ice Age climates. *Paleoceanography* 17: 11–1–11–12.
- Poore RZ and Sloan LC (1996) Climates and climate variability of the Pliocene. *Marine Micropaleontology* 27: 1–326.
- Ravelo AC, Andreasen DH, Lyle M, Lyle AO, and Wara MW (2004) Regional climate shifts caused by gradual global cooling in the Pliocene epoch. *Nature* 429: 263–267.
- Raymo M, Grant B, Horowitz M, and Rau G (1996) Mid-Pliocene warmth: Stronger greenhouse and stronger conveyor. *Marine Micropaleontology* 27: 313–326.
- Thompson RS and Fleming RF (1996) Middle Pliocene vegetation: Reconstructions, paleoclimatic inferences, and boundary conditions for climate modeling. *Marine Micropaleontology* 27: 27–49.
- Wara M, Ravelo A, and Delaney M (2005) Permanent El Niño-like conditions during the Pliocene warm period. *Science* 309: 758–761.
- Zachos J, Pagani M, Sloan L, Thomas E, and Billups K (2001) Trends, rhythms, and aberrations in global climate 65 Ma to present. *Science* 292: 686–693.

Paleodroughts and Society

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Introduction

Severe drought is a recurring natural hazard that has prominent economic and agricultural impacts on society. Many societal planning schemes are based on the modern records of interannual and decadal drought variability (Stockton, 1990). The modern record of drought is restricted temporally and geographically, depending on the availability of systematic instrumental weather records. For example, our observed record of drought for North America covers mostly the last 100 years, whereas such modern records cover substantially longer periods for Europe and East Asia. Some scholars have conducted research on the observed record of severe drought, noting periodic cycles of aridity for some regions, such as the 22-year drought cycle in interior North America. These drought cycles have been somewhat disputed because of the temporal limitation of coverage by the modern instrumental record. Paleoclimatic reconstructions of drought have been conducted from a variety of different proxy data, with some reconstructions translating proxy terms into drought indices (e.g., Stable et al., 1998), and others providing more relative or presence/absence information on drought conditions (e.g., Mason et al., 1994). These paleoclimatic reconstructions provide a longer perspective of drought variability for a region, enabling a clearer analysis of the forcing mechanisms of severe drought, improving probabilistic models of future drought prediction, and contributing to a more comprehensive analysis of drought impacts on past societies. This article presents an overview of the way some of the important types of terrestrial paleoclimatic data are utilized in a drought context at different temporal scales within the Holocene, and the way they are assessed in conjunction with potential forcing mechanisms of past climate, emphasizing examples from North America. The article emphasizes temporal variability at annual to decadal timescales, as this variability is the common element that directly links past drought to the modern instrumental record (Overpeck and Webb, 2000). Data for the paleoclimatic reconstructions used in this article were taken from the NOAA Paleoclimatology website at <http://www.ncdc.noaa.gov>.

Paleoclimatic Proxies of Past Drought

This section does not describe the extreme details of all the different types of proxy data or the numerous complexities of drought signals from different proxy data types. For further information, refer to Bradley (1999) and specific references cited in it related to particular proxy data and dating techniques.

Historical Records

Historical (noninstrumental) evidence of drought consists of accurately dated documentary records such as diaries,

newspapers, and chronicles (Baron, 1989). Documentary records provide the highest resolution of all the paleoclimatic proxy data types, enabling detailed reconstructions of daily weather events that may lead up to severe drought. Some documentary evidence, such as annals, letters, and almanacs, possess less temporal resolution. They are reflective of monthly, seasonal, or annual conditions. However, these lines of evidence still provide some general information on extreme drought years (e.g., Therrell et al., 2004). The documentary materials, however, cannot be taken at face value, as poor data quality from unreliable records makes some materials inappropriate for climate reconstruction. Written descriptions of drought may also reflect individual biases based on diarists' personal background and particular motivation for writing. Thus, a thorough knowledge of the documentary materials is essential, and copious daily records enable more accurate quantitative assessments of past drought, such as by analyzing the number of rain days. Content analysis also provides a means of quantifying documentary information into drought and climate terminology (Baron, 1989).

An example of a composite of warm-season rain days has been compiled for Charleston, South Carolina, from numerous documentary records dating back to the early nineteenth century (Figure 1). A distinctive persistent period of relatively lower number of rain days is evident for the mid-nineteenth century, during a time frame when documentary evidence is clearly plentiful. Additional descriptions of verbal evidence from plantation diaries in South Carolina during this decrease of precipitation frequency reveal that farmers continuously wrote about drought, but note that precipitation frequency can differ from precipitation magnitude, so it should not be interpreted as being necessarily the worst period of drought severity in the entire record.

Tree Rings

Trees that exhibit seasonal variations in light, thin-walled earlywood and dark, thick-walled latewood cells in tree growth have annual rings, thus making tree rings an important proxy source for analyzing climatic change at shorter timescales within the last few thousand years. Tree rings, particularly at sites from lowland moisture-sensitive areas, have been used extensively for reconstructing past precipitation and drought in North America during the last several hundred years, with some sites reconstructing drought history over the last millennium (e.g., Hughes and Funkhouser, 1998). Researchers have examined the widths of tree rings as well as tree-ring densities in relation to climate (Schweingruber, 1987). Some tree-ring reconstructions of past drought focus on site-specific areas at about the size of a drainage basin, and some reconstructions apply multivariate statistical methods on a network of tree-ring sites at regional and national scales (Cook and Krusic, 2004). Reconstructions also vary in their temporal resolution, as some

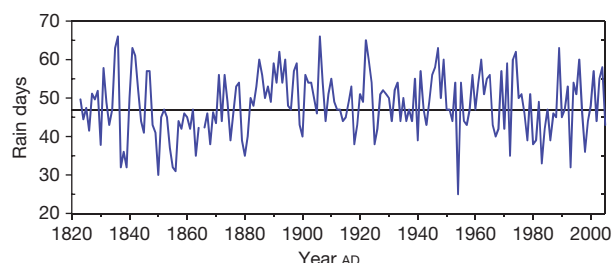


Figure 1 Annual frequencies of June–September rain days in Charleston, South Carolina, USA from 1823 to 2005, as reconstructed from documentary evidence.

studies may be reflective of drought of a particular seasonality, and some reconstructions may generally reflect annual or longer conditions of drought. For example, Woodhouse et al. (2002) reconstructed May–July Palmer Drought Severity Indices (PDSI) back to the 1500s for Eastern Colorado, USA. Interpretation of the PDSI equates decreased negative values as corresponding to increased severity of drought. This reconstruction indicates severe droughts centered around the 1950s, 1930s, 1840s, and a series of more frequent severe droughts during the 1570–1750 period. This reconstruction is typical of those that do not clearly illustrate distinctive specific drought cycles throughout the entire record. The nature of more persistent and frequent droughts in the past prior to the modern instrumental record reveals the importance of obtaining a longer temporal perspective on drought. Cook and Krusic (2004) utilized a network of 835 tree rings to reconstruct annual spatial patterns back several centuries for North America. Their example for 1820 indicates that severe drought is clearly evident over the southern Great Plains and much of the southwestern United States (Figure 2), which corresponds to the Long expedition's travels across the region that created the notion of the Great American Desert (Lawson and Stockton, 1981).

Lake Records

A variety of proxy evidence concerning the paleolimnology of lakes can be applied to detect decadal drought variations as well as those that prevailed up to centennial timescales for the late Quaternary period (Street-Perrott and Harrison, 1985). The geochemistry from lake cores also provides evidence since chemical and isotopic components change in response to salinity changes which are related to past drought. Radiocarbon dating provides most of the dating control for the last 20 ka. However, varves may occasionally be found in some lakes, providing annual dating, while trace elements (e.g., sodium and potassium), pollen, and diatoms can also be examined to obtain paleoclimatic information (Bradbury et al., 1993). Much research has emphasized the reconstruction of past lake levels, including those with closed basins (e.g., Harrison et al., 1996 and Shuman et al., 2002). At larger regional and continental spatial scales, broad synchronous decreases in lake levels from a high number of sites reflect prolonged severe drought. Generally, lake-level fluctuations are due to combined effects of precipitation, evaporation, lake area, runoff, discharge, groundwater, and tectonics. Lower lake levels may generally reflect decreased effective

moisture, which includes decreased precipitation and/or increased evaporation.

Considerable attention has been focused on the reconstruction of spatial patterns of drought in the mid-Holocene, around 6 ka, for central and western North America. Lake levels were interpreted as low relatively to modern day, enabling qualitative interpretations that widespread aridity existed over the Pacific Northwest, extending into the western interior of the Great Basin. This widespread aridity corresponds to the Altithermal, which has been used for a number of decades to describe persistent mid-Holocene drought over parts of North America (Harrison et al., 2003; Mock and Brunelle-Daines, 1999). The causes of widespread drought are largely related to synoptic-scale atmospheric circulation that includes a stronger Pacific subtropical high over the eastern Pacific Ocean, and a strong subtropical ridge over the mid-continent.

Fossil Pollen

The accumulation of fossil pollen over time at a site as reflected in cores of peat bogs and lake muds, aided by radiocarbon dating, enables the reconstruction of vegetation history and related paleoclimatic interpretation, including the study of drought at millennial timescales. Care must be taken when making paleoclimatic interpretations from pollen, because of problems with pollen dispersal and production rates and lake basin characteristics. Vegetation may always not always be in equilibrium with climatic change (Davis and Shaw, 2001), with factors including substrate, migration lag, and disturbances such as fire playing a more dominant role in vegetation change. However, widespread drought and vegetation are generally in equilibrium with one another on late Quaternary timescales within approximately 1.5-ka intervals at the continental scale (Shuman et al., 2002). Evidence of past drought from pollen data can be conducted through qualitative methods such as using an indicator-species approach, presuming that particular taxa respond to the primary climatic changes that occurred in the past. A multiproxy map of paleoclimate for the mid-Holocene shows that pollen data can supplement lake-level data in relative terms. Results for 6 ka from pollen assemblages also support widespread aridity over the Pacific Northwest US and into the western interior (Mock and Brunelle-Daines, 1999), and the pollen data have been taken from more sites than have the lake-level data for the western United States. Quantitative approaches, using modern vegetation/climate relationships and multivariate statistical methods, have also been used to reconstruct the past.

Diatoms

Diatom records can reconstruct paleoclimatic conditions at decadal to millennial timescales within the late Pleistocene (e.g., Laird et al., 2003), given precise dating (e.g., varves, ^{210}Pb dating, ^{14}C dating). Since different species of diatoms thrive in different paleolimnological conditions, some provide specific drought information, based on such variables as salinity, nutrient availability, and habitat availability. Also, as with pollen, statistical methods can be applied to diatom data to quantitatively reconstruct paleoclimate (Moser et al., 1996). An example of a reconstruction of salinity from Moon Lake in the northern Great Plains illustrates that decadal to

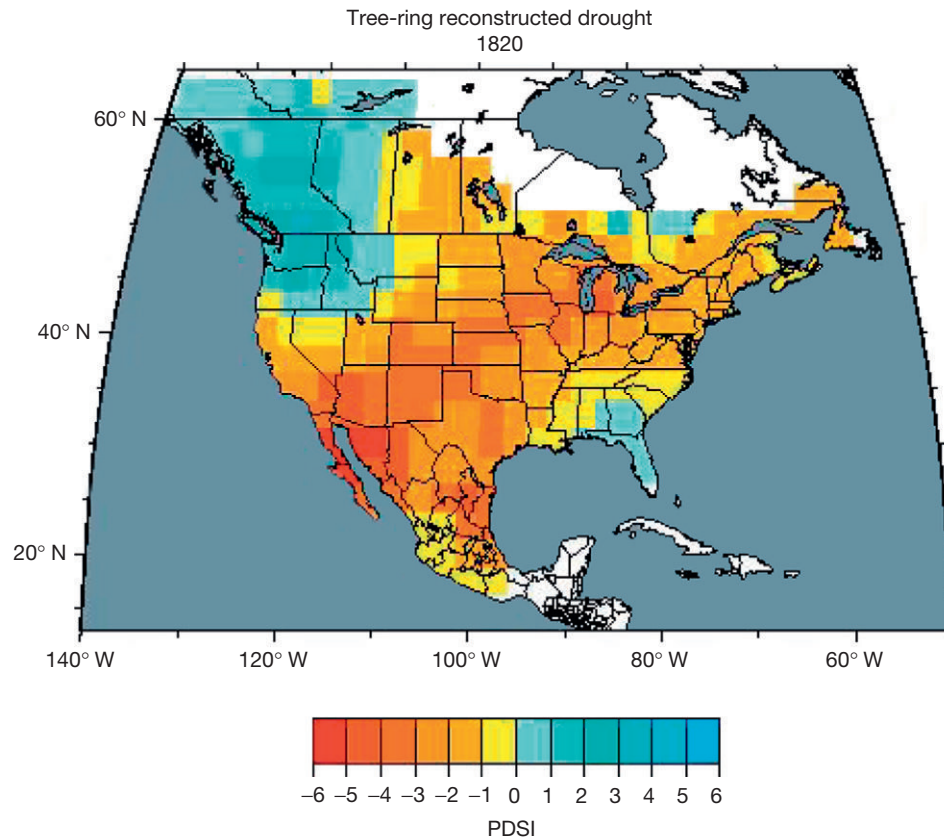


Figure 2 Reconstructed Palmer Drought Severity Indices (PDSI) map for 1820 from tree rings (Cook and Krusic, 2004).

centennial variations of drought were more frequent prior to 1.2 ka (Laird et al., 1996). One of the most severe drought intervals was centered around the Medieval Warm Period from AD 800 to 1200. However, this dry episode may reflect regional changes in synoptic-scale atmospheric circulation or perhaps a difference in seasonal precipitation regimes, as similar diatom reconstructions for other sites in the northern Great Plains differ somewhat in their reconstruction of drought history (Laird et al., 2003). Fossil diatom assemblages from Moon Lake extend back through the Holocene. These data reveal that persistent drought was prevalent from around 7.3 to 4.4 ka, corresponding to the widespread mid-Holocene aridity reflected by other drought proxies (e.g., pollen and lake levels) in North America.

Eolian Records

The stratigraphy of eolian deposits, such as sands, may indicate periods of past aridity and dune reactivation. One example of this comes from aridity in the mid-continent of North America around 6 ka (e.g., Muhs and Zarate, 2001 and Wolfe et al., 2002). Paleowind directions may be inferred from features such as sand dunes, yardangs, and wind furrows. However, nonclimatic factors such as faulting, and site-specific variations in sediment thickness and texture need to be considered. Furthermore, for some locations, single extreme windstorms of a few days' duration may contribute much of the eolian

deposition, as opposed to sustained background deposition over centuries.

Climatic Forcing Mechanisms of Past Drought

General circulation models (GCMs) simulate the geographical distribution of climatic variables at hemispheric and global scales. They are based on fundamental laws that include the ideal gas law, conservation of energy, conservation of mass, and Newton's second law of motion. These fundamental laws explain, theoretically, how the physical processes occur between, within, and among the different components of the climate system. The physical processes are represented by layers of horizontal grids, with the number of layers as well as horizontal resolution varying between different GCMs. Given an appropriate set of boundary conditions of a past interval of interest to a researcher, GCMs can simulate the three-dimensional character of particular episodes of severe drought. Climate modeling of drought for episodes in the modern climate record of North America has received significant attention. For example, Schubert et al. (2004) illustrated that anomalous negative sea-surface temperatures in the tropical Pacific combined with anomalous positive sea-surface temperatures in the tropical Atlantic weaken the low-level jetstream in the southern Great Plains. This situation corresponds with patterns of severe drought over central North America, resembling the 1930s Dust Bowl.

Seager et al. (2005) also modeled the response of low-frequency drought episodes over North America from tropical forcing, not only illustrating the importance of sea-surface temperature forcing but also acknowledging the additional importance of land–atmosphere interactions. In addition, they described the importance of cool ‘La Niña-like’ sea-surface temperatures in the eastern tropical Pacific Ocean on creating the persistence of severe drought. Herweijer et al. (2006) performed modeling simulations of the following North American droughts: 1856–65, 1870–77, and 1890–96, as well as two Medieval drought episodes in the western United States. They further illustrate the importance of a colder tropical Pacific Ocean in creating severe drought. These model simulations generally compare quite well with drought networks of proxy data. Hunt and Elliott (2002) attempted to simulate Mexican megadroughts (persistent droughts of more than a decade in duration), suggesting the importance of El Niño events and various stochastic processes, but their work implied that a more specific understanding of the potential mechanisms is needed to improve the modeling of a megadrought.

Numerous model simulations exist for 6 ka because this time frame was the particular focus of the Paleoclimate Modeling Intercomparison Project (Joussaume et al., 1999). Practically, all of the models agree on the positions and anomalies of large-scale synoptic features over North America (e.g., a stronger Pacific subtropical high during most of the summer). However, some subtle differences in the specific positions of synoptic features, such as the upper-level subtropical ridge over the southern United States, can play important roles on the geographical distribution of monsoonal activity and areal extent of aridity (e.g., Mock and Brunelle-Daines, 1999). The climatic forcing mechanisms at 6 ka are also related to orbital parameters of an increased tilt and changes in precession, plus important ocean feedbacks were also prominent as well (Harrison et al., 2003), and different seasonal precipitation regimes are characteristic of the climate over western and central North America (Mock, 1996). Therefore, one cannot limit the causes of widespread drought in the mid-Holocene to just a few specific parameters.

Impacts of Paleodrought on Past Societies

Although the impacts of drought on earlier societies cannot be assessed in the same ways as can be done in the modern era, some useful lessons can nonetheless be learned from an understanding of premodern drought impacts. The roles of social, economic, and political factors have been studied extensively, using case studies of societal collapses and political unrest, but the role that the climate, including drought, has played in human societies has not received due attention in many cases. A longer temporal perspective of drought and societal impacts can be invaluable in assessing how societies adapt to drought in longer time frames as they also involve technological and ideological changes through time (deMenocal, 2001). Some scholars have long been interested in the role of drought in societal collapses and disappearances, including the case study of the Akkadian collapse in the Near East around 4.2 ka (Cullen et al., 2000). Establishing clear linkages between drought and the decline of past societies requires high-quality

dating control on both the drought events of interest and on evidence of the impacts (e.g., archaeology). For example, some societies may be quite sensitive to single or perhaps a series of extreme drought years, and this requires dating no less precise than annual resolution. Some societies may be quite responsive to decadal drought variations, which can be dated using calibrated radiocarbon.

An 800-year tree-ring record of baldcypress trees was used to reconstruct July Palmer Hydrological Drought Indices from the Tidewater region of coastal Virginia and North Carolina. This record reveals two extremely severe droughts that coincide with prominent historical events. A persistent and severe 7-year drought occurred from 1606 to 1612, which contributed immensely to crop failures and poor water quality at the Jamestown Colony – this eventually contributed to the high mortality rate (Blanton, 2000; Stahle et al., 1998). A shorter but more severe drought is also evident for 1587–89, which coincided with the disappearance of the Lost Colony at Roanoke Island. Although nonclimatic factors clearly played important roles in both of these historical events, the strong negative drought anomalies cannot be ignored as an important factor, for even modern society would face extreme hardships in such extreme drought conditions. Overall, a prolonged drought of several decades in the late sixteenth century is evident in many tree-ring records from much of North America, corresponding to what has been termed the big ‘megadrought’ (Stahle et al., 2000). A number of prominent drought impacts on society have been linked to this megadrought, including epidemics of cocoliztli (indigenous hemorrhagic fevers) in the mid-1500s that decimated around 80% of the native population in Mexico (Acuna-Soto et al., 2002).

Widespread severe drought also occurred around the late 1200s over many parts of North America. Severe drought occurred in the southwestern United States (Cook et al., 2004), which may relate to the Anasazi collapse, and over the middle of the United States, which may explain the disappearance of the Mill Creek Indians in Iowa (Bryson and Murray, 1977). Overall, the western United States experienced more frequent and severe droughts from AD 900 to 1300 as compared to the normal variability of the last century (Cook et al., 2004). Another prominent and extremely severe drought that likely had a dramatic impact on society, as reflected in oxygen isotopes and gypsum from lake cores and sediment records from the Cariaco Basin in the Caribbean Sea, occurred mostly around AD 800–900 in the Yucatan Peninsula, causing the classic collapse of the Mayan civilization (e.g., Curtis et al., 1996; Haug et al., 2003).

Bowden et al. (1981) described two hypotheses that deal with the changing characteristics of climatic impacts on society through time – lessening and catastrophe – both dealing largely with drought hazard. The lessening hypothesis suggests that societies are better able to adapt to minor climatic stress through time given improved technology, increased social organization, and mitigation. The catastrophe hypothesis suggests that societies become more vulnerable to major climatic stress as time progresses due to their increased vulnerability, which is related to increased reliance on technology and social organization. Bowden et al. (1981) tested these hypotheses on three regions using available climatic and population data: India, the US Great Plains, and the African Sahel. They found

some clear indications of lessening, but results are less clear for catastrophe given the limitations on data quality and record length (including paleoclimatic data). Riebsame (1991) continued to examine these two hypotheses for the US Great Plains in extrapolating into a world of future global warming, addressing societal concerns for the catastrophic aspects. This framework of the hypotheses for assessing future climate impacts may prove quite valuable for some regions if they possess well-dated and interpreted paleodrought and archaeological evidence.

Conclusions

Reconstructions of past droughts prior to the era of systematic modern instrumental records provide a valuable longer perspective of drought that allows a fuller understanding of the recurrence of such extreme events and their duration and magnitude. The emphasis in this article has been on examples from North America, but such approaches for understanding past drought have also been employed for other areas of the world where drought is a prominent hazard (e.g., Overpeck and Webb, 2000; Touchan et al., 2003). These aspects of paleodrought at different temporal and spatial scales are critical for societal planning schemes such as dam and water regulation (Meko et al., 1995). Accurate reconstructions of paleodroughts can also aid in the study of famous historical events and the disappearance or disruption of past societies. Establishment of clear relationships between past drought and societal responses requires multidisciplinary approaches from the scientific, social science, and historical fields.

See also: Dendroclimatology; Paleoclimate Relevance to Global Warming. **Diatom Records:** Freshwater Laminated Sequences. **Lake Level Studies:** Overview. **Paleoclimate:** Timescales of Climate Change. **Paleoclimate Modeling:** Paleo-ENSO; Quaternary Environments. **Paleoclimate Reconstruction:** Approaches; Historical Climatology; The Last Millennium. **Plant Macrofossil Methods and Studies:** Dendroarchaeology. **Plant Macrofossil Records:** Arctic North America.

References

- Acuna-Soto R, Stahle DW, Cleaveland MK, and Therrell MD (2002) Megadrought and megadeath in 16th century Mexico. *Emerging Infectious Diseases* 8: 360–362.
- Baron WR (1989) Retrieving climate history: A bibliographical essay. *Agricultural History* 63: 7–35.
- Blanton DB (2000) Drought as a factor in the Jamestown colony, 1607–1612. *Historical Archaeology* 34: 74–81.
- Bowden MJ, Kates RW, Kay PA, et al. (1981) The effect of climate fluctuations on human populations: Two hypotheses. In: Wigley TML, Ingram MJ, and Farmer G (eds.) *Climate and History*, pp. 479–513. Cambridge: Cambridge University Press.
- Bradbury JP, Dean WE, and Anderson RY (1993) *Holocene Climatic and Limnologic History of the North-Central United States as Recorded in the Varved Sediments of Elk Lake, Minnesota: A Synthesis*, pp. 309–328. Boulder, CO: Geological Society of America Special Paper 276.
- Bradley RS (1999) *Paleoclimatology: Reconstructing Climates of the Quaternary*. San Diego: Academic Press.
- Bryson RA and Murray TJ (1977) *Climates of Hunger*. Madison: University of Wisconsin Press.
- Cook ER and Krusic PJ (2004) North American Summer PDSI Reconstructions. *IGBP PAGES/World Data Center for Paleoclimatology Data Contribution Series # 2004-045*. Boulder, CO: NOAA/NGDC Paleoclimatology Program.
- Cook ER, Woodhouse CA, Eakin CM, Meko DM, and Stahle DW (2004) Long-term aridity changes in the western United States. *Science* 306: 1015–1018.
- Cullen HM, deMenocal PB, Hemming S, et al. (2000) Climate change and the collapse of the Akkadian Empire: Evidence from the Deep Sea. *Geology* 28: 379–382.
- Curtis JH, Hodel DA, and Brenner M (1996) Climate variability on the Yucatan Peninsula (Mexico) during the last 3500 years, and implications for Maya cultural evolution. *Quaternary Research* 46: 37–47.
- Davis MB and Shaw RG (2001) Range shifts and adaptive responses to Quaternary climate change. *Science* 292: 673–678.
- deMenocal PB (2001) Cultural responses to climate change during the late Holocene. *Science* 292: 667–673.
- Harrison SP, Kutzbach JE, Liu Z, et al. (2003) Mid-Holocene climates of the Americas: A dynamical response to changed seasonality. *Climate Dynamics* 20: 663–688.
- Harrison SP, Yu G, and Tarasov PE (1996) Late Quaternary lake-level record from northern Eurasia. *Quaternary Research* 45: 138–159.
- Haug GH, Gunther D, Peterson LC, Sigman DM, Hughen KA, and Aeschlimann B (2003) Climate and the collapse of the Maya civilization. *Science* 299: 1731–1735.
- Herweijer C, Seager R, and Cook ER (2006) North American droughts of the mid-to-late nineteenth century: A history, simulation and implication for Mediaeval drought. *The Holocene* 16: 159–171.
- Hughes MK and Funkhouser G (1998) Extremes of moisture availability reconstructed from tree rings for recent millennia in the Great Basin of Western North America. In: Beniston M and Innes JL (eds.) *The Impacts of Climate Variability on Forests*, pp. 99–107. Berlin: Springer.
- Hunt BG and Elliott TI (2002) Mexican megadrought. *Climate Dynamics* 20: 1–12.
- Joussauze S, Taylor KE, Braconnot P, et al. (1999) Monsoon changes for 6000 years ago: Results of 18 simulations from the Paleoclimate Modeling Intercomparison Project (PMIP). *Geophysical Research Letters* 26: 859–862.
- Laird KR, Cumming BF, Wunsam S, et al. (2003) Lake sediments record large-scale shifts in moisture regimes across the northern prairies of North America during the past two millennia. *Proceedings of the National Academy of Sciences USA* 100: 2483–2488.
- Laird KR, Fritz SC, Maasch KA, and Cumming BF (1996) Greater drought intensity and frequency before AD 1200 in the Northern Great Plains, USA. *Nature* 384: 552–554.
- Lawson MP and Stockton CW (1981) Desert myth and climatic reality. *Annals of the Association of American Geographers* 71: 527–535.
- Mason IM, Guzikowska MAJ, Rapley CG, and Street-Perrott FA (1994) The response of lake levels and areas to climate change. *Climatic Change* 27: 161–197.
- Meko DM, Stockton CW, and Boggess WR (1995) The tree-ring record of severe sustained drought. *Water Resources Bulletin* 31: 789–801.
- Mock CJ (1996) Climatic controls and spatial variations of precipitation in the western United States. *Journal of Climate* 9: 1111–1125.
- Mock CJ and Brunelle-Daines AR (1999) A modern analogue of western United States summer palaeoclimate at 6000 years before present. *The Holocene* 9: 541–545.
- Moser KA, MacDonald GM, and Smol JP (1996) Application of freshwater diatoms to geographical research. *Progress in Physical Geography* 20: 21–52.
- Muhs DR and Zarate M (2001) Late Quaternary Eolian records of the Americas and their paleoclimatic significance. In: Markgraf V (ed.) *Interhemispheric Climate Linkages*, pp. 183–216. San Diego: Academic Press.
- Overpeck JT and Webb RS (2000) Nonglacial rapid climate events: Past and future. *Proceedings of the National Academy of Sciences* 97: 1335–1338.
- Riebsame WE (1991) Sustainability of the Great Plains in an uncertain climate. *Great Plains Research* 1: 133–151.
- Schubert SD, Suarez MJ, Pegion PJ, Koster RD, and Bacmeister JT (2004) On the cause of the 1930s Dust Bowl. *Science* 303: 1855–1859.
- Schweingruber FH (1987) *Tree Rings: Basics and Applications of Dendrochronology*. Dordrecht: D. Reidel Publishing Company.
- Seager R, Kushnir Y, Herweijer C, Naik N, and Velez J (2005) Modeling of tropical forcing of persistent droughts and pluvials over western North America: 1856–2000. *Journal of Climate* 18: 4068–4091.
- Shuman B, Webb T III, Bartlein PJ, and Williams JW (2002) The anatomy of a climatic oscillation: Vegetation change in Eastern North America during the Younger Dryas chronozone. *Quaternary Science Reviews* 21: 1777–1791.
- Stahle DW, Cleaveland MK, Blanton DB, Therrell MD, and Gay DA (1998) The Lost Colony and Jamestown droughts. *Science* 280: 564–567.
- Stahle DW, Cook ER, Cleaveland MK, et al. (2000) Tree-ring data document 16th century megadrought over North America. *Eos, Transactions, American Geophysical Union* 81: 121–125.

- Stockton CW (1990) Climatic variability on the scale of decades to centuries. *Climatic Change* 16: 173–183.
- Street-Perrott FA and Harrison SP (1985) Lake levels and climate reconstruction. In: Hecht A (ed.) *Paleoclimate Analysis and Modeling*, pp. 291–340. New York: Wiley.
- Therrell MD, Stahle DW, and Acuna Soto R (2004) Aztec drought and the Curse of One Rabbit. *Bulletin of the American Meteorological Society* 85: 1263–1272.
- Touchan R, Meko DM, Funkhouser G, et al. (2003) Preliminary reconstructions of spring precipitation in southwestern Turkey from tree-ring width. *International Journal of Climatology* 23: 157–171.
- Wolfe SA, Ollerheat J, and Lian OB (2002) Holocene eolian activity in south-central Saskatchewan and the southern Canadian Prairies. *Géographie Physique et Quaternaire* 56: 25–37.
- Woodhouse CA, Lukas JJ, and Brown PM (2002) Mid-19th century Colorado drought. *Bulletin of the American Meteorological Society* 83: 1485–1493.

Relevant Website

<http://www.ncdc.noaa.gov> – NOAA Paleoclimatology website.

Sub-Milankovitch (DO/Heinrich) Events

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The Quaternary has seen a succession of glacial–interglacial cycles lasting from 20 to more than 100 ka driven primarily by Milankovitch forcing (oscillations of the Earth's orbit around the Sun). However, more rapid, large-amplitude climatic changes are now recognized in all high temporal resolution paleoclimate records. The terms Dansgaard–Oeschger (DO) cycles and Heinrich events (HEs) designate specific events linked to climatic changes, which occurred during glacial periods on timescales of decades to centuries. DO cycles are designated in honor of Willy Dansgaard and Hans Oeschger, renowned geochemists and pioneers of ice core studies (Dansgaard et al., 1993). They described the succession of interstadials (warm episodes) and stadials (cold episodes) recorded as oscillations in the oxygen isotopic ratios in Greenland ice cores (Johnsen et al., 1997; North Greenland Ice Core Project, 2004; Figure 1).

Twenty-five such 'events' are described for the period from 120 to 10 ka BP. Each DO corresponds to (1) a relatively gradual cooling of $\sim 5\text{--}10\text{ }^{\circ}\text{C}$ (at the altitude of the Greenland ice sheet) that lasted $\sim 600\text{--}2000$ years; (2) a subsequent, more rapid decline of $\sim 5\text{--}10\text{ }^{\circ}\text{C}$ into peak stadial conditions, lasting another 300–700 years; and (3) a very abrupt return ($\sim 3\text{--}5\text{ }^{\circ}\text{C}$ warming per century) to interstadial conditions. HEs are named after their discoverer Harmut Heinrich, who in 1988 observed a succession of sand layers in glacial sections of sediment cores taken from the top of a deep seamount from the northeast Atlantic, which he rightly attributed to a succession of periods of large-scale iceberg melting and sedimentation of ice-rafted detritus (IRD). Wally Broecker, who coined the terms in 1992, was the first to recognize the implications of Heinrich's discovery, including the probable climatic links between large iceberg surges, meltwater input to the North Atlantic, interruption of the thermohaline circulation, and high-latitude cooling. Gerard Bond and coauthors (1992, 1993) published the first detailed descriptions of these events (Figure 2) and suggested the climatic links between DO cycles and HEs. They have shown in particular that for four of the five HEs recorded in North Atlantic sediments during the last glacial period between 70 and 20 ka BP, peak IRD deposition corresponds to the coldest phase of a DO stadial and is directly followed by a pronounced warm interstadial. The last typical Heinrich event, HE1, occurred during the early phase of the last deglaciation (or 'termination'); this is a period for which precise correlation of marine IRD records with the Greenland ice core records is difficult due to the occurrence of several large-amplitude climatic oscillations within a few millennia (including the Oldest and Older Dryas). However, as for other HEs, HE1 was followed by a major warming, which in this case constituted most of the deglacial temperature change in the North Atlantic region during the last termination. The Younger

Dryas, which interrupted the last deglaciation between 12.5 and 11 ka BP (calendar years), is now generally considered as the last Heinrich event (HE0). Since the early 1990s, it has become increasingly apparent that the impact of these climate events has not been restricted to the North Atlantic region, and that evidence for similar perturbations exists from throughout the world – in the surface and deep ocean, on the continents (low and high latitudes), and in the polar ice sheets of both hemispheres (Sarnthein et al., 2002).

The study of millennial climate events and their global teleconnections is complicated by the difficulty involved in placing them in a sufficiently precise chronology. Uncertainty in the ^{14}C chronology readily exceeds a few hundred years for ages older than ~ 15 ka BP. Calibration of radiocarbon ages is complicated by the nonlinear relationship between calendar age and ^{14}C dates. Dating of marine records involves the added uncertainty of poorly constrained 'reservoir age' variability because marine carbon is not exchanged sufficiently rapidly with the atmosphere to result in 'zero-age' surface water. None of the classical stratigraphic tools are applicable for distant correlations because most of them depend on climate-related proxies (e.g., stable oxygen isotopes in ice and Foraminifera or the distribution of stenotopic species) that will not necessarily have changed in synchrony across different climatic zones or (in the case of global geomagnetic or glacio-eustatic methods) are simply not sufficiently precise relative to the duration of the events being studied. Nevertheless, correlations between proxies at the regional scale are sufficiently precise and allow the detailed study of local interactions during rapid climate changes between ice sheets, ocean, and continental surfaces.

The following sections briefly present some of the proposed origins and global consequences of past submillennial-scale climate change, as inferred from their impacts on various paleoclimate proxies. In this presentation, which is based on the state of knowledge as of mid-2005, and to a certain extent on our own scientific bias, we simplify the current understanding of the climate system a great deal. Therefore, we cannot pretend that this summary will represent a consensus, which in many cases does not yet exist.

High-Latitude Signals

The various Greenland ice core records are strikingly similar, exhibiting the same series of characteristic DO climate events (Figure 1). Each DO cycle represented in the Greenland temperature record has been linked with a large-scale reorganization of the atmospheric circulation around the Northern ice sheets. The approximately 'sawtooth' waveform described by

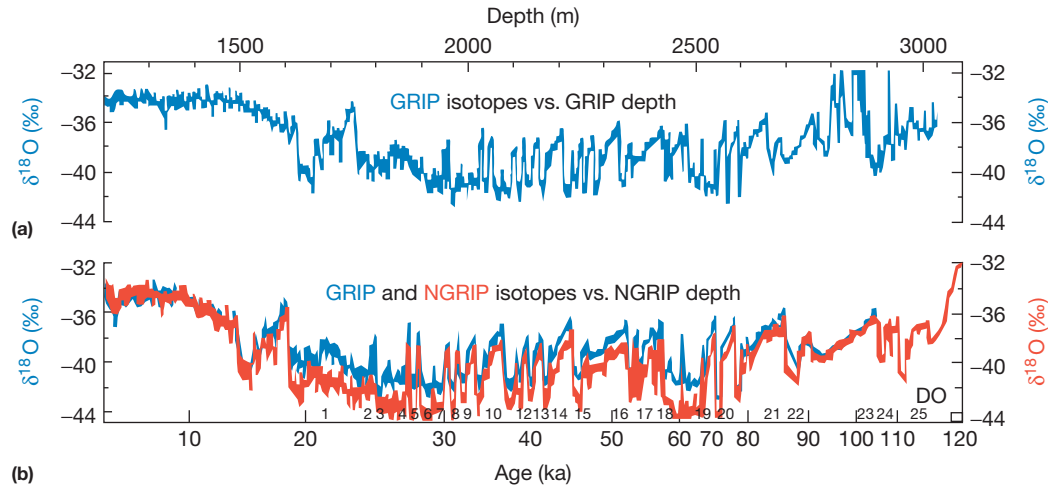


Figure 1 Ice $\delta^{18}\text{O}$ records along central Greenland drillings: the GRIP and North Grip (NGRIP) records. Redrawn from North Greenland Ice Core Project (2004) High-resolution record of Northern Hemisphere climate extending into the last interglacial period. *Nature* 431: 147–151; with the original GRIP timescale of Johnsen et al. (1997) and numbered DO.

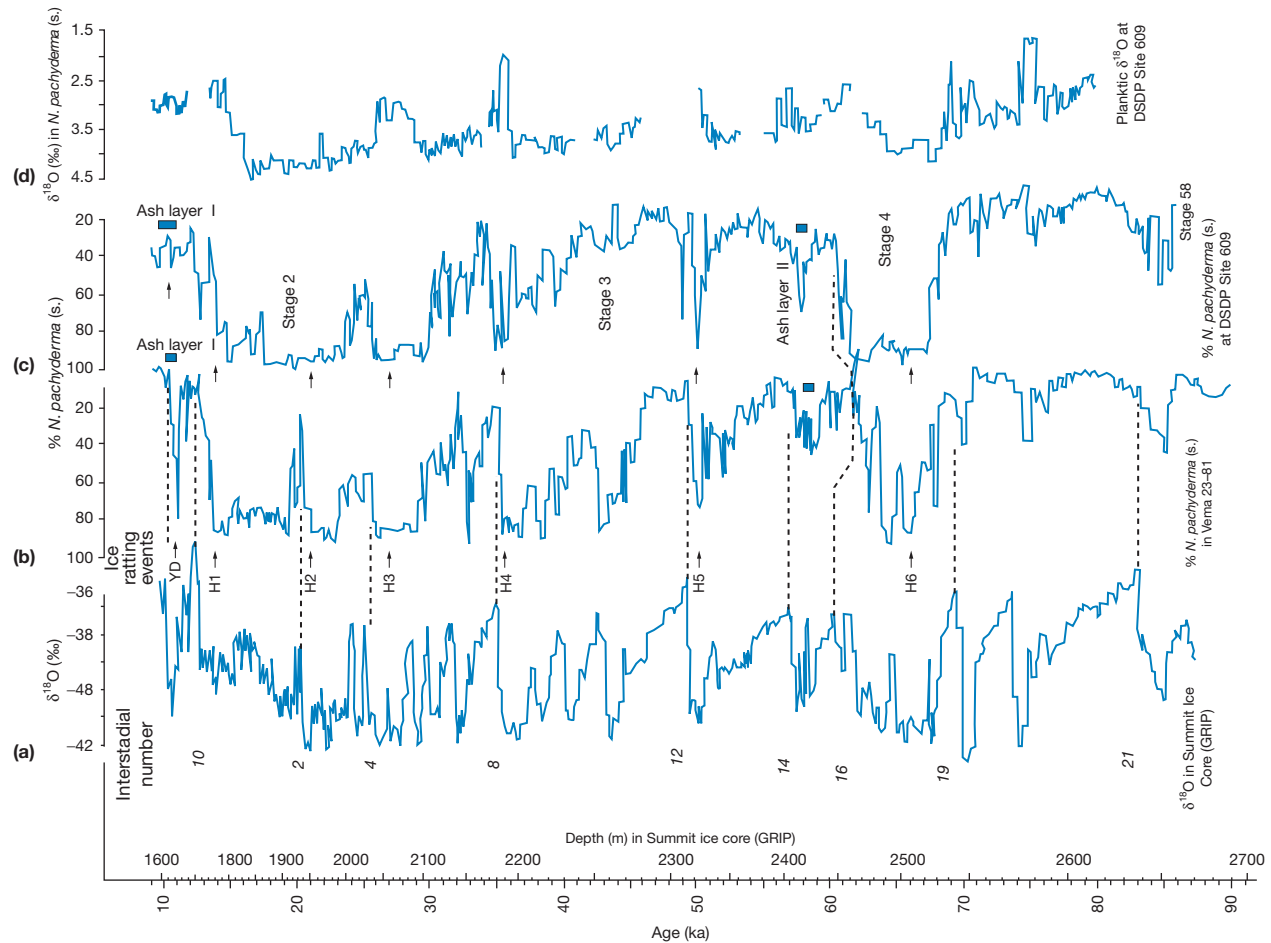


Figure 2 Records of the rapid climatic variability around the northern Atlantic Ocean. (a) The GRIP ice $\delta^{18}\text{O}$ record with numbered interstade cycles. (b) Core V23-81 (54°15' N, 16°50' W) parallel changes in the relative amount of a polar planktic Foraminifera: *Neogloboquadrina pachyderma* (note inverted scale), an indicator of the southern expansion of the polar waters. (c) Same for DSDP site 609 (49°53' N, 24°14' W). (d) *N. pachyderma* $\delta^{18}\text{O}$ (note inverted scale). The negative $\delta^{18}\text{O}$ anomaly linked with cold events tracks the surging iceberg meltwater during Heinrich events H1–H6. Redrawn from Bond G, Broecker W, Johnsen S, et al. (1993) Correlations between climate records from North Atlantic sediments and Greenland ice. *Nature* 365: 143–147.

the Greenland ice core isotopic records is also present in Atlantic records of sea surface temperature at $\sim 45^\circ \text{N}$ (Figures 1, 2, and 6), suggesting that DO cycles and HEs were linked to latitudinal oscillations of the polar front and the associated wind-driven circulation.

However, ocean sediment records may present different signatures depending on their location and proximity to the ice sheets. Much knowledge has been acquired by the study of sediment cores sampled from throughout the world under the IMAGES program (International Marine Past Global Changes Study) with the giant corer aboard the research vessel Marion Dufresne, and with the research vessel Hudson in the Labrador Sea and the Ocean Drilling Program in the North Atlantic.

Exceptional records obtained in the Nordic seas show that a large fraction of the Greenland and Norwegian seas were free of sea ice, at least in summer, during warm interstadials. Sea ice probably extended much further during stadials. Magnetite particles eroded by weathering of old volcanic margins were actively transported during each interstadial to the northern Atlantic, thus indicating significant deep-water production and transport at these times. The magnetic susceptibility records in

key regions of the North Atlantic thus exhibit a notable similarity to the Greenland ice isotopic record (Figure 5).

Parallel changes in surface- and deep-water oxygen isotopic ratios at these high northern latitudes also indicate pulses of low $\delta^{18}\text{O}$ deep-water export, particularly during stadial events. This suggests periods of deep convection along the glaciomarine margins by brine formation, as occurs today along the Ronne-Filchner ice shelves in the Weddell Sea.

Further south in the North Atlantic, between 55° and 40°N , the sedimentary signal is dominated by a succession of major ice-rafting events (HEs), occurring approximately every 5–10 ka (Figures 2, 3, and 6). This zone is called the Ruddiman Belt after Ruddiman (1997) (Figure 6) and constitutes the main melting zone for icebergs on their eastward journey along the polar front. The IRD material associated with the HEs has been clearly identified as deriving in large part from the Laurentide Archæan shield of Canada, which is from 3 to more than 4 billion years old, and with a significant component of metamorphosed carbonate and iron-coated minerals (Bond et al., 1992, 1993).

The material included in HE IRD is present in many sizes, from 1- μm -diameter clay to more than 1-mm-diameter sand

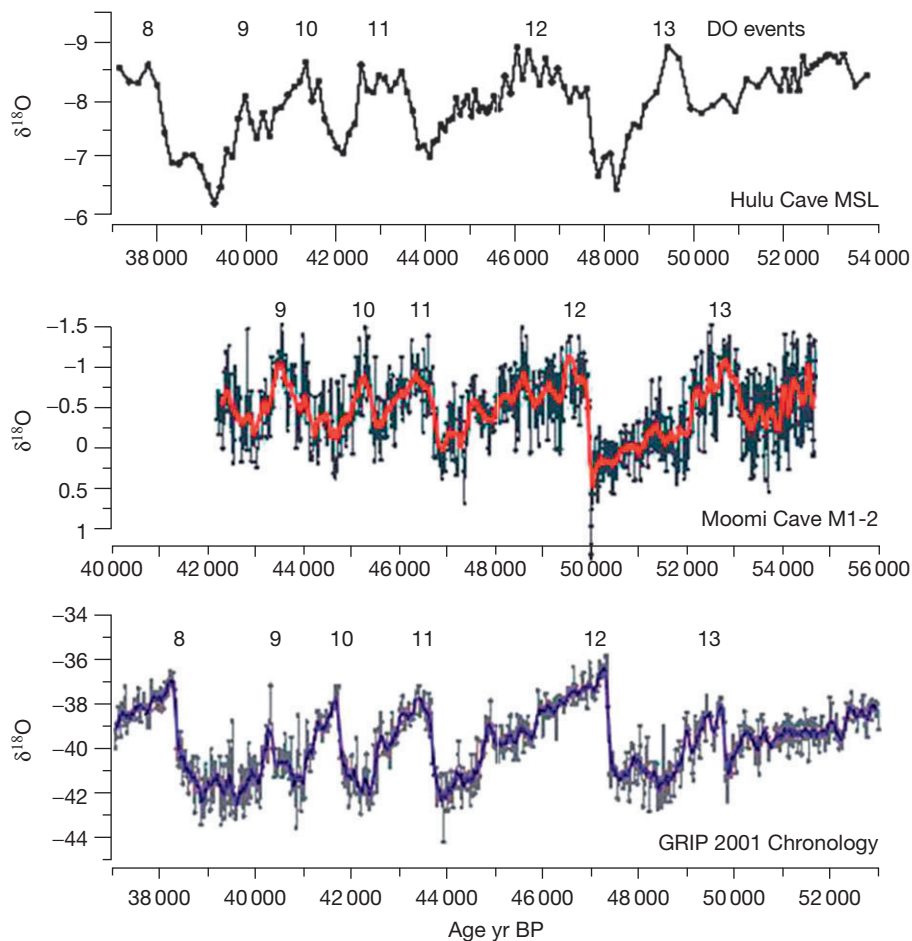


Figure 3 Comparison of the oxygen isotope ratios of stalagmite M1–2 from Socotra Island (Indian Ocean) with oxygen isotopes from the GRIP ice core (Johnsen et al., 2001) and the $\delta^{18}\text{O}$ record of a stalagmite from Hulu Cave in central China (Wang et al., 2001). The timescales are independent and shifted to give the best fit for DO events. The oxygen isotopic scales for the stalagmite records are reversed. The locations of DO events 9–13 as identified in each record are also shown. From Burns SJ, Fleitmann D, Matter A, Kramers J, and Al-Subbary AA (2003) Indian Ocean climate and an absolute chronology over Dansgaard/Oeschger events 9 to 13. *Science* 301: 1365–1367.

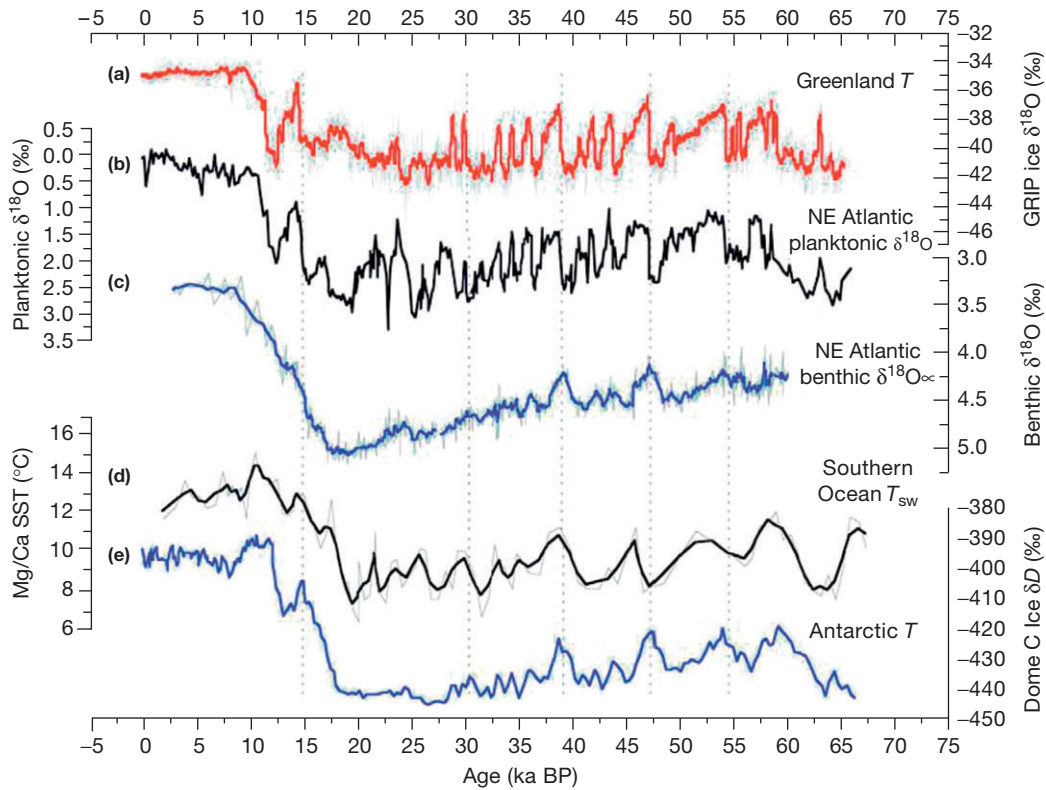


Figure 4 Submillennial asynchrony of Northern and Southern Hemisphere temperature change and of the surface- and deep-water T - $\delta^{18}\text{O}$ signature in the northeast Atlantic, as illustrated by (a) Greenland temperatures (data from Meese et al., 1997); (b) North Atlantic subpolar surface temperatures, recorded by the $\delta^{18}\text{O}$ of planktonic Foraminifera (spliced record from Skinner et al., 2003 and N. J. Shackleton, personal communication); (c) North Atlantic benthic $\delta^{18}\text{O}$ (spliced record from Skinner et al., 2003 and N. J. Shackleton, personal communication); (d) Southern Ocean surface temperatures, recorded by the Mg/Ca ratios of planktonic Foraminifera (Pahnke et al., 2003); and Antarctic temperatures (EPICA Community Members, 2004).

and pebbles. Some HEs, such as HE6 and HE3, are ‘anomalous’ in their characteristics because they contain more clastic material from the Arctic and Western Europe. These events appear to be associated in particular with periods of increased invasion of the ice shelves over the upper continental margins.

In the North Atlantic, each HE is associated with pronounced surface-water cooling, with a parallel $\delta^{18}\text{O}$ anomaly at high latitudes. The observed 1–2‰ drops in foraminiferal $\delta^{18}\text{O}$ would correspond to ~ 0.1 – 0.5 Sv ($1 \text{ Sv} = 10^6 \text{ m}^3 \text{ s}^{-1}$) of meltwater invading the North Atlantic for ~ 500 years. Corresponding sea-level changes are estimated at ~ 5 – 15 m, which is ~ 5 – 10% of the full glacial–interglacial amplitude. Cold events associated with HEs are also observed on the continents at similarly high latitudes. The Oldest, Older, and Younger Dryas climate events have been described in terrestrial pollen series for some time. Records of cold stadials corresponding to HEs during the glacial period have also been observed in European loess sections (indicated by changes in mollusk shells or the $\delta^{13}\text{C}$ of residual plant material), in pollen records from sediment cores on ocean margins, and in the isotopic ratio ($\delta^{18}\text{O}$ or $\delta^{13}\text{C}$) of cave stalagmites (Figure 3). These latter records have the advantage of being precisely dated using $^{230}\text{Th}/^{234}\text{U}$ dating with an error of ~ 500 years at 50 ka BP.

The impacts of DO and HE events have not been restricted to the high latitudes in the North Atlantic region.

Paleotemperature reconstructions from the North Pacific appear to indicate an antiphase relationship with respect to Greenland temperature (Figure 4). At mid-latitudes in the Pacific, highly detailed records from the Santa Barbara Basin indicate a very close coupling between local temperatures and Greenland temperatures. Benthic fauna and carbon isotope excursions in this basin also suggest a coupling between Greenland interstadials and decreased local bottom-water oxygenation, possibly accompanied by pulses of methane clathrate dissociation. In the Southern Ocean, millennial temperature changes have been observed between $\sim 40^\circ$ and 55°S (Figure 3), with cooling events apparently associated with colder conditions over Antarctica and with increased IRD supply from the Antarctic ice sheet. Preliminary chronologies suggest that these Southern Ocean IRD events occurred out of phase with the HEs observed in the North Atlantic. Hence, it appears that the DO cycles and HEs that have been recorded in the North Atlantic region have been associated with ‘southern counterparts,’ such that for each HE (at least) a corresponding ‘A event’ may be identified in Antarctica. These Antarctic A events, in contrast to their DO counterparts, are characterized by a relatively gradual warming (of $\sim 10^\circ \text{C}$ over a few millennia) followed by a slightly more prolonged cooling of similar amplitude. The specific character of Antarctic temperature change, and its relation to temperature variability recorded in Greenland, is discussed elsewhere in this encyclopedia.

Low-Latitude Signals

The best climate information obtained from low-latitude records is linked to changes in temperature, precipitation, and monsoon activity. Rapid temperature shifts, synchronous with Greenland variability across the last deglaciation, have been identified in the Cariaco Basin and the South China Sea, whereas temperature changes that are apparently synchronous with Antarctica have also been reconstructed in the tropical Atlantic just outside of the Cariaco Basin and in the Makassar Strait of the western tropical Pacific. Other apparently 'Antarctic-type' temperature changes have also been reconstructed for the last deglaciation (using the alkenone temperature proxy) in the Angola Basin of the southeast tropical Atlantic, although these have been interpreted as significantly lagging behind the changes in Antarctica and thus appear to be more synchronous with Greenland change (Figure 5).

The direct response of the Asian and African monsoons to the changing seasonality of insolation forcing has been well known for nearly two decades. The response of the monsoon

system to Northern Hemisphere rapid climate change is equally well marked, with dramatic reduction of the Indian summer monsoon (and its associated upwelling system in the Arabian Sea and low-salinity waters around Indonesia) during each DO stadial. Asian summer and winter monsoons were also affected (Figure 6), with significant summer monsoon increase during the main DO interstadials. The main DO stadials have seen a significant decrease in the main trade wind circulation and associated Hadley cell strength over the western Pacific, in a way similar to the conditions associated with El Niño events today. Some of the temperature shifts observed in the intertropical areas may be associated with these changes.

Millennial Changes in Deep- and Intermediate-Ocean Hydrography

The ocean's large-scale circulation is a key part of the Earth's climate system that is directly coupled with atmospheric circulation and composition, as well as the transport of heat and

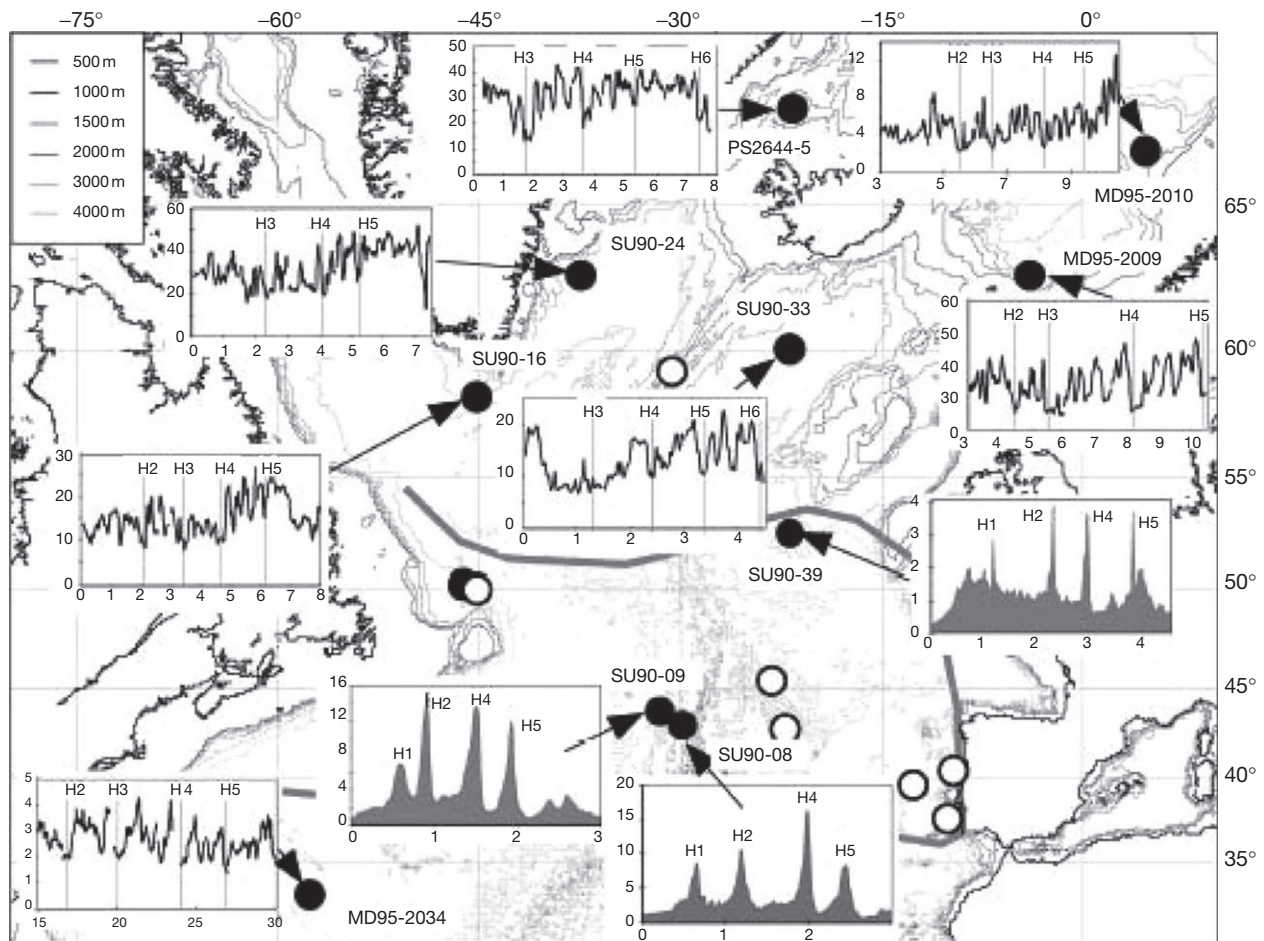


Figure 5 Low-field magnetic susceptibility records (in 10^{-4} SI) reported versus depth (m) for a few representative cores from the northern Atlantic Ocean. The gray curves are typical for the Ruddiman Belt (see Figure 6), with peaks of high susceptibility during the Heinrich events, whereas the others show the short-term variations in which Heinrich layers (vertical lines) coincide with minima. Signals from core MD95-2009 and SU90-33, at $\sim 60^\circ$ N, present particularly strong similarities with the Greenland ice isotopic records (Figures 1, 2, and 6). From Kissel C (2005) Magnetic signature of rapid climate variations in glacial North Atlantic, a review. *Comptes Rendus Geoscience* 337: 908–918.

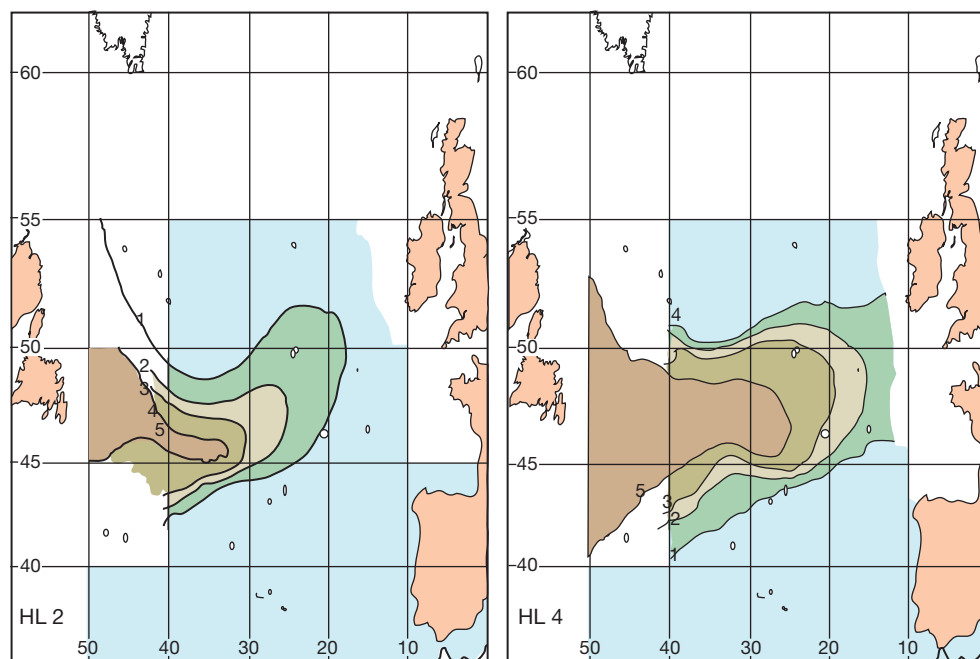


Figure 6 Distribution of high-susceptibility sediment associated with ice surge events HE1 and HE4. The colored area is at approximately the same location as the belt where sediment contains a large fraction of ice-rafted detritus described by Ruddiman (1977). Redrawn from Grousset FE, Labeyrie L, Sinko JA, et al. (1993) Patterns of ice-rafted detritus in the glacial North Atlantic (40–55° N). *Paleoceanography* 8(2): 175–211.

moisture across the globe. A large body of paleoceanographic evidence supports the hypothesis that the geometry of the world's surface-, intermediate-, and deep-water masses was significantly altered during past glacial climate regimes, and that this situation has also varied on millennial timescales (Labeyrie et al., 2002).

Some evidence has suggested that the variability of the Atlantic deep- and intermediate-water mass geometry during HEs and DO cycles has included an inverse relationship between the ventilation of glacial North Atlantic deep water (hereafter GNDW) and glacial North Atlantic intermediate water (GNAIW), with stadials coinciding instead with a reduced northern deep-water export and an increased northern intermediate water export. However, what is apparent from the paleoceanographic record is that both intermediate- and deep-water masses exported from the North Atlantic have been affected at least in part in antiphase with the export of their southern-sourced 'counterparts', the glacial analogues of Antarctic intermediate water (hereafter GAAIW) and Antarctic bottom water (hereafter GSDW). In particular, intense stadial events in the North Atlantic region appear to have been accompanied by an increased incursion of GSDW (at the expense of GNDW) into the deep Atlantic.

Millennial changes in the supply of *Mediterranean outflow water* (MOW) to the intermediate Atlantic are not well constrained, although it appears that a more restricted flow of MOW out of the Mediterranean and northward onto the Iberian Margin (due to lowered sea level and a shallower Mediterranean sill depth) was nonetheless more intensified during past glacial and, perhaps, stadial climate regimes, because it was colder, more saline, and hence more dense. In the Pacific, millennial intermediate- and deep-water variability was

closely connected with the changes experienced in the Atlantic. Thus, along with the apparently increased AAIW export in the intermediate Southern Ocean during North Atlantic DO stadials and HEs (Pahnke et al., 2003), there was a concurrent decrease, within the uncertainties of the age models, in the influence of NADW versus AABW in the deep Southern Ocean (which should also reflect the character of deep water entering the deep Pacific south of New Zealand), as well as an increase (at least across the last deglaciation) in the ventilation of the intermediate North Pacific. Thus, evidence suggests that the ventilation of the intermediate Pacific by both southern and northern sources increased during North Atlantic stadials, in 'antiphase' with the ventilation of the deep Atlantic by NADW and the intermediate Atlantic by GNAIW.

North–South Linkage

Any viable theory that proposes to explain past millennial climate change must be consistent with the geological record and must be capable of accounting for the observed patterns of climate change recorded in the Greenland and Antarctic ice cores and in any other records that can be securely linked to these. The relation between Greenland and Antarctic temperature change on millennial timescales is perhaps best described as 'asynchronous'; indeed, the fact that the two climate signals are distinct (rather than simply offset or in antiphase) is visually obvious but also statistically demonstrable.

Despite the stratigraphic difficulties involved in linking distal paleoclimate records with submillennial precision, it appears that stadial–interstadial temperature changes in Greenland and Antarctica have remained coupled with parallel

surface-water temperature changes recorded in the North Atlantic and Southern Ocean, respectively. The relation of Greenland and Antarctic temperature change to low-latitude climate variability is more ambiguous, although it appears that Greenland stadials (and hence expansions of the North Atlantic polar regime) have been coupled with synchronous increases in the dominance of winter precipitation in the Asian and Indian monsoons and with reduced precipitation in the Northern Hemisphere tropical Atlantic.

Testing the various mechanisms that are proposed to explain the Greenland–Antarctic asynchrony relies in many ways on assigning ‘Arctic’ or ‘Antarctic’ affiliations to records from throughout the globe. However, biases due to proxy seasonality and/or dating errors, as well as offsets between radiometric and ice core chronologies, remain significant hurdles to overcome in making precise ice core correlations for this purpose. One theory that proposes to explain north–south climate connections, the ‘bipolar seesaw’ in its most simplified form (Stocker and Johnsen, 2003), predicts that Atlantic tropical temperatures should either vary in antiphase with respect to Greenland temperatures or be correlated in some way with Antarctic temperatures. This prediction has been inadequately tested, although it is apparent that the precipitation sources of both Greenland and Antarctica (presumably in part of low-latitude origin) warmed significantly during Greenland stadials. It is very likely, if not certain, that the global connections associated with the Greenland–Antarctic asynchrony have been mediated by atmospheric processes and, therefore, that a better understanding of the intensities and positions of the main atmospheric frontal systems is required if we are to accurately interpret the available paleoceanographic evidence.

One of the most intriguing observations that affects the role of the deep-ocean circulation in millennial north–south connections is that the atmospheric coupling between Greenland and northeast Atlantic temperatures appears to have run parallel with an unexplained coupling between Antarctic temperature and benthic foraminiferal $\delta^{18}\text{O}$ from the deep northeast Atlantic (Shackleton et al., 2000; Figure 6). There is good evidence that changes in glacio-eustatic sea level may have played a role in the apparent Antarctic–deep Atlantic link. However, estimates of ice volume change can only partially explain the amplitude of the northeast Atlantic benthic $\delta^{18}\text{O}$ variability, and there is independent evidence that purely local hydrographic variability (i.e., local deep water $\delta^{18}\text{O}$ and temperature change) also contributed to the benthic isotopic excursions recorded in the deep Atlantic. Similar millennial benthic $\delta^{18}\text{O}$ changes to those observed in the deep Northeast Atlantic have been recorded in the intermediate Southern Ocean. Although this may indicate the existence of a strong millennial, global (glacio-eustatic) component in the marine $\delta^{18}\text{O}$ record (which was linked in some way to Antarctic temperature change), such an interpretation would need to be reconciled with parallel $\delta^{13}\text{C}$ evidence for concurrent pulses of (cold, fresh) AAIW export, which will also have left their mark on the Southern Ocean benthic $\delta^{18}\text{O}$ record. In addition, other records from the abyssal Southern Ocean show no millennial benthic $\delta^{18}\text{O}$ variability (hence no evidence of glacio-eustatic variability), although they do include clear indications of changes in local deep-water hydrography. The resolution of these intriguing questions is certain to set much-needed

constraints on the models that we may propose to flesh out or to supersede the bipolar seesaw concept.

Mechanisms of Millennial Climate Change: Modeling Efforts

The rapidly expanding suite of high-resolution paleoclimate records that exhibit millennial-scale variability has provided an enormous stimulus to numerical modeling efforts during the past decade. Much of this modeling has focused on the interaction of the ice sheets and the atmosphere–ocean circulation system, including the influence of high-latitude freshwater discharge on North Atlantic stratification, deep-water formation, and the associated meridional heat transport. Evidence from the geological record has been crucial in directing this modeling focus. Despite many advances, few model simulations have accurately reproduced the characteristic structure and timing of the ‘type events’ that are recorded in the Greenland and North Atlantic temperature records, for example. One difficulty involved in doing so derives from the fact that the characteristic recurrence time of DO events ($\sim 1\text{--}2\text{ ka}$) or HEs ($\sim 5\text{--}7\text{ ka}$) is much longer than the characteristic timescale of the ocean–atmosphere system ($\sim 0.1\text{--}1\text{ ka}$ at most). This suggests that millennial climate changes similar to those observed in the geological archives are not likely to derive from internally sustained oscillation in the ocean–atmosphere system alone and are more likely to have involved additional climate parameters with longer time constants, such as the cryosphere. Furthermore, based on the abrupt character of many climate transitions recorded in the geological archive, it is probable that past millennial climate change has involved the repeated crossing of stability thresholds that emerge from the interaction of the ocean, atmosphere, and cryosphere ‘subsystems.’

The role of the cryosphere in past millennial climate change is directly supported by geological evidence of the occurrence of massive iceberg discharges into the North Atlantic region during each DO stadial. Hence, a minimal model of millennial-scale climate variability should couple together the ocean, atmosphere, and ice sheets. Indeed, it is not clear how the ocean circulation could be correctly modeled without considering the atmospheric circulation, which in turn will respond sensitively to the altitude and extent of ice sheets and sea ice. Generally, there are two approaches to modeling the combined ice–ocean–atmosphere effects on millennial climate change. The first is to prescribe or parameterize the presumed impact of ice sheet variability (generally limited to prescribing the discharge of freshwater into the North Atlantic) and to determine the impact of this on the ocean–atmosphere system. The second approach involves directly modeling ice sheet dynamics and attempting to incorporate this into a coupled ocean–atmosphere model. Using the first approach, it has been possible to simulate entire sequences of millennial climate changes that bear a resemblance to the geological record. This has illustrated the plausibility of ice sheet discharges into the North Atlantic (or indeed the Southern Ocean) as triggers for climate change via their impact on ocean overturning and meridional heat transport. However, such models leave open the question of what physical mechanisms and specific climatic feedbacks ultimately drive ice sheet dynamics.

Direct modeling of ice sheet dynamics has independently confirmed the plausibility of recurrent ice sheet destabilization in principle. However, although the timing of modeled ice sheet discharges (~ 1 every 4500 years) is approximately consistent with the geological evidence, the amount of freshwater associated with the collapse of the Hudson Bay ice stream, for example, is small (order of magnitude ~ 0.01 Sv). Based on most (independent) ocean and ocean-atmosphere models, such a freshwater perturbation would be too small to induce a significant change in the North Atlantic density stratification and hence the Atlantic overturning circulation and regional climate.

In addition, modeled freshwater fluxes associated with individual ice sheet discharges do not agree with our best estimates of rapid sea-level change. Field reconstructions suggest millennial sea-level changes of at least ~ 10 m during the last glaciation, which would correspond to ~ 1 Sv sustained over ~ 100 years. This implies an integrated freshwater flux that is orders of magnitude larger than the regional values simulated by ice sheet models. A positive feedback mechanism that amplifies the effects of individual ice stream discharges must therefore exist, and it is possible that sea-level rise may provide such a mechanism by destabilizing grounded marine ice sheets and thus triggering multiple discharges at approximately the same time.

As discussed previously, the main climatic impact of ice sheet discharges is generally perceived in terms of their effect on the ocean overturning circulation. Perturbations of ocean circulation due to freshwater discharge have been modeled extensively with coupled climate models of varying complexity. In most cases, the fundamental behavior of the ocean circulation is drawn in terms of 'hysteresis loops' (forced oscillations between 'on' and 'off' states via different thresholds for each transition), wherein a primary difficulty lies in determining proximity of the system to a given threshold or, indeed, whether it ever crosses such thresholds. Hence, the degree of ocean circulation 'collapse' in various models is found to be sensitive to not only the amount and location of the freshwater input but also the model parameterization (including that of mixing and diffusive processes). Nevertheless, model experiments generally suggest that ~ 0.1 Sv or more of freshwater is needed to induce a significant change in North Atlantic overturning. Hence, the previous estimate of ~ 1 Sv over 100 years (based on sea-level evidence), if supplied to the North Atlantic, would be sufficient to shut down the Atlantic overturning circulation and cause surface temperature changes of up to $\sim 15^\circ\text{C}$ in the North Atlantic region.

In parallel with these effects (although only if the modeled circulation changes are large enough), models also predict a compensating warming in the Southern Hemisphere. This is consistent in principle with the so-called bipolar seesaw mentioned previously. The term bipolar seesaw has not always been used in the literature to designate the same physical process or even the same temporal relationship between Greenland and Antarctic climate change. Here, it is used to designate the simple principle of partially 'mutually exclusive' heat budgets in the north and south, which are physically connected via a mechanism that involves the Atlantic overturning circulation and the meridional heat transport of the ocean-atmosphere system. Although the bipolar seesaw

mechanism for linking millennial climate change in the Northern and Southern hemispheres remains a subject of intense debate, it is more or less generally held that the apparent north-south connection can be explained by a set of physical mechanisms that involved changes in the large-scale ocean circulation that responded to (and/or incurred) changes in atmospheric circulation, high-latitude sea ice distribution, and ice sheet stability. A variety of more specifically defined models have been built on this central, and rather theoretically sparse, ocean-atmosphere-cryosphere bipolar seesaw notion. Testing these various models with better and more widespread geological data is clearly a priority of current research.

Much has been learned from the modeling efforts of recent years, and yet much remains to be resolved, including the specific effects of mixing and diffusion processes at the ocean boundaries (i.e., due to wind stress) and in the ocean interior. It seems likely that a resolution of the bipolar seesaw controversy in particular (i.e., testing its existence in precisely defined terms) will provide a sound basis for modeling millennial climate change more generally.

See also: Carbonate Stable Isotopes: Speleothems. Glacial Climates: Thermohaline Circulation. Glacial Landforms, Ice Sheets: Growth and Decay. Glacial Landforms, Sediments: Glaciomarine Sediments and Ice-Rafted Debris. Ice Core Methods: Chronologies. Ice Core Records: Antarctic Stable Isotopes; Correlations Between Greenland and Antarctica; Greenland Stable Isotopes. Ice Cores: History of Research, Greenland and Antarctica. Loess Records: Europe. Paleooceanography, Records: Late Pleistocene North Atlantic; Late Pleistocene North Pacific; Late Pleistocene South Atlantic. Paleoclimate: Timescales of Climate Change. Paleoclimate Modeling: Last Glacial Maximum GCMs. Paleoclimate Reconstruction: Younger Dryas Oscillation, Global Evidence. Radiocarbon Dating: Calibration of the ^{14}C Record.

References

- Bond G, Heinrich H, Broecker W, et al. (1992) Evidence for massive discharges of icebergs into the glacial north Atlantic. *Nature* 360: 245–249.
- Bond G, Broecker W, Johnsen S, et al. (1993) Correlations between climate records from North Atlantic sediments and Greenland ice. *Nature* 365: 143–147.
- Dansgaard W, Johnsen SJ, Clausen HB, et al. (1993) Evidence for general instability of past climate from a 250-kyr ice-core record. *Nature* 364: 218–220.
- EPICA Community Members (2004) Eight glacial cycles from an Antarctic ice core. *Nature* 429: 623–628.
- Johnsen SJ, Clausen HB, Dansgaard W, et al. (1997) The d^{18}O record along the Greenland Ice Core Project deep ice core and the problem of possible Eemian climatic instability. *Journal of Geophysical Research* 102(C12): 26397–26410.
- Johnsen SJ, Dahl-Jensen D, Gundestrup N, et al. (2001) Oxygen isotope and palaeotemperature records from six Greenland ice-core stations: Camp Century, Dye-3, GRIP, GISP2, Renland and NorthGRIP. *Journal of Quaternary Science* 16: 299–307.
- Labeyrie L, Cole J, Alverson K, and Stocker T (2002) The history of climate dynamics in the Late Quaternary. In: Alverson RBK and Pedersen T (eds.) *Paleoclimates, Global Change and the Future*, pp. 33–62. Berlin: Springer.
- Meese DA, Gow AJ, Alley RB, et al. (1997) The Greenland Ice Sheet Project 2 depth-age scale: Methods and results. *Journal of Geophysical Research* 102(C12): 26411–26423.
- North Greenland Ice Core Project (2004) High-resolution record of Northern Hemisphere climate extending into the last interglacial period. *Nature* 431: 147–151.

- Pahnke K, Zahn R, Elderfield H, and Schulz M (2003) 340,000-year centennial-scale marine record of Southern Hemisphere climatic oscillation. *Science* 301: 948–952.
- Ruddiman WF (1977) Late Quaternary deposition of ice-rafted sand in the sub-polar North Atlantic (lat 40° to 65° N). *Geological Society of America Bulletin* 88: 1813–1827.
- Sarnthein M, Kennett JP, Allen JRM, et al. (2002) Decadal-to-millennial-scale climate variability – Chronology and mechanisms: Summary and recommendations. *Quaternary Science Reviews* 21: 1121–1128.
- Shackleton NJ, Hall MA, and Vincent E (2000) Phase relationships between millennial-scale events 64,000–24,000 years ago. *Paleoceanography* 15(6): 565–569.
- Skinner LC, Shackleton NJ, and Elderfield H (2003) Millennial-scale variability of deep-water temperature and $\delta^{18}\text{O}_{\text{dw}}$ indicating deep-water source variations in the Northeast Atlantic, 0–34 cal. ka BP. *Geochemistry, Geophysics, Geosystems* 4: 1–17.
- Stocker TF and Johnsen SJ (2003) A minimum thermodynamic model for the bipolar seesaw. *Paleoceanography* 18(4): 1087.
- Wang YJ, Cheng H, Edwards RL, et al. (2001) A high-resolution absolute-dated late Pleistocene monsoon record from Hulu Cave, China. *Science* 294: 2345–2348.

Paleotempestology

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Introduction

Paleotempestology is an emerging field of science that studies past tropical cyclone activity beyond the period of instrumental observations, typically spanning the last several centuries to six millennia. Tropical cyclones are known by different names in different regions of the world – *hurricanes* in North America, *typhoons* in the Northwest Pacific, and *cyclones* in South Asia and Australia. Here, the term *hurricanes* is sometimes used interchangeably with all other types of tropical cyclones.

In the United States, the instrumental record of tropical cyclone activity is essentially confined to the last 150 years and it is much shorter in other parts of the world. This record is too short to fully capture the occurrence of the rare but most destructive hurricanes – the ‘catastrophic’ hurricanes of category 4 and 5 intensity according to the Atlantic and Eastern Pacific Basins-based Saffir–Simpson scale. Therefore, by providing a long-term, empirical record of hurricane activity over the last 6000 years, paleotempestology has shown that it is useful for revealing the spatial and temporal variability of hurricane activity and deciphering its relationship with global climatic changes.

Two main sources of data are available for reconstructing past hurricane activity to beyond the instrumental period – geological proxy records, and historical documentary records. Therefore, they underscore two major approaches to the study of paleotempestology – geological and archival.

Geological Proxy Records

A landfalling hurricane is an important geological and ecological agent that can severely impact the landforms, ecosystems, and sedimentary and hydrological processes on the coast (Figure 1). These geophysical and ecological impacts may leave a geological record that can be deciphered by means of proxy techniques. Several geological or biological proxies have been successfully employed to reconstruct past hurricane strikes. So far, the most widely used proxy is overwash sand layers deposited in the sediments of coastal lakes, marshes, and swamps (Donnelly and Webb, 2004; Liu, 2004; Liu and Fearn, 1993). Storm-built beach ridges in Australia have also been shown to contain a record of cyclone strikes spanning the past 6000 years (Nott, 2011; Nott and Hayne, 2001; Nott et al., 2009). In addition, oxygen isotope values extracted from tree rings and stalagmites have been deciphered to contain a record of past tropical cyclone activity affecting coastal regions (Frappier et al., 2007; Miller et al., 2006; Nott et al., 2007).

Overwash Sand Layers in Coastal Back-Barrier Lakes

When a hurricane makes landfall, the strong onshore winds, especially in the forward-right quadrant of the intense low-pressure system, will generate a storm surge directed toward

the coast. Generally speaking, stronger and larger hurricanes correspond with higher storm surges. For intense hurricanes, the storm surge plus the wave runup may overwash the coastal barrier, causing sand to be eroded from the beach or dunes and deposited in the lake or marsh behind it. Stratigraphically, the overwash fan will exist in the form of a sand layer, which is thickest near the shore and thins out toward the center of the lake (Liu, 2004; Liu and Fearn, 1993, 2000a). In a coastal back-barrier lake that had been subjected to repeated overwash events in the past, the sediment stratigraphy should contain multiple sand layers that contain a stratigraphic record of intense hurricane strikes (Figure 2). These sand layers are composed of fine to coarse sand that is usually well sorted, and have abrupt contacts with the more organic, finer, and darker sediments both above and below. Their thickness may range from a few mm to more than 10 cm (Figure 3). They can be identified either visually or by means of sedimentological techniques such as loss-on-ignition analysis and grain-size analysis of core samples (Donnelly and Woodruff, 2007; Liu and Fearn, 2000a; Yu et al., 2009).

The attribution of these sand layers to overwash processes can be verified by using supplementary proxies such as marine microfossils (e.g., dinoflagellates, marine diatoms, Foraminifera) that suggest seawater intrusion (Hippensteel and Martin, 1999; Liu et al., 2008; Scott et al., 2003; Williams, 2010), opal phytolith assemblages that are derived from sand dunes (Lu and Liu, 2005), or geochemical signals that suggest beach or bay origin of the transported material (Donnelly, 2005; Woodruff et al., 2009). A chronology of past overwash events or hurricane strikes can be established by means of radiometric dating techniques such as radiocarbon (^{14}C), lead-210 (^{210}Pb), or cesium-137 (^{137}Cs) dating, which may be supplemented by using stratigraphic markers such as pollen and lead pollutants (Donnelly and Webb, 2004; Liu et al., 2008). For any particular location, the return period of hurricanes of a specific intensity category can be calculated by tallying up the number of events occurring over a given period of time (Liu, 2004).

The intensity of the paleohurricanes may be harder to infer from the proxy record than their frequencies. As a first approximation, it can be assumed that stronger hurricanes tend to produce higher storm surges and thus more extensive overwash fans. Therefore, within the same core, sand layer thickness can be used as a rough indicator of storm intensity (Figure 4). The sedimentary impact of recent hurricanes of known intensity can be used as a modern analog for calibrating the intensity estimate of paleohurricanes. For example, on the basis of sediment-stratigraphic evidence that the overwash sand layer deposited by Hurricane Frederic, a category 3 hurricane that struck Alabama in 1979, was only confined to the near-shore sediments, Liu and Fearn (1993) inferred that older sand layers that occurred in cores taken from the center of Lake Shelby (where the Frederic sand layer was absent) must have been deposited by prehistoric hurricanes of category 4 or 5 intensity.

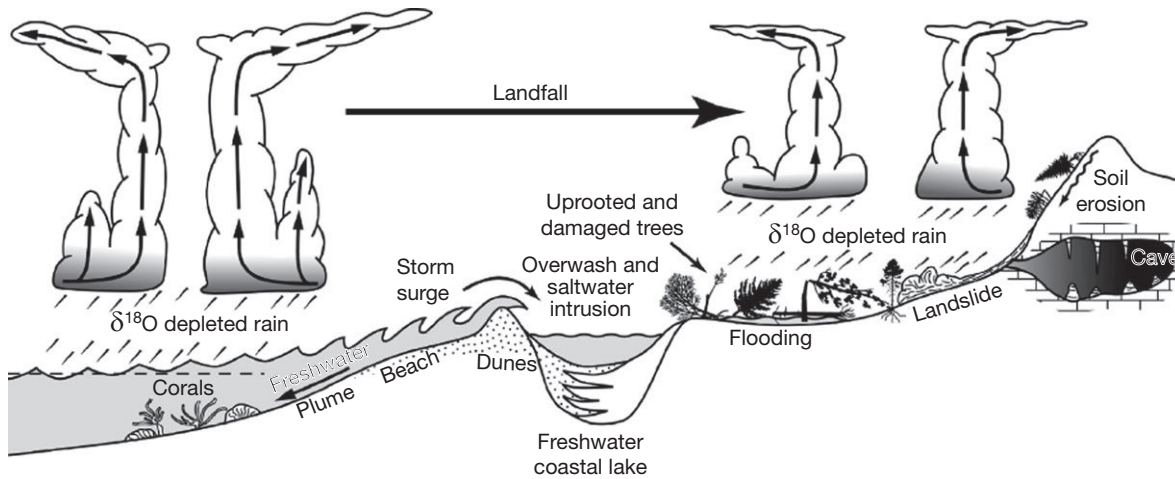


Figure 1 Environmental impacts of catastrophic hurricanes. Modified from Liu K-b (2004) *Paleotempestology: Principles, methods, and examples from Gulf Coast lake sediments*. In: Murnane RJ and Liu K-b (eds.) *Hurricanes and Typhoons: Past, Present, and Future*, pp. 13–57. New York: Columbia University Press. A landfalling hurricane may cause a storm surge that overtops beach barriers, resulting in the formation of an overwash fan and the deposition of a sand layer in the sediments of a back-barrier lake or marsh. The strong wind may cause massive damage or mortality to trees, leaving behind a paleoecological record of disturbance and succession, including the occurrence of posthurricane fires. Heavy precipitation may cause flooding in the lowlands, and soil erosion and landslide in the uplands. The $\delta^{18}\text{O}$ -depleted signal in the hurricane rains may be recorded in the cellulose of tree rings and the calcium carbonate of speleothems and coral skeletons if it is not attenuated by hydrological processes.

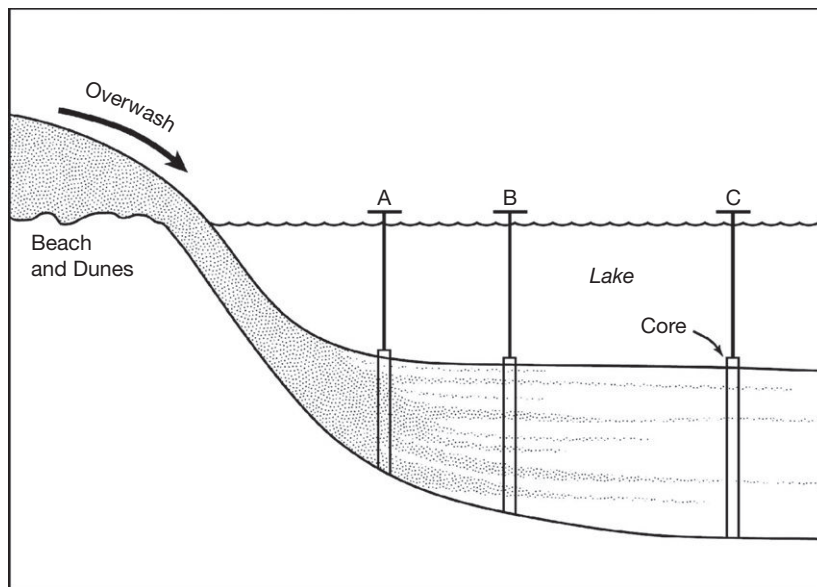


Figure 2 Hypothetical pattern of sand layer deposition in a coastal lake subjected to repeated storm overwash events in the past. The overwash sand layers are thicker near the sand barrier and become thinner toward the lake center. A core taken from site B will contain more and thicker sand layers than the one taken from site C. A core taken from site A, however, may consist of all sand without discrete layers. Reproduced from Liu K-b (2004) *Paleotempestology: Principles, methods, and examples from Gulf Coast lake sediments*. In: Murnane RJ and Liu K-b (eds.) *Hurricanes and Typhoons: Past, Present, and Future*, pp. 13–57. New York: Columbia University Press.

Coastal lakes from Alabama and northwestern Florida have yielded proxy records of catastrophic (category 4 and 5) hurricane strikes that span the last 5000 years (Liu and Fearn, 1993, 2000a).

In addition to storm layer thickness, grain size can also be used as an indicator of storm intensity based on the principle that compared with weaker storms, stronger storms generate bigger waves and higher storm surges, which in turn have

greater energy to transport coarser sediments to the coring site. On the basis of grain-size data of sand layers in cores taken from a back-barrier lake, Woodruff et al. (2008) applied a simple advective-settling model to constrain the coastal flooding intensities (which are taken to be a function of storm intensity) required to transport clastic sediments to various distances behind the barrier during overwash processes. Grain size has been used as a principal indicator for

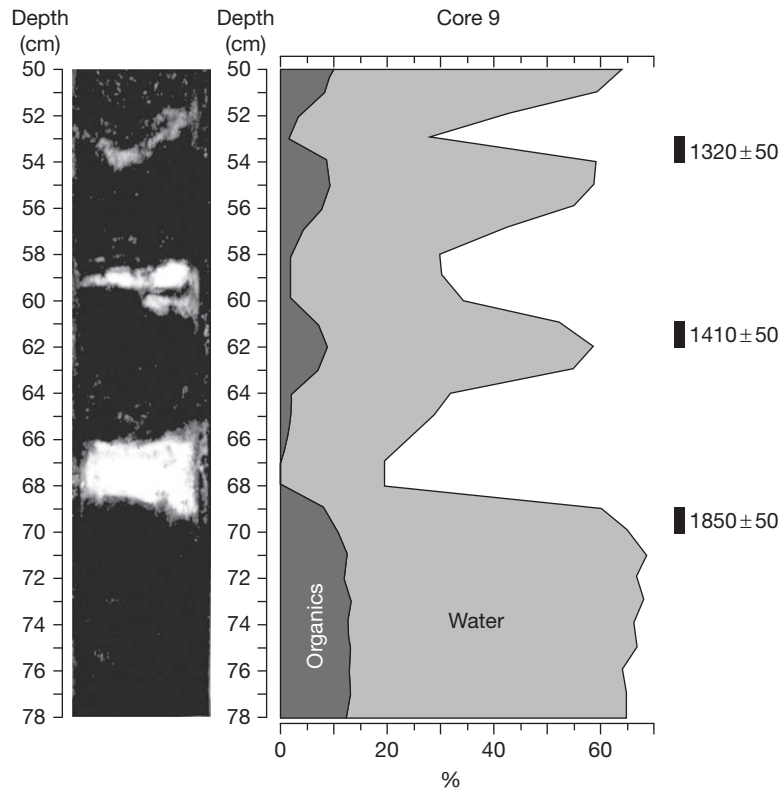


Figure 3 Photograph of three prominent sand layers in the 50–78 cm segment of core 9 from Western Lake, northwestern Florida, and the corresponding water and organic matter content curves determined by loss on ignition. Radiocarbon dates are on the right. Reproduced from Liu K-b and Fearn ML (2000) Reconstruction of prehistoric landfall frequencies of catastrophic hurricanes in northwestern Florida from lake sediment records. *Quaternary Research* 54: 238–245.

paleointensity reconstruction in studies from a back-barrier tidal lagoon in western Texas (Wallace and Anderson, 2010) and an atoll reef lagoon fringed by wave-transported coral boulders in the South China Sea (Yu et al., 2009).

As a rule of thumb, stronger storms tend to be rarer than weaker ones. Thus for any particular site, a statistical relationship exists between hurricane return periods (frequency) and hurricane return levels (intensity) in the historical record of storm activity. Elsner et al. (2008) have developed a statistical model that quantitatively links the frequency and intensity data in the historical record (AD 1851–2005) for the Lake Shelby area. By applying this statistical relationship to the proxy-derived frequency estimate of 11 events over 3500 years (or a return period of 318 years), they inferred that the sand layers in the proxy record of this lake were probably deposited by paleohurricanes with wind speeds of at least 64 m s^{-1} or category 4 or 5 intensity.

Proxy Records from Coastal Marshes and Other Coastal Sites

The same principle and research methods can be applied to coastal back-barrier wetlands (marshes and swamps) for generating proxy records of past hurricane strikes (Donnelly and Webb, 2004; Donnelly et al., 2001a,b, 2004). These back-barrier wetlands are typically peat-accumulating environments situated behind barrier beaches. As in coastal lakes, overtopping of the barrier beach by storm surge will lead to the formation of an

overwash fan behind the barrier, which will be expressed stratigraphically as a sand layer sandwiched between marsh peat (Figure 5). If the barrier beach is breached by the storm surge to form an inlet, a large flood-tidal delta may be formed across the back-barrier marsh, which may even extend into the lagoon or bay behind it (Donnelly and Webb, 2004). Well-dated proxy records of recent and prehistoric hurricane strikes have been obtained from back-barrier marshes in Rhode Island, New Jersey, and New York (Donnelly et al., 2001a,b; Scileppi and Donnelly, 2007). Similarly, millennial proxy records of paleo-hurricane strikes have been identified from back-barrier mangrove or palmetto swamps in Belize (McCloskey and Keller, 2009). In the Chenier Plain of southwest Louisiana, the storm surge deposit laid down by Hurricane Rita in a coastal woodland (ridge) and marsh (swale) environment was studied in detail with regard to its stratigraphy, sedimentology, and Foraminifera content (Williams, 2009), which may provide a modern analog for identifying prehistoric storm deposits in older sediments in similar geomorphic settings.

Estuarine or deltaic marshes formed near the mouth of major rivers and along tidal inlets can also provide proxy records of past hurricane strikes (Liu and Fearn, 2000b; Scott et al., 2003). A beach barrier may or may not be present, but these marshes are inundated by the storm surge during an intense hurricane strike. Stormy waves and the storm surge may stir up the silt and mud from the bottom of the adjacent bays, and a layer of lagoonal mud will be deposited on the

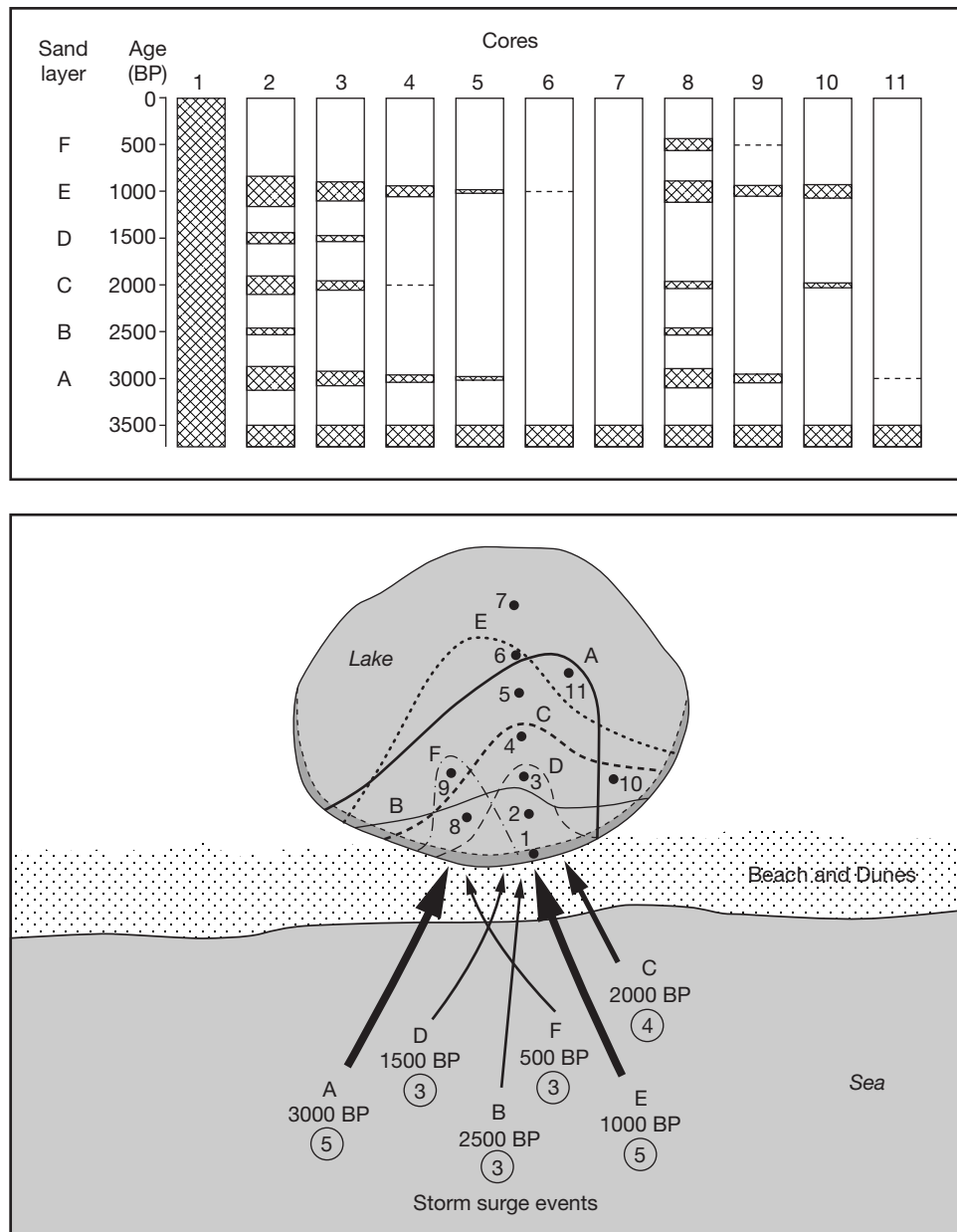


Figure 4 Model showing the positive relationship among hurricane intensities, extent of overwash fans, and thickness of sand layers in sediment cores. Reproduced from Liu K-b and Fearn ML (2000) Reconstruction of prehistoric landfall frequencies of catastrophic hurricanes in northwestern Florida from lake sediment records. *Quaternary Research* 54: 238–245. Top: Hypothetical sedimentary stratigraphies of 11 cores (#1–11) taken from different parts of a lake impacted by intense hurricanes and associated overwash events six times during the past 3000 years. Thick and thin sand layers are represented by cross-shaded bands and dotted lines, respectively. Bottom: Hypothetical pattern of overwash sand deposition in the lake where the 11 cores were taken. Solid, dotted, or dashed lines inside the lake denote the horizontal extents of the six overwash fans (labeled A–F) corresponding to the six hurricane strikes and overwash events (arrows). Thicknesses of the arrows are proportional to the intensity of the hurricanes according to the Saffir–Simpson scale, the latter also designated by the circled number below each arrow. Numbered black dots (#1–11) represent cores taken from the lake.

marsh surface, forming a storm deposit. This storm deposit may consist of sand, silt, or clay, depending on the hydrodynamic conditions of the storm surge and the sediment supply. This layer may also be recognized by the presence of marine microfossils such as dinoflagellates and offshore calcareous Foraminifera. Abruptly terminated oyster-shell layers in a

core may also be interpreted as indicators of storm events, as oyster beds may be buried and killed by thick storm deposition (Scott et al., 2003). Proxy records spanning 5000–6000 years have been obtained from several estuarine/deltaic marshes and tidal inlets in Louisiana, Mississippi, and South Carolina (Liu, 2004; Liu and Fearn, 2000b; Scott et al., 2003).

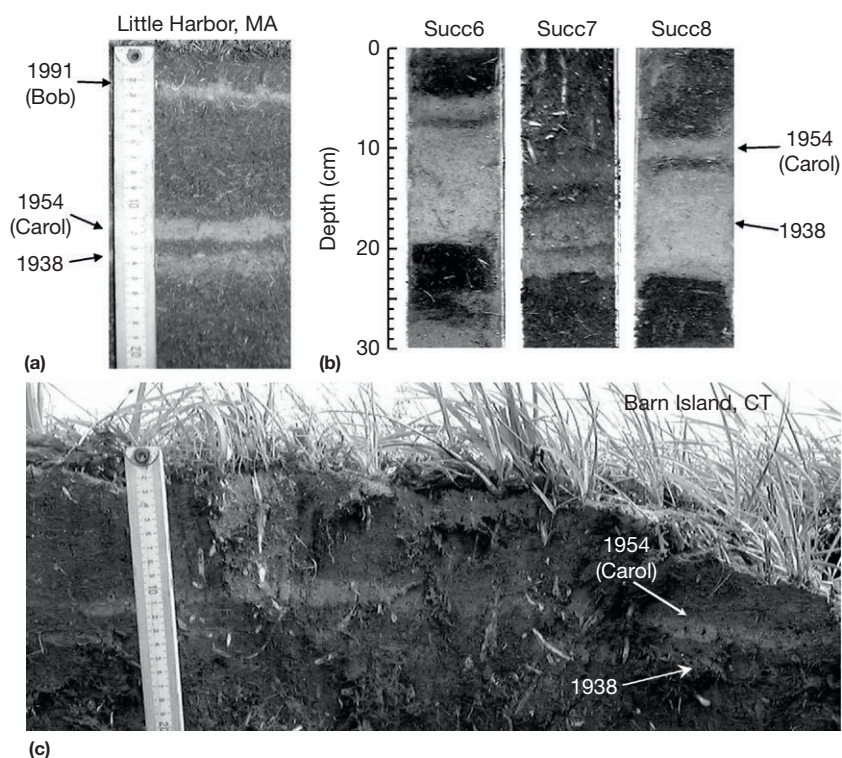


Figure 5 Photographs of overwash sand layers in back-barrier marsh sediments. (a) Three sand layers near the top of a core from Little Harbor Marsh, Massachusetts, deposited by three historic hurricanes: Bob (1991), Carol (1954), and the 1938 Hurricane. (b) Two sand layers in cores taken from Succotash Marsh, Rhode Island, deposited by Hurricane Carol (1954) and the 1938 Hurricane. (c) Two sand layers deposited by the same hurricanes are exposed in a stratigraphic section at Barn Island Marsh, Connecticut. Reproduced from Donnelly JP and Webb T III (2004). Backbarrier sedimentary records of intense hurricane landfalls in the northeastern United States. In: Murnane RJ and Liu K-b (eds.) *Hurricanes and Typhoons: Past, Present, and Future*, pp. 58–95. New York: Columbia University Press.

A rather unusual proxy record was obtained from a currently nontidal meromictic (chemically stratified) lake near Boston, Massachusetts (Besonen et al., 2008). Annually laminated, or varved, sediments deposited in Lower Mystic Lake are interrupted by up to 36 anomalous graded beds that have been interpreted to be deposited by strong hurricane-related flooding events in the lake basin during the past millennium. While their depositional mechanisms remain uncertain and may also involve some nontropical aspects, 10 out of 11 of the most prominent graded beds deposited during the historical period correspond chronologically with the years in which strong hurricanes affected the Boston area. A likely explanation is that strong hurricane winds and intense rainfall caused vegetation disturbance and intensified soil erosion in the watershed, resulting in the formation of these graded, flood deposits (Besonen et al., 2008).

Beach Ridges as Paleohurricane Proxy

While sediments from coastal lakes and wetlands have proven to be the most useful data archive in paleotempestology, other natural archives and proxies have recently been developed. In the coastal zones of northeastern, northern, and western Australia, chronostratigraphic series of shore-parallel beach ridges have been interpreted as proxy records of tropical cyclone strikes spanning up to the last 7000 years (Forsyth

et al., 2010; Nott, 2011; Nott and Hayne, 2001; Nott et al., 2007, 2009; Figure 6). Depending on the supply of source materials, these beach ridges may be composed of pure sand, coral rubble, or pure marine shells or a mixture of these. At each coastal location, typically 10–30 beach ridges are present, and they typically stand between 3 and 6 m above mean sea level (Nott et al., 2007). Their stratigraphy and chronology suggest that they are built by individual storm events as a result of marine inundation and wave transportation of source materials. Dating of successive ridges, usually by the radiocarbon (^{14}C) or optically stimulated luminescence (OSL) method, thus provides a chronological record of past cyclone strikes at that location. The average interval between the emplacement of successive ridges can be taken as a measure of the return period of cyclone events. Intervals generally in the order of 200–300 years have been reported from various coastal locations in Australia (Nott et al., 2007; Nott et al., 2009).

Paleocyclone intensity can be inferred from the elevation of individual ridges based on the assumption that ridge height represents the minimum height of the storm inundation, which is a function of storm intensity. A study using numerical storm surge and wave models suggests that virtually all the beach ridges in northeastern and northern Australia were formed by catastrophic cyclones of category 5 intensity, while those in western Australia were probably emplaced by category 2–4 cyclones (Nott, 2003, 2011; Nott et al., 2007).



Figure 6 Beach ridge plain at Cowley Beach in Queensland, Australia, consisting of 29 shore-parallel sand-dominated beach ridges that provide a proxy record of cyclone activity over the past 5500 years. Reproduced from Nott J (2007) The importance of quaternary records in reducing risk from tropical cyclones. *Palaeogeography, Palaeoclimatology, Palaeoecology* 251: 137–149.

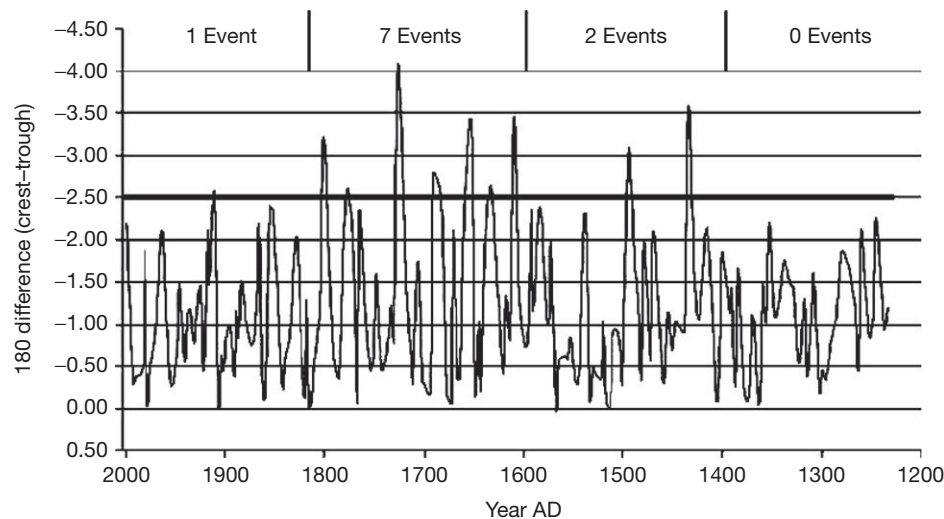


Figure 7 Detrended $\delta^{18}\text{O}$ ‰ record of landfalling cyclones from AD 1228 to 2003 reconstructed from a stalagmite collected from Chillagoe, Queensland, Australia. Reproduced from Nott J (2007) The importance of quaternary records in reducing risk from tropical cyclones. *Palaeogeography, Palaeoclimatology, Palaeoecology* 251: 137–149. The thick horizontal line at -2.50 ‰ denotes a critical value for moderate-to-severe hazard impact attributable to intense cyclones.

Stable Isotopes in Tree Rings and Stalagmites

Another new and promising proxy is the stable isotopic signal preserved in the cellulose of tree rings and the calcium carbonate of stalagmites in limestone caves. Due to intense convection and the fractionation process, rainwater produced by hurricanes has a highly depleted oxygen isotopic signal (i.e., more negative $\delta^{18}\text{O}$ values) compared with that of low-latitude thunderstorms (Lawrence and Gedzelman, 1996). After a hurricane strike, this isotopically depleted signal from the intense hurricane rains may persist in surface or soil water for several weeks before it is attenuated, and may be incorporated into the growth rings of trees and speleothems (cave deposits) in the storm-affected region. A key to applying the

$\delta^{18}\text{O}$ proxy in paleotempestology lies in the ability to sample these annually banded archives and to detect the isotopically depleted signals at subseasonal resolutions, which can now be accomplished through microsampling techniques. Miller et al. (2006) produced a 220-year record of tropical cyclone activity in southern Georgia based on oxygen isotope values extracted from α -cellulose in the latewood portion of longleaf pine tree rings. Similarly, in a feasibility study, Frappier et al. (2007) presented a 23-year (1977–2000) stalagmite record of oxygen isotopic values from Belize that contain multiple negative excursions corresponding to known hurricane rain events in the historical record. A 777-year (AD 1226–2003) stalagmite $\delta^{18}\text{O}$ record obtained from a limestone cave at Chillagoe in

northeastern Queensland, Australia, reveals significant century-scale variability in tropical cyclone activity (Nott et al., 2007; Figure 7).

Paleohurricane Activity and Global Climate Changes

Since the early 1990s, more than two dozen proxy records of paleotropical cyclone activity have been produced from different parts of the world, mainly from the Atlantic and Gulf coasts of North America, the Caribbean region (including Central America), Australia, and the western Pacific (China and Japan). These paleo-records provide a long-term perspective in global hurricane activity, which is vital for understanding the relationships between hurricane activities and global climate changes and their possible controlling mechanisms.

These proxy records reveal that from region to region tropical cyclone activity exhibit long-term variations on timescales of centuries to millennia. On the Gulf of Mexico coast, a hyperactive period existed 3800–1000 years ago, which was preceded and succeeded by relatively quiescent periods during 5000–3800 cal year BP and 1000–0 year BP, respectively (Liu and Fearn, 2000a; Figure 8). A 5400-year proxy record from Puerto Rico in the northern Caribbean suggests that active periods occurred 5400–3600, 2500–1000, and 250–0 years ago, separated by relatively quiet periods in between (Donnelly and Woodruff, 2007). However, a new proxy record from southwestern Japan also suggests increased typhoon activity during 3600–2500 year BP and 1000–300 year BP (Woodruff et al., 2009). Conversely, beach ridge plain chronologies from northeastern and northern Australia exhibit two major gaps in ridge formation, which were interpreted to represent quiescent periods around 3400–2500 year BP and 1500–500 year BP (Forsyth et al., 2010), while a major gap of around 5400–3700 year BP was found in Western Australia (Nott, 2011). On a centennial timescale, the stalagmite $\delta^{18}\text{O}$ record from northeastern Australia shows that over the last 800 years, as many as seven cyclone events occurred during an active period from AD 1600 to 1800, compared with only 1, 2, and 0 events during the other bicentennial periods of AD 1800–2000, 1400–1600, and 1200–1400, respectively (Nott et al., 2007; Figure 7).

Two scenarios have emerged to explain the millennial-scale variations in Atlantic hurricane activity observed in the proxy record. Liu and Fearn (2000a) proposed the ‘Bermuda High hypothesis,’ which evokes long-term shifts in the position of the Bermuda High, the subtropical anticyclone that steers many intense hurricanes (especially the ‘Cape Verde’ type of hurricanes) westward across the tropical Atlantic ocean toward North America. According to this hypothesis, a southwestward shift of the Bermuda High would steer more hurricanes toward the Gulf coast, thereby resulting in hyperactivity such as that detected in the Gulf coast proxy record 3800–1000 years ago, whereas a northeastward shift would steer more hurricanes to hit the Atlantic coast (Liu and Fearn, 2000a; Figure 9). This hypothesis implies an antiphase relationship between the Gulf coast and the Atlantic coast. Such an inverse relationship is supported by a statistical analysis that links the historical records of hurricane activity from the Gulf and Atlantic coasts with the North Atlantic Oscillation (NAO), which is related to the position of the Bermuda High

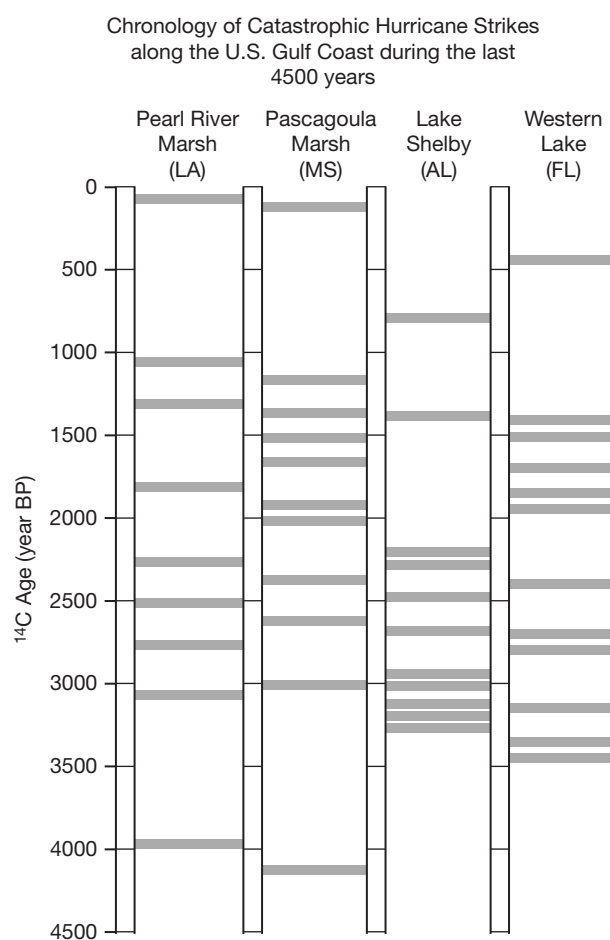


Figure 8 Chronostratigraphic patterns of overwash sand layers (horizontal bars) representing catastrophic hurricane strikes over the last 4500 years documented at four sites along the Gulf coast from west to east: Pearl River Marsh (Louisiana), Pascagoula Marsh (Mississippi), Lake Shelby (Alabama), and Western Lake (Florida). Reproduced from Liu K-b (2004) *Paleotempestology: Principles, methods, and examples from Gulf Coast lake sediments*. In: Murnane RJ and Liu K-b (eds.) *Hurricanes and Typhoons: Past, Present, and Future*, pp. 13–57. New York: Columbia University Press. The clustering of events between approximately 1000 and 3400 ^{14}C years BP (before present) suggests the occurrence of a hyperactive period bracketed by relatively quiet periods. Each site was directly struck by catastrophic hurricanes 9–12 times during the last 3400 ^{14}C years (or 3800 calendar years), suggesting a return period of approximately 350 years.

(Elsner et al., 2000). Paleoclimatic records from North America and adjacent regions are consistent with the scenario of a southwestward shift in the Bermuda High 4000–3000 years ago (Liu and Fearn, 2000a). The proxy record from South Carolina on the Atlantic coast also seems to support the Bermuda High hypothesis (Scott et al., 2003).

The proxy record from Puerto Rico forms the basis of a competing scenario. Donnelly and Woodruff (2007) found that the paleohurricane activities recorded in Laguna Playa Grande correspond negatively with the proxy record of El Niño events, but positively with the proxy record of monsoonal precipitation in tropical Africa, suggesting that Atlantic hurricane activity is strongly modulated by large-scale circulation

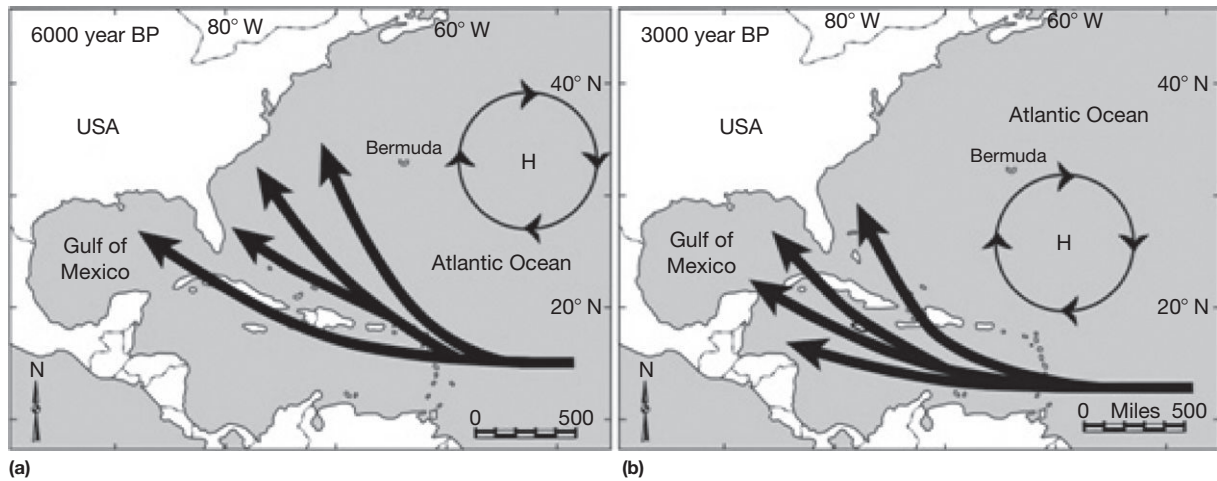


Figure 9 Postulated changes in paleoclimatic conditions and in the position of the Bermuda High around 6000 year BP (a) and 3000 year BP (b) and their impacts on the predominant tracks of catastrophic hurricanes. Reproduced from Liu K-b and Fearn ML (2000) Reconstruction of prehistoric landfall frequencies of catastrophic hurricanes in northwestern Florida from lake sediment records. *Quaternary Research* 54: 238–245. Top: A more northeasterly position of the Bermuda High around 6000 years ago would have resulted in more landfalls on the Atlantic coast. Bottom: A shift of the Bermuda High to a more southeasterly position near the Caribbean around 3000 years ago probably steered more catastrophic hurricanes to hit the Gulf coast.

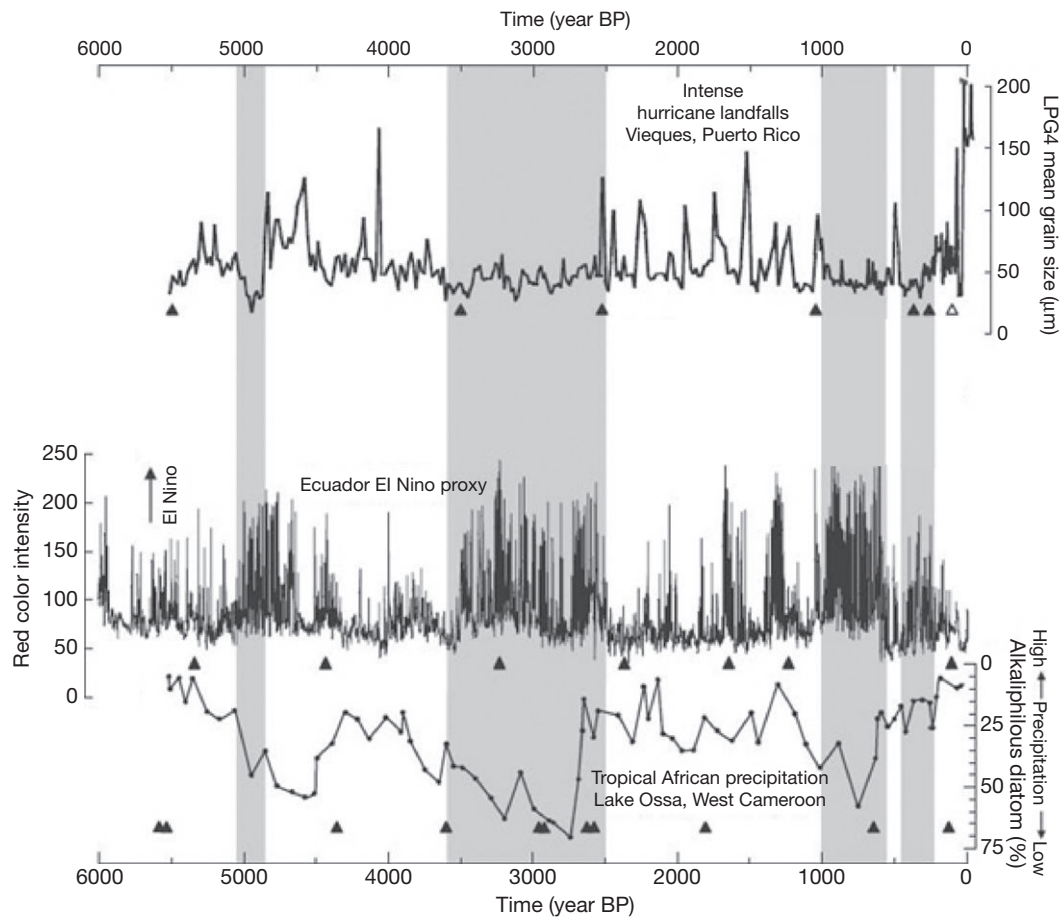


Figure 10 Comparison of reconstructed paleohurricane frequency record from Puerto Rico (top) with reconstructed El Niño frequency record from Ecuador (bottom) and reconstructed West African monsoon precipitation record from West Cameroon. Reproduced from Donnelly JP and Woodruff JS (2007) Intense hurricane activity over the past 5,000 years controlled by El Niño and the West African monsoon. *Nature* 447: 465–468. Periods of low paleohurricane activity in Puerto Rico (gray vertical bars) seem to correspond with the times of frequent El Niño events and low monsoonal precipitation in West Africa.

features such as El Niño/Southern Oscillation (ENSO) and the strength of the West African monsoon (Figure 10). This scenario implies that hurricane activity regimes in different coastal regions across the Atlantic basin would be both in-phase and synchronous, as they are governed by the same set of large-scale climatic factors that affect the rise and fall of hurricane activity. Mann et al. (2009) have produced a curve of basin-wide hurricane frequency indices for the past 1500 years on the basis of a compilation of proxy records in the Atlantic basin.

Historical Documentary Records

For the historical period during which written records are available, information about past tropical cyclone activity can be extracted from documentary sources derived from books or other data archives. The length of time covered by these documentary records varies from one region of the world to another, but it rarely exceeds a few hundred to a thousand years. Though comparatively short in time span, an advantage of these historical documentary records over the geological proxy data is that they usually have very high temporal resolution and precision, typically allowing reconstruction to the year, month, day, and even hour of the event. Many records contain wind direction that can be used to construct storm tracks (e.g., Mock et al., 2010). Some historical records contain reliable barometric pressure data, enabling precise calculations for determining storm intensity.

US Data Sources and Reconstructions

For the Atlantic Basin including the United States, 6-hourly data on the positions and intensities (central pressures and wind speeds) of all known hurricanes and tropical storms in the past one and a half centuries since 1851 are available at the National Hurricane Center's North Atlantic hurricane

database, or HURDAT (Landsea et al., 2004). For the historical period from the sixteenth to the nineteenth century, information about hurricane activity largely comes from documentary sources, including newspapers, plantation diaries, and some instrumental weather records (Chenoweth, 2006; Mock, 2004). Several major inventories of historical hurricanes affecting the United States and the Atlantic Basin (including the Caribbean and Gulf of Mexico region) have been compiled (e.g., Chenoweth, 2006; Fernandez-Partagas and Diaz, 1996; Ludlam, 1963). These documentary data, often narrative and spatially uneven, can be quantified and statistically analyzed to reveal the frequency patterns of hurricane strikes for selected regions. For example, a 223-year time series of tropical cyclone frequencies for South Carolina from AD 1778 to 2000 has been compiled (Mock, 2004). This time series reveals that two periods of unprecedented, very active tropical cyclone activity occurred in the 1830s and from 1880 to 1910 (Figure 11). A similar reconstruction has also been produced for Louisiana for the period 1799–2007 on the basis of a variety of historical records from United States, British, and Spanish sources (Mock, 2008). This time series also shows that Louisiana was impacted by hurricanes more frequently during the nineteenth century than during the twentieth century.

Spanish and British Archives

For the Caribbean region, documentary evidence derived from the aforementioned data sources in the United States or from local newspapers, diaries, or annals in the Caribbean has been used to compile inventories of hurricane activity during the past five centuries (e.g., Chenoweth, 2003; Chenoweth and Divine, 2008; Millas, 1968). However, two vast sources of documentary data have been used for reconstructing the frequency patterns of historical hurricane activity – Spanish archives and British archives (Chenoweth, 2006; Garcia Herrera et al., 2004, 2007). Much unexploited French sources remain to be utilized for this type of research. The Spanish

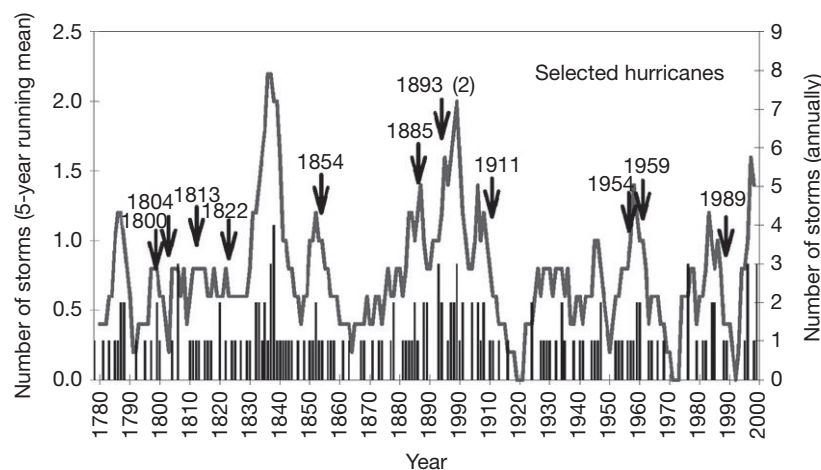


Figure 11 Time series of tropical cyclones affecting Charleston, South Carolina, from AD 1778 to 2000, reconstructed from local newspapers, plantation diaries, and early instrumental records. Reproduced from Mock CJ (2004) Tropical cyclone reconstructions from documentary records: Examples for South Carolina, United State. In: Murnane RJ and Liu K-b (eds.) *Hurricanes and Typhoons: Past, Present, and Future*, pp. 121–148. New York: Columbia University Press. The vertical bars represent annual frequencies (right axis) and the continuous line represents the 5-year running mean (left axis).

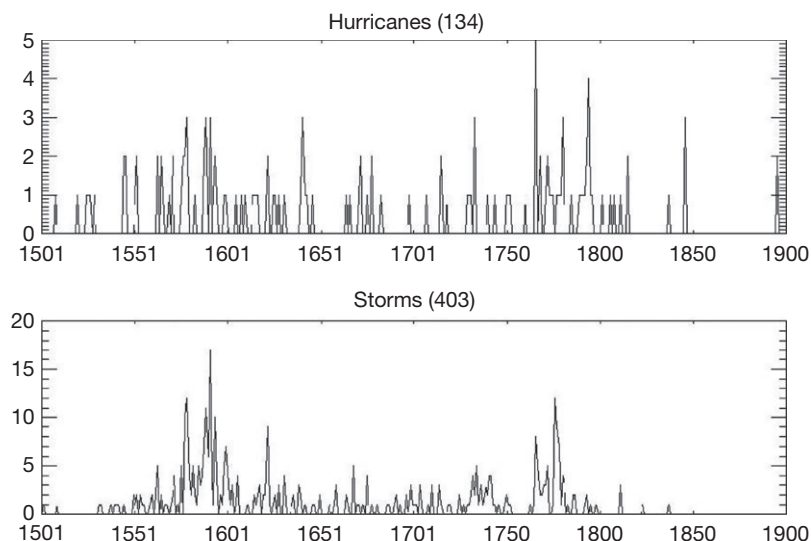


Figure 12 Time series of hurricanes (top) and storms (bottom) occurring in the Caribbean basin from the sixteenth to the nineteenth century, reconstructed from Spanish documentary evidence preserved in the Archivo General de Indias (AGI). The term storm refers to any event of excessive wind that lacks a clear indication of the occurrence of a shift in wind direction typically associated with the vortex of a hurricane. This category may include, among others, less intense tropical storms, extratropical storms, squall lines, and strong frontal passage episodes. Reproduced from Garcia Herrera R, Gimeno L, Ribera P, and Hernandez E (2005) New records of Atlantic hurricanes from Spanish documentary sources. *Journal of Geophysical Research* 110: D03109.

archives consist of millions of official documents recording all aspects of daily life in the Spanish colonies in the Americas from the Columbian contact to the nineteenth century. Most of these colonial documents are preserved in the Archivo General de Indias (AGI) in Seville, and several other archives in Madrid and Valladolid. The British archives, mainly starting in the mid-seventeenth century and covering a shorter time span than their Spanish counterparts, principally consist of mariners' logbooks kept on board Royal Navy ships during the period of English colonization of the Caribbean and related colonial papers found at the United Kingdom National Archives in London. British newspapers, available from the British Library, also based in London, supplement these colonial archives (Chenoweth, 2006; Chenoweth and Divine, 2008). The potential of all these colonial archives as a rich data source for paleotempestology is high, but work has only begun to fully exploit this potential due to the vast amount of documents contained in them, which extends well beyond the Atlantic Basin (Garcia Herrera et al., 2004, 2007; Mock, in press). An example of a reconstruction of Caribbean hurricane and storm activity for the period AD 1500–1900 using the Spanish archives shows two active periods in the sixteenth and the eighteenth centuries separated by a relatively inactive period in the seventeenth century, with the activity peaking during 1766–80 (Garcia Herrera et al., 2005; Figure 12).

Chinese Documentary Records and Guangdong Typhoon Time Series

By far, the longest documentary record of tropical cyclone activity can be found in China, where the historical record of typhoon landfalls spans more than 1000 years. Abundant information on typhoon landfalls can be found in the official histories commissioned by the central imperial government

(*zhengshi*), Veritable Records (*shilu*) of the Emperor's daily activities, semiofficial local gazettes (*fang zhi*) written by scholars at the county, district, or provincial level, and unofficial literary works such as personal diaries, travel logbooks, and poems (Louie and Liu, 2004). These Chinese historical sources reveal that the term *jufeng* was coined as early as AD 470 to designate typhoon (a wind that comes in all four directions) as a specific meteorological phenomenon that is distinct from other kinds of storms (Louie and Liu, 2003). A typhoon that struck the Shandong Peninsula in eastern China in AD 816 is probably the earliest recorded tropical cyclone landfall in the world (Louie and Liu, 2003).

Among the four sources of Chinese documentary data mentioned earlier, the local gazettes or *fang zhi*, especially those kept by coastal counties in southern and eastern China, are the richest data source for historical typhoon landfalls. For the majority of counties in China, there is at least one volume of county gazette that is accessible in a public library in China or overseas. Liu et al. (2001) have produced a 935-year time series of typhoon landfalls for the Guangdong Province of southern China spanning the period from AD 975 to 1909 based on data extracted from local gazettes (Figure 13). This 1000-year record of tropical cyclone time series – the longest in the world – reveals an approximately 50-year periodicity that may be linked to the Pacific Decadal Oscillation (PDO; Liu et al., 2001). The data also show two prominent peaks in Guangdong typhoon activity occurring at AD 1660–80 and AD 1850–80. Remarkably, these two typhoon maxima coincide with two of the coldest and driest episodes in northern and central China during the Little Ice Age (Figure 14). Liu et al. (2001) hypothesized that the apparent increase in typhoon activity in Guangdong during these cold episodes was caused by a southward displacement of the storm tracks associated with an intensification of the westerlies and a southward shift of the

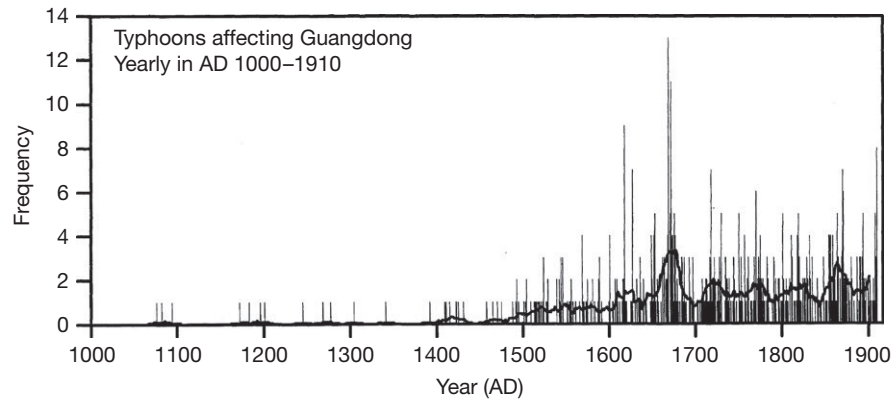


Figure 13 Time series of typhoons striking Guangdong Province, southern China, from AD 1000 to 1909, reconstructed from Chinese local gazette (*fang zhi*) records. The vertical bars represent annual frequencies and the continuous line represents the 21-year running mean. Reproduced from Liu K-b, Shen C, and Louie KS (2001) A 1000-year history of typhoon landfalls in Guangdong, southern China, reconstructed from Chinese historical documentary records. *Annals of the Association of American Geographers* 91: 453–464.

subtropical high-pressure system, resulting in more typhoons making landfall in Guangdong while fewer striking the coastal provinces further north. Statistical analysis of modern and historical typhoon data from the western Pacific suggests that more frequent occurrence of La Nina conditions may also have caused a change in the predominant storm tracks, resulting in more typhoons moving straightly westward to hit Guangdong instead of recurving northwest to hit Japan (Elsner and Liu, 2003).

Research Needs and Future Research Directions

Paleotempestology is a relatively young science that started in the early 1990s (Liu and Fearn, 1993). The term paleotempestology was coined by Kerry Emanuel at the Massachusetts Institute of Technology (MIT) in 1998. It has since developed significantly in scope, in theory and methods, and in the number of practitioners. The following research needs and directions can be envisioned in the near future:

1. Expansion of geographical coverage and spatial data network: Early proxy work in paleotempestology started in the Gulf of Mexico coast, but more recently, proxy records have been obtained from the US Atlantic coast, the Caribbean region, Australia, and China and Japan. However, there is still a strong need for more well-validated and well-dated proxy records from all the tropical cyclone-prone regions of the world. For example, proxy records have yet to be published from the western coast of Mexico, which is impacted by hurricanes from the Eastern North Pacific basin.
2. Development of new proxies and archives: A promising research direction in paleotempestology involves the development of multiproxy techniques in detecting the geophysical and ecological impacts of past hurricanes. To date, the proxy that has been proven to be most effective in paleotempestology is the overwash sediments preserved in coastal lakes and wetlands, while beach ridge chronology has also been developed as a useful tool in Australia. Significant methodological advances have been made in the development of new proxies such as stable isotopic signals in tree rings, speleothems, and corals, but more calibration and sensitivity studies (e.g., Lewis et al., 2011) are needed to refine these promising techniques.
3. Coupling of geological and documentary archives: The voluminous documentary records in the Spanish, British, French, United States, Canadian, Chinese, and other colonial archives have barely been exploited for the reconstruction of tropical cyclone activities during the past centuries. Much more needs to be done. Wherever both geological and documentary records exist, they should be integrated for data calibration and comparison.
4. Improved calibration of proxy records with modern analog studies: While the frequency of past hurricane strikes can be estimated by enumerating the overwash sand layers in a core, the reconstruction of paleohurricane intensity is a more challenging task in paleotempestology. Regardless of the proxies or archives being used, intensity reconstruction for paleohurricane events can be improved by better understanding of the environmental impacts and/or sedimentological signatures of modern hurricanes of known intensities. For example, recent intense hurricanes such as Camille, Andrew, Hugo, Ivan, Katrina, and Rita can be used as modern analogs to aid proxy-based paleohurricane intensity reconstructions.
5. Integration of proxy data with modeling results: Another approach in improving the intensity estimate for paleohurricanes from proxy data is by means of computer modeling experiments. Storm surge modeling can simulate the overwash processes under various boundary conditions with regard to hurricane intensity, oceanographic response, and coastal terrain characteristics (e.g., Cheung et al., 2007). These modeling results can be integrated with the proxy record to help validate the intensity estimates.
6. Statistical applications in paleotempestology: When coupled with the historical hurricane record, proxy data can improve our estimates of the hurricane risk in a region by providing a longer, though chronologically less precise, record of past events, especially the extreme events. These historical and proxy data can be used in the statistical modeling of wind speed exceedance probabilities.

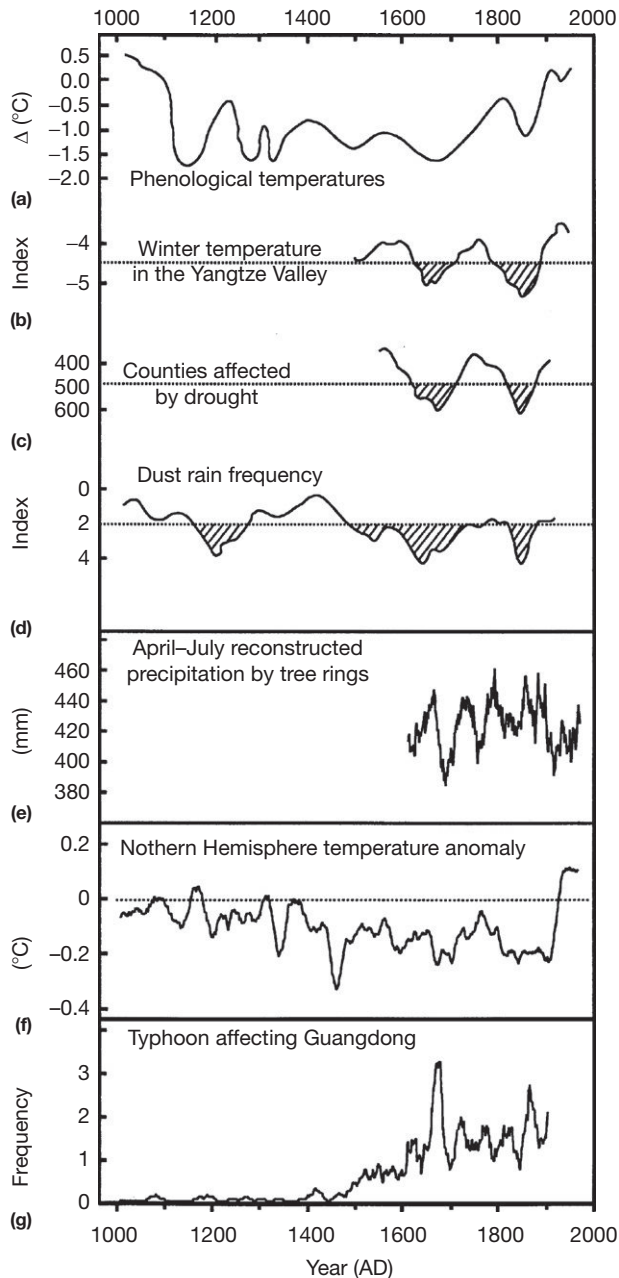


Figure 14 Comparison between (g) the Guangdong typhoon time series (21-year running mean) with other paleoclimatic time series from China and the Northern Hemisphere for the last millennium: (a) phenologically derived temperature in China, (b) winter temperature in the lower Yangtze river valley, (c) number of Chinese counties affected by drought, (d) frequency of dust rains in China, (e) tree-ring-based reconstruction of April-to-July precipitation in Huashan, north-central China, and (f) northern hemisphere temperature anomaly derived from a variety of paleoclimatic data sources. Reproduced from Liu K-b, Shen C, and Louie KS (2001) A 1000-year history of typhoon landfalls in Guangdong, southern China, reconstructed from Chinese historical documentary records. *Annals of the Association of American Geographers* 91: 453–464. The two peaks in Guangdong typhoon activity in AD 1660–80 and 1850–80 correspond to two of the coldest and driest episodes in central and northern China during the Little Ice Age.

Advanced statistical techniques such as Bayesian analysis and extreme value theory can be applied to the modern and proxy data to reduce the uncertainty of error estimates and to improve the assessment of hurricane risks along the coast. Work has just begun in this research direction in paleotempestology (Elsner et al., 2008).

See also: **Paleoclimate Reconstruction: Historical Climatology.**

References

- Besonen MR, Bradley RS, Mudelsee M, Abbott MB, and Francus P (2008) A 1,000-year, annually-resolved record of hurricane activity from Boston, Massachusetts. *Geophysical Research Letters* 35: L14705.
- Chenoweth M (2003) *The 18th Century Climate of Jamaica*. Pennsylvania, PA: American Philosophical Society.
- Chenoweth M (2006) A reassessment of historical Atlantic Basin tropical cyclone activity, 1700–1855. *Climatic Change* 76: 169–240.
- Chenoweth M and Divine D (2008) A document-based 318-year record of tropical cyclones in the Lesser Antilles, 1690–2007. *Geochemistry, Geophysics, Geosystems* 9: Q08013. <http://dx.doi.org/10.1029/2008GC002066>.
- Cheung KF, Tang LJ, Donnelly JP, et al. (2007) Numerical modeling and field evidence of coastal overwash in southern New England from Hurricane Bob and implications for paleotempestology. *Journal of Geophysical Research* 112: F03024. <http://dx.doi.org/10.1029/2006JF000612>.
- Donnelly JP (2005) Evidence of past intense tropical cyclones from back barrier salt pond sediments: A case study from Isla de Culebrita, Puerto Rico, USA. *Journal of Coastal Research* 42: 201–210.
- Donnelly JP, Bryant SS, Butler J, et al. (2001a) A 700 yr sedimentary record of intense hurricane landfalls in southern New England. *Geological Society of America Bulletin* 113: 714–727.
- Donnelly JP, Butler J, Roll S, Wengren M, and Webb T III (2004) A backbarrier overwash record of intense storms from Brigantine, New Jersey. *Marine Geology* 210: 107–121.
- Donnelly JP, Roll S, Wengren M, Butler J, Lederer R, and Webb T III (2001b) Sedimentary evidence of intense hurricane strikes from New Jersey. *Geology* 29: 615–618.
- Donnelly JP and Webb T III (2004) Backbarrier sedimentary records of intense hurricane landfalls in the northeastern United States. In: Murnane RJ and Liu K-b (eds.) *Hurricanes and Typhoons: Past, Present, and Future*, pp. 58–95. New York: Columbia University Press.
- Donnelly JP and Woodruff JS (2007) Intense hurricane activity over the past 5,000 years controlled by El Niño and the West African monsoon. *Nature* 447: 465–468.
- Elsner JB, Jagger TH, and Liu KB (2008) Comparison of hurricane return levels using historical and geological records. *Journal of Applied Meteorology and Climatology* 47: 368–375.
- Elsner JB and Liu K-b (2003) Examining the ENSO-typhoon hypothesis. *Climate Research* 25: 43–54.
- Elsner JB, Liu K-b, and Kocher BL (2000) Spatial variations in major U.S. hurricane activity: Statistics and a physical mechanism. *Journal of Climate* 13: 2293–2305.
- Fernandez-Partagas J and Diaz HF (1996) Atlantic hurricanes in the second half of the nineteenth century. *Bulletin of the American Meteorological Society* 77: 2899–2906.
- Forsyth A, Nott J, and Bateman M (2010) Beach ridge plain evidence of a variable late-Holocene tropical cyclone climate, North Queensland, Australia. *Paleogeography, Palaeoclimatology, Palaeoecology* 297: 707–716.
- Frappier AB, Sahagian D, Carpenter SJ, González LA, and Frappier BR (2007) Stalagmitestable isotope record of recent tropical cyclone events. *Geology* 35: 111–114.
- García Herrera R, Gimeno L, Ribera P, and Hernandez E (2005) New records of Atlantic hurricanes from Spanish documentary sources. *Journal of Geophysical Research* 110: D03109.
- García Herrera R, Gimeno L, Ribera P, Hernandez E, Gonzalez E, and Fernandez G (2007) Identification of Caribbean basin hurricanes from Spanish documentary sources. *Climatic Change* 83: 55–85.
- García Herrera R, Rubio F, Wheeler D, Hernandez E, Prieto MR, and Gimeno L (2004) The use of Spanish and British documentary sources in the investigation of Atlantic hurricane incidence in historical times. In: Murnane RJ and Liu K-b (eds.)

- Hurricanes and Typhoons: Past, Present, and Future*, pp. 149–176. New York: Columbia University Press.
- Hippensteel SP and Martin RE (1999) Foraminifera as an indicator of overwash deposits, barrier island sediment supply, and barrier island evolution: Folly Island, South Carolina. *Palaeogeography, Palaeoclimatology, Palaeoecology* 149: 115–125.
- Landsea CW, Anderson C, Charles N, et al. (2004) The Atlantic hurricane database re-analysis project: Documentation for the 1851–1910 alterations and additions to the HURDAT database. In: Murnane RJ and Liu K-b (eds.) *Hurricanes and Typhoons: Past, Present and Future*, pp. 177–221. New York: Columbia University Press.
- Lawrence JR and Gedzelman SD (1996) Low stable isotope ratios of tropical cyclone rains. *Geophysical Research Letters* 23: 527–530.
- Lewis DB, Finkelstein DB, Grissino-Mayer HD, Mora CI, and Perfect E (2011) A multitree perspective of the tree ring tropical cyclone record from longleaf pine (*Pinus palustris* Mill.), Big Thicket National Preserve, Texas, United States. *Journal of Geophysical Research* 116: G02017. <http://dx.doi.org/10.1029/2009JG001194>.
- Liu K-b (2004) Paleotempestology: Principles, methods, and examples from Gulf coast lake sediments. In: Murnane RJ and Liu K-b (eds.) *Hurricanes and Typhoons: Past, Present, and Future*, pp. 13–57. New York: Columbia University Press.
- Liu K-b and Fearn ML (1993) Lake-sediment record of late Holocene hurricane activities from coastal Alabama. *Geology* 21: 793–796.
- Liu K-b and Fearn ML (2000a) Reconstruction of prehistoric landfall frequencies of catastrophic hurricanes in northwestern Florida from lake sediment records. *Quaternary Research* 54: 238–245.
- Liu K-b and Fearn ML (2000b) Holocene history of catastrophic hurricane landfalls along the Gulf of Mexico coast reconstructed from coastal lake and marsh sediments. In: Ning ZH and Abdollahi K (eds.) *Current Stresses and Potential Vulnerabilities: Implications of Global Change for the Gulf Coast Region of the United States*, pp. 38–47. Baton Rouge, LA: Franklin Press.
- Liu K-b, Lu HY, and Shen C (2008) A 1,200-year proxy record of hurricanes and fires from the Gulf of Mexico coast: Testing the hypothesis of hurricane-fire interactions. *Quaternary Research* 69: 29–41.
- Liu K-b, Shen C, and Louie KS (2001) A 1000-year history of typhoon landfalls in Guangdong, southern China, reconstructed from Chinese historical documentary records. *Annals of the Association of American Geographers* 91: 453–464.
- Louie KS and Liu K-b (2003) Earliest historical records of typhoons in China. *Journal of Historical Geography* 29: 299–316.
- Louie KS and Liu K-b (2004) Ancient records of typhoons in Chinese historical documents. In: Murnane RJ and Liu K-b (eds.) *Hurricanes and Typhoons: Past, Present, and Future*, pp. 222–248. New York: Columbia University Press.
- Lu HY and Liu K-b (2005) Phytolith assemblages as indicators of coastal environmental changes and hurricane overwash deposition. *The Holocene* 15: 965–972.
- Ludlam DM (1963) *Early American Hurricanes, 1492–1870*. Boston, MA: American Meteorological Society.
- Mann ME, Woodruff JD, Donnelly JP, and Zhang Z (2009) Atlantic hurricanes and climate over the past 1,500 years. *Nature* 460: 880–883.
- McCloskey TA and Keller G (2009) 5000 year sedimentary record of hurricane strikes on the central coast of Belize. *Quaternary International* 195: 53–68.
- Millas JC (1968) *Hurricanes of the Caribbean and Adjacent Regions, 1492–1800*. Miami, FL: Academy of the Arts and Sciences of the Americas.
- Miller DL, Mora CI, Grissino-Mayer HD, Mock CJ, Uhle ME, and Sharp Z (2006) Tree-ring isotope records of tropical cyclone activity. *Proceedings of the National Academy of Sciences* 103: 14294–14297.
- Mock CJ (2004) Tropical cyclone reconstructions from documentary records: Examples for South Carolina, United State. In: Murnane RJ and Liu K-b (eds.) *Hurricanes and Typhoons: Past, Present, and Future*, pp. 121–148. New York: Columbia University Press.
- Mock CJ (2008) Tropical cyclone variations in Louisiana, U.S.A., since the eighteenth century. *Geochemistry, Geophysics, Geosystems* 9: Q05V02. <http://dx.doi.org/10.1029/2007GC001846>.
- Mock CJ (2012) Instrumental and documentary evidence of environmental change. In: Matthews JA, Bartlein PJ, Briffa KR, et al. (eds.) *The Sage Handbook of Environmental Change*, pp. 345–360. Thousand Oaks, CA: Sage Press.
- Mock CJ, Chenoweth M, Altamirano I, Rodgers MD, and Garcia-Herrera R (2010) The Great New Orleans Hurricane of 1812. *Bulletin of the American Meteorological Society* 91: 1653–1663.
- Nott J (2003) Intensity of prehistoric tropical cyclones. *Journal of Geophysical Research* 108(D7): 4212–4223.
- Nott J (2011) A 6,000 year tropical cyclone record from Western Australia. *Quaternary Science Reviews* 30: 713–722.
- Nott J, Haig J, Neil H, and Gillieson D (2007) Greater frequency variability of landfalling tropical cyclones at centennial compared to seasonal and decadal scales. *Earth and Planetary Science Letters* 255: 367–372.
- Nott J and Hayne M (2001) High frequency of ‘super-cyclones’ along the Great Barrier Reef over the past 5,000 years. *Nature* 413: 508–512.
- Nott J, Smithers S, Walsh K, and Rhodes E (2009) Sand beach ridges record 6000 year history of extreme tropical cyclone activity in northeastern Australia. *Quaternary Science Reviews* 28: 1511–1520.
- Scileppi E and Donnelly JP (2007) Sedimentary evidence of hurricane strikes in western Long Island, New York. *Geochemistry, Geophysics, Geosystems* 8(6). <http://dx.doi.org/10.1029/2006GC001463>.
- Scott DB, Collins ES, Gayes PT, and Wright E (2003) Records of prehistoric hurricanes on the South Carolina coast based on micropaleontological and sedimentological evidence, with comparison to other Atlantic Coast record. *Geological Society of America Bulletin* 115: 1027–1039.
- Wallace DJ and Anderson JB (2010) Evidence of similar probability of intense hurricane strikes for the Gulf of Mexico over the late Holocene. *Geology* 38: 511–514.
- Williams HFL (2009) Stratigraphy, sedimentology, and microfossil content of Hurricane Rita storm surge deposits in Southwest Louisiana. *Journal of Coastal Research* 25: 1041–1051.
- Williams HFL (2010) Storm surge deposition by Hurricane Ike on the McFaddin National Wildlife Refuge, Texas: Implications for paleotempestology studies. *Journal of Foraminiferal Studies* 40: 510–519.
- Woodruff JD, Donnelly JP, Mohrig D, and Geyer WR (2008) Reconstructing relative flooding intensities responsible for hurricane induced deposits from Laguna Playa Grande, Vieques, Puerto Rico. *Geology* 36: 391–394.
- Woodruff JD, Donnelly JP, and Okusu A (2009) Exploring typhoon variability over the mid-to-late Holocene: Evidence of extreme coastal flooding from Kamikoshiki, Japan. *Quaternary Science Reviews* 28: 1774–1785.
- Yu K, Zhao J, Shia Q, and Meng Q (2009) Reconstruction of storm/tsunami records over the last 4000 years using transported coral blocks and lagoon sediments in the southern South China Sea. *Quaternary International* 195: 128–137.

Younger Dryas Oscillation, Global Evidence

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Introduction

A thorough review of the global Younger Dryas oscillation is a difficult task, not possible to be covered completely in this article. The approach taken here is one of many possible approaches to this topic, namely, a terrestrial geological one. This is, however, adequate because the original identification, naming, and documentation of the Younger Dryas event was carried out by Scandinavian Quaternary terrestrial geologists and paleobotanists. In addition, the Younger Dryas is still probably best manifested by an often distinct lithologic change in lacustrine records of Western Europe (Figure 1).

The concept that the late-glacial interval ended with a distinct cold period originates from paleobotanical and lithostratigraphic studies of Swedish and Danish bog and lake sites (Andersson, 1896; Hartz and Milthers, 1901), such as the Allerød clay pit in Denmark (Hartz and Milthers, 1901). The term 'Dryas' refers to the occasionally abundant finds of *Dryas octopetala* (Mountain Avens) leaves in these often minerogenic-rich lake sediments. The prefix 'Younger' stems from the insight that the original 'Dryas' period was preceded by a warmer stage (Allerød), which in turn was underlain by sediments implying at least another cold period (Hartz and Milthers, 1901). With the onset of pollen analysis in 1916, the later refining of pollen analytical techniques, and a steadily growing number of pollen diagrams, it was obvious that the Younger Dryas represented a distinct vegetational setback in large parts of Europe, and during a time when a general glacial–interglacial plant succession and immigration took place. It was generally interpreted as an effect of a sudden decrease in (annual) temperature, unfavorable for the previously rapidly spreading forest vegetation. The cooling favored the expansion of cold-tolerant, light-demanding plants, and led to glacial advances and a lowering of the equilibrium-line altitude. This is clearly documented in, for example, Norway, Sweden, and Finland by the Ra-Middle Swedish endmoraines–Salpausselkä moraine system, by glacier advances in the Alps, and reglaciation in ice-free areas such as the northern British Isles, Faroe Islands, and southern Sweden.

When carbon-14 (^{14}C) dating became a general dating tool, two of the Dryas periods, the Older and Younger Dryas, were designated Nordic late-glacial chronozones (Mangerud et al., 1974), while the cold period preceding the first late-glacial warming and the Bølling Chronozone was often referred to as the Oldest Dryas stadial. Based on a set of Scandinavian ^{14}C records, the age of the Younger Dryas Chronozone was set to 11 000–10 000 ^{14}C years before the present (BP). However, this Nordic nomenclature was soon to become globally used, although the rather detailed climatostratigraphy upon which it was based, was at that time rarely found outside Europe. The research history of the late-glacial zones/periods is summarized by Björck et al. (1998), who also suggested that the often inconsistent stratigraphic nomenclature of the late-glacial

oscillations, including Younger Dryas, should be termed events and be defined by the Greenland Summit ice core stratigraphies. In this new scheme, the Younger Dryas corresponds to Greenland Stadial 1 (GS-1), and the original Allerød warm period is defined as three events: Greenland Interstadial-1c to 1a (GI-1c to GI-1a).

Although the Younger Dryas oscillation had been a well-known feature among European Quaternary geologists for almost a century, it was not until it was 'discovered' on the North American continent that it became a really 'hot' research topic. In spite of reports published at least half a century ago of complex sediment lithologies in eastern Canada and distinct pollen stratigraphic changes in New England in the late-glacial age, these findings were not regarded as evidence for climatic oscillations. However, in the mid-1980s, researchers began to relate these phenomena to a Younger Dryas cooling; old sites were reinterpreted in New England (Peteet, 1987) and in eastern Canada (Mott et al., 1986), and new sites with a presumed Younger Dryas oscillation were found (summarized by Wright, 1989). Mainly based on marine evidence, Ruddiman and McIntyre (1981) presented a model for oceanic Polar Front shifts during the last deglaciation, in which the Polar Front was displaced far south (northern Spain) in the North Atlantic about the same time as when the Younger Dryas cooling in mainland Europe was presumed to have occurred. The implication was that a large part of the North Atlantic, and thus most of the European Atlantic coasts, was shut off from warm Atlantic waters and that the marine environment north of the front was characterized by abundant icebergs and sea ice. The lack of Younger Dryas records in North America probably prevented the authors from drawing the position of the front all the way to the east coast of America. The Ruddiman–McIntyre model of Polar Front shifts and Boyle and Keigwin's (1987) results from the Bermuda Rise showing reduced North Atlantic deep water (NADW) formation during the Younger Dryas played an important role in the increased understanding of the late-glacial climatic shifts, especially, perhaps, of some of the characteristics of, and mechanisms behind, the Younger Dryas cooling.

In the late 1980s, the evidence for a circum-North Atlantic Younger Dryas cooling became more widely accepted, and the common late-glacial climatic pattern around the North Atlantic was clearly manifested by the regional summaries of the North Atlantic Seaboard Programme (NASP) of International Geological Correlation Programme project 253 as well as by the NASP synthesis (Lowe et al., 1994).

The Northern Hemisphere Younger Dryas

With the overwhelming evidence of an amphi-North Atlantic Younger Dryas cooling, it became clear that it must have had an impact over large regions, perhaps even a hemispheric or



Figure 1 A section in a peat bog in Schleswig-Holstein, northern Germany, dug out by Hartmut Usinger in Kiel. The thick light gray unit is a lacustrine sediment, clay gyttja, formed during the Younger Dryas oscillation. The dark unit below is an organic-rich sediment formed during the Bølling-Allerød warm period (GS-1), and the unit above is Holocene gyttja and peat. Note the distinct lower and upper boundaries of Younger Dryas, and for scaling the spade in foreground.

global influence (Broecker, 1994). Its strong impact on North Atlantic trade-wind intensity and regional precipitation patterns shown in the Cariaco basin of the tropical North Atlantic (Hughen et al., 1996) was a clear indication of its large influence on Northern Hemisphere climate.

Possible Triggering Mechanisms Behind the Younger Dryas

Changes in NADW and thermohaline circulation (THC) were excellent mechanisms for explaining the late-glacial climatic oscillations (Boyle and Keigwin, 1987), most clearly displayed in north-west European records where the North Atlantic current has its strongest impact. However, the exceptionally long and severe Younger Dryas cooling, compared with other late-glacial North Atlantic coolings (Björck et al., 1996), led to a special interest in its underlying mechanisms. The idea that a sudden fresh-water event triggered the onset of Younger Dryas, such as the diversion of the outlet of Lake Agassiz from the Gulf of Mexico to an eastern outlet with a more or less catastrophic drainage through the St. Lawrence River (Broecker et al., 1988) and a simultaneous Baltic Ice Lake drainage (Björck et al., 1996) into the North Sea, is a mechanism that is supported by coupled atmosphere-ocean modeling (Manabe and Stouffer, 1988). A sudden decrease in salinity in areas where NADW normally occurs would have a rapid and disturbing effect on the THC. This would seriously hamper the northward meridional heat advection, and this model of explaining the sudden onset and exceptional length of the Younger Dryas has been widely acknowledged as the most plausible mechanism for

the onset of the oscillation. The problem is, however, that it is only based on circumstantial evidence; unless we can pinpoint the onset of such events to the very same year, the link remains indirect. Furthermore, the Younger Dryas Lake Agassiz drainage through the Great Lakes has recently been questioned (Teller et al., 2005), but the last words on this topic are certainly not written.

Timing of the Younger Dryas

The occurrence of a cold oscillation preceding the warm Holocene in all the deep Greenland ice cores, starting with the Camp Century core (Johnsen et al., 1972), became a strong support for those who claimed that Younger Dryas was a significant climatic oscillation. The Greenland Summit ice cores Greenland Ice Sheet Project-2 and Greenland Ice Core Project (GRIP) imply that the Younger Dryas lasted 1150–1300 years, depending on the ice core that was used for the determination (Alley, 2000), and that it started around 12 800 ice years BP. Likewise, the age of the end of the Younger Dryas varies between different Greenland ice cores and their different age models. However, the most likely age seems to be 11 600–11 650 cal. years BP, and it appears to have been synchronous over large parts of the North Atlantic region (Björck et al., 1996; Hughen et al., 1996).

The suggested ^{14}C age of 11 000–10 000 years BP for the Younger Dryas Chronozone (Mangerud et al., 1974) also meant that this ^{14}C age usually was assigned to the climatic oscillation. However, with the advent of ^{14}C accelerator mass spectrometry (AMS) dating during the early 1980s it became possible to ^{14}C date very small quantities (a few milligrams) of well-defined terrestrial organic material (e.g., leaves and seeds) formed in direct contact with atmospheric carbon dioxide. Previously, the normal type of dated material consisted of bulk sediment material, prone to different dating problems such as hard-water errors and reworked organic material. With the increasing number of geologically ‘more reliable’ AMS ^{14}C measurements, it became clear that atmospheric ^{14}C variations were considerable during the millennia preceding the Holocene. In combination with the idea that the late-glacial oscillations were triggered by a varying strength of the North Atlantic THC, and thus by changes in ocean ventilation, these ^{14}C variations were often interpreted as responses within the carbon cycle of atmosphere-ocean exchange processes. For example, during the early 1990s, an increasing number of high-resolution ^{14}C dates around the Allerød-Younger Dryas boundary began to indicate that this often distinct paleoclimatic change was difficult to ^{14}C date in detail; depending on the side of the boundary the ^{14}C samples originated from, ages could vary between 11 000 and 10 600 ^{14}C years BP. Studies also showed that a ^{14}C age of 11 000 years BP clearly predates the boundary, and that the first (oldest) Younger Dryas ^{14}C samples often obtain ages of 10 700–10 600 ^{14}C years BP (Björck et al., 1996). In fact, this very rapid age shift, which we today know is a robust feature, could be used to answer different questions and hypotheses about Younger Dryas. Was this distinct rise in atmospheric ^{14}C content ($\Delta^{14}\text{C}$ values) an effect of a severe disturbance in the North Atlantic THC resulting in decreased (25–50%) ventilation between the ocean and the atmosphere? If this was the case it could, for example, imply that the onset of Younger Dryas was characterized by a

rapid spread of sea ice. In addition, the large ^{14}C changes at this boundary are an excellent time marker for evaluating if the event is hemispherically and/or globally synchronous.

Although the end of the Younger Dryas was pragmatically set to 10000 ^{14}C years BP in the Nordic chronostratigraphy (Mangerud et al., 1974), the majority of age estimates for the end of the climatic cooling, often based on bulk sediment dates, was usually centered around 10200 ^{14}C years BP. However, with the possibility of AMS dating of identified macrofossils, it gradually became clear that the Younger Dryas–Preboreal transition around the North Atlantic was characterized by a so-called ^{14}C plateau (Ammann and Lotter, 1989), that is, a long period when ^{14}C ages are roughly the same, resulting in declining $\Delta^{14}\text{C}$ values. The new calibration curve (Reimer et al., 2004), with the end of Younger Dryas at 11650–11600 cal. years BP, shows that the last 200 years of Younger Dryas are characterized by a gradual decline in ^{14}C ages from 10200 to 10050 ^{14}C years BP (Figure 2). At the transition, a 200–250-year long ^{14}C plateau starts, with ^{14}C ages varying between 10050 and 10000 ^{14}C years BP; thus a ^{14}C age of 10000–10050 for the end of Younger Dryas is presently the best estimate. The boundary now seems to coincide approximately with the onset of a ^{14}C plateau, implying an increased global ocean ventilation rate; that is, if carbon cycle changes, and not ^{14}C production, are responsible for this decrease in $\Delta^{14}\text{C}$.

We can conclude that the Younger Dryas cool event is bordered by distinct changes in the atmospheric ^{14}C content. By comparisons with the beryllium-10 (^{10}Be) record from ice cores, it shows that the main part of these changes cannot be

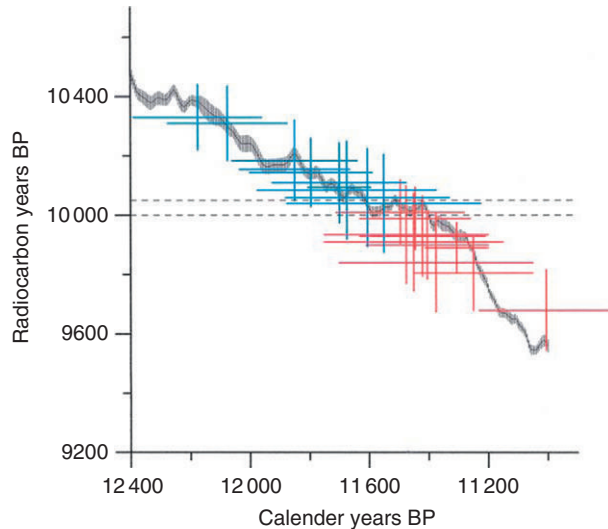


Figure 2 Part of the radiocarbon calibration data set IntCal04 (Reimer et al., 2004), based on radiocarbon-dated tree rings from German pines, between 10500 and 9500 ^{14}C years BP. Each thin vertical line on the curve shows a ^{14}C age of a dated tree ring, and the length of the line is the ^{14}C dating error (1σ). The calibration curve is compared to a set of 20 ^{14}C dates on terrestrial macrofossils from sediments in three south Swedish lakes (Björck et al., 1996). The blue dates come from sediments formed just or slightly before the Younger Dryas–Holocene boundary, and the red dates are from sediments of very early Holocene age. The vertical lines display the radiocarbon error (1σ) and the horizontal lines show the calibration error (1σ). The dashed lines mark the 10050–10000 ^{14}C years plateau.

ascribed to production changes, but must be related to variations within the carbon cycle. Because of the large and sudden changes, it is very likely that the largest carbon reservoir – the ocean – was a key player for these variations. This is an additional indication that the Younger Dryas cooling was triggered by THC changes and a varying strength of NADW.

Younger Dryas Records Outside the Amphi-North Atlantic Region

In considering evidence for a Northern Hemisphere Younger Dryas oscillation it is necessary to focus on the huge North Pacific region. Several studies in this region have shown that climate variability during the last ice-age cycle seems to have been very similar to that reconstructed from terrestrial, ice core, and marine records in the circum-North Atlantic region. These studies constitute marine and lacustrine sediments, speleothems, and loess profiles, ranging from the west coast of North America/Central America to Japan and up onto the Chinese loess plateau. Although the chronologies for some of these records have often been too poor for direct correlations to the often better dated North Atlantic records, the climatic pattern seems very similar. It has, therefore, often been stated that a direct link exists between North Atlantic and North Pacific climates on millennial time scales. In fact, a Younger Dryas-like event seems to be the last of these glacial suborbital climate oscillations in the North Pacific region, but the exact timing has usually not been possible to determine. Although not very well dated, one of the first indications of a Younger Dryas event in the north-eastern Pacific region came from a lacustrine study in south-eastern Alaska (Engstrom et al., 1990), showing a pine parkland being replaced by tundra. However, the climatic ‘signature’ of this event is often very different on both sides of the North Pacific. While the Santa Barbara basin sediments became bioturbated (Kennett and Ingram, 1995) as an effect of locally increased ventilation, the Japan Sea records show a change to laminated sediments, possibly caused by increased sea ice cover/anoxic conditions. The Chinese speleothems and loess profiles suggest an oscillating strength of the East Asian monsoon system during the Last Termination, with a weakening of the Younger Dryas summer monsoon, explaining the more arid conditions during this time, and a strengthening of the winter monsoon with the increased dust load from central Asia. The latter is seen in the Greenland Summit ice cores. Uranium–thorium ages from Tangshan Cave in east China (Zhao et al., 2003) constrain the Younger Dryas fairly well (12500–11540 years BP), but indicate that the onset and end of the Younger Dryas dry event in East Asia may have been delayed in relation to the North Atlantic. There are, however, other less well-dated records in the desert–loess boundary area of northern China, implying more oscillating Younger Dryas conditions with tropical–polar interconnections (Zhou et al., 2001). At Lake Suigetsu in Japan (Nakagawa et al., 2003), the Last Termination in East Asia has probably been studied in the greatest detail. Its annually laminated sediments have been ^{14}C dated and pollen analyzed, and its pollen record has been quantified in terms of mean annual temperature. The reconstruction shows a temperature decline of 2–4 °C between 12300 and 11250 years BP, based on the varve chronology, which implies a substantial delay at

the onset and end of the Younger Dryas in relation to the North Atlantic.

The atmospheric ^{14}C signal is a global signal, with a few decades of lag for the Southern Hemisphere. It is therefore of interest to compare the set of ^{14}C dates at the onset and end of the Younger Dryas in Lake Suigetsu with corresponding ^{14}C dates in the North Atlantic region. As noted above, the very onset of the Younger Dryas is dated to 10 700–10 600 ^{14}C years BP on terrestrial macrofossils and tree rings in Europe, after a shift from ca. 11 000 ^{14}C years BP in less than 50 cal. years. The same shift is seen in the Lake Suigetsu ^{14}C record, but it predates Lake Suigetsu's Younger Dryas by a few hundred years. The Chinese and Japanese data, therefore, confirm each other: the Younger Dryas dry and cold event in East Asia lags the North Atlantic Younger Dryas cooling with at least 200–300 years. It is probable that the reason behind this may lie in the often suggested 'winter dominance' of the Younger Dryas climate, creatively evaluated and reviewed by (Denton et al., 2005). Studies show that late-lasting Eurasian snow cover will hamper the following summer monsoon by cooling down the land mass, decreasing the monsoon driving temperature gradient between the sea and the continent. The several hundred year-long transition into Younger Dryas, in Chinese and Japanese records, could imply that the Younger Dryas cold winter conditions in the North Atlantic region spread eastwards on the Eurasian continent resulting in gradually longer and cooler winters, increasing the intensity of the winter monsoon and gradually weakening the summer monsoon.

Similarly, if we compare the ^{14}C ages at the end of the Younger Dryas in Lake Suigetsu with the ^{14}C ages of the Younger Dryas–Preboreal transition around the North Atlantic and around 11 600 cal. years BP in the calibration data set, we can try to evaluate lags between the records. In European detailed ^{14}C records, for example (Björck et al., 1996), Late Younger Dryas ^{14}C ages are greater than 10 000 years BP and rather around 10 100 cal. years BP, while the early Preboreal ^{14}C ages are usually not older than 10 100 years BP (including 1σ), but rather around 10 000 years BP (Figure 2), a pattern which can be seen in the calibration data set before and after 11 600 cal. years BP. Because of the fairly large spread of the Lake Suigetsu ^{14}C dates at this time period a similar distinct pattern is hard to detect. However, it is impossible to shift the ^{14}C dates of the transition in Lake Suigetsu to more than 100–200 varve years older ages, and therefore the notion of a delay of the Holocene warming in Japan, and perhaps also in other parts of East Asia, seems very likely. The scenario may be the same as for the delayed onset of the Younger Dryas; the Eurasian continent was gradually warmed from the west and the propagation of the marine warming from the North Atlantic into the Pacific Ocean took a few centuries.

The presence of a Younger Dryas-like oscillation in the circum-North Atlantic and circum-North Pacific regions would imply that it also occurs in the region in between: North Africa and the Arabian Sea/northern Indian Ocean. Independent data sets show that northern equatorial Africa was dominated by arid conditions during the time of the Younger Dryas: lake levels fell in the west and east (Gasse, 2000), and mountain forests were reduced. These mostly arid conditions may have been related to decreased intensity of the summer monsoon, which is supported by evidence of decreased summer

upwelling along the Oman and Pakistan coasts during the Younger Dryas; summer upwelling strength is closely linked to the regional summer monsoon.

It is clear that the partial reorganization of the climate system during the Younger Dryas had a fairly substantial impact on the Northern Hemisphere climate, especially the winter conditions; both the atmospheric and marine circulation pattern changed, North Atlantic sea ice extended considerably, the albedo increased, monsoon intensity decreased, and atmospheric aerosols became more abundant, possibly most so in winter.

Is There Evidence for a Younger Dryas Oscillation in the Southern Hemisphere?

Although the Younger Dryas was most likely triggered by changes in the North Atlantic, it would be surprising if this event was not felt by the climate system of the Southern Hemisphere; the large impact of the Younger Dryas on the Northern Hemisphere climate ought to have been propagated, in one way or another, into some of the driving mechanisms for the Southern Hemisphere climate system.

South America is probably the continent where discussions about the presence or absence of a Younger Dryas cold event have been most lively. Most of these conflicting glacial, palynological, and paleoentomological records are from Chile and Argentina, but several records also exist from more northerly areas such as the central Andes and in the Amazon Basin (Clapperton, 1993). Paleoclimatic interpretations of these data sets and the chronology of the records seem to have been the two main obstacles for finding a common Younger Dryas signal. The hitherto best dated series of South American cooling events come from two sites at 41°S in Chile and Argentina (Hajdas et al., 2003) dated to ca. 11 500–10 200 ^{14}C years BP (ca. 13 400–12 000 cal. years BP). This implies that any South American equivalent to the Younger Dryas overlapped half of the European Allerød warm period and half of the Younger Dryas cooling. It should, however, also be remembered that the data sets analyzed and discussed, were formed during a climatically very dynamic time period. Large differences should, therefore, be expected between the Amazon Basin at the Equator and Tierra del Fuego in the south, the Atlantic coast in the east, and the high Andes in the west. The debate was also fueled by reports of well-dated so-called Younger Dryas glacial advances in New Zealand (Denton and Hendsy, 1994) with mean ^{14}C ages of 11 000 ^{14}C years BP. In light of the already existing knowledge that such ages predate the North Atlantic Younger Dryas cooling, it would have been more appropriate to regard these dates as evidence for 'Allerød glacial advances in the Southern Hemisphere.' Time is also needed to build-up glacier ice and form the moraines. The concept of pre-Younger Dryas glacial advances in New Zealand is now generally accepted (Shulmeister et al., 2005), and detailed dating of a climate reversal in Kaipo bog, with an age of 13 600–12 600 cal. years BP (Hajdas et al., 2006), shows that this cooling most likely triggered any glacial advances.

Since the 1980s, there have been indications that Antarctic ice cores show a cold spell, the Antarctic cold reversal (ACR), before the onset of the Holocene. However, the exact timing of this event has been uncertain due to difficulties with obtaining

sufficiently detailed chronologies for Antarctic ice cores. It was regarded by many as a strong indication for an Antarctic Younger Dryas and, as such, also a Southern Hemisphere one. However, when very detailed methane (CH_4) measurements of GRIP and the Antarctic Vostok and Byrd ice cores were produced (Blunier et al., 1997), it became obvious that Antarctic and Greenland climatic changes are asynchronous, especially during glacial times (Blunier et al., 1998). With respect to the comparison between the ACR and the Younger Dryas, the CH_4 correlations have shown that the ACR began around 13 800 ice years BP and ended at ca. 12 200 ice years BP in the Vostok ice core (Petit et al., 1999). Furthermore, the post-ACR rise in temperature occurred over ca. 600 years. This implies that the ACR started ca. 1000 years before the Younger Dryas started in the Northern Hemisphere, that is, at the beginning of the Allerød warming (GI-1c), and ended in the middle of the Younger Dryas (Figure 3). In fact, this type of climatic development seems consistent with many other observations in the Southern Hemisphere, such as the New Zealand climatic reversal/glacial advances, marine cores in the South Atlantic and Indian Ocean, and records south of the equator on the Atlantic side of Africa (Gasse, 2000). However, the Antarctic ice cores do not give us a completely consistent picture; the isotope stratigraphy of the Last Termination from the Taylor Dome ice core is more similar to the Greenland ice cores than are the other Antarctic cores (Steig et al., 1998). A warming peak is reached at ca. 14 000 ice years BP, which seems to be a normal feature in Antarctic ice cores, followed by a gradual cooling with minimum values at 12 900–12 300 ice years BP, corresponding chronologically to the first part of the Younger Dryas. At ca. 11 500 ice years BP a sudden Holocene warming occurred, which is in contrast to the long transition seen in the Vostok ice core. It should, however, be pointed out that these time differences may partly be explained by problems in the various ice chronologies. It is also possible that

different regions of Antarctica responded in different ways to the deglacial warming of the Last Termination. Most of the currently available data do, however, imply that the ACR began more than 1000 years before the onset of the Younger Dryas in the north, but not in direct 'antiphase' with the Bølling–Allerød (GI-1) warming in the north. They also imply that the start of the Holocene warming in most Antarctic ice cores leads the warming in the north with by at least 500 years.

The findings that warmings during glacial time in Antarctic ice cores roughly coincide with cold periods over the Greenland Summit, and vice versa, resulted in the hypothesis of a bipolar seesaw climate pattern (Broecker, 1998). To what degree the Younger Dryas and the ACR are part of this is difficult to judge, at least as long as the chronologies are incomplete. Even before this hypothesis was formulated, climate models had shown that a breakdown of the THC in the North Atlantic would result in a warming of the Southern Ocean. Later models differed on how much such a warming would spread in the Southern Hemisphere: to the entire hemisphere or only to the South Atlantic and southern Indian Ocean? Based on today's knowledge it is, however, tempting to conclude that (partial) breakdown of THC and NADW in the North Atlantic, as a result of sudden fresh-water discharges, resulted in a more or less extensive warming during Younger Dryas time in the Southern Hemisphere. In fact, it is also likely that large parts of the Southern Hemisphere experienced a cold period (ACR) before its entry into the Holocene warming, but clearly before the Northern Hemisphere Younger Dryas.

Conclusions

We can conclude that the present interglacial, the Holocene, was preceded by a distinct cool/dry event in the Northern

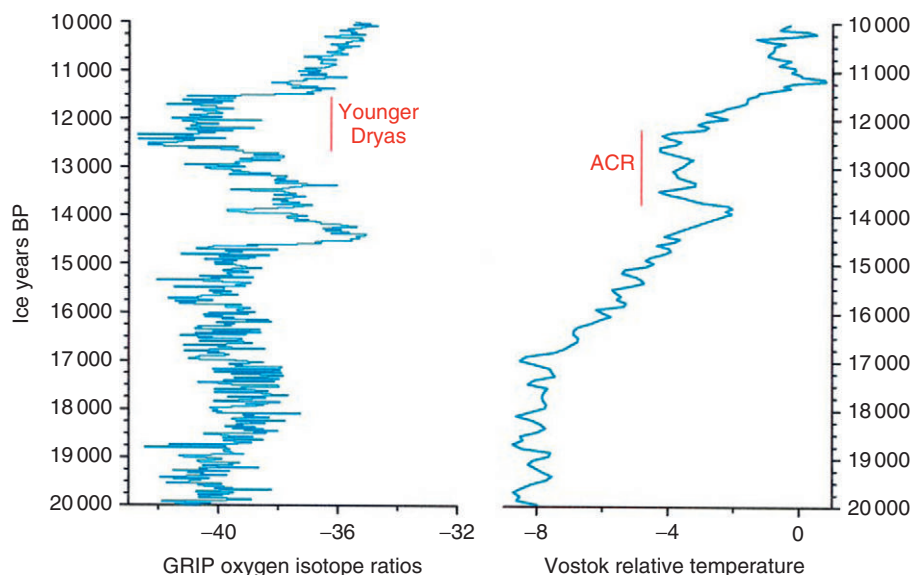


Figure 3 The GRIP oxygen isotope data (Johnsen et al., 2001) displaying temperature shifts over the Greenland Summit compared with the relative temperature record from the Vostok station in Antarctica (Petit et al., 1999), between 20 000 and 10 000 ice years BP. Ice years are approximately the same as calendar years. The positions of the Younger Dryas and the ACR are indicated. Note that the timescale of the GRIP data around Younger Dryas time is about 100 years too young when compared with the latest, but still unpublished chronology from the NorthGRIP core.

Hemisphere, generally designated the Younger Dryas cooling or the GS-1 event in the Greenland ice cores and dated to ca. 12 800–11 600 cal. years BP, and manifested by a winter-dominated climatic signature. In East Asia, there is good evidence that the onset and end of the event was slightly delayed in relation to its North Atlantic counterpart. It is likely that the event was triggered by rapid melt-water outbursts into the North Atlantic from the Laurentide and Fennoscandian ice sheets.

In large parts of the Southern Hemisphere there is clear evidence of a pre-Holocene cooling event. The timing and character of this event are, however, less clear. Most data indicate that it began 500–1000 years before its Northern Hemisphere equivalent and ended at least 400 years before its termination around the North Atlantic. This means that it overlaps most, or at least the later part, of the Allerød warm period (GI-1c–1a) and at least the beginning of the Younger Dryas (GS-1) in the north. It also shows that the Holocene warming began much earlier in the Southern Hemisphere, during the peak of the northern Younger Dryas cooling. It is difficult to judge to what extent this pattern is part of a bipolar seesaw, but the transitional overlap between northern–southern warming and cooling ([Figure 3](#)) is reminiscent of the interhemispheric phase lags during the Dansgaard/Oeschger cycles of glacial time. A key question is of course: which hemisphere leads? With respect to the Younger Dryas oscillation, it is likely that a partial shutdown of the Atlantic conveyor belt decreased northern THC, which led to a warming in the Southern Ocean, explaining the early onset of the interglacial warming in the south. If we postulate that fresh-water forcing in one hemisphere is the main forcing mechanism for warmings in the other hemisphere, the Northern Hemisphere Heinrich-1 event could have triggered the onset of warming in Antarctica at 17 000 ice years BP, and if melt-water peak-1A originated from Antarctica ([Clark et al., 1996](#)), it would imply that it triggered the Bølling warming ([Weaver et al., 2003](#)). Thus, the bipolar seesaw pattern may be very complex, especially during phases of deglaciation, and the Younger Dryas oscillation is certainly an important and globally significant part of the complexities of the Last Termination.

See also: Pollen Analysis, Principles. **Ice Core Methods:** Overview. **Ice Cores:** History of Research, Greenland and Antarctica. **Paleoclimate:** The Younger Dryas Climate Event. **Radiocarbon Dating:** Conventional Method.

References

- Alley RB (2000) The Younger Dryas cold interval as viewed from central Greenland. *Quaternary Science Reviews* 19: 213–226.
- Ammann B and Lotter AF (1989) Late-Glacial radiocarbon stratigraphy and palynostratigraphy on the Swiss Plateau. *Boreas* 18: 109–126.
- Andersson G (1896) *Svenska växtvärldens historia*. Stockholm: P.A. Norstedt & Söner.
- Björck S, Kromer B, Johnsen S, et al. (1996) Synchronised terrestrial–atmospheric deglacial records around the North Atlantic. *Science* 274: 1155–1160.
- Björck S, Walker MJC, Cwynar LC, et al. (1998) An event stratigraphy for the Last Termination in the North Atlantic region based on the Greenland ice-core record: A proposal by the INTIMATE group. *Journal of Quaternary Science* 13: 283–292.
- Blunier T, Schwander J, Stauffer B, et al. (1997) Timing of the Antarctic cold reversal and the atmospheric CO₂ increase with respect to the Younger Dryas event. *Geophysical Research Letters* 24: 2683–2686.
- Blunier T, Chappellaz J, Schwander J, et al. (1998) Asynchrony of Antarctic and Greenland climate change during the last glacial period. *Nature* 394: 739–743.
- Boyle EA and Keigwin LD (1987) North Atlantic thermohaline circulation during the past 20,000 years linked to high-latitude surface temperatures. *Nature* 330: 35–40.
- Broecker WS (1994) Massive iceberg discharges as triggers for global climate change. *Nature* 372: 421–424.
- Broecker WS (1998) Paleocirculation during the last deglaciation: A bipolar seesaw? *Paleoceanography* 13: 119–121.
- Broecker WS, Andree M, Wolfli W, et al. (1988) The chronology of the last deglaciation: Implications to the cause of the Younger Dryas event. *Paleoceanography* 3: 1–19.
- Clapperton C (1993) *Quaternary Geology and Geomorphology of South America*. Amsterdam: Elsevier.
- Clark PU, Alley RB, Keigwin LD, et al. (1996) Origin of the first global meltwater pulse following the last glacial maximum. *Paleoceanography* 11: 563–577.
- Denton GH and Hendy CH (1994) Younger Dryas age advance of Franz-Josef Glacier in the southern Alps of New Zealand. *Science* 264: 1434–1437.
- Denton GH, Alley RB, Comer GC, and Broecker WS (2005) The role of seasonality in abrupt climate change. *Quaternary Science Reviews* 24: 1159–1182.
- Engstrom DR, Hansen BC, and Wright HE Jr. (1990) A possible Younger Dryas record in Southeastern Alaska. *Science* 250: 1383–1385.
- Gasse F (2000) Hydrological changes in the African tropics since the last glacial maximum. *Quaternary Science Reviews* 19: 189–211.
- Hajdas I, Bonania G, Moreno PI, and Ariztegui D (2003) Precise radiocarbon dating of Late-Glacial cooling in mid-latitude South America. *Quaternary Research* 59: 70–78.
- Hajdas I, Lowe DJ, Newnham RM, and Bonani G (2006) Timing of the late-glacial climate reversal in the Southern Hemisphere: A high-precision radiocarbon chronology for Kaipo bog, New Zealand. *Quaternary Research* 65: 340–345.
- Hartz N and Milthers V (1901) Det senglacielle i Allerød tegelvægsgrav. *Meddelelser fra Dansk Geologisk Forening* 8: 31–60.
- Hughen KA, Overpeck JT, Peterson LC, and Trumbore S (1996) Rapid climate changes in the tropical Atlantic region during the last deglaciation. *Nature* 380: 51–54.
- Johnsen SJ, Dansgaard W, Clausen HB, and Langway Jun CC (1972) Oxygen isotope profiles through Antarctic and Greenland ice sheets. *Nature* 235: 429–434.
- Johnsen SJ, Dahl-Jensen D, Gundestrup N, et al. (2001) Oxygen isotope and palaeotemperature records from six Greenland ice-core stations: Camp Century, Dye-3, GRIP, GISP2, Renland and NorthGRIP. *Journal of Quaternary Science* 16: 299–307.
- Kennett JP and Ingram BL (1995) A 20,000 year record of ocean circulation and climate-change from the Santa-Barbara Basin. *Nature* 377: 510–514.
- Lowe JJ, Ammann B, Birks HH, et al. (1994) Climatic changes in areas adjacent to the North Atlantic during the last glacial–interglacial transition (14–9 ka BP) – A contribution to IGCP-253. *Journal of Quaternary Science* 9: 185–198.
- Manabe S and Stouffer RJ (1988) Two stable equilibria of a coupled ocean–atmosphere model. *Journal of Climate* 1: 841–866.
- Mangerud J, Andersen ST, Berglund BE, and Donner JJ (1974) Quaternary stratigraphy of Norden, a proposal for terminology and classification. *Boreas* 3: 109–128.
- Mott RJ, Grant DR, Stea R, and Occhietti S (1986) Late-glacial climate oscillation in Atlantic Canada equivalent to the Allerød/Younger Dryas event. *Nature* 323: 247–250.
- Nakagawa T, Kitagawa H, Yasuda Y, et al. (2003) Asynchronous climate changes in the North Atlantic and Japan during the last termination. *Science* 299: 688–691.
- Peteet DM (1987) Younger Dryas in North America – Modeling, data analysis, and re-evaluation. In: Berger WL (ed.) *Abrupt Climatic Change: Evidence and Implications*, pp. 185–193. Dordrecht: Reidel.
- Petit JR, Jouzel J, Raynaud D, et al. (1999) Climate and atmospheric history of the past 420,000 years from the Vostok ice core, Antarctica. *Nature* 399: 429–436.
- Reimer PJ, Baillie MGL, Bard E, et al. (2004) IntCal04 terrestrial radiocarbon age calibration, 0–26 cal kyr BP. *Radiocarbon* 46: 1029–1058.
- Ruddiman WF and McIntyre A (1981) The North Atlantic Ocean during the last deglaciation. *Paleogeography, Paleoclimatology, Paleocology* 35: 145–214.
- Shulmeister J, Fink D, and Augustinus P (2005) A cosmogenic nuclide chronology of the last glacial transition in North-West Nelson, New Zealand – New insights in Southern Hemisphere climate forcing during the last deglaciation. *Earth and Planetary Science Letters* 233: 455–466.

- Steig EJ, Brook EJ, White JWC, et al. (1998) Synchronous climate changes in Antarctica and the North Atlantic. *Science* 282: 92–95.
- Teller JT, Boyd M, Yanga Z, Korc PSG, and Fardd AM (2005) Alternative routing of Lake Agassiz overflow during the Younger Dryas: New dates, paleotopography, and a re-evaluation. *Quaternary Science Reviews* 24: 1890–1905.
- Weaver AJ, Saenko OA, Clark PU, and Mitrovica JX (2003) Meltwater pulse 1A from Antarctica as a trigger of the Bølling–Allerød warm interval. *Science* 299: 1709–1713.
- Wright HE Jr. (1989) The amphi-Atlantic distribution of the Younger Dryas palaeoclimatic oscillation. *Quaternary Science Reviews* 8: 295–306.
- Zhao J, Wang Y, Collerson KD, and Gagan MK (2003) Speleothem U-series of semi-synchronous climate oscillations during the last deglaciation. *Earth and Planetary Science Letters* 216: 155–161.
- Zhou W, Head M, An Z, et al. (2001) Terrestrial evidence for a spatial structure of tropical–polar interconnections during the Younger Dryas episode. *Earth and Planetary Science Letters* 191: 231–239.

The Last Millennium

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The period of roughly the last millennium is particularly important for characterizing the natural variability of the climate and, in so doing, framing possible, more recent anthropogenic climate changes. Numerous natural archives or ‘proxy’ records, such as tree rings, corals, ice cores, lake sediments, and historical documents, are available to aid in reconstructing the details of climate changes within this time frame (Bradley, 1999; Bradley and Jones, 1995; Fritts, 1976; Huang et al., 2000; Jacoby and D’Arrigo, 1989; Jones and Mann, 2004; Jones et al., 2001; Luterbacher et al., 2004; Mann et al., 1998; Wigley et al., 1981). Moreover, the basic external constraints on the climate system (the configuration of the continents, the extent of any continental ice sheets, and the geometry of Earth’s orbit) have not changed appreciably on the millennial timescale. For this reason, the variability of the climate over the period of the past one or two millennia, but prior to the past two centuries during which anthropogenic impacts are likely to be a dominant factor, can provide insights into the envelope of natural variability of the climate. These insights can, in turn, guide our assessment of the extent to which recent climate changes are anomalous and the extent to which these changes are likely related to nonnatural (i.e., anthropogenic) impacts on climate.

A considerable body of scientific research in recent decades has sought to characterize the nature of climate variability over the past millennium (Bradley and Jones, 1995; Briffa et al., 2001). Numerous groups have used proxy data to reconstruct this history of large-scale changes in temperature, precipitation and drought, and atmospheric circulation in past centuries. These reconstructions place in a richer and significantly more detailed context previous notions of climate change embodied by terms such as the ‘Little Ice Age’ and ‘Medieval Warm Period.’ In particular, much of the climate variation in past centuries is characterized by a complex pattern of regional and seasonal variation that is difficult to describe using simple, broad labels. Some of the past observed changes appear to be consistent with changes in radiative forcing of climate (e.g., due to the radiative impacts of explosive volcanic eruptions and inferred changes in solar output), whereas other changes are probably associated with internal, chaotic oscillations of the climate system. At hemispheric or global scales, late twentieth-century climate changes appear to be anomalous in the context of at least the past millennium. Simulations with theoretical climate models support the conclusion that these changes are likely associated, in large part, with anthropogenic impacts on the climate.

Large-Scale Temperature Trends

Numerous research groups have used networks of climate proxy data to reconstruct large-scale surface temperatures in past centuries at seasonal and annual resolution. These reconstructions emphasize both spatial patterns of past temperature

change and average changes over the globe and both hemispheres. Due to the greater number of available climate proxy data (Figure 1), these reconstructions have generally emphasized temperature changes in the Northern Hemisphere and over a period extending as far back in time as the past two millennia (Jones and Mann, 2004; Moberg et al., 2005; Oerlemans, 2005; Rutherford et al., 2005). Most reconstructions indicate generally cooler conditions during the fifteenth through nineteenth centuries, which includes the intervals usually used to define the Little Ice Age. Warmer conditions are found during the eleventh through the fourteenth centuries, which includes the intervals typically used to define the so-called Medieval Warm Period (Folland et al., 2001).

The reconstructions suggest anomalous late twentieth-century warmth at hemispheric scales (Figure 2). Uncertainties are greater, due to the paucity of Southern Hemisphere proxy data, for the Southern Hemisphere and globe, but it is also likely that the late twentieth-century warmth is unprecedented in these contexts as well (Jones and Mann, 2004). These observations are consistent with evidence that radiative forcing has been considerably enhanced by anthropogenic influences during the twentieth century.

Regional Climate Changes

Surface Temperature

In interpreting past temperature changes, one must carefully distinguish regional temperature changes from large-scale, global, or hemispheric average temperature changes. Large-scale average temperature estimates must be based on the average of temperature estimates over a sufficiently large number of distinct regions and on a contemporaneous basis. Drawing conclusions from any one or small set of regions at any particular time is perilous because relative temperature variations on interannual, decadal, and even millennial timescales are often quite different (Mann et al., 2003). Temperature variations in continental regions are often substantially greater than those in oceanic regions, and they are often different during different seasons (e.g., summer vs. winter). Indeed, relative temperature changes are often observed to be opposite in different locations. This behavior is believed to be due to a number of factors, including long-term changes in the characteristics of the prevailing atmospheric and oceanic circulation. Cyclonic storm systems, for example, typically cause regions in one sector of the storm to warm while causing regions in other sectors of the storm to cool. Increases in the strength of a warm poleward ocean current lead to warming in higher latitude regions at the expense of cooling in lower latitude regions. Moreover, changes over time in these behaviors of the ocean and atmosphere are typically regional, rather than global, in their influence.

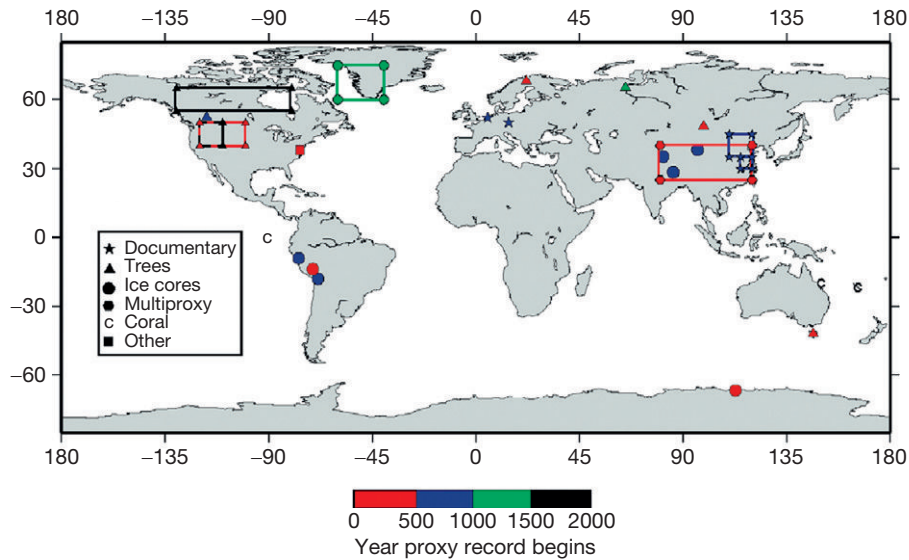


Figure 1 Map of available high-resolution (annual or decadal) temperature proxy indicators available over much of the past two millennia. The boxes represent the individual locations used in series that represent regional composites. Reproduced from Jones PD and Mann ME (2004) *Climate over past millennia*. *Reviews of Geophysics* 42: RG2002. Copyright AGU.

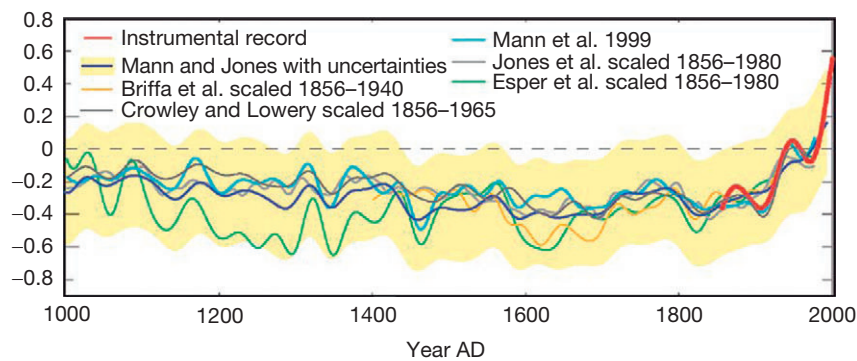


Figure 2 Reconstructions of northern annual temperatures over the past millennium. Shown for comparison is the instrumental Northern Hemisphere record 1856–2003. Also shown is the 95% uncertainty band (shading) for one of the reconstructions. Adapted from Jones PD and Mann ME (2004) *Climate over past millennia*. *Reviews of Geophysics* 42: RG2002.

Considerable spatial variability is observed in the temperature variations over the past millennium. Although the Little Ice Age is observed to have been a relatively cool period for certain regions such as Europe, it clearly was not uniformly cold for all regions. Similarly, the so-called Medieval Warm Period of the ninth through the thirteenth centuries was in fact a cold interval for many regions, despite the fact that it was a warm period for Europe. The notions of the Little Ice Age and Medieval Warm Period as global climate episodes is, to some extent, an artifact of the Eurocentric origins of these terms. Although relative hemispheric warmth during the tenth through the twelfth centuries and cool conditions during the fifteenth through the early twentieth century are indeed evident for hemispheric mean temperature (Figure 1), the changes are small (only a few tenths of a degree C) compared to the larger regional changes (often a degree or more). The specific periods of coldness and warmth differ from region to

region (Figure 3) from those for the Northern Hemisphere as a whole. Temperature changes appear to have been different for continents and oceans and for different seasons. The greatest episodes of cooling during the seventeenth through the nineteenth centuries appear to have occurred over the continents and during summer. Studies suggest that this may be related to the influence of explosive volcanism on climate.

Precipitation and Drought

Changes in hydrological climate variables such as precipitation and drought are equally, if not more, important from a societal standpoint than are past changes in surface temperature. A considerable body of recent work has sought to use paleoclimate proxy indicators, such as tree rings, corals, ice cores, and lake sediments, combined with long historical documentary evidence to document these past changes.

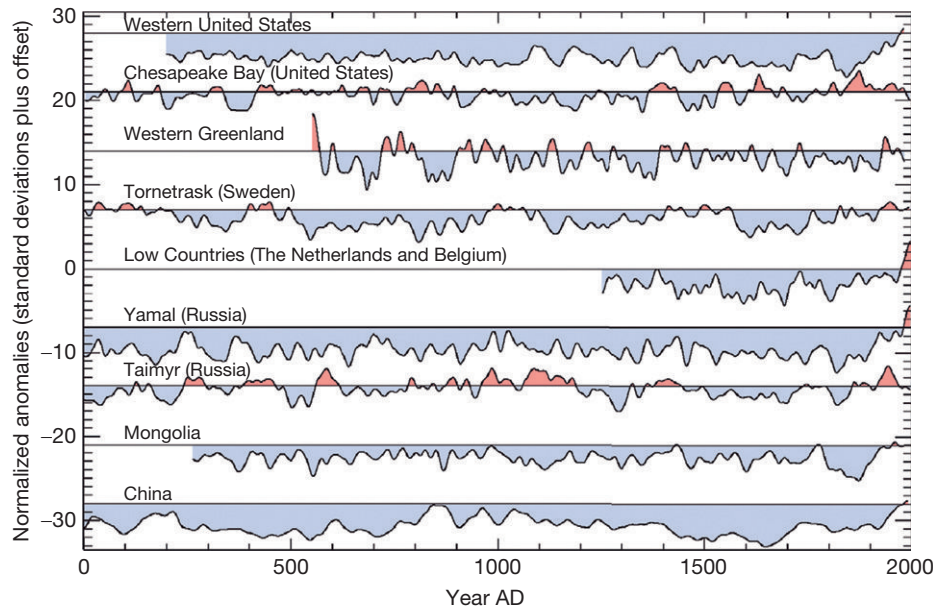


Figure 3 Temporal histories of nine temperature-sensitive proxy records, chosen to illustrate a variety of proxy types, regions, and seasonal windows. All series have been smoothed with a 40-year low-pass filter, then normalized so that the filtered series have unit standard deviation. Reproduced from Mann ME, Ammann CM, Bradley RS, et al. (2003) On past temperatures and anomalous late 20th century warmth. *EOS* 84: 256–258. Copyright AGU.

Much of this work has emphasized changes in North American drought patterns in past centuries. Tree-ring reconstructions suggest that the range of North American drought variability observed during the twentieth century may not fully represent the range of drought evident in earlier centuries. The reconstructions also suggest that the twentieth-century dust bowl still stands out as the most extreme summer drought in large areas of the United States in recent centuries. A sixteenth-century ‘megadrought’ appears to have been important over the United States, western Canada, and northwestern Mexico and may have far exceeded the twentieth-century dust bowl event in scale and magnitude. For the western United States on the whole (Cook et al., 2004), the earlier centuries of the past millennium were relatively dry and later centuries relatively wet (Figure 4). Reconstructed lake levels in equatorial east Africa (Kenya) suggest a similar pattern of dry conditions during the earlier part of the millennium (eleventh through thirteenth centuries) and wet conditions during the later centuries (e.g., fourteenth through nineteenth centuries), with peak wet conditions during the mid-seventeenth to the mid-eighteenth centuries (Verschuren et al., 2000). These long-term changes in drought may be tied to changes in the El Niño/Southern Oscillation (ENSO) phenomenon, as discussed in more detail later. The long-term reconstructions suggest that drought in the western United State during the past decade may be approaching in magnitude the most severe droughts of the past millennium.

Changes in precipitation patterns in Europe in past centuries, including dry conditions in central Europe and the European Alps during the late seventeenth and early eighteenth centuries, may reflect changes in the large-scale atmospheric circulation (e.g., the North Atlantic Oscillation or ‘NAO’ phenomenon) influencing the region. Independent support for this conclusion is obtained from the pattern of mountain glacier expansion and retreat in both northern and southern Europe in past centuries.

Changes in Atmospheric and Oceanic Circulation

At regional scales, climate variability and change in past centuries is closely associated with the behavior of particular modes of climate variability, such as NAO and ENSO. These modes describe the characteristic patterns of reorganization of the atmospheric and oceanic circulations over time. Changes in NAO and ENSO are especially relevant for understanding the complex patterns of climate change associated with the so-called Little Ice Age and Medieval Warm Period.

Proxy climate indicators do not directly measure past changes in atmospheric circulation. Nonetheless, proxy measures of past temperature and precipitation changes are often used to indirectly infer past changes in atmospheric circulation and climate indices.

El Niño/Southern Oscillation

Indices of ENSO such as the Southern Oscillation Index (Figure 5) have been reconstructed based on tree rings in ENSO-sensitive regions and coral data from the tropical Pacific Ocean (Evans et al., 2002; Stahle et al., 1998). In addition, qualitative evidence regarding the history of El Niño events in past centuries is available from long-term historical documents in regions influenced by El Niño (e.g., tropical South America). The various reconstructions show important features in common. The large 1982–83 and 1997–98 El Niño events, for example, seem to be outside the range of variation seen over the past few centuries. The uncertainties in the reconstructions are still substantial. A more extensive network of ENSO-sensitive proxy indicators will likely be required for more confident conclusions regarding the behavior of ENSO in past centuries.

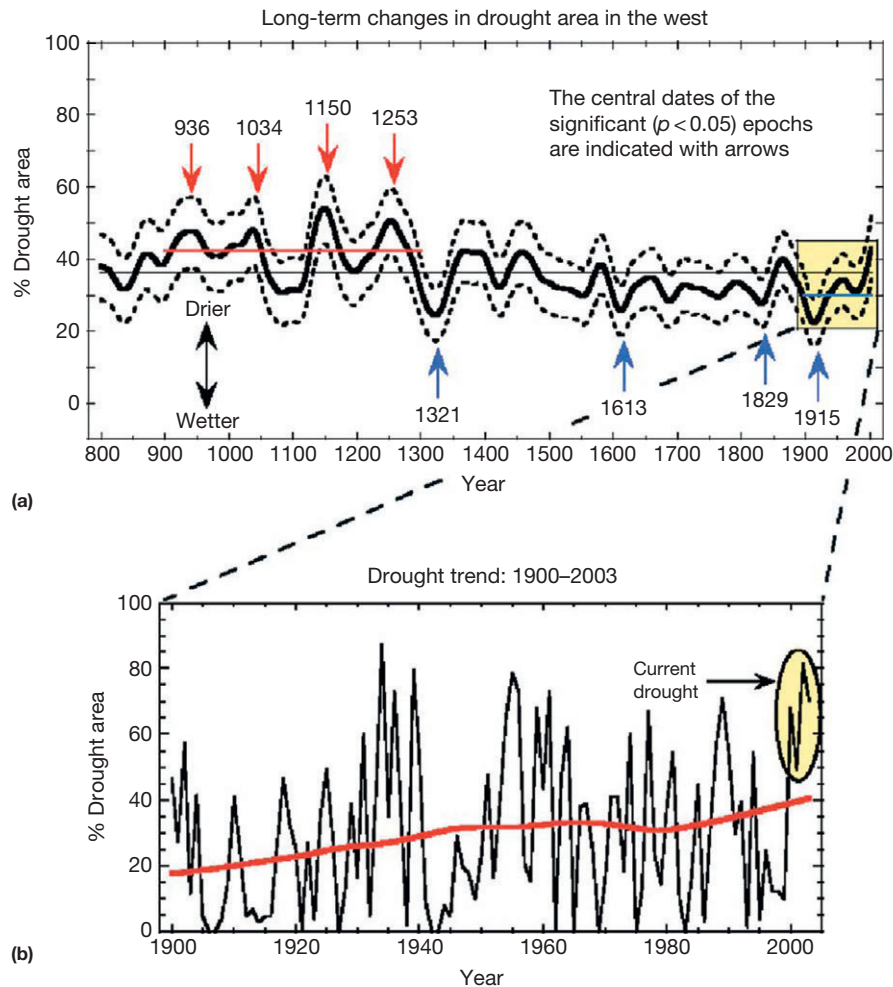


Figure 4 Reconstruction of Drought Area Index from tree-ring data, indicative of changes in the extent of drought over the western United States during the past millennium. Reprinted from Cook ER, Woodhouse C, Meko DM, and Stahle DW (2004) Long-term aridity changes in the western United States. *Science* 306: 1015–1018, with permission from the American Association for the Advancement of Science (AAAS).

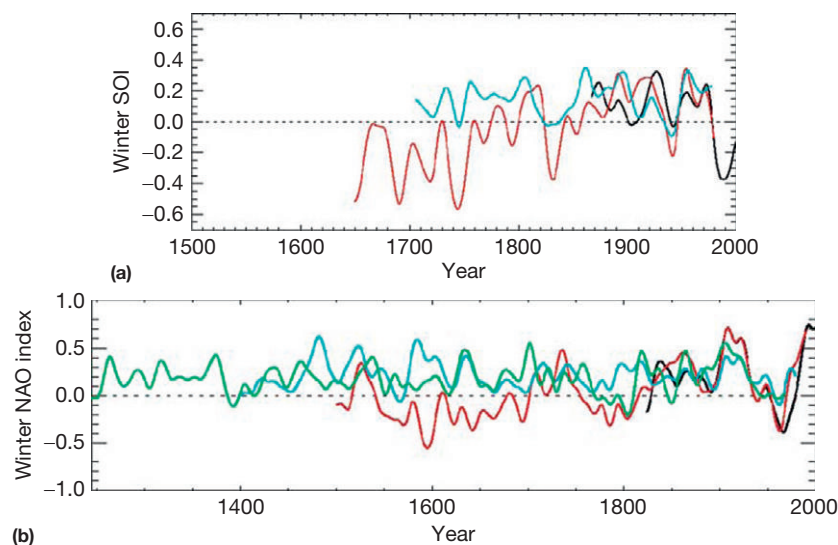


Figure 5 Proxy-based reconstructions of two key atmospheric indices: (a) the Southern Oscillation Index (SOI) and (b) the North Atlantic Oscillation (NAO). Series are smoothed to emphasize variations on 30-year and longer timescales. Reproduced from Jones PD and Mann ME (2004) Climate over past millennia. *Reviews of Geophysics* 42: RG2002. Copyright AGU.

Fossil coral evidence, although incomplete, allows insights into changes in both the interannual variability and the mean state of an ENSO-dominated region of the tropical Pacific over roughly the past millennium (Cobb et al., 2003). Such evidence suggests La Niña-like conditions in the earlier centuries of the past millennium (the tenth through the thirteenth centuries) and El Niño-like conditions later on (e.g., the mid-seventeenth century through the mid-eighteenth century). Such conditions are consistent with the previously discussed evidence for dry and wet conditions, respectively, in equatorial east Africa during these same time intervals. Some studies suggest a relative absence of interannual variability, but a persistence of decadal variability, during the early and mid-nineteenth century. Non-stationarity is evident on multidecadal timescales in the extra-tropical influences of ENSO during past centuries.

North Atlantic Oscillation

A number of studies have sought to reconstruct the boreal winter or cold season NAO (or the related Arctic Oscillation or AO index) in past centuries (Luterbacher et al., 1999). These reconstructions are based on composites of proxy data or combinations of proxy, historical, and instrumental data, calibrated against the modern instrumental NAO index. Most of the NAO reconstructions suggest the trend in recent decades toward an enhanced positive NAO index to be relatively unusual, if not unprecedented, in recent past centuries (see Figure 5). This conclusion seems consistent with insights gained from model studies with regard to both the natural variability of the NAO and the potential response of the NAO to anthropogenic forcing of the climate. As with the reconstruction of El Niño indices discussed previously, sizable uncertainties in the reconstructions still preclude any definitive conclusions. Preliminary reconstructions of the summer AO index have also been attempted. These reconstructions suggest different behavior in past centuries from the winter NAO index.

The NAO pattern of atmospheric variability may have played a prominent role in explaining regional patterns of surface temperature change in past centuries in the extratropical Northern Hemisphere. For example, a tendency toward the negative phase of the NAO seems to explain the enhanced regional winter cooling in Europe during the seventeenth century. Similarly, a tendency toward the positive phase of the NAO may explain the enhanced regional warming in Europe during the medieval times. Such conclusions are consistent with ice borehole data from Greenland and from geochemical and geobiological evidence from North Atlantic and Mediterranean sediment cores. Forward modeling of glacial mass balance in Europe in past centuries suggests that the inverse relationship between glacier expansion and recession in northern and southern/central Europe in past centuries is consistent with NAO impacts on winter precipitation. Long-term evidence from ocean and lake sediments suggests that the NAO has played a prominent role in the climate variability of the North Atlantic and neighboring regions over several millennia.

Other Modes

A number of other climate indices have been reconstructed in past centuries from proxy data. These reconstructions are

typically more preliminary, and even more uncertain in nature, than reconstructions of the NAO and ENSO indices. One such index, the Pacific Decadal Oscillation (PDO) index, describes climate variability in the North Pacific over the past several centuries (Minobe, 1997). The behavior of the PDO has been reconstructed from tree-ring data and ice cores from maritime glaciers in the Pacific Northwest of the North American continent. The various reconstructions of this index suggest robust decadal variability over the past several centuries impacting the climate of the North Pacific and the North American continent. In certain cases, the reconstructions indicate a substantial positive trend in the PDO since the mid-nineteenth century.

Another climate index, the Atlantic Multidecadal Oscillation (AMO), describes a roughly 50- to 70-year pattern of variation in sea surface temperature over the North Atlantic Ocean, with larger scale impacts on atmospheric temperature and circulation over a substantial portion of the extratropical Northern Hemisphere. The behavior of this index in past centuries has been reconstructed from a number of different sources of proxy information (Delworth and Mann, 2000). Model simulations produce a mode of variability similar to the observed AMO in both its spatial pattern and its dominant (multidecadal) timescale of variation. The simulations suggest that the origins of the variability are in coupled ocean-atmosphere processes that interact with the thermohaline circulation of the North Atlantic Ocean on multidecadal timescales (Delworth and Mann, 2000). The reconstructions of the AMO indicate substantial multidecadal variability in North Atlantic climate over several centuries. The twentieth-century oscillations of the AMO have been implicated in the irregularity of the pattern of warming in the Northern Hemisphere during the twentieth century and in some changes over time in the frequency and intensity of Atlantic tropical storms and hurricanes.

Reconstructions from proxy data of the Southern Hemisphere counterpart to the AO index, the Antarctic Oscillation index, have also been attempted in recent years. Due to the paucity of high-resolution Southern Hemisphere proxy data, however, such reconstructions are necessarily preliminary and highly uncertain in nature.

Theoretical Modeling Results

A number of simulations of the climate over the past millennium or longer have been performed using a variety of theoretical climate models and empirical estimates of past natural and anthropogenic radiative ‘forcing’ histories. The natural forcing generally includes changes in solar radiative output and dust aerosols due to explosive volcanic forcing, whereas the more recent anthropogenic radiative forcing generally includes a combination of human increases in greenhouse gas concentrations since the eighteenth century and industrial aerosols produced largely during the twentieth century. In some simulations, the more secondary anthropogenic impacts during the nineteenth and twentieth centuries associated with human land use changes are also included. Temperature changes over the past millennium can be diagnosed directly from the simulation output. These theoretical estimates of

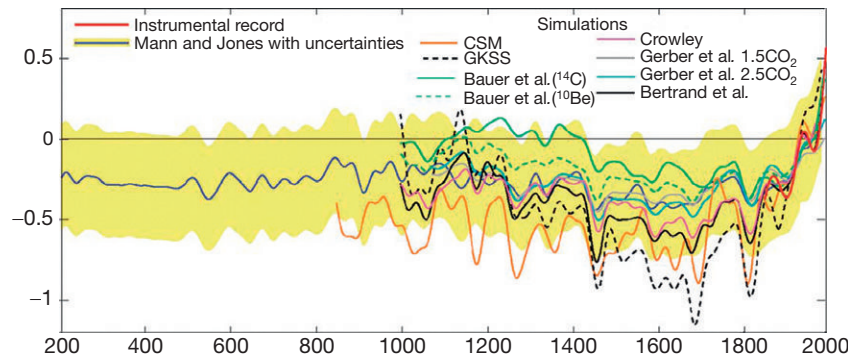


Figure 6 Model-based estimates of Northern Hemisphere temperature variations over the past two millennia. Series have been smoothed to highlight variations on timescales greater than 40 years. The simulations employ a range of climate models with differing sensitivities to radiative forcing and employ various different estimates of radiative forcing histories, including natural (solar + volcanic) and modern anthropogenic (greenhouse gas and sulfur aerosol) impacts. Shown for comparison is the instrumental Northern Hemisphere record for 1856–2003 and an empirical proxy-based estimate of Northern Hemisphere mean temperature changes with its 95% uncertainty band (shading). Adapted from Jones PD and Mann ME (2004) *Climate over past millennia. Reviews of Geophysics* 42: RG2002.

past temperature changes are entirely independent of the proxy-based, empirical estimates. Comparisons between the empirical and the modeled temperature histories can thus greatly inform our interpretation of the factors governing the climate changes of the past millennium. Comparisons of the empirical Northern Hemisphere temperature reconstructions with the modeled Northern Hemisphere mean temperature changes (Figure 6) demonstrate that natural factors appear to explain the major reconstructed surface temperature changes of the past millennium, prior to the twentieth century (Crowley, 2000; Robock, 2000).

Only anthropogenic forcing of climate, however, can explain the recent anomalous large-scale warming in the late twentieth century (Crowley, 2000; Crowley et al., 2003). The simulations reproduce the anomalous late twentieth-century hemispheric warmth and indicate that it can only be explained by anthropogenic (greenhouse gas combined with sulfate aerosol) forcing. Moreover, the natural variability produced by the models is generally consistent with the long-term proxy-based reconstructions within published uncertainties. Some simulations overpredict the observed warming since the early to mid-nineteenth century, apparently due to the absence of important forcings (e.g., anthropogenic aerosols or land use changes). These simulations typically appear too cold during the nineteenth century, relative to the instrumental record and the reconstructions, when they are aligned to agree with the instrumental record during the twentieth century (as in Figure 6). Simulations that take into account nineteenth- and twentieth-century land use changes match the observations more closely, suggesting the importance of including land use changes as a significant, albeit secondary, nineteenth- and twentieth-century external radiative forcing. One model simulation ('GKSS'; see Figure 6) suggests especially large-amplitude variations in past centuries, relative to other simulations and reconstructions, but it appears to be highly unrealistic. The larger variability displayed by this simulation appears to result from a combination of a very large assumed solar forcing amplitude and the absence of twentieth-century sulfate aerosol forcing, and an apparent error in the way the model was

initialized. The initialization error leads to a long-term drift in temperature that is unrealistic in nature.

It is in some respects more informative to compare the spatial patterns of the reconstructed and modeled changes rather than just the simple hemispheric means. An analysis of the spatial patterns of past temperature changes, for example, can help elucidate the role of dynamical responses of the climate related to factors such as NAO and ENSO. The substantial cooling in large areas of Europe during the late seventeenth and early eighteenth centuries appears to be related to long-term variations in NAO that are themselves forced by a combination of changes in solar radiative output and explosive volcanism (Figure 7).

These changes are associated with a large-scale dynamical response of the climate system to natural radiative forcing by explosive volcanic activity and solar output that interacts with the NAO pattern of atmospheric circulation. The moderate apparent lowering of solar irradiance during the seventeenth century led to only moderate decreases in the hemispheric mean temperature (see Figure 6). However, seasonally reinforced patterns of response to this forcing, which include substantial continental summer cooling due to the lesser thermal inertia of land relative to ocean and winter cooling over most land regions due to an induced shift toward the negative phase of the NAO, lead to a tendency for strong annual mean cooling in some regions such as Europe (Schmidt et al., 2004; Shindell et al., 2001). Conversely, some land regions (e.g., northeastern North America) are observed to warm due to the changes in atmospheric circulation associated with the negative phase of the NAO that lead to flow of warmer maritime air regions during times of decreased solar irradiance.

Past volcanic forcing leads to larger annual hemispheric mean temperature changes than does past solar forcing due to the substantially greater associated radiative forcing changes. However, in the case of volcanic forcing, a tendency for seasonally opposite direct radiative and dynamical (i.e., NAO) responses leads to offsetting seasonal temperature changes over extratropical continental regions (cooling and warming, respectively), leading to smaller regional annual mean changes

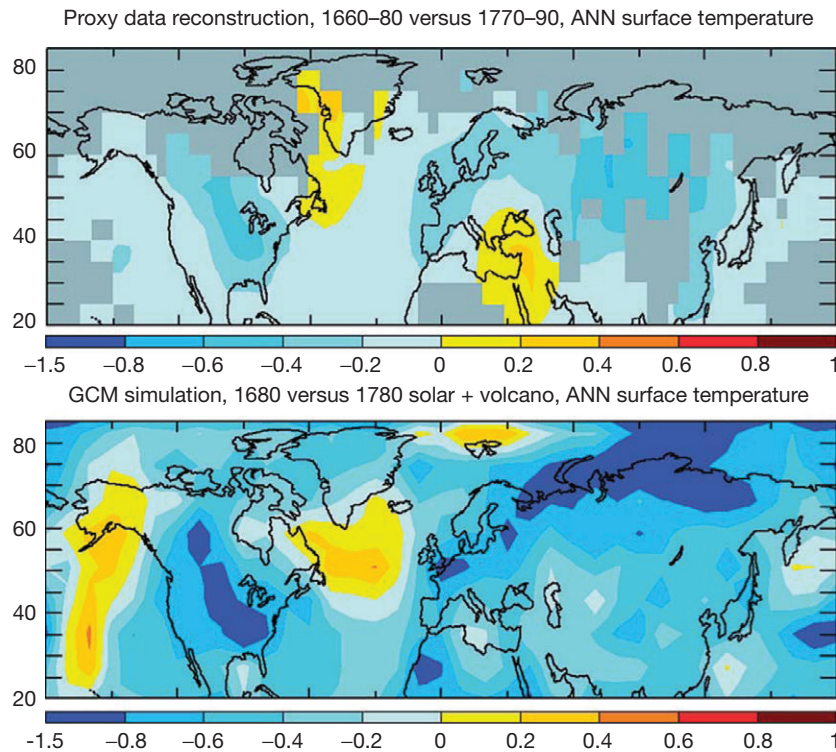


Figure 7 Reconstructed and (bottom) simulated annual average temperature difference between 1660–80 and 1770–90. The reconstructed surface temperatures are based on a multiproxy estimate using tree rings, ice cores, corals, and historical data. Model results are based on the sum of the response in two simulations: one incorporating reconstructed solar irradiance changes during this period and one using volcanic forcing scaled to changes over this time. Reproduced from Schmidt GA, Shindell DT, Miller RL, Mann ME, and Rind D (2004) General circulation modeling of Holocene climate variability. *Quaternary Science Reviews* 23: 2167–2181.

than in the case of solar forcing. The predicted pattern of response by a climate model to the lowering of solar irradiance combined with the response to explosive volcanism matches the empirical, reconstructed pattern of temperature change between the late seventeenth and late eighteenth century (Figure 7). The seasonally opposite responses to volcanic forcing over the extratropical continents explain why some Northern Hemisphere average temperature estimates (Figure 6) emphasizing the summer or warm season and the extratropical continents show greater variation in past centuries than other reconstructions emphasizing the full hemisphere and all seasons.

Recent work suggests that ENSO may also be an important component of the response of the climate to forcing during the past 1000 years. Model simulations using simple models of the coupled tropical ocean–atmosphere to study the response of ENSO to solar and volcanic forcing (Mann et al., 2005) yield results consistent with the proxy evidence for past changes in ENSO discussed previously. The model simulations indicate a counterintuitive tendency toward El Niño (warm eastern and central tropical Pacific) conditions in response to negative radiative forcing (past explosive tropical volcanic eruptions or decreases in solar irradiance) and a tendency for La Niña-like conditions in response to positive radiative forcing (i.e., increases in solar irradiance). This relationship results from the complex pattern of tropical ocean–atmosphere dynamics and feedback responses of ENSO to changes in

radiative forcing. The combined response of ENSO to solar and volcanic forcing in past centuries provides a potential explanation for why the tropical Pacific appears to have been in a cold La Niña-like state during the Medieval Warm Period and a warm El Niño-like state during the Little Ice Age (Adams et al., 2003), as discussed previously. These forced changes in ENSO would likely counteract the larger scale warming and cooling at these times, respectively, leading to smaller changes in global and hemispheric mean temperatures than would otherwise be expected. For this reason, climate models that fail to fully resolve the physics underlying ENSO may not provide a reliable indication of past changes in global or hemispheric temperatures.

Further modeling experiments and model/data comparison focusing on the relationship between past radiative forcing of climate, and the behavior of important dynamical modes of variability such as NAO and ENSO over the past one or more millennia, should increasingly help elucidate the factors that may be responsible for important changes in climate in past centuries.

See also: Paleoclimate Relevance to Global Warming; Varved Lake Sediments; Varved Marine Sediments. **Glacial Climates:** Volcanic and Solar Forcing. **Glacial Landforms:** Evidence of Glacier Recession. **Paleoclimate:** Introduction. **Paleoclimate Reconstruction:** Paleodroughts and Society.

References

- Adams JB, Mann ME, and Ammann CM (2003) Proxy evidence for an El Niño-like response to volcanic forcing. *Nature* 426: 274–278.
- Bradley RS (1999) *Paleoclimatology: Reconstructing Climates of the Quaternary*. San Diego: Academic Press.
- Bradley RS and Jones PD (eds.) (1995) *Climate Since A.D. 1500*. London: Routledge.
- Briffa KR, Osborn TJ, Schweingruber FH, et al. (2001) Low-frequency temperature variations from a northern tree-ring density network. *Journal of Geophysical Research* 106: 2929–2941.
- Cobb KM, Charles CD, Cheng H, and Edwards RL (2003) El Niño–Southern oscillation and tropical Pacific climate during the last millennium. *Nature* 424: 271–276.
- Cook ER, Woodhouse C, Meko DM, and Stahle DW (2004) Long-term aridity changes in the western United States. *Science* 306: 1015–1018.
- Crowley TJ (2000) Causes of climate change over the past 1000 years. *Science* 289: 270–277.
- Crowley TJ, Baum SK, Kim K-Y, Hegerl GC, and Hyde WT (2003) Modeling ocean heat content changes during the last millennium. *Geophysical Research Letters* 30: 1932.
- Delworth TL and Mann ME (2000) Observed and simulated multidecadal variability in the Northern Hemisphere. *Climate Dynamics* 16: 661–676.
- Evans MN, Kaplan A, and Cane MA (2002) Pacific sea surface temperature field reconstruction from coral $\delta^{18}\text{O}$ data using reduced space objective analysis. *Paleoceanography* 17: 1007.
- Folland CK, Karl TR, and Christy JR (2001) Observed climate variability and change. In: Houghton JT, Ding Y, and Griggs DJ, et al. (eds.) *Climate Change 2001: The Scientific Basis*, pp. 99–181. Cambridge, UK: Cambridge University Press.
- Fritts HC (1976) *Tree Rings and Climate*. London: Academic Press.
- Huang S, Pollack HN, and Shen P-Y (2000) Temperature trends over the past five centuries reconstructed from borehole temperature. *Nature* 403: 756–758.
- Jacoby GC and D'Arrigo RD (1989) Reconstructed Northern Hemisphere annual temperature since 1671 based on high-latitude tree-ring data from North America. *Climatic Change* 14: 39–59.
- Jones PD and Mann ME (2004) Climate over past millennia. *Reviews of Geophysics* 42: RG2002.
- Jones PD, Osborn TJ, and Briffa KR (2001) The evolution of climate over the last millennium. *Science* 292: 662–667.
- Luterbacher J, Dietrich D, Xoplaki E, Grosjean M, and Wanner H (2004) European annual and seasonal temperature variability, trends and extremes since 1500. *Science* 303: 1499–1503.
- Luterbacher J, Schmutz C, Gyalistras D, Xoplaki E, and Wanner H (1999) Reconstruction of monthly NAO and EU indices back to AD 1675. *Geophysical Research Letters* 26: 2745–2748.
- Mann ME, Ammann CM, Bradley RS, et al. (2003) On past temperatures and anomalous late 20th century warmth. *EOS* 84: 256–258.
- Mann ME, Bradley RS, and Hughes MK (1998) Global-scale temperature patterns and climate forcing over the past six centuries. *Nature* 392: 779–787.
- Mann ME, Cane MA, Zebiak SE, and Clement A (2005) Volcanic and solar forcing of the tropical Pacific over the past 1000 years. *Journal of Climate* 18: 447–456.
- Minobe S (1997) A 50–80 year climatic oscillation over the North Pacific and North America. *Geophysical Research Letters* 24: 683–686.
- Moberg A, Sonechkin DM, Holmgren K, Datsenko NM, and Karlen W (2005) Highly variable Northern Hemisphere temperatures reconstructed from low- and high-resolution proxy data. *Nature* 433: 613–617.
- Oerlemans J (2005) Extracting a climate signal from 169 glacier records. *Science* 308: 675–677.
- Robock A (2000) Volcanic eruptions and climate. *Reviews of Geophysics* 38: 191–219.
- Rutherford S, Mann ME, Osborn TJ, et al. (2005) Proxy-based Northern Hemisphere surface temperature reconstructions: Sensitivity to methodology, predictor network, target season and target domain. *Journal of Climate* 18: 2308–2329.
- Schmidt GA, Shindell DT, Miller RL, Mann ME, and Rind D (2004) General circulation modeling of Holocene climate variability. *Quaternary Science Reviews* 23: 2167–2181.
- Shindell DT, Schmidt GA, Mann ME, Rind D, and Waple A (2001) Solar forcing of regional climate change during the Maunder Minimum. *Science* 294: 2149–2152.
- Stahle DW, D'Arrigo RD, Krusic PJ, et al. (1998) Experimental dendroclimatic reconstruction of the Southern Oscillation. *Bulletin of the American Meteorological Society* 79: 2137–2152.
- Verschuren D, Laird KR, and Cumming BF (2000) Rainfall and drought in equatorial east Africa during the past 1100 years. *Nature* 403: 410–414.
- Wigley TML, Ingram MJ, and Farmer G (1981) Past climates and their impact on man: A review. In: Wigley TML, Ingram MJ, and Farmer G (eds.) *Climate and History*, pp. 3–50. Cambridge, UK: Cambridge University Press.

Historical Climatology

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Introduction

Historical climatology is the branch of climatology that uses written documents to investigate past climate. Such records can include instrumental or noninstrumental weather records. For paleoclimatological purposes, historical documentary sources are most appropriately thought of as weather records that pre-date the establishment of official meteorological networks. Modern observing networks use standardized observation methods and instruments for recording and measuring meteorological variables at defined levels of precision and accuracy. The earliest organized networks date from the late eighteenth century, but it was not until the establishment of international congresses by government-funded meteorological services in 1873 that systematic efforts began to standardize procedures and instrumentation internationally (Fleming, 1990). Historical data are more likely to be derived from nonstandard methods of instrument exposure, highly variable observing practices and definitions of weather terms, and uncalibrated instruments. Climatologists working with historical documents attempt to extend modern records back in time using statistical methods that account for nonstandard methods of instrument exposure, observing practices, and instrumentation (Bradley and Jones, 1995; Jones and Mann, 2004). Only by such procedures can the knowledge of past climate be accurately quantified and compared with current and future observations.

Written records are most abundant for the past millennium and some provide temporal resolution at the subhourly to daily timescales, which are simply unavailable for any other type of proxy weather record. The availability of written sources in both time and place are the ultimate limits to climate history reconstruction from such records. These limitations restrict their contribution to the last 1 ka, although some fragmentary records in parts of the Mediterranean region, the Middle East, and China are over 2-ka old. China holds probably the largest collection of written documents with climate information, but even these records contain significant gaps (Ge et al., 2003). Both European and Chinese archives hold large collections of climate information from about AD 1400, with the European records covering the largest geographical area due to the extensive European colonization of the Americas, Africa, Oceania, and Asia. While other paleoclimate proxy records can provide timescales extending thousands to millions of years in the past, historical documents provide unmatched subannual resolution not possible from other sources. Historical records, along with other proxy records (e.g., tree rings and corals), are used in proxy-based reconstructions of climate (see Section 1.3 of Jones and Mann (2004) for bibliographical references). Historical records from outside of Europe and East Asia are still needed to improve the spatial coverage of data and to provide independent checks on climate models. With the increasing knowledge of large-scale climate features on the interannual to millennial timescales and with the improved computational

and statistical methods, there is greater interest in the long-term variability of intraannual and intraseasonal weather and climate (Klein Tank et al., 2002), including the frequency and intensity of local-scale phenomena and hazards. Historical documents are beginning to assist in answering some of these questions (e.g., Chuine et al., 2004). The impact of climate on human affairs is also an important feature of the written documentary record. Past interactions between climate and society are of great interest as current uncertainties about the direction of future climate have a direct impact on current responses and planning for future climate change.

Potential and Limitations of Historical Documents

The challenge of producing climatically homogeneous instrumental records is well recognized (IPCC, 2001). In the first place, the absence of suitable records is a major reason why most of the longest time series are rarely more than 150 years or so in length. The discontinuous and fragmentary nature of historical records requires careful analysis and calibration. Historical data should not be uncritically merged together to form data sets based on heterogeneous source data that have no consistent definitions of terms, frequency of observation, and nonstandardized instrumentation, exposure, and procedures. Even the construction of a simple index or chronology is sensitive to the methods used in its construction.

Both the quantity and quality of instrumental data decrease in number further back in time. Long series of homogeneous weather records can be produced from the noninstrumental component of these records, such as wind direction and rain/snow days that can be linked with modern instrumental records. However, careful source selection and robust statistical methods are still required to produce a homogeneous time series that is useful for paleoclimate studies. For example, the definition of a 'frost' is highly variable as the term may be based on any number of factors or thresholds. The same is true for rain and snow day counts. Observer sensitivities to various weather phenomena vary between persons and within the lifetime of a single individual. Some long time series were drawn up by nonspecialists who uncritically used multiple source documents of varying quality and/or secondary sources (e.g., see the critical analysis of past El Niño events by Ortleib (2000)). Jones and Mann (2004) provide a very useful discussion of unreliable conclusions about past climate due to poor interpretation of data sources.

The potential of new data sources, particularly the greater geographical coverage offered by ships' logbooks, is great. Many of the most important global, hemispheric, and regional climate signals are strongest, or located mainly, over ocean areas (e.g., the El Niño/Southern Oscillation). Colonial archives contain records from the land areas of the tropics that are only now beginning to be rediscovered. Even data sets

that have received attention in the past, such as the extensive records of the Hudson's Bay Company in Canada, contain many records of past weather that are yet to be studied.

Sources and Repositories of Documentary Materials

Documentary written records that provide information on past climate are incredibly plentiful but reside largely in undigitized form in thousands of locations worldwide. One of the most important tasks for historical climatologists is ascertaining the locations of such records. Government and ecclesiastical bureaucracies are the main source of historical climate data. The large Chinese government bureaus of the past 2.5 ka generated large masses of data concerned with climate impacts on agriculture, particularly drought and floods, heat waves and cold waves, and extreme weather events such as typhoons, snowstorms, and localized severe weather (Ge et al., 2003, 2005). Most of the original accounts are now no longer extant, but the surviving documents hint at the large number of such documents that originally existed. In Europe, ecclesiastical records from the Christian churches are the main source of surviving records of climate for the past 2 ka, although the number of documents is especially small in the first thousand years or so. Parish records and chronicles provide a wealth of information providing direct or indirect information on past climate (Le Roy Ladurie, 1988). From about AD 1500, secular government bureaucracies begin to supplant the Church as the most important sources of weather data. However, proselytizing religious groups remain important sources for some of the earliest instrumental weather records in many parts of the world (e.g., missionary accounts from Africa (Nash and Endfield, 2002) and German Moravian missionaries in Greenland and Labrador (Macpherson, 1987)). Personal diaries and journals, including those of exploring expeditions, are among the most important sources of weather, particularly for the past 300–500 years. Similarly, corporate records such as the Hudson's Bay Company and the English and Dutch East India companies provide the only source of weather data for large areas of Canada and India, respectively, that otherwise would not exist.

Given the large array of sources for historical weather data, all archives, libraries, museums, and other repositories of documents should be considered to potentially hold valuable climate data. Government archives, such as the UK National Archives (formerly the Public Record Office) in London and Archives Nationales in Paris, often are the single largest repositories of government records and are the best single sources for many types of records. There can be no substitute, however, for an exhaustive search of all types of repositories. The large volume of potential records makes interdisciplinary cooperation between climatologists, historians, librarians, and archivists essential for scientific investigations of historical climate. Data search efforts supported by the United Nations such as Archival Climate History Survey (ARCHISS) provide an excellent model for locating such records (Gioda et al., 2003). Many universities and archives have searchable databases that facilitate such research.

Types of Documentary Materials

Any written record may provide direct or indirect weather information. Church chronicles, parish records, government

surveys and other official accounts, estate and farm records, records of crop yields and harvest dates, phenological records, diaries and journals (including those specifically devoted to weather), astronomical observatory records, almanacs, books, travel and explorer accounts, newspapers, legal documents, harbor master records, ship protest records (legal documents absolving a ship's captain from damage sustained in a voyage due to weather or accident) and ship logbooks are examples of historical weather data sources. Some of the most numerous and important documentary materials for historical climatology are reviewed here.

Ships' Logbooks

Weather data from ships are an integral part of historical climate data, since about 70% of the Earth's surface is covered with water, but the retrieval of pre-1850 weather reports is limited to the Maury Abstract Log Collection (US Department of Commerce, 1998), the Climate of the World's Oceans (CLIWOC) project (Herrera et al., 2004), and some small private data sets. Large amounts of both instrumental and noninstrumental data from ships remain to be used in paleoclimate studies.

The ship's logbook is a documentary record that has evolved over time in both its content and appearance. A logbook is a record of all of the activities related to the navigation of a ship and significant events that play a role in the ship's journey. A logbook documents the measurements to determine the ship's location, estimated direction and speed of movement, and those features of the wind, weather, and sea that are needed to estimate a ship's location. A sample logbook is shown in Figure 1.

The amount and type of weather information in a logbook varies depending upon the type of ship (merchant vs. naval), the keeper of the logbook (masters, captains, lieutenants, or other party), its location (usually less detail in port), and when it was kept (generally fewer observations were recorded per day in the seventeenth and early eighteenth centuries than in later years).

Instrumental weather records first appear in widely scattered logbooks in the second half of the eighteenth century, but it is only in the mid-nineteenth century that the frequency of temperature and barometric pressure data increase so as to be very common. Noninstrumental weather records of wind direction, wind force, and prevailing weather conditions are available in virtually every logbook. In the English-speaking maritime world of England and the United States, merchant ships normally ceased making entries in their logbooks during their time in port. Navy ships made observations both at sea and in port. Observations in port ceased only if the ship was taken out of service. For this reason, long time series are most likely to be derived from naval ports and other anchorages continuously in use by naval forces.

Table 1 reveals the evolution of the most frequently used wind force terms in common use in ships of the British Royal Navy from the 1690s through the 1820s. The constantly changing frequency of individual terms, relative to others simultaneously in use, points to the need to carefully document changing practices before attempting to use such data for reconstruction of historical wind records. Each country's navy, as well as merchant ship fleets, has its own variants of terms that evolved as well.

Diaries and Journals

Many private diaries kept by individuals are available from many parts of the world. Some are devoted solely to weather (Figure 2) while others are oriented toward many topics, the weather of which is only one part of the content (Figure 3). The usefulness of diaries and journals for climate research depends upon the writer's power of observation and willingness to faithfully record these observations on a regular basis. Length of record does not necessarily equal quality; many shorter series may be of much greater reliability.

Nonetheless, many very long series of diaries do serve as anchors for long time series such as those used to produce the Central England temperature (CET) series (Manley, 1974). Multiple overlapping diaries from the same locality can be found in continuously occupied areas to cross-check each

observer for consistency and reliability in a general sense. The observers can be ranked for general reliability, and certain specific weather elements may be observed more keenly by one observer than by others. In general, even the diaries of poorer quality can contain information that is unique due to the particular interests and motivations that drive the writer to keep the record in the first place. A farmer, a shopkeeper, a fisherman, a lighthouse keeper, a plantation owner, a clergyman, and a surveyor will all have different perspectives and motivations for recording weather data. The sensitivity of their particular occupations may well drive the relative frequency of certain types of weather events being recorded in their diaries with respect to other weather events.

Government and Other Official Records

Many crops and foodstuffs grown by man are so important to their local economies that extensive documentation on their production exists in local-, regional-, and national-level government archives. Although no single location has a continuous record, it is possible to piece together similar type documents from the immediate area to produce a long-term climate series. Pfister et al. (1994) document an extensive database of European meteorological data gathered over a period of many years and discuss the methods involved in reconstructing temperature and precipitation indices. A good example of the type of data in official records is from viticultural records. The French wine-growing region of Burgundy contains extensive documentation from a number of vineyards that record the wind harvest dates as far back as the year 1370. Chuine et al. (2004) produced a record of Burgundy spring and summer temperatures estimated from the relationship between wine output and spring–summer temperature in the years 1960–1989. Such records are so far relatively rare as the extensive work of extracting the scattered documents and combining them into a continuous series is time consuming. Such results show how an interdisciplinary approach between historians and climatologists can extract valuable climate data from a

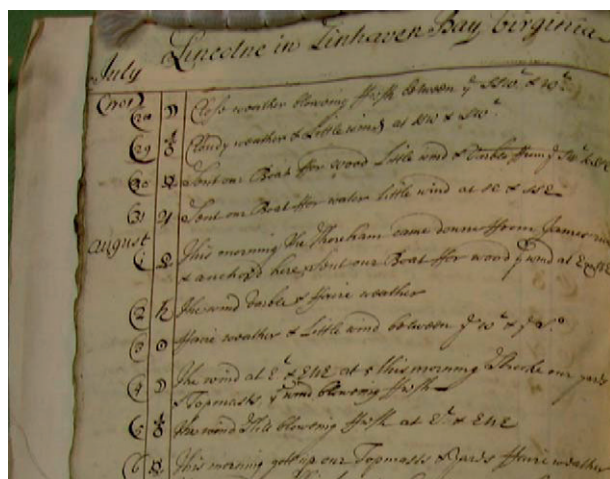


Figure 1 The logbook of HMS Lincoln, moored at Lynhaven Bay, Virginia, 28 Jul–6 Aug 1700 (detail). A typical example of the daily weather and significant activity on a Royal Navy ship provided in a logbook while the ship was at anchor in port. From the UK National Archives.

Table 1 Frequency of the most common wind force terms in the 420 British Royal Navy ships' logbooks, by decade, from 1691 to 1820, expressed as a percentage frequency of each decade's total number of observations

Decade	Percentage of wind force terms										
	Little wind	Small gale	Moderate gale	Fresh gale	Strong gale	Hard gale	Light airs	Light breeze	Moderate breeze	Fresh breeze	Strong breeze
1691–1700	9.4	9.4	20.1	33.5	3.5	24	0	0	0	0	0
1701–1710	8.7	3	26	31.9	5.9	23.7	0	0	0.7	0.2	0
1711–1720	10.6	0.3	26.6	32.3	5.4	24.5	0	0	0.3	0	0
1721–1730	11.8	0	15.3	38	3.6	29.6	0.2	0	0.4	1.1	0
1731–1740	15.8	0	14	42	5.1	20.9	0.2	0.4	0.7	0.9	0
1741–1750	8.8	0.3	7.8	43.9	9.9	26.5	1.9	0	0.8	0.3	0
1751–1760	8.4	0	1.5	38.6	11.5	21	8.8	1.9	2.3	6.1	0
1761–1770	4.6	0	2.6	22.8	14.4	13.7	12.2	3.6	5.5	20.6	0
1771–1780	1.8	0	0.2	23.6	21.8	9.8	6	4	6.6	25.2	1
1781–1790	1.8	0	0	13.3	18	6.3	11	5.2	12.6	29	2.9
1791–1800	0	0	0	10.1	25	3.4	16.4	5.2	8.6	27.2	4.1
1801–1810	0	0	0.3	10.8	17.4	3	10.8	6.5	9.7	29.8	11.8
1811–1820	0	0	0.4	11.6	20.9	1.6	4.7	3.5	10.1	26	21.3

Data are for ship encounters with tropical cyclones in the North Atlantic, so the values are not representative of climatological values in all weather. However, usage of specific wind force terms is accurately captured. Over time, some wind force terms go out of use while others came into use.

	H	D	W	W	H	D	W	W
1	8. 62. Rain	8. 62. Rain	8. 62. Rain	8. 62. Rain	8. 62. Rain	8. 62. Rain	8. 62. Rain	8. 62. Rain
2	8. 60. Rain	8. 60. Rain	8. 60. Rain	8. 60. Rain	8. 60. Rain	8. 60. Rain	8. 60. Rain	8. 60. Rain
3	8. 59. Rain	8. 59. Rain	8. 59. Rain	8. 59. Rain	8. 59. Rain	8. 59. Rain	8. 59. Rain	8. 59. Rain
4	8. 58. Rain	8. 58. Rain	8. 58. Rain	8. 58. Rain	8. 58. Rain	8. 58. Rain	8. 58. Rain	8. 58. Rain
5	8. 57. Rain	8. 57. Rain	8. 57. Rain	8. 57. Rain	8. 57. Rain	8. 57. Rain	8. 57. Rain	8. 57. Rain
6	8. 56. Rain	8. 56. Rain	8. 56. Rain	8. 56. Rain	8. 56. Rain	8. 56. Rain	8. 56. Rain	8. 56. Rain
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Figure 2 A weather diary kept near Camden, South Carolina from the James Kershaw Papers, Manuscripts Division, South Caroliniana Library, University of South Carolina. The page here is from Jan 1796 and documents unusually wet weather. Courtesy of Curator of Manuscripts, South Caroliniana Library, University of South Carolina, Columbia, South Carolina.

combination of modern data, historical documents, and a statistical model. While not all records will extend back more than 600 years, there are still many such records of shorter length that may be produced from historical records. Accounts from both private and local government sources of income loss from revenue sources such as tolls due to the freezing, flooding, and drying up of navigation channels provide examples of other records that are available for use.

Newspapers and Government Gazettes

A fruitful source for weather data from many areas of the world is the newspaper. Newspapers first began to appear in Europe in the seventeenth century, first in the form of officially sanctioned gazettes (which continued into the twentieth century in many parts of the world) and then into privately owned newspapers that proliferated in both Europe and North America and their colonies in the eighteenth century. Descriptions of natural wonders and disasters caused by the weather were common since most food production was local and trade restrictions between nations were often in force. News of weather-induced shortfalls in local food or crop production spread quickly among merchants, plantation owners, and government officials, who often formed the core of the newspaper subscriber base. Newspapers routinely provided the account of the impact of weather on food supplies and availability, shipping, and local damage reports. These include the inability to mill grain due to low river levels, the inability of ships to reach their

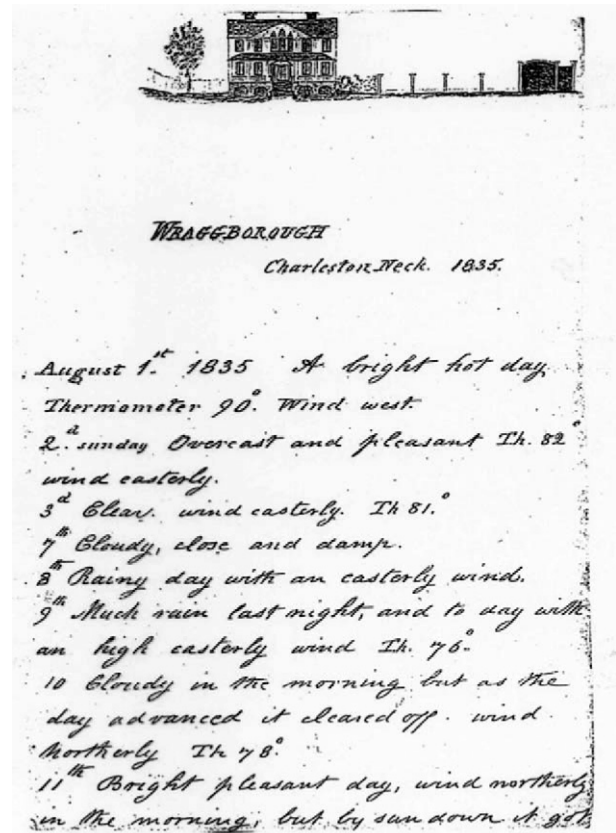


Figure 3 A personal weather diary for 1–11 Aug 1835 taken from the diary of Charles Heyward at his plantation in South Carolina, part of the Heyward family papers. Courtesy of Curator of Manuscripts, South Caroliniana Library, University of South Carolina, Columbia, South Carolina.

destination due to either too little or too much water, crop yields and the weather factors causing a shortfall or windfall; extremes of heat and cold causing excess deaths or raising the price for scarce fuel material; tornados, hail, and thunderstorms damaging crops, killing farm animals, and people; references to earlier instances of similar weather and other impressionistic descriptions of how unusual the weather appeared to the people at that time; weather events in foreign regions that would be of interest to merchants involved in trade with the affected region. In some instances, local weather diaries were published in newspapers providing another source of data for the historical climatologist (Figure 4).

Newspaper accounts provide weather from ships at sea encountering bad weather, since ships were the only source of transoceanic shipment of cargo and people until the twentieth century. Incoming ships described their encounters with gales, calms, lightning strikes, hurricanes, thick fogs, snowstorms, icebergs, heavy seas, and other phenomena which may be otherwise lost as most ships' logbooks are no longer extant.

In many instances, there are other official and private documents that also cover items recorded in newspapers. Initial accounts of ships lost at sea are often followed up by government and merchant records concerned with recording damages and losses for insurance and other claims made by owners,



Figure 4 Daily weather record for May 1786 for Nassau, Bahamas published in The Bahama Gazette of 27 May–3 Jun 1786. Air temperatures at 7 a.m., 2 p.m., and 9 p.m. are provided along with prevailing winds and weather conditions for each day. The original record is not known to be extant so the newspaper provides the only evidence of its existence.

passengers, crew, and other claimants. Newspaper accounts should not be considered to be the last word on weather impacts, but they do offer the widest variety of historical weather and weather impact data that can lead the historian and climatologists to other source documents.

Methods of Historical Climatology Research

Paleoclimate studies relying on historical documents incorporate these sources in different ways. The most common method is to extend a preexisting instrumental record as far back in time as the documents provide by comparing overlapping portions of different records to create a single climatologically homogeneous composite series. The CET series is a good

example of such a record (Manley, 1974; Parker et al., 1992). Production of such long time series involves the comparison of different weather record sources that overlap one another in time, which allows a direct comparison of their differences to be measured and adjusted via statistical techniques (often a regression model) so that nonclimatic biases (from different site exposures, times of observation, different instruments, and other methodological differences) are removed and a continuous time series, with estimated margins of error in the derived values, of one or more meteorological variables are produced. Parker and Horton (2005) provide an admirable example of determining both random and systematic errors in temperature data that should inform any approach to climate data reconstruction. Other studies use historical documents to serve as independent checks on another proxy source climate

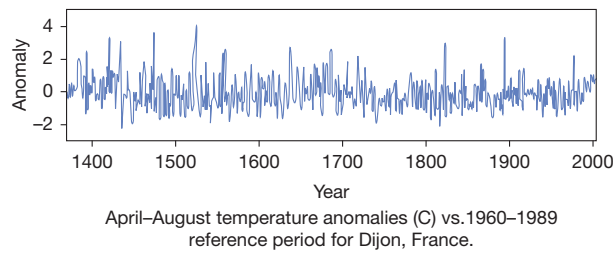


Figure 5 Reconstructed spring–summer temperatures for Dijon, France from grape harvest records, 1370–2003 (Chuine et al., 2004). Figure derived from Pfister C (1993) Historical Weather Indices from Switzerland. IGBP PAGES/World Data Center-A Paleoclimatology Data Contribution Series #93-027. NOAA/NGDC Paleoclimatology Program, Boulder, CO, USA.

record (e.g., tree-ring reconstructions of precipitation and historical and archaeological records (Stahle et al., 1998)). The validity of this method assumes that the historical documentary record has been compiled in a consistent manner and uses reliable original sources. In other work, long records of weather event types, such as grape harvest dates in France (Chuine et al., 2004; Figure 5), El Niño events off the coast of South America (Quinn and Neal, 1983), and drought and flood chronologies in southern Africa (Nash and Endfield, 2002) are gathered to reconstruct a particular aspect of the Earth's climate system. Multivariate calibration methods employing both historical data sources and other proxy records with annual resolution are compared with climate models to assess the climate sensitivity of the models and to compare model-based spatial patterns of climate variability with the proxy-based record.

Discontinuous records can also provide information on past climate. Japanese instrumental temperature records are continuous only back to the 1870s, but colonial Dutch records now provide several decades worth of data back to 1819 which can still be compared with modern data (Können et al., 2003). Sometimes a modern data set is not needed to produce significant results. For example, short episodes of volcanic-induced global cooling were derived from a 21-year series of ships' logbooks from the early nineteenth century that is not continually linked to the modern record but used only a contemporary reference period to measure the short-term volcanic-induced temperature anomalies (Chenoweth, 2001).

Our understanding of history can be altered from historical climatology research. Climate descriptions of the American Great Plains once were thought to be artificial creations of the human mind but are now known from climate reconstructions to reflect actual climate anomalies (Lawson and Stockton, 1981). These early encounters with the weather colored the expectations of subsequent emigrants to the region and helped to shape its political growth and future economic path (Mock, 2000). Historical events of great importance are sometimes shaped in part by the role of weather as the Spanish Armada Storms of 1588 (Douglas et al., 1978). In turn, evidence of unusual climate extremes can sometimes be found by employing a synoptic approach to historical data (Chenoweth, 1996; Kingston, 1988). In this approach, multiple document types (e.g., journals, logbooks, letters, newspaper articles, legal documents) can be applied to map subdaily to monthly scale weather phenomena.

Suggestions for Future Research

Future research using documentary historical weather should focus on several fronts, and be adaptable to new discoveries from other fields of research as well as those in climatology. First, the construction of catalogs of known or likely sources of weather records needs to be compiled for all regions of the world. The National Oceanic and Atmospheric Administration (NOAA) historical climate data catalog is being produced for data from the United States, and the World Meteorological Organization (WMO) ARCHISS data survey project has located new sources in Central and South America. Only by knowing the amount and type of data available can scientists, in cooperation with other social scientists and historians, exploit the sources that have the best potential for filling in missing data from the historical record. Surveys of libraries and archives, including online database searches, are making it easier for such data to be located. Second, given knowledge of the availability of data sources, the records of areas that are known to be colocated in regions where important 'centers of action' are known to exist, such as the monsoon regions of India, the eastern tropical Pacific Ocean, high latitude sites, or areas where data can complement existing climate indices, such as the Caribbean region, where trade wind fluctuations are associated, in part, with fluctuations in the North Atlantic oscillation, should be chosen. Such studies can assist in determining the relative nonstationarity of the climate system and provide invaluable data on natural variability prior to anthropogenic effects on climate. Any record determined to be of good quality, regardless of its geographical location, should be closely studied. By extending the length of existing records, there is additional data available for calibration of proxy data sets. The creation of new data sets provides new data to be used as an independent data source to evaluate previous results, as well as being available for inclusion in future single or multiproxy climate reconstructions. Third, the original source documents should always be used when available. There are no shortcuts to the production of high-quality historical climate records. Fourth, different research communities must continue to reach out and share data and ideas. This process of knowledge diffusion will ensure that the value of the large amounts of historical data still awaiting use is recognized for its cultural, historical, and scientific value.

See also: Paleoclimate Relevance to Global Warming. **Paleobotany:** Overview of Terrestrial Pollen Data. **Paleoclimatology:** Introduction. **Paleoclimatology Reconstruction:** Paleotempestology.

References

- Bradley RS and Jones PD (eds.) (1995) *Recent Developments in Studies of Climate Since A.D. 1500*, revised edition, pp. 666–679. London: Routledge.
- Chenoweth M (1996) Ships' logbooks and 'The Year without a Summer'. *Bulletin of the American Meteorological Society* 77: 2077–2093.
- Chenoweth M (2001) Two major volcanic cooling episodes derived from global marine air temperature, AD 1807–1827. *Geophysical Research Letters* 28(15): 2963–2966.
- Chuine I, Yiou P, Viovy N, Seguin B, Daux V, and Ladurie ELR (2004) Grape ripening as a past climate indicator. *Nature* 432: 289–290.

- Douglas KS, Lamb HH, and Loader CC (1978) A meteorological study of July to October 1588: The Spanish Armada storms. CRU RP6, *Climatic Research Unit, School of Environmental Studies*, p. 44. Norwich: University of East Anglia.
- Fleming JR (1990) *Meteorology in America, 1800–1870*. p. 264. Baltimore: The Johns Hopkins University Press.
- Ge Q, Zheng J, Fang X, et al. (2003) Winter half-year temperature reconstruction for the middle and lower reaches of the Yellow River and Yangtze River, China, during the past 2000 years. *The Holocene* 13(6): 933–940.
- Ge Q-S, Zheng J-Y, Hao Z-X, Zhang P-Y, and Wang W-C (2005) Reconstruction of historical climate in China: High-resolution precipitation data from Qing Dynasty archives. *Bulletin of the American Meteorological Society* 86(5): 671–679.
- Gioda A, Baker M, Garza G, et al. (2003) The ARCHISS (Archival Climate History Survey) project in Latin America: Examples of homogeneous documentary data series (Potosí, Mexico City and Lima). *Geophysical Research Abstracts* 5: 09762.
- Herrera RG, Wheeler D, Können G, Jones P, Koek F, and Prieto MR (2004) Cliwoc Final Report. p. 69. Madrid: Universidad Complutense de Madrid.
- IPCC (2001) Climate change 2001: The scientific basis. In: Houghton JT, Ding Y, and Griggs DJ, et al. (eds.) *Contribution of Working Group I to the Third Assessment Report of the Intergovernmental Panel on Climate Change*, p. 881. New York: Cambridge University Press.
- Jones PD and Mann M (2004) Climate over past millennia. *Reviews of Geophysics* 42(2): RG2002. <http://dx.doi.org/10.1029/2003RG000143>.
- Kington JA (1988) *Daily Weather Mapping for the 1780s*. Cambridge: Cambridge University Press.
- Klein Tank AMG, Wijngaard JB, Können GP, et al. (2002) Daily dataset of 20th-century surface air temperature and precipitation series for the European Climate Assessment. *International Journal of Climatology* 22(12): 1441–1453.
- Können GP, Zaiki M, Baede APM, Mikami T, Jones PD, and Tsukahara T (2003) Pre-1872 extension of the Japanese instrumental meteorological observation series back to 1819. *Journal of Climate* 16(1): 118–131.
- Lawson MP and Stockton CW (1981) Desert myth and climatic reality. *Annals of the Association of American Geographers* 71(4): 527–535.
- Le Roy Ladurie E (1988) *Times of Feast, Times of Famine: A History of Climate Since the Year 1000*. New York: Noonday Press, Farrar, Straus and Giroux.
- Macpherson AG (1987) Early Moravian interest in northern Labrador weather and climate: The beginning of instrumental recording in Newfoundland. In: Steele DH (ed.) *Early Science in Newfoundland and Labrador*, p. 30. Avalon: Sigma Xi.
- Manley G (1974) Central England temperatures, monthly means 1658–1973. *Quarterly Journal of the Royal Meteorological Society* 100: 389–405.
- Mock CJ (2000) Rainfall in the Garden of the US Great Plains during the late nineteenth century. *Climatic Change* 44: 173–195.
- Nash DJ and Endfield GH (2002) A 19th century climate chronology for the Kalahari region of central southern Africa derived from missionary correspondence. *International Journal of Climatology* 22(7): 821–841.
- Ortleib L (2000) The documented historical record of El Niño events in Peru: An update of the Quinn record (sixteenth through nineteenth centuries). In: Diaz HF and Markgraf V (eds.) *El Niño and the Southern Oscillation*, p. 207. Cambridge: Cambridge University Press.
- Parker DE and Horton B (2005) Uncertainties in central England temperature 1878–2003 and some improvements to the maximum and minimum series. *International Journal of Climatology* 25(9): 1149–1171.
- Parker DE, Legg TP, and Folland CK (1992) A new daily Central England temperature series, 1772–1991. *International Journal of Climatology* 12: 317–342.
- Pfister C, Kington J, Kleinlogel G, Schuele H, and Siffert E (1994) The creation of high resolution spatio-temporal reconstructions of past climate from direct meteorological observations and proxy data: Methodological considerations and results. In: Frenzel B, Pfister C, and Glaeser B (eds.) *Climatic Trends and Anomalies in Europe 1675–1715*, pp. 329–376. Stuttgart: Gustav Fischer.
- Quinn WH and Neal VT (1983) Long-term variations in the Southern Oscillation, El Niño, and Chilean subtropical rainfall. *Fishery Bulletin* 81(2): 363–374.
- Stahle DW, Cleaveland MK, Blanton DB, Therrell MD, and Gay DA (1998) The lost colony and Jamestown droughts. *Science* 280: 564–567.
- US Department of Commerce (1998) *The Maury Collection: Global Ship Observations 1792–1910 (Version 1.0)*. Asheville: National Oceanic and Atmospheric Administration (CD-ROM).

Paleoclimate Relevance to Global Warming

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The Earth's climate is vastly different now from what it was at the beginning of the Quaternary period. It is different from what it was 20 000 years ago when ice sheets covered much of the Northern Hemisphere. Climate changes in the past were driven by natural causes, such as variations in the Earth's orbit or the carbon dioxide (CO₂) content of the atmosphere. Current and future climatic changes, however, have another source as well – human activities. Humans cannot directly rival the power of the natural forces driving the climate, such as the immense energy input from the Sun. They can, however, indirectly alter the natural flows of energy enough to create significant climatic changes.

Anthropogenic climate change is a problem of potentially great significance for the Earth's future. Many activities associated with human economic development have changed and are changing our physical environment in unprecedented ways. CO₂, methane (CH₄), and water vapor are the principal heat-trapping greenhouse gases. Carbon, the main element in the top two human-enhanced greenhouse gases (CO₂ and CH₄), is the underpinning of most fuels used in transportation and energy production. Carbon also makes up approximately half of the dry weight of most vegetation. Thus, the carbon that cycles through biological species, air, water, and soils is both an essential nutrient and a potential problem, depending on its location in the system. Human modification of the carbon cycle has far-reaching implications for human welfare and the health of the biosphere.

Although the changes that force anthropogenic climate change are unprecedented – and thus paleoclimatic conditions are not analogous to future anthropogenic changes – paleoclimate is a backdrop against which the understanding of the climate system can be calibrated, an understanding essential to have confidence in climate models, the tools used to project future changes. Many scientists conducting research to gain understanding of specific aspects of the climate system and the cycles or events that drive it benefit from studies of past climate changes in the paleoclimatic record, revealed through climate proxies.

This article focuses on how the paleoclimatic record can help understand anthropogenic climate change and how researchers have used that information to advance the knowledge of human interactions with the climate system.

Anthropogenic Climate Change

Scientific Consensus

It is scientifically well established that the Earth's surface air temperature has warmed significantly during the past 150 years, and an upward trend can be clearly discerned by plotting historical temperatures. [Figure 1](#) shows the observed global temperature record as a plot of the yearly deviations

from the 1961–90 average temperature. A glance at [Figure 1](#) shows that Earth's temperature is highly variable, with year-to-year changes often masking the overall rise of approximately 0.7 °C. Nevertheless, the long-term upward trend is obvious. Especially noticeable is the rapid rise at the end of the twentieth century. [Figure 1](#) shows that the 20 warmest years of the twentieth century include the entire decade of the 1990s and all but three years from the 1980s as well. The beginning of the twenty-first century has not changed this trend. 2001–2004 are among the five warmest years on record, exceeded only by 1998, the all-time record high through 2004. There is good reason to believe that the 1990s would have been even hotter had the eruption of Mt. Pinatubo in the Philippines not put enough dust into the atmosphere to cause global cooling of a few tenths of a degree for several years. Clearly, the recent past has seen substantial surface warming. This surface thermometer record has been matched by shrinking of the vast majority of mountain glaciers, warming in the oceans, shrinking of Arctic Sea ice, and responses of plants and animals (e.g., earlier bloom dates in the spring and earlier arrival dates of migrating birds).

Radiative Forcing

In addition, it is well established that human activities have caused increases in radiative forcing, with radiative forcing defined as a change in the balance between radiation coming into and going out of the Earth–atmosphere system. The annual average atmospheric concentration of carbon dioxide is currently approximately 378 ppm. In the past few centuries, atmospheric carbon dioxide has increased by more than 30%, as indicated in [Figure 2](#), and the predominant cause is human activity – the burning of fossil fuels and, to a considerable extent, land use changes such as deforestation. As shown in [Figure 2](#), concentrations of other greenhouse gases and other radiatively active atmospheric components, such as sulfate aerosols, are rising as well due to human activities. Although there is uncertainty regarding the current level of anthropogenic radiative forcing, it is well established that the greenhouse gas and aerosol concentrations indicated in [Figure 2](#) are affecting the energy balance of the climate system and contributing to the observed warming shown in [Figure 1](#). An in-depth treatment of the various contributors to anthropogenic radiative forcing is beyond the scope of this article. For more information, refer to the references at the end of the article.

'Fingerprints' in the Paleoclimatic Record

Could the warming shown in [Figure 1](#), especially of the past few decades, be a natural occurrence? Might Earth's climate undergo natural fluctuations that could result in the

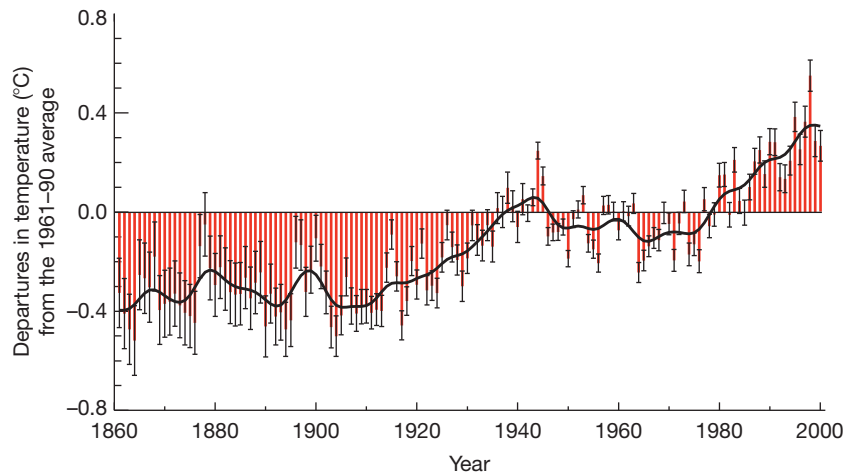


Figure 1 Variation in Earth's average global temperature from 1860 to 1999. Data are taken from global networks of thermometers, corrected for a variety of effects, and combined to produce a global average for each year. Wide bars represent temperature deviations for each year, relative to the 1961–90 average temperature, and narrow bars show uncertainties in the yearly temperatures. The black curve is a best fit to the data. Adapted from Wolfson R and Schneider SH (2002) *Understanding climate science*. In: Schneider et al. (eds.) *Climate Change Policy*, p. 1. Washington, DC: Island Press.

temperature record of [Figure 1](#)? Increasingly, it has been found that the answer to that question is 'no.' One would be in a better position to determine whether the temperature rise of the past century is natural if the record could be extended further back in time. Unfortunately, direct temperature measurements of sufficient accuracy or geographic coverage simply do not exist before the mid-1800s, if then. However, by carefully considering other quantities that do depend on temperature, climatologists can reconstruct approximate temperature records that stretch back hundreds, thousands, and even millions of years.

The paleoclimatic record can tell more about how anthropogenic changes compare to naturally induced changes in the past, both in magnitude and in rate. It can provide more information about the behavior of processes in the climate system under conditions other than present conditions. It can also impart specific information about ecosystems of the past and how ecosystems may change in the future with anthropogenic climate change.

Methods used to attribute observed warming trends to anthropogenic climate change are often called 'fingerprint analysis.' Although each fingerprint may be a circumstantial piece of evidence for human-induced climate changes, taken together, many such lines of evidence have led the Intergovernmental Panel on Climate Change (IPCC, 2001) to conclude that there is a discernible human fingerprint on climate change. For example, the Earth's stratosphere has cooled while the surface has warmed – a fingerprint of changes due to increased atmospheric CO₂ rather than, for example, a fingerprint of an increase in the heat output of the Sun that should have warmed all altitudes in the atmosphere (Schneider, 2003).

Instrumental Record

[Figure 3](#) shows three different attempts, using the same climate model, to reproduce the historical temperature record of [Figure 1](#). In the model runs of [Figure 3\(a\)](#), only estimates

of solar variability and volcanic activity – purely natural radiative forcings – were included. The projected temperature variation, represented by a thick band indicating the degree of uncertainty in the model calculations, does not show an overall warming trend and clearly is a poor fit to the actual surface temperature record. The runs of [Figure 3\(b\)](#) include only anthropogenic forcings, such as emissions of greenhouse gases and aerosols. This clearly does a much better job of reproducing the observed temperatures, especially in the late twentieth century, but deviates significantly from the historical record around midcentury. Finally, [Figure 3\(c\)](#) shows the results from runs that include both natural and anthropogenic forcings. The fit is excellent, and the model runs of [Figure 3](#) taken together strongly suggest that the temperature rise of the past few decades is unlikely to be explained without invoking anthropogenic greenhouse gases as a significant causal factor. [Figure 3](#) provides substantial circumstantial evidence of a discernible human influence on climate and, with the other evidence mentioned, supports the IPCC conclusion that most of the warming observed during the past 50 years is attributable to human activities (IPCC, 2001).

Last Millennium

[Figure 4](#) shows the results of a study that attempts to push the Northern Hemisphere temperature record back a full 1000 years (Mann et al., 1999). In this work, climatologist Michael Mann and colleagues performed a complex statistical analysis involving 112 separate indicators related to temperature, including such diverse factors as tree rings, the extent of mountain glaciers, and changes in coral reefs. The resulting temperature record of [Figure 4](#) is a 'reconstruction' of what one might expect had thermometer-based measurements been available. Although there is considerable uncertainty in the millennial temperature reconstruction, as shown by the error band in [Figure 4](#), the overall trend is most consistent with a gradual temperature decrease during the first 900 years,

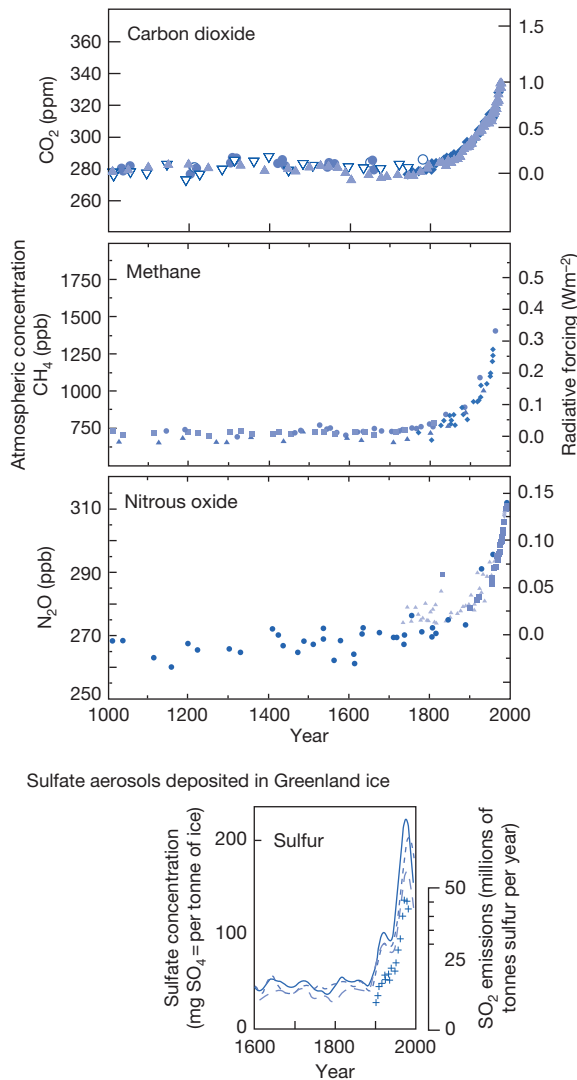


Figure 2 Records of the atmospheric concentrations of three well-mixed greenhouse gases (carbon dioxide, methane, and nitrous oxide) during the past 1000 years and records of sulfate aerosols since 1600. Adapted from IPCC (2001). In: Houghton JT, Ding Y, Griggs DJ, Noguer M, van der Linden PJ, Dai X, Maskell K, and Johnson CA (eds.) *The Scientific Basis – Contribution of Working Group I to the Third Assessment Report of the Intergovernmental Panel on Climate Change* Cambridge, UK: Cambridge University Press.

followed by a sharp upturn in the twentieth century. That upturn is a compressed representation of the thermometer-based temperature record shown in Figure 1. Among other things, Figure 4 suggests that the 1990s was the warmest decade not only of the twentieth century but also of the entire millennium. Taken in the context of Figure 4, the temperature rise of the last century clearly is an unusual occurrence.

Mann et al. (1999) correlated their temperature reconstruction with several factors known to influence climate, including solar activity, volcanism, and humankind's release of greenhouse gases. They found that solar variability and volcanism (natural influences) were the dominant influences in the first 900 years of the millennium but that much of the twentieth-

century variation could be attributed to human activity. Given the indirect, statistical nature of the study, this result can hardly be taken as conclusive evidence that humans are to blame for twentieth-century global warming. However, this result does provide independent corroboration of studies such as the one in Figure 3, which also indicate a human influence on climate. Similar results have also been shown in other studies, such as Jones et al. (2001) and Mann et al. (1998). This record has been pushed back two millennia (Mann and Jones, 2003), yielding similar results.

Ice Cores

Ice cores bored from the Greenland and Antarctic ice sheets provide estimates of both temperature and atmospheric CO₂ dating back hundreds of thousands of years (Jouzel et al., 1987; Petit et al., 2000). The ice cores have yielded a continuous record of the past 740 000 years (EPICA Community Members, 2004), and scientists expect that the record may be extended 1 million years into the past (White, 2004). Variations in ice density associated with seasonal snowfall patterns provide a highly time-resolved calibration of the age of given points in a given ice core, and analysis of air bubbles trapped in ancient ice gives an indication of CO₂ concentration. Temperature inference is accomplished by comparing a ratio of oxygen isotopes that is sensitive to temperature. Temperature differentially affects the concentration of the isotopes in evaporating water, and therefore their concentration in precipitation and in the ice. The result of such an ice core analysis, shown in Figure 5, gives dramatic evidence that temperature and CO₂ concentration are correlated over the long term.

It is not clear from the graph alone that the CO₂ variations in Figure 5 are the cause of the temperature changes. Sometimes, a CO₂ increase precedes a warming, but sometimes it does not. In fact, climatologists suspect a feedback process whereby a slight increase in temperature, probably caused by subtle changes in the Earth's orbit, results in an increase in atmospheric CO₂ through a variety of mechanisms, such as the release of CO₂ dissolved in the oceans (IPCC, 2001). The increased atmospheric CO₂, in turn, leads to greenhouse warming, amplifying the initial temperature increase. The result is a nearly simultaneous and substantial increase in both CO₂ and temperature. Eventually, orbital changes trigger a modest temperature decrease, and again feedback mechanisms amplify the decrease, driving down both CO₂ and temperature. Some paleoclimatologists also posit biotic feedback mechanisms, such as an initial cooling that causes a drying of the continents, producing more windblown dust (IPCC, 2001). This dust contains minerals needed by phytoplankton in the oceans. As dust settles on the ocean surface, it fertilizes these tiny oceanic organisms. The phytoplankton, in turn, increase their productivity by drawing down atmospheric CO₂, thus making the move toward an Ice Age even more rapid and deep. However, despite such complexities, there is still much regularity in the record. The pattern of varying temperature and CO₂ concentration shown in Figure 5 is believed to repeat on a timescale of roughly every 100 000 years over most of the past 1 million years, at least in part as a result of periodic changes in the Earth's orbit and the inclination of its polar axis.

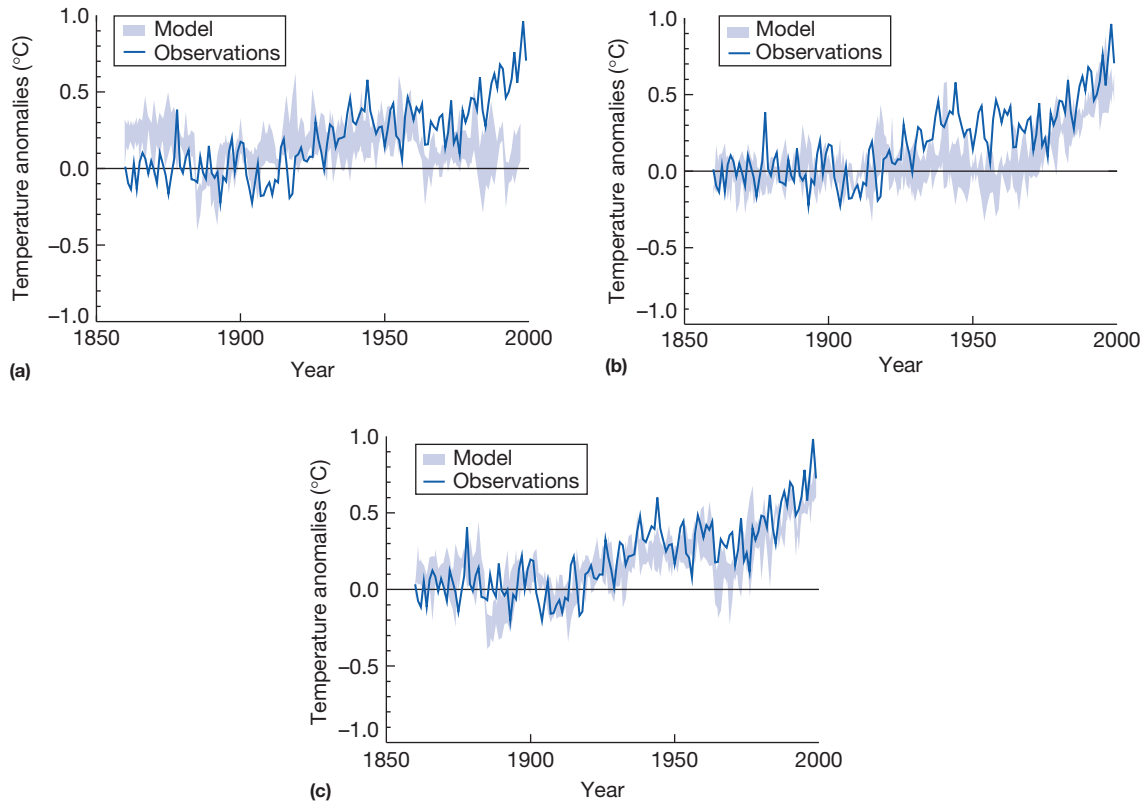


Figure 3 Attempts to model Earth's temperature from the 1860s using different model assumptions. In all three graphs, the solid curve is the observed surface temperature record of Figure 1. Gray bands represent model projections. In each graph, the bands encompass the results of four separate model runs. In (a), only natural forcings – volcanic activity and solar variability – are included. Clearly, this simulation lacks the upward trend in the observed temperature record, suggesting that the temperature rise of the past century and a half is unlikely to have a purely natural explanation. Simulation (b), including only anthropogenic forcings, does much better, especially with the rapid temperature increase of the late twentieth century. Simulation (c), combining both natural and anthropogenic forcings, shows the best agreement with observations. Adapted from Wolfson R and Schneider SH (2002) Understanding climate science. In: Schneider et al. (eds.) *Climate Change Policy*, p. 1. Washington, DC: Island Press. Data from IPCC (2001). In: Houghton JT, Ding Y, Griggs DJ, Noguer M, van der Linden PJ, Dai X, Maskell K, and Johnson CA (eds.) *The Scientific Basis – Contribution of Working Group I to the Third Assessment Report of the Intergovernmental Panel on Climate Change*. Cambridge, UK: Cambridge University Press.

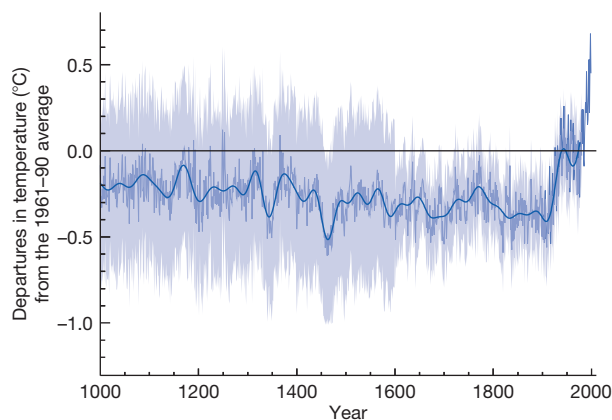


Figure 4 Reconstruction of the 1000-year temperature record for the Northern Hemisphere. The black curve is a best fit to the millennial temperature record; gray is the 95% confidence interval, meaning that there is a 95% chance that the actual temperature falls within this band. Data from the mid-19th century on are from the thermometer-based temperature record of Figure 1. Adapted from Wolfson R and Schneider SH (2002) Understanding climate science. In: Schneider et al. (eds.) *Climate Change Policy*, p. 1. Washington, DC: Island Press.

Note that Figure 5 shows brief periods of warmth punctuated by much longer, cooler Ice Ages, which are characterized by dramatically different climatic conditions, with ice sheets 2 km thick covering what is now Canada, the northeastern United States, and northwestern Europe and engulfing high mountain plateaus throughout the world. Today, humans enjoy the warmth of an interglacial period, but not long ago, geologically speaking, conditions were very different. Figure 5 shows that the difference in temperature between an interglacial and an Ice Age is on the order of 6 °C. Climate models driven by standard assumptions about population, land use, and energy consumption project a warming over the next century of 1.5–6 °C. The difference between the higher and lower ends of this range has substantial implications for sea-level rise, extreme weather, redistribution of species ranges, and other impacts. Policymakers and the general public often ask how a few degrees can matter all that much. Figure 5 provides one striking answer: Downward changes on the same order as the largest projected warming are enough to make the difference between the current climate and an Ice Age.

Note in Figure 5 that the maximum CO₂ concentration in the ice core record of the past 160 000 years is less than

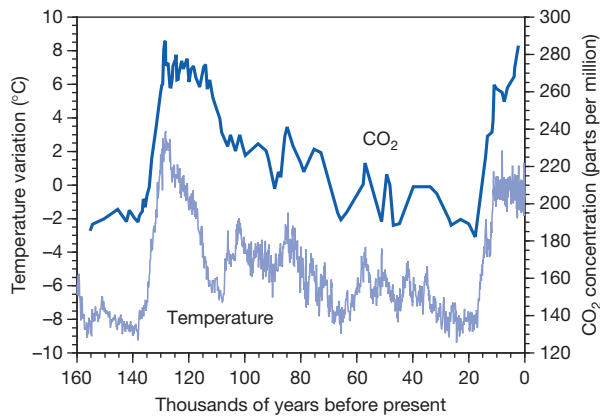


Figure 5 Atmospheric carbon dioxide (top curve) and temperature variation (bottom curve) during the past 160,000 years, from ice cores taken at Vostok, Antarctica. The record shows long stretches of low temperature (Ice Ages) separated by brief, warm interglacial periods. The correlation between CO₂ and temperature is quite obvious. Note also the small change, averaging perhaps 6 °C, between the present warm climate and the recent Ice Age. Data do not extend to the present, but stop well before the industrial era. Adapted from Wolfson R and Schneider SH (2002) *Understanding climate science*. In: Schneider et al. (eds.) *Climate Change Policy*, p. 1. Washington, DC: Island Press. CO₂ data are from Petit JR et al. (2000) *Historical isotopic temperature record from the Vostok ice core*. In: *Trends: A Compendium of Data on Global Change*. Oak Ridge, TN: Oak Ridge National Laboratory, Carbon Dioxide Information Analysis Center. Temperature data are from Jouzel J et al. (1987) *Vostok ice core: A continuous isotope temperature record over the last climatic cycle (160,000 years)*. *Nature* 329: 403–408, as reproduced in the Carbon Dioxide Information Analysis Center.

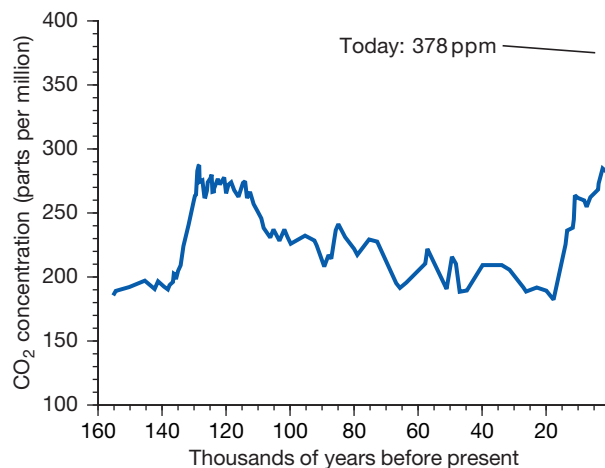


Figure 6 The CO₂ record of Figure 5, with data up to 2004 included. The CO₂ rise of Figure 2 is shown here as a dramatic jump to levels not seen on Earth for hundreds of thousands (and probably millions) of years. Adapted from Wolfson R and Schneider SH (2002) *Understanding climate science*. In: Schneider et al. (eds.) *Climate Change Policy*, p. 1. Washington, DC: Island Press.

300 ppm. This does not include the very recent past, but only the preindustrial period. The present-day concentration of 378 ppm is far above anything the Earth has seen, probably for millions of years. Figure 6 shows the effect of adding the

recent rise in CO₂ to the ice core data. Clearly, the anthropogenic increase in CO₂ concentration is unprecedented in both its size and its rapidity. The human race has made truly dramatic changes in the Earth's atmosphere during the past century or so, and significant climate change can be expected as a result of these changes almost certainly.

Finally, a very unique 'fingerprint' study was conducted by Ruddiman (2003). Ruddiman believes that atmospheric concentrations of CO₂ and CH₄ were first altered by humans thousands of years ago, resulting from the discovery of agriculture and subsequent technological innovations in farming. Ruddiman analyzed the Pleistocene CO₂ record by examining the Vostok ice core and came to the conclusion that the interglacial life cycle of CO₂ in the present interglacial has varied from the other three interglacials occurring in the past 400,000 years. In typical interglacials, CO₂ peaks early and then stabilizes or decreases, but in the Holocene interglacial, CO₂ levels fell early on and then began to recover. Ruddiman believes the main difference between this abnormal interglacial and the others is the presence of land clearing and agriculture. He points to two ways in which land clearing typically increases CO₂ concentrations: burning and decomposition of downed vegetation and increased weathering of soil carbon in a system that has been disturbed (i.e., tilled). These forces, Ruddiman believes, warmed the Earth by a global mean of approximately 0.8 °C in recent millennia (2 °C at higher latitudes) and have contributed to an uncharacteristically stable 10,000 years. Although there are many unresolved uncertainties in this work, this is an area ripe for additional research.

Climate Processes

Beyond changes in mean temperature, the paleoclimatic record can also tell us much about past climate variability, including interannual variability driven by such processes as El Niño–Southern Oscillation (ENSO), North Atlantic Oscillation, and changes in the meridional overturning circulation in the North Atlantic, also known as the thermohaline circulation. Again, other articles in this encyclopedia focus specifically on these issues, and therefore a detailed treatment of these issues has not been presented here. However, study of the past behavior of the climate system may allow scientists to project more confidently the changes in climate processes that could occur under anthropogenic climate change.

For example, a number of studies have used paleoclimatic proxies to produce reconstructions of ENSO variability prior to the instrumental record (Mann et al., 2000; Stahle et al., 1998), focusing on proxies for sea surface temperature in the equatorial Pacific, where key processes determining the evolution of the ENSO cycle occur. Jones et al. (2000) compared these reconstructions back to the early 1700s and compare them to observational data back to the late 1800s. They found that both records exhibit similar interannual and interdecadal variability, and that the recent period marked by frequent warm phase (El Niño) events is unique in one record (Stahle et al., 1998) and is unique since 1800 in the other (Mann et al., 2000), although uncertainty in the reconstructions prevents high confidence in such conclusions. However, such findings lend plausibility to projections of future changes in the ENSO cycle

under anthropogenic climate change, such as Timmermann et al. (1999) suggested. Timmermann et al. used a complex global climate model to simulate the ENSO cycle under future climate change and found that in their model, more frequent warm-phase (El Niño) conditions and more intense cold-phase (La Niña) events occurred.

Climate Modeling and the Paleoclimatic Record

Several ways in which the paleoclimatic record can directly provide information about the climate system and potential responses of the system to anthropogenic climate change have been discussed. Simulation of the climate system is a complex undertaking. The ideal model would include all processes known to have climatological significance and would involve fine-enough spatial and temporal scales to resolve salient phenomena occurring over limited geographical regions and relatively short time spans. Today's most comprehensive models strive to reach this ideal but still entail many compromises and approximations. As mentioned previously, the paleoclimatic record also provides the backdrop against which climate modelers can calibrate and test their models, lending their future projections more credibility. This section provides several examples of ways in which the paleoclimatic record is used in climate modeling, covering model validation, model calibration, and modeling of ecosystem dynamics.

Model Validation

How can modelers be more confident in their model results? How do they know that they have taken into account enough climatologically significant processes, and that they have satisfactorily parameterized processes for which the spatial scale are below that of their models' grid cells? The answer lies in a variety of model validation techniques, most of which involve evaluating a model's ability to reproduce known climatic conditions in response to known forcings.

One way to gain confidence in a model's ability to predict future climate is to evaluate its accuracy in modeling past climates. Can the model, for example, reproduce the temperature variations from 1860 to the present charted in Figure 1? Figure 3 is an 'experiment' with one climate model to do just that. The excellent fit between observed and modeled temperatures, in Figure 3(c), suggests greater confidence in this model's projections of future climate.

In one of the seminal comparisons of paleoclimatic data and climate models, members of the Cooperative Holocene Mapping Project (COHMAP) compiled an extensive library of paleoclimatic data to create a reconstruction of the climate of the past 18 000 years and compared the data to modeled representations of the paleoclimate using the National Center for Atmospheric Research's Community Climate Model (CCM). Their well-known *Science* article (COHMAP Members, 1988) summarizes many of their comparisons. Basing their modeling of the changes in the seasonal and latitudinal distribution of solar radiation predicted by the Milankovitch theory, the COHMAP project was able to reproduce many of the paleoclimatic conditions consistent with their paleoclimatic data set of pollen, plankton, and lake level records. For example, Figure 7

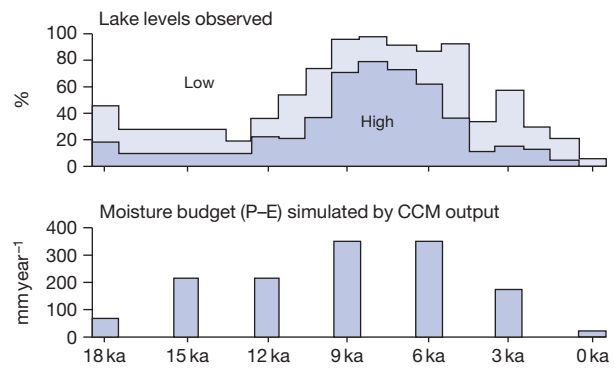


Figure 7 Comparison of observed lake levels in the northern tropics during the past 18 000 years with a model simulation of the moisture budget (precipitation – evaporation). Adapted from COHMAP Members (1988) Climatic changes of the last 18,000 years: Observations and model simulations. *Science* 241: 1043–1052.

shows a comparison of observed lake levels in the northern tropics during the past 18 000 years with a simulation of the moisture budget (precipitation minus evaporation) by CCM. Although the agreement is not perfect, the model output correctly displays a maximum for the moisture budget of 6–9 ka, when lake levels were at their maximum. This represents one of the first regional-scale validations of a complex climate model.

Anthropogenic radiative forcings projected to 2100 are different than any known in the past, and there is no direct way of empirically testing the model's performance – except to wait until the Earth's climate system performs the 'experiment.' However, testing the models against such known paleoclimatic and paleoecological variations gives considerable confidence that these models are treating the essential climate-determining processes with reasonable fidelity, and it can therefore be expected of them to produce moderately realistic projections of future climate, at least at the large scale, given credible emissions scenarios.

Variations in the projections of different models are still expected, and because future greenhouse gas emissions depend on human behavior, future projections will differ depending on what assumptions modelers make about the human policy responses to the prospect of global warming. The uncertainties in projections of human behavior cause about as much spread in estimates of future warming as do uncertainties in the understanding of the sensitivity of the climate system to specified radiative forcings.

Model Calibration

Much of the uncertainty in current understanding of the climate system is captured by the so-called climate sensitivity, defined as the equilibrium global mean surface temperature increase from a doubling of atmospheric CO₂. Using general circulation models, the IPCC has long estimated the climate sensitivity to be between 1.5 and 4.5 °C (IPCC, 2001), although other studies have produced distributions wider than the IPCC range, with significant probability of climate sensitivity above 4.5 °C (Andronova and Schlesinger, 2001; Forest et al., 2001; Stainforth et al., 2005). A detailed

description of the factors that contribute to uncertainty in climate sensitivity is beyond the scope of this article. However, the paleoclimatic record is a key source of information for determining the plausible range for this important parameter.

Given the uncertainties inherent in the feedback processes that must be incorporated into climate models, is there any way to estimate climate sensitivity empirically? One way is to use historical paleoclimatic situations in which both forcing and temperatures were different than those of the modern era. Hoffert and Covey (1992) attempted this estimation and arrived at a climate sensitivity in the range of 2–3 °C for a doubling of CO₂. Of course, precise forcings and temperatures are not available for the distant past, so a considerable degree of uncertainty exists in such empirical attempts. Nevertheless, they do add confidence that current ranges of climate sensitivity are not wildly unrealistic, and they suggest that the probability of climate sensitivities much above or below the long-standing 1.5–4.5 °C estimated range of warming for a doubling of CO₂ is not high.

More precise information on both temperature trends and radiative forcings exists for the past 50 years. If one were to estimate climate sensitivity by tracking how much warming occurred during the past 50 years and scaling this warming by the amount of radiative forcing (i.e., how many watts of radiation energy is added per square meter of earth) caused by greenhouse gas increases, a ratio of temperature rise to forcing (i.e., climate sensitivity) could be estimated. Thus, to empirically estimate the climate sensitivity, all the forcings need to be estimated and aggregated to find the net overall forcing number. Unfortunately, there will continue to be a very large range of uncertainty associated with estimates of the all-important climate sensitivity parameter. In fact, uncertainty in radiative forcings is at least a factor of three.

Andronova and Schlesinger (2000), for example, attempted to create a broad probability distribution for climate sensitivity by scaling the observed temperature trends against a range of possible forcings. Because of the uncertainties in forcing, they obtained a very broad distribution. Approximately 50% of the estimates for climate sensitivity lie outside of the IPCC's 1.5–4.5 °C range. Other groups have attempted semiempirical estimation of climate sensitivity, obtaining similar results to those of Andronova and Schlesinger (Forest et al., 2001; Murphy et al., 2004).

Hegerl et al. (2003) completed a similar analysis using a reconstruction of temperature and radiative forcing during the past millennium, obtaining similar results. During this longer time period, aerosol forcing, which is the dominant source of uncertainty in recent times, is relatively unimportant, and the responses to relevant forcings (greenhouse gases, solar variability, volcanism, and land use change) can be separated more effectively than for the shorter instrumental period.

Ecosystem Dynamics

Since anthropogenic climate change is proceeding, on a sustained globally averaged basis, at an order of magnitude faster than Pleistocene/Holocene climatic changes, dramatic and differential responses of plant, insect, mammal, and bird species can be anticipated. Thus, a very large restructuring and reorientation of biological communities seems likely. Future

climates may not only be quite different from more recent previous climates but also be different from those inferred from paleoclimatic data and from those to which some existing species are evolutionarily adapted. Therefore, these data cannot provide direct analogs for future assemblages under anthropogenic climate change, but they can provide a means to calibrate and verify models to predict future ecosystem dynamics (Crowley, 1993; Schneider, 1993).

For example, Figure 8 displays paleovegetation maps during the past 18 000 years, reconstructed using the method of modern analogs and more than 13 000 samples of fossils and modern pollen. During the last Ice Age, most of Canada was under ice; pollen cores indicate that as the ice receded, boreal trees moved northward chasing the ice cap. One interpretation of this information was that biological communities moved intact with a changing climate. If this were true, the principal ecological concern regarding the prospect of future climate change would be that human land-use patterns might block what had previously been the free-ranging movement of natural communities in response to climate change. However, analyses by COHMAP, mentioned previously, incorporating multiple pollen types found that although nearly all species moved north, as expected, during the transition from the last Ice Age to the present interglacial, the distribution of pollen types provided no analogs to today's vegetation communities during a significant portion of the transition period (Overpeck et al., 1992). That is, whereas all species moved, they did so individually, not as groups of different species. The relevance of this is that in the future ecotypes will not necessarily move as a unit as climate changes (assuming there is time and space for such a migration). Additionally, past vegetation response to climatic change at an average rate of 1 °C per millennium indicates that credible predictions of vegetation changes from comparable or more rapid changes projected from human activities cannot neglect the time-evolving (transient) dynamics of the ecological system.

Conclusions

A brief description of the relevance of paleoclimatic research to global climate change has been given. Paleoclimatic research updates the human understanding of past climatic fluctuations, of the impacts of past climatic changes on physical processes and natural ecosystems, and of human capacity to model the climate system. Even if humans get the greenhouse gas emissions under control quickly, global temperature will continue to rise, and the effects of global climate change, particularly sea-level rise, will almost certainly continue to increase well beyond the end of the twenty-first century. The paleoclimatic record provides a critical source of information for understanding the climate of today and modeling the climate of the future.

See also: [Glacial Climates](#); [Biosphere Feedbacks](#); [Thermohaline Circulation](#). **Glaciation, Causes:** [Astronomical Theory of Paleoclimates](#). **Introduction:** [Societal Relevance of Quaternary Research](#). **Paleoclimate Modeling:** [Data–Model Comparisons](#); [Quaternary Environments](#). **Paleoclimate Reconstruction:** [Approaches](#); [The Last Millennium](#).

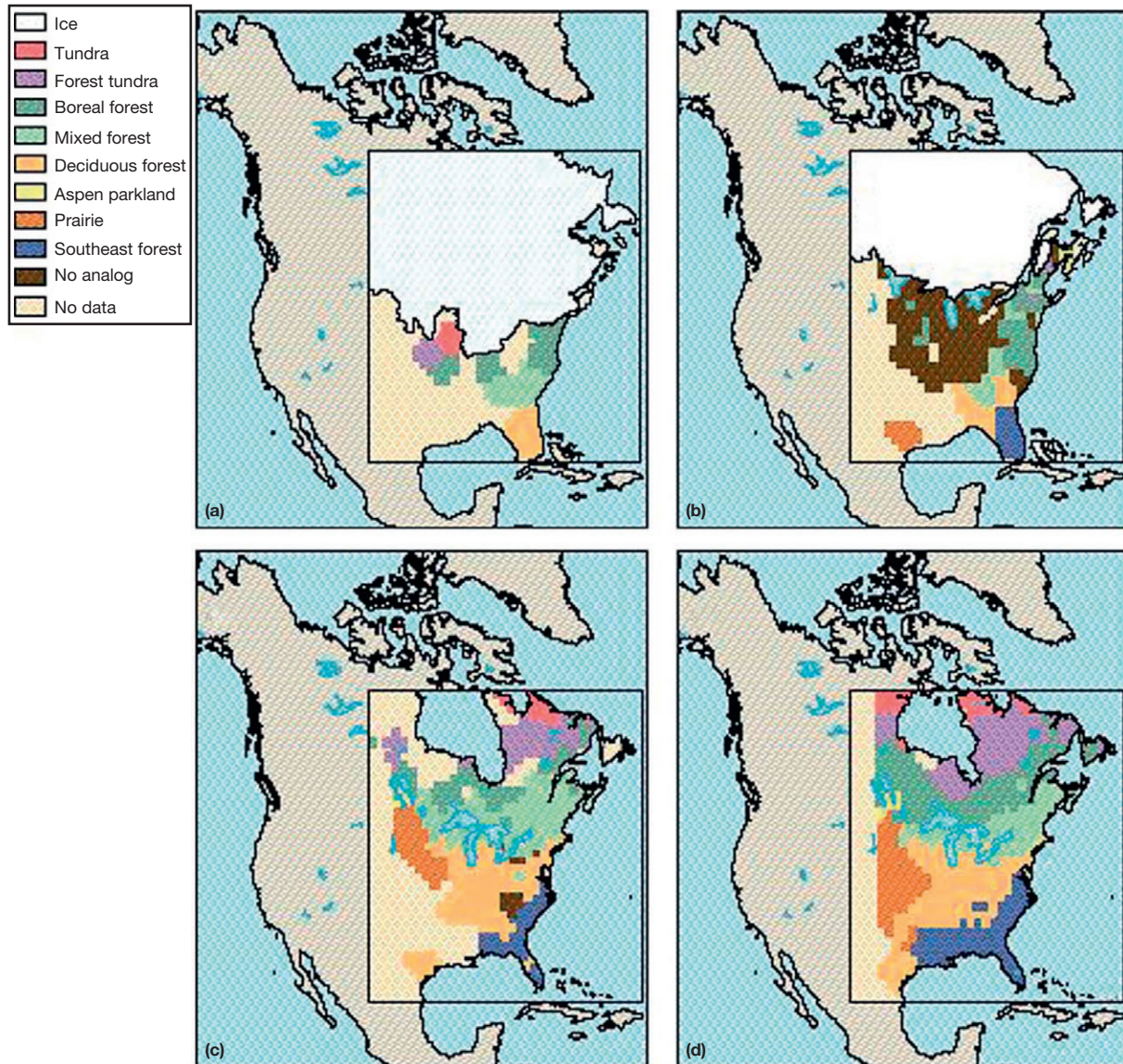


Figure 8 Selected paleovegetation maps reconstructed by using the method of modern analogs and more than 13 000 samples of fossils and modern pollen: (a) 18 000 ka, (b) 12 000 ka, (c) 6000 ka, and (d) modern. No vegetation is mapped in areas without fossil pollen data, and no analog refers to vegetation without any modern analog. Adapted from Root TL and Schneider SH (1998) Factors affecting biological resources – Climate change. In: Mac MJ et al. (eds.) *Status and Trends of the Nation's Biological Resources*, p. 89. Reston, VA: U.S. Geological Survey. Data from Overpeck JT et al. (1992) Mapping eastern North American vegetation change over the past 18,000 years: No analogs and the future. *Geology* 20: 1071–1074.

References

- Andronova NG and Schlesinger ME (2001) Objective estimation of the probability density function for climate sensitivity. *Journal of Geophysical Research* 106(22): 605–612.
- COHMAP Members (1988) Climatic changes of the last 18,000 years: Observations and model simulations. *Science* 241: 1043–1052.
- Crowley T (1993) Use and misuse of the geologic 'analog' concept. In: Eddy JA and Oeschger H (eds.) *Global Changes in the Perspective of the Past*, p. 17. Chichester, UK: Wiley Dahlem Workshop Report ES12.
- EPICA Community Members (2004) Eight glacial cycles from an Antarctic ice core. *Nature* 429: 623–628.
- Forest CE, et al. (2001) Quantifying uncertainties in climate system properties with the use of recent climate observations. *Science* 295: 113–117.
- Hegerl GC, et al. (2003) Detection of volcanic, solar and greenhouse gas signals in paleo-reconstructions of Northern Hemispheric temperature. *Geophysical Research Letters* 30: 1242–1245.
- Hoffert MI and Covey C (1992) Deriving global sensitivity from palaeoclimate reconstructions. *Nature* 360: 573–576.
- Intergovernmental Panel on Climate Change (IPCC) (2001) In: Houghton JT, Ding Y, Griggs DJ, Noguer M, van der Linden PJ, Dai X, Maskell K, and Johnson CA (eds.) *The Scientific Basis – Contribution of Working Group I to the Third Assessment Report of the Intergovernmental Panel on Climate Change*. Cambridge, UK: Cambridge University Press.
- Jones PD, et al. (2000) The evolution of climate over the last millennium. *Science* 292: 662–667.
- Jouzel J, et al. (1987) Vostok ice core: A continuous isotope temperature record over the last climatic cycle (160,000 years). *Nature* 329: 403–408.
- Mann ME, Bradley RS, and Hughes MK (1998) Global-scale temperature patterns and climate forcing over the past six centuries. *Nature* 392: 779–787.
- Mann ME, Bradley RS, and Hughes MK (1999) Northern Hemisphere temperatures during the past millennium: Inferences, uncertainties, and limitations. *Geophysical Research Letters* 26: 759–762.

- Mann ME, Bradley RS, and Hughes MK (2000) Long-term variability in the El Niño/Southern Oscillation and associated teleconnections. In: Diaz HF and Markgraf V (eds.) *El Niño and the Southern Oscillation: Multiscale Variability and Global and Regional Impacts*, p. 357. Cambridge, UK: Cambridge University Press.
- Mann ME and Jones PD (2003) Global surface temperatures over the past two millennia. *Geophysical Research Letters* 30: 1820–1823.
- Murphy JM, et al. (2004) Quantification of modeling uncertainties in a large ensemble of climate change simulations. *Nature* 430: 768–772.
- Overpeck JT, et al. (1992) Mapping eastern North American vegetation change over the past 18,000 years: No analogs and the future. *Geology* 20: 1071–1074.
- Petit JR, et al. (2000) Historical isotopic temperature record from the Vostok ice core. In: *Trends: A Compendium of Data on Global Change*. Oak Ridge, TN: Oak Ridge National Laboratory, Carbon Dioxide Information Analysis Center.
- Ruddiman WF (2003) The anthropogenic greenhouse era began thousands of years ago. *Climatic Change* 61: 261–293.
- Schneider SH (1993) Can paleoclimatic and paleoecological analyses validate future global climate and ecological change projections? In: Eddy JA and Oeschger H (eds.) *Global Changes in the Perspective of the Past*, p. 317. Chichester, UK: Wiley.
- Schneider SH (2003) Imaginable surprise. In: Potter TD and Colman BR (eds.) *Handbook of Weather, Climate, and Water*, p. 947. Chichester, UK: Wiley.
- Stahle DW, et al. (1998) Experimental dendroclimatic reconstruction of the Southern Oscillation. *Bulletin of the American Meteorological Society* 79: 2137–2152.
- Stainforth DA, et al. (2005) Uncertainty in predictions of the climate response to rising levels of greenhouse gases. *Nature* 433: 403–406.
- Timmermann A, et al. (1999) Increased El Niño frequency in a climate model forced by future greenhouse warming. *Nature* 398: 694–697.
- White JWC (2004) Do I hear a million? *Science* 304: 1609–1610.

Introduction

The term paleohydrology was first used in the fluvial literature by Leopold and Miller (1954) in their Geological Survey Water-Supply Paper on the postglacial chronology of alluvial valleys in Wyoming (USA). Leopold and Miller recognized that the interactions of climate, vegetation, stream regime and runoff would have been different under varying past climates and outlined a research agenda that is still relevant today. For example, in relation to the climate versus human impact debate, they stated that an "... understanding of our soil erosion problems and of the efficacy of control measures requires that we sharpen our tools for distinguishing those features of the modern landscape molded by natural forces alone and those related to activities of man. Geomorphology can provide techniques that will help us" (Leopold and Miller, 1954, p. 75). It was Stanley Schumm who developed paleohydrological concepts further, outlining his ideas in the *The Fluvial System* (Schumm, 1977). Therein, he defined paleohydrology as the "science of the waters of the Earth, their composition, distribution, movement on ancient landscapes from the occurrence of the first rainfall to the beginning of continuous hydrological records" (Schumm, 1977, p. 42). This somewhat cumbersome definition contradicts the neat concepts he articulated; Schumm recognized that rivers would respond in predictable ways, at least qualitatively, in response to changes in discharge regime and sediment loads. For the purposes of this article, fluvial paleohydrology can be defined as the study of fluvial processes and their hydrological implications before the onset of instrumental records (modified from Brown, 1997) and covers, therefore, the response to global change of: runoff and discharge regime; sediment transport and storage; and extreme floods.

Throughout the majority of the 50-plus-year history of fluvial paleohydrology, studies were primarily descriptive and qualitative in nature; however, recent methodological advances (Brown et al., 2009), for example, in (a) airborne and terrestrial light detection and ranging (LiDAR); (b) sub-surface remote sensing; (c) geographic information systems (GIS); (d) landscape, hydrological and flood hydraulic models; and (e) geochronology (e.g., accelerator mass spectroscopy (AMS) C-14 and optically stimulated luminescence (OSL) dating techniques), have enabled a quantification revolution in fluvial paleohydrology. This article will review recent developments in our understanding of Holocene fluvial systems through these methodological advances, with a particular focus on: (a) quantification of basin scale sediment transport and storage; (b) the quantification of paleoflood discharges; and (c) key debates regarding the drivers of fluvial response over different spatial and temporal scales, in particular, the relative roles of allogenic drivers (e.g., climate and land use land cover (LULC)) and autogenic drivers. Autogenic processes are particularly important to quantify and understand as they

cloud the links between fluvial system response and the controls of LULC and climate (e.g., Van De Wiel and Coulthard, 2010), identified as key issues in sustainable river management aimed at mitigating negative impacts of future global change. Indeed it is pertinent to note here the role of fluvial paleohydrology in contemporary river management as it is important for river managers to understand the long-term history and geomorphological evolution of fluvial systems (Sear and Arnell, 2006). This is not only due to changing fluvial processes over decadal and centennial timescales but also due to the long history of river management in many regions of the world (Hudson et al., 2008).

Long-Term Sediment Transport and Storage

Fluvial paleohydrological research over the last two decades has demonstrated that natural early Holocene floodplains, certainly in lowland low-energy river systems, were likely to be stable wetland environments (Walter and Merritts, 2008), with systems such as the Narew River in Poland, one of the few remaining anastomosing fluvial wetland environments to have survived channel management and floodplain drainage (Gradzinski et al., 2003), serving as a modern analog for many early Holocene rivers. Subsequent human impact over varying spatial and temporal scales has led to increased sediment loads and floodplain sedimentation rates resulting in floodplain alluviation, especially over recent millennia in a European context (Notebaert and Verstraeten, 2010). It is likely that these changes have fundamentally altered fluvial processes and river hydraulics. This was shown by Knox (2006) in the Upper Mississippi basin, where the relatively recent post-colonial human impacts of mining and agriculture have led to alluviation, while channel bed elevations have remained stable. This resulted in an increase in river channel capacity from 3.6 m² in AD 1832 to 24.8 m² in 1990, thereby increasing bankfull flood discharge and other flow hydraulics, such as bed shear stress and stream power, thus potentially increasing the geomorphic activity of the river. As de Moor and Verstraeten (2008) demonstrate, river systems shifted from a natural sediment supply (detachment)-limited system to an anthropogenic capacity (transport)-limited system with a high proportion of sediment stored as colluvium or in alluvial floodplains. In this section, sediment budget research aimed at quantifying long-term sediment transport pathways and storage will be discussed, as this has implications for the sensitivity of river systems to global change, and in particular climatic variability. For a detailed overview of methodologies, readers are directed to Brown et al. (2009).

It is only recently that there has been sufficient field data, allied to newly available data handling techniques (Brown et al., 2009), to enable the quantification of sediment storage at large spatial scales, such as the Rhine basin (Hoffmann et al.,

2009). Earlier studies tended to be qualitative, or focused on small spatial and/or temporal scales. Some of the most detailed quantitative sediment budgets available have been reconstructed for river basins of the European loess belt (de Moor and Verstraeten, 2008; Notebaert et al. 2009; Verstraeten et al., 2009). In the 55-km² Nethen basin, a tributary of the Dijle basin (Figure 1), Verstraeten et al. (2009) demonstrated that 38% of sediment eroded during the Holocene was stored in colluvium and dry valleys, with 23% as alluvium. In the Geul River catchment (380 km²), de Moor and Verstraeten (2008) applied a two-dimensional erosion and hillslope delivery model (WATEM/SEDEM), combined with geomorphic field data, to demonstrate that over the last millennium >80% of eroded sediment was stored as colluvium, suggesting the basin has been a capacity-limited system over this time span. Prior to this period, less sediment was generated within the basin and a high percentage was transported to the main river valleys or exported out of the system, thus the natural fluvial system was supply limited (de Moor and Verstraeten, 2008). In the Dijle catchment (758 km²), the sediment budget illustrated in Figure 1(b) shows that the total estimated erosion for the Holocene was 817 ± 66 Mt, of which 327 ± 34 Mt was deposited as colluvium and 352 ± 11 Mt on alluvial floodplains

(Notebaert et al., 2009). The sensitivity of sediment dynamics in the catchment was tested by Notebaert et al. (2011) using a modeling approach, the WATEM/SEDEM sediment model complemented through runoff modeling using a spatially distributed model (STREAM). The key findings demonstrate that of the observed increase in soil erosion from the early Holocene to the present (Notebaert et al., 2009), 9% of the increase is due to climatic variations while 6000% is due to land use change. These dramatic shifts account for the changing floodplain characteristics documented from floodplain boring surveys (Figure 1(a)), with a shift from peat-dominated stable wetland systems to alluviated, fine-grained floodplains, as also noted in the United States (Walter and Merritts, 2008). Regarding the modeled runoff, the mean discharge shows a possible slight increase from the early to late Holocene but there is greater change in peak flows (Figure 1(c)). For example, the modeled mean discharge for the Dijle catchment pristine scenario (4950–4851 BC) is 6.8 ± 1.3 m³ s⁻¹, compared to 7.2 ± 1.9 m³ s⁻¹ for the modern period (AD 1900–1999), while maximal discharges increase from 27.7 to 119.9 m³ s⁻¹ over this same period (Notebaert et al., 2011). The influence on extreme events can also be seen in the recurrence interval curves (Figure 1(d)) where the lowest discharges for the

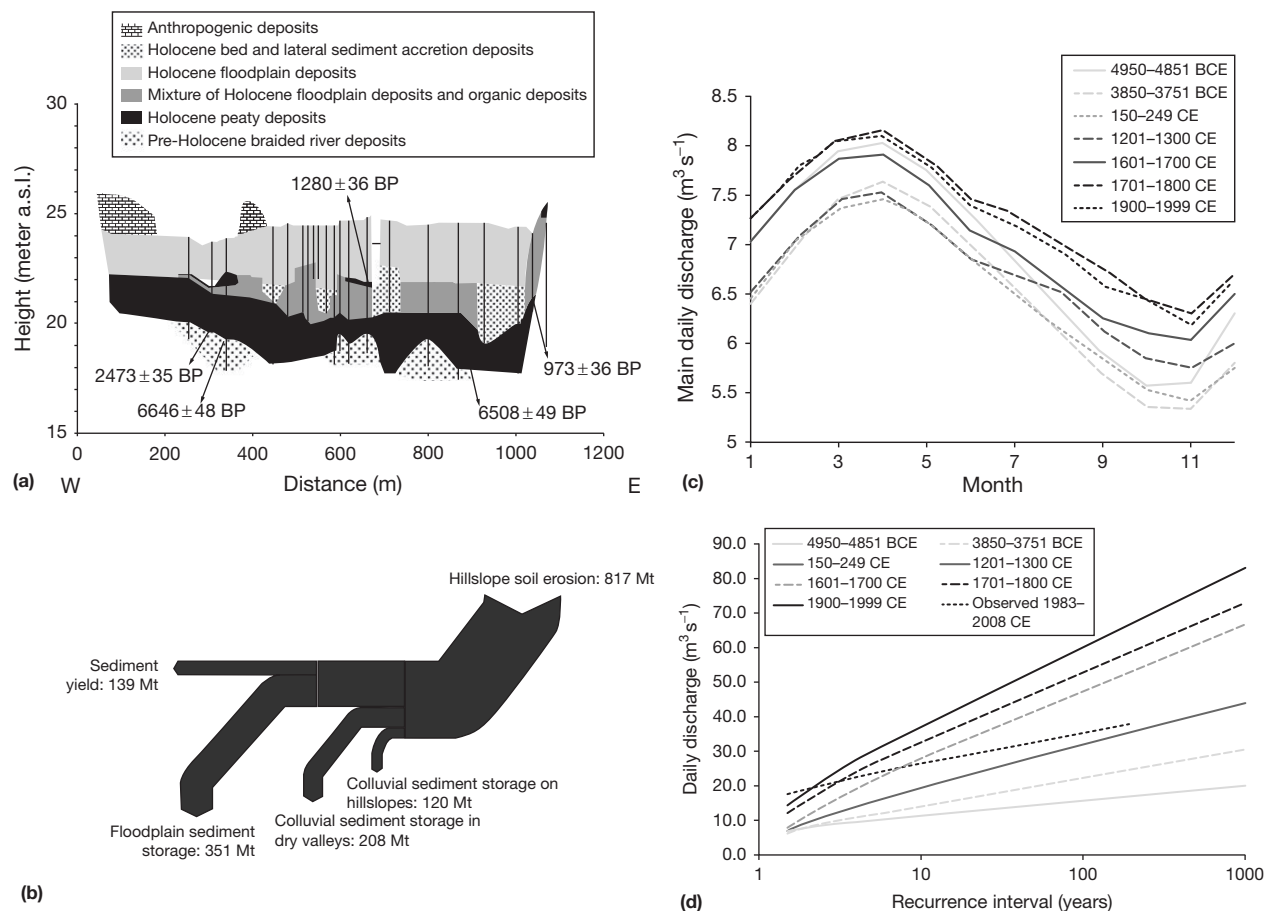


Figure 1 Holocene sediment budget and paleohydrology of the Dijle River, Belgium. (a) A typical valley cross-section showing the peaty nature of the natural floodplain environment and anthropogenic alluviation, which increases in rate over the last millennium (Notebaert et al. 2009). (b) Holocene sediment budget for the Dijle River (Notebaert et al., 2009). (c) Modeled mean daily discharge for various Holocene LULC scenarios (Notebaert et al., 2011). (d) Recurrence interval curves for the same Holocene LULC scenarios as in (c) (Notebaert et al., 2011).

100- and 1000-year recurrence intervals are modeled for the older scenarios (4950–4851 BC and 3850–3751 BC). The increase in sedimentation due to human impact noted in small to medium basins of the loess belt is also evident at larger spatial scales in Europe, for example, in the Rhine basin, though the larger the spatial scale the more problematic it is to link cause and effect (Hoffmann et al., 2009). While a combination of sediment budgets and hydrological modeling attest to the importance of human impact in controlling sediment dynamics and fluvial regime in Western Europe (Notebaert et al., 2011), it is problematic in alluvial floodplain settings to test the variability of extreme hydrological events. Sedimentary flood records located in bedrock gorge settings, where flood magnitude can be reconstructed, will be considered in the next section.

Flood Response to Holocene Climatic Variability

Paleoflood hydrology, a subdiscipline of paleohydrology, is the reconstruction of the magnitude and frequency of floods, recorded in the instrumental flood record, using geomorphological evidence (based on the definition of Baker et al., 2002). Paleoflood hydrological techniques have primarily been developed through two research themes: (a) the quantification of extreme Holocene floods; and (b) the reconstruction of late Pleistocene megafloods (e.g., Carling et al., 2010). Paleoflood research combines geomorphic mapping of flood-related landforms and sediments (e.g., flood benches); the sedimentology and stratigraphy of flood deposits (Benito et al., 2003), including geochronological control; and hydraulic flood modeling (Webb and Jarrett, 2002), as used in conventional hydrology, to quantify paleodischarge. This section will focus on Holocene paleofloods and flood–climate relationships, though the role of human impact on flooding has also been documented (e.g., Benito et al., 2010).

Narrow bedrock gorges are the main geomorphic setting for paleoflood studies (Benito et al., 2003). The reasons for this are twofold: firstly, the channel geometry and valley cross-section is relatively stable so that channel changes are restricted and/or readily quantifiable; and secondly floodwater stage is sensitive to flood magnitude. This latter point can be seen in Figure 2(a), which shows a valley cross-section of the Llobregat River (NE Spain) with three paleoflood sites located at different elevations above the channel and therefore requiring different discharge thresholds to be reached before deposition can occur (Thorndycraft et al., 2005). The most utilized geomorphic flood evidence in paleoflood studies, as shown in Figure 2(a) and 2(b), are slackwater flood deposits. These are fine-grained sediments (silts and sands) deposited in marginal zones of bedrock gorges where flow separation creates eddies and zones of slackwater flow, promoting deposition of flood sediments and their preservation during subsequent events (Benito et al., 2003). Slackwater flood deposits are usually found associated with two distinct sediment–landform assemblages: (a) the infill of valley side alcoves; and (b) flood benches. Alcoves and small caves located along bedrock gorges may become inundated during floods and fine-grained slackwater flood sediments are deposited (Figure 2(a) and 2(b)). These sites usually afford excellent preservation, as the

sediments are protected from slope processes in between successive flood events and may be sealed by cave sediments, which help identify between flood events in the stratigraphic record. Flood benches are terrace-like morphological units which may be formed in areas of canyon expansion or on the inside of gorge meander bends. In appearance they are very similar to fluvial terraces, with a low relief upper surface and a clear scarp slope toward the river; however, the processes of their formation are distinctive. While terraces are abandoned floodplain surfaces, formed after river incision, flood benches result from the aggradation of slackwater flood sediments from successive flood events (Benito et al., 2003), the scarp slope at the edge of the bench formed by flood erosion approximately demarcating the transition from the main flood flow path from the zones of flow separation and slackwater.

For a detailed overview of the sedimentology of slackwater flood deposits, readers are directed to Benito et al. (2003). The key element of the stratigraphy of slackwater sequences, whether preserved in alcoves or flood benches, is the identification of individual flood events. The most common criteria for distinguishing flood events within a stratigraphic sequence are: (a) a distinct clay layer formed during the waning stage of a flood; (b) a layer of nonflood sediments that marks a clear boundary between two successive flood units, for example, a buried soil, or precipitated carbonates in caves; and (c) bioturbation indicative of an exposed sedimentary surface. The chronology of the paleoflood events are determined by standard geochronological techniques, usually AMS radiocarbon dating of organics transported during the flood, or the OSL dating of quartz sands within the deposits.

Webb and Jarrett (2002) outline a number of methods for the quantification of paleodischarge, though the usual approach is through step-backwater hydraulic flood modeling, for example, using the one-dimensional HEC-RAS (Hydrologic Engineering Center's River Analysis System) model developed by the US Army Corps of Engineers for testing the design of structures built on floodplains, for example. In paleoflood hydrology, an iterative approach is used whereby a range of flood discharges are routed through the surveyed study reach with the resulting modeled floodwater elevations matched to the mapped paleostage indicators identified along the reach. While consideration needs to be made to assess the potential model errors, for example, the determination of roughness coefficients, a quantified discharge estimate can be assigned for each geomorphic flood indicator along a reach. For most paleostage indicators such as slackwater flood sediments, this modeled discharge is a minimum discharge, the true floodwater stage was an unknown elevation above the deposited sediments; however, upper bound (nonexceedance) evidence may also be used to provide an upper limit of paleodischarge, for example, a terrace surface with no geomorphic evidence for inundation over a determined time interval (Levish, 2002).

Combining the methodological approaches described above provides data that can be interrogated to ascertain a paleoflood magnitude and frequency record. One example of this is the paleoflood record of the Llobregat River in NE Spain (Thorndycraft et al., 2005, 2006), presented in Figure 2. Figure 2(a) illustrates the paleoflood stratigraphy of three alcove sites located at different elevations above the channel bed, with the associated geochronological and paleodischarge

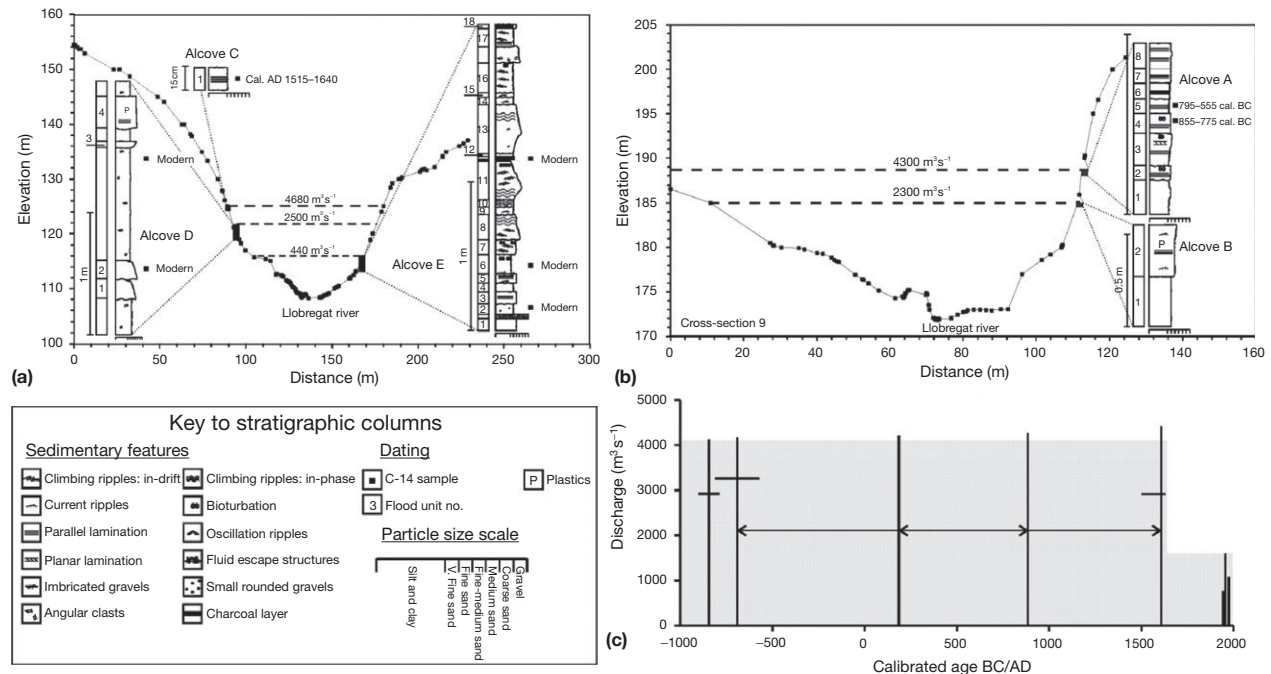


Figure 2 The paleoflood hydrology of the Llobregat River (NE Spain). Paleoflood stratigraphy, chronology and threshold discharges for (a) Monistrol (3370 km²) and (b) Pont de Vilomara (1845 km²). Reproduced from Thorndycraft VR, Benito G, Rico M, Sánchez-Moya Y, Sopena A, and Casas A (2005) A long-term flood discharge record derived from slackwater flood deposits of the Llobregat River, NE Spain. *Journal of Hydrology* 313: 16–31. (c) Extreme floods of the Llobregat River at Pont de Vilomara according to the gauge station record and paleoflood information. The paleoflood events are plotted according to the estimated minimum peak discharge and the midpoint of the radiocarbon error range, with the exception of the event dated to cal AD 1515–1640 which has been assigned to the 1617 event according to documentary evidence. The arrows indicate two undated events.

information annotated. The figure therefore shows flood magnitude frequency relationships for the Llobregat River, with low-magnitude high-frequency events preserved in the lower Alcove E; the largest floods of the instrumental record preserved at Alcove D; and an extreme event identified at Alcove C, the AMS dating of which provided an age of 1580–1740 cal. AD. Eight floods of similar magnitude at the upstream Pont de Vilomara site are illustrated for Alcove A (Figure 2(b)). The only dateable material for this sequence was found for the middle two flood units, but nevertheless this provides sufficient information to illustrate that at least five extreme magnitude events have occurred in the last 3000 years (Figure 2(c)). In addition to the two events that appear to date to the cold phase of regional climate centered on 2850 cal. BP, evident in the Bond curve, it has been assumed for Figure 2(c) that the most recent flood from this sequence is the same event at Alcove C at Monistrol (Figure 2(a)). This flood has been assigned, after cross-checking with the documentary flood record, to the known flood of AD 1617, believed to be the largest event of the last millennium. The combination of paleoflood with archival environmental history is of particular importance, as the 1617 flood can be seen to be one extreme event within a period of high magnitude and frequency flooding that occurred during particularly cold and dry decades of the Little Ice Age (LIA). From anecdotal reports on the timing of rainfall intensities and wind directions, coupled with a nested meteorological approach where events with increasingly less data and information were compared to modern analogs, synoptic conditions associated with the 1617 flood could be hypothesized

(Thorndycraft et al., 2006). The key factors that led to the flood were heavy rainfall centered over the upper basin (4–5 November 1617) followed by convective storms arriving in the littoral mountain range (lower basin) on the 6th November, coinciding with the flood peak arriving from the upstream. It appears that the event was associated with high pressure blocking over Central Europe and low pressure over the Iberian Peninsula drawing air up from North Africa and across the Mediterranean. Such synoptic conditions would have been more prevalent during cold phases of the LIA than at present, hence an increased probability of floods of such magnitude occurring in the LIA. This case study illustrates the importance of atmospheric circulation rather than moisture content of the atmosphere, as also demonstrated in the Upper Mississippi (Knox, 2000), for the generation of extreme floods at the basin scale.

Paleoflood hydrology, therefore, demonstrates that the magnitude and frequency of extreme floods can vary in response to climatic variability, and that in many river basins, gauge station flow records are of insufficient length to have recorded the largest flood flows. The next section considers whether these variations in flood magnitude and frequency can be seen to drive fluvial system response or whether it is LULC and/or autogenic processes that are the dominant control.

Drivers of Change

Much of the literature on fluvial paleohydrology attempts to define the roles of key allogenic drivers on fluvial system

response, in particular climatic variability and LULC changes, which links to the goal of hydrologists and river managers to predict future forcing of climate and LULC on flood hydrology. As we have seen, the human impact on alluviation in the European loess belt is compelling (Notebaert and Verstraeten, 2010), as is the evidence for extreme flood response to late Holocene climatic variability (Thorndycraft et al., 2005). However, the role of autogenic processes on the preserved stratigraphic record also need to be considered. While autogenic controls have long been recognized in the sedimentary record, recent methodological advances (Brown et al., 2009) now allow their improved quantification in Holocene alluvial records (e.g., Daniels, 2008; Erkens et al., 2009). Daniels (2008) suggested that autogenic bed elevation changes that are time-transgressive and propagate at predictable rates through a river basin, can be deciphered from allogenic forcing recorded in alluvial stratigraphic units correlated to a specific time interval and occurring over a larger spatial area. In the Upper Rhine Graben, Erkens et al. (2009) identified autogenic control of Rhine terrace formation during the Mid-Holocene. Due to the low valley slope in the graben reach of the Rhine, terrace incision is believed to have been initiated in response to meander cut-offs, causing a localized increase in gradient and therefore incision in response to higher stream power, rather than by any change in hydrology or sediment load.

It is interesting to note the contrast of opinions the fluvial paleohydrological community has reached with regards fluvial drivers. While Macklin et al. (2010, p. 1573) state that the “application of meta-analysis [of radiocarbon databases] ... is allowing for the first time the ‘signal’ of climate change and agricultural activity in the fluvial record to be quantified and distinguished from the ‘noise’ of autogenic river processes ...,” fluvial landscape evolution models have demonstrated in relation to long-term bedload transport that “identical system inputs ... can lead to drastically different outputs from each event ...” which means that “we may not be able to link cause to effect” (Van de Wiel and Coulthard, 2010, p. 90). This conclusion is reached through the identification of self-organized criticality in fluvial systems (Van de Wiel and Coulthard, 2010). Van de Wiel and Coulthard’s modeling research focuses on bedload transport but many alluvial archives are based on fine-grained overbank sediments, where over centennial and millennial timescales in Europe the impact of anthropogenically increased sediment load on river systems has been demonstrated (Notebaert and Verstraeten, 2010).

The current state of research into distinguishing fluvial drivers is likely to fall between the two opinions expressed above. The future research agenda will be focused on quantifying the relative impacts of autogenic and allogenic drivers through combined alluvial stratigraphic and modeling approaches, the latter including landscape, hydrological, and flood hydraulic models. Furthermore, an essential prerequisite for deciphering past (centennial scale) drivers of fluvial system change are sound geochronologies and while it is beyond the scope of this article to go into detail about the AMS C-14 and OSL dating of alluvial sediments, it is important to mention the recent application of Bayesian age models in fluvial paleohydrology (Chiverrell et al., 2009; Thorndycraft et al., 2012). For example, Chiverrell et al. (2009) used a Bayesian approach to interrogate the chronology of terrace evolution in

the Ribble Valley (NW England), identifying contrasting chronological scenarios of river evolution, while also highlighting diachronous terrace formation along the valley, once again highlighting the difficulties of linking alluvial records to specific causal events.

Summary

To summarize, this article has covered some of the recent findings in sediment budget and paleoflood studies that quantify the impact of changing drivers on fluvial systems. Recent methodological advances, including the combination of LiDAR topographic surveys, geomorphic field data and modeling, have enabled greater spatial coverage and refined temporal resolutions enabling basin scale sediment budgets to be quantified (Brown et al., 2009) even for large rivers such as the Rhine (Hoffmann et al., 2009). The quantification of basin scale sediment budgets enables inferences to be made regarding the main drivers of change, including the role of autogenic processes. Paleoflood research in bedrock geomorphic settings highlights that the magnitude of extreme floods is responsive to past climatic variability, an interpretation not always readily available in alluvial settings. The future challenge for fluvial paleohydrology is to quantify allogenic and autogenic drivers and fluvial response, including flow hydraulics, over decadal to centennial changes at the reach scale so that the discipline can link to the aims of river managers (Sear and Arnell, 2006) and have a greater impact on river management strategies. To achieve this, future research will need to further integrate field stratigraphy, sedimentology, and geomorphology with landscape, hydrological, and hydraulic flood modeling.

See also: **Fluvial Environments:** Deltaic Environments; Responses to Rapid Environmental Change; Sediments; Terrace Sequences. **Luminescence Dating:** Optical Dating. **Paleoclimate Reconstruction:** The Last Millennium. **Radiocarbon Dating:** AMS Radiocarbon Dating.

References

- Baker VR, Webb RH, and House PK (2002) The scientific and societal value of paleoflood hydrology. In: House PK, Webb RH, Baker VR, and Levish DR (eds.) *Ancient Floods, Modern Hazards: Principles and Applications of Paleoflood Hydrology*. *Water Science and Application Series*, vol. 5, pp. 1–19. Washington, DC: American Geophysical Union.
- Benito G, Rico M, Sánchez-Moya Y, Sopena A, Thorndycraft VR, and Barriendos M (2010) Late Holocene climatic variability and land-use change impacts on flood hydrology (Guadalentin River, SE Spain). *Global and Planetary Change* 70: 53–63.
- Benito G, Sánchez-Moya Y, and Sopena A (2003) Sedimentology of high-stage flood deposits of the Tagus River, Central Spain. *Sedimentary Geology* 157: 107–132.
- Brown AG (1997) *Alluvial Geomorphology: Floodplain Archaeology and Environmental Change*. *Cambridge Manuals in Archaeology*. Cambridge: Cambridge University Press.
- Brown AG, Carey C, Erkens G, et al. (2009) From sedimentary records to sediment budgets: Multiple approaches to catchment sediment flux. *Geomorphology* 108: 35–47.
- Carling P, Villanueva I, Herget J, Wright N, Borodavko P, and Morvan H (2010) Unsteady 1D and 2D hydraulic models with ice dam break for Quaternary megaflood, Altai Mountains, southern Siberia. *Global and Planetary Change* 70: 24–34.

- Chiverrell RC, Foster GC, Thomas GSP, Marshall P, and Hamilton D (2009) Robust chronologies for landform development in fluvial environments. *Earth Surface Processes and Landforms* 34: 319–328.
- Daniels JM (2008) Distinguishing allogenic from autogenic causes of bed elevation change in late Quaternary alluvial stratigraphic records. *Geomorphology* 101: 159–171.
- de Moor JJW and Verstraeten G (2008) Alluvial and colluvial sediment storage in the Geul River catchment (The Netherlands) – Combining field and modelling data to construct a Late Holocene sediment budget. *Geomorphology* 95: 487–503.
- Erkens G, Dambeck R, Volleberg KP, et al. (2009) Fluvial terrace formation in the northern Upper Rhine Graben during the last 20 000 years as a result of allogenic controls and autogenic evolution. *Geomorphology* 103: 476–495.
- Gradzinski R, Barya J, Doktor M, et al. (2003) Vegetation-controlled modern anastomosing system of the upper Narew River (NE Poland) and its sediments. *Sedimentary Geology* 157: 253–276.
- Hoffmann T, Erkens G, Gerlach R, Klostermann J, and Lang A (2009) Trends and controls of Holocene floodplain sedimentation in the Rhine catchment. *Catena* 77: 96–106.
- Hudson PF, Middelkoop H, and Stouthamer E (2008) Flood management along the Lower Mississippi and Rhine Rivers (the Netherlands) and the continuum of geomorphic adjustment. *Geomorphology* 101: 209–236.
- Knox JC (2000) Sensitivity of modern and Holocene floods to climate change. *Quaternary Science Reviews* 19: 439–457.
- Knox JC (2006) Floodplain sedimentation in the Upper Mississippi Valley: Natural versus human accelerated. *Geomorphology* 79: 286–310.
- Leopold LB and Miller JP (1954) A post-glacial chronology for some alluvial valleys in Wyoming. Geological Survey Water-Supply Paper 1261.
- Levish D (2002) Paleohydrologic bounds – Non-exceedance information for flood-frequency analysis. In: House PK, Webb RH, Baker VR, and Levish DR (eds.) *Ancient Floods, Modern Hazards: Principles and Applications of Paleoflood Hydrology. Water Science and Application Series*, vol. 5, pp. 175–190. Washington, DC: American Geophysical Union.
- Macklin MG, Jones AF, and Lewin J (2010) River response to rapid Holocene environmental change: Evidence and explanation in British catchments. *Quaternary Science Reviews* 29: 1555–1576.
- Notebaert B and Verstraeten G (2010) Sensitivity of West and Central European river systems to environmental changes during the Holocene: A review. *Earth-Science Reviews* 103: 163–182.
- Notebaert B, Verstraeten G, Rommens T, Vanmontfort B, Govers G, and Poesen J (2009) Establishing a Holocene sediment budget for the river Dijle. *Catena* 77: 150–163.
- Notebaert B, Verstraeten G, Ward P, Renssen H, and Van Rompaey A (2011) Modelling the sensitivity of sediment and water runoff dynamics to Holocene climate and land use changes at the catchment scale. *Geomorphology* 126: 18–31.
- Schumm SA (1977) *The Fluvial System*. New York: Wiley.
- Sear DA and Arnell NW (2006) The application of palaeohydrology in river management. *Catena* 66: 169–183.
- Thorndycraft VR, Barriendos M, Benito G, Rico M, and Casas A (2006) The catastrophic floods of A.D. 1617 in Catalonia (NE Spain) and their climatic context. *Hydrological Sciences Journal* 51: 899–912.
- Thorndycraft VR, Benito G, Rico M, Sánchez-Moya Y, Sopena A, and Casas A (2005) A long-term flood discharge record derived from slackwater flood deposits of the Llobregat River, NE Spain. *Journal of Hydrology* 313: 16–31.
- Thorndycraft VR, Benito G, Sánchez-Moya Y, and Sopena A (2012) A Bayesian age modelling approach for investigating the response of flood magnitude and frequency to Holocene global change. *The Holocene* 22: 13–22.
- Van De Wiel MJ and Coulthard TJ (2010) Self-organized criticality in river basins: Challenging sedimentary records of environmental change. *Geology* 38: 87–90.
- Verstraeten G, Rommens T, Peeters I, Poesen J, Govers G, and Lang A (2009) A temporarily changing Holocene sediment budget for a loess-covered catchment (central Belgium). *Geomorphology* 108: 24–34.
- Walter RC and Merritts DJ (2008) Natural streams and the legacy of water-powered mills. *Science* 319: 299–304.
- Webb RH and Jarrett RD (2002) One-dimensional estimation techniques for discharges of paleofloods and historical floods. In: House PK, Webb RH, Baker VR, and Levish DR (eds.) *Ancient Floods, Modern Hazards: Principles and Applications of Paleoflood Hydrology. Water Science and Application Series*, vol. 5, pp. 111–125. Washington, DC: American Geophysical Union.

PALEOLIMNOLOGY

Contents

Overview of Paleolimnology

Cladocera

Freshwater Mollusks

Lake Chemistry

Physical Properties of Lake Sediments

Contributions of Paleolimnological Research to Biogeography

Pigment Studies

Multiproxy Approaches

Visible and Infrared Spectroscopical Applications

Overview of Paleolimnology

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Introduction

Paleolimnology is the field of science that uses the sediments from inland waters to reconstruct past environmental conditions. Inland waters are variable and range from small shallow ponds to large lakes, from standing waters of marshlands to faster flowing waters from large rivers, and from low conductivity freshwaters to athalassic (saline) lakes. Derived from the Greek *paleo* meaning ancient and *limno* meaning lake, paleolimnological studies provide the means by which it is possible to track how an environment has changed over time, even if no prior measurements exist for longer time scales, that is, hundreds to thousands of years ago. Recognition of the field has developed especially over the course of the past three decades, largely as a result of its explosive development as an applied science. Concern about the degradation of Earth's environmental conditions, largely as a result of the impact of human activities, has sparked an increased need for environmental assessments and the ability to effectively manage anthropogenic activity in a manner that is least detrimental to the environment. Studies focused on the natural development of environments address issues such as global climate change and other related topics, by providing an understanding of the natural cycle and the baseline conditions from which it is possible to discern which effects were caused by human activity and which were natural. Paleolimnological studies can help distinguish between the two. Underlying this research is the requirement to establish the extent of the impact being studied and the rate of environmental change. However, such studies are often hindered from the beginning, because in order to assess the state of current conditions it is necessary to have a perspective on how conditions have changed through time. In

the majority of examples, regardless of the time period being considered, data on background conditions do not exist. The instrumental record is short and in most instances might only exist on a decadal or centennial scale. For example, temperature records may exist for over 100 years in some cases, but these are rare. In the Canadian Arctic, ongoing temperature records only began in the late 1940s. The same applies in the case of acid precipitation. Many lakes have acidified within the last six decades, but no preimpact pH readings exist and in most cases instrumented environmental readings only exist for shorter periods, on the order of half a century or less. Paleolimnology provides the means by which past environmental conditions can be determined as paleolimnological techniques lengthen the instrumental record by using various proxy indicators of the environmental variable being investigated. With these longer records, baseline environmental conditions can be quantified and the natural variability of the system assessed. By reconstructing paleoenvironmental conditions, it is possible to address many topics concerning past and present-day conditions. From these it will be possible to better understand and predict future trajectories of environmental conditions on Earth. At a more immediate level, it is possible to use these data to derive best practices in the management of an environment so as to maintain its ecosystem services. Policy development concerning environmental management is one of the more tangible and rewarding aspects of paleolimnological research.

Although paleolimnology has advanced gradually as a discipline, it is now a rapidly progressing science, reflecting the development of novel tools and applications. The roots of paleolimnology can be traced to DeGeer's study in the late nineteenth century of seasonal sediment deposition in Swedish

lakes (Wright, 2010). Early paleolimnological studies, dating back to about 1920, were largely descriptive (Frey, 1988); however, tremendous advances have been made over the ensuing decades and have transformed the science to one that is quantitative, applied, and still expanding (Birks, 1998; Smol, 2008). This is largely due to the coincidental timing of a serious environmental problem, namely acid precipitation, and the development of personal computers (PCs;) (Battarbee, 1991). Acid precipitation was affecting lakes and paleolimnology was the science responsible for addressing this problem. The increased computing power that PCs provided allowed for the development of computationally complex multivariate analyses that could be applied to other environmental challenges, such as climate change and eutrophication. While early studies mostly focused on one primary proxy indicator, the importance of multiproxy studies has been recognized and it is now typical for multi-investigator, multiproxy international research studies to be undertaken. Paleolimnological developments have been rapid and significant over the last three decades.

The first international paleolimnology symposium was held in 1967. Following its success and subsequent meetings, the International Paleolimnology Association was formalized in 2006 (proposed in 2003). An early compendium of related articles (Haworth and Lund, 1984) provided one of the first collections of paleolimnological papers. Several scientific journals and series (e.g., *Journal of Paleolimnology* and *Developments in Paleoenvironmental Research*) are dedicated to syntheses of paleolimnological methods (e.g., Francus, 2004; Last and Smol, 2001a,b; Smol et al., 2001a,b) and regional data (Battarbee et al., 2004; Pienitz et al., 2004). In-depth textbooks have focused on the scientific applications of paleolimnology and are useful to students at the senior undergraduate and graduate levels (Berglund, 1986; Cohen, 2003; Smol, 2008).

Lake as Archives

Lakes are archives of past environmental conditions, as their sediments represent an accumulation of biological, chemical, and physical components originating from within and without the lake (Figure 1). Materials making up the sedimentary matrix can be divided into one of two groupings: constituent material originating from within the lake itself is termed autochthonous, whereas material originating outside the lake, such as in its watershed (catchment) or air shed, is referred to as allochthonous. Sediment components can be further subdivided into biological, chemical, and physical groupings. These are described in greater detail below and as separate articles in this encyclopedia (see Table 1). Over time, as sediments buildup, a time–depth relationship is established with older sediments underlying younger sediments, as per the Law of Superposition. Through the use of a sediment corer, it is possible to recover a vertical column of sediment that represents past periods of deposition. By obtaining deeper sediments it is possible to sample farther back in time. Figure 1 illustrates a hypothetical example of how sediments from a lake formed in a glaciated terrain might be able to capture past environments. In this example, the lake formed following deglaciation of the region and the sediments captured the glaciolacustrine nature of the early lake through the

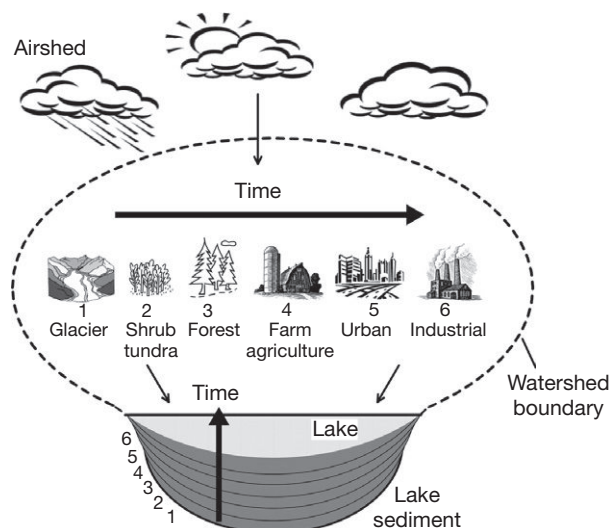


Figure 1 Idealized diagram of a typical progression of events within a postglacial lake and watershed, captured within lake sediment stratigraphies. Deglaciation is first followed by a natural succession of vegetative types, leaving pollen and other remnants behind in the sediment. Agricultural activities introduce modified and/or exotic vegetation and enhance sediment and nutrient runoff from within the watershed. Finally, urbanization and industrialization typically introduce unique, unnatural materials into watersheds, including many forms of organic contaminants, inorganic acids from atmospheric fallout, combustion residuals (soot and metals), and so on. A record of each of these watershed events may be found in the sediment sequence of lake sediments in their order of progression.

various steps of landscape evolution (tundra, forest, and human-induced changes of agriculture and urbanization). From the time of their inception, lakes start accumulating sediments and hence can serve as an archive spanning over several different time scales. For long-lived lakes whose sediments were not affected (or removed) by recent glaciations, for example, Lake Baikal (Russia), Lake Biwa (Japan), and African Rift Valley lakes, sediments have accumulated for millions of years. In areas that were glaciated during the last Ice Age, for example, large areas of North America and Europe, sediments usually represent only the Holocene or, approximately, the past 10 000 ka of accumulation. However, exciting new discoveries have been reported, concerning lakes whose sediments survived the ‘bulldozing’ effects or removal of previous glacial deposits in North America and Russia, including a 250 000-year record from Lake El’gygytyn in NE Russia (Brigham-Grette et al., 2007) and a 200 000-year record from Baffin Island, Canadian Arctic (Axford et al., 2009). These long and continuous records are shedding new light upon the most recent preglacial history of these continents, in light of the effects of global warming which are magnified at higher latitudes.

The Paleolimnological Method

Paleolimnological studies require a series of sampling and analyses, with each requiring varying degrees of expertise

Table 1 Examples of biological, physical, and geochemical proxy indicators used in paleolimnological analyses

Category	Example use or application	Key source references
Biological		
<i>Morphological remains</i>		
Algae		Smol et al. (2001a)
Diatoms	Aquatic chemistry, microhabitats, biodiversity	
Chrysophytes, scales, cysts	"	
Desmids	"	
Insects – arthropods		Smol et al. (2001b)
Coleoptera	Aquatic chemistry and microhabitats	
Chironomids	"	
Trichoptera	"	
Orbitid mites	"	
Crustaceans (Zooplankton)	"	
Ostracods	"	
Cladocera	"	
Copepods		
Mollusks		Smol et al. (2001b)
Gastropods, bivalves	Aquatic chemistry, microhabitats, fluvial conditions	
Testate amoebae	Moisture content, acidity	Smol et al. (2001a,b)
	Aquatic chemistry, hydrology, vegetation	
Sponges	Terrestrial vegetation, aquatic macrophytes	
Pollen		Smol et al. (2001a)
Phytoliths	Terrestrial catchment vegetation, especially grasses	Smol et al. (2001a)
Plant macrofossils		
<i>Biogeochemical remains</i>		Castañeda and Schouten (2011) (biomarkers)
Pigments	Vegetation – terrestrial and aquatic	
Biomarkers	Aquatic chemistry and identification of algal classes present	
DNA	Confirmation of taxon presence	
Alkenes, alkenones	Temperatures	
	Biomarkers identifying:	
	– life forms of, for example, bacteria, archaea, eukarya;	
	– process identification:	

(Continued)

Table 1 (Continued)

Category	Example use or application	Key source references
	e.g., biomass burning, photosynthesis	
Physical		Last and Smol (2001b)
Grain size	Sediment source and processes of sedimentation, turbulence	
Loss-on-ignition (LOI)	Composition of sediment matrix as organic, inorganic carbon fractions	
Mineralogy and elemental composition	Sediment source, water chemistry	
Magnetic properties	Sedimentation, erosion, dating	
Fluid inclusions	Aquatic paleochemistry and paleoclimate	
Fly ash and charcoal	Industrialization: burning of fossil fuels, fires	Smol et al. (2001a), Winkler (1985), Rose (2001)
Visible and infrared spectroscopy (VIS)	Sediment composition/chemistry, mineral phases	
<i>Geochemical</i>		
Organic matter – bulk measurements		
C:N, total organic carbon (TOC), $\delta^{13}\text{C}$, $\delta^{15}\text{N}$	Lake levels, salinity, hydrology	Leng et al. (2006)
Geochemical ratios Sr/Ca, Mg/Ca	Paleoclimate, paleoproductivity	
Biogenic silica		Smol et al. (2001a) (Biogenic silica)
Stable isotopes $^2\text{H}/^1\text{H}$, $^{15}\text{N}/^{14}\text{N}$, $^{13}\text{C}/^{12}\text{C}$, $^{18}\text{O}/^{16}\text{O}$	Contaminant transportation	
POPs (persistent organic pollutants)		

(Figure 2). Termed the paleolimnological method, it includes several steps, starting with the retrieval of a sediment core. Once retrieved, a core is sectioned, either on site or back in a laboratory, into subsections which are partitioned for various analyses including chronological dating, and geochemical, physical, and biological or proxy indicator analyses. From the generated data, multivariate analyses are used to reconstruct the environmental condition being investigated. More often than not, a completed paleolimnological analysis will enlist the use of a variety of laboratories, each specializing in one or more of the steps.

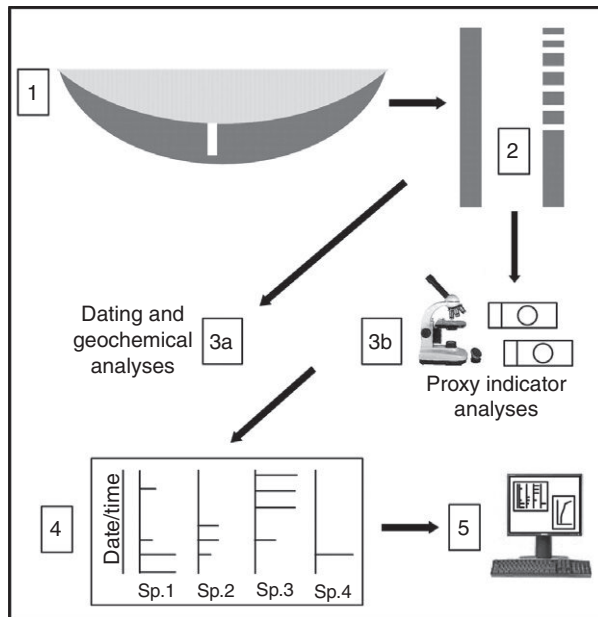


Figure 2 The paleolimnological method. Major steps involved in conducting a paleoenvironmental assessment are as follows: (1) A sediment core is retrieved from a lake and subsectioned into discrete intervals (typically 0.25–1.0 cm) (2) for dating and proxy analyses. The thickness of the sampling interval determines the temporal resolution of the final results. Thinner subsections (0.25–0.5 cm) represent shorter time intervals but also require more time and expense to process the added number of samples. It is common for researchers to sample cores finely near the surface and more coarsely downcore in order to capture recent events with the best possible resolution while also acquiring adequate information about longer-term background or natural conditions. Subsections are then further divided and typically shipped to separate laboratories for analysis of dating parameters, such as ^{210}Pb or ^{137}Cs ; (3a) geochemical analyses, such as mineral composition, total carbon, and so on and (3b) for analyses of specific proxy indicators such as diatoms, pollen, and chemical contaminants. (4) The results from the dating analyses and modeling are then combined with those from the proxy indicators to create a detailed history of each proxy indicator examined within the lake sediment. (5) Results from all of the independent proxy indicators are then combined and further analyzed using multivariate statistics to yield a comprehensive reconstruction of the history of events for the lake and its watershed.

Coring

The first step in the paleolimnological method involves retrieval of the sediment core. Particular attention must be given to this step, as it will be the basis from which the remainder of the study is established. The investigator must ensure that an undisturbed core, representative of the depositional environment, is recovered. Sediment stratigraphy must be intact, otherwise incorrect inferences will be made and erroneous conclusions drawn.

There are a variety of different coring devices available, each one suited to a particular type of lake and/or sediment type. [Glew et al. \(2001\)](#) describe the practice and various techniques needed to obtain an undisturbed sediment core from lakes of various depths. Similar to lakes, but facing its own challenges, peat coring methods have been described by [de Vleeschouwer et al. \(2010\)](#). Over a dozen different kinds of corers exist, many

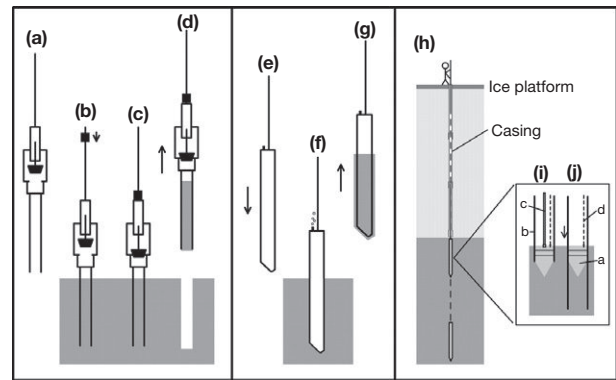


Figure 3 Examples of three commonly used corers: (a–d) gravity corer: (a) A gravity corer with an open breach is lowered down through the lake water slowly into the lake sediment (b). Once seated in the sediment at the proper depth, a weighted messenger is dropped down the coring line in order to trigger a plunger that seals the top of the core (b) and (c). Sealing the top of the gravity core creates a suction that holds the lake sediment within the sampling tube until the core is retrieved to the lake surface (d). (e–g) Freeze corer: A frozen finger (freeze core) is filled with dry ice and ethanol and lowered into the sediment to an appropriate depth (e–f). The core is left in the sediment until a coating of material freezes firmly to the outside of the casing (f). The core is then retrieved to the surface with a representative stratigraphy frozen to the outside casing. (h–j) A piston corer uses suction in the same manner as a gravity corer to capture sediment within a section of core tube. Typically deployed from a stable ice platform (h), a rigid set of tubes is lowered close to the sediment surface (i). The core tube (b) is then forced past the flange tip (a) of the corer by means of a push rod (c) into the sediment (j) while the tip remains locked in place with a control cable (d). Once the core tube has been pushed to its furthest extent into the sediment, the assembly is retrieved to the surface for subsampling. A rigid casing is periodically used to guide piston cores into place through the water column (h) and piston cores may be deployed and retrieved sequentially in order to create one continuous record for the entire sediment stratigraphy.

of which are common, albeit on a smaller scale, to the corers used in oceanographic studies. Each one is designed to retrieve an undisturbed sediment core from lakes of various types. [Figure 3](#) illustrates three commonly used sediment corers in lakes: gravity, freeze, and piston corers. Each one is best suited to a particular kind of lake and sediment. Sediment corers are usually deployed from either a boat or floating platform, or in cases where a lake may freeze over seasonally, from an ice platform ([Figure 4\(a\)](#) and [4\(b\)](#)).

Gravity corers, as the name suggests, use the force of gravity to penetrate the uppermost sediments. The corer, often composed of a sturdy metal such as brass or stainless steel, houses a plunger that can seal the top of a clear plastic or polymer tube. The corer is slowly lowered through the water column, so as to prevent the formation of a bow wave that would disturb the sediment–water interface. The mass of the corer should be sufficient to help force the tube into the sediments. Care must be taken not to overshoot and completely bury the corer and core tube in the sediments, as the sediment–water interface represents the present-day conditions and is an important reference point for the core. Once the person deploying the corer is certain that the corer has settled into the sediment, a messenger (i.e., a small weight) is deployed down the core line to trigger the plunger. The plunger seals the top of the core tube

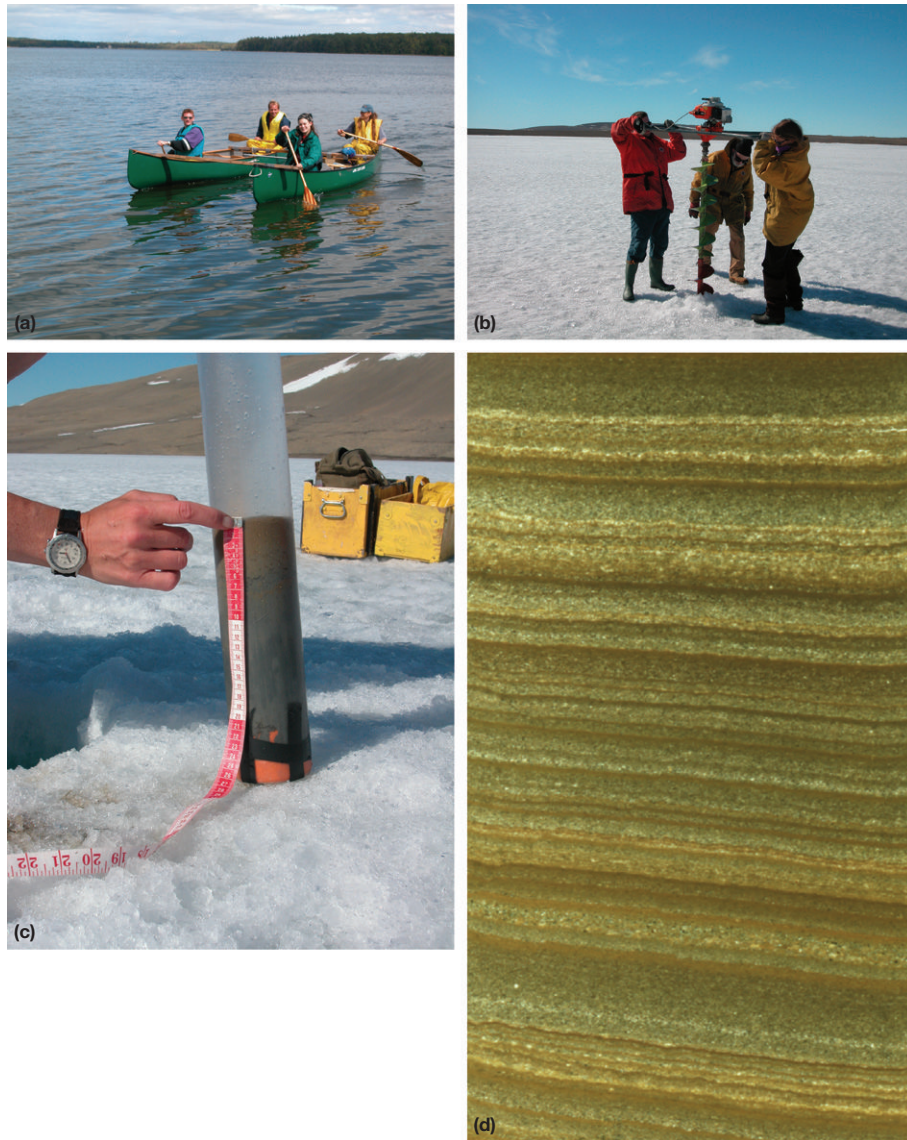


Figure 4 (a) Typical coring platforms can include two canoes lashed together or (b) the frozen ice surface of the lake. (c) The sediments recovered by gravity coring exhibit color changes reflecting different sediment matrices or (d) fine laminations as recovered from a varved lake. Picture photo credits: (a)–(c), M. Douglas and (d) S. Lamoureux.

so that the corer and sediment core can be pulled up to the surface of the lake. Before breaking the surface–water interface, a bung (often made of rubber or closed cell foam) must be inserted at the base of the core tube to plug it and prevent any sediments from falling out. If deployed correctly, these sediment cores will reliably collect an intact sediment–water interface and the underlying sediments, up to 1 m or so in length, depending upon the nature of the sediment matrix and the length of the core tube. Gravity cores are especially well suited for studies that focus on the recent past (the past few centuries or less). So long as the sediments are not too compacted, for example, clays or sand, cores can be successfully obtained.

Frigid finger or frozen coring is specialized and more complicated than gravity coring. In this case, sediments are brought to the surface after having been frozen *in situ*. The corer consists of a box or tube that has been sealed at one end. This end is

either wedge-shaped or conical and has been weighted, often with lead. The corer is then filled with dry ice and ethanol and lowered gently into the sediments where it remains for roughly 5–7 min, depending on the ethanol–dry ice charge, while the sediments freeze to the outside of the corer. The carbon dioxide gas, generated from the dry ice, escapes via a small vent at the top of the corer. Upon retrieval at the surface, the corer and its rind of frozen sediments (usually about 1 cm thick) are separated and the frozen sediment is transported and stored in a freezer prior to further analyses. This kind of coring does not recover large amounts of sediments, but it is especially useful in lakes where the sediments are very loosely consolidated yet finely structured. For example, the sediment of meromictic lakes is often varved (annually laminated). Freezing these sediments *in situ*, as described, preserves these laminations. Again, depending upon the nature of the sediment and length of the

coring apparatus, it is possible to retrieve sediment cores up to 2 m or more in length. Generally, however, shorter cores are obtained. The specialized nature of this coring technique restricts its use to a certain extent. Because of transport regulations regarding dry ice (effectively, solid carbon dioxide), it is difficult, although not impossible, to use this technique in very remote areas.

Piston coring is used when obtaining longer sedimentary sequences. The coring apparatus consists of a tube, which is connected to a handle at the top by a series of rods. Inside the tube is a piston, connected to the top of the device by a strong wire cable that can be positioned at either end of the tube. With the piston positioned at the leading end of the core tube, the corer is pushed into the sediments using the rods, until it is directly above the sediment sequence to be recovered. The piston is then locked into place by securing its connecting cable in place with a clamp, and the core tube is then pushed 1 m down past the piston. This results in a core tube filled with sediment that is brought back to the surface of the lake by pulling up the rods. Another core tube is attached, and the process is repeated for the next depth interval. Generally, a researcher will work downward through the lake sediments, sequentially recovering 1 m of core at a time until an impenetrable layer is met. In order to ensure that a continuous sediment sequence is obtained, and no depths remain unsampled, the researcher will overlap the core intervals so that there are a few centimeters of overlap between the bottom of one core section and the top of the next lower interval. Because of problems with rigidity and strength of the rods, casing is often used to guide and constrain the rods and the attached corer back to the same core hole. In addition, there is a limit to the length of the core sequence that can be obtained. A combined length of water depth and rod length of no more than 20 m is generally the maximum. Piston cores have been used to obtain entire postglacial sequences (up to 11 m or more) in some lakes.

The three methods described can be used on relatively small lakes and require little logistical support. The equipment can even be transported by a coring team to remote areas without mechanical assistance. However, core retrieval from the larger lakes of the world, such as lakes Biwa, Baikal, Titicaca, Great Salt Lake, and even the Laurentian Great Lakes, require much more logistical planning and equipment. Thus, for a long period, these lakes remained uncored. Recent design advances and a concerted push by the paleolimnological community have resulted in long, multimillion-year cores being obtained. In some instances, drilling crews similar to those used on oil rigs are employed to recover the cores (Dean, 2010). Drilling platforms can be deployed from specialized drill ships, such as those used by oceanographers, or from ingenious platforms designed for easy transport and deployment on these large, yet often remote, sites. The challenges and problems associated with these designs and large projects are described by Leroy and Colman (2001). The efforts however, are rewarding as long, millennial-scale core records can be obtained. This is of especial interest to long-term global change studies.

Subsectioning

Upon retrieval of a core, it is extruded and subsectioned into intervals so that the sediments can be partitioned for various

subsequent analyses. This can be done on site, immediately upon retrieval of the core, or it can be completed back in the more controlled conditions of a laboratory. The decision is based upon the nature of the sediments, the kind of core, as well as the location of the coring site. Frozen cores are transported intact, back to a freezer and laboratory, until further analyses. Gravity cores can be subsectioned on site, especially if the sediment matrix is loose and unconsolidated and/or if during transport back to the lab there is a likelihood of the sediments getting mixed. Piston cores are usually returned back to the lab in 1 m sections, where they are split lengthwise. One half is preserved for the archives and the other working half is subsectioned, as necessary.

Dating Analyses – Geochronology

Once a sediment core has been retrieved and the researcher is assured that the core represents an unmixed stratigraphic column of sediments, the age–depth relationship can be used in the reconstructions. Establishment of a reliable, accurate chronology is key for all temporal interpretations. According to the Law of Superposition, younger sediments will overlie older, deeper sediments. This is the basis of relative age dating. In most instances, an absolute chronology is desirable so that specific dates can be linked to past events. Paleolimnologists have a number of dating techniques at their disposal to help resolve depth with respect to time. Each is appropriate for a certain age range and sediment type. These are covered in detail in the dating article and articles in Last and Smol (2001a). For events that have occurred within the past 150 years, ^{210}Pb and ^{137}Cs radioactive isotopes are commonly used. Another radioactive isotope, ^{14}C , has been used to date lake sediments up to roughly 50 000 years of age. Accelerator mass spectroscopy dating extends this age limit and new developments are refining dating chronologies. Other dating techniques include electron spin resonance, paleomagnetism, luminescence dating, varve chronology, amino acid racemization, tephrochronology, and marker beds of pollen, such as *Ambrosia*, that can reflect specific events such as European settlement of eastern North America. In a similar approach, Rose and Appleby (2005) demonstrated how a rise in spheroidal carbonaceous fly-ash particles could be dated to 1850 ± 25 years, for lakes within the United Kingdom.

Proxy Indicator Analyses

Proxy indicators are isolated from the sediment matrix and analyzed so as to infer past environmental conditions. These can be subdivided into three groups, namely, biological, physical, and geochemical. A brief review of these indicators is provided below. Table 1 lists some of the more commonly used proxies and references specific dedicated entries within this publication (*Encyclopedia of Quaternary Science*) which provide a more detailed account of each proxy. While early paleolimnological studies were based solely upon one main proxy indicator, multiproxy studies employing several biological, geochemical, and/or physical proxies are now common practice, as they yield more robust and holistic reconstructions.

Biological indicators

In order to be used as a paleoindicator, also known as a proxy indicator, a biological proxy must possess several important

characteristics. It needs to reproduce quickly, be abundant, easily and consistently identified, readily preserved in sediments, and ecologically constrained to specific environmental conditions. Ideally, the proxy contains either durable 'hard' parts or resilient 'soft' components that do not easily decompose and are readily identified through microscopy or geochemical analyses. Numerous biological indicators are used in paleolimnological analyses to infer past conditions. The remnants of these living organisms are preserved within lake sediments as morphologically identifiable 'hard' parts, that are usually derived from the organism's body, and are easily enumerated using microscopy (such as light microscopy – 5–1000 $\times$ magnification – or scanning electron microscopy for higher magnifications at the sub- and micron level). Many of these cell walls or hard parts are composed of chitin (e.g., insects), biogenic silica (e.g., diatoms and chrysophytes), calcium carbonate (e.g., mollusks), cellulose (e.g., desmids, plant macrofossils), or other biopolymers such as sporopollenin (e.g., pollen grains). Examples of these proxies include algae (diatoms, chrysophytes, and desmids), plant macrofossils, zooplankton, mollusks, and insects. Others are preserved as biochemical remnants derived from their 'soft,' organic parts such as pigments and biomolecular signatures derived, for example, from alkenones and DNA fragments preserved in sediment. These so-called biomarkers are compounds that can be traced to a source organism or process (Castañeda and Schouten, 2011) and require biogeochemical analyses for their identification. Biomarkers have been established for numerous organisms, including green algae (botryococenes), cyanobacteria (heterocysts and glycolipids), bacteria (bacteriopheopigments, BHPs), and for processes such as photosynthesis and biomass burning (Castañeda and Schouten, 2011).

The use of biomarkers is increasing as new techniques for isolating various biochemical components are being developed. Not only is the taxonomic catalog of biomarker signatures increasing, but a new understanding of organisms' distribution is expanding. D'Andrea and Huang (2005) reported the discovery of long-chain alkenones (LCAs) in a sediment core and surface sediments of Greenland lakes. This discovery confirmed the presence of prymnesiophyte (haptophyte algae), that had not been previously reported using conventional light microscopy methods. A similar survey of prairie lake surface sediments from Canada established the presence of LCAs and confirmed that haptophytes typically inhabit sulfate-rich, saline lakes (Toney et al., 2011).

The identification of these specific biomarkers is helpful in cases where the hard, morphological remains of a proxy have not been preserved. For example, in situations where taphonomic processes such as dissolution removes diatom frustules, measurements of the diatom biomarker, loliolide, can provide the missing diatom abundance data, although not at a species level (Castañeda and Schouten, 2011). Identification at the species level can be attained through analyses of fossil DNA fragments, and although much further work remains in cataloging the DNA signature of living organisms and distinguishing this from the fossil signature (Castañeda and Schouten, 2011), some advances have been made. For example, Theroux et al. (2010) used DNA fingerprinting to resolve the issue of species identification in their study on the diversity of lacustrine haptophytes (prymnesiophytes). The presence of alkenones confirmed the presence of haptophytes, but

genetic analyses were needed for identification to species level. This has important implications for alkenone temperature calibrations, which are useful for paleoclimatic reconstructions, yet are best applied to know species assemblages.

Many of these proxies and their derivatives, such as algae, zooplankton, mollusks, and sponges, originate from within the lake (autochthonous) and can be used to reconstruct water chemistry and microhabitat availability throughout the core sequence. Others are transported into the lake from outside the lake (allochthonous) via wind and water currents, and stored within the sediment archives. Allochthonous proxies, including pollen, phytoliths, and insects, can be used to track vegetation shifts and infer past climate variables, including temperature and precipitation, as well as ecosystem dynamics.

The value of an indicator lies in the paleolimnologist's ability to quantify the organisms' environmental optima and tolerances. In so describing each taxon's environmental niche, or space, it is therefore possible to link its occurrence in a stratigraphic profile to specific environmental conditions. Until recent advances in appropriate multivariate statistics, precise quantification of these environmental optima and tolerances was rare and difficult. This calibration of indicators to environmental conditions or variables is achieved by using surface-sediment training sets and is described briefly here and in detail by Birks (1995), Birks et al. (2012), and Smol (2008).

Physical and geochemical indicators

Examination of the physical and chemical sedimentary matrix has revealed a large number of sources for determining paleoenvironmental conditions, including paleoclimates and other past environmental conditions. Table 1 lists some of the more commonly used indicators and some of their applications. These include bulk density, grain size, mineralogy, and magnetic susceptibility.

Progress in analyzing stable isotopes preserved in lake sediments has been rapid over the last decade and is described in the recent literature (Ito, 2001; Leng et al., 2006). Using different sediment fractions, that is, biogenic, precipitate, and so on, various stable isotope ratios can be measured to reconstruct past water chemistry variables (Table 1). These can be related to paleoclimatic conditions. Oxygen, carbon, nitrogen, and hydrogen isotopes from lacustrine sediments contribute to paleoclimate reconstructions; however, a detailed knowledge of processes controlling their abundance is necessary, as even small differences in temperature can cause, for example, a shift in their equilibrium (Leng et al., 2006). Oxygen isotopes are measured from inorganic and organic samples and are used in paleoclimatic reconstructions. Other information on productivity and changes in nutrient supply are obtained from carbon and nitrogen isotopes measured from organic materials in lakes. Isotopic studies are well suited for multiproxy studies as interpretation based solely on isotopes is unconstrained and hence limited.

Data Analyses and Quantification

One of the greatest advances in paleolimnology has been in data analyses leading to the quantification of environmental change and paleoenvironmental assessment. This big step brought paleolimnology forward from a descriptive to a

quantitative science. It then became possible to test hypotheses and quantify environmental responses to ecological perturbations. As indicated by Birks (1998, 2008), Birks et al. (2012), and Smol (2008), the creation of modern surface-sediment calibration or training sets conferred strength and rigor to paleolimnological studies that had previously been lacking. By linking biotic assemblages with their associated environmental data, it is possible to quantify species–environmental relationships. The modern training set is constructed by undertaking a field-intensive regional survey of approximately 50 lakes. From each site, the surface sediments (top 1 cm) are collected and the physical and chemical water chemistry is measured. Two data matrices are then developed from these sample collections. The first is composed of the abundance of the biological proxy indicators in question for each site, for example, diatoms (can include several hundred species). The second matrix consists of the physical and chemical characteristics for each site. Often relative abundance measurements are used. Using multivariate analyses, for example, such as multivariate regression and canonical correspondence analyses, it is possible to identify which environmental variables are contributing the most to the observed species distribution. Once each species' environmental optima and tolerances are calculated, it is possible to construct transfer functions that relate species' abundance throughout a stratigraphic sequence to quantitative values of the environmental variable being reconstructed. A simplified description of the process is provided by Smol (2008).

As longer and finer-resolution stratigraphic records are obtained, it is possible to conduct time-series and rate-of-change analyses on stratigraphic paleolimnological data (Birks, 1998; Birks et al., 2012). Innovative developments in numerical and statistical techniques are providing answers to complex questions. This is especially true of the capacity to handle the large amounts of data generated by multiproxy studies.

Paleoenvironmental Assessment: Paleolimnological Applications

Paleolimnology has developed into a science with a broad spectrum of applications. These range from describing long-term trends in climate change to helping solve environmental crises caused directly or indirectly by anthropogenic activities. Paleolimnological analyses have been conducted on every continent and at nearly every possible latitude on Earth, where lakes or paleolake sediments exist. Paleolimnological analyses have also been proposed by exobiologists for potential work on other planets, including Mars (Doran et al., 2004; Lim and Cockell, 2002). This section provides a brief overview of some of the diverse studies that use paleolimnological applications and makes reference to additional articles in the encyclopedia.

Some of the most important applied paleolimnological work has focused on the human-induced problems of acid precipitation and eutrophication. In both instances, success has largely been due to the usefulness of diatoms in tracking these changes. Diatom species–environment responses are strong along pH and nutrient gradients and hence the sensitivity of this proxy to changing environmental conditions is remarkable. Reviews of these environmental problems are detailed by Smol (2008).

It is possible to track the source of nutrients and pollutants to a lake via various vectors, using paleolimnological techniques. Surprisingly, even very remote lakes are experiencing the consequences of elevated nutrient and pollutant levels. These vectors include animals (zoological vectors), atmospheric currents, and runoff. Some salmon species' life cycles make use of nursery lakes for spawning purposes. Because adult salmon spend most of their time in the ocean, their flesh is enriched in ^{15}N and this can be used as a convenient paleolimnological signal that identifies the source of nutrient additions to freshwaters, of a marine origin. Decomposing adult carcasses add nutrients elevated in ^{15}N to these lakes. Using a combination of ^{15}N isotopes and diatoms, it is possible to reconstruct the long-term population dynamics of these fish populations over several millennia (Finney et al., 2002). In a parallel study of decomposing whale carcasses, elevated levels of ^{15}N and diatom-inferred nutrients were used to track past human occupation (Thule Inuit) of a High Arctic wintering camp for four centuries (AD 1200–1600; Douglas et al., 2004). In the Antarctic, Hodson and Johnston (1997) examined sedimentary seal hairs to track changes in seal populations. In a subsequent paleolimnological reconstruction, fluctuations in historical seal populations could be linked to human harvesting activities such as whaling and sealing. This information is useful for the development of conservation policies on how to deal with recent population increase in fur seals that threaten nearby freshwater aquatic ecosystems (Hodson et al., 1998).

Muir et al. (2009) measured the spatial and temporal distribution of atmospheric deposition of mercury in surface sediments across eastern North America and found a significant trend of deposition of mercury into the isolated Arctic. Wolfe et al. (2001) documented the ecological changes in alpine lakes, in the Rocky Mountains of Colorado, induced by anthropogenic atmospheric nitrogen deposition. The high elevation lakes' algal assemblages were sensitive to the added input of nitrogen and this left a sedimentary signal. Blais et al. (2005) tracked elevated levels of nutrients and pollutants such as persistent organic pollutants (POPs) that were being transported from the ocean environment back to Arctic lakes and ponds via seabirds that were feeding in the adjacent marine system. The POPs had been transported via air currents to high latitudes. In a complementary study, Michelutti et al. (2010) demonstrated how trophic position influenced the levels of metal pollution. Two seabird species that feed from different trophic levels (planktonic fish and invertebrates vs. benthic mollusks) transported different metals derived from their marine food sources back to lakes, thereby providing a means by which particular metals could be used to reconstruct past seabird populations. Zale (1994) used the elevated levels of phosphorus and copper in Antarctic lake sediments to determine past penguin rookery populations. These birds were bioconcentrating nutrients and metals from their marine diet of krill, and some of their guano was leaching into a nearby lake. Sun et al. (2000) conducted a similar study and inferred penguin population fluctuations in response to shifting climatic conditions over the past 3000 years.

Paleolimnological studies have been used to study long-term environmental climate change. Kashiwaya et al. (2001) reconstructed major climatological and limnological changes for the past 3.5 million years of Lake Baikal, Russia. They related

physical properties of the core to the effects of Milankovitch orbital parameters. Using diatoms as the main proxy indicator, [Laird et al. \(2003\)](#) reconstructed the drought intensity and frequency for the past two millennia of the Northern Great Plains of North America by tracking diatom-inferred lake water salinity. Reported at subdecadal resolution, these data identified droughts that lasted for centuries and were related to the same atmospheric circulation anomalies causing modern and potential future drought. The historical perspective provided by these paleolimnological studies has demonstrated the susceptibility of western and central Canada to the effects of warming climates: during drier periods in the past, lake basins closed, resulting in increased salinity and eutrophication ([Schindler, 2009](#)).

[Bigler and Hall \(2003\)](#) were able to directly link diatom assemblages as quantitative indicators of temperatures at a century-scale in northern Sweden. New advances in oxygen isotope measurements are providing the potential for direct temperature inferences from diatoms. [Crespin et al. \(2010\)](#) reported that the temperature-dependent relationship between the $\delta^{18}\text{O}$ value of diatom frustules and lake water is a promising tool for reconstructing paleotemperatures. Recent advances on the uses of biomarkers are contributing toward the reconstruction of past temperatures as well. [Pearson et al. \(2011\)](#) report on the use of glycerol dialkyl glycerol tetraethers (GDGTs) derived from Archaea and bacteria in lake sediments. A calibration of GDGTs from the surface sediments of 90 lakes distributed across a transect from the Scandinavian Arctic to the Antarctic yielded a root mean-squared error of 2.0°C and demonstrated the potential of this chemical proxy as a useful tool for paleothermometry.

Lake sediment analyses contribute important data to large projects such as those defined by large international collaborative efforts. The IGBP-PAGES PEP III (Pole–Equator–Pole) is one such project. Focused on the paleoclimate change of the last 200 000 years in a transect that crossed Africa and Europe, paleolimnological studies were the most common analyses used ([Oldfield and Thompson, 2004](#)) and included studies of sea-level change on the Scandinavian coast ([Hedenstrom and Risberg, 1999](#)), paleolakes of the Sahara ([Gasse et al., 1987](#)), as well as many others. The net result of these efforts was a better understanding of climate variation throughout this long transect, as well as the interactions between humans and climate over this same time period through Europe, the Middle East, and Africa. The complexities of the interactions between the Eurasian ice sheets, the North Atlantic circulation, and tropical monsoons, for example, were illustrated, and although additional research is required, the long time-series recoverable from many of these geographic areas means that answers will be forthcoming.

Paleolimnological studies from across the continents indicate a consistent climate record during the Holocene: moist conditions (high lake levels) during the early Holocene and a subsequent drying trend in the later Holocene. However, regional differences exist and can be refined by local studies. An overall picture of long-term changes can be extracted from these studies to tease out the orbital timescales that underlie the shorter term climatic variability ([Birks, 2008](#)). Compilations of regional climate syntheses have been completed for many regions of the globe. Paleolimnological

reconstructions, including those from ancient or seasonally present lakes, contribute important information to the overall picture of past climates. [Wünnemann et al. \(2007\)](#) examined climate instability for the past 45 000 years as recorded in Chinese desert lakes. [Vimeux et al. \(2009\)](#) performed an extensive investigation of past climates of South America. [Hodgson and Smol \(2008\)](#) and [Pienitz et al. \(2004\)](#) compiled a synthesis of documented environmental change at high latitudes. As predicted by general circulation models, these high latitudes are already exhibiting the effects of a warming climate. In one example, the effects of this recent warming were documented across a circum-arctic latitudinal gradient, based on a diatom and zooplankton study of lake sediments ([Smol et al., 2005](#)). Lakes at higher latitudes exhibited a markedly greater amount of warming than those at lower latitudes. [Fritz \(2008\)](#) discusses the challenges of reconstructing paleoclimates from the complex paleolimnological data derived from regional studies. Given the unique environmental conditions of any lake, a nonlinear response to climatic forcing can occur, resulting in spatial heterogeneity in the timing of inferred change. In order to assess climate at longer temporal scales, including multidecadal, it is important to synthesize large data sets from which patterns of local climate response can be separated from broad, regional scale patterns.

Paleolimnology is playing an important role in bettering our understanding of natural hazards such as tsunamis. Often resulting from earthquake activity, tsunamis generate run-up waves that can inundate coastal areas and hence leave a signal in coastal water bodies. [Hutchinson et al. \(2000\)](#) identified layers of sand in the sediments of a lake situated at the head of a fiord on western Vancouver Island, Canada. These three tsunami events were followed by a fall in relative sea level, confirmed by analyses of protozoan and diatom microfossils. In the Shetland Islands, [Bondevik et al. \(2005\)](#) identified three sand layers in sediments of coastal fens and lakes dating between 8000 and 1500 years ago. The oldest layer could be related to a tsunami generated by a well-studied landslide on the Norwegian continental slope which would have generated a wave of over 20 m. [Pisaric et al. \(2011\)](#) documented the impact and limited recovery of coastal lakes from a storm surge event in 1999 in the Mackenzie Delta, Arctic Canada, using diatom-based paleolimnological techniques. This long-term perspective showed that the extent of the storm surge was unmatched over the past 1000 years and was likely a signal of a higher probability of future surges, due to the impacts of a warming climate and rising sea levels. Knowledge of the frequency of past tsunamis and smaller storm surges is important to improve our understanding of their impacts and the subsequent recovery of coastal environments.

Environmental management practices and policies are increasingly relying on paleolimnological data to provide the reference baseline conditions that need to be met in order to return an ecosystem back to its natural (preimpact) conditions. [Salgado et al. \(2009\)](#) conducted paleolimnological analyses of aquatic macrophytes (macroremains) to determine the predisturbance communities in Loch Leven, Scotland. They established the reference state characterized by soft-water macrophytes and noted that in order to return to preimpact conditions, nutrient reduction to the lake would have to be effected.

Lakes provide a range of ecosystem services to those communities living within their catchments and, in certain instances, where multiple stressors have negatively affected these services, such as food security (provisioning), trade-offs are necessary in order to ensure environmental and/or economic stability. Tropical Lake Victoria in Africa provides such an example of a lake which has experienced rapidly expanding human population growth and changes to agricultural practices. Changes in land-use practices are contributing high nutrient loading into the lake, and this eutrophication is affecting the aquatic ecosystem, threatening the survival of an endemic fisheries population that supports the large human population living within its catchment. [Verschuren et al. \(2002\)](#) conducted a multiproxy paleolimnological study to reconstruct past limnological conditions and confirm the cause of fisheries decline. By examining sedimentary pigments, diatoms, and chironomid remains, they identified the periods of eutrophication and subsequent anoxia that developed in conjunction with human population growth. Deep-water anoxia affected the habitat of certain fish species, resulting in a fisheries collapse and a need for an improved management of land-use practices. In a similar study from Loch Leven, Scotland, [May and Spears \(2012\)](#) used paleolimnological reconstructions to provide environmental managers with key background conditions from which to develop effective policies for conflicting ecosystem service activities including pollution and flood control, fish stocking, and recreation and nature conservation.

Paleolimnological studies are not restricted to sediments from lakes or paleolakes. Large river systems have been successfully cored and yielded stratigraphically intact sediment cores whose data could be applied to tracking environmental changes and subsequent management decisions. [Reavie et al. \(1998\)](#) inferred past shoreline characteristics from sediment cores obtained from fluvial lakes within the large St. Lawrence River, Canada. The diatom transfer functions identified regions whereby water quality had improved, following management decisions to restrict nutrient runoff. These abatement measures decreased macrophyte density nutrient levels. In a second example, [Reavie et al. \(2005\)](#) used diatom and geochemical data (metals) to track environmental degradation in the St. George River draining Lake Superior, Canada. A sediment core obtained from this fluvial lake faithfully recorded the effects of anthropogenic activities such as logging and dam-building, as well as the effects of the steel-, leather-, and paper-manufacturing industries.

Recent advances in drilling hard-to-reach lakes, such as the subglacial lakes in Antarctica, hold great promise of yielding important information on the genesis, longevity, and biological flora and fauna of these isolated ecosystems. Recent efforts to drill through several kilometers of ice to reach the waters of Lake Vostok (up to 4 million years old) and Lake Ellsworth (up to 400 000 years old) provide the opportunity to retrieve sediment cores that will add to our understanding of the history of the East and West Antarctic Ice Sheets, respectively ([Gramling, 2012](#)).

New Developments/Conclusions

Paleolimnology is continuing to develop at a rapid pace. It is clear that the footprint of humans on the environment is not a

delicate one and that environmental managers will continue to rely heavily on this branch of science. Paleolimnology has now established a central role in determining the water framework for many nations ([Dalton et al., 2009](#)). As new environmental proxy indicators are described and calibrated, and new analytical techniques are developed, our understanding of complex environmental interactions will increase. The description and development of new proxy indicators, especially biomarkers, are refining the types of questions that can be resolved and a more holistic reconstruction of past environments will be possible.

New developments in acquiring quantitative data nondestructively are enhancing paleolimnology's ability to infer past environments. Information from digital images of cores can be applied to a number of different research questions ([Francus, 2004](#)). Image analysis is rapid, relatively inexpensive to obtain, and can be applied to a wide range of data. Because digital photography can yield high-resolution images, methods for counting and identifying fossils, such as pollen and calcareous microfossils, are being refined. These methods can greatly assist in studies in which numerous samples are processed. New numerical and statistical methods are providing powerful, complementary tools that will aid in the interpretation of multiproxy studies. These are key to deciphering changes at interannual and decadal scales and modeling biological dynamics.

As paleolimnological studies deliver higher temporal-resolution interpretations, a better understanding of neolimnological processes is becoming possible. [Saros \(2009\)](#) describes how the interface of neo- and paleolimnology is yielding a better understanding of mechanistic links of species–environment relationships. These long-term, high-resolution data are critical in the management of ecosystems and to our understanding of ecosystem resilience to current and future environmental change, such as global warming ([Willis et al., 2010](#)).

See also: [Diatom Introduction](#); [Oribatid Mites](#); [Phytoliths](#); [Plant Macrofossil Introduction](#); [Varved Lake Sediments](#). **Beetle Records:** [Overview](#). **Carbonate Stable Isotopes:** [Lake Sediments](#). **Chironomid Records:** [Chironomid Overview](#). **Paleolimnology:** [Cladocera](#); [Contributions of Paleolimnological Research to Biogeography](#); [Freshwater Mollusks](#); [Lake Chemistry](#); [Multiproxy Approaches](#); [Physical Properties of Lake Sediments](#); [Pigment Studies](#); [Visible and Infrared Spectroscopical Applications](#). **Plant Macrofossil Methods and Studies:** [Treeline Studies](#).

References

- Antoniades D, Michelutti N, Quinlan R, et al. (2011) Cultural eutrophication, anoxia, and ecosystem recovery in Meretta Lake, High Arctic Canada. *Limnology and Oceanography* 56: 639–650.
- Axford Y, Briner JP, Cooke CA, et al. (2009) Recent changes in a remote Arctic lake are unique within the past 200,000 years. *Proceedings of the National Academy of Sciences* 106: 18443–18446.
- Battarbee RW (1991) Recent paleolimnology and diatom-based environmental reconstruction. In: Shane LCK and Cushing EJ (eds.) *Quaternary Landscapes*, pp. 129–174. Minneapolis: University of Minnesota Press.
- Battarbee RW, Gasse F, and Stickley CE (2004) *Past Climate Variability through Europe and Africa*, vol. 6. Dordrecht: Kluwer Academic Publishers.

- Berglund B (1986) *Handbook of Holocene Palaeoecology and Palaeohydrology*. Chichester: John Wiley and Sons.
- Bigler C and Hall RI (2003) Diatoms as quantitative indicators of July temperature: A validation attempt at century-scale with meteorological data from northern Sweden. *Paleogeography Paleoclimatology and Paleoecology* 189: 147–160.
- Birks HJB (1995) Quantitative paleoenvironmental reconstructions. In: Maddy D and Brew JS (eds.) *Statistical Modelling of Quaternary Science Data. Technical Guide 5*, pp. 161–254. Cambridge: Quaternary Research Association.
- Birks HJB (1998) Numerical tools in paleolimnology – Progress, potentialities and problems. *Journal of Paleolimnology* 20: 307–322.
- Birks HJB (2008) Holocene climate research – Progress, paradigms, and problems. In: Battarbee RW and Binney HA (eds.) *Natural Climate Variability and Global Warming: A Holocene Perspective*. Chichester: Wiley-Blackwell.
- Birks HJB, Lotter AF, Juggins S, and Smol JP (2012) *Tracking Environmental Change Using Lake Sediments. Data Handling and Numerical Techniques*, vol. 5. Dordrecht: Kluwer Academic Publishers.
- Blais JM, Kimpe LE, McMahon D, et al. (2005) Arctic seabirds transport marine-derived contaminants. *Science* 309: 445.
- Bondevik S, Mangerud J, Dawson S, Dawson A, and Lohne O (2005) Evidence for three North sea tsunamis at the Shetland Islands between 8000 and 1500 years ago. *Quaternary Science Reviews* 24: 1757–1775.
- Brigham-Grette J, Melles M, and Minyuk P (2007) Scientific Party Overview and significance of a 250 ka paleoclimate record from El'gygytyn Crater Lake, NE Russia. *Journal of Paleolimnology* 37: 1–16.
- Castañeda IS and Schouten S (2011) A review of molecular organic proxies for examining modern and ancient lacustrine environments. *Quaternary Science Reviews* 30: 2851–2891.
- Cohen AS (2003) *Paleolimnology. The History and Evolution of Lake Systems*. Oxford: Oxford University Press.
- Crespin J, Sylvestre F, Alexandre A, Sonzogni C, Pailles C, and Perga M-E (2010) Re-examination of the temperature-dependent relationship between $\delta^{18}\text{O}_{\text{diatoms}}$ and $\delta^{18}\text{O}_{\text{lake water}}$ and implications for paleoclimate inferences. *Journal of Paleolimnology* 44: 547–557.
- D'Andrea WJ and Huang Y (2005) Long chain alkenones in Greenland lake sediments: Low $\delta^{13}\text{C}$ values and exceptional abundance. *Organic Geochemistry* 36: 1234–1241.
- Dalton C, Taylor D, and Jennings E (2009) The role of paleolimnology in implementing the water framework directive in Ireland. *Biology and Environment: Proceedings of the Royal Irish Academy* 109B: 161–174.
- de Vleeschouwer F, Chambers FM, and Swindles GT (2010) Coring and sub-sampling of peatlands for palaeoenvironmental research. *Mires and Peat* 7: 1–10.
- Dean WE (2010) Recent advances in global lake coring hold promise for global change research in paleolimnology. *Journal of Paleolimnology* 44: 741–743.
- Doran PT, Priscu JC, Lyons WB, Powell RD, Andersen DT, and Poreda RJ (2004) Paleolimnology of extreme cold terrestrial and extraterrestrial environments. In: Pienitz R, Douglas MSV, and Smol JP (eds.) *Long-Term Environmental Change in Arctic and Antarctic lakes*, pp. 475–511. Dordrecht: Springer.
- Douglas MSV, Smol JP, Savelle JM, and Blais JM (2004) Prehistoric Inuit whalers affected Arctic freshwater ecosystems. *Proceedings of the National Academy of Sciences* 101: 1613–1617.
- Finney BP, Gregory-Eaves I, Douglas MSV, and Smol JP (2002) Fisheries productivity in the northeastern Pacific Ocean over the past 2,200 years. *Nature* 416: 729–733.
- Francus P (ed.) (2004) *Image Analysis, Sediments and Paleoenvironments. Developments in Paleoenvironmental Research*, vol. 7. Dordrecht: Kluwer Academic Publishers.
- Frey DG (1988) What is paleolimnology? *Journal of Paleolimnology* 1: 5–8.
- Fritz SC (2008) Deciphering climate history from lake sediments. *Journal of Paleolimnology* 39: 5–16.
- Gasse F, Fontes JC, Plaziat JC, et al. (1987) Biological remains, geochemistry and stable isotopes for the reconstruction of environmental and hydrological changes in the Holocene lakes from North Sahara. *Paleogeography, Paleoclimatology and Paleoecology* 60: 1–46.
- Glew JR, Smol JP, and Last WM (2001) Sediment core collection and extrusion. In: Last WM and Smol JP (eds.) *Tracking Environmental Change Using Lake Sediments. Basin Analysis, Coring, and Chronological Techniques*, vol. 1, pp. 73–105. Dordrecht: Kluwer Academic Publishers.
- Gramling C (2012) A tiny window opens into Lake Vostok, while a vast continent awaits. *Science* 335: 788–789.
- Haworth EY and Lund JW (1984) *Lake Sediments and Environmental History. Studies in Palaeolimnology and Palaeoecology in Honour of Winifred Tutin*. UK: Leicester University Press.
- Hedenstrom A and Risberg J (1999) Early Holocene shore displacement in Southern Central Sweden as recorded in elevated isolated basins. *Boreas* 28: 490–504.
- Hodgson DA and Smol JP (2008) High-latitude paleolimnology. In: Vincent WF and Laybourn-Parry J (eds.) *Limnology of Arctic and Antarctic Aquatic Ecosystems*. New York: Oxford University Press.
- Hodson DA and Johnston NM (1997) Inferring seal populations from lake sediments. *Nature* 387: 30–31.
- Hodson DA, Johnston NM, Caulkett AP, and Jones VJ (1998) Palaeolimnology of Antarctic fur seal *Arctocephalus gazella* populations and implications for Antarctic management. *Biological Conservation* 83: 145–154.
- Hutchinson I, Guilbault J-P, Clague JJ, and Bobrowsky PT (2000) Tsunamis and tectonic deformation at the northern Cascadia margin: A 3000-year record from Deserated Lake, Vancouver Island, British Columbia. *The Holocene* 10: 429–439.
- Ito E (2001) Application of stable isotope techniques to inorganic and biogenic carbonates. In: Last WM and Smol JP (eds.) *Tracking Environmental Change Using Lake Sediments. Physical and Geochemical Methods*, vol. 2, pp. 351–371. Dordrecht: Kluwer Academic Publishers.
- Kashiwaya K, Sakai H, Ryugo M, Horii M, and Kawai T (2001) Long-term climatological cycles found in a 3.5 million year continental record. *Journal of Paleolimnology* 25: 271–278.
- Laird KR, Cumming BF, Wunsam S, Rusak JA, Oglesby RJ, Fritz SC, and Leavitt PR (2003) Lake sediments record large-scale shifts in moisture regimes across the northern prairies of North America during the past two millennia. *Proceedings of the National Academy of Science* 100: 2483–2488.
- Last WM and Smol JP (eds.) (2001a) *Tracking Environmental Change Using Lake Sediments. Basin Analysis, Coring, and Chronological Techniques*, vol. 1. Dordrecht: Kluwer Academic Publishers.
- Last WM and Smol JP (eds.) (2001b) *Tracking Environmental Change Using Lake Sediments. Physical and Geochemical Methods*, vol. 2. Dordrecht: Kluwer Academic Publishers.
- Leng MJ, Lamb AL, Marshall JD, et al. (2006) Isotopes in Lake Sediments. In: Leng MJ (ed.) *Isotopes in Paleoenvironmental Research*, pp. 147–184. Dordrecht: Springer.
- Leroy SA and Colman SM (2001) Coring and drilling equipment and procedures for recovery of long lacustrine sequences. In: Last WM and Smol JP (eds.) *Tracking Environmental Change Using Lake Sediments. Basin Analysis, Coring, and Chronological techniques*, vol. 1, pp. 107–135. Dordrecht: Kluwer Academic Publishers.
- Lim DSS and Cockell CS (2002) Paleolimnology in the High Arctic – Implications for the exploration of Mars. *International Journal of Astrobiology* 1: 381.
- May L and Spears BM (2012) Managing ecosystem services at Loch Leven, Scotland, UK: Actions, impacts and unintended consequences. *Hydrobiologia* 681: 117–130.
- Michelutti N, Blais JM, Mallory ML, et al. (2010) Trophic position influences the efficacy of seabirds as metal biovectors. *Proceedings of the National Academy of Sciences* 107: 10543–10548.
- Muir DCG, Wang X, Yang F, et al. (2009) Spatial trends and historical deposition of mercury in Eastern and Northern Canada inferred from lake sediment cores. *Environmental Science and Technology* 43: 4802–4809.
- Oldfield F and Thompson R (2004) Archives and proxies along the PEP III transect. In: Battarbee RW, Gasse F, and Stickley CE (eds.) *Past Climate Variability Through Europe and Asia*, pp. 7–29. Dordrecht: Springer.
- Pearson E, Juggins S, Talbot HM, et al. (2011) A lacustrine GDGT-temperature calibration from the Scandinavian Arctic to Antarctic: Renewed potential for the application of GDGT-paleothermometry in lakes. *Geochimica et Cosmochimica Acta* 75: 6225–6238.
- Pienitz R, Douglas MSV, and Smol JP (eds.) (2004) *Long-Term Environmental Change in Arctic and Antarctic Lakes*, vol. 8. Dordrecht: Springer.
- Pisarcic MFJ, Thienpont JR, Kokelj SV, et al. (2011) Impacts of a recent storm surge on an Arctic delta ecosystem examined in the context of the last millennium. *Proceedings of the National Academy of Sciences* 108: 8960–8965.
- Reavie ED, Robbins JA, Stoermer EF, et al. (2005) Paleolimnology of a fluvial lake downstream of Lake Superior and the industrialized region of Sault Saint Marie. *Canadian Journal of Fisheries and Aquatic Science* 62: 2586–2608.
- Reavie ED, Smol JP, Carignan R, and Lorrain S (1998) Diatom paleolimnology of two fluvial lakes in the St. Lawrence River: A reconstruction of environmental changes during the last century. *Journal of Phycology* 34: 446–456.
- Rose NL (2001) Fly ash particles. In: Last WM and Smol JP (eds.) *Tracking Environmental Change using Lake Sediments, Volume 2, Physical and Chemical Techniques*. Dordrecht: Kluwer Academic Publishers.
- Rose NL and Appleby PG (2005) Regional applications of lake sediment dating by spheroidal carbonaceous particle analysis I: United Kingdom. *Journal of Paleolimnology* 34: 349–361.
- Salgado J, Sayer S, Carvalho L, Davidson T, and Gunn I (2009) Assessing aquatic macrophyte community change through the integration of palaeolimnological

- and historical data at Loch Leven, Scotland. *Journal of Paleolimnology* 43: 191–204.
- Saros JE (2009) Integrating neo- and paleolimnological approaches to refine interpretations of environmental change. *Journal of Paleolimnology* 41: 243–252.
- Schindler DW (2009) Lakes as sentinels and integrators for the effects of climate change on watersheds, airsheds, and landscapes. *Limnology and Oceanography* 54: 2349–2358.
- Smol JP (2008) *Pollution of Lakes and Rivers. A Paleoenvironmental Perspective*, 2nd edn. Oxford: Blackwell Publishing.
- Smol JP, Birks HJB, and Last WM (eds.) (2001a) *Tracking Environmental Change Using Lake Sediments*. In: *Terrestrial, Algal and Siliceous Indicators*, vol. 3. Dordrecht: Kluwer Academic Publishers.
- Smol JP, Birks HJB, and Last WM (eds.) (2001b) *Tracking Environmental Change Using Lake Sediments*. In: *Zoological Indicators*, vol. 4. Dordrecht: Kluwer Academic Publishers.
- Smol JP, Wolfe AP, Birks HJB, et al. (2005) Climate-driven regime shifts in the biological communities of arctic lakes. *Proceedings of the National Academy of Science* 102: 4397–4402.
- Sun L, Xie Z, and Zhao J (2000) A 3,000-year record of penguin populations. *Nature* 407: 855.
- Theroux S, D'Andrea WJ, Toney J, Amaral-Zettler L, and Huang Y (2010) Phylogenetic diversity and evolutionary relatedness of alkenone-producing haptophyte algae in lakes: Implications for continental paleotemperature reconstructions. *Earth and Planetary Science Letters* 300: 311–320.
- Toney JL, Leavitt PR, and Huang Y (2011) Alkenones are common in prairie lakes of interior Canada. *Organic Geochemistry* 42: 707–712.
- Verschuren D, Johnson TC, Kling HJ, et al. (2002) History and timing of human impact on Lake Victoria, East Africa. *Proceedings of the Royal Society of London B* 269: 289–294.
- Vimeux F, Sylvestre F, and Khodri M (eds.) (2009) *Past Climate Variability in South America and Surrounding Regions. From the Last Glacial Maximum to the Holocene. Developments in Paleoenvironmental Research*, Vol. 14. Dordrecht: Springer Science + Business Media BV.
- Willis KJ, Bailey RM, Bhagwat SA, and Birks HJB (2010) Biodiversity baselines, thresholds and resilience: Testing predictions and assumptions using paleoecological data. *Trends in Ecology & Evolution* 25: 583–591.
- Winkler MG (1985) Charcoal analysis for paleoenvironmental interpretation: A chemical assay. *Quaternary Research* 23: 313–326.
- Wolfe AP, Baron JS, and Cornett RJ (2001) Anthropogenic nitrogen deposition induces rapid ecological changes in alpine lakes of the Colorado Front Range (USA). *Journal of Paleolimnology* 25: 1–7.
- Wolfe BB, Hall RI, Edwards TWD, et al. (2008) hydroecological responses of the Athabasca Delta, Canada, to changes in river flow and climate during the 20th century. *Ecohydrology* 1: 131–148.
- Wright HE (2010) High points in paleolimnological studies as viewed by a convert. *Journal of Paleolimnology* 44: 497–503.
- Wünnemann B, Hartmann K, Janssen M, and Zhang HZ (2007) Responses of Chinese desert lakes to climate instability during the past 45,000 years. In: Madsen DB, Fa-Hu C, and Xing G (eds.) *Late Quaternary Climate Change and Human Adaptation in Arid China. Developments in Quaternary Science* 9 Amsterdam: Elsevier.
- Zale R (1994) Changes in the size of the Hope Bay Adelie penguin rookery as inferred from Lake Boeckella sediment. *Ecography* 17: 297–304.

Cladocera

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Glossary

Abiotic: Nonliving.

Anthropogenic: Human-caused.

Benthic: Relating to the lake floor.

Biotic: Living.

Bottom-up control: Food web control related to chemical and physical variables of the lake, also effect of organisms with a lower trophic position than the one in question.

Branchiopoda: The taxonomical class that includes cladocerans.

Carapace: Thick, hard, chitinous shield that covers the body of most cladocerans.

Chydorid: A common group of littoral cladocerans.

Clone: A group of organisms of the same genetic constitution that are descended from a common ancestor by asexual reproduction.

Cyanobacteria: A group of bacteria formerly regarded as blue-green algae. Often in eutrophic conditions form toxic blooms in water bodies.

Dissolved organic carbon: Contributes to defining the color and ultraviolet radiation penetration in the water body.

Ephippium: The hard chitinous case covering cladoceran resting eggs.

Eutrophic: Nutrient- and biomass-rich water body.

Exoskeleton: The protective or supporting structure covering the body of cladocerans.

Gamogenesis: Sexual reproduction.

Head shield: The protective chitinous structure covering the head of a cladoceran.

Herbivore: Plant- or algae-eating organism.

Lacustrine: Refers to lakes and ponds.

Limnology: Study of fresh water.

Littoral: The shallow zone in a lake where plants are rooted, typically the nearshore region of a water body.

Melanin: The dark pigment present in cladoceran carapace, provides protection against ultraviolet radiation.

Molting: To shed the exoskeleton.

Neolimnology: Study of modern freshwaters, as opposed to paleolimnology.

Neuston: The minute organisms inhabiting the surface layer of water or moving on the surface film.

Oligotrophic: Nutrient and biomass-poor water body.

Parthenogenesis: Reproduction where offspring are formed from unfertilized eggs, with no need of males.

Pelagic: Open-water area of a lake.

Planktivore: An organism feeding on plankton.

Postabdomen: The posterior part of the cladoceran body.

Subfossil: A specimen that is found under fossil conditions but is relatively young on the geological time scale.

Top-down control: Food web control from organisms with a lower trophic position than the one in question.

Training set: A set of species collected from modern sites along with relevant environmental data. Form a basis for quantitative paleolimnological reconstruction when used in conjugation with subfossil assemblages.

Transfer function: A mathematical function that describes the relationship between biological species and environmental variables that allows the past values of an environmental variable to be inferred from the composition of a fossil assemblage.

Ultraviolet radiation: Biologically harmful radiation from the sun, wavelengths 280 to 400 nm.

Introduction

The taxon Cladocera (Crustacea: Branchiopoda) was first recognized by [Latreille \(1829\)](#), but it took 100 years more before the study of their subfossil remains was introduced into paleolimnology. The first quantitative study using animal microfossils in North America was published by [Deevey \(1942\)](#), and a few years later, DG Frey extensively promoted the use of cladoceran remains in aquatic research with his numerous publications in the mid-twentieth century (e.g., [Frey, 1960, 1964](#)). These first publications reported the occurrence of one or a few cladoceran taxa, and although they recognized the potential of cladoceran fossils as indicators of past environmental conditions, their main emphasis was to find out which species of cladocerans were preserved in sediments, how they could be

identified, and what was the general sedimentation process in a lake. The taxonomic guidelines that were set by these pioneer studies are still valid and for many species, still represent the best drawings of the remains.

Between the 1960s and 1980s, the use of cladocerans in paleolimnological research increased rapidly. Sediment cores were collected from lakes across Europe and North America with the purpose of resolving community changes in time spans varying from decades to thousands of years ([Goulden, 1964](#); [Hofmann, 1986](#)). Many of the studies also concentrated on species distribution patterns using only surface sediments ([Korhola, 1999](#)), as it was becoming evident that a single sediment sample revealed the cladoceran community composition in a water body better than a careful and extensive limnological sampling over a number of years ([Frey, 1960](#)).

These studies linked cladocerans to environmental variables such as temperature (climate), pH (acidification), and nutrients (eutrophication), initializing the use of cladocerans as indicators of past environmental conditions. However, many studies also concluded that the ecological indicator value of individual cladoceran species was variable and often rather low. Instead, interpretations should be based on the overall structure and composition of the community assemblage. The development of transfer functions in 1990s significantly increased the utility of cladocerans (and other organisms in the fossil record) in reconstructing the past. Transfer functions quantitatively reconstruct past environmental conditions in lakes. They are generated from a set of surface samples (0–1 cm depth) usually collected from >20 training set lakes, representing a large gradient of environmental conditions. Statistical correlations of the subfossil remains to environmental parameters that quantitatively identify the community response to changes of a certain environmental variable are obtained using multivariate analyses.

During the past two decades, subfossil cladoceran research has developed further by incorporating ecological processes in the reconstruction models. This means that in addition to collecting single measurements of physical and chemical variables for transfer function creation, researchers are including information such as planktivorous fish densities in their models (Amsinck et al., 2005a; Jeppesen et al., 1996). Further, the response of zooplankton to environmental conditions is not only measured in terms of species abundance and composition but also as changes in carapace size (Jeppesen et al., 2002), rates between parthenogenetic and gamogenetic reproduction (Sarmaja-Korjonen, 2004), and as shifts in clonal composition of a species (Hairston et al., 1999). Including such process, studies in paleolimnology provide new ways of studying past environments; however, the traditional method of identifying and counting remains still provides the basis for all paleolimnological work and has not lost its value.

From Living Animal to a Subfossil Remain

Organisms that are abundant in many kinds of aquatic habitats are most suitable for paleolimnological studies. Their high abundance also allows for the use of statistical methods, thus assuring a better understanding of the reliability of the reconstructions. However, living organisms become available for paleolimnological reconstructions only after their death or following processes that leave behind identifiable remains, such as molting.

The flora and fauna of lakes are made up of numerous groups of organisms that all contribute to the structure and functioning of the food web and the state of the ecosystem. Many of these organisms provide quick and reliable information on the water body in which they live. However, not all organisms are suitable as paleolimnological indicators; for instance, the remains of mobile adult insects that travel across a large set of water bodies cannot be used in the site-specific interpretation of environmental variables. Cladocerans are one of the few aquatic animal groups that are routinely used in reconstructing the past. The Cladocera are some of the primary

herbivores in lakes and contribute to the recycling of nutrients in the water column. In addition, they form a large portion of planktivorous fish and invertebrate diet. Because of their trophic position in the middle of the food web, they are subjected to both bottom-up and top-down forces and have a critical role in carbon and energy transfer through the food web.

The use of any group of organisms in paleolimnology depends primarily on their resistance to degradation after death and secondly, on the availability of features that can be used for identification. Cladocerans are moderately well preserved in sediment, and specimens of all 11 existing families leave recognizable remains. After death, cladocerans break into small fragments such as exoskeletons and ephippia, which are modified carapaces enclosing and protecting the eggs (Figure 1). Both exoskeletons and ephippia are composed of chitin, are chemically inert, and have features useful for species identification (Korhola and Rautio, 2001). Among taxonomic groups, there are great differences in the number of body parts that preserve and can be specifically identified. For most pelagic cladocerans with soft chitinous cover, the identification becomes difficult after death and many important features required for accurate identification are lost. For instance, the carapaces of *Daphnia* break into small, irregular pieces and sizes that in many cases can only be used as qualitative information of the presence of this genus in the lake. And ephippia, that are often the only remains used to quantify the *Daphnia* abundance, cannot be identified to the species level. This is unfortunate because most ecological information on living cladocerans comes from pelagic species, especially the genus *Daphnia*, and cannot be effectively used in paleolimnological interpretation. Conversely, for many littoral and benthic species, once the remains have been removed from the sediment, they reveal structures and pores that allow more accurate identification than that of intact whole individuals. Chydorid head shields are especially easy to identify to species level using the number and shape of pores that become visible in the remains after a sample preparation (Box 1).

Despite the large number of taxonomic keys available for contemporary Cladocera, the identification of their remains still relies largely on individual research papers that each scientist needs to assemble to make an identification guide. Important exceptions are the identification guides of cladoceran remains in North American lake sediments (Korosi and Smol, 2012a,b; Sweetman and Smol, 2006) and of Central and Northern Europe (Szeroczyńska and Sarmaja-Korjonen, 2007). The review paper of Korhola and Rautio (2001) also gives an extensive list of papers that serve as a good starting point in subfossil cladoceran identification.

The Indicator Value of the Cladocera

The high abundance of cladocerans in most of the world's freshwater habitats and the large range of different optimal tolerances to environmental conditions among species make cladocerans very powerful in the detection of changes in environmental conditions. The order Cladocera makes an important and abundant contribution to lake and slow running river food webs worldwide. Few species are found in brackish water, and the majority of the about 400 species are strictly lacustrine.

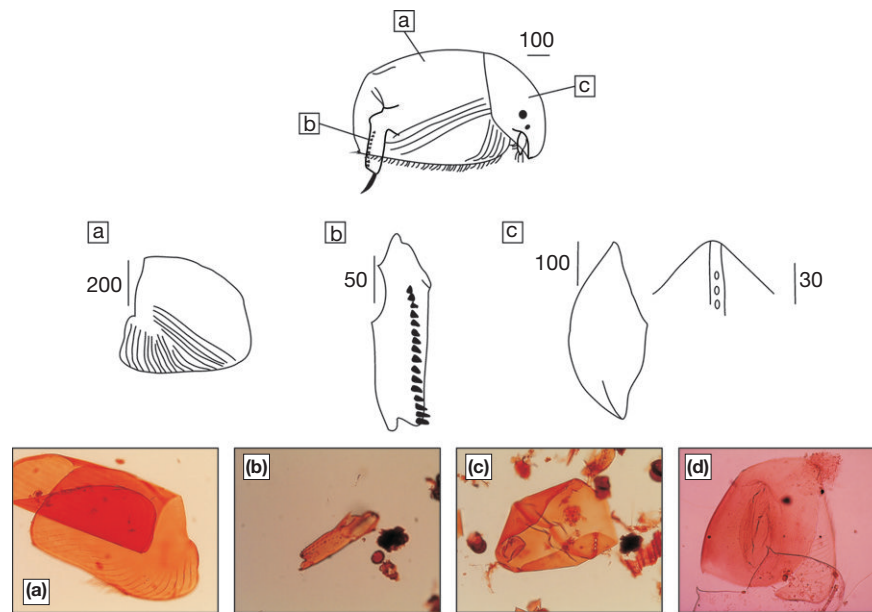


Figure 1 *Acroperus harpae* (Crustacea, Cladocera) and its body parts that are used in the identification of a subfossil animal: (a) carapace, (b) post abdomen, (c) head shield, (d) ephippium. Scale bar unit μm . Photo courtesy Cladocera workshop, Helsinki, 2000.

BOX 1 Sample preparation

Standard procedure for cleaning the sediment samples and concentrating the cladoceran remains for subfossil Cladocera analysis includes heating and stirring the sediment (usually a known volume, e.g., 1 cm^3 or 1 g) in 10% KOH for 15–30 min, washing the KOH-treated sample through a 40–50- μm mesh, centrifuging the residue, and finally staining the residue with, for example, safranin. For microscopic slide preparation, the residue is mounted on slides usually in glycerol or glycerol gelatin. Quite often, the results of the analysis are illustrated as relative proportions (%) of individual species of the total Cladocera community because proportions are irrespective of variations in sedimentation rate and water content. However, concentrations and, if the sediment core is properly dated, accumulation rates may provide additional paleoecological data on past environmental changes. To be able to evaluate concentration of the remains or individuals in the sediment samples, the residue of a known sediment volume should be brought up to a known volume (with distilled water and/or glycerol, e.g., 10 ml) after centrifuging, and a known volume of the sample (e.g., $100\text{ }\mu\text{l}$) should be placed on a slide. An alternative method to gain concentration data is to add marker grains, for example, *Lycopodium* spores within the residue, count the marker grains during the microscopic analysis, and afterward, calculate concentrations for individual taxa. A comprehensive protocol for sample preparation is presented by Szeroczyńska and Sarmaja-Korjonen (2007).

The usefulness of cladoceran remains in aquatic research is twofold. The remains in the surface sediment can be used to complement the limnological water column sampling of the modern community assemblage in lakes, and when combined with physical–chemical measurements, of environmental control of those assemblages. The down-core studies, on the other hand, use the natural lake archives to reveal the past environmental conditions in lakes.

Because cladoceran abundance varies seasonally and they occupy a variety of different habitats within a water body, their adequate representative collection with limnological sampling techniques is difficult, and many species are often overlooked. As a result, surface sediment coring from the middle of a lake

has great advantage over limnological sampling, as it integrates this seasonal and spatial heterogeneity. Cladocerans from different habitats in the littoral zone are transported offshore via currents and mixed with remains of offshore species (Frey, 1988). A surface sediment sample collected from the middle of a lake represents the entire community from all different habitats of the lake, and usually, just one sample is representative of the community structure of those species that leave sufficient recognizable remains after death. Korhola (1999) reported the cladoceran distribution pattern in subarctic Fennoscandia from single surface sediment samples per lake. In 53 lakes, he identified 29 cladoceran species, the identical number found in 17 ponds in the same geographical area as a result of extensive seasonal sampling with an average of 12 samples per site (Figure 2). Similarly, Nevalainen (2010) compared chydorid species richness and evenness in single surface sediment samples taken from sublittoral zones from nine boreal lakes with contemporary limnological sweep net samples collected weekly throughout the active period from three different habitats of the same lakes. The results showed that sedimentary assemblages clearly collect higher amounts of biodiversity with lower counting efforts than the exhaustive net sampling and analysis. Therefore, it is possible, and even more suitable, to use surface sediment assemblages in biodiversity evaluations of such aquatic organisms which preserve adequately as fossils, such as cladocerans. However, some studies have shown that there can be significant variations in species assemblages among surface sediment samples in different habitat types and depths within one lake (Kattel et al., 2007). The 33 surface sediment samples from an oligotrophic mountain loch in Scotland showed that the remains of benthic chydorid species were mostly deposited close to their original littoral habitats and that planktonic taxa accumulate to pelagic deep-water locations. Similar results are now also available from a dystrophic boreal lake from Finland (Figure 3) and further support the

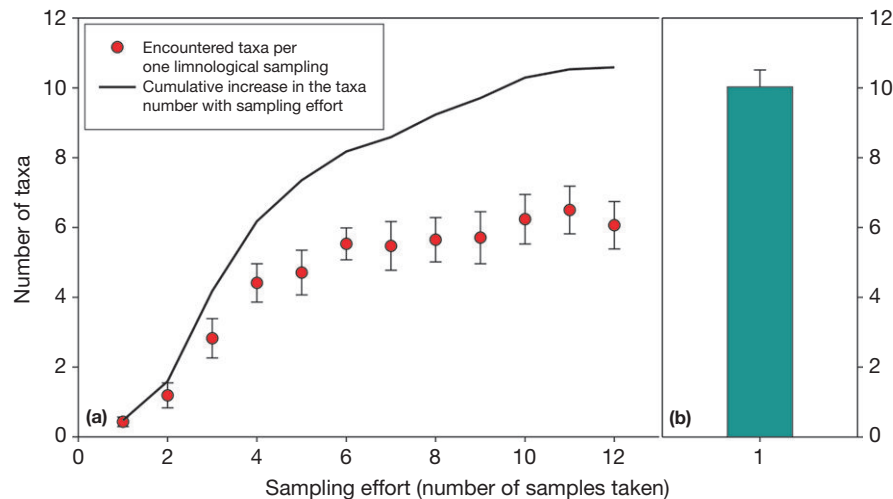


Figure 2 Comparison of (a) limnological and (b) paleolimnological sampling efficiency. Limnological data is from 17 water bodies that each were sampled 12 times from ice-out to ice-in; paleolimnological data is from 53 lakes that each were sampled once. Encountered cladoceran species number per sampling (a, b) and the cumulative increase in total species number (a) are shown. Errors bars are standard errors, cumulative line is mean for the 17 sites. Limnological data from [Rautio \(1998\)](#) and paleolimnological data from [Korhola \(1999\)](#).

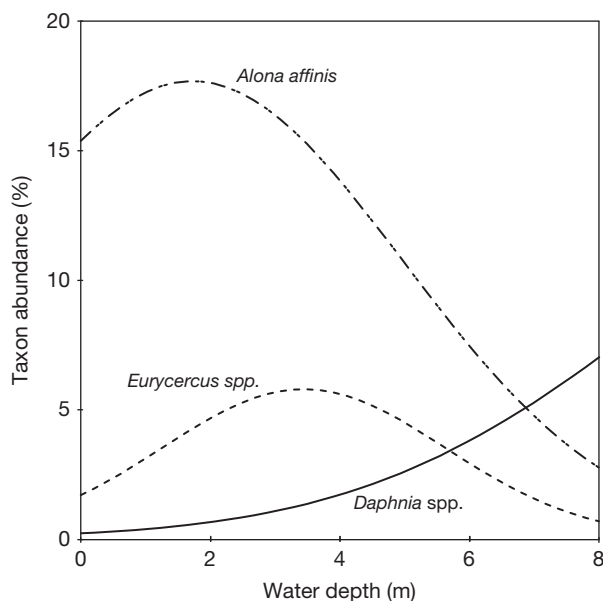


Figure 3 Taxon response curves (generalized linear models, GLMs) to water depth in 31 surface sediment samples in Lake Pieni-Kauro, Eastern Finland, showing the divergent depth preferences of *Alona affinis*, *Eurycercus*, and *Daphnia*. Modified from Nevalainen L (2011) Intra-lake heterogeneity of cladoceran (Crustacea) assemblages forced by local hydrology. *Hydrobiologia* 676: 9–22.

idea of heterogeneous and habitat-specific distribution of cladoceran taxa, at least in some basins ([Nevalainen, 2011](#)). Although such variability may interrupt the development of transfer functions and quantitative paleoenvironmental reconstructions, the surface sediment assemblages still remain a more efficient way of identifying the modern Cladocera community of a lake than limnological water column samples, with the restriction that they only capture species that leave remains in the sediment.

Down-core studies open the view into the past and show the greatest potential of the science of paleolimnology: reconstruction of the past. The Cladocera have existed more than 100 million years, and at least for the past 200 000 years, no morphological changes have been observed ([Frey, 1962](#); [Hann and Karrow, 1993](#)) which is essential when comparing fossil remains with modern samples. Similarly, an important factor is the measurable interaction between the Cladocera and lake abiotic and biotic characteristics. Many such interactions are well defined for cladocerans and contribute to the application of Cladocera in paleolimnology. Some of the best-defined interactions are between cladocerans and lake physical variables, especially lake perimeter, watershed area, lake depth, water temperature, and sediment organic content (e.g., [Korhola, 1999](#)). Physical and chemical parameters related to water turbidity, temperature, and conductivity control the entire cladoceran community in the water body. However, thermal structure and related nutrient and algal stratification tend to influence more pelagic populations, and macrophytes and habitat diversity influence more littoral populations, thereby making the indicator value of each species to environmental changes different.

In general, cladoceran species numbers are highest in lakes where a large range of environmental conditions allow a high number of niches and hence species, making these sites most optimal for the quantification of species–environment relationships and reconstruction of the past. Among water bodies, the greatest diversity of cladocerans is found in lakes with vast open-water areas and abundant littoral vegetation. Such habitat diversity greatly increases the number of cladoceran species present. In particular, the diverse members of the family Chydoridae are common in sites with variable substrata (macrophytes, mud, sand). The more homogenous pelagic environment usually supports fewer species than littoral and benthic habitats ([Table 1](#)). Species from the suborders Daphniidae and Bosminidae are the most commonly used pelagic species in paleolimnological interpretation.

Table 1 The most typical distribution of cladoceran genera among the major habitats of a lake

<i>Pelagic</i>	<i>Littoral</i>	<i>Benthic</i>				
Planktonic	Vegetation	Open water	Neuston	Rock	Sand	Mud
<i>Daphnia</i>	<i>Graptoleberis</i>	<i>Simocephalus</i>	<i>Scapholeberis</i>	<i>Alona</i>	<i>Alona</i>	<i>Ilyocryptus</i>
<i>Bosmina</i>	<i>Eurycerus</i>	<i>Bosmina</i>		<i>Alonopsis</i>	<i>Chydorus</i>	<i>Chydorus</i>
<i>Diaphanosoma</i>	<i>Alona</i>	<i>Polyphemus</i>		<i>Rhynchotalona</i>		<i>Leydigia</i>
<i>Leptodora</i>	<i>Pleuroxus</i>	<i>Holopedium</i>		<i>Chydorus</i>		
<i>Bythotrephes</i>	<i>Oxyurella</i>			<i>Monospilus</i>		
	<i>Acroperus</i>					
	<i>Camptocercus</i>					
	<i>Disparalona</i>					
	<i>Ophryoxus</i>					
	<i>Sida</i>					
	<i>Latona</i>					
	<i>Acantholeberis</i>					

Source: Korhola and Rautio (2001).

In addition to habitat diversity and number of ecological niches available, temperature is another important controlling factor of cladoceran ecology and the most important environmental variable controlling their global-scale distributions. The abundance of Cladocera is highest in the mid-latitudes at temperatures between 15 °C and 20 °C (Gillooly and Dodson, 2000). This is the temperature range for most cladocerans that maximizes parthenogenetic reproduction, that is, a female's ability to asexually produce offspring of unfertilized eggs (Box 2). A population can multiply many times in just a few days under favorable temperatures or only produce one generation per season under marginal temperatures or not reproduce at all if the temperature is too low. For instance, *Ceriodaphnia quadrangularis* requires temperatures over 8 °C to reproduce in boreal North America. As a result of direct and indirect effects of temperature, Arctic tundra lakes are inhabited by fewer than ten cladoceran species (Bennike et al., 2005; Brancelj et al., 2009). Some species, however, are adapted to cold environments. *Daphnia middendorffiana* is restricted at the southern end of its range by a maximum temperature of around 15 °C (Patalas, 1990), and *Daphnia umbra* also favors cold mountain and tundra lakes (Schwenk et al., 2004). Temperature-mediated effects on species, composition, and population abundance are detectable in the sediments.

It is recognized that the more evolved an organism, the more complex is its interaction with the biotic and abiotic variables of its habitat. Therefore, recognizing and quantifying response signals to one specific environmental variable becomes more difficult. The presence or absence of a cladoceran in a given environment depends on two very different factors: dispersal limitation and the suitability of local environmental conditions, making the estimation of population or community response to a single environmental variable challenging. This is a source of error in predictions and interpretations. Jeppesen et al. (2001) discussed some of the pitfalls when aiming to interpret cladoceran response to a single environmental variable. Their paper gives an example of the links among climate, water level, and lake productivity that should be taken into consideration when reconstructing any one of these variables. Hence, direct effects of water level change on Cladocera are rare and need to be substantial before any significant effect can be detected. Rather, the cladoceran

BOX 2 Parthenogenetic life cycle provides a tool for paleolimnology

The lifecycle of cladocerans consists of active and dormant stages as reproduction is cyclical parthenogenesis where two reproduction modes alternate: parthenogenesis (asexual) and gamogenesis (sexual). Parthenogenesis is used under favorable environmental conditions as an efficient mechanism for population growth and maintenance. Parthenogenetic eggs develop without fertilization and genetic recombination, and consequently, the young are identical clones of their mothers (Frey, 1982; Hebert and Ward, 1972). Gamogenesis, on the other hand, occurs during periods of unfavorable environmental conditions and produces diapausal resting eggs for dormancy. Gamogenesis starts with the appearance of males (from parthenogenetic eggs) and gamogenetic females (switch from parthenogenetic females), continues with copulation, and finishes in fertilization of one or two diploid resting eggs. Accordingly, it provides a mechanism for reestablishment of populations after seasonal (e.g., winter in northern regions) or irregular (e.g., human disturbances) stresses, means for passive dispersal, and production of new genotypes via meiosis and fertilization. Although there are obvious advantages in reproducing with cyclical parthenogenesis, it is a very rare lifecycle and has extremely strict requirements for its maintenance (Lynch and Gabriel, 1983). Cladoceran resting eggs within most taxa are enclosed and protected by a morphologically modified carapace, the ephippium, and the eggs store the genetic material of the species over centuries, possibly even for longer time periods (Hairston, 1996). The resting eggs tolerate extreme events of desiccation, freezing, and digestion, and they form an egg bank in the sediment. Thus, the resting eggs in this sedimentary egg bank present an analogous pool of biodiversity to the terrestrial seed bank in plants and, additionally, provide interesting material for paleogenetic examinations (Brede et al., 2009; Limburg and Weider, 2002) and resurrection ecology (Cousyn et al., 2001; Kerfoot and Weider, 2004). Techniques are evolving to extract DNA from resting eggs. In addition, markers have been developed that allow the genotype information to be used for purposes other than just to resolve taxonomic uncertainties (Hairston et al., 1999; Schwenk et al., 2004). In favorable conditions, the resting eggs can hatch, and an ancient clone of the Cladocera can be brought back to life. This fascinating technique of resurrection ecology allows both the estimation of cladocerans to adapt genetically to changing environmental conditions and the potential to study and experiment with apparently extinct clones. So far, *Daphnia* with a big resistant ephippium has served as a test genus for such paleogenetic studies.

community responds to indirect effects of water level changes, such as substrata availability, that can be measured as a ratio of littoral habitat to pelagic habitat (macrophytes, mud, and sand vs. open water). However, in addition to changes in water level,

this ratio is dependent on temperature, which affects littoral production. Any changes in watershed dynamics that control the amount and quality of runoff also control the productivity of the littoral zone. As a consequence, changes in cladoceran community composition can be misinterpreted as an indication of water level change, when they actually arise from changes in watershed land use or temperature shifts. Inclusion of multiple environmental variables and the use of multivariate analyses help to overcome such problems in interpretation and are the norm in modern paleolimnological research.

Applications of Subfossil Cladocera in Aquatic Research

The following case studies describe the successful use of cladocerans as indicators of various environmental conditions and changes, and demonstrate the use of cladoceran subfossils in various fields of biological research. A summary of the use of cladoceran subfossil remains in paleolimnological interpretation is given in the adjoining figure (Figure 4).

Assemblage of All Cladoceran Remains

Cladoceran responses to prehistoric human settlements

Human settlement on lake watersheds has been identified as one of the key factors regulating lake production and species

community structure in lakes after the last glacial period. Cladoceran subfossil remains preserved in lake sediments were used by Szeroczyńska (1991) to estimate impacts of prehistoric settlements in three lakes in Poland. Analysis of cladoceran community changes during a time period of >10 000 years revealed several eutrophication episodes in the lakes. The first of these reflected favorable climatic conditions after the glacial period and hence allowed the initial development of cladoceran communities. Later eutrophication episodes, especially the ones that occurred during Roman and Medieval times, reflected human influence and agricultural activities on the watershed. In this study, eutrophication was inferred from the relative abundance of eutrophic–oligotrophic indicator species and from a percentage distribution of pelagic-to-littoral species. The results were cross-calibrated with the pollen record obtained from the same lakes, which indicated more frequent human settlements to have taken place on the watershed than was revealed from the cladoceran assemblage. This disparity in the results is most likely based on the different nature of pollen and cladocerans as environmental indicators. Pollen can be carried out from distant settlements, and therefore, their indicator values of the lake's trophic status may be insignificant, while cladocerans reflect actual conditions in a given water body.

Cladoceran responses to recent watershed disturbances

Cladoceran community analysis from lake sediments has also proven to have great value in evaluating lake response to

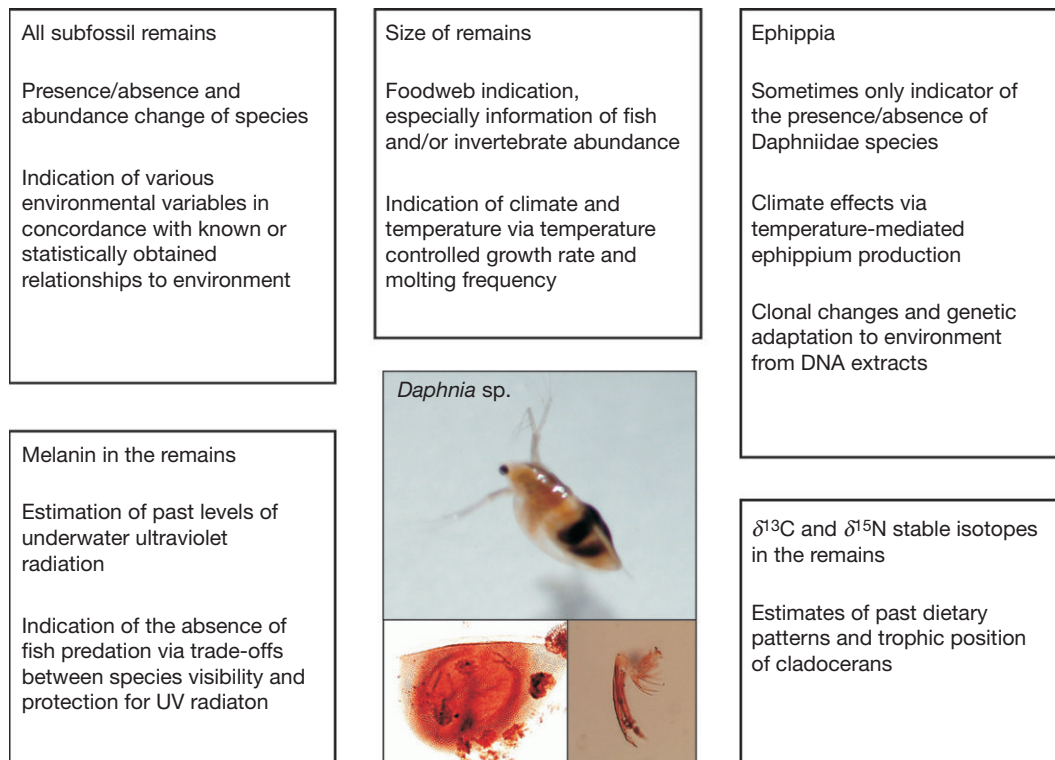


Figure 4 The use of cladoceran remains in paleolimnology. The top three boxes indicate the most commonly used applications (apart from the ephippia DNA part) and the bottom two boxes show developing techniques. More detailed information for each application is given in the text. *Daphnia* photo courtesy Patrick Van Hove (modern) and Cladocera workshop, Helsinki, 2000 (remains).

anthropogenic disturbances in shorter time scales. [Bredesen et al. \(2002\)](#) quantified effects of forest harvesting on cladoceran community assemblages in four lakes in British Columbia, Canada, that were subjected to watershed clearance ranging between 28% and 70% in the 1960s. In all lakes, the cladoceran species composition showed statistically significant changes following forest harvesting with eutrophic cladoceran taxa, for instance, *Daphnia pulex* (total phosphorus optima = $53 \mu\text{g l}^{-1}$) increasing in abundance. Based on these observed community changes, they concluded that forest-harvesting activities on the lake watershed caused nutrient changes in the water column that affected the cladoceran community via bottom-up processes. The severity of these effects depends on the topography of the area being logged and the amount of vegetation regrowth that controls the runoff of material into the lake.

Cladoceran responses to declines in lake water calcium concentrations

Calcium (Ca) concentrations are declining in many soft water lakes of the Canadian Shield and likely elsewhere ([Jeziorski et al., 2008](#)) due to complex interaction with acidification and accelerated uptake by watershed forest regrowth following tree harvesting. The Cladocera requires Ca through the use of calcium carbonate minerals as structural materials in their exoskeletons. For example, laboratory experiments have shown that *D. pulex* reproduction and survival were negatively impacted below 1.5 mg Ca l^{-1} ([Ashforth and Yan, 2008](#)). Using the Cladocera from the surface sediments from lakes in Ontario, Canada, [Jeziorski et al. \(2012\)](#) showed that within the cladoceran community, daphniids were the most responsive to Ca decline; however, responses differed within daphniid species complexes and Ca thresholds in nature appeared to be higher than those observed in the laboratory.

Size of Remains

Cladoceran size as an indicator of temperature and planktivorous fish abundance

In addition to species composition and abundance, ecological process parameters of Cladocera can be used to interpret changes in abiotic and biotic variables in the habitat. Size is one of the most commonly used process parameters. [Manca and Comoli \(2004\)](#) showed how the reduction in the range of *Daphnia* body size, measured from a head shield length, indicated time periods with cold climate and short growing season. They studied fishless mountain lakes in the Himalayas where the number of molts and *Daphnia* body size depended on the duration of the ice-free period and on summer temperatures. The results obtained from *Daphnia* stratigraphy showed a relatively strong correlation with the instrumental record of climate.

In lakes with planktivorous fish, the zooplankton size structure is influenced more by predation than temperature, with small species dominating lakes with high-predation pressure. [Jeppesen et al. \(2002\)](#) have created a transfer function to infer fish density from the size of *Daphnia* ephippia in the sediment and recently [Amsinck et al. \(2005b\)](#) tested the validity of this method in a down-core paleolimnological study in a coastal Danish lake. Their results show how the progressive eutrophication of the lake coincided with increased fish density and decreased mean size of *Daphnia* ephippia. With some

restrictions, they were also able to successfully reconstruct known fish kill incidents.

Ephippia

Ephippia production as indicator of temperature

Gamogenesis and production of resting eggs (ephippia) are considered to be under environmental control, mainly by climatic cues, although the multiple density- and predation-related factors controlling gamogenesis are only partially understood. The preservation of parthenogenetic carapaces and gamogenetic ephippia is good in the families Bosminidae and Chydoridae, allowing for a detection of variation in these two reproduction modes in subfossil assemblages. [Jeppesen et al. \(2003\)](#) showed that the ratio of ephippia and carapaces of *Bosmina* was significantly related to mean summer air temperature in a surface sediment dataset through a wide geographical transect from Greenland to New Zealand. A higher proportion of gamogenesis (ephippia) was found in colder climates. A similar relationship between mean annual temperature and *Bosmina* and chydorid ephippia was detected along a latitudinal gradient in Pan-European lakes by [Bjerring et al. \(2009\)](#). New results from a surface sediment dataset along a climate gradient from the south-boreal to the subarctic in Finland, based solely on chydorids, indicate a strong relationship between chydorid ephippia and summer temperatures ([Figure 5; Kultti et al., 2011](#)). Whether the effect of temperature on the production of ephippia is direct or indirect (food/habitat/oxygen availability) remains to be seen, but the results so far suggest that bosminid and chydorid ephippia have a powerful indicator value in the paleoclimatological research.

Using chydorids, [Sarmaja-Korjonen \(2004\)](#) used the two reproduction modes of cladocerans to estimate the length of the open-water season; high ephippium proportion is an indicator of cold climates and short open-water seasons and vice versa. Utilizing ephippium analysis and comparing the results to pollen-based temperature reconstruction curve, [Sarmaja-Korjonen and Seppä \(2007\)](#) suggested a considerable reduction in the length of the open-water season during the cold 8200 cal years BP event in Finland. Furthermore, chydorid ephippia have proven to increase and consequently indicate the harsh climate of the Little Ice Age in Finland ([Luoto et al., 2008](#)).

Subfossil genetical material reveals responses to eutrophication

To address the impact of eutrophication and the resultant increase in the abundance of nutritionally poor and toxic cyanobacteria on *Daphnia* growth rates, [Hairston et al. \(1999\)](#) combined paleolimnological methods with genetic experiments. This ingenious approach allowed them to measure *Daphnia* adaptation to eutrophication from diapausing (resting) eggs preserved in the sediments in Lake Constance in Central Europe that were hatched in the laboratory for the growth experiments. Changes in *Daphnia* clonal structure and the resultant different resistance to cyanobacteria, measured as the effect of diet on the somatic juvenile growth rate, showed that the *Daphnia* population evolved an increased ability to cope with a diet containing cyanobacteria. Genotypes from years of extensive eutrophication had higher resistance to the cyanobacteria than genotypes before the appearance of

cyanobacterial blooms. The earlier genotypes that were most heavily affected by cyanobacteria were nearly eliminated within the years of continued summertime exposure to cyanobacteria. Apart from showing that rapid adaptive evolution may significantly affect the cladoceran genetic pool, these results have important implications for paleolimnological studies because they draw attention to the still largely undiscovered potential of subfossil zooplankton remains. This is the remarkable capability of zooplankton resting eggs to store genetic

material for decades, and possibly centuries, thus providing information on species evolutionary–ecological responses to environmental change. On the other hand, these results, and the adaptive genetic response of cladocerans, raise an important question concerning the identification of fossil assemblages based on comparisons with the external morphology of modern species. Our interpretation of fossil cladoceran assemblages to reconstruct past environmental conditions rests on the assumption that these organisms have not evolved in the Late Quaternary, although increasing amount of paleogenetic evidence shows rapid evolutionary changes under environmental perturbations (e.g., Brede et al., 2009; Cousyn et al., 2001; Kerfoot and Weider, 2004; Limburg and Weider, 2002).

Chemical Composition of Cladoceran Remains

Carapace stable isotopes indicate changes in the past food web

Stable isotope analysis of different groups of organisms is routinely used in studies of food webs and ecosystem structure. Stable carbon isotopes in particular are commonly used to quantify food sources and energy flow in aquatic ecosystems, since carbon stable isotopes are known to fractionate little between each trophic transfer (Peterson and Fry, 1987). Stable nitrogen isotopes fractionate more and are typically used to infer trophic positions of consumers in food webs (Minagawa and Wada, 1984). While the technical aspects of stable isotope analysis have become easier and more affordable in recent years due to instrumental developments, this technique has only recently started to emerge in paleolimnological cladoceran studies (Perga, 2010; Perga et al., 2010; Struck et al., 1998). The main reason for the limited use of stable isotopes with subfossil organism is the relatively large sample size required per sample (minimum 0.2 mg dry weight) which means hours of microscope work picking individual remains.

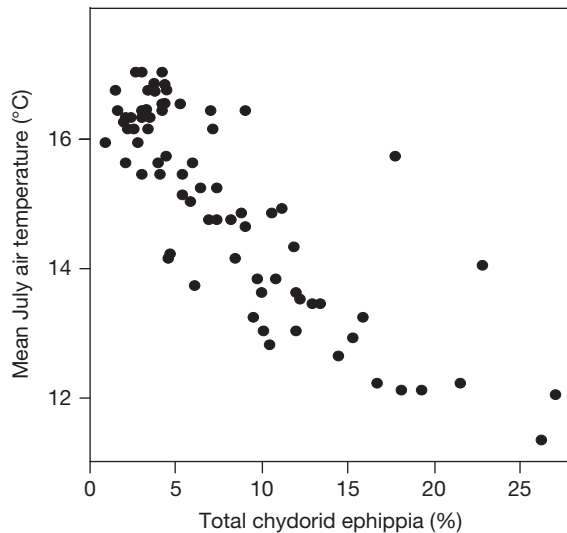


Figure 5 Relationship between total chydorid ephippia (% as calculated from the sum of carapaces + ephippia) and mean July air temperature across a latitudinal climate gradient (60–70° N) in Finland. Modified from Kultti S, Nevalainen L, Luoto TP, and Sarmaja-Korjonen K (2011) Subfossil chydorid (Cladocera, Chydoridae) ephippia as paleoenvironmental proxies: Evidence from boreal and subarctic lakes in Finland. *Hydrobiologia* 676: 23–37.

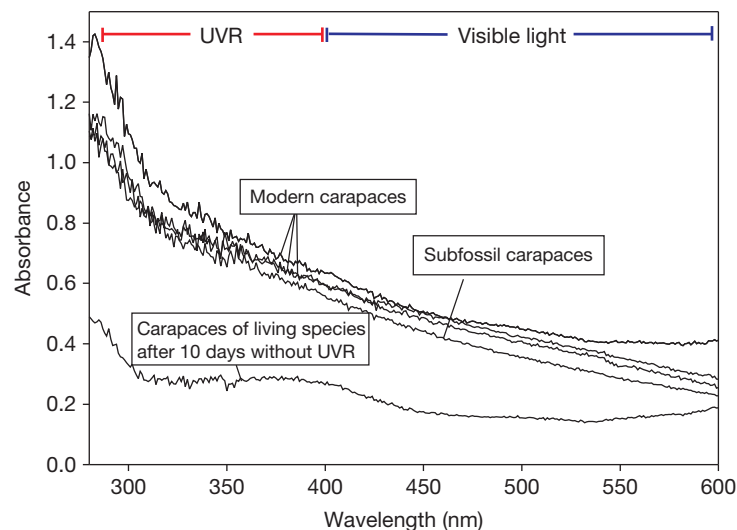


Figure 6 The difference in the *Daphnia middendorffiana* carapace melanin absorbance among modern carapaces (dissected from living freshly collected *Daphnia*), subfossil carapaces (after 2.5 years under sediment), and carapaces from individuals without ultraviolet radiation exposure. The similar spectra of the modern and subfossil carapaces suggest that melanin remains chemically intact in the sediment and provides a new application for estimating past UVR conditions in lakes.

The method also still needs better validation to ensure that $\delta^{13}\text{C}$ and $\delta^{15}\text{N}$ values of the carapaces reflect those of the whole body.

Perga (2010) investigated the C- and N-isotope composition of *Daphnia* and *Bosmina* carapaces and whole bodies, and found that they were strongly correlated. Carapace $\delta^{13}\text{C}$ was close to the whole body $\delta^{13}\text{C}$ for both species while carapaces of *Daphnia* were heavily (-7.9‰) depleted in $\delta^{15}\text{N}$ (carapaces of *Bosmina* were not analyzed). Using this validation, Perga et al. (2010) reconstructed the planktonic food web of an oligotrophic subalpine lake over the past 100 years. They also used other proxies and long-term biological data sets to better understand how different environmental drivers had affected the lake. An increase in the $\delta^{15}\text{N}$ values of *Bosmina* remains indicated a change in the *Bosmina* diet toward higher trophic levels, and the authors suggested that this was caused by a shift from phytoplankton diet to a diet dominated by heterotrophic flagellates. Additional studies from multiple lakes and species are required to confirm the usefulness of stable isotopes of cladoceran remains. It will be especially important to confirm that the nearly eight unit difference in $\delta^{15}\text{N}$ values between carapaces and the whole body applies to all lakes and species. Stable isotopes potentially provide another method to size structure to reveal changes in food web processes and when used with other proxies, allow a more accurate paleolimnological reconstruction.

Carapace melanin as indicator of past optical environments in lakes

Cladocerans in clear and shallow lakes have often dark-colored carapaces. The color indicates the presence of melanin pigment that the animal synthesizes when exposed to ultraviolet radiation (UVR). Rautio and Korhola (2002) have shown how the concentration of melanin in *Daphnia* is correlated to the water body-dissolved organic carbon concentration which largely defines the underwater UVR. They and Jeppesen et al. (2001) have suggested that carapace melanin could be measured in subfossil *Daphnia* remains and used as an indicator of changes in the past optical conditions in lakes. In order to use cladoceran melanin as a paleolimnological indicator, it is essential to know whether the melanin is chemically inert in the carapace after death. This information is now available (Figure 6) and provides a new challenging cladoceran indicator, ready to be applied in down-core studies.

Conclusions

The usefulness of cladocerans as environmental indicators and their success in paleolimnological interpretation arises from the diversity and abundance of their fossil remains, from the high intraspecies variability in reaction to different environmental changes and from their short generation time and consequent rapid response (days to weeks) to any shifts in the environment they inhabit. There are many mechanisms that link environmental parameters to cladoceran community composition in a water body. Some are caused by regional climate and temperature-induced changes and work both via direct and indirect mechanisms, while many are associated

with components of a single water body, notably with changes in water chemistry, and in physical parameters associated with watershed and water column properties. As shown in the many examples summarized here, these effects are likely to be reflected in cladoceran species composition and abundance or in more subtle ways in clonal structure or pigmentation of the species, that advancing technology has only recently been able to quantify.

See also: Diatom Introduction. Chironomid Records: Chironomid Overview. Paleolimnology: Overview of Paleolimnology. Pollen Methods and Studies: Numerical Analysis Methods.

References

- Amsinck SL, Jeppesen E, and Landkildehus F (2005a) Relationships between environmental variables and zooplankton subfossils in the surface sediments of 36 shallow coastal brackish lakes with special emphasis on the role of fish. *Journal of Paleolimnology* 33: 39–51.
- Amsinck SL, Jeppesen E, and Landkildehus F (2005b) Inference of past changes in zooplankton community structure and planktivorous fish abundance from sedimentary subfossils – A study of a coastal lake subjected to major fish kill incidents during the past century. *Archiv für Hydrobiologie* 162: 363–382.
- Ashforth D and Yan ND (2008) The interactive effects of calcium concentration and temperature on the survival and reproduction of *Daphnia pulex* at high and low food concentrations. *Limnology and Oceanography* 53: 420–432.
- Bennike O, Brodesen KP, Jeppesen E, and Walker I (2005) Aquatic invertebrates and high latitude paleolimnology. In: Pienitz R, Douglass MSV, and Smol HJB (eds.) *Long-Term Environmental Change in Arctic and Antarctic Lakes*, pp. 159–185. Dordrecht: Springer.
- Bjerring R, Becares E, Declerck S, et al. (2009) Subfossil Cladocera in relation to contemporary environmental variables in 54 Pan-European lakes. *Freshwater Biology* 54: 2401–2417.
- Brancelj A, Kernan M, Jeppesen E, et al. (2009) Cladocera remains from the sediment of remote cold lakes: A study of 294 lakes across Europe. *Fundamental and Applied Limnology, Advances in Limnology* 62: 269–294.
- Brede N, Dandrock C, Straile D, et al. (2009) The impact of human-made ecological changes on the genetic architecture of *Daphnia* species. *Proceedings of the National Academy of Sciences* 106: 4758–4763.
- Bredesen EL, Bos DG, Laird KR, and Cumming BF (2002) A cladoceran-based paleolimnological assessment of the impact of forest harvesting on four lakes from the central interior of British Columbia, Canada. *Journal of Limnology* 28: 389–402.
- Cousyn C, De Meester L, Colbourne JK, et al. (2001) Rapid, local adaptation of zooplankton behavior to changes in predation pressure in the absence of neutral genetic changes. *Proceedings of the National Academy of Sciences* 98: 6256–6260.
- Deevey ES (1942) Studies on Connecticut lake sediments. III. The biostratonomy of Lindsey Pond. *American Journal of Science* 240: 233–264.
- Frey DG (1960) The ecological significance of cladoceran remains in lake sediments. *Ecology* 41: 684–698.
- Frey DG (1962) Cladocera from the Eemian Interglacial of Denmark. *Journal of Palaeontology* 36: 1133–1154.
- Frey DG (1964) Remains of animals in Quaternary lake and bog sediments and their interpretation. *Archives für Hydrobiologie Supplement Ergebnisse der Limnologie* 2: 1–116.
- Frey DG (1982) Contrasting strategies of gamogenesis in northern and southern populations of Cladocera. *Ecology* 63: 223–241.
- Frey DG (1988) Littoral and offshore communities of diatoms, cladocerans and dipterous larvae, and their interpretation in paleolimnology. *Journal of Paleolimnology* 1: 179–191.
- Gillooly JF and Dodson SI (2000) Latitudinal patterns in the size distribution and seasonal dynamics of new world, freshwater cladocerans. *Limnology and Oceanography* 45: 22–30.
- Goulden CE (1964) The history of the cladoceran fauna of Esthwaite Water (England) and its limnological significance. *Archives of Hydrobiologie* 60: 1–52.

- Hairton NG Jr. (1996) Zooplankton egg banks as biotic reservoirs in changing environment. *Limnology and Oceanography* 41: 1087–1092.
- Hairton NG Jr., Lampert W, Cáceres CE, et al. (1999) Rapid evolution revealed by dormant eggs. *Nature* 401: 446.
- Hann BJ and Karrow P (1993) Comparative analysis of cladoceran microfossils in the Don and Scarborough Formations, Toronto, Canada. *Journal of Paleolimnology* 9: 223–241.
- Hebert PDN and Ward RD (1972) Inheritance during parthenogenesis in *Daphnia magna*. *Genetics* 71: 639–642.
- Hofmann W (1986) Developmental history of the Großer Plöner See and the Schöhsee (north Germany), cladoceran analysis, with special reference to eutrophication. *Archiv für Hydrobiologie, Supplement* 74: 259–287.
- Jeppesen E, Agerbo Madsen E, Jensen JP, and Anderson NJ (1996) Reconstructing the past density of planktivorous fish and trophic structure from sedimentary zooplankton fossils: A surface sediment calibration data set from 30 predominantly shallow lakes. *Freshwater Biology* 36: 115–127.
- Jeppesen E, Jensen JP, Amsinck S, Landkildehus F, Lauridsen T, and Mitchell SF (2002) Reconstructing the historical changes in *Daphnia* mean size and planktivorous fish abundance in lakes from the size of *Daphnia* ephippia in the sediment. *Journal of Paleolimnology* 27: 133–143.
- Jeppesen E, Jensen JP, Lauridsen TL, et al. (2003) Sub-fossil cladocerans in the surface sediment of 135 lakes as proxies for community structure of zooplankton, fish abundance and lake temperature. *Hydrobiologia* 491: 321–330.
- Jeppesen E, Leavitt P, De Meester L, and Jensen JP (2001) Functional ecology and palaeolimnology: Using cladoceran remains to reconstruct anthropogenic impact. *Trends in Ecology & Evolution* 16: 191–198.
- Jeziorski A, Paterson AM, and Smol JP (2012) Crustacean zooplankton sedimentary remains from calcium-poor lakes: Complex responses to threshold concentrations. *Aquatic Sciences* 74(1): 121–131.
- Jeziorski A, Yan ND, Paterson AM, et al. (2008) The widespread threat of calcium decline in fresh waters. *Science* 322: 1374–1377.
- Kattel GR, Battarbee RW, Mackay A, and Birks HJB (2007) Are cladoceran fossil in lake sediment samples a biased reflection of the communities from which they are derived? *Journal of Paleolimnology* 38: 157–181.
- Kerfoot CF and Weider LJ (2004) Experimental paleoecology (resurrection ecology): Chasing Van Valen's Red Queen hypothesis. *Limnology and Oceanography* 49: 1300–1316.
- Korhola A (1999) Distribution patterns of Cladocera in subarctic Fennoscandian lakes and their potential in environmental reconstruction. *Ecography* 22: 357–373.
- Korhola A and Rautio M (2001) Cladocera and other branchiopod crustaceans. In: Smol JP, Birks HJB, and Last WM (eds.) *Tracking Environmental Change Using Lake Sediments: Volume 4, Zoological indicators*, pp. 5–41. Dordrecht: Kluwer Academic Publishers.
- Korosi J and Smol JP (2012a) An illustrated guide to the identification of cladoceran subfossils from lake sediments in northeastern North America: Part 1 – the Daphniidae, Leptodoridae, Bosminidae, Polyphemidae, Holopedidae, Sididae, and Macrothricidae. *Journal of Paleolimnology* 48(3): 571–586.
- Korosi J and Smol JP (2012b) An illustrated guide to the identification of cladoceran subfossils from lake sediments in northeastern North America: Part 2 – the Chydoridae. *Journal of Paleolimnology* 48(3): 587–622.
- Kultti S, Nevalainen L, Luoto TP, and Sarmaja-Korjonen K (2011) Subfossil chydorid (Cladocera, Chydoridae) ephippia as paleoenvironmental proxies: Evidence from boreal and subarctic lakes in Finland. *Hydrobiologia* 676: 23–37.
- Latreille PA (1829) Les crustaceans, les arachnides, les insectes. In: Cluvier G (ed.) *Le regle animal distribue d'apres son organisation, pour servir base a l'histoire naturelle des animaux et d'introduction a l'anatomie*. Paris: Deterville.
- Limburg PA and Weider LJ (2002) 'Ancient' DNA in the resting egg bank of a microcrustacean can serve as a paleolimnological database. *Proceedings of the Royal Society of London. Series B* 269: 281–287.
- Luoto TP, Nevalainen L, and Sarmaja-Korjonen K (2008) Multi-proxy evidence for the 'Little Ice Age' from Lake Hamträsk, Southern Finland. *Journal of Paleolimnology* 40: 1097–1113.
- Lynch M and Gabriel W (1983) Phenotypic evolution and parthenogenesis. *American Naturalist* 122: 745–764.
- Manca M and Comoli P (2004) Reconstructing long-term changes in *Daphnia's* body size from subfossil remains in sediments of a small lake in the Himalayas. *Journal of Paleolimnology* 32: 95–107.
- Minagawa M and Wada E (1984) Stepwise enrichment of ^{15}N along food chains: Further evidence and the relation between $\delta^{15}\text{N}$ and animal age. *Geochimica et Cosmochimica Acta* 48: 1135–1140.
- Nevalainen L (2010) Evaluation of microcrustacean (Cladocera, Chydoridae) biodiversity based on sweep net and surface sediment samples. *Ecoscience* 17: 356–364.
- Nevalainen L (2011) Intra-lake heterogeneity of cladoceran (Crustacea) assemblages forced by local hydrology. *Hydrobiologia* 676: 9–22.
- Patalas K (1990) Diversity of the zooplankton communities in Canadian lakes as a function of climate. *Verhandlungen der Internationalen Vereinigung für Theoretische und Angewandte Limnologie* 24: 360–368.
- Perga M-E (2010) Potential of $\delta^{13}\text{C}$ and $\delta^{15}\text{N}$ of cladoceran subfossil exoskeletons for paleoecological studies. *Journal of Paleolimnology* 44: 387–395.
- Perga M-E, Desmer M, Enters D, and Reyss J-L (2010) A century of bottom-up and top-down-driven changes on a planktonic food web: A paleoecological and paleoisotopic study of Lake Annecy, France. *Limnology and Oceanography* 55: 803–816.
- Peterson BJ and Fry B (1987) Stable isotopes in ecosystem studies. *Annual Review of Ecological Systems* 18: 293–320.
- Rautio M (1998) Community structure of crustacean zooplankton in subarctic ponds – effects of altitude and physical heterogeneity. *Ecography* 21: 327–335.
- Rautio M and Korhola A (2002) UV-induced pigmentation in subarctic *Daphnia*. *Limnology and Oceanography* 47: 295–299.
- Sarmaja-Korjonen K (2004) Chydorid ephippia as indicators of past environmental changes – A new method. *Hydrobiologia* 526: 129–136.
- Sarmaja-Korjonen K and Seppä H (2007) Abrupt and consistent responses of aquatic and terrestrial ecosystems to the 8200 cal yr cold event – A lacustrine record from Lake Arapisto, Finland. *The Holocene* 17: 457–467.
- Schwenk K, Junttila P, Rautio M, et al. (2004) Ecological, morphological, and genetic differentiation of *Daphnia* (*Hyalodaphnia*) from the Finnish and Russian subarctic. *Limnology and Oceanography* 49: 532–539.
- Struck U, Voss M, von Bodungen B, and Mumm N (1998) Stable isotopes of nitrogen in fossil cladoceran exoskeletons: Implications for nitrogen sources in the central Baltic Sea during the past century. *Naturwissenschaften* 85: 597–603.
- Sweetman JN and Smol JP (2006) A guide to the identification of cladoceran remains (Crustacea, Branchiopoda) in Alaskan lake sediments. *Archives of Hydrobiology* 151(supplement): Monographic Studies, 353–394.
- Szeroczyńska K (1991) Impact of prehistoric settlements on the Cladocera in the sediments of Lakes Suszek, Bledowo, and Skrzetuszewskie. *Hydrobiologia* 225: 105–114.
- Szeroczyńska K and Sarmaja-Korjonen K (2007) *Atlas of Subfossil Cladocera from Central and Northern Europe p. 84*. Świecie: Friends of the Lower Vistula Society.

Freshwater Mollusks

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Introduction

Freshwater mollusks (Phylum Mollusca) are among the most ubiquitous macroscopic remains preserved in continental Quaternary sediments. They are exclusively represented by gastropods (Class Gastropoda) and bivalves (Class Bivalvia). With the only exception of opisthobranchs (Class Gastropoda), all have an external calcareous shell (of one or two pieces, respectively) which is generally the only fossilizable structure in the sedimentary record (Figure 1). Indeed, the freshwater mollusk evidence from the Quaternary usually involves unaltered shells, in contrast to fossils from older deposits that mainly consist of molds or casts.

The fossil record of freshwater mollusks is characterized by low diversity compared to the marine record due, in part, to poorer preservation of organisms as well as the ephemeral nature of many freshwater habitats (Taylor, 1988). Fossilization is biased toward lowland and lake deposits, which are low-energetic environments where fossilization is favored. Fossiliferous river deposits are fewer than those of lakes, and the record is negligible or absent for such habitats as upland seepages, springs, and subterranean streams (Gray, 1988). This incomplete record is also due to the poor preservation potential of the often light, thin shells of most freshwater taxa. In fact, the fossil record of freshwater gastropods is patchy at best and likely to significantly underestimate the age and diversity of freshwater lineages (Strong et al., 2008).

Despite these difficulties, assemblages of mollusk shells can add important information to paleoecological and paleolimnological reconstructions (Gray, 1988). They have been widely used to reconstruct ancient local habitats and climatic conditions in combination with other biological and sedimentary proxy records. A number of modern studies have focused on the analysis of stable isotopes ($\delta^{18}\text{O}$ and $\delta^{13}\text{C}$) for inferring past temperatures, evaporative effects, and productivity levels. Much recent work has been directed to the evaluation of the processes and agents involved in the genesis of fossil assemblages, such as their compositional fidelity and taphonomy. The aim of this chapter is to review the main research directions associated with Quaternary freshwater mollusks that contribute to make them effective paleolimnological indicators. An overview of the present status of research, advances in methodology, as well as the problems and possibilities of every research direction are presented and discussed. Some case studies are also presented that exemplify various applications in fossil reconstructions.

Taxonomy

Most taxa encountered as fossils in Quaternary freshwater deposits are referable to extant species (Figure 2). Consequently, their reliable identification requires a good knowledge of the

taxonomy of modern fauna. Bivalves and gastropods include a variety of groups derived independently from marine ancestry, and varying in range of habitat, reproduction, and distribution.

Mollusk taxonomy is rather complicated and is at present being substantially revised. New suites of anatomical, ultrastructural, and molecular characters developed in the past 30 years have fuelled a revolution in our understanding of gastropod phylogenetics (see Bouchet and Rocroi, 2005 and references therein). The current understanding of the phylogeny of bivalves is still developing (Bogan, 2008).

Bivalves

Bogan (2008) listed a total of 19 families with 206 genera and an estimated 1026 species of freshwater bivalves (Table 1). Freshwater bivalves are found in three subclasses: Pteriomorpha, Paleoheterodonta, and Heterodonta. The Pteriomorpha, mainly a marine group, is only represented in freshwater by four species of Arcidae (genus *Scaphula*) in India, Burma, and Thailand and by five species of Mytilidae. The Paleoheterodonta include the order Unioniformes, which is one of the groups with larger radiations in freshwater. Within this order, the most diverse family is the Unionidae. Bogan (2008) estimated a total of 621 species worldwide, while Graf and Cummings (2007) calculated approximately 674 species. Two main areas exhibiting high diversity and endemism are the southeastern United States and southern Asia. In fact, a maximum of 42 genera and 271 species of unionids were reported for the southeastern United States. On the other hand, the Hyriidae and the Mycetopodidae account for the majority of bivalve faunal diversity in South America. The subclass Heterodonta encompasses numerous families, each having only 1–5 genera and species. Exceptions are the Sphaeriidae and Corbiculidae, with larger radiations. A maximum of 196 species of Sphaeriidae has been described, most of them in South America (Bogan, 2008).

Gastropods

Gastropods are more diverse than bivalves, with a total fauna estimated at approximately 4000 valid species. By far, the most diverse group is the Rissooidea. While most families are probably known within 70–90% of actual diversity, the estimated 1000 species of Rissooidea may represent as little as 25% of their actual diversity, as evidenced by the fact that they comprise about 80% of current new species descriptions (Strong et al., 2008).

The freshwater gastropod fauna is dominated by two main components: the Caenogastropoda and pulmonate heterobranchs (Table 2; Bouchet and Rocroi, 2005). Some Neritimorpha (Neritiliidae, Neritidae; previously a subgroup of 'archaeogastropods'; Bouchet and Rocroi, 2005) have also invaded freshwater.

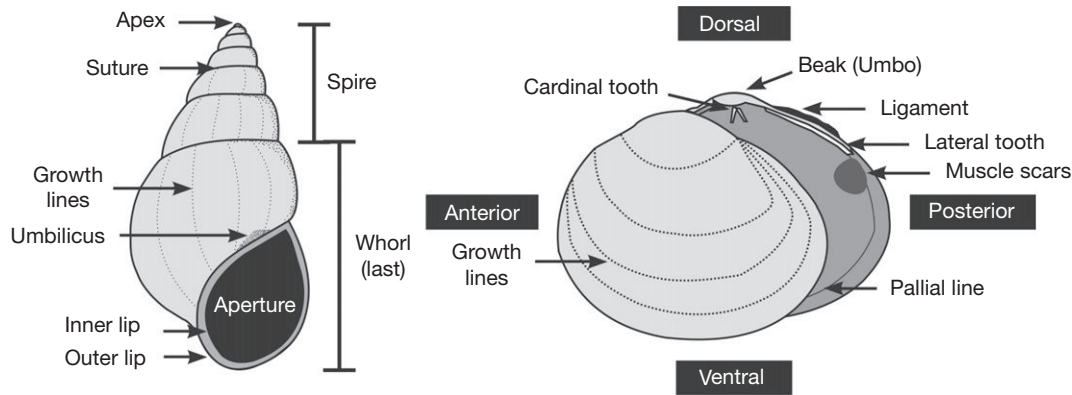


Figure 1 Shell characteristics of gastropods (left) and bivalves (right).

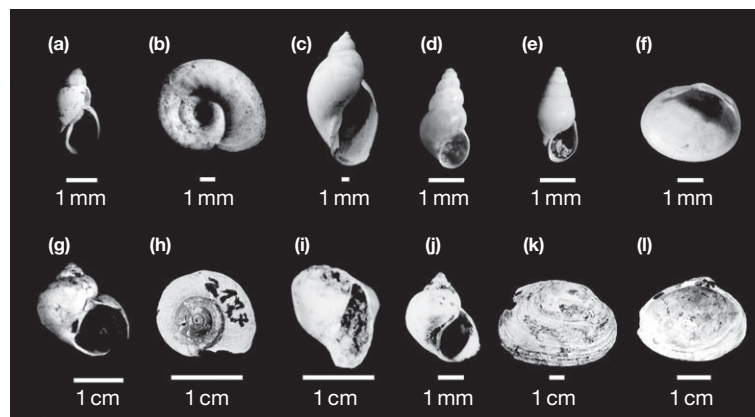


Figure 2 Some representative mollusks from the continental Quaternary of Argentina (top row) and Uruguay (bottom row). (a) *Lymnaea viator*, (b) *Biomphalaria peregrina*, (c) *Chilina mendozana*, (d) *Heleobia parchappii*, (e) *Heleobia* cf. *kuesteri*, (f) *Pisidium chiquitanum*, (g) *Pomacea canaliculata*, (h) *Biomphalaria tenagophila*, (i) *Chilina fluminea*, (j) *Potamolithus* sp., (k) *Diplodon paraeformis*, and (l) *Neocorbicula limosa*. Modified from Martínez SA and Rojas A (2004) Quaternary continental molluscs from Northern Uruguay: distribution and paleoecology. *Quaternary International* 114: 123–128; De Francesco CG and Hassan GS (2009) The significance of molluscs as paleoecological indicators of freshwater systems in central-western Argentina. *Palaeogeography, Palaeoclimatology, Palaeoecology* 274: 105–113, with permission.

The Caenogastropoda (for many years named Prosobranchia and containing most of the former Mesogastropoda and all the Neogastropoda) share a few (probably ancestral) characters: they breathe through gills, carry an operculum (which sometimes fossilizes), and are usually gonochoristic (separated into two distinct sexes) and occasionally parthenogenetic (reproduction through the development of unfertilized eggs), but only rarely hermaphroditic (having both male and female reproductive organs in the individual). Caenogastropods have invaded freshwaters on at least as many occasions as the number of their superfamilies. The families Viviparidae, Pleuroceridae, Hydrobiidae, and Pomatiopsidae are worldwide in distribution. The Melanopsidae and Bithyniidae are both restricted to the Eastern Hemisphere. The Ampullariidae and Thiaridae are circumtropical, with distributions reaching to the subtropics (Dillon, 2000).

The Heterobranchia (containing the former Opisthobranchia and Pulmonata, as well as the so-called Allogastropoda or 'lower Heterobranchia'; Bouchet and Rocroi, 2005) include all the remaining gastropod families. Within this group,

pulmonates are the most abundant. They have lost their gills and now respire over the inner surface of their mantle, effectively a lung (some groups also developed a secondary gill). They typically have much lighter shells than caenogastropods and lack an operculum. In addition, they are hermaphroditic. All freshwater pulmonate families belong to the Basommatophora, so named because their eyes are at the base of their tentacles. They seem to have derived from a single ancient invasion of freshwater, dating at the latest, to the Jurassic period. Minor families (including the limpet-shaped Acroloxidae) are held to be the most ancestral of the freshwater pulmonates, by virtue of anatomical detail. The worldwide Lymnaeidae, with their rather ordinary looking, dextral shells of medium to high spire, are believed the next most ancestral. Members of the primarily holarctic Physidae are distinguished by their inflated shells coiled in a left-handed ('sinistral') fashion, quite unusual in the class Gastropoda. Members of the worldwide Planorbidae are also sinistral, but often planispiral (coiled in one plane, like a flat spiral; Dillon, 2000). They include the limpets previously separated as Ancyliidae.

Table 1 Taxonomic classification and total number of valid, described genera and species of freshwater bivalves

<i>Subclass</i>	<i>Order</i>	<i>Family</i>	<i>Genera/species</i>	<i>Examples</i>
Pteriomorpha	Arcoida	Arcidae	1/4	<i>Scaphula pinna</i>
	Mytiloida	Mytilidae	4/6	<i>Limnoperna fortunei</i>
Paleoheterodonta	Unioniformes	Etheriidae	1/1	<i>Etheria elliptica</i>
		Hyriidae	17/83	<i>Diplodon patagonicus</i>
		Iridinidae	6/41	<i>Mutela rostrata</i>
		Margaritiferidae	3/12	<i>Margaritifera margaritifera</i>
		Mycetopodidae	12/39	<i>Anodontites trapesialis</i>
		Unionidae	142/621	<i>Elliptio complanata</i>
Heterodonta	Veneroida	Cardiidae	2/5	<i>Adacna laeviscula</i>
		Corbiculidae	3/6	<i>Corbicula fluminalis</i>
		Sphaeriidae	5/196	<i>Sphaerium simile</i>
		Dreissenidae	2/4	<i>Dreissena polymorpha</i>
		Solenidae	1/1	<i>Pharella waltoni</i>
		Donacidae	2/2	<i>Egeria paradoxa</i>
		Novaculidae	1/2	<i>Novaculina siamensis</i>
	Myoida	Corbulidae	1/1	<i>Guianadesma sinuosum</i> ^a
		Erodonidae	2/2	<i>Erodona afra</i>
		Teredinidae	1/1	<i>Nausitora dunlopei</i>
		Lyonsiidae	1/1	<i>Anticorbula fluviatilis</i>

^aHuber (2010) considered this species a junior synonym of *Anticorbula fluviatilis*. This lowers the total number of families represented in freshwater to 18.

Data from Bogan, 2008 and Freshwater Animal Diversity Assessment (FADA) Project (<http://fada.biodiversity.be>).

Table 2 Taxonomic classification and total number of valid, described species of freshwater gastropods

<i>Clade</i>	<i>Informal group</i>	<i>Superfamily</i>	<i>Family</i>	<i>Species</i>	<i>Examples</i>
Neritimorpha		Helicinoidea	Neritillidae	5	<i>Neritilia succinea</i>
		Neritoidea	Neritidae	110	<i>Vittina natalensis</i>
Caenogastropoda	Architaenioglossa	Ampullarioidea	Ampullariidae	105–170	<i>Pomacea canaliculata</i>
		Viviparoidea	Viviparidae	125–150	<i>Vivipara dissimilis</i>
		Cerithioidea	Melanopsidae	25–50	<i>Fagotia wuesti</i>
			Paludomidae	100	<i>Cleopatra africana</i>
			Pachychilidae	165–225	<i>Pachychilus indiorum</i>
			Pleuroceridae	200	<i>Doryssa cachoeirae</i>
			Thiaridae	135	<i>Melanoidea tuberculata</i>
		Littorinoidea	Littorinidae	2	<i>Cremnoconchus syhadrensis</i>
		Rissooidea	Amnicolidae	200	<i>Amnicola limosa</i>
			Assimineidae	20	<i>Assiminea infima</i>
			Bithyniidae	130	<i>Bithynia tentaculata</i>
			Cochliopidae	246	<i>Heleobia parchappii</i>
			Helicostoidae	1	<i>Helicostoa sinensis</i>
			Hydrobiidae	1250	<i>Potamopyrgus antipodarum</i>
			Lithoglyphidae	100	<i>Potamolithus lapidum</i>
			Moitessieriidae	55	<i>Moitessieria rayi</i>
			Pomatiopsidae	170	<i>Oncomelania hupensis</i>
			Stenothyridae	60	<i>Stenothyra hunanensis</i>
		Buccinoidea	Buccinidae	8–10	<i>Clea helena</i>
		Muricoidea	Marginellidae	2	<i>Rivomarginella morrisoni</i>
Heterobranchia	Allogastropoda	Glacidorboidea	Glacidorbidae	20	<i>Glacidorbis bicarinatus</i>
		Valvatoidea	Valvatidae	71	<i>Valvata cristata</i>
	Opisthobranchia	Acochlidioidea	Acochlididae	4	<i>Acochlidium fijiensis</i>
		Hedylopsoidea	Tantulidae	1	<i>Tantulum elegans</i>
		Strubellioidea	Strubelliidae	1	<i>Strubellia paradoxa</i>
	Pulmonata	Chilinoidea	Chiliniidae	15	<i>Chilina parchappii</i>
			Latiidae	1	<i>Latia neritoides</i>
		Acroloxoidea	Acroloxidae	40	<i>Acroloxus lacustris</i>
		Lymnaeoidae	Lymnaeidae	100	<i>Lymnaea viator</i>
		Planorboidea	Physidae	80	<i>Physa acuta</i>
			Planorbidae	250	<i>Ancylus fluviatilis</i>

Classification follows Bouchet and Rocroi (2005). Data on species abundance from Strong et al. (2008).

It is very common that a single species may occur in a wide variety of water bodies because it is eurytopic (or a generalist) and has a wide tolerance to a particular variable. Even species considered to be stenotypic, or limited in their environmental tolerances, may occasionally occur well outside their normal range (Ellis, 1940). An example of this is the limpet *Ancylus fluviatilis*, which lives on clean stones in quickly running streams. This led Sparks (1964) to conclude that the interpretation of freshwater mollusk paleoecology becomes a matter of relative probability. The corbiculid clam *Corbicula fluminalis*, a conspicuous bivalve present in the Pleistocene of North-West Europe, principally inhabits freshwater habitats, but there is evidence that it could also tolerate mildly brackish conditions (Meijer and Preece, 2000). The thiarid gastropod *Melanoidea tuberculata* occurs abundantly in a wide range of modern and Quaternary fresh and brackish water habitats throughout Africa and Asia. In highly evaporated lakes, too saline for most freshwater organisms, *M. tuberculata* is often the only mollusk present (Leng et al., 1999; Pointier et al., 1992). Similarly, the rissoiidean snail *Heleobia parchappii* lives both in fresh and saline water. When this species occurs in saline water, it is the only mollusk present; however, when in freshwater, it is usually accompanied by other mollusk species (De Francesco and Hassan, 2009). These euryhaline species (having a wide range of tolerance to salinity) still bring some important advantages as paleobioindicators. Although the species from freshwater habitats cannot be morphologically distinguished from those inhabiting saline waters, the overall composition of the whole assemblage may be indicative of these two habitat extremes. In fact, the presence of monotypic deposits of any of these species (Figure 4) would be indicative of elevated salinity in the past.

Examples of paleolimnological reconstructions based on the paleoecology of freshwater mollusks are found worldwide. Yang et al. (2001) studied the mollusk assemblages recovered from two Holocene marl deposits in southwestern Ontario (Canada), indicating cool shallow vegetated water bodies. As the authors found some differences between cores obtained at different depths, they recommended a minimum of two sampling sites from any given sedimentation basin, near shore and mid-basin, to maximize the assessment of assemblages and to obtain a more reliable paleolimnological reconstruction. Similarly, Martínez and Rojas (2004) found differences in species diversity in Pleistocene fossil assemblages of northern Uruguay that correlated with the variety of sub-environments represented. Oswald (2006) reconstructed the variations in depth of an arctic shallow lake during the

mid-Holocene based on the presence and abundance of two species of Sphaeriidae indicative of shallow water (*Pisidium nitidum* and *P. rotundatum*). De Francesco et al. (2007) reconstructed the hydrological dynamics of a very shallow vegetated water body in a Pleistocene alluvial succession from Mendoza (Argentina). According to the alternation between assemblages of aquatic and hygrophilous land snails, they recognized four main stages indicating occasional submersions within a semi-temporary regime (Figure 5). More recently, De Francesco and Hassan (2009) discriminated between two mollusk assemblages indicative of different energetic environmental conditions related to the stability of the flow regime: (1) lentic habitats (shallow lakes) or calm backwaters (pools) within lotic systems such as streams and rivers and (2) lotic habitats where a directional water flow exists. These same groups were recognized in the fossil record, allowing the reconstruction of habitats that differed in the magnitude of water flow. Similar differences in hydrological regime have been reported by Grimley et al. (2009) for a Pleistocene lacustrine section in western Tennessee, USA. Here, a change from a permanent lake phase at ca. 24–23.6 ka to a marsh/shallow lake environment at circa 23.6–23.1 ka is suggested by the replacement of aquatic snails (e.g., *Ammicola limosa*, *Gyraulus parvus*, *Valvata tricarinata*) by amphibious species (*Pomatiopsis lapidaria*, *P. scalaris*, *Fossaria dalli*, *F. parva*).

Stable Isotopes

Freshwater mollusks are also very useful to evaluate paleoenvironmental and climatic conditions through the analysis of the variations in stable isotope composition recorded in the shell carbonate. To allow paleoenvironmental inference, shell carbonate should be deposited at or near isotope equilibrium with the ambient water. This enables differentiation between environmental and intrinsic organism variation (a condition referred to as ‘vital effects’). Examples of both isotope equilibrium and disequilibrium precipitation have been reported for freshwater mollusks. Shanahan et al. (2005) illustrated the potential complexities of interpreting stable isotopic data from fossil gastropod shells, even when environmental conditions are nearly constant, and placed limitations on the paleoenvironmental deductions that can be made. They also suggested that explaining the differences between the shell isotopic compositions of lung- and gill-breathing snails requires a combination of both behavioral and physiological factors.

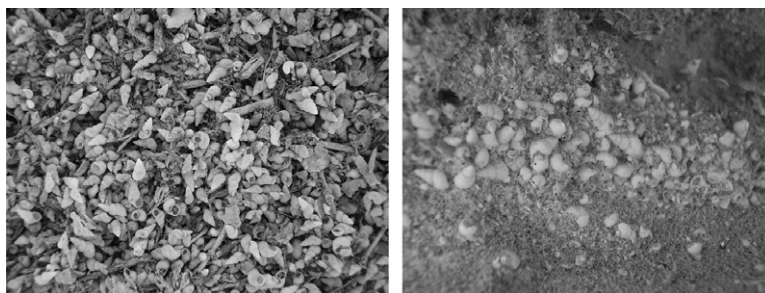


Figure 4 Monotypic shell concentrations of *Heleobia parchappii* as indicators of elevated salinity in the past. Left: Modern death shell assemblages in a saline stream. Right: Holocene assemblages in an outcropping alluvial succession.

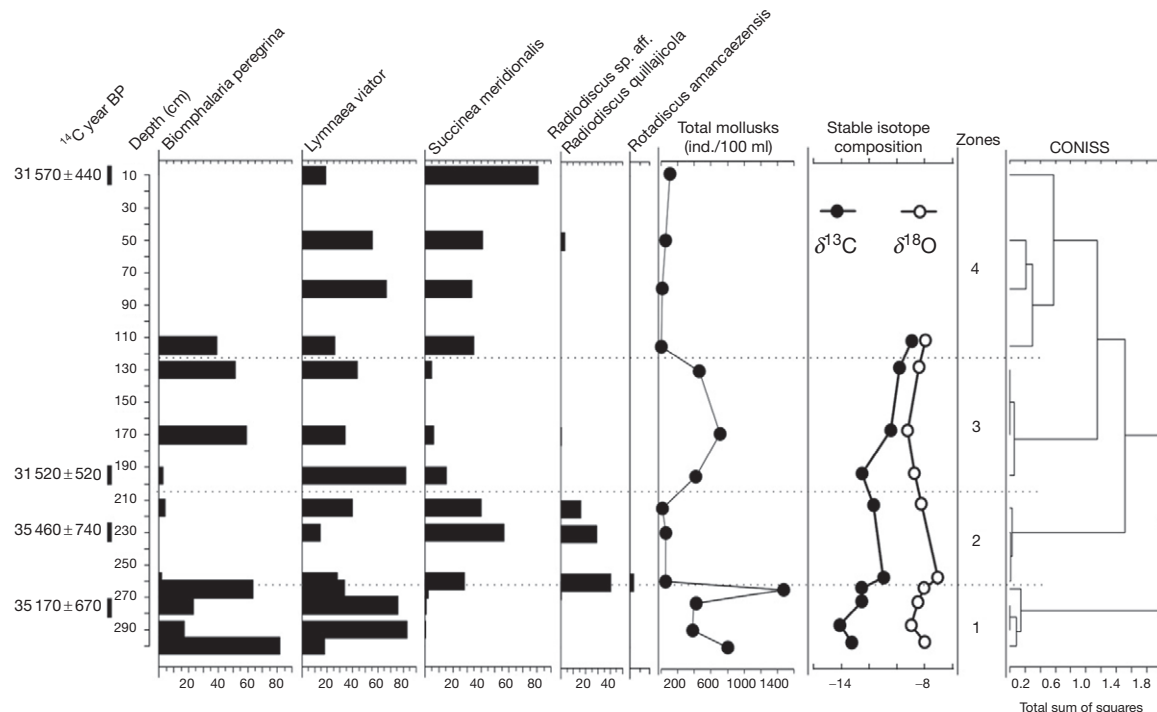


Figure 5 Relative abundances of species, and total densities of mollusk shells from La Bomba sequence (Pleistocene). The alternation between levels dominated by freshwater pulmonate snails (*Biomphalaria peregrina* and *Lymnaea viator*) and levels dominated by hygrophilous land snails (*Succinea meridionalis*, *Radiodiscus* aff. *Radiodiscus quillajicola*, and *Rotadiscus amancaezensis*) allows one to recognize the evolution of an ephemeral shallow water body subject to periodical droughts. Adapted from De Francesco CG, Zárate MA, and Miquel SE (2007) late Pleistocene mollusk assemblages and inferred paleoenvironments from the Andean piedmont of Mendoza, Argentina. *Palaeogeography, Palaeoclimatology, Palaeoecology* 251: 461–469, with permission.

Oxygen Stable Isotopes ($\delta^{18}\text{O}$)

As the oxygen isotope composition of shell carbonate usually precipitates in isotopic equilibrium with ambient water, and because physiological effects are often minor, oxygen isotopes provide useful estimates of environmental temperature change by indicating the effect of temperature on evaporation. Relatively high values, for instance, may be interpreted as indicating high evaporation/precipitation ratios, which are considered to reflect relatively warm, dry conditions, while relatively low values could indicate cooler and moister times (Stuiver, 1968). Yet, Fastovsky et al. (1993) observed that the unionid bivalve *Elliptio complanata* does not secrete aragonite in oxygen isotopic equilibrium with the water in which it grows. Similarly, Shanahan et al. (2005) found that even in springs with uniform environmental conditions, variation in isotopic composition of several species of freshwater snails exceeded predicted values. Therefore, paleoclimatic inferences based on oxygen isotopes should be performed with great caution, taking into consideration the vital effects of particular species.

The application of $\delta^{18}\text{O}$ values of freshwater shells as paleoclimatic indicators has resulted in some significant contributions. As an example, a succession of $\delta^{18}\text{O}$ values in a 37-m core obtained from the crater lake of Valle di Castiglione (near Rome) allowed discrimination between two climatic phases. The succession of $\delta^{18}\text{O}$ values in the core interval 37–2.3 m ranged from -2.8 to $+6.9\text{‰}$ with only six samples below 0‰ , indicating arid climatic phases. In contrast,

only negative $\delta^{18}\text{O}$ values (from -13.3 to -0.5) were found above this level, corresponding to wet climatic periods, in agreement with a strong evaporative control on the lake water isotopic composition (Zanchetta et al., 1999). In another example, the existence of arid conditions between 35 and 15 ka BP and warmer conditions around 35–25 ka BP was suggested for the Pampa and Patagonia regions of Argentina from the analysis of $\delta^{18}\text{O}$ composition of freshwater gastropod shells (Bonadonna et al., 1999).

Carbon Stable Isotopes ($\delta^{13}\text{C}$)

Stable carbon isotope data of freshwater shells have historically been used as proxies of photosynthesis, respiration, decay, and paleoproductivity (see Miller and Tevesz, 2001 and references therein). However, the interpretation of $\delta^{13}\text{C}$ composition remains controversial. This is because both respired CO_2 and ambient inorganic carbon contribute to mollusk shells and the relative importance of each source will determine whether $\delta^{13}\text{C}$ records mainly dietary $\delta^{13}\text{C}$, or the $\delta^{13}\text{C}$ of ambient dissolved inorganic carbon (DIC). Shanahan et al. (2005) found that for lung-breathing pulmonate gastropods, shell $\delta^{13}\text{C}$ mainly reflects diet. On the other hand, shell $\delta^{13}\text{C}$ of gill-breathing mollusks mainly reflects DIC and, therefore, may record changes in pH, temperature, or salinity (Fritz and Poplawski, 1974; McConnaughey and Gillikin, 2008; Shanahan et al., 2005). However, in some gill-breathing

unionids as well as in the pearl mussel *Margaritifera margaritifera*, the use of $\delta^{13}\text{C}$ shells as a recorder for environmental signals is questionable since they are strongly influenced by individual metabolic activity (Dettman et al., 1999; Geist et al., 2005). Wurster and Patterson (2001) suggested that $\delta^{13}\text{C}_{(\text{shell})}$ values appear to be related to physiological variables such as metabolic activity, trophic position, and/or reproduction, which may have significant implications for evolutionary and physiological studies, for both extant and extinct species. Because of the many different possible origins of shell $\delta^{13}\text{C}$ (environmental CO_2/O_2 ratio, DIC content, or the animal's physiology), the proper interpretation of records depends on context (McConnaughey and Gillikin, 2008).

Strontium Stable Isotopes ($^{87}\text{Sr}/^{86}\text{Sr}$)

Another stable isotope ratio that has received less attention than $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ but is at times employed in paleolimnological reconstructions is the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio. Mollusk shells are rich in calcium carbonate and strontium, the latter documenting the strontium ratio of water from which they were deposited if samples were not subject to geochemical reactions after mineral formation (Reinhardt et al., 1998). Because of similar atomic weights, ^{87}Sr and ^{86}Sr do not fractionate

measurably during natural chemical reactions or physical transformations. The study of strontium ratios has been successfully applied to the study of ancient lake evolution in a number of places (e.g., Fan et al., 2010; Hart et al., 2004; Rosenthal et al., 1989).

Sample Size Effects

The number of shells analyzed may contribute to mask interpretations of environmental change. Jones et al. (2002) found differences in $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values between two coexisting snails (*G. piscinarum* and *Valvata cristata*) in an 8-m sediment core taken at Lake Gölhisar, Turkey (Figure 6). Covariation between $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ suggested that carbon incorporated into the shells was not substantially derived from metabolic carbon but rather from DIC in the lake water. They interpreted these results as a spatial (i.e., microhabitat) or temporal (time of calcification) effect. Lake water $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ have been shown to change annually. Consequently, Jones et al. (2002) suggested that only by analyzing a sufficient number of individual shells from each level in a given sequence can a reliable interpretation of environmental change be obtained. Studies where only one or two shells per level are analyzed are unlikely to provide a true indication of the variability within the period

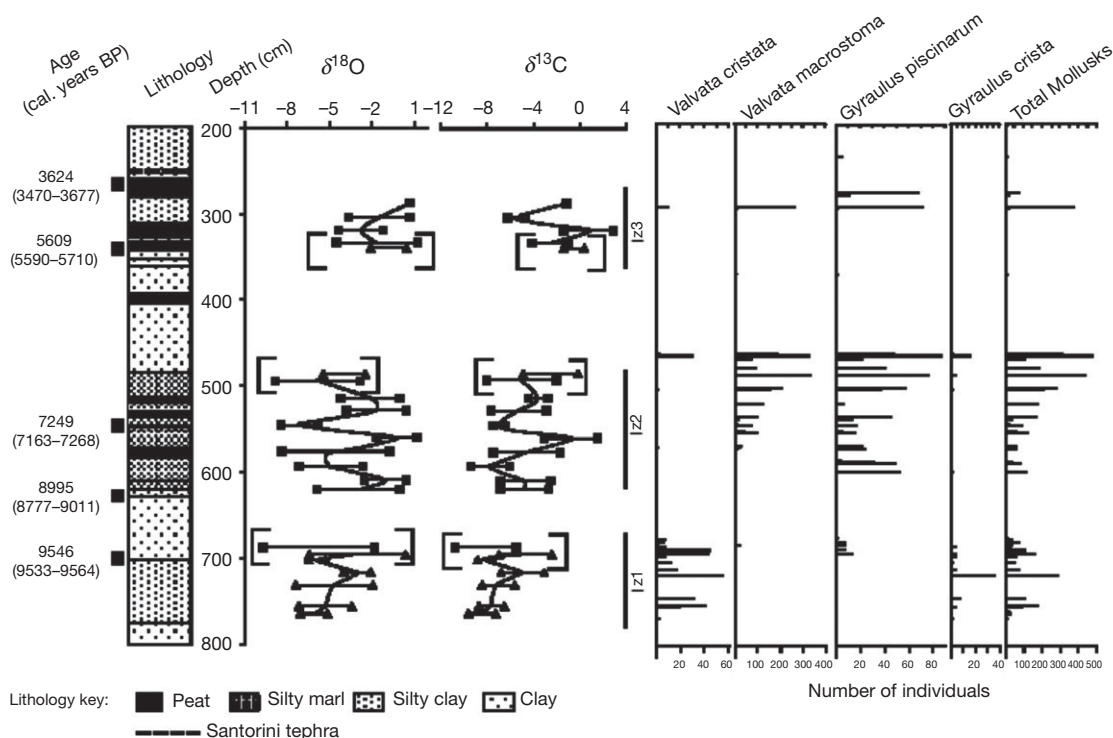


Figure 6 Variations in $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ shell composition of two snail species in an 8-m sediment core from Lake Gölhisar (Turkey). The range of values for *Gyraulus piscinarum* (squares) and *V. cristata* (triangles) at each sample level are shown with a trend line running through the midpoint of each level for isotope zones 1–3 (Iz 1–3). Sample levels where significant numbers (≥ 4) of both species were analyzed are bracketed (bracketed values are one standard deviation either side of the given values). The more positive $\delta^{18}\text{O}$ values recorded in isotope zones 2 and 3 reflect periods of greater aridity, when lake water was more strongly evaporated. On the other hand, the more negative values observed in isotope zone 1 reflect more positive water balance within the lake system and, probably, higher humidity. In the three levels where both species were analyzed in significant numbers, $\delta^{13}\text{C}$ and to a lesser extent $\delta^{18}\text{O}$ values for *V. cristata* were always more positive than those of *G. piscinarum*. This difference was interpreted as a spatial (i.e., habitat) or temporal (time of calcification) effect. Adapted from Jones MD, Leng MJ, Eastwood WJ, Keen DH, and Turney CSM (2002) Interpreting stable-isotope records from freshwater snail-shell carbonate: A Holocene case study from Lake Gölhisar, Turkey. *The Holocene* 12: 543–548, with permission.

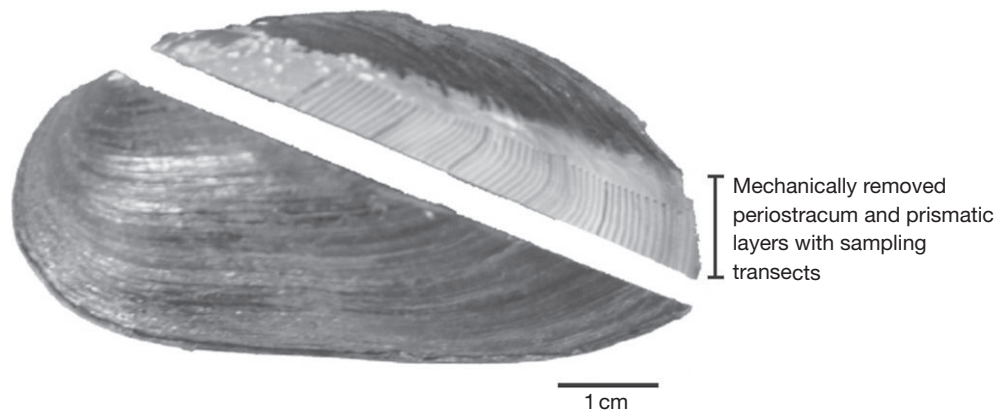


Figure 7 Micromilling of a valve of *Elliptio* sp. for sclerochronological analysis. The valve surface is physically cleaned to remove the periostracum, prismatic layer, and any exterior contamination. Adapted from Carroll M, Romanek C, and Paddock L (2006) The relationship between the hydrogen and oxygen isotopes of freshwater bivalve shells and their home streams. *Chemical Geology* 234: 211–222, with permission.

represented by the sample. This is especially true of small lakes in which there are potentially large temporal variations in lake water-isotope values.

Sclerochronology

As shell carbonate is deposited continually over the lifetime of the organisms, the analysis of the ontogenetic variations in $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values along skeletal growth lines of fossil and recent mollusks can be used as a high-resolution proxy to reconstruct variations in aquatic environments, as well as to characterize the physiology of the CaCO_3 secreting organism. Recent technological advances permit microsampling of carbonate to obtain records with subannual and even daily resolution (Figure 7). This is the main focus of sclerochronology, a discipline originally defined as the skeletal equivalent of dendrochronology (Hudson et al., 1976). Sclerochronology may be referred as the study of physical and chemical variation in the accretionary skeletons of organisms.

Sclerochronology has mainly been developed in marine and estuarine organisms. Yet, in the last decade, a growing number of studies have included freshwater mollusks. Wurster and Patterson (2001) investigated the suitability of two lacustrine bivalves, *Sphaerium simile* and *Dreissena polymorpha*, for recovering paleoenvironmental records on daily to weekly time scales (Figure 8). As relatively short-term temperature variability was faithfully recorded, they concluded that the seasonal pattern in the $\delta^{18}\text{O}_{(\text{shell})}$ values is a powerful tool for interpreting past climates. Variations in annual shell growth of the pearl mussel *M. margaritifera* allowed the reconstruction of summer (June–August) air temperatures for each year over the period AD 1777–1993 (Schöne et al., 2004). Helama and Nielsen (2008) demonstrated that sclerochronological reconstruction based on statistically reliable growth variability is also possible, not only through the analysis of live-collected material but also using subfossil shells. In fact, Leng et al. (1999) analyzed the ontogenetic variations in $\delta^{18}\text{O}$ values along skeletal growth lines of modern and fossil shells of the freshwater-brackish snail *M. tuberculata*. Results indicated that the isotopic composition of early Holocene rainfall in the Ethiopian Rift Valley was slightly more negative than that of the present day,

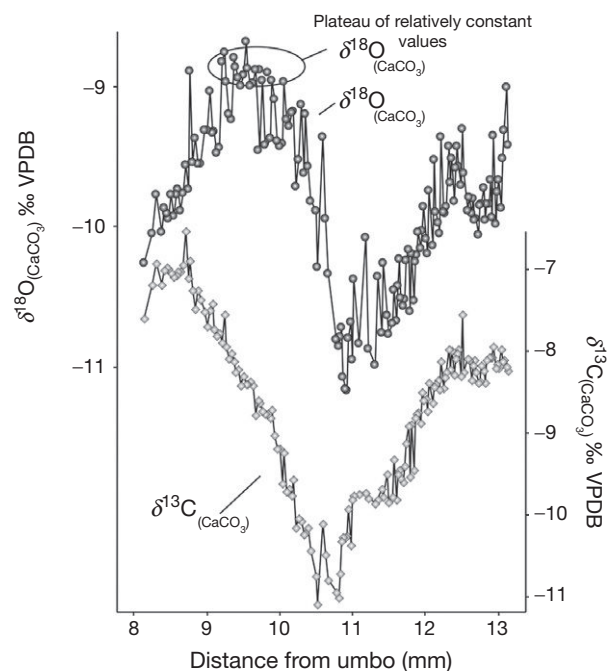


Figure 8 Short-term temperature variability recorded in the shell of the fingernail clam *Sphaerium simile*. The $\delta^{18}\text{O}$ values show a seasonal pattern. A plateau with minor variation in $\delta^{18}\text{O}$ at 9.3–10.1 mm indicates that the organism does not accrete CaCO_3 during the winter months. The increase in $\delta^{18}\text{O}$ toward most recent times (13 mm) reflects a brief warming period that took place toward the end of 1999. VPDB = Vienna Pee Dee Belemnite. Adapted from Wurster CM and Patterson WP (2001) Seasonal variation in stable oxygen and carbon isotope values recovered from modern lacustrine freshwater molluscs: Paleoclimatological implications for sub-weekly temperature records. *Journal of Paleolimnology* 26: 205–218, with permission.

suggesting a greater influence of summer rainfall derived from the Atlantic monsoon. Also by analyzing isotope variations in shells of *M. tuberculata*, Smith et al. (2004) concluded that, during the Pleistocene, the currently hyperarid Western Desert of Egypt was subject to significantly wetter conditions (high precipitation) than present.

Hydrogen isotope ratios (as δD : the ratio of 2H to 1H in a sample vs. in a standard) are another proxy used to assess paleoenvironmental conditions. Hydrogen isotopes are fractionated in natural waters by the same mechanisms as oxygen isotopes, though to a lesser extent, due to the lower relative mass difference in isotope species. The δD of the organic matrix in mollusk shells potentially offers an independent proxy for the $\delta^{18}O$ value of water and therefore a method to improve $\delta^{18}O$ paleoenvironmental interpretations in freshwater systems. Although Carroll et al. (2006) demonstrated the potential of this technique to reconstruct paleoenvironments through the analysis of freshwater mussel shells, it has not been applied at present in paleolimnological research.

Taphonomy

Since the early 1980s, extensive research conducted in contemporary marginal marine settings demonstrated a good correlation between taphonomic conditions of shelly remains accumulating in local death assemblages and the type of depositional environment (see review in Best and Kidwell, 2000). These actualistic studies fundamentally advanced our understanding of fossilization processes and enhanced our knowledge of how shell preservation varies, both among environments, as well as within single homogenous habitats.

The development of taphonomic studies in freshwater habitats has been scarce and restricted to certain regions of the globe, while other regions have remained practically unexplored. The scarce taphonomic studies carried out thus far on freshwater mollusks have focused on the quantitative fidelity, taphonomic signatures, and spatial fidelity of assemblages found in lacustrine and fluvial environments.

Quantitative Fidelity

The quantitative fidelity is the quantitative faithfulness of the record of morphs, age classes, species richness, species abundances, trophic structure, etc., to the original biological signal (Behrensmeier et al., 2000). In other words, it is the degree of similarity between the life assemblage and the potential death assemblage. A high fidelity between unionid live and dead assemblages has been reported in a variety of streams and reservoirs in east-central Ohio (Cummins, 1994). The death assemblage also preserved the rank orders (live/dead) of abundance and biomass of the preservable mollusk components in some environments, but not in those that were most environmentally disturbed. Similarly, surveys of living mollusks and dead assemblages undertaken at River Lech (Austria) also showed a high fidelity and indicated that most accumulating death assemblages were of entirely local provenance (Briggs et al., 1990). More recently, the analysis of the present-day mollusk assemblages occurring in straight and meandering sectors of the Touro Passo River (Brazil) indicated high ecological fidelity (Martello et al., 2006). The taxonomic composition and some features of the structure of communities, especially the dominant species, also reflected some ecological differences between the two main sectors sampled, such as the complexity of habitats in the meandering sector.

In a recent contribution, Erthal et al. (2011) found that fossil assemblages (Pleistocene–Holocene) from the Touro Passo Formation differed strongly from both living and death assemblages of the Touro Passo River in species composition (identities), rank abundance, and proportional abundance of species. This poor fidelity is driven by (1) taphonomic processes, which favor the long-term preservation of thick-shelled species; (2) human-induced changes in freshwater environments, such as the introduction of invasive alien species (e.g., *C. fluminea*); and, to a lesser extent, (3) the differences in the magnitude of time averaging. The results obtained by Erthal et al. (2011) have implications for conservation paleobiology. Comparing the compositional fidelity among living, death, and Quaternary fossil assemblages allows us to detect the decline of some species and/or to determine baselines for recovering their original biodiversity, contributing to programs in conservation biology.

Taphonomic Signatures and Spatial Fidelity

Shells of freshwater mollusks are expected to acquire diverse postmortem alteration because they are vulnerable to processes that operate in high-energy fluvial or lacustrine systems (Figure 9). The type of taphonomic signature that can be quantified in shells varies from author to author, depending on the taxa analyzed. Among the most commonly used are presence of articulation (only for bivalves), degree of fragmentation, shell texture (the alteration of shell surface by abrasive or dissolutive processes), encrustation (percentage of encrusting organisms on the shell), and bioerosion (percentage of bioerrosive marks in the shell). The presence of 'proteinaceous parts' (Kotzian and Simões, 2006), that is, the presence of periostracum, ligament, and operculum, is another very useful signature for freshwater shells. As abrasion and dissolution may act together in shell destruction, they are sometimes grouped in one category, namely, 'corrosion' (see Brett and Baird, 1986).

The taphonomic signature of dead shells has provided useful insights into the extirpation of freshwater mussels associated with mercury pollution in the North Fork Holston River, Virginia, USA. In fact, the degree of shell alteration proved to be a reliable predictor for distinguishing affected and unaffected sites along the river. Upstream areas that have had a continuous input of recently dead shell material exhibited younger taphonomic signatures. A site located directly below the contamination source exhibited the oldest taphonomic signature. This site had been devoid of viable populations for at least 30 years and thus was likely to be dominated by old specimens that have been altered by the prolonged action of various taphonomic factors operating in fluvial systems. The areas downstream exhibited an intermediate taphonomic signature, likely reflecting the fresh-dead shell input from mussels reintroduced in postindustrial restoration efforts (Brown et al., 2005).

Differences in some taphonomic signatures (disarticulation, fragmentation, edge rounding, proteinaceous remains, encrustation, precipitation, and corrosion) were compared between two fluvial subenvironments (straight versus meandering to braided sectors) in the Touro Passo River (Brazil) (Kotzian and Simões, 2006). The preliminary results indicated that dissolution appeared to be the most important taphonomic process affecting freshwater shells, and its magnitude appeared to be

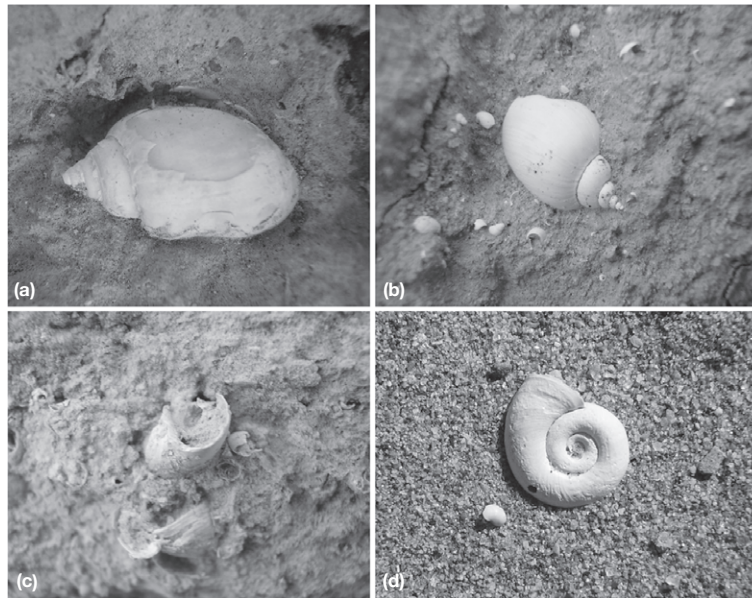


Figure 9 Shell preservation in Holocene alluvial successions from central-western Argentina. (a–c) *Chilina mendozana* showing different degrees of breakage and dissolution; (d) *Biomphalaria peregrina* showing incipient dissolution.

quantitatively similar to those recorded in marine assemblages. As the great majority of shells were preserved within the living habitat, these assemblages also preserved good spatial fidelity.

Conclusions

The high abundance, diversity, and relative size exhibited by freshwater mollusks make them a very valuable proxy for paleolimnological and taphonomic studies. The relatively large shell size (compared to that of microfossils) allows easy recognition of any alteration in shell texture by taphonomic processes, which is of relevance for outlining the origin of fossil deposits as well as for understanding the processes of fossilization in freshwater. From preliminary taphonomic studies, it can be determined whether there is high taxonomic and ecological fidelity in a freshwater habitat. Mollusk death assemblages preserved in the littoral zone of freshwater bodies appear to be good indicators of the degree of biocoenosis, which confirms the utility of freshwater fossil shells as paleolimnological indicators. Thus, it becomes very important to include some cores from near shore sites when sampling lakes (paleolimnological studies usually involve only cores from the deepest point of the lakes). In addition, the carbonate shell allows very precise paleoclimatic reconstructions (mainly temperature and related hydrological processes) through the analysis of the oxygen stable isotope composition. Micromilling of accretionary shell carbonate yields ontogenetic variation in $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values that can be used to characterize the environment and physiology of the CaCO_3 secreting organism at high resolution.

There are several questions to be addressed in order to define paleolimnological reconstructions from freshwater mollusks. For instance, the scale of time averaging (the amount of time represented within fossil assemblages) is still unknown for freshwater habitats. Cohen (1989) estimated a value >2000 years for modern death assemblages of Lake Tanganyika (Africa). There is a huge need to increase the number of

actualistic taphonomic studies in different continental settings, in order to know how patterns and processes vary at local and regional scales. It is also very important to combine contemporary ecology and paleoecology to gain a fuller understanding of changing ecological patterns at multiple timescales. At present, it can be said that our knowledge on paleoecological and taphonomic aspects of Quaternary freshwater mollusks is at an early stage of development, with preliminary results showing great potential for future research directions.

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See also: **Carbonate Stable Isotopes:** Lake Sediments.
Paleolimnology: Cladocera; Multiproxy Approaches; Overview of Paleolimnology.

References

- Behrensmeier AK, Kidwell SM, and Gastaldo RA (2000) Taphonomy and paleobiology. *Paleobiology* 26: 103–147.
- Best MMR and Kidwell SM (2000) Bivalve taphonomy in tropical mixed siliciclastic-carbonate settings. I. Environmental variation in shell condition. *Paleobiology* 26: 80–102.

- Birks HJB (2003) Quantitative paleoenvironmental reconstructions from Holocene biological data. In: Mackay AW, Battarbee R, Birks HJB, and Oldfield F (eds.) *Global Change in the Holocene*, pp. 107–123. London: Arnold Publishers.
- Bogan AE (2008) Global diversity of freshwater mussels (Mollusca, Bivalvia) in freshwater. *Hydrobiologia* 595: 139–147.
- Bonadonna FP, Leone G, and Zanchetta G (1999) Stable isotope analyses on the last 30 ka molluscan fauna from Pampa grassland, Bonaerense region, Argentina. *Palaeogeography, Palaeoclimatology, Palaeoecology* 153: 289–308.
- Bouchet P and Rocroi J-P (2005) Classification and nomenclator of gastropod families. *Malacologia* 47: 1–397.
- Brett CE and Baird GC (1986) Comparative taphonomy: A key to paleoenvironmental interpretation based on fossil preservation. *Palaios* 1: 207–227.
- Briggs DJ, Gilbertson DD, and Harris AL (1990) Molluscan taphonomy in a braided river environment and its implications for studies of Quaternary cold-stage river deposits. *Journal of Biogeography* 17: 623–637.
- Brown ME, Kowalewski M, Neves RJ, Cherry DS, and Schreiber ME (2005) Freshwater mussel shells as environmental chronicles: Geochemical and taphonomic signatures of mercury-related extirpations in the North Fork Holston River, Virginia. *Environmental Science and Technology* 39: 1455–1462.
- Bush MB (1988) The use of multivariate analysis and modern analogue sites as an aid to the interpretation of data from fossil mollusc assemblages. *Journal of Biogeography* 15: 849–861.
- Carroll M, Romanek C, and Paddock L (2006) The relationship between the hydrogen and oxygen isotopes of freshwater bivalve shells and their home streams. *Chemical Geology* 234: 211–222.
- Cohen AS (1989) The taphonomy of gastropod shell accumulations in large lakes: An example from Lake Tanganyika, Africa. *Paleobiology* 15: 26–45.
- Cummins RH (1994) Taphonomic processes in modern freshwater molluscan death assemblages: Implications for the freshwater fossil record. *Palaeogeography, Palaeoclimatology, Palaeoecology* 108: 55–73.
- De Francesco CG (2007) Las limitaciones a la identificación de especies de *Heleobia* Stimpson, 1865 (Gastropoda: Rissoidae) en el registro fósil del Cuaternario tardío y sus implicancias paleoambientales. *Ameghiniana* 44: 631–635.
- De Francesco CG and Hassan GS (2009) The significance of molluscs as paleoecological indicators of freshwater systems in central-western Argentina. *Palaeogeography, Palaeoclimatology, Palaeoecology* 274: 105–113.
- De Francesco CG, Zárate MA, and Miquel SE (2007) Late Pleistocene mollusc assemblages and inferred paleoenvironments from the Andean piedmont of Mendoza, Argentina. *Palaeogeography, Palaeoclimatology, Palaeoecology* 251: 461–469.
- Dettman DL, Reische AK, and Lohmann KC (1999) Controls on the stable isotope composition of seasonal growth bands in aragonitic fresh-water bivalves (unionidae). *Geochimica et Cosmochimica Acta* 63: 1049–1057.
- Dillon RT Jr. (2000) *The Ecology of Freshwater Molluscs*. New York: Cambridge University Press.
- Dodd JR and Stanton RJ Jr. (1990) *Paleoecology: Concepts and Applications*, 2nd edn. New Jersey: Wiley-Interscience Publication.
- Ellis AE (1940) The Mollusca of a Norfolk broad. *Journal of Conchology* 21: 224–243.
- Erthal F, Kotzian CB, and Simões MG (2011) Fidelity of molluscan assemblages from the Touro Passo Formation (Pleistocene–Holocene), southern Brazil: Taphonomy as a tool for discovering natural baselines for freshwater communities. *Palaios* 26: 433–446.
- Fan Y-X, Chen F-H, Wei G-X, et al. (2010) Potential water sources for Late Quaternary Megalake Jilantai-Hetao, China, inferred from mollusk shell $^{87}\text{Sr}/^{86}\text{Sr}$ ratios. *Journal of Paleolimnology* 43: 577–587.
- Fastovsky DE, Arthur MA, Strater NH, and Foss A (1993) Freshwater bivalves (Unionidae), disequilibrium isotopic fractionation, and temperatures. *Palaios* 8: 602–608.
- Fritz P and Poplawski S (1974) ^{18}O and ^{13}C in shells of freshwater molluscs and their environments. *Earth and Planetary Science Letters* 24: 91–98.
- Geist J, Auerswald K, and Boom A (2005) Stable carbon isotopes in freshwater mussel shells: Environmental record or marker for metabolic activity? *Geochimica et Cosmochimica Acta* 69: 3545–3554.
- Graf DL and Cummings KS (2007) Review of the systematics and global diversity of freshwater mussel species (Bivalvia: Unionoida). *Journal of Molluscan Studies* 74: 291–314.
- Gray J (1988) Evolution of the freshwater ecosystem: The fossil record. *Palaeogeography, Palaeoclimatology, Palaeoecology* 62: 1–214.
- Grimley DA, Larsen D, Kaplan SW, Yansa CH, Curry BB, and Ochse BB (2009) A multi-proxy paleoecological and paleoclimatic record within full glacial lacustrine deposits, western Tennessee, USA. *Journal of Quaternary Science* 24: 960–981.
- Hart WS, Quade J, Madsen DB, Kaufman DS, and Oviatt CG (2004) The $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of lacustrine carbonates and lake-level history of the Bonneville paleolake system. *Geological Society of America Bulletin* 116: 1107–1119.
- Helama S and Nielsen JK (2008) Construction of statistically reliable sclerochronology using subfossil shells of river pearl mussel. *Journal of Paleolimnology* 40: 247–261.
- Huber M (2010) *Compendium of Bivalves. A Full-Color Guide to 3,300 of the World's Marine Bivalves. A status on Bivalvia After 250 Years of Research*. Hackenheim: ConchBooks.
- Hudson JH, Shinn EA, Halley RB, and Lidz B (1976) Sclerochronology: A tool for interpreting past environments. *Geology* 4: 361–364.
- Jones MD, Leng MJ, Eastwood WJ, Keen DH, and Turney CSM (2002) Interpreting stable-isotope records from freshwater snail-shell carbonate: A Holocene case study from Lake Göllühisar, Turkey. *The Holocene* 12: 543–548.
- Kotzian CB and Simões MG (2006) Taphonomy of recent freshwater molluscan death assemblages, Touro Passo Stream, southern Brazil. *Revista Brasileira de Paleontologia* 9: 243–260.
- Leng MJ, Lamb AL, Lamb HF, and Telford RJ (1999) Palaeoclimatic implications of isotopic data from modern and early Holocene shells of the freshwater snail *Melanoides tuberculata*, from lakes in the Ethiopian Rift Valley. *Journal of Paleolimnology* 21: 97–106.
- Martello AR, Kotzian CB, and Simões MG (2006) Quantitative fidelity of recent freshwater mollusk assemblages from the Touro Passo River, Rio Grande do Sul, Brazil. *Iheringia Série Zoologia* 96: 453–465.
- Martínez SA and Rojas A (2004) Quaternary continental molluscs from Northern Uruguay: Distribution and paleoecology. *Quaternary International* 114: 123–128.
- McConnaughey TA and Gillikin DP (2008) Carbon isotopes in mollusk shell carbonates. *Geo-Marine Letters* 28: 287–299.
- Meijer T and Preece RC (2000) A review of the occurrence of *Corbicula* in the Pleistocene of North-West Europe. *Geologie en Mijnbouw* 79: 241–255.
- Miller BB and Bajc AF (1990) Non-marine mollusks, Vol. 8. In: Warner BG (ed.) *Methods in Quaternary Ecology. Geoscience Canada Reprint Series* 16(3): 165–175.
- Miller BB and Tevesz MJS (2001) Freshwater mollusks. In: Smol JP, Birks HJB, and Last WM (eds.) *Tracking Environmental Change Using Lake Sediments. Zoological Indicators*, vol. 4, pp. 153–171. Dordrecht: Kluwer Academic Publishers.
- Oswald WW (2006) Mollusks in a Holocene lake-sediment core from the Arctic Foothills of northern Alaska. *The Nautilus* 120: 30–33.
- Pointier JP, Delay B, Toffart JL, Lefevre M, and Romero-Alvarez R (1992) Life history traits of three morphs of *Melanoides tuberculata* (Gastropoda: Thiariidae), an invading snail in the French West Indies. *Journal of Molluscan Studies* 58: 415–423.
- Reinhardt EG, Stanley DJ, and Patterson RT (1998) Strontium isotopic paleontological method as a high-resolution paleosalinity tool for lagoonal environments. *Geology* 26: 1003–1006.
- Rosenthal Y, Katz A, and Tchernov E (1989) The reconstruction of Quaternary freshwater lakes from the chemical and isotopic composition of gastropod shells: The Dead Sea Rift, Israel. *Palaeogeography, Palaeoclimatology, Palaeoecology* 74: 241–253.
- Schöne BR, Dunca E, Mutvei H, and Norlund U (2004) A 217-year record of summer air temperature reconstructed from freshwater pearl mussels (*M. margaritifera*, Sweden). *Quaternary Science Reviews* 23: 1803–1816.
- Shanahan TM, Pigati JS, Dettman DL, and Quade J (2005) Isotopic variability in the aragonite shells of freshwater gastropods living in springs with nearly constant temperature and isotopic composition. *Geochimica et Cosmochimica Acta* 69: 3949–3966.
- Smith JR, Giegengack R, and Schwarcz HP (2004) Constraints on Pleistocene pluvial climates through stable-isotope analysis of fossil-spring tufas and associated gastropods, Kharga Oasis, Egypt. *Palaeogeography, Palaeoclimatology, Palaeoecology* 206: 157–175.
- Sparks BW (1964) Non-marine mollusca and Quaternary ecology. *The Journal of Animal Ecology* 33: 87–98.
- Strong EE, Gargominy O, Ponder WF, and Bouchet P (2008) Global diversity of gastropods (Gastropoda; Mollusca) in freshwater. *Hydrobiologia* 595: 149–166.
- Stuiver M (1968) Oxygen-18 content of atmospheric precipitation during the last 11,000 years in the Great Lakes region. *Science* 162: 994–997.
- Taylor DW (1988) Aspects of freshwater mollusc ecological biogeography. *Palaeogeography, Palaeoclimatology, Palaeoecology* 62: 511–576.
- Wurster CM and Patterson WP (2001) Seasonal variation in stable oxygen and carbon isotope values recovered from modern lacustrine freshwater molluscs: Paleoclimatological implications for sub-weekly temperature records. *Journal of Paleolimnology* 26: 205–218.
- Yang J, Karrow PF, and Mackie GL (2001) Paleoecological analysis of molluscan assemblages in two marl deposits in the Waterloo region, southwestern Ontario, Canada. *Journal of Paleolimnology* 25: 313–328.
- Zanchetta G, Bonadonna FP, and Leone G (1999) A 37-meter record of paleoclimatological events from stable isotope data on continental molluscs in Valle di Castiglione, near Rome, Italy. *Quaternary Research* 52: 293–299.

Introduction

In recent decades, paleolimnological reconstructions of long-term changes in lake water chemistry have proven to be valuable contributors to the resolution of scientific problems, many of which are at the forefront of public debates regarding environmental management. Paleolimnological techniques have been effective in reconstructing past changes in water chemistry variables because these methods rely on biota and on geochemical processes that respond in predictable ways to changes in lake water chemistry. As long-term measured data from lakes are exceedingly rare in most regions of the world, these proxy records are valuable in assessing the nature and trajectories of water chemistry changes in many freshwater systems (Smol, 2008). Paleolimnological records developed from sediment profiles can generate information about the magnitude, timing, and causes of historical changes in the water chemistry of freshwater systems. Perhaps the most valuable contribution that paleolimnological studies make is through their ability to separate long-term, naturally occurring ecosystem processes from changes related to anthropogenic activities.

Early paleolimnological studies attempted to qualitatively infer changes in water chemistry, based on biotic changes in lakes. While these studies could identify patterns of community change, and thus made important scientific contributions, they were limited by their inability to generate quantitative information. Recent refinements, particularly in the area of multivariate statistical methods and geochemical analytical techniques, have resulted in great advances in the precision, accuracy, and reliability of paleolimnological inferences drawn from the sedimentary record in modern studies. Within the last several decades, paleolimnological contributions have provided the insights needed to inform public debate about water chemistry issues with widespread impacts, most notably those surrounding eutrophication, acidification, and climate change resulting from anthropogenic activities. As paleolimnological techniques for reconstructing water chemistry continue to grow in diversity and sophistication, they will remain important contributors to scientific and public discussions regarding key societal issues.

Acidification

Background

Widespread public awareness of acid precipitation and lake acidification began in the 1970s and early 1980s when studies linked changes in lakes, particularly the loss of fish populations, to acidification in these sites and attributed these changes to industrial pollution. However, the political influence and economic clout of the industries involved in this pollution resulted in a prolonged debate about the causes of acidification and the certainty of the underlying science. A need arose for scientists to show conclusively that these lakes had been acidified by anthropogenic pollution, and

that they were neither naturally acidic nor had been acidified through natural processes. Counter arguments and denials of responsibility by industry led to additional research, and it was in these responses to industry that paleolimnology truly demonstrated the unique contributions it could make in drawing conclusions about such historical questions (reviewed in Smol, 2008).

The Effects of Acidification on Lake Ecosystems

Acidification has profound repercussions for lake ecosystems, and the changes in lake water chemistry associated with the acidification process affect aquatic organisms through both direct and indirect mechanisms. Acidification occurs due to the deposition of sulfur and nitrogen compounds from industrial activity in precipitation. In catchments receiving acidic precipitation, metals in soils (most notably Al) can be increasingly mobilized and transported into lakes, elevating concentrations in these waters. Dissolved organic carbon (DOC) concentrations in lake waters can also be greatly reduced as a result of acidification (see below). Acidification also has profound effects on lake biota and can lead to the complete alteration of lake food webs. Many of these effects were documented during whole-lake experiments in the Experimental Lakes Area in northern Ontario (Schindler et al., 1985). Acidification can cause large changes in phytoplankton communities, including reductions in the diversity of certain algal groups, and a change in the dominance of different phytoplankton and periphyton groups. Similar impacts affect aquatic plants which typically exhibit reduced diversity. In addition, acidification typically results in changes to invertebrate communities, as some invertebrate groups are unable to survive at low pH values, while those that remain may prosper due to decreased competition. Declining lake water calcium is also linked to the long-term effects of acid deposition. Paleolimnological studies have shown that the reduction of Ca concentrations below important thresholds in North American lakes has led to species replacements and indeed the disappearance of sensitive zooplankton species (Jeziorski et al., 2008). Near the top of the food web, acidification results in the disruption or cessation of reproduction among fish populations, and eventually to the extirpation of certain fish species in acidified lakes (Schindler et al., 1985).

Paleolimnological Studies of Acidification

Large-scale paleolimnological investigations were launched into the acidification of lakes in North America (Paleoecological Investigation of Recent Lake Acidification (PIRLA), Charles and Whitehead, 1986) and Europe (the Surface Water Acidification Programme (SWAP), Battarbee et al., 1990). The goals of each of these studies were to conduct regional assessments of lakes to determine the natural background pH conditions of these ecosystems and to determine whether their acidic state

resulted from natural processes, such as long-term postglacial soil acidification, soil changes due to land use or catchment vegetation changes, or if anthropogenic impacts were responsible for recent changes in the acidity of the lakes in these regions. Many of the methodologies now viewed as standard practice by paleolimnologists were developed as a result of these projects. Although these studies primarily used diatoms for reconstructions, other biological indicators were examined, and they have continued to be developed as indicators of acidity, including chrysophyte scales and cysts (Figure 1, see Smol and Cumming, 2000), chironomids and other midges, and cladocerans.

The reconstruction of acidification from microfossil groups is typically done through the application of a transfer function, which is developed using a calibration set (also called a training set) (outlined in Birks, 1998). In this approach, a series of lakes are sampled, physical and chemical measurements are taken from lake water, and modern bioindicator communities are analyzed, normally from surface sediments. Detailed datasets of these indicator groups are then compared with lake chemistry using multivariate statistics, and so-called transfer functions are developed that can generate inferences of water chemistry variables (e.g., pH, alkalinity, phosphorus) based on the analysis of microfossil communities (see Battarbee et al., 2008).

Geochemical proxies have also been used to reconstruct acidification, including measurements of total sulfur and of sulfur stable isotope changes (see Nriagu and Coker, 1983). Isotopic measurements of sulfur in these studies may additionally allow the source of sulfur pollution to be pinpointed (Nriagu and Coker, 1983). More recently, measurements of sediment copper and nickel, in addition to biological proxy indicators, have been used to track the recovery of industrially acidified lakes toward preimpact conditions (Tropea et al., 2010). Another promising geochemical technique that has recently been applied to paleolimnological problems, including acidification, is near-infrared spectroscopy (NIRS). In NIRS analyses, the properties of lake sediment samples are determined by their transmittance of radiation in the near-infrared wavelengths (700–2500 nm). The frequencies of radiation absorbed by sediments are dependent on their chemical composition, and the spectra produced by these analyses can be

used to infer the chemistry of sediments (reviewed in Korsman et al., 2001). The same training set approach used for other paleolimnological indicators (see above) must also be used to calibrate NIRS methods prior to performing quantitative reconstructions from sediment profiles. NIRS has shown great promise as a tool for developing accurate yet relatively simple and inexpensive inferences of water chemistry variables, including pH, in paleolimnological studies (Korsman et al., 2001; Rosén et al., 2011).

Changes in the pH status of lakes are not always due to anthropogenic effects. In certain cases, such as in high-altitude or high-latitude lakes, climate and related factors may exert primary control over lake acidity. Reconstructions of lake water pH can thus also be useful as indicators of climate change. Variations in pH have been shown to be linked with climate in remote, alpine lakes in Europe (Psenner and Schmidt, 1992). Diatom-based pH reconstructions have also been successful at mirroring measured climate records at time-scales from several hundred years in alpine lakes in Austria (Sommaruga-Wögrath et al., 1997) to recent decades in the Canadian High Arctic (Antoniades et al., 2005), demonstrating the potential of these techniques to utilize changes in pH to reconstruct longer term climatic change in such remote regions.

Eutrophication

Background

Nutrients, primarily phosphorus and nitrogen, play a vital role in determining the biological dynamics of a lake and are therefore of special interest to paleolimnologists and the public at large. Eutrophication, which is the process by which lakes become more productive due to increases in nutrients, can happen as a result of natural or anthropogenic causes (see Harper, 1991; Smol, 2008). Eutrophication can occur due to natural increases in the nutrient flux from several sources, as lakes receive inputs of nutrients from rivers, overland drainage in their catchments, and atmospheric deposition. These natural variations may be caused by changes in climate or changes in vegetation, among other factors. A far more common concern is cultural (i.e., human-induced) eutrophication. Cultural eutrophication is typically caused by point source pollution and may result from increases in nutrient inputs to lakes due to fertilizer runoff, untreated sewage, or industrial outputs (Harper, 1991). The deterioration of water quality in many lakes due to increased nutrients became a widely acknowledged problem in the 1960s and 1970s and continues to be among the most serious problems facing freshwater ecosystems today. Schindler (1974) convincingly showed through whole-lake manipulation experiments that phosphorus, often derived from detergents and other human sources, was responsible for this degradation.

The Consequences of Eutrophication for Lakes

The effects of eutrophication in lakes can include the development of algal blooms (including toxic forms), decreases in water clarity, increases in the growth of aquatic plants, decreases in hypolimnetic oxygen concentrations that affect fish populations, and taste and odor problems associated with

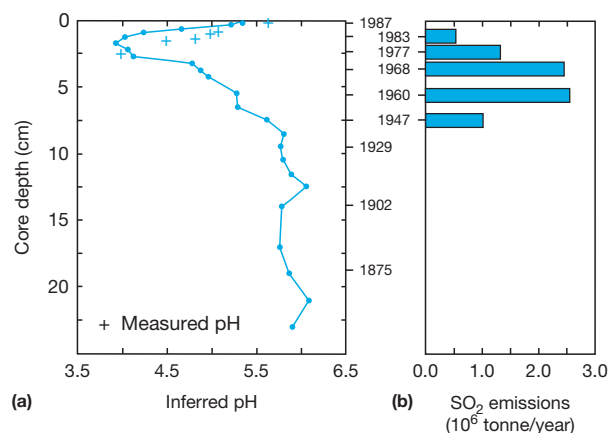


Figure 1 Stratigraphic diagram from Swan Lake, ON, Canada, showing paleolimnological reconstruction of (a) trends in pH inferred from chrysophyte scales and their agreement with measured values for lake water pH recovery and (b) how these changes reflect nearby emissions from smelting (adapted from Dixit et al., 1989).

drinking water (Figure 2, reviewed in Smol, 2008). These changes can also profoundly alter the food webs of freshwater ecosystems. Paleolimnology can play an important role in determining the timescale and magnitude of historical changes in trophic status and also in providing information on whether these changes result from natural processes or are human induced. The effects of eutrophication are pervasive in lake ecosystems; therefore, paleolimnologists have a number of options when selecting a proxy indicator for reconstructing past nutrient levels, and these can be used either independently or in complementary approaches. Because P is the most common limiting nutrient in temperate freshwater systems, it is by far the most frequent nutrient level reconstructed in paleolimnological study and will be the focus of this discussion. However, efforts to reconstruct nitrogen levels from lake sediments typically have similar strengths and limitations to those of phosphorus. Nitrogen reconstructions can be undertaken in N-limited water bodies, or in sites where nitrogen is of particular interest, using similar approaches to those described here (see Smol, 2008).

Paleolimnological Reconstructions of Lake Eutrophication

Intuitively, the simplest and most direct way to reconstruct lake P levels would be to measure the phosphorus concentrations of the sediments themselves. However, there are several



Figure 2 Algal blooms and excessive growth often result from eutrophication in freshwater sites. Top, cyanobacterial blooms on Lac St. Augustin, Québec, Canada (photograph by Martin Bouchard-Valentine). Bottom, diversion of River Cher at Chenonceaux Castle, Chenonceaux, France (photograph by R. Pienitz).

complicating factors that hinder the development of historical lake water phosphorus records from sediments. First, it is critical to identify the proportions of sedimentary phosphorus that are derived from autochthonous and allochthonous sources in order to draw any conclusions about past lake water chemistry. In addition, rates of phosphorus accumulation in lake sediments are affected by the iron content of water, and changes in redox status can affect the rates of phosphorus accumulation in sediments. Both of these processes can alter phosphorus exchange between the water column and sediments over time so that they may not consistently reflect the concentrations of lake P at the time of deposition. Also, in sites with anoxic bottom water, the net flux of phosphorus may be from the sediments to the overlying water (reviewed in Boyle, 2001). The complex nature of these depositional and diagenetic processes implies that a thorough understanding of a lake's depositional system is required, and even then, paleolimnological records of phosphorus derived from sediment measurements must be carefully interpreted.

Analysis of biogenic silica offers another geochemical approach to inferring eutrophication histories through the reconstruction past production levels of organisms with siliceous components, and so the approach is most useful for tracking past changes in diatom algal populations. Biogenic silica analysis is a measure of the concentrations of amorphous silica in sediments, which is primarily a proxy for the abundance of diatoms and other siliceous microfossils in lakes (reviewed in Conley and Schelske, 2001). Although biogenic silica analyses can only provide information regarding siliceous organisms, these inferences have been shown to be reliable reconstructions of productivity and are commonly used to infer past levels of phosphorus from sedimentary profiles.

Several different bioindicators have been successfully used to derive robust inferences of past phosphorus levels in lakes. Of these, diatoms have been by far the most commonly used microfossil group used in studies that reconstruct past lake water phosphorus (Hall and Smol, 2010). Because phosphorus is so closely tied to primary productivity in most freshwater aquatic systems, algal indicators typically respond quickly and consistently to changes in phosphorus levels, and phosphorus levels are commonly among the factors most directly linked with changing community composition. Other bioindicator communities may respond directly to changes in phosphorus, or as a result of changes in other organisms within their food webs. As such, phycological and zoological indicators have both proven effective in reconstructing lake water phosphorus. Using transfer function approaches, paleolimnologists have been successful in quantitatively reconstructing the eutrophication histories of lakes using different indicator fossils, including diatoms (see Hall and Smol, 2010), chrysophyte scales and cysts (see Zeeb and Smol, 2001), and cladocera. In addition, recent studies have shown that NIRS analyses are capable of detecting and reconstructing trophic status from measurements of bulk sediment (see discussion above, Korsman et al., 2001).

Another common effect of lake eutrophication is the development of anoxic conditions in the hypolimnion of lakes. Hypolimnetic anoxia develops from increases in the rate of bacterial decomposition of organic matter due to the higher rates of productivity and organic sedimentation that result

from eutrophic conditions. The higher consumption of oxygen near the sediment–water interface can lead to anoxia in lake bottom waters. Chironomids are particularly sensitive to this change in lake water oxygen concentration, as a large part of the life cycles of some taxa are spent in benthic habitats. Models have been developed to infer hypolimnetic oxygen status or total phosphorus from chironomid remains (e.g., Quinlan et al., 1998), and these models have been applied to sediment cores in historical reconstructions of hypolimnetic oxygen and eutrophication (see reviews in Porinchu and MacDonald, 2003).

Fossil pigments in sediment cores have also been used to reconstruct the development of eutrophication-induced anoxia. Specific marker pigments such as okenone and isorenieratene permit reconstructions of anoxic conditions because the sulfur bacteria that produce them are obligate anaerobes. The presence of such pigments has been used in paleolimnological studies to infer the development of anoxia in a lake subject to eutrophication from sewage inputs (Antoniades et al., 2011).

Salinity

Background

Records of lake water salinity and/or specific conductivity changes are of interest to paleolimnologists because these variables are typically controlled by climate. As a result, the ultimate goal of paleolimnological studies concerned with salinity is often the indirect reconstruction of climatic change (Smol and Cumming, 2000). The majority of paleosalinity studies have focused on closed-basin lakes (i.e., those without an outlet), since the cause of salinity changes in such lakes is essentially limited to the difference between precipitation and evaporation, allowing for the reconstruction of drought histories (Figure 3). These drought records are of particular interest in arid regions such as the continental interiors of North America or Australia, where frequent and/or severe droughts can have great economic consequences, as well as in the Antarctic. Salinity changes are recorded in the paleolimnological record by their effects on the community structure of bioindicators, as well as by geochemical signals left in the structure of microfossil remains.

Studies of Paleosalinity

Diatoms have also been the most commonly used paleoindicator group for inferring past salinity levels. Their species distributions have been shown to be strongly correlated with salinity in many lakes. They have therefore been successfully and extensively used as indicators of paleosalinity in numerous studies. Another algal group that has been used to reconstruct changes in salinity is fossil chrysophytes. Although chrysophyte scales can be used in such approaches, preservation problems exist in high-salinity waters. However, chrysophyte cysts can be used effectively to infer paleosalinity and are not prone to the same preservation problems (reviewed in Zeeb and Smol, 2001). Salinity changes are also reflected in chironomid communities, and transfer functions and salinity reconstructions have been completed from fossil chironomids in diverse regions including Europe, Canada, and Africa.



Figure 3 Dramatic changes in the chemistry of closed-basin lakes are preserved by multiple indicators in the paleolimnological record. Top, lake PC36, British Columbia, Canada in spring 1991 (photograph by Susan Wilson). Bottom, lake PC36 in fall 1991, showing the effects of evaporative concentration (photograph by B. Cumming).

Freshwater ostracods have also proven to be reliable indicators of paleosalinity. Past salinity levels can be derived from ostracods using two approaches (see Holmes, 2001 for a review). Ostracod community composition is sensitive to the chemical composition of saline waters as well as to any change in salinity levels. In addition, ostracod shells are secreted over very short time periods, and an analysis of trace element geochemistry, most importantly concentrations and ratios of Sr and Mg, can be used to infer salinity levels at the time of secretion (Chivas et al., 1985). These geochemical approaches can also be applied to fossil mollusks.

Dissolved Organic Carbon

Background

DOC is an important component of freshwaters, and it plays a central role in many limnological processes (Williamson et al., 1999). DOC is often largely derived from terrestrial vegetation (so-called allochthonous sources) and deposited from lake catchments either by streams or by overland flow (Figure 4). In some environments, however, *in situ* algal and macrophyte production can give rise to large quantities of DOC (autochthonous sources). As such, on local scales, the degree of vegetation and the moisture regimes of catchments control the supply, delivery, and thus lake water DOC concentration. On broader scales, studies suggest that lake water DOC concentration in

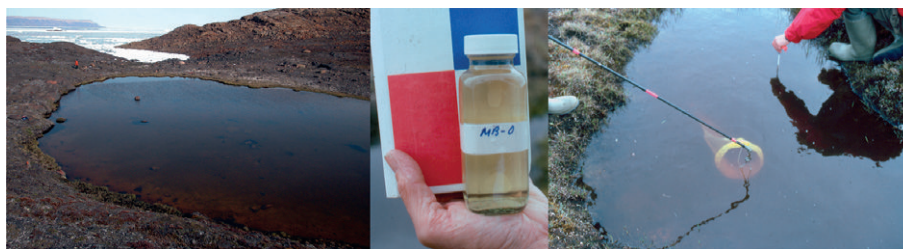


Figure 4 DOC attenuates UV radiation and protects organisms from its harmful effects. Left, a high DOC site, Skraeling Island, Canada (photograph by Bronwyn Keatley). Middle and right, pictures show the effects of high DOC on water color, Mould Bay, Prince Patrick Island, Canada (photographs by J. Smol).

Europe and North America increased during the late twentieth and early twenty-first centuries (Evans et al., 2005). The cause of this increase is the subject of debate, with suggested causes including climate change, deforestation, and changes in atmospheric deposition chemistry (Monteith et al., 2007; Roulet and Moore, 2006; Weyhenmeyer and Karlsson, 2009).

The Effects of Lake Water DOC Concentrations

Exposure to ultraviolet (UV) radiation (UV-A: 320–400 nm, UV-B: 280–320 nm), and especially to UV-B radiation, can produce harmful effects on lake biota (Rautio and Tartarotti, 2010; Vincent and Roy, 1993). However, DOC acts as a natural sunscreen and is a strong attenuator of UV radiation (UVR) in freshwaters. It most strongly absorbs UVR in the wavelengths that are most harmful to aquatic organisms (reviewed in Vincent and Pienitz, 1996). Although concerns exist about increased UVR exposure due to thinning of the stratospheric ozone layer, Pienitz and Vincent (2000) showed that the effects of changing lake water DOC levels on underwater UVR exposure were over 100 times greater than those caused by ozone thinning. Lakes with low levels of DOC (i.e., $<4 \text{ mg l}^{-1}$), most commonly located in sparsely vegetated regions such as at high altitudes or latitudes, are particularly transparent to UV radiation. They are thereby vulnerable to biotic damage as a result of any increase in incoming UV radiation (e.g., Laurion et al., 1997).

Paleolimnology and Reconstruction of DOC Histories

An understanding of past DOC dynamics is important to paleolimnologists (and limnologists) for two major reasons. First, because of its critical role in protecting lake biota from UV radiation, DOC histories can help to determine the vulnerability of a lake due to changes in incident UV radiation. Second, because DOC is affected by other variables of interest to paleolimnologists, including changes in acidity, vegetation, and hydrological regime, DOC changes can also be used as proxies for other influences, such as climate and pollution. Under predicted future climate change scenarios, DOC in lakes is expected to decrease in regions subject to reductions in precipitation and greater drought frequency but could also increase in the lakes of certain ecotonal regions as a result of shifting vegetation zones, such as the northward movement of treeline in subarctic regions.

There are two main approaches that paleolimnologists have successfully used to reconstruct DOC levels. In certain regions, DOC is among the strongest controlling factors influencing

diatom community compositions. The majority of such DOC reconstructions to date have focused on reconstructing the effects of treeline shifts in subarctic Canada and Russia (e.g., Pienitz and Vincent, 2000). Because of the relationship between DOC and pH, diatom-inferred DOC records have also been used as a proxy for acidification in studies of industrial pollution (Dixit et al., 2001). More recently, visible NIRS has been used to develop models to infer DOC concentrations directly from sediments (Rouillard et al., 2011).

The second major approach to reconstructing changes in historical DOC levels has been through analysis of sediments to identify and quantify fossil pigments. In such studies, measurements are taken of the abundances in sediments of various pigments produced by algae and cyanobacteria to protect themselves from UV exposure. Periods of higher exposure to UV radiation cause increases in the production of these UV-shielding pigments, and thus their concentrations in lake sediments are increased. Inferences of historical DOC levels can then be made on the basis of variations of key pigments in the sedimentary record (see also Leavitt and Hodgson, 2001).

Trace Metals

Background

Trace metals, such as copper, selenium, and zinc, are micronutrients required by biota. However, as concentrations increase, these metals can become toxic, while other heavy metals such as cadmium, mercury, lead, chromium, and thallium are toxic even at low concentrations. As is the case of almost all water chemistry variables, metal concentrations in lake water are at least partially derived from their catchments, and their natural (i.e., background) concentrations are dependent on local geology and hydrological regimes, among other factors. Inputs in lakes of trace metals resulting from anthropogenic activities can be derived from point source pollution, although deposition through the atmosphere from diffuse sources is a more common source (reviewed in Smol, 2008).

Paleolimnological Studies of Metal Concentrations

Paleolimnologists have relied primarily on geochemical analysis of bulk sediments to reconstruct changes in past concentrations of metals. Although metal concentrations in the sedimentary record may be complicated by postdepositional processes, paleolimnological studies can, at the very least, infer

trends in metal concentrations (reviewed in Boyle, 2001; Smol, 2008). The effect of metal inputs may vary between lakes, as the natural water chemistry for a site can determine the biological availability of metals and the ability of lake sediments to act as sinks for metals. Depending on the focus of scientific investigation, diverse metals have been the subject of paleolimnological study, including Pb, Hg, Cu, S, Ni, and Zn. Changes in these metals in the sedimentary record have provided evidence of the effects of local industrial or mining activity, combustion of fossil fuels, and long-range deposition (Smol, 2008). Isotopic methods using lake sediments can also be used to differentiate between natural metal deposition and anthropogenic sources and in certain cases can pinpoint the sources of pollution. Although geochemical methods have been the primary means of reconstructing metal concentrations, cladocerans, sedimentary pigments, and diatoms (through both compositional change and an increase in deformities) have also been used to infer trends in past metal concentrations (Figure 5, reviewed in Smol, 2008).

Persistent Organic Pollutants

Background

Persistent organic pollutants (POPs), including polycyclic aromatic hydrocarbons (PAHs), are a group of chemicals that share certain characteristics: they persist in the environment (from weeks to decades), they are toxic (to various levels) to organisms, and they are biomagnified through successive levels of food webs (reviewed in Blais and Muir, 2001). Many different substances with diverse uses fall within this definition, including such well-known substances as dichlorodiphenyltrichloroethane (DDT), polychlorinated biphenyls (PCBs), PAHs, and dioxins. A large number of these chemicals are human-made. Due to their persistence in the environment, POPs may be transported over long distances and therefore can be found in surprisingly high concentrations in remote lakes, biota, and sediments that are far removed from POP use (see Muir and Rose, 2004). Although POPs are often found in lakes in concentrations of parts per billion or even parts per trillion, these substances are of great concern to

ecologists and the public because they can have damaging effects on organisms at extremely low concentrations. The situation is complicated further as many POPs may be biomagnified up the food chain. Concern over the health effects of POPs has resulted in bans on the use of many of these substances. However, given their environmental persistence, paleolimnological studies can provide valuable insights into the fates of POPs subsequent to their deposition in freshwater ecosystems.

Paleolimnological Studies of POPs

Due to their often low concentrations, there are few reliable direct records of POPs in natural waters. In recent decades, improvements in analytical techniques have allowed for more reliable measurements of POPs to be made (see Blais and Muir, 2001). However, the general lack of historical measurements of POP concentrations makes information derived from paleolimnological studies particularly valuable. As with all paleolimnological techniques, an understanding of the processes involved in POP deposition is required for accurate interpretation of the sedimentary record. Although concerns also exist about diagenetic processes, indications are that the concentrations of POPs in sediments often provide important archives of past contaminant loads, although diffusion and other postdepositional processes may occur within sediments (Muir and Rose, 2004; Smol, 2008). POPs have also been observed to be removed from the water column more quickly in eutrophic lakes due to increases in both sedimentation and scavenging by phytoplankton (reviewed in Blais and Muir, 2001).

Studies using lake sediments have provided insights into the transport and fate of POPs. Such studies have demonstrated the pervasive nature of these compounds in global ecosystems, clarified the processes involved in the cycling and transportation of these substances to the polar regions (known as global distillation), and produced information about the sources of POP in lake sediments. Recent research has concentrated on wildlife populations, particularly seabirds, as 'biovectors' that can deposit contaminants in high concentrations to aquatic ecosystems (Figure 6; Blais et al., 2007). For example, it was demonstrated that northern fulmars caused greatly elevated PCB concentrations in Arctic lakes and surrounding food webs (Choy et al., 2010), and sediment core studies were used to reconstruct the historical linkages between seabird populations and contaminant deposition (Michelutti et al., 2009).

Summary

Changes in lake chemistry have profound effects on the biotic communities and environmental processes that occur within lakes. As a result, changes in lake chemistry are often preserved in the sedimentary record of these lakes. Paleolimnological proxy records, constructed using both biological and geochemical indicators, can provide important information about the histories of lakes and changes in their chemistry that could not be derived otherwise from other sources.

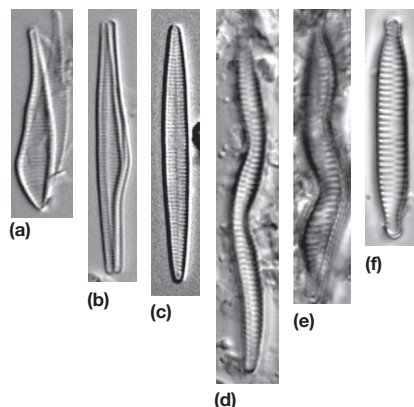


Figure 5 Teratogenic (deformed) diatoms (a, b, d, e) are commonly found in lakes impacted by metal pollution. Normal specimens are shown for comparison (c, f).

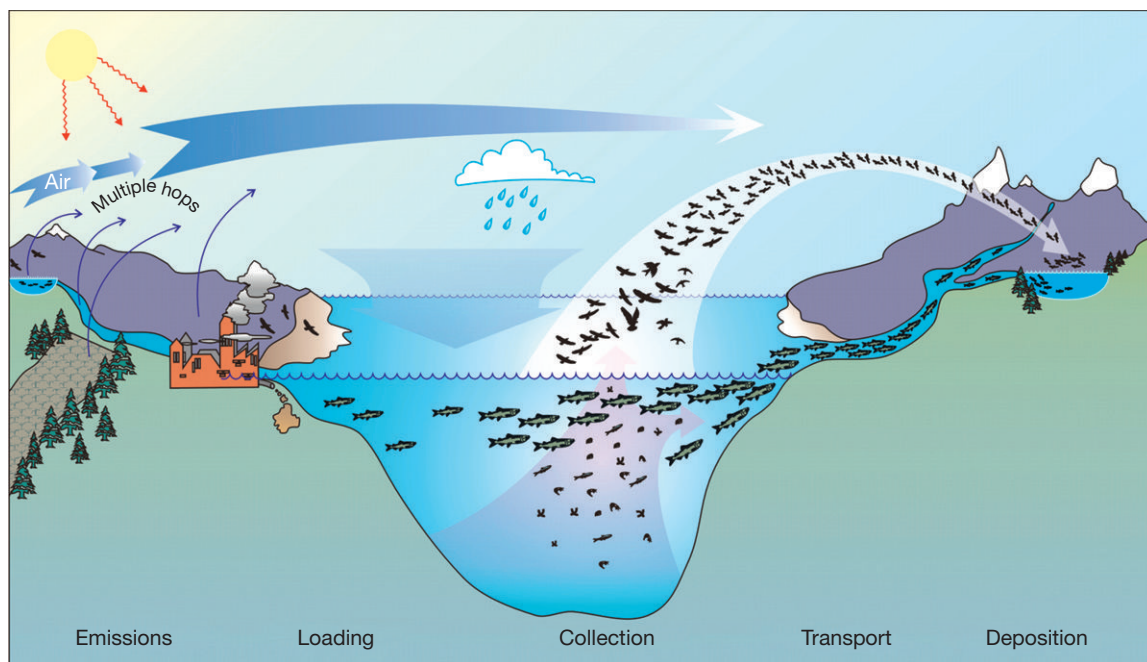


Figure 6 Diagram showing the flow of contaminants such as POPs and the role of wildlife as 'biovectors.' Chemicals are released from industrial or agricultural applications and pass through the atmosphere or hydrosphere where they are incorporated into food webs and biomagnified in animals. These animals then congregate in areas such as hatcheries and deposit contaminants in high concentrations. Reprinted with permission from Blais JM, MacDonald RW, MacKay D, Webster E, Harvey C, and Smol JP (2007) Biologically mediated transport of contaminants to aquatic systems. *Environmental Science and Technology* 41: 1075–1084. Copyright 2007 American Chemical Society.

See also: [Diatom Introduction](#). [Chironomid Records: Africa](#); [Chironomid Overview](#); [Late Pleistocene of Europe](#); [Postglacial Europe](#); [Postglacial Southern Hemisphere](#). [Diatom Methods: Salinity and Climate Reconstructions from Continental Lakes](#). [Paleolimnology: Cladocera](#); [Freshwater Mollusks](#); [Multiproxy Approaches](#); [Pigment Studies](#); [Visible and Infrared Spectroscopical Applications](#).

References

- Antoniades D, Douglas MSV, and Smol JP (2005) Quantitative estimates of recent environmental changes in the Canadian High Arctic inferred from diatoms in lake and pond sediments. *Journal of Paleolimnology* 33: 349–360.
- Antoniades D, Michelutti N, Quinlan R, et al. (2011) Cultural eutrophication, anoxia, and ecosystem recovery in Merrett Lake, High Arctic Canada. *Limnology and Oceanography* 56: 639–650.
- Battarbee RW, Mason J, Renberg I, and Talling JF (1990) *Palaeolimnology and Lake Acidification*. London: Royal Society of London.
- Battarbee RW, Monteith DT, Juggins S, et al. (2008) Assessing the accuracy of diatom-based transfer functions in defining reference pH conditions for acidified lakes in the United Kingdom. *The Holocene* 18: 57–67.
- Birks HJB (1998) Numerical tools in palaeolimnology – Progress, potentialities, and problems. *Journal of Paleolimnology* 20: 307–332.
- Blais JM, MacDonald RW, MacKay D, Webster E, Harvey C, and Smol JP (2007) Biologically mediated transport of contaminants to aquatic systems. *Environmental Science and Technology* 41: 1075–1084.
- Blais JM and Muir DCG (2001) Paleolimnological methods and applications for persistent organic pollutants. In: Smol JP, Birks HJB, and Last WM (eds.) *Tracking Environmental Change Using Lake Sediments. Physical and Geochemical Methods*, vol. 2, pp. 271–298. Dordrecht: Kluwer.
- Boyle JF (2001) Inorganic geochemical methods in paleolimnology. In: Smol JP, Birks HJB, and Last WM (eds.) *Tracking Environmental Change Using Lake Sediments. Physical and Geochemical Methods*, vol. 2, pp. 83–141. Dordrecht: Kluwer.
- Charles DF and Whitehead DR (1986) The PIRLA project: Paleoeological investigation of recent lake acidification. *Hydrobiologia* 143: 13–20.
- Chivas AR, DeDecker P, and Shelley JMG (1985) Strontium content of ostracods indicates lacustrine paleosalinity. *Nature* 316: 251–253.
- Choy ES, Kimpe LE, Mallory ML, Smol JP, and Blais JM (2010) Contamination of an arctic terrestrial food web with marine-derived persistent organic pollutants transported by breeding seabirds. *Environmental Pollution* 158: 3431–3438.
- Conley DJ and Schelske CL (2001) Biogenic silica. In: Smol JP, Birks HJB, and Last WM (eds.) *Tracking Environmental Change Using Lake Sediments. Terrestrial, Algal, and Siliceous Indicators*, vol. 3, pp. 281–293. Dordrecht: Kluwer.
- Dixit SS, Dixit AS, and Smol JP (1989) Lake acidification recovery can be monitored using Chrysophycean microfossils. *Canadian Journal of Fisheries and Aquatic Sciences* 46: 1309–1312.
- Dixit SS, Keller W, Dixit AS, and Smol JP (2001) Diatom-inferred dissolved organic carbon reconstructions provide assessments of past UV-B penetration in Canadian shield lakes. *Canadian Journal of Fisheries and Aquatic Sciences* 58: 543–550.
- Evans CD, Monteith DT, and Cooper DM (2005) Long-term increases in surface water dissolved organic carbon: Observations, possible causes and environmental impacts. *Environmental Pollution* 137: 55–71.
- Hall RI and Smol JP (2010) Diatoms as indicators of lake eutrophication. In: Smol JP and Stoermer EF (eds.) *The Diatoms: Applications for Environmental and Earth Sciences*, 2nd edn., pp. 122–151. Cambridge: Cambridge University Press.
- Harper DM (1991) *Eutrophication of Fresh Waters: Principles, Problems and Restoration*. Dordrecht: Kluwer.
- Holmes JA (2001) Ostracoda. In: Smol JP, Birks HJB, and Last WM (eds.) *Tracking Environmental Change Using Lake Sediments. Zoological Indicators*, vol. 4, pp. 125–151. Dordrecht: Kluwer.
- Jeziorski A, Yan ND, Paterson AM, et al. (2008) The widespread threat of calcium decline in fresh waters. *Science* 322: 1374–1377.
- Korsman T, Dabakk E, Nilsson MB, and Renberg I (2001) Near-infrared spectrometry (NIRS) in paleolimnology. In: Smol JP, Birks HJB, and Last WM (eds.) *Tracking Environmental Change Using Lake Sediments. Physical and Geochemical Methods*, vol. 2, pp. 299–317. Dordrecht: Kluwer.
- Laurion I, Vincent WF, and Lean DRS (1997) Underwater ultraviolet radiation: Development of spectral models for northern high latitude lakes. *Photochemistry and Photobiology* 65: 107–114.

- Leavitt PR and Hodgson DA (2001) Sedimentary pigments. In: Smol JP, Birks HJB, and Last WM (eds.) *Tracking Environmental Change Using Lake Sediments. Terrestrial, Algal, and Siliceous Indicators*, vol. 3, pp. 295–325. Dordrecht: Kluwer.
- Michelutti N, Liu H, Smol JP, et al. (2009) Accelerated delivery of polychlorinated biphenyls (PCBs) in recent sediments near a large seabird colony in Arctic Canada. *Environmental Pollution* 157: 2769–2775.
- Monteith DT, Stoddard JL, Evans CD, et al. (2007) Dissolved organic carbon trends resulting from changes in atmospheric deposition chemistry. *Nature* 450: 537–540.
- Muir DCG and Rose NL (2004) Lake sediments as records of Arctic and Antarctic pollution. In: Pienitz R, Douglas MSV, and Smol JP (eds.) *Long-term Environmental Change in Arctic and Antarctic Lakes*, pp. 209–239. Dordrecht: Kluwer.
- Nriagu JO and Coker RD (1983) Sulfur in sediments chronicles past changes in lake acidification. *Nature* 303: 692–694.
- Pienitz R and Vincent WF (2000) Effect of climate change relative to ozone depletion on UV exposure in subarctic lakes. *Nature* 404: 484–487.
- Porinchu DF and MacDonald GM (2003) The use and application of freshwater midges (Chironomidae: Insecta: Diptera) in geographical research. *Progress in Physical Geography* 27: 378–422.
- Psenner R and Schmidt R (1992) Climate-driven pH control of remote alpine lakes and effects of acid deposition. *Nature* 356: 781–783.
- Quinlan R, Smol JP, and Hall RI (1998) Quantitative inferences of past hypolimnetic anoxia in south-central Ontario lakes using fossil midges (Diptera: Chironomidae). *Canadian Journal of Fisheries and Aquatic Sciences* 55: 587–596.
- Rautio M and Tartarotti B (2010) UV radiation and freshwater zooplankton: Damage, protection and recovery. *Freshwater Reviews* 3: 105–131.
- Rosén P, Bindler R, Korsman T, Mighall T, and Bishop K (2011) The complementary power of pH and lake water organic carbon reconstructions for discerning the influences on surface waters across decadal to millennial time scales. *Biogeosciences Discussions* 8: 2349–2466.
- Rouillard A, Rosén P, Douglas MSV, Pienitz R, and Smol JP (2011) A model for inferring dissolved organic carbon (DOC) in lakewater from visible-near-infrared spectroscopy (VNIRS) measures in lake sediment. *Journal of Paleolimnology* 46: 187–202.
- Roulet N and Moore TR (2006) Environmental chemistry: Browning the waters. *Nature* 444: 283–284.
- Schindler DW (1974) Eutrophication and recovery in experimental lakes: Implications for lake management. *Science* 184: 897–899.
- Schindler DW, Mills KH, Malley DF, et al. (1985) Long-term ecosystem stress: The effects of years of experimental acidification of a small lake. *Science* 228: 1395–1401.
- Smol JP (2008) *Pollution of Lakes and Rivers: A Paleoenvironmental Perspective*, 2nd edn. Oxford: Blackwell.
- Smol JP and Cumming BF (2000) Tracking long-term changes in climate using algal indicators in lake sediments. *Journal of Phycology* 36: 986–1011.
- Sommaruga-Wögrath S, Koinig KA, Schmidt R, Sommaruga R, Tessadri R, and Psenner R (1997) Temperature effects on the acidity of remote alpine lakes. *Nature* 387: 64–67.
- Tropea AE, Paterson AM, Keller W, and Smol JP (2010) Sudbury sediments revisited: Evaluating limnological recovery in a multiple-stressor environment. *Water, Air, and Soil Pollution* 210: 317–333.
- Vincent WF and Pienitz R (1996) Sensitivity of high-latitude freshwater ecosystems to global change: Temperature and solar ultraviolet radiation. *Geoscience Canada* 23: 231–236.
- Vincent WF and Roy S (1993) Solar ultraviolet-B radiation and aquatic primary production: Damage, protection, and recovery. *Environmental Reviews* 1: 1–12.
- Weyhenmeyer GA and Karlsson J (2009) Nonlinear response of dissolved organic carbon concentrations in boreal lakes to increasing temperatures. *Limnology and Oceanography* 54: 2513–2519.
- Williamson CE, Morris DP, Pace ML, and Olson AG (1999) Dissolved organic carbon and nutrients as regulators of lake ecosystems: Resurrection of a more integrated paradigm. *Limnology and Oceanography* 44: 795–803.
- Zeeb BA and Smol JP (2001) Chrysophyte scales and cysts. In: Smol JP, Birks HJB, and Last WM (eds.) *Tracking Environmental Change Using Lake Sediments Terrestrial, Algal, and Siliceous Indicators*, vol. 3, pp. 203–223. Dordrecht: Kluwer.

Physical Properties of Lake Sediments

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Physical Properties of Lacustrine Sediments

As most Quaternary scientists use lacustrine sediments as a paleoenvironmental proxy, the focus of this article is on those physical properties of the lacustrine sedimentary record diagnostic in a proxy context. Most attention is placed here upon the sediments, with additional reference to the methods of assessing physical properties. We have also sought to catalog the data available from the literature on the physical properties of lakes for reference and comparative purposes. Despite the vast number of measurements performed to assess the physical properties of lacustrine sediment, the majority of those data are not publicly accessible; results are reported by way of summary statistics, descriptive text, or figures.

Sedimentary processes in lakes are determined by a process network within, and beyond, the lake. Since lakes can preserve near-continuous records of change, both within the lake, and the contributing watershed (Last and Smol, 2001), frequent use has been made of the physical properties of lacustrine deposits as paleoenvironmental records; such use has included varve thickness (e.g., Nederbragt and Thunrow, 2001), grain size (e.g., Sundborg and Calles, 2001), grain density (e.g., Kashiwaya et al., 1999), and mineralogy (e.g., Solotchina et al., 2004). Indeed, reliance on physical properties of sedimentary sequences as paleoenvironmental proxy records is a well-established theme in the literature; some of these records span 10^5 years (e.g., Kashiwaya et al., 1999; Vogel et al., 2010). All lacustrine sediments exhibit physical properties that are a composite product of processes occurring at a range of temporal and spatial scales. In recognition of this complexity, Hodder et al. (2007) defined the 'process network' as the system of temporally and spatially connected processes involved in the transfer of an environmental signal that is expressed in physical (or other) properties of lacustrine sediment (Figure 1). Although our emphasis is placed on physical properties, links with the geochemical and biological properties of sediments are similarly related to the process network but are omitted from Figure 1.

Although the process network can be considered conceptually identical for all lakes, the network at any one lake must be operationally unique, as no two lakes experience an identical suite of variables. Even for a single lake, the network is likely to be temporally dynamic as the various component systems evolve. That each process network is operationally unique, and temporally dynamic, is a mixed blessing in the use of lacustrine sediments and their physical properties as paleoenvironmental proxies. On the one hand, sensitivity to environmental processes is requisite, as it is vital that lacustrine sediment preserves a composite record of network interactions to enable their use as multivariate environmental proxies. On the other hand, should the causative pathways evolve over time, it is possible that unobserved processes have contributed

in unknown ways to the sedimentary record, a challenge elaborated upon elsewhere in this volume.

Basin-Wide Properties

Whole-lake patterns of lacustrine sediment accumulation are controlled by factors including the location and nature of inflow(s), Coriolis effect, wind, basin geometry, bathymetry, basin trap efficiency, and spatiotemporal evolution of these factors. The allometric properties of the resultant lacustrine sedimentary packages are fundamental physical properties and influence any paleoenvironmental interpretation derived from physical parameters.

Remote sensing techniques offer researchers the ability to assess the nature of both (1) whole-lake and (2) long-term sedimentary records, unencumbered by the practical limitations of deep and/or spatially intensive coring. In the absence of remote sensing techniques, assessment of the quantity and locations of core sampling sites is often determined by a mix of operational considerations (equipment, water depth, remoteness), cost and complexity of the intended analysis (Gilbert, 2003c). Workers seeking to core lakes can, therefore, only benefit from additional knowledge of site specifics: bathymetry, sediment distribution, sediment depth (cf. Glew et al., 2001). Remote sensing techniques have been successfully used to assess lake coring programs (e.g., Gilbert, 2003c), particularly for selecting core sites that are as historically representative as possible and/or show evidence of continuous, regular sedimentation. Remote sensing commonly involves the acoustic reflection technique, which relies on the generation of normal-incidence acoustic waves that penetrate the sediment package, paired with detection of any reflected waves generated at the sediment–water interface and at boundaries within the sediment package. Repeated generation of acoustic pulses from a transmitter along a longitudinal profile, and detection of reflected returns, can be used to generate an acoustic profile of the sub-bottom in the form of a continuous profile (Figure 2). Acoustic profiles are not geologic cross sections, as their appearance depends on the acoustic impedance of sediment packages and the variation of sound velocity in sediment. Reflection of acoustic waves from the lake floor, and from within bottom sediments, is determined primarily by thresholds in acoustic impedance. Acoustic impedance for normal-incidence waves is a function of the density of the media and therefore related to sediment–water content, porosity, organic content, and grain-size distribution. Impedance discontinuities of sufficient magnitude, encountered as sound passes through the sediment, result in reflection of a portion of wave energy, termed a reflector. Because acoustic impedance is a function of the physical characteristics of the sediment, acoustic reflectors permit visualization of the nature of the deposit and, therefore, inference of depositional processes.

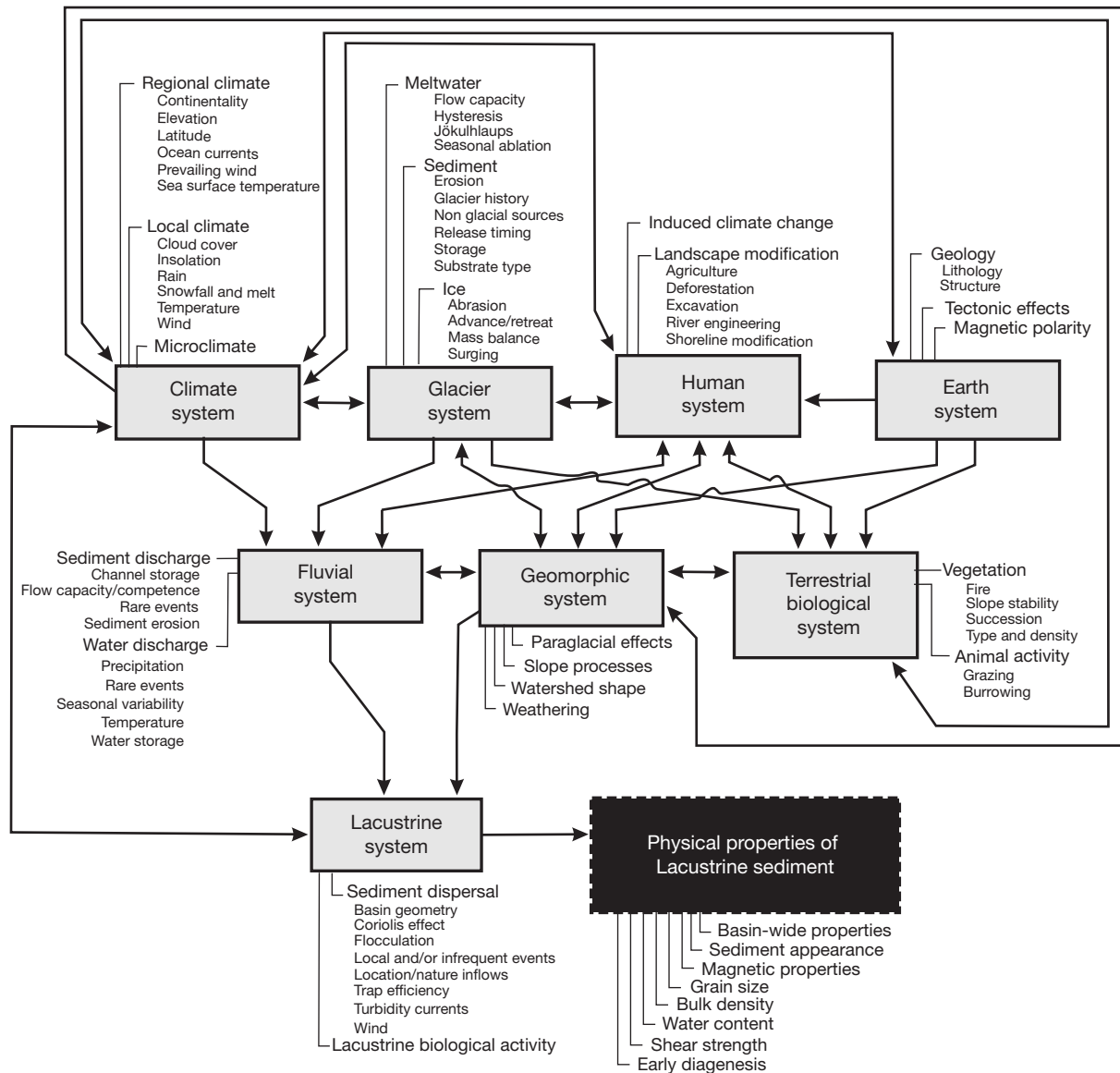


Figure 1 Conceptual model of the process network contributing to the physical properties of lacustrine sediments. Modified from: Gilbert R (2003) Lacustrine sedimentation. In: Middleton GE (ed.), *Encyclopedia of Sedimentology and Sedimentary Rocks*, pp. 404–408. Dordrecht: Kluwer; and Hodder KR, Gilbert R, and Desloges J (2007) Glaciolacustrine varved sediment as an alpine hydroclimatic proxy. *Journal of Paleolimnology* 38: 365–394. Reproduced with kind permission from Springer Science+Business Media.

An important advantage of this procedure is that it provides an overview of the stratigraphy of the entire lacustrine system in a way that single or multiple cores cannot.

Remote sensing techniques have helped demonstrate that bottom sediment packages can show considerable variation in thickness (Figure 2). Focusing (Likens and Davis, 1975) of sediment in certain regions of lake basins has been attributed to the action of turbidity currents (Francus et al., 2008; Ludlam et al., 1996), slope instability (Pickrill and Irwin, 1983), the depth of overlying water (Blais and Kalff, 1995), and modification of sediment dispersal patterns by wind (Girardclos et al., 2003). The implications of sediment focusing for recovery of records of environmental change are clear: portions of the lake basin may undergo enhanced sediment accumulation, or

punctuated accumulation, and may therefore record different paleoenvironmental signals.

Lake trap efficiency describes the portion of sediment input trapped in the lake basin and therefore also the corresponding portion lost to outflow from the lake. Greater residence time of water in the basin allows for greater sediment trapping. The ratio relating lake volume to inflow (Brune, 1953) is an important property of the lake and associated sedimentary record. According to this ratio, trap efficiency will vary somewhat with inflow: the lower the volume-to-inflow ratio, the greater the susceptibility of trap efficiency to inflow variability. Thus, particularly in small basins, changes in inflow can modify the magnitude and grain-size distribution of the sediment load received at the lake floor. Where stratification restricts

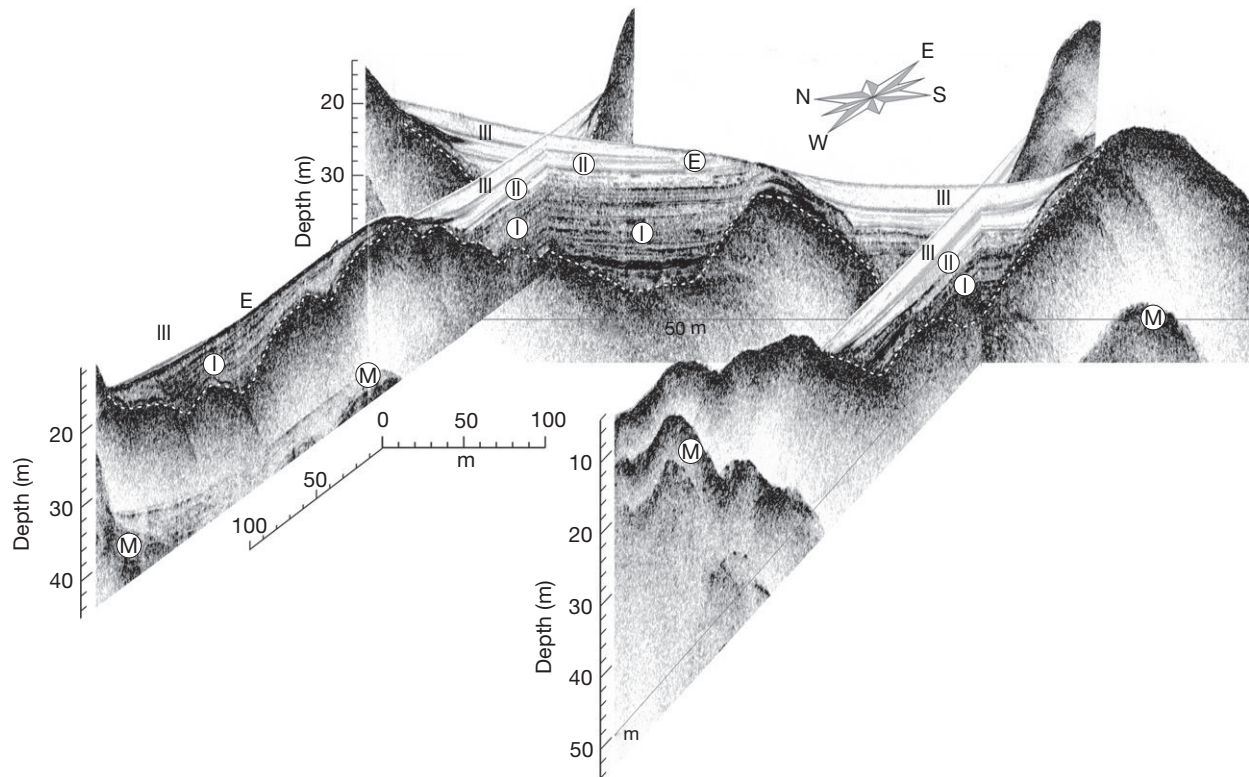


Figure 2 Fence diagram of a sub-bottom acoustic record from Devil Lake, a morphologically complex basin on the Canadian Shield (45° N 76° W) showing considerable spatial variation in sediment thickness. Depths in sediment assume sound velocity as that in water: 1455 m s^{-1} . Dotted line traces bedrock surface. Acoustic facies shown are (I) late Pleistocene glaciolacustrine sediment, (II) postglacial sediment deposited during lowering of glacial Lake Iroquois, and (III) Holocene gyttja. E refers to erosion surfaces in the record; M refers to multiple reflectors from hard surfaces. Reproduced with kind permission from Springer Science+Business Media. Gilbert R (2003) Spatially irregular sedimentation in a small, morphologically complex lake: Implications for paleoenvironmental studies. *Journal of Paleolimnology* 29: 209–220. (License Number 2791480399320).

inflow mixing, trap efficiency also varies with the nature of mixing in addition to inflow rate. The paleoenvironmental consequences of varying trap efficiency are somewhat less explored but Axellson (1967) estimated that an order of magnitude increase in flow rate results in a two-thirds reduction in silt trapping (Figure 3). Under a set of simplifying assumptions, trap efficiency has been used to infer past inflow volumes using the difference in primary grain size between sampling points in the lake (Sundborg and Calles, 2001). This approach relies on assumptions around (1) the relationship between grain size and settling velocity, and (2) modification of the period available for suspended sediment settling under conditions of fluctuating inflow rate.

Appearance of the Sedimentary Record

Detailed notes on sediment stratigraphy inform subsequent interpretation of physical, chemical, and biological proxies. Cohen (2003, p. 163) reminds workers that accurate interpretation "... relies on sound core or outcrop descriptions, and interpretations using more sophisticated analytical techniques than the qualitative assessments of the human eye can go wildly astray when supporting core and outcrop observations are not available." Schnurrenburger et al. (2003) present a lake sediment classification scheme based, in part, on macroscopic classification of the freshly exposed sediment surface into

discrete lithologic units. Such units are described, as appropriate, by color, sedimentary structure, thickness, inclination, bedding planes, presence of macroscopic fossils, and notes on the degree of stratigraphic disturbance by the coring process. Similar observations can be made on the material trapped at or below the core catcher, if available, to aid interpretation of the region where the core was refused further entry into the sediment. Assessment of color is best performed on wet sediment as representing the material as it occurs *in situ* or on partially dried sediment where rapidly (slowly) drying coarse (fine) sediment leads to the greatest contrast in color and tone. This technique has been shown to be particularly effective in distinguishing clastic varves that are indistinct or poorly visible in wet sediment. In a partially dried core or section, fine sediment deposited in winter remains moist and darker, while the coarser sediment dries more quickly and appears lighter (Gilbert, 2003b). As well, changes in color during exposure to air and drying may provide important information, for example, as iron oxidizes from black in a reduced state to red in the oxidized state. The overriding goals of the core logging process are to (1) help guide the interpretation of any subsequent analysis, (2) describe the overall compositional variability present in the core to guide subsampling, and (3) to identify potential unconformities in the sedimentary record.

The color reflected from exposed sediment surfaces is ultimately a function of the geometry of the exposed surface

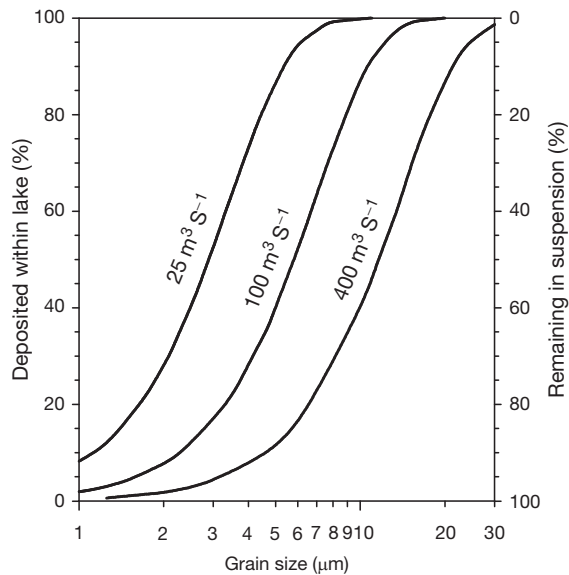


Figure 3 Sediment deposited in the outlet section of Lake Laitaure (Sweden), or remaining in suspension, in percentage of the amounts received at the lake delta over a 16:1 range of inflow rate. Plots based on calculations presented in Axellson (1967), and assume a water temperature of 8 °C, sediment particle density of 2.65 g cm⁻³, and that one-third of the lake volume is taking part in mixing (i.e., stratified lake). For a given grain size, a greater proportion are deposited at lower inflow rates. For a given deposit, a greater proportion is composed of coarse grains as inflow rate increases.

(roughness), water content, sediment mineralogy, and the intensity and spectrum of incident light. As a nondestructive sampling technique, description of color can be undertaken alongside logs of sediment type, structure, and bedding. The traditional reliance on qualitative comparison of samples with standardized Munsell color charts has been criticized for being time consuming and of questionable utility due to variations in the spectral response of the human eye (Kemp et al., 2001). Systematic downcore measurements of color reflectance using a spectrophotometer can provide reproducible and objective measurements that contribute to a quantitative, or semiquantitative, analysis. Recent application to lacustrine sediment includes the use of ratios in reflectance between wavelength bands (Debret et al., 2011), and narrow wavelength ranges (Wolfe et al., 2006), as proxy indicators of chlorophyll, carbonate, and iron concentrations. As freshly exposed core faces may oxidize rapidly, with a concomitant change in color, the use of a protective film during automated measurement may be necessary.

Radiography is less often used and more difficult to do than color analysis but is a nondestructive, and relatively rapid, technique applicable to whole cores. Radiography provides insights into the layering and internal structures of sediments (Kemp et al., 2001), as it can reveal stratigraphy otherwise not evident (Dean et al., 1999). Differential attenuation via absorption and scattering of x-radiation as it passes through the sediment sample is the result of variations in density, quantity (thickness), grain size, and mineralogy. The resulting radiograph, recorded via gray-scale intensity, is proportional to attenuation and must be calibrated with materials of known

and contrasting density (Principato, 2004). Reliance on radiographs obtained using conventional film has now migrated to digital x-radiographic techniques (St-Onge et al., 2007) and computed tomography (Ketcham and Carlson, 2001; Støren et al., 2010).

Sediment fabric describes the orientation and packing of sediment particles at the microscopic scale (Last, 2001). Although the paleoenvironmental importance of fabric has long been recognized, the available literature in application to lake sediments appears somewhat more modest (Last, 2001) – especially in contrast with marine sediments (Bennett et al., 1981; Hein, 1991). Nevertheless, mention of fabric continues to contribute to paleoenvironmental interpretations of lacustrine sediment sequences (Kemp et al., 2001). The orientation and packing of sediment particles are innately related to other physical parameters, including porosity, water content, strength, and bulk density. Fabric is controlled by the shape of the particles (e.g., single platy grains settling in calm water tend to accumulate horizontally) and by processes (especially currents) during deposition. However related these physical parameters are, the concept of sedimentary fabric is a distinct physical parameter, and use of the term ‘fabric’ has been invoked where ‘texture,’ ‘color,’ or ‘lithology’ is intended. Fabric naturally depends on the population characteristics of the grain size and shape being deposited, as well as the processes and rates of deposition, slope, and bed roughness. Analysis of fabric relies on high-resolution, optical or scanning electron microscopy (SEM) techniques to resolve grain-to-grain orientation, the presence and size of void space, and graded bedding (Dean et al., 1999; O’Brien and Pietraszek-Mattner, 1998). O’Brien and Pietraszek-Mattner (1998) used clay fabric to distinguish deposition as individual grains (Figure 4(a)) from suspension settling of flocculated sediment (Figure 4(b)).

Magnetic Properties

The paleomagnetic record of intensity, orientation, and polarity and the record of magnetic susceptibility (MS) of lacustrine sediments have been used to characterize the physical properties of the sediment, the processes of deposition, and the history of the deposit. Natural remnant magnetization in sediments is caused by the alignment of magnetic grains, particularly magnetite, in the Earth’s magnetic field at the time of deposition (Barendregt et al., 1998). Variation in the Earth’s magnetic field has occurred as a result of migration of the North Magnetic Pole and reversals in polarity. At any given location, the declination with respect to True North, the dip with respect to a horizontal plane (inclination), and the direction of the field are a function of these changes. Paleomagnetic records of inclination and declination have been used with some qualifications to date of lacustrine sedimentary records and thus establish the late Pleistocene and Holocene histories of lakes and their drainage basins (Carmichael et al., 1990; Mothersill, 1985, 1988) by comparison to type curves for a region (e.g., Creer and Tucholka, 1982). Assessment of correspondence between paleomagnetic records can reveal striking similarities among lakes within the same region (Figure 5). The most suitable materials are silt and clay-sized grains of magnetite free from secondary magnetization, postdepositional deformation, or weathering.

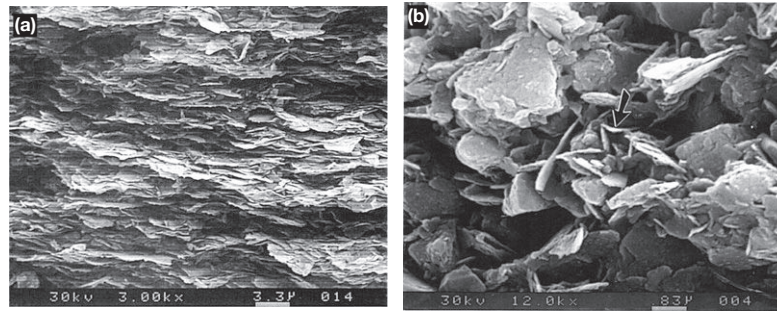


Figure 4 Fabric of (a) dispersed Pennsylvanian fireclay from a sedimentation experiment where grains accumulated horizontally ('face to face') and (b) flocculated clay formed in Pleistocene glacial Lake Albany where grains accumulate in a random ('cardhouse'; 'edge to face') arrangement. Sediments in (a) were dispersed in hexametaphosphate solution and allowed to settle for approximately 6 months. Reproduced from O'Brien NR and Pietraszek-Mattner S (1998) Origin of the fabric of laminated fine-grained glaciolacustrine deposits. *Journal of Sedimentary Research* 68: 832–840, with permission from the Society for Sedimentary Geology/SEPM.

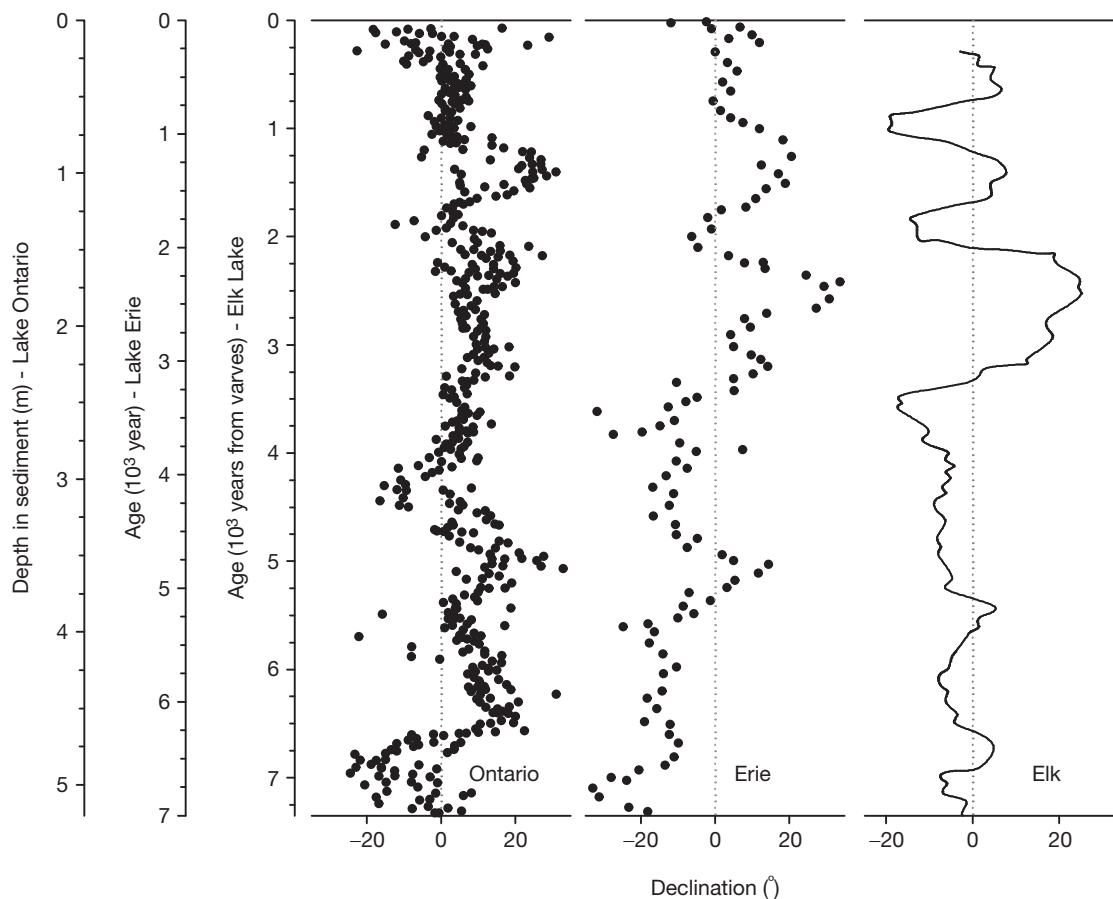


Figure 5 Magnetic declination in sediments from Lake Ontario, Lake Erie, and Elk Lake based on the comparison made by Carmichael et al. (1990). This plot compiled from data presented in Carmichael et al. (1990; Lake Ontario), Creer and Tucholka (1982; Lake Erie), and Sprowl and Banerjee (1989; Elk Lake composite curve). No attempt has been made to account for variations in sedimentation rate.

These fine grains are more likely to be oriented by the magnetic field than by the processes of transport and deposition, particularly in comparison to larger particles.

Paleomagnetic dating may be particularly useful where organic materials for radiocarbon dating are absent or where ancient carbonate (e.g., as marl) contaminates contemporary carbon or in massive sediments in the absence of varves as an

absolute dating record (Larsen et al., 2000). Records of magnetic reversals in sediments that include a lacustrine component have been used to date much longer records, especially of glaciation, extending back to the Pliocene (Barendregt et al., 1998; Froese et al., 2000). The absence of a recognizable magnetic signal or a pattern through time has been used to indicate the presence of other processes that prevent the orientation

of particles to the magnetic field, including currents in the lake during both summer and winter deposition of varves (Gravenor and Coyle, 1985).

Measurement of MS involves subjecting the sediment to a low-intensity magnetic field. This induces a small magnetization that lasts only while the field is present (Nowaczyk, 2001). Measurement of MS produces a record of magnetic minerals (Bradshaw and Thompson, 1985) and so may be useful in establishing provenance of a deposit (Foster et al., 1998). Comparison of MS profiles allows correlation among sites in one deposit where the stratigraphy may otherwise be ambiguous (Loizeau et al., 1997; Figure 6). Because most volcanic ash is highly magnetically susceptible, MS is also an effective method of locating tephra layers or zones in sediment, particularly where they cannot be readily recognized by color or texture. MS may also serve as a surrogate for measurement of bulk density and organic carbon content of the sediment, as the more compact the sediment and the lower the carbon content, the higher the concentration of magnetically susceptible minerals. Measurement of MS can be carried out *in situ* on exposures or on whole cores in nonmetallic tubes and thus provides a rapid, inexpensive, nondestructive technique particularly suitable to preliminary assessment of a deposit (Dearing, 1999).

Grain Size

The size of a clastic sediment particle and the size distribution of a group of particles are fundamental physical properties. Grain-size distribution influences other physical properties of bulk sediment, such as surface area, shear strength, and bulk

density and therefore related parameters including porosity and water content. It is a key clue in the interpretation of sedimentary process within a lake basin, the provenance of the sediment, and ultimately, the paleoenvironmental conditions in the surrounding watershed (Last, 2001). Although considerable scatter exists in the relationship between water depth and mean grain size, the deepest portions of the basin generally show the smallest grain size (Figure 7). Grain size is a widely applied technique in the interpretation of sedimentary records from modern, or former, glacier-fed lakes dominated by clastic inputs.

Although the size of an individual grain is a clear concept for simple geometries, in practice, its determination for natural particles is not straightforward. Only a sphere can be described by a single 'size' measurement; natural grains vary in geometry, roughness, and volume. To some degree, the size and distribution of grains in a particular sample depend on the technique used, because the size of individual grains is defined differently in each technique. The consequence is that different techniques of analysis can produce different measures of grain size for the same sediment sample (Goossens, 2008). Such techniques include analysis by pipette, laser diffraction, image analysis, electrical sensing, and x-ray absorption during sedimentation (an extensive review of techniques is provided in Last, 2001).

Grain-size analysis is traditionally directed at characterization of discrete clastic grains that require chemical dispersion to be analyzed as individuals, termed 'primary particles.' However, natural particles are more likely to be composites including two or more primary particles associated by flocculation or aggregation. Therefore, the *in situ* or effective grain-size

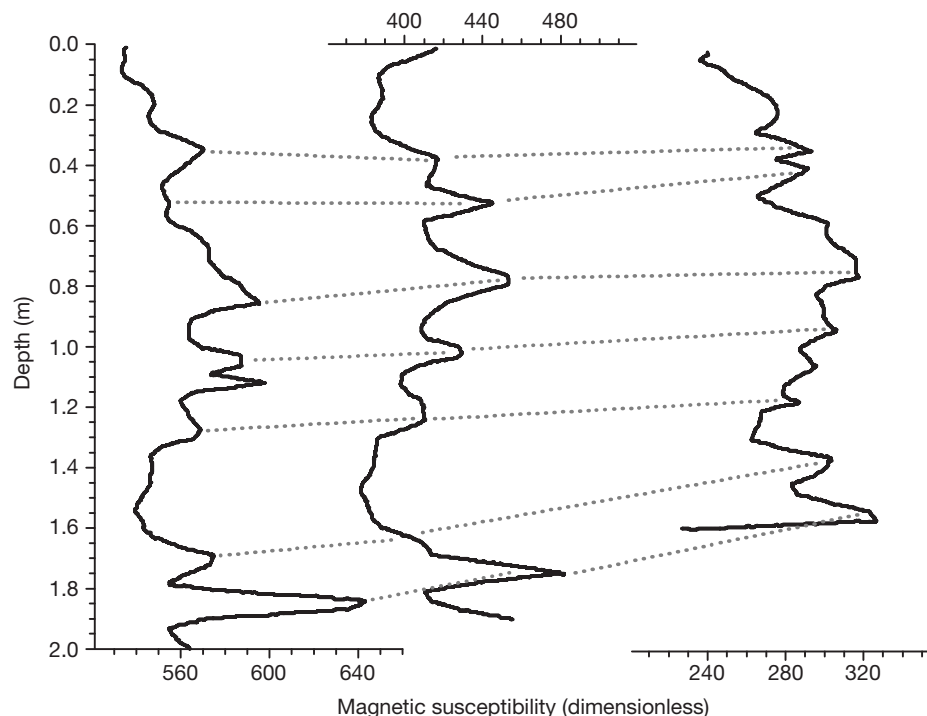


Figure 6 Magnetic susceptibility of three cores from Phewa Tal, Nepal, illustrating the interpreted correlation of deposits between sites (dotted lines). Sediment stratigraphy among cores from this lake has been described as 'poorly correlated' and 'ambiguous' by Ross and Gilbert (1999). Data reinterpreted from Ross and Gilbert (1999) and permission kindly provided by Elsevier.

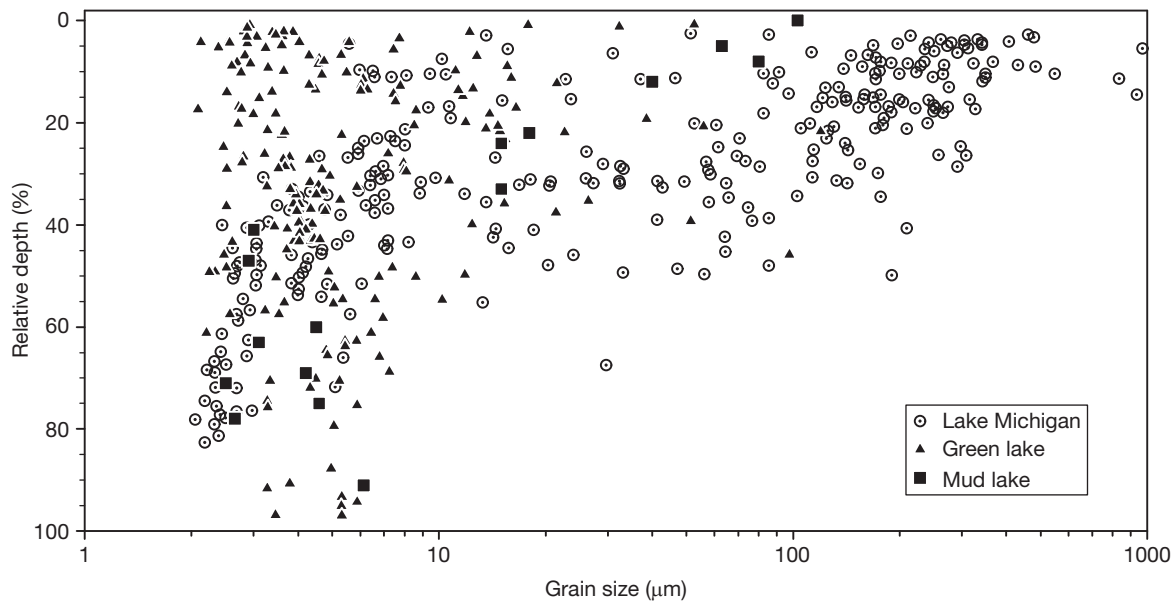


Figure 7 Mean grain-size relationship with relative water depth. Relative water depth (=100% for maximum depth) is used here to facilitate interlake comparison. Glacier-fed lakes are presented with solid filled symbols. This plot compiled from data presented in Schiefer (2006; Green Lake), Cahill (1981, Lake Michigan), and Hodder et al. (2006; Mud Lake). A larger mean grain size is generally found in shallow water.

distribution of the composite particles differs from the absolute grain-size distribution of the primary particles. The *in situ* hydrodynamics of the particle depend on the effective particle size, which may remain unknown, following laboratory treatments directed at characterization of primary particles. Field techniques that permit quantification of *in situ* grain size include laser diffraction (Hodder, 2009) and cameras (Grossart and Simon, 1993).

Attempts to interpret the processes that produce a particular sample are often based on suites of statistical measures of varying complexity. Size distribution is used to infer sediment deposition process, especially in comparison with other size distributions, to reveal time-transgressive trends. Unimodal distributions are rare because most deposits are the product of multiple simultaneous agents that confound a single process interpretation. Nevertheless, given a priori knowledge of the general depositional environment, the interpretation of lacustrine grain-size distributions is eased to some extent. The production, transport, and storage of sediment through the process network lead to partitioning of grain sizes within deposits from differing sedimentary environments. Lakes can usually be considered 'ultimate deposits' (cf. Stravers et al., 1991) as they undergo net deposition, and the sediment package is largely stable over the Holocene. As ultimate deposits, lakes are dominated by a background of silt and clay particles punctuated by event layers with markedly different grain-size distribution. Events interpreted from grain-size distributions in sedimentary records include turbidity currents (Gilbert et al., 2006), mass movements (Vogel et al., 2010), hurricanes (Donnelly and Woodruff, 2007), ice rafting (Vogel et al., 2010) or accumulation on lake ice platforms by niveo-eolian processes (Lewis et al., 2002), and snow avalanches (Vasskog et al., 2011).

Bulk Density

Bulk density describes the total volume of sediment relative to the dry or wet mass and therefore represents an index of the mineral, organic, gas, and water contributions within a sample. Bulk density also varies with fabric, as the grain-size distribution within a sediment sample modifies the packing of particles per unit volume. Considerable scatter may be present in the relationship between grain-size distribution and dry bulk density (Figure 8), but it is generally found that fine silt correlates strongly with dry bulk density. When used in combination with measurements of area, such as the surface area of a core sample, bulk density is a crucial physical parameter in the calculation of accumulation rates. Bulk density has also been directly applied as a proxy indicator, such as for the area covered by glacier ice (Bakke et al., 2005).

As organic material contains a significant quantity of water and in some lake sediments occupies a relatively minor volume, its specific contribution to bulk density is often omitted. In the absence of organic material, the wet bulk density (ρ_{bw}) of lake sediment can be calculated as

$$\rho_{bw} = \frac{m_w + m_s}{V_t} = \rho_w \frac{W}{100} + \rho_s \left[1 - \frac{W}{100} \right] \quad [1]$$

and is derived mass of water (m_w), mass of sediment (m_s), volume of the raw sediment sample (V_t), water content (W), density of water (ρ_w), and density of sediment particles (ρ_s). The equation assumes uniform density of sediment particles, and a value of 2.65 g cm^{-3} is typically invoked. The contributions via dissolved salts to the density of pore water are omitted. Dry bulk density (ρ_{bd}) is described by omitting the first term, assuming no water remains in pore spaces. The

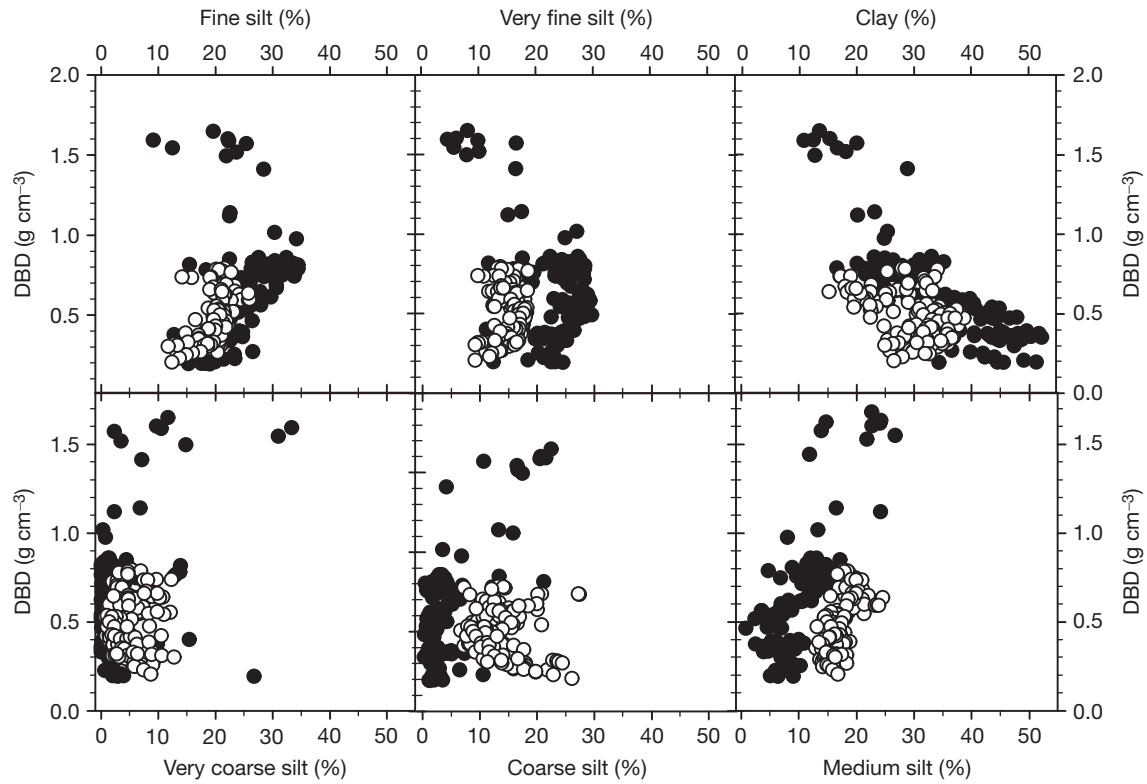


Figure 8 Scatterplot matrix showing the relationship between grain size and dry bulk density (DBD; g cm^{-3}) in two Norwegian lakes: Folgefonna and Aspvatnet. Grain-size analysis using SediGraph 5100. This plot is compiled from data presented in Bakke et al. (2006a; Folgefonna; hollow symbols) and Bakke et al. (2006b; Aspvatnet; filled symbols).

density of particles is overestimated in the presence of organic material, for which wet density values between 1.0 and 1.5 g cm^{-3} are typically invoked. Dewatering of organic matter and water-bearing minerals and loss of volatile organics during the drying process may be erroneously interpreted as water lost from pore spaces. In the presence of organic material, the wet bulk density (ρ_w) of lake sediment can be calculated as

$$\rho_{bw} = \frac{m_w + m_s + m_o}{V_t} \quad [2]$$

which includes mass of water (m_w), mass of sediment (m_s), mass of organic material (m_o), and total volume of the raw sediment sample (V_t).

Water Content

Water trapped within lacustrine sediment occupies the void space between solid mineral or organic material, and measures of water content (W) are derived using the mass of a raw sediment sample (m_r) compared to the dry mass of the same sediment sample (m_d):

$$W = 100 \left[\frac{m_r - m_d}{m_r} \right] \quad [3]$$

The resulting measure of water content is a percentage by mass of the raw sample. Dried sediment samples are typically

oven dried (105°C) for 12–24 h (Heiri et al., 2001) or freeze dried, in both cases for a period sufficient to arrive at constant dry mass. This technique does not partition the source of water lost during drying, such as evaporation from pore spaces, dewatering of minerals, or dehydration of organic matter (Santisteban et al., 2004). Oven drying may also result in loss of volatile organic matter components (Meyers and Teranes, 2001). Collectively, the overlap of these effects renders W a composite measure of mass loss from multiple repositories.

An overall decline in water content with depth is typical (Figure 9) as deeper sediment has been subject to greater period and mass of loading and therefore greater dewatering than sediment above. The portion of the sedimentary record near the sediment–water interface may be more liquid than solid (Glew et al., 2001), and water content can be over 50% in this zone (Figure 9). A water content of 50% does not correspond with half of the sample volume occupied by water, as water content is calculated by mass and the density difference between water and clastic sediment necessitates a corresponding water volume much greater than 50% (Figure 10). A variety of factors, including grain size (Figure 11), organic matter content, and dry sediment density, influence the water content of lacustrine sediments. Glacier-fed lakes appear to show lower water content than other lakes (Figure 9), perhaps owing to lower productivity, rock-flour sediment inputs, and/or the focusing action of turbidity currents. Strong empirical

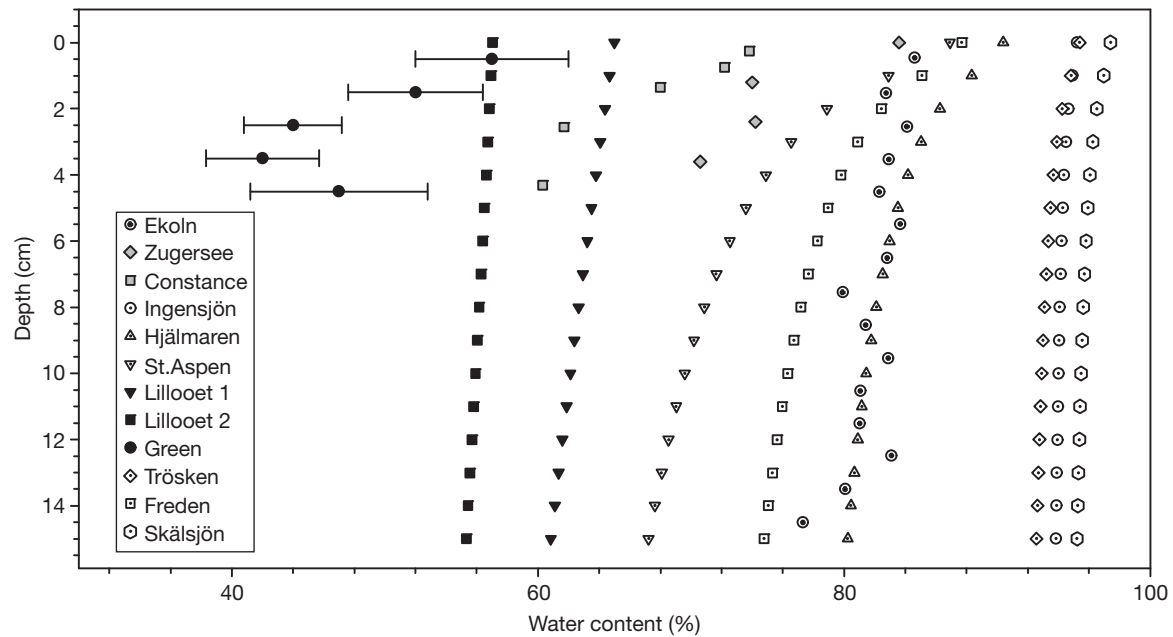


Figure 9 Water content with depth below the sediment–water interface. Glacier-fed lakes are presented with solid filled symbols, other lakes with hollow or dotted symbols. This plot is compiled from data presented in Schiefer (2006; Green), Larsen and Chilingar (1983; ZugerSee, Constance), Håkanson and Jansson (1983; Ekoln, Ingensjön, Tröskén, Skälsjön, Hjälmaren, St. Aspen, Vänern), and unpublished data from Lillooet Lake, British Columbia. Error bars for Green Lake are single standard deviation reported by Schiefer (2006).

associations have been demonstrated between water content and grain size (Schiefer, 2006), organic content (Menounos, 1997), water depth (Håkanson, 1981), sediment depth (Håkanson, 1981; Schiefer, 2006), and varve thickness (Gälman et al., 2006), respectively. As such, water content has been proposed as a potential surrogate measure for other physical properties (Håkanson, 1977, 1981; Menounos, 1997). Given that water content can be measured more readily, and perhaps at lesser expense, than other parameters (e.g., grain-size distribution, MS), it has also been proposed as an indicator variable useful in tuning the sampling for other analyses.

The water content immediately below the sediment–water interface shows characteristic spatial patterns within a basin, although such patterns vary widely between sites depending on the chemical, physical, and biological characteristics of each lake (Figure 12). The nonglacial pattern shows low water content values in shallow-water sediments subject to wind- and inflow-induced mixing, and with associated larger mean grain size (Bengtsson and Enell, 1986; Håkanson and Jansson, 1983). More isolated, deeper waters are often described as having a smaller mean grain size and correspondingly high water content (Figure 11). The water content of bottom sediments from glacier-fed Green Lake shows an opposite pattern – sediments retrieved from shallow water exhibit the highest water content values. Both the nonglacial and glacier-fed patterns can be distinguished in Figure 12. Glacier-fed lakes are typically highly turbid, cold, and rapidly flushed systems with correspondingly low productivity (Hodder, 2009), factors that may collectively correspond with low organic content, higher bulk density, and enhanced dewatering.

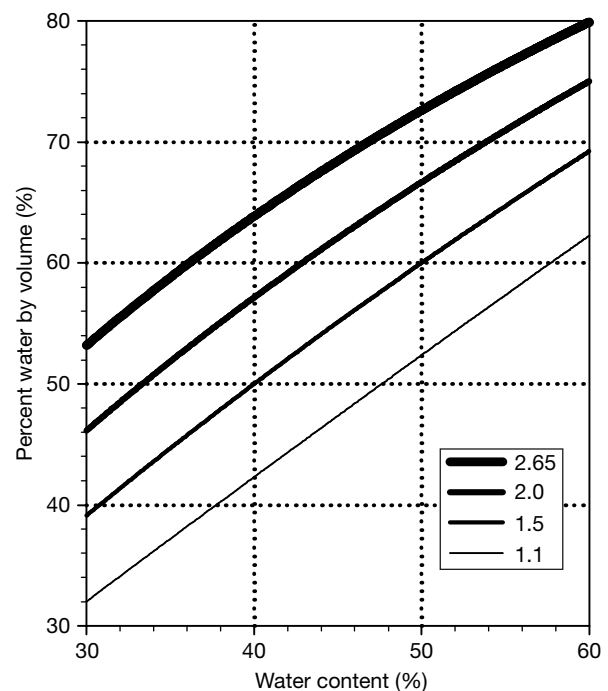


Figure 10 Relationship between water content (percentage by mass of raw sample) and percent water by volume in a two-phase mixture plotted for four values of sediment particles density (ρ_s ; g cm^{-3}). Density of water (ρ_w) assumed constant at 1.0 g cm^{-3} .

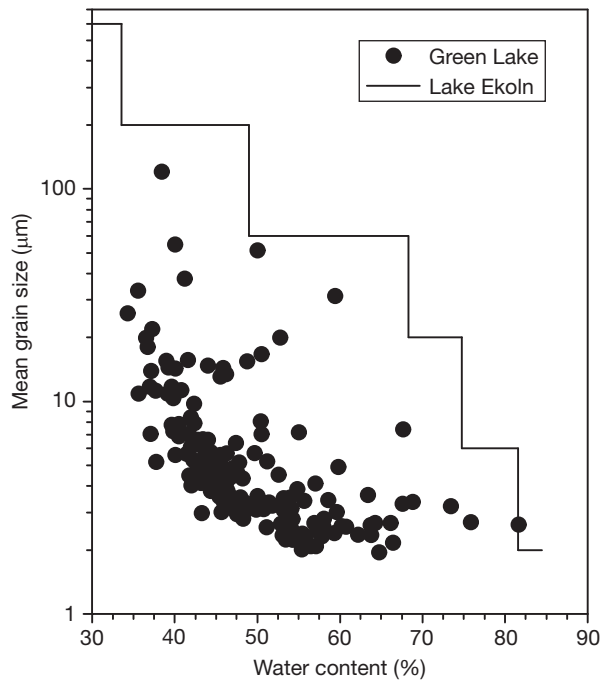


Figure 11 Water content relationship with mean grain size of bottom sediment. This plot is compiled from data presented in Håkanson (1977; Lake Ekoln) and Schiefer (2006; Green Lake).

Shear Strength

A fundamental property of a sedimentary deposit is its ability to resist deformation, commonly expressed as shear strength. Shear strength is especially important in determining the processes of slope failure and sediment focusing in a lake, the generation of other gravity flow processes such as turbidity currents, the life processes of the aquatic benthos, and the stability of engineered structures resting on the lake floor. The nature and success of sediment recovery using a corer also relate to variations in shear strength with depth, as this controls the combined nature of friction along the core walls and deformation at the leading edge of the corer (Glew et al., 2001). The ensuing stick-slip entry of the corer into the sediment is the result of changes in shear strength at, or near, the base of the corer. Penetrometers, widely applied in marine settings, rely on this property to measure shear strength *in situ* (Håkanson, 1986), within core samples (Backman et al., 2005), or to assess substrate suitability for coring in the field (Spooner et al., 2004). Shear strength will vary with depth below the sediment surface, and with position within a lake basin (Figure 13).

Shear strength (τ) is described according to

$$\tau = C + (\sigma - p) \tan \phi \quad [4]$$

where C is the cohesion between particles, σ is the pressure normal to the surface, p is the excess pore-water pressure, and ϕ is the angle of internal friction between particles. Excess pore-water pressure develops as sediment consolidates and packing

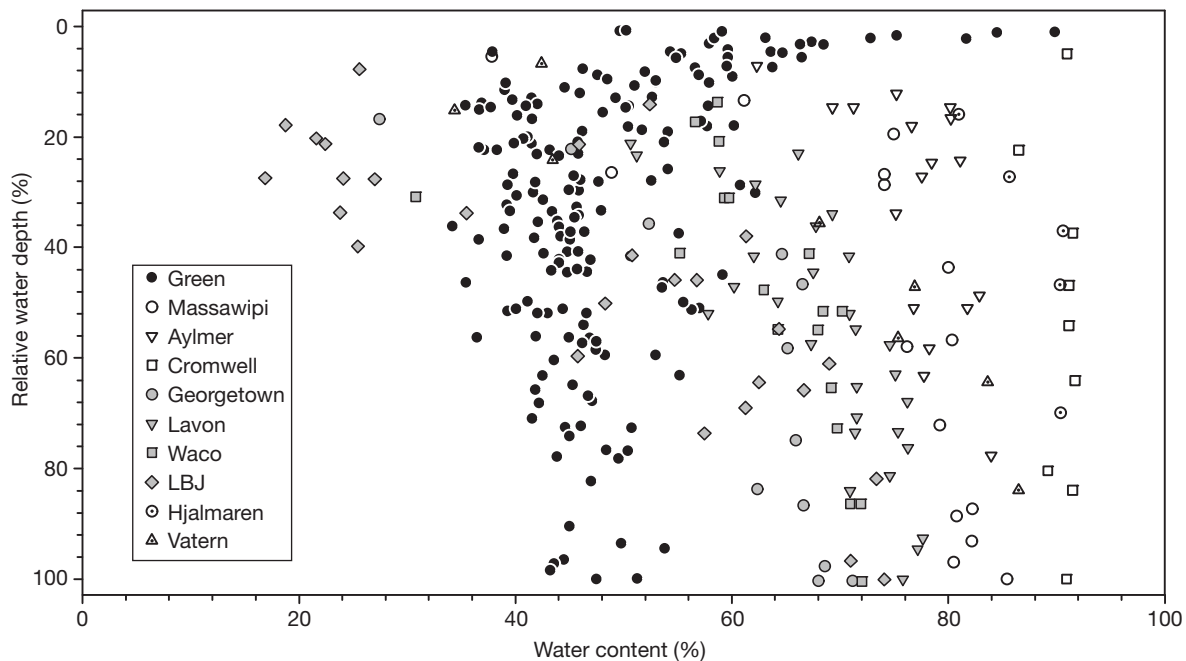


Figure 12 Water content of bottom sediments plotted against overlying relative water depth at various sites within lake and reservoir basins. Relative water depth (= 100% for the maximum depth) is used here to facilitate interlake comparison. This plot is compiled from data presented in Håkanson (1981; Hjälmaren and North Vättern, dotted symbols), Blais and Kalff (1995; Massawippi, Aylmer, and Cromwell, open symbols), Abraham et al. (1999; Georgetown, Lavon, Waco, and LBJ reservoirs, gray symbols), and Schiefer (2006; Green, black symbols). Glacier-fed Green Lake presented with black-filled symbols, other lakes with hollow, dotted, or gray-filled symbols.

increases, so the volume of water-filled voids between particles decreases and pore-water exit is limited. This situation is common in rapidly deposited sediments where grains deposited almost simultaneously form an open-network architecture which may collapse under applied stress such as loading or vibration. In some cases, this raises the pore-water pressure to near or above the normal pressure due to the mass of sediment. Particularly in sand, where cohesion is near zero, and if escape of the excess pore water is blocked by impervious fine-grained sediment above, the strength of the sediment may fall to zero. Fine-grained platy sediments, for example, clay minerals, deposited in edge-to-edge contact in saline water due to van der Waals forces may also collapse due to increased pore water and pressure. This spontaneous liquefaction (Terzaghi, 1955) changes sediment under the influence of gravity to a low-viscosity fluid. As the pore water escapes, the pressure decreases sufficiently to allow the shear strength to exceed the shear stress and the sediment ceases to move. As sediment deforms, the cohesion between particles decreases as they move one against another. This decrease in strength is referred to as the sensitivity, S , of the sediment and is expressed as the measurement of the ratio of the strength of undisturbed sediment to the strength of sediment that has been mixed mechanically, or remolded. Where $4 < S < 8$, the sediment is sensitive; where $S < 16$, the sediment is said to be quick (Terzaghi, 1955).

Early Diagenesis

Diagenesis, in the context of lacustrine sediment, refers to *in situ* changes in sediment that occur following deposition. Diagenetic modification of sediment can significantly alter the sedimentary record and therefore must be assessed in any paleoenvironmental interpretation. This includes all of the chemical, physical, and biological changes to the sediment following primary deposition: compaction, bioturbation, precipitation, and dissolution. Although chemical and biological diagenetic processes are beyond the scope of this article, some biological and chemical processes can leave behind physical features. For example, the movement of animals or insects through soft sediment can also create markings that are later visible as trace fossils (*lebensspuren*); widths can range from 10^{-4} (e.g., nematodes; O'Brien and Pietraszek-Mattner, 1998) to 10^{-2} m (e.g., water beetles; Uchman et al., 2009). Compaction occurs where sediment volume is reduced as individual grains are brought closer together, particularly where sediment is deposited rapidly. The spontaneous adjustment between sediment grains via compaction is promoted by the continued and progressive loading of overlying sediment through time. Water content is often used as a proxy for compaction (Håkanson and Jansson, 1983; Menounos, 1997). Although the porosity of the sediment is the primary control on the degree of possible compaction, other factors such as grain-size distribution and particle shape can also contribute. The overall result is the same: the density of the sediment is increased while volume is reduced.

Future Research

The physical properties of all lacustrine sediments are a composite product of processes occurring at a range of temporal

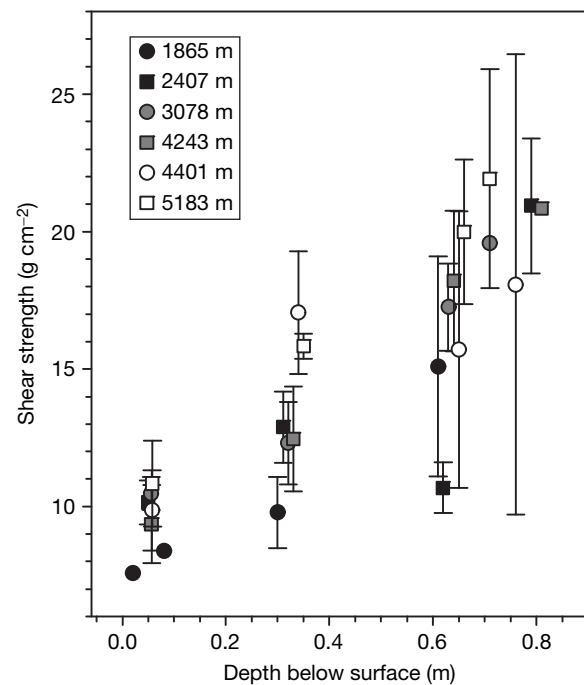


Figure 13 Shear strength in Lillooet Lake sediments plotted according to depth below sediment surface and distance from the Lillooet River delta (the primary inflow to the lake). Whiskers are single standard deviation. Shear strength generally increases with depth below surface and with distance from primary inflow. Unpublished data from Lillooet Lake, British Columbia.

and spatial scales. The composite character of the sedimentary record is precisely what makes lacustrine sediment valuable as a paleoenvironmental proxy: a variety of processes, both within the drainage basin (or lake) and beyond, contribute to the record. These sediment packages are a mix of clastic particles, organic matter, water, and other constituents. Determinations of physical properties are fundamental in any attempt to interpret a deposit as a paleoenvironmental proxy. This applies, for example, to bedding thickness or bulk density in the determination of sedimentation rate(s) or to determination of basin infill patterns to contextualize sediment accumulation and therefore the type(s) of paleoenvironmental signal recorded. Recent advances have included very-high-resolution proxies (e.g., $<100 \mu\text{m}$; Cuven et al., 2011), the generation of new proxies from physical properties (Debret et al., 2011; Wolfe et al., 2006), and deeper consideration of the nature, and consequences, of spatial variability in proxy signals (e.g., Cuven et al., 2011; Stewart and Lamoureux, 2011).

Future work will undoubtedly be aimed at exploiting a fuller understanding of the link between multiple, simultaneous processes, the nature of the associated physical properties preserved in lacustrine sediments, and the interpretation of corresponding proxies. Growing recognition of site or region specificity in lacustrine proxies promises to reveal new sources of paleoenvironmental data, some of which record unique or conflicting proxy relationships. Given the significant investments of time, effort, and funding required to produce research data, we also expect to see greater participation in data archiving/sharing following future research. Such an approach promises to promote the original research that created the data but

may also lead to new collaborations between new data users and original data creators. Data sharing is somewhat more common for temporal series of physical properties (e.g., paleolimnological proxies, World Data Center for Paleoclimatology) than spatial series of similar data. In both cases, the data represent a research product that is arguably of greater value than the associated publications in which they appear.

See also: [Paleolimnology: Multiproxy Approaches; Overview of Paleolimnology.](#)

References

- Abraham J, Allen PM, Dunbar JA, and Dworkin SI (1999) Sediment type distribution in reservoirs: Sediment source versus morphometry. *Environmental Geology* 38(2): 101–110.
- Axelsson W (1967) The Laitaure Delta, a study of deltaic morphology and processes. *Geografiska Annaler* 49A: 1–127.
- Backman J, Moran K, McInroy D, and IODP Expedition 302 Scientists (2005) IODP Expedition 302 Arctic Coring Expedition (ACEX): A first look at the Cenozoic paleoceanography of the central Arctic Ocean. *Scientific Drilling* 1: 12–17.
- Bakke J, Lie Ø, Nesje A, Olaf Dahl S, and Paasche Ø (2005) Utilizing physical sediment variability in glacier-fed lakes for continuous glacier reconstructions during the Holocene, northern Fjellfonna, western Norway. *The Holocene* 15: 161–176.
- Bakke J, Lie Ø, Nesje A, Olaf Dahl S, and Paasche Ø (2006a) Fjellfonna, Western Norway Glacial Lake Sediment Data and ELA Reconstruction. *IGBP PAGES/World Data Center for Paleoclimatology Data Contribution Series # 2006-028*. NOAA/NCDC Paleoclimatology Program, Boulder, CO, USA.
- Bakke J, Olaf Dahl S, Paasche Ø, Løvlie R, and Nesje A (2006b) Northern Norway Aspvatnet Glacial Lake Sediment Data. *IGBP PAGES/World Data Center for Paleoclimatology Data Contribution Series # 2006-027*. NOAA/NCDC Paleoclimatology Program, Boulder, CO, USA.
- Barendregt RW, Vincent JS, Irving E, and Baker J (1998) Magnetostratigraphy of Quaternary and late Tertiary sediments on Banks Island, Canadian Arctic Archipelago. *Canadian Journal of Earth Sciences* 35: 147–161.
- Bengtsson L and Enell M (1986) Chemical analysis. In: Berglund B (ed.) *Handbook of Holocene Palaeoecology and Palaeohydrology*, pp. 423–451. Chichester: Wiley.
- Bennett RH, Bryant WR, and Keller GH (1981) Clay fabric of selected submarine sediments: Fundamental properties and models. *Journal of Sedimentary Petrology* 51: 217–232.
- Blais J and Kalff J (1995) The influence of lake morphometry on sediment focusing. *Limnology and Oceanography* 40(3): 582–588.
- Bloxham J and Gubbins D (1985) The secular variations of Earth's magnetic field. *Nature* 317: 777–781.
- Bradshaw R and Thompson R (1985) The use of magnetic measurements to investigate the mineralogy of Icelandic lake sediments. *Boreas* 14: 203–215.
- Brune GM (1953) Trap efficiency of reservoirs. *Transactions of the American Geophysical Union* 34: 407–418.
- Cahill RA (1981) Geochemistry of recent Lake Michigan sediments. Illinois State Geological Survey Division Circular 51. Champagne: Illinois State Geological Survey.
- Carmichael CM, Mothersill JS, and Morris WA (1990) Paleomagnetic and pollen chronostratigraphic correlations of the lake glacial and postglacial sediments in Lake Ontario. *Canadian Journal of Earth Sciences* 27: 131–147.
- Cohen AS (2003) *Paleolimnology: The History and Evolution of Lake Systems*. New York: Oxford University Press.
- Creer KM and Tucholka P (1982) Construction of type curves of geomagnetic secular variation for dating lake sediments from east central North America. *Canadian Journal of Earth Sciences* 19: 1106.
- Cuven S, Francus P, and Lamoureux S (2011) Mid to Late Holocene hydroclimatic and geochemical records from the varved sediments of East Lake, Cape Bounty, Canadian High Arctic. *Quaternary Science Reviews* 30(19–20): 2651–2665.
- Dean JM, Kemp AES, Bull D, Pike J, Petterson G, and Zolitschka B (1999) Taking varves to bits: Scanning electron microscopy in the study of laminated sediments and varves. *Journal of Paleolimnology* 22: 121–136.
- Dearing JA (1999) Holocene environmental change from magnetic proxies in lake sediments. In: Maher BA and Thompson R (eds.) *Quaternary Climates, Environments and Magnetism*, pp. 231–278. Cambridge: Cambridge University Press.
- Debret M, Sebag D, Desmet M, et al. (2011) Spectrocolorimetric interpretation of sedimentary dynamics: The new Q7/4 diagram. *Earth-Science Reviews* 109: 1–19.
- Donnelly JP and Woodruff JD (2007) Intense hurricane activity over the past 5,000 years controlled by El Niño and the West African monsoon. *Nature* 447: 465–468.
- Foster IDL, Lees JA, Owens PN, and Walling DE (1998) Mineral magnetic characterization of sediment sources from an analysis of lake and floodplain sediments in the catchments of the Old Mill reservoir and Slapton Ley, south Devon, UK. *Earth Surface Processes and Landforms* 23: 685–703.
- Francus P, Bradley RS, Lewis T, Abbott M, Retelle M, and Stoner JS (2008) Limnological and sedimentary processes at Sawtooth Lake, Canadian High Arctic, and their influence on varve formation. *Journal of Paleolimnology* 40(3): 963–985.
- Froese DG, Barendregt RW, Endin RJ, and Baker J (2000) Paleomagnetic evidence for multiple Late Pliocene–early Pleistocene glaciations in the Klondike area, Yukon Territory. *Canadian Journal of Earth Sciences* 37: 863–877.
- Gälmán V, Petterson G, and Renberg I (2006) A comparison of sediment varves (1950–2003 AD) in two adjacent lakes in Northern Sweden. *Journal of Paleolimnology* 35(4): 837–853.
- Gilbert R (1975) Sedimentation in Lillooet Lake, British Columbia. *Canadian Journal of Earth Sciences* 12(10): 1697–1711.
- Gilbert R (2003a) Lacustrine sedimentation. In: Middleton GE (ed.) *Encyclopedia of Sedimentology and Sedimentary Rocks*, pp. 404–408. Dordrecht, MA: Kluwer.
- Gilbert R (2003b) Varves. In: Middleton GE (ed.) *Encyclopedia of Sedimentology and Sedimentary Rocks*, pp. 764–766. Dordrecht, MA: Kluwer.
- Gilbert R (2003c) Spatially irregular sedimentation in a small, morphologically complex lake: Implications for paleoenvironmental studies. *Journal of Paleolimnology* 29: 209–220.
- Gilbert R, Crookshanks S, Hodder KR, Spagnoli J, and Stull R (2006) The record of an extreme flood in the sediments of montane Lillooet Lake, British Columbia: Implications for paleoenvironmental assessment. *Journal of Paleolimnology* 37: 737–745.
- Girardclos S, Baster I, Wildi W, Pugin A, and Rachoud-Schneider AM (2003) Bottom-current and wind-pattern changes as indicated by Late Glacial and Holocene sediments from western Lake Geneva (Switzerland). *Eclogae Geologicae Helvetiae* 96: S39–S48.
- Glew JR, Smol JP, and Last WM (2001) Sediment core collection and extrusion. In: *Tracking Environmental Change Using Lake Sediments*, vol. 1 (Part II), pp. 73–105.
- Goossens D (2008) Techniques to measure grain-size distributions of loamy sediments: A comparative study of ten instruments for wet analysis. *Sedimentology* 55: 65–96.
- Gravenor CP and Coyle DA (1985) Origin and magnetic fabric of glacial varves, Nottawasaga River, Ontario, Canada. *Canadian Journal of Earth Sciences* 22: 291–293.
- Grossart HP and Simon M (1993) Limnetic macroscopic organic aggregates (lake snow): Occurrence, characteristics, and microbial dynamics in Lake Constance. *Limnology and Oceanography* 38(3): 532–546.
- Håkanson L (1977) The influence of wind, fetch, and water depth on the distribution of sediments in Lake Vänern, Sweden. *Canadian Journal of Earth Sciences* 14(3): 397–412.
- Håkanson L (1981) Determination of characteristic values for physical and chemical lake sediment parameters. *Water Resources Research* 17(6): 1625–1640.
- Håkanson L (1986) A sediment penetrometer for in situ determination of sediment type and potential bottom dynamic conditions. *International Review of Hydrobiology* 71(6): 851–858.
- Håkanson L and Jansson M (1983) *Principles of Lake Sedimentology*. Berlin: Springer.
- Hein FJ (1991) The need for grain size analyses in marine geotechnical studies. In: Syvitski JPM (ed.) *Principles, Methods, and Application of Particle Size Analysis*, pp. 346–362. New York: Cambridge University Press.
- Heiri O, Lotter AF, and Lemcke G (2001) Loss on ignition as a method for estimating organic and carbonate content in sediments: Reproducibility and comparability of results. *Journal of Paleolimnology* 25: 101–110.
- Hodder KR (2009) Flocculation: A key process in the sediment flux of a large, glacier-fed lake. *Earth Surface Processes and Landforms* 34: 1151–1163.
- Hodder KR, Desloges JR, and Gilbert R (2006) Pattern and timing of sediment infill at glacier-fed Mud Lake: Implications for Late-Glacial and Holocene Climate in the Monashee Mountain region of British Columbia, Canada. *The Holocene* 16: 705–716.
- Hodder KR, Gilbert R, and Desloges J (2007) Glaciolacustrine varved sediment as an alpine hydroclimatic proxy. *Journal of Paleolimnology* 38: 365–394.
- Kashiwaya K, Ryuguo M, Horii M, Sakai H, Nakamura T, and Kawai T (1999) Climato-limnological signals during the past 260,000 years in physical properties of bottom sediments from Lake Baikal. *Journal of Paleolimnology* 21: 143–150.
- Kemp AES, Dean J, Pearce RB, and Pike J (2001) Recognition and analysis of bedding and sediment fabric features. In: Last WM and Smol JP (eds.) *Tracking*

- Environmental Change Using Lake Sediments. Physical and Chemical Techniques*, vol. 2, pp. 7–22. Dordrecht, MA: Kluwer.
- Ketcham RA and Carlson WD (2001) Acquisition, optimization and interpretation of X-ray computed tomographic imagery: Applications to the geosciences. *Computers and Geosciences* 27(4): 381–400.
- Lamoureux SF (1994) Embedding unfrozen lake sediments for thin section preparation. *Journal of Paleolimnology* 10: 141–146.
- Larsen G and Chilingar GV (1983) *Diagenesis in Sediments and Sedimentary Rocks*. Amsterdam: Elsevier.
- Larsen CPS, Morris WA, and MacDonald GM (2000) Records of geomagnetic secular variation since 1200 AD and the potential for chronological control of lake sediments in northern Alberta. *Canadian Journal of Earth Sciences* 37: 1711–1722.
- Last WM (2001) Textural analysis of lake sediments. In: Last WM and Smol JP (eds.) *Tracking Environmental Change Using Lake Sediments. Physical and Geochemical Methods*, vol. 2, pp. 41–81. Dordrecht, MA: Kluwer.
- Last WM and Smol JP (2001) An introduction to basin analysis, coring, and chronological techniques used in paleolimnology. In: Last WM and Smol JP (eds.) *Tracking Environmental Change Using Lake Sediments. Basin Analysis, Coring, and Chronological Techniques*, vol. 1, pp. 1–4. Dordrecht, MA: Kluwer.
- Lewis T, Gilbert R, and Lamoureux SF (2002) Spatial and temporal changes in sedimentary processes at proglacial Bear Lake, Devon Island, Nunavut, Canada. *Arctic, Antarctic, and Alpine Research* 34(2): 119–129.
- Likens GE and Davis MB (1975) Post-glacial history of Mirror Lake and its watershed in New Hampshire, USA: An initial review. *Verhandlungen der Internationalen Vereinigung für Theoretische und Angewandte Limnologie* 19(2): 982–993.
- Loizeau JL, Dominik J, Luzzi T, and Vernet JP (1997) Sediment core correlation and mapping of sediment accumulation rates in Lake Geneva (Switzerland, France) using volume magnetic susceptibility. *Journal of Great Lakes Research* 23: 391–402.
- Ludlam SD, Feeney S, and Douglas MSV (1996) Changes in the importance of lotic and littoral diatoms in a high arctic lake over the last 191 years. *Journal of Paleolimnology* 16(2): 187–204.
- Menounos B (1997) The water content of lake sediments and its relationship to other physical parameters: An alpine case study. *The Holocene* 7: 207–212.
- Meyers PA and Teranes JL (2001) Sediment organic matter. In: Last WM and Smol JP (eds.) *Tracking Environmental Change Using Lake Sediments. Physical and Geochemical Methods*, vol. 2, pp. 239–269. Dordrecht: Kluwer.
- Michelutti N, Blais JM, Cumming BF, et al. (2010) Do spectrally inferred determinations of chlorophyll a reflect trends in lake trophic status? *Journal of Paleolimnology* 43: 205–217.
- Mothersill JS (1985) Batchawana Bay, Lake Superior: Late Quaternary sedimentary fill and paleomagnetic record. *Canadian Journal of Earth Sciences* 22: 39.
- Mothersill JS (1988) Paleomagnetic dating of late glacial and postglacial sediments in Lake Superior. *Canadian Journal of Earth Sciences* 25: 1791–1799.
- Nederbragt AJ and Thurow JW (2001) A 6000 yr varve record of Holocene climate in Saanich Inlet, British Columbia, from digital sediment colour analysis of ODP Leg 196S cores. *Marine Geology* 174: 95–110.
- Nowaczyk NR (2001) Logging of magnetic susceptibility. In: Last WM and Smol JP (eds.) *Tracking Environmental Change Using Lake Sediments. Basin Analysis, Coring, and Chronological Techniques*, vol. 1, pp. 155–170. Dordrecht, MA: Kluwer.
- O'Brien NR and Pietraszek-Mattner S (1998) Origin of the fabric of laminated fine-grained glaciolacustrine deposits. *Journal of Sedimentary Research* 68: 832–840.
- Pickrill RA and Irwin J (1983) Sedimentation in a deep glacier-fed lake – Lake Tekapo, New Zealand. *Sedimentology* 30: 63–75.
- Principato SM (2004) X-ray radiographs of sediment cores: A guide to analyzing diamicton. In: Francus P (ed.) *Image Analysis, Sediments and Palaeoenvironments*, pp. 165–185. Dordrecht, MA: Kluwer.
- Ross J and Gilbert R (1999) Lacustrine sedimentation in a monsoon environment: The record from Phewa Tal, middle mountain region of Nepal. *Geomorphology* 27: 307–323.
- Santisteban JL, Mediavilla R, López-Pamo E, et al. (2004) Loss on ignition: A qualitative or quantitative method for organic matter and carbonate mineral content in sediments? *Journal of Paleolimnology* 32: 287–299.
- Schiefer E (2006) Predicting sediment physical properties within a Montane Lake Basin, Southern Coast Mountains, British Columbia, Canada. *Lake and Reservoir Management* 22(1): 69–78.
- Schnurrenberger D, Russell J, and Kelts K (2003) Classification of lacustrine sediments based on sedimentary components. *Journal of Paleolimnology* 29: 141–154.
- Solotchina EP, Prokopenko AA, Kuzmin MI, Solotchin PA, and Zhdanova AN (2004) Climate signals in sediment mineralogy of Lake Baikal and Lake Hovsgol during the LGM-Holocene transition and the 1-Ma carbonate record from the HDP-04 drill core. *Quaternary International* 205(1–2): 38–52.
- Spooner IS, Williams P, and Martin K (2004) Construction and use of an inexpensive, lightweight free-fall penetrometer: Applications to paleolimnological research. *Journal of Paleolimnology* 32: 305–310.
- Sprowl DR and Banerjee SK (1989) The Holocene paleosecular variation record from Elk Lake, Minnesota. *Journal of Geophysical Research* 94(B7): 9369–9388.
- Stewart KA and Lamoureux SF (2011) Connections between river runoff and limnological conditions in adjacent high Arctic Lakes: Cape Bounty, Melville Island, Nunavut. *Arctic* 64(2): 169–182.
- St-Onge G, Mulder T, and Francus P (2007) Continuous physical properties of cored marine sediments. In: Hillaire-Marcel C and de Vernal A (eds.) *Proxies in late Cenozoic Paleoclimatology*, pp. 63–98. Amsterdam: Elsevier.
- Støren EN, Dahl SO, Nesje A, and Paasche O (2010) Identifying the sedimentary imprint of high-frequency Holocene river floods in lake sediments: Development and application of a new method. *Quaternary Science Reviews* 29(23–24): 3021–3033.
- Stravers JA, Syvitski JPM, and Praeg DB (1991) Application of size sequence data to glacial–paraglacial sediment transport and sediment partitioning. In: Syvitski JPM (ed.) *Principles, Methods, and Application of Particle Size Analysis*, pp. 293–310. Cambridge: Cambridge University Press.
- Sundborg Å and Calles B (2001) Water discharges determined from sediment distributions: A palaeohydrological method. *Geografiska Annaler* 83A(1–2): 39–54.
- Terzaghi K (1955) Influence of geological factors on the engineering properties of sediments. *Economic Geology* 50: 557–618.
- Uchman A, Kazakauskas V, and Gaigalas A (2009) Trace fossils from Late Pleistocene varved lacustrine sediments in eastern Lithuania. *Palaeogeography, Palaeoclimatology, Palaeoecology* 272(3–4): 199–211.
- Vasskog K, Nesje A, Støren AN, Waldmann N, Chapron E, and Ariztegui D (2011) A Holocene record of snow-avalanche and flood activity reconstructed from a lacustrine sedimentary sequence in Oldevatnet, western Norway. *The Holocene* 21(4): 597–614.
- Vogel H, Wagner B, Zanchetta G, Sulpizio R, and Rosén P (2010) A paleoclimate record with tephrochronological age control for the last glacial–interglacial cycle from Lake Ohrid, Albania and Macedonia. *Journal of Paleolimnology* 44(1): 295–310.
- Wolfe AP, Vinebrooke RD, Michelutti N, Rivard B, and Das B (2006) Experimental calibration of lake-sediment spectral reflectance to chlorophyll a concentrations: Methodology and paleolimnological validation. *Journal of Paleolimnology* 36(1): 91–100.

Contributions of Paleolimnological Research to Biogeography

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Introduction

Biogeography uses the spatial and temporal distributions of organisms to address a wide range of ecological and environmental questions including the following: (1) What are the environmental variables controlling the distribution of a group of organisms? (2) Why has the composition of this community changed over time? (3) What is the impact of climate change on ecosystems? (4) How do human activities (e.g., land clearance agriculture, mining, and other industry) affect the distribution of species? (5) How do communities respond to disturbance, such as fire? (6) What causes changes in biodiversity? (7) How do species spread from one region to another and establish populations in a new area, and (8) How does the introduction of exotic species affect ecosystem functioning? The question of scale is import to biogeographers – and these problems are addressed on spatial scales ranging from microscopic to global and on temporal scales from hours to millions of years. This article will discuss several different spatial and temporal scales but will focus on local to global scales and the last 10 000 years (the Anthropocene and Holocene).

Paleolimnology uses the physical, chemical, and biological information stored in lake, reservoir, wetland, and to a lesser extent, river sediments to reconstruct past environmental conditions. Underlying paleolimnological research is the principal that, in many lake systems, sediments accumulate layer upon layer, year after year, such that younger sediments overlie older sediments. A variety of dating techniques, including ^{210}Pb and ^{14}C , are available to date sediment layers (Appleby, 2001; Björck and Wohlfarth, 2001; see [History of Dating Methods](#)). Much paleolimnological research has focused on addressing a wide spectrum of environmental change issues, which have depended on an understanding of modern biogeographic distributions of aquatic organisms to reconstruct past environmental conditions (Smol, 2008, 2010). Although in the last 20–30 years, remains from many different aquatic organisms, including ostracods, diatoms, chironomids, *Chaoborus*, cladocera, algal pigments, siliceous protozoans, and chrysophytes, have been used, this review will focus on two groups, diatoms (class Bacillariophyceae; see [Diatom Introduction](#)) and midges (families Chironomidae and Chaoboridae; see [Chironomid Overview](#)).

Diatoms are unicellular algae that are characterized by a cell wall composed of opaline silica ([Figure 1](#)). The diatom cell (frustule) is comprised of two halves (valves) that fit together like a petri dish, and are held together by siliceous bands. The opaline silica composition of the cell wall ensures that diatoms are generally well preserved in lake sediments. It has been suggested that there are at least 12 000 different diatom species (Willén, 1991), but new estimates are as great as 200 000 (Mann and Droop, 1996). Based on the size, shape, and ornamentation of the valve, diatoms can be identified to species or even subspecies level ([Figure 1](#)). Diatoms are useful

bioindicators because they occur in large numbers, making statistical analyses possible, in almost every aquatic environment where there is sufficient light for photosynthesis.

Chironomids, commonly referred to as the non-biting midges, are ubiquitous in freshwater ecosystems. Their life cycle is characterized by four stages (egg, larvae, pupae, and adult), with most of their life spent in the aquatic larval stage. Larval stages have a head capsule composed of chitin and are preserved in lake sediments. Using characteristics of the mandibles and ventromental plate, chironomids can be identified to genus or species level ([Figure 2](#)). *Chaoborus*, commonly referred to as the phantom midges, are also widespread, and because the larvae are covered by a rigid, chitinous cuticle, they are represented in lake sediments by body parts such as antennal segments, premandibular fans, and head capsule fragments, although they are most commonly represented by their often well-preserved and easily identifiable chitinous mandibles (Frey, 1964). The mandible structures are usually retained in fossil assemblages and can typically be identified to species level (Uutala and Smol, 1996). Using diatom-, chironomid- and *Chaoborus*-based research, this chapter reviews how paleolimnological research is contributing to answering biogeographic questions. The basis for some of this research is the ‘paleolimnological approach.’

Following the 1980s, when rapid advances in paleolimnological techniques occurred as a result of the ‘acid rain debate’ (see [Lake Chemistry; Multiproxy Approaches](#)), paleolimnology shifted from being largely qualitative and descriptive to quantitative and analytical (Birks, 1998). Many researchers now use a ‘paleolimnological approach’ to quantitatively infer past environmental conditions ([Figure 3](#); Cohen, 2003; Smol, 2008). This approach assumes that the modern relationships between abiotic conditions and biota remain unchanged through time. Transfer functions (i.e., mathematical relationships) relate modern community composition to abiotic conditions, and these transfer functions can then be applied to fossil assemblage data to infer past abiotic variables ([Figure 4](#)). Modern community composition data are determined from the upper 0–1 cm of lake sediments, which represents a spatially and temporally integrated sample, from a large lake training or calibration set (typically more than 50 lakes). Such samples allow for direct comparison to fossil assemblages preserved in deeper sediments. A wide range of statistical techniques have been developed to determine transfer functions (Birks, 2010; Cohen, 2003). For any given study, the selection of lakes is critical, as the calibration or training set must span a gradient of the environmental variable of interest. For example, if a researcher is interested in determining past pH, the calibration set must include lakes with a wide range of pH values. Once the biotic and abiotic data are in hand, it is critical to ensure that the diatom or chironomid distribution is, indeed, controlled by the environmental variable of interest. If it is, then species optima and tolerances of the variable of interest

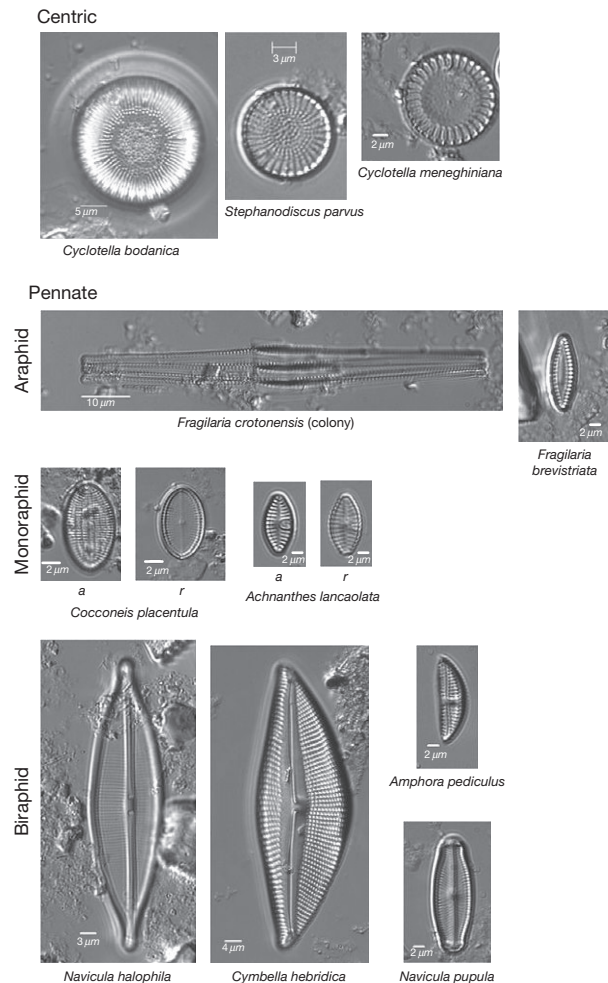


Figure 1 Diatom photomicrographs illustrating the two main morphologic groups: the centrics and the pennates. The pennate diatoms are subdivided into three groups based on the presence or absence of a raphe structure, which is a slit along the long axis of the diatom from which a living diatom secretes mucilage that can be used for attachment or movement. The three groups are the following: (1) the araphids, which are diatom taxa that do not have a raphe structure; (2) the monoraphids, which are diatom taxa that have a raphe structure on one valve and not the other; and (3) the biraphids, which are diatom taxa that have a raphe structure on both valves. These photographs also illustrate the incredible diversity of diatom valve structure and ornamentation, which are used to identify diatoms to species level. Photomicrographs are taken using a Leica transmission light microscope equipped with differential interference contrast optics at 1000 $\times$.

can be determined, which is an important contribution to the understanding of the biogeography of these organisms. Transfer functions have been developed to quantitatively infer pH, salinity, nutrient levels, dissolved organic carbon (DOC), metals, light, temperature, and lake level variation from diatoms (Moser, 2004; Smol and Stoermer, 2010), and temperature, lake level, salinity, phosphorus, and dissolved oxygen from chironomids (Brodersen and Quinlan, 2006; Brooks, 2006; Porinchu and MacDonald, 2003). These models have been applied to fossil assemblages to address a broad spectrum of research in biogeography including the following: (1) climate change,

(2) effects of disturbance and ensuing succession, (3) impacts of atmospheric pollution on water quality, (4) biodiversity, (5) invasive species, and (6) species extinction and evolution.

Critical to the ability to apply paleolimnological techniques to address many biogeographic issues has been the advancement of new statistical techniques, which have allowed for improved accuracy and precision in inferences of past environmental conditions (Birks, 1998). The ability to obtain high-temporal resolution records has also allowed paleolimnologists to address questions that they could not have attempted in the past. New methods for coring and subsampling sediment cores at finer intervals (Glew et al., 2001; Leroy and Colman, 2001), as well as improved dating methods (Last and Smol, 2001), have allowed researchers to infer changes over higher resolution intervals. In some cases, when annual laminations or varves are present, it is also possible to use calibration-in-time instead of calibration-in-space methods to quantitatively infer past variables, such as temperature (Larocque-Tobler et al., 2011).

Paleoclimate

Given the present period of rapid climate change, long-term (hundreds to thousands of years), high-temporal resolution (annual to centennial), quantitative records of climate are required to determine natural variability and trends in climate. Such records will help to test climate models and better predict future climate scenarios. There is also an increasing need to understand the impact of climate warming on aquatic systems in order to better predict changes to water resources.

Biogeographers have traditionally used the strong links between vegetation and climate to infer past climates, using tree rings and fossil pollen. In the last few decades, paleolimnological proxies, including diatoms and chironomids, have provided new opportunities to reconstruct paleoclimates and the impacts of climate change on aquatic systems. Inference models for temperature have been developed from diatoms (Bloom et al., 2003; Lotter et al., 1997) and chironomids (Luoto, 2009; Walker and Cwynar, 2006). However, it has been suggested that chironomid-inference models are more accurate than diatom-based models (Battarbee, 2000; Bloom et al., 2003). The reconstruction of past temperatures (both air and water) from subfossil chironomid assemblages has contributed to our understanding of past climate change, which should help us to more accurately predict the future.

One of the most studied and debated climate periods of the Quaternary is the Younger Dryas oscillation, a cold period that occurred ~13–11.5 calendar ka and punctuated a general warming trend that began at the end of the last glaciation (Broecker et al., 2010). The Younger Dryas was first reported from the North Atlantic and was hypothesized to be the result of cold glacier meltwater slowing oceanic circulation (Broecker et al., 1989). Regional temperature records of the Younger Dryas interval based on chironomid-inference models for temperature and other proxies help us understand whether changes in the North Atlantic were globally teleconnected. For example, chironomid records show that California was 4 °C colder than immediately prior to the YD and that this cold period was followed by rapid warming (Figure 5; MacDonald et al., 2008). Based on these data, a teleconnection was

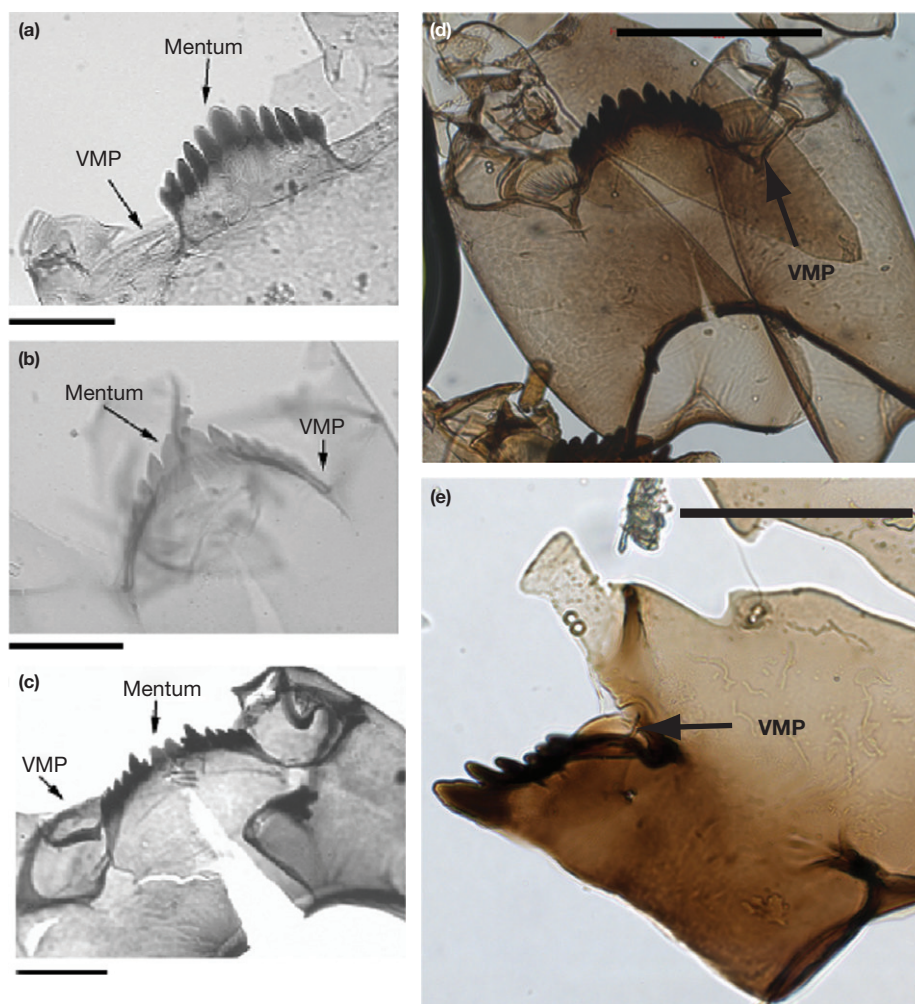


Figure 2 Chironomid photomicrographs illustrating different ventromental plate and mandibular structures of subfossil chironomid remains recovered from the Sierra Nevada, United States. VMP, ventromental plate. (a) *Tanytarsina* (Chironominae: Tanytarsini) has an elongate bar or sausage-shaped, striated VMP and a flat mentum with one median tooth and five lateral teeth. (b) *Psectrocladius* (*Psectrocladius*) *sordidellus*-type (Orthoclaudiinae) has a reduced, unstriated VMP, with beard present and a strongly arched mentum with two median teeth and five lateral teeth. (c) *Microtendipes* (Chironominae: Chironomini) has a striated, fan-shaped VMP and a weakly pigmented tridentate median tooth and six lateral teeth with first tooth fused to second lateral. (d) *Dicrotendipes nervosus*-type (Chironominae: Chironomini), has a striated, fan-shaped VMP with crenulate outer margin and one median tooth and six lateral teeth with the first lateral tooth larger than the rest. (e) *Heterotrissocadius marcidus*-type (Orthoclaudiinae) has a bulbous VMP and paired median teeth with a heavily pigmented gula. Photomicrographs are taken using a Nikon transmission light microscope at 20 $\times$. Scale bar = 100 μ m. Photo credit: D.F. Porinchu.

suggested between the Atlantic and Pacific oceans. Chironomid and pollen records from Chile support this idea (Massaferro et al., 2009). However, chironomid-based temperature records from other parts of the world (e.g., Tasmania, Australia, and Yukon, Canada) indicate no temperature change during this period (Kurek et al., 2009; Rees and Cwynar, 2010). By developing a global network of sites that investigate temperature variations during this important time period and others of similar importance, it is possible to gain insights into the mechanisms driving these rapid climate change events and their translation across the globe.

Although diatoms have been used less frequently than chironomids to determine past records of temperature, diatom-based research provides striking evidence of the impacts of recent warming on aquatic systems. Research from the Arctic

has shown dramatic increases in diatom diversity (Figure 6; Douglas et al., 1994; Smol et al., 2005) that are unprecedented for at least the last 200 000 years (Axford et al., 2009). Evidence clearly links these changes to recent warming (Smol, 2010). In lakes from the boreal forest, there are coincident changes in diatom community composition, which are related to the effects of warming temperatures on ice cover, growing seasonal and thermal stratification, and the internal cycling of nutrients (Rühland and Smol, 2003). In a study from Europe, Clarke et al. used a top-bottom approach to determine changes in many lakes in response to recent warming (Clarke et al., 2005). The top-bottom approach compares a single modern sample (top 0–1 cm of sediment) to a sample representing pre-1850 (single bottom sample, usually determined to be pre-1850 using ^{210}Pb dating), allowing many lakes to be evaluated

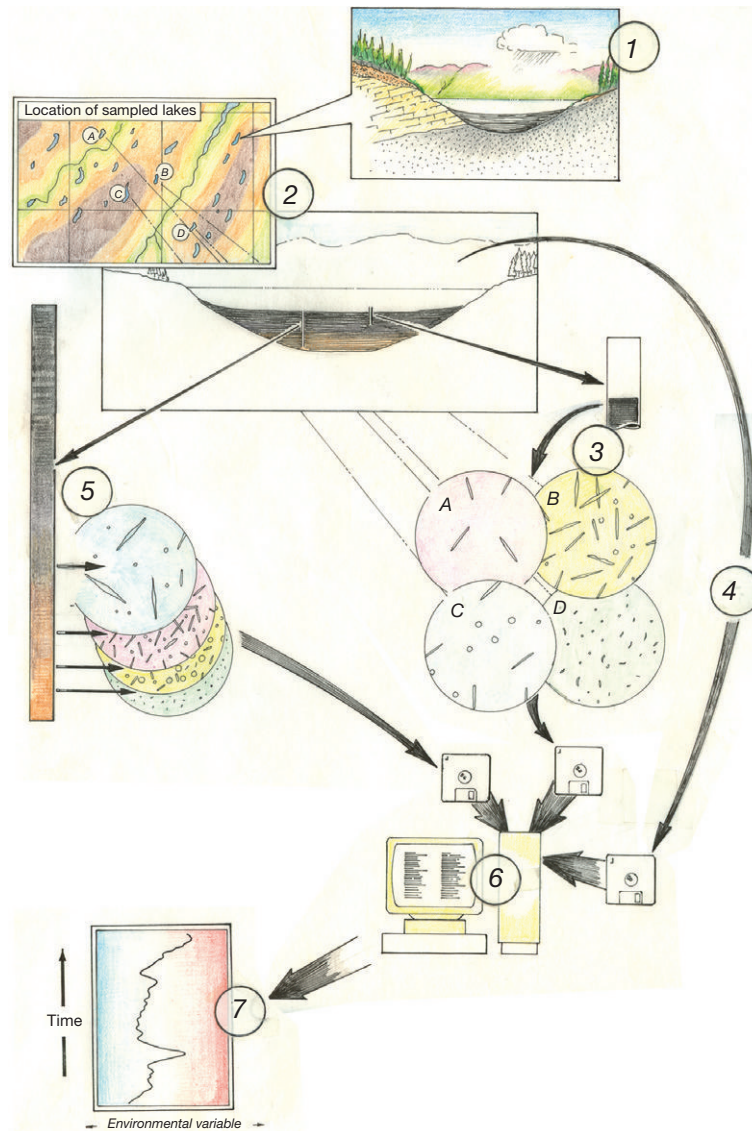


Figure 3 The paleolimnological approach. (1) Lake sediments provide a holistic record of environmental change because they receive inputs from atmospheric, aquatic, and terrestrial systems. The sediments are laid down year after year and can be dated using a variety of dating techniques. (2) Past environmental variables can be estimated using the modern relationships between species assemblages and environmental variables from a suite of lakes, called a calibration or lake training set. (3) To determine a modern diatom assemblage, the upper 0–1 cm of sediment is collected using a gravity corer. This sediment sample represents a spatially and temporally integrated sample of the last few years. These samples are processed following standard techniques and the proxies are identified and enumerated. Information is noted about each lake, including morphometry, Secchi disc depth, temperature, pH, conductivity, and numerous other water variables (e.g., nutrients, metals, ions, etc.). (5) Records of environmental change are retrieved from a subset of lakes. A variety of different coring techniques are available and selected depending on the sediment characteristics and lake depth. These cores are subsampled and proxies are enumerated in the same manner as for the sediment surface samples. (6) Numerical analyses are used to study relationships between the modern proxies (3) and environmental data (4) in order to develop transfer functions (Figure 4) to quantitatively infer past environmental variables from fossil data (5). (7) Environmental variables can be inferred for thousands of years in the past and plotted to show past environmental change. Reproduced from Moser KA (2004) Paleolimnology and the frontiers of biogeography. *Physical Geography* 25: 453–480.

quickly. Clarke et al.'s (2005) research showed that modern day diatom community composition had changed markedly from pre-industrial communities. They analyzed diatoms in over 209 lakes from 11 different districts and found that the ratio of planktonic to benthic diatoms had increased in 7 of the 12 districts. They hypothesized that these changes were the result of temperature-related changes in nutrients, ice cover, and erosion (Figure 7). More recently, Rühland et al. (2008)

combined data from several data sets and demonstrated that changes in diatom community composition in response to recent warming may be global in scale.

In closed lake systems, increased aridity results in lower lake levels and increased salinity, which affect diatom community composition (see [Salinity and Climate Reconstructions from Continental Lakes](#)). Therefore, diatom-inference models for salinity and salinity-related variables (Fritz et al., 2010) and

lake depth (Wolin and Stone, 2010) have been used to infer past drought. On the Great Plains, paleolimnological research has indicated that during the last two millennia, droughts were much more severe than the worst droughts experienced during the historical record (Laird et al., 1996). Even longer records, spanning the entire Holocene, show that the early to mid-Holocene intervals were even drier than the last 2000–3000 years (Clark et al., 2002; Laird et al., 1996). This research shows that records spanning the last century are too short to

accurately represent the full range of the natural variability of drought on the Great Plains and thus are inadequate for reconstructing drought severity during the past, or for predicting the future (Clark et al., 2002).

Similar approaches are now being used in alpine areas of the southwestern United States (Bloom et al., 2003). Here, changes in lake depth and salinity follow steep elevational gradients, and lake calibration sets from across these gradients have provided data to develop diatom-inference models. These models, applied to fossil diatom assemblages from the Younger Dryas chronozone, show marked changes in effective moisture (precipitation minus evaporation) during this period of rapid temperature change (Figure 5; MacDonald et al., 2008). This research indicates that the Younger Dryas was not only a period of rapid temperature variability but was also marked by hydrologic change.

Finally, as dating techniques (e.g., amino acid racemization, U-series dating) are developed for the period not covered by ^{14}C dating (prior to 40 000 years ago), there is great potential for determining long records (hundreds of thousands to millions of years) of regional climate change from ancient lakes. For example, records from Bear Lake, Utah (USA; Kaufman et al., 2009) and Lake Ohrid (Macedonia/Albania) (Reed et al., 2010), used diatoms and other proxies to determine regional climate change and link these regional records to global records of glacial–interglacial cycles. Such records provide data critical for understanding global climate connections past and future.

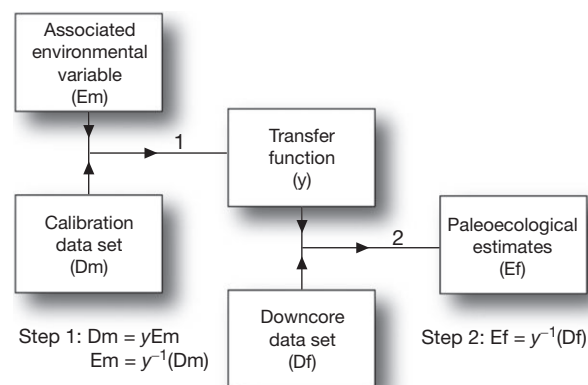


Figure 4 Schematic diagram showing that a transfer function is a mathematical relationship established between modern species (calibration data set) and environmental data (associated environmental variables). This mathematical relationship can then be applied to fossil data (downcore data) to infer past environmental variables (paleoecological estimates). Reproduced from Bradley RS (1983) *Paleoclimatology: Reconstructing Climates of the Quaternary*. San Diego: Harcourt-Academic Press.

Terrestrial–Aquatic Links: Disturbance and Succession

In recent years, with the recognition of the profound impact of human activities on biogeochemical cycles (Schlesinger,

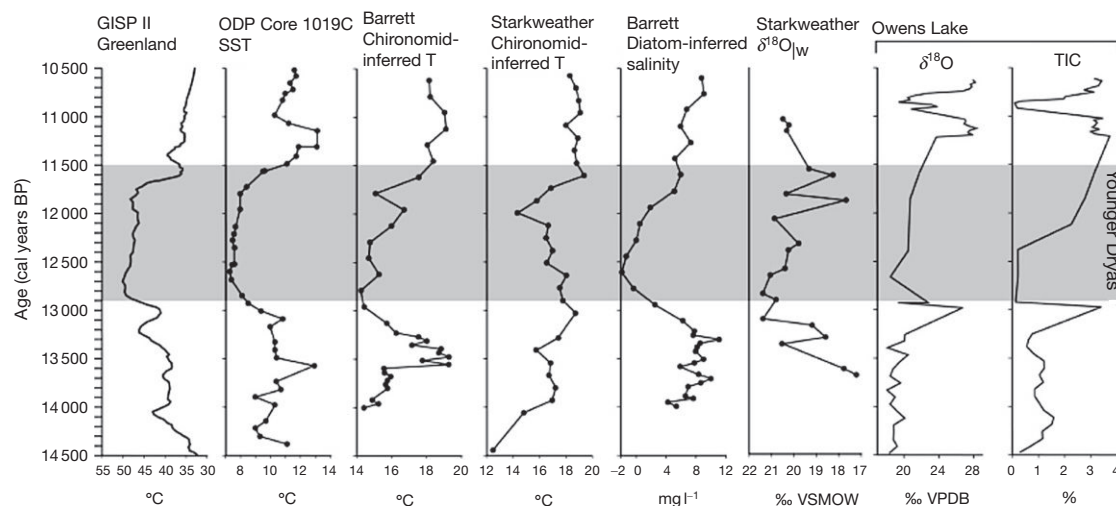


Figure 5 Graphic representing climate changes during the Younger Dryas chronozone determined from Sierra Nevada lake sediments. Chironomid-inferred temperatures for two lakes, Barrett and Starkweather, are compared to oxygen isotope-derived temperature records from GISP II (Greenland) ice cores (Alley, 2000) and sea surface temperature inferred from ODP CORE 1019 C off the coast of northern California (Barron et al., 2003). Diatom-inferred salinity from Starkweather Lake, which tracks changes in effective moisture (precipitation–evaporation), shows that the Younger Dryas was characterized by first wet and then dry conditions. This is corroborated by the cellulose $\delta^{18}\text{O}$, also from Starkweather Lake, and $\delta^{18}\text{O}$ and TIC from nearby Owens Lake (Benson et al., 1997, 2002). The shaded band shows the timing of the Younger Dryas chronozone. Reproduced from MacDonald GM, Moser KA, Bloom AM, et al. (2008) Evidence of temperature depression and hydrological variations in the eastern Sierra Nevada during the Younger Dryas stage. *Quaternary Research* 70: 131–140.

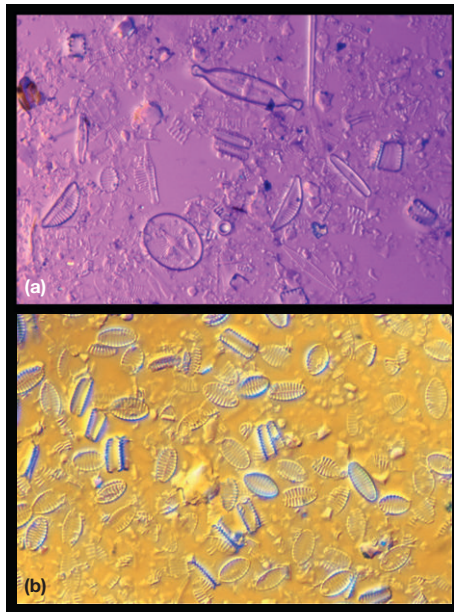


Figure 6 Photomicrographs showing recent, marked changes in diatom biodiversity, typical of arctic lakes. (a) Represents modern, diverse diatom assemblages and (b) Represents nearly monospecific communities typical of pre-1850 assemblages. This change is caused by recent and rapidly warming temperatures, which are resulting in shorter ice cover duration and greater habitat availability (Douglas et al., 1994).

1997), there has been increased interest in the development of a better understanding of the flux of chemicals between the terrestrial, aquatic, and atmospheric systems. One way that humans are altering biogeochemical cycles is through landscape change, but dramatic landscape change can also occur due to natural phenomena such as climate change, fire, disease, wind, and flooding. Natural and anthropogenic landscape changes may result in a transfer of nutrients and sediments from the terrestrial to aquatic systems, potentially altering these ecosystems. Biogeographic research has determined vegetation succession following natural disturbances and human-caused landscape change and paleolimnology provides the opportunity to determine the effects of vegetation and soil succession on aquatic systems.

Early biogeographic research hypothesized that after a disturbance, vegetation succession followed predictable, unidirectional pathways and that the vegetation of earlier stages altered the environment to allow for the establishment of subsequent vegetation (Clements, 1936). Similarly, aquatic succession was suggested to follow a unidirectional pathway of increasing eutrophication and eventual infilling of lakes and ponds with sediments (Deevey, 1942; Rohde, 1969). Recent research from Glacier Bay, Alaska (USA), used paleolimnological methods to show that this model of aquatic succession is oversimplistic, and that aquatic succession can follow many different pathways (Engstrom et al., 2000). Using fossil diatoms and a chronostratigraphic approach to follow lake development after deglaciation, Engstrom et al. (2000) found that lakes became more dilute and acidic over time, accumulated DOC, and sometimes showed a relatively early, short-term rise in nitrogen. The inferred limnological changes were largely

attributed to vegetation and soil succession, but differences in each lake's trajectory were attributed to differences in hydrology.

The Glacier Bay research examined successional changes following deglaciation; however, soil and vegetation succession also follow other natural disturbances, including, fire, windstorms, and outbreaks of pathogens. Such catchment changes are hypothesized to impact limnological conditions through changes in geochemical pathways. For example, deforestation can have a pronounced effect on a lake ecosystem through increasing runoff from the catchment that is enriched in dissolved nutrients and particulate matter (Likens, 1985). Paleolimnological research has determined a wide range of lake responses to vegetation succession. Hall and Smol (1993) showed that the rapid demise of hemlock ~4800 ¹⁴C years BP in eastern North America resulted in a minor increase in lake trophic status. Some investigators have reported an increase in nutrient export to lakes following fires (Enache and Prairie, 2000; Wright, 1976), whereas other researchers have observed no change in nutrient inputs (Paterson et al., 1998). Paleolimnological studies have also indicated reductions in lake water acidity due to increased erosion and export of base cations from the catchment (Korhola et al., 1996; Rhodes and Davis, 1995). Undoubtedly, the different lake responses that have been observed are related to the initial conditions of the lake, including the catchment characteristics and the lake chemistry, as well as the severity of the fire. Studies of the natural links between aquatic and terrestrial systems provide a context for understanding the impacts of human-caused landscape change on lake ecosystems.

In Europe, the landscape has been modified and shaped for thousands of years by human activities, including forest clearance and agriculture (Birks, 1988). Land use change can directly affect lake water quality through modification of the hydrology, nutrient loading, and increased export of other material from the catchment (Dillon and Kircher, 1975; Harper and Stewart, 1987). Similar to investigations of natural deforestation, research shows a wide range of lake responses to human deforestation. For example, in Switzerland, Lotter (2001) used pollen to show that major land clearance occurred around Soppensee during the Bronze and Iron Age, although, a diatom-inferred TP reconstruction indicated that nutrient enrichment only occurred within the last 50 years. At Diss Mere, England, however, nutrient enrichment inferred from diatoms was linked to Neolithic, Bronze Age, and Iron Age forest clearance (Birks et al., 1995). The impact of land use change will depend on the extent of deforestation, grazing pressure, tillage, and fertilizer use, as well as the starting conditions of the lake itself (Heathwaite, 1997).

In North America, research has tended to focus on the impacts of land clearance by Europeans; however, pre-European communities also altered the landscape and, in turn, water quality. For example, in southern Ontario, land clearance by the Iroquois for settlement and agriculture between AD 1268 and 1486 led to an increase in nutrient inputs, which fundamentally and permanently altered lake ecosystems from natural baseline conditions (Figure 8; Ekdahl et al., 2007). This research indicates that initial perturbations by humans may significantly affect how a lake recovers from subsequent disturbance, which has important implications for water management,

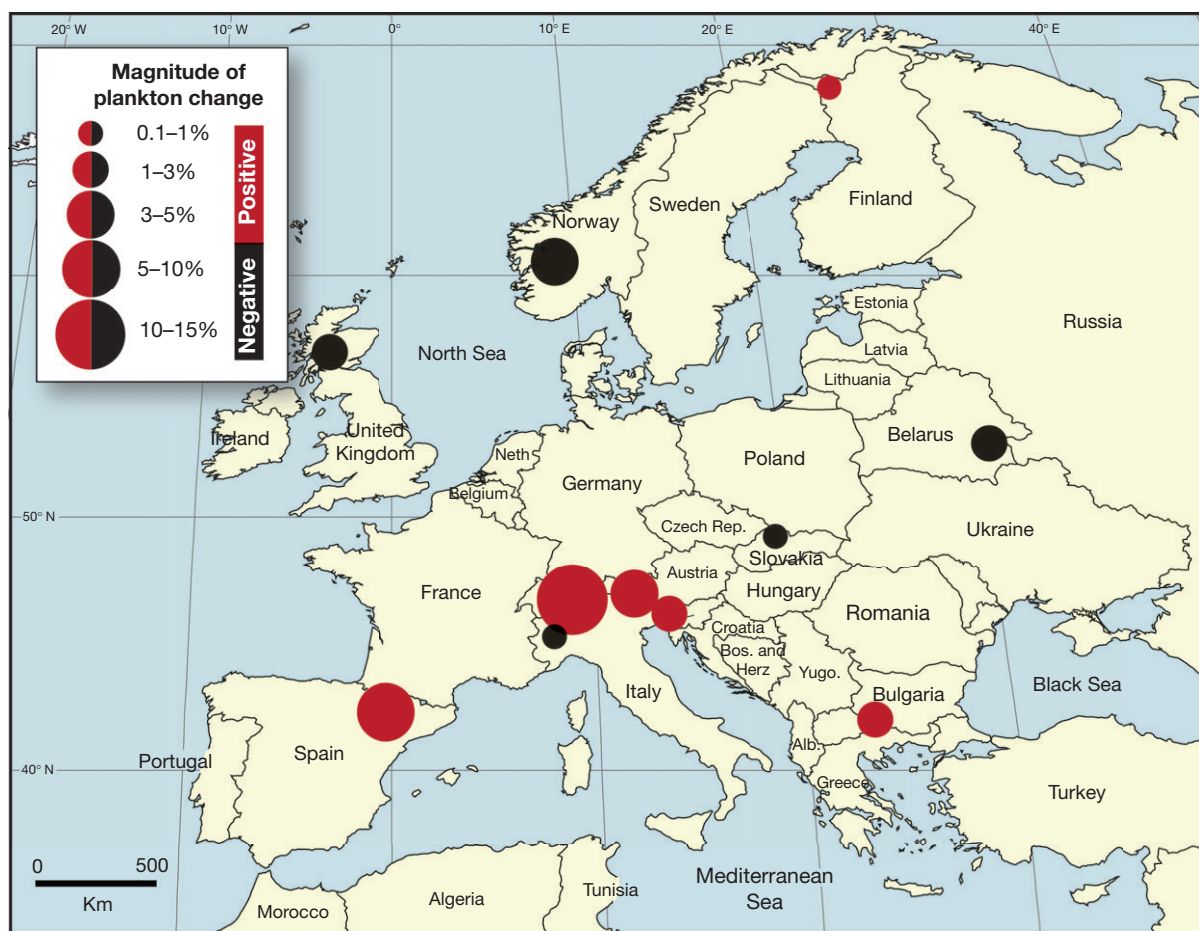


Figure 7 Map showing changes in the ratio of planktonic to benthic diatoms in 11 lake districts from across Europe. Note that the largest positive changes are reported from alpine lakes. Reproduced from Clarke G, Kernan M, Marchetto A, Sorvari S, and Catalan J (2005) Using diatoms to assess geographical patterns of change in high-altitude European lakes from pre-industrial times to the present day. *Aquatic Sciences* 67: 224–236.

particularly where prehistoric populations were larger and people less nomadic.

Owing to lower population densities of early peoples, the magnitude of change in water quality caused by landscape changes is generally much less than that caused by recent humans. Changes in diatom community composition show dramatic increases in nutrient inputs to lakes resulting from human activities, including deforestation, land clearance and agriculture, road construction, and urbanization (Smol, 2008). In some studies, changes in chironomid community composition have also shown a concomitant decrease in oxygen following increased nutrients and algal productivity (Clerk et al., 2000; Little et al., 2000). Regional assessments of eutrophication using diatom-based approaches are also possible. In Minnesota, a diatom-based transfer function was used to infer pre-European trophic status in comparison to post-settlement trends in 55 lakes (Ramstack et al., 2004). The results indicated that 30% of urban and 30% of agricultural lakes showed a statistically significant increase in phosphorus between pre-settlement levels and 1970.

The historical perspective provided by these records has important implications for determining restoration goals. For example, in Europe, the EU Water Framework Directive

has a goal to achieve a chemical and ecological level comparable to background conditions for all surface waters by the year 2015 (Räsänen et al., 2006). Here, background conditions are considered to be those existing prior to intensive farming or forestry and before large-scale industrial disturbances, and can be determined by paleolimnological reconstructions. However, if prehistoric peoples did impact lake water quality, an earlier time period may provide a more accurate representation of baseline conditions (Bradshaw et al., 2006; Räsänen et al., 2006).

Impacts of Atmospheric Pollution on Water Quality

Pollution resulting from industrial activities has caused acid rain and increased metal and nutrient emissions to the atmosphere. For example, it has been reported that industrial emissions of Pb, Cd, and Zn exceed the flux from natural sources by factors of 18, 5, and 3, respectively, and that for As, Hg, Ni, Sb, and V, the contributions from human sources may exceed those from natural sources by 100–200% (Nriagu, 1989). Emissions of nutrients, in particular nitrogen, have also increased as a result of the production of synthetic fertilizers, fossil fuel burning, and increased cultivation of leguminous

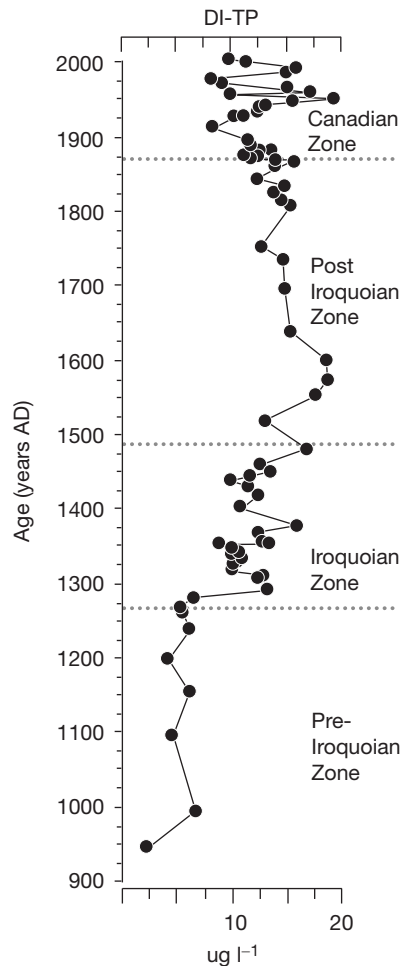


Figure 8 Diatom-inferred total phosphorus for Crawford Lake, Milton, Ontario (Canada), based on a 64-lake calibration set from southern Ontario (Reavie and Smol, 2001). Note that phosphorus increases with the onset of agriculture by the Iroquois and never returns to pre-Iroquoian levels. The greatest increase in phosphorus, however, occurs with European settlement of the catchment. The most recent decrease in phosphorus may indicate a recovery with reforestation. Reproduced from Ekdahl EJ, Terranes JL, Wittkop CA, Stoermer, EF, Reavie ED, and Smol, JP (2007) Diatom assemblage response to Iroquoian and Euro-Canadian eutrophication of Crawford Lake, Ontario, Canada. *Journal of Paleolimnology* 37: 233–246. Top photo shows reconstructed Iroquois village at Crawford Lake (bottom photo). Photo credit K.A. Moser.

crops (Schlesinger, 1997). Anthropogenic production of fixed nitrogen is now creating more biologically available nitrogen than natural ecosystem processes (Hessen et al., 2009). Approximately 60% of deposited atmospheric nitrogen is from human-derived sources (Schlesinger, 1997). Biogeographers are interested in how atmospheric pollution from human activities has affected species distributions and ecosystem functioning. Using paleolimnological techniques, changes in atmospheric pollution and its impacts on ecosystems can be determined.

One of the most widely studied pollution problems in paleolimnological research has been acid rain (Battarbee et al., 2010; Smol, 2008), and it is this research that has formed the basis for paleolimnological investigations of other pollution problems (see also [Lake Chemistry](#) and [Multiproxy Approaches](#)). In the 1980s, acid rain was identified as a serious environmental problem in Europe and the United States; however, no long-term measurements were available from which to determine the impacts of acid rain on lake ecosystems. The problem of acid rain required high-quality, historical data to

determine acidification trends. Key questions that needed to be addressed were the following: (1) Were lakes acidic because of acid deposition, or were they naturally acidic? (2) Was acidification part of a natural long-term trend? (3) Which lakes were most susceptible to acidification? (4) If lakes were being acidified, then by how much? (5) Are lakes recovering from reduced acid deposition? (Smol, 2008).

However, there was little monitoring of lake water pH prior to 1980, so paleolimnology was used to provide long-term, quantitative reconstructions of lake water acidity (Battarbee et al., 1990; Charles and Whitehead, 1986). Diatom-based models were used to track changes in lake water pH in the United States, Canada, and Europe. One area of particularly intense research was Adirondack Park, in upstate New York, where many of the 3000 lakes were acidic. In the Adirondacks, paleolimnological records showed that most lakes had become more acidic beginning ca. 1900–50, coincident with increases in the deposition of strong acids, as evidenced by increases in coal and oil soot particles and polycyclic aromatic hydrocarbons, lead (Pb), and vanadium (V) (Cumming et al., 1994).

In addition, regional assessments of acidification were undertaken in Adirondack Park using the top–bottom approach (Cumming et al., 1992). Diatom-inferred pH from recent sediments was compared to pH from pre-1850 sediments from selected lakes, and the changes were mapped. The map showed that lakes that underwent the greatest changes in pH from pre-1850 to the present day were those in the southwestern part of the study area where the lakes were poorly buffered and acid deposition was greatest. In this research, paleolimnological data were also used to estimate critical load values, that is, the highest load of a substance that will not cause chemical changes resulting in long-term, harmful effects (Smol, 2008). By comparing known sulfate deposition to diatom-inferred pH, the sulfate load required to induce a change in lake water pH can be determined. Using this approach, Cumming et al. (1994) showed that many Adirondack lakes acidified with wet sulfate loads of only 5–25 kg ha⁻¹ year⁻¹, which was less than reductions proposed by legislation.

These techniques can be used to track other kinds of pollution. Alpine lakes are relatively remote and as a result are often considered to be pristine and relatively untouched by humans. Paleolimnological research has shown this perception to be largely untrue, mainly owing to the deposition of long-distance pollutants. For example, paleolimnological data have been used to track changes in metal emissions and their effects on aquatic ecosystems. Research in the Swiss Alps (Kamenik et al., 2005) and Uinta Mountains, Utah, USA (Moser et al., 2010), showed significant metal enrichment in alpine lakes related to mining, although in Europe, the increase in metals occurred during the Early Medieval period, whereas in Utah, it occurred in the mid- to late-1800s with the arrival of Europeans and onset of mining. Both of these investigations showed marked changes in diatom community composition coincident with increased Pb emissions, suggesting aquatic ecosystem change in response to metal enrichment.

Recent research in alpine areas of Colorado (USA) showed striking changes in diatom community composition, indicative of increased nutrients from the 1950s (Wolfe et al., 2001, 2003). By comparing these diatom records to nitrogen isotope records, it was argued that nutrient enrichment is occurring as a result of increased nitrogen deposition from agriculture and industry to the east of the mountains. Similar changes have now been reported from many alpine regions of North America (Holtgrieve et al., 2011).

Biodiversity

Broadly defined, biodiversity is the number, variety, and variability of living organisms. Considerable biogeographic research has been directed at understanding the spatial patterns of biodiversity, and two general categories of explanations have been proposed (MacDonald, 2003). The first category is referred to as historical theories of biodiversity, which state that modern geographical distributions are controlled by historical processes, such as allopatric speciation (i.e., formation of new species by geographic isolation) and geographic dispersal. The second category assumes that modern patterns of diversity are the result of environmental conditions and biological processes (e.g., predation, competitive exclusion, and stochastic variations). For macroorganisms, it is generally accepted that

both of these categories of processes determine species distributions and diversity (Cox and Moore, 2005). However, for microorganisms there is considerable debate. It has been argued that microorganisms have cosmopolitan distributions because they are so abundant and easily dispersed that biodiversity is only limited by species environmental tolerances (Finlay, 2002). This theory is referred to as the ubiquity hypothesis (i.e., ‘everything is everywhere’ and ‘the environment selects’; Finlay, 2002). If this hypothesis is correct, biogeographic patterns (e.g., latitudinal diversity gradients) would be absent (Vyverman et al., 2007). Other research, however, indicates that microorganism, particularly diatom, distributions show biogeographic patterns similar to macroorganism distributions, indicating that historical processes are important (Martiny et al., 2006). Recently, paleolimnological research has been contributing to this debate, and results support that historical processes are important (Vanormelingen et al., 2008).

First, lake calibration or training sets developed for determining relations between diatom community composition and environmental variables provide an excellent opportunity to establish global and regional patterns of species distributions and diversity to test the ubiquity theory (Telford et al., 2006; Vyverman et al., 2007). By combining several lake training sets to develop a global lake training set, Vyverman et al. (2007) showed that variation in both regional and global genus richness was more strongly influenced by geographic variables such as connectivity between lakes, degree of lake district isolation, and habitat availability than by local environmental variables, such as pH and conductivity, implying that historical processes are important in determining diatom distributions.

Secondly, paleolimnology can be used to track the introductions of exotic species by humans, which further provides strong evidence that natural dispersal of diatom species is limited by geographic barriers. Documentary evidence of diatom introductions is rare; however, the paleolimnological record provides an opportunity to determine the history of species introductions and occurrences (Willis and Birks, 2006). For example, sediment records in New Zealand showed that prior to 1880, *Asterionella formosa*, a widespread, planktonic diatom, was absent (Vanormelingen et al., 2008). Today, this diatom is widespread, occurring in 45% of lakes with known phytoplankton records and it is suspected that this diatom arrived in New Zealand with the introduction of salmon ova. These results show that there were barriers to the expansion of *A. formosa* to New Zealand, and it was only when its dispersal was facilitated by humans that its range expanded.

Invasive Species

Biogeographers are interested in changes in species ranges, and dispersal and establishment are important controls on a species distribution. As more and more humans have traveled greater and greater distances, humans have become increasingly important vectors of dispersal of both micro- and macroorganisms. The introduction of exotic species by humans has been both purposeful and accidental, but regardless, the introduction of non-native species can have devastating impacts on ecosystems, economy, and human health (Vitousek et al.,

1997). Paleolimnology can provide a long-term context to evaluate the consequences of species introductions.

The Great Lakes, which has been referred to as an 'underwater zoo,' provides many examples of species invasions (Smol, 2008). At least 139 species have been introduced to the Great Lakes since the 1800s, of these 24 were algal species and of these 16 were diatom taxa (Mills et al., 1993). Probably all of the diatom species were introduced from Europe or Eurasia through ship ballast waters (Mills et al., 1993). Owing to limitations in monitoring the expansive Great Lakes, our capacity to track the arrival and proliferation of exotic species is quite restricted (Edlund et al., 2000).

Paleolimnological techniques provide opportunities to determine the timing and spread of invasive species and to understand their repercussions. Stoermer et al. (1985) used a dated sediment core from Lake Ontario to track the arrival and expansion of the diatom *Stephanodiscus binderanus*. This research showed that this species arrived in the Great Lakes in 1952, and Reavie et al. (1998) found a peak in this diatom correlating to the 1950s in sediment cores from the St. Lawrence River. This timing is similar to observations from Montreal (Brunel, 1956) and Chicago (Vaughn, 1961). More recently, Edlund et al. (2000) showed that the exotic diatom, *Thalassiosira baltica*, was present in Lake Ontario by 1988, even though it was not reported until 2000. In 1993, mandatory ballast water exchange, which was thought to greatly decrease the introduction of invasive species, was implemented. Because the 'observational' arrival date of 2000 occurs after 1993, the reported entry of *T. baltica* could have been used as evidence that the ballast water exchange is ineffective; however, the paleolimnological data shows that it arrived prior to mandatory ballast exchange, negating this suggestion (Smol, 2008).

Although humans have inadvertently introduced many taxa, they have also purposely introduced certain species. In freshwaters, this has typically been fish, and although fish in lakes may always seem beneficial, adding fish to naturally 'fishless' lakes alters food webs and communities that have developed over thousands of years; however, the practice of fish stocking remains widespread. It is estimated that nonnative species have been introduced in 60% of all western US mountain lakes since intensive stocking of sport fishes began in the mid-1800s (Bahls, 1992). In the Canadian Rocky Mountains, nonnative trout have been introduced to at least 20% of all lakes (Donald, 1987). Most of the time, stocking was done without collecting baseline (i.e., pre-introduction) data, and follow-up monitoring is rare (Smol, 2008). Paleolimnological data can be used to reconstruct the missing data. An example of this comes from the Canadian Rocky National Parks, where fish stocking was prevalent since the turn of the century (Lamontagne and Schindler, 1994). For present-day park managers, it was almost impossible to say whether fish communities in a lake were native, recently introduced, or a combination of the two. Three lakes in Jasper Park, which were stocked with salmonids in the 1920s and 1930s, were selected because it was unclear what their 'fish status' was prior to stocking. In order to restore the Jasper lakes, 'target conditions' were required (Smol, 2008). Using methods established in eastern North America, *Chaoborus* mandibles preserved in dated sediment cores were used to determine baseline

conditions. *Chaoborus americanus* does not survive in lakes with fish because its instars are particularly susceptible to predation by fish, and *C. trivittatus* is vulnerable to fish predation (Uutala, 1990). These two species are, therefore, indicative of 'fishless' lakes. Paleolimnological analyses of sediment cores from the three stocked Jasper lakes revealed that these two species were present in prestocking sediments from two of the lakes, indicating that these lakes were naturally fishless. This research provided park managers with critical information for successful restoration of Jasper lakes.

Extinctions and Evolution

Species evolution and extinction have occurred throughout much of Earth's history, usually at steady and gradual rates, but punctuated by periods of rapid extinction (MacDonald, 2003). Some scientists believe that the rapid rates of extinction occurring today represent a mass extinction event (Wilson, 1989). Therefore, efforts to understand factors controlling extinction and evolution are of great interest to biogeographers. Recent paleolimnological research has investigated both, but with particular attention to microorganisms.

Extinctions or extirpations (i.e., local extinctions) can occur for many reasons, including resource limitation and the introduction of exotic species. In paleolimnological research from the Great Lakes, the diatom *Stephanodiscus niagarae* was in serious decline by 1980 and had disappeared from Lake Ontario by 1988 (Julius et al., 1998). The obvious explanation for the loss of this diatom species was the continued and varied environmental stresses, including pollution and invasive species. However, a second explanation has since been suggested. By measuring valve size in sediment records, Stoermer et al. (1989) and Julius et al. (1998) showed that the mean diameter of *S. niagarae* valves began to decrease in the late 1940s to early 1950s. They hypothesized that valve size decreased owing to intense silica depletion due to increased phosphorus from pollution. Increased phosphorus led to intense diatom blooms that depleted silica resources. Eventually, diatom valve size fell below the minimum required for cells to develop.

Sediment records have also been used to study extirpations of larger organisms, such as fish. Similar to the North American Great Lakes, large lakes in Africa have recently endured multiple stresses, including human population growth, increased cultivation of land, meteorological variability, resource extraction, intensive fishing, introduction of exotic species, and more recently, climate change. In Lake Victoria, the world's largest tropical lake, almost half of its 500+ species of endemic cichlid species have gone extinct (Verschuren et al., 2002). There are two hypotheses to explain this loss. The first is that the introduction of Nile perch resulted in top-down effects to Lake Victoria's food web, which led to a loss of cichlids and increased primary productivity, and the second is that landscape change, particularly intensification of agriculture, resulted in increased productivity, which resulted in bottom-up changes (Hecky et al., 2010). The latter hypothesis argues that increased productivity led to increased inshore turbidity. Because cichlids are highly dependent on visibility for feeding and reproduction, they are less competitive in turbid waters. If the former hypothesis is correct, then primary productivity should have increased after the Nile perch populations expanded in the

1980s. Owing to a paucity of records of limnological variables for Lake Victoria, the paleolimnological record has been used to understand the loss of cichlid biodiversity. Sediment records clearly show an increase in eutrophic diatoms and increases in other measures of productivity beginning in the late 1960s. These records indicate that increased primary productivity (caused by increased phosphorus from growing populations and intensified agriculture) predates the transition to Nile perch dominance. It is, therefore, suggested that eutrophication alone can lead to extinctions and loss of biodiversity. The management implications of this research are that increasing fishing mortality of Nile perch to reverse changes in water quality will fail and the only way for any type of recovery is through decreases in phosphorus loading.

Although paleolimnological data have been used less frequently to study evolution than extinction, there is great potential for tracking evolutionary change and testing evolutionary hypotheses using lake sediment records. An example from Yellowstone Lake, Wyoming, tests the idea that large population sizes, high dispersal rates, and low reproductive barriers result in gradual rates of protist evolution (Theriot et al., 2006). *Stephanodiscus yellowstonensis*, an endemic diatom to Yellowstone Lake, was known to have evolved from *Stephanodiscus niagarae*, which does not occur in Yellowstone Lake today (Theriot, 1992). A sediment record from Yellowstone Lake revealed that dramatic morphological changes from *S. niagarae* to *S. yellowstonensis* were coincident with rapid warming and environmental change following the end of the last glaciation, as evidenced by diatom, pollen and geochemical records from the same lake sediments. These results showed that diatom evolution can occur more rapidly (one species evolved to another in ~2 ka) than previously shown and that evolutionary change in this example is directionally driven rather than random. Using similar approaches, there is great potential for addressing questions about microorganism evolution using the particularly long sediment records spanning millions of years that are preserved in ancient lakes.

Conclusion

In the past 20–30 years, paleolimnology has changed from a qualitative to a mainly quantitative science reflecting rapid technical and statistical advances. As a result, paleolimnologists have been able to effectively use fossils from diatoms, chironomids, and *Chaoborus* preserved in Quaternary lake sediments to contribute to our understanding of biogeographic issues. Using the paleolimnological approach, which uses mathematical relationships between modern assemblages and environmental variables, it has been possible to determine long-term records of temperature and drought, which provide data necessary for accurately forecasting future climate change and its impacts. Some researchers have also used paleolimnology to show how recent warming affects aquatic ecosystems around the world. Paleolimnology has furthered our understanding of the links between the terrestrial and aquatic systems and shown that previous ideas about unidirectional change and lake succession have been oversimplified. Investigators have reported the significant impact of human activities on lake water quality, both directly in the catchment and

remotely via long-distance transport of pollutants. For example, metals and nutrients derived from mining activities, agriculture, and industry are dramatically altering alpine lake ecosystems. Paleolimnological data are also being used to track the entry and spread of invasive species, which are difficult to monitor in large lake systems, such as the Great Lakes of North America and Africa. Research on invasive species provides managers with the necessary information for successful restoration and conservation practices. Finally, paleolimnological research is being used to test hypotheses about biodiversity, extinction, and evolution, particularly with respect to microorganisms. Paleolimnology has made significant contributions to biogeography but also to our understanding of the significant impacts of humans on aquatic ecosystems. Perhaps most optimistically, this review shows that paleolimnological research offers a historical perspective that provides managers and policy makers with directions for successfully protecting, conserving, and restoring Earth systems.

See also: Dating Techniques; Diatom Introduction; Paleoclimate Relevance to Global Warming; Varved Lake Sediments. **Chironomid Records:** Chironomid Overview. **Diatom Methods:** $\delta^{18}\text{O}$ Records; Diatomites: Their Formation, Distribution, and Uses; Salinity and Climate Reconstructions from Continental Lakes. **Diatom Records:** Large Lakes. **Glaciation, Causes:** Tectonic Uplift and Continental Configurations. **Lake Level Studies:** Overview. **Paleobotany:** Overview of Terrestrial Pollen Data. **Paleoclimate:** Introduction; The Younger Dryas Climate Event; Timescales of Climate Change. **Paleoclimate Modeling:** Data–Model Comparisons. **Paleoclimate Reconstruction:** Younger Dryas Oscillation, Global Evidence. **Paleolimnology:** Multiproxy Approaches; Overview of Paleolimnology.

References

- Alley RB (2000) The Younger Dryas cold interval as viewed from central Greenland. *Quaternary Science Reviews* 19: 213–226.
- Appleby PG (2001) Chronostratigraphic techniques in recent sediments. In: Last WM and Smol JP (eds.) *Tracking Environmental Change Using Lake Sediments: Basin Analysis, Coring and Chronological Techniques*, vol. 1, pp. 171–204. Dordrecht: Kluwer Academic Publishers.
- Axford Y, Briner JP, Cooke CA, et al. (2009) Recent changes in a remote arctic lake are unique within the past 200,000 years. *Proceedings of the National Academy of Sciences of the United States of America* 106: 18443–18446.
- Bahls PF (1992) The status of fish populations and management of high mountain lakes in the western United States. *Northwest Science* 66: 183–193.
- Barron JA, Heusser L, Herbert T, and Lyle M (2003) High-resolution climatic evolution of coastal northern California during the past 16,000 years. *Paleoceanography* 18: 1020–1034.
- Battarbee RW (2000) Palaeolimnological approaches to climate change, with special regard to the biological record. *Quaternary Science Reviews* 19: 107–124.
- Battarbee RW, Charles DF, Bigler C, Cumming BF, and Renberg I (2010) Diatoms as indicators of surface-water acidity. In: Smol JP and Stoermer EF (eds.) *The Diatoms: Applications for the Environmental and Earth Sciences*, 2nd edn., pp. 174–185. Cambridge: Cambridge University Press.
- Battarbee RW, Mason J, Renberg I, and Talling JF (eds.) (1990) *Paleolimnology and Lake Acidification*. London: The Royal Society.
- Benson LV, Burdett JW, Lund SP, Kashgarian M, and Mensing S (1997) Nearly synchronous climate change in the Northern Hemisphere during the last glacial termination. *Nature* 388: 263–265.
- Benson L, Kashgarian M, Rye R, et al. (2002) Holocene multidecadal and multicentennial droughts affecting Northern California and Nevada. *Quaternary Science Reviews* 21: 659–682.

- Birks HJB (1988) Introduction. In: Birks HH, Birks HJB, Kaland PE, and Moe D (eds.) *The Cultural Landscape – Past, Present and Future*, pp. 179–187. Cambridge: Cambridge University Press.
- Birks HJB (1998) D.G. Frey and E.S. Deevey review #1: Numerical tools in paleolimnology: Progress, potentialities and problems. *Journal of Paleolimnology* 20: 307–322.
- Birks HJB (2010) Numerical methods for the analysis of diatom assemblage data. In: Smol JP and Stoermer EF (eds.) *The Diatoms: Applications for the Environmental and Earth Sciences*, 2nd edn., pp. 23–56. Cambridge: Cambridge University Press.
- Birks HJB, Anderson NJ, and Fritz SC (1995) Post-glacial changes in total phosphorus at Diss Mere, Norfolk inferred from fossil diatom assemblages. In: Patrick ST and Anderson NJ (eds.) *Ecology and Paleoecology of Lake Eutrophication, an Informal Workshop*, pp. 48–49. Copenhagen: Geological Survey of Denmark and Greenland.
- Björck S and Wohlfarth B (2001) ^{14}C chronostratigraphic techniques in paleolimnology. In: Last WM and Smol JP (eds.) *Tracking Environmental Change Using Lake Sediments: Basin Analysis, Coring and Chronological Techniques*, vol. 1, pp. 205–246. Dordrecht: Kluwer Academic Publishers.
- Bloom AM, Moser KA, Porinchu DF, and MacDonald GM (2003) Diatom-inference models for surface-water temperature and salinity developed from a 57-lake calibration set from the Sierra Nevada, California, USA. *Journal of Paleolimnology* 29: 235–255.
- Bradshaw EG, Nielsen AB, and Anderson NJ (2006) Using diatoms to assess the impacts of prehistoric, pre-industrial and modern land-use on Danish lakes. *Regional Environmental Change* 6: 17–24.
- Brodersen KP and Quinlan R (2006) Midges as palaeoindicators of lake productivity, eutrophication and hypolimnetic oxygen. *Quaternary Science Reviews* 25: 1995–2012.
- Broecker WS, Denton GH, Edwards RL, Cheng H, Alley RB, and Putnam AE (2010) Putting the Younger Dryas cold event into context. *Quaternary Science Reviews* 29: 1078–1081.
- Broecker WS, Kennett JP, Flower BP, et al. (1989) Routing of meltwater from the Laurentian ice-sheet during the Younger Dryas cold episode. *Nature* 341: 318–321.
- Brooks SJ (2006) Fossil midges (Diptera: Chironomidae) as palaeoclimatic indicators for the Eurasian region. *Quaternary Science Reviews* 25: 1894–1910.
- Brunel J (1956) Addition du *Stephanodiscus binderanus* à la flore diatomique de l'Amérique du Nord. *Le Naturaliste Canadien* 83: 89–95.
- Charles DF and Whitehead DR (1986) The PIRLA project: Paleoeological investigation of recent lake acidification. *Hydrobiologia* 143: 13–20.
- Clark JS, Grimm EC, Donovan JJ, Fritz SC, Engstrom DR, and Almendinger JE (2002) Drought cycles and landscape responses to past aridity on prairies of the northern Great Plains, USA. *Ecology* 83: 595–601.
- Clarke G, Kernan M, Marchetto A, Sorvari S, and Catalan J (2005) Using diatoms to assess geographical patterns of change in high-altitude European lakes from pre-industrial times to the present day. *Aquatic Sciences* 67: 224–236.
- Clements FE (1936) Nature and the structure of the climax. *Journal of Ecology* 24: 252–284.
- Clerk S, Hall R, and Smol JP (2000) Quantitative inferences of past hypolimnetic anoxia and nutrient levels from a Canadian Precambrian Shield lake. *Journal of Paleolimnology* 23: 319–336.
- Cohen S (2003) *Paleolimnology: The History and Evolution of Lake Sediments*. Oxford: Oxford University Press.
- Cox CB and Moore PD (2005) *Biogeography: An Evolutionary and Ecological Approach*. Malden, MA: Blackwell Publishing.
- Cumming BF, Davey K, Smol JP, and Birks HJB (1994) When did Adirondack lakes begin to acidify and are they still acidifying? *Canadian Journal of Fisheries and Aquatic Sciences* 49: 128–141.
- Cumming BF, Smol JP, Kingston JC, et al. (1992) How much acidification has occurred in Adirondack region lakes (New York, USA) since preindustrial times? *Canadian Journal of Fisheries and Aquatic Sciences* 49: 128–141.
- Deevey ES Jr. (1942) Studies on Connecticut lake sediments III. The biostratigraphy of Linsley Pond. *American Journal of Science* 240: 233–264.
- Dillon PJ and Kircher WB (1975) The effects of geology and land use on the export of phosphorus from watersheds. *Water Research* 9: 135–148.
- Donald DB (1987) Assessment of the outcome of eight decades of trout stocking in the mountain national parks, Canada. *North American Journal of Fisheries Management* 7: 545–553.
- Douglas MSV, Smol JP, and Blake W (1994) Marked post-18th century environmental change in high arctic ecosystems. *Science* 266: 416–419.
- Edlund MB, Taylor CM, Schleske CL, and Stoermer EF (2000) *Thalassiosira baltica* (Grunow) Ostenfled (Bacillariophyta), a new exotic species in the Great Lakes. *Canadian Journal of Fisheries and Aquatic Sciences* 57: 610–615.
- Ekdahl EJ, Terranes JL, Wittkop CA, Stoermer EF, Reavie ED, and Smol JP (2007) Diatom assemblage response to Iroquoian and Euro-Canadian eutrophication of Crawford Lake, Ontario, Canada. *Journal of Paleolimnology* 37: 233–246.
- Enache M and Prairie YT (2000) Paleolimnological reconstruction of forest fire induced changes in lake biogeochemistry (Lac Francis, Abitibi, Quebec, Canada). *Canadian Journal of Fisheries and Aquatic Sciences* 57: 146–154.
- Engstrom DR, Fritz SC, Almendinger JE, and Juggins S (2000) Chemical and biological trends during lake evolution in recently deglaciated terrain. *Nature* 408: 161–166.
- Finlay BJ (2002) Global dispersal of free-living microbial eukaryote species. *Science* 296: 1061–1063.
- Frey DG (1964) Remains of animals in Quaternary lake and bog sediments and their interpretation. *Archiv für Hydrobiologie, Ergebnisse der Limnologie* 2: 1–114.
- Fritz SC, Cumming BF, Gasse F, and Laird KR (2010) Diatoms as indicators of hydrologic and climate change in saline lakes. In: Smol JP and Stoermer EF (eds.) *The Diatoms: Applications for the Environmental and Earth Sciences*, 2nd edn., pp. 186–199. Cambridge: Cambridge University Press.
- Glew JR, Smol JP, and Last WM (2001) Sediment core collection and extrusion. In: Last WM and Smol JP (eds.) *Tracking Environmental Change Using Lake Sediments: Basin Analysis, Coring and Chronological Techniques*, vol. 1, pp. 73–106. Dordrecht: Kluwer Academic Publishers.
- Hall RI and Smol JP (1993) The influence of catchment size on lake trophic status during the hemlock decline and recovery (4800–3500 BP) in southern Ontario lakes. *Hydrobiologia* 269(270): 371–390.
- Harper DM and Stewart WDP (1987) The effects of land use upon water chemistry, particularly nutrient enrichment in shallow lowland lakes: Comparative studies of three lochs in Scotland. *Hydrobiologia* 143: 425–431.
- Heathwaite AI (1997) Sources and pathways of phosphorus loss from agriculture. In: Tunney H, Carton OT, Brookes PC, and Johnston AE (eds.) *Phosphorus Loss from Soil to Water*, pp. 205–223. Oxford: CAB International.
- Hecky RE, Mugidde R, Ramlal PS, Talbot MR, and Kling GW (2010) Multiple stressors cause rapid ecosystem change in Lake Victoria. *Freshwater Biology* 55: 19–42.
- Hessen DO, Anderson T, Larsen S, Skjelkvale BL, and de Wit HA (2009) Nitrogen deposition, catchment productivity, and climate as determinants of lake stoichiometry. *Limnology and Oceanography* 54: 2520–2528.
- Holtgrieve GW, Schindler DE, Hobbs H, et al. (2011) A coherent signature of anthropogenic nitrogen deposition to remote watersheds of the northern hemisphere. *Science* 334: 1545–1548.
- Julius ML, Stoermer EF, Taylor CM, and Schleske CL (1998) Local extirpation of *Stephanodiscus niagarae* (Bacillariophyceae) in the recent limnological record of Lake Ontario. *Journal of Phycology* 34: 766–771.
- Kamenik C, Koinig KA, and Schmidt R (2005) Potential effects of pre-industrial lead pollution on algal assemblages from an alpine lake. In: Jones J (ed.) *International Association of Theoretical and Applied Limnology*, vol. 29, pp. 535–538. Stuttgart: Schweizerbart Science Publishers.
- Kaufman DS, Bright J, Dean WE, et al. (2009) A quarter-million years of paleoenvironmental change at Bear Lake, Utah and Idaho. In: Rosenbaum JG and Kaufman DS (eds.) *Paleoenvironments of Bear Lake, Utah and Idaho and its Catchment*, *Geological Society of America Special Papers* 450, pp. 311–331.
- Korhola A, Virkanen J, Tikkanen M, and Blom T (1996) Fire-induced pH rise in a naturally acid hill-top lake, Finland: A palaeoecological survey. *Journal of Ecology* 84: 257–265.
- Kurek J, Cwynar LC, and Vermaire JC (2009) A late Quaternary paleotemperature record from Hanging Lake, northern Yukon Territory, eastern Beringia. *Quaternary Research* 72: 246–257.
- Laird KR, Fritz SC, Maasch KA, and Cumming BF (1996) Greater drought frequency between AD 1200 in the northern Great Plain, USA. *Nature* 384: 552–554.
- Lamontagne S and Schindler DW (1994) Historical status of fish populations in Canadian Rocky Mountain lakes inferred from subfossil Chaoborus (Diptera: Chaoboridae) mandibles. *Canadian Journal of Fisheries and Aquatic Sciences* 51: 1376–1383.
- Larocque-Tobler I, Grosjean M, and Kamenik C (2011) Calibration-in-time versus calibration-in-space (transfer function) to quantitatively infer July air temperature using biological indicators (chironomids) preserved in lake sediments. *Palaeogeography, Palaeoclimatology, Palaeoecology* 299: 281–288.
- Last WM and Smol JP (eds.) (2001) *Tracking Environmental Change Using Lake Sediments: Basin Analysis, Coring and Chronological Techniques*, vol. 1. Dordrecht: Kluwer Academic Publishers.
- Leroy SAG and Colman SM (2001) Coring and drilling equipment and procedures for recovery of long lacustrine sequences. In: Last WM and Smol JP (eds.) *Tracking Environmental Change Using Lake Sediments: Basin Analysis, Coring and Chronological Techniques*, vol. 1, pp. 107–136. Dordrecht: Kluwer Academic Publishers.

- Likens GE (ed.) (1985) *An Ecosystem Approach to Aquatic Ecology: Mirror Lake and Its Environment*. New York: Springer-Verlag.
- Little JL, Hall RI, Quinlan R, and Smol JP (2000) Past trophic status and hypolimnetic anoxia during eutrophication and remediation of Gravenhurst Bay, Ontario: A comparison of diatoms, chironomids and historical records. *Canadian Journal of Fisheries and Aquatic Sciences* 57: 333–341.
- Lotter AF (2001) The paleolimnology of Soppensee (central Switzerland) as evidenced by diatom, pollen and fossil pigment analyses. *Journal of Paleolimnology* 25: 65–79.
- Lotter AF, Birks HJB, Hofmann W, and Marchetto A (1997) Modern diatom, cladocera, chironomid, and chrysophyte cyst assemblages as quantitative indicators for the reconstruction of past environmental conditions in the Alps.1. Climate. *Journal of Paleolimnology* 18: 395–420.
- Luoto TP (2009) Subfossil Chironomidae (Insecta: Diptera) along a latitudinal gradient in Finland: Development of a new temperature inference model. *Journal of Quaternary Science* 24: 150–158.
- MacDonald GM (2003) *Biogeography: Introduction to Space, Time and Life*. New York: John Wiley and Sons Inc.
- MacDonald GM, Moser KA, Bloom AM, et al. (2008) Evidence of temperature depression and hydrological variations in the eastern Sierra Nevada during the Younger Dryas stage. *Quaternary Research* 70: 131–140.
- Mann DG and Droop SJM (1996) Biodiversity, biogeography and conservation of diatoms. *Hydrobiologia* 336: 19–32.
- Martiny JBH, Bohannan BJM, Brown JH, et al. (2006) Microbial biogeography: Putting microorganisms on the map. *Nature Reviews Microbiology* 4: 102–112.
- Massaferro JL, Moreno PI, Denton GH, Vandergoes M, and Dieffenbacher-Krall A (2009) Chironomid and pollen evidence for climate fluctuations during the Last Glacial Termination in NW Patagonia. *Quaternary Science Reviews* 28: 517–525.
- Mills EL, Leach JH, Carlton JT, and Secor CL (1993) Exotic species in the Great Lakes: A history of biotic crises and anthropogenic introductions. *Journal of Great Lakes Research* 19: 1–54.
- Moser KA (2004) Paleolimnology and the frontiers of biogeography. *Physical Geography* 25: 453–480.
- Moser KA, Mordecai JS, Reynolds JL, Rosenbaum JG, and Ketterer ME (2010) Diatom changes in two Uinta Mountain, Utah lakes: Responses to anthropogenic and natural atmospheric inputs. *Hydrobiologia* 648: 91–108.
- Nriagu JO (1989) A global assessment of natural sources of atmospheric trace metals. *Nature* 338: 47–49.
- Paterson AM, Cumming BF, and Smol JP (1998) Assessment of the effects of logging, forest fire and drought on lakes in northwestern Ontario: A 30-year paleolimnological perspective. *Canadian Journal of Forest Research* 28: 1546–1556.
- Porinchu DF and MacDonald GM (2003) The use and application of freshwater midges (Chironomidae: Insecta: Diptera) in geographical research. *Progress in Physical Geography* 27: 378–422.
- Ramstack JM, Fritz SC, and Engstrom DR (2004) Twentieth century water quality trends in Minnesota lakes compared with pre-settlement variability. *Canadian Journal of Fisheries and Aquatic Sciences* 61: 561–576.
- Räsänen J, Kauppi T, and Salonen VP (2006) Sediment-based investigation of naturally or historically eutrophic lakes – Implications for lake management. *Journal of Environmental Management* 79: 253–265.
- Reavie ED and Smol JP (2001) Diatom-environmental relationships in 64 alkaline southeastern Ontario Canada lakes: a diatom-based model for water quality reconstructions. *Journal of Paleolimnology* 25: 25–42.
- Reavie ED, Smol JP, Carnignan R, and Lorrain S (1998) Diatom paleolimnology of two fluvial lakes in the St Lawrence River: A reconstruction of environmental changes during the last century. *Journal of Phycology* 34: 446–456.
- Reed JM, Cvetkoska A, Levkov Z, Vogel H, and Wagner B (2010) The last glacial-interglacial cycle in Lake Ohrid (Macedonia/Albania): Testing diatom response to climate. *Biogeosciences* 7: 3083–3094.
- Rees ABH and Cwynar LC (2010) Evidence for early postglacial warming in Mount Field National Park, Tasmania. *Quaternary Science Reviews* 29: 443–454.
- Rhodes TE and Davis RB (1995) Effects of late Holocene forest disturbance and vegetation change on acidic Mud Pond, Maine, USA. *Ecology* 76: 734–746.
- Rohde W (1969) Crystallization of eutrophication concepts in North Europe. In: Hutchinson GE (ed.) *Eutrophication*, pp. 50–64. Washington, DC: National Academy of Sciences.
- Rühland K, Paterson AM, and Smol JP (2008) Hemispheric-scale patterns of climate-related shifts in planktonic diatoms from North American and European lakes. *Global Change Biology* 14: 2740–2754.
- Rühland KM and Smol JP (2003) Freshwater diatoms from the Canadian arctic treeline and development of paleolimnological inference models. *Journal of Phycology* 38: 249–264.
- Schlesinger WH (1997) *Biogeochemistry: An Analysis of Global Change*, 2nd edn. San Diego: Academic Press.
- Smol JP (2008) *Pollution of Lakes and Rivers: A Paleoenvironmental Perspective*, 2nd edn. Chichester, GB: John Wiley and Sons.
- Smol JP (2010) The power of the past: Using sediments to track the effects of multiple stressors on lake ecosystems. *Freshwater Biology* 55(supplement 1): 43–59.
- Smol JP and Stoermer EF (eds.) (2010) *The Diatoms: Applications for the Environmental and Earth Sciences*, 2nd edn. Cambridge: Cambridge University Press.
- Smol JP, Wolfe AP, Birks HJB, et al. (2005) Climate-driven regime shifts in the biological communities of arctic lakes. *Proceedings of the National Academy of Sciences of the United States of America* 102: 4397–4402.
- Stoermer EF, Emmert G, and Schelske CF (1989) Morphological variation in *Stephanodiscus niagarae* (Bacillariophyta) in a lake Ontario sediment core. *Journal of Paleolimnology* 2: 227–236.
- Stoermer EF, Wolin JA, Schelske CL, and Conley DJ (1985) An assessment of ecological changes during the recent history of Lake Ontario based on siliceous algal microfossils preserved in the sediments. *Journal of Phycology* 21: 257–276.
- Telford RJ, Vandvik V, and Birks HJB (2006) Dispersal limitations matter for microbial morphospecies. *Science* 312: 1015.
- Theriot EC (1992) Clusters, species concepts and morphological evolution of diatoms. *Systematic Biology* 41: 141–157.
- Theriot EC, Fritz SC, Whitlock C, and Conley DJ (2006) Late Quaternary rapid morphologic evolution of an endemic diatom in Yellowstone Lake, Wyoming. *Paleobiology* 32: 38–54.
- Uutala AJ (1990) *Chaoborus* (Diptera: Chaoboridae) mandibles – Paleolimnological indicators of the historical status of fish populations in acid-sensitive lakes. *Journal of Paleolimnology* 4: 139–151.
- Uutala AJ and Smol JP (1996) Paleolimnological reconstructions of long-term changes in fisheries status in Sudbury area lakes. *Canadian Journal of Fisheries and Aquatic Sciences* 53: 174–180.
- Vanormelingen P, Verleyen E, and Vyverman W (2008) The diversity and distribution of diatoms: From cosmopolitanism to narrow endemism. *Biodiversity and Conservation* 17: 393–405.
- Vaughn JC (1961) Coagulation difficulties of the south district filtration plant. *Pure Water* 13: 45–49.
- Verschuren D, Johnson TC, Kling H, et al. (2002) History and timing of human impact on Lake Victoria, East Africa. *Proceedings of the Royal Society of London B* 269: 289–294.
- Vitousek PM, D'Antonio CM, and Loope LL (1997) Introduced species: A significant component of human-caused global change. *New Zealand Journal of Ecology* 21: 1–16.
- Vyverman W, Verleyen E, Sabbe K, et al. (2007) Historical processes constrain patterns in global diatom biodiversity. *Ecology* 88: 1924–1931.
- Walker IR and Cwynar LC (2006) Midges and paleotemperature reconstruction – The North American experience. *Quaternary Science Reviews* 25: 1911–1925.
- Willén E (1991) Planktonic diatoms – An ecological review. *Algalological Studies* 62: 69–106.
- Willis KJ and Birks HJB (2006) What is natural? The need for a long-term perspective in biodiversity conservation. *Science* 314: 1261–1265.
- Wilson EO (1989) Threats to biodiversity. *Scientific American* 261: 108.
- Wolfe AP, Baron JS, and Cornett J (2001) Anthropogenic nitrogen deposition induces rapid ecological changes in alpine lakes of the Colorado Front Range (USA). *Journal of Paleolimnology* 25: 1–7.
- Wolfe AP, Van Gorp AC, and Baron JS (2003) Recent ecological and biogeochemical changes in alpine lakes of Rocky Mountain National Park (Colorado, USA) a response to anthropogenic deposition. *Geobiology* 1: 153–168.
- Wolin JA and Stone JR (2010) Diatoms as indicators of lake level change in freshwater lakes. In: Smol JP and Stoermer EF (eds.) *The Diatoms: Applications for the Environmental and Earth Sciences*, 2nd edn., pp. 174–185. Cambridge: Cambridge University Press.
- Wright HE (1976) The impact of forest fire on the nutrient influxes to small lakes in northeastern Minnesota. *Ecology* 57: 649–662.

Pigment Studies

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Introduction

Pigments of photosynthetic organisms including chlorophylls (Chls), carotenoids, photoprotective compounds, and their derivatives are common in sediments of aquatic environments. They are produced by algae, phototrophic bacteria, and aquatic plants and may also be present in detritus from terrestrial or resuspended material and some invertebrate animals. Many groups of photosynthetic organisms are soft bodied and so leave few morphological remains in sediments. Therefore, biochemical fossils such as pigments may be the only signatures left behind. Pigments can be used to quantify past primary production in aquatic systems, and because many pigments show a degree of taxonomic specificity, they can provide information about past communities of algae or photosynthetic bacteria. Modern analytical techniques allow rapid and accurate separation, identification, and quantification of pigments in sediment stratigraphies, and through improved understanding of biogeochemistry and depositional processes, pigments are routinely used in the reconstruction of paleoenvironments with a wide range of applications (Figure 1).

Pigments as Biomarkers

Of the numerous pigments produced in the aquatic environment, those that are lipid soluble have superior preservation in the sedimentary record. Therefore, sedimentary Chls, carotenoids, and certain ultraviolet (UV)-absorbing lipids are more widely used than water-soluble compounds such as phycobilins. Chls, and carotenoids have several characteristics that make them suitable as Quaternary biomarkers:

1. They are widely distributed and are constituents of all photosynthetic organisms.
2. They are taxonomically specific, although the degree of specificity depends on the pigment. Most phototrophs produce a number of Chls and carotenoids. Some pigments such as Chl *a* are found in all algal groups, whereas others such as the carotenoid alloxanthin are found solely in cryptophytes, and oscillaxanthin is confined mostly to the cyanobacterial group *Oscillatoriaceae*. Affinities of pigments among taxonomic groupings of photosynthetic organisms are shown in Table 1.
3. They are relatively simple to isolate and quantify using a variety of analytical techniques dependent on the degree of detail and precision required.
4. They persist in the sediment environment, provided conditions are favorable (see later).

Structure and Nature of Pigments

Chlorophylls

Chls are magnesium coordination complexes of four pyrrole units (Figure 2(a); A–D) cyclically arranged with a fifth

isocyclic ring (Figure 2(a); E; reviewed in Jeffrey et al., 1997; Grimm, et al., 2006). There are four major Chl groups, of which three, including Chls *a*, *b*, and *c*, derive from algae; one group, the bacteriochlorophylls (Bchls), is produced by photosynthetic bacteria. Each is modeled on the same basic structure but distinguished by various modifications. In Chls *a* and *b* and Bchls *c*, *d*, and *e*, pyrrole D is reduced, and the structure of the ring is referred to as a chlorin; in Chl *c*, D remains unsaturated (called a porphyrin), whereas in Bchls *a* and *b*, both rings D and B are reduced (called a bacteriochlorin). Another feature that distinguishes the Chls is the presence of a long-chain alcohol appended to ring D (phytol present in Chls *a* and *b* and Bchls *a* and *b*; farnesol in Bchls *c*, *d*, and *e*; absent in Chls *c*). However, all Chls have a magnesium atom in the central ring.

Chls are susceptible to oxidative degradation leading to the formation of a number of colored breakdown products. For example, loss of the central magnesium atom from Chls leads to the formation of pheophytins. Loss of the phytol chain, usually from the action of chlorophyllase enzymes, leads to the formation of chlorophyllide. Further degradation causing loss of both the Mg and phytol chain leads to the formation of pheophorbides. Each of these breakdown products is readily detectable in sedimentary pigment extracts and can aid in the interpretation of stratigraphies. Bchls *c*, *d*, and *e* are especially susceptible to attack by substituent groups at ring B or C, leading to the formation of a number of homologues that can make the quantification of Bchl from pigment mixtures particularly challenging. Recently, a further group of compounds, steryl chlorin esters (SCEs), has been identified. SCEs are formed when sterols react with Chl molecules during passage through the gut of herbivores such as zooplankton. The nature of their formation has led to their use as an indicator of zooplankton grazing, but identification of these compounds usually involves the use of more complicated techniques such as mass spectrometry (MS; see later).

Carotenoids

Carotenoids are a larger group of compounds characterized by a long aliphatic polyene chain composed of eight isoprene (C_5H_8) units to give a long conjugated double-bond structure, the chromophore (Figure 2(b); reviewed in Britton et al., 1995; Jeffrey et al., 1997; Frank et al., 1999). Modifications to the end groups (Figure 2(b); A and B) of this basic structure allow different carotenoids to be formed. The two main divisions of carotenoids are the carotenes, which are composed solely of carbon and hydrogen atoms (e.g., β -carotene), and the more numerous xanthophylls, which carry one or more oxygen atom (e.g., fucoxanthin). Common end group modifications include the addition of a 5,6-epoxide to a cyclic end group as in fucoxanthin or the addition of a ketone group as in echinenone but can also include the addition of aldehyde, carboxy, carbomethoxy, methoxy acetate, and lactone groups.

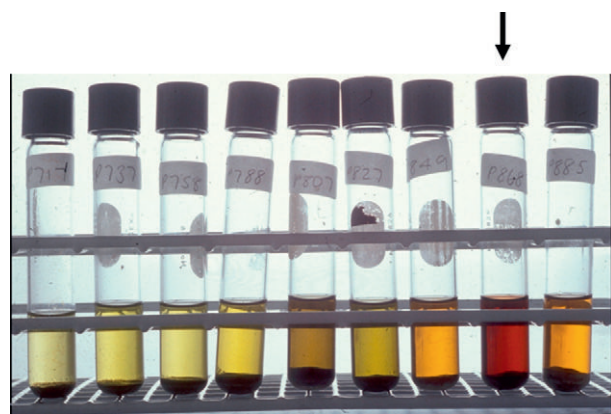


Figure 1 Pigment mixtures extracted from a lake sediment core from Greenland composed mostly of carotenoids and pheophytins. Moving left to right, samples become increasingly dominated by orange carotenoids as the relative abundance of yellow pheophytins declines. Sample indicated with the arrow contains a high abundance of the red carotenoid okenone produced by purple sulfur bacteria. Photograph by S. McGowan.

Because of the extended double-bond system, carotenoids are susceptible to oxidation. Saturation of the double-bond system produces colorless compounds, and so, in contrast to the Chls, carotenoid derivatives are mostly undetectable by spectrophotometric techniques. However, early stages of carotenoid degradation may involve conversion to colored *cis*-isomers. Carotenoid chlorin esters, such as fucoxanthinol pheophorbide ester, may be formed when carotenoids and chlorins esterify on exposure to enzymes in herbivore digestive tracts, and so, they are of potential use as a grazing indicator.

UV-Absorbing Compounds

Compounds that absorb in the UV region (280–400 nm) have been isolated from aquatic organisms. These compounds are produced primarily by cyanobacteria in lake sediments and may function as photoprotectants because they are produced in response to UV exposure. One such pigment, scytonemin, is a yellow-brown, lipid-soluble dimeric pigment with a structure that is based on indolic and phenolic subunits (Figure 2(c); Squier et al., 2004). Scytonemin and its reduced counterparts have been used to infer past UV exposure in aquatic environments (Leavitt et al., 1997).

Light-Absorbing Properties

Pigments in phototrophs capture light either for photosynthesis or for protection from damaging levels of light. In both Chls and carotenoids, light absorption is caused by excitation of the electrons on the double-bond system in the chlorin/porphyrin ring and conjugated chain, respectively (called the chromophore). Absorbance of light at particular wavelengths of the visible spectrum causes Chls and carotenoids to have distinctive absorption spectra (and so colors) that are a useful aid in pigment identification (Figure 3; reviewed in Jeffrey et al., 1997; Britton et al., 1995).

Table 1 Photosynthetic pigments commonly found in the water column and sediments and relative stability in sedimentary environments rated from most stable (1)–(4) least stable (pigments rated 4 are rarely preserved in sediments) or unknown (–) stability

Pigment	Affinity	Stability
Chl <i>a</i>	All photosynthetic algae, higher plants	3
Chl <i>b</i>	Green algae, euglenophytes, higher plants	2
Chl <i>c</i>	Dinoflagellates, diatoms, chrysophytes	4
Bchls <i>a</i> and <i>b</i>	Purple sulfur and nonsulfur bacteria	–
Bchls <i>c</i> , <i>d</i> , and <i>e</i>	Green sulfur bacteria	–
Pheophytin <i>a</i>	Chl <i>a</i> derivative	1
Pheophytin <i>b</i>	Chl <i>b</i> derivative	2
Pheophorbide <i>a</i>	Grazing, senescent diatoms	3
Pyro-pheo (pigments)	Derivatives of <i>a</i> and <i>b</i> phorbins	2
Steryl chlorin esters	Chlorophyll derivatives from grazing	–
β-Carotene	Most algae and plants	1
α-Carotene	Cryptophytes, prochlorophytes, rhodophytes	2
Alloxanthin	Cryptophytes	1
Fucoxanthin	Diatoms, prymnesiophytes, chrysophytes, raphidophytes, several dinoflagellates	2
Diadinoxanthin	Diatoms, dinoflagellates, prymnesiophytes, chrysophytes, raphidophytes, euglenophytes, cryptophytes	3
Diatoxanthin	Diatoms, dinoflagellates, chrysophytes	2
Peridinin	Dinoflagellates	4
Canthaxanthin	Colonial cyanobacteria, herbivore tissues	1
Echinenone	Colonial cyanobacteria	1
Myxoxanthophyll	Colonial cyanobacteria	2
Oscillaxanthin	Cyanobacteria (<i>Oscillatoriaceae</i>)	2
Scytonemin	Colonial cyanobacteria	–
Zeaxanthin	Cyanobacteria	1
Lutein	Green algae, euglenophytes, higher plants	1
Neoxanthin	Green algae, euglenophytes, higher plants	4
Violaxanthin	Green algae, euglenophytes, higher plants	4
Isorenieratene	Green sulfur bacteria	1
Okenone	Purple sulfur bacteria	1
Rhodopinal	Purple sulfur bacteria	–
Astaxanthin	Herbivore tissue, N-limited green algae	4

Stability estimates from Leavitt and Hodgson (2001).

Chls absorb predominantly over blue and red wavelengths (Figure 3(a)) and are thus green in color, whereas carotenoids absorb blue and green light and so appear yellow, orange, or red (Figure 3(b)). Because light absorption properties are determined by the chlorin/porphyrin ring in Chls, loss of the phytol chain to form chlorophyllide has little impact on the spectrum, whereas loss of the magnesium atom changes the structure of the ring and causes a shift in the spectrum of pheophytin which is therefore yellow. In carotenoids, typical spectra have three absorbance maxima (Figure 3(b)); lutein and myxoxanthophyll), and because the chromophore rather than

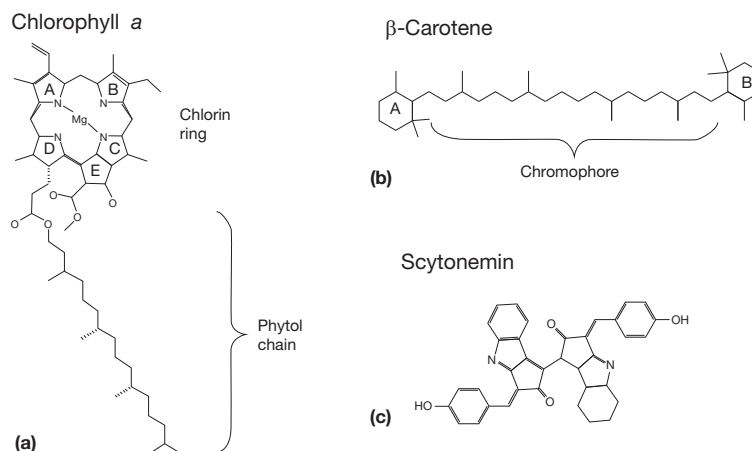


Figure 2 Chemical structures of (a) Chl *a* consisting of cyclic tetrapyrroles with a fifth isocyclic ring (E) and phytol chain, (b) β -carotene showing typical carotenoid structure of a double-bond chromophore bound at each end by functional groups, and (c) scytonemin illustrating the dimeric structure based on indolic and phenolic subunits.

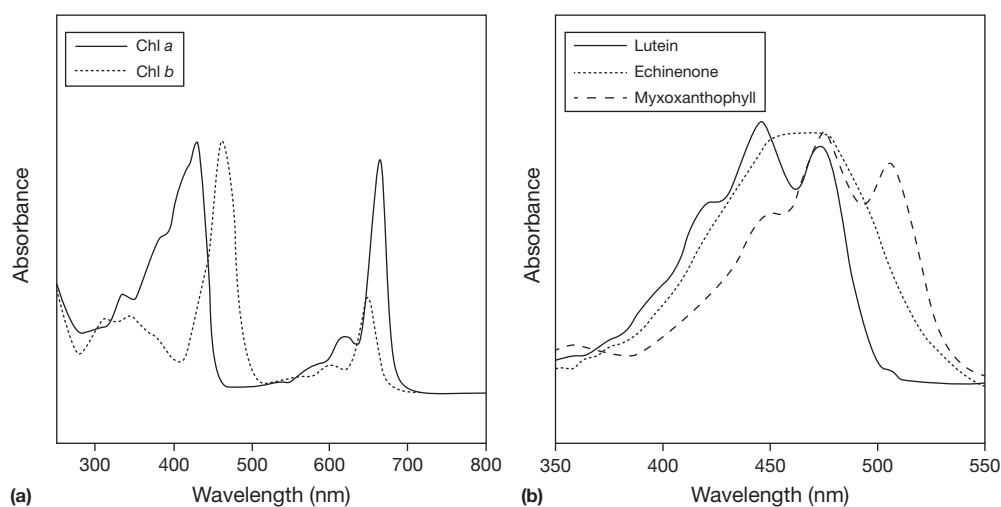


Figure 3 Absorption spectra of (a) Chl *a* and *b* and (b) carotenoids lutein showing the three absorption maxima typical of many carotenoid spectra: myxoxanthophyll showing how the presence of a sugar group in the structure has lengthened the chromophore and caused absorbance at longer wavelengths, and echinenone showing how the presence of the ketone end group has caused the loss of spectral fine structure.

the functional group determines light absorbance properties, carotenoids with very different end groups may have very similar absorbance spectra. This can impede the identification of certain carotenoids using absorbance characteristics alone. However, the addition of some end groups can affect light absorbance characteristics, for example, the addition of a ketone group can destroy the fine structure of the spectrum (Figure 2, echinenone) and the addition of a glycoside group can cause the chromophore to absorb at longer wavelengths (Figure 2, myxoxanthophyll). Therefore, although spectra cannot be used to conclusively identify pigments, they are a valuable aid to initial identification. Chls and carotenoids in solution obey Beer–Lambert law which states that absorbance is linearly proportional to their concentration. Thus, spectrophotometry is commonly used as a means of determining concentrations of pigments based on pigment-specific absorption coefficients.

Analysis

Pigment degradation can occur on exposure to light, oxygen, and heat, so care must be taken with pigment handling and storage. Sedimentary pigment samples should be frozen at $\leq -20^\circ\text{C}$ and freeze-dried in the dark shortly before analysis (Reuss and Conley, 2005). Analysis of the pigments within a sediment sample involves (i) extraction in suitable solvents, (ii) separation of the pigment mixture, (iii) identification of the pigments, and (iv) quantification. Today, this is typically achieved by using high-performance liquid chromatography (HPLC) and online photodiode array (PDA) spectrophotometry because they provide rapid and accurate analytical capabilities (reviewed in Leavitt and Hodgson, 2001).

(i) Extraction of pigments is achieved using a variety of organic solvents that are commonly acetone-based. A good

extraction solvent will be capable of dissolving pigments that range in polarity, and freeze-dried sediment is preferable to minimize interference from water in the sediments. Extraction usually involves soaking the sediments in solvent overnight at -4°C , followed by filtration and sequential washing until all pigments are removed.

- (ii) HPLC separates pigments primarily on the basis of polarity and mass. This is achieved by forcing solvents (the mobile phase) under pressure ($>2000\text{ kPa}$) through a separation column consisting of fine ($<10\text{ }\mu\text{m}$) densely packed particles (the stationary phase). Pigment separation is usually achieved by reversed-phase HPLC (RP-HPLC) where the stationary phase is nonpolar (i.e., lipid rather than water soluble) and the mobile phase is polar (water soluble). When the pigment mixture is carried through, polar pigments that have less affinity for the stationary phase are carried through the column more rapidly because of their greater affinity for the solvent (e.g., polar fucoxanthin elutes early in the chromatogram; see Figure 4). By slowly changing the solvent composition of the mobile phase so that it becomes less polar, more lipid-soluble pigments will be gradually released one by one, as their affinity for the mobile phase exceeds that of the stationary phase. Although polarity is a prime driver for elution order, mass of each molecule will further decide how long the molecule takes to be released from the column (the retention time). There are numerous conditions, solvent mixes, and methods suitable for the separation of pigments (see Leavitt and Hodgson, 2001), and choice of method is usually dependent on the separation efficiency, time taken, and pigments present.
- (iii) Identification of pigments is usually achieved using a combination of retention times and spectral characteristics. Once

eluted from the HPLC column, individual pigments are detected and retention times identified by scanning with a spectrophotometer at a fixed wavelength of maximal pigment absorbance (e.g., 450 nm) to produce a chromatogram (Figure 4). The position and retention time of the pigment on the chromatogram can be used to tentatively identify the pigment based on comparison with known commercial pigment standards under the same separation conditions. However, use of a PDA detector allows eluted pigments to be scanned simultaneously at multiple wavelengths to generate an absorbance spectrum, and this can be used to aid identification. For further identification of unknowns, the collection of the sample and its identification by MS (sometimes online) can be helpful in identifying problematic compounds such as Bchl homologues and degradation compounds.

- (iv) Quantification of pigments is achieved from spectroscopic properties using Beer–Lambert law. Internal calibration standards may be added to the extracts at the extraction or injection stage to correct for errors associated with analysis and handling. Calibration curves are constructed by injecting a known quantity of each pigment to allow conversion of chromatogram peak areas to concentrations. Such calibrations may be automated within the analytical software, but complex sedimentary mixtures usually require the expertise of an experienced operator to check that interpretations are correct. Final concentrations of pigments are normally expressed in nanomoles (nm) to correct for differences among pigment molecular weights and calculated relative to the organic matrix of sediment to reduce the effects of dilution by allochthonous mineralogenic material and because mineralization of labile bulk organics partly compensates for pigment degradation.

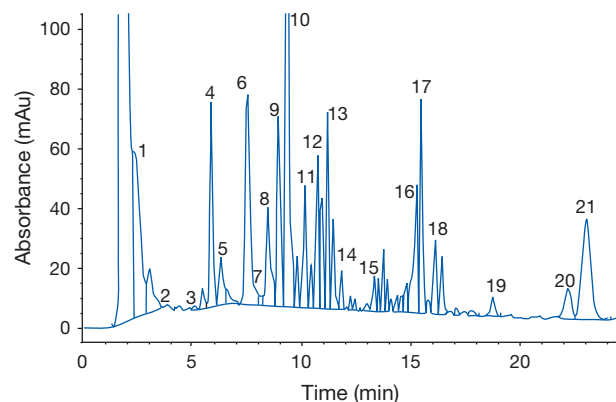


Figure 4 Typical RP-HPLC chromatogram of a pigment mixture. Pigments include (1) scytonemin, (2) Chl *c*, (3) fucoxanthin, (4) UV-absorbing compound *b* and (5) *a*, (6) internal standard Sudan II dye, (7) myxoxanthophyll, (8) alloxanthin, (9) diatoxanthin, (10) lutein–zeaxanthin, (11–13) carotenoid isomers, (14) canthaxanthin, (15) Chl *b*, (16) okenone, (17) Chl *a*, (18) echinenone, (19) pheophytin *b*, (20) pheophytin *a*, and (21) β -carotene. Note: Chromatograph has been loaded with excessive pigments for demonstration purposes. Reproduced from Leavitt PR and Hodgson DA (2001) Sedimentary pigments. In: Smol JP, Birks HJB, and Last WM (eds.), *Terrestrial, Algal and Siliceous Indicators, Vol. 3: Tracking Environmental Changes using Lake Sediments*. With kind permission from Springer Science+Business Media B.V.

Nonchromatographic analytical techniques for bulk unseparated pigments were used regularly prior to the development of HPLC (Swain, 1985). Chl derivatives can be measured on bulk extracts using a spectrophotometer while total carotenoids are measured on extracts that are first exposed to potassium hydroxide (saponification), which results in the removal of Chls. Because specific cyanobacterial carotenoids such as oscillaxanthin and myxoxanthophyll are relatively polar, they can be removed from pigment mixtures by phase separation and quantified using spectrophotometry that takes advantage of the carotenoids' distinctive spectra (Figure 3(b); myxoxanthophyll). Although the increased use of HPLC has meant that these techniques are now becoming obsolete, because they do not require specialist equipment, they can be useful in providing initial estimates of primary production.

An alternative nonchromatographic technique of estimating sedimentary Chl *a* concentrations is the use of reflectance spectroscopy (Das et al., 2005; cf. *Visible and Infrared Spectroscopical Applications*). This approach measures the size of troughs in reflectance spectra of sediments in either the visible near-infrared range ($600\text{--}760\text{ nm}$) or in the visible range (visible reflectance spectroscopy; VRS $650\text{--}700\text{ nm}$) to estimate changes in sedimentary Chl abundance (Michelutti et al., 2010). Accurate quantification of sedimentary Chl concentration can be achieved by calibration with samples measured by

HPLC, but overall, down-core pigment trends can be ascertained without the calibration stage. These techniques have been successfully used to record shifts in primary production in lakes from a wide range of trophic states (Michelutti et al., 2010). Therefore, because of its nondestructive nature, this novel analytical technique has great potential for the rapid screening of core sequences, where taxonomic detail about primary producers is not required.

Recently, there has been a growth in studies investigating the chemical structure of individual pigment isolates using MS techniques. MS analysis may be conducted directly on the HPLC eluent (online HPLC-MS), and techniques such as electron ionization, liquid secondary ion, and tandem MS are used to identify pigments by fragmenting molecules and identifying ionized fragments based on mass. Such techniques are usually employed as an identification tool in complex or unusual pigment mixtures that are difficult to identify based solely on spectral and retention characteristics. Another approach that is currently developing is the analysis of stable isotopes of carbon, nitrogen, and hydrogen (deuterium) in individual pigments using isotope-ratio MS (IRMS). This method requires a more lengthy extraction procedure to provide sufficient isolated pigment for IRMS analysis. Pigments are usually extracted using a combination of solid phase chromatography, reversed- and normal-phase HPLC, so this procedure can require a large amount of sediment to yield sufficient pigment for analysis. Pigment-specific isotope studies have the potential to tell us about changes in productivity and nutrient processing in individual algal taxa (Enders et al., 2008). Thus, interpretation is more straightforward than bulk sediments, where isotopic shifts can be caused by numerous factors. Stable carbon isotopes in pigments have also been used to investigate degradation processes in pheopigments by analyzing the isotopic composition of individual molecular fragments to determine the sources of phytol fragments added to molecules during degradation (Chikaraishi et al., 2007). Stable isotopes have been analyzed in Chl derivatives (porphyrins) from mid-Cretaceous shales and used to infer cyanobacterial nitrogen fixation. Therefore, taxonomic information may be inferred even in very heavily degraded samples that contain no intact diagnostic pigments (Kashiyama et al., 2008). The barriers to research in this area are the analytical time involved and the need for large samples, but as these problems are overcome, they have great potential in future paleoenvironmental research.

Preservation in Aquatic Environments

Some pigments degrade more readily than others (see Table 1; reviewed in Leavitt, 1993). Chls are relatively labile, but because they produce detectable breakdown products, they still leave identifiable remains in sedimentary records. Carotenoids, particularly carotenes, such as β -carotene are generally more stable, but some xanthophylls can be especially degradation prone; for example, fucoxanthin and peridinin are very labile because of the presence of a 5,6-epoxy-end group. Consequently, fucoxanthin is sometimes only present in the uppermost part of sediment records, whereas peridinin is rarely found in sediments. Although variability in degradation among pigments makes it difficult to reconstruct shifts in

relative algal abundance among groups, down-core patterns in individual pigments give good estimates of the abundance of specific algal groups. Thanks to past experimental work, there is now a thorough understanding of how degradation affects the interpretation of pigment stratigraphies. Therefore, changes in pigment preservation can give useful information about changes in past environmental conditions.

The general rules governing pigment degradation involve exposure to oxygen, light, and heat (reviewed in Leavitt, 1993; Cuddington and Leavitt, 1999). Because these conditions characterize the water column of many aquatic ecosystems, the majority of pigments (>95%) usually degrade before reaching the sediment. Degradation processes in the water column include chemically or microbially mediated oxidation, digestion during grazing by invertebrates, bacterial degradation, cell lysis, and enzymatic metabolism during senescence. These processes occur rapidly, with a half-life of days (Figure 5). Sedimentary pigment preservation is therefore largely dependent on how quickly pigments are degraded in the water column, which depends on lake morphometry and depth, sinking velocities of phototrophs, and grazing by zooplankton (which may negatively affect pigments by destruction from digestion but encourage preservation by increasing the sinking velocity of those pigments that become incorporated into heavier feces). Once deposited in the surficial sediments, pigment degradation is usually slower, although this will be dependent on conditions at the sediment–water interface. In oxygenated sediments where burrowing invertebrates are common, degradation via oxidation and digestion will continue. However, the conditions that characterize the sediments of many lakes tend to slow down pigment degradation such that it occurs over longer timescales (years). Evidence of such degradation in the upper sediments (~10 cm) is demonstrated by concentrations of labile pigments (e.g., fucoxanthin, Chl *a*) in sediment cores that commonly increase exponentially in the uppermost sediments. Once incorporated into the sediment below the zone of bioturbation, conditions are usually favorable for preservation, and most degradation involves slow saturation reactions that occur over timescales of millennia. Therefore, provided that pigments can pass the initial stages of deposition, they have been known to preserve intact for thousands of years and are therefore of considerable value to the paleoenvironmental record.

Pigments in Quaternary Deposits

Pigments have been used for paleoenvironmental reconstruction in a range of aquatic deposits. Because adequate preservation of pigments favors environments where pigments are rapidly buried and removed from light, heat, and oxygen, many lake sediments yield excellent pigment records. Records tend to be especially well preserved in lakes where bottom waters are anoxic or in shallow lakes where pigments are more rapidly incorporated into the sediment because sinking depths are small and production is dominated by algae growing on sediment surfaces. Most lake-based pigment reconstructions are recent (last few centuries) and some span the Holocene period, but pigments are increasingly being used for environmental reconstruction dating back to the Pleistocene

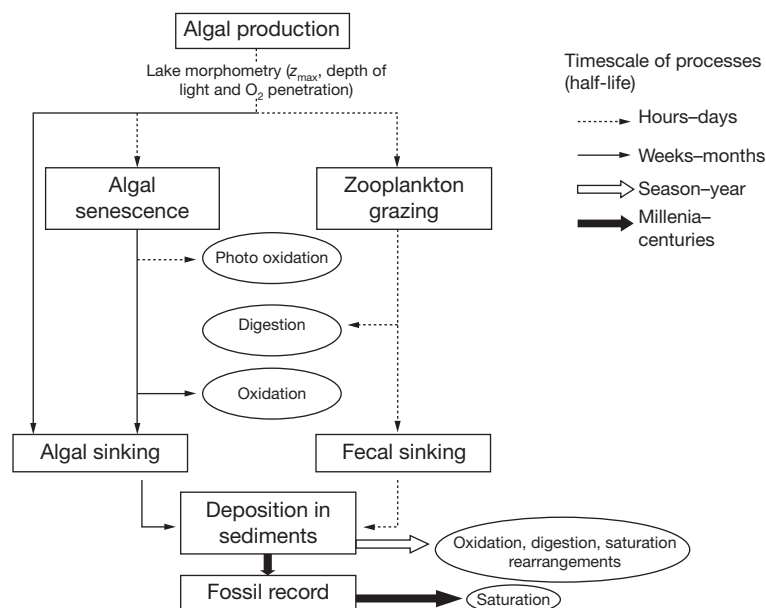


Figure 5 Pathways and rate of pigment degradation during transport through the water column and incorporation into the sediment. Adapted from Leavitt PR (1993) A review of factors that regulate carotenoid and chlorophyll deposition and fossil pigment abundance. *Journal of Paleolimnology* 9: 109–127; Cuddington K and Leavitt PR (1999) An individual-based model of pigment flux in lakes: Implications for organic biogeochemistry and paleoecology. *Canadian Journal of Fisheries and Aquatic Sciences* 56: 1964–1977.

and earlier (Tani et al., 2009; see later). In marine environments, lower rates of burial, deeper water columns, tidal exposure, and lower productivity may compromise preservation in some sites (Reuss et al., 2005). However, pigments have been used as effective indicators in coastal embayments and estuaries. For example, pigments were useful in reconstructing the eutrophication history of an anoxic estuary in Denmark over the past century (Ellegaard et al., 2006), and a >30-ka record extracted from Antarctica has yielded information about the development of marine flora during glacial–interglacial cycles (Hodgson et al., 2003). Geochemists are continuing to develop new ways of identifying and analyzing derivatives of Chl and carotenoid pigments to extract paleoenvironmental information from samples dating back millions of years (Kashiyama et al., 2008).

Sediment Pigments and Monitoring Records

Sedimentary pigments integrate algal production from the entire lake, including algae that live on the lake bottom (Buchaca and Catalan, 2008). They are therefore uniquely placed to quantify whole-ecosystem production which is usually not possible by plankton monitoring alone. However, comparison of sedimentary pigments with long-term phytoplankton monitoring records has revealed generally good correlations between the two in some sites. Comparisons of a 27-year algal monitoring record from Lough Neagh (Northern Ireland) showed strong correlations between the dominant cyanobacteria and measures of diagnostic pigments (Bunting et al., 2007). Similarly, a 20-year algal monitoring record compared with annually laminated (varved) sediments in an experimentally enriched Canadian boreal lake showed that pigments correlated well with algal standing crop and recorded the

striking increase in green algae and cyanobacteria that resulted from eutrophication (Leavitt and Findlay, 1994). Correlations with labile pigments from chrysophytes, diatoms, and dinoflagellates were weaker, but this was caused by the invariant nature of these algal groups as eutrophication progressed and the lability of some of these pigments.

In Himmerfjärden Bay (Swedish Baltic) when pigments from varved sediment cores were compared with >20 years of monitoring data, pigments from the predominant algal group (diatoms) correlated well with known records and registered past episodes of diatom blooms (Figure 6). Pigments from cyanobacteria and total algae did not match monitored records, possibly because picoplanktonic cyanobacteria were not enumerated in algal counts or because gas-vacuolated cyanobacteria did not sink into sediments as readily (Bianchi et al., 2002). This work also showed that sediment resuspension and differential settling of diatoms are important in this site because, although sediments were annually laminated, correlations among monitoring records and sediments were only evident on 5-year smoothed averages. Therefore, although pigments faithfully record changes in dominant algal groups in the water column, the fossil record may be biased due to preferential deposition of certain algal taxa, differential degradation of some pigments, and standard algal monitoring techniques that rarely sample the entire water column. Thus far, comparisons of pigments and contemporary records are disadvantaged by the short length of monitoring records that reduces the statistical power of correlation analyses and necessitate the analysis of the uppermost parts of the sediment record where degradation processes are most active. As long-term records continue to be gathered, better assessment of how phytoplankton pigments are integrated into the sediment record should become possible.

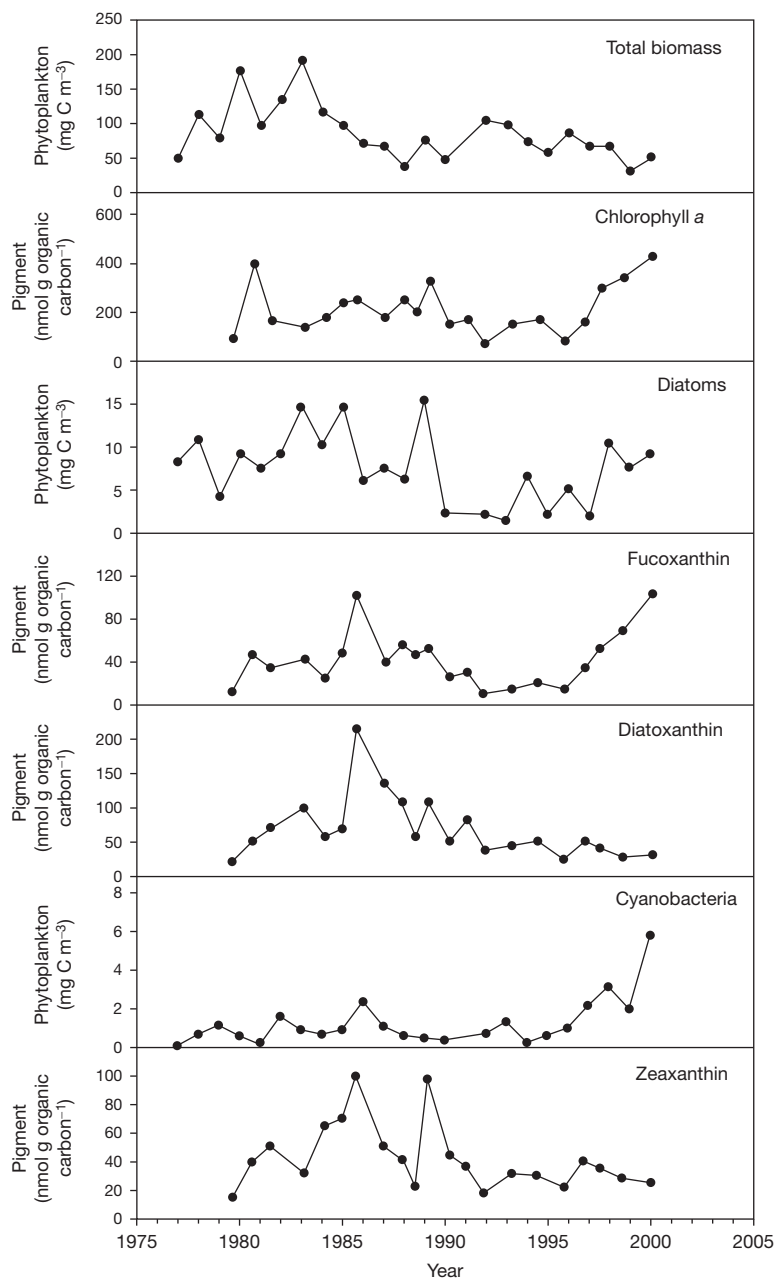


Figure 6 Comparison of indicator pigments extracted from annually laminated sediments in Himmerfjärden Bay (Baltic Sea) with monitored phytoplankton records for total phytoplankton biomass versus Chl *a*, total diatoms versus diatoxanthin and fucoxanthin, and total cyanobacteria versus zeaxanthin. Reproduced from Bianchi TS, Engelhaupt E, McKee BA, et al. (2002) Do sediments from coastal sites accurately reflect time trends in water column phytoplankton? A test from Himmerfjärden Bay (Baltic Sea proper). *Limnology and Oceanography* 47: 1537–1544, with permission. Copyright (2002) by the American Society of Limnology and Oceanography, Inc.

Eutrophication

Pigments have been used effectively in deeper lakes to record changes in water transparency associated with lake eutrophication (nutrient enrichment). In deep lakes where wind energy is insufficient to completely mix lake waters, bottom waters may become permanently anoxic (termed meromixis). If light penetrates into the anoxic layer, it can provide a habitat for photosynthetic sulfur bacteria that require both anoxia and light for photosynthesis. Specific purple and green sulfur

bacterial pigments (okenone and isorenieratene, respectively) declined in meromictic alpine Lake Tovel over the past 75 years, and this was interpreted to be caused by decreasing water transparency due to nutrient enrichment and increased algal growth in the upper water column (Lami et al., 1991).

One consequence of eutrophication in lakes and estuaries has been the widespread increase in cyanobacterial blooms. Because morphological fossils of cyanobacteria may not always preserve reliably in sediments, pigments are particularly useful in

assessing past blooms because these taxa have several indicator pigments. The cyanobacterial carotenoid myxoxanthophyll increased markedly after 1965, indicating increases in *Planktothrix agardhii* cyanobacteria in Lough Neagh as the lake became enriched (Bunting et al., 2007). In contrast, sedimentary pigments have also demonstrated that some lakes in the North American prairies and UK lowlands, as well as estuaries in the Baltic have sustained abundant cyanobacterial communities for millennia, suggesting that they are natural features of these systems (Bianchi et al., 2000; McGowan et al., 1999). Such information is valuable in determining appropriate restoration targets for ecosystems subjected to the effects of anthropogenic nutrient enrichment.

Aquatic Food Web Changes

Fossil pigments are important in demonstrating that changes in lake food web structure rather than nutrients can control algal communities in some circumstances. This work was pioneered in lakes in Michigan where three complete changes in fish communities (trout, cyprinid, bass) resulted in striking changes in body size of the zooplankton communities (Leavitt et al., 1989). Because zooplankton graze on algae in the water

column, when larger, more efficient zooplankton grazers were replaced with smaller species, most pigments increased in sediment cores, but there was a decline in deepwater blooms of chrysophytes. When larger zooplankton returned, deep-water chrysophyte blooms increased.

Sedimentary pigments are of particular value in monitoring algal community change in shallow lakes, where the majority of production may be associated with benthic algae that grow attached to surfaces and are difficult to monitor quantitatively. Over the past century, many shallow lakes in northern Europe have exhibited regime shifts and changed from a situation where there are few algae in the water column and there are abundant aquatic plants to one where the water is turbid with algal growth and the lake is devoid of aquatic plants. The effect of previous state changes on lake algal communities has been investigated using pigments and other proxies in sediment cores from a lake from Denmark. Analyses showed that remains of aquatic plants declined, while planktonic diatoms and cladoceran zooplankton that reside in the water column simultaneously increased, indicating that the lake changed to the turbid state ca. 1940 (McGowan et al., 2005; Figure 7). Pigments showed that when this happened, cyanophytes (myxoxanthophyll) in the water column increased, but

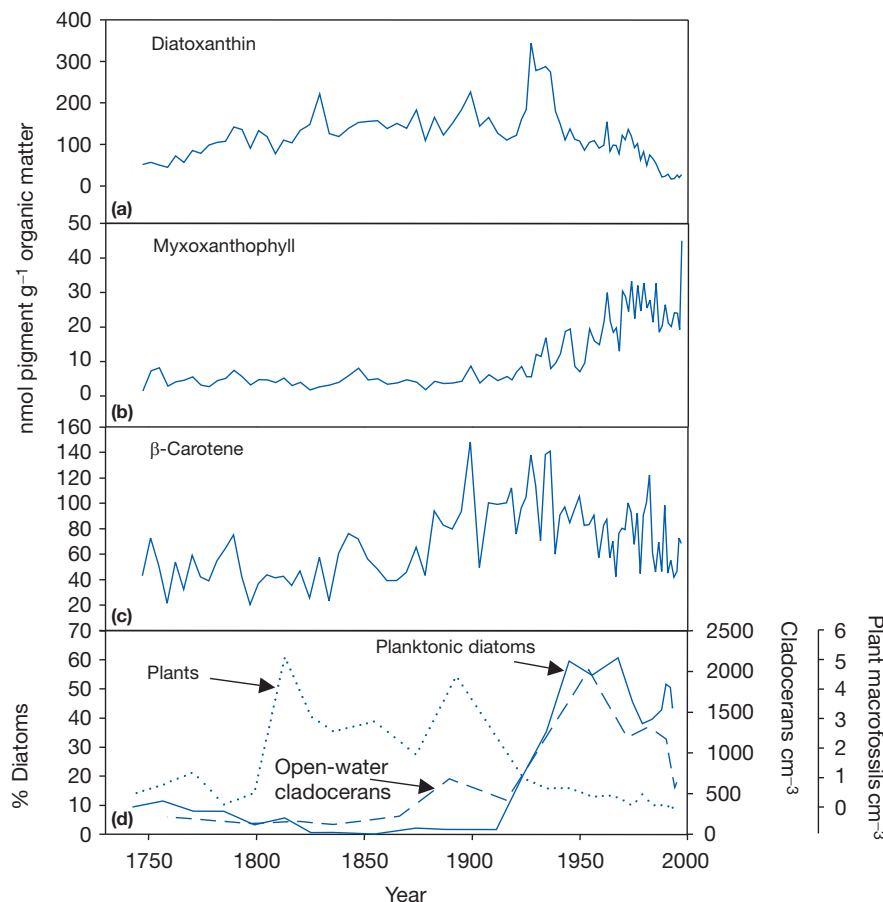


Figure 7 Changes in pigments from (a) diatoms (diatoxanthin), (b) cyanobacteria (myxoxanthophyll), and (c) total algae (β -carotene) compared with (d) other sediment remains of plant macrofossils, planktonic diatoms, and open-water cladocerans from sediments of shallow Lake Lading, Denmark. Adapted from McGowan S, Leavitt PR, Hall RI, et al. (2005) Controls of algal abundance and community composition during ecosystem state change. *Ecology* 86: 2200–2211.

diatoxanthin from benthic diatoms that grow attached to surfaces declined. This compensatory switch in the habitat of algal production resulted in little overall change in whole-lake algal abundance (indicated by β -carotene). Therefore, because sedimentary pigments integrate algal production from the entire lake basin, they can answer questions that would be difficult to approach by conventional monitoring techniques.

Atmospheric Deposition of Contaminants

Pigments have been used both in the detection of atmospheric contamination and in tracking lake response and recovery from its effects. Industrial contaminants containing sulfur and nitrogen have polluted lakes in many remote regions through deposition of 'acid rain.' Because acidification of Chl *a* (maximum absorbance 430 nm) leads to the formation of pheophytin *a* (maximum absorbance 410 nm), the 430 nmE: 410 nm absorbance ratio of bulk pigment extracts has been used to estimate lake water pH (Guilizzoni et al., 1992). Strong correlations between this ratio in surface sediments of alpine lakes and lake water pH and good down-core agreement with chrysophyte scale reconstructions corroborate its use, but because of degradative transformations from Chl *a* to pheophytin *a* in the uppermost sediments, such ratios must be interpreted cautiously.

Pigments also play an important role in detecting changes in water clarity associated with lake acidification. When lakes acidify, brown-colored dissolved organic carbon (DOC) concentrations decline and the water becomes more transparent. Experimental acidification recorded in sediments of a lake in western Ontario led to a sixfold increase in fucoxanthin from deepwater chrysophytes and elevated concentrations of UV-screening pigments when acidification was moderate, but at very low pH, UV penetration caused algal production to be suppressed (Leavitt et al., 1999). Similar lakes exposed to acidification from atmospheric pollution during the twentieth century also showed elevated levels of alloxanthin and fucoxanthin, indicating deepwater blooms of cryptophytes, chrysophytes, and dinoflagellates, and Chl *b* indicating proliferation of filamentous green algae (Vinebrooke et al., 2002; Figure 8). As well as acidifying lakes, nitrogen in industrial emissions has the potential to fertilize lakes that are less acid sensitive. In comparison to the effects of acid deposition, the effects of nitrogen deposition are only just becoming fully realized. Compound-specific stable isotope analysis of chloropigments from sediments in the Rocky Mountains of Colorado has provided an important piece of evidence that atmospheric deposition of nitrogen is fertilizing lakes in this remote area (Enders et al., 2008). The marked decline of the $\delta^{15}\text{N}$ signature of sedimentary Chl derivatives provides

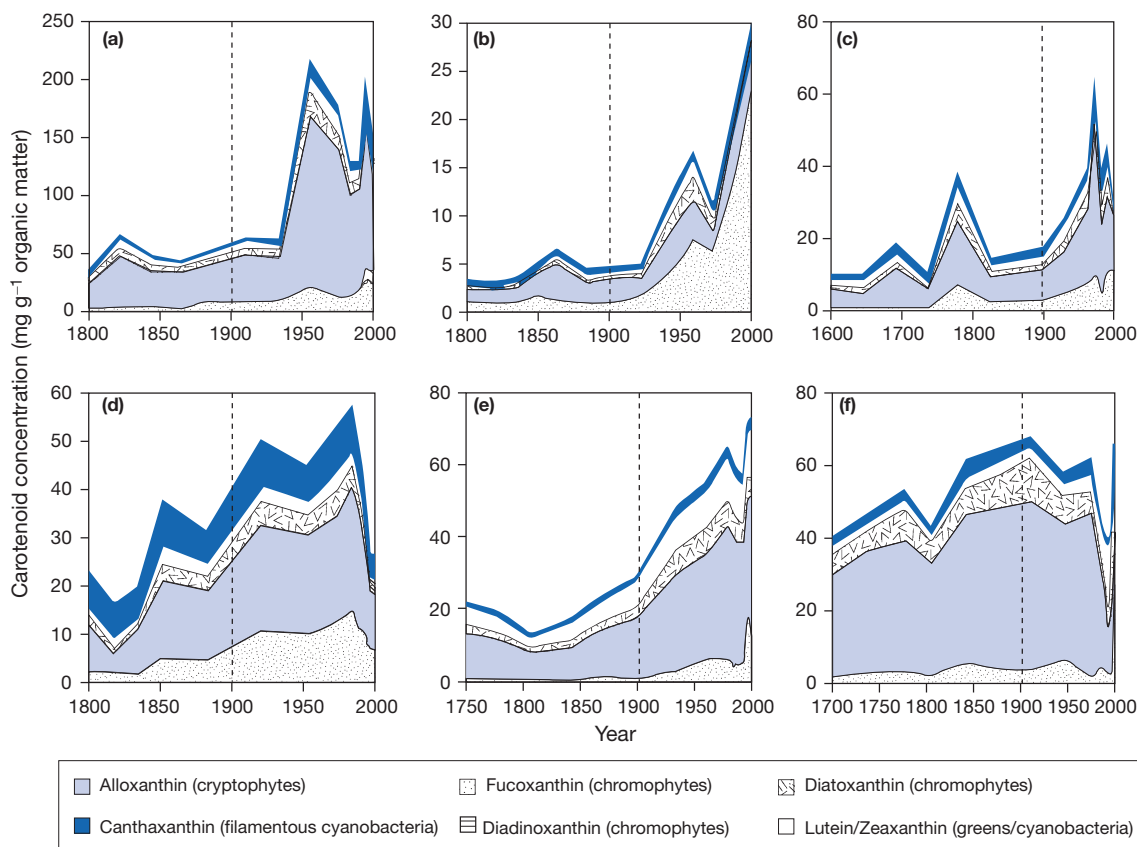


Figure 8 Pigments in sediment cores of six lakes from Killarney Park, Ontario, Canada. Lakes vary in severity of acidification ranging from severe in (a) Acid Lake and (b) OSA Lake to moderate in (c) George Lake and (d) Bell Lake to negligible in reference lakes: (e) Teardrop and (f) Helen. The onset of industrial deposition over the lakes (ca. 1900) is indicated by dashed vertical lines. Reproduced from Vinebrooke RD, Dixit SS, Graham MD, et al. (2002) Whole-lake algal responses to a century of acidic industrial deposition on the Canadian Shield. *Canadian Journal of Fisheries and Aquatic Sciences* 59: 483–493, with permission from NRC Research Press.

convincing evidence that algal production was stimulated by an increase in nitrogen availability since 1950.

Hydrological Change

Pigments are excellent proxies in tracking past hydrological change in lakes because their production is strongly affected by water level changes. The hydrological regime of Lake Kinneret in Israel has been strongly managed over the past century for hydroelectric power schemes, wetland drainage, and conversion to a water storage reservoir. Paleopigments show that the increase in water level fluctuations has been associated with elevated algal production and degradation of water quality (Hambright et al., 2008).

Pigments have also been used to assess the response of delta lakes to changes in flood frequency in the Peace-Athabasca Delta (PAD; northern Alberta). Declines in the regularity of flooding cause ions and nutrients in PAD lake waters to become more concentrated because of lower inputs of dilute river waters. Lake levels also decline when flooding is infrequent, leading to the expansion of submerged and emergent aquatic vegetation which provides a habitat for algal growth. Pigments from lake sediments in the PAD demonstrate how these changes in lake conditions during periods of reduced flood frequency lead to increased lake primary production over decadal timescales (Figure 9; McGowan et al., 2011).

In closed-basin lakes, hydrological change may be accompanied by changes in salinity. In lakes of western Canada, pigments in surface sediments showed that there was a turnover of algal pigments along a salinity gradient (lutein-zeaxanthin and diatoxanthin replaced fucoxanthin as salinity increased), implying that changes in algal groups can infer past salinity changes (Vinebrooke et al., 1998). However, because salinity changes covary with water level, pigments in sediment-core stratigraphies of closed-basin lakes are likely to register a combination of the two factors. A further level of complexity arises in saline systems which become meromictic

(permanently stratified). Saline lakes in West Greenland stratify because dense saline water becomes overlain by less dense fresh water formed during winter ice cover (McGowan et al., 2008). These conditions favor anaerobic purple sulfur bacteria with characteristic pigments (e.g., okenone). Therefore, climate-driven changes in the water balance in closed-basin lakes in West Greenland lead to marked shifts in phototrophic community structure caused by a combination of changes in water level, lake mixing, and salinity.

Catchment Vegetation Changes

Changes in catchment vegetation can influence the flux of colored terrestrial DOC into lakes. Therefore, pigments can be indirect proxies for catchment vegetation by recording DOC-related fluctuations which influence light penetration into the water. DOC screens harmful UV radiation (UVR) from the water, and thus, lake sediments from Antarctic areas devoid of vegetation contain abundant UVR-screening scytonemin pigments produced to protect algae when DOC is absent. Similarly, lakes in montane environments where terrestrial sources of DOC are scarce are poorly buffered against the effects of droughts. In the Canadian Rockies, a period of drought resulted in an increase in abundance of UV-absorbing pigments in lake sediment cores because of a decline in DOC (Leavitt et al., 1997). Longer records spanning the Holocene indicate that UV-absorbing pigments are abundant in newly deglaciated lakes, reflecting the lack of terrestrial DOC in the catchment and the properties of lake waters turbid with suspended glacial flour that attenuate UVR less efficiently. Lakes situated at treeline may experience dramatic changes in DOC inputs; for example, Crowfoot Lake was above treeline >10000 BP, below treeline for the next 6000 years, and returned to treeline ~4000 BP (Leavitt et al., 2003; Figure 10). When below treeline, greater development of terrestrial vegetation (as indicated by pollen from N_2 -fixing plants)

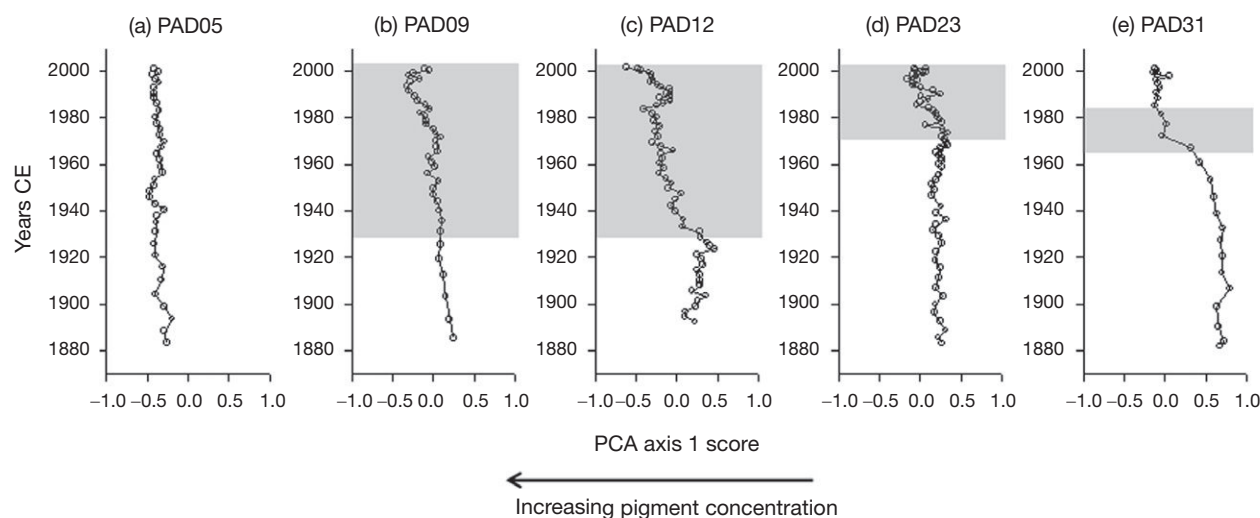


Figure 9 Principal components analysis axis 1 scores from pigment assemblages of sediment cores from five PAD lakes. Shaded areas demarcate the periods of reduced flood frequency (b–e), and there was no change in water flux in ‘reference’ lake PAD05 (a). Axis 1 scores moving to the left of the diagram indicate an increase in pigment concentrations. Reproduced from McGowan S, Leavitt PR, Hall RI, et al. (2011) Interdecadal declines in flood frequency increase primary production in lakes of a northern river delta. *Global Change Biology* 17: 1212–1224.

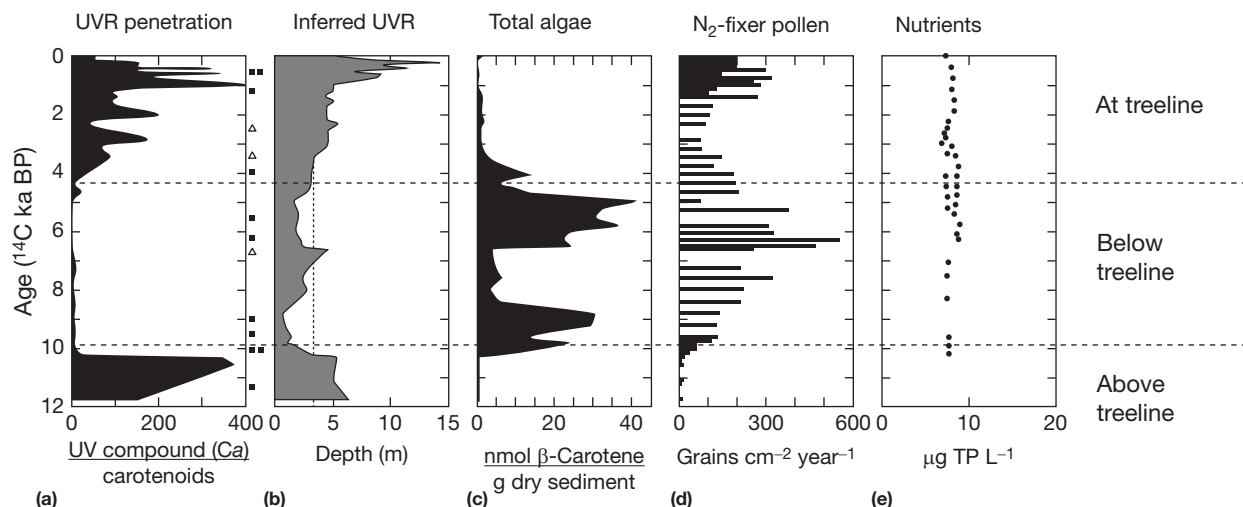


Figure 10 Multiple proxies in sediments from Crowfoot Lake including UVR reconstruction using (a) the ratio of UV-absorbing pigment A: the sum of major carotenoids and (b) diatoms compared with (c) β -carotene as a measure of total production, (d) pollen from N_2 fixers, and (e) nutrients (total phosphorus) reconstructed from diatoms. Reproduced from Leavitt PR, Cumming BF, Smol JP, et al. (2003) Climatic control of ultraviolet radiation effects on lakes. *Limnology and Oceanography* 48: 2062–2069, with permission. Copyright (2003) by the American Society of Limnology and Oceanography, Inc.

increased inputs of DOC and augmented the attenuation of UVR, leading to a lower relative abundance of UV-screening pigments. This interpretation was corroborated by diatom-based UVR reconstructions. Further analysis of other algal pigments indicated that during periods of high UV penetration, algal production was 10–25-fold lower, and because lake nutrient concentrations did not change, UV was likely the causal factor in suppressing algal production. Pigments in surface sediments of sites spanning the treeline in Sweden appear to confirm that UV-screening compounds are more abundant in tundra than forested sites, highlighting the importance of DOC as a control on aquatic production and the sensitivity of pigments to ecotonal boundary changes (Reuss et al., 2009).

Long-Term Records

As cores are recovered from deeper lakes with records spanning thousands to millions of years, pigments have begun to be used in interpreting these continental-scale climate records through glacial and interglacial periods. Changes in rhodospinal, a pigment from anaerobic purple sulfur bacteria, were interpreted as being caused by alternating periods of lake anoxia and oxygenation in sediments of an Italian crater lake during the last glacial period (24–15 000 BP; Chondrogianni et al., 2004). Together with evidence from other proxies, including stable isotopes, sediment laminations, diatoms, and ostracods, it was suggested that the lake became well oxygenated during cooler drier periods because it was shallower and production declined (rhodospinal declines), whereas during warmer, wetter intervals, lake level and production (total carotenoids) increased, leading to lake anoxia (Figure 11).

One key issue that pigments can help to address is the comparison of climatic conditions in the previous (Eemian; Marine Isotope Stage 5e or MIS 5e) and present (Holocene) Interglacial periods. Pigments analyzed from lake sediment cores in Antarctica and Siberia (Lake Baikal) appear to confirm that the

Eemian was generally warmer than the Holocene (Fietz et al., 2007; Hodgson et al., 2006). In Lake Baikal, this was inferred from the higher concentrations of Chl *a* in Eemian sediments, which was interpreted as being caused by greater availability of phosphorus nutrient during the warmer Eemian (Fietz et al., 2007). The Lake Baikal pigment records also assisted in defining the start and termination of the MIS 5e Interglacial period (termed the Kazantsevo Interglacial) in this area of Russia. Probably the longest pigment sequences yet recovered were obtained from cores spanning the Plio–Pleistocene (4.5 Ma) period in Lake Baikal (Tani et al., 2009; Figure 12). Over these timescales, pigments are largely degraded, and so, the degradation products SCEs were analyzed to infer past phytoplankton composition. SCEs were categorized into indicators of diatoms, chlorophytes (SCEs deriving from Chl *b*), and dinoflagellates (SCEs formed with the dinoflagellate marker dinostanol). Using these markers, the pigments showed not only that there was generally good agreement with the marine benthic oxygen isotope record but also that there are periods of elevated dinoflagellate (4.5–4.2 and 1.7–1.3 Ma) and chlorophyte (2.5–2.6 Ma) abundance associated with periods of regional cooling and lower diatom abundance. This is the first time that phytoplankton other than diatoms have been used to infer past change in these long records.

Conclusions

Recent decades have seen a great expansion in analytical capabilities and improved understanding of how pigments and their degradation products are incorporated into the sediment record. Pigment analysis has become a routine part of multi-proxy studies and contributes information that would be impossible to derive from other proxies. As these techniques continue to be used in an increasing range of environments, it is predicted that pigment analysis should make a valuable contribution to the Quaternary sciences in the future.

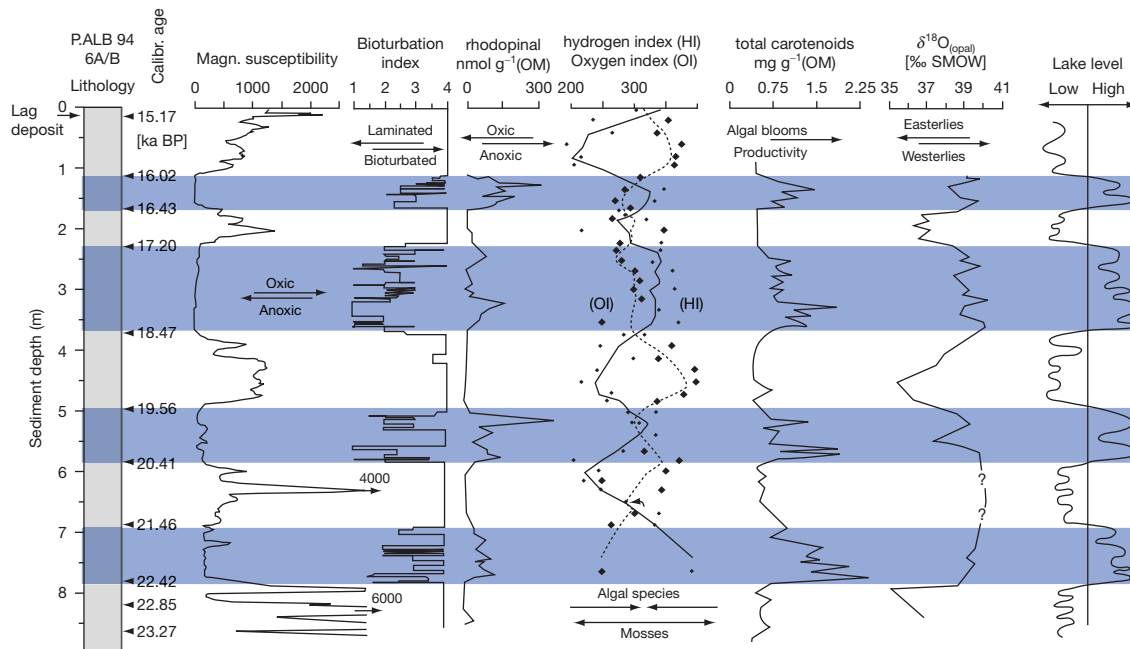


Figure 11 A combination of proxies including lithology, magnetic susceptibility, bioturbation index, the concentration of purple sulfur bacterial pigment rhodopinal, hydrocarbon, and oxygen content of the organic matrix (oxygen and hydrogen index), total carotenoids, and $\delta^{18}\text{O}$ stable isotopes from diatom silica were used to infer historic changes in water level in Lake Albano. Reproduced from Chondrogianni C, Ariztegui D, Rolph T, et al. (2004) Millennial to interannual climate variability in the Mediterranean during the Last Glacial Maximum. *Quaternary International* 122: 31–41, with permission from Elsevier.

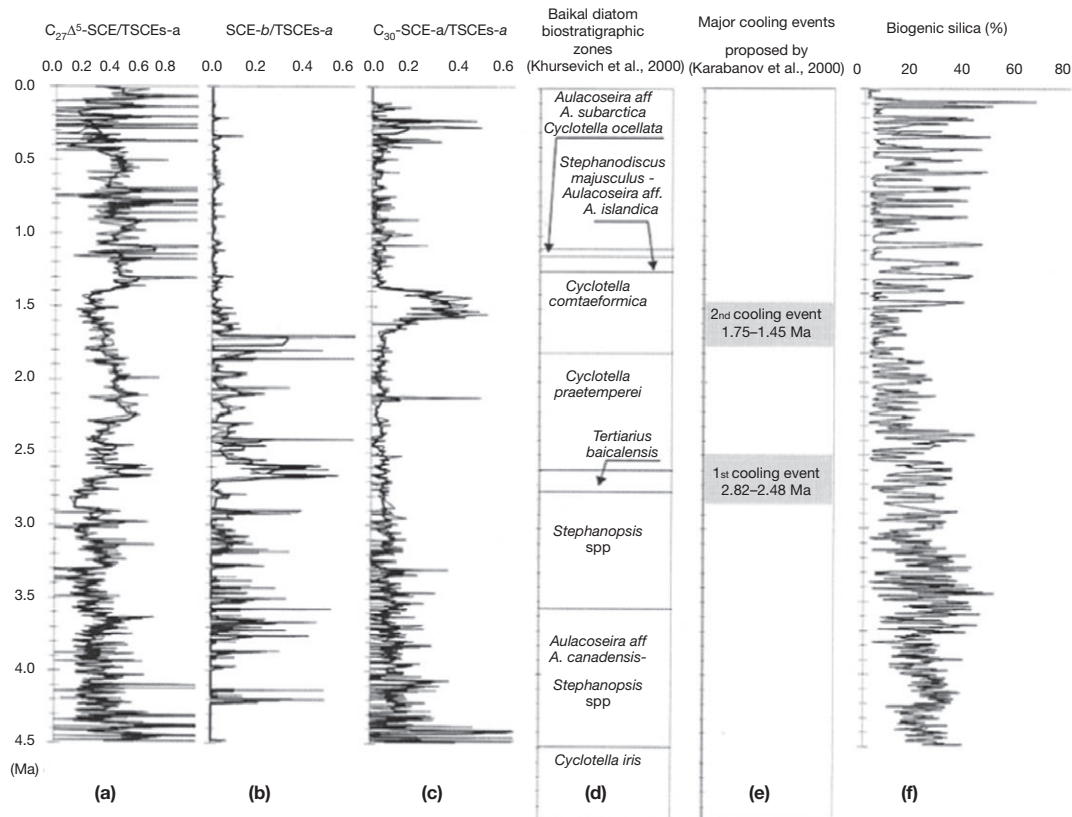


Figure 12 Depth profiles of ratios of $\text{C}_{27}\Delta^5\text{-SCE-a}$ (possible diatom marker) (a), SCE-b (b) (chlorophyte marker), and $\text{C}_{30}\text{-SCE-a}$ (dinoflagellate marker) (c) to total SCEs in the upper 200 m of BDP-98 drill core from Lake Baikal. Baikal diatom biostratigraphic zones (d); major periods of regional cooling (e) and % biogenic silica (f) are shown for comparison. Reproduced from Tani Y, Nara F, Soma Y, et al. (2009) Phytoplankton assemblage in the Plio–Pleistocene record of Lake Baikal as indicated by sedimentary sterol chlorin esters. *Quaternary International* 205: 126–136, with permission from Elsevier.

See also: Diatom Introduction. **Diatom Methods:** Salinity and Climate Reconstructions from Continental Lakes. **Paleobotany:** Overview of Terrestrial Pollen Data. **Paleoceanography, Biological Proxies:** Biomarker Indicators of Past Climate. **Paleolimnology:** Cladocera; Contributions of Paleolimnological Research to Biogeography; Lake Chemistry; Multiproxy Approaches; Overview of Paleolimnology; Physical Properties of Lake Sediments; Visible and Infrared Spectroscopical Applications. **Plant Macrofossil Methods and Studies:** Treeline Studies. **Quaternary Stratigraphy:** Continental Biostratigraphy.

References

- Bianchi TS, Engelhaupt E, McKee BA, et al. (2002) Do sediments from coastal sites accurately reflect time trends in water column phytoplankton? A test from Himmerfjärden Bay (Baltic Sea proper). *Limnology and Oceanography* 47: 1537–1544.
- Bianchi TS, Engelhaupt E, Westman P, et al. (2000) Cyanobacterial blooms in the Baltic Sea: Natural or human-induced? *Limnology and Oceanography* 45: 716–726.
- Britton G, Liaen-Jensen S, and Pfander H (1995) *Carotenoids*. Basel: Birkhäuser.
- Buchaca T and Catalan J (2008) On the contribution of phytoplankton and benthic biofilms to the sediment record of marker pigments in high mountain lakes. *Journal of Paleolimnology* 40: 369–383.
- Bunting L, Leavitt PR, Gibson CE, et al. (2007) Degradation of water quality in Lough Neagh, Northern Ireland, by diffuse nitrogen flux from a phosphorus-rich catchment. *Limnology and Oceanography* 52: 354–369.
- Chikaraishi Y, Matsumoto GI, Kitazato H, et al. (2007) Sources and transformation processes of pheopigments: Stable carbon and hydrogen isotopic evidence from Lake Haruna, Japan. *Organic Geochemistry* 38: 985–1001.
- Chondrogianni C, Ariztegui D, Rolph T, et al. (2004) Millennial to interannual climate variability in the Mediterranean during the Last Glacial Maximum. *Quaternary International* 122: 31–41.
- Cuddington K and Leavitt PR (1999) An individual-based model of pigment flux in lakes: Implications for organic biogeochemistry and paleoecology. *Canadian Journal of Fisheries and Aquatic Sciences* 56: 1964–1977.
- Das B, Vinebrooke RD, Sanchez-Azofeifa A, et al. (2005) Inferring sedimentary chlorophyll concentrations with reflectance spectroscopy: A novel approach to reconstructing changes in the trophic status of mountain lakes. *Canadian Journal of Fisheries and Aquatic Sciences* 62: 1067–1078.
- Ellegaard M, Clarke AL, Reuss N, et al. (2006) Multi-proxy evidence of long-term changes in ecosystem structure in a Danish marine estuary, linked to increased nutrient loading. *Estuarine, Coastal and Shelf Science* 68: 567–578.
- Enders SK, Pagani M, Pantoja S, et al. (2008) Compound-specific stable isotopes of organic compounds from lake sediments track recent environmental changes in an alpine ecosystem, Rocky Mountain National Park, Colorado. *Limnology and Oceanography* 53: 1468–1478.
- Fietz S, Nicklisch A, and Oberhänsli H (2007) Phytoplankton response to climate changes in Lake Baikal during the Holocene and Kazantsevo interglacials assessed from sedimentary pigments. *Journal of Paleolimnology* 37: 177–203.
- Frank HA, Young A, and Britton G, et al. (eds.) (1999) *The Photochemistry of Carotenoids*. In: *Advances in Photosynthesis and Respiration Series* 8: Dordrecht: Springer.
- Grimm B, Porra RJ, and Rüdiger W, et al. (eds.) (2006) *Chlorophylls and Bacteriochlorophylls: Biochemistry, Biophysics, Functions and Applications*. In: *Advances in Photosynthesis and Respiration Series* 25: Dordrecht: Springer.
- Guilizzoni P, Lami A, and Marchetto A (1992) Plant pigment ratios from lake sediments as indicators of recent acidification in alpine lakes. *Limnology and Oceanography* 37: 1565–1569.
- Hambricht KD, Zohary T, Eckert WW, et al. (2008) Exploitation and destabilization of a warm, freshwater ecosystem through engineered hydrological change. *Ecological Applications* 18: 1591–1603.
- Hodgson DA, McMinn A, Kirkup H, et al. (2003) Colonization, succession and extinction of marine floras during a glacial cycle: A case study from the Windmill Islands (east Antarctica) using biomarkers. *Paleoceanography* 18: 1067.
- Hodgson DA, Verleyen E, Squier AH, et al. (2006) Interglacial environments of coastal east Antarctica: Comparison of MIS 1 (Holocene) and MIS 5e (Last Interglacial) lake-sediment records. *Quaternary Science Reviews* 25: 179–197.
- Jeffrey SW, Mantoura RFC, and Wright SW (1997) *Phytoplankton Pigments in Oceanography*. Paris: UNESCO.
- Kashiyama Y, Ogawa NO, Kuroda J, et al. (2008) Diazotrophic cyanobacteria as the major photoautotrophs during the mid-Cretaceous oceanic anoxic events: Nitrogen and carbon isotopic evidence from sedimentary porphyrin. *Organic Geochemistry* 39: 532–549.
- Lami A, Guilizzoni P, and Maserferro J (1991) Record of fossil pigments in an alpine lake (L. Tovel, N. Italy). *Memorie dell'Istituto Italiano di Idrobiologia* 49: 117–126.
- Leavitt PR (1993) A review of factors that regulate carotenoid and chlorophyll deposition and fossil pigment abundance. *Journal of Paleolimnology* 9: 109–127.
- Leavitt PR, Carpenter SR, and Kitchell JF (1989) Whole-lake experiments: The annual record of fossil pigments and zooplankton. *Limnology and Oceanography* 34: 700–717.
- Leavitt PR, Cumming BF, Smol JP, et al. (2003) Climatic control of ultraviolet radiation effects on lakes. *Limnology and Oceanography* 48: 2062–2069.
- Leavitt PR and Findlay DL (1994) Comparison of fossil pigments with 20 years of phytoplankton data from eutrophic Lake 227, Experimental Lakes Area, Ontario. *Canadian Journal of Fisheries and Aquatic Sciences* 51: 2286–2299.
- Leavitt PR, Findlay DL, Hall RI, et al. (1999) Algal responses to dissolved organic carbon loss and pH decline during whole-lake acidification: Evidence from paleolimnology. *Limnology and Oceanography* 44: 757–773.
- Leavitt PR and Hodgson DA (2001) Sedimentary pigments. In: Smol JP, Birks HJB, and Last WM (eds.) *Tracking Environmental Changes using Lake Sediments. Terrestrial, Algal and Siliceous Indicators* 3: Dordrecht: Kluwer.
- Leavitt PR, Vinebrooke RD, Donald DB, et al. (1997) Past ultraviolet radiation environments in lakes derived from fossil pigments. *Nature* 388: 457–459.
- McGowan S, Britton G, Haworth EY, et al. (1999) Ancient blue-green blooms. *Limnology and Oceanography* 44: 436–439.
- McGowan S, Juhler RK, and Anderson NJ (2008) Autotrophic response to lake age, conductivity and temperature in two West Greenland lakes. *Journal of Paleolimnology* 39: 301–317.
- McGowan S, Leavitt PR, Hall RI, et al. (2005) Controls of algal abundance and community composition during ecosystem state change. *Ecology* 86: 2200–2211.
- McGowan S, Leavitt PR, Hall RI, et al. (2011) Interdecadal declines in flood frequency increase primary production in lakes of a northern river delta. *Global Change Biology* 17: 1212–1224.
- Michelutti N, Blais JM, Cumming BF, et al. (2010) Do spectrally inferred determinations of chlorophyll a reflect trends in lake trophic status? *Journal of Paleolimnology* 43: 205–217.
- Reuss N and Conley DJ (2005) Effects of sediment storage conditions on pigment analyses. *Limnology and Oceanography: Methods* 3: 477–487.
- Reuss N, Conley DJ, and Bianchi TS (2005) Preservation conditions and the use of sediment pigments as a tool for recent ecological reconstruction in four Northern European estuaries. *Marine Chemistry* 95: 283–302.
- Reuss N, Leavitt PR, Hall RI, et al. (2009) Development and application of sedimentary pigments for assessing effects of climatic and environmental changes on subarctic lakes in northern Sweden. *Journal of Paleolimnology* 43: 149–169.
- Squier AH, Hodgson DA, and Keely BJ (2004) A critical assessment of the analysis and distributions of scytonemin and related UV screening pigments in sediments. *Organic Geochemistry* 35: 1221–1228.
- Swain EB (1985) Measurement and interpretation of sedimentary pigments. *Freshwater Biology* 15: 53–75.
- Tani Y, Nara F, Soma Y, et al. (2009) Phytoplankton assemblage in the Plio–Pleistocene record of Lake Baikal as indicated by sedimentary sterol chlorin esters. *Quaternary International* 205: 126–136.
- Vinebrooke RD, Dixit SS, Graham MD, et al. (2002) Whole-lake algal responses to a century of acidic industrial deposition on the Canadian Shield. *Canadian Journal of Fisheries and Aquatic Sciences* 59: 483–493.
- Vinebrooke RD, Hall RI, Leavitt PR, et al. (1998) Fossil pigments as indicators of phototrophic response to salinity and climatic change in lakes of western Canada. *Canadian Journal of Fisheries and Aquatic Sciences* 55: 668–681.

Multiproxy Approaches

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Introduction

Paleolimnology is a multidisciplinary science, and thus it naturally lends itself to multiproxy studies. Nonetheless, most of the early paleolimnological research had a single-proxy focus, and, although single-proxy studies can be very informative and provide defensible conclusions, multiproxy approaches are becoming increasingly popular because they allow for more complete and robust environmental reconstructions. A search in the *Journal of Paleolimnology* using the keyword 'multiproxy' (with no filters) shows a steadily rising number of manuscripts that report on multiproxy data over the past 12 years (Figure 1). Multiproxy studies in the current paleolimnological arena are commonplace, and data spanning a wide breadth of topics from aquatic chemistry, catchment vegetation, and atmospheric deposition can often be found in a single manuscript.

Birks and Birks (2006) identify eight basic requirements for a multiproxy study, including (1) the need for clear research questions at the outset, (2) a strong leader, (3) carefully selected study sites and coring locations, (4) proper data storage and coordination, (5) a reliable sediment chronology, (6) clear presentation of the results, (7) the proper use of numerical techniques, and (8) interpretation and publication of large amounts of data. These are useful guidelines for any paleolimnological study but are especially critical requirements for a successful multiproxy investigation, as problems can become exacerbated when dealing with many collaborators and large amounts of data. There is no doubt that multiproxy studies require individuals who are detail oriented in the initial planning stages but keep the broad perspective of the study goals in mind, especially when writing manuscripts. Despite these apparent difficulties, the benefits of a multiproxy approach typically far outweigh the challenges involved.

Over the last several decades, there have been tremendous advancements in paleolimnology in both the development and interpretation of new proxies as well as in numerical techniques for analyzing complex stratigraphic data (Cohen, 2003; Smol, 2008). The number of analyses that can be performed on lake sediments to address a wide range of pressing environmental concerns is continually increasing. In fact, a limiting step in many multiproxy studies is the finite amount of sediment that is available for analyses. Reviews detailing the vast array of sedimentary proxies are given in Cohen (2003), Last and Smol (2001a,b), Smol (2008), and Smol et al. (2001a,b). In a properly designed study, the inclusion of several proxies provides for a greater understanding and often additional insights into different aspects of limnological processes.

Multiple Proxies to Obtain a Holistic Approach to Paleolimnological Reconstructions

The most common reason to adopt a multiproxy approach in a paleolimnological study is simply to gain a broader perspective

on environmental change than can be offered by using a single proxy. For example, a multiproxy study can provide a more complete history of change that includes independent data on the limnological regime, catchment, and atmospheric deposition. Often such robust paleoenvironmental reconstructions are required to refute possible competing hypotheses believed responsible for change, and it is only then that legislators and governments will take action to curb environmental threats. Perhaps one of the best examples of how a multiproxy paleolimnological approach was applied to tackle a common environmental issue is the example of acidification of lakes and rivers (see also Lake Chemistry).

Acid rain, or more appropriately, acidic deposition, occurs when emissions of sulfur and nitrogen oxides are converted into sulfuric and nitric acids in the atmosphere and return to the land in either dry or wet deposition. Although there are natural sources of acidic precipitation, the largest sources in many regions are anthropogenic, including the burning of coal and other fossil fuels and the smelting of sulfur-rich ores. The acidification of lakes and waterways was first recognized as a potential environmental problem in the 1950s, although acidic deposition had been occurring on a large scale since the onset of the Industrial Revolution, over a century earlier. The effects of acidic deposition on lakes and rivers are complex but typically involve a marked decrease in lake water pH in poorly buffered systems. The acidification of catchment soils can also impact aquatic chemistry. For example, inorganic monomeric aluminum (Al^{3+}), which, at certain concentrations, is toxic to fish and other biota, is mobilized from acidifying catchment soils and eventually leaches into lakes and waterways. Other impacts on aquatic chemistry include declines in the concentration of dissolved organic carbon (DOC), which acts as a natural sunscreen protecting aquatic biota from harmful ultraviolet rays. These changes in water chemistry have dramatic impacts on the size and composition of algal and invertebrate communities, as well as noticeable signs of stress on fish populations, including decreased fecundity and in some instances complete extirpation. Many of these effects have been well documented during an experimental acidification of a boreal softwater lake (Lake 223) in the Experimental Lakes Area of northwestern Ontario, Canada (Schindler et al., 1985).

Acid rain was a highly politicized issue, especially during the peak of its public concern in the 1980s. Governments were initially reticent in their actions to pass legislation that curbed acid-causing emissions until it was unequivocally demonstrated that such emissions were indeed responsible for acidifying lakes and rivers. Opponents of taking legislative action claimed (often with no real evidence) that the acid lakes were either always acidic or that the acidification was part of a natural long-term soil acidification process that had been occurring since lake formation. Others argued that it was changes in catchment vegetation that led to the acidification, not the emissions. With little to no long-term monitoring data,

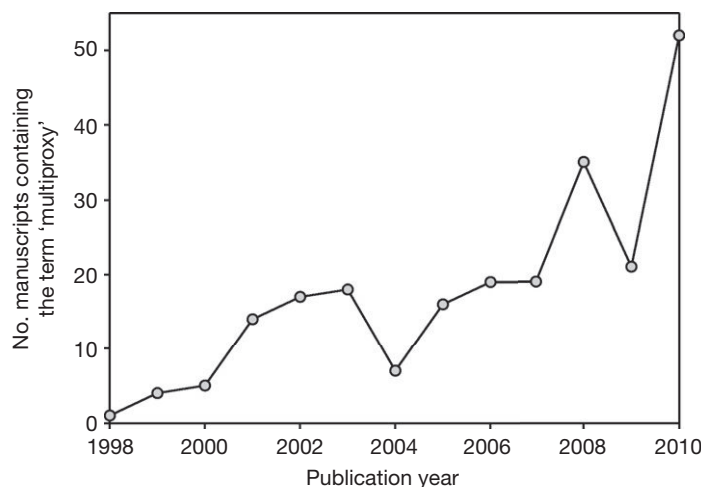


Figure 1 Search results, with no filters, showing the number of manuscripts that contain the term 'multiproxy' over the past 12 years in the *Journal of Paleolimnology*.

a multiproxy paleolimnological approach was the only way to answer critical questions such as: Have the lakes acidified? If so, by how much? And what are the causes of acidification?

Large-scale paleolimnological acidification programs were undertaken in the United States (Charles and Smol, 1990; Charles and Whitehead, 1986; Whitehead et al., 1990), Canada (Dixit et al., 1995), and the United Kingdom and Scandinavia (Battarbee et al., 1990). The first step was to identify and develop proxies that could be used to infer lake water acidity. This was accomplished through surface sediment training sets, or calibrations sets, mainly involving diatoms and, to a lesser extent, chrysophytes. Modern-day assemblages of diatoms and chrysophytes were shown to have distributions that were strongly controlled by lake water pH (Battarbee et al., 2010). This allowed for the computation of diatom and chrysophyte-based pH transfer functions that could be applied to fossil assemblages preserved in sediment cores to quantitatively infer past lake water pH.

An oft-cited example of diatom- and chrysophyte-inferred pH change is from Baby Lake, located near Sudbury, Ontario, Canada (Dixit et al., 1992). For several decades, the Sudbury region was one of the largest point sources of sulfur emissions in the world, due to the massive mining and smelting activities that occurred in the area. The large output of sulfur emissions, coupled with the poor buffering capacity of the underlying geology, made Sudbury area lakes ideal study sites to investigate the impacts of acidic deposition. More importantly, local governments and scientists had the foresight to undertake water monitoring programs, given the potential for damage from the sulfur emissions. Baby Lake was one such site where pH measurements were available, beginning in 1968. The historical pH measurements provided ground-truthing data for the diatom and chrysophyte transfer functions. The diatom and chrysophyte pH inferences showed that, prior to mining activities, Baby Lake had a circumneutral pH, near pH 6.5. The inferred pH showed a dramatic decline beginning around 1940, reaching a low of pH 4.2 by the mid-1970s, coincident with the period of intensive smelting activities and massive sulfur emissions. With the closure of a nearby smelter, the lake water pH started to show signs of recovery back to pre-industrial pH levels. Importantly, the trends in the measured

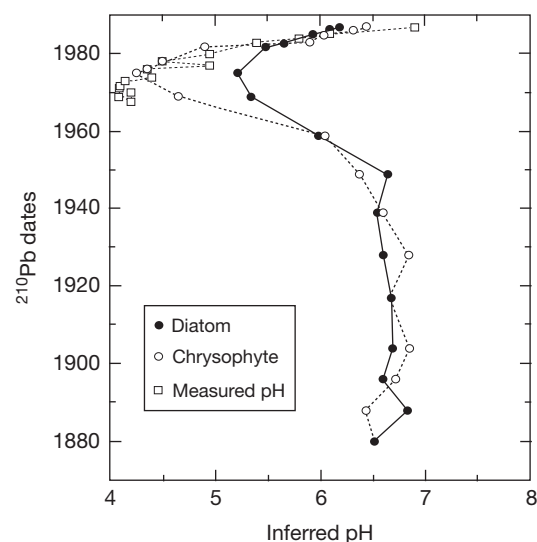


Figure 2 Diatom (solid circles) and chrysophyte (open circles) inferred pH reconstructions from Baby Lake, near Sudbury, Ontario. Also shown are the measured pH data (squares) dating back to 1968. Reproduced from Hutchinson TC and Havas M (1986) Recovery of previously acidified lakes near Coniston, Ontario, Canada, following reductions in atmospheric sulphur and metal emissions. *Water, Air and Soil Pollution* 29: 319–333; Dixit SS, Dixit AS, and Smol JP (1992) Assessment of changes in lakewater chemistry in the Sudbury area lakes since preindustrial times. *Canadian Journal of Fisheries and Aquatic Sciences* 49: 8–16, with permission.

pH data closely matched those of the paleolimnological data (Figure 2). Similar profiles of marked pH declines, coincident with the onset of large-scale industrial activities, were reconstructed for many other Sudbury area lakes and indeed in many acid-sensitive regions around the world (Smol, 2008).

The timing of species changes and lake water pH inferences have also been used in conjunction with geochemical analyses to confirm the source of pollution. The paleolimnological acidification program in the United States focused on acid-sensitive regions that received large amounts of acidic

deposition, such as the Adirondack Mountains in New York State. One of the study sites in the Adirondack Mountains was Deep Lake, a headwater lake that was known to be acidic (pH ~ 4.8) and fishless in the 1980s. Lake water pH reconstructions showed that Deep Lake was initially acidic in preindustrial times, but became increasingly acidic during the 1900s, coincident with increases in acidic emissions to this region (Charles et al., 1990). Sedimentary analyses of coal and oil soot particles, polycyclic aromatic hydrocarbons (PAHs), and metals such as lead and vanadium showed clear increases coincident with the onset of large-scale fossil fuel combustion and also with the timing of aquatic species changes and the beginning of inferred pH declines (Figure 3). The multiproxy approach of using both species and geochemical data further implicated acidic emissions as the cause of widespread lake acidification.

There was mounting evidence demonstrating the linkages between industrial emissions and acidifying waterways, but there were still claims that long-term changes in catchment soils were wholly or partly responsible for lake acidification. An example from Scandinavia involves the argument that acidification had occurred as a result of a decline in traditional forms of agriculture, which led to afforestation and hence an increase in the humus content of soil, which in turn resulted in the acidification of lakes due to shifts in ion exchange processes in watershed soils (Rosenqvist, 1977, 1978). To test this hypothesis, Anderson and Korsman (1990) analyzed pollen grains preserved in lake sediments in order to investigate the role that vegetation shifts and land-use changes may have had on lake water pH. They used an historical analog approach where they argued that a well-documented decline of farming activities in the Hälsingland region of Northern Sweden during $\sim$ AD 500 (the so-called migration period) should result in similar afforestation and subsequent pH declines in lakes. Their pollen analyses showed that, similar to present-day, afforestation did occur $\sim$ AD 500; however, the diatom data did not show any signs of acidification. This was in sharp contrast to modern times, where one study lake showed a recent acidification, as predicted with acidic deposition.

The combination of aquatic, terrestrial, and geochemical proxies used in the acid rain programs resulted in holistic

paleolimnological reconstructions and provided clear evidence that acidic emissions were largely responsible for lake acidification. Alternative hypotheses suggested by the opponents of legislative action were strongly and consistently refuted in many studies. The strength of this multiproxy approach ultimately resulted in legislation that led to marked reductions in acid-forming emissions.

Multiple Proxies to Identify Stressors and Effects

Many environmental or ecological studies can identify either stressors or effects, but rarely both. A multiproxy paleolimnological approach has the advantage of being able to identify environmental stressors and the associated effects, as was clearly shown in the aforementioned acid rain programs of the 1980s and 1990s. In the preceding section, it was discussed how multiple proxies identified acidic emissions as the key stressor involved in widespread lake acidification. However, one of the most disconcerting effects of lake acidification, at least to the public at large, was the disappearance of fish species. The effects of acidification on fish reproduction and survival were well documented during the artificial acidification of a softwater boreal lake (Lake 223) in the Experimental Lakes Area of northwestern Ontario, Canada (Schindler et al., 1985). However, in lakes with little or no monitoring data, it proved difficult to link the decline in fish populations with acidification. It is possible that fish were never present in some naturally acidic lakes. If fish populations could be reconstructed from lake sediments, it would be possible to link their demise with the onset of acidification, or even to determine if a lake was naturally fishless.

Historical fish populations are difficult to reconstruct from lake sediments because fish leave few fossil remains in the sedimentary record. Often, more indirect techniques are required. One common approach is the use of subfossil remains (mandibles) of *Chaoborus* larvae (Figure 4), which live in the limnetic zone of lakes, feeding on zooplankton, and can tolerate a wide range of pH levels. *Chaoborus* larvae differ in their patterns of vertical migrations, as well as size and color, all of

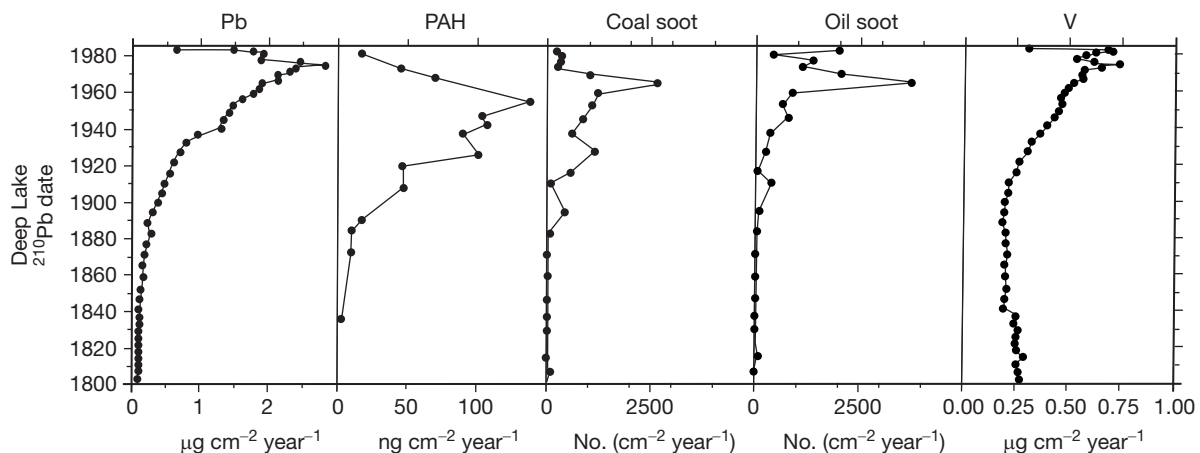


Figure 3 Measured concentrations of lead (Pb), polycyclic aromatic hydrocarbons (PAHs), coal soot, oil soot, and vanadium (V) measured in the Deep Lake sediment core from the Adirondack Mountains (New York, USA). Reproduced from Charles DF, Binford MW, Fry BD, et al. (1990) Paleocological investigation of recent lake acidification in the Adirondack Mountains, N.Y. *Journal of Paleolimnology* 3: 195–241, with permission.



Figure 4 Photo of *Chaoborus* larva (a), and preserved mandibles of *C. americanus* (b), an indicator of fishless conditions used in the Uutala (1990) study in Adirondack Park (New York, USA). Images courtesy of Joshua Kurek.

which can affect their susceptibility to fish predation. Thus, the presence or absence of planktivorous fish is one of the main factors that can affect the distribution of *Chaoborus*. Certain taxa (e.g., *Chaoborus punctipennis*, *Chaoborus albatus*, *Chaoborus trivittatus*, and *Chaoborus flavicans*) exhibit diurnal migration and only enter the upper water column to feed at night when fish, which are visual predators, have limited vision. The only known taxon in the Northeastern United States and Canada that does not exhibit diurnal vertical migration is *Chaoborus americanus*, and thus, it rarely occurs in lakes with planktivorous fish. *C. americanus* has colonized lakes from which fish have been extirpated, and it has been extirpated itself in lakes from which fish have been reintroduced (Evans, 1989; Nilssen, 1984).

In the Adirondack region of New York State, USA, Uutala (1990) examined *Chaoborus* remains to assess the status of fish populations in five lakes over the past several centuries. One of the study sites, Windfall Pond, was a lake that had not acidified and currently had viable fish populations. The sediment core from this site contained the remains of only *Chaoborus* taxa that practiced vertical migration, indicating the long-term presence of fish in this lake. Other study sites including Brooktrout Lake, Deep Lake, and Upper Wallace Pond had all acidified in recent years and were all currently fishless. The appearance of the nonmigratory *C. americanus* in the recent sediments of Brooktrout Lake (~1970s) and Deep Lake (~1940s) indicates that these two lakes lost their fish populations in recent decades. Upper Wallace Pond showed a dominance of *C. americanus* throughout its entire core, indicating this lake likely never supported planktivorous fish.

The combination of *Chaoborus* analysis with diatom-inferred pH reconstructions and/or analyses of metal chemistry from the Adirondack lake sediment cores clearly identified

both stressor (acidification) and effects (fish extirpations). This type of multiproxy approach has direct management implications as it identifies which lakes should qualify for fish restocking programs.

Multiple Proxies to Resolve Competing Hypotheses of Environmental Change

Aquatic ecosystems change through time in response to natural and anthropogenic stressors. Multiple proxies are often required in order to bring resolution to competing hypotheses of change. Several proxies can be used in support of one another (i.e., the weight of evidence approach), thereby strengthening paleoenvironmental interpretations, or can be used in what Birks and Birks (2006) refer to as 'data splitting' where one proxy is used to interpret stratigraphic changes in another independent proxy.

An example of 'data splitting' and the use of supporting proxies is provided by Wolfe et al. (2001) who recorded marked and near-synchronous change in fossil diatom assemblages in two alpine lakes in the Colorado Front Range (USA). The diatom assemblage prior to the ~1900s was characteristic of oligotrophic alpine lakes and was composed primarily of *Aulacoseira* taxa, small benthic *Fragilaria sensu lato* taxa, and various *Achnanthes* taxa (Figure 5). After ~1900, the planktonic taxa *Asterionella formosa* and *Fragilaria crotonensis* increased in abundance, and post ~1950, *A. formosa* became the dominant taxon in both study sites. Prior research conducted in Arctic regions showed marked shifts in diatom assemblages (reviewed in Smol and Douglas, 2007) that were attributed to lengthened growing seasons associated with recent climate warming. The increased percentage of planktonic taxa recorded in the alpine lakes could plausibly be explained by climate warming. This is especially true given the remote location of the alpine lakes and the fact that high-altitude ecosystems, similar to high-latitude ecosystems, are among the most susceptible to climate change. However, Wolfe et al. had reason to believe that the recorded diatom changes in their specific study lakes were also driven by nutrient additions, specifically atmospheric deposition of fixed nitrogen from anthropogenic sources. The reasons for this suspicion included the fact that *A. formosa* and *F. crotonensis* are well-known indicators of nutrient enrichment, and their increase suggests a response to nutrients rather than climate. In addition, the study sites were located along an air mass trajectory that carries atmospheric pollution from the highly populated Denver-Fort Collins urban corridor (~2.4 million people at the time of the study) in an easterly manner up the Colorado Front Range.

To test their hypothesis, Wolfe et al. analyzed stable isotopes of nitrogen ($\delta^{15}\text{N}$) in the sediments. In both study lakes, they recorded a trend toward lighter isotopic values post ~1950 coincident with the shift from oligotrophic to mesotrophic diatom taxa (Figure 5). The decline in sedimentary $\delta^{15}\text{N}$ values was suggested to reflect increasing additions of isotopically depleted N compounds from areas east of the Colorado Front Range. The isotopic depletions are consistent with fertilizer produced from the Haber-Bosch process, which results in NH_3 that has $\delta^{15}\text{N}$ values of $\sim 0 \pm 3\%$. In addition,

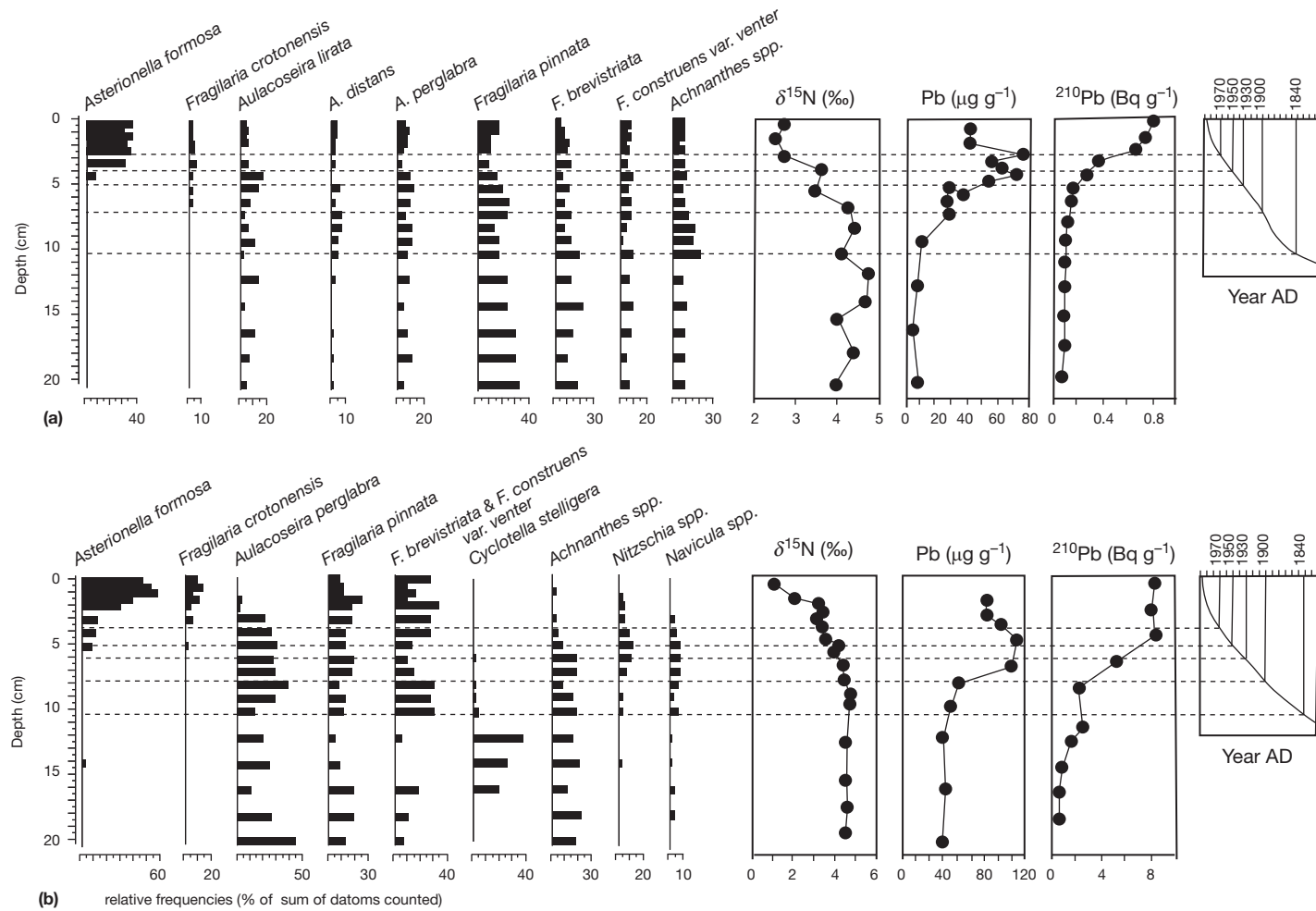


Figure 5 Fossil diatom stratigraphies and profiles of sedimentary $\delta^{15}\text{N}$, Pb, and ^{210}Pb dating profiles for Sky Pond (a) and Lake Louise (b), two lakes in the Colorado Front Range that have been affected by anthropogenic atmospheric N-deposition. Reproduced from Wolfe AP, Baron JS, and Cornett RJ (2001) Anthropogenic nitrogen deposition induces rapid ecological change in alpine lakes of the Colorado Front Range (USA). *Journal of Paleolimnology* 25: 1–7, with permission.

confined animal feeding operations, which rose from 1.4 to 3.7 million head between 1940 and 1970, have NH_3 emission that range from -15 to -9% . These isotopically depleted N sources are consistent with the $\delta^{15}\text{N}$ declines recorded in the study cores.

Although sediment diagenetic effects can alter $\delta^{15}\text{N}$ values, previous research has suggested that the $\delta^{15}\text{N}$ in sediments is typically representative of the isotopic content in the water column (Brenner et al., 1999). Moreover, the timing of the isotopic depletion is consistent with the rise in population, which tripled between 1950 and 1995, and the associated increase in industrial and agricultural activities. Still, it was previously suggested that the study sites, at elevation >3300 m above sea level, were above the limit for deposition of anthropogenic emissions by upslope air masses. To further dispel this notion, Wolfe et al. analyzed the sediment for stable lead (Pb) concentrations. The lead profile shows an increase in concentration in pre-1930s sediments consistent with mining and smelting activities in the area (Figure 5). Further increases occur after this time associated with the addition of alkyl Pb additives to gasoline in the early 1920s. Both profiles also record a synchronous decline in lead concentrations that mark the placement of strict controls on leaded gasoline beginning in the 1970s. The lead data, together with $\delta^{15}\text{N}$ isotopic data, clearly demonstrate that anthropogenic emissions are capable of reaching the elevation of the study sites. The Wolfe et al. study illustrates some of the advantages of ‘data splitting’ as well as using proxies in support of one another to identify the key environmental stressor causing ecological change.

Multiple Proxies to Detect Differential Limnological Responses in Spatially Separated Lake Habitats

Lakes are complex ecosystems, and a single proxy does not always reveal sufficient information to demonstrate how a lake responds to an environmental stressor. For example, in some instances, a decoupling in limnological processes can occur between spatially separated habitats. By selecting proxies that characterize different lake strata, it is possible to track changes in distinct lake zones. For example, planktonic diatoms will reflect epilimnetic conditions, and profundal chironomids will record changes in the hypolimnion.

Little et al. (2000) used a multiproxy approach involving (1) diatoms to reconstruct epilimnetic nutrient concentrations and (2) chironomids to infer hypolimnetic anoxia in Gravenhurst Bay (Ontario, Canada), a water body that has well-documented eutrophication and recovery events. A typical lake response to eutrophication is that the added nutrients increase epilimnetic primary production, resulting in increased organic matter settling to the hypolimnetic water, which in turn becomes anoxic due to decomposition of the excess organic material. Little et al. used a diatom-based total phosphorus (TP) inference model to track the onset of eutrophication at Gravenhurst Bay, which began with the arrival of European settlers in the mid-1800s. As expected, hypolimnetic oxygen levels, inferred from chironomid assemblages, indicated that anoxic conditions became more prevalent with the onset of this eutrophication. The highest diatom-inferred TP (DI-TP) values occurred around 1950, which also corresponded with the period of the most severe

hypolimnetic oxygen deficits, as marked by the disappearance of profundal chironomid taxa. Following 1972, sewage inputs to the bay declined and the diatom assemblages tracked this recovery, showing decreases in inferred TP and a shift back to oligotrophic taxa. Interestingly, the benthic chironomids showed no parallel recovery. The proxy data are entirely consistent with monitoring data that show continued declines in surface water TP concentrations and persistent anoxia in the deepwater. The reason for this decoupling remains unclear, but the study serves as an example of the importance of using different proxies in order to track changes in spatially separated lake habitats.

MacDonald et al. (2009) provide another example of a decoupling between chironomids and indicators of lake production in a multiproxy investigation in a small lake in the Northwest Territories of Canada. In this study, MacDonald et al. used organic matter content by loss-on-ignition (LOI) and biogenic silica (BSi) as indicators of aquatic production in conjunction with chironomid-inferred July air temperature and summer water temperatures to reconstruct climatic conditions over the past ~ 2000 years. Their records of LOI, BSi, and chironomid assemblages are generally in good agreement, showing shifts in lake production that are consistent with warmer and more productive conditions prior to the Little Ice Age (LIA), as well as declines in lake production and inferred temperatures during the LIA. The increased warming of the twentieth century is marked by LOI and BSi concentrations that equal or exceed that of any other time during the past 2000 years. However, the chironomid-inferred temperatures show no evidence of any recent warming trend. It is hypothesized that enhanced lake water stratification during twentieth century warming has led to decreased oxygen and lower temperatures of the hypolimnion of the lake, resulting in an ambiguous chironomid response with respect to warming. This study demonstrates the value of a multiproxy approach to better understand how different lake zones may respond to recent climate change.

Multiple Proxies to Track Multiple Environmental Stressors

Lakes are being impacted by multiple stressors with the pervasive effects of climate warming superimposed on more regional stressors, including catchment development, eutrophication, acidification, and deposition of atmospheric contaminants, as well as newly identified threats such as calcium depletion (Jeziorski et al., 2008). All of these stressors lead to widespread concern over the general health of lakes and can frustrate lake managers and government agencies with respect to identifying the most important stressors affecting lakes. Multiproxy studies can pinpoint the principle stressor affecting a given set of lakes, which greatly assists regulating agencies in effectively spending their often limited budgets for lake remediation projects. Below, we discuss two examples of using multiple proxies to assess the role of multiple stressors on ecological change. The first example concerns the issue of increasing algal blooms in Lake of the Woods (LoW), Ontario, Canada, and the second involves assessing the role of animals as biological vectors of nutrients and contaminants into freshwaters.

Assessing the Factors Leading to Increased Algal Blooms in Lake of the Woods (Ontario, Canada)

Cyanobacterial blooms are increasingly reported as a water quality concern in lakes globally. These blooms obviously degrade the recreational enjoyment of the lakes but also have serious ecological ramifications as they increase turbidity and kill off macrophytes that provide important habitats for fish and invertebrates. When the blooms die off, they can create anoxic conditions that can lead to fish kills. Moreover, cyanobacteria constitute a public health threat as some taxa produce toxins that can lead to serious and sometimes fatal diseases (Pearl and Huisman, 2008). Nutrient enrichment of water by intensive shoreline development, as well as industrial and agricultural activities, often drives the growth of these algal blooms, but other factors such as climate warming and increased thermal stratification can also be important.

Lake of the Woods (LoW) is an international water body spanning the Canadian provinces of Ontario and Manitoba and the state of Minnesota in the United States. The lake is large (3850 km²) and contains numerous basins with nearly 14 000 islands and hundreds of inlets and bays. It has a long history of human disturbance dating back to the 1850s, including land clearance by European settlers, logging and sawmill activities, and at the turn of the twentieth century, the construction of dams for water level regulation and hydroelectric power. The lake is popular for recreational activities, and there is extensive cottage development around its shoreline. In recent years, lake residents have reported an increase in the frequency and intensity of late-summer cyanobacterial blooms in the northern portion of the lake (DeSellas et al., 2009). However, there are little long-term monitoring data to assess whether these blooms represent an actual deviation from natural variability.

Rühland et al. (2010) conducted a paleolimnological study to investigate the role that multiple stressors, including nutrient enrichment, dam construction, and recent warming, may have had on exacerbating nuisance algal blooms in LoW. To do so, they recovered sediment cores from three separate basins that varied in trophic status, due to differences in the degree of their connectivity to the Rainy River, which accounts for over 70% of the inflow to LoW and is the main source of nutrient inputs into the lake. Two of the coring sites, PP1 and Bigstone Bay, are meso-to-eutrophic basins and are sites where recent algal blooms have been recorded. The other coring site, Whitefish Bay, is hydrologically isolated from the nutrient-rich flow from the Rainy River and is currently oligotrophic; no recent algal blooms have been documented there. DI-TP models were used to estimate the timing, magnitude, and direction of water quality change in these three basins. In addition, sedimentary chlorophyll *a* concentrations were inferred, using spectral reflectance (Michelutti et al., 2010), to provide an indication of whole-lake production.

Diatom assemblages at all sites showed marked changes in response to hydro-management activities and recent climate warming, indicating an ecological response to multiple stressors. However, at the nutrient-rich sites (PP1 and Bigstone Bay) where nuisance blooms occur, DI-TP showed either little-to-no change over time or even decreases over the last few decades, which is entirely consistent with remediation efforts since the mid-1970s, resulting in a steady decline in TP loading

from the Rainy River. This indicates that increased TP concentration is not the main trigger of recent nuisance algal blooms in LoW. The chlorophyll *a* reconstructions show increases over the past several decades in the two nutrient-rich sites (PP1 and Bigstone Bay), but no change in the oligotrophic Whitefish Bay (Figure 6), and therefore are likely tracking increased intensity of cyanobacterial blooms. The chlorophyll *a* reconstructions provide strong evidence to support the lake residents' observations that nuisance algal blooms have been increasing in recent years in the northern portion of the lake. Significant ($p \leq 0.05$) correlations between temperature and chlorophyll *a* reconstructions indicate that temperature likely plays a key role in the recent algal blooms. The possible mechanisms by which temperature is driving an increase in algal blooms include increased duration and strength of thermal stratification, which favors the buoyant cyanophytes (e.g., Pearl and Huisman, 2008), longer ice-free seasons affecting nutrient availability (e.g., seasonal N-limitation), and warmer temperatures that may influence internal TP loading.

The multiple proxies used in the Rühland et al. study answered several important lake management questions pertaining to cyanobacterial blooms in LoW. First, the data indicated that lake water TP may not be the ideal indicator of aquatic ecosystem health. Second, the chlorophyll *a* reconstructions confirmed the lake residents' perception that nuisance algal blooms were indeed increasing in recent years. Third, the diatom data (along with supporting water quality monitoring data) demonstrated that the recent algal blooms were not driven by increases in lake water TP alone, and the

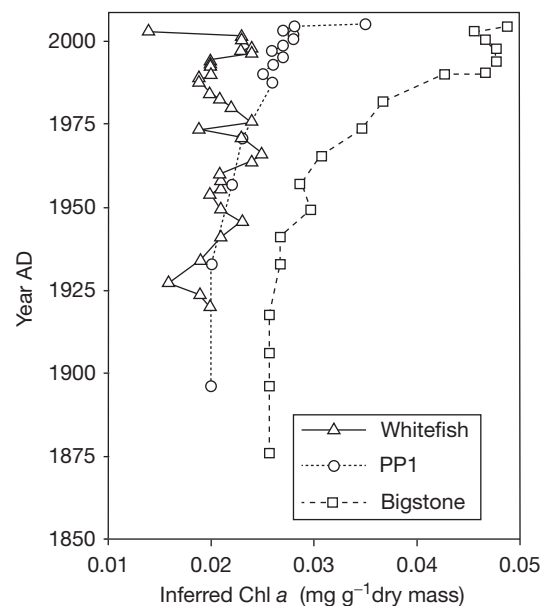


Figure 6 Inferred sedimentary chlorophyll *a* profiles in the nutrient-rich sites PP1 and Bigstone Bay, as well as for the oligotrophic basin, Whitefish Bay in Lake of the Woods. An increase in recent algal blooms has been documented in PP1 and Bigstone Bay, but not in Whitefish Bay. Reproduced from Michelutti N, Blais JM, Cumming BF, et al. (2010) Do spectrally inferred determinations of chlorophyll *a* reflect trends in lake trophic status? *Journal of Paleolimnology* 43: 205–217, with permission.

triggers for nuisance blooms are complex, with climate change being an important contributor to the increased blooms.

Animals as Biological Vectors of Nutrients and Contaminants

Animals that cross ecosystem boundaries can transport nutrients and other substances from one food web to another, often with beneficial consequences for the recipient ecosystem. In many instances, this transport occurs from marine to terrestrial environments. For example, sea turtles that spend most of their life feeding in the ocean subsidize nutrient-poor dune ecosystems, when they come ashore to nest, via marine-derived nutrients from waste products that enhance vegetation and promote dune stabilization (Hannan et al., 2007). Other examples of 'biovectors' that transport and focus nutrients from the ocean to inland sites include sea lions, whales, salmon, and seabirds (Blais et al., 2007; Fariña et al., 2003; Michelutti et al., 2009). When the waste products from these biovectors are deposited in or around lakes, paleolimnology can be used to track the associated limnological effects as well as changes in population dynamics of the biovectors themselves.

The decline of fish populations in many coastal regions around the world has been attributed to many factors including overfishing, pollution, habitat destruction, and climate change. The lack of long-term monitoring data for most fish stocks makes it impossible to identify the potential stressors responsible for the declines. Sockeye salmon (*Oncorhynchus nerka*) represent a commercially and culturally important salmon species on the west coast of North America. Sockeye salmon hatch in nursery lakes, where they spend the first 1–3 years of their lives before heading out to the ocean where they spend the next 2–3 years and accumulate >95% of their biomass (Naiman et al., 2002). In the ocean, the salmon feed on plankton, squid, and small fish, and due to their high trophic position, they preferentially accumulate greater levels of the heavy ^{15}N isotope, relative to the light ^{14}N isotope. The adult salmon return upstream to the same nursery lakes where they hatched in order to spawn, and then die shortly thereafter. Upon their death, all of the marine-derived nutrients, including the elevated $\delta^{15}\text{N}$ signal, are released to the nursery lake. The elevated $\delta^{15}\text{N}$ signature in the salmon tissues is preserved in the sediments and thus provides an indirect proxy of past salmon abundances.

Finney et al. (2000) tracked past salmon fluctuations in five Alaskan nursery lakes (including two control lakes with no salmon) using a multiproxy approach that included (1) diatoms, to infer nutrient levels; (2) cladoceran concentrations, to infer food availability for the juvenile salmon; and (3) sedimentary $\delta^{15}\text{N}$, as an indirect proxy of salmon abundances. They showed a relationship between salmon abundances and climate in the North Pacific, with greater abundances (as inferred by $\delta^{15}\text{N}$) recorded during periods of warm sea surface temperatures in the Gulf of Alaska (Figure 7). Although climate was identified as an important factor, Finney et al. also showed how overfishing contributed to a lengthy twentieth century collapse of a commercial fishery in one of their study sites. Specifically, the harvesting of salmon in the ocean reduces the amount of carcass-derived nutrient subsidies to the nursery lakes. Lower nutrient inputs into the oligotrophic nursery lakes translates into lower algal abundances and thus lower abundances of zooplankton, which

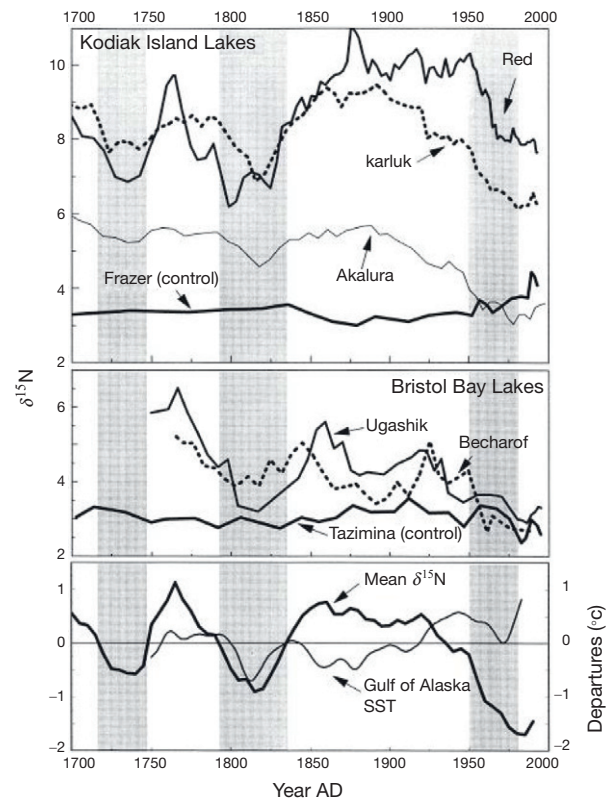


Figure 7 Sedimentary $\delta^{15}\text{N}$ profiles from the sockeye salmon nursery lakes (Red, Karluk, Akalura, Ugashik, and Becharof) and two control lakes (Frazer and Tazimina) from the Kodiak Island and Bristol Bay region of Alaska. The bottommost panel shows the mean $\delta^{15}\text{N}$ signature for the nursery lakes plotted against the sea surface temperature (SST) from the Gulf of Alaska. Reproduced from Finney BP, Gregroy-Eaves I, Sweetman J, Douglas MSV, and Smol J (2000) Impacts of climatic change and fishing on Pacific salmon abundance over the past 300 years. *Science* 290: 795–799, with permission.

represent a critical food source for the juvenile salmon. The overharvesting of adult salmon in the ocean before they can return to their nursery lake effectively decouples this delicate feedback cycle between nutrients, food availability, and juvenile salmon abundances.

The Finney et al. study illustrated the importance of biovector nutrient subsidies to the salmon nursery lakes. In fact, marine-derived nutrients delivered by biovectors often shape the ecosystem structure and function of many coastal and inland sites. An unfortunate irony is that, in addition to life sustaining nutrients, biovectors can also shunt contaminants to receptor sites, sometimes at concentrations that are detrimental to aquatic life (Blais et al., 2007). This occurs because biovectors typically occupy a high trophic position in the marine food web and thus they biologically accumulate and concentrate contaminants, which can be transferred to recipient ecosystems via waste products and carcasses. Krümmel et al. (2003, 2005) analyzed polychlorinated biphenyl (PCB) concentrations in the sediments from nursery lakes that receive a high return of sockeye salmon. They showed that salmon represent a more potent transport pathway for PCBs relative to atmospheric processes.

Seabirds may be the most globally relevant biovectors because they are found on every continent, often in large colonies of tens of thousands of individuals and some that even number millions. Adopting a similar approach to that developed in the salmon research, Michelutti et al. (2009) examined the impact of a large colony of seabirds on a series of freshwater ponds located near Cape Vera on Devon Island (High Arctic Canada). The study site at Cape Vera is home to 20 000+ Northern fulmars (*Fulmarus glacialis*) that nest on steep-sided cliffs above a small coastal foreland containing numerous freshwater ponds. Using a multiproxy approach, Michelutti et al. showed that the northern fulmars had undergone a climate-induced population increase over the last several decades (as inferred by sedimentary $\delta^{15}\text{N}$) and that contaminants including PCBs and metals such as cadmium can be directly linked to biotransport by seabirds. Biotransported contaminants thus provide an additional, independent means of tracking past seabird population dynamics.

Typically, research on biovector transport focuses on gregarious animals that congregate in dense and populous communities because large numbers are required for the biotransported nutrients and other substances to have any significant ecological effect. However, one large animal, such as a whale, can have a similar effect to thousands of smaller-bodied species. Douglas et al. (2004) used a multiproxy approach to track the influence of Thule Inuit whaling activities on the ecology of Arctic ponds. The Thule people were the first whalers in the High Arctic, and they would butcher their kills on land, often near freshwater ponds. This is evidenced by the numerous bowhead whale bones that often litter the landscape near former Thule camps. Douglas et al. recovered a sediment core from a pond on Southeastern Somerset Island that predated Thule occupation. They were able to track the arrival of the whalers over eight centuries ago by showing an enrichment in sedimentary $\delta^{15}\text{N}$ (bowhead whales are enriched in $\delta^{15}\text{N}$) and a coincident shift in diatom assemblages, namely, the rise of *Pinnularia balfouriana*, a moss epiphyte that increased in abundance as mosses increased in abundance in response to nutrient increases from the flensed whale carcasses.

The effects of biovector transport on aquatic ecosystems include changes to water quality, nutrient availability, contaminant exposure, and biological assemblages. In order to document the multiple impacts that biovectors have on lakes and ponds, the inclusion of multiple proxies that provide independent, but related, ecological information is required.

Conclusions

The last several decades have witnessed tremendous advancements in the field of paleolimnology with respect to proxy development, the application of new proxies to environmental problems, and statistical techniques for analyzing complex datasets. These advancements have paved the way for multiproxy approaches, which are becoming increasingly popular because they offer many advantages over more traditional single-proxy studies. These advantages include the ability to (1) obtain holistic environmental reconstructions, (2) identify both environmental stressors and associated ecological effects, (3) resolve competing hypotheses of environmental change, (4) track

differential limnological responses within spatially separated lake habitats, and (5) track multiple environmental stressors.

Multiproxy studies in paleolimnology are now the norm. This is driven in part by the increasing complexity of environmental issues facing freshwater ecosystems, and society in general. The combinations of threats from many well-known stressors including climate change, eutrophication, and acidification are having heretofore unpredicted and often complex effects, and newly discovered limnological threats such as Ca depletion loom large. Multiproxy studies present several challenges, the biggest of which include coordinating all participants involved and a willingness and capacity to synthesize large and complex datasets. Carefully designed and well-executed multiproxy studies have the potential to address newly emerging environmental issues all within a paleolimnological framework.

See also: **Paleolimnology:** Contributions of Paleolimnological Research to Biogeography; Lake Chemistry.

References

- Anderson NJ and Korsman T (1990) Land-use change and lake acidification: Iron age de-settlement in Northern Sweden as a pre-industrial analogue. *Philosophical Transactions of the Royal Society of London, B* 327: 373–376.
- Battarbee RW, Charles DF, Bigler C, Cumming BF, and Renberg I (2010) Diatoms as indicators of surface-water acidity. In: Smol JP and Stoermer EF (eds.) *The Diatoms: Applications for the Environmental and Earth Sciences*, 2nd edn., pp. 98–121. Cambridge: Cambridge University Press.
- Battarbee RW, Mason J, Renberg I, and Talling JF (eds.) (1990) *Paleolimnology and Lake Acidification*. London: The Royal Society.
- Birks HH and Birks HJB (2006) Multi-proxy studies in paleolimnology. *Vegetation History and Archaeobotany* 15: 235–251.
- Blais JM, Macdonald RW, Mackey D, Webster E, Harvey C, and Smol JP (2007) Biologically mediated transport of contaminants to aquatic systems. *Environmental Science & Technology* 41: 1075–1084.
- Brenner M, Whitmore TJ, Curtis JH, Hodell DA, and Schelske CL (1999) Stable isotope ($\delta^{13}\text{C}$ and $\delta^{15}\text{N}$) signatures of sedimented organic matter as indicators of historical lake trophic state. *Journal of Paleolimnology* 22: 205–221.
- Charles DF, Binford MW, Fry BD, et al. (1990) Paleoeological Investigation of Recent Lake Acidification in the Adirondack Mountains, N.Y. *Journal of Paleolimnology* 3: 195–241.
- Charles DF and Smol JP (1990) The PIRLA II project: Regional assessment of lake acidification trends. *Verhandlungen der Internationalen Vereinigung von Limnologen* 24: 474–480.
- Charles DF and Whitehead DR (1986) The PIRLA project: Paleoeological Investigation of Recent Lake Acidification. *Hydrobiologia* 143: 13–20.
- Cohen AS (2003) *Paleolimnology: The history and evolution of lake systems*. New York: Oxford University Press.
- DeSellas AM, Paterson AM, Clark BJ, Baratono NG, and Sellers TJ (2009) State of the basin report for Lake of the Woods and Rainy River Basin. *Multi-Agency Technical Report [Internet]*. Available from <http://lowwsf.com/progress-we-are-making/state-of-the-basin-report.html>.
- Dixit SS, Dixit AS, and Smol JP (1992) Assessment of changes in lake water chemistry in the Sudbury area lakes since preindustrial times. *Canadian Journal of Fisheries and Aquatic Sciences* 49: 8–16.
- Dixit SS, Dixit AS, Smol JP, and Keller W (1995) Reading the records stored in lake sediments: A method of examining the history and extent of industrial damage to lakes. In: Gunn JM (ed.) *Restoration and Recovery of an Industrial Region*, pp. 33–44. New York: Springer.
- Douglas MSV, Smol JP, Savelle JM, and Blais JM (2004) Prehistoric Inuit whalers affected Arctic freshwater ecosystems. *Proceedings of the National Academy of Sciences* 101: 1613–1617.
- Evans RA (1989) Response of limnetic insect populations of two acidic, fishless lakes to liming and brook trout (*Salvelinus fontinalis*). *Canadian Journal of Fisheries and Aquatic Sciences* 46: 342–351.

- Fariña JM, Salazar S, Wallem KP, Witman JD, and Ellis JC (2003) Nutrient exchanges between marine and terrestrial ecosystems: The case of the Galapagos sea lion *Zalophus wollebaecki*. *Journal of Animal Ecology* 72: 873–887.
- Finney BP, Gregroy-Eaves I, Sweetman J, Douglas MSV, and Smol J (2000) Impacts of climatic change and fishing on Pacific salmon abundance over the past 300 years. *Science* 290: 795–799.
- Hannan LB, James JD, Ehrhardt LM, and Weishampel JF (2007) Dune vegetation fertilization by nesting sea turtles. *Ecology* 88: 1053–1058.
- Jezorski A, Yan ND, Paterson AM, et al. (2008) The widespread threat of calcium decline in fresh waters. *Science* 322: 1374–1377.
- Krümml EM, Gregory-Eaves I, Macdonald W, et al. (2005) Concentrations and fluxes of salmon-derived polychlorinated biphenyls (PCBs) in lake sediments. *Environmental Science & Technology* 39: 7020–7026.
- Krümml EM, Macdonald RW, Kimpe LE, et al. (2003) Delivery of pollutants by spawning salmon – Fish dump toxic industrial compounds in Alaskan lakes on their return from the ocean. *Nature* 425: 255–256.
- Last WM and Smol JP (eds.) (2001a) *Tracking Environmental Change Using Lake Sediments. Basin Analysis, Coring, and Chronological Techniques*, vol. 1. Dordrecht: Kluwer.
- Last WM and Smol JP (eds.) (2001b) *Tracking Environmental Change Using Lake Sediments. Physical and Geochemical Methods*, vol. 2. Dordrecht: Kluwer.
- Little J, Hall R, Quinlan R, and Smol JP (2000) Past trophic status and hypolimnetic anoxia during eutrophication and remediation of Gravenhurst Bay, Ontario: Comparison of diatoms, chironomids, and historical records. *Canadian Journal of Fisheries and Aquatic Sciences* 57: 333–341.
- MacDonald GM, Porinchu DF, Rolland N, Kremenetsky KV, and Kaufman DS (2009) Paleolimnological evidence of the response of the Central Canadian treeline zone to radiative forcing and hemispheric patterns of temperature change over the past 2000 years. *Journal of Paleolimnology* 41: 129–141.
- Michelutti N, Blais JM, Cumming BF, et al. (2010) Do spectrally inferred determinations of chlorophyll *a* reflect trends in lake trophic status? *Journal of Paleolimnology* 43: 205–217.
- Michelutti N, Keatley BE, Brimble S, et al. (2009) Seabird-driven shifts in Arctic pond ecosystems. *Proceedings of the Royal Society B* 276: 591–596.
- Naiman RJ, Bilby RE, Schindler DE, and Helfield JM (2002) Pacific salmon, nutrients, and the dynamics of freshwater and riparian ecosystems. *Ecosystems* 5: 399–417.
- Nilssen JP (1984) An ecological jig-saw puzzle: Reconstructing aquatic biogeography and pH in an acidified region. *Reports of the Institute of Freshwater Research, Drottningholm* 61: 138–147.
- Pearl HW and Huisman J (2008) Blooms like it hot. *Science* 320: 57–58.
- Rosenqvist IT (1977) *Acid Soil – Acid Water*. Oslo: Ingeniørforlaget.
- Rosenqvist IT (1978) Alternative sources of acidification of river waters in Norway. *Science of the Total Environment* 10: 39–49.
- Rühland KM, Paterson AM, Hargan K, Jenkin A, Clark BJ, and Smol JP (2010) Reorganization of algal communities in the Lake of the Woods (Ontario, Canada) in response to turn-of-the-century damming and recent warming. *Limnology and Oceanography* 55: 2433–2451.
- Schindler DW, Mills KH, Malley DF, et al. (1985) Long-term ecosystem stress: The effects of years of experimental acidification of a small lake. *Science* 228: 1395–1401.
- Smol JP (2008) *Pollution of Lakes and Rivers: A Paleoenvironmental Perspective*. Oxford: Wiley-Blackwell.
- Smol JP, Birks HJB, and Last WM (eds.) (2001a) *Tracking Environmental Change Using Lake Sediments. Terrestrial, Algal, and Siliceous Indicators*, vol. 3. Dordrecht: Kluwer.
- Smol JP, Birks HJB, and Last WM (eds.) (2001b) *Tracking Environmental Change Using Lake Sediments. Zoological Indicators*, vol. 4. Dordrecht: Kluwer.
- Smol JP and Douglas MSV (2007) From controversy to consensus: Making the case for recent climate change in the Arctic using lake sediments. *Frontiers in Ecology and the Environment* 5: 466–474.
- Uutala AJ (1990) *Chaoborus* (Diptera: Chaoboridae) mandibles-paleolimnological indicators of the historical status of fish populations in acid-sensitive lakes. *Journal of Paleolimnology* 4: 139–151.
- Whitehead DR, Charles DF, and Goldstein RA (1990) The PIRLA project (Paleoecological Investigation of Recent Lake Acidification): An introduction to the synthesis of the project. *Journal of Paleolimnology* 3: 187–194.
- Wolfe AP, Baron JS, and Cornett RJ (2001) Anthropogenic nitrogen deposition induces rapid ecological change in alpine lakes of the Colorado Front Range (USA). *Journal of Paleolimnology* 25: 1–7.

Visible and Infrared Spectroscopical Applications

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Introduction

Infrared (IR) radiation was first discovered by Herschel in the nineteenth century, but it was not until late in the nineteenth century that the term 'infrared' was introduced (Herschel, 2000). From its first discovery to spectroscopic applications that make use of the particular characteristics of IR radiation, another century passed. In fact, it was not until the 1950s when analytically sophisticated devices made their way from research laboratories to industry. IR spectroscopic techniques have since been improved and today applications are manifold.

Visible (VIS) and IR spectroscopies utilize a wide range of the electromagnetic spectrum between 25 000 and 1 cm⁻¹ (0.38–1000 µm). Because of the wide range covered and the difference of interactions between radiation and different materials, the IR region is commonly subdivided into near-infrared (NIR; 0.78–3 µm), mid-infrared (MIR; 3–50 µm), and far-infrared (FIR; 50–1000 µm) regions. Because of their relatively simple, reliable, and nearly maintenance-free instrument setups, VIS and IR spectroscopies are widely used in industrial and scientific research applications that commonly include process monitoring and quality control (Davis and Williams, 1997).

Because of the advantages of VIS and IR spectroscopy over other methods, applications of these techniques are also becoming popular in Quaternary and environmental sciences. Pioneering work utilizing the VIS and NIR regions in paleolimnology include quantitative inferences of lake-water pH, total phosphorus (TP) concentration, and total organic carbon (TOC) concentration (Korsman et al., 1992; Nilsson et al., 1996); inferences of carbon, nitrogen, phosphorus, and diatoms in lake sediments (Malley et al., 1996, 1999, 2000); and assessment of the origin and spatial variability of sediments (Korsman et al., 1999). The MIR region has long been utilized for the identification of different mineral phases (Farmer, 1964, 1974) and to date commercial and computerized spectral libraries exist for almost any chemical compound (e.g., Sadler standard spectra). Pioneering work on both lacustrine and marine sediments utilizing the MIR region includes qualitative and quantitative analyses of different mineral phases (Bertaux et al., 1996, 1998; Chester and Elderfield, 1968; Rosén et al., 2010, 2011a; Sifeddine et al., 1994; Vogel et al., 2008).

Theoretical Background

IR spectroscopy utilizes the fact that molecules absorb specific frequencies characteristic of their three-dimensional structure, atomic weight, and bond type. Spectra in the visible-NIR

(VNIR) region absorb radiation between polar bonds of light atoms such as those in the first period (row) of the periodic table. The characteristic VNIR absorbance spectrum of a sample arises thus from a combination of fundamental C–H, O–H, N–H, and S–H vibrations. In VNIR spectroscopy (VNIRS), overtones of molecular vibrations are primarily measured (Figure 1). As the number of combinations and overtones in VNIR spectra are large, spectra from complex materials can contain a high number of overlapping bands from different bonds, making direct interpretation from these spectra difficult (Bokobza, 1998). Multivariate modeling is therefore commonly applied to extract chemical information from complex VNIR spectra (Figure 2). VNIRS is primarily used today for analyzing samples that contain organic and hydrogen bounds such as carbohydrates, cellulose and hemicellulose, proteins, and lipids (Davis and Williams, 1997; Osborne and Fearn 1988). Lake sediments constitute a mixture of these and other VNIR-active substances derived from organisms formerly living in a lake and its catchment and IR spectroscopy is therefore of interest for sediment studies (Meyers and Ishiwatari, 1993).

In contrast to VNIR, spectra in the MIR region directly monitor fundamental vibrations (stretching, rocking, bending, wagging, and twisting) that are dependent on the mass of the atoms, three-dimensional structure, and degree of symmetry of the molecule (Figure 2). Spectra in the MIR region measured from solid, liquid, or gas-phased samples contain very detailed structural and compositional information on both organic and inorganic sample compounds. A great advantage of the MIR over the VNIR region is that compound-specific absorption bands are usually better separated, facilitating visual interpretation of the spectra, as well as identification and quantification of MIR-active substances. The MIR region is divided into the group frequency region (1500–4000 cm⁻¹) and the fingerprint region (400–1500 cm⁻¹; Figure 1). Absorbance in the fingerprint region is generally due to intramolecular phenomena and thus is highly specific to each compound. The specificity in the fingerprint region makes it easier to identify specific components or use computerized data searches within reference libraries. The vibrational bands in the group frequency region are generally broader and are typically due to functional groups (e.g., CH, CH₂, CH₃, –OH, C=O, –CHO, N–H).

Quantification of different compounds from MIR spectra is possible because absorbance directly relates to concentration of the compound of interest according to the Beer–Lambert law:

$$A = \varepsilon Cl \quad [1]$$

where ε is the molecular absorptivity, C is the concentration, and l is the path length. However, peak integration, as a

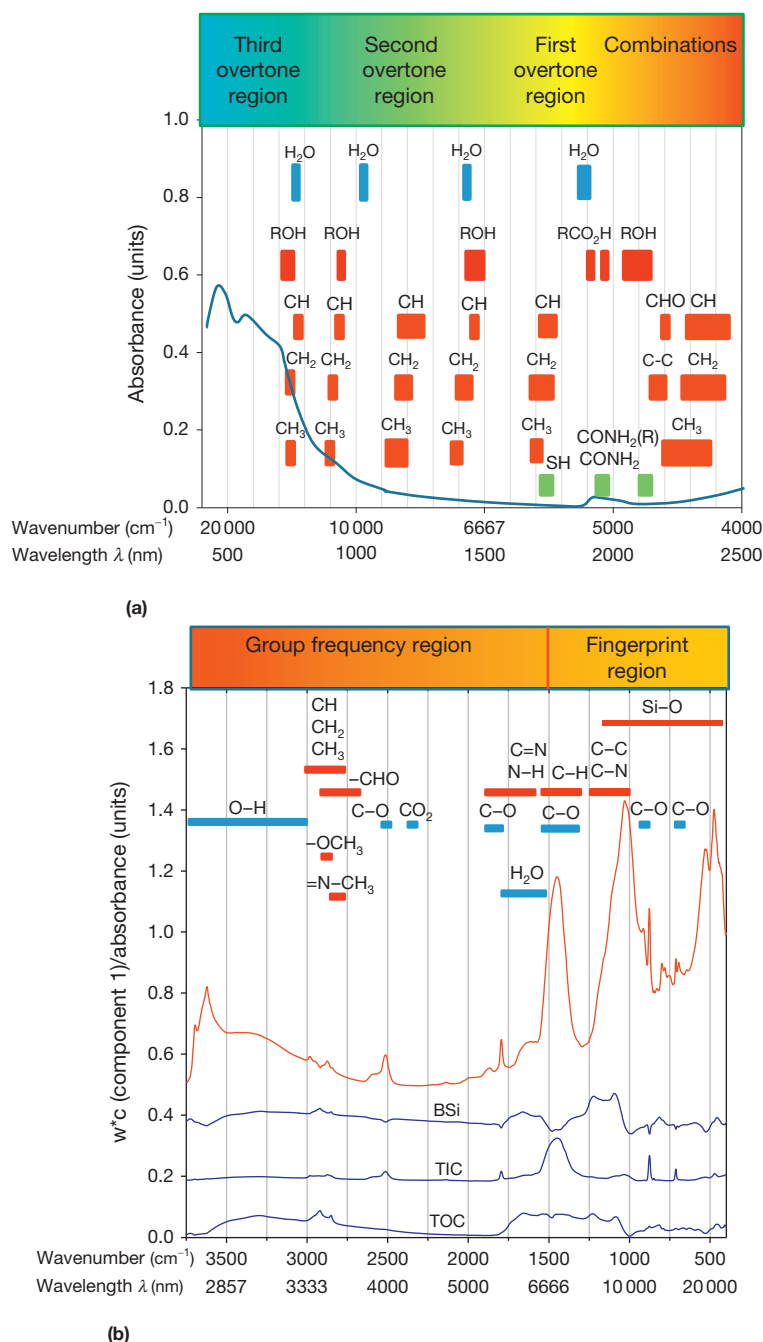


Figure 1 Example of spectra and molecular vibrations for some common organic and inorganic bond types in (a) the visible and near-infrared region and (b) the mid-infrared region. The spectra (red line) are an example of a sediment sample from Lake Ohrid (Macedonia/Albania), with total inorganic carbon (TIC) and total organic carbon (TOC) concentrations of 3% and 1.5%, respectively. Loading plots (bottom blue curves) show the relative contribution of different wave numbers to the partial least squares regression (PLSR) models for biogenic silica (BSi), TIC, and TOC using a large number of sediment samples (879–3164) of varying age (0–340,000 years) and from very diverse lake settings in Antarctica, Argentina, Canada, Macedonia/Albania, Siberia, and Sweden (Figure 3; Rosén et al., 2011a). High values indicate wave numbers positively correlated to the Y variable and low values indicate wave numbers negatively correlated to the Y variable. For better visualization, spectra have been shifted vertically from their common baseline by adding a constant value to each spectrum. Reproduced with permission from Bruker Optics.

means of quantifying a certain compound, is only possible when overlapping with other MIR-active compounds in the same spectral region can be ruled out. In the case of marine or lacustrine sediments that are mixtures of a wide

variety of MIR-active compounds, multivariate data analysis and calibration need to be applied in order to quantify certain compounds of interest (see the section 'Theoretical Background').

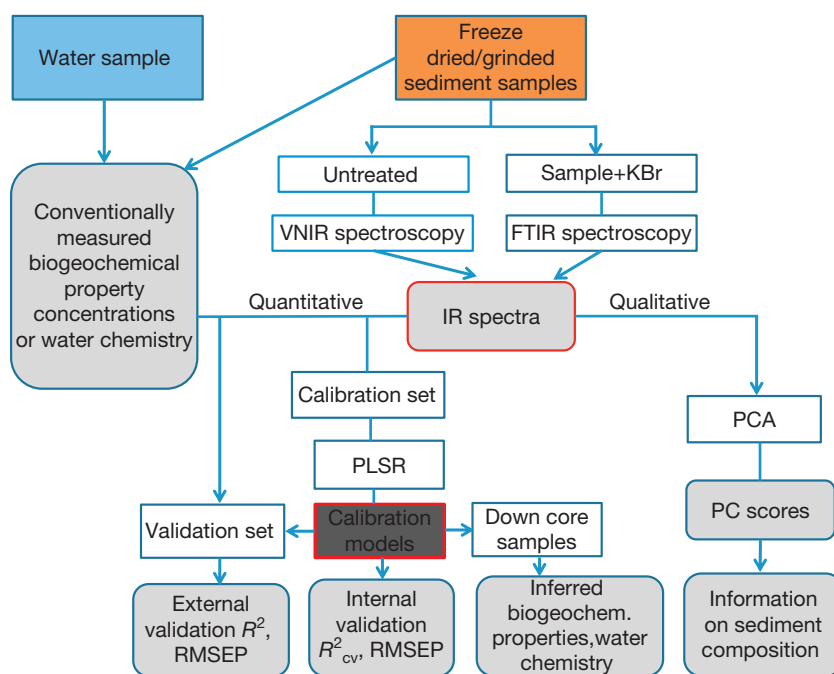


Figure 2 Flow chart documenting work flow for qualitative and quantitative analysis of sediment samples using visible and infrared spectroscopy.

Sample Preparation and Statistical Approaches in VIS and IR Spectroscopy

VIS and IR spectroscopy are quick methods that only require minor sample pretreatment. Prior to analysis, samples need to be (freeze-)dried, ground to particle size in the fine-silt fraction (2–63 μm), and homogenized in order to achieve material suitable for reproducible and representative measurements (Figure 2). Dried and ground sediment samples can be measured using VNIR devices without any further sample preparation. Small sample quantities (ca. 100 mg) are placed in sample cups and measured manually, depending on the instrument setup, by placing the sample into the IR beam or automatically using multi-samplers. In contrast to VNIRS, measurements in the MIR region, commonly by means of Fourier transform infrared spectroscopy (FTIRS), require dilution of the sample material by IR transparent substances such as KBr, at an appropriate ratio. This is because FTMIR measurements on solids can produce very high absorbance and thus noisy data. For sediment samples, a commonly applied ratio of sediment to KBr is 0.02, that is, 10 mg sediment mixed with 500 mg KBr. Similarly to VNIRS, powdered and mixed samples are placed in sample cups and measured either manually or automatically using multi-samplers. If a multi-sampler is used, up to 500 samples can be analyzed in 1 day.

If samples are measured on different IR spectrometers, but results need to be compared afterward, special attention has to be paid to ensure that the analytical environment and measurement conditions (constant laboratory temperature and humidity, vacuum, etc.) are as similar as possible. When these criteria are matched, calibrations developed for one instrument can be transferred to another and vice versa. This has been tested for Bruker FTIR spectrometers operating under

vacuum (Bruker IFS 66v/S) and nitrogen cooled (Vertex 70) conditions. Measurements between both instruments are comparable, despite different detector and instrument setups.

As combinations of absorption bands seen in the VIS to MIR regions are typically broad, leading to complex and overlapping spectra, it can be difficult to assign distinct features to specific chemical components. One way to extract qualitative and quantitative information on sediment components is through multivariate data analysis, such as principal component analysis (PCA) and partial least squares regression (PLSR) (Martens and Næs, 1989). Prior to multivariate analysis, the spectra need to be baseline-corrected. A commonly used technique is multiple scatter correction (Geladi and Dabakk, 1995; Geladi et al., 1985). The aim of this method is to remove variation in spectra caused by between-sample scatter variations in order to linearize the spectra and eliminate variation caused by noise. The remaining variation is only supposed to contain chemical information about the samples. Spectra in the MIR region can also be baseline-corrected by setting to zero certain wave numbers that do not contain chemical information. PCA on the baseline-corrected VIS to MIR spectra can yield a first impression on major changes in sediment composition, outliers can be identified, sediment horizons where large changes occur can be assessed, and the spectral region and sediment properties that cause most of the variation can be identified.

Based on a first evaluation using PCA, other, more time-consuming and expensive techniques can be applied to selected sections for more detailed multiproxy analyses. After calibration using multivariate regression techniques such as PLSR, spectral information can be used to quantitatively assess compositional changes of the sediment (e.g., Rosén et al., 2011a). Reliable calibrations can be established using an

internal calibration. An internal calibration is, for instance, a calibration for a stratigraphically connected sediment succession (e.g., Vogel et al., 2008) or a regionally confined surface sediment sample set (e.g., Rosén et al., 2010). In such an attempt, preferably at least 100 (VIS–MIR) spectra are regressed against conventionally analyzed sediment properties such as organic carbon, inorganic carbon, and biogenic silica (BSi) concentrations. The calibration is statistically evaluated by using cross-validated coefficient of correlation (R^2_{cv}) and root mean squared error of prediction based on cross-validation ($RMSEP_{cv}$). The most important spectral regions used by the model can be assessed by using the loading values from the PLSR (Figure 1). For optimal performance of the model, there should be a good correlation between the important spectral region used by the calibration model and the actual spectral fingerprint of the compound of interest. When a robust calibration model is established, it can be applied in a rapid and cost-efficient way to infer concentrations of important sediment components solely from spectral information, by using only small sediment quantities. This approach can be of great advantage when very long sediment sequences, several hundred meters in length, need to be analyzed in high resolution (Rosén et al., 2010; Vogel et al., 2008).

VNIRS Applications for the Qualitative and Quantitative Inference of Sediment Properties

To date, a number of applications exist that utilize the advantages of measurements in the VNIR region to extract environmental information and information on sediment composition. VNIR applications for the quantification of sediment constituents such as TP, total carbon (TC), and total nitrogen (TN) exist and show good statistical performance compared to data measured using conventional techniques (r^2 's of 0.99 for TP, and >0.93 for TC and TN; Malley et al., 1999, 2000). In contrast, inferences of carbonates and the amount of diatoms/BSi using VNIRS lack similarly good correlations, with r^2 values of 0.8 for the inference of carbonate content (Malley et al., 1999) and r^2 values ranging from 0.54 to 0.42 for the inference of diatom amounts/BSi concentrations (Hahn et al., 2011; Malley et al., 1999). This is primarily because carbonate and BSi molecules are not excited by radiation in the VNIR region and thus direct relations between absorbance and concentrations of these components cannot be established.

Besides quantitative inference of particular sediment components, NIR spectral information has been used to describe changes in sediment composition in a more qualitative way. Korsman et al. (1999) used NIR spectra to show the influence of catchment properties on surface sediment quality in a small lake in northern Sweden. The sediment properties could be explained only to a minor extent by water depth and organic matter concentration, and the NIR spectra revealed the importance of different types of land use and inlets for the sediment properties. A study by Rosén et al. (2000) documented the impact of catchment vegetation along an altitudinal gradient on compositional changes in surface sediment composition. In this study, a surface sediment set consisting of 76 samples was collected along an altitudinal (hence climatic) and vegetation gradient. The results show that surficial

sediments from lakes situated at different locations along the gradient could be distinguished and that a statistically significant relationship is observed between measured altitude of the lakes and altitude inferred from VNIR spectra. The altitude transfer function was converted to July temperature and successfully applied to a Holocene sediment sequence from an alpine lake in northern Sweden. The inferred temperature from the VNIRS altitude model showed similar trends to inferred July temperatures derived from chironomids, diatoms, and pollen. However, similar to transfer functions based on biological indicators, the correlation between the VNIR spectral information and climate is not direct, but probably an effect of qualitative changes in sediment composition, depending on climate-controlled catchment dynamics.

The initial ideas by Nilsson et al. (1996), Korsman et al. (1999), and Rosén et al. (2000) and the fact that a substantial part of organic carbon in small headwater lakes may originate from catchment sources led to the application of VNIRS that aims at inferring TOC concentrations of lake water using sediment archives (Rosén, 2005). It is unlikely that there is a direct relationship between NIR spectra of sediments and TOC concentration of ambient lake water. Instead, the correlation between VNIR spectra and lake-water TOC concentration is more likely an effect of correlated source-related compositional changes of organic matter deposited at the bottom of the lake. As a first attempt, Rosén (2005) could show the potential of VNIRS to infer lake-water TOC in different lake settings. The study demonstrates the impact of Holocene changes in catchment dynamics on the inferred lake-water TOC values in boreal, tree-line, and alpine lakes and thus emphasizes the role of catchment properties for lake-water TOC concentration (Battin et al., 2009). In close correspondence to the study by Korsman et al. (1999), it was demonstrated that organic carbon concentrations in lacustrine sediment only explained a minor part of the VNIRS-inferred lake-water TOC. This is because VNIRS spectra contain more detailed information on organic matter (Figure 1) enabling, for instance, characterization and identification of humic substances in lacustrine sediment (e.g., Belzile et al., 1997). The application of VNIRS lake-water TOC has been further tested and applied to lakes affected by human impact (Cunningham et al., 2011; Rosén et al., 2011b) and has been used to assess the effect of mire development, permafrost dynamics, and tree-line changes in Sweden, Finland, Russia, and Canada (Jones et al., 2011; Kokfelt et al., 2009; Reuss et al., 2010; Rosén and Hammarlund, 2007; Rosén et al., 2009; Rouillard et al., 2011; Rydberg et al., 2010).

Pigments and their degradation products in lacustrine sediment can be applied to infer past primary productivity (Leavitt et al., 1994). Analysis of pigments commonly requires wet chemical sample preparation and high-performance liquid chromatography (HPLC) and is thus time-consuming. A simpler means of analysis might be through VNIRS. Chlorophyll *a*, its derivatives, and degradation products exhibit a reflectance minimum close to 675 nm in the VIS region. Das et al. (2005) and Wolfe et al. (2006) found a strong correlation between the area of the reflectance trough in the 650–700 nm region and summed concentrations of chlorophyll *a*, its derivatives, and degradational products. Further applications of VNIRS to assess pigments yielded promising results, which are,

of course, not as detailed and precise as pigment inferences using HPLC, but on the other hand open up the possibility of obtaining a rough idea of historical changes in lake trophic status in a nondestructive and rapid way (Keatley et al., 2011; Michelutti et al., 2010).

VNIRS has also been applied to ornithogenic (bird-deposited) sediments to reconstruct the extent of penguin populations (Liu et al., 2011). Synthetic mixtures of penguin guano and weathered soils demonstrated that reflectance, in particular in the NIR regions, increased with guano concentration. PCA and linear mixing modeling were applied to spectral data in order to quantify the amount of penguin guano in downcore lacustrine sediment samples. Based on this approach and by assuming that the amount of penguin dropping input to lake sediments reflects the size of the penguin population, the size of penguin populations on the Ardley Islands, Antarctica, during the past 2700 year was reconstructed (Liu et al., 2011).

VNIR spectroscopy is commonly used in the industry for *in situ* process monitoring. A great advantage is that the measurements are done directly on the moving sample in the process stream without any sample pretreatment. *In situ* scanning using visible light (380–730 nm) was also tested on laminated, siliciclastic sediments from proglacial Lake Silvaplana in the eastern Swiss Alps (Trachsel et al., 2010). Scanning was performed at 2 mm resolution to assess a correlation between spectral information and climate variables. By using multiple linear regression, a calibration model was established that explains 84% of the variance of summer (JJAS) temperature during the calibration period 1864–1950. When applied downcore to quantitatively infer summer temperature back to AD 1177, the temperature reconstruction was in good agreement with independent historical and tree-ring datasets that extend back to AD 1500 and AD 1177, respectively. This approach indicates that valuable information on sediment composition can be derived from *in situ* scanning. This implies that VNIRS can be used as a nondestructive technique to rapidly acquire quantitative paleoenvironmental information in high resolution without tedious sample pretreatment.

MIRS Applications for the Qualitative and Quantitative Inference of Sediment Constituents

Due to the very confined character of absorption bands (Figure 1), the MIR region offers the opportunity to provide detailed qualitative and quantitative information on a variety of organic and inorganic sedimentary constituents. Early MIR spectroscopy (MIRS) applications were developed for the detection and quantification of a variety of amorphous and crystalline mineral phases, such as opal, quartz, kaolinite, and carbonates in deep-sea sediment and consolidated sedimentary rocks from different depositional sources (Chester and Elderfield, 1968, 1973; Chester and Green, 1968; Ji et al., 2009; Mecozzi et al., 2001). Quantification of mineral phases in these studies has been based on single peak absorbance/transmission calibration by regressing absorbance in phase-specific regions onto concentration by using synthetic sediment mixtures. This approach was found to perform well only if other phases absorbing in the same spectral regions are absent (Chester

and Elderfield, 1968; Chester and Green, 1968; Mecozzi et al., 2001).

An early attempt to use the MIR technique in paleolimnology was to quantitatively assess quartz, kaolinite, gibbsite, and amorphous silica concentrations in sediment (Bertaux et al., 1996, 1998). The approach was based on end-member modeling of MIR spectral information from synthetic multicomponent sediment mixtures. The technique was subsequently used in a number of environmental reconstructions using lake sediment (Bertaux et al., 1996; Wirrmann et al., 2001, 2006). Another attempt to extract quantitative information on BSi, TOC, and total inorganic carbon (TIC) from MIR spectra of lake sediment was introduced by Vogel et al. (2008). In this approach, data on property concentration, measured by conventional analytical techniques, was regressed onto MIR spectral information by applying PLSR. The results indicate that calibrations for stratigraphically connected sediment successions can be developed and applied downcore (Vogel et al., 2008, 2010). In order to test the general applicability of these calibrations independent of a stratigraphic connection, larger datasets from lakes worldwide were included into the PLSR calibrations. The results showed that BSi, TOC, and TIC can be assessed independently by using universally applicable calibrations based on samples from around the world (Figure 3; Rosén et al., 2010, 2011a). MIRS also enables very detailed studies of important sediment components such as BSi (Fröhlich, 1989; Gendron-Badou et al., 2003; Schmidt et al., 2001) and allows for rapid purity control of BSi samples prior to stable isotope analysis (Swann and Patwardhan, 2011).

MIRS has furthermore been proven useful for the characterization of specific organic substances in lake sediments. In particular, studies aiming at deciphering the composition and abundance of humic and fulvic acids in lacustrine (Belzile et al., 1997; Calace et al., 1999) and marine sediment (Braguglia et al., 1995; Calace et al., 2006; Mecozzi and Pietrantonio, 2006) produced promising results. Moreover, diagenetic/degradational changes of humic acids that, for example, show loss of carboxylic and nitrogen-containing groups may be identified from MIRS spectral information (Calace et al., 2006). Due to the wealth of information on organic and inorganic sedimentary components in the MIR region, it is more than likely that other important constituents found in lacustrine sediment can also be qualitatively and quantitatively assessed using MIRS. Due to small sample size and simple preparation and measurement setups, these applications could significantly contribute to studies aiming at deciphering past climatic and environmental changes over very long (Ma) time scales and/or in very high temporal resolution, for example, annually laminated sediments. Apart from applications aimed at identification and quantification of specific components in lacustrine sediment, MIRS has also been applied to track changes in environmental variables. Based on similar ideas to VNIRS (Rosén, 2005; Rosén et al., 2000), MIRS has been applied to assess past tree-line and lake-water TOC changes (Rosén and Persson, 2006).

In recent years, the development of FTIRS microscopy devices has made it possible to assess component-specific spectra of components identified under the microscope. One such example is demonstrated in the study by Rosén et al. (2010) where spectra from single diatom valves were assessed. Thanks to the continuous technical development of FTIR spectroscopy,

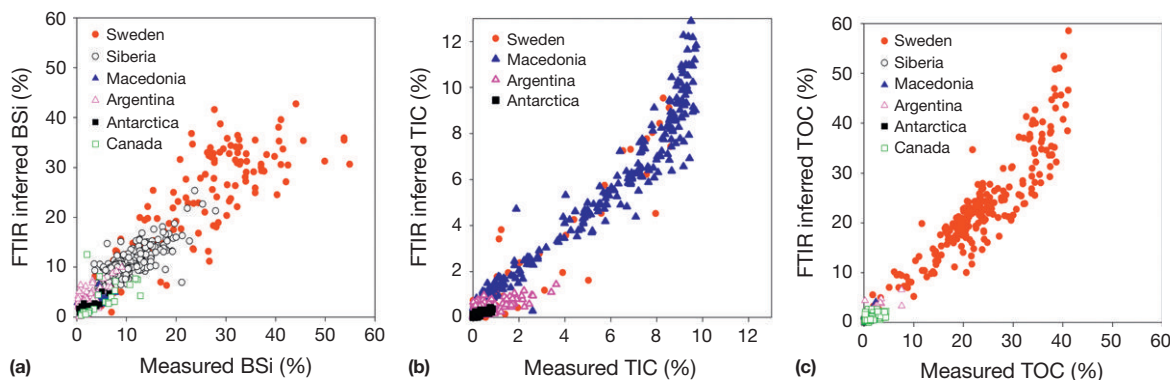


Figure 3 Scatter plots showing the relationships between conventionally measured and Fourier transform infrared spectroscopy (FTIRS)-inferred biogenic silica (BSi), total inorganic carbon (TIC), and total organic carbon (TOC) using sediment from ca. 100 Swedish lakes, Lake El'gygytyn (Siberia), Lake Ohrid (Albania/Macedonia), Laguna Potrok Aike (Argentina), Lake Hoare (Antarctica), and Lake Pingualuit (Canada). The inferred values are based on partial least square regression (PLSR) models using the region of 400–3750 cm^{-1} . Reproduced from Rosén P, Bindler R, Korsman T, Mighall T, and Bishop K (2011) The complementary power of pH and lake-water organic carbon reconstructions for discerning the influences on surface waters across decadal to millennial time scales. *Biogeosciences* 8: 2717–2727.

potential applications of the technique also continue to increase. Considering the variety of applications presented in this article, we encourage more researchers to further develop both the VNIRS and FTIRS techniques for applications in marine or lake sediment science.

See also: [Paleolimnology: Overview of Paleolimnology.](#)

References

- Battin TJ, Luyssaert S, Kaplan LA, Aufdenkampe AK, Richter A, and Tranvik LJ (2009) The boundless carbon cycle. *Nature Geoscience* 2: 598–600.
- Belzile N, Joly HA, and Li H (1997) Characterization of humic substances extracted from Canadian lake sediments. *Canadian Journal of Chemistry* 75: 14–27.
- Bertaux J, Fröhlich F, and Ildefonse P (1998) Multicomponent analysis of FTIR spectra: Quantification of amorphous and crystallized mineral phases in synthetic and natural sediments. *Journal of Sedimentary Research* 68: 440–447.
- Bertaux J, Ledru M-P, Soubié F, and Sondag F (1996) The use of quantitative mineralogy linked to palynological studies in paleoenvironmental reconstruction: The case study of the 'Lagoa Campestre' lake, Salitre, Minas Gerais, Brazil. *Comptes Rendus de l'Académie des Sciences Paris* 323(Series II): 65–71.
- Bokobza L (1998) Near infrared spectroscopy. *Journal of Near Infrared Spectroscopy* 6: 3–17.
- Braguglia CM, Campanella L, Petronio BM, and Scerbo R (1995) Sedimentary humic acids in the continental margin of the Ross Sea (Antarctica). *International Journal of Environment Analytical Chemistry* 60: 61–70.
- Calace N, Capolei M, Lucchese M, and Petronio BM (1999) The structural composition of humic compounds as indicator of organic carbon sources. *Talanta* 49: 277–284.
- Calace N, Cardellicchio N, Petronio BM, Pietrantonio M, and Pietroletti M (2006) Sedimentary humic substances in the northern Adriatic sea (Mediterranean sea). *Marine Environmental Research* 61: 40–58.
- Chester R and Elderfield H (1968) The infra-red determination of opal in siliceous deep-sea sediments. *Geochimica et Cosmochimica Acta* 32: 1128–1140.
- Chester R and Elderfield H (1973) An infrared study of clay minerals. 2. The identification of kaolinite-group clays in deep-sea sediments. *Chemical Geology* 12: 281–288.
- Chester R and Green RN (1968) The infra-red determination of quartz in sediments and sedimentary rocks. *Chemical Geology* 3: 199–202.
- Cunningham L, Bishop K, Lind Mettävainio E, and Rosén P (2011) Trends in total organic carbon (TOC) content of lake water: Long-term reconstructions from southern Sweden. *Journal of Paleolimnology* 45: 507–518.
- Das B, Vinebrooke RD, Sanchez-Azofeifa A, Rivard B, and Wolfe AP (2005) Inferring sedimentary chlorophyll concentrations with reflectance spectroscopy: A novel approach to reconstructing historical changes in the trophic status of mountain lakes. *Canadian Journal of Fisheries and Aquatic Sciences* 62: 1067–1078.
- Davis ANC and Williams P (1997) Near infrared spectroscopy: The future waves. *The Proceedings of the 7th International Conference on Near Infrared Spectroscopy*, Chichester: NIR Publications.
- Farmer VC (1964) Infrared spectroscopy of silicates and related compounds. In: Taylor HFW (ed.) *The Chemistry of Cements*, vol. 2. London: Academic press.
- Farmer VC (ed.) (1974) *The infrared spectra of minerals*. Mineralogical Society Monograph 4: Dorking, Surrey: Adlard & Son.
- Fröhlich F (1989) Deep-sea biogenic silica: New structural and analytical data from infrared analysis – Geological implications. *Terra Nova* 1: 267–273.
- Geladi P and Dabakk E (1995) An overview of chemometrics applications in NIR spectroscopy. *Journal of Near Infrared Spectroscopy* 3: 119–132.
- Geladi P, MacDougall D, and Martens H (1985) Linearization and scatter-correction for near-infrared reflectance spectra of meat. *Applied Spectroscopy* 39: 491–500.
- Gendron-Badou A, Coradin T, Maquet J, Fröhlich F, and Livage J (2003) Spectroscopic characterization of biogenic silica. *Journal of Non-Crystalline Solids* 316: 331–337.
- Hahn A, Rosén P, Kliem P, Ohlendorf C, and Zolitschka B (2011) Comparative study of infrared techniques for fast biogeochemical sediment analyses. *Geochemistry, Geophysics, Geosystems* 12(10) <http://dx.doi.org/10.1029/2011GC003686>.
- Herschel W (2000) Herschel's paper (reprint). *Journal of Near Infrared Spectroscopy* 8: 75–86.
- Ji J, Ge Y, Balsam W, Damuth JE, and Chen J (2009) Rapid identification of dolomite using a Fourier Transform Infrared Spectrophotometer (FTIR): A fast method for identifying Heinrich events in IODP Site U1308. *Marine Geology* 258: 60–68.
- Jones VJ, Solovieva N, Self AE, et al. (2011) The influence of Holocene tree-line advance and retreat on an arctic lake ecosystem: A multi-proxy study from Kharine Lake, North Eastern European Russia. *Journal of Paleolimnology* 46: 123–137.
- Keatley BE, Blais JM, Douglas MSV, et al. (2011) Historical seabird population dynamics and their effects on Arctic pond ecosystems: A multi-proxy paleolimnological study from Cape Vera, Devon Island, Arctic Canada. *Fundamental and Applied Limnology* 179: 51–66.
- Kokfelt U, Rosén P, Schoning K, et al. (2009) Ecosystem responses to increased precipitation and permafrost decay in subarctic Sweden inferred from peat and lake sediments. *Global Change Biology* 15: 1652–1663.
- Korsman T, Nilsson MB, Landgren K, and Renberg I (1999) Spatial variability in surface sediment composition characterised by near-infrared (NIR) reflectance spectroscopy. *Journal of Paleolimnology* 21: 61–71.
- Korsman T, Nilsson MB, Öhman J, and Renberg I (1992) Near-infrared reflectance spectroscopy of sediments: A potential method to infer the past pH of lakes. *Environmental Science and Technology* 26: 2122–2126.
- Leavitt PR, Schindler DE, Paul AJ, Hardie AK, and Schindler DW (1994) Fossil pigment records of phytoplankton in trout-stocked alpine lakes. *Canadian Journal of Fisheries and Aquatic Sciences* 51: 2411–2423.

- Liu X, Sun J, Sun L, Liu W, and Wang Y (2011) Reflectance spectroscopy: A new approach for reconstructing penguin population size from Antarctic ornithogenic sediments. *Journal of Paleolimnology* 45: 213–222.
- Malley DF, Hauser BW, Williams PC, and Hall J (1996) Prediction of organic carbon, nitrogen and phosphorous in freshwater sediments using near infrared reflectance spectroscopy. In: Davies AMC and Williams P (eds.) *Near Infrared Spectroscopy: The Future Waves*, pp. 691–699. Chichester: NIR Publications.
- Malley DF, Lockhart L, Wilkinson P, and Hauser B (2000) Determination of carbon, carbonate, nitrogen, and phosphorus in freshwater sediments by near-infrared reflectance spectroscopy: Rapid analysis and a check on conventional analytical methods. *Journal of Paleolimnology* 24: 415–425.
- Malley DF, Röncke H, Findlay DL, and Zippel B (1999) Feasibility of using near-infrared reflectance spectroscopy for the analysis of C, N, P, and diatoms in lake sediment. *Journal of Paleolimnology* 21: 295–306.
- Martens H and Næs T (1989) *Multivariate Calibration*. New York: John Wiley & Sons.
- Mecozi M and Pietrantonio E (2006) Carbohydrates proteins and lipids in fulvic and humic acids of sediments and its relationships with mucilaginous aggregates in the Italian seas. *Marine Chemistry* 101: 27–39.
- Mecozi M, Pietrantonio E, Amici M, and Romanelli G (2001) Determination of carbonate in marine solid samples by FTIR-ATR spectroscopy. *Analyst* 126: 144–146.
- Meyers PA and Ishiwatari R (1993) Lacustrine organic geochemistry – An overview of indicators of organic-matter sources and diagenesis in lake-sediments. *Organic Geochemistry* 20: 867–900.
- Michelutti N, Jules MB, Cumming BF, et al. (2010) Do spectrally inferred determinations of chlorophyll a reflect trends in lake trophic status? *Journal of Paleolimnology* 43: 205–217.
- Nilsson MB, Dabakk E, Korsman T, and Renberg I (1996) Quantifying relationships between near-infrared reflectance spectra of lake sediment and water chemistry. *Environmental Science and Technology* 30: 2586–2590.
- Osborne BG and Fearn T (1988) *Near Infrared Spectroscopy in Food Analysis*. New York: Wiley.
- Reuss N, Hammarlund D, Rundgren M, and Rosén P (2010) Lake ecosystem responses to Holocene climate change at the subarctic. *Ecosystems* 13: 393–409.
- Rosén P (2005) Total organic carbon (TOC) of lake water during the Holocene inferred from lake sediments and near-infrared spectroscopy (NIRS) in eight lakes from northern Sweden. *Biogeochemistry* 76: 503–516.
- Rosén P, Bindler R, Korsman T, Mighall T, and Bishop K (2011a) The complementary power of pH and lake-water organic carbon reconstructions for discerning the influences on surface waters across decadal to millennial time scales. *Biogeosciences* 8: 2717–2727.
- Rosén P, Cunningham L, Vonk J, and Karlsson J (2009) Effects of climate on organic carbon and ratio of planktonic to benthic primary producers in a subarctic lake during the past 45 years. *Limnology and Oceanography* 54(5): 1723–1732.
- Rosén P, Dabakk E, Renberg I, Nilsson M, and Hall RI (2000) Near-infrared spectrometry (NIRS): A new tool for inferring past climatic changes from lake sediments. *The Holocene* 10: 161–166.
- Rosén P and Hammarlund D (2007) Effects of climate, fire and vegetation development on Holocene changes in total organic carbon concentration in three boreal forest lakes in northern Sweden. *Biogeosciences* 4: 975–984.
- Rosén P and Persson P (2006) Fourier-transform infrared spectroscopy (FTIRS), a new method to infer past changes in tree-line position and TOC using lake sediment. *Journal of Paleolimnology* 35: 913–923.
- Rosén P, Vogel H, Cunningham L, et al. (2011b) A universally applicable model for the quantitative determination of lake sediment composition using Fourier transform infrared spectroscopy. *Environmental Science and Technology* <http://dx.doi.org/10.1021/es200203z>.
- Rosén P, Vogel H, Cunningham L, Reuss N, Conley D, and Persson P (2010) Fourier transform infrared spectroscopy, a new method for rapid determination of total organic and inorganic carbon and opal concentration in lake sediments. *Journal of Paleolimnology* 43: 247–259.
- Rouillard A, Rosén P, Douglas MSV, Pienitz R, and Smol JP (2011) Development and application of a VNIRS model for inferring lake water dissolved organic carbon (DOC) using a Canadian Arctic lake sediment calibration set. *Journal of Paleolimnology* 46: 187–202.
- Rydborg J, Klaminder J, Rosén P, and Bindler R (2010) Climate driven release of carbon and mercury from permafrost mires increases mercury loading to sub-arctic lakes. *Science of the Total Environment* 408: 4778–4783.
- Schmidt M, Botz R, Rickert D, Bohrmann G, Hall SR, and Mann S (2001) Oxygen isotopes of marine diatoms and relations to opal-A maturation. *Geochimica et Cosmochimica Acta* 65(2): 201–211.
- Sifeddine A, Fröhlich F, Fournier M, et al. (1994) La sédimentation lacustre indicateur de changements de paléoenvironnements au cours des 30 000 dernières années (Carajas, Amazonie, Brésil). *Comptes Rendus de l'Académie des Sciences Paris* 318(2): 1645–1652.
- Swann GEA and Patwardhan SV (2011) Application of Fourier Transform Infrared Spectroscopy (FTIR) for assessing biogenic silica sample purity in geochemical analyses and paleoenvironmental research. *Climate of the Past* 7: 65–74.
- Trachsel M, Grosjean M, Schnyder D, Kamenik C, and Rein B (2010) Scanning reflectance spectroscopy (380–730 nm): A novel method for quantitative high-resolution climate reconstructions from minerogenic lake sediments. *Journal of Paleolimnology* 44: 979–994.
- Vogel H, Rosén P, Wagner B, Melles M, and Persson P (2008) Fourier Transform Infrared Spectroscopy, a new cost-effective tool for quantitative analysis of biogeochemical properties in long sediment records. *Journal of Paleolimnology* 40: 689–702.
- Vogel H, Wagner B, Zanchetta G, Sulpizio R, and Rosén P (2010) A paleoclimate record with tephrochronological age control for the last glacial-interglacial cycle from Lake Ohrid, Albania and Macedonia. *Journal of Paleolimnology* 44: 295–310.
- Wirrmann D, Bertaux J, and Kossoni A (2001) late Holocene paleoclimatic changes in western central Africa inferred from mineral abundances in dated sediments from Lake Ossa (southwestern Cameroon). *Quaternary Research* 56: 275–287.
- Wirrmann D, Sémah A-M, and Chacornac-Rault M (2006) late Holocene paleoenvironment in northern Caledonia, southwestern Pacific, from a multiproxy analysis of lake sediment. *Quaternary Research* 66: 213–232.
- Wolfe AP, Vinebrooke RD, Michelutti N, Rivard B, and Das B (2006) Experimental calibration of lake-sediment spectral reflectance to chlorophyll a concentrations: Methodology and paleolimnological validation. *Journal of Paleolimnology* 36: 91–100.

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PALEOSOLS AND WIND-BLOWN SEDIMENTS

Contents

Overview

Nature of Paleosols

Mineral Magnetic Analysis

Soil Micromorphology

Weathering Profiles

Soil Morphology in Quaternary Studies

Biogeochemical Role of Dust in Quaternary Climate Cycles

Overview

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Introduction

Eolian (windblown) sediments, soils, and paleosols constitute some of the most important components of the Quaternary stratigraphic record. Low-latitude arid or semiarid regions, under the influence of subtropical high-pressure cells, have significant areas of eolian sand (Goudie, 2002; Lancaster, 1995; McKee, 1983). ‘Loess’ is eolian silt that mantles preexisting topography and is thick enough to form a recognizable unit when observed in the field. As much as 10% of Earth’s land surface may be covered by loess, primarily in mid-latitude regions (Pye, 1987). A distinct advantage of eolian sand and loess over other Quaternary sediments is that they can be dated directly, using luminescence methods, requiring only the sediment (Aitken, 1998).

Although eolian sand and loess occupy large areas of Earth’s land surface, eolian sediments are not limited to these areas only. It is now known that many parts of the deep-ocean floor are mantled with long-range-transported (LRT) eolian particles, mostly <10 µm in diameter. In addition, polar ice caps, Greenland and Antarctica, also contain important records of LRT dust. Records of periodic dust deposition onto polar ice can be retrieved from cores that now span much of the middle and late Quaternary.

Soils are naturally occurring bodies that mantle most of the land surface of Earth. Soil geography is a function of the combined effects of climate and vegetation, as well as parent material composition, parent material age, and topography (Birkeland, 1999; Jenny, 1941; Schaetzl and Anderson, 2005). Because of their close links to climate, vegetation, and age of parent material, soils are powerful tools in Quaternary stratigraphy, geomorphology and landscape evolution, archaeology, and paleoclimatology. Soils of the past, called paleosols, are particularly important in stratigraphy, interpreting past conditions of climate or vegetation, and the duration of land surface stability.

Eolian Sand in the Quaternary

Sand consists of mineral particles with diameters >50 µm but <2000 µm. Eolian sand can take the form of dunes or sheets. Dune fields are eolian sand bodies that not only contain many dunes but may also occupy significant areas, on the order of 1000–10 000 km². Some areas of eolian sand in desert basins are so large (on the order of 50 000–100 000 km² or more) that they are referred to as ‘sand seas’ or ‘ergs.’ The largest sand seas in the world are found in subtropical latitudes of Africa, Arabia,

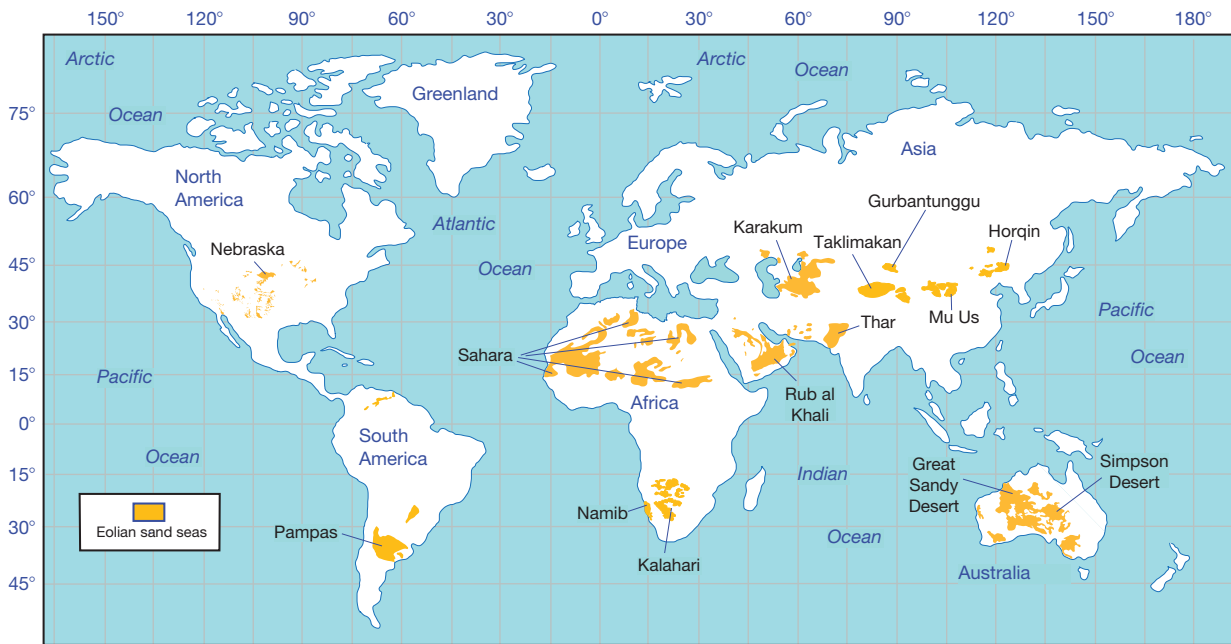


Figure 1 Map showing the global distribution of eolian sand, whether active or stabilized. Sources: North America: [Muhs and Zárate \(2001\)](#), [Muhs and Holliday \(1995\)](#), and [Thorpe and Smith \(1952\)](#); South America: [Muhs and Zárate \(2001\)](#) and [Zárate \(2003\)](#) and sources therein; Africa: [Wilson \(1973\)](#), [Lancaster \(1989\)](#), and [Thomas and Shaw \(1991\)](#); Central Asia and India: [Wilson \(1973\)](#); Sun and Muhs (this section); Australia: [Bowler et al. \(2001\)](#).



Figure 2 Photographic gallery of active dunes in (a) the Algodones Dunes, California, USA; (b) the northern Sinai Desert, Egypt; (c) Kelso Dunes, California, USA; and (d) Great Sand Dunes National Park, Colorado, USA. All photographs by D.R. Muhs, USGS.

and Australia ([Figure 1](#)). The main reason for this is that these regions are situated under subtropical high-pressure cells, where subsiding air suppresses precipitation and deserts develop. The large sand seas in these areas are found mostly in topographic basins, where sand accumulates from surrounding uplands. Lancaster reviews the origin, ages, and paleoclimatic

significance of low-latitude sand seas. Large sand seas, both active and stable, also occur in the mid-latitudes of Asia, South America, and North America. Although mid-latitude regions are usually not as dry as the areas under subtropical high-pressure cells, the winds are usually stronger and large sand bodies can accumulate if there is a sufficient supply of

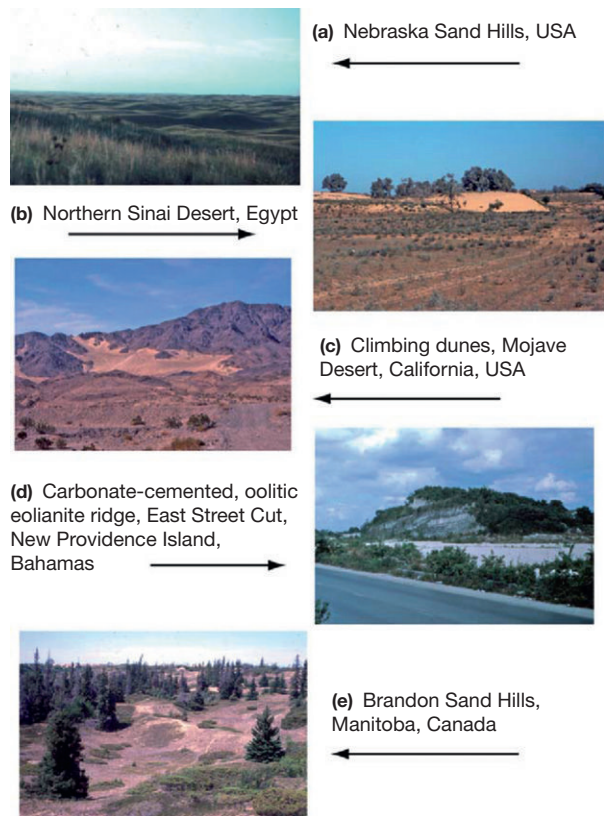


Figure 3 Photographic gallery of inactive or mostly inactive dunes in the (a) Nebraska Sand Hills, USA; (b) northern Sinai Desert, Egypt; (c) Mojave Desert, California, USA; (d) New Providence Island, Bahamas; and (e) Brandon Sand Hills, Manitoba, Canada. All photographs by D.R. Muhs, USGS.

sand. Sun and Muhs review the origins of mid-latitude dune fields. At still higher latitudes, dune fields are very small, but they are widespread over North America and Eurasia. Most high-latitude dune fields are related, either directly or indirectly, to the abundance of sand supplies that are ultimately linked to glacial sources. Wolfe reviews the origins of high-latitude dune fields.

Dune fields are significant for understanding past climates of the Quaternary for at least two very important reasons. One is that degree of dune activity is, to a great extent, a function of the amount of vegetation cover, which in turn is often conditioned by overall moisture balance (Lancaster, 1995). In many low-latitude dune fields and some mid-latitude dune fields, eolian sand is currently active (Figure 2). Active dunes are those that lack a vegetation cover and whose particles are transported on a seasonal to daily basis. In contrast, in a climate with a positive moisture balance, a stabilizing vegetation cover can be sustained. Inactive or stable dunes usually have a vegetation cover and experience little or no movement (Figure 3). Thus, in areas where inactive dunes occur, eolian sediments often reflect a past period of greater-than-present aridity or vegetation disturbance (e.g., from fire, overgrazing, etc.). Furthermore, stratigraphic sections that contain alternations of eolian sand and buried soils may record shifts between relatively arid and relatively humid climates.

A second reason for dunes being important for Quaternary paleoclimates is that they are sensitive indicators of wind

direction. The geomorphology of dunes is variable (Figure 4). Some dunes have a crescentic form (barchan, barchanoid-ridge, and parabolic dunes), others have a linear form (linear, longitudinal, or sief dunes), and still others have a multiple-ridged form (star dunes). The wind regimes associated with barchan, linear, and star dunes are reasonably well established (Fryberger and Dean, 1979; Lancaster, 1995; Livingstone and Warren, 1996). Thus, stabilized dunes can yield past wind directions if the original dune form is well preserved. Some studies have shown dune geomorphic evidence of multiple shifts in past wind directions that reflect changes in synoptic-scale climatology.

Loess in the Quaternary

Loess occupies large areas of China, Central Asia, Russia, Europe, Argentina, New Zealand, and North America. Most loess is concentrated in mid-latitude regions (Figure 5), and its texture and mineralogy make it one of the most favorable parent materials for agricultural soils. Indeed, loess-derived soils produce crops that support literally millions of people in Asia, South America, North America, and Europe. Loess is dominated by silt-sized particles (50–2 μm diameter), although most loess contains smaller amounts of sand and clay. This particle size distribution, along with slight carbonate cementation, produces interparticle binding that makes loess stand in vertical bluffs in exposures (Figure 6).

The importance of loess in Quaternary studies is that it is one of the few terrestrial deposits that may contain nearly complete records of glacial-interglacial cycles. A classical model of loess is that it forms primarily during glacial periods, when continental ice sheets or valley glaciers produce silt-sized rock 'flour' by grinding. This silt is in turn transported and sorted by fluvial processes as outwash, deposited on floodplains, and ultimately entrained by the wind. It is now also known that nonglacial processes also produce silt-sized particles. Nevertheless, even in areas known to contain nonglacial loess, it appears that the most favorable times for loess entrainment, transportation, and deposition are during glacial periods. Mahowald et al. (1999) have pointed out several factors, including generally drier climates, decreased vegetation cover, stronger winds, and a decreased intensity of the hydrological cycle, that enhance loess and dust transport during glacial periods. The latter factor can account for loess or finer grained dust staying in suspension longer than is the case under nonglacial climates. Thus, Quaternary stratigraphic studies have shown that loess is characteristic of glacial periods in most regions (Figure 7).

Another important characteristic of loess for Quaternary paleoclimatic studies is that it can aid in reconstruction of past atmospheric circulation. Because loess is deposited directly by the wind, it is one of the most straightforward paleoclimate proxies that can be used to test atmospheric general circulation models. Loess properties that can be used to reconstruct paleowinds include decreasing thickness and particle size downwind from a source. In some regions, such as North America, paleowinds generated from loess deposits are at variance with modeled atmospheric circulation for the last glacial period (Muhs and Bettis, 2000). In other regions, such as South America, loess patterns are in good agreement with model reconstructions (Muhs and Zárate, 2001).

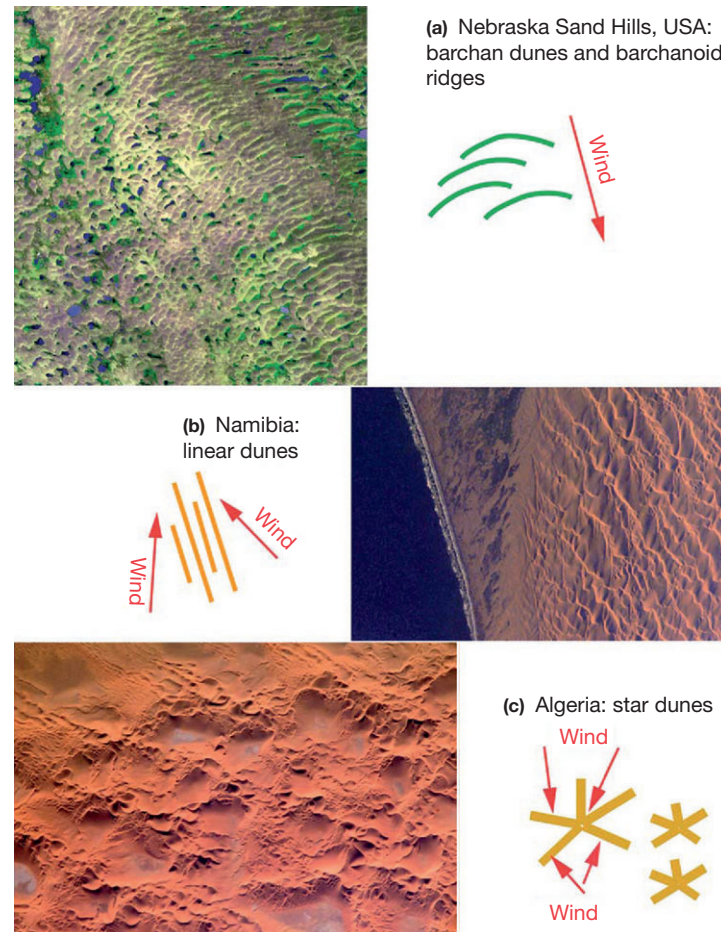


Figure 4 Space-based image gallery of dune forms and diagrams of predominant dune-forming winds: (a) simulated natural-color image of inactive (stable) barchan and barchanoid-ridge dunes in the Nebraska Sand Hills, USA, from the Advanced Spaceborne Thermal Emission and Reflection Radiometer (ASTER) on NASA's Terra satellite, acquired on 10 September 2001 (image courtesy NASA/GSFC/METI/ERSDAC/JAROS and US/Japan ASTER Science Team); (b) Namib sand sea, Namibia, near Conception Bay (astronaut photograph ISS011-E-9756 taken on 28 June 2005 from the International Space Station; photograph courtesy of the ISS Crew Earth and the Image Science and Analysis Group, Johnson Space Center); (c) star dunes in the Issaouane Erg, eastern Algeria (astronaut photograph ISS010-E-13539 taken on 16 January 2005 from the International Space Station; photograph courtesy of the ISS Crew Earth and the Image Science & Analysis Group, Johnson Space Center).

Fine-Grained Eolian Sediments in the Quaternary

Far-traveled dust (also called aerosolic dust and LRT dust) is one of the most interesting and best studied sediments in the Quaternary paleoclimate record. LRT dust consists largely of particles $<10\ \mu\text{m}$ in diameter; thus, it has some overlap with loess, and indeed some LRT dust is probably the distal, or farthest traveled, component of loess. Aerosolic or LRT dust is archived principally in three kinds of settings: (i) deep-sea sediments, (ii) lake sediments, and (iii) ice caps. Rea reviews the important records of LRT dust deposition in deep-sea sediments. One distinct advantage of studying LRT dust in deep-sea sediments is that the eolian record can be placed directly within the global oxygen isotope stratigraphic record (Figure 7). There are important records of long-distance transport of aerosolic dust to the Pacific Ocean (largely from Asia) and the Atlantic Ocean (largely from Africa) throughout much of the Quaternary (Rea, 1994). Aerosolic, or LRT, dust is important for a number of reasons:

1. it is an indicator of source-area aridity;
2. it is an important tracer for atmospheric circulation;

3. it has the potential for altering Earth's heat balance (either cooling or warming) via effects on radiation transfer; and
4. it can provide iron and other nutrients to primary producers, such as phytoplankton, in the world's oceans. LRT dust can also be an important component of soils and may contribute significantly to soil fertility.

Soils and Paleosols in Quaternary Records

Soils are found on virtually every part of the Earth's land surface, other than areas covered by water bodies (lakes and rivers), glacial ice, or steep mountain slopes. The evolution of soil morphology is a function of the interacting factors of climate, organisms (plants and animals), relief or topography, parent material, and time (Birkeland, 1999; Jenny, 1941). There is a rich diversity of soils in North America (Figure 8), and it is the close association of soils with climate, vegetation, and age of parent material that gives soils their utility in Quaternary stratigraphic studies. Soils form in part by weathering of the host sediment or rock. Bettis reviews this topic in an article on weathering profiles. McFadden reviews soil

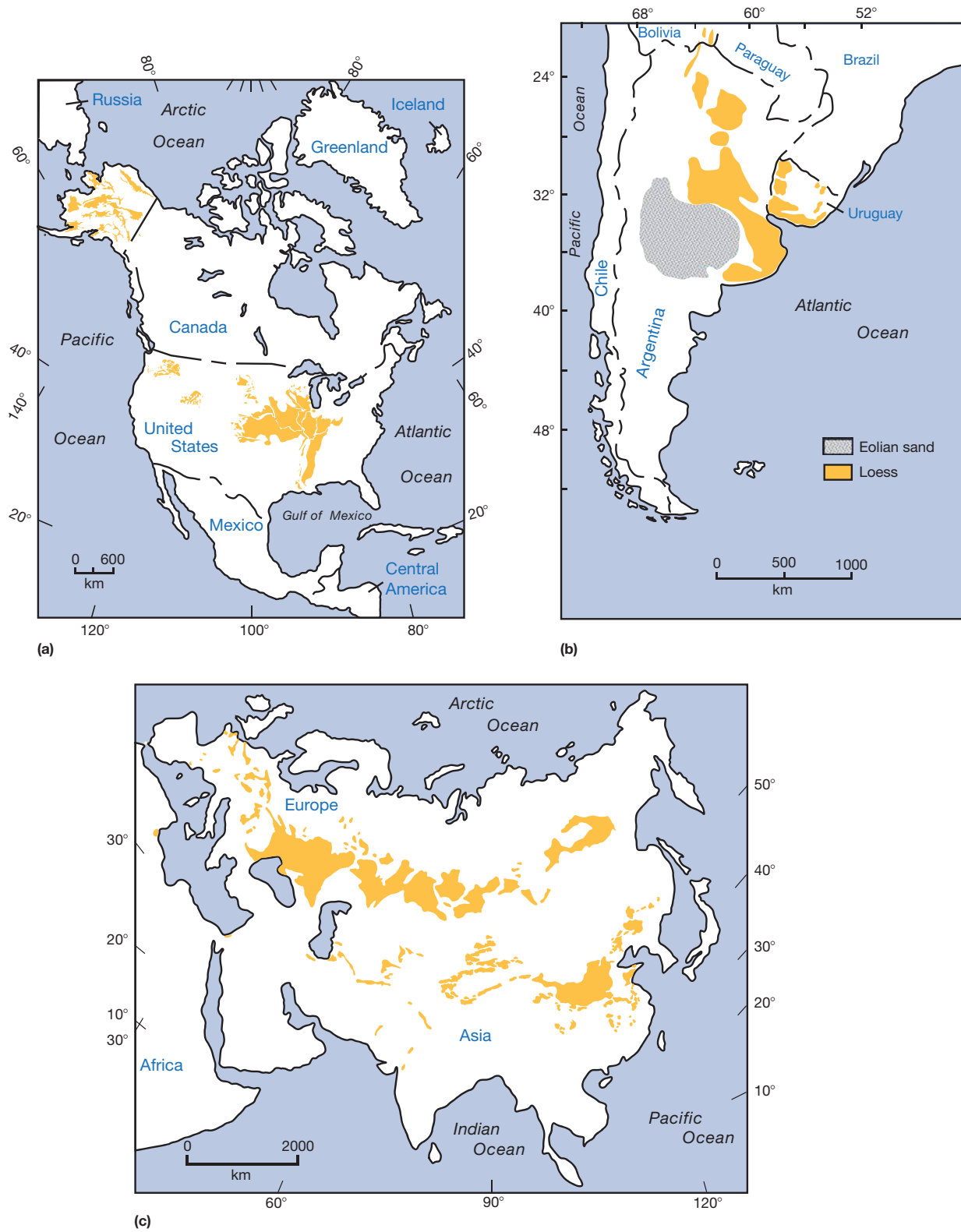


Figure 5 (a) Map showing the distribution of loess in North America, from compilations in Péwé (1975), Bettis et al. (2003), and Busacca et al. (2004), and references therein. (b) Map showing the distribution of loess and eolian sand in southern South America, redrawn from Zárate (2003) and sources therein. (c) Map showing the distribution of loess in Eurasia and localities or regions referred to in the text. Compiled from Rozycki (1991) and Liu (1988) and redrawn from figure in Muhs and Bettis (2003).



Figure 6 Photographic gallery of late Quaternary loess deposits in North America: (a) Birch Hill, near Fairbanks, Alaska, USA; (b) Eustis, Nebraska, USA; (c) Loveland paratype locality near Council Bluffs, Iowa, USA; (d) Crowley's Ridge, Arkansas, USA. All photographs by D.R. Muhs, USGS.

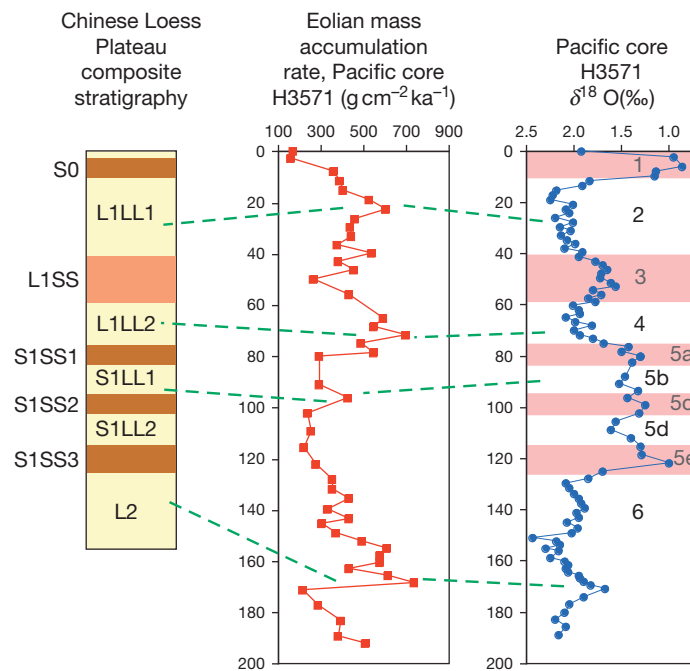


Figure 7 Plots of eolian mass accumulation rates as a function of age (estimated from oxygen-isotope stratigraphy, also shown) in western Pacific core H3571 (drawn from data tabulated in Kawahata et al., 2000). Also shown is generalized stratigraphy for the late Quaternary in the Chinese Loess Plateau (stratigraphy redrawn from Porter SC (2001) Chinese loess record of monsoon climate during the last glacial–interglacial cycle. *Earth-Science Reviews* 54: 115–128).

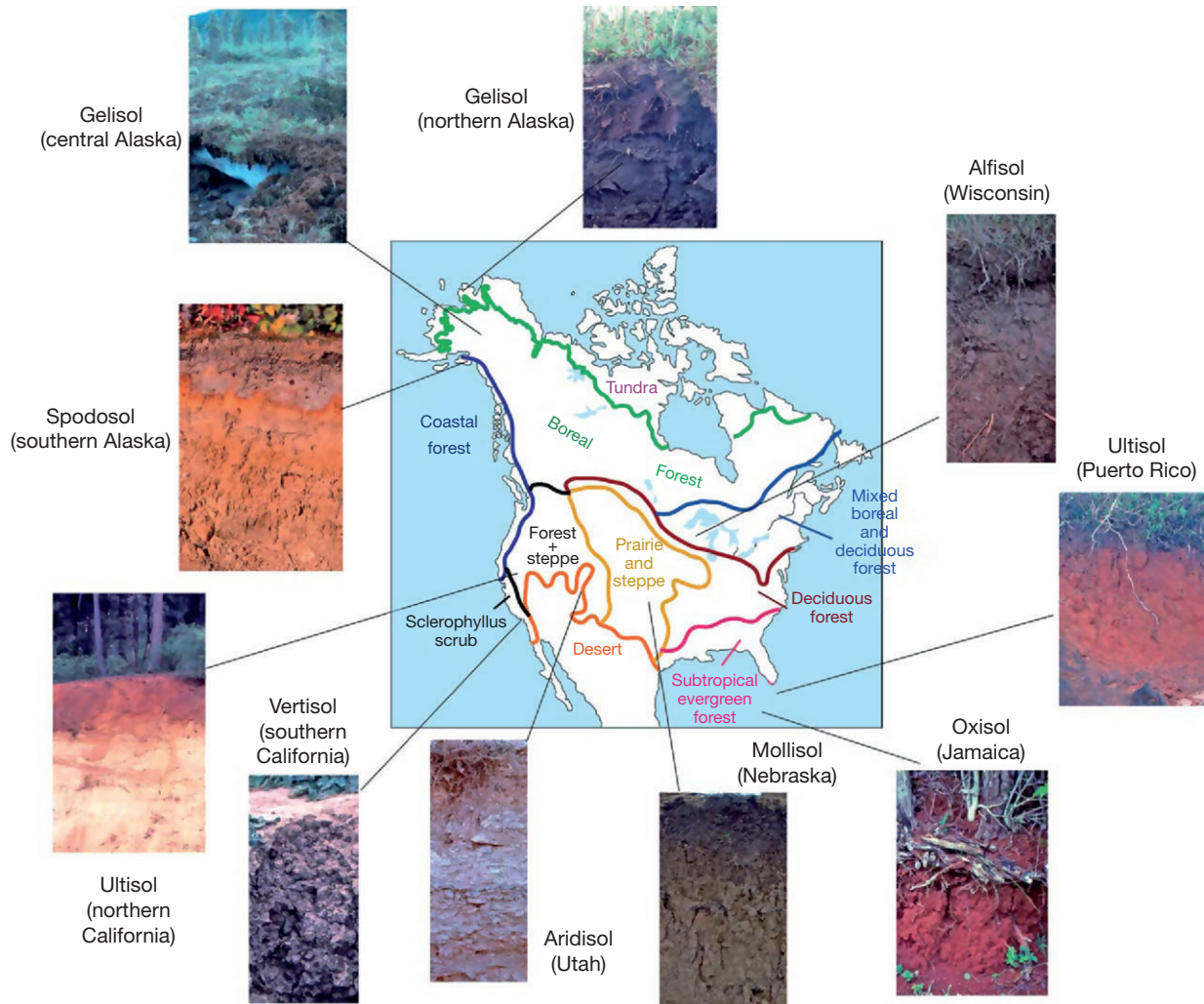


Figure 8 Photographic gallery of typical soils in different orders of the US Soil Taxonomy, shown as a function of major biomes of North America. All photographs by D.R. Muhs, USGS.

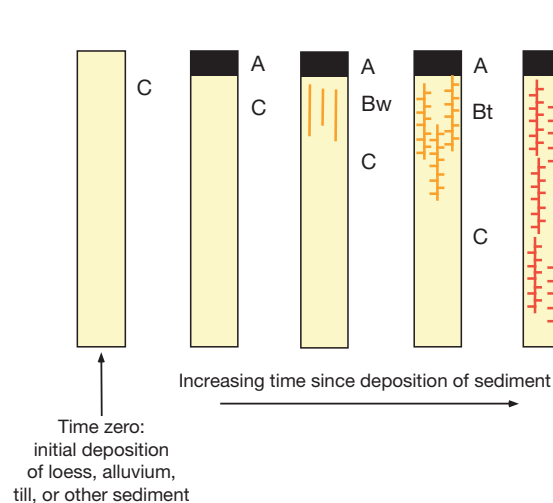


Figure 9 Hypothetical changes in soil morphology over time in a humid, mid-latitude climate, showing increasing thickness, clay content, and redness of B horizons. Horizontal black bars represent soil A horizons (topsoil); vertical brown (younger soils) or red (oldest soil) lines represent soil B horizons (subsoil).

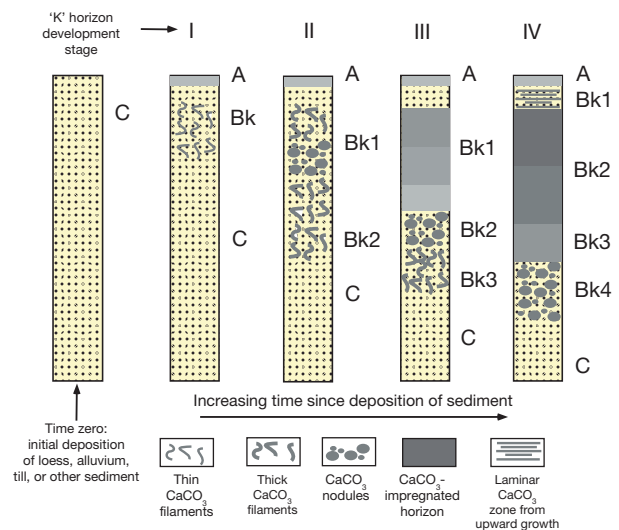


Figure 10 Hypothetical changes in soil morphology over time in an arid climate, showing increasing thickness, carbonate content, and complexity of Bk (carbonate-rich) horizon. Based on concepts in Gile et al. (1966).

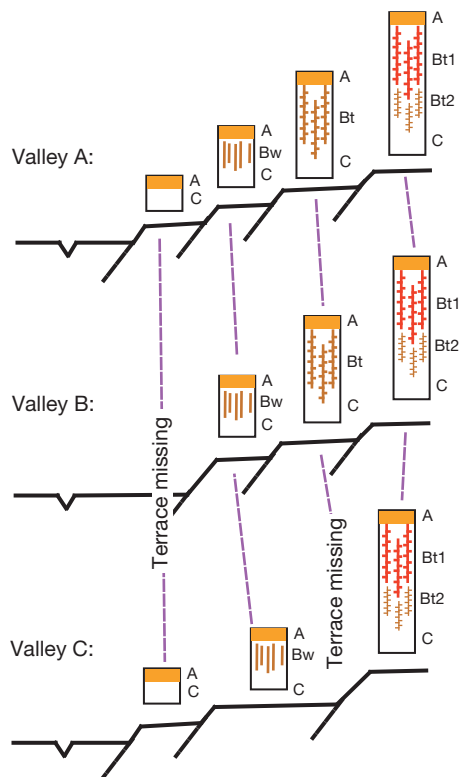


Figure 11 Hypothetical example of the application of soil morphology to correlation of stream terraces between valleys. See text for discussion.

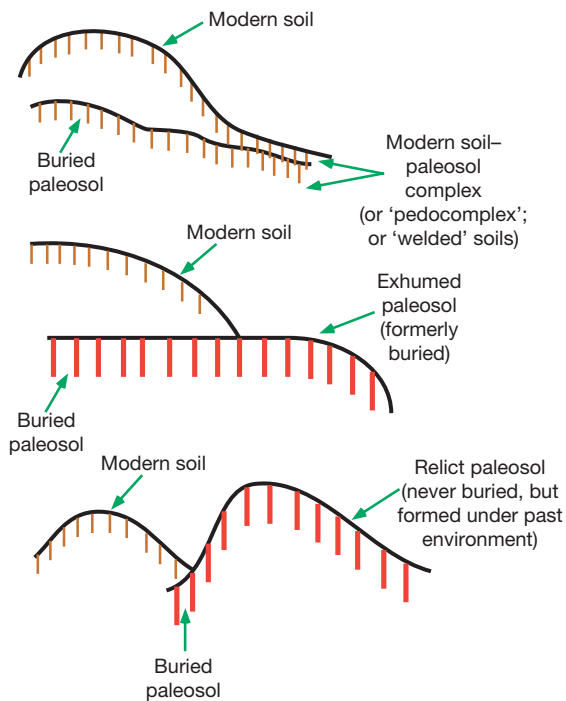


Figure 12 Hypothetical landscapes illustrating three types of paleosols. Upper: buried paleosol on left and 'welded' (joined) buried paleosol and modern soil; middle: buried paleosol on left and exhumed (formerly buried, but now at surface) paleosol; and lower: modern soil developed on recent landform, formed under present climate and vegetation, on left; on right, relict paleosol on surface, never buried, but formed primarily in different climate and vegetation than in the present.

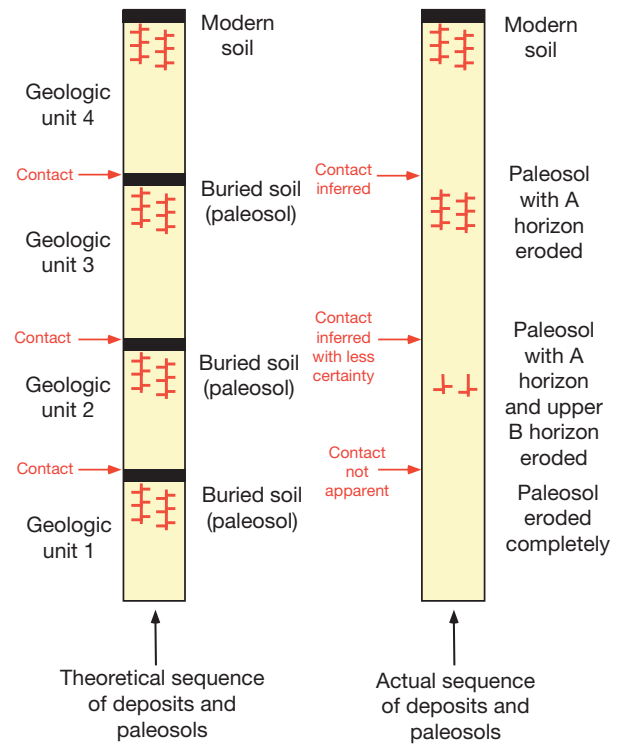


Figure 13 Theoretical (left) versus commonly observed (right) sequences of geologic deposits and paleosols. Horizontal black bars on the left represent soil A horizons (topsoil); vertical red lines represent soil B horizons (subsoil). See text for discussion.

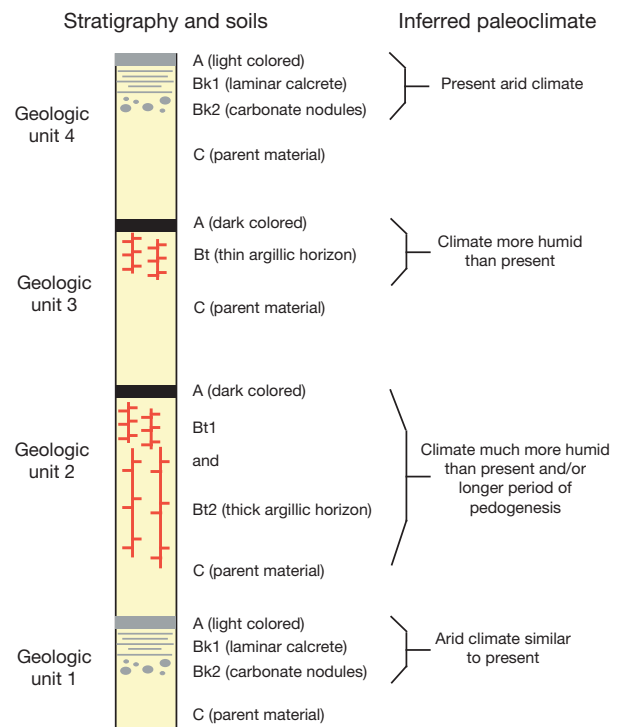


Figure 14 Hypothetical sequence of geologic units and paleosols, showing paleoclimatic interpretations of past soil-forming environments compared to present ones.

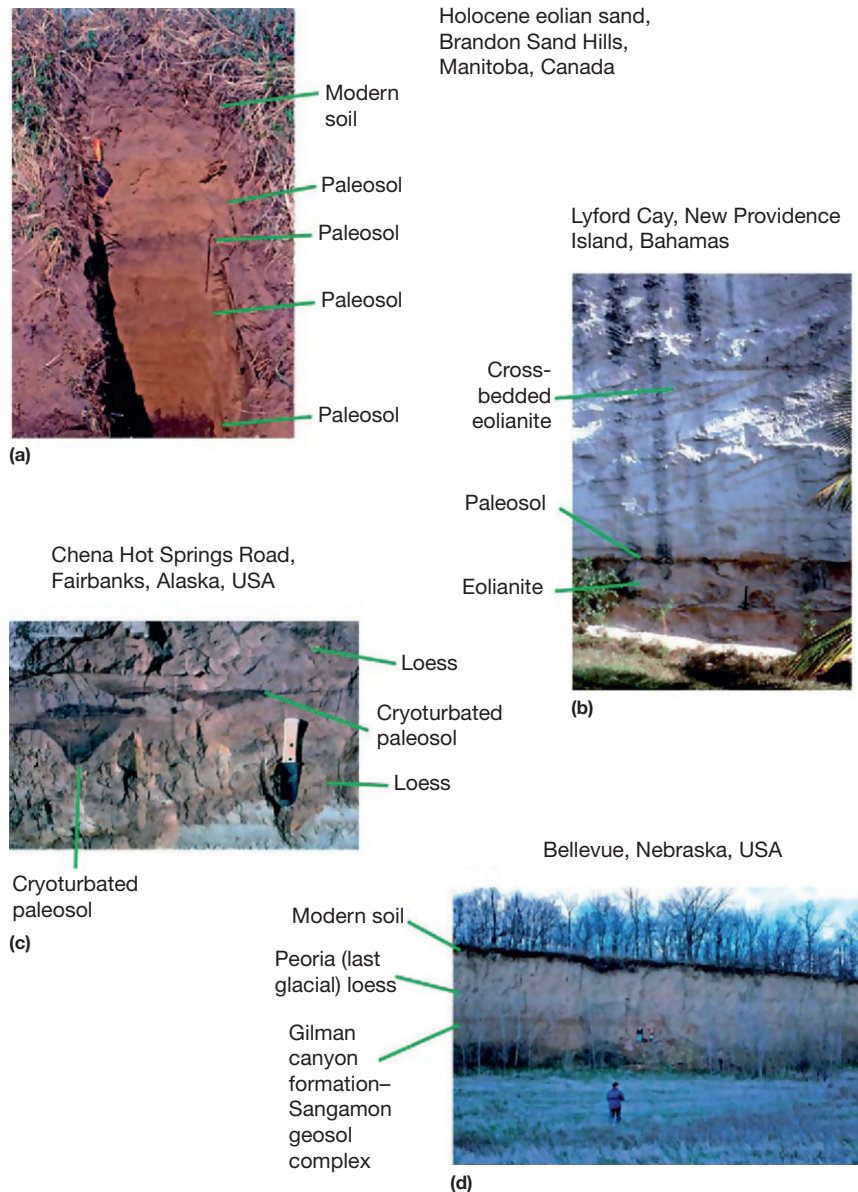


Figure 15 Photographic gallery of buried paleosols: (a) weakly developed paleosols in eolian sand in the Brandon Sand Hills, Manitoba, Canada; (b) paleosol developed in carbonate eolianite, Lyford Cay area, New Providence Island, Bahamas; (c) cryoturbated paleosol fragments in late Quaternary loess, Chena Hot Springs Road exposure, near Fairbanks, Alaska, USA; (d) Gilman Canyon Formation and Sangamon Geosol (paleosol) in Quaternary loess at Bellevue, near Omaha, Nebraska, USA. All photographs by D.R. Muhs, USGS.

morphology in Quaternary studies and Kemp discusses soils at a finer scale elsewhere in this encyclopedia.

Soil morphology proceeds along evolutionary pathways that differ from climate to climate (Figure 8). In the humid climates of the mid-latitudes, relatively fine-textured parent materials (fine-grained till, loess, alluvium, or lacustrine sediment) yield soil B horizons that tend to become thicker, more clay-rich, and redder over time (Figure 9). In contrast, in semi-arid and arid regions, where precipitation is too low for chemical leaching, carbonates build up over time (Figure 10). Thus, increasing overall carbonate content, and thickness and complexity of carbonate horizons are characteristic of soil evolution over time in dry regions (Gile et al., 1966).

Evolution of soils through time can be applied to Quaternary stratigraphic studies. Considerable success has been achieved in

this application by the career-long efforts of Birkeland (1999). A hypothetical example of the use of soils in stratigraphy, specifically valley-to-valley stream-terrace correlation, illustrates the approach. In Figure 11, Valley A has four terraces with better developed soils on successively higher terraces, whereas Valleys B and C each have only three terraces. Soil morphology indicates that the first terrace of Valley A is missing in Valley B and the third terrace of Valley A is missing in Valley C. This simple example can be extended to valley-to-valley correlation of glacial sequences, colluvial deposits, eolian sands, marine terraces, and ancient lacustrine deposits. The approach is reliable and inexpensive and allows stratigraphic correlation when materials suitable for numerical dating are absent.

Paleosols are soils of the past (Figure 12). What constitutes a paleosol has been the subject of considerable debate, and

diverse points of view are presented in Morrison (1998). Morrison proposed that 'geosol' be used to denote buried paleosols and succeeded in placing this term in the North American Stratigraphic Code. Despite this, the term has not gained general acceptance by many Quaternary stratigraphers and pedologists, and the term 'buried paleosol' has continued to be used by most investigators.

Two important aspects of paleosols are that they are key stratigraphic markers and can be used in paleoenvironmental interpretations. As a stratigraphic marker, a buried paleosol's uppermost boundary provides the contact between geologic deposits of two different ages (Figure 13). As such, paleosols are often crucial in interpreting Quaternary stratigraphy. It is important to remember, however, that many buried paleosols are remnants of once-thicker soils, and in some cases, they have been eroded completely (Figure 13). Nevertheless, many Quaternary sections contain buried paleosols with sufficient preservation to make paleoenvironmental interpretations possible. A hypothetical example of such a situation is shown in Figure 14, where the present arid climate has produced a modern soil with a well-developed Bk horizon. Below this is a buried paleosol that lacks a Bk horizon but has a thin Bt horizon, suggesting a wetter past climate that was sufficient to leach carbonates and allow clay illuviation. Still deeper is a buried paleosol with an even thicker Bt horizon, which could suggest a very humid climate and/or a much longer period of pedogenesis than that of the paleosol immediately above it. Finally, at the base of the hypothetical section is a buried paleosol with a Bk horizon, suggesting a time when the climate was arid, as is the case today. Thus, the overall sequence represents a history as follows:

- an arid climate, much like today;
- a long humid period, much wetter than today;
- a shorter, humid period; and
- the present arid climate.

This kind of application of soil morphology to paleoclimatic reconstruction has been particularly fruitful in loess sequences, where paleosols are often well preserved (Figure 15).

Conclusion

Eolian sediments (sand, loess, and LRT or aerosolic dust) are important archives of the Quaternary geologic and paleoclimatic record. They are extensive on Earth's surface at all latitudes and can be dated using luminescence methods that require only the sediment. Unlike many Quaternary sediments, eolian deposits are direct indicators of atmospheric circulation. Furthermore, they are robust indicators of past aridity. In addition, fine-grained eolian sediments may have effects on climate itself, both directly, in terms of radiative transfer via albedo effects, and indirectly, caused by fertilization of primary producers that affect the global carbon cycle.

Soils and paleosols are also key components of the Quaternary stratigraphic record. They cover a majority of the Earth's land surface and are closely linked to climate, vegetation, and the length of time that a surface or deposit has been exposed. Paleosols, soils of the past, are important stratigraphic markers in the Quaternary record. Under conditions of good preservation, paleosols contain rich records of climate history.

See also: Eolian Records, Deep-Sea Sediments; Loess Deposits: Origins and Properties. **Dune Fields:** High Latitudes; Low Latitudes; Mid-Latitudes. **Loess Records:** Central Asia; China; Europe; North America; South America. **Luminescence Dating:** Optical Dating. **Paleosols and Wind-Blown Sediments:** Mineral Magnetic Analysis; Nature of Paleosols; Soil Micromorphology; Soil Morphology in Quaternary Studies; Weathering Profiles.

References

- Aitken MJ (1998) *An Introduction to Optical Dating*. Oxford: Oxford University Press.
- Bettis EA III, Muhs DR, Roberts HM, and Wintle AG (2003) Last glacial loess in the conterminous U.S.A. *Quaternary Science Reviews* 22: 1907–1946.
- Birkeland PW (1999) *Soils and Geomorphology*. New York: Oxford University Press.
- Bowler JM, Wyrwoll K-H, and Lu Y (2001) Variations of the northwest Australian summer monsoon over the last 300,000 years: The paleohydrological record of the Gregory (Mulan) Lakes System. *Quaternary International* 83–85: 63–80.
- Busacca AJ, Begét JE, Markewich HW, Muhs DR, Lancaster N, and Sweeney MR (2004) Eolian sediments. In: Gillespie AR, Porter SC, and Atwater BF (eds.) *The Quaternary Period in the United States*, pp. 275–309. Amsterdam: Elsevier.
- Fryberger SG and Dean G (1979) Dune forms and wind regime. In: McKee ED (ed.) *A Study of Global Sand Seas*, pp. 137–169. Washington: U.S. Geological Survey. Professional Paper 1052.
- Gile LH, Peterson FF, and Grossman RB (1966) Morphological and genetic sequences of carbonate accumulation in desert soils. *Soil Science* 101: 347–360.
- Goudie AS (2002) *Great Warm Deserts of the World*. Oxford: Oxford University Press.
- Jenny H (1941) *Factors of Soil Formation*. New York: McGraw-Hill.
- Kawahata H, Okamoto T, Matsumoto E, and Ujiiie H (2000) Fluctuations of eolian flux and ocean productivity in the mid-latitude North Pacific during the last 200 kyr. *Quaternary Science Reviews* 19: 1279–1291.
- Lancaster N (1989) *The Namib Sand Sea: Dune Forms, Processes and Sediments*. Rotterdam: A.A. Balkema.
- Lancaster N (1995) *Geomorphology of Desert Dunes*. London: Routledge.
- Liu T (1988) *Loess in China*, 2nd edn. Beijing: China Ocean Press. Berlin: Springer.
- Livingstone I and Warren A (1996) *Aeolian Geomorphology: An Introduction*. Essex: Addison Wesley Longman Limited.
- Mahowald N, Kohfeld K, Hansson M, et al. (1999) Dust sources and deposition during the last glacial maximum and current climate: A comparison of model results with paleodata from ice cores and marine sediments. *Journal of Geophysical Research* 104: 15895–15916.
- McKee ED (1983) Eolian sand bodies of the world. In: Brookfield ME and Ahlbrandt TS (eds.) *Eolian Sediments and Processes. Developments in Sedimentology*, vol. 38, pp. 1–25. Amsterdam: Elsevier.
- Morrison RB (1998) How can the treatment of pedostratigraphic units in the North American Stratigraphic Code be improved? *Quaternary International* 51–52: 30–33.
- Muhs DR and Bettis EA III (2000) Geochemical variations in Peoria loess of western Iowa indicate paleowinds of midcontinental North America during last glaciation. *Quaternary Research* 53: 49–61.
- Muhs DR and Bettis EA III (2003) Quaternary loess–paleosol sequences as examples of climate-driven sedimentary extremes. *Geological Society of America Special Paper* 370: 53–74. Special Paper.
- Muhs DR and Holliday VT (1995) Evidence for active dune sand on the Great Plains in the 19th century from accounts of early explorers. *Quaternary Research* 43: 198–208.
- Muhs DR and Zárte M (2001) Late Quaternary eolian records of the Americas and their paleoclimatic significance. In: Markgraf V (ed.) *Interhemispheric Climate Linkages*, pp. 183–216. San Diego, CA: Academic Press.
- Péwé TL (1975) Quaternary geology of Alaska. *U.S. Geological Survey Professional Paper* 835: 1–145.
- Pye K (1987) *Aeolian Dust and Dust Deposits*. San Diego, CA: Academic Press.
- Rea DK (1994) The paleoclimatic record provided by eolian deposition in the deep sea: The geological history of wind. *Reviews of Geophysics* 32: 159–195.
- Rozyczki SZ (1991) *Loess and Loess-Like Deposits*. Warsaw: Ossolineum Press, Polish Academy of Sciences.
- Schaezel RJ and Anderson S (2005) *Soils: Genesis and Geomorphology*. Cambridge: Cambridge University Press.
- Thomas DSG and Shaw PA (1991) *The Kalahari Environment*. Cambridge: Cambridge University Press.
- Thorp J and Smith HTU (1952) Pleistocene eolian deposits of the United States, Alaska, and parts of Canada. *National Research Council Committee for the Study of Eolian Deposits*, Geological Society of America, scale 1:2,500,000.
- Wilson IG (1973) Ergs. *Sedimentary Geology* 10: 77–106.
- Zárte MA (2003) Loess of southern South America. *Quaternary Science Reviews* 22: 1987–2006.

Nature of Paleosols

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Introduction

Paleosols are soils ‘formed in a landscape of the past,’ often but not always buried, which provide crucial evidence on past land surfaces and environments. Quaternary paleosols were recognized in the nineteenth century, if not earlier. In 1866, A.H. Worthen described a soil horizon separating glacial till from overlying loess in central Illinois, United States, later named the Sangamon Soil. The term ‘soil’ was poorly defined in this early work, however, and often referred to sediments rich in organic matter rather than to well-defined soil profiles as we recognize them today. In Russia, [Dokuchaev \(1883, pp. 94–95, 162\)](#) described soils buried by dune sand and slope-wash. By the 1920s and 1930s, pedologists increasingly drew much more specific analogies between surface soil profiles and buried paleosols. Since that time, Quaternary researchers in many parts of the world have used such analogies to infer paleoclimate from paleosol morphology. Finally, in the late twentieth century, it was recognized that the great loess–paleosol sequences of China and other regions represent the only terrestrial records of Quaternary climate change comparable in length and completeness to those available from marine cores.

Paleosols still play a crucial role in Quaternary studies, both as stratigraphic markers and as sources of paleoclimatic data. More recent work has revealed the complexity of paleosol genesis, however, often leading to qualification of earlier interpretations. Quaternary paleosols, especially those in loess successions, rarely represent a simple sequence of parent material deposition followed by top-down pedogenesis. In fact, many paleosols reflect the interaction of sedimentary and pedogenic processes, with pedogenesis occurring in actively accumulating parent sediments. Thus, the morphological features of many Quaternary paleosols reflect not only the climate under which they formed, but also parent material sedimentation rate relative to the rate of soil-forming processes. Finally, recent decades have seen increasing research on paleoenvironmental reconstruction based on magnetic or geochemical proxy data extracted from paleosols.

Issues of Terminology

The term ‘paleosol’ is applied to buried soils, soils that were once buried but are now exhumed by erosion and form the modern surface soil, and even relict soils that preserve evidence of ancient soil-forming environments but have never been buried. The focus of this article is primarily on buried paleosols. The North American Stratigraphic Code uses the term ‘geosol’ (introduced by [Morrison, 1967](#)) for formally defined pedomorphological units, which must be buried paleosols, but this term has seen little use elsewhere.

Where an existing soil is buried by a thin increment of sediment, subsequent surface soil development may extend downward into the buried paleosol. The two overlapping profiles are then referred to as either a ‘composite soil’ or ‘welded soils’; ‘pedocomplex’ is also used to refer to this phenomenon. In some settings, a composite paleosol may diverge laterally into two or more discrete paleosols. Such divergence may be observed over short distances in alluvial or hillslope deposits; for example, a single thick surface soil may split into several distinct surface and buried soils with distance down a single footslope. Over greater distances, multiple buried soils in source-proximal loess can converge into a single composite paleosol at distal sites.

Most researchers restrict the term paleosol to horizons representing a significant period of surface exposure, and significant alteration from the original condition of the parent material. However, there are no simple, widely agreed-upon criteria for the threshold of pedogenesis above which a paleosol can be identified.

This article uses the horizon designations adopted by US pedologists, including A (surficial horizons of organic matter accumulation), E (eluvial horizon below the A, often light colored), and B (subsoil horizons).

Recognition of Paleosols

Analogy with surface soils remains the essential type of evidence used in identifying buried paleosols. Macro- and micromorphological features diagnostic of specific pedogenic processes are often most convincing, but multiple lines of evidence are generally needed to identify paleosols that represent a significant period of surface exposure and pedogenesis. For example, some fine-grained sediments, including unoxidized glacial tills, quickly develop a form of blocky structure on exposure in road- or streamcuts, which might be mistaken as evidence for more protracted pedogenesis. A sequence of horizons in which the structure varies in a pattern common to many surface soils (e.g., a downward transition from fine granular to blocky structure that coarsens with depth) provides more convincing evidence for a significant period of pedogenesis, and microscopic evidence of extensive clay illuviation or mineral weathering probably indicates thousands of years of surface exposure. Geochemical, mineralogical, or rock-magnetic data can be valuable in characterizing the degree of parent material alteration by weathering and other pedogenic processes, but in themselves are rarely useful in paleosol recognition. Few paleopedologists would accept the identification of a buried soil simply on the basis of measurements such as magnetic susceptibility, in the absence of morphological evidence visible in the field.

Some paleosols are so similar in appearance to modern soils that there is little debate over their pedogenic origin. This is particularly true when a dark-colored A horizon is preserved, a common characteristic of Holocene paleosols in dune sand, loess, tephra, or alluvium (Figure 1). Dark-colored beds of sedimentary origin do occur in alluvium, but generally can be identified by lack of pedogenic features or the presence of undisturbed sedimentary structures. In paleosols without a dark-colored A horizon, strongly expressed pedogenic structure and color may still allow general consensus on the presence of a buried soil (Figures 2 and 3(a)–3(c)), but there is often disagreement on both the assignment of specific horizon labels (A, B, etc.) and the relationship of the paleosol to modern soil taxa. These issues are even more contentious in the case of weakly expressed, light-colored paleosols. Sediment that accumulated on vegetated surfaces, such as loess, often displays evidence of root growth or animal burrowing but this is rarely considered sufficient for identification as a paleosol (Figure 3(b)).



Figure 1 Paleosols with well-preserved, dark-colored A horizons. (a) Buried soils in a loess exposure in western Nebraska, United States. The most prominent dark band (marked by arrow) is the A horizon of the Brady Soil, which separates late Pleistocene and Holocene loess units across much of the central Great Plains region. Lighter bands above the Brady Soil are weak soils developed during the more humid intervals of the Holocene, separated by increments of more rapidly deposited loess. (b) Dark buried A horizon within dune sand in eastern Inner Mongolia, China.

Another problem arises when paleosols are identified in stratified sediments in which individual beds have highly contrasting colors and/or textures. This commonly leads to disagreement over whether particular horizons are pedogenic or sedimentary in origin. To some extent, this is a false dichotomy, since the boundaries of pedogenic horizons may tend to develop at lithologic discontinuities that influence soil water movement and related pedogenic processes.

Exhumed paleosols are generally only recognizable where they can be traced into the subsurface. This is the case across large areas of the midwestern United States outside the limits of the last glaciation (Marine Isotope Stage (MIS) 2). The Sangamon Geosol formed during the last interglaciation (MIS 5) and the subsequent transition back to glacial conditions during MIS 4. It is buried in many places but has been exposed by erosion on sideslopes and in effect forms the surface soil there. As discussed below, relict paleosols are clearly identifiable only where their morphology records paleoenvironmental conditions far different from those of the present.

Paleosols as Stratigraphic Markers and Numerical Dating of Paleosols

Paleosols by their nature mark former land surfaces. If a paleosol contains horizons that typically take thousands to tens of thousands of years to develop, then the surface it formed beneath was eroding or aggrading quite slowly, if at all. Thus, significant portions of the time represented by an entire stratigraphic column may be compressed into the paleosol profile.

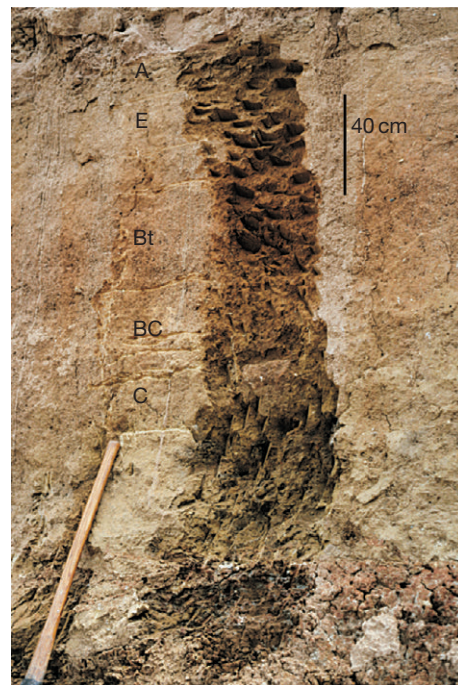


Figure 2 Profile of the Sangamon Geosol in southern Illinois, formed in Illinoian (MIS 6) glacial till, and buried by loess of the last glaciation (MIS 3 and 2). Although a dark A horizon is not preserved, the reddish brown Bt horizon and overlying light-colored E horizon form a profile closely analogous with modern forest soils of the region. Illinoian till (C horizon) overlies an older paleosol visible near bottom of photo.

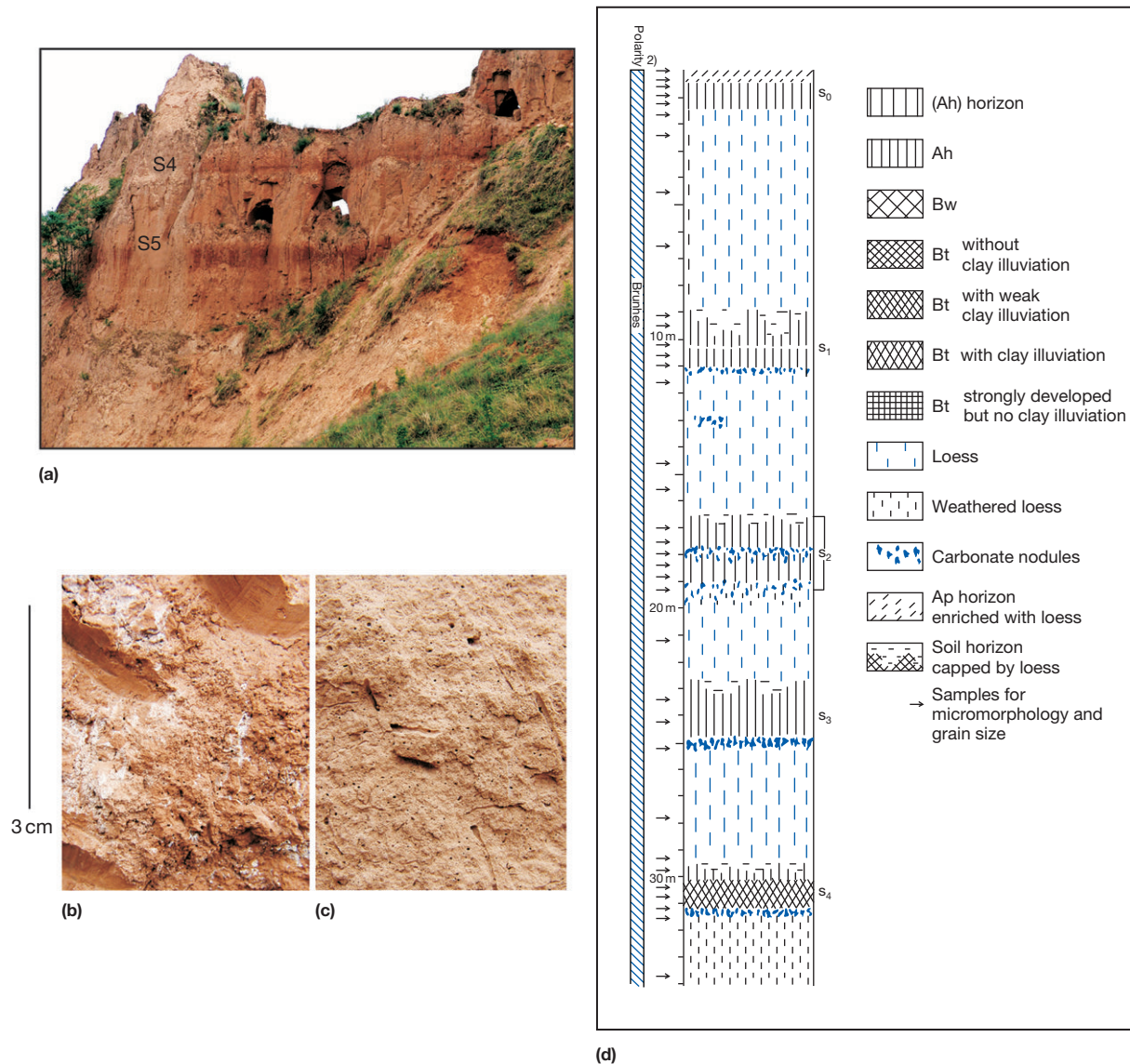


Figure 3 Paleopedology of the Luochuan loess section, Chinese Loess Plateau. (a) Photo of part of the section, with paleosols S4 and S5 marked. (b) Close-up of paleosol S1, which has reddish brown color similar to many B horizons, but micromorphological characteristics of a bioturbated A horizon. (c) Close-up of loess below S1 (L2), with displays abundant biopores indicating accumulation on a vegetation-covered surface. (d) Interpretation of paleosols S1 through to S4 in terms of their closest modern analogs and environment of formation, based mainly on inference of pedogenic processes from micromorphology. Adapted from Bronger A and Heinkele T (1989) Micromorphology and genesis of paleosols in the Luochuan loess section, China: Pedostratigraphic and environmental implications. *Geoderma* 45: 123–143.

Such relatively stable surfaces marked by well-developed paleosols are often of great interest within Quaternary deposits. In glaciated regions of North America and Europe, each major glaciation in effect created a new land surface underlain by glacial till, outwash, loess, and related deposits. That surface was then exposed to pedogenesis for tens of thousands of years before being buried by sediments of the next glaciation. Through this process, paleosols can potentially mark the boundaries of each glacial–interglacial cycle. In reality, subglacial erosion must often remove paleosols that are directly overridden by ice, since they are only preserved in some of the regions that have experienced multiple glaciations. In the central United States, the Sangamon Geosol is extensively

preserved within and beneath loess deposits beyond the ice margin of the last glaciation (MIS 2). This paleosol is also often present beneath glacial sediment from MIS 2 in Illinois, near the southern limit of the MIS 2 ice sheet. In Michigan, Wisconsin, and Minnesota, states of the northern United States covered by thicker and longer-lasting glacial ice during MIS 2, there are few reported localities where the Sangamon Geosol is preserved under younger glacial sediment. Prominent paleosols in the loess successions of China and Central Asia also mark global interglacial periods, based on the fairly sparse age control now available, mainly from magnetostratigraphy.

In theory, paleosols that form stratigraphic markers may be time transgressive. For example, as continental ice sheets grew

during the last glaciation, soils farther south would have been buried later. However, age control on paleosols rarely has sufficient resolution to clearly demonstrate time transgression. Uncertain age control also makes it difficult to test the assumption that most strongly expressed paleosols form during interglacial periods, and especially during short intervals of peak interglacial warmth. In some cases, geochronological data clearly disprove this assumption. For example, a prominent paleosol in the Palouse Loess of eastern Washington, United States, formed during the last glaciation, based on the ages of bracketing tephras (**Figure 4**).

This example illustrates the importance of numerical dating methods in the interpretation of paleosols. The most straightforward application of numerical dating is to bracket the time of soil formation, by dating deposition of both the paleosol's parent material and the sediment that buries it. Techniques useful for this purpose include luminescence dating; radiocarbon dating of detrital wood, charcoal, or other material; and tephrochronology. The duration of pedogenesis, and in certain cases the numerical age of overlying sediments, can be estimated from measurements of cosmogenic nuclides (beryllium-10, aluminum-26, etc.) within a paleosol. These nuclides are either produced *in situ* within soil minerals or produced in the atmosphere and added to the soil in precipitation. Radiocarbon dating of soil organic matter in paleosols generally underestimates the age of the parent material, at least

slightly overestimates the age of overlying sediment, and provides little information on the duration of pedogenesis. Nonetheless, this technique has provided valuable information in settings where other dating methods are not applicable or produce questionable results.

Inference of Paleoenvironments from Paleosol Morphology

Surface soils display regional or local patterns of variation that can be used to infer the paleoenvironments in which paleosols developed. Such analogies can provide information on paleoclimate, paleoecology, and even paleohydrology.

Long-recognized regional and global patterns of soil geography are especially useful in the interpretation of paleosols. For example, as first clearly documented by **Dokuchaev (1883)**, soils on stable upland summits in temperate grasslands commonly display relatively thick, dark colored, highly bioturbated A horizons, while temperate forest soils typically have thin A horizons underlain by light-colored eluvial horizons. **Bronger (1991)** provides a comprehensive overview of changes in soil macro- and micromorphology across the forest to grassland transition in Eurasia and North America. These changes can potentially be used to identify former vegetation boundaries as recorded by regionally extensive paleosols.

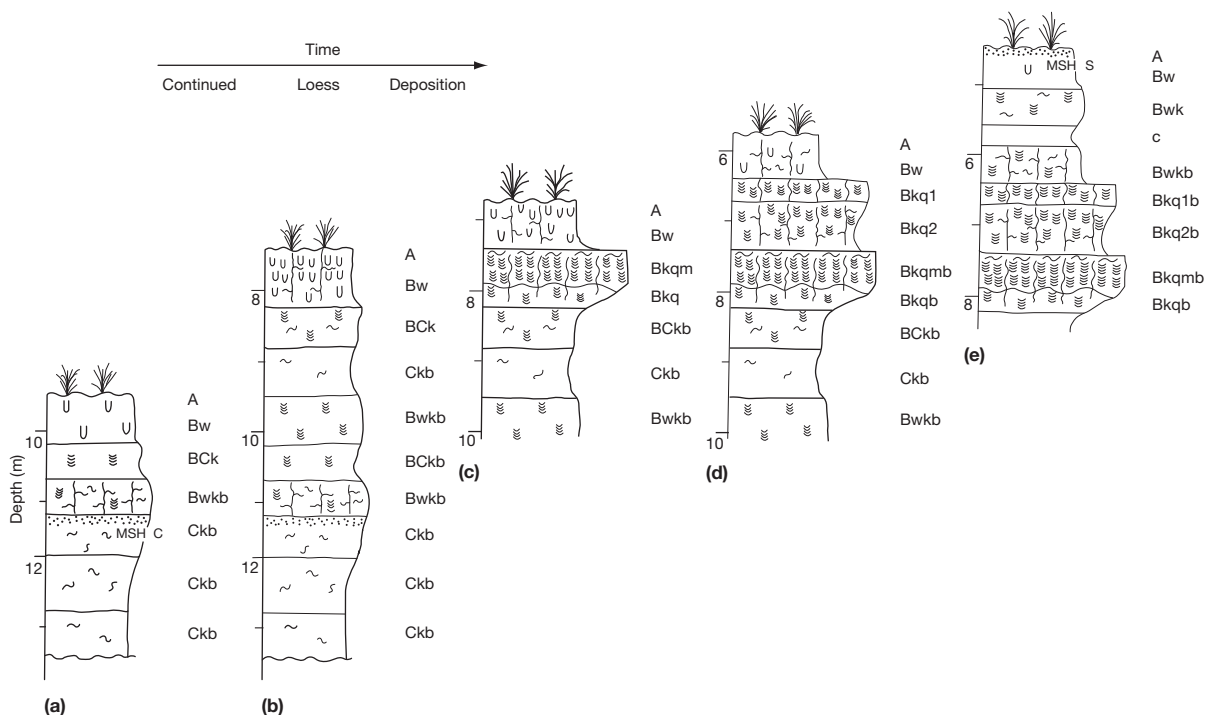


Figure 4 Aggradational pedogenesis in loess of the Palouse region, northwestern United States. Former A horizons burrowed by insects (cicadas) are buried by additional loess sedimentation, and are transformed into calcium carbonate- and silica-cemented B horizons. In this case, aggradational pedogenesis produced a prominent composite soil with two peaks of carbonate and silica content. Development of this major paleosol occurred within the last glaciation, not under interglacial conditions, because it is bracketed by tephra from the Mount St. Helens (MSH) volcano. Tephra are labeled: MSH-C (erupted about 43 000 years ago), and MSH-S (about 15 400 years ago). Reproduced from McDonald EV and Busacca AJ (1990) Interaction between aggrading geomorphic surfaces and the formation of a late Pleistocene paleosol in the Palouse Loess of eastern Washington State. In: Kneupper PLK and McFadden LD (eds.) *Soils and Landscape Evolution. Proceedings of the 21st Annual Binghamton Symposium in Geomorphology*, pp. 449–470. Amsterdam: Elsevier, Geomorphology.

Secondary calcium carbonate accumulation and clay illuviation can also provide evidence of the paleoenvironments in which buried soils formed. In well-drained soils, secondary carbonate accumulation is much more common in regions where evaporation exceeds precipitation than in more humid areas, and the depth to secondary carbonate decreases with decreasing rainfall. In temperate Eurasia, illuvial clay pedofeatures are generally believed to be limited to forest-covered regions and the forest-steppe transition zone, where leaching of base cations has been sufficient to allow clay dispersion and translocation. This pattern has often been disputed in North America, however, on the basis of clay-rich B horizons in dry regions.

Weathering, oxidation, and reduction can also leave evidence of past environments of soil formation. Significant depletion of weatherable minerals and formation of kaolinite-rich clay mineral suites is generally associated with warm humid climates of the tropics and subtropics, although deep weathering may also occur in temperate regions given enough time. In fact, the most convincing examples of relict paleosols are soil profiles displaying mineralogical evidence of intense weathering in regions that are now relatively dry and/or cool. Poorly drained soils commonly display features indicative of episodic or sustained reducing conditions. In parts of the midwestern United States where the Sangamon Geosol formed in glacial sediments, it can be assigned to drainage classes through morphological and mineralogical analogies with surface soils. Finally, distinctive soil faunal burrows and root traces may help identify the ecosystems within which paleosols developed.

These spatial patterns and their relation to paleoenvironmental factors are often difficult to quantify. Nonetheless,

there are many good examples of qualitative or semiquantitative paleoenvironmental reconstruction using paleosol morphology. In the classic Chinese loess section of Luochuan (Figure 3), neither the Holocene soil (S0) nor the upper three Pleistocene soils (S1 through to S3) display micromorphological evidence of significant clay translocation. However, illuvial clay is present in many of the older paleosols. The exact climatic contrast recorded by this upward change in pedogenic regime is difficult to specify, but it clearly suggests that precipitation during interglacial periods was greater before MIS 9 (correlated with S3), allowing more effective leaching of base cations and perhaps sustaining forest or forest-steppe vegetation. Soils S0 through to S3 are all interpreted as having formed under drier conditions and steppe vegetation. Consistent with this interpretation, other researchers have found illuvial clay features in S1 near Xi'an, in the southeastern Loess Plateau where modern precipitation is greater than at Luochuan (Figure 5).

One of the most important complications in the paleoenvironmental interpretation of paleosols is the time dependence of soil morphological evolution. Boardman (1985) considered the case of interglacial paleosols in the United States and Europe, which are often more strongly developed and redder than soils formed during the Holocene. This could indicate that past interglaciations were warmer than the present one, but probably also reflects a greater duration of pedogenesis. In the subtropics of central Mexico, a sequence of paleosols formed in tephra provides evidence of the time dependence of soil morphological evolution. Paleosols with a shorter duration of pedogenesis have the chemical, mineralogical, and morphological characteristics of andosols (volcanic

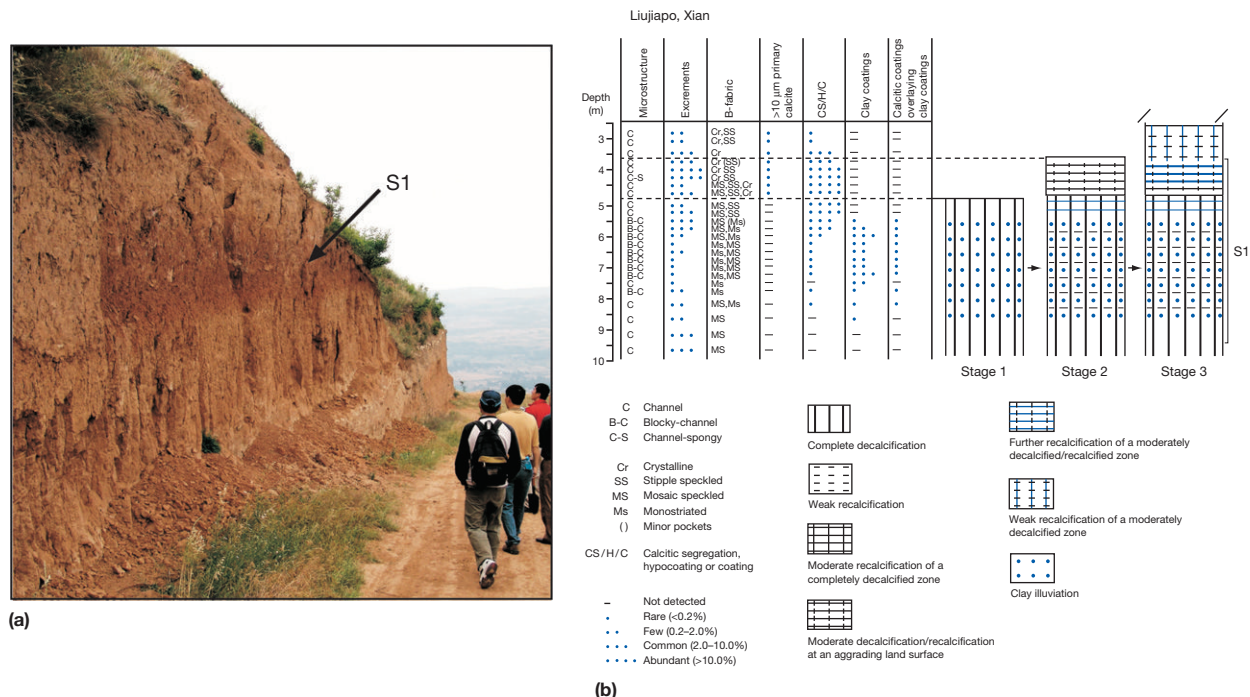


Figure 5 Pedosedimentary reconstruction of the development of S1, the paleosol representing the last interglacial period, at a site near Xi'an in the humid southeastern Loess Plateau, China. (a) Photo of S1 near Xi'an. (b) Diagram from Kemp et al. (1997), illustrating the reconstructed sequence of pedogenic and sedimentary events (right), and the supporting micromorphological data (upper left).

ash-derived soils with low bulk density, high organic matter, and abundant allophane), but with increasing duration of pedogenesis mineral weathering and clay illuviation become more extensive until gross morphology resembles Luvisols (forest soils with clay-enriched B horizons) that were previously interpreted as evidence of more humid climate.

Aggradational Pedogenesis and Postburial Alteration of Paleosols

Traditional conceptual models of soil genesis often assume top-down soil formation below a static land surface. In contrast, many paleosols, especially those in loess, volcanoclastic sediment, alluvium, or colluvium, developed beneath an episodically or continually aggrading land surface. In essence, soil-forming processes competed with, and in some cases were strongly influenced by, sedimentation. This phenomenon has previously been called 'upbuilding,' and 'cumulization' refers to the specific case in which sedimentation causes overthickening of one or more horizons. Here the term 'aggradational pedogenesis' is used to emphasize the active roles played by both sedimentation and pedogenic processes in the upward growth of soil profiles. Soils formed by aggradational pedogenesis form part of a continuum, from sedimentation too rapid for any significant pedogenesis, to top-down pedogenesis beneath a completely stable land surface. Stratigraphic successions in which recognizable paleosols are separated by less altered sediment represent variation along this continuum over time, in response to changing rates of both sedimentation and pedogenesis.

The concept of aggradational pedogenesis implies that paleosol horizons may have a complex history, reflecting initial formation just below the ground surface followed by burial at increasing depths over time. As a simple example, consider a highly bioturbated A horizon with relatively high organic matter content and granular structure, that was buried by several tens of centimeters of loess or alluvium. With much less fresh input, the organic matter content of this horizon could be depleted by decomposition, resulting in a lighter color. Blocky structure might develop under the lower frequency of bioturbation and wetting and drying cycles at the horizon's new depth, although granular aggregates could still be preserved within the larger blocky peds. Either clay illuviation or secondary carbonate accumulation could occur within the former A horizon. In effect, the A horizon would be transformed into a B horizon through shallow and/or slow burial. This transformation can explain why many paleosols have been interpreted as lacking A horizons, in fact, aggradational pedogenesis should probably be the preferred explanation unless there is clear evidence of truncation by an erosional event. Former B horizons can also be transformed, for example, by superimposition of secondary carbonates on illuvial clay.

Research in several regions of the world supports this model of aggradational pedogenesis. A well-documented example is illustrated in [Figure 4](#). Insect burrows that formed within or just below surficial A horizons were buried at a greater depth by loess sedimentation, and then overprinted by calcium carbonate and silica cementation. This resulted in B horizons characterized by relict A horizon features, such as the insect burrows. In the central Great Plains, United States, the

Sangamon Geosol is often capped by what appears to be an unusually thick, dark-colored A horizon, with well-preserved granular structure and other evidence of soil faunal activity ([Figure 6](#)). As early as 1951, James Thorp recognized that much of this horizon is in fact a loess unit deposited on top of the existing Sangamon Geosol profile in the early stages of the last glaciation. The existing Sangamon A horizon was built upward by slow accumulation of this loess. The loess unit responsible for aggradation of the Sangamon land surface is now identified as the Gilman Canyon Formation, which accumulated mainly during MIS 3, based on radiocarbon and luminescence dating. One key piece of evidence supporting this interpretation is provided by soil stratigraphy along transects from sites near loess sources to more distal localities. Near loess source areas, the Gilman Canyon Formation is several meters thick and contains two or more buried A horizons ([Figure 6\(a\)](#)). With greater distance downwind, the loess unit thins and the A horizons merge, resulting in the appearance of a single unusually thick A horizon at the most distal sites ([Figure 6\(b\)](#)).

Micromorphological investigations can often provide crucial evidence in reconstructing the pedosedimentary history of Quaternary deposits such as loess–paleosol sequences. This approach is discussed elsewhere in this encyclopedia, but one instructive example is shown in [Figure 5\(b\)](#). The development of the S1 paleosol in the southern Loess Plateau of China is inferred to have included three stages, beginning with a relatively humid phase with slow loess deposition and clay illuviation. Two subsequent stages marked by increasingly dry conditions and more rapid loess sedimentation, during which secondary carbonate features were superimposed on illuvial clay deposited during the earlier more humid phase. Another noteworthy example is the micromorphology-based reconstruction of sedimentary events and climatic changes affecting a complex polygenetic paleosol in Iowa, United States.

The possibility of aggradational pedogenesis must be taken into account when paleosols are used as stratigraphic markers. Consider a buried paleosol that marks the boundary between the deposits of two glaciations. If only top-down pedogenesis is recognized, the upper boundary of sediment assigned to the older glaciation should be placed at the top of the paleosol profile. If, however, this profile has been built upward by loess deposition as it developed, its upper horizons may in fact contain sediment from the interglacial period or from early stages of the younger glaciation. In that case, pedogenesis may well have obscured the actual upper boundary of sediment assignable to the older glaciation. This complication is well-established in the case of the Sangamon Geosol in the central United States, and is probably not uncommon elsewhere.

Aggradational pedogenesis also has implications for the paleoenvironmental interpretation of paleosols. If the overthickening of an A horizon by loess deposition (e.g., [Figure 6](#)) is not recognized, it could be inferred that the entire underlying soil profile were formed in a cool, humid grassland environment, similar to areas where thick, dark A horizons are found today. From a broader perspective, aggradational pedogenesis can produce features in paleosols that are difficult to explain when using simple surface soil profiles as analogs. These can include overthickened A or B horizons, B horizons with multiple peaks of clay or secondary carbonate, and the occurrence

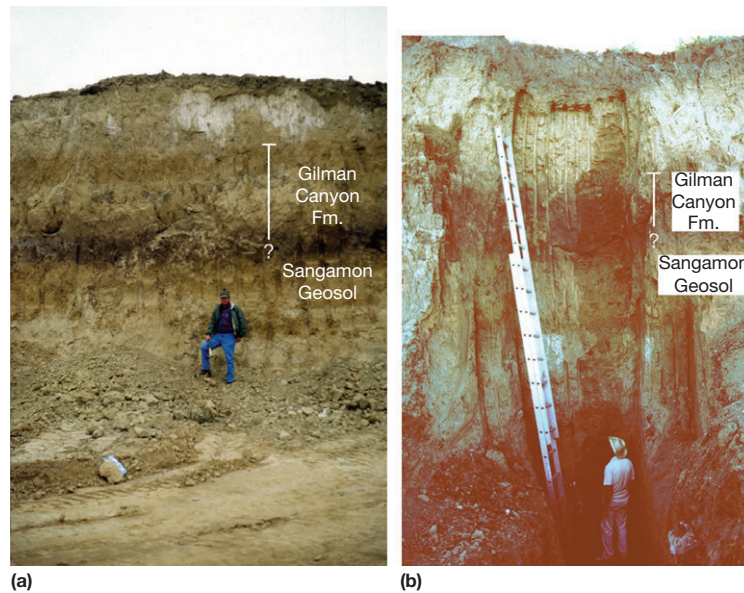


Figure 6 Sangamon Geosol and overlying Gilman Canyon Formation loess on the Great Plains. (a) Section close to loess sources, in which Gilman Canyon Formation is thicker and contains two recognizable buried A horizons, the lower of which is darker. (b) Section relatively far from loess sources, in which Gilman Canyon Formation is contained within a single overthickened A horizon at top of Sangamon Geosol. Total Gilman Canyon Formation thickness is indicated by the vertical bar in both parts, with question mark indicating uncertainty of its lower boundary.

of both secondary carbonate and illuvial clay in the same B horizon. Some surface soils contain features best explained by aggradational pedogenesis, and recent insights on that process from paleopedology may well be applicable in explaining surface soil genesis.

Transformation of A horizons into B horizons, and superimposition of secondary carbonates over illuvial clay can both be considered forms of postburial alteration in paleosols, a widely recognized but poorly understood phenomenon. Possible forms of postburial alteration include loss of organic matter, compaction, precipitation of secondary carbonates in formerly carbonate-free horizons, and formation or destruction of redoximorphic features. Some of these processes may in fact be most effective when burial is shallow, as in aggradational pedogenesis. For example, an A horizon buried in the well-oxidized shallow subsurface may lose organic matter more rapidly than one buried quickly to a much greater depth, where either reducing conditions or infrequent wetting inhibit decomposition. Alternatively, compaction generally increases with overburden thickness.

Postburial alteration can easily lead to misinterpretation of paleosols. For example, many paleosol horizons in the Chinese loess superficially resemble reddish brown B horizons, but in at least some cases have micromorphological features characteristic of bioturbated A horizons. Presumably, the color of these horizons has been altered by postburial loss of organic matter (Figure 3(b)).

Magnetic and Geochemical Proxy Data from Quaternary Paleosols

Recent decades have seen extensive efforts to extract paleoclimatic reconstructions from the rock-magnetic properties and

stable isotope geochemistry of paleosols. Pedogenesis clearly affects the rock-magnetic properties of soil horizons, although these properties are also affected by parent material provenance and sorting during transport. The most commonly measured rock-magnetic property is magnetic susceptibility, the degree of magnetization induced in a material by a magnetic field. On the Chinese Loess Plateau, magnetic susceptibility is greatly enhanced in paleosols, but interpretations remain controversial. Various studies suggest that magnetic susceptibility is related to both paleoprecipitation and duration of pedogenesis. In contrast, loess-derived paleosols in some other regions display depletion of magnetic susceptibility relative to unweathered loess. Taken together, these observations suggest that much more information on the processes that alter rock-magnetic properties during pedogenesis is required, before these properties can be considered broadly applicable, unambiguous proxies of paleoclimate.

Ratios of the stable isotopes of carbon, oxygen, and other elements can be measured in paleosol organic matter and pedogenic carbonates. The theory and methodology of reconstructing paleoenvironments from isotopic records are beyond the scope of the chapter, but the great potential of stable isotope data extracted from paleosols can be summarized as follows. Stable carbon isotopes in both soil organic matter and pedogenic carbonates often record variation in the relative abundance of plants using the C_3 , C_4 , and Crassulacean acid metabolism photosynthetic pathways, which generally represents a response to climate change; changing atmospheric C isotope ratios and moisture stress or other factors influencing photosynthesis also may be recorded. In shallow horizons, C isotopes in carbonates are sensitive to soil characteristics affecting exchange of soil carbon dioxide (CO_2) with the atmosphere. Oxygen isotope ratios in carbonates can be used to reconstruct past changes in air temperature and atmospheric

circulation. Oxygen isotope ratios in pedogenic iron oxides can provide similar information, and may be less prone to post-burial alteration, but have received little attention in Quaternary paleopedology.

Here we emphasize the need for interpreting isotopic records from soils in light of the complexities of paleosol genesis described above. For example, when isotopic analyses have been carried out on organic carbon collected from a range of depths down through a paleosol, it is important to know whether this sequence can be interpreted as a simple time series. This could clearly be the case for a soil developing through aggradational pedogenesis. It would not be a safe assumption in the case of top-down pedogenesis, even if radiocarbon ages from organic matter systematically increased with depth, because such depth trends may simply record slower carbon cycling at greater depths. Landscape-scale soil stratigraphic work, combined with high-resolution luminescence dating at selected sites, can potentially resolve this issue.

Future Directions in Quaternary Paleosol Research

It is now clear that the stratigraphic and paleoenvironmental interpretation of Quaternary paleosols is a problem of great complexity, because of aggradational pedogenesis and the dependence of soil properties on time as well as environmental factors. New cosmogenic nuclide and luminescence dating methods hold great promise as tools in resolving both of these sources of complexity, but must be integrated with the methodology and insights developed over more than a century of research on paleosols and their modern analogs.

See also: Loess Deposits: Origins and Properties. **Loess Records:** China. **Paleoceanography, Physical and Chemical Proxies:** Terrigenous Sediments. **Paleosols and Wind-Blown Sediments:** Mineral Magnetic Analysis; Soil Micromorphology; Weathering Profiles. **Plant Macrofossil Methods and Studies:** Treeline Studies.

References

- Almond PC and Tonkin PJ (1999) Pedogenesis by upbuilding in an extreme leaching environment and slow loess accretion, south Westland, New Zealand. *Geoderma* 92: 1–36.
- Balco G, Stone JOH, and Mason JA (2005) Numerical ages for Plio-Pleistocene sediment sequences by $^{26}\text{Al}/^{10}\text{Be}$ dating of quartz in buried paleosols. *Earth and Planet Science Letters* 232: 179–191.
- Begét J (2001) Continuous Late Quaternary proxy climate records from loess in Beringia. *Quaternary Science Reviews* 20: 499–507.
- Birkeland PW (1999) *Soils and Geomorphology*, 3rd edn. New York: Oxford University Press.
- Boardman J (1985) Comparison of soils in midwestern United States and Western Europe with the Interglacial Record. *Quaternary Research* 23: 62–75.
- Bronger A (1991) Argillic horizons in modern loess soils in an ustic soil moisture regime: Comparative studies in forest-steppe and steppe areas from Eastern Europe and the United States. *Advances in Soil Science* 15: 41–90.
- Bronger A and Bruhn N (1989) Relict and recent features in tropical Alfisols from south India. *Catena Supplement* 16: 107–128.
- Bronger A and Catt JA (1989) Paleosols: Problems of definition, recognition, and interpretation. *Catena Supplement* 16: 1–7.
- Bronger A and Heinkele T (1989) Micromorphology and genesis of paleosols in the Luochuan loess section, China: Pedostratigraphic and environmental implications. *Geoderma* 45: 123–143.
- Busacca AJ (1989) Long Quaternary record in eastern Washington, USA interpreted from multiple buried paleosols in loess. *Geoderma* 45: 105–122.
- Curry BB and Pavich MJ (1996) Absence of glaciation in Illinois during marine isotope stages 3 and 5. *Quaternary Research* 46: 19–26.
- Dokuchaev VV (1883) *Russian Chernozem*. Translation of 1948 reprint (Moscow), published in 1967 by Israel Program for Scientific Translations, Jerusalem.
- Follmer LR (1982) The geomorphology of the Sangamon surface: Its spatial and temporal attributes. In: Thorn CE (ed.) *Space and Time in Geomorphology*, pp. 117–146. London: Allen and Unwin.
- Follmer LR (1983) Sangamon and Wisconsinan pedogenesis in the midwestern United States. In: Porter SC and Wright HJ (eds.) *The Late Pleistocene*, pp. 138–144. Minneapolis, USA: University of Minnesota Press.
- Jacobs PM (1998) Morphology and weathering trends of the Sangamon Soil Complex in south-central Indiana in relation to paleodrainage. *Quaternary Research* 50: 221–229.
- Jacobs PM and Mason JA (2005) Impact of Holocene dust aggradation on A horizon characteristics and carbon storage in loess-derived Mollisols of the Great Plains, USA. *Geoderma* 125: 95–106.
- Kemp RA (2001) Pedogenic modification of loess; significance for palaeoclimatic reconstructions. *Earth-Science Reviews* 54: 145–156.
- Kemp RA, Derbyshire E, and Meng X (1997) Micromorphological variation of the S1 paleosol across northwest China. *Catena* 31: 77–90.
- Koch PL (1998) Isotopic reconstruction of past continental environments. *Annual Review of Earth and Planetary Sciences* 26: 573–613.
- Kukla GJ (1977) Pleistocene land-sea correlations. I. Europe. *Earth-Science Reviews* 13: 307–374.
- Leverett F (1898) The weathered zone (Sangamon) between the Iowan loess and Illinoian till sheet. *Journal of Geology* 6: 171–181.
- Liu T and Ding Z (1998) Chinese loess and the paleomonsoon. *Annual Review of Earth and Planetary Sciences* 26: 111–145.
- Maher BA and Thompson R (1995) Paleorainfall reconstructions from pedogenic magnetic susceptibility variations in the Chinese loess and Paleosols. *Quaternary Research* 44: 383–391.
- McDonald EV and Busacca AJ (1990) Interaction between aggrading geomorphic surfaces and the formation of a late Pleistocene paleosol in the Palouse Loess of eastern Washington State. In: Kneupper PLK and McFadden LD (eds.) *Soils and Landscape Evolution*, pp. 449–470. Proceedings of the 21st Annual Binghamton Symposium in Geomorphology, Amsterdam: Elsevier Geomorphology.
- Morrison RB (1967) Principles of Quaternary soil stratigraphy. In: Morrison RB and Wright HJ (eds.) *Quaternary Soils*, pp. 1–69. Reno, NV: Desert Research Institute.
- North American Commission on Stratigraphic Nomenclature (1983) North American Stratigraphic Code. *American Association of Petroleum Geologists Bulletin* 67: 841–875.
- Ruhe RV (1956) Geomorphic surfaces and the nature of soils. *Soil Science* 82: 441–455.
- Ruhe RV, Daniels RB, and Cady J (1967) *Landscape Evolution and Soil Formation in Southwestern Iowa*. Washington, DC: US Department of Agriculture. Technical Bulletin No. 1349.
- Ruhe RV and Olson CG (1980) Soil welding. *Soil Science* 130: 132–139.
- Schaezel RJ and Anderson S (2005) *Soils: Genesis and Geomorphology*. New York: Cambridge University Press.
- Sedov S, Solleiro-Rebolledo E, Gama-Castro JE, et al. (2001) Buried paleosols of the Nevado de Toluca: An alternative record of late Quaternary environmental change in central Mexico. *Journal of Quaternary Science* 16: 375–389.
- Solleiro-Rebolledo E, Sedov S, Gama-Castro J, Flores Roman D, and Escamilla-Sarabia G (2003) Paleosol sedimentary sequences of the Glacis de Buenavista, central Mexico: Interaction of Late Quaternary pedogenesis and volcanic sedimentation. *Quaternary International* 106–107: 185–201.
- Thorp J, Johnson W, and Reed EC (1951) Some post-Pliocene buried soils of the central United States. *Journal of Soil Science* 2: 1–19.
- Vidic NJ, Singer MJ, and Verosub KL (2004) Duration dependence of magnetic susceptibility enhancement in the Chinese loess-paleosols of the past 620 ky. *Palaeogeography, Palaeoclimatology, Palaeoecology* 211: 271–288.
- Woida K and Thompson ML (1993) Polygenesis of a Pleistocene paleosol in southern Iowa. *Geological Society of America Bulletin* 105: 1445–1461.
- Worthen AH (1866) *Geology of Illinois*. Champaign, IL: Illinois Geological Survey.

Mineral Magnetic Analysis

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Glossary

Antiferromagnetic: Anything that is not magnetic because of equal and opposite sources of magnetism within the material.

Chronosequence: A group of soils in which time is the predominant soil-forming factor. Parent material, climate, biota, and topography do not vary significantly among the soils of a chronosequence.

Diamagnetism: Weak magnetization due to the orbital motion of electrons.

Ferrimagnetism: A form of ferromagnetism in which the electron spin is opposite but unequal.

Ferromagnetism: A form of magnetism in which there is strong interaction among atoms in a crystal.

Frequency-dependent susceptibility: The difference in induced magnetization at two frequencies that differ by tenfold.

Loess: Wind-blown silt that is deposited and stabilized so that it becomes soil parent material.

Magnetic susceptibility: A measure of the temporary induced magnetization of a sample placed in a weak magnetic field.

Oxyhydroxides: A broad classification of minerals found in soils and sediments that are a combination of iron, aluminum, oxygen, and hydroxyls.

Paramagnetism: Magnetism arising from the spin of electrons in an atom.

Parent material: The starting material from which soil develops. It may be residuum or transported material.

Remanent magnetization: The permanent magnetization remaining after a sample is removed from a magnetic field.

Introduction

Measurement of magnetic properties of natural materials (soils, sediments, and rocks) is teaching Earth scientists about the Earth and Earth processes. The purpose of this article is to discuss what magnetism is and how magnetic measurements are used to learn about soils and sediments.

What Is Magnetism?

Magnetism is a force produced by a magnetic field. The force acts at a distance from the source of the magnetic field. For ferromagnetic materials, such as those containing certain compounds of iron, nickel, and cobalt, the interaction is particularly strong. Unlike electric charges, magnetic 'charges' always come in pairs, called poles, which have opposing charges. In a typical magnet, the magnetic force from a given pole strongly attracts the opposite pole of another magnet and repels the similar pole. The most common application of this characteristic is the use of a compass to determine direction.

The Chinese were probably the first to discover that an iron needle, when rubbed against a special rock known as a 'lode-stone' pointed in a north–south direction when freely suspended. In 1600, William Gilbert, the personal physician to Queen Elizabeth I, published the first scientific study of magnetism in a book entitled *De Magnete* (On the Magnet). In 1821, Hans Christian Oersted, a Danish scientist, discovered that an electric current in a wire produced a magnetic field. André-Marie Ampère determined that magnetism was a force between electric currents. He found that two parallel electrical currents flowing in the same direction attracted each other, and

two parallel electrical currents flowing in opposite directions repelled each other.

In space, in the Sun and in the Earth's core, electric currents produce magnetism. In the Earth, movement of the molten iron–nickel core creates a dynamo that produces the magnetic field observed at the surface of the Earth (Figure 1). To first approximation, the magnetic field at the surface of the Earth is the same as the magnetic field of a bar magnet with one pole near the geographic North Pole and the other near the geographic South Pole.

This article is primarily concerned with magnetism on a much smaller scale, namely the magnetism of certain minerals, and the use of the magnetic properties of these minerals to better understand earth processes and earth history. On this much smaller scale, elements and minerals possess three kinds of magnetism: diamagnetism, paramagnetism, and ferromagnetism. In each case, the source of the magnetism is the spin of each electron around its axis and the rotation of the electrons around the nucleus of an atom. The magnetic moment of an element is the sum of the spin magnetic moment and orbital magnetic moment.

Diamagnetism is the normal response of any material to an applied magnetic field. The response is transient, meaning the magnetization disappears when the applied field disappears. The magnitude of the response is small, and because of this diamagnetism has little value in environmental magnetism. For purely diamagnetic material, the orientation of the spin and orbital magnetic moments cancel each other, resulting in a zero magnetic moment. If the spin and orbital magnetic moments do not cancel each other entirely, the atom has a permanent magnetic moment and is said to be paramagnetic. In the presence of an applied magnetic field, a paramagnetic

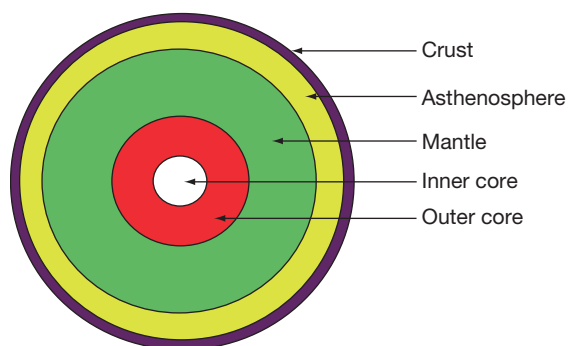


Figure 1 The model Earth. Surrounding the solid Earth is the atmosphere and hydrosphere. The thin active zone at the surface, upon which all terrestrial life depends is the pedosphere that is a few meters thick. The outer part of the solid Earth below the pedosphere is the 100 km thick, rigid crust, the lithosphere. Below the crust is the mantle. The upper portion of the mantle is the soft, plastic (deformable) asthenosphere, which varies in thickness but is generally thought to be about 200 km thick. The remainder of the 2600 km thick mantle is solid. The mantle surrounds the core, which has a radius of 3500 km, the densest part of the solid Earth. It is believed to consist of two parts, a solid inner core and an outer liquid core. It is the motion of the liquid portion of the core that produces the Earth's magnetic field.

material also acquires an induced magnetization, and the paramagnetic response is about 1000 times greater than the diamagnetic response. This is strong enough to make paramagnetic minerals of interest to Earth scientists.

In contrast to the transient magnetization of diamagnetic and paramagnetic materials, ferromagnetic materials possess a permanent or remanent magnetization. Various forms of remanent magnetization are explained later in this article. The permanent magnetization results from the interaction and alignment of electrons between different atoms in a mineral lattice. The lattice is the arrangement of atoms in a crystal. The most common natural ferromagnetic materials are oxides and oxyhydroxides of iron, cobalt, and nickel. There are different types of ferromagnetism. In ferrimagnetic materials, the aligned spins form two interpenetrating lattices with opposite alignment but different strengths. This produces a net permanent magnetization. In antiferromagnetic materials, the two lattices have the same strength so the net magnetization should be zero. However, defects in the crystal structure (lattice defects) or a slight tilting of the two lattices with respect to each other (spin canting) can lead to antiferromagnetic material with a net magnetization. Dunlop and Özdemir (1997) present many details on these concepts of magnetization.

The most important naturally occurring magnetic minerals are magnetite, maghemite, and hematite. All are iron oxides that may contain varying amounts of titanium that substitutes for iron during mineral formation. Thus, technically, we should be writing about magnetite and titanomagnetites, and about hematite and titanohematites but for simplicity, we use the simpler terms. Magnetite is found in igneous, metamorphic, and sedimentary rocks and in the unconsolidated sediments that arise from the decomposition and erosion of these rocks. Among soil scientists, magnetite, especially coarse-grained magnetite, is considered an inherited mineral that is part of soil parent material.

Hematite gives soil a red color and is a common weathering product of other iron-bearing minerals. Goethite is another weathering product that contributes brown and yellow-brown colors to soil. Both are weakly magnetic compared to magnetite and maghemite but are typically more abundant in soils than either of the more magnetic iron oxides. The red color imparted to soils from hematite implies strongly weathered parent material and is often a clue that such soils will also have strong magnetic properties due to the presence of magnetite or maghemite.

Maghemite is strongly magnetic (like magnetite) and common in soils. Unlike magnetite, it is considered a secondary iron oxide, produced in soils and sediments by chemical changes to inherited minerals. Its abundance and distribution in soil profiles is used to indicate intensity of soil formation and of weathering of parent material.

Magnetic Measurements

The simplest magnetic measurement is magnetic susceptibility, which is defined as the proportionality constant between the induced (paramagnetic) magnetization produced by a weak magnetic field and the applied field itself. The magnetic susceptibility is related to the concentration and grain size of ferromagnetic materials in a sample. The magnetic susceptibility is determined by placing a small (<10 g) sample in a weak uniform magnetic field with strength (H) and measuring the magnetization per unit volume (M) acquired by the sample. The volumetric magnetic susceptibility of the sample (κ) is the magnetization acquired per unit field (eqn [1]).

$$\kappa = M/H \quad [1]$$

In the MKS (meters, kilograms, seconds) system, volumetric susceptibility (κ) is dimensionless because both M and H have units of $A\ m^{-1}$ (ampere per meter). Much of the literature on mineral magnetic properties of earth materials uses mass susceptibility (χ), which is volume susceptibility (κ) divided by sample density (ρ ; eqn [2]). In the MKS system, mass magnetic susceptibility has units of $m^3\ kg^{-1}$.

$$\chi = \kappa/\rho \quad [2]$$

The measurement of magnetic susceptibility is a rapid, nondestructive technique that can be used in the laboratory or field with inexpensive instrumentation. The Bartington susceptibility bridge is a commonly used instrument for both laboratory and field studies.

With a Bartington bridge, the magnetic susceptibility can be measured at two frequencies that differ by one order of magnitude; these frequencies are 0.45–0.50 and 4.5–5.0 kHz. The difference in susceptibility at the two frequencies is the frequency-dependent susceptibility (χ_{fd}). Submicron size magnetic particles respond differently to different frequencies than do larger particles, making frequency-dependent susceptibility a simple (indirect) measure of the presence of these fine-grained magnetic particles. Low-field magnetic susceptibility is temperature dependent, and some laboratories are equipped with instruments that can measure κ and χ over a range of temperatures from near absolute zero to 1000 °K. Examples of these measurements and interpretations are given in the next section of this article.

As noted above, ferromagnetic materials can acquire a remanent magnetization that is retained after an external field is removed. Natural remanent magnetization (NRM), anhysteretic remanent magnetization (ARM), and isothermal remanent magnetization (IRM) are three types of magnetizations of use to Earth scientists. The name of each type is based on the conditions under which the remanence was acquired. NRM is the magnetization acquired by a sample through natural processes without application of a laboratory magnetic field. ARM is the magnetization acquired from an applied (direct) magnetic field in the presence of an external alternating magnetic field that is reduced to zero from a predetermined maximum. IRM is the magnetization acquired in the presence of a strong external field at ambient temperature. The response of a sample to a very strong magnetic field can also be studied. At the peak value of the applied field, the sample is magnetized as much as it can possibly be. At this point, the spins are said to be saturated and the magnetization is called the saturation magnetization (M_s). When the applied field is decreased to zero, the M_s decreases but does not fall to zero.

This residual permanent magnetization is the saturation remanence (M_r). Typically M_r and M_s are determined by repeated cycling of the external magnetic field, which produces a hysteresis loop (Figure 2). The direction of the arrows shows that the reverse path is not the same as the forward path as M and H change. The lack of correspondence of the two paths is termed hysteresis. The coercive force (H_c on Figure 2) is the reverse field that is needed to reduce the overall magnetization, M , to zero. The coercivity of remanence (H_{cr}) is the somewhat stronger reverse field that is needed to leave the sample with zero magnetization after the applied field has been removed. These four parameters derived from the hysteresis measurement and the shape of the hysteresis curve provide important information about magnetic grain size and magnetic mineralogy. Like magnetic susceptibility, the remanence measurements are not physically destructive. Thompson and Oldfield (1986) and Evans and

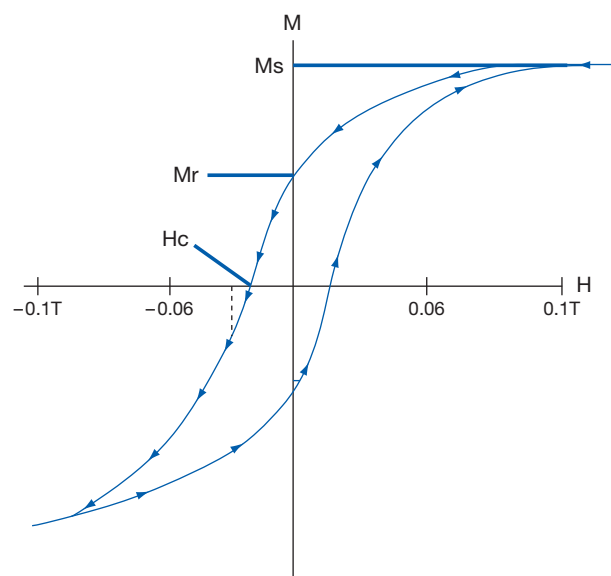


Figure 2 Example of a simple hysteresis loop.

Heller (2003) provide detailed discussions of hysteresis curves. A more sophisticated approach, based on families of partial hysteresis curves known as first-order reversal curves, is described by Roberts et al. (2000).

An important consequence of the nondestructive nature of magnetic measurements is that once these properties have been measured, samples can be subsequently treated to remove specific magnetic fractions prior to making additional magnetic measurements. For example, magnetic properties of soil samples can be measured prior to and following treatment with acids that remove carbonates, and with citrate-bicarbonate-dithionite (CBD) to remove secondary iron oxides and iron oxyhydroxides. In addition, bulk soil samples can be separated into particle size fractions, such as silt ($2\text{--}50\text{ }\mu\text{m}$) or clay ($<2\text{ }\mu\text{m}$), and the magnetic properties of the fractions can be determined.

Interpretation of Magnetic Properties

Soils

Soils are derived from geologic parent materials through biogeochemical processes that act under the influence and control of climate, biota, and topography over time. Mineral magnetic measurements are made on soils and paleosols to determine what the conditions were at the time the soil or soil horizon was formed. This is the 'inverse problem' in which a soil property can be measured and the condition under which the property was formed is inferred from the property. Most often interpretations are made from a group of properties rather than any single property. Mineral magnetic studies of soil are particularly useful when the effect of one of the five soil-forming factors (time, parent material, climate, biota, and topography) can be isolated from the other four.

Soil scientists assume that the C horizon of a soil formed in wind- or water-transported parent material is unaltered (or weakly altered) and is representative of the initial state of that material. This assumption can readily be checked by examining the mineralogy and chemistry of the C horizon. A further assumption is that differences in the amount, kind and vertical distribution of the pedogenic magnetic minerals between the C horizon and other horizons reflect the soil-forming conditions at the sample location. Similarly, soils formed *in situ* from rock have inherited and pedogenic magnetic minerals, and similar interpretations are made on these soils.

Magnetic Susceptibility

In well-drained soils, magnetic susceptibility is often higher in A horizons than in B or C horizons as shown in Figure 3.

The increase in magnetic susceptibility (magnetic enhancement) in the upper soil horizon has been attributed to the conversion of less magnetic or nonmagnetic minerals to more magnetic magnetite (Fe_3O_4) or maghemite ($\gamma\text{-Fe}_2\text{O}_3$) either through rapid oxidation by fire or by slower pedogenic processes. Magnetite and maghemite are the dominant contributors to soil magnetic susceptibility because their magnetic susceptibilities are 10 000 times greater than other commonly occurring iron minerals in soils. Magnetite is the most magnetic naturally occurring mineral. Maghemite has the same

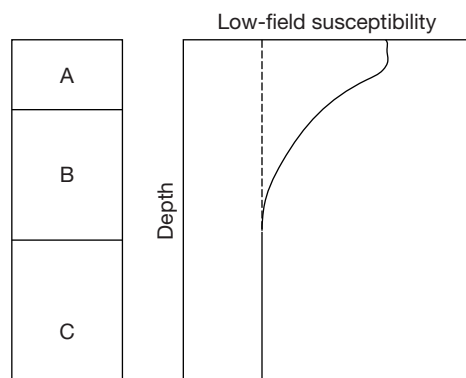


Figure 3 Hypothetical low-field magnetic susceptibility for a soil with three horizons. The total thickness of the profile is 1.5 m. Note the difference in magnetic susceptibility between the original parent material susceptibility (dashed line) and the magnetic enhancement in the A and upper B horizon (solid line).

chemical formula as hematite but is significantly more magnetic because of its crystal structure. It is distinguished from hematite by the Greek gamma before the formula ($\gamma\text{-Fe}_2\text{O}_3$). Among the other common iron minerals in soil are goethite ($\alpha\text{-FeOOH}$), ferrihydrite ($5\text{Fe}_2\text{O}_3 \cdot 9\text{H}_2\text{O}$), and lepidocrocite ($\gamma\text{-FeOOH}$). Goethite, hematite, ferrihydrite, and lepidocrocite are considered 'secondary' soil minerals because they are rarely inherited from the parent material and are usually produced by biogeochemical processes that change the inherited iron minerals into these forms. When magnetite is in large grains ($>1\text{ }\mu\text{m}$) it is typically interpreted to be an inherited mineral in soil. When it is found in small grains, it is interpreted to be a secondary iron mineral.

The minerals that contribute to the enhancement shown in **Figure 3** are usually different than the minerals of the C horizon. Typically, more maghemite is found in the magnetically enhanced horizons than in the C horizon. Of equal interest is that the thickness of the zone of enhancement increases with time and varies with climate. Among the five soil-forming factors, the two with the greatest interest for Earth scientists are climate and time. Magnetic susceptibility enhancement tends to be higher in soils that are more intensely weathered than in less intensely weathered materials. Intense weathering can result from either higher mean annual temperature (MAT) and mean annual precipitation (MAP) or a longer time period of soil formation. Magnetic susceptibility alone cannot be used to clearly separate the factors controlling the intensity of parent material weathering, but it is a useful measurement. **Figure 4** illustrates how increasing the time of soil formation affects magnetic susceptibility enhancement.

The four profiles sampled were developed on the same transported parent material, under the same present MAT and MAP and vegetation, and in the same topographic position along the north coast of California. The main difference between the profiles is the duration of parent material weathering. Such a group of soils is known as a chronosequence. All of the profiles in this chronosequence are not shown in **Figure 4**. The difference in duration of the weathering between each profile is about 100 000 years. The youngest profile (Cabrillo series) is about 100 000 years old and shows an

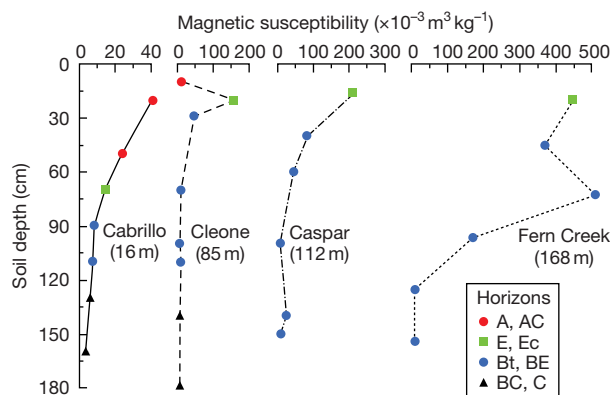


Figure 4 Vertical distribution of magnetic susceptibility (χ) in a chronosequence formed on marine terrace deposits in northern coastal California. Terraces 1, 2, 3, and 5 are represented by the Cabrillo (1), Cleone (2), Caspar (3), and Fern Creek (5) soils. Reprinted from Singer MJ and Fine P (1989) Pedogenic factors affecting magnetic susceptibility of northern California soils. *Soil Science Society of America Journal* 53: 1119–1127, with permission from the Soil Science Society of America.

enhancement of about 30 units of susceptibility compared to the parent material. The oldest profile shown (the Fern Creek series), which is five times as old, has about 480 units of susceptibility enhancement in the uppermost horizon. The enhancement also extends much deeper into the profile (**Figure 4**). These data illustrate that magnetic susceptibility within a chronosequence changes over time as a function of soil-forming processes.

The exact processes that produce the enhancement and distribution of magnetic susceptibility in soil profiles are not fully known and are subjects of current research. One common soil-forming process is dissolution of inherited minerals and formation of new 'secondary' or 'pedologic' minerals. Weakly magnetic or nonmagnetic iron-containing minerals are common inherited minerals in most soils, and although they have limited solubility, over time, these minerals dissolve, and the iron is released to the soil solution where it may form strongly magnetic secondary minerals such as maghemite. The secondary minerals may remain where they form or be transported to lower horizons. Soluble products of mineral dissolution may also be carried downward as water flows through the soil. Recombination of the dissolution products may form strongly magnetic minerals in subsoil horizons.

CBD treatment of soil samples is used to separate fine-grained from coarse-grained iron-containing minerals. The assumption is that the fine-grained magnetic minerals are formed in soils from transformations of the coarse-grained minerals. **Figure 5** illustrates the effect of CBD extraction on the magnetic properties of a series of paleosols (S-1 to S-5) from the central loess plateau in the People's Republic of China.

This vertical section represents the upper 50 m of a 130-m thick section of buried soils separated by partially weathered loess. The Chinese loess–paleosol sequence has been of great interest to Earth scientists because it is one of the best records of terrestrial climate change yet discovered. In this material, magnetic susceptibility is much higher in the paleosols than in

the loess. Furthermore, the CBD extraction removes most of the low-field susceptibility from the paleosols. This susceptibility has been attributed to fine-grained magnetic minerals although the exact composition of these minerals is still in dispute. These fine-grained iron minerals are formed in soils from inherited coarser-grained minerals. Figure 5 clearly shows that the majority of magnetic minerals in the Chinese paleosols are pedogenic. The magnetic minerals in the loess parent material are inherited and pedogenic. The small amount (compared to the paleosols) of fine-grained pedogenic magnetic minerals which exist in the loess indicate that the loess has either weathered in place after deposition or that it weathered at the source before it was transported to the present site. In our thinking, loess deposition and pedogenesis occur simultaneously at different rates. Thus, it is expected that the loess reflects periods of weaker pedogenesis than the paleosols either because of more rapid deposition of loess or a climate less conducive to soil formation.

The second curve (χ_{post}) also illustrates that the amount of coarse (lithogenic) component is not constant, implying that there are either differences in pre- or postdeposition weathering or different sources of the loess. In either case, the usefulness of CBD in differentiating between the coarse-grained and fine-grained magnetic materials is illustrated in Figure 5.

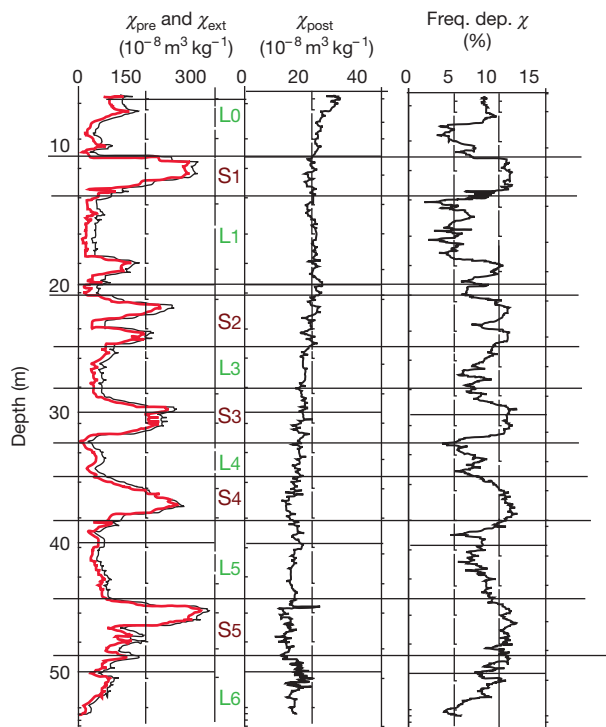


Figure 5 Total low-frequency magnetic susceptibility (χ_{pre} , black line) and CBD-extractable, (χ_{ext} , red line) and the lithogenic, coarse-textured fraction remaining (χ_{post}) when χ_{ext} is subtracted from χ_{pre} . The frequency-dependent magnetic susceptibility as a function of depth is shown on the far loess and paleosol layers are indicated by green and brown letters. Figure adapted from Vidic NJ, Singer MJ, and Verosub KL (2004) Duration dependence of magnetic susceptibility enhancement in the Chinese loess-paleosols of the past 620 ky. *Palaeogeography, Palaeoclimatology, Palaeoecology* 211: 271–288.

Frequency Dependence

Frequency-dependent susceptibility provides a way of assessing the amount of the finest-grained magnetic minerals in a sample. As noted earlier, submicron size magnetic particles respond differently to different frequencies, making frequency-dependent susceptibility a simple measure of the presence of these fine-grained magnetic particles. The presence of these fine-grained magnetic particles is interpreted to indicate soil development. In Figure 5, the frequency-dependent susceptibility varies like the total susceptibility with the highest values in the soils and lowest values in those parts of the section that are less weathered. This observation provides additional evidence that the soils have more very fine-grained magnetic minerals and that pedogenic processes form these minerals.

In persistent anaerobic environments, magnetite is produced by bacteria, where it occurs in particles that are on the order of a nanometer ($10^{-3} \mu\text{m}$). Biogenic magnetite has not been shown to occur in well-drained terrestrial environments, so it is not expected in most soils. In poorly drained soils that remain saturated or nearly saturated with water for long periods, iron is reduced to mobile Fe^{2+} from immobile Fe^{3+} and is often leached from the soil profile. Under these conditions, there is little magnetic susceptibility enhancement and magnetic mineralogy has limited usefulness in pedogenic studies (Figure 6). The Capay soil series is wetter more frequently than the Yolo and much of the iron has been reduced and removed from the profile, leaving the Capay profile with a lower magnetic susceptibility than the Yolo.

The differences between the magnetic susceptibilities of the two soils can be attributed to the different hydrology of the two

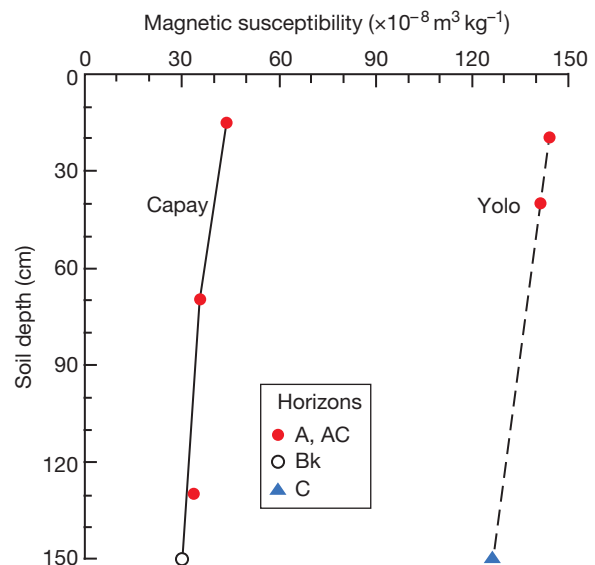


Figure 6 Vertical distribution of low-field mass magnetic susceptibility (χ) in a well-drained soil (Yolo series, dashed line) and a somewhat poorly drained soil (Capay series, solid line) formed on recent alluvium in the Sacramento Valley, CA. Reprinted from Singer MJ and Fine P (1989) Pedogenic factors affecting magnetic susceptibility of northern California soils. *Soil Science Society of America Journal* 53: 1119–1127, with permission from the Soil Science Society of America.

profiles because all other soil-forming factors are the same. The soils have formed in the same age alluvium from the same source under the same climate and biotic conditions. Only their topographic positions differentiate the soil-forming processes at the two sites. Both soils exhibit small enhancement in the uppermost horizon compared to the parent material, but the Capay profile has significantly lower low-frequency susceptibility than the Yolo profile because of persistent wetness. The loss of susceptibility, although not useful in reconstructing past climates, has been used as a method for delineating the extent of wetlands.

Natural Remanent Magnetization, Anhysteretic Remanent Magnetization, and Saturation Isothermal Remanent Magnetization

With the exception of NRM, which is the remanent magnetization of natural samples that have not been exposed to a laboratory magnetic field, all forms of remanent magnetism measure how much magnetization a sample retains after it is removed from a magnetic field. The different forms are used to obtain information about the concentration, mineralogy, and grain size of the magnetic material. Sometimes more than one magnetic property is used to try to separate the effect of grain size from the effect of concentration. For example, the ARM intensity acquired per unit of applied field is designated as the ARM susceptibility (χ_{ARM}). It is particularly sensitive to the presence of small magnetite or maghemite grains. Alternatively, mass magnetic susceptibility (χ) is particularly sensitive to the presence of larger magnetite or maghemite grains. The same is true for SIRM (saturation IRM) which is the IRM acquired in a very strong magnetic field. Thus, the ratio χ_{ARM}/χ or the ratio of the ARM intensity to the SIRM intensity provides information about the relative proportions of fine-grained to coarse-grained magnetite or maghemite. The use of these ratios and others are described in detail in [Evans and Heller \(2003\)](#) and [Maher \(1986\)](#).

Each of these techniques provides information about the minerals that contribute to the magnetic signal from a sample. Climate interpretations made from these measurements are based on our concepts of what kinds of minerals we expect to find, where they occur in soils, their particle size, and magnetic behavior. Interpretations need to be made carefully, because many factors contribute to the measured magnetic property and much is yet to be learned about mineral magnetic properties of soils and sediments.

Conclusions

Naturally occurring magnetic minerals such as magnetite, maghemite, hematite, and goethite are useful for understanding earth materials and for interpreting the climate history of the Earth. Magnetic susceptibility, frequency-dependent susceptibility, various types of remanent magnetizations, and ratios of these quantities are used by Earth scientists to determine the concentrations, compositions, and grain size distributions of naturally occurring magnetic minerals. When these magnetic measurements are applied to soils and sediment, they

provide answers to questions about soil processes and the history of the Earth.

See also: Eolian Records, Deep-Sea Sediments; Loess Deposits: Origins and Properties. **Glaciation, Causes:** Astronomical Theory of Paleoclimates. **Ice Core Methods:** Chronologies. **Loess Records:** China; North America. **Paleoclimate Reconstruction:** Approaches. **Paleosols and Wind-Blown Sediments:** Nature of Paleosols; Soil Micromorphology; Soil Morphology in Quaternary Studies; Weathering Profiles. **Quaternary Stratigraphy:** Pedostratigraphy.

References

- Birkeland PW (1999) *Soils and Geomorphology*, 3rd edn. Oxford: Oxford Press.
- Buol SM, Southard RJ, Graham RC, and McDaniel PA (2003) *Soil Genesis and Classification*, 5th edn. Iowa City: Iowa State Press.
- Cioppa MT and Kodama KP (2003) Environmental magnetic and magnetic fabric studies in Lake Waynewood, northeastern Pennsylvania, USA: Evidence for changes in watershed dynamics. *Journal of Paleolimnology* 29: 61–78.
- Dunlop DJ and Özdemir Ö (1997) *Rock Magnetism*. Cambridge: Cambridge University Press.
- Evans ME and Heller F (2003) *Environmental Magnetism. Principles and Applications of Enviromagnetics*. San Diego, CA: Academic Press.
- Eyre JK and Shaw J (1994) Magnetic enhancement of Chinese loess – The role of $\gamma\text{Fe}_2\text{O}_3$? *Geophysical Journal International* 117: 265–271.
- Fassbinder JW, Stanek EH, and Hojatollah V (1990) Occurrence of magnetic bacteria in soil. *Nature* 343: 161–163.
- Fine P and Singer MJ (1989) Contribution of ferromagnetic minerals to oxalate- and dithionite-extractable iron. *Soil Science Society of America Journal* 53: 191–196.
- Gautam P, Blaha U, and Appel E (2005) Magnetic susceptibility of dust-loaded leaves as a proxy of traffic-related heavy metal pollution in Kathmandu city, Nepal. *Atmospheric Environment* 39: 2201–2211.
- Geiss C, Banerjee SK, Camill P, and Umbanhowar CE (2004) Sediment-magnetic signature of land-use and drought as recorded in lake sediment from south-central Minnesota, USA. *Quaternary Research* 62: 117–125.
- Grimley DA, Arruda NK, and Bramstedt MW (2004) Using magnetic susceptibility to facilitate more rapid, reproducible and precise delineation of hydric soils in the midwestern USA. *Catena* 58: 183–213.
- Hamblin WK (1989) *The Earth's Dynamic Systems*, 5th edn. New York: Macmillan Publishing Company.
- Jenny H (1980) *The Soil Resource, Origin and Behavior*. New York: Springer.
- Maher BA (1986) Characterization of soils by mineral magnetic measurements. *Physics of the Earth and Planetary Interiors* 42: 76–92.
- Parfitt RL and Childs CW (1988) Estimation of forms of Fe and Al: A review, and analysis of contrasting soils by dissolution and Mossbauer methods. *Australian Journal of Soil Research* 26: 121–144.
- Roberts AP, Pike CR, and Verosub K (2000) First-order reversal curve diagrams: A new tool for characterizing the magnetic properties of natural samples. *Journal of Geophysical Research* 105: 28461–28475.
- Singer MJ and Fine P (1989) Pedogenic factors affecting magnetic susceptibility of northern California soils. *Soil Science Society of America Journal* 53: 1119–1127.
- Singer MJ, Verosub KL, Fine P, and TenPas J (1996) A conceptual model for the enhancement of magnetic susceptibility in soils. *Quaternary International* 34–36: 243–248.
- Thompson R and Oldfield F (1986) *Environmental Magnetism*. London: Allen and Unwin.
- Verosub KL and Roberts AP (1995) Environmental magnetism: Past, present, and future. *Journal of Geophysical Research* 100: 2175–2192.
- Vidic NJ, Singer MJ, and Verosub KL (2004) Duration dependence of magnetic susceptibility enhancement in the Chinese loess-paleosols of the past 620 ky. *Palaeogeography, Palaeoclimatology, Palaeoecology* 211: 271–288.
- Vlag PA, Kruiver PP, and Dekkers MJ (2004) Evaluating climate change by multivariate statistical techniques on magnetic and chemical properties of marine sediments (Azores region). *Palaeogeography, Palaeoclimatology, Palaeoecology* 212: 23–44.
- Williams RD and Cooper JR (1990) Locating soil boundaries using magnetic susceptibility. *Soil Science* 150: 889–895.

Soil Micromorphology

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Soil micromorphology is 'concerned with the description, interpretation, and measurement of components, features, and fabrics in soils' at a scale 'beyond that which can be seen readily with the naked eye' (Bullock et al., 1985, p. 9). Although it may include the examination of undisturbed soil aggregates with optical or scanning electron microscopes, it is generally restricted to the study of thin sections using polarizing microscopes. Thin section production requires initial drying and resin impregnation of an undisturbed block of unconsolidated soil material under vacuum. Once hardened, the block is sawn in half and one side polished and affixed to a glass slide. The soil slice is then ground to a thickness of approximately 25–30 µm and, unless microchemical analysis is intended, the polished surface is overlain with a coverslip. Thin section dimensions are typically on the order of 5 × 7 cm, although larger formats (e.g., 13 × 6 cm) are common particularly in archaeological studies.

Soil micromorphology came to prominence with the pioneering work of Kubiena (1938, 1970), who recognized different types of micromorphological fabrics and related them to particular suites of pedogenic processes and soil-forming environments. Although Kubiena's terminology has been superseded by more morphologically based systems (Brewer, 1976; Bullock et al., 1985; FitzPatrick, 1993; Stoops, 2003), the essence of his genetically based approach is still maintained within micromorphological contributions to pedological and paleopedological studies. Soil micromorphology is a valuable technique for the examination of buried, polygenetic, welded, and accretionary paleosols or pedocomplexes, because the recognition of individual or suites of micromorphological features, in tandem with other data, can provide the basis for relatively detailed reconstructions of formative processes and associated paleoenvironmental conditions (Kemp, 1998, 1999). The technique has also been extended to examine materials such as glacial till, lacustrine laminated sediments, and 'cultural' deposits in caves and beneath buildings or archaeological sites (Courty et al., 1989; French, 2003); however, these nonsoil (*sensu stricto*) examples are beyond the scope of this review and are not considered further here.

The aim of this article is to review the use of soil micromorphology in Quaternary paleopedology. It focuses initially on the ways that micromorphological data can aid in the identification of buried soils and also their interpretation in terms of pedogenic processes and environmental controls, notably climate and human activity. It then considers how microstratigraphic associations of features may cast light on changes in intrinsic or extrinsic environmental controls, and it concludes with a discussion of how ordered sequences of pedosedimentary processes can be inferred from micromorphological depth functions within welded and accretionary paleosols or pedocomplexes and used to reconstruct landscape evolution and associated environmental changes. In all cases, it must be realized that soil micromorphology is not a 'magical means of solving pedological problems' (Brewer, 1972, p. 83): there

may be situations in which patterns or variabilities of features are so complex or variable that they are impossible to unravel (Kemp, 1998). Furthermore, it should be regarded as part of a scale continuum of observation and should rarely, if ever, be used in isolation. Utilization of field descriptions and data from other analyses frequently resolves discrepancies, uncertainties, or difficulties regarding micromorphological interpretation.

Micromorphology and the Identification of Buried Soils

Buried soils within sedimentary sequences are normally recognized on the basis of diagnostic horizon sequences and other field criteria. However, in places where the soils are weakly developed, limited in lateral exposure, truncated, or altered by postburial oxidation, groundwater, or loading processes, identification may not be so straightforward. In such cases, the occurrence of a buried soil may be confirmed by the presence of micromorphological features assumed to be of pedogenic origin (e.g., floral or faunal channels, faunal excrements, root pseudomorphs, *in situ* weathered mineral grains, calcitic biospheroids (produced by earthworms), and illuvial clay coatings; Figure 1). It is generally safer to rely on a combination or assemblage of such features, however, because individually they are not necessarily diagnostic of a buried soil. Well-oriented clay coatings, for instance, have been reported from fluvial sediments and glacial tills, whereas diagenetic groundwater concentrations of iron oxides or calcium carbonate may have similar characteristics to precipitates formed in soils (Kemp, 1985). Furthermore, some rapidly formed pedological features, such as channels and excrements, may have limited paleopedological significance where they survive because they may only reflect short-lived vegetation cover and bioturbation at ephemeral aggrading surfaces (e.g., within a floodplain environment), whereas other more resistant features, such as sharply bounded, rounded 'foreign' aggregates, nodules, and fragmented clay coatings (Figure 1(h)), may not be *in situ*, having been eroded from soils formed elsewhere. Depth functions of micromorphological features may support a pedogenic interpretation and also help to differentiate *in situ* soils from transported soil material, provided that similar functions have been established for soils forming at present surfaces.

Paleoenvironmental Significance of Micromorphological Features Within Buried Soils

Once identified, a buried soil is generally interpreted in terms of the pedogenic processes and environments assumed to be currently responsible for that type of soil forming at a present surface. Field and bulk analytical criteria normally form the main bases, although micromorphology has also played a significant role by providing important insights into a number

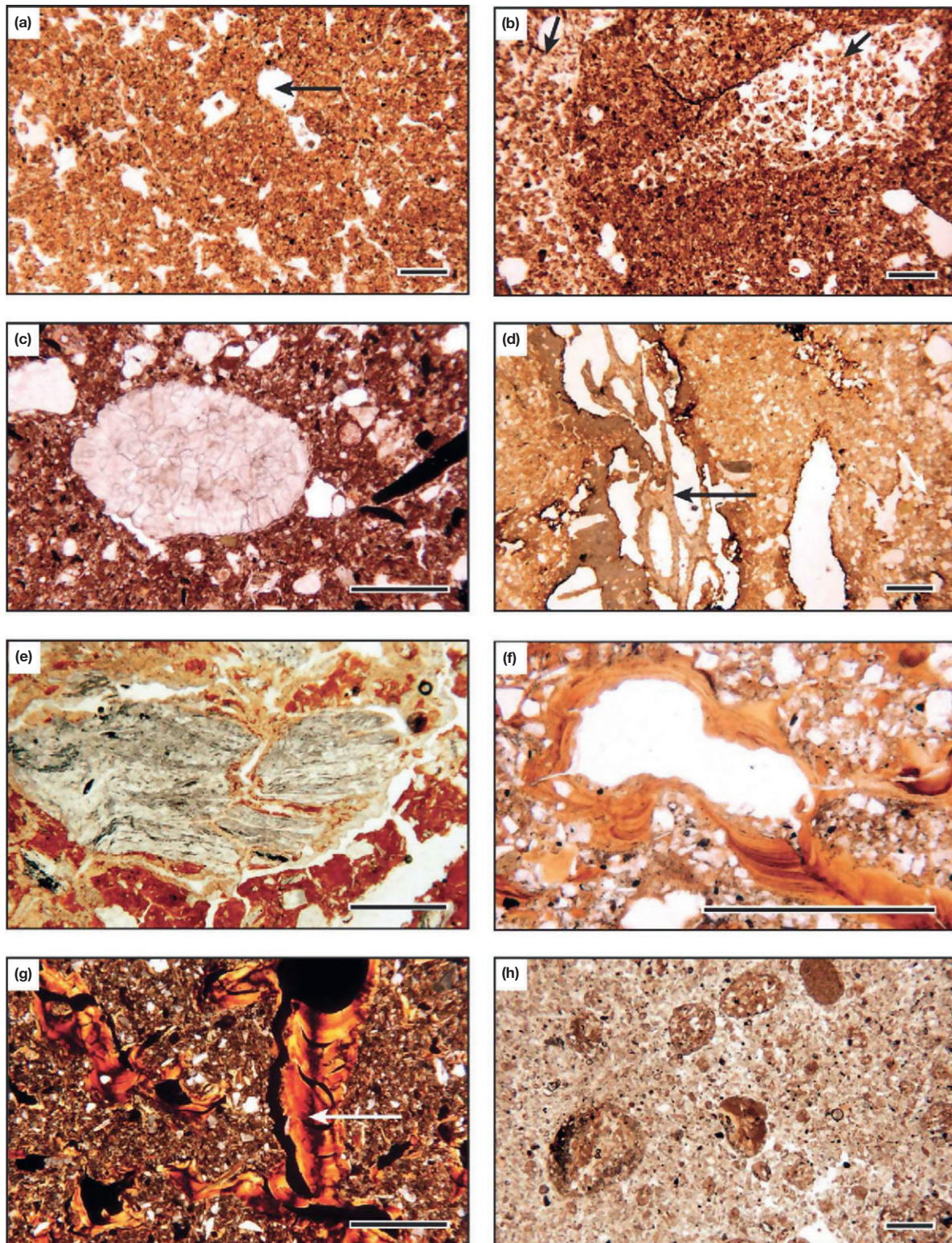


Figure 1 Micromorphological features that may help to confirm the presence of a buried soil (a–f) or identify a sediment containing reworked soil material (h). PPL, plane polarized light; XPL, crossed polarized light; scale bars = 500 μm . (a) Spongy microstructure comprising coalesced faunal excrements separated by irregular voids and more regular root channels (black arrow); PPL; Tianshui, China. (b) Loose infilling of channels by faunal excrements (black arrows); PPL; Southland, New Zealand. (c) Calcitic biospheroid produced by earthworms; PPL; Isle of Wight, UK. (d) Calcitic root pseudomorph (black arrow) within a root channel; PPL; Kent, UK. (e) Schist clast showing signs of chemical weathering to red or yellow clay along zones of weakness with the presence of fitting disaggregated fragments, which is the evidence of the *in situ* nature of the process; PPL; Almeria, Spain. (f) Microlaminated (illuvial) clay coatings around a channel; PPL; Essex, UK. (g) Microlaminated (illuvial) clay coatings showing continuous orientation with characteristic extinction lines (white arrow); XPL; Tucumán, Argentina. (h) Rounded aggregates that have been eroded from another soil and incorporated within the parent material of a new soil; PPL; Santa Fe, Argentina.

of key processes and associated environmental controls, particularly calcite redistribution, clay illuviation, and the generation of cryogenic microstructures. Such an approach is acceptable provided it is tempered by an appreciation of equifinality, namely that some pedofeature–pedogenic process–environment relations may not be unique and that different local and regional combinations of pedogenic processes and/or environmental controls may potentially produce similar micromorphological features (Brewer, 1972). Thus, it is advisable to base overall interpretations on suites of micromorphological features displaying characteristic depth functions supported by field and analytical data.

Calcite Coatings and Infillings

Some of the most common micromorphological features reported from buried soils comprise concentrations of ‘secondary’ calcite crystals within coatings, infillings, impregnative hypocoatings, and nodules (Figure 2) and reflect processes of dissolution, solute migration, and precipitation. Crystal sizes vary considerably (micrite, <10 µm; microsparite, 10–50 µm; and sparite, >50 µm) depending on conditions affecting growth, such as composition of percolating solutions, evaporation rate, and space constraints (Courty et al., 1989). Pure calcite infillings or coatings around grains and along voids often grade into adjacent zones of calcitic impregnation (hypocoatings) of the groundmass (Figure 2(c)). Some coatings, particularly associated with the surfaces of excrements, are composed of needle-fiber calcite (Figure 2(d)) that is thought to form in association with fungal activity (Becze-Déak et al., 1997). Calcitic nodules have a variety of forms ranging from dense, discrete, sharply bounded features to irregular, diffuse segregations (Figure 2(e) and 2(f)) that may even grade into complete micrite-impregnated and cemented groundmasses (Bullock et al., 1985).

The source of the calcite, direction of mobilization, and precipitation mechanism may vary considerably according to climate, parent material, level of biological activity, and groundwater conditions, so assignment of specific environmental significance to individual features is not always straightforward without subsidiary evidence (Fedoroff et al., 1990). For example, upward or lateral migration from a high groundwater table enriched in calcium bicarbonate and vertical leaching from surface to subsurface horizons provide two very contrasting mechanisms (Courty et al., 1989). Surface horizons of soils developed in calcareous parent materials under subhumid or even humid conditions may be decalcified with reprecipitation of leachates occurring lower in the sola. Similar concentrations developed in noncalcareous or weakly calcareous parent materials, however, may reflect vertical leaching processes under more semiarid environments, particularly where there have been inputs of calcareous dust (Becze-Déak et al., 1997). Small-scale changes in pH or partial CO₂ pressures associated with biological activity may also be important controlling influences (Herrero and Porta, 1987). Some compound calcitic depletion–concentration features in paleosols (Figure 2(g)), reflecting localized dissolution–reprecipitation processes (Herrero and Porta, 1987), have been attributed to relatively arid conditions (Kemp et al., 1999, 2001). In such environments, it is also common to find secondary gypsum crystals

within the groundmass (Figure 2(h)) or as individual or clusters of lenticular crystals along voids.

Iron, Manganese, and Aluminum Coatings

Other crystalline pedofeatures reported from soils include (tropical) gibbsite and (hydromorphic) goethite coatings (Stoops, 2003), although periodic anaerobic conditions more often lead to the development of ‘amorphous’ or ‘cryptocrystalline’ ferruginous and manganiferous coatings, hypocoatings, segregations, nodules, and root pseudomorphs, often with associated zones of depletion (Figure 3). Such features in paleosols are invariably taken to imply an imperfectly drained pedoenvironment, although interpretations often need to take into account the potential effects of postburial groundwater oscillations and, in the case of sharply bounded nodules, the possibility of inheritance from the parent material rather than *in situ* formation (Figure 3(d); Courty et al., 1989; Stoops, 2003). Other amorphous isotropic coatings and hypocoatings contain high proportions of organic material, particularly those formed in spodic environments (Stoops, 2003) or where agricultural and urban soils have had large quantities of slurry added.

Anthropogenic Features

Considerable emphasis has been placed by some authorities on the ability to recognize anthropogenic features in thin sections of some buried Holocene soils (Courty et al., 1989; French, 2003; MacPhail, 1998). Often, these features may simply comprise dumped components, such as fragments of pottery and building mortar (Figure 4(a)). Large concentrations of charcoal have sometimes been attributed to deliberate burning during land clearance prior to agricultural development, whereas the presence of ‘exotic’ materials such as carbonized organic matter, turf, ash, and bone (Figure 4(a) and 4(b)) has been used to support the notion that a soil was manured to increase fertility and organic matter content (Carter and Davidson, 1998). Increasing emphasis is being placed on controlled experimental trials using ancient agricultural techniques so as to understand the formation of anthropogenic soil micromorphological features and their preservation after abandonment and burial (MacPhail, 1998). Even then, concerns have been expressed regarding the uniqueness of some microfabrics ascribed to tillage or animal grazing activities (Carter and Davidson, 1998), although the problem of equifinality is not just confined to this area of micromorphology.

The most commonly reported micromorphological indicators of cultivation in buried arable soils are crusts and ‘agricutans’ (Courty et al., 1989) around voids. Such features have been shown to develop when unstable aggregates associated with cultivation at sparsely vegetated surfaces are broken down by intense rainfall. Crusts, typically moderately sorted layered units of clay grading into fine and coarse silt and then into groundmass at their base, get formed following slaking, transport of particles by overland flow, and differential settling at, or close to, the surface during water ponding. These are often subsequently disrupted and mixed by pedoturbation, leaving a random clustering of fragments embedded within the groundmass (Figure 4(c)). Linked to these features are a variety of poorly sorted coarse or silty clay coatings or infillings (agricutans;

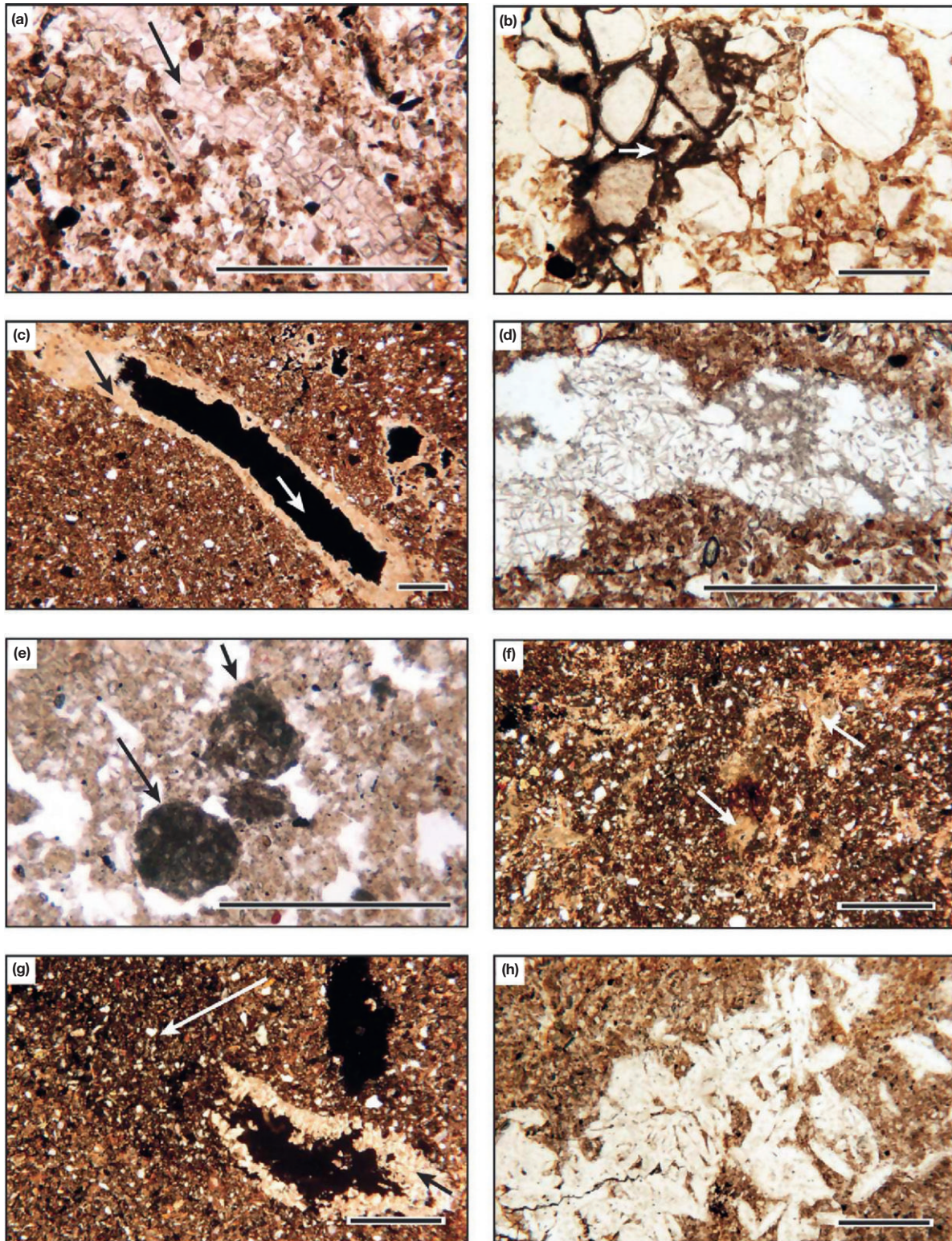


Figure 2 Crystalline pedofeatures recorded in thin sections from buried soils. PPL, plane polarized light; XPL, crossed polarized light; scale bars = 500 μm . (a) Large crystals of calcite (sparite) infilling a channel (black arrow); PPL; Siberia, Russia. (b) Small crystals of calcite (micrite) infilling packing voids (white arrow) between sand grains; PPL; Essex, UK. (c) Calcitic hypocoatings (black arrow) comprising concentrations of micrite superimposed on the groundmass adjacent to a channel (white arrow); XPL; Tucumán, Argentina. (d) Loose infilling of channel with needle-fiber calcite, which is thought to form in association with fungal activity; PPL; Siberia, Russia. (e) Micrite-impregnated nodules (black arrows) with sharp boundaries to adjacent groundmass; PPL; Buenos Aires, Argentina. (f) Segregations of micrite (white arrows) with diffuse boundaries to adjacent groundmass; XPL; Washington, USA. (g) Compound calcitic depletion-concentration feature with zone of calcite depletion (white arrow) adjacent to channel coating comprising sparite crystals (black arrow); XPL; Xining, China. (h) Clusters of secondary gypsum crystals within groundmass; PPL; Siberia, Russia.

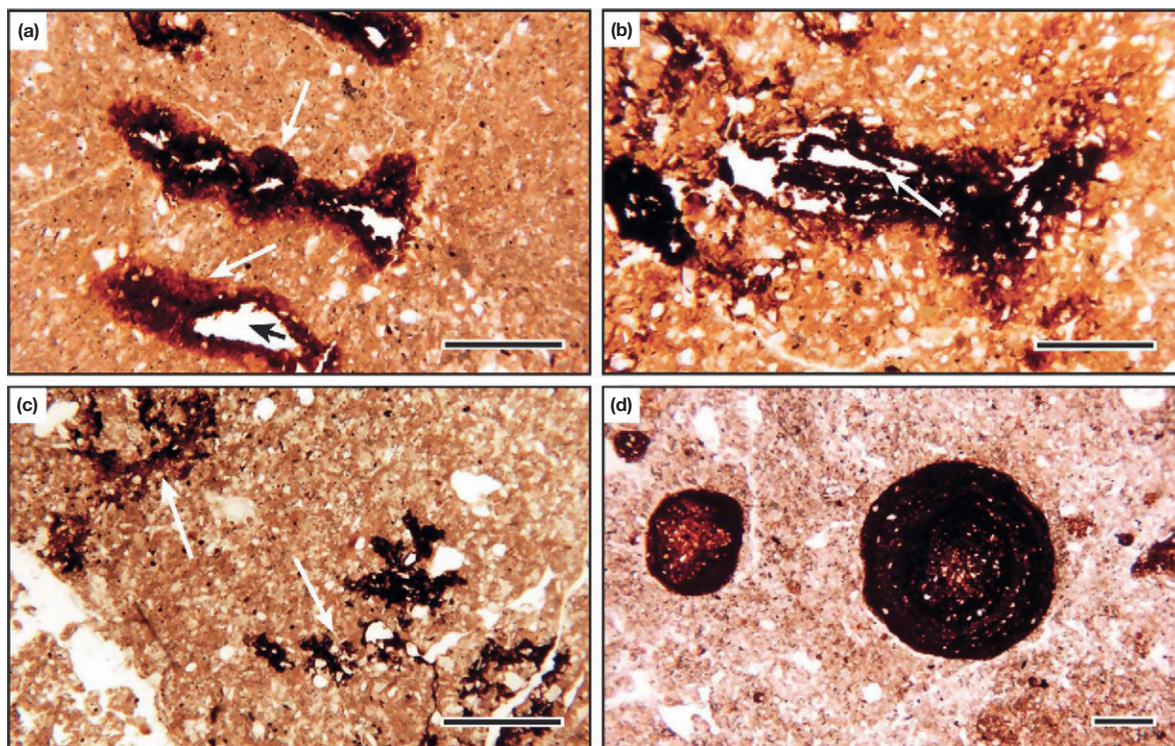


Figure 3 Amorphous pedofeatures recorded in thin sections from buried soils. Plane polarized light; scale bar = 500 μm . (a) Amorphous hypoc coatings comprising concentrations of iron/manganese oxides (white arrows) superimposed on the groundmass adjacent to channels (black arrow); Kent, UK. (b) Amorphous hypoc coating that links into an amorphous (ferrimanganiferous) pseudomorph of a degraded root (white arrow); Kent, UK. (c) Amorphous segregations comprising diffuse concentrations of iron/manganese oxides (white arrows) irregularly superimposed on the groundmass; Santa Fe, Argentina. (d) Amorphous nodules, possibly transported from elsewhere and thus inherited from the parent material, comprising high concentrations of iron/manganese oxides with very sharp boundaries to the adjacent groundmass; Idaho, USA.

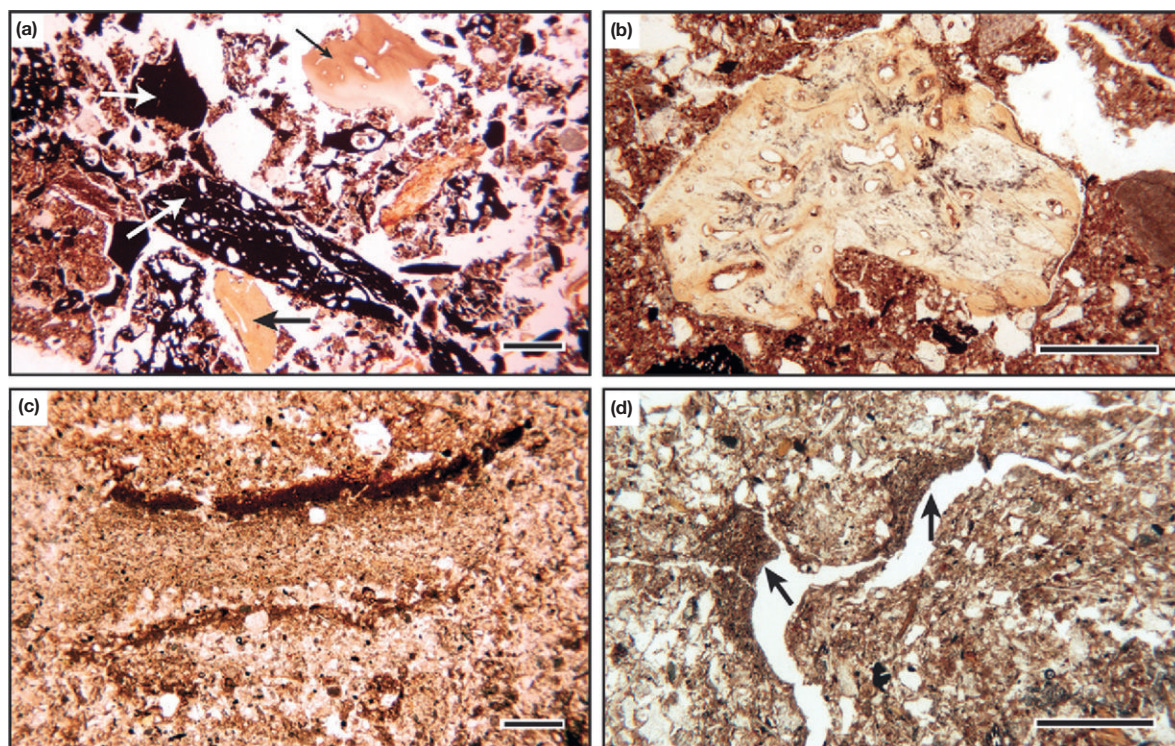


Figure 4 Anthropogenic features recorded in thin sections from buried soils. Plane polarized light; scale bars = 500 μm . (a) Clustering of charcoal (white arrow), building mortar (thick black arrow), and bone (thin black arrow); London, UK. (b) Weathered bone; central Peruvian Andes. (c) Fragment of a crust formed by surface slaking; Lanzhou, China. (d) Agricutans (black arrows) comprising coatings of silty clay around voids; Tucumán, Argentina.

Figure 4(d)), which result from rapid translocation of particles in suspension from the slaked surfaces (Courty et al., 1989).

Crusts and silty clay coatings provide classic examples of the need to base micromorphological interpretations on overall context and with reference to all other information. Although they undoubtedly can be associated with arable activity, they primarily reflect a response to slaking of exposed unstable aggregates and therefore may be naturally induced under a range of environments. For instance, the sparse vegetative cover typical of arid and semiarid zones provides soils, often having low organic matter contents, with limited protection from the effects of infrequent and intense rainstorms. Slaking and development of crusts and coarsely textured coatings are therefore common phenomena (Courty and Fedoroff, 1985).

Clay and Silt Coatings

Other silty clay and silt coatings, particularly in mid-latitude European paleosols, are often assigned cold climate significance and are regarded as having formed under relatively non-vegetated surfaces subjected to large inputs of water from snow or ice melt (Fedoroff and Goldberg, 1982). Clearly, such an interpretation is favored where there is subsidiary evidence of cryogenic conditions having existed, either in the field (e.g., ice-wedge casts or periglacial involutions) or in thin sections. Various micromorphological features, similar to ones recorded from present-day Arctic soils and reproduced experimentally in the laboratory by freezing or cycles of freezing and thawing,

have been recorded from Quaternary loess–paleosol sequences. These include banded fabrics, frost-shattered stones, platy structure, and cryoturbated granular aggregates (Figure 5), and they are generally accepted as sound indicators of cryogenic activity and thus periglacial environments (Fedoroff et al., 1990; van Vliet-Lanoë, 1998).

Micro laminated, continuously oriented clay coatings in soils (Figure 1(f)) have been traditionally considered in mid-latitudes as representing the gradual accumulation of clay translocated in suspension under (interglacial) temperate, seasonally dry climates on stable, forest-covered landscapes (Bronger, 1991). Van Vliet-Lanoë (1998), however, maintained that clay coatings in argillic horizons of northwest Europe formed under Late Glacial boreal environments, thus questioning the interglacial status attached to such units. Although Late Glacial interstadial argillic horizons (clay enriched, compared to the overlying horizon) associated with birch forest cover have been reported from southeast England, fine clay illuviation is still largely regarded as an interglacial phenomenon in Britain (Catt, 1996). The previous discussion emphasizes the dangers of assigning too narrow a paleoenvironmental significance to any micromorphological feature in isolation.

Microstratigraphic Relations Within Paleosols

In addition to highlighting individual soil features beyond the range of the human eye that are of paleoenvironmental significance, micromorphology also enables the reconstruction

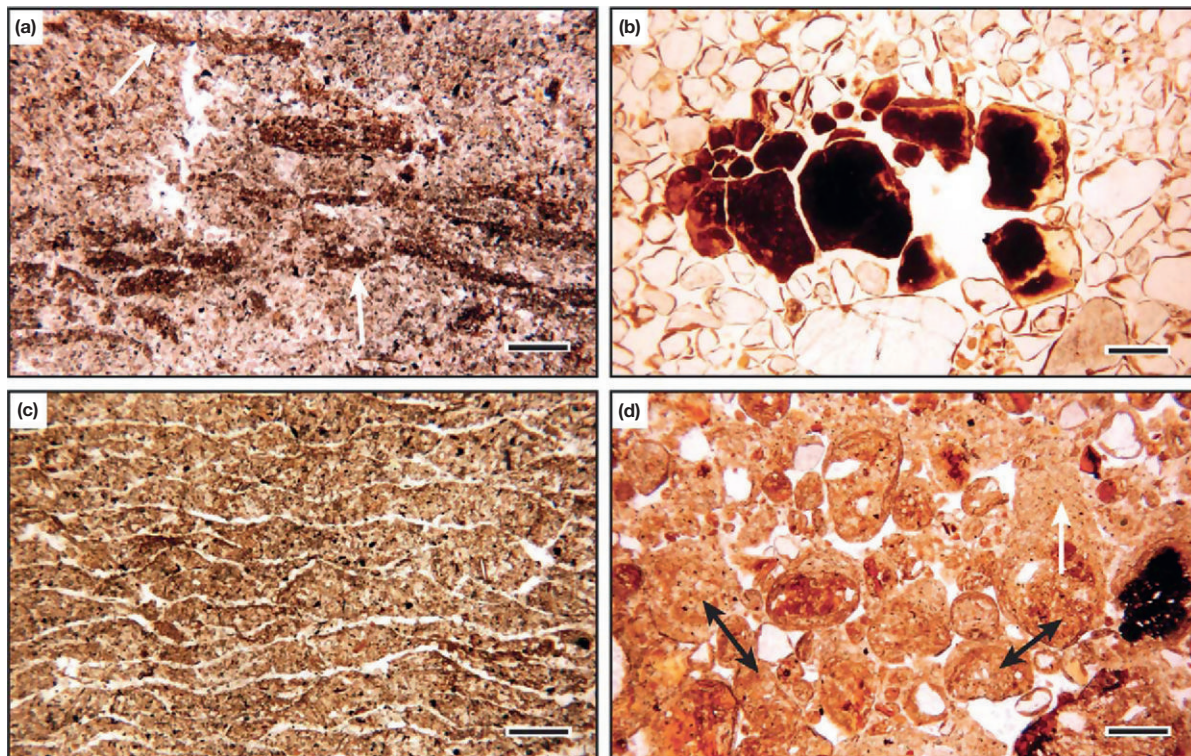


Figure 5 Cryogenic features recorded in thin sections from buried soils. Plane polarized light; scale bars = 500 μm . (a) Partially disrupted banded fabric, formed by freeze–thaw cycles, comprising alternating subhorizontal bands of silty clay (white arrows) and silt; Lanzhou, China. (b) Frost-shattered flint clast with fitting disaggregated fragments is evidence of the *in situ* nature of the process; Suffolk, UK. (c) Platy structure, formed by ice segregation parallel to the freezing front, comprising platy aggregates separated by horizontal planar voids; Siberia, Russia. (d) Granular aggregates (black arrows) created by cryogenic stresses associated with cryoturbation and solifluction (Van Vliet-Lanoë, 1998), some with cappings of silt grains (white arrow); Bavaria, Germany.

of ordered sequences of pedogenic events from the superposition of components within compound features (Brewer, 1972; Kemp, 1985). Sometimes these microstratigraphic sequences reflect intrinsic changes in the pedoenvironment as the soil develops. For instance, progressive leaching of a calcareous parent material may lead to accumulation of secondary carbonate in the form of calcitic hypocoatings around voids in subsoil horizons. As pH and base saturation drop, conditions may become conducive for clay translocation with consequential accumulation of illuvial clay on top of (and therefore

postdating) the calcitic concentrations (Figure 6(a)). Alternatively, extensive clay illuviation may result in blockage of voids, causing restricted drainage, periodic anaerobic conditions, and development of ferrimanganiferous concentration and depletion features that are superimposed on the clay coatings and infillings (Figure 6(b)).

Perhaps, of more use to Quaternary scientists are microstratigraphic sequences linked to changes in extrinsic, particularly climatic, controls. For instance, some clay coatings in soils from present-day semiarid regions have been interpreted as

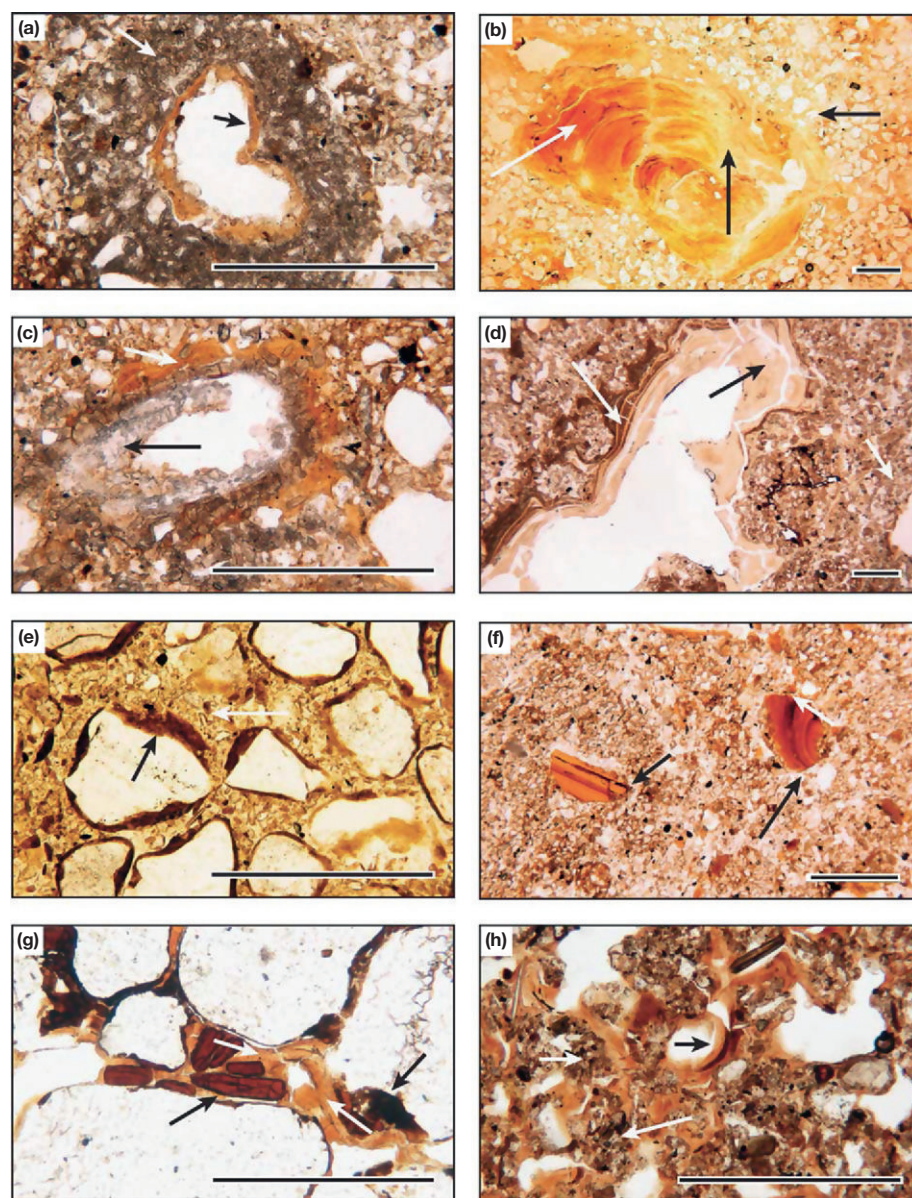


Figure 6 Microstratigraphic associations recorded in thin sections from buried soils. Plane polarized light; scale bars = 500 μ m. (a) Illuvial clay (black arrow) postdating a calcitic hypocoating (white arrow) around a channel; Kent, UK. (b) Illuvial clay (white arrow) completely infilling and blocking a channel, leading to gleying and iron/manganese oxide removal with development of gray depletion zones (black arrows) across part of the infilling and adjacent groundmass; Essex, UK. (c) Calcitic coating (black arrow) postdating a clay coating (white arrow) around a channel; Kent, UK. (d) Compound laminated coating around a channel comprising coarse clay (white arrow) followed by fine clay (black arrow); Buenos Aires, Argentina. (e) Sand grains coated by coarse clay (black arrow) with intervening packing voids subsequently infilled by silt grains (white arrow); Suffolk, UK. (f) Fragments of clay coatings (black arrows) embedded within groundmass; Normandy, France. (g) Undisturbed and fragmented red clay coatings (black arrows) engulfed within, and therefore postdated by, orange-brown clay coatings (white arrows); Essex, UK. (h) Illuvial clay (black arrow) coating (and therefore postdating) granules (white arrows) and voids within a spongy microstructure of an old A horizon; Tucumán, Argentina.

forming during a previous more humid period, and superimposed secondary carbonate features (Figure 6(c)) are taken to reflect a reduction in moisture availability and leaching, with an associated increase in calcareous dust inputs at the surface, in response to drying of climate (Nettleton et al., 1989). In certain situations, compound coatings or infillings containing progressively coarser or finer textured layers (Figure 6(d) and 6(e)) have been interpreted in terms of environmental deterioration or amelioration, respectively, at an interglacial–glacial stage boundary (Fedoroff and Goldberg, 1982; Fedoroff et al., 1990). Fragmented clay coatings (Figure 6(f)) within polygenetic or ‘paleoargillic’ buried and nonburied paleosols in temperate latitudes, particularly when associated with identifiable cryogenic features, are traditionally interpreted in terms of a period of clay illuviation during an interglacial followed by freeze–thaw disruption during a cold stage (Avery, 1985). Undisturbed clay coatings in clear microstratigraphic association with fragmented clay coatings (Figure 6(g)) have been used to imply subsequent reversion to interglacial conditions (Kemp and Faulkner, 1998). However, considerable care must be taken in assigning too strict a climatic significance to such microstratigraphies, bearing in mind uncertainties regarding the environmental factors controlling particle translocation and the possibility of other disruptive processes being responsible for fragmentation. Furthermore, when dealing with polygenetic paleosols potentially spanning several interglacial–glacial cycles, there are undoubtedly situations in which the pattern of micromorphological features is so variable and complex that it is impossible to unravel and provide a coherent and reliable reconstruction (Kemp and Faulkner, 1998).

Micromorphological Reconstructions of Pedosedimentary and Landscape Evolution

The realization of the need to take into account the changing forms and balances of pedogenic and sedimentary inputs over time in the interpretation of many pedocomplexes and alluvial, colluvial, or eolian sequences containing paleosols has led to a move away from traditional approaches that tend to encourage the examination of paleosol units as discrete static entities formed at single stable land surfaces (Kemp, 1998). From a micromorphological standpoint, this pedosedimentary approach requires consideration of depth trends in type and abundance of micromorphological features throughout the stratigraphic column, not just within field-designated paleosol horizons. Quantification of the key pedological and sedimentary features in each thin section allows the construction of micromorphological depth functions and the identification of levels or zones marking changes in balances and/or types of inferred pedogenic and sedimentary processes. Interpretation of these depth functions and any associated microstratigraphic relations of key features, independently or along with existing field and bulk analytical data, provides the basis for the reconstruction of the landscape evolution in terms of a series of pedosedimentary stages (Kemp, 1999).

Following this approach, relatively simple microstratigraphic associations such as secondary carbonate postdating clay within compound coatings (Figure 6(c)) may be interpreted to reflect overprinting of a paleosol through leaching and redistribution of calcite from soils higher up in the stratigraphic sequence. Other micromorphologically based reconstructions of welded paleosols may involve more complex sequences of sedimentation, erosion, and pedogenic phases (Figure 7). In some

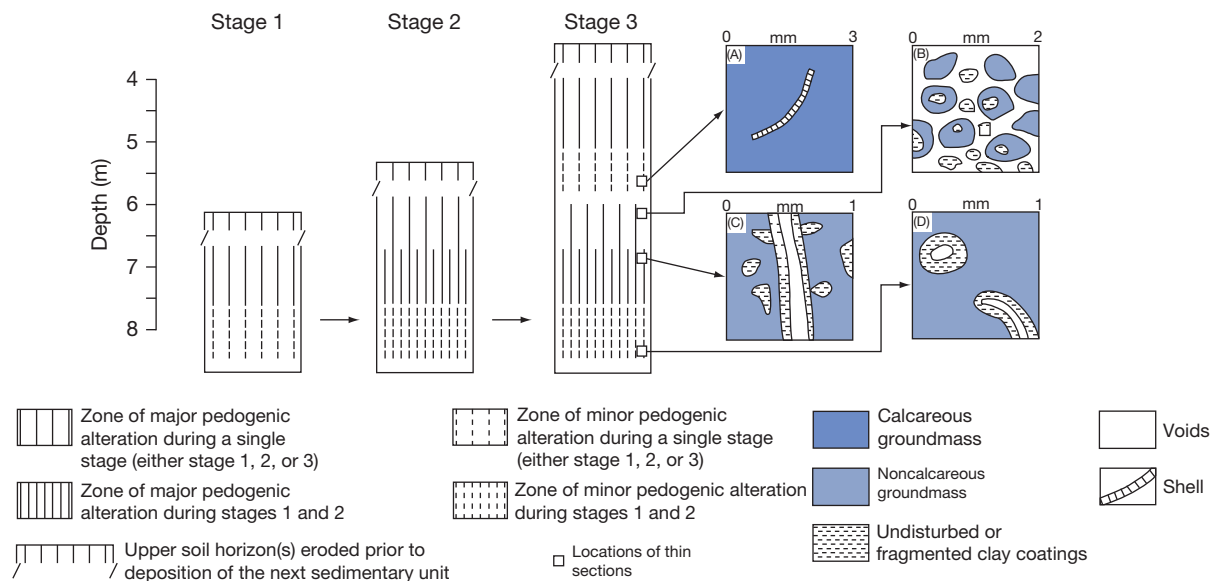


Figure 7 Key micromorphological fabrics in a thin section from a buried welded paleosol at Attenfeld (Germany) and the schematic reconstruction of its pedosedimentary development. Stage 1: deposition of solifluction sediments, soil development with accumulation of clay coatings in lower horizons (thin section D), followed by cryogenic disturbance with fragmentation of clay coatings toward the top of the soil (thin section C). Stage 2: truncation, deposition of a thin unit of solifluction sediments, soil development with accumulation of clay coatings in both the new deposits and those lower down where superimposed on the existing cryogenic fabric (thin section C), followed by cryogenic disturbance with fragmentation of clay coatings and formation of cryoturbated granular aggregates toward the top of the soil (thin section B). Stage 3: truncation, deposition of a thick unit of calcareous loess, and soil development extending through loess but not into underlying welded paleosol (as evidenced by preservation and lack of dissolution of shells in the lower part of calcareous loess; thin section A). Reprinted from Kemp RA (1999) *Micromorphology of loess–paleosol sequences: A record of paleoenvironmental change Catena* 35: 179–196. © 1999, with permission from Elsevier.

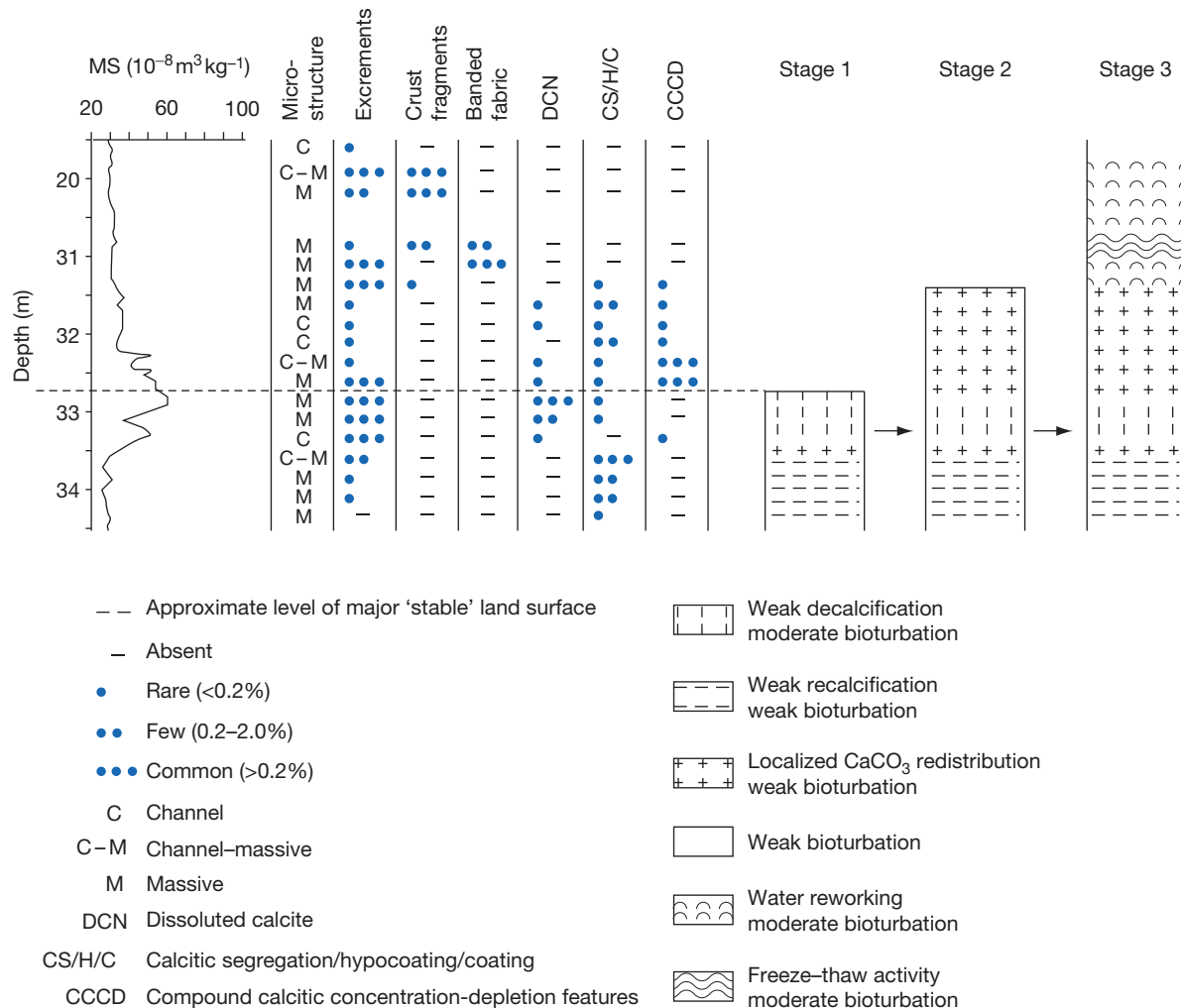


Figure 8 Depth function of the main micromorphological features and the associated reconstruction of pedosedimentary stages responsible for the development of part of a loess sequence at Gaolanshan in western China. A paleosol had been identified between 32.2 and 33.5 m on the basis of peak values of magnetic susceptibility, but pedogenic activity of environmental significance is clearly revealed in the thin sections within the loess above. Stage 1: deposition of loess, establishment of a relatively stable land surface, and development of a bioturbated weak leaching profile (zone of excrements and partially dissolved calcite overlying concentrations of secondary carbonate in the form of calcitic segregations and hypocoatings) equivalent to the lower part of the paleosol unit identified on the basis of magnetic susceptibility. Stage 2: deposition of loess with sufficient moisture available for localized redistribution of calcium carbonate as the surface accreted (compound calcitic concentration-depletion features, partially dissolved calcite and coincidental calcitic segregations and hypocoatings). Stage 3: deposition of loess under generally drier conditions (no calcitic features), with ephemeral sparsely vegetated surfaces subjected to infrequent intense rain (crusts) and with freeze-thaw processes (banded fabric) active at times. Reprinted from Kemp RA (1999) Micromorphology of loess-paleosol sequences: A record of paleoenvironmental change. *Catena* 35: 179–196. © 1999, with permission from Elsevier.

colluvial, alluvial, or loess environments, pedogenesis and sediment deposition effectively take place continuously and simultaneously within a geological time framework. Micromorphology can be very useful in identifying relict biofabrics of A horizons that have been transformed to B horizons as sediments have accreted (Figure 6(h)) and also recognizing pedogenic activity of environmental significance within field-designated sedimentary units (Figure 8). Indeed, in situations in which sedimentation rates are rapid, the pedosedimentary sequence

may be so detailed as to provide a semicontinuous record of fine-scale paleoenvironmental changes (Figure 9).

See also: **Glacial Landforms, Sediments:** Micromorphology of Glacial Sediments. **Paleosols and Wind-Blown Sediments:** Nature of Paleosols; Soil Morphology in Quaternary Studies. **Quaternary Stratigraphy:** Pedostratigraphy.

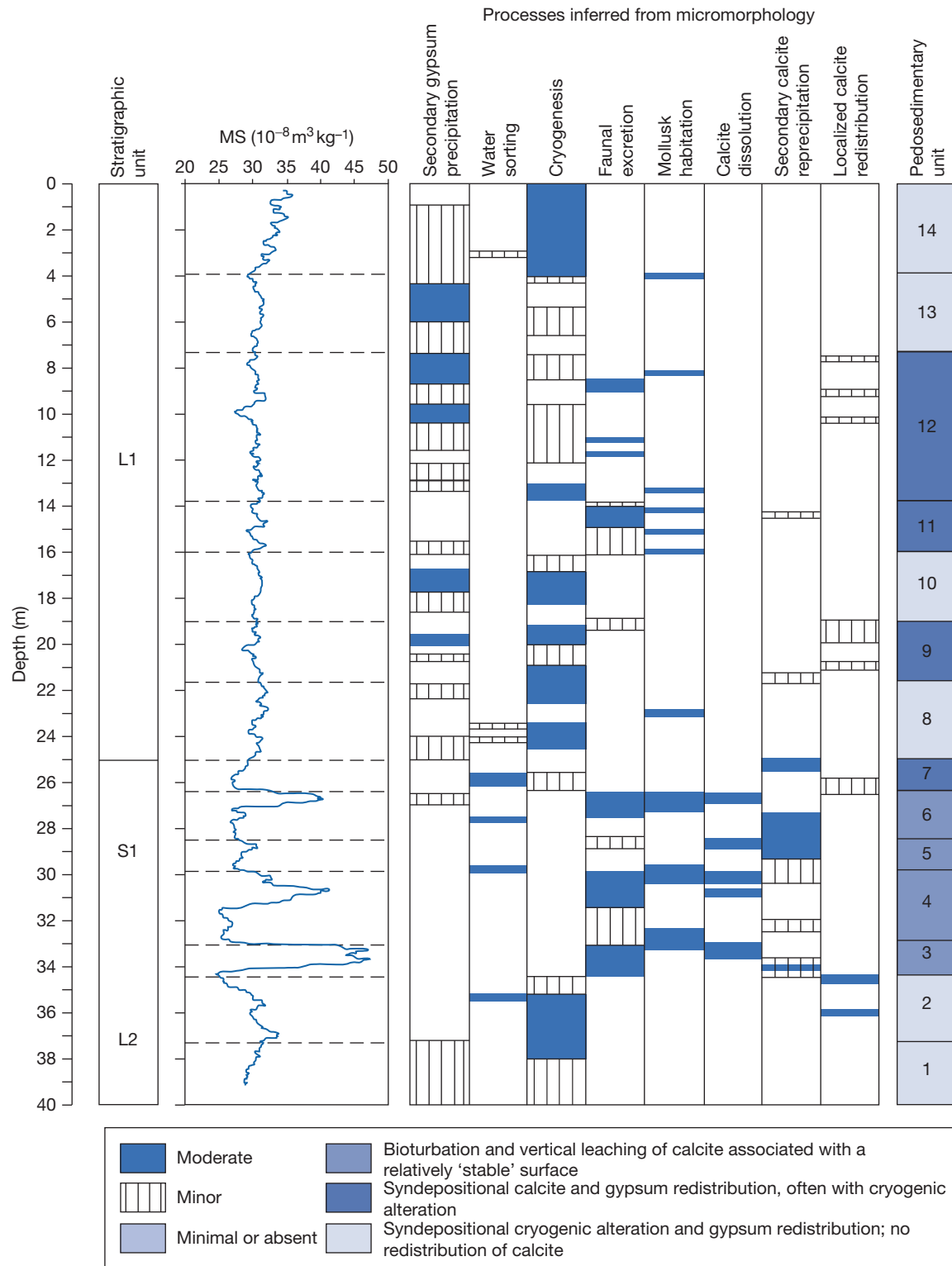


Figure 9 Pedosedimentary log based on the vertical changes in relative influences of different post- and syndepositional processes inferred from thin sections spaced at close vertical intervals through the rapidly accreting loess (L)–paleosol (S) sequence at Jiouzhoutai in western China. Each pedosedimentary unit may be interpreted to reflect specific environmental conditions (e.g., unit 14 is dominated by cryogenic features and secondary gypsum formed as loess was accumulating under a cold dry environment; unit 9 represents slightly moister conditions, supporting syndepositional bioturbation and localized redistribution of calcite; and unit 3 marks a reduction in loess accretion and establishment of a relatively stable land surface under a warmer and moister environment, supporting mollusk habitation, significant bioturbation, and vertical leaching and redistribution of carbonate). Thus, the high-resolution pedosedimentary sequence provides a relatively detailed semicontinuous record of paleoenvironmental change. Reproduced from Kemp RA, Derbyshire E, and Meng XM (1999) Comparison of proxy records of late Pleistocene climate change from a high-resolution loess–paleosol sequence in north-central China. *Journal of Quaternary Science* 14: 91–96. © John Wiley & Sons Limited.

References

- Avery BW (1985) Argillic horizons and their significance in England and Wales. In: Boardman J (ed.) *Soils and Quaternary Landscape Evolution*, pp. 69–86. New York: Wiley.
- Becze-Déak J, Langohr R, and Verrecchia EP (1997) Small scale secondary CaCO_3 accumulations in selected sections of the European loess belt. Morphological forms and potential for paleoenvironmental reconstruction. *Geoderma* 76: 221–252.
- Brewer R (1972) The basis of interpretation of soil micromorphological data. *Geoderma* 8: 81–94.
- Brewer R (1976) *Fabric and Mineral Analysis of Soils*. Huntington, New York: Krieger.
- Bronger A (1991) Argillic horizons in modern loess soils in an ustic soil moisture regime: Comparative studies in forest–steppe and steppe areas from Eastern Europe and the United States. *Advances in Soil Science* 15: 41–90.
- Bullock P, Fedoroff N, Jongerius A, Stoops G, and Tursina T (1985) *Handbook for Thin Section Description*. Wolverhampton: Waine.
- Carter SP and Davidson DA (1998) An evaluation of the contribution of soil micromorphology to the study of ancient arable agriculture. *Geoarchaeology* 13: 535–547.
- Catt JA (1996) Recent work on Quaternary paleosols in Britain. *Quaternary International* 34–36: 183–190.
- Courty M-A and Fedoroff N (1985) Micromorphology of recent and buried soils in a semi-arid region of northwestern India. *Geoderma* 35: 287–332.
- Courty M-A, Goldberg P, and MacPhail R (1989) *Soils and Micromorphology in Archaeology*. Cambridge, UK: Cambridge University Press.
- Fedoroff N, Courty M-A, and Thompson ML (1990) Micromorphological evidence of paleoenvironmental change in Pleistocene and Holocene paleosols. In: Douglas LA (ed.) *Soil Micromorphology: A Basic and Applied Science*, pp. 653–665. New York: Elsevier.
- Fedoroff N and Goldberg P (1982) Comparative micromorphology of two late Pleistocene paleosols (in the Paris Basin). *Catena* 9: 227–251.
- FitzPatrick EA (1993) *Soil Microscopy and Micromorphology*. New York: Wiley.
- French C (2003) *Geoarchaeology in Action*. London: Routledge.
- Herrero J and Porta J (1987) Gypsiferous soils in the northeast of Spain. In: Fedoroff N, Bresson L-M, and Courty M-A (eds.) *Soil Micromorphology*, pp. 187–192. Plaisir, France: Association Française pour l'Etude du Sol.
- Kemp RA (1985) Soil Micromorphology and the Quaternary. *Quaternary Research Association*. Technical Guide 2.
- Kemp RA (1998) Role of micromorphology in paleopedological research. *Quaternary International* 51–52: 133–141.
- Kemp RA (1999) Micromorphology of loess–paleosol sequences: A record of paleoenvironmental change. *Catena* 35: 179–196.
- Kemp RA, Derbyshire E, and Meng XM (1999) Comparison of proxy records of late Pleistocene climate change from a high-resolution loess–paleosol sequence in north-central China. *Journal of Quaternary Science* 14: 91–96.
- Kemp RA, Derbyshire E, and Meng XM (2001) A high resolution micromorphological record of changing landscapes and climates on the western Loess Plateau of China during Oxygen Isotope Stage 5. *Palaeogeography, Palaeoclimatology, Palaeoecology* 170: 157–169.
- Kemp RA and Faulkner JS (1998) Short-range variations in the micromorphology of a paleosol from Wivenhoe in southeast England. *Journal of Quaternary Science* 13: 233–243.
- Kubiena WL (1938) *Micropedology*. Ames, IA: Collegiate Press.
- Kubiena WL (1970) *Micromorphological Features of Soil Geography*. New Brunswick, NJ: Rutgers University Press.
- MacPhail RI (1998) A reply to Carter and Davidson's an evaluation of the contribution of soil micromorphology to the study of ancient arable agriculture. *Geoarchaeology* 13: 549–564.
- Nettleton WD, Gamble EE, Allen BL, Borst G, and Peterson FF (1989) Relict soils of subtropical regions of the United States. *Catena Supplement* 16: 59–93.
- Stoops G (2003) *Guidelines for Analysis and Description of Soil and Regolith Thin Sections*. Madison, WI: Soil Science Society of America.
- van Vliet-Lanoë B (1998) Frost and soils: Implications for paleosols, paleoclimates and stratigraphy. *Catena* 34: 157–183.

Weathering Profiles

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Introduction

Weathering is a process of conversion of rock-forming minerals toward an equilibrium state. Rocks and sediments at or near the Earth's surface are subjected to physical, chemical, and biological environments that differ from those present where constituent minerals formed, and by various processes they change toward equilibrium with Earth-surface conditions (Taylor and Eggleton, 2001). Weathering processes are grouped into two broad categories: mechanical (physical) weathering and chemical weathering. The former involves changes that do not directly result in chemical or mineralogical changes to the affected geological materials, whereas the latter result in a variety of mineralogical and chemical changes to the original materials. Mechanical and chemical weathering can and often do occur together, and both can be mediated by biota. These changes to rocks and sediments are organized into weathering profiles that reflect the interplay of weathering, erosion, and redistribution of material at and below the land surface over time. Alterations of geologic materials that take place in weathering profiles produce physical, chemical, and hydrologic characteristics that influence urban and agricultural land use, water resources, the occurrence of some economically important deposits, and landscape evolution. Weathering profiles range from centimeters to tens of meters in thickness. They begin developing upon exposure at the land surface and may form over tens of millions of years in some stable continental interiors. Fossil-weathering profiles occur throughout the geologic record, becoming more common with the spread of land plants in the Carboniferous. Because weathering profiles form from the land surface, some of their properties may reflect climatic conditions during their formation and thus provide information useful for paleoenvironmental reconstructions. Weathering profiles are the 'roots' of surface soils and reflect the cumulative effects of near-surface alterations of geologic materials over extended periods of time.

The literature on weathering profiles is replete with terms for weathered geologic material. The most all-encompassing term is regolith – the mantle of fragmented and usually un lithified geologic material that overlies bedrock and that includes residual and transported materials. Some definitions include soil as the upper zone of the regolith, whereas engineering geologists often call all unconsolidated material above bedrock 'soil.' The regolith consists of several distinct materials that may be intimately related (Ollier and Pain, 1996):

1. *Weathered rock*. Weathered rock in place is saprolite, a soft, earthy, typically clay-rich, thoroughly decomposed rock formed in place by chemical weathering of igneous, metamorphic, or sedimentary rocks (Figure 1). Structures present in the original rock, such as veins and fractures, are preserved in saprolite.

2. *Residuum*. Material originally formed by weathering of rock that has subsequently moved a short distance, either by creep or through removal of finer weathering products.
3. *Transported sediments and chemical deposits*. Materials transported by various agencies from a place of origin, where they were weathered to become available for transport, to a place where they have been deposited.
4. *Soil*. Physical, chemical, and biological alterations of geologic materials that occur immediately below a land surface and that grade downward into a weathering profile.

Weathering profiles are formed in regolith. They may be defined as the vertical extent of a weathered sequence from the initiating land surface or the originating surface to the unweathered parent rock. Weathering profiles form by chemical and mechanical weathering and vary considerably on several spatial scales because of variations in materials they are formed in, topography, hydrology, surficial processes, and climate.

Weathering Processes

Both mechanical and chemical weathering processes act to form weathering profiles. The boundary between the weathered and unweathered material in a weathering profile is called the weathering front (Mabbutt, 1961). The weathering front marks important differences in permeability, porosity, strength, and behavior, and it is the locus of initial changes in the unweathered geologic material that begin the weathering process. Advance of the weathering front occurs along and is dependent on the presence of fractures that allow the penetration of meteoric fluids into the unweathered material (Figure 2). Thus, rock types, tectonic and climatic settings, and landscape conditions that favor the formation of extensive interconnected fracture networks facilitate the development of weathering profiles (Figure 3). Some of the more important mechanical weathering processes include expansion from release of confining pressure (unloading), exfoliation, frost shattering, pressure induced by root growth, and pressure resulting from the growth of salt crystals. Water enters unweathered material on the macroscale by penetration along interconnected pores, primarily fractures in rocks, lithified sediments, and unlithified fine-grained sediments. Also of great importance is the presence of intra- and transgranular microcracks formed by a variety of forces, including crystallization, tectonics, and unloading. The microcrack systems are the loci of water-mineral interactions that are the initial stages of chemical weathering (Bisdom, 1967).

The effects of rock or sediment texture on penetration of the weathering front are fundamental (Thomas, 1974). Disaggregation of coarse-grained rocks such as granite into constituent

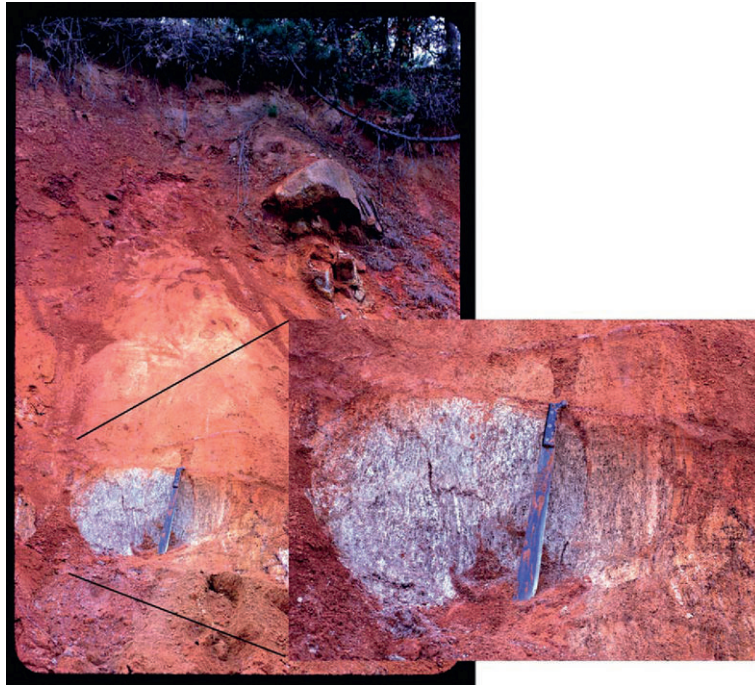


Figure 1 Clay-rich saprolite formed by in-place chemical weathering of biotite granite–gneiss. Protolith texture and structures are preserved in core stones such as the one in the lower part of this profile. Photograph by M.J. Pavich.



Figure 2 Oxidation, marked by brown and reddish brown colors, advances into gray, unweathered middle Pleistocene glacial diamicton along subvertical fractures. The fractures are avenues along which meteoric water and the solutes it carries penetrate the deposit, thereby advancing the weathering front. Photograph by E.A. Bettis.

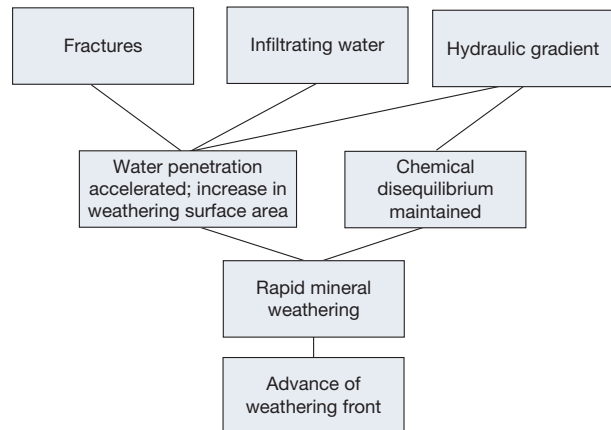


Figure 3 Flowchart showing the primary factors acting to advance the weathering front in geologic materials.

mineral grains by oxidation of biotite and hydration of other minerals often takes place before much fundamental chemical change has occurred. Fine-grained rocks of similar mineral composition weather more slowly, most likely because a system of microcracks and larger, interconnected fractures is usually not as extensively developed in these rocks. In arid, polar, and some coastal environments, crystallization from supersaturated solutions of adventitious salts (salts derived from sea spray or saline water bodies) can produce expansive stresses

leading to granular disintegration or scaling. The various physical and chemical processes acting at the weathering front act to make this a zone of relatively high hydraulic conductivity in a weathering profile. This promotes downward progression of the weathering front by transporting weathering products away from the loci of weathering (Figure 3).

In the early stages of weathering, hydration (the addition of water to a mineral) produces an increase in volume. As weathering progresses, this gives way to a volume loss as solutes are transported away from the locus of weathering by soil water and/or groundwater movement (leaching). Other chemical weathering processes, including oxidation, reduction, hydrolysis, carbonation, and chelation, act to bring about significant changes to weathering materials. In weathering, oxidation simply means a reaction of materials with oxygen to form oxides or, if water is incorporated, hydroxides. The oxygen can be derived directly from the atmosphere, from oxygen dissolved in water, or from bacterial activity. Reduction is the opposite of oxidation and usually takes place in anaerobic (oxygen-deficient), water-logged sites. Carbonation is the reaction of carbonate and bicarbonate ions with minerals. Carbonate and bicarbonate ions are common intermediate products in certain types of weathering, particularly the breakdown of feldspars. In one of the most common carbonation reactions, carbon dioxide (CO_2) reacts with water to form carbonic acid (H_2CO_3), a potent solvent in weathering environments. In the process of chelation, an ion, usually a metal, is held within an organic ring structure. Chelating agents allow the removal of ions from otherwise insoluble solids and enable the movement and precipitation of metallic ions in weathering profiles. Hydrolysis is a very effective chemical weathering process in which ions of minerals and the H^+ or OH^- ions of water react. The concentration of hydrogen ions is of fundamental importance in all weathering reactions (Ollier and Pain, 1996). In weathering profiles, the roots of living plants and acid hydrogen clay (clay with a high proportion of H^+ ions on its cation exchange sites) are primary sources of H^+ ions, which create an acid environment that promotes mineral weathering.

As a consequence of increasing efficacy of mechanical and chemical weathering toward the land surface, most weathering profiles exhibit trends of decreasing particle size, increasing clay content, and increasing voids ratio toward the surface (Gerrard, 1988). Chemical weathering of common rock-forming minerals produces only a few abundant groups of secondary minerals. Because the secondary minerals produced during weathering develop preferentially after specific primary minerals, the secondary mineralogy of most weathering profiles is dependent on the primary mineralogy and its distribution in the source rock (Nesbitt and Young, 1989). The application of weathering indices to weathering profile evaluation emphasizes a continuum of change from fresh to highly altered materials and provides numerical expression to the chemical or mineralogical changes imparted by weathering. The basis of chemical weathering indices is that some elements (e.g., Ca, Na, and K) found in weatherable minerals (e.g., feldspars, pyroxenes, amphiboles, and micas) are relatively mobile, whereas others (e.g., Al, Fe, and Ti) are not. As weathering progresses, the proportion of mobile elements remaining in the weathering materials decreases while the proportion of immobile elements increases. The chemical index of alteration

(CIA) developed by Nesbitt and Young (1982, 1989) is based on the molecular proportions of major element oxides:

$$\text{CIA} = [\text{Al}_2\text{O}_3 / (\text{Al}_2\text{O}_3 + \text{CaO} + \text{Na}_2\text{O} + \text{K}_2\text{O})] \times 100$$

The CaO in the equation is attributable only to silicates, necessitating the subtraction of nonsilicate CaO in materials in which they occur in significant quantities. Values of CIA increase from 50 to 55 for primary minerals, such as biotite and feldspar, to 100 for secondary minerals, such as kaolinite and gibbsite, with intermediate values (70–85) for smectites. Sueoka (1988) proposed the chemical weathering index (CWI), also based on molecular weight proportions in which

$$\text{CWI} = [\text{Al}_2\text{O}_3 + \text{Fe}_2\text{O}_3 + \text{TiO}_2 + \text{H}_2\text{O}(\pm) / \text{all chemical components}] \times 100\%$$

Several studies have shown progressive increase in both CWI and CIA toward the land surface in weathering profiles developed on granitic rocks. The character of weathering products within a weathering profile and among weathering profiles depends on the weathering environment because the mobility of the cations released during weathering of silicate rocks varies widely and is affected by the pH and, in the case of iron, the electron potential (E_h) of groundwater (Thomas, 1974). The most mobile elements are usually lost to the site of weathering in solution; elements with intermediate mobility remain in solution for short periods and often are precipitated as new minerals close to the site of weathering. Aluminum and iron in the ferric state, the least mobile of the common elements, generally remain as residual products of weathering, usually as sesquioxides.

Water is a prerequisite to chemical weathering because it provides and carries away reactants, thus playing a significant role in the direction and progression of weathering (Figure 3). In general, weathering progresses to deeper levels and at faster rates where groundwater gradients are steepest. Organic matter and biological processes also play critical roles in the development of weathering profiles. Plants and burrowing animals mix and cycle materials through weathering profiles. Organic acids produced during organic matter degradation form organic acids that lower pH, and they also form complexes (chelates) with metal cations that increase the solubility of important elements bringing some elements such as iron into solution in weathering profiles. Microbes, primarily bacteria, also reduce iron(III) and Mn(III) and Mn(IV) in saturated to near-saturated parts of weathering profiles to reduced ions that move through or out of the weathering profile in solution.

Oxidation, another important process in weathering profiles, occurs when geologic materials exist in an environment in which the oxygen supply is high or exceeds the biogeochemical oxygen demand (BOD). Most oxidation in weathering profiles occurs when oxygenated waters come in contact with mineral grains. Iron is the most commonly oxidized element. The oxidation of iron, from the ferrous (Fe(II)) to the ferric (Fe(III)) state, disrupts the electrical neutrality of the crystal lattice, promoting the formation of an oxide, hematite (Fe_2O_3), or a hydrous oxide such as goethite. These iron minerals impart brown, red, or yellow color to the weathered material (Figure 4).

Reduction (sometimes referred to as 'deoxidation') occurs in an environment in which the oxygen supply is limited or the



Figure 4 Yellowish brown colors in the upper, oxidized part of this exposure of loess give way to gray matrix colors and reddish brown mottles in the lower reduced (deoxidized) part of the exposure. Iron ions in the matrix of the oxidized zone are dominantly ferrous (Fe(II)), whereas in the reduced zone the matrix contains abundant ferric iron (Fe(III)). Iron oxides and hydrous oxides are abundant in the mottles where ferric iron, mobilized from the reduced matrix, moved to pores where higher Eh promoted the formation of iron oxides. Photograph by E.A. Bettis.

BOD is high. Saturation or near saturation and the presence of organic matter and microbes are prerequisites for this process. In this environment, iron is reduced to its mobile ferrous form. Once in this form, iron may be lost through net movement of the groundwater; it may remain in the sediment matrix and react with sulfides; or it may move into fractures, pores, or other voids and be oxidized. Matrix colors in a reduced zone usually range from gray to greenish gray, a condition reflecting relatively low free iron oxide content (Figure 4). In contrast to the grayish matrix, fractures and pores in reduced zones, where iron has migrated and become oxidized, have brown and reddish brown stains, streaks, and mottles.

If geologic material has not been exposed to atmospheric oxygen or oxygenated water, it may be unoxidized. In this state, most iron is in the ferrous form, and the matrix is a uniform dark gray color without stains, streaks, and mottles (Figure 5).

Weathering processes lead to the development of a weathering profile if the rate of weathering exceeds the rate of removal of weathered material by surface erosion. The upper part of the weathering profile is dominated by processes that remove soluble weathering products in solution and that lead to the concentration of more inert minerals and secondary products, such as iron oxides, that accumulate as residues. The soil profile is located at the top of this part of the weathering profile. The materials lost in solution may be removed entirely by groundwater flow or reprecipitated, either lower in the weathering profile or downslope elsewhere on the landscape.



Figure 5 Unoxidized glacial diamicton with uniform dark gray matrix color and lack of brown or red mottles, streaks, and stains. Unoxidized materials have not been exposed to the atmosphere or to oxygenated water for a period of time sufficient for oxidation of their matrix iron to begin. Photograph by T.J. Kemmis.

Weathering Zones

Weathering profiles can be subdivided into zones on the basis of physical, chemical, and mineralogical properties of the regolith. The simplest subdivision of the weathering profile includes two zones: an upper mobile or 'active' zone, which has been affected by biological and abiotic disturbance of weathered material, and a lower zone of *in situ* weathered material recognized by little or no volume change from the unweathered geologic material, the presence of joint planes, veins of resistant minerals, or other rock structures. The lower zone is usually referred to as 'saprolite.'

Walther (1916) is credited as the first to formalize the sequence of zones found from the surface downward in deep-weathering profiles:

1. soil and duricrust,
2. mottled zone,
3. pallid zone, and
4. fresh bedrock.

This sequence has subsequently come to be called a 'laterite' profile. The pallid zone has preserved rock structure, is depleted in metallic cations, and usually contains kaolinite. The mottled zone contains patches of iron oxide in a pallid zone matrix and may extend up into the active zone. Several variations on Walther's original scheme have been presented, and the reader is referred to Gerrard (1988) and Ollier and Pain (1996) for examples. Ruxton and Berry (1957) developed a physical and chemomineralogical classification scheme for weathering zones developed in granite:

1. residual debris composed of structureless sandy clay or clayey sand, 1–25 m thick with up to 30% clay, composed

- predominantly of quartz and kaolin, reddish brown when clayey and light brown or orange when less clayey;
2. residual debris with less than 10% free, rounded corestones (spheroidal boulders of fresh rock), less than 5% clay but common clay-forming minerals, generally light color;
 3. similar to (2) but with much of the original rock structure preserved and 10–50% corestones;
 4. rectangular and locked corestones (50–90% solid rock) set in a matrix of residual debris; and
 5. partially weathered rock with minor amounts of residual debris along major fracture planes, more than 90% solid rock, significant iron staining and biotite decomposition may be present.

Figure 6 illustrates some of the more commonly used classification schemes for weathering profiles in crystalline rocks.

The specific features of weathering profiles and the zones they comprise vary, depending on the physical and chemical nature of the materials in which they form, as well as across landscapes. The weathering zone schemes of [Walther \(1916\)](#) and [Ruxton and Berry \(1957\)](#) are characteristic of those formed in crystalline rocks. Weathering profiles in sedimentary rocks seldom exhibit the distinct zoning, corestones, and well-defined basal surface of weathering associated with crystalline rocks. Relatively pure carbonates such as chalks typically have very thin weathering profiles. The residuum of carbonate rock weathering profiles contains large amounts of insoluble rock

constituents, such as quartz, chert, iron, and manganese oxides and clay minerals. Large fractures, bedding planes, and solutional openings in carbonates foster extreme lateral variability of these weathering profiles (**Figure 7**). Weathering profiles formed in siltstones and mudstones are characterized by a



Figure 7 Weathering profile formed in limestone. Note variations in the weathering profile associated with the occurrence of fractures and bedding planes in the rock. Also note the presence of residual chert clasts in the upper part of the profile. Photograph by E.A. Bettis.

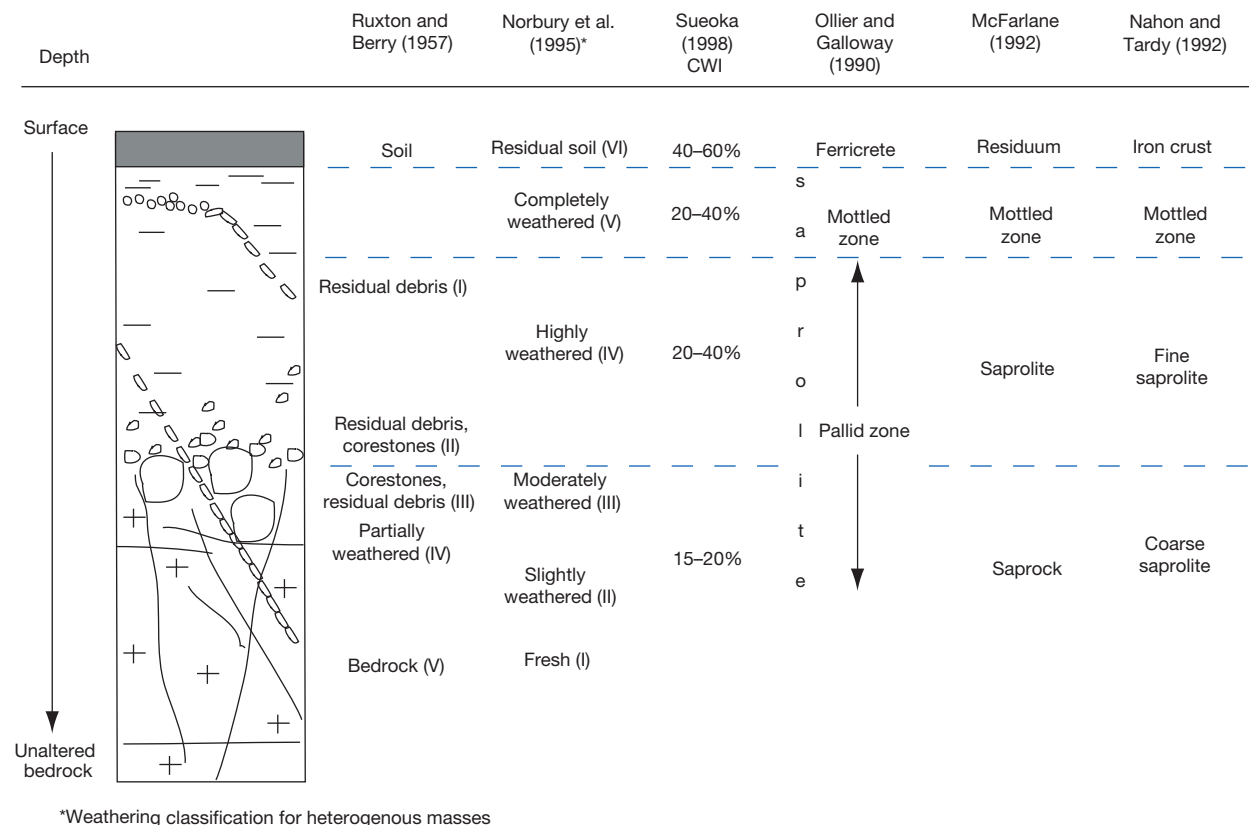
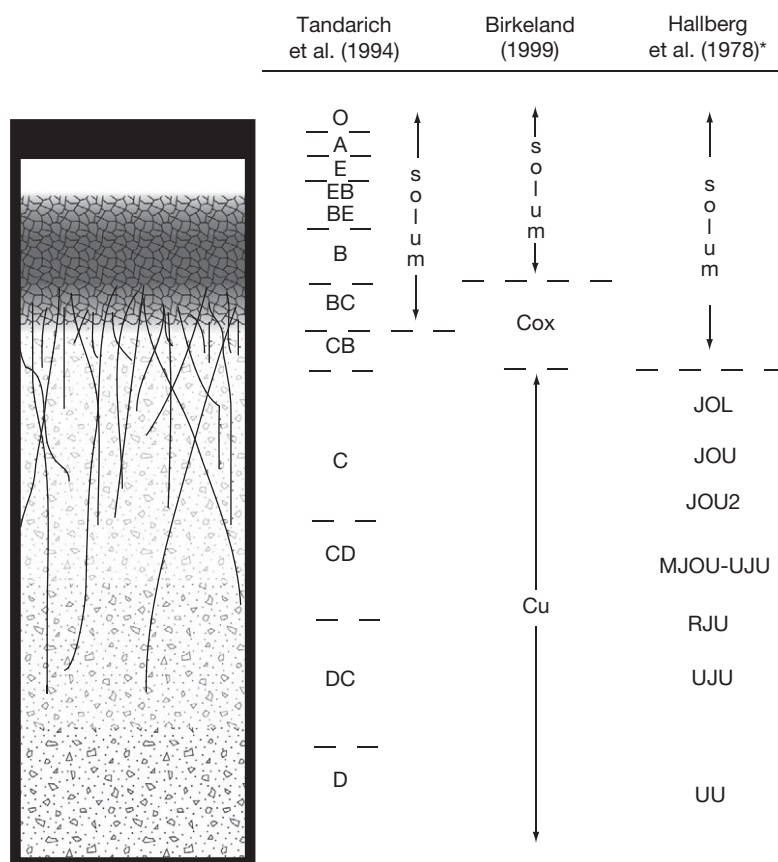


Figure 6 Six weathering zone classifications in common use for weathering profiles formed in crystalline rock. Modified from Taylor RG and Howard KWF (1999) Lithological evidence for the evolution of weathered mantles in Uganda by tectonically controlled cycles of deep weathering and stripping. *Catena* 35: 65–94; Figure 4.



*All possible weathering zones or weathering zone combinations not indicated

Figure 8 Weathering zone schemes commonly applied to unlithified deposits in the United States.

relatively thin low-strength and low-permeability zone near the surface that grades downward into a jointed and fissured zone of higher permeability (Deere and Patton, 1971). Fissures are extremely important features in the weathering of shale, siltstone, and mudstone. They often develop along bedding planes, from structural modifications, and as a response to unloading and land surface processes.

Weathering profiles in unlithified sediments have generally received less attention than those formed in rock. Glacial sediments, loess, eolian sand, alluvium, and exposed marine sediments are widespread in many areas of the world. Weathering zones in these materials are typically identified by color changes, mottling patterns, the presence/absence of fractures, and sometimes chemical changes such as carbonate leaching or accumulation. Three weathering zone schemes for unlithified sediments are in general use (Figure 8):

1. Birkeland (1999) used the designation 'Cu' for unoxidized material and 'Cox' for oxidized material.
2. Hallberg et al. (1978) used a more elaborate scheme that can be applied to materials whose unaltered state is oxidized or unoxidized (Table 1).
3. Tandarich et al. (1994) provided a classification that unifies unaltered deposits and the overlying soil profile into a single 'pedo-weathering profile.'

A conceptual problem common to all these classification schemes is that they imply a genesis beyond their simple description – an unfortunate circumstance given the great uncertainty surrounding the mechanisms by which weathering mantles develop. Taylor and Howard (1999) proposed the use of strictly descriptive graphic logs, similar to those employed to describe lithostratigraphy, as a way to objectively describe weathering profiles. Practical problems with applying weathering zone schemes to unlithified sediments usually center around two issues: (1) determination of the physical and chemical properties of the initial material that has undergone weathering and (2) complications with using color as a proxy for oxidation state in materials that contain significant amounts of detrital carbon.

Spatial Patterns of Weathering Profiles

Much discussion has focused on the distribution of thick weathering profiles, their age, and conditions fostering their development. Common assumptions are that deep weathering is associated with warm, wet climates and that the development of thick weathering profiles implies great age of the associated geomorphic surface. Although greater amounts of

Table 1 Weathering zone terminology proposed by Hallberg et al. (1978) for Quaternary materials

<i>First symbol</i>	<i>Munsell color reference</i>
O (oxidized)	60% of matrix with hues redder than 2.5Y; hues of 2.5Y with values of 5 or higher; may have segregation of secondary Fe and Mn compounds into mottles, nodules, etc.
R (reduced)	60% of matrix with hues of 2.5Y with values of 3 or less; hues of 2.5Y with value of 4 and chroma of 2 or less; hues of 5Y, N, 5GY, 5G, 5BG, 10Y, and 10GY with values of 4 or higher; colors are usually mixed as mottles, bands, or diffuse blends of colors; may have considerable segregation of secondary Fe and Mn compounds into mottles, nodules, etc.
D (deoxidized – applied only to loess)	60% of matrix with hues of 10YR, 2.5Y, and/or 5Y, values of 5 and 6, and chromas of 1 or 2; segregation of Fe compounds into tubules and/or nodules
U (unoxidized)	Uniform matrix color; hues of 5Y and N with values of 5 or less; 5GY, 5G, 5BG, 10Y, and 10GY with values of 6 or less; no secondary segregation of Fe and Mn compounds
<i>Second symbol</i>	<i>Fractures (if present)</i>
J (jointed – fractured)	Well-defined vertical to subvertical fractures present; often identified by colors that contrast with the matrix and coatings and/or rinds of secondary minerals
<i>Second or third symbol</i>	<i>Leached or unleached state – used only with materials having primary matrix carbonate</i>
U (unleached)	Reacts with dilute HCl
L (leached)	Does not react with dilute HCl
L2	Leached of primary carbonates but secondary carbonates present as coatings, nodules, or other accumulations
<i>Modifier symbol</i>	<i>When used precedes first symbol</i>
M (mottled)	Has zones with 20–50% contrasting mottles; when used with the unoxidized zone it infers 20% or less mottles of reduced colors
<i>Examples of descriptions using weathering zone terminology</i>	
OL	Oxidized and leached; yellowish brown (10YR5/6) matrix color, does not react with weak HCl
MOU	Mottled, oxidized, and unleached; strong brown (7.5YR4/4) matrix color with common gray (5Y5/1) and light olive brown (2.5Y5/4) mottles, reacts with weak HCl
UJU	Unoxidized, jointed, and unleached; uniform dark greenish gray (5GY4/1) matrix with few thin vertical fractures which have mottled light olive brown (2.5Y5/6) and olive gray (5Y5/2) faces, reacts with weak HCl

water passing through a material and higher temperatures will speed chemical weathering, the depth of weathering is a function of the rate of weathering profile advance and the rate of erosion. Thus, weathering profile thickness is related to rock/sediment properties, environmental history, and geomorphic history. Since the effects of rock and sediment properties have already been discussed, this section focuses on the role of

environmental and geomorphic history on weathering profiles. Strakhov (1967) proposed five global-scale zones of weathering related to latitudinal variations in climate and vegetation (Figure 9). He also recognized three major disturbances to the latitudinal pattern: glacial sedimentation, arid sedimentation, and active tectonics. Weathering profiles on intrusive igneous rocks, such as granite, have been described and studied in many areas of the world and have provided a 'standard' for comparing global patterns. The development of thick weathering profiles has traditionally been associated with humid tropical and subtropical environments, and their occurrence in areas with other climatic regimes has been cited as evidence of relict weathering under past wetter and/or warmer climates.

With several caveats, the assemblage of clay minerals formed in weathering profiles in granitic rocks generally reflects the climatic regime or annual rainfall average of the area (Gerrard, 1994). In arid regions, where leaching is not intense, smectite is the dominant alteration product. Illite (mica) is probably a poor indicator of macroclimatic conditions because it rarely forms in weathering environments. Kaolinite can be produced under a variety of climatic conditions. It can dominate very wet temperate regions, and it is the most commonly reported secondary clay mineral in wet tropical environments. In addition to kaolinite, smectite can occur in drier tropical regions, and gibbsite and boehmite can occur in the wettest regions. Seasonality of precipitation is also important, and weathering profiles in areas with monsoons often contain illite and montmorillonite.

Other characteristics of weathering profiles, such as particle-size distribution, have also been suggested as indicators of climate-driven weathering. Two end member types of profiles formed on granitic rocks, 'arenaceous' or sandy profiles and 'argillaceous' or clayey, have been defined. Sandy regoliths are widely distributed in temperate regions, contain common to abundant weatherable minerals, and form primarily by the alteration of plagioclase feldspars to clay, accompanied by mechanical fracture of quartz and orthoclase to sand-size material. In contrast, argillaceous regoliths, most commonly found in wet tropical and subtropical regions, contain few weatherable minerals and have high clay and low silt content due to the rapid transformation of all feldspars directly to clay minerals.

Weathering profile characteristics also vary across landscapes. Although the primary control on weathering profile thickness appears to be the intensity and direction of jointing in rocks, landscape-related patterns are also apparent. Thomas (1974) summarized evidence relating deep-weathering profiles to topography:

- The weathering front is usually highly irregular and usually exhibits a pattern of discrete basins and domes.
- Weathering patterns are aligned with the fracture system developed in the rock.
- Weathering depth is usually shallow beneath streams.
- Where marked inequalities in the resistance of rock do not occur, the depth of weathering usually increases irregularly away from streams toward interfluvies in the humid tropics.
- In drier savannas and semiarid regions, deep-weathering profiles are usually restricted to depressions in the landscape.

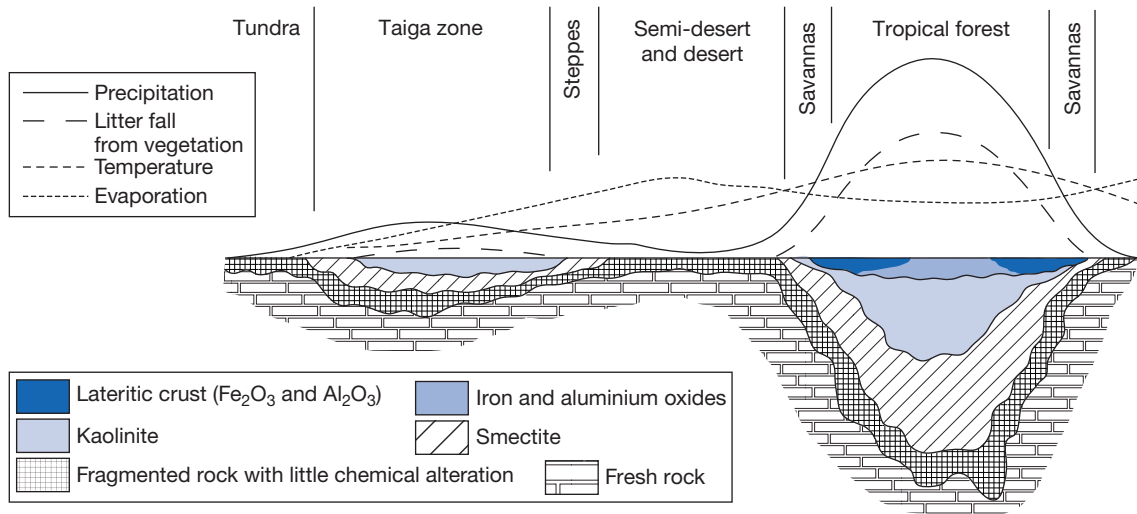


Figure 9 Pole-to-equator variation of weathering profiles as proposed by Strakhov (1967). Adapted from Figure 9.10 in Schaetzl RJ and Anderson S (2005) *Soils: Genesis and Geomorphology*. New York: Cambridge.



Figure 10 Weathering profile developed beneath a slope formed in late Pleistocene glacial diamicton. The oxidized zone is thick beneath the summit, thins beneath the sideslope, then thickens in the lower part of the slope where sandy zones promote deep penetration of oxygenated water. Weathering zone terminology of Hallberg et al. (1978): OL, oxidized and leached; JOU, jointed oxidized and unleached; UU, unoxidized and unleached; MJOU, mottled jointed oxidized and unleached. Photograph by T.J. Kemmis.

- High relief and steep slopes foster a pattern of rocky hills and weathered valley floors in both humid and arid areas.
- Patterns of surface relief may offer few clues to deep-weathering patterns.

Weathering profiles formed in Quaternary sediments that have been dated provide insights into spatial patterns of weathering profiles during the early stages of their development (Figure 10). General trends include:

- advance of the weathering front along fractures and other pores in the sediment matrix;
- increasing thickness of weathering profiles with increasing coarseness of the sediment matrix;
- deeper penetration of the weathering front on uplands near valley margins than beneath either uplands distant from

valleys or beneath valley floors except in arid and semiarid regions, where the thickest weathering profiles are found in valleys;

- weathering profiles that are thicker in well-drained than in poorly drained landscape positions, again with the exception of arid and semiarid regions;
- increasing thickness of weathering profiles in well-drained landscape positions with increasing mean annual infiltration;
- correspondence between oxidation state of dominant ferromanganese oxides (and therefore sediment matrix color) and the position of the water table;
- great complexity in the details related primarily to variations in sediment properties (bedding, structure, porosity, density, and mineralogy), water movement through the sediment, biotic effects, and the nature of surface erosion.

Dating Weathering Profiles

There are many problems surrounding generalizations about the relationship among weathering profile thickness and morphology and the age and stability of geomorphic surfaces; however, as Thomas (1974) suggested, deep lateritic profiles with thick pallid zones, as well as distinctly zoned weathering profiles, are probably reasonable indicators of surface stability and prolonged development. Deep-weathering profiles are not the product of a single weathering event in a well-defined time in the past but are the combined result of long-lasting superimposed weathering processes during extended periods of time. Changes in climate, vegetation, and landscapes over the length of time during which weathering has taken place complicate the interpretation of weathering profile histories. Most deep-weathering profiles occur in relatively stable cratonic regions that were not directly affected by glaciation during the Quaternary, or are in stratigraphic positions indicating development prior to the Quaternary.

Three methods have been used to determine the age of deep-weathering profiles: K–Ar, $^{40}\text{Ar}/^{39}\text{Ar}$, and cosmogenic isotope dating. The age of neoformed alunite-group sulfates (alunite and jarosite) and hollandite-group manganese oxides (cryptomellane and hollandite) precipitated through weathering reactions can be determined using K–Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ geochronology (Vasconcelos, 1999). Supergene alunite and jarosite, because of the relative simplicity of their mineral structures, can be dated equally well using either method, provided that large enough quantities of pure mineral separates can be obtained. On the other hand, the complex nature of supergene Mn oxides and the large number of mineral structures encompassed by the Mn-oxide group make application of the $^{40}\text{Ar}/^{39}\text{Ar}$ incremental-heating methodology better suited to dating these phases (Vasconcelos, 1999). The K–Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ methods can be applied across the full sweep of geologic time, but age measurements at the young end of the spectrum are more likely to be compromised by contamination with atmospheric Ar or primordial ^{40}Ar . The $^{40}\text{Ar}/^{39}\text{Ar}$ method allows for assessment of such contamination and therefore performs better in the younger (0.25–1.0 Ma) age range.

Cosmogenic ^{10}Be inventories have been used to estimate the length of accumulation time and hence the minimum age of weathering profiles formed during the past 5 Myr (Pavich et al., 1985). Complexities such as surface erosion and shielding effects of glacial ice limit the application of this technique in some areas, but several studies have successfully applied this and other cosmogenic isotope dating techniques to Quaternary weathering profiles (Pavich, 1986; Phillips et al., 1998).

Applications

Knowledge of the distribution, properties, origin, and age of weathering profiles has a wide variety of geological, engineering, hydrological, and land-use applications. Initially, interest in weathering profiles was fueled by geologists seeking to understand the origin and distribution of economically important supergene minerals and ores associated with some deep-weathering profiles. Subsequently, geologists also began to realize that information on the nature and distribution of past climates could also be gleaned from weathering profiles. In many cases, truncated weathering profiles are the only surviving evidence of ancient geomorphic surfaces (Bettis, 1998; Figure 11). Geochemical, isotopic, and geochronological studies of weathering profiles provide important insights into long-term erosion rates and landscape evolution. Engineers and engineering geologists have long appreciated the variations in physical properties of the regolith that occur through weathering profiles. Mapping the thickness and properties of weathering profiles has added a new dimension to land-use planning, construction engineering, and landslide hazard assessment (Hencher and McNicholl, 1995). Infiltrating water moves through weathering profiles on its way to both discharge areas and aquifers. Variations in hydrologic and hydrochemical properties of weathering profiles control many important aspects of water movement and quality in the vadose (unsaturated) and shallow phreatic (water-saturated) zones that have critical impacts on water supplies in many areas of the world.



Figure 11 (a) A weathering profile formed in late Pleistocene glacial diamicton overlying another weathering profile formed in older glacial diamicton. The contact between the gray unoxidized and unleached (UU) diamicton and the underlying brown oxidized and leached (OL) diamicton marks a significant stratigraphic break during which the lower weathering profile developed. Photograph by T.J. Kemmis. (b) Abrupt contact between unoxidized and unleached and oxidized and leached glacial diamictons. The presence of an unoxidized zone overlying an oxidized zone marks a significant stratigraphic break. Upper parts of the oxidized unit's weathering profile were removed by glacial erosion during advance of the glacier that deposited the upper diamicton. Photograph by E.A. Bettis.

Conclusions

Weathering profiles form over extended periods of time as geologic materials are altered in response to physical and chemical changes in the near-surface environment. Weathering profiles are the long-lasting 'roots' of surface soils, and their properties reflect the cumulative effects of weathering under changing climatic, biotic, and geomorphic conditions at a given location. Variations in the factors that affect weathering from place to place on a landscape, around the globe, and through time, have produced patterns in the properties of weathering profiles. Deciphering the history preserved in these patterns can provide information crucial to understanding long-term processes that shape the Earth's surface and that affect the environment in which we live.

See also: K/Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ Dating; Loess Deposits: Origins and Properties. **Paleosols and Wind-Blown Sediments:** Nature of Paleosols.

References

- Bettis EA III (1998) Subsoil weathering profile characteristics as indicators of the relative rank of stratigraphic breaks in till sequences. *Quaternary International* 51–52, 72–73.
- Birkeland PW (1999) *Soils and Geomorphology*, 3rd edn. New York: Oxford University Press.
- Bisdom EBA (1967) The role of micro-crack systems in the spheroidal weathering of an intrusive granite in Galicia (N.W. Spain). *Geologie en Mijnbouw – Netherlands Journal of Geosciences* 46: 333–340.
- Deere DU and Patton ED (1971) Slope stability in residual soils. In: *Proceedings of the 4th Panamerican Conference on Soil Mechanics and Foundation Engineering*, vol. 1, pp. 87–170. San Juan, Puerto Rico. New York: American Society of Civil Engineers.
- Gerrard AJ (1988) *Rocks and Landforms*. London: Unwin Hyman.
- Gerrard AJ (1994) Weathering of granitic rocks: Environment and clay mineral formation. In: Robinson DA and Williams RBG (eds.) *Rock Weathering and Landform Evolution*, pp. 3–20. New York: Wiley.
- Hallberg GR, Fenton TE, and Miller GA (1978) Standard weathering zone terminology for the description of Quaternary sediments in Iowa. In: Hallberg GR (ed.) *Standard Procedures for the Evaluation of Quaternary Materials in Iowa*, pp. 75–109. Iowa City: Iowa Geological Survey. Technical Information Series No. 8.
- Hencher SR and McNicholl DP (1995) Engineering in weathered rock. *Quarterly Journal of Engineering Geology* 28: 253–266.
- Mabbutt JA (1961) Basal surface or weathering front? *Proceedings of the Geological Association* 72: 357–358.
- McFarlane MJ (1992) Groundwater movement and water chemistry associated with weathering profiles of the African Surface in Malawi. In: Wright EP and Burgess WG (eds.) *Hydrogeology of Crystalline Basement Aquifers in Africa*, pp. 101–129. London: Geological Society of London. Special Publication 66.
- Nahon D and Tardy Y (1992) The ferruginous laterites. In: Butt CRM and Zeegers H (eds.) *Regolith Exploration Geochemistry in Tropical and Sub-tropical Terrains. Handbook of Exploration Geochemistry*, vol. 4, pp. 41–55. New York: Elsevier.
- Nesbitt HW and Young GM (1982) Early Proterozoic climates and plate motions inferred from major element chemistry of lutites. *Nature* 299: 715–717.
- Nesbitt HW and Young GM (1989) Formation and diagenesis of weathering profiles. *Journal of Geology* 97: 129–147.
- Norbury DR, Hencher SP, Cripps JC, and Lumsden A (1995) The description and classification of weathered rocks for engineering purposes. Geological Society Engineering Group Working Party Report. *Quarterly Journal of Engineering Geology* 28: 207–242.
- Ollier CD and Galloway RW (1990) The laterite profile, ferricrete and unconformity. *Catena* 17: 97–109.
- Ollier CD and Pain C (1996) *Regolith, Soils and Landforms*. Chichester: Wiley.
- Pavich MJ (1986) Processes and rates of saprolite production and erosion on a foliated granitic rock of the Virginia Piedmont. In: Colman SM and Dethier DP (eds.) *Rates of Chemical Weathering of Rocks and Minerals*, pp. 551–590. Orlando: Academic Press.
- Pavich MJ, Brown L, Valette-Silver JN, Klein J, and Middleton R (1985) ^{10}Be analysis of a Quaternary weathering profile in the Virginia Piedmont. *Geology* 13: 39–41.
- Phillips WM, McDonald EV, Reneau SL, and Poths J (1998) Dating soils and alluvium with cosmogenic ^{21}Ne depth profiles: Case studies from the Pajarito Plateau, New Mexico, USA. *Earth and Planetary Science Letters* 160: 209–223.
- Ruxton PB and Berry L (1957) Weathering of granite and associated features in Hong Kong. *Bulletin of the Geological Society of America* 68: 1263–1292.
- Strakhov NM (1967) *Principles of Lithogenesis*, vol. 1. Edinburgh: Oliver and Boyd. Fitzimmons P (trans.).
- Sueoka T (1988) Identification and classification of granitic residual soils using chemical weathering index. In: *Proceedings of the 2nd International Conference on Geomechanics in Tropical Soils. Geomechanics in Tropical Soils*, vol. 1, pp. 55–61.
- Tandarich JP, Darmody RG, and Follmer LR (1994) The pedo-weathering profile: A paradigm for whole-regolith pedology from the glaciated midcontinental United States of America. In: Cremeens DL, Brown RB, and Huddelson JH (eds.) *Whole Regolith Pedology*, pp. 97–117. Madison: Soil Science Society of America. Special Publication 34.
- Taylor G and Eggleton RA (2001) *Regolith Geology and Geomorphology*. Chichester: Wiley.
- Taylor RG and Howard KWF (1999) Lithological evidence for the evolution of weathered mantles in Uganda by tectonically controlled cycles of deep weathering and stripping. *Catena* 35: 65–94.
- Thomas M (1974) *Tropical Geomorphology: A Study of Weathering and Landform Development in Warm Climates*. New York: Wiley.
- Vasconcelos PM (1999) K-Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ geochronology of weathering processes. *Annual Review of Earth and Planetary Sciences* 27: 183–229.
- Walther J (1916) Das geologische Alter und die Bildung des Laterits. *Petermanns Geographische Mitteilungen* 62(1–7): 46–53.

Soil Morphology in Quaternary Studies

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Introduction

One of the most fundamental characteristics of all soils that enable their recognition is that they exhibit 'horizon' development as the most fundamental aspect of a soil profile (i.e., A–B–C profiles of 'master horizons'; [Figure 1](#)). This aspect of soils also allows them to be distinguished from other Earth materials that have some features that strongly resemble those of soil. Soil horizons are recognized on the basis of key attributes that distinguish them from the materials in which they are formed, called soil 'parent materials.' Chemical, mineralogical, and biological aspects of soils are used to define different soil horizons, but soil morphologic features are the main basis for recognizing and describing them in field settings.

Soil scientists formally recognize literally dozens of soil morphological properties ([Buol et al., 1997](#)). Some of the more important properties include soil texture, structure, color, mottling, surface coatings, and horizon thickness. These properties can be used to identify the master horizons because certain key properties strongly characterize them. For example, the A horizon typically is brown to dark brown, reflecting the accumulation of humus (organic matter in various states of chemical decomposition), one of the hallmarks of nearly all A horizons. In contrast, the E horizon is generally pale, largely reflecting the loss of iron minerals through certain chemical weathering reactions. Characterization of these and other properties (signified by the use of small case letters associated with specific pedogenic features) is used to identify soil 'genetic' horizons and 'subhorizons' (e.g., Bs horizon; 's' signifies presence of iron and aluminum oxides that formed in the soil). When combined with other types of data (e.g., textural, chemical, mineralogical, and micromorphological), 'diagnostic' horizons (e.g., argillic horizon) are designated ([Figure 2](#)). Such horizons are the primary basis for the classification of soils in the United States Department of Agronomy Soil Taxonomy.

The Relation of Soil Properties and Soil-Forming Processes

The emergence of soil properties is the result of a large number of soil-forming 'processes.' For example, soil texture, a morphological property defined as the relative proportions of sand, silt, and clay in the 'fine earth' (i.e., particles less than 2 mm in diameter), changes because of certain processes. One is the movement of rainwater through a soil, which causes a change in soil texture of the fine fraction of the soil parent material by the following:

1. Causing 'translocation' of fine soil particles from shallower to deeper parts of the soil. Different kinds of translocation include transport of mass as colloids, solid particles, and solutes. So, changes in soil texture that largely reflect

increases in clay content (a key morphological feature of argillic horizons in important soil orders like Aridisols, Alfisols, and Ultisols) are related to processes of translocation.

2. Chemical weathering, which can cause the formation of clay by alteration of soil parent material minerals (e.g., feldspar).
3. Physical weathering (also known as mechanical weathering), which is the physical breakdown of large particles into smaller particles.

Dissolved salts in soil water may move into larger parent material grains, causing them to disintegrate by process of salt weathering. These examples show that the formation of many soil morphological properties is often complicated, because essentially all soil properties are the expression of the influences of a very large number of different soil-forming processes that may operate at varying times (daily or seasonal) and magnitudes. These processes may also affect the character of A or C horizons, but they mainly operate in and help produce many of the important features that are fundamental to 'B' horizons.

Another important morphological property is soil color. It is often one of the most obvious and easily recognized soil morphological properties. Studies of soils in many landscapes show that they exhibit great variation in color, both among soils of different types ([Figure 3\(a\)–3\(c\)](#)) or between different horizons ([Figure 3\(a\)](#) and [3\(c\)](#); also see [Figure 2](#)). Like soil texture, the development of soil color also reflects many soil-forming processes. In one very common environment (well-drained and oxidizing conditions), certain soil horizons (especially the B horizon) become redder, or 'rubified.' This is most obvious if the soil parent material is one of the more abundant, typically pale brown soil parent materials with abundant quartz and feldspar (e.g., granite, many sandstones, and some metamorphic rocks). The principal soil-forming process involved is the oxidation of ferrous, divalent Fe to ferric, trivalent Fe, which typically precipitates as various iron oxyhydroxides (hereafter referred to as Fe oxides) that are generally reddish to reddish orange or reddish brown in color (e.g., hematite, goethite). A large body of literature provides explanations of the processes of chemical reduction and oxidation (chemical reactions that involve electron transfer and changes in valence state), solid-state transformation, and dehydration that ultimately produce these minerals (cf. [Schwertmann, 1994](#)). In relatively warm, oxidizing soil environments, hematite (typically red) is ultimately the thermodynamically favored mineral, whereas cooler temperatures often favor goethite formation (brown or yellow). This implies that soil Fe oxide mineralogy, which in turn is strongly correlated with soil color, should at least partly reflect processes that are in turn strongly related to climate (see below). But studies of soil morphology and soil Fe oxide mineralogy also show that the

relative content of hematite is also correlated with a degree of redness, which in turn may be related to the type of initially formed, poorly crystalline iron mineral. The accumulated mass of Fe oxides in soils is both time and climate dependent (Birkeland, 1999; Schwertmann, 1994), that is, soil color reflects processes clearly strongly related to climate, but that evolve over time.

Soil Morphology Studies and the State Factor (CLORPT) Approach

The climate in which a soil is forming or has formed is an important 'factor' that strongly influences the types, rates, and relative importance of various soil-forming processes and their relationship to different morphological properties. For example, the process of soil infiltration is closely related to the

timing, duration, and magnitude of rainfall, as well as soil temperature, humidity, and plant canopy characteristics, all of which are directly related to or essentially define the climate of a given area. Climate, however, is only one 'factor' that has a direct influence on processes of soil formation. Several centuries of soil research show that a large majority of literally hundreds of soil-forming processes can be linked in very fundamental ways to five principal 'factors.' In addition to climate (C), they include 'organisms' (O), perhaps better identified as the 'biotic' factor; 'relief' (R), or sometimes called the topographic factor; 'parent material' (P), the type of materials from which the soil has formed; and 'time' (T), or the duration of time over which the soil has formed (Jenny, 1941). Some aspects of soils cannot be easily evaluated using only these factors (e.g., spatial variation in soil development that reflect landscape evolution), but in most cases these five factors are the most critical ones, and they have come to be popularly known as 'CLORPT.' Birkeland (1999), Holliday (2004), and Schaetzl and Anderson (2004) and other researchers discuss criticisms of this approach and the recent emergence of other conceptual approaches in pedology; nevertheless, a large body of literature demonstrates the enormous impact the CLORPT approach has had in many Quaternary studies.

The real power of the CLORPT approach lies in its remarkable ability to elucidate the relative impacts of these five factors on soil formation. This is achieved through careful selection of groups of soils in circumstances such that the influences of one factor can be isolated as varying, while the others are essentially held 'constant' (Birkeland, 1999; Buol et al., 1997; Jenny, 1941). To identify differences that are related to soil age, a 'soil chronosequence' is established, and the time-dependent change in soil morphology (or a given property) is called a chronofunction (Figure 4). A large number of studies describe

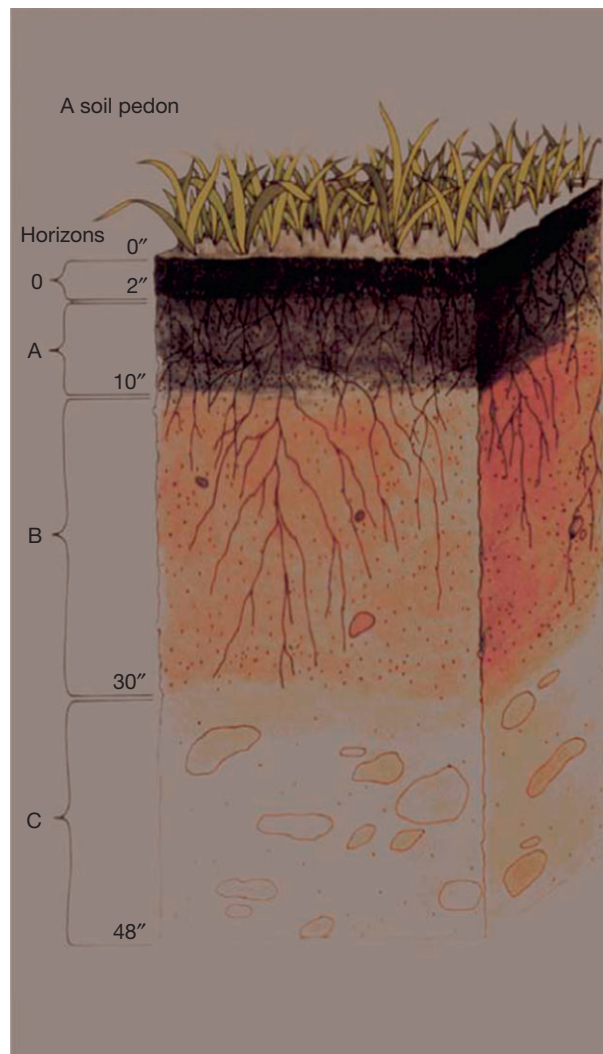


Figure 1 Schematic diagram of an idealized soil profile that exhibits a vertical sequence of soil horizons. Reproduced from The Soil Survey Staff (2006) Soil – Fundamental concepts. *Soils – Tools for Educators*, version 3.0 National Soil Survey Web Page. Lincoln: National Soil Survey Center, USDA-NRCS. <http://soils.usda.gov>.



Figure 2 A soil profile with a clay-enriched and reddened 'argillic horizon.' The horizons consists of an A horizon (0–7 inch.), and E, B/E, and Bt horizons from 11 to 33 inch. depth. Well-developed structure in this horizon is indicated by the strongly expressed blocky 'peds' that are outlined by cracks. Well-expressed A and E horizons over the argillic horizon are also evident. The base of the B horizon and top of the C horizon is at a depth of 38 inch. Scale in feet. From the SSSA Marbut Slide Collection (1968).

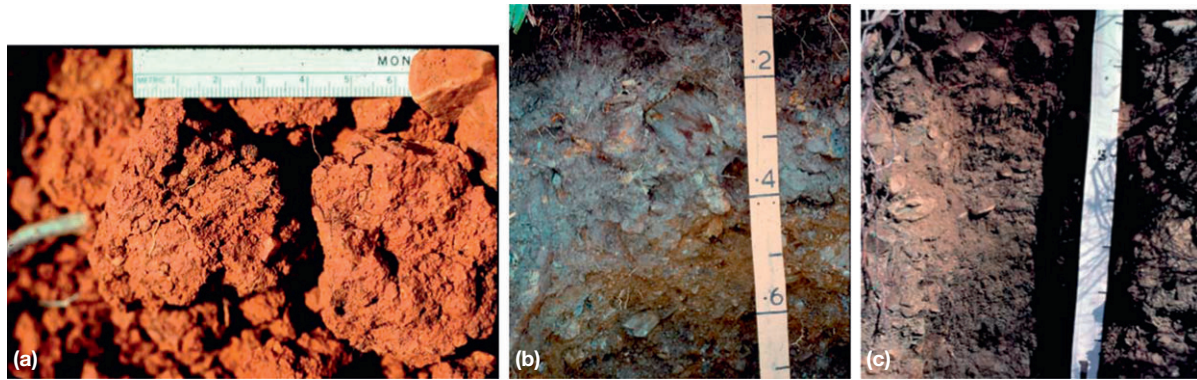


Figure 3 Soil materials with strongly contrasting soil colors. (a) Reddish brown (2.5 YR 5/4) peds from a B horizon in a very strongly developed soil formed in alluvial fan deposits in the Transverse Ranges, southern California. (b) Olive yellow (2.5 Y 6/6) B horizon in alluvial terrace deposits in the Transverse Ranges, southern California. The olive color of the soil reflects the presence of abundant chlorite in the parent materials, a mineral common in certain types of schists, a type of metamorphic rock. Scale in tenths of meters. (c) Strongly contrasting soil colors in a soil forming on the lower slope of a glacial moraine in New Zealand. The dark colors reflect the presence of humus, the reddish colors reflect the presence of oxidized iron-bearing iron oxyhydroxides, and the pale colors reflect the loss of iron oxides and humic materials by various types of leaching processes common in a humid, but seasonably cool climate. Scale in tenths of meters. Photograph by Bruce Harrison.



Figure 4 A suite of alluvial terraces associated with the North Fork of the San Gabriel River, San Gabriel Mountains, southern California. The ages of the terraces shown range from only a few centuries old (1) to a hundred thousand years old (4). Older eroded topography above terrace 4 is associated with a several hundred thousand year old terrace that is more well preserved elsewhere in this area. The soils present on each of these terraces would constitute a soil chronosequence, and changes in the development of selected properties of these soils would constitute a chronofunction, given the generally similar alluvial parent materials, topographic position, and vegetation of these soils.

soil chronosequences in various climate regimes, parent materials and landforms, and the mathematical nature of the chronofunctions (Birkeland, 1999; Bockheim, 1980). To show the influence of climate on soil development, a 'soil climosequence' is established, which generates 'climofunctions.' Similarly, 'biosequences,' 'toposequences,' and 'lithosequences' (and their associated factor-related functions) can be developed in appropriate circumstances.

Considering the key research questions and goals of Quaternary science, the most typical application of morphological studies of soils in Quaternary studies reflects the desire to use soils to provide important deposit and landform age information or paleoclimate/paleoenvironmental information. Therefore, soil chronosequence and climosequence studies have

been of particular interest to Quaternary scientists, although toposequences, not discussed in this article, are also used in studies of landform development. Studies of soil morphology and associated Quaternary sediments and landforms have been conducted in virtually all climatic regimes (cf. Birkeland, 1999; Buol et al., 1997), but a substantial fraction of soil morphology studies associated with chronosequence and climosequence development have been conducted in the temperate, arid, and semiarid regions. The discussion below mainly focuses on these latter studies.

There are some unavoidable problems with the CLORPT approach. An important one is that the CLORPT factors are not independent, but instead are often somewhat too strongly interdependent, such as climate and organisms. Therefore, it is sometimes difficult to determine the relative degree of importance with respect to two factors and their respective influence on the formation of a given property. Another important problem, especially in the study of Quaternary soils, is the global climate changes on timescales ranging from tens to hundreds of thousands of years (interglacial-to-glacial changes) to shorter timescales. Some of the larger climate changes of the Quaternary almost certainly caused changes in the rates, processes, and magnitudes of soil development through time. Therefore, there can be no chronofunctions that are completely 'time dependent,' that is, the increasingly older, pre-Holocene soils in a chronosequence will almost surely have morphological properties that must unavoidably reflect both the results of climate-change-induced processes and straightforward age-dependent properties. Such soils are called 'polygenetic' soils and they are abundant in many landscapes.

There may also be relatively few (if any) true climosequences, because time-dependent changes in soil development may inevitably lead to dramatic and rapid changes in soil development (which may be expressed as changes in morphologic expression) as 'soil thresholds' are crossed, potentially after varying times of soil development and reflecting different fundamental changes in soil-forming processes (Muhs, 1984). These problems are not insurmountable, as discussed in

Birkeland (1999), Holliday (2004), and elsewhere, and we need to be aware of these complexities. Significant increases in our understanding of the impact of climate changes on soils and their associated landscapes have increased our ability to isolate the most strongly time-dependent soil-forming processes and associated properties, and analogously, climate-related properties.

Relations Between Soil Morphology and Soil Age

Although there are a very large number of soil morphological properties, a relatively small number of them have proved especially useful in Quaternary studies. This partly reflects the fact that the relations between only a small handful of morphological properties, their related soil-forming processes, and their connection to CLORPT are reasonably well understood. For example, the origin and evolution of soil carbonate morphology is generally well understood with respect to age and climate, whereas the origins of soil structure with respect to climate and age are more complex and less well understood. Also, some morphologic properties are generally easier to quickly recognize and objectively characterize in the field than others (Reheis et al., 1989). There are many other seldom-used properties such as pore and root size, type, and density, but they are not unimportant and should also be included as possible in soil descriptions. For example, soil pore type, frequency, and spatial relationships (e.g., connectedness) critically affect the internal movement of soil water, certain soil materials, and soil gases.

An excellent opportunity to maximize the relationship between soil age and soil morphology using the CLORPT approach is provided by landscapes with flights of terraces of fluvial origin (see Figure 4). The composition of the alluvium, the topography of the surfaces and their vegetation communities are often broadly similar. Soils on pre-Holocene terraces have been subjected to different climate(s) earlier in the Quaternary, but some strategies can be used to recognize, and therefore explicitly consider, the probable impacts of the former climate regimes on soil-forming rates and processes. Other groups of landforms have also proved to be useful for the purpose of development of chronosequences, including lava flow surfaces (cf. McFadden et al., 1986), dunes (cf. Muhs et al., 1996), alluvial fans (cf. Harden et al., 1991), glacial moraines (cf. Burke and Birkeland, 1979), and marine terraces (Chadwick et al., 1990). Some landforms are generally unsuitable for chronosequence studies, such as those dominated by surface erosion as a process in their development. Examples include 'pediments,' which may require as many as several million years to form in certain rock types, and landforms characterized by inherently moderate to very steep gradients (e.g., hill slopes). Strong soils may form in loess, but continued episodes of loess accumulation slowly bury previously developed soils, promoting the development of a complicated, cumelic, probably polygenetic soil profile (Muhs and Bettis, 2003), making it difficult in most circumstances to directly use the CLORPT approach.

In addition to a favorable suite of landforms, recognizing and isolating strongly time-dependent properties are facilitated when the related soil-forming processes are generally well

understood. When this is the case, the potential influences of climate and/or changes in climate on a soil-forming process can be explicitly considered, and the impact on the process rate and associated property understood. A good example is provided by the strongly time-dependent changes in soil carbonate morphology, first described formally by Gile et al. (1966) and the subject of numerous subsequent studies (Birkeland, 1999). These studies provide a robust understanding of the key processes that control the rate, magnitude, and depth of carbonate accumulation, even enabling the development of numerical models for the simulation of calcic soils in different climates and parent materials (McFadden and Amundson, 2002).

Soil morphology does provide, however, a powerful means of determining 'relative ages'; for many types of landforms and it can also be used to temporally correlate landforms ('correlated ages'; Colman et al., 1987). Deposit or landform ages based on soil morphology can also be used in some favorable circumstances to provide provisional age-range estimates for certain landforms and deposits. One benefit of this is that such estimates can be used as a basis for selecting the most appropriate numerical dating technique. They may also be used to critically evaluate the results of more recently developed, but often difficult to interpret geochronological methods (e.g., cosmogenic radionuclide (CRN) dating).

Selected Examples of Time-Dependent Changes and Applications in Quaternary Studies

The development of soil chronosequences associated with different types of landforms shows that certain soil properties change systematically in strongly time-dependent ways in many different climatic settings. In most cases, key soil morphological properties require a minimum of a few centuries to millennia to form, and in some cases may require tens to hundreds of millennia (Birkeland, 1999). Chronosequence studies also show that rates of property development may vary considerably in different climate regimes. Examples of Quaternary studies that have greatly benefited from the use of soil morphology include the impacts of climate and climate change on fluvial systems (cf. Bull, 1991), eolian processes (cf. Muhs et al., 1996), and slopes (cf. McAuliffe et al., 2006); the impact of active tectonics on landscapes ranging from the fault scarp scale (Eppes et al., 2002; Machette, 1978; Menges, 1990), to the much larger scale of nearly continental scale, arc-related mountain belts (Pazzaglia and Brandon, 2001). Soil research has also been applied in important ways in archaeological research (Holliday, 2004; Waters, 1992) and landscape ecology (McAuliffe, 2003). Happily, the increased understanding of landscape evolution partly assisted by soil morphology studies has often subsequently provided important new insights into soil-forming processes.

Soil color is one soil property that, in many environments, changes in a strongly time-dependent manner. The Munsell Soil Color Chart is the most widely used means of determining soil color semiquantitatively. Soil color varies greatly, but many studies have shown that one of the strongest time-dependent changes in soil color, ranging from tens to hundreds of thousands of years, is the reddening of soils, the result of the accumulation of ferric iron oxyhydroxides such as

hematite (Schwertmann, 1994; Figure 5(a)–5(d)). Often the reddest color occurs in the most well-developed part of the soil B horizon. Once a chronofunction for soil ‘reddening’ has been developed in a certain region, it may be one of the most important, relatively easily measured soil properties that can be used as a basis for correlation of associated surfaces or deposits or, in appropriate circumstances, estimated age ranges.

Other properties of the soil B horizon that sometimes also change in a strongly time-dependent fashion are soil clay content, the number and thickness of soil clay films, and the expression of soil structure. The thickness of the soil B horizon (in which the strongest expression of these soil properties nearly always occurs) has been shown to change in a very strongly time-dependent fashion. The rate of change of B horizon thickness and other soil properties can be expressed mathematically in a few different ways, such as exponential or power functions.

Many explanations have been provided to explain these different mathematical trends. For example, the generally linear rate of soil carbonate accumulation over time spans of tens to hundreds of thousands of years may be attributable to the main process by which this material accumulates in soils (discussed below), the continual incorporation of calcium present in windborne dust or dissolved in rainwater. In contrast, a curvilinear function may reflect time-dependent changes in the accumulation of soil materials that mainly accumulate by *in situ* chemical weathering, a process that in many circumstances depletes the supply of the associated parent materials.

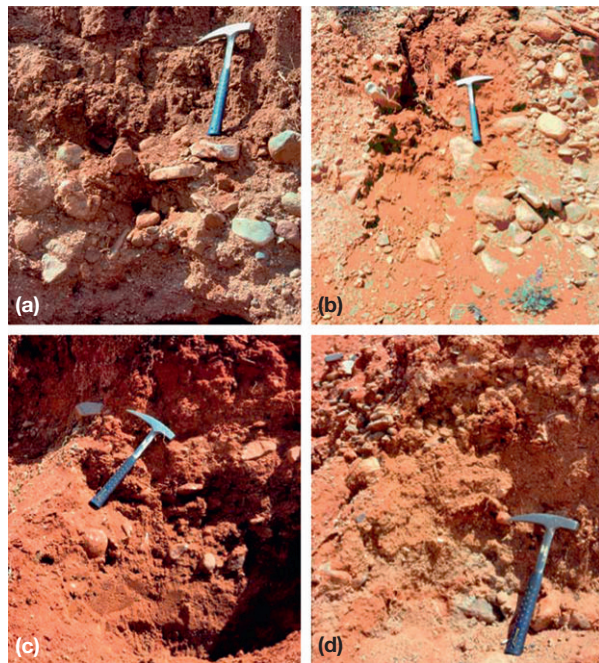


Figure 5 The increasingly redder colors of the B horizons of soils in a soil chronosequence associated with alluvial and fan surfaces of the Santa Catalina Mountains, near Tucson, Arizona. Age ranges of the soils are estimated on the basis of the similarities of the degree of their soil property development with that observed in other soils of the southwestern United States that have formed in a similarly arid, desert climate. (a) 1–5 ka old (Holocene); (b) 10–25 ka old (latest (?) Pleistocene soil); (c) 50–100 ka old (late Pleistocene); (d) 100–500 ka old (late middle (?) Pleistocene).

A few researchers also have proposed that some unexpected trends may be the result of past times during which global climate was uniquely favorable to rapid soil development (see discussion below). Field-based Quaternary studies show, however, that in some cases the cause of the development of unusually thick B horizons in some settings does not require remarkable changes in soil development rates or the unpredictable ‘emergence’ of some ostensibly anomalous property. For example, studies of Harrison and his colleagues (1990) showed that even very small contributions of material from hill slope sources above and well distant from a terrace tread can dramatically increase their B horizons to thicknesses double or more than those associated with topographically isolated surfaces. This is accomplished while the soil A horizon is converted slowly to a B horizon, in a fashion somewhat similar to that proposed for soils in loess (Busacca, 1989; Muhs and Bettis, 2003). Thus, substantial differences in the rate of change of a B horizon can be caused by seemingly subtle differences in surface slope, distance to slope sources, differences in composition of materials (including soils) on these slopes, and processes of bioturbation.

The time-dependent change in soil carbonate morphology is the one chronofunction probably most well-known by geoscientists who work in arid and semiarid regions (Figure 6(a) and 6(b)). Other studies also show that carbonate coatings on gravel (also called rinds) also thicken in a strongly time-dependent fashion, at least in some apparently favorable, but not particularly well-understood circumstances (Amoroso, 2006; Treadwell-Steitz and McFadden, 2000; Vincent et al., 1994). Chronosequence studies have highlighted the essential role that desert dust plays in the accumulation of increasing mass of carbonate, silica, and gypsum and its morphological

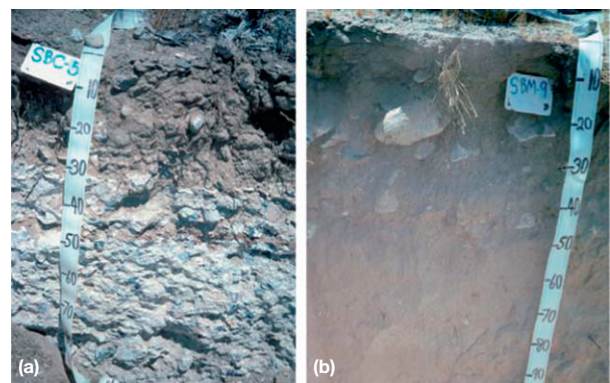


Figure 6 Expression of soil carbonate morphology in soils of different ages associated with alluvial fan deposits of the northern flank of the San Bernardino Mountains. (a) A relatively young ‘Holocene’ soil in gravelly alluvium showing ‘stage 1’ morphology. Note the carbonate coatings visible as the white areas on the bottom sides of the boulders between a depth of 20–50 cm. This is the most common expression of this morphologic stage in gravel-rich parent materials. (b) Old ‘Pleistocene’ soil, showing ‘stage 4’ morphology (massively cemented, strongly indurated horizon). Laminar morphology is also a hallmark of this advanced stage of carbonate accumulation and is visible at a depth of 45–50 cm. A diagnostic horizon with these features is called a ‘petrocalcic’ horizon, which, in this case, is buried by a much younger alluvial deposit, which implies loss of a once overlying B horizon by erosion of the fan surface (Eppes, 2002). Photograph by M. Eppes.

evolution (Harden et al., 1991). In some landscapes, the deposition of dust also greatly influences the evolution of many spatially extensive landforms and associated surficial deposits in arid regions. With relatively high rates of dust deposition, desert 'loess' deposits may form, whereas lower rates allow accumulation of the dust below the soil surface into the B horizon, which favors stronger soil development (Muhs and Bettis, 2003). In drier parts of deserts, these processes also favor desert pavement, because the fine material is trapped below any large, stable rocks, which stay at the surface as the slowly accumulating fine material is converted to soil.

In many landscapes, soils are rarely older than the Quaternary, usually younger than Middle Quaternary, and are often of Holocene age. Minimally developed soils may also dominate landscapes that have been tectonically very active during the Quaternary. In these landscapes, steep slopes, generated by active tectonics, favor high erosion rates and active accumulation of sediment on floodplains and often extensive alluvial fans. The soils in both cases are weakly developed, either because erosion removes the sediments faster than soils can form, or the deposition of sediment is too rapid to allow soils to form. Alpine glaciation is typically more extensive in the highest parts of mountain ranges, extending to lower elevations with increases in latitude. Many of the soils in landscapes dominated by glacial landforms reflect recent glaciations of the Quaternary. Only in the most tectonically stable regions in favorable latitudes do processes favor continued soil development (or at least preservation) over millions of years on geomorphically stable landforms. The best studied of these are very old, pre-Quaternary, highly polygenetic soils of Australia. These soils have a complicated morphology that reflects soil development over several million years under numerous global climate changes. They also reflect major changes in the climate of Australia that were caused by substantial changes in this continent's latitude associated with the northward movement of the Australian plate during the Cenozoic (Birkeland, 1999). Such old, complex soils are extremely rare and not typical of most continents.

Relations of Soil Morphology and Climate

As discussed above, a basic process that drives much of soil development is weathering, and the interaction of soil with infiltrating soil water is one of the underlying causes of many types of weathering processes. The most important soil-forming factor that dictates how much and when water will infiltrate and interact with soil solids, soil pore water, and the soil atmosphere is the climate factor. Because a very important goal of many Quaternary studies is to understand how climate changes affected the Earth during the Quaternary, the strong soil-climate relationship has attracted the interest of many Quaternary scientists. One important question that is asked concerns which aspects of soils can be evaluated specifically for the purpose of obtaining climate proxy information. Put another way, information from soils is sought that can be used to reconstruct quantitative aspects of the climate during some past time.

Some answers to these questions have been provided through studies of 'climosequences' and derived climofunctions using the CLORPT approach. Such studies have demonstrated

that there is a significant relation between the development of certain important diagnostic horizons and climate. For example, a very strongly chemically altered B horizon enriched in certain clay minerals, iron and aluminum oxides, and virtually devoid of the original parent materials (the oxic B horizon) forms only in generally seasonably warm, generally humid climate. In contrast, a B horizon enriched in calcium carbonate, a relatively soluble mineral, usually only forms in a generally arid or semiarid climate (a calcic horizon). The Mollic A epipedon usually forms in a subhumid to semiarid climate under grass, whereas usually other types of surface soil horizons typically form in different climate and vegetation settings (Buol et al., 1997). There are also important relations between soil mineralogy, soil chemistry, and climate (Birkeland, 1999). For example, soil formation in a generally 'humid' climate often favors the formation of kaolinite and aluminum oxyhydroxides, but not smectite, a clay mineral typical of soils of drylands. The rate of Fe oxide accumulation increases more rapidly for soils forming in progressively more humid climates in southern California. Certain aspects of B horizon morphological properties reflect B horizon mineralogy, partly reflecting these climate-soil property relationships.

Quaternary scientists, however, seek relatively 'high-resolution' climate proxy information, that is, they seek information such as the annual quantity and seasonal timing of rainfall or annual variation in air temperature. Climosequence studies show that although climate-soil relationships can be observed, these relationships are complicated in many respects, so that soil features provide mainly 'low-resolution' proxy data. Recalling the example concerning clay content discussed above, a single morphological property is the result of many underlying soil-forming processes, and these processes are affected by different aspects of 'climate' in potentially very different ways. For example, the magnitude of chemical weathering may increase in increasingly humid climate regimes, but such soils in the landscapes are also subject to eolian dust and its incorporation (cf. Muhs et al., 1990). The clay in this dust may well have formed in an entirely different kind of climate, thereby greatly complicating efforts to correlate B horizon mineralogy and overall clay content with climate type. In other words, there is no simple climate 'transfer function' that enables one to equate, for example, soil clay percent or mineralogy as a function of depth with a region's annual (or monthly) rainfall, temperature, relative humidity, actual and potential evapotranspiration, and wind velocity and/or direction.

One important aspect of soil carbonate morphology that does reflect a moderately good relation with 'climate' is the depth of carbonate accumulation and mean annual rainfall. Since Arkley (1963) first formally recognized this relation, many more recent studies have generally corroborated that relationship (cf. Retallack, 1994). The climate-depth relation is complicated and not as strong as might otherwise be expected because many soil-forming processes that are linked to 'climate' in varying ways determine the depth of soil carbonate accumulation. Numerical models of carbonate accumulation in soils have helped elucidate the relationship between carbonate morphology and climate (McFadden and Amundson, 2002). Model results show, for example, that climate changes during the Quaternary should have significantly changed the rate and depth of carbonate accumulation in soils,

and the morphology of some calcic soils seems to reflect changes simulated by such models. These models also show, however, that the specific changes in rates and depths of carbonate accumulation would depend on several important variables, some of which are difficult to specify quantitatively. No model yet exists that is capable of simulating the carbonate accumulation and morphological changes in natural settings, and until one does, carbonate soil morphology cannot be used as a source of high-resolution climate proxy data.

Morrison (1967) has argued that during certain periods of the Quaternary when climate was relatively warm (e.g., interglacial or interstadial periods), conditions were especially conducive for chemical weathering and very rapid soil development. He calls such periods 'soil-forming intervals' and the associated soils, 'geosols.' Some Quaternary scientists have accepted many of the key arguments that make up the conceptual basis of geosols, and, given the presumed circumstances under which they form, their identification in Quaternary sequences would have important climatostratigraphic significance (Catt, 1986). Many other studies strongly indicate, however, that the concept is very problematical (Birkeland, 1999).

Soil Morphology and Quaternary Deposits

Some soil morphology studies in the Quaternary sciences lay a strong emphasis on the recognition and interpretation of buried soils (Figure 7). The recognition of buried soils is



Figure 7 A well-developed buried soil in alluvial terrace deposits associated with Duncan Canyon Wash, a river associated with a drainage basin of the south flank of the San Gabriel Mountains, southern California. The top of the buried soil is indicated by the arrow. Geologist's hammer for scale.

particularly important with respect to the development and interpretation of stratigraphic sequences in Quaternary sediments. For example, the presence of a buried soil in a sequence of Quaternary sediments unambiguously demonstrates the presence of a stratigraphic unconformity, which in turn shows that the surface associated with the buried soil was geomorphologically stable over the period of time needed to allow the soil to form. Such soils can, for example, reveal how both local and regional base level has changed in the past, and the pattern of the buried soils may then help to determine the underlying causes of these changes. The degree of development of the buried soil also can be useful to estimate the duration of time over which it formed; in many cases, buried soils can be used as a basis to temporally correlate deposits over relatively large areas. Many times there is little evidence other than the presence of buried soils to allow correlation of gravelly non-marine deposits (e.g., those typical of alluvial fans and other high-energy depositional environments) since they often lack fossils or other datable materials.

There are some problems or disadvantages with respect to the recognition and interpretation of buried soils. Some features formed many meters to tens of meters below the ground surface by processes in the vadose (aerated) zone or groundwater table fluctuations may strongly resemble some features formed through soil development (Figure 8). Such misidentifications could easily lead to serious errors in the interpretation of the Quaternary history of a given region and so it is essential to avoid such errors. Also, the geomorphic processes that cause the burial of landscapes often lead to erosion and/or modification of the uppermost horizons of soils of those landscapes, thereby significantly diminishing the extent to which they can be used to provide the information often sought by Quaternary scientists.



Figure 8 Massively calcite-cemented alluvium, Marble Mountains, Mojave Desert in southern California. This material resembles an advanced stage of soil carbonate morphology, but it is formed by processes largely unrelated to soil development, including bank infiltration of stream water that contains large concentrations of dissolved calcite. The process occurs during large floods in these normally dry washes, which with time causes calcite cementation of the initially very porous gravelly materials along the bank with calcite. Field hand shovel for scale.

Quantitative Evaluation of Soil Morphological Development: Soil Profile Indices

The determination of a soil profile index, or profile development index (PDI), enables conventional soil morphological data (e.g., consistence, Munsell soil color, soil fine earth texture) to be transformed to a 'numerical' form. This is another way to quantitatively evaluate soil morphological information, besides laboratory analysis of soil materials. The method used most frequently is that developed from studies of soils in unconsolidated alluvial deposits of Quaternary age located in the San Joaquin Valley, California (Harden, 1982). Indices are calculated by rather arbitrarily assigning numerical values to different categories of properties (e.g., change in rubification, soil consistence, structure or increases in clay film thickness and content). Determination of both horizon and the PDI is made on the basis of quantitative comparisons of observed horizon property values with both those of the parent materials and a 'maximum value' for the given property.

Many studies show that this method is useful for highlighting strongly time-dependent trends in the morphological development of a chronosequence of soils (Figure 9), although PDI-derived age estimates should not be equated with conventional numerical ages. In some circumstances, the use of the PDI may be problematic. For example, the maximum rubification hue is 10R (pale red to red). This is a color that can ultimately form in soils that are over a million years old, but that have been subject to a soil moisture regime that has never (or only very rarely) favored chemically reducing conditions. These conditions would terminate the continued

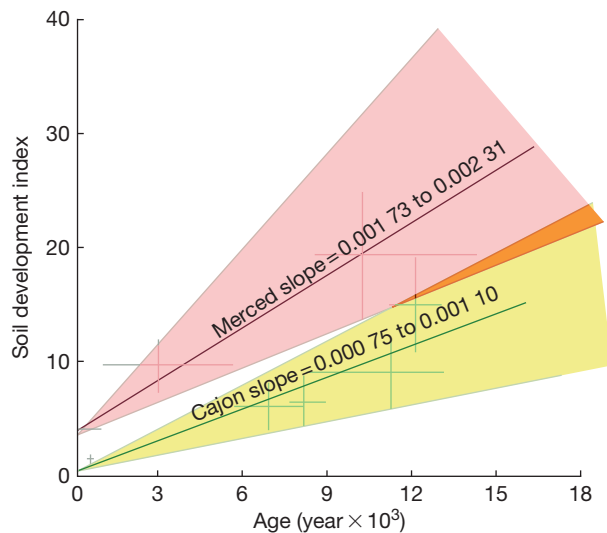


Figure 9 The change in the Profile Development Index values observed in two soil chronosequences in California from the Cajon Pass area of the Transverse Ranges and the Merced area of the Central Valley. The crosses show the estimated age ranges of the soil parent materials and the estimated range in the associated values of the PDI. The colored regions show the statistically determined range of uncertainty associated with the ordinary least-squares linear regression.

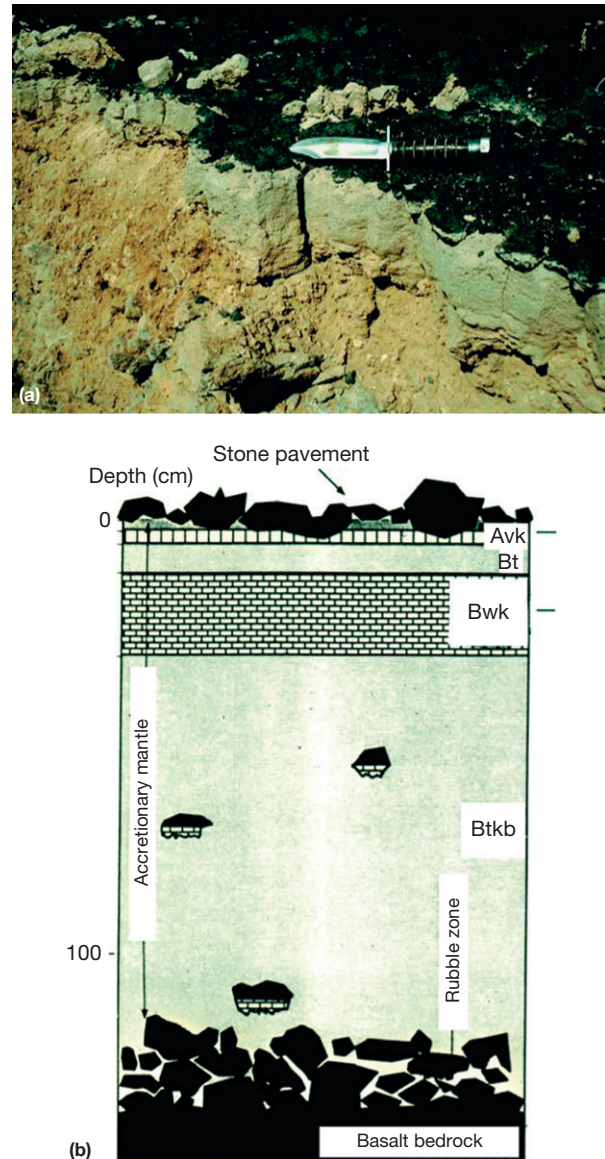


Figure 10 (a) A typical desert pavement and the soil profile beneath it formed on a late Pleistocene basaltic lava flow in the Cima Volcanic Field, Mojave Desert, California. Field knife for scale. (b) Schematic diagram depicting the stone pavement, soil, and underlying basalt flow bedrock. Age determination of the flow based on Ar 40–Ar 39 dating is 212–239 ka (unpublished age date; New Mexico Technical Institute Geochronology Laboratory, Heizler, 1999); the pavement clasts, from CRN dating, range from 140 to 180 ka (Poeths et al., 1998). The ages of the surface pavement rocks are consistent with pavement genesis via entirely subaerial, mechanical weathering of surface materials formed during flow emplacement. The ages of the uppermost materials based on thermoluminescence (TL) dating are much younger (5 ka in Avk; 13.5 ka in Bk; McFadden et al., 1998). The much younger TL ages are consistent with the morphologic development of the soil that immediately underlies the much older, darkly varnished stone pavement, illustrating how soil development faithfully reflects both its age and the desert surficial processes.

accumulation of iron minerals that promote strong rubification, or even cause their loss entirely from the soil, and such conditions may arise in the strongly weathered soils of landscapes with climate regimes that are very humid compared to California. Therefore, in a region associated with a relatively humid climate, it would be inappropriate to use the maximum value for rubification in the determination of the PDI of soils of that region. Also, many aspects of soil descriptions are unavoidably subjective and highly dependent on the experience and skills of the person describing the soil. Thus, a PDI incorporates this subjectivity. Finally, the numerical data assigned to the property categories represent ordinal-scale data. Calculation of the PDI essentially upgrades this data to interval-scale data which in turn is used in the determination of various statistics; this, however, is also problematic because ordinal-scale data cannot be used in the calculation of such statistics.

The Future of Soil Morphology Studies in Quaternary Science

Soil morphology studies have proved to be very useful in diverse areas of Quaternary research. The recent development of new methods for quantitative analysis of soil materials and numerical models of soil development that have helped elucidate the nature of soil-forming processes provide new opportunities for the study and interpretation of soil morphology. The development of new geochronological methods provides additional means to independently determine the numerical age of soils, which in some places like deserts was very difficult. CRN age dating (or 'surface exposure' dating) is an example of a method that is particularly useful in the dating of soils and surface features (Gosse and Phillips, 2001; Wells et al., 1995).

The emergence of some of these new methodologies has also somewhat diminished the use of soil morphology in some areas of Quaternary research. Many Quaternary scientists continue to recognize, however, the utility of soils in their research,

because they recognize that the character of soil morphologic development yields key insights into many processes of landform evolution on both short and long geologic timescales. Studies of the origin and evolution of extensive rocky desert surfaces are a good example of this. Studies of soil morphology help show that desert pavements form very differently than was thought for many decades (McFadden et al., 1987, 2005; Wells et al., 1995; Figure 10). CRN dating provides additional support for the new model of desert surface formation, because this method can be used to determine the 'first exposure ages' of the rocks in the pavement. The morphology of the soils immediately beneath the pavement, however, reflects the processes that have most recently dominated their development in the past few thousands of years, which CRN ages alone could not reveal. Similarly, on some land surfaces that might otherwise appear to be geologically young, the presence of even small fragments of soil materials that can only form over long periods of time (e.g., massively cemented soil carbonate or deeply rubified, clay-rich, near-surface B horizons associated with very different soil parent materials) reveal their true antiquity.

CRN dating of surficial deposits and soils can provide important information about basin-wide weathering and erosion rates, but soils also progressively incorporate eolian materials, so obtaining meaningful CRN ages in such circumstances may prove difficult, if not impossible. If soil chronosequence information is available, however, the soil morphology will almost always accurately reflect at least a general age range for the associated surface or deposit. Moreover, obtaining the CRN dates is ultimately a very expensive proposition; the description and evaluation of soils is not.

A soil is a highly complex natural system that probably becomes even more complex with increasing age, as it is periodically and unavoidably subjected to global climate changes of various magnitudes. These changes alter soil water balance and hydrology, plant communities, eolian accumulation rates and composition, and the types and rates of other important surficial processes that influence soil development (e.g., rate

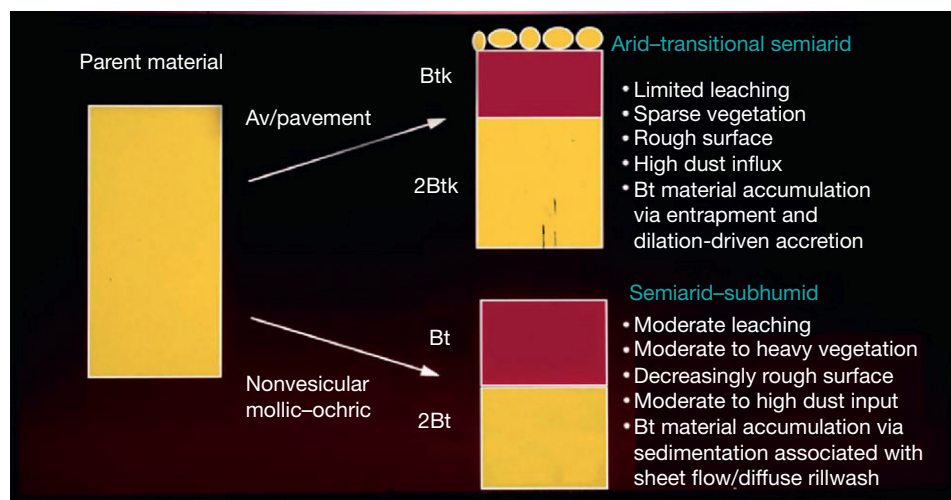


Figure 11 Two fundamentally different evolutionary pathways of soil and surface development in gravelly, alluvial parent materials in different climate regimes. The very different morphologic features of these soils reflect the different types and rates of various soil-forming processes and surficial processes that culminate in profoundly contrasting soils and landforms.

and magnitude of infiltration of rainwater relative to surface runoff, which strongly influences the nature of flow over the soil surface). Studies of some desert soil chronosequences in the American Southwest that have revealed very anomalous, poorly understood patterns of soil development may well reflect this soil system complexity (Figure 11). Perhaps these 'anomalies' actually do represent certain characteristic aspects of soils that are better understood as dynamical, complex systems, such as emergent properties, nonlinear behavior, or self-organization (cf. Sprott, 1997). The CLORPT approach, which has proved its great usefulness in soil morphological research in Quaternary sciences for nearly 70 years, may be the one used in those future studies that answer these questions.

See also: Loess Deposits: Origins and Properties. Paleosols and Wind-Blown Sediments: Mineral Magnetic Analysis; Nature of Paleosols; Soil Micromorphology. Quaternary Stratigraphy: Chronostratigraphy.

References

- Amoroso L (2006) Age calibration of carbonate rind thickness in Pleistocene soils for surficial deposit age estimation, Southwest USA. *Quaternary Research* 65: 172–178.
- Birkeland PW (1994) Variation in soil catena characteristics of moraines with time and climate, South Island, New Zealand. *Quaternary Research* 42: 49–59.
- Birkeland PW (1999) *Soils and Geomorphology*. New York: Oxford University Press.
- Bockheim JG (1980) Solution and use of chronofunctions in studying soil development. *Geoderma* 24: 71–85.
- Bull WB (1991) *Geomorphic Responses to Climatic Change*. New York: Oxford University Press.
- Bull SW, Southard RJ, Graham RC, and McDaniel PA (1997) *Soil Genesis and Classification*. Ames, IA: Iowa State Press.
- Burke RM and Birkeland PW (1979) Reevaluation of multiparameter dating techniques and their application to the glacial sequence along the eastern escarpment of the Sierra Nevada, California. *Quaternary Research* 11: 21–51.
- Busacca AJ (1989) Long Quaternary record in eastern Washington, USA, interpreted from multiple buried paleosols in loess. *Geoderma* 45: 105–122.
- Catt JA (1986) *Soils and Quaternary Geology – A Handbook for Field Scientists*. Oxford: Clarendon Press.
- Chadwick OA, Brimhall GH, and Hendricks DM (1990) From a black to a gray box – A mass balance interpretation of pedogenesis. *Geomorphology* 2: 369–390.
- Colman SM, Pierce KL, and Birkeland PW (1987) Suggested terminology for Quaternary dating methods. *Quaternary Research* 28: 314–319.
- Eppes MC (2000) *Soil Geomorphology of the North Flank of the San Bernardino Mountains, Southern California*. University of New Mexico. PhD Dissertation.
- Eppes MC, McFadden LD, Matti J, and Powell R (2002) Influence of soil development on the geomorphic evolution of landscapes: An example from the Transverse Ranges of California. *Geology* 30: 195–198.
- Gile LH, Peterson FF, and Grossman RB (1966) Morphological and genetic sequences of carbonate accumulation in desert soils. *Soil Science* 101: 347–360.
- Gosse JC and Phillips FM (2001) Terrestrial in situ cosmogenic nuclides: Theory and application. *Quaternary Science Reviews* 20: 1475–1560.
- Harden JW (1982) A quantitative index of soil development from field descriptions examples from a chronosequence in central California. *Geoderma* 28: 1–28.
- Harden JW, Taylor EM, McFadden LD, and Reheis MC (1991) Calcic, silicic and gypsic soil chronosequences in arid and semi-arid environments. Soil Science Society of America. Special Publication 26, pp. 1–16.
- Harrison JBJ, McFadden LD, and Weldon RJ (1990) Spatial soil variability in the Cajon Pass chronosequence: Implications for the use of soils as a geochronological tool. *Geomorphology* 3: 399–416.
- Holliday VT (2004) *Soils in Archeological Research*. Oxford, UK: Oxford University Press.
- Jenny H (1941) *Factors of Soil Formation*. New York: McGraw-Hill.
- Machette MN (1978) Dating Quaternary faults in the southwestern United States by using buried calcic paleosols. *US Geological Survey Journal of Research* 6: 369–381.
- McAuliffe JR (2003) The interface between precipitation and vegetation – The importance of soils in arid and semi-arid environments. In: Weltzin JF and McPherson JF (eds.) *Changing precipitation Regimes and Terrestrial Ecosystems*, pp. 2–26. Tucson, AZ: The University of Arizona Press.
- McAuliffe JR, Scuderi LA, and McFadden LD (2006) Tree-ring record of hillslope erosion and valley floor dynamics: Landscape responses to climate variations during the last 400 year on the Colorado Plateau, northeastern Arizona. *Global and Planetary Change* 50: 184–201.
- McFadden LD and Amundson R (2002) *Inorganic Carbon Modeling. Encyclopedia of Soil Science*, pp. 718–721. New York: Dekker.
- McFadden LD, Eppes MC, Gillespie AR, and Hallet B (2005) Physical weathering in arid landscapes due to diurnal variation in the direction of solar heating. *Geological Society of America Bulletin* 117: 161–173.
- McFadden LD, Wells SG, and Jercinovic MJ (1987) Influences of eolian and pedogenic processes on the evolution and origin of desert pavements. *Geology* 15: 504–508.
- Menges CM (1990) Soils and geomorphic evolution of bedrock facets on a tectonically active mountain front, western Sangre de Cristo Mountains, New Mexico. In: Kruepfer PLK and McFadden LD (eds.) *Soils and Landscape Evolution*, pp. 301–332. Amsterdam: Elsevier.
- Morrison RB (1967) Principles of Quaternary stratigraphy. In: Morrison RB and Wright HB (eds.) *Quaternary Soils*, pp. 1–69. International Association of Quaternary Research.
- Muhs DR (1984) Intrinsic thresholds in soil systems. *Physical Geography* 5: 99–110.
- Muhs DR and Bettis DR (2003) Quaternary loess-paleosol sequences as examples of climate-driven sedimentary extremes. Geological Society of America. Special Paper 370, pp. 53–74.
- Muhs DR, Bush CA, Stewart KC, Rowland TR, and Crittenden RC (1990) Geochemical evidence of Saharan dust material for soils developed in Quaternary limestones of the Caribbean and western Atlantic islands. *Quaternary Research* 33: 157–177.
- Muhs DR, Stafford TW, Cowherd SA, et al. (1996) Origin of the late Quaternary dune fields of northeastern Colorado. *Geomorphology* 17: 129–149.
- Pazzaglia FJ and Brandon MT (2001) A fluvial record of long-term steady-state uplift and erosion across the Cascadia forearc high, western Washington State. *American Journal of Science* 301: 385–431.
- Poths J, McFadden LD, McLain A, Anderson K, and Wells SG (1998) *Insights into stone pavements from cosmogenic helium dates*. Abstracts with Programs, Geological Society of America Annual Meeting 30: A–299.
- Reheis MC, Harden JW, McFadden LD, and Shroba RR (1989) Development rates of Late Quaternary soils, Silver Lake Playa, California. *Soil Science Society of America Journal* 53: 1127–1140.
- Retallack GJ (1994) The environmental factor approach to the interpretation of paleosols. Soil Science Society of America Special Publication 33, pp. 31–64.
- Schaetzl RJ and Anderson SN (2004) *Soils: Genesis and Geomorphology*. Cambridge: Cambridge University Press.
- Schwertmann U (1994) Relations between iron oxides soil color and soil formation. In: *Soil Color*, pp. 51–70. Madison, WI: Soil Science Society of America Special Publication Number 31.
- Soil Survey Staff (2006) Soil – Fundamental concepts. *Soils – Tools for Educators, version 3.0 National Soil Survey Web Page*. Lincoln: National Soil Survey Center, USDA-NRCS <http://soils.usda.gov>.
- Sprott JC (1997) *The Science of Complexity*. <http://sprott.physics.wisc.edu>.
- Treadwell-Steitz C and McFadden LD (2000) Influence of parent material and grain size on carbonate formation in gravelly soils in a desert Piedmont, Sevilleta LTER, Palo Duro Canyon, New Mexico. *Geoderma* 94: 1–22.
- Vincent KR, Bull WB, and Chadwick OA (1994) Construction of a soil chronosequence using the thickness of pedogenic carbonate coatings. *Journal of Geosciences Education* 42: 316–324.
- Waters MR (1992) *Principles of Geoarchaeology: A North American Perspective*. Tucson: The University of Arizona Press.
- Wells SG, McFadden LD, Poths J, and Olinger CT (1995) Cosmogenic ³He exposure dating of stone pavements: Implications for landscape evolution in deserts. *Geology* 23: 613–616.

Biogeochemical Role of Dust in Quaternary Climate Cycles

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Introduction

The geologic record reveals to us that the activity of dust on Earth has varied considerably through time, in response to large reorganizations of the climate system. For example, major increases in the deposition of dust downwind of Asia began 6 to 2 Ma, mostly likely associated with the uplift of the Himalayas and the subsequent evolution of the East Asian monsoon system (e.g., [An et al., 2001](#)). We see a similar intensification of the dust production downwind of Africa with the onset of Northern Hemisphere glaciations and cooling of North Atlantic surface temperatures after 2.8 Ma ([DeMenocal, 2004](#)).

Over the past 2.6 My, the Earth has gradually transitioned into a period of regular, glacial–interglacial cycles, and the global dust cycle has developed in pace with it. Over the past 800 000 years, the Earth's dustiest periods have coincided with the periods of maximum ice sheets and minimum temperatures, which come every 100 000 years. Well-constrained records of dust deposition in marine sediments from the Pacific Ocean ([Hovan et al., 1991](#); [Stancin et al., 2008](#); [Winckler et al., 2008](#)), the equatorial Atlantic and Indian Oceans ([Clemens and Prell, 1990](#); [DeMenocal, 2004](#)), the mid-latitude South Atlantic Ocean ([Martinez-Garcia et al., 2009](#)), in loess deposits on the continents (e.g., [Ding et al., 1994, 2001](#); [Guo et al., 2002](#); [Karimi et al., 2011](#); [Muhs et al., 2003](#); [Rousseau et al., 2011](#); [Schmidt et al., 2011](#); [Tófaló et al., 2011](#); [Újvári et al., 2010](#); [Zech et al., 2011](#)), and in ice at remote locations of the polar ice caps ([Lambert et al., 2008](#)) suggest that these cycles of dust deposition occur in a very similar fashion throughout the world during the Quaternary ([Figure 1](#)).

Superb records of changes in dust deposition have been retrieved from Antarctic ice cores over the past 800 000 years ([Lambert et al., 2008](#)). A close examination of these records suggests that fluctuations in dust, atmospheric trace gases, and temperature exhibit remarkably cyclic behavior, suggesting that dust is a key feedback within the Earth system ([Figure 2](#)). As the Earth enters an Ice Age, dust concentrations begin to rise only after temperature and CO₂ concentrations have begun to drop toward their full glacial conditions. In other words, the dust cycle is invigorated about half-way through the Earth's descent into glaciation, and thus may behave as some form of feedback that could amplify glacial conditions ([Kohfeld and Ridgwell, 2009](#)).

Global compilations of dust deposition have been used as a means of evaluating our ability to simulate glacial–interglacial changes in the dust cycle using Earth system models, and until recently focused on the Last Glacial Maximum (LGM) period, approximately 21 000 years ago ([Figure 3](#)). During the last Ice Age, dust deposition increased by approximately two to four times globally, and by as much as 20 times in the polar regions when compared with warmer interglacial periods ([Kohfeld and Harrison, 2001](#); [Kohfeld and Tegen, 2007](#); [Mahowald et al., 2006a](#); [Rea, 1994](#)).

Dust as a Link Between Land, Atmosphere, and Sea

These geologic records demonstrate that dust deposition has changed dramatically in the past, and may have behaved as a rhythmic feedback that responds to changes in the Earth's climate, and subsequently impacts different parts of the Earth system. Dust is emitted from the land surface in response to changes in land cover, but also in response to climate, which controls changes in temperature, precipitation, wind, and soil moisture. Dust enters the atmosphere where it affects the regional and global radiative budget – and therefore climate – before it is deposited. Dust deposition delivers important trace elements and nutrients to land surfaces and has contributed to long-term soil formation and development throughout the Quaternary in several parts of the world ([Chadwick and Davis, 1990](#); [Chadwick et al., 1999](#); [Muhs, 2001](#); [Muhs et al., 1990, 2010a,b](#); [Swap et al., 1992](#); [Vitousek et al., 1999](#)). When dust is deposited over the sea, it may act as a fertilizer to marine organisms, influencing ocean carbon sequestration and the production of several trace gases and chemical constituents that can affect climate.

We can also anticipate that dust will continue changing in the future, with natural and human-induced perturbations to climate and the Earth's land surface. Our ability to simulate these changes with dust cycle models can be evaluated based on comparisons between Earth system models and the geologic record, and help us to assess what role dust could play in modifying future climate and biogeochemical systems. In the past 12 years, Quaternary dust observations have been used as a means of validating dust cycle models, and have helped to quantify the impacts of changes on dust on Quaternary climate and carbon cycles ([Cakmur et al., 2006](#); [Kohfeld and Ridgwell, 2009](#); [Mahowald et al., 1999, 2006a](#); [Reader et al., 1999](#); [Takemura et al., 2009](#); [Tegen et al., 2002](#); [Werner et al., 2003](#)).

Land-Dust Emissions

Simulated estimates of total modern global dust emissions range from 1019 to 3321 Tg year⁻¹, with optimized multimodel estimates yielding a tighter range of 1500–2600 Tg year⁻¹ ([Cakmur et al., 2006](#)). A comparison of all modeling studies suggests convergence at a value of 2000 Tg year⁻¹, although this convergence does not imply accuracy given large uncertainties associated with our ability to quantify dust emission, deposition, and particle size characteristics at a global scale ([Shao et al., 2011](#)). Several factors influence the emission of dust from the land surface. These factors include (a) the characteristics of the soil surface, including surface roughness, particle size, soil mineralogy, (b) soil moisture, (c) soil availability, (d) the degree of vegetative cover, and (e) the degree of particle cohesion, crusting, or cementation. Once these basic conditions are met, dust emissions occur when the wind speeds exceed a threshold

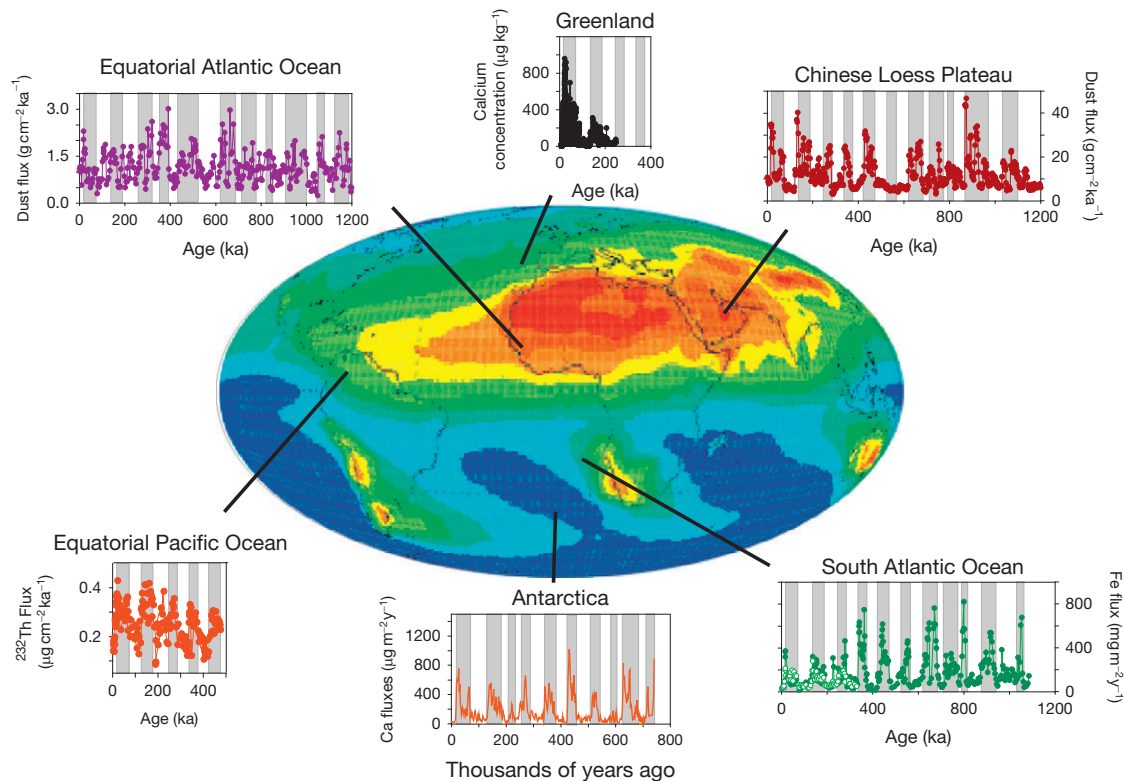


Figure 1 Dust deposition on land, ocean, and ice surfaces all around the globe has changed in pace with the ice ages over the course of the Quaternary. Data shown here represent a variety of proxies for dust and include (moving clockwise from the top): Calcium concentrations ($\mu\text{g kg}^{-1}$) from the Greenland ice sheet GRIP core (De Angelis et al., 1997); aeolian fluxes ($\text{g cm}^{-2} \text{ka}^{-1}$) calculated for the Zhaojiachuan (35.45°N , 107.49°E , 1250 m) and Lingtai 35.04°N , 107.39°E , 1350 m) loess sections from the central China Loess Plateau (Sun and An, 2005); Iron fluxes (green circles) and ^{230}Th -normalized iron fluxes (white circles, $\text{mg m}^{-2} \text{year}$) from deep-sea core ODP Site 1090 (42.91°S , 8.9°E , -3700 m) in the South Atlantic Ocean (Martinez-Garcia et al., 2009); nonsea salt Calcium fluxes to the EPICA Dome Concordia ice core in East Antarctica (Lambert et al., 2008); normalized ^{230}Th fluxes from deep-sea core TT013-PC72 (0.11°N , 139.4°W , -4298 m) in the Equatorial Pacific Ocean (Winckler et al., 2008); and dust fluxes to ODP Site 108-659A (18.08°N , 21.03°W , -3082 m) in the Equatorial Atlantic Ocean (Tiedemann et al., 1994). Glacial and cold stadial periods are shaded in gray. Central figure shows average atmospheric dust concentrations taken from Jickells et al. (2005).

velocity that allows soil particles to be mobilized. A look at satellite images demonstrates that dust emissions happen today mostly in arid and semiarid regions around the world (Figure 4), in dust emission ‘hotspots’ found over lake, fluvial, and dune deposits, but also in areas of active glaciation such as Alaska (Crusius et al., 2011). The majority of these dust emission sources (63%) are located in North Africa and the Middle East, with Australia and Asia holding 11% and 8% of these hotspots, respectively (Mahowald et al., 2009).

Given that many factors controlling dust emissions are likely to change with climate (e.g., temperature, precipitation, and winds), it is not surprising that dust deposition appears to have changed throughout the Quaternary. For example, increased meridional temperature gradients at the LGM compared to today were likely to have increased the strength of the storm tracks, increasing surface winds and gustiness in some locations (McGee et al., 2010). Temperature reconstructions suggest that tropical sea-surface temperatures were approximately $2\text{--}3^\circ \text{C}$ cooler at the LGM compared to today, and the difference was as large as -2 to -6°C in certain polar regions (MARGO Project Members, 2009). Terrestrial surface temperatures were on the order of -5.5°C cooler in the tropics (Ballantyne et al., 2005) and as much as $10\text{--}23^\circ \text{C}$ lower in

Antarctica and Greenland, respectively (McGee et al., 2010). The thermal wind relationship suggests that increased glacial meridional temperature gradients were likely to be related to increased pressure gradients and intensified jet streams. These may have intensified gusts associated with cold fronts near mid-latitude sources (McGee et al., 2010). Observational evidence also supports increases in wind speeds over key regions. Evidence of increases in the median grain size during glacial periods is found near source regions such as in China (Ding et al., 2001) and North Africa (Sarnthein et al., 1981). Measurements of changes in the median grain size data cannot always be interpreted simply as changes in wind speeds. Nevertheless, analysis of model simulations also supports increases in wind speeds near glacial source regions. Analysis of extreme winds from LGM simulations suggest that the frequency of extreme wind speeds increased over key source regions in Patagonia, Asia, and Africa (McGee et al., 2010).

Changes in climate likely resulted in expanded dust source areas during the LGM for a number of possible reasons. First, given that sea level was lower by approximately 120 m (Peltier and Fairbanks, 2006), exposed continental shelves might have served as new dust sources. Some model simulations have implicated exposed shelves in high-latitude regions as contributing

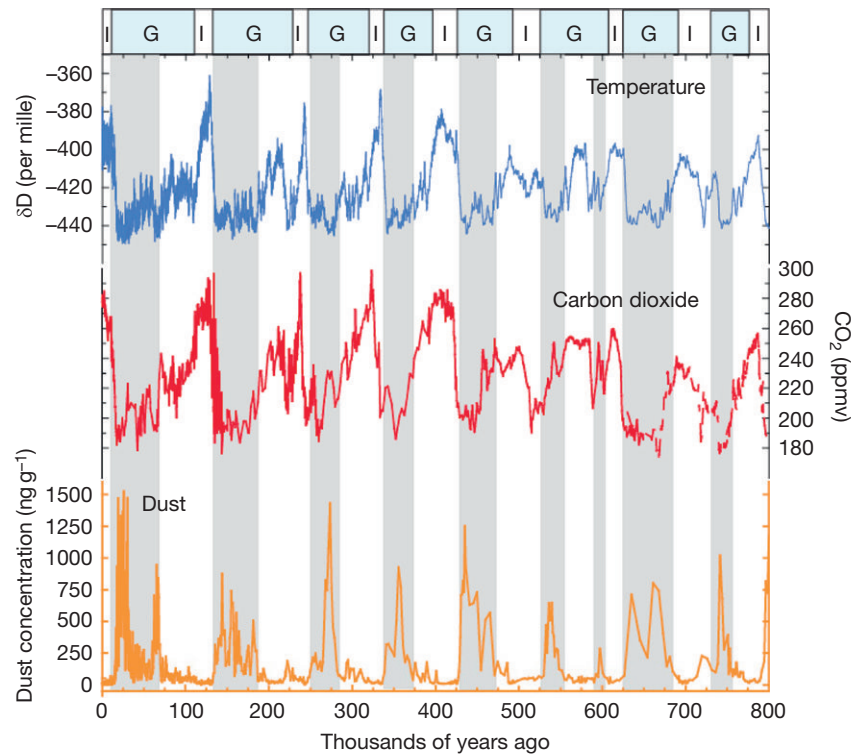


Figure 2 Temperature and carbon dioxide concentrations measured in Antarctic ice cores have demonstrated tightly coupled variations over the last 800,000 years. While dust deposition exhibits a regular rhythm as well, it only increases during the later stages of each of the glacial cycles. Redrawn from Kohfeld KE and Ridgwell A (2009). Glacial–interglacial variability in Atmospheric CO₂. In: Le Quere C and Saltzman ES (eds.) *The Surface Ocean – Lower Atmosphere Processes, Geophysical Monograph Series*. AGU: Washington, DC. <http://dx.doi.org/10.1029/2008GM000845>.

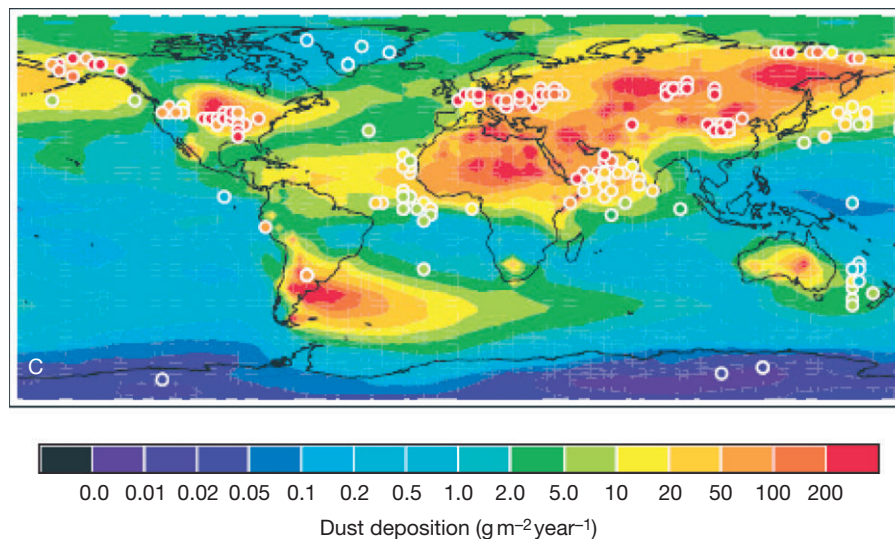


Figure 3 Comparison between observed and modeled changes in dust deposition demonstrate that the Earth was two to three times dustier during the last glacial period. Reproduced from Mahowald NM, Muhs DR, Levis S et al. (2006) Change in atmospheric mineral aerosols in response to climate: Last glacial period pre-industrial, modern and doubled carbon dioxide climates. *Journal of Geophysical Research – Atmospheres* 111: doi:10.1029/2005JD006653.

to dust production (e.g., Andersen et al., 1998; Werner et al., 2003). Werner et al. (2003) suggested that exposed shelves (primarily in Australia, South America, eastern Siberia, and Alaska) may have contributed as much as 25% to the observed

increase in dust during the glacial period. Evidence suggests that lowered sea level may have enhanced potential sources for coastal sand dunes (Hanebuth and Lantzsch, 2008; Muhs et al., 2008, 2010a). Enhanced glacial sources of dust from

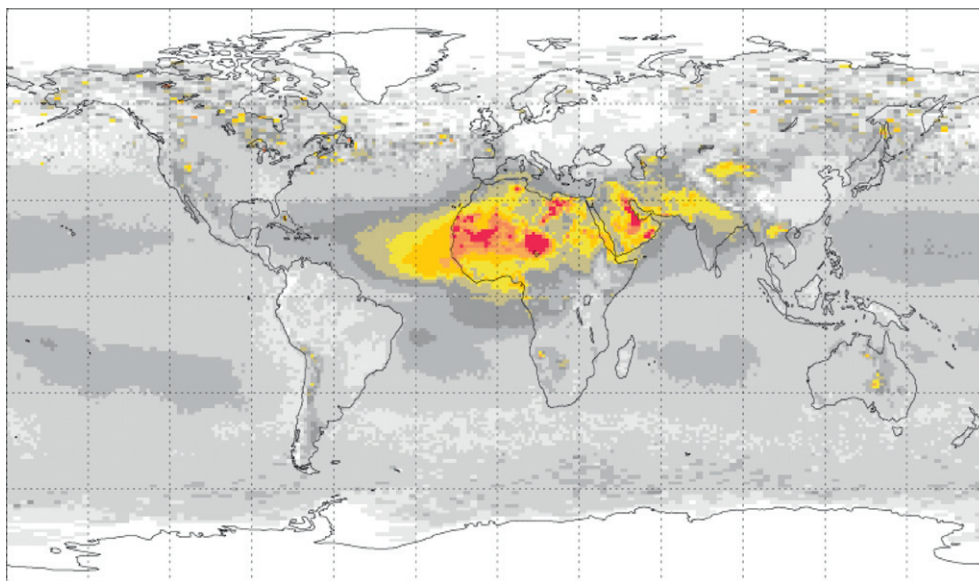


Figure 4 The Aerosol Index averaged from 1985 to 1990 from the Total Ozone Mapping Spectrometer (TOMS) provides a first-order approximation of where dust sources are located today (NASA Total Ozone Mapping Spectrometer Ozone Processing Team, NASA/GSFC Code 613.3, Data Product: Aerosol Index, http://toms.gsfc.nasa.gov/aerosols/aerosols_v8.html).

continental shelves are difficult to infer (Walter et al., 2000), but have been suggested for parts of Australia, South America, and Antarctica (Bigler et al., 2006; Gingele et al., 2004).

Second, in addition to climatic changes, lower CO₂ levels may also have inhibited photosynthetic activity and restricted vegetation near potential dust source areas. Several reconstructions of paleo-vegetation, lake level status, and paleo-humidity suggest that conditions during the LGM were likely more arid near source regions (see, e.g., Kohfeld and Harrison, 2001; Tjallingii et al., 2008). Vegetation reconstructions indicate that the boreal forests were greatly fragmented during the LGM across Europe, Asia, Beringia, and North America, and replaced by more cold, arid types such as tundra and steppe vegetation (Prentice et al., 2000) or by large ice sheets. Deserts in eastern Asia were expanded during the LGM (Yu et al., 2000), conditions in and near Asian dust source areas were more arid during glacial periods, and deposition was greatly increased near these sources (e.g., Ding et al., 2001). Today the highest rates of dust deposition are found downwind of North Africa, and have been shown to be two to five times greater at the last glacial period compared with interglacial periods (Kohfeld and Harrison, 2001; Ruddiman, 1997; Sarnthein et al., 1981). Provenance studies suggest that glacial and interglacial dust deposited downwind of Africa has similar sources (Cole et al., 2009; Grousset et al., 1998). Changes in dust deposition have also been primarily attributed to colder, drier, and windier conditions (Harrison et al., 2001), and recent work suggests that the Sahara and Sahel regions of Africa experienced reduced vegetation cover during the last glacial period as well (Tjallingii et al., 2008). Another notable, important source region during the last glacial period was the region near Patagonia, South America. Provenance studies have repeatedly demonstrated that this region became an important source of dust to the Antarctic Ice Sheet during glacial periods (Basile et al., 1997; Delmonte et al., 2004; Gabrielli et al., 2010; Grousset et al., 1992).

During ice ages when large ice sheets covered large portions of the continents in the Northern Hemisphere, the increase in

frozen ground cover is likely to have inhibited some dust emissions from high latitude sources. At the same time, glacial grinding associated with expanded glaciers also provided an important source of dust to the atmosphere. In fact, inclusion of inferred, glaciogenic dust sources in one model simulation increased LGM emission estimates by 57% and helped to account for massive dust deposition rates observed downwind of sources in North America, Europe, and Asia (see Mahowald et al., 2006a and references therein).

Several modeling studies have demonstrated that the impact of glacial climate on soil moisture, expanded continental shelves, vegetation, and winds resulted in increases in dust emissions and deposition that were comparable to the two- to threefold increases found in observational records at the LGM compared to today (Table 1). However, a closer look at those simulations demonstrates that while these simulations all reproduce glacial–interglacial changes in dust emissions that are similar to those expected from observed changes, the reasons responsible for the changes in dust deposition are very different. Simulated increases in LGM dust source regions range from 35% to 200%. The contribution of enhanced wind strengths over existing source regions to increased glacial dust emissions range from 3% to 65%. Only one of these model simulations has even inferred the impact of glaciogenic sources (Mahowald et al., 2006a). This wide range of differences in model simulations suggest that our ability to constrain future changes in dust source areas and dust emissions in response to anthropogenic climate change remains a challenge.

Atmosphere – The Impact of Dust on the Radiative Budget in the Past, Present, and Future

The overall radiative impacts of dust have been difficult to quantify. Estimating a global impact of dust on the radiative budget is difficult because both the sign and the magnitude of

Table 1 Simulated changes in dust emissions during the Last Glacial Maximum

Study	Simulated changes	Total change in emissions (LGM/Today)	% Change in source area extent	% Contribution to LGM emission increases	
				Winds over current sources	Expanded source areas
Mahowald et al. (1999)	Vegetation + CO ₂	3.0	+200	3	97
Werner et al. (2003)	Vegetation + CO ₂ + shelf	2.2	+270	65	35
Mahowald et al. (2006a)	Vegetation + CO ₂	2.1	+35	36	64
Mahowald et al. (2006a)	Vegetation + CO ₂ + glaciogenic	3.3	+35 + glaciogenic	18	82
Takemura et al. (2009)	Vegetation + shelf	2.4	n/a	60	40

Percent changes in emissions are expressed as percent increase at the LGM relative to today.

the impact of dust particles on the radiative balance depends on the optical properties of dust particles as well as the underlying albedo of clouds or the land surface. Dust can either increase or decrease planetary albedo, or surface reflectivity, depending on both the single scattering albedo of the dust particle as well as the albedo of the underlying surface. Dust can have a cooling effect over optically dark surfaces such as forests and ocean by reflecting incoming solar radiation to space and reducing radiative fluxes at the surface. In contrast, over highly reflective surfaces such as ice sheets, deserts, and snow, dust can increase the absorption of solar radiation and decrease snow surface albedo (e.g., Clarke and Noone, 2007; Wagenbach et al., 1996). For example, the incorporation of dust into surface snow layers can decrease the surface albedo by half (Aoki et al., 2006). To complicate matters further, dust particles may also alter the radiative budget via indirect effects such as providing condensation nuclei for ice cloud formation (e.g., Andreae and Rosenfeld, 2008; Rosenfeld et al., 2001; Su et al., 2008; Yin and Chen, 2007).

Higher atmospheric dust loadings during the last Ice Age are likely to have impacted the planet's radiative budget as well as surface temperatures. Three studies have estimated changes in the impact of glacial dust on radiative forcing during the LGM (Claquin et al., 2003; Mahowald et al., 2006b; Takemura et al., 2009). All three have suggested that the overall global impact of dust ranged from -0.88 to -1.04 W m^{-2} at the top of the atmosphere. Placed in perspective, this change is small relative to local radiative cooling estimated for Northern Hemisphere ice sheets at the LGM (i.e., -20 W m^{-2} ; Hewitt and Mitchell, 1998), but of the same order or magnitude as that estimated for modern radiative forcing associated with all anthropogenic activities (i.e., $+1.6 \text{ W m}^{-2}$; IPCC, 2007). The largest impacts were observed in the tropics in all studies. Claquin et al. (2003) estimated that the negative radiative forcing due to dust reached values of -2.2 to -3.2 W m^{-2} while Mahowald et al. (2006b) saw values as great as -4 W m^{-2} at the top of the atmosphere. Takemura et al. (2009) suggest that the direct radiative impacts of glacial dust were close to zero at the tropopause and only achieve a -0.9 W m^{-2} decrease in the radiative forcing when indirect effects of dust as ice cloud condensation nuclei are included. Although these studies produce different spatial patterns of radiative forcing and for different reasons, all three studies suggest that the radiative feedbacks associated with enhanced atmospheric dust loading were likely to contribute to a cooler climate during the LGM. The globally averaged response of the radiative forcing due to

glacial dust was a reduction in surface temperatures by 0.85°C (Mahowald et al., 2006b).

Sea-Dust as a Fertilizer

In addition to their impact on the global radiative budget, dust particles contain materials such as iron, phosphorus, and silica that are essential to the productivity of marine ecosystems. In many parts of the ocean, the critical concentrations of dissolved iron are very low relative to other necessary nutrients for phytoplankton production. Dust is frequently considered to contain approximately 3.5 wt% of iron on average (Duce, 1986; Duce and Tindale, 1991; Fung et al., 2000; Taylor and McClennan, 1985; Zhu et al., 1997), although this number can vary by region, material type, and particle size (Baker and Jickells, 2006; Muhs et al., 2010b). Nevertheless, additional inputs of iron via atmospheric dust could lead to an increase in phytoplankton activity and growth, and in some cases increased fixation and storage of carbon in the ocean. This hypothesis has been the subject of extensive modeling work, data studies, and multiple large-scale field experiments (de Baar et al., 2005). In fact, the impacts of sprinkling iron on the Southern Ocean as part of one of these experiments was visible from space satellites (Figure 5). Given that atmospheric dust deposition was two to five times greater during glacial periods, Martin (1990) suggested that the effect of iron fertilization from enhanced dust deposition to the oceans during Ice Ages would be large enough to reduce atmospheric carbon dioxide concentrations to their glacial levels. Indeed, modeling and observational studies that have focused on the impact of dust during the last glacial period have found that the feedback of dust upon atmospheric CO₂ is likely to contribute in the range of 10–40 ppm to the glacial–interglacial change in CO₂, with a best estimate of 15 ppm (Kohfeld and Ridgwell, 2009). However, sensitivity studies investigating the impact of aeolian iron inputs on carbon sequestration in the ocean have suggested that even a 50% reduction in aeolian inputs to the ocean would have a limited effect, increasing atmospheric CO₂ by only 14 ppm (Parekh et al., 2006). A fivefold increase in aeolian iron input would decrease atmospheric CO₂ by approximately the same amount (Mahowald et al., 2009). At the same time, dust inputs can have large regional importance to biogeochemical cycling within the ocean that is particularly dependent on intermittent aeolian sources of soluble iron. For example, Crusius et al. (2011) estimate that local contributions

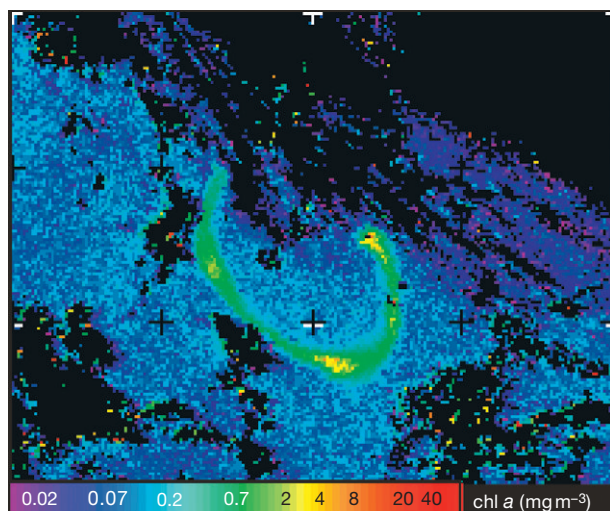


Figure 5 Dust can be a source of nutrients such as silica and iron. In many parts of the ocean, the critical concentrations of dissolved iron are very low relative to other necessary nutrients for phytoplankton production. As a result, we find many parts of the ocean where additional aeolian inputs of iron via dust would lead to an increase in phytoplankton activity and growth (Boyd et al., 2007), and in some cases carbon export (Buesseler et al., 2004; Pollard et al., 2009). Ocean color satellite (SeaWiFS) image of surface ocean chlorophyll concentrations, taken approximately 6 weeks after iron was released in the Southern Ocean. The patch of green affected by the iron is ~ 100 km across. (SeaWiFS data provided by the NASA DAAC/GSFC and copyright of Orbital Imaging Corps and the NASA SeaWiFS project, and processed at CCMS-PML.)

of soluble iron from one glacial flour dust storm event to the Gulf of Alaska was 60–400 tons, comparable in magnitude to annual iron sources from coastal eddies. The impact of this previously unrecognized source of iron deposition from dust, particularly on local primary producers, remains to be determined. These results taken together suggest not only that aeolian inputs have a substantial impact on the ocean carbon cycle, but also that our current knowledge of the sensitivity of the ocean carbon cycle to iron from dust requires further analysis, particularly for understanding regional influences of dust upon marine ecosystems.

Implications for Future Changes in the Dust Cycle

One might also anticipate that future changes in climate attributable to human activities are likely to perturb dust emissions, although the climate-related changes in future atmospheric dust loadings are much less certain. Several previous studies have predicted that dust emissions may decrease by 20–60% from today's levels by 2090 while others suggest that emissions may increase by as much as 200% (Mahowald et al., 2006a; Tegen et al., 2004; Woodward et al., 2005). Analysis of simulations of source areas from seventeen different models shows that they predict increases and decreases of up to 50% in desert area (Mahowald, 2007), emphasizing that our predictions of changes in desert areas (and therefore ultimately dust emissions) are rather uncertain.

Our estimations of the global and regional radiative impacts of dust also contain large uncertainties. The IPCC (2007) has recently provided a best estimate of -0.1 W m^{-2} for the direct radiative effect of anthropogenic dust, with values ranging from -0.3 to $+0.1 \text{ W m}^{-2}$. This range of values appears considerably smaller than estimates for the Third Assessment Report in which values ranged from -0.6 to $+0.4 \text{ W m}^{-2}$ (Houghton et al., 2001). Between the publication of these two reports, new constraints have been placed not only on the estimated magnitude of anthropogenic emissions (dropping the value from 50% to 20% of the global loading), but also on estimates of the optical properties of dust. In fact, when one examines the summed natural and anthropogenic radiative forcing for dust, our newest computed values range from -1.5 to $+0.5 \text{ W m}^{-2}$. This range, although slightly more negative, is not much different from the values published by the Third Assessment Report (-1.3 to $+0.8 \text{ W m}^{-2}$). Regionally, the radiative impact can be two orders of magnitude larger, more than 100 W m^{-2} . Temporal changes in the anthropogenic contribution to desert dust in the atmosphere are equally difficult to quantify, but recent attempts combining observations and modeling suggest that changes in anthropogenic desert dust have reduced the direct radiative forcing at the top of the atmosphere by $-0.14 \pm 0.11 \text{ W m}^{-2}$ over the twentieth century (Mahowald et al., 2010). Nevertheless, our uncertainty about the overall radiative forcing of dust remains large due to the paucity of data sets, uncertainties in the optical properties of mineral aerosols, and our ability to quantify indirect effects.

Estimating the future impact of dust on iron fertilization in the ocean is faced with some of the same uncertainties encountered when estimating the future emissions and radiative impacts of atmospheric dust. For example, modeled estimates of future dust emissions resulting from climate-induced changes in winds and vegetation predict both positive to negative changes (Mahowald et al., 2006a,b; Tegen et al., 2004; Woodward et al., 2005). Understanding the processes controlling the solubility and bioavailability of iron is currently a field of extensive study, and several aspects of the iron cycle are greatly simplified in model experiments. Iron solubility, for example, is normally expressed as a constant 3.5%, whereas measured solubilities of atmospheric iron range from less than 1% to 81% (Baker et al., 2006a,b; Johansen et al., 2000; Schroth et al., 2009; Siefert et al., 1999). Furthermore, anthropogenic emissions of iron from combustion sources tend to increase the solubility. Confounding this impact even further is the impact of climate changes on ocean circulation, which may affect the ability of regions of the ocean to utilize iron (for review see, e.g., Mahowald et al., 2009).

Conclusion

In conclusion, several advances have been made in our understanding of the dust cycle over the past few decades. Our ability to model dust within the Earth system has greatly improved, and our quantification of changes in dust during the last glacial cycle has also led to understandings of some of the first-order controls on dust at a global scale. We have a reasonable understanding of the first-order contribution of iron fertilization of dust to changes in atmospheric CO_2 on glacial–interglacial timescales.

Several challenges remain. Our understanding of the radiative impact of dust today remains uncertain, hindered by uncertainties in our understanding of the optical properties of dust as well as our quantification of emissions and the anthropogenic contribution to the atmosphere dust loading. These uncertainties increase as we attempt to predict what the future changes in dust emissions, atmospheric dust loadings, and dust impacts on global biogeochemical systems are likely to be.

See also: Eolian Records, Deep-Sea Sediments. **Glacial Climates:** Biosphere Feedbacks; Effects of Atmospheric Dust. **Introduction:** Understanding Quaternary Climatic Change. **Loess Records:** Central Asia; China; Europe; North America; South America. **Paleoclimate Modeling:** Data–Model Comparisons; Last Glacial Maximum GCMs; Quaternary Environments. **Sea Level Studies:** Eustatic Sea-Level Changes – Glacial–Interglacial Cycles; Eustatic Sea-Level Changes Since the Last Glacial Maximum.

References

- An ZS, Kutzbach JE, Prell WL, and Porter SC (2001) Evolution of the Asian monsoons and phased uplift of the Himalaya–Tibetan plateau since Late Miocene times. *Nature* 411: 62–66.
- Andersen KK, Armengaud A, and Genthon C (1998) Atmospheric dust under glacial and interglacial conditions. *Geophysical Research Letters* 25: 2281–2284.
- Andreae MO and Rosenfeld D (2008) Aerosol–cloud–precipitation interactions. Part 1: the nature and sources of cloud-active aerosols. *Earth-Science Reviews* 89: 13–41.
- Aoki T, Motoyoshi H, Kodama Y, Yasunari TJ, Sugiura K, and Kobayashi H (2006) Atmospheric aerosol deposition on snow surfaces and its effect on albedo. *Scientific Online Letters on the Atmosphere* 2: 13–16.
- Baker A, French M, and Linge K (2006a) Trends in aerosol nutrient solubility along a west–east transect of the Saharan dust plume. *Geophysical Research Letters* 33: L07805.
- Baker A, Jickells T, and Linge K (2006b) Trends in the solubility of iron, aluminum, manganese and phosphorus in aerosol collected over the Atlantic Ocean. *Marine Chemistry* 98: 43–58.
- Baker A and Jickells T (2006) Mineral particle size as a control on aerosol iron solubility. *Geophysical Research Letters* 33: L17608.
- Ballantyne AP, Lavine M, Crowley TJ, Liu J, and Baker PB (2005) Meta-analysis of tropical surface temperatures during the Last Glacial Maximum. *Geophysical Research Letters* 32: <http://dx.doi.org/10.1029/2004GL021217>.
- Basile I, Grousset FE, Revel M, Petit JR, Biscaye PE, and Barkov NI (1997) Patagonian origin of glacial dust deposited in East Antarctica (Vostok and Dome C) during glacial stages 2, 4 and 6. *Earth and Planetary Science Letters* 146: 573–589.
- Bigler M, Rothlisberger R, Lambert F, Stocker TF, and Wagenbach D (2006) Aerosol deposited in East Antarctica over the last glacial cycle: Detailed apportionment of continental and sea-salt contributions. *Journal of Geophysical Research* 111: <http://dx.doi.org/10.1029/2005JD006469>.
- Boyd PW, Law CS, Wong CS, et al. (2004) The decline and fate of an iron-induced subarctic phytoplankton bloom. *Nature* 428: 549–553.
- Buesseler KO, Andrews JE, Pike SM, and Charette MA (2004) The effects of iron fertilization on carbon sequestration in the Southern Ocean. *Science* 304: 414–417.
- Cakmur RV, Miller RL, Perlwitz J, et al. (2006) Constraining the global dust emission and load by minimizing the difference between the model and observations. *Journal of Geophysical Research–Atmospheres* 111: D06207.
- Chadwick AO and Davis JO (1990) Soil-forming intervals caused by eolian sediment pulses in the Lahontan Basin, northwestern Nevada. *Geology* 18: 243–246.
- Chadwick OA, Derry LA, Vitousek PM, Huebert BJ, and Hedin LO (1999) Changing sources of nutrients during four million years of ecosystem development. *Nature* 397: 491–497.
- Claquin T, Roelandt C, Kohfeld KE, et al. (2003) Radiative forcing of climate by ice-age atmospheric dust. *Climate Dynamics* 20: 193–202. <http://dx.doi.org/10.1007/s00382-002-0269-1>.
- Clarke AD and Noone KJ (2007) Soot in the arctic snowpack: A cause for perturbations in radiative transfer. *Atmospheric Environment* 41: 64–72.
- Clemens SC and Prell WL (1990) Late Pleistocene variability of Arabian Sea summer monsoon winds and continental aridity: Eolian records from the lithogenic component of deep-sea sediments. *Paleoceanography* 5: 109–145.
- Cole JM, Goldstein SL, deMenocal PB, Hemming SR, and Grousset FE (2009) Contrasting compositions of Saharan dust in the eastern Atlantic Ocean during the last deglaciation and African Humid Period. *Earth and Planetary Science Letters* 278: 257–266.
- Crusius J, Schroth AW, Gassó S, Moy CM, Levy RC, and Gatica M (2011) Glacial flour dust storms in the Gulf of Alaska: Hydrologic and meteorological controls and their importance as a source of bioavailable iron. *Geophysical Research Letters* 38: <http://dx.doi.org/10.1029/2010GL046573>.
- De Angelis M, Steffensen JP, Legrand MR, Clausen HB, and Hammer CU (1997) Primary aerosol (sea salt and soil dust) deposited in Greenland ice during the last climatic cycle: Comparison with east Antarctic records. *Journal of Geophysical Research* 102: 26681–26698.
- de Baar H, Boyd P, Coale K, et al. (2005) Synthesis of iron fertilization experiments: From the iron age in the age of enlightenment. *Journal of Geophysical Research – Oceans* 110: Art. No. C09S16.
- Delmonte B, Basile-Doelsch I, Petit JR, et al. (2004) Comparing the Epica and Vostok dust records during the last 220,000 years: Stratigraphical correlation and provenance in glacial periods. *Earth-Science Reviews* 66: 63–87.
- DeMenocal PB (2004) African climate change and faunal evolution during the Pliocene–Pleistocene. *Earth and Planetary Science Letters* 220: 2–24.
- Ding Z, Yu Z, Rutter NW, and Liu T (1994) Towards an orbital time scale for Chinese loess deposits. *Quaternary Science Reviews* 13: 39–70.
- Ding ZL, Yu ZW, Yang SL, Sun JM, Xiong SF, and Liu TS (2001) Coeval changes in grain size and sedimentation rate of eolian loess, the Chinese Loess Plateau. *Geophysical Research Letters* 28: 2097–2100.
- Duce RA (1986) The impact of atmospheric nitrogen, phosphorus, and iron species on marine productivity. In: Buat-Menard P (ed.) *The Role of Air–Sea Exchange in Geochemical Cycling*, pp. 497–529. Norwell, Mass: Reidel.
- Duce RA and Tindale NW (1991) Atmospheric transport of iron and its deposition in the ocean. *Limnology and Oceanography* 36: 1715–1726.
- Fung IY, Meyn SK, Tegen I, Doney SC, John JG, and Bishop JKB (2000) Iron supply and demand in the upper ocean. *Global Biogeochemical Cycles* 14: 281–295.
- Gabrielli P, Wegner A, Petit JR, et al. (2010) A major glacial–interglacial change in aeolian dust composition inferred from Rare Earth Elements in Antarctic ice. *Quaternary Science Reviews* 29: 265–273.
- Ginge FX, De Deckker P, and Hillenbrand C-D (2004) Late Quaternary terrigenous sediments from the Murray Canyons area, offshore South Australia and their implications for sea level change, palaeoclimate and palaeodrainage of the Murray–Darling Basin. *Marine Geology* 212: 183–197.
- Grousset FE, Biscaye PE, Ravel M, et al. (1992) Antarctic (Dome C) ice-core dust at 18 k.y. B. P.: Isotopic constraints on origins. *Earth and Planetary Science Letters* 111: 175–182.
- Grousset FE, Parra M, Bory A, et al. (1998) Saharan wind regimes traced by Sr–Nd isotopic composition of subtropical Atlantic sediments: Last Glacial Maximum vs. today. *Quaternary Science Reviews* 17: 395–409.
- Guo ZT, Ruddiman WF, Hao QZ, et al. (2002) Onset of Asian desertification by 22 Myr ago inferred from loess deposits in China. *Nature* 416: 159–163.
- Hanebuth TJJ and Lantsch H (2008) A Late Quaternary sedimentary shelf system under hyperarid conditions: Unravelling climatic, oceanographic and sea-level controls (Golfe d’Arguin, Mauritania, NW Africa). *Marine Geology* 256: 77–89.
- Harrison SP, Kohfeld KE, Roelandt C, and Claquin T (2001) The role of dust in climate changes today, at the last glacial maximum and in the future. *Earth-Science Reviews* 54: 43–80.
- Hewitt CD and Mitchell JFB (1998) A fully coupled GCM simulation of the climate of the mid-Holocene. *Geophysical Research Letters* 25: 361–364.
- Houghton J, Ding Y, Griggs DJ, et al. (2001) *Climate Change 2001: The Scientific Basis. Contribution of Working Group I to the Third Assessment Report of the Intergovernmental Panel on Climate Change*. Cambridge, United Kingdom and New York, USA: Cambridge University Press.
- Hovan SA, Rea DK, and Pisias NG (1991) Late Pleistocene continental climate and oceanic variability recorded in Northwest Pacific sediments. *Paleoceanography* 6: 349–370.
- IPCC (2007) *Climate Change 2007: The Physical Science Basis*. In: Solomon S, Qin D, Manning M, Chen Z, Marquis M, Averyt KB, Tignor M, and Miller HL (eds.) *Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change*. Cambridge, UK and New York, NY, USA: Cambridge University Press.

- Jickells TD, An ZS, Andersen KK, et al. (2005) Global iron connections between desert dust, ocean biogeochemistry, and climate. *Science* 308: 67–71.
- Johansen A, Siefert R, and Hoffmann M (2000) Chemical composition of aerosols collected over the tropical North Atlantic Ocean. *Journal of Geophysical Research* 105: 15277–15312.
- Karimi A, Frechen M, Khademi H, Kehl M, and Jalalian A (2011) Chronostratigraphy of loess deposits in northeast Iran. *Quaternary International* 234: 124–132.
- Kohfeld KE and Harrison SP (2001) DIRTMAP: The geologic record of dust. *Earth-Science Reviews* 54: 81–114.
- Kohfeld KE and Ridgwell A (2009) Glacial-interglacial variability in atmospheric CO₂. In: Le Quere C and Saltzman ES (eds.) *The Surface Ocean – Lower Atmosphere Processes, Geophysical Monograph Series*. AGU: Washington, DC. <http://dx.doi.org/10.1029/2008GM000845>.
- Kohfeld KE and Tegen I (2007) The record of soil dust aerosols and their role in the earth system. In: Holland HD and Turekian KK (eds.) *Treatise on Geochemistry*, pp. 1–26. Elsevier Ltd: Oxford.
- Lambert F, Delmonte B, Petit JR, et al. (2008) Dust-climate couplings over the past 800,000 years from the EPICA Dome C ice core. *Nature* 452: 616–619.
- Mahowald NM (2007) Anthropocene changes in desert area: Sensitivity to climate model predictions. *Geophysical Research Letters* 34: <http://dx.doi.org/10.1029/2007GL030472>.
- Mahowald N, Engelstaedter S, Luo C, et al. (2009) Atmospheric iron deposition: Global distribution, variability and human perturbations. *Annual Reviews of Marine Sciences* 1: 245–278.
- Mahowald NM, Kloster S, Engelstaedter S, et al. (2010) Observed 20th century desert dust variability: Impact on climate and biogeochemistry. *Atmospheric Chemistry and Physics* 10: 10875–10893.
- Mahowald N, Kohfeld KE, Hansson M, et al. (1999) Dust sources and deposition during the Last Glacial Maximum and current climate: A comparison of model results with palaeodata from ice cores and marine sediments. *Journal of Geophysical Research* 104: 15,895–15,916.
- Mahowald NM, Muhs DR, Levis S, et al. (2006a) Change in atmospheric mineral aerosols in response to climate: Last glacial period pre-industrial, modern and doubled carbon dioxide climates. *Journal of Geophysical Research-Atmospheres* 111: <http://dx.doi.org/10.1029/2005JD006653>.
- Mahowald NM, Yoshioka M, Collins WD, Conley AJ, Filmore DW, and Coleman DB (2006b) Climate response and radiative forcing from mineral aerosols during the last glacial maximum, pre-industrial, current and doubled-carbon dioxide climates. *Geophysical Research Letters* 33: <http://dx.doi.org/10.1029/2006GL026126>.
- Martin J (1990) Glacial-interglacial CO₂ change: The iron hypothesis. *Paleoceanography* 5: 1–13.
- Martinez-Garcia A, Rosell A, Geibert W, et al. (2009) Links between iron supply, marine productivity, sea surface temperature, and CO₂ over the last 1.1 Ma. *Paleoceanography* 24: PA1207.
- McGee D, Broecker WS, and Winckler G (2010) Gustiness: The driver of glacial dustiness? *Quaternary Science Reviews* 29: 2340–2350.
- MARGO Project Members (2009) Constraints on the magnitude and patterns of ocean cooling at the Last Glacial Maximum. *Nature Geoscience* 2: 127–132.
- Muhs DR (2001) Evolution of soils on Quaternary reef terraces of Barbados, West Indies. *Quaternary Research* 56: 66–78.
- Muhs DR, Ager TA, Bettis EA III, et al. (2003) Stratigraphy and palaeoclimatic significance of Late Quaternary loess–palaeosol sequences of the Last Interglacial–Glacial cycle in central Alaska. *Quaternary Science Reviews* 22: 1947–1986.
- Muhs DR, Budahn J, Avila A, Skipp G, Freeman J, and Patterson D (2010a) The role of African dust in the formation of Quaternary soils on Mallorca, Spain and implications for the genesis of Red Mediterranean soils. *Quaternary Science Reviews* 29: 2518e2543.
- Muhs DR, Budahn J, Skipp G, Prospero JM, Patterson D, and Bettis EA III (2010b) Geochemical and mineralogical evidence for Sahara and Sahel dust additions to Quaternary soils on Lanzarote, eastern Canary Islands, Spain. *Terra Nova* 22: 399–410.
- Muhs D, Budahn J, Johnson D, et al. (2008) Geochemical evidence for airborne dust additions to soils in Channel Islands National Park, California. *Geological Society of America Bulletin* 120: 106–126.
- Muhs DR, Bush CA, Stewart KC, Rowland TR, and Crittenden RC (1990) Geochemical evidence of Saharan dust parent material for soils developed on Quaternary limestones of Caribbean and western Atlantic Islands. *Quaternary Research* 33: 157–177.
- Parekh P, Dutkiewicz S, Follows MJ, and Ito T (2006) Atmospheric carbon dioxide in a less dusty world. *Geophysical Research Letters* 33: L03610.
- Peltier WR and Fairbanks RG (2006) Global glacial ice volume and Last Glacial Maximum duration from an extended Barbados sea level record. *Quaternary Science Reviews* 25: 3322–3337.
- Pollard RT, Salter I, Sanders RJ, et al. (2009) Southern Ocean deep-water carbon export enhanced by natural iron fertilization. *Nature* 457: 577–U81.
- Prentice IC, Jolly D, and BIOME 6000 participants (2000) Mid-Holocene and glacial maximum vegetation geography of the northern continents and Africa. *Journal of Biogeography* 27: 507–519.
- Rea DK (1994) The paleoclimatic record provided by eolian deposition in the deep sea: The geologic history of wind. *Reviews of Geophysics* 32: 159–195.
- Reader MC, Fung I, and McFarlane N (1999) The mineral dust aerosol cycle during the Last Glacial Maximum. *Journal of Geophysical Research-Atmospheres* 104: 9381–9398.
- Rosenfeld D, Rudich Y, and Lahav R (2001) Desert dust suppressing precipitation: A possible desertification feedback loop. *Proceedings of the National Academy of Sciences of the United States of America* 98: 5975–5980.
- Rousseau D-D, Antoine P, Gerasimenko N, et al. (2011) North Atlantic abrupt climatic events of the last glacial period recorded in Ukrainian loess deposits. *Climate of the Past* 7: 221–234.
- Ruddiman WF (1997) Tropical Atlantic terrigenous fluxes since 25,000 yrs B.P. *Marine Geology* 136: 189–207.
- Sarnthein M, Tetzlaff G, Koopmann B, Wolter K, and Pflaumann U (1981) Glacial and interglacial wind regimes over the eastern subtropical Atlantic and North-West Africa. *Nature* 293: 193–196.
- Schmidt ED, Frechen M, Murray AS, Tsukamoto S, and Bittmann F (2011) Luminescence chronology of the loess record from the Tönchesberg section: A comparison of using quartz and feldspar as dosimeter to extend the age range beyond the Eemian. *Quaternary International* 234: 10–22.
- Schroth AW, Crusius J, Sholkovitz ER, and Bostick BC (2009) Iron solubility driven by speciation in dust sources to the ocean. *Nature Geoscience* 2: 337–340.
- Shao Y, Wyrwoll K-H, Chappell A, et al. (2011) Dust cycle: An emerging core theme in Earth system science. *Aeolian Research* 2: 181–204.
- Siefert R, Johansen A, and Hoffman M (1999) Chemical characterization of ambient aerosol collected during the southwest monsoon and intermonsoon seasons over the Arabian Sea: Labile-Fe(II) and other trace metals. *Journal of Geophysical Research* 104: 3511–3526.
- Stancin A, Gleason J, Hovan S, et al. (2008) Miocene to recent eolian dust record from the Southwest Pacific Ocean at 40 degrees S latitude. *Paleogeography, Palaeoclimatology, Palaeoecology* 261: 218–233.
- Su J, Huang J, Fu Q, Minnis P, Ge J, and Bi J (2008) Estimation of Asian dust aerosol effect on cloud radiation forcing using Fu-Liou radiative model and CERES measurements. *Atmospheric Chemistry and Physics* 8: 2763–2771.
- Sun Y and An ZS (2005) Late Pliocene-Pleistocene changes in mass accumulation rates of eolian deposits on the central Chinese Loess Plateau. *Journal of Geophysical Research* 110(D23101) <http://dx.doi.org/10.1029/2005JD006064>.
- Swap R, Garstang M, Greco S, Talbot R, and Kallberg P (1992) Saharan dust in the Amazon Basin. *Tellus, Series-B* 144B: 133–149.
- Takemura T, Egashira M, Matsuzawa K, Ichijo H, Oishi R, and Abe-Ouchi A (2009) A simulation of the global distribution and radiative forcing of soil dust aerosols at the Last Glacial Maximum. *Atmospheric Chemistry and Physics* 9: 3061–3073.
- Taylor SR and McLennan SM (1985) *The Continental Crust: Its Composition and Evolution*. Cambridge, MA: Blackwell Science.
- Tegen I, Harrison S, Kohfeld K, Prentice I, Coe M, and Heimann M (2002) Impact of vegetation and preferential source areas on the dust aerosol cycle. *Journal of Geophysical Research* 107: 4576. <http://dx.doi.org/10.1029/2001JD000963>.
- Tegen I, Werner M, Harrison SP, and Kohfeld KE (2004) Relative importance of climate and land use in determining present and future global soil dust emission. *Geophysical Research Letters* 31: <http://dx.doi.org/10.1029/2003GL019216>.
- Tiedemann R, Sarnthein M, and Shackleton NJ (1994) Astronomic timescale for the Pliocene Atlantic $\delta^{18}\text{O}$ and dust flux records of Ocean Drilling Program site 659. *Paleoceanography* 9(4): 619–638.
- Tjallingii R, Claussen M, Stuut J-BW, et al. (2008) Coherent high- and low-latitude control of the northwest African hydrological balance. *Nature Geoscience* 1: 670–675.
- Tófaló O, Orgeira MJ, Compagnucci R, Alonso MS, and Ramos A (2011) Characterization of a loess-paleosols section including a new record of the last interglacial stage in Pampean plain, Argentina. *Journal of South American Earth Sciences* 31: 81–92.
- Újvári G, Kovács J, Varga G, Raucsik B, and Markovic SB (2010) Dust flux estimates for the Last Glacial Period in East Central Europe based on terrestrial records of loess deposits: A review. *Quaternary Science Reviews* 29: 3157–3166.
- Vitousek PM, Kennedy MJ, Derry LA, and Chadwick OA (1999) Weathering versus atmospheric sources of strontium in ecosystems on young volcanic soils. *Oecologia* 121: 255–259.

- Wagenbach D, Preunkert S, Schafer J, Jung W, and Tomadin L (1996) Northward transport of Saharan dust recorded in a deep Alpine ice core. In: Guerzoni S and Chester R (eds.) *The impact of African dust across the Mediterranean*, pp. 291–300. Dordrecht: Kluwer Academic Publishers.
- Walter HJ, Hegner E, Diekmann B, Kuhn G, and Rutgers van der Loeff MM (2000) Provenance and transport of terrigenous sediment in the South Atlantic Ocean and their relations to glacial and interglacial cycles: Nd and Sr isotopic evidence. *Geochimica et Cosmochimica Acta* 64: 3813–3827.
- Werner M, Tegen I, Harrison SP, et al. (2003) Seasonal and interannual variability of the mineral dust cycle under present and glacial climate conditions. *Journal of Geophysical Research* 108: <http://dx.doi.org/10.1029/2002JD002365>.
- Winckler G, Anderson RF, Fleisher MQ, McGee D, and Mahowald N (2008) Covariant Glacial-Interglacial Dust Fluxes in the Equatorial Pacific and Antarctica. *Science* 320: 93–96.
- Woodward S, Roberts DL, and Betts RA (2005) A simulation of the effect of climate change-induced desertification on mineral dust aerosol. *Geophysical Research Letters* 32: <http://dx.doi.org/10.1029/2005GL023482>.
- Yin Y and Chen L (2007) The effects of heating by transported dust layers on cloud and precipitation: A numerical study. *Atmos. Chem. Phys.* 7: 3497–3505.
- Yu G, Chen XD, Ni J, et al. (2000) Palaeovegetation of China: A pollen-data-based synthesis for the mid-Holocene and the last glacial maximum. *Journal of Biogeography* 27: 635–664.
- Zech W, Zech R, Zech M, et al. (2011) Obliquity forcing of Quaternary glaciation and environmental changes in NE Siberia. *Quaternary International* 234: 133–145.
- Zhu XR, Prospero JM, and Millero FJ (1997) Diel variability of soluble Fe(II) and soluble total Fe in North African dust in the trade winds at Barbados. *Journal of Geophysical Research* 102: 21297–21305.

Patterned Ground *see* **Permafrost and Periglacial Features:** Patterned Ground

Peat Studies *see* **Plant Macrofossil Methods and Studies:** Mire and Peat Macros

Pedostratigraphy *see* **Quaternary Stratigraphy:** Pedostratigraphy

PERMAFROST AND PERIGLACIAL FEATURES

Contents

Active Layer Processes

Cryoturbation Structures

Ice Wedges and Ice-Wedge Casts

Patterned Ground

Permafrost

Frost Mounds: Active and Relict Forms

Slope Deposits and Forms

Periglacial Fluvial Sediments and Forms

Rock Weathering

Permafrost and Glacier Interactions

Block/Rock Streams

Blockfields (Felsenmeer)

Rock Glaciers and Protalus Forms

Yedoma: Late Pleistocene Ice-Rich Syngenetic Permafrost of Beringia

Paraglacial Geomorphology

Talus Slopes

Thermokarst Topography

Active Layer Processes

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The Active Layer

Meteorological parameters constitute the dominant control over temperature at the ground surface at most locations on Earth. Many of these (e.g., incoming and outgoing radiation, air temperature) follow one or more periodic courses, including daily, annual, and longer (e.g., glacial–interglacial) fluctuations. These fluctuations cause heating and cooling of the ground to depths proportional to the period of fluctuations. In most temperate and cold regions, as well as in areas of high elevation, periodic annual ground temperature transitions

through 0 °C (32 °F) result in seasonal freezing and thawing of the upper ground layer. In areas where the ground remains frozen only during the colder part of the year, this layer is termed ‘seasonally frozen ground.’ In areas underlain by perennially frozen strata (permafrost), the overlying layer subject to seasonal thaw and freezing is called the ‘active layer’ (**Figure 1**). The active layer is of considerable importance in permafrost regions because most geomorphic, hydrological, and biological processes occur in it. Changes in its thickness induced by human action or climatic change can have profound consequences for ecological systems and human infrastructure.

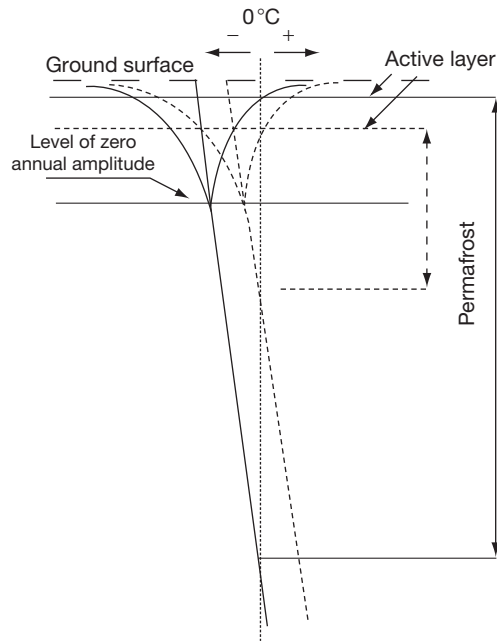


Figure 1 Ground temperature ‘envelope’ in a typical contemporary permafrost region (solid lines). Higher temperatures are to the right and lower to the left; 0 °C is represented by a vertical dashed line. The thermal profile in summer forms the right-hand component of the ‘trumpet curve’; its intersection with the 0 °C line defines the thickness of the active layer. The depth of zero annual amplitude is the level below which no significant season temperature fluctuations are registered. The dashed trumpet curve depicts changes in the thermal profile introduced by a simple increase in the mean annual temperature at the ground surface. Prominent among these changes are thickening of the active layer and a decrease in the thickness of permafrost.

Definition

The term ‘active layer’ was first proposed by Sumgin, who defined it as “the layer of seasonal thaw in permafrost areas” (Sumgin, 1927, p. 56). The term was later introduced to English scientific literature by Muller (1947). The International Permafrost Association’s *Multi-Language Glossary of Permafrost and Related Ground Ice Terms* (van Everdingen, 1998) defines active layer as “the layer of ground that is subject to annual thawing and freezing in areas underlain by permafrost.” Conceptually, the definition of the active layer is straightforward, although its precise meaning is a subject of some scientific debate. The main source of contention is the meaning of the term ‘frozen ground,’ which can indicate either solidification by ice or cooling below 0 °C (Burn, 1998). Although the distinction between these meanings seems minor, it can result in major differences in describing the overall permafrost system. For example, the ice-nucleation temperature in soils can be depressed due to salinity, capillary effects, and pressure, causing ground to remain unfrozen at temperatures below 0 °C. Because permafrost is defined as ground that remains below 0 °C for 2 or more consecutive years, the bottom portion of the active layer defined on the basis of phase changes of water can be considered as a part of the permafrost layer. Alternatively, the temperature-based definition of the active layer clearly separates it from permafrost. Sumgin’s (1927) original definition did not

differentiate between these two meanings. In contrast, Muller (1947) defined the active layer as a “layer of ground above permafrost which thaws in summer and freezes again in winter” (Muller, 1947 p. 213), indicating that the base of the active layer occurs at the depth of maximum seasonal penetration of the 0 °C isotherm into the ground. Many permafrost researchers (e.g., Burn, 1998; Nelson and Hinkel, 2003) prefer the definition based entirely on thermal criteria, without regard to material composition or properties, as it clearly differentiates between the components of the permafrost system and is more easily applicable to field and modeling studies.

The active layer is the only part of the permafrost system that experiences substantial changes in its physical properties during each annual cycle. These changes involve variations of ice/liquid water content, thermal conductivity, density, and mechanical properties, all of which are of critical importance for many natural phenomena and processes. Most geomorphic, biological, and hydrological activities in permafrost-affected soils are confined to the layer of seasonal thaw.

Physical Processes

With the exception of locations where no appreciable moisture is present, seasonal thawing and freezing of the ground is a complicated physical process involving temperature change, phase transition, moisture migration, ground heave and subsidence, and changes in the mechanical, chemical, and thermal properties of the ground. Appreciation of several physical processes is essential for understanding the dynamics of the layer.

Propagation of Thawing and Freezing Fronts

The process of seasonal thawing is initiated when sufficient heat arrives at the ground surface (i.e., when snowmelt is complete) and continues until the end of the warm season, with the onset of refreezing. Although heat transfer occurs primarily by conduction at most locations, nonconductive processes (e.g., coupled heat and moisture flow) can be significant (Hinkel et al., 1997; Outcalt et al., 1990). Convective heat transfer by infiltration of precipitation can also have a significant effect (Hinkel et al., 2001). During autumn freezeback, the active layer cools rapidly to 0 °C and becomes isothermal. Subsequent cooling initiates propagation of the freezing front from the surface downward. One of the main features distinguishing many permafrost environments from the seasonally frozen ground of nonpermafrost regions is the existence of bodies of segregated ice in positions corresponding to the long-term base of the active layer. The frozen ground acts as a heat sink and, depending on the local situation, progressive freezing occurs simultaneously from the surface downward and upward from the base of the active layer (Figure 2). This situation, termed ‘two-sided freezing,’ results in layers of segregated ice just below the surface and at the base of the active layer. Freezing from below is widespread in areas with permafrost temperatures below –5 °C, where it is usually precedes freezing from the surface (Shur, 1988). Two-sided freezing is a highly significant phenomenon because it controls many important physical and geomorphic processes in permafrost environments (Mackay, 1981).

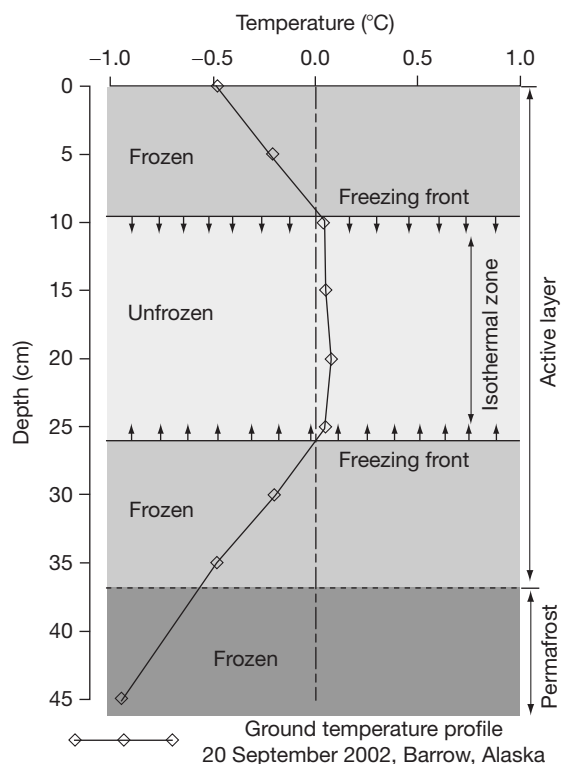


Figure 2 Active-layer temperature profile during freezing as observed on 20 September 2002 in Barrow, Alaska. The mean annual permafrost temperature at the base of the active layer is approximately -8°C . Note the two-sided freezing, indicated by freezing fronts moving downward from the surface and upward from the permafrost table. The maximum thaw depth of 37 cm was reached on 28 August, followed by progressive freezing from the bottom. Freezing from the surface began on 15 September. The position of freezing fronts stabilized at approximately the depths shown on **Figure 3** until 29 October, due to release of the latent heat associated with moisture migration and ice formation at the freezing fronts. Note the isothermal zone with temperatures close to 0°C . Once the isothermal zone became desiccated, the progression of the freezing fronts was very rapid and the entire active layer was frozen to temperatures below -1°C by 3 November.

Zero-Curtain Effect

The permafrost table just below the active layer is relatively impermeable, although the upper layers of permafrost can contain appreciable amounts of unfrozen water (Romanovsky and Osterkamp, 2000), and migration of moisture through the active-layer/permafrost boundary has been documented (Mackay, 1983). Complemented by low evaporation, a characteristic feature of cold regions, this results in accumulation of substantial amounts of water within the active layer, especially in areas of level terrain such as the Arctic coastal plain of northern Alaska. The presence of water significantly modifies heat propagation within the active layer due to latent heat effects associated with transitions between the liquid and solid phases. During freezing, large quantities of heat are released by ice nucleation, whereas during thaw, large amounts of heat are absorbed by the melting of ice. Because release or absorption of latent heat does not result in temperature change, the active layer can remain near 0°C for relatively long periods of time while the water within it

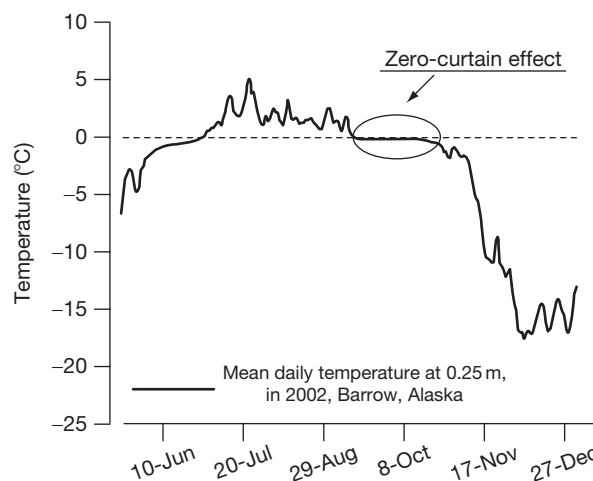


Figure 3 Temperature time series within the isothermal zone of **Figure 2**, at the depth of 25 cm. Note the prolonged period (48 days, 10 September until 28 October) when temperature remained within 0.1°C of the freezing point (zero-curtain effect).

undergoes phase change (**Figure 3**). Sumgin (1927) called this phenomenon the 'zero-curtain effect.' Outcalt et al. (1990) demonstrated that vapor transport and internal distillation can play an important role in forming and maintaining the zero-curtain effect, which can also occur in freezing soils outside permafrost regions. Although it has been defined in various ways, the term 'zero curtain' should be regarded only as an effect; units of measurement should not be assigned to it (Outcalt et al., 1990).

Ice Segregation

If unfrozen water is present in the ground, the process of active-layer freezing is accompanied by water migration toward the freezing front, due to gradients of potential (free energy) between frozen and unfrozen water. This process is sometimes referred to as 'cryosuction.' The process of water migration and subsequent freezing is known as 'ice segregation.' The resulting segregation ice is usually arranged in horizontal layers or lenses oriented perpendicular to the propagation of the freezing fronts, with desiccated layers between. If vertical cracks occur, however, the ice lenses can form within them as the freezing front proceeds, resulting in vertical ice layers. Although ice segregation is common within the seasonally frozen layer of nonpermafrost regions, it is more pronounced under slow upward freezing from the base of the active layer in areas underlain by permafrost. In this case, water migration toward a freezing front coincides with gravitational migration, resulting in a relatively thick layer of ice accumulation at the bottom of the active layer. The amount of segregated ice depends on freezing intensity, prewinter moisture content, the existence of external supplies of water, and soil texture. The latter parameter is a major control over ice segregation, which occurs predominantly in medium-textured soils (silt), due to capillarity. The low permeability of clay-rich sediments limits water movement through the soil and inhibits ice segregation. Although sands and other coarse-grained materials do not contain segregation ice in areas of seasonal freezing (due to their high permeability), in permafrost areas, it can occur as a result of favorable

freezing conditions over long periods of time. Peat is not particularly susceptible to ice segregation, although its unusual thermal properties may enhance the process in underlying fine-grained soils. The vertical distribution of ice lenses is controlled by the composition and distribution of sediments, cracks, and voids within the active layer, as well as by the geometry of the freezing process.

Frost Heave and Thaw Subsidence

Ice segregation and volumetric expansion of water during freezing can result in heaving of the ground surface, a process referred to as 'frost heave.' Subsequent melting of these ice layers in spring causes thaw subsidence (settlement) of the ground surface. Because ice segregation depends on factors that vary spatially and temporarily, frost heave and thaw subsidence are not uniform, and can vary from place to place and from year to year. Frost heave and thaw subsidence can displace buildings, roads, airport runways, railways, pipelines, drainage systems, and other structures (Figure 4). These processes are also of great importance in the development of many of the periglacial geomorphic features of cold regions, including patterned ground, palsas, pingos, seasonal frost mounds, and a wide array of mass-movement features. Techniques for measuring the magnitude of frost heave and thaw subsidence in permafrost regions were reviewed extensively by Tait et al. (2005). Global positioning systems technology appears to have considerable potential for monitoring these processes (Little et al., 2003).

Cryoturbation

Periodic freezing and thawing of soils and the associated processes of ice segregation, frost heave, and thaw subsidence all affect soil structure. Changes in soil structure induced by these processes include enlargement of pores and voids, loosening, aggregation, and sorting of soil particles. The collective term used to describe all soil movements due to frost action is 'cryoturbation' (van Everdingen, 1998). This term is also used to describe irregular structures formed in earth material by frost-action processes, characterized by folded, broken, and dislocated beds and lenses of unconsolidated deposits, including organic horizons and even bedrock.

Active-Layer Thickness

Factors Influencing Active-Layer Thickness

The impacts of annual thaw depend, to a very large degree, on the depth to which it penetrates. The term 'active-layer thickness' (ALT) refers to the thickness of the thawed layer (measured normal to the ground surface) that is reached at the end of the warm season (van Everdingen, 1998). This is distinct from the term 'thaw depth,' which refers to the thickness of the thawed layer at any time during its development in summer. In other words, thaw depth is essentially an instantaneous value that is always less than or equal to the thickness of the fully developed active layer (Nelson and Hinkel, 2003).

Factors influencing the annual thickness of the active layer and its spatial and temporal distribution include



Figure 4 Bicycle trail affected by differential thaw settlement near Fairbanks, Alaska. An increase in ALT, induced by increased absorption of solar radiation by the dark paving material, has penetrated into ice-rich permafrost, causing thaw consolidation and subsidence at the surface. Much of the permafrost near Fairbanks is very close to 0 °C and relatively small disturbances render it susceptible to thaw. Although this example was not induced by climatic warming, it illustrates (a) the type of disturbance that could result under a sustained increase of air temperature and (b) the great geographic variability in thaw settlement possible over even short distances.

1. mean annual surface temperature,
2. the amplitude of the annual temperature cycle at the ground surface, which reflects the degree of climatic continentality,
3. the composition and thermal properties of the subsurface material, and
4. the ground moisture regime.

The 'Stefan Solution,' a simple analytic tool for estimating the depth of thaw, illustrates the roles of these parameters. The Stefan estimate of the depth to which thaw penetrates is given by

$$Z_t = [(2\lambda_t s T t)/(\rho w L)]^{1/2} \quad [1]$$

where Z_t is thaw depth (m); λ_t the substrate's thermal conductivity in the thawed state ($\text{J s}^{-1} \text{m}^{-1} \text{°C}$); T , the mean temperature of the thawing season (°C); t , the length of the thawing season (days); s a scaling factor (s d^{-1}); ρ , soil density

(kg m^{-3}); w , water content (expressed in dimensionless form on a volumetric basis); and L is the volumetric latent heat of fusion (J kg^{-1}). The quantity (Tt), known as the 'thawing index,' is also referred to as 'cumulative degree-days of thawing' (DDT). The thawing index is a time–temperature integral usually computed as the sum of mean daily temperatures above 0°C over the course of the thaw season.

Owing to its neglect of heat capacity and to several other simplifying assumptions, the Stefan Solution usually overestimates the depth of thaw. Numerous improvements to this simple analytic expression (reviewed in Andersland and Ladanyi, 2004) have been developed, but the structure of [1] illustrates clearly the most important factors affecting the depth of thaw. The proportionality between ALT and the square root of the thawing index is demonstrated in Figure 5. The Stefan equation and its variants can also be used to predict the depth to which frost penetrates in regions of seasonally frozen ground. Estimates of thermal conductivity based on a given soil's texture, density, and moisture content are widely available as tables or nomograms (e.g., Andersland and Ladanyi, 2004). More sophisticated numerical models of soil thermal evolution (e.g., Goodrich, 1982, and the many variants of his model) are preferred over analytic expressions when data sufficient to parameterize such models are available.

The composition and texture of subsurface materials determine their thermal conductivity and heat capacity which, in turn, influence the rate and depth of heat propagation into the ground. Fine-grained soils generally have smaller values of thermal conductivity than coarser sediments do. Under similar

conditions, the thickness of the active layer will be greatest in coarse-grained sediments such as sands and gravels. The presence of ice and water close to their transition temperature has a major effect on the thermal properties of soils within the active layer. When unfrozen water changes to ice, its conductivity increases by approximately four times so that frozen soils have higher thermal conductivity than unfrozen soils do. Owing to this effect, known as the 'thermal offset,' under equal absolute values of temperature gradient, the active layer loses more heat in winter than it gains through the summer, and hence permafrost can exist at locations where the mean annual surface temperature is slightly positive. The thermal offset is especially pronounced in natural peat. The heat capacity of unfrozen water is approximately double that of ice, which leads to higher heat capacity in unfrozen soil. Although this effect tends to be smaller than variations of thermal conductivity, significant changes of heat capacity can be associated with the period when phase changes occur. Changes of heat storage in the soil include a large latent heat component over the range of phase-change temperatures, leading to significant increases of heat capacity.

Soil moisture affects the thickness of active layer in two ways: through changes in the thermal properties of soil and through the amount of phase transition, which generally has a more pronounced effect. In general, the greater the amount of water in the ground the greater the amount of heat consumed for phase transition and the thinner the active layer. In some situations, however, this effect can be offset by the increased thermal conductivity of wet soils. The volume fraction of ice and water depend on temperature and specific site conditions.

It should be noted that increase in precipitation can sufficiently alter the ground thermal regime and promote increase in the ALT, even during periods of climatic cooling. For example, marked increases in the ALT and upper permafrost temperatures occurred in the central Lena River basin, Russia in association with abrupt increases in soil moisture following an extremely wet summer of 2005 (Iijima et al., 2010). Summer precipitations influence the active layer in two ways: (1) through increase in thermal conductivity of soils as discussed earlier and (2) through convective heat transfer associated with the percolation of relatively warm meteoric water through the thawed portion of the soil column (e.g., Hinkel and Outcalt, 1994).

Geographic Patterns and Variability of the Active Layer

Zonal trends in ALT are discernable, if similar soils and surface cover are considered at all sample locations. ALT decreases poleward and increases in continental areas because mean annual temperature at the ground surface and its annual amplitude are arranged across the surface of the Earth in broad geographic zones. At local and regional scales, however, ALT is highly variable over space. Variations of subsurface properties, soil moisture, and snow and vegetation cover modify the zonal pattern radically, resulting in a high degree of geographic variability. Topography also plays an important role in determining ALT because it is an important determinant of the amount of heat received by the ground surface, it affects the distribution of snow and vegetation, and it influences the surface and subsurface moisture regimes. The concept of 'topoclimate' (Geiger, 1969) is, therefore, a very important consideration in studies of ALT.

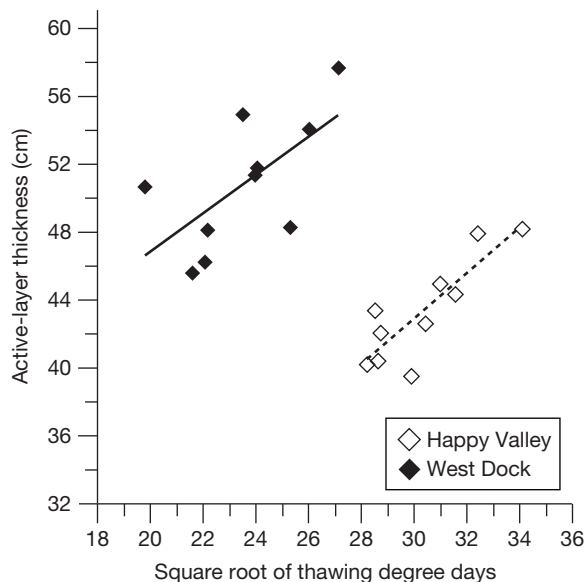


Figure 5 Ten-year record of ALT, plotted against $\sqrt{\text{DDT}}$ at two 1 km^2 CALM sampling locations in northern Alaska. The West Dock location is on the Arctic Coastal Plain physiographic region, an area of low relief and abundant ponds and lakes. Happy Valley is in the Arctic Foothills, with substantial relief and large variations in vegetation and snow cover. Despite large differences in the scale of maximum variability of ALT at these locations (Nelson et al., 1999), the average ALT value from 121 sample points at each location conforms to the structure of the Stefan relation (eqn [1]). Mean ALT values at both locations closely track seasonal air temperature.

The spatial variability of soil and ground moisture, amplified by cryoturbation and ice-segregation processes, can also result in great variability of ALT over small lateral distances. Interannual variations, however, indicate a strong dependence on air temperature (Figure 5), which may be subject to change under sustained climatic change. Estimating and monitoring changes in the thickness of the active layer, a major component of the Arctic terrestrial ecosystem, is of primary importance for most practical applications in engineering and ecology. An international active-layer monitoring program, currently comprising about 200 sites, operates under the name 'Circumpolar Active Layer Monitoring' (CALM) network (Nelson et al., 2008; Shiklomanov et al., 2010).

The Transition Zone

Numerous studies have demonstrated that the genesis and properties of the uppermost permafrost differ from those of the underlying permafrost. The uppermost layer is subject to episodic thaw at subdecadal to multicentennial scales and is referred to as the 'transition zone' (Shur et al., 2005). Its lower bound is the long-term permafrost table and coincides with the top of primary ice wedges. The transition zone is important for permafrost stability and serves as a buffer between the active layer and ice-rich permafrost, especially permafrost containing massive ice. The zone can aggrade and change properties over time through ice segregation (Mackay, 1983) and by infiltration of meteoric water or meltwater from ground ice in the overlying active layer, with subsequent refreezing (Hinkel et al., 2001). The existence of the transition zone explains the fact that wedge ice and other massive ice formations are frequently located well below the active layer (Shur, 1975). Ice veins or rejuvenated ice wedges (secondary or tertiary growth stages) have been observed extending from primary ice wedges upward into the transition zone, indicating reactivation of ice-wedge growth (Lewkowicz, 1994).

The transition zone is ice-rich, with the ice content often much greater than in the permafrost immediately beneath it (e.g., Kanevskiy, 2003; Kokelj and Burn, 2003; Shur, 1988). Ice occurs as segregation lenses, ice veins, and contemporary ice wedges (Lewkowicz, 1994; Shur, 1988). In the continuous permafrost zone, the cryogenic structure of this layer differs significantly from that of the underlying permafrost. In Russian permafrost literature, the characteristic net-like cryogenic structure found in the transition zone is known as 'atactic'; Murton and French (1994) refer to it as 'suspended.' Owing to latent heat effects, the ice-rich transition zone resists thaw and tends to promote thermal stability at greater depths (Shur, 1975; Smith, 1988). Although the transition zone is enriched with ice relative to the active layer, the upper region contains less ice than underlying regions because it thaws more frequently. The uppermost part of the transition zone is referred to as the 'transient layer.' Deep thaw is largely driven by changes in the surface energy equation due to interannual variations and climatic fluctuations, and occurs at the subdecadal to multicentennial scale. During most warm years, the transient layer protects the underlying ice-rich materials from thaw. Deeper in the transition zone, thaw is less frequent and occurs in response to climate change or a major disturbance (Figure 6).

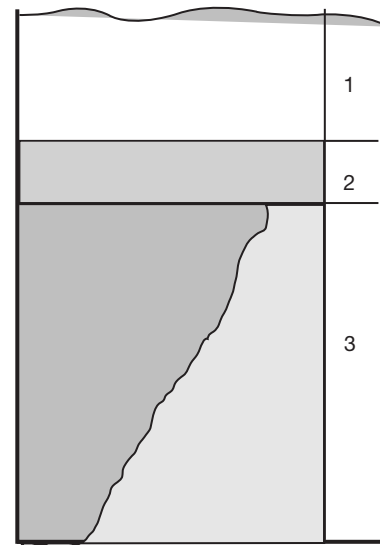


Figure 6 Schematic diagram of a three-layer model of the permafrost-active layer system: (1) active layer (seasonal freezing and thawing), (2) transient layer with high and quasiuniform ice content and freezing/thawing at decadal to century scales, and (3) long-term permafrost (freezing and thawing at century to millennial scales), with spatially inhomogeneous ice content. Shur YI (1988) The upper horizon of permafrost soils. In: Senneset K (ed.) *Proceedings of the Fifth International Conference on Permafrost*, pp. 867–871. Trondheim, Norway: Tapir.

The Active Layer and Climate Change

Roles in Climate-Change Science

Because most geomorphic, biological, and hydrological processes in permafrost regions are restricted to the active layer, long-term changes in its thickness can have important consequences for natural and human systems. As with permafrost (US Arctic Research Commission, 2003), the active layer assumes three important roles in climate-change science: translator, facilitator, and record keeper.

Translator

The active layer translates climatic change to natural and human systems because increases in ALT can lead to thaw settlement, development of thermokarst terrain, and destabilization of human infrastructure at the surface (Nelson et al., 2002). Depending on subsurface conditions, thaw settlement induced by melting ground ice can have very disparate hydrological changes at the surface. Jorgenson et al. (2001) described widespread ecological changes in central Alaska, in which birch forests established on ice-rich 'uplands' have been transformed into extensive wetlands by thaw settlement. In western Alaska, Smith et al. (2005) found that thaw of ice-rich permafrost established subsurface drainage ways, ultimately leading to the draining of large lakes and drying at the surface.

Facilitator

Widespread increases in ALT induced by climatic warming have the potential to facilitate and amplify that warming. Large quantities of organic carbon are known to be stored in near-surface layers of permafrost. A widespread increase in ALT could

mobilize much of this carbon, releasing it to the atmosphere in the form of methane and carbon dioxide and leading to amplified climatic warming. Uncertainties about the amount of carbon sequestered in high-latitude regions, details of hydrological responses to thaw, and the evolution of vegetation preclude precise estimation of the contribution of thawing permafrost to climate change at present (Hobbie et al., 2000; Stokstad, 2004; Sturm et al., 2001).

Record keeper

Permafrost contains a great deal of proxy information about past climatic changes. Cryostratigraphic techniques (French, 1998; Murton, 2001), used in concert with isotopic analysis, provide information about increases in ALT that occurred millennia ago. Such evidence includes truncated ice wedges, ice- and sediment-wedge casts, and thaw unconformities.

Simulation studies indicate that widespread increases in ALT will occur in conjunction with climatic warming (Anisimov et al., 1997; Nelson et al., 2001). One of the most difficult challenges facing permafrost scientists evaluating the impacts of climate change is to differentiate between climate-induced and anthropogenic effects (Hinzman et al., 2005; Nelson, 2003).

Observed ALT Trends

Thaw-depth observations in permafrost regions indicate substantial interannual fluctuations in ALT, primarily in response to variations in summer air temperature (Popova and Shmakin, 2009; Smith et al., 2009). Decadal trends in ALT vary by region. A progressive increase in ALT has been observed in some Nordic countries (Åkerman and Johansson, 2008; Callaghan et al., 2010). However, the increased thaw propagation has stopped during the past 3 (2007–2010) years, as ALT has been more or less stable, mainly due to dry summer conditions (Christiansen et al., 2010). Observations on Svalbard and Greenland show increased ALT since the late 1990s, although the increase is not spatially and temporarily uniform (Christiansen et al., 2010).

Significant permafrost degradation has been reported in the Russian European North. As a result of recent climatic warming, ALT has increased significantly (Mazhitova, 2008; Mazhitova et al., 2008) and permafrost patches that were 10–15 m thick have thawed completely in the Vorkuta area (Oberman, 2007). Permafrost degradation is also evident in the discontinuous permafrost zone of West Siberia (Romanovsky et al., 2010). However, areas of West Siberia underlain by continuous permafrost do not show pronounced ALT trends (Vasiliev et al., 2008). Increased ALT since the mid-1990s has been observed in East Siberia (Fyodorov-Davydov et al., 2008) and Chukotka (Zamolodchikov, 2008). The frequently reported long-term permafrost thawing in Eastern Siberia, Russia, near the city of Yakutsk, is related directly to natural or anthropogenic disturbances such as forest fires, agriculture, and infrastructure development (Federov, 1996; Federov and Konstantinov, 2008) and does not correlate with climate (Romanovsky et al., 2010).

The permafrost areas of Central Asia have experienced a general increase in ALT. On the Qinghai–Tibet Plateau, an increase of 7.5 cm year⁻¹ was reported from 1995 to 2007 in response to an increase in air temperature (Wu and Zhang, 2010). Rates of ALT increase up to 40 cm year⁻¹ over the past

decade were observed at some Mongolian locations with warm permafrost.

ALT trends are different for North American sites. There, a progressive increase of ALT is evident only at sites in the Alaskan Interior, where in 2007, the maximum ALT for the 18-year observation period occurred (Viereck et al., 2008). On the Alaskan North Slope, ALT is relatively stable, lacking pronounced trends over the 1995–2008 period (Streletskiy et al., 2008). Similar results are reported from the Western Canadian Arctic. Smith et al. (2009) found that there was no definite trend in ALT over the last 15 years in the Mackenzie Valley. At several locations, the active layer has thinned after 1998. Results from long-term monitoring in the northern portion of the Mackenzie Valley indicate that there was an increase of 8 cm in thaw depth between 1983 and 2008; but the maximum was in 1998, and since then, thaw depths have generally been shallower (Burn and Kokelj, 2009). In the eastern portion of the Canadian Arctic, ALT has increased since the mid-1990s, with the largest increase occurring in bedrock in the discontinuous permafrost zone (Smith et al., 2010).

Thaw penetration into an ice-rich layer at the base of the active layer is accompanied by loss of volume (thaw consolidation), manifested as subsidence at the ground surface. Results from monitoring of the position of the ground surface at representative locations on the North Slope of Alaska indicate that up to 11 cm in subsidence had occurred over the 2001–2008 period (Streletskiy et al., 2008). Similar results have been obtained from the Russian European North, where increases in ALT caused subsidence of 20 cm over the 1998–2008 period (Mazhitova and Kaverin, 2007). A subsidence of 1–4 cm/decade was estimated by Liu et al. (2010), who have used interferometric synthetic aperture radar to measure surface deformation from 1992 to 2000 over permafrost on the North Slope of Alaska.

Concluding Comments

The active layer is an important component of the permafrost system. Its thermal behavior and thickness constitute an important influence on geomorphic and hydrologic processes, ecosystems, and human activities. Although numerical simulation studies indicate that widespread thickening of the active layer could occur in conjunction with global warming, the existing record of ALT shows diverse regional patterns. Numerous feedback mechanisms involving vegetation succession (Sturm et al., 2001) and hydrological changes (Smith et al., 2005) could act to limit the magnitude of future increases in ALT, even under sustained climatic warming. The monitoring programs implemented by the International Permafrost Association are critical for assessing the evolution of permafrost and the active layer.

See also: **Permafrost and Periglacial Features:** Cryoturbation Structures; Frost Mounds: Active and Relict Forms; Ice Wedges and Ice-Wedge Casts; Paraglacial Geomorphology; Patterned Ground; Permafrost; Slope Deposits and Forms; Thermokarst Topography. **Quaternary Stratigraphy:** Overview.

References

- Åkerman HJ and Johansson M (2008) Thawing permafrost and thicker active layers in sub-arctic Sweden. *Permafrost and Periglacial Processes* 19: 279–292. <http://dx.doi.org/10.1002/ppp.626>.
- Andersland OB and Ladanyi B (2004) *Frozen Ground Engineering*. Hoboken, NJ: Wiley.
- Anisimov OA, Shiklomanov NI, and Nelson FE (1997) Effects of global warming on permafrost and active-layer thickness: Results from transient general circulation models. *Global and Planetary Change* 15: 61–77.
- Burn CR (1998) The active layer: Two contrasting definitions. *Permafrost and Periglacial Processes* 9: 411–416.
- Burn CR and Kokelj SV (2009) The environment and permafrost of the Mackenzie Delta area. *Permafrost and Periglacial Processes* 20(2): 83–105.
- Callaghan TV, Bergholm F, Christensen TR, Jonasson C, Kokfelt U, and Johansson M (2010) A new climate era in the sub-Arctic: Accelerating climate changes and multiple impacts. *Geophysical Research Letters* 37: L14705. <http://dx.doi.org/10.1029/2009GL042064>.
- Christiansen HH, Etzelmüller B, Isaksen K, et al. (2010) The thermal state of permafrost in the Nordic area during the international polar year 2007–2009. *Permafrost and Periglacial Processes* 21: 156–181. <http://dx.doi.org/10.1002/ppp.687>.
- Fedorov AN (1996) Effects of recent climate change on permafrost landscapes in central Sakha. *Polar Geography* 20: 99–108.
- Fedorov AN and Konstantinov PY (2008) Recent changes in ground temperature and the effect on permafrost landscapes in Central Yakutia. In: Kane DL and Hinkel KM (eds.) *Proceedings of the Ninth International Conference on Permafrost*, vol. 1, pp. 433–438. Fairbanks, AK: Institute of Northern Engineering, University of Alaska Fairbanks.
- French HM (1998) An appraisal of cryostratigraphy in north-west arctic Canada. *Permafrost and Periglacial Processes* 9: 297–312.
- Fyodorov-Davydov DG, Kholodov AL, Ostroumov VE, et al. (2008) Seasonal thaw of soils in the north Yakutian ecosystems. In: Kane DL and Hinkel KM (eds.) *Proceedings of the Ninth International Conference on Permafrost*, vol. 1, pp. 481–486. Fairbanks, AK: Institute of Northern Engineering, University of Alaska Fairbanks.
- Geiger R (1969) Topoclimates. In: Flohn N (ed.) *World Survey of Climatology, 2. General Climatology*, pp. 105–138. Amsterdam: Elsevier.
- Goodrich LE (1982) The influence of snow cover on the ground thermal regime. *Canadian Geotechnical Journal* 19: 421–432.
- Hinkel KM and Outcalt SI (1994) Identification of heat transfer processes during soil cooling, freezing, and thaw in central Alaska. *Permafrost and Periglacial Processes* 5(4): 217–235.
- Hinkel KM, Outcalt SI, and Taylor AE (1997) Seasonal patterns of coupled flow in the active layer at three sites in northwest North America. *Canadian Journal of Earth Sciences* 34: 667–678.
- Hinkel KM, Paetzold R, Nelson FE, and Bockheim JG (2001) Patterns of soil temperature and moisture in the active layer and upper permafrost at Barrow, Alaska: 1993–1999. *Global and Planetary Change* 29: 293–309.
- Hinzman L, Bettes N, Bolton WR, et al. (2005) Evidence and implications of recent climate change in northern Alaska and other arctic regions. *Climatic Change* 72: 251–298.
- Hobbie SE, Schimel JP, Trumbore SE, and Randerson JR (2000) Controls over carbon storage and turnover in high-latitude soils. *Global Change Biology* 6: 196–210.
- Iijima Y, Fedorov AN, Park H, et al. (2010) Abrupt increases in soil temperatures following increased precipitation in a permafrost region, central Lena River basin, Russia. *Permafrost and Periglacial Processes* 21(1): 30–41.
- Jorgenson MT, Racine CH, Walters JC, and Osterkamp TE (2001) Permafrost degradation and ecological changes associated with a warming climate in central Alaska. *Climatic Change* 48: 551–571.
- Kanevskiy MZ (2003) Cryogenic structure of mountain slope deposits, northeast Russia. *Proceedings of the Eighth International Conference on Permafrost*, vol. 1, pp. 513–518. Lisse, The Netherlands: AA Balkema.
- Kokelj SV and Burn CR (2003) 'Drunken forest' and near surface ground ice in Mackenzie Delta, Northwest Territories, Canada. *Proceedings of the Eighth International Conference on Permafrost*, vol. 1, pp. 567–572. Lisse, The Netherlands: AA Balkema.
- Lewkowicz AG (1994) Ice-wedge rejuvenation, Fosheim Peninsula, Ellesmere Island, Canada. *Permafrost and Periglacial Processes* 5: 251–268.
- Little JD, Sandall H, Walegur MT, and Nelson FE (2003) Application of differential global positioning systems to monitor frost heave and thaw subsidence in tundra environments. *Permafrost and Periglacial Processes* 14: 349–357.
- Liu L, Zhang T, and Wahr J (2010) InSAR measurements of surface deformation over permafrost on the North Slope of Alaska. *Journal of Geophysical Research-Earth Surface* 115: F03023. <http://dx.doi.org/10.1029/2009JF001547>.
- Mackay JR (1981) Active layer slope movement in a continuous permafrost environment, Garry Island, Northwest Territories, Canada. *Canadian Journal of Earth Sciences* 18: 1666–1680.
- Mackay JR (1983) Downward water movement into frozen ground, western arctic coast, Canada. *Canadian Journal of Earth Sciences* 20: 120–134.
- Mazhitova GG (2008) Soil temperature regimes in the discontinuous permafrost zone in the East European Russian Arctic. *Eurasian Soil Science* 41(1): 48–62.
- Mazhitova GG and Kaverin DA (2007) Thaw depth dynamics and soil surface subsidence at a Circumpolar Active Layer Monitoring (CALM) site in the East European Russian Arctic. *Kriosfera Zemli* XI(4): 20–30 (in Russian).
- Mazhitova G, Malkova G, Chestnykh O, and Zamolodchikov D (2008) Recent decade thaw depth dynamics in the European Russian Arctic based on the Circumpolar Active Layer Monitoring (CALM) data. In: Kane DL and Hinkel KM (eds.) *Proceedings of the Ninth International Conference on Permafrost*, vol. 1, pp. 1155–1160. Fairbanks, AK: Institute of Northern Engineering, University of Alaska Fairbanks.
- Muller SW (1947) *Permafrost or Permanently Frozen Ground and Related Engineering Problems*. Ann Arbor, MI: J.W. Edwards.
- Murton JB (2001) Thermokarst sediments and sedimentary structures, Tuktoyaktuk Coastlands, western arctic Canada. *Global and Planetary Change* 28: 175–192.
- Murton JB and French HM (1994) Cryostrutures in permafrost, Tuktoyaktuk coastlands, western arctic Canada. *Canadian Journal of Earth Sciences* 31: 737–747.
- Nelson FE (2003) (Un)frozen in time. *Science* 299: 1673–1675.
- Nelson FE, Anisimov OA, and Shiklomanov NI (2001) Subsidence risk from thawing permafrost. *Nature* 410: 889–890.
- Nelson FE, Anisimov OA, and Shiklomanov NI (2002) Climate change and hazard zonation in the circum-Arctic permafrost regions. *Natural Hazards* 26: 203–225.
- Nelson FE and Hinkel KM (2003) Methods for measuring active-layer thickness. In: Humlum O and Matsuoka N (eds.) *A Handbook on Periglacial Field Methods*. Longyearbyen, Norway: University of the North in Svalbard Available online at http://www.unis.no/RESEARCH/GEOLOGY/Geo_research/Ole/PeriglacialHandbook/ActiveLayerThicknessMethods.htm.
- Nelson FE, Shiklomanov NI, Hinkel KM, and Brown J (2008) Decadal results from the Circumpolar Active Layer Monitoring (CALM) program. In: Kane DL and Hinkel KM (eds.) *Proceedings of the Ninth International Conference on Permafrost*, vol. 2, pp. 1273–1280. Fairbanks, AK: Institute of Northern Engineering, University of Alaska Fairbanks.
- Nelson FE, Shiklomanov NI, and Mueller GR (1999) Variability of active-layer thickness at multiple spatial scales, north-central Alaska, U.S.A. *Arctic, Antarctic, and Alpine Research* 31: 158–165.
- Oberman NG (2007) Global warming and permafrost changes in Pechora-Urals region. *Exploration and Protection of Natural Resources* 4: 63–68.
- Outcalt SI, Nelson FE, and Hinkel KM (1990) The zero-curtain effect: Heat and mass transfer across an isothermal region in freezing soil. *Water Resources Research* 26: 1509–1516.
- Popova VV and Shmakina AB (2009) The influence of seasonal climatic parameters on the permafrost thermal regime, West Siberia, Russia. *Permafrost and Periglacial Processes* 20: 41–56. <http://dx.doi.org/10.1002/ppp.640>.
- Romanovsky VE, Drozdov DS, Oberman NG, et al. (2010) Thermal state of permafrost in Russia. *Permafrost and Periglacial Processes* 21: 136–155. <http://dx.doi.org/10.1002/ppp.683>.
- Romanovsky VE and Osterkamp TE (2000) Effects of unfrozen water on heat and mass transport processes in the active layer and permafrost. *Permafrost and Periglacial Processes* 11: 219–239.
- Shiklomanov NI, Streletskiy DA, Nelson FE, et al. (2010) Decadal variations of active-layer thickness in moisture-controlled landscapes, Barrow, Alaska. *Journal of Geophysical Research* 115: G00I04. <http://dx.doi.org/10.1029/2009JG001248>.
- Shur YL (1975) On the transient layer. In: Shvetsov PF and Chistotinov LV (eds.) *Methods of Geocryological Studies*, pp. 82–85. Moscow: USSR Institute of Hydrogeology and Engineering Geology (in Russian).
- Shur YI (1988) The upper horizon of permafrost soils. In: Senneset K (ed.) *Proceedings of the Fifth International Conference on Permafrost*, pp. 867–871. Trondheim, Norway: Tapir.
- Shur Y, Hinkel KM, and Nelson FE (2005) The transient layer: Implications for geocryology and climate-change science. *Permafrost and Periglacial Processes* 16: 5–17.
- Smith MW (1988) The significance of climatic change for the permafrost environment. In: Senneset K (ed.) *Proceedings of the Fifth International Conference on Permafrost*, pp. 18–23. Trondheim, Norway: Tapir.

- Smith S, Romanovsky V, Lewkowicz A, et al. (2010) Thermal state of permafrost in North America: A contribution to the international polar year. *Permafrost and Periglacial Processes* 21: 117–135. <http://dx.doi.org/10.1002/ppp.690>.
- Smith LC, Sheng Y, MacDonald GM, and Hinzman LD (2005) Disappearing Arctic lakes. *Science* 308: 1429.
- Smith SL, Wolfe SA, Riseborough DW, and Nixon FM (2009) Active-layer characteristics and summer climatic indices, Mackenzie Valley, Northwest Territories, Canada. *Permafrost and Periglacial Processes* 20: 201–220. <http://dx.doi.org/10.1002/ppp.651>.
- Stokstad E (2004) Defrosting the carbon freezer of the north. *Science* 304: 1618–1620.
- Streletskiy DA, Shiklomanov NI, Nelson FE, and Klene AE (2008) 13 years of observations at Alaskan CALM sites: Long-term active layer and ground surface temperature trends. In: Kane DL and Hinkel KM (eds.) *Proceedings of the Ninth International Conference on Permafrost*, vol. 2, pp. 1727–1732. Fairbanks, AK: Institute of Northern Engineering, University of Alaska Fairbanks.
- Sturm M, Racine C, and Tape K (2001) Climate change: Increasing shrub abundance in the Arctic. *Nature* 411: 546–547.
- Sumgin MI (1927) *Vechnaya merzlota pochv v predelakh SSSR (Permafrost Soils in the USSR)*. Vladivostok: Nauka.
- Tait M, Moorman B, and Li S (2005) The long-term stability of survey monuments in permafrost. *Engineering Geology* 79: 61–79.
- US Arctic Research Commission Permafrost Task Force (2003) Climate Change, Permafrost, and Impacts on Civil Infrastructure. Arlington, VA: US Arctic Research Commission Special Report 01–03.
- van Everdingen RO (ed.) (1998) *Multi-Language Glossary of Permafrost and Related Ground-Ice Terms*. Calgary: University of Calgary.
- Vasiliev AA, Leibman MO, and Moskalenko NG (2008) Active layer monitoring in West Siberia under the CALM II Program. In: Kane DL and Hinkel KM (eds.) *Proceedings of the Ninth International Conference on Permafrost*. Fairbanks, AK: Institute of Northern Engineering, University of Alaska Fairbanks, Fairbanks, AK, June 29–July 3, vol. 2, pp. 1815–1821.
- Viereck LA, Werdin-Pfisterer NR, Phyllis CA, and Yoshikawa K (2008) Effect of wildfire and fireline construction on the annual depth of thaw in a black spruce permafrost forest in interior Alaska: A 360 year record of recovery. In: Kane DL and Hinkel KM (eds.) *Proceedings of the Ninth International Conference on Permafrost*. Fairbanks, AK: Institute of Northern Engineering, University of Alaska Fairbanks, Fairbanks, AK, June 29–July 3, vol. 2, pp. 2021–2027.
- Wu Q and Zhang T (2010) Changes in active layer thickness over the Qinghai-Tibetan Plateau from 1995 to 2007. *Journal of Geophysical Research* 115: D09107. <http://dx.doi.org/10.1029/2009JD012974>.
- Zamolodchikov D (2008) Recent climate and active layer changes in Northeast Russia: Regional output of Circumpolar Active Layer Monitoring (CALM). In: Kane DL and Hinkel KM (eds.) *Proceedings of the Ninth International Conference on Permafrost*. Fairbanks, AK: Institute of Northern Engineering, University of Alaska Fairbanks, Fairbanks, AK, June 29–July 3, vol. 2, pp. 2021–2027.

Relevant Websites

- <http://www.udel.edu/Geography/calm/> – The Circumpolar Active Layer Monitoring program's web site contains information about the CALM network, its goals and achievements, its constituent sites, and links to data.
- <http://ipa.arcticportal.org/> – The International Permafrost Association's web site explains goals, organization, membership, current activities, and a large number of links to related programs and organizations.
- <http://nsidc.org/fgdc/> – The Frozen Ground Data Center (FGDC), part of the National Snow and Ice Data Center in Boulder, Colorado, USA, contains a large number of data sets pertaining to permafrost, the active layer, and seasonally frozen ground.
- <http://www.gtnp.org/> – The Global Terrestrial Network for Permafrost (GTN-P) is a group of permafrost monitoring programs operating under the World Meteorological Organization's Global Climate Monitoring program. The GTN-P website contains information about efforts to monitor permafrost environments around the globe.
- <http://www.sciencemag.org/cgi/content/full/299/5613/1673> – A web site supporting a perspective article in the journal *Science* contains hyperlinks to a wide variety of sources describing the role of permafrost and the active layer in climate-change science.

Cryoturbation Structures

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Introduction

'Cryoturbation' is a term widely applied to sedimentary deformations that develop under periglacial conditions (Edelman et al., 1936). Its use, however, should be limited to nonsorted deformation features of cryogenic origin, that is, genetically caused by direct or indirect freezing or thawing action (Vandenberghe, 1988). From a descriptive standpoint, cryoturbations are characterized by fluidized structures with smooth boundaries and laminations that point to dominantly liquid or plastic deformation. They generally occur beneath low-angle surfaces and show little or no surface morphological expression. When the genesis of the deformation structure is uncertain, the use of a purely descriptive term ('involution') is recommended.

The great variability of cryoturbation types indicates the wide range of processes involved. Understanding the different cryogenic processes that have produced them is important because the processes are often bound to specific environments and climates. In turn, it is often possible to infer the climatic and environmental conditions from specific cryoturbation types at the time and place where they developed. Therefore, a descriptive classification is presented first, followed by a discussion of the origins of the different types of cryoturbations and, finally, their paleoclimatic significance.

Classification of Cryoturbations

A classification that assigns the wide range of cryoturbation structures to six morphological types has been proposed by Vandenberghe (1988). The deformations are classified according to their diagnostic sedimentary architecture, comprising symmetry and amplitude–wavelength ratio of the folds, and frequency and patterns of occurrence. Because this classification has proved its general applicability during the last two decades, it may be retained with some slight adaptations (Figure 1):

• Type 1 Cryoturbations

Undulations or individual folds of relatively small amplitude (in the range of decimeters) and large wavelength (in the range of a few meters).

• Type 2 Cryoturbations

Closely spaced series of repetitive, generally regular, symmetrical, intensely folded forms. Most reach a common depth and, in horizontal plan view, occur in polygonal patterns over large areas (Figure 2). Downward-extending parts may be flat bottomed. Amplitudes are mostly between 0.5 and 2 m (or occasionally much more). The volume of upward- and downward-extending parts must of course be equal, but the

width–depth ratios of upward- and downward-extending parts, may differ (Figure 1). The width–depth ratios and the amplitude may differ according to the lithological (and thus physical) properties and the relative thickness of the sediment layers taking part in rising or sinking movements.

• Type 3 Cryoturbations

Identical forms and patterns as in type 2, but of distinctly smaller dimensions (drops in the range of centimeters to pockets of a few decimeters; Figure 3).

• Type 4 Cryoturbations

Symmetrical forms, as in types 2 and 3, with different amplitudes; however, they do not occur in series but, rather, as individual forms. The pocket form is the most common one, in contrast to individual dome forms (Figure 4).

• Type 5 Cryoturbations

Apparently similar to type 2, but the lower sediment has risen along platy 'dykes' in contrast to type 2, where the uprising sediments form structures shaped like 'diapirs' or cauliflower flowers (Figure 5). In turn, the upper sediment of type 5 has subsided 'sack-wise,' while in types 2–3, it has subsided along platy 'dykes' (Kasse, 1993; Figure 6).

• Type 6 Cryoturbations

Irregular structures of all dimensions and degrees of contortion.

The Origin of Cryoturbations

Periglacial Loading

Loading is a process often invoked to explain involutions of both cryogenic and noncryogenic origin. In general terms, sediments with relatively high density may sink down into underlying, lighter sediments, while the latter sediments may rise into the upper, heavier ones. This process takes place purely under gravitational forces, without any lateral heterogeneity or force being required. Apart from this 'reversed vertical density gradient,' the intergranular contacts must be released in the lower sediments, leading to liquefaction. This can occur only in saturated soils with high void ratios because the frictional strength of most unconsolidated sediments is too high for plastic deformation (Vandenberghe and Van den Broek, 1982). In flat areas, such prerequisites necessitate an (non-lateral) excess water supply in poorly drained conditions.

The two requirements of general validity (reversed density gradient and liquefaction of the lower layer) are only rarely met in natural conditions. Cryoturbations of type 2 are independent of lithology and vertical sediment successions, and mostly show no original reversed density gradient. However, the existence of

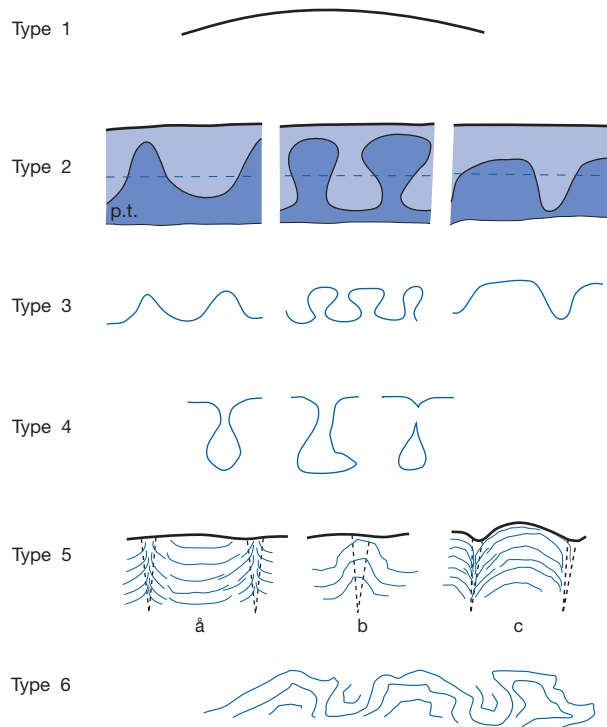


Figure 1 Types of cryoturbation structures, as modified from the classification of [Vandenberghe \(1988\)](#). p.t., permafrost table.

permafrost may create the required conditions ([Vandenberghe, 1988](#); [Vandenberghe and Van den Broek, 1982](#)). The upper part of the permafrost is often the focus of ice segregation and is therefore enriched in ice lenses, which results in a general decrease of the density of these sediments in comparison with the overlying (water-saturated) layer (first requirement). In addition, thaw of this ice-rich upper layer of permafrost can induce liquefaction (second requirement). This type of loading thus operates only in the described specific conditions and is termed 'periglacial loading' (according to [Gullentops and Paulissen, 1978](#)). It explains the type 2 involutions.

A crucial point is the maintenance of nondrained conditions during thaw of near-surface permafrost ([Van Vliet-Lanoë, 1991](#)). Moist, low-lying and flat areas without lateral drainage and underlain by fine-grained sediments (with low permeability) are most favorable for periglacial loading. The independence of the vertical lithological succession is illustrated by the fact that fine-grained sediments (e.g., silt) may sometimes sink down into coarser-grained sediments (e.g., sands or gravels), but the opposite case occurs with the same frequency. The deformations may even develop in lithologically homogeneous sediments ([Vandenberghe, 1988](#)).

The presence of an impervious layer at the time of loading is indicated by the flat-bottomed appearance of the downsinking parts ([Vandenberghe and Van den Broek, 1982](#)) and their common depth over large areas, although their amplitude may vary ([Kasse, 1999](#)). In the absence of lithological discontinuities, there is no other explanation for such an impervious horizon than the top of a frozen soil. Furthermore, the existence of permafrost at the time of loading is occasionally demonstrated by the occurrence of the top of ice-wedge

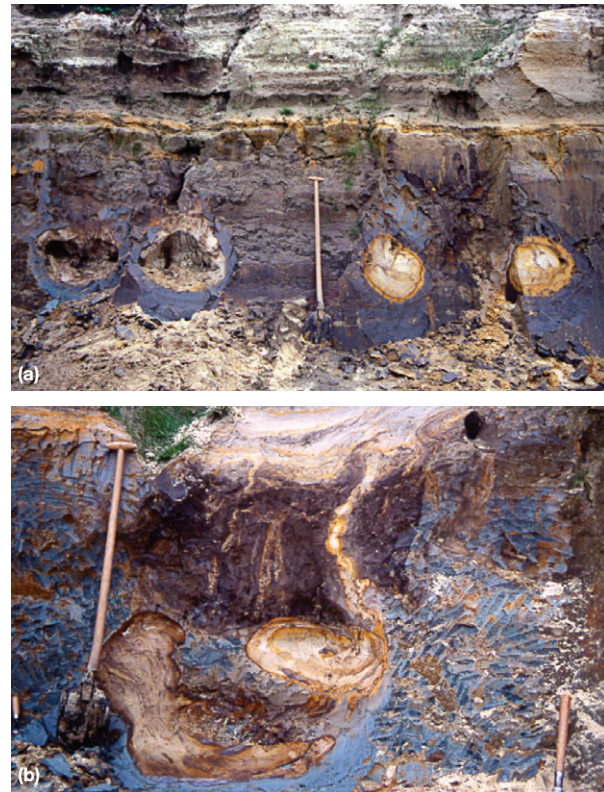


Figure 2 Type 2 cryoturbations caused by loading processes above degrading permafrost; the amplitude of involutions is ca. 1.60 m. Weichselian sands penetrate down into the underlying and presently compact Lower Pleistocene clays (Meerle, Belgium).

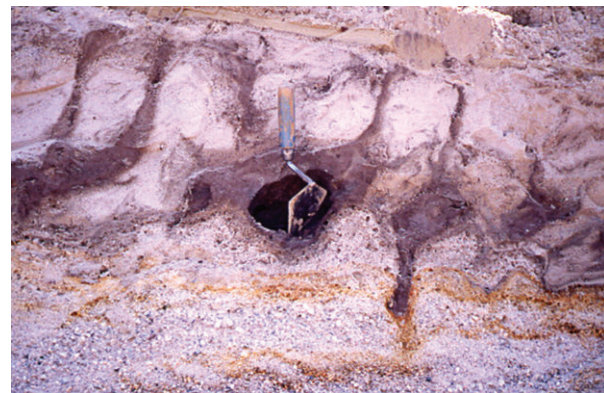


Figure 3 Type 3 cryoturbations caused by loading; the amplitude of the involutions is 30–40 cm. Slightly humic loams penetrate in underlying sands and gravely sands (Weichselian deposits, Nochten, eastern Germany; cf. [Kasse et al., 2003](#)).

pseudomorphs at the same level as the base of cryoturbations ([Vandenberghe, 1983a](#)). The position of that level determines the maximum amplitude of the periglacial loading structures.

There is a striking similarity between cryoturbated three-dimensional (3D) structures and those resulting from loading experiments in the laboratory, including primary reversed density gradients above an impervious horizon ([Anketell et al., 1970](#); [Dzulinski, 1966](#); [Harris et al., 2000](#); [Talbot and Jackson,](#)

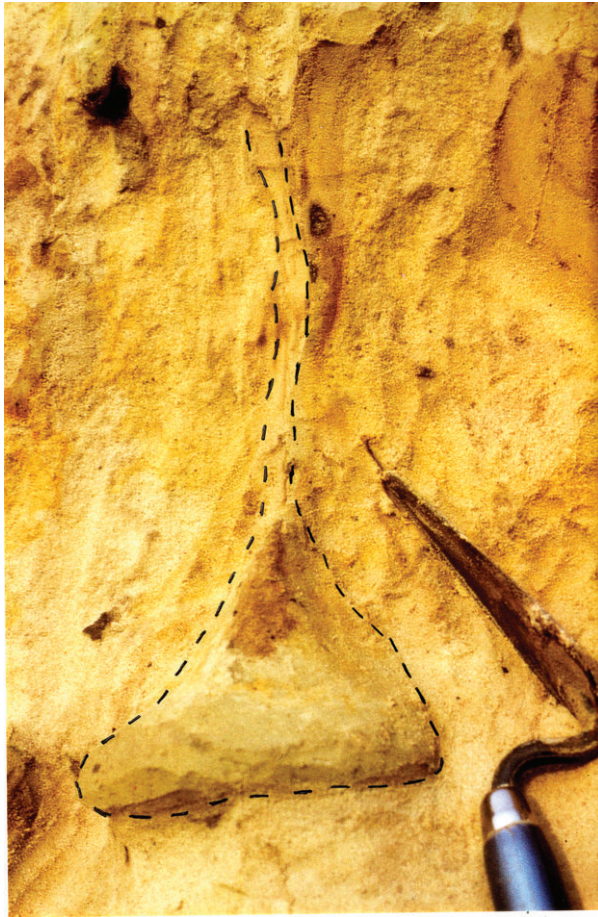


Figure 4 Type 4 cryoturbation due to a process of local loading; the amplitude is between 30 and 40 cm (Weichselian Late Pleniglacial; Weelde, northern Belgium).

1987). Although such a ‘similarity’ does not prove a common origin, it does show that it is possible. Loading processes in unconsolidated sediments, as a result of supersaturation, are not always of periglacial origin, even when showing a primary normal density gradient. Such conditions may occur whenever there is an impervious horizon with a local, bowl-shaped topography that creates supersaturation. Examples include Holocene podzolic soils developed in sands, where the impervious layer is formed by the iron-stained and humic illuviation horizon of the soil (Kasse, 1999; Schwan, 1990).

Some involutions have a more limited areal extent (type 3) or even a single occurrence (type 4), and a smaller amplitude (both types 3 and 4). Otherwise, their 3D-architecture (including occasional flat bottoms) is very similar to that of type 2 and, by analogy, also suggests an origin by loading. Their characteristic amplitude and areal extent, however, point to a shallow deformation layer, which may be interpreted as the (seasonally) thawed soil at the time of loading and not the degrading upper layer of permafrost. Oversaturation and liquefaction may be induced by more local conditions of lateral water supply (determined by lithology and topography) instead of the presence of ice lenses in the upper permafrost layer for type 2 cryoturbations. Type 3 cryoturbations often form in topographic depressions that functioned previously



Figure 5 Type 5 cryoturbation formed by injection due to cryohydrostatic pressure. (Weichselian loesses, Maastricht-Belvédère, The Netherlands; Vandenberghe, 1992; Vandenberghe et al., 1987). Height of photograph is 100 cm.

as pools in which low-permeability silts and clays were deposited. Lateral water flow in sands toward the center of the depression may be confined between the overlying fine-grained sediments of low permeability and a lower impervious layer that is represented by the frozen subsoil. This may result in a similar supersaturation and reversed density gradient as in the situation where type 2 deformations originated. Thus, it is not surprising that identical 3D structures develop, although with smaller dimensions.

The amplitude of cryoturbations may vary over a relatively short distance and thus shows a gradual transition from type 2 to type 3 (Kasse, 1999). This may be due to different thicknesses of the oversaturated layer arising from local variations in topography, vegetation, or lithology. This proves that type 3 cryoturbations may occur above permafrost, but they may also be associated with the thaw of a seasonally frozen layer.

Cryohydrostatic Flow

Processes similar to those described earlier for type 3 and 4 cryoturbations may cause sufficiently high pore pressures, resulting not only in fluidization of the sediments but also in injection. Considerable amounts of water may be supplied laterally and trapped between impervious parts of the soil, for example, between the permafrost table below and the downward freezing layer above. Owing to the resulting increased pore water pressure, the unfrozen and fluidized sediment-water mixture may be expelled by hydrostatic pressure into places of lesser resistance. As this process is limited to periglacial conditions, the term ‘cryohydrostatic’ is used (Vandenberghe, 1988). Frost and desiccation cracks may function as zones of obvious soil weakness, and hence polygonal networks of upward-displaced sediment may develop, leading to type 5a cryoturbations (Figure 5). Generally, the original crack is completely destroyed by the intrusion. The deformation is mostly restricted to the surroundings of the crack, which contrasts with the approximately equally alternating up-doming and downsinking structures of type 2 cryoturbations.

The process of upward injection of sediment into a pre-existing frost-crack polygon produces the 3D sedimentary structures shown in Figure 6. In the horizontal section at the

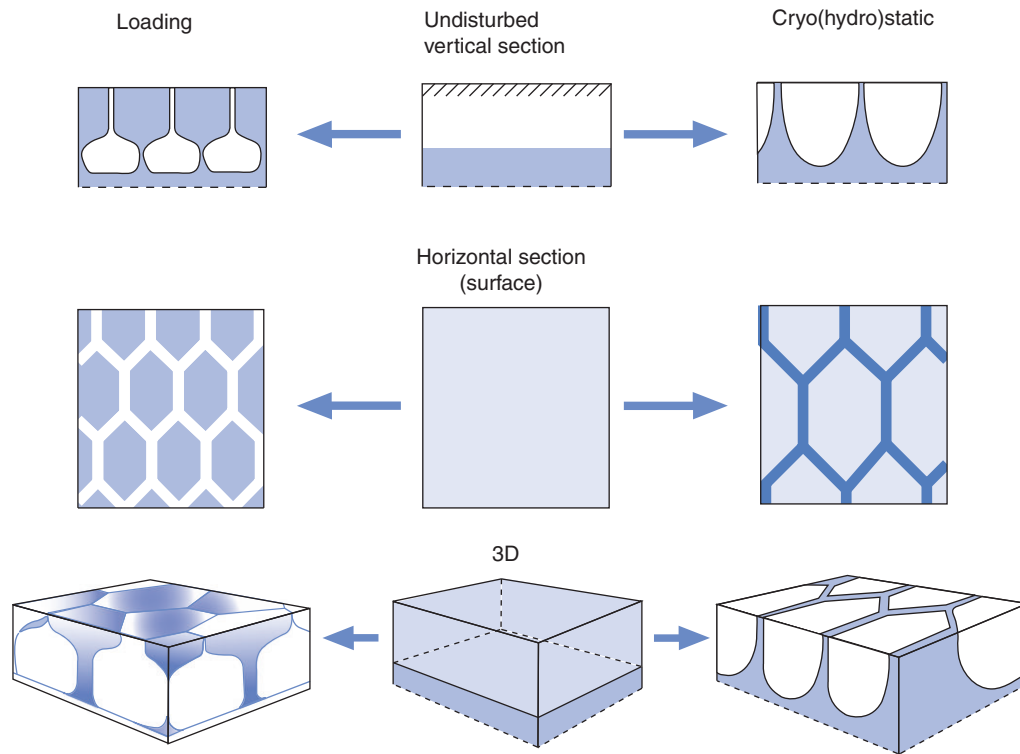


Figure 6 Horizontal, vertical, and 3D diagrams illustrating the different sedimentary structures formed by the mechanisms of loading and cryo(hydro)static pressure. Reproduced from Vandenberghe J (1992) Cryoturbations: A sediment structural analysis. *Permafrost and Periglacial Processes* 3: 343–352, with permission from Wiley & Sons.

top of the structures, a polygonal pattern appears, with a linear network consisting of the lower, intruded material, whereas the space in the core of the cells is composed of the upper sediment; this is in contrast to type 2 and 3 involutions. The 3D patterns therefore make it easy to distinguish these cryoturbation mechanisms from each other (Figure 6; Vandenberghe, 1992). Different frost susceptibility of the soil in a vertical profile (Van Vliet-Lanoë, 1991) is not necessary to create type 5 cryoturbations because identical structures may occur in homogeneous sediments and in layered sediments with different lithologies.

Cryohydrostatic injection may occur along narrow fissures and also as wider intrusions. In the vertical section, the latter resemble loading structures (type 5b in Figure 1). Again, the difference is that here a preexisting polygonal crack pattern is present along which the lower, liquefied sediment moves upward. A polygonal pattern of broad standing plates develops, whereas the original crack forms are generally lost (Kasse, 1993). Many transitional forms may occur between 'fissure injections' and 'intrusions.'

As in type 2 cryoturbations, the generation of sufficient pressure in the liquefied sediment is necessary to overcome the overburden pressure. Therefore, an impervious horizon is required below the unfrozen material, for example, a deeply frozen layer or the permafrost table.

Cryostatic Heave

In contrast to the fluid state (of unfrozen sediment) in gravitational loading and cryohydrostatic flow, plastic deformation

(in freezing sediment) is involved in the process of cryostatic heave. Cryostatic pressure develops as a result of both the volumetric expansion of freezing water and the growth of ice lenses due to migration of water. Although disputed by theoretical considerations and by experimental and field evidence (French, 1988; Mackay, 1980), differential frost susceptibility and consequent differential frost penetration and heaving may cause sedimentary deformations (Van Vliet-Lanoë, 1991). Van Vliet-Lanoë also used this mechanism to explain patterned-ground development. According to experimental data (Corte, 1971; Pissart, 1970, 1982), repeated freeze–thaw cycles may induce considerable movements of sediment. Kokelj et al. (2007) attributed a main role to the increased thickness of organic material in troughs around mud hummocks, which induces thinning of the underlying active layer and results in a bowl-shaped permafrost table below hummock centers. Ice enrichment at the top of the permafrost in the bowl leads to thrusting of soil radially inward and upward. When hummocks thaw and degrade, outward spreading of sediment at the surface completes the circular movement of the sediment.

In laterally homogeneous soils, cryostatic pressures and frost disturbance of the soil occur randomly and therefore should result in irregularly spaced structures (types 1 and 6). It is difficult to imagine how regular deformation structures could develop without preexisting regular, horizontal spatial patterns of heterogeneity. For example, frost fissures could allow differential frost penetration and induce a regular pattern of cryoturbations as defined by the fissure polygons. Likewise,

regularly spaced pockets of sediment formed by other processes, for example, type 2 loading, may also have different rates of freezing in comparison with the surrounding sediment and consequently create regular patterns of cryoturbation (Vandenberghe, 1983b).

Cryostatic pressures induce upheaval of soils in cell-like patterns controlled by the initial thermal or desiccation fissure network. Freezing of the sediment in the fissures results in lateral pressure, which may cause doming of the upper sediments in the center of the cells. In the horizontal section, the preserved structure is identical to the pattern caused by contraction polygons. This process may result in the formation of hummocks (Bohncke et al., 1993; Vandenberghe, 1988). The corresponding sedimentary deformations are classified as type 5c (Figure 1).

Cryoturbations resulting from differential cryostatic pressure apparently require a high frequency of freeze–thaw alternation rather than the existence of permafrost, except in the case of the processes reported by Van-Vliet-Lanoë (1991) and Kokelj et al. (2007).

Paleoclimatic Significance

All types of cryoturbation, except type 2, require cold conditions but not necessarily permafrost. They may be favored by repeated freeze–thaw cycles, in the case of cryostatic pressure. But with load casting (types 3 and 4) and cryohydrostatic pressure (types 5a and 5b), an impervious layer is required in order to build up the necessary conditions of oversaturation or artesian pressure. The impervious boundary in unconsolidated sediments generally corresponds with the top of a frozen subsoil. Apart from type 2 cryoturbations, this frozen subsoil may correspond with the remains of seasonally frozen ground that was still present at the time of the cryoturbation process. Therefore, all these types of cryoturbation have a limited applicability in paleoclimatic reconstruction: they simply point to frequently alternating freezing and thawing or to long-lasting and deep seasonal freezing.

Type 2 cryoturbations are particularly significant due to their large amplitude of folds. The loading process in type 2 is identical to that causing type 3 and 4 cryoturbations; but in the latter cases, it is possible, considering their small amplitude, that the deformation took place in a layer that thawed in spring. However, the large amplitudes of type 2 deformations often exceed the maximum thickness of a seasonally thawing layer. The loading process occurs when permafrost degrades and the overlying thaw layer is thicker than a seasonal thaw layer. Arguments against formation within an annual active layer (French, 1986) are the general lack of excess water in the active layer and the regularity of the type 2 deformations that would probably be destroyed if the loading process were repetitive (possibly annually). Other arguments for the presence of permafrost have been discussed above (e.g., the presence of ice-wedge pseudomorphs, the wide lateral extent, and the flat bottom of the pockets). These arguments not only point to the occurrence of a perennially frozen subsoil but also to the widespread or spatially continuous occurrence of permafrost. In conclusion, type 2 cryoturbations point to the

former presence of continuous permafrost and to formation during the degradation of that permafrost.

Paleoclimatic conditions necessary for the formation of type 2 cryoturbations are thus the same as those required for the development of continuous permafrost. Originally, a mean annual air temperature (MAAT) of less than -6°C was proposed by Péwé (1966). Later, this condition was specified in terms of subsoil lithology by Romanovskii (1985), Vandenberghe and Pissart (1993), and Huijzer and Vandenberghe (1998). A MAAT of -4°C was generally accepted as the upper thermal limit for fine-grained soils, whereas for coarse-grained subsoils, an upper limit of -8°C was assumed. It has since been argued that other factors besides temperature and subsoil lithology are influential, such as vegetation and snow thickness (Murton and Kolstrup, 2003). Van Huissteden et al. (2003) discussed in detail the potential factors that may modify the minimum paleoclimatic conditions required for the development of continuous permafrost and thus for the development of type 2 cryoturbations. It may be argued that an error bar of up to $\pm 2.5^{\circ}\text{C}$ might be applied to these cryoturbations, and thus to permafrost occurrence. In this way, the MAATs suggested earlier should be sufficient to accommodate local variations in vegetation and snow, and other deviations from present-day analogs.

Redistribution of Organic Material as a Result of Cryoturbation

The development of cryoturbations, accompanied by thickening of the active layer, moves organic matter downward, resulting in increased storage of organic carbon within the subsoil. In addition, microbial decomposition of buried organic material is delayed in comparison with that in topsoil horizons (Bockheim, 2007). The rate of carbon storage in soils is highest during the early stages of soil formation (Schlesinger, 1990). Humic-rich topsoil horizons that have subsided by cryoturbation may retain approximately their original carbon content, while the uppermost sediments may continue to undergo soil formation and accumulate new carbon (Hugelius et al., 2010; Kaiser et al., 2007). Thus, cryoturbation may promote long-term storage of carbon in the subsoil, and as such, the downsinking of C-rich topsoil may contribute significantly to the sequestration of terrestrial carbon (Kaiser et al., 2007).

A key question is: what will happen to this added store of organic carbon once cryoturbation ceases but warming continues and permafrost degrades further or disappears? Schuur et al. (2008) recognized the vulnerability of the large amounts of carbon within organic pools in permafrost, because active-layer thickening will promote microbial decomposition of previously frozen organic carbon released from the uppermost layer of permafrost. The nature and rate of this carbon release, however, are major uncertainties that are only beginning to be understood quantitatively but are beyond the scope of this contribution (Schuur et al., 2008). One approach might be to consider the evolution of the carbon pool from the last glacial cryoturbated organic layers when permafrost degraded after the Last Glacial Maximum.

Conclusion

The cryoturbation structures described earlier are linked to specific processes and environmental conditions. They may originate from (1) gravitational forces during thaw of frozen subsoil, (2) confined and pressurized water–sediment injection from between frozen parts of the subsoil, or (3) pressures due to differential frost penetration and heave. The first process occurs during degradation of underlying frozen subsoil, indicating a warm period that follows the cold period when freezing of the subsoil occurred. The other two processes occur during soil freezing, indicating cold-climate conditions.

Type 2, 3, and 4 cryoturbations, all originate by periglacial loading at the time of thawing of the underlying frozen soil. Their amplitude is a function of the depth of the impervious horizon below the surface at the time of their formation. Large amplitudes can develop above the remnant of perennially frozen ground, whereas smaller amplitudes may locally occur above the remnant of seasonally frozen subsoil.

Type 2 cryoturbations point to widespread or spatially continuous permafrost. For type 5a and 5b cryoturbations, permafrost is not necessary because an impervious horizon below the liquefied sediment is sufficient to create the required pressure in that sediment. However, in unconsolidated sediments, an impervious layer is generally absent and the permafrost table may act as such. The other cryoturbation types do not require permafrost conditions but may also occur with deep seasonal frost or frequent freeze–thaw cycles.

Cryoturbations associated with increasing active-layer thickness may contribute considerably to the redistribution of organic material and ultimately to supplementary storage of organic carbon.

See also: Permafrost and Periglacial Features: Active Layer Processes; Ice Wedges and Ice-Wedge Casts; Permafrost; Slope Deposits and Forms.

References

- Anketell JM, Cegla J, and Dzulinski S (1970) On the deformational structures in systems with reversed density gradients. *Annales de la Société Géologique de Pologne* 40: 3–30.
- Bockheim JG (2007) Importance of cryoturbation in redistributing organic carbon in permafrost-affected soils. *Journal of the American Society of Soil Science* 71: 1335–1342.
- Bohncke S, Vandenberghe J, and Huijzer AS (1993) Periglacial palaeoenvironments during the Weichselian Late Glacial in the Maas valley, the Netherlands. *Geologie en Mijnbouw* 72: 193–210.
- Corte A (1971) Laboratory formation of extrusion features by multi-cyclic freeze-thaw in soils. *Bulletin Centre de Géomorphologie* 13–15: 117–131.
- Dzulinski S (1966) Sedimentary structures resulting from convection-like pattern of motion. *Annales de la Société Géologique de Pologne* 36: 3–21.
- Edelman CH, Florschütz F, and Jeswiet J (1936) Ueber spätpleistozäne frühholozäne kryoturbate Ablagerungen in den ostlichen Niederlanden. *Verhandelingen Geologische Serie* 11: 301–336.
- French HM (1986) Periglacial involutions and mass displacement structures, Banks Island, Canada. *Geografiska Annaler* 68A: 167–174.
- French HM (1988) Active layer processes. In: Clark MJ (ed.) *Advances in Periglacial Geomorphology*, pp. 151–177. New York: Wiley.
- Gullentops F and Paulissen E (1978) The drop soil of the Eisden type. *Biuletyn Peryglacjalny* 27: 105–115.
- Harris C, Murton J, and Davies MCR (2000) Soft-sediment deformation during thawing of ice-rich frozen soils: Results of scaled centrifuge modelling experiments. *Sedimentology* 47: 687–700.
- Hugelius G, Kuhry P, Tarnocai C, and Virtanen T (2010) Soil organic carbon pools in a periglacial landscape: A case study from the Central Canadian Arctic. *Permafrost and Periglacial Processes* 21: 16–28.
- Huijzer A and Vandenberghe J (1998) Climatic reconstruction of the Weichselian Pleniglacial in north-western and central Europe. *Journal of Quaternary Science* 13: 391–417.
- Kaiser C, Meyer H, Biasi C, Rusalimova O, Barsukov P, and Richter A (2007) Conservation of soil organic matter through cryoturbation in arctic soils in Siberia. *Journal of Geophysical Research* 112: G2017. <http://dx.doi.org/10.1029/2006JG00258>.
- Kasse C (1993) Periglacial environments and climatic development during the Early Pleistocene Tiglian stage (Beerse Member) in northern Belgium. *Geologie en Mijnbouw* 72: 107–123.
- Kasse C (1999) Can involutions be used as palaeotemperature indicators? *Biuletyn Peryglacjalny* 38: 95–109.
- Kasse C, Vandenberghe J, Van Huissteden J, Bohncke SJP, and Bos JAA (2003) Sensitivity of Weichselian fluvial systems to climate change (Nochten mine, eastern Germany). *Quaternary Science Reviews* 22: 2141–2156.
- Kokelj SV, Burn CR, and Tarnocai C (2007) The structure and dynamics of earth hummocks in the subarctic forest near Inuvik, Northwest Territories, Canada. *Arctic, Antarctic, and Alpine Research* 39: 99–109.
- Mackay JR (1980) The origin of hummocks, western Arctic coast, Canada. *Canadian Journal of Earth Sciences* 17: 996–1006.
- Murton J and Kolstrup E (2003) Ice-wedge casts as indicators of palaeotemperatures: Precise proxy or wishful thinking? *Progress in Physical Geography* 27: 155–170.
- Péwé TL (1966) Paleoclimatic significance of fossil ice wedges. *Biuletyn Peryglacjalny* 15: 65–73.
- Pissart A (1970) Les phénomènes physiques essentiels liés au gel, les structures périglaciaires qui en résultent et leur signification climatique. *Annales de la Société Géologique de Belgique* 93: 7–49.
- Pissart A (1982) Déformation de cylindres de limon entourés de graviers sous l'alternance de gel-dégel. *Biuletyn Peryglacjalny* 29: 275–285.
- Romanovskii NN (1985) Distribution of recently active ice and soil wedges in the USSR. In: Church M and Slaymaker O (eds.) *Field and Theory: Lectures in Geocryology*, pp. 154–165. Vancouver, BC: University of British Columbia Press.
- Schlesinger W (1990) Evidence from chronosequence studies for a low carbon-storage potential of soils. *Nature* 348: 232–234.
- Schuur EAG, et al. (2008) Vulnerability of permafrost carbon to climate change: Implications for the global carbon cycle. *BioScience* 58: 701–714.
- Schwan J (1990) Non-cryogenic deformations in Loch Lomond Stadial to Early Flandrian coversands in North Lincolnshire, England. *Geologie en Mijnbouw* 69: 173–178.
- Talbot CJ and Jackson MPA (1987) Salt tectonics. *Scientific American* 257(2): 70–79.
- Van Huissteden J, Vandenberghe J, and Pollard D (2003) Palaeotemperature reconstructions of the European permafrost zone during marine oxygen isotope stage 3 compared with climate model results. *Journal of Quaternary Science* 18: 453–464.
- Van Vliet-Lanoë B (1991) Differential frost heave, load casting and convection: Converging mechanisms; a discussion of the origin of cryoturbations. *Permafrost and Periglacial Processes* 2: 123–139.
- Vandenberghe J (1983a) Ice-wedge casts and involutions as permafrost indicators and their stratigraphic position in the Weichselian. *Proceedings of the Fourth International Conference on Permafrost, Fairbanks*, pp. 1298–1302. Washington, DC: National Academy Press.
- Vandenberghe J (1983b) Some periglacial phenomena and their stratigraphical position in Weichselian deposits in the Netherlands. *Polarforschung* 53: 97–107.
- Vandenberghe J (1988) Cryoturbations. In: Clark MJ (ed.) *Advances in Periglacial Geomorphology*, pp. 179–198. New York: Wiley.
- Vandenberghe J (1992) Cryoturbations: A sediment structural analysis. *Permafrost and Periglacial Processes* 3: 343–352.
- Vandenberghe J and Pissart A (1993) Permafrost changes in Europe during the Last Glacial. *Permafrost and Periglacial Processes* 4: 121–135.
- Vandenberghe J, Roebroeks W, Van Kolfschoten T, Mûcher H, and Meijer T (1987) Sedimentary processes, periglacial activity and stratigraphy of the loess and fluvial deposits at Maastricht-Belvédère. In: Pecs M and French HM (eds.) *Loess and Periglacial Phenomena*, pp. 51–62. Budapest: Akademiai Kiado.
- Vandenberghe J and Van den Broek P (1982) Weichselian convolution phenomena and processes in fine sediments. *Boreas* 11: 299–315.

Ice Wedges and Ice-Wedge Casts

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Introduction

Ice wedges belong to a group of vein- and wedge-shaped structures that are characteristic of periglacial environments. The structures form by infilling of thermal contraction cracks in frozen ground with ice, clastic sediment or organic material. If ice within the infilled cracks melts, the resulting voids sometimes fill with soil, producing a pseudomorph or cast of the original ice structure. Such ice-wedge casts have been widely used since the 1960s to reconstruct Quaternary paleotemperatures. But recent studies of thermal contraction cracking and of the processes and structures of primary and secondary infilling indicate that the paleoenvironmental significance of relict veins and wedges is more complex than was initially understood and that many paleotemperature estimates in the Quaternary literature must therefore be treated with caution.

Thermal Contraction Cracking

When frozen ground cools, it contracts by an amount determined by the drop in temperature and the linear coefficient of thermal expansion. If the resulting tensile stress cannot be relieved by creep, the ground cracks. Thermal contraction cracking occurs in winter when ground that is already cold rapidly cools (Lachenbruch, 1962, 1966). In addition to rapid drops in air temperature, many site-specific factors influence cracking (e.g., snow cover, vegetation, creep of frozen ground and particle size; Mackay, 1993) because they affect the coefficients of thermal expansion and the rates of ground temperature change (Mackay and Burn, 2002; Yershov, 1998).

In a vertical cross section, thermal contraction cracks in old ice wedges often form narrow V shapes ~2–4 m deep and ~2–20 mm wide at the ground surface. Exceptionally wide cracks (≤ 20 cm) probably result from unusually high coefficients of thermal contraction. Cracks tend to be vertical because contraction is constrained laterally, and they taper downward because horizontal tensile stresses decline with depth (Mackay, 1974). Cracking tends to recur in the same location because previous cracks form lines of weakness. The frequency of ice-wedge cracking is highly variable, site-specific and time-dependent (Mackay, 1992).

Primary Infilling

Mechanisms and Conditions

Thermal contraction cracks may fill with snow, ice, mineral particles or organic material, according to near-surface conditions. In winter, snow may blow into cracks. In spring, meltwater from snow or ground ice may flow into them and refreeze, or crystals of hoar ice may grow on crack walls when

the air becomes warmer than the ground. But in midwinter, when temperatures in the snowpack tend to be colder than those in the underlying permafrost, sublimation may remove ice from the sides of open cracks. In polar deserts, sand often blows into open cracks, and in tundra regions, organic material may contribute to crack infilling. The cracks may narrow before infilling or close without infilling because of thermal expansion of ground in summer or plugging of cracks with ice or sediment (Mackay, 1975). As a result, the infilling materials and volumes may vary over short distances, from year to year or during a single year.

Veins and Wedges of Primary Infilling

The infilling of thermal contraction cracks produces structures of variable size and shape (Table 1). Veins are more or less uniform in width and generally < 10 cm wide, whereas wedges taper downward and have a maximum width generally > 10 cm. An important distinction is made between the infilling of an open crack (primary infilling) and infilling of a void created by melt of ice veins or wedges (secondary infilling). Wedges restricted to the active layer or to seasonally frozen ground (SFG) in nonpermafrost areas are prefixed accordingly.

Ice veins and ice wedges

An 'ice vein' constitutes the ice that fills a single thermal contraction crack, whereas an 'ice wedge' constitutes the ice that fills multiple cracks in more or less the same location within permafrost. Ice wedges usually taper downward because of their tendency to crack near their centers, the frequent confinement of cracks to wedge ice, and the downward taper of the original cracks (Mackay, 1974). Actively growing ice wedges are most common in the continuous permafrost zone (Romanovskij, 1985). Wedges that are rarely active also occur in discontinuous permafrost, remaining dormant during most years but cracking during winters that are unusually cold or deficient in snow cover. Ice wedges tend to be larger and more common in fine-grained and ice-rich sediments (or icy peats) than in ice-poor sand or gravel because of the higher coefficients of thermal expansion of icy silty clays and peats. Accordingly, smaller ice-wedge polygons often occur in finer-textured material, in ice-rich soils, or where temperature changes are marked (Black, 1974).

Three types of ice wedges can be distinguished according to ground-surface stability (Mackay, 1990, 1995). On stable surfaces, ice wedges are superimposed (epigenetic) and grow wider; on surfaces rising by sedimentation or peat aggradation, the wedges grow upward (syngenetic); and on surfaces lowering by denudation, they grow downward (anti-syngenetic; Figure 1).

Table 1 Terminology of wedges and veins^a

Ground thermal regime	Primary infilling of thermal contraction crack	Primary structure	Secondary infilling (replacement of ice with soil)	Secondary structure	Genetic interpretation
Permafrost	Ice	Ice wedge	Soil	Soil wedge ^b	Ice-wedge pseudomorph
Active layer	Ice	Seasonal ice vein	Soil	Soil vein	Active-layer ice-vein pseudomorph
				Soil wedge	Active-layer ice-wedge pseudomorph
Seasonally frozen ground (SFG) in nonpermafrost area	Ice	Seasonal ice vein	Soil	Soil vein	SFG ice-vein pseudomorph
Permafrost	Ice and soil ^c	Composite wedge	Soil replaces icy part	Soil wedge	Composite-wedge pseudomorph
Active layer	Ice and soil	Seasonal composite vein	Soil replaces icy part	Soil vein	Active-layer composite-vein pseudomorph
SFG	Ice and soil	Seasonal composite vein	Soil replaces icy part	Soil vein	SFG composite-vein pseudomorph
Permafrost	Soil	Soil wedge (e.g., sand wedge)	None		Primary permafrost soil wedge
Active layer	Soil	Soil vein (e.g., sand vein)	None		Primary active-layer soil wedge
SFG	Soil	Soil vein	None		Primary SFG soil vein

^aVeins are more or less uniform in width and generally <10 cm wide; wedges taper downward and have a maximum width generally >10 cm.

^bThe term 'soil wedge' rather than 'ground wedge' is used to describe non-icy infillings because 'ground' (as in perennially frozen ground) includes ice.

^cSoil includes all non-icy material (mineral particles, rock fragments, organic material).

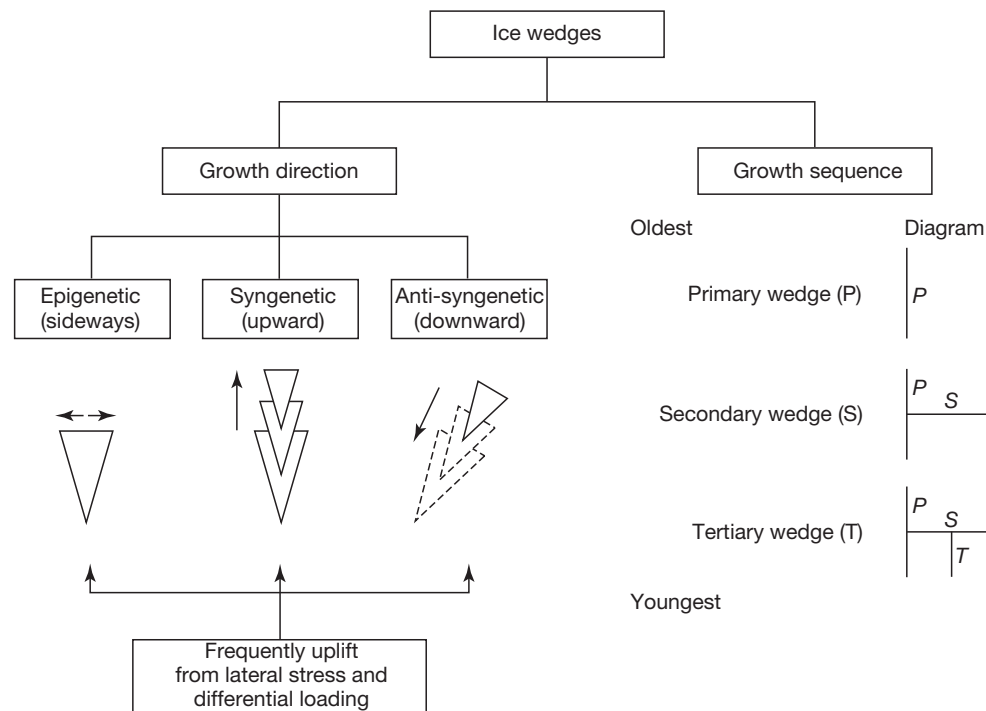


Figure 1 Classification of ice wedges according to growth direction and growth sequence. Reproduced from Mackay JR (2000) Thermally induced movements in ice-wedge polygons, western Arctic coast: A long-term study. *Geographie Physique et Quaternaire* 54: 41–68; Figure 2.

Epigenetic wedges

Epigenetic wedges grow in preexisting permafrost and are younger than the host material. They grow progressively wider but little if any higher or deeper, resulting in a downward-tapering body of wedge ice. Black (1974) suggested, however, that wedges cannot widen beyond that state where their sides are inclined less than 45°, because compressive stresses cause flow and shear to occur within the wedge. The age of ice on the edges of epigenetic wedges changes little from bottom to top. Epigenetic ice wedges beneath areas of tundra commonly attain maximum widths of 0.5–2 m and depths of 3–5 m (Figure 2(a) and 2(b)). Ice wedges beneath polar deserts are rarely described in vertical section but may be smaller, due to the arid conditions and slow growth, for example, by hoar-frost accretion.

Syngenetic wedges

Syngenetic wedges grow upward as the permafrost table rises due to accumulation of sediment or peat on the ground surface. Hence, a new ice vein in a syngenetic wedge is of similar age to that of the new material in which it grows, and the age of ice on the edges of a syngenetic wedge decreases upward. Syngenetic wedges are often vertically nested in a chevron pattern, their shape determined by both the vertical and horizontal rates of wedge growth. Tall and wide syngenetic wedges develop where sedimentation continues for long periods of time, as in major floodplains (Dostovalov and Popov, 1966) or regions of thick loess (Kanevskiy et al., 2011). The largest reported syngenetic wedges, from Siberia, are 50–80 m or more high and ≤8–10 m wide (Yershov, 1998). Narrow

syngenetic wedges may form during accumulation of peat in ice-wedge troughs and polygon centers (Figure 3(b); Mackay, 1992, 2000) or aggradation of windblown sand (Figure 2(c)).

Anti-syngenetic wedges

Anti-syngenetic wedges develop where erosion progressively lowers the ground surface. Favored locations include convex sites near hilltops where the rate of active-layer movement increases downslope, and steep, bare hillslopes that recede rapidly by river undercutting. As the active layer is eroded, ice veins grow to progressively greater depths, perpendicular to the ground surface. In tandem with removal of surface material, active-layer deepening truncates the top of anti-syngenetic wedges. Thus, the age of ice on the edges of the wedges increases upward, and the growth record of old wedges is always incomplete. In hilly topography along the western Arctic coast, Canada, anti-syngenetic wedges are probably more abundant than either epi- or syngenetic ones (Mackay, 1995). They also tend to be much wider than epigenetic wedges in adjacent flattish areas; maximum true widths are commonly a few meters, with the widest >8.4 m (Mackay, 2000). Such widths may develop because less winter snow accumulates on windblown hillslopes than on most flattish areas, and so the hillslopes are probably colder and subject to more frequent cracking.

Ice-wedge growth forms may vary between adjacent sites and through time. Hilltop portions of active ice-wedge systems could be epigenetic, receding hillslope portions anti-syngenetic, and depositional areas of valley bottoms syngenetic. Epigenetic wedges growing near hilltop edges may evolve into anti-syngenetic wedges through hillslope recession, while anti-syngenetic



Figure 2 Ice wedges from shrub tundra terrain of the Mackenzie Delta region, western Arctic Canada. (a) Epigenetic ice wedge exposed in vertical section perpendicular to axial plane of wedge, Crumbling Point, Summer Island. Person for scale. (b) Epigenetic ice wedge exposed in a vertical and horizontal section oblique to axial plane of wedge, northeast Richards Island. Person for scale. A second ice wedge occurs on the left of the photograph. (c) Syngenetic ice wedges in eolian sand, Crumbling Point. Person for scale. (d) Rejuvenated ice wedge showing evidence for several growth stages at its top, North Head, Richards Island. Spade for scale. © 2006, Photographs by J. B. Murton.

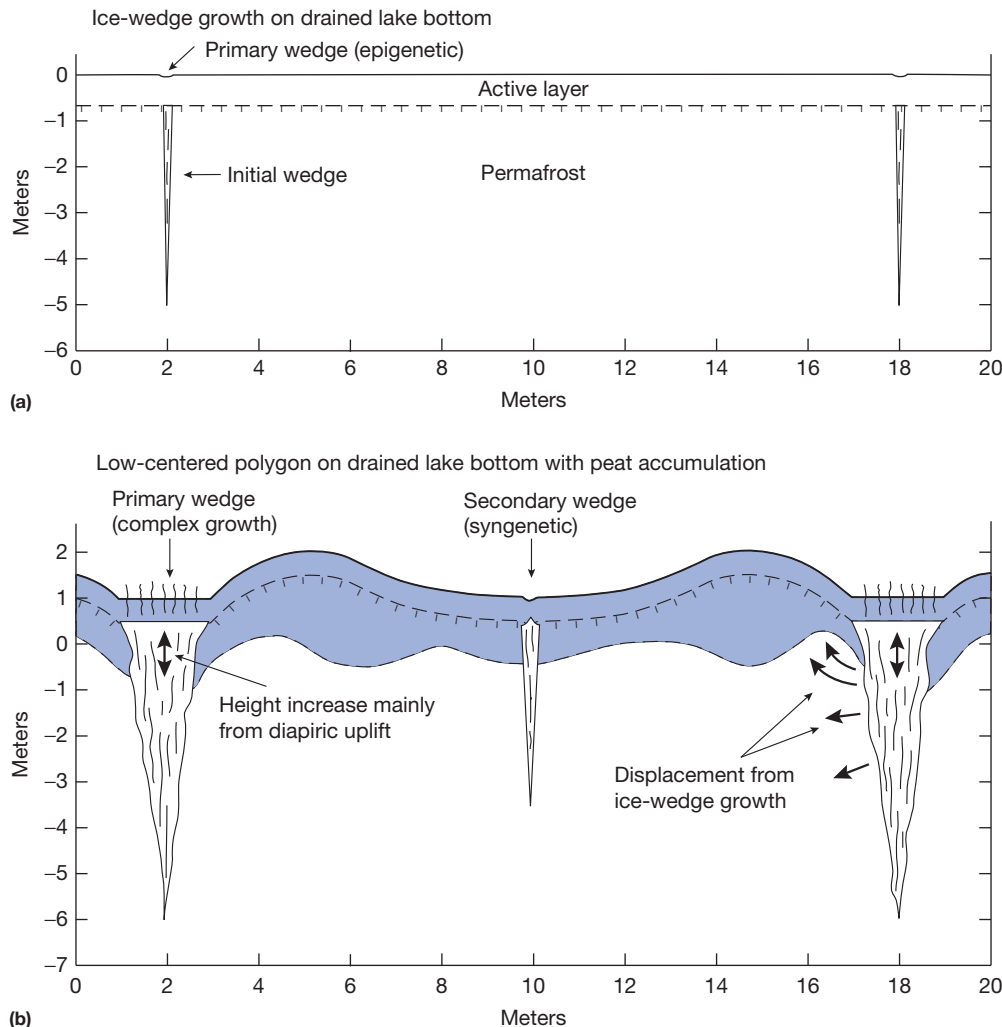


Figure 3 (a) Schematic diagram showing the growth of primary epigenetic ice wedges beneath a newly drained lake bottom. The tops of the wedges commence growth 0.7 m below zero datum. (b) After accumulation of 1–2 m of peat, the tops of the primary ice wedges are now 0.5 m above zero datum (the mean level of the peat–soil contact). Thus, the tops have moved upward 1.2 m, although there is no evidence for syngenetic growth. A secondary, syngenetic ice wedge has commenced growth in the polygon center. Reproduced from Mackay JR (1992) The frequency of ice-wedge cracking (1967–1987) at Garry Island, western Arctic coast, Canada. *Canadian Journal of Earth Sciences* 29: 236–248.

wedges growing near the base of receding hillslopes may evolve into syngenetic wedges if they are buried by deposition from upslope.

This threefold classification is undoubtedly useful but can be difficult to apply to some wedges. For example, many so-called epigenetic ice wedges beneath peaty, low-centered polygons grow both wider and higher due to uplift of wedge ice rather than syngenetic growth (Mackay, 1992). The ground surface adjoining epigenetic wedges rises progressively by the added volume of wedge ice and sometimes by the accumulation of peat. Field evidence for upward growth occurs where the tops of ice wedges are above the original ground surface (de Leffingwell, 1915), which in areas of long-term peat accumulation approximates to the contact between the soil and overlying peat (Figure 3(b)). Another problem sometimes arises where little is known about ground-surface stability, which can therefore make it difficult to distinguish between epi- and

syngenetic wedge growth. In this case, the term 'rejuvenated' wedge is sometimes appropriate.

Rejuvenated wedges

Rejuvenated wedges have two or more growth stages evident at their tops (Lewkowicz, 1994; Mackay, 1974). They comprise a series of nested wedges, in which a lower wedge has one or more raised tops, usually extending up from its center and often forming a pyramidal shape (Figure 2(d)). Rejuvenated wedges form when the permafrost table rises, such that ice veins in the active layer are not completely melted by summer thaw but add an increment of ice to the wedge top. Rejuvenation results from several factors, including climate cooling, accumulation of colluvial or peat deposits, unusually warm summers followed by cooler ones, or by lake drainage and regrowth of old ice wedges partially thawed beneath the former lake.

Soil veins and soil wedges of primary infilling

Soil veins and wedges of primary infilling form where mineral particles, rock fragments or organic materials accumulate in open thermal contraction cracks (Table 1). The best-known type of primary soil wedge is a permafrost sand wedge (Murton et al., 2000; Péwé, 1959). Active sand wedges are most abundant in polar deserts, where vegetation and snow cover tend to be meager or absent, promoting eolian sand transport when thermal contraction cracks are open (Bockheim et al., 2009a; Sletton et al., 2003). But sand wedges may form locally in sandy tundra with little or no winter snow cover. Small, active primary sand wedges (~ 2 m high) and veins are not restricted to permafrost but also form in active layers and, occasionally, in seasonally frozen ground.

Individual sand veins are typically 1–10 mm thick, 10–400 cm deep and occur singly or in groups known as 'bundles' and sand wedges (Figure 4). Each vein forms where sand falls into an open thermal contraction crack, often blown there. Sand wedges often vary more in shape than ice wedges because the sandy materials in which they frequently occur have tensile strengths similar to that of the sandy infill. Hence, the location of cracking in sand tends to be more variable than cracking in wedge ice, resulting in sand veins branching out from the sides and toes of sand wedges and in complex, cross-cutting relationships between bundles of sand veins within sand wedges. Epigenetic sand wedges develop in stable substrates; syngenetic bundles and wedges develop where cracking and infilling coincide with sediment aggradation, for example, within alluvial fans and eolian or fluvio-eolian sand sheets; and anti-syngenetic wedges form where the ground surface is lowering, for example, through deflation of sand and silt or sublimation of buried glacier ice (Murton and Bateman, 2007).

Soil veins and wedges of primary infilling have a high preservation potential in Quaternary sediment sequences because they contain little ground ice and therefore suffer little if any modification during permafrost degradation. As a result, relict sand wedges dating back to the Middle or early Pleistocene have been observed, for example, in southern Patagonia, where four horizons of wedges are dated from ~ 200 ka to ~ 1.0 – 1.1 Ma (Bockheim et al., 2009b). Relict wedges can often be distinguished from ice-wedge pseudomorphs (discussed later) by the occurrence of vertical to steeply dipping lamination within the wedge and by upturned strata in the host material. But there are no unique criteria to identify primary sand veins and wedges because of similarities between sand veins and wedges of diverse origins (Murton et al., 2000). Identification depends on both the occurrence of distinctive features, which are not always present (Table 2), and upon evaluation of their lithofacies context. This procedure should exclude subaqueous and warm-climate-related deformation products.

Dating of the cracking and infilling can be established in some late Quaternary sand veins and wedges of primary infilling by measuring the luminescence of their sand grains (e.g., Bateman et al., 2010; Kolstrup and Mejdahl, 1986). This method assumes that sunlight during grain transport totally bleaches any preexisting luminescence. After burial of the sand, the luminescence gradually regenerates through exposure to natural ionizing radiation. Luminescence dating of ancient or relict primary sand wedges can provide valuable

geochronological constraints to Quaternary histories (Buylaert et al., 2009; Murton et al., 1997).

Composite veins and composite wedges

Composite veins and wedges form by infilling of thermal contraction cracks with both ice and soil (Figure 5). Ice–soil mixtures may fill individual cracks, or the infill may vary over time due to changes in snow cover and eolian sand transport. Active permafrost composite wedges have been reported from Victoria Land, Antarctica, and vertical sections through active and inactive ones from western Arctic Canada. Composite veins may form seasonally in the active layer but experience thaw disturbance in summer.

Wedge development

Rates of wedge development vary substantially. Reported rates of widening of established ice wedges and sand wedges in some arctic and antarctic regions are ~ 1 – 4 mm a year, but the high variability of cracking frequency indicates large interannual variations in growth. Young ice wedges sometimes grow rapidly. For example, for the first 10 years after the draining of Lake Illisarvik, western Arctic Canada, growth rates ranged from ~ 1 to 3 cm a year, but ice-wedge growth had ceased in most of the lake bottom within 12 years because growing vegetation trapped deep snow (Mackay and Burn, 2002). Much lower growth rates occur where conditions for wedge development are marginal, as in sub-arctic peatlands.

Ice wedges, sand wedges and composite wedges develop over timescales measured in a few years to thousands of years. Their development often involves episodes of growth, inactivity and partial thaw because of changes in climate and ground-surface conditions. Fluctuations in active-layer depth produce complex histories of ice-wedge thaw and rejuvenation. Evidence for two phases of ice-wedge growth separated by early Holocene climate warming and active-layer deepening is common along the western Arctic coast, Canada (Mackay, 1995), and more complex stages of wedge growth, involving both climate change and variation in the primary infilling, are inferred from composite wedges here (Figure 5(c)).

Secondary Infilling

Ice within primary veins and wedges is susceptible to melting and replacement (secondary infilling) with soil or rock fragments. Melting results from a variety of local or regional factors that trigger thermokarst subsidence or thermal erosion. In turn, melting may initiate secondary infilling of voids left from ice or composite veins and wedges in permafrost (ice-wedge casting), in the active layer or in areas of seasonally frozen ground. Because secondary infilling processes affect the size and shape of ice-wedge-related structures in Quaternary sediments, we must consider the infilling mechanisms and conditions before we consider the paleoenvironmental significance of veins and wedges.

Mechanisms and Conditions

Thaw-consolidation processes influence sediment strength and thus the amount of infilling and type of deformation

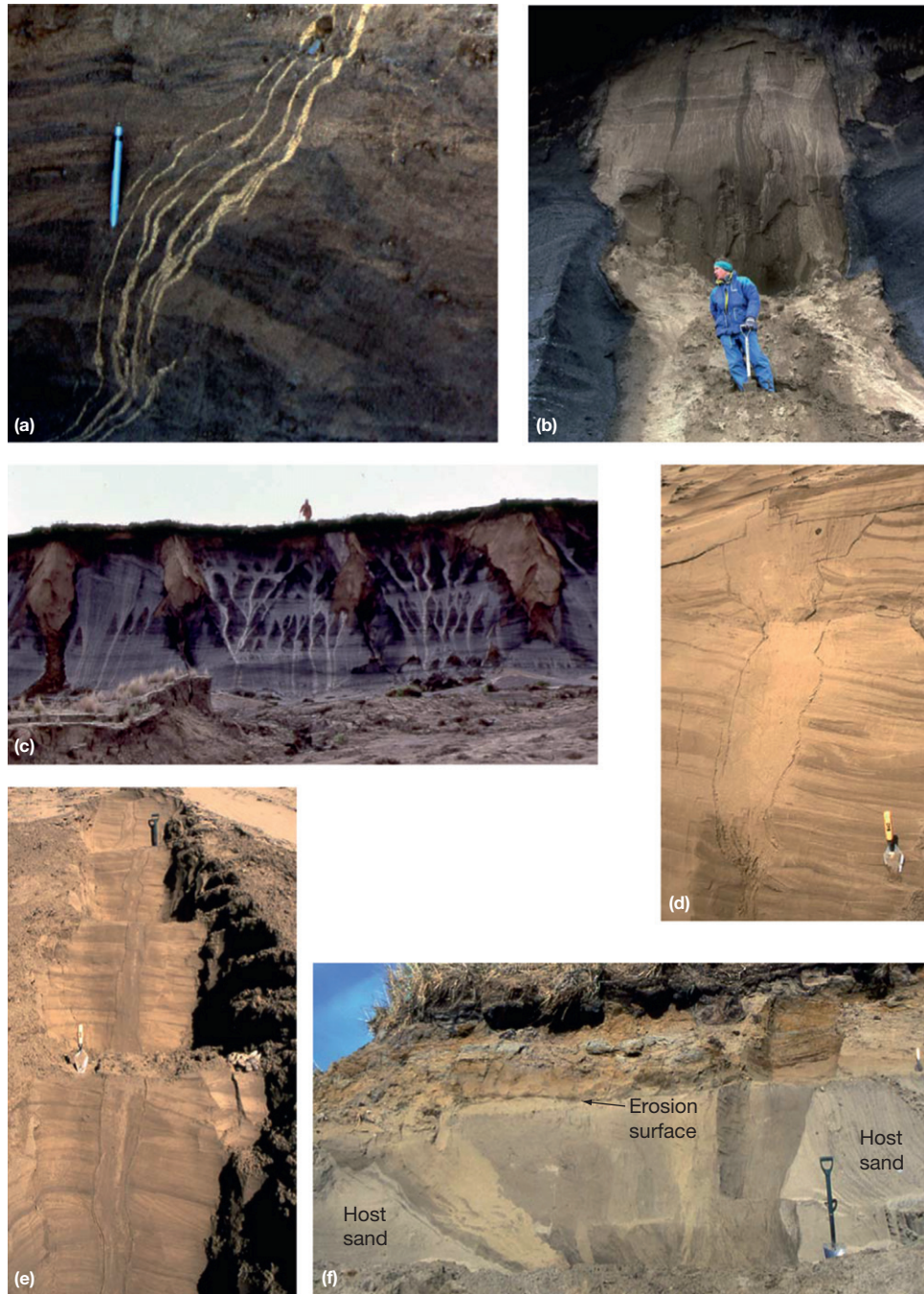


Figure 4 Sand veins, bundles and wedges of primary infilling in vertical cross section, Mackenzie Delta region, western Arctic Canada. (a) Sand veins in sand, Crumbling Point. Pencil for scale. (b) Bundles of sand veins within an epigenetic composite wedge (with a predominant sand infilling – see caption to [Figure 5\(c\)](#)), Crumbling Point. Person for scale. (c) Four epigenetic sand wedges in massive ice and icy sediments, Crumbling Point. Person for scale. (d) Rejuvenated sand wedge in sand, Hadwen Island. The main part of the wedge penetrates eolian dune sand, whereas the rejuvenated top is in eolian sand-sheet deposits. Trowel for scale. (e) Syngenetic sand-vein bundle in eolian sand sheet, Summer Island. The height of the bundle exceeds 4.9 m, and the true width of the bundle varies from 7 to 21 cm. Trowel and spade for scale. (f) Anti-syngenetic sand wedge, Johnson Bay, northeast Tuktoyaktuk Peninsula. The maximum true width of the wedge is 3.9 m. The top of the wedge is truncated by an erosion surface. Bundles of sand veins occur within the wedge. Spade for scale. © 2006, Photographs by J. B. Murton.

associated with ice-wedge casting. In a series of centrifuge modeling experiments to investigate casting processes and structures, model ice wedges melted more slowly than ground ice in the adjacent soil because of their larger latent heat effects. The experiments produced a variety of artificial ice-wedge

pseudomorphs, tunnels and involutions ([Figure 6](#); [Harris and Murton, 2005](#)). They showed that if the soil beside a melting wedge has time to consolidate, then the soil will be much stiffer when the wedge melts than it would be if the wedge melted at the same rate as the soil. Such soil is less likely to

Table 2 Characteristic features of primary sand veins and wedges

<i>Feature</i>	<i>Origin</i>	<i>Comment</i>
Sand vein (commonly 1–10 mm thick, 10–400 cm deep, sharp planar to curved sides, internally massive, truncating stratification in host sand and, commonly, other sand veins; veins may bifurcate and rejoin and are commonly laterally and vertically discontinuous)	Filling of individual thermal contraction crack	Host strata adjacent to sand veins are generally undeformed because of the negligible volume of sand infill added to the ground
Vertical to steeply dipping lamination, sometimes arranged in bundles	Filling of numerous thermal contraction cracks	Group of sand veins, sometimes as mineralogically or texturally distinct bundles within sand wedges. But some sand wedges or parts of them appear on close field examination to be massive, even in wind-etched exposures, because sand source was texturally and mineralogically very uniform
Sand veins extending from the sides or toes of wedges in sand (apophyses)	Variation in location of infilled thermal contraction cracks	Tensile strength of the wedge and host sand is similar, allowing more variation in the location of thermal contraction cracking than is typical of cracking of ice wedges in soil
Inclusions of host material are sometimes cross-cut by sand veins	Variation in location of thermal contraction cracking	As above
Sand grains are often more rounded (some like millet seeds) and, seen under scanning electron microscope, may contain more upturned plates, dish-shaped concavities and an 'orange-peel texture' than grains in host material	Eolian abrasion of sand infill is greater than that of host material	Sand grains of infill often resemble those of eolian sand sheets or dunes. Sand is commonly fine- to medium-grained and moderately well sorted
Pebbles or granules, where present, are generally concentrated at or near the top of sand wedges and any gravel content very distinctly decreases with depth	(1) Smaller particles fall furthest down thermal contraction cracks; (2) the limited crack width can act as a filter mechanism, restricting the larger-sized material to the upper part of the wedge	
Upturned strata in host material adjacent to sand wedges	Addition of wedge material to permafrost and ground expansion in summer	Downturned and undeformed strata are more unusual. Downturned strata in massive ice may reflect creep of ice in response to greater density of sand. Downturned strata in sand may result from folding associated with lateral compression during wedge growth. Sometimes both up- and downturned strata occur beside different parts of the same wedge. Undeformed strata adjacent to sand wedges are difficult to explain. Initial loose packing and high void ratios might contribute toward accommodation space
Ventifact lags commonly extend laterally from tops of sand wedges	Eolian abrasion	Strong winds that transported sand to thermal contraction cracks caused deflation and abrasion of adjacent material
Large-scale polygonal network of wedges	Thermal contraction cracking	

deform than unconsolidated soil with porewater pressures in excess of hydrostatic. Sand consolidates almost instantaneously, and therefore excess pore pressures are unlikely. Clay consolidates more slowly and may generate excess pore pressures during thaw, making the clay soft and deformable. Hence, secondary infilling in the experiments occurred to a greater degree in clayey and icy soils (Figure 6(e) and 6(f)) than in sandy or silty soils that contained less ice (Figure 6(a)–6(d)).

Secondary infilling of tunnels thermally eroded through ice wedges occurs by subsidence or collapse of material from tunnel walls and roofs, or by transport of material along tunnels by flowing water or colluvial processes (Figure 7). Secondary infilling is often episodic and interrupted or

counteracted by refreezing of water. For example, after episodes of thermal erosion of ice-wedge terrain (e.g., Fortier et al., 2007), water often refreezes in underground cavities within wedge ice to form 'pool' ice. Furthermore, active-layer deepening during unusually warm summers or episodes of warmer climate may be counteracted later by ice-wedge rejuvenation during cooler summers or climates. Where the wedge ice melts completely, infilling of tunnels and surface troughs may continue after melting has ceased, as indicated by voids within ice-wedge pseudomorphs and troughs above them (Romanovskij, 1973). Thus, the final pseudomorph may develop over a variable and sometimes long period of time, complicating attempts to date ice-wedge melting and secondary infilling.

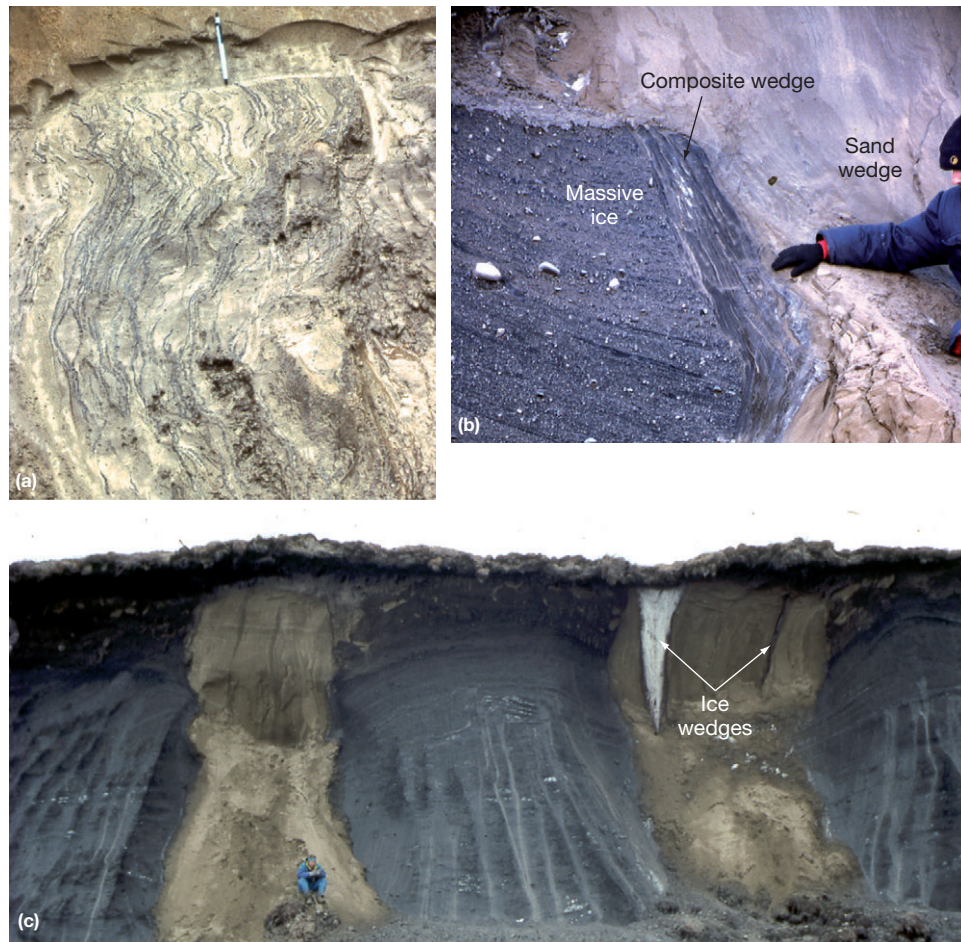


Figure 5 Composite wedges of epigenetic type, Mackenzie Delta region, western Arctic Canada. (a) Alternating ice veins and sand veins within a composite wedge, Hadwen Island. Pencil for scale. (b) Composite wedge along the side of a sand wedge penetrating massive ice, Mason Bay, Richards Island. Hand for scale. (c) Composite wedges penetrating massive ice and icy sediments, Crumbling Point. Ice veins and sand veins similar to those in (b) occur along the sides of both wedges, truncated at a depth of ~ 2 m. Both wedges commenced growth as composite wedges, continued as sand wedges and were then interrupted by active-layer deepening during the early Holocene. The wedge on the right recommenced growth by ice-wedge development at a stratigraphically higher level during the late Holocene. Person for scale. (a and b) © 2006, Photograph by J. B. Murton. (c) Reprinted from Murton JB, Worsley P, and Gozdzik J (2000) Sand veins and wedges in cold aeolian environments. *Quaternary Science Reviews* 19: 899–922.

Melting of ice veins or wedges does not always lead to secondary infilling (Dylik, 1966). For example, syngenetic ice wedges tens of meters high in very ice-rich silts are unlikely to produce pseudomorphs, except perhaps in their toes, because the silts become liquid like on thawing and prone to reworking and erosion.

Soil Veins and Wedges of Secondary Infilling (Pseudomorphs)

Soil veins and wedges of secondary infilling preserve, to varying degrees, a record of the size and shape of the original ice structures. Preservation varies due to thaw consolidation, lateral movement of host material or tunnel collapse. Hence, structures of secondary infilling preserve imperfectly the original ice-wedge features and so are best termed ‘pseudomorphs’ (Table 1; Péwé et al., 1969). The term ‘cast,’ which is widely used in the Quaternary and periglacial literature, is rather misleading because it implies exact preservation of the size and shape of the ice wedge (Harry and Gozdzik, 1988). Such

preservation is of course impossible to determine where no ice remains. Moreover, exact preservation of an original roof of sediment after the underlying wedge has melted remains to be demonstrated.

Characteristics of Cenozoic-age pseudomorphs

The characteristic features of ice-wedge pseudomorphs and composite-wedge pseudomorphs of Cenozoic age are listed in Table 3 and illustrated in Figure 8. Seldom, however, are all of the characteristics found together in a single stratigraphic section.

Pseudomorphs have infills comprising material similar to that beside or above them. This similarity results from subsidence of the host or overlying material into space left by melting wedge ice. The effects of subsidence are clearest where a distinctive horizon such as peat, paleosol or tephra downturns toward or into a pseudomorph. Depending on the nature of the host sediment and the processes of ice-wedge casting, pseudomorph infills can be massive or stratified, the

strata comprising indistinct vertical to steeply dipping layers or concave-up layers. The layering in ice-wedge pseudomorphs can often be readily distinguished from the fine vertical lamination associated with soil wedges of primary infilling (Figure 4). In composite-wedge pseudomorphs, however, layering formed at different times by primary and secondary infilling may be preserved. Some pseudomorphs contain clasts or faulted blocks of sediment too large to have entered into thermal contraction cracks no wider than a few centimeters; such clasts are therefore attributed to secondary infilling. Elongate clasts may become vertically aligned by slipping down into the narrow gap beside inward-melting ice wedges. Water

seeping down such gaps may carry fine sediment with it, depositing a 'lining' along the sides of pseudomorphs. Some pseudomorphs may contain tunnels or tunnel infills. Normal faults or downturned layers commonly occur in stratified sediments adjacent to pseudomorphs, as a result of loss of lateral support of the host sediment adjacent to melting ice wedges. Finally, pseudomorphs often form a polygonal network, sometimes clearly depicted on aerial photographs by patterns of differential crop growth.

Distinguishing ice-wedge pseudomorphs from wedges of primary infilling or wedge-like structures unrelated to thermal contraction cracking is based on criteria listed in Tables 2 and 3.

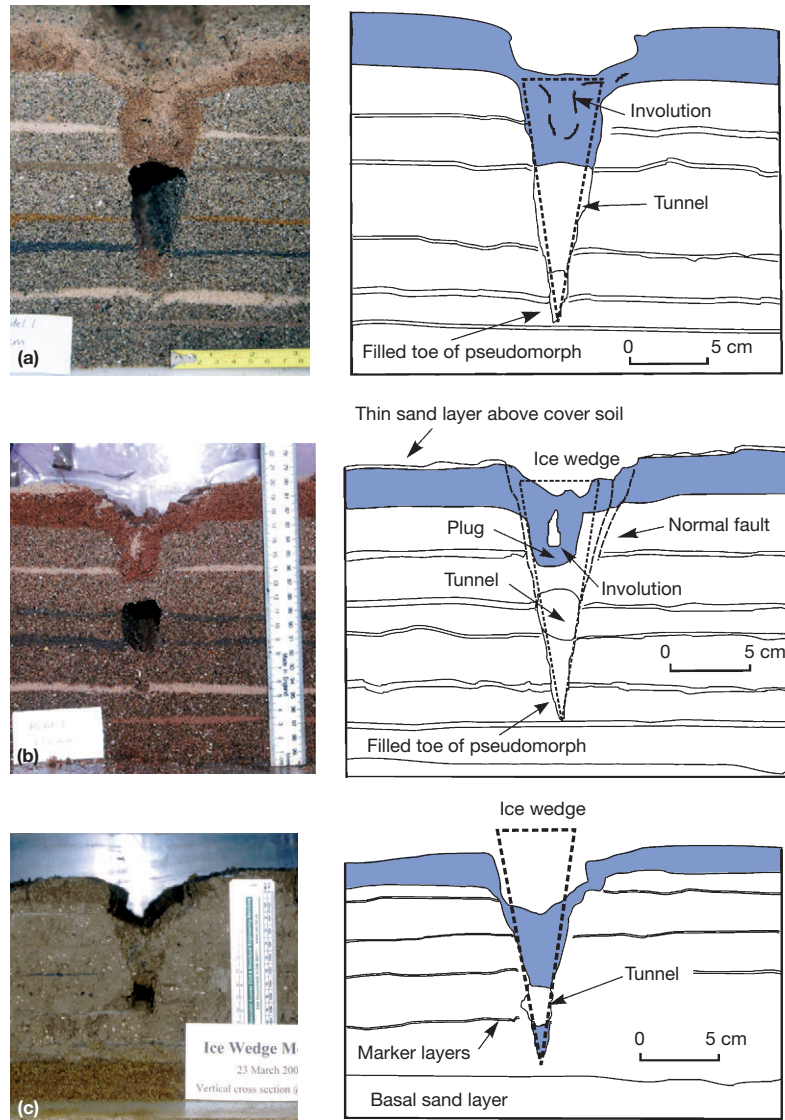


Figure 6 Photographs and annotated drawings of artificial ice-wedge pseudomorphs, tunnels and involutions excavated in vertical sections following simulation of ice-wedge casting in a large geotechnical centrifuge. (a) Model 1 (medium sand, 15% gravimetric moisture content). (b) Model 2 (medium sand, 20% moisture, undrained). (c) Model 3 (Pegwell Bay silt, 40% moisture content). (d) Model 4 (Pegwell Bay silt, 20% moisture content). (e) Model 5 (2:1 by weight silt-clay mixture, 30% moisture content). (f) Model 6 (2:1 by weight silt-clay mixture, 60% moisture content). The models indicate that the degree of soil deformation during casting increased as the host sediments became finer-grained and more ice-rich. In other words, the ice-wedge shape was generally best preserved in sand, and most modified in clayey silt. Reproduced from Harris C and Murton JB (2005) Experimental simulation of ice-wedge casting: Processes, products and paleoenvironmental significance. In: Harris C and Murton JB (eds.) *Cryospheric Systems: Glaciers and Permafrost*, pp. 131–143. Special Publication 242. London: Geological Society.

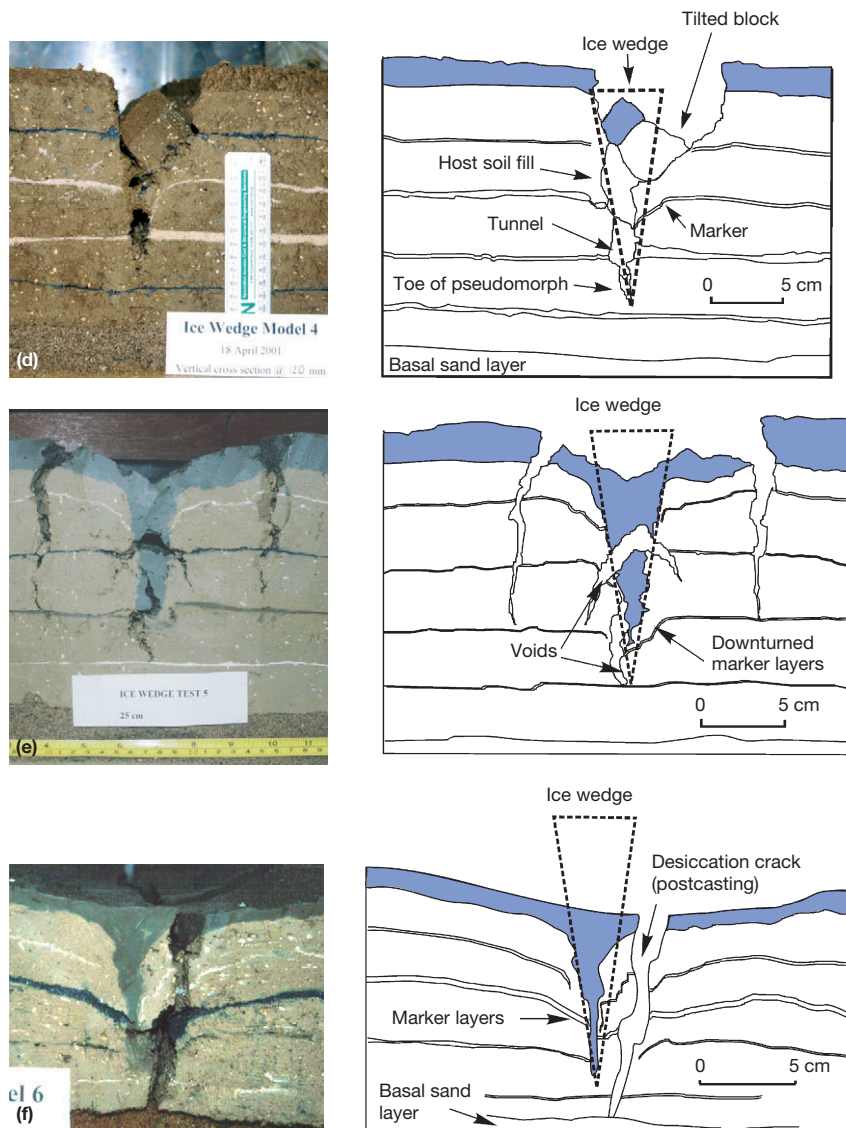


Figure 6 (Continued)

Structures superficially similar to ice-wedge pseudomorphs – for example, solution pipes and tree-root casts – are easily identified by excavating the host material to determine their three-dimensional geometry and composition. But problems may arise when distinguishing between infilled thermal contraction cracks and desiccation cracks in the case of small soil veins and wedges ($\sim < 1$ m high) in silts or clays. The distinction between sand wedges of primary infilling and sand wedges of secondary infilling where upturned strata flank wedges whose fill is massive, well-sorted sand is also problematic (Black, 1976).

Some pseudomorphs develop through a combination of infilling conditions or mechanisms. Active-layer soil wedges ~ 0.1 – 0.5 m wide and ~ 0.3 – 2.3 m deep from Siberia, Spitsbergen and northern Quebec have been attributed to secondary infilling during melt of seasonal ice veins, without a contribution from melting of underlying ice wedges. Some infilling can certainly occur in summer as meltwater allows

soil or rock fragments to flow or slide into voids, producing indistinct vertical layers within soil wedges; consistent with this interpretation are remnants of seasonal ice veins observed at the summer thaw front in soil wedges. But the seasonal interpretation has rightly been questioned where soil wedges overlie ice veins or wedges within permafrost, because the bottoms of the soil wedges may develop where the tops of the ice wedges have melted. Distinguishing between melt of seasonal ice veins and melt of underlying permafrost ice wedges is difficult because active-layer depth varies from year to year and ice-wedge rejuvenation is common. Two conclusions arise. First, active-layer soil wedges form by secondary infilling in different ways, including (1) melt of seasonal ice veins, (2) melt of underlying ice-wedge tops, (3) complete melt of ice wedges and erosion of the tops of ice-wedge pseudomorphs, or (4) a combination of them. Second, some so-called ice-wedge pseudomorphs are probably compound structures,

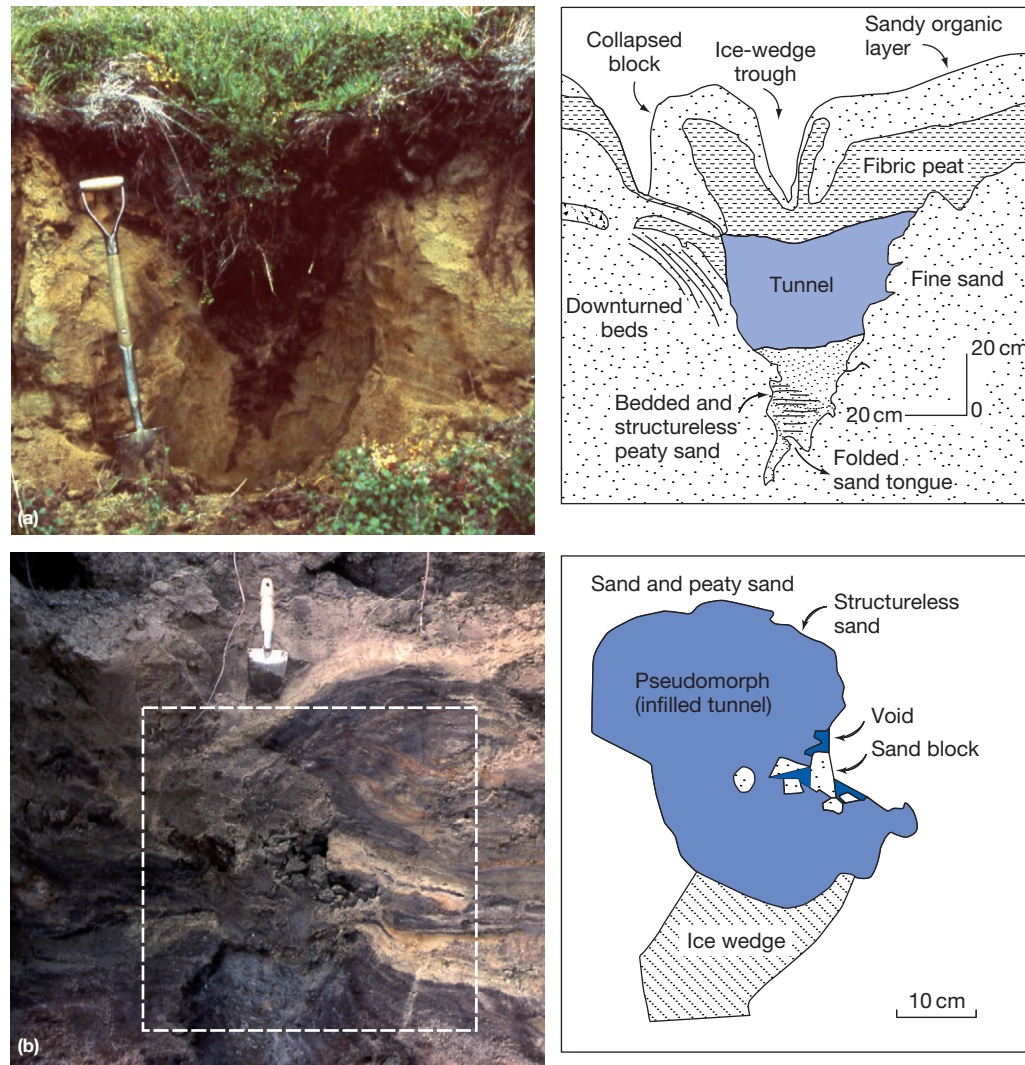


Figure 7 Partially thawed ice wedges and associated pseudomorphs, Mackenzie Delta region, western Arctic Canada. (a) Photograph and diagram of a partly filled tunnel into which a mat of peat and tundra vegetation has subsided where an ice wedge has melted completely at a coastal bluff, Hendrickson Island. Just 3 m inland (not shown), the top of the same ice wedge is at a depth of 45 cm. (b) Photograph and diagram of a tunnel infilled by blocks of fallen sand above a partially thawed ice wedge, Hadwen Island. Reproduced from Murton JB and French HM (1993) Thaw modification of frost-fissure wedges, Richards Island, Pleistocene Mackenzie Delta, western Canadian Arctic. *Journal of Quaternary Science* 8: 185–196.

the lower part comprising a pseudomorph after a permafrost ice wedge, and the upper part an active-layer soil wedge formed by melt of seasonal ice veins.

Composite-wedge pseudomorphs form by a combination of primary and secondary infilling, and therefore share features with both primary soil wedges and ice-wedge pseudomorphs (Figure 9; Ghysels and Heyse, 2006; Gozdzik, 1973). Where composite wedges contain little ice, their pseudomorphs are difficult to distinguish from primary sand wedges. It is likely that some structures interpreted as ice-wedge pseudomorphs in the Quaternary literature are instead composite-wedge pseudomorphs, especially where former vegetation cover was sparse, favoring a combination of blowing soil and snow during winter.

The variable shape and size of pseudomorphs complicates their identification. Pseudomorphs that taper uniformly

downward like a typical epigenetic ice wedge (Figure 8(a)–8(f)) are easier to identify than pseudomorphs that are irregular (Figure 8(g) and 8(h)). Such irregularity often results from slumping, thermal erosion and rejuvenation of wedges. In terrain subject to thermal erosion and refreezing of water-filled cavities, melting of ice wedges truncated by irregular bodies of pool ice may produce irregular pseudomorphs, although their preservation potential is uncertain. Melting and secondary infilling of rejuvenated ice wedges may also produce irregular and complex pseudomorphs. In terms of size, pseudomorphs vary in maximum width from a few millimeters to a few meters and in height from a few tens of centimeters to several meters, although truncated pseudomorphs are common. Small, narrow pseudomorphs are often the most difficult to identify because melt of ice veins in cohesive materials does not always lead to downturning or faulting of adjacent sediment.

Table 3 Characteristic features of ice-wedge pseudomorphs and composite-wedge pseudomorphs

<i>Pseudomorph characteristics</i>	<i>Suggested origin</i>	<i>Comment</i>
Infill is similar to material above or beside it	Subsidence of host or overlying material into space formed by melting ice wedge	
Massive infill, often poorly sorted and lacking systematic vertical trends in particle size. Vertical to steeply dipping layers, often indistinct, within some pseudomorphs	Collapse of overlying or adjacent sediment into void	Infill often has a chaotic appearance and may be looser than host material
Vertical to steeply dipping layers, often indistinct, within some pseudomorphs	Subsidence of adjacent, stratified host material	Layers tend to be thicker than veins of primary infilling and can often be traced upward and then sideways, joining the adjacent host strata, which therefore precludes primary infilling
Concave-up sediment layers, stacked within gravelly pseudomorphs	Gradual subsidence or slumping of overlying material during downward melt of ice wedge	Observed in some large ice-wedge pseudomorphs within stratified fluvial sand and gravel
Large clasts or faulted blocks of sediment within pseudomorph	Subsidence or slumping	Material is too large to have entered into original crack by primary infilling
Elongate clasts are vertically to steeply aligned within pseudomorph	Realignment of elongate clasts in narrow gap along sides of inward-melting ice wedge	Common in gravel or diamicton sequences
Lining: a layer of fine sediment along the sides of pseudomorphs	Deposited by water seeping down the sides of melting ice wedges	Tends to form near bottom of pseudomorphs
Tunnels or irregular voids infilled to various degrees	Tunnels often develop by thermal erosion of ice-wedge ice and then infill to varying degrees with sediment	Tunnel infill is often poorly consolidated
Polygonal configuration in plan	Thermal contraction cracking	Polygons are sometimes seen in vegetation patterns on aerial photographs. Polygons are characteristic of wedges of both primary and secondary infilling but are not always present during early stages of wedge development
<i>Adjacent host material</i>		
Downturned strata	Subsidence due to loss of lateral support as ice wedge melts	Downturning does not always occur, and traces of original upturned strata are sometimes preserved, especially in thaw-stable or very cohesive hosts (e.g., brecciated bedrock, peat, compact clay); infilling occurs mainly from above
Normal faults, often stepped	Brittle failure associated with subsidence or collapse of host material	Common in sand

Paleoenvironmental Significance of Veins and Wedges

Relict Ice Wedges

Ice wedges can persist in permafrost environments for decades to many thousands of years after they cease to crack and grow. Such inactive or relict wedges are widespread in permafrost sedimentary sequences in arctic and sub-arctic lowlands of Canada, Alaska and Siberia, with the oldest known wedges probably exceeding 740 000 years in age (Froese et al., 2008), in contrast to some Holocene wedges that have only been inactive during the twentieth century (Kokelj et al., 2007). Analysis of the crystallography and stable oxygen and hydrogen isotopes of the ice and of the gas composition (O_2 , N_2 , Ar) of air entrapped in the ice can elucidate the source and mechanism of infill of the ice wedges. From these analyses, inferences can be drawn about past moisture and winter

temperature conditions (e.g., Popp et al., 2006; St-Jean et al., 2011; Streletskaia et al., 2011).

Ice-Wedge Pseudomorphs

Ice-wedge pseudomorphs are frequently used to estimate former maximum values of mean annual air temperature (MAAT) and the mean temperature of the coldest month (MTCM). Widely used estimates are $\leq -4^\circ\text{C}$ and $\leq -8^\circ\text{C}$ (to -6°C) for the MAAT associated with fine-grained and coarse-grained substrates, respectively, and $\leq -20^\circ\text{C}$ for the MTCM. The widespread use of ice-wedge pseudomorphs as paleotemperature indicators arose from studies of ice wedges in Alaska (Péwé, 1966). There, the area of active ice wedges approximates the continuous permafrost zone, and has a MAAT ranging from $\sim -6^\circ\text{C}$ to -8°C in the south to $\sim -12^\circ\text{C}$ in the north. Péwé (1966) concluded that “[i]f ice wedges can be

proven to have existed in the geologic past in a particular area, it can be broadly assumed that, at the time of ice-wedges formation, a rigorous permafrost environment existed with a [MAAT] in the vicinity of -6 to -8 °C, or colder."

Unfortunately, several problems bedevil the Holy Grail of a simple relationship between ice-wedge growth and MAAT (Harry and Goździk, 1988; Murton and Kolstrup, 2003). First, many site-specific and time-dependent factors influence cracking and wedge development. Second, it is the rapidity and amount of temperature change in the upper few meters of frozen ground, rather than the air temperature, that is important to cracking. Because ground-surface temperatures in northern latitudes differ from MAATs by variable amounts (often ~ 1 – 6 °C), depending on near-surface factors such as snow cover and vegetation, "no simple relationship exists (between cracking and MAAT) except that colder temperatures regionally promote more contraction cracking" (Black, 1976). Third, ice-wedge cracking occurs in winter, whereas MAAT also takes into account temperatures in summer. Fourth, the long periods over which MAATs are compiled may conceal significant short-term climatic variation. Rare, unusually cold winters allow intermittent ice-wedge growth in the discontinuous

permafrost zone. This explains why the maximum proposed value of MAAT at which ice-wedge cracking can occur has increased from Péwé's original value of -6 to -8 °C to values of -4 °C or -3.5 °C. In conclusion, ice-wedge pseudomorphs simply indicate the former occurrence of permafrost: most likely, but not inevitably, continuous permafrost.

Relict Primary Sand Wedges

Relict primary sand wedges indicate the former occurrence of (1) sand transport, often by wind, during winter or spring; and (2) little or no snow cover, for otherwise the cracks may fill with ice or ice–sand mixtures. These conditions have generally been attributed to (1) climatic aridity in polar deserts or steppe tundra; (2) dry, windswept environments marginal to Pleistocene ice sheets; (3) edaphic aridity due to an absence of permafrost; (4) locally dry surface conditions such as exposed, elevated areas and well-drained, coarse-grained sediments or simply local areas where plenty of wind and a sand source are available; and (5) widespread ground instability and sparse vegetation cover due to active geomorphic processes and rapid climatic change (Murton et al., 2000).



Figure 8 Ice-wedge pseudomorphs in deposits of Cenozoic age. Large pseudomorphs are common in periglacial fluvial sand and gravel, as shown from (a) Whisby quarry, near Lincoln, England (spade for scale); and (b) Cassington quarry near Oxford, England (survey tripod for scale). In relatively homogeneous sediments, pseudomorphs can be marked out by distinctive downturned layers, such as (c) gray gleysol in reworked loess, Kesselt, Belgium (spade for scale); and (d) white tephra in gravel, ~ 35 km southeast of Dawson City, Yukon, Canada (knife and axe for scale). (e) Vertically aligned to steeply dipping clasts within pseudomorph from Sewerby, East Yorkshire, England (2.75 m high section). (f) The smallest pseudomorphs are veins of secondary infilling, as shown by sand veins from the Dinkel valley, the Netherlands (pen for scale). (g and h) Deformed, irregular pseudomorphs in gravel, Donnelly Dome area, Alaska. (a–f) © 2006, Photographs by JB Murton. (g and h) Reproduced from Pewe TL, Church RE, and Andresen MJ (1969) *Origin and Paleoclimatic Significance of Large-Scale Patterned Ground in the Donnelly Dome Area, Alaska*. Special Publication 103. Boulder, CO: Geological Society of America.

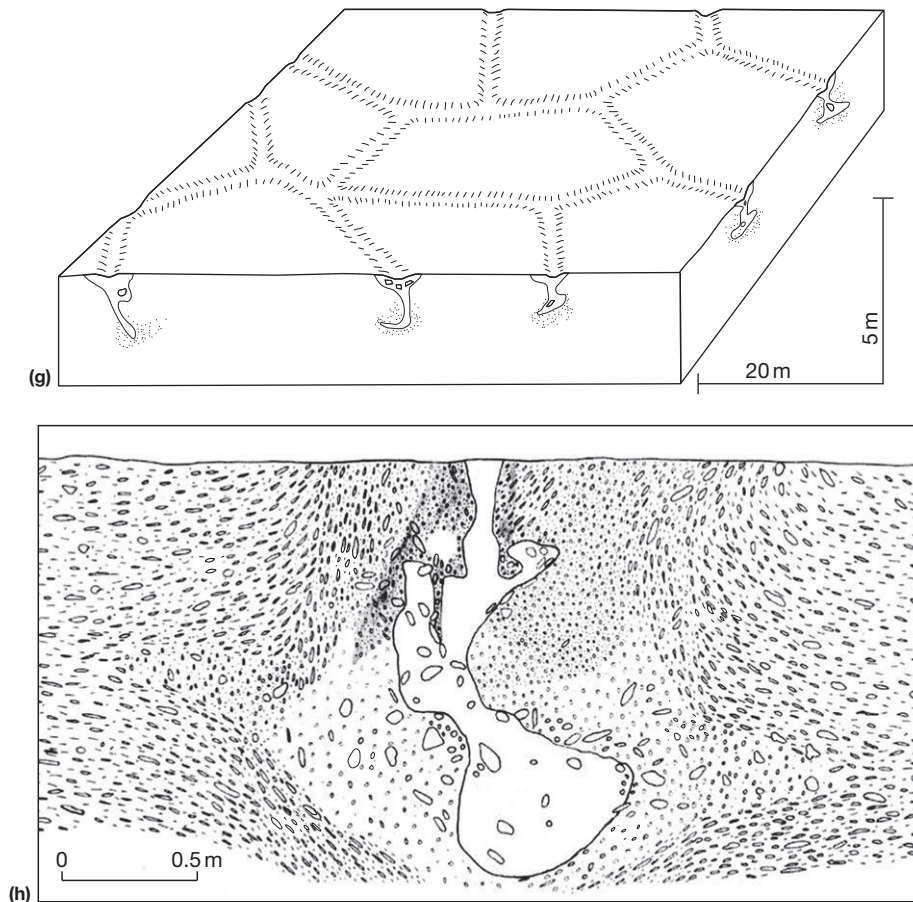


Figure 8 (Continued)

Relict primary sand wedges >2 m high and with well-developed vertical lamination usually indicate the former occurrence of continuous permafrost. This inference is based on the maximum reported depth of seasonal frost cracks (~2 m) and the occurrence of repeated cracking and infilling, which is most likely where permafrost is continuous.

Soil Veins and Small Soil Wedges

Relict soil veins and small wedges indicate the former occurrence of thermal contraction cracking but do not distinguish between permafrost, active layers or seasonally frozen ground. Maarleveld (1976) inferred a MAAT of 0 °C for narrow frost cracks (soil veins) in Pleistocene sediments of the central Netherlands, and a MAAT of between 0 and –6 °C for larger frost cracks transitional with ice-wedge pseudomorphs. Such estimates, however, are of questionable validity, not only for reasons discussed earlier, but also because soil veins and small soil wedges are found throughout contemporary cold regions.

Selective Preservation

Paleoenvironmental interpretation of relict veins and wedges is biased by their selective preservation in the geological record. Primary soil veins and wedges have the highest preservation

potential due to the paucity or absence of ice within the infill; thaw-related disturbance occurs only where soil wedges penetrate ice-rich materials. Ice wedges tend to be larger or more common in ice-rich silts, clays or peats, whereas their pseudomorphs form preferentially and are better preserved in ice-poor sand and gravel, which deform little during melt of constituent ice wedges (Black, 1976). Such selective preservation has probably led to an underestimation of the number of ice-wedge pseudomorphs in fine-grained soils that were originally ice-rich. This biases reconstructions toward colder temperatures.

Another form of selective preservation favors identification of former ice wedges with simple shapes. Because of thaw disturbance, the preservation potential of syngenetic ice wedges with serrated edges or of rejuvenated ice wedges with pyramidal tops is probably much less than wedges with a simple V shape.

Conclusions

Thermal contraction cracks may fill with surface materials, narrow before infilling, or close without infilling. Infills comprise ice, mineral particles or organic material, and form a variety of veins and wedges of primary infilling (Table 1). Different types of wedges develop according to the stability of the ground surface and permafrost table. Melt of ice veins and wedges sometimes initiates secondary infilling of the space left by

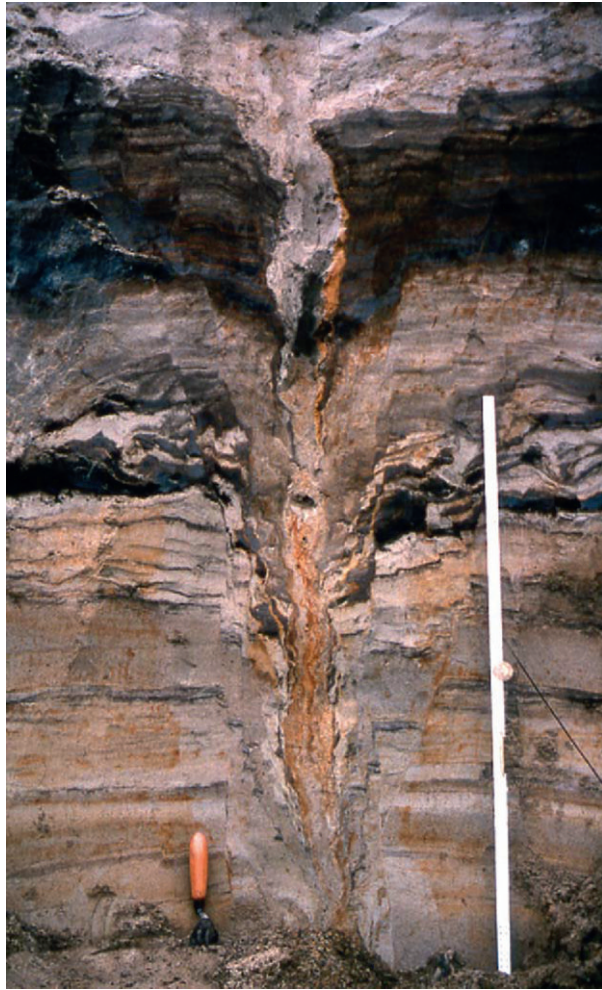


Figure 9 Composite-wedge pseudomorph in sands of the Cromer Forest Bed Formation, near Trimingham, north Norfolk, England. The central (orange) part of the wedge comprises well-sorted, vertically laminated sands typical of a sand wedge of primary infilling, whereas the outer part – in the host sand – comprises step normal faults typical of secondary infilling associated with the formation of an ice-wedge pseudomorph. The black horizon near the top is the Cromer Forest Bed. The exposed wedge is ~2 m high. © 2006, Photograph by R.J.O. Hamblin. Published with permission of the British Geological Survey. The wedge is described in Fish PR, Carr SJ, Rose J, Hamblin RJO, and Eissmann L. 1996. A periglacial composite wedge cast from the Trimingham area, North Norfolk, England. *Bulletin of the Geological Society of Norfolk* 46: 11–16.

melting ice. The resulting ice-wedge pseudomorphs provide reliable evidence for the former occurrence of permafrost but they do not provide reliable quantitative estimates of MAATs. A more promising field of quantitative paleoenvironmental study that remains to be fully explored concerns the inferences that could be drawn from detailed analyses of the primary infillings of large syngenetic and anti-syngenetic wedges, particularly in terms of isotopes, organic material and eolian sand.

See also: **Permafrost and Periglacial Features: Active Layer Processes; Patterned Ground; Permafrost.**

References

- Bateman MD, Murton JB, and Boulter C (2010) The source of De variability in periglacial sand wedges: Depositional processes versus measurement issues. *Quaternary Geochronology* 5: 250–256.
- Black RF (1974) Ice-wedge polygons of northern Alaska. In: Coates DR (ed.) *Glacial Geomorphology*, pp. 247–275. Binghamton: State University of New York.
- Black RF (1976) Periglacial features indicative of permafrost: Ice and soil wedges. *Quaternary Research* 6: 3–26.
- Bockheim J, Coronato A, Rabassa J, Ercolano B, and Ponce J (2009a) Relict sand wedges in southern Patagonia and their stratigraphic and paleo-environmental significance. *Quaternary Science Reviews* 28: 1188–1199.
- Bockheim J, Kurz MD, Soule SA, and Burke A (2009b) Genesis of active sand-filled polygons in Lower and Central Beacon Valley, Antarctica. *Permafrost and Periglacial Processes* 20: 295–308.
- Buylaert JP, Ghysels G, Murray AS, et al. (2009) Optical dating of relict sand wedges and composite-wedge pseudomorphs in Flanders, Belgium. *Boreas* 38: 160–175.
- de Leffingwell EK (1915) Ground-ice wedges – The dominant form of ground-ice on the north coast of Alaska. *Journal of Geology* 23: 635–654.
- Dostovalov BN and Popov AI (1966) Polygonal systems of ice-wedges and conditions of their development. In: *Permafrost International Conference, Proceedings, Lafayette, Indiana, 11–15 November 1963*, pp. 102–105. Washington, DC: National Academy of Sciences–National Research Council Publication 1287.
- Dylik J (1966) Problems of ice-wedge structures and frost-fissure polygons. *Biuletyn Peryglacjalny* 15: 241–291.
- Fortier D, Allard M, and Shur Y (2007) Observation of rapid drainage system development by thermal erosion of ice wedges on Bylot Island, Canadian Arctic Archipelago. *Permafrost and Periglacial Processes* 18: 229–243.
- Froese DG, Westgate JA, Reyes AV, Enkin RJ, and Preece SJ (2008) Ancient permafrost and a future, warmer Arctic. *Science* 321: 1648.
- Ghysels G and Heyse I (2006) Composite-wedge pseudomorphs in Flanders, Belgium. *Permafrost and Periglacial Processes* 17: 145–161.
- Gozdzik JS (1973) Geneza i pozycja stratygraficzna struktur peryglacjalnych w Polsce (summary: Origin and stratigraphical position of periglacial structures in Middle Poland). *Acta Geographica Lodziensis* 31: 104–117.
- Harris C and Murton JB (2005) Experimental simulation of ice-wedge casting: Processes, products and palaeoenvironmental significance. In: Harris C and Murton JB (eds.) *Cryospheric Systems: Glaciers and Permafrost*, pp. 131–143. London: Geological Society Special Publication 242.
- Harry DG and Gozdzik JS (1988) Ice wedges: Growth, thaw transformation, and palaeoenvironmental significance. *Journal of Quaternary Science* 3: 39–55.
- Kanevskiy M, Shur Y, Fortier D, Jorgenson MT, and Stephani E (2011) Cryostratigraphy of late Pleistocene syngenetic permafrost (yedoma) in northern Alaska, Itkillik River exposure. *Quaternary Research* 76(3): 584–596.
- Kokelj SV, Pisarcic MFJ, and Burn CR (2007) Cessation of ice-wedge development during the 20th century in spruce forests of eastern Mackenzie Delta, Northwest Territories, Canada. *Canadian Journal of Earth Sciences* 44: 1503–1515.
- Kolstrup E and Mejdahl V (1986) Three frost wedge casts from Jutland (Denmark) and TL-dating of their infill. *Boreas* 15: 311–321.
- Lachenbruch AH (1962) *Mechanics of Thermal Contraction Cracks and Ice-Wedge Polygons in Permafrost*. Baltimore, MD: Geological Survey of America Special Paper 70.
- Lachenbruch AH (1966) Contraction theory of ice-wedge polygons: A qualitative discussion. In: *Permafrost International Conference, Proceedings, Lafayette, Indiana, 11–15 November 1963*, pp. 63–71. Washington, DC: National Academy of Sciences–National Research Council Publication 1287.
- Lewkowicz AG (1994) Ice-wedge rejuvenation, Fosheim Peninsula, Ellesmere Island. *Permafrost and Periglacial Processes* 5: 251–268.
- Maarleveld GC (1976) Periglacial phenomena and the mean annual temperature during the Last Glacial Time in the Netherlands. *Biuletyn Peryglacjalny* 26: 57–78.
- Mackay JR (1974) Ice-wedge cracks, Garry Island, Northwest Territories. *Canadian Journal of Earth Sciences* 11: 1366–1383.
- Mackay JR (1975) The closing of ice-wedge cracks in permafrost, Garry Island, Northwest Territories. *Canadian Journal of Earth Sciences* 12: 1668–1674.
- Mackay JR (1990) Some observations on the growth and deformation of epigenetic, syngenetic and anti-syngenetic ice wedges. *Permafrost and Periglacial Processes* 1: 15–29.
- Mackay JR (1992) The frequency of ice-wedge cracking (1967–1987) at Garry Island, western Arctic coast, Canada. *Canadian Journal of Earth Sciences* 29: 236–248.
- Mackay JR (1993) Air temperature, snow cover, creep of frozen ground, and the time of ice-wedge cracking, western Arctic coast. *Canadian Journal of Earth Sciences* 30: 1720–1729.

- Mackay JR (1995) Ice wedges on hillslopes and landform evolution in the late Quaternary, western Arctic coast, Canada. *Canadian Journal of Earth Sciences* 32: 1093–1105.
- Mackay JR (2000) Thermally induced movements in ice-wedge polygons, western arctic coast: A long-term study. *Geographie Physique et Quaternaire* 54: 41–68.
- Mackay JR and Burn CR (2002) The first 20 years (1978–1979 to 1998–1999) of ice-wedge growth at the Illisarvik experimental drained lake site, western Arctic coast, Canada. *Canadian Journal of Earth Sciences* 39: 95–111.
- Murton JB and Bateman MD (2007) Syngenetic sand veins and anti-syngenetic sand wedges, Tuktoyaktuk Coastlands, Western Arctic Canada. *Permafrost and Periglacial Processes* 18: 33–47.
- Murton JB and French HM (1993) Thaw modification of frost-fissure wedges, Richards Island, Pleistocene Mackenzie Delta, western Canadian Arctic. *Journal of Quaternary Science* 8: 185–196.
- Murton JB, French HM, and Lamothe M (1997) Late Wisconsinan erosion and aeolian deposition, Summer and Hadwen Islands, Mackenzie Delta area, western Canadian Arctic: Optical dating and implications for glacial chronology. *Canadian Journal of Earth Sciences* 34: 190–199.
- Murton JB and Kolstrup E (2003) Ice-wedge casts as indicators of palaeotemperatures: Precise proxy or wishful thinking? *Progress in Physical Geography* 27: 155–170.
- Murton JB, Worsley P, and Gozdzik J (2000) Sand veins and wedges in cold aeolian environments. *Quaternary Science Reviews* 19: 899–922.
- Péwé TL (1959) Sand-wedge polygons (tessellations) in the McMurdo Sound Region, Antarctica – A progress report. *American Journal of Science* 257: 545–552.
- Péwé TL (1966) Palaeoclimatic significance of fossil ice wedges. *Biuletyn Peryglacjalny* 15: 65–73.
- Péwé TL, Church RE, and Andresen MJ (1969) *Origin and Paleoclimatic Significance of Large-Scale Patterned Ground in the Donnelly Dome Area, Alaska*. Boulder, CO: Geological Society of America Special Publication 103.
- Popp S, Diekmann B, Meyer H, Siegert C, Syromyatnikov I, and Hubberten HW (2006) Palaeoclimate signals as inferred from stable-isotope composition of ground ice in Verkhoyansk foreland, Central Yakutia. *Permafrost and Periglacial Processes* 17: 119–132.
- Romanovskij NN (1973) Regularities in formation of frost-fissures and development of frost-fissure polygons. *Biuletyn Peryglacjalny* 23: 237–277.
- Romanovskij NN (1985) Distribution of recently active ice and soil wedges in the U.S.S.R.. In: Church M and Slaymaker O (eds.) *Field and Theory: Lectures in Geocryology*, pp. 154–165. Vancouver: University of British Columbia Press.
- Sletton RS, Hallet B, and Fletcher RC (2003) Resurfacing time of terrestrial surfaces by the formation and maturation of polygonal patterned ground. *Journal of Geophysical Research* 108(E4): 8044. <http://dx.doi.org/10.1029/2002JE001914>.
- St-Jean M, Lauriol B, Clark ID, Lacelle D, and Zdanowicz C (2011) Investigation of ice-wedge infilling processes using stable oxygen and hydrogen isotopes, crystallography and occluded gases (O₂, N₂, Ar). *Permafrost and Periglacial Processes* 22: 49–64. <http://dx.doi.org/10.1002/ppp.680>.
- Streletskaia I, Vasiliev A, and Meyer H (2011) Isotopic composition of syngenetic ice wedges and palaeoclimate reconstruction, western Taymyr, Russian Arctic. *Permafrost and Periglacial Processes*, <http://dx.doi.org/10.1002/ppp.707>.
- Yershov ED (1998) *General Geocryology*. Cambridge: Cambridge University Press.

Patterned Ground

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Introduction

The term *patterned ground* describes terrain that exhibits regular or irregular surface patterning, most commonly in the form of circles, polygons, irregular networks, or stripes. Step-like, oval, lobate, and garland patterns are also sometimes referred to as patterned ground. Following Washburn (1956), most investigators have recognized a distinction between *sorted patterned ground* produced by the alternation of soil and coarse debris at the ground surface, and *nonsorted patterned ground* defined by microrelief (alternating mounds and depressions, or ridges and furrows) or vegetation cover (alternating vegetated and unvegetated ground). Although various forms of ground patterning occur in a range of environmental settings, most patterned ground is associated with periglacial environments where recurrent freezing and thawing of moist soil is the dominant formative mechanism, though neither permafrost nor severe winter cold is required for the formation of all periglacial patterned ground types. Active patterned ground produced by ground freezing and thawing occurs in high-latitude permafrost environments, in subarctic areas subject to deep seasonal ground freezing, on alpine and high subtropical mountains, and on mountains in cool temperate environments. Certain forms of patterned ground produced by shallow ground freezing also occur at low altitudes in cool temperate environments (e.g., Ballantyne, 1996), but these are usually small-scale features that have formed under exceptional circumstances. On exposed plateaus, the interplay of wind scour, loosening of soil particles by frost, and selective vegetation growth also results in the development of surface patterns in the form of clumps, crescents, ribbons, and stripes of vegetation amid unvegetated stony ground (Ballantyne and Harris, 1994, pp. 260–261), but such wind-patterned ground is genetically unrelated to patterned ground created by freezing and thawing of the soil and so, it is not considered further here.

Most research on periglacial patterned ground has involved the study of mature landforms, and the processes responsible for pattern initiation have often been inferred from present-day pattern characteristics, internal structure, and measurement of soil and clast displacement. There has also been a tendency for studies to focus on individual pattern types, leading to numerous theories of pattern origin. Washburn (1956), for example, examined 19 hypotheses for pattern initiation and development and inferred a multiplicity of formative mechanisms. Recent numerical modeling of ground freezing and thawing, however, suggests that most periglacial patterned ground has been initiated by differential frost heave of frost-susceptible soil, and that the differences between sorted patterns and nonsorted patterns may reflect initial conditions, particularly the presence or absence of vegetation cover and the abundance of clasts. *Frost heave* is uplift of soil due to the formation of lenses of segregated ice in fine-textured soils and *differential frost heave* is laterally nonuniform frost heave,

which produces an undulating surface microtopography. The application of linear stability analysis to freezing soils (Kessler et al., 2001; Peterson, 2011; Peterson and Krantz, 2003, 2008) has demonstrated that one-dimensional frost heave is unstable under a range of boundary conditions and tends to develop into differential frost heave, with the eventual development of a dominant modal wavelength that determines pattern spacing. Various feedback processes, still incompletely understood, are thought to enhance the initial perturbation caused by differential frost heave, permitting the eventual development of both sorted and nonsorted ground patterns.

Sorted Patterned Ground

Characteristics

Sorted patterns develop on unvegetated or sparsely vegetated terrain where abundant clasts overlie or are embedded in fine soil. On level or gently sloping ground, sorted patterns usually consist of cells or domains of predominantly fine soil surrounded by stony borders that in the thawed state may be depressed or raised relative to the fine centers (Figures 1–3), though in active patterns, the fine cell is often heaved above the level of surrounding clasts during winter freezing. These concentrations of clasts often take the form of regularly spaced sorted circles or polygons, or an irregular sorted net or labyrinth of clasts separating fine cells of varying size and shape. Less commonly, stone pits develop, characterized by concentrations of clasts surrounded by raised borders of fine-grained sediment. All sorted patterns tend to become increasingly elongated along the direction of maximum gradient as slope angle increases, grading downslope into sorted stripes consisting of alternating bands of fine and coarse debris (Figure 4).

Reported pattern widths of sorted nets, circles, polygons, and stripes range from about 10 to 20 m, though accounts of forms exceeding 4 m in width are rare and probably reflect the concentration of clasts into the troughs of ice-wedge polygons (see below). Pattern width is often related to the coarseness of clasts that define the pattern margins (Goldthwait, 1976): patterns defined by small pebbles rarely exceed 0.4 m in width, and patterns defined by bouldery margins usually exceed 1–2 m in width. Within the coarse borders, clast size tends to diminish with depth, and in active patterns, clasts in pattern borders are often erect (standing on end) as a result of lateral pressures exerted during freezing and expansion of fine centers in winter (Figure 5). In permafrost environments, striking examples of large sorted circles (Figure 6) consist of a central slightly domed plug or cylinder of fine soil 2–3 m wide surrounded by an annulus of gravel that rises up to 0.5 m above the soil domain (Hallet and Prestrud, 1986; Washburn, 1989), with both the central plug and surrounding gravel rings extending downward to the underlying permafrost table. Elsewhere, sections excavated through both small- and large-scale sorted



Figure 1 Active sorted circles developed in floodplain deposits, Fauldalen, northern Norway. Pattern width 1.0–1.5 m.



Figure 2 Large-scale sorted net (pattern widths 2–4 m) developed on a periglacial blockfield at 1700 m on Mount Bitterness, St Mary's Range, South Island, New Zealand.



Figure 3 Sorted nets 0.6–1.1 m wide at a site of intermittent flooding, Jotunheimen Massif, south-central Norway.



Figure 4 Active sorted stripes with a pattern width averaging 0.6 m developed on felsite regolith at 600 m on Tinto Hill, southern Scotland. These stripes are believed to have formed through a combination of shallow ground freezing and differential needle-ice growth with concurrent downslope migration of both clasts and fine soil. When this site was dug over to a depth of 0.3 m, stripes reformed within 3 years.



Figure 5 Erected clasts at the margin of a large sorted net on Mount Bitterness, St Mary's Range, South Island, New Zealand. Clast erection is due to lateral pressures exerted on the coarse border during winter freezing and expansion of adjacent fine sediments.

patterns often show that clasts forming coarse borders occupy V-shaped troughs in finer soil. Sorted patterns may be active (largely unvegetated and subject to displacement of both fine sediment and clasts as a result of recurrent ground freezing and thawing; [Figures 1–6](#)) or relict features with vegetated fine-grained cells and lichen-covered bouldery margins ([Figure 7](#)). Relict large-scale patterns defined by boulders have often survived intact throughout the Holocene ([Ballantyne and Harris, 1994](#); [Grab, 2002](#)), but small-scale patterns have limited preservation potential. Some sorted patterns exhibit secondary



Figure 6 Large active sorted circles comprising a central soil plug surrounded by a raised annulus of gravel, Spitsbergen. Photograph by Bernard Hallet.



Figure 7 Relict elongated sorted nets 1–2 m wide, Grampian Highlands, Scotland.

sorting, with decimetric patterns defined by smaller clasts forming inside the fine domain of a larger sorted circle, net, or polygon.

Formation

Most researchers have envisaged the development of sorted patterns in terms of a circulatory model involving upward and/or outward movement of sediment in cells, migration of surface and near-surface sediment from cells to adjacent borders, and compensatory movement of soil toward cell centers at depth. Sorting occurs when sediment moving outward toward the margins at or near the soil surface is coarser than that moving inward toward cell centers at depth. Measurements of sediment displacement on and within active sorted circles (Hallet, 1998; Washburn, 1989) appear consistent with this general model. Sorted stripes develop on slopes because a downslope component of sediment movement is superimposed on the soil circulation. There is, however, less consensus regarding the processes driving sediment movement. Three main groups of theories have emerged: sorting by differential frost heave, sorting by buoyancy-driven soil circulation, and sorting associated with free convection of soil porewater during thaw.

Differential frost heave

A long-established view is that lateral displacement of clasts toward pattern margins and fines toward pattern centers reflects an uneven penetration of freezing planes into the ground, resulting in differential frost heave and migration of clasts away from zones of impeded freezing (Nicholson, 1976). Once established, this effect should be self-perpetuating, as the freezing plane will tend to descend more rapidly below zones of clast concentration than below intervening cells of fine sediment, where formation of ice lenses and associated release of latent heat slows the downward passage of the freezing front. The initial development of inclined freezing planes in nonsorted ground has traditionally been explained in two ways: The first involves contraction cracking of the ground surface as a result of rapid cooling or desiccation. The resulting contraction cracks act as loci of rapid ground freezing so that clasts migrate toward and become concentrated in the cracks, forming sorted polygons (Ballantyne and Matthews, 1983). The second, termed ‘primary frost sorting’ by Goldthwait (1976), proposes that randomly spaced concentrations of finer sediment freeze more slowly than surrounding soil, generating inclined freezing planes. Recurrent upward and outward frost sorting of clasts away from such concentrations of finer soil then progressively enhances the textural contrasts between these and adjacent ground. Additionally, because of their greater frost-susceptibility (propensity for ice-lens formation), such concentrations of finer sediment are liable to greater frost heave than surrounding soil so that upheaved clasts undergo gravity-induced creep toward cell margins. Sorted circles or sorted nets may initially be formed, but it is possible that the interaction of adjacent fine cells may ultimately produce sorted polygons. These concepts have been further developed by Van Vliet-Lanoë (1988), who proposed that formation of sorted polygons and contiguous sorted circles is associated with soil cracking and positive frost-susceptibility gradients (reduction in frost-susceptibility with soil depth), whereas isolated or coalescing sorted circles form on soils characterized by negative frost-susceptibility gradients.

An alternative explanation for the role of differential frost heave has been developed by Kessler et al. (2001), who proposed that large-scale sorted circles are self-organized patterns that emerge spontaneously due to the inherent instability of one-dimensional annual frost heave of the active layer, the zone of seasonal freezing and thawing that overlies permafrost. Their model assumes an initial state with a 1-m-thick active layer initially comprising a layer of stones overlying a layer of frost-susceptible soil. They argued that slight positive (i.e., updomed) perturbations at the stone–soil interface are progressively enhanced by the movement of soil inward under the site of the perturbation during freezing (Figure 8), whereas soil movements during thaw are vertical (because of lateral confinement) so that there is a small net increment of soil added under the site of the perturbation with each freeze–thaw cycle. Kessler et al. (2001) argued that this positive feedback is modulated by a concurrent negative feedback, because a descending freezing front will penetrate positive soil perturbations before adjacent regions, pushing soil toward unfrozen regions that surround the perturbation. This negative feedback favors the development of long-wavelength perturbations, achieved through the merger of initial perturbations, until wavelength

(spacing) stabilizes, limited by soil redistribution rate. Through a process of net soil accretion driven by recurrent annual freeze–thaw cycles, the dominant perturbations form soil plugs that slowly rise through the overlying stones toward the surface, expanding both upward and outward. Emergence of soil plugs is followed by updoming and radial expansion of fine soil at the surface, pushing stones to the margins of the plugs until all the soil from the underlying layer is depleted. After formation, sorted circles are maintained by differential frost heave-driven circulation within the soil and stone domains (Figure 9). Kessler et al. (2001) showed that their model successfully replicates large-scale sorted circles with a mean spacing of 3.6 m, comprising a 2.4-m-wide soil domain surrounded by a 1.0-m-wide, 0.3-m-high ring of stones, representing 750 years of annual freeze–thaw cycles, dimensions,

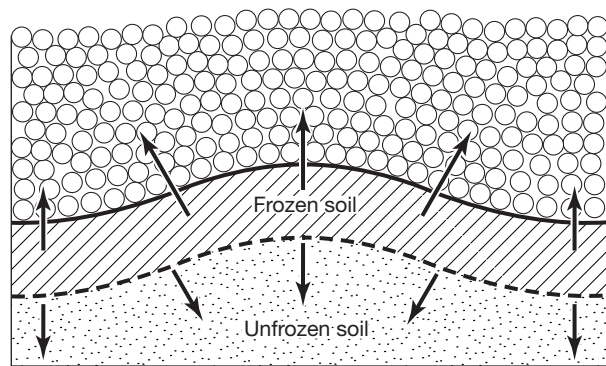


Figure 8 Conceptual model of patterned ground initiation by differential frost heave proposed by Kessler et al. (2001). During annual freezing of the active layer, frost heave drives material away from the freezing front (dashed line) at a small positive perturbation between a layer of soil and an overlying layer of stones. Movement is both outward toward the ground surface and inward toward unfrozen soil. As soil displacements are not reversed during thaw, this mechanism promotes upward growth of a plug of soil over recurrent freeze–thaw cycles. Reproduced, with permission, from Kessler MA, Murray AB, Werner BT, and Hallet B (2001) A model for sorted circles as self-organized patterns. *Journal of Geophysical Research* 106(B7): 13287–13306. ©2001 American Geophysical Union.

and timing that are consistent with those of the large-scale sorted circles investigated by Hallet and Prestrud (1986) in Spitsbergen (Figure 6).

The validity of differential frost heave as a sorting mechanism is well established for shallow sorted patterns with widths <1.0 m, depths <25 cm, and borders dominated by clasts <15–20 cm in length. For such small-scale patterns, the dominant sorting mechanism often appears to be differential needle-ice growth (e.g., Ballantyne, 1996, 2001; Matsuoka et al., 2003). Surface or near-surface ice needles (elongated ice crystals that grow perpendicular to the surface slope) form where the temperature of moist surface soil is reduced to freezing point by radiative cooling. Needle-ice spicules, commonly 2–20 cm in length, frequently form in environments dominated by short-term (mainly diurnal) freezing cycles. The development of small-scale sorted patterns in such environments is constrained by vegetation cover and soil texture: patterns develop only in unvegetated, fine-grained, frost-susceptible soils that permit porewater to be drawn upward to feed needle-ice growth, and where clast size is small relative to pattern dimensions. During freezing, needle ice grows under surface clasts, raising them above the level of the ground surface. As needle ice develops preferentially above patches of fine-grained soil, the layer of upraised clasts forms alternating hummocks and depressions. During thaw, there is preferential movement of clasts into depressions so that over successive cycles of needle-ice growth and decay, the textural differences between fine and coarse soil domains become progressively enhanced until sorted patterns are formed (Figure 10). This may happen quite rapidly. At a site in Scotland, for example, a small-scale sorted net developed in frost-susceptible soil as a result of differential needle-ice growth and differential frost heave in eight freeze–thaw cycles (Ballantyne, 1996), and small-scale sorted stripes attributed to a combination of differential needle-ice growth and downslope migration of sediment have been found to re-form on cleared ground within three winters (Ballantyne, 2001; Figure 4).

Werner and Hallet (1993) have developed a two-dimensional computer simulation of sorted stripe development on the basis of differential needle-ice growth. Their model assumes preferential growth of needle ice in stone-free regions

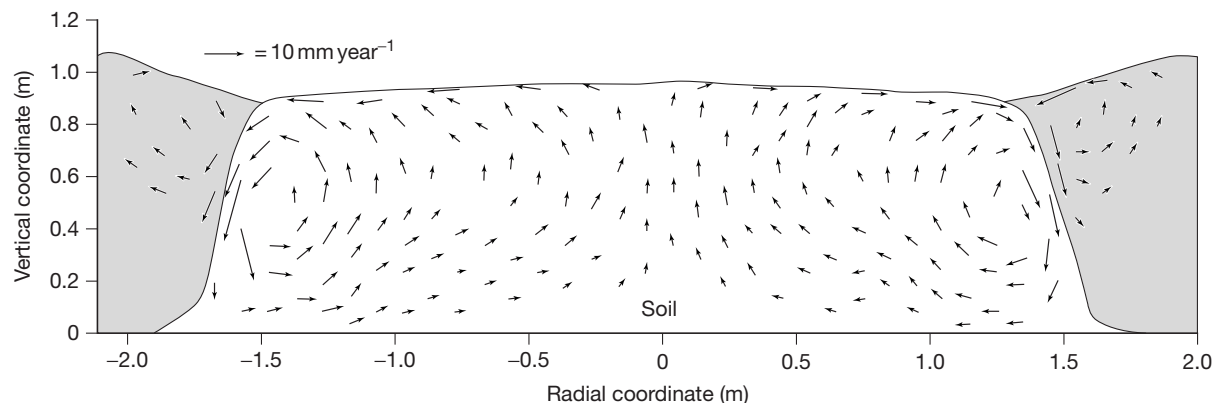


Figure 9 Velocity vectors within a sorted circle modeled by Kessler et al. (2001), simplified from the original diagram. Movement of soil is driven by differential frost heave. Reproduced, with permission, from Kessler MA, Murray AB, Werner BT, and Hallet B (2001) A model for sorted circles as self-organized patterns. *Journal of Geophysical Research* 106(B7): 13287–13306. ©2001 American Geophysical Union.

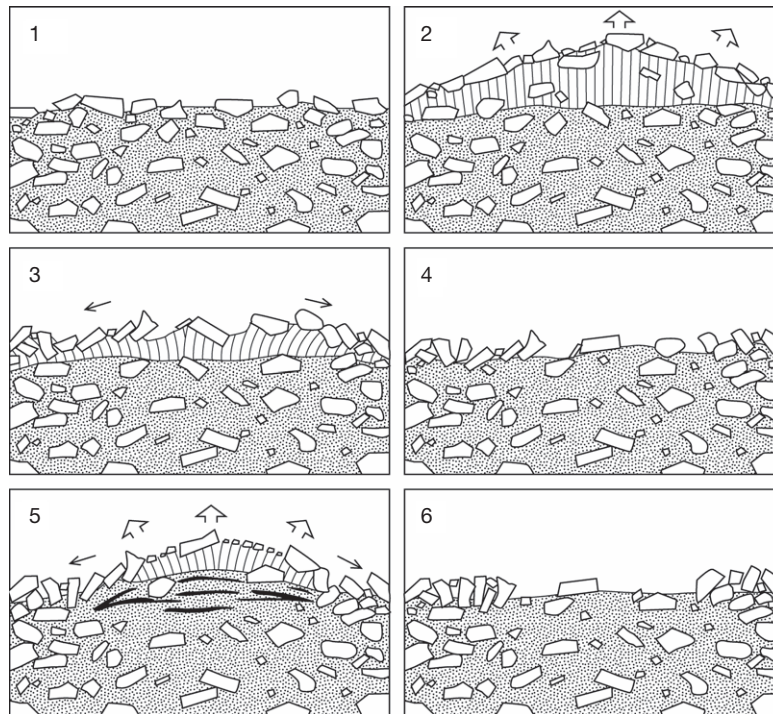


Figure 10 Sequence of sorted pattern development by differential needle-ice growth. 1. Initially unsorted, frost-susceptible soil. 2. Differential growth of needle ice, causing updoming of surface clasts. 3. Thaw of needle ice, with lateral migration of clasts to the margins of domes. 4. Complete thaw and emergence of incipient lateral sorting. 5. Renewed freezing, with preferential needle-ice growth above, and preferential ice-lens formation within, incipient cells of fine soil; lateral movement of clasts toward the margins of the updomed soil. 6. Miniature sorted net, with concentrations of clasts surrounding cells of predominantly fine soil. From Ballantyne CK (1996) Formation of miniature sorted patterns by shallow ground freezing: A field experiment. *Permafrost and Periglacial Processes* 7: 409–424. ©1996 John Wiley & Sons Limited. Reproduced with permission.

during each freezing cycle and calculates the trajectory of iterative clast movements along fall lines determined partly by slope gradient and partly by the location of unfrozen soil domains. They showed that random spatial variations in clast concentration evolve initially into irregular stripes of alternating high and low clast concentration and ultimately into a stable, regularly spaced stripe pattern that is related, as in the field (Figures 4 and 11) to clast size. They also reported that the model generates sorted circles and nets where the surface slope is zero.

Although the development of shallow small-scale sorted patterns by needle-ice growth and differential frost heave is supported by both numerical modeling and observational evidence, the development of sorted patterns >1 m wide by differential frost heave operating alone is less certain. Measurements of soil movements on and within large sorted circles developed above permafrost (Hallet, 1998; Washburn, 1989) indicate the existence of long-term soil circulation with slow ($\sim 10 \text{ mm a}^{-1}$) upwelling of soil at the center of fine cells, radial movement of surface and near-surface soil toward the margins of fine cells, and net subsidence of sediment at their periphery, implying deep-seated return movement of soil toward cell centers at depth. Hallet (1998) noted that differential frost heave produces large seasonal soil displacements, but these are largely compensated by thaw settlement: it is a small annual residual displacement of sediment that contributes to long-term soil circulation. Whether such residual displacement is achieved by differential frost heave alone or depends, in part, on buoyancy forces in thawing soil (see below) is more contentious.

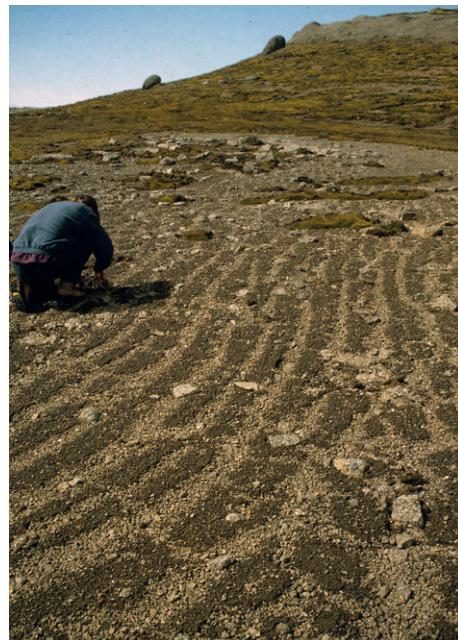


Figure 11 Miniature sorted stripes developed on basalt regolith at 700 m, Faroe Islands.

An important contribution to this debate has been made by Kessler and Werner (2003), who developed a numerical model capable of simulating development of all common forms of periglacial sorted patterned ground. Their model is based on the interaction between two feedback mechanisms: first, the development of inclined freezing planes, with resultant textural segregation between soil-rich and clast-rich domains, as proposed by Kessler et al. (2001); and second, lateral squeezing of clast domains during seasonal freezing and expansion of soil-rich domains. Squeezing and confinement of clast domains are assumed to drive clast movement along the axes of coarse borders, producing clast-dominated margins of regular thickness and width. These two feedback mechanisms were used to drive clast displacements over repeated iterations. As three boundary conditions (clast concentration, slope gradient, and degree of lateral confinement) were varied within the iterative procedure, different forms of sorted patterned ground evolved spontaneously. Decreases in stone concentration produced a transition from sorted circles to nets and ultimately to stone islands (stone pits); increasing slope produced sorted stripes; and stone islands proved transitional to stone polygons with increased lateral confinement (decreased soil compressibility). This model represents a remarkable

advance in understanding how sorted patterns might evolve and self-regulate their form and dimensions under a set of simplifying assumptions, but it may not incorporate the full range of processes involved.

Soil circulation

The theory that sorted patterns may be generated by buoyancy-driven convection cells operating within thawing soils also has a long history and received support in a study by Hallet and Prestrud (1986) of the dynamics of large-scale (3–6 m wide) sorted circles developed in a 1.0–1.5-m-thick active layer in Spitsbergen. Hallet and Prestrud inferred from the updoming of the fine cell centers and subsidence at the fine cell margins that the sorted circles had originated by free convection (circulatory motion caused by the buoyant rise of low-density fluid and the corresponding descent of denser fluid), arguing that the characteristics of sediment movement within the circles could be explained by the establishment of a double circulation cell in the fine domain of each circle (Figure 12(a)). They attributed such soil circulation to a decrease in soil bulk density with depth during seasonal thaw of ice-rich soil, though they stressed that convective soil movement would occur intermittently during thaw (i.e., not simultaneously throughout

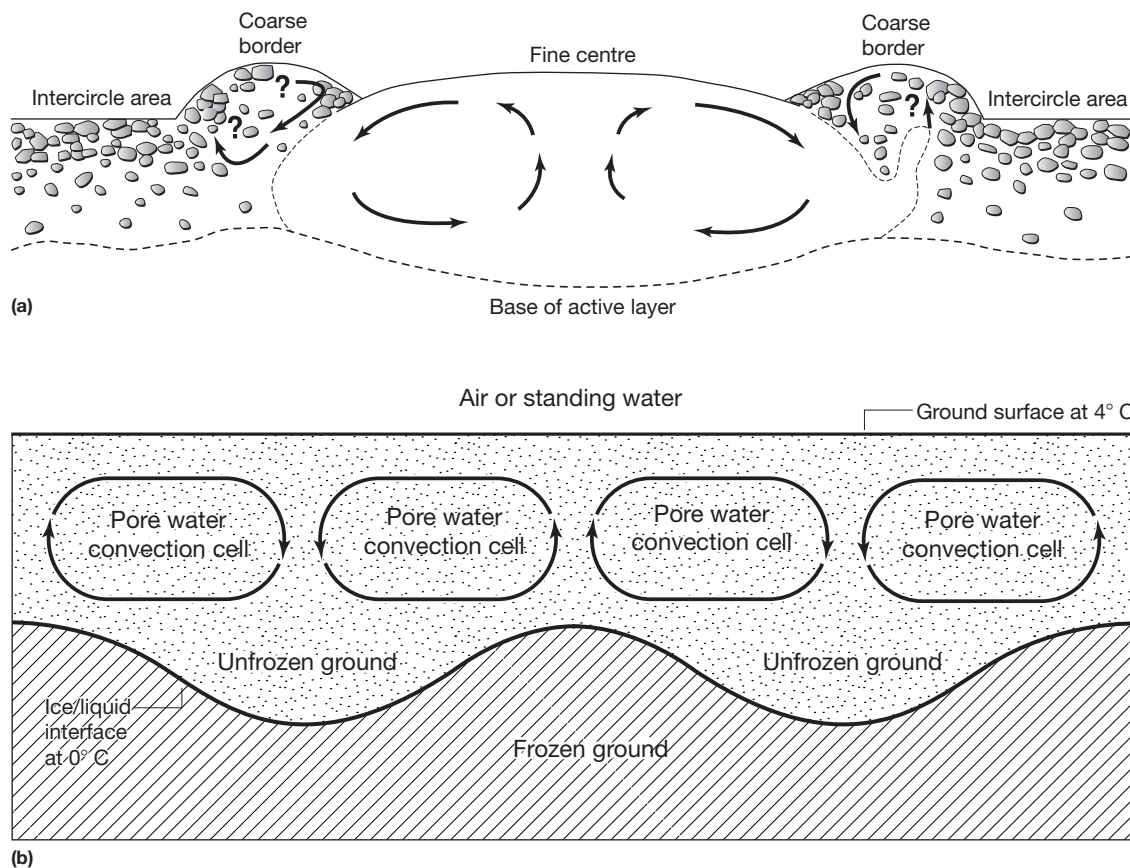


Figure 12 (a) Model of sediment displacements within sorted circles proposed by Hallet and Prestrud (1986). From Hallet B and Prestrud S (1986) Dynamics of periglacial sorted circles in Spitsbergen. *Quaternary Research* 26: 81–99. ©1986 the University of Washington. Reproduced with permission. (b) Model of free convection of porewater during active layer thawing, with resultant development of corrugations in the permafrost table. Based on Krantz WB (1990) Self-organization manifest as patterned ground in recurrently frozen soils. *Earth Science Reviews* 29: 117–130. ©1990 Elsevier Science Ltd. Reproduced with permission.

the full depth of thawed soil) and would be superimposed on cyclic soil displacement due to seasonal freezing and thawing.

Subsequent theoretical analysis by [Hallet and Waddington \(1992\)](#) of thaw consolidation in an ice-rich active layer showed that though the density profile in fine-grained soil above the thaw front may be gravitationally unstable for much of the thaw season, the buoyancy forces generated by density inversion during thaw are insufficient to drive wholesale soil circulation and thus insufficient to initiate lateral sorting. However, their analysis also indicated that buoyant soil generated seasonally at the base of the active layer tends to rise diapirically into the overlying sediment. In preexisting large-scale sorted circles, this may explain evidence for upwelling soil at the centers of fine cells ([Hallet, 1998](#)). They also showed that buoyancy-driven diapirism is likely to be enhanced (1) by a high concentration of ice lenses (due to 'two-sided' freezeback above cold permafrost) at the base of the active layer (cf. [Swanson et al., 1999](#)) and (2) in stratified soils with low-permeability basal layers. It is therefore possible that though buoyancy-driven soil circulation does not drive pattern initiation, it may produce incremental, long-term soil displacements that account, at least in part, for evidence of circulatory soil movements within existing sorted circles, nets, polygons, and stripes developed above permafrost.

Free convection of porewater

An alternative model for the initiation, spacing, and self-regulation of sorted patterns is based on density-driven movements of porewater rather than soil ([Krantz, 1990](#)). Liquid water increases slightly in density as its temperature rises from 0 to 4 °C; during thaw of frozen soil, meltwater at the thaw plane (at 0 °C) is therefore slightly less dense than overlying soil porewater. This model invokes establishment of convection cells during thaw, with descending plumes of relatively dense warm water and ascending plumes of cooler, less dense water. Downflow of warmer water is envisaged as enhancing thaw at the base of the active layer, so that a pattern of alternating peaks and troughs develops at the permafrost table ([Figure 12\(b\)](#)). Although this model predicts pattern width-depth ratios that are consistent with field observations, the link between the postulated development of subsurface corrugations in frozen substrate and that of sorted patterns at the surface has not been established. Moreover, density differences between water at 4 °C and water at 0 °C are very small and produce buoyancy effects two orders of magnitude less than those associated with density inversion in thawing soils ([Hallet and Waddington, 1992](#)). It is questionable whether they are sufficient to drive convective movement of porewater except possibly in very permeable soils.

Synthesis

As outlined earlier, theoretical considerations favor differential frost heave rather than buoyancy-driven soil circulation as the driving mechanism for the initiation of sorted patterned ground. Numerical modeling also suggests that differential frost heave may evolve from uniform one-dimensional frost heave without the need to invoke desiccation cracking, frost cracking, or lateral or vertical variations in the frost-susceptibility of soils, though such preconditions may influence the initial loci of differential frost heave and determine initial pattern spacing. It is widely

accepted that small-scale decimetric patterns can develop in areas of shallow ground freezing as a result of differential frost heave and/or needle-ice growth, with feedback-driven self-regulation producing regular spacing in sorted patterns ([Werner and Hallet, 1993](#)). Differential frost heave may also drive circulatory soil and clast movements to produce large-scale sorted patterns ([Kessler et al., 2001](#)), but it is possible that density inversion during thaw may cause buoyancy-induced soil circulation within such larger patterns.

Environmental Significance of Sorted Patterned Ground

Small-scale sorted patterns produced by differential frost heave or needle-ice growth are known to form as a consequence of shallow (often nocturnal) soil freezing and are therefore of little consequence for retrodiction of former periglacial conditions. Larger forms are diagnostic of deep annual ground freezing and thawing, and [Goldthwait \(1976\)](#) regarded sorted patterns >2 m wide as indicative of permafrost. Theoretically, however, there is no reason why large-scale sorted patterns produced by differential frost sorting cannot occur in areas of deep (>1 m) seasonal ground freezing in the absence of permafrost, provided that the soil contains abundant moisture during freezeback; the main role played by permafrost in promoting differential frost heave is that it impedes downwards movement of soil water. Reviewing the limited climatic data on the conditions under which active large-scale sorted patterns occur, [Grab \(2002\)](#) noted that all sites with active sorted patterns >1 m wide are associated with at least localized permafrost, and suggested that patterns >0.8 m wide occur where mean annual air temperature is –1.6 °C or less. Given the paucity of relevant data, this estimate must be regarded as provisional. There are, moreover, records of active sorted circles and nets exceeding 1–2 m in width in nonpermafrost areas where edaphic, hydrological, or microclimatic conditions are particularly favorable for patterned ground development, such as areas of seasonal waterlogging or glacier margins. Large-scale relict sorted patterns can be used as indicative of former permafrost or deep seasonal freezing if such circumstances can be ruled out.

Nonsorted Patterned Ground

Unlike the various forms of periglacial sorted patterns, all of which have genetic affinities, nonsorted patterns are more diverse in character and in some cases genetically distinct. Some of the most commonly occurring types are considered in the next section.

Ice-Wedge Polygons

The origin and significance of ice-wedge polygons are dealt with in detail in the article on ice wedges and ice-wedge casts, so these features are described only briefly here.

Ice wedges take the form of vertical dykes, V-shaped in cross-section, that penetrate downward into permafrost from the base of the active layer. They are initiated by thermal contraction cracking of near-surface permafrost during rapid winter cooling. Meltwater penetrating such cracks refreezes to

form vertical ice veins. Subsequent thermal contraction reopens cracks along the network of ice veins, so that ice wedges thicken incrementally with each successive episode of contraction cracking and associated meltwater refreezing. In some circumstances, open thermal contraction cracks fill with sediment rather than ice, and the resulting structures are termed primary soil wedges or primary sand wedges, depending on the nature of the infill.

The surface expression of ice wedges, primary soil wedges, and sand wedges is a polygonal network of troughs at the ground surface. Such networks, referred to collectively as frost polygons or frost-crack polygons, define two main types of pattern: low-center polygons, where polygon centers are depressed relative to ridges on either side of the ice-wedge troughs, and high-center polygons, which rise above the troughs as gentle domes. Orthogonal and hexagonal patterns are most common, though pentagonal and irregular forms also occur. Polygon diameters generally range from 5 to 50 m, and larger polygons may be subdivided by a smaller set. Relict frost polygons of late Pleistocene age often show up as crop marks on aerial photographs. Because active ice-wedge polygons are associated with continuous permafrost and severe winter cooling, relict forms indicate similar conditions in the past, though it is often difficult to establish the age of such relict features and their precise implications in terms of former air temperature regime (Murton and Kolstrup, 2003).

Vegetation-Defined Nonsorted Circles (Frost Boils) and Stripes

Nonsorted circles (circular patches of bare or sparsely vegetated soil surrounded by vegetation cover) are widespread in arctic permafrost environments, particularly well-vegetated mid- and low-arctic areas (Walker et al., 2004). Such circles are commonly termed frost boils or mud boils, are usually 0.5–3.0 m in diameter, and occur in isolation, in irregular clusters or as regularly spaced patterns (Figure 13). On gentle to moderate slopes, the equivalent landforms are nonsorted stripes consisting of alternating bands of vegetated and bare soil



Figure 13 Vegetation-defined nonsorted circles (frost boils) on Howe Island, Alaska, in summer. Unvegetated cells of mineral soil are 1–2 m in diameter and slightly domed. Photograph by Anja Kade.

aligned downslope. In winter, nonsorted circles tend to be frost heaved above the level of surrounding vegetated ground (Kade and Walker, 2008; Nicolsky et al., 2008), but in summer, they characteristically form gentle domes. The permafrost table under such nonsorted circles forms a saucer-shaped depression as a result of greater depths of thaw under unvegetated ground. Excavation of some frost boils has revealed diapiric structures indicative of injection of soil from near the permafrost table into overlying layers, and freshly exposed soil at the centers of some circles also suggests injection from below.

Peterson and Krantz (2003) developed a numerical model of nonsorted circle initiation through spontaneous generation of differential frost heave: an initial perturbation in the freezing soil results in conduction of heat from crest regions to adjacent troughs, increasing frost heave under crests and, thus, the amplitude of the perturbation, and driving water movement from troughs to feed ice-lens growth under crests. They employed linear stability analysis to predict the limiting conditions (soil texture and permeability, unfrozen moisture content and overburden pressure) under which differential frost heave occurs, and the characteristic spacing of resultant patterns. In a later paper, Peterson and Krantz (2008) also addressed the subsequent development of nonsorted circles, taking into account the insulating effects of vegetation cover in intercircle areas. They showed that the evolution of differential frost heave throughout the freezing of the active layer leads to the selection of a preferred wavelength that dictates pattern size so that the microtopography of the ground surface evolves away from the initial random configuration of crests and troughs toward a more regular pattern spacing. Peterson (2011) showed that the characteristic size and spacing of emerging nonsorted circles is positively related to active-layer depth, but is also affected by snow depth during ground freezing. Interestingly, the model developed by Peterson and Krantz (2008) suggests that permafrost is not a prerequisite for frost-boil development, provided that sufficient moisture is available; as with large-scale sorted patterns, the main effect of permafrost is to provide an impermeable substrate, so that soil water is abundant during freezing. A different modeling approach employed by Nicolsky et al. (2008) to describe soil and water movement within extant nonsorted circles emphasized the importance of waterlogged conditions and the role of nonuniform vegetation cover in initiating and maintaining differential frost heave, with movement of both water and soil toward the locus of ice-lens formation under bare or thinly vegetated circle centers (Figure 14). The importance of vegetation cover in moderating differential frost heave of nonsorted circles has been demonstrated by Kade and Walker (2008), who showed that removal of vegetation from circle centers resulted in a 6% increase in active-layer depth and a 26% increase in frost heave, whereas the addition of a 10-cm-thick moss layer to circle centers resulted in a 15% reduction in active-layer depth and a 52% decrease in frost heave.

Although differential frost heave appears to offer an adequate explanation for circle initiation (Peterson, 2011; Peterson and Krantz, 2003), evolution (Peterson and Krantz, 2008), and maintenance (Daanen et al., 2008; Nicolsky et al., 2008), the evidence for diapiric upward movement in some frost boils suggests that thaw-related soil displacements may also contribute to the development of nonsorted circles. Mackay

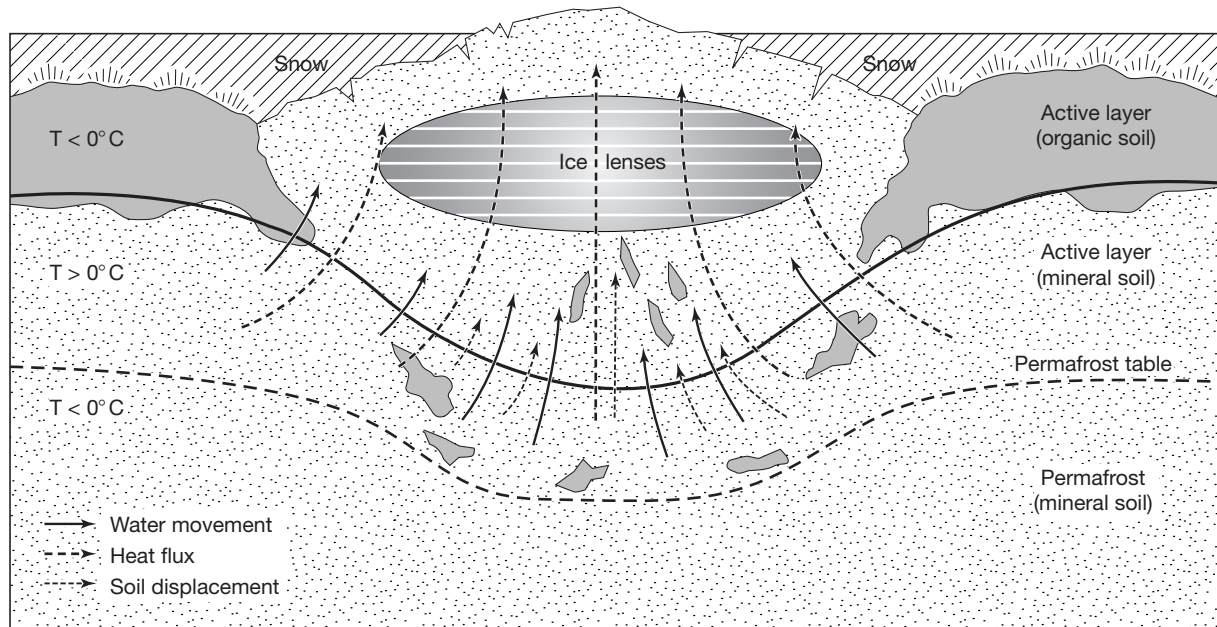


Figure 14 Physical processes taking place within a vegetation-defined nonsorted circle (frost boil) during autumn freezeback. Reproduced, with permission, from [Nicolson DJ, Romanovsky VE, Tipenko GS, and Walker DA \(2008\)](#) Modeling biophysical interactions in nonsorted circles in the low arctic. *Journal of Geophysical Research* 113: G03S05. ©2008 American Geophysical Union.

(1980) proposed that, during thaw, low-density soil with high water content rises upward through overlying soil layers in the center of frost boils, being replaced by gravity-induced soil movement down the sides of the underlying saucer-shaped depression in the permafrost table, and thus driving circular movement of soil upward at circle centers and downward at circle margins. The feasibility of soil diapirism has been established by [Swanson et al. \(1999\)](#), who showed that thawed soils near the permafrost table have a bulk density up to 800 kg m^{-3} less than that of overlying dry soil. Diapiric movements of saturated soil are most likely to occur in areas of cold permafrost where upward freezing from the permafrost table in winter results in ice-rich soil at the base of the active layer, producing excess water content in this zone during thaw. It is not known whether diapiric movements of wet soil during thaw operate independently of differential frost heave to initiate frost boils through extrusion of soil at the surface, but the emergence of frost-susceptible soil at the surface is likely to result in differential frost heave, so both processes act synergistically to expand the area of the fine domain.

Earth Hummocks

A widespread phenomenon in arctic lowlands is nonsorted patterned ground consisting of roughly circular dome-shaped earth hummocks. These features are typically 0.2–0.6 m high, 0.5–2.0 m wide, and separated by a network of narrow troughs and depressions ([Figure 15](#)). Similar features occur in permafrost-free periglacial environments such as alpine plateaus, and in maritime periglacial environments such as Iceland, where they are referred to by the collective term *thúfur* ([Figure 16](#)) and are known to form on disturbed ground within a few years or decades. Earth hummocks usually occur as closely



Figure 15 Earth hummocks developed above permafrost, Ellesmere Island, arctic Canada.

spaced assemblages on flat or gently sloping terrain of frost-susceptible soils containing few or no stones, in areas of tundra, grassland, or open subarctic woodland. Both hummocks and the intervening troughs are usually completely vegetated, apart from the localized erosion of hummock crests. On gentle slopes, earth hummocks are sometimes elongate downslope, and where gradient exceeds about 6° earth hummocks sometimes grade into nonsorted stripes (relief stripes) that form a ridge-and-furrow pattern on slopes ([Ballantyne and Harris, 1994; Nicholson, 1976; Figure 17](#)). In high-arctic permafrost environments, however, hummocky microforms referred to as *slope hummocks* occur on slopes of $6\text{--}27^\circ$ ([Figure 18](#)). Such slope hummocks resemble the earth hummocks found on more level ground, but have a distinctive origin ([Lewkowicz, 2011; see below](#)).



Figure 16 Thúfur field developed on permafrost-free terrain, northern Iceland.



Figure 17 Nonsorted 'relief stripes' at 900 m on Seana Braigh, northern Scotland. These stripes are inactive under present conditions.



Figure 18 Section excavated through a slope hummock on a slope of 8°, Ellesmere Island, Canadian Arctic Archipelago. Burial of organic soil at the downslope edge of the hummock suggests rotation of the hummock as it moved downslope over a deforming substrate during thaw of ice-rich soil (gelifluction).

Sections excavated through earth hummocks usually reveal disruption and deformation of underlying sediments, often in the form of updomed soil horizons and/or diapiric intrusions of fine-grained, sometimes organic-rich soil, though hummock

stratigraphy may be complex (Van Vliet-Lanoë et al., 1998). In permafrost areas, the permafrost table underlying hummocks takes the form of bowl-shaped depressions. These reflect slower summer thaw under interhummock troughs, which retain snow longer and often contain a thicker insulating cover of vegetation and organic-rich soil than hummock crests (Kokelj et al., 2007).

Although it is widely accepted that earth hummocks owe their development to soil displacement induced by recurrent freezing and thawing of the ground, there are several theories as to how they are initiated (Grab, 2005). A long-held view is that hummock growth in permafrost areas is generated by upward displacement of soil due to generation of cryostatic (effectively hydrostatic) pressure in confined pockets of saturated, unfrozen soil during freezeback, an effect that is likely to be self-sustaining once surface microrelief has formed (Nicholson, 1976). Mackay (1980), however, pointed out that pockets of unfrozen soil are likely to be desiccated as a result of water movement to the freezing plane to feed growing ice lenses. Mackay proposed an alternative equilibrium model for hummocks that involves radial displacement of sediment down the flanks of hummocks during thaw, inward movement of sediment toward the floors of bowl-shaped depressions in the permafrost table, and buoyancy-driven diapiric rise of mud toward the hummock crest toward the end of the thaw season. This circulatory motion is initiated by the development of bowl-shaped depressions in the permafrost table, which Mackay attributed to the formation of unvegetated areas ('mud patches') under which seasonal thawing is deeper.

Differential frost heave offers an alternative solution to the problem of hummock initiation. Lewis et al. (1993) calculated that differential frost heave may generate a characteristic longitudinal wavelength compatible with hummock dimensions, and the differential frost heave model of Peterson and Krantz (2003) may also apply to hummock initiation. Once a pattern of crest and depressions has become established through differential frost heave, preferential vegetation growth and organic soil formation in depressions may result in deeper frost penetration under crests than troughs, and consequent formation of bowl-shaped depressions in the permafrost table. Whether subsequent hummock evolution reflects buoyancy-driven soil circulation, as advocated by Mackay (1980), or gradual ice enrichment beneath an aggrading permafrost table, as proposed by Kokelj et al. (2007), is unclear. Both appear plausible, though diapiric structures in some hummocks favor the former.

Models that depend on the establishment of depressions in the permafrost table, however, cannot explain the formation of thúfur in permafrost-free areas. These may be initiated by differential frost heave, as advocated by Lewis et al. (1993), but the evolution of mature hummocks from differentially heaved crests and depressions has not been demonstrated. Van Vliet-Lanoë et al. (1998) have suggested that hummocks in nonpermafrost areas develop through differential frost penetration initiated by desiccation cracking, with subsequent upward injection of subsurface sediment along dilation cracks during thaw. At present, there is no single model that appears to account for hummock formation in both permafrost and nonpermafrost environments. Differential frost heave is probably widely responsible for hummock initiation, but

the subsequent evolution of earth hummocks in different environments may encompass different mechanisms, potentially including differential frost heave, buoyancy-driven soil diapirism, soil uplift above aggrading ice-rich permafrost, and dilation cracking and soil injection. The development of relief stripes on slopes seems to imply an overall cellular circulation of soil, on which gradient imposes a gravity-induced downslope vector of soil movement, but not all earth hummocks grade into relief stripes as slope increases.

The slope hummocks that occupy gentle to moderate slopes in high-arctic permafrost environments appear to develop in a distinctive way that is largely independent of the processes outlined earlier. Lewkowicz (2011) argues that slope hummocks evolve through accumulation of niveo-eolian sediment on hummock crests, removal of sediment between hummocks by slopewash, and slow downslope movement of hummocks by solifluction. Some slope hummocks even exhibit structures that indicate rotation of individual hummocks as they move downslope over a deforming substrate (Figure 18).

Because earth hummocks form in nonpermafrost environments, relict earth hummocks such as those on British mountains (Ballantyne and Harris, 1994) are not necessarily diagnostic of former permafrost. There is some indication, however, that the formation of *thúfur* in Iceland may be related to Holocene climatic shifts: Van Vliet-Lanoë et al. (1998) have employed tephrostratigraphy to relate periods of late Holocene hummock growth to periods of climatic deterioration.

Conclusion

Explanation of sorted and nonsorted periglacial patterned ground has traditionally been based on the concept of multiple origins. Recent numerical models complemented by field and laboratory experiments suggest that most if not all patterns apart from frost polygons are initiated by differential frost heave, either generated spontaneously from one-dimensional frost heave or by some perturbation such as ground cracking. Differing initial boundary conditions (particularly vegetation cover) probably explain the contrast between sorted and nonsorted forms. The subsequent evolution of some patterned ground forms, however, may involve additional mechanisms, of which buoyancy-induced soil movement is the most important. Understanding of the full range of processes operating and their environmental controls is a research priority if relict patterned ground forms are to be employed in Quaternary environmental reconstruction.

See also: **Paleoclimate Reconstruction: Approaches.**
Permafrost and Periglacial Features: Active Layer Processes;
Blockfields (Felsenmeer); Cryoturbation Structures; Ice Wedges and
Ice-Wedge Casts; Permafrost; Slope Deposits and Forms.

References

- Ballantyne CK (1996) Formation of miniature sorted patterns by shallow ground freezing: A field experiment. *Permafrost and Periglacial Processes* 7: 409–424.
- Ballantyne CK and Harris C (1994) *The Periglaciation of Great Britain*. Cambridge: Cambridge University Press.
- Ballantyne CK and Matthews JA (1983) Desiccation cracking and sorted polygon development, Jotunheimen, Norway. *Arctic and Alpine Research* 15: 339–349.
- Daanen RP, Misra D, Epstein HE, Walker DA, and Romanovsky V (2008) Simulating nonsorted circle development in arctic tundra ecosystems. *Journal of Geophysical Research* 113: G03S06.
- Goldthwait RP (1976) Frost-sorted patterned ground: A review. *Quaternary Research* 6: 27–35.
- Grab S (2002) Characteristics and palaeoenvironmental significance of relict sorted patterned ground, Drakensberg plateau, southern Africa. *Quaternary Science Reviews* 21: 1729–1744.
- Grab SW (2005) Aspects of the geomorphology, genesis and environmental significance of earth hummocks (*thúfur*, *pounus*): Miniature cryogenic mounds. *Progress in Physical Geography* 29: 139–155.
- Hallet B (1998) Measurements of soil motion in sorted circles, western Spitsbergen. In: Lewkowicz AG and Allard M (eds.) *Permafrost: Proceedings of the Seventh International Conference*, pp. 415–420. Quebec: Université Laval.
- Hallet B and Prestrud S (1986) Dynamics of periglacial sorted circles in western Spitsbergen. *Quaternary Research* 26: 81–99.
- Hallet B and Waddington ED (1992) Buoyancy forces induced by freeze–thaw in the active layer: Implications for diapirism and soil circulation. In: Dixon JC and Abrahams AD (eds.) *Periglacial Geomorphology*, pp. 252–279. Chichester: Wiley.
- Kade A and Walker DA (2008) Experimental alteration of vegetation on nonsorted circles: Effects of cryogenic activity and implications for climate change in the arctic. *Arctic, Antarctic, and Alpine Research* 40: 96–103.
- Kessler MA, Murray AB, Werner BT, and Hallet B (2001) A model for sorted circles as self-organized patterns. *Journal of Geophysical Research* 106(B7): 13287–13306.
- Kessler MA and Werner BT (2003) Self organization of sorted patterned ground. *Science* 299: 380–383.
- Kokelj SV, Burn CR, and Tarnocai C (2007) The structure and dynamics of earth hummocks in the subarctic forest near Inuvik, Northwest Territories, Canada. *Arctic, Antarctic, and Alpine Research* 39: 99–109.
- Krantz WB (1990) Self-organization manifest as patterned ground in recurrently frozen soils. *Earth-Science Reviews* 29: 117–130.
- Lewis GC, Krantz WB, and Caine N (1993) A model for the initiation of patterned ground owing to differential frost heave. *Permafrost: Proceedings of the Sixth International Conference*, pp. 1044–1049. Wushan Guangzhou, China: South China University of Technology Press.
- Lewkowicz AG (2011) Slope hummock development, Fosheim Peninsula, Ellesmere Island, Nunavut, Canada. *Quaternary Research* 75: 334–346.
- Mackay JR (1980) The origin of hummocks, western arctic coast, Canada. *Canadian Journal of Earth Sciences* 17: 996–1006.
- Matsuoka N, Abe M, and Ijiri M (2003) Differential frost heave and sorted patterned ground: Field measurements and a laboratory experiment. *Geomorphology* 52: 73–85.
- Murton JB and Kolstrup E (2003) Ice wedge casts as indicators of palaeotemperatures: Precise proxy or wishful thinking? *Progress in Physical Geography* 27: 155–170.
- Nicholson FH (1976) Patterned ground formation and description as suggested by low arctic and subarctic examples. *Arctic and Alpine Research* 8: 329–342.
- Nicolisky DJ, Romanovsky VE, Tipenko GS, and Walker DA (2008) Modeling biophysical interactions in nonsorted circles in the low arctic. *Journal of Geophysical Research* 113: G03S05.
- Peterson RA (2011) Assessing the role of differential frost heave in the origin of non-sorted circles. *Quaternary Research* 75: 325–333.
- Peterson RA and Krantz WB (2003) A mechanism for differential frost heave and its implications for patterned ground formation. *Journal of Glaciology* 49: 69–80.
- Peterson RA and Krantz WB (2008) Differential frost heave model for patterned ground formation: Corroboration with observations along a North American arctic transect. *Journal of Geophysical Research* 113: G03S04.
- Swanson DK, Ping C-L, and Michaelson GJ (1999) Diapirism of soils due to thaw of ice-rich material in the permafrost table. *Permafrost and Periglacial Processes* 10: 349–367.
- Van Vliet-Lanoë B (1988) The significance of cryoturbation phenomena in environmental reconstruction. *Journal of Quaternary Science* 3: 85–96.
- Van Vliet-Lanoë B, Bourgeois O, and Dauterive O (1998) *Thufur* formation in northern Iceland and its relation to Holocene climate change. *Permafrost and Periglacial Processes* 9: 347–365.

Walker DA, Epstein HE, Gould WA, et al. (2004) Frost-boil ecosystems: Complex interactions between landforms, soils, vegetation and climate. *Permafrost and Periglacial Processes* 15: 171–188.

Washburn AL (1956) Classification of patterned ground and review of suggested origins. *Bulletin of the Geological Society of America* 67: 823–865.

Washburn AL (1989) Near-surface soil displacement in sorted circles, Resolute area, Cornwallis Island, Canadian High Arctic. *Canadian Journal of Earth Sciences* 26: 941–955.

Werner BT and Hallet B (1993) Numerical simulation of self-organized stone stripes. *Nature* 361: 142–145.

Permafrost

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Introduction

Permafrost is the ground that remains at or below 0 °C for 2 years or more. Permafrost forms a layer, centimeters to kilometers thick, sandwiched between a seasonally thawed active layer at the surface of the ground, and unfrozen ground at depth. The thickest permafrost formed during the Pleistocene in unglaciated Siberia, and is over 1500 m deep (French, 2007). Permafrost in Alaska and western Arctic Canada is up to about 600 m thick, again due to permafrost aggradation during the Pleistocene under subaerial conditions. The persistence of permafrost increases with its thickness so that deep permafrost may not only take many millennia to form, but its degradation would occur over similar timescales.

Permafrost is a product of climate. At continental scale, the two most important controlling variables are air temperature and winter snowfall. As a result, permafrost is found at relatively high latitudes and at high altitudes (Brown et al., 1997). In western North America, permafrost is found at more southerly latitudes than in western Eurasia because the Cordillera blocks weather systems originating in the Pacific, creating a continental climate inland. This climate not only provides low winter temperatures, but also a rain shadow, leading to relatively low snowfall, and hence relatively little insulation for the ground from the frigid conditions. Permafrost is widespread in Alaska and Canada's Yukon Territory, but there is little permafrost in northwest Russia. The extent of permafrost in the Yukon increased during the Quaternary interglacials, as the St. Elias Mountains were uplifted, and the climatic effectiveness of this coastal barrier increased (Burn, 1994).

The top of permafrost is characteristically ice rich, so that if the active layer deepens and permafrost thaws, the surface of the ground may become unstable. Subsidence is common in flat terrain after disturbance, but may not be spatially uniform, and water may accumulate in the resulting depressions. These ponds perpetuate the thermal disturbance, and grow into water bodies known as thermokarst, or thaw, lakes. Lakes are common in permafrost terrain because the frozen ground around and beneath them inhibits seepage. On hillslopes, thawing of the ice-rich ground at the top of permafrost may lead to landslides, due to reduced friction on the surface of melting ice lenses, and loss of shear strength following an increase in pore-water pressure (French, 2007). The principal natural disturbance currently encountered is forest fire, but most infrastructure construction also leads to active-layer deepening, and therefore requires careful management and application of special techniques to prevent ground instability.

Thermal Regime

Figure 1 illustrates the thermal regime of permafrost. The two principal features are an envelope of ground temperature near the surface and an increase in ground temperature with depth.

The mean annual temperature at the top of permafrost, T_p , is one of three variables that governs the thickness of permafrost, z , which also depends on the geothermal flux, G , and the thermal conductivity of the frozen ground, λ (Williams and Smith, 1989). G is related to z by:

$$G = \lambda T_p / z \quad [1]$$

Of these variables, the geothermal flux is spatially the most constant, ranging in North America between 0.05 and 0.2 W m⁻², through the stable craton of the Canadian Shield to the tectonically active terrain of the Cordillera. The thermal properties vary with water and ice content. Air, water, and ice have λ of 0.022, 0.56, and 2.2 W m⁻¹ °C⁻¹, respectively, and specific heat capacities, c , of 0.86, 4200, and 1900 kJ m⁻³ °C⁻¹, respectively. Mineral materials have λ of about 2.0 W m⁻¹ °C⁻¹ (except for quartz, 8.0 W m⁻¹ °C⁻¹), and c of about 2200 kJ m⁻³ °C⁻¹ (Burn, 2004).

The two sides of the ground temperature envelope represent the annual minimum and maximum temperatures in the ground. These values do not occur simultaneously with depth, but are reached progressively later from the surface downwards. In practical terms, the depth at which the difference between annual minimum and maximum temperatures is less than 0.1 °C represents the depth of zero annual amplitude, z_0 , and the temperature at this depth, T_g , is the mean annual ground temperature. In most locations, z_0 is at a depth of 15–25 m. The variation in z_0 depends primarily on the thermal diffusivity, λ/c . In bedrock and sand, λ/c is relatively high, and z_0 relatively large, but in peat and where the active layer is wet, the temperature envelope attenuates near the surface, and z_0 is relatively small. In thin permafrost, z_0 may be at the base of the perennially frozen ground.

T_p depends on the mean annual ground surface temperature, T_s , and the seasonal change in λ within the active layer due to freezing and thawing. The difference between T_s and T_p is the thermal offset (Romanovsky and Osterkamp, 1995). The thermal offset increases with water (ice) content in the active layer, and is therefore greatest in saturated peat, which has a high porosity.

Aggradation and Degradation

Freezing and thawing of the ground occur in response to the seasonal variation in climate at the ground surface. Permafrost forms where seasonal frost penetration exceeds thawing during the following summer. The thermal offset is a product of the asymmetry in seasonal values of λ , and is critical in permafrost development at the limit of perennially frozen ground. The principal variables governing the development of permafrost are T_p , λ , L , the latent heat of fusion of unfrozen ground, and t , the time since initiation of freezing. The depth of aggrading permafrost, z , is related to these variables by (Williams and Smith, 1989):

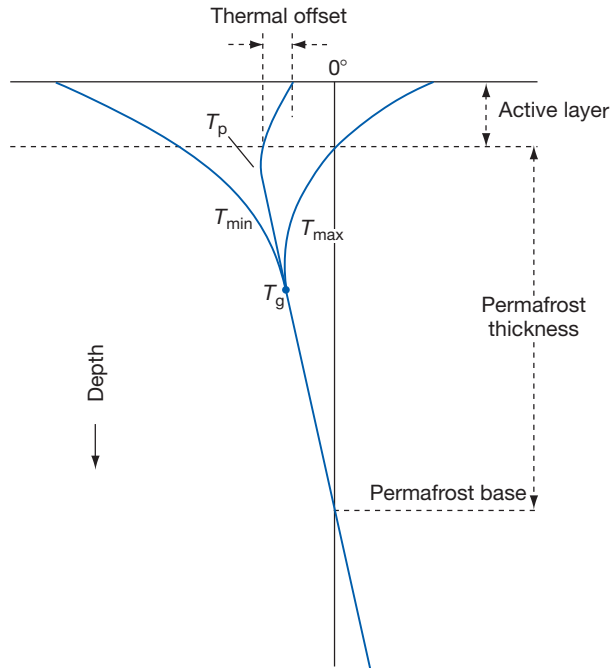


Figure 1 Thermal regime of permafrost.

$$z = \sqrt{(2\lambda T_p t / L)} \quad [2]$$

where λ is the value for frozen ground. Equation [2] is the Stefan solution, a model that assumes freezing follows a step change in surface temperature, and that the ground is uniformly at 0 °C before freezing begins. The former assumption is a valid approximation of conditions following drainage of lakes in permafrost regions, where permafrost aggrades into the unfrozen lake bottom. The latter is also valid, because c is two orders of magnitude less than L , so the heat budget is dominated by freezing of water, not cooling of the ground. Equation [2] is useful for estimating the time required to form deep permafrost. For example, permafrost thickness varies in association with the glacial limits near the western Arctic coast of Canada over distances of less than 50 km. At unglaciated locations, the thickness is over 500 m, but at glaciated sites the thickness is 300 m or less. Significant permafrost did not develop beneath glacial ice in this area, and these depths represent the combination of differences in time and surface temperature history for the area.

The Stefan solution is also valid for determination of the rate of permafrost thaw following surface disturbance, but in this case λ is for thawed ground, and T_p is the new ground temperature above 0 °C. An important aspect of this model is the relation between depth of thaw and square root of time. This implies that even relatively thin ice-rich permafrost, on the order of 10 m, requires centuries to degrade following the perturbations to ground temperature accompanying natural disturbances (Burn, 1998). There are no landscape-scale data to test the model for permafrost degradation over Quaternary time, but as for aggradation, the model has been verified for natural disturbances such as forest fires, thaw slumps, and thaw lakes. In the case of developing thaw lakes, horizontal expansion may be equivalent to thaw deepening, where t is 1 year.

Upward thawing of permafrost from its base is a very slow process, because of the low geothermal flux and high L . The rates of thaw are 0.5–2.0 cm year⁻¹, except where additional heat is supplied by groundwater convection.

Ground Surface Temperature

T_s is a function of the surface energy balance, so while it is broadly governed by the climate, at local scale it is a function of aspect, shade, ground surface wetness, and the winter snow cover (Williams and Smith, 1989). Shade and soil moisture reduce the ground temperature in summer by lowering incident radiation and promoting evaporation, respectively. Aspect is important in mountainous areas, for in the Northern Hemisphere, south-facing slopes receive critically more radiation in spring and autumn, keeping these surfaces above 0 °C for longer than slopes with other aspects. Snow cover provides insulation at the ground surface in winter, trapping the ground heat flux and raising surface temperatures. In combination, these factors are responsible for a variation in T_s of about 5 °C in any locality. The influence of the snow cover is responsible for raising T_s above the mean annual air temperature, in general by about 3 °C, but in places by as much as 10 °C (Smith et al., 1998). As a result of such variation, permafrost is spatially discontinuous through a considerable extent of its distribution, for in places, T_g may locally vary above and below 0 °C.

The importance of snow cover is well demonstrated by data collected from peatlands in Canada's western Arctic at Inuvik, south of treeline, and Parson's Lake, in the tundra, 60 km north of Inuvik, during 1993–94. T_s at Inuvik was approximately –1.7 °C, while at Parson's Lake, T_s was –5.5 °C (Smith et al., 1998). The snow cover at Inuvik is commonly 50 cm deep by the end of March, with a density of about 0.2 g cm⁻³, while at Parson's Lake there is about half the snow depth, and the snow pack has twice the density, raising its λ .

The relation between air and ground surface temperature at a site may be obtained through detailed flux measurement and determination of the surface energy balance. Such an approach is inappropriate for many practical purposes and for predicting permafrost conditions over large areas. Instead, the relation may be summarized through the n -factor, or ratio of air-to-ground surface temperature. The n -factors for a site are determined for the freezing, n_f , and thawing, n_t , seasons. These indices vary with surface and subsurface characteristics. In winter, n_f depends on snow depth, but also the amount of unfrozen ground in the soil profile. Where the active layer is deep or wet, release of latent heat during soil freezing maintains a high T_s , and reduces n_f (Karunaratne and Burn, 2004). In summer, n_t depends on the rate of thawing, because energy initially entering the soil is used to melt ice rather than warm the ground. In the latter half of summer, air and ground surface temperatures are almost equal. For sites in the boreal forest, values between 0.1 and 0.5 have been measured for n_f , and between 0.6 and 1.1 for n_t . The values may be substantially higher for artificial surfaces.

Forest fires cause the most widespread natural disturbances to surface conditions in permafrost regions. These fires are not often extinguished rapidly because of their remote locations. Permafrost is rarely affected by the fire itself, but degrades

following the change in surface conditions (Burn, 1998). In particular, ground temperatures increase because of loss of the soil's organic layer and the shade of trees, drying of the soil surface, and the increase in snow accumulation following destruction of the canopy. As a result, landslides are common in sloping terrain following fires.

Distribution

At the continental scale, latitudinal permafrost is mapped according to its extent beneath the land surface (Figure 2). Continuous permafrost is defined as regions where 90% or more of the land surface is underlain by perennially frozen ground. Discontinuous permafrost is divided into categories of widespread, sporadic, and isolated patches of permafrost, where, respectively, 90–50%, 50–10% and less than 10% of the surface is underlain by permafrost. In addition, subsea permafrost occurs in the continental shelves of the Arctic Ocean at locations that were dry land during glacial periods, when sea level was lower (Brown et al., 1997; for instance, the Bering Land Bridge). Alpine permafrost is classified at high elevations in areas where latitudinal permafrost is not prevalent, such as

the extensive permafrost of the Tibetan plateau. Lake and river bottoms are not included in the measure, because permafrost is normally absent immediately beneath water bodies that do not freeze through. Unfrozen ground surrounded by permafrost, such as beneath lakes and rivers, is known as talik.

The distribution of permafrost is fundamentally a function of climate, but is modified by surface characteristics. Near its limit, permafrost is characteristically present in peatlands, because the thermal offset induces mean ground temperatures below 0 °C, even though the surface conditions may be several degrees Celsius higher (Smith and Riseborough, 2002). The continental boundary between discontinuous and continuous permafrost follows treeline, as a result of forest and shrub vegetation enhancing the snow pack by reducing ground-level wind speed and, hence, drifting. In the Northern Hemisphere, the southern limit of continuous permafrost has been considered to correspond with a mean annual air temperature of –8 °C, or T_g of –5 °C. Permafrost in this region is continuous in both space and time, because a considerable climate change is required to degrade it. Even submersion by rising sea level has not yet led to the eradication of the continuous permafrost established during the Pleistocene in the coastal shelves of the Arctic Ocean.

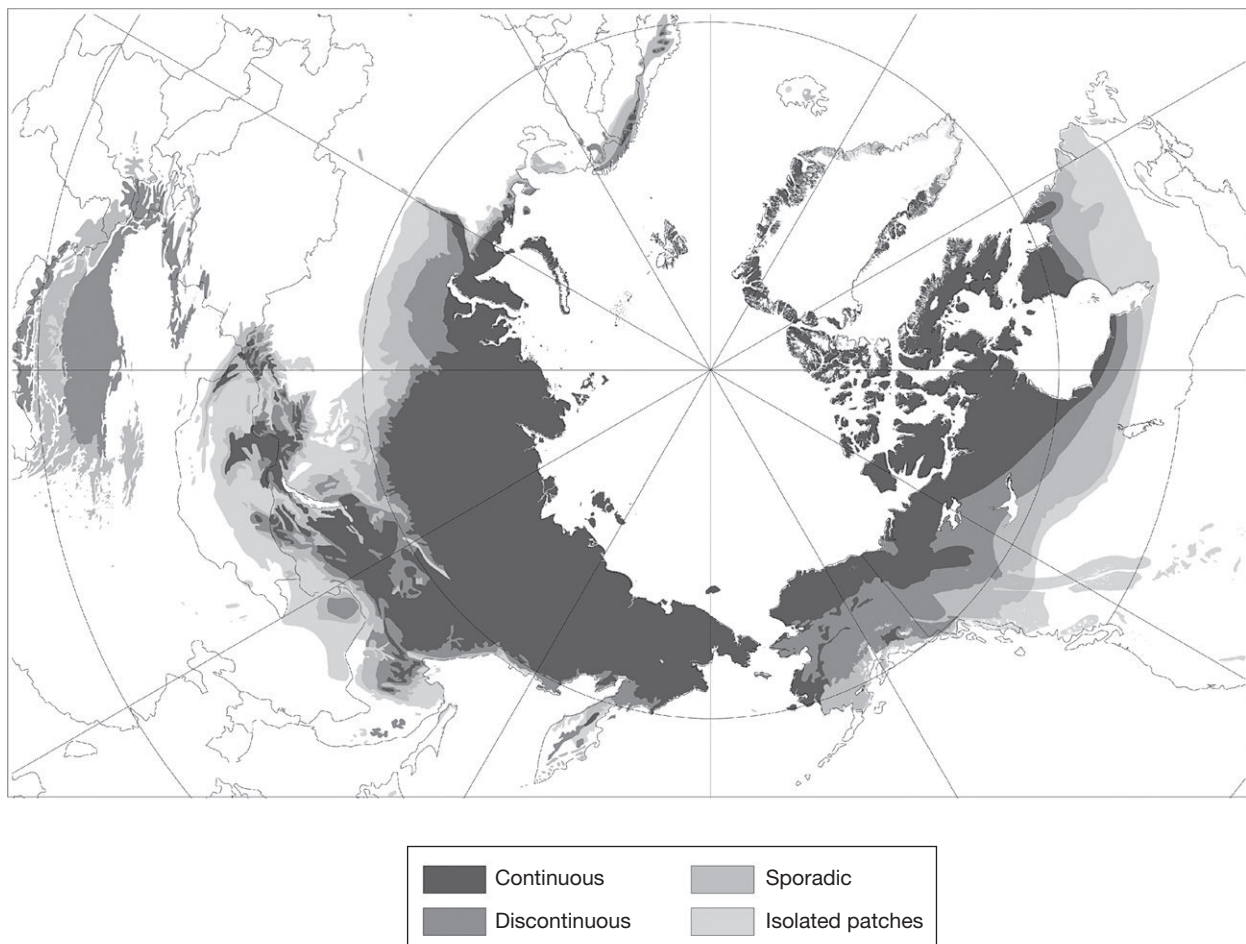


Figure 2 Distribution of permafrost in the Northern Hemisphere. Reproduced from Brown J, Ferrians OJ, Heginbottom JA, and Melnikov ES (1997) International Permafrost Association circumpolar map of permafrost and ground ice conditions. *United States Geological Survey Circum-Pacific Map Series CP-45 1*: 100000000. www.nsidc.org.

Ground Ice

The geotechnical significance of permafrost is derived from the presence of excess ice, that is ice content greater than the saturation water content. Excess ice occurs in the ground following burial of surface ice, such as river, lake or glacial ice, or it may form in the ground itself (French, 2007). There has been considerable discussion surrounding the origin of extensive tabular beds of massive ice, which are bodies of almost pure ice found near the western Arctic coast of Canada (Figure 3) and near the Siberian coast. Within the glacial limits, some glacier ice is preserved (Murton, 2009), but other massive ice formed within sediments. There are two principal processes responsible for the growth of ice in sediments, and these depend on the particle-size composition of the materials.

In coarse-grained sediment, such as sand, freezing of saturated ground is accompanied by expulsion of water, as the specific volume of ice is greater than that of water, and so the porosity is exceeded. In confined systems, the expelled water raises local water pressure, and so there is potential for water to be injected into cracks in the frozen ground, or to lift the ground and form a pressurized, subterranean pool (Mackay, 1998). The water may then freeze in place to form massive ice. This is the mechanism responsible for the development of pingos in drained lake basins, and for some frost blisters, or ice-cored mounds, within the active layer.

In silts and clays, however, films of adsorbed water remain unfrozen at the surface of soil particles, and during freezing these conduct water along soil moisture potential gradients into freezing soil (Williams and Smith, 1989). The potential gradient is proportional to the temperature gradient, but a suction of $1.2 \text{ MPa } ^\circ\text{C}^{-1}$ develops, and over time moves water through the soil even though the hydraulic conductivity is relatively low. As the water moves into freezing ground the volume rises above saturation, and soil particles are pushed apart to form layers of pure ice (Figure 4). The process is called ice segregation. Lenses of segregated ice are common in most freezing soils, and are responsible for the uplift, or frost heave, of the ground. Large massive-ice bodies in permafrost have been formed by a combination of freezing in coarse sediments, which provided a ready supply of water under pressure, and

overlying silty clay, in which segregation led to development of large ice lenses (Figure 3; Mackay and Dallimore, 1992).

Segregated ice is almost ubiquitous at the top of permafrost in unconsolidated sediments and some bedrock. This ice-rich zone is commonly 0.5–1.0 m thick, and contains ice lenses that have formed with water drawn downward into permafrost from the base of the active layer at the end of summer and in early autumn. This ice-rich layer and the occurrence of ice wedges at the base of the active layer combine to make permafrost terrain sensitive to disturbance.

Ice Wedges

Ice wedges are one of the few geomorphic features of diagnostic of permafrost (Mackay, 2000). These wedge-shaped masses of relatively pure ice form at the base of the active layer following repeated cracking of the ground due to thermal contraction in winter (Figure 5). The thermal contraction cracks are infiltrated by snowmelt, which freezes to form relatively weak vertical planes of ice in the near surface that become loci for cracking in subsequent winters. In dry environments, or where sand is abundant, the cracks may be infilled with eolian deposits to form sand wedges or composite sand and ice wedges (French, 2007). After repetition of the cracking and infilling over many winters, the vertical veins of ice grade into wedge-shaped structures because the cracks vary annually in depth. The frequency of crack penetration, and of the depth of infiltration, declines below the active layer. The addition of volume to the permafrost through the growth of wedges leads to deformation of adjacent stratigraphy, so that bedding is upturned next to wedges, and characteristically two ridges form at the surface on either side of the wedge. The wedge underlies a central trough, formed as the ground separates above the growing wedge. The troughs intersect to create a network of irregular, four- to seven-sided polygons, which are common in surficial sediments throughout continuous permafrost (Figure 6). If permafrost degrades, the primary structures of sand wedges may be preserved in stratigraphy, but ice wedges melt and may be replaced by casts formed from material that infills the void.



Figure 3 Massive tabular ice exposed at Peninsula Point, near Tuktoyaktuk, Northwest Territories, Canada. Photograph taken in June 2001 by C.R. Burn.



Figure 4 Lenses of segregated ice in glaciolacustrine sediments near Mayo, Yukon Territory, Canada. Photograph taken in January 1986 by C.R. Burn.



Figure 5 Ice wedge exposed near Tuktoyaktuk, Northwest Territories, Canada. Photograph taken in April 1987 by C.R. Burn.

Three types of ice wedge are recognized from different sedimentary environments (Mackay, 1990). Epigenetic wedges form under a stable surface, syngenetic wedges in surfaces of deposition, and antisynthetic, or hillslope ice wedges, where the surface is eroding. Epigenetic wedges are V-shaped, with flat tops that have been truncated at the base of the active layer. Since the thermal contraction crack penetrates downwards into the ground from near the top of permafrost, syngenetic wedges are characteristically long and thin, for the apex of the crack migrates upwards with the point of crack initiation, as the surface is raised.

The size of an ice wedge is not a reliable guide to its age, as the rate of growth of wedges varies with time. Initially, growth is rapid, but as the ridges on each side of the wedge grow and trap snow, the drifts reduce the thermal stress. Networks of thermal contraction cracks have been presumed to become denser as the climate becomes colder and conditions for cracking improve. However, field observations indicate that wedges cease to crack frequently as they age, so secondary wedges may develop to divide primary polygons and relieve thermal stress under constant conditions. In field stratigraphy, the size of an ice wedge also depends on its angle of exposure, for few wedges are exposed in true cross section.

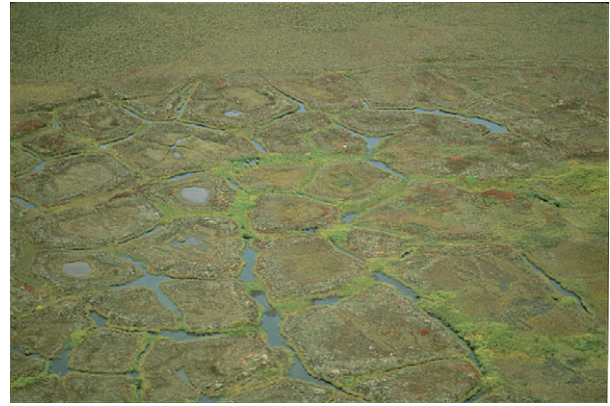


Figure 6 Ice-wedge polygons at Garry Island, Northwest Territories, Canada. Photograph taken in August 2003 by C.R. Burn.



Figure 7 Thermokarst lake near Mayo, Yukon Territory, Canada, showing trees that have become submerged as the lake has expanded by thawing of permafrost. Photograph taken in July 1988 by C.R. Burn.

Lakes

Lakes alter ground temperature substantially from patterns determined by climate (Burn, 2002). Ground beneath lakes is warmer than the surrounding permafrost because lake-bottom temperatures are at or above 0 °C throughout the year, unless the water body freezes through. In bedrock areas, many lakes occupy depressions formed by glacial erosion, but in sediments the lakes are commonly thermokarst features (Figure 7). Once a lake is established, its development continues until the ground ice surrounding the lake is exhausted, the water balance cannot sustain the lake, or it drains by thermal erosion of surrounding permafrost. Down-cutting and enlargement of a small channel through ice-rich ground is well-established as the primary mechanism of rapid lake drainage in northwest North America (Mackay, 1997).

In the boreal forest, some thaw lakes are sheltered from the wind and mixing of the water body in summer is restricted to the upper 2 m or so. In contrast, most tundra lakes, and larger lakes in the boreal forest, are well mixed in summer, and temperatures are uniform to depths of about 30 m. Many tundra lakes have well-developed littoral terraces, beneath water depths of less than 1 m, in which permafrost is

preserved, and a central pool of deeper water where the mean annual lake-bottom temperature is between 2 and 4 °C (Figure 8). Beneath the deep pool, a talik (unfrozen ground) develops whose dimensions depend on the width of the lake as well as the temperature at the lake bottom and in the surrounding permafrost. The width of the lake is the most locally variable factor, and so controls whether the talik penetrates the full thickness of permafrost or is closed beneath the lake.

The flat tundra regions of western Arctic Canada, Alaska, and Russia contain extensive areas where lakes of all sizes are oriented in a common direction, perpendicular to the prevailing winds. Several theories have been proposed to account for the consistency of orientation, but the most convincing is that in summer, when the lakes have no ice cover, the strongest winds establish a two-cell current circulation. This leads to preferential erosion at the ends of the lakes by water driven across the lake and returning around the lake shore. The mechanism has been replicated in the laboratory, and is consistent with observations from western Arctic Canada, where, within a few hundred kilometers, lake orientations vary by 90° in association with changes in wind direction.

Hillslopes

In addition to catastrophic movements on slopes induced by thawing of near-surface permafrost, slower downslope movement occurs due to ice in the ground. At the surface, the active layer commonly moves downslope relative to permafrost at the end of summer as the depth of thaw approaches the ice-rich zone at the base of the active layer. The movement, called solifluction, may be at a rate of up to a few centimeters per year. The moving soil forms lobes that spill onto flat ground in the valley floor when they reach the bottom of the slope. The heights of the solifluction lobes are about 60 cm at the foot, and they grade back into the hillslope. The front of each lobe rolls gradually over onto itself, so that vegetation is buried, and this layer is progressively older upslope (French, 2007).

Ice-rich permafrost itself deforms in response to gravity, because at ambient temperatures the ice is close to its melting point and so tends to creep. Downslope deformation of

permafrost has been measured both in massive ice and ice-rich clay in northwest Canada (Dallimore et al., 1996), and the development of rock glaciers in boulder fields with a matrix of pore ice, is also due to creep of the interstitial ice. In Alpine regions, rock glaciers are a common product of permafrost creep, but built structures may also cause such deformation because of the additional load they apply to hillslope materials.

Other Landforms

Two other sets of landforms are characteristic of permafrost terrain. If ice-rich permafrost is exposed by erosion in river banks, lake shores, or at the coast, then ablation of ground ice may initiate development of retrogressive thaw slumps. These features consist of a steep headwall of thawing permafrost, connected to the locus of erosion by a footslope of low gradient (Figure 9).

Thaw slumps are active during summer, when air temperatures are above 0 °C, and are among the most rapidly developing geomorphic features, with measured rates of headwall retreat up to 16 m year⁻¹.

Perhaps the most well-known permafrost landforms are pingos, conical hills that contain a core of ice (Mackay, 1998). The ice accumulates from water pushed into ground beneath the pingo by pressure from a hydraulic head developed down a hillslope, or due to the growth of permafrost in saturated sand. Pingos formed at the base of hillslopes are best documented from Greenland, Alaska, Yukon Territory, and Svalbard, while pingos forming from aggradation of permafrost are best known from Canada's western Arctic coastlands (Figure 10).

Permafrost and Climate Change

Since permafrost is both a function of climate and a structural component of cold-regions' terrain, its response to climate change anticipated during the twenty-first century is significant for infrastructure planning and management of natural resources. The time involved in thawing permafrost completely is great, so over the next hundred years it is unlikely there will

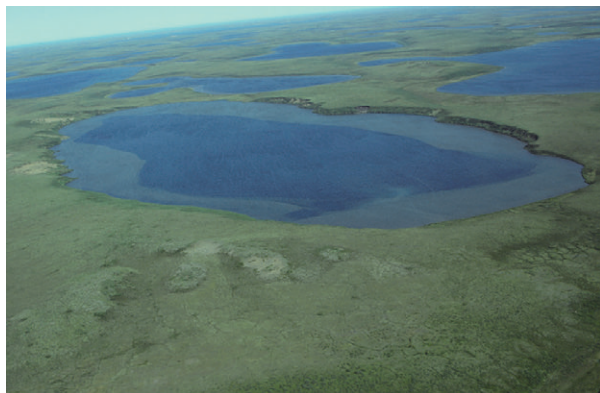


Figure 8 Thaw lake on Richards Island, Mackenzie delta area, western Arctic coast, Canada, showing littoral terraces surrounding a central pool. Photograph taken in August 1991 by C.R. Burn.



Figure 9 Retrogressive thaw slump near Mayo, Yukon Territory, Canada. Photograph taken in August 1984 by C.R. Burn.

be widespread eradication of perennially frozen ground, except where the current thickness is only a few meters. At present, warming of polar regions is expected to be significant, but as important for permafrost is the projected increase in snow fall. Two principal effects of regional climatic warming are an increase in the temperature of permafrost, and deepening of the active layer, with associated consequences for ground stability.

The thermal response of permafrost to climate warming involves an inflection of the ground temperature profile, so that the coldest part of the ground is located progressively deeper in the profile. Curved ground temperature profiles indicating warming of permafrost since the 1850s have been obtained in northern Alaska, and indicate the response to climate warming since the Little Ice Age (Lachenbruch and Marshall, 1986). During the 2007–2009 International Polar Year, an integrated circumpolar program to determine the current temperature in permafrost reported warmer conditions than in prior assessments at almost every location (Romanovsky et al., 2010). Many of the reports contained near-surface observations only, and so the time scale of permafrost disturbance remains elusive. However, where deeper measurements have become available, the time scale of permafrost disturbance is interpreted to be similar to that in Alaska (e.g., Burn and Zhang, 2009). There has been some monitoring of near-surface permafrost in northwest North America during the last 20 years which indicates ground warming in general, but cooling at locations where snow depths have been low (Osterkamp and Romanovsky, 1999). In terms of ground temperatures, the response is most easily detectable in relatively cold permafrost, because the additional energy is used to warm the ground, while in warm permafrost much of the additional energy is consumed by the latent heat required to melt ice. As a result, the thermal response of discontinuous permafrost to climate warming is slower and more difficult to detect than the response of continuous permafrost, even though considerable thawing may be occurring.

A thicker active layer is anticipated under climate warming, because of warmer summer conditions and deeper snow cover in winter. Evidence of a regionally deeper active layer in northwest Canada is preserved in cryostratigraphy by a thaw unconformity about 1.5 m below the base of the present active layer (Burn, 1997). The deep seasonal thaw has been associated with the Holocene thermal optimum, which occurred there

9000 years ago, and a more northerly treeline because the coast was 100 km further north at the end of glaciation. The unconformity has been recognized from truncated ice wedges and a change in the form and concentration of ground ice (Figures 4 and 11). The base of the active layer has aggraded upwards from this unconformity, and considerable ice has been incorporated in the uppermost meter of permafrost. Thaw of this ground would liberate this ice, and would also make organic matter currently contained in permafrost available for decomposition by microbial processes in the active layer. It is not yet clear whether a more biologically productive boreal landscape would act as a net carbon sink, from increased vegetation growth, or a net source from decomposition of organic matter currently unavailable in permafrost (Mack et al., 2004). There is considerable research in progress on this topic, because preliminary results suggest a massive amount of carbon is stored in the upper 3 m of permafrost, on the order of 10^{15} kg, or over 100 times the world's annual industrial emissions (Kuhry et al., 2009; Tarnocai et al., 2009).

An area of particular interest with respect to carbon release from permafrost terrain concerns methane production from thermokarst lakes. Graphic demonstrations of methane bubbles caught by lake ice igniting when punctured have illustrated the contribution of this process to the carbon budget of the atmosphere (e.g., Walter et al., 2006). As a result, there is interest in whether permafrost degradation beneath thermokarst lakes will increase as climate warming accelerates. The key issues are whether the rate of lake initiation will accelerate, for which there is some evidence (Jorgenson et al., 2006), and whether the rate of lake expansion will change. There are few studies yet of the latter, but Burn and Smith (1990), in finding lake expansion was governed by the Stefan solution, noted the consistent rate of increase in lake radius, and hence an accelerating change in lake area. A complicating factor is that lake expansion may lead to lake drainage and cessation of permafrost degradation (Mackay, 1992). In summary, the evidence, particularly from Alaska, of increasing numbers of thaw lakes, supports the concern that degrading permafrost will emit increasing amounts of methane into the atmosphere.



Figure 10 Ibyuk Pingo near Tuktoyaktuk, Northwest Territories, Canada. Photograph taken in August 1990 by C.R. Burn.

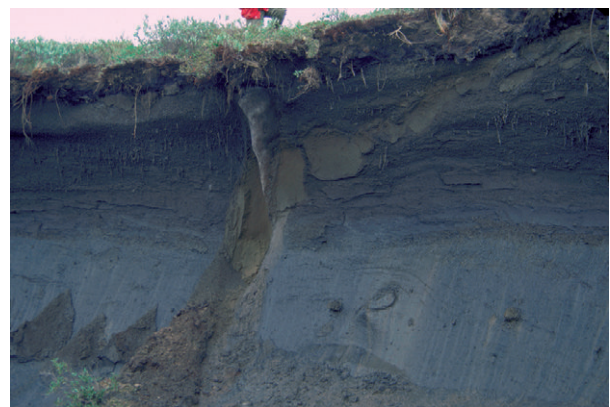


Figure 11 The early Holocene thaw unconformity at a site on Richards Island, western Arctic coast, Canada. The top of the ice wedge marks the base of the active layer. Rhizomes above the unconformity indicate the ground was part of a paleoactive layer. Photograph taken in August 1990 by C.R. Burn.

See also: **Ice Core Methods:** Borehole Temperature Records. **Permafrost and Periglacial Features:** Active Layer Processes; Frost Mounds: Active and Relict Forms; Ice Wedges and Ice-Wedge Casts; Rock Glaciers and Protalus Forms; Slope Deposits and Forms; Thermokarst Topography.

References

- Brown J, Ferrians OJ, Heginbottom JA, and Melnikov ES (1997) International Permafrost Association circum-Arctic map of permafrost and ground ice conditions. *United States Geological Survey Circum-Pacific Map Series CP-45* 1: 10000000. www.nsidc.org.
- Burn CR (1994) Permafrost, tectonics and past and future regional climate change, Yukon and adjacent Northwest Territories. *Canadian Journal of Earth Sciences* 31: 182–191.
- Burn CR (1997) Cryostratigraphy, paleogeography, and climate change during the early Holocene warm interval, western Arctic coast Canada. *Canadian Journal of Earth Sciences* 34: 912–925.
- Burn CR (1998) The response of permafrost and near-surface ground temperatures to forest fire, Takhini River valley, southern Yukon Territory. *Canadian Journal of Earth Sciences* 35: 184–199.
- Burn CR (2002) Tundra lakes and permafrost, Richards Island, western Arctic coast, Canada. *Canadian Journal of Earth Sciences* 39: 1281–1298.
- Burn CR (2004) The thermal regime of cryosols. In: Kimble JM (ed.) *Cryosols Permafrost-Affected Soils*, p. 391. Berlin: Springer.
- Burn CR and Smith MW (1990) Development of thermokarst lakes during the Holocene at sites near Mayo, Yukon Territory. *Permafrost and Periglacial Processes* 1(2): 161–176.
- Burn CR and Zhang Y (2009) Permafrost and climate change at Herschel Island (Qikiqtaruk), Yukon Territory, Canada. *Journal of Geophysical Research (Earth Surface)* 114: F02001. <http://dx.doi.org/10.1029/2008JF001087>.
- Dallimore SR, Nixon FM, Egginton PA, and Bisson JG (1996) Deep-seated creep of massive ground ice, Tuktoyaktuk, NWT, Canada. *Permafrost and Periglacial Processes* 7: 337–347.
- French HM (2007) *The Periglacial Environment*, 3rd edn. Chichester: John Wiley.
- Jorgenson MT, Shur YL, and Pullman ER (2006) Abrupt increase in permafrost degradation in Arctic Alaska. *Geophysical Research Letters* 33: L02503. <http://dx.doi.org/10.1029/2005GL024960>.
- Karunaratne KC and Burn CR (2004) Relations between air and surface temperature in discontinuous permafrost terrain near Mayo, Yukon Territory. *Canadian Journal of Earth Sciences* 41: 1437–1451.
- Kuhry P, Ping C-L, Schurr EAG, Tarnocai C, and Zimov S (2009) Report from the International Permafrost Association: Carbon pools in permafrost regions. *Permafrost and Periglacial Processes* 20: 229–234.
- Lachenbruch AH and Marshall BV (1986) Changing climate: Geothermal evidence from permafrost in the Alaskan arctic. *Science* 234: 689–696.
- Mack MC, Schuur EAG, Bret-Harte MS, Shaver GR, and Chapin FS III (2004) Ecosystem carbon storage in arctic tundra reduced by long-term nutrient fertilization. *Nature* 431: 440–443.
- Mackay JR (1990) Some observations on the growth and deformation of epigenetic, syngenetic, and anti-syngenetic ice wedges. *Permafrost and Periglacial Processes* 1: 15–29.
- Mackay JR (1992) Lake stability in an ice-rich permafrost environment: Examples from the western arctic coast. In: Roberts RD and Bothwell ML (eds.) *Aquatic Ecosystems in Semi-arid Regions: Implications for Resource Management National Hydrology Research Institute Symposium Series* 7, pp. 1–26. Saskatoon, SK: Environment Canada.
- Mackay JR (1997) A full-scale field experiment (1978–1995) on the growth of permafrost by means of lake drainage, western Arctic coast: A discussion of the method and some results. *Canadian Journal of Earth Sciences* 34: 17–33.
- Mackay JR (1998) Pingo growth and collapse, Tuktoyaktuk Peninsula area, western Arctic coast, Canada: A long-term study. *Géographie physique et Quaternaire* 52: 271–323.
- Mackay JR (2000) Thermally induced movements in ice-wedge polygons, western Arctic coast: A long-term study. *Géographie physique et Quaternaire* 54: 41–68.
- Mackay JR and Dallimore SR (1992) Massive ice of the Tuktoyaktuk area, western Arctic coast, Canada. *Canadian Journal of Earth Sciences* 29: 1235–1249.
- Murton JB (2009) Stratigraphy and paleoenvironments of Richards Island and the eastern Beaufort Continental Shelf during the last glacial–interglacial cycle. *Permafrost and Periglacial Processes* 20: 107–125.
- Osterkamp TE and Romanovsky VE (1999) Evidence for warming and thawing of discontinuous permafrost in Alaska. *Permafrost and Periglacial Processes* 10: 17–37.
- Romanovsky VE and Osterkamp TE (1995) Interannual variations of the thermal regime of the active layer and near-surface permafrost in northern Alaska. *Permafrost and Periglacial Processes* 6: 313–335.
- Romanovsky VE, Smith SL, and Christiansen HH (2010) Permafrost state in the northern polar hemisphere during the International Polar Year 2007–2009: A synthesis. *Permafrost and Periglacial Processes* 21: 106–116.
- Smith MW and Riseborough DW (2002) Climate and the limits of permafrost. *Permafrost and Periglacial Processes* 13: 1–15.
- Smith CAS, Burn CR, Tarnocai C, and Sproule B (1998) Air and soil temperature relations along an ecological transect through the permafrost zones of northwestern Canada. In: *Proceedings of the Seventh International Conference on Permafrost*, Yellowknife, NWT, 23–27 June 1998. Québec, Nordica, pp. 1009–1015.
- Tarnocai C, Canadell JG, Schuur EAG, Kuhry P, Mazhitova G, and Zimov S (2009) Soil organic carbon pools in the northern circumpolar permafrost region. *Global Biogeochemical Cycles* 23: GB2023. <http://dx.doi.org/10.1029/2008GB003327>.
- Walter KM, Zimov SA, Chanton JP, Verbyla D, and Chapin FS (2006) Methane bubbling from Siberian thaw lakes as a positive feedback to climate warming. *Nature* 443: 71–75.
- Williams J and Smith MW (1989) *The Frozen Earth: Fundamentals of Geocryology*. Cambridge: Cambridge University Press.

Frost Mounds: Active and Relict Forms

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Introduction

Frost mounds are ice-cored hills or hummocks found in areas of permafrost and/or seasonally frozen ground. They comprise pingos, palsas, lithalsas, and seasonal frost mounds and blisters. There is significant diversity in their scale, host material, mechanism of formation, ground-ice type, and origin. One process, however, is common to all frost mounds: the subsurface formation of ground ice (injection or segregation ice) generates upward forces able to heave and deform overlying frozen materials, resulting in the upward doming of the ground surface.

The collapse of frost mounds can result from mechanical processes (e.g., extension of the overburden material to expose the ice core) or climate forcing (e.g., thickening of the active layer into the ice core). Melting of an ice core over a prolonged period is liable to lead to the collapse of a frost mound. As the ice core melts out, significant sediment redistribution toward the margins of the landform can occur through mass movement and surface wash. These processes (thaw and redistribution of sediment) have the potential to generate annular ring-like ridges or ramparts surrounding central water-filled depressions. Not all frost mounds, however, generate a geomorphological signature of their demise; preservation potential is highly dependent on local factors, particularly the host material. Nevertheless, numerous examples of circular or near-circular enclosed depressions have been recognized in present-day temperate areas. Many of these landforms were initially interpreted as the remains of pingos formed during Quaternary cold periods. Wisely, however, some researchers were more cautious (Sparks et al., 1972), and many landforms once interpreted as pingo 'scars' have since been reinterpreted as the remains of other types of frost mounds (see Pissart, 2000, 2003 and references therein), or even as landforms formed by glacial processes (Iannicelli, 2003).

This article describes (i) the growth and collapse of active frost mounds in present-day periglacial environments, and (ii) Quaternary landforms ('ramparted depressions') interpreted as relict frost mounds. It focuses on those contemporary frost mounds that are likely to have a long-term geomorphological signature (i.e., pingos, mineral palsas, and lithalsas).

Modern-Day Frost Mounds

Pingos

Pingos are perennial frost mounds found in the continuous and discontinuous permafrost zones (Figures 1 and 2). They are typically produced by the injection and subsequent freezing of water within near-surface sediments or bedrock, resulting in the formation of a massive ground-ice body and the heave of overlying materials.

The injection of water into near-surface permafrost to form a pingo ice core capable of deforming units of overlying frozen sediments (often several meters thick) and dome the land surface requires high subsurface water pressures. These pressures can develop in two ways: either (i) in a closed system, as a result of water expulsion from saturated coarse-grained sediments during refreezing of a talik (a zone of unfrozen sediment within continuous permafrost); or (ii) in an open system, because of artesian water pressures within a subpermafrost aquifer (Holmes et al., 1968; Mackay, 1998; Müller, 1959). In permafrost environments, both processes lead to the upward injection and freezing of water. Pingos are categorized on the basis of their process of formation: the first type is referred to as 'closed-system pingos' (or hydrostatic pingos) and the second type as 'open-system pingos' (or hydraulic pingos). The former occur in lowland settings within the continuous permafrost zone, generally in localities where surface lake drainage has occurred, while the latter are more common in valley-bottom and footslope localities in both discontinuous and continuous permafrost.

Closed-system pingo formation

Much of our understanding of the nature and origin of closed-system pingos derives from the seminal studies of Mackay in the area northeast of the Mackenzie Delta, Canada (Mackay, 1998). Here, permafrost is developed in a thick sequence of deltaic sands, often mantled by clays and silts. In these lowland areas, surface water accumulates in summer to form pools and lakes. If the lake depth is greater than the winter ice thickness, then the lower parts of the lake will remain liquid throughout the year and the mean annual lake bottom temperature will be above zero. This will have a strong thermal influence on the underlying permafrost, and a thaw bulb or talik will develop below the lake (Figure 3).

The lakes of this region are typically impounded by perennially frozen but unconsolidated sediments. As a result, the lakes are highly susceptible to sudden drainage events generated by thermal erosion around the lake shore and by mass-wasting processes. When lakes drain, the underlying sediments are no longer insulated by the overlying water body, and so the talik begins to refreeze. Permafrost aggrades both from the surface downwards and from the lateral margins of the talik inwards (Figure 3). When the unfrozen saturated sands in the talik freeze, the 9% volume increase associated with the phase change from liquid water to ice results in expulsion of approximately 9% of the porewater ahead of the advancing freezing front. This expelled water migrates, under hydrostatic pressure, to the zone of least confining stress, usually below a residual pond within the drained lake basin (Figure 3). Freezing from the surface downwards then initiates the formation of the pingo ice core.



Figure 1 View of a closed-system pingo, Mackenzie Delta, Canada. Photograph by Charles Harris.



Figure 2 Open-system pingo, Adventdalen, Svalbard. The pingo has a height of ~30 m and a diameter of around 200 m. Photograph by Neil Ross.

If the rate at which expelled porewater accumulating beneath the bottom of the pingo ice core exceeds the downward rate of freezing of the water, then a subpingo water lens develops (Figure 3). In this instance, the hydrostatic pressure is able to support the overburden stress and cause further uplift and deformation of the overlying ice core and frozen overburden (Mackay, 1998). Monitoring of closed-system pingos indicates that high hydrostatic pressures in subpingo water lenses can mechanically fracture the overlying materials, leading to (i) rupturing of the pingo sides, (ii) high-angle faulting of overlying frozen sediments, (iii) formation of surface springs, and (iv) development of winter icings.

Closed-system pingos in northwest Canada may exceed 40 m in height, though the vast majority are <20 m high (Mackay, 1998). Basal diameters up to about 250 m have also been recognized. The diameter and plan form of closed-system pingos depend on the size and shape of the original talik, which is often circular or elliptical beneath thaw ponds, but can be markedly linear beneath elongated large water bodies, leading to the formation of linear, ridge-like pingos (Pissart and French, 1976).

Importantly, closed-system pingos have the potential to grow extremely rapidly. Data collected from a pingo in northwest Canada, which began growing soon after 1950, showed rapid initial growth that then slowed (Mackay, 1998). However, Mackay was careful to warn that generalizations on pingo

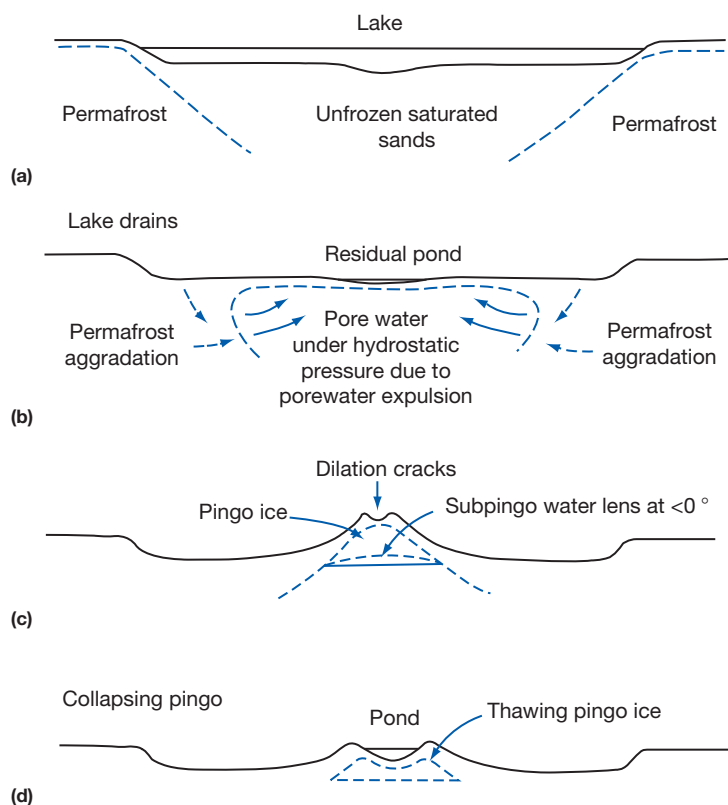


Figure 3 Formation of closed-system pingos. (a) Development of a talik beneath a thaw lake. (b) Lake drainage leading to refreezing of the talik and pore water expulsion as the permafrost table advances inwards. (c) Progressive freezing of a pressurized subpingo water lens leading to thickening of the pingo ice core and pingo growth. (d) Pingo collapse due to partial thaw of pingo ice beneath a central pond. Reproduced from Mackay JR (1998) Pingo growth and collapse, Tuktoyaktuk Peninsula area, Western Arctic Coast, Canada: A long term field study. *Géographie Physique et Quaternaire* 52(3): 271–323, with permission.

growth rates based on this sole example were probably inappropriate. Only some of the pingos monitored in his studies showed a progressive uplift: some pingos ruptured, causing water loss from subpingo water lenses and subsequent subsidence, while others showed up-and-down oscillations (Mackay, 1998).

Open-system pingo formation

In contrast to the closed-system type, open-system pingos are most common in areas of marked relief, often in lower valley sides or in valley bottoms (Figure 2). The first detailed description of open-system pingos was by Müller in East Greenland (Müller, 1959), but long-term studies of open-system pingo growth and decay have not been undertaken. Much less is therefore known, from direct observations, about their origins and internal structure than for closed-system forms.

Many open-system pingos occur in areas with thin and/or discontinuous permafrost. They have been described in central Alaska (Holmes et al., 1968), the Yukon Territory, Canada, the Tibetan Plateau, and central Yakutia, Russia. However, they are also found in areas of continuous permafrost, and have been recorded in Greenland (Müller, 1959; Worsley and Gurney, 1996) and Svalbard (Liestøl, 1977). Open-system pingos develop within a range of superficial sediments, including fluvial and slope deposits, and some even develop in weathered bedrock (Holmes et al., 1968; Liestøl, 1977; Müller, 1959). Open-system pingos are often somewhat smaller than their closed-system counterparts and are reported as single, isolated features and in clusters or groups of mutually interfering landforms of different ages.

The most widely cited model of open-system pingo formation is that of Holmes et al. (1968), who observed that nearly all pingos of Central Alaska lie on gentle to moderate slopes near the slope base in minor valleys and in gently sloping valley-bottom locations. Their preferential development on south- and southeast-facing slopes was attributed to thinner permafrost on slopes with these aspects. Holmes et al. (1968) suggested that local groundwater flow was crucial in determining the location and likelihood of open-system pingo development (Figure 4). Pressurized groundwater at intermediate depths beneath the valley sides has a tendency to flow toward the ground surface within subpermafrost unfrozen sediments or through bedrock fractures. Groundwater flow is restricted, however, by the base of the overlying permafrost, which acts like an aquiclude, confining upward flow (Figure 4). But near the base of the slopes – where the artesian pressure of groundwater sourced from the more elevated parts of the topography is greatest, and where thicker permafrost below the valley floor blocks the lateral flow of subpermafrost groundwater – water is forced upwards toward the ground surface. This water will either freeze in the near-surface to form a pingo, or – if it is too warm or the permafrost is locally absent – it will issue from the ground surface as a spring. Whether one or the other develops appears to be a very delicate balance of local thermal and geological conditions: Holmes et al. (1968) reported active and apparently perennial springs in association with some open-system pingos, a feature also characteristic of open-system pingos in Greenland and Svalbard.

In Eastern Greenland, open-system pingos are formed in glaciogenic materials, alluvial sediments, and bedrock in the

continuous permafrost zone, where mean annual air temperature is around -10°C and permafrost is recorded to a depth of 220 m. Here, the hydraulic system leading to pingo growth is probably fed, to some degree, by deep intra- or subpermafrost groundwater (Müller, 1959; Worsley and Gurney, 1996), rather than by significant amounts of recharge from seasonal surface melt. The postulated deep origins for these waters is supported by measurements of warm ($>0^{\circ}\text{C}$) and highly mineralized water discharging from perennial groundwater springs close to pingos in northeast Greenland; in winter, the springs form large surface icings. These characteristics suggest geothermal warming of ground waters, perhaps as they percolate along structurally controlled seepage zones and through taliks determined by zones of high geothermal heat flow (Worsley and Gurney, 1996). Interestingly, some Greenland pingos occur in localized clusters. This tendency has been attributed not only to the importance of regional hydrogeology in determining localities suitable for pingo development, but also to the evolution of very localized groundwater pathways through time.

On Svalbard, open-system pingos are also common in valleys. A simple but elegant model that suggested a close link between the pingos and polythermal glaciers on Svalbard was proposed by Liestøl (1977). Although permafrost is thick ($\sim 100\text{--}200\text{ m}$) and continuous beneath the inland parts of Svalbard, Liestøl (1977) recognized that it thins rapidly beneath the margins of the polythermal glaciers, and is probably entirely absent beneath the thicker warm-based ice beneath glacier accumulation areas. Subglacial meltwaters are therefore able to recharge groundwater where permeable strata underlie the warm-based parts of these glaciers. This provides a major source of meltwater to maintain (and pressurize) the groundwater throughout a significant part of the year, despite the continuous nature of the surrounding permafrost. Beyond the glacier margins – in a system akin to that responsible for the development of Alaskan open-system pingos (Figure 4) – this glacier-sourced groundwater is able to migrate through confined subpermafrost aquifers, within which considerable artesian pressures can be generated. In large, broad valleys such as Reindalen (Figure 5), this artesian pressure is sufficient for groundwaters to reach the near-surface, where freezing can generate open-system pingo growth. As suggested for Greenland, groundwater flow paths in Svalbard may also be influenced by the geological structure and geothermal heat. Deep percolation and geothermal heating of groundwater are indicated by the warm surface springs in parts of western Svalbard (Liestøl, 1977). Many pingos in Svalbard also show surface discharge of water (Figure 2) whose elevated salt content significantly lowers the freezing point. Other similarities may exist between the open-system pingos of Greenland and Svalbard, inasmuch as deep groundwaters in Greenland are also likely to have their initial origins in glacial meltwaters, probably from the Greenland Ice Sheet.

A subtype of open-system pingo has been identified in low-lying areas below the maximum Holocene sea level in Svalbard and Greenland. These landforms presumably date from the late to mid-Holocene and formed where, following relative sea-level fall, permafrost has aggraded in newly exposed and highly frost-susceptible fine-grained marine muds that favor

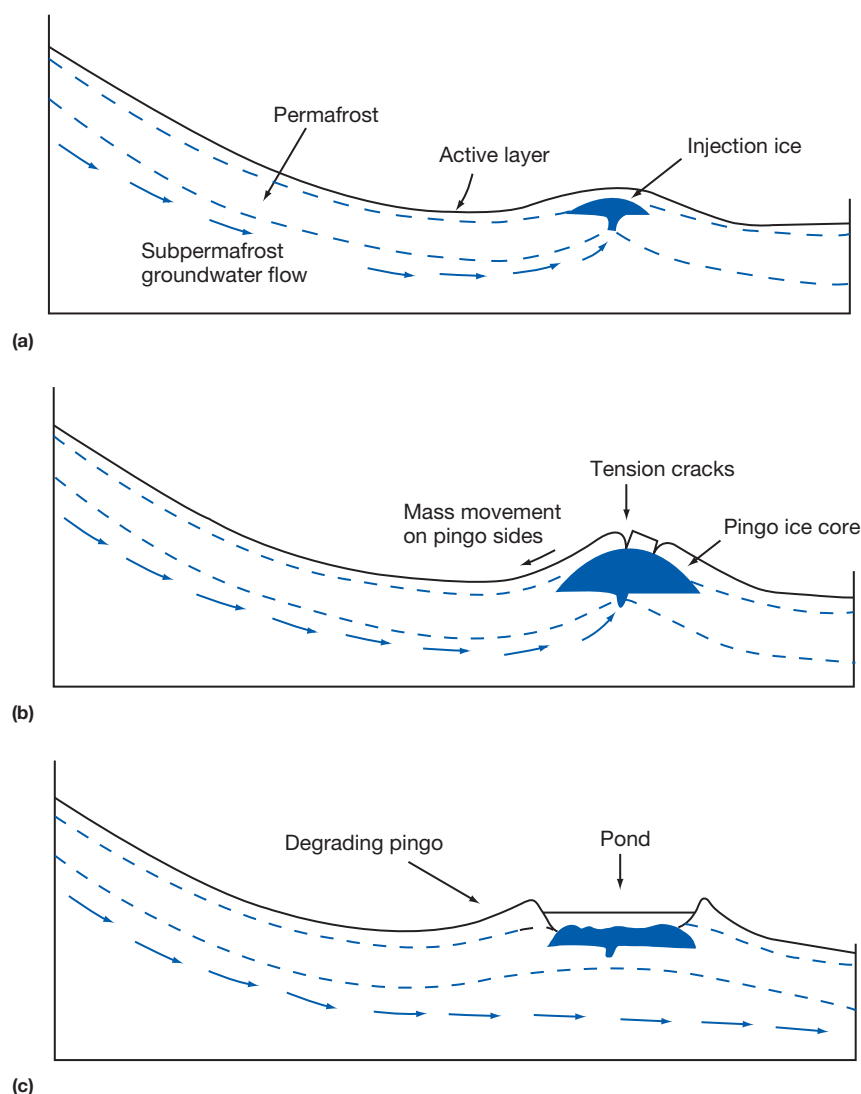


Figure 4 Cycle of formation and decay of open-system pingos. (a) Percolation of subpermafrost groundwater within a confined aquifer and injection into thin permafrost, initiating formation of a pingo ice core. (b) Formation of cracks due to tensional stresses developed as the surface is stretched by updoming of pingo center. (c) Development of central crater with pond as the ice core decays. Modified from Holmes GW, Hopkins DM, and Foster HL (1968) Pingos in central Alaska. *United States Geological Survey Bulletin* 1241-H: H1–H40; Ballantyne CK and Harris C (1994) *The Periglacialization of Great Britain*. Cambridge, UK: Cambridge University Press.



Figure 5 Open-system pingo, Reindalen, Svalbard. Photograph by Neil Ross.

ice segregation and frost heave (Ross et al., 2007). As a consequence, these pingos may be intermediate between closed- and open-system types. Such near-shore pingos, however, also share many characteristics with lithalsas.

Pingo collapse

Pingo degradation is normally initiated by the development of radial dilation cracks and concentric cracks caused by tensional stresses arising from bending and extension of the frozen overburden during pingo growth (Figures 4 and 6; Mackay, 1988). Sediment covering the ice core is lost because of slumping and solifluction of the active layer on the steep pingo sides (Babinski, 1982; Mackay, 1988), leading to the exposure of the pingo ice core, which is then highly susceptible to melting in summer months.

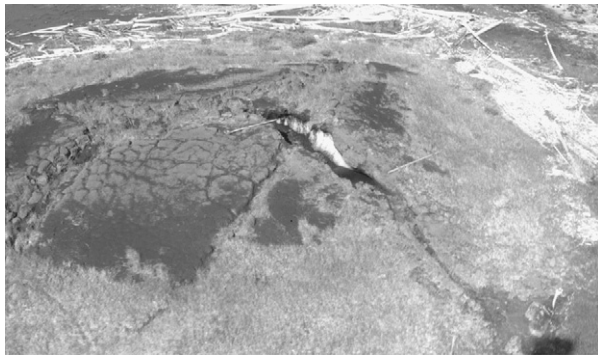


Figure 6 Closed-system pingo in the Mackenzie Delta, Canada, showing development of radial dilation cracks caused by tensional stress generated by upward growth. The exposed ice core is susceptible to thawing, initiating pingo collapse. Photograph by Charles Harris.



Figure 8 A collapsed closed-system pingo, Mackenzie Delta area, Canada. The pingo has an external diameter of approximately 250 m. Photograph by Ross Mackay.



Figure 7 Frozen lake within the crater of an open-system pingo, Reindalen, Svalbard. Photograph by Neil Ross.

Once melting of the ice core begins, the landform can be susceptible to an important positive feedback. As the ice core melts, a summit crater is generally formed, surrounded by a complex arrangement of ridges and mounds. These ridges may impound meltwater from the melting ice core, to form a lake. Crater lakes, many with perennial groundwater springs and outflow streams, have been reported from active pingos in Canada (Mackay, 1988), Greenland (Müller, 1959), Svalbard (Liestøl, 1977), Mongolia (Babinski, 1982), and Alaska (Holmes et al., 1968) (Figure 7). Because of the high thermal conductivity of water, melting of the underlying ice core can occur rapidly beneath these water bodies. The lake will also generally increase the rate at which lateral (thermal) erosion of the surrounding frozen ridge occurs. As degradation progresses, the lake increases in diameter, further increasing (i) the surface area of the ice core subject to melt and (ii) the length of crater-lake shoreline subject to thermal erosion.

Some processes, however, slow the rate of degradation. Intermittent, sometimes seasonal, breaching of crater lakes will result in outflow from, and possibly drainage of, the enclosed basin. If a crater-lake drains completely, then the rate of landform degradation may reduce significantly. Inward movement of thawing sediment from the crater rim is also

possible once degradation is well advanced, and this may also slow the degradation of the ice core by insulating it from warm air temperatures.

Complete landform degradation often leaves a pond surrounded by a circular, sometimes ice-cored, ridge of sediment. Although mature collapsed pingos with ramparts have been described from the contemporary permafrost zone (Figure 8; Babinski, 1982; Holmes et al., 1968; Mackay, 1988; Müller, 1959; Worsley and Gurney, 1996), there have been few investigations of their internal structures. Solifluction and slumping are clearly important processes in their development, as they result in a net transfer of sediment to the base of the mound where it accumulates to form a circular ridge. In addition to mass-movement processes, however, short-lived streamflow events may also play an important role in the redistribution of sediment necessary to form rampart ridges around collapsing pingos. Small-scale outlet channel forms, associated with small alluvial fans, have been observed around the periphery of several collapsing pingos (Mackay, 1988; Müller, 1959).

Other Frost Mounds

In addition to closed- and open-system pingos, arrays of smaller seasonal and perennial ground-ice mounds have been recognized. This section describes seasonal frost mounds, palsas, mineral palsas, and lithalsas. Because seasonal frost mounds and palsas have limited long-term preservation potential, emphasis is given here to mineral palsas and lithalsas, which are more likely to be represented in the Quaternary record.

Seasonal frost mounds

Ephemeral frost and icing mounds or blisters (Pollard, 1988), usually less than 5 m high, form in permafrost regions as a result of winter freeze-back of the active layer, trapping supra-permafrost groundwater fed by springs. Mounds are mantled with a thin layer of peat and soil, underlain by mineral soils, and contain bubble-rich or clear ice cores. These are sometimes layered and sometimes arched over a water-filled chamber within which sufficiently high hydrostatic pressures are developed in winter to uplift and deform the frozen overburden. Such landforms have a low preservation potential in the



Figure 9 Palsa mire, northern Finland, with low palsa mounds in the background and a decayed and collapsed mound in the foreground. Photograph by Matti Seppälä.

geological record and are unlikely to be found, or easily identified, in areas formerly affected by permafrost.

Palsas

Palsas are low permafrost mounds, up to around 7 m high and commonly up to 30 m in diameter (Figure 9). [Pissart \(2002\)](#) described palsas as “perennial mounds covered in peat, situated in the discontinuous permafrost zone and due chiefly to segregation ice fed by cryosuction.” Palsas develop in peatland mires in the circumpolar discontinuous or sporadic permafrost zones, and are normally observed in groups or ‘fields’ ([Gurney, 2001](#)). They form as a result of differential winter frost penetration. Deeper frost penetration in areas with thinner snow cover leads to frost heave due to ice segregation. The more deeply frozen peat is frost-heaved above the surrounding landscape, and in summer its surface layers dry, forming an effective insulating cover that limits the depth of thaw and preserves a frozen core beneath the palsa. During succeeding winters, snow is blown from the upper surface of the growing palsa but accumulates in the low-lying wetter areas, causing further differential frost penetration, ice segregation, and palsa growth ([Gurney, 2001](#)).

Palsas eventually decay when erosion of their sides and loss of peat cover from their top cause the frozen core to melt and collapse ([Gurney, 2001](#)). This often forms ponds (Figure 9) that are occasionally surrounded by a low ridge. Such ponds, however, rapidly fill with peat and their preservation potential is extremely low. All stages of cyclic palsa evolution (growth, maturity, and collapse) can commonly be observed in the same mire, though there is recent evidence for climatically driven degradation of palsa mires in Scandinavia and North America.

Mineral palsas and lithalsas

The frozen core of a palsa may extend through the peat cover into underlying frost-susceptible sediments. These sediments are heaved upwards by the formation of segregation ice, fed by cryosuction, to form a domed mineral core ([Gurney, 2001](#); [Pissart, 2002](#)). The peat cover of such ‘mineral palsas’ may be <10 cm thick and they appear to form part of a continuum of landforms caused by local redistribution of winter snow, differential frost penetration, and associated differential heave. At one end member are palsas formed entirely in peat, and at the

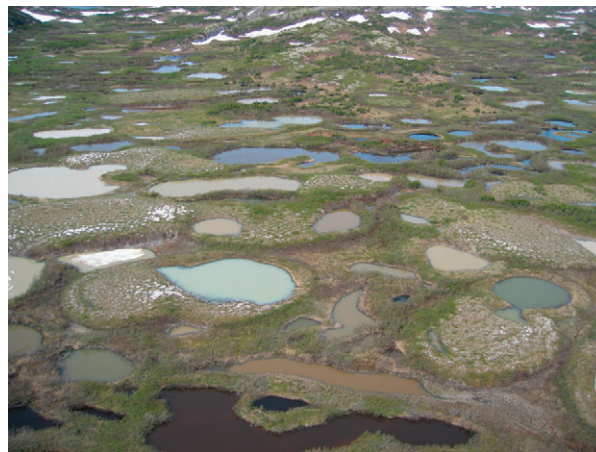


Figure 10 Decaying mineral palsas or lithalsas with central ponds and raised rims, 20 km inland from the eastern coast of Hudson Bay, near Umiujaq, northern Quebec, Canada. The features are developed in marine silts. Photograph by Michel Allard.

other are low (<10 m high) perennial frost mounds developed in sediments with little or no peat cover ([Gurney, 2001](#); [Pissart, 2002](#)). To distinguish them from other types of palsa, circular to oval mounds formed in sediments with no peat cover have been named ‘lithalsas’ ([Pissart et al., 1998](#)).

Mineral palsas and lithalsas occur in extensive contiguous fields, with younger features commonly disturbing older ramparts and depressions (Figure 10). They are particularly common in Scandinavia and North America. Although normally developed in fine-grained lacustrine and marine sediments ([Allard et al., 1987](#); [Pissart et al., 1998](#)), they can also form in areas with marine sands, in gravel, till, and stony glaciomarine diamicton ([Allard et al., 1987](#)). Besides being associated with low-lying zones of latitudinal permafrost, lithalsas have also been recognized in high-mountain environments. [Wünnemann et al. \(2008\)](#) mapped lithalsas at elevations >4500 m in the Himalayas. Unlike palsas developed entirely in peat, mineral palsas and lithalsas are more likely to generate enclosed rim-ridges during their collapse ([Gurney, 2001](#); [Pissart, 2002](#)). Because they are composed of mineral sediments, these landforms, like collapsed pingos, have much greater preservation potential than palsas.

Mineral palsa and lithalsa collapse

The growth of radial and concentric cracks, small-scale landsliding, thaw consolidation, and solifluction have all been reported from mineral palsas and lithalsas ([Allard et al., 1987](#); [Calmels et al., 2008a,b](#)). Like the collapse of pingos, the collapse of these smaller landforms results in the development of thermokarst depressions surrounded by ring-ridge ramparts (Figure 10). These are composed of mineral sediments accumulated by mass movement, as well as through lateral compression due to the growth of segregation ice ([Calmels and Allard, 2008](#)).

There is some evidence for a general trend of mineral palsa and lithalsa degradation in response to twentieth-century warming, although some studies emphasize the influence of local hydrology on their growth or decay. For example, [Lewkowicz and Coultish \(2004\)](#) showed that mineral palsa formation was controlled by the influence of beaver on the

local environment: frost-mound degradation followed flooding due to dam construction, and growth occurred after dams were breached and the ground surface was exposed. This unusual example clearly shows the influence that local hydrology has on frost-mound development, particularly in the sporadic or discontinuous permafrost zones.

Active frost mounds: a summary

Frost-mound growth depends on a complex interplay of processes and environmental conditions. Growth is strongly controlled not just by the presence or absence of permafrost (and hence the climatic regime), but also by the regional and local hydrogeological conditions, glacier meltwaters, geological structure, and the geothermal heat flux. Such complexity must be considered when interpreting Quaternary landforms that have all the hallmarks of relict frost mounds.

Relict Frost Mounds

Enclosed depressions surrounded by annular ring-ridges or ramparts have been interpreted as remnant landforms that mark the former location of Quaternary frost mounds. They are frequently found in clusters, and, in Western Europe, are sited on valley bottoms, at slope–valley interfaces, and in some plateau locations. Various terms have been introduced to describe these relict features. Given the complexities of interpretation, however, terms such as ‘relict frost mound’ or ‘ramparted depression’ are preferred to those with genetic connotations such as ‘pingo scar.’ Useful reviews of ramparted ground-ice depression sites of Pleistocene age from northwest Europe and North America are given by [Flemal \(1976\)](#) and [Pissart \(2003\)](#).

In Quaternary ramparted depressions, the ramparts may contain overturned, deformed, and tilted stratified sediments overlain by mass-movement deposits showing crude radially dipping stratification ([Pissart, 2000](#); [Ross et al., 2011](#)). The presence of these structures often provides important evidence for an origin related to the decay of ground-ice mounds.

The classic site where Pleistocene ramparted depressions were first detailed is on the Hautes Fagnes Plateau, Belgium. Here, enclosed, peat-filled depressions surrounded by distinct ramparts are widespread. The ramparts are commonly less than 1 m high, but some reach a height of 5 m ([Pissart, 2000, 2003](#)), and the thickness of basin infill varies between 1 and 7.5 m, with diameters of up to several hundred meters. These features were first interpreted as the remains of open-system pingos of the Younger Dryas age, dated by radiocarbon, tephra, and pollen stratigraphy methods ([Pissart, 2000, 2003](#)). The absence of an obvious aquifer to supply groundwater subsequently led Pissart to conclude that segregation ice rather than injection ice was responsible for the development of these landforms, and that they represent former lithalsas rather than open-system pingos. Their ramparts are thought to have formed by lateral compression and accumulation of slumped sediment ([Pissart, 2000](#)).

In Scandinavia and the British Isles, relict ramparted depressions ([Figure 11](#)) formerly interpreted as open-system pingos have also been reinterpreted as lithalsa remnants ([Pissart, 2000, 2003](#)). [Pissart \(2002\)](#) argued that the known spatial characteristics of contemporary pingos (widely spaced,



Figure 11 Relict ramparted depression thought to mark the former location of a Quaternary frost mound, Llanpumsaint, Wales, UK. Photograph by Neil Ross.

low number per unit area) and lithalsas (groups of closely spaced landforms, often very numerous, with a high concentration per unit area) allow simple interpretation of relict Pleistocene landforms. Interpretation of ramparted depressions becomes more complicated, however, when it is accepted that in past and present glaciated regions burial and subsequent melting of marginal glacier ice can form landforms topographically similar to those produced by the collapse of frost mounds. Extensive areas with ramparted depressions in the northern North American Prairies have long been interpreted not as ground-ice-related phenomena but as the product of subglacial and supraglacial processes ([Flemal, 1976](#); [Iannicelli, 2003](#); [Mollard, 2000](#)). Similar interpretations have been proposed in Scandinavia. Icebergs, grounded by lake drainage, have also been used to explain ring-ridges ([Mollard, 2000](#)). Therefore, while many European researchers have favored a periglacial origin for ramparted depressions, the North American literature tends to favor a glacial origin ([Flemal, 1976](#)). Contemporary studies in permafrost regions aimed at assessing the processes of pingo decay and collapse are therefore critical for the accurate identification and classification of relict Pleistocene ramparted ground-ice depressions. Recent studies of ramparted depression in Europe have been more aware of the risks associated with assuming that ramparted depressions represent a relict permafrost landform and have also assessed potential glacial origins ([Ross et al., 2011](#)). Even when detailed sedimentological and geophysical data are available from particular sites, however, precise interpretation is frequently difficult, given (i) the range of frost-mound types identified in modern cold environments and (ii) the array of glacial processes that can lead to the burial and melt-out of glacier ice, and therefore the formation of topographic depressions with enclosing ridges or ramparts.

Recent Developments in Frost-Mound Research

Several key factors have limited our understanding of frost-mound-related processes. These are (i) a lack of natural exposure (of both the internal structures of frost mounds and the surrounding host material), (ii) the cost and logistical challenges associated with creating artificial exposures (e.g., by drilling) in remote high-latitude permafrost environments, (iii) too great a focus on site-specific studies, and (iv) a lack of long-term monitoring. This section outlines a series of recent studies that have begun to address these research shortcomings.

To overcome the lack of natural exposure of frost-mound internal structure, and the costs associated with drilling, several recent studies have applied geophysical tools (ground penetrating radar, electrical resistivity, seismic, and electromagnetic

techniques) to image the internal structures of frost mounds (Fortier et al., 2008; Ross et al., 2007; Yoshikawa et al., 2006). Because of the contrasting physical properties of unfrozen water, ice, and host geological materials, these techniques offer significant potential for characterizing the internal structures, ice content, and host materials of frost mounds, which can then be used to infer the likely mechanisms of formation. The most comprehensive of these studies is that of Yoshikawa et al. (2006), who integrated a suite of ground and airborne geophysical data with borehole data to constrain the internal structures of open-system pingos in Alaska. These authors showed that two-dimensional (2D) electrical resistivity tomography seems to be the most successful single technique for imaging pingo internal structures (although it is best combined with others). Because pure ice is highly resistive, electrical resistivity is normally extremely useful for delineating the massive ice core in pingos; values of $>10\,000\ \Omega\ \text{m}$ are common for pingo cores (Yoshikawa et al., 2006). Ross et al. (2007) have shown, however, that the response of the pingo core can be highly dependent on the host material and the type of ice (injection or segregation ice). Near-shore open-system pingos in Svalbard are characterized by very low values of resistivity ($<2000\ \Omega\ \text{m}$) for permanently frozen ground. These results were attributed to pingo growth by segregation-ice development within fine-grained, saline marine clays (Ross et al., 2007), a mechanism not dissimilar to that by which lithalsas form.

Research into present-day frost mounds has tended to be rather focused on site-specific investigations and has, arguably, ignored the 'bigger picture.' Small-scale Geographic Information System (GIS) and remote-sensing studies of high-latitude environments show considerable potential for addressing this shortcoming and for contributing toward our understanding of frost-mound development. By far, the best example of this is the study of Grosse and Jones (2011), who compiled a geodatabase of over ~6000 pingos from openly accessible Russian map and remote-sensing datasets, and used GIS tools to analyze the spatial distribution of these landforms across Northern Asia. By integrating open-access permafrost, geological, climatological, hydrological, and elevation datasets, they were able to (i) quantify the key factors for pingo development across northern Asia (near-surface geology and hydrology), (ii) delimit the range of mean annual ground and air temperatures that the pingos in the region of study were found in, (iii) propose that 34% of the pingos studied are located in areas where climate models predict permafrost thaw by 2100, and which may initiate pingo degradation, and (iv) provide an up-to-date figure for the number of pingos on Earth ($>11\,000$).

Thanks to the work of Mackay (1998), closed-system pingos have been the subject of long-term monitoring studies, revealing key data concerning both internal structures and the way in which this type of frost mound forms and degrades. Until recently, however, comparable monitoring studies had not been undertaken on other types of frost mound. In 2000, a long-term monitoring and investigative study of a lithalsa was initiated in Quebec (Calmels et al., 2008a,b and references therein). During the period of monitoring (2000–2005), the lithalsa showed a pattern of long-term degradation, including the initial stages of rim-ridge formation (Calmels et al., 2008a). Tomodensitometric scanning of cores from the monitored lithalsa revealed unparalleled information regarding the ice

and gas content within the landform, while ground temperature measurements were used to infer the processes of groundwater migration and ice-lens growth (Calmels et al., 2008b). Long-term monitoring studies such as this provide critical information, and if we are ever to fully understand the mechanisms that drive the evolution of other frost mounds (e.g., open-system pingos), it will only be achieved through systematic investigations similar to those undertaken by Calmels et al. (2008a,b) on lithalsas.

The quantitative studies outlined above demonstrate that frost-mound research has clearly explored new and exciting avenues in recent years, not just by embracing and applying modern and novel techniques (geophysics and GIS) but also by doing the 'basics' correctly (through the setting up of long-term monitoring programs). These studies have provided important information regarding the processes responsible for the growth and collapse of modern-day frost mounds, information that is critical for making accurate interpretations of the likely mechanisms responsible for the formation of Quaternary ramparted depressions.

See also: [Permafrost and Periglacial Features: Permafrost; Thermokarst Topography.](#)

References

- Allard M, Seguin MK, and Lévesque R (1987) Palsas and mineral permafrost mounds in northern Quebec. In: Gardiner V (ed.) *International Geomorphology*, pp. 285–309. London: John Wiley.
- Babinski Z (1982) Pingo degradation in the Bayan-Nuurin-Khotnor Basin, Khangai Mountains, Mongolia. *Boreas* 11: 291–298.
- Calmels F and Allard M (2008) Segregated ice structures in various heaved permafrost landforms through CT scan. *Earth Surface Processes and Landforms* 33: 209–225.
- Calmels F, Allard M, and Delisle G (2008a) Development and decay of a lithalsa in Northern Quebec: A geomorphological history. *Geomorphology* 97: 287–299.
- Calmels F, Delisle G, and Allard M (2008b) Internal structure and the thermal and hydrological regime of a typical lithalsa: Significance for permafrost growth and decay. *Canadian Journal of Earth Sciences* 45: 31–43.
- Fleming RC (1976) Pingos and pingo scars: Their characteristics, distribution and utility in reconstructing former permafrost environments. *Quaternary Research* 6: 37–53.
- Fortier R, LeBlanc A-M, Allard M, Buteau S, and Calmels F (2008) Internal structure and conditions of permafrost mounds at Umiujaq in Nunavik, Canada, inferred from field investigation and electrical resistivity tomography. *Canadian Journal of Earth Sciences* 45: 367–387.
- Grosse G and Jones BM (2011) Spatial distribution of pingos in northern Asia. *The Cryosphere* 5: 13–33.
- Gurney SD (2001) Aspects of the genesis, geomorphology and terminology of palsas: Perennial cryogenic mounds. *Progress in Physical Geography* 25: 249–260.
- Holmes GW, Hopkins DM, and Foster HL (1968) Pingos in central Alaska. *United States Geological Survey Bulletin* 1241-H: H1–H40.
- Iannicelli M (2003) Reinterpretation of the original DeKalb Mounds in Illinois. *Physical Geography* 24: 170–182.
- Lewkowicz AG and Coulthill TL (2004) Beaver damming and palsa dynamics in a subarctic mountainous environment, Wolf Creek, Yukon Territory, Canada. *Arctic, Antarctic, and Alpine Research* 36: 208–218.
- Liestøl Ø (1977) Pingos, springs, and permafrost in Spitsbergen. *Norsk Polarinstitutt Årbok* 1975: 7–29.
- Mackay JR (1988) Pingo collapse and paleoclimatic reconstruction. *Canadian Journal of Earth Sciences* 25: 495–511.
- Mackay JR (1998) Pingo growth and collapse, Tuktoyaktuk Peninsula area, Western Arctic Coast, Canada: A long term field study. *Géographie Physique et Quaternaire* 52: 271–323.
- Mollard JD (2000) Ice-shaped ring-forms in Western Canada: Their airphoto expressions and manifold polygenetic origins. *Quaternary International* 68–71: 187–198.

- Müller F (1959) Beobachtungen über Pingos. *Meddelelser om Grønland* 153: 1–127. (translated by Sinclair DA, National Research Council of Canada, Ottawa, TT-1073, 1963).
- Pissart A (2000) Remnants of lithalsas of the Hautes Fagnes, Belgium: A summary of present day knowledge. *Permafrost and Periglacial Processes* 11: 327–355.
- Pissart A (2002) Palsas, lithalsas and remnants of these periglacial mounds: A progress report. *Progress in Physical Geography* 26: 605–621.
- Pissart A (2003) The remnants of Younger Dryas lithalsas on the Hautes Fagnes Plateau in Belgium and elsewhere in the world. *Geomorphology* 52: 5–38.
- Pissart A and French HM (1976) Pingo investigations, north-central Banks Island, Canadian Arctic. *Canadian Journal of Earth Sciences* 13: 937–946.
- Pissart A, Harris S, Prick A, and Van Vliet-Lanoë B (1998) La signification paleoclimatique des lithalsas (palses minerales). *Biuletyn Peryglacjalny* 37: 141–154.
- Pollard WH (1988) Seasonal frost mounds. In: Clark MJ (ed.) *Advances in Periglacial Geomorphology*, pp. 201–229. Chichester: Wiley.
- Ross N, Brabham PJ, Harris C, and Christiansen HH (2007) Internal structure of open system pingos, Adventdalen, Svalbard: The use of resistivity tomography to assess ground-ice conditions. *Journal of Environmental and Engineering Geophysics* 12: 113–126.
- Ross N, Harris C, Brabham PJ, and Sheppard TH (2011) Internal structure and geological context of ramparted depressions, Llanpumsaint, Wales. *Permafrost and Periglacial Processes* 22: 291–305.
- Sparks BW, Williams RBG, and Bell FG (1972) Presumed ice depressions in East Anglia. *Proceedings of the Royal Society A* 327: 329–343.
- Worsley P and Gurney SD (1996) Geomorphology and hydrogeological significance of the Holocene pingos in the Karup Valley area, Traill Island, northern east Greenland. *Journal of Quaternary Science* 11: 249–262.
- Wünnemann B, Reinhardt C, Kottlia BS, and Riedel F (2008) Observations on the relationship between lake formation, permafrost activity and lithalsa development during the last 20 000 years in the Tso Kar Basin, Ladakh, India. *Permafrost and Periglacial Processes* 19: 341–358.
- Yoshikawa K, Leuschen C, Ikeda A, et al. (2006) Comparison of geophysical investigations for detection of massive ground ice (pingo ice). *Journal of Geophysical Research* 111: E06S19.

Slope Deposits and Forms

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Periglacial slope processes are largely controlled by ground freezing and thawing leading to rock weathering, frost heave, and thaw consolidation. Rock weathering is predominantly mechanical and releases rockfalls on steep slopes that form scree deposits. Frozen sediments within permafrost and within the seasonally frozen active layer are frequently ice rich and therefore produce high moisture contents when they thaw. Periglacial regions are therefore very susceptible to both rapid slope failures and accelerated soil creep or solifluction, and mass movement deposits frequently mantle both modern and relict periglacial slopes.

Where scree materials lying within the permafrost zone contain interstitial ice—for instance, through burial of snow by rockfall—creep deformation of the frozen ground may generate distinctive flow structures, termed rock glaciers. The presence of a shallow active layer over a largely impermeable permafrost table promotes overland flow and slope wash. In steeper terrain, debris flows are common where either rapid snowmelt or high-intensity rainfall events release large volumes of water into the active layer. Thus, the range of slope processes and landforms in periglacial regions is high although in general dominated by gravitational mass movement. Periglacial processes frequently operate on pre-existing glacial terrain and contribute to the rapid ‘paraglacial’ landform adjustment that follows deglaciation.

Significance of Ground Freezing and Thawing

Permafrost regions are characterized by seasonal thawing and refreezing of the active layer, whereas in non-permafrost periglacial regions, winter frost penetrates into the ground from the surface. Where winter temperatures are variable, such as many mid-latitude mountain regions, freezing may be interrupted by periods of thaw. In high mountains at low latitudes, diurnal freezing and thawing cycles frequently dominate.

Water occupying pore spaces in sediments (and in porous rocks) does not all freeze at 0 °C. Thin films that wet the soil particles form a layer of bound water that freezes at temperatures well below zero (Williams and Smith, 1989; Yershov, 1998). Much of the pore water in fine-grained sediments may therefore remain unfrozen at subzero temperatures. Ice crystallization begins within slightly larger pore spaces that contain free or only weakly bound water, and here phase change leads to formation of lenticular ‘segregation ice.’ Progressive phase change sets up a cryosuction below the freezing front, causing water to migrate toward it from unfrozen zones below, to supply continued ice lens growth. This ice segregation process leads to an increase in volume of the frozen soil and frost heave of the surface, with ice contents in the frozen ground well in excess of the natural saturated moisture content. Silts and clays are prone to ice segregation when they freeze and are termed ‘frost-susceptible’ soils.

In seasonally frozen ground where no permafrost is present, and in the active layer above warm permafrost, segregation ice is likely to be evenly distributed provided there is a supply of water. However, where water is limited, ice segregation is concentrated in the upper part of the frozen profile. Within the active layer above cold permafrost, water can migrate downwards in summer to supply the growth of ice lenses in the base of the active layer and top of the permafrost. Thus, the phenomenon of ‘summer frost heave’ may be observed, associated with the development of an ice-rich zone at the permafrost–active layer interface (Mackay, 1983). In winter, a freezing front penetrates upwards from the permafrost table as well as downwards from the ground surface, leading to water migration toward both freezing fronts, leaving the central active layer relatively dry and ice-free.

Thawing of ice-rich frozen sediments is accompanied by closure of voids previously occupied by ice lenses and expulsion of the excess water they contain. This thaw consolidation process causes a reduction in soil volume and settlement at the surface. However, low permeability of the overlying thawed soil limits the rate at which excess water can be expelled (the still-frozen substrate below is largely impermeable). During this period of thaw settlement, self-weight stresses are transferred from particle contacts to the soil water, raising pore water pressures and reducing effective stress. The soil becomes soft and easily deformed, promoting downslope gravitational mass movement by solifluction and, in extreme cases, landslides.

The generation of pore water pressures during soil thaw depends largely on the ratio between the rate of thaw (which governs the rate of meltwater release) and the rate of consolidation (which is controlled by the rate of expulsion of excess meltwater). This ratio, the thaw-consolidation ratio, forms the basis of the thaw-consolidation theory (McRoberts and Morgenstern, 1974). Thus, a high value of R (approaching or greater than 1) indicates rapid thaw and slow drainage, a condition likely to raise pore pressures and reduce effective stresses, whereas a low value of R indicates slow thaw penetration, rapid drainage, and little tendency for excess pore pressures to develop.

Slow Mass Movement

Solifluction

The term solifluction was defined originally by Andersson (1909) as “slow flowing from higher to lower ground of masses of waste saturated with water” but was not restricted to cold climates. Washburn (1979) suggested the term gelifluction be used for saturated soil movement specifically associated with ground thawing. Since both frost creep and gelifluction are caused by frost heaving and subsequent thaw settlement, in practice they are often impossible to differentiate

in field measurements (Harris, 1981) and the term solifluction is now commonly used to include all such processes contributing to slow periglacial mass wasting in which surface displacement rates range from several millimeters to several tens of millimeters per year (Ballantyne and Harris, 1994). The process is commonly the dominant mechanism of sediment transport on periglacial slopes (Matsuoka, 2001).

Frost Creep

Frost creep, one of the main components of solifluction, is defined as the slow downslope gravitational deformation of slope materials caused by diurnal and annual cycles of freezing and thawing that lead to frost heave perpendicular to the slope surface followed by more nearly vertical thaw settlement (Washburn, 1979). Surface tension between soil grains creates apparent cohesion that prevents truly vertical resettlement, but during each freeze–thaw cycle there is a net downslope displacement of the disturbed soil. Maximum potential frost creep is given by vertical settlement and equals the product of frost heave and the tangent of the slope angle. Thus, the annual rate of surface movement depends on the amount of heave, the slope angle, and the frequency of frost heave cycles. Since diurnal freezing cycles generally penetrate only a few centimeters into the ground, associated frost creep affects only the uppermost few centimeters of soil (Matsuoka, 1998). Surface frost creep rates generally increase toward the surface, are generally less than 1 or 2 cm per year, and the depth of displacement depends largely on the depth of frost penetration but is usually less than 30 cm (Matsuoka, 2001).

Certain periglacial slopes may be affected by the diurnal formation of needle-like ice crystals a few centimeters in length that grow in clusters upward from the soil surface. Needle ice requires a frost-susceptible soil, a good supply of water from below, and sufficient permeability to allow rapid moisture migration to the freezing front. The rate of release of latent heat during phase change prevents downward advance of the freezing front and allows ice crystals to grow upward, perpendicular to the ground surface. Diurnal needle ice growth can lift soil particles and stones 50 mm or more. Higashi and Corte (1971) showed that surface displacements may exceed the potential frost creep value as a result of particle rolling and deformation of the supporting ice needles prior to thaw so that where needle ice cycles are frequent, rates of surface displacement may total several hundred millimeters per year.

Gelifluction

Gelifluction is the slow gravitational shear deformation of fine-grained frost-susceptible sediments (although such sediments may also contain abundant larger clasts) associated with high moisture contents and high pore pressures generated during thaw consolidation of ice-rich frozen ground. In a series of laboratory experiments simulating gelifluction in a silty ice-rich soil, Harris et al. (1997, 2001, 2003) demonstrated that pore pressures rise rapidly immediately following thaw (Figure 1), at which time the newly thawed soil has a low density, high moisture content, and an extremely low shear strength, making it susceptible to deformation under gravitational stresses below the threshold for slope failure.

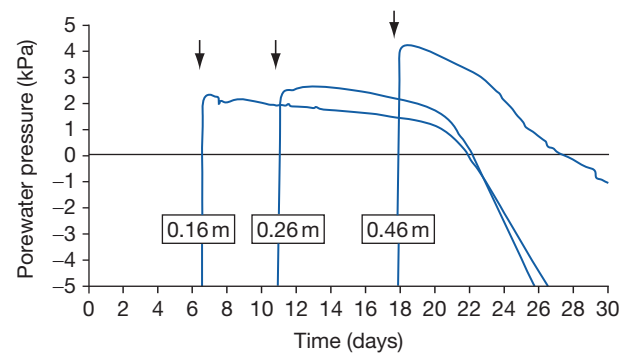


Figure 1 Pore pressures recorded during experimental simulation of solifluction. Data relate to the second thaw cycle in a 1/10-scale planar slope model of gradient 12° constructed of frost-susceptible natural fine sandy silt collected from Vire in Normandy, France. The model was thawed under a gravitational acceleration of 10 gravities in the geotechnical centrifuge (Harris et al., 2001). Time and depth are scaled to the equivalent prototype scale. Arrows indicate time of thawing at the respective transducer depths.

Field measurements in Greenland by Washburn (1967) demonstrated surface movement through an annual cycle of surface-downward freezing and thawing (Figure 2a). Winter frost penetration from the surface caused frost heave (displacement from P1 to P2 in Figure 2a), and in spring, thaw of the ice-rich active layer from the surface resulted initially in combined surface settlement and downslope displacement (P2 to P3 in Figure 2a) and, finally, in thaw settlement approximately perpendicular to the soil surface, reflecting the closure of the voids but no further downslope movement (P3 to P4 in Figure 2a). Full-scale laboratory simulation of gelifluction by Harris and Davies (2000) demonstrated very similar soil surface movements through a cycle of downward freezing and downward thawing (Figure 2b). In cold permafrost regions, however, little active-layer displacement may occur until late in the summer, when the ice-rich basal zone begins to melt. At this time, the active layer moves downslope en masse across the deforming basal zone softened by high pore pressures generated during thaw consolidation (Lewkowicz and Clarke, 1998). Such late summer gelifluction was termed ‘plug-like flow’ by Mackay (1981) and has been modelled in large-scale laboratory experiments that simulated a sloping active layer above ice-rich permafrost (Harris et al., 2008).

Several authors have described gelifluction as the slow viscous flow of soil that occurs annually during thawing when field moisture content exceeds the liquid limit (Smith, 1988; Harris et al., 1997). However, scaled simulation modeling in a geotechnical centrifuge by Harris et al. (2003) demonstrated that soil deformation during gelifluction is not a viscosity-controlled process but, rather, the prefailure elastoplastic deformation of a frictional soil whose strength is controlled largely by pore water pressures. Profiles of soil movement have been observed through burial and reexcavation of marker columns and tubes. Matsuoka (2001) identified four styles of slow periglacial mass movement associated with (a) frost creep due to diurnal needle ice, (b) frost creep due to diurnal segregation ice, (c) frost creep and gelifluction due to one-sided

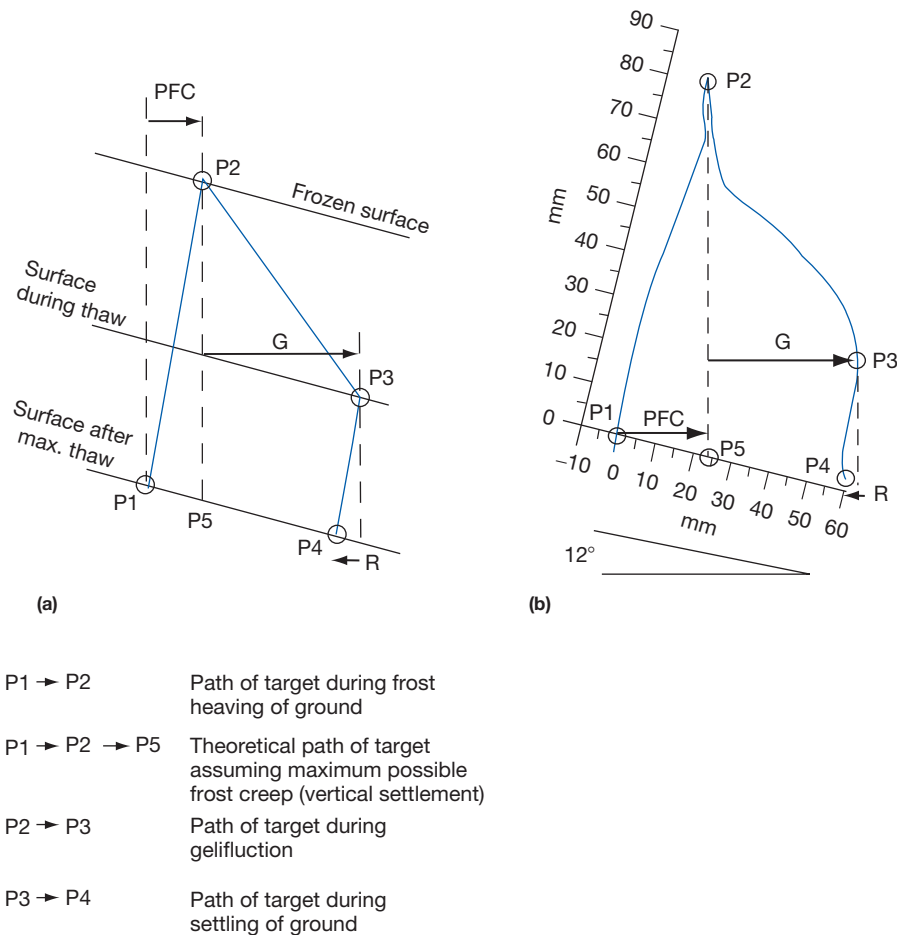


Figure 2 (a) Measured surface displacements, Mesters Vig, Greenland, over an annual cycle of ground freezing and thawing (Washburn, 1967). PFC, horizontal component of potential frost creep (with vertical resettlement of soil); G, horizontal component of gelifluction (downslope displacement in excess of potential frost creep); R, horizontal component of retrograde movement due to resettlement perpendicular to the soil surface. (b) Observed surface displacement over a cycle of ground freezing and thawing in an experimental simulation of solifluction processes. From Harris C and Davies MCR (2000) Gelifluction: Observations from large-scale laboratory simulations, *Arctic, Antarctic and Alpine Research* 32(2): 202–207. Reproduced with permission.

seasonal ground freezing, and (d) plug-like flow due to two-sided active-layer freezing (Figure 3).

Sediment Transport Rates

Measured solifluction transport rates in present-day non-permafrost periglacial regions are generally between 20 and $50 \text{ cm}^3 \text{ cm}^{-1} \text{ yr}^{-1}$, with a maximum close to $120 \text{ cm}^3 \text{ cm}^{-1} \text{ yr}^{-1}$ (Matsuoka, 2001). In permafrost regions, transport rates are slightly higher, ranging up to $230 \text{ cm}^3 \text{ cm}^{-1} \text{ yr}^{-1}$, mainly due to the greater thickness of active layer affected by plug-like solifluction. Thus, sediment transport rates due to solifluction are one or more orders of magnitude higher than transport rates due to soil creep in humid temperate regions. State-of-the-art field instrumentation for monitoring of solifluction is described by Harris et al. (2007) and Matsuoka (2010).

Many hill slopes in the nonglaciated temperate zone, such as southern Britain, show clear evidence of Quaternary periglacial solifluction processes, both in their slope profiles and in their mantle of relict periglacial slope deposits dating to late Pleistocene cold stages (Ballantyne and Harris, 1994).

Landforms

On mountain slopes, solifluction rates vary in response to changing soil properties, vegetation, drainage, snow cover, and topography. Sediment accumulation in zones of slower movement leads to formation of distinctive step-like landforms referred to as solifluction lobes or terraces (Harris, 1981). Frost sorting and differential movements often result in a concentration of clasts in the risers, producing 'stone-banked' lobes and terraces, in contrast to those with vegetation-covered risers termed 'turf-banked' (Benedict, 1970). Solifluction lobes have risers up to approximately 1.5 m high, whereas tread lengths are a function of the frequency of lobes on a given profile but generally range up to several tens of meters (Figure 4a). Solifluction terraces are larger, with risers up to 3 m high, although risers up to 6.5 m high have been reported (Harris, 1981). Detailed morphological analyses of solifluction lobes have been carried out by Hugenholtz and Lewkowicz (2002) and Matsuoka et al. (2005). In continuous permafrost regions, solifluction rates are more uniform, and lobes and terraces are generally not observed. Here, solifluction tends to produce smooth convexo-concave

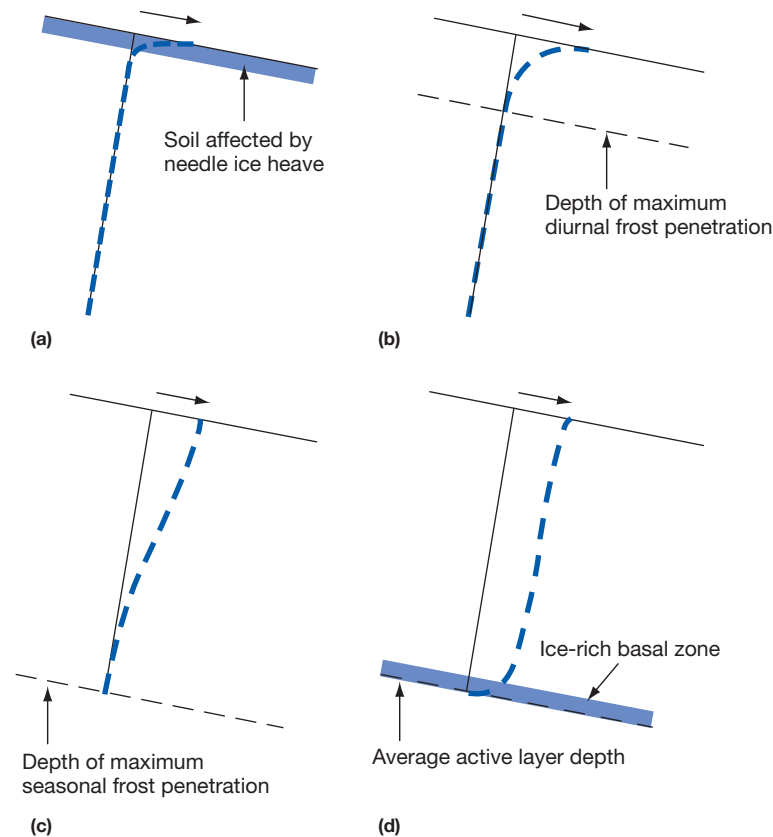


Figure 3 Typical frost creep and gelifluction displacement profiles: (a) diurnal needle ice, (b) diurnal frost penetration and formation of segregation ice, (c) seasonal frost penetration and formation of segregation ice, and (d) two-sided active-layer freezing above cold permafrost. Note the depths of soil displacement vary with depth of frost penetration or, in the case of permafrost, depth of summer thaw. Surface movements shown represent accumulated movement over several years. Classification based on Matsuo (2001).

slopes, sometimes with streamlines marked by elongation of surface patterned ground (Figure 4b).

During solifluction, coarser clasts are carried downslope within a deforming fine soil matrix, and as a result, distinctive shear fabrics develop in which elongates show a strong preferred downslope orientation, often with a slightly imbricate dip relative to the ground surface (Harris, 1981). At lobe and terrace fronts, transverse clast orientations approximately parallel to the alignment of the riser are observed, with b-axes dipping upslope. In many turf-banked solifluction lobes, downslope movement over the ground surface below leads to burial of soil and vegetation beneath the advancing lobe front. Excavation often reveals a complex series of buried soils extending beneath the lobe upslope from the front, and radiocarbon dating has allowed reconstruction of past rates of lobe frontal advance and, therefore, past variations in intensity of solifluction (Benedict, 1970; Matthews et al., 2005).

Rapid Mass Movements

Active-Layer Detachment Slides

These are shallow landslides developed in thawing soil overlying permafrost (Figure 5a). They occur on gentle to moderate slopes and are reported from North America and parts of

Siberia (Lewkowicz and Harris, 2005). Related ribbon-like flow-dominated slope failures common in the discontinuous permafrost zone following forest fires were referred to as 'skin-flows' by McRoberts and Morgenstern (1974). Observational evidence suggests that active-layer detachment slides may form rapidly, within a day, but may continue to be sporadically active over several days or even years (Lewkowicz, 1992, 2007). Slides are translational in nature, with the displaced mass maintaining its coherence over a basal shear zone and displaying lateral and terminal compression ridges containing multiple thrust blocks (Harris and Lewkowicz, 1993).

Detailed investigations of detachment slides in Canada, in the continuous permafrost zone of Ellesmere Island and the discontinuous permafrost zone of the Mackenzie Valley, have shown median length-to-width ratios of between 2.4 and 3.6 (Lewkowicz and Harris, 2005) and median depth-to-length ratios of 1 to 2%. Such extremely shallow landslides would be associated with flow-type failure in non-permafrost regions, but the presence of an ice-rich zone immediately above the permafrost table coupled with rapid thaw rates initiated by high late summer temperatures in Ellesmere Island and forest fire in the Mackenzie Valley (Harris and Lewkowicz, 2000) result in sufficiently high basal pore pressures to initiate slope failure even under the extremely low stress conditions on slopes as gentle as 5–10°. Coherence of the displaced active

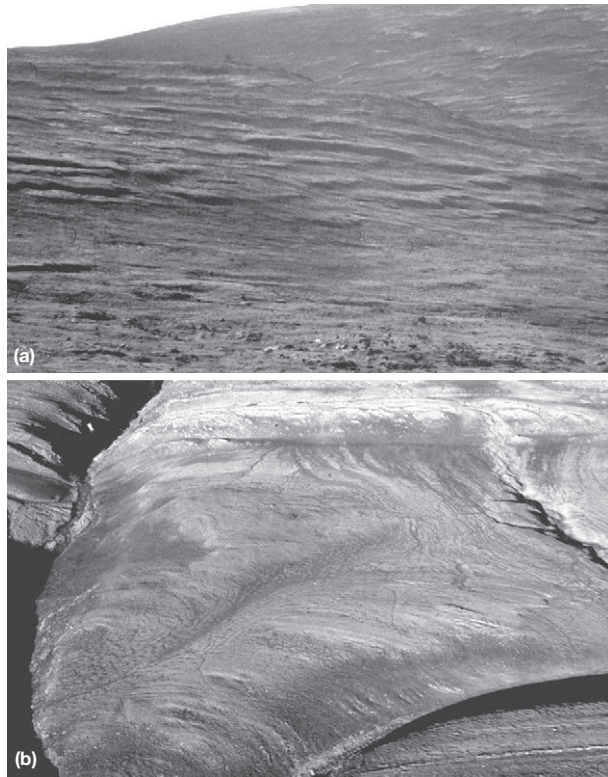


Figure 4 (a) Solifluction lobes in Jotunheimen, southern Norway, an area of seasonal frost penetration without permafrost. Lobe fronts are up to 1.5 m high. (b) Solifluction sheet, Ellesmere Island, Canada. This is an area of continuous permafrost, with active-layer depths less than 1 m, where with two-sided active-layer freezing causes plug-like flow (see Lewkowicz and Clark, 1998).

layer in Ellesmere Island detachment slides is due to two-sided freezing that generates cryosuction, causing water migration toward both the surface and the base of the active layer, leaving the central zone overconsolidated, dry, and stiff.

Ground-Ice Slumps

These are deeper slope failures that develop when ice-rich permafrost (as opposed to the active layer) becomes exposed to rapid thaw degradation, usually as a result of lateral erosion by rivers, lakes or coasts. The exposed ice-rich permafrost thaws back to form an embayment, releasing mudflows that carry away soft and very wet mud (**Figure 5b**). The rate of retreat of the ice-rich headwall depends on the air temperature, exposure to solar radiation, ground ice content, and sediment cover and features may persist for several years. Such slump structures developed in ice-rich permafrost are a component of thermokarst.

Debris Flows

Debris flows are common in mountain regions where gradients are steep. Although not restricted to the periglacial zone (**Coussot and Meunier, 1996**), debris-flow initiation is facilitated by the presence of thin debris covers on steep slopes that

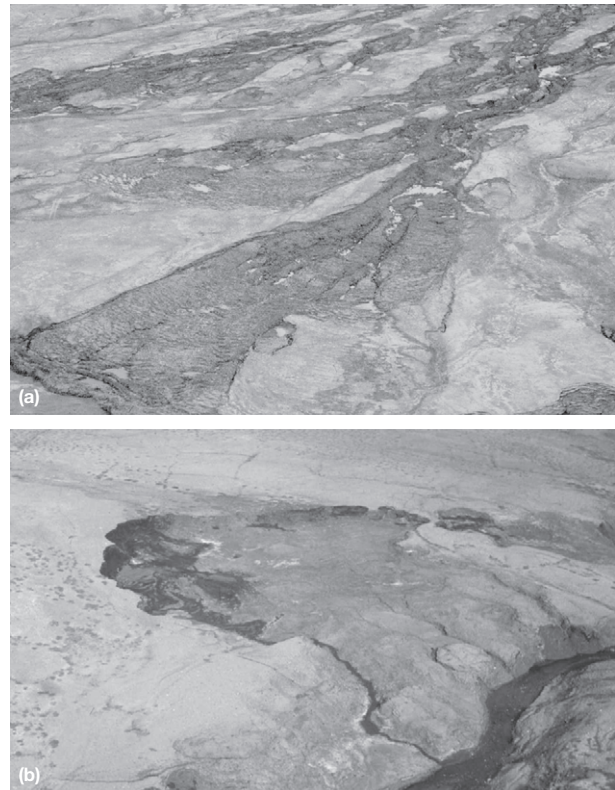


Figure 5 (a) Active-layer detachment slides, Ellesmere Island, Canada. The slide depth is generally less than 1 m and the length of the largest slide shown is approximately 750 m. (b) Ground-ice slump caused by lateral erosion by river exposing near-surface ground ice. Note rectangular-shaped ice-wedge polygons in the background.

are underlain by an impermeable permafrost table or shallow bedrock. Heavy rainfall events lead to rapid saturation of the debris, the development of seepage pressures, and initiation of rapid flows with high water contents (**Larsson, 1982**). The flow track is ribbon-like, flanked by bouldery levees, and terminates in a lobate fan-shaped frontal zone.

Relict Periglacial Slope Deposits

Where Quaternary cold stages were associated only with periglacial environments rather than glaciation, such as in southern England and northern France, slope deposits are locally derived (although sometimes with the addition of windblown loessic silts) and therefore closely reflect the lithology of the underlying bedrock. **Harris (1987)** and **Hutchinson (1991)** identified two distinct classes of periglacial slope deposits on the basis of their parent materials: coarser granular deposits derived from non-argillaceous bedrocks and clay-rich deposits derived from argillaceous bedrocks. Where Quaternary glaciation has occurred, tills are frequently reworked by subaerial cold-climate slope processes immediately following deglaciation, and they may closely resemble granular periglacial slope deposits. The presence of erratic clasts with surface striations and edge rounding, and a higher proportion of fines, may be used to identify such paraglacial deposits (**Harris, 1998**).

Granular Periglacial Slope Deposits

In Britain, granular periglacial slope deposits are often mapped as 'head deposits.' These generally comprise angular frost-weathered clasts derived from the underlying bedrock, set in a frost-susceptible silty to fine sandy matrix (Figure 6), and occur beneath smooth convexo-concave hill slopes, often thickening in the foot-slope zone to form several meters of valley fill. In coastal locations in both southern England and northern France, solifluction deposits mantle relict cliffs and bury raised beaches (Ballantyne and Harris, 1994). It is generally assumed that these deposits accumulated by solifluction, although rapid slope failures such as mudflows and detachment slides may



Figure 6 Head deposits, Wembury, south Devon. Bedrock is Devonian slate and the rockhead is progressively weathered and overturned as it merges upwards into the base of the head deposits.

also have been active, especially where slopes were steeper. Clasts show low-angle dips approximately parallel to the slope surface and are strongly oriented downslope.

Occasional layers and lenses of well-sorted silts and sands are thought to reflect episodes of slope wash and reworking of eolian silts. A common characteristic is a merging rockhead boundary in which bedrock becomes increasingly shattered, gradually merging upward into the base of the slope deposit and becoming overturned in a downslope direction (Figure 6).

In the chalklands of southern England, frost weathering driven primarily by seasonal freeze-thaw cycles has reduced near-surface chalk to a fine calcareous mud, which was apparently highly susceptible to solifluction. The resulting chalk head is termed 'coombe rock.' It forms a thin mantle over valley sides but thickens rapidly into the valley bottoms (Ballantyne and Harris, 1994). Coombe rock consists of silt-sized chalk powder containing occasional chalk clasts that become larger and more frequent closer to the rockhead and show strong downslope preferred orientations.

Periglacial Slope Deposits Derived from Clay Bedrock

Where bedrocks consist of clays and mudstones (e.g., the Lias clays of the Cotswolds and the Paleogene clays of southeast England), slope deposits comprise several meters of partially remolded clay, often containing fragments of more resistant sedimentary rocks derived from further upslope. Underlying and within these deposits are low-angle shear planes often lying approximately parallel to the ground surface (Chandler, 1972; Hutchinson, 1991; Ballantyne and Harris, 1994). Hutchinson (1991) and Spink (1991) illustrated the range of shears commonly observed in clayey periglacial solifluction deposits (Figure 7). The most extensive are the basal shears

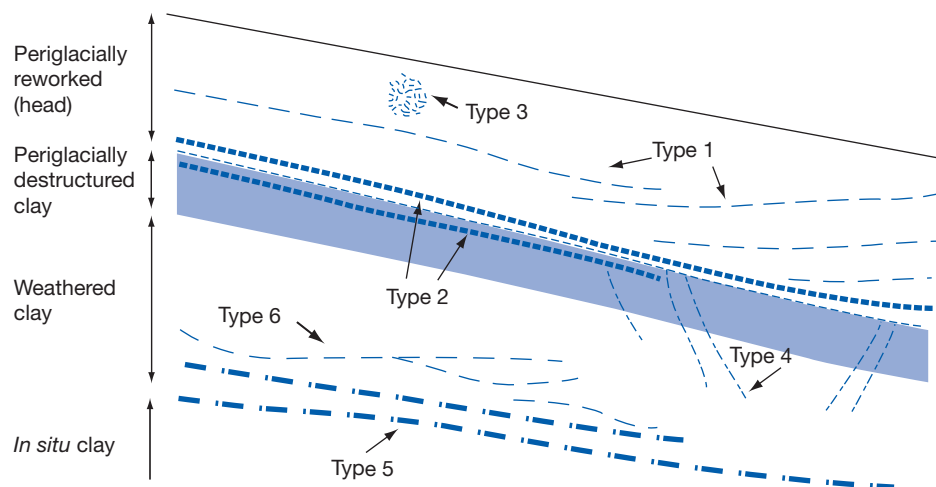


Figure 7 Schematic diagram showing classification of periglacial shears based on Hutchinson (1991) and Spink (1991). Type 1 are discontinuous, subhorizontal solifluction shears (Spink, 1991) and include emergent shears (Hutchinson 1991). Type 2 are extensive basal shears approximately parallel to the ground surface. These occur at the base of the clay head and may also be found close to the surface of, but within, the disturbed clay below. Type 3 are small-scale discontinuous shears, termed accommodation shears by Spink, within the clay head material. Hutchinson describes the clay head as having a mudslide fabric. Type 4 are high-angle thaw collapse shears associated with thermokarstic ground subsidence. Type 5 are subhorizontal principal underthaw or deep-seated landslide shears. Type 6 are undulose discontinuous secondary underthaw shears. Spink also includes as type 7 subvertical cracks thought to be associated with contraction of the clay during drying, but these are not shear surfaces and they probably postdate the periglacial phase.

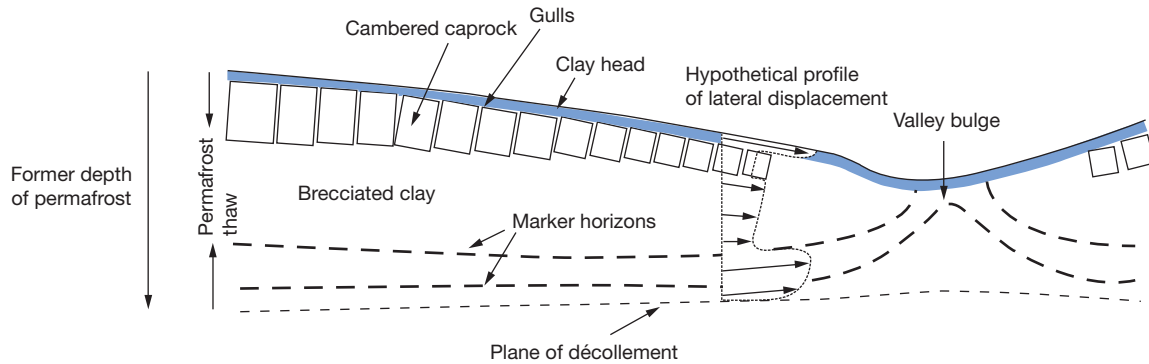


Figure 8 Model of ground disturbance associated with periglacial cambering based on the interpretation of Hutchinson (1991). The figure shows a hypothetical profile of lateral displacement. The clay head deposits are likely to be underlain by, and to contain, periglacial shear surfaces. Extrusion of clay above the basal décollement is considered to have resulted from upward thawing of the ice-rich permafrost base, resulting in softening by thaw consolidation.

over which the slope deposits have slipped. Smaller scale discontinuous shears within the displaced soil mass reflect internal deformation during mass movement.

These relict periglacial shear surfaces underlying clayey slope deposits indicate shallow translational landsliding, equivalent to the active-layer detachment slides of the present-day Arctic. Hutchinson (1991) illustrated the probable mechanism of failure (Figure 8). Since shear surfaces are observed beneath slopes as gentle as 4° , and very high pore pressures would be necessary to reduce the shear strength sufficiently to induce failure, it is concluded that the only mechanism capable of generating such high pore pressures is thaw consolidation of an ice-rich soil (active layer) above a permafrost table. The presence of these relict active-layer detachment slides is best known from engineering problems that they have caused, and possibly the best documented is near Sevenoaks in Kent, where road construction reactivated slope instability in sheared periglacial clays (Skempton and Weeks, 1976). Here, careful mapping revealed lobate landslide frontal zones, and boreholes showed at least two generations of slope instability, the most recent of which probably occurred during the Late Glacial Interstadial, after approximately 12 ka. The presence of relict shallow periglacially induced landslides has caused widespread engineering problems elsewhere in southeast England, the Cotswolds, and Lincoln Edge and further north in the Carboniferous mudrocks of Staffordshire and Derbyshire, where the failure of an earth dam at Carsington in 1984 was directly attributable to sheared soliflucted clay that extended beneath the dam foundation (Skempton et al., 1991).

Clay bedrocks are often affected not only by near-surface shallow active-layer landsliding but also by deeper-seated disturbance associated with permafrost growth and decay. Steep shears (Figure 7) extending many meters below the ground surface are interpreted as the result of thaw subsidence during melting of ice-rich permafrost, whereas extensive subhorizontal and less extensive undulose shears at a depth of up to 7 m were explained by Spink (1991) as deep-seated slides within degrading permafrost. It is postulated that upward thawing of the permafrost base due to geothermal heat flux led to high pore pressures as ice-rich permafrost degraded.

Cambering and Associated Structures

Although not widely reported from modern periglacial environments, deep-seated structures observed in Britain and termed cambering have been interpreted as originating under permafrost (Hutchinson, 1991; Ballantyne and Harris, 1994). Cambering occurs where large-scale valley incision has exposed more competent caprocks overlying less competent clay or mudrocks in the valley sides. Caprocks show gravitational extension down the valley sides, forming an attenuating drape, and extension is accommodated by the development of deep fractures termed gulls. Gulls run parallel to the contours and separate intact caprock blocks (Figure 8). Cambering is often associated with the development of anticlinal deformation, termed valley bulging, within the less competent strata beneath the valley axis. The development of cambering is summarized in Figure 8, which identifies the following stages:

1. Initial incision of the river leading to unloading and valley rebound.
2. Permafrost develops, with ice segregation processes increasing ice contents of the frozen clay.
3. Valleyward creep of the frozen clay in response to lateral stresses. This causes extension and initial cambering of caprocks.
4. Permafrost degradation due to large-scale climate warming leads to thaw from the surface downward, enhanced surficial solifluction, and displacement of caprock valleywards over the thaw-softened clays beneath.
5. Thawing of the permafrost base due to geothermal heat flux, thaw consolidation, and high pore pressures within an effectively confined thawed stratum.
6. Lateral extrusion of the thawed clay at the permafrost base leading to compression along the valley axis and pronounced increase in the valley bulge structure.

Most cambering and valley bulging in Britain apparently predates the Devensian glacial stage (Oxygen Isotope Stages 4–2). Hutchinson (1991) postulated that the Anglian stage might have corresponded with the most severe periglacial conditions, with widespread deep permafrost, although this

remains speculative. It seems likely that deep-seated bedrock disturbance took place during more than one Quaternary cold stage.

Periglacial Slope Evolution

An important consequence of the effectiveness of periglacial slope processes is that in the periglacial zone that extended beyond the Pleistocene glacial limits, the landscape is essentially periglacial in origin. Evidence takes the form of hill slope and valley floor deposits and the morphology of the slopes. Kirkby (1987) modeled slope evolution under periglacial conditions (dominated by frost weathering and solifluction). Using evidence from field measurements, he assumed that sediment transport by periglacial solifluction is at least an order of magnitude faster than temperate soil creep. The model predicts progressive decline in valley side gradients and development of rounded summits. Where sediment accumulates in the footslope zone (as in coastal areas during periods of low sea level), an extensive basal concavity develops as slope sediment accumulates. Estimated timescales for this slope degradation are well within the time span of cold stages during the Quaternary.

See also: **Permafrost and Periglacial Features: Active Layer Processes; Patterned Ground; Rock Weathering; Talus Slopes; Thermokarst Topography.**

References

- Andersson JG (1906) Solifluction, a component of subaerial denudation. *Journal of Geology* 14: 91–112.
- Ballantyne CK and Harris C (1994) *The Periglaciation of Great Britain*. Cambridge, UK: Cambridge University Press.
- Benedict JB (1970) Downslope soil movement in a Colorado alpine region: Rates, processes, and climatic significance. *Arctic and Alpine Research* 2: 165–226.
- Chandler RJ (1972) Periglacial mudslides in Vestspitsbergen and their bearing on the origin of fossil 'solifluction shears' in low angled clay slopes. *Quarterly Journal of Engineering Geology* 5: 223–241.
- Coussot P and Meunier M (1996) Recognition, classification and mechanical description of debris flows. *Earth-Science Reviews* 40: 209–227.
- Harris C (1981) *Periglacial Mass Wasting: A Review of Research*. Norwich, UK: Geo Books, BGRG Research Monograph.
- Harris C (1987) Solifluction and related deposits in England and Wales. In: Boardman J (ed.) *Periglacial Processes and Landforms*, pp. 209–224. Cambridge, UK: Cambridge University Press.
- Harris C (1998) The micromorphology of paraglacial and periglacial slope deposits: A case study from Morfa-Bychan, West Wales. *Journal of Quaternary Science* 13: 73–84.
- Harris C and Davies MCR (2000) Gelifluction: Observations from large-scale laboratory simulations. *Arctic, Antarctic and Alpine Research* 32(2): 202–207.
- Harris C and Lewkowicz AG (1993) Form and internal structure of active-layer detachment slides, Fosheim Peninsula, Ellesmere Island, N.W.T., Canada. *Canadian Journal of Earth Sciences* 30: 1708–1714.
- Harris C and Lewkowicz AG (2000) An analysis of the stability of thawing slopes, Ellesmere Island, Nunavut, Canada. *Canadian Geotechnical Journal* 37: 449–462.
- Harris C, Davies MCR, and Coutard J-P (1997) Rates and processes of periglacial solifluction: An experimental approach. *Earth Surface Processes and Landforms* 22: 849–868.
- Harris C, Kern-Luetsch M, Murton JB, Font M, Davies M, and Smith F (2008) Solifluction processes on permafrost and non-permafrost slopes: Results of a large scale laboratory simulation. *Permafrost and Periglacial Processes* 19: 359–378.
- Harris C, Luetsch MA, Davies MCR, and Smith FW (2007) Field instrumentation for real-time monitoring of periglacial solifluction. *Permafrost and Periglacial Processes* 18: 105–114.
- Harris C, Rea B, and Davies MCR (2001) Scaled physical modelling of mass movement processes on thawing slopes. *Permafrost and Periglacial Processes* 12(1): 125–136.
- Harris C, Davies MCR, and Rea BR (2003) Gelifluction: Viscous flow or plastic creep? *Earth Surface Processes and Landforms* 28: 1289–1301.
- Higashi A and Corte AE (1971) Solifluction: A model experiment. *Science* 171: 480–482.
- Hugenholtz CH and Lewkowicz AG (2002) Morphometry and environmental characteristics of turf-banked solifluction lobes, Klunane Range, Yukon Territory, Canada. *Permafrost and Periglacial Processes* 13: 301–313.
- Hutchinson JN (1991) Periglacial slope processes. In: Forster A, Culshaw MG, Cripps JC, Little JA, and Moon CF (eds.) *Quaternary Engineering Geology*, pp. 283–331. London: Geological Society Engineering Geology Special Publication 7.
- Kirkby MJ (1987) General models of long-term slope evolution through mass movement. In: Anderson MG and Richards KS (eds.) *Slope Stability*, pp. 357–379. Chichester, UK: Wiley.
- Larsson S (1982) Geomorphological effects on slopes of Longyearvalley, Spitsbergen, after a heavy rainstorm in July 1972. *Geografiska Annaler* 64A: 105–125.
- Lewkowicz AG (1992) Factors influencing the distribution and initiation of active-layer detachment slides on Ellesmere Island, Arctic Canada. In: Dixon JC and Abrahams AD (eds.) *Periglacial Geomorphology, Proceedings of the 22nd Annual Binghamton Symposium in Geomorphology*, pp. 223–250. Chichester, UK: Wiley.
- Lewkowicz AG (2007) Dynamics of active-layer detachment failures, Fosheim Peninsula, Ellesmere Island, Nunavut, Canada. *Permafrost and Periglacial Processes* 18: 89–103.
- Lewkowicz AG and Clarke S (1998) Late-summer solifluction and active layer depths, Fosheim Peninsula, Ellesmere Island, Canada. In: Lewkowicz AG and Allard M (eds.) *Proceedings of the 6th International Conference on Permafrost, Yellowknife, Canada*, pp. 641–666. Quebec, Canada: Centre d'études nordiques, University of Laval, Sainte-Foy.
- Lewkowicz AG and Harris C (2005) Morphology and geotechnique of active-layer detachment failures in discontinuous and continuous permafrost, northern Canada. *Geomorphology* 69: 275–297.
- Mackay JR (1981) Active layer slope movement in a continuous permafrost environment, Garry Island, Northwest Territories, Canada. *Canadian Journal of Earth Sciences* 18: 1666–1680.
- Mackay JR (1983) Downward water movement into frozen ground, western Arctic coast, Canada. *Canadian Journal of Earth Sciences* 20: 120–134.
- Matsuoka N (1998) The relationship between frost heave and downslope soil movement: Field measurements in the Japanese Alps. *Permafrost and Periglacial Processes* 9: 121–133.
- Matsuoka N (2001) Solifluction rates, processes and landforms: A global review. *Earth-Science Reviews* 55: 107–133.
- Matsuoka N (2010) Solifluction and mudflow on a limestone periglacial slope in the Swiss Alps: 14 Years of monitoring. *Permafrost and Periglacial Processes* 21: 219–240.
- Matsuoka N, Ikeda A, and Date T (2005) Morphometric analysis of solifluction lobes and rock glaciers in the Swiss Alps. *Permafrost and Periglacial Processes* 16: 99–113.
- Matthews JA, Seppälä M, and Dresser PQ (2005) Holocene solifluction, climate variation and fire in a subarctic landscape at Pippokangas, Finnish Lapland, based on radiocarbon-dated buried charcoal. *Journal of Quaternary Science* 20(6): 533–548.
- McRoberts EC and Morgenstern NR (1974) The stability of thawing slopes. *Canadian Geotechnical Journal* 11: 447–469.
- Skempton AW and Weeks AG (1976) The Quaternary history of the Lower Greensand escarpment and Weald Clay vale near Sevenoaks, Kent. *Philosophical Transactions of the Royal Society A283*: 493–526.
- Skempton AW, Norbury D, Petley DJ, and Spink TW (1991) Solifluction shears at Carsington, Derbyshire. In: Forster A, Culshaw MG, Cripps JC, Little JA, and Moon CF (eds.) *Quaternary Engineering Geology*, pp. 381–397. London: Geological Society Engineering Geology Special Publication 7.
- Smith DJ (1988) Rates and controls of soil movement on a solifluction slope in the Mt. Rae area, Canadian Rocky Mountains. *Zeitschrift für Geomorphologie Supplementband* 71: 25–44.

- Spink TW (1991) Periglacial discontinuities in Eocene Clay near Denham, Buckinghamshire. In: Forster A, Culshaw MG, Cripps JC, Little JA, and Moon CF (eds.) *Quaternary Engineering Geology*, pp. 389–396. London: Geological Society Engineering Geology Special Publication 7.
- Washburn AL (1967) Instrumental observations of mass-wasting in the Mesters Vig district, Northeast Greenland. *Meddelelser om Grønland* 166: .
- Washburn AL (1979) *Geocryology: A Survey of Periglacial Processes and Environments*. London: Arnold.
- Williams PJ and Smith MW (1989) *The Frozen Earth*. Cambridge, UK: Cambridge University Press.
- Yershov ED (1998) *General Geochryology*. Cambridge, UK: Cambridge University Press.

Periglacial Fluvial Sediments and Forms

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Introduction

Fluvial deposits in Quaternary sedimentary successions have often been characterized as 'periglacial' when indications exist that they have been deposited under cold-climate conditions. An older, now less frequently used term, mainly in the European literature, is the adjective 'fluvioperiglacial,' which has often been used to denote deposits of small-scale fluvial activity in periglacial environments (Van Huissteden, 1990).

Rivers inside the periglacial zone, just like those outside it, show a wide variety of morphology, a feature that depends on discharge regime, sediment supply, and gradient. In Arctic areas, examples of all the broad classes of river-channel planform, including braided, meandering, and anastomosing, occur (Vandenberghe, 2001, 2003; Vandenberghe and Woo, 2002; Figure 1). The physiographic diversity within the periglacial zone is considerable (French, 2007) and exerts a large influence on river type. Geology and tectonics influence basin bedrock and topography; conspicuous large-scale features are the shield areas of Canada and Scandinavia, the mountain ranges of the Eurasian and western American continents, and the vast sedimentary lowlands of northern Eurasia. Superposed on this is climatic variation, ranging from moist maritime climates with thick winter snow cover, ample summer precipitation and moderate yearly temperature variation, cold and dry desert conditions in the high Arctic, and continental interiors with semiarid conditions and extreme yearly temperature variation, for example, Central Yakutia. This in turn influences vegetation, which ranges from polar desert, tundra, and steppe to boreal forest. Present-day mountain glaciers and ice caps influence river hydrology and sediment supply, while past glaciation may have left behind large sediment stores that still supply sediment to the rivers (paraglacial landscapes; Slaymaker, 2009).

What rivers in the periglacial zone have in common is a discharge regime that is dominated by a pronounced snowmelt discharge peak in spring (Serreze and Barry, 2005), ice cover during part of the year and, in permafrost areas, the interaction with permafrost and ground ice (see section 'Landforms and Processes'). Interaction with aeolian deposition is common, for example, the cold-climate fluvio-aeolian deposits described by Bryant (1983).

Fluvial sediments deposited under periglacial conditions comprise a considerable part of the terrestrial stratigraphic record. Under periglacial conditions, sediment supply tends to be higher than in temperate climates (Vandenberghe, 2003; see below), and interglacials during the Quaternary were short compared to glacials. In fluvial successions in present-day temperate regions, sediments deposited under cold climates therefore tend to dominate. For instance, at

least 87% of the thickness of a 22-m-thick post-Saalian glacial basin fill in the Netherlands contained cold-climate fluvial deposits dating from the Last Glacial (Van Huissteden, 1990).

The effects of anthropogenic climate change are expected to be the strongest in the Arctic (ACIA, 2005), affecting the evolution of periglacial river systems. Many Arctic regions have already experienced increased precipitation and river discharge, and both factors are expected to increase further in the future as a result of climate warming (Walsh et al., 2005). In combination with enhanced permafrost degradation and sediment delivery to rivers, this may cause changes in river morphology. Periglacial river floodplains are also hotspots in the Arctic carbon cycle because they provide sinks for carbon, by storing organic carbon in sediments (e.g., Tarnocai et al., 2009; Van Huissteden, 1990).

Landforms and Processes

Morphology

The terms 'periglacial rivers' or 'periglacial river deposits' may suggest a clearly distinguished fluvial morphology or sedimentary facies linked to periglacial environments. But the variety of forms described above suggests that this is not the case. Vandenberghe and Woo (2002) have therefore argued that 'periglacial rivers' do not exist. By themselves, neither river planform nor sedimentary facies are diagnostic of periglacial environments; for instance, one cannot say that braided-river deposits indicate periglacial conditions. Instead, only smaller-scale morphological or sedimentary features are diagnostic of periglacial environments. One example is beaded drainage (see section 'Permafrost'). A second is hillslope drainage lines in low Arctic (vegetated) tundra, which have a singular appearance (mediated by solifluction), sometimes including recognizable channels, sometimes not. Sufficiently small streams may be strongly influenced by the frozen ground condition. As the river becomes larger, the special thermal conditions appear to matter less.

In all climates, it is the sediment supply relative to transport capacity that determines river morphology (Vandenberghe, 2003; Figure 2). Vandenberghe and Woo (2002) listed braided proglacial sandar, braided rivers, meandering rivers, transitional braided-meandering, anastomosing rivers, and wetland channels. Lunt et al. (2004) presented a depositional model for a braided-anastomosing river. These different types of channel may coexist within the same drainage basin (Figure 1). Between wet and arid cold-climate conditions, a range of fluvial planforms is seen which is similar to that of warmer climates, with braided rivers predominantly in arid conditions. The distinction between braided and meandering rivers is a function of stream power, bank strength, and grain size. With the exception of bank strength, which is influenced by

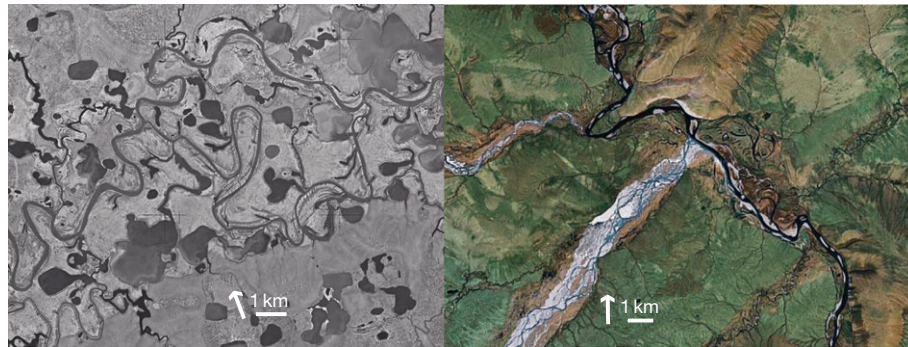


Figure 1 Left: Multithreaded meandering channels of the Berelegh river, Indigirka lowlands, Northeastern Siberia. Keyhole image 1977, EROS Data Center, USGS. The area shows many thaw (thermokarst) lakes. Several of the floodplain lakes are thaw lakes that have become connected to the river channel; not connected (closed) thaw lakes occur to the north and south of the floodplain. Right: Braided, meandering, and meandering-anastomosed river channels in Eastern Siberia, approx. 68° 54' N, 14° 4' E. ©2010 Google™, ©2011 Geocentre Consulting, Image ©2011 DigitalGlobe, Image ©2011 TerraMetrics.

		No vegetation	Patchy vegetation	Continuous vegetation
Continuous to discontinuous permafrost	Low Stream power Sediment supply High	Braided	Braided → meander	Meandering
Discontinuous to sporadic permafrost and deep seasonal frost	Low Stream power Sediment supply High		Braided	Anabranching
		Anabranching		
		Braided ↔ Meandering		

Figure 2 Relation of river morphology to permafrost and vegetation characteristics. Modified from Vandenberghe J (2003) Climate forcing of fluvial system development: An evolution of ideas. *Quaternary Science Reviews* 22: 2053–2060.

permafrost (see section ‘Landforms and Processes’), these have no unique link to periglacial conditions.

Present-day anastomosed rivers are not often reported in the literature, but fine examples do exist in periglacial regions (see [Figure 1](#)). A study of the Lena basin rivers showed the importance of log jams and ice jams ([Gautier and Costard, 2000](#)). The annually active main channels occupy a stable position, and avulsion activity is clearly recognizable. Reworking of aeolian material is prominent. A number of delta plains occur on the Arctic coast, for example, the Lena, Mackenzie, and Colville. Examples of Arctic delta morphology and deposits have been described by [Walker et al. \(1987\)](#) and [Hill et al. \(2001\)](#).

Vegetation is a climate-controlled factor, and several authors have attributed a significant role to vegetation in the interpretation of past fluvial succession changes ([Huisink, 1997](#); [Van Huissteden et al., 2001](#); [Figure 2](#)). Shifts in vegetation occur at both warm–cold and humid–arid climate transitions. On the one hand, vegetation determines, sediment delivery – by controlling slope and bank stability – and on the other, it determines discharge – through interception and evapotranspiration. From a theoretical model tested against field data, [Millar \(2000\)](#) concluded that bank vegetation, expressed as a friction angle, exerts a significant and quantifiable control on alluvial channel patterns. [French \(2007\)](#) listed

the availability of coarse-grained bed material in combination with discharge regime as a decisive factor for the presence of braided channels in the north Canadian archipelago. Here, braided rivers may develop into ‘periglacial sandar,’ rivers with all the characteristics of proglacial sandar, but without a glacial sediment source.

Hydrology

The hydrology of periglacial rivers is strongly influenced by the cold climate. There is much uncertainty about values of precipitation and evaporation in the Arctic as the observation network is sparse and the instrumentation for quantifying precipitation and evaporation is subject to uncertainty, particularly in cold climates. Net annual precipitation in the present-day Arctic is typically 150–300 mm, but may reach 1000 mm in the humid North Atlantic sector. Although precipitation over land areas peaks in summer, evapotranspiration in summer is also high. The net precipitation surplus is therefore small or even negative over many Arctic drainage basins; typical values for the precipitation surplus are in the order of 0–200 mm. For the large Eurasian and Canadian river basins (Ob, Yenesei, Lena, and Mackenzie), runoff ranges between 182 mm (Lena) and 396 mm (Ob) basin, while the runoff–

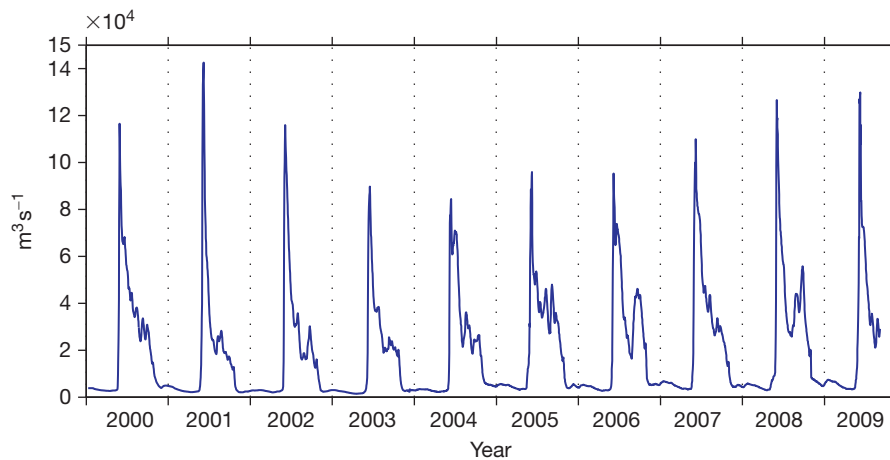


Figure 3 Discharge of the Lena river at Kusur, close to the river delta between 2000 and autumn 2009. Data from Russia Arctic Net/Roshydromet, <http://www.russia-arcticnet.sr.unh.edu/index.html>. The river shows a pronounced spring snowmelt discharge peak.

precipitation ratio ranges between 0.45 and 0.79 (Serreze and Barry, 2005).

The storage of precipitation in the snowpack during a large part of the year and its sudden release in spring strongly influences the annual distribution of the discharge. Moreover, in spring, infiltration into the still frozen ground is restricted (Clark, 1988; Shanley and Chalmers, 1999). Characteristically, the rivers show a rapid increase of discharge in May and June, culminating in a discharge peak in June (Figure 3). The snowmelt runoff takes most of the yearly precipitation, except in wetland basins; the runoff-precipitation ratio may be as high as 0.7–0.8 (Woo, 1990). The importance of snowmelt runoff decreases southward. Concentration of discharge can be expected to occur in regions where snow cover persists for longer than 1 month in 50% of the years (Church, 1988). This includes regions outside the periglacial zone. The magnitude of spring breakup floods depends strongly on weather conditions. With slow melting of the snow in the drainage basin, discharge increase will be less prominent. Usually, however, rainstorms accompany snowmelt and augment discharge. Discharge during snowmelt may display a distinct diurnal pattern (McCann et al., 1972).

Church (1988) distinguished four types of Arctic and subarctic discharge regimes: the subarctic nival regime, the Arctic nival regime, the proglacial regime, and the muskeg (wetland) regime. The subarctic and Arctic nival discharge regimes are characterized by the presence of a pronounced spring flood and decreasing flow in summer. The spring floods may be exacerbated by ice jams (Livingston et al., 2009). In high Arctic watersheds underlain by continuous permafrost, groundwater discharge is virtually absent and rivers are dry in winter. In subarctic river basins, groundwater contributes to the river discharge, in particular in winter, and summer rainstorms may produce flood peaks of higher magnitude than the spring flood. Proglacial rivers derive much of their discharge from melting glacier ice or extensive ice fields and snow banks; the discharge typically continues to rise throughout the summer and is highest in August when glacier ablation is most pronounced. The muskeg discharge regime occurs in poorly drained basins and wetland areas. In summer, high water

retention in lakes, wetland soil, and vegetation contributes to a slow response of the river to rainstorms (Roulet and Woo, 1988).

There are many catchments that do not fit into these four hydrographic categories (Clark, 1988). Superimposed on this general pattern are modifications by local climate and other drainage basin characteristics. Vegetation influences response time of rivers (McCann et al., 1972). In forested regions, evapotranspiration takes a larger part of the summer precipitation. The presence or absence of permafrost is also an important factor. Groundwater contribution ranges from practically zero in continuous permafrost areas to 40% in the discontinuous permafrost zone (Mackay and Løken, 1974). In continuous permafrost areas of the Arctic, only the larger rivers sustain significant winter discharge, particularly if they originate in more southerly areas or lakes (Figure 3). The presence of permafrost also promotes quick response of rivers to storm events and snowmelt (McCann et al., 1972). However, shallow subsurface runoff may also be dominant in peat-covered landscapes as the very conductive surface peat layers offer little attenuation to water drainage (Quinton and Marsh, 1999).

For many rivers in periglacial regions, the nival flood is the major sediment-transporting event (Church, 1988). Pronounced snowmelt discharge causes river channels to have a higher capacity than other rivers with comparable values of annual discharge. This will be expressed in channel dimensions: width, depth, meander wavelength, and gradient. Paleo-discharge estimates of rivers in periglacial regions therefore characterize mainly the nival flood discharge during spring.

Sediment

Sediment delivery to cold-climate rivers is often considered to be high as a result of rapid physical rock weathering, intense slope processes due to the presence of permafrost, and limited vegetation cover in the more arid cold environments. In addition, permafrost degradation may destabilize slopes and increase sediment supply greatly (Bowden et al., 2008). Even in lowlands, mass wasting may be intense. Rapid mass wasting

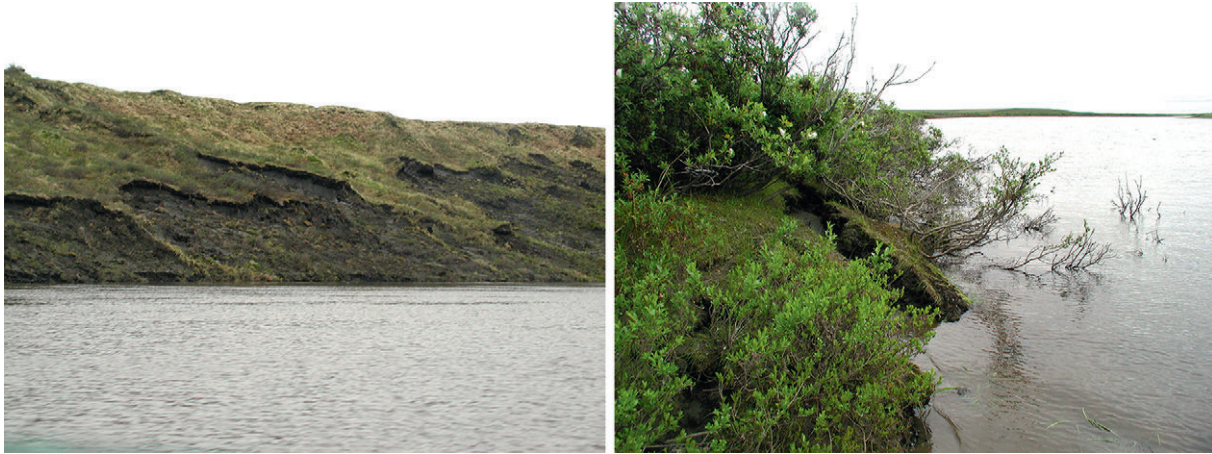


Figure 4 Left: Active-layer detachment slides on hillslope, Berelegh river, Indigirka lowlands, northeast Siberia. Right: Block slumping along the river bank of the Berelegh river caused by thermal erosion. Photographs by J. van Huissteden.

includes active-layer detachment slides, and thaw slumping (Lewkowicz, 1988; Figure 4). Slow mass-wasting processes include frost creep, gelifluction, and slope wash. Frost creep probably reaches its highest movement rate in the subarctic and higher middle latitudes because it depends on freeze-thaw alternations (Lewkowicz, 1988). The factors that influence gelifluction rate are soil moisture, vegetation cover, slope gradient, and soil grain-size distribution. A well-developed vegetation cover retards gelifluction, while high soil moisture by sufficient precipitation or melt of ground ice enhances it. Materials with a high silt content are most sensitive. Slope wash varies, depending on vegetation, material, and slope hydrology. The concentration of precipitation and runoff in a short time of the year enhances erosion in areas of sparse vegetation (French, 2007; McCann et al., 1972). The presence of large amounts of unconsolidated glacial sediments in catchments that were glaciated during the Pleistocene also enhances sediment supply considerably. Such landscapes in transition are known as ‘paraglacial’ (Church and Ryder, 1972; Slaymaker, 2009).

By contrast, in unglaciated periglacial river basins, measurements of sediment yield do not indicate exceptionally high values. French (2007) summarized total sediment yield from a number of larger periglacial catchments; these range between 0.8 and 85 tonnes $\text{km}^{-2} \text{year}^{-1}$, with partly glaciated Baffin Island as an outlier with 400–790 tonnes $\text{km}^{-2} \text{year}^{-1}$. Data compiled on suspended sediment yield by Bogen and Bønsnens (2003) range between 12 and 82.5 tonnes $\text{km}^{-2} \text{year}^{-1}$ for high Arctic nonglaciated catchments in Svalbard and Norway; higher values up to 200 tonnes $\text{km}^{-2} \text{year}^{-1}$ were measured in Iceland. By comparison, the North American average is 130 tonnes $\text{km}^{-2} \text{year}^{-1}$. Apparently, unglaciated periglacial catchments are modest sediment sources, despite specific factors that potentially contribute to a high sediment yield.

Glaciated catchments, however, produce more sediment. Bogen and Bønsnens (2003) compared an unglaciated and two glaciated catchments in Svalbard, reporting annual suspended sediment yields of 82.5, 359, and 586 tonnes $\text{km}^{-2} \text{year}^{-1}$, respectively. Likewise, sediment transport peak values (5.3, 11.4, and 9.0 g l^{-1}) and averages (0.16, 0.31, and 0.55 g l^{-1}) varied strongly between the catchments. For the unglaciated

catchment, the sediment was also coarser (0.040 mm vs. 0.014 and 0.021 mm). Within glaciated catchments, orders of magnitude differences in sediment yield relate to the thermal regime of the glaciers (cold vs. temperate glaciers) and bedrock erosion susceptibility. Cold-based glaciers (with frozen bed) have lower sediment yields than temperate glaciers (Bogen and Bønsnens, 2003).

Sediment discharge is often largely restricted to the nival discharge peak (e.g., Forbes and Lamoureux, 2005). Observations on the percentages of suspended and bedload sediment transport are limited and point to high interannual variation. Solute transport is relatively minor (Woo and McCann, 1994). Bedload accounts for 77–90% of the annual load of Baffin Island rivers (Church and Gilbert, 1975). Next to the transport processes, the catchment substratum obviously determines these percentages. However, rivers that are fed largely by groundwater (e.g., wetland rivers outside regions with continuous permafrost) tend to have a sediment concentration that is an order of magnitude lower than rivers fed by surface runoff (Woo and McCann, 1994).

Aeolian activity and reworking of wind-blown sediment can also supply sediment to rivers in periglacial regions (e.g., Bryant, 1983). Van Huissteden (1990) and Mol et al. (2000) demonstrated this for Pleistocene rivers in Europe. Periglacial river systems can also act as source areas for loess, which may influence ecosystems distant from the source river (Walker and Everett, 1991). For instance, the widespread occurrence of loess in ‘yedoma’ or ice-complex deposits in eastern Siberia consists largely of floodplain-derived loess.

Boreal and Arctic soils are estimated to store the equivalent of twice the amount of carbon stored in the atmosphere (Schuur et al., 2009). Periglacial floodplains are of specific importance because of their net primary productivity of the vegetation and high carbon sequestration rates in sediments as peat or organic-rich silt and clay (e.g., Van Huissteden, 1990; Zimov et al., 2006). However, emission of carbon and greenhouse gases may also be high. Van Huissteden et al. (2005) reported methane emissions from floodplain wetlands which are of similar magnitude as those of thaw lakes in northern Siberia. In addition, fluvial erosion may remobilize stored carbon.

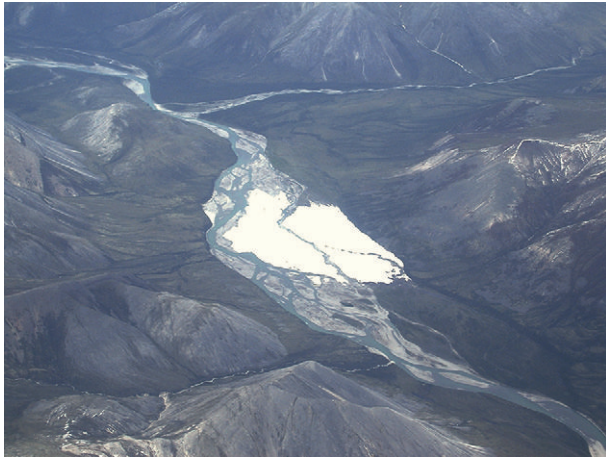


Figure 5 River icing, Werchojansk mountains, Eastern Siberia. The river itself is an example of a periglacial sandur. Photograph by J. van Huissteden.

Ice

Freezing of rivers during winter raises the water level because of the increased flow resistance created by the ice cover. Near channel obstructions, where flowing water is forced above the ice cover and freezes, river icings or icing mounds (naleds) can develop. These obstructions are most frequent in shallow braided channels (French, 2007). Icings can survive much of the summer melting (Figure 5). In channels with bottom-fast ice, the river beds and banks are protected against scouring during the spring breakup flood as most of the water will initially flow over the ice. By contrast, scouring tends to be more pronounced beneath floating ice, exacerbated by ice floes. The transition to floating ice is associated with a stepwise increase in channel geometry on the Alaskan North Slope (Best et al., 2005).

During spring breakup, ice dams may form, particularly in Arctic rivers with headwaters in the south, where ice melting starts earlier. Ice jams raise flood height of the spring flood and may cause considerable damage (Church, 1988); they are not restricted to rivers in periglacial regions. The high flow velocities, high stage, and mechanical action of moving ice make breakup an important agent of change for river channels, banks, and riparian vegetation (Prouse and Culp, 2003). Near the Yukon River, ice-jam flood deposits can be recognized in its overbank deposits, by the presence of isolated clasts resulting from ice rafting in fine-grained deposits. The deposits indicate a recurrence interval of 25–38 years (Livingston et al., 2009). In Arctic nival rivers, channels become filled with drift snow throughout the winter, which retards flow until the snow becomes saturated with water (Church, 1988).

Permafrost

In areas with ice-rich permafrost, river and floodplain morphology is strongly influenced by permafrost degradation. In smaller headwater streams, 'beaded drainage' is common and is probably the only typical periglacial example of channel planform (Figure 6). The channels exhibit a pattern of small



Figure 6 Beaded drainage channel in a small lowland headwater channel near Kytalyk research station, Indigirka lowlands, Northeast Siberia. The scale bar is 100 m. GeoEye Image, August 2010, ©2008 by GeoEye.

ponds, connected by the channel. The ponds result from thawing where channels cross intersecting ice wedges (French, 2007).

For larger rivers, permafrost also influences river-bank stability. Freezing of unconsolidated river bank and bed material raises its shear strength. Before bank erosion can occur, ground ice in river banks needs to melt, but it is uncertain to what extent this affects channel migration (Clark, 1988). A classic study by Scott (1978) shows a dual effect. During spring breakup, still frozen banks are protected from erosion, but subsequent thawing saturates the banks with water, resulting in destabilization and stronger erosion (see also Figure 4). This may cause high bank erosion rates (Costard et al., 2003). The river water tends to undercut banks by creating thermo-erosional niches, which sometimes can penetrate for several meters below overhanging frozen bank material (Walker et al., 1987). Block slumping is favored by this type of bank undercutting and is a very common feature along lowland rivers in permafrost terrain. On higher river banks, active-layer detachment slides may contribute extra sediment to the river (Figure 4). Higher river banks in unconsolidated frozen sediment tend to show a typical 'baydzherakh' relief, consisting of small hills created by preferential thaw along ice wedges.

The distinction between thaw lakes and floodbasin lakes is not sharp, as floodplain lakes may also show features of expansion by permafrost thaw (**Figure 1**). For example, the majority of lakes in the Mackenzie Delta appear to have formed as thaw lakes and they are partly maintained by permafrost thaw. The lakes become filled in with sediment through over-bank flooding of the river. When the lake is tapped by a river channel, the sedimentation rate increases and the fill may have a delta-like appearance, connected to levees (**Hill et al., 2001**). **Van Huissteden (1990)** documented thaw-lake infilling with fluvial sediment in a Pleistocene succession.

Rivers and lakes in permafrost regions are often underlain by taliks (**French, 2007**). These unfrozen layers or bodies within a permafrost area result from heat transfer from river and lake water to the underlying permafrost; in areas of discontinuous permafrost, the taliks are likely to reach the base of the permafrost. How large the contribution of groundwater flow to the river discharge is, depends on the size and inter-connectedness of the taliks.

Pleistocene Periglacial Environments

Rivers in periglacial environments can be extremely varied, with braided, meandering, straight, and anastomosing channel systems (**Figure 1**). This is also true of cold-climate river sediments in Pleistocene periglacial environments. Additionally there may be episodes of river incision within cold-climate phases. There are currently differences of opinion on the timing of such episodes with respect to climate change during glacial–interglacial cycles.

In sedimentary successions, diagnostic criteria for identifying ‘periglacial’ or ‘cold-climate’ river deposits and forms are not simple, and periglacial conditions during deposition must be deduced from accessory sedimentary features. These are generally periglacial deformation structures, for example, ice-wedge pseudomorphs and cryoturbations. The interpretation of these, however, may be ambiguous, and it is necessary to demonstrate that these features are syndepositional rather than postdepositional. More rarely, features such as preserved thermoerosional niches, frigites (blocks of unconsolidated sediments that have been transported in a frozen condition), or ice-rafted material may be found (**Van Huissteden, 1990**).

On a larger scale, lithology and inferred river-channel plan-form have been used to distinguish Pleistocene periglacial rivers, leading to circular reasoning. Generally, the coarser-grained, supposedly braided-river sediments and fluvial aggradation were assumed to represent deposition during cold climates, while meandering rivers and incision were restricted to warmer climates. This can be traced back to the climatic geomorphology of **Büdel (1977)**. Recent studies in fluvial stratigraphy also refer to this model, to explain the formation of Pleistocene terrace staircases (e.g., **Bridgland and Westaway (2008)** and references therein). However, the link between coarse-grained sediments and cold conditions has been falsified on many occasions (**Vandenberghe, 2003**; see below). In addition, fine-grained deposits occur abundantly in present-day floodplains in periglacial regions (e.g., **Figure 1**, with deposits consisting of silt and fine sand only). Moreover,

climate-controlled terrace successions also occur widely beyond the Pleistocene periglacial zone.

Since the beginning of the 1990s, new conceptual models have contested those simple relationships. The models have focused more on the changing processes at climate transitions, resulting in particular in short episodes of highly increased discharge–sediment ratios caused by the delayed response of vegetation to temperature changes (**Gibbard and Lewin, 2002**; **Vandenberghe, 1995**). The preservation potential of former floodplains, however, seems to be lower at warm-to-cold transitions than at cold-to-warm ones (**Vandenberghe, 2008**). This may explain the apparent rarity of interglacial terraces in Pleistocene terrace staircases and thus the equal number of terraces and glacial–interglacial cycles (and not a twofold higher number of terraces). Moreover, fluvial system changes may depend on the time scale. Many authors have assumed that climate change translates itself directly into fluvial depositional changes. However, **Van Huissteden et al. (2001)** have shown that this is not always the case, particularly on the time scale of the millennial climate changes that occurred during glacial climates of the Quaternary.

River sedimentation in cold conditions is believed to be especially sensitive in regions where rivers have been close to metastable equilibrium or thresholds (**Kasse et al., 2003**). Expression of river activity in sedimentation is also related to local conditions. **Rose and Boardman (1983)** emphasized the importance of relief contrast and the rate of sediment supply to the channel. Properties such as vegetation cover, particle size of the transported material, and gradient were mentioned by **Vandenberghe (1995)**, **Kasse (1998)**, and **Mol et al. (2000)**. Consequently, delayed responses to climatic conditions may provoke varying fluvial processes, sediments, and patterns depending on these local conditions. Lastly, rapid climate change may have caused conditions for which a present-day analog is absent. For instance, the frequent and large amplitude climate variations of Marine Isotope Stage (MIS) 3 contrast with the relatively stable Holocene. **Van Huissteden and Kasse (2001)** attributed rapid fluvial basin filling during the Middle Weichselian in Western Europe to frequent cycles of permafrost aggradation and degradation related to the MIS interstadials.

Two examples illustrate the variety of Pleistocene cold-climate fluvial environments. First, in southern Britain, lowland river terrace sediments predominantly consist of waterlain, sorted gravel and sand (**Figure 7**) occurring in suites of sedimentary structures that indicate deposition under broadly similar fluvial conditions. The most common facies is massive or horizontally crudely bedded gravel (**Figure 8(a)**). The clasts have a great size range, their form determined by local source lithologies, but the most frequent are medium to fine gravel (1–30 cm). Larger clasts do occur and may reach over 50 cm, but these are rare. Periglacial conditions are indicated by cryoturbation structures, ice-wedge pseudomorphs, physically weathered materials, and (rarely) frigites.

The gravels often show poorly developed horizontal bedding and are generally matrix supported and matrix rich. Clast-supported gravel is common, particularly in the coarser particle sizes. Interstices are filled with silt, sand, and granules. Secondary clay may also be present. Individual gravel units are typically up to ca. 90–125 cm thick but are frequently superimposed to produce multistorey units of greater thickness. Laterally, the

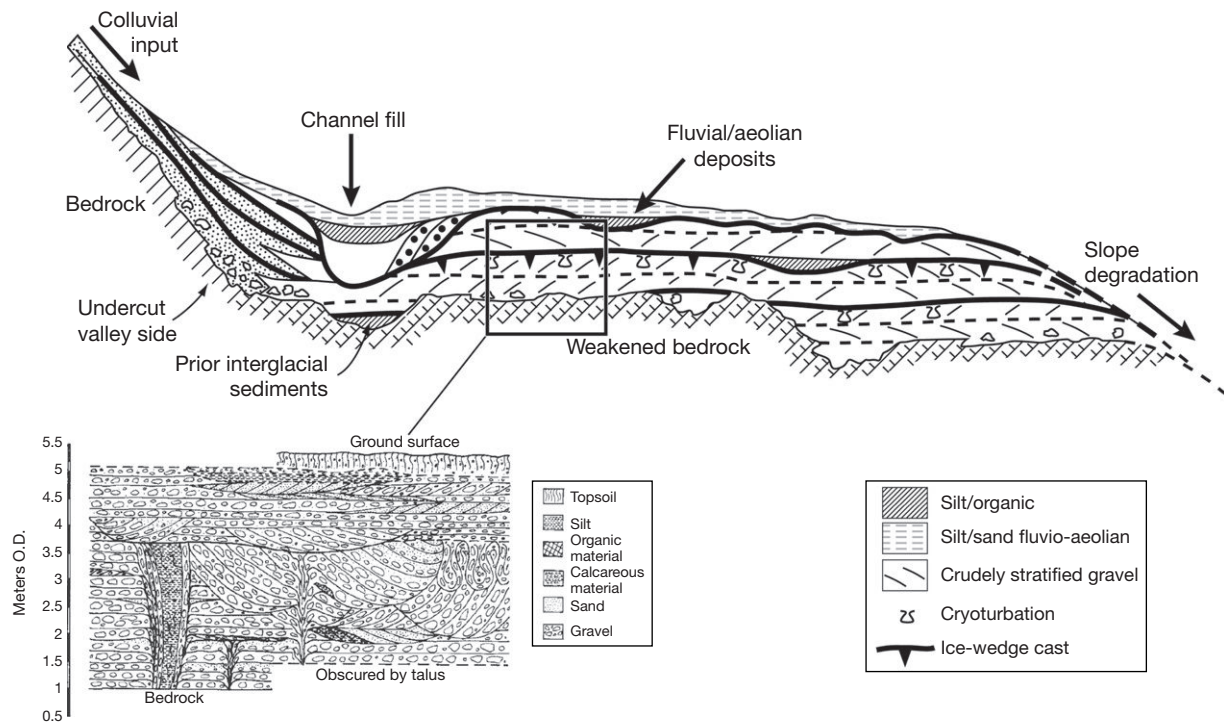


Figure 7 Schematic section of coarse-grained terrace sediments in lowland Britain. Single- or multistorey, matrix-rich gravels rest on an erosional substrate base. These may be capped by colluvial and aeolian sediments. Inset section showing sedimentary structures and periglacial features is based on [Briant et al. \(2004\)](#).

gravel units may be persistent and can be traced across exposures for over 30 m. They usually rest on an erosional base and often occupy channels cut into the underlying substrate.

'Channel' scours with erosional bases within gravels may be filled by interbedded sands and gravels at a range of scales. Some probably represent the truncated lower levels of channel fills abandoned by switching, other smaller ones to scouring related to migrating bedforms. Sandy facies occur as lenses and occasionally as massive sand beds. Finer materials occur infrequently as beds of silty clay, clayey silt, and organic sediment, either in channel fills or in former floodplain depressions. Organic materials include both autochthonous aquatic plants and animal fossils and allochthonous material washed in during flood events from vegetated stable tracts and neighboring slopes. Occasionally there are vertically accreting fining-upwards sequences.

Adjacent to contemporaneous valley sides, lenticular wedge- or tongue-shaped bodies of locally derived diamicton are often found, both interstratified with and overlying fluvial sediments. These indicate a dominance of catchment-wide slope-sediment contributions which may contrast to point-sourced glacier-melt sediment inputs into ice-marginal river systems. Rivers in former periglacial environments commonly only occupied part of the valley bottom at any particular time, and large areas of alluvial plains became temporarily stable and colonized by vegetation. During exposure, these surfaces experienced subaerial weathering, and ice-wedge polygons, and cryoturbation developed in favorable localities, particularly on higher stable tracts. Infill and truncation of ice-wedge casts, together with disturbance, truncation, and burial of valley-side mass-flow deposits are found, in combination

with abandoned channel fills. Extensive planar erosional surfaces have been attributed to truncation of fluvially inactive areas, raised slightly above the active areas of the river by large-scale channel migration. The association of truncated infilled ice-wedge casts and other periglacial phenomena, channel fills, and earlier sediments beneath such surfaces could be interpreted as a normal feature of cold-climate autogenic reworking ([Figure 8\(a\)](#) and [8\(b\)](#)).

The assemblage of sedimentary facies and structures present in British Pleistocene river terraces closely approximates those of braided-river environments (e.g., [Bryant, 1983](#)). This is also true of extensive areas of central and eastern Europe. The consistent association with climatic indicators such as ice-wedge casts, solifluction deposits, frozen block transport, and cryoturbation structures suggests deposition under a predominantly periglacial regime.

A second example consists of river sediments in European lowland basins, for example, in the Netherlands, northern Germany, and Poland ([Figure 9](#)). These sediments consist of alternating beds of coarse sand with varying amounts of gravel, fine sand, silt and clay, and organic beds (peat and gyttja). Usually, the beds have been periglacially disturbed and show a large range of cryogenic structures, including ice-wedge casts, frost fissures, and cryoturbations. The fine-grained beds indicate alluvial plains with well-developed marshy flood basins. These flood basins may have been occupied by shallow lakes for much of the year. These lakes may have developed partly as thaw lakes, as in many contemporary Arctic environments.

In addition, sand was deposited in overbank environments. Sedimentary structures in many sand beds indicate deposition

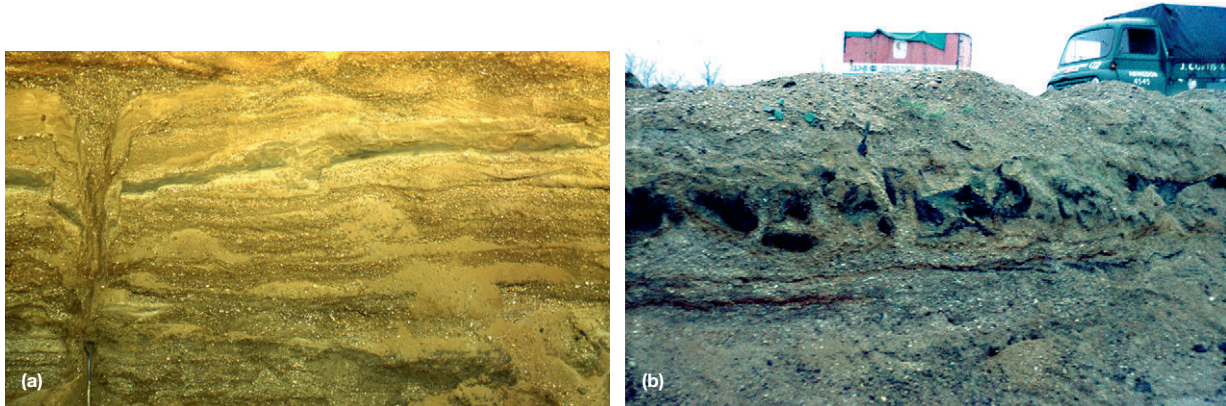


Figure 8 (a) Crudely stratified gravel at Block Fen, Cambridgeshire, England; prominent ice-wedge cast at left side. (b) Cryoturbated gravel and fine organic sediment at Thrupp House Farm, near Abingdon, Berkshire, England. Photographs by P.L. Gibbard.

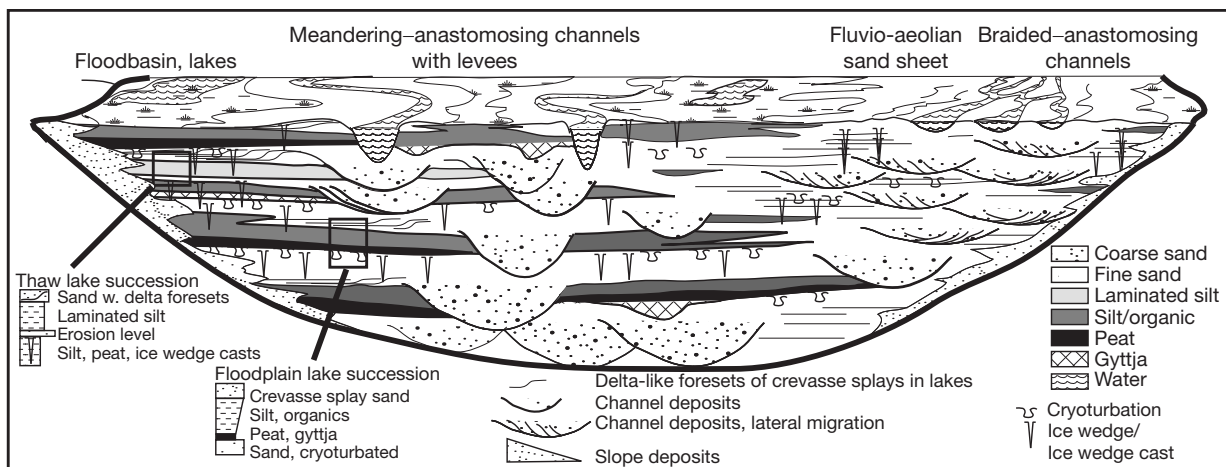


Figure 9 Schematic section and block diagram of fine-grained fluvial deposits of Middle Weichselian age in northwest European lowlands, based on the Dinkel river valley fill in the eastern Netherlands (Van Huissteden, 1990). The valley fill shows two distinct fluvial facies, occurring contemporaneously: anastomosing-meandering rivers with extensive floodbasin deposition (left) and anastomosing-braided rivers with fluvio-aeolian overbank deposition (right). The coarsest material in the deposits is coarse sand with minor amounts of gravel.

in levees and extensive crevasse-splay systems that prograded to shallow lakes or marshy floodbasins. Periglacial features including ice-wedge casts and involutions occur abundantly at the top of these sand bodies, underlying the silts and peats, which suggests that permafrost degradation may have created the depressions in which the fines were deposited; lacustrine sediments (gyttja or laminated silts) are common. Channel deposits form a minor component of alluvial sediment assemblages, suggesting minor lateral channel migration. Channels were stabilized by abundant riparian vegetation and cohesion of the fine-grained beds. Channel migration occurred mainly by avulsion (Figure 9). The large volume of overbank deposits, sedimentary structures, and geometry of the channel deposits suggest anastomosing patterns, with more than one active channel on the floodplain, separated by marshy flood basins (e.g., Mol, 1997; Van Huissteden, 1990).

Facies variations do occur. A lateral transition toward fluvio-aeolian sands has been described from the Dinkel valley

in the Netherlands (Van Huissteden, 1990; Figure 9). In this fluvio-aeolian facies, the silty overbank deposits have been partially reworked into aeolian sand-sheet deposits. The aeolian deposits in the fluvio-aeolian facies are assumed to have been derived from periodically dry channel beds and bars. In continental environments, fluvially deposited, but reworked aeolian materials may form a very important contribution to sediment assemblages.

Future Perspective

Future climate warming is expected to be strongest in the Arctic (ACIA, 2005). Consequently, many periglacial rivers today are systems in transition. Permafrost zones will shift northward, and warming will tend to increase active-layer thickness and talik extent, increasing groundwater contribution to the discharge (Bense et al., 2009; Woo and McCann,

1994). Second, expansion of thaw ponds and lakes will increase water storage in catchments. Modeled thaw-lake expansion in response to future warming indicates a threefold increase in lake area for northern Siberia in the next century, which is subsequently reduced by lake drainage (Van Huissteden et al., 2011).

Climate models also indicate a general increase of precipitation over the large Arctic drainage basins, resulting in an increased precipitation surplus as well (ACIA, 2005). This precipitation increase is concentrated in winter. For the summer, the tendency in precipitation surplus is less certain; particularly for evaporation, the models show contradictory results. The increased winter precipitation will induce higher spring discharge peaks and enhance permafrost degradation (ACIA, 2005). In the Eurasian Arctic, the annual discharge from the six largest rivers increased by 7% between 1936 and 1999 (Peterson et al., 2002).

Permafrost degradation, in turn, may increase slope instability, and sediment and nutrient delivery to rivers (Bowden et al., 2008). On the other hand, the extent of snow cover has decreased since approximately 1980. Climate models predict reductions between 9% and 18% by the end of the twenty-first century. These reductions are most prominent in the spring and autumn months, resulting in an earlier spring discharge peak and less prominent nival discharge. Spring breakup has shifted to earlier dates during the last century for many Arctic rivers (ACIA, 2005).

A further effect of Arctic warming is vegetational change. Throughout the Arctic, an expansion of shrub vegetation in tundra areas is reported (e.g., Tape et al., 2006). This affects not only hillslopes, but also river beds and banks; Tape et al. (2006) identified shrub expansion in braided-river beds. This may result in a complete transformation of rivers, by affecting bed morphology, bank stability, channel and floodplain roughness, slope stability, and sediment delivery, and by increasing evapotranspiration.

Woo and McCann (1994) stressed the complex response of (fluvial) environmental processes as a result of climate warming. Anisimov et al. (2008) reported model experiments on the effects of increasing discharge for rivers in northern Russia, resulting in a tendency toward multithreaded channels. However, the effects of other changing factors such as increasing vegetation result in large uncertainty. Apart from the development of simulation models, spatial and climatic analogs may be applied for predictions of future hydrological and fluvial consequences (Woo, 1990). However, caution must be exercised in the use of poleward shifting climatic zones to predict changes in present-day periglacial rivers because of nonanalog and transient effects such as delayed vegetation migration.

See also: Loess Deposits: Origins and Properties; Paleohydrology. **Dune Fields:** High Latitudes. **Fluvial Environments:** Responses to Rapid Environmental Change; Sediments; Terrace Sequences. **Permafrost and Periglacial Features:** Cryoturbation Structures; Ice Wedges and Ice-Wedge Casts; Paraglacial Geomorphology; Slope Deposits and Forms; Thermokarst Topography; Yedoma: Late Pleistocene Ice-Rich Syngenetic Permafrost of Beringia.

References

- ACIA (2005) *Arctic Climate Impact Assessment*. Cambridge: Cambridge University Press.
- Anisimov O, Vandenbergh J, Lobanov V, and Kondratiev A (2008) Predicting changes in alluvial channel patterns in North-European Russia under conditions of global warming. *Geomorphology* 98: 262–274.
- Bense VF, Ferguson G, and Kooi H (2009) Evolution of shallow groundwater flow systems in areas of degrading permafrost. *Geophysical Research Letters* 36: L22401. <http://dx.doi.org/10.1029/2009GL039225>.
- Best H, McNamara JP, and Liberty L (2005) Association of ice and river channel morphology determined using ground-penetrating radar in the Kuparuk River, Alaska. *Arctic, Antarctic, and Alpine Research* 37: 157–162.
- Bogen J and Bønsnes TE (2003) Erosion and sediment transport in High Arctic rivers, Svalbard. *Polar Research* 22: 175–189.
- Bowden WB, Gooseff MN, Balser A, Green A, Peterson BJ, and Bradford J (2008) Sediment and nutrient delivery from thermokarst features in the foothills of the North Slope, Alaska: Potential impacts on headwater stream ecosystems. *Journal of Geophysical Research* 113: G02026. <http://dx.doi.org/10.1029/2007JG000470>.
- Briant RM, Coope GR, Preece RC, et al. (2004) Fluvial system response to Late Devensian (Weichselian) aridity, Baston, Lincolnshire, England. *Journal of Quaternary Science* 19: 479–496.
- Bridgland D and Westaway R (2008) Climatically controlled river terrace staircases: A worldwide Quaternary phenomenon. *Geomorphology* 98: 285–315.
- Bryant ID (1983) The utilization of Arctic river analogue studies in the interpretation of periglacial river sediments from southern Britain. In: Gregory KJ (ed.) *Background to Palaeohydrology*, pp. 413–431. Chichester: Wiley.
- Büdel J (1977) *Klima-Geomorphologie*. Berlin: Gebrüder Bornträger.
- Church M (1988) Floods in cold climates. In: Baker VR, Kochel RC, and Patton PC (eds.) *Flood Geomorphology*, pp. 205–231. New York: Wiley.
- Church MA and Gilbert R (1975) Proglacial fluvial and lacustrine environments. In: Jopling AV and McDonald BC (eds.) *Glaciofluvial and Glaciolacustrine Sedimentation*, pp. 22–100. Tulsa, OK: Society of Economic Paleontologists and Mineralogists.
- Church MA and Ryder JM (1972) Paraglacial sedimentation: A consideration of fluvial processes conditioned by glaciation. *Geological Society of America Bulletin* 83: 3059–3072.
- Clark MJ (1988) Periglacial hydrology. In: Clark MJ (ed.) *Advances in Periglacial Geomorphology*, pp. 415–462. Chichester: Wiley.
- Costard F, Dupeyrat L, Gautier E, and Carey-Gailhardis E (2003) Fluvial thermal erosion investigations along a rapidly eroding river bank: Application to the Lena river (central Siberia). *Earth Surface Processes and Landforms* 28: 1349–1359.
- Forbes AC and Lamoureux SF (2005) Climatic controls on streamflow and suspended sediment transport in three large middle Arctic catchments, Boothia Peninsula, Nunavut, Canada. *Arctic, Antarctic, and Alpine Research* 37: 304–315.
- French HM (2007) *The Periglacial Environment*. New York: Wiley.
- Gautier E and Costard F (2000) Les systèmes fluviaux à chenaux anastomosés en milieu périglaciaire: la Léna et ses principaux affluents (Sibérie centrale). *Géographie physique et Quaternaire* 54: 327–342.
- Gibbard PL and Lewin J (2002) Climate and related controls on interglacial fluvial sedimentation in lowland Britain. *Sedimentary Geology* 151: 187–210.
- Hill PR, Lewis CP, Desmarais S, Kauppaymuthoo V, and Rais H (2001) The Mackenzie Delta: Sedimentary processes and facies of a high-latitude, fine-grained delta. *Sedimentology* 48: 1047–1078.
- Huisink M (1997) Late glacial sedimentological changes in a lowland river in response to climatic change: The Maas, southern Netherlands. *Journal of Quaternary Science* 12: 209–223.
- Kasse C (1998) Depositional model for cold-climate tundra rivers. In: Benito G, Baker VR, and Gregory KJ (eds.) *Palaeohydrology and Environmental Change*, pp. 83–97. Chichester: Wiley.
- Kasse C, Vandenbergh J, Van Huissteden J, Bohncke SJP, and Bos JAA (2003) Sensitivity of Weichselian fluvial systems to climate change (Nochten mine, eastern Germany). *Quaternary Science Reviews* 22: 2141–2156.
- Lewkowicz AG (1988) Slope processes. In: Clark MJ (ed.) *Advances in Periglacial Geomorphology*, pp. 325–368. Chichester: Wiley.
- Livingston JM, Smith DG, Froese DG, and Hugenholz CH (2009) Floodplain stratigraphy of the ice jam dominated middle Yukon River: A new approach to long-term flood frequency. *Hydrological Processes* 23: 357–371.
- Lunt IA, Bridge JS, and Tye RS (2004) A quantitative, three-dimensional depositional model of gravely braided rivers. *Sedimentology* 51: 377–414.

- Mackay DK and Løken OH (1974) Arctic hydrology. In: Ives JD and Barry RG (eds.) *Arctic and Alpine Environments*, pp. 111–132. London: Methuen & Co.
- McCann SB, Howarth PJ, and Cogley JG (1972) Fluvial processes in a periglacial environment. *Transactions of the Institute of British Geographers* 55: 69–82.
- Millar RG (2000) Influence of bank vegetation on alluvial channel patterns. *Water Resources Research* 36: 1109–1118.
- Mol J (1997) Fluvial response to Weichselian climate changes in the Niederlausitz (Germany). *Journal of Quaternary Science* 8: 15–30.
- Mol J, Vandenberghe J, and Kasse C (2000) River response to variations of periglacial climate in mid-latitude Europe. *Geomorphology* 3: 131–148.
- Peterson BJ, Holmes RM, McClelland JW, et al. (2002) Increasing river discharge to the Arctic Ocean. *Science* 298: 2171–2173.
- Prouse TD and Culp JM (2003) Ice breakup: A neglected factor in river ecology. *Canadian Journal of Civil Engineering* 30: 128–144.
- Quinton WL and Marsh P (1999) A conceptual framework for runoff generation in a permafrost environment. *Hydrological Processes* 13: 2563–2581.
- Rose J and Boardman J (1983) River activity in relation to short-term climatic deterioration. *Quaternary Studies in Poland* 4: 189–198.
- Roulet NJ and Woo MK (1988) Runoff generation in a low arctic drainage basin. *Journal of Hydrology* 101: 213–226.
- Schuur EAG, Vogel JG, Crummer KG, Lee H, Sickma JO, and Osterkamp TE (2009) The effect of permafrost thaw on old carbon release and net carbon exchange from tundra. *Nature* 459: 556–559. <http://dx.doi.org/10.1038/nature0803>.
- Scott KM (1978) Effects of permafrost on stream channel behavior in Arctic Alaska. *Geological Survey Professional Paper 1068*, US Government Printing Office, Washington, p. 19.
- Serreze MC and Barry RG (2005) *The Arctic Climate System*. Cambridge Atmospheric and Space Science Series. Cambridge: Cambridge University Press.
- Shanley JB and Chalmers A (1999) The effect of frozen soil on snowmelt runoff at Sleepers River, Vermont. *Hydrological Processes* 13: 1843–1857.
- Slaymaker O (2009) Proglacial, periglacial or paraglacial? *Geological Society Special Publications (London)* 320: 71–84. <http://dx.doi.org/10.1144/SP320.6>.
- Tape K, Sturm M, and Racine C (2006) The evidence for shrub expansion in Northern Alaska and the Pan-Arctic. *Global Change Biology* 12: 686–702. <http://dx.doi.org/10.1111/j.1365-2486.2006.01128.x>.
- Tarnocai C, Canadell JG, Schuur EAG, Kuhry P, Mazhitova G, and Zimov S (2009) Soil organic carbon pools in the northern circumpolar permafrost region. *Global Biogeochemical Cycles* 23: GB2023. <http://dx.doi.org/10.1029/2008GB003327>.
- Van Huissteden J (1990) Tundra rivers of the Last Glacial: Sedimentation and geomorphological processes during the Middle Pleniglacial in the Dinkel valley (eastern Netherlands). *Mededelingen Rijks Geologische Dienst* 44(3): 3–138.
- Van Huissteden J, Berrittella C, Parmentier FJW, Mi Y, Maximov TC, and Dolman AJ (2011) Methane emissions from permafrost thaw lakes limited by lake drainage. *Nature, Climate Change* 1: 119–123.
- Van Huissteden J, Gibbard PL, and Briant RM (2001) Periglacial fluvial systems in northwest Europe during marine isotope stages 4 and 3. *Quaternary International* 79: 75–88.
- Van Huissteden J and Kasse C (2001) Detection of rapid climate change in Last Glacial fluvial successions in the Netherlands. *Global and Planetary Change* 28: 319–339.
- Van Huissteden J, Maximov TC, Kononov AV, and Dolman AJ (2005) High methane flux from an Arctic floodplain (Indigirka lowlands, eastern Siberia). *Journal of Geophysical Research* 110: G02002. <http://dx.doi.org/10.1029/2005JG000010>.
- Vandenberghe J (1995) Timescales, climate and river development. *Quaternary Science Reviews* 14: 631–638.
- Vandenberghe J (2001) A typology of Pleistocene cold-based rivers. *Quaternary International* 79: 111–121.
- Vandenberghe J (2003) Climate forcing of fluvial system development: An evolution of ideas. *Quaternary Science Reviews* 22: 2053–2060.
- Vandenberghe J (2008) The fluvial cycle at cold–warm–cold transitions in lowland regions: A refinement of theory. *Geomorphology* 98: 275–284.
- Vandenberghe J and Woo M (2002) Modern and ancient periglacial river types. *Progress in Physical Geography* 26: 479–506.
- Walker J, Arnborg L, and Peippo J (1987) Riverbank erosion in the Colville delta, Alaska. *Geografiska Annaler* 69A: 61–70.
- Walker DA and Everett KR (1991) Loess ecosystems of Northern Alaska: Regional gradient and toposequence at Prudhoe Bay. *Ecological Monographs* 61: 437–464.
- Walsh JE, Anisimov O, Ove J, et al. (2005) Cryosphere and hydrology. In: Symon C (ed.) *Arctic Climate Impact Assessment, ACIA*, pp. 183–242. Cambridge: Cambridge University Press.
- Woo MK (1990) Consequences of climatic change for hydrology in permafrost zones. *Journal of Cold Regions Engineering* 4: 15–20.
- Woo MK and McCann SB (1994) Climatic variability, climatic change, runoff, and suspended sediment regimes in northern Canada. *Physical Geography* 15: 201–226.
- Zimov SA, Schuur EAG, and Chapin FS III (2006) Permafrost and the global carbon budget. *Science* 312: 1612–1613.

Rock Weathering

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Angular, shattered rock is abundant in many cold regions, where it forms a major source of gravel-size clasts to hillslopes, rivers and glaciers. The shattering has often been attributed to frost action, particularly *in situ* freezing and expansion of water trapped in spaces within rock. However, increasing evidence suggests that ice segregation is a major process in fracture of fine-grained and porous rock, and that other mechanical, chemical and biological weathering processes contribute significantly to rock weathering in periglacial and permafrost environments. This article focuses on rock weathering processes, particularly mechanical ones, about which there has been much uncertainty concerning the role of frost. It also summarizes data on weathering rates and products.

Mechanical Weathering Processes

Mechanical weathering processes in periglacial and permafrost regions result from freezing and thawing of water (frost weathering; reviewed by Matsuoka and Murton, 2008) and from changes in unfrozen water content or temperature. Two processes of frost weathering can be distinguished (volumetric expansion and ice segregation), as can two processes that may be well developed in periglacial regions but are not confined to them (hydration shattering and thermally induced stress).

Azonal processes such as salt weathering and pressure release are not discussed, although they may be widespread in polar deserts, maritime regions and recently deglaciated mountains.

Volumetric Expansion

Upon freezing, water expands in volume by 9%. If the water completely fills spaces in rock and freezes *in situ*, then, theoretically, at a temperature of $-22\text{ }^{\circ}\text{C}$ it can exert a maximum tensile stress of 207 MPa in a rock strong enough to withstand it, as the water becomes ice. This ice-induced stress is easily great enough to fracture any rock because the maximum tensile strength of rock is one or two orders of magnitude lower than 207 MPa. Furthermore, the other conditions of (1) a high degree of rock saturation to avoid compression of air, (2) rapid cooling rates to allow *in situ* freezing of water within a closed system, and (3) freezing to temperatures below $-5\text{ }^{\circ}\text{C}$, seldom coincide at sites supposed to favor frost weathering, such as bergschrunds, snow patches and rockwalls (McGreevy, 1981). Thus, the dominant role in frost weathering often attributed to volumetric expansion has been questioned (Walder and Hallet, 1985, 1986). This process, however, does appear to operate within a few centimeters of the surface of rocks whose pores or cracks are water-filled (or nearly so) and subject to rapid freezing, resulting in prying away of mineral grains and

rock flakes. At a larger, outcrop scale, volumetric expansion within existing water-filled joints in bedrock can widen joints and heave bedrock.

A second way in which volumetric expansion may break up rock is by hydrofracture. Just as the expansion of water freezing within saturated pores in sand or gravel expels water ahead of the freezing front, so will does freezing of a rock that is saturated (or nearly so) and contains relatively large pores (e.g., medium- to coarse-grained sandstones). Likewise, freezing of saturated rock with pores small enough to favor cryosuction and ice segregation can instead expel water if freezing is sufficiently rapid (Walder and Hallet, 1985, 1986). If the water cannot drain away as quickly as it is expelled, pore water pressure may rise sufficiently to deform or hydrofracture the rock. Therefore, intact rock may potentially hydrofracture during rapid inward freezing of boulders (causing the rock to burst; Lautridou, 1988) or, perhaps, during two-sided freezing of an active layer. Nonetheless, hydrofracture of rock during freezing remains to be investigated in detail, particularly its relationship with bedrock heave by volumetric expansion in water-filled joints in the active layer (Dyke, 1984).

At the scale of sand-sized particles, quartz disintegrates more readily than feldspar, contrary to its behavior in temperate and warm-climate conditions (Konishchev and Rogov, 1993). A substantial body of experimental data indicates that quartz sand subjected to numerous freeze–thaw cycles tends to break down more readily than feldspar and produce finer particles, within the 0.05- to 0.1-mm (silt) size fraction. The causes of breakdown, inferred in part using scanning electron microscopy, are attributed to (1) the wedging effect of ice forming in microcracks, which makes them widen and lengthen, and (2) freezing of water within gas–liquid inclusions.

Ice Segregation

A theoretical model of rock fracture by ice segregation attributes the fracture to progressive microcracking as a result of growing ice lenses (Walder and Hallet, 1985, 1986). Ice segregation occurs when liquid water migrates through a porous medium toward freezing sites where lenses of ice grow, segregated from adjacent mineral particles. The migration results from temperature-gradient-induced suction gradients in freezing or frozen ground. As porous soil or rock freezes and cools, a suction (dP_w) develops in the liquid water held in capillaries and adsorbed on the surfaces of mineral particles (Williams and Smith, 1989):

$$dP_w = -dTl/VT \quad [1]$$

where dT is the lowering of the freezing point, l is the latent heat of fusion, V is the specific volume of water, and T is the absolute temperature. Equation [1] indicates that a temperature drop of $1\text{ }^{\circ}\text{C}$ induces a ‘cryosuction’ of 1.2 MPa (12 atm).

Suction gradients drive liquid water toward freezing sites. The water nourishes ice lenses at sites where release of latent heat from freezing of soil water balances loss of soil heat toward the ground surface or permafrost table. While this balance persists, ice lenses continue to grow on the cold side of a partially frozen fringe, with the lenses aligned perpendicular to the direction of heat flow. However, when heat loss exceeds heat supply, the fringe cools, which promotes pore-ice formation. This in turn reduces the permeability of the fringe, effectively shutting off the water supply to the ice lens, and thus the fringe advances through the soil until a new thermal balance is reached between heat supply to and heat loss from the cold side of the fringe, at which point a new ice lens develops (Williams and Smith, 1989). In this way, the fringe episodically moves, producing a rhythmic banding of segregated ice and pore ice (Figure 1). Detailed discussions of the microphysics of the ice-segregation process and of the spacing and thickness of ice lenses are given by Rempel (2007, 2011).

There is increasing evidence that ice segregation is an important cause of fracture of moist, porous rocks (Matsuoka, 2001; Matsuoka and Murton, 2008), just like it is in the fissuring and heave of frost-susceptible soils. Unlike the stringent conditions needed for volumetric expansion to fracture natural rocks, the conditions needed for ice segregation (slow rates of freezing or sustained subzero temperatures in moist, porous fine-grained rock) are common in natural bedrock. Laboratory experiments have demonstrated that ice segregation can fracture fresh blocks of tuff and sandstone under constant temperatures and temperature gradients (Akagawa and Fukuda, 1991; Hallet et al., 1991), and simulation of natural uni- and bidirectional freezing regimes resulted in fracture of fine-grained, porous rock by ice-lens growth in near-surface permafrost and seasonal frost (Figure 1; Murton et al., 2001; 2006). The fractures and ice lenses resemble those in arctic permafrost and provide an explanation for some Ice Age weathering profiles in the mid-latitudes. Ice segregation and the resulting frost cracking may also contribute to a frost “buzzsaw” mechanism for erosion of some mountains, for example the eastern Southern Alps of New Zealand (Hales and Roering, 2009).

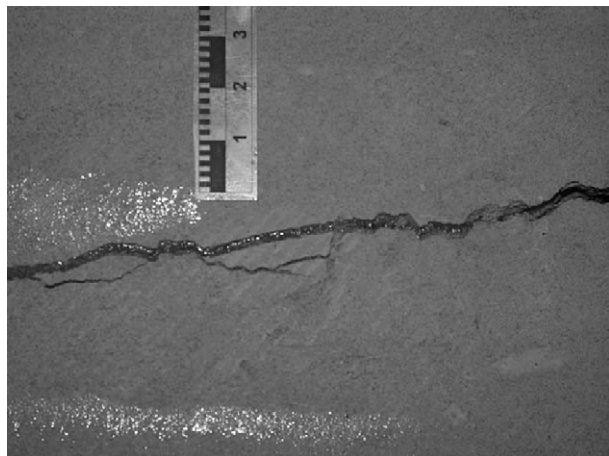


Figure 1 Segregated ice within artificial permafrost developed in a siliceous chalk (Tuffeau). Scale in centimeters. Satellite cracks have begun to develop beneath the main crack. The rock between the ice lenses contains pore ice.

Hydration Shattering

Hydration shattering results from the pressure exerted on the walls of pores with diameters of a few micrometers or less by water molecules adsorbed onto the surfaces of clay particles. Because water comprises polar molecules, the positive charge of the two hydrogen atoms ($2H^+$) at one end of the molecule is attracted to the negative charge of clay surfaces, leaving the negatively charged oxygen atom (O^{2-}) at the other end pointing away from the surfaces. Where the pores between clay particles are large ($>5\ \mu m$ diameter) and filled with ‘bulk’ (nonadsorbed) water, projecting oxygen atoms on opposite sides of the pore are too far apart for the similar negative charges to repel each other. However, where the pores are small ($<5\ \mu m$ diameter), they are dominated by adsorbed water, which is difficult to freeze, rigid, and capable of exerting force against pore walls. In such pores, adjacent negatively charged O^{2-} atoms repel each other. As the temperature falls below $4\ ^\circ C$, the adsorbed water expands; the degree of alignment (ordering) of water molecules increases (and hence the repulsive force between negative charges at the free ends of adsorbed layers); and the relative humidity increases, supplying water to the adsorbed water layer. Rock failure is attributed to fatigue due to expansion and contraction as water molecules are sorbed and desorbed on the adsorbed layer. Large temperature changes and cycling within the freezing and thawing range may promote sorption and desorption and, therefore, rock fracture (Hudec, 1973).

Laboratory experiments suggest that hydration shattering at subzero temperatures can be effective in breaking down clay-rich rocks (Dunn and Hudec, 1966). It is unclear, however, whether this process is effective in fracturing coarser-grained rocks or in producing angular fragments of sound rock that are commonly observed in alpine and polar regions (Washburn, 1979, p. 74).

Thermally Induced Stress

Rock has a tendency to expand on warming and to contract on cooling to an amount determined by the thermal coefficient of linear expansion of its constituent minerals. The coefficient denotes the relative change in length resulting from a $1\ ^\circ C$ temperature change—approximately $1\text{--}12 \times 10^{-6}$ length $^\circ C^{-1}$ for common rock-forming minerals. If we consider a body of intact rock that is laterally extensive and subjected to surface cooling, the rock surface is unable to contract freely in a horizontal direction because of (1) the physical constraints imposed by the lateral continuity of the rock surface and (2) the thermal stability of the underlying rock. Thus, the rock resists thermal contraction in a horizontal direction and, as a result, experiences tensile stress in this direction. The stress can be relieved by either fracture or creep of the rock—the same responses to tensile stress as those exhibited by winter cooling of frozen soil in permafrost regions.

Rock fracture may result in two situations, determined by the magnitude and frequency of thermally induced stresses. Repetitive, low-magnitude stress events, each one insufficient to cause immediate failure of the rock, may collectively cause failure over time (thermal stress fatigue), whereas a single, high-magnitude stress event generated by very rapid

temperature change ($>2^{\circ}\text{C min}^{-1}$) may trigger cracking immediately if tensile stress exceeds tensile strength (thermal shock). The temperature changes believed to favor rock fracture typify high-altitude and high-latitude regions, where large inputs of solar radiation may occur when air temperatures are below 0°C . During such times, rock surfaces often have above-zero temperatures until cloud blocks out solar radiation and the surface temperature plummets to many degrees below 0°C . Frequent and sometimes rapid changes in net radiation at rock surfaces produce frequent and rapid temperature changes as well as steep temperature gradients in the outermost centimeters of rock—conditions that favor thermally induced stress.

Centimeter-scale polygonal crack patterns observed in rocks of the High Andes and ice-free regions of Antarctica have been attributed to such stress (Hall, 1999), which seems reasonable in these extremely dry (ice-poor) environments. The efficacy of this weathering process, however, is probably limited to the outer millimeters or few centimeters of the rock surfaces because surface temperature waves are exponentially dampened with depth (Williams and Smith, 1989).

Chemical Weathering Processes

Although chemical weathering in periglacial regions has often been regarded as of minor significance (because of low temperatures), there is increasing evidence that contests this view (Pope et al., 1995). In alpine periglacial regions, dissolution of mineral grains commences soon after fresh bedrock is exposed to surface or near-surface environments (Dixon and Thorn, 2005). In the Colorado Rocky Mountains and the Jotunheimen mountains of southern Norway, for example, solutional processes promote disintegration of coarse debris. Dissolution of quartz, feldspars and micas within the debris results in an absolute loss of alkalis (Na and K), alkali earths (Ca and Mg), and silicon (Si) relative to resistant chemical species (Fe, Al and Ti). In Rapp's (1960) seminal study of mountain slopes in Karkevagge, arctic Sweden, solution was identified as the single most important transport mechanism. Recent work, summarized by Thorn et al. (2001), supports the importance of chemical weathering at this locality, emphasizing that pyrite-rich bedrock oxidizes to produce sulfuric acid, in turn favoring dissolution of muscovite and solution of calcite.

Oxidation of iron-bearing minerals produces reddish-colored weathering rinds on bedrock or clast surfaces. Even the dominantly mechanical process of *grus* formation on a nunatak in the Juneau Icefield, Alaska, is associated with measurable chemical weathering as biotite transforms to vermiculite, which is accompanied by a loss of K and Mg and by oxidation of Fe. From such studies, it is clear that the chemical weathering processes in alpine periglacial regions are similar to those in mid-latitude environments (Hall et al., 2002; Dixon and Thorn, 2005). This conclusion is consistent with the observation that the solute contents of alpine streams (ions dominated by Ca, Si and, sometimes, Mg, with smaller abundances of Na, K or Mg; anions typically dominated by bicarbonate and, in some locations, sulfate) are generally similar to those in other climatic regions. Solute concentrations in alpine streams, however, are typically lower than those in other environments, probably

because exposed bedrock rather than thick regolith dominates most alpine basins, thus limiting the opportunities for chemical weathering by slowly percolating groundwater.

Chemical weathering also takes place within sediment in arctic and sub-arctic active layers. As ice within active layers thaws in summer releasing water into the soil, weathering of labile minerals takes place above the water table. Oxidation of sulfides (e.g., pyrite) and solution of carbonates (e.g., siderite) has been inferred from studies of heavy minerals in paleo-active layers in Canada, with distinct mineralogical differences sometimes found between sediment that has thawed in the active layer and underlying sediment that has remained frozen within permafrost. Differences in clay mineralogy and soil color (the presence or absence of oxidation) have also been noted across some thaw unconformities (Burn et al., 1986), providing further evidence of chemical weathering in permafrost regions.

Biological Weathering Processes

There is a growing appreciation of the importance of biological and biochemical weathering in cold regions (Hall et al., 2002). These processes involve a range of rock-colonizing organisms (e.g., bacteria, fungi, lichens, mosses and trees) that exert physical stresses and/or induce chemical processes (e.g., chelation, oxidation and reduction). Tree roots, for example, can pry apart jointed rock on rocky mountain slopes, whereas micro-organisms living within or beneath rocks in polar deserts may cause both the removal and precipitation of minerals at or close to the rock surface. For example, lichens growing on and within rocks in the dry valleys of Southern Victoria Land, Antarctica (Friedmann, 1982), and the Jotunheimen mountains of southern Norway (McCarroll and Viles, 1995) cause exfoliation of the rock surface as a result of solubilization of cement and by mechanical processes involving penetration of hyphae into the rock and freeze-thaw-induced expansion and contraction of the thallus of endolithic lichens. Likewise, fungal growth can induce biochemical processes that appear to be the main agents of weathering rind development on boulders on Icelandic moraines (Etienne, 2002).

Weathering Rates

Weathering rates in periglacial regions are poorly known because weathering tends to be a slow, subsurface process and it is often difficult to measure fracture growth and to date weathered surfaces. Occasionally, explosive shattering or 'bursting' of boulders exposed to cold air temperatures in the arctic and sub-arctic has been inferred from descriptions of sounds resembling a rifle discharge and from angular rock fragments scattered over snow around a broken boulder. These observations suggest that such shattering results from an explosive release of tensional stresses, although whether these stresses result from hydrofracturing, thermal shock or ice segregation is often unclear (Mackay, 1999). By contrast, much slower rates of crack propagation seem to be characteristic of ice segregation and bedrock heave (Murton et al., 2001, 2006), consistent with slow migration of unfrozen water toward freezing sites (Walder and Hallet, 1985, 1986).

The rates of weathering rind formation in mid-latitude alpine environments appear to be comparable to, or greater than, those in non-alpine environments, with 6- to 45-mm thick weathering rinds forming in 10,000–20,000 years in the Colorado Alpine compared to rates of ~ 0.18 mm over the same period from non-Alpine California (Dixon and Thorn, 2005). In Lapland, weathering rinds ~ 1 -mm thick can develop in 5 years, indicating how rapidly chemical weathering commences. However, such growth rates may decay exponentially through time, making it difficult to extrapolate from short-term field experiments to long-term natural rates.

Chemical erosion (mass removal) and chemical denudation (surface lowering) rates in alpine settings are often substantial (Dixon and Thorn, 2005). In mid-latitude alpine periglacial environments of Scandinavia, North America and central Europe, rates of chemical erosion range from ~ 9 to $100 \text{ t km}^{-2} \text{ year}^{-1}$, comparable to the global average for all rock types ($45 \text{ t km}^{-2} \text{ year}^{-1}$) and to values from the Amazon and Mississippi rivers ($35\text{--}45 \text{ t km}^{-2} \text{ year}^{-1}$). Moreover, chemical erosion rates in alpine settings often substantially exceed mechanical erosion rates. Rates of chemical denudation measured in alpine Colorado and Scandinavia are $\sim 0.2\text{--}5$ mm per 1,000 years (higher in carbonates than crystalline rocks), similar to those in other montane and polar environments. In general, the denudation of alpine landscapes is strongly influenced by chemical processes because of their strong coupling to the drainage network, facilitating rapid and effective removal of solutes from the drainage basins (Dixon and Thorn, 2005).

Steep rocky slopes in periglacial regions undergo headwall retreat as a result of weathering and rockfall. Estimates of rock-wall retreat rates in Holocene and present-day alpine environments have an overall mean of 1.1 m ka^{-1} , which is distinctly greater than that of 0.3 m ka^{-1} from arctic environments (Table 1). As indicated by the large standard deviations in Table 1, the retreat rates are highly variable, in part due to differences in lithology, microclimate and measurement techniques. Although such rates must also vary through time, simple extrapolation of the overall means during the Holocene (past 11,500 years) indicates, to a first approximation, net rockwall retreats of ~ 3.5 m in arctic environments and ~ 12.5 m in alpine ones. The implications of such rates must be borne in mind before carrying out cosmogenic dating of such slopes.

Weathering Products

The products of periglacial weathering vary substantially in scale and nature. At the microscale are silt-sized particles,



Figure 2 Grus developed on granite, Mount Sedgwick, Barn Mountains, Northern Yukon, Canada. Ice axe for scale.

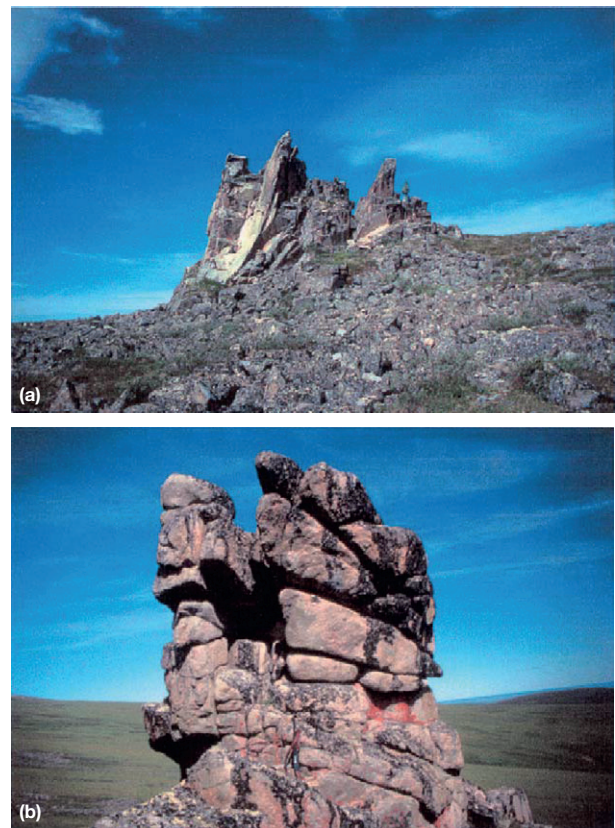


Figure 3 Tors varying from angular (a) to rounded (b) near Mount Sedgwick, Northern Yukon, Canada. Person for scale in (a), and shotgun in (b). Note blockfield surrounding tor in (a).

Table 1 Rockwall retreat rates in arctic and alpine environments^a

	Rockwall retreat rates (m ka^{-1})		
	Minimum	Mean	Maximum
Arctic mean	0.1(0.1, $n=13$) ^b	0.3(0.4, $n=7$)	1.0(1.2, $n=13$)
Alpine mean	0.7(0.8, $n=9$)	1.1(0.9, $n=10$)	2.1(1.4, $n=9$)

^aCalculated from summary data in Curry and Morris (2004, Table 6b).

^bValues in parentheses indicate standard deviation and number of observations.

often quartz, formed within cryogenic regoliths. Such particles probably constitute important sources of silt to periglacial rivers and eolian systems and, during the Pleistocene, to the ice-rich yedoma deposits of Beringia—a major archive of Quaternary environmental change. Within periglacial soils and regoliths, ferromagnesian minerals are often oxidized on grain surfaces and along cleavage planes, as is common in alpine regions (Dixon and Thorn, 2005). Disaggregated bedrock material, consisting of sand- to pebble-sized particles and

sometimes clay (grus), is common in periglacial areas of crystalline rocks. Although grus is attributed primarily to mechanical weathering, strong reddish brown colors indicative of oxidation are sometimes present (Figure 2).

Weathering rinds (~ 0.5 – 45 -mm thick) are often developed on bedrock or coarse debris. The rinds generally have a reddish-colored outer surface (due to oxidation of iron-bearing minerals) underlain by a lighter colored bleached zone. Some rinds have chemical rock coatings; in the Scandes Mountains of

Swedish and Norwegian Lapland, for example, such coatings are dominated by Fe, Si, Al, Ca and heavy metals.

Pebble- to boulder-sized rock fragments often litter hill-slopes in rocky periglacial regions, forming gravelly surfaces (Figure 2) and bouldery blockfields (Figure 3(a)). Many, but not all, of these fragments are angular. However, occasionally angular and rounded clasts occur side by side (Hall et al., 2002) as a result of lithological differences influencing the style of weathering (e.g., rock fracture versus granular

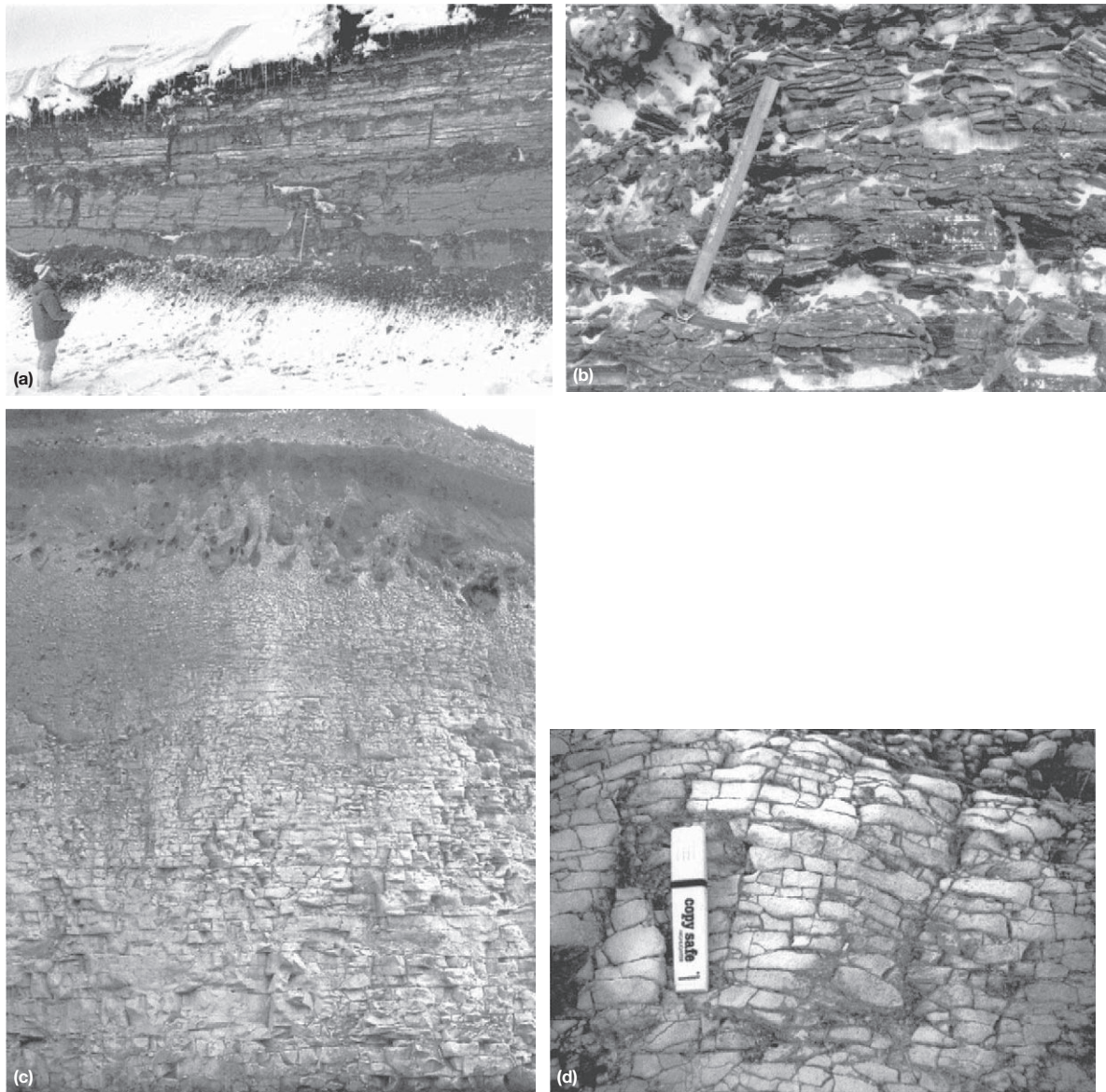


Figure 4 Weathering profiles formed by ice segregation within bedrock permafrost. The profiles comprise mainly horizontal to subhorizontal fractures that delimit angular, platy fragments of rock, with the overall fracture density decreasing with depth. (a) A profile in continuous permafrost developed in Lower Cretaceous shale on the Sabine Peninsula ($\sim 76^\circ$ N), eastern Melville Island, High Arctic Canada. Person for scale. A close-up view (b) shows that lenses of segregated ice (white) intersperse tabular rock fragments; ice axe for scale (reproduced from French HM, Bennett L, and Hayley DW (1986) Ground ice conditions near Rea Point and on Sabine Peninsula, eastern Melville Island. *Canadian Journal of Earth Sciences* 23: 1389–1400, with permission). (c) and (d) A relict weathering profile developed, probably in large part during Marine Isotope Stage 2, in Upper Cretaceous Chalk near Beresford Gap ($\sim 51^\circ$ N), Thanet, eastern Kent, UK. Vertical section in (c) is ~ 7 m high. Pen for scale in (d) (From Murton JB (1996) Near-surface brecciation of Chalk, Isle of Thanet, southeast England: A comparison with ice-rich brecciated bedrocks in Canada and Spitsbergen. *Permafrost and Periglacial Processes* 7: 153–164, reproduced with permission).

disintegration). The same variation in angularity of form appears to occur in weathering forms, which range from rounded pits, pans, grooves and tafoni to angular rockwalls. Even within the same small region, tors may vary in shape from rounded to angular (Figure 3).

Fractures in clasts and bedrock can result from a variety of mechanical weathering processes. Fractures formed by ice segregation are usually more or less parallel to the rock surface (i.e., cooling surface; Mackay, 1999), whereas those formed by hydrofracture or thermal shock may have a variable orientation, often perpendicular to the rock surface.

Weathering profiles comprising fractured (brecciated), porous bedrock can result from ice segregation within permafrost and seasonally frozen ground. The profiles are characterized by pebble- to cobble-sized rock fragments that are tabular, angular to subangular, aligned more or less parallel to the cooling surfaces (ground surface and permafrost table), and interspersed with ice lenses (Figure 4(a) and 4(b); French et al., 1986). The rock types include limestones, fine-grained sandstones and shales, all of which contain an interconnected network of fine pores through which unfrozen water can migrate toward the upper meters of permafrost. Relict profiles are common in limestones and mudstones in areas of former Pleistocene permafrost, such as southern England and northern France (Figure 4(c) and 4(d); Murton, 1996), and were probably a significant influence on periglacial landscape evolution, supplying rock fragments to hillslopes and rivers.

In the case of weathering profiles showing joints and fractures in low-porosity crystalline bedrock beneath autochthonous blockfields, ice segregation is unlikely. A more feasible process of fracture growth and clast supply is that of frost wedging. It is hypothesized that bedrock disintegration at the weathering front beneath certain autochthonous blockfields is mainly due to frost wedging that enlarges pre-existing stress-release joints at the base of a water-saturated active layer above permafrost (Ballantyne, 2010; Goodfellow et al., 2008). This hypothesis remains to be tested by field monitoring or laboratory experiments.

Conclusions

Of the mechanical weathering processes known to occur in periglacial and permafrost regions, only ice segregation is at present strongly implicated as a widespread and volumetrically important cause of rock fracture. Because fine-grained and porous bedrock is frost-susceptible, ice segregation is a major cause of both soil fissuring and bedrock fracture. Ice segregation is most common in the active layer and near-surface permafrost and in seasonally frozen ground in nonpermafrost regions; thus, the resulting fissures and fractures tend to be most common in the upper meters of soil and bedrock. Additional mechanical weathering processes certainly occur in some cold regions, but their extent and geomorphic significance remain to be clearly established.

It is clear that both chemical and biological weathering processes are widespread and important in many periglacial and permafrost regions. However, their main contribution to landform development, at least in alpine and Antarctic regions, seems limited to small-scale forms (e.g., pits, pans, grooves and

tafoni) and to microscale features, such as rinds and coatings. It is also clear that mechanical, chemical and/or biological processes in cold regions often operate concurrently at the weathering boundary layer (Pope et al., 1995). The combined effects of frost and salt near periglacial coasts, for example, can promote rock disintegration, whereas interactions of mechanical and chemical weathering processes often contribute to the formation of weathering rinds (Thorn et al., 2001). Even in the polar deserts of Antarctica, a variety of weathering processes interact (Guglielmin et al., 2005). Indeed, chemical weathering may occur wherever liquid water is present; thus, dissolution at the ends of microcracks (stress corrosion cracking) might enhance crack propagation during ice segregation. In turn, ice segregation prepares fresh mineral surfaces for chemical weathering (Hoch et al., 1999), indicating feedbacks between weathering processes.

See also: [Cosmogenic Nuclide Dating: Landscape Evolution.](#)
[Permafrost and Periglacial Features: Block/Rock Streams; Slope Deposits and Forms.](#)

References

- Akagawa S and Fukuda M (1991) Frost heave mechanism in welded tuff. *Permafrost and Periglacial Processes* 2: 301–309.
- Ballantyne CK (2010) A general model of autochthonous blockfield evolution. *Permafrost and Periglacial Processes* 21: 289–310.
- Burn CR, Michel FA, and Smith MW (1986) Stratigraphic, isotopic and mineralogical evidence for an early Holocene thaw unconformity at Mayo, Yukon Territory. *Canadian Journal of Earth Sciences* 23: 795–803.
- Curry AJ and Morris CJ (2004) Lateglacial and Holocene talus slope development and rockwall retreat on Mynydd Du, UK. *Geomorphology* 58: 85–106.
- Dixon JC and Thorn CE (2005) Chemical weathering and landscape development in mid-latitude alpine environments. *Geomorphology* 67: 127–145.
- Dunn JR and Hudec PP (1966) Water, clay and rock soundness. *Ohio Journal of Science* 66: 65–78.
- Dyke LD (1984) Frost heaving of bedrock in permafrost regions. *Bulletin of the Association of Engineering Geologists* 21: 389–405.
- Etienne S (2002) The role of biological weathering in periglacial areas: A study of weathering rinds in south Iceland. *Geomorphology* 47: 75–86.
- French HM, Bennett L, and Hayley DW (1986) Ground ice conditions near Rea Point and on Sabine Peninsula, eastern Melville Island. *Canadian Journal of Earth Sciences* 23: 1389–1400.
- Friedmann EI (1982) Endolithic microorganisms in the Antarctic cold desert. *Science* 215: 1045–1053.
- Goodfellow BW, Fredin IO, Derron M-H, and Stroeven AP (2008) Weathering processes and Quaternary origin of an alpine blockfield in Arctic Sweden. *Boreas* 38: 379–398.
- Guglielmin M, Cannone N, Strini A, and Lewkowicz AG (2005) Biotic and abiotic processes on granite weathering of landforms in a cryotic environment, northern Victoria Land, Antarctica. *Permafrost and Periglacial Processes* 16: 69–87.
- Hales TC and Roering JJ (2009) A frost 'buzzsaw' mechanism for erosion of the eastern Southern Alps, New Zealand. *Geomorphology* 107: 241–253.
- Hall K (1999) The role of thermal stress fatigue in the breakdown of rock in cold regions. *Geomorphology* 31: 47–63.
- Hall K, Thorn CE, Matsuoka N, and Prick A (2002) Weathering in cold regions: Some thoughts and perspectives. *Progress in Physical Geography* 26: 577–603.
- Hallet B, Walder JS, and Stubbs CW (1991) Weathering by segregation ice growth in microcracks at sustained sub-zero temperatures: Verification from an experimental study using acoustic emissions. *Permafrost and Periglacial Processes* 2: 283–300.
- Hoch AR, Reddy MM, and Drever JI (1999) Importance of mechanical disaggregation in chemical weathering in a cold alpine environment, San Juan Mountains, Colorado. *Geological Society of America Bulletin* 111: 304–314.
- Hudec PP (1973) Weathering of rocks in Arctic and sub-Arctic environments. In: Aitken JD and Glass DJ (eds.) *Canadian Arctic Geology*, pp. 313–335.

- Saskatoon, Saskatchewan: Geological Association of Canada/Canadian Society of Petroleum Geologists.
- Konishchev VN and Rogov VV (1993) Investigations of cryogenic weathering in Europe and northern Asia. *Permafrost and Periglacial Processes* 4: 49–64.
- Lautridou J-P (1988) Recent advances in cryogenic weathering. In: Clark MJ (ed.) *Advances in Periglacial Geomorphology*, pp. 33–47. Chichester: Wiley.
- Mackay JR (1999) Cold-climate shattering (1974–1993) of 200 glacial erratics on the exposed bottom of a recently drained Arctic lake, Western Arctic Coast, Canada. *Permafrost and Periglacial Processes* 10: 125–136.
- Matsuoka N (2001) Microgelivation versus macrogelivation: Towards bridging the gap between laboratory and field frost weathering. *Permafrost and Periglacial Processes* 12: 299–313.
- Matsuoka N and Murton JB (2008) Frost weathering: Recent advances and future directions. *Permafrost and Periglacial Processes* 19: 195–210.
- McCarroll D and Viles HA (1995) Rock-weathering by the lichen *Lecidea auriculata* in an arctic alpine environment. *Earth Surface Processes and Landforms* 20: 199–206.
- McGreevy JP (1981) Some perspectives on frost shattering. *Progress in Physical Geography* 5: 56–75.
- Murton JB (1996) Near-surface brecciation of Chalk, Isle of Thanet, southeast England: A comparison with ice-rich brecciated bedrocks in Canada and Spitsbergen. *Permafrost and Periglacial Processes* 7: 153–164.
- Murton JB, Coutard J-P, Ozouf J-C, et al. (2001) Physical modelling of bedrock brecciation by ice segregation in permafrost. *Permafrost and Periglacial Processes* 12: 255–266.
- Murton JB, Peterson R, and Ozouf J-C (2006) Bedrock fracture by ice segregation in cold regions. *Science* 314: 1127–1129.
- Pope GA, Dorn RI, and Dixon JC (1995) A new conceptual model for understanding geographical variations in weathering. *Annals of the Association of American Geographers* 85: 38–64.
- Rapp A (1960) Recent development of mountain slopes in Karkevagge and surroundings, northern Scandinavia. *Geografiska Annaler* 42: 73–200.
- Rempel AW (2007) The formation of ice lenses and frost heave. *Journal of Geophysical Research* 112: F020S21. <http://dx.doi.org/10.1029/2006JF000525>.
- Rempel AW (2011) Microscopic and environmental controls on the spacing and thickness of segregated ice. *Quaternary Research* 75: 316–324. <http://dx.doi.org/10.1016/j.yqres.2010.07.005>.
- Thorn CE, Darmody RG, Dixon JC, and Schlyter P (2001) The chemical weathering regime of Karkevagge, arctic-alpine Sweden. *Geomorphology* 41: 37–52.
- Walder JS and Hallet B (1985) A theoretical model of the fracture of rock during freezing. *Geological Society of America Bulletin* 96: 336–346.
- Walder JS and Hallet B (1986) The physical basis of frost weathering: Toward a more fundamental and unified perspective. *Arctic and Alpine Research* 18: 27–32.
- Washburn AL (1979) *Geocryology: A Survey of Periglacial Processes and Environments*. London: Arnold.
- Williams PJ and Smith MW (1989) *The Frozen Earth*. Cambridge: Cambridge University Press.

Permafrost and Glacier Interactions

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Introduction

In terms of their extent and seasonal persistence, glaciers and permafrost constitute the two most important elements of the global cryosphere. Combined, they currently cover an area of over 35 million km², and they were more widespread still during Quaternary glacial periods. However, in spite of their extent and contiguity in both modern and Pleistocene environments and their responsiveness to climatic variations, their potential for active interactions has traditionally received little attention. First, glaciers and permafrost are often viewed as being mutually exclusive because thick ice insulates the land surface from the prevailing climatic conditions. Second, key processes such as basal sliding and subglacial sediment deformation are often considered inactive beneath cold-based ice. Third, with most work in glacial geomorphology and geology undertaken in areas characterized by hard bedrock (e.g., the Laurentian Shield) and/or glaciated by warm-based ice, there has been little need to consider the interaction of cold-based glaciers with permafrozen sediments. A growing body of research, however, has challenged these assumptions by recognizing that, in certain circumstances, glaciers and permafrost can couple and interact and that these interactions are more extensive and glaciologically more significant than was previously thought (Astakhov et al., 1996; Bennett et al., 2003; Cuffey et al., 1999, 2000; Echelmeyer and Zhongxiang, 1987; Murton et al., 2004).

This article considers the key characteristics, formation, and likely extent of glacier–permafrost interactions, their influence on the dynamic behavior of glaciers and ice sheets, and their geomorphological and geological signatures. Specific emphasis is placed on sedimentary sequences and landforms in lowland environments influenced by continental ice sheets as these tend to have a greater preservation potential than those in uplands and are, therefore, the situations most commonly examined in Quaternary paleoglaciological reconstructions. Reviews focusing on glacier–permafrost interactions in high mountain areas are provided by Etzelmüller and Hagen (2005) and Haeberli (2005), and climate-change impacts in glacier and permafrost regions are reviewed by Haeberli and Burn (2002).

The Nature, Formation, and Extent of Permafrost in Glacial Environments

Permafrost is defined as “ground (i.e., soil and/or rock) that remains at or below 0 °C for at least two consecutive years” (French, 2007, p. 83). Defined primarily by temperature and time, permafrost is associated with a range of physical properties and can occur within bedrock, unconsolidated sediments, and organic materials such as peat. Permafrost often contains water either as a liquid or as a solid (ground ice), and it is

important therefore to differentiate between its temperature and state. The terms ‘cryotic’ (≤ 0 °C) and ‘noncryotic’ (> 0 °C) relate solely to temperature, whereas the terms ‘frozen’ and ‘unfrozen’ refer to the state of any included water. In this respect, numerous factors, including pressure, soil particle size, and the presence of impurities, can depress the temperature at which ice starts to nucleate, and so the thermal boundaries between cryotic/noncryotic and frozen/unfrozen can differ by several degrees Celsius (Williams and Smith, 1989). Hence, permafrost is ‘perennially cryotic’ and may be frozen (e.g., ice-bonded) or unfrozen. In comprising a varied mixture of ice, water, and sediment, and being associated with genetic processes essentially identical to those associated with subaerial permafrost, subglacial permafrost can be physically and structurally indistinguishable from the debris-rich ice in the basal layers of glaciers and ice sheets (Hughes, 1973). Consequently, some confusion exists within the literature as to whether such mixtures should be considered part of the glacier (i.e., basal ice) or part of the glacier bed (i.e., frozen subglacial sediment) (Fitzsimons, 2006).

Most permafrost develops subaerially in response to climatic regimes where the mean annual air temperature remains below 0 °C. Glacier–permafrost interactions are therefore commonly associated with ice margins advancing into areas of pre-existing permafrost. However, ice-bonded permafrost can develop or persist subglacially if the temperature of the basal ice remains below the pressure melting point, that is, beneath cold-based glaciers. Conversely, beneath warm-based glaciers, where the basal ice remains at the pressure melting point, the heat flux associated with both active basal processes and the production of large volumes of meltwater is inimical to the long-term survival of ice-bonded permafrost, although unfrozen permafrost may persist.

While direct observations of the extent of subglacial permafrost are sparse, numerical and geomorphological reconstructions suggest that there are three principal situations in which it can occur:

- *Ice-marginal zones where cold basal temperatures are promoted by thin ice cover:* While steady-state models and field observations initially suggested that this zone is restricted to a few kilometers in width (e.g., Mooers, 1990; Moran et al., 1980), transient modeling indicates that subglacial permafrost may extend for tens to hundreds of kilometers during periods of ice advance (Cutler et al., 2000).
- *The interior of ice masses, where air temperatures are at a minimum:* Cold firn can be advected to the base of the ice mass, and flow-related strain heating is minimized by low ice-surface slopes (Dyke, 1993).
- *Areas of strong flow divergence:* The ice rapidly thins and strain heating is minimal (Dyke, 1993).

The spatial extent of cold-based ice and subglacial permafrost can change significantly during a glacial cycle, thereby

influencing rates of ice-sheet growth and decay. Ice temperature modeling of the late Pleistocene Laurentide Ice Sheet has suggested that the growth phase up to the Last Glacial Maximum (LGM) was associated with widespread frozen-bed conditions, with the cold-based area remaining at ~65% for 70 000 years (Marshall and Clark, 2002). Following the LGM, the area of warm-based ice increased exponentially in response to the increased insulation promoted by thicker ice and a combination of geothermal and strain heating. This led to increased flow velocities and the rapid decay of the ice sheet. Geomorphological studies have also indicated that cold-based conditions and renewed growth of subglacial permafrost can occur during deglaciation, due to reductions in ice thickness and basal shear stress (Dyke, 1993).

Basal Processes Beneath Cold-Based Ice

Traditional Views

Glacier–permafrost interactions are generally attributed to cold-based or polythermal glaciers in which basal temperatures remain below the pressure melting point. In this situation, glaciologists have traditionally assumed that basal processes are of little significance due to the limited supply of meltwater to the glacier bed, which is thought to be required to drive both basal sliding and subglacial sediment deformation. Cold-based ice resting on permafrost is therefore often conceptualized as ‘frozen to the bed,’ such that ice flow occurs only through internal creep (e.g., Boulton, 1972). This assumption is commonly used as a boundary condition in numerical ice-sheet models, which prescribe a velocity of zero if the basal temperature is calculated to be below the pressure melting point. This zero velocity at the ice bed was seemingly corroborated by empirical measurements beneath Meserve Glacier, a small cirque glacier in Antarctica with a mean basal temperature of -18.8°C , where a variety of techniques failed to record any movement across the ice–bed contact (Holdsworth and Bull, 1970). As basal processes are associated with erosion and sediment transfer, their absence is also thought to limit the erosivity of cold-based glaciers, although strong adhesion between the ice and the bed may promote plucking and removal of rafts of bedrock or frozen sediment (Boulton, 1972). Consequently, cold-based glaciers have typically been viewed as slow moving and geomorphologically impotent, and of limited research interest.

Evidence for Active Basal Processes Beneath Cold-Based Ice

The assumption that basal processes remain inactive beneath cold-based ice and therefore are always geomorphologically ineffective has been challenged (Waller, 2001). Central to this is the realization that liquid water can persist at temperatures well below the pressure melting point as nanometer-thick interfacial films at ice–bed or ice–particle interfaces, thereby facilitating both sliding and sediment deformation. Gilpin (1980) demonstrated the occurrence of regelation around a weighted wire at temperatures down to -35°C , providing a model that was subsequently used by Shreve (1984) to demonstrate the potential for non-zero sliding speeds at subfreezing temperatures. However, it was Echelmeyer and Zhongxiang’s

(1987) direct observation of basal motion beneath a predominantly cold-based glacier in China that forced glaciologists to reappraise their views on the dynamics of glaciers resting on permafrost. Using a tunnel excavated in the base of the Urumqi Glacier Number 1, they observed both sliding over bedrock and deformation of a layer of ‘ice-laden drift’ equivalent to 85% of the surface motion, in spite of recorded temperatures remaining below -1.75°C . Direct observations of the subglacial zone of the Suess Glacier in Antarctica (basal temperature of -17°C) suggest that ice lenses and lenses lacking pore ice can provide potential lines of weakness, permitting bed deformation and the entrainment of debris at very low temperatures (Fitzsimons, 2006). Additionally, repeat measurements at the Meserve Glacier have demonstrated small but measurable sliding rates of between 2 and 8 mm a $^{-1}$ at a temperature of -17°C (Cuffey et al., 1999).

Stratigraphic observations of formerly glaciated regions where permafrost has remained intact since its deformation by Pleistocene ice sheets provide a vital additional insight into the nature of glacier–permafrost interactions. Work in Western Siberia (Astakhov et al., 1996) and the Western Canadian Arctic (Murton et al., 2004) has demonstrated their extensive nature, with Astakhov et al. (1996) concluding that, “our inspection of frozen relicts of the Pleistocene glaciation in Siberia shows ubiquitous traces of this process over vast expanses, sediment sheets tens to hundreds of metres thick being involved in subglacial deformation” (p. 186). These studies highlight the crucial role played by a combination of temperature, ice content, and liquid water content. While coarse-grained sediments freeze at temperatures close to the pressure melting point, fine-grained sediments can retain significant quantities of liquid water at temperatures as low as -20°C (Williams and Smith, 1989). Consequently, ice-rich and clay-rich sediments remain deformable at temperatures below the pressure melting point. By providing zones of weakness, such layers can result in a migration of the zone of maximum shear and the mobilization of the entire overlying sequence (Figure 1). Consequently, the thickness of partially frozen subglacial deforming layers is thought to be substantially greater than that of entirely unfrozen deformation sequences (usually <1 m thick) (Waller et al., 2009) (see Section ‘Geological Evidence’).

Geomorphological Evidence

Proglacial and Ice-Marginal Landforms

Geomorphological investigations of the ice-marginal environments surrounding former Pleistocene ice masses suggest that glacier–permafrost interactions were particularly active within these areas. Cold-based ice interacted with pre-existing frozen sediments to generate distinctive landform assemblages comprising push moraines or ice-thrust complexes, hummocky or controlled moraines, and meltwater channels. These constitute a landform assemblage potentially indicative of glacier–permafrost interactions (Dyke and Evans, 2005; Etzelmüller and Hagen, 2005).

Push moraines

Geomorphologists have long considered the importance of permafrost in the formation of push moraines (Boulton

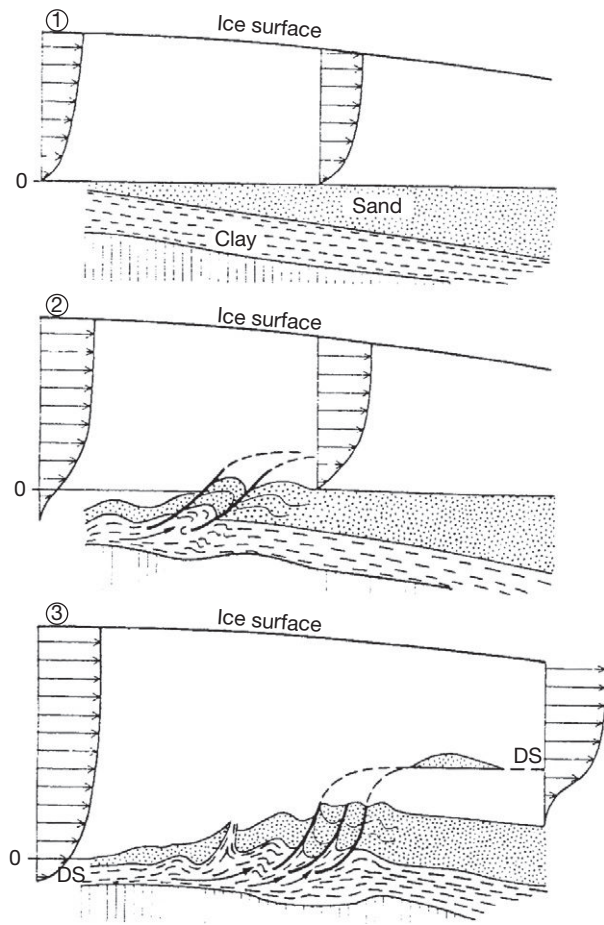


Figure 1 Deformations associated with the interaction of the Kara Ice Sheet with pre-existing and deep-seated permafrost composed of clay and sand in West Siberia as described by Astakhov et al. (1996, p. 186). Stages 1–3 relate to progressive thickening of the ice during an advance. Parallel arrows indicate reconstructed velocity profiles, '0' marks the original ice-bed interface, and 'DS' represents the dynamic sole of the glacier.

et al., 1999; Haeberli, 1979; Kälin, 1971; Kupsch, 1962). Permafrost is considered particularly influential in the development of large, multicrested push moraines (Bennett, 2001; Figure 2), although Haeberli (1979) suggested that its influence extends to the development of smaller features in high mountain regions. Permafrost is considered influential for two reasons: First, ice-bonded permafrost is often thought necessary to increase sediment cohesion and explain the preservation of primary bedding within unconsolidated sedimentary units that have been heavily deformed (e.g., Kupsch, 1962). In combination with the development of strong adhesion between the advancing ice and the pre-existing permafrost, the enhanced rigidity facilitates stress propagation into the foreland. Second, the development of very high pore-water pressures at the base of pre-existing permafrost due to the restriction of drainage from subglacial aquifers is thought to provide an ideal décollement along which displacement can occur (Mooers, 1990) and, in some cases, result in the entrainment of large bedrock rafts (Moran et al., 1980). Research within

modern-day environments has clarified the role of proglacial permafrost. For example, Boulton et al. (1999) concluded that the development of a 1.5-km-wide push moraine at Holmstrømbreen, Spitsbergen, required cementation of the foreland by ice-bonded permafrost and the development of an artesian head beneath this aquiclude of around 40 m.

Hummocky, controlled, and ribbed moraines

Glacier–permafrost interactions can also promote the formation of debris-rich ice margins and the subsequent development of hummocky or controlled moraines (Attig et al., 1989; Evans, 2009). Polythermal glaciers and ice sheets with cold-based margins terminating on permafrost are commonly associated with warm-based interiors where meltwater is generated. This water is driven to the cold-based margin by the gradient in ice overburden pressure where it refreezes, resulting in the net transport of water and the entrainment of thick sequences of basal ice (Weertman, 1961). Flow compression, folding, and thrusting can further thicken the sequence. The subsequent melt-out of this basal ice typically produces extensive areas of chaotic hummocky moraine or controlled moraines (where the melt-out of debris-rich structures imparts a linearity to the moraine) composed of melt-out tills, flow tills, and glaciofluvial sediments. The presence of hummocky terrain has consequently been used in Quaternary settings to delineate the extent of ice-marginal permafrost (e.g., Attig et al., 1989; Moran, et al., 1980). In regions where the permafrost has persisted since deglaciation, large volumes of ice buried within hummocky moraine can be preserved. Such 'retarded' deglaciation provides an extensive and potentially immensely valuable paleoglaciological archive the importance of which has yet to be explored (Grosvald et al., 1986).

Kleman and Hättestrand (1999) have suggested that a switch from frozen- to thawed-bed conditions can result in the extension, fracture, and boudinage of frozen till sheets to produce ribbed or Rogen moraines. They are therefore an additional landform considered indicative of changes in basal thermal regime.

Meltwater channels

The presence of permafrost clearly limits the generation of subglacial meltwater. But meltwater can be derived from other sources and locations. For example, in mapping the landscape impact of former ice caps in the Canadian High Arctic, Dyke and Evans (2005) highlighted the development of extensive networks of lateral meltwater channels eroded by supraglacial meltwater along cold-based ice margins. Such channels can provide a high-resolution record of ice recession. Attig et al. (1989) suggested that ice-marginal 'frozen-bed' zones can feature large tunnel channels associated with the episodic drainage of meltwater from warm-based areas located up ice.

Subglacial Landforms

Many glacial geomorphologists have assumed that cold-based glaciers resting on permafrost protect and preserve preglacial landforms. This view is commonly employed in



Figure 2 Proglacial moraine complex at the margin of the Leverett Glacier, western Greenland (67°03'28"N, 50°10'50"W), associated with ice advance into pre-existing permafrost.

geomorphological inverse models that relate the occurrence of preglacial features such as tors and patterned ground to cold-based conditions, in contrast to flow lineations such as drumlins generated beneath areas of warm-based ice (Dyke, 1993; Kleman and Borgström, 1996). As areas lacking clear evidence of landscape modification can also be interpreted as having been unglaciated (i.e., nunataks), this has led to significant variations in Quaternary ice-sheet and sea-level reconstructions, although cosmogenic isotope dating is helping to resolve these debates.

Even where processes at the ice-bed interface itself are inactive, glacial erosion and entrainment may still occur. Boulton (1972) suggested that englacial debris can come into contact with the bed if the flow lines intersect protuberances. Adhesion of the glacier sole to the bed can also result in quarrying or plucking of large lithified or unlithified erratics if the pressure melting point isotherm is at a shallow depth and the bed contains favorably orientated discontinuities.

Geomorphological mapping of glacier forelands in East Antarctica has revealed a range of distinctive landforms produced by cold-based glaciers that provide evidence of their ability to erode, transport, and deposit. While the basal temperatures beneath nearby glaciers were estimated at -24°C , Atkins et al. (2002) documented a range of features reflecting the plucking of clasts from the stoss side of bedrock outcrops, their dragging across the bed to create abrasional features such as striae, and their deposition as comminuted breccia, isolated boulders, or ice-cored cones. Working on the Meserve Glacier in Antarctica, Cuffey et al. (2000) measured sediment transfer rates of $10\text{--}30\text{ m}^3\text{ a}^{-1}$ associated with the entrainment and transport of eroded material within a basal ice layer. While the corresponding denudation rates of 9×10^{-7} and $3 \times 10^{-6}\text{ m a}^{-1}$ are very low compared to warm-based glaciers, they demonstrate the potential of cold-based glaciers to create significant landforms, given long time periods.

Geological Evidence

Cold-based ice can actively couple with and deform seasonally frozen ground or permafrost. Bennett et al. (2003), for example, described the occurrence of attenuated ice blocks and ice-filled shear planes in a frozen diamicton exposed following the winter surge of an Icelandic outlet glacier. The ice blocks

were attributed to the collapse of a rapidly advancing and heavily crevassed ice margin and entrainment into an initially unfrozen subglacial sediment. Subsequent attenuation to form augen-like structures highlighted a transition to subfreezing conditions during which deformation remained active (Figure 3(a)). While the short-lived nature of these conditions means they do not reflect glacier–permafrost conditions *sensu stricto*, they provide an insight into the style of deformation associated with deformation under subfreezing conditions.

Investigations of permafrost sedimentary sequences in Western Siberia (Astakhov et al., 1996) and the Western Canadian Arctic (Murton et al., 2004, 2005), which were deformed by the high-latitude margins of Kara and Laurentide Ice Sheets and have subsequently remained frozen, provide clear evidence for the widespread occurrence of Pleistocene glacier–permafrost interactions. They also highlight the related geological processes and products:

- The rheological complexity of permafrost sequences associated with variations in ice content and particle size. Astakhov et al. (1996) documented structures indicative of both ductile and brittle deformation. Sands are comparatively rigid, deforming via brittle failure (unless ice rich) to create boundinaged sand lenses or boudins. In contrast, both ice-rich and clay-rich sediments deform comparatively easily and in a ductile manner. Murton et al. (2004) described recumbent and S-shaped folds, sand lenses, and layers within glacially deformed permafrost, inferring the ductile deformation of warm, ice-rich materials (Figure 3(a)). Collectively, these structures reflect simple shear and extension consistent with submarginal rather than proglacial glacio-tectonic deformation. Significantly, the thickness of glacially deformed permafrost is substantial. Astakhov et al. (1996) reported thicknesses of tens to hundreds of meters (Figure 1), while Murton et al. (2004) reported thicknesses of at least 5–20 m (Figure 3(c)).
- Deformation was accompanied by erosion of permafrost. Murton et al. (2004) described blocks of ice up to 15 m long and 1.5 m high, some folded, dispersed throughout the glaciectonite (Figure 3(c)). Submarginal erosion of ice also explains the occurrence of an angular unconformity (décollement surface) along the top of the underlying massive ice. Preservation of these ‘ice clasts’ and the massive ice indicates that erosion occurred at subfreezing temperatures.

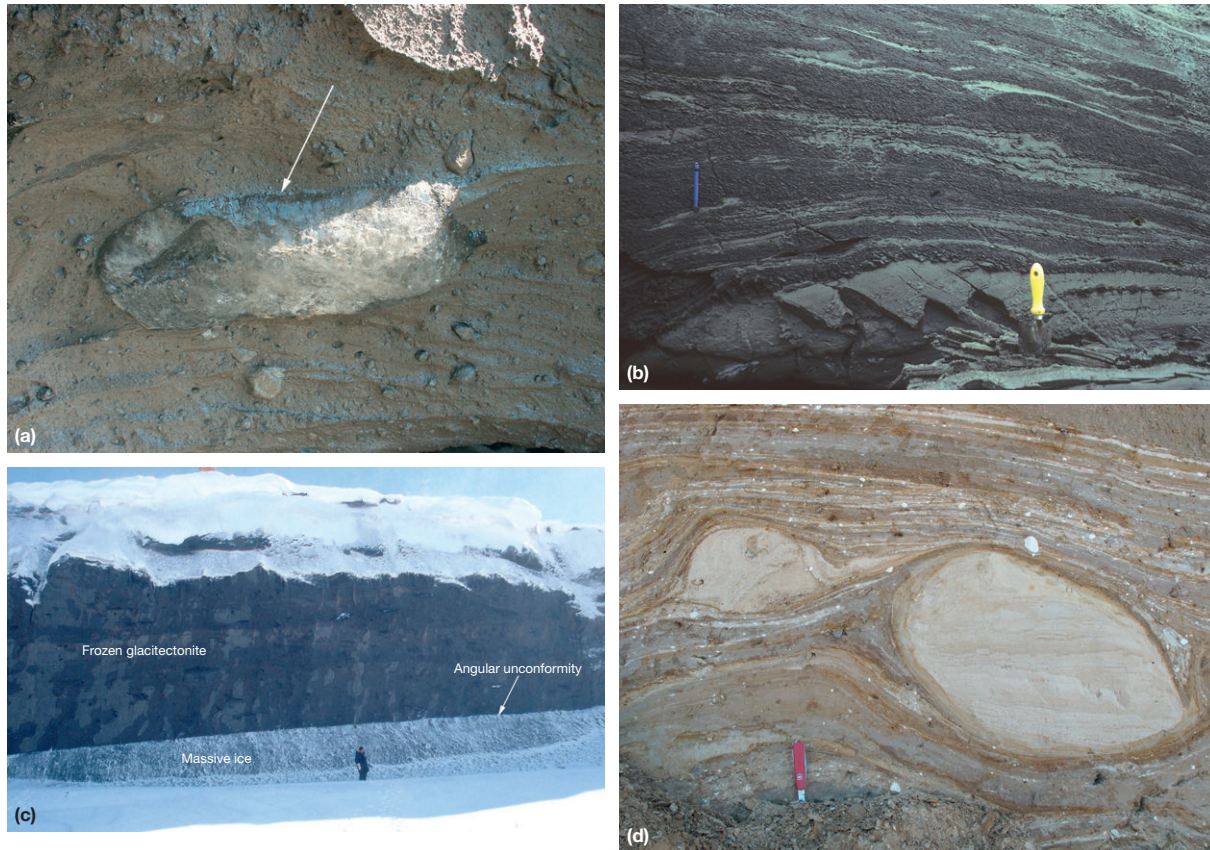


Figure 3 Geological evidence of glacier–permafrost interactions: (a) ice augen (~60 cm long, ice flow from left to right, highlighted by the arrow) situated within frozen diamict at the margin of Hagafellsjökull-Eystri, Iceland ($64^{\circ}31'19''$ N, $20^{\circ}9'29''$ W); (b) attenuated sand lenses and fold noses within silty clay (trowel for scale), North Head, Western Canadian Arctic ($69^{\circ}43'26''$ N, $134^{\circ}25'40''$ W); (c) frozen glaciectonite overlying massive ice (person for scale), Pullen Island, Western Canadian Arctic ($69^{\circ}46'27''$ N, $134^{\circ}24'26''$ W); and (d) streamlined sand intraclasts within laminated diamict (penknife for scale) at West Runton, North Norfolk, UK ($52^{\circ}56'36''$ N, $1^{\circ}13'41''$ E).

- Relict basal ice from the Laurentide and Kara Ice Sheets is also widely preserved in both glaciated regions. This is because permafrost there has existed continuously since before the last glaciation, with the result that deglaciation has been arrested, and will not be completed until the permafrost that preserves the glacial ice has degraded.
- The injection of pressurized water into proglacial permafrost occurred during the retreat of the Laurentide Ice Sheet (Morton, 2005). Superimposed on some glacially deformed permafrost are undeformed dykes and sills of intrusive ice and some undeformed massive segregated–intrusive ice which clearly postdates deformation.

Widespread permafrost development during Pleistocene glacial periods suggests that glacier–permafrost interactions are likely to have occurred around some mid-latitude margins of ice sheets during their initial advances. Subsequent permafrost degradation would, however, have destroyed or disrupted ice-hosted structures that provide incontrovertible evidence of the interactions within modern-day permafrost regions. Waller et al. (2011) have suggested that sand intraclasts

commonly found within till sequences may provide diagnostic sedimentological evidence for former glacier–permafrost interactions. Numerous authors have questioned the likely survival of these unconsolidated sand masses that frequently contain undisturbed primary bedding within otherwise heavily deformed sequences. Waller et al. (2011) argued that particle-size induced variations in the quantity of liquid water associated with ‘warm permafrost’ provide a simple explanation for the rheological heterogeneities between the comparatively rigid sand intraclasts and the surrounding, more easily deformed fine-grained tills observed within glaciectonic sequences at West Runton, North Norfolk, UK (Figure 3(d)). Their entrainment is also consistent with Boulton’s (1972) prediction that the subglacial quarrying and entrainment of unlithified rafts is a process most likely associated with cold-based ice. If correct, this will allow sedimentological evidence to be used in combination with geomorphological evidence (e.g., Kleman and Borgström, 1996) to reconstruct the basal thermal regimes of Quaternary ice sheets.

Permafrost aggradation during deglaciation may also be associated with distinctive sedimentological evidence. A progressive cooling of the subglacial permafrost could lead to a

switch from ductile to brittle deformation and the development of polyphase tectonic structures (Murton et al., 2004).

Conclusion

Although glaciers and permafrost have often been considered largely mutually exclusive, recent work has led to an increasing awareness of the important role played by active glacier–permafrost interactions, particularly in ice-marginal settings. Transient ice-sheet model outputs suggest that subglacial permafrost may extend for tens to hundreds of kilometers up-glacier of the margin during advances into areas of thick permafrost. In addition, field observations within both modern and Pleistocene glacial environments have falsified the assumption that basal processes are inactive beneath cold-based ice. In certain circumstances, the persistence of interfacial films of liquid water adsorbed to particles and bedrock surfaces at temperatures well below the pressure melting point can sustain non-zero sliding velocities and allow partially frozen sediments to deform. Where soft-bedded glaciers overlie ‘warm permafrost,’ these processes can dominate total surface motion. Even beneath very cold ice, this can lead to some erosion, entrainment, and sediment transport.

Glaciers overlying permafrost generate a variety of distinctive landforms and sediments. The presence of ice- and clay-rich layers within permafrost or, alternatively, the attainment of high pore-water pressures at the base of permafrost can provide potential décollement surfaces and result in the mobilization of bedrock rafts or sequences tens to hundreds of meters thick as the dynamic bed is displaced from the ice-bed interface into the permafrozen substrate. This can generate complexly deformed sequences or push moraines at the ice margin. Net basal adfreezing, thrusting, and folding can also lead to the creation of thick basal ice sequences beneath the cold-based margins of polythermal glaciers, which can subsequently melt out to produce large tracts of hummocky or controlled moraine. Alternatively, in permafrost regions, they can lead to the burial and preservation of large volumes of massive ice. Finally, the presence of transitions from cold- to warm-based ice can lead to the development of Rogen moraines.

Further research is required to determine the extent and rheology of subglacial permafrost. Identification of diagnostic sedimentological evidence for glacier–permafrost interactions in particular will complement geomorphological inverse models and help constrain their spatial extent beneath Quaternary ice sheets and by analogy, modern-day ice masses. Further work is also required to elucidate the conditions under which basal processes remain active at subfreezing temperatures and to constrain their varying contribution to glacier flow and geomorphic activity at different temperatures and in different substrates. This will both clarify the role of glacier–permafrost interactions and allow their inclusion into ice-sheet models.

See also: **Cosmogenic Nuclide Dating:** Methods; Overview. **Glacial Landforms:** Introduction.

References

- Astakhov VI, Kaplyanskaya FA, and Tarnogradskiy VD (1996) Pleistocene permafrost of West Siberia as a deformable glacier bed. *Permafrost and Periglacial Processes* 7: 165–191.
- Atkins CB, Barrett PJ, and Hicock SR (2002) Cold glaciers erode and deposit: Evidence from Allan Hills, Antarctica. *Geology* 30: 659–662.
- Attig JW, Mickelson DM, and Clayton L (1989) Late Wisconsin landform distribution and glacier-bed conditions in Wisconsin. *Sedimentary Geology* 62: 399–405.
- Bennett MR (2001) The morphology, structural evolution and significance of push moraines. *Earth-Science Reviews* 53: 197–236.
- Bennett MR, Waller RI, Midgley NG, et al. (2003) Subglacial deformation at sub-freezing temperatures? Evidence from Hagafellsjökull-Eystrí, Iceland. *Quaternary Science Reviews* 22: 915–923.
- Boulton GS (1972) The role of thermal régime in glacial sedimentation. In: Price RJ and Sugden DE (eds.) *Polar Geomorphology*, pp. 1–19. London: Institute of British Geographers Special Publication No. 4.
- Boulton GS, van der Meer JJM, Beets DJ, Hart JK, and Ruegg GHJ (1999) The sedimentary and structural evolution of a recent push moraine complex: Holmströmbreen, Spitsbergen. *Quaternary Science Reviews* 18: 339–371.
- Cuffey KM, Conway H, Gades AM, et al. (2000) Entrainment at cold glacier beds. *Geology* 28: 351–354.
- Cuffey KM, Conway H, Hallet B, Gades AM, and Raymond CF (1999) Interfacial water in polar glaciers and glacier sliding at -17°C . *Geophysical Research Letters* 26: 751–754.
- Cutler PM, MacAyeal DR, Mickelson DM, Parizek BR, and Colgan PM (2000) A numerical investigation of ice-lobe-permafrost interaction around the southern Laurentide ice sheet. *Journal of Glaciology* 46: 311–325.
- Dyke AS (1993) Landscapes of cold-centred Late Wisconsinan ice caps, Arctic Canada. *Progress in Physical Geography* 17: 223–247.
- Dyke AS and Evans DJA (2005) Ice-marginal terrestrial landsystems: Northern Laurentide and Innuitian Ice Sheet margins. In: Evans DJA (ed.) *Glacial Landsystems*, pp. 143–165. London: Arnold.
- Echelmeyer K and Zhongxiang W (1987) Direct observation of basal sliding and deformation of basal drift at sub-freezing temperatures. *Journal of Glaciology* 33: 83–98.
- Etzelmüller B and Hagen JO (2005) Glacier-permafrost interaction in Arctic and alpine mountain environments with examples from southern Norway and Svalbard. In: Harris C and Murton JB (eds.) *Cryospheric Systems: Glaciers and Permafrost*, vol. 242, pp. 11–28. London: Geological Society of London. Geological Society Special Publication.
- Evans DJA (2009) Controlled moraines: Origins, characteristics and palaeoglaciological implications. *Quaternary Science Reviews* 28: 183–208.
- Fitzsimons SJ (2006) Mechanical behaviour and structure of the debris-rich basal ice layer. In: Knight PG (ed.) *Glacier Science and Environmental Change*, pp. 329–334. Oxford: Blackwell Publishing.
- French HM (2007) *The Periglacial Environment*, 3rd edn. Chichester: Wiley.
- Gilpin RR (1980) Wire regelation at low temperatures. *Journal of Colloid and Interface Science* 77: 435–448.
- Grosvald MG, Vtyurin BI, Sukhodrovskiy VL, and Shishorina ZG (1986) Underground ice in Western Siberia: Origin and geological significance. *Polar Geography and Geology* 10: 173–183.
- Haeberli W (1979) Holocene push-moraines in Alpine Permafrost. *Geografiska Annaler* 61A: 43–48.
- Haeberli W (2005) Investigating glacier-permafrost relationships in high-mountain areas: Historical background, selected examples and research needs. In: Harris C and Murton JB (eds.) *Cryospheric Systems: Glaciers and Permafrost*, vol. 242, pp. 29–38. London: Geological Society of London. Geological Society Special Publication.
- Haeberli W and Burn C (2002) Natural hazards in forests – Glacier and permafrost effects as related to climate changes. In: Sidle RC (ed.) *Environmental Change and Geomorphic Hazards in Forests. IUFRO Research Series* 9: 167–202. Oxfordshire: CABI.
- Holdsworth G and Bull C (1970) The flow law of cold ice; investigations at Meserve Glacier, Antarctica. In: Gow AJ, Keeler C, Langway CC, and Weeks WF (eds.) *International Symposium on Antarctic Glaciological Exploration*, pp. 204–216. Gentbrugge, Belgium: International Association of Hydrological Sciences IAHS Publication No. 86.
- Hughes T (1973) Glacial permafrost and the pleistocene ice ages. *North American Contribution to the 2nd International Conference on Permafrost*, pp. 213–222. Washington, DC: National Academy of Sciences.

- Kälin M (1971) The active push moraine of Thompson Glacier, Axel Heiberg Island, Canadian Arctic Archipelago, Canada. *Axel Heiberg Island Research Report*. Montreal: McGill University. <http://dx.doi.org/10.3929/ethz-a-000088589>.
- Kleman J and Borgström I (1996) Reconstruction of palaeo-ice sheets: The use of geomorphological data. *Earth Surface Processes and Landforms* 21: 893–909.
- Kleman J and Hättestrand C (1999) Frozen-bed Fennoscandian and Laurentide Ice Sheets during the Last Glacial Maximum. *Nature* 402: 63–66.
- Kupsch WO (1962) Ice-thrust ridges in western Canada. *Journal of Geology* 70: 582–594.
- Marshall SJ and Clark PU (2002) Basal temperature evolution of North American ice sheets and implications for the 100-kyr cycle. *Geophysical Research Letters* 67: 1–67–4.
- Mooers HD (1990) Ice-marginal thrusting of drift and bedrock: Thermal regime, subglacial aquifers, and glacial surges. *Canadian Journal of Earth Sciences* 27: 849–862.
- Moran SR, Clayton L, Hooke R, Le B, Fenton MM, and Andriashek LD (1980) Glacier-bed landforms of the Prairie region of North America. *Journal of Glaciology* 25: 457–476.
- Murton JB (2005) Ground-ice stratigraphy and formation at North Head, Tuktoyaktuk Coastlands, Western Arctic Canada: A product of glacier–permafrost interactions. *Permafrost and Periglacial Processes* 16: 31–50. <http://dx.doi.org/10.1002/ppp.513>.
- Murton JB, Waller RI, Hart JK, Whiteman CA, Pollard WH, and Clark ID (2004) Stratigraphy and glaciotectionic structures of permafrost deformed beneath the northwest margin of the Laurentide Ice Sheet, Tuktoyaktuk Coastlands, Canada. *Journal of Glaciology* 50: 399–412.
- Murton JB, Whiteman CA, Waller RI, Pollard WH, Clark ID, and Dallimore SR (2005) Basal ice facies and supraglacial melt-out till of the Laurentide Ice Sheet, Tuktoyaktuk Coastlands, Western Arctic Canada. *Quaternary Science Reviews* 24: 681–708.
- Shreve RL (1984) Glacier sliding at subfreezing temperatures. *Journal of Glaciology* 30: 341–347.
- Waller RI (2001) The influence of basal processes on the dynamic behaviour of cold-based glaciers. *Quaternary International* 86: 117–128.
- Waller RI, Murton JB, and Whiteman CA (2009) Geological evidence for subglacial deformation of Pleistocene permafrost. *Proceedings of the Geologists' Association* 120: 155–162.
- Waller RI, Phillips E, Murton JB, Lee JR, and Whiteman CA (2011) Sand intraclasts as evidence of subglacial deformation of Middle Pleistocene permafrost, North Norfolk, UK. *Quaternary Science Reviews* 30(23–24): 3481–3500.
- Weertman J (1961) Mechanism for the formation of inner moraines found near the edge of cold ice caps and ice sheets. *Journal of Glaciology* 3: 965–978.
- Williams PJ and Smith MW (1989) *The Frozen Earth: Fundamentals of Geocryology*. Cambridge: Cambridge University Press.

Relevant Websites

<http://ipa.arcticportal.org/> – International Permafrost Association.
http://gsc.nrcan.gc.ca/permafrost/whatis_e.php – Natural Resources Canada: Permafrost.

Block/Rock Streams

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Introduction

Block streams (or rock streams) result from the accumulation by mass wasting of coarse rock debris and extend a greater distance downslope than across slope. Thus, they are linear deposits with downslope alignment and occupy some hillslopes and valley axes in areas of past or present periglacial conditions (White, 1976, 1981). Most of the block streams described in the literature are relict features. Active block streams are those that continue to move downslope under the influence of the locally prevailing climate.

Block streams have often been confused with block fields, block slopes, talus, and rock glaciers, probably because these features can occur in close proximity and may grade into one another, making it difficult to isolate and delimit individual forms. In the Falkland Islands, South Atlantic, where (relict) block streams are known as stone runs, Clark (1972) recognized that they were components of a former cold-climate erosion–transport–deposition system that included (1) hilltops (debris source areas) characterized by block fields, altiplanation terraces, and residual tors; (2) hillsides (zones of debris transport) dominated by frost-riven crags, scree, block terraces, and stone runs; and (3) valleys (zones of debris accumulation) occupied by stone runs and sheets of head. It is likely that block streams in many other regions are also part of a local continuum of blocky materials, and this is probably the reason for some of the terminological confusion encountered in the literature. Further confusion may have arisen because (active) block streams and stone runs have been considered to be distinctly different features (Harris, 1994). Adopting the basic morphotopographical definition given previously enables a range of block-stream forms of various dimensions, gradients, and composition to be accommodated and allows both relict and active features to be included.

Geographical Distribution of Block Streams

Block streams have been recorded in many areas of the world and are generally found at low to high altitudes in the mid- to high latitudes or at high altitudes in the low latitudes. These are the regions that either currently experience periglacial climatic conditions or did so at various times during the Pleistocene. However, the total area occupied by block streams is not known, although it is undoubtedly very small. In the Northern Hemisphere, block streams have been recorded in parts of the United States and Canada (Harris and Pedersen, 1998; Nelson et al., 2007; Potter and Moss, 1968; Smith, 1949, 1953), China (Harris et al., 1998), Russia (where they are called kurums; Perov, 1969; Romanovskij and Tyurin, 1986), Italy (Firpo et al., 2006), and England (Bennett et al., 1996; Clark, 1994). In the Southern Hemisphere, block streams have been reported from Australia (Caine, 1983; Caine and Jennings, 1968), South Africa (Boelhouwers, 1999; Boelhouwers and Sumner, 2003), and the Falkland Islands (Andersson, 1906, 1907; Clark, 1972).

Characteristics of Block Streams

Plan Form

Block streams can assume a range of plan forms (Figure 1). A single-thread form (straight or sinuous) may occur in part or along the entire block-stream length; such patterns are usually confined to valley axes. Some axial block streams with numerous tributary feeder block streams have a dendritic plan. On more open slopes, braided or anastomosing plans can develop, and with increasing distance downslope, this pattern may grade into a single-thread form. In the Falkland Islands, many block streams consist of parallel block stripes. These can diverge from and/or converge into more extensive hillside or axial block spreads.

Dimensions

There is great variation in the reported dimensions of block streams. They range from widths of up to a few tens of meters and lengths of up to a few hundred meters to widths of several hundred meters and lengths of up to 5 km (Barrows et al., 2004; Clark, 1972; Harris et al., 1998; White, 1976). Quoted dimensions for some relict block streams are likely to be minima. Several reports refer to coarse debris being overlain by peat and vegetation along block-stream margins, indicating a continuation of debris beyond the apparent limits (Andersson, 1906; Caine and Jennings, 1968; Smith, 1949).

Morphology

A surface consisting of openwork rock debris is common to all block streams, but there is usually considerable morphological diversity both across individual block streams and between them. Some block-stream surfaces are diversified by random or aligned pits and elongated depressions, whereas others have longitudinal furrows extending for considerable distances along their central tracts. Such features can be 2 or 3 m in depth. Along the longitudinal axis of many block streams, boulder steps, arcuate ridges, and lobes occur. These rise between 0.5 and 9 m above the areas immediately downslope, and it is not always clear whether they are controlled by underlying rock scarps or represent overriding masses (flow structures) of coarse debris (Barrows et al., 2004; Boelhouwers, 1999; Clark, 1994; Potter and Moss, 1968).

Although block-stream longitudinal gradients are generally $<15^\circ$, with values as low as 1° or 2° being reported (Andersson, 1907; Caine, 1983; Clark, 1972; Smith, 1953), gradients of approximately $30\text{--}35^\circ$ have also been recorded (Barrows et al., 2004; Harris and Pedersen, 1998; Harris et al., 1998; Perov, 1969). On relict block streams, gradients in excess of 15° are usually, but not exclusively, associated with the distal slopes of steps, ridges, and lobes and also with the terminal zone (Boelhouwers, 1999; Caine, 1983; Potter and Moss, 1968).

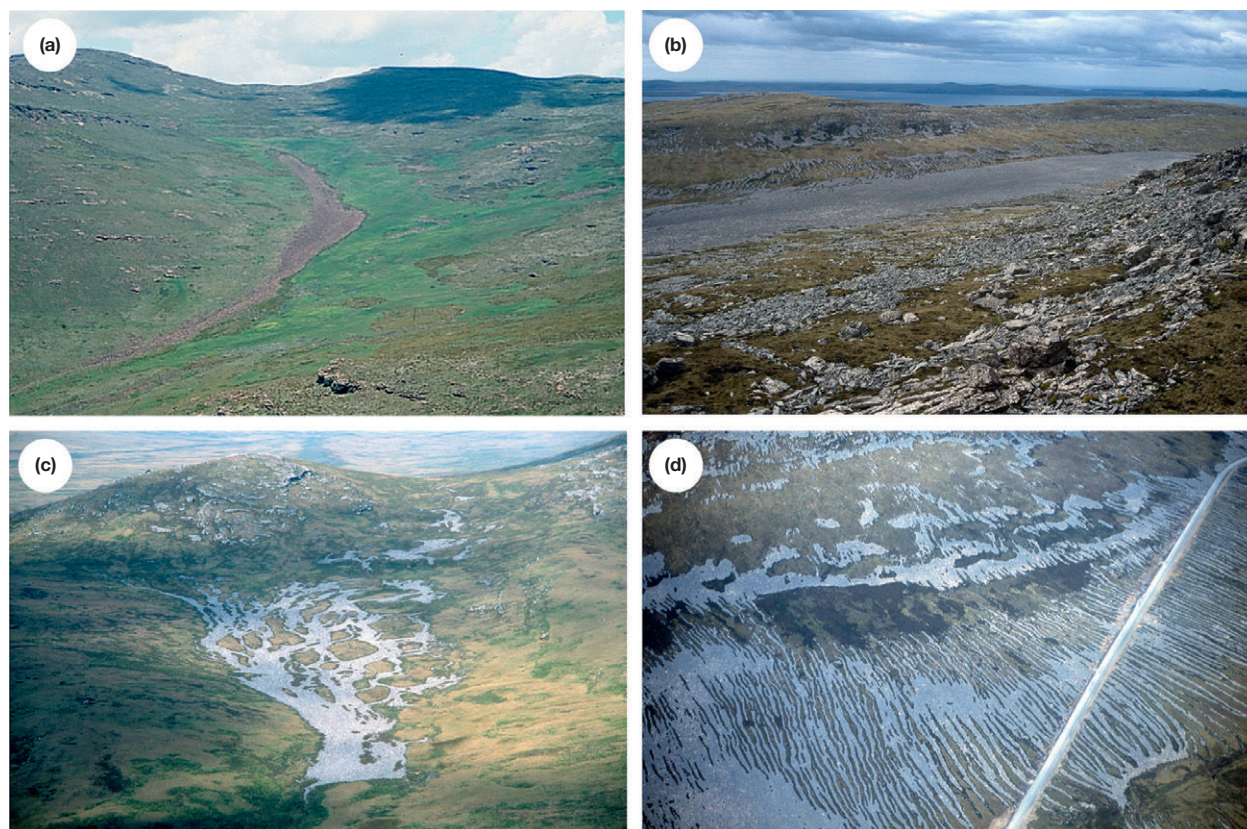


Figure 1 Situations and plan forms of block streams. (a) Single-thread valley-axis block stream near Sani Pass, Lesotho. Reproduced from Boelhouwers JC, Holness S, Meiklejohn I, and Sumner P (2002) Observations on a block stream in the vicinity of Sani Pass, Lesotho Highlands, Southern Africa. *Permafrost and Periglacial Processes* 13: 251–257. © John Wiley & Sons. (b) Valley-axis block stream with tributary feeder block streams (dendritic plan form), Falkland Islands. (c) Block stream with braided or anastomosing plan form, Falkland Islands. (d) Parallel block stripes, Falkland Islands.

Some block streams are downslope extensions to talus and/or block fields/block slopes; these block streams are usually considerably steeper in their upper reaches than in their lower reaches. Other block streams originate from rock scarps or tor-like outcrops, whereas some are sourced from the upper parts of convex ridges that lack upstanding outcrops (Figure 2; Boelhouwers, 1999; Clapperton, 1975; Potter and Moss, 1968; Romanovskij and Tyurin, 1986), deposits of till (Harris et al., 1998), debris terraces (Clark, 1972), and low-gradient solifluction slopes (Barrows et al., 2004).

Block-stream margins may display either a smooth or an irregular form in different parts and can be sharp or diffuse, the latter usually being associated with the encroachment of peat and vegetation from adjacent ground. Terminal areas of block streams have often been reported as being lobate in plan and convex in profile, but Smith (1968) described a block-stream toe in the Beartooth Mountains of Wyoming and Montana as having a V-shaped reentrant descending 10 m to a lower boulder zone. A similar stepped reentrant has been observed in the front of a Falkland Islands block stream.

Structure and Composition

Some relict block streams have attained thicknesses of 9 or 10 m in part (Boelhouwers, 1999; Potter and Moss, 1968; White, 1981), although average thicknesses tend to be in the

range of 1–3 m. Romanovskij and Tyurin (1986) cited examples of active kurums in Arctic regions of Russia as being 0.3 or 0.4 m thick, and Harris et al. (1998) described an active block stream in China with a thickness of 0.15 m.

Block streams, and other coarse blocky accumulations, are usually found on well-jointed igneous and metamorphic rocks, presumably because these lithologies are susceptible to macrogelivation, thermal stress fracturing, and/or deep chemical weathering along existing joints. These processes, acting on well-jointed rocks, usually produce numerous large blocks. The English language block-stream literature is dominated by examples developed on quartzite, basalt, and dolerite.

Several records exist of relict block-stream stratigraphy, as observed in road cuts or natural exposures. These generally indicate a marked decrease in block size with increasing depth (inverse grading) and a lack of interstitial fine material in at least the uppermost part of the debris (Figure 3(a); Boelhouwers, 1999; Clapperton, 1975; Clark, 1972, 1994; Potter and Moss, 1968; Smith, 1953). The most detailed descriptions of block-stream structure are those from southeastern Australia by Caine and Jennings (1968) and Caine (1983). Three stratigraphic units have been recognized. The uppermost is a 0.5–3-m-thick matrix-free layer of openwork blocks with minor quantities of angular gravel. Below this occurs a block layer 0.2–0.6 m thick with void space occupied by organic-rich fines and capped by a 0.01–0.03-m-thick layer of gravel derived



Figure 2 Examples of block-stream source areas, in the Falkland Islands. (a) Block stream originating from a 2-m-high rock scarp. (b) Block stream originating from near the crest of a convex ridge without upstanding outcrops.

from weathering of the overlying blocks. The upper surface of this layer is usually at the level of the local water table. The basal layer consists of blocks with intervening void space filled by yellow brown/dark brown silty sand and is up to 0.5 m thick. Beneath the block streams, deeply weathered bedrock is present in which some original structures are preserved.

Sections across block stripes in the Falkland Islands show that some display inverse grading and occupy depressions in the underlying fine-grained material. Beneath the adjacent vegetated stripes, fine-grained material containing relatively few boulders is found (**Figure 3(b)**). Although the stratigraphy of active block streams is little known beyond the statement given by **Harris et al. (1998)** that a sandy loam underlies a 0.15-m-thick layer of blocks, it indicates that materials of considerably finer grain size underlie even shallow block accumulations.

Measures of the size and shape of clasts contained within block streams reveal that marked variations exist. Surface clasts between 0.2 m and several meters in length have been reported (**Bennett et al., 1996; Clark, 1994; Perov, 1969**). Blocks 6–8 m in length were recorded by **Andersson (1907)**, **Smith (1949, 1953)**, and **Potter and Moss (1968)**. A 9-m-long block occurs in a Falkland Islands stone run (**Figure 4**). **Aldiss and Edwards (1999)** noted local variations in block size. Statements to the effect that average block size decreases with increasing depth (inverse grading) in block streams have rarely been supported by detailed measurements, although photographic evidence

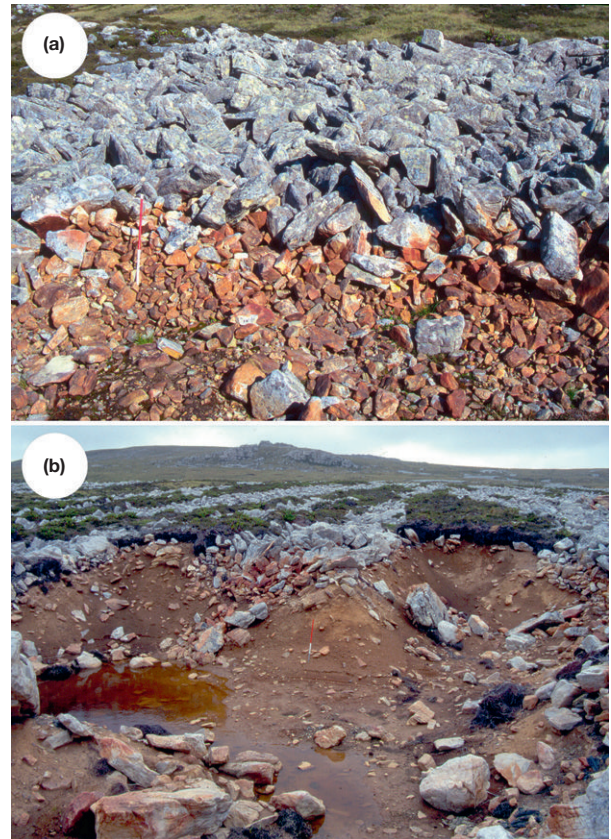


Figure 3 Vertical exposures beneath block streams. (a) Inverse grading and an absence of fine-grained material in a Falkland Islands stone run. Survey pole (left center) is 1 m in length. (b) Block stripes in the Falkland Islands. The central stripe has inverse grading and occupies a shallow depression. Note the relative lack of blocks in fine-grained material beneath adjacent vegetated ground and block stripes. Survey pole (center) is 1 m in length.



Figure 4 A 9-m-long block in a Falkland Islands stone run. Survey pole (center) is 1 m in length, and arrows indicate where the *a*-axis of the block is partly concealed by smaller blocks.

has been presented (**Boelhouwers, 1999; Clapperton, 1975**). Few studies have considered the downslope variation in size of surface blocks, and among those that have there is no general consensus. **Caine and Jennings (1968)** noted a reduction in

block size downstream, whereas [Potter and Moss \(1968\)](#) and [Boelhouwers et al. \(2002\)](#) found little variation in block size along block-stream axes.

Clast shapes have frequently been referred to but rarely quantified. Shape terminology includes tabular, platy, elongate, equant, slab, and diamond-shaped, angular, subangular, subrounded, and edge rounded, without clear definitions. Therefore, it is difficult to compare block streams in different locations. Some of the variation in clast size and shape is probably related to characteristics inherited from the parent rock, particularly the spacing of joints, and some may be a function of weathering and erosion during and since block accumulation.

Assessments of block-stream clast fabrics have yielded a wide range of results. A preferred downslope orientation of surface clasts is mentioned in several publications, but supporting fabric diagrams have not always been presented. Where such diagrams do occur, they show (1) blocks to be well oriented with either upslope or downslope dips ([Harris et al., 1998](#); [Potter and Moss, 1968](#)), (2) blocks to have a distinct cross-slope orientation or a random pattern in the vicinity of block-stream confluence sites ([Boelhouwers et al.,](#)

[2002](#); [Harris et al., 1998](#)), (3) blocks to have a primary downslope orientation and a secondary cross-slope orientation ([Boelhouwers et al., 2002](#); [Caine, 1983](#)), and (4) blocks with no preferred orientation ([Bennett et al., 1996](#)). In addition, blocks with a transverse orientation and upslope dip have been recorded by [Boelhouwers \(1999\)](#) at the front of a block stream, and a tendency for clasts to be arranged on edge has been noted in certain areas of some block streams ([Aldiss and Edwards, 1999](#); [Clapperton, 1975](#); [Clark, 1994](#)). Two-dimensional fabric diagrams presented by [Boelhouwers et al. \(2002\)](#) for a South African block stream are an indication of the variations that can occur in clast orientations ([Figure 5](#)). These variations are probably, in part, a reflection of the distance and mode of downslope transport.

Thermal Regime of Active Block Streams

Pronounced differences exist between the thermal regime of coarse blocky materials and adjacent mineral soils. Mean annual ground temperatures 4–7 °C cooler than in adjacent

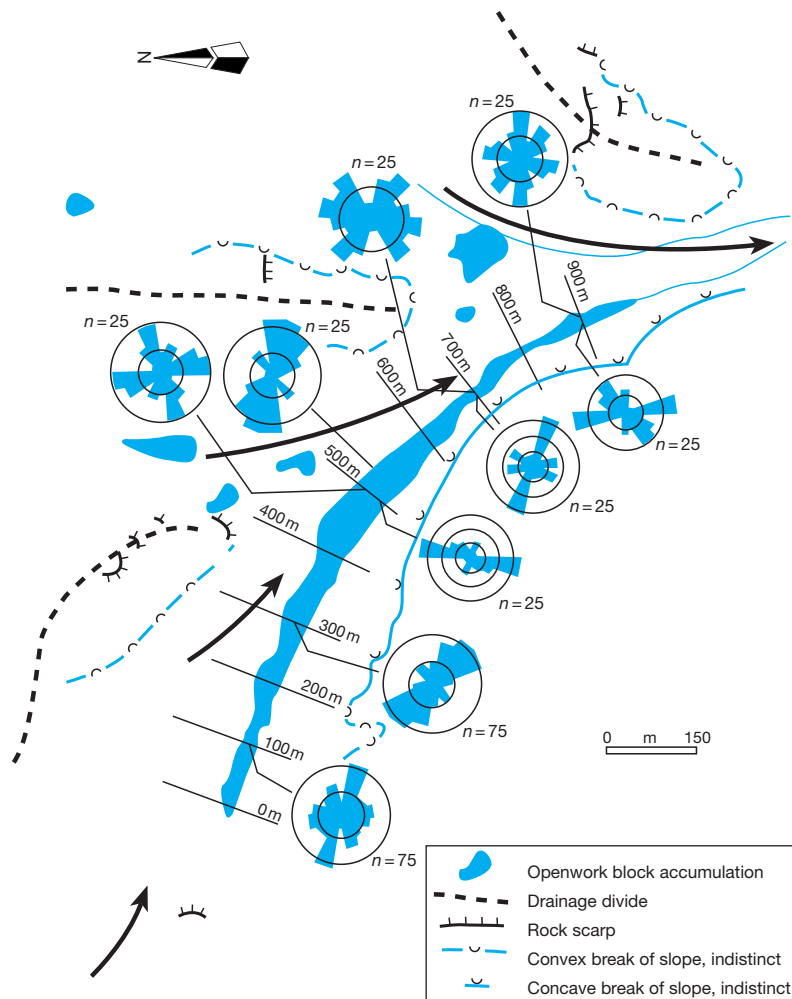


Figure 5 Two-dimensional fabric diagrams for surface blocks on a block stream in Lesotho. Reproduced from [Boelhouwers JC, Holness S, Meiklejohn I, and Sumner P \(2002\) Observations on a block stream in the vicinity of Sani Pass, Lesotho Highlands, Southern Africa. *Permafrost and Periglacial Processes* 13: 251–257. © John Wiley & Sons.](#)

mineral soils were recorded by [Harris and Pedersen \(1998\)](#) during a 2-year period in a block stream in southwestern Alberta. The cause of the disparity was attributed to continuous air exchange between the atmosphere and the block-stream void space, assisted by strong winds that favored evaporative cooling. This mechanism was thought to explain the occurrence of permafrost and ice masses in blocky deposits below the margin of regional continuous permafrost.

In central Asia, [Gorbunov et al. \(2004\)](#) found that mean annual air temperatures (MAATs) inside blocky deposits were 2.5–4 °C cooler than the air temperature above the surface. It was concluded that in such situations, permafrost could develop under positive MAATs of 4 °C. Similarly, [Guodong et al. \(2007\)](#) have demonstrated that in periglacial areas, coarse rock layers can behave like a ‘thermal semiconductor’ and lower the temperature of the underlying ground. Cooling is attributed to the ‘chimney effect’ (a natural form of convection driven by density buoyancy) and wind-forced convection. As well as having significant engineering applications, these findings have implications for understanding the processes involved in the creation of relict block streams.

Processes of Block-Stream Formation

Block streams are a consequence of several processes acting in combination or succession. [Harris et al. \(1998\)](#) found that blocks on an active block stream were derived by frost shattering of rock outcrops and by frost heaving from till. The downslope movement of active block streams has been attributed to the creeping and sliding of blocks in association with void-space ice. During the winter, ice forms in block streams and accompanying frost heave and thaw settlement cause some downslope movement of blocks. Other movement occurs when near-surface ice and snow begin to melt; friction between blocks is reduced and lubrication by water enables them to slide across the underlying ice-covered/cemented blocks ([Harris et al., 1998](#); [Romanovskij and Tyurin, 1986](#)).

The processes that contributed to the development of relict block streams have generated considerable discussion. A four-stage evolution has been envisaged, comprising (1) the mechanism(s) of block production, (2) downslope movement by mass wasting, (3) frost heaving and vertical sorting of blocks during transport, and (4) immobilization of the blocks as a consequence of the removal of interstitial fine-grained material. Elements of this model were proposed by [Andersson \(1906, 1907\)](#) based on his observations of block streams in Bear Island and the Falkland Islands, and from which he introduced the term solifluction to explain the downslope movement of blocks.

It has frequently been inferred that blocky debris results from the frost shattering (macrogelivation) of rock outcrops. This is because many blocks are very angular and, in some areas, can be traced upslope to degraded outcrops around which openwork accumulations of the most recently released blocks occur ([Boelhouwers, 1999](#); [Clark, 1972](#)). The absence of evidence for deep chemical weathering at such sites has served to reinforce the view that block production was a function of macrogelivation. However, very angular blocks occur in regions that have never experienced periglacial climates; in these situations, thermal stress

fracturing may account for the angularity ([Boelhouwers, 2004](#)). Therefore, block angularity cannot be taken as a diagnostic characteristic of macrogelivation. Rounded blocks have also been observed in block streams, and their origin as corestones, exhumed from deep chemically weathered mantles, has been proposed by [Caine and Jennings \(1968\)](#), [Caine \(1983\)](#), and [Boelhouwers et al. \(2002\)](#). Block fields on some Norwegian mountain plateaus consist of rounded blocks, and evidence for chemical weathering at these sites is given by [Rea et al. \(1996\)](#). Thus, block production may result from at least three different processes and, in certain areas, it is possible that each has contributed material to block streams.

Following [Andersson \(1906, 1907\)](#), some researchers have attributed the downslope movement of relict block streams either to solifluction or to the less specific process of mass wasting ([Aldiss and Edwards, 1999](#); [Boelhouwers, 1999](#); [Boelhouwers et al., 2002](#); [Caine, 1983](#); [Caine and Jennings, 1968](#); [Clapperton, 1975](#); [Potter and Moss, 1968](#); [Smith, 1949, 1953](#)). At most locations, it has been considered that the blocks moved in association with a matrix of fine-grained sediment. Indeed, [Aldiss and Edwards \(1999\)](#) considered it “mechanically unfeasible that stone runs could have moved without any matrix to separate the blocks.” However, at the head of some block streams, there are block piles adjacent to their source outcrops, and neither blocks nor outcrops show evidence for fine material ever having been present ([Boelhouwers, 1999](#); [Caine and Jennings, 1968](#); [Clark, 1972, 1994](#)). In these cases, frost wedging and heaving may have caused block movement away from outcrops, especially on very gentle slopes.

[Caine \(1983\)](#) reviewed two mechanisms of movement for block streams in Tasmania: a frictional mechanism and a viscous flow mechanism. These were not viewed as being mutually exclusive; both could have been involved in varying degrees. For a frictional model to apply, finer sediments and/or ice must have been present within the block-stream void space in order to facilitate high artesian pressures and shearing. Deep seasonal freezing and impervious weathered clays beneath block streams were thought to favor this mechanism, and permafrost was considered to be unnecessary. A viscous flow model based on frost heave and creep was thought capable of inducing some downslope block transport, whereas alternative viscous models were not likely to account for block movement. In conclusion, [Caine \(1983\)](#) considered that most of the movement resulted from sliding failure in remolded clays beneath the block streams when seasonal freezing occurred and high pore water pressures developed during thawing. Frost creep was regarded as a secondary process but one that could account for some block-stream fabrics.

The inverse grading evident in many block streams is generally attributed to block sorting as a consequence of frost heave during downslope transport ([Aldiss and Edwards, 1999](#); [Boelhouwers, 1999](#); [Potter and Moss, 1968](#)). Given the size of many surface blocks, it has been argued that heaving and sorting are likely to have been of greatest efficacy where block streams possessed a frost-susceptible matrix of fine sediment.

The absence of a fine-grained matrix from the upper thicknesses of relict block streams has been attributed to its removal following the cessation of downslope movement. Because running water can be heard and/or seen at the base of many block streams, it has been assumed that surface streams gradually

worked their way down through the blocks and removed the matrix. This has been referred to as the washing or flushing out of fines (Andersson, 1906, 1907; Boelhouwers, 1999; Boelhouwers et al., 2002; Caine, 1983; Caine and Jennings, 1968; Clapperton, 1975; Clark, 1994; Potter and Moss, 1968; Smith, 1949, 1953). Accumulations of fine-grained alluvium downstream of block fields were reported by Caine (1983) and Clark (1994), lending support to the notion of fluvial reworking.

Smith (1968) challenged the view that surface streams working downward had removed matrix materials. He proposed that piping (subsurface flow) was capable of removing material from situations and depths that could not be easily accounted for by surface flow, but this was recognized to be a speculative hypothesis requiring further testing.

In contrast, Aldiss and Edwards (1999) favored a process of matrix replacement by ice during freezing and flushing out of matrix by water during thaw. Because blocks have a greater thermal conductivity and a lower heat capacity compared to those of the matrix, it was argued that a coating of ice would develop around the blocks as the freezing front advanced into the block stream. Thus formed, the ice was thought to push the matrix particles away from the blocks, effectively separating the coarse and fine components. During times of thaw, some fine-grained material was likely to be flushed away by water and smaller blocks would be transported by water or gravity to a greater depth in the block stream, explaining the inverse grading that has frequently been reported. This mechanism, if valid, would have operated during downslope movement rather than following cessation of movement. Ultimately, an ice-bound block stream would develop and continue to move downslope until such time as the climate ameliorated.

The fact that block streams often possess strongly developed fabrics has been taken as evidence for downslope movement and fabric development prior to the removal of matrix (Boelhouwers and Sumner, 2003). However, it has not been explained satisfactorily how the fabric was retained when the removal of matrix seems likely to have resulted in some rearrangement of blocks (Potter and Moss, 1968). Strongly developed fabrics in an active block stream, consisting of blocks without interstitial material but with a sandy loam beneath, have been reported by Harris et al. (1998). Therefore, not all block streams require a fine-grained matrix in order for such fabrics to form. It appears that the development of fabrics in both active and relict block streams is an aspect that has not been fully elucidated.

From examination of some of the block streams of the Falkland Islands, André et al. (2008) proposed an alternative to the periglacial model of formation (Figure 6). They argued, from micromorphological, scanning electron microscopy (SEM), X-ray diffraction (XRD), and grain-size analyses, that the block streams are complex polygenetic landforms that have evolved through a six-stage development sequence. First, they suggested that the fine-grained material observed beneath the block streams was a product of intense and/or prolonged subtropical chemical weathering because the material had characteristics similar to those recorded in quartzite-derived tropical soils in Africa and South America. A Miocene age (23–5 Ma) for the weathering and regolith production was envisaged. Second, in the late Miocene–Early Quaternary (10–2 Ma), solifluction and debris sliding led to stripping of the upper part of the regolith from hillcrests and slopes,

resulting in accumulation of the material on the valley floors. Third, a phase of soil development then occurred, possibly associated with the temperate conditions of an early Quaternary interglacial stage. In the fourth stage, stripping of the remainder of the hillside regolith occurred. This material, essentially the middle and lower parts of the chemically weathered profile, consisted in down-profile sequence of cobbles embedded in fine-grained material and then larger blocks of quartzite; the latter now form the upper part of the block streams. André et al. (2008) were uncertain as to whether this phase of regolith stripping had occurred as a result of gelifluction under periglacial conditions or by other (nonperiglacial) mass-movement processes. Nevertheless, the removal (washing out) of the fine material was believed to have been either concurrent with or subsequent to debris emplacement in Quaternary cold episodes (fifth stage of the development sequence), and periglacial processes (frost heave and frost sorting) were implicated to account for some rearrangement of block-stream boulders in the final (sixth) stage. In essence, André et al. (2008) consider Falkland block streams to have strong affinities to inverted weathering profiles, with periglacial processes having played a relatively minor role at a late stage in their evolution.

The competing models of block-stream formation find similarities in the alternative models proposed to account for relict autochthonous block-field development. However, Ballantyne (2010) argued that evidence supporting both a pre-Quaternary chemical weathering hypothesis and a Quaternary periglacial (mechanical) hypothesis for block fields can be accommodated within a single evolutionary framework. Although plateau block fields were the focus of this inclusive model, the surface lowering envisaged by stripping of chemically weathered regolith cover and downslope transport by mass movement, followed by frost wedging of bedrock and downslope movement of blocks by frost creep and/or gelifluction may, in some circumstances, explain the occurrence of hillside and valley floor block streams. As with the hypothesis advanced by André et al. (2008), a variety of processes operating over a long period would therefore account for block-stream formation.

Rates of Movement of Active Block Streams

Harris et al. (1998) provided the most detailed results concerning rates of block movement on an active block stream. Seven lines painted across slope on a block stream 4800 m above sea level on a southwest-facing slope of 31° in Qinghai Province, China, were monitored for up to 5 years. MAAT of the site was estimated to be −6 °C and mean annual precipitation (MAP) 320 mm. Block movement was found to vary both across and along the block stream, with movement being greatest in the center and least near the source and the toe. Movement was only noted for surface blocks; in the subsurface, abundant plant stems were taken to indicate block stability. The maximum amount of movement of an individual block during a 12-month period was 2.1 m, but the maximum mean rate of movement along any line was considerably less at up to 0.6 m per year. Marked variations in mean rates of movement were recorded for most of the lines as well as between individual lines. A seasonal pattern to block movement was reported,

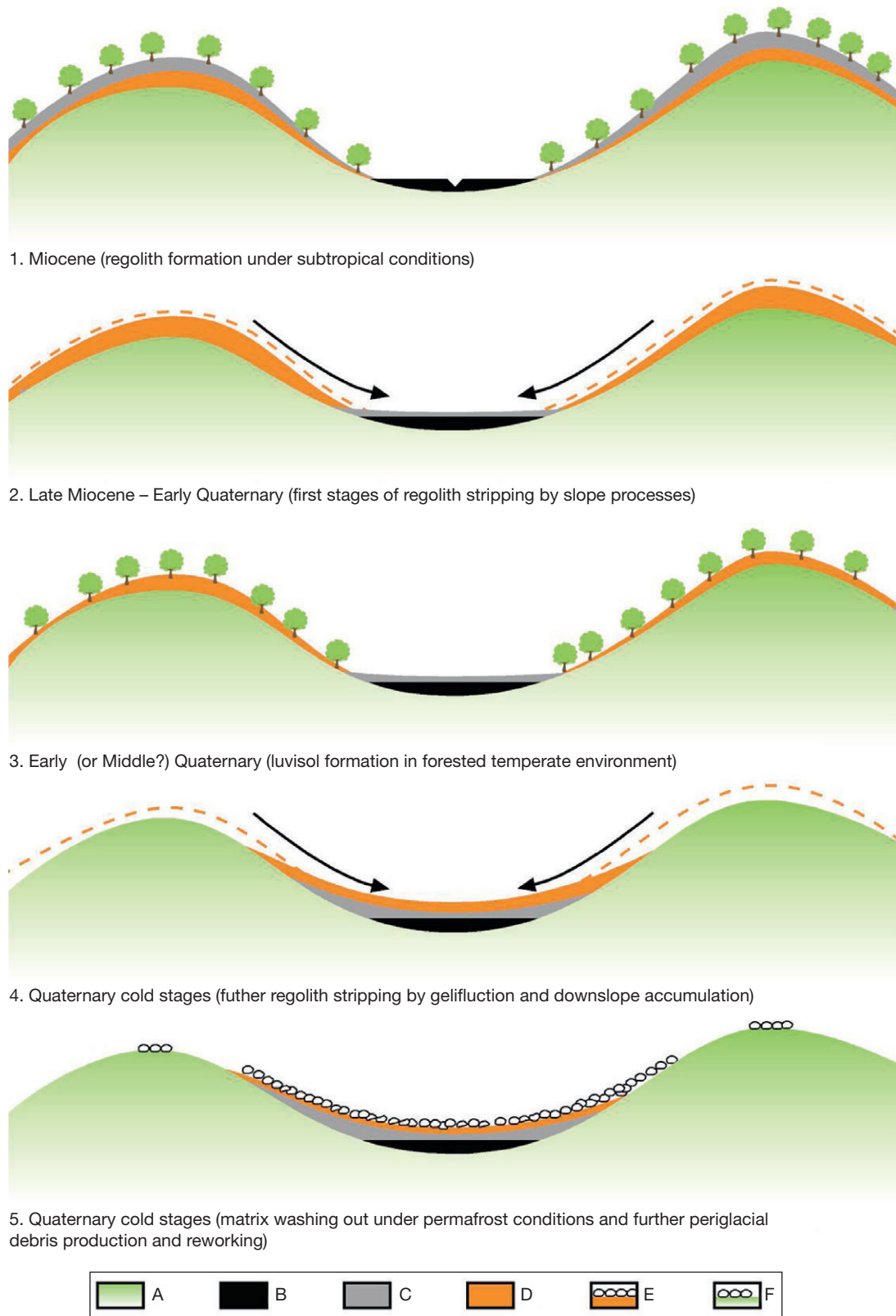


Figure 6 Tentative scenario of block-stream formation in the Falkland Islands. A: Bedrock (mainly quartzite with interbedded siltstones and mudstones). B: Alluvial deposits. C: Grus (initially top of regolith). D: Block-rich material with fine-grained matrix (initially bottom of regolith). E: Valley-bottom block stream derived from washing out of matrix in the upper parts. F: Hillslope block field. Reproduced from André MF, Hall K, Bertran P, and Arocena J (2008) Stone runs in the Falkland Islands: Periglacial or tropical? *Geomorphology* 95: 524–543. © Elsevier.

with most occurring between October and May when the ground was frozen but lacking a continuous snow cover. Some movement also occurred in June and July, probably in association with snow and freezing temperatures, but movement was minimal in August and September. Downslope movement resulted in block rotation, and small lobate clusters of moved blocks were observed close to the toe.

Climatic and Paleoclimatic Significance of Block Streams

Harris (1994) assessed the relationship of active block streams in China, Siberia, and North America to key climatic parameters. It was shown that they are associated with cold, dry climates (MAAT range, -6°C to -20°C ; MAP range, 50–500 mm) and occupy a discrete climatic zone that has limited overlap with zones characterized by active rock glaciers and active gelifluction landforms (Figure 7). However, the climatic conditions that currently favor active block streams are not necessarily the same as those under which relict block streams developed. Furthermore, it was argued that these zones display variations with respect to latitude, altitude, and aspect, with block streams being associated with higher latitudes and altitudes and with southwest and south slope aspects.

Attempts have been made to draw paleoclimatic inferences from relict block streams because they are thought to have undergone downslope movement associated with periglacial climatic conditions. Boelhouwers (1999) estimated that a MAAT of approximately 0°C was necessary for block-stream development in the Hex River Mountains, South Africa, implying a reduction of 7 or 8°C from present-day MAAT. Permafrost was not considered a prerequisite for block-stream mobilization; rather, deep seasonal frost penetration to 2-m depth was favored. In turn, this implies the lack of a sufficiently thick insulating snow cover during the autumn/winter freezing period. Again, however, the parameters that governed block-

stream development in one location are unlikely to have been identical to those that operated elsewhere. If more than one mechanism of downslope movement can account for block streams, then it may be the case that these mechanisms are controlled to a degree by the local climatic regime.

Age of Relict Block Streams

Many relict block streams occur in areas that lay adjacent to but beyond the margins of the (global) Last Glacial Maximum (LGM; 23 ± 4 ka). It has been inferred that these areas were characterized by severely cold periglacial climates at that time, and that block-stream formation represents one response of the landscape to those conditions (Boelhouwers, 1999; Boelhouwers et al., 2002; Clapperton, 1975; Clark, 1972, 1994; Potter and Moss, 1968).

A maximum age for block-stream formation in the Snowy Mountains, southeastern Australia, was provided by a ^{14}C age of $35.2 \pm 1.6 / -2.15$ ka obtained from *Nothofagus* wood from beneath a block stream (Caine and Jennings, 1968). Block-stream formation was attributed to a subsequent cold stage, coeval with the LGM. Caine (1983) reported ^{14}C dates of 3.08–4.7 ka from the base of soil and peat developed on block fields in Tasmania. These were taken to indicate that the block fields (and by inference the block streams) had been stable since at least the mid-Holocene. Using weathering rind thickness measurements of block-field clasts, Caine (1983) calculated that the block fields had probably been stable for approximately the past 13 ka.

Cosmogenic isotope (^{36}Cl) surface exposure dating of block streams and block slopes in southeastern Australia has shown a clustering of dates within the period of the LGM (Barrows et al., 2004). A weighted mean age of 21.9 ± 0.5 ka was obtained and is broadly equivalent to the timing of the maximum ice advance during the LGM as determined from cosmogenic dates from moraines in the region. These dates support earlier age estimates for relict block streams based on morphostratigraphy, ^{14}C assays, and weathering rind data. However, two boulders returned ages of 157 and 498 ka respectively, indicating that not necessarily all block-stream boulders were produced during the LGM.

Both optically stimulated luminescence (OSL) dating and cosmogenic isotope (^{10}Be and ^{26}Al) surface exposure dating have been applied to the block streams of the Falkland Islands. Hansom et al. (2008) collected samples for OSL dating from fine-grained materials from beneath and close to the downslope margins of block streams in order to constrain the emplacement age of the fine sediments and to provide maximum ages for deposition of the overlying blocks. Age estimates ranging from >54 to 16 ka were obtained and suggest that some block streams were probably active throughout the LGM and remained so until 16 ka or later.

The Falkland Islands block-stream dating undertaken by Wilson et al. (2008) involved sixteen ^{10}Be and ^{26}Al exposure ages and the results stand in marked contrast to the OSL dates reported by Hansom et al. (2008). Samples for dating were taken from one of the largest block streams (Prince's Street: ~ 4 km long and up to 400 m wide) and the smaller Estancia Brook block stream (~ 2.5 km long, 250 m wide), and were drawn

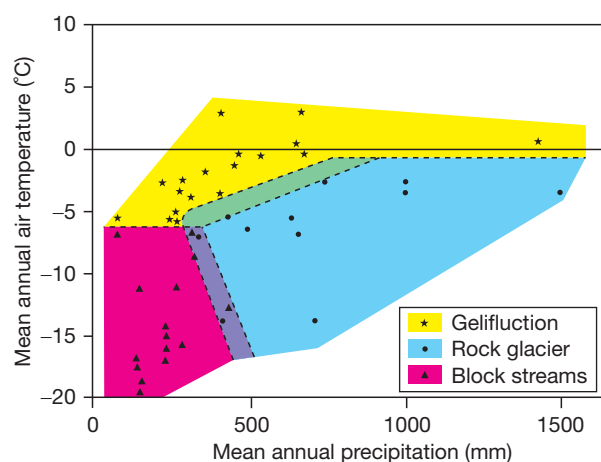


Figure 7 Relationship between climatic parameters (mean annual air temperature and mean annual precipitation) and distribution of gelifluction landforms, active rock glaciers, and active block streams in alpine landscapes. Reproduced from Harris SA (1994) Climatic zonality of periglacial landforms in mountain areas. *Arctic* 47: 184–192. © Arctic Institute of North America.

from different components of block-stream terrain: (1) valley-axis block accumulations, (2) narrow hillslope block streams that converge downslope with the valley-axis blocks and represent relict flow lines of block transport, and (3) hillcrest outcrops from which the blocks derived.

The range of ages obtained (731–42 ka) indicate that the samples cover a long period, at least 690 ka, and most have isotopic concentrations consistent with a simple exposure history. Ages of valley-axis blocks from Prince's Street are generally older than valley-axis blocks from Estancia Brook, and valley-axis blocks are generally older than samples derived from hillslope block streams. Two samples from hillcrest outcrops have isotopic concentrations indicative of a complex exposure history, most likely due to periods of burial and/or shielding. Overall, the ages demonstrate that the block streams date from more than one cold stage and can be considered as composite features that most likely formed over several cold stages. The block streams are long-lived landforms, as suggested by Aldiss and Edwards (1999). But just how close the earliest age obtained is to the start of block-stream activity is not known. In marked contrast to the OSL ages reported by Hansom et al. (2008), and some of the cosmogenic isotope ages for Australian block streams (Barrows et al., 2004), the ^{10}Be and ^{26}Al data of Wilson et al. (2008) provide no evidence for block-stream activity during the LGM.

The ages reported by these studies constitute a major advance in establishing a time scale for block-stream development, and may stimulate further and focused investigations on these enigmatic landscape features.

See also: [Permafrost and Periglacial Features: Blockfields \(Felsenmeer\)](#); [Cryoturbation Structures](#); [Permafrost](#); [Rock Weathering](#).

References

- Aldiss DT and Edwards EJ (1999) *The Geology of the Falkland Islands*. British Geological Survey Technical Report WC/99/10.
- Andersson JG (1906) Solifluction: A component of subaerial denudation. *Journal of Geology* 14: 91–112.
- Andersson JG (1907) Contributions to the geology of the Falkland Islands. *Wissenschaftliche Ergebnisse der Schwedischen Südpolar-Expedition, 1901–1903* 3: 1–38.
- André MF, Hall K, Bertran P, and Arocena J (2008) Stone runs in the Falkland Islands: Periglacial or tropical. *Geomorphology* 95: 524–543.
- Ballantyne CK (2010) A general model of autochthonous blockfield evolution. *Permafrost and Periglacial Processes* 21: 289–300.
- Barrows TT, Stone JO, and Fifield LK (2004) Exposure ages for Pleistocene periglacial deposits in Australia. *Quaternary Science Reviews* 23: 697–708.
- Bennett MR, Mather AE, and Glasser NF (1996) Earth hummocks and boulder runs at Merrivale, Dartmoor. In: Charman DJ, Newham RM, and Croot DG (eds.) *The Quaternary of Devon and east Cornwall: Field Guide*, pp. 81–96. London: Quaternary Research Association.
- Boelhouwers JC (1999) Relict periglacial slope deposits in the Hex River Mountains, South Africa: Observations and paleoenvironmental implications. *Geomorphology* 30: 245–258.
- Boelhouwers JC (2004) New perspectives on autochthonous blockfield development. *Polar Geography* 27: 67–80.
- Boelhouwers JC, Holness S, Meiklejohn I, and Sumner P (2002) Observations on a blockstream in the vicinity of Sani Pass, Lesotho Highlands, Southern Africa. *Permafrost and Periglacial Processes* 13: 251–257.
- Boelhouwers JC and Sumner PD (2003) The paleoenvironmental significance of southern African blockfields and blockstreams. In: Phillips M, Springman SM, and Arenson LU (eds.) *Proceedings of the 8th International Conference on Permafrost*, pp. 73–78. Lisse, The Netherlands: Swets & Zeitlinger.
- Caine N (1983) *The Mountains of Northeastern Tasmania*. Rotterdam, The Netherlands: Balkema.
- Caine N and Jennings JN (1968) Some blockstreams of the Toolong Range Kosciusko State Park, New South Wales. *Journal and Proceedings, Royal Society of New South Wales* 101: 93–103.
- Clapperton CM (1975) Further observations on the stone runs of the Falkland Islands. *Biuletyn Peryglacjalny* 24: 211–217.
- Clark R (1972) Periglacial landforms and landscapes in the Falkland Islands. *Biuletyn Peryglacjalny* 21: 33–50.
- Clark R (1994) Tors, rock platforms and debris slopes at Stiperstones, Shropshire, England. *Field Studies* 8: 451–472.
- Firpo M, Guglielmin M, and Queirolo C (2006) Relict blockfields in the Ligurian Alps (Mount Beigua, Italy). *Permafrost and Periglacial Processes* 17: 71–78.
- Gorbunov AP, Marchenko SS, and Seversky EV (2004) The thermal environment of blocky materials in the mountains of central Asia. *Permafrost and Periglacial Processes* 15: 95–98.
- Guodong C, Yuanming L, Zhizhong S, and Fan J (2007) The 'thermal semi-conductor' effect of crushed rocks. *Permafrost and Periglacial Processes* 18: 151–160.
- Hansom JD, Evans DJA, Sanderson DCW, Bingham RG, and Bentley MJ (2008) Constraining the age and formation of stone runs in the Falkland Islands using Optically Stimulated Luminescence. *Geomorphology* 94: 117–130.
- Harris SA (1994) Climatic zonality of periglacial landforms in mountain areas. *Arctic* 47: 184–192.
- Harris SA, Cheng G, Zhao X, and Yongqin D (1998) Nature and dynamics of an active block stream, Kunlun Pass, Qinghai Province, People's Republic of China. *Geografiska Annaler* 80A: 123–133.
- Harris SA and Pedersen DE (1998) Thermal regimes beneath coarse blocky materials. *Permafrost and Periglacial Processes* 9: 107–120.
- Nelson KJP, Nelson FE, and Walegur MT (2007) Periglacial Appalachia: Palaeoclimatic significance of blockfield elevation gradients, eastern USA. *Permafrost and Periglacial Processes* 18: 61–73.
- Perov VF (1969) Block fields in the Khibiny Mts. *Biuletyn Peryglacjalny* 19: 381–387.
- Potter N and Moss JH (1968) Origin of the Blue Rocks block field and adjacent deposits, Berks County, Pennsylvania. *Geological Society of America Bulletin* 79: 255–262.
- Rea BR, Whalley WB, Rainey MM, and Gordon JE (1996) Blockfields old or new? Evidence and implications from some plateaus in northern Norway. *Geomorphology* 15: 109–121.
- Romanovskij NN and Tyurin AI (1986) Kurums. *Biuletyn Peryglacjalny* 31: 249–259.
- Smith HTU (1949) Periglacial features in the driftless area of southern Wisconsin. *Journal of Geology* 57: 196–215.
- Smith HTU (1953) The Hickory Run boulder field, Carbon County, Pennsylvania. *American Journal of Science* 251: 625–642.
- Smith HTU (1968) Piping in relation to periglacial boulder concentrations. *Biuletyn Peryglacjalny* 17: 195–204.
- White SE (1976) Rock glaciers and block fields, review and new data. *Quaternary Research* 6: 77–97.
- White SE (1981) Alpine mass movement forms (noncatastrophic): Classification, description and significance. *Arctic and Alpine Research* 13: 127–137.
- Wilson P, Bentley MJ, Schnabel C, Clark R, and Xu S (2008) Stone run (block stream) formation in the Falkland Islands over several cold stages, deduced from cosmogenic isotope (^{10}Be and ^{26}Al) surface exposure dating. *Journal of Quaternary Science* 23: 461–473.

Blockfields (Felsenmeer)

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Introduction

Blockfields are found on plateaux, summits, and ridges and have also been described as blockmeer, felsenmeer, boulder fields, stone fields, and mountain-top detritus. There has long been an interest in these landforms, and interpretations have varied over time as views on the erosional efficacy of Pleistocene glaciers and ice sheets and understanding of geomorphic processes have changed (Goddard, 1989). Blockfields are to be found predominantly in high to mid-latitudes (Arctic Canada, Greenland, Fennoscandia, Iceland, Svalbard, UK, and USA) and are the focus of discussion in this article, though they are present at lower latitudes, for example, the Lesotho Highlands and Namibia (Boelhouwers, 2004) and Tasmania (Caine, 1983). They are characterized, as their name suggests, by a block-covered surface, which may extend to rockhead, or overlie a clast- or matrix-supported profile. The matrix may comprise sand-, silt-, and clay-size fractions.

The main subject of this article is autochthonous blockfields (formed *in situ* and confined to tectonically stable environments; Goddard, 1989), though para-autochthonous blockfields may be of relevance but only if the material is sourced locally and from weathering products produced contemporaneously, for example, up-slope (Nesje, 1989; Nesje et al., 1988). As material transport distance increases, the transport mechanism itself defines the character of the landform, with boulder-banked solifluction lobes, or larger boulder/block streams, making them irrelevant to this article. True allochthonous blockfields, derived for example from an old till cover (Roaldset et al., 1982), are not considered here.

The contentious issues regarding blockfields relate to their age and origin. For blockfields in mid- to high latitudes, both issues are interlinked: that is, the age can to some degree dictate the origin, and vice versa (Ballantyne, 1998; Goddard, 1989; Nesje, 1989; Rea et al., 1996a,b; Roaldset et al., 1982). Age ranges of <10 ka to >2 Ma, for different blockfields, have been postulated in the literature, suggesting a chemical weathering origin, a frost weathering origin, or some combination of both. This range of ages and thus weathering origins may have significant implications for the dynamics and geometry of Pleistocene ice sheets. For example, a pre-Pleistocene age (>1.8 Ma) indicates nonglaciation or glaciation beneath ice that was cold-based and nonerosive or experienced minimal erosion. To add further complexity, blockfield-mantled surfaces are often found in juxtaposition with deeply incised glacial valleys. For these reasons, the identification and interpretation of the age and origin of blockfields is of particular interest, especially in middle to high latitudes, within the regions affected by Pleistocene ice sheets. Their study may provide details on paleoclimates, glacier/ice sheet geometries and dynamics, and landscape evolution over short or longer time periods. In the following sections, we present the key characteristics, age and origin, implications for ice dynamics and landscape evolution, paleoclimate, and future prospects for studies on blockfields.

Blockfield Characteristics

Surface

The formation/preservation of blockfields is obviously controlled at a first order by slope angles. As the slope angle increases, the downslope shear stress on a particle increases, initiating ever greater rates of movement, thus preventing blockfield formation. Dahl (1966) reports an upper limit of 25° for the presence of blockfield slopes; beyond this, talus slopes form, and Rea et al. (1996a) report blockfield cover on slopes of <10° on plateaux in north Norway. Blockfields may be found as extensive spreads (Figure 1(a)) often in association with tors (Dyke, 1993; Ives, 1958; Kleman and Borgstrom, 1990; Rea et al., 1996a; Sugden and Watts, 1977; Figure 1(b)) or small areas confined to summits and ridges (Figure 1(c)). Generally, the surface cover is either an openwork of blocks ranging from cobble through to boulder sizes (Figure 2(a)), or a similar blocky surface set in a matrix of fines (Figure 2(b)). An openwork blocky surface will tend to display crude sorting patterns with perhaps some imbricate boulders. Much better developed sorting is found on surfaces where there are more fines (Figure 3). The removal of fine-grained matrix sediments is more effective on steeper slopes, making openwork blockfields more prevalent as the slope angle increases (Rea et al., 1996a,b). Large boulder sizes favor the formation of an openwork blockfield. Lower slope angles will tend to retain fines and may accumulate them from up-slope. The exposed surfaces of blockfield boulders and associated tors tend to be heavily weathered, often showing a significant degree of breakdown through granular disintegration, and forms such as weathering pits may be common (Ballantyne, 1998; Dahl, 1966).

Profile

The depth of blockfields is obviously controlled by a number of variables, including age, slope angle, lithology, and weathering versus erosion rate. An average seems to be on the order of 1 m, with excavated depths exceeding 1.5 m in places without reaching rockhead (Ballantyne, 1998; Dahl, 1966; Rea et al., 1996a,b). Fjellanger et al. (2006) and Fjellanger and Sørbel (2007) observed depths >2 m and inferred depths of >3 m for the Varanger Peninsula, north Norway. The nature of the deposits also often precludes excavation to rockhead, making accurate depths difficult to determine, although admirable digging efforts have demonstrated that this can be achieved (Goodfellow et al., 2008).

Ballantyne (1998) reported three primary types of blockfield profile:

- Type 1 is clast-supported, and has an openwork blocky surface with fines forming a matrix at some depth below the surface (Figures 4(a), 1(b) foreground and Figure 2(a)).
- Type 2 is matrix-supported, and the surface has blocks set in a cohesionless sandy matrix that continues down through the profile (Figure 4(b)).



Figure 1 (a) Typical blockfield-mantled plateau surface on the summit of Fjelltindnase, Øksfjordjøkelen, north Norway. Reprinted from Rea BR, Whalley WB, Rainey MM, and Gordon JE (1996) Blockfields, old or new? Evidence and implications from some plateaus in northern Norway. *Geomorphology* 15: 109–121. (b) Blockfield-mantled slope running up to a tor (Øksfjordjøkelen, north Norway). Reprinted from Rea BR, Whalley WB, and Porter EM (1996) Rock weathering and the formation of summit blockfield slopes in Norway: Examples and implications. In: Anderson MG and Brooks SM (eds.) *Advances in Hillslope Processes*, pp. 1257–1275. Chichester: Wiley. (c) Flat-topped blockfield-mantled ridge (1420 m) on the northern side of Kangigdleq, Uummannaq Region, west Greenland, with the calving margin of Rink Isbræ in the background. Photograph by B. R. Rea.

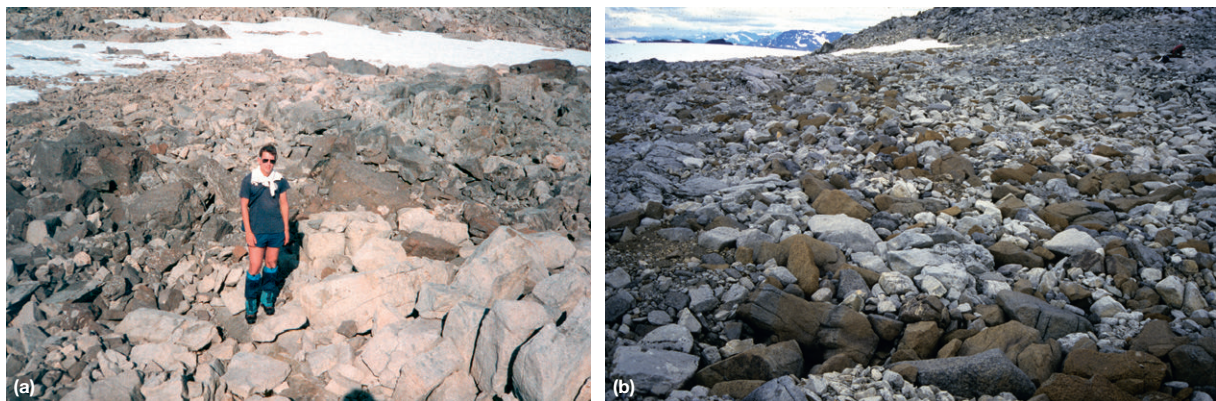


Figure 2 (a) Openwork type 1 blockfield surface (Øksfjordjøkelen, north Norway). Photograph by B.R. Rea. (b) Mixed blockfield, with openwork surface (type 1) and blocks set within matrix (left side of photograph) type 3 (Øksfjordjøkelen, north Norway). Photograph by B.R. Rea.

- Type 3 is matrix-supported, and the surface has blocks set in a silt/clay matrix (here silt/clay is used in order to be more comprehensive than Ballantyne's (1998) silt matrix) and shows evidence for frost sorting (Figures 4(c) and 3).

Particle Size

In autochthonous blockfields, the particle size is controlled by the lithology of the underlying bedrock and by weathering

mechanisms. At the boulder scale, the joint spacing in the bedrock is the controlling factor, and larger joint spacings produce larger blocks, while smaller joint spacings produce smaller blocks, irrespective of weathering process (mechanical or chemical). Joint spacing will also control the weathering rate under any weathering regime, as smaller joint spacings provide a greater surface area of rock susceptible to water ingress. Massive lithologies and outcrops are less susceptible to weathering (Ballantyne, 1998). Analysis of the matrix sediments potentially provides a more diagnostic fingerprint regarding

the degree of chemical alteration, and thus on the weathering history. This is most applicable to coarse-grained igneous and metamorphic rocks, as sedimentary rock mineralogies are closer to equilibrium with conditions (temperature and pressure) at the Earth's surface (Bowen's reaction series). Additionally, sedimentary rocks may already contain mineral assemblages diagnostic of former weathering environments, but not those responsible for their most recent breakdown. Fine-grained metamorphic rocks may suffer similar inheritance problems.

Mechanical weathering of exposed rock surfaces and joint-controlled blocks will proceed by granular disintegration. The mode of mechanical weathering is irrelevant: it may occur through classic freeze-thaw (expansion and contraction) of *in situ* porewater, or via stressing and enhanced ice growth by porewater migration to a freezing front (segregation ice formation), or through thermal stress/shock produced by rapid surface temperature fluctuations (Hall et al., 2002). Lines of mechanical weakness are defined mostly by intercrystalline microcracks and thus exert control on the particle-size range of the fines produced. The minimum grain size in the fines is equivalent to the minimum crystal size, while larger grains are composed of multiple crystals. Mechanical weathering is unable to reduce the particle size further. Indeed, experimental studies and field evidence suggest that many lithologies are extremely resistant to mechanical weathering below the joint scale, exemplified by the preservation of striae and friction cracks on bedrock surfaces following subaerial exposures of >10 ka (Ballantyne, 1998; Lautridou and Seppälä, 1986). Thus, mechanical action tends to produce matrix sediments that are sand-sized and larger (Rea et al., 1996b).

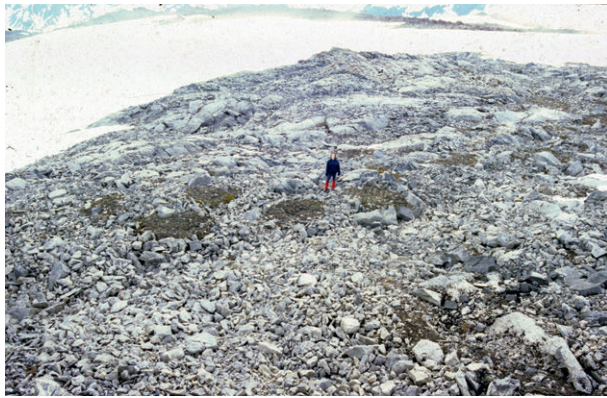


Figure 3 A mixed blockfield surface, with type 3 showing well-developed patterned ground and type 1 (in the bottom center/left) showing crude sorting (Øksfjordjøkelen, north Norway). Photograph by B.R. Rea.



Figure 4 (a) Type 1 blockfield, developed on fine-grained Cambrian Quartzite, An Teallach, Scotland. (b) Type 2 blockfield developed on Torridon Sandstone, An Teallach. (c) Type 3 blockfield, developed on Moine Schist, Fannich Mountains. Reproduced from Ballantyne CK (1998) Age and significance of mountain-top detritus. *Permafrost and Periglacial Processes* 9: 327–345.

Chemical weathering can reduce particle sizes below the crystal size of the parent bedrock, with the silt (63–2 mm) and clay (<2 mm) size fractions increasing as the degree of alteration proceeds (van Wambeke, 1962). The reduction in size is effectively the long-term result of destabilization and dissolution of the original mineral, with only a portion remaining *in situ* as a clay mineral. Based on original French terminology, Hall et al. (1989) present a tripartite granulometry classification most applicable to coarse-grained rocks, which can be used as a proxy for the degree of chemical alteration (Table 1). Although mostly applied to low-latitude

Table 1 Granulometric classification, applicable to coarse-grained rocks, indicating the different degrees of chemical weathering

Classification	Median grain size (μm)	Silt content (%)	Clay content (%)	Increasing chemical alteration
Granular grus	>1000	<15	<3	
Grus	<1000	15–25	<6	
Clayey grus	<1000	>25	>6	↓

Adapted from Hall AM, Mellor A, and Wilson MJ (1989) The clay mineralogy and age of deeply weathered rock in north-east Scotland. *Zeitschrift für Geomorphologie, Supplementband 72*: 97–108.

in situ weathering profiles (Hall et al., 1989), this classification has been used to infer a dominant role for chemical weathering in the production of plateau blockfields in north Norway (Rea et al., 1996a). Figure 5(a) shows the particle-size data for two blockfield sites in North Norway, and indicates that Pit 1 (Figure 5(b)) is a type 3 clayey grus and Pit 3 (Figure 5(c)) is a type 3 grus/clayey grus (Rea et al., 1996a). Given sufficient time even in cold environments, chemical weathering can produce fines that are significantly enriched in silt- and clay-size fractions, compared with the parent bedrock.

Fine-Grained Matrix Geochemistry

The mineralogy of the weathering products is potentially a diagnostic tool for the weathering origin and thus the age of blockfields, but it must be treated with caution. Of greatest importance are the clays that result from chemical weathering processes occurring at the Earth's surface – not those produced by hydrothermal alteration or recrystallization at depth. It is also important to note that for sedimentary and metamorphic bedrock it is possible that weathering products from previous subaerial exposure may become re-exposed at the surface as a result of exhumation. For example, Fjellanger and Nystuen (2007) demonstrated a complex weathering history for quartzites on the Varanger Peninsula. During the Neoproterozoic,

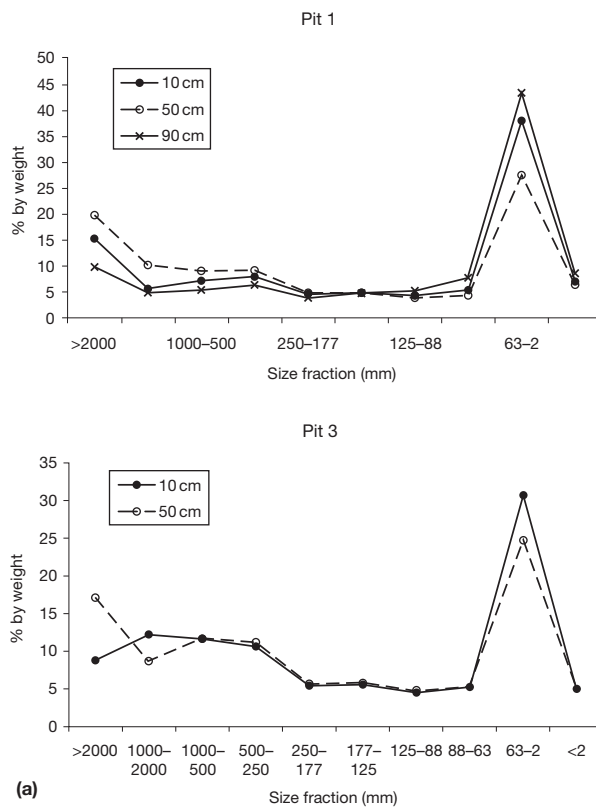


Figure 5 (a) Particle-size distributions of blockfield fines. Pit 1 is defined as a clayey grus, while Pit 3 is a grus/clayey grus, based on the classification shown in Table 1. (b) Pit 1 displays a matrix-supported profile and is representative of a type 3 blockfield. Note the water in the base of the pit and the overall depth. Rockhead has not been reached. (c) Pit 3 displays a matrix-supported profile and is representative of a type 3 blockfield. Rockhead has again not been reached. Reproduced from Rea BR, Whalley WB, and Porter EM (1996) Rock weathering and the formation of summit blockfield slopes in Norway: Examples and implications. In: Anderson MG and Brooks SM (eds.) *Advances in Hillslope Processes*, pp.1257–1275. Chichester: Wiley.

subaerial weathering produced clastic kaolinite and other mixed-layer clay minerals which coated quartz sand grains. Subsequent burial, to shallow depth, produced diagenetic kaolinite and illite, which are found today on the re-exposed 'paleic-surface.'

Controlling factors on the weathering process are:

1. parent lithology, which dictates the mineralogy;
2. climate, subdivided into precipitation and temperature;
3. topography, combined with (2), controlling the water to rock contact area (drainage); and
4. age, which controls the length of time exposed to weathering.

Parent lithology and the water/rock ratio control the clay minerals that are produced, temperature controls the reaction rates, and time determines how far along a particular path the weathering sequence has reached (Velde, 1992). The importance of time must be emphasized and is highlighted in Figure 6.

Even under unfavorable conditions of low reaction intensity and low temperature (e.g., at high latitudes/altitudes), time alone may progress the chemical weathering process toward clay mineral production. However, time has significant implications for the origin of blockfields in mid- to high latitudes, as discussed below.

Some generalizations are useful for understanding the formation and presence of clay minerals, which may be found in blockfield matrix sediments. Table 2 shows the major clay mineral types and associated climates, but it must be

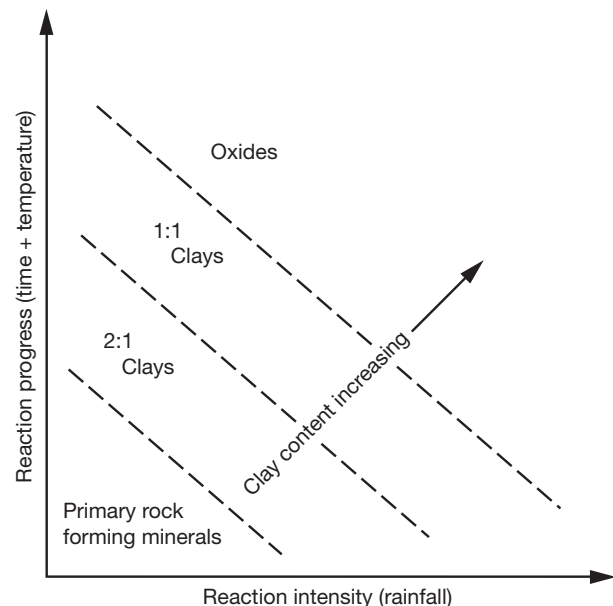


Figure 6 A weathering sequence shown as a function of reaction intensity and reaction progress. Given a higher reaction intensity (e.g., greater water/rock ratio), less time is required to progress from minerals through the weathering sequence. Alternatively, for a given reaction intensity, increasing age will ultimately progress the weathering process, while increasing temperature may counter a lower reaction intensity or decrease the exposure time. Reproduced from Velde B (1992) *Introduction to Clay Minerals: Chemistry, Origins, Uses and Environmental Significance*. London: Chapman and Hall.

remembered that local factors, including parent lithology, inheritance, and age, will all influence the actual minerals found in any location. Excluding dissolution of carbonate, in high-latitude/-altitude environments, low temperatures and often limited availability of moisture ensure that chemical weathering is curtailed. However, it is not absent, as surface and near-surface temperatures can be sufficient to allow chemical weathering, but the production rates are likely to be extremely low (Hall et al., 2002). This means simply that, even if hydrothermal, diagenetic, metamorphic, and exhumation of older weathering products have been excluded, the presence of clay minerals is not sufficient to interpret blockfield matrix sediments as inherited from a different (warmer) climatic regime. Rather, it is a combination of clay types and concentrations that better represent inheritance from a previous weathering climatic regime (e.g., Dahl, 1987; Fjellanger and Nystuen, 2007; Hall et al., 1989; Marquette et al., 2004; Rea et al., 1996a,b; Roaldset et al., 1982). Given that even in the coldest and driest of environments limited chemical and mechanical weathering will still occur, no matter how slowly, relict autochthonous blockfields do not really exist (Ballantyne, 2010a). It should also be mentioned that aeolian deposition could deliver clay minerals to a site, though no reports of such a source in blockfields are known from the literature (Goodfellow et al., 2008).

The application of more advanced geochemical interpretative techniques has begun to contribute to blockfield research (e.g., Goodfellow et al., 2008). These comprise a number of techniques whereby elemental mobility, elemental mass balance, and dimensionless estimates of the degree of weathering may be determined in order to elucidate *in situ* weathering activity and to enhance understanding of the potential longevity of the weathering history. Goodfellow et al. (2008) analyzed samples from two sites in the vicinity of Tarfala-tjärro, Northern Sweden. They used the isocron technique to demonstrate that there had been little movement of labile elements within the profile. This technique cross-plots elemental concentrations from the matrix and the bedrock, with scaling factors allotted to elements to ensure that they

Table 2 Idealized climate, weathering profile, and clay mineralogy relationships. It should be noted that the parent lithology, local conditions, and inheritance will control the actual weathering products found *in situ*

<i>Climate</i>	<i>Depth to rockhead (m)</i>	<i>Predominant clays</i>
Desert (hot and cold)	0.1	Mostly inherited (mica chlorite)
Semiarid steppe	1.0	Smectite (salts, carbonate concentrations, sepiolite-palygorskite)
Temperate	1.0	Illite, kaolinite, soil vermiculite, hydrated oxides (minor smectite)
Tropical wet	15	Kaolinite, hydrated iron aluminum oxides
Tropical dry	1.5	Smectite

Data from Velde B (1992) *Introduction to Clay Minerals: Chemistry, Origins, Uses and Environmental Significance*. London: Chapman and Hall.

are spread across the diagram. Immobile elements should plot on a straight line (i.e., no change in concentration between parent and weathered matrix), with elements enriched in the matrix plotting above and those that are depleted plotting below (Taylor and Eggleton, 2001). Goodfellow et al. (2008) also applied a mass balance analysis (mass of elements lost or gained by the matrix relative to the parent) to show that there had been only minor elemental losses from both the sites. Finally, they applied a chemical index of alteration (CIA – one of many possible such analyses; Taylor and Eggleton, 2001) analysis which demonstrated the absence of vertical weathering intensity (due to either limited chemical weathering and/or profile mixing due to periglacial activity, e.g., heave and sorting) and limited production of secondary weathering minerals. It also demonstrated that at least some additional matrix material had been added to the site because the CIA of the matrix was lower than that of the ‘parent’ bedrock; this was interpreted to be from up-slope.

Clast Angularity

The traditional view of cold-climate mechanical weathering is that it produces angular clast forms. However, granular disintegration in cold climates would appear to produce rounded forms (Ballantyne, 1998). Hall et al. (2002) and Boelhouwers (2004) argue that angularity of weathering products and landforms in cold climates is not universal, and in many instances rounded clasts and landforms are produced. This has important implications, as angularity of blocks and clasts cannot be taken as evidence of either a cold or a warm climate: rather, it represents a mechanical origin.

Grain-scale characteristics can be treated in the same manner as the larger material, so that mechanical grain textures are likely to be sharp and angular, while chemical signatures will etch and round grains by dissolution, alteration, and overgrowths. Visualization of blockfield matrix sediments is generally best achieved using techniques in scanning electron microscopy, and grain textures may be used as supporting evidence for weathering origin interpretations (Mellor and Wilson, 1989; Roaldset et al., 1982; Whalley et al., 1996).

Observations on half-buried clasts are extremely useful for determining the frost-heave activity in a blockfield. For example, if a clast sitting on the blockfield surface shows evidence of well-developed granular disintegration and rounding on the exposed (upper) surface, with an angular character on the subsurface, this indicates stability of the profile. In other words, the clast has been *in situ* for some considerable length of time for the granular disintegration to have developed (Ballantyne, 1998).

Age and Origin

It is important to note that the term ‘blockfield’ should only be applied as a generic descriptor of the landform and does not define genesis (age or weathering origin). There has been long-standing debate in the scientific literature regarding the age and origin of blockfields, which has focused on three potential scenarios, namely, recent (e.g., late- to postglacial; Dahl, 1966; Dredge, 1992), Pleistocene (Ballantyne, 1998; Kleman

and Borgstrom, 1990; Nesje et al., 1988), and pre-Pleistocene (Dahl, 1987; Dyke, 1993; Kleman and Stroeven, 1997; Nesje, 1989; Nesje et al., 1988; Roaldset et al., 1982).

Recent <10 ka

Of necessity, rapid formation is related to recently evolved blockfields but may also have applicability to old preserved forms, though this would be difficult to determine. Most early interpretations of blockfields assumed rapid formation based on the assumption that the last ice sheets were pervasively erosive, so any blockfields not sourced from till (Roaldset et al., 1982) must have formed during the postglacial period. Such views have been superseded with a greater understanding of ice dynamics and knowledge that glaciers and ice sheets can be entirely, or in part, cold-based and nonerosive and that distinct thermal boundaries may be represented in the landform record (Ballantyne, 2010b; Kleman and Borgstrom, 1990; Kleman and Stroeven, 1997). However, evidence exists that blockfields have formed rapidly, both in the Late Glacial and postglacial periods.

Dredge (1992) presents evidence for postglacial formation of carbonate blockfields on Igloolik Island and the Melville Peninsula, in the Eastern Canadian Arctic. Currently a permafrost climate, conditions are such that mechanical and chemical weathering (given the bedrock lithology) are active today. These assumptions are supported by field observations; for example, mudlines (linear rows of mud circles) associated with active-layer processes cross-cut 1-ka-old beach ridges (Dredge, 1992). The angular nature of the material in beach ridges up to 140 m a.s.l. has been used to cite rapid formation, potentially by freeze–thaw weathering in a shallow marine environment.

Ballantyne (1998) reports blockfields developed during the ‘Late Glacial’ in northwest Scotland between the limits of the Last Glacial Maximum ice sheet and the Loch Lomond Stadial (LLS) (Younger Dryas) ice-field limits; it should be noted, however, that some trimlines previously identified by Ballantyne and coworkers have now been reinterpreted as representing internal ice-sheet thermal boundaries, which may have implications for the formation age of some of the Scottish examples (Ballantyne, 2010b). Exposure to intense active-layer and periglacial processes was of limited duration, and formation is assumed to have ceased at the end of the LLS (i.e., there are no blockfields developed inside the limits of the LLS glaciers). Development is lithologically controlled, with thick blockfields found only on microgranite and quartzite, and only thin blockfields developed on some granites and schists (Ballantyne, 1998).

These two examples suggest that, given favorable lithologies, blockfields can develop rapidly in permafrost/periglacial climates, probably within a few millennia, becoming more developed with time (thicker). Importantly, in both examples, rapid development only occurs under cold-climate conditions, which should have an impact upon the particle-size range, quantity of fines, clay mineralogy, and clast angularity. Particle-size range should be controlled by granular disintegration, clay minerals (excluding inheritance) should be absent or representative of limited surface chemical alteration, and quantities should be low, and a type 1 openwork structure is expected,

given the limited production of fines. Account, of course, must be taken of the parent lithology; for example, Dredge (1992) reports a terminal grade of weathering products as 35% sand, 45% silt, and 20% clay, but this reflects the grain size of the original dolomitic limestone, namely, 30% fine sand, 50% silt, and 20% clay. Site 4 in Figure 7 (Ballantyne, 1998) shows a type 1 blockfield, while site 11 is a type 3 blockfield, both resulting from rapid development (Ballantyne, 1998). The relative abundance of fines at site 11 is assumed to reflect the grain size and joint spacing in the mica schist.

Pleistocene

Blockfield formation within the Pleistocene is of broad time frame, covering at least 1.8 Ma, so this hypothesis can be viewed as making more time available for weathering to progress. Given that no 'recent' blockfields show significant formation of clay minerals, it may be assumed that the major

agent of formation during the Pleistocene is mechanical weathering, with some minor component of chemical activity, presumably most intense during interglacials. Even for mid-latitude sites, based on Holocene interglacial weathering, the chemical weathering rate will be low (Goodfellow et al., 2008), although temperatures during the last interglacial were higher than the present one. It is expected that particle size will be controlled by granular disintegration, and that clay minerals (excluding inheritance) should be present but may be representative of limited chemical alteration depending upon the location, for example, high or mid-latitude (Goodfellow et al., 2008). Longer periods of exposure should produce more fines, but the size range will be controlled significantly by the bedrock grain size. Type 2 (most likely) or type 3 (lithology-dependent) profiles are expected, given the greater quantity of fines and role of frost action in sorting the profile.

This age range requires that the blockfield has survived through at least one glacial-interglacial cycle, and, provided the blockfield lies within the limits of one or more Pleistocene

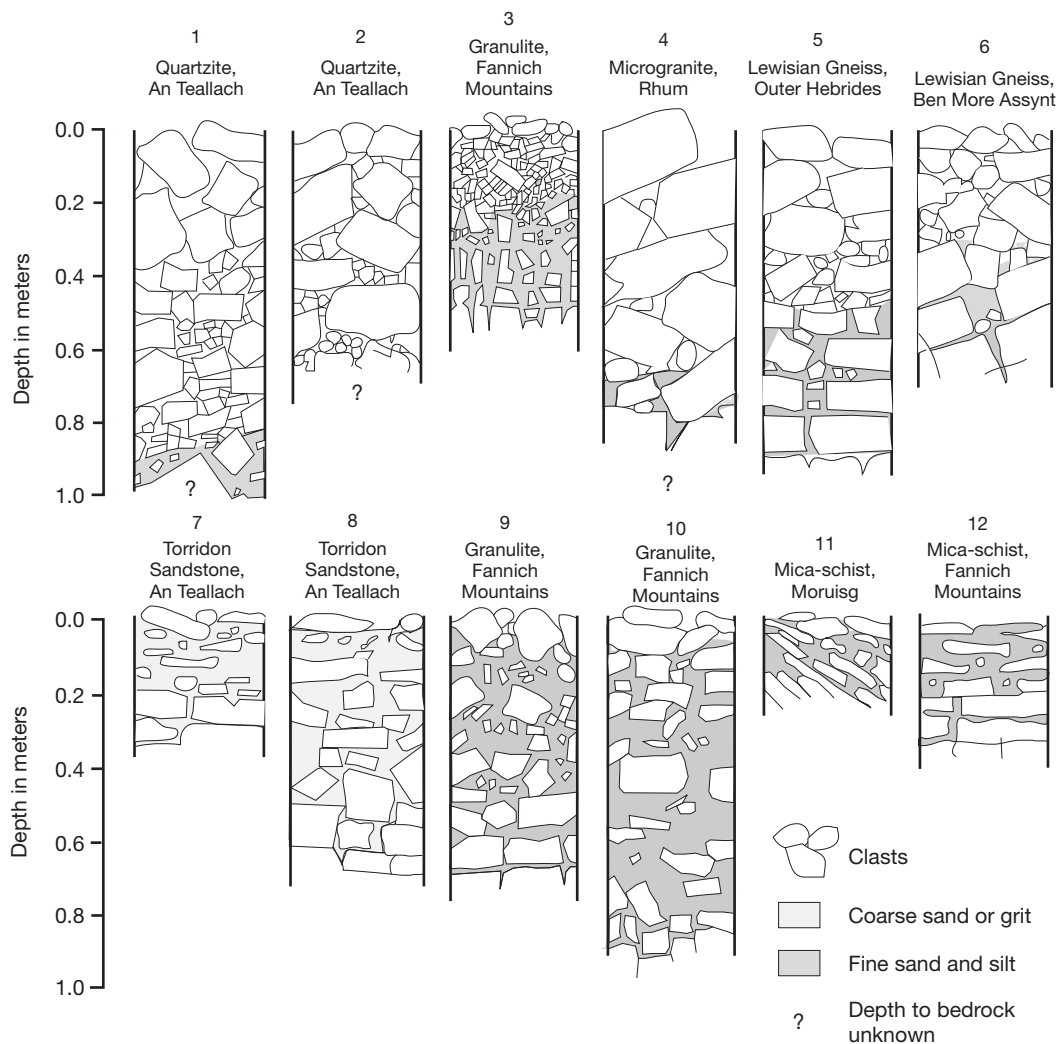


Figure 7 Profiles through blockfields developed on mountains in the northwest Highlands of Scotland. Type 1 are represented by 1–6, type 2 by 7–8, and type 3 by 9–12. Reproduced from Ballantyne CK (1998) Age and significance of mountain-top detritus. *Permafrost and Periglacial Processes* 9: 327–345.

glaciation, there are implications for ice dynamics and ice sheet/glacier geometries. Blockfields may survive through glaciation either as nunataks (Dahl, 1987) in which case there may be trimlines defining a lower limit to their distribution (Ballantyne, 1998; McCarroll, et al., 1995; Nesje, 1989; Nesje et al., 1988), or beneath cold-based protective ice (Hättestrand, and Stroven, 2002; Ives, 1958; Kleman and Borgstrom, 1990; Kleman and Stroeven, 1997; Rea et al., 1996a,b; Sugden and Watts, 1977).

Pre-Pleistocene

A pre-Pleistocene age means that the blockfield has been developing for probably a significantly longer time than 1.8 Ma and has also probably experienced warmer climates than the present day. As conditions deteriorated into the Pleistocene, local high-level cirque glaciation, then regional icefields, and ultimately ice sheets grew and disappeared repeatedly in mid- and high latitudes (Kleman and Stroeven, 1997). During the onset and termination phases, the relict weathering profiles would have been exposed to intense periglacial and permafrost processes which would have moved the basal corestones upward through frost-heave mechanisms (Hättestrand and Stroven, 2002; Kleman and Borgstrom, 1990; Kleman and Stroeven, 1997; Rea et al., 1996a,b; Whalley et al., 1996). Frost-heaving of clasts through the profile will destroy any former weathering horizons, homogenizing the matrix sediments (Figure 5; Rea et al., 1996a). As chemical weathering was dominant during the initial formation, this should be reflected, to some degree at least, in the particle size and clay mineralogy of the matrix sediments. Depending upon the parent lithology and degree of alteration, the matrix sediments may be classified under any of the granulometric classes shown in Table 1 (Dahl, 1966; Rea et al., 1996a,b; Roaldset et al., 1982; Whalley, et al., 1996). Type 2 or type 3 profiles are expected (Figures 2(b), 3, 4(b, c), and 5(b, c)) unless a significant amount of fines have been removed. The matrix sediment may also contain a coarser component of Pleistocene age, derived from granular disintegration of exposed surface blocks.

Again, if the blockfield is located within the limits of any former glaciations, then it has survived intact, either by remaining unglaciated, for example, as a nunatak or enclave like the Antarctic Dry Valleys, or by protection beneath a cold-based ice cover.

A Unifying Model

As noted above, there has been long-standing debate regarding the age and origin of blockfields, with convincing evidence presented by different authors, for different sites, to support the full spectrum of suggested formation ages. Ballantyne (2010a) presented a model that eloquently unifies the pre-Pleistocene and Pleistocene models. Utilizing the sediment stripping/tor exhumation rates determined from cosmogenic exposure ages and tor relief, he clearly demonstrates that in many instances, up to a few tens of meters of sediment have been stripped from many landsurfaces which contain tors and blockfields. It is assumed that unless a landsurface has previously been completely denuded (e.g., under warm-based ice), it must have evolved from a pre-Pleistocene regolith cover,

which depending upon geographic location, lithology, and age may have had weathering fronts at considerable depths below the surface and contained substantial amounts of clays and clay minerals indicative of weathering longevity and warmer climates. Once this simple concept is established, it then follows that (Figure 8):

- if Pleistocene regolith stripping is limited, then intact roots of the saprolitic profile, including *in situ* corestones, will be preserved;
- if Pleistocene regolith stripping approaches the depth of the initial weathering front, then evidence of this will be preserved in the clay minerals and matrix particle size, but the profile is modified by periglacial frost wedging and heaving of blocks to the surface;
- if Pleistocene regolith stripping exceeds the depth of the initial weathering front, then evidence of this former profile is lost and a periglacial-derived blockfield is formed containing small amounts of clay produced by limited chemical weathering and possibly some remnant from the original profile.

This model provides a framework for interpreting the vast majority of blockfields described in the literature, resolving the apparent dichotomy of blockfields with significant quantities of evolved clays and those with limited and less evolved clays.

Implications for Ice Geometry and Dynamics

The lower limits of blockfields, if marked by a trimline, represent an important tool in paleoglaciology and provide an accurate marker of ice thickness (Figure 9). Examples include those in southern Norway (Nesje, 1989; Nesje et al., 1988) and the Scottish Highlands (Ballantyne, 1998; McCarroll, et al., 1995), although, as noted above, it has recently been suggested some of these may actually represent internal thermal, and thus erosional, boundaries. However, dating of blockfields and trimlines is a significant problem which will be discussed briefly below.

Proximity to a present-day ice mass (Rea et al., 1996a,b), spatial relationships with ice limits and ice-modified landscapes (Ballantyne, 1998; Dyke, 1993; Hättestrand and Stroven, 2002; Kleman and Borgstrom, 1990), or the presence of erratics (Ives, 1958; Sugden and Watts, 1977) provide evidence which may indicate that a blockfield has been formerly covered by ice. In such situations, it is clear that the ice cover has been nonerosive, and in all likelihood cold-based. The presence of scattered erratics presents no significant problems of interpretation, as erratics may be transported in cold-based ice (Ives, 1958; Marquette et al., 2004; Rea et al., 1996b; Sugden and Watts, 1977). However, large numbers of erratics may indicate that the blockfield is sourced from till, having been sorted sometime after deposition (Roaldset et al., 1982). Such an origin for blockfields places them in a different category from the autochthonous or para-autochthonous blockfields discussed in this article.

The juxtaposition of plateaux containing preserved relict landsurfaces including blockfields and tors, with adjacent deeply incised glacial valleys and fjords, is interpreted as a landscape of selective linear erosion. From local, through

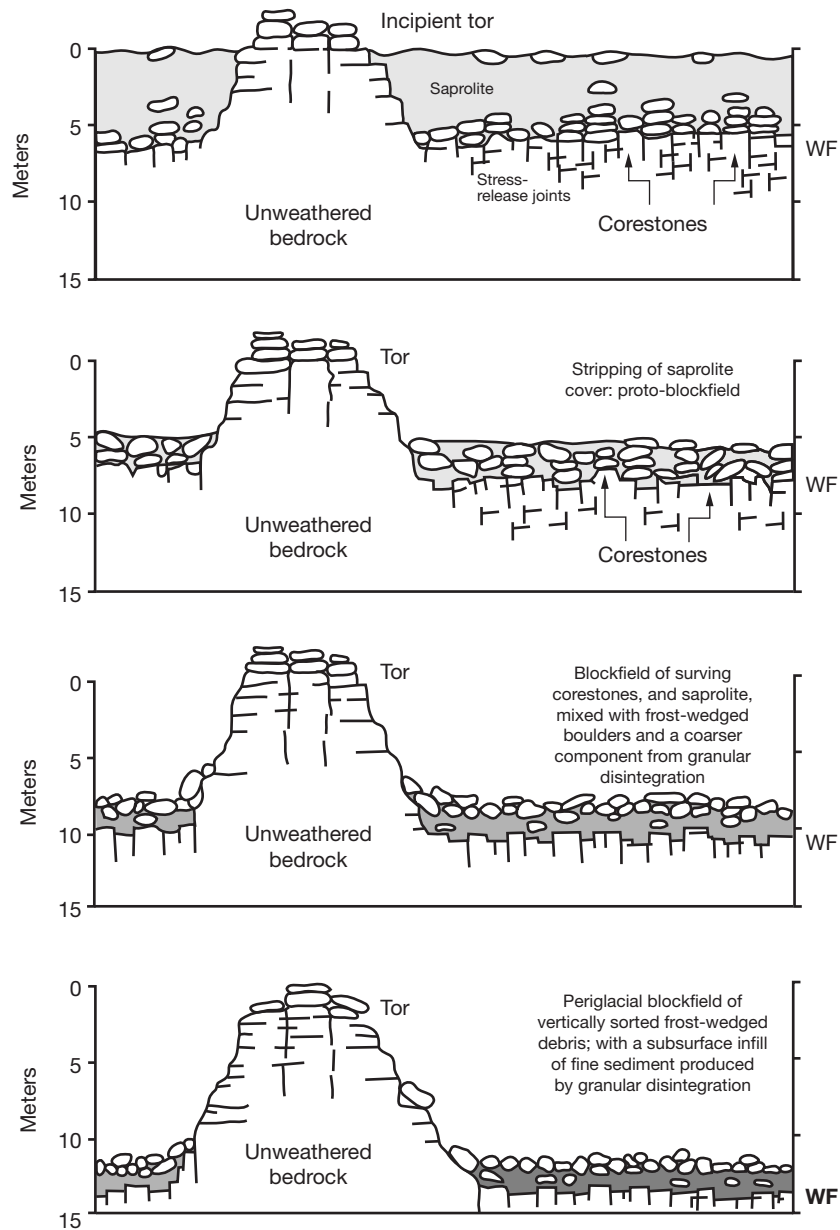


Figure 8 A model for autochthonous blockfield formation illustrating the various levels of modification resulting from preglacial regolith stripping and Pleistocene cold-climate weathering. WF, weathering front. Stage 1 represents the preglacial situation where there is deep chemically weathered saprolite cover. During Stage 2 in the early Pleistocene, changed conditions see the initiation of saprolite stripping (with tor exhumation) and concomitant concentration of core stones with the rate of surface lowering reduced once a corestone lag 'armors' surface. Stage 3 follows continued stripping and increased modification by permafrost/periglacial weathering, with the weathering front depth controlled by the maximum depth reached by either seasonal freezing or active-layer melting. The matrix will have a mixed mineralogy and particle sizes comprising some relict saprolite and some granular disintegration products. Stage 4 is reached when surface lowering has removed all traces of the former weathering, the profiles are entirely dominated by frost-wedged and heaved boulders with a granular disintegration-derived coarse matrix. There may be traces, or minor amounts, of clay minerals present but these are the products of interglacial weathering. Modified from Ballantyne CK (2010) A general model of autochthonous blockfield evolution. *Permafrost and Periglacial Processes* 21: 289–300. <http://dx.doi.org/10.1002/ppp.700>.

regional, to full ice-sheet glaciation, ice in the valleys is more likely to be warm-based, due to factors such as lower elevation, thicker ice, and increased ice discharge. The corollary is that on the intervening plateaux/interfluvies ice remains thinner, colder, and less or nonerosive (Rea et al., 1996b; Sugden and Watts, 1977). Through this topographic control, preservation of landsurfaces and extensive erosion can coexist. From the

model presented in Figure 8, it is clear that these landsurfaces are not truly relict or preserved (e.g., if 10 m of regolith has been stripped). However, they do represent a guide to a preglacial surface (though not necessarily one graded to a former sea level) and a marker for interpreting evolution of a landscape of selective linear erosion. Preservation–modification relationships without vertical incision have also been found



Figure 9 Trimline with blockfield above, delimiting former glacier ice thickness, Mount Hoffmann, Yosemite, California, USA. Photograph by C. K. Ballantyne.

for the centers of ice sheets and ice caps in parts of Scandinavia and the Canadian Arctic and are also related to basal thermal boundaries (Dyke, 1993; Hättestrand and Stroven, 2002; Kleman and Borgstrom, 1990; Kleman and Stroeven, 1997).

Dating Blockfields

While the age of blockfields is of major interest, its determination is complicated by the likely duration of protection/preservation by cold-based ice. Clay mineralogy has traditionally been used as a proxy for longevity and aggressiveness of the weathering environment, though, as cautioned by Goddard (1989), care must be taken with this approach. The clay mineral species, the quantity of clays, and potential for a non-*in situ* weathering origin, for example, bedrock or exhumed sources (Fjellanger and Nystuen, 2007) or aeolian deposition, must all be assessed before making an age interpretation. For example, there is uncertainty regarding the age of gibbsite in montane soils in Scotland; is it truly pre-Pleistocene or the result of interglacial weathering through the Pleistocene (Ballantyne, 1998; Hall et al., 1989)? Even when the evidence is convincing that the clay mineralogy represents a long-term *in situ* history, this is generally poorly constrained (Rea et al., 1996a,b; Roaldset et al., 1982).

The development of cosmogenic dating represents a major breakthrough in the study of relict landsurfaces. From the discussion above, it seems that there are blockfields in mid- to high-latitude locations that have experienced considerable longevity in the landscape, with little modification, despite repeated coverage during Pleistocene glaciations. Stone et al. (1998) used ^{36}Cl and ^{10}Be to date sites above and below trimlines in northwest Scotland. The authors clearly demonstrated the exposure age differences between the sites, suggesting that the older ages (22.6–70 cal ka BP) above the trimline indicate exposure throughout the last glaciation and the younger ages below the trimline (13.9–17.6 cal ka BP) indicate erosion and resetting of the cosmogenic clock. However, complications exist in interpreting exposure ages, as there are inheritance effects associated with multiple exposure–burial events, which are likely to affect the measured age (Ballantyne, 1998; Bierman et al., 1999; Marquette et al., 2004; Stone et al., 1998). The use

of cosmogenic radionuclide ratios with different half-lives (e.g., $^{26}\text{Al}/^{10}\text{Be}$) must be used to indicate whether single or multiple phases of burial–exposure have occurred. Bierman et al. (1999) used paired ratios to interpret minimum total histories for glaciated bedrock surfaces on Baffin Island up to 735 ka (Table 3). Dating exposure ages/histories of tors, blockfield blocks and erratics can also help constrain the dimensions and duration of former ice cover (e.g., Marquette et al., 2004). Additionally, combining paired ratios with exposure history proxy records derived independently may further improve the modeling of exposure histories (e.g., Phillips et al., 2006; Stroven et al., 2002). Table 3 presents some example ages that offer convincing support of the concept of longevity of blockfields and tors in repeatedly glaciated landscapes.

Small et al. (1997, 1999) determined bedrock erosion and regolith production rates for weathering-rate-limited summits in some mountain ranges in western United States. Their data indicated well-mixed profiles due to active frost-heaving processes. They could find no evidence for former ice cover, and they inferred that weathering rates beneath the regolith (blockfield) were almost double those of bare rock ridges and tors. The summit flats, crest ridges and tors were thought to have evolved over the past 2–3 Ma by more rapid weathering and surface lowering beneath the blockfield and reduced weathering on crest ridges and tors. Surface lowering was accomplished by slow creep of material down to, and over, the sides of the summit flats (Anderson, 2002). The results of this work are interesting, as they question the widely held interpretation that these summits represent former land surfaces (though not necessarily graded to a former base level). The model remains to be tested in other localities where there is clear evidence of a former ice cover (which could reduce considerably the time for weathering and mass movement) and on blockfields where bedrock lithological changes are retained within the blockfield. For example, where an underlying bedrock lithological boundary is retained within the overlying blockfield, and provided that the boundary is not oriented directly downslope, this indicates no removal of material by mass movement.

Conclusion

Long-standing debates regarding the age of blockfields are now, through the advent of new dating techniques such as cosmogenic nuclide dating, beginning to be resolved. Cosmogenic dating has confirmed age ranges (from blockfield and/or tors, see Table 3) from the recent through to 845 ka (Fabel et al., 2002), representing long-term evolution and ‘stability’ within the landscape. However, cosmogenic nuclide dating cannot provide all the answers, as multiple burial–exposure histories make interpretation of ages complex and the heaving of fresh boulders to the surface may mean a range of exposure ages are found, and any date can be considered a minimum (Bierman et al., 1999; Marquette et al., 2004). Geomorphology and clay mineralogy remain vital for understanding the origin of blockfields in terms of their weathering origins. The history of blockfields can provide data on glacial, interglacial, and preglacial earth-surface processes and an integration of these through time, and also provide marker surfaces from which to quantify glacial incision and rates of landscape

Table 3 Examples of cosmogenic nuclide exposure ages and total histories for blockfields, tors, and associated trimlines

Source	Sample	Comment	Nuclide	Exposure age (ka)
Stone et al. (1998)	ST-1	Below trimline	^{36}Cl	17.4 ± 1.3
Stone et al. (1998)	ST-2	Below trimline	^{36}Cl	17.6 ± 1.4
Stone et al. (1998)	ST-3	Above trimline	^{36}Cl	70.3 ± 5.1
Stone et al. (1998)	ST-5	Above trimline	^{36}Cl	50.2 ± 3.3
Stone et al. (1998)	AT-1	Below trimline	^{10}Be	14.4 ± 1.3
Stone et al. (1998)	AT-3	Below trimline	^{10}Be	14.7 ± 1.3
Stone et al. (1998)	AT-5	Above trimline	^{10}Be	49.4 ± 4.7
Stone et al. (1998)	AT-6	Above trimline	^{10}Be	31.2 ± 2.9
Bierman et al. (1999)	KM-95-16	—	Paired ^{10}Be and ^{26}Al	558 ^a
Bierman et al. (1999)	KM-95-108	—	Paired ^{10}Be and ^{26}Al	450 ^a
Bierman et al. (1999)	KM-95-112	—	Paired ^{10}Be and ^{26}Al	735 ^a
Gualtier and Brigham-Grette (2001)	97LD7	Tor	^{10}Be	254
Gualtier and Brigham-Grette (2001)	97LD8	Blockfield	^{10}Be	143
Fabel et al. (2002)	Alddasčorru (1380)	Bedrock in relict patch	^{10}Be	$\sim 210^b$
Fabel et al. (2002)	Tjuolmma (920)	Bedrock in relict patch	Paired ^{10}Be and ^{26}Al	845 ^c
Stroven et al. (2002)	99-01	Naakakarhakka – tor	Paired ^{10}Be and ^{26}Al	$> 485^d$
Stroven et al. (2002)	99-04	Lamuvaara – tor	Paired ^{10}Be and ^{26}Al	$> 765^d$
Marquette et al. (2004)	LAB-00-01B	Felsenmeer zone – bedrock outcrops and tors	^{10}Be	157 ± 15.2
Marquette et al. (2004)	LAB-00-22A	Felsenmeer zone – bedrock outcrops and tors	^{10}Be	88.6 ± 7.0
Marquette et al. (2004)	KAU-99-01	Felsenmeer zone – blocks	^{10}Be	144 ± 11.4
Phillips et al. (2006)	DDL3	Da Dhruim Lom erratic	Paired ^{10}Be and ^{26}Al	123
Phillips et al. (2006)	CA 2	An Cnoidh pitted surface	Paired ^{10}Be and ^{26}Al	471 ^e
Phillips et al. (2006)	RF1	Clach na Gnù is tor summit	Paired ^{10}Be and ^{26}Al	675 ^e
Phillips et al. (2006)	CB 2	Clach Bhàn tor summit	Paired ^{10}Be and ^{26}Al	307 ^e
Nesje et al. (2007)	COSMO 5	Andoya – outcrop above trimline	^{10}Be	31.52 ± 2.57^f
Nesje et al. (2007)	COSMO 6	Andoya – perched boulder above trimline	^{10}Be	38.56 ± 3.02^f
Nesje et al. (2007)	COSMO 18	Skånland – outcrop above trimline	^{10}Be	74.99 ± 4.06^f
Nesje et al. (2007)	COSMO 19	Skånland – outcrop above trimline	^{10}Be	56.85 ± 3.14^f

^aMinimum total history ages assuming periods of burial under cold-based ice, identified from discordant ^{10}Be and ^{26}Al ages.

^bMinimum total history assuming the site is only covered at full glacial.

^cMean age estimate obtained using a cosmogenic accumulation and decay mode assuming prior exposure and a single shielding event with subsequent reexposure.

^dModeled using a burial–exposure history constrained by the DSDP 607 $\delta^{18}\text{O}$ benthic foraminiferal record.

^eModeled using a burial–exposure history constrained by the DSDP 607 $\delta^{18}\text{O}$ benthic foraminiferal record.

^fPre-Last Glacial Maximum.

evolution under repeated cycles of glaciation. Preservation may define the geometries and thermal regimes of Pleistocene ice masses. Mapping, dating, and analyses of blockfields still have much to offer Quaternary science, glaciology, and landscape-evolution research.

See also: Glacial Landforms, Erosional Features: Major Scale Forms. Glacial Landforms, Ice Sheets: Evidence of Glacier and Ice Sheet Extent; Trimlines and Paleonunataks. Permafrost and Periglacial Features: Active Layer Processes; Block/Rock Streams; Permafrost; Slope Deposits and Forms.

References

- Anderson RS (2002) Modeling the tor-dotted crests, bedrock ridges, and parabolic profiles of high alpine surfaces of the Wind River Range, Wyoming. *Geomorphology* 46: 35–58.
- Ballantyne CK (1998) Age and significance of mountain-top detritus. *Permafrost and Periglacial Processes* 9: 327–345.
- Ballantyne CK (2010a) A general model of autochthonous blockfield evolution. *Permafrost and Periglacial Processes* 21: 289–300. <http://dx.doi.org/10.1002/ppp.700>.
- Ballantyne CK (2010b) Extent and deglacial chronology of the last British–Irish Ice Sheet: Implications of exposure dating using cosmogenic isotopes. *Journal of Quaternary Science* 25: 515–534.
- Bierman PR, Marsella KA, Patterson C, Davis PT, and Caffee M (1999) Mid-Pleistocene cosmogenic minimum-age limits for pre-Wisconsinan glacial surfaces in southwestern Minnesota and southern Baffin Island: A multiple nuclide approach. *Geomorphology* 27: 25–39.
- Boelhouwers J (2004) New perspectives in autochthonous block-field development. *Polar Geography* 28: 133–146.
- Caine N (1983) *The Mountains of Northeastern Tasmania. A Study of Alpine Geomorphology*. Rotterdam and Salem NH, USA: Balkema.
- Dahl E (1987) The nunatak theory reconsidered. *Ecological Bulletins* 38: 77–94.
- Dahl R (1966) Block fields, weathering pits and tor-like forms in the Narvik Mountains, Nordland, Norway. *Geografiska Annaler A* 48: 55–85.
- Dredge LA (1992) Breakup of limestone bedrock by frost shattering and chemical weathering, eastern Canadian Arctic. *Arctic and Alpine Research* 24: 314–323.
- Dyke AS (1993) Landscapes of cold-centered late Wisconsin ice caps, Arctic Canada. *Progress in Physical Geography* 17: 223–247.
- Fabel D, Stroven AP, Harbor J, Kleman J, Elmore D, and Fink D (2002) Landscape preservation under Fennoscandian ice sheets determined from in situ produced ^{10}Be and ^{26}Al . *Earth and Planetary Science Letters* 201: 397–406.

- Fjellanger J and Nystuen JP (2007) Diagenesis and weathering of quartzite at the palaeic surface on the Varanger Peninsula, northern Norway. *Norwegian Journal of Geology* 87: 133–145.
- Fjellanger J and Sørbel L (2007) Origin of the palaeic landforms and glacial impact on the Varanger Peninsula, northern Norway. *Norwegian Journal of Geology* 87: 223–238.
- Fjellanger J, Sørbel L, Linge H, Brook EJ, Raisbeck GM, and Yiou F (2006) Glacial survival of blockfields on the Varanger Peninsula, northern Norway. *Geomorphology* 82: 255–272.
- Goddard A (1989) Les vestiges des manteaux d'alteration sur les socles des hautes latitudes? Identification, signification. *Zeitschrift für Geomorphologie, Supplementband* 72: 1–20 (translation in Evans DJA (ed.) (1996) *Cold Climate Landforms*, pp. 397–411. Chichester: Wiley).
- Goodfellow BW, Fredin O, Derron M-H, and Stroeven AP (2008) Weathering processes and Quaternary origin of an alpine blockfield in Arctic Sweden. *Boreas* 38: 379–398. <http://dx.doi.org/10.1111/j.1502-3885.2008.00061.x>.
- Gualtier L and Brigham-Grette J (2001) The age and origin of the Little Diomed Island upland surface. *Arctic* 54: 12–21.
- Hall AM, Mellor A, and Wilson MJ (1989) The clay mineralogy and age of deeply weathered rock in north-east Scotland. *Zeitschrift für Geomorphologie, Supplementband* 72: 97–108.
- Hall K, Thorn CE, Matsuoka N, and Prick A (2002) Weathering in cold regions: Some thoughts and perspectives. *Progress in Physical Geography* 26: 577–603.
- Hättestrand C and Stroven AP (2002) A relict landscape in the centre of the Fennoscandian glaciation: Geomorphological evidence of minimal Quaternary glacial erosion. *Geomorphology* 44: 127–143.
- Ives JD (1958) Glacial geomorphology of the Torngat Mountains northern Labrador. *Geographical Bulletin* 12: 47–75.
- Kleman J and Borgstrom I (1990) The boulder fields of Mt. Fulufjället, west-central Sweden. *Geografiska Annaler A* 72: 63–78.
- Kleman J and Stroeven AP (1997) Preglacial surface remnants and Quaternary glacial regimes in northwestern Sweden. *Geomorphology* 19: 35–54.
- Lautridou J-P and Seppälä M (1986) Experimental frost shattering of some Pre-Cambrian rocks. *Geografiska Annaler A* 68: 89–100.
- Marquette GC, Gray JT, Gosse JC, et al. (2004) Felsenmeer, persistence under non-erosive ice in the Torngat and Kaumajet mountains, Quebec and Labrador, as determined by soil weathering and cosmogenic nuclide exposure dating. *Canadian Journal of Earth Sciences* 41: 19–38.
- McCarroll D, Ballantyne CK, Nesje A, and Dahl SO (1995) Nunataks of the last ice sheet in north-west Scotland. *Boreas* 24: 305–323.
- Mellor A and Wilson MJ (1989) Origin and significance of gibbsitic montane soils in Scotland, UK. *Arctic and Alpine Research* 21: 417–424.
- Nesje A (1989) The geographical and altitudinal distribution of block fields in southern Norway and its significance to the Pleistocene ice sheets. *Zeitschrift für Geomorphologie* 71: 41–53.
- Nesje A, Dahl SO, Anda E, and Rye N (1988) Block fields in southern Norway: Significance for the Late Weichselian ice sheet. *Norsk Geologisk Tidsskrift* 68: 149–169.
- Nesje A, Dahl SO, Linge H, et al. (2007) The surface geometry of the Last Glacial Maximum ice sheet in the Andøya-Skånland region, northern Norway, constrained by surface exposure dating and clay mineralogy. *Boreas* 36: 227–239.
- Phillips WM, Hall AM, Mottran R, Fifield LK, and Sugden DE (2006) Cosmogenic ^{10}Be and ^{26}Al exposure ages of tors and erratics, Cairngorm Mountains, Scotland: Timescales for the development of a classic landscape of selective linear glacial erosion. *Geomorphology* 73: 222–245.
- Rea BR, Whalley WB, and Porter EM (1996a) Rock weathering and the formation of summit blockfield slopes in Norway: Examples and implications. In: Anderson MG and Brooks SM (eds.) *Advances in Hillslope Processes*, pp. 1257–1275. Chichester: Wiley.
- Rea BR, Whalley WB, Rainey MM, and Gordon JE (1996b) Blockfields, old or new? Evidence and implications from some plateaus in northern Norway. *Geomorphology* 15: 109–121.
- Roaldset E, Pettersen E, Longva O, and Mangerud J (1982) Remnants of preglacial weathering in western Norway. *Norsk Geologisk Tidsskrift* 62: 169–178.
- Small EE, Anderson RS, and Hancock GS (1999) Estimates of the rate of regolith production using ^{10}Be and ^{26}Al from alpine hillslope. *Geomorphology* 27: 131–150.
- Small EE, Anderson RS, Repka JL, and Finkel R (1997) Erosion rates of alpine bedrock summit surfaces deduced from in situ ^{10}Be and ^{26}Al . *Earth and Planetary Science Letters* 150: 413–425.
- Stone JO, Ballantyne CKB, and Fifield K (1998) Exposure dating and validation of periglacial weathering limits, northwest Scotland. *Geology* 26: 587–590.
- Stroven AP, Fabel D, Hättestrand C, and Harbor J (2002) A relict landscape in the centre of Fennoscandian glaciation: Cosmogenic radionuclide evidence of tors preserved through multiple glacial cycles. *Geomorphology* 44: 145–154.
- Sugden DE and Watts SH (1977) Tors, felsenmeer, and glaciation in northern Cumberland Peninsula, Baffin Island. *Canadian Journal of Earth Sciences* 14: 2817–2823.
- Taylor G and Eggerton RA (2001) *Regolith Geology and Geomorphology*. Chichester: Wiley.
- van Wambeke AR (1962) Criteria for classifying tropical soils by age. *Journal of Soil Science* 13: 124–132.
- Velde B (1992) *Introduction to Clay Minerals: Chemistry, Origins, Uses and Environmental Significance*. London: Chapman and Hall.
- Whalley WB, Rea BR, Rainey MM, and McAlister J (1996) Rock weathering in blockfields: Some preliminary data from mountain plateaus in north Norway. In: Widdowson M (ed.) *Palaeosurfaces: Recognition, Reconstruction, and Interpretation*, pp. 133–145. London: Geological Society of London. Special Publication No. 120.

Rock Glaciers and Protalus Forms

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Definition and Distribution

Perennially frozen debris in cold mountains may, under the force of gravity, form so-called rock glaciers, which often show a shape similar to that of lava streams (Figures 1–7). Formerly, rock glaciers were often thought to be a form of debris-covered glaciers, which led to the term ‘rock glaciers.’ Since then, a number of fundamental differences between rock glaciers and glaciers related, for example, to their thermal conditions and mass balance have been established. This has led to the use of the term ‘rock glaciers’ to distinguish their individual character from that of glaciers (Barsch, 1996). This meaning of ‘rock glacier’ is used in this article, although the equivalent term ‘rock glacier’ is also established in the scientific literature.

Rock glaciers occur in most cold mountains of Earth. These include the Andes, the Rocky Mountains, the European Alps, the Pyrenees, the Caucasus region, the Central Asian mountain ranges, the Siberian mountain ranges, the Himalayas, the New Zealand Alps, Greenland, Antarctica, and the Arctic and Antarctic Islands (Barsch, 1996; Martin and Whalley, 1987). Rock glaciers often occur in similar, though slightly colder and drier climatic settings as glaciers (Brazier et al., 1998; Humlum, 1998). It is also speculated, on the basis of space imagery, that rock glacier-like features exist on Mars. Under suitable climatic and topographic conditions – that is, mean annual air temperatures between roughly -2 and -6 °C, total annual precipitation between 500 and 1500 mm, and sufficient debris production from rock walls (Etzelmueller and Frauenfelder, 2009) – rock glaciers may be abundant, with their number often exceeding that of glaciers in the same mountain range, but with significantly smaller areas for individual rock glaciers.

Thermal Conditions

A fundamental characteristic of rock glaciers is their thermal state (Haeberli, 2000; Haeberli et al., 2006). They consist of a perennially frozen mixture of debris and ice, that is, their long-term existence requires permafrost conditions. Permafrost is defined as lithospheric material with year-round negative ground temperatures (Harris et al., 2009). A surface layer of depth ranging from several decimeters to meters thaws during summer and is termed the ‘active layer.’ Below the ‘permafrost table’ – the lower boundary of the active layer – the temperatures in the ‘permafrost body’ remain negative down to the ‘permafrost base’ (Haeberli, 1985; Haeberli et al., 2006; Vonder Muehl et al., 2003). Thus, the mixture of debris and ice within the permafrost body remains frozen throughout the year, possibly over centuries and millennia if the thermal conditions do not change substantially over time.

The surface layer of rock glaciers consists of debris with typical diameters ranging from several centimeters or decimeters to meters, depending on the geological setting, weathering

conditions, and processes of debris reworking. If this layer is at least as thick as the local active layer, that is, the seasonal thaw depth under the specific local conditions, then positive ground temperatures during summer cannot reach down to the ground ice containing permafrost body. Thus, ground ice can persist for very long periods of time (Haeberli et al., 2006).

From the aforementioned principle, it becomes clear that the long-lasting persistence of the frozen core within a rock glacier requires permafrost conditions. The difference between rock glaciers, dead ice, or debris-covered glaciers can thus be described by the thermal state of the material.

Transformations between these conditions are possible. For instance, glacier ice under an insulating debris cover and lying within a permafrost environment is subject to seasonal ice melt, though substantially damped, wherever the debris cover is thinner than the local seasonal thaw depth. The ice body will continuously lose mass if not compensated by the advection of glacier ice. The resulting meltwater, and with it the contained energy, can percolate into and warm the underlying ice body. If the ice body contains sufficient debris, this material will accumulate at the top of the ice core, leading to a thickening of the debris cover. When the thickness of the debris cover equals the local seasonal thaw depth, the underlying ice is protected against surface melt unless lateral processes such as erosion occur. Thus, the debris within rock glaciers may stem from periglacial and glacial processes and sources, and the rock glacier ice content may originate, for example, from refreezing rain and meltwater, glaciers, buried ice and snow patches, or combinations thereof (Haeberli and Vonder Muehl, 1996; Haeberli et al., 2006). Temporal transformations and spatial transitions between rock glaciers, debris-covered glaciers or dead ice, and ice-cored moraines are often found in nature.

Being closely related to permafrost conditions, rock glaciers indicate the existence of discontinuous or continuous mountain permafrost on a regional scale (Harris et al., 2009). The lower boundary of regional rock glacier distribution, however, does not generally coincide exactly with the lower boundary of permafrost distribution (Frauenfelder et al., 2008). Rock glaciers may end above the regional permafrost limit due to topographic constraints, limited debris supply, or insufficient time to fully develop. However, the coarse debris cover on top of a rock glacier favors ground cooling due to enhanced heat transfer to the atmosphere, allowing rock glaciers to extend downslope below the regional permafrost limit.

Composition

The composition of rock glaciers is investigated by boreholes and geophysical surveys, particularly geoseismics, direct current resistivity soundings, or georadar. So far, only a few boreholes have been drilled through rock glaciers (Arenson et al., 2002; Haeberli et al., 1998, 2006; Harris et al., 2009; Hoelzle



Figure 1 Active rock glacier in the Muragl valley, Upper Engadine, Swiss Alps. Photograph by R. Frauenfelder.



Figure 4 Inactive rock glaciers in the Upper Engadine, Swiss Alps. Note the yellow color from lichen cover and the vegetated rock glacier fronts.

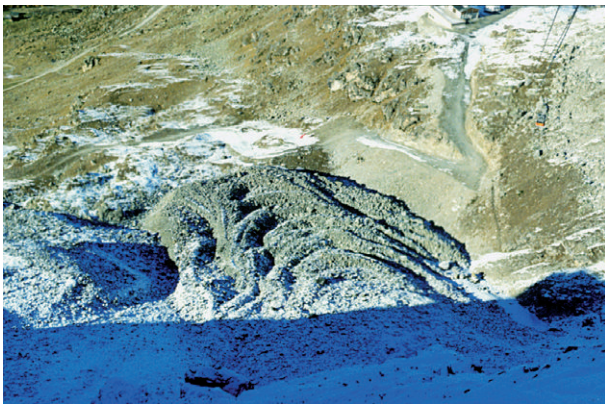


Figure 2 Active rock glacier Murtel, Upper Engadine, Swiss Alps.



Figure 5 Relict rock glacier at the Julier Pass, Swiss Alps. Note the complete vegetation cover and the concave forms from ice loss.



Figure 3 Active rock glacier in the Suvretta valley, Upper Engadine, Swiss Alps.

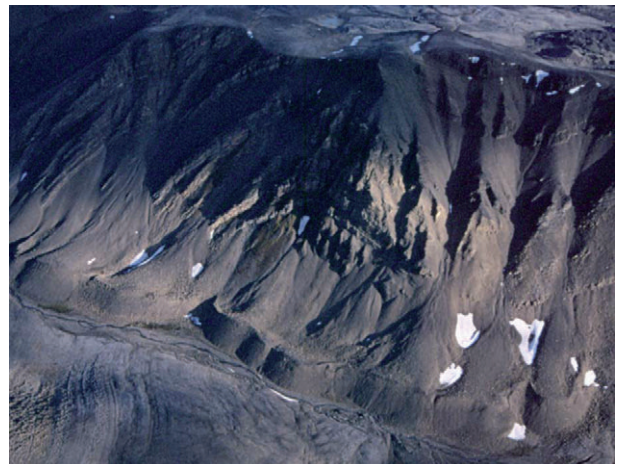


Figure 6 Typical ramp-type rock glaciers at Nordenskjöldkysten, Svalbard.



Figure 7 Rock glacier in the Tien Shan. Photograph by S. Titkov.

et al., 1998; Steig et al., 1998) and, therefore, the extent to which their findings can be generalized is uncertain. All the drillings have revealed a surface layer of ice-free debris with a thickness on the order of meters (active layer; Haeberli et al., 2006). The underlying layer, a few tens of meters thick, consists of a frozen mixture of ice and debris with an ice content ranging from close to 100% to a few tens of percent. In rock glaciers where deep core drilling was conducted, a third layer of frozen coarse debris with little ice content was encountered. This layer is believed to originate from surface blocks that had fallen over the rock glacier front and were overridden by the advancing rock glacier (see section 'Kinematics and Advance'; Haeberli et al., 1998).

In the rock glaciers with deep core drillings, a meter-thick layer of sand-rich ice was found within the second ice-rich layer, at depths between 15 and 30 m. The majority of the horizontal deformation within the rock glacier was found to be concentrated in these sandy ice layers (see section 'Kinematics and Advance'; Arenson et al., 2002).

Many geophysical soundings have confirmed the general layering and composition described earlier for other rock glaciers (Berthling et al., 1998, 2000; Bucki et al., 2004; Isaksen et al., 2000). In addition, some high-resolution soundings have revealed fine layers within the ice-rich subsurface layer. These microlayers were attributed to seasonal sequences of snow/ice and debris deposition at the rock glacier surface, and their continuous incorporation and deformation within the creeping body (Berthling et al., 2000; Humlum et al., 2007).

Kinematics and Advance

Besides their special thermal conditions, the second fundamental characteristic of rock glaciers is their mode of displacement. Given sufficient terrain slope, the ice–debris mixture within a rock glacier deforms under the force of gravity and the stress of the vertical column. The resulting surface speeds on rock glaciers may be several meters per year (Kääb et al., 2003; Kaufmann and Ladstädter, 2002; Figure 8). Rock glacier surface speed depends, among other things, on surface slope, composition and internal structure, thickness of the ice-rich body, and ground temperature (Haeberli, 1985). The observed minimum surface speeds of a few millimeters per year are not

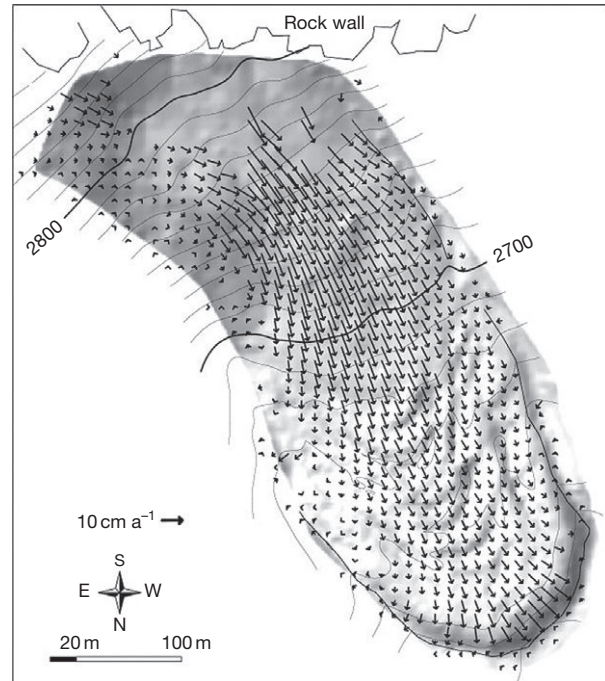


Figure 8 Surface displacement vectors on Murtel rock glacier, Swiss Alps, measured from air photographs of 1987 and 1996.

limited by the deformation process itself, but rather by the available measurement technique (e.g., digital photogrammetry, global navigation satellite system, radar interferometry; Lambiel and Delaloye, 2004; Strozzini et al., 2004).

A coherent ice content within rock glaciers leads to the lateral transfer of stresses and thus to a coherent velocity field. Owing to effects such as lateral friction and variations in viscosity, the highest horizontal speeds on a rock glacier are usually found along its centerline. In the few available boreholes through rock glaciers, the horizontal deformation was concentrated in horizontal layers some decimeters to meters thick, rather than distributed throughout the body of the rock glacier, with a clear parabolic variation of horizontal speeds with depth such as found for glacier ice (Arenson et al., 2002).

The basic concept of mass transport within a rock glacier can be derived from the characteristic thermal and kinematic conditions: debris is deposited at the rock glacier surface, either directly from weathering processes at the headwall or from other secondary sources, in most cases presumably glaciers and avalanches (Humlum, 1998; Humlum et al., 2007). Together with ice from different sources (see section 'Thermal Conditions'), this mass represents the mass input to the rock glacier system. Following a basic law of fluid dynamics, the surface debris is transported downslope, with the highest creep speeds at the rock glacier surface. The advected surface material then falls over the steep rock glacier front. The grain-size sorting along the frontal slope occurring during this process leads to the typical appearance of rock glacier fronts. The surface debris that accumulates at the front is then overridden by the advancing rock glacier and again incorporated into its base, in a 'caterpillar' or 'conveyor belt' fashion (Kääb and Reichmuth, 2005; Figure 9). On account of their nonice sediment content

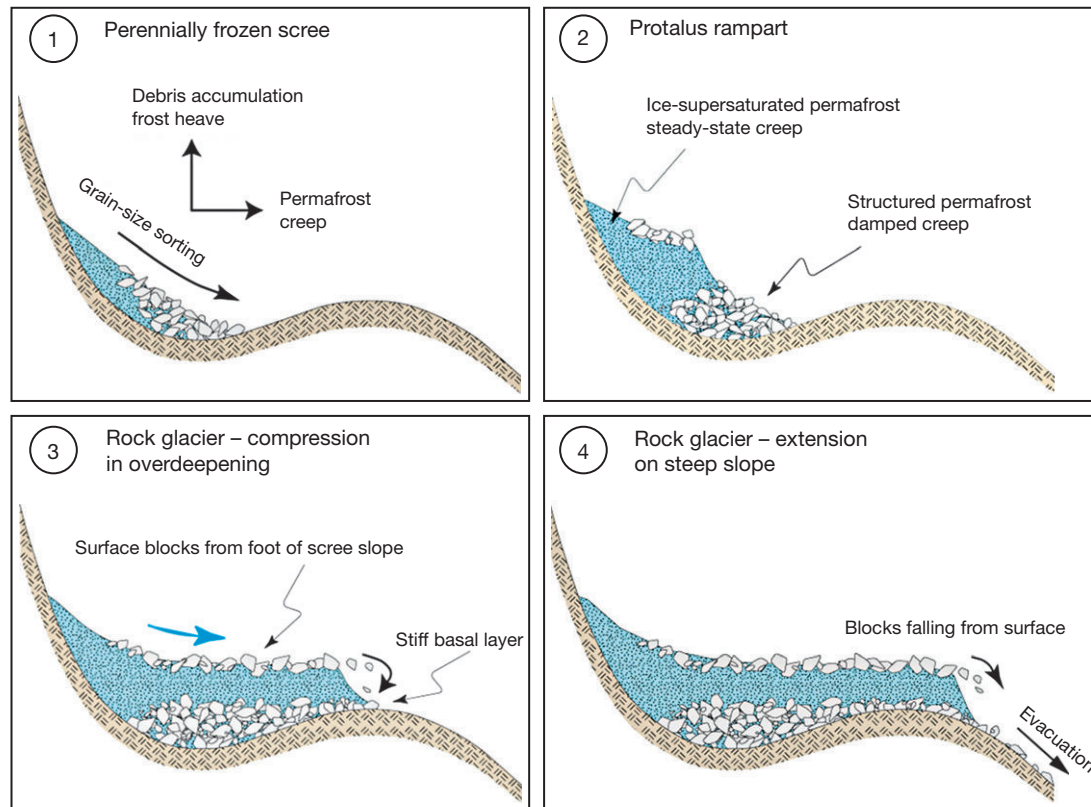


Figure 9 Development stages and structure of an idealized rock glacier. Particle-size sorting on scree slopes concentrates finer particles on the upper slope sections (1) that become supersaturated with ice and start to deform at higher rates than coarser particles, which are not ice-supersaturated, at the slope base (2). Further debris from the headwall accumulates preferentially on the resulting ramp (2) and rafts downslope above the creeping frozen body (3). This blocky upper layer is transported over the rock glacier front, overridden, and eventually reincorporated into the rock glacier as its basal layer (3), unless debris falls over a terrain ridge and thus leaves the rock glacier system (4).

and thermal conditions, rock glaciers, in contrast to glaciers, cannot retreat. In the case of a total cessation of mass supply, a rock glacier will remain stagnant in its horizontal extent, but it will not retreat. The main process of mass loss is the melt-out of ice at the rock glacier front, where the thermally protecting debris cover becomes thinner than the thaw depth due to frontal erosion. In addition, the frontal grain-size sorting leads to smaller grain sizes in the upper part of the front, which reduces the insulation effect in comparison to the coarser debris at the rock glacier surface. Further processes of mass loss of rock glaciers include evacuation of debris, erosion and rockfall over steep terrain, debris flows, or fluvial and coastal erosion. Owing to the melt-out of ice, the potential loss of nonice debris, and the decrease of horizontal speeds with depth, the advance rate of rock glaciers is significantly smaller than their surface speed. Advance rates measured so far range from a few centimeters to decimeters per year.

Shape and Surface Morphology

Rock glaciers can be up to several hundred meters wide and a few kilometers long. Their altitudinal extent may attain several hundred meters, and their thickness may attain several tens of meters. Their overall shape and extent reflects the creep of the

mass, usually down the steepest slopes. The ratios between rock glacier length and width vary significantly, with values both above and below 1. The latter ratio reflects a number of internal conditions such as viscosity, and external factors such as topography or mass supply.

Many rock glaciers show a characteristic surface morphology consisting of transverse and/or longitudinal ridges and furrows. This morphology, in addition to the overall shape, makes rock glaciers resemble lava streams. So far, little is known about the origin of the rock glacier surface morphology.

According to the main hypotheses, the formation of transverse ridges and furrows on rock glaciers might be attributed to (a) external factors, such as variations in debris input or climate conditions; or (b) internal factors, such as the dynamic evolution of ridges under compressive flow, or disturbance propagation from differential movement of discrete layers. In addition, it is unknown whether (c) transverse ridges and furrows represent dynamic structures, for example, forming from emergence under compressive flow regime (i.e., active formation); or (d) if these forms are thermally derived due to, for instance, differential melt (thermokarst) or frost heave processes (i.e., passive formation). It is important to note that these four elements of hypotheses do not necessarily contradict one another. One possible process might enhance the other, or a combination might be necessary for the formation of

microtopographic structures. A number of studies suggest that compressive flow is a major process involved in the formation of transverse ridges (Kääb and Weber, 2004).

The high thermal inertia of rock glaciers allows these frozen bodies to continuously deform and preserves their topography over centuries and millennia (Haeblerli, 2000; Kääb et al., 2003). In addition, the debris content and the ice-free, blocky active layer of rock glaciers contribute to the conservation of surface features even over periods of ice melt. The surface topography of rock glaciers cumulatively reflects the dynamic history and thus, in a complex way, their present and past internal conditions and environment.

Activity and Age

The degree of activity of rock glaciers is usually classified into three different types: (i) active, (ii) inactive, and (iii) relict (Ikeda and Matsuoka, 2002). It should, however, be noted that this classification is a theoretical concept. The transition between the three stages is actually continuous.

Active rock glaciers have a coherent core of ice-rich frozen debris, the deformation of which leads to the rock glacier creep, with rates of several centimeters to meters per year. The permafrost conditions of the body are intact, with perennially negative ground temperatures throughout the entire column, except for the seasonally thawed active layer. The continuous creep and related debris transport along the surface leads to the characteristic steep front with freshly exposed fines in the upper part and coarser debris in the lower part, vertically sorted by grain size due to debris fall along the front (see section 'Kinematics and Advance'). The continuous deformation of the rock glacier surface, rotation and possibly related uplift of individual blocks, and the outwash of fines toward the permafrost table allow only a few specialized plant species to grow and prevent lichens from reaching great age and size (Burga et al., 2004; Haeblerli et al., 1979; Sloan and Dyke, 1998).

Inactive rock glaciers have no coherent ice core and thus no coherent velocity field due to the lack of stress transfer. They show no significant horizontal speed, except perhaps local movement, for example, due to thermokarst processes, patches of ice-rich ground, or erosion processes. Inactive rock glaciers contain a thinner permafrost body and a thicker active layer than active ones (see section 'Environmental Change and Climatic Significance'). They can be transitional forms between active and relict rock glaciers. However, indications have also been found that active rock glaciers might become inactive for certain periods – for example, due to climatic changes – but reactivate after environmental conditions become suitable again. Owing to their kinematic inactivity, inactive rock glaciers or inactive parts of rock glacier complexes usually show more vegetation cover, and more and larger lichens compared to active rock glaciers under a similar environmental setting. In particular, the frontal parts with fines exposed are preferred locations for the invasion of vegetation. Owing to the lack of material supply toward and over the front, erosion is able to reduce the frontal slope.

No permafrost or ice is present within relict rock glaciers. On account of this volume loss of excess ice, rock glacier sections with formerly higher debris content, such as ridges

and the rock glacier margins, are elevated above the rock glacier sections with formerly higher ice contents (Figure 5); thus, the original debris concentration of the active phase is still reflected in the surface topography. The characteristic ridges on rock glaciers can often also be recognized on relict rock glaciers. Similarly, the frontal zones, which in active rock glaciers contain less ice, become visible as terminal ridges. Relict rock glaciers might be covered by dense vegetation, even trees. In some rock glaciers with clear relict characteristics, patchy permafrost and ground-ice occurrences and even gravitational movements on the order of millimeters per year have been found, pointing to continuous transitions between the stages of rock glacier activity (Strozzi et al., 2004).

A number of studies have shown that the age of active rock glaciers is on the order of some to many millennia. The dating methods used range from surface dating techniques (Frauenfelder et al., 2004; Laustella et al., 2003), ^{14}C dating of organic material recovered from boreholes (Haeblerli et al., 1999), flowline calculations (Frauenfelder et al., 2004; Kääb et al., 1997, 1998), and lichen measurements (Haeblerli et al., 1979; Sloan and Dyke, 1998) to paleoclimatic considerations and numerical models (Frauenfelder et al., 2001, 2008). In the European Alps, today's relict rock glaciers are believed to have formed during the last Late Glacial period, though much older forms might exist at places where the climate conditions permitted continuous evolution, and glaciations did not interrupt this evolution.

Environmental Change and Climatic Significance

Rock glaciers are intimately dependent on their geoecological environment (Barsch, 1996). Thus, changes in these conditions have a potentially strong effect on the rock glacier system. Millennium-scale climatic changes – for example, temperature shifts such as between the Last Glacial Maximum, the Late Glacial period, and the Holocene – affect, among others, the thermal conditions, and the production and availability of debris. An increase in ground temperatures, possibly above the 0°C threshold, and/or a reduction of headwall weathering and rockfall onto underlying rock glaciers (Olyphant, 1987) are, for instance, most likely responsible for the decline in activity of rock glaciers at the beginning of the Holocene. Similarly, the atmospheric warming trends since the end of the Little Ice Age observed in most cold mountain regions may be responsible for the recent decay of rock glaciers. Indeed, the altitudinal belts where active, inactive, and relict rock glaciers are found suggest that temperature changes are an important driver of changes in rock glacier activity (Frauenfelder et al., 2001, 2008).

Significantly less well investigated, but potentially of similar importance to temperature, is the influence of long-term changes in precipitation regime. Related changes in mass supply or ground temperature may also be due to spatiotemporal changes in snow cover. Changes in the climatic regime may also indirectly favor or hinder rock glacier development. Thus, rock glaciers may be overridden by glaciers in times of climate change or rock glaciers may develop or recover at locations previously occupied by glaciers (see section 'Thermal Conditions'; Kääb and Kneisel, 2006; Maisch et al., 2003).

As yet, too little is known about the exact climatic significance of rock glaciers in order to use them reliably for paleoclimate reconstruction (Brazier et al., 1998; Frauenfelder et al., 2001; Humlum, 1998). The typical climate of modern active rock glaciers seems similar to the conditions found at the equilibrium line altitude of modern glaciers, though it is somewhat drier. Thus, the altitudinal belts where active, inactive, and relict rock glaciers are found can be used as proxies for paleoclimatic conditions that were suitable for active rock glaciers to develop (Frauenfelder et al., 2001). Such reconstruction is, however, substantially complicated by the fact that rock glacier evolution depends not only on climatic conditions, but also on debris supply and interactions with glaciation. For instance, insufficient debris production may inhibit formation of a rock glacier under otherwise suitable geoecological conditions. Or, the topographic and local climate conditions may favor a glacier to occupy the potential geoecological niche of a rock glacier. In addition, former rock glacier forms may have been removed by glacial erosion or covered by modern glaciers. As a result, reconstruction of ancient rock glaciers and their paleoclimate requires the consideration of the complete and complex spatiotemporal history of periglacial, glacial, and mass-wasting processes throughout the existence of the rock glaciers investigated (Frauenfelder et al., 2001; Maisch et al., 2003).

Geostatistical investigations on the basis of surface velocity measurements and climatic parameters, laboratory tests, and numerical modeling show that the deformation rate of rock glaciers is dependent, among other factors (e.g., slope, composition, or thickness), on the ground temperature (Kääb et al., 2007). Warmer rock glaciers show, in general, higher surface velocities than colder ones do. Consequently, ground-temperature warming as a result of atmospheric warming or changes in snow cover is expected to increase rock glacier deformation. Indeed, early investigations confirm a significant recent acceleration of many rock glaciers in the European Alps where an air temperature increase of close to +1 °C has been observed since the 1980s (Kääb et al., 2007).

From theoretical considerations, rock glacier permafrost will react in three stages to atmospheric warming (Haeberli, 1992): (i) increase in seasonal thaw depth (i.e., active-layer thickness), (ii) ground-temperature warming, and (iii) thermal adjustment toward a new thermal equilibrium accompanied by reduced thickness of the permafrost body (descending permafrost table and rising permafrost base).

See also: **Permafrost and Periglacial Features: Active Layer Processes; Block/Rock Streams; Permafrost; Permafrost and Glacier Interactions; Slope Deposits and Forms.**

References

- Arenson L, Hoelzle M, and Springman S (2002) Borehole deformation measurements and internal structure of some rock glaciers in Switzerland. *Permafrost and Periglacial Processes* 13: 117–135.
- Barsch D (1996) Rockglaciers. *Indicators for the Present and Former Geoecology in High Mountain Environments*. Berlin: Springer.
- Berthling I, Etzelmüller B, Eiken T, and Sollid JL (1998) Rock glaciers on Prins Karls Forland, Svalbard. I. Internal structure, flow velocity and morphology. *Permafrost and Periglacial Processes* 9: 135–145.
- Berthling I, Etzelmüller B, Isaksen K, and Sollid JL (2000) Rock glaciers on Prins Karls Forland. II. GPR soundings and the development of internal structures. *Permafrost and Periglacial Processes* 11: 357–369.
- Brazier V, Kirkbride MP, and Owens IF (1998) The relationship between climate and rock glacier distribution in the Ben Ohau Range, New Zealand. *Geografiska Annaler* 80A: 193–205.
- Bucki AK, Echelmeyer KA, and MacInnes S (2004) The thickness and internal structure of Fireweed rock glacier, Alaska, USA, as determined by geophysical methods. *Journal of Glaciology* 50: 67–75.
- Burga C, Frauenfelder R, Ruffet J, Hoelzle M, and Kääb A (2004) Vegetation on Alpine rock glacier surfaces: A contribution to abundance and dynamics on extreme plant habitats. *Flora* 199: 505–515.
- Etzelmüller B and Frauenfelder R (2009) Factors controlling the distribution of mountain permafrost in the northern hemisphere and their influence on sediment transfer. *Arctic, Antarctic, and Alpine Research* 41: 48–58.
- Frauenfelder R, Haeberli W, Hoelzle M, and Maisch M (2001) Using relict rockglaciers in GIS-based modelling to reconstruct Younger Dryas permafrost distribution patterns in the Err-Julier area, Swiss Alps. *Norwegian Journal of Geography* 55: 195–202.
- Frauenfelder R, Laustela M, and Kääb A (2004) Velocities and relative surface ages of selected Alpine rockglaciers. In: Bezzola G and Sernadeni N (eds.) *Turbulenzen in der Geomorphologie. Jahrestagung der Schweizerischen Geomorphologischen Gesellschaft, Erstfeld*, vol. 184, pp. 103–118. Zurich: Mitteilungen der Versuchsanstalt für Wasserbau, Hydrologie und Glaziologie, ETH Zürich.
- Frauenfelder R, Schneider B, and Kääb A (2008) Using dynamic modelling to simulate the distribution of rockglaciers. *Geomorphology* 93: 130–143.
- Haeberli W (1985) *Creep of Mountain Permafrost*, vol. 77. Zurich: Mitteilungen der Versuchsanstalt für Wasserbau, Hydrologie und Glaziologie der ETH Zürich.
- Haeberli W (1992) Possible effects of climatic change on the evolution of Alpine permafrost. In: Boer M and Koster E (eds.) *Greenhouse-impact on Cold-climate Ecosystems and Landscapes, Catena Supplement*, vol. 22, pp. 23–35. Reiskirchen, Germany: Catena.
- Haeberli W (2000) Modern research perspectives relating to permafrost creep and rock glaciers. *Permafrost and Periglacial Processes* 11: 290–293.
- Haeberli W, Hallet B, Arenson L, et al. (2006) Permafrost creep and rock glacier dynamics. *Permafrost and Periglacial Processes* 17: 189–214.
- Haeberli W, Hoelzle M, Kääb A, Keller F, Vonder Mühll D, and Wagner S (1998) Ten years after drilling through the permafrost of the active rock glacier Murtèl, Eastern Swiss Alps: Answered questions and new perspectives. In: Lewkowicz AG and Allard M (eds.) *Proceedings of the Seventh International Conference on Permafrost*, pp. 403–410. Québec: Centre d'études nordiques, Université Laval.
- Haeberli W, Kääb A, Wagner S, et al. (1999) Pollen analysis and ¹⁴C-age of moss remains recovered from a permafrost core of the active rock glacier Murtèl/Corvatsch (Swiss Alps): Geomorphological and glaciological implications. *Journal of Glaciology* 45: 1–8.
- Haeberli W, King L, and Flotron A (1979) Surface movement and lichen-cover studies at the active rock glacier near the Grubengletscher, Wallis, Swiss Alps. *Arctic and Alpine Research* 11: 421–441.
- Haeberli W and Vonder Mühll D (1996) On the characteristics and possible origins of ice in rock glacier permafrost. *Zeitschrift für Geomorphologie* 104: 43–57.
- Harris C, Arenson LU, Christiansen HH, et al. (2009) Permafrost and climate in Europe: Monitoring and modelling thermal, geomorphological and geotechnical responses. *Earth-Science Reviews* 92: 117–171.
- Hoelzle M, Wagner S, Kääb A, and Vonder Mühll D (1998) Surface movement and internal deformation of ice-rock mixtures within rock glaciers in the Upper Engadin, Switzerland. In: Lewkowicz AG and Allard M (eds.) *Proceedings of the Seventh International Conference on Permafrost*, pp. 465–472. Québec: Centre d'études nordiques, Université Laval.
- Humlum O (1998) The climatic significance of rock glaciers. *Permafrost and Periglacial Processes* 9: 375–395.
- Humlum O, Christiansen HH, and Juliussen H (2007) Avalanche-derived rock glaciers in Svalbard. *Permafrost and Periglacial Processes* 18: 75–88.
- Ikeda A and Matsuoka N (2002) Degradation of talus-derived rock glaciers in the Upper Engadin, Swiss Alps. *Permafrost and Periglacial Processes* 13: 145–161.
- Isaksen K, Ødegard RS, Eiken T, and Sollid JL (2000) Composition, flow and development of two tongue-shaped rock glaciers in the permafrost of Svalbard. *Permafrost and Periglacial Processes* 11: 241–257.
- Kääb A, Frauenfelder R, and Roer I (2007) On the response of rockglacier creep to surface temperature increase. *Global and Planetary Change* 56: 172–187.

- Kääb A, Gudmundsson GH, and Hoelzle M (1998) Surface deformation of creeping mountain permafrost. Photogrammetric investigations on rock glacier Murtèl, Swiss Alps. In: Lewkowicz AG and Allard M (eds.) *Proceedings of the Seventh International Conference on Permafrost*, pp. 531–537. Québec: Centre d'études nordiques, Université Laval.
- Kääb A, Haeberli W, and Gudmundsson GH (1997) Analysing the creep of mountain permafrost using high precision aerial photogrammetry: 25 years of monitoring Gruben rock glacier, Swiss Alps. *Permafrost and Periglacial Processes* 8: 409–426.
- Kääb A, Kaufmann V, Ladstädter R, and Eiken T (2003) Rock glacier dynamics: Implications from high-resolution measurements of surface velocity fields. *Proceedings of the Eighth International Conference on Permafrost*, vol. 1, pp. 501–506. Lisse, The Netherlands: AA. Balkema.
- Kääb A and Kneisel C (2006) Permafrost creep within a recently deglaciated glacier forefield: Muragl, Swiss Alps. *Permafrost and Periglacial Processes* 17: 79–85.
- Kääb A and Reichmuth T (2005) Advance mechanisms of rockglaciers. *Permafrost and Periglacial Processes* 16: 187–193.
- Kääb A and Weber M (2004) Development of transverse ridges on rockglaciers. Field measurements and laboratory experiments. *Permafrost and Periglacial Processes* 15: 379–391.
- Kaufmann V and Ladstädter R (2002) Spatio-temporal analysis of the dynamic behaviour of the Hohebenkar rock glaciers (Oetztal Alps, Austria) by means of digital photogrammetric methods. *Grazer Schriften der Geographie und Raumforschung* 37: 119–140.
- Lambiel C and Delaloye R (2004) Contribution of real-time kinematic GPS in the study of creeping mountain permafrost: Examples from the Western Swiss Alps. *Permafrost and Periglacial Processes* 15: 229–241.
- Laustella M, Egli M, Frauenfelder R, Kääb A, Maisch M, and Haeberli W (2003) Weathering rind measurements and relative age dating of rockglacier surfaces in crystalline regions of the Eastern Swiss Alps. *Proceedings of the Eighth International Conference on Permafrost*, vol. 1, pp. 627–632. Lisse, The Netherlands: AA. Balkema.
- Maisch M, Haeberli W, Frauenfelder R, and Kääb A (2003) Lateglacial and Holocene evolution of glaciers and permafrost in the Val Muragl, Upper Engadine, Swiss Alps. *Proceedings of the Eighth International Conference on Permafrost*, vol. 1, pp. 717–722. Lisse, The Netherlands: AA. Balkema.
- Martin HE and Whalley WB (1987) Rock glaciers. Part I. Rock glacier morphology: Classification and distribution. *Progress in Physical Geography* 11: 260–282.
- Olyphant GA (1987) Rock glacier response to abrupt changes in talus production. In: Giardino JR, Shroder JF, and Vitek JD (eds.) *Rock Glaciers*, pp. 55–64. Winchester, MA: Allen and Unwin.
- Sloan VF and Dyke LD (1998) Decadal and millennial velocities of rock glaciers, Selwyn Mountains, Canada. *Geografiska Annaler* 80A: 237–249.
- Steig EJ, Fitzpatrick JJ, Potter N, and Clark DH (1998) The geochemical record in rock glaciers. *Geografiska Annaler* 80A: 277–286.
- Strozzi T, Kääb A, and Frauenfelder R (2004) Detecting and quantifying mountain permafrost creep from in-situ, airborne and spaceborne remote sensing methods. *International Journal of Remote Sensing* 25: 2919–2931.
- Vonder Mühll DS, Arenson LU, and Springman SM (2003) Temperature conditions in two Alpine rock glaciers. *Proceedings of the Eighth International Conference on Permafrost*, vol. 2, pp. 1195–1200. Lisse, The Netherlands: AA. Balkema.

Yedoma: Late Pleistocene Ice-Rich Syngenetic Permafrost of Beringia

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Introduction

Beringia represents the largest contiguous area of the Arctic that remained unglaciated during the late Pliocene and Pleistocene. [Hulten \(1937\)](#), the Swedish biogeographer, invoked the concept of Beringia to explain the distribution of Arctic and boreal plants around the Bering Strait. He originally considered Beringia to include only the area of the continental shelf exposed when lowered sea level connected eastern Asia with North America. Today, Beringia is considered more broadly to extend from the Taymyr Peninsula through Alaska to the Yukon Territory, united by aspects of the cold-adapted flora and fauna of the region and the lack of glaciation. This area was located between the Eurasian Ice Sheet to the west and the Laurentide and Cordilleran ice sheets to the east ([Figure 1](#)). Extreme continental climate conditions were connected to the presence of a largely exposed continental shelf, a likely perennial Arctic sea-ice cover, and the existence of large ice sheets to the east and west. By virtue of the lack of glaciation, much of Beringia preserves an exceptional sedimentary record of Pleistocene environmental change.

Across this broad region, vast tracts of ice-rich and largely fine-grained sediments are present. These deposits are found in the Arctic and Subarctic areas of Beringia, and locally exceed 50 m in thickness. The deposits, owing to the presence of permafrost, preserve a diversity of exceptional paleoenvironmental archives, including mammals (e.g., [Guthrie, 1990](#)), paleobotanical remains (e.g., [Gitterman et al., 1982](#)), ancient DNA (e.g., [Shapiro and Cooper, 2003](#)), and relict ice (e.g., [Kaplina and Lozhkin, 1984](#)), and host a globally significant, and potentially vulnerable, reservoir of organic carbon (e.g., [Zimov et al., 2006](#)). While many aspects of their composition and paleoenvironmental history are agreed upon, details of their formation lack consensus, particularly regarding the strict role of eolian processes. These differences are most pronounced among workers in western Beringia (Siberia), while in eastern Beringia (Alaska and Yukon), a general consensus exists. This article reviews the history of research, terminology, characteristics, and differing ideas on the formation of the ice-rich deposits and their paleoenvironmental significance.

Terminology

The terminology for the ice-rich, syngenetically frozen deposits varies across Beringia. These deposits are often referred to as 'Ice Complex' (in Russian: *ledovyy kompleks*) or 'Yedoma' in Siberia, while in North America, other terms such as 'muck' or retransported silt could be partially considered as equivalent terms ([Froese et al., 2009](#); [Péwé, 1975](#); [Shur et al., 2004](#)).

[Solov'ev \(1959\)](#) first defined Ice Complex as frozen deposits of various age, composition, genesis, and thickness, with numerous ice wedges. Ice Complex deposits of late Pleistocene age are widely distributed on the East Siberian coastal plains (Lena-Anabar, Yana-Indigirka, and Kolyma lowlands) as well as on the New Siberian Archipelago, but they also occur in more southerly regions of East Siberia along large river valleys ([Figure 1](#)).

According to [Sher \(1997\)](#), at least three different meanings of the term 'Yedoma' exist in the Russian literature: (1) a 'Yedoma surface' in the geomorphic sense, (2) a 'Yedoma Suite' in the stratigraphic sense, or (3) a cryolithological feature implying a special kind of frozen sediment, widely distributed in Beringia ('cryolithology' is a Russian term, not widely used in North America, for the study of the genesis, structure, and lithology of frozen earth materials; [van Everdingen, 1998](#)). The latter concept of Yedoma is the one used in this article, encompassing distinctive ice-rich silts and silty sand penetrated by large ice wedges, resulting from sedimentation and syngenetic freezing, and driven by certain climatic and environmental conditions during the late Pleistocene. The term 'Ice Complex' is often used synonymously with 'Yedoma' in this sense.

The term 'Yedoma' was originally used in the geomorphic sense to describe the hills separating thermokarst depressions in northeastern Siberia, especially in the Yana-Indigirka and Kolyma lowlands (e.g., [Tomirdiario, 1996](#)). These surfaces were considered to be former accumulation plains, composed of late Pleistocene Ice Complex deposits, cross-cut to varying extents by thermokarst basins. In this sense, Yedoma can also describe a periglacial relief type formed by thermokarst processes ([Solov'ev, 1959](#)).

Initially, the stratigraphic term 'Yedoma Suite' was adopted for middle Pleistocene horizons in the northeastern Siberian lowlands ([Vas'kovsky, 1963](#)). Based on cryolithological, paleoecological, and geochronological studies at the key site of Duvanny Yar on the lower Kolyma River ([Figure 1](#)), the stratigraphic position of the 'Yedoma Suite' was considered to be only late Pleistocene ([Sher et al., 1987](#)). The upper late Pleistocene horizon of Duvanny Yar represents the stratotype of the 'Yedoma Suite' (e.g., [Zanina et al., 2011](#)), recorded by a 40- to 50-m-thick Ice Complex sequence that is characterized by heterogeneous gray-brown sandy silt with a layered cryostructure and syngenetic ice wedges up to tens of meters high. The sediments contain grass roots and lenses of fine-grained plant detritus as well as paleosols ([Zanina et al., 2011](#)). Fossil teeth and bones of the late Pleistocene megafauna are common ([Sher et al., 1979](#)). Radiocarbon dates from Duvanny Yar suggest that much of the Ice Complex formed during the late Pleistocene from ca. 40 and 13 ka (^{14}C) BP ([Vasil'chuk et al., 2001](#)).

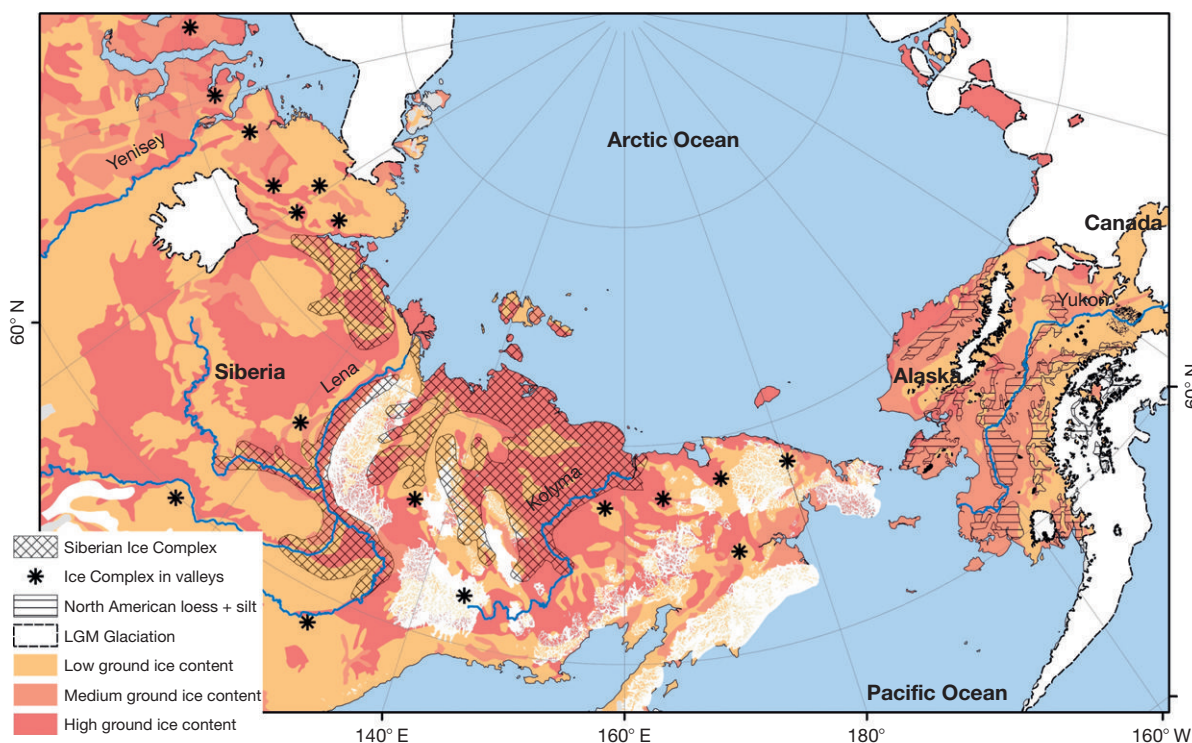


Figure 1 Distribution of ice-rich permafrost deposits in Arctic and Subarctic lowlands in the region of late Pleistocene Beringia (according to Romanovskii, 1993, for Siberian Ice Complex; Wolfe et al., 2009, for North American loess; Ehlers and Gibbard, 2003, for LGM Glaciation; and Brown et al., 1997, for ground ice content).

In North America, where the research history of ice-rich deposits is closely related to placer gold-mining activities, they are often given the name ‘muck’ as they are a common overburden for gold-bearing fluvial deposits (Fraser and Burn, 1997; Péwé, 1975). However, the term ‘muck’ originates from the gold mining vernacular and is generally not used in a strict genetic or stratigraphic sense, and thus, it is not restricted to late Pleistocene ice-rich deposits. For example, Péwé (1975) indicated that ‘muck’ can also encompass Holocene ice-rich silts, which would be inconsistent with Yedoma as commonly defined. In recent work, the term ‘muck’ is therefore not frequently used, but instead these deposits are typically referred to as simply ice-rich silts or perennally frozen silts, emphasizing both their typical grain size and the presence of permafrost.

In North America, syngenetic ice-rich silt deposits, thought to be broad equivalents to the Siberian examples, occur in lower segments of the Arctic Foothills, in the northern part of Seward Peninsula, in numerous areas in Interior Alaska (Kanevskiy et al., 2011), and the Klondike region of Yukon Territory (Froese et al., 2009). The Klondike sites, by virtue of the gold-mining activities, have created excellent exposures for study (Figure 2). Similar to the Russian examples, Yedoma deposits in Yukon and Alaska are dominated by silt-sized particles, abundant organic matter, ice-rich syngenetic permafrost, and ice wedges and may reach 40 m in thickness. These deposits, when wet, are commonly nearly black in color from the abundant organic matter. Perhaps their most distinctive property, however, is the pungent odor they give off from the incompletely decomposed organic matter that is exposed



Figure 2 Syngenetic ice wedges with full and partial thaw surfaces. The presence of shoulders on the ice wedges (white arrows) indicates the position of the paleoactive layer and the nature of episodic sedimentation of the loessal silts on the surface of an ice-wedge polygonal ground network; to either side of the shoulders, some ice wedges are fully truncated at these levels. Sediments date to ca. 30 ka BP, Quartz Creek, Klondike area, central Yukon. Photo shows a vertical section ca. 5–6 m high. Photograph by D. Froese.

as they thaw. Although these deposits are largely of late Pleistocene age, the presence of distal, dateable volcanic ash (tephra) within some of them indicates a middle Pleistocene age (Froese et al., 2008, 2009; Reyes et al., 2010).

Following Sher (1997) and Kanevskiy et al. (2011), this article uses the term ‘Yedoma’ to describe ice-rich syngenetic permafrost that developed extensively under late Pleistocene

cold-climate conditions in unglaciated regions of Eurasia and North America.

Distribution

Yedoma deposits are widely distributed across Beringia (Figure 1). In eastern Siberia, their presence in lowlands is reflected in the modern distribution of thermokarst lakes developed on ice-rich permafrost (Figure 3). Thermokarst depressions with lakes and thermoerosional valleys shape the modern lowland relief and dissect remnants of late Pleistocene accumulation plains into Yedoma hills and uplands. Other indicators of the presence of Yedoma deposits are thaw slumps on coastal bluffs several hundred meters wide and several tens of meters deep (Figure 4), lake shores, and river banks. These sites

expose thermokarst mounds as well as large ice wedges 2–5 m wide and up to tens of meters high. These ice wedges, associated sediments, and thermokarst mounds (Figures 5 and 6) are parts of former polygonal ice-wedge systems that developed over long time periods. Generally, these exposures are formed by active thermoerosion and slumping of ice-rich permafrost, exposing sediments and ground ice in rapidly retreating steep bluffs and terraces, and allowing access to still-frozen sediments in vertical profiles.

The distribution of Yedoma in Northeast Siberia relates to a number of distinct geomorphological settings. One type of setting fringes low-elevation coastal mountains, which served as the main sediment sources for Yedoma. Heavy-mineral studies suggest a dominantly local origin for the sediments rather than long-distance transport of the silts which make up the main clastic component of Yedoma. Gently inclined

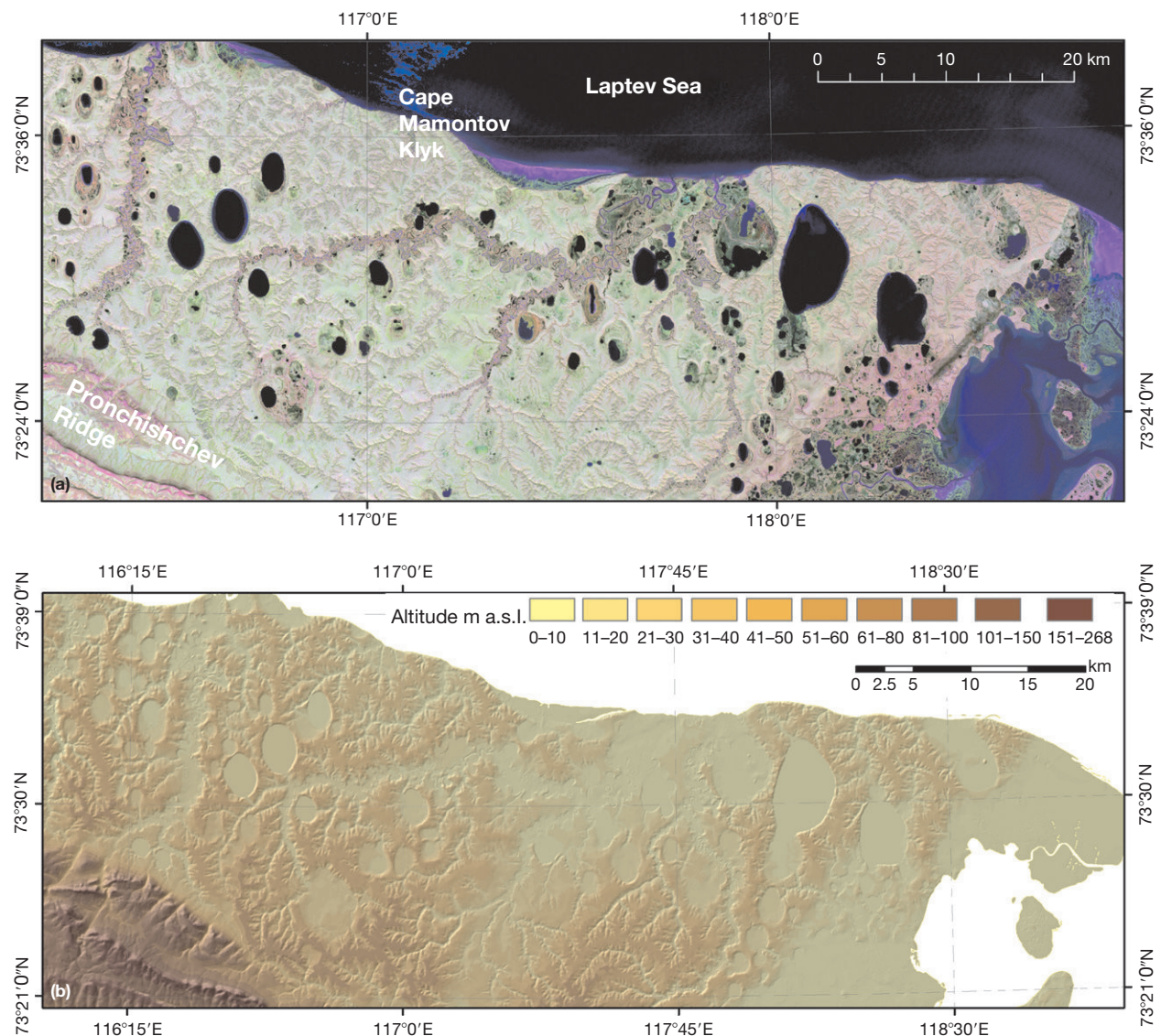


Figure 3 Yedoma landscape at the western Laptev Sea coastal plain in front of the Pronchishchev Ridge characterized by Yedoma hills, thermokarst basins, and numerous small-branched thermoerosional valley systems and larger river valleys: (a) satellite image Landsat-7 ETM+ and (b) digital elevation model (adapted from Grosse et al., 2006).



Figure 4 Yedoma hill with a retrogressive thaw slump ca. 300 m long and 80 m wide at the southern coast of Bol'shoy Lyakhovsky Island (New Siberian Archipelago). The slump headwall exposing large ice wedges is ca. 25 m high. Photograph by S. Wetterich.



Figure 5 Syngenetic ice wedges and intrapolygon frozen sediment columns of a Yedoma exposure at the southern coast of Bol'shoy Lyakhovsky Island (New Siberian Archipelago). Photograph by V. Tumskoy. Person for scale.



Figure 6 Syngenetic ice wedges and intrapolygon frozen sediment columns of a Yedoma exposure in the Colville River Valley (Alaska North Slope). Photograph by G. Grosse. People for scale.

foreland accumulation plains underlain by widespread ice-wedge polygon systems were the original landscape for the accumulation of Yedoma deposits in the North Siberian lowlands during the late Pleistocene. A second geomorphologic setting in the hinterland of Yedoma areas comprises separate bedrock elevations (100–400 m high), for example, the granite domes on Bol'shoy Lyakhovsky Island (New Siberian Archipelago) and Cape Svyatoy Nos (Laptev Sea), and short fault ranges as found on Bel'kovsky, Stolbovoy, and Kotel'ny islands of the New Siberian Archipelago. These elevations are often shaped by cryoplanation terraces (**Figure 7(a)**) and framed by cryopediment areas, indicating strong periglacial weathering processes (**Figure 7(b)**). A third setting is found in extended lowland areas further away from mountain ranges. Although these distal areas are currently dominated by large and numerous thermokarst basins and lakes, Yedoma hills are commonly present between them, indicating their former widespread occurrence, such as in the Yana-Indigirka lowland. Yedoma deposits are also present in Central Yakutia in terraces of the lower parts of large mountain valleys.

All aggradational landscape types where Yedoma occurs are characterized by large, relatively flat surfaces with low hydrological gradients that are conducive to the long-term formation of syngenetic ice-wedge polygon systems. Poorly developed drainage in the late Pleistocene environment of Beringia, with its cold-arid climate, was presumably an important factor for Yedoma formation. It is likely that accumulation of Yedoma deposits also took place on vast areas of the Beringian shelf that were exposed during the late Pleistocene ([Romanovskii et al., 2000](#)).

Broad equivalents to Siberian Yedoma also occur in the Arctic and Subarctic regions of unglaciated Alaska and Yukon, or eastern Beringia. In tundra areas of northern Alaska,



Figure 7 Hinterland of Ice Complex accumulation areas: (a) Cryoplanation terraces of the Cretaceous granite dome at Cape Shalaurova on Bol'shoy Lyakhovsky Island, New Siberian Archipelago, shaped by intense periglacial weathering, and (b) Cretaceous granite domes on Cape Svyatoy Nos, Laptev Sea, surrounded by cryopediments. Photographs by S. Wetterich and H. Meyer, AWI Potsdam.

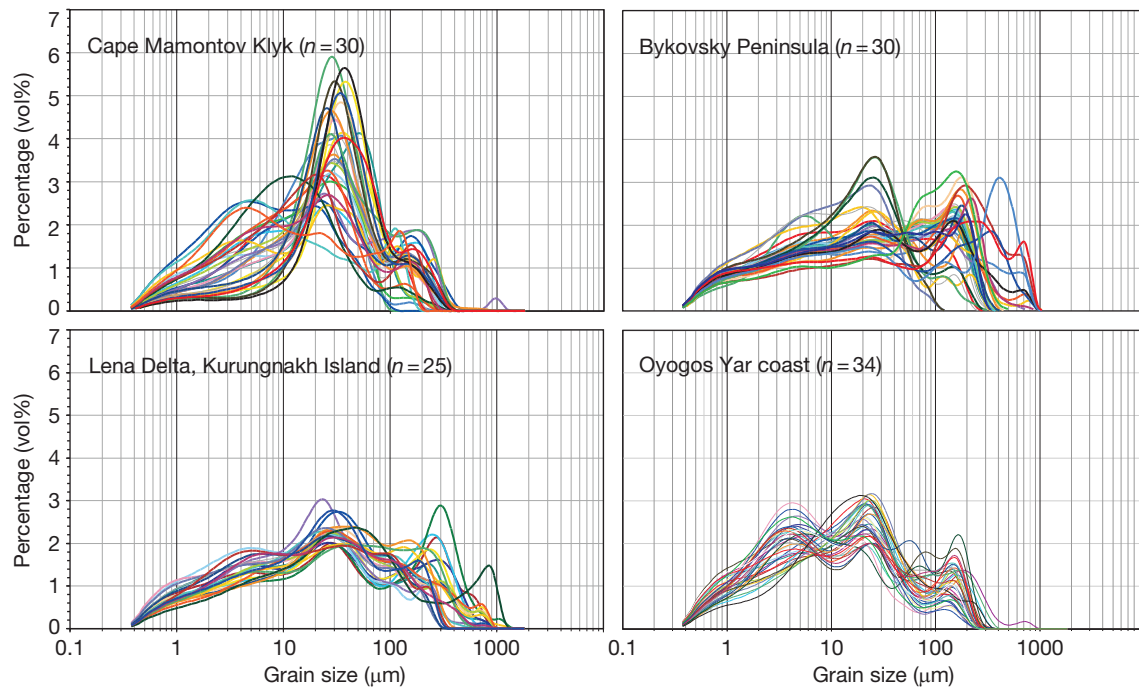


Figure 8 Grain-size distribution curves from Yedoma deposits of different sites in the Laptev Sea region according to Schirmer et al. (2011b).

Yedoma is readily recognized where deep thermokarst lake basins are inset into thick ice-rich permafrost. In areas further south, within the modern boreal forest, Yedoma is obscured by forest cover. In general, the interior of Yukon and Alaska includes Yedoma deposits in areas of greater relief than their counterparts in Siberia, and Yedoma occurs further south into the present-day discontinuous permafrost zone (50–90% underlain by permafrost). In the Fairbanks and Klondike regions, where these deposits are best characterized, their distribution strongly reflects local site conditions. In these areas, Yedoma is found on north- and east-facing slopes and within narrow valleys along hillslopes.

Cryolithology

Yedoma sequences include buried cryosols marked by brownish horizons as well as the inclusion of organic matter and peat (Figures 5 and 6). Evidence of past cryoturbation, highlighted by the presence of the organic matter, is common. The poorly sorted clastic sediments range in grain size dominantly from silt to fine-grained sand and sometimes contain coarser-grained sand and gravel (Trask sorting 1.5–9, mean grain size 10–300 μm). Grain-size characteristics differ from site to site as well as within horizons, but modal grain size of nearly all samples is within the silt and fine sand range (Fraser and Burn, 1997; Sanborn et al., 2006). The presence of multimode grain-size distributions (Figure 8) suggests a range of transport processes and underscores the importance of redeposition of silts with coarser grain sizes. Heavy-mineral analyses of Siberian Yedoma suggest significant differences in detrital composition between sites, indicating different local sediment sources of the hinterland (Schirmer et al., 2011b).



Figure 9 Mineral sediment with ice bands and fine lens-like reticulate cryostructures and some brownish plant detritus-rich patches typical of Yedoma deposits from the Buor Khaya Peninsula, Laptev Sea. Photograph by J. Strauss, AWI Potsdam.

Cryogenic structures in Yedoma deposits, in particular, the presence of syngenetic ice wedges, are similar at most study sites. Ice wedges often have pronounced shoulders and partial thaw surfaces, indicating the episodic nature of sedimentation and permafrost aggradation with varying active-layer depth on the polygonal ground network (Figure 2). Thin horizontal ice lenses or net-like reticulated ice veins subdivided by up to 2–4-cm-thick ice bands are common (Figure 9); they indicate enrichment in segregated ice near the permafrost table under subaerial conditions, including inundated polygonal ponds, by a slowly aggrading soil surface, changing active-layer thickness, and freezing under poorly drained conditions (e.g., Romanovskii, 1977).

The frozen sediment sequences commonly contain excess ice, with gravimetric ice contents (ratio of the mass of liquid water and ice in a sample to the dry mass of the sample, expressed as a mass percentage, [van Everdingen, 1998](#)) of 70 to >100 wt% ([Schirmer et al., 2011b](#)); this corresponds to an ice content of 30 to more than 60 wt% for the sediment columns of Yedoma exposures. Considering that ice wedges



Figure 10 Graminoid-rich paleosol with roots overlying ice wedge, marking the paleoactive layer when the soil was formed. The paleosol is cross-cut by a syngenetic ice wedge (vertical arrow at left) and includes beds of Dominion Creek tephra (82 ± 9 ka; horizontal arrow), marking early MIS 4 cold conditions in Yukon. Ice axe is 80 cm long. Adapted from Froese et al. (2009).

account for about 50% by volume of most Yedoma sequences, the total volumetric ground ice content likely ranges from 65 to 90%.

Stable isotope signatures of ground ice ($\delta^{18}\text{O}$, δD , d-excess) are used as a proxy for winter paleotemperatures (the most negative $\delta^{18}\text{O}$ and δD values reflect the coldest temperatures) and moisture sources. Late Pleistocene ice-wedge ice reflects very cold winter temperatures and moisture sources that are isotopically distinct from Holocene values (e.g., [Meyer et al., 2002](#); [Vasil'chuk, 1992](#); [Wetterich et al., 2011](#)).

Total organic carbon (TOC) content in Yedoma deposits, based on more than 700 analyses from 14 Siberian sites, is variable (<1 to >20 wt%) and relatively high (site-specific averages from 1.2 to 4.8 wt%; [Schirmer et al., 2011a,b](#)). Woody macrofossils and peat are present, along with numerous small filamentous rootlets and dispersed organic detritus. Detailed studies of paleosols within Yedoma deposits indicate that much of the organic matter accreted as root detritus from graminoid vegetation under periglacial conditions ([Figure 10](#); [Sanborn et al., 2006](#)). Thaw unconformities truncating ice wedges, syngenetic ice wedges with pronounced shoulders ([Figure 2](#)), and cryoturbated paleosols, all record episodic changes in sedimentation and paleoactive-layer depths. The $\delta^{13}\text{C}$ values of bulk organic material range from about -29 to -24 ‰, indicating that only freshwater aquatic and subaerial terrestrial environments contributed this material to the formation of Yedoma deposits ([Schirmer et al., 2011b](#)). Variations in TOC content, C/N ratio, and $\delta^{13}\text{C}$ values ([Figure 11](#)) relate to changes in bioproductivity, intensity and character of cryosol formation, and different degrees of organic matter

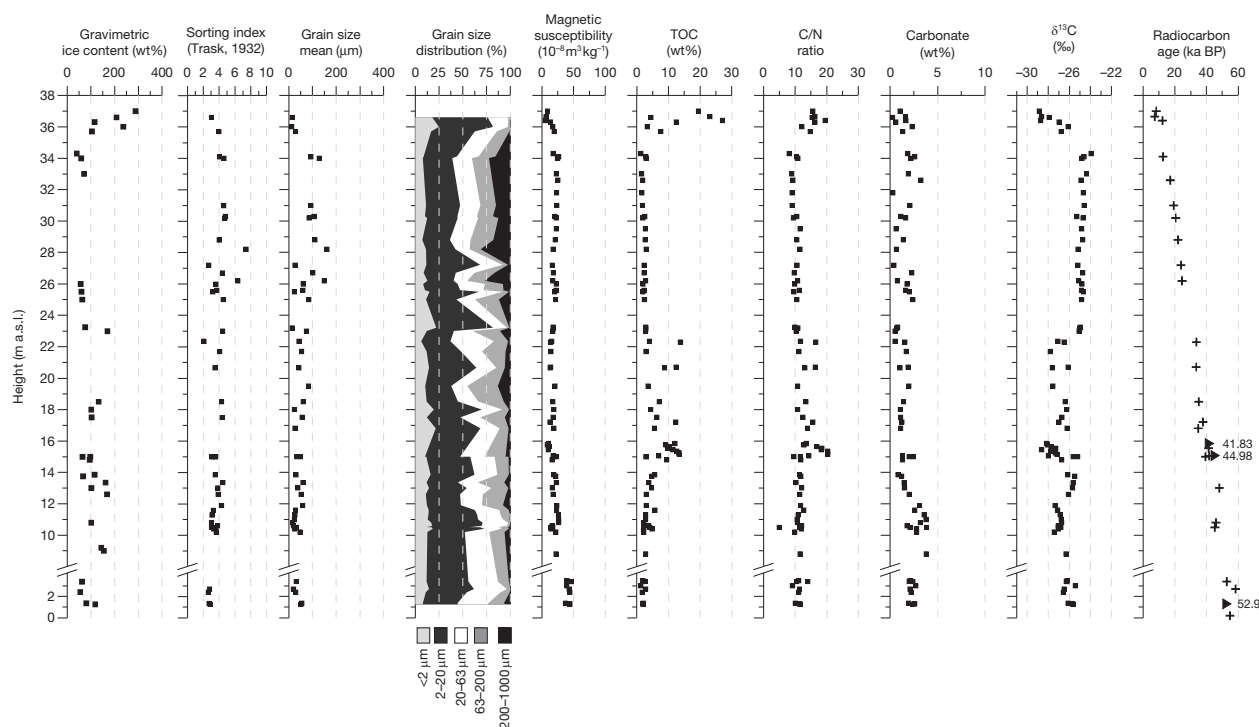


Figure 11 Compilation of cryolithological parameters of the Yedoma sequence and the Holocene cover from Bykovsky Peninsula, Laptev Sea according to [Schirmer et al. \(2011b\)](#).

decomposition, as well as variations in plant associations. High TOC contents, high C/N ratios, and low $\delta^{13}\text{C}$ values reflect less-decomposed organic matter under anaerobic conditions, which were characteristic of the interstadial Marine Isotope Stage (MIS) 3. During cold stages, such as MIS 2, TOC was less variable and generally low, indicating more stable environments with reduced bioproductivity and low C/N ratios. High $\delta^{13}\text{C}$ values reflect relatively dry, aerobic conditions.

Fossils and Paleoenvironmental Archives

Yedoma deposits are among the most productive hosts for preserving Pleistocene fossils in the northern hemisphere (Guthrie, 1990), and because of the presence of permafrost, they provide exceptional preservation conditions, unequaled elsewhere (Shapiro and Cooper, 2003). Intense research over the past decades has already been undertaken to use the fossil record to describe past environments of Beringia through a combination of proxies such as pollen, soils, plant macrofossils, insects, mammals, ostracods, or testate amoebae for quantitative and qualitative reconstructions (e.g., Sher et al., 2005; Wetterich et al., 2008; Zazula et al., 2007).

Across Alaska, Yukon, and Siberia, the mammoth fauna remains preserved in Yedoma are dominated by mammoth (*Mammuthus primigenius*), horse (*Equus caballus*), bison (*Bison priscus*), and caribou (*Rangifer tarandus*). Fossils of less common species are recovered occasionally, including camel (*Camelops hesternus*), musk ox (*Ovibos pallantis*), woolly rhino (*Coelodonta antiquitatis*), and saiga antelope (*Saiga* sp.). Carnivores such as cave lion (*Panthera leo spelaea*), short-faced bear (*Arctodus simus*), and wolf (*Canis lupus*) are also rare. Some of these species occurred only on one side of the Bering Land Bridge (such as the wholly rhino in Siberia, and camel and short-faced bear in eastern Beringia), while others were widely distributed across Beringia. The grazing fauna, much of it collected from detrital fossils recovered from Yedoma exposures across Beringia, have been linked to late Pleistocene environments of the region through radiocarbon dating (e.g., Sher et al., 2005). Mummified or freeze-dried carcasses recovered from these sites, although rare, highlight the role of permafrost in preserving the late Pleistocene paleontological record (Guthrie, 1990; Shapiro and Cooper, 2003).

Paleobotanical materials are equally well preserved within Yedoma and, in particular, have become better understood through plant macrofossils recovered from these deposits. For example, in eastern Beringia, plant macrofossils from Arctic ground squirrel (*Spermophilus parryii*) nests, seed caches, and burrows are dominated by grasses, dryland sedges, sage, and a wide variety of flowering forbs (e.g., Zazula et al., 2007). Together, these plants indicate open grass- and forb-rich steppe tundra in eastern Beringia, developed on largely well-drained substrates during MIS 2 (Froese et al., 2009).

Palynological analyses from the Yedoma indicate a dominance of herbs (e.g., *Artemisia*, Caryophyllaceae, Asteraceae), grasses (e.g., Poaceae, Cyperaceae), and to a lesser extent, dwarf shrubs (e.g., *Salix*, *Alnus*, *Betula*). According to Andreev et al. (2011), the period of Yedoma formation in Siberia was characterized by sparse grass-sedge tundra growing in an extremely cold and dry climate during MIS 4, tundra-steppe

vegetation with higher productivity in a more moderate and humid climate during MIS 3, and sparse grass-tundra present in the cooler and drier climate of MIS 2. Nevertheless, the presence of algal spores and remains of other aquatic organisms (e.g., hydrophytes, ostracods) found in many samples indicates the existence of small freshwater ponds in low-center polygons. In addition, spores from dung-inhabiting fungi indicate the widespread occurrence of mammoth fauna species during the time of Yedoma formation (Sher et al., 2005). Floral and faunal communities that existed when Yedoma formed disappeared approximately at the Pleistocene–Holocene transition. The paleoecosystem associated with Yedoma, often called Mammoth Steppe or steppe tundra in eastern Beringia (e.g., Guthrie, 1990) or Tundra-Steppe in western Beringia (e.g., Yurtsev, 1982), combined elements found in both cold tundra and dry steppe environments. The climate was more continental in the late Pleistocene Arctic than today, and reconstructions from Siberia indicate colder winters and warmer summers, suggesting stronger seasonal gradients in temperature and precipitation (e.g., Andreev et al., 2011). Because of the polygonal patterned ground relief, the tundra-steppe landscape was characterized by a patchwork-like distribution of vegetation communities at a local scale.

Formation Processes

Yedoma in Northeast Siberia (Western Beringia)

The processes that formed Yedoma in western Beringia are disputed, and some differences of interpretation remain between researchers of western and eastern Beringia. These differences largely reflect the prominence attributed to eolian processes, and in particular, the interpretation of the silts as being primarily loess by researchers in Yukon and Alaska. In contrast, workers in Siberia have, in recent decades, proposed several hypotheses about the origin of Yedoma, including alluvial, glaciolacustrine, deltaic, proluvial and colluvial, cryogenic-eolian, nival, and polygenetic processes (for an overview, see Schirmer et al., 2011b). Zhestkova (1982) and Sher (1997) proposed a polygenetic origin in which Yedoma deposits accumulated under different sedimentation regimes, but were largely controlled by similar landscape and relief characteristics, climate conditions, periglacial processes, and sediment sources. An overarching similarity is the presence of large syngenetic ice wedges and remains of the late Pleistocene mammoth fauna and tundra-steppe flora.

A broad concept of Yedoma formation that combines processes of cryogenic weathering, material transport and accumulation, and relief shaping under cold-arid climate conditions was proposed by Kunitsky et al. (2002). According to this cryolithogenetic concept (cryolithogenesis: formation of sedimentary materials in the permafrost regions, according to Yershov, 2004), Yedoma represents a characteristic periglacial facies whose formation is controlled by the interaction of several climatic, landscape, and geological preconditions typical for nonglaciated Arctic lowlands (Figure 12). Initially, mixtures of windblown snow, plant, and fine-grained mineral detritus accumulated in numerous perennial snow fields that were topographically protected not only among hills and low mountain ranges, but also on terrain edges (e.g., steep slopes,

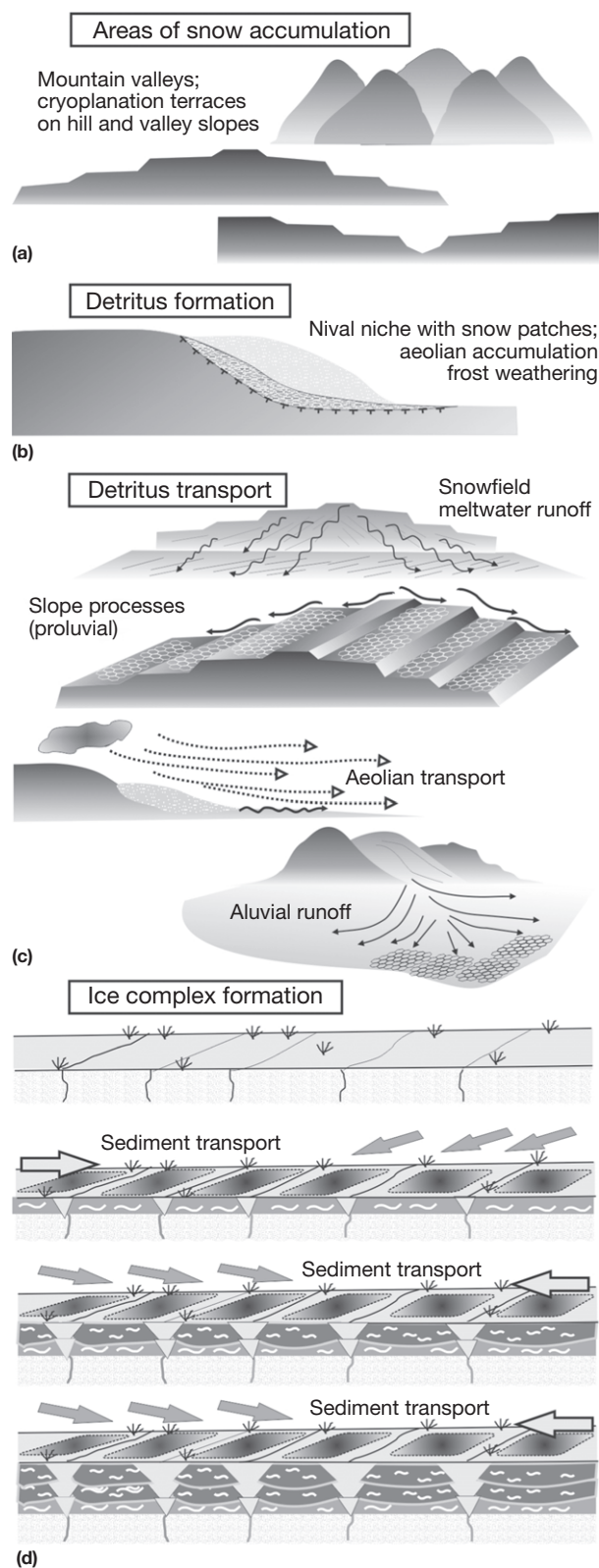


Figure 12 Landscape characteristics and development stages (a–d) during late Pleistocene Yedoma formation in Siberian Arctic lowlands (according to Schirrmeister et al. (2011b)). The arrows in part d indicate the various types of sediment transport described in the text (e.g., alluvial, proluvial, colluvial, eolian).

valleys, and cryoplanation terraces; Figure 12(a)). Around their margins and beneath the snowfields, intense freeze-thaw cycles associated with wet conditions (frost weathering) formed additional fine-grained sediment (Figure 12(b); Konishchev and Rogov, 1993; Kunitsky et al., 2002). Some of this sediment and associated organic matter was transported downslope by runoff from seasonal meltwater (Figure 12(c)). This fine-grained detritus was then transported further by a variety of processes (alluvial, proluvial, colluvial, slope wash, solifluction, permafrost creep, and eolian) to foreland plains, cryopediments, or large alluvial fans. As a result, different types of Yedoma deposits accumulated, with variable grain-size distributions, dependent on local environmental conditions. As sediment aggraded in flat accumulation areas, polygonal ice-wedge systems developed syngenetically (Figure 12(d)). Over thousands of years, these recurrent processes of snowfield accumulation, snow melt, and meltwater transport and secondary sediment transport processes led to the accumulation of thick Yedoma deposits on wide plains. On a smaller scale, similar phenomena can be observed today on numerous slopes and terrain steps in northeastern Siberia, where nivation is an essential geomorphological factor (Kunitsky et al., 2002).

The typical Ice Complex formation therefore consisted of several concurrent cryogenic processes, including sediment accumulation and freezing, ice segregation, syngenetic ice-wedge growth, sediment reworking, peat aggradation, cryosol formation, and cryoturbation, all promoted by enduring and harsh continental climate conditions. The formation of large polygonal systems of ice wedges and thick continuous sequences of frozen deposits recorded the persistence of accumulation areas that were stable and poorly drained, and had low topographic gradients. In short, the proposed cryolithogenic model integrates several previous concepts and emphasizes the polygenetic character of Yedoma deposits. It also identifies several sediment sources, weathering processes, and pathways by which sediments in typical periglacial landscapes can build up such fine-grained ice-rich deposits under extreme cold-climate conditions.

In Siberia, Yedoma deposits predominantly date from MIS 3 and 2. The initiation of Yedoma formation, based on radiocarbon dates, varies between about >55 ka BP on the New Siberian Islands and 27 ka BP at the western Laptev Sea coast. The latest deposition is dated between 28 ka BP on the New Siberian Islands and 17–13 ka BP in the western Laptev Sea. Unconformities within Yedoma sequences are common, encompassing up to 20 ka, and likely record thermokarst and thermoerosion processes. In general, the basal layers of Yedoma contrast sharply with the underlying deposits – often fluvial sands with peat layers, or loess-like flood plain deposits – dated by $^{230}\text{Th}/^{234}\text{U}$ and luminescence to an age of 100–60 ka (Schirrmeister et al., 2011b). The upper boundary of Yedoma is characterized by a thaw unconformity and separates locally confined Holocene deposits on top of Yedoma hills.

Alaska and Yukon ‘Muck’: Eastern Beringia Equivalents of Yedoma

As in Siberia, early work in northwestern North America was controversial, and workers argued that the silts, or later ‘muck’ deposits, in Alaska and Yukon may have been produced by

lacustrine, estuarine, eolian, residual weathering or fluvial processes, or their interaction (e.g., [Taber, 1943](#)). Following work by [Péwé \(1955\)](#), who argued strongly for an eolian origin, all workers have accepted a predominantly eolian origin for these sediments. In fact, no other ideas have been seriously entertained since the late 1950s. This likely reflects differences in the overall setting of these deposits, and perhaps a more generous consideration of loess in the North American literature relative to the concept of Yedoma in Siberia.

Loess is the most widely distributed surficial material in the nonglaciated areas of Alaska and Yukon ([Péwé, 1975](#)). Many of the areas with Yedoma in North America are associated with uplands and valleys, and to a lesser degree with extensive accumulation plains similar to the North Siberian shelves, resulting in the common differentiation into 'upland silt' and valley-bottom 'muck' deposits. These differences have been important in the development of ideas concerning the genesis of these deposits. Upland silts mantle hilltop exposures and are commonly found on alluvial terraces or sheltered behind bedrock obstacles, and have been unambiguously interpreted as loess ([Muhs et al., 2008](#); [Péwé, 1975](#)). However, it should be noted that these upland silt deposits are not ice-rich and differ significantly from Yedoma deposits as described in valley-bottom settings (e.g., [Bray et al., 2006](#); [Froese et al., 2009](#); [Péwé, 1975](#)).

In Péwé's original work and that of earlier researchers, the valley-bottom 'muck' deposits were considered to be 'reworked loess' derived from mixing of upland silts with organic material, accumulated syngenetically with permafrost growth in valley-bottom settings, and interbedded with primary airfall-derived loess. At some sites, well-developed bedding and sandy lenses indicate fluvial influence during deposition, but these beds are relatively minor at most sites. Commonly, the loess-adjective 'loessal' is used to describe material dominated by coarse silt and fine sand grain sizes, much of which is likely primary airfall loess, though evidence of secondary, short-distance redeposition by snowmelt or sheet flows may also be present (e.g., [Froese et al., 2009](#); [Sanborn et al., 2006](#)). Loess from MIS 2 is relatively thin on upland sites ([Muhs et al., 2008](#)), whereas thick accumulations dating to that time are present in valley-bottom sites ([Guthrie, 1990](#); [Péwé et al., 1975](#)). This suggests that either the sediments accumulated thickly in valley bottoms or these valleys acted as sediment traps, while upland surfaces, with sparse vegetation cover and, thus, low friction, may have been bypassed by these sediments (cf. [Muhs et al., 2008](#)).

The degree of mixing of the valley-bottom 'muck' deposits has been an open question, and throughout the 1970s and 1980s likely discouraged much paleoenvironmental work on these sites. However, increasingly large numbers of radiocarbon ages, typically on plant macrofossils, show few inversions in sedimentary sequences and argue for the stratigraphic coherence of these archives and their suitability for paleoenvironmental reconstruction ([Froese et al., 2009](#); [Zazula et al., 2007](#)).

A detailed study of paleosols within Yedoma deposits in the Klondike area, correlated between sites by distinctive distal tephras dating to MIS 2 and 4, confirms a consistent record of sedimentation, with syngenetic permafrost aggradation interrupted by periods of soil development between sites in valleys 20 km apart ([Sanborn et al., 2006](#)). These authors

showed that the valley-bottom silts underwent only limited weathering. Micromorphological evidence from the soils indicated that the organic matter within the sediments, primarily fine graminoid detritus, accreted mainly by below-ground inputs of root detritus as the surface sediments aggraded with loessal inputs ([Sanborn et al., 2006](#)). Importantly, the correlation of these consistent soil properties and timing of loess and syngenetic permafrost aggradation argue for regional controls on sedimentation rather than local site conditions.

In eastern Beringia, Yedoma formation largely dates to the late Pleistocene. [Péwé \(1975\)](#) argued for a consistent climatically driven period of ice-rich loessal sedimentation, which he named the Goldstream Formation. This formation postdated a significant thaw of permafrost during the last interglacial (MIS 5e) and likely dates through the Wisconsin. Subsequent workers have dropped the formal definition, but based on radiocarbon ages and tephrochronology, most of the Yedoma deposits date to late MIS 3 and MIS 2. Radiocarbon ages are fairly consistent between interior Yukon and Alaska for a major period of Yedoma development during MIS 2, beginning around 30 ka BP and continuing until ca. 11.5 ka BP ([Fraser and Burn, 1997](#); [Hamilton et al., 1988](#); [Zazula et al., 2007](#)). A pronounced unconformity at ca. 10 ka BP marks the degradation of these surfaces and is characterized by organic-detritus-rich sedimentation above the thaw unconformity sediments ([Fraser and Burn, 1997](#)). Earlier Yedoma sedimentation during MIS 4 is well documented in central Yukon through the presence of the Sheep Creek-K tephra (ca. 80 ka; [Froese et al., 2009](#)). At each time slice where these deposits are recognized, and paleoenvironmental reconstructions are available, the reconstructions indicate the aggradation of the loessal silts in an open, grass- and forb-rich steppe-tundra community, suggesting that these deposits are a recurrent feature of cold stages during the Pleistocene of Beringia ([Froese et al., 2009](#)).

In Alaska, more recent studies of Yedoma deposits have focused on the well-known CRREL permafrost tunnel near Fairbanks that exposed fine-grained frozen sequences containing syngenetic ice wedges ([Bray et al., 2006](#); [Hamilton et al., 1988](#); [Kanevskiy et al., 2008](#); [Shur et al., 2004](#)). A Yedoma site with very similar characteristics to Northeast Siberian sites is reported from the Itkillik River in the Arctic Foothills ([Kanevskiy et al., 2011](#)). These Yedoma deposits cover the period between >48 and 14.3 ka BP.

Conclusion

During the late Pleistocene cold stages in Beringia, Yedoma deposition was favored by a cold, dry climate that precluded widespread glaciation of lowlands and low mountain ranges, and resulted in intense periglacial weathering, sediment transport, accumulation, and syngenetic permafrost growth for several tens of thousands of years. The overall characteristics of Yedoma in Siberia (West Beringia) and North America (East Beringia) are given in [Table 1](#). The formation of large polygonal ice-wedge systems and sequences of ice-rich silty deposits tens of meters thick was closely related to the persistence of stable, poorly drained, and flat to gently sloping accumulation areas in the lowlands of Siberia and northern Alaska. Within the interior of Yukon and Alaska, Yedoma deposits are more

Table 1 Typical features of the Yedoma deposits

	<i>West Beringia (Siberia)</i>	<i>East Beringia (Alaska and Yukon)</i>
Formation age	80–12 ka BP	
Stratigraphy	MIS 4–MIS 2	
Terminology	Ice Complex, Yedoma Suite	'Muck' deposits or syngenetic late Pleistocene perennially frozen ice-rich silts
Landscape	Stable, poorly drained areas with low topographic gradient	
Distribution	East Siberian Arctic lowlands and large river valleys, foothills	Foothills and lowlands in northern Alaska; north easterly exposures and within narrow valleys in interior Yukon and Alaska
Sediment	Poorly sorted, organic-rich silt and silty sand	Organic-rich silt and silty sand; largely related to eolian sedimentation
Ground ice	Ice saturated or supersaturated, syngenetic ice wedges, segregation ice	
Paleoclimate	Highly continental, cold arid	
Paleoecology	Tundra steppe/Mammoth steppe, cryoxeric vegetation, Mammoth Fauna	
Genesis	Syngenetic permafrost of multiple origins depending on region: nival, alluvial, proluvial–colluvial, sheet flow, soil creep, or eolian sedimentation	Largely eolian sedimentation (airfall loess) with syngenetic permafrost, but locally, other deposits may be interbedded

variable, reflecting the variation of local site conditions with topography. Paleontological reconstructions of the environment during Yedoma formation indicate the presence of cryoxeric steppe-tundra communities. Yedoma deposits are defined by distinct cryolithological and sedimentological characteristics, and are preserved today in elevated erosional remnants in thermokarst landscapes termed Yedoma hills in Siberia, while in the previously unglaciated region of Alaska and the Yukon Territory, these deposits are found mostly in foothill regions, on north-facing slopes, and as fill in narrow valleys. In summary, Yedoma must be considered as a special type of periglacial or cryogenic facies typical of cold stages of late Pleistocene Beringia.

See also: [Paleoclimate Reconstruction: Pliocene Environments. Plant Macrofossil Records: Arctic Eurasia.](#)

References

- Andreev AA, Schirrmeister L, Tarasov PE, et al. (2011) Vegetation and climate history in the Laptev Sea region (Arctic Siberia) during Late Quaternary inferred from pollen records. *Quaternary Science Reviews* 30(17–18): 2182–2199.
- Bray MT, French HM, and Shur Y (2006) Further cryostratigraphic observations in the CRREL Permafrost Tunnel, Fox, Alaska. *Permafrost and Periglacial Processes* 17(3): 233–243. <http://dx.doi.org/10.1002/ppp.558>.
- Brown J, Ferrians OJ Jr, Heginbottom JA, and Melnikov ES (eds.) (1997) *Circum-Arctic Map of Permafrost and Ground-Ice Conditions*. Washington, DC: US Geological Survey in Cooperation with the Circum-Pacific Council for Energy and Mineral Resources. Circum-Pacific Map Series CP-45, scale 1:10,000,000, 1 sheet.
- Ehlers J and Gibbard PL (2003) Extent and chronology of glaciation. *Quaternary Science Reviews* 22: 1561–1568. [http://dx.doi.org/10.1016/S0277-3791\(03\)00130-6](http://dx.doi.org/10.1016/S0277-3791(03)00130-6).
- Fraser TA and Burn CR (1997) On the nature and origin of 'muck' deposits in the Klondike area, Yukon Territory. *Canadian Journal of Earth Sciences* 34: 1333–1344. <http://dx.doi.org/10.1139/e17-106>.
- Froese DG, Westgate JA, Reyes AV, Enkin RJ, and Preece SJ (2008) Ancient permafrost and a future, warmer arctic. *Science* 321: 1648. <http://dx.doi.org/10.1126/science.1157525>.
- Froese DG, Westgate JA, Sanborn PT, Reyes AV, and Pearce NJG (2009) The Klondike goldfields and Pleistocene environments of Beringia. *GSA Today* 19(8): 4–10. <http://dx.doi.org/10.1130/GSATG54A.1>.
- Gilerman RE, Sher AV, and Matthews JV (1982) Comparison of the development of tundra-steppe environments in West and East Beringia: Pollen and macrofossil evidence from key sections. In: Hopkins DM, Matthews JV, Schweger CE, and Young SB (eds.) *Paleoecology of Beringia*, pp. 43–73. New York: Academic Press.
- Grosse G, Schirrmeister L, and Malthus TJ (2006) Application of Landsat-7 satellite data and a DEM for the quantification of thermokarst-affected terrain types in the periglacial Lena-Anabar coastal lowland. *Polar Research* 25(1): 51–67. <http://dx.doi.org/10.1111/j.1751-8369.2006.tb00150.x>.
- Guthrie RD (1990) *Frozen Fauna of the Mammoth Steppe: The Story of Blue Babe*, p. 323. Chicago: University of Chicago Press.
- Hamilton TD, Craig JL, and Sellmann PV (1988) The Fox permafrost tunnel: A late Quaternary geologic record in central Alaska. *Geological Society of America Bulletin* 100: 948–969.
- Hulten E (1937) *Outline of the History of Arctic and Boreal Biota During the Quaternary Period*. Stockholm: Bokfrlags Aktiebolaget Thule.
- Kanevskiy M, Fortier D, Shur Y, Bray M, and Jorgenson T (2008) Detailed cryostratigraphic studies of syngenetic permafrost in the winze of the CRREL permafrost tunnel, Fox, Alaska. In: Kane DL and Hinkel KM (eds.) *Proceedings of the Ninth International Conference on Permafrost*, Fairbanks, Alaska, 29 June to 3 July 2008, vol. 1, pp. 889–894. Institute of Northern Engineering, University of Alaska Fairbanks.
- Kanevskiy M, Shur Y, Fortier D, Jorgenson MT, and Stephani E (2011) Cryostratigraphy of late Pleistocene syngenetic permafrost (yedoma) in northern Alaska, Itkillik River exposure. *Quaternary Research* 75(3): 584–596. <http://dx.doi.org/10.1016/j.yqres.2010.12.003>.
- Kaplina TN and Lozhkin AV (1984) Age and history of accumulation of the 'Ice Complex' of the maritime lowlands of Yakutiya. In: Velichko AA, Wright HE Jr., and Barnosky CW (eds.) *Late Quaternary Environments of the Soviet Union*, pp. 147–151. Minneapolis, MN: University of Minnesota Press.
- Konishchev VN and Rogov VV (1993) Investigations of cryogenic weathering in Europe and Northern Asia. *Permafrost and Periglacial Processes* 4: 49–64. <http://dx.doi.org/10.1002/ppp.3430040105>.
- Kunitsky VV, Schirrmeister L, Grosse G, and Kienast F (2002) Snow patches in nival landscapes and their role for the Ice Complex formation in the Laptev Sea coastal lowlands. *Polarforschung* 70: 53–67.
- Meyer H, Dereviagin A, Siegert C, Schirrmeister L, and Hubberten H-W (2002) Palaeoclimate reconstruction on Big Lyakhovsky Island, North Siberia – Hydrogen and oxygen isotopes in ice wedges. *Permafrost and Periglacial Processes* 13: 91–105. <http://dx.doi.org/10.1002/ppp.416>.
- Muhs DR, Ager TA, Skipp G, Beann J, Budahn J, and McGeethin JP (2008) Paleoclimatic significance of chemical weathering in loess-derived paleosols of subarctic central Alaska. *Arctic, Antarctic, and Alpine Research* 40: 396–411. [http://dx.doi.org/10.1657/1523-0430\(07-022\)\[MUHS\]2.0.CO;2](http://dx.doi.org/10.1657/1523-0430(07-022)[MUHS]2.0.CO;2).
- Péwé TL (1955) Origin of the upland silt near Fairbanks, Alaska. *Geological Society of America Bulletin* 66: 699–724.
- Péwé TL (1975) Quaternary geology of Alaska. U.S. Geological Survey Professional Paper 835, p. 143.
- Reyes AV, Froese DG, and Jensen BJ (2010) Permafrost response to last interglacial warming: Field evidence from unglaciated Yukon and Alaska. *Quaternary Science Reviews* 29: 3256–3274. <http://dx.doi.org/10.1016/j.quascirev.2010.07.013>.
- Romanovskii NN (1977) *Formation of Polygonal-Wedge Systems*, p. 214. Novosibirsk: Nauka. (in Russian).
- Romanovskii NN (1993) *Fundamentals of the Cryogenesis of the Lithosphere*, p. 336. Moscow: Moscow State University Press. (in Russian).
- Romanovskii NN, Hubberten H-W, Gavrilov AV, et al. (2000) Thermokarst and land – Ocean interactions, Laptev Sea Region, Russia. *Permafrost and Periglacial Processes* 11: 137–152.

- Sanborn PT, Smith CAS, Froese DG, Zazula GD, and Westgate JA (2006) Full-glacial paleosols in perennially frozen loess sequences Klondike goldfields, Yukon Territory, Canada. *Quaternary Research* 66: 147–157. <http://dx.doi.org/10.1016/j.yqres.2006.02.008>.
- Schirmermeister L, Grosse G, Wetterich S, et al. (2011a) Fossil organic matter characteristics in permafrost deposits of the Northeast Siberian Arctic. *Journal of Geophysical Research* 116(G00M02): 16 pp.
- Schirmermeister L, Kunitzky VV, Grosse G, et al. (2011b) Sedimentary characteristics and origin of the Late Pleistocene Ice Complex on North-East Siberian Arctic coastal lowlands and islands – A review. *Quaternary International*. 241(1–2): 3–25.
- Shapiro B and Cooper A (2003) Beringia as an Ice Age genetic museum. *Quaternary Research* 60: 94–100. [http://dx.doi.org/10.1016/S0033-5894\(03\)00009-7](http://dx.doi.org/10.1016/S0033-5894(03)00009-7).
- Sher AV (1997) Yedoma as a store of paleoenvironmental records in Beringia. In: Elias S and Brigham-Grette J (eds.) *Beringia Paleoenvironmental Workshop*, September 1997, pp. 92–94. Abstracts and Program.
- Sher AV, Kaplina TN, Giterman RE, et al. (1979) Late Cenozoic of the Kolyma Lowland. Paper presented at 14th Pacific Science Congress, Academy of Science, USSR, Khabarovsk.
- Sher AV, Kuzmina SA, Kuznetsova TV, and Sulerzhitsky LD (2005) New insights into the Weichselian environment and climate of the East Siberian Arctic, derived from fossil insects, plants, and mammals. *Quaternary Science Reviews* 24: 533–569. <http://dx.doi.org/10.1016/j.quascirev.2004.09.007>.
- Sher AV, Kaplina TN, and Ovander MG (1987) Unified regional stratigraphic chart for the Quaternary deposits in the Yana-Kolyma Lowland and its mountainous surroundings. Explanatory Note. In: *Decisions of Interdepartmental Stratigraphic Conference on the Quaternary of the Eastern USSR*, Magadan, 1982. USSR Academy of Sciences, Far-Eastern Branch, North-Eastern Complex Research Institute, Magadan, USSR, pp. 29–69 (in Russian).
- Shur Y, French HM, Bray MT, and Anderson DA (2004) Syngenetic permafrost growth: Cryostratigraphic observations from the CRREL Tunnel near Fairbanks, Alaska. *Permafrost and Periglacial Processes* 15: 339–347.
- Solov'ev PA (1959) *The Cryolithozone in the Northern Part of the Lena-Amga-Interfluve*, p. 144. Moscow, Russia: Academy of Science of the USSR. (in Russian).
- Taber S (1943) Perennially frozen ground in Alaska: Its origin and history. *Bulletin of the Geological Society of America* 54: 1433–1548.
- Tomirdiario SV (1996) Palaeogeography of Beringia and Arctida. In: West CF (ed.) *American Beginnings: The Prehistory and Palaeoecology of Beringia*, pp. 58–69. Chicago and London: University of Chicago Press.
- van Everdingen RO (ed.) (1998) *Multi-Language Glossary of Permafrost and Related Ground-Ice Terms*. Calgary: University of Calgary.
- Vas'kovsky AP (1963) Stratigraphic outline of Quaternary deposits in northeastern Asia. In: Egiazarova VKh (ed.) *Geology of the Koryaksky Mountains*, pp. 24–53, 143–168. Moscow: Gostoptekhizdat, (in Russian).
- Vasil'chuk YuK (1992) Oxygen Isotope Composition of Ground Ice. Application to Paleogeocryological Reconstructions. Moscow, Russia: Geological Faculty of Moscow State University, Russian Academy of Sciences vol. 1, p. 420 (in Russian).
- Vasil'chuk YK, Vasil'chuk AC, Rank D, Kutschera W, and Kim JC (2001) Radiocarbon dating of delta O-18-delta D plots in Late Pleistocene ice-wedges of the Duvanny Yar (Lower Kolyma River, Northern Yakutia). *Radiocarbon* 43(2B): 541–553.
- Wetterich S, Kuzmina S, Andreev AA, et al. (2008) Palaeoenvironmental dynamics inferred from late Quaternary permafrost deposits on Kurungnakh Island, Lena Delta, Northeast Siberia, Russia. *Quaternary Science Reviews* 27(15–16): 1523–1540. <http://dx.doi.org/10.1016/j.quascirev.2008.04.007>.
- Wetterich S, Rudaya N, Andreev A, Meyer H, Opel T, and Schirmermeister L (2011) *Last Glacial Maximum records in permafrost of the East Siberian Arctic*. *Quaternary Science Reviews* 30: 3139–3151.
- Wolfe SA, Gillis A, and Robertson L (2009) Late Quaternary eolian deposits of northern North America: Age and extent. Geological Survey of Canada, Open File 6006.
- Yershov ED (2004) *General Geocryology*, p. 608. Cambridge, UK: Cambridge University Press.
- Yurtsev BA (1982) Relics of the xerophyte vegetation of Beringia in Northeastern Asia. In: Hopkins DM, Matthews JV, Schweger CE, and Young SB (eds.) *Paleoecology of Beringia*, pp. 157–177. New York: Academic Press.
- Zanina OG, Gubin SV, Kuzmina SA, Maximovich SV, and Lopatin DA (2011) Late-Pleistocene (MIS 3–2) palaeoenvironments as recorded by sediments, palaeosols, and ground-squirrel nests at Duvanny Yar, Kolyma lowland, northeast Siberia. *Quaternary Science Reviews*. 30(17–18): 2107–2123.
- Zazula GD, Froese DG, Elias SA, Kuzmina S, and Mathewes RW (2007) Arctic ground squirrels of the mammoth-steppe: Paleoecology of Late Pleistocene middens (24 000–29 450 14C yr BP), Yukon Territory, Canada. *Quaternary Science Reviews* 26: 979–1003. <http://dx.doi.org/10.1016/j.quascirev.2006.12.006>.
- Zhestkova TN (1982) *Formation of the Cryogenic Structure of Ground*, p. 209. Moscow: Nauka. (in Russian).
- Zimov SA, Schuur EAG, and Chapin FS III (2006) Permafrost and the global carbon budget. *Science* 312: 1612–1613. <http://dx.doi.org/10.1126/science.1128908>.

Paraglacial Geomorphology

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Introduction and Definition

Paraglacial geomorphology is the study of the ways in which glaciated landscapes adjust to nonglacial conditions during and after deglaciation. The term paraglacial was introduced by Ryder (1971) to describe 'nonglacial processes that are directly conditioned by glaciation' (Church and Ryder, 1972), but it has subsequently been adopted to describe glacially conditioned landforms, landscapes, and sediment accumulations, as well as processes. This widened usage has prompted redefinition of the term as a descriptor of 'nonglacial earth-surface processes, sediment accumulations, landforms, land systems, and landscapes that are directly conditioned by glaciation and deglaciation' (Ballantyne, 2002a). Initial research on paraglacial geomorphology focused on fluvial responses to late Pleistocene or early Holocene deglaciation. The past two decades, however, have seen an upsurge in research on paraglacial landscape adjustment following recent glacier retreat, together with a wider focus that incorporates rock slopes, drift-mantled slopes and glacier forelands, alluvial and lacustrine systems, and glaciated coasts. Developments in paraglacial geomorphology before 2001 are reviewed in Ballantyne (2002a). A remarkably large body of pertinent literature has appeared since then, largely devoted to landscape changes following recent (post-Little Ice Age) deglaciation.

Theory

The Paraglacial Sediment Cascade

The retreat of glacier ice often reveals landscapes in an unstable or metastable state, and in particular, results in the exposure of glaciogenic or glacially conditioned sediment sources that are subsequently released and reworked by a variety of processes over a wide range of timescales. Rock masses stressed by the weight of glacier ice experience decompression, opening of joints, and adjustment in the form of rockfalls, rock topples, rock slides, rock avalanches and gravitational slope deformation. Glaciogenic sediments on valley-side slopes may be reworked by translational landslips, debris flows, snow avalanches, and running water. Valley-floor glaciogenic sediments (till, glaciofluvial deposits, and glacial-lacustrine sediments) are eroded, transported, and redeposited by rivers, forming alluvial fans and valley fills, or deltas or bottom deposits in lakes and fjords. Glaciofluvial deposits exposed at the coast are eroded by waves and currents, and redeposited as barrier deposits and shelf sediments.

The paraglacial system may thus be envisaged as a sediment cascade in which nonrenewable sediment sources (unstable rockwalls, drift-mantled slopes, valley-floor glaciogenic deposits, and coastal glaciogenic deposits) are released or entrained by a range of processes and transported toward terminal sediment sinks (Figure 1). Source-to-sink transport is often interrupted,

however, by sediment redeposition in paraglacial sediment stores such as talus accumulations, debris cones, alluvial fans and valley fills, deltas, beaches, spits, and bars. Some paraglacial sediment stores continue to accumulate for several millennia, but others experience degradation within a few centuries or millennia of formation.

Models of Paraglacial Sediment Yield

Ballantyne (2002b) has argued that for primary paraglacial systems in which sediment sources are not replenished, the trajectory of sediment reworking may be approximated by an exhaustion model in which the rate of sediment yield is related to the remaining available sediment by a negative exponential function:

$$S_t = S_0 e^{-\lambda t} \quad [1]$$

where t is time elapsed since deglaciation, S_t the proportion of sediment available for reworking at time t , S_0 the sediment available for reworking at $t=0$, and λ is the rate of sediment loss by release and/or stabilization (Figure 2(a)). The value of λ varies over several orders of magnitude for particular paraglacial systems. On glacier forelands and drift-mantled slopes, for example, most landform adjustment and associated sediment reworking is completed within a few decades or centuries, whereas rock-mass adjustment to glacial unloading may continue to affect slope stability for many millennia after deglaciation (Figure 2(b)).

Secondary paraglacial systems (primarily fluvial systems), in which sediment inputs include both glaciogenic sediment sources and remobilization of upstream paraglacial sediment stores, behave in a more complex way. Church and Slaymaker (1989) interpreted a downstream increase in the specific non-dissolved sediment yield of rivers draining deglaciated terrain as indicating a delayed peak in paraglacial sediment yield in large catchments, an argument developed further by Church (2002), who envisaged a 'wave' of reworked sediment progressing down the fluvial system over millennial timescales. Alternatively, a downstream increase in specific sediment yield may be consistent with the simple exhaustion model, providing (1) that initial sediment availability (S_0) is greater in small tributary basins and (2) that the rate of decline in sediment supply (λ) is more gradual with increasing catchment size (Figure 2c). The latter assumption is likely to be met in most cases because of the contribution of reworked sediment from upstream sources, but the validity of the former assumption probably varies from site to site.

The models illustrated in Figure 2(a)–2(c) assume steady-state conditions in which there is no change in process-generating mechanisms or other boundary conditions. Over millennial timescales, this is rarely, if ever, the case, and paraglacial sediment flux may be influenced by such factors as climate change (advance or retreat of glaciers, changes in

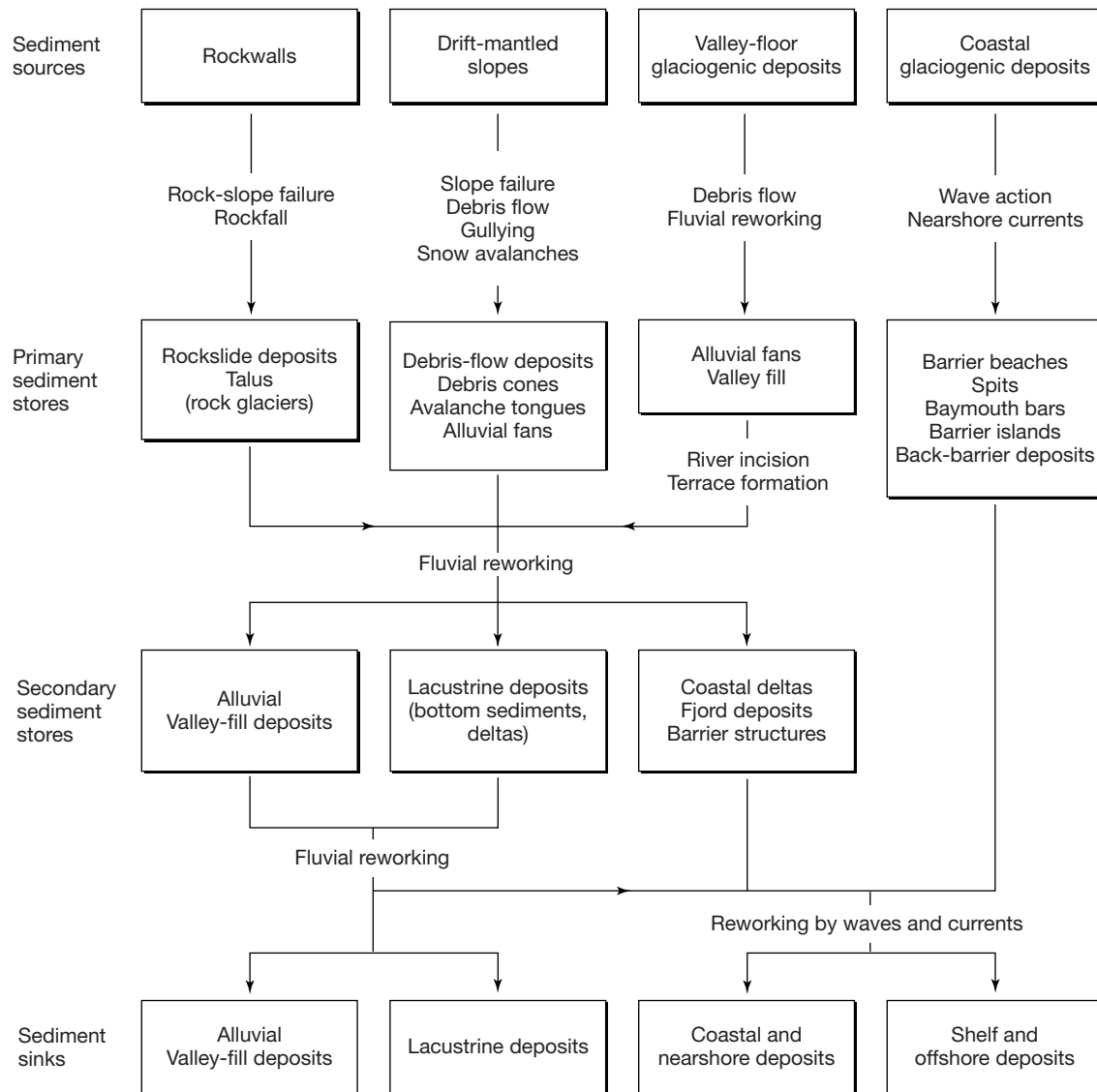


Figure 1 Paraglacial sediment cascade. Sediment may enter storage several times during source-to-sink transport. Reprinted from Ballantyne CK (2002) Paraglacial geomorphology. *Quaternary Science Reviews* 21: 1935–2017, © 2002, with permission from Elsevier.

catchment hydrology or in the magnitude and frequency of extreme rainstorm events), tectonic activity (uplift, tilting, or seismic events), sea-level change, or anthropogenic influences such as vegetation burning, deforestation, or cultivation. Such extrinsic perturbations tend to prolong, rejuvenate, or reinitiate paraglacial sediment reworking (Figure 2(d)). Lowering of base level by glacioisostatic uplift or deglaciation, for example, may cause renewed incision and reworking of glaciogenic deposits, paraglacial fans, and valley fills (e.g., Meigs et al., 2006), and glacioeustatic sea-level rise permits rising seas to tap pristine sources of littoral glaciogenic sediment, rejuvenating nearshore paraglacial sediment flux (Forbes and Syvitski, 1994). Extreme climatic events such as exceptional rainstorms may cause reworking of valley-side drift deposits by debris flows and torrents several millennia after the initial phase of paraglacial sediment reworking has ended.

Modeling Paraglacial Sediment Storage

As paraglacial sediment stores such as talus cones, debris cones, alluvial fans, and alluvial valley fills accumulate, they are subject to erosional losses. If it is assumed that rates of non-dissolved sediment loss from paraglacial sediment stores are related to sediment availability (i.e., the volume of erodible sediment in the sediment store), then the volume of sediment in storage (S) at time t is approximated by

$$S = (S_0 - S_0 e^{-\lambda t}) e^{-\kappa t} \quad [2]$$

where κ is the rate of sediment loss from the store. If S_0 is set at unity, this simplifies to

$$S = (1 - e^{-\lambda t}) e^{-\kappa t} \quad [3]$$

Figure 3 illustrates the change in sediment storage for $\lambda = 1.0 \text{ ka}^{-1}$ and a range of values of κ . One implication of

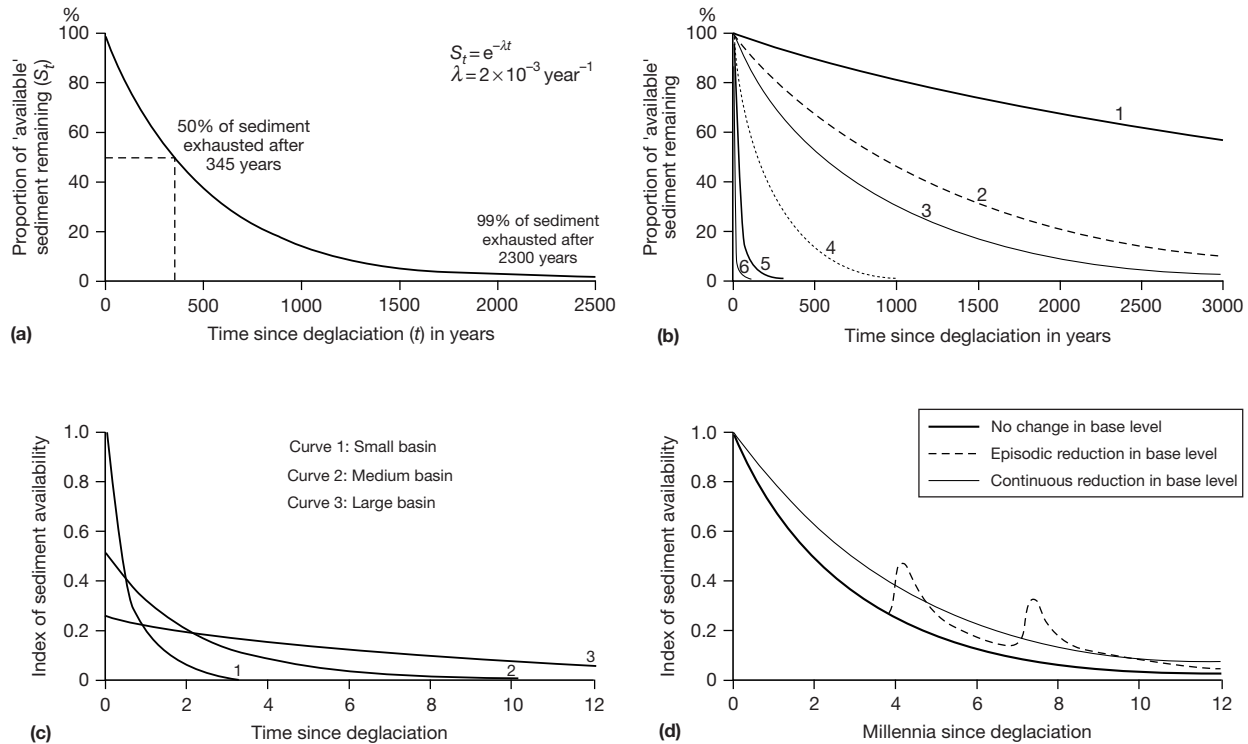


Figure 2 Models of paraglacial sediment flux. (a) Simple exhaustion model of paraglacial sediment release, in which rate of release (λ) is related to the proportion of 'available' sediment remaining. (b) Illustrative exhaustion curves for different primary paraglacial processes: (1) rock-slope failure, (2) rockfall and talus accumulation, (3) accumulation of large alluvial fans, (4) deep-seated gravitational slope deformation, (5) modification of drift-mantled slopes, and (6) modification of glacier forelands. (c) Exhaustion model of paraglacial sediment release for drainage basins of different size. (d) Influence of external perturbation on the temporal pattern of paraglacial sediment release: effect of base-level change of sediment release in fluvial systems. Reproduced from Ballantyne CK (2003) Paraglacial landform succession and sediment storage in deglaciated mountain valleys: Theory and approaches to calibration. *Zeitschrift für Geomorphologie Supplementband* 132: 1–18.

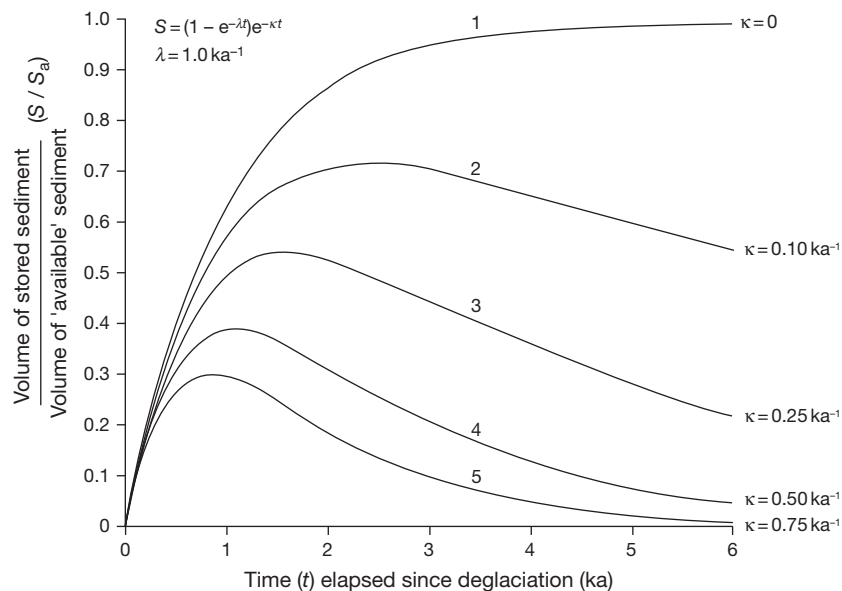


Figure 3 General model of paraglacial sediment storage. Sediment stores such as debris cones, alluvial fans, or valley fills experience an initial period of net accumulation, succeeded by a longer period of net sediment loss. This occurs because the sediment sources feeding such sediment stores are nonrenewable. Reproduced from Ballantyne CK (2003) Paraglacial landform succession and sediment storage in deglaciated mountain valleys: Theory and approaches to calibration. *Zeitschrift für Geomorphologie Supplementband* 132: 1–18.

this model is that for all paraglacial sediment stores that experience output losses ($\kappa > 0$), an eventual change from net sediment accumulation to net sediment loss is an intrinsic property of paraglacial sediment storage; this reflects the nonrenewable nature of paraglacial sediment sources. Another implication is that both the maximum amount of sediment in storage ($\max(S)$) and the timing of maximum sediment storage (t') are related to the relative values of λ and κ , and can be expressed as

$$\max(S) = \left[\frac{\lambda}{\lambda + \kappa} \right] \left[\frac{\kappa}{\lambda + \kappa} \right]^{\kappa/\lambda} \quad [4]$$

and

$$t' = -\ln \frac{[\kappa/(\lambda + \kappa)]}{\kappa} \quad [5]$$

These relationships potentially allow the history of net sediment accumulation and loss to be retrodicted for individual paraglacial sediment stores, provided that time since deglaciation and time of maximum sediment storage are known, and both the maximum and present volumes of stored sediment can be evaluated (Ballantyne, 2003). On short (decadal and centennial) timescales, however, the exhaustion model outlined earlier may be delayed by sediment accumulating behind moraine barriers (Cossart and Fort, 2008), and over long timescales, fluvial response is likely to be complicated by the fact that sediments of different sizes move at different rates through the fluvial system (Church, 2002).

Paraglacial Succession

Because paraglacial subsystems relax over widely differing timescales (Figure 2(b)), some paraglacial sediment stores experience net erosion and degradation even as others continue to accumulate. In consequence, paraglacial landform assemblages are increasingly dominated by sediment stores with the greatest formative longevity and persistence in the landscape (Figure 4). This evolutionary process is termed paraglacial succession (Ballantyne, 2003). Owen et al. (1995) have illustrated this concept by mapping the paraglacial landforms occupying valleys deglaciated at different times in the Lahul Himalaya, showing that these exhibit a marked increase through time in areas occupied by alluvial fills, alluvial fans, and slopes modified by mass movement.

Paraglacial Land Systems

Ballantyne (2002a) has identified six paraglacial land systems characterizing particular geomorphological contexts: rock slopes, drift-mantled slopes, glacier forelands, alluvial settings, lacustrine deposits, and coastal areas. It is important to appreciate, however, that these frequently overlap spatially. Deglaciated mountain valleys, for example, may incorporate rock-slope failures, reworked drift slopes, and paraglacially modified glacier forelands and alluvial fans, and lacustrine and coastal responses are strongly linked to the passage of reworked glaciogenic sediment through the fluvial system.

Rock Slopes

Glaciation and deglaciation affect rock-slope stability in two ways. First, glacial erosion tends to steepen and lengthen rock slopes, increasing and reorientating the self-weight stresses acting behind the rock face. Second, during glacial episodes, the weight of overlying or adjacent glacier ice increases stresses within rock masses. As some of the resulting rock deformation is elastic and stored within the rock as strain energy, deglaciation results in gradual rock-mass dilation, often termed stress release. This process causes propagation of the internal joint network, together with reduced cohesion along joints and reduction of internal locking stresses. These changes may initiate three types of responses on steep rockwalls, namely, catastrophic rock-slope failure, deep-seated gravitational rock-mass deformation, and slow recovery of equilibrium through episodic rockfalls. On lower slopes and valley floors, stress release is manifest in the development of decompression structures (stress-release joints) on erosional bedforms such as roches moutonnées (Cossart et al., 2008). The nature of rockwall response is determined mainly by joint density, and orientation and inclination of the principal joint sets relative to rock faces.

Catastrophic rock-slope failure

There are numerous documented examples of catastrophic rock-slope failures on recently deglaciated terrain (Ballantyne, 2002a). These have generally been attributed to debuttressing of glacially steepened rockwalls during deglaciation and associated stress release, though in some cases, permafrost degradation and thaw of ice in joints may also have contributed to the failure. There is also evidence for widespread paraglacial rock-slope failure following late Pleistocene deglaciation (e.g., Ballantyne et al., 2009). Cossart et al. (2008) have demonstrated that major rockfalls and rock avalanches in the Upper Durance catchment (French Alps) are concentrated in lower valley-side slopes within the limits of glacial occupancy at the Last Glacial Maximum, and that their locations coincide with areas of inferred high glacial loading stress, consistent with the interpretation of these rock-slope failures as a response to deglacial stress release.

Cruden and Hu (1993) proposed that the pattern of paraglacial rock-slope failure might be approximated by an exhaustion model (Figure 2(a)). Their model assumes that there are a finite number of potential failure sites following deglaciation, that each site fails only once, and that the probability of individual failures remains constant. The overall probability of failure within a given area therefore declines exponentially with time as the number of potential failure sites diminishes. Although this model has not been rigorously tested, cosmogenic isotope exposure dating of rock-slope failures in formerly glaciated mountain environments is consistent with a general pattern of frequent failures during or soon after deglaciation followed by less frequent failures thereafter, the latter probably conditioned by rock-mass weakening due to stress release but triggered by earthquakes or extreme rainstorms (Ballantyne et al., 2009; Cossart et al., 2008; Figure 5).

Deep-seated gravitational slope deformations

Deep-seated gravitational slope deformations (rock-mass deformations) are widespread on steep slopes in formerly

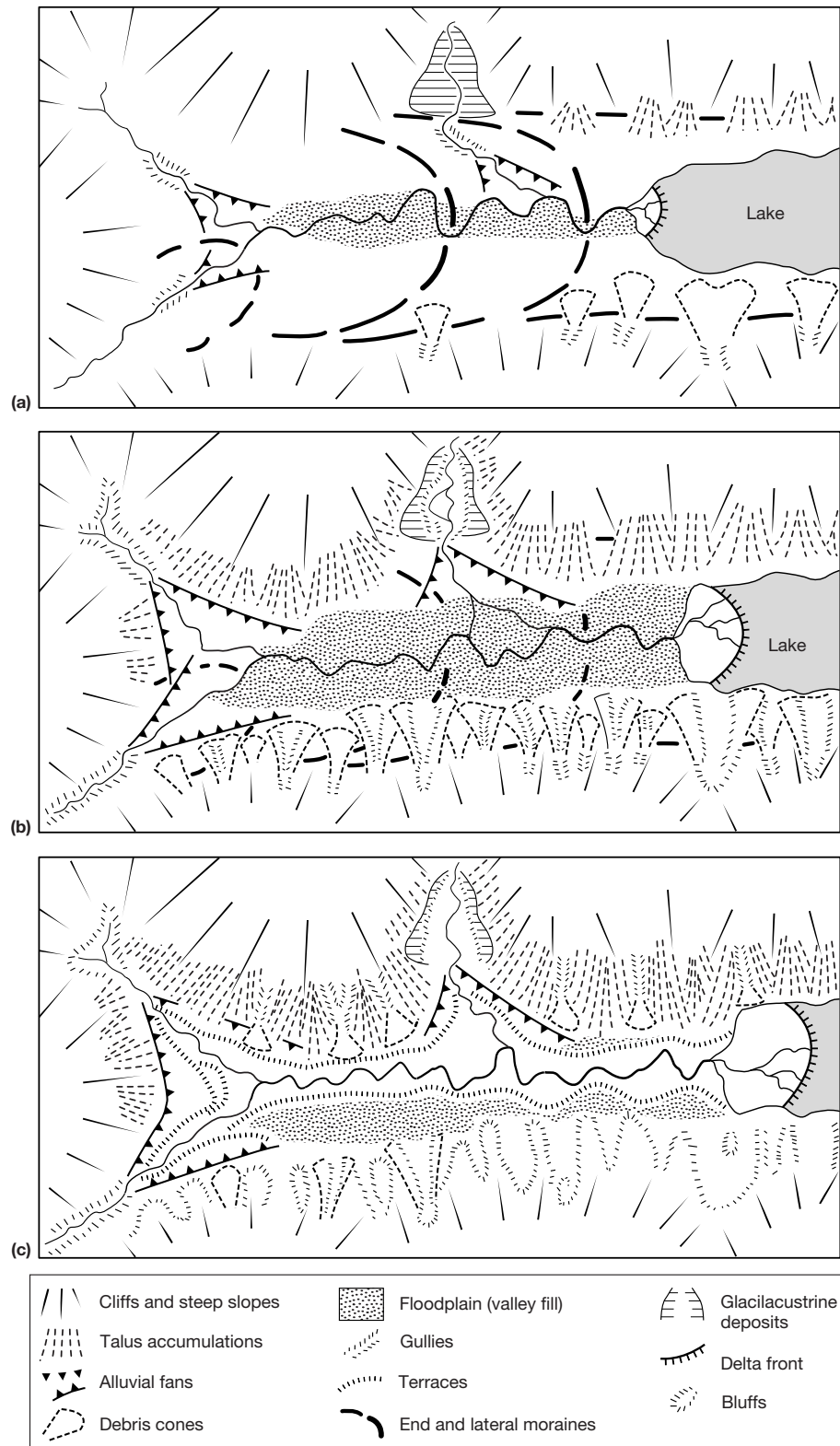


Figure 4 Paraglacial succession (schematic). (a) After deglaciation, glacial landforms and sediments are largely intact. (b) After a few millennia, alluvial fans and valley fills approach their maximum dimensions, though paraglacial debris cones are already degraded; talus continues to accumulate and the delta has prograded into the lake. (c) After ~10–15 ka, little evidence of glacial deposition remains, paraglacial debris cones are reduced to eroded bluffs, paraglacial fans and valley fills have been entrenched, erosion of talus has commenced, and only the delta has a net positive sediment budget and continues to prograde. Reproduced from Ballantyne CK (2003) Paraglacial landform succession and sediment storage in deglaciated mountain valleys: Theory and approaches to calibration. *Zeitschrift für Geomorphologie Supplementband* 132: 1–18.

glaciated mountains and have been widely attributed to glacial steepening, debuttreasing, decompression, and paraglacial stress release. Such deformation often involves downslope displacement of entire mountainside slopes, sometimes covering areas over 1 km² in area, but usually without debris runoff. Landforms associated with rock-slope deformation include ridge-top graben (trenches), tension cracks, minor rock top-ple, anticarps, and bulging slopes (Figure 6). Recent downwastage of the Affliction Glacier in British Columbia, for example, has resulted in slow gravitational displacement of roughly 30 million cubic meters of rock with surface strain rates of a few millimeters to a few centimeters per year, and associated development of tension cracks, anticarps, graben, and collapse pits (Bovis, 1990; Figure 7). The widespread nature of similar features on mountains that now lack glacier

ice suggests that paraglacial rock-slope deformation commonly accompanied late Pleistocene deglaciation, but dating control is weak and it is not established whether deformation typically occurred as a single rapid event, a series of intermittent shifts, or gradually through rock-mass creep. The nature of failure is also unclear in most cases: this may have occurred along a continuous shear surface, or over two or more shear planes separated by zones of fractured rock, or over a deep-seated zone of crushed rock.

Paraglacial rockfall and talus accumulation

Another response of glacially steepened rockwalls to deglaciation is pronounced rockfall activity and consequent accumulation of talus. Several authors have noted that recent rockfall rates are inadequate to account for the large volumes of talus



Figure 5 The Beinn Alligin rock avalanche, NW Scotland, a paraglacial rock-slope failure dated using cosmogenic ¹⁰Be to ~4.0 ka BP, ~7.5 ka after final deglaciation of this location.



Figure 6 Bulging slopes and anticarps on Sgurr a'Bhealaich Dheirg, NW Scotland, produced by deep-seated gravitational slope deformation on the flank of a glacial trough. The highest anticarp is about 8 m high. Photograph by D. Jarman.

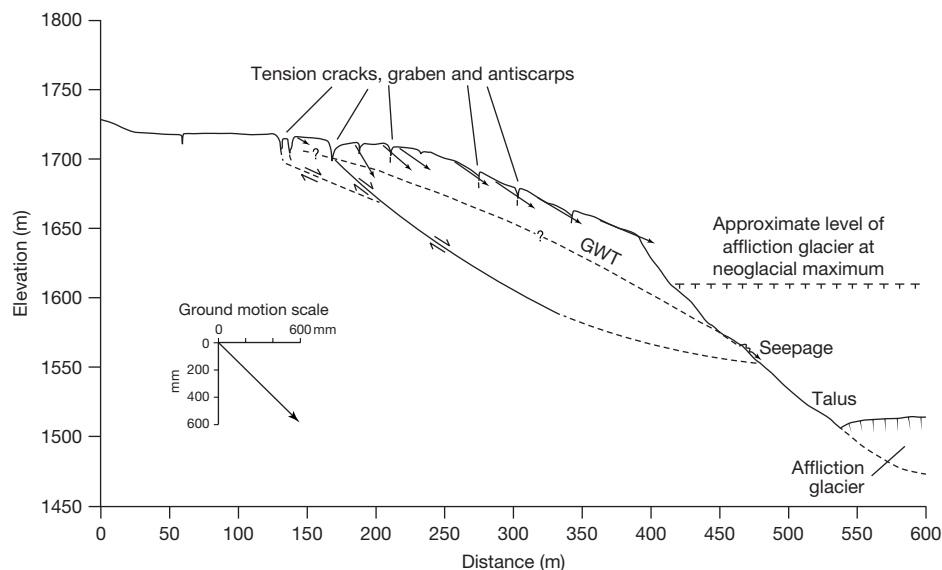


Figure 7 Cross-section of rock-slope displacement caused by downwastage of Affliction Glacier, British Columbia. The ground motion vectors represent displacement over 4 years. GWT: inferred water table. Reprinted from Bovis MJ (1990) Rock-slope deformation at Affliction Creek, southern Coast Mountains, British Columbia. *Canadian Journal of Earth Sciences* 27: 243–254. © 1990, with permission from NRC Research Press.

below rockwalls deglaciated at the end of the Pleistocene, and have suggested that rockfall was enhanced by stress release immediately after deglaciation (Ballantyne, 2002a). Curry and Morris (2004), for example, calculated that 84% of rock-wall retreat due to rockfall at a site in south Wales occurred between 17.3 cal ka BP and 11.5 ka BP following late Pleistocene deglaciation, and estimated that roughly 50% of rockfall occurred within 1000 years after deglaciation. Rapid talus accumulation following recent glacier retreat has also been observed in a range of mountain environments. Frequent rock-falls following late Pleistocene deglaciation were probably instrumental in the development of now-relict talus rock glaciers in former permafrost areas.

Drift-Mantled Slopes

Glacier retreat often leaves lower valley-side slopes mantled in thick glaciogenic deposits, sometimes composed of stacked lateral moraines. Such deposits are vulnerable to erosion by shallow translational failure, debris flow, snow avalanches, streamflow, and surface wash. In extreme cases, these processes completely rework drift-mantled slopes, forming a land system of deep gullies, intervening drift arêtes, coalescing debris cones, and slope-foot fans (Figure 8). In the Himalayas, this landform association achieves spectacular dimensions (Iturrizaga, 2008; Owen et al., 1995), and in high-latitude environments, gully-ing of valley-side glaciogenic sediments may be triggered by rapid melt of buried dead-ice cores (Mercier et al., 2009). In mountain environments where recent ice downwastage has exposed steep-sided lateral moraines, proximal moraine slopes may be extensively modified by slumping and debris falls, with rapid evacuation of debris by recurrent debris flows (Curry et al., 2006). Rates of gully erosion can be very rapid. On recently deglaciated terrain in the Jostedal area in Norway, Ballantyne and Benn (1994) calculated a minimum incision rate of 50–100 mm year⁻¹ for gullies excavated on valley-side drift deposits between AD 1943 and AD 1991, and Curry et al. (2006) inferred average minimum incision rates of ~95 mm year⁻¹ for gullies cutting into a lateral moraine in the Swiss Alps. In both cases, gullies appear to have reached



Figure 8 Paraglacial modification of drift-mantled slopes, Fåbergstølsdalen, Norway. Less than 50 years after glacier retreat, erosion by debris flows has removed most of the original valley-side drift deposits (stacked lateral moraines), creating deep unstable gullies and depositing coalescing debris cones at the slope foot.

their maximum dimensions within about 50 years after deglaciation, after which gully relief has been reduced by reduction of intergully slopes and gully infilling. Although complete transformation of drift-mantled slopes may be achieved within a few decades or centuries (Ballantyne and Benn, 1994; Curry et al., 2006; Mercier et al., 2009), renewed sediment remobilization long after the initial period of paraglacial modification sometimes occurs, usually in response to extreme rainstorm events. Degraded vegetation-covered gully systems and associated inactive debris cones are common in steep-sided valleys that were deglaciated at the end of the Pleistocene, and the aforementioned evidence suggests that many of these formed and atrophied within a century or two of ice-margin retreat.

Glacier Forelands

Valley-floor deposits exposed by glacier retreat tend to experience rapid modification by mass movement, frost sorting, wind erosion, and surface runoff. The slopes of steep-sided moraines decline to stable gradients within a few decades through slumping, and solifluction sometimes modifies the original form and near-surface structure of sediment bedforms (Ballantyne, 2002a). Recently deglaciated forelands often exhibit periglacial microforms, including sorted nets and stripes, boulder-cored frost boils, surface cracks, miniature solifluction lobes, and ploughing boulders, all of which imply disruption by freeze–thaw of near-surface glaciogenic deposits (Matthews et al., 1998). Deflation of unvegetated glaciogenic sediments by down-glacier katabatic winds causes loss of surface fines, and development of a surficial stony lag deposit, with associated accumulation of aeolian sands and silty sands farther downwind. During the retreat of Pleistocene ice sheets, wind erosion operated on a vast scale across outwash plains, resulting in the deposition of thick proximal coversands and distal loess accumulations of essentially paraglacial origin. Finally, the sediment load of proglacial meltwater streams is often substantially enriched by the addition of remobilized foreland sediments, mainly through surface wash, bank collapse, and sediment contributions from debris flows and slopewash from adjacent slopes. Though such effects may be comparatively short-lived due to sediment exhaustion and surface armoring (Orwin and Smart, 2004), elevated rates of sediment input into distal proglacial lakes often accompanies glacier retreat, raising the question of whether mineral-rich layers in the bottom sediments of such lakes represent episodes of glacier advance (as is often assumed) or sediment-enhanced proglacial runoff during glacier retreat (Ballantyne, 2002a). The main effects of paraglacial adjustment of recently deglaciated terrain are thus sediment redistribution, relief modification, and changes in near-surface sediment characteristics. Most modification is complete within a few decades or centuries due to stabilization of foreland sediments by vegetation, though sharp-crested moraines may experience progressive degradation over longer timescales.

Alluvial Land Systems

Paraglacial fans

Among the most widespread paraglacial landforms are debris cones and alluvial fans composed mainly of glaciogenic sediment reworked by debris flows and tributary streams.

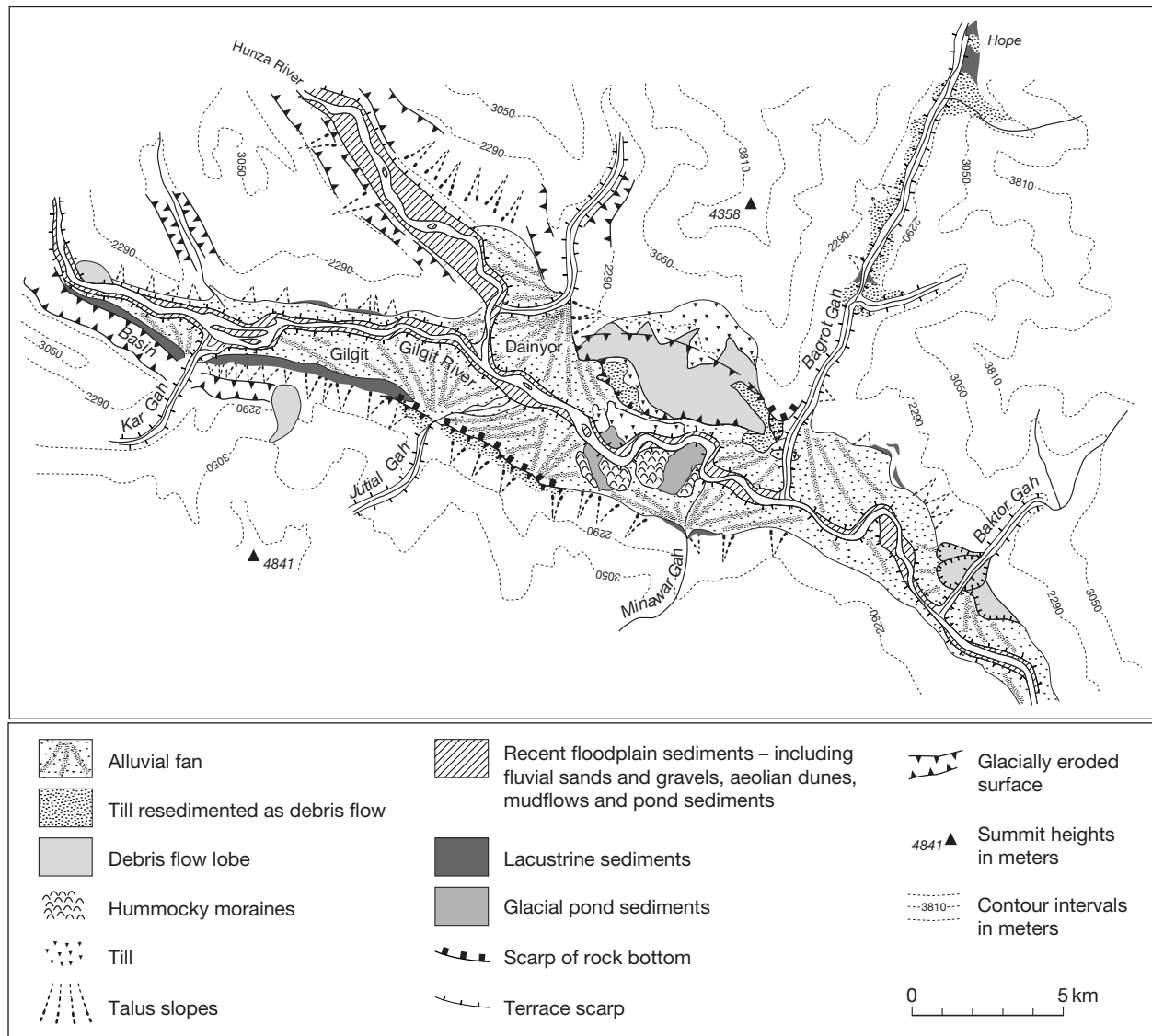


Figure 9 Landforms and surficial sediments of the Gilgit Valley, Karakoram Mountains. The surface of the valley fill is dominated by huge paraglacial fans. Adapted from Owen LA (1989) Terraces, uplift and climate in the Karakoram Mountains, northern Pakistan: Karakoram intermontane basin evolution. *Zeitschrift für Geomorphologie Supplementband* 76: 117–146.

On recently deglaciated terrain, small debris cones and alluvial fans have accumulated and then degraded (due to depleted sediment input) within a few centuries (Ballantyne, 2002a). Large relict fans that accumulated following late Pleistocene deglaciation, however, often took several millennia to achieve their maximum dimensions before experiencing net erosion. Ryder (1971), for example, demonstrated that relict alluvial fans in British Columbia began to accumulate soon after deglaciation and continued to aggrade until shortly after ~ 6.6 ^{14}C ka BP. These fans were subsequently eroded by fan-head trenching due to reduction in sediment input, or incised due to base-level lowering, resulting locally in the formation of nested multilevel fans. The lithofacies of paraglacial fans are often complex, incorporating alluvial gravels, diamictic sediments deposited by debris flows and hyperconcentrated flows, and sometimes sediments of aeolian

and lacustrine origin. Some have impressive dimensions: the lower Cheekye fan in British Columbia, for example, occupies 8.3 km^2 . Roughly 1.4 km^3 of sediment accumulated on this fan prior to ~ 6.0 ^{14}C ka BP, but subsequent sediment accumulation has been an order of magnitude less (Friele et al., 1999). The floors of major valleys in the Hindukush, Karakoram, and Himalayan mountains are often dominated by huge paraglacial fans, commonly several tens of meters thick (Figure 9), produced by a combination of debris flows and hyperconcentrated flows (Iturrizaga, 2008; Owen, 1989; Owen et al., 1995). Barnard et al. (2006) have demonstrated through the use of surface exposure dating that such fans and associated terraces in the Khumbu Himal are chronologically related to Late Quaternary glacier fluctuations, and proposed that they represent rapid large-scale reworking of glaciogenic sediments during periods of glacier retreat.

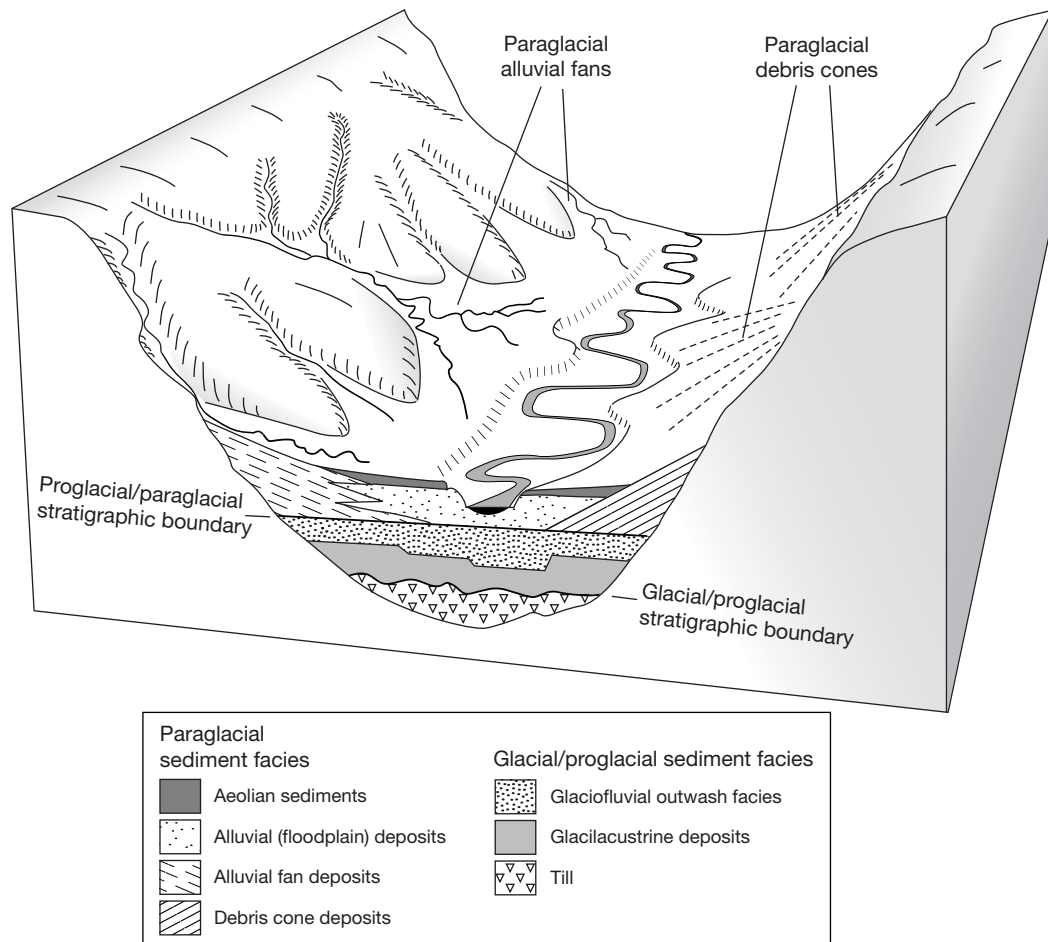


Figure 10 Glacial, proglacial, and paraglacial sediment sequence within a paraglacial valley fill (schematic). Reprinted from Ballantyne CK (2002) Paraglacial geomorphology. *Quaternary Science Reviews* 21: 1935–2017. © 2002, with permission from Elsevier.

Paraglacial valley fills

In glaciated areas, valley-fill deposits often comprise a lower sequence of glaciogenic deposits overlain by an upper sequence of paraglacial deposits. The paraglacial sequence may be exclusively alluvial, or include debris flow, floodplain, alluvial fan, lacustrine, and/or aeolian units in some combination (Owen, 1989; Figures 9 and 10). In the Walensee Valley (Switzerland), for example, the valley fill comprises patchy till overlain in sequence by glacilacustrine sediments, post-glacial lacustrine sediments, and then alluvial gravels. For this site, Müller (1999) calculated average sedimentation rates of 70–100 mm year⁻¹ immediately after deglaciation, reducing to only 3–4 mm year⁻¹ after ~12.3 ¹⁴C ka BP as the surrounding landscape stabilized. Many paraglacial valley fills have been deeply incised by trunk and tributary rivers, reflecting initial aggradation due to abundant sediment supply and then fluvial incision and terrace formation as sediment inputs diminished (Ballantyne, 2002a). Church (2002) has argued that as sediment supply in headward areas of large deglaciated basins is progressively reduced, floodplains in such areas experience net incision, whereas floodplains farther downstream continue to aggrade in response to sediment supply from farther upstream. He concluded that in British Columbia, floodplains up to about 175 km from headwaters are currently

degrading, whereas beyond this distance, aggradation is the norm, though the paraglacial signal in floodplain evolution is now overwhelmed in many catchments by more recent disturbances, particularly the effects of human land use.

Paraglacial Lacustrine Deposits

During deglaciation, proglacial lakes experience a transition from proximal, ice-contact sediment accumulation to ice-distal and then dominantly paraglacial sedimentation. The last is characterized by deltaic progradation (Figure 11) and the accumulation of fine-grained bottom deposits. In mountainous areas, reworking of valley-side glaciogenic sediments by rainstorm-triggered debris-flow events may interrupt the depositional sequence.

Studies of the inland 'fjord lakes' of British Columbia using geophysical profiling indicate that vast quantities of sediment were deposited on lake floors within decades during and immediately after glacier retreat. The largest, Okanagan Lake, is occupied by about 90 km³ of sediment up to 800 m thick. Overlying these deglacial deposits are units up to about 30 m thick, interpreted as representing slower paraglacial sediment influx throughout the Holocene (e.g., Gilbert and Desloges, 1992). A study by Hinderer (2001) of postglacial sediment

accumulation in major valleys and lake basins in the Alps similarly highlights rapid sediment accumulation during and immediately after deglaciation, with a progressive reduction in sediment input during most of the Holocene.

Paraglacial Coasts

Paraglacial coasts are those where remobilization of glaciogenic sediment significantly affects nearshore and offshore sediment budgets. There are two main sediment inputs. First, delivery of reworked glacial sediment by rivers enhances sediment flux in estuarine and fjord-head settings. Second, where glaciogenic sediments crop out at the coast, they are readily eroded by waves and tidal currents, introducing vast quantities of sediment into the nearshore zone. The location and



Figure 11 Paraglacial valley fill and delta progradation, Loch Achtriochan, Glencoe, Scotland.

duration of paraglacial sediment delivery at coasts is determined both by the disposition of glaciogenic deposits relative to the coast and by changes in relative sea level (Forbes and Syvitski, 1994). Where glaciogenic sediments are reworked by coastal erosion, rising sea levels allow wave action to access to new sediment sources; falling sea levels tend to isolate glaciogenic sediment sources from direct wave erosion, but may also result in enhanced sediment input from rivers where lowering base level stimulates river incision into floodplain and valley-fill sediments.

Fjords

In fjords, paraglacial sedimentation commences when retreating ice becomes land-based, and has four main components: (1) delta progradation from the fjord head and lateral tributary valleys, (2) sediment reworking by submarine mass movements and turbidity currents, (3) settling of suspended sediment on the fjord floor, and (4) reworking of glaciogenic sediments around fjord margins (Figure 12). Steep-walled, deep fjords tend to be dominated by (1)–(3); (4) may dominate in shallow fjords such as those of eastern Canada. Rising sea levels may submerge paraglacial deltas but encourage shoreface erosion; falling sea levels trigger erosion of deltas and upvalley outwash deposits, rejuvenating fluvial sediment supply. Glacier retreat from fjords may also result in a drastic fall in base level in tributary valleys, triggering a sequence of rapid landscape responses. Meigs et al. (2006), for example, have described how the recent retreat of the Tyndall Glacier and the opening of Taan Fjord in Alaska resulted in a 400-m drop in base level at the outlets of four tributary valleys, causing rapid fluvial erosion and evacuation of valley-fill sediments, incision of slot gorges in bedrock, and localized

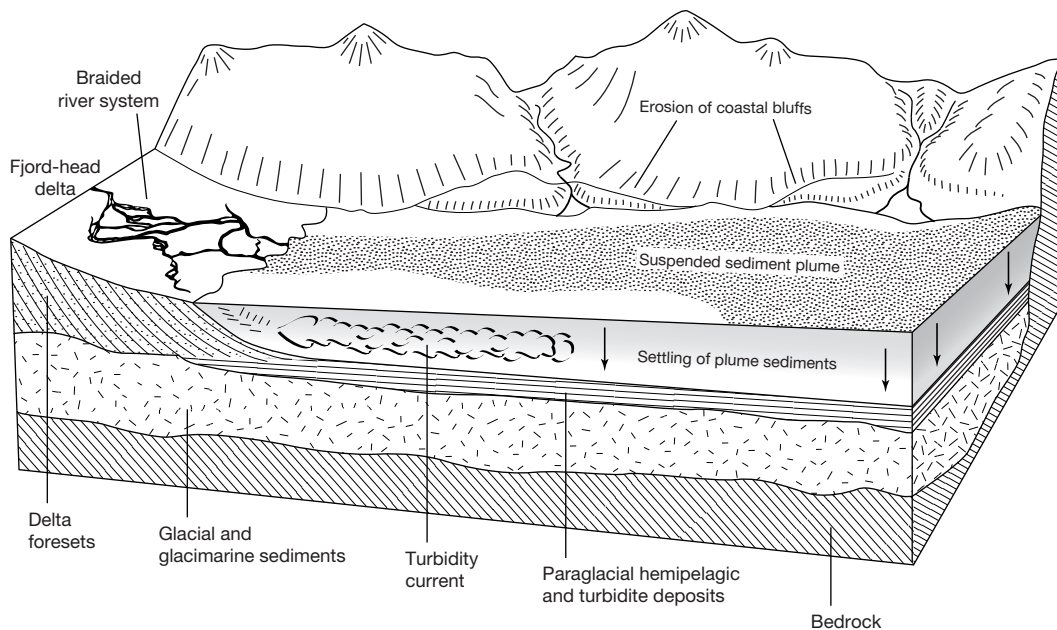


Figure 12 Paraglacial sedimentation in fjords (schematic). The dominant elements are (1) deposition of sand and gravel foresets at the front of a prograding fjord-head delta; (2) slumping of the delta front, generating turbidity currents; (3) settling of suspended particles from a surface sediment plume, and the resultant accumulation of hemipelagic bottom sediments; and (4) reworking of shoreface deposits by wave action. Reprinted from Ballantyne CK (2002) Paraglacial geomorphology. *Quaternary Science Reviews* 21: 1935–2017. © 2002, with permission from Elsevier.

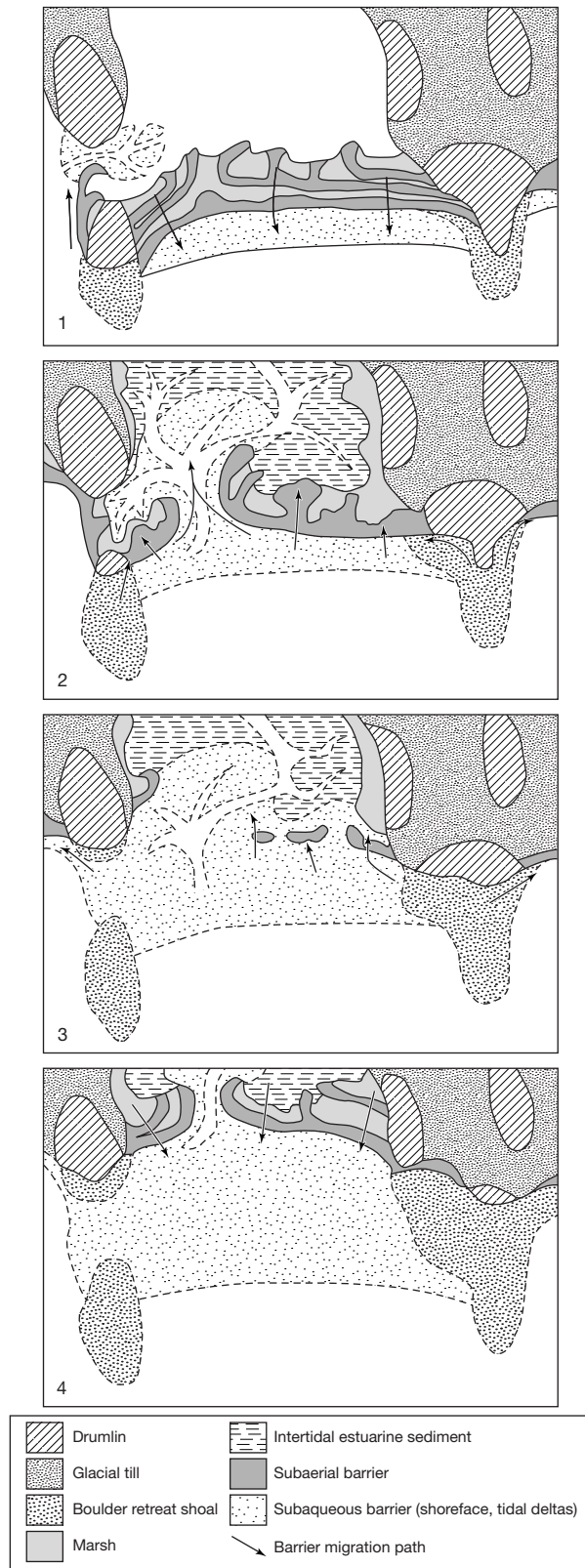


Figure 13 Cycle of formation, destruction, and re-formation of paraglacial barrier systems. (1) Marine reworking of glaciogenic deposits into a prograding barrier system. (2) As sediment supply diminishes, barriers retreat, losing sediment to washover, dune, and tidal inlet sinks.

landsliding on valley walls, with associated progradation of fan deltas from tributary outlets. Fjord sedimentation rates generally decline following deglaciation.

Barrier coasts

Paraglacial barrier coasts are those characterized by nearshore sediment accumulations (beaches, baymouth bars, spits, and barrier islands) fed by remobilized glaciogenic sediments. There is a transition from small barriers in rocky inlets where glaciogenic sediment sources are limited to coasts dominated by continuous barrier forms, such as the Beaufort Sea coast of Alaska or the southern Baltic coast. Both progradational and retrogradational barriers occur, depending usually on the balance between sediment supply and sea-level change; under conditions of diminishing sediment supply, barrier retreat, erosion, and destruction occur, and a new barrier system may develop landward of the original (Forbes et al., 1995; Figure 13). Subtypes of paraglacial coasts include those where marine transgression has partly drowned drumlins, which form discrete sources of sediment, and active outwash coasts such as SE Iceland. On barrier coasts where erosion of bouldery till has formed the sediment source, boulder lag deposits have accumulated, forming boulder barricades, pavements, and beaches that armor the shoreface and reduce the sediment supply to barrier structures.

Paraglacial shelf deposits

On many glaciated shelves, the majority of postglacial sedimentation comprises paraglacial deposits overlying distal glacial facies. Paraglacial shelf sediments are characterized by prograding deltaic wedges fronted by mass-flow and hemipelagic deposits. On the Alaskan shelf, for example, paraglacial sediment fines and thins seaward for ~75 km, and has an average thickness of ~55 m.

Wider Implications of Paraglacial Landscape Modification

All glaciated landscapes exhibit evidence of paraglacial modification. In many areas, such modification has been localized, and glacial landforms and sediments are preserved in near-pristine state. Elsewhere, particularly in steeplands, rock-slope adjustment and remobilization of glaciogenic sediment has transformed glacial landscapes. Most postglacial landscape evolution in glaciated environments, whether accomplished by rock-slope failure, landslides, debris flows, rivers, wind action, or coastal erosion, involves glacial conditioning, and hence can be described as paraglacial.

The widespread extent of Holocene paraglacial landscape modification has implications for understanding the long-term evolution of glacial landscapes. Such landscapes are not the result of Quaternary glaciation alone, but of recurrent cycles of alternating glacial and paraglacial erosion and sediment

(3) Destruction of original barrier. (4) Barrier regrowth from new sediment sources. Reprinted from Ballantyne CK (2002) Paraglacial geomorphology. *Quaternary Science Reviews* 21: 1935–2017. © 2002, with permission from Elsevier.

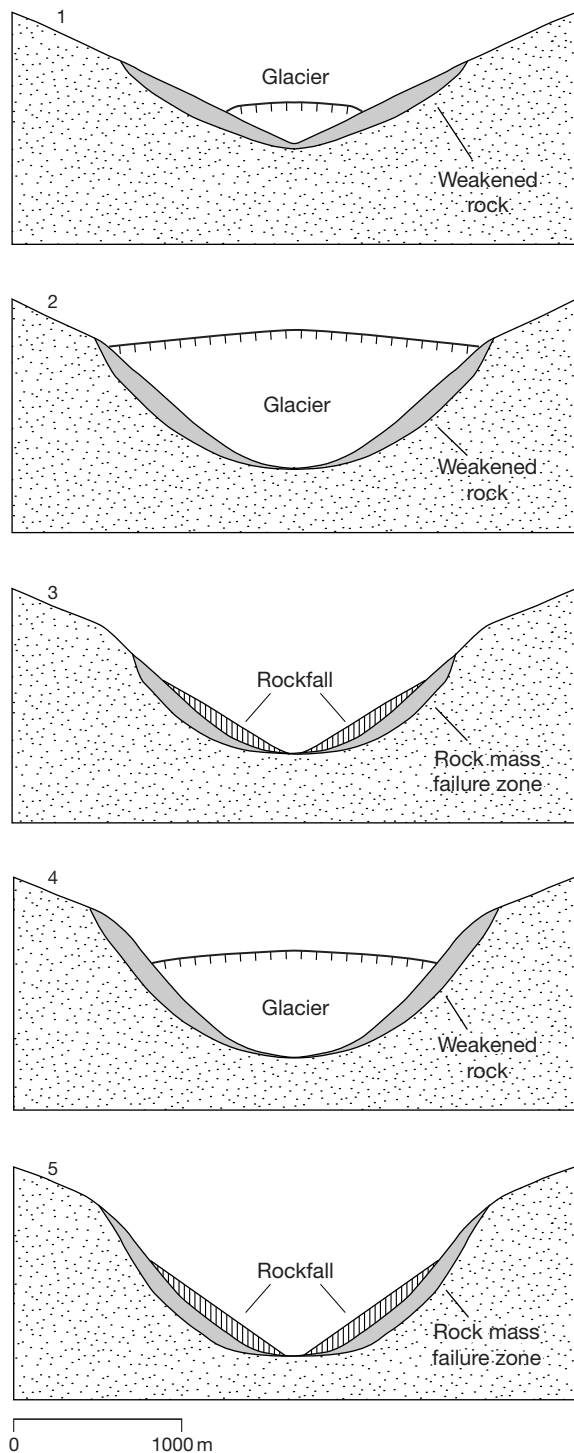


Figure 14 Model of glacial trough development on bedrock with low rock-mass strength during consecutive glacial and paraglacial phases. The evolving form of the valley is dependent not only on subglacial erosion, but also on paraglacial rock-slope adjustment during nonglacial intervals. Reprinted from Augustinus PC (1995) Glacial valley cross-profile development: The influence of *in situ* rock stress and rock mass strength, with examples from the Southern Alps, New Zealand. *Geomorphology* 14: 87–97. © 1995, with permission from Elsevier.

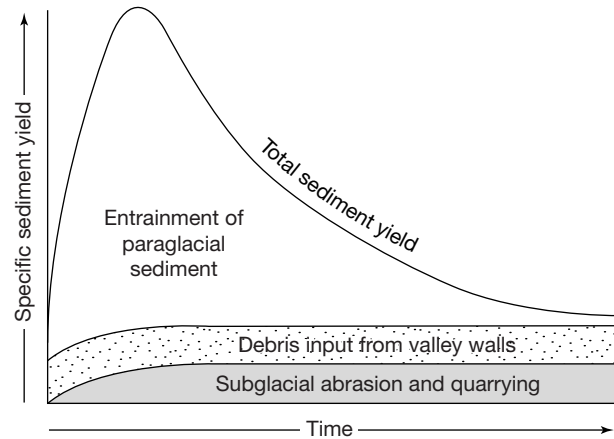


Figure 15 Rates of glacial sediment transport during glacier expansion. In the initial stages, sediment transport rates are high because of the abundance of readily entrainable sediment in paraglacial stores overridden by the advancing ice. As such sources are progressively exhausted, rates fall to lower levels sustained by subglacial erosion of sound bedrock and sediment delivery to the ice surface. Reprinted from Ballantyne CK (2002) Paraglacial geomorphology. *Quaternary Science Reviews* 21: 1935–2017. © 2002, with permission from Elsevier.

remobilization, transport, and redeposition. In glaciated mountains, paraglacial rock-slope failure, rock-slope deformation, and rockfall alter the form of rockwalls during interglacials; in addition, rock-mass weakening due to paraglacial stress release may determine the foci of glacial erosion during later periods of glacial advance (Augustinus, 1995; Figure 14). Thus, some archetypal forms of glacial erosion such as cirques, arêtes, and glacial troughs may owe their form as much to successive episodes of paraglacial rock-mass adjustment as to glacial quarrying and abrasion. Of more widespread importance is the role played by paraglacial sediment stores during the initial stages of glacier expansion. As such sediment stores provide an abundant source of readily entrainable debris, much of the sediment transported during the initial stages of glacial advance may be of paraglacial origin, rather than eroded from intact bedrock (Ballantyne, 2002a; Ballantyne and Benn, 1994). Where this is the case, the initial stages of glacier expansion are likely to be marked by enhanced rates of glacial sediment transport; when glacial recycling of paraglacial debris is complete, glaciers transport only sediment supplied by ‘normal’ entrainment processes such as subglacial erosion and gravitational debris delivery from valley-side slopes (Figure 15). Just as paraglacial processes are, by definition, glacially conditioned, so some aspects of glacial sediment transport reflect paraglacial conditioning or inheritance.

See also: **Fluvial Environments:** Responses to Rapid Environmental Change; Terrace Sequences. **Glacial Landforms:** Glacial Landforms, Sediments: Glacial Sequence Stratigraphy. **Permafrost and Periglacial Features:** Rock Glaciers and Protalus Forms; Slope Deposits and Forms; Talus Slopes. **Quaternary Stratigraphy:** Lithostratigraphy; Morphostratigraphy/Allostratigraphy.

References

- Augustinus PC (1995) Glacial valley cross-profile development: The influence of in situ rock stress and rock mass strength, with examples from the southern Alps, New Zealand. *Geomorphology* 14: 87–97.
- Ballantyne CK (2002a) Paraglacial geomorphology. *Quaternary Science Reviews* 21: 1935–2017.
- Ballantyne CK (2002b) A general model of paraglacial landscape response. *The Holocene* 12: 371–376.
- Ballantyne CK (2003) Paraglacial landform succession and sediment storage in deglaciated mountain valleys: Theory and approaches to calibration. *Zeitschrift für Geomorphologie Supplementband* 132: 1–18.
- Ballantyne CK and Benn DI (1994) Paraglacial slope adjustment and resedimentation following glacier retreat, Fåbergstølsdalen, Norway. *Arctic and Alpine Research* 26: 255–269.
- Ballantyne CK, Schnabel C, and Xu S (2009) Exposure dating and reinterpretation of coarse debris accumulations ('rock glaciers') in the Cairngorm Mountains, Scotland. *Journal of Quaternary Science* 24: 19–31.
- Barnard PL, Owen LA, and Finkel RC (2006) Quaternary fans and terraces in the Khumbu Himal south of Mount Everest: Their characteristics, age and formation. *Journal of the Geological Society of London* 163: 383–399.
- Bovis MJ (1990) Rock-slope deformation at Affliction Creek, southern Coast Mountains, British Columbia. *Canadian Journal of Earth Sciences* 27: 243–254.
- Church M (2002) Fluvial sediment transfer in cold regions. In: Hewitt K, et al. (ed.) *Landscapes of Transition*, pp. 93–117. Amsterdam: Kluwer.
- Church M and Ryder JM (1972) Paraglacial sedimentation: A consideration of fluvial processes conditioned by glaciation. *Geological Society of America Bulletin* 83: 3059–3071.
- Church M and Slaymaker O (1989) Disequilibrium of Holocene sediment yield in glaciated British Columbia. *Nature* 337: 452–454.
- Cossart E, Braucher R, Fort M, Bourlès DL, and Carcaillet J (2008) Slope instability in relation to glacial debuitressing in alpine areas (Upper Durance catchment, southeastern France): Evidence from field data and ^{10}Be cosmic ray exposure ages. *Geomorphology* 95: 3–26.
- Cossart E and Fort M (2008) Sediment storage and release in early deglaciated areas: Towards an application of the exhaustion model from the case of the Massif des Écrins (French Alps) since the Little Ice Age. *Norsk Geografisk Tidsskrift – Norwegian Journal of Geography* 62: 115–131.
- Cruden DM and Hu XQ (1993) Exhaustion and steady-state models for predicting landslide hazards in the Canadian Rocky Mountains. *Geomorphology* 8: 279–285.
- Curry AM, Cleasby V, and Zukowsky P (2006) Paraglacial response of steep, sediment-mantled slopes to post-'Little Ice Age' glacier recession in the central Swiss Alps. *Journal of Quaternary Science* 21: 211–225.
- Curry AM and Morris CJ (2004) Lateglacial and Holocene talus slope development and rockwall retreat on Mynydd Du, UK. *Geomorphology* 58: 85–106.
- Forbes DL, Orford JD, Carter RWG, Shaw J, and Jennings SC (1995) Morphodynamic evolution, self-organization, and instability of coarse-clastic barriers on paraglacial coasts. *Marine Geology* 126: 63–85.
- Forbes DL and Syvitski JPM (1994) Paraglacial coasts. In: Carter RWG and Woodroff CD (eds.) *Coastal Evolution: Late Quaternary Shoreline Morphodynamics*, pp. 373–424. Cambridge: Cambridge University Press.
- Friele PA, Ekes C, and Hicken EJ (1999) Evolution of Cheekye fan, Squamish, British Columbia: Holocene sedimentation and implications for hazard assessment. *Canadian Journal of Earth Sciences* 36: 2023–2031.
- Gilbert R and Desloges JR (1992) The late Quaternary sedimentary record of Stave Lake, southwestern British Columbia. *Canadian Journal of Earth Sciences* 29: 1997–2006.
- Hinderer M (2001) Late Quaternary denudation of the Alps, valley and lake fillings and modern river loads. *Geodinamica Acta* 13: 1–33.
- Iturrizaga L (2008) Paraglacial landform assemblages in the Hindukush and Karakoram Mountains. *Geomorphology* 95: 27–47.
- Matthews JA, Shakesby RA, Berrisford MS, and McEwen LJ (1998) Periglacial patterned ground in the Styggedalsbreen glacier foreland, Jotunheimen, southern Norway: Micro-topographical, paraglacial and geochronological controls. *Permafrost and Periglacial Processes* 9: 147–166.
- Meigs A, Krugh WC, Davis K, and Bank G (2006) Ultra-rapid landscape response and sediment yield following glacier retreat, Icy Bay, southern Alaska. *Geomorphology* 78: 207–221.
- Mercier D, Étienne S, Sellier D, and André M-F (2009) Paraglacial gullying of sediment-mantled slopes: A case study of Colletthøgda, Kongsfjorden area, West Spitsbergen (Svalbard). *Earth Surface Processes and Landforms* 34: 1772–1789.
- Müller BU (1999) Paraglacial sedimentation and denudation processes in an Alpine valley of Switzerland An approach to the quantification of sediment budgets. *Geodinamica Acta* 12: 291–301.
- Orwin JF and Smart CC (2004) The evidence for paraglacial sedimentation and its temporal scale in the deglaciating basin of Small River Glacier, Canada. *Geomorphology* 58: 175–202.
- Owen LA (1989) Terraces, uplift and climate in the Karakoram Mountains, northern Pakistan: Karakoram intermontane basin evolution. *Zeitschrift für Geomorphologie Supplementband* 76: 117–146.
- Owen LA, Benn DI, Derbyshire E, et al. (1995) The geomorphology and landscape evolution of the Lahul Himalaya, Northern India. *Zeitschrift für Geomorphologie* 39: 145–174.
- Ryder JM (1971) The stratigraphy and morphology of para-glacial alluvial fans in south-central British Columbia. *Canadian Journal of Earth Sciences* 8: 279–298.

Talus Slopes

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Introduction

Talus slopes are coarse clastic landforms produced by mass wasting processes. They may be broadly defined as distinctive accumulations of loose, coarse, usually angular rock debris at the foot of steep bare rock slopes. The terms talus (North American) and scree (English) are synonymous and refer to both the landform and its constituent material. Alternative terms include debris slopes (Gardner, 1980) and colluvial fans (Blikra and Nemec, 1998). Talus slopes occur in a wide range of environments but most significantly in those where physical weathering processes dominate and gravitational processes deliver freshly weathered products downslope. The production and/or accumulation of debris must be greater than its subsequent weathering or removal and the accumulations must be of sufficient thickness to develop a characteristic morphology independent of the underlying slope. Simple debris veneers only a few particles thick are termed 'debris mantled slopes' (Church et al., 1979). Most studies of talus slopes and talus slope processes have taken place in periglacial or alpine environments. The coarseness of most talus deposits makes them resistant to subsequent erosion: they are often stable, long-lasting elements of the landscape and preserved as relict or inactive forms. Hales and Roering (2005) mapped the distribution of scree slopes in a 80×40 km wide transect across the Southern Alps of New Zealand. They estimate that screes only cover 10% of slopes in the wetter areas west of the divide but may mantle up to 70% of the slopes in the drier eastern part, although only 20–25% are presently active.

Talus slopes vary considerably in morphology, composition, and geomorphic setting. Their basic characteristics depend primarily on the morphology and composition of the flanking cliffs, the geomorphic processes involved in their development, and the topography of the cliff foot zone. Although the dominant process is usually assumed to be rock fall, accumulation of talus material and its subsequent reworking may be the result of many different processes acting singly or in combination. This results in a gradational range of forms with several distinctive end members that nevertheless share many common characteristics.

Rock Fall and Rockfall Talus

Rock fall is the free or bounding fall of rock debris down steep slopes. Rock falls vary in size from individual pebbles to catastrophic failures of several million cubic meters (sturzstrom or rock avalanches). Rubble derived from very large single events is usually unsorted, extends much further from the cliff, and may have distinctive morphological features (e.g., ridges and furrows). Normal talus may subsequently develop below the failure surface, covering the upper parts of the feature. Smaller rock falls ($<10^{-1}$ – 10^2 m³) are the primary process associated

with the formation of talus slopes and may be classified into two types (Rapp, 1960). Massive vertical cliffs are dominated by primary rock falls where detachment is followed by direct transfer to the slope below. Primary rock falls are triggered mainly by pressure release, freeze–thaw activity, or weathering on the cliff face. However, on more complex cliff forms, debris may accumulate on irregularities in the cliff (e.g., benches, gullies) and subsequently be dislodged by other rock falls, snow avalanches, surface water flow, or animals. These secondary rock falls have different magnitude–frequency characteristics than primary rock falls. In an interesting discussion of rockfall provision to talus slopes, Krautblatter and Dikau (2007) identify cliff storage as a significant intermediate step between backwearing (detachment of rock from the host cliff) and rockfall deposition (i.e., delivery to the talus), noting that each may have different triggers.

Rockfall Triggers

The triggering mechanisms for rock falls have been inferred from inventory studies that compare the pattern of rock falls with simultaneous observations of temperature and precipitation. Diurnal and, to a lesser extent, seasonal variations in activity are usually studied by direct inventory or observational methods that record the timing of individual events, often some measure of their magnitude, in combination (where available) with meteorological observations from a proximal site. In some cases, they involve periods of continuous observations (e.g., Gardner, 1969, 1980; Luckman, 1976; Rapp, 1960). More generally, they consist of measurement of the accumulation of debris on marked surfaces over time. These have included boulders, plastic sheets (Krautblatter and Moser, 2009; Luckman, 1978, 1988), wooden traps (Prior et al., 1971), annual snow cover (Matsuoka and Sakai, 1999; Rapp, 1960), or netting (Gray, 1972). Decadal-scale inventory data are usually acquired on the basis of the data collected for other purposes, for example, rock falls onto railways (Lan et al., 2010; Rapp, 1960) or onto roads (Bjerrum and Jorstad, 1968; Luckman, 2008; Pack and Boie, 2002; Weiczorek et al., 1992), or through documentary archives. Such studies can yield rockfall volumes, seasonal, and, sometimes, daily records of activity, and again strongly emphasize the seasonal nature of the processes. Recently, several studies have used dendrochronology to address questions of the frequency and spatial distribution of rock falls from sites in Switzerland and elsewhere (e.g., Moya et al., 2010; Schneuwly and Stoffel, 2008a,b; Stoffel, 2006; Stoffel et al., 2010). These tree-ring studies are usually focused on the runout zones or peripheral areas of talus slopes and use scars and anomalous ring characteristics to identify rockfall events, determine rockfall hazards, or evaluate the efficacy of forests as a protection against rock fall. Examination of the position of damage within the annual ring may allow determination of the seasonality of impacts, although

predictably, the majority of impacts occur within the dormant period of growth (Schneuwly and Stoffel, 2008a,b).

At longer timescales, inventory data have been gathered from accumulated volumes on dated surfaces, for example, by lichenometric studies that date individual boulders (Luckman and Fiske, 1995; McCarroll et al., 1998) or the use of lichens to define a dated surface on which new arrivals can be clearly identified (e.g., André, 1997). These studies may allow determination of rates of rockfall activity on decadal or longer timescales and allow inferences concerning the variation of rockfall rates with changing climates, usually the Little Ice Age (e.g., Luckman and Fiske, 1997; McCarroll et al., 1998) or recent global warming (e.g., Gruber et al., 2004).

These inventory and observational data have led to the almost universal acceptance of frost splitting in bedrock and possibly subsequent ice melt as major trigger mechanisms for primary rock falls. This is based on observations of the diurnal or seasonal distribution of rock falls in the mountain and periglacial environments in which rock falls and talus are most frequently studied. In some cases, degradation (thaw) of mountain permafrost with increasing temperatures has also been identified as a potential rockfall trigger in the alpine zone of mountains (Davies et al., 2001; Gruber et al., 2004). However, detailed observation and modeling of the process is difficult and controversial and has generally been addressed through the context of experimental and theoretical studies on freeze-thaw weathering in periglacial environments. Recent work by Hallet and others (reported extensively by Hales and Roering, 2007, 2009) has suggested that simple volumetric expansion of freezing water within voids or planes of weakness in the rock (the traditional model of freeze-thaw weathering) is not capable of developing the confining pressures necessary to split rocks apart. They suggest that rocks are split by segregated ice growth resulting from migration of water through the rock to the ice lenses at subfreezing temperatures. This process is facilitated by thin interfacial water layers called premelted films and operates at stable temperatures between -3 and -8 °C, well below the fluctuations around 0 °C suggested as significant for conventional freeze-thaw activity. Frequent temperature oscillations across the freezing threshold are therefore unlikely to be a major rockfall trigger under this mechanism. Hales and Roering (2007) present a simple model of the depth and intensity of segregation ice growth in cliffs above screes in New Zealand to show that sites with the predicted maxima in intensity of segregation ice growth correspond with areas exhibiting maximum rockfall erosion rates and scree formation. This process controversy leads directly into the whole literature on the nature and controls of frost or freezing process in periglacial environments which is discussed elsewhere in this volume. Available data on the timing of rock falls usually remain sufficiently imprecise to determine whether falls are associated with the exertion of pressure during freezing (around 0 °C or at lower temperatures of -3 to -8 °C) or related to the release of frost-separated rocks following subsequent melting of the ice. However, the observed spring and fall timing for rockfall events indicates strong temperature dependence of the rockfall-triggering process.

Although less extensively reported, inventory data also indicate that rock falls are often associated with major precipitation events that destabilize loose materials on the slope.

Recently, Krautblatter and Moser (2009) studied accumulation over a 4-year period onto large areas of plastic sheeting (940 m^2) placed on talus slopes in the German Alps. They report that over 90% of the rock falls are associated with heavy precipitation events. However, these are mainly secondary rock falls mobilizing previously weathered materials from the cliff zone. Other studies also note a greater frequency of falls associated with precipitation events (e.g., Luckman, 1976) or relationships with the diurnal temperature regime on the cliff face, particularly periods of solar illumination (Gardner, 1980). These observations may also include records of fairly randomly distributed triggers of secondary falls, including root penetration and wedging, animal or human activity on the cliff, in addition to entrainment of material by downslope movement of rock falls, snow avalanches, or water that dislodges loose surface materials.

In addition to these more-conventionally discussed, climate-related, rockfall triggers, two other major trigger mechanisms are discussed in the literature. Several authors (e.g., Bull and Brandon, 1998; Bull et al., 1994; Guzzetti and Reichenbach, 2010; Mills, 1991) have noted patterns of considerable regional rockfall activity triggered by earthquakes. Additionally, several authors have suggested that unloading or pressure release may be a significant cause of higher rock fall, rock avalanche, or landslide activity. This is particularly cited as an explanation for greater rates of rock fall and talus accumulation inferred to have occurred immediately following deglaciation when the downwasting of ice cover removed lateral support from valley sides previously buttressed by ice, leading to a period of accelerated rates of rock fall or rockslide activity (e.g., Hinchliffe and Ballantyne, 1999; Wieczorek and Jäger, 1996).

Rockfall Movement

When rock falls reach the talus, their subsequent downslope movement may be arrested completely or the rock falls may continue to travel downslope in a series of bouncing trajectories of diminishing length and height, or by rolling or sliding over the surface (see Dorren, 2003; Dorren et al., 2006). These bouncing boulders may produce characteristic 'bump holes' in the surface materials of the slope (Rapp, 1960) and/or considerable damage to vegetation or human infrastructures beyond the talus (see, e.g., Dorren et al., 2006; Schneuwly, 2010).

Movement paths may be very irregular depending on the nature of the impacts and the size and shape of the boulder. Boulders often shatter on their first impact and the independent fragments travel on separate paths. Boulder size and shape are critical. Approximately equidimensional boulders tend to roll or bounce following impact, whereas less compact forms may break up on impact. The downslope movement of disc-shaped boulders can vary considerably. Discoid boulders landing on edge (i.e., with their long-axis vertical) may generate considerable angular momentum, rotating about their center of gravity, and travel 'wheel-like' for long distances downslope. However, should they land on their side (the 'a-b' face) or lose their vertical orientation, they will be quickly arrested or slide only short distances over the surface. Downslope trajectories will also vary with the properties of the surface. For example, where discoid boulders land on snow, the boulders may mire

(landing on edge) or slide depending on their orientation on impact. This erratic behavior is dependent on individual boulder characteristics and the detailed configuration of obstacles in the boulders' path and makes modeling of rockfall paths and runout distances challenging because of the many potentially random elements involved.

Rockfall Hazards

The largest boulders may come to rest some distance below the talus foot, and many talus slopes have a runout zone or a basal fringe of 'out runners' (Blikra and Nemec, 1998). As a result of the obvious hazard associated with rock fall, considerable work has been carried out in modeling rockfall activity. Much of this work is summarized by Dorren (2003), Guzzetti and Reichenbach (2010), and Lan et al. (2010). Dorren (2003) divides these models into empirical, process, and GIS-based models (e.g. Marquinez et al., 2003) and they may further be subdivided into two- or three-dimensional models depending on whether the rock falls are allowed to deviate laterally from a straight downslope path. Many talus slopes have a basal runout fringe of boulders (known as the rockfall shadow zone; Evans and Hungr, 1993) extending beyond the lower limit of the talus. Particular attention has been paid to modeling maximum runout distances in this zone because of the obvious hazard to human structures. Early, two-dimensional, empirical models used Heim's (1932) concept of rockfall fahrböschung (defined as the angle between the top of the slope and the furthest traveled boulder) to define this rockfall hazard zone. Subsequent work from British Columbia by Evans and Hungr (1993) suggests that the outer limit of the shadow zone is best defined by the intersection between the ground surface and a line projected downslope at an angle of 27.5° from the top of the appropriate talus slope.

Three-dimensional modeling of rockfall trajectories has become increasingly prevalent, and numerical, physics-based, models of rockfall trajectories, draped over digital terrain models of potential downslope pathways and supplemented by GIS analyses, have been used to produce sophisticated models of rockfall trajectories (see Guzzetti and Reichenbach, 2010; Guzzetti et al., 2003; Lan et al., 2010 and references therein) that predict or reconstruct rockfall paths and runout distributions. These allow a more accurate prediction of runout distances and may be used to model activity in a variety of settings and at scales ranging from individual sources to regional patterns of activity. For example, Dorren et al. (2006) compare three-dimensional modeling with the results of real-size experiments (boulders ~ 0.95 m diameter) to evaluate the effectiveness of forest cover as protective devices to mitigate rockfall hazards.

Rockfall Talus Morphology

Rockfall talus slopes result from the accumulation of rockfall debris from discrete rockfall events over long periods of time. The plan morphology of talus slopes depends on the form of the cliffs supplying the debris and the morphology of the surface on which the debris accumulates. Relatively simple cliffs, straight in plan view and of uniform or similar geology, tend to produce sheet (straight) rockfall talus slopes lacking

significant lateral variation in their characteristics (Figure 1). However, as the cliffs become dissected, rock falls moving downslope are channeled into gullies or basins, and deposition is focused below couloirs, leading to differential accumulation and the eventual development of talus cones. As well as funneling rock falls, these couloirs channel other gravitational transfer processes down the cliff face (e.g., surface stream flow and snow avalanches) that may result in significant modification of the talus below. Therefore, cones developed below dissected cliffs are rarely simple, single-process (rock fall) forms. The upper slopes of the talus may be gullied or reworked by other processes, resulting in more complex landforms (Figure 2).

Many talus slope profiles are characterized by long straight segments at or about $33\text{--}35^\circ$, often with a strong basal concavity. These slopes have been the focus of considerable field measurement, experimentation, theoretical analyses, and discussion. Initial talus models assumed that these straight slopes were developed by sliding and failure of loose unconsolidated materials (termed dry avalanching or, more recently, grain



Figure 1 A classic straight sheet talus slope developed primarily by rock fall at Small River, British Columbia, Canada. Note the straight slope, slight basal concavity, and 'run out' fringe of boulders beyond the talus foot.



Figure 2 Coalescing multiprocess talus cones, resting on raised beaches, Templefjorden, Spitzbergen. Note the lower angle, debris-flow-modified cone on the left.

flows) and that the slopes were at the angle of repose of these materials (see Carson, 1977; Statham, 1976). This view was challenged by Statham (1976), who developed a model based on energy expenditure and loss by individual rockfall particles moving over the surface. Extensive reports on measurements of talus slope angles (e.g., Francou and Manté, 1990; Sauchyn, 1986) indicate that measured angles of the straight upper portions of talus slopes are usually between $\sim 32\text{--}37^\circ$ and up to 40° . However, mean talus angles may be as low as $25\text{--}30^\circ$ due to the basal profile concavity. Several authors report differences in mean slope angles as a result of different process combinations (e.g., Church et al., 1979; Sauchyn, 1986). Francou and Manté (1990) propose that talus slopes have a segmented profile: an upper, usually straight slope and a lower section with a basal concavity, separated by a critical angle of $\sim 33\text{--}34^\circ$. They consider that the upper slope is essentially a transport surface, whereas the lower slope characteristics result from both transport and depositional processes. The length and angle of the basal concavity depend on site characteristics, maturity of talus development, and the depositional processes involved.

Direct observation at many sites suggests that dry avalanches (grain flows) on talus slopes tend to be quite small and are confined to relatively small-sized gravel material on the upper slopes (e.g., Van Steijn et al., 2002). Moreover, examination of many talus slopes indicates that the characteristic coarse openwork surface texture is merely a veneer. The openwork surface acts like a sieve: smaller particles or rock chips landing on the slope are trapped or washed into open voids, which become filled with fines and rock chips at depth. Most active rock falls raise dust clouds from the cliffs and talus, and there may also be considerable fine material deposited on the talus from niveo-aeolian processes (Héty and Gray, 2000). Therefore, despite their coarse surface appearance, most talus slopes have considerable quantities of interstitial fine material at depth or exposed on the higher parts of the slope. They are not exclusively formed of the coarse cohesionless material that early models assumed and many subsequent experimental studies simulated.

Surface Processes

Talus creep is the downslope movement of individual particles resting on the talus surface. It results from a combination of many processes, including frost action, running water, rockfall impact, needle ice, animal disturbance, and snow avalanches. This process has frequently been monitored using painted stones or lines on the talus in several long-term studies (e.g., Gardner, 1980; Perez, 1993). Rates of movement on upper talus slopes where fines and smaller clasts are mixed typically average $1\text{--}10\text{ cm year}^{-1}$ (greater where more fines are present), but individual stones may move much greater distances as a result of rock fall or avalanche impacts. The movement of large clasts or those on the lower slopes is much less and generally due to burial or dislodgement by direct impact. Several authors have described a process of dry avalanching of materials or grain flows (van Steijn et al., 2002) – though often seen in modeled talus or in gravel piles, this is usually fairly small-scale on natural talus slopes. On upper talus slopes, frost-susceptible fines at or near the surface favor creep through frost action or,

in some extreme cases, solifluction. Several authors have reported miniature sorted stripes on the upper portion of talus slopes.

Talus Materials

Talus deposits consist of angular, irregular rock fragments with a wide range of sizes. The nature of the debris depends to a large extent on the lithology (or lithologies) exposed in the cliff, as joint and bedding characteristics are primary determinants of the shape and size of talus particles. Most talus slopes show a distinctive increase in particle size of surface materials downslope. This is known as ‘fall sorting’, is often logarithmic in form (e.g., Church et al., 1979; Statham, 1976) and is most marked at the base of the slope. Fall sorting is common on rockfall talus slopes, though not universal, and primarily results from two mechanisms. Larger boulders have greater momentum and therefore tend to travel greater distances downslope. Secondly, the frictional resistance (roughness) of the surface over which the boulder slides, rolls, or bounces is defined by the relationship between the size of the moving boulder and the irregularities of the surface (boulders and voids) over which it is passing. Big boulders are only effectively trapped in large ‘holes’ or when they impact other large boulders or obstacles and lose momentum. Therefore, these boulders come to rest in areas of the talus with boulders (holes) of a similar size. The degree of sorting depends on the slope length, cliff height, and the size and shape of dominant particles. It is best developed on slopes with a wide size range of boulders. Locally random effects or differences in particle shapes may result in the absence of or anomalous patterns of sorting. For example, the frictional properties of snow-covered talus are quite different than snow-free talus: large boulders may be trapped in wet snow or slide over the surface and result in local pockets of coarser materials. Processes other than rock fall may modify these sorting patterns (see ‘Modification of Talus Forms’, below).

Several studies of particle shape indicate a tendency for larger particles on the lower slopes to be more equidimensional (e.g., Perez, 1989). Unless they are small fragments, platy or elongate clasts tend to break up with repeated impacts. Also, ‘blocky’ clasts rolling or bouncing downslope can rotate about any axis with similar efficiency, whereas ‘discs’ or ‘rods’ can lose momentum rapidly if their axis of rotation is changed by impact, for example, from rolling to sliding. Talus fabric is also, to some extent, lithology-dependent. Where clasts are elongate or platy, the fabrics tend to dip downslope and plunge less steeply than the talus slope, giving an upslope imbrication (Perez, 1998). Elongate clasts may show a ‘rolling’ fabric (long axis across slope) or a sliding fabric (long axis downslope; Blikra and Nemec, 1998). However, given greater equidimensionality of debris on many lower slopes, fabrics are often, as might be expected, fairly random, unless the talus material is predisposed to an elongated shape by the lithologic characteristics of the material or emplaced by a process favoring sliding; for example, Perez (1989) describes a strong downslope orientation for blocks at Lassen Peak where emplacement is dominated by sliding over a seasonal snow cover.

In recent years, there have been detailed studies of the sedimentology of talus and related deposits (Bertran et al.,

1997; Blikra and Nemec, 1998; van Steijn et al., 1995, 2002) that have developed criteria to differentiate between talus, snow avalanche, debris flow, and other mass-wasting deposits found within talus-like formations. Blikra and Nemec characterize talus as a colluvial deposit consisting of “tongue shaped beds with marked upslope fining, normal grading and mainly openwork texture” that may show a ‘rolling,’ downslope or random fabric depending on the dominant process and the lithologically determined shape of clasts. The deposits of most other processes that influence talus slope development contain a matrix of fines. However, the differences in sedimentological characteristics in these sediments are subtle and difficult to categorize precisely in the absence of surface morphological evidence.

Bedded or ‘stratified screes’ (grèzes litées) were first described from periglacial deposits in Northern France (see van Steijn et al., 1995). They are composed of platy or friable calcareous rocks that show dipping beds of coarser openwork gravels alternating with matrix-filled gravels. Studies of these deposits suggest that the openwork beds result from dry grain flows, whereas the matrix-supported deposits are associated with debris flows, sheet flow, or quasialluvial processes. Recently, Hétu and others (Hétu and Gray, 2000; van Steijn et al., 2002) have described contemporary analogs for these deposits from talus slopes developed from friable, shaley materials in Gaspésie, Québec. Hétu has also described small ‘frost coated clastic flows’ (FCCFs) that are similar to grain flows, but the clasts are ice-coated and move downslope as small 0.5–1.5 m wide flows at $2\text{--}6\text{ m s}^{-1}$ on slopes between $\sim 36\text{--}24^\circ$. The alternation of matrix-rich beds with finer clasts with these coarser openwork deposits can lead to the development of stratified screes. This may be a realistic analog for periglacial conditions in Northern France during the last glaciation. Van Steijn et al. (2002) indicate that dry flows are seen in all climates and are not necessarily periglacial phenomena. FCCF are, however, formed in periglacial environments, with suitably friable lithologies, a high frequency of freeze-thaw cycles and a limited snow cover. They suggest that dry avalanche deposits and FCCFs may be differentiated based on the angle of slope: dry flows cease at angles below $\sim 34^\circ$, whereas FCCFs may flow on slopes as low as 24° (Van Steijn et al., 2002).

Modification of Talus Forms

As talus deposits accumulate, geomorphic processes other than rock fall may rework loose surface materials, modifying talus morphology and surface characteristics. These changes are most frequently seen in periglacial or alpine environments where long-continued modification of talus can lead to the development of distinctive landforms associated with snow avalanches, debris flows, and permafrost creep.

Snow Avalanches

In arctic or alpine areas, snow avalanches running down the cliff face may entrain material and deposit it on the talus below (Rapp, 1960). However, as these snow avalanches move over the talus, they may also erode or incorporate loose surface



Figure 3 Talus slopes modified by extensive snow avalanche activity resulting in avalanche boulder tongues, Pallenvagge, Abisko Mountains, Sweden. The darker talus is more stable and heavily lichen-covered, and the lighter areas show surface reworking by snow avalanches. The basal profiles of the well-developed avalanche tongues (center and left) have strong basal concavities.

material, frequently exposing fines at the surface. This material is transported downslope, where it is deposited on coarser material as a scattered cover of precariously positioned smaller boulders or rock fragments, let down from an ablating snow cover. These phenomena are known as avalanche or perched boulders (Jomelli, 1999; Luckman, 1988; Rapp, 1960). The upper, eroded slopes show little size sorting and may be stripped of loose materials, but there is a rapid increase in mean grain size toward the base, where avalanched material is mixed with coarse blocky rockfall debris. Long-continued or intense avalanche activity may carry debris well beyond the limits of the talus, resulting in a loose scatter of coarse debris or the development of an avalanche boulder tongue (Jomelli and Francou, 2000; Luckman, 1978; Rapp, 1959). These ‘roadbank tongues’ can be recognized by a raised tongue of debris extending beyond the original talus, often with an asymmetric cross-section and a smoothed beveled top, flanked by steep side slopes and a lobate front. Talus slopes modified by avalanches exhibit a pronounced basal concavity, a strong size sorting of surface debris at their lower end and often extend a considerable distance beyond the toe of adjacent, normal rockfall talus across slopes of as little as $8\text{--}10^\circ$ (Figure 3).

Debris Flows

The upper portion of talus slopes contains considerable fines mixed with small rock clasts. Many talus cones are dissected by small gullies that head at the base of the cliff and decrease in size downslope. During heavy, intense rainstorms, water is channeled down the couloirs onto the talus below, mobilizing debris within the gullies on both rock wall and talus, saturating the apex area and generating small hillside debris flows. Single or multiple examples of these features with their characteristic channels, levee margins, and terminal lobes may be generated on the talus, partially covering earlier rockfall-deposited material (Figure 4). Once initiated, these headward gullies will continue to function as pathways for subsequent events, ultimately producing an irregular, often somewhat radial, pattern

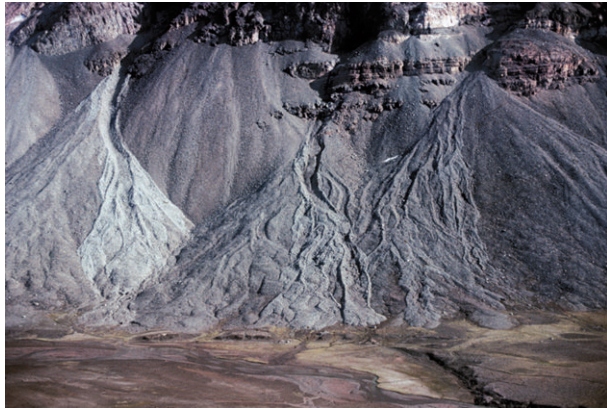


Figure 4 Talus cones, extensively modified by debris-flow activity, Tanquary Fjord, Ellesmere Island, Canada. Note the characteristic levees and terminal lobes of the debris flows and the lower mean slopes of the heavily modified 'alluvial' talus.

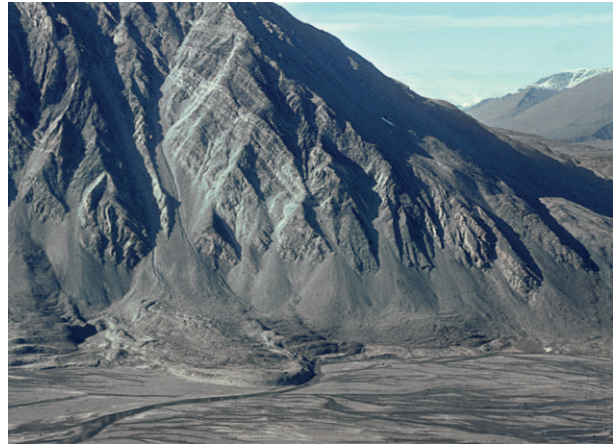


Figure 6 Talus cones showing modification into a variety of rock-glacier forms by permafrost creep, Tanquary Fjord, Ellesmere Island, Canada.



Figure 5 Talus slopes, partially relict and grass-covered (left) but modified by snow avalanche activity to create avalanche boulder tongues (center and right) and subsequently reworked by hillside debris-flow activity, Lairig Ghru, Cairngorm Mountains, Scotland.

of debris-flow sediments extending downslope across the lower slopes of the talus. Subsequent erosion or dispersal of these materials takes place by later debris-flow events or snow-avalanche activities ([Figure 5](#); [Luckman, 1992](#)). These forms have been christened 'alluvial talus' and over long periods of time, they may develop into high-angle debris-flow or alluvial cones.

Protalus and Rock Glaciers

Talus slopes are most frequently studied in periglacial environments. At some talus sites, a perennial firn (snow) patch may develop at the base of the slope. Rockfall or avalanche debris landing on this icy surface slides to the base, accumulating as a ridge which has been termed a protalus rampart or nivation ridge.

In many relatively dry, cold alpine areas, thick talus accumulations contain permafrost. The continued buildup of talus and ice in the basal talus may lead to the development of a convex bulge (protalus bulge) in the lower talus. Subsequent

deformation and creep of this rock-ice mixture leads to rock-glacier development that transports the coarse talus material away from the base of the slope ([Figure 6](#)). Rates of movement are rarely more than a centimeter per year (see [Rock Glaciers and Protalus Forms](#)).

Rates of Development

Talus slopes below rock cliffs are a measure of the accumulated removal of debris from these cliffs and therefore have been the focus of many attempts to determine rates of rockwall recession and landscape change in a variety of environments. Attempts at measurement have been carried out using a number of techniques at different timescales. Short-term measurements involve determination of the accumulation on marked surfaces for periods of up to 10–15 years (e.g., [Luckman, 1988](#); [Rapp, 1960](#); [Sass and Wollny, 2001](#)). Such estimates are hampered by spatial (the relatively small proportion of the talus sampled) and temporal (only a few years) sampling problems. An extension of these studies has used lichenometry to estimate accumulation rates over longer periods of time (>1–300 years; [André, 1997](#); [Luckman and Fiske, 1995](#); [McCarroll et al., 1998](#)). Finally, there have been several attempts to calculate volumes based on well-defined cones, by interpolating the sub-talus bedrock profile from adjacent slopes ([André, 1997](#); [Hales and Roering, 2007](#); [Rapp, 1958](#)). In recent years, the potential for error has been reduced by using ground-penetrating radar to estimate the sub-talus surfaces ([Sass and Wollny, 2001](#)). The rates determined by all of these measures are invariably fairly low, often <1 cm per millennium but exceptionally up to 1 m. They vary with environment, techniques used, and cliff geology, with little systematic pattern (see summaries in [André, 1997](#); [Krautblatter and Dikau, 2007](#); [Sass and Wollny, 2001](#)). In an attempt to examine these controls, André's studies in Spitzbergen indicate that rates of rockwall recession rarely exceed 1–2 m during the Holocene, except for sites with glacier-related stress relaxation on rock walls, highly frost-susceptible bedrock, or

exceptionally high cliffs. Krautblatter and Dikau (2007) present an extensive discussion of these estimates and discuss a series of theoretical models of the relationships between rock-wall recession (backwearing), intermediate storage on the cliff, and rockfall supply to the talus below.

The low contemporary estimates of debris production and the presence of many inactive talus slopes in formerly glaciated areas have led to the suggestion that much of the talus accumulation in these areas is paraglacial in origin; that is, these features accumulate rapidly during a period of accelerated mass-wasting activity immediately following deglaciation due to the release of confining pressure in the rock walls and exposure to atmospheric conditions (André, 1997; Ballantyne, 2002; Luckman and Fiske, 1995). Mills (1992) has documented talus slopes, 80–100 m thick with well-developed morphology and fall sorting that have formed over an 8-year period in the crater of Mt St. Helens. This indicates that classical talus forms can develop quite rapidly, although in this case, the process is considerably accelerated by abundant freshly shattered bedrock and continued earthquake activity. Nevertheless, the presence of relict or vegetated talus in many areas and relatively low rates of accumulation in some contemporary studies have led to an interesting discussion as to whether the high rates of accumulation reflect a paraglacial (induced by accelerated rockwall failures following glacier recession) or periglacial (severe freeze-thaw activity in a colder postglacial climate) origin (see Ballantyne, 2002).

Paleoenvironmental Significance

In recent years, some attention has been directed to the paleoenvironmental significance of talus slopes and deposits. Many of these studies involve the examination of gully sections or trenches cut in the upper parts of talus that are presently inactive (Curry and Black, 2003; Hinchliffe, 1999). Alternating layers of openwork clasts interbedded with finer grade deposits, peats, or paleosols are interpreted as intervals of renewed rockfall activity and stability, respectively, related to cooler and/or wetter climate intervals that can be dated by ^{14}C . Hétu and Gray (2000) describe colluvial sequences interbedded with paleosols that reflect alternating process regimes during the Little Ice Age and are, in part, conditioned by the changes in the forest cover. Similarly, explanations for grèzes litées deposits invoke alternating periods of gravitational (dry avalanching or FCCF deposits) and debris-flow/slopewash deposits attributed to periods of wetter and/or colder intervals. However, although these sediments reflect different processes, deposition on talus slopes is discontinuous and episodic, and individual stratigraphic units may be quite thin. The dating control is often limited and the deposits may only represent a single, almost instantaneous event, for example, a debris flow or single depositional event (see Krautblatter and Moser, 2009), rather than decade-century intervals with a different climatic regime. Contemporary studies of rainfall-triggered debris-flow activity indicate that relatively low-frequency storms may trigger synchronous events over areas of $>10\text{ km}^2$ (Luckman, 1992; Rapp, 1960), producing similar stratigraphic evidence on different talus slopes. Similarly, although heavily lichen-covered rock fall or avalanche debris that occurs beyond

the limits of contemporary activity indicates that there have been larger events in the past, it does not necessarily indicate that there have been changes in the rate or intensity of activity. It may simply indicate that these larger scale events have recurrence intervals that are greater than the effective range of the dating technique used.

Therefore, given that the magnitude-frequency spectrum of processes that contribute debris to talus slopes is poorly defined and it is difficult to determine whether individual deposits represent a single event or multiple events, paleoclimate inferences from these deposits should be made with caution. In many cases, the deposits may represent individual high-magnitude, infrequent events that have little or no relationship to 'average climate' conditions.

See also: [Permafrost and Periglacial Features: Paraglacial Geomorphology; Rock Glaciers and Protalus Forms; Slope Deposits and Forms.](#)

References

- André M-F (1997) Holocene rockwall retreat in Svalbard: A triple rate evolution. *Earth Surface Processes and Landforms* 22: 423–440.
- Ballantyne CK (2002) Paraglacial geomorphology. *Quaternary Science Reviews* 21: 1935–2017.
- Bjerrum L and Jorstad FA (1968) *Stability of Rock Slopes in Norway*, pp. 1–11. Norwegian Geotechnical Institute Publication 79. Oslo: Norwegian Geotechnical Institute.
- Blikra LH and Nemec W (1998) Postglacial colluvium in western Norway: Depositional processes, facies and paleoclimatic record. *Sedimentology* 45: 909–959.
- Bertran P, Francou B, Texier J-P, and van Steijn H (1997) Fabric characteristics of subaerial slope deposits. *Sedimentology* 44: 1–16.
- Bull WB and Brandon MT (1998) Lichen dating of earthquake-generated regional rockfall events, Southern Alps, New Zealand. *Bulletin of the Geological Society of America* 110: 60–84.
- Bull WB, King J, Kong F, Moutoux T, and Phillips WM (1994) Lichen dating of coseismic landslide hazards in alpine mountains. *Geomorphology* 10: 253–264.
- Carson MA (1977) Angles of repose, angles of shearing resistance and angles of talus slopes. *Earth Surface Processes* 2: 363–380.
- Church M, Stock RF, and Ryder JM (1979) Contemporary sedimentary environments on Baffin Island, N.W.T., Canada: Debris slope accumulations. *Arctic and Alpine Research* 11: 371–402.
- Curry AM and Black R (2003) Structure, sedimentology and evolution of rockfall talus, Mynydd Du, south Wales. *Proceedings of the Geologists' Association* 114: 49–64.
- Davies MCR, Hamza O, and Harris C (2001) The effect of rise in mean annual temperature on the stability of rock slopes containing ice-filled discontinuities. *Permafrost and Periglacial Processes* 12: 137–144.
- Dorren LKA (2003) A review of rockfall mechanics and modeling approaches. *Progress in Physical Geography* 27: 69–87.
- Dorren LKA, Berger F, and Putters US (2006) Real size experiments and simulation of rockfall on forested and non-forested slopes. *Natural Hazards and Earth System Sciences* 6: 145–153.
- Evans SG and Hungri O (1993) The assessment of rockfall hazard at the base of talus slopes. *Canadian Geotechnical Journal* 30: 620–636.
- Francou B and Manté C (1990) Analysis of the segmentation in the profile of Alpine talus slopes. *Permafrost and Periglacial Processes* 1: 53–60.
- Gardner JS (1969) Rockfall: A geomorphic process in high mountain terrain. *Albertan Geographer* 6: 15–20.
- Gardner JS (1980) Frequency, magnitude and spatial distribution of mountain rockfalls and rockslides in the Highwood Pass area, Alberta, Canada. In: Coates DR and Vitek J (eds.) *Thresholds in Geomorphology*, pp. 267–295. London: Allen and Unwin.

- Gray JT (1972) Debris accretion on talus slopes in the central Yukon Territory. In: Slaymaker O and McPherson HJ (eds.) *Mountain Geomorphology*, pp. 75–84. Vancouver: Tantalus Press.
- Gruber S, Hoelzle M, and Haeberli W (2004) Permafrost thaw and destabilization of Alpine rock walls in the hot summer of 2003. *Geophysical Research Letters* 31: L13504.
- Guzzetti F and Reichenbach P (2010) Rockfalls and their Hazard. In: Stoffel M, Bollschweiler M, Butler DR, and Luckman BH (eds.) *Tree Rings and Natural Hazards: A State-of-the-Art*, pp. 129–137. Berlin: Springer.
- Guzzetti F, Wieczorek GF, and Reichenbach P (2003) Rockfall hazard and risk assessment in the Yosemite Valley, California, USA. *Natural Hazards and Earth System Science* 3: 491–503.
- Hales TC and Roering JJ (2005) Climate controlled variations in scree production, Southern Alps, New Zealand. *Geology* 33: 701–704.
- Hales TC and Roering JJ (2007) Climatic controls on frost cracking and implications for the evolution of bedrock landscapes. *Journal of Geophysical Research* 112: F02033 1029/2006JF000616.
- Hales TC and Roering JJ (2009) A frost “buzzsaw” mechanism for erosion of the eastern Southern Alps, New Zealand. *Geomorphology* 107: 241–253.
- Heim A (1932) Bergsturz und mens. *Beiblatt zur Vierteljahrschrift der Naturforschenden Gesellschaft in Zürich* 77: 1–217.
- Héty B and Gray JT (2000) Effects of environmental change on scree slope development throughout the postglacial period in the Chic-Choc Mountains in the northern Gaspé Peninsula, Québec. *Geomorphology* 32: 135–155.
- Hinchliffe S (1999) Timing and significance of talus slope reworking, Trotternish, Skye, northwest Scotland. *The Holocene* 9: 483–494.
- Hinchliffe S and Ballantyne CK (1999) Talus accumulation and rockwall retreat, Trotternish, Isle of Skye, Scotland. *Scottish Geographical Journal* 115: 53–70.
- Jomelli V (1999) Dépôts d’avalanches dans les Alpes françaises: Géométrie, sédimentologie et géodynamique depuis le Petit Âge glaciaire. *Géographie physique et Quaternaire* 53: 199–209.
- Jomelli V and Francou B (2000) Comparing the characteristics of rockfall talus and snow avalanche landforms in an Alpine environment using a new methodological approach: Massif des Ecrins, French Alps. *Geomorphology* 86: 130–143.
- Krautblatter M and Dikau R (2007) Towards a uniform concept for the comparison and extrapolation of rockwall retreat and rockfall supply. *Geografiska Annaler* 89A: 21–40.
- Krautblatter M and Moser M (2009) A nonlinear model coupling rockfall and rainfall intensity based on a four year measurement in a high Alpine rock wall (Reintal, German Alps). *Natural Hazards and Earth System Sciences* 9: 1425–1432.
- Lan H, Martin CD, Zhou C, and Lim CH (2010) Rockfall hazard analysis using LIDAR and spatial modeling. *Geomorphology* 118: 213–223.
- Luckman BH (1976) Rockfalls and rockfall inventory data: Some observations from Surprise Valley, Jasper National Park, Canada. *Earth Surface Processes* 1: 287–298.
- Luckman BH (1978) Geomorphic work of snow avalanches in the Canadian Rockies. *Arctic and Alpine Research* 10: 261–276.
- Luckman BH (1988) Debris accumulation patterns on talus slopes in Surprise Valley, Alberta, Canada. *Géographie physique et Quaternaire* 42: 247–278.
- Luckman BH (1992) Debris flow and snow avalanche landforms in the Lairig Ghru, Cairngorm Mountains, Scotland. *Geografiska Annaler* 74a: 109–121.
- Luckman BH (2008) Forty years of rockfall accumulation at the Mount Wilcox Site, Jasper National Park, Alberta, Canada. *Geographica Polonica* 81(1): 79–91.
- Luckman BH and Fiske CJ (1995) Estimating long-term rockfall accretion rates by lichenometry. In: Slaymaker HO (ed.) *Steepland Geomorphology*, pp. 233–255. London: Wiley.
- Luckman BH and Fiske CJ (1997) Holocene development of coarse debris landforms in the Canadian Rockies. In: Matthews J, Brunsden D, Frenzel B, and Weiss M (eds.) *Rapid Mass Movement and Climatic Variation During the Holocene* *Paläoklimaforschung* 19: 283–297.
- Marquinez J, Menéndez-Duarte R, Farias P, and Jiménez-Sánchez M (2003) Predictive GIS-based model of rockfall activity in mountain cliffs. *Natural Hazards* 30: 341–360.
- Matsuoka N and Sakai H (1999) Rockfall activity from an alpine cliff during thawing periods. *Geomorphology* 28: 309–328.
- McCarroll D, Matthews JA, and Shakesby RA (1998) Spatial and temporal patterns of Holocene rockfall activity on a Norwegian talus slope estimated by lichenometry. *Arctic and Alpine Research* 30: 51–60.
- Mills HH (1991) Temporal variation of mass-wasting activity in Mount St. Helens Crater, Washington, U.S.A. indicated by seismic activity. *Arctic and Alpine Research* 23: 417–423.
- Mills HH (1992) Post-eruption erosion and deposition in the 1980 crater of Mount St. Helens, Washington, determined from digital maps. *Earth Surface Processes and Landforms* 17: 739–754.
- Moya J, Corominas J, and Arcas JP (2010) Assessment of rockfall frequency for hazard analysis at Sòla D’Andorra (Eastern Pyrenees). In: Stoffel M, Bollschweiler M, Butler DR, and Luckman BH (eds.) *Tree Rings and Natural Hazards: A State-of-the-Art*, pp. 161–175. Berlin: Springer.
- Pack RT and Boie K (2002) Utah Rockfalls Hazard Inventory: Phase 1. Salt Lake City: Utah Department of Transport.
- Perez FL (1989) Talus fabric and Particle Morphology on Lassen Peak, California. *Geografiska Annaler* 71A: 43–57.
- Perez FL (1993) Talus movement in the high equatorial Andes: A synthesis of ten years of data. *Permafrost and Periglacial Processes* 4: 199–215.
- Perez FL (1998) Talus morphology, clast fabric, and botanical indicators of slope processes on the Chaos Crags (California Cascades). *Géographie physique et Quaternaire* 52: 47–68.
- Prior DB, Stephens N, and Douglas GR (1971) Some examples of mudflow and rockfall activity in north-east Ireland. In: Brunsden D (ed.) *Slopes: Form and Process*, pp. 93–110. Institute of British Geographers, Special, Publication No.3. Oxford: Alden Press.
- Rapp A (1958) *Talus Slopes and Mountain Walls at Templefjorden, Spitzbergen*. Oslo: Norsk Polarinstitutt Skrifter 119.
- Rapp A (1959) Avalanche boulder tongues in Lapland. *Geografiska Annaler* 41: 34–48.
- Rapp A (1960) Recent development of mountain slopes in Kärkevagge and surroundings. *Geografiska Annaler* 42: 73–200.
- Sass O and Wolny K (2001) Investigations regarding Alpine talus slopes using ground penetrating radar (GPR) in the Bavarian Alps, Germany. *Earth Surface Processes and Landforms* 28: 1071–1088.
- Sauchyn DJ (1986) Particle size and shape variation on alpine debris fans, Canadian Rocky Mountains. *Physical Geography* 7: 191–217.
- Schneuwly DM (2010) Reconstruction and spatial analysis of rockfall frequency and bounce heights derived from tree rings. In: Stoffel M, Bollschweiler M, Butler DR, and Luckman BH (eds.) *Tree Rings and Natural Hazards: A State-of-the-Art*, pp. 177–180. Berlin: Springer.
- Schneuwly DM and Stoffel M (2008a) Spatial activity analysis of rockfall activity, bounce heights and geomorphic changes over the last 50 years—a case study using dendrogeomorphology. *Geomorphology* 102: 522–531.
- Schneuwly DM and Stoffel M (2008b) Tree-ring based reconstruction of the seasonal timing, major events and origin of rockfall on a case-study slope in the Swiss Alps. *Natural Hazards and Earth System Science* 3: 491–503.
- Statham I (1976) A scree slope rockfall model. *Earth Surface Processes* 1: 43–62.
- Stoffel M (2006) A review of studies dealing with tree rings and rockfall activity: The role of dendrogeomorphology in natural hazards research. *Natural Hazards* 39: 51–70.
- Stoffel M, Schneuwly DM, and Bollschweiler M (2010) Assessing rockfall activity in mountain forest—implications for hazard assessment. In: Stoffel M, Bollschweiler M, Butler DR, and Luckman BH (eds.) *Tree Rings and Natural Hazards: A State-of-the-Art*, pp. 139–155. Berlin: Springer.
- van Steijn H, Bertran B, Francou B, and Texier J-P (1995) Models for the genetic and environmental interpretation of stratified slope deposits: A review. *Permafrost and Periglacial Processes* 6: 125–146.
- van Steijn H, Boelhouwers J, Harris S, and Héty B (2002) Recent research on the nature, origin and climatic relations of blocky and stratified slope deposits. *Progress in Physical Geography* 26: 551–575.
- Wieczorek GF and Jäger S (1996) Triggering mechanisms and depositional rates of postglacial slope-movement processes in the Yosemite Valley. *Geomorphology* 15: 17–31.
- Wieczorek GF, Snyder JB, Alger CS, and Issacson KA (1992) Rock falls in Yosemite Valley, California. U.S.G.S. Open File Report 93–387.

Thermokarst Topography

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Introduction

Thermokarst terrain exhibits a pitted relief formed from the melting of ground ice. 'Thermokarst' is used widely to describe all processes and landforms that are associated with melting ground ice, but thermokarst, or thaw, lakes are probably the most well-known features (Figure 1; French, 2007). Thermokarst processes are generally only considered with respect to permafrost terrain, and soil displacement due to melting of seasonal frost is not included. Thermokarst topography and landforms are found throughout the permafrost regions, wherever ground ice is exposed or melted. Thermokarst features are uncommon in bedrock areas, but may occur locally in depressions where a sedimentary cover, including ground ice, has accumulated.

The geotechnical significance of permafrost derives from the presence of ice in the ground, and the terrain instability that follows thawing of ice-rich permafrost. The volume of water released as ice melts is commonly higher than the saturation water content of the soil, and so pore-water pressures rise. In flat terrain, consolidation of thawed soil leads to subsidence of the ground and an undulating topography, because the rate and extent of thawing is spatially variable. This variability stems primarily from the heterogeneous distribution of ground ice. In sloping terrain, high pore-water pressures lead to a reduction in effective stress and loss of soil strength so that slopes may fail. Landslides are common in permafrost terrain that has been burned by forest fire, because the resulting change in surface characteristics, not the heat of the fire, raises ground temperatures and leads to permafrost degradation (Burn, 1998). Thermokarst processes may be problematic for infrastructure and design of construction projects, which modify surface conditions and warm the ground. There are numerous examples of structures that have become unserviceable due to melting of ground ice beneath them (Figure 2; Crawford and Johnston, 1971; French, 1975). While the thermal effects of artificial surfaces can be modeled reliably (Andersland and Ladanyi, 1994), the ground-ice distribution at a site is commonly less well known and the source of most unforeseen incidents.

Permafrost and the Active Layer

Permafrost is the ground that remains at or below 0 °C for 2 or more years. Permafrost terrain consists of a seasonally thawed active layer underlain by perennially frozen ground. The base of the active layer is at the top of permafrost. Continental permafrost is found where the climate sustains frozen ground, with vast extents in northern regions of Canada, Alaska, and Siberia. Alpine or plateau permafrost occurs at high elevation, in areas where continental permafrost is not found. There are areas of permafrost in the major mountains of the world, and an extensive distribution in the Tibetan Plateau. In total, about

15% of the land surface of the Northern Hemisphere is underlain by permafrost (Zhang et al., 2000).

Permafrost is classified by the proportion of the exposed land area of a region that contains perennially frozen ground. By definition, permafrost is considered continuous where it underlies more than 90% of the terrain. In these areas, the spatial mean ground temperature is well below 0 °C, commonly <−5 °C, and only a few locations have a microclimate that does not support permafrost. The amount of unfrozen ground increases in the discontinuous permafrost zones, so that the proportions of perennially frozen ground in the extensive and sporadic discontinuous permafrost zones are 90–50% and 50–10%, respectively (Zhang et al., 2000). Isolated patches of permafrost occur where less than 10% of the ground supports permafrost. The spatial mean ground temperature increases with the proportion of unfrozen ground. Thermokarst features occur in all permafrost regions because the critical requirement is ground ice, while surface disturbances may raise ground temperatures above 0 °C at any location if either permafrost is exposed to the atmosphere, or water accumulates to depths exceeding the winter-ice thickness.

The thickness of the active layer depends on a variety of factors, principally the temperature and duration of the summer, thermal properties of near-surface materials, ice content of the soil, and the nature of heat transfer in the near surface (Williams and Smith, 1989). However, conditions during the preceding freezing season are also significant, particularly the average winter temperature and snow depth, which control the temperature of the active layer before thawing begins. The deepest active layers are found in coastal beaches and lake shores, where convection enhances the transfer of heat to the thawing front, and in bedrock where there is relatively little ice to melt each summer. Deep active layers are also common in well-drained dry ground, such as gravel. Relatively shallow active layers are found in peatlands, where evaporation within damp moss maintains cool ground temperatures and preserves permafrost. In the Northern Hemisphere, the southern limits of continental permafrost are found in peatlands.

Near-Surface Ground Ice

Characteristically, there is an ice-rich zone at the top of permafrost, just below the active layer. This layer may have particularly high ice contents in fine-grained sediments in moist environments (Figure 3), but it has also been observed in bedrock. If the active layer deepens and permafrost thaws, then the near-surface ground ice will melt. The presence of this ice-rich horizon is the basis for thermokarst development, and the reason that permafrost terrain is considered sensitive to disturbance. Commonly, the ice occurs in subhorizontal lenses of centimeter thickness, segregating similarly sized beds of sediment (segregated ice lenses), or as bodies of almost pure 'massive' ice.

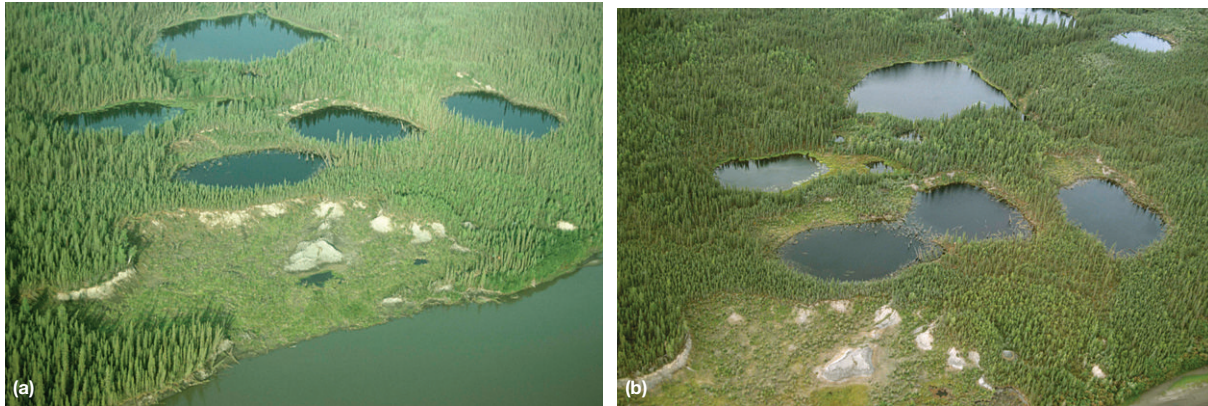


Figure 1 Thermokarst lakes near Mayo, Yukon Territory, Canada (see [Burn and Smith, 1990](#)). (a) In July 1988; (b) in July 1998, showing expansion that led to the merging of two lakes. Photographs by C.R. Burn.



Figure 2 Two buildings in Dawson City, Yukon Territory, that are no longer in use because of thaw settlement beneath them. Photograph taken in July 1994 by C.R. Burn.



Figure 3 Ice-rich near-surface permafrost exposed in a slumped block at Pelly Island, western Arctic coast, Canada. The block is an ice-wedge polygon that was undercut by shoreline erosion, and split along the ice wedges. There is a black organic horizon at the top of permafrost, an ice vein has been truncated by the early Holocene thaw unconformity, and an ice wedge is exposed on the left side of the block. Determinations of the near-surface volumetric ice content of permafrost on the island ranged up to 58% ([Burn, 1997](#)). Photograph taken in August 1993 by C.R. Burn.

The most common forms of massive ice are ice wedges that form from repeated infilling of thermal contraction cracks by snow melt. These linear bodies are V-shaped in cross section, are commonly capped by the active layer, and extend downward for about 5 m ([Figure 3](#)). They form outline networks of tundra polygons, which are easily seen in lowlands, but are obscured on hill slopes by soil movement. Other massive icy bodies may be formed near the base of the active layer by injection of water during the autumn freezeback, due to pressures generated during freezing of coarse-grained sediments. In some locations, bodies of massive ice several meters thick may be buried glacier ice, remnant from the last glaciation, or masses formed during regional aggradation of permafrost.

Where the ice is distributed within the mineral soil, the ground is particularly ice rich in the upper 2 or 3 m of permafrost ([Figure 3](#)). The ice-rich zone has been called the ‘intermediate layer’ of permafrost, because it may thaw episodically due to climate variation ([Shur et al., 2005](#)). However, in some locations, these sediments were formerly within the active layer only during the Holocene climatic optimum, and have accumulated ice during the subsequent cooling and thinning of the active layer ([Burn, 1997](#)). Similarly ice-rich horizons are found just below the active layer in depositional environments, where the top of permafrost has aggraded with sedimentation or peat growth. The ice enrichment occurs due to water being drawn along the ground-temperature profile into freezing soil and permafrost at the end of summer and during freeze back of the active layer ([Mackay, 1983](#)).

Massive ice and individual ice lenses may be nearly 100% ice, although air bubbles are commonly distributed through the ice body. The ice-rich zone of segregated ice lenses is generally also of high ice content. In sandy materials, the volume of ice may be 30% greater than the saturated volume of thawed soil, but in fine-grained sediments, the super saturation may be greater ([Burn, 1997](#)).

Disturbances to Permafrost

The ground surface temperature, T_s , and thermal properties of ground materials control the heat flux through the active layer

into permafrost. T_s is controlled by the surface energy balance, providing a solution for

$$Q^* = Q_H + Q_E + Q_G \quad [1]$$

where Q^* is the net radiation at a site, Q_H is the flux of sensible heat between the surface and the atmosphere, Q_E is the latent heat flux (evaporation or condensation) between the surface and the atmosphere, and Q_G is the ground heat flux between the surface and underlying soil. Changes to surface characteristics alter these fluxes. For instance, drying of the surface reduces Q_E , and a higher T_s is then required to increase Q_H and Q_G to maintain their equivalence with Q^* . In permafrost regions, changes in surface conditions characteristically lead to either deepening of the active layer, long-term degradation of permafrost, or both (Mackay, 1970). Deepening of the active layer is a result of surface warming in summer, but increases in snowfall or winter temperatures may also lead to an increase in thaw depth.

In addition to T_s , the thermal properties of surface materials are critical to the stability of permafrost, because they control the heat flux through the active layer. In a substantial portion of the permafrost region, especially in the discontinuous zones, perennially frozen ground is covered by a blanket of peat and mosses. This low-density organic layer has a low thermal conductivity and is commonly moist. As a result of its damp state, heat supplied from the surface is used in evaporation rather than warming the ground, thus maintaining cool conditions at depth. The organic layer is critical to the stability of the active layer, and disturbance of the surficial organic cover leads to thawing of near-surface permafrost (Mackay, 1970; Zoltai, 1993).

Forest fires create the most widespread surface disturbances in permafrost regions. Fires change several surface characteristics to influence the surface energy balance (Figure 4).

First, Q^* increases at the ground surface following a reduction in shade and a decrease in surface albedo from charred debris. Second, the organic horizon may be incinerated, removing its effective insulation, and drying the surface. Third, the reduction in canopy cover and accumulation of deadfalls increases snow depth, reducing heat loss from the ground in winter. Burn (1998) measured an increase in mean annual soil temperature, at 20-cm depth, of 1.25 °C between proximal

burned and unburned sites in southern Yukon Territory, where an intense fire in 1958 had incinerated the surface organic horizon. The increase in ground temperature prompted long-term permafrost degradation at the burned site. A 25-year record of active-layer development following fire at Inuvik, Northwest Territories, detected initial deepening of the active layer, but aggradation of permafrost as vegetation recovered (Mackay, 1995).

Deepening of the active layer leads to subsidence of the ground where near-surface permafrost is ice rich (Mackay, 1970). Figure 5 shows that active-layer deepening leads to lowering of the surface of permafrost by the sum of the increase in active-layer thickness and ground subsidence. As the ice content of permafrost increases, for a given increase in active-layer depth, the surface subsidence increases. Where the degree of surface disturbance or near-surface ice content is spatially variable, subsidence will vary and thermokarst topography will develop. Mackay (1970) and French (1975) provide several examples of surface disturbances that led to ground subsidence.

Permafrost Degradation

Long-term degradation of permafrost will occur if the mean annual ground temperature rises above 0 °C. Such long-term thaw is a function of the mean surface temperature, T , the volumetric latent heat in permafrost, L , and the thermal conductivity of thawed materials, λ . The Stefan solution has been successfully used under field conditions to approximate the relation between the depth of thaw, z , and the time, t , since thawing began (Burn, 1998, 2000; Burn and Smith, 1990):

$$z = \sqrt{(2\lambda Tt/L)} \quad [2]$$

or,

$$z = b\sqrt{t} \quad [3]$$

$$b = \sqrt{(2\lambda T/L)} \quad [4]$$

This model assumes that all heat entering the ground is used to thaw permafrost, rather than warm thawed soil above

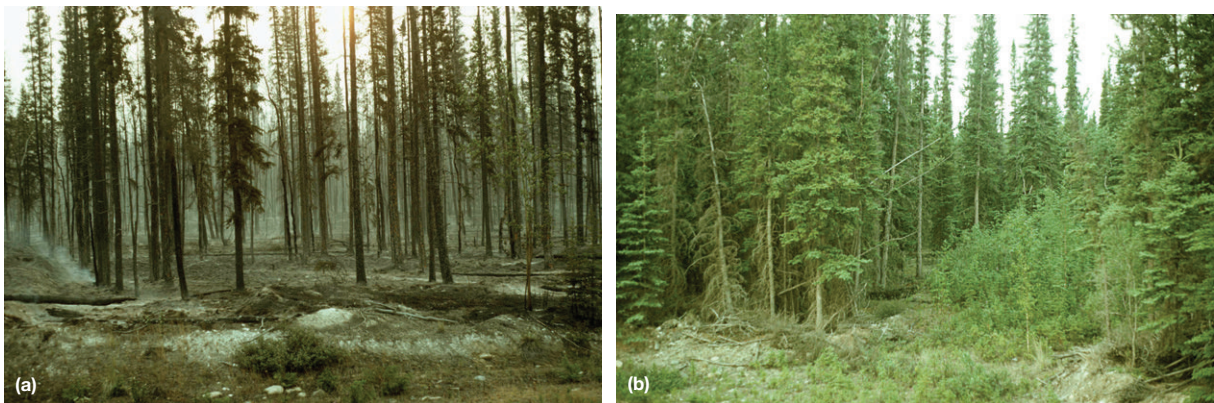


Figure 4 Change in surface conditions wrought by a forest fire in the boreal forest of central Yukon Territory, Canada, during July 1995. The Klondike Highway was used to control the fire. (a) East of the highway where the forest was burned; (b) to west of the highway, with undisturbed forest. Photographs by C.R. Burn.

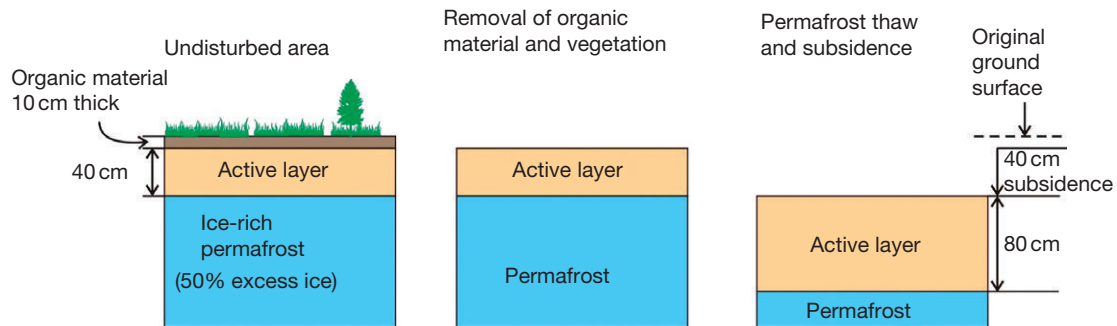


Figure 5 Thaw consolidation of permafrost terrain due to deepening of the active layer. The presence of excess ice in near-surface permafrost leads to lowering of the ground surface as the active layer deepens, and an absolute lowering of the permafrost table greater than the increase in active-layer depth. Diagram by S. Smith, Geological Survey of Canada. Reproduced from Mackay JR (1970) Disturbances to the tundra and forest tundra environment of the western Arctic. *Canadian Geotechnical Journal* 7: 420–432.

the permafrost or the permafrost itself, and that the surface temperature history involves a step change to the value that generates thawing. The assumptions are reasonable, because the latent heat of fusion is one to two orders of magnitude greater than the heat required to raise the same volume of soil materials by 1 °C, and because natural changes in surface conditions require decades for recovery of vegetation to former conditions. The rate of permafrost degradation may be obtained by differentiating eqn [3] with respect to time:

$$\frac{dz}{dt} = \left(\frac{b}{2\sqrt{t}} \right) \quad [5]$$

Therefore, the rate of permafrost degradation declines with time since disturbance.

The Stefan solution is a one-dimensional analysis of permafrost degradation, and may be applied to the rate of thaw beneath a disturbance. However, at the edge of disturbances heat flow is both vertical and horizontal, in response to the temperature gradients across the boundary of the disturbance. In this case, the rate of horizontal expansion of the disturbance may be obtained from eqn [5], with $t=1$ (Burn and Smith, 1990). The relation may be applied when ground-ice conditions are spatially continuous. Commonly thermokarst disturbances are halted as local ground ice is exhausted.

Thermokarst Lakes

Ponding of water in depressions formed by melting of near-surface ground ice commonly leads to deepening of the active layer and degradation of permafrost. While the depth of a lake is less than one-half to two-thirds of the maximum winter-ice thickness, permafrost may be maintained beneath the lake, but once the pool exceeds this amount, the mean annual water-bottom temperature rises above 0 °C, and permafrost is unsustainable beneath the pond (Figure 1). As a result, the pond deepens and expands. The rate of expansion depends on the ground-ice content of near-surface permafrost. Figure 1 shows actively expanding thermokarst lakes in central Yukon Territory, growing in ice-rich glaciolacustrine sediments (Burn and Smith, 1990). These lakes are enlarging at their shores at a rate of about 0.5 m year⁻¹.



Figure 6 Tundra lake on Richards Island, western Arctic coast, Canada, showing the dark water of the deep central pool, and scars of retrogressive thaw slumps around the lake. Photograph taken in August 1991 by C.R. Burn.

There are numerous thaw lakes in permafrost regions because the frozen ground inhibits drainage. The Arctic coastal plains of northern Alaska, Siberia, and Canada's western Arctic contain thousands of lakes (Figure 6), but most of these are thousands of years old (Hinkel et al., 2003; Rampton, 1988). In many cases, the lakes have reached a steady state with the surrounding permafrost, so that the taliks of unfrozen ground beneath the lakes are of constant size (Burn, 2002). The taliks may be perched above permafrost for small lakes, but for large lakes, the talik may penetrate permafrost. The width of lakes where the taliks penetrate permafrost depends on the temperatures at the lake bottom and in the surrounding permafrost, and the geothermal gradient (Mackay, 1963). Many tundra lakes have shallow littoral terraces containing permafrost (Figure 6), which may extend for over 100 m from the shore. The terraces shield permafrost around the lake from thaw, so these lakes are quasistable. Degradation of permafrost around these lakes may be initiated by terrestrial processes such as mass movements, rather than lacustrine effects.

The expansion of thermokarst lakes follows eqn [2] (Burn and Smith, 1990). The rate of deepening depends on the ground-ice content, thermal conductivity of thawed lake-bottom sediments, and on the lake-bottom temperature.

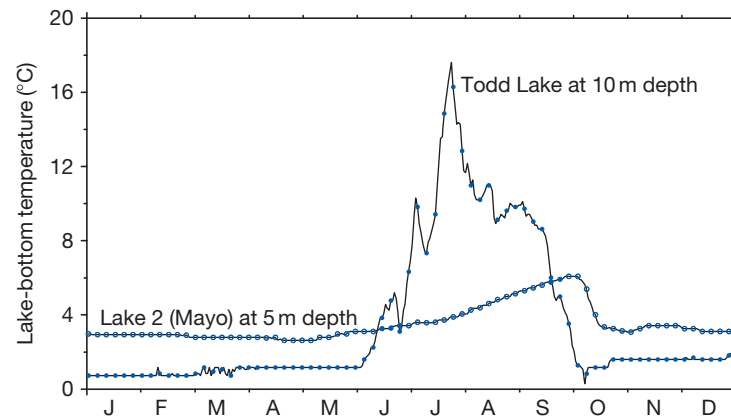


Figure 7 Daily mean lake-bottom temperatures recorded in 2001 at 10-m depth in the central pool of Todd Lake, a tundra lake near the western Arctic coast of Canada (Burn, 2002), and at 5-m depth in Lake 2 at Mayo (Burn and Smith, 1990). Todd Lake is well mixed by forced convection in summer, but Lake 2 is shielded from the wind and is poorly mixed. The series are plotted from daily data, but the symbols are plotted only every 5 days for clarity. The annual mean temperature at both of these depths is 3.5 °C.

Lake-bottom temperature is sensitive to the extent of mixing in the water column. Figure 7 indicates the annual temperature cycle in a small thermokarst lake in the boreal forest near Mayo, Yukon Territory, and a tundra lake near Canada's western Arctic coast. The tundra lake is well mixed in summer, so that warming of the lake bottom occurs rapidly. The forest surrounding the other lake acts as a wind break, and the lake is not mechanically mixed. Hence, warming at the lake bottom occurs by conduction, although cooling in fall is by density-induced free convection. A result is that the mean annual lake-bottom temperatures are similar at the sites. These data illustrate the independence of lake-bottom temperatures from climate, for the mean annual air temperature at Mayo is about -3°C , while at the coast it is about -10°C .

Thermokarst lakes may cease growing if the surrounding ground ice is exhausted, the basin becomes infilled by sedimentation or plant growth, or the lake drains. Drainage, or partial drainage, of thaw lakes is well documented, and commonly occurs if the active layer surrounding a lake deepens to below water level, so that once flow begins, it accelerates from a trickle to a torrent (Mackay, 1997). Drainage is rarely complete, and residual ponds are common in drained basins. Extensive depressions may be exposed by drainage of large lakes. These are best-known from Siberia, where they are called 'alasses.'

Retrogressive Thaw Slumps

Thaw slumps are the most active geomorphological features of permafrost terrain. Erosion of river banks, along the shores of lakes, or at the coast, may expose permafrost and initiate ablation of ice-rich ground. In river banks, high water during the spring flood may remove covering sediments and overhanging mats of peat, or undercut banks that subsequently collapse to reveal ice-rich ground. Along lake shores and at the coast, erosion during storms may similarly undercut ice-rich ground which is subsequently exposed (Figure 3). Once exposed to air temperatures above 0°C , or to direct insolation, the ice-rich ground may thaw and turn to a mud slurry due to the high water content. This mud falls to the base of the

exposure and flows down to the water body. The exposed permafrost commonly forms a steep headwall ranging from 20° to 90° (Figure 8(a)), while the thawed materials accumulate in a more gentle foot slope of inclination between 3° and 15° , stretching from the foot of the headwall to the water body (Figures 8 and 9). Headwall ablation rates of up to 16 m year^{-1} have been measured, but more commonly the headwall retreat is about 10 m year^{-1} in a range of environments (Burn and Lewkowicz, 1990). Thaw slumps continue to develop until the headwall enters sediments of low ice content, as with the slump on Figure 1; the headwall is covered by an insulating layer of thawed debris, commonly an overhanging peat mat, or the terrain geometry allows the foot slope to rise and cover the permafrost. The longevity of thaw slumps therefore depends on ground-ice distribution and local topography. Many thaw slumps near the coast are rejuvenated by continuing coastal erosion, for instance, the slumps in Figure 9 are within a larger stabilized area.

Direct thawing of permafrost occurs in the ablating headwall, where the gradient is too steep to retain thawed material on the surface. In contrast, on the foot slope, thawed material accumulates to shield the underlying permafrost. The disturbance of surface conditions from the previously undisturbed vegetation cover may be sufficient to cause long-term thawing of the permafrost beneath the foot slope. The foot slope commonly develops a deeper snow cover than the surrounding undisturbed ground because the rough accumulation of surface debris prevents drifting. In the boreal forest, trees are toppled by the thaw slump, so shading of the surface is less on the foot slope than in undisturbed ground. In the boreal forest, such disturbance raises the mean ground temperature above 0°C , and permafrost is only reestablished after vegetation succession restores the original forest. An increase in mean annual ground temperature of about 4°C over undisturbed conditions has been measured at the thaw slump illustrated in Figure 8 (Burn, 2000).

Kokelj et al. (2009) pointed out that at many tundra lakes in western Arctic Canada retrogressive thaw slumps reinitiate after stabilization, and that there may be a depression in the lake bottom close to the lake shore adjacent to thaw slumps.

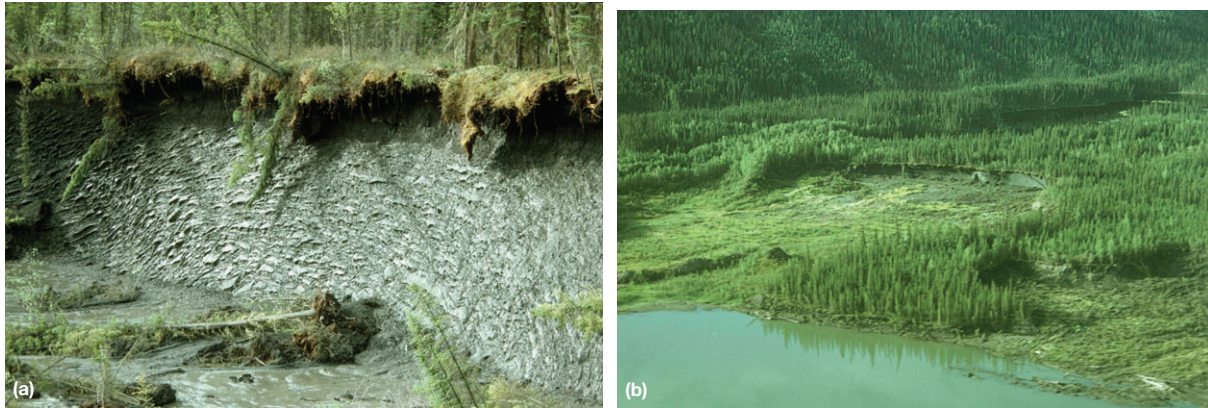


Figure 8 A retrogressive thaw slump near Mayo, central Yukon Territory, Canada. (a) Ice-rich glaciolacustrine sediments exposed in the 10-m-high headwall (June 1984); (b) overview of the slump showing the headwall and footslope of about 350 m between the headwall and Stewart River (August 1984). River erosion of ice-rich permafrost initiated the slump in 1949, and the headwall retreated from the river at up to 16 m year^{-1} . Photographs by C.R. Burn.



Figure 9 Retrogressive thaw slumps initiated by coastal erosion of Pelly Island, western Arctic Canada. Photograph taken in August 1993 by C.R. Burn.

They demonstrated that the annual mean ground temperature beneath the footslope may be higher than in the adjacent tundra due to snow accumulation on the rough footslope in winter. As a result, lake talik configuration adjacent to the slump deepens locally. The deepening increases thawing of permafrost at the foot of the slump and leads to reinitiation of the slump in the foot print of the original feature.

Active-Layer Detachments

In permafrost regions, failures commonly occur on low-angled slopes (McRoberts and Morgenstern, 1974). Characteristically, the failures occur during relatively warm summers or after forest fires (Harris and Lewkowicz, 2000). The failure involves detachment of the active layer from the underlying permafrost so that it may slide downslope (Figure 10).

Such active-layer detachments have also been called 'skin flows' (McRoberts and Morgenstern, 1974) and 'active-layer glides' (Mackay and Mathews, 1973). The detachment accumulates at the base of the slope as an involuted sheet, and may override and fan out there.

The development of active-layer detachments follows thawing of ice-rich ground at the bottom of, and below, the active layer. Water released by thawing excess ice is, by definition, of greater volume than the pore space in mineral soil, so the pore-water pressure increases. This reduces the effective stress at the thaw front, and may lead to failure on even gentle slopes (Harris and Lewkowicz, 2000). Slope stability may be maintained if the excess pore water drains as the thawed soil consolidates, reducing the pore pressure. The critical soil parameter is the thaw consolidation ratio, R (McRoberts and Morgenstern, 1974):

$$R = b/(2\sqrt{c_v}) \quad [6]$$

where c_v is the coefficient of consolidation. The shear strength of the soil may decline over time due to slow deformation at the base of the active layer toward the end of summer, and reduce the excess pore pressures required for detachment (Harris and Lewkowicz, 2000). Active-layer detachments are common in fine-grained soil and diamicton, where the hydraulic conductivity is low and consolidation is slow, but they may occur in other materials if the rate of thaw is sufficiently high. They occur in clusters, both in space and time, due to their association with particular events and sedimentary materials.

Sedimentary Record

The sedimentary sequences in thermokarst basins contain primary deposits laid down during initial subsidence and secondary deposits transported into the basin. The primary deposits may contain disorganized remnant vegetation, particularly wood pieces, that grew before thermokarst initiation, and shells from pond fauna (Burn and Smith, 1990). In the boreal forest, trees either topple into thermokarst lakes or are gradually submerged (Burn and Smith, 1990), and such organic matter is well preserved in the anaerobic lake bottom. Primary deposits of thaw slumps are massive clastic sediments, due to thawing and consolidation of soil in the foot slope, containing dispersed organic accumulations of peat or

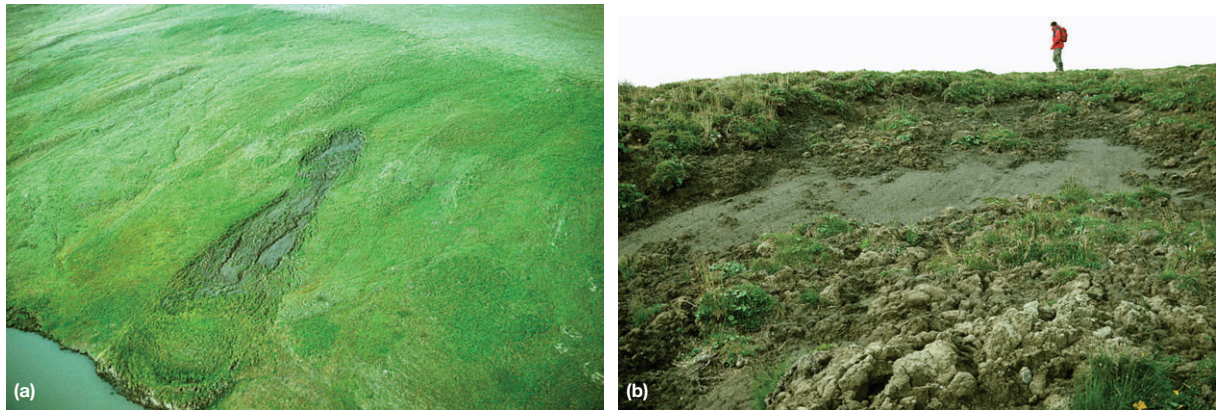


Figure 10 Active-layer detachment failure on Pelly Island, western Arctic coast, Canada, August 1993. (a) Aerial view of failure; (b) failure surface showing slickensides. The slickensided surface of exposed ground indicates that the active layer has moved intact over permafrost. Photographs by C.R. Burn.

woody material, representing debris that fell into the slump. The sedimentary structure left by active-layer detachment is convoluted, as a result of deformation of the active layer at the base of the slope.

Secondary deposits accumulate in thermokarst basins from mass wasting of the surrounding slopes (Murton, 1996). These follow a three-part sequence governed by the evolution of the basin. First, clastic sediments accumulate in the basin from thaw slumping or detachment around the edges of the basin or lake. Second, as ground ice around the basin is exhausted and as slope gradients decline, deposition shifts to sediments entrained around the lake edges and transported to the center in suspension. Third, after drainage, further slow mass wasting into the basin accompanies organic accumulation and eolian deposition. Thermokarst sequences may be observed in the headwalls of retrogressive thaw slumps (Burn and Smith, 1990), and, particularly, in coastal exposures (Murton, 1996).

Thermokarst and Climate Change

Thawing of permafrost is a consequence of climate warming (Jorgenson et al., 2006), and deepening of the active layer increases emissions of methane and carbon dioxide due to increased respiration by soil microfauna (Christensen et al., 2004). The widespread climate warming in Arctic regions since the 1980s has been associated with thermokarst development in some areas (Jorgenson et al., 2006; Payette et al., 2004), and, in the past, periods of climate warming have also been periods of thermokarst lake initiation (Burn and Smith, 1990; Rampton, 1988). In western Arctic Canada, the rate of thaw-slump development around tundra lakes almost doubled during 1973–2004 in comparison with 1950–73 (Lantz and Kokelj, 2008). However, as discussed above, initiation of thermokarst features may occur on a site-specific basis, and may follow surface disturbance, irrespective of climate warming. For example, the lakes illustrated in Figure 1 appear to have been initiated at various times during the last 800 years (Burn and Smith, 1990). Furthermore, once a thermokarst basin is filled with water, the thermal regime of the lake bottom is disconnected from climatic conditions by the properties of the water, and thawing of ice-rich permafrost around the lake may proceed regardless of the climate (Mackay, 1978).

The prospect of thermokarst development accelerating during the twenty-first century has been a key factor stimulating research into permafrost conditions, especially with renewed interest in northern hydrocarbon development. Wholesale eradication of permafrost will take many millennia, but the upper layers of the ground appear to exhibit some response to climate warming already, for North America's longest continuous active-layer record, from the Illisarvik field experiment in western Arctic Canada, shows a gradual increase in annual thaw depth from 1983 to 2008 (Burn and Kokelj, 2009). The long-term stability of infrastructure above permafrost is now widely recognized as problematic for northern development projects. There is also widespread concern that acceleration of thermokarst lake development will lead to significant emission of CH_4 from lake sediments (Walter et al., 2006). While there is no doubt that anaerobic decomposition of organic material in lake-bottom sediments commonly leads to release of methane, the relation between aggregate lake expansion and climate change is not yet clear. This is because numerous factors influence shoreline recession, especially the local ground-ice content, and because thermokarst lakes may also drain if the surrounding permafrost no longer encloses them. However, the key observation at present is probably the accelerating initiation of tundra lakes due to degradation of ice wedges, reported from northern Alaska (Jorgenson et al., 2006), which suggests the methane flux from ice-rich tundra regions is probably increasing at present.

See also: **Ice Core Methods:** Borehole Temperature Records.
Permafrost and Periglacial Features: Active Layer Processes;
 Frost Mounds: Active and Relict Forms; Ice Wedges and Ice-Wedge
 Casts; Permafrost; Slope Deposits and Forms.

References

- Andersland OB and Ladanyi B (1994) *An Introduction to Frozen Ground Engineering*. New York: Chapman and Hall.
- Burn CR (1997) Cryostratigraphy, paleogeography, and climate change during the early Holocene warm interval, western Arctic coast, Canada. *Canadian Journal of Earth Sciences* 34: 912–925.

- Burn CR (1998) The response (1958–1997) of permafrost and near-surface ground temperatures to forest fire, Takhini River valley, southern Yukon Territory. *Canadian Journal of Earth Sciences* 35: 184–199.
- Burn CR (2000) The thermal regime of a retrogressive thaw slump near Mayo, Yukon Territory. *Canadian Journal of Earth Sciences* 37: 967–981.
- Burn CR (2002) Tundra lakes and permafrost, Richards Island, western Arctic coast, Canada. *Canadian Journal of Earth Sciences* 39: 1281–1298.
- Burn CR and Kokelj SV (2009) The environment and permafrost of the Mackenzie Delta area. *Permafrost and Periglacial Processes* 20: 83–105. <http://dx.doi.org/10.1002/ppp.655>.
- Burn CR and Lewkowicz AG (1990) Retrogressive thaw slumps. *The Canadian Geographer* 34: 273–276.
- Burn CR and Smith MW (1990) Development of thermokarst lakes during the Holocene at sites near Mayo, Yukon Territory. *Permafrost and Periglacial Processes* 1: 161–176.
- Christensen TR, Johansson T, Åkerman HJ, et al. (2004) Thawing sub-arctic permafrost: Effects on vegetation and methane emissions. *Geophysical Research Letters* 31: L04501. <http://dx.doi.org/10.1029/2003GL018680>.
- Crawford CB and Johnston GH (1971) Construction on permafrost. *Canadian Geotechnical Journal* 8: 236–251.
- French HM (1975) Man-induced thermokarst, Sachs Harbour airstrip, Banks Island, Northwest Territories. *Canadian Journal of Earth Sciences* 12: 132–144.
- French HM (2007) *The Periglacial Environment*, 3rd edn., pp. 109–126. Chichester: John Wiley & Sons.
- Harris C and Lewkowicz AG (2000) An analysis of the stability of thawing slopes, Ellesmere Island, Nunavut, Canada. *Canadian Geotechnical Journal* 37: 449–462.
- Hinkel KM, Eisner WR, Bockheim JG, Nelson FE, Peyerson KM, and Dai X (2003) Spatial extent, age, and carbon stocks in drained thaw lake basins on the Barrow Peninsula, Alaska. *Arctic, Antarctic, and Alpine Research* 35: 291–3000.
- Jorgenson MT, Shur YL, and Pullman ER (2006) Abrupt increase in permafrost degradation in arctic Alaska. *Geophysical Research Letters* 33: L04501. <http://dx.doi.org/10.1029/2005GL024960>.
- Kokelj SV, Lantz TC, Kanigan J, Smith SL, and Coutts R (2009) Origin and polycyclic behaviour of tundra thaw slumps, Mackenzie Delta region, Northwest Territories, Canada. *Permafrost and Periglacial Processes* 20: 173–184. <http://dx.doi.org/10.1002/ppp.642>.
- Lantz TC and Kokelj SV (2008) Increasing rates of retrogressive thaw slump activity in the Mackenzie Delta region, N.W.T., Canada. *Geophysical Research Letters* 35: L06502. <http://dx.doi.org/10.1029/2007GL032433>.
- Mackay JR (1963) *The Mackenzie Delta Area, N.W.T. Geographical Branch Memoir 8*. Ottawa: Department of Mines and Technical Surveys.
- Mackay JR (1970) Disturbances to the tundra and forest tundra environment of the western Arctic. *Canadian Geotechnical Journal* 7: 420–432.
- Mackay JR (1978) Freshwater shelled invertebrate indicators of paleoclimate in northwestern Canada during late glacial times: Discussion. *Canadian Journal of Earth Sciences* 15: 461–462.
- Mackay JR (1983) Downward water movement into frozen ground, western arctic coast, Canada. *Canadian Journal of Earth Sciences* 20: 120–134.
- Mackay JR (1995) Active layer changes (1968–1993) following the forest-tundra fire near Inuvik, N.W.T., Canada. *Arctic and Alpine Research* 27: 323–336.
- Mackay JR (1997) A full-scale field experiment (1978–1995) on the growth of permafrost by means of lake drainage, western Arctic coast: A discussion of the method and some results. *Canadian Journal of Earth Sciences* 34: 17–33.
- Mackay JR and Mathews WH (1973) Geomorphology and Quaternary history of the Mackenzie River Valley near Fort Good Hope, N.W.T., Canada. *Canadian Journal of Earth Sciences* 10: 26–41.
- McRoberts EC and Morgenstern NR (1974) The stability of thawing slopes. *Canadian Geotechnical Journal* 11: 447–469.
- Murton JB (1996) Thermokarst-lake-basin sediments, Tuktoyaktuk Coastlands, western Arctic Canada. *Sedimentology* 43: 737–760.
- Payette S, Delwaide A, Caccianiga M, and Beauchemin M (2004) Accelerated thawing of subarctic peatland permafrost over the last 50 years. *Geophysical Research Letters* 31: L04501. <http://dx.doi.org/10.1029/2004GL020358>.
- Rampton VN (1988) Quaternary geology of the Tuktoyaktuk coastlands, Northwest Territories. *Geological Survey of Canada Memoir* 423.
- Shur Y, Hinkel KM, and Nelson FE (2005) The transient layer: Implications for geocryology and climate-change science. *Permafrost and Periglacial Processes* 16: 5–17.
- Walter KM, Zimov SA, Chanton JP, Verbyla D, and Chapin FS (2006) Methane bubbling from Siberian thaw lakes as a positive feedback to climate warming. *Nature* 443: 71–75.
- Williams PJ and Smith MW (1989) *The Frozen Earth: Fundamentals of Geocryology*. Cambridge: Cambridge University Press.
- Zhang T, Heginbottom JA, Barry RG, and Brown J (2000) Further statistics on the distribution of permafrost and ground ice in the northern hemisphere. *Polar Geography* 24: 126–131.
- Zoltai SC (1993) Cyclic development of permafrost in the peatlands of northwestern Alberta, Canada. *Arctic and Alpine Research* 25: 240–246.

Phytoliths

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What Are Phytoliths?

Phytoliths (from Greek *φυτό* *phyto*, 'plant,' and *λίθος* *lithon*, 'stone') are a class of microfossils commonly produced in higher plants (Pearsall and Piperno, 1993; Piperno, 2006). They are composed of the noncrystalline (amorphous) form of hydrogenated silica called opal and are sometimes referred to as 'silica bodies,' 'plant opal,' or 'opaline silica.' Calcium oxalate crystals and starch grains produced by some plants may also be considered phytoliths in the broad sense (Hart and Wallis, 2003) but are not discussed here. Opal phytoliths are commensurate in size with typical plant cells and range from 5 to about 200 μm . They are therefore most commonly found in the silt fraction of sediments (Figure 1). Phytoliths are a valuable proxy for reconstructing past plant communities, especially graminoid-dominated types such as grasslands, semideserts, and tundras. Although phytoliths' taxonomic resolution is lower than that of macrofossils (i.e., rarely to the species level), they may provide subfamily or genus-level information for some families (e.g., Pteridaceae (maidenhair ferns), Pinaceae (pines), Poaceae (grasses), Cyperaceae (sedges), and Asteraceae (asters, daisies, sunflowers)). They also accumulate more locally than pollen and are remarkably durable in dry, acidic, or aerobic conditions (e.g., in paleosols, dung, or cave deposits), where pollen is scarcely preserved (Blinnikov et al., 2002; Piperno, 2006; Strömberg, 2005).

Phytolith Formation

Silicon is the second most common element in the Earth's crust. Phytoliths are formed inside plant tissues from soluble silica carried up by plant vessels from groundwater in the process of transpiration (Raven, 1983). The weathering of silicate minerals such as quartz at the Earth's surface provides large amounts of labile silica, some of which is absorbed by growing plants (Saccone et al., 2007). In solution, silica may exist as monosilicic acid ($\text{Si}(\text{OH}_4)$) when pH values range between 2 and 9. It is carried upward in the vascular system and becomes concentrated because of transpiration around the leaf stomata and in epidermis in general. The supersaturated solution begins to polymerize or gel and then solidifies and forms solid opaline silica ($\text{SiO}_2 \times n\text{H}_2\text{O}$) bodies (phytoliths) within and between some of the plant cells. Although in many plants greater amount of transpiration leads to higher levels of silicification,

the exact relationship between transpiration and deposition of phytoliths remains poorly understood (Piperno, 2006).

The first documented phytoliths (Struve, 1835) were those from grasses, and for many years, these were the most extensively studied (Figure 2; Mulholland, 1989; Smithson, 1958; Twiss et al., 1969). Phytoliths are also described from pteridophytes (ferns), gymnosperms (Klein and Geis, 1978), and selected families of angiosperms (Geis, 1973; Morris et al., 2009), throughout the world. Nevertheless, it should be noted that not all plants produce phytoliths. Piperno (2006) lists about 50 families of higher plants as confirmed phytolith producers as compared to 78 families in which phytoliths have not been observed, despite extensive studies. It is important to note that some of the latter families do accumulate some silica but no observable phytolith morphotypes.

Phytoliths are found in nearly all plant structures: stems, leaves, roots, and inflorescences. In general, many more phytoliths form in the aerial tissue than subterranean tissue probably because of the role played by aboveground evapotranspiration in precipitating silica from solution near the epidermal surface (Figure 3). Patterns of phytolith formation appear to be consistent within families and species from continent to continent (Piperno, 2006); however, only a fraction of the global flora has been investigated for phytolith presence. The formation of phytoliths often involves the filling of the entire cell interior. Cell interiors act as molds, producing casts of solid phytoliths of different repeatable three-dimensional (3D) shapes, more or less corresponding to the cell wall outline. In contrast, the most studied diagnostic phytoliths of grasses and sedges are actually produced inside the cells under an active enzyme control (Blackman, 1969; Sangster et al., 2001) and do not fill the entire cell as in a mold. Such solid phytoliths are denser, more resistant to dissolution, and more durable than those produced by the molding/infilling process. For example, Fraysse et al. (2009) found that an average phytolith of a horsetail or larch has only a 6-year half-life in a soil solution (10–12 years at $\text{pH}=2-3$ to <1 year at $\text{pH}>6$). Grass phytoliths, in contrast, may have half-lives of many hundreds or even thousands of years, especially in dry conditions.

Phytoliths from woody plant species (trees and shrubs) have been much less studied. One reason for this is poor preservation of the most common tree phytoliths (polyhedral and anticlinal epidermal) produced in the cell walls of leaves of these plants. These phytoliths are poorly preserved because of their fragile flat structure with long edges and high surface-to-volume ratios.

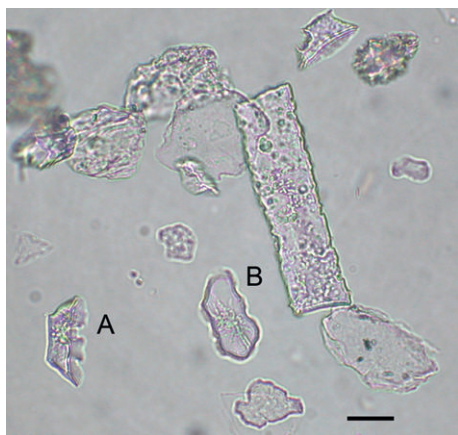


Figure 1 Mainly grass phytoliths from late MIS 3 stage sediments (ca. 29 000 years BP) from central Alaska. Lobate trapezoids, cf. *Poa*, are visible from two different angles in A and B. Scale bar is approximately 10 μm .

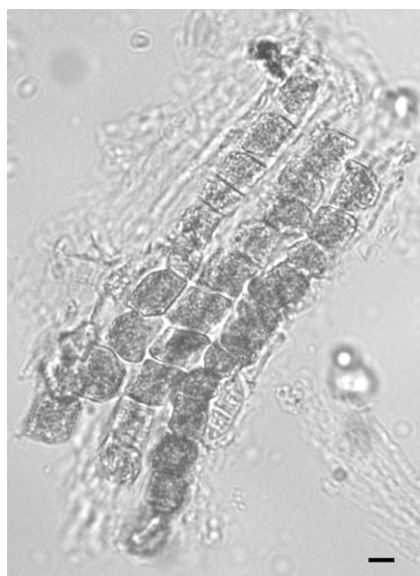


Figure 2 Polygonal phytoliths from *Quercus macrocarpa* leaf. Scale bar is approximately 10 μm .

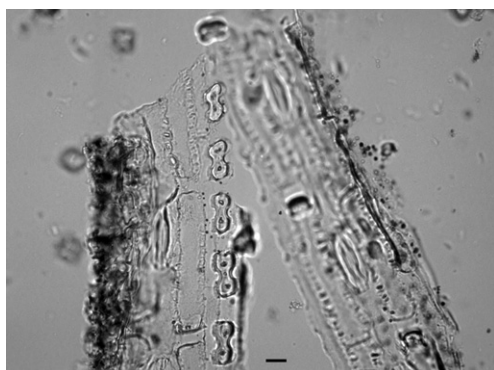


Figure 3 Bilobate phytoliths from the leaf epidermis of *Panicum virgatum*, a typical C_4 panicoid grass from the tall grass prairie in the American Midwest. Scale bar is approximately 10 μm .

However, spherical tree phytoliths, produced in large numbers in the center of groups of silicified epidermal cells of trees and shrubs, tend to be better preserved (Bozarth, 1992). These are especially common in the tropical plants and are in fact used in indices diagnostic of specific tropical biomes (Bremond et al., 2005, 2008a,b). Modern collections of tropical phytoliths do exist, but there are fewer well-known diagnostic forms as compared to graminoids (Kealhofer and Piperno, 1998; Runge, 1999).

Phytolith Composition, Characteristics, and Function

Phytoliths are composed of amorphous (noncrystalline) silicon dioxide (SiO_2) and about 4–9% water with admixture of other chemical elements. Measurable amounts of microcrystalline structures have been reported within phytoliths, creating the possibility of direct dating by optically stimulated luminescence (Rowlett and Pearsall, 1993). Phytoliths are optically isotropic, with refractive indices between 1.41 and 1.47 and a specific gravity between 1.5 and 2.3. Under transmitted light, they range from colorless to slightly brown or pink to opaque. The coloration may be related to the presence of iron and other elements as well as to the residual organics and/or exposure to fire (charred carbon). Phytoliths frequently contain significant amounts of occluded, chemisorbed, or solid elements such as aluminum, iron, titanium, manganese, phosphorus, copper, nitrogen, and carbon. It is thought that the presence of carbon within phytoliths is the result of the trapping of plant cellular material during phytolith formation. The presence of carbon allows directed ^{14}C dating of phytoliths (see the Section 'Phytoliths and Chronology'). It has been demonstrated that more than 50% of this encapsulated carbon is protected from oxidation and thus provides another source of paleoenvironmental and paleoclimatic information (Kelly et al., 1991).

Moreover, because phytolith-occluded carbon is of plant cellular origin, it is theoretically possible that deoxyribonucleic acid (DNA) is present and could provide direct evidence of floral origin. Recent attempts at isolating DNA from phytoliths have unfortunately been unsuccessful (Elbaum et al., 2009).

The functional significance of silica in plants is not obvious. In some families, the mechanism of phytolith formation appears passive, and so they represent no more than 'waste products' (Sangster et al., 2001). Alternatively, phytoliths provide mechanical support, preventing the collapse of cell walls under tension during transpiration (Raven, 1983). Moreover, phytoliths provide plants with increased resistance to browsing by herbivorous mammals and insects and attack by pathogenic fungi (Massey et al., 2007; McNaughton et al., 1985).

An interesting observation is that plant families that use organic materials (e.g., cork, lignin) for internal structural support do not accumulate as much silica as families that rely primarily on inorganic material such as silica. Silica is heavy but energetically cheap to absorb from soil and redeposit. In contrast, organic structures must be manufactured by plants internally at a considerable energy cost, but they are more lightweight and versatile. Thus, there must be an evolutionary trade-off between silica 'absorbers' versus silica 'refusers.' In spite of our incomplete understanding as to why some plants prefer silica and some do not, it does appear that silicon is an

essential element in the development of many species (Piperno, 2006; Raven, 1983).

Phytolith Morphology

Phytolith analysis is based on the premise that shapes of phytoliths correspond to the taxonomy of plants in which they formed. In this sense, phytoliths are just another class of microfossils, like pollen. What is different about them, however, is the imprecise correspondence between the shapes of phytoliths and their taxonomic position. Each species produces only one kind of pollen. This is not the case with phytoliths. Each plant produces many different phytoliths, depending on the tissue (multiplicity of shapes). In a thorough study of phytolith production in wild rice (*Zizania palustris*), Yost and Blinnikov (2011) found 23 different morphotypes in *Zizania* that were locally diagnostic of the species. As a rule, only a few morphotypes are abundant within the plant, and in practice, many analysts choose to lump some morphotypes together for faster counts. Still, the sheer diversity of morphotypes, even within a single species, represents the first challenge of the method because there are few comprehensive studies of phytolith production at the species or genus level (Twiss, 2001).

The second, and more serious, challenge is that of redundancy. To understand redundancy, consider pollen: all oaks produce *Quercus*-type pollen diagnostic of the genus, and no other genera produce this type of pollen. Thus, when *Quercus*-type pollen grains are encountered in sediments, they suggest presence of oak. With phytoliths, identical morphotypes can be found in many taxa. For example, all grasses produce flat rectangular epidermal cell phytoliths ('short plates'), which are common across all subfamilies and are thus redundant. Such phytoliths do suggest Poaceae family in a sediment record but provide no information about what subfamily or genus of grasses they may have come from. These two problems (multiplicity and redundancy) make phytolith studies considerably more challenging than pollen analysis and explain why the method has been lagging behind palynology, although phytoliths were first described by botanists around the same time as pollen (Struve, 1835).

Representativeness and Preservation

Phytolith analysis is a rapidly evolving discipline, and there are many basic identification problems to be resolved. Again, these problems are numerous, such as the many different phytoliths that are found in a single taxon, and redundancy, when the same form can occur in many taxa. Yet another problem is associated with lack of representation of some phytolith forms in assemblages extracted from sediments (differential preservation). As mentioned above, some phytoliths produced by plants do not persist for long and are poorly represented in sediments. For example, in a study of sedge and grass phytolith production in Minnesota wetlands, Yost and Blinnikov (2011) discovered that sedges were greatly underrepresented relative to grasses in modern lake sediments. Even in cases where sedges represented over half of all vegetation cover around the lake, more than 90% of identifiable phytoliths in

sediments were those of grasses, and very few sedge phytoliths were encountered in modern samples. Sedges do produce phytoliths, but they are apparently much more fragile forms, prone to rapid dissolution. Cabanes et al. (2011), in a controlled experiment in Israel, proved that modern inflorescence wheat phytolith morphotypes (which are long and fragile dendritic forms) were less stable than those from leaves or stems. Furthermore, burned phytoliths were less stable than unburned, and a fossil phytolith assemblage ca. 3000 years old was more stable than the modern wheat assemblage, due to the removal of the fragile forms over time. The study also showed that individual morphotypes have different stabilities, and as a result of dissolution and abrasion, some morphotypes may resemble others over time. Fishkis et al. (2010) set up an experiment to test downward transport of phytoliths in soil profiles and found that in some soils (Cambisols), addition of water significantly increased vertical movement of phytoliths from <1 to >2 cm. The addition of earthworms, on the other hand, did not substantially increase vertical transport. Clearly, more studies are needed to assess the degree of vertical translocation in soils.

With respect to horizontal translocation, Blinnikov et al. (2011) demonstrated that phytolith assemblages, under control mixtures on 9 × 9-m plots, do in fact reflect specific phytolith producers on these plots after 8 years of experiment. However, a small proportion of phytoliths (<5%) could not be attributed to the producers on the plots themselves and in all likelihood came from neighboring plots a few meters away. Fredlund and Tieszen (1997) found that close to half of all phytoliths in native prairies may be translocated from a few to a few hundred meters away, due to fire, wind, or animal movements, on the scale of hundreds of years.

Despite these problems, many reference collections of modern phytoliths and soil assemblages are being compiled within specific regions. One such example is the reference collection produced by Wallis (2003) who documented the occurrence of distinctive phytoliths from the leaves of 177 non-Poaceae plant species from northwestern Australia. The study showed that although only 50% of the specimens examined were phytolith producers, there was enough morphological variability to distinguish between fossil phytolith assemblages from the region. Blinnikov (2005) documented phytoliths in more than 30 species of most common grasses, forbs, and trees in Oregon and Washington states. Morris et al. (2009) produced a reference collection of grass and nongrass phytoliths for the Great Basin in the United States. Other locations that have seen much research on regional phytolith production have been the Neotropics (Piperno, 1988), Argentina (Fernández Honaine et al., 2009; Gallego and Distel, 2004), South Africa (Cordova and Scott, 2010), west and central Africa (Bremond et al., 2005; Eichhorn et al., 2010; Runge, 1999), the Middle East (Ball et al., 2007), the Russian Caucasus (Blinnikov, 1994), southeast Asia (Kealhofer and Piperno, 1998), China (Lu et al., 2005), and India (Kajale and Eksambekar, 2007).

Terminology and Classification

Historically, the terminology and classification of phytoliths have depended on the researcher. The same basic shape could be variously called a 'hat,' 'conical,' 't-phytolith patelliform,' or

'rondel,' for example. While some researchers only used the geometry of the shape in its name, others gave phytoliths names linked to their presumed anatomical source (e.g., tracheid, epidermal cell, trichome). To make things even more complicated, only fairly recently did researchers begin to use true 3D shapes in phytolith analysis, as seen under rotation in a liquid medium (Figure 4). Much of the research done in the 1960s through 1980s relied primarily on fixed media for phytolith counts, making some shapes impossible to identify correctly. For example, bilobate and some trapezoidal lobate forms may appear identical in certain orientations but completely different in others (Rapp and Mulholland, 1992).

The main reason for this overall confusion is the genuine problem of continuity of shapes exhibited by phytoliths and the lack of exact 'types,' unlike in pollen or diatom analysis. A number of independent schemes were devised which were used to deal with phytoliths extracted from plants or sediments, generally from a specific region. There was a certain amount of overlap between these schemes, and over the years, a number of attempts have been made to develop a universally accepted standard classification scheme (Bowdery et al., 2001; Pearsall, 2000). In order to establish standards and to harmonize the naming and description of phytoliths, a working group has been formed that has developed a standard protocol for naming phytolith types and a glossary of descriptors: *'The first International Code for Phytolith Nomenclature 1.0'* (Madella et al., 2005). It is suggested that when naming a particular phytolith type, there should be up to three descriptors. The first descriptor should be for the 2D or 3D shape. The second descriptor should be texture and/or ornamentation and should be added if they are characteristic or diagnostic. A third descriptor can be added when the anatomical origin is clear and beyond doubt. For example, bulliform is an established botanical term used to describe a particular type of cell found in grass leaf epidermis. In this particular case, the word conveys both an anatomical and a descriptive meaning. It is expected that as the scheme is adopted, it would undergo a number of changes, as more phytoliths are discovered and described (Table 1).

An alternative solution would be to abolish typological approach entirely and rely on some image recognition protocol that could automate phytolith shape and size counts for certain projects where only few basic forms are observed (Ball et al., 1996). Computer image recognition software may assist in quickly estimating proportion of certain shapes and sizes on the cover slip and then assign probability of each sample belonging to a particular standard from a specific plant community. However, most analysts work with true 3D shapes of phytoliths and rotate grains under the microscope, while typical image recognition software can only work in 2D, so much more research in this area is necessary.

Phytolith Analytical Techniques

Phytoliths are composed of opal, that is, hydrated non-crystalline silica, a material extremely resistant to acids and high temperatures (up to 1000 °C), but soluble in very alkaline solutions and mildly soluble in water if immersed for long intervals. Opal is significantly less dense than quartz (1.5–2.3 vs. 2.6 specific gravity, respectively). Therefore, extraction of

phytoliths from plants and sediments relies on either heat (ashing in a muffle furnace) or strong acidic digestions (e.g., with hot concentrated nitric acid) to remove organic matter. Additional treatments include mild hydrochloric acid to remove carbonates and flotation in heavy liquid (e.g., solutions of zinc iodide, sodium polytungstate, or toluene, usually at 2.3 specific gravity) to separate phytoliths (<2.3) from heavier minerals (e.g., quartz is about 2.6). The techniques of phytolith extraction are basically the same as those of diatom extraction. One of the challenges of working with aquatic sediments is precisely in separating phytoliths from diatoms, with the latter commonly outnumbering similarly composed phytoliths by 10:1. Although both are made from the same biogenic silica, the fine difference in shape and density, and therefore buoyancy, may allow separation of most common short cell phytoliths of grasses (compact and dense) from most common diatoms (usually more elongated and less dense; Yost and Blinnikov, 2011).

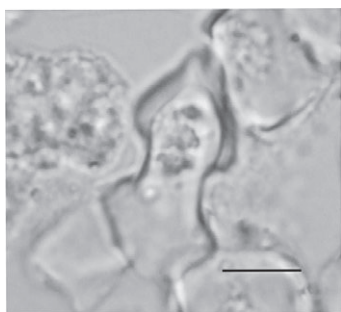
For a full range of suitable phytolith field techniques, research design, laboratory analytical methods, and techniques, as well as interpretation methods and theory, refer to Piperno (1988, 2006) and Pearsall (2000). Parr (2002) described an alternative microwave digestion technique for extracting phytoliths from sediments. The technique has obvious advantages in reducing the preparation time for phytolith analysis, as well as reducing the chance of contamination, by reducing the number of steps involved. For example, the traditional technique of acid digestion requires 6–8 h. Microwave digestion may take only an hour.

Advances in scanning electron microscopy (SEM) allow very accurate images of phytoliths to be obtained. However, most phytolith analysts prefer to rely on traditional optical microscopes for counting shapes, for faster and cheaper processing. Additionally, standard SEM images are static and do not allow for easy 3D rotation. Nevertheless, with the advent of digital photography, even optical microscopy can now provide reliable high-quality images, including both stills and videos. For fine analysis of surface sculpturing and other close-up details, SEM is the tool of choice. However, few of the details visible in SEM images appear to be diagnostic.

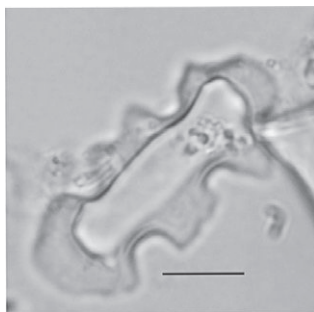
Phytoliths and Chronology

Radiocarbon

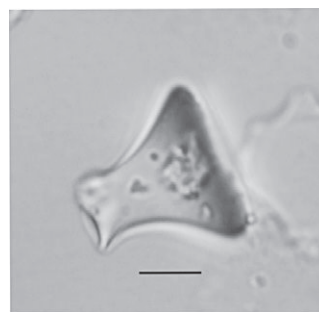
An important aspect of late Quaternary research has been the ability to reliably date changes in the relevant sequences and processes. In addition to providing vegetation information, phytoliths contain datable ¹⁴C. Moreover, because phytolith-occluded carbon is effectively sealed against postdeposition contamination at the time of the phytolith formation, they provide an ideal time capsule for use in radiocarbon dating. Wilding (1967) was the first to determine a radiocarbon age from phytolith-occluded carbon. However, because of the large samples (45 kg of sediment) needed prior to the advent of accelerator mass spectrometry (AMS) dating, it was not practical to replicate his study. Following the development of the AMS dating method (requiring far smaller samples) in the late 1970s, more radiocarbon dates have been obtained from phytoliths (Mulholland and Prior, 1993). As with most new techniques, issues have been raised with regard to preparation, accuracy, and reliability, and more research developments are expected.



Bilobate short cell
(grass origin)



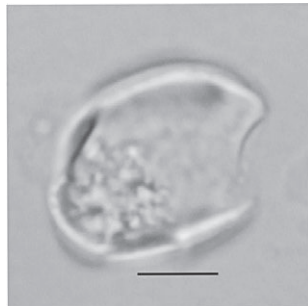
Trapeziform polylobate
(grass origin)



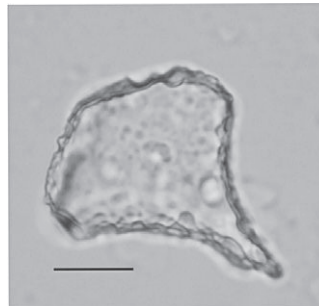
Conical short cell
(grass origin)



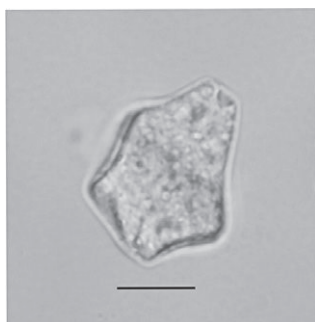
Elongate long cell
(grass origin)



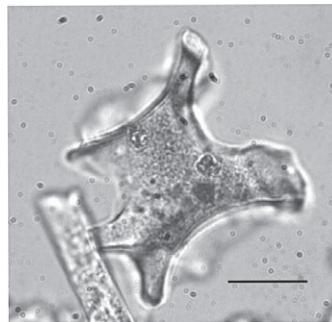
Saddle short cell
(grass origin)



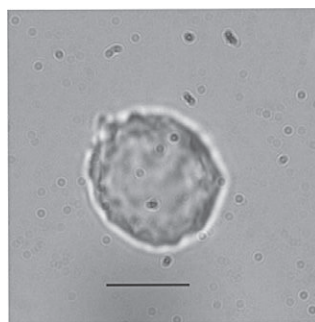
Cuneiform bulliform cell
(grass origin)



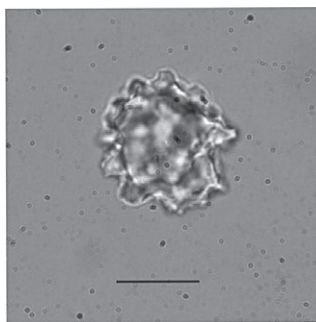
Polyhedral
(tree/shrub origin)



Anticlinal
(fern origin)



Globular granular
(tree/shrub origin)



Globular echinate
(palm origin)

Figure 4 Various morphotypes extracted from aeolian sediments. Scale bar is 10 μm .

Table 1 Naming according to International Code for Phytolith Nomenclature (ICPN) Working Group

Schematic drawings ^a	ICPN names	Former nicknames
 or  ...etc	Bilobate short cell	Dumbbell or bilobate
	Trapeziform short cell	Square or rectangle
	Cylindrical polylobate	Polylobate
	Trapeziform polylobate	Polylobate
	Trapeziform sinuate	
	Elongate echinate long cell	Elongate spiny or elongate sinuous
	Cuneiform bulliform cell	Bulliform or fan shaped
	Parallelepipedal bulliform cell	Bulliform
	Acicular hair cell	Point shaped
	Unciform hair cell	Point shaped
	Globular granulate	Spherical rugose
	Globular echinate	Spherical crenate
	Cylindric sulcate tracheid	Tracheid

Source: Madella M, Alexandre A, and Ball T (2005) International code for phytolith nomenclature 1.0. *Annals of Botany* 96: 253–260, with permission from Oxford University Press.

^aSeveral drawings are made after Fredlund and Tieszen (1994).

Optically Stimulated Luminescence

In spite of being generally described as amorphous, phytoliths appear to contain microcrystal structures. Theoretically, these microcrystal structures should be able to trap electrons and, as such, offer potential material for thermoluminescence (TL) dating. Rowlett and Pearsall (1993) examined the feasibility of TL analysis of phytoliths. They dated a number of phytoliths from sediments associated with ceramics of known cultural affiliation and radiocarbon-dated organic material. However, although their TL dates were in general agreement with the previously

accepted ages, indicating the potential of this method, the phytolith matrix was shown to be unstable at high temperatures.

Phytolith Chemical Analysis

In recent years, previously unstudied chemical components of phytoliths and their occlusions have been analyzed and documented, following the development of new instruments and techniques. Instruments such as the inductively coupled plasma source mass spectroscope, capable of measuring elements in parts per billion, have opened up the possibility of measuring trace elements present in minute quantities in phytoliths. Hart (2001) used this technique to examine trace elements present within phytoliths from the leaves of *Actinotus helianthi* and *Triodia mitchellii*. She was able to confirm the presence of more than double the previously documented number of phytolith-occluded trace elements. Smith and Anderson (2001) used tetramethylammonium hydroxide (TMAH) thermochemolysis and gas chromatography–mass spectrometry (GC–MS) to characterize the organic compounds in C₃ and C₄ phytoliths.

In an attempt to improve on phytolith taxonomic resolution, Carnelli et al. (2002) have proposed a new method of distinguishing between similar morphological forms from different taxa. Using X-ray analysis paired with phytolith morphology, they have been able to demonstrate that aluminum in phytoliths from Ericaceae (heaths) and conifers was found in significantly higher concentrations than in those from Cyperaceae and Poaceae. The technique will require further investigations of different taxa to confirm that the codeposition of aluminum and silicon in phytoliths is more frequent in woody plants than herbaceous plants.

Stable Isotope Analysis

It has been demonstrated that the stable isotope analysis of phytoliths can contribute considerable paleoenvironmental and paleoclimate evidence (Kelly et al., 1991; Smith and Anderson, 2001; Smith and White, 2004; Webb and Longstaffe, 2003). Tests carried out by Kelly et al. (1991) confirmed that the carbon occluded within phytoliths is protected by the nonporous glasslike microstructure of the silica body and is therefore unlikely to be vulnerable to decay.

Information about climatic conditions during plant growth may be inferred from the $\delta^{18}\text{O}$ signature preserved in oxygen isotopes from phytoliths, as discussed by Webb and Longstaffe (2003). Their hypothesis was that climate factors which affect the isotopic composition of oxygen of water (H₂O) within growing plants should leave a record in the oxygen isotopic ratios of phytolith silica. They were able to determine that the plant growing temperature is recorded in the $\delta^{18}\text{O}$ of phytoliths formed within the nontranspiring tissue of grasses. However, the signal from nontranspiring tissue can be masked by phytoliths from transpiring tissue, which in contrast are variably enriched in $\delta^{18}\text{O}$, depending on relative humidity.

Phytoliths trap trace amounts of organic carbon (0.09–1.3 wt%; Kelly et al., 1991). Organic compounds can be attached to the silicates by hydrogen bonding, or alternatively parts of the cytoplasm may become occluded within the phytolith as they silicify (Fredlund, 1993). Wilding's (1967) isolation and radiocarbon analysis of phytolith carbon created the

opportunity to extract more paleoenvironmental information. Fredlund (1993) demonstrated that careful analysis of phytolith-occluded carbon had the potential to reconstruct record of the range, variability, and changes of C_3 (around 27‰ $\delta^{13}C$) and C_4 grasses (around 12‰ $\delta^{13}C$) of past grassland ecosystems. However, Smith and Anderson (2001) showed that the method was limited by the compression of the phytolith $\delta^{13}C$ scale (6.6‰) relative to the whole plant scale (11.2‰). Their study demonstrated that it was possible to characterize the phytolith organic compounds in C_3 and C_4 grass phytoliths using TMAH and GC-MS analysis. These analyses record the presence of lipids, which are generally depleted in $\delta^{13}C$, and suggested, therefore, that lipid presence explains the depletion in $\delta^{13}C$ in phytolith carbon, relative to whole plant $\delta^{13}C$. Moreover, they suggested that the isotopic differences in C_3 and C_4 grass phytoliths were most likely caused by differential fractionation associated with the formation of lipids in C_3 and C_4 plants. Smith and White (2004) have gone on to resolve some modern calibration issues raised in previous work and show that the carbon isotope ratios of fossil phytoliths present a unique tool for reconstructing the proportion of C_3 and C_4 grasses in ancient grasslands.

The occluded carbon within phytoliths is formed by photosynthesis from atmospheric CO_2 when the plant was growing. Therefore, it should be possible to relate the carbon isotope analysis of phytolith-occluded carbon with carbon isotope concentration in atmospheric CO_2 . Carter (2009) has extracted a last glacial cycle record from phytoliths extracted from a 120 000-year-old loess sediment core. Phytoliths were extracted from the 7.4-m core, and the occluded carbon was analyzed morphologically and isotopically. Rates of isotopic fractionation between plant and phytolith were determined by measurements from a number of modern tree, fern, and grass species. There was little correlation between the $\delta^{13}C$ record and the phytolith morphological assemblages, suggesting that the $\delta^{13}C$ signal from occluded carbon was largely determined by the atmospheric CO_2 carbon isotopic ratio.

Although stable isotope studies of phytoliths show promise, Webb and Longstaffe (2010) have not found a significant correlation ($r < 0.3$) between climate variables and the $\delta^{13}C$ values of *Calamovilfa longifolia* tissues (average $\delta^{13}C = -13.1 \pm 0.6$ ‰; $n = 70$) in a robust study across the North American prairies. The strongest correlations were obtained between the $\delta^{13}C$ values of stem or sheath phytoliths and humidity ($r = 0.3$), latitude ($r = 0.4$), and amount of precipitation ($r = 0.5$). However, use of these relationships was limited by the wide spread in $\delta^{13}C$ values of phytoliths from different plant tissues at the same location.

Paleoenvironmental Reconstruction Using Phytolith Analysis

Some of the earliest known phytoliths come from the time of dinosaurs in the late Cretaceous, dated about 80 million years BP (Prasad et al., 2005). Already at that time, shapes attributable to modern subfamilies of grasses could be detected. Phytolith research in the Miocene (Strömberg, 2005) demonstrated the high utility of the method when applied to times of rapid grass evolution, coupled with strong environmental change. In Pleistocene records, phytoliths allow analysis of

long records, particularly from loess (Blinnikov et al., 2002; Lu et al., 2005) that span over 100 000 years and potentially the entire 2.6 million years of the epoch. Analysis of paleosols, lake sediments, cave deposits, and animal middens allows the application of phytoliths to focus on the most dramatic times of climate change, such as the Last Glacial Maximum and Late Glacial to interglacial transition, as well as short-term fluctuations such as the Younger Dryas oscillation and events in the mid-Holocene.

Because of the lack of clear ‘types’ and high redundancy of many forms, it is best to combine phytolith analysis with other proxy studies, such as macrofossils, pollen, chironomids, and paleogenetic methods, done on the same sediments using a multiproxy approach (Blinnikov et al., 2002; Cordova et al., 2010; Wooller et al., 2011).

There are a few common approaches to interpreting phytolith paleorecords. They include (a) detection of specific, highly diagnostic forms in order to infer presence or absence of a specific taxon (Yost and Blinnikov, 2011); (b) matching modern and paleoassemblages to interpret paleovegetation (Blinnikov et al., 2002); (c) matching modern and paleoassemblages to infer paleoclimate via a transfer function; or (d) looking for changes in ratios of different plant functional groups (e.g., trees vs. grass) as interpreted on the basis of an index of different morphotypes (Bremond et al., 2008a,b).

An example of the direct diagnostic match is that of Yost (2007), who studied wild rice in Minnesota lakes. He was able to demonstrate the existence of wild rice in Lake Ogechie as early as 3500 years BP, based on highly diagnostic rondels from the wild rice inflorescence, considerably earlier than previously inferred from pollen (2500 years BP).

Bremond et al. (2005, 2008a,b) proposed and tested a few indices that allowed interpretation of paleovegetation in Africa, based on ratios of nonarboreal and arboreal forms. Fredlund and Tieszen (1994, 1997) developed a robust modern dataset that allows direct matching of any paleosols from the Great Plains in North America to their modern analogs, both in terms of climate and vegetation type. Cordova et al. (2010) and Morris et al. (2009) tested the applicability of a burned phytolith index, a ratio of charred versus clean phytoliths, as a proxy for paleofires that can complement charcoal records.

Many paleoenvironmental reconstructions involving phytolith analysis have used qualitative or quantitative methods and a combination of the above-listed approaches (Powers-Jones and Padmore, 1993). For example, Carter and Lian (2000) used the morphological differences in phytolith assemblages extracted from a well-dated 6.4-m loess sediment core to infer changes in the relative abundance of forest versus grass cover. The record showed correlations between the tree-shrub phytolith fluctuations and the oxygen isotope curve between marine oxygen isotope stages (MISs) 1 and 5, suggesting that changes in the ratio of arboreal to nonarboreal phytoliths directly result from changes in climate (Figure 5).

Since 2000, more rigorous statistical and image analysis methods have been employed to provide better statistical procedures, dating techniques, and analytical methods. For example, in order to establish a more quantifiable relationship between vegetation and phytolith assemblages, various quantitative statistical methods have been applied, such as

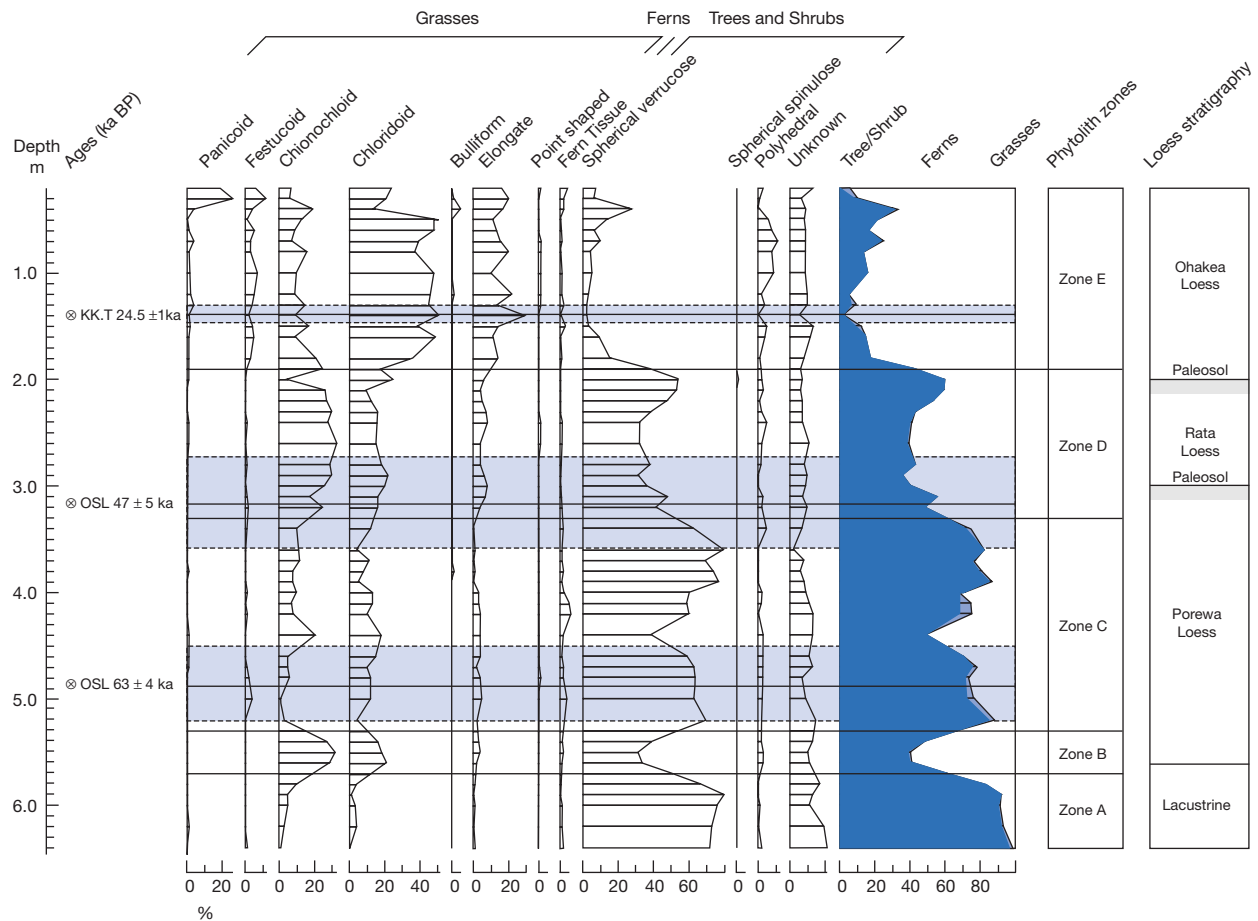


Figure 5 Example of a phytolith stratigraphic frequency diagram showing vegetation changes through time from southeastern North Island, New Zealand. ka, thousand years before present. Reprinted from [Carter JA and Lian OB \(2000\)](#). Paleoenvironmental reconstruction from the last interglacial using phytolith analysis, south eastern North Island, New Zealand. *Journal of Quaternary Science* 15: 733–743, with permission.

constrained classification techniques and cluster analysis and correspondence analysis (see review in [Piperno, 2006](#)). [Bobrov et al. \(2001\)](#) used cluster and canonical correspondence techniques to distinguish among the phytoliths from 69 Cyperaceae and Poaceae species. A transfer function technique used by [Prebble et al. \(2002\)](#) quantified the relationship between modern phytolith assemblages and environmental parameters for the quantification of a long late Quaternary paleoenvironmental record. [Blinnikov et al. \(2002\)](#) used a squared chord distance dissimilarity measure to successfully match modern and paleoassemblages for eight common vegetation types of the Pacific Northwest. For example, their study showed that it was possible to distinguish a ponderosa pine-dominated forest from a Douglas-fir-dominated forest or mesic *Agropyron*–*Festuca* grassland from xeric *Stipa*–*Poa*-dominated grassland. They also attempted direct interpretations of paleoclimate parameters based on their modern phytolith dataset. However, the range of modern climates from the region was not sufficiently broadly tested to distinguish much paleoclimatic diversity. Part of the problem was the necessary concentration of modern sample sites in only the few modern grasslands in the region, not fully representative of the modern climate range. Another

problem was an inability to match nonanalog paleoclimates of the late Pleistocene. [Bremond et al. \(2008a\)](#) demonstrated high precision of phytoliths in correctly detecting four key biomes in west Africa. In their study, phytoliths outperformed pollen in detecting regional biomes ([Figure 6](#)).

Phytoliths in Archaeological Research

The study of the Quaternary inevitably touches on human history, and many methods are employed in common in archaeology-, anthropology-, and geology-based Quaternary research. As the climate warmed following the last glacial interval, ca. 12 ka BP, humans began experiments in cultivating wild grasses, leading to the development of agriculture. Distinctive phytoliths have been described from wheat (*Triticum* spp.), barley (*Hordeum* spp.), millet (*Panicum* spp.), rice (*Oryza sativa*), maize (*Zea mays*), and wild rice (*Zizania palustris*). Other families that produce phytoliths important in archaeological research include bananas (Musaceae), squashes (Cucurbitaceae), euphorbias (Euphorbiaceae), bromeliads (Bromeliaceae), and dayflowers (Commelinaceae). Presence of these and other phytoliths in ancient sedimentary deposits

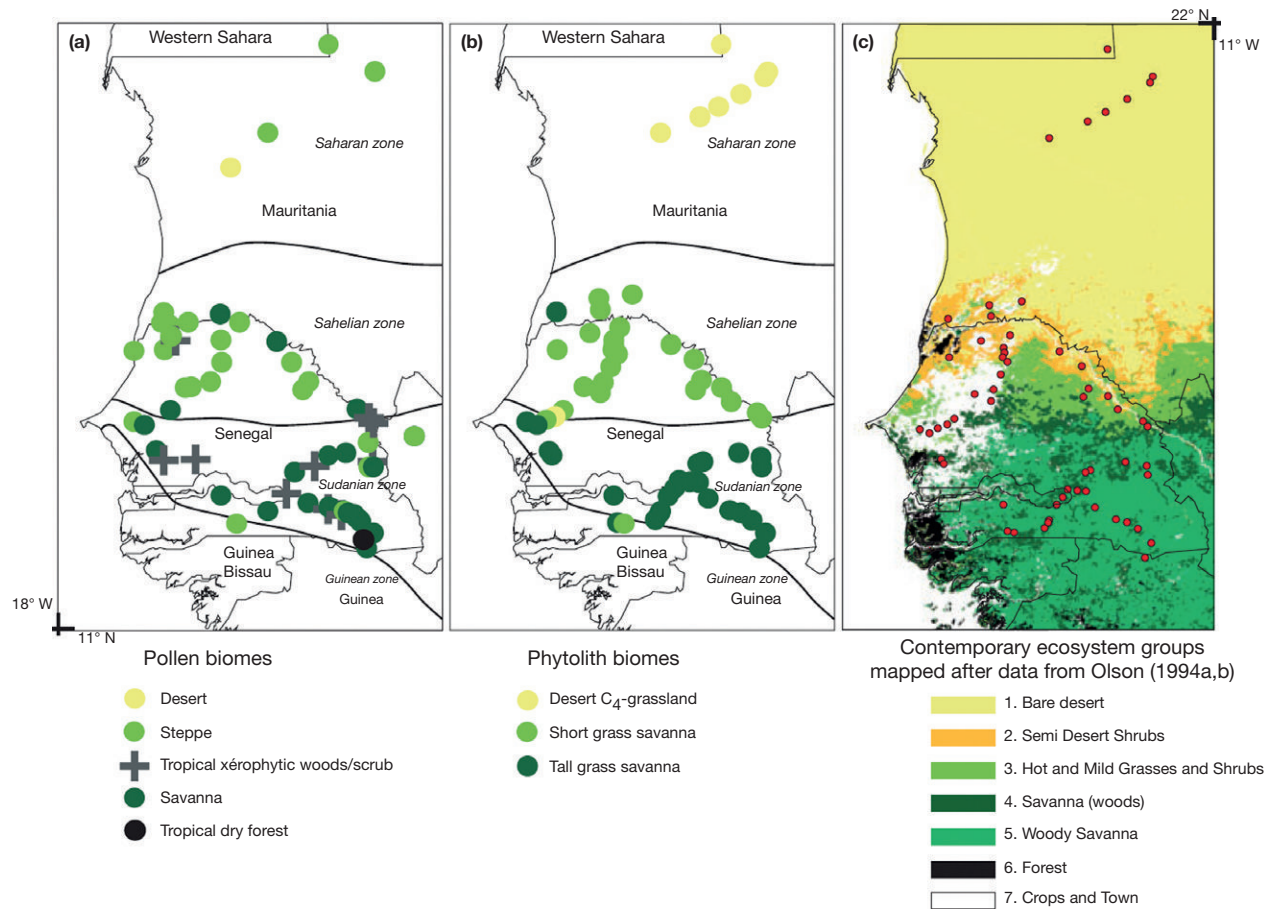


Figure 6 Assignment of the samples to (a) pollen and (b) phytolith biomes (Bremond et al., 2008a,b). (c) Olson's ecosystem classes (16 in this study area) were combined and reduced to seven groups representing more than 94% of the area. Reprinted from Bremond L, Alexandre A, Peyron O, and Guiot J (2008) Definition of grassland biomes from phytoliths in West Africa. *Journal of Biogeography* 35: 2039–2048, with permission.

has been used as evidence for the development of agriculture and landscape change in several independent centers, including North and South America, the Middle East, and south and southeast Asia. Accordingly, phytolith analysis is used extensively in archaeology, with particular respect to the early exploitation and development of grain crops (Meunier and Colin, 2001). Asian rice (*Oryza sativa* L. var. *japonica*) was domesticated in the middle Yangtze River Valley of southern China shortly after the termination of the Pleistocene. Distinctive morphological differences between phytoliths found in wild and domesticated *Oryza* species are particularly suitable in documenting change from early exploitation of a wild resource to its subsequent cultivation (Zhao and Piperno, 2000).

Analysis of phytoliths from sediments has also proved useful in describing the early cultivation of nongrain crops such as banana and taro. Hybridization of varieties of banana and of taro has resulted in near-sterile plants, producing little or no seed. However, these plants still produce distinctive phytoliths and starch, which have been extracted from prehistoric sediments and used to document the deliberate planting of taro (*Colocasia esculenta*) and banana (*Musa* spp.). This demonstrates that agriculture arose independently in New Guinea between 6950 and 6440 years BP (Denham et al., 2003). One

of the more interesting questions regarding the domestication of banana is how and when it got to Africa. No wild bananas have ever been reported in the natural state in Africa; therefore, they must have been deliberately introduced by humans. *Musa* phytoliths, dated from the first millennium BC, from an archaeological site in South Cameroon, document the first recorded presence of domesticated *Musa* bananas in central Africa (Mbida et al., 2000).

See also: Loess Deposits: Origins and Properties. **Luminescence Dating:** Optical Dating. **Radiocarbon Dating:** ^{14}C of Plant Macrofossils.

References

- Ball TB, Baird GI, al-Farsi A, Ghanzanfar SA, and Woolstenhulme LA (2007) A survey of phytoliths produced by the vegetation of Dhofar, Oman. In: Madella M and Zurro D (eds.) *Plants, People and Places: Recent Studies in Phytolith Analysis*, pp. 29–40. Oxford: Oxbow Books.
- Ball TB, Gardner JS, and Brotherson JD (1996) Identifying phytoliths produced by the inflorescence bracts of three species of wheat (*Triticum monococcum* L., *T.*

- dicoccon* Schrank, and *T. aestivum* L.) using computer-assisted image and statistical analysis. *Journal of Archaeological Science* 23: 619–632.
- Blackman E (1969) Observations on the development of the silica cells of the leaf sheath in wheat (*Triticum aestivum*). *Canadian Journal of Botany* 47: 827–838.
- Blinnikov MS (1994) Phytolith analysis and Holocene dynamics of alpine vegetation. *Experimental Investigation of Alpine Plant Communities in the Northwestern Caucasus*, vol. 115, pp. 23–42. Zurich: Veröffentlichungen des Geobotanischen Inst. Stiftung Rübel.
- Blinnikov MS (2005) Phytoliths in plants and soils of the interior Pacific Northwest, USA. *Review of Palaeobotany and Palynology* 135: 71–98.
- Blinnikov MS, Bagent CM, and Reyerson PE (2011) Phytolith assemblages and opal concentrations from modern soils differentiate temperate grasslands of controlled composition on experimental plots at Cedar Creek, Minnesota. *Quaternary International*. <http://dx.doi.org/10.1016/j.quaint.2011.12.023>.
- Blinnikov M, Busacca A, and Whitlock C (2002) Reconstruction of the late Pleistocene grassland of the Columbia basin, Washington, USA, based on phytolith records in loess. *Palaeogeography, Palaeoclimatology, Palaeoecology* 177: 77–101.
- Bobrov AA, Bobrova EK, and Alexeev JE (2001) Biogenic silica in biosystematics-potential uses. In: Meunier JD and Colin F (eds.) *Phytoliths: Applications in Earth Sciences and Human History*, pp. 279–288. Leiden: Balkema.
- Bowdery D, Hart DM, Lentfer C, and Wallis LA (2001) A universal phytolith key. In: Meunier JD and Colin F (eds.) *Phytoliths: Applications in Earth Sciences and Human History*, pp. 267–278. Leiden: Balkema.
- Bozarth SR (1992) Classification of opal phytoliths formed in selected dicotyledons native to the Great Plains. In: Rapp J and Mulholland SC (eds.) *Phytolith Systematics: Emerging Issues*, p. 193. New York: Plenum Press.
- Bremont L, Alexandre A, Hely C, and Guiot J (2005) A phytolith index as a proxy of tree cover density in tropical areas: Calibration with leaf area index along a forest-savanna transect in southeastern Cameroon. *Global and Planetary Change* 45: 277–293.
- Bremont L, Alexandre A, Peyron O, and Guiot J (2008a) Definition of grassland biomes from phytoliths in West Africa. *Journal of Biogeography* 35: 2039–2048.
- Bremont L, Alexandre A, Wooller MJ, Hely C, Williamson D, and Schafer PA (2008b) Phytolith indices as proxies of grass subfamilies on East African tropical mountains. *Global and Planetary Change* 61: 209–224.
- Cabanes D, Weiner S, and Shahack-Gross R (2011) Stability of phytoliths in the archaeological record: A dissolution study of modern and fossil phytoliths. *Journal of Archaeological Science* 38: 2480–2490.
- Carnelli AL, Madella M, Theurillat JP, and Ammann B (2002) Aluminium in the opal silica reticule of phytoliths: A new tool in palaeoecological studies. *American Journal of Botany* 89: 346–351.
- Carter JA (2009) Atmospheric carbon isotope signatures in phytolith-occluded carbon. *Quaternary International* 193: 20–29.
- Carter JA and Lian OB (2000) Paleoenvironmental reconstruction from the last interglacial using phytolith analysis, south eastern North Island, New Zealand. *Journal of Quaternary Science* 15: 733–743.
- Cordova C, Mandel R, Johnson W, and Palmer M (2010) Late Quaternary environmental change inferred from phytoliths and other soil-related proxies: Case studies from the central and southern Great Plains, USA. *Catena* 85: 87–108. <http://dx.doi.org/10.1016/j.catena.2010.08.015>.
- Cordova CE and Scott L (2010) The potential of Poaceae, Cyperaceae, and Restionaceae phytoliths to reflect past environmental conditions in South Africa. *Palaeoecology of Africa* 30: 107–133.
- Denham TP, Haberle SG, and Lentfer C (2003) Origins of agriculture at Kuk Swamp in the Highlands of New Guinea. *Science* 301: 189–193.
- Eichhorn B, Neumann K, and Garnier A (2010) Seed phytoliths in West African Commelinaceae and their potential for palaeoecological studies. *Palaeogeography, Palaeoclimatology, Palaeoecology* 298: 300–310.
- Elbaum R, Melamed-Bessudo C, Tuross N, Levy AA, and Weiner S (2009) New methods to isolate organic materials from silicified phytoliths reveal fragmented glycoproteins but no DNA. *Quaternary International* 193: 11–19.
- Fernández Honaine M, Osterrieth ML, and Zucol AF (2009) Plant communities and soil phytolith assemblages relationship in native grasslands from southeastern Buenos Aires province, Argentina. *Catena* 76: 89–96.
- Fishkis O, Ingwersen J, Lamers M, Denysenko D, and Streck T (2010) Phytolith transport in soil: A laboratory study on intact soil cores. *European Journal of Soil Science* 61: 445–455.
- Frayse F, Pokrovsky OS, Schott J, and Meunier J (2009) Surface chemistry and reactivity of plant phytoliths in aqueous solutions. *Chemical Geology* 258: 197–206.
- Fredlund GG (1993) Paleoenvironmental interpretations of stable carbon, hydrogen, and oxygen isotopes from opal phytoliths, Eustis Ash Pit, Nebraska. In: Pearsall DM and Piperno DR (eds.) *Current Research in Phytolith Analysis in Archaeology and* *Paleoecology*, pp. 37–46. Philadelphia: The University Museum of Archaeology and Anthropology, MASCA.
- Fredlund GG and Tieszen LL (1994) Modern phytolith assemblages from the North American Great Plains. *Journal of Biogeography* 21: 321–335.
- Fredlund GG and Tieszen LL (1997) Calibrating grass phytolith assemblages in climatic terms: Application to late Pleistocene assemblages from Kansas and Nebraska. *Palaeogeography, Palaeoclimatology, Palaeoecology* 136: 199–211.
- Gallego L and Distel RA (2004) Phytolith assemblages in grasses native to Central Argentina. *Annals of Botany* 94: 865–874.
- Geis JW (1973) Biogenic silica in selected species of deciduous angiosperms. *Soil Science* 116: 113–130.
- Hart DM (2001) Elements occluded within phytoliths. In: Meunier JD and Colin F (eds.) *Phytoliths: Applications in Earth Sciences and Human History*, pp. 313–316. Leiden: Balkema.
- Hart DM and Wallis LA (2003) *Phytolith and Starch Research in the Australian-Pacific-Asian Regions: The State of the Art*. Canberra: Pandanus Books.
- Kajale MD and Eksambekar SP (2007) Phytolith analytical study on a late Chalcolithic – Early historical archaeostratigraphical sequence from Balathal, south Rajasthan, India. In: Madella M and Zucol D (eds.) *Plants, People and Places: Recent Studies in Phytolith Analysis*, pp. 118–133. Oxford: Oxbow Books.
- Kealhofer L and Piperno D (1998) *Opal Phytoliths in Southeast Asian Flora*. *Smithsonian Contributions to Botany* Vol. 88. Washington, DC: Smithsonian Institution Press.
- Kelly EF, Amundson RG, Marino BD, and Deniro MJ (1991) Stable isotope ratios of carbon in phytoliths as a quantitative method of monitoring vegetation and climate change. *Quaternary Research* 35: 222–233.
- Kelly EF, Blecker SW, Yonker CM, Olson CG, Wohl EE, and Todd LC (1998) Stable isotope composition of soil organic matter and phytoliths as paleoenvironmental indicators. *Geoderma* 82: 59–81.
- Klein RL and Geis JW (1978) Biogenic silica in the Pinaceae. *Soil Science* 126: 145–155.
- Lu H, Wu N, Yang X, Jiang H, Liu K, and Liu T (2005) Phytoliths as quantitative indicators for the reconstruction of past environmental conditions in China I: Phytolith-based transfer functions. *Quaternary Science Reviews* 25: 945–959.
- Madella M, Alexandre A, and Ball T (2005) International code for phytolith nomenclature 1.0. *Annals of Botany* 96: 253–260.
- Massey FP, Ennos A, and Hartley SE (2007) Grasses and the resource availability hypothesis: The importance of silica-based defences. *Journal of Ecology* 95: 414–424.
- Mbida CM, Van Neer W, Doutrelepon H, and Vrydaghs L (2000) Evidence for banana cultivation and animal husbandry during the first millennium BC in the forest of Southern Cameroon. *Journal of Archaeological Science* 27: 151–162.
- McNaughton SJ, Tarrants JL, McNaughton MM, and Davis RH (1985) Silica as a defense against herbivory and a growth promoter in African grasses. *Ecology* 66: 528–535.
- Meunier JD and Colin F (2001) *Phytoliths: Applications in Earth Sciences and Human History*. Leiden: Balkema.
- Morris LR, Baker FA, Morris C, and Ryel RJ (2009) Phytolith types and type-frequencies in native and introduced species of the sagebrush steppe and pinyon-juniper woodlands of the Great Basin, USA. *Review of Palaeobotany and Palynology* 157: 339–357.
- Mulholland SC (1989) Phytolith shape frequencies in North Dakota grasses: A comparison to general patterns. *Journal of Archaeological Science* 16: 489–511.
- Mulholland SC and Prior C (1993) AMS radiocarbon dating of phytoliths. In: Pearsall DM and Piperno DR (eds.) *Current Research in Phytolith Analysis in Archaeology and Paleoecology*, pp. 21–23. Philadelphia: The University Museum of Archaeology and Anthropology, MASCA.
- Parr JF (2002) A comparison of heavy liquid floatation and microwave digestion techniques for the extraction of fossil phytoliths from sediments. *Review of Palaeobotany and Palynology* 120: 315.
- Pearsall DM (2000) *Paleoethnobotany: A Handbook of Procedures*. San Diego, CA: Academic Press.
- Pearsall DM and Piperno DR (1993) *Current Research in Phytolith Analysis: Applications in Archaeology and Paleoecology*. Philadelphia: The University Museum of Archaeology and Anthropology, MASCA.
- Piperno DR (1988) *Phytolith Analysis: An Archaeological and Geological Perspective*. San Diego, CA: Academic Press.
- Piperno DR (2006) *Phytoliths: A Comprehensive Guide for Archaeologists and Paleoecologists*. Lanham, MD: Altamira Press.
- Powers-Jones AH and Padmore J (1993) The use of quantitative methods and statistical analyses in the study of opal phytoliths. In: Pearsall DM and Piperno DR (eds.) *Current Research in Phytolith Analysis: Applications in Archaeology and*

- Paleoecology*, p. 47. Philadelphia: The University Museum of Archaeology and Anthropology, MASCA.
- Prasad V, Strömberg CE, Alimohammadian H, and Sahni A (2005) Dinosaur coprolites and the early evolution of grasses and grazers. *Science* 310: 1177–1180.
- Prebble M, Schallenburg M, Carter J, and Shulmeister J (2002) An analysis of phytolith assemblages for the quantitative reconstruction of late Quaternary environments of the Lower Taieri Plain, Otago, South Island, New Zealand I. Modern assemblages and transfer functions. *Journal of Paleolimnology* 27: 393–413.
- Rapp J and Mulholland SC (1992) *Phytolith Systematics: Emerging Issues*. New York: Plenum Press.
- Raven JA (1983) The transport and function of silica in plants. *Biological Reviews of the Cambridge Philosophical Society* 58: 179–207.
- Rowlett RM and Pearsall DM (1993) Archaeological age determinations derived from opal phytoliths in thermoluminescence. In: Pearsall DM and Piperno DR (eds.) *Current Research in Phytolith Analysis: Applications in Archaeology and Paleoecology*, p. 25. Philadelphia: The University Museum of Archaeology and Anthropology, MASCA.
- Runge F (1999) The opal phytolith inventory of soils in Central Africa—quantities, shapes, classifications, and spectra. *Review of Palaeobotany and Palynology* 107: 23–53.
- Saccone L, Conley DJ, Koning E, et al. (2007) Assessing the extraction and quantification of amorphous silica in soils of forest and grassland ecosystems. *European Journal of Soil Science* 58: 1446–1459.
- Sangster AG, Hodson MJ, and Tubb HJ (2001) Silicon deposition in higher plants. In: Datnoff IE, Snyder GH, and Korndorfer GH (eds.) *Silicon in Agriculture*, pp. 85–113. Amsterdam: Elsevier.
- Smith FA and Anderson KB (2001) Characterization of organic compounds in phytoliths: Improving the resolving power of phytolith $\delta^{13}\text{C}$ as a tool for paleoecological reconstruction of C_3 and C_4 grasses. In: Meunier JD and Colin F (eds.) *Phytoliths: Applications in Earth Sciences and Human History*, pp. 317–327. Leiden: Balkema.
- Smith FA and White JWC (2004) Modern calibration of phytolith carbon signatures for C_3/C_4 paleograssland reconstruction. *Palaeogeography, Palaeoclimatology, Palaeoecology* 207: 277–304.
- Smithson F (1958) Grass opal in British soils. *Journal of Soil Science* 9: 148–154.
- Strömberg CE (2005) Decoupled taxonomic radiation and ecological expansion of open-habitat grasses in the Cenozoic of North America. *Proceedings of the National Academy of Sciences of the United States of America* 102: 11980–11984.
- Struve GA (1835) *De Silica in Plantis Nonnullis*. Dissertation, Berlin.
- Twiss PC (2001) A curmudgeon's view of grass phytolithology. In: Meunier JD and Colin F (eds.) *Phytolith: Applications in Earth Sciences and Human History*, pp. 7–25. Leiden: Balkema.
- Twiss PC, Suess E, and Smith RM (1969) Morphological classification of grass phytoliths. *Soil Science Society of America Proceedings* 33: 109–115.
- Wallis LA (2003) An overview of leaf phytolith production patterns in selected northwest Australian flora. *Review of Palaeobotany and Palynology* 125: 201–248.
- Webb EA and Longstaffe FJ (2003) The relationship between phytolith and plant water $\delta^{18}\text{O}$ values in grasses. *Geochimica et Cosmochimica Acta* 67: 1437–1449.
- Webb EA and Longstaffe FJ (2010) Limitations on the climatic and ecological signals provided by the $\delta^{13}\text{C}$ values of phytoliths from a C_4 North American prairie grass. *Geochimica et Cosmochimica Acta* 74: 3041–3050.
- Wilding LP (1967) Radiocarbon dating of biogenic opal. *Science* 156: 66–67.
- Wooler MJ, Zazula GD, Blinnikov MS, et al. (2011) The detailed palaeoecology of a mid-Wisconsinan interstadial (ca. 32 000 ^{14}C a BP) vegetation surface from interior Alaska. *Journal of Quaternary Science* 26: 746–756. <http://dx.doi.org/10.1002/jqs.1497>.
- Yost CL (2007) *Phytolith Evidence for the Presence and Abundance of Wild Rice (Zizania palustris) from Central Minnesota Lake Sediments*. Cloud, MN: St. Cloud State University MS Thesis.
- Yost CL and Blinnikov MS (2011) Locally diagnostic phytoliths of wild rice (*Zizania palustris* L.) from Minnesota, USA: Comparison to other wetland grasses and usefulness for archaeobotany and paleoecological reconstructions. *Journal of Archaeological Science* 38: 1977–1991.
- Zhao Z and Piperno DR (2000) Late Pleistocene/Holocene environments in the middle Yangtze River Valley, China and rice (*Oryza sativa* L.) domestication: The phytolith evidence. *Geoarchaeology* 15: 203–222.

Relevant Websites

- <http://www.mnh.si.edu/highlight/phytoliths/index.html> – Phytoliths: An Archaeobotanist's Best Friend.
- <http://www.paleoresearch.com/services/phytoliths.html> – Phytolith and Starch Analysis.
- <http://phytolith.missouri.edu/index.shtml> – Phytolith Database from Pearsall's Lab.
- <http://www.phytolithresearch.com/> – Phytolith research in India.
- <http://www.phytolithsociety.org/> – International Society for Phytolith Research.

Plant Macrofossil Introduction

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Introduction

Plant macrofossil analysis is being increasingly used in Quaternary science. Quantitative paleoecological macrofossil analysis was developed by W.A. Watts in the United States (Birks and Birks, 1980; Watts and Winter, 1966) and his work showed the value of combining macrofossils with pollen analysis. Macrofossil analysis is a 'low-tech' process, but it requires patience and a good botanical knowledge of plant identification and ecology. This subsection describes the up-to-date developments in macrofossil analysis and illustrates how diverse its uses and applications have become. It has become an essential tool in paleoecological studies and, of course, has long been the foundation of paleoethnobotany and archaeobotany. This article is intended to provide a background and a survey of the uses of plant macrofossil analysis as presented in the literature. Further background information is available from the list of works found in the References section.

What Is a Plant Macrofossil?

A plant macrofossil can be considered to be a plant fossil that is visible to the naked eye and that can be manipulated by hand. The sizes of macrofossils range from 0.5 mm to enormous tree trunks (usually termed 'megafossils'). The median size is 0.5–2.0 mm. Macrofossils are usually studied using a stereomicroscope at magnifications between 10× and 40×. Greater magnifications (e.g., 100×) can be obtained using a high-power microscope.

Plant remains are not only fruits and seeds but also vegetative parts, such as leaves, cuticles, buds, bud scales, anthers, flower parts, rhizomes, twigs, wood, bark, etc. Lower-plant remains include sporangia and megaspores, mosses (but seldom liverworts, as they are usually not preserved), occasionally lichens and seaweeds, and *Characeae* oospores. During plant macrofossil analysis, animal remains are frequently found, especially in lake sediments (e.g., chironomids, *Cladocera ephippia*, bryozoa, oribatida, Trichoptera, and fish). Quaternary plant macrofossils are seldom true fossils where the organic material is replaced by inorganic minerals. They are most often the original organic material and thus can be termed subfossils.

Occasionally, the remains are trace fossils, for example, impressions in sediments or 'negative' impressions in pottery.

Why Study Plant Macrofossils?

What Can Plant Macrofossils Add to a Paleoecological Study?

Macrofossils are often studied in conjunction with pollen and add new dimensions to pollen studies.

- Macroremains can often be identified to a higher taxonomic resolution than pollen, often to species level.
- Macroremains can be recovered of taxa that produce little or no pollen, including cryptograms.
- Macroremains are relatively heavy and are generally not dispersed far from the parent plant. Exceptionally, streams and rivers may transport remains far from their source, and winter winds can disperse seeds and fruits many kilometers over a smooth winter-snow surface. These processes aid dispersal and colonization of recently deglaciated landscapes. In the Arctic, they have contributed to the formation of the relatively uniform circumarctic flora. However, most plant remains are deposited near their sources, which allows past local vegetation to be reconstructed in some detail.
- When local pollen production is sparse, the pollen signal may be confused by the relatively large amounts of pollen that have been transported from great distances. In such situations, plant macrofossils are needed to provide realistic reconstructions of the local vegetation (see Birks and Birks, 2000, 2003).

Where Are Macrofossils Found?

Being organic, plant remains are found only in places where decay is inhibited, usually in anoxic, waterlogged, arid, or sometimes frozen environments. They are best preserved in low-energy environments because of their delicate nature.

Macrofossils are most often studied from lake or peat sediments. Sequential deposition means that temporal series can be analyzed. Rapid accumulation can result in finely detailed sequences showing changes in vegetation and environment

(e.g., Birks, 2000; Birks et al., 2000). Mire communities produce peat *in situ*, and the plant remains thus reflect the local peat-forming vegetation. Fruits and seeds can be sparse in peat deposits, especially bog peat, and thus the identification and quantification of vegetative parts and mosses, including *Sphagnum*, play an important role. Macrofossils in lake sediments also reflect local vegetation but the assemblages also contain remains from neighboring communities, such as littoral swamps and nearby upland vegetation, especially trees (e.g., Eide et al., 2006). Remains can be transported into lakes by wind, birds, inflowing streams, solifluction, and slope-wash (an extreme example is avalanches; Figure 1; Birks, 2001).

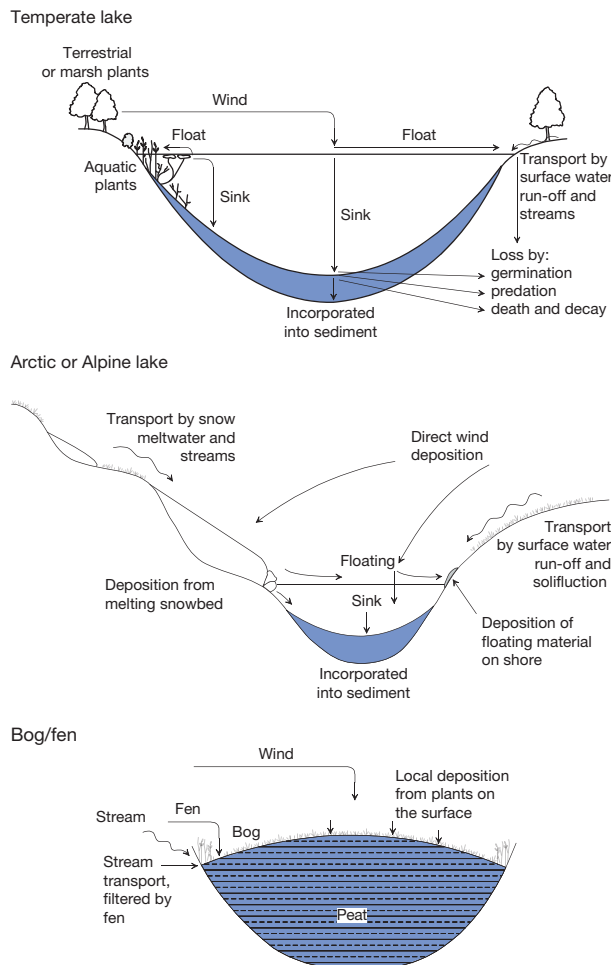


Figure 1 Illustration of the taphonomic processes leading to the preservation of macrofossil assemblages in different sedimentary environments. Top: A temperate lake (reproduced from Birks HH (1980) Plant macrofossils in Quaternary lake sediments. *Archiv für Hydrobiologie* 15: 1–60). Middle: An arctic or alpine lake (reproduced from Birks HH (1991) Holocene vegetation history and climatic change in west Spitsbergen – Plant macrofossils from Skardtjørna, an arctic lake. *The Holocene* 1: 209–218). Bottom: A bog or fen. Reproduced from Birks HH (2001) Plant macrofossils. In: Smol JP, Birks HJB, and Last WM (eds.) *Tracking Environmental Change Using Lake Sediments. Terrestrial, Algal, and Siliceous Indicators*, vol. 3, pp. 49–74, Figure 2, © Kluwer Academic Publishers, Dordrecht, The Netherlands, with kind permission of Springer Science and Business Media.

Elsewhere, plant remains are usually deposited as single events. Plant macrofossils can be recovered from soils, and they may contribute to seed-bank studies. Old soils can be preserved by permafrost, and long sequences may be built up by loess deposition. Complete vegetation can be preserved under volcanic ash (e.g., Baales et al., 2002). About 18 000 years ago, arctic tundra was preserved under tephra in the Seward Peninsula, Alaska. Most soft parts disappeared before permafrost penetrated the ash layer, so seeds and fruits, some vegetative remains, and mosses were the major sources of evidence for a local vegetation mosaic determined by moisture availability and snow lie (Goetcheus and Birks, 2001). Preservation in tufa can likewise reflect local vegetation.

Alluvial sediments can be productive sources of plant macrofossils. Macrofossils from Holocene alluvial deposits allowed Baker and colleagues (this encyclopedia) to reconstruct riverine and upland plant communities within the forest and prairie biomes that are poorly differentiated by pollen. Outwash plains of glacial rivers contain organic deposits laid down in calm backwaters, which are subsequently buried (e.g., Holyoak, 1984). The organic lenses in glaciofluvial deposits are important sources of Pleistocene vegetation history. In embayments of forest streams, tree leaves may be well preserved (e.g., Chaney, 1924). Leaf layers may also be preserved in lake sediments, both temperate (e.g., Spicer, 1989) and glacial (e.g., Kråkenes Lake; Gulliksen et al., 1998).

Another source of macrofossil material is archaeological excavations (See Use in Environmental Archaeology). Knowledge of the taphonomy of the samples can yield important information on the lifestyle and economy of past human populations. Samples can be preserved by water logging, such as in wells, latrines, and ditches, and by high water tables in lakeshore dwellings (e.g., in Switzerland) and seashore buildings (e.g., Medieval Bryggen in Bergen; Krzywinski and Soltevedt, 1988). Dry environments preserve samples associated with dwellings (e.g., middens, post holes, hearths, tombs, etc.). Carbonized remains are preserved by their conversion to charcoal. Pottery can contain negative impressions of plant remains that were incorporated in the clay and subsequently incinerated.

An unusual source of plant macrofossils is bodies preserved in peat bogs (e.g., Tollund Man in Denmark, Glob, 1971; Lindow Man in England, Stead et al., 1986), in permafrost (e.g., in Siberia and Alaska), and in ice (e.g., The Italian Ice Man). Frozen corpses of large grazing mammals from full-glacial times have been widely studied from Siberia (e.g., Ukraintseva, 1993) and Alaska (e.g., Zazula et al., 2003). Their diet, reconstructed from stomach contents and dung, reflects the full-glacial vegetation and the nature of the so-called Mammoth Steppe. Analyses of extinct ground-sloth dung from a cave in Patagonia, Argentina, revealed the character of the glacial vegetation before trees arrived in the Holocene (Moore, 1978).

How Do We Sample Plant Macrofossils?

Most lake and bog sediments are sampled by coring. As a larger quantity of sediment is required for macrofossil analysis than for pollen analysis, wide-diameter cores are preferred. With wide cores, high-resolution sampling is possible. Monoliths or spot samples can be excavated from open sections in dry bogs, alluvial

banks, glacial gravel deposits, and archaeological contexts. The monoliths or samples are subsampled in the laboratory or screened on site. All sediments should be refrigerated at 4 °C until required or air- or freeze-dried. For accelerator mass spectrometry (AMS) radiocarbon dating, it is best to dry or deep freeze the samples at –20 °C to prevent microorganism metabolism.

How Do We Do Plant Macrofossil Analysis?

The whole process of macrofossil analyses is summarized in [Figure 2](#) (Birks, 2001). After the material is obtained from the field, subsamples are taken depending on the abundance of

macrofossils and the resolution required. Subsamples for quantitative analysis are ideally about 50 cm³. However, samples of ~25 cm³ or less have been successfully used when the amount of available material is small. The subsamples are stored as above until needed.

For quantitative analyses, subsample volume is measured by displacement of water. Sometimes, weight may be used for the dry sediment. Sediments may need to be disaggregated. Sodium pyrophosphate (Na₄P₂O₇ · 10H₂O) is useful for disaggregating clay-rich samples. Sometimes, gyttja needs to be soaked for a short time (1–2 h) in 10% potassium hydroxide. Calcareous samples can be treated with dilute acid. Archaeological samples can undergo various treatments.

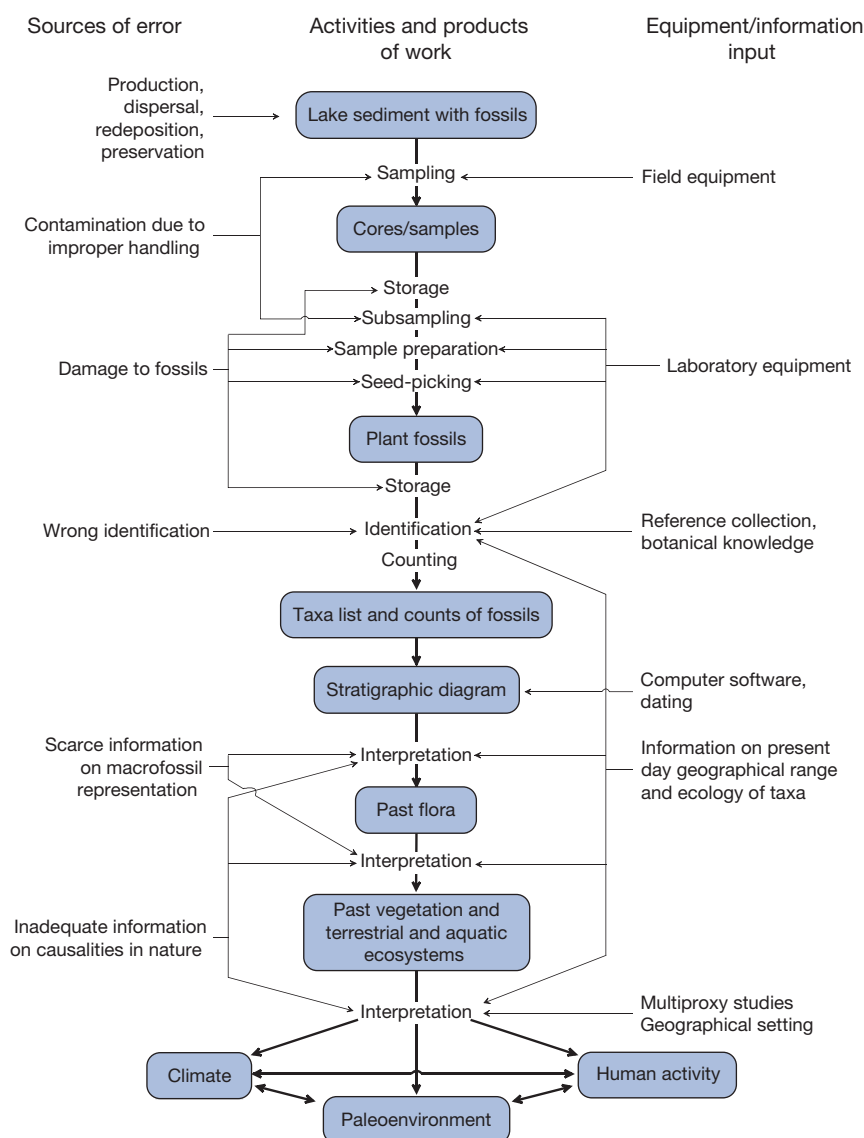


Figure 2 Flowchart showing the processes in a plant macrofossil study. The central column shows the successive phases from sampling to the reconstruction of paleoenvironment and climate. The equipment and information needed are shown on the right. The potential sources of error are shown on the left. Modified from Wasylkova K (1986) Analysis of fossil fruits and seeds. In: Berglund BE (ed.) *Handbook of Holocene Palaeoecology and Palaeohydrology*, pp. 571–590. Chichester: Wiley by Birks HH (2001) Plant macrofossils. In: Smol JP, Birks HJB, and Last WM (eds.) *Tracking Environmental Change Using Lake Sediments. Vol. 3: Terrestrial, Algal, and Siliceous Indicators*, pp. 49–74, Figure 1, © Kluwer Academic Publishers, Dordrecht, The Netherlands, with kind permission of Springer Science and Business Media.

The sample is washed through sieves with a gentle spray of water. Sieves can be stacked if necessary, but the smallest mesh diameter should be about 125 μm if tiny seeds are to be collected. The residue is washed gently off the sieve into a container and kept cold until analyzed. Small quantities of the residue are suspended in 2–3 mm of water in a shallow dish (e.g., Petri dish) and examined systematically under a stereomicroscope at about $12\times$ magnification until the whole sample has been examined. Remains of interest are picked out and sorted, identified, counted, and tabulated.

Results may be presented as lists with presence/absence or numbers, as spots indicating abundance next to the appropriate pollen taxon on a pollen diagram from the same core, or as stratigraphic concentration diagrams (number in a constant volume of sediment calculated from the numbers counted in known sediment volumes). Influx values can be calculated if the sediment accumulation rate is known, but this is rarely done. Normally, percentage diagrams are not used, as there is no consistent 'macrofossil rain' comparable to 'pollen rain,' but they have proved useful in several cases. TILIA.GRAPH has proved to be an excellent computer tool for drawing stratigraphic macrofossil diagrams (<http://www.demeter.museum.state.il.us/pub/grimm>). The program C2 is ideal for plotting various proxies from the same or different cores together.

Identification of Macrofossils

A comprehensive reference collection is necessary for macrofossil identification. For material collected in the field, voucher plant specimens should be made for taxonomically difficult taxa. Alternatively, material may be collected from herbaria. Herbarium specimens are usually collected in flower and may not have seeds present, but they can provide vegetative material. Botanic garden seed exchanges can provide useful material whose provenance is known. Seed banks such as the Kew Millennium Seed Bank and the US Department of Agriculture can also provide material. All items in a reference collection should be numbered and cataloged.

Published photographs and drawings are useful for narrowing down an identification before consulting the reference collection. There are illustrated seed atlases and articles about single genera (see also Birks, 2001). A selection of modern seeds, fruits, and other parts that are often found fossilized or that have proved to be important indicators in macrofossil studies has been photographed for this article (Figures 3–13).

Fossils and reference material are compared under a stereomicroscope. Higher magnification with transmitted or overhead light or a combination of both can reveal cell patterns and other small details. Scanning electron micrographs can be taken, but this is usually impractical for routine identification. Modern material can be 'fossilized' by soaking or heating in 10% potassium hydroxide to remove waxes, fleshy tissue, etc. that normally disappear in sediments. Seed surfaces can usefully be examined while they are drying out; damp or dry surfaces often reveal patterns invisible in wet specimens. Identifications can be easily recorded photographically with the digital equipment now available for both stereo- and high-powered microscopes.

How Do We Interpret Plant Macrofossil Data?

The occurrence of a taxon in an assemblage can be interpreted in terms of its modern ecological tolerances, using it as an indicator species. This approach is used in paleoecology and also in archaeobotany where plant remains are associated with human activity rather than natural vegetation. Assemblages can be interpreted as analogs with modern vegetation and environment, which can strengthen interpretations made from single taxa (Birks and Birks, 1980).

In a quantitative analysis, how can we interpret quantitative changes in the abundance of taxa? A macrofossil assemblage, however local, is derived from vegetation at different scales. Its composition depends on the processes of production, dispersal, deposition (including sorting and redeposition), and preservation (Figure 14). Seed or fruit dispersal mechanism may be unspecialized, but most plants have adaptations that aid dispersal: wings and parachutes for wind dispersal, burrs and reversed spines for external animal dispersal, succulent parts for internal animal dispersal, and corky tissue or small seeds with reticulate patterns that trap air for water dispersal. In reality, many processes may influence seed/fruit dispersal, as well as the apparent adaptation; for example, many propagules float. Distribution of plant remains may be affected by stream or river dynamics, sheet-wash, storms, fires, and smooth snow or ice in winter facilitating wind dispersal. Animal dispersal is influenced by foraging behavior and food preference (e.g., packrats and humans). The arrival of plant remains in a deposit is largely by chance, as few such places are suitable for germination and seedling establishment. Subsequent preservation depends partly on the deposition environment (waterlogged and anoxic, dry, acid, or alkaline) and partly on the structure and composition of the potential macrofossil. Some seeds and fruits are rarely found fossilized, even though they were abundantly produced. A most frustrating example is *Artemisia* fruits. Alternatively, some remains can be superabundant, for example *Juncus* seeds and *Characeae* oospores. This kind of huge imbalance makes the calculation of percentages rather meaningless in most situations.

Taphonomic processes result in a fossil assemblage in which the abundance of the fossils is related somehow to their abundance in the vegetation (Figure 14). Each taphonomic process can be studied independently. However, the pragmatic approach integrates all the processes and relates the fossil assemblages to modern assemblages through modern surface samples. A series of surface samples will indicate the variability of assemblages produced by a vegetation type due to the variability in taphonomic processes.

It is not feasible to relate macrofossil assemblages precisely to modern plant occurrences and abundances, but general relationships can be established (e.g., Birks, 1973). Although macrofossils can be counted or their abundance estimated and stratigraphic diagrams plotted, the actual values are rarely considered as a quantitative representation of the parent vegetation. Gradually increasing knowledge about the representation of plant remains in different depositional environments allows us to estimate in general terms the abundance of taxa in the past and to relate this to modern situations (Birks, 1973).

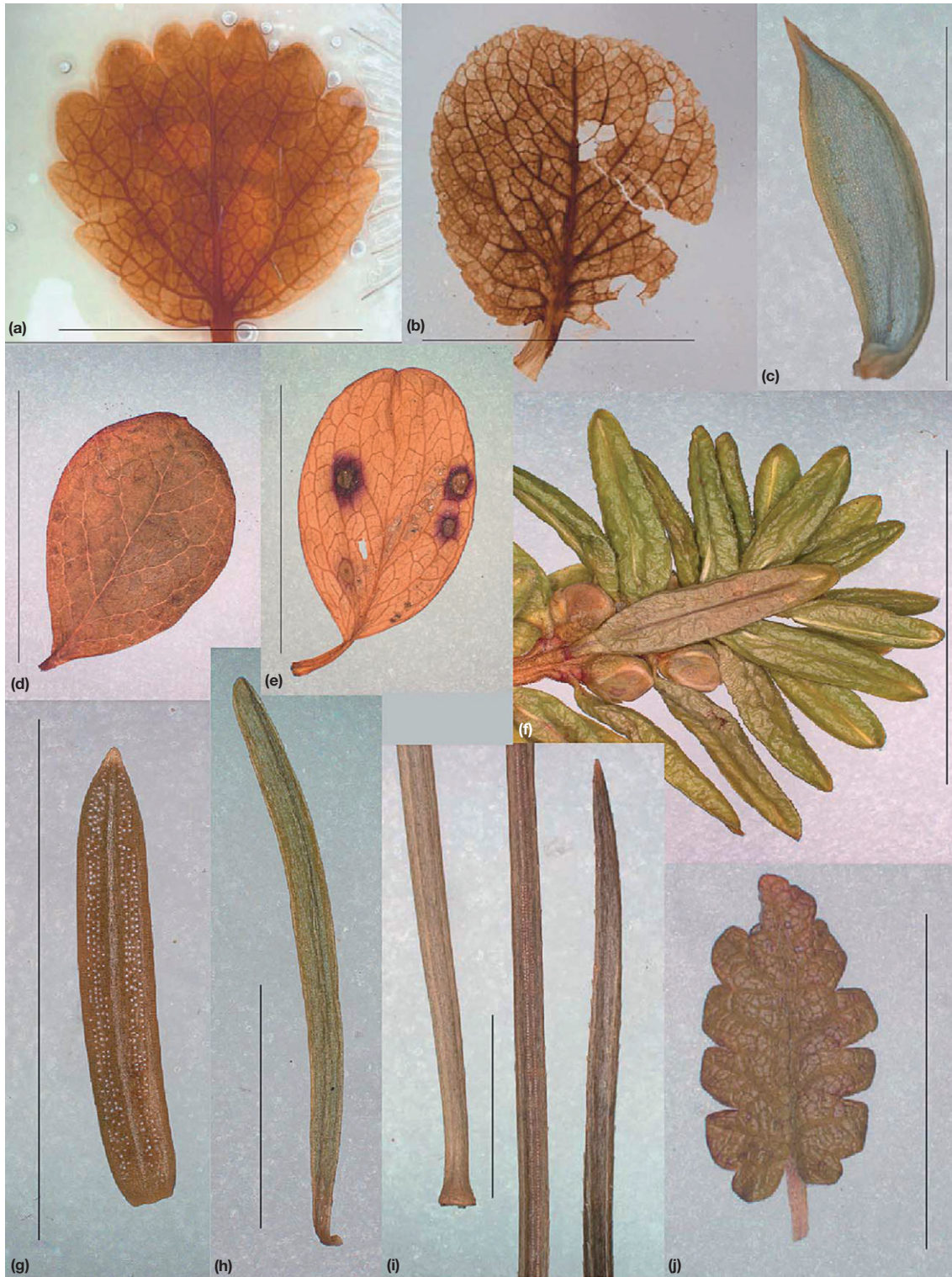


Figure 3 Leaves: (a) *Betula nana*; (b) *Salix herbacea* (fossil); (c) *Juniperus communis* showing stomata; (d) *Vaccinium uliginosum* var. *alpinum*; (e) *Salix polaris*; (f) *Empetrum nigrum*; (g) *Picea abies*; (h) *Larix europaea*; (i) *Pinus strobus*; and (j) *Dryas octopetala*. Scale lines are 1 cm.



Figure 4 Trees and shrubs: (a) *Betula pubescens* fruit; (b) *Betula nana* fruit; (c) *Alnus glutinosa* fruit; (d) *Betula pubescens* female catkin scale; (e) *Betula nana* female catkin scale; (f) *Picea abies* seed; (g) *Pinus sylvestris* male cone; (h) *Pinus sylvestris* bud scale; (i) *Actinidia chinensis* (Kiwi fruit) seed; (j) *Populus tremula* catkin scale; and (k) *Prunus spinosa* endocarp. Scale lines are 1 mm.

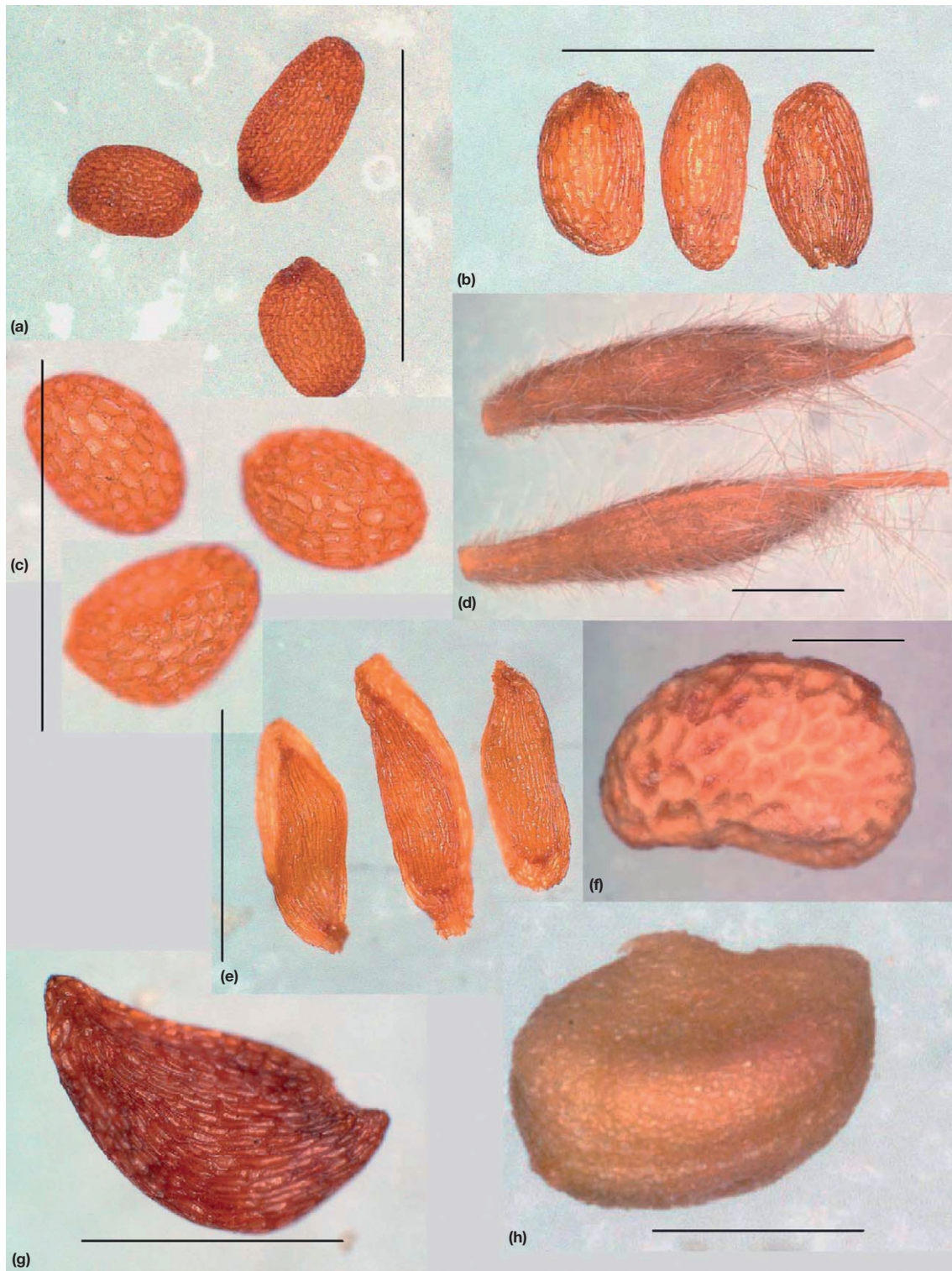


Figure 5 Dwarf shrubs: (a) *Bruckenthalia spiculifolia* seeds; (b) *Cassiope hypnoides* seeds; (c) *Loiseleuria procumbens* seeds; (d) *Dryas octopetala* fruits; (e) *Rhododendron ferrugineum* seeds; (f) *Rubus idaeus* seed; (g) *Vaccinium myrtillus* seed; and (h) *Empetrum nigrum* seed. Scale lines are 1 mm.

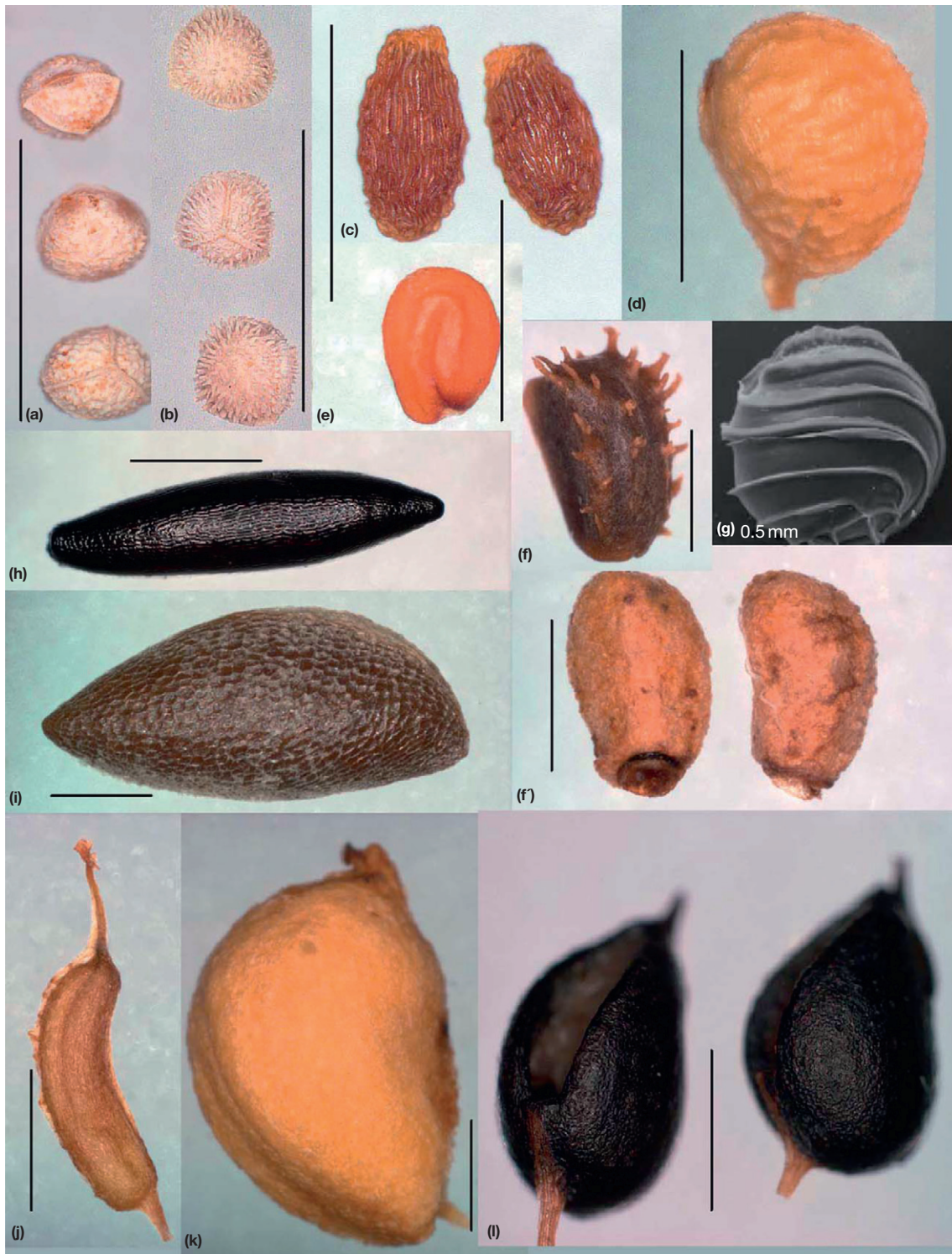


Figure 6 Aquatic plants with submerged leaves: (a) *Isoetes lacustris* megaspores; (b) *Isoetes echinospora* megaspores; (c) *Lobelia dortmanna* seeds; (d) *Ranunculus trichophyllus* (subgenus *Batrachium*) fruit; (e) *Subularia aquatica* seed; (f) *Myriophyllum spicatum* (extreme spiny seed); (f') *M. spicatum* (smooth seeds); (g) *Nitella opaca* oospore (scanning electron micrograph, SEM); (h) *Naias flexilis* seed; (i) *N. marina* seed; (j) *Zannichellia palustris* fruit; (k) *Potamogeton pectinatus* endocarp; and (l) *Ruppia maritima* fruits. Scale lines are 1 mm.

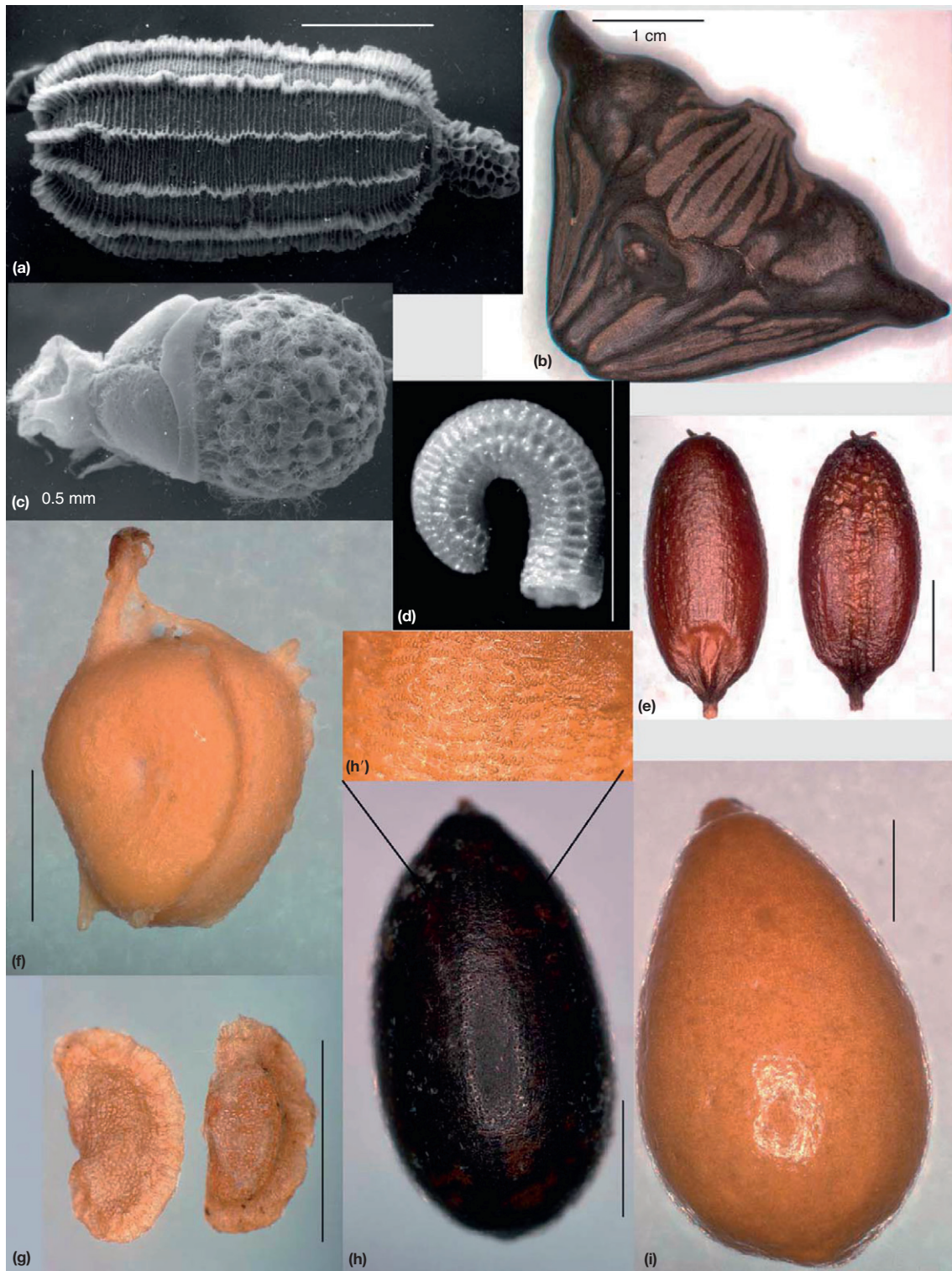


Figure 7 Aquatic plants with floating leaves: (a) *Eichhornia crassipes* seed (SEM); (b) *Trapa natans* fruit; (c) *Azolla filiculoides* megaspore (SEM); (d) *Elatine hydropiper* seed; (e) *Glyceria fluitans* caryopses; (f) *Potamogeton natans* endocarp; (g) *Callitriche stagnalis* fruits; (h) *Nymphaea alba* seed, with detail of cell pattern; and (i) *Nuphar lutea* seed. Scale lines are 1 mm.



Figure 8 Emergent aquatic and fen plants: (a) *Hippuris vulgaris* fruit and seed; (b) *Alisma plantago-aquatica* seed; (c) *Eupatorium cannabinum* fruits; (d) *Bidens cernua* fruit; (e) *Epilobium anagallidifolium* fruits; (f) *Epilobium (Chamaenerion) angustifolium* fruits; (g) *Epilobium hirsutum* fruits. (h) *Trollius europaeus* seed; (i) *Caltha palustris* seed; and (j) *Filipendula ulmaria* fruits. Scale lines are 1 mm.

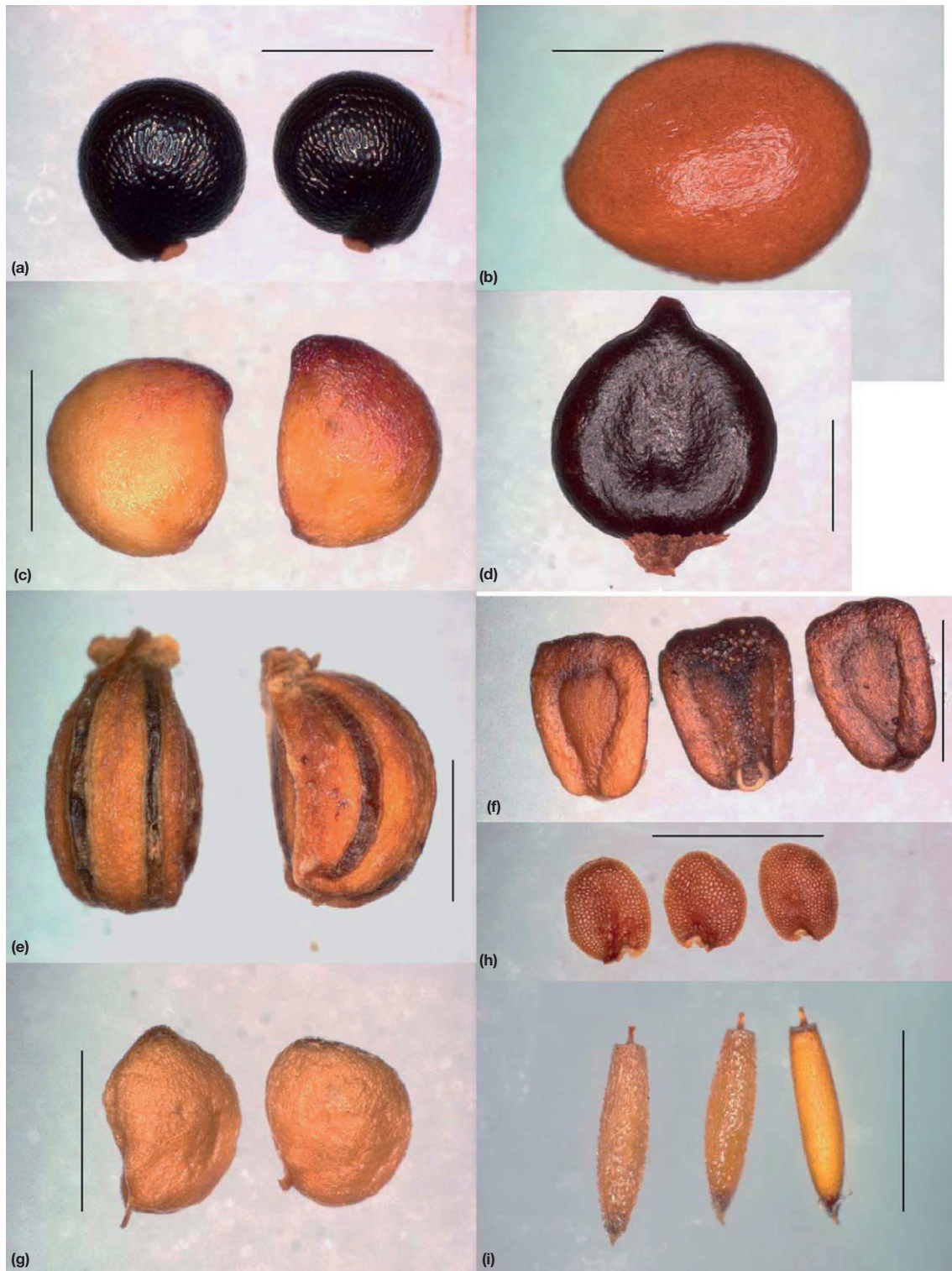


Figure 9 Emergent aquatics and fen plants: (a) *Montia fontana* seeds; (b) *Menyanthes trifoliata* seed; (c) *Potentilla* (*Comarum*) *palustre* fruits; (d) *Polygonum lapathifolium* seed; (e) *Cicutia virosa* fruits; (f) *Lycopodium europaeus* fruits; (g) *Ranunculus hyperboreus* fruits; (h) *Rorippa islandica* seeds; and (i) *Typha latifolia* fruits. Scale lines are 1 mm.



Figure 10 Cyperaceae: (a) *Eleocharis palustris*; (b) *Eleocharis uniglumis*; (c) *Carex rostrata*; (d) *Carex nigra*; (e) *Scirpus lacustris*; (f) *Trichophorum caespitosum*; (g) *Cyperus engelmannii*; (h) *Eriophorum vaginatum*; (i) *Cladium mariscus*; (j) *Isolepis setaceus*; and (k) *Kobresia myosuroides*. Scale lines are 1 mm.

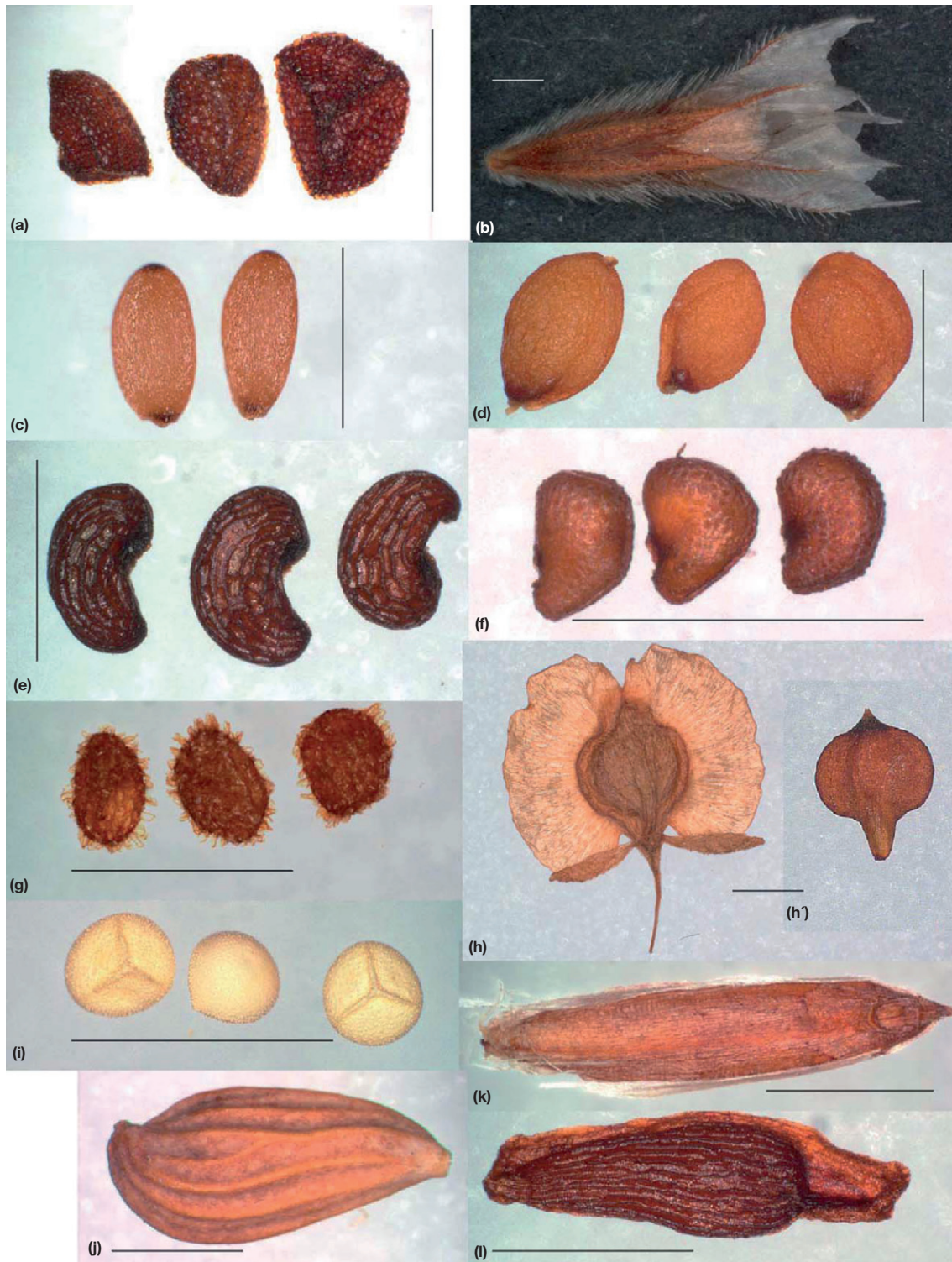


Figure 11 Some arctic and alpine plants: (a) *Androsace septentrionalis* seeds; (b) *Armeria maritima* fruit (calyx); (c) *Campanula rotundifolia* seeds; (d) *Draba incana* seeds; (e) *Papaver radicum* ssp. *ovatilobum* seeds; (f) *Sagina intermedia* seeds; (g) *Gentianella detonsa* seeds; (h) *Oxyria digyna* fruit and (h') seed; (i) *Selaginella selaginoides* megaspores; (j) *Thalictrum alpinum* fruit; (k) *Festuca rubra* caryopsis and (l) *Sedum rosea* seed. Scale lines are 1 mm.

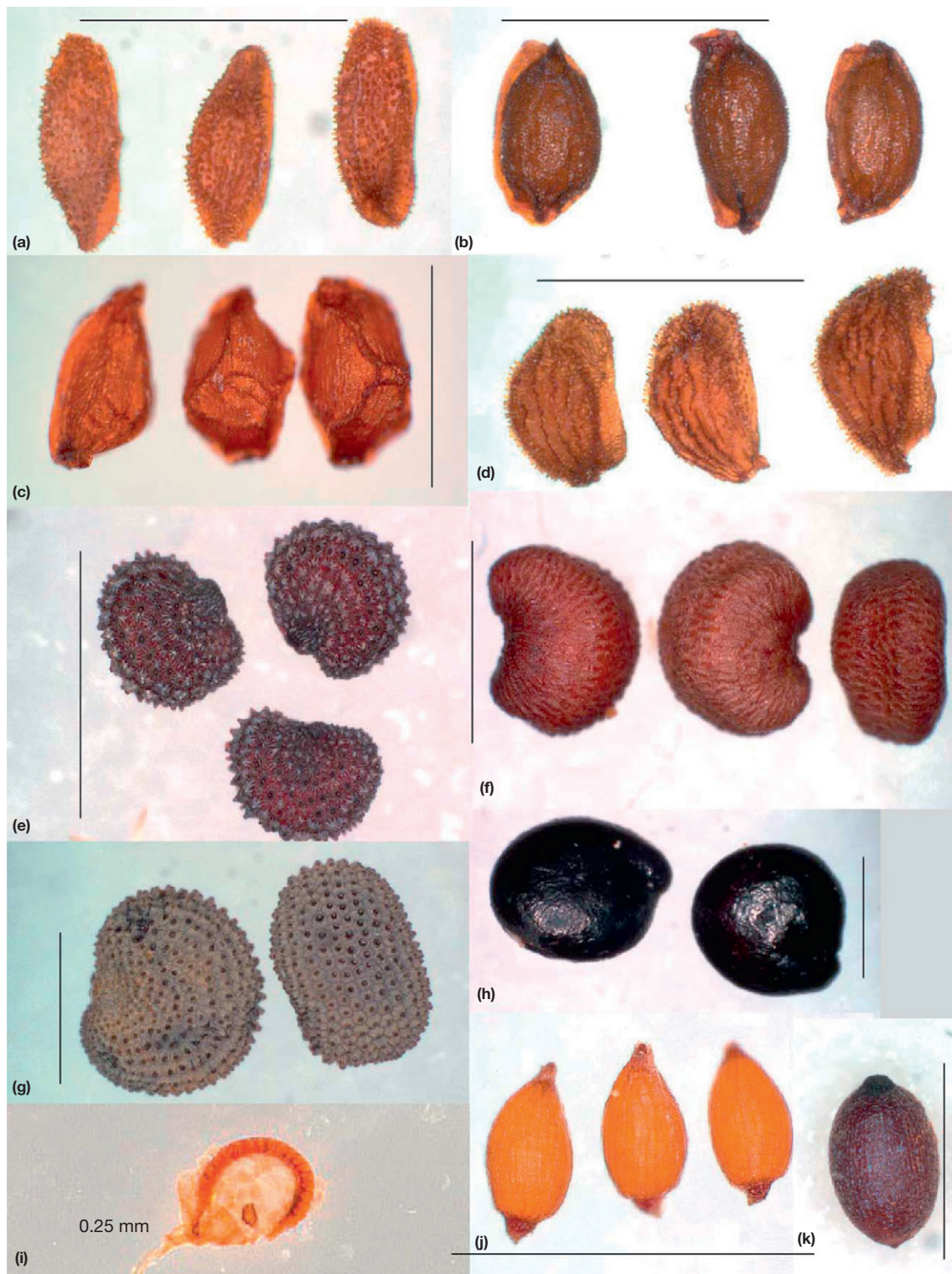


Figure 12 Saxifraga, Caryophyllaceae, and others: (a) *Saxifraga aizoides* seeds; (b) *S. cespitosa* seeds; (c) *S. oppositifolia* seeds; (d) *S. stellaris* seeds; (e) *Lychnis flos-cuculi* seeds; (f) *Silene acaulis* seeds; (g) *Melandrium (Silene) album* seeds; (h) *Chenopodium album* seeds; (i) Fern sporangium (*Asplenium septentrionale*); (j) *Juncus articulatus* seeds; and (k) *Luzula multiflora* seed. Scale lines are 1 mm.

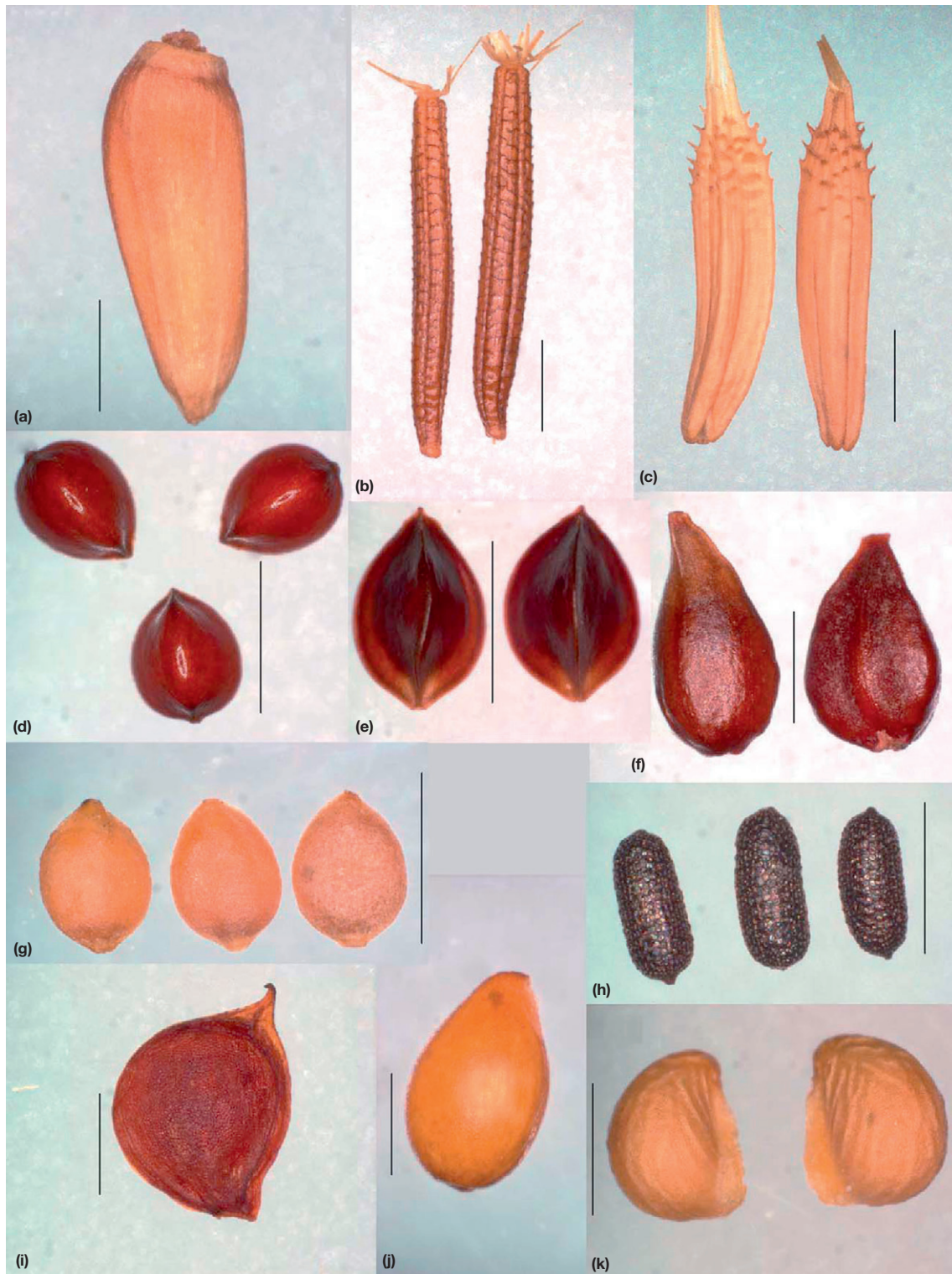


Figure 13 Weeds: (a) *Cirsium arvense* fruit; (b) *Leontodon autumnnalis* fruit; (c) *Taraxacum croceum* fruit; (d) *Rumex acetosella* seeds; (e) *R. acetosa* seeds; (f) *Polygonum aviculare* seeds; (g) *Urtica dioica* fruits; (h) *Hypericum perforatum* seeds; (i) *Ranunculus acris* fruit; (j) *Viola palustris* seed; and (k) *Potentilla erecta* fruits. Scale lines are 1 mm.

Taphonomic processes in the formation of a macrofossil assemblage

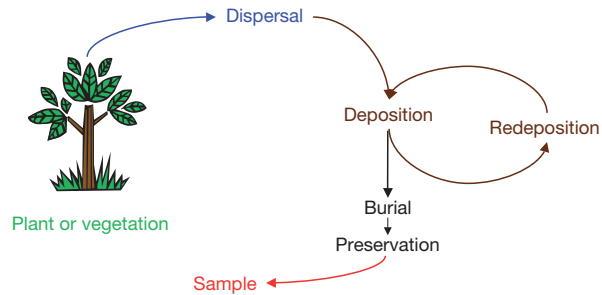


Figure 14 The taphonomic processes involved in the formation of a plant macrofossil assemblage. Each process can be studied individually, or the resulting assemblage can be studied using surface samples relating the assemblage to the modern vegetation.

Plant Macrofossils in Paleocology

Plant macrofossil analysis is rarely conducted in isolation. It is usually combined with pollen analysis, and it can profitably contribute to multiproxy studies. A combination of pollen and macrofossil analyses provides more detailed and realistic interpretations than either alone, each filling gaps left by the other. The articles in this 'macrofossil' subsection provide overviews of the value of macrofossils in vegetation history, in some paleoecological situations where pollen evidence is particularly uninformative, and in other Quaternary applications.

Macrofossils in Vegetation History

Macrofossils are important in the study of vegetation history because they can indicate the local occurrence of taxa with abundant, widely dispersed pollen, such as *Betula* and *Pinus* (Birks and Birks, 2000, 2003). This information could have climatic implications (Birks and Birks, 2003). The identification of bud scales provides records of some deciduous trees (e.g., *Quercus*, *Ulmus*, *Corylus*, *Ostrya/Carpinus*) that otherwise produce few macrofossils (e.g., Eide et al., 2006). Some trees and herbs produce small quantities of fragile or nondescript pollen. Macrofossils can provide more detailed paleoecological records (e.g., *Larix*, *Populus*, Cupressaceae, *Juncus*, *Dryas*, etc.). In addition, macrofossils can frequently distinguish taxa below the level of pollen taxa which are typically identified to the family or genus level. Thus they can provide greater detail about forest and other vegetational successions. For example, many conifer needles can be distinguished to species (e.g., Dunwiddie, 1987; Watts, 1979). *Betula* fruits and catkin scales distinguish dwarf from tree birch and between many tree-birch species. Many Cyperaceae can be identified to species (Figure 10) as well as members of the Poaceae, Caryophyllaceae, Ericaceae, Rosaceae, Saxifragaceae, Apiaceae, Potamogetonaceae, Polygonaceae, Juncaceae, and others (see Figures 3–13). The combination of macrofossil and pollen studies of vegetation history and successions, both aquatic and terrestrial, provides much more information than provided by pollen analysis alone (e.g., Eide et al., 2006).

Situations Where Pollen Is Uninformative

Macrofossils can provide paleoecological insight where pollen alone is uninformative or confusing. Most obvious are treeless situations, such as arctic or alpine tundra vegetation or prairie or steppe grassland. The local pollen production of arctic and alpine tundra vegetation is low. Large amounts of long-distance tree pollen can confuse the local vegetation signal, which macrofossils can clarify (Birks, 1991). In alpine regions, fluctuations of treeline are of special interest, but the pollen signal is confounded by abundant tree pollen from the forest below. Macrofossils can reconstruct the occurrence of alpine vegetation and indicate when trees became locally present. Birks (1993, 2003) showed that large percentages of tree-*Betula* pollen in the Late Glacial of western Norway and Scotland were an artifact caused by low local pollen production and overrepresentation of long-distance pollen. Macrofossils indicated that tree-*Betula* was not present and that the pollen originated from trees growing further south in Europe or from dwarf birch (*Betula nana*). Likewise, the high percentages of tree pollen found in samples from Svalbard must derive from long-distance transport. Birks (1991) used macrofossils to reconstruct the tundra vegetation and to relate the changes in vegetation density and composition to climate change, which had been impossible to deduce from pollen analyses. Pollen influx data may indicate the local presence of trees, but near treeline pollen production may be reduced in the marginal environment. Eide et al. (2006) compared Holocene tree pollen percentages, pollen influx, and macrofossil concentrations along an altitudinal transect up a south Norwegian valley. Pollen influx values of trees present near treeline were much lower than in their main area of distribution, and an estimate of tree presence and abundance could only be obtained from the macrofossil record (Figure 15). Macrofossils and megafossils have been used to document movements in arctic treelines and tree colonization and response to late-glacial climate changes, particularly in Switzerland. The Holocene macrofossil record at Sägistalsee, Switzerland, was used to test hypotheses about the relative impacts of catchment vegetation and climate changes on the paleolimnology of the alpine lake (Lotter and Birks, 2003).

In grassland biomes, Poaceae, *Artemisia*, and Cyperaceae pollen predominate. Lower taxonomic resolution is not feasible. However, macrofossil records can be used to reconstruct Holocene prairie communities in detail and to track the movements of the prairie forest ecotone through time. In arid lands, macrofossil deposits are sparse, but accumulations of plant material in middens by rodents such as packrats are a rich source of paleovegetation information. Phytogeographical changes between the full glacial and the Holocene, and changes in distribution of vegetation types such as juniper communities in response to change in Holocene humidity have been documented. Pollen records from full-glacial Beringia have been interpreted as an extinct grassland biome, 'Mammoth Steppe.' Plant macrofossil analyses from small pond sediments, from vegetation buried by tephra, and from stomach contents or dung of animals preserved in permafrost revealed a mosaic of tundra communities differentiated largely by moisture availability. Thus 'Mammoth Steppe' could be characterized. Its components were found to have analogs with modern vegetation (Guthrie, 2001).

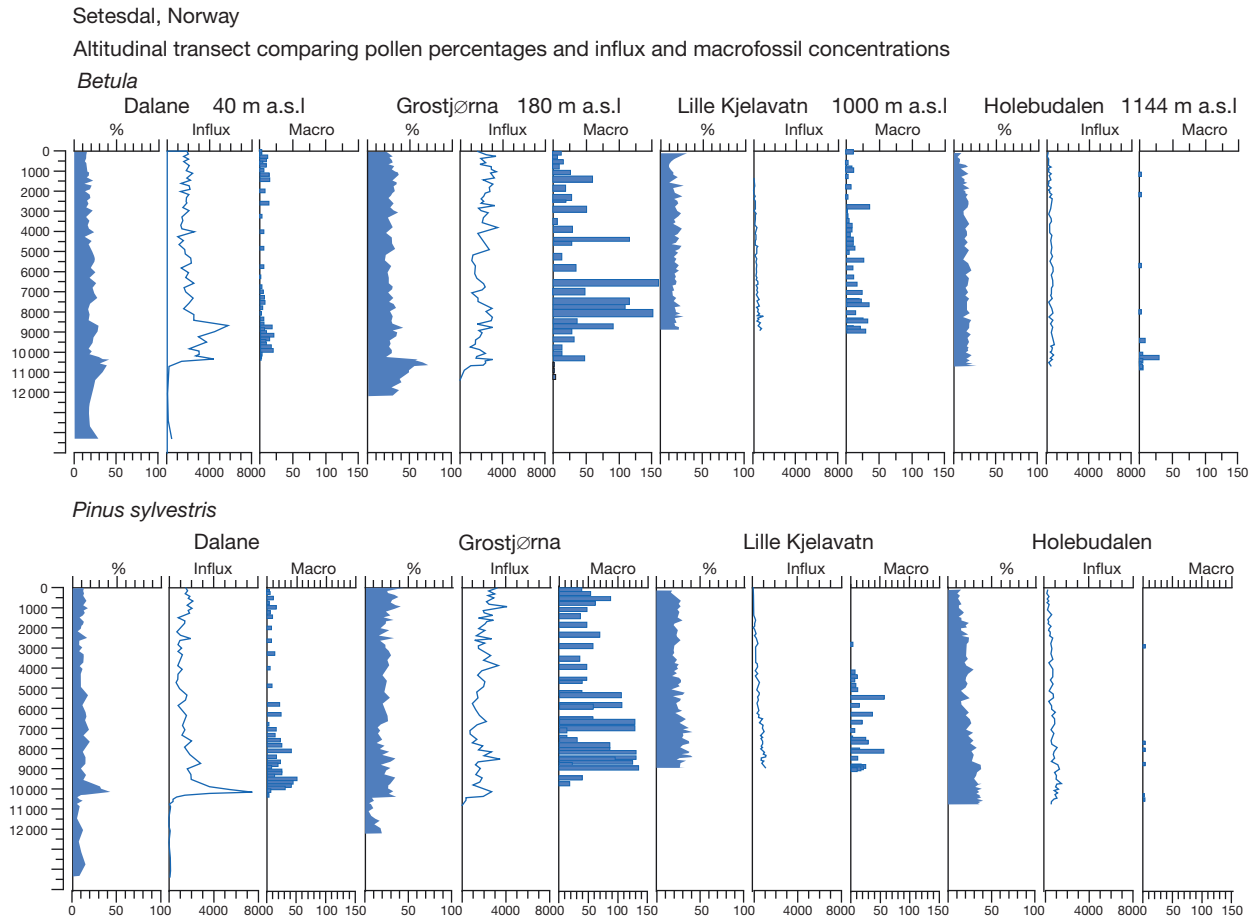


Figure 15 Comparison of four Holocene lake-sediment sequences along an altitudinal transect from 40 m above sea level (a.s.l.) to 1144 m a.s.l. in Setesdal valley, south Norway. Pollen percentages (filled curves), pollen influx (open curves), and plant macrofossil concentrations (histograms) of *Betula* and *Pinus* are plotted against calibrated years BP, using the same x-axis scales for each proxy for all the lakes. Data from Eide W, Birks HH, Bigelow NH, Peglar SM, and Birks HJB (2006) Holocene forest development along the Setesdal valley, southern Norway, reconstructed from macrofossil and pollen evidence. *Vegetation History and Archaeobotany* 15: 65–85, Figure 6, with kind permission of Springer Science and Business Media.

Mire vegetation often contains quantities of bryophytes that are diagnostic habitat indicators. These are infrequently represented by spores, except *Sphagnum*, but mosses are frequently recognizable in the peat and lake sediments. Together with identifications of vegetative plant remains, details of the mire communities can be reconstructed and changes in moisture availability inferred. Aquatic plants are often poorly represented by pollen, either because of low pollen production (e.g., *Potamogeton* spp.) or poorly preserved pollen often adapted for water pollination (e.g., *Najas*, *Ruppia*, *Callitriche*, *Ceratophyllum*, *Zannichellia*, *Elatine*, *Juncus*). Characeae algae can be important constituents of aquatic vegetation, but they do not produce microspores. All these taxa produce readily identifiable, often abundant, macrofossils which allow hydroseres to be tracked through time (e.g., Birks, 2000; Birks et al., 2001). In addition, many marsh taxa within Cyperaceae and Poaceae can be identified by their macrofossils. The importance of macrophytes in aquatic ecosystems is now being recognized, and there are increasing numbers of multiproxy paleolimnological studies involving macrofossil analyses.

Other Applications of Macrofossil Analysis

Macrofossils are necessary for AMS dating of terrestrial plant material. Securely identified macrofossils can be selected at fine temporal resolution. The resulting radiocarbon (^{14}C) chronology can be converted to calendar years, or it can be wiggle-matched against calibrated annual data such as tree-ring chronologies or annually laminated lake sediments thus avoiding radiocarbon plateaux (Björck and Wohlfarth, 2001). Gulliksen et al. (1998) wiggle-matched a high-resolution series of AMS dates on macrofossils and gyttja at Kråkenes, western Norway, and determined the age of the Younger Dryas–Holocene boundary at 11530 ± 40 –60 calendar years before present (BP). By using terrestrial plant material, errors produced by old carbon reservoirs in lake sediments and by noncontemporaneous material such as deep-growing roots in peat can be avoided. When terrestrial and marine fossils are dated from the same marine sediment core, the marine reservoir age and its changes through late-glacial time can be estimated (Bondevik et al., 2006). Terrestrial macrofossils can also provide precise



Figure 16 *Salix herbacea*, female plant in Norway. Photograph by H.J.B. Birks.

^{14}C ages of tephra layers, which, once dated, provide time markers in terrestrial and marine sequences (Birks et al., 1996).

The density of stomata on leaves of some species today is sensitive to atmospheric CO_2 concentration. This can be calibrated and then used to reconstruct past levels and changes in CO_2 concentration by measuring stomatal frequency on fossil leaves of the same species. This technique has highlighted shortcomings in the detail of CO_2 records measured in ice cores.

The Remarkable Versatility of *Salix herbacea*

Salix herbacea leaves are small and leathery and are frequently preserved in cold-climate deposits (Figure 16). *S. herbacea* grows in open treeless vegetation with late snow-lie, in an arctic or alpine climate with plentiful snow. Its leaves are often frequent in full- and late-glacial sediments and can form part of the early succession after the Holocene warming (e.g., Gulliksen et al., 1998). At high altitudes, they can provide a continuous Holocene record. They are often sufficiently abundant to provide material for AMS dating (e.g., Gulliksen et al., 1998). Finally, they are used for reconstructing atmospheric CO_2 changes.

How Will Plant Macrofossil Analysis Develop in the Future?

Macrofossil analysis requires good botanical knowledge of plant morphology, identification, taxonomy, and ecology, and people with these skills are rare and decreasing in numbers (Birks and Birks, 2000). However, the use of plant macrofossil analysis continues to develop. Together with artifacts made of plant material, macrofossils are the basis for paleoethnobotany and archaeobotany, and they will continue in this important role. In the future, more sophisticated studies of vegetative plant material will extend our knowledge, for example from peat and fen deposits and also from preserved animal (and human) bodies. Knowledge of the representation of vegetation and species by macrofossils is slowly increasing, in both aquatic and terrestrial (packrat) locations. More modern representation studies are needed from all types of ecosystems, to aid in the interpretation of past macrofossil records.

Plant macrofossils are being increasingly used in multiproxy studies, usually in conjunction with pollen analysis. One exciting new development is their use in hypothesis testing as at Sägistalsee (Lotter and Birks, 2003). Another new development is the testing of climate-dependent models of vegetation (forest) development through the Holocene driven by independently reconstructed temperature, against the actual forest development reconstructed by macrofossils. Discrepancies helped to differentiate between different agencies of vegetation change, namely climate and human impact (Heiri et al., 2006). Plant macrofossils will probably play a pivotal role in these newly developing modeling approaches. In a multiproxy study, they provide an essential record of vegetation development either in a mire or bog, or in a lake ecosystem (lake plus its catchment).

The role of plant macrofossils in radiocarbon chronology will continue to be essential in Quaternary geology and paleoecology. High-resolution AMS ^{14}C dates around terrestrial tephra horizons will establish these as time markers wherever the tephrae are found, especially in marine cores. Dating of terrestrial and marine macrofossils from the same sediment cores will help to establish the magnitude of the marine reservoir age and how it has varied. This will be enormously useful in correcting the marine ^{14}C chronology especially during the late-glacial climatic transition. The reconstructions of atmospheric CO_2 concentrations from stomatal density on fossil leaf cuticles need to be continued, to establish detailed CO_2 changes during the Late Glacial that are not discerned by ice-core CO_2 records, and also to reconstruct CO_2 levels in previous Quaternary and older geological periods. There have been various attempts to retrieve DNA from fossil plant material, and as DNA extraction techniques improve, fossil DNA assemblages may give information about past vegetation or the diet of extinct animals which is not already apparent from the macrofossils themselves.

Macrofossils from arid and treeless areas, where lake or peat deposits are rare, are perhaps the only means of obtaining a phytogeographic and vegetation history. Plant remains are preserved in rodent middens, alluvial deposits, and also in archaeological contexts in such areas. Macrofossils are also the best means of studying arctic and alpine phytogeography and vegetational development, as the pollen record is fraught with difficulties. Knowledge about these areas of the world, both about the natural vegetation development and also the impact of human activities on it, is steadily increasing through the application of macrofossil analysis. Macrofossil analyses, usually in the context of multidisciplinary studies, are important in evaluating human impact on lake ecosystems, revealing the history of acidification, eutrophication, and pollution and their effects on macrophyte communities and hence on whole aquatic ecosystems. The astonishing rapidity of aquatic ecosystem responses to human impact, especially in arid lands, was demonstrated through macrofossil analyses, and it is hoped that this signal has been received by organizations concerned with the management of hydrological systems and water sustainability.

See also: **Chironomid Records:** Chironomid Overview.
Paleolimnology: Cladocera. **Plant Macrofossil Methods and Studies:** CO_2 Reconstruction from Fossil Leaves; Megafossils; Mire and Peat Macros; Paleolimnological Applications; Rodent Middens;

Surface Samples, Taphonomy, Representation; Treeline Studies; Use in Environmental Archaeology; Validation of Pollen Studies. **Plant Macrofossil Records:** Arctic Eurasia; Arctic North America; Greenland; Holocene North America; Late Glacial Multidisciplinary Studies. **Pollen Methods and Studies:** Archaeological Applications. **Radiocarbon Dating:** ^{14}C of Plant Macrofossils.

References

- Baales M, Jörns O, Street M, Bittmann F, Weninger B, and Wiethold J (2002) Impact of the late glacial eruption of the Laacher See Volcano, Central Rhineland, Germany. *Quaternary Research* 58: 273–288.
- Birks HH (1973) Modern macrofossil assemblages in lake sediments in Minnesota. In: Birks HJB and West RG (eds.) *Quaternary Plant Ecology*, pp. 173–189. Oxford: Blackwell.
- Birks HH (1991) Holocene vegetation history and climatic change in west Spitsbergen – Plant macrofossils from Skardtjørna, an arctic lake. *The Holocene* 1: 209–218.
- Birks HH (1993) The importance of plant macrofossils in late-glacial climatic reconstructions: An example from western Norway. *Quaternary Science Reviews* 12: 719–726.
- Birks HH (2000) Aquatic macrophyte vegetation development in Kråkenes Lake, western Norway, during the late-glacial and early-Holocene. *Journal of Paleolimnology* 23: 7–19.
- Birks HH (2001) Plant macrofossils. In: Smol JP, Birks HJB, and Last WM (eds.) *Tracking Environmental Change Using Lake Sediments. Terrestrial, Algal, and Siliceous Indicators*, vol. 3, pp. 49–74. Dordrecht, The Netherlands: Kluwer Academic Publishers.
- Birks HH (2003) The importance of plant macrofossils in the reconstruction of Lateglacial vegetation and climate: Examples from Scotland, western Norway, and Minnesota, USA. *Quaternary Science Reviews* 22: 453–473.
- Birks HJB and Birks HH (1980) *Quaternary Palaeoecology*, ch. 5. London: Edward Arnold. (reprinted 2005, Blackburn Press, New Jersey, USA).
- Birks HH and Birks HJB (2000) Future uses of pollen analysis must include plant macrofossils. *Journal of Biogeography* 27: 31–35.
- Birks HH, Battarbee RW, and Birks HJB (2000) The development of the aquatic ecosystem at Kråkenes Lake, western Norway, during the late-glacial and early-Holocene – A synthesis. *Journal of Paleolimnology* 23: 91–114.
- Birks HH and Birks HJB (2003) Reconstructing Holocene climate from pollen and plant macrofossils. In: Mackay A, Battarbee R, Birks J, and Oldfield F (eds.) *Global Change in the Holocene*, pp. 342–357. London: Arnold.
- Birks HH, Gulliksen S, Hafliadason H, Mangerud J, and Possnert G (1996) New radiocarbon dates for the Vedde Ash and the Saksunarvatn Ash from western Norway. *Quaternary Research* 45: 119–127.
- Birks HH, Peglar SM, Boomer I, Flower RJ, and Ramdani M (2001) Palaeolimnological responses of nine North African lakes in the CASSARINA Project to recent environmental changes and human impact detected by plant macrofossil, pollen, and faunal analyses. *Aquatic Ecology* 35: 405–430.
- Björck S and Wohlfarth B (2001) ^{14}C chronostratigraphic techniques in paleolimnology. In: Last WM and Smol JP (eds.) *Tracking Environmental Change using Lake Sediments. Basin Analysis, Coring, and Chronological Techniques*, vol. 1, pp. 205–245. Dordrecht, The Netherlands: Kluwer Academic Publishers.
- Bondevik S, Mangerud J, Birks HH, Gulliksen S, and Reimer P (2006) Changes in North Atlantic radiocarbon reservoir ages during the Allerød and Younger Dryas. *Science* 312: 1514–1517.
- Chaney RW (1924) Quantitative studies of the Bridge Creek Flora. *American Journal of Science* 8: 127–144.
- Dunwiddie PW (1987) Macrofossil and pollen representation of coniferous trees in modern sediments from Washington. *Ecology* 68: 1–11.
- Eide W, Birks HH, Bigelow NH, Peglar SM, and Birks HJB (2006) Holocene forest development along the Setesdal valley, southern Norway, reconstructed from macrofossil and pollen evidence. *Vegetation History and Archaeobotany* 15: 65–85.
- Glob PV (1971) *The Bog People*. London: Paladin.
- Goetcheus VG and Birks HH (2001) Full-glacial upland tundra vegetation preserved under tephra in Beringia National Park, Seward Peninsula, Alaska. *Quaternary Science Reviews* 20: 135–147.
- Gulliksen S, Birks HH, Possnert G, and Mangerud J (1998) A calendar age estimate of the Younger Dryas – Holocene boundary at Kråkenes, western Norway. *The Holocene* 8: 249–259.
- Guthrie RD (2001) Origin and causes of the mammoth steppe: A story of cloud cover, woolly mammoth tooth pits, buckles, and inside-out Beringia. *Quaternary Science Reviews* 20: 549–574.
- Heiri C, Bugmann H, Tinner W, Heiri O, and Lishke H (2006) A model-based reconstruction of Holocene treeline dynamics in the Central Swiss Alps. *Journal of Ecology* 94: 206–216.
- Holyoak DT (1984) Taphonomy of prospective plant macrofossils in a river catchment on Spitsbergen. *New Phytologist* 98: 405–423.
- Krzywinski K and Soltvedt EC (1988) A medieval brewery (1200–1450) at Bryggen, Bergen. *The Bryggen Papers*, pp. 1–68. Norway: University of Bergen. Supplementary Series 3.
- Lotter AF and Birks HJB (2003) The Holocene palaeolimnology of Sägistalsee and its environmental history – A synthesis. *Journal of Paleolimnology* 30: 333–342.
- Moore DM (1978) Post-glacial vegetation in the South Patagonian territory of the giant ground sloth, Mylodon. *Botanical Journal of the Linnean Society* 77: 177–202.
- Spicer RA (1989) The formation and interpretation of plant fossil assemblages. *Advances in Botanical Research* 16: 95–191.
- Stead IM, Bourke JB, and Brothwell D (1986) *Lindow Man: The Body in the Bog*. London: British Museum Publications.
- Ukrainseva VV (1993) *Vegetation Cover and Environment of the 'Mammoth Epoch' in Siberia*. Hot Springs, SD: The Mammoth Site of Hot Springs.
- Watts WA (1979) Late Quaternary vegetation of central Appalachia and the New Jersey coastal plain. *Ecological Monographs* 49: 427–469.
- Watts WA and Winter TC (1966) Plant macrofossils from Kirchner Marsh, Minnesota – A paleoecological study. *Geological Society of America Bulletin* 77: 1339–1360.
- Zazula GD, Froese DG, Schweger CE, et al. (2003) Ice-age steppe vegetation in east Beringia. *Nature* 423: 603.

General Reading About Plant Macrofossils

- Birks HH (1980) Plant macrofossils in Quaternary lake sediments. *Archiv für Hydrobiologie* 15: 1–60.
- Dickson CA (1970) The study of plant macrofossils in British Quaternary deposits. In: Walker D and West RG (eds.) *Studies in the Vegetational History of the British Isles*, pp. 233–254. Cambridge: Cambridge University Press.
- Warner BG (1988) Methods in Quaternary Ecology #3. Plant macrofossils. *Geoscience Canada* 15: 121–129.
- Wasylikowa K (1986) Analysis of fossil fruits and seeds. In: Berglund BE (ed.) *Handbook of Holocene Palaeoecology and Palaeohydrology*, pp. 571–590. Chichester: Wiley.
- Watts WA (1978) Plant macrofossils and Quaternary paleoecology. In: Walker D and Guppy JC (eds.) *Biology and Quaternary Environments*, pp. 53–67. Canberra: Australian Academy of Science.

Macrofossil Identification Manuals

- Anderberg A-L (1994) *Atlas of Seeds. Part 4. Resedaceae–Umbelliferae*, 281 pp. Stockholm: Swedish Museum of Natural History.
- Beijerinck W (1976) *Zadenatlas der Nederlandsche Flora*, 316 pp. Amsterdam: Backhuys & Meesters (in Dutch).
- Berggren G (1964) *Atlas of Seeds. Part 2. Cyperaceae*, 68 pp. Stockholm: Swedish Museum of Natural History.
- Berggren G (1981) *Atlas of Seeds. Part 3. Salicaceae–Cruciferae*, 259 pp. Stockholm: Swedish Museum of Natural History.
- Bertsch K (1941) *Fruchte und Samen. Ein Bestimmungsbuch zur Pflanzendecke der vorgeschichtlichen Zeit*. Stuttgart: Ferdinand Enke.
- Davis LW (1993) *Weed Seeds of the Great Plains: A Handbook for Identification*. Kansas, USA: University Press.
- Elias S and Pollak O (1987) Photographic atlas and key to windblown seeds of alpine plants from Niwot Ridge, Front Range, Colorado, USA. INSTAAR Occasional Paper 45, 28 pp.
- Gale R and Cutler D (2000) *Plants in Archaeology: Identification Manual of Artifacts of Plant Origin from Europe and the Mediterranean to c. 1500*. Otley: Westbury and Royal Botanic Gardens, Kew. ISBN 1-84103-002-3 ISBN.
- Grigas A (1986) Lietuvos augalu vaisiai ir seklos (Fruits and seeds of Lithuanian plants). Lithuanian: Vilnius 'Mokslas'.
- Katz NJ, Katz SV, and Kipiani MG (1965) *Atlas and Keys of Fruits and Seeds Occurring in the Quaternary Deposits of the USSR*, 365 pp. Moscow: Nauka (in Russian).

- Martin JB and Barkley WD (1961) *Seed Identification Manual*. Berkeley: California University Press. Reprinted: 2000 by The Blackburn Press, ISBN 1-930665-03-2.
- Montgomery FH (1977) *Seeds and Fruits of Plants in Eastern Canada and Northeastern United States*. Toronto: University of Toronto Press.
- Schoch WH, Pawlik B, and Schweingruber FH (1988) *Botanische makroreste*. Bern and Stuttgart: P. Haupt. [German, English, French].
- USDA (1948) *Woody-Plant Seed Manual*. Washington, DC: Forest Service, US Department of Agriculture Miscellaneous. [Publication 654].
- Webb CJ and Simpson MJA (2001) *Seeds of New Zealand Gymnosperms and Dicotyledons*. Christchurch, NZ: Manuka Press.

Papers on Macrofossil Identification: Seeds and Fruits

- Aalto M (1970) Potamogetonaceae fruits. 1. Recent and subfossil endocarps of Fennoscandian species. *Acta Botanica Fennica* 88: 1–55.
- Berggren G (1974) Seed morphology of some Epilobium species in Scandinavia. *Svensk Botanisk Tidskrift* 68: 164–169.
- Bright RC and Woo R (1969) Coating seeds with Ammonium chloride: A technique for better photographs. *Turtox News* 47: 226–229.
- Cappers RTJ (1993) The identification of potamogetonaceae fruits found in The Netherlands. *Acta Botanica Neerlandica* 42: 447–460.
- Chang ER, Dickinson TA, and Jefferies RL (2000) Seed flora of La P  rou Bay, Manitoba, Canada: A DELTA database of morphological and ecological characters. *Canadian Journal of Botany* 78: 481–496.
- Conolly AP (1976) The use of the electron microscope for the identification of seeds, with special reference to Saxifraga, and Papaver. *Folia Quaternaria* 47: 29–32.
- Fowler K and Stennett-Willson J (1978) Sporoderm architecture in modern Azolla. *Fern Gazette* 11: 405–412.
- Grosse-Brauckmann G and Streitz B (1992) Pflanzliche Makrofossilien mitteleurop  ischer Torfe III. Fr  chte, Samen und einige Gewebe (Fotos von fossilen Pflanzenresten). *Telma* 22: 53–102.
- Haas JN (1994) First identification key for charophyte oospores from central Europe. *European Journal of Phycology* 29: 227–235.
- Huckerby E, Marchant R, and Oldfield F (1972) Identification of fossil seeds of Erica and Calluna by scanning electron microscopy. *New Phytologist* 71: 387–392.
- Jacom  t S, Brombacher C, and Martin D (1979–1988) Arch  obotanik am Z  richsee. In: Punchakunnel P, Wagner C, and Felice N (eds.) *Berichter der Z  rcher Denkmalpflege*, Monographie 7.
- Jessen K (1955) Key to subfossil Potamogeton. *Botanisk Tidskrift* 52: 1–7.
- Kaplan K (1981) Embryologische, pollen- und samenmorphologische untersuchungen zur systematik von Saxifraga (Saxifragaceae). In: *Bibliotheca Botanica*, vol. 134, 56 pp.
- Krause W (1986) Zur Bestimmungsm  glichkeit subfossiler Characeen-Oosporen an Beispielen aus Schweizer Seen. *Vierteljahrsschrift der Naturforschenden Gesellschaft in Z  rich* 131: 295–313.
- K  rber-Grohne U (1964) Bestimmungsschl  ssel f  r subfossile Juncus-samen und Gramineen-Fr  chte. *Probleme der K  stenerforschung im S  dlichen Nordseegebiet*, vol. 7. Hildesheim: August Lax.
- Kuzniewska E (1974) Studia systematyczne nad nasionami srodkowo – i palnocyfocnoeuropejskich gatunkow rodzaju Saxifraga L. *Prace Opolskiego Towarzystwa Przyjaciol Nauk Wydzial* 111: 1–142.
- Martin AC (1951) Identifying pondweed seeds eaten by ducks. *Journal of Wildlife Management* 15: 253–258.
- Nilsson    and Hjelmquist H (1967) Studies on the nutlet structure of South Scandinavian species of Carex. *Botaniska Notiser* 120: 460–485.
- d'Olivat HJP and Pals JP (1974) *Determination Tables and Descriptions of the Seeds and Fruits of the Ranunculaceae Occurring in the Netherlands*. Amsterdam: University of Amsterdam. I. P. P. Publicatie 149.
- van Dinter M and Birks HH (1996) Distinguishing fossil Betula nana and B. pubescens using their wingless fruits: Implications for the late-glacial vegetational history of western Norway. *Vegetation History and Archaeobotany* 5: 229–240.

Mosses: Mosses Are Identified Using Modern Moss Floras

- Dickson JH (1973) *Bryophytes of the Pleistocene*. Cambridge: Cambridge University Press.
- Janssens JA (1990) Methods in Quaternary Ecology 11. Bryophytes. In: *Geosciences Canada*, vol. 17, pp. 13–24.

Wood

- Barefoot AC and Hankins FW (1982) *Identification of Modern and Tertiary Woods*. Oxford: Clarendon Press.
- Grosser D (1977) *Die H  lzer Mitteleuropas*. Berlin: Springer.
- Jane FW (1956) *The Structure of Wood*. A. and C. Black. London.
- Mork E (1966) *Vedanatomi. With an Identification Key for Microscopic Wood Sections*, 2nd edn. Oslo: Forlag Johan Grundt Tanum (in Norwegian).
- Schweingruber FH (1990) *Anatomie europ  ischer H  lzer (Anatomy of European Woods)*. Bern and Stuttgart: Verlag Paul Hapt (in German and English).

Cuticles – Bud Scales – Leaves

- Grosse-Brauckmann G (1972)   ber pflanzliche makrofossilien mitteleurop  ischer Torfe – I. Gewebereste Krautiger Pflanzen und ihre Merkmale. *Telma* 2: 19–55 (in German).
- Grosse-Brauckmann G (1973)   ber pflanzliche makrofossilien mitteleurop  ischer Torfe – II. Weitere Reste (Fr  chte und Samen, Moose, U.A.) und ihre Bestimmungsm  glichkeiten. *Telma* 4: 51–117.
- Grosse-Brauckmann G (1986) Analysis of vegetative plant macrofossils. In: Berglund BE (ed.) *Holocene Handbook of Palaeoecology and Palaeohydrology*, pp. 591–618. Chichester: Wiley.
- Jones TP and Rowe NP (eds.) (1999) *Fossil Plants and Spores: Modern Techniques*, 396 pp. London: Geological Society of London (a collection of papers).
- Katz NJ, Katz SV, and Skovejeva EI (1977) *Atlas of Plant Remains in Peat-Soil*. Moscow: Nedra (in Russian).
- Martinetto E (2003) Leaves of terrestrial plants from the Pliocene shallow marine and transitional deposits of Asti. *Bollettino della Soci  t   Paleontologica Italiana* 42: 75–111 (Piedmont, NW Italy).
- Palmer PG (1976) Grass cuticles: A new palaeoecological tool for East African lake sediments. *Canadian Journal of Botany* 54: 1725–1734.
- Tomlinson P (1985) An aid to the identification of fossil buds, bud-scales and catkin-bracts of British trees and shrubs. *Circaea* 3: 45–130.
- Westerkamp C and Demmelmeyer H (1997) *Blattoberfl  chen mitteleurop  ischer laubgeh  lze: Leaf Surfaces of Central European Woody Plants. Atlas and Keys*. Berlin, Stuttgart: G. Borntraeger.

Relevant Websites

- http://www.ars.usda.gov/main/site_main.htm – Center for Genetic Resources Preservation.
- http://www.usgs.nau.edu/global_change/macrodigitalibrary.asp – Macrodigital library, southwest USA.
- <http://www.rbgkew.org.uk/msbp> – Millennium Seed Bank, Wakehurst Place, Kew.
- <http://www.uwyo.edu/Botany/NAPMD/index.htm> – North American Plant macrofossil database.
- <http://www.ncdc.noaa.gov/paleo/plantmacros.html> – WDC for Paleoclimatology Plant Macrofossils.

PLANT MACROFOSSIL METHODS AND STUDIES

Contents

CO₂ Reconstruction from Fossil Leaves

Megafossils

Dendroarchaeology

Mire and Peat Macros

Paleolimnological Applications

Rodent Middens

Surface Samples, Taphonomy, Representation

Treeline Studies

Use in Environmental Archaeology

Validation of Pollen Studies

CO₂ Reconstruction from Fossil Leaves

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Introduction

The global carbon cycle is closely linked to the Earth's climate system by a large number of processes acting over a wide range of temporal scales. Changes in the operation of the carbon cycle can be expected to result in variations in the atmospheric concentration of carbon dioxide (CO₂) and, because CO₂ is a major greenhouse gas, such variations are likely to influence global temperatures. At the same time, atmospheric CO₂ levels are to a large extent determined by climatic conditions, which makes it difficult to separate cause and effect within the coupled carbon cycle–climate system.

Documenting interactions over short (submillennial) timescales is important to improve our understanding of the role of atmospheric CO₂ in Quaternary climate change. This requires CO₂ records at very high resolution, but detection of such rapid variations in ice cores is limited by inherent smoothing resulting from molecular diffusion during air enclosure. The degree to which original CO₂ amplitudes are attenuated is mainly controlled by temperature and accumulation rate at the drilling site, and conditions in Antarctica are generally unfavorable for detection of CO₂ fluctuations lasting less than one or two centuries. Unfortunately, ice-core CO₂ records from the higher-accumulation Greenland sites are not considered reliable because of artifacts related to elevated impurity (mostly carbonate) levels in the ice.

An alternative approach, with the potential to provide CO₂ reconstructions at the necessary resolution, is based on stomatal frequency analysis of leaves of terrestrial plants preserved in Quaternary lake sediments and peat. In recent years, this

method has provided a number of CO₂ reconstructions with trends similar to those shown by ice-core records but with clear evidence for centennial-scale variability and larger amplitude of change. Many of the detected CO₂ fluctuations are temporally related to climate events of global significance, which points to the importance of carbon cycle–climate interaction also on submillennial timescales during the Quaternary.

Basis of the Method

The outer surface of vascular plants is covered by an impermeable layer called the cuticle, and gas exchange with the atmosphere is primarily restricted to small pores – stomata – that are primarily located on leaves ([Figure 1](#)). Through the control of stomatal aperture and number of stomata that develop on the leaf surface, this arrangement allows the plant to maximize CO₂ supply for photosynthesis and minimize water loss from transpiration, that is, to optimize water use efficiency. Stomata form early in leaf development ([Tichá, 1982](#)), and their initiation is largely determined by ambient CO₂ conditions. The relationship between atmospheric CO₂ concentration and stomatal frequency is an inverse one; more leaf surface (epidermal) cells develop into stomata at low concentrations and vice versa. Although the mechanism governing stomatal initiation is not fully understood, stomatal frequency represents an important physiological plant response that is detectable in the fossil record.

Adaptation of naturally growing plants to changing CO₂ levels was first noticed by [Woodward \(1987\)](#) for herbarium

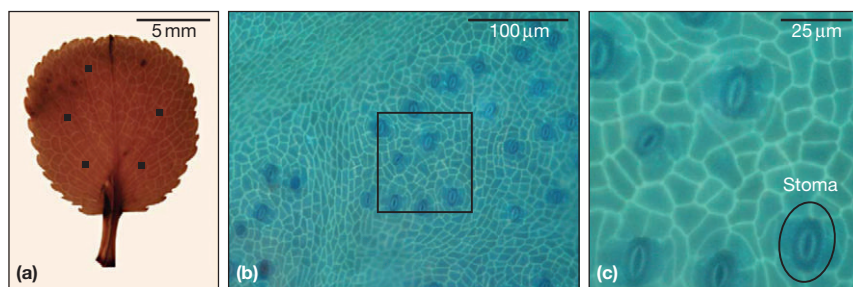


Figure 1 (a) Leaf of *Salix herbacea*. The five black squares distributed over the leaf surface indicate suitable field areas for counting of stomata and epidermal cells. (b) Field area of a *S. herbacea* leaf seen through an epifluorescence microscope. The upper left corner of the picture where no stomata are seen covers part of a major vein. The black square indicates an interveinal area suitable for counting. (c) Close-up of the black square in (b). Four entire stomata, and another three that cross the border of the field area, are seen together with epidermal cells.

leaves of temperate tree species collected since the mid-eighteenth century, and a CO₂ response has later been shown for more than a hundred species, both in herbarium studies and controlled experiments (Royer, 2001). Basically, each new leaf that develops on a plant becomes adjusted to the ambient conditions, which means that stomatal frequency analysis of fossil leaves has the potential to reconstruct annual CO₂ changes. In practice, however, very few Quaternary leaf sequences have the necessary resolution to allow CO₂ reconstruction on subdecadal timescales.

Stomatal frequency data may be quantified in two different ways. Stomatal density (SD) is simply a measure of the number of stomata per unit leaf area (expressed in mm⁻²). Stomatal index (SI) quantifies the proportion of the total number of leaf surface cells (stomata and epidermal cells) that are stomata, where a stoma consists of the stomatal pore and two flanking guard cells (Figure 1). The SI, which is expressed as a percentage, was introduced by Salisbury (1927) and is often defined as $SI = SD / (SD + ED) \times 100$, where ED is epidermal cell density (mm⁻²). Because SI is a proportional measure, it is unaffected by leaf expansion and therefore the preferred parameter for CO₂ reconstruction. In comparison, SD is influenced both by stomatal initiation and leaf expansion, which makes it sensitive to environmental factors other than CO₂, particularly light, temperature, and water conditions. Many experiments and field studies have shown a SD response to these parameters, while there are few indications of an influence on SI (Beerling, 1999; Royer, 2001; Wagner et al., 2004). Consequently, CO₂ reconstructions based on SD are considered less secure.

Outline of the Method

Site Selection

The prime requisite for a site to be used in a stomatal frequency study is, of course, that it should show a high concentration of fossil leaves. Such sites are hard to find, and this is a major limiting factor. As for plant macrofossil studies in general, it is difficult to predict if a site contains an abundant leaf record, but some recommendations for the selection of suitable lake sites can be made (Birks, 2001).

Because the CO₂ response is species specific (Rundgren and Björck, 2003), it is desirable to find leaf sequences with

continuous presence of a single species over long periods of time. When two or more species are combined to extend a reconstruction, the individual leaf records should preferably overlap in some interval (Rundgren and Björck, 2003; Rundgren et al., 2005). Although it is possible to estimate CO₂ concentrations from single leaf occurrences, precision increases significantly with the number of leaves per sample. Ideally, three to five leaves of a species should be available from each level. *Betula pendula*, *Betula pubescens*, *Salix herbacea*, *Quercus robur*, and *Quercus petraea* are among the species that have most frequently been used for CO₂ reconstruction, which partly reflects their relative abundance and good preservation potential in Quaternary lake sediments and peat.

Most CO₂ reconstructions have been based on lake sediments, but leaves extracted from bog and swamp peat and leaf litter have also been used. So far, stomatal frequency studies have predominantly been conducted in temperate and Arctic environments. Low-latitude areas are, however, receiving increased attention (Wagner et al., 2005), partly because their generally more stable climatic conditions during glacial–interglacial cycles imply a potential for long and continuous leaf sequences.

Sampling

The strategy for sampling is consistent with plant macrofossil studies in general and depends on the type of site and source material (Birks, 2001). When coring in lake sediments and peat, several parallel cores may be needed to obtain sufficient leaf material. Cutting of cores, especially at intervals of <1 cm, must be done carefully to minimize destruction of leaves.

Sample Preparation

Samples are processed according to standard methods for plant macrofossil analysis (e.g., Birks, 2001). If treatment with sodium hydroxide (NaOH) is needed, precautions should be taken that there is no damage to the leaves, and tests should be performed to design a suitable treatment before applying it to all samples (Rundgren et al., 2005). A sieve with mesh size of 500 μm is sufficient to catch leaves and leaf fragments. Any fragment passing through this mesh size will be too small for stomatal frequency analysis. Leaves are generally delicate structures and require gentle treatment in the sieving process and during subsequent handling.

Identification

Whole leaves and large fragments are usually identifiable during sorting of the sieving residue using a stereomicroscope with the aid of standard floras and a reference collection. Small leaf fragments do, however, often require closer examination and may be identified later at the counting stage based on cuticular characteristics. Cuticle examination may also support identification in ordinary plant macrofossil studies. If not immediately subjected to further analysis, leaves should be stored in water in a cool place or frozen to prevent bacterial and fungal growth over a long period.

Leaf Preparation

If an epifluorescence microscope is to be used for counting, no further preparation is needed, except cleaning leaf surfaces from any remaining mud with a brush. After that, each leaf fragment to be counted may be placed together with glycerine on a microscope slide and covered by a cover slip. If a normal light source is used, cuticles need to be isolated from the fossil leaves. Areas of about 0.5 cm × 0.5 cm are cut out with a scalpel and bleached in 4% NaHClO₂ solution to remove the mesophyll. This may take from a few minutes up to many hours. The remaining cuticles are separated, stained with safranin, and mounted in glycerine jelly on microscope slides.

Counting

Counting of stomata and epidermal cells is conducted using epifluorescence or transmitted light microscopy. In the first method, fluorescent light reaches the leaf from above through the objective, and excitation highlights cell walls and other structures on its surface. In the second method, normal light from below passes through the isolated and stained cuticle and makes its morphological features visible.

The size of the field area used for counting varies among species and is optimized with regard to cuticular morphology, in particular cell size. A larger field of view results in more representative values but, in practice, field size is limited by the distribution of veins on the leaf surface (Figure 1). Field area size should be selected so that, in general, at least five stomata (including two flanking guard cells) and 50 epidermal cells are visible, and the same size should be used throughout the study. Typically, field areas of 0.03–0.06 mm² and magnifications of 200–600× are used.

Stomata and epidermal cells can be tallied directly at the microscope using, for example, a hand counter. In this case, a graticule in one of the eyepieces is normally used to define the field area and facilitate counting. Alternatively, if a camera is attached to the microscope and linked to a computer, counting may be made (immediately or at a later time) with the aid of image analysis software. Although slightly more time consuming, this arrangement enhances counting accuracy and allows images of all counted fields to be conveniently archived for later study. Irrespective of counting procedure, stomata and epidermal cells are usually counted within a rectangular or quadratic area, and standardized criteria for how to deal with stomata (including two flanking guard

cells) and epidermal cells along the field area margin are needed. An appropriate approach is that described by Poole and Kürschner (1999), where all stomata and epidermal cells crossing two predefined sides of the field area and the corner between them are included in the count, while those crossing the two remaining sides and the three remaining corners are not (Figure 1).

To minimize the effect of intraleaf variability, counts should be made from several field areas distributed over the leaf surface and used to calculate leaf SI and SD mean values. These areas should, however, not be placed along margins, or near the leaf base or tip, where stomatal frequency values often are found to be anomalous (Poole et al., 1996; Tichá, 1982; Figure 1). To minimize variability caused by the presence of veins that do not carry stomata, counting should also be restricted to interveinal areas. The specific number of counts needed to produce reliable mean values varies from species to species and should be determined in a pilot study (Poole and Kürschner, 1999). This may be done through successive calculation of the mean as more counts are added. Fluctuation of the mean is typically found to be within acceptable limits after 7–10 counts per leaf. The number of counts needed is partly dependent on stomatal arrangement, with hypostomatous species (having stomata only on the lower leaf surface) normally found to require fewer counts than amphistomatous species (having stomata on both surfaces). To account for interleaf variability, sample SI and SD mean values are calculated based on leaf (or leaf fragment) mean values and used in CO₂ calibration. It should, however, be noted that it is not always possible to determine if small-sized fragments are derived from separate leaves.

The counting procedure described above only applies to broad-leaved angiosperm species. Gymnosperms exhibit a different mode of leaf development and stomatal patterning, which requires alternative approaches. Different parameters have been used to quantify stomatal frequency in conifer needles, including SD and number of stomatal rows. In contrast to the situation in broad-leaved species, stomatal initiation is usually more accurately reflected by SD rather than SI in conifers, but the number of stomata per millimeter of needle length is considered to be the most sensitive parameter (Kouwenberg et al., 2003). The often weak CO₂ responses found in earlier studies of conifer species (Royer, 2001) may partly be due to inappropriate quantification of stomatal frequency.

Calibration

Atmospheric CO₂ concentration at the time of growth is reconstructed through calibration of stomatal frequency data based on a modern training set (Figure 2). Although stomatal responses generally are species specific, similarities in CO₂ response may allow reconstruction with training sets including closely related species (Jessen et al., 2005; Rundgren et al., 2005; Wagner et al., 1999). Calibration datasets are generally obtained from modern or herbarium leaves collected from plants growing in natural environments, or a combination of both. Also leaves from well-dated shallow peat cores are sometimes used (Kouwenberg et al., 2005; Wagner et al., 1999).

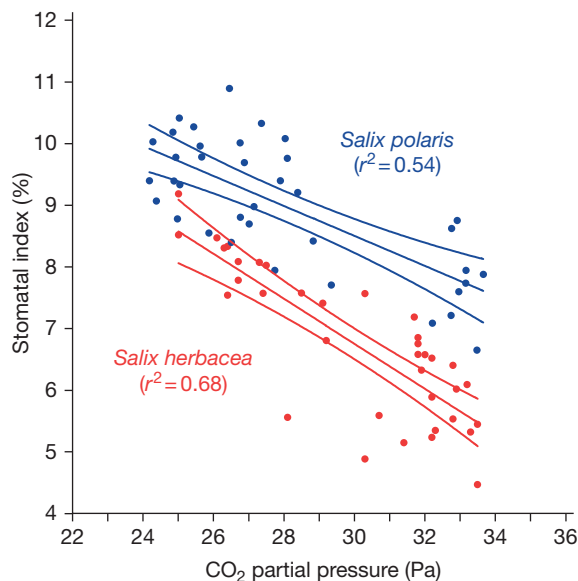


Figure 2 Examples of modern training sets used to reconstruct past CO₂ levels from fossil stomatal frequency data. The calibration data shown are based on a combination of herbarium material (dating back to 1840) and modern leaves collected along altitudinal gradients and were applied to stomatal index (SI) data obtained from *S. herbacea* and *Salix polaris* leaves preserved in Swedish late-glacial and Holocene lake sediments (Rundgren and Beerling, 1999; Rundgren and Björck, 2003). Lines through the data represent linear regressions (with 95% confidence intervals). Adapted from Rundgren M and Björck S (2003) Lateglacial and early Holocene variations in atmospheric CO₂ concentration indicated by high-resolution stomatal index data. *Earth and Planetary Science Letters* 213: 191–204, with permission.

Leaves from plants grown in controlled CO₂ environments are, however, usually discarded because of concerns that stomatal responses to the often abrupt and high-amplitude CO₂ changes applied in experiments are unrepresentative of those recorded by fossil leaves, which developed under natural conditions.

The use of modern leaves in calibration is based on the observation that plants respond to changes in CO₂ partial pressure rather than CO₂ concentration (Woodward and Bazzaz, 1988). Because CO₂ partial pressure decreases with altitude, as a consequence of lowered air pressure, leaves collected along an altitudinal gradient provide a series of calibration data points. The herbarium approach takes advantage of the marked industrial CO₂ increase documented over the last two centuries from atmospheric measurements and high-resolution ice cores. With information about the year (and altitude) of collection, the CO₂ conditions experienced by the plant represented by the herbarium specimen may be calculated.

To capture the nature of the response and minimize error estimates on reconstructed CO₂ values, modern leaves should be selected to cover a CO₂ gradient as wide as possible. This also reduces the need for extrapolation in the calibration process. Because detailed CO₂ data are only available for the last few hundred years, however, reconstruction of very low (e.g., glacial) CO₂ values will inevitably involve extrapolation. Calibration datasets applied in CO₂ reconstruction generally show r^2 (coefficient of determination) values, as determined by

linear regression, in the range 0.5–0.8. In theory, however, any stomatal frequency response to CO₂ is expected to follow a sigmoidal rather than a linear function (Kürschner et al., 1997). This is because leaves may neither lack nor attain an infinite number of stomata. Although there are indications that some species are currently approaching, or already have reached, their response limit (Kürschner et al., 1997; Wagner et al., 2004), most modern training sets are adequately described by linear models (Figure 2). Calibration datasets must, of course, be produced following the same analytical protocol used to obtain fossil data. Consequently, to enable a calibration dataset to be applied in other fossil studies, it is important to provide details about how it was produced.

Calibration of stomatal frequency data has been performed both by inverse and classical (linear) regression. In some instances, stomatal frequency and CO₂ data have been log-transformed prior to calibration in order to account for the partial nonlinear CO₂ response indicated for some species at the high CO₂ concentrations experienced in recent decades (Wagner et al., 2004, 2005). Calibration also involves calculation of confidence intervals for the obtained CO₂ estimates. Both a high r^2 value for the modern training set used, and good replication of fossil data (many leaves counted per sample), contributes to produce narrow intervals. If fossil leaves are derived from a high-altitude site where partial pressure is significantly lower than at sea level, all reconstructed CO₂ values need to be adjusted for this pressure difference before they are converted to concentrations.

Presentation of Results

CO₂ estimates should be displayed together with their calculated confidence intervals. In high-resolution studies, some degree of smoothing is often applied to raw sample values in order to account for the inherent scatter in biological systems and to illustrate more realistic rates of change. When new indicator species are introduced, the calibration data used should be presented graphically.

Overview of CO₂ Reconstructions Based on Stomatal Frequency

Fossil Responses

Following the demonstration, based on herbarium material, that plants have adjusted their stomatal frequency in response to increasing CO₂ concentrations over the last two centuries (Woodward, 1987), a number of fossil studies have also provided evidence for responses on Quaternary timescales (Beerling, 1993; Beerling et al., 1992, 1993; McElwain et al., 1995). These early studies on Saalian, Weichselian, late-glacial, and Holocene leaves were, however, based on isolated samples. Stratigraphic data supporting a CO₂ response was later presented by Wagner et al. (1996), who demonstrated marked SD and SI reductions for an individual *B. pendula* tree growing on a Dutch peat bog in the period 1952–95. Although not stratigraphically constrained, the well-dated SD data obtained from *Pinus flexilis* needles preserved in packrat middens in Western United States (van de Water et al., 1994) clearly illustrate

a stomatal frequency response to CO₂ changes on glacial-interglacial timescales.

Late-Glacial and Holocene Reconstructions

The first attempt to actually reconstruct Quaternary CO₂ levels was made by Beerling et al. (1995) based on late-glacial and early Holocene lake sediments from western Norway with a high concentration of *S. herbacea* leaves. Their study, which was based on SD data, reconstructed CO₂ concentrations around 275 parts per million by volume (ppmv) during the late Allerød (late AL equivalent to the end of Greenland Interstadial 1 (GI-1) (Björck et al., 1998)), followed by a rapid decrease to ~210 ppmv at the beginning of the Younger Dryas (YD, equivalent to Greenland stadial 1 (GS-1)). CO₂ concentrations increased gradually through the YD to reach about 270 ppmv at the opening of the Holocene and then continued to rise for another ~300 years. Another CO₂ reconstruction covering a similar interval (12 800–10 800 cal years BP) was presented by Rundgren and Björck (2003), who combined partly overlapping high-resolution sequences of *Salix polaris*, *S. herbacea*, and *Betula nana* leaves from the sediments of a lake in southwestern Sweden (Figure 3). This SI-based record

exhibits the same pattern of late-glacial CO₂ change, and similar absolute values, as found by Beerling et al. (1995). In addition, the Swedish study clearly indicates a short-lived reduction in CO₂ concentrations a few centuries into the Holocene that was suggested to be related to the Preboreal oscillation (an early Holocene cooling event; Björck et al., 1997). A third late-glacial-to-early Holocene CO₂ reconstruction based on SD data has been obtained from *Dryas integrifolia* leaves and *Picea mariana*, *Picea glauca*, and *Larix laricina* needles extracted from two ponds in Atlantic Canada (McElwain et al., 2002). In accordance with the other studies, this record shows a clear CO₂ decrease at the AL/YD (GI-1/GS-1) transition. It differs, however, by indicating consistently low YD concentrations (~235 ppmv), followed by a CO₂ rise at the transition to the Holocene, back to levels similar to those at the end of the Allerød (~300 ppmv). A temporary CO₂ decrease, possibly related to the one detected by Rundgren and Björck (2003), is seen in the record with the highest early Holocene resolution.

The first evidence of a century scale, early Holocene (Preboreal) CO₂ oscillation was, however, presented by Wagner et al. (1999), who analyzed the SI of *B. pendula* and *B. pubescens* leaves in lake sediments and peat deposited in an abandoned river channel in the Netherlands. According to results from a

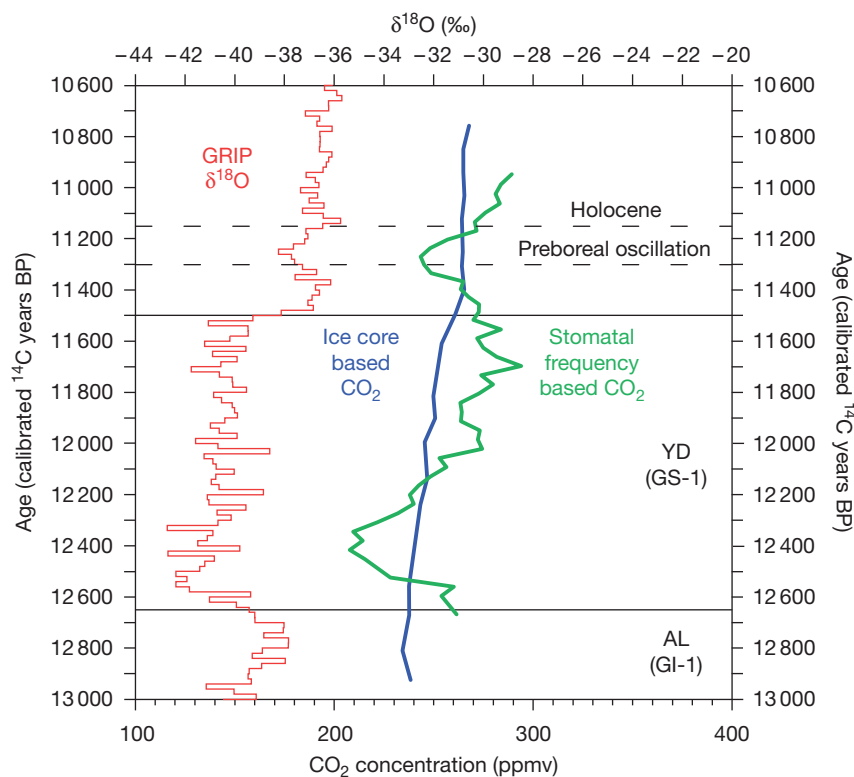


Figure 3 Stomatal-based CO₂ reconstruction through the late-glacial and early Holocene (Rundgren and Björck, 2003). The reconstruction is based on SI data for partly overlapping leaf sequences of three species (*S. herbacea*, *S. polaris*, and *Betula nana*) preserved in the sediments of a Swedish lake, and the curve shown represents the smoothed combined CO₂ record (5-point running mean values). Also shown is the CO₂ record from the Antarctic Dome C ice core (Monnin et al., 2001). The oxygen isotope record from the Greenland Ice Project (GRIP) ice core (Johnsen et al., 2001) illustrates the general temperature development in the North Atlantic region. AL, Allerød interstadial; YD, Younger Dryas stadial. GI-1 and GS-1 are the equivalents to AL and YD in the INTIMATE event stratigraphy (Björck et al., 1998). Age of Preboreal oscillation according to Björck et al. (1997). Adapted from Rundgren M and Björck S (2003) Lateglacial and early Holocene variations in atmospheric CO₂ concentration indicated by high-resolution stomatal index data. *Earth and Planetary Science Letters* 213: 191–204, with permission.

later attempt to date the sequence in detail (van der Plicht et al., 2004), their data indicate a temporary CO₂ decrease (from a general Preboreal level around 340 ppmv) between ca. 11 250 and 11 100 cal years BP. In addition, CO₂ increased distinctly following the YD/Holocene transition. A compilation of all stomatal evidence for a Preboreal CO₂ fluctuation was presented by Wagner et al. (2004). Following its publication, yet another record showing a clear early Preboreal CO₂ decrease has been produced from a temporally detailed SI study of *S. herbacea* leaves preserved in the sediments of a lake in the Faroe Islands (Jessen et al., 2007). As shown in this study, the short-lived Preboreal CO₂ fluctuations indicated by the Dutch, Swedish, and Faroese records all overlap in age when errors in radiocarbon dates are taken into account (Figure 4), which suggests that they all reflect the same event. In the Faroese record, which covers the period ca. 11 200–10 300 cal years BP, the early Preboreal event is followed by a series of CO₂ oscillations within the range 320–340 ppmv. When compared with records spanning the late-glacial-to-early Holocene transition (Beerling et al., 1995; McElwain et al., 2002; Rundgren and Björck, 2003), the two Preboreal

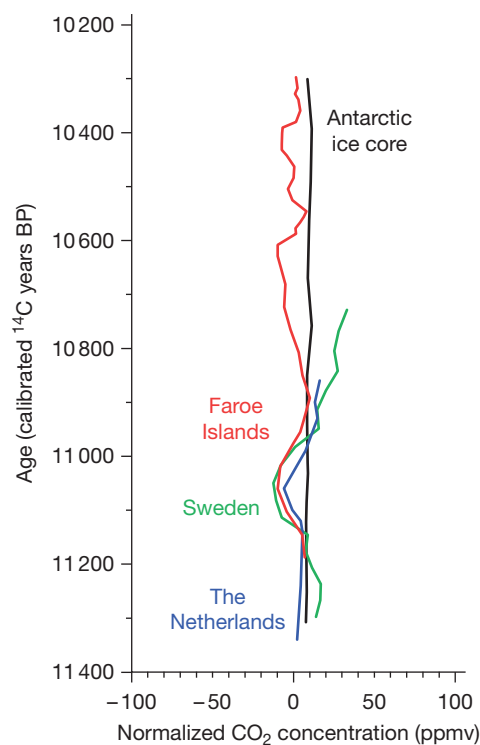


Figure 4 Comparison of three early Holocene stomatal-based CO₂ reconstructions from sites in the Netherlands (Wagner et al., 1999, 3-point running mean values), Sweden (Rundgren and Björck, 2003, 5-point running mean values), and the Faroe Islands (Jessen et al., 2007, 5-point running mean values). The figure illustrates that a century-scale CO₂ decrease centered at ca. 11 100 cal years BP (according to the Faroese age model) and believed to be related to the Preboreal oscillation (Björck et al., 1997) is reproduced by three independent stomatal frequency studies. Also shown is the CO₂ record from the Antarctic Dome C ice core (Monnin et al., 2001), which does not show any evidence of the CO₂ decrease reconstructed by stomatal frequency analysis.

reconstructions detailed above show overall unexpectedly high CO₂ concentrations. Potential explanations for such divergent and consistently elevated CO₂ estimates – also relative to ice-core records – are discussed toward the end of this review.

Rundgren and Beerling (1999) presented a CO₂ record for the past 9000 years based on SI data obtained from a remarkably long sequence of *S. herbacea* leaves preserved in the sediments of a lake in northern Sweden. Although the leaf record is sparse prior to ca. 5000 cal years BP, the reconstruction clearly indicates an overall CO₂ increase from levels of ~270 ppmv at this time up to the present day. The high-resolution data available for the remaining part of the Holocene shows, however, that this millennial-scale trend was punctuated by a marked period of lower CO₂ levels between ca. 1000 and 500 cal years BP (AD 950–1450), and the industrial CO₂ rise is clearly visible in the record. The period 8700–6800 cal years BP was studied in more detail by Wagner et al. (2002), who analyzed *B. pendula* and *B. pubescens* leaves from a Danish lake sediment sequence. They found evidence for a ~25 ppmv CO₂ drop (from a level of ~300 ppmv) between ca. 8400 and 8000 cal years BP that was suggested to be related to the so-called ‘8.2 ka BP event,’ a cooling event of possibly global significance (Rohling and Pälike, 2005). A short-lived CO₂ decrease at this time may also be present in the Holocene record from northern Sweden (Rundgren and Beerling, 1999; Wagner et al., 2004).

The CO₂ evolution of the last ~1200 years was studied in detail by Kouwenberg et al. (2005) based on stomatal frequency (number of stomata per millimeter of needle length) analysis of *Tsuga heterophylla* needles preserved in the sediments of a pond in Northwestern United States. They reconstructed a series of centennial-scale CO₂ fluctuations around a preindustrial average of 280–290 ppmv, with the most pronounced minima centered on ca. AD 850 and 1150, and with maxima at ca. AD 1000, 1300, and 1700. The pattern of change was compared with paleoclimatic records from various regions, and it was concluded that a coupling has existed between atmospheric CO₂ and climate during the last millennium. Another CO₂ record covering part of the same period (c. AD 1000–1500) has been produced using SI data from *Q. robur* leaves derived from a Dutch river channel deposit. This record is consistent with that of Kouwenberg et al. (2005) in showing a distinct CO₂ increase between ca. AD 1200 and AD 1300, but absolute values are generally higher (~290–320 ppmv).

Eemian Reconstructions

S. herbacea leaves extracted from sediment samples collected in East Greenland and known to be of last interglacial age were analyzed for SI by Rundgren and Bennike (2002). A majority of these samples, for which there was no stratigraphic control, yielded typical interglacial CO₂ values (250–280 ppmv), but two lower estimates were taken to indicate century-scale variability during the Eemian. CO₂ levels during the last interglacial were investigated in more detail by Rundgren et al. (2005), who analyzed SI of *B. pendula*, *Q. robur*, and *Q. petraea* leaves from a Danish lake deposit. They reconstructed concentrations between 250 and 290 ppmv, and a marked centennial to millennial variability, during the first 3000 years of the Eemian. CO₂ levels were more stable, and slightly higher (290–300 ppmv), during the following three millennia. Based on

pollen stratigraphic correlations, the observed shift in CO₂ dynamics was coincident with a major climatic transition (the end of the Eemian climatic optimum) in northwestern Europe.

Potential and Limitations of the Method

It is clear from the overview presented above that stomatal frequency analysis has become increasingly and successfully applied in Quaternary science over the past two decades. These studies, which have been based on leaf sequences derived from a range of species and sampled from sites with a wide geographic distribution, have contributed important information about CO₂ dynamics during various parts of the Quaternary. In particular, they have provided ample evidence for century-scale CO₂ variability, which is rarely detected in more traditional ice-core studies. In a Quaternary perspective, this ability to detect short-term variations is probably the largest merit of the method because it enables high-resolution CO₂ and climate records to be compared in order to better understand carbon cycle–climate interactions on submillennial timescales. The potential of this type of exercise is particularly large when stomatal frequency and climate proxy data are available from the same sediment or peat sequence (Jessen et al., 2005; Rundgren et al., 2005). In addition, stomatal frequency analysis has the potential to provide high-resolution CO₂ reconstructions for Quaternary (and earlier) intervals that are beyond the reach of ice cores (e.g., McElwain et al., 2005; Royer et al., 2001).

It should, however, be acknowledged that the stomatal frequency method is still in a stage of development. A fundamental test of the reliability of any method is, of course, its performance in relation to other methods. Quaternary CO₂ records are traditionally based on analysis of air bubbles enclosed in polar ice cores (Stauffer), and this is presently the most widely accepted method. On a multimillennial timescale during the Holocene, stomatal-based CO₂ reconstructions show trends very similar to ice cores (Rundgren and Beerling, 1999). For the late Holocene, a good agreement in trends is also found on centennial timescales (Kouwenberg et al., 2005; Rundgren and Beerling, 1999; van Hoof et al., 2005). The century-scale CO₂ fluctuations seen in early Holocene stomatal-based reconstructions (Jessen et al., 2007) (Rundgren and Björck, 2003; Wagner et al., 1999, 2002) are, however, not evident in ice-core records (Figures 3 and 4). The situation is similar during the YD (GS-1), where stomatal frequency reconstructions show a CO₂ decrease, at the same time as ice-core records are characterized by a slow and steady rise in CO₂. Overall, the late-glacial trend is similar for stomatal and ice-core-based data (McElwain et al., 2002; Rundgren and Björck, 2003; Figure 3), but the stomatal reconstructions show a temporally more detailed record connected to the climate changes around the YD.

In summary, stomatal frequency analysis has the capacity to reconstruct millennial-scale CO₂ trends consistent with ice-core records. Moreover, centennial-scale trends are similar for the two methods during the late Holocene. This suggests that stomatal frequency analysis is an adequate method for CO₂ reconstruction and that there is only limited influence from

other environmental factors on the results, at least for SI-based reconstructions. For the early Holocene and older periods, however, century-scale CO₂ variability is not apparent in ice-core records but is seen in stomatal-based reconstructions. This discrepancy does not necessarily reflect flaws in the stomatal frequency method; it may as well be an indication that the effect of diffusional smoothing on CO₂ records from deep ice cores is currently underestimated. As shown by Wagner et al. (2004), both late and early Holocene century-scale CO₂ fluctuations are reproduced in a number of stomatal frequency reconstructions from different species and geographic regions, which strengthens the credibility of the method.

The consistently larger CO₂ amplitude in stomatal-based reconstructions, compared to ice-core records, may partly be explained by a combination of the generally larger scatter in biological proxy data and diffusional smoothing of CO₂ concentrations measured in ice cores. The contribution of the latter process was investigated by van Hoof et al. (2005) by applying a firn diffusion model, constrained by site-specific parameters for a high-resolution Antarctic ice core, to a stomatal reconstruction between AD 1000 and AD 1500. This resulted in a 25% reduction of the amplitude and a CO₂ record consistent with the ice core.

When the CO₂ reconstructions reviewed above are compared with ice-core records, it is clear that many stomatal frequency studies provide Holocene CO₂ estimates consistently slightly higher than shown by ice cores. This discrepancy is difficult to explain. Higher CO₂ concentrations at the time of bud-burst at high-latitude sites relative to the mean annual levels reflected by ice cores can only account for a minor part. Possibly, the answer lies in partly inadequate calibration datasets, and efforts are now being made to improve CO₂ inference models (Wagner et al., 2005). Difficulties in obtaining stomatal frequency data representative of very low CO₂ concentrations are, however, likely to continue to obstruct quantification of CO₂ levels during glacial periods.

See also: Plant Macrofossil Introduction. **Ice Core Methods:** CO₂ Studies.

References

- Beerling DJ (1993) Changes in the stomatal density of *Betula nana* leaves in response to increase in atmospheric carbon dioxide concentrations since the late-glacial. *Special Papers in Palaeontology* 49: 181–187.
- Beerling DJ (1999) Stomatal density and index: Theory and application. In: Jones TP and Rowe NP (eds.) *Fossil Plants and Spores: Modern Techniques*, pp. 251–256. London: The Geological Society.
- Beerling DJ, Birks HH, and Woodward FI (1995) Rapid late-glacial atmospheric CO₂ changes reconstructed from the stomatal density record of fossil leaves. *Journal of Quaternary Science* 10: 379–384.
- Beerling DJ, Chaloner WG, Huntley B, Pearson JA, and Tooley MJ (1993) Stomatal density responds to the glacial cycle of environmental change. *Proceedings of the Royal Society of London Series B* 251: 133–138.
- Beerling DJ, Chaloner WG, Huntley B, et al. (1992) Variations in the stomatal density of *Salix herbacea* L. under the changing atmospheric CO₂ concentrations of late- and post-glacial time. *Philosophical Transactions of the Royal Society of London Series B* 336: 215–224.
- Birks HH (2001) Plant macrofossils. In: Smol JP, Birks HJB, and Last WM (eds.) *Tracking Environmental Change Using Lake Sediments, Vol. 3: Terrestrial, Algal, and Siliceous Indicators*, pp. 49–74. Dordrecht: Kluwer Academic Publishers.

- Björck S, Rundgren M, Ingólfsson Ó, and Funder S (1997) The Preboreal oscillation around the Nordic Seas: Terrestrial and lacustrine responses. *Journal of Quaternary Science* 12: 455–465.
- Björck S, Walker MJC, Cwynar LC, et al. (1998) An event stratigraphy for the Last Termination in the North Atlantic region based on the Greenland ice-core record: A proposal by the INTIMATE group. *Journal of Quaternary Science* 13: 283–292.
- Jessen CA, Rundgren M, Björck S, and Hammarlund D (2005) Abrupt climatic changes and an unstable transition into a late Holocene Thermal Decline: A multiproxy lacustrine record from southern Sweden. *Journal of Quaternary Science* 20: 349–362.
- Jessen CA, Rundgren M, Björck S, and Muscheler R (2007) Climate forced atmospheric CO₂ variability in the early Holocene: A stomatal frequency reconstruction. *Global and Planetary Change* 57 (3–4): 247–260.
- Johnsen SJ, Dahl-Jensen D, Gundestrup N, et al. (2001) Oxygen isotope and palaeotemperature records from six Greenland ice-core stations: Camp Century, Dye-3, GRIP, GISP2, Renland and North GRIP. *Journal of Quaternary Science* 16: 299–307.
- Kouwenberg LR, McElwain JC, Kürschner WM, et al. (2003) Stomatal frequency adjustment of four conifer species to historical changes in atmospheric CO₂. *American Journal of Botany* 90: 610–619.
- Kouwenberg LR, Wagner F, Kürschner WM, and Visscher H (2005) Atmospheric CO₂ fluctuations during the last millennium reconstructed by stomatal frequency analysis of *Tsuga heterophylla* needles. *Geology* 33: 33–36.
- Kürschner WM, Wagner F, Visscher EH, and Visscher H (1997) Predicting the response of leaf stomatal frequency to a future CO₂-enriched atmosphere: Constraints from historical observations. *Geologische Rundschau* 86: 512–517.
- McElwain JC, Mayle FE, and Beerling DJ (2002) Stomatal evidence for a decline in atmospheric CO₂ concentration during the Younger Dryas stadial: A comparison with Antarctic ice core records. *Journal of Quaternary Science* 17: 21–29.
- McElwain J, Mitchell FJG, and Jones MB (1995) Relationship of stomatal density and index of *Salix cinerea* to atmospheric carbon dioxide concentrations in the Holocene. *The Holocene* 5: 216–219.
- McElwain JC, Wade-Murphy J, and Hesselbo SP (2005) Changes in carbon dioxide during an anoxic event linked to intrusion into Gondwana coals. *Nature* 435: 479–482.
- Monnin E, Indermühle A, Dällenbach A, et al. (2001) Atmospheric CO₂ concentrations over the last glacial termination. *Science* 291: 112–114.
- Poole I and Kürschner WM (1999) Stomatal density and index: The practice. In: Jones TP and Rowe NP (eds.) *Fossil Plants and Spores: Modern Techniques*, pp. 257–260. London: The Geological Society.
- Poole I, Weyers JDB, Lawson T, and Raven JA (1996) Variations in stomatal density and index: Implications for palaeoclimatic reconstructions. *Plant, Cell and Environment* 19: 705–712.
- Rohling EJ and Pälike H (2005) Centennial-scale climate cooling with a sudden cool event around 8,200 years ago. *Nature* 434: 975–979.
- Royer DL (2001) Stomatal density and stomatal index as indicators of paleoatmospheric CO₂ concentration. *Review of Palaeobotany and Palynology* 114: 1–28.
- Royer DL, Scott LW, Beerling DJ, et al. (2001) Paleobotanical evidence for near present-day levels of atmospheric CO₂ during part of the tertiary. *Science* 292: 2310–2313.
- Rundgren M and Beerling D (1999) A Holocene CO₂ record from the stomatal index of subfossil *Salix herbacea* L. leaves from northern Sweden. *The Holocene* 9: 509–513.
- Rundgren M and Bennike O (2002) Century-scale changes of atmospheric CO₂ during the last interglacial. *Geology* 30: 187–189.
- Rundgren M and Björck S (2003) Lateglacial and early Holocene variations in atmospheric CO₂ concentration indicated by high-resolution stomatal index data. *Earth and Planetary Science Letters* 213: 191–204.
- Rundgren M, Björck S, and Hammarlund D (2005) Last interglacial atmospheric CO₂ changes from stomatal index data and their relation to climate variations. *Global and Planetary Change* 49: 47–62.
- Salisbury EJ (1927) On the causes and ecological significance of stomatal frequency, with special reference to the woodland flora. *Philosophical Transactions of the Royal Society of London Series B* 216: 1–65.
- Tichá I (1982) Photosynthetic characteristics during ontogenesis of leaves: 7. Stomata density and sizes. *Photosynthetica* 16: 375–471.
- Van de Water PK, Leavitt SW, and Betancourt JL (1994) Trends in stomatal density and ¹³C/¹²C ratios of *Pinus flexilis* needles during last glacial-interglacial cycle. *Science* 264: 239–243.
- van der Plicht J, van Geel B, Bohncke SJP, et al. (2004) The Preboreal climate reversal and a subsequent solar-forced climate shift. *Journal of Quaternary Science* 19: 263–269.
- van Hoof TB (2004) *Coupling Between Atmospheric CO₂ and Temperature During the Onset of the Little Ice Age*. The Netherlands: Utrecht University PhD Thesis, LPP Contribution Series No. 18.
- van Hoof TB, Kaspers KA, Wagner F, et al. (2005) Atmospheric CO₂ during the 13th century AD: Reconciliation of data from ice-core measurements and stomatal frequency analysis. *Tellus* 57B: 351–355.
- Wagner F, Aaby B, and Visscher H (2002) Rapid atmospheric CO₂ changes associated with the 8,200-years-B.P. cooling event. *Proceedings of the National Academy of Sciences of the United States of America* 99: 12011–12014.
- Wagner F, Below R, De Klerk P, et al. (1996) A natural experiment on plant acclimation: Lifetime stomatal frequency response of an individual tree to annual atmospheric CO₂ increase. *Proceedings of the National Academy of Sciences of the United States of America* 93: 11705–11708.
- Wagner F, Bohncke SJP, Dilcher DL, et al. (1999) Century-scale shifts in early Holocene atmospheric CO₂ concentration. *Science* 284: 1971–1973.
- Wagner F, Dilcher DL, and Visscher H (2005) Stomatal frequency responses in hardwood-swamp vegetation from Florida during a 60-year continuous CO₂ increase. *American Journal of Botany* 92: 690–695.
- Wagner F, Kouwenberg LR, van Hoof TB, and Visscher H (2004) Reproducibility of Holocene atmospheric CO₂ records based on stomatal frequency. *Quaternary Science Reviews* 23: 1947–1954.
- Woodward FI (1987) Stomatal numbers are sensitive to increases in CO₂ from pre-industrial levels. *Nature* 327: 617–618.
- Woodward FI and Bazzaz FA (1988) The responses of stomatal density to CO₂ partial pressure. *Journal of Experimental Botany* 209: 1771–1781.

Introduction

The northern treeline zone has long been the focus of study by Quaternary paleoclimatologists and paleoecologists. The geographic position of the circum-Arctic treeline is controlled by global climate patterns. As trees growing in the treeline zone are at their physiological range limits and sensitive to environmental perturbations, studies of past treeline shifts can provide information on the form and rate of response of trees and forest plant communities to climatic changes, and reconstructions of past treeline movement provide insights into past variations in climate. In addition, special attention has focused upon the northern treeline because large geographic shifts in this zone can produce feedbacks that amplify the impacts of climate change (Foley et al., 1994). Quaternary scientists have used several lines of evidence to study past climate and vegetation at treeline, including fossil pollen and fossil stomata from lake sediments and peats, macrofossils such as preserved conifer needles, and paleopedological records. One of the most striking and direct forms of evidence pertaining to past northward advances of the northern treeline comes from wood megafossils that have been recovered and analyzed from a number of locales in North America and Eurasia.

Current Treeline Location, Vegetation, and Climatic Controls

The northern treeline is best thought of as a zone of transition across which continuous forest cover of the boreal forest gives way to increasingly scattered fragments of forest and smaller and smaller individual trees (Figure 1). Eventually, the ultimate species' limits for trees are reached and the vegetation becomes a treeless tundra. The extent of this transitional zone can range from a few kilometers to over a hundred kilometers. At the extreme northern limits of their distributions, coniferous species often grow in a low, contorted and ground-hugging form called *krummholz* (twisted wood; Figure 2). The current position of the northern treeline zone (Figure 3) lies generally between 60° and 70° N latitude. The circumpolar treeline descends southward to below 60° N in central Canada and ascends northward to 73° N in the Taimyr Peninsula region of central Siberia. In Europe, the dominant treeline conifers include *Pinus sylvestris* L. (scots pine), *Picea abies* (L.) Karst. and *Picea obovata* Ledeb. (Norway and Russian spruce), and *Larix sibirica* Ledeb. (Siberian larch). In northern Asia, the dominant treeline conifers include larches (*L. sibirica* and *Larix dahurica* Turcz. ex Trautv.) and *Pinus pumila* (Pall.), Regel (stone pine) in the far east. *Pinus sylvestris* and *Picea* are found in the boreal forest of northern Asia, but are not generally present as far north as the treeline in that region. In North America, the dominant coniferous trees include *Picea glauca* (Moench), Voss (white spruce), *Picea mariana* (P. Mill.) BSP (black

spruce), and larch (*Larix laricina* (Du Roi) K. Koch). *Pinus banksiana* Lamb. (jack pine) is found in the northern boreal forest, but does not grow at treeline. Broadleaf trees such as birch (*Betula*) and poplar (*Populus*) can also be found in the North American and Eurasian treeline zones. In northern Scandinavia, the treeline is formed by *Betula pubescens* ssp. *czerepanovii* (Orlova) Hamet-Ahti.

Cool summer temperatures coupled with a short growing season appear to be the ultimate reason why trees do not grow further north in the Arctic. Air and soil temperatures are too cool and the growing season is too short for trees to conduct adequate photosynthesis to meet growth and reproductive requirements. Sexual reproduction is usually rare for *krummholz* trees and propagation is often by vegetative layering. Very cold winter temperatures can limit tree occurrence for some species such as *Pinus sylvestris*, and damage to buds and needles by blowing snow in the winters can also limit growth or establishment of trees. The treeline zone typically occurs (Figure 3) where the average July temperatures lie between 12.5 and 10.0 °C (MacDonald, 2002). The treeline also roughly corresponds to the average July position of the Arctic Frontal Zone. Areas north of this frontal boundary are dominated by cold and dry polar air masses. Thus, the geographic position of the treeline is related to both general latitudinal patterns of decreasing insolation and general atmospheric circulation. Average winter temperatures have a much greater range than summer temperatures when treeline sites in North America and Eurasia are compared. Average January temperatures range from only –10 to –15 °C at the Fennoscandian treeline in northwestern Europe, between –20 and –25 °C for much of the North American treeline and below –40 °C for some regions of central Siberia.

The location of treeline is ultimately controlled by climate, but the presence or absence of trees also affects the climate. The dark foliage and rough canopy of forested areas absorb more sunlight than light colored and smooth tundra surfaces, and this is important for planetary albedo and radiation fluxes. Climatological models suggest that if the treeline were to significantly advance northward, this could actually produce a significant increase in the temperature for the entire earth surface due to the increased capture of solar energy by the dark forest cover. Model experiments suggest that pronounced northward extensions of treeline in the Quaternary might have played a role in global patterns of climate warming (Foley et al., 1994).

The general correspondence between summer temperatures and the geographic position of the modern treeline leads to the conclusion that past shifts in treeline likely reflect changes in summer thermal conditions in the Arctic and sub-Arctic. In simplest terms the presence of trees might be used to infer minimum average July temperatures of 12.5–10.0 °C at that location. The use of treeline to infer winter conditions is more problematic. Due to the relationship between

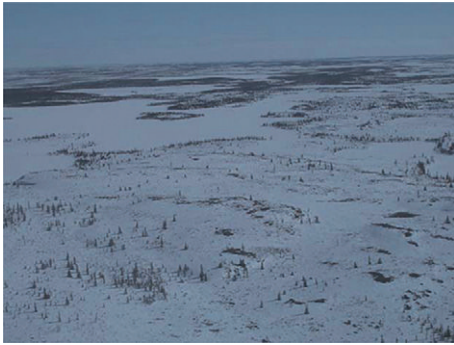


Figure 1 An aerial view of the *Picea* dominated northern treeline zone of central Canada showing the transition from denser forest cover in the south to scattered small trees and krummholz at treeline (© Glen M. MacDonald).



Figure 2 Krummholz *Larix dahurica* near the Lena River delta in eastern Siberia with a pocket knife for scale (© Glen M. MacDonald).

summer temperatures and the Arctic Frontal Zone, reconstructions of treeline position can cast light upon both insolation-driven thermal changes and those attributable to changes in atmospheric circulation patterns (MacDonald et al., 1993, 2000b; Ritchie and Hare, 1971; Ritchie et al., 1983).

Wood Megafossil Analysis and the Interpretation of Past Treeline Shifts

The occurrence of ancient tree stumps on the presently treeless tundra has been known since at least the reports of A.F. Middendorf from Siberia in the 1860s. Megafossils of wood, which includes trunks, branches, and roots of conifer and broadleaf trees, have now been recovered from a number of areas beyond the modern treeline in North America and Eurasia. In some cases, these megafossils are found in rooted growth position or lying near their former point of growth on the tundra surface (Figure 4(a)). Such wood might be preserved for thousands of years in the Arctic tundra for several reasons:

1. The cold temperatures and brevity of the warm season in the Arctic and northern sub-Arctic restrict microbiological and macrobiological decomposition rates.
2. The relatively dry conditions of many Arctic and sub-Arctic sites further restrict decomposition rates.
3. The radial growth rates for northern trees are often very slow, producing narrow growth rings and very dense wood, which is more resistant to water infiltration and rot than fast growing, less dense wood.
4. Fire frequency in sparsely wooded treeline and tundra sites is generally extremely low, limiting the chance that wood will be consumed by fire.

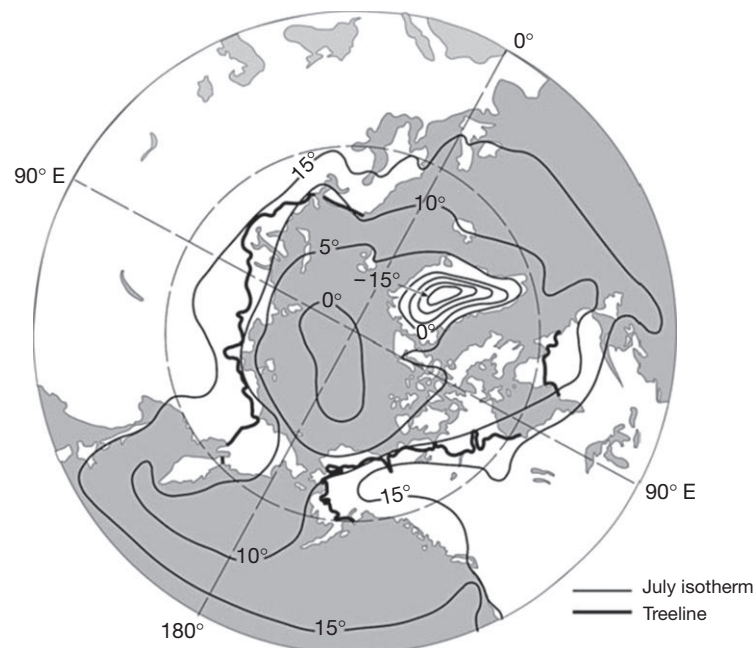


Figure 3 The relationship between the location of the northern treeline and mean July temperature.



Figure 4 Early to mid-Holocene wood megafossils: (a) *Larix* log lying on the tundra surface on the Taimyr Peninsula, Russia, (b) *Pinus* stump in the shallow waters of a small lake on the Kola Peninsula, Russia, and (c) *Larix* stumps covered with colluvium and subsequently exposed by stream erosion near the mouth of the Lena River, Russia (© Glen M. MacDonald).

5. Human population densities are generally low in the tundra zone, limiting, although not completely restricting, the removal or use of wood by people.

In addition to being found on the surface, wood megafossils are commonly preserved in the waters and sediments of lakes, peatbogs, fluvial deposits, and colluvium (Figure 4(b) and 4(c)).

Field sampling strategy to collect megafossils varies. In some cases, systematic foot surveys and transects are done of the tundra surface or along stream channels and other exposures. The waters of lakes can also be surveyed to find material

lying on top of the sediment. Megafossils can be recovered from some lake or peat cores. Finally, newly exposed landscapes created by the recent melting of perennial snowbanks or glacier fronts can reveal megafossils from previous warm periods. The use of geographic positioning systems (GPS) to provide exact coordinates of the finds, careful note taking, and photographic documentation of the finds are important components of the field work.

Identification of wood megafossils can typically be done to the level of genus using thin sections of wood to examine the characteristics of the xylem cells and associated structures

(Core et al., 1979; Hoadley, 1990). The absence of vessels identifies the wood as a gymnosperm (Pinophyta). An example of a simple key for some typical boreal gymnosperm genera is as follows:

1. Resin ducts normally present.
 - (a) Wood with abrupt latewood–earlywood transition.
 - (b) Ray tracheids present – *Larix*.
 - (c) Wood with gradual transition.
 - (d) Cross-field pits large with indistinct borders – *Pinus*.
 - (e) Cross-field pits small and bordered – *Picea*.
2. Resin ducts normally absent.
 - (a) Ray tracheids absent and walls of ray cells conspicuously pitted – *Abies*.

Two approaches are generally used to provide chronological control for the wood megafossils. First, samples which are more than 200 years and less than 40 000 years in age can be radiocarbon dated. Typically, the precision of such estimates ranges from 100 to 500 years. If at all possible, a portion of any sample submitted for radiocarbon dating should be archived to serve as a voucher specimen should further analysis or verification be required. Dendrochronological cross-dating has been used to assign ages to wood megafossil samples dating from the past 300 to 1000 years in many studies of treeline dynamics in the Western and Eastern hemispheres. In certain areas, such as northern Fennoscandia and the Polar Urals–Yamal Peninsula region, dendrochronological techniques have begun to be used to produce tree-ring chronologies that can be used to cross-date megafossils back millennia (e.g., Shiyatov et al., 1996; Zetterberg et al., 1996), and such efforts will likely bear fruit elsewhere in the treeline zone. Such cross-dated samples provide exact calendar year dating control for the samples.

When care is taken in the collection, identification, and dating of the megafossil, it can be used as evidence for the past minimum northern position of the treeline. Megafossils can be collected from currently treed sites at treeline to provide a minimum age for the establishment of tree populations at the site. When large collections are made of many samples, the changes in the number of samples recovered from different time periods can potentially provide evidence of changes in forest density in the past (MacDonald et al., 2000b).

Fossil pollen studies of lake sediments and peat sections have long been used to reconstruct treeline history for the Holocene. However, this approach suffers from the often poor spatial resolution of pollen records, which are generally dominated by wind dispersed taxa that can be carried tens to hundreds of kilometers from parent trees. This is exacerbated by the low pollen productivity of tundra vegetation, which allows the pollen signal from tundra sites to be dominated by long-distance dispersed tree pollen from the south. For example, in Canada it is not unusual to record *Picea* pollen percentages of over 10% from lakes located in the tundra over 100 km or more, north of any living spruce trees (Gajewski, 2002). So long as sampling is conducted away from beaches or north flowing rivers where long-distance transported wood might occur, the use of wood megafossils can provide a more spatially explicit means of recording past treeline location because the tree likely grew at the site the megafossil is recovered.

Although wood megafossils are not prone to the long-distance transport problems that affect pollen-based estimates

of past treeline, because of their inherent scarcity in many regions, wood megafossils might sometimes provide only minimum age estimates of the timing of tree species' arrival in a region or of actual range limits. A comparison of the fossil pollen record, the fossil conifer stomata record, and the wood megafossil record from a small lake lying beyond the modern conifer treeline on the Kola Peninsula of Russia illustrates this point (Gervais et al., 2002; MacDonald et al., 2000a). The wood megafossil record (Figure 5) suggests the establishment of Scots pine at the site at about 7400 cal years BP (6500 ^{14}C years BP). In contrast, the fossil pollen record shows an earlier large increase in *Pine* pollen, commencing at about 8200 cal years BP (7500 ^{14}C years BP). Studies on the Kola Peninsula have shown that *Pinus* conifer stomata preserved in lake sediments can provide a relatively spatially precise indication of the position of treeline, as such stomata initially arrive in the lake on needles which are likely not to be transported far from the tree (Gervais and MacDonald, 2001). The stomata record from the same lake indicates that Scots pine was established around the lake by 8.8 cal ka BP (8000 ^{14}C years BP) – significantly earlier than is indicated by the wood megafossil record. Although lags between the megafossil record and the palynological record are not universal – and in some cases megafossils can yield similar or earlier arrival dates than palynological records (Barnekow, 1999; Kullman, 1998a; 1999) – care should be taken in implying a maximum age for the presence of a species based upon the megafossil record alone.

Studies From Selected Regions

Tuktoyaktuk Peninsula

Surface finds of spruce wood have been reported from north of the modern treeline in Tuktoyaktuk Peninsula region of Canada (Figure 6). Wood has also been found in exposures of organic sediments. The wood samples come from sites that extend as far as 40 km north of the modern limits of spruce. The samples range in age (Table 1) from ~7500 to 5600 cal years BP (6600–4900 ^{14}C years BP). The magnitude of the range extension of spruce trees beyond their modern limits, and comparison of the tree-ring widths of the fossil wood with the ring growth rates of living spruce trees to the south, suggest that summer temperatures on the Tuktoyaktuk Peninsula were as much as 3 °C warmer during the early to mid-Holocene than today. The ages of the spruce megafossils correspond with fossil pollen evidence from lakes on the peninsula that show markedly increased amounts of spruce pollen during the early to mid-Holocene (Ritchie, 1984; Spear, 1983).

Northern Quebec

Wood megafossils have been used in a number of studies to investigate the age and dynamics of treeline forest and krummholz populations of spruce in northern Quebec. In many cases, these studies have focused on using megafossils cross-dated with living trees to extend dendrochronological records of climate change and ecological response (e.g., Wang et al., 2001). In other cases, wood megafossils have been used in concert with radiocarbon-dated charcoal from soils which is

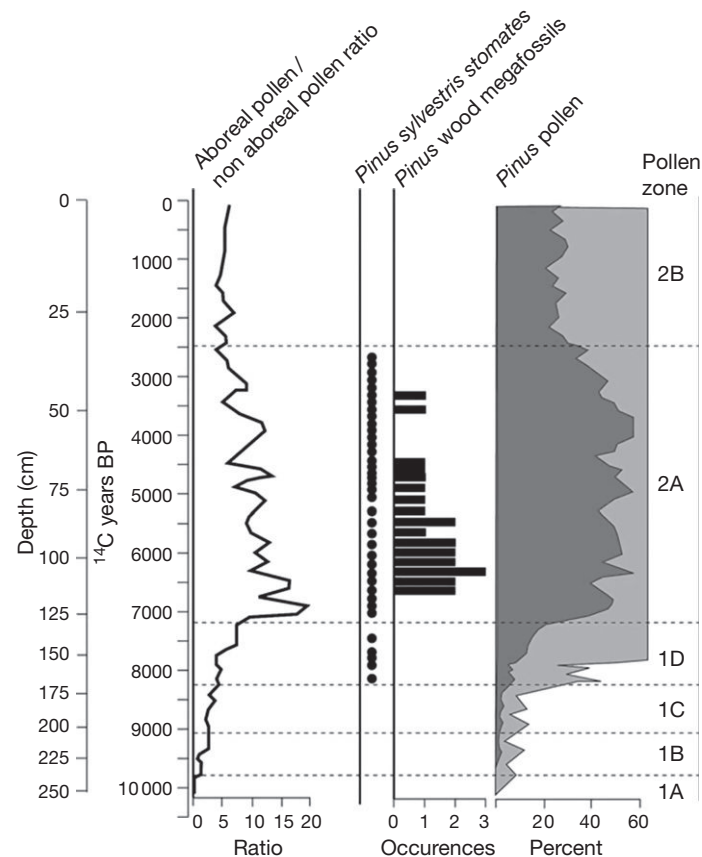


Figure 5 A comparison of evidence for the arrival of *Pinus sylvestris* on the northern Kola Peninsula based on *Pinus* pollen, wood megafossils, and *Pinus* stomata. (Data from Gervais et al., 2002)

indicative of past forest presence and fires. Dated megafossils indicate that spruce trees arrived in the vicinity of the present treeline by 5000 cal years BP. There is also megafossil evidence of several minor readvances in treeline or infilling of the tree-line zone following 3000 cal years BP (Arseneault and Payette, 1997). The older megafossils in this region appear to be present in peats.

Fennoscandia–Kola Peninsula

Radiocarbon-dated and dendrochronologically dated wood macrofossils have been reported from a number of high-latitude and high-elevation tundra sites in Fennoscandia and the adjacent Kola Peninsula of Russia. Megafossils of *Pinus sylvestris* beyond its present species' range limits are particularly common and have long-provided evidence (Eronen, 1979) that pines grew both higher in latitude and higher in elevation during the early to mid-Holocene (Figure 7). The abundance of *Pinus sylvestris* wood has allowed for the development of a >7000 year tree-ring chronology for Finnish Lapland (Eronen et al., 2002; Zetterberg et al., 1996). Much of the wood for that chronology was retrieved from lakes. The abundance of drowned trees in lakes was combined with other evidence to infer that conditions in northern Fennoscandia were drier than present between 8000 and 4000 cal years BP (Eronen et al.,

1999). Indeed the rising water levels during the latter part of the Holocene likely played a role in the preservation of abundant wood megafossils in the small lakes of the region.

Some of the most remarkable reports of megafossil wood come from the south in the Scandes Mountains of Sweden (Table 2). Megafossils of *Pinus sylvestris*, *Picea abies*, *B. pubescens*, and other woody species have been found well above their modern range limits and dated to as far back as the early Holocene and Late Glacial (e.g., Kullman, 1998a,b, 1999, 2004). *B. pubescens* wood has been recovered from melting snowbanks and glaciers as much as 650 m above the present elevational range limits of the species, indicating a much higher treeline for birch between 9700 and 7000 cal years BP (8700 and 6200 ¹⁴C years BP). It has been estimated that summer temperatures were likely 3 °C warmer than present during this time period (Kullman, 2004). Another remarkable detail of the finds in the Scandes Mountains is the occurrence of *L. sibirica* wood dating from the early Holocene (Kullman, 1998b). Today, the species does not occur naturally in Fennoscandia. The presence of *Larix* in Fennoscandia during the early post-glacial implies both rapid rates of range expansion at the close of the last Ice Age and the development of a vegetation that has no modern analog in the region. Intriguingly, *Larix* pollen has not been found in regional sedimentary records, and there is vigorous discussion about the dates and interpretation of some

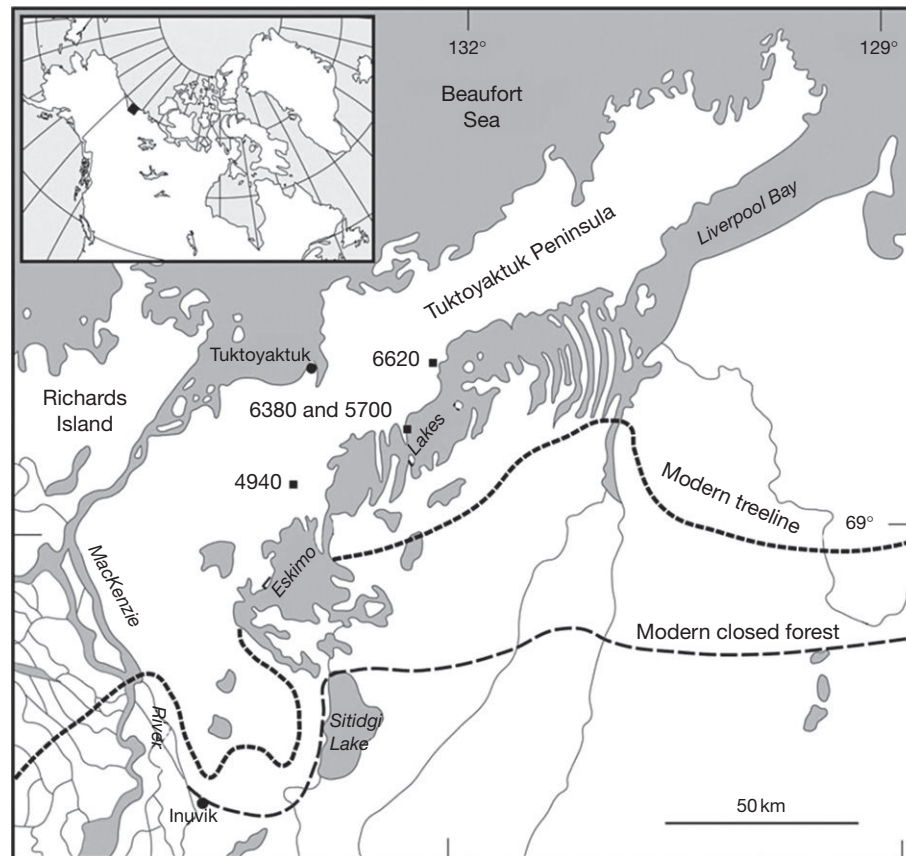


Figure 6 *Picea* megafossil locations and radiocarbon dates from the Tuktoyaktuk Peninsula, Canada. (Data from Ritchie, 1984).

Table 1 Ages of dated *Picea* wood from the Tuktoyaktuk Peninsula, Canada

Location	Age (^{14}C years BP)	Age (cal years BP (1 σ))	Lab number
69° 07' N 133° 16' W	4940 $\pm$ 140	5607–5716	GSC-1265
69° 27' N 132° 10' W	6620 $\pm$ 70	7469–7524	GSC-3239
69° 16' N 132° 20' W	6380 $\pm$ 70	7261–7332	GSC-3242
69° 16' N 132° 20' W	5700 $\pm$ 100	6410–6548	BGS-215

Data from Ritchie JC (1984) *Past and Present Vegetation of the far Northwest of Canada*. Toronto: University of Toronto Press.

of the early Holocene and Late Glacial megafossil finds from the Scandes (Birks et al., 2005; Kullman, 2005).

Higher-frequency climatic changes and forest response are also evident in the wood megafossil records of Fennoscandia and the adjacent Kola Peninsula. For example, warmer summer temperatures during AD 1000–1300 (coeval with the 'Medieval Warm Period' observed in other parts of Europe) are evident in a megafossil stumps from the Khibiny Mountains of the Kola Peninsula which indicate a 100–140 m upward shift in the *Pinus sylvestris* limit compared to today. Rooted stumps of *Pinus sylvestris* from a small pond suggest drier conditions during the same period (Kremenetski et al., 2004).

Northern European Russia and Siberia

As mentioned earlier, there is a long history of scientific investigation of wood megafossils from the tundra regions of the

Russian Federation. Aside from the finds of *Pinus sylvestris* from the Kola Peninsula which were discussed earlier, wood from *Pinus*, *Picea*, *Larix*, and *Betula* has been reported from sites extending eastward to the Polar Urals and thence onward to the Chukotka region in the far east of the Russian Federation (Khotinsky, 1984; Kremenetski et al., 1998; MacDonald et al., 2000b). Many of the samples recovered thus far come from surface finds or fallen and *in situ* stumps that can date back to the early to mid-Holocene. Wood has also been recovered from organic soil exposures, exposures of colluvium, cut-banks along river and stream channels, and from small lakes. A broad synthesis of the many available dates suggests that boreal trees were present at or beyond their modern range limits across northern Eurasia by about 11400 cal years BP (10000 ^{14}C years BP). Between the period 10000 and 3200 cal years BP (9000 and 3000 ^{14}C years BP), the range limits of boreal trees extended north of their modern ranges in most regions, reaching the current Arctic coastline in some areas (Figure 8).

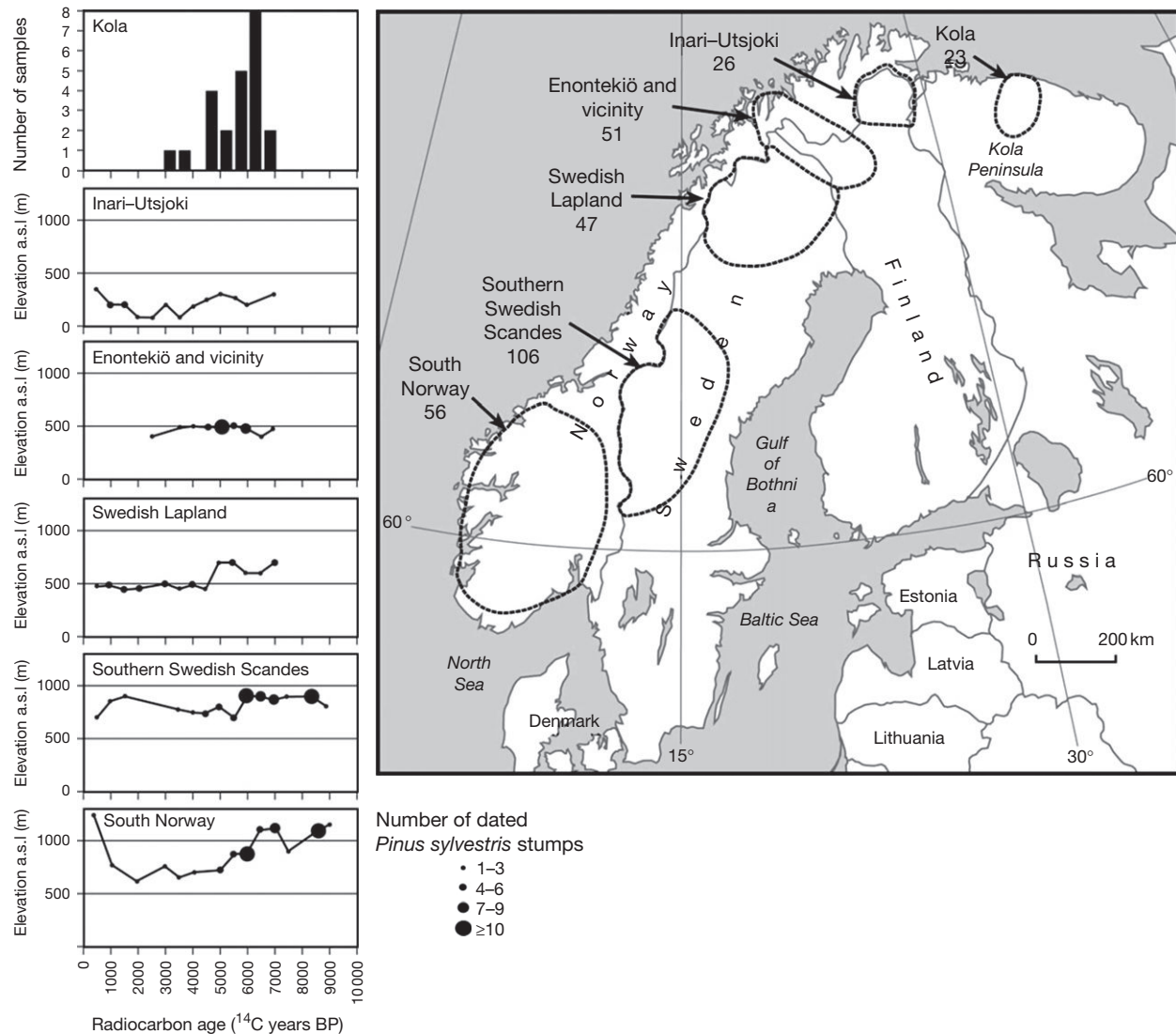


Figure 7 *Pinus sylvestris* megafossil records from the Kola Peninsula compared with evidence from Fennoscandia from studies compiled by Eronen and Huttunen (1993). Reproduced from MacDonald GM, Gervais BR, Snyder JA, Tarasov GA, and Borisova OK (2000) Radiocarbon dated *Pinus sylvestris* L. wood from beyond treeline on the Kola Peninsula, Russia. *The Holocene* 10: 143–147.

Table 2 Ages of dated wood from two sites in the Scandes Mountains, Sweden

Location	Species	Age (^{14}C years BP)	Age (cal years BP (1 σ))	Lab number
Mt Sylarna	<i>Pinus sylvestris</i>	8010 $\pm$ 60	8930–9002	Beta-99191
Krabbfjällnäset	<i>Larix sibirica</i>	7550 $\pm$ 60	8330–8410	Beta-99414
	<i>Pinus sylvestris</i>	8240 $\pm$ 60	9121–9302	Beta-96580
	<i>Picea abies</i>	7410 $\pm$ 90	8231–8314	Beta-99201
	<i>Picea abies</i>	4750 $\pm$ 70	5333–5347	Beta-96579
	<i>Betula pubescens</i>	8190 $\pm$ 90	9072–9144	Beta-96581

Data from Kullman L (1998) Palaeoecological, biogeographical and palaeoclimatological implications of early Molocene immigration of *Larix sibirica* Ledeb. into the Scandes Mountains, Sweden. *Global Ecology and Biogeography Letters* 7: 181–188.

The range extension was over 100 km in some instances, and summer temperatures were likely at least 2.5 °C warmer than present (MacDonald et al., 2000b). In some regions such as central Siberia it is possible that summer temperatures were

as much as 5–7 °C warmer than present. Between 4500 and 3200 cal years BP (4000 and 3000 ^{14}C years BP) there was a seemingly synchronous retreat of treeline to near modern positions.

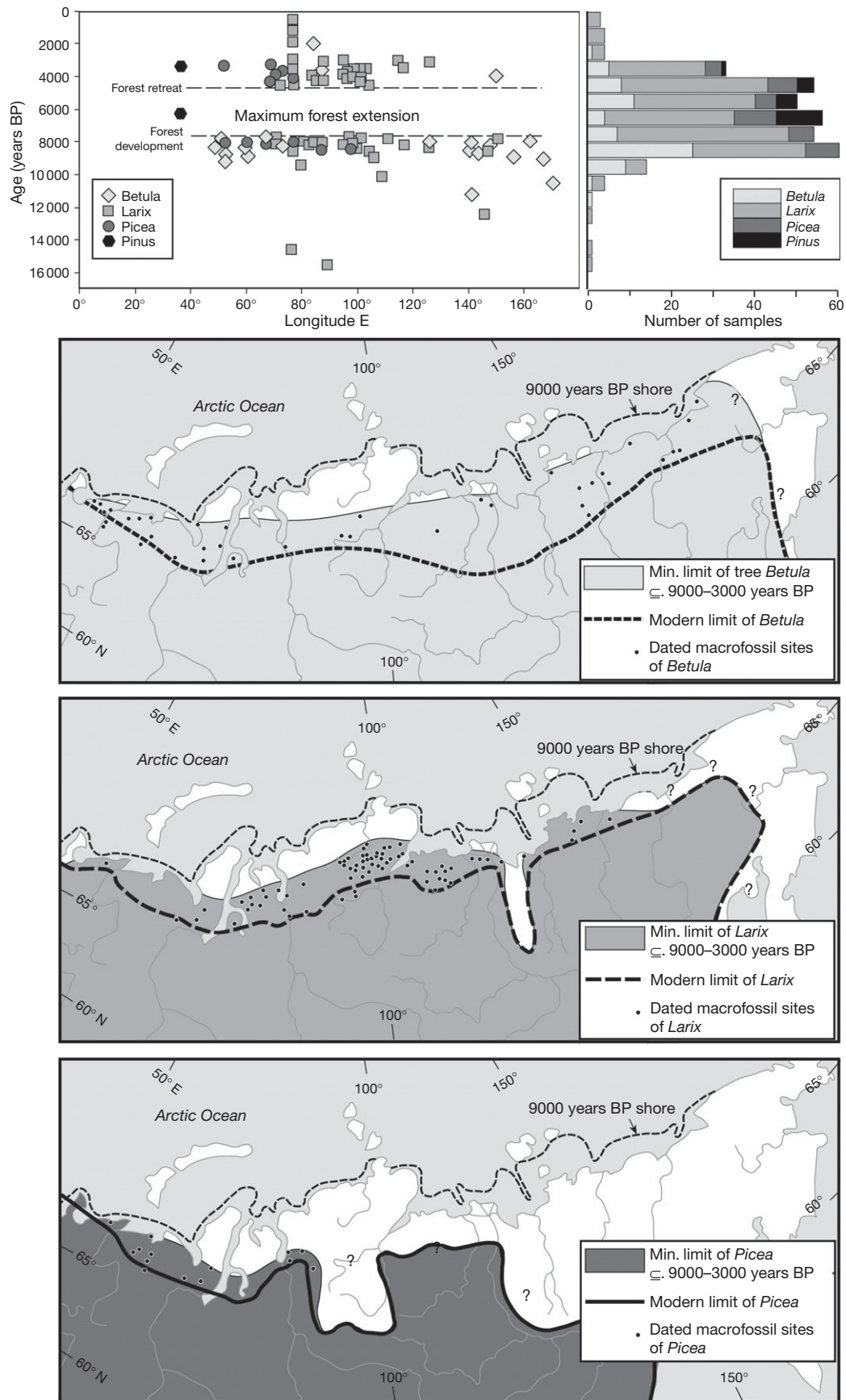


Figure 8 Holocene treeline advance and retreat from northern Russian Eurasia based upon wood megafossils. Reproduced from MacDonald GM, Velichko AA, Kremenetski CV, et al. (2000) Holocene treeline history and climate change across northern Eurasia. *Quaternary Research* 53: 302–311.

The abundance of megafossils in portions of northern Siberia allow for multicentennial tree-ring records to be produced (e.g., Shiyatov et al., 1996). Recently, for example, a detailed dendrochronological and radiocarbon-based study of subfossil *L. sibirica* megafossils was reported from the southern edge of the Yamal Peninsula in western Siberia (Hantemirov and Shiyatov, 2002). That study produced a remarkable 4000-year-long dendrochronological record. Most of the surface finds in the region date back only about 750 years, but older megafossils are abundant in river bank exposures. The abundance of wood and documented presence of additional wood, dated at more than 9000 years old, provide the possibility of developing a Holocene length record of tree-ring-based summer temperature estimates from the region.

See also: Dendrochronology; Plant Macrofossil Introduction. **Paleoclimate:** Paleoclimate History of the Arctic. **Plant Macrofossil Methods and Studies:** Surface Samples, Taphonomy, Representation; Treeline Studies.

References

- Arseneault D and Payette S (1997) Reconstruction of millennial forest dynamics tree remains in a subarctic tree line peatland. *Ecology* 78: 1873–1883.
- Barnekow L (1999) Holocene tree-line dynamics and inferred climatic changes in the Torneträsk area, northern Sweden, based on macrofossil and pollen records. *The Holocene* 9: 253–265.
- Birks HH, Larsen E, and Birks HJB (2005) Did tree-*Betula*, *Pinus* and *Picea* survive the last glaciation along the west coast of Norway? A review of the evidence, in light of Kullman (2002). *Journal of Biogeography* 32: 1461–1471.
- Core HA, Cote WA, and Day AC (1979) *Wood Structure and Identification*, p. 124. Syracuse: Syracuse University Press.
- Eronen M (1979) The retreat of pine forest in Finnish Lapland since the Holocene climatic optimum: A general discussion with radiocarbon evidence from subfossil pines. *Fennia* 157: 93–114.
- Eronen M and Huttunen P (1993) Pine megafossils as indicators of Holocene climatic changes in Fennoscandia. *Paläoklimaforschung-Palaeoclimate Research* 9: 29–40.
- Eronen M, Hyvärinen H, and Zetterberg P (1999) Holocene changes in humidity inferred from lake sediments and submerged Scots pines dated by tree rings. *The Holocene* 9: 569–580.
- Eronen M, Zetterberg P, Briffa KR, Lindholm M, Meriläinen J, and Timonen M (2002) The supra-long Scots pine tree-ring record for Finnish Lapland. Part 1: Chronology construction and initial inferences. *The Holocene* 12: 673–680.
- Foley J, Kutzbach JE, Coe MT, and Levis S (1994) Feedbacks between climate and boreal forests during the Holocene epoch. *Nature* 371: 52–54.
- Gajewski K (2002) Modern pollen deposition in lake sediments from the Canadian Arctic. *Arctic, Antarctic, and Alpine Research* 34: 26–32.
- Gervais BR and MacDonald GM (2001) Modern pollen and stomate deposition in lake surface sediments from across the treeline on the Kola Peninsula, Russia. *Review of Palaeobotany and Palynology* 114: 223–237.
- Gervais BR, MacDonald GM, Snyder JA, and Kremenetski CV (2002) *Pinus sylvestris* treeline development and movement on the Kola Peninsula of Russia: Pollen and stomate evidence. *Journal of Ecology* 90: 627–638.
- Hantemirov RM and Shiyatov SG (2002) A continuous multimillennial ring-width chronology in Yamal, northwestern Siberia. *The Holocene* 12: 717–726.
- Hoadley RB (1990) *Identifying Wood; Accurate Results with Simple Tools*, p. 224. Newtown: Taunton Press.
- Khotinsky NA (1984) Holocene vegetation history. In: Velichko AA (ed.) *Late Quaternary Environments of the USSR*, pp. 179–200. Minneapolis: University of Minnesota.
- Kremenetski KV, Boettger T, MacDonald GM, Vaschalova T, Sulerzhitsky L, and Hiller A (2004) Medieval climate warming and aridity as indicated by multiproxy evidence from the Kola Peninsula, Russia. *Palaeogeography, Palaeoclimatology, Palaeoecology* 209: 113–125.
- Kremenetski CV, Sulerzhitsky LD, and Hantemirov R (1998) Holocene history of the northern range limits of some trees and shrubs in Russia. *Arctic and Alpine Research* 30: 317–333.
- Kullman L (1998a) Non-analogous tree flora in the Scandes Mountains, Sweden, during the early Holocene: Macrofossil evidence of rapid geographic spread and response to palaeoclimate. *Boreas* 27: 153–161.
- Kullman L (1998b) Palaeoecological biogeographical and palaeoclimatological implications of early Holocene immigration of *Larix sibirica* Ledeb. into the Scandes Mountains, Sweden. *Global Ecology and Biogeography Letters* 7: 181–188.
- Kullman L (1999) Early Holocene tree-growth at a high-elevation site in the northernmost Scandes of Sweden (Lapland). A palaeobiogeographical case study based on megafossil evidence. *Geografiska Annaler* 81A: 63–74.
- Kullman L (2004) Early Holocene appearance of mountain birch (*Betula pubescens* ssp. *tortuosa*) at unprecedented high elevations in the Swedish Scandes: Megafossil evidence exposed by recent snow and ice recession. *Arctic, Antarctic, and Alpine Research* 36: 172–180.
- Kullman L (2005) On the presence of late-glacial trees in the Scandes. *Journal of Biogeography* 32: 1499–1500.
- MacDonald GM (2002) The boreal forest. In: Orme AR (ed.) *The Physical Geography of North America*, pp. 270–290. Oxford: Oxford University Press.
- MacDonald GM, Edwards TWD, Moser KA, Pienitz R, and Smol JP (1993) Rapid response of treeline vegetation and lakes to past climate warming. *Nature* 361: 243–246.
- MacDonald GM, Gervais BR, Snyder JA, Tarasov GA, and Borisova OK (2000a) Radiocarbon dated *Pinus sylvestris* L. wood from beyond treeline on the Kola Peninsula, Russia. *The Holocene* 10: 143–147.
- MacDonald GM, Velichko AA, Kremenetski CV, et al. (2000b) Holocene treeline history and climate change across northern Eurasia. *Quaternary Research* 53: 302–311.
- Ritchie JC (1984) *Past and Present Vegetation of the Far Northwest of Canada*. Toronto: University of Toronto Press.
- Ritchie JC, Cwynar LC, and Spear RW (1983) Evidence from northwest Canada for an early Holocene Milankovitch thermal maximum. *Nature* 305: 126–128.
- Ritchie JC and Hare FK (1971) Late-Quaternary vegetation and climate near the arctic tree line of northwestern North America. *Quaternary Research* 1: 331–342.
- Shiyatov SG, Hantemirov RM, Schweingruber FH, Briffa KR, and Moell M (1996) Potential long-chronology development on the Northwest Siberian Plain: Early results. *Dendrochronologia* 14: 13–29.
- Spear RW (1983) Paleoeecological approaches to a study of treeline fluctuations in the Mackenzie Delta Region, Northwest Territories: Preliminary results. In: Morisset P and Payette S (eds.) *Tree-Line Ecology. Nordicana*, vol. 47, pp. 61–72. Québec, Canada: Presses de l'université Laval.
- Wang L, Payette S, and Bégin Y (2001) 1300-year tree-ring width and density series based on living, dead and subfossil black spruce at tree-line in Subarctic Québec, Canada. *The Holocene* 11: 333–341.
- Zetterberg P, Eronen M, and Lindholm M (1996) Construction of a 7500-year tree ring based record for Scots pine (*Pinus sylvestris* L.) in northern Fennoscandia and its application to growth variation and paleoclimatic studies. In: Spiecker H, Mielikäinen K, Kohl M, and Skovsgaard JP (eds.) *Growth Trends in European Forests*, pp. 7–18. Berlin: Springer.

Dendroarchaeology

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Introduction

Dendrochronology, the science of tree-ring dating, is the most accurate and precise nondocumentary dating method available to researchers studying terrestrial records of the recent past. Tree-ring dates are accurate and precise to the year, and sometimes season, and have no associated statistical uncertainty or standard error. All other archaeological dating techniques that use natural materials (e.g., radiocarbon and archeomagnetism) have been calibrated using dendrochronological samples (Struiver, 1978). It is this precision and accuracy that has allowed dendroarchaeologists to construct detailed cultural chronologies in many parts of the world, and to explore a plethora of environmental, social, and behavioral questions regarding past human adaptations.

Developed in the early twentieth century by Andrew Ellicott Douglass (Figure 1) in the southwestern US (Nash, 1998; Webb, 1983), dendrochronology is probably best known as a technique that provides absolute Christian calendar dates for archaeological sites in the US southwest. Dendroarchaeological dating has been used to date sites, structures, artifacts, and such cultural phenomena as phases and periods (Dean, 1996). Dendroarchaeological is practiced in many areas, but is used most in the US southwest due to factors of history and preservation.

Tree-Ring Requirements

Trees must exhibit four attributes in order to be dendrochronologically useful (Fritts, 1976, 1991). First, they must produce distinguishable annual growth layers (rings) (Figure 2). Second, an individual ring must grow in a uniform manner around the bole of the tree, or branch (i.e., it must exhibit circuit uniformity). Third, the rings must exhibit some type of annual variability, such as ring width, ring density, etc. Finally, a sample must contain a sufficient number of rings to permit the identification of variability patterns; in southwestern conifers, 50–100 rings are usually sufficient; in other areas and with other species like European oak, 200+ rings may be necessary (Baillie, 1982).

The most fundamental principle of dendrochronology is crossdating — matching patterns of annual variability among local or regional tree populations (Douglass, 1941). In the US southwest, the skeleton plot method of crossdating has been used successfully for decades. Dendrochronologists in eastern North America, Europe, and the Middle East who examine angiosperms, however, use a quantitative method of crossdating, based on precise ring-width or ring-density measurements (Baillie, 1995). Both methods work equally well in their respective areas. If the variability patterns on a sample do not exactly match the tree population, the sample does not cross-date and cannot be used in dendrochronology. In this critical

respect, crossdating is fundamentally different than counting rings or estimating tree age.

Crossdating allows dendrochronologists to assign a single calendar year to each and every ring on a sample, but not all trees or archaeological samples produce dates. Some trees produce cambium in response to microenvironmental factors that are not reflected in the tree population; others respond to nonclimatic factors such as nutrient availability; and still others do not produce annual rings.

In order to construct a chronology, the skeleton-plotted samples from living trees, standing snags, remnant wood, and archaeological samples are combined into a master skeleton plot (Stokes and Smiley, 1968). By overlapping the plots of individual specimens, the chronology can be extended backwards in time (Figure 3) until a lack of sample depth precludes crossdating. Specific small ‘marker rings’ that occur on a large proportion (usually >75 percent) of samples in a collection help establish the basic pattern which is then tested against additional samples and nearby chronologies. In addition to small size, marker rings may be identified by internal features such as frost-damaged cells, false or double rings, unusually wide or narrow latewood bands, or other microscopically visible attributes.

Dendroarchaeological Date Attributes

Even when samples produce dates, they must still be interpreted, and not all archaeological tree-ring dates are the same. Because dendroarchaeological samples are the result of past human behaviors, specific sample attributes and the archaeological context must be considered in any interpretation of the chronological materials. Sample attributes can be used to identify cutting, near-cutting, and noncutting tree-ring dates.

Cutting dates, also known as tree-death dates, retain the last cambial layer grown by the tree and indicate that the tree was cut in a specific year (assuming tree death resulted from human harvesting). Evidence on the sample that it is a cutting date includes the presence of bark, beetle galleries, a shiny patina, or a continuous ring around the sample (Ahlstrom, 1985; Towner, 1997). In the literature, the cutting date is denoted by the symbol ‘v,’ — noting that in the opinion of the analyst, the last ring on the sample is the last ring grown by the tree.

Samples that retain the last cambial layer can also be assigned a cutting (death) season. Those that exhibit a terminal ring with a full complement of latewood were cut after the growing season for that particular year. Samples that show only earlywood cells were cut during the growing season. Because different tree species have different growing seasons, cutting dates in a structure in the same year from different species can confine construction to a relatively small time period, as little as 4–6 weeks in some cases (Dean and Warren, 1983; Towner and Dean, 1992).

Samples that yield near-cutting dates also retain the last ring grown by the tree; however, these particular samples may (or may not) contain a locally absent or missing ring near the end of the sample ring sequence, and are denoted by a '+' symbol (Ahlstrom, 1985; Towner, 1997). There is no way to verify the absence (or presence) of the missing rings near the end of the ring sequence, so near-cutting dates should be considered within a year or two of tree-death dates.

Noncutting dates result from two different processes: exterior ring loss or ring counts near the outside of the sample ring

sequence (and are denoted by 'vv' or '++', respectively) (Ahlstrom, 1985; Bannister, 1962; Haury, 1935). Samples that do not retain the last ring grown by the tree (vv dates) have suffered exterior ring loss either through natural processes such as erosion, or cultural processes such as beam shaping. Thus, a sample dated AD 790–957 vv could not have been cut prior to AD 957, but because it may be missing one, 10, or even 100 exterior rings, the harvesting date cannot be determined with any degree of confidence. Partially ring-counted specimens (++ dates) result from a lack of crossdating on the sample beyond a specific year. Ahlstrom (1985) suggests that ++ dates may indicate deadwood use because as trees die a slow natural death, they respond less and less to macroenvironmental conditions and produce more sporadic cambial growth layers. Noncutting dates (both vv and ++) provide only a *terminus post quem*, a date before which tree death could not have occurred; a noncutting date may predate the actual use of a beam by years, decades, or even centuries.

Dendroarchaeological Sampling and Sample Preparation

Dendroarchaeological samples may be collected in several different ways. Most commonly, they are recovered as charcoal pieces during excavations. Such specimens need to be handled gently during transport to the laboratory. They should be wrapped in cotton batting and string to protect them, but need no special stabilization. Another common method is extracting cores from intact beams (Figure 4), using a specially made drill bit similar to an elongated hole saw. The resulting cores ($\frac{1}{2}$ –1" diameter) retain the rings in the beam, although they may exhibit more locally absent rings than cross sections due to their small size. Finally, dendroarchaeological wood samples can be collected as loose pieces or cut into cross sections, but such techniques have a greater impact on the archaeological record.

After transport to the laboratory, dendroarchaeological samples are prepared for analysis under the microscope. Charcoal specimens require no surfacing and are simply broken to expose a fresh surface. Wood samples are sanded with increasingly finer grit sandpaper, sometimes up to 500 grit, to expose the individual cells and ring structure. All samples are then examined under a binocular microscope at 10–40× power.



Figure 1 Andrew Ellicott Douglass, the founder of dendrochronology.

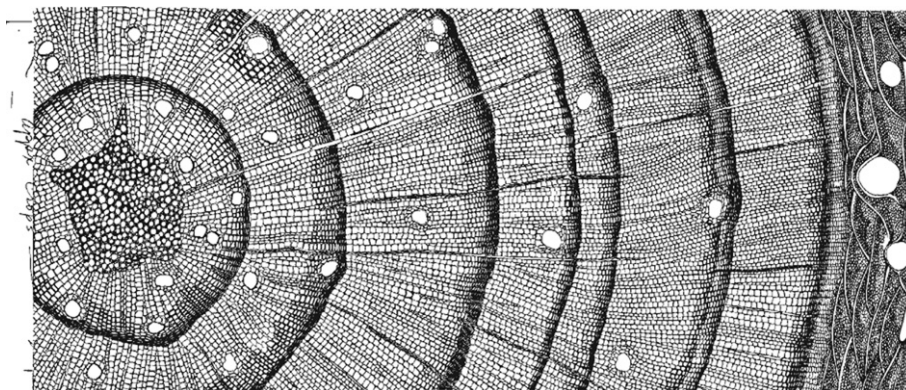


Figure 2 Schematic of conifer tree rings.

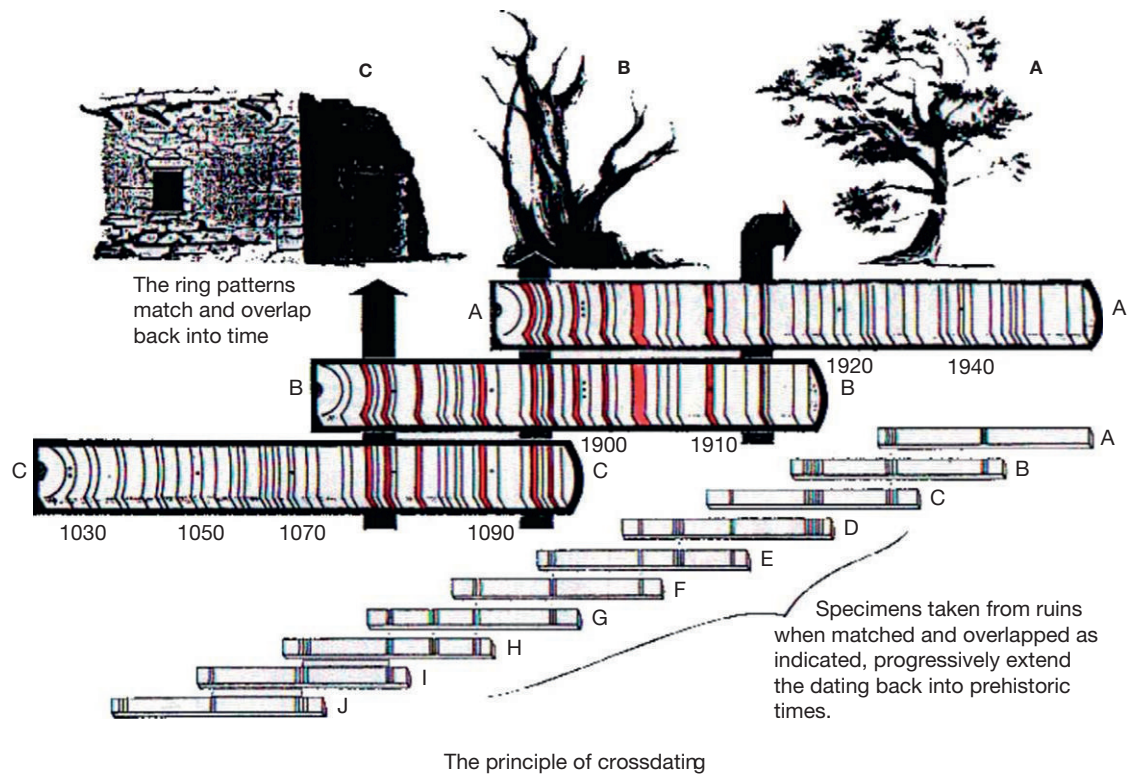


Figure 3 The process of chronology-building.



Figure 4 Ronald H. Towner collecting a dendroarchaeological sample.

Chronological Information from Dendroarchaeological Samples

The keys to determining when a room, site, or artifact was constructed are: (1) identifying date clusters; and, (2) explaining anomalous dates. Identifying anomalous dates requires (1) an adequate sample of dates and (2) defining date clusters. An adequate number of dates is a relative term that depends on the number of samples available for collection, the number of samples collected, and the number of dates derived. Single dates provide limited information of any kind. Ahlstrom (1985) has defined a date cluster as 'three or more dates falling in a brief time interval,' but definitions vary among researchers and from collection to collection. Ahlstrom has adapted the stem-and-leaf technique to plotting tree-ring dates in order to provide a detailed visual representation of each date in a distribution. Applying the technique to archaeological tree-ring dates, he identified an 'ideal' date distribution as a single line on the plot consisting of entirely cutting dates in the same year (Ahlstrom, 1985). The author enlarged this concept (Figure 5) and proposed that different date distributions could be used to identify different wood-use behaviors. Although stem-and-leaf plots separate the dates from their archaeological contexts, they delineate the temporal attributes of samples and allow an initial assessment of the relationships among them. Because the stem-and-leaf technique ignores the archaeological context of tree-ring dates, however, it does not fully explain past human behavior. It is important to note, however, that although the tree-ring dates are precise and accurate, it is the

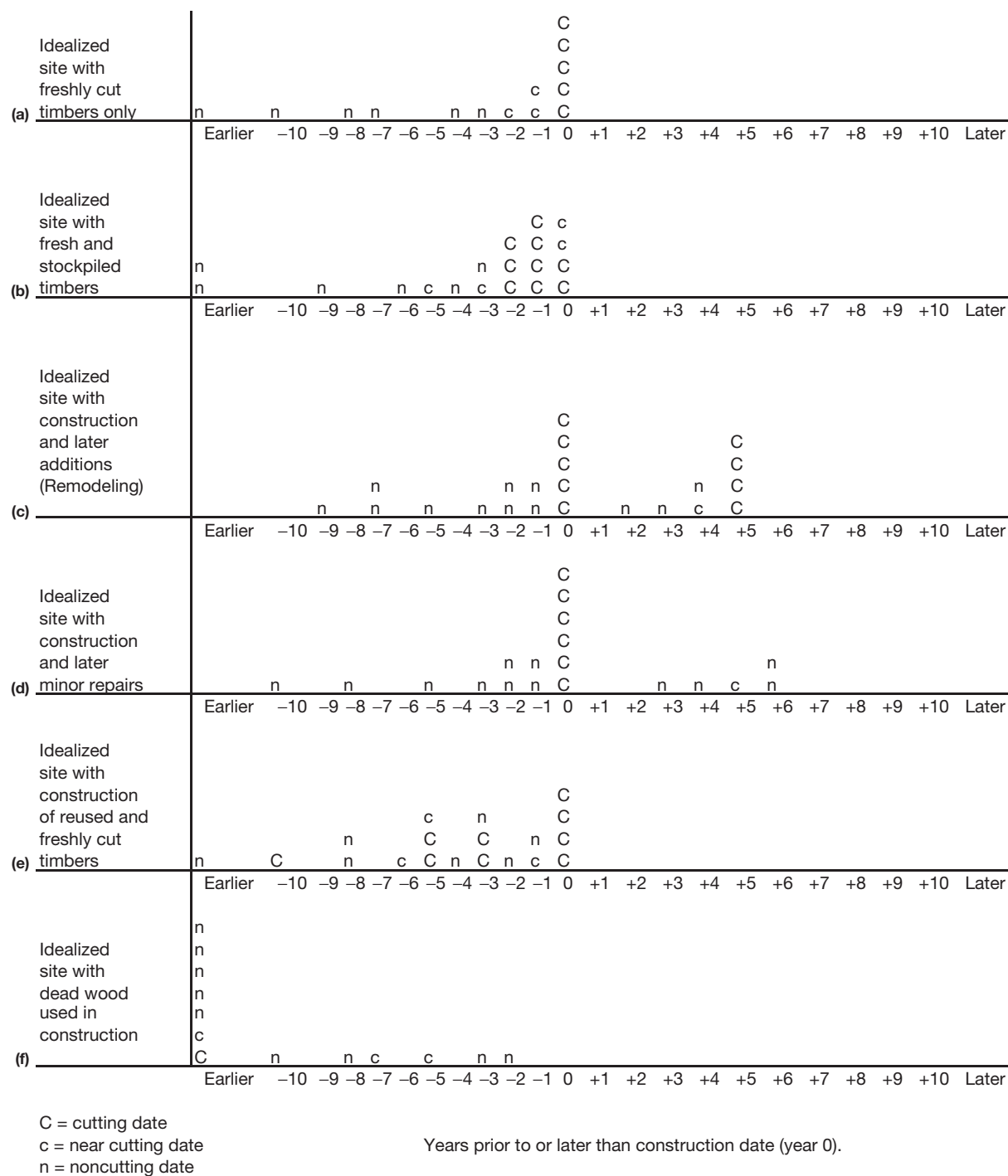


Figure 5 Idealized histograms of different wood use behaviors.

archaeological context and interrelationships between samples and date clusters that facilitates behavior interpretations.

Environmental Information from Dendroarchaeological Samples

The precise chronological data provided by tree-ring dates — even well provenienced cutting-date clusters — illuminate more than simply the temporal aspects of an occupation.

Embedded in the samples is also important environmental information which can be derived from dendroarchaeological samples in two ways. First, the tree species exploited by past peoples may indicate aspects of local species availability and thus, environmental composition. Second, dendroarchaeological samples can be used to reconstruct past precipitation and/or temperature regimes and identify past 'extreme' climatic events.

The tree species present in archaeological tree-ring collections are the result of several factors. First, site occupants often

exploited the most available species; thus, archaeological samples can be used to assess past species distributions and forest composition (at least at the 'generic' level). For example, the Black Mesa Archaeological Project (BMAP) in northern Arizona collected more than 5000 tree-ring samples from Navajo archaeological sites that were occupied between 1800 and 1972 (Dean and Russell). The modern vegetation community of Black Mesa consists predominately of pinyon-juniper forest, isolated stands of ponderosa pine and Douglas-fir, and various nonconiferous species. The Navajo BMAP dendroarchaeological samples mirror this distribution both in terms of species availability and spatial distribution, that is, in areas with higher concentrations of one species, that species was most often used in structure construction (Dean and Russell). These data indicate little ecological change in the composition of Black Mesa forests over the past 200 years.

Species distributions in archaeological collections are affected by more than species availability, however. Human social and economic systems also affect the presence and proportions of species in collections. The example below demonstrates the influence of human tree selection that does not reflect local environmental variability.

Bonde and Christensen (1993) have dated several Viking ships built with oak timbers that demonstrate the impact of economic systems on the species distribution in archaeological sites from northern Europe. The sunken ships are located in Scandinavia, primarily Sweden and Denmark, but do not crossdate with local tree-ring chronologies. Through comparisons with several other northern European oak chronologies, Bonde and Christensen (1993) identified the source of the ship timbers as northern Ireland. They suggest that whereas ship-building was conducted in Denmark, the raw materials for construction were imported from across the North Sea, a process that certainly required significant organizational and technological skills.

Dendroclimatic Analyses

Climatic and cultural information inherent in archaeological tree-ring samples have been linked since the 1930s. When Douglass 'bridged-the-gap' between his modern and archaeological chronologies, it became apparent that the 'Great Drought' of the late 1200s in western North America had caused difficulties in crossdating because of many missing and microrings during the AD 1270–1300 period (Douglass, 1935; Nash, 1997). The cliff dwellings of Mesa Verde and other sites in the Four Corners area were abandoned at about the same time, which also significantly reduced the sample size available for analysis. The temporal correlation between the drought and abandonment suggested to many archaeologists a causal relationship between the former and the latter.

Since Douglass' early efforts, the use of dendroarchaeological samples to examine past environmental variability has contributed significantly to our understanding of past human-environment interaction (Dean et al., 1985; Van West, 1994). The low-elevation conifers found in most southwestern archaeological sites, including pinyon pine, juniper, Douglas fir, and ponderosa pine, typically respond to moisture (Fritts, 1991) and, therefore, precipitation reconstructions have been developed for much of the area.

In general, two different types of dendroclimatic studies have been conducted using dendroarchaeological samples: qualitative and quantitative. Qualitative studies use tree-ring chronologies as a direct measure of the relative variability in past climates; such qualitative reconstructions provide accurate indications of the relative deviation of annual precipitation (Dean and Robinson, 1977). Quantitative reconstructions require modern environmental data by which to calibrate tree growth and rigorous statistical testing to validate the results. Quantitative reconstructions use mathematical models to define the relationship between tree growth and specific environmental parameters, and statistically estimate the past variability of the environmental factors, such as inches of rainfall, degrees of temperature, etc. Such estimates are typically smoothed to identify long-term trends, but can be presented as annual or even seasonal values (Fritts, 1991).

These brief examples demonstrate the value of dendroarchaeological samples for eliciting information about past environmental conditions. By their very nature, dendroarchaeological samples also contain information about past human activities. It is a goal of dendroarchaeology to combine these data sets to further our understanding of how past peoples adapted to and modified their environments.

Behavioral Information from Dendroarchaeological Samples

The key to deriving behavioral information from dendroarchaeological samples is the identification of anomalous dates. Anomalous dates are defined as those dates that do not date the event of interest (Dean, 1978) and are therefore dependent upon the archaeological context and research question. The tree-ring dates themselves are precise and accurate; they date the last ring on the sample, but may not date the event of interest. For example, if one is interested in dating the construction of Balcony House in Mesa Verde, Colorado, one of the most visited cliff dwellings in the world, the 1910 cutting date from Room 3 is clearly anomalous. If on the other hand, the event of interest is dating the stabilization work of Jess Nusbaum, the 100+ dates in the AD 1200s are considered anomalous (Parks and Dean, 1990). Obviously, dates are only anomalous within a context, and all dates may be anomalous in different contexts.

The behavioral information inherent in archaeological tree-ring samples is directly affected by two major factors: the behavior of past peoples and the behavior of archaeologists and dendrochronologists. Past peoples must have: (1) used wood as a resource for building structures, making artifacts, and as fuel; (2) they must have exploited species that are appropriate for dendrochronological analysis; and, (3) they must have used wood in ways that insured its preservation in the archaeological record. Archaeologists influence tree-ring data by: (1) selecting sites for study that contain dendrochronological materials, (2) precisely recording the provenience and surface attributes of samples, and, (3) properly collecting and submitting the samples for analysis. Dendrochronologists must correctly date the samples and describe their microscopic attributes, i.e., terminal ring characteristics, false and microrings, etc.

Certainly, the above-described species selection and economic procurement systems represent past human behaviors. It is at the individual beam, room and structure, and site level, however, that most detailed human behaviors can be identified. Such behaviors include species-selection preferences, deadwood use, beam stockpiling, beam reuse, structure repair and remodeling, structure abandonment, and beam harvesting, preparation, and modification. When data from many beams, rooms, and sites are combined, it is possible to identify broad-scale patterns of how specific groups treated wood as a resource.

This is one area where dendroarchaeology differs from other subfields of dendrochronology. Single samples or site collections are related to specific past human activities and may not contain information relative to larger spatial and temporal issues. For example, the collection and analysis of 130 high-altitude bristlecone pine specimens from the San Francisco Peaks of northern Arizona has had important ramifications for reconstructing temperature variability in the entire Southwest and across western North America for the past two thousand years (Salzer, 2000). In contrast, the collection of 4294 archaeological samples from Pueblo Bonito in Chaco Canyon informs us about the activities of a specific group of people in a specific time and place (although environmental information is contained in those samples as well) (Windes and Ford, 1996). Dendroarchaeological specimens are part of a specific human behavioral context that may (or may not) be directly related to broader patterns.

The first level of analysis for illuminating past human behaviors is at the individual beam level. Precise provenience and sample attribute information can reveal aspects of beam function, procurement, and preparation. Beam function includes information about the architectural element and its use, such as a roof primary or secondary beam, and a door lintel or jamb. In many cases, specific species and sizes may have been preferred for specific architectural elements (cf. Dean and Warren (1983) and Windes and Ford (1996)). Beam procurement methods include cutting with various types of stone or metal axes or saws, breaking, or burning. Noting the procurement method on a sample may indicate the use of deadwood (breaking, burning) or provide a temporal framework for undated samples (i.e., different saw marks relate to technological innovations in saw technology). Beam preparation may include removing limbs, bark, shaping a beam, and preparing the beam ends using various tools. Beam preparation tools include various types of axes and saws as well as draw knives, axes, and grinding tools. Identifying the types of tools used to procure and prepare timbers may have implications for interpreting anomalous dates, distinguishing technological traditions, or recognizing cultural interaction.

At the room and structure level, it is necessary to document the architecture as completely as possible. Such characteristics as the bond-abut relationships of walls, changes in wall construction materials or plastering, changes in room function denoted by sealed doorways or covered hearths, and other architectural attributes can help in determining when the room was built (as opposed to when the beams were harvested) and provide clues to the nature of the occupation. When combined with beam attributes and dates, such information can help to identify anomalous dates, the use of

deadwood, the reuse of older beams from other rooms or sites, the stockpiling of beams, the repair and remodeling of rooms and structures, and can yield information relative to the duration of occupation and mode of abandonment of a room or structure. These combined data can help illuminate aspects of human social organization, such as the use of a room or structure by a family or supra-family group; they may also contribute to understanding the dynamics of that group through time, such as changing structure use in response to generational changes in family size, the immigration of new families into a settlement, and many others.

At the site and regional levels, dendroarchaeological data provide the temporal control necessary to delineate broad patterns of human behavior. At the site level, the initial founding of a site can often be determined through tree-ring analysis, and the duration of the occupation may also be delineated. An excellent example of the former is Douglass' dating of the Mesa Verde cliff dwellings (Douglass, 1929). Prior to the advent of dendroarchaeology, archaeologists debated whether the structures were hundreds or thousands of years old. Douglass' precise dating of these ruins to the thirteenth century had profound implications for archaeological and anthropological theories of the rate of human cultural development in the New World. Similarly, before large-scale sampling efforts and detailed analysis documented the duration of occupation of the Kayenta Anasazi cliff dwelling of Betatakin at less than 40 yr (Dean, 1969), many people assumed such structures had been occupied for hundreds of years. Again, such temporal compression has required anthropologists to reexamine their theories of cultural evolution, and has forced archaeologists to consider the importance of settlement mobility patterns in interpretations of population demographics and history.

At the regional level, dendroarchaeological data contribute to understanding population dynamics, culture change, and interaction in a variety of ways. For example, recent dendroarchaeological analysis of several Navajo pueblito structures in Palluche Canyon, New Mexico, reveals very similar initial construction, remodeling, and abandonment dates (Ababneh et al., 2002). These small three to five room structures were all apparently constructed, temporarily abandoned, reoccupied, and finally abandoned within a period of about 20 yr. The archaeological and dendrochronological data suggest all the sites were occupied by the same extended family group, and this inference is supported by ethnographic data and Navajo oral traditions (Ababneh et al., 2000). Thus, these sites can now be related to a specific period in Navajo history, a specific form of social organization, and perhaps to a particular descendent group of Navajos.

On a larger scale, dendroarchaeological data have been used to illuminate aspects of significant migrations both during the prehistoric period and during the early historic period. Recently, the author has suggested that the Navajo immigration out of their Dinétah homeland in northwest New Mexico and into northeastern Arizona during the 1700s was a social and economic process unrelated to the single-year drought of 1748 (Towner, 2003). These large databases enable archaeologists and dendrochronologists to investigate important questions concerning how humans, as individuals, families, and supra-family groups, colonized landscapes, exploited

their environments, interacted with their neighbors, and responded to changes in their physical, social, and technological environments.

Conclusions

Dendroarchaeology is a powerful tool for studying the recent past, but it has certain limitations. As noted above, the two most important factors in dendroarchaeological research are the behavior of past peoples toward wood as a resource and the behavior of archaeologists and dendrochronologists. Fortunately, the US southwest was occupied by groups who used those tree species to build structures, make artifacts, cook, and heat their homes. In some areas, such as the extreme deserts of North and South America, North Africa, and Australia, people simply did not use wood (or did not use dendrochronologically useful species). Dendroarchaeological research in those areas may be extremely limited. Even in areas with appropriate tree species, past human behavior still constrains the application of dendroarchaeology. For example, the longest archaeological tree-ring record in the southwest extends back only to 323 BC. This limit is not imposed by tree species or preservation, but results from the fact that Archaic period populations did not build substantial structures or use large conifer beams for fuel.

The most important factor in the success or failure of dendroarchaeological research in a given area is institutional support. The Southwest, the United Kingdom, Germany, France, and Switzerland have been in the forefront of dendroarchaeological research for decades simply because they have had well-qualified dendrochronologists and funded research programs. The importance of such programs cannot be overstated.

Dendrochronology is the most precise dating method available to scientists studying the recent past. Dendroarchaeology is best known as a technique that provides Christian calendar dates for large archaeological sites. Dendroarchaeology is much more than simply a chronometric technique however; it also provides data relevant to understanding past environmental conditions and change, as well as information important for illuminating aspects of past human behavior at various spatial scales. When these chronological, environmental, and behavioral data are combined, dendroarchaeology can help illuminate aspects of past human interaction with the physical, social, and technological environments.

See also: Dendrochronology; Dendroclimatology. **Glacial Landforms, Tree Rings:** Dendrogeomorphology; Dendroglaciology.

References

- Ababneh LN, Towner RH, Prasciunas MM, and Porter KT (2000) The dendrochronology of Palluche Canyon, Dinéah. *Kiva* 66: 267–290.
- Ahlstrom RVN (1985) The Interpretation of Archaeological Tree-Ring Dates. PhD Dissertation University of Arizona, Tucson, Ann Arbor: University Microfilms.
- Baillie MGL (1982) *Tree-Ring Dating and Archaeology*. Chicago: The University of Chicago Press.
- Baillie MGL (1995) *A Slice Through Time*. London: BT Batsford.
- Bannister B (1962) The interpretation of tree-ring dates. *American Antiquity* 27: 508–514.
- Betancourt JL, Dean JS, and Hull HM (1986) Prehistoric long-distance transport of construction beams, Chaco Canyon, New Mexico. *American Antiquity* 51: 370–374.
- Bonde N and Christensen AE (1993) Dendrochronological dating of the viking ship burials at Oseberg, Gokstad, and Tune, Norway. *Antiquity* 67: 575–583.
- Dean JS (1969) A chronological analysis of Tsegi phase sites in northeastern Arizona. *Papers of the Laboratory of Tree-Ring Research No. 3*. Tucson: University of Arizona.
- Dean JS (1978) Independent dating in archaeological analysis. In: Shiffer MB (ed.) *Advances in Archaeological Method and Theory*, vol. 1, pp. 223–255. New York: Academic Press.
- Dean JS (1996) Behavioral sources of error in archaeological tree-ring dating: Navajo and Pueblo wood use. In: Dean JS, Meko DM, and Swetnam TS (eds.) *Tree-Rings, Environment, and Humanity: Proceedings of the International Conference*, pp. 497–503. Tucson, AZ 17–21 May 1994. Tucson: Radiocarbon.
- Dean JS and Robinson WJ (1977) *Dendroclimatic Variability in the American Southwest, A.D.*, pp. 680–1970. Tucson: Laboratory of Tree-Ring Research, University of Arizona.
- Dean, J. S. and Russell, S. C. *Navajo Wood Use behavior*. Tucson: Laboratory of Tree-Ring Research, University of Arizona.
- Dean JS and Warren RL (1983) Dendrochronology. In: Lekson SH (ed.) *The Architecture and Dendrochronology of Chetro Kettle, Reports of the Chaco Center No. 6*, pp. 105–240. Albuquerque: National Park Service, Division of cultural Research.
- Dean JS, Euler RS, Gemerman GJ, Piog F, Hevely RH, and Karlstrom TNV (1985) Human behavior, demography, and paleoenvironment on the Colorado plateau. *American Antiquity* 50: 537–554.
- Douglass AE (1929) The secret of the southwest solved by talkative tree rings. *National Geographic Magazine* 56: 736–770.
- Douglass AE (1935) Dating Pueblo Bonito and other ruins in the southwest. *National Geographic Society Contributed Technical Papers, Pueblo Bonito Series No. 1*. Washington.
- Douglass AE (1941) Crossdating in dendrochronology. *Journal of Forestry* 39: 825–831.
- Fritts HC (1976) *Tree-Rings and Climate*. London: Academic Press.
- Fritts HC (1991) Reconstructing Large-Scale Climatic Patterns from Tree-Ring Data. Tucson: University of Arizona Press.
- Hauray EW (1935) Tree-rings—the archaeologist's time piece. *American Antiquity* 1: 98–108.
- Nash SE (1998) Time for collaboration: A.E. Douglass, archeologists, and the development of tree-ring dating in the American Southwest. *Journal of the Southwest* 40: 261–305.
- Nash SE (1999) *Time, Trees, and Prehistory: Tree-Ring Dating and the Development of North American Archaeology*. Salt Lake City: University of Utah Press.
- Parks JA and Dean JS (1990) Tree-ring dating of Balcony House: A chronological architectural, and social interpretation. In: Fiero K (ed.) *Balcony House: A History of a cliff Dwelling, Mesa Verde National Park, Colorado*, pp. 91–104. Colorado: Mesa Verde Museum Association, Mesa Verde National Park.
- Salzer MW (2000) Dendroclimatology of the San Francisco Peaks Region of Northern Arizona. PhD Dissertation, Tucson: Geosciences Department, University of Arizona.
- Stokes MA and Smiley TL (1968) *An Introduction to Tree-Ring Dating*. Chicago: University of Chicago Press.
- Stuiver M (1978) Radiocarbon timescale tested against magnetic and other dating methods. *Nature* 273: 271–274.
- Towner RH, Salzer MW, Parks JA, and Barlow RK (2009) *Assessing the Importance of Past Human Behavior in Dendroarchaeological Research: Examples from Range Creek Canyon, Utah, U.S.A Tree-Ring Research* 65: 117–127.
- Towner RH (1997) The Dendrochronology of the Navajo Pueblos of Dinéah. PhD Dissertation, Department of Anthropology, University of Arizona, Tucson. Ann Arbor: University Microfilms.
- Towner RH (2003) *Defending the Dinéah*. Salt Lake City: University of Utah Press.
- Towner RH and Dean JS (1992) LA 2298: The oldest Pueblo revisited. *Kiva* 59: 315–331.
- Van West CR (1994) Modeling prehistoric agricultural productivity in southwestern Colorado: A GIS approach. *Reports of Investigations No. 67. Department of Anthropology*. Pullman: Washington State University.
- Webb GS (1983) *Tree-rings and Telescopes: The Scientific Career of A.E. Douglass*. Tucson: University of Arizona Press.
- Windies TC and Ford D (1996) The Chaco Wood Project: The chronometric reappraisal of pueblo Bonito. *American Antiquity* 61: 295–310.

Mire and Peat Macros

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Introduction

The potential scope of ‘mire and peat macrofossils’ is enormous, given both the substantial range of plant macrofossils preserved in peat sequences and the ~400 million hectare global extent of peat deposits (Barber and Charman, 2003). Peat is defined as accumulated sediment which contains at least 30% dry mass of dead organic material, and mires as peatlands where peat is currently being formed. Therefore this article is focused on Holocene northwest European raised peat bog (ombrotrophic) and fen (minerotrophic peatland) deposits (see Wheeler and Proctor, 2000, for a detailed review of peatland terminology), subjects in which the authors have greatest expertise and experience. In this article, how mire and peat macrofossil analyses have been used to investigate Quaternary environmental change, and the methodologies used to prepare and analyze the samples are discussed. A review of the paleoecological indicator value and identification notes of a representative selection of mire and peat plant macrofossils forms a major part of this article. We include a series of photographic figures as an aid to macrofossil identifications. It is certainly not a comprehensive selection, but serves to extend and complement the identification keys and photographs/illustrations currently available. The figures will not only be of value to paleoecological analyses of mire and peat macrofossils in northwest Europe. Some of the species we describe here have also been recovered from peatlands in southern Québec, Canada (Pellerin and Lavoie, 2003); the Western Great Lakes Region, USA (Booth et al., 2004); Tierra del Fuego, Argentina (Mauquoy et al., 2004a); and the East-European Russian Arctic (Välranta et al., 2003).

Applications

Environmental change reconstructions based on mire and peat macrofossil analyses are limited, in that full plant assemblages cannot be reconstructed and directly compared with detailed neophytosociological studies. This problem occurs due to the selective nature of vegetation decomposition in peatlands, since even decay-resistant *Sphagnum* mosses have different decomposition rates (Johnson and Damman, 1991). Due to the potential for differential preservation and representation of the bog vegetation, successional and/or paleoclimate reconstructions produced from subfossil assemblages need to be interpreted cautiously. Furthermore it is not always possible to identify *Sphagnum* spp. to the lowest taxonomic level, particularly where their stem leaves are absent. Cyperaceous (sedge and allies) remains are also difficult to identify consistently, since in many instances seeds and diagnostic epidermal tissues are simply not present.

Subfossil plant macrofossils preserved in peat samples have been used to reconstruct Holocene raised peat bog vegetation

succession at microform scales (analyses of a single borehole to reconstruct vegetation changes in hummock/hollow microform complexes see McMullen et al., 2004; van der Molen, 1988; van Geel, 1978) and mesoform scales (transects of boreholes analyzed across single raised peat bogs see Barber et al., 1998; Hughes et al., 2000; Svensson, 1988). The aim of much of this research on ombrotrophic (rain-fed) peat bog stratigraphy has been paleoclimate reconstructions, since changes in bog surface wetness recorded by plant macrofossils are highly likely to represent changes in precipitation and temperature (Barber and Charman, 2003; Barber et al., 2004). The macrofossil data produced from this research are equally valuable in peatland carbon accumulation rate studies (Oldfield et al., 1997) and the identification of successional processes in peatlands, for example, the nature and timing of the fen/bog transition (Hughes and Barber, 2004). Applied paleoecological research using plant macrofossil data has provided long-term baseline data for the origin and nature of vegetation changes, which can be used to guide conservation management in environmentally sensitive areas (Chambers et al., 1999). The nature and timing of recent human disturbance to ombrotrophic bogs, for example burning and drainage, has also been assessed (Pellerin and Lavoie, 2003). Dating peat samples using ^{14}C accelerator mass spectrometry has now become the standard method, and it offers the possibility of dating above-ground parts of identified plants, for example, *Sphagnum* leaves or fruits of *Rhynchospora alba*. This is highly advantageous, since cyperaceous roots and the stem bases and attached lower parts of *Eriophorum vaginatum* leaves can be avoided. These have been shown to be younger (because of downward growth) than bryophyte fragments in the same ^{14}C -dated stratigraphic levels (Kilian et al., 2000). Plant macrofossil expertise can be used to pinpoint the best samples possible (above-ground macrofossils, preferably free of fungal contamination) for ^{14}C -dating peat sequences in order to create accurate and precise chronologies (Mauquoy et al., 2004b).

Methods

Analyses of plant macrofossils in peat deposits requires samples of ~5 cm³, and these can be recovered using a 7- or 9-cm diameter ‘Russian’ pattern corer, a Wardenaar peat sampler (to a maximum depth of 1 m), or where peat sections are exposed, 50 (100) × 15 × 10 cm metal boxes can be used. Analyses of the field stratigraphy of the peat bog (Barber et al., 1998) is highly desirable prior to taking the master core that will be subjected to detailed paleoecological analyses. This can be undertaken using a 5 cm diameter ‘Russian’ pattern corer (either 50 or 100 cm length). When sampling a ‘living’ peat bog, care should be taken to avoid compression of the surface layers of the peat deposits. Before subsampling, 5 mm of peat from the core should be trimmed to avoid contamination and

only clean instruments used to avoid the introduction of 'modern' carbon and/or contamination with older or younger macro/microfossils.

Plant macrofossil samples should be gently boiled with 5% KOH (deflocculation to dissolve humic and fulvic acids) and disaggregated. When sieving (on a 125 or 150 µm sieve) the plant remains must be kept just under the water surface (to avoid too much damage of delicate subfossils, e.g., *Sphagnum* stems with their stem leaves still attached). Stem leaves can be very important for *Sphagnum* identification, because they can help considerably in identification to species level. Macrofossils retained on the sieve are then transferred to a petridish and only enough distilled water should be added to float the remains, which are scanned using a low power ($\times 10$ –50) stereo zoom microscope. Moss leaves, cyperaceous epidermal tissues, and small seeds (*Juncus* spp.) need to be mounted onto temporary slides and examined at high magnifications ($\times 100$ –400).

Reference collections are essential in aiding and confirming the identification of mire and peat macrofossils. Above-ground (leaves, stems, flowers, fruits, and seeds) and below-ground (rhizomes and roots) parts of peatland plants can be collected during fieldwork or from a herbarium. The best reference collection, however, is one derived from fossil remains (Grosse-Brauckmann, 1986), where the identifications have been confirmed by an experienced analyst.

A range of techniques are available to present the data. The simplest technique is to use ordinal values, for example, 1 = rare, 2 = occasional, 3 = frequent, 4 = common, and 5 = abundant (Barber, 1981). Alternatives to this are to make quantitative estimates using volume percentages (see the quadrat and leaf count macrofossil analysis technique widely adopted by Barber and Charman (2003)) or absolute numbers, calculated as either concentrations (number of objects per unit volume, Booth et al., 2004; Janssens, 1983) or influx (based on age-depth models). All three techniques can be combined to present different macrofossil components from a single peat sequence, for example, mosses, dwarf shrubs, and cyperaceous roots can be expressed as percentages; bark, wood, and bud scales as ordinal values; and seeds as absolute numbers (Välranta et al., 2003).

Macrofossil Indicator Value and Descriptions of Northwest European Mire and Peat Macrofossils

Calluna vulgaris

Hammond et al. (1990) investigated the position of *Calluna vulgaris* (Figure 1(a)–1(f)) in relation to local raised peat bog water tables, and found that it occurred under minimum and maximum water table depths of 10 cm and 52 cm, respectively, with 90% of occurrences at 30 cm water table depth. It is restricted to drier microforms, because the roots and their endomycorrhizas need an aerated layer about 10–20 cm deep (Lindholm and Markkula, 1984).

Erica tetralix

Erica tetralix (Figure 1(g)–1(k)) can withstand some waterlogging under experimental conditions (Bannister, 1964). Hammond et al. (1990) investigated the position of *Erica*

tetralix in relation to local raised peat bog water tables and found that it occurred under minimum and maximum water table depths of 5–52 cm below the surface, with 90% of its occurrences recorded at sites with a 35 cm water table.

Empetrum nigrum

Empetrum nigrum (Figure 1(l)–1(o)) is intolerant of prolonged waterlogging and is restricted to well-drained hummock microforms on raised peat bogs (Bell and Tallis, 1973) and can occur abundantly in mire margins (Økland, 1990).

Andromeda polifolia

The highest shoot frequencies of *Andromeda polifolia* (Figure 2(a)–2(d)) occur in raised peat bog low hummock and lawn microforms (Lindholm and Vasander, 1981). It can extend into wet hollow microforms, but does not occur in the wettest locations (Jacquemart, 1998).

Vaccinium oxycoccus (*Oxycoccus palustris*)

Growth of *Vaccinium oxycoccus* (Figure 2(e)–2(g)) is optimal where average groundwater levels are 25–30 cm below the root–stem junction (Gronskis and Snickovskis, 1989). It occurs predominantly in moist hollow microforms on raised peat bogs (Lindholm and Vasander, 1981).

Drosera

Drosera spp. (Figure 2(i) and 2(j)) are insectivorous, short-lived herbaceous perennials of open, oligotrophic bogs. The seeds of *Drosera rotundifolia* and *Drosera anglica* are similar but strongly differ in morphology from *Drosera intermedia* seeds.

Eriophorum vaginatum

Eriophorum vaginatum (Figure 2(k)–2(m)) can grow over a large range of moisture conditions, and forms a dominant component of peat communities that experience surface water levels in the spring and subsequent reduced water levels in the summer (Wein, 1973). Drought conditions are not unduly harmful, as the species can survive due to its deep-rooting habit, since the rhizomes may extend to 60 cm below the surface (Gore and Urquhart, 1966). *Eriophorum vaginatum* is capable of invading pools, and paleoecological evidence supporting this observation, has been presented by Barber (1981). Hammond et al. (1990) most frequently encountered this species in peat with a water table at 24 cm depth, although it had a range spanning 0–28 cm, which demonstrates its wide tolerance to water table depth. Økland (1989, 1990) also concluded that it was 'ubiquitous and omnipresent,' although it was less common in carpets and low lawn microforms.

Eriophorum angustifolium

Godwin and Conway (1939) describe *Eriophorum angustifolium* (Figure 3(a) and 3(b)) as a species which characteristically colonizes drying pools. In a similar manner to *Eriophorum vaginatum*, the well-protected stem apex of *Eriophorum*



Figure 1 (a)–(f) *Calluna vulgaris*: (a) remains *in situ*; (b) stem with flower; (c) leaf epidermis; (d) twig without leaves; (e) stem; (f) seed. (g)–(k) *Erica tetralix*: (g) stems and leaves *in situ*; (h) leaf; (i) leaf epidermis; (j) and (k) seed. (l)–(o) *Empetrum nigrum*: (l) and (m) leaves; (n) leaf epidermis; (o) seed.

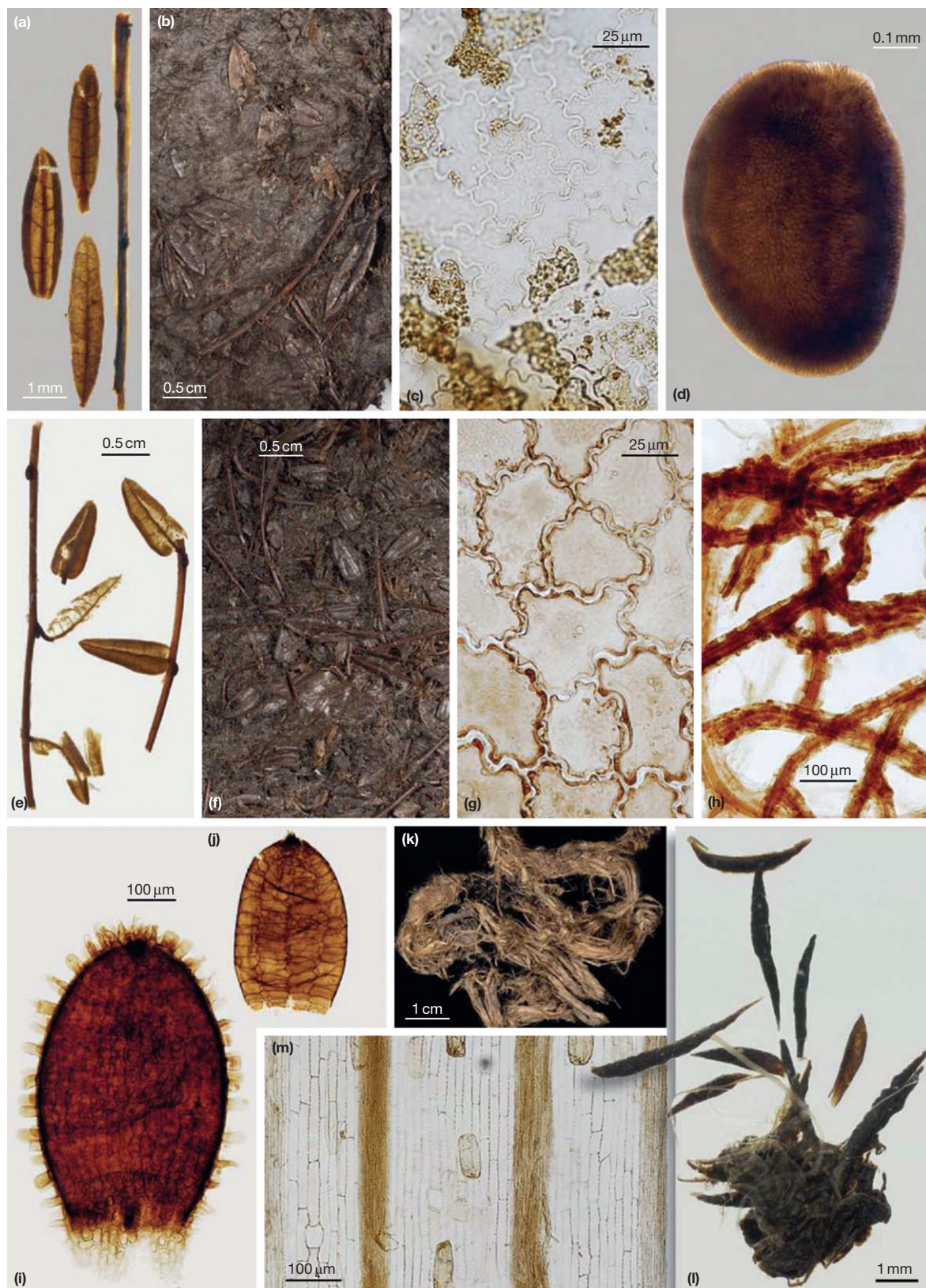


Figure 2 (a)–(d) *Andromeda polifolia*: (a) leaves and twig; (b) remains *in situ*; (c) leaf epidermis; (d) seed. (e)–(g) *Vaccinium oxycoccus*: (e) stems with leaves; (f) remains *in situ*; (g) leaf epidermis. (h) Ericales, rootlets. (i) *Drosera intermedia*, seed; (j) *Drosera rotundifolia* or *D. anglica*, seed. (k)–(m) *Eriophorum vaginatum*: (k) stem and leaf remains; (l) part of stem with some *in situ* spindles and separate spindles; (m) epidermis.

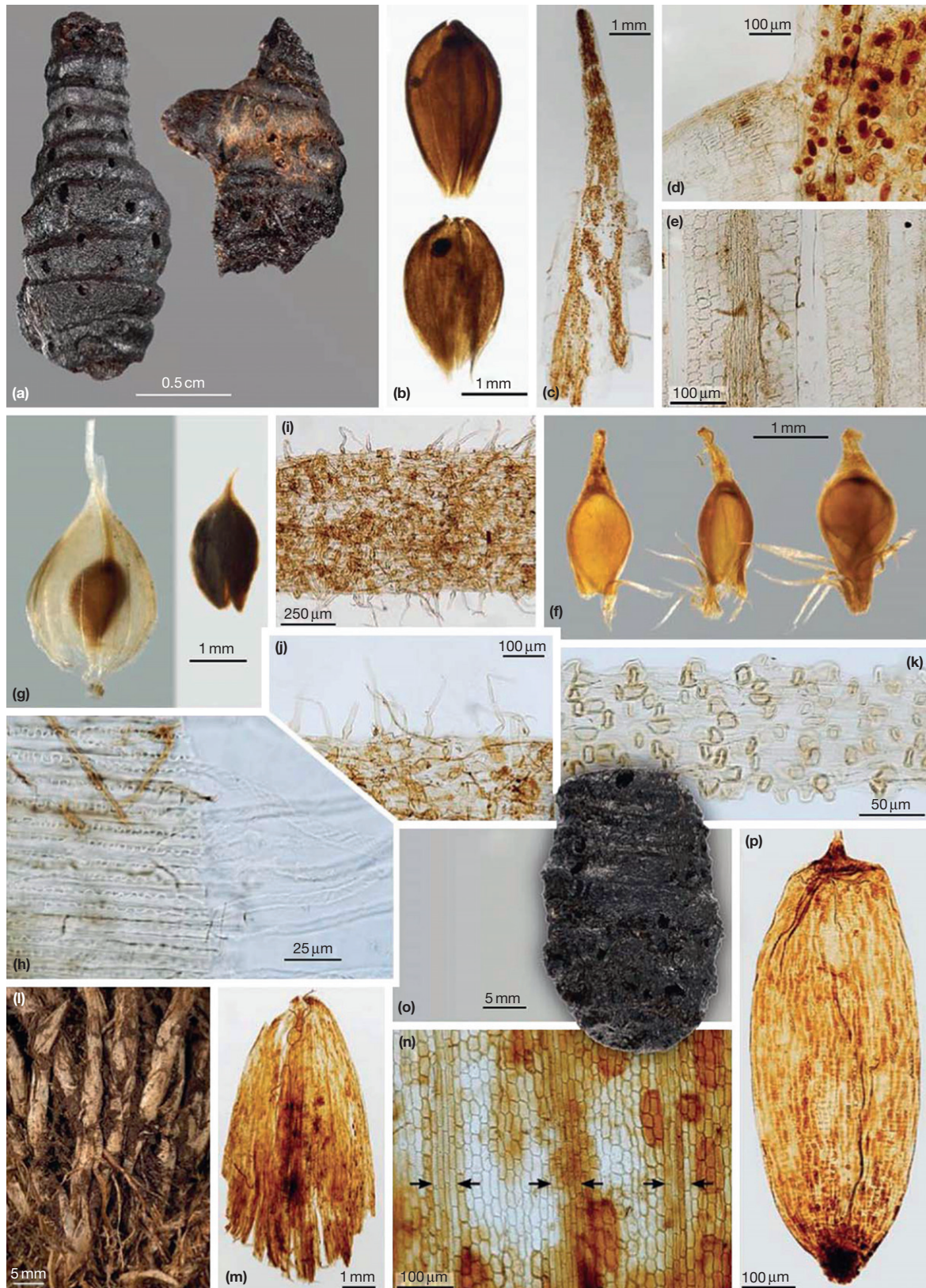


Figure 3 (a) *Eriophorum angustifolium*, lower parts of stems; (b) *Eriophorum* spp., seeds. (c)–(f) *Rhynchospora alba*: (c) leaf; (d) detail of leaf with sheath; (e) leaf epidermis; (f) seeds; (g) *Carex rostrata*, perigynium and achenes. (h)–(j) *Carex* cf. *limosa*: (h) rhizome fragment; (i) and (j) root. (k) *Carex* spp., root. (l)–(n) *Scirpus cespitosus*: (l) remains *in situ*; (m) leaf; (n) leaf epidermis. (o) *Cladium mariscus*, stem base. (p) *Typha* spp., seed.

angustifolium offers it a degree of fire resistance (Phillips, 1954). Boatman and Tomlinson (1977) and Rodwell (1991) confirmed its position at the edges of pools (where it can dominate), in permanent shallow pools, and in dried-up hollow bottoms, as it can tolerate a range of moisture conditions. Van der Molen (1988) noted its occurrence in hollow microforms, with a mean height above the water table of 8–10 cm.

Rhynchospora alba

R. alba (Figure 3(c)–3(f)) is a good indicator of high-mire water tables as it occurs in low lawns and at pool margins (Godwin and Conway, 1939). Its mean height above the water table appears to be 12–14 cm (van der Molen, 1988). Hammond et al. (1990) and Økland (1990) report slightly lower values of 8 and 5–7 cm, respectively, to the water table, with the range extending from 0 to 10 cm. Rodwell (1991) describes the position of *R. alba* as being confined between low lawns and the shallower water around pool margins, from where it may, sometimes, extend across smaller mire pools.

Carex spp. (*Carex rostrata* and *Carex limosa*)

These Cyperaceae (Figure 3(g)–3(k)) occur in mesotrophic and poor fens, although *Carex limosa* can also be found in ombrotrophic bog pool communities. *Carex* spp. macrofossils have been used to identify the fen/bog transition in UK peatlands (Hughes and Barber, 2004).

Scirpus cespitosus (*Trichophorum cespitosum*)

Scirpus cespitosus (Figure 3(l)–3(n)) is typically found on drier areas of raised and blanket peat bogs (Rodwell et al., 1991). On the better drained Rand (sloping margins) of raised peat bogs it is frequently recorded (Godwin, 1975).

Cladium mariscus

Cladium mariscus (Figure 3(o)) grows in calcium rich and nutrient poor, shallow water bodies, for example, in fens and swamps. Macrofossils of this sedge have been recorded with *Phragmites australis* to indicate a reedswamp phase (Hughes and Barber, 2003).

Typha angustifolia and *Typha latifolia*

Typha spp. (Figure 3(p)) produce enormous quantities of wind-dispersed seeds and the presence of low numbers of seeds does not necessarily indicate its local presence. In combination with the pollen record its local presence in terrestrialization phases becomes clear (rooting in nutrient-rich substrates).

Scheuchzeria palustris

Scheuchzeria palustris (Figure 4(a)–4(f)) grows in permanent standing water or in places where local water tables are high (Moore, 1955; Tallis and Birks, 1965). Its occurrence in very wet and slightly minerotrophic peatland microforms has also been confirmed by Økland (1990) using ordination

techniques. Casparie (1972) also mentions its presence can indicate slight minerotrophy in subfossil peat samples. Thick layers of *Scheuchzeria palustris* 'precursor peat' can precede the deposition of later *Sphagnum*-rich levels in peat bog deposits.

Menyanthes trifoliata

Menyanthes trifoliata (Figure 4(g)–4(i)) is an aquatic perennial herb that occurs in a range of aquatic mesotrophic/eutrophic/oligotrophic peatland communities, where water levels are at or above the ground surface (Godwin, 1975; Hewett, 1964).

Potentilla palustris

Potentilla palustris (Figure 4(j)) grows in fens, wet meadows, and lake margins.

Myrica gale

Myrica gale (Figure 4(k)) is a deciduous shrub that chiefly grows on both acidic fens and bogs (marginal slopes of raised bogs) over a pH range of 3.8–6.1 (Skene et al., 2000).

Juncus

Various *Juncus* spp. (Figure 4(l)–4(n)) occur in wet meadows, along shores and in gradient and pioneer situations. The plants produce enormous amounts of seeds. Körber-Grohne (1964) made a useful illustrated key for *Juncus* seeds in northwest-European deposits.

Molinia caerulea

Molinia caerulea (Figure 5(a) and 5(b)) is a perennial grass that has a bimodal pH distribution (pH < 4 and pH > 7, Grime et al., 1988) and can grow in both bog or fen sites where total waterlogging is absent (Godwin, 1975). It is most abundant in sites where there is ground-water movement, good soil aeration during the growing season, and an enriched nutrient supply. It is tolerant of burning (Taylor et al., 2001).

Phragmites australis

Phragmites australis (Figure 5(c)–5(f)) grows in fens, shallow lakes, salt marshes, and flushed areas of bog edges, in low-lying areas intermittently or permanently flooded with shallow or still water (pH range for well-grown stands of 5.5–7.5, but can occur where pH is as low as 3.6, Haslam, 1972). The roots can extend to 100 cm in depth.

Tree Macrofossils – *Alnus glutinosa*, *Betula* spp., *Pinus sylvestris*, *Populus tremula*

Alnus, *Betula*, and *Pinus* (Figures 5(g)–5(n) and 6(a)–6(d)) grow in 'bog woodland,' and base-rich mires ('fen carr'). Buds, bud scales, catkin scales, fruits, leaf fragments, and needles of these trees have been recorded in peat profiles (Hannon et al., 2000; Hughes and Barber, 2004), particularly in deposits spanning the fen/bog transition (Hughes et al., 2000). Remains of *Populus tremula* (Figure 5(o) and 5(p)) often occur in early

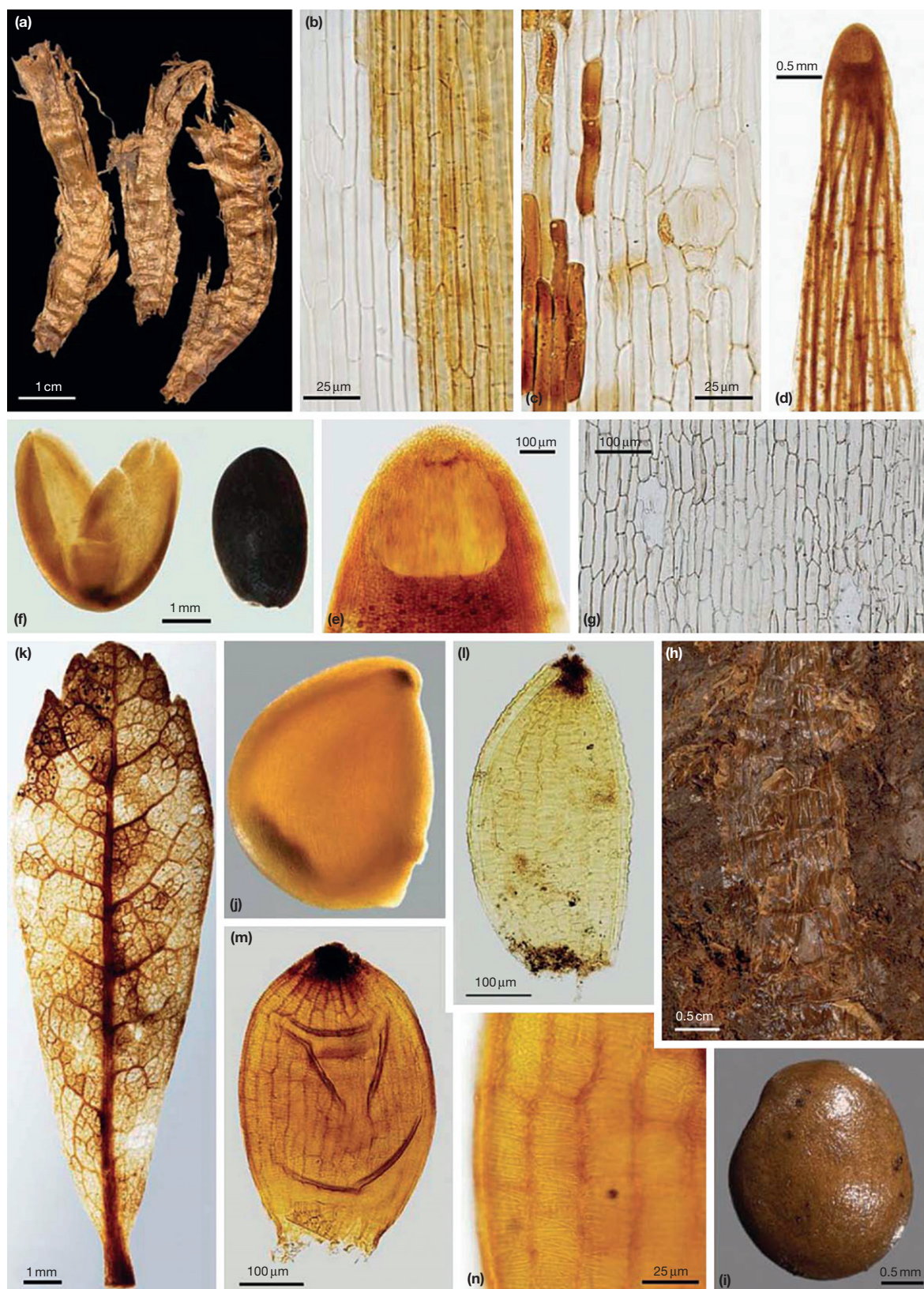


Figure 4 (a)–(f) *Scheuchzeria palustris*: (a) rhizomes and leaf remains; (b) rhizome epidermis; (c) leaf epidermis with stoma; (d) and (e) hydathode; (f) seeds. (g)–(i) *Menyanthes trifoliata*: (g) epidermis; (h) remains *in situ*; (i) seed. (j) *Potentilla palustris*, seed. (k) *Myrica gale*, leaf. (l) *Juncus effusus*-type, seed; (m) and (n) *Juncus articulatus*-type, seed.



Figure 5 (a) and (b) *Molinia caerulea*: (a) thickened lower stem bases; (b) epidermis. (c)–(f) *Phragmites australis*: (c) rhizomes; (d) and (e) rhizome epidermis; (f) leaf epidermis with stoma. (g) and (h) *Alnus* spp.: (g) seeds; (h) remains of female catkin. (i)–(n) *Betula* spp.: (i) seed; (j) *Betula nana*, female catkin scale; (k) tree birch, female catkin scale; (l) detail of leaf; (m) *B. nana*, leaf; (n) *Betula* spp., periderm. (o) and (p) *Populus tremula*: (o) catkin bracts; (p) bud scale.



Figure 6 (a)–(d) *Pinus sylvestris*: (a) needles; (b) needle epidermis with stomata; (c) periderm; (d) mycorrhizal roots. (e) and (f) *Potamogeton*, leaf fragment. (g)–(i) *Ceratophyllum*: (g) leaf with spines *in situ*; (h) spine; (i) seed. (j)–(n) *Nymphaeaceae*: (j) leaf epidermis; (k) leaf fragment with trichosclereids; (l) trichosclereid; (m) *Nymphaea*, seed; (n) *Nymphaea*, seed epidermis.

Holocene deposits, for example, in small depressions, indicating stands of *Populus* on drier soils (van Geel et al., 1981).

Potamogeton

Seeds of *Potamogeton* can be identified with the study by Aalto (1970), but also leaf remains with their characteristic venation (parallel, with connections (Figure 6(e) and 6(f))) can be common in lake deposits.

Ceratophyllum

Ceratophyllum spp. (Figure 6(g)–6(i)) are underwater rootless perennial plants which occur in nutrient-rich water. Pollen of *Ceratophyllum* does not preserve, and seeds are quite rare. However, the characteristic leaf spines are common in pollen slides and leaf fragments with *in situ* leaf spines can be found in macrofossil samples of lake deposits.

Nymphaeaceae

Asterosclereids and the suberized basal cells of Nymphaeaceae mucilaginous hairs (Figure 6(j)–6(n)) are frequently found in pollen slides from lake and pool deposits and indicate their local presence (Pals et al., 1980).

Trapa natans

Water chestnut (*Trapa natans*; Figure 7(a)) is an annual aquatic plant that grows at a depth of 1–2 m in lakes and former river channels with a muddy bottom and fairly warm water. The fruit has a very hard, rough shell with four sharp spines (Figure 7(a) shows a spine). Fossil fruits have been identified by Korhola and Tikkanen (1997), most abundantly in shallow, eutrophic water, prior to final infilling of lake basins (Schofield and Bunting, 2005).

Equisetum

The above-ground remains of *Equisetum fluvatile* (Figure 7(b)–7(f)) in peat deposits are black, while rhizomes and roots are dark-brown. Rhizome epidermis cells are long and show a very regular pattern with zip-connections between the cells. Dark parts with lighter spots in the epidermis are a characteristic feature. The outer cell layer of the roots show long and short cells, with simple cell structure (no zip-connections).

Thelypteris palustris

Thelypteris palustris (Figure 7(g)–7(k)) can be an important peat former in Holocene deposits formed under eutrophic conditions. Often leaf fragments, curved ends of leaf axes, roots, and tracheids are preserved (Grosse-Brauckmann, 1972a). Fern sporangia are common in *Thelypteris* peat.

Characeae

Oospores of Characeae (Figure 7(l)) are often numerous in lake deposits, especially when these deposits were formed in pioneer conditions (inputs of minerogenic material). A key for

central European charophyte oospores and a discussion of their ecology is available (Haas, 1994).

Gloeotrichia

Globular colonies consisting of hyaline sheaths of *Gloeotrichia*-type (Figure 7(m) and 7(n)) regularly occur in shallow freshwater deposits (van Geel et al., 1989, 1994, 1996). Increases of Cyanobacteria can indicate eutrophication (increased phosphorus loading which may cause nitrogen-limited growth conditions). In such conditions Cyanobacteria capable of nitrogen fixation can thrive. *Gloeotrichia*-type was also observed in nutrient-poor pioneer conditions at the start of the Late-Glacial period (van Geel et al., 1989).

Sphagnum austinii (*Sphagnum imbricatum* spp. *austinii*)

This *Sphagnum* spp. (Figure 8(a) and 8(b)) is particularly striking since its hyaline cells possess numerous transverse lamellae, which appear to be comb-like (Smith, 2004). The ecology of modern examples of this species indicates that it is a bryophyte capable of growing under relatively xeric conditions. Like other members of section *Sphagnum*, it has both inrolled cucullate (hooded at the apex) leaves, and grows in a compact form with a very close and imbricate arrangement of the branch leaves. This serves to reduce the surface area for evaporation (Flatberg, 1986). Green (1968) experimentally demonstrated its ability to retain water, while Flatberg (1986) confirmed its presence in (high) hummocks. Detailed paleoecological macrofossil analysis undertaken by van der Molen and Hoekstra (1988) highlight the ability of this bryophyte to grow under dry mire surface conditions in the past.

Sphagnum imbricatum spp. *austinii* does, however, have morphologically distinct ecads or ecophenes, which allow it to grow in both hummocks and (low) lawns. The lawn ecad has a lax growth form, which has been experimentally induced by Green (1968). The ability of *Sphagnum imbricatum* to grow under wetter mire surface conditions has been demonstrated in the fossil record by Casparie (1972), van Geel (1978), and Barber (1981). The disappearance of *Sphagnum imbricatum* from the stratigraphy of British and Irish peatlands has been noted by many paleoecologists (see the review in Mauquoy and Barber, 1999). The reasons for the disappearance of *Sphagnum imbricatum* in northwest European peat deposits are still not clear. It may result from a possible combination of climatic changes and eutrophication (deposition of nutrients bound to soil dust following human disturbance).

Sphagnum magellanicum

Subfossil leaves of *Sphagnum magellanicum* (Figure 8(c)–8(f)) are readily recognized by their photosynthetic cells, which are oval and largely enclosed by hyaline cells (Smith, 2004). Andrus et al. (1983) recorded *Sphagnum magellanicum* in lawns and low hummocks (25.8 ± 9.5 cm above the spring water table). Hammond (1990) found that 90% of the occurrences of *Sphagnum magellanicum* occurred at a water table depth of 24 cm. Økland (1990), however, suggests it has a broad habitat niche. The species became dominant in many raised bogs during the last millennium.



Figure 7 (a) *Trapa natans*, barbed appendage of seed. (b)–(f) *Equisetum fluviatile*: (b) rhizome epidermis; (c) leaf; (d) and (e) stem diaphragm; (f) root. (g)–(k) *Thelypteris palustris*: (g) leaf fragment; (h) young fossilized stems with folded leaves; (i) fern sporangium (*Thelypteris*-type); (j) tracheids; (k) root. (l) Characeae, oospore. (m) and (n) *Gloeotrichia*-type, colony. (o) *Sphagnum* peat.

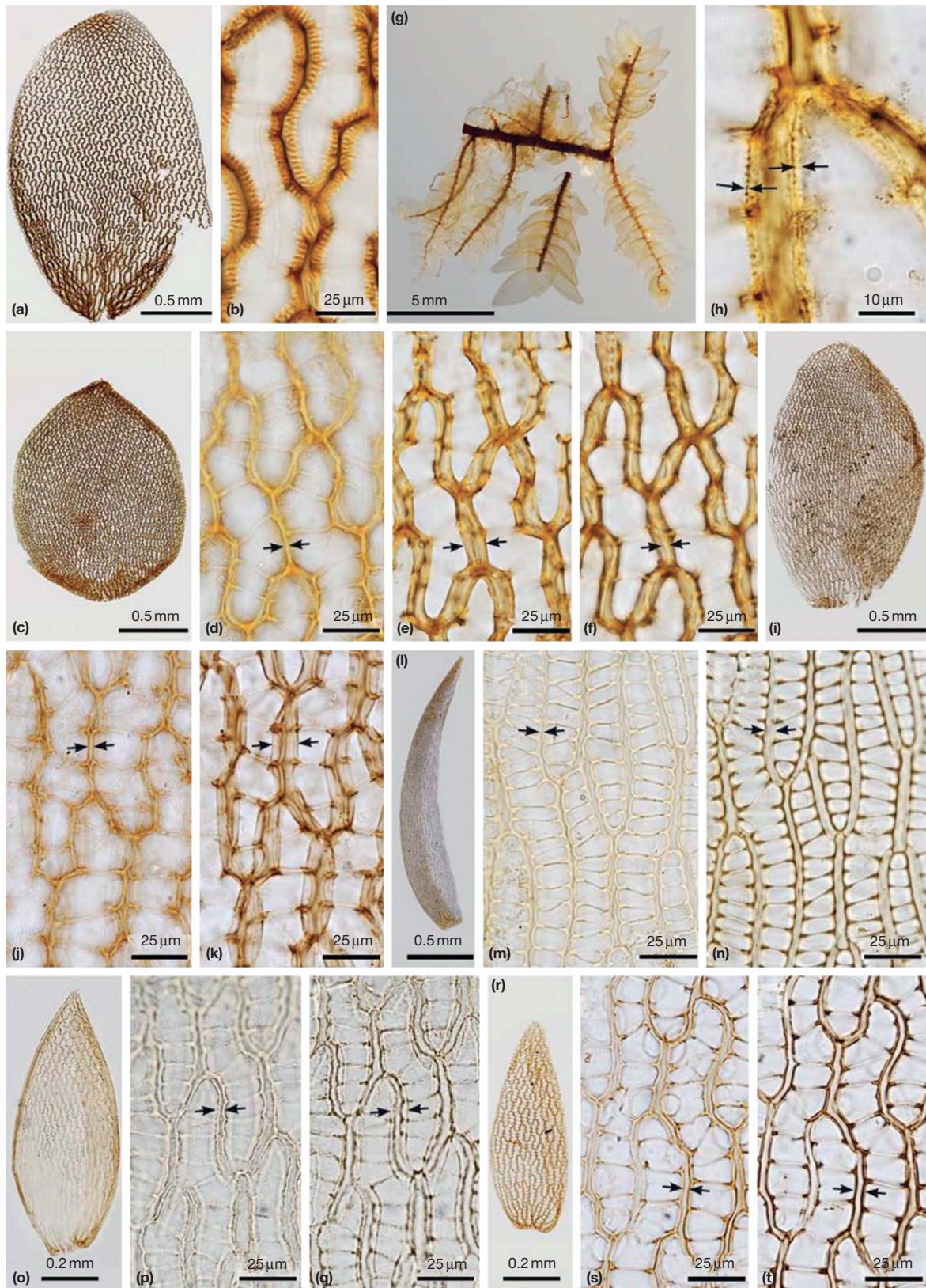


Figure 8 (a) and (b) *Sphagnum austinii* (*S. imbricatum* spp. *austinii*): (a) leaf; (b) transverse lamellae in hyaline cells. (c)–(f) *Sphagnum magellanicum*: (c) leaf; (d), (e), and (f) high, middle, and low focus. (g) and (h) *Sphagnum papillosum*: (h) papillae on cell walls. (i)–(k) *Sphagnum palustre*: (i) leaf; (j) and (k) dorsal high and low focus. (l)–(n) *Sphagnum* sect. *Cuspidata*: (l) leaf, (m) and (n) ventral high and low focus. (o)–(q) *Sphagnum* cf. *tenellum*: (o) leaf; (p) and (q) ventral high and low focus. (r)–(t): *Sphagnum* sect. *Acutifolia*: (r) leaf; (s) and (t) dorsal high and low focus.

Sphagnum papillosum

The internal commissural walls of *Sphagnum papillosum* (Figure 8(g) and 8(h)) are rough since they possess projecting papillae (Smith, 2004). This distinctive feature of the hyaline cells makes identification of *Sphagnum papillosum* straightforward. Boatman (1983) found it predominantly in low lawns and pools in the Silver Flowe complex of bogs in southwest Scotland. Its 10–12 cm mean height above the water table also suggests an optimal low lawn position (Janssens, 1992; van der Molen et al., 1994). Conversely, Hammond et al. (1990) found that 90% of the occurrences of the species were at relatively high positions on lawns (23 cm), just below the optimum value for *Sphagnum magellanicum* (24 cm). Økland (1990) made similar findings since it was recorded largely in lawn microforms, and only rarely in hummocks as single shoots. This species appears to have a wide ecotop breadth, like *Sphagnum magellanicum*, although its optimum position would appear to be centered on low lawn positions.

Sphagnum palustre

The branch leaves of *Sphagnum palustre* (Figure 8(i)–8(k)) are ovate or broadly ovate and the apex has a hood-shaped tip. The photosynthetic cells are variable in shape and can be either triangular and fully exposed on the ventral surface, barrel-shaped and broadly exposed ventrally, or lens-shaped with negligible ventral exposure (Smith, 2004). This species can be found in mesotrophic marshes, stream sides, and in wet woodlands. Macrofossils of *Sphagnum palustre* in poor fen peat deposits have been identified by Hughes et al. (2000).

Sphagnum Section *Cuspidata*

Species within section *Cuspidata* (Figure 8(l)–8(n)) can be identified by their ovate-to-lanceolate shape, narrow hyaline cells, and their triangular-to-trapezoidal-shaped photosynthetic cells broadly exposed on the dorsal surface. The leaves of *Sphagnum cuspidatum* are particularly distinctive, as they may extend to 3.5 mm in length. Section *Cuspidata* leaves are predominantly found in pool and low lawn ecotopes, and are therefore a good indicator of raised mire water levels.

Sphagnum Section *Acutifolia*

Leaves from *Sphagnum* section *Acutifolia* (Figures 8(r)–8(t) and 9(a)–9(d)) are readily recognized by their ovate-to-narrow ovate branch leaves (Smith, 2004), which are small in *Sphagnum fuscum* (1.1–1.3 mm in length) and *Sphagnum capillifolium* (0.8–1.4 mm) and rather bigger in *Sphagnum subnitens* (1.2–2.7 mm) and *Sphagnum molle* (1.4–2.3 mm). All of the species within the section possess triangular photosynthetic cells broadly exposed on the ventral surface and possess similarly sized dorsal pores (up to 30 µm). The overlapping size ranges of the branch leaves, makes positive identification to species level difficult (unless stem leaves are present, as these are very valuable in aiding identification, see Johansson, 1995).

The difficulty of consistently and accurately identifying leaves to species level also poses problems in determining the degree of mire surface wetness. This is not such a problem for

either *Sphagnum fuscum* and *Sphagnum capillifolium* var. *rubellum* leaves, since they occupy hummocks; Andrus et al. (1983) found the two species growing at mean heights of 30.9 ± 10.6 and 24.8 ± 9.4 cm above the spring water table, respectively. Rydin and McDonald (1985) similarly found the two species at respective heights of 35 and 25 cm above the water table, while Janssens (1992) derived an optimum height of 27.5 cm above the mean water table for *Sphagnum fuscum*.

Sphagnum molle and *Sphagnum subnitens* are, however, lawn species (Smith, 2004), which reduces the accuracy of any paleoecological reconstruction made using branch leaves identified to *Sphagnum* section *Acutifolia* alone.

Sphagnum Section *Subsecunda*

Branch leaves in *Sphagnum* section *Subsecunda* (Figure 9(e) and 9(f)) are typically 0.7–2.5 mm long and 0.4–1.5 mm wide. Numerous small dorsal/ventral pores are present along the commissures (the junction between the hyaline and the green cells), with up to 40 and 20 per cell, respectively. Species in this section can be found in depressions (often submerged), in base-rich flushes, fens, woods, and moors.

Aulacomnium palustre

The leaf cells of *Aulacomnium palustre* (Figure 9(i) and 9(j)) have rounded lumens. This moss grows at relatively high positions above the water table, for example, Andrus et al. (1983) recorded it at positions 30.0–31.4 cm above the spring water table, in small patches on hummock sides. Økland (1990) similarly confirms the presence of this species within upper hummocks and, additionally, as ‘single shoots’ among *Sphagnum* species. Nicholson and Gignac (1995) found it growing at a greater range of heights above the water table (10–40 cm).

Polytrichum commune and *Polytrichum juniperinum*/ *alpestre*-Type

Polytrichum commune (Figure 10(a)–10(c)) grows under a wide range of acidic habitats, including heaths, woods, wet moorlands, bogs, lake margins, pools, and stream sides in damp-to-wet microhabitats. It is commonly associated with a range of *Sphagnum* spp., especially *Sphagnum palustre* and *Sphagnum recurvum*. *Polytrichum juniperinum* (Figure 9(k)–9(o)) grows on well-drained acidic soils on heaths and moors (Dickson, 1973; Smith, 2004). *Polytrichum alpestre* similarly displays a preference for drier microforms and has been interpreted as a hummock-level moss (Hughes et al., 2000), since it forms dense tussocks on ombrotrophic bogs (Dickson, 1973).

Paludella squarrosa

Paludella squarrosa (Figure 10(d)–10(f)) is a moss that occurs in fens and subfossil leaves identified in a peat profile collected from the east-European Russian Arctic have been used to show a wet and mesotrophic phase in the local peatland succession (Välranta et al., 2003).

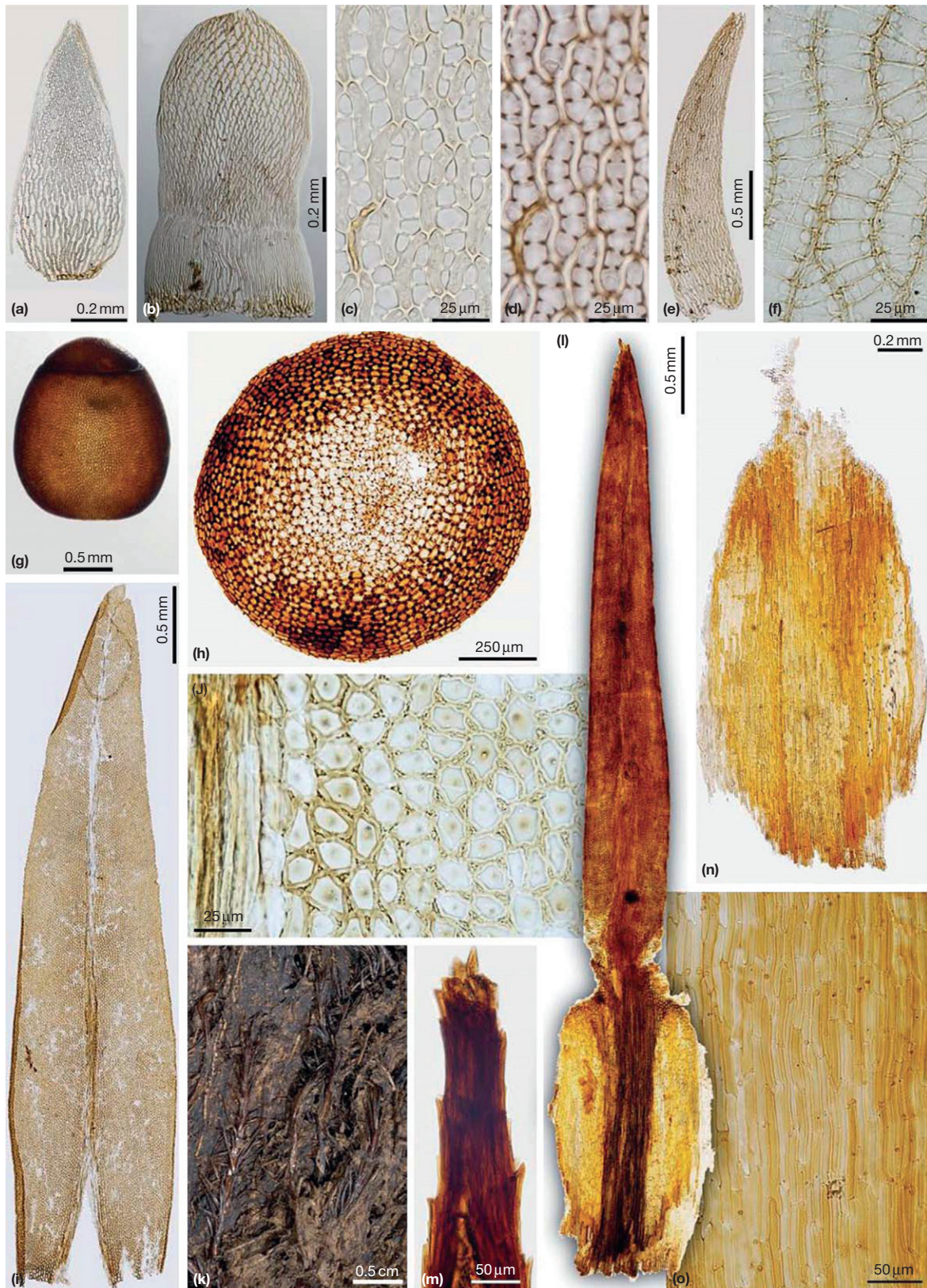


Figure 9 (a)–(d) *Sphagnum fuscum* (recent): (a) and (b) leaf and stem leaf; (c) and (d) ventral high and low focus. (e) and (f) *Sphagnum* sect. *Subsecunda*. (g) and (h) *Sphagnum* spp., sporangium and operculum. (i) and (j) *Aulacomnium palustre*, leaf and detail of leaf. (k)–(o) *Polytrichum juniperinum/alpestre*-type: (k) *Polytrichum* in situ; (l) leaf; (m) leaf tip; (n) leaf base; (o) detail of leaf base.

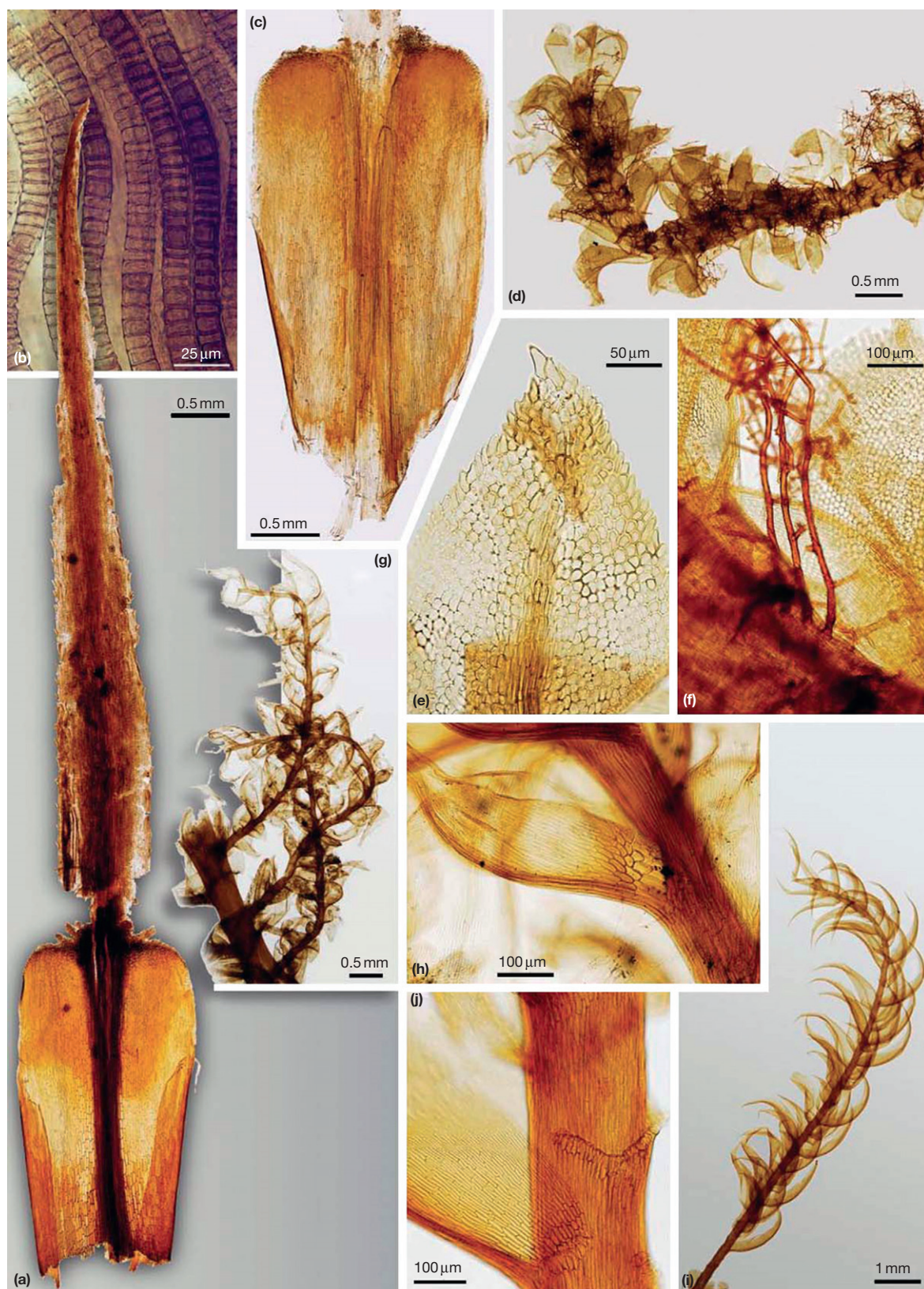


Figure 10 (a)–(c) *Polytrichum commune*: (a) leaf; (b) detail of leaf with lamellae; (c) leaf base. (d)–(f) *Paludella squarrosa*; (g) and (h) *Calliergon giganteum*; (i) and (j) *Drepanocladus* cf. *aduncus*.

Calliergon giganteum

Calliergon giganteum (Figure 10(g) and 10(h)) is a moss that primarily occurs in rich fens. Subfossil leaves of *Calliergon giganteum* were identified by van Geel et al. (1986) and used to indicate stagnant, eutrophic, and base-rich aquatic environments in an Eemian lake deposit.

Drepanocladus spp. (Figure 10(i) and 10(j))

Drepanocladus aduncus is an aquatic moss which occurs in moderate-rich fens, small lakes, ponds, or streams. *Drepanocladus fluitans* is found in pools in blanket bogs, sometimes with *Sphagnum cuspidatum*. Watson (1968) describes it as a species that can tolerate both aquatic and terrestrial habitats, since it is found both in deep pools and in sites which are only periodically wet. Dickson (1973), however, places *Drepanocladus fluitans* as a species primarily found in base-poor pools. Ohlson and Malmer (1990) corroborate this, as they recorded *Drepanocladus fluitans* with *R. alba* growing in mud-bottom communities.

Fungi

Fungal fruit bodies (~125–500 µm in diameter) are regularly recorded in raised bog macrofossil samples (van Geel, 1978). The dark-brown fruit bodies are often still filled with ascospores. *Meliola ellisii* (Figure 11(b)) is an obligate parasite on *Calluna vulgaris*. The currently unidentified Type 18 (Figure 11(a)) occurs on roots of *Eriophorum vaginatum*. Mycelium with chlamydospores of *Glomus* cf. *fasciculatum* (Figure 11(c)) can be common in various types of (nonoligotrophic) peat deposits. It grows as a vesicular-arbuscular endomycorrhizal fungus on a variety of host plants. In lake sediments the presence of *Glomus* indicates soil erosion in the catchment of the lake (Anderson et al., 1984).

Freshwater Sponges

Spicules and gemmules of freshwater sponges (Figure 11(e) and 11(f)) are of regular occurrence in macrofossil samples, especially in shallow water deposits at transitions from open water in the first stage of terrestrialization (Harrison, 1988; van Geel et al., 1989).

Chironomidae and Trichoptera

The chitinous larval remains of chironomids (Figure 11(g) and 11(h)) are very useful in paleoenvironmental studies of lake sediments (Walker, 2001). For the use of caddis fly larvae remains we refer to Elias (2001).

Acari (Oribatid Mites)

The majority of oribatid mites (Figure 11(i)) have a sclerotized, chitin-rich cuticle. They are soil dwellers, common in peat deposits, but normally no records are made and for identifications specialist literature and the help of an acarologist is essential (Schelvis and van Geel, 1989). For an overview of the potential use of mites in paleoecology see Solhøy (2001).

Piscicola geometra

Piscicola geometra (Figure 11(j)) is an external parasite of freshwater fish. They lay their eggs in cocoons on the pond bottom, plants, stones, etc. Cocoons can be common in lake deposits (Pals et al., 1980).

Bryozoa

Bryozoa (Figure 11(k)–11(m)) are small, sessile, colonial, and filter-feeding animals. There are more than 65 freshwater species worldwide. Bryozoa of the class Phylactolaemata produce asexual buds called statoblasts. These consist of two chitinous valves with regenerative cells within, and function as reproductive, survival, and dispersal agents. Statoblasts can be identified and provide information on past aquatic environments (Francis, 2001).

Foraminiferae

In peat deposits and lake sediments, the presence of linings of foraminiferae (Figure 11(n)) indicate temporary inundation with salt or brackish water.

Ostracods

Ostracods (Figure 11(o)) are bivalved aquatic crustaceans with shells, common in nonmarine waters with neutral-to-alkaline pH. Ostracods are sensitive to a range of ecological factors (Holmes, 2001).

Callidina angusticollis

The loricas of *Callidina angusticollis* (Figure 11(p)) are of regular occurrence in Holocene raised bog deposits, but to date no specific ecological information can be derived from the occurrence of this representative of the Rotifera (van Geel, 1978).

Diaptomus cf. *castor*

According to Bennike (1998) the egg sacs of *Diaptomus* cf. *castor* (Figure 11(q)) might indicate low arctic or subarctic climates, but that idea must be incorrect, as van Dam et al. (1988) recorded them from late Holocene moorland pool deposits in The Netherlands.

Conclusions

High-quality macrofossil datasets from peat deposits can be used to reconstruct the former degree of mire surface wetness, successional processes of the local peat-forming vegetation, trophic/pH status, and assess the effect of human disturbance (e.g., burning, drainage, and rapid forest expansion; Pellerin and Lavoie, 2003). A thorough knowledge of the present-day ecology of the plants growing in bogs and fens is essential in order to make accurate paleoenvironmental reconstructions, and provides the basis for evaluating the subfossil presence of these plants (the individualistic approach, Birks and Birks, 2003). We have, therefore, compiled and reviewed a selection

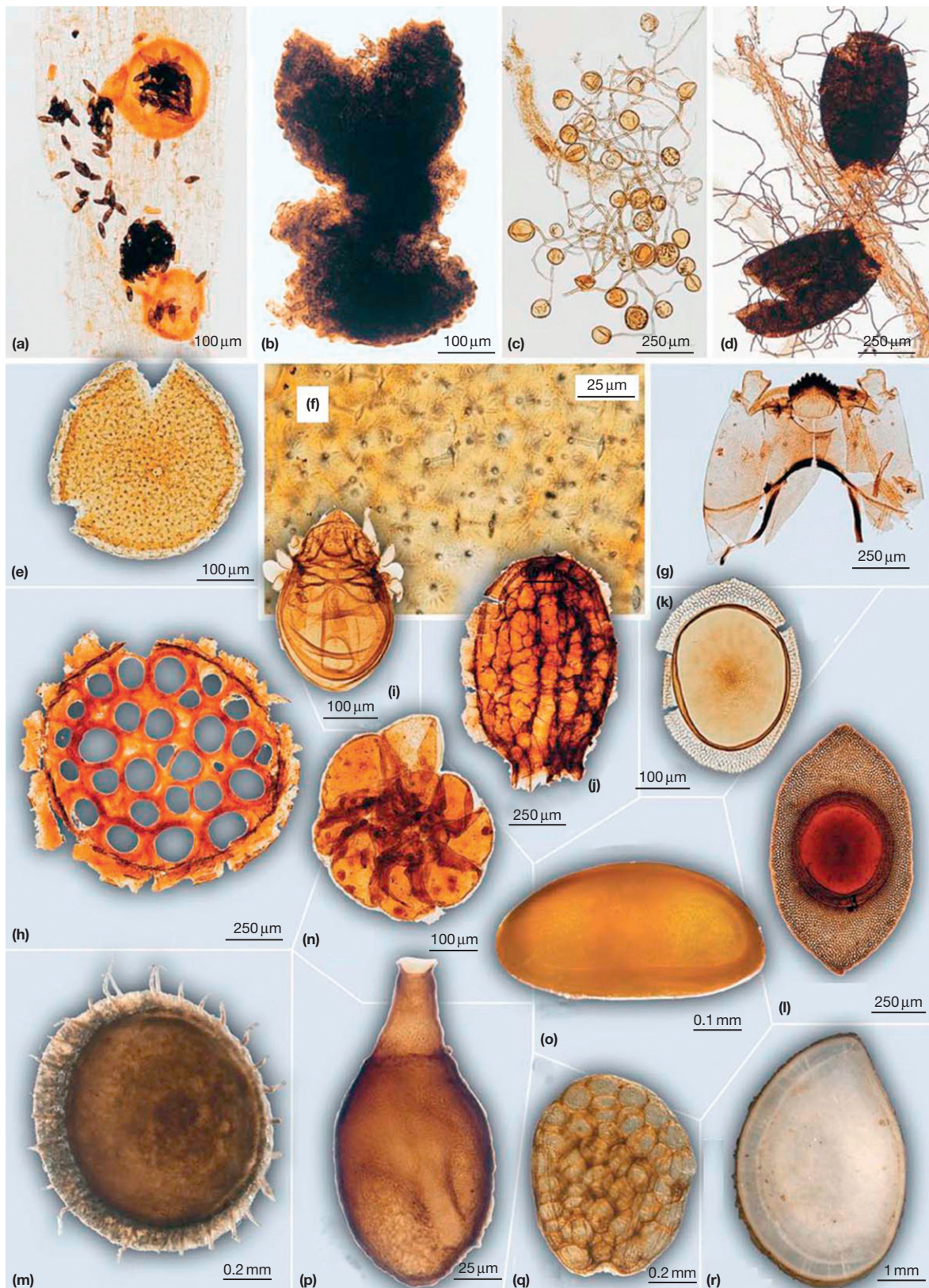


Figure 11 (a) Fungal fruit body (Type 18; not identified) in roots of cf. *Eriophorum vaginatum*. (b) *Meliola ellisii*, fruit body with ascospores. (c) *Glomus*, mycelium with chlamydospores. (d) mycorrhiza tissue (*Cenococcum geophilum*) on root tips (not preserved). (e) and (f) Freshwater sponge, gemmule. (g) Chironomid, larval head capsule. (h) Trichoptera, membrane of pupal case. (i) Acari, Oribatida. (j) *Piscicola geometra*, cocoon. (k)–(m) Bryozoa statoblasts: (k) *Plumatella*-type; (l) *Lophopus crystallinus*; (m) *Cristatella mucedo*. (n) Foraminifera. (o) Ostracod. (p) *Callidina angusticollis*, lorica. (q) *Diaptomus castor*, egg sac. (r) *Bithynia*, operculum.

of literature on the actuo- and paleoecology of peat-forming plants to guide the interpretation of subfossil peat sequences. Identifications of mire and peat macrofossils can often be made to species level, and is the key advantage of the technique compared to pollen analysis, since grains and spores are usually only identifiable to family or genus level and may come from a variety of habitats outside the peat-forming wetland. Our illustrations of a range of mire and peat macrofossils may help to identify remains of peat-forming plants, but we stress that type material (preferably fossil) is the best way to confirm correct identifications.

See also: Plant Macrofossil Introduction. **Plant Macrofossil Methods and Studies:** Validation of Pollen Studies. **Pollen Methods and Studies:** Reconstructing Past Biodiversity Development.

References

- Aalto M (1970) Potamogetonaceae fruits. 1. Recent and subfossil endocarps of Fennoscandian species. *Acta Botanica Fennica* 88: 1–55.
- Aerts R, Wallén B, and Malmer N (1992) Growth limiting nutrients in *Sphagnum* dominated bogs subject to low and high atmospheric nitrogen supply. *Journal of Ecology* 80: 131–140.
- Anderberg AL (1994) *Atlas of Seeds and Small Fruits of Northwest European Plant Species with Morphological Descriptions*. Uddevalla: Risbergs Tryckeri. Part 4 Resedaceae-Umbelliferae. Swedish Museum of Natural History, Stockholm.
- Anderson RS, Homola RL, Davis RB, and Jacobson GL (1984) Fossil remains of the mycorrhizal fungal *Glomus fasciculatum* complex in postglacial lake sediments from Maine. *Canadian Journal of Botany* 62: 2325–2328.
- Andrus RE, Wagner DJ, and Titus JE (1983) Vertical zonation of *Sphagnum* mosses along hummock–hollow gradients. *Canadian Journal of Botany* 61: 3128–3139.
- Barber KE (1981) *Peat Stratigraphy and Climatic Change: A Palaeoecological Test of the Theory of Cyclic Peat Bog Regeneration*. Rotterdam: Balkema.
- Barber KE and Charman DJ (2003) Holocene palaeoclimate records from peatlands. In: Mackay A, Battarbee R, Birks J, and Oldfield F (eds.) *Global Change in the Holocene*, pp. 210–226. London: Arnold.
- Barber KE, Dumayne-Peaty L, Hughes PDM, Mauquoy D, and Scaife RG (1998) Replicability and variability of the recent macrofossil and proxy-climate record from raised bogs: Field stratigraphy and macrofossil data from Bolton Fell Moss and Walton Moss, Cumbria, England. *Journal of Quaternary Science* 13: 515–528.
- Barber KE, Zolitschka B, Tarasov P, and Lotter AF (2004) Atlantic to Urals – The Holocene climatic record of mid-latitude Europe. In: Battarbee RW, Gasse F, and Stickley CE (eds.) *Past Climate Variability Through Europe and Africa*, pp. 417–442. Dordrecht: Springer.
- Barry TA and Synnott DM (1987) Further studies into bryophyte occurrence and succession in the Hochmoor peat types of Ireland. *Glasra* 10: 1–21.
- Bell JNB and Tallis JH (1973) Biological flora of the British Isles. *Empetrum nigrum* L. *Journal of Ecology* 61: 289–305.
- Bennike O (1998) Fossil egg sacs of Diaptomus (Crustacea: Copepoda) in Late Quaternary lake sediments. *Journal of Paleolimnology* 19: 77–79.
- Berggren G (1964) *Atlas of Seeds. Part 2. Cyperaceae*. Stockholm: Swedish Natural Science Research Council.
- Birks HH (2001) Plant macrofossils. In: Smol JP, Birks HJB, and Last WM (eds.) *Tracking Environmental Change Using Lake Sediments*, pp. 49–74. Dordrecht: Kluwer Academic Publishers.
- Birks HH and Birks HJB (2003) Reconstructing Holocene climates from pollen and plant macrofossils. In: Mackay A, Battarbee RW, Birks HJB, and Oldfield F (eds.) *Global Change in The Holocene*, pp. 342–357. London: Arnold.
- Blaauw M, van Geel B, Heuvelink GBM, Mauquoy D, and van der Plicht J (2003) A numerical approach to ¹⁴C wiggle-match dating of organic deposits: Best fits and confidence intervals. *Quaternary Science Reviews* 22: 1485–1500.
- Boatman DJ (1983) The Silver Flowe National Nature Reserve, Galloway, Scotland. *Journal of Biogeography* 10: 163–274.
- Boatman DJ and Tomlinson RW (1977) The Silver Flowe: Some structural and hydrological features of Brishie bog and their bearing on pool formation. *Journal of Ecology* 65: 531–546.
- Bohncke SJP, Van Haaster H, and Wiegers J (1984) *Paludella squarrosa* (Hedw) bried in a late subboreal Holland peat sequence. *Journal of Bryology* 13: 219–226.
- Booth RK, Jackson ST, and Gray ED (2004) Paleocology and high-resolution paleohydrology of a kettle peatland in upper Michigan. *Quaternary Research* 61: 1–13.
- Bos JAA, Dickson JH, Coope GR, and Jardine WG (2004) Flora, fauna and climate of Scotland during the Weichselian Middle Pleniglacial – Palynological, macrofossil and coleopteran investigations. *Palaeogeography, Palaeoclimatology, Palaeoecology* 204: 65–100.
- Casparie WA (1972) Bog development in Southeastern Drenthe. *Vegetatio* 25: 1–271.
- Chambers FM, Mauquoy D, and Todd PA (1999) Recent rise to dominance of *Molinia caerulea* in environmentally sensitive areas: New perspectives from palaeoecological data. *Journal of Applied Ecology* 36: 719–733.
- Charman DJ, Hendon D, and Packman S (1999) Multiproxy surface wetness records from replicate cores on an ombrotrophic mire: Implications for Holocene palaeoclimate records. *Journal of Quaternary Science* 14: 451–463.
- Dickson JH (1973) Bryophytes of the Pleistocene. *The British Record and its Chronological and Ecological Implications*. Cambridge: Cambridge University Press.
- Eddy A, Welch D, and Rawes M (1969) The vegetation of the Moor House National Nature Reserve in the Northern Pennines, England. *Vegetatio* 16: 239–284.
- Elias SA (2001) Coleoptera and Trichoptera. In: Smol JP, Birks HJB, and Last WM (eds.) *Tracking Environmental Change Using Lake Sediments, Zoological Indicators*, vol. 4, pp. 67–80. Dordrecht: Kluwer Academic Publishers.
- Flatberg I (1986) Taxonomy, morphovariation, distribution and ecology of the *Sphagnum imbricatum* complex with main reference to Norway. *Gunneria* 54: 1–118.
- Francis DR (2001) Bryozoan statoblasts. In: Smol JP, Birks HJB, and Last WM (eds.) *Tracking Environmental Change Using Lake Sediments. Zoological Indicators*, vol. 4, pp. 105–123. Dordrecht: Kluwer Academic Publishers.
- Godwin H (1975) History of the British Flora. *A Factual Basis for Phytogeography*, 2nd edn. Cambridge: Cambridge University Press.
- Godwin H and Conway VM (1939) The ecology of a raised bog near Tregaron, Cardiganshire. *Journal of Ecology* 27: 313–363.
- Gore AJP and Urquhart C (1966) The effects of waterlogging on the growth of *Molinia caerulea* and *Eriophorum vaginatum*. *Journal of Ecology* 54: 617–633.
- Green BH (1968) Factors influencing the spatial and temporal distribution of *Sphagnum imbricatum* Hornsch. ex Russ in the British Isles. *Journal of Ecology* 56: 47–58.
- Grime JP, Hodgson JG, and Hunt R (1988) Comparative Plant Ecology. *A Functional Approach to Common British Species*. London: Unwin Hyman.
- Gronskis I and Snickovskis A (1989) Organization of cranberry growing in Latvia. Fourth International Symposium on *Vaccinium* culture. *Acta Horticulturae* 241: 77–80.
- Grosse-Brauckmann G (1972a) Über pflanzliche Makrofossilien mitteleuropäischer Torfe – I. Gewebereste krautiger Pflanzen und ihre Merkmale. *Telma* 2: 19–55.
- Grosse-Brauckmann G (1972b) Über pflanzliche Makrofossilien mitteleuropäischer Torfe – II. Weitere Reste (Früchte und Samen, Moose, UA.) und ihre Bestimmungsmöglichkeiten. *Telma* 4: 51–177.
- Grosse-Brauckmann G (1986) Analysis of vegetative plant macrofossils. In: Berglund BE (ed.) *Handbook of Holocene Palaeoecology*, pp. 591–618. Chichester: Wiley.
- Grosse-Brauckmann G and Streitz B (1992) Pflanzliche Makrofossilien mitteleuropäischer Torfe – III. Früchte, Samen und einige Gewebe (Fotos von fossilen Pflanzenresten). *Telma* 22: 53–102.
- Haas JN (1994) First identification key for charophyte oospores from central Europe. *European Journal of Phycology* 29: 227–235.
- Hammond RF, van der Krogt G, and Osinga T (1990) Vegetation and water tables on two raised bog remnants in county Kildare. In: Doyle GJ (ed.) *Ecology and Conservation of Irish Peatlands*, pp. 121–134. Dublin: Royal Irish Academy.
- Hannon GE, Bradshaw R, and Emborg J (2000) 6000 years of forest dynamics in Suserup Skov, a seminatural Danish woodland. *Global Ecology and Biogeography Letters* 9: 101–114.
- Harrison F (1988) Utilization of freshwater sponges in paleolimnological studies. *Palaeogeography, Palaeoclimatology, Palaeoecology* 62: 387–397.
- Haslam SM (1972) Biological flora of the British Isles: *Phragmites communis* Trin. (*Arundo phragmites* L., ? *Phragmites australis* (Cav.) Trin. ex Steudel). *Journal of Ecology* 60: 585–610.
- Hewett DG (1964) *Menyanthes trifoliata* L. *Journal of Ecology* 52: 723–735.
- Holmes JA (2001) Ostracoda. In: Smol JP, Birks HJB, and Last WM (eds.) *Tracking Environmental Change Using Lake Sediments, Zoological Indicators*, vol. 4, pp. 125–151. Dordrecht: Kluwer Academic Publishers.

- Hughes PDM and Barber KE (2003) Mire development across the fen–bog transition on the Teifi floodplain at Tregaron Bog, Ceredigion, Wales, and a comparison with 13 other raised bogs. *Journal of Ecology* 91: 253–264.
- Hughes PDM and Barber KE (2004) Contrasting pathways to ombrotrophy in three raised bogs from Ireland and Cumbria, England. *The Holocene* 14: 65–77.
- Hughes PDM, Mauquoy D, Barber KE, and Langdon P (2000) Mire-development pathways and palaeoclimatic records from a full Holocene peat archive at Walton Moss, Cumbria, England. *The Holocene* 10: 465–479.
- Jacquemart AL (1998) Biological flora of the British Isles: *Andromeda polifolia* L. *Journal of Ecology* 86: 527–541.
- Janssens JA (1983) A quantitative method for stratigraphic analyses of bryophytes in Holocene peat. *Journal of Ecology* 71: 189–196.
- Janssens JA (1992) Development of a raised bog complex. In: Wright HE, Coffin BA, and Aaseng NE (eds.) *The Patterned Peatlands of Minnesota*, pp. 189–221. Minneapolis: University of Minnesota Press.
- Johansson P (1995) *Vit Mossor i Norden*. Göteborg, Sweden: Vasastadens Bokbinderi AB.
- Johnson LC and Damman AWH (1991) Species controlled *Sphagnum* decay on a south Swedish raised bog. *Oikos* 61: 234–242.
- Katz NJ, Katz SV, and Kipiani MG (1965) Atlas and keys of fruits and seeds occurring in the Quaternary deposits of the USSR. *Nauka* 365. [in Russian].
- Katz NJ, Katz SV, and Skobeyeva EI (1977) Atlas of plant remains in peat soil. *Nedra*. [in Russian].
- Kilian MR, van der Plicht J, and van Geel B (1995) Dating raised bogs: New aspects of AMS ^{14}C wiggle matching, a reservoir effect and climatic change. *Quaternary Science Reviews* 14: 959–966.
- Kilian MR, van Geel B, and van der Plicht J (2000) ^{14}C AMS wiggle matching of raised bog deposits and models of peat accumulation. *Quaternary Science Reviews* 19: 1011–1033.
- Körber-Grohne U (1964) Bestimmungsschlüssel für subfossile Juncus-Samen und Gramineen-Früchte. Schriftenreihe des Niedersächsischen Landesinstitutes für Marschen und Wurtenforschung, Bd 7.W. Haarnagel, Hildesheim.
- Korhola AA and Tikkanen MJ (1997) Evidence for a more recent occurrence of water chestnut (*Trapa natans* L.) in Finland and its palaeoenvironmental implications. *The Holocene* 7: 39–44.
- Li Y and Vitt DH (1994) The dynamics of moss establishment: Temporal responses to nutrient gradients. *Bryologist* 97: 357–364.
- Li Y and Vitt DH (1995) The dynamics of moss establishment: Temporal responses to a moisture gradient. *Journal of Bryology* 108: 677–687.
- Lindholm T (1980) Dynamics of the height growth of the hummock dwarf shrubs *Empetrum nigrum* L. and *Calluna vulgaris* (L.) Hull on a raised bog. *Annales Botanici Fennici* 17: 343–356.
- Lindholm T and Markkula I (1984) Moisture conditions in hummocks and hollows in virgin and drained sites on the raised bog Laaviosuo, southern Finland. *Annales Botanici Fennici* 21: 241–255.
- Lindholm T and Vasander H (1981) The effect of summer frost damage on the growth and production of some raised bog dwarf shrubs. *Annales Botanici Fennici* 18: 155–167.
- Madden B and Doyle GJ (1990) Primary production on Mongan Bog. In: Doyle GJ (ed.) *Ecology and Conservation of Irish Peatlands*, pp. 147–161. Dublin: Royal Irish Academy.
- Mauquoy D and Barber KE (1999) Evidence for climatic deteriorations associated with the decline of *Sphagnum imbricatum* Hornsch. Ex Russ. in six ombrotrophic mires from Northern England and the Scottish Borders. *The Holocene* 9: 423–437.
- Mauquoy D, Blaauw M, van Geel B, et al. (2004a) Late Holocene climatic changes in Tierra del Fuego based on multi-proxy analyses of peat deposits. *Quaternary Research* 61: 148–158.
- Mauquoy D, van Geel B, Blaauw M, Speranza A, and van der Plicht J (2004b) Changes in solar activity and Holocene climate shifts derived from ^{14}C wiggle-match dated peat deposits. *The Holocene* 14: 45–52.
- McMullen JA, Barber KE, and Johnson B (2004) A palaeoecological perspective of vegetation succession on raised bog microforms. *Ecological Monographs* 74: 45–77.
- Meade R (1992) Some early changes following the rewetting of a vegetated cutover peatland surface at Danes Moss, Cheshire, UK, and their relevance to conservation management. *Biological Conservation* 61: 31–40.
- Moore JJ (1955) The distribution and ecology of *Scheuchzeria palustris* on a raised bog in Offaly. *Irish Nationalists' Journal* 12: 321–329.
- Nicholson BJ and Gignac LD (1995) Ecotope dimensions of peatland bryophyte indicator species along environmental and climate gradients in the Mackenzie River Basin. *Bryologist* 98: 437–451.
- Nilsson Ö and Hjelmquist H (1967) Studies on the outlet structure of south Scandinavian species of *Carex*. *Botaniska Notiser* 120: 460–485.
- Nilsson M, Klarqvist M, Bohlin E, and Possnert G (2001) Variation in ^{14}C age of macrofossils and different fractions of minute peat samples dated by AMS. *The Holocene* 11: 579–586.
- Nyholm E (1954–1969) *Illustrated Moss Flora of Fennoscandia*, vols. 1–6. Lund: Gleerup and Swedish National Research Council.
- Ohlson M and Malmer N (1990) Total nutrient accumulation and seasonal variation in resource allocation in the bog plant *Rhynchospora alba*. *Oikos* 58: 100–108.
- Økland RH (1989) A phytoecological study of the mire Northern Kisselbergmosen, SE Norway. I. Introduction, flora, vegetation and ecological conditions. *Sommerfeltia* 8: 1–172.
- Økland RH (1990) A phytoecological study of the mire Northern Kisselbergmosen, SE Norway. III. Diversity and habitat niche relationships. *Nordic Journal of Botany* 10: 191–220.
- Oldfield F, Thompson R, Crooks PRJ, et al. (1997) Radiocarbon dating of a recent high-latitude peat profile: Stor Åmyrån, northern Sweden. *The Holocene* 7: 283–290.
- Pals JP, van Geel B, and Delfos A (1980) Paleoeecological studies in the Klokkeveel bog near Hoogkarspel (Province of Noord-Holland). *Review of Palaeobotany and Palynology* 30: 371–418.
- Pellerin S and Lavoie C (2003) Reconstructing the recent dynamics of mires using a multitechnique approach. *Journal of Ecology* 91: 1008–1021.
- Phillips ME (1954) Biological Flora of the British Isles: *Eriophorum angustifolium* Roth. *Journal of Ecology* 42: 612–622.
- Robertson KP and Woolhouse HW (1984) Studies of the seasonal course of carbon uptake of *Eriophorum vaginatum* in a moorland habitat. II. The seasonal course of photosynthesis. *Journal of Ecology* 72: 685–700.
- Rodwell JS (ed.) (1991) *British Plant Communities. Mires and Heaths*, vol. 2. Cambridge: Cambridge University Press.
- Rudolph H and Voigt JU (1986) Effects of $\text{NH}_4^+\text{-N}$ and $\text{NO}_3^-\text{-N}$ on growth and metabolism of *Sphagnum magellanicum*. *Physiological Plant Pathology* 66: 339–343.
- Rydin H (1986) Competition and niche separation in *Sphagnum*. *Canadian Journal of Botany* 64: 1817–1824.
- Rydin H and McDonald AJS (1985) Tolerance of *Sphagnum* to water level. *Journal of Bryology* 13: 571–578.
- Schelvis J (1990) The reconstruction of local environments on the basis of remains of oribatid mites (Acari; Oribatida). *Journal of Archaeological Science* 17: 559–571.
- Schelvis J and van Geel B (1989) A palaeoecological study of the mites (Acari) from a Late-glacial deposit at Usselo (The Netherlands). *Boreas* 18: 237–243.
- Schoch WH, Pawlik B, and Schweingruber FH (1988) *Botanical Macro-Remains*. Berne and Stuttgart: Paul Haupt Publishers.
- Schofield JE and Bunting MJ (2005) Mid-Holocene presence of water chestnut (*Trapa natans* L.) in the meres of Holderness, East Yorkshire, UK. *The Holocene* 15: 687–697.
- Skene KR, Sprent JI, Raven JA, and Herdman L (2000) Biological flora of the British Isles: *Myrica gale* L. *Journal of Ecology* 88: 1079–1094.
- Smith AJE (2004) *The Moss Flora of Britain and Ireland*, 2nd edn. Cambridge: Cambridge University Press.
- Solhøy T (2001) Oribatid mites. In: Smol JP, Birks HJB, and Last WM (eds.) *Tracking Environmental Change Using Lake Sediments. Zoological Indicators*, vol. 4, pp. 81–104. Dordrecht: Kluwer Academic Publishers.
- Svensson G (1988) Fossil plant communities and regeneration patterns on a raised bog in South Sweden. *Journal of Ecology* 76: 41–59.
- Sweeney CA (2004) A key for the identification of stomata of the native conifers of Scandinavia. *Review of Palaeobotany and Palynology* 128: 281–290.
- Tallis JH and Birks HJB (1965) The past and present distribution of *Scheuchzeria palustris* L. in Europe. *Journal of Ecology* 53: 287–298.
- Taylor K, Rowland AP, and Jones HE (2001) Biological flora of the British Isles: *Molinia caerulea* (L.) Moench. *Journal of Ecology* 89: 126–144.
- Tolonen K (1966) Stratigraphic and rhizopod analyses on an old raised bog, Varrassuo, in Hollola, South Finland. *Annales Botanici Fennici* 3: 147–166.
- Tomlinson P (1985) An aid to the identification of fossil buds, bud-scales and catkin-bracts of British trees and shrubs. *Circaea* 3: 45–130.
- Väliranta M, Kaakinen A, and Kuhry P (2003) Holocene climate and landscape evolution east of the Pechora delta, East-European Russian Arctic. *Quaternary Research* 59: 335–344.
- van Dam H, van Geel B, van der Wijk A, et al. (1988) Palaeolimnological and documented evidence for alkalization and acidification of two moorland pools (The Netherlands). *Review of Palaeobotany and Palynology* 55: 273–316.
- van der Molen PC (1988) Palaeoecological reconstruction of the regional and local vegetation history of Woodfield Bog Co. Offaly. *Proceedings of the Royal Irish Academy* 88B: 69–97.

- van der Molen PC and Hoekstra SP (1988) A palaeoecological study of the hummock–hollow complex from Engbertsdyksveen in the Netherlands. *Review of Palaeobotany and Palynology* 56: 213–274.
- van der Molen PC, Schalkoort M, and Smit R (1994) Vegetation and ecology of hummock–hollow complexes on an Irish raised bog. *Biology and Environment: Proceedings of the Royal Irish Academy* 94B: 145–175.
- van Dinter M and Birks HH (1996) Distinguishing fossil *Betula nana* and *B. pubescens* using their wingless fruits: Implications for the late-glacial vegetational history of western Norway. *Vegetation History and Archaeobotany* 5: 229–240.
- van Geel B (1978) A palaeoecological study of Holocene peat bog sections in Germany and the Netherlands. *Review of Palaeobotany and Palynology* 25: 1–120.
- van Geel B, Bohncke SJP, and Dee H (1981) A palaeoecological study of an upper Late Glacial and Holocene sequence from “De Borchert”, The Netherlands. *Review of Palaeobotany and Palynology* 31: 367–448.
- van Geel B, Coope GR, and van der Hammen T (1989) Palaeoecology and stratigraphy of the Late Glacial type section at Usselo (the Netherlands). *Review of Palaeobotany and Palynology* 60: 25–129.
- van Geel B, Hallewas DP, and Pals JP (1983) A late Holocene deposit under the Westfriese Zeedijk near Enkhuizen (Prov. of N-Holland, The Netherlands): Palaeoecological and archaeological aspects. *Review of Palaeobotany and Palynology* 38: 269–335.
- van Geel B, Klink AG, Pals JP, and Wiegers J (1986) An Upper Eemian lake deposit from Twente, eastern Netherlands. *Review of Palaeobotany and Palynology* 47: 31–37.
- van Geel B and Middelorp AA (1988) Vegetational history of Carbury Bog (Co. Kildare, Ireland) during the last 850 years and a test of the temperature indicator value of $^{2}\text{H}/^{1}\text{H}$ measurements of peat samples in relation to historical sources and meteorological data. *New Phytologist* 109: 377–392.
- van Geel B, Mur LR, Ralska-Jasiewiczowa M, and Goslar T (1994) Fossil akinetes of *Aphanizomenon* and *Anabaena* as indicators for medieval phosphate-eutrophication of Lake Gosciarz (Central Poland). *Review of Palaeobotany and Palynology* 83: 97–105.
- van Geel B, Odegaard BV, and Ralska-Jasiewiczowa M (1996) *Cyanobacteria as Indicators of Phosphate-Eutrophication of Lakes and Ponds in the Past*. Rixensart, Belgium: PACT 50 (European Center for the Cultural Heritage).
- Walker IR (2001) Midges: Chironomidae and related Diptera. In: Smol JP, Birks HJB, and Last WM (eds.) *Tracking Environmental Change Using Lake Sediments. Zoological Indicators*, vol. 4, pp. 43–66. Dordrecht: Kluwer Academic Publishers.
- Watson EV (1968) *British Mosses and Liverworts*. Cambridge: Cambridge University Press.
- Wein RW (1973) Biological flora of the British Isles: *Eriophorum vaginatum* L. *Journal of Ecology* 61: 601–615.
- Wheeler BD and Proctor MCF (2000) Ecological gradients, subdivisions and terminology of north-west European mires. *Journal of Ecology* 88: 187–203.

Paleolimnological Applications

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Plant macroremains are usually well preserved in anaerobic lake sediments. The macrofossil assemblages are usually dominated by aquatic and shore plants and the plants in the catchment. The most common subfossil plant remains are seeds, fruits, and leaves, but other characteristic parts may also be identified and used in reconstructions. For a more comprehensive overview of plant macroremains in lake sediments, see Birks (1980, 2001).

This article focuses on the contribution of plant macrofossil analysis to paleolimnology. We discuss the past development of the macrophyte flora (submerged, floating-leaved, and emergent lakeshore plants) and of lake ecosystems in general, including water characteristics (pH, nutrients, and temperatures), and eutrophication; we also describe the use of plant macrofossils in reconstructions of past lake level changes.

The Role of Macrophytes in a Lake Ecosystem and Factors Affecting Their Distribution in a Lake

Macrophytes are an important component of lake ecosystems and their processes (Jeppesen et al., 1998). They play a primary role in the formation and stabilization of lake sediments, such as gyttja and marl. They recycle nutrients from the sediment and the lake water, which influences the trophic status of water bodies. They provide food and habitats for many other lake organisms (algae and animals). Thus, the analysis of plant remains in lake sediments provides information on these aspects of the lake ecosystem through time.

The principal factors affecting the structure and composition of aquatic vegetation within a lake are water depth (and related water transparency; Hannon and Gaillard, 1997), water chemistry and nutrient-status (pH, alkalinity, base cations, nitrogen, and phosphorus; Jackson and Charles, 1987; Scheffer et al., 1993), and temperature (Iversen, 1954; Kolstrup, 1980) (Table 1 and Figures 1 and 2).

Lakes can be classified into four main ecological types (Lang, 1994; Hannon and Gaillard, 1997), which have characteristic littoral vegetation zones (Figure 3):

1. Eutrophic lakes (high phosphorus and nitrate content, pH 7 or above, calcareous or not) with well-developed belts of lakeshore vegetation (supralittoral and eulittoral zones) and of submerged and floating-leaved species (sublittoral zone), with each belt being related to a range of water depths.
2. Oligotrophic lakes (low phosphorus content, calcareous rich with pH 7 or above, or calcareous poor with pH 6 or 7) with less dense and diverse littoral vegetation than eutrophic lakes and few distinct vegetation zones.
3. Mesotrophic lakes are intermediate between eutrophic and oligotrophic lakes.

4. Dystrophic lakes (very low phosphorus, noncalcareous, pH 6 or lower) often lack macrophytes. They may have floating marginal mats of *Sphagnum* and *Carex* spp.

Newly formed lakes are rapidly colonized by macrophytes, often in a few decades. A hydrosere (aquatic succession) then proceeds, controlled by internal factors in the lake. The hydrosere may be disrupted or diverted by external factors, such as local or regional human-induced or climatic factors, that result in changes in lake nutrient status and/or water chemistry (Figure 4). Natural hydroseres progress over centuries and the paleolimnological macrofossil record is the only way of documenting long-term hydrosere processes in lake ecosystems (Birks, 2000) and is necessary even if historical records are available (Davidson et al., 2005). Hence, paleolimnology relates the ecological history of lakes to their present ecosystems (Smol, 2002).

Examples of Plant Macrofossil Analysis in Paleolimnological Studies

Late Glacial Paleolimnology

The contribution of plant macrofossil analysis to multidisciplinary studies of Late Glacial lake sediments is discussed elsewhere in this encyclopedia. Iversen (1954) emphasized the indicator value of aquatics as proxies of climate change because they behave as pioneer species, not requiring mature soil to grow. Because they find ideal conditions in a lake relatively soon after its formation, Iversen argued that aquatic plants should react more rapidly to climate warming than terrestrial plants. Aquatics have been used as Late Glacial temperature indicators in The Netherlands (Kolstrup, 1980) and in southwest Switzerland (Gaillard, 1984a, 1985; Table 1). In Switzerland, aquatics were already present during the oldest Late Glacial period from ca. 15 000 uncal ¹⁴C BP (Figure 5). Many of the species present during that period (*Myriophyllum spicatum*, *Potamogeton gramineus*, *P. praelongus*, *P. filiformis*, and *Ranunculus* sect. *Batrachium*) became more abundant from the time of expansion of the dwarf birch (from ca. 14 000 uncal ¹⁴C BP). *Typha latifolia*, *Nymphaea alba*, *Utricularia* spp. *Potamogeton lucens*, *P. perfoliatus*, *P. pusillus*, *P. pectinatus*, *Alisma plantago-aquatica*, *Myriophyllum verticillatum*, and *Hippuris* first appeared with the development of tree birch from 12 700 uncal ¹⁴C BP. On the basis of that data and the indicator value of aquatics (Table 1), Gaillard (1985) suggested that July temperatures had already exceeded 10 °C at the end of the Oldest Dryas at approximately 14 000 uncal ¹⁴C BP. A further increase to more than 13 °C took place ca. 12 700 uncal ¹⁴C BP.

In contrast, in Norway at Kråkenes, *Ranunculus* sect. *Batrachium* and *Nitella* were the earliest pioneers after deglaciation

Table 1 Minimum and optimum mean July temperatures of some lake macrophytes^a

Macrophyte taxa	Minimum mean July temperature (°C)	Optimum mean July temperature (°C)
<i>Alisma plantago aquatica</i>	>10	>13
<i>Ceratophyllum demersum</i>	13	15
<i>Cladium mariscus</i>	>15	17
<i>Hippuris vulgaris</i>	>10	>13
<i>Menyanthes trifoliata</i>	>10	>13
<i>Myriophyllum alterniflorum</i>	10	>10
<i>Myriophyllum spicatum</i>	10	>10
<i>Myriophyllum verticillatum</i>	>10	>13
<i>Nuphar lutea</i>	>15	17
<i>Nymphaea alba</i>	>10	>13
<i>Potamogeton filiformis</i>	8	
<i>Potamogeton gramineus</i>	8	
<i>Potamogeton lucens</i>	>10	>13
<i>Potamogeton pectinatus</i>	>10	>13
<i>Potamogeton perfoliatus</i>	>10	>13
<i>Potamogeton praelongus</i>	8	
<i>Potamogeton pusillus</i>	>10	>13
<i>Schoenoplectus lacustris</i>	>10	>13
<i>Typha latifolia</i>	12–13	15
<i>Typha angustifolia</i>	>14	

^aAccording to Kolstrup (1980) and Gaillard (1984a).

ca. 12 300 uncal ¹⁴C BP (Figure 6; Birks, 2000). The vegetational succession was very slow between 12 000 and 10 900 uncal ¹⁴C BP, inhibited by cool conditions similar to those above treeline in Norway today. The rapid cooling at the start of the Younger Dryas stadial resulted in almost complete extermination of limnic plants and animals.

The Late Glacial and Holocene (18 000–6000 cal BP) plant macrofossil record at Lake Zeribar, Zagros Mountains, Iran, documents increased salinity (occurrence of *Ruppia maritima*) during cold periods. The presence of *Ceratophyllum demersum* during the Bølling/Allerød period was ascribed to rising temperatures. The macrofossil record from Zeribar also provides information on lake level fluctuations.

Holocene Paleolimnology

A feature of the early Holocene macrofossil flora is the abundance and rapid successions of macrophytes. In central Europe (Gaillard, 1984a; Ammann and Tobolski, 1983), Late Glacial taxa such as *Ranunculus* sect. *Batrachium*, *Potamogeton* spp. (e.g. *P. gramineus*, *P. pusillus*, *P. lucens*, and *P. filiformis*), and *Myriophyllum spicatum*, are often replaced by *Nymphaea alba*, *Nuphar lutea*, *Ceratophyllum demersum*, and *Trapa natans* and emergents such as *Comarum palustre*, *Cladium mariscus*, *Menyanthes trifoliata*, and *Carex* spp. (Figure 5).

In southwestern Switzerland (Gaillard, 1984a, 1985), the early Holocene is characterized by a renewed development of *Potamogeton* spp., *M. spicatum*, and *Nymphaea alba*, and by the first occurrence of *Ceratophyllum demersum*. The occurrence of the thermophilous lakeshore species *Cladium mariscus* suggests

that minimum July temperatures had reached ca. 17 °C in the earliest Holocene (Table 1).

Kråkenes, western Norway (Birks, 2000) provides a contrast. The biotic response to the rapid warming of the early Holocene (10 000 uncal ¹⁴C BP; 11 530 cal BP) was immediate. The pioneers *Ranunculus* sect. *Batrachium* and *Nitella* expanded within one to three decades, closely followed by other elodeids. From 11 430 to 10 720 cal BP, a remarkable isoetid succession took place, with phases dominated by *Limosella aquatica*, *Subularia aquatica*, *Elatine hydropiper*, and *Isoetes echinospora*. Approximately 800 years into the Holocene, most of the macrophytes declined, which was probably related to nutrient depletion in the base-poor catchment (Birks, 2000). Approximately 550 years later, *I. lacustris* and *Nymphaea* colonized, and a stable flora developed (Figure 6).

At Lake Zeribar, Iran, the replacement of *Najas marina* by *Najas minor* approximately 11 000 cal yr BP was interpreted in terms of increasing temperatures and decline in salinity (Wasylikowa, 2005).

During the middle and late Holocene, changes in the macrophyte flora were generally due to processes such as infilling, eutrophication, or acidification, or to lake level fluctuations. Although it was prominent in the early Holocene, *Cladium mariscus* often declined from 7000 to 6000 cal yr BP (Gaillard, 1984a). This strongly calciphilous species was affected by the gradual leaching of carbonates and the progressive acidification of soils. In Scandinavia, *Cladium* is still common on the islands of Öland and Gotland, both characterized by calcareous bedrocks.

Väliiranta (2005) investigated changes in aquatic and terrestrial environments of northeastern European Russia and Finnish Lapland using plant macrofossil analysis, pollen, and other proxy data. Similar long-term patterns in early Holocene expansion of aquatics, a subsequent maximum in aquatic species richness, and a decline or disappearance of aquatics after the middle Holocene were recorded regardless of the lakes' locations in different vegetation and climate zones (Figure 7; Kultti et al., 2003, 2004). Väliiranta suggests that the early Holocene warming and sufficient nutrient status enabled the immigration and establishment of aquatic plant communities. Subsequently, as summer temperatures gradually declined through the Holocene, the length of the open-water season (i.e. temperature-related parameters) probably controlled the absence or presence of limnophytes.

Coastal Environments

The history of coastal environments may provide information on past changes in sea level. There are many examples of such studies in the literature; however, seldom are they supported by plant macrofossil records.

In the context of an archaeological study of the Mesolithic settlements (ca. 7000–5500 uncal yr BP) at Skateholm, Skåne, south coast of Sweden, the sediments from an ancient lagoon were studied for pollen, plant macrofossils, insects, and diatoms (Gaillard and Lemdahl, 1988). The record of plant remains was remarkably species rich (Figures 8 and 9). The local vegetation around the settlements and the trophic state, the salinity, and the water level of the lagoon were reconstructed. Between the settlements and the sea, a wetland landscape was characterized

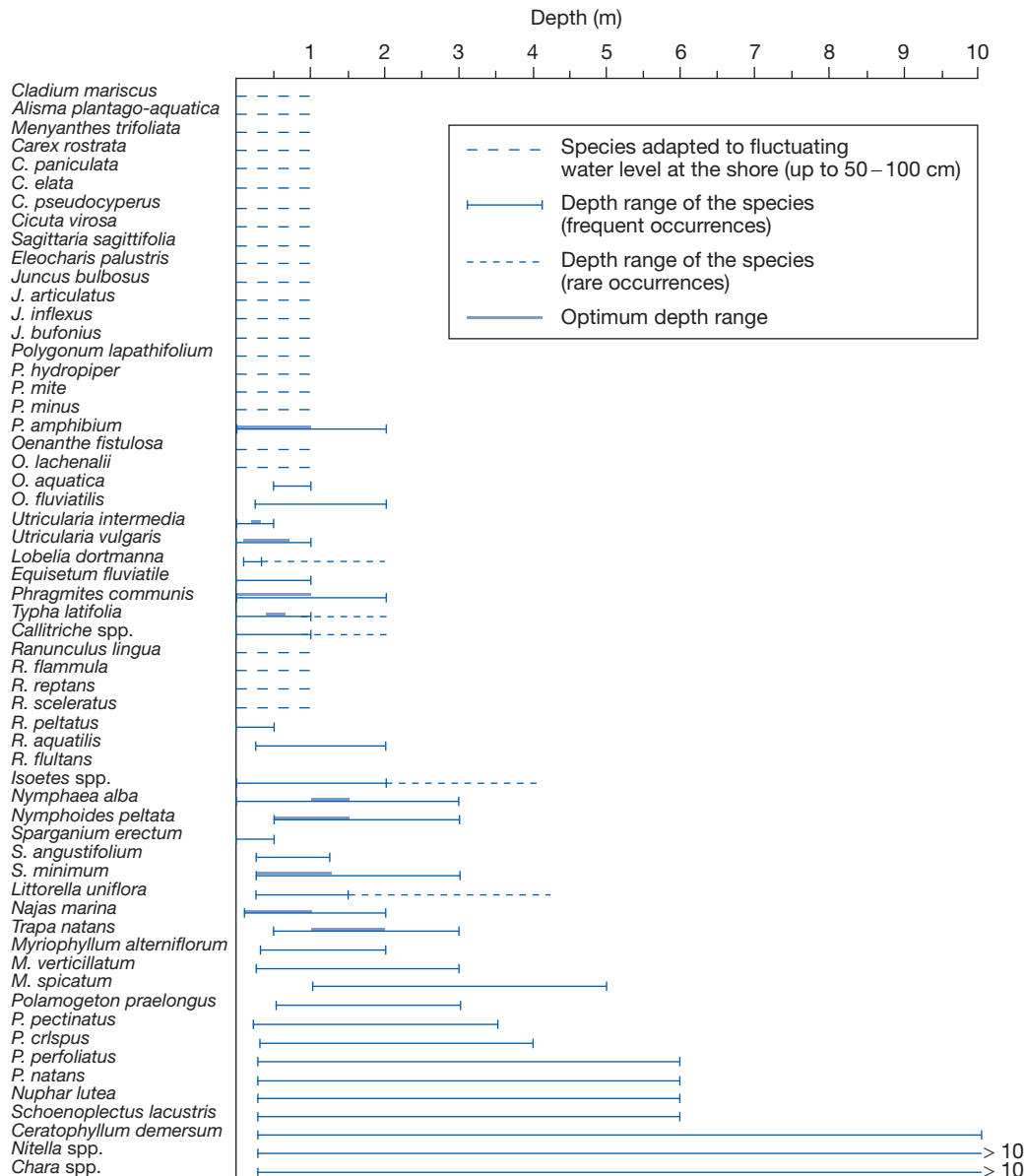


Figure 1 Potential depth ranges of emergent, floating-leaved, and submerged lakeshore and aquatic taxa. After Hannon GE and Gaillard M-J (1997) The plant-macrofossil record of past lake-level changes. *Journal of Paleolimnology* 18:15–28.

by zones of swampy alder forest; reed swamp (*Phragmites*); and open, shallow brackish, mesotrophic to eutrophic water (with *Ruppia maritima*, *Zannichellia palustris*, *Najas marina*, and *Chara*). The lagoon was particularly shallow and sheltered between 6900 and 6200 BP. The sea progressively invaded the basin (*Littorina* transgression), which resulted in the retreat of the marginal alder woods. Between 5800 and 4000 BP, the lagoon was more open to the sea, becoming a shallow bay.

Eutrophication

Very few modern lake ecosystems are unaffected by human activity. A common source of ecosystem disruption is eutrophication through the addition of nutrients (nitrates and phosphates). They may derive from agricultural runoff or from

sewage. In several cases, the effects have been highly undesirable. Macrophyte productivity increases. Boat propellers and fishing lines become entangled. The decay of the increased bioproduction results in anaerobic conditions, which can lead to fish kills and the production of unpleasant smells from sulfides in the sediments. Dense macrophyte growth shades out smaller species and alters the habitats and balance of invertebrates and fish. If phytoplankton and epiphytic algal growth increases, macrophytes may be shaded out and a complete reorganization occurs (alternative stable state; Scheffer et al., 1993) in which, during summer, the water is full of cyanobacteria (green slime) and macrophytes are absent.

How are the presence and abundance of macrophytes during cultural eutrophication represented by macrofossils? Davidson et al. (2005) tested the macrofossil record at Groby

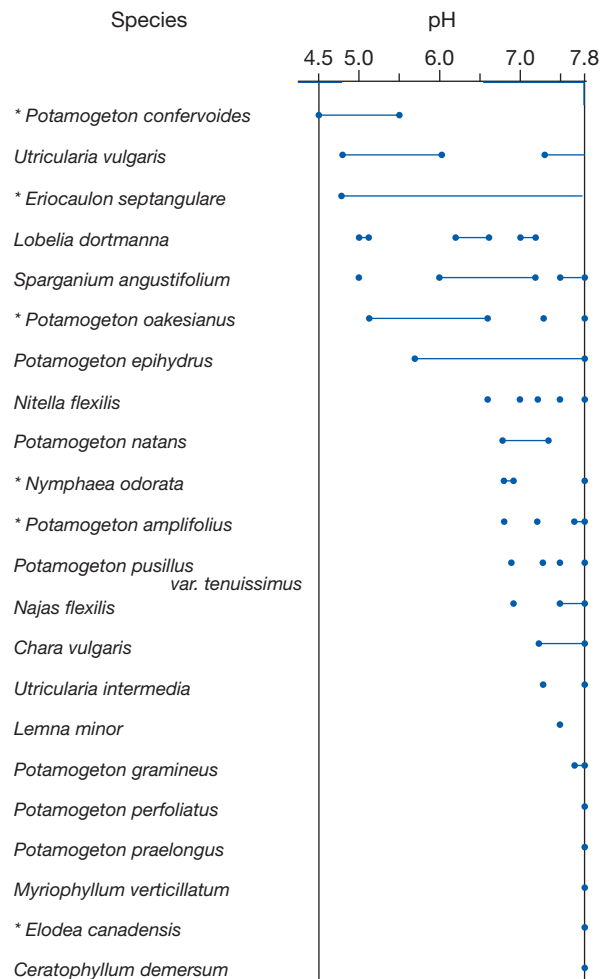


Figure 2 Water pH ranges of aquatic macrophyte taxa from 31 unproductive lakes in the Adirondack Mountains (New York). Non-European species are indicated with an asterisk. After Jackson ST and Charles DF (1987) Aquatic macrophytes in Adirondack (New York) lakes: Patterns of species composition in relation to environment. *Canadian Journal of Botany* 66: 1449–1460; simplified by Hannon GE and Gaillard M-J (1997) The plant-macrofossil record of past lake-level changes. *Journal of Paleolimnology* 18: 15–28.

Pool, central England, against an exceptional floristic record spanning 250 years. Approximately 40% of the macrophytes were recorded as macrofossils, highlighting their irregular production and poor dispersal. However, several taxa were found as macrofossils that were not recorded in the lake, showing that visual recording of macrophytes is also fraught with difficulties. In general, the sediments recorded more detailed changes than the rather irregular recording visits, but both records showed similar changes over time, suggesting that, even if it is incomplete, the macrofossil record provides good insight into macrophyte community dynamics and successions.

A 7000-year-long macrofossil record from Gundsømagle Sø, a lake in eastern Denmark currently dominated by *Nymphaea alba* (Rasmussen and Anderson, 2005), shows limnological responses to changes in external factors such as climate, land use, and changing chemical composition of the inflow water and also to internal (autogenic) factors such as competition and

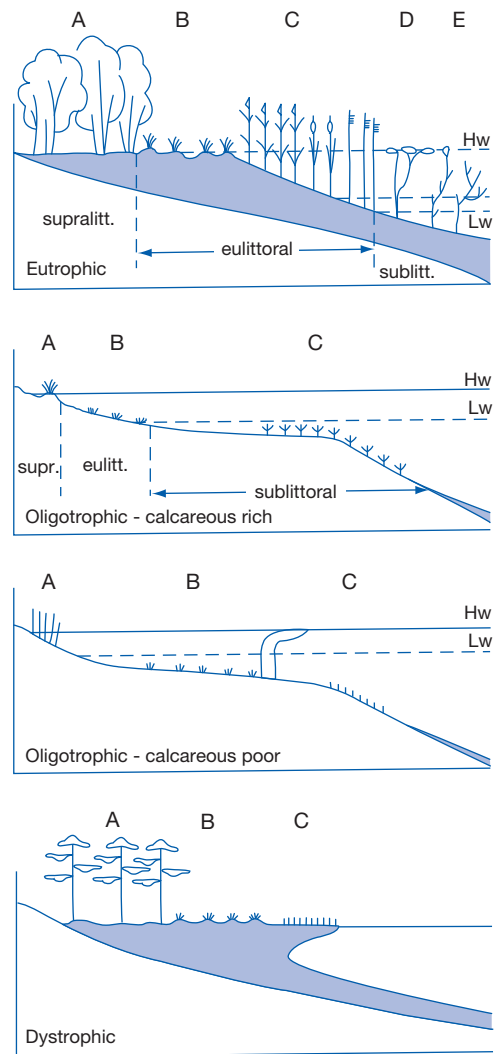


Figure 3 The three major ecological lake types of the paleoarctic and neoarctic regions and their shore vegetation. Eutrophic lakes: A, alder swamps (*Alnus glutinosa*); B, sedge vegetation (Magnocaricion); C, reed vegetation (Phragmition); D, floating-leaved vegetation (Nymphaeion); and E, submerged vegetation (Potamogetonion p.p.). Oligotrophic lakes (calcareous-rich): A, nitrophilous vegetation on wet ground; B, *Eleochariton acicularis* and *Deschampsion litoralis*; and C, Charetaillia and Potamogetonion p.p. Oligotrophic lakes (calcareous-poor): A, sedge vegetation (*Carex rostrata*, *Caricion nigrae*); B, *Isoetion lacustris* and *Lobelia*; and C, mats of *Nitella* (Nitelletalia). Dystrophic lakes: A, pine wood or wooded mire (*P. sylvestris* or *P. rotundata*, Dicrano Pinion); B, *E. vaginatum*, *Sphagnum magellanici*; and C, floating Cyperaceae mats. supr. or supralitt., supralittoral; sublitt., sublittoral; eulitt., eulittoral; Hw, high water level; Lw, low water level. After Hannon GE and Gaillard M-J (1997) The plant-macrofossil record of past lake-level changes. *Journal of Paleolimnology* 18: 15–28.

sediment accumulation (Figure 10). The macrofossils record sparse macrophyte vegetation characteristic of clear-water alkaline mesotrophic lakes before ca. 3200 BP, probably limited by low light in the relatively deep (>5 m) water at the coring site. Subsequent erosion from extensive Bronze and Iron Age clearances and agriculture increased the sediment load to the lake. Submerged and emergent macrophytes expanded in the

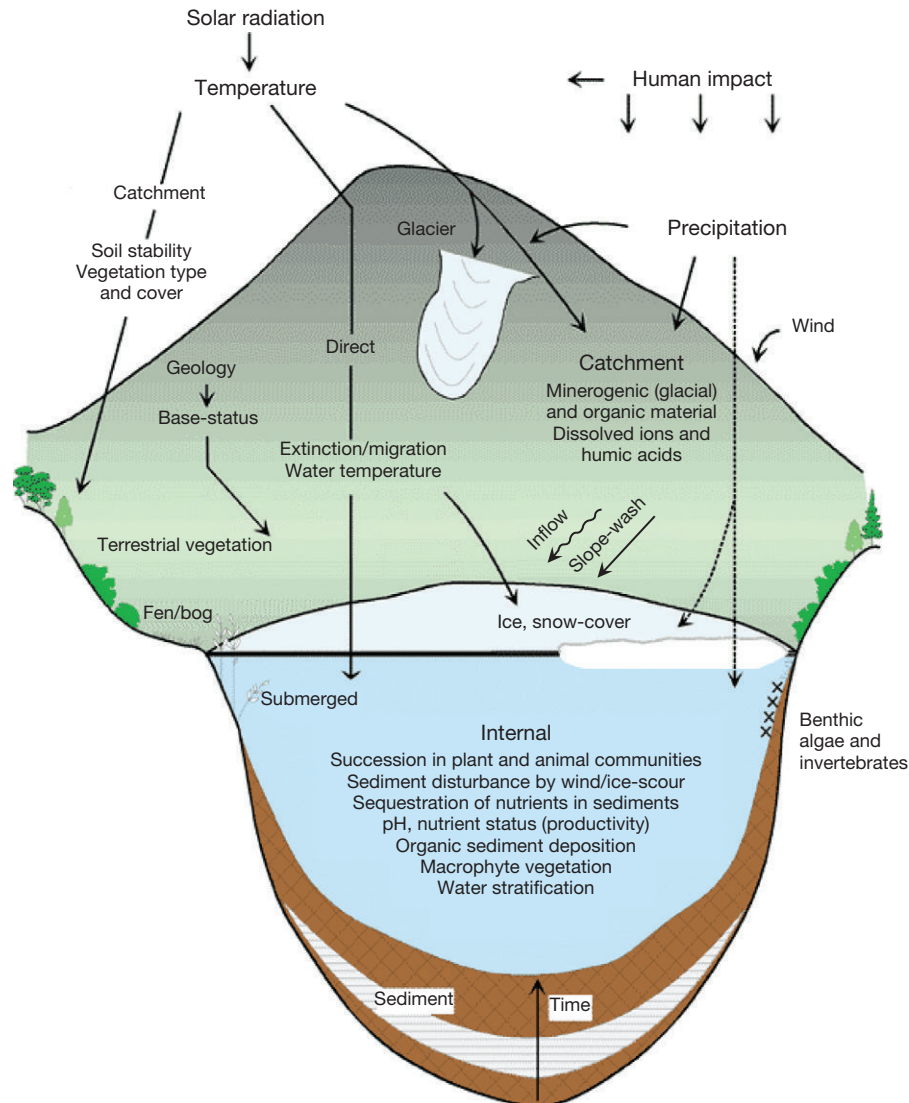


Figure 4 A diagrammatic representation of the aquatic ecosystem. The ecosystem includes the lake and the catchment. External factors affect both the catchment and the lake. Internal (autogenic) factors act in the catchment and the state of the catchment affects the lake. The lake is also affected by its own numerous internal factors. Modified from Birks HH, Battarbee RW, and Birks HJB (2000) The development of the aquatic ecosystem at Kråkenes Lakes, Western Norway, during the late-glacial and early Holocene—a synthesis. *Journal of Paleolimnology* 23: 91–114.

shallower water due to a combination of sediment accumulation and lake-level lowering during a dry climatic phase. Later, taxa of eutrophic conditions expanded (*Ceratophyllum*, *Potamogeton crispus*, *P. pectinatus*, *Zannichellia palustris*, and *Stratiotes aloides*) at the expense of the weakly competitive mesotrophic taxa *Nitella*, *Najas flexilis*, and *N. marina*. This change coincided with retting (soaking) of flax and hemp in the lake, especially between 800 and 150 BP (1150–1800 AD). At approximately 0 BP (1950 AD), submerged and emergent macrophytes disappeared from the macrofossil record and were recorded as absent in lake surveys. This switch to an alternative stable state (Scheffer et al., 1993) was probably triggered by light limitation by the mass of phytoplankton and floating-leaved *N. alba* that had increased in response to the ever-increasing nutrient load from farming and sewage. In Lake Søbygaard, central Denmark, the process has gone a stage further (Brodersen et al.,

2001). The macrofossil record shows that, as in Gundsømagle Sø, *N. alba* and *Potamogeton crispus* replaced submerged aquatics (mainly *Ceratophyllum*) in the early 20th century. After approximately 1970, no submerged macrophytes were recorded and *N. alba* became sparse. Today, the shallow water is turbid with suspended phytoplankton and submerged macrophytes are absent.

A multiproxy paleolimnological investigation of the eutrophicated Lake Sallie in North Dakota, USA (Birks et al., 1976) revealed stages in its eutrophication that coincided with urban and agricultural development and sewage treatment. After European settlement in 1870, macrofossil analysis showed that a hard-water flora dominated by *Chara* and *Najas flexilis* was replaced by increasing quantities of *Ceratophyllum demersum* as phosphate increased, followed by *Potamogeton* spp. and *Ruppia maritima*, characteristic of high-conductivity lakes, until

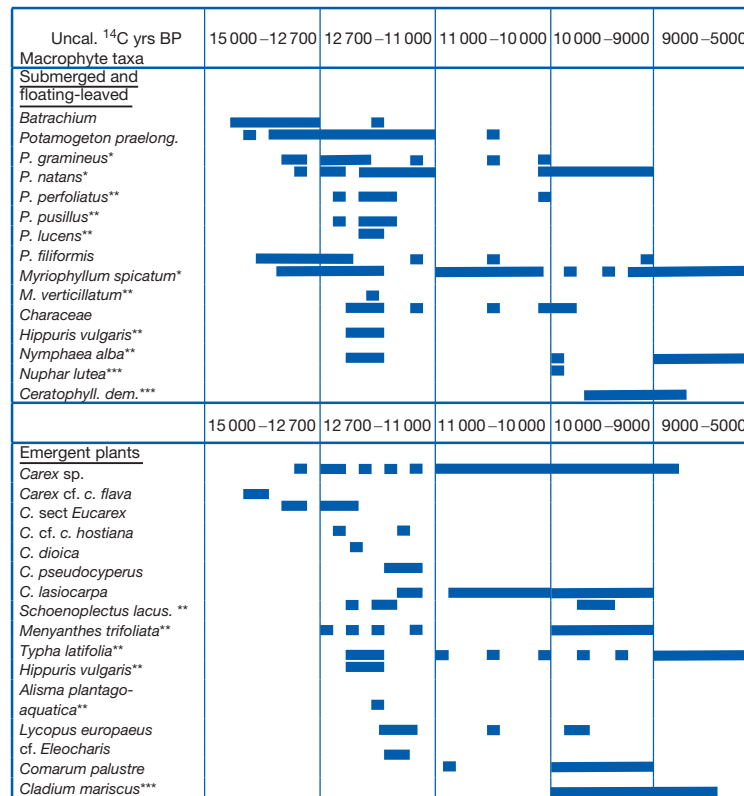


Figure 5 Late Glacial and Holocene occurrence of macrophytes in lakes of the southwestern Swiss Plateau. *First occurrence at the beginning of the last regional pollen assemblage zone (rPAZ) of the Oldest Dryas, pre-12 700 uncal BP, characterized by *Betula nana*. **First occurrence during the *Juniperus-Hippophae* rPAZ or the *Betula-Gramineae* rPAZ (i.e., post 12 700 uncal BP). ***First occurrence at the beginning of the Holocene, from 10 000 uncal BP (11 500 cal BP). After Gaillard M-J (1984a) Etude palynologique de l'évolution tardi- et postglaciaire de la végétation du Moyen-Pays romand (Suisse). Dissertationes Botanicae 77.

Kråkenes lake Aquatic Macrofossils

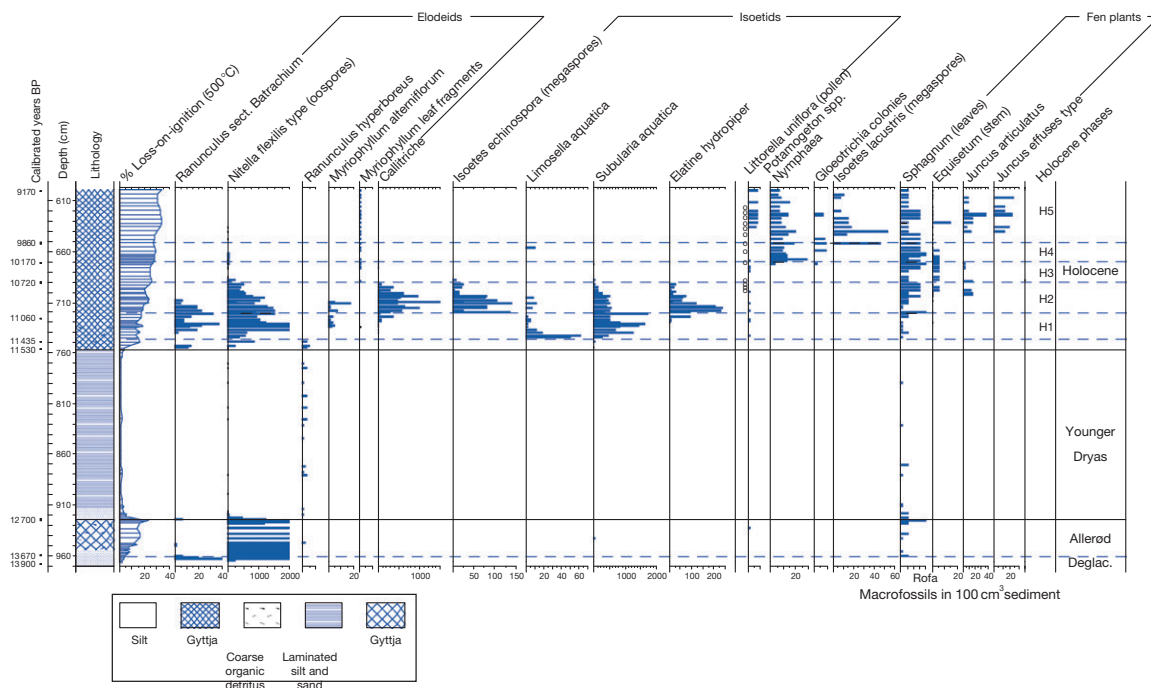


Figure 6 Selected macrofossil taxa from the Late Glacial and early Holocene sediments of Kråkenes Lake, western Norway. The shallow-water isoetids are *Limosella aquatica*, *Subularia aquatica*, *Elatine hydropiper*, and *Littorella uniflora*. H3 is the phase of sparse macrophytes. Adapted from Birks HH (2001) Plant macrofossils. In: Smol JP, Birks HJB, and Last WM (eds.) *Tracking Environmental Change Using Lake Sediments*, pp. 49–74. Dordrecht, The Netherlands: Kluwer.

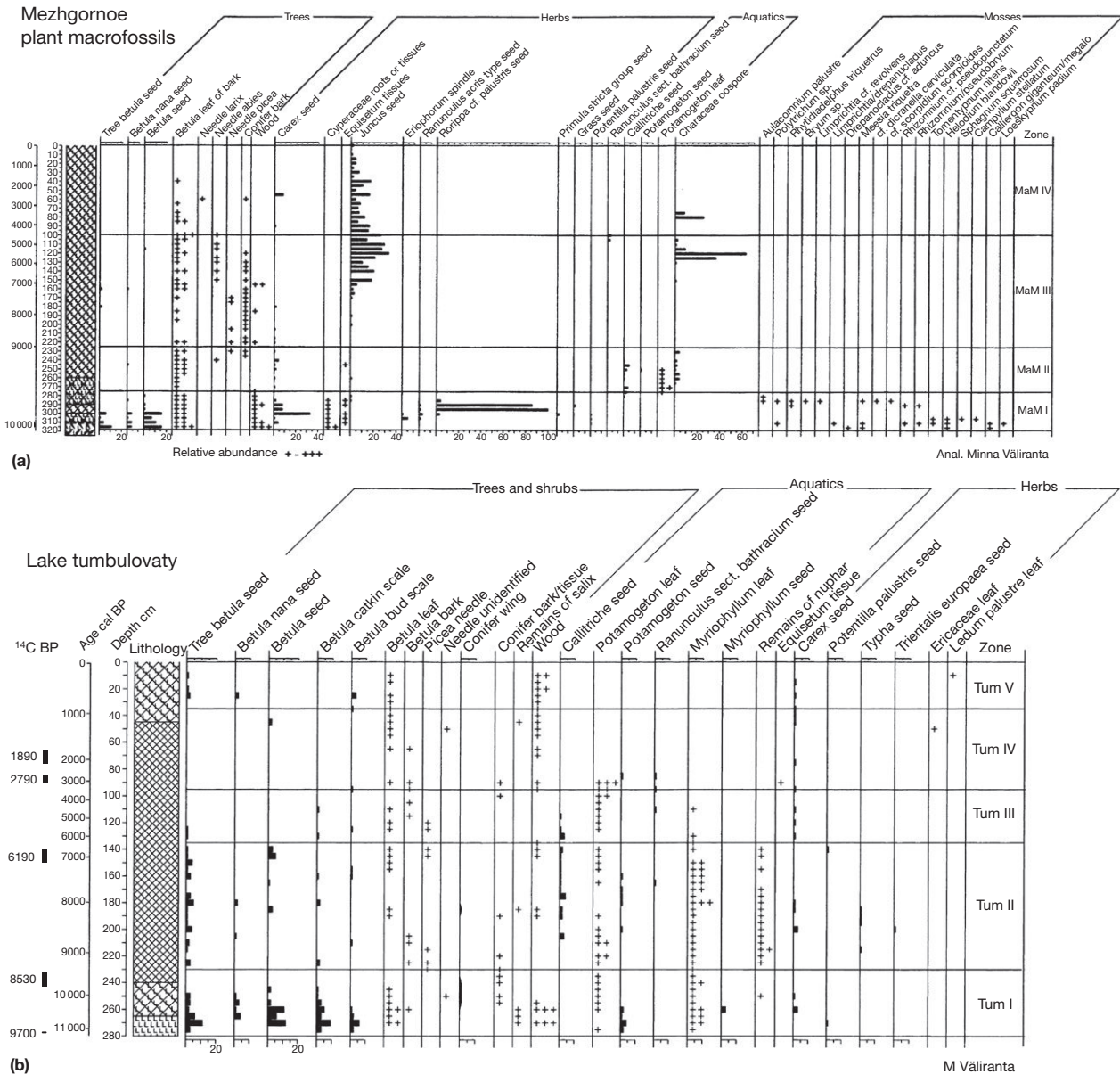


Figure 7 (a) Plant macrofossils from Lake Mezhgornoe, northeast European Russia. The results are presented as concentrations per 20 cm³ sediment or as relative abundance, i.e. +, rare; ++, occasional; and +++, abundant. Sediments are silt with detritus in the bottom, followed by gyttja silt and silt gyttja. (b) Plant macrofossils from Lake Tumbulovaty, northeast European Russia. Results presented as in (a). Sediments are silt, followed by silt gyttja and gyttja, with silt gyttja again at the top. From [Väliaranta M \(2005\) Plant Macrofossil Evidence of Changes in Aquatic and Terrestrial Environments in North-eastern European Russia and Finnish Lapland since Late Weichselian](#). PhD thesis No. 181, Department of Geology, University of Helsinki. ISSN 1795-3499, ISBN 952-10-2153-5 (PDF).

a dense macrophyte growth prevailed, coarse fish replaced the game fish, H₂S evolved from the anaerobic sediments, and cyanobacteria blooms dominated the phytoplankton in summer. These changes occurred in 100 years between European settlement and 1970 when the core was taken.

Rapid changes during the past century were also documented in nine North African lakes and lagoons (Birks et al., 2001) within the multidisciplinary CASSARINA Project (Flower, 2001). The lakes in Morocco and Tunisia were impacted by the withdrawal of freshwater for agriculture and

cities. However, in the Nile Delta, Egypt, the seasonal Nile flood was progressively controlled by dams, leading to a year-round supply of freshwater. The paleolimnological sequence from Edku Lake ([Figure 11](#)) shows natural variations in salinity before ca. 1940. Macrophytes and freshwater molluscs, including biharzia-carrying *Biomphalaria alexandrina*, expanded rapidly after 1964. Eutrophication was aggravated by the introduction of nitrogen-fixing *Azolla filiculoides* in 1992. The ecosystem has been further modified by the spread of a floating blanket of *Eichhornia crassipes* (water hyacinth).

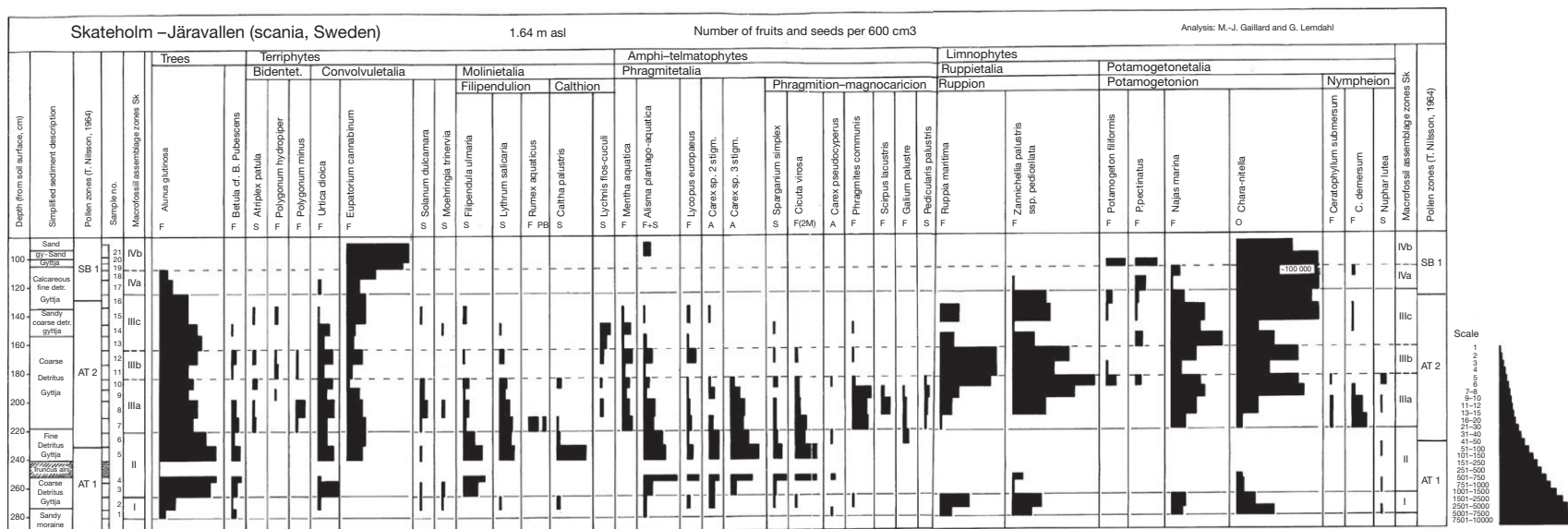


Figure 8 Plant macrofossil diagram from Skateholm-Järvavallen, southern coast of Skåne, southern Sweden. A, achene; F, fruit; F(2 M), 1 fruit = 2 mericarps; O, oospore; PB, pericarp bract; S, seed; 2st, 2 stigmas. From Gaillard M-J and Lemdahl G (1988) Plant macrofossil analysis (seeds and fruits) at Skateholm-Järvavallen, southern Sweden—A lagoonal landscape during the Atlantic and early Subboreal time. In: Larsson L (ed.) *The Skateholm Project I. Man and Environment*, pp. 34–38. Stockholm: Almqvist & Wiksell.

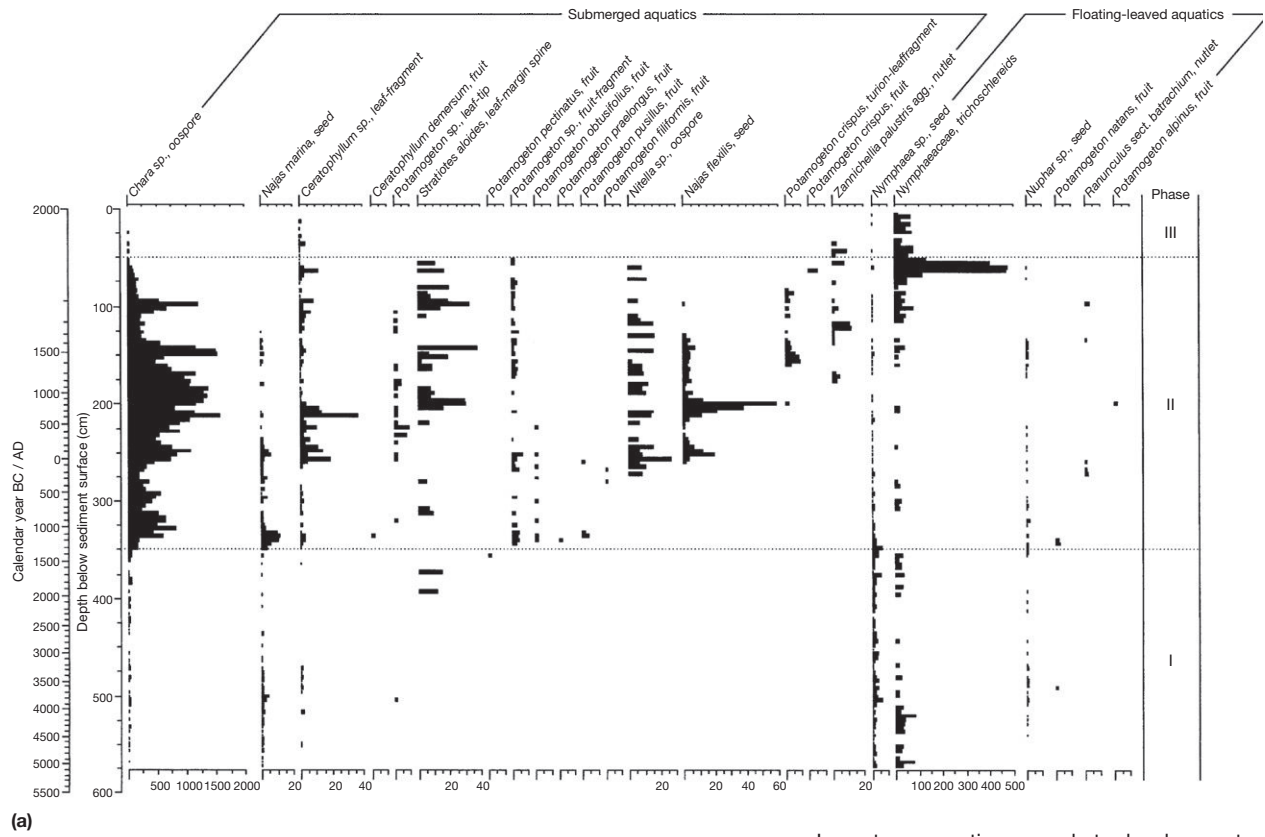
	Past plant assemblages	Macrofossil zones	Modern plant associations	T°	Trophic state	Water depth
		I II IIIa IIIb IIIc IVa IVb	1 2 3 4 5 6 7 8 9 10		a b c	A B C D E
A	<i>Alnus glutinosa</i>	• • • • • •	*			
A	<i>Betula cf. B. pub.</i>	• • • • •	*			
B	<i>Eupatorium cann.</i>	• • • • • •	* *			
B	<i>Urtica dioica</i>	• • • • • •	* * *			
C	<i>Filipendula ulmaria</i>	• • • • •	* * * * *		x	
C	<i>Alisma p.-aquat.*</i>	• • • • • •	* * *	x		
C	<i>Lycopus europaeus</i>	• • • • •	* * * *	x		
C	<i>Lythrum salicaria*</i>	• • • • •	* * * *			
C	<i>Sparganium simpl.*</i>	• • • • •	* *		x	
C	<i>Cicuta virosa*</i>	• • • • •	* * * *	x	x	
C	<i>Carex spp.*</i>	• • •	* * * *			
C	<i>Caltha palustris</i>	• • •	* * * * *			
C	<i>Galium palustre</i>	• •	* * * * *			
C	<i>Moehringia trin.</i>	• • • • •	*			
D	<i>Solanum dulcamara</i>	• • • • •	* * *		x	x
D	<i>Nuphar lutea*</i>	• • •	*			
D	<i>Lysimachia vulg.</i>	•	* *			
D	<i>Rumex aquaticus*</i>	•	* * *			
D	<i>Scutellaria galeric.</i>	•	* * * * *		x	
D	<i>Pedicularis pal.</i>	•	* *			
D	<i>Phragmites comm.*</i>	• • • • •	* * * * *	x	x	x
D	<i>Scirpus lac.*</i>	•	* * * *	x		x
D	<i>Ceratophyll. dem.*</i>	• • • • •	* *	x	x	x
E	<i>Mentha aquatica</i>	• • • • •	* * * * *			
E	<i>Atriplex patula</i>	• • •	*			
E	<i>Polygonum hydrop.</i>	• • •	* * *		x	
E	<i>Polygonum minus</i>	• • •	* * *	x		
E	<i>Ranunculus scel.</i>	• • •	*			
E	<i>Lychnis flos-cuculi</i>	• • •	* *			
F	<i>Potamogeton filif.*</i>	• • • • •	* * *		x	x
F	<i>Potamogeton pect.*</i>	• • • • •	* * *		x	x
G	<i>Najas marina*</i>	• • • • •	* * * *	x	x	x
G	<i>Ruppia maritima*</i>	• • • • •	*			x
G	<i>Zannichellia pal.*</i>	• • • • •	*			x

Figure 9 The distribution of plant macrofossils in the profile from Skateholm-Järvallen, southern coast of Skåne, southern Sweden, with the taxa arranged in order of similar occurrence, grouped into past assemblages A–G, and compared with phytosociological associations. The modern ecological requirements of the taxa for summer temperature (T°), water trophicity, and water depth are shown as well. Macrophyte taxa are shown with an asterisk. From Gaillard M-J and Lemdahl G (1988) Plant macrofossil analysis (seeds and fruits) at Skateholm-Järvallen, southern Sweden—A lagoonal landscape during the Atlantic and early Subboreal time. In: Larsson L (ed.) *The Skateholm Project I. Man and Environment*, pp. 34–38. Stockholm: Almqvist & Wiksell.

Plant macrofossils formed the basis of the aquatic ecosystem reconstructions in the CASSARINA Project. Detrended correspondence analyses of the sequences (Birks and Birks, 2001) showed unprecedented amounts of ecosystem turnover within 100 years in all the lakes, much more than any paleolimnologically documented natural rates of change. As the external stresses are maintained, the rates of change are increasing, emphasizing the pressures on wetland ecosystems in arid countries.

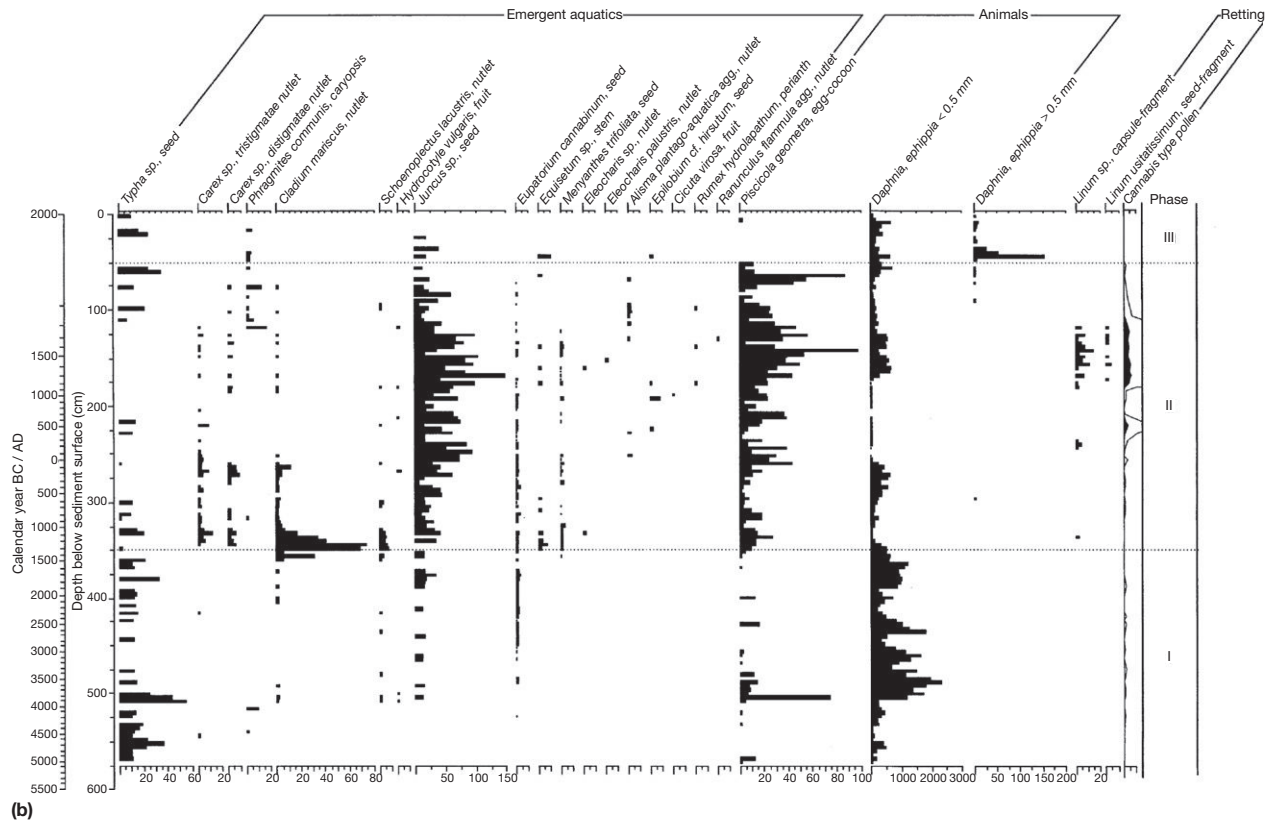
The Contribution of Macrofossil Analysis to Studies of Past Lake-Level Changes

Plant macrofossils provide the most useful biostratigraphical data for reconstruction of former lake-level changes because the distribution of macrophytes depends mainly on water depth. In some species, the water-depth range is sufficiently narrow to determine water-level changes, but most species have relatively broad ranges (Figure 1). Water chemistry and



(a)

Long-term aquatic macrophyte development



(b)

Figure 10 (Continued)

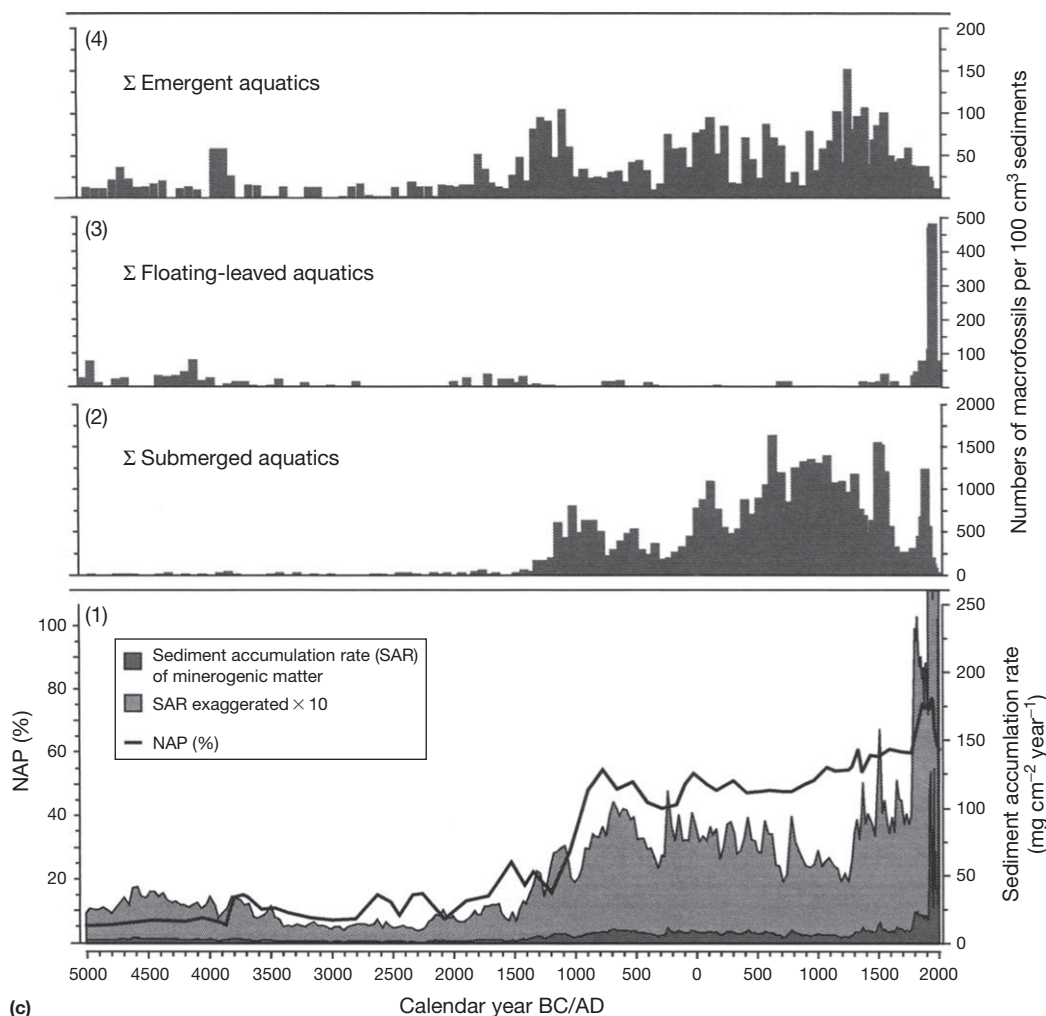


Figure 10 Plant macrofossils from Lake Gundsømagle, Zealand, Denmark. (a) Submerged and floating-leaved aquatics. (b) Emergent aquatics. Concentrations in 100 cm³ of sediment. Note: The curve for *Cannabis* type is a pollen percentage curve (percentages as dark silhouettes; values exaggerated 10 times as lines). The diagrams are plotted on a depth scale with time as a side axis. (c) 1, Sediment accumulation rate (SAR) of minerogenic matter (proxy for soil erosion) and nonarboreal pollen (NAP) (proxy for open land); 2, summary of total number of macrofossil remains of submerged aquatics; 3, summary of total number of macrofossil remains of floating-leaved aquatics; and 4, summary of total number of macrofossil remains of emergent aquatics. The diagrams are plotted on a horizontal timescale. From Rasmussen P and Anderson NJ (2005) Natural and anthropogenic forcing of aquatic macrophyte development in a shallow Danish lake during the last 7000 years. *Journal of Biogeography* 32: 1993–2005.

nutrient status must also be considered when interpreting water-level changes (Figure 2). Water-level changes have climatic significance if they are determined by temperature and precipitation changes.

Most emergent species (Figure 3) tolerate interannual water depth fluctuations of 1 m but do not grow in water deeper than 1 m. *Schoenoplectus lacustris* and *Phragmites communis* may occupy the shoreline or the sublittoral zone (Figure 3) outside the *Nymphaea*–*Nuphar* zone. Sublittoral species generally have broad water-depth ranges, favoring deeper water (Figure 1). *Nymphaea alba*, *Najas marina*, and *Nuphar lutea* prefer water depths of 0.5–2.0 m, whereas *Myriophyllum spicatum*, several *Potamogeton* species, Characeae, and *Ceratophyllum demersum* frequently grow in 6 m of water (10 m for *Ceratophyllum*).

The overlap of habitats for most species listed in Figure 1 means that the results from a single core are difficult to interpret unequivocally in terms of water-depth fluctuations. However,

within one lake at a given stage in its development, depth ranges of the macrophytes should be relatively consistent. Thus, fluctuations in abundance and composition in the stratigraphical macrofossil record at a littoral coring point should primarily reflect spatial movements of vegetation belts and, therefore, changes in water depth. This phenomenon has been demonstrated in several European and North American lakes (Hannon and Gaillard, 1997). Movements of vegetation zones are most easily detected in shallow, species-rich eutrophic and mesotrophic lakes (Figure 3) (Hannon and Gaillard, 1997).

The spread of macrophytes toward the center of the lake may result from lake-level lowering or from sediment accumulation. In a single core, the plant macrofossil record of these two processes is often indistinguishable. Therefore, analyses of a transect of cores perpendicular to the shoreline is essential for reliable lake-level interpretations (Gaillard, 1984b; Digerfeldt, 1986). In addition, sediment accumulation can be

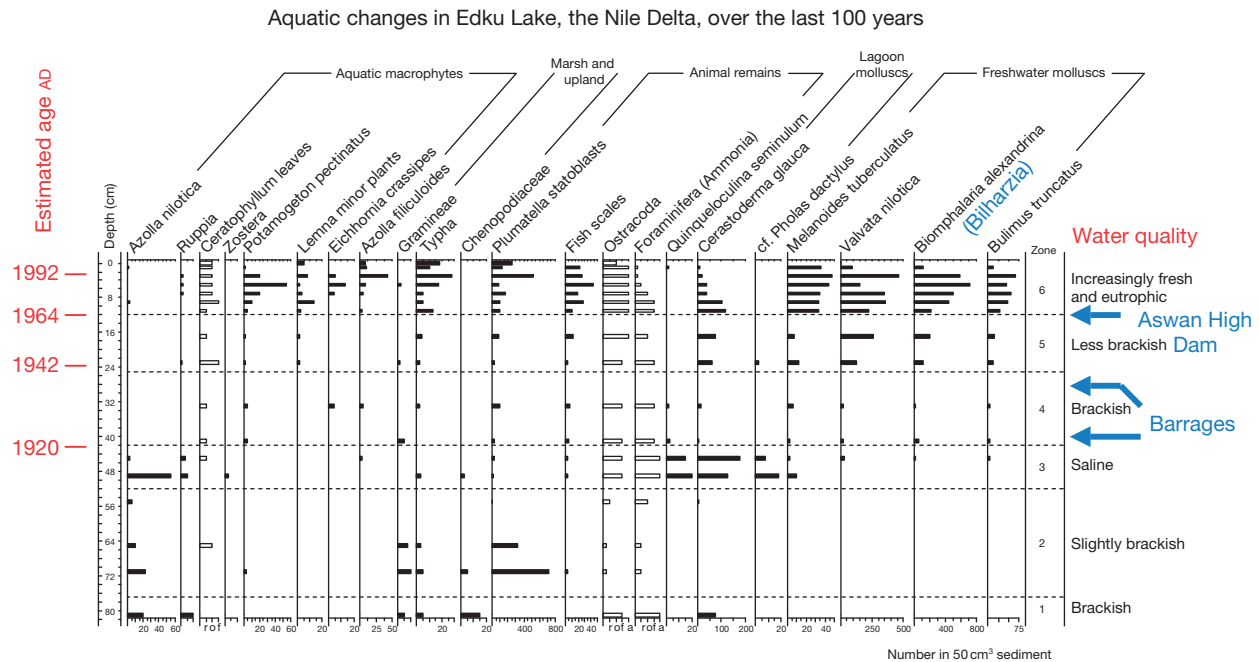


Figure 11 A plant and animal macrofossil diagram through recent sediments in Edku Lake, the Nile Delta, Egypt. Approximate ages are shown on the left and inferred water-quality changes are shown on the right. The timing of dam constructions is shown in blue. Data from Birks HH, Peglar SM, Boomer I, Flower RJ, and Ramdani M (2001) Palaeolimnological responses of nine North African lakes in the CASSARINA Project to recent environmental changes and human impact detected by plant macrofossil, pollen, and faunal analyses. *Aquatic Ecology* 35: 405–430.

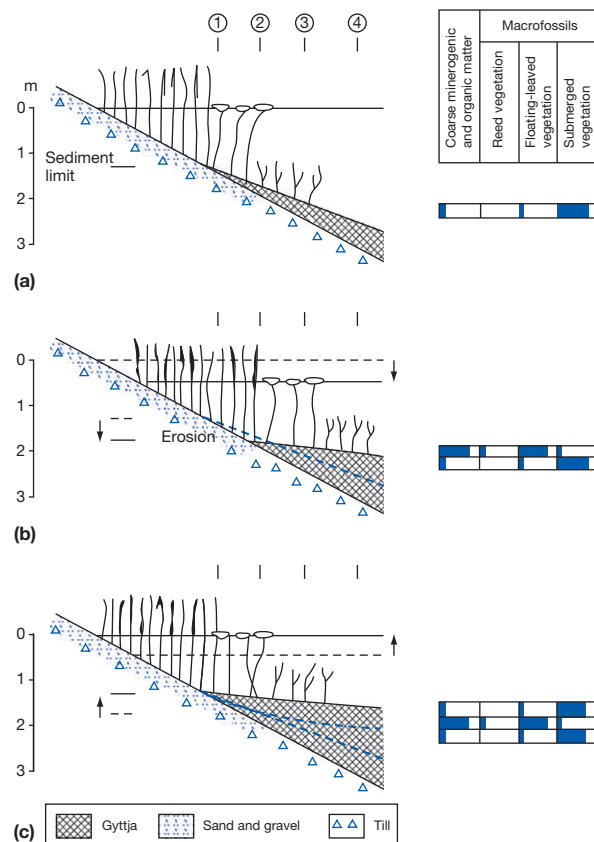


Figure 12 The recommended methodology for reconstruction of past lake-level changes—that is, a combination of evidence inferred from

distinguished from a lake-level lowering by changes in the level of the sediment limit (Figure 12) (Gaillard and Digerfeldt, 1991). The use of several cores also enables changes in water depth to be distinguished from effects of chemistry and nutrient status changes (Digerfeldt, 1986; Hannon and Gaillard, 1997). This is particularly important when records of past lake-level changes are used as proxy records of past climate (humidity) change.

the analysis of a series of cores taken on a transect perpendicular to the shore toward the center of the lake. The most useful evidence is inferred from the core sediment stratigraphy—and the identification of the ‘sediment limit’ for various time periods—and from the plant macrofossil record. The figure shows the effects of lake-level changes on the distribution of shore vegetation, the level of the ‘sediment limit’ (boundary between organic and pure minerogenic sediments), and sediment composition. (a) The outward spread of the lake vegetation is determined primarily by water depth, and the level of the sediment limit is determined by exposure. The content of coarse minerogenic and organic matter in the sediment decreases outwards from the shore. The schematic diagram on the right shows the macrofossil and sediment composition in core 3. (b) A lowering in lake level will cause a lowering of sediment limit and erosion of the marginal sediment. As recorded in the schematic macrofossil diagram, the lowering will also result in an outward spread of the lake vegetation and an increase in the content of coarse matter in core 3. (c) A subsequent rise in lake level will produce a rise in the sediment limit, an inward displacement of the lake vegetation, and a decrease in the content of coarse matter in core 3. From Gaillard M-J and Digerfeldt G (1991) Palaeohydrological studies and their contribution to paleoecological and paleoclimatic reconstructions. *Ecological Bulletin* 41: 275–282.

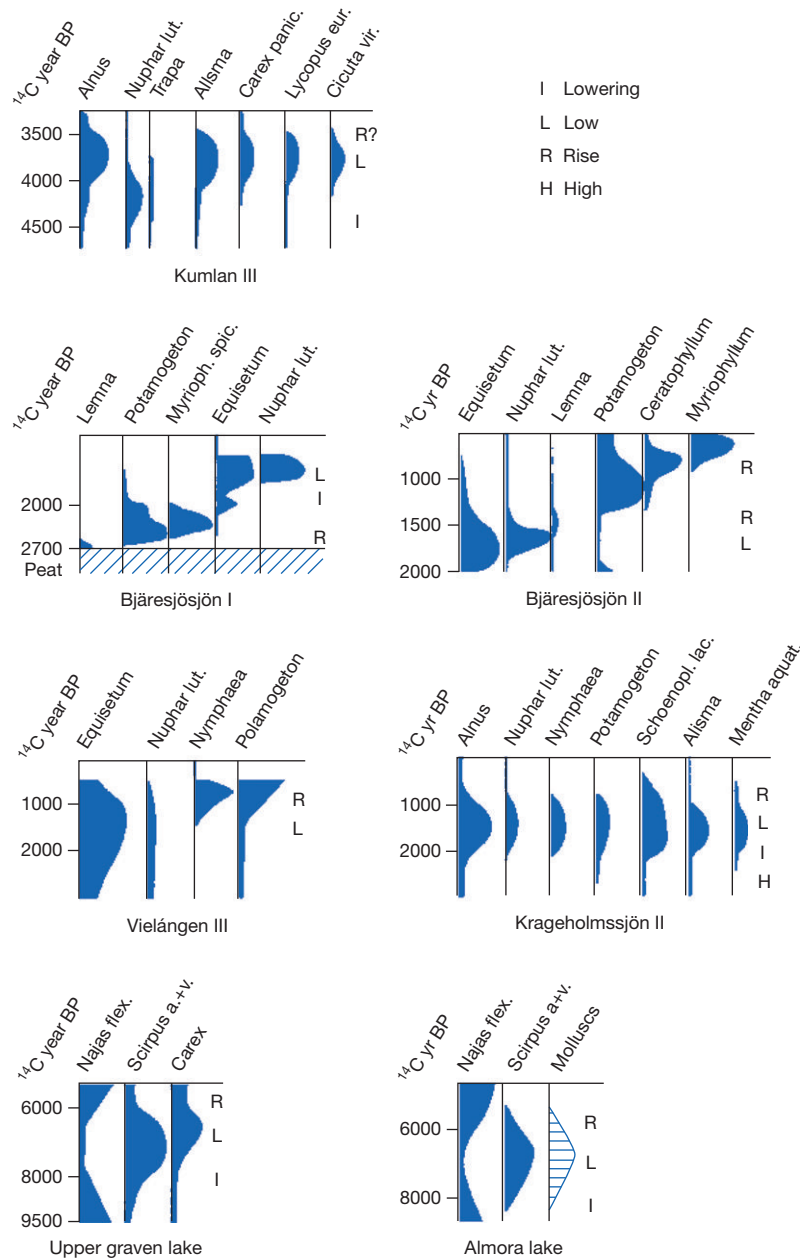


Figure 13 Simplified plant macrofossil diagrams from sites investigated for lake-level changes in Europe (Sweden) and the United States (Minnesota), showing successions interpreted in terms of lake-level changes. The interpretation in terms of lake-level changes (shown at the right of each diagram) is inferred from the original publications and the syntheses published by Gaillard and Digerfeldt (1991). The diagrams are plotted against timescales. a, acutus; aquat, aquatilis; Equiset, Equisetum; eur, europaeus; flex, flexilis; fl, fluviatile; lac, lacustris; lat, latifolia; lut, lutea; m, mariscus; Menyanth, Menyanthes; Myrioph, Myriophyllum; panic, paniculata; Schoenopl, Schoenoplectus; spic, spicatum; trif, trifoliata; v, validus; LG, Late Glacial. From Hannon GE and Gaillard M-J (1997) The plant-macrofossil record of past lake-level changes. *Journal of Paleolimnology* 18: 15–28.

The movement of macrophyte zones toward the shore can only be interpreted as a rise in lake level. Note, however, that increased exposure in large lakes may cause a similar change. This is one reason why not all lakes are suitable for lake-level studies.

Figures 13 and 14 show examples of two major types of Holocene plant macrofossil successions associated with lake-level changes (Hannon and Gaillard, 1997): (1) the ideal succession of sublittoral species followed by eulittoral species (or vice versa) (Vielångan I and II, Torreberga I and II, Trummen,

Bjäresjön I and II, Kumlan I and III, Upper Grave lake, and Almora Lake) and (2) the simultaneous increase (or decrease) of all lakeshore plants (eulittoral and sublittoral) (Krageholmssjön II, Bökeberg, and Kumlan II). The first type of record is the most elegant demonstration of a change in water depth at the coring point, but the second type is also useful. A reduction in abundant macrophytes remains indicates that the lakeshore is situated further from the coring point (Figure 15).

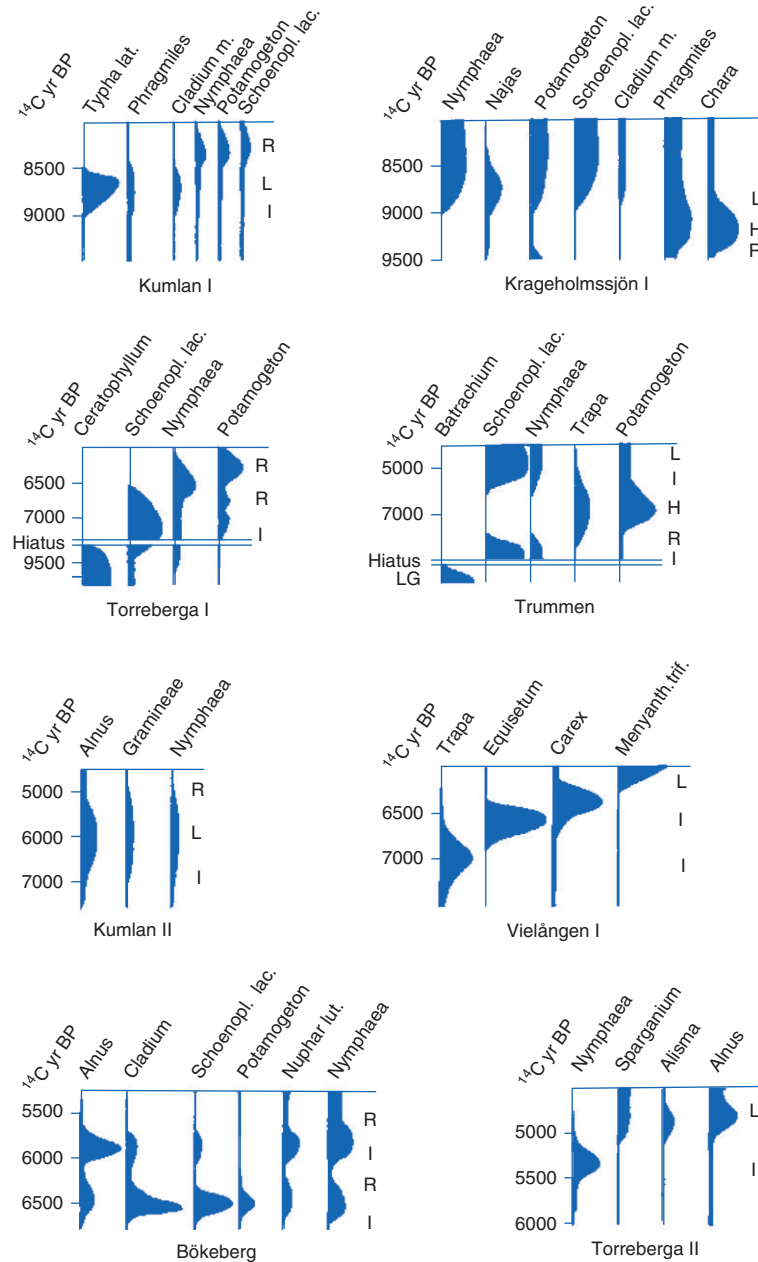


Figure 14 Simplified plant macrofossil diagrams from sites investigated for lake-level changes in Europe (Sweden) and the United States (Minnesota), showing successions interpreted in terms of lake-level changes. For abbreviations, see the legend to [Figure 13](#). From [Hannon GE and Gaillard M-J \(1997\)](#) The plant-macrofossil record of past lake-level changes. *Journal of Paleolimnology* 18: 15–28.

The Role of Macrofossils in Elucidating Causes of Paleolimnological Changes—Hypothesis Testing

Usually, the causes of paleolimnological change are inferred from the data in a ‘narrative’ way (e.g. the causes of eutrophication). Alternative factors are rarely tested. Within a well-designed multiproxy study, hypotheses about the causes of paleolimnological change can be made from part of the data set and tested against an independent part of the data set.

This novel approach was used in a multiproxy study of Holocene paleolimnology at Sägistalsee, Swiss Alps ([Lotter](#)

and [Birks, 2003](#)). The lake is at the present treeline, an important ecotone where the vegetation physiognomy changes. Changes in treelines and tree species limits can be detected in a sediment record using macrofossils. All the Sägistalsee proxies were recorded on one core, thus avoiding problems of multiple-core correlation. Principal components analysis was used to summarize the changes in all the proxies. Plant macrofossils were used to reconstruct vegetation changes in the catchment as well as in the lake, particularly the occurrence of trees and grazing indicators.

Table 2 Significance of the responses by terrestrial pollen and lake proxies to the predictors, climate and catchment vegetation, during the Holocene at Sägistalsee^a

'Responses' (proxies)	Scale	Is climate a significant predictor?	Is catchment vegetation a significant predictor?
Terrestrial Pollen	Catchment and regional	Y	Y
Macrofossils	Catchment	—	—
Lake biota			
Chironomids	Lake	N	Y
Cladocera	Lake	N	Y
Lake abiotic proxies			
Sediment grain size	Lake	—	Y
Magnetic parameters	Lake	—	Y
Geochemistry	Lake	—	(Y)

^aClimate parameters used were the insolation curve, the central European cold phases, and the Atlantic IRD record. The macrofossil record was used to reconstruct catchment vegetation. Statistically significant positive relationships are indicated by Y and significant negative relationships by N. Nonsignificance is indicated by dashes. Geochemistry was only slightly significantly related to catchment vegetation (Y). The most influential catchment changes were the development of closed *Picea* forest and stable soils and the forest clearances and catchment use by Bronze and Iron Age people. Results are from Lotter AF and Birks HJB (2003) The Holocene paleolimnology of Sägistalsee and its environmental history—A synthesis. *Journal of Paleolimnology* 30: 333–342.

Changes in a lake ecosystem can result from autogenic changes or changes in its catchment. Both can be affected by external factors, particularly climate or human activity (Figure 4). Two hypotheses were tested at Sägistalsee that might explain the Holocene paleolimnological changes: (1) Climate exerted significant control and (2) catchment vegetation changes have been a major influence. The macrofossil record was used as the basis for the catchment vegetation reconstructions. The Holocene climate factors used were the insolation curve, the central European cold phases, and the Atlantic ice-rafted detritus record. The summarized responses of the other proxies were tested against the two predictors using partial redundancy analyses and restricted Monte Carlo permutation tests. The results (Table 2) showed that climate had no significant effect on lake variables, but that it did influence regional vegetation development inferred from pollen data. In contrast, the catchment vegetation was a significant predictor for all the lake variables except geochemistry, influencing grain size and magnetism (which reflect soil stabilization and erosion) and dominating the effects on the lake biota.

This study showed how intimately a lake is linked to its catchment. It also demonstrated the key role played by plant macrofossils within this multiproxy study. The hypothesis-testing approach will undoubtedly become more widespread in paleolimnology in attempts to determine the factors influencing lake ecosystem changes.

See also: Plant Macrofossil Introduction. Lake Level Studies: West-Central Europe. Plant Macrofossil Methods and Studies: Surface Samples, Taphonomy, Representation; Treeline Studies. Plant Macrofossil Records: Late Glacial Multidisciplinary Studies.

References

- Ammann B and Tobolski K (1983) Vegetation development during the Late-Würm at Lobsigensee (Swiss Plateau). *Studies in the Late Quaternary of Lobsigensee 1. Rev. Paléobiol* 2: 163–180.
- Birks HH (1980) Plant macrofossils in Quaternary lake sediments. *Archiv für Hydrobiologie. Ergebnisse der Limnologien* 15: 60.
- Birks HH (2000) Aquatic macrophyte vegetation development in Kråkenes Lake, western Norway, during the late-glacial and early-Holocene. *Journal of Paleolimnology* 23: 67–76.
- Birks HH (2001) Plant macrofossils. In: Smol JP, Birks HJB, and Last WM (eds.) *Tracking Environmental Change Using Lake Sediments*, vol. 3, pp. 49–74. The Netherlands: Kluwer, Dordrecht.
- Birks HH and Birks HJB (2001) Recent ecosystem dynamics in nine North African lakes in the CASSARINA Project. *Aquatic Botany* 35: 461–478.
- Birks HH, Whiteside MC, Stark D, and Bright RC (1976) Recent paleolimnology of three lakes in northwestern Minnesota. *Quaternary Research* 6: 249–272.
- Birks HH, Peglar SM, Boomer I, Flower RJ, and Ramdani M (2001) Palaeolimnological responses of nine North African lakes in the CASSARINA Project to recent environmental changes and human impact detected by plant macrofossil, pollen, and faunal analyses. *Aquatic Ecology* 35: 405–430.
- Brodersen KP, Odgaard BV, Vestergaard O, and Anderson NJ (2001) Chironomid stratigraphy in the shallow and eutrophic Lake Søbygaard, Denmark: Chironomid–macrophyte co-occurrence. *Freshwater Biology* 46: 253–267.
- Davidson TA, Sayer CD, Bennion H, et al. (2005) A 250 year comparison of historical, macrofossil and pollen records of aquatic plants in a shallow lake. *Freshwater Biology* 50: 1671–1686.
- Digerfeldt G (1986) Studies on past lake-level fluctuations. In: Berglund BE (ed.) *Handbook of Holocene Palaeoecology and Palaeohydrology*, pp. 127–144. New York: Wiley.
- Flower RJ (2001) Change, stress, sustainability and aquatic ecosystem resilience in north African wetland lakes during the 20th century: An introduction to integrated biodiversity studies within the CASSARINA Project. *Aquatic Ecology* 35: 261–280.
- Gaillard M-J (1984a) Etude palynologique de l'évolution tardi- et postglaciaire de la végétation du Moyen-Pays romand (Suisse). *Diss. Bot.* 77.
- Gaillard M-J (1984b) *A Palaeohydrological Study of Krageholmssjön (Scania, South Sweden). Regional Vegetation History and Water-Level Changes*. Lund, Sweden: University of Lund. Lundqua Report No. 25.
- Gaillard M-J (1985) Lateglacial and Holocene environments of some ancient lakes in the western Swiss Plateau. *Diss. Bot.* 87: 273–336.
- Gaillard M-J and Digerfeldt G (1991) Palaeohydrological studies and their contribution to palaeoecological and palaeoclimatic reconstructions. *Ecological Bulletin* 41: 275–282.
- Gaillard M-J and Lemdahl G (1988) Plant macrofossil analysis (seeds and fruits) at Skateholm-Järavallen, southern Sweden—A lagoonal landscape during the Atlantic and early Subboreal time. In: Larsson L (ed.) *The Skateholm Project I. Man and Environment*, pp. 34–38. Stockholm: Almqvist & Wiksell.
- Hannon GE and Gaillard M-J (1997) The plant-macrofossil record of past lake-level changes. *Journal of Paleolimnology* 18: 15–28.
- Iversen J (1954) The Late-glacial flora of Denmark and its relation to climate and soil. *Danm. Geol. Unders* 2(80): 87–119.
- Jackson ST and Charles DF (1987) Aquatic macrophytes in Adirondack (New York) lakes: Patterns of species composition in relation to environment. *Canadian Journal of Botany* 66: 1449–1460.
- Jeppesen E, Søndergaard M, Søndergaard M, and Christoffersen K (eds.) (1998) *The Structuring Role of Submerged Macrophytes in Lakes, Ecological Studies No. 131*. New York: Springer.
- Kolstrup E (1980) Climate and stratigraphy in northwestern Europe between 30 000 BP and 13 000 BP, with special reference to The Netherlands. *Med. Rijks Geol. Dienst* 32(15): 181–253.
- Kultti S, Välranta M, Sarmaja-Korjonen K, et al. (2003) Palaeoecological evidence of changes in vegetation and climate during the Holocene in the pre-Polar Urals, northeast European Russia. *Journal of Quaternary Science* 18(6): 503–520.

- Kulti S, Oksanen P, and Väliranta M (2004) Holocene tree line, permafrost, and climate dynamics in the Nenets region, East European Arctic. *Canadian Journal of Earth Sciences* 41: 1141–1158.
- Lang G (1994) *Quartäre Vegetationsgeschichte Europas*. Jena, Germany: Fischer Verlag.
- Lotter AF and Birks HJB (2003) The Holocene palaeolimnology of Sägistalsee and its environmental history—A synthesis. *Journal of Paleolimnology* 30: 333–342.
- Rasmussen P and Anderson NJ (2005) Natural and anthropogenic forcing of aquatic macrophyte development in a shallow Danish lake during the last 7000 years. *Journal of Biogeography* 32: 1993–2005.
- Scheffer M, Hosper SH, Meijer M-L, Moss B, and Jeppesen E (1993) Alternative equilibria in shallow lakes. *TREE* 8: 275–279.
- Smol JP (2002) *Pollution of Lakes and Rivers. A Paleoenvironmental Perspective*. London: Arnold.
- Väliranta M (2005) Plant Macrofossil Evidence of Changes in Aquatic and Terrestrial Environments in North-eastern European Russia and Finnish Lapland since Late Weichselian. PhD thesis No. 181, Department of Geology, University of Helsinki. ISSN 1795-3499, ISBN 952-10-2153-5 (PDF).
- Wasylikowa K (2005) Palaeoecology of Lake Zeribar, Iran, in the Pleniglacial, Lateglacial and Holocene, reconstructed from plant macrofossils. *The Holocene* 15: 720–735.

Rodent Middens

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Introduction

The study of Pleistocene terrestrial environments has tended to focus mainly on the middle and high latitude regions where water-lain deposits (lake sediments, peat bogs, fluvial deposits) tend to be well preserved. By their very nature, most desert and semi-arid regions offer few if any archives of waterlain sediments for such studies. However, an important, if unexpected archive of Pleistocene desert environments does exist, in the form of rodent middens. Rodent middens are produced by *Neotoma* (packrats) and *Erethizon* (porcupine) in North America; *Lagidium* (viscacha), *Phyllotis* (leaf-eared mice), *Abrocoma* (chinchilla rats) and *Octodontomys* (mountain degu) in South America; *Procavia* (rock hyrax) and *Petromus* (Dassie rat) in Africa and the Middle East; *Alticola* (mountain voles) and *Ochotona* (pikas) in central Asia, and *Leporillus* (stick-nest rats) in Australia. The studies of South American middens have mostly concerned pollen analysis, although a few plant macrofossil records have been developed (Betancourt et al., 2000; Latorre et al., 2002). The studies of hyrax middens from the Middle East and desert regions of Africa have focused mainly on pollen (Scott and Vogel, 2000). The few studied Stick-nest rat middens in Australia are all of late Holocene age (Pearson, 1999). Little research has been done on porcupine middens in North America, although these middens are known to persist as long as 25 000 years in arid conditions (Betancourt et al., 1986). The middens of North American packrats have yielded by far the most well-studied records of late Pleistocene plant macrofossils. This article focuses primarily on this remarkable record.

North American Packrat Middens

Packrats, or woodrats, are the native North American rats of the genus *Neotoma* (Figure 1). There are 21 species in this genus. These rats are broadly distributed, ranging from Arctic Canada to Nicaragua. They are a very successful group, and only a few species inhabit desert and semi-arid regions. While packrats are not strict herbivores, plants are by far the most important part of their diets. Packrats live in dens. Regardless of size, each den is occupied by one individual, or the female and her young. These dens range from piles of sticks under shrubs or trees to more elaborate shelters in caves and rock shelters. In the arid Southwest, these dens often include tightly-packed bundles of cactus stems that deter predators. Besides bedding material and food plants, packrats also accumulate other objects from the surrounding landscape. These objects, plus packrat feces, are cemented together by successive layers of dried rat urine to form a midden (Figure 2). The crystalized urine layers encapsulate plant materials and rat feces to form a hardened mass,

called amberat. When protected from direct rainfall and high humidity, amberat is extremely resistant to decomposition. It secures biological specimens (plant parts, bones, insects, etc.) into a kind of desiccated time capsule, documenting the ancient life of the region. Middens of late Holocene age have been found throughout the western United States and southwestern Canada (Hebda et al., 1990). In the arid regions of the southwestern United States and northern Mexico, many packrat middens have remained intact beyond the limit of the radiocarbon dating method (>45 000 years).

Packrat ecology has been described by Vaughan (1990). Packrats require succulent plants for food, and they must have shelter, as they do not have the metabolic mechanisms for water retention found in some other rodents. The dens are often built in rock shelters, crevices, and caves, where there is a microclimate that is significantly buffered from ambient climatic conditions. The annual variation of temperatures in packrat dens from mid- to high latitudes is as little as 15 °C. Packrat middens accumulate as long as a den is occupied. In sheltered localities, these middens may accumulate for millennia, as favorable denning sites are inhabited by succeeding generations of packrats. Packrat middens are an unusual source of fossils, because the packrat is the primary agent that accumulates the materials found in middens. They eat a wide variety of plants, and the unconsumed pieces of these plants eventually form the bulk of the plant macrofossils in midden records. Because of the rodent's food preferences, plant macrofossil records from middens do not represent the full spectrum of regional vegetation; nevertheless, the variety of plant species found in these assemblages can be truly remarkable. Dial and Czaplewski (1990) discussed the problems of potential bias in packrat midden plant macrofossil assemblages. Their study concluded that if several middens are studied from a given region, then the contemporaneous plant species diversity of that region may be fairly represented in the suite of midden samples. A study of paired samples of modern vegetation communities and paired samples of ancient communities was used to investigate the probability that the absence of a plant taxon from a packrat midden implies absence from the ancient plant community where the packrat lived (Nowak et al., 2000). The probability of the occurrence of a false inference was calculated to be from 7–11%, depending on the taxon. Lyford et al. (2004) studied the contents of 59 modern packrat (*Neotoma cinerea*) middens in escarpments and canyons in three states in the Rocky Mountain region. They found that these middens are highly reliable sensors for the presence of key woody plant species, such as *Juniperus osteosperma* (Utah juniper) and *J. scopulorum* (Rocky Mountain juniper), but that other conifers were occasionally absent from modern middens if their presence in the local vegetation cover was less than 20% of the canopy cover.



Figure 1 Photograph of a packrat (*Neotoma*), the principle agent of plant materials that become incorporated into packrat middens (photo courtesy of Corel Corporation).



Figure 2 Packrat (*Neotoma*) midden sample from a rock shelter at Emery Falls, Grand Canyon, Arizona. Photo by Thomas Van Devender. The shiny surface is amberat (dried packrat urine). Note fossil conifer needles that have weathered out from the broken edge of the midden.

How Middens Are Analyzed

Packrat midden analysis methods have been described by Spaulding et al. (1990). Most ancient packrat middens are indurated (hardened almost to the consistency of rock), and resemble blocks of asphalt. Amberat forms a tough, outer rind that resists erosion and decomposition (Figure 1). When moist, amberat is sticky. It is this stickiness that often cements middens to the walls of caves or rock shelters. Midden samples are taken by the removal of discrete horizontal sections of midden, often using hammers and chisels. Each sample is radiocarbon-dated, because successive layers in a midden may have been deposited at widely-spaced time intervals, or occasionally in random order. Radiocarbon dates are often obtained from packrat fecal pellets, or from plant macrofossils. The majority of middens are of Holocene and late Wisconsin age (Figure 3).

Following rehydration in a laboratory, the samples are wet-screened to remove fine sediment, then dried and sorted under

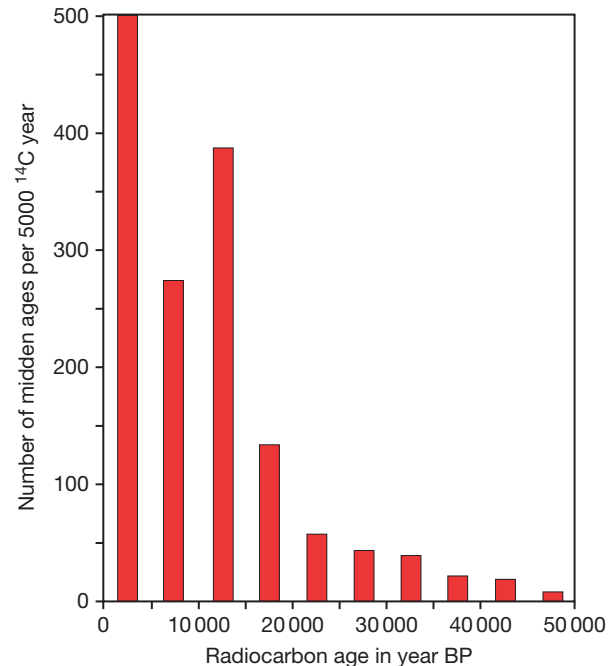


Figure 3 Distribution of radiocarbon ages of packrat middens from the southwestern United States and northern Mexico, after Thompson et al., 2004.

a low-power microscope. The plant macrofossils are picked out and sorted from this dried matrix. Macrofossils from individual species of plants are sometimes submitted for AMS radiocarbon dating (see AMS Method).

Plant macrofossils from middens are frequently identified to the species level. Midden plant macrofossil analysis is generally done on a presence/absence basis for each identified taxon. Occasionally, a subjective quantification of the remains is conducted, either through application of a relative abundance scale (rare, common, etc.), or through specimen counts and dry weights of materials. The results of plant macrofossil analyses are often tabulated in terms of relative species abundance. They are separated into categories such as rare, common, abundant, and very abundant.

Distribution of Fossil Sites in the Arid Southwest

Figures 4 through to 7 illustrate published packrat midden records from southwestern North America. Dozens of sites have been studied in each of the major desert regions discussed below, and more than a thousand radiocarbon dates have been obtained from regional midden samples. Most of the midden samples discussed here have been taken from canyons and rocky hillsides, just above the desert floor. In the American Southwest, no Pleistocene middens have been sampled at elevations greater than 2500 m. The elevational distribution of midden samples reflects regional physiography. Sites south of 36° latitude generally occur at elevations below 1500 m, while more northerly sites reflect a broader range of elevations.

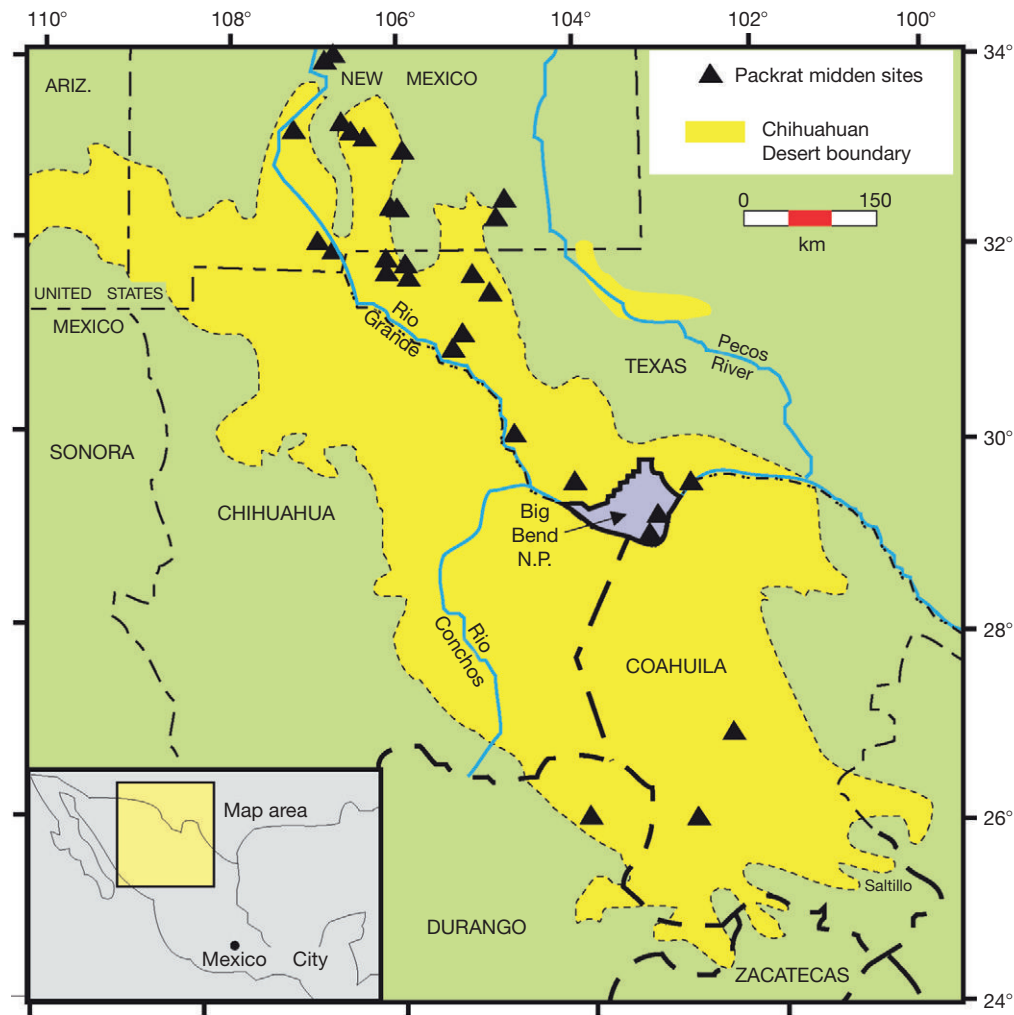


Figure 4 Map of the Chihuahuan Desert region, showing location of packrat middens from which fossil analyses have been published (triangles).

Regional Studies

Chihuahuan Desert

The Chihuahuan Desert is an interior continental desert, stretching from southeastern Arizona in the northwest to northern Zacatecas, Mexico in the southeast (Figure 4). The desert lies in the rainshadow of the Sierra Madre Occidental to the west and the Sierra Madre Oriental to the east. Sub-freezing temperatures are not uncommon in winter, and Arctic air mass incursions may bring very cold weather to even the southernmost parts of the desert. Today the southern Chihuahuan Desert supports subtropical vegetation, rich in succulent scrub species. This flora diminishes to the north, giving way to more cold-tolerant desert grassland. The plant macrofossil record developed from Chihuahuan Desert packrat middens reveals a series of environmental changes through the late Pleistocene and the Holocene. These changes brought about major shifts in plant distributions, both within the desert (i.e., elevational shifts) and beyond its boundaries (i.e., regional extirpations and the invasion of new species). The plant macrofossil samples studied from this desert range in age

from >43 300 years BP to recent. Regional plant macrofossil records are discussed by Van Devender (1990a).

In the northern Chihuahuan desert, pinyon-juniper-oak woodland dominated limestone slopes during the middle to late Wisconsin interval (42 000 to 12 750 cal years BP). At the northern edge of the desert (Rough Canyon, New Mexico), Rocky Mountain juniper (*J. scopulorum*) began to decline at 14 000 cal years BP, as Holocene warming started and it was regionally extirpated by 12 400 cal years BP (Betancourt et al., 2001). An early Holocene oak-juniper woodland was replaced by desert-grassland after about 9250 years BP. At Rough Canyon, elements of the modern desert vegetation began dominating packrat midden floras after 5750 years BP (Betancourt et al., 2001). By about 4700 cal years BP, the common elements of the Chihuahuan Desert flora had become established throughout the northern Chihuahuan Desert region, forming a relatively modern desert-scrub vegetation on rocky slopes. This vegetation continues to be regionally dominant today.

In the central Chihuahuan Desert, in and around Big Bend National Park, Texas, the middle Wisconsin vegetation (40 000–26 500 years BP) was a pinyon-juniper-oak

woodland on the uplands, and juniper grassland on the lowlands. By late Wisconsin times (prior to 25 000 years BP), papershell pinyon (*Pinus remota*) became important in the upland flora, and pinyon-juniper-oak woodland spread into lowland regions, indicating relatively mild winters, cool summers, and greater precipitation than today. The boundary between late Wisconsin and early Holocene environments (between 13 100 and 12 100 cal years BP) is well marked in the vegetation, with the demise of papershell pinyon and increase of shrub oak (*Quercus pungens*) before 9200 cal years BP, followed by a succulent desert-scrub vegetation, including lechuguilla (*Agave lechuguilla*), catclaw acacia (*Acacia greggii*), honey mesquite (*Prosopis glandulosa*), and prickly pear cactus (*Opuntia* spp.). Vegetation diversity increased in the mid-Holocene, signaling the shift to subtropical climatic conditions. More mesic vegetation communities developed here in the last 1000 years. By the middle Holocene, the vegetation in the lowlands of Big Bend was essentially modern desert-scrub.

In the southern Chihuahuan Desert of Mexico, the late Wisconsin vegetation on limestone slopes was a woodland dominated by juniper and papershell pinyon in association with succulents such as sotol (*Dasyllirion*) and beaked yucca (*Yucca rostrata*). Pinyon pine declined after 16 000 cal years BP, as succulents increased. Between 13 600 and 10 550 cal years BP the regional vegetation shifted to Chihuahuan desert-scrub. The paleobotanical record from this region shows unique combinations of floristic elements. Late Glacial floras included both conifers and desert succulents, but macrofossils of temperate plants disappeared from regional records during the middle and late Holocene. A mosaic of temperate and xeric habitats has been available to the regional biota, even during the last 1000 years, when other regions of the Chihuahuan Desert have experienced extremes of aridity.

Sonoran Desert

This desert comprises the arid, subtropical region centered around the head of the Gulf of California (Figure 5), encompassing parts of Sonora and Baja California in Mexico, and southern Arizona and California in the United States. All parts of the Sonoran Desert are occasionally subject to winter freezes, although rarely for more than one night at a time. Most of the large, columnar cacti that characterize the Sonoran Desert are not able to tolerate prolonged freezing. Rainfall ranges from a bi-seasonal pattern with summer monsoon moisture in Sonora and Arizona to a winter rainfall pattern along the west coast of Baja California. Changes in large-scale atmospheric circulation patterns during the Wisconsin glaciation caused changes in Sonoran Desert regional climates, especially in precipitation patterns and the frequency of winter freezes.

Unlike the Chihuahuan Desert, the plant macrofossil record from Sonoran Desert packrat middens indicates that regional vegetation succession from the late Wisconsin through the Holocene made a single, continuous progression toward modern plant communities. The regional paleobotanical record has been summarized by Van Devender (1990b). The chronological sequence of plant communities during the last 20 000 years is similar to the modern vegetational gradient from the tops of mountains to valley floors. This history is summarized in Table 1. Central and southern Arizona were clothed with a variety of woodlands in the late Wisconsin. In upland regions, juniper or juniper-oak woodlands persisted until the early Holocene. Juniper died out in the Sonoran Desert after 10 000 cal years BP. The vegetation of lowland Sonoran regions has been desert-scrub since the early Holocene. Desert-scrub communities, including species found today in the Mojave Desert, persisted in the harsh lowlands of the lower Colorado River valley throughout the late Pleistocene. It is thought that this region may have served as a refuge for North American desert plants for much of the Quaternary. Typical Mojave Desert species migrated north along the Colorado River and crossed over it only during the warmest (interglacial) periods. The ecotone (ecological boundary) between the Sonoran and Mojave Deserts has essentially remained in its current position since 10 000 cal years BP. After 4500 cal years BP, modern vegetation became established throughout the Sonoran Desert.

Based on the plant macrofossil evidence, late Wisconsin climate in the Sonoran Desert had cooler-than-modern summers and a dominance of winter precipitation. While winters may have been generally cooler, it appears that freezing conditions were actually rarer than today, a remarkable finding, given the degree of climatic cooling associated with North American glacial climates in regions further north. Early Holocene Sonoran climates were transitional between late Wisconsin and modern climates. Regional summer temperatures probably remained cooler, based on the persistence of conifers in the uplands. Parts of south-central Arizona received up to 30% more precipitation than today in the early Holocene. A relatively modern climate was established after 10 000 cal years BP, characterized by hot summers. Mid-Holocene plant macrofossil assemblages indicate greater species richness than today, probably because of increased rainfall. While the packrat midden

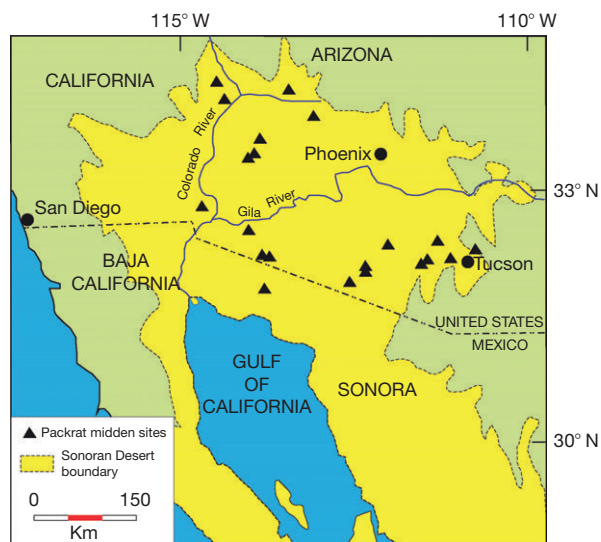


Figure 5 Map of the Sonoran Desert region, showing location of packrat middens from which fossil analyses have been published (triangles).

Table 1 Summary of Changes of Sonoran Desert Vegetation Inferred from Packrat Midden Data (after Elias, 1995)*Region time interval vegetation***Central Arizona**

Uplands late Wisconsin Pinyon-juniper-oak woodland (550–1550 m)

Early Holocene Juniper-oak woodland with Sonoran desert-scrub

Late Holocene Sonoran desert-scrub with increasing subtropical elements

Lowlands late Wisconsin Juniper woodlands and/or Mohave desert-scrub (<600 m)

Early Holocene onwards Sonoran desert-scrub

Southern Arizona

Uplands late Wisconsin Pinyon-oak-juniper woodland with Joshua tree (550–1550 m)

After 13 000 cal years BP Xeric juniper woodland with shrub oak or chaparral

After 10 000 cal years BP Juniper disappears from entire region

Lowlands late Wisconsin Xeric juniper woodland with Joshua tree, yucca, beargrass (600–300 m)

Early Holocene desert-scrub common

After 4500 cal years BP Modern Sonoran desert-scrub

Baja California

Lowlands early Holocene abundant Boojam tree; buckwheats and Mormon tea (0–400 m) in desert-scrub community

Late Holocene Boojam tree less abundant; buckwheats and Mormon tea disappear; increasing succulents

Table 2 Summary of Changes of Mojave Desert Vegetation Inferred From Packrat Midden Data (after Elias, 1995)*Region time interval vegetation***Southern and Central Mojave**

Lowlands late Wisconsin Juniper woodland (<1000 m)

Early Holocene desert-scrub widespread by 11 500 cal years BP; succulents and grasses increasing

Uplands late Wisconsin Juniper-pinyon woodland (1000–1800 m)

Early Holocene Woodland persists at higher elevations and moist slopes

Mid-Holocene desert-scrub rich in species

Late Holocene downward shift of blackbrush desert-scrub

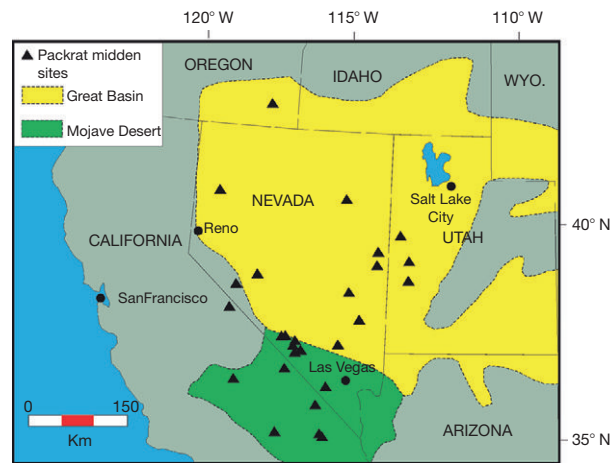
Northern Mojave

Uplands late Wisconsin desert-scrub in dry locations, with Mormon tea, shadscale, (1000–1800 m) rubber rabbitbrush, and snowberry; juniper woodland in most regions; at higher elevations, limber pine in drier sites and white fir in mesic sites

Early Holocene Woodland persists on some slopes; after 8700 cal years BP, then replaced by creosote bush and Joshua tree scrub

Mid-Holocene desert-scrub lacking in diversity

Late Holocene rich desert-scrub communities at lower elevations and pinyon-juniper woodlands at higher elevations

**Figure 6** Map of the Mojave Desert and Great Basin regions, showing location of packrat middens from which fossil analyses have been published (triangles).

results are internally consistent throughout the Sonoran Desert, none of the regional paleoclimatic reconstructions based on these fossil data have supported the results of general circulation model simulations for the early and middle Holocene (Van Devender et al., 1994). By 4500 cal years BP, a completely modern climatic regime was established in the Sonoran Desert.

Mojave Desert

The Mojave is a small, topographically complex desert, situated in southern Nevada and adjacent California and Arizona

(Figure 6). The modern vegetation of the southern Mojave Desert (south of 36° N latitude) is dominated by creosote bush (*Larrea divaricata*), white bursage (*Ambrosia dumosa*), and pygmy-cedar (*Peucephyllum schottii*). The northern Mojave region is generally higher in elevation, and the modern vegetation here is a temperate desert-scrub, including blackbrush (*Coleogyne ramosissima*), Joshua tree (*Yucca brevifolia*), ground-thorn (*Menodora spinescens*), boxthorn (*Lycium* spp.), and goldenbushes (*Haplopappus* spp.). The higher mountains in southern Nevada support pinyon-juniper woodlands above 1800 m elevation, giving way to fir-pine forest above 2200 m. Bristlecone pine (*Pinus longaeva*) dominates subalpine landscapes above 2700 m.

The late Wisconsin and Holocene vegetation history of the Mojave Desert has been discussed by Spaulding (1990, 1995), and is summarized in Table 2. The northern Mojave region has yielded more packrat midden studies than the southern region. During the late Wisconsin period, woodlands extended down-slope on to the Mojave lowlands. These were 'pygmy' woodlands, composed primarily of juniper (mixed with pinyon pine at elevations below 1500 m), shrub oak, and Joshua tree in some regions. During the Last Glacial Maximum in the northern Mojave, elevations above 1600 m had populations of limber pine (*Pinus flexilis*), replaced at times by white fir (*Abies concolor*). The lower reaches of the Grand Canyon lie in the southeastern corner of the Mojave Desert. During the late Wisconsin, this region supported more succulent species, including barrel cacti and yuccas.

By 11 600 cal years BP, a major shift in Mojave vegetation was underway: desert-scrub vegetation spread across the lowlands, at the expense of semi-arid woodland. However, the transition from glacial to postglacial vegetation was not smooth; it was marked by multiple incursions of new taxa and extirpations of others. Not all of the newly invading species remained throughout the Holocene. The immigration of warm-adapted species from their southern refugia took place in stages. On the other hand, widespread desert vegetation

apparently developed earlier in the Mojave than in the Sonoran or Chihuahuan deserts, with the earliest occurrences documented from middens dating to 18 000 cal years BP. By 8600 years BP, woodlands had been replaced by desert vegetation throughout the lower elevations of the Mojave. An interval of increased aridity is marked by vegetation changes in most regions of the Mojave during the mid-Holocene, but regional packrat midden evidence suggests that effective moisture subsequently increased between 4200 and 1400 cal years BP. This increase was probably due to a combination of increased precipitation and decreased temperatures. Conifers retreated up mountain slopes in the Mojave region during the last 1400–500 years, signaling a return to hotter, drier conditions.

Great Basin

The Great Basin is a large, arid region between the Rocky Mountains and the Sierra Nevada (Figure 6). This enormous, closed basin is made up of broad valleys of sagebrush steppe, dissected by more than one hundred narrow mountain chains that trend north to northeast. During the last glaciation, immense lakes covered the Great Basin lowlands, and glaciers flowed down from high mountain valleys. In the modern vegetation, sagebrush and shadscale (*Atriplex confertifolia*) dominate the low elevation steppe vegetation. Pinyon-juniper woodland covers most low mountain slopes in the southern two-thirds of the region, while juniper woodland characterizes the northern third. Montane forests of ponderosa pine (*Pinus ponderosa*), white fir, and Douglas fir grow in the mountains of the eastern and southern sectors. The montane zone in the central and northern Great Basin is covered by sagebrush grassland. Some isolated mountain ranges are covered by mountain mahogany (*Cercocarpus*) woodland. Sub-Alpine forests of bristlecone pine, limber pine, and Englemann spruce (*Picea engelmannii*) occur in the higher mountain ranges, and small pockets of Alpine tundra grow on isolated peaks.

The late Pleistocene and Holocene vegetational history of the Great Basin has been discussed by Thompson (1990), and is summarized in Table 3 (see Western North America). Many plant species that lived in the Great Basin during the Pleistocene have their northern limits today in the Mojave Desert; other plants grew as much as 1000 m downslope from their modern ranges. Unlike some other deserts, there is no evidence that extra-limital plant species immigrated into the Great Basin during the late Pleistocene. Instead, regional floras suffered a reduction in species diversity, as many of the modern taxa were absent.

The packrat midden records show that sub-Alpine conifers, especially limber pine, expanded their ranges downslope by as much as 1000 m during the late Wisconsin interval. Curiously, the modern dominants of the montane zone (spruce and fir) were absent from the region during much of the late Pleistocene. Many persisted south of the Great Basin, in southern Nevada and southeastern California. In the western Great Basin, juniper dominated lower elevation woodlands in areas adjacent to pluvial lakes. Although Utah juniper (*Juniperus osteosperma*) was likely the dominant species at these sites, possible hybrids of western juniper (*Juniperus occidentalis*), Rocky Mountain juniper (*Juniperus scopulorum*) and California juniper (*Juniperus californica*) have been found in some

Table 3 Summary of Changes in Great Basin Vegetation Inferred From Packrat Midden Data (after Elias, 1995)

Time interval vegetation

Late Wisconsin (26 000–17 000 cal years BP)

Uplands:

Sub-Alpine conifers grow as much as 1000 m downslope from modern limits; dominant trees: bristlecone pine, limber pine, prostrate juniper, and Englemann spruce in some locations; many shrubs and trees common today are absent, including Douglas fir, white fir, and Rocky Mountain juniper.

These shift south, into the Mojave Desert and elsewhere

Steppe regions:

Sagebrush and other steppe species present across entire range of elevations in Great Basin, up to sub-Alpine (as ground cover under coniferous forests).

Steppe species include sagebrush, rabbitbrush, horsebrush, snake-weed, and snowberry steppe species spread south into Mojave and Sonoran deserts, and other regions.

Late Glacial (17 000–11 500 cal years BP)

Limber pine, prostrate juniper, Englemann spruce, and some shrubs decline from 16 000–13 000 cal years BP, replaced by snake-weed, rock spiraea Bristlecone and limber pine persist at late Wisconsin levels until after 11 500 cal years BP

Mesic-adapted plants at many elevations replaced by thermophiles starting at 16 000 cal years BP In southern Great Basin, a major shift to desert vegetation before 13 000 cal years BP

Early Holocene (11 500–7800 cal years BP)

Bristlecone and limber pines absent from western Utah and eastern Nevada by 9600 cal years BP. Limber pine and Douglas fir replaced by Utah juniper and Gambel oak in southeastern Nevada by 10 000 cal years BP.

Mid-Holocene (7800–4500 cal years BP)

Rapid transition to modern vegetation zones between 7800 and 6900 cal years BP. Pinyon-juniper woodland established in several mountain ranges, characterized by Utah juniper and single-needle pinyon pine. Midden assemblages include a number of species that are rare or absent from Great Basin today.

Late Holocene (4500 cal years BP–present)

Few changes in regional vegetation. Ranges of sagebrush and shadscale steppe contract; upper treeline retreats to mid-Holocene levels Lower treeline drops about 170 m between 4500 and 2000 cal years BP, then retreats to modern levels during last 2000 years

middens. Sagebrush and other steppe species expanded their ranges in the late Wisconsin, both in latitude and elevation. They spread south to the Mojave and Sonoran deserts, and into adjacent mountain ranges. The drier regions of the northern Great Basin were dominated by shadscale.

The downslope extension of bristlecone and limber pines might have produced a contiguous forest zone stretching from the eastern shore of pluvial Lake Lahontan to the Rockies, but edaphic factors probably prevented conifer establishment in some regions. These vast coniferous woodlands became fragmented as lower treeline advanced upslope in response to climatic warming at the end of the last glaciation. The shift from Pleistocene to Holocene vegetation regimes took place in different ways at different sites. In the boundary region between the Mojave Desert and Great Basin, the major vegetation change took place by 13 000 cal years BP. In the north, this shift was completed between 10 000 and 7800 cal years BP.

The Great Basin has a large number of endemic grasses and other herbs. The packrat midden record suggests that this is due to a combination of factors. Range fragmentation led to genetic isolation and rapid evolution, at least at the subspecific level. Also, new habitats opened up as the huge pluvial lakes receded and climates changed at the end of the last glaciation. During

the Holocene, semi-arid lowlands have acted as biotic barriers, much as pluvial lakes did during the Pleistocene. Lyford et al. (2002) extended Holocene midden research into the northern edge of the Great Basin in southern Montana. They were able to document the timing of a regional shift from cooler and moister to hotter and drier climates, based on the replacement of boreal and montane juniper species by the Great Basin juniper, *J. osteosperma*. This replacement occurred by 4700 ^{14}C years BP, and was followed by another cool, moist interval (4400–2700 ^{14}C years BP) during which the lower treeline shifted downslope once again.

The Colorado Plateau

This plateau is a well-marked physiographic province that covers nearly 400 000 km² of the Four Corners region (Figure 7). Plateau elevations range from 1000 to 4600 m, with much of the plateau below 1800 m. Topography strongly affects regional climates, now and in the late Pleistocene. Modern vegetation zones are shown in Figure 8. These zones do not fall in simple elevation bands, because local bedrock type strongly affects the vegetation.

More than 180 packrat middens have been analyzed from sites on the Colorado Plateau. The middens occur only in regions with sandstone and limestone bedrock. The plant macrofossil record of the region has been discussed by Betancourt (1990) and is summarized in Figure 8. During the late Wisconsin, upper treeline on the Colorado Plateau was significantly depressed, and Alpine tundra regions may have supported more sage cover than is found there today. Today Alpine tundra is found in only three high mountain regions, at elevations above 3660 m. Late Glacial coniferous woodlands extended down to 2000 m elevation in

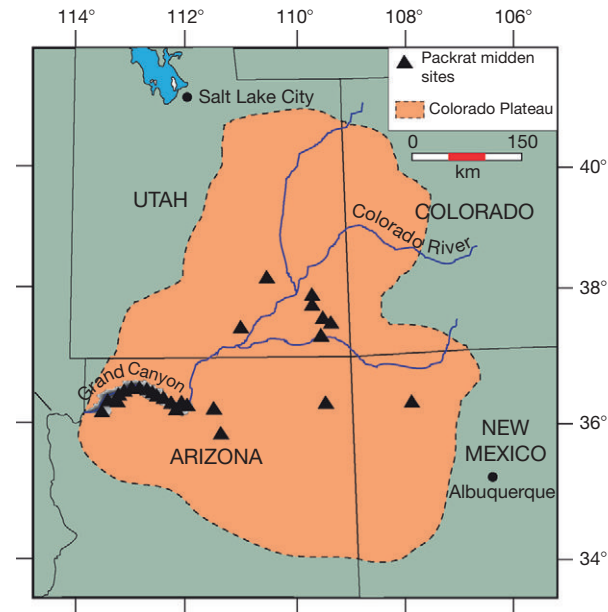


Figure 7 Map of the Colorado Plateau region, showing location of packrat middens from which fossil analyses have been published (triangles).

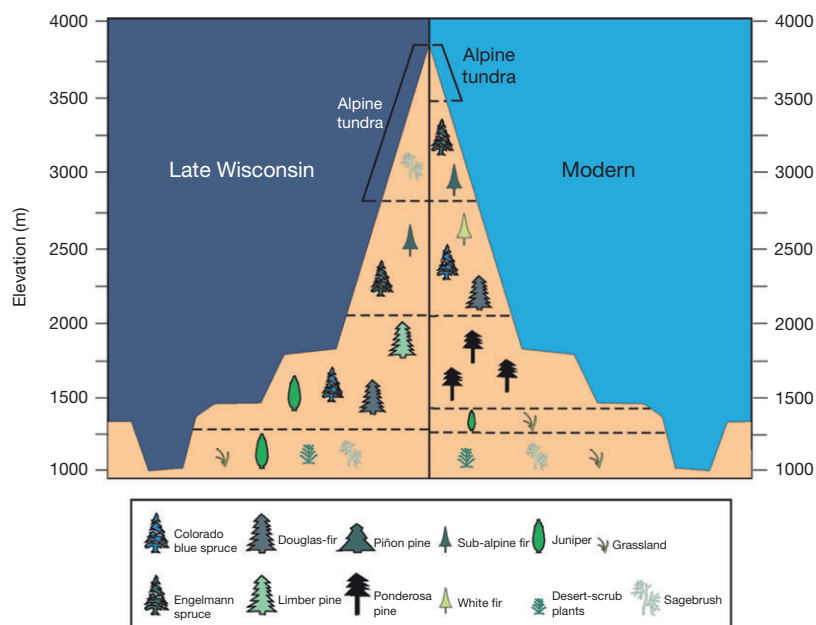


Figure 8 Diagrammatic drawing showing principal vegetation zones of the Colorado Plateau during late Wisconsin and modern times, after Betancourt, 1990.

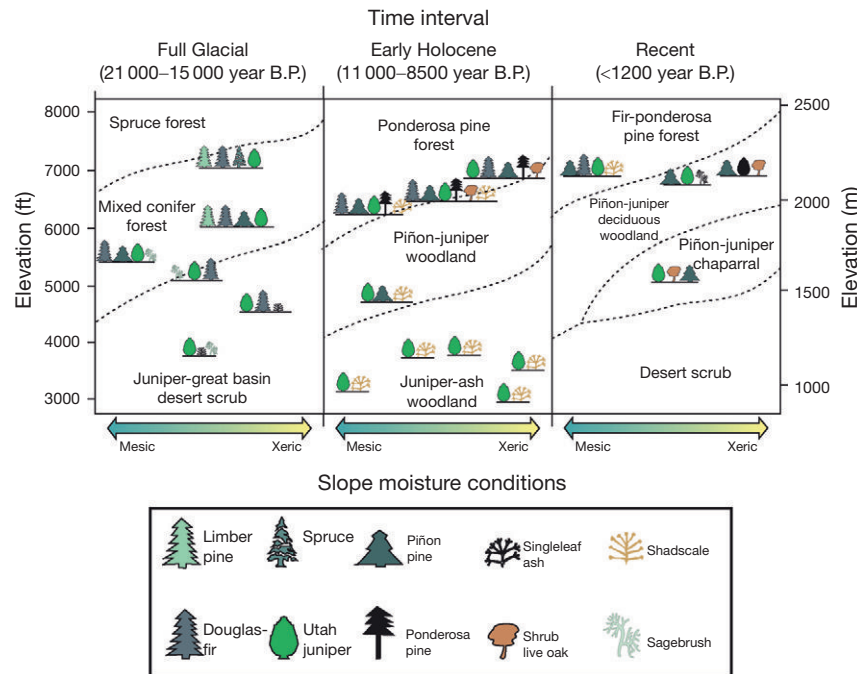


Figure 9 Diagrammatic drawing showing principal vegetation zones of the Grand Canyon during Full Glacial, early Holocene, and modern times, after Cole, 1990b.

mesic canyon habitats. In the San Juan Mountains and Glen Canyon, coniferous woodlands with limber pine, Douglas fir (*Pseudotsuga menziesii*), junipers, and Colorado blue spruce (*Picea pungens*) grew down to 1300 m elevation on rocky sites. Regional lowlands (<1300 m elevation) were covered in desert-scrub with junipers and sagebrush.

Plant macrofossil records from the Grand Canyon indicate that the ranges of regional plant species are now 700–900 m higher and 400–700 km further upstream than they were in the late Wisconsin interval (Cole, 1990a) (Figure 9). Cole's (1985) work on the paleobotany of the Grand Canyon also suggested that plant migration lagged significantly behind climate change at the end of the last glaciation. For instance, the upslope immigration of some species to their modern zones was not completed until the mid-Holocene; some plants, such as ocotillo (*Fouquieria splendens*), may still be adjusting to past climate fluctuations. These assertions sparked a heated debate among paleobotanists (Markgraf, 1986; Cole, 1986). The unexpected associations of species in early Holocene plant macrofossil assemblages may be partly due to unique climatic conditions, involving the dynamics of monsoonal moisture and shifts in the timing of summer warmth. Betancourt (1990) suggested that the length and difficulty of the various migration corridors leading to the Grand Canyon may have played a significant role in the postglacial establishment of new species. Cole (1990a) concluded that the biological phenomena, rather than unusual climatic conditions, are the cause of the peculiar plant associations of the Grand Canyon in the early Holocene. In his view, 'as the complexities of plant population dynamics in the Grand Canyon become better understood, the biologically-oriented theories will gain ground.'

Major Implications of Packrat Midden Studies

The plant macrofossil record is providing valuable data on past and present ecosystems of the arid southwest. The fossil record shows that environmental change during the Pleistocene was substantial, even in regions such as the arid southwest that were far-removed from continental ice sheets. The midden record offers good evidence for individualistic responses to major climate changes that brought about uneven displacement of species, and anomalous associations in which some species do not appear to belong in the communities where they were found.

The age and origin of southwestern desert biological communities are important questions in the field of biogeography. About 1200 of the 5500 species of southwestern plants are considered endemic to that region, a far higher proportion than elsewhere in North America. Even though many of these plants evolved locally, the packrat midden data suggest that desert plant communities are very recent and very dynamic. One of the reasons the midden data are so important is that they offer tangible evidence of the history of species in this region. For instance, based on modern data, creosote bush was thought by some to have immigrated from South America during the Holocene. However, the fossil evidence shows that it grew at least as far north as the Arizona-Sonora border during the middle of the last glaciation. We are not currently able to verify the exact whereabouts of southern refugia for the desert biota in Mexico, because many regions hypothesized to have been refugia are now in the tropical or subtropical zones, where preservation of middens is extremely limited.

Many intriguing questions remain to be addressed by the plant macrofossil record. For instance, the late Wisconsin pygmy conifer woodlands that covered much of the lower elevations of the American Southwest remained intact until after 11 000 years. Was this long-term ecosystem stability a product of climatic stability across a wide region? What brought about the demise of the lowland woodlands? One way of approaching these questions is the analysis of stable isotopes from the cellulose preserved in woody plant macrofossils. In particular, analysis of changes in the concentration of the deuterium may help to clarify the climate regimes under which these plants grew, and to define the climatic changes that forced their retreat from the desert lowlands. Another difficult question concerns a major switch in pinyon pines at the end of the Pleistocene. Colorado pinyon (*Pinus edulis*) and single-needle pinyon (*Pinus monophylla*) became the dominant pinyon pine species in the Holocene, at the expense of papershell pinyon and another single-needle pinyon, *Pinus californiarum* var. *fallax*, that were dominant in the late Pleistocene. Why this reversal of roles? One approach to the problem of population dynamics of pinyon pines is through the study of ancient DNA. Gene amplification techniques are being applied to pinyon wood, cones, and needles preserved in middens, to unravel the genetic relationships between ancient populations and their modern descendants (Betancourt et al, 1991). Modern pinyon pine ranges are far from static. Both single-needle pinyon and Colorado pinyon are still migrating. The northernmost population of Colorado pinyon became established in north-central Colorado only 400 years ago. The same is true of single-needle pinyon, which arrived at its current northwestern limit near Reno, Nevada within the last 700 years.

Summary

The fossil record developed from packrat middens in the arid southwest is a unique resource for investigating the pace and direction of biotic changes. Indeed, without the packrat midden data, we would know little about the biotic history of this region where lake sediments are scarce and peat bogs almost non-existent. The midden data have shown that late Pleistocene landscapes of the southwest were radically different from today, with coniferous woodlands covering much of the lowland regions that now support only desert-scrub. This change was not just a downslope shift of intact communities; it brought together new associations of plants and animals in an ever-changing dynamic that is still at work today (Van Devender, 1985). Arid-land biotic communities viewed as 'stable' from our short-term perspective are actually in a constant state of flux. Each new environmental change brings a different cast of characters to the biological stage. In some regions, like the Great Basin, the same players appear in succeeding intervals. In other regions, such as the Sonoran and Chihuahuan deserts, new players appear from time to time, often at the expense of others. The fossil record has much to say to both the paleontological and modern ecological research communities; we are just starting to come to grips with its message.

See also: Plant Macrofossil Introduction. **Pollen Methods and Studies:** Reconstructing Past Biodiversity Development. **Radiocarbon Dating:** AMS Radiocarbon Dating.

References

- Betancourt JL (1990) Late Quaternary Biogeography of the Colorado Plateau. In: Betancourt JL, Van Devender TR, and Martin PS (eds.) *Packrat Middens, the Last 40 000 Years of Biotic Change*, pp. 259–292. University of Arizona Press.
- Betancourt JL, Milton JR, and Anderson RS (1991) Fossil and genetic history of a pinyon pine (*Pinus edulis*) isolate. *Ecology* 72: 1685–1697.
- Betancourt JL, Rylander KA, Peñalba C, and McVickar JL (2001) Late Quaternary vegetation history of Rough Canyon, south-central New Mexico, USA. *Palaeogeography, Palaeoclimatology* 165: 71–95.
- Betancourt JL, Van Devender TR, and Rose M (1986) Comparison of plant macrofossils in woodrat (*Neotoma* sp) and porcupine (*Erethizon dorsatum*) middens from the western United States. *Journal of Mammalogy* 67: 266–273.
- Cole KL (1986) In defense of inertia. *American Nature* 127: 727–728.
- Cole KL (1990a) Reconstruction of past desert vegetation along the Colorado River using packrat middens. *Palaeogeography, Palaeoclimatology* 76: 349–366.
- Cole KL (1990b) Late Quaternary Vegetation Gradients Through the Grand Canyon. In: Betancourt JL, Van Devender TR, and Martin PS (eds.) *Packrat Middens, the Last 40 000 Years of Biotic*, pp. 240–258. University of Arizona Press.
- Dial KP and Czaplewski NJ (1990) In: Betancourt JL, Van Devender TR, and Martin PS (eds.) *Do Woodrat Middens Accurately Represent the Animal's Environments and Diets? The Woodhouse Mesa Study. Packrat Middens, the Last 40 000 Years of Biotic Change*. pp. 43–58. University of Arizona Press.
- Elias SA (1995) Packrat Middens, Archives of Desert Biotic History. *Encyclopedia of Environmental Biology* 3: 19–35.
- Hebda RJ, Warner BG, and Cannings RA (1990) Pollen, plant macrofossils, and insects from fossil woodrat (*Neotoma cinerea*) middens in British Columbia. *Géog. Phys. Quatern.* 44: 227–234.
- Kuch M, Rohland N, Betancourt JL, Latorre C, Stepan S, and Poinar HN (2002) Molecular analysis of a 11 700-year-old rodent midden from the Atacama Desert. *Chilean Molecular Ecology* 11: 913–924.
- Lyford ME, Betancourt JL, and Jackson ST (2002) Holocene vegetation and climate history of the northern Bighorn Basin, southern Montana. *Quaternary Research* 58: 171–181.
- Lyford ME, Jackson ST, Gray ST, and Eddy RG (2004) Validating the use of woodrat (*Neotoma*) middens for documenting natural invasions. *Journal of Biogeography* 31: 333–342.
- Markgraf V (1986) Plant inertia reassessed. *American Nature* 127: 725–726.
- Nowak RS, Nowak CL, and Tausch RJ (2000) Probability that a fossil absent from a sample is also absent from the paleolandscape. *Quaternary Research* 54: 144–154.
- Pearson S (1999) Late Holocene biological records from the middens of stick-nest rats in the central Australian arid zone. *Quaternary International* 59: 39–46.
- Scott L, Marais E, and Brook GA (2004) Fossil hyrax dung and evidence of Late Pleistocene and Holocene vegetation types in the Namib Desert. *Journal of Quaternary Science* 19: 829–832.
- Spaulding WG (1990) In: Betancourt JL, Van Devender TR, and Martin PS (eds.) *Vegetational and Climatic Development of the Mojave Desert: The Last Glacial Maximum to the Present. Packrat Middens, the Last 40 000 Years of Biotic Change*, pp. 166–169. University of Arizona Press.
- Spaulding WG (1995) Environmental change, ecosystem responses, and the Late Quaternary development of the Mojave Desert. In: Steadman DW and Mead JL (eds.) *Late Quaternary Environments and Deep History: A Tribute to Paul S. Martin*, pp. 139–164. South Dakota: The Mammoth Site of Hot Springs.
- Spaulding WG, Betancourt JL, Croft LK, and Cole KL (1990) Packrat Middens: Their Composition and Methods of Analysis. In: Betancourt JL, Van Devender TR, and Martin PS (eds.) *Packrat Middens, the Last 40 000 Years of Biotic Change*, pp. 59–84. University of Arizona Press.
- Thompson RS (1990) Late Quaternary Vegetation and Climate in the Great Basin. In: Betancourt JL, Van Devender TR, and Martin PS (eds.) *Packrat Middens, the Last 40 000 Years of Biotic Change*, pp. 200–239. University of Arizona Press.
- Thompson RS, Shafer SL, Strickland LE, Ven de Water PK, and Anderson KH (2004) Quaternary vegetation and climate change in the western United States: Developments, perspectives, and prospects. In: Gillespie AR, Porter SC, and Atwater BF (eds.) *The Quaternary Period in the United States*, pp. 403–426. Elsevier.

- Van Devender TR (1985) Climatic cadences and the composition of Chihuahuan desert communities: the Late Pleistocene packrat record. In: Diamond J and Case TJ (eds.) *Community Ecology*, pp. 285–299. Harper and Row.
- Van Devender TR (1990a) Late Quaternary Vegetation and Climate of the Chihuahuan Desert, United States and Mexico. In: Betancourt JL, Van Devender TR, and Martin PS (eds.) *Packrat Middens, the Last 40 000 Years of Biotic Change*, pp. 104–133. University of Arizona Press.
- Van Devender TR (1990b) Late Quaternary Vegetation and Climate of the Sonoran Desert, United States and Mexico. In: Betancourt JL, Van Devender TR, and Martin PS (eds.) *Packrat Middens, the Last 40 000 Years of Biotic Change*, pp. 134–165. University of Arizona Press.
- Van Devender TR, Birgess TL, and Turner RM (1994) Paleoclimatic implications of plant remains from the Sierra Bacha, Sonora, Mexico. *Quaternary Research* 41: 99–108.
- Vaughan TA (1990) Ecology of living packrats. In: Betancourt JL, Van Devender TR, and Martin PS (eds.) *Packrat Middens, the Last 40 000 Years of Biotic Change*, pp. 14–27. University of Arizona Press.
- Webeck K and Pearson S (2005) Stick-nest rat middens and a late-Holocene record of White Range, central Australia. *Holocene* 15: 466–471.

Surface Samples, Taphonomy, Representation

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Plant macrofossils preserved in lake sediment are frequently identified and counted as a means of reconstructing the vegetational communities of a lake and surrounding areas. However, many factors affect the relationships between macrofossil assemblages and plant communities. This article explores those factors and the types of studies that clarify them.

In an ideal system, each individual plant of a species would produce a standard quantity of preservable macrofossils. (The term macrofossil is here used to indicate identifiable plant parts that are large enough to be visible without magnification and have the potential to become incorporated into lake or peatland sediment.) These macrofossils would disperse in a predictable fashion, uniformly incorporate into sediment across a lake basin, and be lost to predators and decomposition in consistent proportions, all regardless of individual basin morphology. Macrofossil assemblages could then be used to reconstruct past vegetational communities by means of precise mathematical inference models for each macrofossil type, which might appear as follows:

$$V_A = (N_A \div (P_A - L_A)) \times D_A \times b \times S$$

where V_A is the proportion of local vegetation represented by plant taxon A; N_A is the number of macrofossil type A found in standard volume of sediment; P_A is the annual production of macrofossil type A; L_A is the mean annual loss from sediment of macrofossil type A by decomposition, predation, etc.; D_A is the mean dispersal distance from source plants for macrofossil type A; b is the basin coefficient accounting for basin characteristics such as fetch, wind direction, and slope; and S is the sample location within the basin.

Alas, natural systems do not work with mathematical uniformity but are notoriously quirky and constantly changing. An additional factor could be added to the previous formula:

U = all unknown factors that paleoecologists have not even begun to anticipate.

Macrofossils do not meet the first and most important requirement for quantitative paleoenvironmental reconstruction (Birks, 1995); they are not systematically (linearly or unimodally) related to the variable (proportion of a species within a vegetational community) of interest. Thus, macrofossils are more commonly called upon as qualitative indicators of past environments. Ecological conditions may be inferred based on the known ecological tolerances of species whose remains are found as macrofossils (indicator species approach), or vegetational composition may be reconstructed based on known community associations of multiple species represented in sediment (assemblage approach) (Birks and Birks, 1980).

Surface Sample Studies

Significance

Studies of surface sediment samples are unlikely to resolve the idiosyncrasies of individual macrofossil types to the extent that precise translation into former vegetational communities is possible. However, such studies are useful to determine what factors affect macrofossil deposition; which plant taxa are over-, under-, or not represented by macrofossils; what catchment is represented by the macrofossil assemblage; what type of lake is best suited for macrofossil analysis; what part of the lake should be cored for sediment collection; and so forth. Because of biases inflicted on macrofossil assemblages by many taphonomic effects, ancient macrofossil assemblages should be compared to modern macrofossil assemblages rather than to plant communities.

Review of Surface Sample Studies

Numerous studies of surface sediment macrofossils have increased the body of knowledge concerning macrofossil taphonomy and representation of vegetational communities. H. H. Birks conducted one of the earliest surface macrofossil studies (Birks, 1973). Her work set the standard for such studies, and her ongoing analysis and discussion of macrofossil work has significantly shaped the science (Birks, 2003; Birks and Birks, 2000).

Several studies have emphasized macrofossils generated by aquatic or wetland plant species. Birks (1973) analyzed sediment samples from 32 small lakes in Minnesota and studied the relationship between macrofossils and source vegetation with respect to distance between source plants and incorporation of macrofossils into sediment. Greatrex (1983) studied seeds and fruits found in fen, fen-carr, and swamp carr sediment in Great Britain. In an attempt to develop a mathematical model to infer past water depths from macrofossil assemblages, Dieffenbacher-Krall and Halteman (2000) analyzed surface sediment samples from numerous alkaline lakes in New England. Zhao et al. (2006) performed extensive sediment and vegetation sampling of a British lake to assess spatial patchiness of plant remains in relation to community structure. This study is particularly notable for the researchers' efforts to identify leaves of aquatic plant species that typically produce little seed and also for a comparison of both macrofossil and pollen assemblages with vegetation structure.

Studies by Dunwiddie (1987) and Wainman and Mathewes (1990) examined the extent to which conifer needles found in ponds from high-relief areas of Washington state and British Columbia represented the composition of Pacific Northwest forests. These studies also compared needle deposition with the pollen record. In a study that provides an excellent analog for early postglacial environments of the Northern

Hemisphere, Glaser (1981) observed transport of plant macrofossils across frozen tundra ponds in Alaska and their post-melt-off deposition.

Several studies have focused on higher energy environments not typically chosen for Late Glacial and Holocene paleoecological studies. Holyoak (1984) conducted research on a river catchment in Spitsbergen. Davis (1985) analyzed sediment from Upper Chesapeake Bay and conducted seed settling experiments to explain the assemblages found in sediment. Spicer and Wolfe (1987) examined conifer needles, leaflets, and seeds in a high-energy environment in mountainous northern California.

Most Southern Hemisphere surface sample macrofossil studies have been conducted in Tasmania and New Zealand, where the macrofossil assemblages are typically dominated by leaves. The flora of this region includes many broad-leaved, evergreen species whose leaves have thick cuticles and are less prone to rapid decomposition than those of most broad-leaved Northern Hemisphere species. Sediment is often seed poor, and the seeds tend to be small and difficult to identify. Hill and Gibson (1986) and Drake and Burrows (1980) studied the relationships between macrofossil assemblages and vegetation composition from lakes in Tasmania and New Zealand, respectively. McQueen (1969) tested whether macrofossils in a New Zealand lake represented multiple climate zones due to downhill transport in a high-relief environment.

Methods for Surface Sample Studies

Researchers typically select sediment sampling points along a predetermined transect or transects. A transect may be parallel to wind direction, along major lake axes, or perpendicular to shore. Samples may be taken at predetermined distances from shore, at select water depths, or within distinct vegetation zones. A grab sampler (Figure 1), surface corer (Figure 2), or dredge is generally used to collect the uppermost 2–4 cm of sediment, although sediment may be scooped by hand where the water is sufficiently shallow. Limiting collection to the uppermost 2–4 cm secures only the most recent few years of sediment, although studies have collected as much as the top 10 cm.

There are several quantitative methods for sampling aquatic vegetation: (1) surveys of the entire lake bottom using transects, releve plots, and divers; (2) quadrats in different vegetative units; (3) cover abundance in the immediate vicinity of sediment collection points, using a grapple to collect vegetation where water is insufficiently clear to identify species from the surface; and (4) distance to nearest mature individual of each plant species.

Lakeside vegetation is surveyed using standard methods, such as releves along a transect, cover abundance, or tree and shrub diameters. A researcher may use aerial surveys or record all species within a certain distance of the lakeshore. Understory plants, which typically have minimal representation within macrofossil assemblages, are recorded in releves along a transect.

Comparison of macrofossils with vegetation is complex and no standard methodology emerges from the research literature. Multivariate techniques may be used to compare complex data

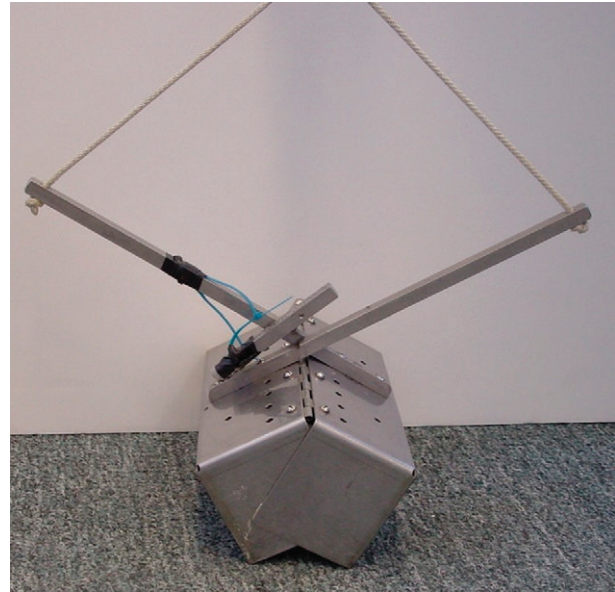


Figure 1 A grab sampling device (shown here) is lowered in the open position. As it sinks into sediment, tension on the drop line is released, closing the sampler. A grab sampler has the advantages of being inexpensive and simple to use. However, the depth of sediment collected cannot be controlled and sediment is mixed to an unknown depth. By contrast, a Birge–Ekman dredge is triggered when the operator drops a messenger down the line, releasing the spring-closing jaws. By measuring the depth to the sediment surface, the operator can lower the dredge to the proper depth before triggering the jaws to ensure collection of the sediment–water interface.



Figure 2 A surface sediment corer is a highly accurate method of collecting surface sediment. The surface–water interface is collected in a clear, polycarbonate barrel. By removing the coring device from the top of the barrel, and pushing the bottom plug up from below, the sediment can be extruded in precise increments, ensuring collection of the uppermost sediment without mixing.

sets. Percentages of representation within the macrofossil assemblage and vegetation community may be compared. Typically, researchers report which species in the vegetation are represented in the macrofossil assemblages and to what degree

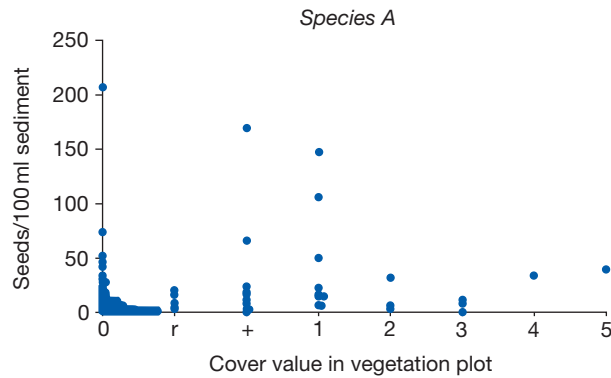


Figure 3 Sample graph comparing abundance of macrofossil type A to the presence of species A in the vegetational community. After Birks HH (1973) Modern macrofossil assemblages in lake sediment in Minnesota. In: Birks HJB and West RG (Eds.) *Quaternary Plant Ecology*, pp. 172–189. Oxford: Blackwell.

taxa are over- or underrepresented. Graphic presentation of these measurements enhances these comparisons (Birks, 1973; Figure 3). Researchers generally discuss which environments—aquatic, lakeside, or upland—are responsible for most macrofossil deposition. Studies also report the percentage of a macrofossil type found within a given distance of potential source plants.

Aquatic Seed Bank Studies

Aquatic seed bank studies provide important information about seed dispersal and sedimentation, although they are conducted for entirely different purposes than surface sediment macrofossil studies. Seed bank studies are generally conducted to determine what propagules are available for recolonization of a site or to ascertain what role the seed bank plays in maintaining established vegetation. In a seed bank study, sediment is collected and placed in a greenhouse under ‘ideal’ germination conditions. After a set amount of time, emerging seedlings are identified and counted.

Because of the different methodologies, seed bank studies are not directly comparable to surface sediment macrofossil studies, and caution must be exercised in using seed bank data for interpretation of macrofossil assemblages. Germination studies strive to measure ‘active’ components of the seed bank, those seeds or other propagules that retain the potential for growth. Only seedlings that survive to identifiable size are counted. Nonviable seeds, seed fragments, and vegetative fragments in the sediment are not represented. Germination studies are somewhat problematic for determining active components of the seed bank because greenhouse conditions may not precisely match conditions of the site from which the sediment was taken. Important germination cues such as light may vary, or seed burial depths may be incorrect. Potentially viable seed may be missed if specific germination requirements, such as chilling, or pregermination processing, such as passing through the gut of a bird, did not occur.

Published aquatic seed bank studies are uncommon. Keddy and Reznicek (1982) studied the relationship between existing vegetation and the seed bank of a coastal plain in Ontario. In a

study from a large, calcareous lake in Alberta, Haag (1983) grew seeds from sediment in a greenhouse and made interesting observations about seed production at different water depths. Kautsky (1990) examined seed and tuber banks from a bay of the Baltic Sea. Bonis and Lepart (1994) examined differences in the seed bank associated with depth in sediment for temporary saline marshes in France. Additionally, Bonis and Grillas (2002) wrote a useful review about charophyte propagule banks.

Macrofossil Production

Annual production of macrofossil type A, factor P_A of the previously discussed macrofossil:vegetation inference model, is very much dependent on the plant part in question. For example, plant parts that are intentionally shed on an annual basis (e.g., deciduous leaves, *Larix* needles, and bud scales), as well as reproductive parts such as seeds, staminate and pistillate strobili, cone scales, and oospores, can be expected to be found in sediment in significantly greater quantities than plant parts that are shed only occasionally or in small quantities (e.g., vegetative parts including twigs and evergreen leaves or needles). However, species with parts that break into identifiable fragments may be overrepresented in the fossil record depending on how fragments are counted (Holyoak, 1984).

P_A depends on reproductive and overwintering strategies of species A. Annual plants, which overwinter as seed, typically produce many times more seed per year than perennial species. Many obligate aquatic plant species successfully produce seed only under specific conditions. The primary reproductive strategy for these species is vegetative. They overwinter as tubers or dormant buds, which decay more readily than seed, reducing representation of these species in the macrofossil record. In a seed production experiment, Yeo (1966) found that the production of seeds by different aquatic species varied from zero to 2 million per plant. Production of tubers, turions, and dormant buds by species that rely on these reproductive methods varied from less than 100 to nearly 70 000. For low producers, such as *Nuphar*, just a few seeds may represent a large mass of plants, whereas for high producers, such as *Najas*, many seeds may in fact represent few plants.

To further complicate factor P_A , seed production may vary from year to year depending on environmental conditions. Some obligate aquatic species, for example, only produce seeds during dry growing seasons that leave the plants stranded above the water line. Bonis and Lepart (1994) found seed production differences of up to 600% within a single species. Production of charophyte oospores may differ from 4 to 70% from one year to the next (Bonis and Lepart, 1994; Bonis and Grillas, 2002).

Annual seed production may also vary among individuals of the same species in the same lake depending on site characteristics. Haag (1983) stated that seed production in some species decreases with increasing water depth because of reduced irradiance and a longer amount of time required to grow to the surface, where many aquatic species carry out pollination. Bonis and Grillas (2002) observed that decreasing water depth triggers sexual maturation of *Chara australis*, with

individuals remaining infertile in water over 54 cm deep; *Nitella sonderi* successfully set oospores at all water depths.

Taphonomy

Taphonomy, the study of processes by which fossils become incorporated into sediment, encompasses the other factors of the macrofossil:vegetation inference model: loss (L_A), dispersal distance (D_A), basin characteristics (b), and within-lake sample location (S).

Loss

Potential macrofossils are lost as a consequence of decomposition, predation, and seed germination, although germinated seeds may still leave behind identifiable fragments of seed coats. Long-term preservation is greatest for plant parts that are meant to be long-lasting, such as seeds with strong seed coats and evergreen leaves with thick cuticles. Although soft leaves decay more rapidly, even some of these may leave behind well-preserved parts, such as the spines of *Ceratophyllum* leaves. Spicer and Greer (1986) speculated that basin morphology can affect decomposition rate within sediment. A lake with a low surface area relative to its depth has less oxygen entering the water by means of diffusion across the water–air interface, creating a less aerobic environment and decreased decomposition.

Dispersal Distance

Dispersal distance of potential macrofossils from source plants is influenced by both species and basin/environment characteristics. Tiny seeds or parts, for example, are likely to be more widely dispersed than large ones. They may be carried a greater distance before settling and become resuspended and winnowed into deeper water as a consequence of seasonal overturn or wind-generated wave action. Dieffenbacher-Krall and Halteman (2000) found that, in general, potential macrofossils are deposited at water depths equal to or greater than the depth at which the source plants grow. Essentially, macrofossils are more likely to travel downhill, into deeper water, than they are to travel uphill. However, the study found exceptions to this generalization because wave action can push buoyant seeds back toward shore before they settle.

Macrofossil buoyancy has great significance with respect to dispersal distance, and a fair amount of data exist about seed buoyancy. In an article written in 1893, H. B. Guppy shared observations of seeds floating down the River Thames. Guppy pondered whether seed transportation by water might be a possible explanation for multicontinental distribution of some aquatic plant species. His curiosity led to greenhouse experiments in which he tested the seed buoyancy of 320 British plant species in both fresh and salt water (Guppy, 1906). Praeger (1913) studied seed buoyancy of 786 British plant species. Guppy's studies led him to conclude that long-distance transport of species via seawater was probably not a significant factor in circumboreal distribution of plant species, and he reverted to the usual suspect, waterfowl, as a more probable explanation. However, his and Praeger's studies still

constitute the greatest body of knowledge regarding seed buoyancy.

Guppy (1893, 1906) and Praeger (1913) found that the seeds of many species sink almost immediately, whereas others can remain buoyant for months in either fresh or salt water. Praeger reported that 44.3% of species tested had seeds that sunk within 1 minute, 12.6% of species had seeds that sunk within 1 day, and 30.2% had seeds that sunk within 1 week. The remainder were buoyant for longer periods of time. Guppy generalized that riverside and seaside species tended to have more buoyant seeds. He found that grasses possess little buoyancy, although air bubbles trapped in glumes may cause floating for a few days. *Scirpus* seeds sink rapidly, but *Carex* seeds may float for several months if they remain trapped within the buoyant utricle. It is important to note that Guppy used seeds in a condition nearly like that at the time of natural dispersal. Praeger thoroughly dried his seeds, which would increase their buoyancy. Thus, the results of the two researchers are sometimes in disagreement regarding the same species.

Davis (1985) tested the specific gravity and settling velocity of seeds of eight obligate aquatic species as part of his aforementioned surface sample study. He found that all seeds tested had specific gravities >1.0 and sank rapidly in distilled water, with the exception of *Potamogeton*, which floated. Davis speculated that seeds with slower settling velocities would tend to disperse more broadly; his surface sample analysis confirmed this hypothesis.

Little is known about buoyancy of macrofossils other than seeds. Hill and Gibson (1986) conducted floatation studies on leaves of 11 Tasmanian species. Most leaves floated for long periods of time, but 3 species had leaves that sank quickly. In their surface sediment samples, Hill and Gibson found that the more rapidly sinking leaves were deposited in nearshore sediments, whereas leaves of other species were dispersed more broadly.

Dispersal mechanisms also affect distance of dispersal from growing source plants. Kautsky (1990) found that *Potamogeton pectinatus* produced abundant quantities of seed, but the seeds were not found in sediment from immediately below the plants. The seeds remained attached to parent plants and drifted away with them as the plants senesced in autumn to be deposited on the shoreline in abundance. Wind-dispersed seeds, such as those of birch and conifers, are often found in great numbers in sediment samples from the center of lakes and in low numbers near shore. Seeds may also be transported away from parent plants by waterfowl.

Season of dispersal of potential macrofossils can greatly influence the distance of macrofossil settling from source plants. Macrofossils that disperse after snowfall, such as *Betula* seeds and pistillate bracts or *Larix* needles, may blow great distances over snow before coming to rest. Macrofossils blowing across a lake may become incorporated in the ice. The seeds melt down into the ice as a consequence of greater solar radiation absorption because of their dark color. Guppy (1893) observed many seeds being transported within ice and found that ice often damages seed coats. Although not affecting germination, seed coat damage causes the seeds to sink immediately after ice melt. Any macrofossils floating at the time of freeze-over become incorporated into ice and sink in place when the ice melts. Glaser (1981) found that macrofossils

may be sorted as they sweep across snow according to different aerodynamic properties.

Basin and Environmental Characteristics

Additional features of the local environment may affect dispersal distance, particularly the energy of the environment determined by the slope of the catchment basin and lake bottom. Shoreline vegetation impacts representation of terrestrial vegetation within the macrofossil record by filtering in-wash. The mode of entry into a basin by a macrofossil affects how far from shore the macrofossil settles into sediment. For example, macrofossils falling from overhanging branches tend to come to rest in nearshore sediment. Those carried by streams would have a distribution related to the energy dissipation pattern of the stream water. Sporadic events, including periods of high runoff, erosion, or local fires, can also change patterns of macrofossil distribution.

Sample Location within a Basin

Macrofossil deposition is not consistent across a basin. Assemblage composition varies with the sediment core location within a basin. More macrofossils are found in upwind sediment (Wainman and Mathewes, 1990), and nearshore sediment generally contains many more macrofossils than lake center sediment (Watts and Winter, 1966). Davis (1985) and Hill and Gibson (1986) also noted high spatial variability of macrofossil composition across a lake, with different samples providing very different pictures of the surrounding vegetation. Samples from different water depths reflect different vegetation zones within and around a lake, making macrofossils a useful proxy for past water level changes (Digerfeldt, 1986). Dieffenbacher-Krall and Halteman (2000) demonstrated that macrofossils from some species are deposited in a limited range of water depths, whereas others spread broadly across a lake.

Representation

All factors considered, how reliably can V_A be inferred from the macrofossil content of sediment? Certainly, a one-to-one relationship between macrofossils and vegetation community is not to be expected. Macrofossil abundance is not a direct measurement of plant abundance, and the absence of a macrofossil type does not confirm the absence of its associated plant in the local vegetation. However, Watts and Winter (1966) argued that quantitative analysis of macrofossils is valid. A predictable relationship may be made possible by evaluating individual macrofossil taphonomy and which species tend to be over-, under-, or not represented in the macrofossil record. Macrofossil presence within sediment is a definitive indicator of the local presence of particular plant species. Occasionally, one may find seeds of species not noted in a lake or surrounding vegetation, but these tend to be small plants easily overlooked in a vegetation survey, such as *Callitriche* or *Najas gracillima*. The results of surface sample studies provide strong evidence against long-distance transport as a possible confounding factor, finding very few incidences of macrofossil transport from substantial distances.

The macroscopic remains of aquatic and wetland species tend to originate near their sedimentation point. Greatrex (1983) estimated that 50–100% of aquatic and wetland macrofossils originated within 1 m of the sample sites. Zhao et al. (2006) determined that for some species, 100% of macrofossils derived from plants within 20 m of the sample sites. However, representation is inconsistent. In some cases, a macrofossil type is abundant where the plants are abundant. Often, however, a macrofossil type may be abundant with no potential source plants in the near vicinity (Birks, 1973; Dieffenbacher-Krall and Halteman, 2000; Zhao et al., 2006). Species abundance does not ensure abundance of its macrofossils in the sedimentary record. Greatrex (1983) found few *Carex* seeds despite the fact that the bulk of the sediment was composed of the vegetative remains of sedge. Rare species are unlikely to be represented in the macrofossil record, and some species simply do not appear to produce preservable macrofossils.

Surface sample studies confirm that macrofossil assemblages do not provide one-to-one representation of upland communities. Some woody genera, particularly *Betula*, *Alnus*, and *Larix*, are excessively prominent in macrofossil assemblages. Their abundance or near proximity to a sample site may better be determined by a combination of seed and vegetative remains rather than simply seed alone. Other genera, such as *Salix*, may not be observed in the seed record, but they may contribute other plant parts, such as scales, leaves, twigs, catkins, or stipules. Upland macrofossil assemblages may be swamped by abundant remains of obligate and facultative hydrophyte species. For these reasons, macrofossil data are usually presented as concentrations per unit of sediment rather than as percentages of macrofossil sum.

Macrofossil samples may contain remains of all the major forest species in a catchment (Spicer and Wolfe, 1987), or as few as 4 of 14 tree species and 10 of 16 shrub species may be represented (Wainman and Mathewes, 1990). Nonetheless, macrofossils provide a valuable means of characterizing local vegetation if relationships between macrofossil abundance and plant abundance are known. Dunwiddie (1987) concluded that needle assemblages in sediment from small ponds could provide estimates of past species composition in Pacific Northwest coniferous forest, although herbaceous plant representation was sporadic. He found that the macrofossil record represented 85% of tree species, but when herbaceous species were included, macrofossils represented only 22% of the total terrestrial flora. In a region of shifting ecosystem boundaries across Minnesota, Birks (1973) determined that macrofossil assemblages were specific to distinct ecotones.

Macrofossil and pollen analysis together greatly enhance accurate vegetational community reconstruction. Pollen assemblages often contain significant proportions of taxa that are not present in local vegetation (Dunwiddie, 1987), and they poorly reflect the aquatic species composition of a lake (Zhao et al., 2006). However, the long-distance transport issues that complicate pollen analysis are absent from macrofossil analysis. Some species, such as *Larix*, produce small quantities of pollen that is poorly represented in the pollen rain. However, *Larix* is generally abundant in the macrofossil record if the species was present in the local vegetation. Other plants, such as *Artemisia*, are rare in the macrofossil record but adequately

represented by pollen. Species living at ecological extremes of their geographic ranges or at treeline may not produce much pollen, but their presence is recorded by macrofossil remains. In these situations, the pollen assemblage may be misleadingly dominated by long-distance transport (Eide et al., 2005). Macrofossils can confirm the presence of some species, such as *Thuja plicata* in the Pacific Northwest, because macrofossils can often be identified to a finer taxonomic level than pollen. Birks (2003) and Birks and Birks (2000) cite multiple cases in which macrofossil analysis was essential for accurate interpretation of pollen records.

Conclusion

Plant macrofossils preserved in lake sediment do not provide a direct means of estimating proportions of species within local vegetation. Deposition of individual macrofossil taxa is affected by such factors as annual production of the macrofossil, annual loss of the macrofossil, dispersal distance from source plants, basin characteristics, and sample location within the lake. However, study of plant macrofossils in surface sediment samples, and comparison with vegetation surveys, provide the data necessary to make qualitative, if not quantitative, estimates of past vegetational community composition based on macrofossils. Such studies are the only means of discovering biases within macrofossil assemblages that result from taphonomic processes and individual species characteristics.

See also: Plant Macrofossil Introduction. **Plant Macrofossil Methods and Studies:** Paleolimnological Applications; Validation of Pollen Studies.

References

- Birks HH (1973) Modern macrofossil assemblages in lake sediment in Minnesota. In: Birks HJB and West RG (eds.) *Quaternary Plant Ecology*, pp. 172–189. Oxford: Blackwell.
- Birks HH (2003) The importance of plant macrofossils in the reconstruction of Lateglacial vegetation and climate: Examples from Scotland, western Norway, and Minnesota, USA. *Quaternary Science Reviews* 22: 453–473.
- Birks HH and Birks HJB (2000) Future uses of pollen analysis must include plant macrofossils. *Journal of Biogeography* 27: 31–35.
- Birks HJB (1995) Quantitative palaeoenvironmental reconstructions. In: Maddy D and Brew JS (eds.) *Statistical Modeling of Quaternary Science Data*, pp. 161–254. Cambridge, UK: Quaternary Research Association.
- Birks HJB and Birks HH (1980) *Quaternary Palaeoecology*. London: Arnold.
- Bonis A and Grillas P (2002) Deposition, germination, and spatio-temporal patterns of charophyte propagule banks: A review. *Aquatic Botany* 72: 235–248.
- Bonis A and Lepart J (1994) Vertical structure of seed banks and the impact of depth of burial on recruitment in two temporary marshes. *Vegetatio* 112: 127–139.
- Davis FW (1985) Historical changes in submerged macrophyte communities of Upper Chesapeake Bay. *Ecology* 66: 981–993.
- Dieffenbacher-Krall AC and Halteman WA (2000) The relationship of modern plant remains to water depth in alkaline lakes in New England, USA. *Journal of Paleolimnology* 24: 213–229.
- Digerfeldt G (1986) Studies on past lake-level fluctuations. In: Berglund BE (ed.) *Handbook of Holocene Palaeoecology and Palaeohydrology*, pp. 127–143. Chichester, UK: Wiley.
- Drake H and Burrows CJ (1980) The influx of potential macrofossils into Lady Lake, north Westland, New Zealand. *New Zealand Journal of Botany* 18: 257–274.
- Dunwiddie PW (1987) Macrofossil and pollen representation of coniferous trees in modern sediments from Washington. *Ecology* 68: 1–11.
- Eide W, Birks HH, Bigelow NH, Peglar SM, and Birks HJB (2005) Holocene forest development along the Setesdal valley, southern Norway, reconstructed from macrofossil and pollen evidence. *Vegetation History and Archaeobotany* 15: 65–85.
- Glaser PH (1981) Transport and deposition of leaves and seeds on tundra: A late-glacial analog. *Arctic and Alpine Research* 13: 173–182.
- Greatrex PA (1983) Interpretation of macrofossil assemblages from surface sampling of macroscopic plant remains in mire communities. *Journal of Ecology* 71: 773–791.
- Guppy HB (1893) The River Thames as an agent in plant dispersal. *Journal of the Linnean Society* 29: 333–346.
- Guppy HB (1906) *Observations of a Naturalist in the Pacific between 1896 and 1899*. London: Macmillan.
- Haag RW (1983) Emergence of seedlings of aquatic macrophytes from lake sediments. *Canadian Journal of Botany* 61: 148–156.
- Hill RS and Gibson N (1986) Distribution of plant macrofossils in Lake Dobson, Tasmania. *Journal of Ecology* 74: 373–384.
- Holyoak DT (1984) Taphonomy of prospective plant macrofossils in a river catchment on Spitsbergen. *New Phytologist* 98: 405–423.
- Kautsky L (1990) Seed and tuber banks of aquatic macrophytes in the Askö area, northern Baltic proper. *Holarctic Ecology* 13: 143–148.
- Keddy PA and Reznicek AA (1982) The role of seed banks in the persistence of Ontario's coastal plain flora. *American Journal of Botany* 69: 13–22.
- McQueen DR (1969) Macroscopic plant remains in recent lake sediments. *Tuatara* 17: 13–19.
- Praeger RL (1913) On the buoyancy of seeds of some Britannc plants. *Scientific Proceedings of the Royal Dublin Society* 14: 13–62.
- Spicer RA and Greer AG (1986) Plant taphonomy in fluvial and lacustrine systems. In: Broadhead TW (ed.) *Land Plants: Notes for a Short Course*, pp. 10–26. Knoxville: University of Tennessee.
- Spicer RA and Wolfe JA (1987) Plant taphonomy of late Holocene deposits in Trinity (Clair Engle) Lake, northern California. *Paleobiology* 13: 227–245.
- Wainman N and Mathewes RW (1990) Distribution of plant macroremains in surface sediments of Marion Lake, southwestern British Columbia. *Canadian Journal of Botany* 68: 364–373.
- Watts WA and Winter TC (1966) Plant macrofossils from Kirchner Marsh, Minnesota—A paleoecological study. *Geological Society of America Bulletin* 77: 1339–1360.
- Yeo RR (1966) Yields of propagules of certain aquatic plants I. *Weeds* 14: 110–113.
- Zhao Y, Sayer CD, Birks HH, Hughes M, and Peglar SM (2006) Spatial representation of aquatic vegetation by macrofossils and pollen in a small and shallow lake. *Journal of Paleolimnology* 35: 335–350.

Treeline Studies

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Introduction

Treelines and timberlines are eye-catching landscape and biological boundaries beyond which trees cannot grow for various reasons (Figures 1, 2, and 3). Temperature and precipitation thresholds are the most important factors in determining whether a site is treeless or not, but vegetation type, soil type, snow cover, topography, and wind can locally codetermine the limit of tree growth. For instance, low temperatures at Alpine or polar treeline significantly reduce production of plant tissues (Körner, 1999). Thus in comparison to other vegetational life-forms, trees are particularly affected by temperature deficits because of their strategy to accumulate large amounts of biomass to overgrow other plants. Moisture deficits determine the transition between forest and grassland communities. This treeline is often called 'lower treeline' or 'dry treeline' to distinguish it from 'Alpine', 'polar', or 'upper treeline'.

Most ecological studies prefer to differentiate between tree-line and timberline. Treeline (or tree limit) designates the limit of tree growth, and timberline (or forest line, forest limit) the limit of closed forests. In mountainous areas the transitional zone between these two vegetational boundaries is usually rather narrow under natural conditions (i.e., 100–200 m of vertical extent). Anthropogenic impact (e.g., pastoral activities or timber exploitation) tends to lower timberline more than treeline, causing an extension of the transitional zone (e.g., up to 300–400 m in the Alps). Polar and drought-caused timberline and treelines can be separated by very broad transition zones such as parklands with widely spaced trees (Arno and Hammerly, 1993).

The spatial position of treeline depends on the definition of 'tree'. The most common criterion to define 'tree' at treeline is a minimum height between 2 and 8 m, and this criterion is sometimes used to separate timberline from treeline (e.g., 8 m versus 2 m of tree height, Tinner and Theurillat, 2003). Above treeline, tree species may grow as low, prostrate, or stunted individuals (<2 m, Figures 4 and 5). This zone is often referred to as the krummholz zone, and its upper limit as krummholz limit or tree-species limit. In the zone between timberline and the krummholz limit, total biomass drops dramatically (e.g., at the boreal forest limit from ca. 20 kg m⁻² to 0.6 kg m⁻², Strasburger, 1991) since tall trees are gradually substituted by shorter shrubs and herbs. Net primary productivity also declines, though to a less extent (e.g., in the Alps from ca. 1000 g/(m²·yr) to <200 g/(m²·yr), Ellenberg, 1996). Vegetational changes at treeline strongly influence many ecosystem elements and processes (e.g., animal life, soil formation, water balance, local climate, energy, and nutrient fluxes) and are, therefore, an exceptionally attractive subject of ecological and paleoecological research.

The main aim of paleoenvironmental treeline studies is to reconstruct past treeline positions and compositions. The

significance of such reconstructions lies mainly in two factors: (i) the paleoclimatic approach uses the climatic sensitivity of treelines to quantify past climatic changes at decadal to millennial scales. Due to the local to regional spatial resolution such studies provide valuable information about climatic changes at mountainous or polar sites; and (ii) the paleoecological approach uses independent environmental proxies (e.g., oxygen isotopes as climatic proxy) to provide insights into biosphere responses to rapid and/or extreme environmental changes.

Reconstructing Treelines by Using Macrofossils

Treeline changes can be reconstructed by consulting natural environmental archives such as lake or mire deposits. In such deposits, pollen is usually ubiquitous and, therefore, easy to be analyzed and quantified. However, because pollen is easily transported by winds over long distances and great elevation differences, pollen records at treeline have rather low spatial resolution and are, therefore, difficult to interpret. Macrofossil and megafossil analyses are considered to be more reliable tools for reconstructing treelines, since they have higher spatial resolution (meters to decameters) than pollen studies (decameters to kilometers; Birks, 2001). Megafossil and macrofossil records have also higher taxonomic resolution. If compared with megafossil studies, macrofossil studies have two advantages: (i) due to their small sizes (0.1 mm to cm-sized) macrofossils are deposited in stratigraphic order and can, therefore, provide complete Holocene and/or Glacial sequences with high-temporal resolution from single sites; and (ii) in contrast to megafossils, macrofossil numbers are sometimes high enough to be evaluated quantitatively, in a manner similar to that for pollen, diatoms, or chironomids. However, megafossils have the advantage of higher spatial resolutions, since they are often found *in situ* where they were growing. In addition, logs may be used for temporally precise dendroecological or dendroclimatic studies (e.g. Nicolussi et al., 2005).

To gain treeline-inferred temperature reconstructions it is assumed that the present occurrence of trees at treeline is in equilibrium with local climate of recent decades. After having determined the past spatial position of treeline by macrofossil analysis, past treeline fluctuations (m) can be converted to (summer) temperature changes (°C) by using today's lapse rates (under the assumption of, for example, Holocene stability). In regions affected by strong human impact (e.g., the Alps) single trees in remote areas may still indicate the potential altitudinal limit of forest and tree growth (Figure 4). A second assumption is often made when interpreting pollen and macrofossil records at treeline. Changing values of tree species in pollen or macrofossil records are usually considered to reflect tree-population dynamics rather than productivity changes of



Figure 1 Lower and upper timberlines in the continental climate of Idaho, USA. Sagebrush (*Artemisia* steppe) in foreground. Tree growth is prevented by lack of moisture on the lower slopes. Photograph by HH Birks.



Figure 2 Upper natural timberline in the Rocky Mountain National Park (CO, USA, established 1915). Timberline is formed by *Abies lasiocarpa*, *Picea engelmannii*, and *Pinus flexilis*. The transitional zone (ecotone) between sub-Alpine forests and Alpine meadows is narrow and forests rather closed up to treeline. Photograph by W Tinner.



Figure 3 Timberline in the Swiss Central Alps (Valais region) is located at about 2000–2200 m and if compared with the natural position it is lowered by about 200–400 m by anthropogenic activities. It is formed by *Picea abies* (up to ca. 2100 m), *Pinus cembra*, and *Larix decidua* trees. The lower parts of the forests are disrupted for economic purposes. In the foreground, the study site Gouillé Rion (2343 m a.s.l.) which is situated at today's treeline. Photograph by W Tinner.



Figure 4 Highest occurrences of *Pinus cembra* trees near the site Gouillé Rion, Central Swiss Alps. The size of the trees is decreasing with increasing altitude (ca. 2320–2400 m a.s.l. (above sea level) in the picture). Uppermost individuals at about 2380 m a.s.l. reach a size of ca. 0.5 m. Trees at these altitudes are restricted to rocky slopes which are not easily accessible. Photograph by W Tinner.



Figure 5 Highest *Pinus cembra* individual near Gouillé Rion (Swiss Alps) at 2415 m a.s.l. The *Pinus cembra* krummholz shows winter damage by desiccation which is identifiable by the orange–brown color of dead shoots. During winter the low portions of the shrub were encased in snow packs and thus protected from winter drought. In the background single trees and tree groups of Swiss stone pine at 2350–2300 m a.s.l. (= treeline). Photograph by W Tinner.

the individuals alone such as higher pollen or plant-organ production in response to warmer climate. For instance, reduced macrofossil concentrations of tree species are conventionally interpreted as the consequence of a local decline in treeline elevation.

Macrofossil data are usually presented as concentration data, i.e., as macrofossil numbers or areas (e.g., mm²) per volume of sediment (e.g., 50 cm³, Birks, 2001). The reason for such a procedure is that the numbers found in the sediments are generally too low to allow reliable estimates of accumulation rates or calculation of percentages. However, in contrast to percentage values, concentration values presuppose uniform sedimentation rates. Another constraint of concentration values is that, in contrast to accumulation rates and percentages, they cannot be directly linked to modern vegetational data or observations, although the establishment of such a link

is compelling for a correct interpretation of any kind of paleoecological data. In treeline studies one of the most important questions is whether a site was below, at, or above treeline. Due to the above-mentioned restrictions, it is not easy to address this question in detail by using macrofossil concentration values alone. Macrofossil percentages may help to address this issue where numbers are high enough and provided that the range of values does not differ greatly among the taxa (Briks, 2001). Different studies have shown that sums should reach 45–50 to produce reliable percentage results (Hofmann, 1986; Heiri and Lotter, 2001). In the best case, macrofossil sums should be dominated by only one kind of macrofossil type (e.g., needles or leaves) to reduce the range of values. A recent treeline study from the Alps (Kaltenrieder et al. 2005, Tinner and Kaltenrieder 2005) illustrates this approach (Figure 6). In a first step, macrofossil-concentration values from two independently radiocarbon-dated cores (10 m apart in a small lake) were compared to assess whether the macrofossil results can be reproduced temporally and spatially. For this purpose the temporal resolution of the counts of core GR6/7 was lowered by always amalgamating six adjacent samples. Even minor fluctuations in macrofossil numbers, such as the different peaks of *Larix decidua* needles, occurred in both stratigraphies, indicating that low macrofossil counts and local tafonomic processes did not significantly affect the spatial and temporal reproducibility of the data. In a second step, macrofossils (e.g., needles, leaves, or fruits) of taxa that could be unambiguously attributed to trees (indicating forest), shrubs (shrublands), or creeping shrubs and upland herbs (Alpine meadows, debris vegetation, or snowbeds) were considered for percentage calculation, whereas wood fragments, twigs, charcoal, bark, aquatic plants, and Cyperaceae were excluded

from the macrofossil sum (Tinner and Theurillat, 2003). Since macrofossils are not transported very far from their point of origin, the subsums indicating growth of forests, shrublands, or Alpine meadows (Figure 6) provide valuable information about the past dominant vegetational types at this site. According to the percentage data the site was densely forested from about 10 000–4500 calibrated years before present (cal yr BP; sums of trees usually between 90% and 100%). Alpine meadows were predominant before 11 500 cal yr BP and admixed with forests prior to 11 000 cal yr BP, whereas a special vegetation type dominated by shrublands and meadows (treeline ecotone) became established after 4500 cal yr BP. This enlargement of the treeline ecotone was primarily a result of human activity (Tinner and Theurillat, 2003).

Late-Glacial Treeline Dynamics

In the temperate biomes of the Northern Hemisphere, (dense) forests and thus treelines were established during the early Late-Glacial, ca. 18 000–14 500 cal yr ago. Only the southern regions of Europe were forested during this early period of the Late-Glacial, and macrofossil records show that treeline depressions there reached at least 1000–1500 m in comparison with today. For instance, radiocarbon-dated macrofossil records unambiguously document rather closed *Pinus cembra* treeline stands at 400–500 m in the southern Alps ca. 16 000–14 500 cal yr ago (e.g., site Balladrum, near Locarno; Hofstetter et al., 2006 in press). Summer temperatures increased abruptly after 14 500 cal yr BP (Heiri and Millet, 2005), and treelines in southern Europe subsequently reached higher positions, although they were still about 600 m lower

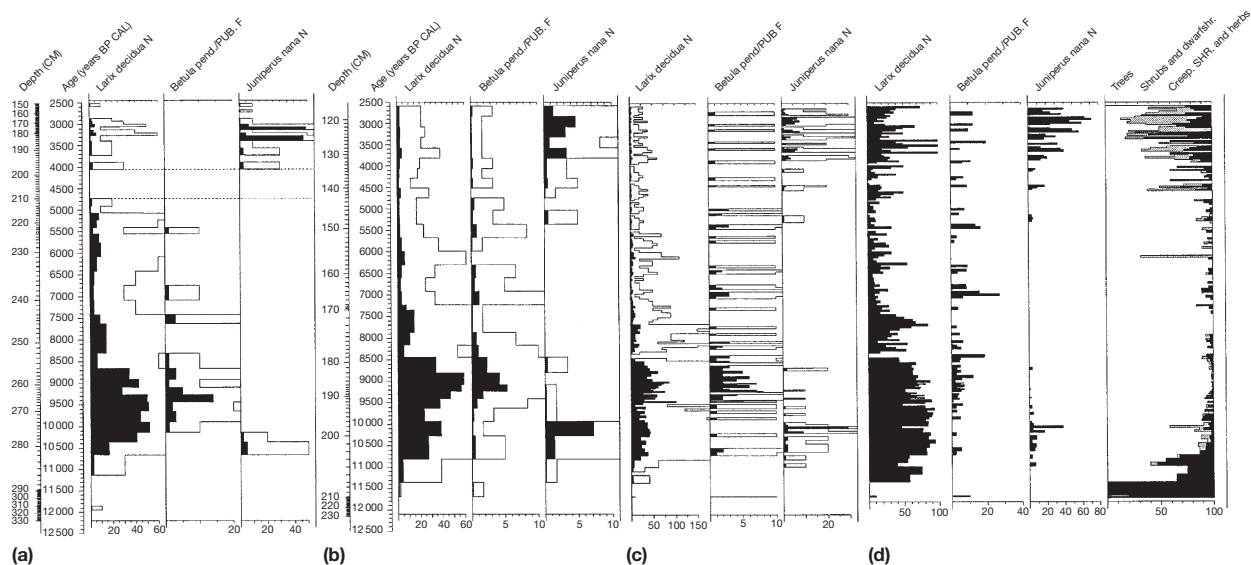


Figure 6 Comparison of macrofossil records from different cores with independent chronologies from the treeline pond Gouillé Rion (2,343 m a.s.l., Swiss Alps): (a) Macrofossil concentrations of core GR2. The age–depth model used for the chronology of GR2 relies on 14 mass spectrometry ^{14}C -measurements on identified terrestrial plant macrofossils. (b) Macrofossil concentrations of core GR6/7, always six adjacent samples of 0.5 cm were amalgamated to reach temporal resolutions comparable with GR2. Cores GR2 and GR6/GR7 are 10 m apart. The age–depth model used for the chronology of GR6/7 relies on 12 mass spectrometry ^{14}C -measurements on identified terrestrial plant macrofossils. (c) Core GR6/7, same data as (b) but with sixfold higher resolution. (d) Macrofossil percentages for core GR6/7, same resolution as (c). For details on macrofossil percentage calculations see Tinner and Kaltenrieder (2005).

than they are today. Similarly, macrofossil studies document that after having been established at about 15 000–14 500 years ago, treelines in and north of the Alps were depressed by about 600–800 m during the period 14 500–11 500 cal yr ago (Tobolski and Ammann, 2000, Gobet et al., 2005).

Strong treeline depressions during the Late Glacial were also characteristic for Northern America. Macrofossil studies document that treeline was lowered by at least 600 m during the period 17 000–13 500 cal yr BP in the Yellowstone and Grand Teton National Park areas (Whitlock, 1993). After 13 500 cal yr BP treelines moved upwards, but today's forest limits were reached only after 11 500 cal yr BP at the beginning of the Holocene. These results from the Northern Rocky Mountains are corroborated by pollen studies in the Southern Rocky Mountains, suggesting that before 13 000 cal yr BP treeline was lowered by about 300–700 m (Fall, 1997). Detailed macrofossil studies covering the Late-Glacial period, however, are still very rare. Considering the high climatic variability of this period, new macrofossil series with a high chronological resolution and precision would allow a better assessment of local impacts of strong climatic change on treeline vegetation.

Holocene Treeline Dynamics

At the beginning of the Holocene treelines moved rapidly upwards, as shown by the analysis of macrofossils from lake and mire deposits along altitudinal transects. In the Alps, the timberline ascended 800 m in 200–300 years (11 550–11 350 cal yr BP, see Figure 6 for expansion of trees at today's treeline at about 11 350–11 250 cal yr BP; Tinner and Kaltenrieder, 2005). These rapid vegetational changes were caused by abrupt climate warming of about 4–6 °C at around 11 550–11 350 cal yr BP. During the Holocene, the position of timberline and treeline in the Alps fluctuated within a band of 100–180 m, and treeline was at its uppermost position (ca. 180 m higher than today) during the period 10 000–6000 cal yr BP (e.g. Tinner and Theurillat, 2003). A virtual transect from the central Alps (Figure 7) starting at about 300 m below (2017 m) and ending at about 200–250 m above the today's tree limit (2557 m) illustrates how sharply and unambiguously past forest-Alpine transitions can be traced by macrofossil analysis. Before the onset of large-scale agricultural activities at ca. 2200 BC (Bronze Age), forests were rather closed up to the tree limit at ca. 2400 m (sites Simplon and Goullé Rion in Figure 7). Above the forest limit, the treeline ecotone encompassed about 100–150 m and was characterized by the mixed occurrence of tree, shrub, dwarf-shrub, and herbaceous species (site Goullé Loéré). Natural meadows dominated by creeping shrubs and herbaceous species formed the alpine belt above the treeline ecotone (site Lengi Egga, Figure 7). After 6000–5000 cal yr BP timberline and treeline progressively declined by about 180 m and 300–400 m, respectively. The regression of densely forested areas (timberline) in the Alps during the past 5000 years was primarily caused by human impact, whereas the course of treeline was primarily driven by climatic change (Tinner and Theurillat, 2003). Macrofossil studies suggest that severe diebacks of treeline vegetation occurred during cold-humid periods (CE-1–CE-8, see Figure 8) such as

the 8200 cal BP event. During the early and mid-Holocene treeline oscillated by about ± 50 –100 m. A severe treeline decline lasted from 8400 to 7500 cal yr BP (CE-3), with two minima at 8000 and 7600 cal yr BP. Assuming constant lapse rates of 0.7 °C/100 m it is possible to estimate the Holocene temperature oscillations in the Alps as 0.8–1.2 °C between 10 500 and 4000 cal yr BP, when (summer) temperatures were about 0.8–1.2 °C higher than today. After 4000 cal yr BP summer temperature successively declined to values comparable to the average values of the twentieth century.

At the beginning of the Holocene at ca. 11 550 cal yr BP most areas of Scandinavia were still glaciated. Coastal areas, however, were partially ice-free and thus provide evidence for the earliest afforestation. Macrofossil records show that treeline reached the coast of southwestern Norway (Kråkenes, 62° N) at ca. 10 900 cal yr BP, about 650 years after the beginning of the Holocene (Birks, 2003). Megafossil finds suggest that at about the same time (ca. 11 000 cal yr BP) the tree limit in interior Scandinavia (63° N) had abruptly reached very high positions 100–300 m (uncorrected for glacioisostatic land upheaval) above the modern tree limit (formed by *Betula pubescens* at ca. 800–1000 m a.s.l.), briefly or immediately after the deglaciation (e.g. Kullman, 1995; Bergman et al. 2005). This would imply an almost synchronous expansion of trees over about 1000–1200 altitudinal meters and over distances of hundreds of kilometers. However, recent macrofossil studies along altitudinal transects (Bergman et al. 2005; Eide et al., 2006) suggest that these very early megafossil finds may represent scattered individuals in extremely favorable microclimatic settings and that the regional expansion of forests and thus the establishment of continuous treelines occurred later, at around 9500–9000 cal yr BP (Figure 9). The general subsequent course of Holocene treeline fluctuations in Scandinavia is similar to that reconstructed for the Alpine sites. Megafossil, macrofossil, and pollen studies suggest that treeline was at its uppermost limit between 9500–9000 and 6500–6000 cal yr BP in Norway and Sweden, about 200–400 m above the present treeline position. At some study sites, treeline declined temporarily between 8300 and 8000 cal yr BP (e.g. Dahl and Nesje, 1996; Barnekow, 1999, Barnett et al. 2001). Use of today's lapse rates implies that during the period of highest treeline positions summer temperatures were 1.3–2.0 °C higher than today. A consistent pattern emerges from Asian macrofossil studies (e.g. Kremenetski et al., 1998; MacDonald et al., 2000). Forests across northern Russia developed at the onset of the Holocene at 11 500 cal yr BP. Maximum polar treeline positions were reached at 10 000–5000 cal yr BP, probably in response to summer temperatures 2.5–7.0 °C warmer than modern.

Megafossil and macrofossil records from North America (e.g., Yellowstone and Grand Teton National Park: Whitlock, 1993; San Juan Mountains: Carrara et al., 1991; and New Hampshire White Mountains: Miller and Spear, 1999) suggest that today's treeline positions were reached at the end of the Late Glacial and the beginning of the Holocene at about 12 500–11 000 cal yr BP in response to climatic warming and/or more moisture availability. If compared with Eurasia, today's treeline positions were reached about 1000 years earlier at some of the North American sites. This early development may have resulted from warmer conditions during, for

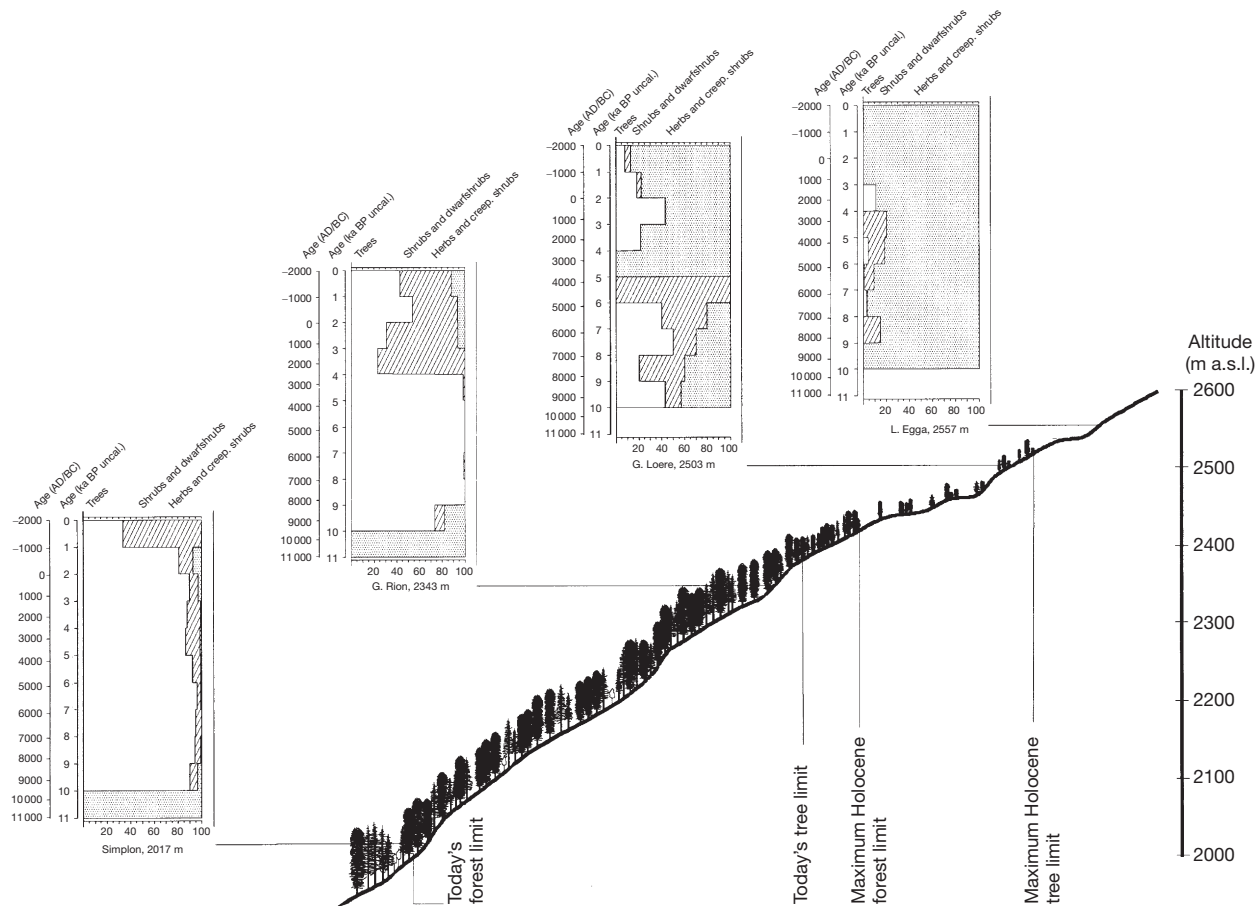


Figure 7 Virtual macrofossil and vegetational transect along an altitudinal gradient in the Swiss Central Alps (Valais region). The macrofossil data are in percentages. Modified from [Tinner and Theurillat \(2003\)](#), with permission. The tree macrofossil sum shows the decreasing relevance of trees (e.g., *Larix decidua*, *Pinus cembra*, and *Betula*) with increasing altitude. In the treeline ecotone shrubs are important. Above the treeline ecotone herbs and creeping shrubs are predominant (Alpine meadows, snow-bed, and debris vegetation). The vegetational transect shows a reconstruction of treeline vegetation for the period 10,000–5,000 cal yr BP, when forests were dominated by *Pinus cembra* and *Larix decidua*. Today Simplon and Gouillé Rion are located in the treeline ecotone, with a mixture of shrub (e.g., *Calluna vulgaris* and *Juniperus nana*), tree and herb species, whereas Gouillé Loéré and Lengi Egga are situated in the Alpine belt where herbs (e.g., *Carex curvula*) and creeping shrubs (e.g., *Salix herbacea*) are predominant.

example, the Younger Dryas (see Late-Glacial treeline dynamics above). Unfortunately, there are only a few North American studies available, and in addition they have rather low temporal resolution and precision. In combination with the radiocarbon plateau at the Late-Glacial/Holocene transition (e.g., 10 500 ^{14}C yr BP = 12 570 cal yr BP; 10 100 ^{14}C yr BP = 11 660 cal yr BP) this may have affected the establishment of a reliable chronology. However, in western North America radiocarbon-dated megafossils and macrofossils that have been preserved in alpine environments provide direct and unequivocal evidence for past high stands of Alpine timberline (e.g. [Carrara et al., 1991](#); [Fall, 1997](#); [Reasoner et al., 2001](#); [Figure 10](#)). The recovery of wood and macrofossils ranging in age from ca. 10 000 to 5000 cal yr BP from sites situated 100–270 m above present treeline clearly indicates that sub-Alpine forests reached at least this elevation during the early and mid-Holocene and suggests that mean (summer) temperatures were 1–2 °C warmer than today. Macrofossil records from across southwestern Canada ([Reasoner et al., 2001](#)) indicate two clear periods of higher-than present treeline during

the Holocene: an early phase from ca. 11 000–10 000 to ca. 8,200 cal yr BP and a later phase between 7300 and 5700 cal yr BP. These data are in general agreement with the European records, suggesting that the 8.2 event may have impacted treelines on a northern hemispherical scale. Moreover, dendrochronological analyses of living trees near alpine treeline and dead snags *in situ* in North America document abrupt synchronous terminations of tree-ring records in response to Holocene climatic oscillations such as the Little Ice Age (e.g. [Luckman, 1994](#)).

It is clear that the position and composition of the timberline ecotone has been sensitive to Late-Glacial and Holocene climate change. The millennial-scale trends generally reflect gradual decline in northern hemisphere insolation from approximately 11% higher-than-present at the end of the Late Glacial and in the early Holocene. Superimposed on this long-term trend are higher-frequency fluctuations related to changes in oceanic circulation, volcanic activity, and solar irradiance or a combination of these factors. Most of these factors influenced climate at continental to global scales, but

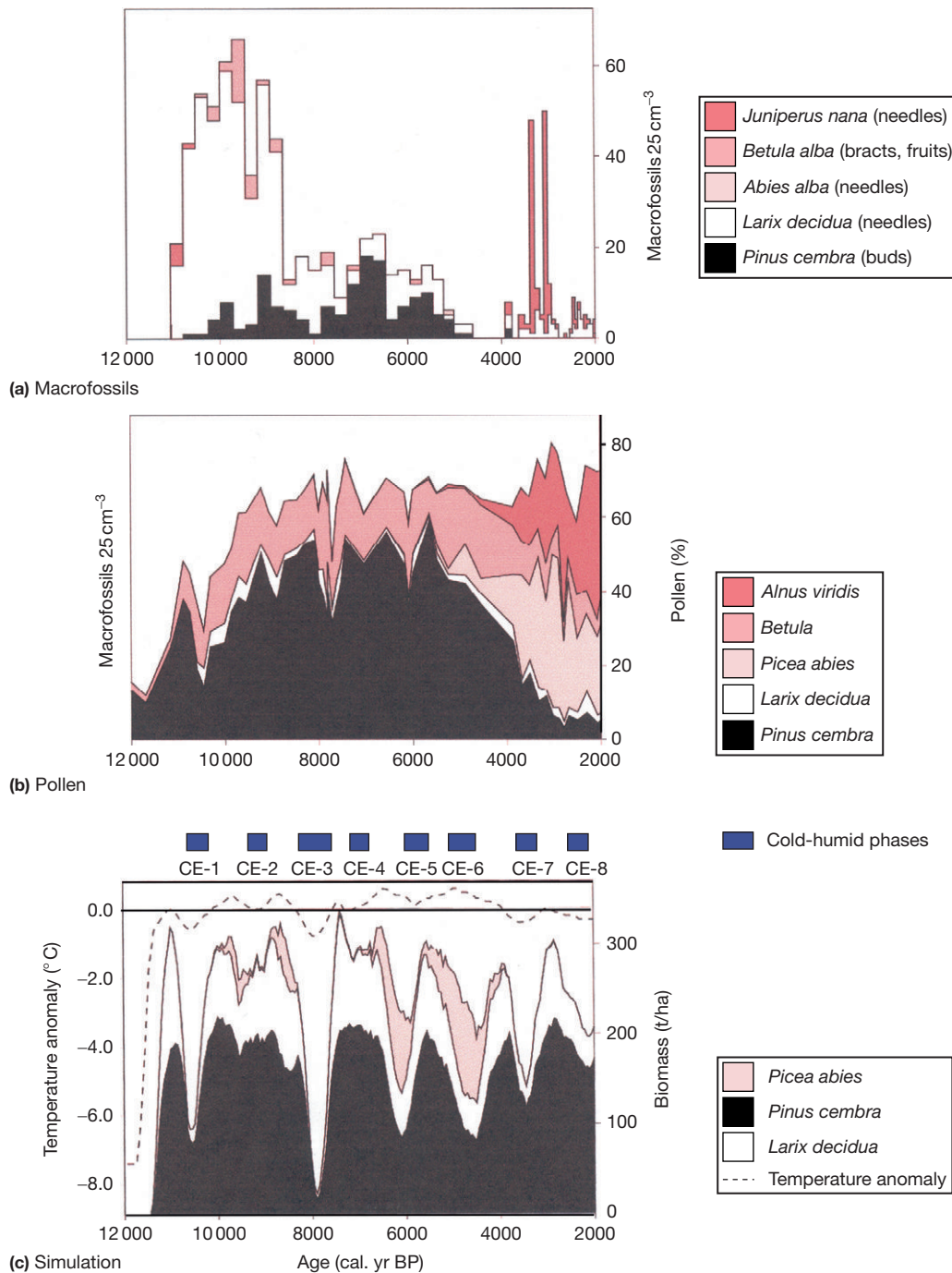


Figure 8 Comparison between (a) macrofossil concentrations, (b) pollen percentages, and (c) simulation results for Gouillé Rion (2,343 m a.s.l., Central Swiss Alps). No correction factors were applied to the pollen data. The order of the species in the legend corresponds to the order in which they are plotted in the figures. Modified from Heiri et al. (2006), with permission. Macrofossil and pollen data from Wick and Tinner (1997). Vegetation-independent Holocene temperature scenario for the Swiss Alps from Heiri et al. (2004). Central European cold phases (CE) according to Haas et al. (1998).

currently it is unclear whether they generated synchronous oscillations of treelines over such wide ranges. The similarity of the general Holocene treeline trends in North America and Europe suggests at least the possibility of such intercontinental linkages, though new high-resolution records are necessary to resolve this issue.

Fossil Evidence and Models: Linking the Past to the Future

It is clear that anthropogenic climate forcing over the next 100 years is likely to rival or exceed the warmest conditions of the Holocene. Concerns about global warming involve

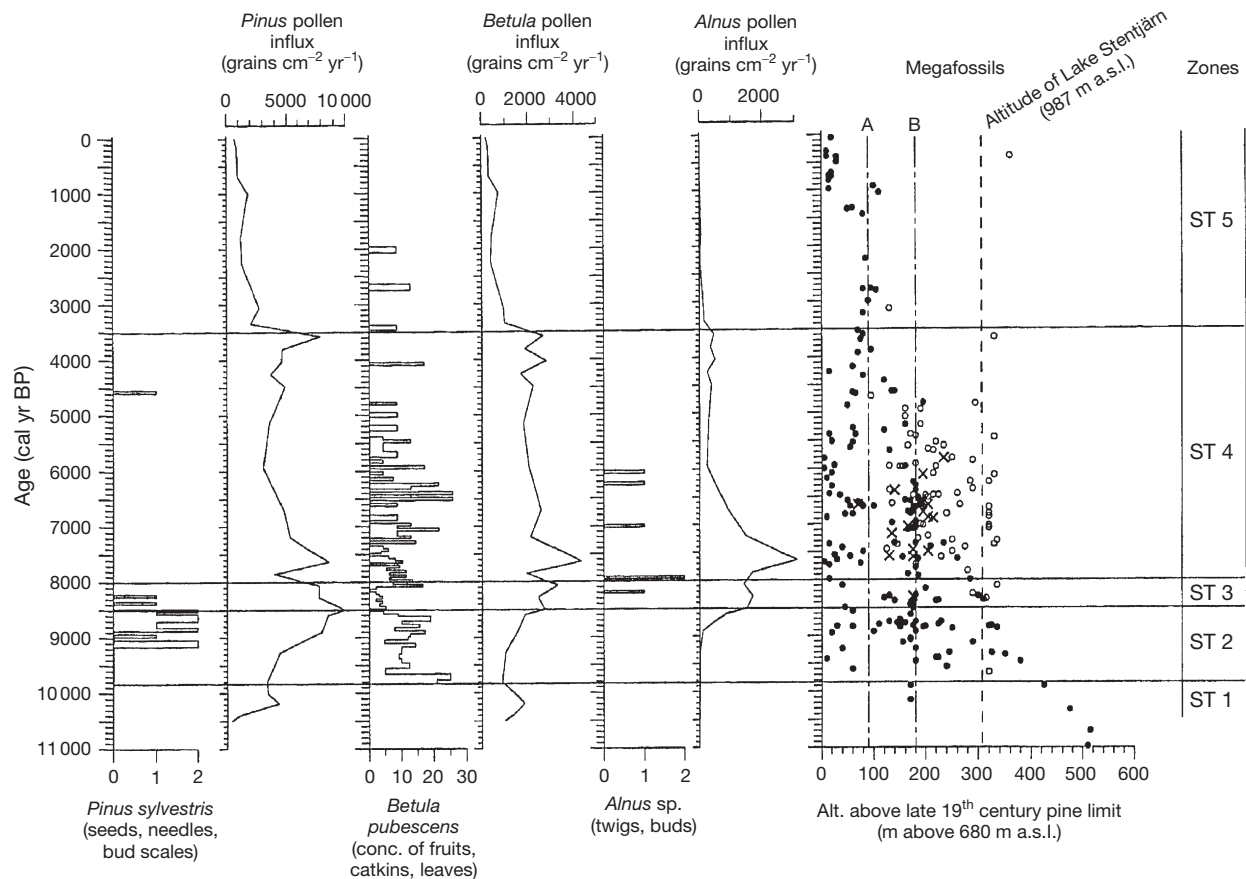


Figure 9 Comparison between records of macrofossils, pollen influx, and radiocarbon-dated finds of subfossil wood remains (megafossils) in the Scandes Mountains, west-central Sweden from Bergman et al. (2005), with permission. The macrofossil and pollen data are from Lake Stentjärn (987 m asl). The original megafossil data were presented in different studies (e.g., Kullman, 1995). The local limit of *Pinus sylvestris* at the end of the nineteenth century is used as reference level. (●) *Pinus sylvestris*, (○) *Betula*, (x) *Alnus incana*. The vertical lines labeled A and B represent upper limits of *Alnus incana* and *Betula pubescens*, respectively.

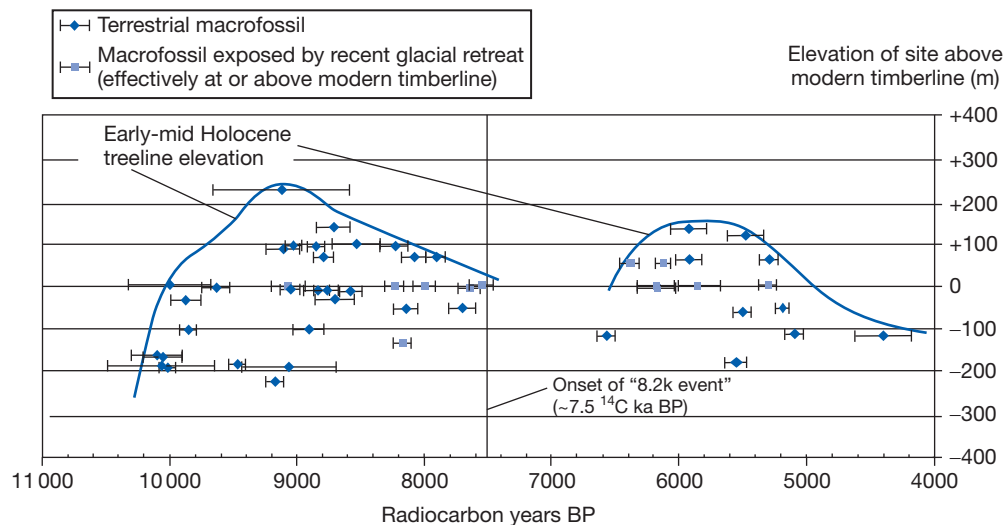


Figure 10 The approximate early-to-mid-Holocene treeline elevation in the southern Canadian Cordillera based on radiocarbon-dated wood macrofossils recovered from high-elevation sites. Modified from Reasoner et al. (2001), with permission.

adjustments, collapses, migrations, or extinctions of boreal and Alpine life. A promising way to assess the environmental consequences of future global change is to use experimental and long-term monitoring data to establish quantitative models and to check the model-based results by paleoenvironmental records. For this purpose, Holocene treeline dynamics along an elevational transect in the Central Alps were simulated by using the forest succession model FORCLIM (Heiri et al., 2006; Figure 8). An independent (chironomid-inferred) temperature reconstruction from the northern Alps (Heiri et al., 2004) was used to drive the model. The simulation results, yielding treeline fluctuations of about ± 100 m above the today's treeline (2350–2400 m, data of altitudinal fluctuations not shown, see Heiri et al., 2006), show a good agreement with pollen and macrofossil data for the period between 11 500 and 4500 cal yr BP (see Holocene treeline dynamics above). Moreover, the timing of treeline lowerings as suggested by the model corresponds to decreases of pollen and macrofossil values of the predominant tree timberline species *Pinus cembra* (e.g., during CE-1–CE-6, Figure 8). These results corroborate the interpretation of most paleoecologists that decadal-to-centennial-scale Holocene fluctuations of pollen and macrofossils reflect treeline shifts rather than productivity changes alone. After 4500 cal yr BP the macrofossil data suggest a distinct lowering of the treeline (Figures 7 and 8), while the model projects the presence of forests up to an altitude of 2400 m a.s.l. This discrepancy between paleobotanical data and simulation is explained by human impact on treeline, a factor that is not considered in the model. In fact, concomitant with the macrofossil-inferred lowering of treeline, pollen types indicative of human impact (e.g., *Plantago alpina*, *Cerealia* from fields in the lowlands) increased markedly at this site (Tinner et al., 1996). Other discrepancies such as the presence of *Picea abies* in the model and its absence in the macrofossil data are explained by precipitation changes that are not considered in the model because such realistic reconstructions are still lacking (e.g., less moisture availability during the early Holocene impeding the expansion of *Picea*). The validation of the model with paleobotanical data was thus successful for the natural or quasinalural period before 4500 cal yr BP, when the treeline position was mainly controlled by climate. This is a first important step towards applying forest-succession models for assessing future vegetation dynamics at treeline.

Independent from their function to validate quantitative models, paleotree line records provide valuable insights in long-term environmental processes that are not easily recognizable, since they exceed the human life span. If the treeline reconstructions are compared with independent proxies of environmental change it is possible to reconstruct past response dynamics. Surprisingly, relatively few studies have addressed past responses of ecosystems such as treeline communities to, for example, climatic change. One reason for avoiding this topic in paleoecology is that accurate studies are extremely labor-intensive and thus require strong efforts to reach very high temporal resolutions (<10–20 yr/sample) and precisions. To assess how treelines could respond to global change new high-resolution studies emphasizing macrofossil analysis are urgently needed. The importance is underscored by recent studies suggesting that polar and Alpine treelines are probably already responding to global warming over recent decades (Kullman, 2002; Holtmeier and Broll, 2005).

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See also: Plant Macrofossil Introduction. **Plant Macrofossil Methods and Studies:** Surface Samples, Taphonomy, Representation.

References

- Arno SF and Hammerly RP (1993) *Timberline – Mountain and Arctic Forest Frontiers*, p. 304. Seattle: The Mountaineers.
- Barnekow L (1999) Holocene tree-line dynamics and inferred climatic changes in the Abisko area, northern Sweden, based on macrofossil and pollen records. *The Holocene* 9: 253–265.
- Barnett C, Dumayne-Peaty L, and Matthews JA (2001) Holocene climatic change and tree-line response in Leirdalen, central Jotunheimen, south central Norway. *Review of Palaeobotany and Palynology* 117: 119–137.
- Bergman J, Hammarlund D, Hannon G, Barnekow L, and Wohlfarth B (2005) Deglacial vegetation succession and Holocene tree-limit dynamics in the Scandes Mountains, west-central Sweden: stratigraphic data compared to megafossil evidence. *Review of Palaeobotany and Palynology* 134: 129–151.
- Birks HH (2001) Plant macrofossils. In: Smol JP, Birks HJB, and Last WM (eds.) *Tracking Environmental Change Using Lake Sediments*, vol. 3, pp. 49–74. Dordrecht: Kluwer.
- Birks HH (2003) The importance of plant macrofossils in the reconstruction of Lateglacial vegetation and climate: Examples from Scotland, western Norway, and Minnesota, USA. *Quaternary Science Reviews* 22: 453–473.
- Carrara PE, Trimble DA, and Rubin M (1991) Holocene treeline fluctuations in the Northern San-Juan mountains, Colorado, USA, as indicated by radiocarbon-dated conifer wood. *Arctic and Alpine Research* 23: 233–246.
- Dahl SO and Nesje A (1996) A new approach to calculating Holocene winter precipitation by combining glacier equilibrium-line altitudes and pine-tree limits: A case study from Hardangerjøkulen, central-southern Norway. *The Holocene* 6: 381–398.
- Eide W, Birks HH, Bigelow NH, Peglar SM, and Birks HJB (2006) Holocene forest development along the Setesdal valley, southern Norway, reconstructed from macrofossil and pollen evidences. *Vegetation History and Archaeobotany* 15: 65–85.
- Ellenberg H (1996) *Vegetation Mitteleuropas mit den Alpen in ökologischer Sicht*, p. 1096. Stuttgart: Ulmer.
- Fall PL (1997) Timberline fluctuations and late Quaternary paleoclimates in the Southern Rocky Mountains, Colorado. *Geological Society of America Bulletin* 109: 1306–1320.
- Gobet E, Tinner W, Bigler C, Hochuli PA, and Ammann B (2005) Early-Holocene afforestation processes in the lower subalpine belt of the Central Swiss Alps as inferred from macrofossil and pollen records. *The Holocene* 15: 672–686.
- Haas JN, Richoz I, Tinner W, and Wick L (1998) Synchronous Holocene climatic oscillations recorded on the Swiss Plateau and at timberline in the Alps. *The Holocene* 8: 301–309.
- Heiri C, Bugmann H, Tinner W, Heiri O, and Lischke H (2006) A model-based reconstruction of Holocene treeline dynamics in the Central Swiss Alps. *Journal of Ecology* 94: 206–216.
- Heiri O and Lotter AF (2001) Effect of low count sums on quantitative environmental reconstructions: An example using subfossil chironomids. *Journal of Paleolimnology* 26: 343–350.
- Heiri O and Millet L (2005) Reconstruction of Late Glacial summer temperatures from chironomid assemblages in Lac Lautrey (Jura, France). *Journal of Quaternary Science* 20: 33–44.
- Heiri O, Tinner W, and Lotter AF (2004) Evidence for cooler European summers during periods of changing meltwater flux to the North Atlantic. *Proceedings of the National Academy of Sciences of the United States of America* 101: 15285–15288.
- Hofmann W (1986) Chironomid analysis. In: Berglund BE (ed.) *Handbook of Holocene Palaeoecology and Palaeohydrology*, pp. 715–727. Chichester: John Wiley & Sons.

- Hofstetter S, Tinner W, Valsecchi V, Carraro G, and Conedera M (2006) Lateglacial and Holocene vegetation history in the Insubrian Southern Alps – New indications from a small-scale site. *Vegetation History and Archaeobotany* 15: 89–98.
- Holtmeier FK and Broll G (2005) Sensitivity and response of northern hemisphere altitudinal and polar treelines to environmental change at landscape and local scales. *Global Ecology and Biogeography* 14: 395–410.
- Kaltenrieder P, Tinner W, and Ammann B (2005) Zur Langzeitökologie des Lärchen-Arvengürtels in den südlichen Walliser Alpen. *Botanica Helvetica* 115: 137–154.
- Körner C (1999) *Alpine Plant Life*, p. 338. Berlin: Springer.
- Kremenetski CV, Sulerzhitsky LD, and Hantemirov R (1998) Holocene history of the northern range limits of some trees and shrubs in Russia. *Arctic and Alpine Research* 30: 317–333.
- Kullman L (1995) Holocene Tree-Limit and Climate History from the Scandes-Mountains, Sweden. *Ecology* 76: 2490–2502.
- Kullman L (2002) Rapid recent range-margin rise of tree and shrub species in the Swedish Scandes. *Journal of Ecology* 90: 68–77.
- Luckman BH (1994) Evidence for climatic conditions between ca. 900–1300 A.D. in the southern Canadian Rockies. *Climatic Change* 26: 171–182.
- MacDonald GM, Velichko AA, Kremenetski CV, et al. (2000) Holocene treeline history and climate change across Northern Eurasia. *Quaternary Research* 53: 302–311.
- Miller NG and Spear RW (1999) Late Quaternary history of the alpine flora of the New Hampshire White Mountains. *Géographie Physique et Quaternaire* 53: 137–157.
- Nicolussi K, Kaufmann M, Patzelt G, van der Plicht J, and Thurner A (2005) Holocene tree-line variability in the Kauner Valley, Central Eastern Alps, indicated by dendrochronological analysis of living trees and subfossil logs. *Vegetation History and Archaeobotany* 14: 221–234.
- Reasoner MA, Davis PT, and Osborn G (2001) Evaluation of proposed early-Holocene advances of alpine glaciers in the North Cascade Range, Washington State, USA: Constraints provided by palaeoenvironmental reconstructions. *The Holocene* 11: 607–611.
- Strasburger E, Noll F, Schenck H, et al. (1991) *Lehrbuch der Botanik*, p. 1030. Stuttgart: Gustav Fischer.
- Tinner W and Kaltenrieder P (2005) Rapid responses of high-mountain vegetation to early Holocene environmental changes in the Swiss Alps. *Journal of Ecology* 93: 936–947.
- Tinner W and Theurillat JP (2003) Uppermost limit, extent, and fluctuations of the timberline and treeline ecocline in the Swiss Central Alps during the past 11,500 years. *Arctic Antarctic and Alpine Research* 35: 158–169.
- Tinner W, Ammann B, and Germann P (1996) Treeline fluctuations recorded for 12,500 years by soil profiles, pollen, and plant macrofossils in the central Swiss Alps. *Arctic and Alpine Research* 28: 131–147.
- Tobolski K and Ammann B (2000) Macrofossils as records of plant responses to rapid Late Glacial climatic changes at three sites in the Swiss Alps. *Palaeogeography Palaeoclimatology Palaeoecology* 159: 251–259.
- Whitlock C (1993) Postglacial vegetation and climate of Grand Teton and Southern Yellowstone National Parks. *Ecological Monographs* 63: 173–198.
- Wick L and Tinner W (1997) Vegetation changes and timberline fluctuations in the Central Alps as indicator of Holocene climatic oscillations. *Arctic and Alpine Research* 29: 445–458.

Use in Environmental Archaeology

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Introduction

Environmental archaeology concerns the study of vegetation (flora) and animals (fauna), which lived in association with people of the past, and the way in which humans interacted with these other living organisms (Wilkinson and Stevens, 2003). The study of once-living organisms in archaeology is termed bioarchaeology and forms part of the 'archaeological sciences' (Figure 1).

Within these, archaeobotany or paleoethnobotany is the study of ancient plant remains (as proxy-data) preserved on, or in association with, archaeological sites ('on-site data'; Kreuz, 1995; article of Barker in Albarella, 2001; Jacomet and Kreuz, 1999; Pearsall, 2000). This is a contrast to paleoenvironmental approaches because plant remains from archaeological sites are 'ecofacts' and there as a result of human action. As paleoeconomic studies are carried out on archaeological material that happens to be biological, the approach is a combination of archaeology and the biological sciences. It is for this reason that developments in archaeological theory have much greater relevance to paleoeconomic studies (Wilkinson and Stevens, 2003). Today therefore, environmental archaeologists are both archaeologists and natural scientists and should have trained in both areas.

Environmental archaeology is mainly interested in the activities carried out by past populations (Tables 1 and 2). It usually covers the time period in which modern humans interacted somehow with their environment (ca. the last 40 000 years). Here again, the period in which sedentarism arose – at around 10 000 calendar years (cal yr) ago in different parts of the world (Diamond, 2002) – is by far the best represented. This covers mainly the present interglacial period, the Holocene (Table 3).

The timescale used in this article is that which is most familiar to archaeologists, namely calendar years AD resp. BC. Dating is partly based on calibrated radiocarbon dating and partly on dendrochronology.

Methodological Basics

Types of Remains

The most important macroremains (average size >0.1–0.2 mm) from archaeological sediments are seeds, fruits, parts of infructescences (like cereal chaff; Figures 2 and 3), and wood. In certain cases they are found as imprints (Figure 4) or in 'products' such as bread or coprolites. Theoretically every part of a plant can be found as an ecofact. Although in the following we deal only with macroremains, it should be emphasized that also microremains (pollen, fungal spores, etc.) in archaeological structures are ecofacts and their spectra and amounts are to a high degree a result of

human action. For more details see Dimbleby (1985); Pearsall (2000); and Jacomet and Kreuz (1999). (see also Table 6).

Taphonomy

Archaeologists use the term taphonomy to explain the processes that lead to the preservation (often fossilization) of biological remains (Schiffer, 1987; Jacomet and Kreuz, 1999, figure 4.8, p. 79). There are two aspects to consider when evaluating archaeological data: structural and depositional evidence. Structures are what people construct (buildings, fences, wells, ditches, and pits). Deposits result either from deliberate human action, as for example storage of grains in pits or in disposing of rubbish, or they may result from indirect processes that are not intentional by humans, such as silting of ditches. Deposits will contain the vast majority of biological remains that are studied by environmental archaeologists. These remains belong to different categories depending on how they reached the site (Figure 5). Negative features (like pits) commonly become infilled following their abandonment (Figures 6 and 7).

The processes by which a site is buried will have a major impact on how well archaeological and biological remains survive. Generally, material will be best preserved when burial is rapid and the energy levels by which the burying medium is deposited are low. A good example is the Roman town of Pompeii which was very rapidly buried by the eruption of the volcano Vesuvius in AD 79, but we can find rapid burial also in Neolithic lake-shore and mire settlements in the lowlands surrounding the Alps (Figure 8). The converse of the rapidity/energy correlation is also true: slow burial and high energies make for poor preservation and biological preservation is, therefore, likely to be minimal (Schiffer, 1987).

Preservation

Preservation of plant material from excavations may be different from that in natural sediments (Figure 9 and Table 4). Charring is the most common of all archaeobotanical preservation types. It preserves plant material after it has been subjected to burning (or smoldering), transforming the plant parts (particularly the dense or woody parts) from a carbon-based compound to almost pure carbon (Figure 10; Jacomet and Kreuz, 1999; Wilkinson and Stevens, 2003). Potentially charred material can be recovered from any archaeological site where fires have burnt or to which fire waste has been brought (so most commonly in settlement sites, but also, for example, in incineration graves). The most common type of charred plant remains is wood charcoal because wood is usually used to fuel fires. In addition, the most frequent charred remains are those resulting from the processing of food plants (like cereal grains, chaff, and the weed seeds that were accidentally harvested). Charred

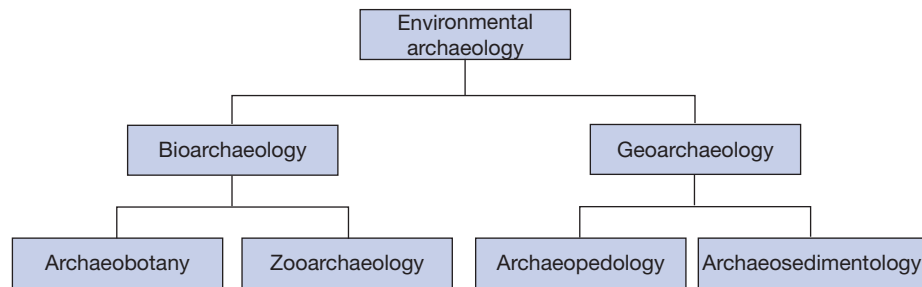


Figure 1 Subdivisions of Environmental Archaeology. Reproduced with permission from Wilkinson and Stevens 2003, p. 17, figure 3.

Table 1 The stages of food-provision, indicating for each phase the processes involved, the location where activities take place, and the key influencing cultural factors. The precise order of stages, particularly those between procurement and consumption can vary. Also the preparation stage may occur in more than one step. For example, raw resources may undergo preliminary modification prior to storage or distribution

<i>Processes</i>	<i>Stages</i>	<i>Location</i>	<i>Dominant cultural factors</i>
Collecting/ hunting/ fishing	Procurement	Hunting/gathering/ fishing area	Economics: primary production; work organization; Technology of food production.
Cultivating/growing/ animal husbandry		Farm/fishery/garden	Belief systems: what is/is not food.
Allocating/storing	Distribution	Granary/market	Politics: rent/tribute/tax/potlatch; divisions in domestic unit; decisions-on seed, sale, consumption; control of resources.
Processing	Modification/ Preparation	Inside/outside dwelling	Social: division and stratification of labour.
Eating	Consumption	Kitchen/other areas Table/eating area	Economics: technology of preparation. Social: how organized; who participates; what is served.
Clearing up	Disposal	Eating area/kitchen/ rubbish area	Belief systems: allocation of particular foods; prohibitions. Social: what is disposed of vs. consumed. Belief systems

Reproduced with permission from Samuel D (1999). Bread making and social interactions at the Amarna Workmen's village, Egypt. *World Archaeology* 31: 121–144 (Taylor and Francis Ltd, <http://www.tandf.co.uk/journals>).

Table 2 The archaeological and supporting data which relate to each stage of food-provision. The table shows the potential abundance of the evidence which can be applied to the study of ancient food systems

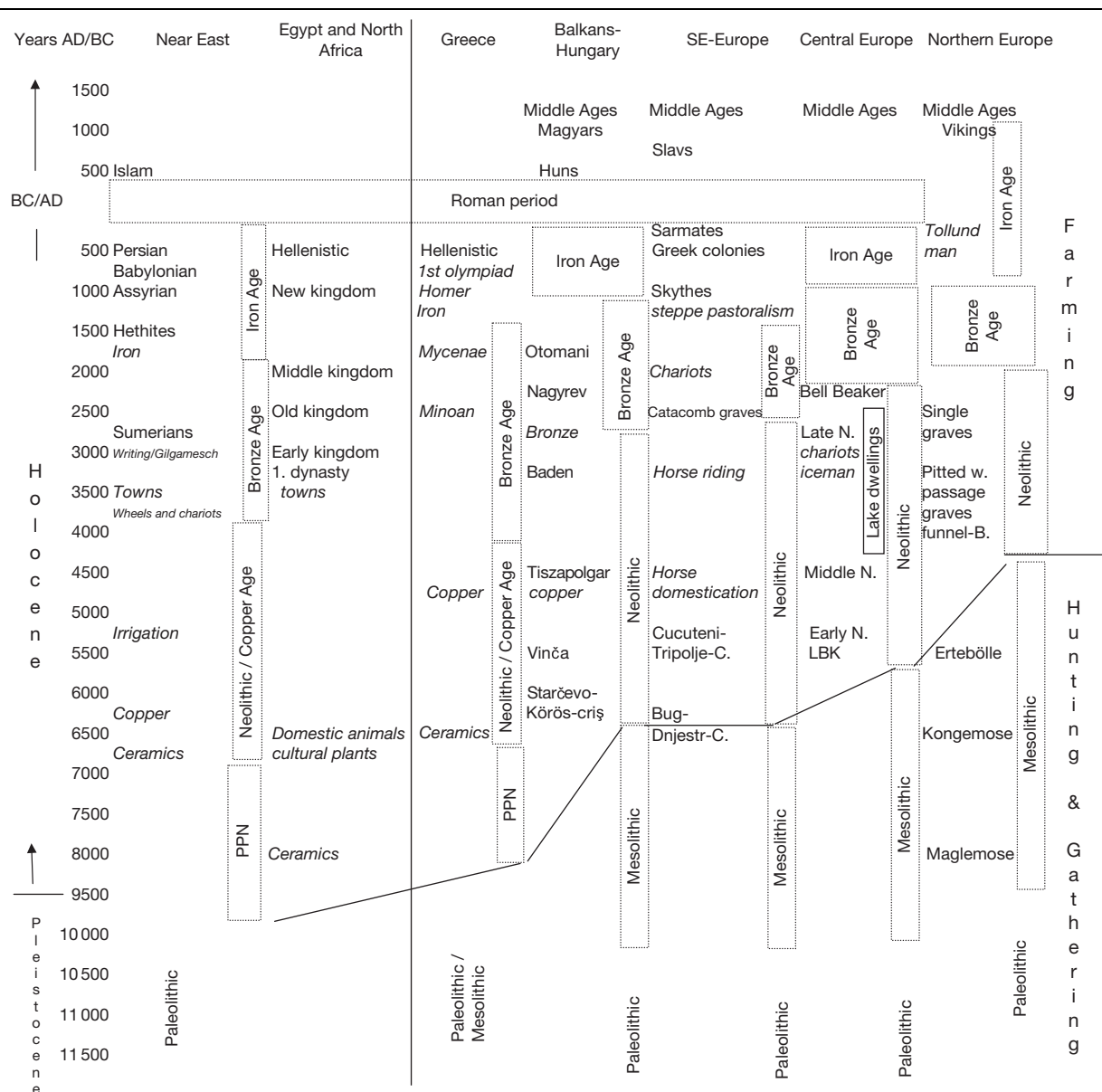
<i>Food provision stage</i>	<i>Types of material remains</i>	<i>Supporting data</i>
Procurement	– plant remains – animal remains – resource procurement tools – human remains	– ethnography – experimental replication – biology
Storage	– silos – bins – storage vessels	– ethnography
Preparation	– hearths, ovens – other preparation installations – tools e.g. mortars, querns, flint & metal blades – vessels: pottery, metal, basketry – house layouts – food residues and remains – butchery marks	– ethnography – experimental replication – biology
Consumption	– vessels – house layouts – gut contents: known individual, short time – coprolites: unknown individual(s), short time – cess: known/unknown individual/group, variable time span – human remains: individuals, variable span	– ethnography – experimental replication (for digestive processes) – biology

(Continued)

Table 2 (Continued)

Food provision stage	Types of material remains	Supporting data
Disposal	Remains disposed of: – vessels – preparation by-products – leftovers Condition of remains – e.g. whole vs. cracked bones Disposal patterns within/around settlement	– ethnography

Reproduced with permission from Samuel D (1999). Bread making and social interactions at the Amarna Workmen's village, Egypt. *World Archaeology* 31: 121–144 (Taylor and Francis Ltd, <http://www.tandf.co.uk/journals>).

Table 3 Chronology table: The last ca. 10 000 years are covered by most of the archaeobotanical investigations

Compiled by S. Jacomet © IPNA Basel University, based partly on Bogucki P and Crabtree PJ (eds.) 2004. *Ancient Europe 8000 B.C. – A.D. 1000. Encyclopedia of the barbarian world, Vol. 1, The Mesolithic to the Copper Age (c. 8000–2000 B.C.)*.



Figure 2 (a) Taking samples on an excavation (Biesheim-Kunheim, Alsace, France, Roman period, mainly 1st/2nd century AD): visible is a pit lined with wooden planks and filled with rubbish and cess, the preservation is waterlogged in the deeper areas of the pit (brownish sediments with high organic content). In the foreground, the wooden planks are sampled, in the background a white bucket is visible in which the soil samples are filled. (b–d) Examples of subfossil macroremains from the same site. (b) seed of *Vaccaria pyramidata*. (c) seed of *Agrostemma githago*. Both are weeds of winter-sown crops. As far as we know, *Vaccaria* was introduced by the Romans to the northern parts of Europe, whereas *Agrostemma* reached this region in late Neolithic times. (d) olive-stone (*Olea europaea*). Olives were imported from the Mediterranean in Roman times. Photographs made by S. Jacomet (a) and G. Haldimann (b, c and d), © IPNA Basel University.

plant parts belong to different classes, according to the nature of their deposition (Table 5). Identifying which type is present on a site is a prerequisite in interpreting charred plant remains.

Mineralization usually occurs when minerals carried in solution are deposited around plant cell surfaces or within inner voids, effectively encasing the plant structure (Wilkinson and Stevens, 2003; Jacomet and Kreuz, 1999). The most common form of mineralization on archaeological sites is as a result of calcium phosphate precipitated from cess. Therefore, mineralized remains are commonly found in middens, cesspits, latrines, sewerage systems, and comparable conditions and help to reconstruct what people were eating (Figure 11). However, mineralization by other materials such as the metals bronze and iron have also been recorded. Note that plant remains are differentially preserved by mineralization, and some remains are not preserved at all. Those of edible species whose seeds and pips are often ingested and so associated with cess are overrepresented; examples are seeds of figs, grapes, blackberries/raspberries, *Apiaceae* (like fennel or coriander), and *Rosaceae* (like plum and apple pips). However, cereal grains are rarely preserved by this mechanism.

Waterlogged plant remains are recovered from archaeological features that contained water when they were dug, such as wells, cesspits, sewers, moats, and ditches (Figure 7). But also whole layers may be preserved under the water table, like in the case of the circumalpine Neolithic and Bronze Age lake dwellings (Figure 8). Waterlogging has the potential to preserve a much greater range of plant material than where plant material

is preserved ‘only’ by charring or mineralization (Figure 9 and Table 4).

Plant remains preserved by desiccation are commonly found in arid situations like deserts, but also in totally dry caves or rock shelters or inside houses (e.g., in false ceilings; Jacomet and Kreuz, 1999). Desiccation occurs when moisture levels are too low for the organisms responsible for organic decomposition to survive. Desiccation can also preserve remains that are not otherwise found in the archaeobotanical record, including whole fruits, flowers, leaves, vegetables (e.g., onion skins in graves of Pharaonic Egypt).

In frozen environments, low temperatures render most decomposing organisms inactive, therefore well-preserved plant remains are common finds. Probably the most famous such find is the Iceman from the Italian/Austrian border near the Similaun Glacier in a height of over 3200 m above sea level (Dickson et al., 2003). As with other examples of burials, the information from frozen bodies is often personal and concerns the last meal of the deceased, his last journey, the use of plant material for clothing, and artifacts (Figure 12). However, the Iceman is preserved with his every-day equipment and clothing, thus shedding light on his lifestyle, in contrast to graves that are specially prepared ‘death assemblages’.

Types of Plant Assemblages, Sampling

Most of the assemblages found on archaeological sites are death assemblages (thanatocoenoses), containing materials from

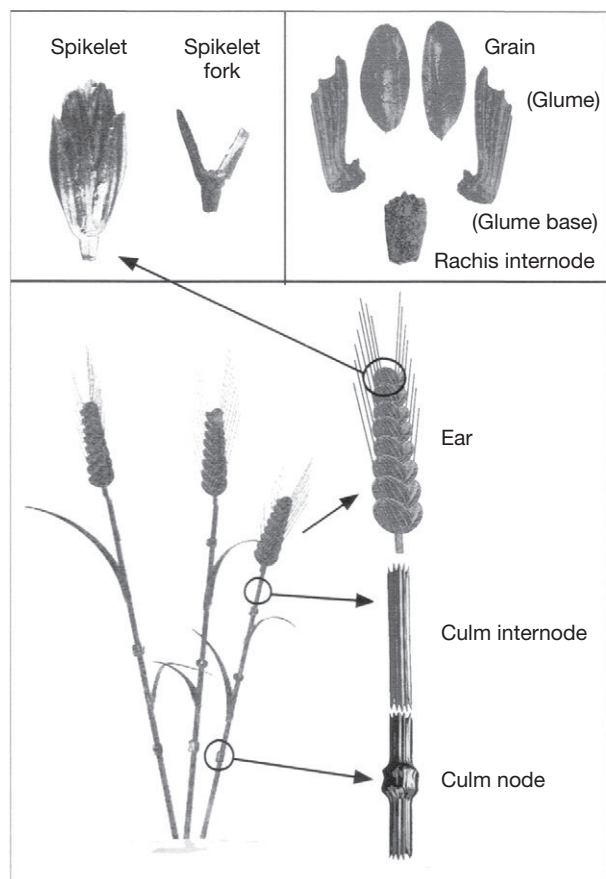


Figure 3 Cereal remain types, the example shows the component parts of hulled wheat. Reproduced with permission from [Wilkinson and Stevens 2003](#), figure 62, p. 159.

different places around a site and altered by human (or animal) action. Only rarely paleobiocoenoses may be preserved, for example an uncleaned harvest in a burnt layer just after harvesting.

Until it is understood how an archaeological deposit formed, and how it was modified since its original placement, none of the paleoecological interpretation models can be used. Before these ideas can be applied we have to be able to assess the effects of processes operating during deposition and following burial, on the preservation, and integrity of archaeobiological assemblages (i.e., the likelihood of all material at a particular level having been deposited at the same time).

Knowledge of the formation of archaeological plant assemblages is especially important at the stage of sampling. Plant macroremains on excavations are recovered from bulk samples of often many liters of volume, depending on the find densities of the structures excavated ([Figure 2a](#); for more details see [Jacomet and Kreuz, 1999](#); and [Wilkinson and Stevens, 2003](#)). Samples should not be taken where their purpose is not understood or contexts are mixed or disturbed. Sampling of the deposits finally has to take into account the spatial distribution of the material, to look at variation across a site or a structure (spatial sampling; [Figure 13](#)). For mummified or frozen burials/bodies, small samples from the stomach or clothing may be taken (e.g., see Holden in [Brothwell and Pollard, 2001](#) or [Dickson et al., 2003](#)).



Figure 4 Impression of a cereal ear at the bottom of a pot. The example shows a tetraploid naked wheat (*Triticum durum* Desf./*turgidum* L.) or emmer (*T. dicoccum* Schübl.) from the Neolithic lake shore settlement Arbon Bleiche 3 (3384–3370 BC) at Lake Constance (Switzerland). Similar remains may also be preserved in burnt daub. The original biological fragment has been combusted during firing. Reproduced with permission from [Jacomet S, Leuzinger U and Schibler J \(2004\). Die neolithische Seeufersiedlung Arbon Bleiche 3. Umwelt und Wirtschaft. \(Archäologie im Thurgau 12\). Frauenfeld: Amt für Archäologie des Kantons Thurgau. Photograph made by D. Steiner, © AATG.](#)

Recovery of the Plant Remains, Analysis, Identification, and Quantification

In several textbooks and papers the recovery of plant remains is described ([Figure 14](#); [Pearsall, 2000](#); [Jacomet and Kreuz, 1999](#)). For the recovery of charred remains flotation may be the suitable procedure, however, wet-sieving (combined with a wash-over-technique, see [Kenward et al., 1980](#)) may also be appropriate. A pretreatment is necessary when the sediments contain clay (e.g., slow drying under low temperatures in a drying chamber and a subsequent soaking in water or deep-freezing and thawing). Sieving of waterlogged remains must be performed carefully to prevent the destruction of plant material. The best method is the use of the wash-over-technique with a gentle water jet. For strongly compacted organic sediments experiments showed that deep-freezing and subsequent



Figure 5 Reconstructed farm houses ('pile-dwellings') of the 4th millennium BC at the open air Museum in Unteruhldingen (Germany): Primary material consists of building materials and tools. Secondary material would include charcoal from fires and by-products from construction, while tertiary materials would consist of re-worked primary and secondary material. Photograph © S. Jacomet.

thawing gives excellent results, whereas the use of potassium hydroxide should be avoided because it destroys fine organic parts (Figure 15). In addition, the smallest sieve mesh should be not larger than 250–300 μm . In some cases (high density of desiccated plant remains or charred seeds) the samples may be put straight under the microscope for identification without processing. Those containing some sediment may be first dry sieved. For information about analysis techniques (including subsampling) see the textbooks of Jacomet and Kreuz (1999) and Pearsall (2000). For identification see Körber-Grohne in van Zeist et al. (1991). For the identification of wood see Neumann et al. (2001) and Schweingruber (1990). Of particular importance is a good reference collection. After identification, the grains, seeds, chaff, nutshell fragments, tubers, etc., are quantified for each sample that was investigated. Before a reliable quantification is possible, units which are counted have to be defined (e.g., one spikelet fork of a glume wheat counts as two glume bases). The resultant data are tabulated on a sample-by-sample basis with individual counts for each identified component. The best way to store data is a relational database, which forms also the basis for evaluations (e.g., see Kreuz and Schäfer, 2002).

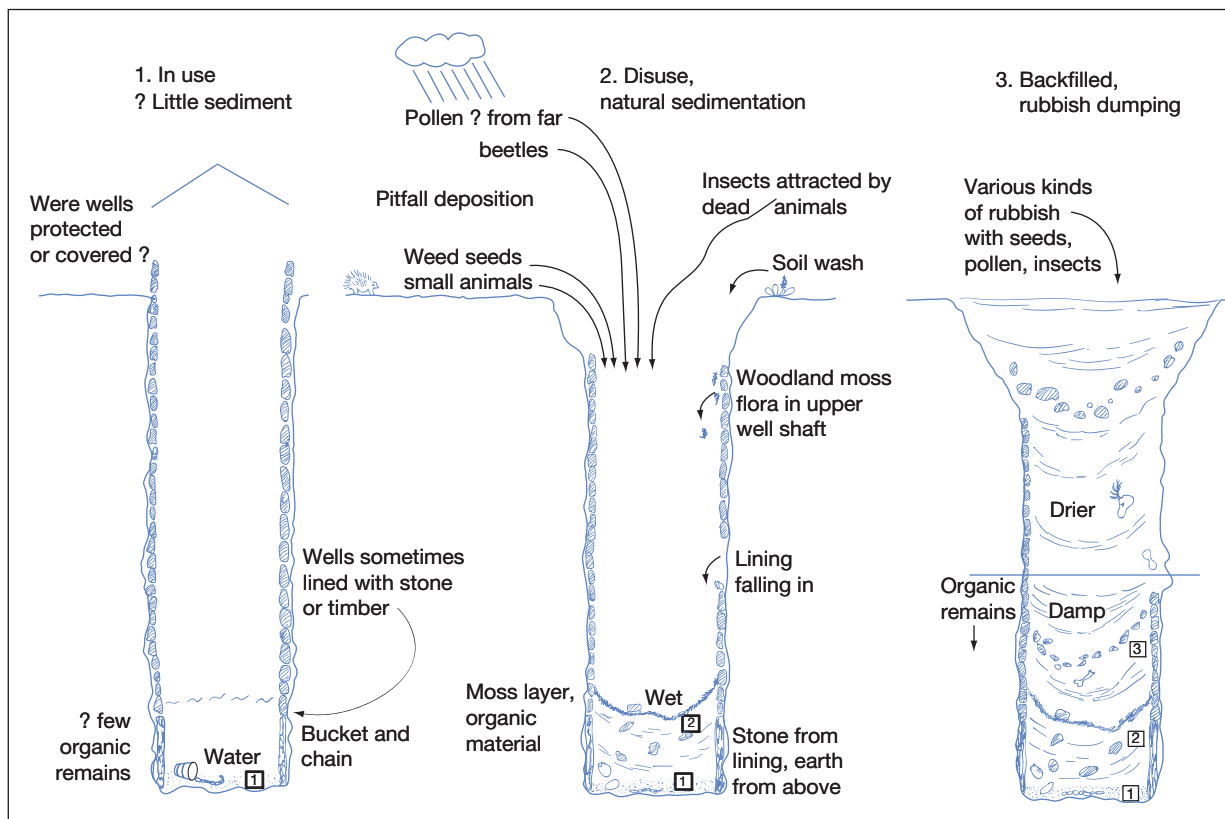


Figure 6 Sediments which accumulate within negative features left open at abandonment appear to form three distinct stages, here the example of the filling history of a Roman well: The primary fill forms from the weathering of the sides and the contemporary surface soil, very soon after the feature is dug (and is thus the deposit that could usefully be sampled in order to reconstruct the environment at the time of construction). The secondary fill forms after abandonment from material either blowing in, or eroding from, a contemporary soil. Tertiary fills are often deliberately filled by humans with rubbish, dung etc., e.g. to level the feature; they may be plus/minus contemporary to the secondary fill, or much later than the original feature. Reproduced with permission from Greig J (1988). The interpretation of some Roman well fills from the midlands of England. In: Küster H (ed.) *Der prähistorische Mensch und seine Umwelt. Festschrift für Udelgard Körber-Grohne zum 65. Geburtstag*, (Forschungen und Berichte zur Vor- und Frühgeschichte in Baden-Württemberg Vol. 31), p. 367–378. Stuttgart: Konrad Theiss Verlag.



Figure 7 Wooden figure of a he-goat, made of oak-wood from a late Celtic (Iron-Age) pit at Fellbach-Schmiden (Baden-Württemberg, Germany). Reproduced with permission from Körber-Grohne U (1999). Der Schacht in der keltischen Viereckschanze von Fellbach-Schmiden (Rems-Murr-Kreis) in botanischer und stratigraphischer Sicht. In: Wieland G (ed.) *Die keltischen Viereckschanzen von Fellbach-Schmiden und Ehningen*, (Forschungen und Berichte zur Vor- und Frühgeschichte in Baden-Württemberg Vol. 80), p. 85–149. Stuttgart: Konrad Theiss Verlag.

Interpreting Proxy-Data from Archaeological Sites

There are many differences in how those who study paleoenvironment and those studying paleoeconomy interpret their data. For the latter, it is of fundamental importance to compare sample data spatially and temporally both across a site, and between sites by examining the proportions of key components. Comparison may be made by using densities, percentages, ubiquities, or more complex statistical techniques (see the textbooks of Jacomet and Kreuz, 1999; Pearsall, 2000; Wilkinson and Stevens, 2003 and Jones 1987; for different examples see the proceedings volumes of the International Work Group for Paleoethnobotany conferences like Behre and Oeggl, 1996; Buxó et al., 2005; and Jacomet et al., 2002). Also worth mentioning is the concept of indicator groups (Hall and Kenward, 2003). In doing an evaluation it is important that samples are examined by class (Table 5), considering taphonomical processes.



Figure 8 Parts of the buildings (wooden shingles, made of silver fir, *Abies alba*) in a destruction layer of Neolithic lakeshore settlement Arbon Bleiche 3 (3384–3370 BC), Lake Constance, Switzerland; these remains were embedded very rapidly after the site burned down, therefore they are preserved. Reproduced with permission from Jacomet S, Leuzinger U and Schibler J (2004). *Die neolithische Seeufersiedlung Arbon Bleiche 3. Umwelt und Wirtschaft*. (Archäologie im Thurgau 12). Frauenfeld: Amt für Archäologie des Kantons Thurgau. Photograph made by D. Steiner, © AATG.

Studying ancient environments based on on-site data uses similar approaches to other paleoecological studies. However, like natural environments, not all archaeological environments have modern analogs as we know from many archaeobotanical investigations of class A samples (Table 5). Especially problematic is that of premodern arable environments (G. Jones, 2002). As a result of these problems, few scientists apply uniformitarian principles rigidly. They use it as a method that can be applied when deemed appropriate rather than a piece of underlying theory which is universally applicable (e.g., see Behre and Jacomet in van Zeist et al., 1991). It also recognizes that the scale of processes that operated in the past, both spatially and temporally, may have been very different from those observed at the present. Some new approaches are mentioned below (see Table 6).

Some Examples of Results

One of the prime goals in archaeology is the deciphering of ancient society: for this, the study of past economies is of fundamental importance. By ‘economy’ archaeobotanists are usually referring to the way people organized the production of food and how foodstuffs were distributed amongst the people and used (consumption; see Wilkinson and Stevens, 2003; how economy can be understood; see Charles and Halstead, 2001, in Brothwell and Pollard, 2001).

Wild Versus Domesticated Plants

Until around 10 000 years ago humans lived as hunters and gatherers; plant fruits such as almond, pistachio, etc., but also roots and tubers were collected from the wild. Around the transition from the Late Glacial to the Holocene, in several parts of the world people began to domesticate plants (Diamond, 2002). For example in the Near East (‘Fertile Crescent’) there were many places where sedentary life began in the upper Paleolithic when

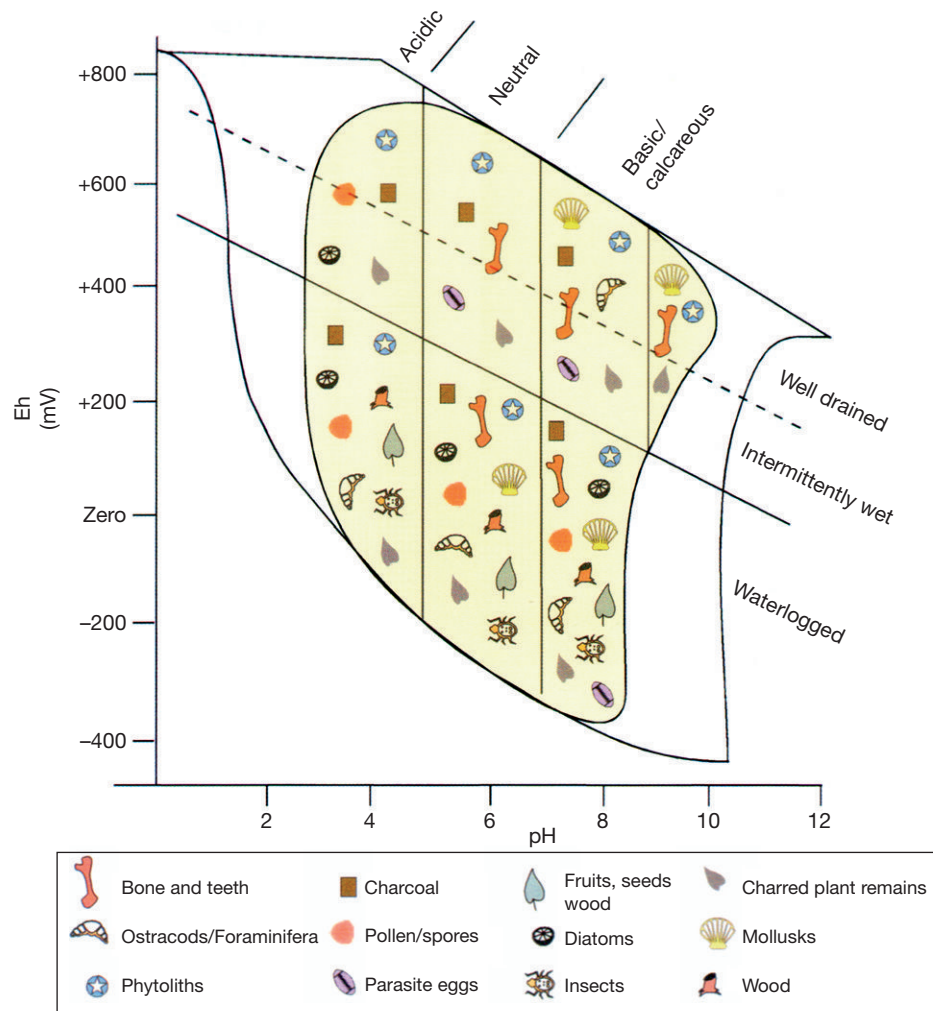


Figure 9 Schematic representation indicating under which depositional environments specific categories of environmental remains can be expected to survive and hence be recovered. If the material is preserved in more-or-less original form and only slightly altered, preservation is called sub-fossil. If the original material is replaced totally, e.g. by charring, preservation is called fossil. The former only survive under specially favorable conditions like waterlogging, the latter under very different conditions, ranging from mineral soils to waterlogged and other conditions. The filled area marks an envelope into which most sediments fit. Reproduced with permission from Jones DM (2002). *Environmental Archaeology, A Guide to the Theory and Practice of Methods, from Sampling and Recovery to Post-Excavation*. Swindon: English Heritage. © English Heritage 2002.

people still lived as hunter-gatherers (e.g., Tell Abu Hureyra in Syria; Hillman et al., 2001; Figure 16). The regular presence of domesticated plants, detectable on the basis of their morphology (Figure 17), is usually noticed in the 'Pre-Pottery-Neolithic B' stage, from 8500 cal yr BC onwards (Figure 3 and Table 7; e.g., Cappers and Bottema, 2002). Recently, new insights into plant domestication have been gained through the results of molecular biology (e.g., see Salamini et al., 2002). Many of these results point to single domestication events. However, based on archaeobotanical data, there is no evidence for a single center of origin; agriculture arose in widely separated geographic and climatic regions. What forced people to begin cultivation is a matter of debate. (see also Table 8).

History of Plant Growing and the Spread of Agriculture

First of all, in the Near East short-lived (annual) plants like cereals (different wheat and barley varieties), flax, and pulses

(pea, lentil, and others) were domesticated (Zohary and Hopf, 2000). This stock of 'founder crops' then spread westwards and reached Central Europe around 5500 cal yr BC (Figure 18). Not all the early domesticated plants reached different parts of Europe together, and the spread of single cultivars is far from being fully understood (for details see Colledge and Conolly, 2007). Concerning wheat species, the agriculture of the early Neolithic Linear Pottery Culture was based on einkorn (*Triticum monococcum*) and emmer (*T. dicoccum*). During the fifth millennium BC hexaploid naked wheat (*Triticum aestivum*) was an important crop in some settlements. In the fourth millennium BC, in the circumalpine Lake Dwellings, a tetraploid naked wheat (*Triticum durum/turgidum*, Figure 10c) played a major role. First sure traces of spelt (*Triticum spelta*) cultivation come from the end of the Neolithic period in Central Europe. This robust glume wheat then became very widespread in Europe during the Bronze Age. The same holds also for millets (*Panicum miliaceum*, *Setaria italica*), domesticated plants originating from

Table 4 Different representation of plant remains under different preservation conditions

Preservation plant parts	Preservation			
	Subfossil		Charred	
	Seeds	Other remains	Seeds	Other remains
<i>Triticum</i> (glume wheat)	+	ch sp st	+++	ch sp st
<i>Triticum</i> (naked wheat)	+	ch sp st	+++	ch sp st
<i>Hordeum</i> (barley)	+	ch sp st	+++	ch sp st
<i>Secale cereale</i> (rye)	+	ch sp st	+++	ch sp st
<i>Avena</i> (oat)	+	ch sp st	+++	ch sp st
<i>Panicum, Setaria</i> (millets)	++	ch	+++	
<i>Fagopyrum</i> (buckwheat)	+		++	
<i>Pisum sativum</i> (pea)	+	f	+++	
<i>Lens culinaris</i> (lentil)	+		+++	
<i>Vicia faba</i> (horsebean)	+	f st	+++	f
<i>Linum usitatissimum</i> (flax)	+++	f st	+	f
<i>Papaver somniferum</i> (poppy)	+++	f	+	f
Wild fruits (strawberry, blackberry...)	+++	f	+	f
Cultivated fruits (peach, cherry...)	+++	f	+	f
Vegetables	++		+	
Spices	+++	l	+	
Medicinal plants	+++	l	+	
Dye plants	+++	f	+	f
Weeds	+++	st f	+++	f
Other wild plants	+++		+	

Adapted from Willerding in [van Zeist et al. 1991](#), with permission. ch = chaff; st = straw, stems; sp = spikelets, ears etc.; l = leaves; f = fruit remains like capsule parts etc. +++ = commonly found; ++ = often found; + = rarely found. In Neolithic lake dwelling settlements e.g. over 99% of poppy seeds are preserved in uncarbonized state.



Figure 10 Some examples of carbonised plant remains: (a) Grain of emmer (*Triticum dicoccum*). (b) spikelet fork of emmer. (c) Rachis of tetraploid naked wheat (with bulbs below the insertion of the glumes; *Triticum durum/turgidum*). (d) Rachis of six-rowed barley (*Hordeum vulgare*). All remains come from the Neolithic lake shore settlement Arbon Bleiche 3 (3384–3370 BC), Lake Constance, Switzerland. Reproduced with permission from [Jacomet S, Leuzinger U and Schibler J \(2004\). Die neolithische Seeufersiedlung Arbon Bleiche 3. Umwelt und Wirtschaft. \(Archäologie im Thurgau 12\). Frauenfeld: Amt für Archäologie des Kantons Thurgau. Photographs made by G. Haldimann. © IPNA Basel University.](#)

Table 5 Charred plant remain classes, according to Hubbard and Clapham

Class A	Plant material that has burnt in the same location from which it was recovered. This category would encompass several types of burnt storages (see Figure 13), including storage pits that catch fire and hearths and kilns, where the feature and its spent fuel survive in the archaeological record. In these cases the relationship between the context and the plant remains is very strong.
Class B	Plant material that is charred during a single burning event, but where the burnt material has been redeposited, either deliberately or accidentally. The act of redeposition means that the material may not specifically relate to the context in which it was found.
Class C	Material that comes from a number of temporally and spatially distinct burning events and activities. The material from these events is mixed together with settlement waste to become incorporated in archaeological features such as middens, rubbish pits etc. The remains have little if any relationship to the context from which they are recovered.

Adapted from [Wilkinson and Stevens 2003](#), Table 14, p. 152 with permission. Class C forms the so called “background noise” after [Bakels CC \(1991\)](#). Tracing crop processing in the Bandkeramik culture. In: Renfrew JM (ed.) *New light on early farming. Recent developments in palaeoethnobotany*, p. 281–288. Edinburgh: Edinburgh University Press.

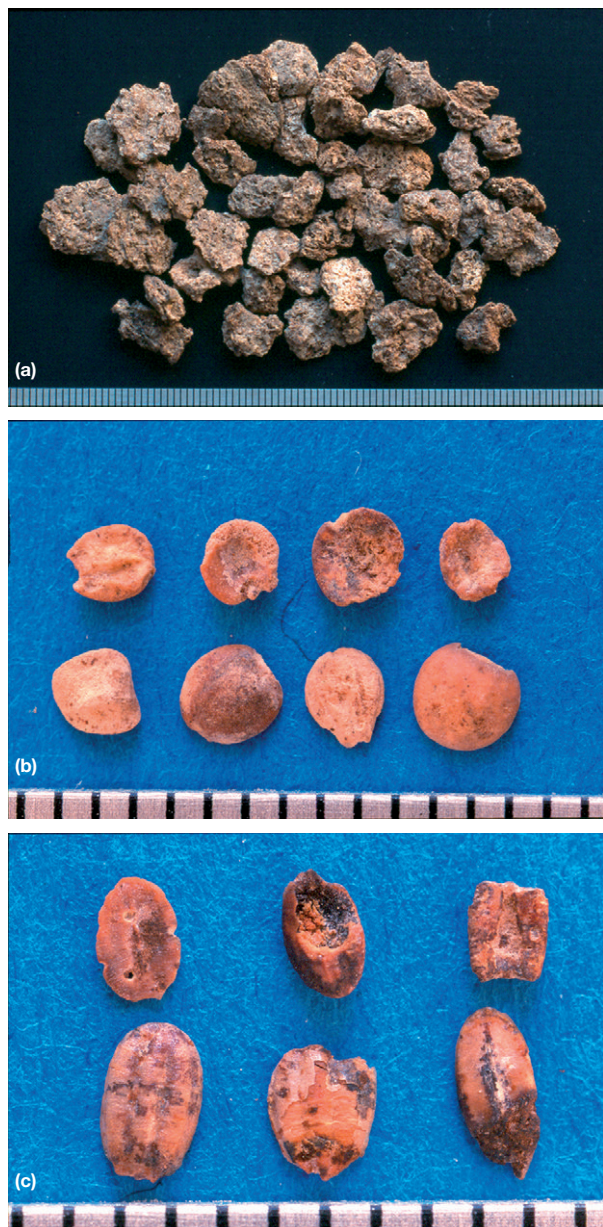


Figure 11 Mineralized cess and plant remains from Roman cesspits in the legionary camp of Vindonissa (Windisch, Switzerland), early 1st century AD. (a) fragments of cess. (b) seeds of coriander (*Coriandrum sativum*). (c) seeds of Apiaceae, probably dill (*Anethum graveolens*). Photographs U. Weber © IPNA Basel University.

China. There are large spatial and temporal differences in the importance of the single cultivated plant species as well as their diversity in the Old World ([Colledge and Conolly, 2007](#)); one of the reasons for this is probably cultural traits.

Only some 1000 years after short-lived plants, woody plants also show traits of domestication. In the Near East, from the Early Bronze Age onwards (about the fourth millennium BC) figs, vines, dates, and olives belong to the domesticates ([Zohary and Hopf, 2000](#)). However, collecting wild plants did not stop after domestication: particularly in northern parts of Europe collecting played a very important role ([Figure 19](#)) until the widespread introduction of gardening during the Roman period. (see also [Table 9](#)).

Applying Biological Evidence to Economic Questions

Processing Techniques, Storage, and Food Preparation

The aim of growing a crop is to obtain a clean product that may be further processed into food. Information about how crops were processed and stored comes from both weed seeds and remains of the cultivated plant itself. After harvest of cereals, for example, the farmer is left with a great deal of mostly undesirable weed seeds, chaff, and straw. Ethnographic work (e.g., [Hillman, 1984](#); [Figure 20](#)) has demonstrated that there are relatively few ways (in the absence of modern machinery) in which the processing sequence of cereals can be carried out. Archaeobotany successfully tried to assign charred assemblages to a single processing stage ([Figure 21](#)). However, often assemblages result from an amalgam of all of the stages needed to obtain clean grain for storage. Then the charred plant remains will only provide information on how much processing had occurred prior to that stage. So, if there were only glumes, grain, and large weed seeds – as in many Roman settlements – it may be concluded that storage was as semiclean spikelets. Alternatively, if there are large numbers of small weed seeds, straw culms, and stems it may be concluded that cereals were stored either as sheaves or as partially threshed ears. After assessing how crops were stored, scientists can begin to examine, for example, the spatial distribution of processing activities and storage. One of many examples is the investigation of a thick layer of carbonized seeds in an eleventh century AD granary from southern France ([Figure 13](#)). In the western part of the granary, naked wheat was stored in bulk. In the eastern part, various crops (at least naked wheat, barley, rye, oat, and grape) were stored in small amounts, most of which were probably separated by light, wooden structures

What the Iceman and his belongings tell us

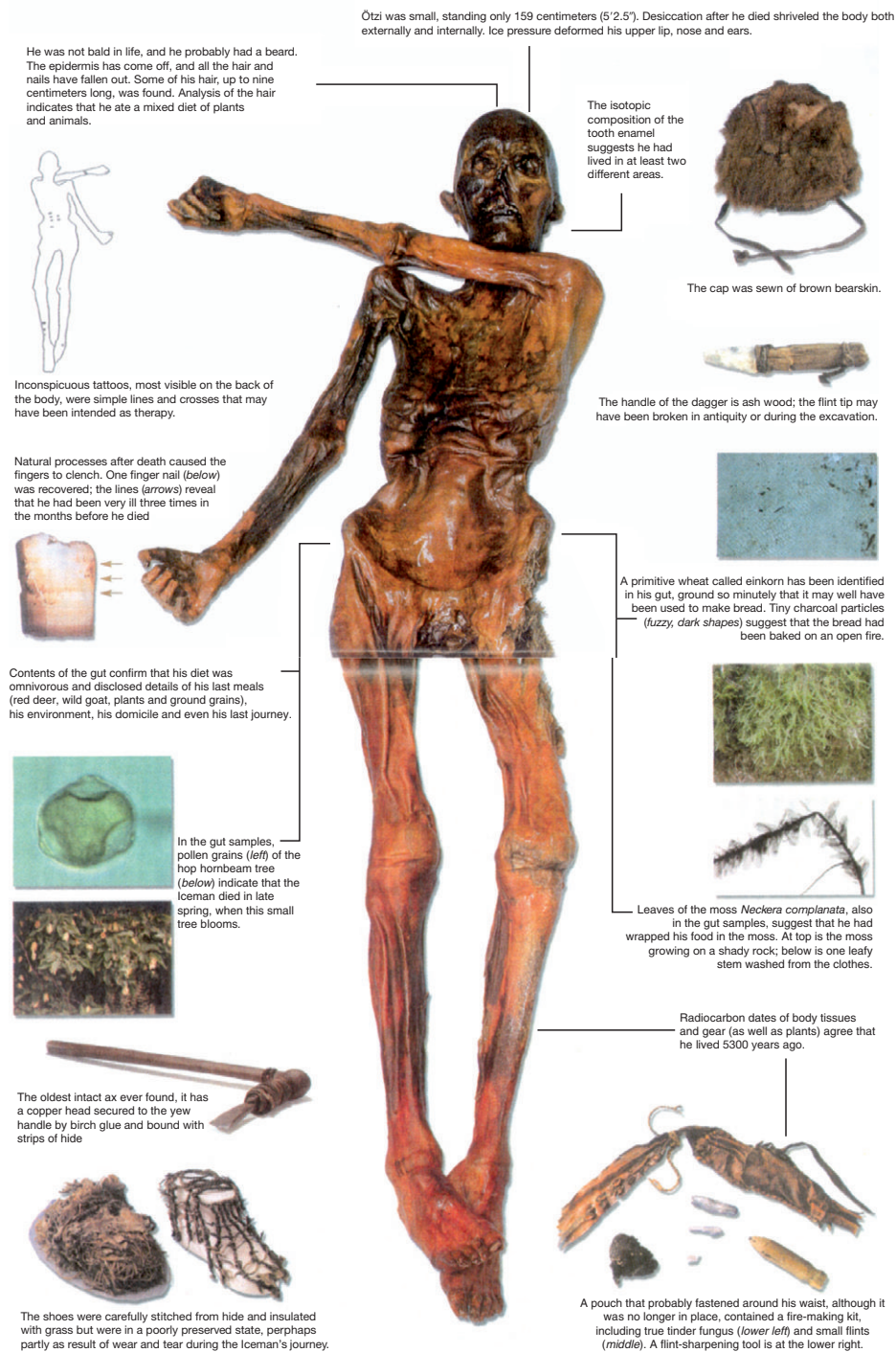


Figure 12 What the Iceman and his belongings can tell us (clothes, equipment etc.). A: The Iceman was small, standing only 159 cm. Desiccation after he died shriveled the body both externally and internally. Ice pressure deformed his upper lip, nose and ears. He was not bald in life, and he probably had a beard. Some of his hair, up to 9 cm long was found. B: The oldest ax ever found, it has a copper head secured to the yew handle by birch glue and bound with strips of hide. C: The handle of the dagger is ash wood, the sheath is made of limebast-fibers and the lace of grasses. D: Flint sharpening tool made of lime-wood. E: A pouch (made of leather), probably fastened around his waist, contained a fire making kit, including true tinder fungus (*Fomes fomentarius*) and small flints. F: The shoes were carefully stitched from hide (of red deer, the sole of bearskin), inside the net and the isolation was made of grasses. G: The cap was sewn of brown bearskin. The Iceman can be taken as representative of how Neolithic people around 3350 BC would have been dressed and the objects that they would have used. Very few of them are recovered in more conventional archaeological settings. Reproduced with permission from [Dickson et al. 2003](#) (for more details see there).

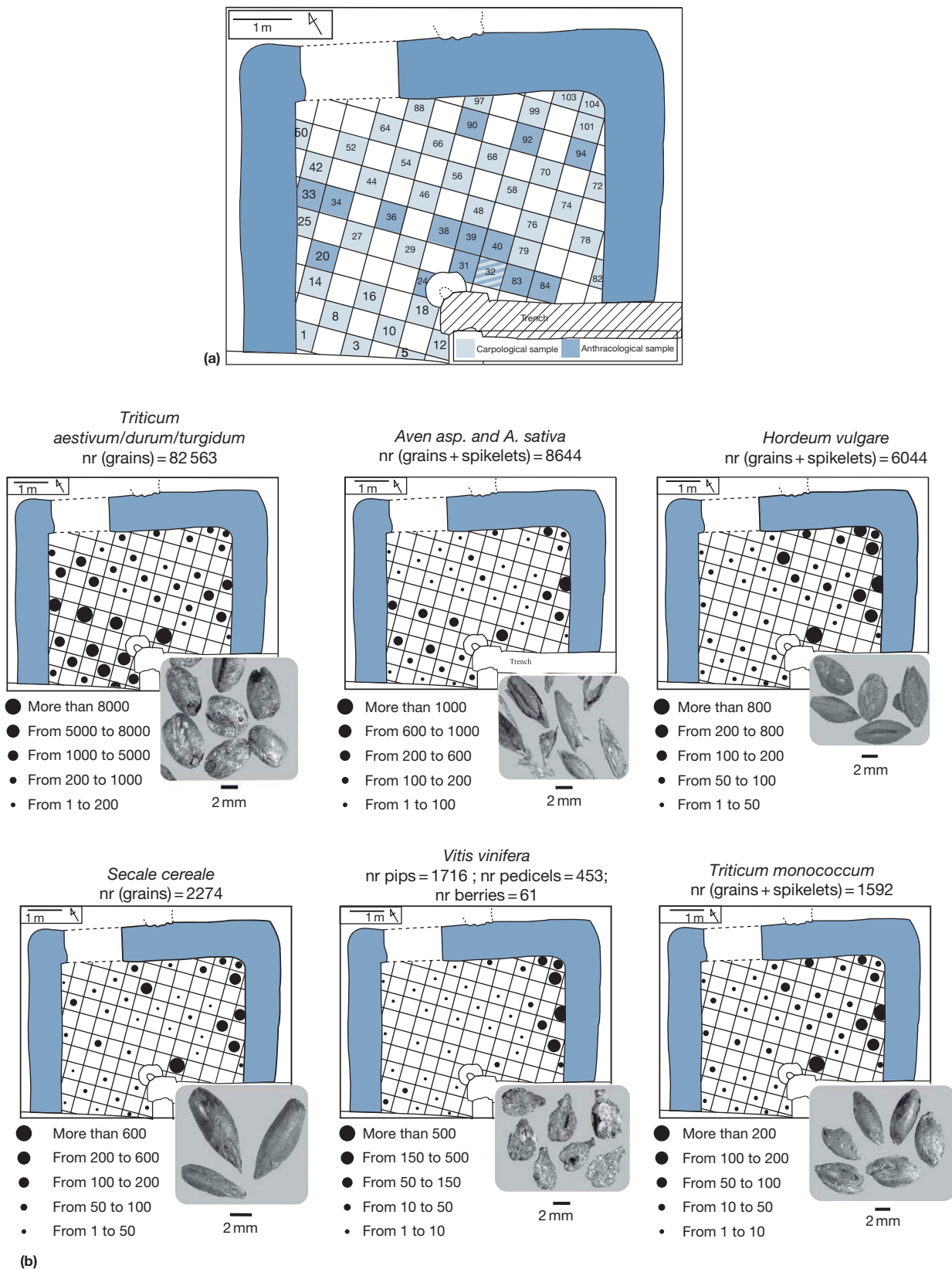


Figure 13 Example of spatial sampling and results: (a) location of the samples analyzed for seed and charcoal remains of a thick layer of carbonized material in an 11th century AD granary from southern France (La Gravette). (b) Spatial distribution of the most important foodplants in the granary (number of plant remains per liter). Reproduced with permission from Ruas M-P, Bouby L, Py V and Cazes J-P (2005). An 11th century AD burnt granary at La Gravette, south-western France: Preliminary archaeobotanical results. *Vegetation History and Archaeobotany* 14.



Figure 14 Sieving of a sample, using the wash-over technique. Photograph by D. Hager, © IPNA, Basel University.



Figure 15 Pictures of deep frozen (left side) and KOH-treated (right side) subfossil plant remains (rachises of rye, *Secale cereale*). They come from Roman (1st/2nd century AD) waterlogged deposits near Biesheim, Alsace (F). They will be published in Vanderpe P and Jacomet S (2007). Comparing different pre-treatment methods for strongly compacted organic sediments prior to wet-sieving: A case study on Roman waterlogged deposits. *Environmental Archaeology*. Photographs made by G. Haldimann, © IPNA Basel University.

Table 6 Further Reading: Methodological Basics

- Boardman, S. and Jones, G. 1990. Experiments on the Effects of Charring on Cereal Plant Components. *Journal of Archaeological Science* 17, 1–11.
- Chabal, L., Fabre, L., Terral, J.-F. and Théry-Parisot, I. 1999. L'antracologie. In: *La Botanique* (ed. C. Bourquin-Mignot, J. E. B., L. Chabal, S. Crozat, L. Fabre, F. Guibal, P. Marinval, H. Richard, J.F. Terral, I. Théry), (Collection Archéologiques, p. 43–104. Paris: Errance.
- Dincauze, D., F. 2000. *Environmental Archaeology. Principles and Practice*. Cambridge: Cambridge University Press.
- Ernst, M. and Jacomet, S. 2006. The value of the archaeobotanical analysis of desiccated plant remains from old buildings: methodological aspects and interpretation of crop weed assemblages. *Vegetation History and Archaeobotany* 15, 45–56.
- Evans, J. and O'Connor, T. 1999. *Environmental Archaeology. Principles and Methods*. Phoenix Mill: Sutton Publishing Limited.
- Germer, R. 1985. *Flora des pharaonischen Ägypten*. (Deutsches Archäologisches Institut Abteilung Kairo, Sonderschrift 14). Mainz: Ph. von Zabern.
- Green, F. J. 1979. Phosphatic mineralization of seeds from archaeological sites. *Journal of Archaeological Science* 6, 279–284.
- Hather, J. G. 1993. *An archaeobotanical guide to root and tuber identification. Vol. 1: Europe and South West Asia*. (Oxbow Monograph 28). Oxford: Oxbow Books.
- Hillman, G. C., Mason, S., de Moulins, D. and Nesbitt, M. 1996. Identification of archaeological remains of wheat: The 1992 London workshop. *Circaea* 12, 195–210.
- Hosch, S. and Jacomet, S. 2001. New aspects of archaeobotanical research in Central European Neolithic Lake Dwelling Sites. *Environmental Archaeology* 6, 59–71.
- Hosch, S. and Zibulski, P. 2003. The influence of inconsistent wet-sieving procedures on the macroremains concentration in waterlogged sediments. *Journal of Archaeological Science* 30, 849–857.
- Jacomet, S. and Brombach, Ch. 2005. Reconstructing intra-site patterns in Neolithic lakeshore settlements: the state of archaeobotanical research and future prospects. In: *WES'04 – Wetland Economies and Societies. Proceedings of the International Conference in Zurich, 10–13 March 2004*. (ed. Della Casa, P. and Trachsel, M.), (Collectio Archaeologica Vol. 10, 69–94), Zürich: Chronos.
- Kenward, H. and Hall, A. 1997. Enhancing bioarchaeological interpretation using indicator groups: Stable manure as a paradigm. *Journal of Archaeological Science* 24, 663–673.
- Kenward, H., Engleman, C., Robertson, A. and Large, F. 1985. Rapid scanning of urban archaeological deposits for insect remains. *Circaea* 3, 163–172.
- Renfrew, C. 2005. *Archaeology: The key concepts*. London: Routledge.
- Schibler, J. and Jacomet, S. 2005. Fair-weather archaeology? A possible relationship between climate and the quality of archaeological sources. In: *Climate variability and Culture change in Neolithic societies of Central Europe 6700–2200 cal BC* (ed. Gronenborn, D.), (RGZM Tagungen Vol. 1), p. 27–39. Mainz: Römisch-Germanisches Zentralmuseum.
- van der Veen, M. 2004. The merchants' diet: food remains from Roman and medieval Quseir al-Qadim. In: *Trade and Travel in the Red Sea Region. Proceedings of Red Sea Project I* (ed. Lunde, P. and Porter, A.), (BAR International Series Vol. 1269), p. 123–130. Oxford: Archaeopress.
- van der Veen, M. and Fjeller, N. R. J. 1982. Sampling Seeds. *Journal of Archaeological Science* 9, 287–298.

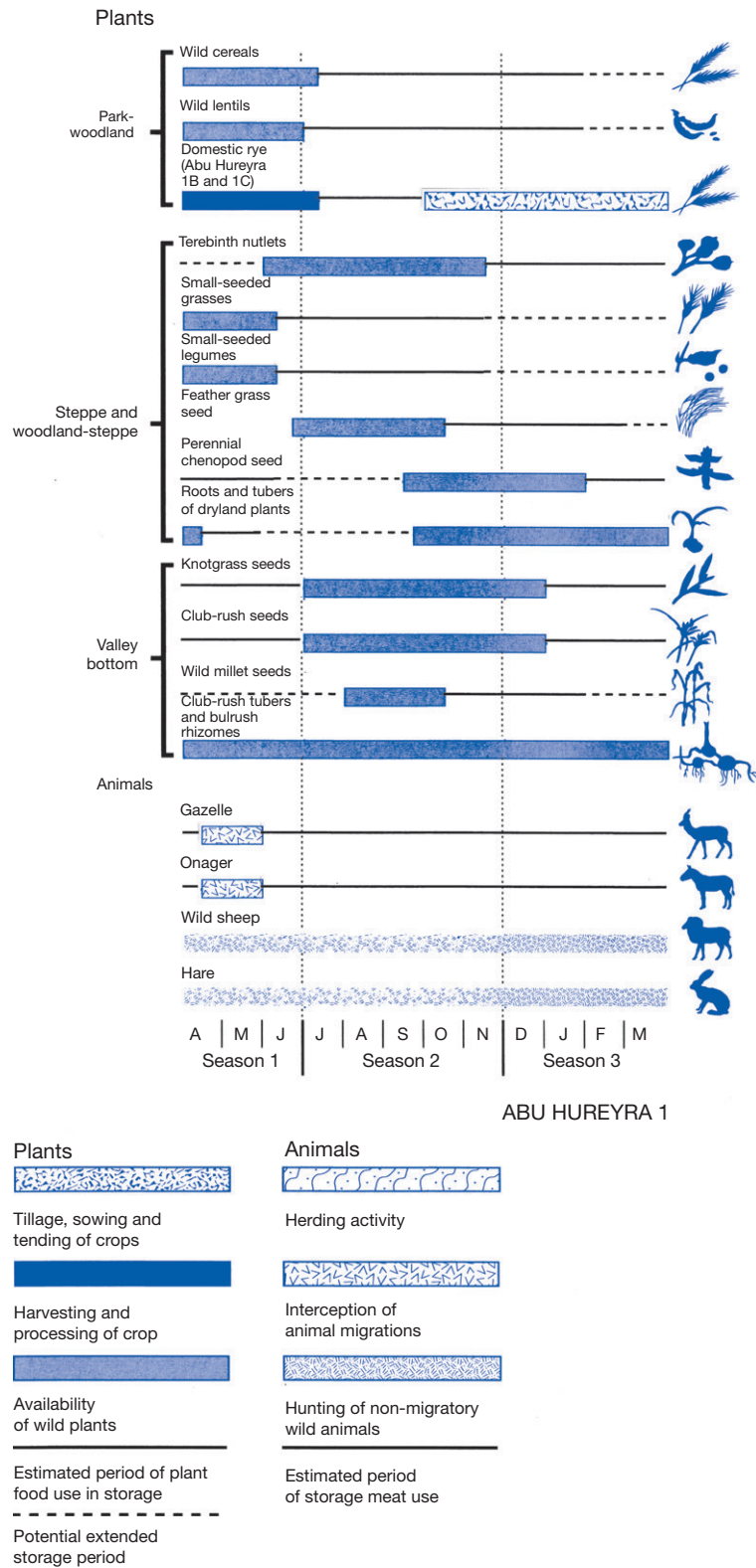


Figure 16 Tell Abu Hureyra, Syria: The seasons of availability and use of the main plant and animal foods, and crops used by the inhabitants during Phase I (before 11 000 BP uncal.), when this was a hunter-gatherer site. The investigation of the plant remains showed that the settlement was occupied year-round. Reproduced with permission from Moore AMT, Hillman GC and Legge AJ (2000). *Village on the Euphrates. From foraging to farming at Abu Hureyra*. Oxford: Oxford University Press.

made from hazelnut, maple, etc. The cereal crops had largely been processed and cleaned. The stored products probably represent taxes paid to the lord who owned the granary.

Although the procedures of processing seem to be efficient, analyses carried out on archaeological bread, dough, excrement, and the contents of the stomachs of bog bodies often show that the operations did not result in clean grain, and bits of chaff, unground grain, and weed seeds still remained in the product eaten by people. Waterlogged plant remains from many bog bodies (at least partly dating to the Late Iron Age) have a rather high amount of weed seeds in their stomachs, as well as cereal chaff; the Grauballe man from Denmark had even whole spikelets of wheat in his stomach (Holden in Brothwell and Pollard, 2001). The excrement of the workers in the salt mines of Hallstatt (Austria; Iron Age) also contained many chaff remains. (see also Table 10).

Production and Consumption

Production and consumption are an integral mesh of activities and relationships (Wilkinson and Stevens, 2003). The most important aspect of production is that of food. Almost all foodstuffs derive from plants (salt is an exception). It was once argued that sites involved in the growing (producers) of plants could be distinguished from those receiving (consumers) plant products. On the one hand it was proposed that grain-rich assemblages from Iron Age Britain represent producer sites and weed/chaff-rich assemblages consumer sites. On the other hand it was argued – based on ethnographical data – that producer sites are characterized by the waste from early stages of crop processing. Therefore, a differentiating of consumer- and producer-sites, based on archaeobotanical evidence only, is difficult. Variation in charred plant assemblages

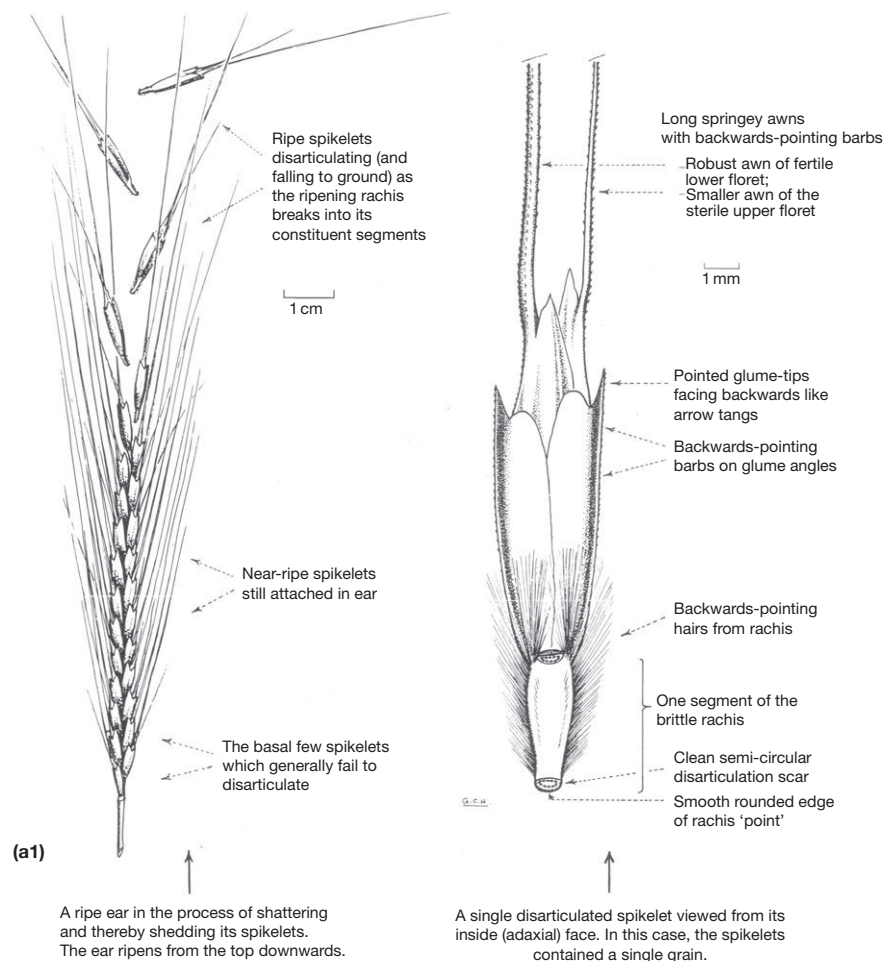


Figure 17 (a) Differences between wild and domesticated cereals. In the case of cereals, the former have brittle rachises and clean, semicircular disarticulation scars, whereas the latter show tough rachises and a rough breakage of rachis, leaving sharp corners which impede penetration. Reproduced with permission from Hillman GC and Davies MS (1992). Domestication rate in wild wheats and barley under primitive cultivation: preliminary results and archaeological implications of field measurements of selection coefficient. In: Anderson P (ed.) *Préhistoire de l'agriculture: Nouvelles approches expérimentales et ethnographiques*, (Monographie du CRA Vol. 6), p. 113–158. Paris: Editions du CNRS. (b) examples of wild (left side) and domesticated (right side) cereal grains (here carbonized specimen of the glume wheat emmer, *Triticum dicoccum*, from Cayönü, Turkey): The former are smaller, the latter larger (Reproduced with permission from Zohary and Hopf, 2000). The mentioned differences are visible in the archaeobotanical material. However, in early sites it is often difficult to decide to which group a remain should be attributed.

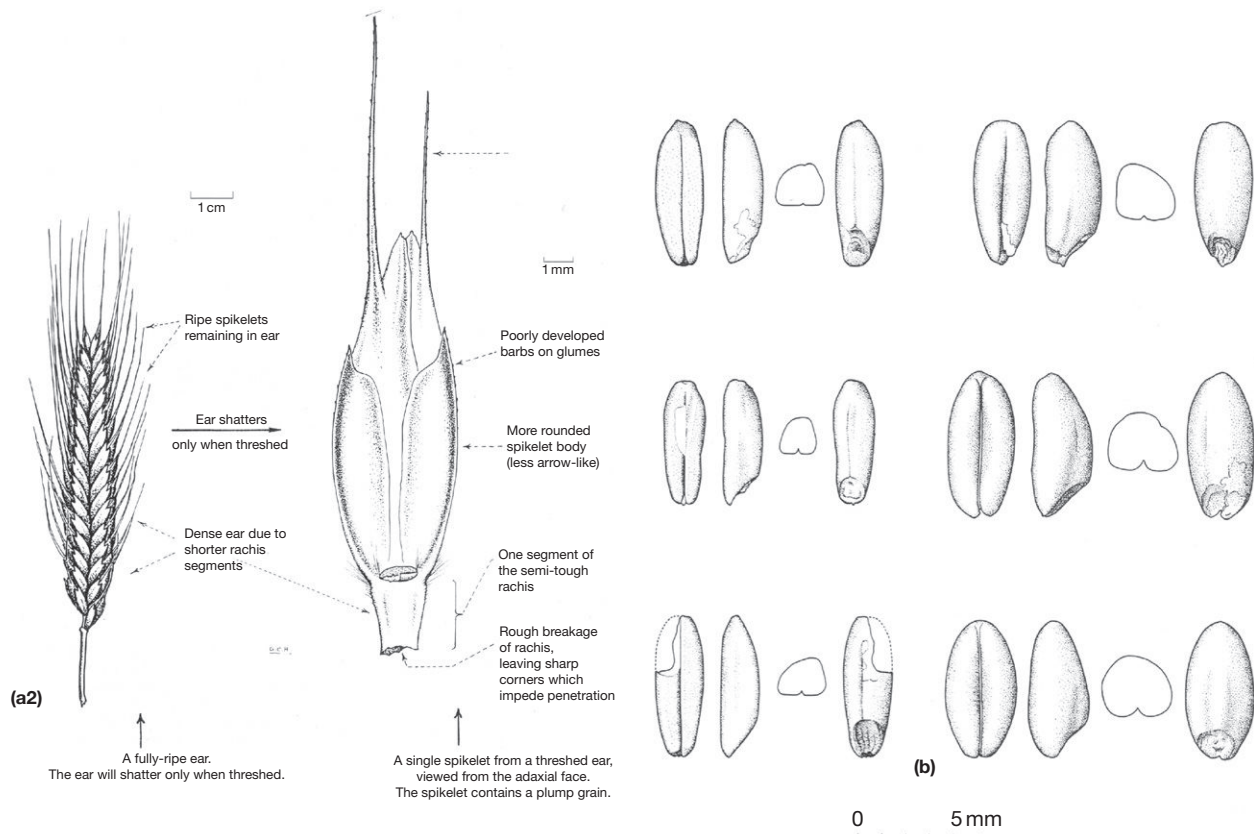


Figure 17 (Cont'd)

from the Late Iron Age in England is, for example, more likely to relate to differing storage practices, for example, of surplus-grain than consumption/production. Only when other – archaeological – evidence is found (Table 2) is a more definitive interpretation possible. However, there are also clear cases of producer units like the Roman *Villae rusticae*. (see also Table 11).

Cultivation Techniques

Much of the information about the cultivation techniques comes from weed seeds. Unfortunately, at each stage of cleaning, processing removes more and more weed seeds, so the cleaner the grain when charred, the less evidence there is of cultivation practices. In any case the first step will be to analyze from which stage of cleaning the weed spectrum derives (Figures 20 and 21). Helpful tools in interpreting weed assemblages are also experimental farming or an ethnoarchaeological approach.

Weed species may be characteristic of environmental factors such as soil types, drainage, climate, and the specific cultivation practices employed by the farmer like manuring, tillage, sowing times, and weeding. This therefore provides a proxy for determining where past crops were grown (G. Jones, 2002). Based on the application of FIBS (Functional Interpretation of Botanical Surveys; Jones et al., 2005) to Neolithic weed spectra, Bogaard (2004) concluded that in those times cereal farming was practiced on small,

intensively worked and manured plots. There are other interpretations, however, including considering shifting cultivation. Larger fields seem to become widespread only from the Metal Ages onwards.

The presence or absence of seeds of weed species from the harvest will depend on the method of harvest. For example, harvesting by sickle will remove cereals, rachises, some culms, and weeds. Harvesting low on the culm will bring in not only these weeds but also seeds of low-growing species such as clover (*Trifolium* sp.). Finds of many low-growing weeds from Late Bronze Age and Iron Age settlements in Europe suggest that harvesting was carried out with sickles or scythes (of which also evidence is found) low down the stem whereas in early Neolithic times high-growing weeds prevailed which points to totally different harvesting methods.

Archaeobotanical investigations will allow the determination of the weed species composition of an arable field to some extent. However, an interpretation of the ratio of perennial-to-annual weeds concerning the type of tillage is not as straightforward (Bogaard, 2002) as thought. Newest results point to the fact that already in Neolithic times fields were intensely worked even though they contained more species of perennial weeds (e.g., creeping buttercup *Ranunculus repens*); these reproduce from fragments of root stem and so in fact benefit by more effective tillage. Only later, for example in medieval times, were crops with dominating annual species found, pointing to a permanent cultivation

Table 7 Archaeobotanical records for cereals in the Epipaleolithic and Pre-Pottery Neolithic periods. Shaded box indicates identification based on chaff (and thus, most reliable); dotted box indicates identification based on grain alone; ? indicates uncertain identification. Abbreviations: EIN Einkorn; EM Emmer; BAR Barley; NAK WHT Free-threshing (naked) wheat; DOM domesticated.

Site (phase)	Country	Period	Date (uncal. years BP)	Crop status	Wild EIN	Wild EM	Wild BAR	Wild RYE	DOM EIN	DOM EM	NAK WHT	DOM BAR	NAK BAR	DOM RYE
Ohalo II	Israel	Epipal. (Kebaran)	19400	Wild										
Wadi al-Hammeh 27	Jordan	Epipal. (Natufian)	12200–11900	Wild										
Iraq ed-Dubb (below)	Jordan	Epipal. (Natufian)	11200–10800	?										
Hayonim Cave and Terrace	Israel	Epipal. (Natufian)	12300–10000	Wild										
Abu Hureyra (I)	Syria	Epipal. (Natufian)	11500–10000	Wild										
Mureybit (I–III)	Syria	Epipal. & PPNA	10500–9600	Wild										
Qermez Dere	Iraq	PPNA	10150–9600	Wild										
Netiv Hagdud	Israel		10000–9400	Wild										
M'lefaat	Iraq		9900–9650	Wild										
Iraq ed-Dubb (structures)	Jordan		9950	?										
Jerf al Ahmar	Syria		9800–9700	Wild										
Tell Aswad (I)	Syria		9700–9300	?										
Jericho (VIIA–X)	Palestine		9400–9200	?										
Dja'de	Syria	Early PPNB	9600–9000	Wild	?									
Wadi el-Jilat 7	Jordan		9500–9200	?										
Nevali Çon	Turkey		9250	Dom.										
Çayönü (grill-cobble)	Turkey		9200–8600	Dom.										
Nahal Hemar (3–4)	Israel		9200–8100	Dom.										
Cafer Höyük (XIII–IX)	Turkey		9200–7900	Dom.										
'Ain Ghazal	Jordan	Middle PPNB	9200–8000	Dom.										
Beidha	Jordan		9100–8550	Dom.										
Ganj Dareh	Iran		9000–8400	?										
Abdul Hosein	Iran		9000–8400	Dom.										
Abu Hureyra (2A)	Syria		9000–8300	Dom.										
Cafer Höyük (III–IV)	Turkey		9000–8500	Dom.										
Tell Aswad (II)	Syria		8900–8500	Dom.										
Asıklı Höyük	Turkey		8800–8500	Dom.										
Jericho (Tr I: XII–XXIII)	Palestine		8800–8650	Dom.										
Wadi el-Jilat 7	Jordan		8800–8500	Dom.										
Ghoreife	Syria		8800–8300	Dom.										
Jarmo	Iran		8750	Dom.										
Ali Kosh (BM)	Iran		8750	Dom.										
Halula	Syria		8700–7900	Dom.										
Tell Sabi Abyad II	Syria	Late/Final PPNB	8500–8000	Dom.										
Wadi Fidan A	Jordan		8500–8100	Dom.										
Ras Shamra (Vc)	Syria		8500–8000	Dom.										
Gritille	Turkey		8500–7800	Dom.										
Can Hasan III	Turkey		8400–7700	Dom.										
Azraq 31	Jordan		8350–8300	Dom.										
Dhuweila (I)	Jordan		8350–8200	Wild										
Tell Bougras	Syria		8350–7850	Dom.										
Tell Ramad (I)	Syria		8300–8100	Dom.										
Abu Hureyra (2B)	Syria		8300–8000	Dom.										
Wadi Fidan C	Jordan		8100–7600	Dom.										
El Kowm II - Caracol	Syria		7800–7700	Dom.										
Atlit-Yam	Israel		7700	Dom.										

Reproduced with permission from Nesbitt M (2002). When and where did domesticated cereals first occur in southwest Asia? In: Cappers RTJ and Bottema S (eds.) *The Dawn of Farming in the Near East*, (Studies in Early Near Eastern Production, Subsistence, and Environment Vol. 6), p. 113–132. Berlin: ex oriente.

of larger fields. The presence of winter annuals from Neolithic times onwards points to winter- and summer-cropping.

Usually, there are many lines of evidence that crops were grown separately. However, there are also hints on mixtures of crops in several periods. This seems to be reasonable as a risk-minimizing strategy. However, another possibility is that those crops were grown as fodder (see also Table 12).

Some Other Aspects of Archaeobotanical Research

There are many more aspects of archaeobotanical research of which only some can be mentioned here. The analysis of plant remains from animal dung allows a thorough

reconstruction of their fodder. Wood was carefully chosen depending on its properties. First traces of widespread grassland occurred in northern parts of Europe from the Bronze and mainly the Iron Age onwards. In parts of Europe, north of the Alps since the Roman period there was widespread trade. 'Exotic' species like pomegranates, dates, pepper, anis, etc. appear already in early Roman contexts. It is also obvious that gardening was introduced by the Romans in regions outside the Mediterranean. Cultivated cherry, apple, peach, walnut, etc., but also legumes and spices like coriander are regularly found north of the Alps from Roman times onwards. There are hints that some 'luxurious' foodstuffs were eaten only by socially higher ranked people like officers of the army (van der Veen, 2003). There were also fundamental

Table 8 Further reading: Hunter/gatherer diet, domestication of plants

- Aurenche, O. and Kozłowski, S. K. 1999. *La naissance du Néolithique au Proche Orient ou le paradis perdu*. Paris: Editions Errance.
- Badr, A., Müller, K., Schäfer-Pregl, R., El Rabey, H., Effgen, S., Ibrahim, H. H., Pozzi, C., Rohde, W. and Salamini, F. 2000. On the origin and domestication history of barley (*Hordeum vulgare*). *Molecular Biology and Evolution* 17, 499–510.
- Benz, M. 2000. *Die Neolithisierung im Vorderen Orient. Theorien, archäologische Daten und ein ethnologisches Modell*. (Studies in Early Near Eastern Production, Subsistence, and Environment 7). Berlin: ex oriente.
- Damania, A. B., Valkoun, J., Willcox, G. and Qualset, C. O. (ed.) 1998. *The Origins of Agriculture and Crop Domestication. Proceedings of the Harlan Symposium, 10–14 May 1997, Aleppo, Syria*. Aleppo: International Centre for Agricultural Research in the Dry Areas ICARDA.
- Dollfus, G. 1997. Paléoenvironnement et sociétés humaines au Moyen-Orient de 20000 BP à 6000 BP. *Paléorient (special issue)* 23.
- Gebauer, A. B. and Price, D. T. 1992. *Transition to agriculture in prehistory*. (Monographs in World Archaeology 4). Madison (Wisconsin): Prehistory Press.
- Guilaine, J. (ed.) 2000. *Premiers paysans du monde. Naissance des agricultures*. Paris: Editions Errance.
- Harlan, J. R. 1995. *The living fields. Our agricultural heritage*. New York: Cambridge University Press.
- Harris, D. (ed.) 1996. *The Origins and Spread of Agriculture and Pastoralism in Eurasia*. London: Bookcraft.
- Harris, D. R. and Hillman, G. C. (ed.) 1989. *Foraging and farming. The evolution of plant exploitation*. London: Unwin Hyman.
- Heun, M., Schäfer-Pregl, R., Klawan, D., Castagna, R., Accerbi, M., Borghi, B. and Salamini, F. 1997. Site of Einkorn wheat domestication identified by DNA fingerprinting. *Science* 278, 1312–1314.
- Lu, H., Liu, Z., Wu, N., Berne, S., Saito, Y., Liu, B. and Wang, L. 2002. Rice domestication and climatic change: Phytolith evidence from East China. *Boreas (Oslo)* 31, 378–385.
- Marshall, F. and Hildebrand, E. 2002. Cattle Before Crops: The Beginnings of Food Production in Africa. *Journal of World Prehistory* 16, 99–143.
- Martinoli, D. 2004. Food plant use, temporal changes and site seasonality at Epipalaeolithic Öküzini and Karain B Caves, Southwest Anatolia, Turkey. *Paléorient* 30, 61–80.
- Martinoli, D. and Jacomet, S. 2004. Identifying endocarp remains and exploring their use at Epipalaeolithic Öküzini in Southwest Anatolia, Turkey. *Vegetation History and Archaeobotany* 13, 45–54.
- Nesbitt, M. 2001. Wheat evolution: integrating archaeological and biological evidence. *The Linnean* 3, 37–60.
- Nesbitt, M. and Samuel, D. 1996. From staple crop to extinction? The archaeology and history of the hulled wheats. In: *Hulled wheats. Promoting the conservation and use of underutilized and neglected crops 4. Proceedings of the First International Workshop on Hulled Wheats, 21–22 July 1995, Castelvecchio Pascoli, Tuscany, Italy* (ed. Padulosi, S., Hammer, K. and Heller, J.), p. 41–100. Rome: International Plant Genetic Resources Institute.
- Piperno, D. R., Weiss, E., Holst, I. and Nadel, D. 2004. Processing of wild cereal grains in the Upper Palaeolithic revealed by starch grain analysis. *Nature* 430, 670–673.
- Smith, B. D. 1998. *The emergence of agriculture*. New York: Scientific American Library.
- Weiss, E., Wetterstrom, W., Nadel, D. and Bar-Yosef, O. 2004. The broad spectrum revisited: Evidence from plant remains. *Proceedings of the National Academy of Sciences of the United States of America* 101, 9551–9555.
- Willcox, G. 2005. The distribution, natural habitats and availability of wild cereals in relation to their domestication in the Near East: multiple events, multiple centres. *Vegetation History and Archaeobotany* 14, 534–541.
- Willcox, G. 2004. Measuring grain size and identifying Near Eastern cereal domestication: evidence from the Euphrates valley. *Journal of Archaeological Science* 31, 145–150.

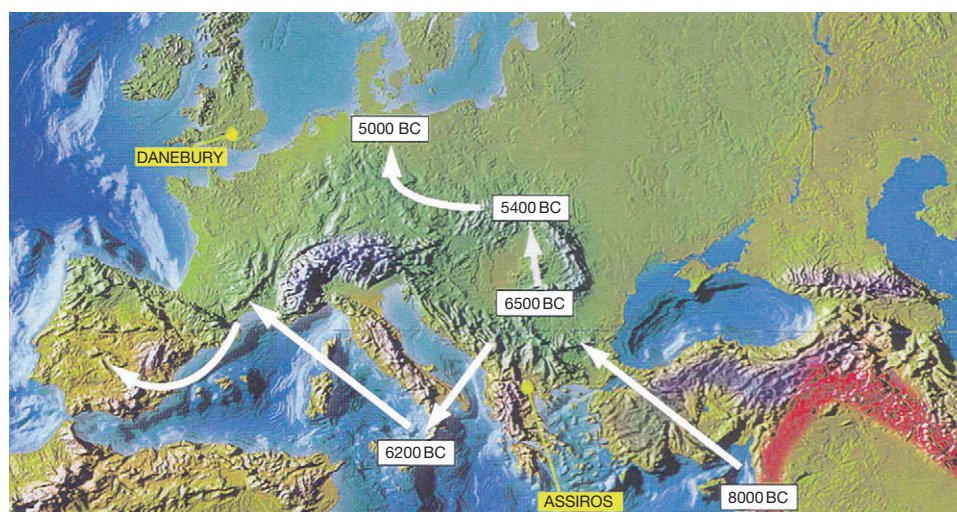


Figure 18 The origins and spread of agriculture from the Near East to Europe. The red shaded area is the Fertile Crescent, where agriculture began some 10 000 years ago. Danebury and Assiros are two archaeological sites where charred wheat remains containing Ancient DNA have been recovered. Reproduced with permission from Brown T and Jones GEM (2001) *New ways with old wheats*. Molecular Signatures from the Past (ed. by the Natural Environment Research Council) 9, 1–2.

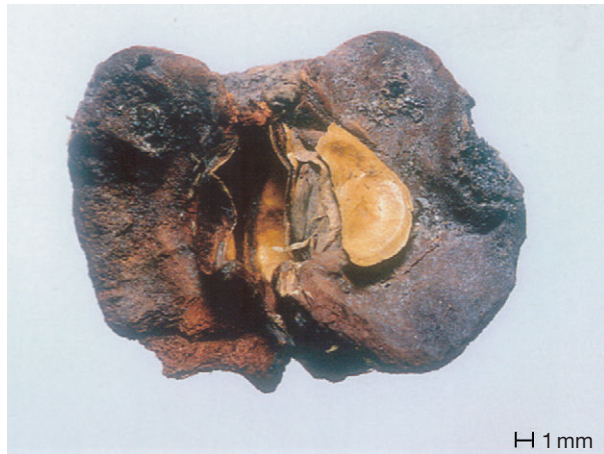


Figure 19 Find of a halved wild apple from the Neolithic Lakshore settlement Arbon Bleiche 3 (3384–3370 BC), Lake Constance, Switzerland. Reproduced with permission from [Jacomet S, Leuzinger U and Schibler J \(2004\). *Die neolithische Seeufersiedlung Arbon Bleiche 3. Umwelt und Wirtschaft*. \(Archäologie im Thurgau 12\). Frauenfeld: Amt für Archäologie des Kantons Thurgau. Photograph made by D. Steiner, © AATG.](#)

differences between the ‘mass production’ of food in the roman villas and agriculture outside the Roman empire (e.g., in the *Germania libera*).

The interpretation of bioarchaeological material may allow also the reconstruction of aspects of rituals ([Wilkinson and Stevens, 2003](#)). These are actions that are associated consciously or subconsciously with symbolic meaning. For instance, many burials have ecofacts that are buried with them, such as plant remains that accompany artifacts and personal belongings. These may be remains of funeral feasts, food to sustain the individual on their journey to or within the afterlife, or offerings to buy their way in. Desiccated offerings of plant foods are a frequent feature of Egyptian burials. Many plant offerings including figs, dates, grain, sesame, and wreaths of lotus, olive, willow, cornflowers, mayweeds, and various fruits were found in Tutankhamen’s tomb. In incineration graves from the Roman period many carbonized food remains can be found. From temples, remains of burnt food offerings have been recorded from Roman Europe. (see also [Table 13](#)).

Table 9 Further reading: History of plant growing, spread of agriculture

- Akeret, Ö. 2005. Bell Beaker Plant Remains from Cortaillod, Sur les Rochettes Est, Switzerland. *Vegetation History and Archaeobotany* 14, 279–286.
- Bakels, C. C., Alkemade, M. J. and Vermeeren, C. E. 1993. Botanische Untersuchungen in der Rössener Siedlung Maastricht-Randwijck. In: *7000 Jahre bäuerliche Landschaft: Entstehung, Erforschung, Erhaltung. 20 Aufsätze zu Ehren von Karl-Heinz Knörzer* (ed. Kalis, A. J. and Meurers-Balke, J.), (Archaeo-Physika Vol. 13), p. 49–62.
- Behre, K.-E. 1992. The history of rye cultivation in Europe. *Vegetation History and Archaeobotany* 1, 141–156.
- Buxó, R. 1997. *Arqueologia de las plantas. La explotación económica de las semillas y los frutos en el marco mediterráneo de las Península Iberica*. Barcelona: Critica.
- Colledge, S. M., Conolly, J. and Shennan, S. J. 2004. Archaeobotanical Evidence for the Spread of Farming in the Eastern Mediterranean. *Current Anthropology* 45, S35–S58.
- Halstead, P. and O’Shea, J. (ed.) 1989. *Bad year economics: cultural responses to risk and uncertainty*. Cambridge: Cambridge University Press.
- Jacomet, S. 2006-a. Neolithic plant economies in the northern alpine foreland (Central Europe) from 5500–3500 BC cal. In: *The origins spread and use of domestic plants in Southwest Asia and Europe*. (ed. Colledge, S. and Conolly, J.), London: UCL Press.
- Jacomet, S. 2006-b. Plant Economy in the Northern Alpine Lake Dwelling area – 3500–2400 BC cal. In: Karg, S., Baumeister, R., Schlichtherle, H. and Robinson, D.E. (Eds.) *Economic and Environmental Changes during the 4th and 3rd Millennia BC*. Proceedings of the 25th Symposium of the AEA Sept. 2004 in Bad Buchau, Germany. *Environmental Archaeology*, 11, 65–65.
- Janssen, W. and Willerding, U. 1997. Gartenbau und Gartenpflanzen. *Reallexikon der Germanischen Altertumskunde* 10, 449–462.
- Kreuz, A. 1990. Die ersten Bauern Mitteleuropas – eine archäobotanische Untersuchung zu Umwelt und Wirtschaft der ältesten Bandkeramik. *Analecta Praehistorica Leidensia* 23, 1–256.
- Kreuz, A., Marinova, E., Schäfer, E. and Wiethold, J. 2005. A comparison of early Neolithic crop and weed assemblages from the Linearbandkeramik and the Bulgarian Neolithic cultures: differences and similarities. *Vegetation History and Archaeobotany* 14, 237–258.
- Maier, U. 1996. Morphological studies of free-threshing wheat ears from a Neolithic site in southwest Germany, and the history of the naked wheats. *Vegetation History and Archaeobotany* 5, 39–55.
- Nesbitt, M. 2003. Obst und Gemüse. *Reallexikon der Germanischen Altertumskunde* 10, 26–30.

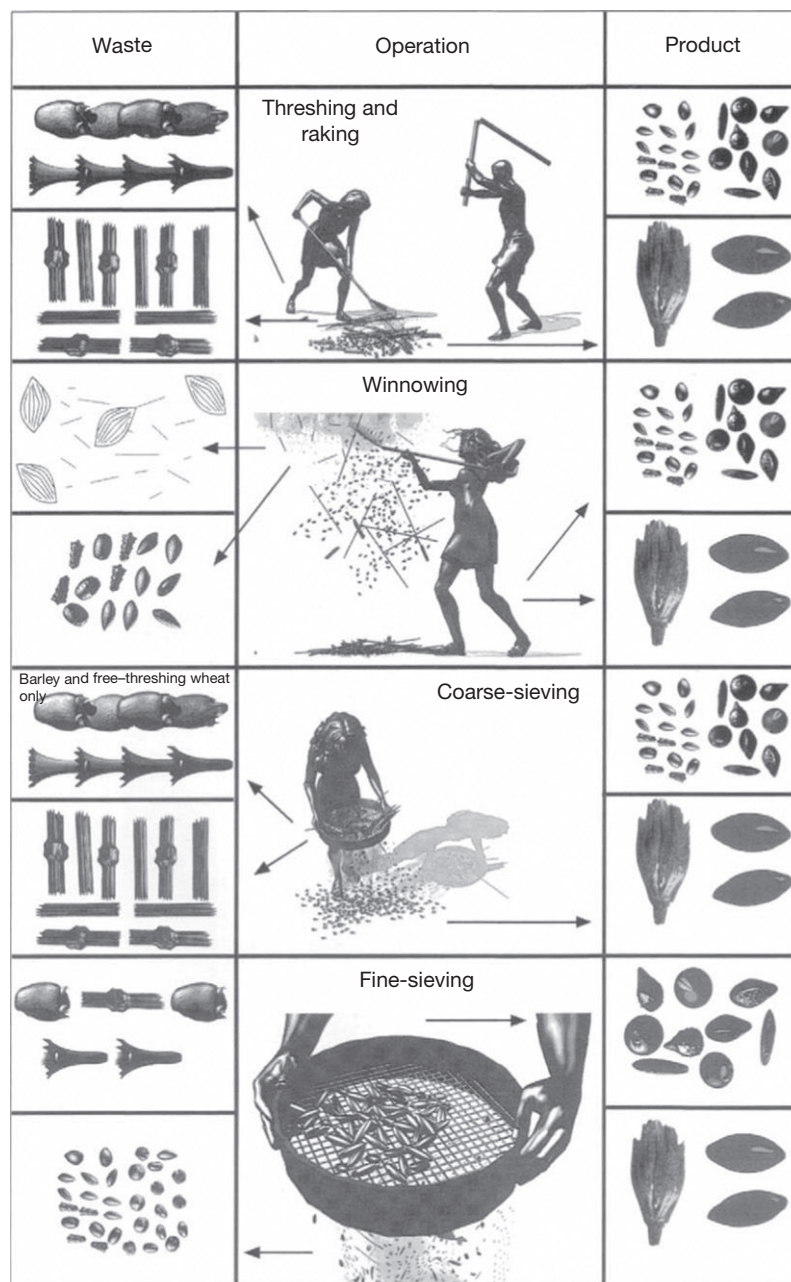


Figure 20 Crop processing stages for wheat and barley showing their effect on charred macrofossil assemblages, based on ethnographic work undertaken by Glynis Jones (Jones GEM (1983). The ethnoarchaeology of crop processing: Seeds of a middle-range methodology. *Archaeological Review from Cambridge* 2: 17–26) and Gordon Hillman (1984) in Greece and Turkey respectively. Each stage produces both a well defined ‘product’ and ‘waste’. Reproduced with permission from Wilkinson and Stevens 2003, figure 74, p. 196–197.

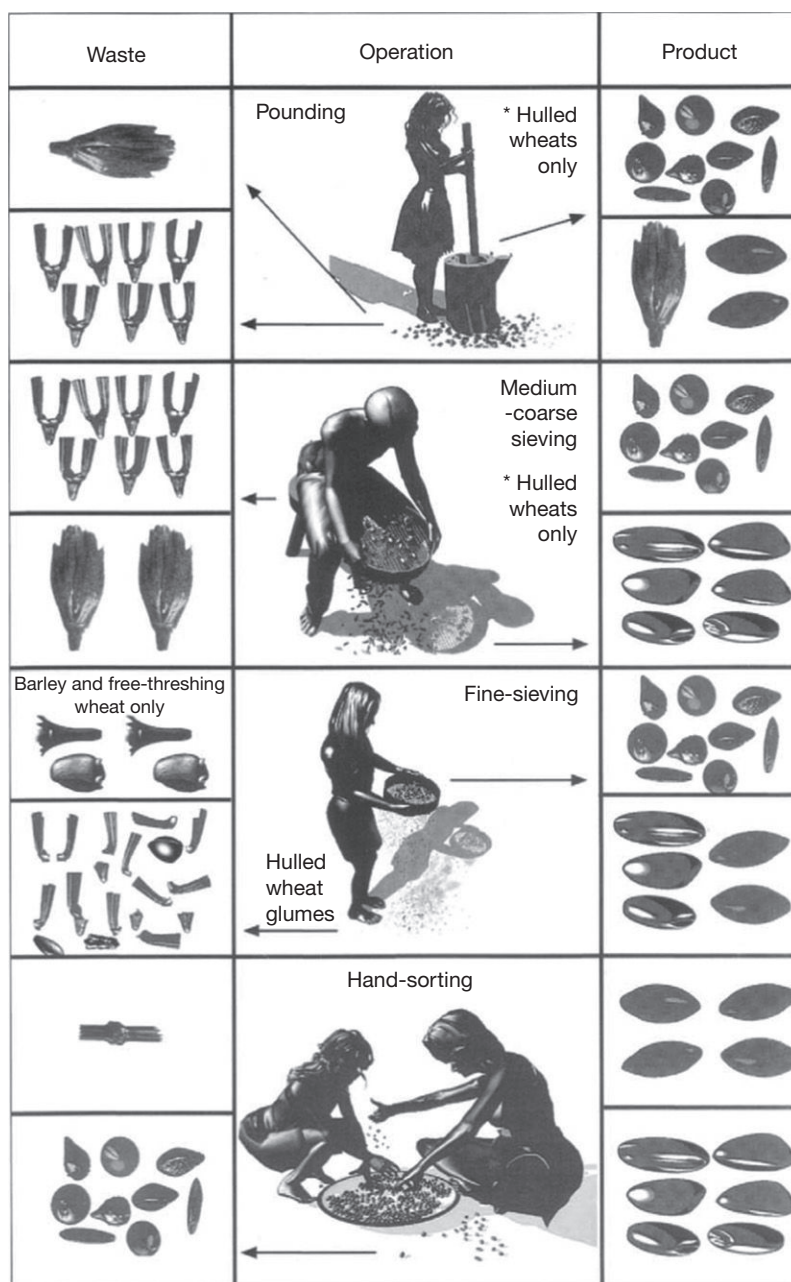


Figure 20 (Cont'd)

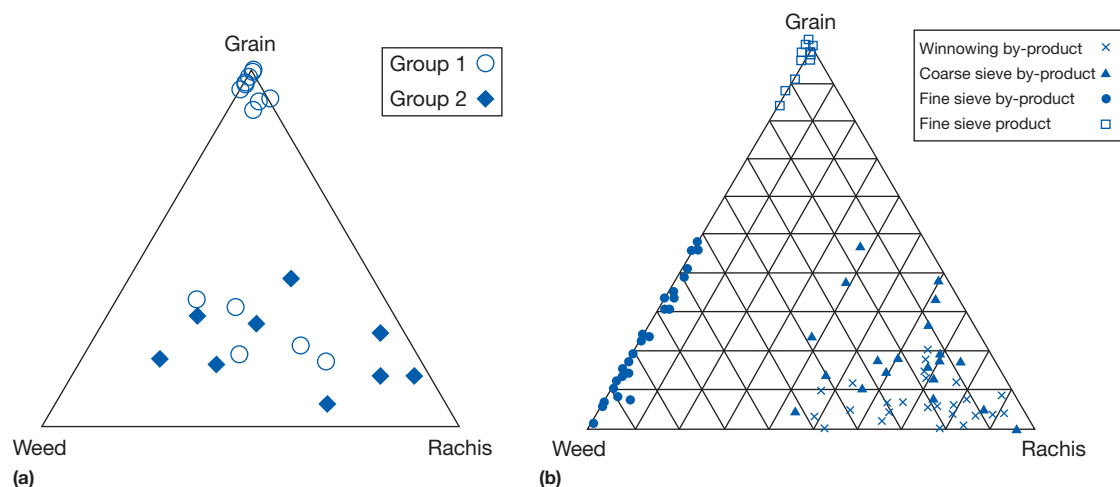


Figure 21 Assessing crop-processing stage from grain, rachis and weed-proportions. Diagrams (a) showing the relative proportions of grains (free threshing cereal only), rachis internodes and weed seeds for two groups of samples: 1: samples with at least 80% of hulled barley, $n=16$ and 2: samples with less than 80% hulled barley, but at least 80% hulled barley and free-threshing wheat ($n=9$). (b) Diagram showing the proportions of grains, rachis internodes and weed seeds in ethnobotanical samples from Amorgos (Early Bronze Age, Greece) of crop-processing products and by-products for free-threshing cereals (after G. Jones). Reproduced with permission from Charles M and Bogaard A (2003). Third-millennium BC Charred Plant Remains from Tell Brak. In: Oates D, Oates J and McDonald H (ed.) *Excavations at Tell Brak, Vol 2: Nagar in the third millenium BC*, (McDonald Institute Monographs Vol. 2), pp. 302–326. Cambridge: MCDonald Institute for Archaeological Research and British School of Archaeology in Iraq.

Table 10 Further reading: Processing techniques, storage, food preparation

- Boenke, N. 2003. Organic resources: Food supply and raw materials (in T. Stöllner: *The Economy of Dürrnberg-Bei-Hallein: an Iron Age Salt-mining Centre in the Austrian Alps*). *The Antiquaries Journal* 83, 146–149.
- Dickson, C. 1990. Experimental Processing and Cooking of Emmer and Spelt Wheats and the Roman Army Diet. In: *Experimentation and Reconstruction in Environmental Archaeology* (ed. Robinson, D. E.), (Symposia of the Association for Environmental Archaeology Vol. 9), p. 33–39. Oxford: Oxbow Books.
- Hansson, A. M. 1997. *On plant food in the Scandinavian peninsula in Early Medieval times*. (Theses and Papers in Archaeology B5). Stockholm: Stockholm University, Archaeological Research Laboratory.
- Hillman, G. C. 1984. Evidence for spelt malting at Catsgore. In: *Excavations at Catsgore 1970–1973. A Romano-British Village* (ed. Leech, R.), (Western Archaeological Trust. Excavation Monograph Vol. 2), p. 137–141.
- Holden, T. G. and Nuñez, L. 1993. An analysis of the gut contents of five well-preserved human bodies from Tarapacá, northern Chile. *Journal of Archaeological Science* 20, 595–611.
- Jones, G. E. M. 1990. The application of present-day cereal processing studies to charred archaeobotanical remains. *Circaea* 6, 91–96.
- Jones, G. E. M., Wardle, K., Halstead, P. and Wardle, D. 1986. Crop Storage at Assiros. *Scientific American* 254, 96–103.
- Samuel, D. 1996a. Investigation of Ancient Egyptian baking and brewing methods by correlative microscopy. *Science* 273, 488–490.
- Samuel, D. 1996b. Archaeology of Ancient Egyptian beer. *Journal of the American Society of Brewing Chemistry* 54, 3–12.
- Valamoti, S. M. 2002. Food remains from Bronze Age Archondiko and Mesimeriani Tomba in northern Greece? *Vegetation History and Archaeobotany* 11, 17–22.
- Valamoti, S. M. 2003. Neolithic and Early Bronze Age “food” from northern Greece: the archaeobotanical evidence. In: *Food, Culture and Identity in the Neolithic and Early Bronze Age* (ed. Parke Pearson, M.), (BAR International Series Vol. 1117), p. 97–111. Oxford: Archaeopress.

Table 11 Further reading: Production and consumption

- Bakels, C. 2001. Producers and consumers in archaeobotany. A comment on “When method meets theory: the use and misuse of cereal producer / consumer models in archaeobotany”. In: *Environmental Archaeology: Meaning and Purpose* (ed. Albarella, U.), (Environmental Science and Technology Library Vol. 17), p. 299–304. Dordrecht / Boston / London: Kluwer Academic Publishers.
- Jones, M. K. 1985. Archaeobotany beyond Subsistence Reconstruction. In: *Beyond domestication in prehistoric Europe* (ed. Barker, G. and Gamble, C.), London: Academic Press Inc.
- Kooistra, L. I. 1996. *Borderland Farming. Possibilities and limitations of farming in the Roman Period and Early Middle Ages between the Rhine and Meuse*. Amersfoort: Van Gorcum.
- van der Veen, M. and Jones, G. E. M. in press. Are-analysis of production and consumption implications for understanding the British Iron Age. *Vegetation History and Archaeobotany*, 15, 217–228.

Table 12 Further Reading: Cultivation techniques

- Hillman, G. C. 1984. Traditional Husbandry and Processing of Archaic Cereals in Recent Times: The Operations, Products and Equipment which might feature in Sumerian Texts. Part I: The Glume Wheats. *Bulletin on Sumerian Agriculture* 1, 114–151.
- Hosch, S. and Jacomet, S. 2004. Ackerbau und Sammelwirtschaft. Ergebnisse der Untersuchung von Samen und Früchten. In: *Die neolithische Seeufersiedlung Arbon Bleiche 3: Wirtschaft und Umwelt* (ed. Jacomet, S., Schibler, J. and Leuzinger, U.), (Archäologie im Thurgau Vol. 12), p. 112–157. Frauenfeld: Amt für Archäologie des Kantons Thurgau.
- Karg, S. 1995. Plant diversity in late medieval cornfields of northern Switzerland. *Vegetation History and Archaeobotany* 4, 41–50.
- Kreuz, A., Marinova, E., Schäfer, E. and Wiethold, J. 2005. A comparison of early Neolithic crop and weed assemblages from the

(Continued)

Table 12 (Continued)

- Linearbandkeramik and the Bulgarian Neolithic cultures: differences and similarities. *Vegetation History and Archaeobotany* 14.
- Kühn, M. 1996. *Spätmittelalterliche Getreidefunde aus einer Brandschicht des Basler Rosshof-Areales (15. Jahrhundert A.D.)*. (Materialhefte zur Archäologie in Basel 11). Basel: Archäologische Bodenforschung des Kantons Basel-Stadt.
- Peña-Chocarro, L. 1999. *Prehistoric Agriculture in Southern Spain during the Neolithic and the Bronze Age. The application of ethnographic models*. (BAR International Series 818). Oxford: Archaeopress.
- Reynolds, P. J. 1999. The nature of experiment in archaeology. In: *Archaeology of the Bronze and Iron Age. Experimental Archaeology. Proceedings of the International Archaeological Conference Szazhalombatta, 3–7 October 1996* (ed. Jerem, E. and Poroszlai, I.), (Archaeolingua Vol. 9), p. 387–396. Budapest: Archaeolingua.
- Rösch, M., Ehrmann, O., Herrmann, L., Schulz, E., Bogenrieder, A., Goldammer, J. P., Hall, M., Page, H. and Schier, W. 2002. An experimental approach to Neolithic shifting cultivation. *Vegetation History and Archaeobotany* 11, 143–154.
- Terral, J.-F. 2000. Exploitation and Management of the Olive Tree During Prehistoric Times in Mediterranean France and Spain. *Journal of Archaeological Science* 27, 127–134.
- van der Veen, M. 1992. *Crop Husbandry Regimes: An Archaeobotanical Study of Farming in northern England 1000 BC–AD 500*. (Sheffield Archaeological Monographs 3). Sheffield: J. R. Collins Publications.

Table 13 Further reading: Other aspects (paleoenvironment, animal fodder, wood use, trade, social structure, ritual etc.)

- Akeret, O. and Rentzel, P. 2001. Micromorphology and plant macrofossil analysis of cattle dung from the Neolithic lake shore settlement of Arbon-Bleiche 3. *Geoarchaeology* 16, 687–700.
- Akeret, Ö., Haas, J.-N., Leuzinger, U. and Jacomet, S. 1999. Plant macrofossils and pollen in goat/sheep faeces from the Neolithic lake-shore settlement Arbon Bleiche 3, Switzerland. *The Holocene* 9, 175–182.
- Albrecht, H., Schlumbaum, A. and Jacomet, S. 1999. Das archäobotanische Fundmaterial: Die Holzkohlen. Ein Beitrag zur mittelalterlichen Holznutzung in der Nordwestschweiz. In: *Laufen Rathausplatz. Eine hölzerne Häuserzeile in einer mittelalterlichen Kleinstadt: Hausbau, Sachkultur und Alltag. Die Ergebnisse der Grabungskampagnen 1988 und 1989* (ed. Pfrommer, J. and Gutscher, D.), p. 249–260. Bern, Stuttgart, Wien: Berner Lehrmittel- und Medienverlag c/o Paul Haupt.
- Bottema, S., Entjes-Nieborg, G. and van Zeist, W. 1990. *Man's Role in the Shaping of the Eastern Mediterranean Landscape*. Rotterdam: Balkema.
- Bouby, L. and Marinval, P. 2004. Fruits and seeds from Roman cremations in Limagne (Massif Central) and the spatial variability of plant offerings in France. *Journal of Archaeological Science* 31, 77–86.
- Jacomet, S., Kučan, D., Ritter, A., Suter, G. and Hagendorn, A. 2002. *Punica granatum* L. (Pomegranates) from early Roman contexts in Vindonissa (Switzerland). *Vegetation History and Archaeobotany* 11, 79–92.
- Körber-Grohne, U. 1990. *Gramineen und Grünlandvegetationen vom Neolithikum bis zum Mittelalter in Mitteleuropa*. Bibliotheca Botanica 139.
- Kreuz, A. 2004. Landwirtschaft im Umbruch? Archäobotanische Untersuchungen zu den Jahrhunderten um Christi Geburt in Hessen und Mainfranken. *Bericht der Römisch-Germanischen Kommission* 85, 97–292, 9 Tafeln.

(Continued)

Table 13 (Continued)

- Kučan, D. 1992. Die Pflanzenreste aus dem römischen Militärlager Oberaden. In: *Das Römerlager in Oberaden III. Die Ausgrabungen im nordwestlichen Lagerbereich und weitere Baustellenuntersuchungen der Jahre 1962–1968* (ed. Kühlborn, J. S.), (Bodenaltertümer Westfalens Vol. 27), p. 237–265. Münster: Aschendorff.
- Kühn, M. and Hadorn, P. 2004. Pflanzliche Makro- und Mikroreste aus Dung von Wiederkäuern. In: *Die jungsteinzeitliche Seeufersiedlung Arbon Bleiche 3. Umwelt und Wirtschaft* (ed. Jacomet, S., Leuzinger, U. and Schibler, J.), (Archäologie im Thurgau Vol. 12), p. 327–350. Frauenfeld: Amt für Archäologie des Kantons Thurgau.
- Leveau, P., Trément, F., Walsh, K. and Barker, G. (ed.) 1999. *Environmental Reconstruction in Mediterranean Landscape Archaeology*. (The Archaeology of Mediterranean landscapes Vol. 2). Oxford: Oxbow Books.
- Petrucchi-Bavaud, M., Schlumbaum, A. and Jacomet, S. 2000. Samen, Früchte und Fertigprodukte. In: *Der Südfriedhof von Vindonissa. Archäologische und naturwissenschaftliche Untersuchungen im römerzeitlichen Gräberfeld Windisch-Dägerli* (ed. Hintermann, D.), (Veröffentlichungen der Gesellschaft Pro Vindonissa Vol. 17), p. 151–159. Brugg: Aargauische Kantonsarchäologie.
- Robinson, D. E. 2002. Domestic burnt offerings and sacrifices at Roman and pre-Roman Pompeii, Italy. *Vegetation History and Archaeobotany* 11, 93–99.
- Schweingruber, F. H. 1976. *Prähistorisches Holz. Die Bedeutung von Holzfunden aus Mitteleuropa für die Lösung archäologischer und vegetationskundlicher Probleme*. (Academica Helvetica 2). Bern: Verlag Paul Haupt.
- Willerding, U. 1999a. Heilmittel und Heilkräuter. *Reallexikon der Germanischen Altertumskunde* 14, 208–233.
- Willerding, U. 1999b. Heu. *Reallexikon der Germanischen Altertumskunde* 14, 510–526.
- Zach, B. 2002. Vegetable offerings on the Roman sacrificial site in Mainz, Germany – short report on the first results. *Vegetation History and Archaeobotany* 11, 101–106.

See also: [Dendrochronology](#); [Plant Macrofossil Introduction](#).
[Quaternary Stratigraphy](#): Overview.

References

- Akeret Ö (2005) Bell Beaker Plant Remains from Cortaillod, Sur les Rochettes Est, Switzerland. *Vegetation History and Archaeobotany* 14.
- Akeret O and Rentzel P (2001) Micromorphology and plant macrofossil analysis of cattle dung from the Neolithic lake shore settlement of Arbon-Bleiche 3. *Geoarchaeology* 16: 687–700.
- Akeret Ö, Haas JN, Leuzinger U, and Jacomet S (1999) Plant macrofossils and pollen in goat/sheep faeces from the Neolithic lake-shore settlement Arbon Bleiche 3, Switzerland. *The Holocene* 9: 175–182.
- Albarella U (ed.) (2001) *Environmental Archaeology: Meaning and Purpose*. *Environmental Science and Technology Library*, vol. 17. Dordrecht: Kluwer Academic Publishers.
- Albrecht H, Schlumbaum A, and Jacomet S (1999) Das archäobotanische fundmaterial: Die Holzkohlen. Ein Beitrag zur mittelalterlichen Holznutzung in der Nordwestschweiz. In: Pfrommer J and Gutscher D (eds.) *Laufen Rathausplatz. Eine hizerne Häuserzeile in einer mittelalterlichen Kleinstadt: Hausbau, Sachkultur und Alltag. Die Ergebnisse der Grabungskampagnen 1988 und 1989*, pp. 249–260. Bern, Stuttgart, Wien: Berner Lehrmittel- und Medienverlag c/o Paul Haupt.
- Aurenche O and Kozłowski SK (1999) *La Naissance du Néolithique au Proche Orient ou le Paradis Perdu*. Paris: Editions Errance.
- Badr A, Müller K, Schäfer-Pregl R, et al. (2000) On the origin and domestication history of barley (*Hordeum vulgare*). *Molecular Biology and Evolution* 17: 499–510.

- Bakels CC (1991) Tracing crop processing in the Bandkeramik culture. In: Renfrew JM (ed.) *New light on early farming. Recent developments in palaeoethnobotany*, pp. 281–288. Edinburgh: Edinburgh University Press.
- Bakels C (2001) Producers and consumers in archaeobotany. A comment on “When method meets theory: The use and misuse of cereal producer/consumer models in archaeobotany”. In: Albarella U (ed.) *Environmental Archaeology: Meaning and Purpose*, (Environmental Science and Technology Library Vol. 17), pp. 299–304. Dordrecht: Kluwer Academic Publishers.
- Bakels CC, Alkemade MJ, and Vermeeren CE (1993) Botanische Untersuchungen in der Rössener Siedlung Maastricht-Randwijck. In: Kalis AJ and Meurers-Balke J (eds.) *7000 Jahre Bäuerliche Landschaft: Entstehung, Erforschung, Erhaltung. 20 Aufsätze zu Ehren von Karl-Heinz Knörzer*, (Archaeo-Physika Vol. 13) pp. 49–62.
- Behre KE and Oeggl K (1996) Early farming in the old world. *Recent advances in archaeobotanical research. Proceedings of the 10th IWGP Symposium, Innsbruck 1995. Vegetation History and Archaeobotany* 5: 1–200, (special issue).
- Behre KE (1992) The history of rye cultivation in Europe. *Vegetation History and Archaeobotany* 1: 141–156.
- Benz M (2000) *Die Neolithisierung im Vorderen Orient. Theorien, archäologische Daten und ein ethnologisches Modell*. Berlin: ex oriente (Studies in Early Near Eastern Production, Subsistence, and Environment 7).
- Boardman S and Jones G (1990) Experiments on the effects of charring on cereal plant components. *Journal of Archaeological Science* 17: 1–11.
- Boenke N (2003) Organic resources: Food supply and raw materials (in T. Stöllner: The Economy of Dürenberg-Bei-Hallein: An Iron Age Salt-mining Centre in the Austrian Alps). *The Antiquaries Journal* 83: 146–149.
- Bogaard A (2002) Questioning the relevance of shifting cultivation to Neolithic farming in the loess belt of Europe: Evidence from the Hambach Forest experiment. *Vegetation History and Archaeobotany* 11: 155–168.
- Bogaard A (2004) *Neolithic Farming in Central Europe. An Archaeobotanical Study of Crop Husbandry Practices*. London: Routledge.
- Bogucki P and Crabtree PJ (eds.) (2004) *Ancient Europe 8000 B.C. – A.D. 1000*. Encyclopedia of the Barbarian World, Vol. 1, The Mesolithic to the Copper Age (c. 8000–2000 B.C.).
- Bottema S, Entjes-Nieborg G, and van Zeist W (1990) *Man's Role in the Shaping of the Eastern Mediterranean Landscape*. Rotterdam: Balkema.
- Bouby L and Marinval P (2004) Fruits and seeds from Roman cremations in Limagne (Massif Central) and the spatial variability of plant offerings in France. *Journal of Archaeological Science* 31: 77–86.
- Brothwell DR and Pollard AM (eds.) (2001) *Handbook of Archaeological Sciences*. Chichester: John Wiley & Sons.
- Brown T and Jones GEM (2001) New ways with old wheats. Molecular Signatures from the Past. *Natural Environment Research Council* 9: 1–2.
- Buxó R, Jacomet S, and Bittmann F (2005) Interaction between man and plants. New Progress in Archaeobotanical Research, *Proceedings of the 13th IWGP Symposium, Girona 2004. Vegetation History and Archaeobotany* 14: 235–236.
- Buxó R (1997) *Arqueología de las Plantas. La Explotación Económica de las Semillas y los Frutos en el Marco Mediterráneo de las Península Ibérica*. Barcelona: Critica.
- Cappers RTJ and Bottema S (eds.) (2002) *The Dawn of Farming in the Near East*. (Studies in Early Near Eastern Production, Subsistence, and Environment Vol. 6) Berlin: ex oriente.
- Chabal L, Fabre L, Terral JF, and Théry-Parisot I (1999) L'anthracologie. In: Bourquin-Mignot C, Brochier JE, and Chabal L, et al. (eds.) *La Botanique*, (Collection Archéologiques), pp. 43–104. Paris: Errance.
- Charles M and Bogaard A (2003) Third-millennium BC charred plant remains from Tell Brak. In: Oates D, Oates J, and McDonald H (eds.) *Excavations at Tell Brak. Vol. 2: Nagar in the Third Millennium BC (McDonald Institute Monographs Vol. 2)*, pp. 302–326. Cambridge: McDonald Institute for Archaeological Research and British School of Archaeology in Iraq.
- Colledge S and Conolly J (2007) *The origins, spread and use of domestic plants in Southwest Asia and Europe*. London: UCL Press.
- Colledge SM, Conolly J, and Shennan SJ (2004) Archaeobotanical evidence for the spread of farming in the eastern Mediterranean. *Current Anthropology* 45: S35–S58.
- Damania AB, Valkoun J, Willcox G, and Qualset CO (eds.) (1998) *The Origins of Agriculture and Crop Domestication*. Proceedings of the Harlan Symposium, 10–14 May 1997, Aleppo, Syria. Aleppo: International Centre for Agricultural Research in the Dry Areas ICARDA.
- Diamond J (2002) Evolution, consequences and future of plant and animal domestication. *Nature* 418: 700–707.
- Dickson JH, Oeggl K, and Handley LL (2003) The Iceman reconsidered. *Scientific American May 70–79*.
- Dickson C (1990) Experimental Processing and Cooking of Emmer and Spelt Wheats and the Roman Army Diet. In: Robinson DE (ed.) *Experimentation and Reconstruction in Environmental Archaeology (Symposia of the Association for Environmental Archaeology Vol. 9)*, pp. 33–39. Oxford: Oxbow Books.
- Dimbleby GW (1985) *The Palynology of Archaeological Sites*. London: Academic Press.
- Dincauze DF (2000) *Environmental Archaeology. Principles and Practice*. Cambridge: Cambridge University Press.
- Dollfus G (1997) Paléoenvironnement et sociétés humaines au Moyen-Orient de 20000 BP à 6000 BP. *Paléorient* 23: (special issue).
- Ernst M and Jacomet S (2006) The value of the archaeobotanical analysis of desiccated plant remains from old buildings: Methodological aspects and interpretation of crop weed assemblages. *Vegetation History and Archaeobotany* 15: 45–56.
- Evans J and O'Connor T (1999) *Environmental Archaeology. Principles and Methods*. Phoenix Mill: Sutton Publishing Limited.
- Gebauer AB and Price DT (1992) *Transition to Agriculture in Prehistory*. Madison: Prehistory Press (Wisconsin) (Monographs in World Archaeology 4).
- Germer R (1985) *Flora des pharaonischen Ägypten*. Mainz: Ph. von Zabern (Deutsches Archäologisches Institut Abteilung Kairo, Sonderschrift 14).
- Green FJ (1979) Phosphatic mineralization of seeds from archaeological sites. *Journal of Archaeological Science* 6: 279–284.
- Greig (1988) The interpretation of some Roman well fills from the midlands of England. In: Küster H (ed.) *Der prähistorische Mensch und seine Umwelt. Festschrift für Udelgard Körber-Grohne zum 65. Geburtstag*. (Forschungen und Berichte zur Vor- und Frühgeschichte in Baden-Württemberg Vol. 31), pp. 367–378. Stuttgart: Konrad Theiss Verlag.
- Guilaine J (ed.) (2000) *Premiers Paysans du Monde. Naissance des Agricultures*. Paris: Editions Errance.
- Hall A and Kenward H (2003) Can we identify biological indicator groups for craft, industry and other activities? In: Murphy P and Wiltshire PEJ (eds.) *The Environmental Archaeology of Industry. Symposia of the Association for Environmental Archaeology No. 20*, pp. 114–130. Oxford: Oxbow Books.
- Halstead P and O'Shea J (eds.) (1989) *Bad Year Economics: Cultural Responses to Risk and Uncertainty*. Cambridge: Cambridge University Press.
- Hansson AM (1997) *On plant food in the Scandinavian peninsula in Early Medieval times. (Theses and Papers in Archaeology B5)*. Stockholm: Stockholm University, Archaeological Research Laboratory.
- Harlan JR (1995) *The Living Fields. Our Agricultural Heritage*. New York: Cambridge University Press.
- Harris D (ed.) (1996) *The Origins and Spread of Agriculture and Pastoralism in Eurasia*. London: Bookcraft.
- Harris DR and Hillman GC (1989) *Foraging and Farming. The Evolution of Plant Exploitation*. London: Unwin Hyman.
- Hather JG (1993) *An Archaeobotanical Guide to Root and Tuber Identification*. Vol. I: Europe and South West Asia Oxford: Oxbow Books (Oxbow Monograph 28).
- Heun M, Schäfer-Pregl R, Klawan D, et al. (1997) Site of Einkorn wheat domestication identified by DNA fingerprinting. *Science* 278: 1312–1314.
- Hillman GC (1984) Interpretation of archaeological plant remains: The application of ethnographic models from Turkey. In: van Zeist WA and Casparie WA (eds.) *Plants and Ancient Man*, pp. 1–41. Rotterdam: Balkema.
- Hillman GC and Davies MS (1992) Domestication rate in wild wheats and barley under primitive cultivation: Preliminary results and archaeological implications of field measurements of selection coefficient. In: Anderson P (ed.) *Préhistoire de L'Agriculture: Nouvelles Approches Expérimentales et Ethnographiques (Monographie du CRA Vol. 6)*, pp. 113–158. Paris: Editions du CNRS.
- Hillman GC, Hedges R, Moore A, Colledge S, and Pettitt P (2001) New evidence of lateglacial cereal cultivation at Abu Hureyra on the Euphrates. *The Holocene* 11: 383–393.
- Hillman GC (1984) Evidence for spelt malting at Catsgore. In: Leech R (ed.) *Excavations at Catsgore 1970–1973. A Romano-British Village*, (Western Archaeological Trust. Excavation Monograph Vol. 2) 137–141.
- Hillman GC (1984) Traditional husbandry and processing of archaic cereals in recent times: The operations, products and equipment which might feature in Sumerian texts. Part I: The glume wheats. *Bulletin on Sumerian Agriculture* 1: 114–151.
- Hillman GC, Mason S, de Moulins D, and Nesbitt M (1996) Identification of archaeological remains of wheat: The 1992 London workshop. *Circaea* 12: 195–210.
- Holden TG and Nuñez L (1993) An analysis of the gut contents of five well-preserved human bodies from Tarapacá, northern Chile. *Journal of Archaeological Science* 20: 595–611.
- Hosch S and Jacomet S (2001) New aspects of archaeobotanical research in Central European Neolithic Lake Dwelling Sites. *Environmental Archaeology* 6: 59–71.
- Hosch S and Jacomet S (2004) Ackerbau und sammelwirtschaft. Ergebnisse der untersuchung von Samen und Früchten. In: Jacomet S, Schibler J, and Leuzinger U (eds.) *Die Neolithische Seeufersiedlung Arbon Bleiche 3: Wirtschaft und Umwelt*.

- Archäologie im Thurgau (Vol. 12), pp. 112–157. Frauenfeld: Amt für Archäologie des Kantons Thurgau.
- Hosch S and Zibulski P (2003) The influence of inconsistent wet-sieving procedures on the macroremains concentration in waterlogged sediments. *Journal of Archaeological Science* 30: 849–857.
- Jacomet S and Kreuz A (1999) *Archäobotanik. Aufgaben, Methoden und Ergebnisse Vegetations- und Agrargeschichtlicher Forschungen*. Stuttgart: Eugen Ulmer.
- Jacomet S, Jones G, Charles M, and Bittmann F (2002) Archaeology of plants. Current research in archaeobotany. *Proceedings of the 12th IWGP Symposium, Sheffield, 2001. Vegetation History and Archaeobotany* 11: 1–179.
- Jacomet S, Leuzinger U, and Schibler J (2004) *Die Neolithische Seeufersiedlung Arbon Bleiche 3. Umwelt und Wirtschaft*. Frauenfeld: Amt für Archäologie des Kantons Thurgau (Archäologie im Thurgau 12).
- Jacomet S (2005) Reconstructing intra-site patterns in Neolithic lakeshore settlements: The state of archaeobotanical research and future prospects. In: Della Casa P and Trachsel M (eds.) *WES'04 – Wetland Economies and Societies, Proceedings of the International Conference in Zurich, 10–13 March 2004. Chronos, Zürich* (Collectio Archaeologica Vol. 10).
- Jacomet S (2007) Neolithic plant economies in the northern alpine foreland (Central Europe) from 5500–3500 BC cal. Colledge S and Conolly J (eds.) *Archaeobotanical Perspectives on the Origin and Spread of Agriculture in Southwest Asia and Europe*. London: UCL Press.
- Jacomet S (2006) Plant economy in the northern alpine lake dwelling area – 3500–2400 BC cal. In: Karg S, Baumeister R, Schlichtherle H, and Robinson, DE (eds.) *Economic and Environmental Changes during the 4th and 3rd Millennia BC*. Proceedings of the 25th Symposium of the AEA September 2004 in Bad Buchau, Germany. *Environmental Archaeology*, 65–85.
- Jacomet S, Kucan D, Ritter A, Suter G, and Hagendorn A (2002) *Punica granatum L.* (Pomegranates) from early Roman contexts in Vindonissa (Switzerland). *Vegetation History and Archaeobotany* 11: 79–92.
- Janssen W and Willerding U (1997) Gartenbau und gartenpflanzen. *Reallexikon der Germanischen Altertumskunde* 10: 449–462.
- Jones DM (2002) *Environmental Archaeology: A Guide to the Theory and Practice of Methods, from Sampling and Recovery to Post-Excavation*. Swindon: English Heritage.
- Jones GEM (1983) The ethnoarchaeology of crop processing: Seeds of a middle-range methodology. *Archaeological Review from Cambridge* 2: 17–26.
- Jones GEM (1987) A statistical approach to the archaeological identification of crop processing. *Journal of Archaeological Science* 14: 311–323.
- Jones GEM (2002) Weed ecology as a method for the archaeobotanical recognition of crop husbandry practices. *Acta Palaeobotanica* 42: 185–193.
- Jones GEM, Charles M, Bogaard A, Hodgson J, and Palmer C (2005) The functional ecology of present-day arable weed floras and its applicability for the identification of past crop husbandry. *Vegetation History and Archaeobotany* 14: 493–504.
- Jones GEM (1990) The application of present-day cereal processing studies to charred archaeobotanical remains. *Circaea* 6: 91–96.
- Jones GEM, Wardle K, Halstead P, and Wardle D (1986) Crop storage at Assiros. *Scientific American* 254: 96–103.
- Jones MK (1985) Archaeobotany beyond subsistence reconstruction. In: Barker G and Gamble C (eds.) *Beyond Domestication in Prehistoric Europe*. London: Academic Press.
- Karg S (1995) Plant diversity in late medieval cornfields of northern Switzerland. *Vegetation History and Archaeobotany* 4: 41–50.
- Kenward HK, Hall AR, and Jones AKC (1980) A tested set of techniques for the extraction of plant and animal macrofossils from waterlogged archaeological deposits. *Science and Archaeology* 22: 3–15.
- Kenward H and Hall A (1997) Enhancing bioarchaeological interpretation using indicator groups: Stable manure as a paradigm. *Journal of Archaeological Science* 24: 663–673.
- Kenward H, Engleman C, Robertson A, and Large F (1985) Rapid scanning of urban archaeological deposits for insect remains. *Circaea* 3: 163–172.
- Kooistra LI (1996) *Borderland Farming. Possibilities and limitations of farming in the Roman Period and Early Middle Ages between the Rhine and Meuse*. Amersfoort: Van Gorcum.
- Körper-Grohne U (1999) Der Schacht in der keltischen Viereckschanze von Fellbach-Schmidlen (Rems-Murr-Kreis) in botanischer und stratigraphischer Sicht. In: Wieland G (ed.) *Die keltischen Viereckschanzen von Fellbach-Schmidlen und Ehningen*, (Forschungen und Berichte zur Vor- und Frühgeschichte in Baden-Württemberg Vol. 80), pp. 85–149. Stuttgart: Konrad Theiss Verlag.
- Körper-Grohne U (1990) Gramineen und Grünlandvegetationen vom Neolithikum bis zum Mittelalter in Mitteleuropa. *Bibliotheca Botanica* 139: 1–104.
- Kreuz A (1995) On-site and off-site data – interpretative tools for a better understanding of Early Neolithic environments. In: Kroll H and Pasternak R (eds.) *Res Archaeobotanicae. International Workgroup for Palaeoethnobotany*. Proceedings of the 9th Symposium, Kiel 1992, pp. 117–134. Kiel: Oetker-Voges.
- Kreuz A and Schäfer E (2002) A new archaeobotanical database program. *Vegetation History and Archaeobotany* 11: 177–179.
- Kreuz A (1990) Die ersten Bauern Mitteleuropas – eine archäobotanische Untersuchung zu Umwelt und Wirtschaft der ältesten Bandkeramik. *Analecta Praehistorica Leidensia* 23: 1–256.
- Kreuz A (2004) Landwirtschaft im Umbruch? Archäobotanische Untersuchungen zu den Jahrhunderten um Christi Geburt in Hessen und Mainfranken. *Bericht der Römisch-Germanischen Kommission* 85.
- Kreuz A, Marinova E, Schäfer E, and Wiethold J (2005) A comparison of early Neolithic crop and weed assemblages from the Linearbandkeramik and the Bulgarian Neolithic cultures: Differences and similarities. *Vegetation History and Archaeobotany* 14: 237–258.
- Kučan D (1992) Die Pflanzenreste aus dem römischen Militärlager Oberaden. In: Kühlborn JS (ed.) *Das Römerlager in Oberaden III. Die Ausgrabungen im nordwestlichen Lagerbereich und weitere Baustellenuntersuchungen der Jahre 1962–1968*. (Bodenaltertümer Westfalens Vol. 27), pp. 237–265. Münster: Aschendorff.
- Kühn M (1996) *Spätmittelalterliche Getreidefunde aus einer Brandschicht des Basler Rosshof-Areales (15. Jahrhundert AD)*. (Materialhefte zur Archäologie in Basel 11). Basel: Archäologische Bodenforschung des Kantons Basel-Stadt.
- Kühn M and Hadorn P (2004) Pflanzliche makro- und mikroreste aus dung von wiederkäuern. In: Jacomet S, Leuzinger U, and Schibler J. (eds.) *Die Jungsteinzeitliche Seeufersiedlung Arbon Bleiche 3. Umwelt und Wirtschaft*, (Archäologie im Thurgau Vol. 12), pp. 327–350. Amt für Archäologie des Kantons Thurgau, Frauenfeld.
- Leveau P, Trément F, Walsh K, and Barker G (eds.) (1999) *Environmental Reconstruction in Mediterranean Landscape Archaeology*. The Archaeology of Mediterranean Landscapes (Vol. 2). Oxford: Oxbow Books.
- Lu H, Liu Z, Wu N, Berne S, Saito Y, Liu B, and Wang L (2002) Rice domestication and climatic change: Phytolith evidence from East China. *Boreas (Oslo)* 31: 378–385.
- Maier U (1996) Morphological studies of free-threshing wheat ears from a Neolithic site in southwest Germany, and the history of the naked wheats. *Vegetation History and Archaeobotany* 5: 39–55.
- Marshall F and Hildebrand E (2002) Cattle before crops: The beginnings of food production in Africa. *Journal of World Prehistory* 16: 99–143.
- Martinoli D (2004) Food plant use, temporal changes and site seasonality at Epipalaeolithic Öküzini and Karain B Caves, southwest Anatolia, Turkey. *Paléorient* 30: 61–80.
- Martinoli D and Jacomet S (2004) Identifying endocarp remains and exploring their use at Epipalaeolithic Öküzini in Southwest Anatolia, Turkey. *Vegetation History and Archaeobotany* 13: 45–54.
- Moore AMT, Hillman GC, and Legge AJ (2000) *Village on the Euphrates. From Foraging to Farming at Abu Hureyra*. Oxford: Oxford University Press.
- Nesbitt M (2003) Obst und Gemüse. *Reallexikon der Germanischen Altertumskunde* 10: 26–30.
- Nesbitt M (2002) When and where did domesticated cereals first occur in southwest Asia? In: Cappers RTJ and Bottema S (eds.) *The Dawn of Farming in the Near East (Studies in Early Near Eastern Production, Subsistence, and Environment Vol. 6)*, pp. 113–132. Berlin: Ex Oriente.
- Nesbitt M (2001) Wheat evolution: Integrating archaeological and biological evidence. *The Linnean* 3: 37–60.
- Nesbitt M and Samuel D (1996) From staple crop to extinction? The archaeology and history of the hulled wheats. In: Padulosi S, Hammer K, and Heller J (eds.) *Hulled Wheats. Promoting the Conservation and Use of Underutilized And Neglected Crops 4*. Proceedings of the First International Workshop on Hulled Wheats, 21–22 July 1995, International Plant Genetic Resources Institute, Rome, pp. 41–100. Italy: Castelveccchio Pascoli, Tuscany.
- Neumann K, Schoch W, Détéienne P, and Schweingruber FH (2001) *Woods of the Sahara and the Sahel: An Antomical Atlas*. Bern: Paul Haupt Verlag.
- Pearsall DM (2000) *Palaeoethnobotany. A Handbook of Procedures*. San Diego: Academic Press.
- Peña-Chocarro L (1999) *Prehistoric Agriculture in Southern Spain during the Neolithic and the Bronze Age. The Application of Ethnographic Models*. Oxford: Archaeopress (BAR International Series 818).
- Petrucchi-Bavaud M, Schlumberg A, and Jacomet S (2000) Samen, Früchte und Fertigprodukte. In: Hintermann D (ed.) *Der Südfriedhof von Vindonissa. Archäologische und Naturwissenschaftliche Untersuchungen im Römerzeitlichen Gräberfeld Windisch-Dägerli*. (Veröffentlichungen der Gesellschaft Pro Vindonissa Vol. 17), pp. 151–159. Brugg: Aargauische Kantonsarchäologie.
- Piperno DR, Weiss E, Holst I, and Nadel D (2004) Processing of wild cereal grains in the Upper Palaeolithic revealed by starch grain analysis. *Nature* 430: 670–673.

- Renfrew C (2005) *Archaeology: The Key Concepts*. London: Routledge.
- Reynolds, P. J. (1999). The nature of experiment in archaeology. In: Jerem E and Poroszlai I (eds.) *Archaeology of the Bronze and Iron Age. Experimental Archaeology*. Proceedings of the International Archaeological Conference Szazhombatta, 3–7 October 1996, (Archaeolingua Vol. 9), pp. 387–396. Archaeolingua, Budapest.
- Robinson DE (2002) Domestic burnt offerings and sacrifices at Roman and pre-Roman Pompeii, Italy. *Vegetation History and Archaeobotany* 11: 93–99.
- Rösch M, Ehrmann O, Herrmann L, et al. (2002) An experimental approach to Neolithic shifting cultivation. *Vegetation History and Archaeobotany* 11: 143–154.
- Ruas MP, Bouby L, Py V, and Cazes JP (2005) An 11th century AD burnt granary at La Gravette, south-western France: preliminary archaeobotanical results. *Vegetation History and Archaeobotany* 14: 416–426.
- Salamini F, Ozkan H, Brandolini A, Schafer-Pregl R, and Martin W (2002) Genetics and geography of wild cereal domestication in the Near East. *Nature Reviews Genetics* 3: 429–441.
- Samuel D (1996) Investigation of Ancient Egyptian baking and brewing methods by correlative microscopy. *Science* 273: 488–490.
- Samuel D (1996) Archaeology of Ancient Egyptian beer. *Journal of the American Society of Brewing Chemistry* 54: 3–12.
- Samuel D (1999) Bread making and social interactions at the Amarna Workmen's village, Egypt. *World Archaeology* 31: 121–144. (Taylor and Francis Ltd, <http://www.tandf.co.uk/journals>).
- Schibler J and Jacomet S (2005) Fair-weather archaeology? A possible relationship between climate and the quality of archaeological sources. In: Gronenborn D (ed.) *Climate Variability and Culture Change in Neolithic Societies of Central Europe 6700–2200 cal BC* (RGZM Tagungen Vol. 1), pp. 27–39. Mainz: Römisch-Germanisches Zentralmuseum.
- Schiffer MB (1987) *Formation Processes of the Archaeological Record*. Albuquerque, NM: University of New Mexico Press.
- Schweingruber FH (1976) *Prähistorisches Holz. Die Bedeutung von Holzfunden aus Mitteleuropa für die Lösung Archäologischer und Vegetationskundlicher Probleme*. Bern: Verlag Paul Haupt (Academica Helvetica 2).
- Schweingruber FH (1990) *Anatomie Europäischer Hölzer – Anatomy of European Woods*. Bern: Paul Haupt Verlag.
- Smith BD (1998) *The Emergence of Agriculture*. New York: Scientific American Library.
- Terral JF (2000) Exploitation and management of the olive tree during prehistoric times in Mediterranean France and Spain. *Journal of Archaeological Science* 27: 127–134.
- Valamoti SM (2002) Food remains from Bronze Age Archondiko and Mesimeriani Toumba in northern Greece? *Vegetation History and Archaeobotany* 11: 17–22.
- Valamoti SM (2003) Neolithic and Early Bronze Age “food” from northern Greece: The archaeobotanical evidence. In: Parke Pearson M (ed.) *Food, Culture and Identity in the Neolithic and Early Bronze Age*, pp. 97–111. Oxford: Archaeopress (BAR International Series Vol. 1117).
- van der Veen M (ed.) (2003) *Luxury Foods*. London: Routledge (World Archaeology Vol. 34/3).
- van der Veen M (1992) *Crop Husbandry Regimes: An Archaeobotanical Study of Farming in Northern England 1000 BC–AD 500*. Sheffield: JR Collins Publications (Sheffield Archaeological Monographs 3).
- van der Veen M (2004) The merchants' diet: Food remains from Roman and medieval Quseir al-Qadim. In: Lunde P and Porter A (eds.) *Trade and Travel in the Red Sea Region. Proceedings of Red Sea Project I*, pp. 123–130. Oxford: Archaeopress (BAR International Series Vol. 1269).
- van der Veen M and Fjeller NRJ (1982) Sampling seeds. *Journal of Archaeological Science* 9: 287–298.
- Van der Veen M and Jones GEM The production and consumption of cereals: A question of scale. *Vegetation History and Archaeobotany* (In press).
- van Zeist W, Wasylikowa K, and Behre KE (1991) *Progress in Old World Palaeoethnobotany. A Retrospective View on the Occasion of 20 years of the International Work Group for Palaeoethnobotany*. Rotterdam: Balkema.
- Vandorpe P and Jacomet S (2007) Comparing different pre-treatment methods for strongly compacted organic sediments prior to wet-sieving: A case study on Roman waterlogged deposits. *Environmental Archaeology* 12: 207–214.
- Weiss E, Wetterstrom W, Nadel D, and Bar-Yosef O (2004) The broad spectrum revisited: Evidence from plant remains. *Proceedings of the National Academy of Sciences of the United States of America* 101: 9551–9555.
- Wilkinson K and Stevens C (2003) *Environmental Archaeology: Approaches. Techniques & Applications*: Tempus, Stroud.
- Willcox G (2004) Measuring grain size and identifying Near Eastern cereal domestication: Evidence from the Euphrates valley. *Journal of Archaeological Science* 31: 145–150.
- Willcox G (2005) The distribution, natural habitats and availability of wild cereals in relation to their domestication in the Near East: Multiple events, multiple centres. *Vegetation History and Archaeobotany* 14: 534–541.
- Willerding U (1999) Heilmittel und Heilkräuter. *Reallexikon der Germanischen Altertumskunde* 14: 208–233.
- Willerding U (1999) Heu. *Reallexikon der Germanischen Altertumskunde* 14: 510–526.
- Zach B (2002) Vegetable offerings on the Roman sacrificial site in Mainz, Germany – short report on the first results. *Vegetation History and Archaeobotany* 11: 101–106.
- Zohary D and Hopf M (2000) *Domestication of Plants in the Old World. The Origin and Spread of Cultivated Plants in West Asia, Europe and the Nile Valley*. Oxford: Clarendon Press.

Validation of Pollen Studies

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Plant macrofossils and pollen play complementary roles in paleoecology. Each has its own advantages and shortcomings. Used separately, they provide valuable information about past vegetation. Used together, the two types of data can corroborate or, occasionally, challenge each other, and they can provide more detailed and refined understanding of past vegetation at multiple spatial scales. In this article, we briefly compare and contrast the flora- and vegetation-sensing properties of pollen and macrofossil data and then present a series of case studies in which the two data sets have been used together profitably at spatial scales ranging from local to continental. We focus on macrofossils from ‘wet’ sediments (lakes, peatlands, buried soils, fluvial deposits, etc.) and do not explicitly treat macrofossils from rodent middens and archaeological sites.

Vegetation-Sensing Properties of Pollen and Plant Macrofossil Data

Pollen grains comprise a distinct and uniform stage in the life history of seed plants—the gametophyte generation. Pollen grains are small, with a restricted range of sizes (5–150 μm), and most are packaged in sporopollenin, a refractory organic compound. Plant macrofossils from seed plants represent the sporophyte generation. However, because seed plant sporophytes are multicellular, differentiated entities, plant macrofossils can take any of a variety of forms: seeds, fruits, anthers, microsporangia, megaspores, ovulate cones, leaves, needles, petioles, bud scales, buds, twigs, branches, etc. The size range is potentially vast, from seeds $<1\text{ mm}^3$ to buried trees $>20\text{ m}^3$. Materials are also variable, although macrofossils preserved in sediments tend to be composed of the more refractory compounds (particularly cellulose and lignins).

Pollen grains are typically morphologically invariant among species within most anemophilous genera and among many genera within certain families. In contrast, interspecific morphological variation of sporophyte plant organs is typically large since plant species are generally described and distinguished based on sporophyte morphology. Accordingly, many plant macrofossils can be identified to species, in contrast to the corresponding pollen grains. This imparts one of the three unique features of plant macrofossil data compared to pollen: high taxonomic resolution. Macrofossils allow us to surmount the taxonomic smoothing frequently imposed by pollen data.

The second unique feature of plant macrofossils is their local nature. Plant macrofossils generally do not travel far when airborne; the primary mode of transport beyond 10^1 – 10^2 m is by surface waters or animals. For depositional basins with no inflowing streams, most plant macrofossils

derive from within 10^1 – 10^2 m of the shore. For depositional environments with a fluvial component, macrofossils must have been derived from plants growing somewhere within the catchment above the point of deposition. In contrast, pollen grains can be transported by wind for distances up to 10^4 km , although more typical travel distances are on the order of 10^1 – 10^5 m . Pollen assemblages in a lake or wetland basin represent a distance-weighted integration of vegetation surrounding the basin, with significant representation of distant vegetation (i.e., $>10^3\text{ m}$). In contrast, plant macrofossils provide a more-local sampling of vegetation, undoubtedly distance-weighted, but with little or no representation beyond ca. 10^2 m (Table 1).

Many plant taxa are poorly represented in pollen assemblages due to low productivity, poor dispersal, poor preservation, or adaptations for zoophilous transport. Many of these taxa—*Pseudotsuga*, *Larix*, *Abies*, *Dryas*, *Populus*, *Oxyria*, *Polygonum*, and *Nymphaeaceae*—are well represented in plant macrofossil assemblages from lakes and other depositional environments. Thus, plant macrofossils can provide information about taxa for which the pollen record may be unreliable—the third unique feature of macrofossil data. This complementarity does not hold for all taxa, however; some plants are poorly represented in both pollen and macrofossil assemblages. For example, *Liriodendron*, an important tree species of eastern North America, is entomophilous and hence almost completely absent from pollen records, regardless of local or regional abundance. Fruits, bud scales, and leaves of *Liriodendron* are large and poorly dispersed, and hence show up only rarely in macrofossil assemblages. Some taxa well represented in pollen assemblages (e.g., *Quercus*, *Carya*, and *Corylus*) are poorly represented as macrofossils in many depositional settings.

Macrofossil and Pollen Representation Within and Among Depositional Environments

Representation of plant macrofossils is far more subject to influences of site type and local depositional environment than pollen. Pollen grains may be subjected to some secondary sorting by fluvial transport and sediment focusing. However, pollen assemblages from diverse depositional environments (lakes, wetlands, forest-floor moss polsters, soil samples, alluvium, caves, and rodent middens) are generally similar within a region due to widespread airborne transport and mixing prior to deposition. In contrast, macrofossils are not subject to the same degree of mixing due to large size and poor dispersal. Furthermore, macrofossils are more subject to sorting by airborne and fluvial dispersal processes than pollen (Spicer, 1989; Greenwood, 1991).

Table 1 Systematic studies of modern representation of terrestrial plant macrofossils in lake sediments

Birks HH (1973) Modern macrofossil assemblages in lake sediments in Minnesota. In: Birks HJB and West RG (eds.) <i>Quaternary Plant Ecology</i> , pp. 173–189. Oxford, UK: Blackwell.
Drake H and Burrows CJ (1980) The influx of potential macrofossils into Lady Lake, north Westland, New Zealand. <i>New Zealand Journal of Botany</i> 18: 257–274.
Dunwiddie PW (1987) Macrofossil and pollen representation of coniferous trees in modern sediments from Washington. <i>Ecology</i> 67: 58–68.
Hill RS and Gibson N (1986) Distribution of potential macrofossils I: Lake Dobson, Tasmania. <i>Journal of Ecology</i> 74: 373–384.
Jackson ST (1989) Postglacial vegetational changes along an elevational gradient in the Adirondack Mountains (New York): A study of plant macrofossils. <i>New York State Museum and Science Service Bulletin</i> 465.
Wainman N and Mathewes RW (1990) Distribution of plant macroremains in surface sediments of Marion Lake, southwestern British Columbia. <i>Canadian Journal of Botany</i> 68: 364–373.

Lakes

Lake sediments are generally the most widely used sources of both pollen and plant macrofossil data. Within-lake variation in pollen assemblages is significant, but the variation is generally small relative to variation among lakes. In contrast, plant macrofossil assemblages within a single lake vary substantially, depending on water depth, distance from shore, and vegetation composition on the nearest shore. Most macrofossil deposition to the water surface will occur close to shore, whereas the best environment for macrofossil preservation in sediments will be in deep waters, which tend to be offshore (Watts, 1978). Moderate-sized lakes (5–100 ha) with relatively gentle bottom slopes will have few macrofossils preserved near shore, where influx is highest, because of high decomposition and low sediment accumulation rates. Deeper waters with good preservation and high sediment accumulation rates will also have few macrofossils because of poor offshore dispersal. Highest macrofossil concentrations will occur in intermediate depths, deep enough to preserve macrofossils and close enough to shore to ensure high macrofossil influx. Since most paleoecologists obtain cores from the deepest portions of lakes, it should come as no surprise that macrofossils are often absent or rare in sediment cores.

Significant sorting of macrofossils occurs with distance from shore due to differential dispersal (and, in some cases, flotation) properties of various plant organs. Large, coarse materials (large, gravity- or animal-dispersed seeds and fruits, ovulate cones, large leaves, branches, etc.) are rarely dispersed more than a few tens of meters from shore, where they are likely to sink and then decompose in shallow benthos. Smaller organs (small seeds and fruits, conifer needles, and bud scales) are dispersed farther from shore and hence are better represented in offshore sediments. In deep waters more than 100 m from shore, only small fragments of small-diameter conifer needles, papery bud scales, and small winged fruits (e.g., *Betula papyrifera* and *B. pubescens*) are typically present. The relative contributions to offshore deposition of airborne dispersal, flotation, and skating across winter ice are inadequately known.

Selection of coring sites for macrofossils thus represents a trade-off between maximizing macrofossil preservation (deep water) and deposition (near shore). For large lakes, it is advisable to run a transect of short cores or dredge samples to determine the ideal site for macrofossil deposition. Alternatively, careful selection of lakes can yield good macrofossil results. In particular, small or moderate-sized lakes with deep waters 10–50 m offshore and steep sides can have very high macrofossil concentrations, yielding excellent records of upland terrestrial vegetation.

Lake sediments are typically sampled using piston-corers with diameters ranging from 5 to 10 cm. Thus, sample volume is limited, and some plant organs of moderate size or intermediate dispersal properties (e.g., *Acer* samaras (one seeded, winged fruit), most deciduous leaves, and coarse conifer needles) are not reliably sampled: Absence from sediments cannot be attributed to absence from vegetation. However, other plant organs (winged seeds of *Betula*, *Alnus*, and many conifers, conifer needles, and many bud scales) are consistently represented in modern lake sediments (Figure 1) and between adjacent stratigraphic samples within a core (Figure 2). In such cases, absence from sediments can be inferred to indicate absence from or rarity in vegetation near the lakeshore. Careful selection of lakes and coring sites within lakes, together with use of large-diameter coring devices (7.5 or 10 cm), can yield records in which the presence/absence of macrofossils in the stratigraphic column can be applied directly to the presence/absence in adjacent vegetation. Also, as discussed later, macrofossil presence/absence and abundance patterns often show close stratigraphic correspondence to pollen abundance of the corresponding taxa.

Wetlands

Macrofossils in wetland sediments can provide excellent records of local wetland vegetation. Shallow water or subsurface water tables limit the extent of waterborne dispersal of macrofossils, and floating-leaved and emergent plants further reduce waterborne dispersal by providing a series of traps and baffles. This has the advantage of amplifying local representation of macrofossils, which can be used to examine local vegetational and hydrological dynamics within a wetland (Watts and Winter, 1966; Singer et al., 1996; McMullen et al., 2004). In fact, species richness of wetland/aquatic macrofossil assemblages can be very high in wetland sediments. In general, macrofossils from surrounding uplands are restricted to plant organs that are particularly well dispersed by airborne means (small samaras, winged seeds, bud scales, and some conifer needles). Sediments are usually sampled using coring devices, with the sample size limitations described previously for lakes. Wetlands with low sediment accumulation rates, oxygenated waters, and/or sustained drawdown periods can have sediments depauperate in macrofossils, as can humified bog peats.

Fluvial Sediments

Fluvial sediments have been underutilized as sources of plant macrofossil data. Concerns with fluvial transport and discontinuous deposition at individual sites have inhibited their use. However, fluvial deposits from low-order (i.e., headwater) streams restrict the potential source area for macrofossils, and

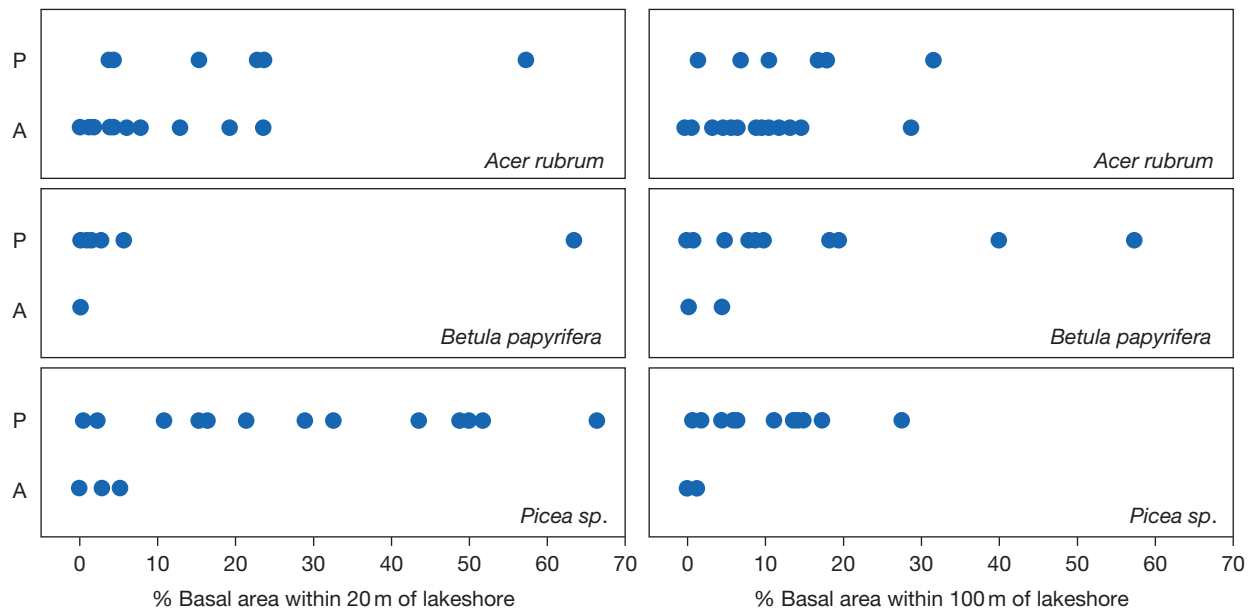


Figure 1 Scatter plots showing the relationship between the presence (P) and absence (A) of macrofossils in modern sediments of 22 lakes and abundance of trees within a 20- and 100-m radius of the lake margin. All sites are small lakes (0.1–2.0 ha) in the northeastern United States (New York, Rhode Island, Massachusetts, and Connecticut). Surface samples comprise 100–200 cm³ obtained using an Ekman dredge from the center of the lake. *Acer rubrum* macrofossils comprise samaras and leaves. *Betula papyrifera* macrofossils comprise samaras and catkin bracts. This taxon includes *B. populifolia* at several sites. Samaras and bracts of the two species are morphologically similar but distinguishable. Because the two species are closely related, have similar growth form, and their macrofossils have similar dispersal properties, we lumped them for purposes of this figure. *Picea* sp. macrofossils include needles, sterigmata (needle stems), seeds, microsporangia, and twigs. Species at the sites include *P. rubens* and *P. mariana*, which are closely related and morphologically indistinguishable for macrofossil types other than cones. *Acer rubrum* macrofossils are large and poorly dispersed relative to the other two species. The occurrence of *A. rubrum* macrofossils in sediment samples of this size (typical of sediment cores) is inconsistent. The absence of macrofossils cannot be construed as indicating the absence of nearby populations. Larger sediment samples might increase the probability of macrofossil occurrence. *Betula papyrifera* and *B. populifolia* macrofossils are small, abundantly produced, and well dispersed. Macrofossils are consistently represented in sediments even when local populations are very small. Some sites lacking any *B. papyrifera* or *B. populifolia* trees within 20 m of shore have macrofossils in sediments, dispersed from trees between 20 and 100 m from shore. *Picea* sp. macrofossils are also very well represented in sediments. In general, if small populations are present near shore, macrofossils are present in sediments. *Picea* trees produce a variety of small organs that are well dispersed and well preserved in sediments. In the case of both *Picea* and *B. papyrifera*, the absence of macrofossils in sediment samples of this size, in lakes of this size and morphometry, can be interpreted as indicating the absence or scarcity of nearby populations. Consistent absence among adjacent samples in a core can be construed to indicate the absence of nearby populations. Source: S. T. Jackson, unpublished data.

individual ‘snapshot’ assemblages can be dated and stacked. Fluvial deposits are usually collected from excavations or from streamcut exposures, and hence sample size is not limited by a coring device. Furthermore, fluvial transport and deposition can concentrate large plant organs (leaves, large seeds and fruits, branches, etc.) that are rare in sediments of most lakes. Significantly, the only macrofossil records of several North American trees (e.g., *Quercus*, *Juglans*, *Liriodendron*, and *Carya*) are from fluvial deposits. Paleoecologists occasionally turn to fluvial deposits in the absence of better sites. The exemplary studies of Richard Baker and colleagues in the Upper Mississippi River Valley indicate that macrofossil and pollen studies of fluvial deposits can potentially yield far richer records of upland vegetation change than lakes or wetlands. (see Holocene North America).

Other Depositional Settings

Plant macrofossils are also preserved in a variety of other depositional settings, ranging from glacial and glaciomarine to former land surfaces buried *in situ* with contemporary

vegetation. In most of these depositional settings, samples are collected from exposures, and sample size is not as limited as for cores. Buried *in situ* vegetation can provide a unique window on past community composition and population density, and accompanying pollen assemblages are comparable to samples from modern soils.

Data Expression

Quaternary pollen data are most commonly expressed as percentages of a defined pollen sum. Certainly, virtually all applications of Quaternary pollen data are quantitative; presence/absence applications are generally pre-Quaternary. Plant macrofossil data, by contrast, are more heterogeneous in expression. Data range from presence (anecdotal occurrences) to presence/absence to quantitative. Macrofossil data from sediment cores are typically presented as concentrations (number of specimens per unit volume of sediment) or, occasionally, accumulation rates (numbers deposited per unit area per year). These data can, of course, be reduced to presence/absence data.

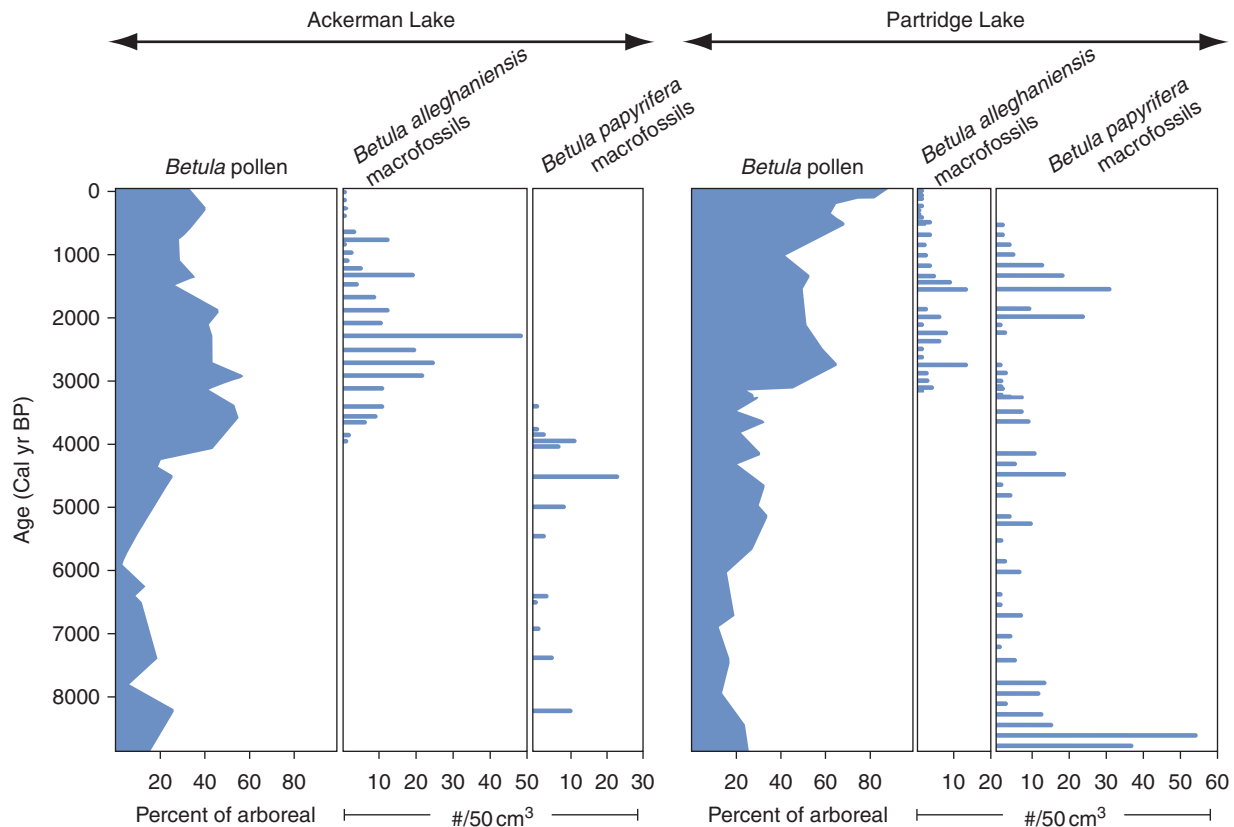


Figure 2 Pollen and macrofossil sequences for *Betula* from two lakes in the Upper Peninsula of Michigan. Ackerman Lake (6 ha) is south of Munising Bay. Steep slopes on the south and east sides, near the coring site, are forested by *Tsuga canadensis*, *Acer saccharum*, *Betula alleghaniensis*, and *Fagus grandifolia*. Partridge Lake (3 ha) is situated near the tip of the Keweenaw Peninsula, 160 km northwest of Ackerman Lake. The steep slopes surrounding Partridge Lake are forested by *A. saccharum*, *B. alleghaniensis*, *Quercus rubra*, *Picea glauca*, and *Pinus strobus*. *Betula* pollen undergoes a dramatic late Holocene increase at both sites, but the increase at Ackerman Lake occurs approximately 1000 years before the increase at Partridge Lake. This would appear to represent a time-transgressive westward migration or expansion of *Betula*, but the pollen data alone render interpretation difficult. Plant macrofossil data clarify the interpretation: The pollen increase at both sites is coincident with first appearance of macrofossils of *B. alleghaniensis*, which persists in both records until the present. Thus, the pollen pattern represents westward expansion of *B. alleghaniensis* populations. *Betula papyrifera* was the sole or dominant *Betula* species before the *B. alleghaniensis* expansion, and it was associated with lower pollen percentages. *Betula papyrifera* is absent from Ackerman Lake sediments after the *B. alleghaniensis* invasion. This probably represents a climate-driven change in disturbance regime rather than competition; *B. papyrifera* regeneration is dependent on large-scale, stand-replacing disturbances (crownfire and extensive windthrow). *Betula papyrifera* persists at Partridge Lake throughout the late Holocene, co-occurring with *B. alleghaniensis*. Stand-replacing disturbances (particularly crownfires) were probably more prevalent near Partridge in the late Holocene; this is consistent with continued high percentages of *Pinus* pollen. Modified (with updating) from Jackson ST and Booth RK (2002) The role of late Holocene climate variability in the expansion of yellow birch in the western Great Lakes region. *Diversity and Distributions* 8: 275–284.

Stratigraphic changes in macrofossil concentrations within a sediment core are generally interpreted as indicating changes in species abundance and vegetation composition. Such interpretations assume that the depositional setting has remained constant. This assumption can be violated in particular if water level changes in the basin have resulted in changes in the preservation environment, distance from the source vegetation along the lakeshore, or density of intervening emergent and aquatic vegetation. However, coring of small deep lakes, or of deep water close to shore, can ensure that this assumption is justified.

Greater difficulties are encountered in comparing assemblages between cores within a lake, or particularly among lakes. Differences in pollen percentages among lakes can be attributed in large part to vegetational differences, whereas differences in pollen concentrations may be attributable to

differences in depositional setting or other factors. Macrofossil concentrations can be compared among lakes along environmental transects, and such comparisons often give sensible results in terms of vegetational patterns. However, such comparisons should be standardized carefully for morphometric and other variables such as distance from shore. Alternatively, a macrofossil sum can be designated and used to calculate macrofossil percentages (Watts and Winter, 1966; Tinner and Kaltenrieder, 2005), facilitating between-site comparison (Jackson and Whitehead, 1991).

For broader comparisons, particularly among different site types, any quantitative expression poses an “apples and oranges” problem, so data are generally reduced to presence/absence. For plant taxa that are particularly well represented in macrofossil assemblages, mapped presence/absence of modern macrofossils closely approximates modern geographic

distributions, and hence snapshot maps of past time intervals can be used to map range dynamics of these species (Jackson et al., 1997).

Tandem Applications of Pollen and Plant Macrofossil Data

Macrofossil and Pollen Data from Individual Sites

Innumerable studies comparing pollen and macrofossil stratigraphy from individual sediment cores have been published, and records of the same taxa generally show parallel stratigraphic changes in abundance. In these cases, the pollen and macrofossils provide mutual corroboration. The two data types are subject to completely different taphonomic processes and record surrounding vegetation at vastly different spatial scales. The fact that pollen and macrofossils of the same taxa so often track each other temporally gives us confidence that both data sets are tracking changes in upland vegetation.

Macrofossils provide far more than corroboration of pollen data, and vice versa: The differential taphonomic biases, differential spatial integration, and differential taxonomic resolution can be used in complementary ways to refine our interpretations of past flora, vegetation, and environment. We provide a few representative examples here.

Tandem studies of pollen and plant macrofossil analyses from the same sites are being used to study late Holocene dynamics of birch species in the western Great Lakes region. Pollen records from the region document increasing birch (*Betula*) pollen in the late Holocene. However, two species of birch trees (*B. papyrifera* and *B. alleghaniensis*) with morphologically similar pollen grains occur in the region, hampering detailed ecological and climatic inferences from the birch pollen record. Both species produce morphologically distinct, highly dispersible samaras and catkin bracts in large numbers. These macrofossils are well represented in sediments of lakes and small peatlands in the region. Detailed studies of pollen and macrofossils from lake sediment cores indicate that *B. papyrifera* has been an important component of regional upland vegetation throughout the Holocene, and that *B. alleghaniensis* has migrated into the region only during the past 4500 years (Figure 2). This latter migration resulted in a time-transgressive increase in *Betula* pollen percentages from east to west across the region (Jackson and Booth, 2002). Similar integrated studies of pollen and macrofossils have helped refine interpretation of pollen records from numerous other regions. The increased taxonomic precision imparted by macrofossil records coupled with the regional smoothing of pollen records can lead to detailed reconstructions of vegetation dynamics that could not have been inferred using either method independently.

An extreme example of how the increased taxonomic resolution of macrofossils can alter interpretations of pollen assemblages comes from an extinct species, *Picea critchfieldii* (Jackson and Weng, 1999). Pollen assemblages from much of the unglaciated southeastern United States show high percentages of *Picea* pollen, routinely ascribed to the boreal and montane *Picea* species that grow in eastern North America today. *Picea* macrofossils found at many of these sites corroborated this interpretation, which, together with other evidence, led to

the perception of widespread boreal forest or mixed boreal conifers/cool-temperate hardwoods at low latitudes (to 30° N) with accompanying drastically reduced temperatures. More detailed morphological and anatomical studies of the *Picea* cones and needles, respectively, led to the discovery that an extinct species, *P. critchfieldii*, occurred over much of the region and was the only *Picea* species at some sites (Jackson and Weng, 1999). This has led to reevaluation of the Pleistocene biogeography, paleoecology, and paleoclimatology of the entire region (Jackson et al., 2000). Whether the pre-Holocene demise of *P. critchfieldii* is one of many plant extinctions or a singular occurrence remains obscure. However, Pleistocene macrofossils of other taxa should be scrutinized to determine whether they belong to extant species.

Macrofossils are also important in detecting palynologically cryptic populations or taxa. Obvious examples include the taxa that are poorly represented by pollen in modern assemblages due to low productivity, poor dispersal, or rapid decomposition (e.g., *Larix*, *Pseudotsuga*, and *Populus*). However, macrofossil data reveal that pollen assemblages, by virtue of their wide spatial integration, may miss important aspects of vegetation and flora. This has become particularly obvious in glacial-age and late-glacial settings (Jackson et al., 2000; Willis et al., 2000; Jackson and Williams, 2004; Willis and Van Andel, 2004). Pollen data for the last glacial period in Europe indicate extensive treeless vegetation; low representation of arboreal pollen types suggests the absence of trees anywhere north of the southern Balkans and Iberian and Italian peninsulas (Huntley et al., 2003). However, charcoal of woody plants (*Picea*, *Pinus*, and *Betula*) has been found in glacial-age paleosols and archaeological sites in central Europe (Willis et al., 2000; Willis and Van Andel, 2004). Similarly, macrofossils of *Pinus strobus*, *B. papyrifera*, *Fagus grandifolia*, *Quercus* spp., and *Carpinus caroliniana* occur in glacial-age sediments of the unglaciated southeastern United States, even though pollen is absent or occurs in trace amounts at the same sites (Jackson et al., 2000). All of these taxa, both European and North American, are prolific pollen producers that are well represented in modern pollen assemblages. Their poor representation in the Pleistocene assemblages remains enigmatic, potentially attributable to their occurrence in small, scattered populations (e.g., in locally wet sites) or to differential effects of reduced CO₂ on physiognomy and reproductive allocation (Jackson and Williams, 2004). Phylogeographic evidence, particularly from North America, suggests that glacial-age populations of temperate trees may have been substantially farther north than suggested by pollen records (McLachlan et al., 2005).

The apparent discrepancies between glacial-age records of pollen and macrofossils arise in part from the vast differences in spatial integration and source area of the respective data types. These differences have manifestations in the Holocene as well and can be used to tease apart scale-dependent patterns in the paleoecological record. We discuss some examples in the following sections that involve spatial arrays or synoptic networks of sites. Here, we discuss a few site-based case studies from the Holocene. Pollen data are effective in indicating population changes, including invasion/expansion, integrated over a large area. The efficacy of pollen data in detecting small, scattered colonizing populations has been the subject of considerable discussion (McLachlan and Clark, 2004).

Comparison of Holocene pollen and macrofossil records indicates that pollen data are in many (and perhaps most) cases ineffective at determining the timing of colonization. Macrofossil occurrences have been observed to precede increases in pollen percentages beyond trace levels by hundreds or thousands of years (Peteet, 1991). Peteet's study, together with the glacial-age anomalies discussed previously, indicates that small, scattered populations of tree species can go undetected in pollen records—an observation consistent with what is currently known about pollen dispersal and representation.

Because of long-distance pollen dispersal, plant macrofossil occurrences seldom precede high pollen percentages in migrating populations. For example, in the Adirondack Mountains of New York, *Pinus strobus* pollen percentages increased at least several hundred years before *Pinus strobus* macrofossils first appeared in sediments (Jackson and Whitehead, 1991). However, in northern Michigan, the first occurrence of *P. strobus* macrofossils in lakes is synchronous with the first occurrence of pollen beyond trace amounts (Booth et al., 2004). The difference may stem from the occurrence of extensive lowlands surrounding the Adirondack highlands, which may have fostered extensive *P. strobus* populations before the species migrated upslope.

Macrofossils can also indicate leads and lags in local population response during periods of decline. Macrofossil and pollen data reveal, for example, that as prairie expanded to replace mesic forest in uplands of southeastern Minnesota during the mid-Holocene, local populations of *Ulmus* persisted on floodplains for at least 500 years after the regional shift (Baker et al., 2002). This may represent persistence of one or two tree generations on the more favorable floodplain sites or gradual climate change that passed thresholds for forest growth on uplands first and wetter sites later. If local populations persist, they become 'relict' populations, and their development can be observed by comparing pollen and macrofossil data from individual sites (Watts, 1979) and in mapped networks (Figure 3).

Differences in the source area of pollen and macrofossil records also have enabled comparative investigations of the dynamics of wetland and upland systems. For example, by comparing the timing of changes in pollen derived from upland trees and changes in the concentrations of locally deposited wetland plant macrofossils, several studies have assessed the relative importance of climatic versus autogenic controls on peatland development and dynamics (Singer et al., 1996; Campbell et al., 1997; Booth et al., 2004).

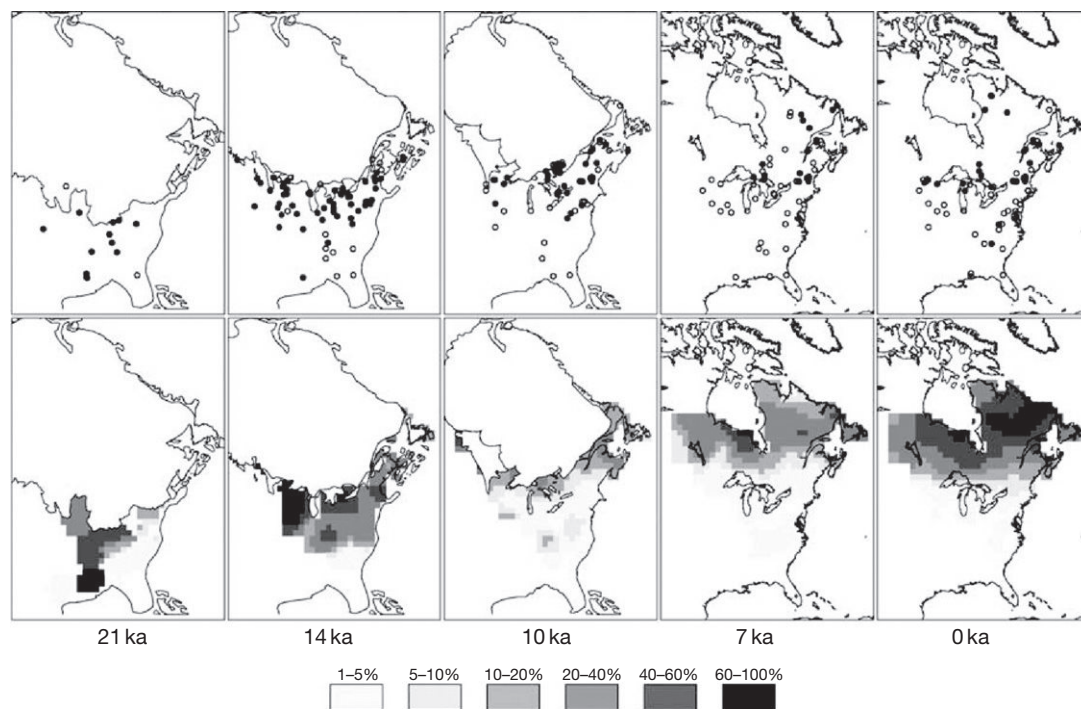


Figure 3 Mapped patterns of macrofossil presence/absence (top) and pollen percentages (bottom) of *Picea* in eastern North America since the Last Glacial Maximum. Open and closed circles in the top map respectively represent the absence and presence of *Picea* macrofossils at sites corresponding to the respective time periods. Time windows for the mapping are broad, encompassing approximately 1000 years for the Holocene (10 ka, 6 ka, and the present), 2000 years for the Late Glacial (14 ka), and 3000 years for the Last Glacial Maximum (21 ka). Comparison of the maps shows general correspondence between pollen and macrofossils. Disappearance of macrofossils from some individual sites lags behind pollen declines during the northward retreat of the southern range limit of *Picea* in the Late Glacial and early Holocene. Local *Picea* populations occurred across a wide swath south of the ice margin during the early Holocene (10 ka), despite the poor representation of pollen in sediments of this period. These patterns probably arise from the persistence of local *Picea* populations in favorable sites (e.g., in wetlands and along lake margins) after regional, upland populations had declined. Modified from Jackson ST, Overpeck JT, Webb T III, Keatts SE, and Anderson KH (1997) Mapped plant macrofossil and pollen records of Late Quaternary vegetation change in eastern North America. *Quaternary Science Reviews* 16: 1–70; and Jackson ST, Webb RS, Anderson KH, et al. (2000) Vegetation and environment in unglaciated eastern North America during the Last Glacial Maximum. *Quaternary Science Reviews* 19: 489–508.

Regional Studies: Site Transects and Networks

Plant macrofossils and pollen have been used together in a number of regional studies, nearly all involving elevational gradients. In typical montane regions, most of the land surface is at low elevations, and frequently the highest pollen productivity is in low-elevation vegetation (e.g., forests). Widespread pollen dispersal tends to smooth vegetational patterns along elevational gradients, and discrimination of elevation-specific vegetation composition and elevational boundaries of individual species and forest types is difficult. Site transects using both pollen and plant macrofossil data have led to elevationally precise reconstructions of vegetation patterns and species ranges (Table 2). Intersite comparisons impose challenges of macrofossil data expression. Montane studies have variously used presence/absence, concentrations in sediments, and macrofossil percentages based on a defined sum (Table 2). In general, these studies confirm the differential spatial smoothing of pollen and macrofossil data, with the macrofossils providing spatially detailed records of vegetation composition and species ranges.

Macrofossils have not been used as extensively in regional studies focused on landscape patterns (e.g., edaphic mosaics) or geographic gradients (e.g., studies of species invasions). Macrofossils from packrat middens show the potential power of these kinds of studies (Betancourt et al., 1990; Lyford et al., 2003), and similar study designs should be developed using

macrofossils from lakes and other sites (Peñalba and Payette, 1997). (see Rodent Middens).

Synoptic-Scale Studies: Mapped Networks

Synoptic-scale studies utilizing pollen data have been a tradition in Quaternary ecology since the pioneering work of Szafer and Sears, and mapping of pollen data at regional to continental scales is now routine (Williams et al., 2004). Plant macrofossils, in contrast, have not been widely applied in synoptic-scale studies. One of the challenges in doing so, of course, is determining data expression. Presence/absence has worked well in recent applications, and the most comprehensive applications to date have been in eastern North America (Jackson et al., 1997, 2000). These studies show generally corresponding patterns between pollen and macrofossil data (Figure 3). For several taxa (e.g., *Pinus*, *Picea*, and *Betula*), the higher taxonomic resolution of the macrofossil data clarifies interpretational uncertainties in the pollen mapping. Also, the maps reveal patterns of occurrence of taxa that are cryptic in the pollen record, particularly for the Last Glacial Maximum. These studies of a select number of taxa provide an indication of the potential power of systematic comparison of mapped patterns of pollen and macrofossil data.

Other mapping efforts involving pollen and macrofossil data include those by Tarasov et al. (1998), Gunin et al. (1999), McLachlan and Clark (2004), and Edwards et al. (2005). The latter study shows how pollen and macrofossils can give contrasting and complementary views of regional vegetation composition and physiognomy.

Table 2 Systematic studies using pollen and plant macrofossil data along elevational gradients in mountainous regions

Anderson RS (1990) Holocene forest development and paleoclimates within the central Sierra Nevada, California. <i>Journal of Ecology</i> 78: 470–489.
Anderson RS (1996) Postglacial biogeography of Sierra lodgepole pine (<i>Pinus contorta</i> var. <i>murrayana</i>) in California. <i>Écoscience</i> 3: 343–351.
Barnekow L and Sandgren P (2001) Paleoclimate and tree-line changes during the Holocene based on pollen and plant macrofossil records from six lakes at different altitudes in northern Sweden. <i>Review of Paleobotany and Palynology</i> 117: 109–118.
Davis MB, Spear RW, and Shane LCK (1980) Holocene climate of New England. <i>Quaternary Research</i> 14: 240–250.
Dunwiddie PW (1986) A 6000-year record of forest history on Mount Rainier, Washington. <i>Ecology</i> 67: 58–68.
Jackson ST (1989) Postglacial vegetational changes along an elevational gradient in the Adirondack Mountains (New York): A study of plant macrofossils. <i>New York State Museum and Science Service Bulletin</i> 465.
Jackson ST and Whitehead DR (1991) Holocene vegetation patterns in the Adirondack Mountains. <i>Ecology</i> 72: 641–653.
Spear RW (1989) Late-Quaternary history of high-elevation vegetation in the White Mountains of New Hampshire. <i>Ecological Monographs</i> 59: 125–151.
Spear RW, Davis MB, and Shane LCK (1994) Late Quaternary history of low- and mid-elevation vegetation in the White Mountains of New Hampshire. <i>Ecological Monographs</i> 64: 85–109.
Tinner W and Kaltenrieder P (2005) Rapid responses of high-mountain vegetation to early Holocene environmental changes in the Swiss Alps. <i>Journal of Ecology</i> 93: 936–947.
Weng C and Jackson ST (1999) Late-glacial and Holocene vegetation and climate history of the Kaibab Plateau, northern Arizona. <i>Paleogeography, Paleoclimatology, Paleoecology</i> 153: 179–201.

Future Prospects

Plant macrofossils can serve as equal or even superior partners with pollen data in many contexts and for many questions. However, despite widespread application in Quaternary paleoecology and paleoclimatology, plant macrofossils continue to play a subservient role to pollen in most studies. The primary exceptions are fossil rodent middens and some studies of fluvial deposits, ombrotrophic bogs, and buried soils. Plant macrofossils have been largely underutilized in other contexts, particularly lakes. Carefully and creatively designed studies using macrofossils and pollen together can greatly advance our understanding of vegetational and biogeographic history. Macrofossil analysis should be incorporated at early stages of paleoecological study design, site selection, and core location rather than as an ad hoc addition to studies centering on pollen and/or paleolimnology. This may require more sites (the best macrofossil sites may not be the most suitable for pollen records and vice versa) and multiple cores (to maximize macrofossil representation and temporal resolution simultaneously). Utilization of macrofossil-rich deposits, such as fluvial sediments, can lead to great advances. Such deposits record species poorly represented in lacustrine sediments and can yield large quantities of megascopic macrofossils capable of species identification. Pre-Quaternary paleobotanists focus on these kinds of sediments, with excellent results, and Quaternary paleoecologists would benefit from adopting similar approaches when possible. Plant macrofossil study in subtropical and tropical regions is vastly underdeveloped; few studies have assessed the

potential for tropical macrofossil analysis. Continued development of public-access macrofossil databases (e.g., www.ncdc.noaa.gov/paleo/plantmacros.html) is needed to support synoptic-scale studies, and macrofossil analysts must continue to be active contributors and users of these databases. Finally, systematic studies of macrofossil taphonomy are needed to support inferences and test assumptions.

Acknowledgments

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See also: Plant Macrofossil Introduction. **Plant Macrofossil Methods and Studies:** Mire and Peat Macros; Surface Samples, Taphonomy, Representation. **Plant Macrofossil Records:** Greenland; Holocene North America.

References

- Baker RG, Bettis EA III, Denniston RF, et al. (2002) Holocene paleoenvironments in southeastern Minnesota—Chasing the prairie–forest ecotone. *Palaeogeography, Palaeoclimatology, Palaeoecology* 177: 103–122.
- Betancourt JL, Van Devender TR, and Martin PS (eds.) (1990) *Packrat Middens: The Last 40,000 Years of Biotic Change*. Tucson: University of Arizona Press.
- Booth RK, Jackson ST, and Gray CED (2004) Paleoeecology and high-resolution paleohydrology of a kettle peatland in Upper Michigan. *Quaternary Research* 61: 1–13.
- Campbell DR, Duthie HC, and Warner BG (1997) Post-glacial development of a kettle-hole peatland in southern Ontario. *Ecoscience* 4: 404–418.
- Edwards ME, Brubaker LB, Lozhkin AV, and Anderson PM (2005) Structurally novel biomes: A response to past warming in Beringia. *Ecology* 86: 1696–1703.
- Greenwood DR (1991) The taphonomy of plant macrofossils. In: Donovan SK (ed.) *The Processes of Fossilization*, pp. 141–169. New York: Columbia University Press.
- Gunin PD, Vostokova EA, Dorofeyuk NI, Tarasov PE, and Black CC (1999) *Vegetation Dynamics of Mongolia*. Dordrecht, The Netherlands: Kluwer.
- Huntley B, Alfano MJ, Allen JRM, et al. (2003) European vegetation during Marine Oxygen Isotope Stage 3. *Quaternary Research* 59: 195–212.
- Jackson ST and Booth RK (2002) The role of late Holocene climate variability in the expansion of yellow birch in the western Great Lakes region. *Diversity and Distributions* 8: 275–284.
- Jackson ST and Weng C (1999) Late Quaternary extinction of a tree species in eastern North America. *Proceedings of the National Academy of Sciences, U.S.A.* 96: 13847–13852.
- Jackson ST and Whitehead DR (1991) Holocene vegetation patterns in the Adirondack Mountains. *Ecology* 72: 641–653.
- Jackson ST and Williams JW (2004) Modern analogs in Quaternary paleoecology: Here today, gone yesterday, gone tomorrow? *Annual Review of Earth and Planetary Sciences* 32: 495–537.
- Jackson ST, Overpeck JT, Webb T III, Keatts SE, and Anderson KH (1997) Mapped plant macrofossil and pollen records of Late Quaternary vegetation change in eastern North America. *Quaternary Science Reviews* 16: 1–70.
- Jackson ST, Webb RS, Anderson KH, et al. (2000) Vegetation and environment in unglaciated eastern North America during the Last Glacial Maximum. *Quaternary Science Reviews* 19: 489–508.
- Lyford ME, Jackson ST, Betancourt JL, and Gray ST (2003) Influence of landscape structure and climate variability on a late Holocene plant migration. *Ecological Monographs* 73: 567–583.
- McLachlan JS and Clark JS (2004) Reconstructing historical ranges with fossil data at continental scales. *Forest Ecology and Management* 197: 139–147.
- McLachlan JS, Clark JS, and Manos PS (2005) Molecular indicators of tree migration capacity under rapid climate change. *Ecology* 86: 2088–2098.
- McMullen JA, Barber KE, and Johnson B (2004) A paleoecological perspective of vegetation succession on raised bog microforms. *Ecological Monographs* 74: 45–77.
- Peñalba MC and Payette S (1997) Late-Holocene expansion of eastern larch (*Larix laricina* [Du Roi] K. Koch.) in northwestern Québec. *Quaternary Research* 48: 114–121.
- Peteet DM (1991) Postglacial migration history of lodgepole pine near Yakutat, Alaska. *Canadian Journal of Botany* 69: 786–796.
- Singer DK, Jackson ST, Madsen BJ, and Wilcox DA (1996) Differentiating climatic and successional influences on long-term development of a marsh. *Ecology* 77: 1765–1778.
- Spicer RA (1989) The formation and interpretation of plant fossil assemblages. *Advances in Ecological Research* 16: 95–191.
- Tarasov PE, Webb T III, and Andreev AA (1998) Present-day and mid-Holocene biomes reconstructed from pollen and plant macrofossil data from the former Soviet Union and Mongolia. *Journal of Biogeography* 25: 1029–1053.
- Tinner W and Kaltenrieder P (2005) Rapid responses of high-mountain vegetation to early Holocene environmental changes in the Swiss Alps. *Journal of Ecology* 93: 936–947.
- Watts WA (1978) Plant macrofossils and Quaternary paleoecology. In: Walker D and Guppy JC (eds.) *Biology and Quaternary Environments*, pp. 53–67. Canberra: Australian Academy of Science.
- Watts WA (1979) Late Quaternary vegetation of central Appalachia and the New Jersey coastal plain. *Ecological Monographs* 49: 427–469.
- Watts WA and Winter TC (1966) Plant macrofossils from Kirchner Marsh, Minnesota—A paleoecological study. *Geological Society of America Bulletin* 77: 1339–1360.
- Williams JW, Shuman BN, Webb T III, Bartlein PJ, and Leduc P (2004) Late Quaternary vegetation dynamics in North America: Scaling from taxa to biomes. *Ecological Monographs* 74: 309–334.
- Willis KJ, Rudner E, and Sümegi P (2000) The full-glacial forests of central and southeastern Europe. *Quaternary Research* 53: 203–213.
- Willis KJ and Van Andel TH (2004) Trees or no trees? The environments of central and eastern Asia during the Last Glaciation. *Quaternary Science Reviews* 23: 2369–2387.

PLANT MACROFOSSIL RECORDS

Contents

Arctic Eurasia

Arctic North America

Greenland

Holocene North America

Late Glacial Multidisciplinary Studies

Arctic Eurasia

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Arctic Eurasia

Compared with Arctic North America, there are relatively few macrofossil records available in Arctic Eurasia. In Western Eurasia, repeated glaciations (Figure 1) abraded older Pleistocene organic material. The most important archives in North Europe and West Siberia are therefore post glacial peat and lake sediments. Plant macrofossil studies focus there on the chronology of recolonization of deglaciated areas during the Late Glacial and the Holocene (Alm and Birks, 1991; Birks, 1991, 2000; MacDonald et al., 2006).

Siberia, east of the Taymyr Peninsula, is situated outside the range of Pleistocene inland glaciations (Figure 1). As a result of the extreme shallowness of the wide northeast Siberian shelf seas, this region is affected by huge coastline shifts during Quaternary sea-level fluctuations. During Pleistocene sea-level lowstands, East Siberia together with Alaska and western Yukon formed a large land bridge and a common bioregion – Beringia (Figure 1). Instead of large ice sheets, there formed thick ice-bonded permafrost deposits, which perfectly preserve organic material over tens of thousands of years and are therefore an excellent archive of fossil organism remains (Figure 2). Not only plant remains but also carcasses of large herbivores are exceptionally well preserved in perennially frozen conditions. They sometimes include even the gastrointestinal content, which gives unique insights into the dietary spectrum of Pleistocene animals (Tomskaya, 2000; van Geel et al., 2008). Also, burrows of ancient rodents may reveal the former food base of herbivores (Gubin et al., 2001).

Several Quaternary key sequences in the Northeast Siberian coastal lowlands became known as recently as the second half of the last century (Giterman et al., 1982; Kaplina, 1980; Romanovskii, 1961); initially, however, macrofossils were sparsely studied. In recent years, systematic multidisciplinary studies have resulted in a considerable increase in available macrofossil records. The results have confirmed that plant macrofossils are key proxies for the reconstruction of

Quaternary terrestrial environments, compensating for the limitations of pollen analyses in the far North. In this article, results and climatic implications of Quaternary plant macrofossil studies in arctic Eurasia are discussed, with focus on Northeast Siberia.

The Arctic Today

The Arctic is regarded as the region north of the polar circle. In consequence of the Earth's axial precession, the high latitudes are characterized by extreme seasonal fluctuations of incoming solar radiation, culminating in the semiannual alternation of polar day and polar night. Due to the low altitude of the sun in the sky, the radiant flux density is very low. Nevertheless, the radiation balance is positive in summer and, owing to the polar day, the radiation supply during the growing season is among the highest in Eurasia. In a geobotanical sense, the Arctic is the region of tundra beyond northern treeline, which correlates approximately with the 10 °C mean July isotherm in West Eurasia or with the 12 °C July isotherm in the more continental East.

In today's Arctic tundra, a cool and short growing season is assumed to be the cause of treelessness and the main factor limiting arctic life. Both factors depend highly on the degree of maritime influence on climate. Proximity to the sea results in smaller seasonal temperature amplitudes, thus relatively mild winters, cool summers, and increased humidity. Due to the proximity of the Arctic Ocean and the North Eurasian shelf seas, the whole Eurasian Arctic tundra region is more or less under maritime climate influence today (Figure 3). Maritime influence involves increased cloudiness, which hampers the warming of the Earth's surface by direct solar radiation. Heightened precipitation brings thick snow cover in winter and saturated soils in summer. Because of the late snow melt, the growing season is very short and a high percentage of received solar energy is lost to snow albedo or is transformed in latent, not sensible, heat, because of waterlogged soils. In regions with

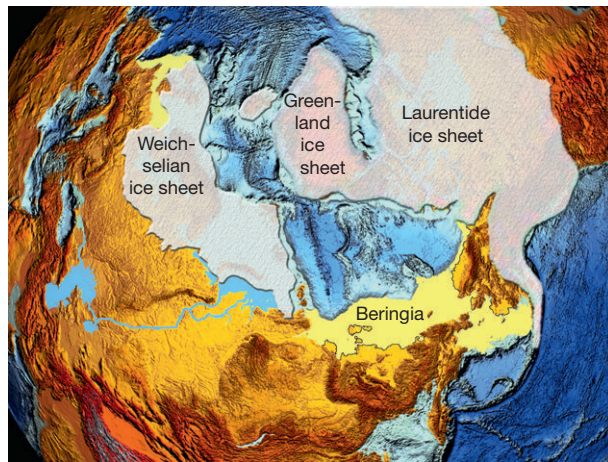


Figure 1 The northern hemisphere during the last cold stage. The figure illustrates the positions and maximum extents of northern inland ice sheets and the dimension of shelf exposure in North East Siberia (yellow signature) resulting in the formation of the subcontinent Beringia. Note the presence of ice-dammed lakes in West Siberia and the rerouting of rivers via the Turgay Pass into the Aral and Caspian Seas. Reproduced from Mangerud J, Jakobsson M, and Alexanderson H (2004) Ice-dammed lakes and rerouting of the drainage of northern Eurasia during the Last Glaciation. *Quaternary Science Reviews* 23 (11–13): 1313–1332. Bathymetric map adapted from the NGDC, NOAA Satellite and Information Service, available at <http://www.ngdc.noaa.gov/mgg/global/>.

a more continental climate, both insolation during summer and energy emission during winter can pass through the cloudless atmosphere much more easily. This results in higher seasonal temperature amplitude. Despite lower mean annual and winter temperatures than in the more maritime Europe and West Siberia, the polar treeline is shifted farther northward in East Siberia, owing to the relatively high summer mean temperature and lengthened growing season (Figure 3).

The composition of tree species that constitute the northern treeline changes longitudinally. The westernmost treeline constituent is the Scandinavian *Betula pubescens* ssp. *tortuosa*, replaced by *Picea obovata* eastward to the Ural Mountains, *Larix sibirica* in West Siberia, and *Larix dahurica* in Yakutia – the most continental part of the Arctic. In the Far East, the stone birch (*Betula ermanii*) is the northernmost tree species. The boreal coniferous forests in North Europe and West Siberia mainly composed of spruce and, in the south, fir are termed dark taiga. In Central and East Siberia, boreal forests are named light taiga owing to the appearance of the foliage of the constituting deciduous larch species. Other tree species that advance close to the northern treeline are *Pinus sylvestris* in Europe and *Chosenia arbutifolia* and *Populus suaveolens* in the Far East. Northern boreal coniferous forests fade to tundra gradually via the forest-tundra ecotone, where trees slowly diminish northwards. Trees advance farthest northward along river valleys, which are favored climatically due to energy transport by river water from the south and faster warming of the sandy floodplain soils (Walter, 1974). Furthermore, soils are better drained owing to the unfrozen ground (formation of taliks) in river valleys. Finally, the impact of winter storms on vegetation is alleviated in river valleys.



Figure 2 Quaternary permafrost exposures at the NE-Siberian coastal district Oyogos Yar. (a) Ice-bonded permafrost (Yedoma) exposed at the sea coast in up to 40 m high cliffs. (b) Warm-stage deposits of a former shallow lake trapped in an ice wedge cast. Fossil organism remains are well preserved in permafrost: *in situ* mammoth tusk (c), plant material accumulated in lacustrine deposits (d), including fruits of boreal aquatics (e) – *Potamogeton* sp. *in situ*. Photographs by F. Kienast.

The vegetation north of the treeline is relatively uniform. Many genera and even species in tundra have a circumpolar distribution. As refuge and center of dispersal, Beringia is floristically, however, more diverse than Arctic Europe and West Siberia – a result of the lack of inland glaciations and a more stable continental climate throughout the Quaternary. Due to the northward decrease of summer temperature, tundra is differentiated into two subzones by coverage and floral composition (Aleksandrova, 1980). The border between the arctic and subarctic tundra subzones is approximately determined by the 6 °C July isotherm. Subarctic tundra includes the forest-tundra ecotone and the southern tundra, which is characterized by the presence of boreal floral elements, such as *Duschetia fruticosa*, Ericaceae, *Empetrum nigrum*, *Comarum palustre*, *Rubus*, etc. Dwarf shrubs are the typical life form in subarctic tundra. Among them, dwarf birches (*Betula nana*, *Betula exilis*, *Betula glandulosa*) are most characteristic of this subzone. They are, in the arctic tundra, replaced by arctic willows such as *Salix polaris*, *Salix reticulata*, cold adapted cushion plants, arctic

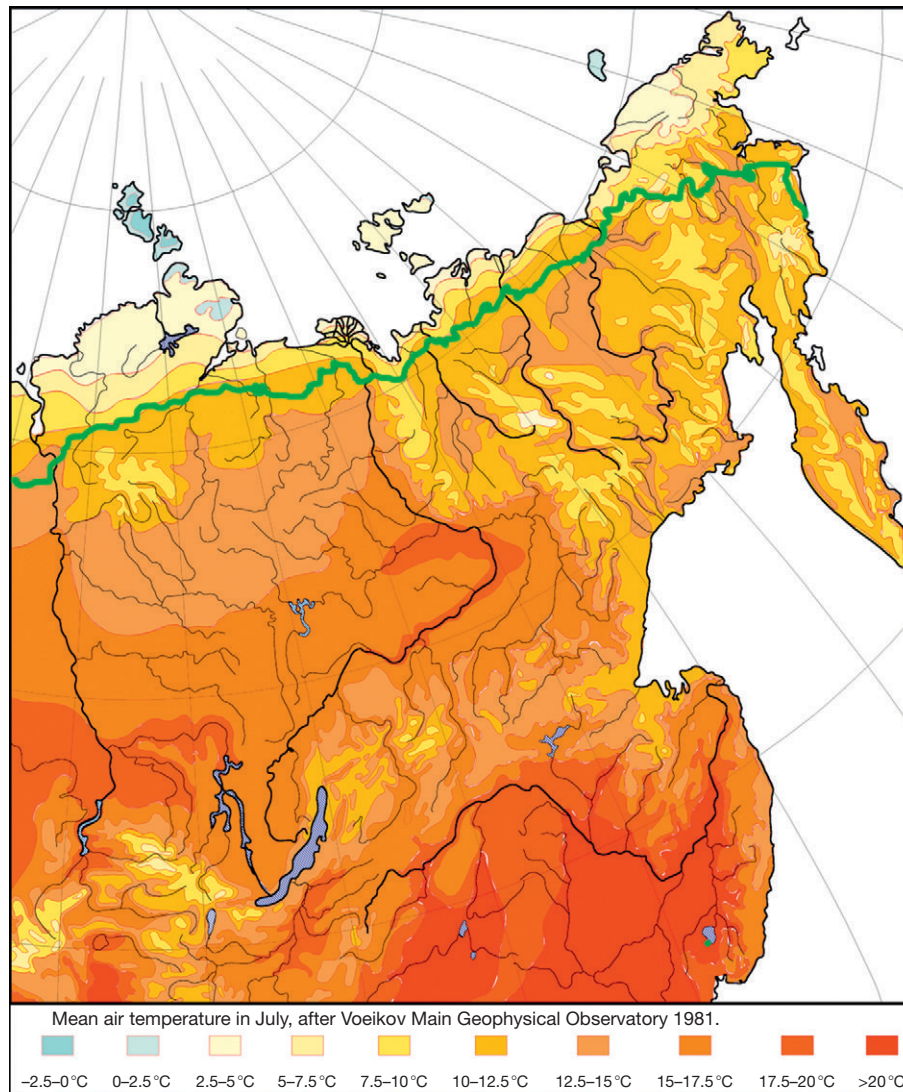


Figure 3 Mean July air temperature (after Voeikov, 1981) and northern treeline in NE-Asia (green, after CAVM Team, 2003). Both July isotherms and treeline are correlated with the course of the coast line, which determines the maritime influence on climate. For example, on the maritime influenced Chukotka Peninsula in the Far East, the mean July temperature is about 8 °C, whereas at the same latitude (65° N) in continental Yakutia it is about 16 °C. The boreal treeline approximately follows the 12 °C mean July isotherm. The July temperature is highest in the most continental part of northern Asia – in Central Yakutia, where winter temperatures are among the lowest in the northern hemisphere.

sedges (*Carex stans*), and mosses. Boreal floral elements are largely absent in the subzone of arctic tundra.

As in every vegetation zone, tundra is composed of a mosaic of various plant communities which are controlled by a variety of local environmental factors. Due to the frozen ground, tundra soils are often saturated. Consequently, wetland vegetation, mainly composed of sedges, cotton grass, grasses, mosses, and dwarf shrubs, mostly on leached and acidic soils or on thick peat, dominates the landscape (CAVM Team, 2003). Small-scale vegetation differentiation is the result of uneven moisture or drainage, as a result of irregular snow cover thickness and active layer depths.

Permafrost affects soils and thus vegetation by stabilization and damming groundwater. Under humid climate, permafrost highly amplifies soil paludification. Today even in Arctic Northeast Siberia, where precipitation is less than 200 mm,

humid conditions prevail since temperature and evaporation are low. The landscape is characterized by widespread wetlands because of the damming effect and the melting of ice-rich permafrost deposits (Figure 4). Under arid climate conditions farther inland in contrast, permafrost is accompanied by high seasonal fluctuations of soil moisture due to high evaporation during summer. Here, permafrost supports salt accumulation in the soil by damming groundwater and supplying additional moisture and solutes (Yelovskaya et al., 1966; Figure 4).

Quaternary Key Sites in Arctic Eurasia

As already pointed out, permafrost deposits are among the most important paleoenvironmental archives in the Arctic. In

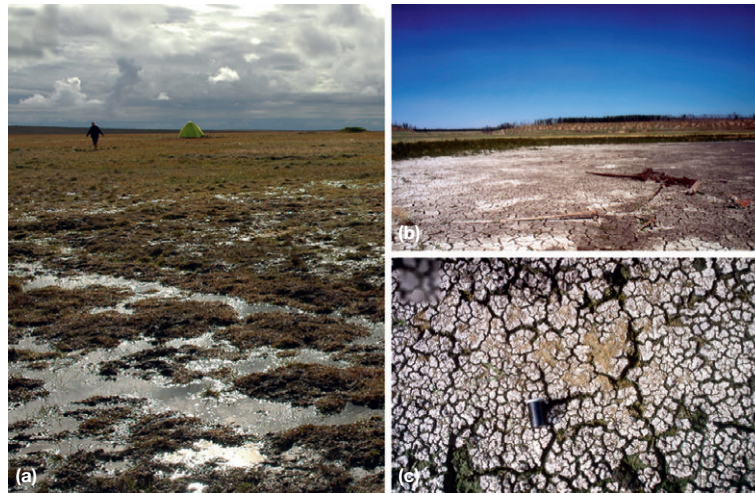


Figure 4 Permafrost impact on soils. Near-surface permafrost can amplify extreme soil conditions by damming and supplying melt water and solutes – wetness in humid situations (a) and salinification under arid conditions (b) and (c). Paludified arctic tundra at Oyogos Yar (a). Shore of a shrinking thermokarst lake in the extremely continental Central Yakutia with a seasonally fluctuating water level (b) and salt crust formation in the range that falls dry in summer (c). Photographs by F. Kienast (a) and C. Siegert (b) and (c).

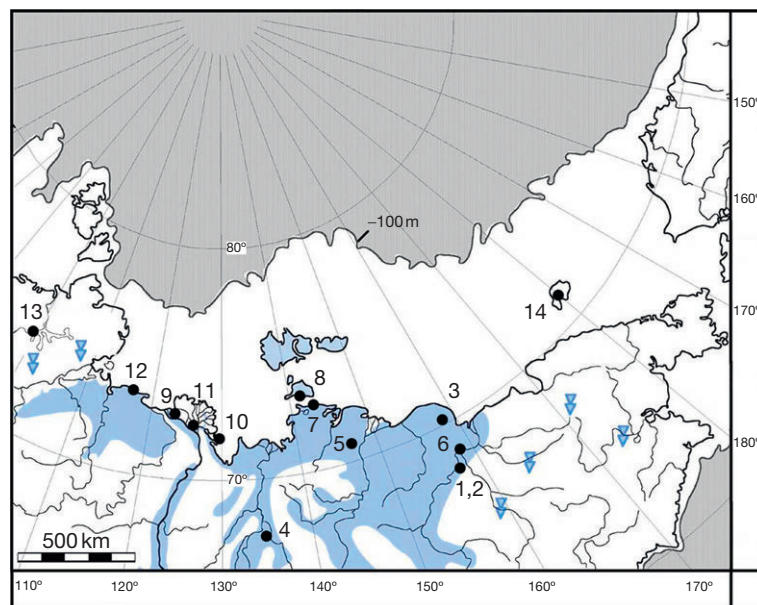


Figure 5 Plant macrofossil key sites (black dots) and the distribution of ice-bonded permafrost (Yedoma, blue signature) in Arctic Eurasia. The occurrence of Yedoma is correlated with extremely low winter temperature and maximum continentality at the time of its formation during the last glacial, when the coast line was farther north (note the 100 m isobaths). Blue wedges indicate the presence of Yedoma restricted to river valleys. Site numbers correlate with Table 1. Yedoma distribution adapted from Romanovsky NN (1993) *Osnovy Kriogeneza Litostery*. Moscow: Izdatelstvo Moscow State University.

East Siberia, large areas are covered by ice-bonded permafrost (Yedoma, Ice Complex, Figures 2(a) and 5), which formed under very low winter temperatures during the last cold stage, when climate was extremely continental due to sea-level lowering and coast line regressions. Such ice-rich permafrost deposits probably formed likewise during older cold stages. Owing to their high ice content, ice-rich permafrost deposits are however thermally very susceptible and eroded repeatedly during warm stages. As a result, cold-stage deposits older than Late Quaternary are scarcely available. From the time prior to the Early Weichselian cold stage (MIS 4) therefore, only interglacial

macrofossil records are known. Moreover, owing to uneven permafrost degradation, the stratigraphical correlation of permafrost deposits in arctic regions is often complicated and records recognized without doubt as older than MIS 5 are rare. For the ecological interpretation of paleobotanical data, the stratigraphical correlation of fossiliferous deposits is crucial. Table 1 gives an overview of Quaternary plant macrofossil key sites in Siberia. Please note that many Russian records processed during the last century seem to be not fully published or publications are unavailable for the international audience (e.g., records discussed in Kaplina (1981) and Giterman et al. (1982)).

Table 1 List of plant macrofossil key sites in Arctic Eurasia with details on age, key taxa, paleovegetation, and references. The number of each site corresponds to the site numbers in [Figure 5](#)

Site no.	Sequence	Age	Location	Key taxa	Inferred paleovegetation	Modern vegetation	References
1	Begunov suite	early Pliocene	Krestovka River, lower course of Kolyma River	Some plesiomorphic (archaic) conifers: <i>Larix</i> sp., <i>Picea</i> sp., <i>Pinus</i> sp., <i>Potamogeton jacuticus</i> , <i>Potamogeton omoloicus</i> , <i>Epipremnum crassum</i>	Larch–spruce–birch taiga with pine	Forest tundra	Giterman et al. (1982) , Kaplina (1981) , and Nikitin (2006)
2	Kutuyakh beds	late Pliocene	Krestovka River, lower course of Kolyma River	<i>Epipremnum crassum</i> , <i>Nuphar pumila</i> , <i>Brasenia</i> sp., <i>Picea</i> sp., <i>Larix</i> sp.	Open taiga	Forest tundra	Giterman et al. (1982) , Kaplina (1981) , and Nikitin (2006)
3	Olyor suite	early Pleistocene	Bol'shaya Chukoch'ya, Alazeya Rivers	<i>Larix</i> sp., <i>Potamogeton</i> spp., <i>Menyanthes trifoliata</i> , <i>Carex</i> spp., <i>Ranunculus sceleratus</i>	Forest tundra	Subarctic tundra	Kats et al. (1982)
4	Ulakhan Sular suite	middle Pleistocene	Tuostakh sand sequence, Adycha River, influent of the middle Yana River	<i>Larix dahurica</i> , <i>Alnus</i> sp., <i>Salix</i> sp., <i>Potamogeton</i> spp., <i>Damasonium stellatum</i> , <i>Sagittaria natans</i> , <i>Rumex acetosella</i> , <i>Atriplex</i> sp., <i>Chenopodium</i> sp., <i>Corispermum</i> spp., <i>Linum perenne</i> , <i>Allium</i> sp.	Aquatic vegetation and alder–willow gallery forests within a larch taiga–dry grassland mosaic	Open larch taiga	Kaplina et al. (1983)
5	Achchagyy suite	MIS 5	Allaikha river near the confluence into the Indigirka River	<i>Larix dahurica</i> , <i>Duschekia fruticosa</i> , <i>Betula alba</i>	Northern taiga, larch–birch forest tundra	Subarctic tundra	Kaplina et al. (1980)
6	Duvanny Yar	MIS 5	Lower course of Kolyma River	<i>Larix dahurica</i>	Northern larch taiga	Northern larch taiga	Giterman et al. (1982)
7	Krest Yuryakh	MIS 5	Bol'shoy Lyakhovsky Island, Oyogos Yar, Dmitri Laptev Strait coast	<i>Larix dahurica</i> , <i>Alnus incana</i> s.l., <i>Duschekia fruticosa</i> , <i>Betula fruticosa</i> , <i>Betula divaricata</i> , <i>Potamogeton perfoliatus</i>	Forest tundra, erect shrub tundra	Arctic tundra, graminoid–cryptogram tundra (northern coast) sedge–grass–moss wetland tundra (southern coast)	Kienast et al. (2008, 2010)
8	Bol'shoy Lyakhovsky Island	MIS 4 to MIS 3	New Siberian Archipelago	<i>Artemisia frigida</i> , <i>Androsace septentrionalis</i> , <i>Potentilla stipularis</i> , <i>Puccinellia</i> sp., <i>Draba</i> sp.	Tundra steppe	Arctic tundra (graminoid–cryptogram tundra)	Andreev et al. (2009)
9	Nagym	MIS 4 to MIS 3	Western Lena River Delta, Olenyok channel	<i>Chamaedaphne calyculata</i> , <i>Stellaria longifolia</i> , <i>Kobresia myosuroides</i> , <i>Ranunculus affinis</i> , <i>Dryas octopetala</i> s.l., <i>Potentilla stipularis</i>	Tundra steppe in proximity of subarctic shrubbery (MIS 4)	Subarctic tundra, Wetland tundra	Schirrmeister et al. (2010)

(Continued)

Table 1 (Continued)

Site no.	Sequence	Age	Location	Key taxa	Inferred paleovegetation	Modern vegetation	References
10	Mamontovy Khayata	MIS 4 to early Holocene	Bykovsky Peninsula	<i>Koeleria cristata</i> , <i>Linum perenne</i> , <i>Alyssum obovatum</i> , <i>Potentilla stipularis</i> , <i>Hordeum brevisubulatum</i> , <i>Puccinellia</i> sp.	Tundra steppe during the last glacial, subarctic tundra during Holocene	Subarctic tundra	Kienast et al. (2005)
11	Kurungnakh Island	MIS 3 to Holocene	Lena River Delta	<i>Phlox sibirica</i> , <i>Kobresia myosuroides</i> , <i>Potentilla stipularis</i> , <i>Allium schoenoprasum</i> , <i>Ranunculus affinis</i> , <i>Hierochloa</i> cf. <i>glabra</i>	Tundra steppe during the last glacial, subarctic tundra during Holocene	Subarctic tundra	Wetterich et al. (2008)
12	Mamontovy Klyk	MIS 3 to Holocene	Laptev Sea mainland coast between Anabar and Olenyok river mouths	<i>Juncus biglumis</i> , <i>Draba</i> sp., <i>Minuartia rubella</i> , <i>Papaver</i> Sect. <i>Scapiflora</i> , <i>Dryas</i> sp., <i>Artemisia</i> sp., <i>Poa</i> sp., early Holocene: <i>Duschekia fruticosa</i> , <i>Betula nana</i> s.l., <i>Empetrum nigrum</i>	Barren with scattered steppe elements, subarctic shrub tundra during the early Holocene	Arctic tundra (sedge–dwarf shrub–moss tundra)	Schirrmeister et al. (2008)
13	Cape Sabler	MIS 3 to Holocene	Western coast of Taymyr Lake, Taymyr Peninsula	<i>Draba</i> sp., <i>Minuartia rubella</i> , <i>Papaver</i> Sect. <i>Scapiflora</i> , <i>Dryas octopetala</i> s.l.	Barren with scattered steppe elements, wetland tundra during Holocene	Subarctic tundra, sedge, moss, dwarf shrub wetland tundra	Kienast et al. (2001)
14	Mamontovaya River	early Holocene	Central part of Wrangel Island	<i>Kobresia myosuroides</i> , <i>Kobresia sibirica</i> , <i>Rumex beringensis</i> , <i>Rumex</i> sect. <i>Acetosa</i> , <i>Silene repens</i> , <i>Ranunculus grayi</i> , <i>Potentilla</i> sp., <i>Androsace septentrionalis</i>	Tundra steppe	Arctic tundra, moist nonacidic tundra	Vazhenina (1996)

Quaternary Vegetation History

Macrofossil studies are generally part of multidisciplinary approaches. Through the combination of other paleoecological proxies such as sediment characteristics, pollen, and remains of herbivorous insects and mammals, we know much about the character of landscape and vegetation during the Quaternary by now, even though indicative upland plant taxa are often rare in the record. Especially, records older than middle Pleistocene are biased in composition through the dominance of aquatic plants, as they were deposited during warm stages exclusively in aquatic environments.

Pliocene

The Pliocene key sites available in the Eurasian Arctic prove that boreal coniferous forests existed in northern Siberia until the inception of northern hemisphere inland glaciations at the

beginning of the Quaternary about 2.6 Ma ago. These taiga forests were more diverse floristically, composed of now extinct conifers, several larch species, spruce, and pine (Giterman et al., 1982; Kaplina, 1981; Nikitin, 2006). Many extinct or extralimital thermophilous plants including *Epipremnum cras-sum*, *Nuphar pumila*, *Brasenia* sp., and *Potamogeton omoloicus* indicate conditions distinctly warmer than the present. However ice wedge casts in the upper Pliocene Kutuyakh beds indicate that permafrost already existed (Kaplina, 1981). Also the dominance of larch species might indicate harsh winter conditions. Due to their situation inland of the greatest land-mass on Earth and completely cut off from moist and mild Pacific air masses by the upfolding of the Himalaya Mountains, the northeast Siberian taiga forests had already evolved from the late Miocene onwards connected with progressive global cooling and increasing continentality. The transformation of northern vegetation in response to colder and more continental conditions during the Quaternary occurred in

Northeast Siberia on a greater scale and earlier than elsewhere in the northern hemisphere (Frenzel, 1968). The subsequent, increasingly abrupt Quaternary climate fluctuations were alleviated in Northeast Siberia because of its already established climatic continentality that more or less endured during the Quaternary, during both cold and warm phases.

Early and Middle Pleistocene

Many Pliocene conifers did not survive the onset of the Quaternary in NE Siberia, resulting in the domination of *Larix dahurica* in the Yakutian coastal lowlands (Giterman et al., 1982; Kaplina, 1981; Nikitin, 2006), probably connected with the very low winter temperature. Early and middle Pleistocene paleobotanical records in the arctic lowlands of Siberia reflect the existence of an open larch taiga – a dry grassland mosaic with lakes that were populated with thermophilous aquatics (e.g., *Damasonium stellatum*, *Sagittaria natans*) and salt-tolerant riparian plants, which are typical of shores at shrinking lakes under arid conditions (*Ranunculus sceleratus*, *Atriplex* sp., *Chenopodium* sp.). In contrast to modern tundra or northern taiga, dry habitats were widespread during early and middle Pleistocene interglacials as is indicated by the presence of steppe plants (*Linum perenne*, *Allium* sp.; Kaplina et al., 1983). The early Pleistocene cold stages were probably not as harsh as the later Saalian and Weichselian glaciations. This is indicated by the diversity of thermophilous, extralimital plants and the absence of extremely cold resistant mammals during the first cold stages (Sher, 1997). Tundra steppe was the dominant vegetation during cold stages, closely connected with the Mammoth faunal complex. As apparent from pollen records, the vegetation was much more uniform across broad regions than it is today, due to smaller latitudinal and longitudinal climate gradients in Eurasia. This seems to have been true not only during glacials but even during interglacials. The kind of today's latitudinal zonation of vegetation apparently did not exist (Sher, 1997).

The Last Interglacial – MIS 5

The last interglacial vegetation in Arctic Eurasia is represented here by two recent macrofossil records from the Krest Yuryakh Suite at about 73° N. The sequence consists of lake sediments filling large ice wedge casts and outcrops along the coasts of the Dmitry Laptev Strait (Figure 2) at both sides, the coastal mainland district Oyogos Yar (Kienast et al., 2011) and the Bol'shoi Lyakhovsky Island (Kienast et al., 2008). The coastal exposures are the first sites in NE Siberia, where last interglacial thermokarst deposits were described (Romanovskii, 1961). Both Eemian macrofossil records are only about 80 km apart, but they already reflect a floristic latitudinal gradient. Whereas trees were absent at today's coast of Bol'shoi Lyakhovsky Island, macrofossils of larch and tree alder have proven the existence of trees at the mainland coast Oyogos Yar at 72.68° N during the last interglacial (Figure 6). Accordingly, the northern treeline was at that time situated between both sites and had shifted by almost 300 km north of its current position. Thus, there existed trees at a place that is today (during the subsequent interglacial) too cold even for the survival of dwarf birches. Larch remains at other last interglacial exposures in

the tundra lowlands of Yakutia, for example, in Achchaggy deposits at the Allaikha River west of the Indigirka River (Kaplina et al., 1980) and the Bolshoy Khomus Yuryakh site east of the Indigirka River, (Lozhkin and Anderson, 1995) suggest that the treeline shift close to the arctic sea coast was a widespread phenomenon. Quantitative reconstructions resulted in a mean temperature of the warmest month of about 12 °C at the Bol'shoi Lyakhovsky Island and exceeding 13 °C at Oyogos Yar, thus about 9 °C warmer than today. Apart from the evidence of trees, warming is indicated by the diverse occurrence of Ericaceae such as *Arctostaphylos uva-ursi*, *Rhododendron* sp., *Chamaedaphne calyculata*, and *Andromeda polifolia* as well as aquatic plants, all of them far beyond their modern limit, for example, *Potamogeton perfoliatus*, *Potamogeton vaginatus*, and *Myriophyllum spicatum* (Figure 6). The last interglacial in western Beringia was not only warmer but also drier than today as is indicated by the remains of steppe plants such as *Allium*, *Artemisia*, *Androsace septentrionalis*, and *Potentilla stipularis*; xerophilous arctic plants such as *Kobresia*, *Ranunculus affinis*, and *Potentilla nivea*; and salt-tolerant pioneer plants of lake littorals, for example, *Chenopodium*, *Tripleurospermum maritimum*, and *Puccinellia* (Figure 7). Corresponding



Figure 6 Macrofossils of warmth-indicator plants. 1–14, forest–tundra plants; 1 *Larix dahurica*, needle tip; 2 *Larix dahurica*, short shoot; 3 *Alnus incana*, nutlet; 4 *Betula divaricata*, scale; 5 *Duschekia fruticosa*, scale; 6 *Betula nana*, scale; 7 *Ledum palustre*, leaf fragments; 8 *Empetrum nigrum*, seed; 9 *Vaccinium vitis-idaea*, seed; 10 *Andromeda polifolia*, seed; 11 *Moehringia laterifolia*, seed; 12 *Chamaedaphne calyculata*, seed; 13 *Stellaria longifolia*, seed; 14 *Chamaenerion angustifolium*, seed; 15–22 boreal aquatics; 15 *Potamogeton perfoliatus*, fruitstone; 16 *P. vaginatus*, fruitstone; 17 *P. filiformis*, fruitstone; 18 *Sparganium minimum*, fruit; 19 *Batrachium* sp., nutlet; 20 *Callitriche hermaphrodita*, seed; 21 *Myriophyllum spicatum*, fruit; 22 *Menyanthes trifoliata*, seed.

to cold-stage records but in contrast to the current warm stage, alkali grass (*Puccinellia*) was among the most abundant plant taxa in the Eemian macrofossil spectra. All the indicators of dryness and fluctuating groundwater and lake levels indicate low net precipitation and, in conjunction with high summer temperature, a climate distinctly more continental than today.

The Last Cold Stage

The last cold stage (Weichselian) was the time of maximum continentality, when NE-Siberian coastlines retreated as far as 800 km northward (Figure 5) due to a global sea-level fall of up to 120 m. As the Laptev and East Siberian Seas are not deeper than 30 m over most of their extent, the shelves remained exposed and continentality persisted during the Mid-Weichselian interstadial (MIS 3) also, when there was a slight rise in the global sea level. Correlated with extreme low winter temperature, ice-bonded permafrost formed in the arctic lowlands and in large river valleys (Figure 5). Yedoma

deposits are frequently exposed and accessible along coastal and river bluffs in the arctic lowlands (Figure 2(a)). From there, a number of sites have been studied for plant macrofossils in recent years: Cape Sabler (Site 13, Figure 5; Kienast et al., 2001), Mamontovy Klyk (Site 12, Figure 5; Schirrmeister et al., 2009), Nagym (Site 9, Figure 5; Schirrmeister et al., 2010), Kurungnakh Sise (Site 11, Figure 5; Wetterich et al., 2008), Mamontovy Khayata (Site 10, Figure 5; Kienast et al., 2005), and Bol'shoi Lyakhovsky Island (Site 8; Andreev et al., 2009). The macrofossil records are more or less similar in species composition and partly resemble also assemblages from last interglacial Krest Yuryakh deposits at the Dmitri Laptev Strait. In contrast to the latter, thermophilous forest or aquatic plant species are however mostly lacking. Some boreal aquatics such as *Callitriche hermaphroditica*, *Potamogeton vaginatus*, and *Batrachium* sp. were still present during the Mid-Weichselian interstadial. The most characteristic feature of arctic vegetation during the last cold stage is, however, the coexistence of steppe



Figure 7 Macrofossils of aridity-indicator plants. 1–10 steppe plants; 1 *Linum perenne*, seed; 2 *Phlox sibirica*, capsule valve; 3 *Artemisia* sp., achene; 4 *Allium* sp., seed; 5 *Koeleria cristata*, floret; 6 *Potentilla stipularis*, nutlet; 7 *Androsace septentrionalis*, seed; 8 *Alyssum obovatum*, seed; 9 *Rumex acetosella*, fruit; 10 *Silene repens*, seed; 11–15 xerophilous arctic plants; 11 *Kobresia myosuroides*, nutlet; 12 *Thalictrum alpinum*, fruit; 13 *Ranunculus affinis*, nutlet; 14 *Potentilla nivea*, nutlet; 15 *Dryas octopetala* s.l., leaf fragment; 16–22 indicators of fluctuating lake levels and salinification; 16 *Carex duriuscula*, nutlet; 17 *Tripleurospermum maritimum*, achene; 18 *Stellaria crassifolia*, seed; 19 *Rorippa palustris*, seed; 20 *Puccinellia* cf. *tenuiflora*, caryopsis; 21 *Rumex maritimus*, fruit; 22 *Chenopodium* sp., seed.



Figure 8 (a) Cold-arid 'Goltsy' region in the alpine belt of the Chersky Mountains (Northeast Siberia). The precipitation is too low to support a perennial snow cover or even a glacier. Note the vegetation differentiation dependent on exposure and soil moisture. (b) Steppe and alpine plants occur there side by side (here: *Festuca kolyomensis* and *Dryas punctata*). Photographs by F. Kienast.



Figure 9 Littoral ruderals at the shore of a shrinking thermokarst lake used as a cattle watering place, in Central Yakutia, with *Rumex maritimus*, *Stellaria crassifolia*, *Puccinellia tenuiflora*, *Senecio congestus*, *Peucedanum salinum*, *Chenopodium rubrum*, among others. A salt crust has formed at the surface of the hardly disturbed soil. Photograph by F. Kienast.

plants such as *Linum perenne*, *Artemisia frigida*, *Allium schoenoprasum*, *Alyssum obovatum*, *Potentilla stipularis*, *Silene repens*, *Androsace septentrionalis*, *Phlox sibirica*, and *Koeleria cristata* with xerophilous arctic plants such as *Kobresia myosuroides*, *Dryas*, *Potentilla nivea*, *Thalictrum alpinum*, and *Ranunculus affinis* (Figure 7). This plant species combination is diagnostic for the Pleistocene tundra-steppe, mammoth steppe or steppe tundra, which is intrinsically tied to the grazing mammoth faunal complex. Tundra-steppe has no biome-sized modern analog, but there are many places with tundra-steppe relict vegetation, especially in mountainous regions of Yakutia, where continental climate encounters cool temperature (Figure 8; Yurtsev, 1982, 2001). Tundra-steppe vegetation was mosaic-like and comprised potentially productive meadows with *Puccinellia*, *Hordeum brevisubulatum*, *Elytrigia*, *Festuca*, and *Poa*, as is indicated by macrofossils in yedoma deposits (Figures 7 and 10). These grasses are typical of saline floodplains and indicate seasonally fluctuating groundwater and salinification due to high evaporation. Remains of plants occurring at the littoral of shrinking lakes especially under saline influence, for example, *Rumex maritimus*, *Chenopodium*, *Tephrosia palustris*, and *Stellaria crassifolia* also indicate fluctuating lake water level due to high evaporation under arid conditions (Figure 9). Other steady components of the cold-stage plant macrofossil records are arctic pioneer plants such as *Papaver* Sect. *Scapiflora*, *Draba*, *Cerastium*, *Minuartia rubella*, and *Selaginella*, indicating locally open, disturbed ground. Northern wetland vegetation represented by macrofossils of *Carex*



Figure 10 *Puccinellia tenuiflora* (loose panicles) and *Hordeum brevisubulatum* (large spikes) growing in a floodplain meadow around a thermokarst lake in Central Yakutia. The productive meadow is intensively used as pasture. Photograph by F. Kienast.

Sect. *Phacocystis*, *Eriophorum scheuchzeri*, and *Eriophorum brachyantherum* played rather a minor role in the vegetation cover.

Late Glacial and Holocene

This period marks a fundamental alteration of northern ecosystems connected with the disappearance of tundra steppe and the extinction of many large herbivores such as mammoth, woolly rhinoceros, steppe bison, and Irish elk in high latitudes. The inception of the early Holocene began with a rapid temperature rise about 11.6 ka ago. The decay of ice caps was however delayed by about 5 ka and the global sea level reached modern values no earlier than 6 ka ago (CAPE-Last Interglacial Project Members, 2006). In consequence of delayed sea-level rise, the wide Laptev and Northeast Siberian Shelves remained long exposed (Bauch et al., 2001) and today's coastal sites were farther inland and remained influenced by persisting continentality during the early Holocene. Consistently, xerophilous plants continued to exist in the Arctic beyond the Weichselian/Holocene transition, for example, *Potentilla*, *Kobresia myosuroides*, and *Androsace septentrionalis* (Kienast et al., 2005; Wetterich et al., 2008). The increase of wetland plants such as *Carex* Sect. *Phacocystis* and *Eriophorum* spp. in the record, however, reflects paludification connected with the onset of thermokarst processes. The start of the accumulation of peat deposits on Chukotka and at the European ice sheet margins documents increasing moistening already during the Late Glacial between 14 and 12 ka (MacDonald et al., 2006). Wetlands expanded thereafter rapidly from 11.5 to 8 ka succeeding the retreating ice sheets in Fennoscandia and supported by increasing precipitation and thermokarst processes in Siberia (MacDonald et al., 2006). In the second half of the Holocene, wetland plants obtained absolute domination in arctic vegetation. Remains of xerophilous plants are no longer detectable in late Holocene macrofossil records. Northern latitude tundra steppe is intrinsically linked to the mammoth faunal complex. A macrofossil record with numerous remains of xerophilous plant taxa such as *Kobresia*, *Potentilla nivea*, and *Selaginella rupestris*, radiocarbon dated to about 10 ka BP, reveals that during the early Holocene, tundra-steppe vegetation dominated on Wrangel Island (Vazhenina, 1996),

the place where mammoth survived the longest (Stuart et al., 2002).

The early Holocene summer temperature shift was the consequence of high summer insolation combined with initially persisting continentality. As a result, boreal shrubs and trees returned into the far North. The expansion of woody taxa and the chronology of reforestation in the whole Eurasian North during the last glacial and the Holocene was recently reconstructed using a newly created macrofossil database (Binney et al., 2009). Accordingly, trees and shrubs expanded spreading from refugia that existed through the last cold stage in northern latitudes, as is suggested by scattered macrofossils, for example, larch remains in a mammoth stomach at 74°N on the Taymyr Peninsula dated about 21 ka BP (Mol et al., 2006). Adaptable ambiguous woody taxa such as birches and shrub alder, which persisted in northern refugia, rapidly occupied potential forest habitats and reached tree sizes (Edwards et al., 2005). Linked to continentality, deciduous trees such as larch and birch expanded rapidly, extending far beyond their present limit (Binney et al., 2009; Kremenetski et al., 1998; MacDonald et al., 2000). Reforestation peaked between 10 and 8 ka ago, when all treeline species including the evergreen spruce and pine reached their maximum Holocene distribution, probably connected with decreasing seasonality and increasing moisture (Binney et al., 2009). Further increasing maritime influence resulted in paludification. Cooler summers led to a new treeline retreat beginning after about 7 ka cal BP in eastern Siberia and culminating between 4 and 3 ka cal BP in Europe and West Siberia. From then on the modern low-diverse wetland tundra was established as the prevailing vegetation at high latitudes.

Reconstruction of Quaternary Vegetation

The basic character of Arctic vegetation during Pleistocene cold stages differed substantially from modern tundra. Within the Pleistocene tundra-steppe vegetation, there coexisted a wide range of plant communities, reflecting a variety of habitat conditions. Due to a greater spatial environmental gradient, the set of plant communities, constituent plant taxa, and habitat conditions was much more diverse than modern tundra and depended on the availability of warmth and moisture. During the Pleistocene these two factors fluctuated considerably, both seasonally and spatially. Modern tundra in contrast is rather shaped by consistently cool and wet conditions. The most important plant communities that occurred in the Arctic during the Quaternary are briefly described below.

Aquatic Vegetation

Most aquatic plants found in high-latitude Pleistocene cold-stage assemblages do not occur so far north today. The occurrence of boreal or even temperate aquatics in the Arctic indicates warm summers during cold stages, a result of increased continentality. Aquatic plant diversity peaked during interglacials as a result of high summer temperature and lengthened growing season. The occurrence of species tolerant of brackish water (e.g., sheathed pondweed (*Potamogeton vaginatus*)) can be traced

back to fluctuating salt contents in freshwater due to high evaporation under arid conditions inland. These species have very restricted modern ranges, but were widespread in the Pleistocene. According to Mai (1985), warmth-indicating southern aquatic plants had already expanded rapidly during the initial stages of interglacials or during treeless interstadials in Central Europe, even exceeding their spread in earlier warm stages. These phenomena can be ascribed to the rapid dispersal of aquatics by birds after the reestablishment of favorable conditions. Apparently, summer temperatures rose abruptly at the onset of warm stages. Freshwater aquatics are pioneers and can occupy suitable habitats very quickly, whereas most tree species disperse slowly and can only become established following sufficient soil formation.

Littoral Pioneers and Riparian Ruderals

Littoral pioneer plants, which occur on wet, bare, erosive soils along the shores of shallow lakes and ponds with seasonally fluctuating water levels, indicate unstable moisture conditions. Fluctuating moisture is characteristic of continental climate as occurred in the Arctic during the Pleistocene. This vegetation shows that lakes and pools, filled with water after snow melt, decreased in size or even dried up in the course of summer owing to high aridity (Figure 4(b) and 4(c)). Similar to aquatics, some littoral plants are facultative halophytes. Their tolerance of high-salt content in soils further strengthens the indication of high evaporation given by their pioneer character. Some representatives, such as saltmarsh grass (*Puccinellia* sp.), golden dock (*Rumex maritimus*), *Chenopodium* sp., and fleshy starwort (*Stellaria crassifolia*), found in Arctic East Siberian cold-stage deposits (Figure 7; Kienast et al., 2005) occur today primarily on the coast, and otherwise are only found in continental inland areas of Asia. Littoral ruderals cannot endure competition and disappear after the establishment of stable wet conditions fostering vegetational succession. In contrast, frequent disturbances by animals, for example, at watering places, promote this type of vegetation (Figure 9).

Meadows

Intense seasonal fluctuations of moisture and salt content in soil are also shown by the presence of saline meadows, which in Northeast Siberia were composed of saltmarsh grass (*Puccinellia* sp.) and wild barley (*Hordeum brevisubulatum*) as well as needle-leaf sedge (*Carex duriuscula*; Kienast et al., 2005). The constituent grasses are characteristic of floodplains around lakes outside the range of seasonal inundation under arid conditions (Figure 10). This vegetation type is productive due to its proximity to seasonally fluctuating groundwater. Also, in the Norwegian periglacial, meadows with saltmarsh grass (*Puccinellia*) existed during the Weichselian cold stage (Alm and Birks, 1991).

Steppe Vegetation

Fossil remains of grazing, steppe-characteristic mammals such as bison, horse, and saiga antelope together with extinct grazers such as mammoth and woolly rhinoceros in high-latitude cold-stage deposits in North America (Guthrie, 1990) and



Figure 11 Pingo on Kurungnakh Sise Island, Lena River Delta with xerophilous vegetation including *Potentilla stipularis*, *Ranunculus affinis*, *Dryas punctata*, *Delphinium middendorffii*, *Trisetum sibiricum*, *Artemisia* spp., *Draba* spp., and *Salix glauca* among others. Photograph by F. Kienast.

Eurasia (Sher et al., 2005) have suggested that grasslands must have been an important component of cold-stage environments in the Eurasian high latitudes during the Pleistocene. This presumption was supported by disjunctive arctic and subarctic occurrences of steppe vegetation in northern Siberia, which are considered to be relicts of a formerly closed and more northerly extended distribution area of those species (Yurtsev, 1982; Figure 8). Finally, the discovery of steppe plant macrofossils such as *Linum perenne*, *Artemisia frigida*, *Allium schoenoprasum*, *Alyssum obovatum*, *Potentilla stipularis*, *Silene repens*, *Androsace septentrionalis*, *Phlox sibirica*, and *Koeleria cristata* in arctic permafrost deposits dated to the last cold stage provided the ultimate proof that steppe plants typical of arid, continental climate were an important component of arctic cold-stage vegetation (Kienast et al., 2005; Zazula et al., 2006). The spectrum of steppe plant macrofossils in Arctic Siberia is illustrated in Figure 7.

Xerophilous Arctic Upland and Pioneer Vegetation

Macrofossils of plants that occur mainly in dry calcareous mountain tundra today likewise reflect dry conditions and climatic continentality during the Pleistocene. Such plants as *Kobresia myosuroides*, *Dryas octopetala*, *Potentilla nivea*, *Thalictrum alpinum*, and *Ranunculus affinis* (Figure 7) indicate little or no snow cover and occur today at exposed sites that are frequently disturbed by deflation in winter. Since restricted on dry, exposed places, such plants play a minor role in modern arctic lowland tundra and occur there only at isolated well-drained microhabitats such as pingos (Figure 11; Walker et al., 2001). Arctic pioneer plants, for example, *Draba* sp., *Minuartia rubella*, *Papaver* Sect. *Scapiflora*, and *Cerastium beeringianum* are also characteristic components of arctic vegetation during cold stages. They indicate locally open, disturbed ground and are able to withstand adverse environmental conditions. The northernmost vascular plants occurring today in Arctic deserts are characteristic species of arctic pioneer vegetation.

Shrub Tundra and Forest Tundra

As we have seen, woodlands with larch (*Larix*), birch (*Betula*), and alder (*Alnus*) extended during warm stages considerably northward far beyond their present limit. Forest tundra shifted up to the modern coast line in eastern Siberia. Some taxa actually forming shrubs today reached tree sizes during early stages of the Holocene and possibly also during older warm stages. The expansion of boreal trees and shrubs is correlated with a warm and lengthened growing season and the presence of well-drained soil. Shrubs, especially willow shrubs, are important and diverse constituents of Arctic vegetation during both cold and warm stages. Willow species composition of cold-stage floras, and their ecological significance, differs from those of warm stages. We are unable, however, to give environmental implications since most Arctic willows are impossible to identify by macrofossils except for leaves. Contrary to this, shrubs of black crowberry (*Empetrum nigrum*), dwarf birches, and Ericaceae such as berries (*Vaccinium* sp.), wild rosemary (*Ledum palustre*), and bearberry (*Arctostaphylos uva-ursi*) in Siberia give clear implications of a locally thick, protective snow cover and acidic soils, both characteristic of interglacials.

Snow-Bed Vegetation

Plant communities that are typical of snow beds occur at sites with thick snow cover and late snow melt on soils permanently soaked and cooled by melting water. Due to the long snow cover, the growing season is considerably shortened at such sites. Among vascular plants, *Ranunculus nivalis*, *Ranunculus pygmaeus*, *Saxifraga nivalis*, and *Saxifraga foliolosa* were found as macrofossils primarily in late Holocene records correlated with increased maritime influence, for example, low summer temperature and thick snow cover.

Wetland Vegetation

Plant communities that require constant moisture played a minor role during cold stages but became abundant during warm stages, not only because of increasing humidity but also due to thermal erosion of ice-rich permafrost, which resulted in extensive wetland formation. But as recently as during the late Holocene, wetland vegetation obtained absolute domination in high latitudes. Sedges (*Carex bigelowii*, *Carex stans*, *Carex arctisibirica*) and cotton grass (*Eriophorum scheuchzeri*) are characteristic plants of permanently wet mineral soils with circumneutral substrates. Dwarf birches, yellow marsh saxifrage (*Saxifraga hirculus*) and hares-tail cotton grass (*Eriophorum vaginatum*) occur at rather acidic mires after peat formation.

Discussion

The Quaternary is characterized by enormous changes of plant distribution rather than by evolutionary processes (Lang, 1994). According to the rule of relative habitat constancy (Walter, 1954), both individual plant species and plant communities compensate for climate shifts by relocation so that

their habitat conditions remain relatively constant. According to pollen and macrofossil records, all warm stages since the middle Pleistocene have been alike in tree composition (Frenzel, 1968). Coniferous forests decreased in diversity but remained conservative in their basic species composition during the Quaternary. Only their distribution has changed in response to climate oscillations, which resulted, during cold stages, in the expansion of steppe vegetation. The response of vegetation to global temperature fluctuations depends highly on changes in humidity. This is exemplified by the arctic vegetation history in the Quaternary.

Tundra is an interglacial vegetation type. As tundra became the prevailing vegetation in northern latitudes during the Holocene, the Pleistocene large herbivores became extinct. During the Pleistocene, wetland tundra was more constricted, also in warm stages. During thermal optima, forests expanded far north toward the coastline. Tundra existed primarily within the context of forest tundra. This is suggested by Eemian macrofossil records from the East Siberian Arctic, which have indicated that forest tundra with an admixture of steppe and meadow communities, combined with warmth-requiring aquatics, prevailed on the Laptev Shelf during the last warm stage (Kienast et al., 2008, 2010). The last interglacial paleovegetation indicates a greater seasonal temperature gradient, with summers distinctly warmer than today, and more variable moisture conditions. All the climatic implications point to more continental climate conditions persisting beyond the Saalian cold stage and giving rise to the assumption that the study site was inland during the Eemian interglacial. Consequently, the marine transgression during the last interglacial proceeded less than during the Holocene. It seems that the Holocene coast line advance exceeded all previous ones in Northeast Siberia. This might be caused by a subsidence of the shelf due to neotectonical spreading of the Laptev shelf and of parts of the East Siberian shelf. The large-scale transgression that took place in the course of the Holocene involved increasing maritime influence onto the climate of the whole region. Beringia of all Arctic regions was the major refuge for Arctic plants and animals. The increased maritime influence resulted in the transformation of vegetation toward modern tundra and in the loss of the mammoth faunal complex, which survived earlier warm stages in Beringia. It seems that proximity to the sea exerts more influence on regional climate in high latitudes than do global temperature trends. Summer temperature and growing-season length are crucial for life in the Arctic. Both depend on the degree of continentality.

See also: Plant Macrofossil Introduction. **Glaciations:** Late Pleistocene Glacial Events in Beringia; Late Pleistocene in Eurasia. **Permafrost and Periglacial Features:** Thermokarst Topography; Yedoma: Late Pleistocene Ice-Rich Syngenetic Permafrost of Beringia. **Plant Macrofossil Methods and Studies:** Megafossils; Rodent Middens. **Plant Macrofossil Records:** Arctic North America. **Vertebrate Records:** Early and Middle Pleistocene of Northern Eurasia; Late Pleistocene Megafaunal Extinctions.

References

- Aleksandrova VD (1980) *The Arctic and Antarctic: Their Division into Geobotanical Areas*. Cambridge: Cambridge University Press.
- Alm T and Birks HH (1991) Late Weichselian flora and vegetation of Andøya, northern Norway – Macrofossil (seed and fruit) evidence from Nedre Æråsvatn. *Nordic Journal of Botany* 11: 465–476.
- Andreev AA, et al. (2009) Weichselian and Holocene palaeoenvironmental history of the Bol'shoy Lyakhovsky Island, New Siberian Archipelago, Arctic Siberia. *Boreas* 38(1): 72–110.
- Bauch HA, Mueller-Lupp T, Taldenkova E, et al. (2001) Chronology of the Holocene transgression at the North Siberian margin. *Global and Planetary Change* 31(1–4): 125–139.
- Binney HA, Willis KJ, Edwards ME, et al. (2009) The distribution of late-Quaternary woody taxa in Eurasia: Evidence from a new macrofossil database. *Quaternary Science Reviews* 28(23–24): 2445–2464.
- Birks HH (1991) Holocene vegetational history and climatic change in west Spitsbergen – Plant macrofossils from Skardtjørna, an Arctic lake. *The Holocene* 1(3): 209–218.
- Birks HH (2000) Aquatic macrophyte vegetation development in Kråkenes Lake, western Norway, during the late-glacial and early-Holocene. *Journal of Paleolimnology* 23(1): 7.
- CAPE-Last Interglacial Project Members (2006) Last Interglacial Arctic warmth confirms polar amplification of climate change. *Quaternary Science Reviews* 25(13–14): 1383–1400.
- CAVM Team (2003) Circumpolar arctic vegetation map. *Conservation of Arctic Flora and Fauna (CAFF) Map No. 1*. Anchorage, AK: U.S. Fish and Wildlife Service.
- Edwards ME, Brubaker LB, Lozhkin AV, and Anderson PM (2005) Structurally novel biomes: A response to past warming in Beringia. *Ecology* 86(7): 1696–1703.
- Frenzel B (1968) *Grundzüge der pleistozänen Vegetationsgeschichte Nord-Eurasiens (Basic Principles of North Eurasian Vegetation History During the Pleistocene)*. Wiesbaden: Franz Steiner Verlag (in German).
- Gubin SV, Maksimovich SV, and Zanina OG (2001) Composition of seeds from fossil Gopher burrows in ice-loess deposits of Zelyony Mys as environmental indicator. *Kryosfera Zemli (Earth's Cryosphere)* 5(2): 76–82 (in Russian).
- Guthrie RD (1990) *Frozen Fauna of the Mammoth Steppe*, p. 323. Chicago: University of Chicago Press.
- Kaplina TN (1981) Istoriya merzlykh tolshech cevernoy Yakutii v pozdnyom kaynozoye (The history of permafrost in North Yakutia during the late Cenozoic). In: Dubikov GI and Baulin VV (eds.) *Istoriya razvitiya mnogolyetnemmerzlykh porod Yevrazii (The History of the Development of Perennially Frozen Deposits in Eurasia)*, pp. 153–181. Moscow: Nauka.
- Kaplina TN, Kartashova GG, Nikitin VP, and Shilova GH (1983) New data about the sand sequence in the Tuostakh depression. *Byulleten komissii po izucheniyu chetvertichnogo perioda* 52: 107–122 (in Russian).
- Kaplina TN, Sher AV, Giterman RE, et al. (1980) The key Section of Pleistocene deposits on the Allaikha River (lower reach of the Indigirka River). *Byulleten komissii po izucheniyu chetvertichnogo perioda* 50: 73–95 (in Russian).
- Kats SV, Kats NYA, Giterman RE, and Sher AV (1982) Lower Pleistocene flora of the eastern Primore lowland. In: Tolmachev AI (ed.) *The Arctic Ocean and its Coast in the Cenozoic Era*, pp. 484–492. New Delhi: Amerind Publishing.
- Kienast F, Schirmermeister L, Siegert C, and Tarasov P (2005) Palaeobotanical evidence for warm summers in the East Siberian Arctic during the last cold stage. *Quaternary Research* 63(3): 283–300.
- Kienast F, Siegert C, Dereviagin A, and Mai DH (2001) Climatic implications of Late Quaternary plant macrofossil assemblages from the Taymyr Peninsula, Siberia. *Global and Planetary Change* 31(1–4): 265.
- Kienast F, Tarasov P, Schirmermeister L, Grosse G, and Andreev AA (2008) Continental climate in the East Siberian Arctic during the last interglacial: Implications from palaeobotanical records. *Global and Planetary Change* 60(3–4): 535–562.
- Kienast F, Wetterich S, Kuzmina S, et al. (2011) Paleontological records indicate the occurrence of open woodlands in a dry inland climate at the present-day Arctic coast in western Beringia during the last interglacial. *Quaternary Science Reviews* 30(17/18): 2134–2159. <http://dx.doi.org/10.1016/j.quascirev.2010.11.024>.
- Kremenetskiy CV, Sulerzhitsky LD, and Hantemirov R (1998) Holocene history of the northern range limits of some trees and shrubs in Russia. *Arctic and Alpine Research* 30(4): 317–333.
- Lang G (1994) *Quartäre Vegetationsgeschichte Europas (Quaternary Vegetation History of Europe)*. Jena: Gustav Fischer Verlag (in German).

- Lozhkin AV and Anderson PM (1995) The last interglaciation in Northeast Siberia. *Quaternary Research* 43(2): 147–158.
- MacDonald GM, Beilman DW, Kremenetski KV, Sheng Y, Smith LC, and Velichko AA (2006) Rapid early development of circumarctic peatlands and atmospheric CH₄ and CO₂ variations. *Science* 314(5797): 285–288.
- MacDonald GM, Velichko AA, Kremenetski CV, et al. (2000) Holocene treeline history and climate change across Northern Eurasia. *Quaternary Research* 53(3): 302–311.
- Mai DH (1985) Entwicklung der Wasser- und Sumpfpflanzen-Gesellschaften Europas von der Kreide bis ins Quartär (The development of the water and swamp plant associations in Europe from the Cretaceous to the Quaternary). *Flora* 176: 449–511 (in German).
- Mangerud J, Jakobsson M, Alexanderson H, et al. (2004) Ice-dammed lakes and rerouting of the drainage of northern Eurasia during the Last Glaciation. *Quaternary Science Reviews* 23(11–13): 1313–1332.
- Mol D, Tikhonov A, van der Plicht J, et al. (2006) Results of the CERPOLEX/ Mammuthus expeditions on the Taimyr Peninsula, Arctic Siberia, Russian Federation. *Quaternary International* 142–143: 186–202.
- Nikitin VP (2006) *Palaeocarpology and Stratigraphy of the Palaeogene and Neogene in Asian Russia*, p. 229. Novosibirsk: Academic Publishing House 'Geo' (in Russian).
- Romanovskii NN (1961) On the Structure of the Alluvial Plain in the Yana-Indigirka Coastal Lowlands and the Conditions of its Formation. *Merzlotnye Issledovaniya (Geocryological Studies)*. Moscow: Moscow State University.
- Schirmermeister L, Grosse G, Kunitsky V, et al. (2009) Periglacial landscape evolution and environmental changes of Arctic lowland areas for the last 60,000 years (Western Laptev Sea coast, Cape Mamontov Klyk). *Polar Research* 27: 249–272.
- Schirmermeister L, Grosse G, Schnelle M, et al. (2010) Late Quaternary paleoenvironmental records from the western Lena Delta Arctic Siberia. *Palaeogeography, Palaeoclimatology, Palaeoecology* 299(1–2): 175–196. <http://dx.doi.org/10.1016/j.palaeo.2010.10.045>.
- Sher AV (1997) A brief overview of the late-Cenozoic history of the western Beringian lowlands. In: Edwards ME, Sher AV, and Guthrie RD (eds.) *Terrestrial Paleoenvironmental Studies in Beringia, Proceedings of a joint Russian-American Workshop*, June 1991, pp. 3–6.
- Sher AV, Kuzmina SA, Kuznetsova TV, and Sulerzhitsky LD (2005) New insights into the Weichselian environment and climate of the East Siberian Arctic, derived from fossil insects, plants, and mammals. *Quaternary Science Reviews* 24(5–6): 533–569.
- Stuart AJ, Sulerzhitsky LD, Orlova LA, Kuzmin YV, and Lister AM (2002) The latest woolly mammoths (*Mammuthus primigenius* Blumenbach) in Europe and Asia: A review of the current evidence. *Quaternary Science Reviews* 21(14–15): 1559–1569.
- Tomskaya AI (2000) *Food Base of the Mammoth During the Late Pleistocene of Yakutia*, p. 59. Yakutsk: Academy of Sciences of Yakutia (in Russian).
- van Geel B, Aptroot A, Baittinger C, et al. (2008) The ecological implications of a Yakutian mammoth's last meal. *Quaternary Research* 69(3): 361–376.
- Vazhenina LN (1996) Seed flora of a Wrangel Island peat bog. In: Bychkov YuM (ed.) *Quaternary Climates and Vegetation of Beringia*, pp. 43–49. Magadan: Russian Academy of Sciences (in Russian).
- Voeikov Main Geophysical Observatory (1981) Climatic atlas of Asia. In: WMO and UNESCO (eds.) *Goscomgidromet USSR*. Leningrad: Gidrometeoizdat.
- Walker DA, Bockheim JG, Chapin III FS, Eugster W, Nelson FE, and Ping CL (2001) Calcium-rich tundra, wildlife, and the 'Mammoth Steppe'. *Quaternary Science Reviews* 20(1–3): 149–163.
- Walter H (1954) Grundlagen der Pflanzenverbreitung, vol. 2: Arealkunde (floristisch-historische Geobotanik) (Basics of Plant Distribution, vol. 2: Floristic-Historical Geobotany). Stuttgart: Gustav Fischer Verlag (in German).
- Walter H (1974) Die vegetation osteuropas, Nord- und Zentralasiens. The Vegetation of Eastern Europe, North and Central Asia, p. 452. Stuttgart: Gustav Fischer Verlag (in German).
- Wetterich S, Kuzmina S, Andreev AA, et al. (2008) Palaeoenvironmental dynamics inferred from late Quaternary permafrost deposits on Kurungnakh Island, Lena Delta, Northeast Siberia, Russia. *Quaternary Science Reviews* 27(15–16): 1523–1540.
- Yelovskaya LG, Konorovsky AK, and Savinov DD (1966) *Salt-Rich Soils Above Permafrost (Kryosols) in Central Yakutia*. Moscow: Nauka (in Russian).
- Yurtsev BA Jr. (1982) Relics of the xerophyte vegetation of Beringia in Northeastern Asia. In: Hopkins DM, Matthews JV, Schweger CE, and Young SB (eds.) *Paleoecology of Beringia*, pp. 157–177. New York: Academic Press.
- Yurtsev BA (2001) The Pleistocene 'Tundra-Steppe' and the productivity paradox: The landscape approach. *Quaternary Science Reviews* 20(1–3): 165–174.
- Zazula GD, Schweger CE, Beaudoin AB, and McCourt GH (2006) Macrofossil and pollen evidence for full-glacial steppe within an ecological mosaic along the Bluefish River, eastern Beringia. *Quaternary International* 142–143: 2–19.

Arctic North America

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This article presents a summary of our current knowledge of the vascular plant macrofossil record for northern North America from the onset of late Cenozoic glaciations (ca. 2.5 Mya) through the Holocene (all ages in calibrated or calendar years ago (ya)). The geographic range for this article extends from Alaska to eastern Canada, from the high Arctic to approximately 60° N ([Figure 1](#)). Where possible, we have used common names for the plants. [Table 1](#) lists the common names, scientific names, and ecological affinities of the taxa mentioned in this article. Citations and site details are included in [Table 2](#).

For the purposes of this article, we define plant macrofossils as vegetative and reproductive parts of vascular plants that may be seen with the naked eye. Vegetative parts include leaves or twigs, whereas reproductive parts include fruits, seeds, and flowers. Macrofossils contrast with megafossils, which typically refer to larger remains such as logs. Plant macrofossils may be recovered within, but are not restricted to, lacustrine cores, peat, colluvium, and alluvium. In rare circumstances, plant macrofossils are found as *in situ* buried vegetation where whole or partial plants are preserved in their former growth position.

Northern Climate and Landscapes

The broadest patterns of northern climate, now and in the past, are a function of latitude, proximity to oceans, and the influence of continental-scale topography ([Figure 2](#)). Coastal regions have maritime-type climates with relatively high precipitation totals and a lack of summer/winter temperature extremes. Interior regions have continental-type climates with lower precipitation totals and greater summer/winter extremes. In the High Arctic, precipitation is very low with cold winters and cold summers. In all regions, most precipitation occurs in the late summer to early winter, with a maximum on the southeastern Alaska coast (>2000 mm per year) and a minimum in the High Arctic (<50 mm per year).

The modern vegetation of northern North America is a mixture of forest and muskeg south of treeline and a variety of tundra types to the north (Flora of North America, 1993+; [CAVM, 2003](#)). In the forested region (boreal forest or taiga), coniferous or mixed forests dominate. In the western regions, white and black spruce are the dominant conifers, whereas in eastern regions jack pine and larch become dominant. In some areas, especially those affected by frequent forest fires, deciduous trees, such as resin birch, cottonwood, and aspen, become important forest components. The boreal forest also has a significant shrub component of willow, alder, birch, and various ericaceous shrubs.

Much of the forested region is underlain by discontinuous permafrost, which is often found on north-facing slopes or in valley bottoms. In these areas, the vegetation is a scattered scrubby black spruce or larch woodland interspersed with grasses and tussock-forming sedges. Mosses and acidophilic forbs and shrubs are also common. The soils are cold and poorly drained, often with standing water in the summer. The permafrost is easily disturbed in this region, resulting in pond formation (thermokarst) and catastrophic ground surface collapse.

Regionally, the northern limit of trees ([Figure 3](#)) is controlled by climate, but on a local scale, factors such as aspect and drainage exert a strong influence. Spruce and pine are the dominant treeline species across the continent, although stands of extralimital cottonwood are present in Alaska. Tree-line in central Canada is a gradational boundary, with a wide swath of lichen woodland and forest–tundra separating the closed forest from the tundra ([Elliott-Fisk, 2000](#)).

The tundra to the north is variable, ranging from a shrub-dominated vegetation to graminoid tundra and barrens. The key climatic variables controlling tundra vegetation are summer temperature and winter snow cover, but local factors, such as topography and substrate (because they affect soil temperature, snow depth, soil nutrient status, and drainage), also strongly influence the tundra vegetation ([CAVM, 2003](#)). Permafrost is continuous throughout the tundra zone, which affects the vegetation through a combination of cold soil temperatures, impeded drainage, and thermokarst. The shrub tundra is a mix of birch, alder, willow, and ericaceous shrubs with graminoids (sedge and grasses) and mosses. Graminoid tundra is dominated by herbs and forbs, whereas barrens are dominated by herbs, lichens, and mosses. Large areas of the tundra zone are very poorly drained and are dominated by wet sedge (often tussock forming) and grass meadows (sometimes with low shrubs).

Summary of Macrofossil Sites

Plant macrofossil analysis, conducted alone or in conjunction with other paleoecological methods, has greatly added to our understanding of Arctic vegetation history. Our discussion is limited to well-documented macrofossil sites for key time intervals in the Quaternary. Due to space limitations, we have not attempted to provide an exhaustive list of all sites that contain plant macrofossils. An important feature of all northern macrofossil sites is their excellent preservation, even for very old sites. This is due, especially in the High Arctic, to the presence of permafrost.

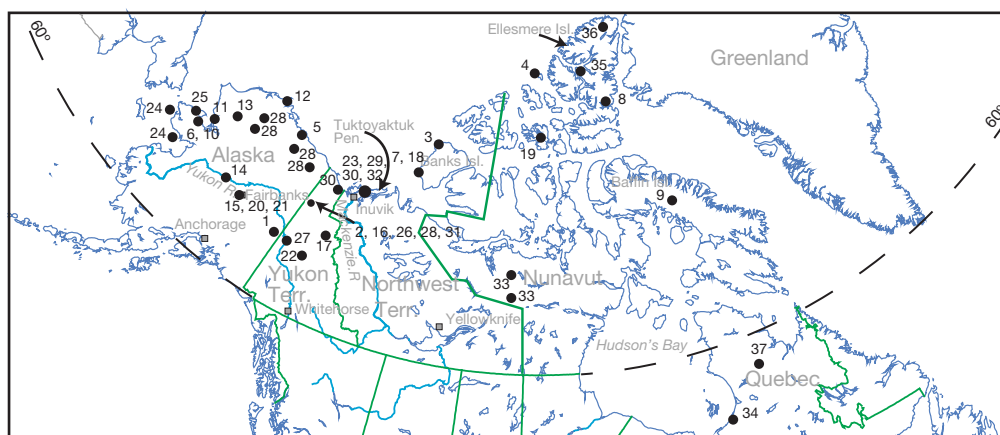


Figure 1 Localities mentioned in the text. Sites are marked by circles; site numbers correlate with [Table 2](#).

Table 1 Taxa mentioned in the text^a

<i>Taxon</i>	<i>Common name</i>	<i>Plant growth form</i>	<i>Typical site and biome</i>
Pteridophyta	Ferns	Herb	Boreal forest and tundra
Pinaceae			
<i>Picea</i> sp.	Spruce	Tree	Boreal forest
<i>Larix</i> sp.	Larch, tamarack	Tree	Muskeg, wet places, boreal forest
<i>Larix laricina</i> (Du Roi) K. Koch	Larch, tamarack	Tree	Muskeg, wet places, boreal forest
<i>Picea glauca</i> (Moench) Voss s.l.	White spruce	Tree	Well-drained to bogs, boreal forest
<i>Picea mariana</i> (Mill.) B.S.P.	Black spruce	Tree	Muskeg, poorly drained, boreal forest
<i>Pinus</i> (<i>Strobus</i>) subsect. <i>Cembrae</i>	Old World five-needled pine	Tree	Extinct in North America; east Asian origin
<i>Pinus</i> subsect. <i>Eustropei</i>	Old World two-needled pine	Tree	Extinct in North America; east Asian origin
<i>Pinus matthewsii</i> sp. Nov.		Tree	Extinct in North America; cone morphology similar to <i>P. contorta</i>
McKown, Stockey, and Schweger			
<i>Abies</i> sp.	Fir	Tree	Mountain slopes, acidic bedrock; forest
Typhaceae			
<i>Typha latifolia</i> L.	Cattail	Herb	Shallow water marshes; forest
Potamogetonaceae			
<i>Potamogeton vaginatus</i> Turcz.	Giant pondweed	Herb	Lakes, slow-moving rivers to 3 m depth; forest and tundra
<i>Zannichellia palustris</i> L.	Horned pondweed	Herb	Shallow (often brackish) water on edge of ponds; forest
Alismataceae			
<i>Alisma</i> sp.	Water-plantain	Herb	Emergent aquatic; southern boreal forest
<i>Sagittaria latifolia</i>	Common arrowhead, wapato, duck potato	Herb	Southern boreal forest
Poaceae			
<i>Deschampsia</i> sp.	Hair grass	Herb	Wet meadows, lakeshores, gravel bars; tundra and forest
<i>Elymus</i> sp.	Rye grass	Herb	Grassy slopes, well-drained, riverbanks; tundra and forest
<i>Glyceria</i> sp.	Manna grass	Herb	Shallow water, lakeshores, riparian, wet meadows; probably forest
<i>Hierochloë hirta</i> ssp. <i>arctica</i> G. Weim.	Sweet grass	Herb	Sandy stream banks, meadows, lakeshores; low Arctic & Alpine tundra
(= <i>H. odorata</i>)			
<i>Poa</i> sp.	Blue grass	Herb	Open slopes, stream banks; boreal forest and tundra
Cyperaceae			
<i>Carex</i> spp.	Sedge	Herb	Forest to tundra to grassland
<i>Kobresia myosuroides</i> (Vill.) Fiori & Paol.		Herb	Dry, calcareous sites; tundra
Araceae			
<i>Aracites globosa</i> (C. & E. Reid) Benn.		Herb	Extinct, wetland?
<i>Epipremnum crassum</i> C. & E. Reid		Herb	Extinct, wetland?

(Continued)

Table 1 (Continued)

<i>Taxon</i>	<i>Common name</i>	<i>Plant growth form</i>	<i>Typical site and biome</i>
Liliaceae			
<i>Allium</i> sp.	Wild onion	Herb	Grassy slopes, riverbanks, shorelines; forest and tundra
Salicaceae			
<i>Populus balsamifera</i> L.	Balsam poplar, cottonwood	Tree	Floodplains, alluvial flats, boreal forest
<i>Populus tremuloides</i> Michx.	Aspen	Tree	Open woods and along creeks, pioneer on burns; boreal forest
<i>Salix</i> sp.	Willow	Tree or shrub	Boreal forest or shrub tundra
<i>Salix arctica</i> Pall.	Arctic willow	Prostrate dwarf shrub	Sedge meadows to dry tundra and heath
Myricaceae			
<i>Myrica arctogale</i> Benn	Extinct sweet gale	Shrub	Extinct, bogs?
<i>Myrica gale</i> L.	Sweet gale	Low shrub	Mesic sites, boreal forest
Betulaceae			
<i>Alnus</i> sp.	Alder	Low or tall shrub	Stream banks, slopes, lakeshores; boreal forest and shrub tundra
<i>Betula glandulosa</i> Michx.	Shrub birch	Low shrub	Muskeg; forest and shrub tundra
<i>Betula nana</i> L.	Dwarf birch	Dwarf shrub	Muskeg; shrub tundra
<i>Betula neoalaskana</i> Sarg.	Alaska birch	Tree	Acid and peaty sites; boreal forest
<i>Betula papyrifera</i> Marsh.	Resin birch, paper birch	Tree	Open woodland; boreal forest
Polygonaceae			
<i>Polygonum</i> sp.	Knotweed	Herb	Pioneer, mesic sites, turfy sites, terrestrial and aquatic; forest and tundra
<i>Polygonum viviparum</i> L.	Alpine bistort	Herb	Turfy, rocky barrens, moist grassy herb mats; forest and tundra
Chenopodiaceae			
<i>Chenopodium</i> sp.		Herb	Disturbed areas, sandy, dry, often saline soils; forest and tundra
<i>Chenopodium simplex</i> (Torr.) Raf. (= <i>C. gigantospermum</i> Aellen)	Maple leaf goosefoot	Herb	Disturbed areas, hardwood slopes, wooded tallus slopes; forest
<i>Corispermum hyssopifolium</i> L.	Bugseed	Herb	Sandy sites, gravel bars on rivers; forest
Caryophyllaceae			
<i>Cerastium</i> sp.	Mouse-ear chickweed	Herb	Disturbed, grassy slopes, gravelly, meadows; forest and tundra
<i>Minuartia</i> spp.	Sandwort	Herb	Mesic or well-drained stony, gravelly, dry slopes; forest and tundra
<i>Silene involucreta</i> (Cham. & Schlecht.) Bocquet (= <i>Melandrium affine</i> J. Vahl)	Campion	Herb	Stony, gravelly, sandy, rocky sites; forest and tundra
Nymphaeaceae			
<i>Nymphaea</i>	Water lily	Herb	Emergent aquatic; boreal forest
Ranunculaceae			
<i>Ranunculus</i> spp.	Buttercup	Herb	Aquatic and terrestrial herbs; forest and tundra
<i>Ranunculus hyperboreas</i> type	Arctic buttercup	Herb	Shallow fresh or brackish ponds, mud flats, pond borders; forest and tundra
<i>Ranunculus aquatilis</i> L. (= <i>Ranunculus trichophyllus</i> Chaix)	White water buttercup	Herb	Submerged aquatic, sometimes calcareous ponds; forest and tundra
<i>Ranunculus lapponicus</i> L.	Lapland buttercup	Herb	Damp moss, willow thickets, bogs; mainly tundra
Papaveraceae			
<i>Papaver</i> sp.	Arctic poppy	Herb	Turfy sites, herb mats, gravelly, calcareous outcrops; tundra
Brassicaceae			
<i>Draba</i> sp.		Herb	Open slopes, calciphilous, mainly tundra
<i>Erysimum</i> cf. <i>cheiranthoides</i> L.	Wormseed mustard	Herb	Moist turf sites, creek banks, lake shores; forest and tundra
<i>Lepidium densiflorum</i> Schrad.	Common pepper-grass	Herb	Dry, open disturbed sites; forest and tundra
Rosaceae			
<i>Chamaerhodos erecta</i> L.	Little rose	Herb	Dry sandy gravelly slopes; forest
<i>Dryas</i> sp.	Mountain avens	Prostrate dwarf shrub	Disturbed, well-drained, dry, gravelly sites, calciphilous; forest and tundra

(Continued)

Table 1 (Continued)

<i>Taxon</i>	<i>Common name</i>	<i>Plant growth form</i>	<i>Typical site and biome</i>
<i>Potentilla</i> sp.	Cinquefoil	Herb or shrub	Often in well-drained disturbed sites, but also moist, turfy sites; forest and tundra
<i>Rubus spectabilis</i>	Salmonberry	Low shrub	Moist woods; coastal boreal forest
<i>Rubus idaeus</i> L.	Raspberry	Dwarf shrub	Thickets, edge of woods, riverbanks; forest
Linaceae			
<i>Linum lewisii</i> Pursh(= <i>Linum perenne</i> L.)	Flax	Herb	Meadows, gravelly banks, sandy areas; mainly forest
Empetraceae			
<i>Empetrum nigrum</i> L.	Crowberry	Dwarf shrub	Heath, swamps, bogs; forest and tundra
Hippuridaceae			
<i>Hippuris vulgaris</i> L.	Mare's tail	Herb	Emergent aquatic, shallow ponds, slow-moving streams, lakes; forest and tundra
Apiaceae			
<i>Bupleurum americanum</i> Coult. & Rose (= <i>B. triradiatum</i> Adams)	Thoroughwort	Herb	Meadows, moist sand, talus slopes, gravel banks; forest and tundra
<i>Sium suave</i> Walt.	Water parsnip	Herb	Emergent aquatic, muddy shores, stream and lake margins; mainly forest
Ericaceae			
<i>Andromeda polifolia</i> L.	Bog rosemary	Low shrub	Muskeg, moist meadows, wet tundra and boreal forest
<i>Arctostaphylos uva-ursi</i> (L.) Spreng. s.l.	Bearberry, kinnikinnick	Dwarf shrub	Exposed rocks, riverbanks, sandy sites, open woods; mainly forest
<i>Cassiope tetragona</i> (L.) D. Don		Dwarf shrub	Dry, rocky sites; tundra
<i>Chamaedaphne calyculata</i> (L.) Moench	Leatherleaf, cassandra	Low or dwarf shrub	Bogs, muskegs, lake margins; forest and tundra
<i>Ledum decumbens</i> (Ait.) Lodd.		Dwarf shrub	Heaths, dry, rocky sites; forest and tundra
<i>Ledum groenlandicum</i> Oeder	Labrador tea	Low shrub	Peat soils, bogs, muskeg; forest and tundra
<i>Vaccinium vitis-idaea</i> L.	Low bush cranberry, lingon berry	Dwarf shrub	Acid soils, muskeg; forest and tundra
Primulaceae			
<i>Androsace septentrionalis</i> L.	Pygmy flower	Herb	Dry rocky, sandy sites, disturbed sites, calcophile tundra and grasslands
Menyanthaceae			
<i>Menyanthes trifoliata</i> L.	Buckbean	Herb	Emergent aquatic, bogs, lake and stream margins; forest and tundra
Polemoniaceae			
<i>Phlox hoodii</i> Richards.	Moss phlox	Herb	Open slopes, dry tundra and grasslands
Lamiaceae			
<i>Lycopus</i> sp.	Tuberous water-horehound	Herb	Wet sand, lakeshores and along streams; forest
<i>Stachys</i> sp.	Hedge-nettle	Herb	Fields, clearings, hot springs, moist woods; forest
Caprifoliaceae			
<i>Sambucus</i> sp.		Shrub	Extinct species; modern sp. in moist woods and sub-Alpine meadows
Asteraceae			
<i>Artemisia frigida</i> Willd.	Prairie sagewort	Low or dwarf shrub	Open sandy slopes, disturbed sites; dry grasslands, rarely tundra
<i>Artemisia</i> sp.	Wormwood	Herb or shrub	Dry and mesic sites; forest and tundra
<i>Bidens frondosa</i> L.	Burr	Herb	Disturbed areas; shorelines; forest
<i>Taraxacum ceratophorum</i> (Ledeb.) DC. s.l.	Dandelion	Herb	Meadows and moist areas; forest and tundra

^aNomenclature follows Cody (1996). Common names, growth form, and typical site/biome are from personal observations, floras (Hultén, 1968; Cody, 1996), or, in the case of extinct taxa, from the cited reference (see Table 2). Shrub classifiers (tall, >2 m; low, 40 cm to 2 m; and dwarf, <40 cm) are from CAVM (2003).

Late Pliocene (3.6–1.8 Mya)

Although not Quaternary, the late Pliocene is an important time interval in Arctic North America because Quaternary-style continental glaciation began at this time and the modern flora

became established. Plant macrofossils from several sites across Arctic and sub-Arctic North America provide evidence for a floristically rich boreal forest that persisted until the onset of glaciation and Northern Hemisphere cooling approximately 2.5 Mya.

Table 2 Key sites mentioned in the text^a

Site No.	Age	Site name	Location	Unit and/or section	Key taxa	Inferred past vegetation	Modern vegetation	Source
1	Late Pliocene	Lost Chicken Mine	Central Alaska	LC-II: Unit L7	<i>Picea</i> , <i>Larix</i> , extinct <i>Pinus</i> (5-needle), <i>Aricites globosa</i> , <i>Sambucus</i>	Forest	Boreal forest	Matthews et al. (2003) <i>Quaternary Research</i> 60: 9–18
2	Late Pliocene	Ch'jee's Bluff	Northern Yukon Territory	Units 1 and 2	<i>Abies</i> , <i>Larix</i> , <i>Picea</i> , extinct <i>Pinus</i> (5-needle), <i>Pinus matthewsii</i> , <i>Aricites globosa</i> , <i>Menyanthes</i> , <i>Sambucus</i>	Forest	Boreal forest	Matthews and Ovenden (1990) <i>Arctic</i> 43: 364–392
3	Late Pliocene	Ballast Brook	Banks Island	Upper unit	Extralimital <i>Larix</i> , <i>Picea</i> , <i>Alnus</i> , <i>Betula</i> (tree type); extinct <i>Pinus</i> (2- and 5-needle), <i>Epipremnum crassum</i> , <i>Aricites globosa</i>	Forest	Barren or prostrate shrub tundra	Matthews and Ovenden (1990) <i>Arctic</i> 43: 364–392
4	Late Pliocene	Beaufort Formation	Meighen Island		Extralimital <i>Larix</i> , <i>Thuja occidentalis</i> , <i>Alnus</i> , <i>Betula</i> (tree type); extinct <i>Picea</i> , <i>Pinus</i> (2- and 5-needle), <i>Epipremnum crassum</i>	Forest	Barren or prostrate shrub tundra	Matthews and Ovenden (1990) <i>Arctic</i> 43: 364–392
5	Late Pliocene/early Pleistocene	Fish Creek	Northern Alaska		Extralimital <i>Larix</i> , <i>Alnus</i>	Boreal forest	Shrub tundra	Matthews and Ovenden (1990) <i>Arctic</i> 43: 364–392
6	Early Pleistocene	Cape Deceit	Seward Peninsula	Cape Deceit Formation	Extralimital <i>Larix laricina</i>	Northern boreal forest	Wetland tundra	Matthews and Ovenden (1990) <i>Arctic</i> 43: 364–392
7	Mid-Pleistocene	Duckhawk Bluffs	Banks Island	Morgan Bluff Formation	Extralimital <i>Picea</i> , <i>Larix</i> , <i>Betula</i> shrubs, <i>Menyanthes trifoliata</i>	Boreal forest	Graminoid prostrate dwarf shrub tundra	Matthews et al. (1986) <i>Géographie Physique et Quaternaire</i> 40: 279–298
8	Pleistocene	Makinson Inlet	Ellesmere Island		Extralimital <i>Menyanthes trifoliata</i> , <i>Andromeda polifolia</i>	Erect shrub tundra	Barren or prostrate shrub tundra	Blake and Matthews (1979) <i>Geological Survey Paper 79-1a, Current Research Part A</i>
9	Late Pliocene to mid-Pleistocene	Isortoq/ Flitaway	Baffin Island		Extralimital <i>Betula nana/glandulosum</i> , <i>Ledum groenlandicum</i> , <i>Vaccinium vitis-idaea</i>	Erect shrub tundra	Cryptogam barren	Morgan et al. (1993) <i>Canadian Journal of Earth Science</i> 30: 954–974
10	MIS 5?	Cape Deceit	Seward Peninsula	Peat 4, Deering Formation	Extralimital <i>Picea</i> and <i>Betula papyrifera</i>	Boreal forest	Wetland tundra	Matthews (1974) <i>Geological Society of America Bulletin</i> 85: 1353–1384
11	MIS 5	Baldwin Peninsula	Kotzebue Sound, Alaska		Extralimital <i>Picea</i>	Woodland?	Wetland tundra	Hamilton and Brigham-Grette (1991) <i>Quaternary International</i> 10–12: 49–71
12	MIS 5	Pelukian shorelines	Barrow, Alaska		Extralimital <i>Picea</i>	Woodland?	Wetland tundra	Hamilton and Brigham-Grette (1991) <i>Quaternary International</i> 10–12: 49–71
13	MIS 5	Nk-37	Northwest Alaska		Extralimital <i>Picea</i>	Open spruce woodland	Erect low-shrub tundra	Edwards et al. (2003) <i>Arctic, Antarctic, and Alpine Research</i> 35: 460–468

14	MIS 5	Palisades	Central Alaska	Upper peat/ forest bed		Boreal forest	Boreal forest	Matheus et al. (2003) <i>Quaternary Research</i> 60: 33–43
15	MIS 5	Eva Creek	Central Alaska		Incr. Polypodiaceae	Boreal forest	Boreal forest	Bigelow, unpublished; Muhs et al. (2003) <i>Quaternary Science Reviews</i> 22: 1947–1986
16	MIS 5	Ch’ijee’s Bluff	Northern Yukon Territory		Extralimital <i>Chenopodium gigantospermum</i> , <i>Rubus spectabilis</i> , <i>Sium suave</i> , <i>Stachys</i> , and <i>Bidens frondosa</i>	Boreal forest	Boreal forest, near its northern limit	Matthews et al. (1990) <i>Géographie Physique et Quaternaire</i> 44: 341–362
17	MIS 5	Hungry Creek	Central Yukon Territory		Extralimital <i>Glyceria</i> and <i>Nymphaea</i>	Boreal forest	Boreal forest	Schweger and Matthews (1991) <i>Quaternary International</i> 10–12: 85–94
18	MIS 5	Duckhawk Bluffs	Banks Island	Cape Collinson Formation	Extralimital <i>Ranunculus lapponicus</i> , <i>Betula glandulosa</i>	Erect shrub tundra?	Graminoid prostrate dwarf shrub tundra	Matthews et al. (1986) <i>Géographie Physique et Quaternaire</i> 40: 279–298
19	MIS 5?	Several sites	Bathurst Island		Extralimital or currently rare <i>Cassiope tetragona</i> , <i>Ranunculus trichophyllus</i> , and <i>Hippuris vulgaris</i>	Prostrate shrub tundra?	Graminoid tundra, with or without prostrate shrubs	Blake (1974) <i>Canadian Journal of Earth Sciences</i> 11: 1025–1042
20	MIS 3	Isabella Basin	Central Alaska		<i>Picea</i> , <i>Arctostaphylos uva-ursi</i> , <i>Alisma</i> , <i>Chenopodium</i>	Forest–tundra with some steppic elements	Boreal forest	Matthews (1974) <i>Canadian Journal of Earth Sciences</i> 11: 828–841
21	MIS 3	Fox Permafrost Tunnel	Central Alaska		<i>Carex</i> , <i>Salix</i> , <i>Chenopodium</i> , <i>Potentilla</i>	Mesic to dry tundra with isolated spruce	Boreal forest	Hamilton et al. (1988) <i>Geological Society of America Bulletin</i> 100: 948–969
22	MIS 3	Mayo Indian Village	Northern Yukon Territory		<i>Corispermum hyssopifolium</i> , <i>Draba</i> , <i>Potentilla</i>	Forest–tundra	Boreal forest	Matthews et al. (1990) <i>Géographie Physique et Quaternaire</i> 44: 15–26
23	MIS 3	Kittigazuit Formation	Tuktoyaktuk Peninsula		<i>Corispermum hyssopifolium</i> , extralimital or currently rare <i>Allium</i> , <i>Linum</i> , <i>Chamaerhodos erecta</i> , <i>Lycopus</i> , <i>Stachys</i> , <i>Zanichellia palustris</i> , <i>Picea Menyanthes trifoliata</i> , <i>Carex</i>	Dry steppe–tundra, possibly scattered spruce	Low-shrub tundra	Dallimore et al. (1997) <i>Canadian Journal of Earth Sciences</i> 34: 1421–1444
24	MIS 2	Submerged Shelf Cores	Bering Sea			Mesic to wet birch-graminoid tundra	Submerged	Elias et al. (1996) <i>Nature</i> 382: 60–63
25	MIS 2	Kitluk Soil	Seward Peninsula		<i>Kobresia myosuroides</i> , <i>Potentilla</i> , <i>Draba</i>	Dry herb-graminoid tundra	Sedge, moss, dwarf shrub wetland	Goetcheus and Birks (2001) <i>Quaternary Science Reviews</i> 20: 135–147
26	MIS 2	Bluefish Basin	Northern Yukon Territory		<i>Poa</i> , <i>Elymus</i> , <i>Draba</i> , <i>Androsace septentrionalis</i> , <i>Artemisia frigida</i> , <i>Phlox</i>	Steppe–tundra	Boreal forest	Zazula et al. (2006) <i>Quaternary International</i> 142–143: 2–16

(Continued)

Table 2 (Continued)

Site No.	Age	Site name	Location	Unit and/or section	Key taxa	Inferred past vegetation	Modern vegetation	Source
27	MIS 2	Klondike 'mucks'	Central Yukon Territory		<i>Poa</i> , <i>Elymus</i> , <i>Draba</i> , <i>Androsace septentrionalis</i> , <i>Artemisia frigida</i> , <i>Phlox</i>	Steppe–tundra	Boreal forest	Zazula et al. (2005) <i>Quaternary Research</i> 63: 189–198
28	Early Holocene	Various sites	Western and northern Alaska		Extralimital <i>Populus</i>	Forest–tundra	Tundra	Mann et al. (2002) <i>Quaternary Science Reviews</i> 21: 997–1021
29	Early Holocene	Sleet Lake	Tuktoyaktuk Peninsula		Extralimital <i>Picea</i>	Forest–tundra	Low-shrub tundra	Spear (1993) <i>Review of paleobotany and Palynology</i> 79: 99–111
30	Early Holocene	Various sites	Arctic Ocean shoreline		Extralimital <i>Populus</i>	Forest–tundra	Tundra	Ovenden (1988) <i>Geological Survey of Canada Paper</i> 88-22; Ritchie (1984)
31	Early to Late Holocene	Polybog	Northern Yukon		Aquatics, graminoids, <i>Sphagnum</i>	Fen-to-bog evolution	Boreal forest, near its northern limit	Ovenden (1982) <i>Boreas</i> 11: 209–224
32	Early to Late Holocene	Kukjuk Peatland	Tuktoyaktuk Peninsula		Extralimital <i>Larix</i> and <i>Picea</i> (early Holocene)	Forest tundra (early Holocene) and tundra (fen-to-bog evolution)	Low-shrub tundra	Vardy et al. (1997) <i>Quaternary Research</i> 47: 90–104
33	Early to Late Holocene	Thelon-Kazan peatland and Baillie Bog	Nunavut		Aquatics, graminoids, <i>Betula glandulosa</i> , <i>Vaccinium vitis-idaea</i> , <i>Ledum decumbens</i> , <i>Arctostaphylos</i>	Tundra (fen-to-heath evolution)	Shrub and graminoid tundra	Vardy et al. (2005) <i>Boreas</i> 34: 324–334
34	Middle to Late Holocene	Saspima-kwananistikw peatland	Eastern Hudson's Bay		Aquatics, graminoids, <i>Salix</i> , <i>Andromeda</i> , <i>Chamaedaphne calyculata</i> , <i>Myrica gale</i> , <i>Larix</i> , <i>Picea</i> , <i>Sphagnum</i>	Tundra and forest–tundra (fen-to-bog evolution)	Forest–tundra	Arlen-Pouliot and Bhiry (2005) <i>The Holocene</i> 15: 408–419
35	Middle to Late Holocene	Hot Weather Creek	Fosheim Peninsula, Ellesmere Island		Peat-forming mosses	More bogs?	Prostrate dwarf shrub tundra	Garneau (2000) <i>Environmental Response to Climatic Change in the Canadian High Arctic</i> , GSC Bulletin 529
36	Middle to Late Holocene	Piper Pass	Northern Ellesmere Island		Peat-forming mosses	More bogs?	Cryptogam herb barren	La Farge-England et al. (1991) <i>Arctic and Alpine Research</i> 23: 80–98
37	Late Holocene	Various sites	Northern Quebec		<i>Larix laricina</i> 5 km beyond modern limit	Forest–tundra	Forest–tundra	Payette et al. (2002) <i>Ambio Special Report; Tundra-Taiga Treeline Research</i> 12: 15–22

^aModern tundra classification is based on CAVM (2003).

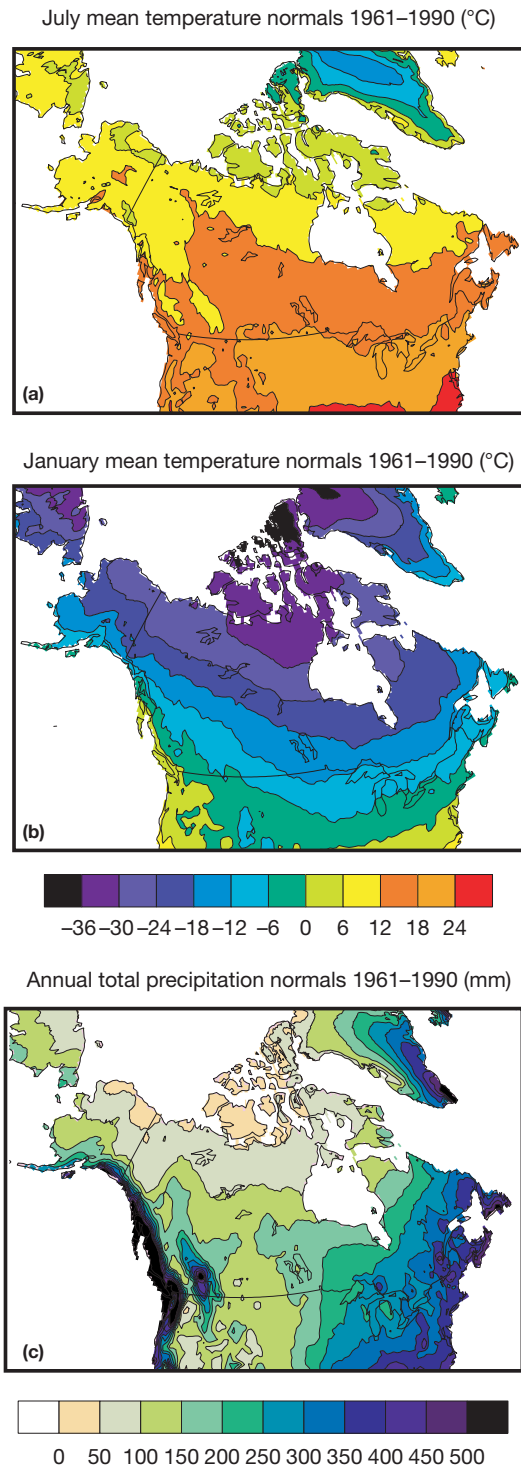


Figure 2 Modern climate of northern North America. (a) Mean July temperature. (b) Mean January temperature. (c) Mean annual precipitation. The western Cordillera and highlands of southern Alaska help establish prevailing wind patterns from northwest to southeast, resulting in the temperature pattern exhibited here. The Cordillera also heavily influences the precipitation regime, largely preventing Pacific Ocean moisture from reaching interior lowlands. Maps are based on data obtained from the Climate Research Unit, University of East Anglia, and drawn using National Center for Atmospheric Research command language.

Key sites (Lost Chicken Mine in east central Alaska, Ch'jee's Bluff and Bluefish in the northern Yukon, Ballast Brook Upper Unit of Banks Island, NWT, the Beaufort Formation on Meighen Island, several localities on Ellesmere Island; **Figure 1**, sites 1–4; **Table 2**) document a boreal forest dominated by five-needle and two-needle pines, spruce, fir, and larch, all with northern boundaries much beyond their present positions. These assemblages typically also include several now extinct taxa (*Myrica arctogale* and *Aracites globosa*) or those currently restricted much further to the south (elderberry (*Sambucus*) and *Epipremnum crassum* (Aracaceae), suggesting much higher mean January temperatures (possibly up to 10–15 °C warmer) and greater precipitation than at present. Climatic transition by 2 Mya saw the loss of many Pliocene tree and shrub taxa, with spruce or larch (Fish Creek, **Figure 1**, site 5; **Table 2**) replacing pine as the dominant conifer in the boreal forest of Quaternary interglaciations.

Early to Middle Pleistocene (ca. 1.8–0.125 Mya)

This period marks the onset of the Quaternary, the hallmark of which is the climatic shifts between glaciations and interglaciations. Interglacial periods were similar to our own climate, at times warmer than today. Glacial periods were cold and dry, with continental-scale ice sheets isolating Alaska and north-west Canada from the rest of North America. One major effect of the enormous ice sheets is the drawdown of the world's oceans, exposing the Bering Land Bridge (BLB) between Alaska and Russia. The effects of the presence or absence of the BLB cannot be overemphasized. An exposed BLB allowed migration of terrestrial plants and animals between Russia and North America but blocked migrations and water flow between the Pacific and Arctic Oceans. In addition, because of greater distance from the ocean, this subcontinental-scale landmass increased continentality and aridity in regions bordering the modern Bering Sea.

Due to biases in preservation (macrofossils are more likely to survive in moist peaty situations typical of interglaciations), the macrofossil sites discussed in this section all date to interglacial periods, when climate was at least as warm as today. The Cape Deceit Formation on the Seward Peninsula (**Figure 1**, site 6) yielded abundant plant macrofossils, which presumably dates between the early and middle Pleistocene. Plant macrofossils include larch needles (**Table 2**), and microfossils include spruce pollen. The site today is west of treeline, indicating climate was slightly warmer than modern during the early to middle Pleistocene.

Several sites from the High Arctic of Canada, although not well dated, probably date to interglaciations of the early to middle Pleistocene. All these sites (**Table 2**) contain extralimital taxa, with strong evidence that treeline was significantly farther north than today. The Morgan Bluffs Formation on Banks Island (**Figure 1**, site 7) contains spruce and larch macrofossils in addition to shrub birch (which barely grows on Banks Island today). On Baffin Island, the Isortoq and Flitaway sites (**Figure 1**, site 9) contain extralimital low Arctic and tree-line shrubs (but no trees). On Ellesmere Island, peat deposits near Makinson Inlet (**Figure 1**, site 8) contain an ericaceous shrub and an emergent aquatic that today do not grow on the island.

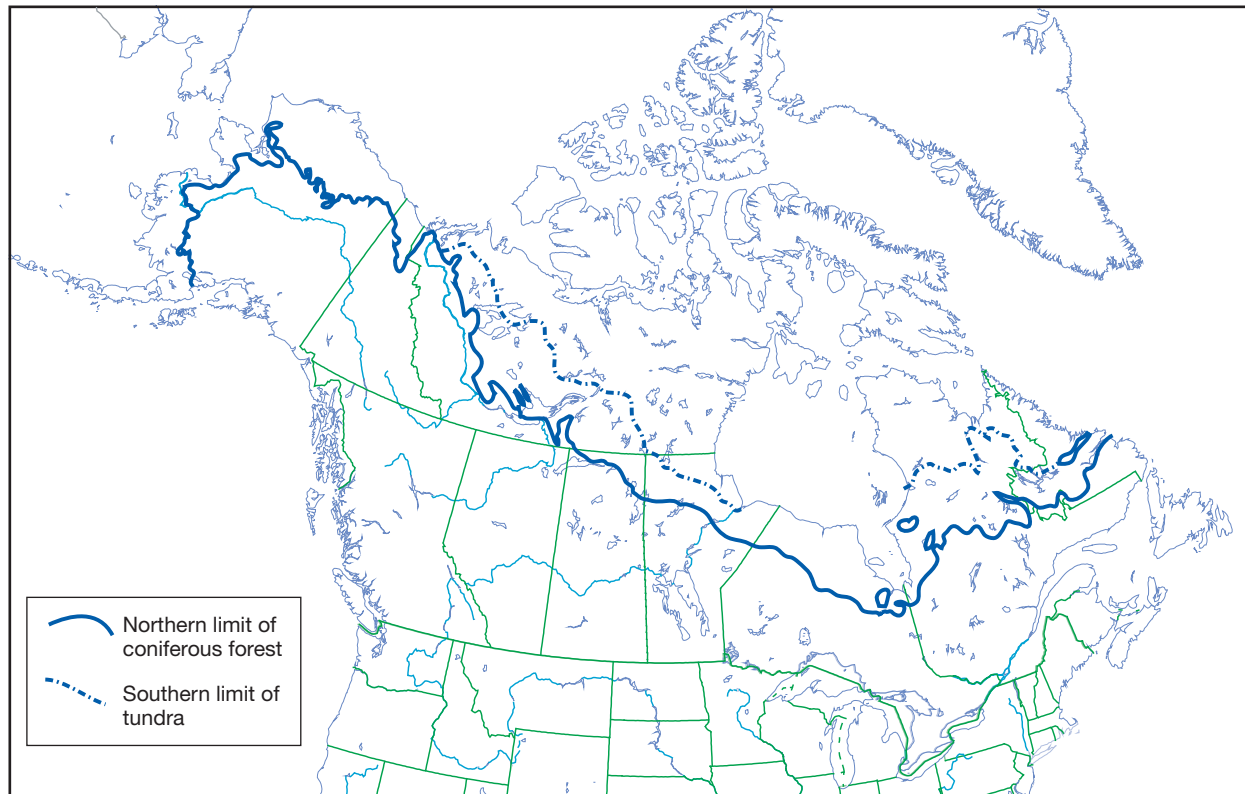


Figure 3 Treeline across Alaska and Canada. The Alaskan section is based on the vegetation map provided in Viereck and Little (1972). The Canadian section is based on the satellite-based Canada Vegetation Cover map in the *National Atlas of Canada*, fifth edition (Energy, Mines, and Resources Canada, 1993).

Last Interglaciation (ca. 125–75 ka)

Long records from marine and ice cores indicate the last interglaciation (Marine Isotope Stage 5 (MIS 5)) was a period of shifting climate from warmer than modern to modern or slightly cooler than modern climate (Petit et al., 1999). The warmest part of the interglaciation was its onset (MIS 5e), when continental glaciers had largely disappeared, the Greenland icecap was significantly smaller than today (Koerner, 1989), and sea level was slightly higher than modern (Esat et al., 1999). In the north, MIS 5 is documented in lake cores, relict shorelines, and buried peat deposits (LIGA, 1991; CAPE Last Interglacial project members, 2006). Although it is often assumed that the sites indicating the warmest temperatures probably dated to MIS 5e, the possibility that some may date to one of the warm intervals after MIS 5e (i.e., MIS 5c or 5a) cannot be excluded.

In Alaska, macrofossil remains, especially spruce, indicate forest expansion beyond current limits to the Chukchi Sea and northwards to the Arctic Ocean. The sites include the Baldwin Peninsula (Figure 1, site 11), the Pelukian shorelines east of Barrow (Figure 1, site 12) (Hamilton and Brigham-Grette, 1991), and probably Cape Deceit (Figure 1, site 10) (Giterman et al., 1982). In northwest Alaska, fossil needles indicate spruce also migrated approximately 80 km beyond its current limit on the Noatak River (Figure 1, site 13; Table 2). Although treeline expansion in the north and west clearly indicates warmer than modern summer temperatures, the expansion on the Noatak

River is less clear and local factors such as permafrost distribution or disturbance may have been important.

In interior Alaska, at Eva Creek (Figure 1, site 15), the Palisades (Figure 1, site 14), and work in progress at Birch Creek indicate that the vegetation was not dramatically different from that of today. Minor differences include more abundant ferns (suggesting relatively warm, moist conditions) at some sites during MIS 5e and an extralimital aquatic (*Sagittaria latifolia*) that today grows only as far north as central British Columbia.

In the northern Yukon Territory, plant macrofossil analyses at Ch'ijee's Bluff and Hungry Creek (Figure 1, sites 16 and 17, respectively; Table 2) also indicate the presence of boreal forest similar to today's vegetation but with taxa close to or beyond their modern northern limits. Ch'ijee's Bluff, approximately 30 km south of the modern forest limit, supports a boreal forest in well-drained settings and black spruce bog in poorly drained areas. Macrofossils include abundant spruce needles, as well as several extralimital herbs and shrubs; these indicate not only the presence of a closed boreal forest but also that the forest was similar to forests growing in the southern Yukon today. Farther south, MIS 5-aged alluvium at Hungry Creek also contains herbaceous taxa near or slightly beyond their modern limits.

In the Canadian High Arctic, samples from the Collinson Formation on Banks Island (Figure 1, site 18) indicate northern migration of shrub tundra, but treeline was still south of the island. Sites from Bathurst Island (Figure 1, site 19) are

poorly dated but may date to MIS 5. These sites contain a few extralimital or rare taxa that are more common in tundra to the south.

MIS 3 (60–25 ka)

MIS 3 was a long period of climatic amelioration and reduced glacial ice volume over North America; however, the BLB remained a conspicuous part of Arctic paleogeography, at least during cool phases. Pollen assemblages and other proxy data suggest complex environmental fluctuations, with some intervals dominated by glacial herb–tundra, whereas others indicate spruce-dominated woodland or mixtures of tundra and forest (Anderson and Lozhkin, 2001).

MIS 3 sites in central Alaska include the loess/colluvial sediments of the Isabella Basin and loess and peat within the Fox Permafrost tunnel (Figure 1, sites 20 and 21, respectively; Table 2); however, chronological uncertainties make the interpretation of their paleoecological data troublesome. Although spruce needles were recovered from Isabella Basin, these may represent interglacial (MIS 5) rather than interstadial deposits (Hamilton and Brigham-Grette, 1991). Plant macrofossils from the Fox permafrost tunnel include graminoids, shrubs, aquatic herbs, and upland tundra forbs, which together with pollen and insect data suggest a mesic to dry tundra mosaic with open spruce stands in lower parts of sheltered valleys. Taken together, plant macrofossils indicate that interstadial conditions between 35 and 30 ka were less arid and less cold than during the early (MIS 4) and late Wisconsinan (MIS 2) intervals, enabling some local spruce to persist within tundra–forest vegetation.

Sites from the Stewart River, central Yukon, indicate the MIS 3 vegetation was variable. The older locality (Mayo section, ca. 38 ka) contains a rooted willow stump with spruce needles and significant amounts of spruce pollen, suggesting the climate may have been similar to modern. However, by approximately 29.6 ka, macrofossils from alluvium at the Mayo Indian Village (MIV; Figure 1, site 22) (just downstream of the Mayo section) include shrubs and diverse tundra herbs typical of well-drained sites, suggesting a tundra environment that was probably no colder than present low Arctic tundra, although it was probably drier (Figure 1 and Table 2). Although a single spruce needle at MIV was considered to be reworked, pollen data suggest there may have been some small groves of trees in sheltered localities, indicating that altitudinal treeline was depressed by at least 850 m from present.

Periglacial dune deposits of the Kittigazuit Formation (Figure 1, site 23), dated to 37–33 ka on the Tuktoyaktuk coast, contain a macrofossil assemblage including diverse dry land steppe–tundra herbs, plants of moist meadows, emergent aquatics, shrubs, and some spruce needles (Figure 1 and Table 2). This assemblage suggests aridity with saline ponds and possibly warmer than present summers. It is unclear whether the spruce needles were found *in situ*. If they are not reworked, this would indicate that scattered trees may have also been present; today the northern limit of spruce is approximately 50 km south of the site.

The available macrofossil evidence from MIS 3 suggests that, with the exception of Kittigazuit Formation, the northern

limit of the boreal forest was somewhat south of today's limit and tundra formed the most widespread vegetation type. The Kittigazuit Formation and MIV both contain *Corispermum hys-sopifolium* and other herbs more typical of dry sites, indicating at least local aridity.

MIS 2 (25–16 ka)

Due to the buildup of glacial ice in the Northern Hemisphere, worldwide sea levels were lowered by approximately 120 m 21 ka, during the Last Glacial Maximum (LGM) (Fairbanks, 1989). In North America, this resulted in the exposure of the BLB between Alaska and Russia, forming Beringia, which extends from eastern Siberia to the Mackenzie River in Canada (Hopkins et al., 1982). Climate reconstructions based on a variety of climate proxy data indicate the Northern Hemisphere cooled at least 5 °C (Ruddiman, 2002). In Beringia, extensive active dune fields, loess deposition, and, in interior Alaska, a near absence of lakes also indicate a very dry climate (Hopkins et al., 1982).

Much previous discussion on full-glacial environments in Beringia has centered on the so-called mammoth steppe or the productivity paradox (Hopkins et al., 1982). This debate has attempted to resolve how the diverse Pleistocene mammoth fauna could survive harsh full-glacial conditions at high latitudes when much paleoecological data suggest the presence of a low-productivity tundra vegetation. Most of what we know about Beringian vegetation during MIS 2 is based on fossil pollen data, indicating the vegetation was a tree- and probably shrub-less herb tundra. However, the presence of grazing large mammals suggests widespread grassland or steppe habitats to provide adequate fodder. Since the dominant pollen taxa (grass (Poaceae), sedge (Cyperaceae), and sage (*Artemisia*)) cannot be identified with greater taxonomic resolution, they have been interpreted as indicative of either steppe or tundra. Thus, the floristic composition, structure, physiognomy, productivity, and spatial distribution of Beringian vegetation cannot be recognized by pollen alone. Recent macrofossil studies from diverse settings have shed light on this discussion and suggest that Beringian vegetation is more complex than the tundra versus steppe dichotomy as postulated previously.

Macrofossils preserved in cores from the Bering Shelf (Figure 1, site 24; Table 2) document this now drowned landscape. Shallow freshwater bog plants, sedges, and pondweed dominate MIS 2- and Late Glacial-aged (ca. 21–16 ka) samples. The vegetation was clearly mesic at the coring localities, with few arid-adapted upland taxa recovered. No arboreal or shrub macrofossils were recovered, although common birch pollen suggests shrubs may have been growing nearby. Taken together, these plant macrofossils represent low-lying mesic to wet birch–graminoid tundra with interspersed ponds.

Probably the best LGM aged plant macrofossil site in Alaska is the buried Kitluk soil found under a 21-ka tephra on the Seward Peninsula (Figure 1, site 25; Table 2). This *in situ* vegetation provides a unique view of a plant community that inhabited an upland, calcareous loess soil near central Beringia during the height of the last glaciation. The plant macrofossil assemblage is dominated by graminoids, specifically the xerophilous sedge *Kobresia myosuroides*. The forb flora on the

buried surface is diverse and includes many taxa typical of dry to mesic calcareous substrates. Prostrate willows (including *Salix arctica*) are present but very rare. Taken as a whole, macrofossils on the LGM upland vegetation on the Seward Peninsula indicate a treeless vegetation mosaic dominated by closed, dry herb tundra–grassland but with some low-growing willows in moister localities.

MIS 2-aged sites from the boreal forested regions of Yukon Territory provide information on the vegetation that occupied the continental interior of eastern-most Beringia. Key sites are ice-rich loess deposits in the Klondike valleys of west-central Yukon (Figure 1, site 27) that contain valley bottom peat and fossil Arctic ground squirrel (*Spermophilus parryi*) middens (dated ca. 27 ka) and low-energy alluvial deposits at the Bluefish exposure (ca. 21–18 ka) in the northern Yukon (Figure 1, site 26; Figure 4; Table 2). Together, well-preserved remains of graminoids, diverse forbs, as well as leaves and/or flowers of

the prairie sage dominate these deposits. These floristically diverse assemblages suggest that plants that are currently common in dry to mesic Alpine tundra and azonal steppe communities formed local steppe–tundra ecotonal mosaics.

Holocene Macrofossil Records (ca. 11 ka to the Present)

The early Holocene (ca. 11–8 Ka) is marked primarily by warmer than modern summer temperatures but colder winters (COHMAP members, 1988). The vegetation reflects this climatic regime by the northward and westward advance of tree-line in Alaska and Canada (Table 2). Wood and leaves from cottonwood or aspen have been collected beyond treeline in western Alaska (ca. 50 km to the west) and on the Arctic coastal plain (ca. 100 km to the north) (Mann et al., 2002). In western Canada, the northward migration of treeline is recorded by spruce and larch needle and/or wood fragments on the

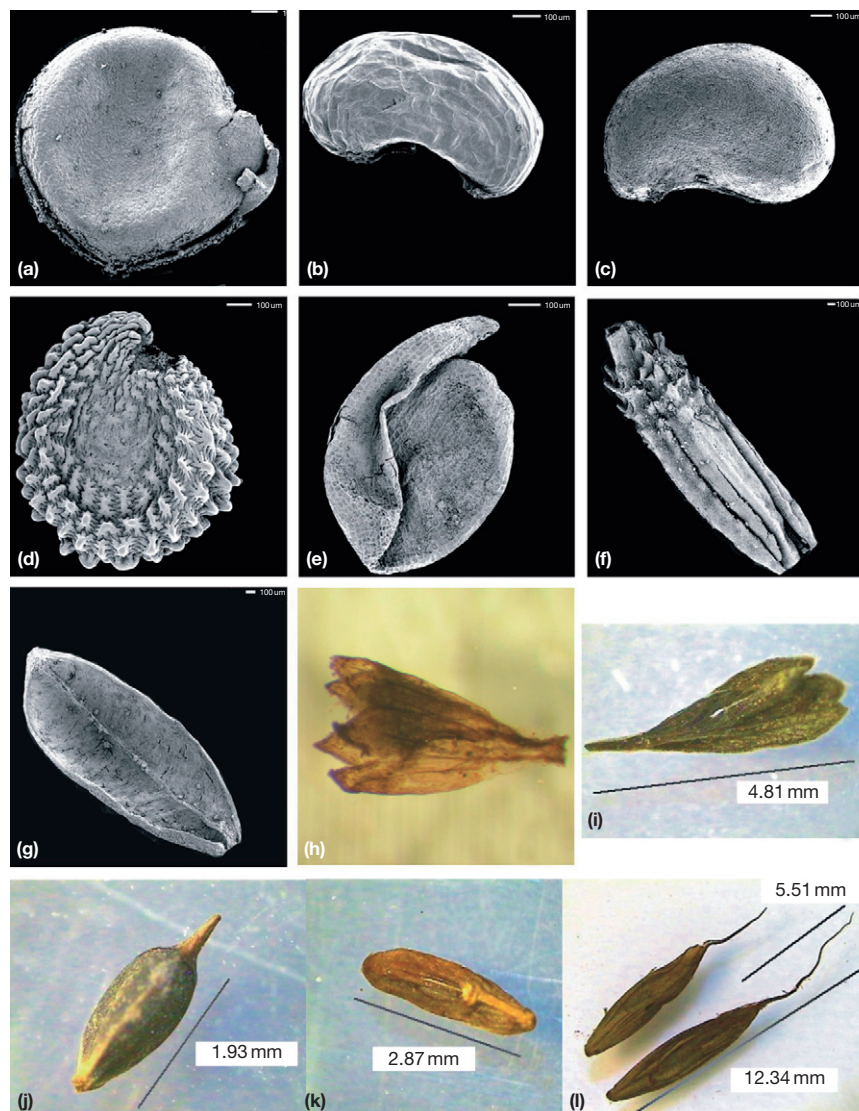


Figure 4 Selected macrofossils from MIS 2 in Yukon Territory (see [Zazula et al., 2006](#)). Scale bars in A–G are 0.1 mm. (a) *Chenopodium* sp. seed; (b) *Papaver* sp. seed; (c) *Potentilla* sp. achene; (d) *Cerastium* sp. seed; (e) *Draba* sp. seed; (f) *Taraxacum ceratophorum* achene; (g) *Phlox* cf. *hoodii* capsule fragment; (h) *Artemisia* flower, ca. 3 mm; (i) *Artemisia frigida* leaves; (j) *Kobresia myosuroides* achene; (k) *Festuca* sp. floret; (l) *Elymus trachycaulus* florets.

Tuktoyaktuk Peninsula (Figure 1, sites 29 and 30) and *Populus* wood from sites on the shores of the Arctic Ocean in the Yukon and Northwest Territories.

The Laurentide Ice Sheet persisted in eastern Canada until the mid-Holocene (Dyke et al., 2002). In northern Quebec, macrofossil records indicate that treeline has remained stable since deglaciation; however, larch needles preserved in peat do indicate a slight expansion beyond its current limit approximately 3.5–2.7 and 2.0–1.6 ka (Figure 1, site 37; Table 2) (Payette et al., 2002).

Peatland development was widespread during the Holocene (Ovenden, 1990), but the extent to which it reflects climate is still debated. Holocene-aged macrofossil records from throughout the north (Figure 1 and Table 2 sites 31, 32, 33 and 34) document site drainage and the development of a relatively rich fen, followed by a transition to a comparatively poor bog or heath vegetation. The time of fen development is dependent on when the site became ice-free. Deglaciation was earliest in the northwest (ca. 15 ka in the northern Yukon) and latest in the east (ca. 6 Ka at Hudson's Bay), so fen development was earliest in the Yukon and western NWT (ca. 12.5–10 Ka) and later in central Nunavut (ca. 8.0 Ka) and at Hudson's Bay (ca. 5.6 Ka).

The transition to bog or heath reflects changes in the local hydrology at the sites. The primary drivers of these changes is either permafrost expansion or autochthonous peat development that may be independent of permafrost. Permafrost expansion can lead to either drying and a heath vegetation (due to the development of high-centered polygons) or wetting and a bog vegetation (due to the development of low-centered polygons). Autochthonous peat development may also dry a site by raising the vegetation above the water table, resulting in a heath vegetation.

Holocene-aged peat deposits in the High Arctic may provide some information on regional climates (Figure 1 and Table 2, sites 35 and 36). Dates of peat development in areas that no longer support thick peat deposits may provide evidence for moister climate in the past (Garneau, 2000); however, local topography and hydrology may be more important factors (La Farge-England et al., 1991).

Discussion

Range Extensions

Much of the macrofossil data presented here document range extensions, especially treeline shifts. Plant macrofossils are a particularly robust proxy for documenting treeline because (1) species-level identification is possible and (2) compared to pollen, macrofossils travel comparatively short distances. At sites where larch is the dominant conifer, larch needles are more easily recognized than larch pollen, which is fragile and nondescript. Another example is *Populus*; *Populus* pollen does not travel far from its source and is easily degraded so that although Alaskan pollen records indeed indicate expansion of *Populus* during the early Holocene (Anderson et al., 2004), it is the macrofossil evidence that clearly shows a range extension.

Much of the modern treeline is defined by the northern limit of spruce and pine. Pollen from both of these taxa travel widely so that 10% spruce and, in places, 30% pine pollen

mark the northern limit of trees (Anderson et al., 1991). Macrofossils (wood, needles, and stomata) provide excellent evidence that the tree was actually growing at the site (Payette et al., 2002).

Knowing the location of treeline in the past is critical for reconstructing the climate and ecology of the region. The modern treeline is influenced by the thermal regime, usually characterized as mean July temperature or accumulated radiation (Elliott-Fisk, 2000). During the late Pliocene, pines grew on Meighen Island, implying at least an 8 °C increase in summer temperatures. Pleistocene interglacial peat deposits also indicate northward expansion of treeline, although not as extreme as during the late Pliocene.

MIS 5 is of particular importance because it is the most recent interglaciation that we can compare with our own. Range extensions of spruce trees near the Arctic Ocean, boreal forest herbs in the Yukon, and tundra shrubs in the High Arctic (if the Bathurst Island site dates to MIS 5) indicate warmer than modern temperatures. However, compared with MIS 5-aged sites elsewhere in the Arctic (LIGA, 1991; CAPE Last Interglacial project members, 2006), the amount of summer warming across northern North America was modest—probably on the order of 1° or 2 °C.

Vegetation Reconstructions

In addition to documenting the local presence of a taxon, because of their high taxonomic resolution, macrofossils provide more exact vegetation reconstruction than is possible with pollen analysis alone. This is especially important when analyzing the Beringian full-glacial vegetation of MIS 2.

The dominance of grass, sedge, and diverse tundra and steppe forbs at MIS 2-aged macrofossil sites in the Yukon suggests plants that are currently common on dry to mesic alpine tundra and azonal steppe communities formed local steppe–tundra ecotonal mosaics. The local composition and distribution of this vegetation was closely related to landscape position, aspect, available moisture, and disturbance (Figure 5). The presence of *Artemisia frigida* leaves is significant because this suggests at least some of the *Artemisia* pollen from MIS 2-aged sites derives from this dry-adapted taxon.

The detail afforded by plant macrofossils has helped establish a picture of substantial floristic diversity during the LGM, at both the local and the regional scale. Tundra herbs were more abundant in some areas, whereas steppe plants were common in others. Vegetation in Beringia during MIS 2 cannot be easily classified as either steppe or tundra because floristic analogs can be drawn with several vegetation types. These studies demonstrate how plant macrofossil-based vegetation reconstructions can resolve local-scale floristic composition that cannot be achieved by pollen analysis alone.

Conclusion

Plant macrofossil records provide important details on climate and vegetation. The exact placement of treeline during different times in the past and the detailed full-glacial vegetation reconstruction are two examples. However, because of biases in deposition and preservation, macrofossil records should also

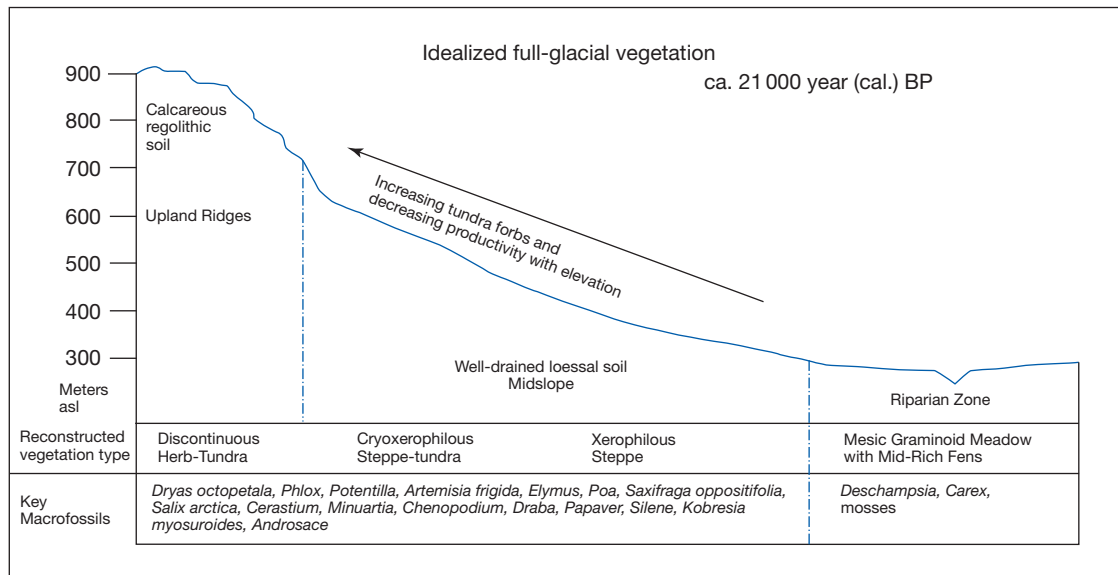


Figure 5 Idealized vegetation–topography sequence for continental eastern Beringia during MIS 2, relating the effects of altitude, drainage, and substrate on vegetation composition. Adapted from Zazula GD, Schweger CE, Beaudoin AB, and McCourt GH (2006) Macrofossil and pollen evidence for full-glacial steppe within an ecological mosaic along the Bluefish River, eastern Beringia. *Quaternary International* 142–143: 2–16.

be interpreted conservatively and modern macrofossil taphonomy studies should be a priority. The absence of a particular macrofossil at a particular site does not mean the taxon is absent from the landscape, highlighting the importance of paired pollen and macrofossil analyses for well-rounded paleoecological reconstructions (Birks, 2001).

See also: **Paleoclimate:** Introduction. **Plant Macrofossil Records:** Holocene North America. **Pollen Records, Late Pleistocene:** Northern North America. **Sea Level Studies:** Coral Records of Relative Sea-Level Changes; Eustatic Sea-Level Changes – Glacial–Interglacial Cycles; Eustatic Sea-Level Changes Since the Last Glacial Maximum. **Paleoclimate Modeling:** Last Glacial Maximum GCMs.

References

- Anderson PM and Lozhkin AV (2001) The stage 3 interstadial complex (Karginskii/middle Wisconsinan interval) of Beringia: Variations in paleoenvironments and implications for paleoclimatic interpretations. *Quaternary Science Reviews* 20: 93–126.
- Anderson PM, Bartlein PJ, Brubaker LB, Gajewski K, and Ritchie JC (1991) Vegetation–pollen–climate relationships for the arcto-boreal region of North America and Greenland. *Journal of Biogeography* 18: 565–582.
- Anderson PM, Edwards ME, and Brubaker LB (2004) Results and paleoclimate implications of 35 years of paleoecological research in Alaska. In: Gillespie AR and Porter SC (eds.) *The Quaternary Period in the United States*, pp. 427–440. Amsterdam: Elsevier.
- Birks HH (2001) Plant macrofossils. In: Smol JP, Birks HJB, and Last WM (eds.) *Tracking Environmental Change Using Lake Sediments, Vol. 3: Terrestrial, Algal, and Siliceous Indicators*. Dordrecht, The Netherlands: Kluwer.
- CAPE Last Interglacial Project Members (2006) Last interglacial Arctic warmth confirms polar amplification of climate change. *Quaternary Science Reviews* <http://dx.doi.org/10.1016/J.quascirev.2006.01.033>.
- CAVM (2003) *Circumpolar Arctic Vegetation Map. Conservation of Arctic Flora and Fauna Map No. 1*. AK: U.S. Fish and Wildlife Service, Anchorage.
- Cody WJ (1996) *Flora of the Yukon Territory*. Ottawa, Ontario, Canada: NRC Research Press.
- COHMAP members (1988) Climatic changes of the last 18,000 years: Observations and model simulations. *Science* 241: 1043–1052.
- Dyke AS, Andrews JT, Clark PU, et al. (2002) The Laurentide and Innuitian ice sheets during the Last Glacial Maximum. *Quaternary Science Reviews* 21: 9–31.
- Elliott-Fisk DL (2000) The taiga and boreal forest. In: Barbour MG and Billings WD (eds.) *North American Terrestrial Vegetation*, 2nd edn., pp. 41–73. Cambridge, UK: Cambridge University Press.
- Energy, Mines, and Resources Canada (1993) *National Atlas of Canada*, 5th edn. Energy, Mines, and Resources Canada, Ottawa, Ontario, Canada.
- Esat TM, McCulloch MT, Chappell J, Pillans B, and Omura A (1999) Rapid fluctuations in sea level recorded at Huon Peninsula during the penultimate deglaciation. *Science* 283: 197–201.
- Fairbanks RG (1989) A 17,000-year glacio-eustatic sea level record: Influence of glacial melting rates on the Younger Dryas event and deep-ocean circulation. *Nature* 342: 637–642.
- Flora of North America Association editorial committee (1993) *Flora of North America North of Mexico*. Available at <http://hua.huh.harvard.edu/FNA/>.
- Garneau M (2000) Peat accumulation and climatic change in the High Arctic. In: Garneau M and Alt BT (eds.) *Environmental Response to Climatic Change in the Canadian High Arctic*. Ottawa, Ontario, Canada: Geological Survey of Canada. GSC Bulletin No. 529.
- Giterman RE, Sher AV, and Matthews JV Jr. (1982) Comparison of the development of steppe–tundra environments in west and east Beringia: Pollen and macrofossil evidence from key sections. In: Hopkins DM, Matthews JV, Schweger CE, and Young SB (eds.) *Paleoecology of Beringia*, pp. 43–74. New York: Academic Press.
- Hamilton TD and Brigham-Grette J (1991) The last interglaciation in Alaska stratigraphy and paleoecology of potential sites. *Quaternary International* 10–12: 49–71.
- Hopkins DM, Matthews JV, Schweger CE, and Young SB (eds.) (1982) *Paleoecology of Beringia*. New York: Academic Press.
- Hultén E (1968) *Flora of Alaska and Neighboring Territories*. Stanford, CA: Stanford University Press.
- Koerner RM (1989) Ice core evidence for extensive melting of the Greenland icesheet in the Last Interglacial. *Science* 244: 964–968.
- La Farge-England C, Vitt DH, and England J (1991) Holocene soligenous fens on a High Arctic fault block, northern Ellesmere Island (82° N), N.W.T., Canada. *Arctic and Alpine Research* 23: 80–98.
- LIGA (1991) Report of 1st discussion group: the last interglacial in high latitudes of the Northern Hemisphere terrestrial and marine evidence. *Quaternary International* 10–12: 9–28.

- Mann DH, Peteet DM, Reanier RE, and Kunz ML (2002) Responses of an arctic landscape to Lateglacial and early Holocene climatic changes: The importance of moisture. *Quaternary Science Reviews* 21: 997–1021.
- Payette S, Eronene M, and Jasinski JJP (2002) The circumboreal tundra–taiga interface: Late Pleistocene and Holocene changes. Ambio Special Report. *Tundra–Taiga Treeline Research* 12: 15–22.
- Petit JR, Jouzel J, Raynaud D, et al. (1999) Climate and atmospheric history of the past 420,000 years from the Vostok ice core, Antarctica. *Nature* 399: 429–436.
- Ovenden L (1990) Peat accumulation in northern wetlands. *Quaternary Research* 33: 377–386.
- Ruddiman WF (2002) *Earth's Climate: Past and Future*. New York: Freeman.
- Viereck LA and Little EL (1972) *Alaska Trees and Shrubs*. Washington, DC: U.S. Department of Agriculture.
- Zazula GD, Schweger CE, Beaudoin AB, and McCourt GH (2006) Macrofossil and pollen evidence for full-glacial steppe within an ecological mosaic along the Bluefish River, eastern Beringia. *Quaternary International* 142–143: 2–16.

Greenland

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Introduction

The Greenland subcontinent is the largest island on Earth, covering an area of 2 166 086 km². The ice-free land between the central Greenland ice sheet (the Inland Ice) and the surrounding waters is up to 250 km broad, and covers 410 449 km². Numerous local ice caps and other types of glaciers are found, as well as large fjords (Figure 1).

Major parts of ice-free Greenland are mountainous, with Alpine topography or plateau landscapes dissected by valleys. Many parts of the landscape show strong influence from glacial erosion, with rounded mountains. But in other areas hardly any sign of glacial erosion is seen between the valleys or fjords.

The deep ice cores from the Inland Ice have aroused a large interest among Quaternary geologists and paleoclimatologists, because they provide a wealth of detailed information about the climatic development in the North Atlantic region during the last Ice Age and the Holocene. However, these investigations provide little information about the biotic reactions to the climatic changes.

The ice-free parts of Greenland only give glimpses of the pre-Holocene vegetation history. Due to the repeated glaciations many deposits have been removed by eroding glaciers and meltwater rivers. In spite of this, the pre-Holocene deposits may give a fairly detailed picture of plant life. Holocene deposits are much more widespread and detailed studies of changes of vegetation and flora history have been conducted.

A review on the vegetation history of West Greenland was provided by Fredskild (1985a) and a review of the Quaternary history of the ice-free parts of Greenland was published by Funder (1989). Analyses of the Holocene vegetation history were started by J. Iversen and were continued primarily by Fredskild (1967, 1973, 1978, 1983, 1985b, 1995, 2000), Funder (1978), Bennike (1999, 2000), and Bennike et al. (1999). Pre-Holocene floras have been reported by Bennike (1990), Bennike and Böcher (1992, 1994); and Björck et al. (2002). In addition to pollen analysis, plant macrofossil analyses have also been conducted. Studies of macroscopic plant remains have been especially important for reconstruction of pre-Holocene and early Holocene vegetation, mainly because pollen is poorly preserved and because plant macrofossils can often be identified to species, whereas pollen can usually only be identified to genus or family.

Present-Day Climate

Greenland spans 24° of latitude, corresponding to 2675 km, which means that marked climatic differences are seen from south to north. The mean summer temperature shows a fall from 11 °C in the south to 2–3 °C in the far north, and the mean winter temperature falls from –4 °C to –30 °C. The precipitation also decreases toward the north, from about

2500 mm yr^{–1} in the far south to below 200 mm yr^{–1} in parts of northern Greenland. In addition, a strong climatic gradient exists from an oceanic type of climate at the outer coast to a more continental climate near the margin of the Inland Ice.

In the central parts of North Greenland, the fjords are covered by land-fast sea-ice all year round, but farther south a period of open water occurs, and the harbors in the southwestern part of Greenland are navigable throughout the year. West Greenland is strongly influenced by the relatively warm, northward flowing West Greenland Current, whereas East and South Greenland are influenced by the cold southward flowing East Greenland Current.

Modern Flora and Vegetation

Considering the size of the ice-free land areas in Greenland, the present species diversity is relatively small. This is primarily due to the northern geographical location and the low temperatures, but it also reflects the rather isolated position of Greenland. The flora comprises ca. 480 species of seed plants, 35 species of pteridophytes (ferns and allied), ca. 600 species of bryophytes (mosses), and ca. 950 species of lichens. Thirty-two woody plant species are found.

Shrubs and heaths are widespread in the lowlands of Greenland, and furthest south, at 60° N, small areas with birch forest are found. In South Greenland, some farming takes place. Furthest to the north, at 83° N, areas of polar desert, without any woody plants, are found. Fell-fields with scattered plants are found on windy places where the snow is blown away during the winter. Herb slopes, snow beds and mires are also widespread.

Terminology and Dating

The terms early, mid-, and late Holocene are used informally, and Late Glacial is used for the time span between 14 and 11.7 ka BP, corresponding to Greenland interstadial 1 and Greenland stadial 1. The last Ice Age corresponds to marine isotope stages 2–5d, and the last interglacial stage to marine isotope stage 5e. Pre-Holocene deposits have mostly been dated by paleomagnetism, luminescence dating, amino acid analyses, and strontium isotope analyses. Younger deposits have been dated by radiocarbon dating, and dates have been calibrated to calendar years BP, which are used here.

Chronology of the Last Deglaciation

During recent decades a lot of data have been gathered that throw light on the chronology of the deglaciation of the ice-free land parts of Greenland. On this basis a chronology for the last recession of the Inland Ice was proposed by Bennike and Björck (2002). At the outer coast the oldest dates are between

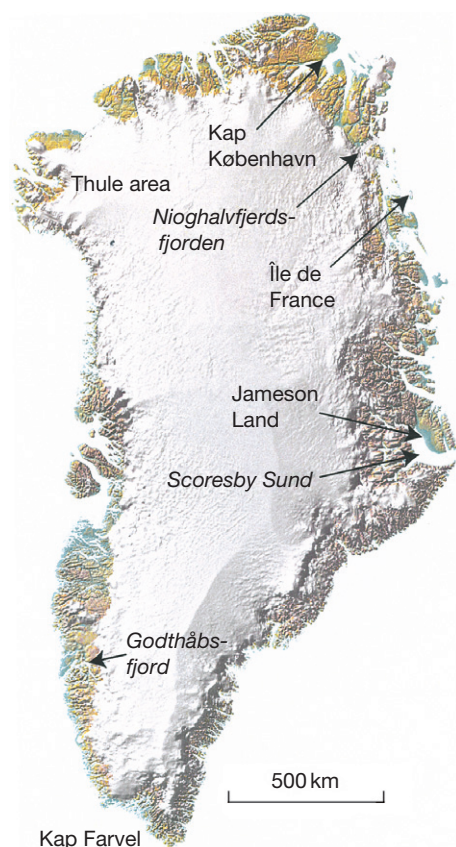


Figure 1 Map of Greenland showing the location of place names mentioned in the text.

11.5 and 10 ka whereas areas near the margin of the Inland Ice were deglaciated between 8000 and 6000 ka. In the far south, the oldest sediments recovered date to 14.1 ka, and the interior of the fjords were deglaciated in the early Holocene.

Late Glacial and Holocene Climate Changes

As mentioned above, on-shore Late Glacial deposits have only been located in South Greenland. Analyses of one sequence indicate that the Younger Dryas was relatively mild in this region (Björck et al., 2002). The abrupt warming at the Younger Dryas to Holocene boundary was followed by a more gradual warming, which was interrupted by short cooling events in the early Holocene. The Holocene is characterized by a marked thermal maximum, during which many plants extended their northern geographical range limit northwards. The Holocene thermal maximum is broadly dated to between 8 and 6 ka, preceding cooling that culminated in the Little Ice Age.

At many sites in Greenland evidence is found that both the Inland Ice and local glaciers were smaller than at present during the mid-Holocene. The best-dated glacial history comes from Nioghalvfjærdsfjorden in northeastern Greenland. The fjord is, at present, covered by a 60 km long floating outlet glacier from the Inland Ice, but the fjord was ice free during a continuous period from 7.8 to 4.5 ka BP (Bennike and Weidick, 2001).

Signs of mid-Holocene declining temperatures are also seen in many pollen diagrams from Greenland. For example,

decreasing *Betula nana* (dwarf birch) pollen percentage values in East Greenland at around 5.6 ka BP and decreasing *Juniperus communis* (juniper) pollen percentage values in southwest Greenland around 4.5 ka BP reflect falling summer temperatures (Funder, 1978; Fredskild, 1983).

The Kap København Formation

Cooling during the Tertiary led to buildup of ice sheets in the polar regions, first in Antarctica, and later in the Arctic. Deposits in the circum-Arctic region show that Pliocene temperatures were relatively high. Rare plant macrofossils on Île de France in NE Greenland give a glimpse of the vegetation at around 3.4 Ma (Bennike et al., 2002). The material comprises remains of *Picea* cf. *mariana* (black spruce) and *Thuja* sp. (cedar). These taxa indicate that the land was covered by coniferous forest or forest-tundra, with a mean temperature for the warmest month of the year over 10 °C, corresponding to at least 6 °C more than at present. Occurrences of old wood at other sites in northern Greenland are also referred to the Pliocene. Such wood is particularly common in Washington land, where *Pinus* (pine) wood is common.

The Kap København Formation is a sequence of deltaic, fluvial and marine clay, silt, and sand deposits in eastern North Greenland. The formation contains a wealth of well-preserved remains of nonmarine plants and animals, with many different groups represented. The most diverse group is beetles, of which at least 210 species are present, a surprising number when compared with the modern-day beetle fauna of Greenland that comprises around 40 species (Böcher, 1995). Vascular plants from the Kap København Formation include a mixture of boreal and Arctic species (Bennike (1990); Figure 2). Taxa such as *Larix groenlandii* (Greenland larch), *Picea mariana*, *Thuja occidentalis* (northern white-cedar), *Taxus* sp. (yew), and *Betula* sect. *Albae*

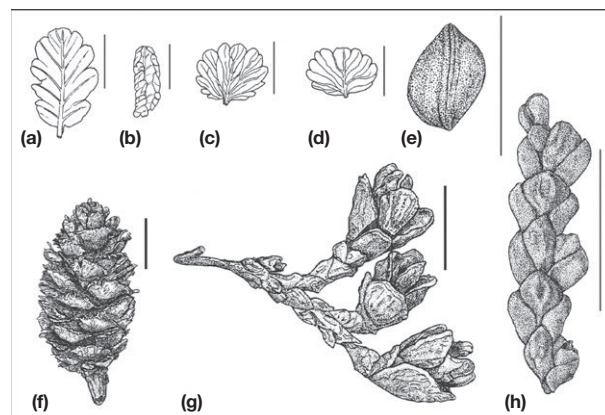


Figure 2 Drawings of plant remains from the Kap København Formation. (a), (b): Leaves of *Dryas octopetala* (mountain avens). (a) Shows a leaf that comes from a plant that grew on a protected site; (b) shows a leaf with re-curved leaf margin, which comes from a plant that grew on a wind exposed site. (c), (d): Leaves of *Betula nana* (dwarf birch). (e) Fruit stone of *Cornus stolonifera* (red-osier dogwood). (f) Cone of *Larix groenlandii* (Greenland larch). (g) Twig of *Thuja occidentalis* (northern white-cedar) with three small cones. (h) Twig of *Thuja occidentalis* with scaly leaves. Thick scale bars: 10 mm, thin scale bars: 5 mm.

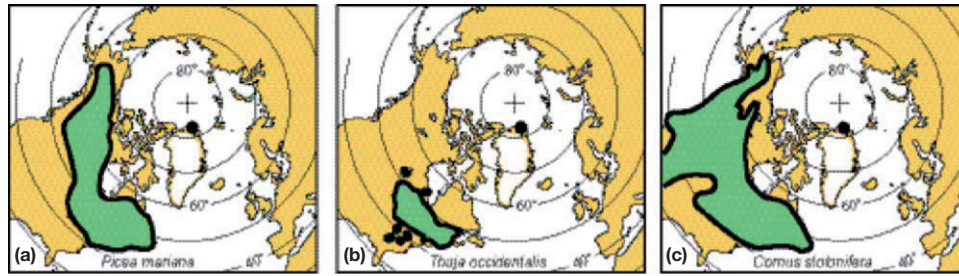


Figure 3 Map of the northern parts of the Earth, showing the present geographical range of (a) *Picea mariana* (black spruce); (b) *Thuja occidentalis* (northern white-cedar); and (c) *Cornus stolonifera* (red-osier dogwood). Remains of these species occur in the Kap København Formation, the location of which is shown by a black dot in northern Greenland.

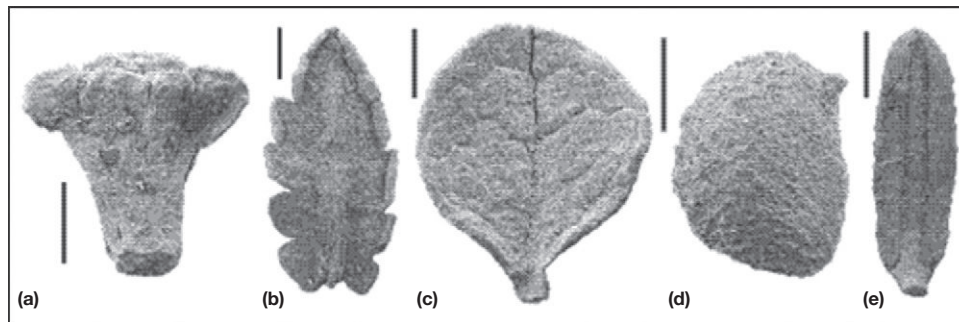


Figure 4 Scanning electron microscope photographs of plant remains from the last Interglacial. (a) Catkin scale of *Alnus* cf. *crispa* (green alder); (b) Leaf of *Dryas octopetala* (mountain avens); (c) Leaf of *Vaccinium uliginosum* (Arctic blueberry); (d) Endocarp of *Arctostaphylos alpina* (Alpine bearberry); (e) Leaf of *Phyllodoce coerulea* (mountain-heath). The scale bars correspond to 1 mm.

(tree birch) belong to the first group (Figure 3), whereas *Dryas octopetala* (mountain avens) and *Oxyria digyna* (mountain sorrel) belong to the second. All remains of wood come from small trees or shrubs, and the maximal trunk diameter is 18 cm. Growth rings are narrow to extremely narrow, which together with the composition of the flora, indicate that the mean temperature for the warmest month of the year was 10–11 °C. This corresponds to 7 °C higher than at present. *Thuja* and *Taxus* cannot tolerate very cold winters, and the mean temperature for the coldest month was probably not below –17 °C. It appears that the area was dominated by forest-tundra and an oceanic type of sub-Arctic climate.

The dating of the Kap København Formation is based on a number of different methods, of which the most important are biostratigraphy, paleomagnetic analyses and amino acid analyses. The biostratigraphically most important groups are foraminifers, marine ostracodes, marine mollusks, and mammals. The occurrence of the extinct rabbit *Hypolagus* sp. and the extant hare *Lepus* sp. have been particularly important. These genera co-occurred during the time period from ca. 2.3 to 2.0 Ma. A correlation with marine isotope stages 91–101, corresponding to 2.35–2.52 Ma, has been proposed (Funder et al., 2001).

The Last Interglacial

A rather large number of mid-Pleistocene marine interglacial occurrences have been located in Greenland, but these do not provide information about the vegetation. In contrast, layers from the last interglacial are found on many localities along

the south and west coast of Jameson Land in East Greenland. The deposits are mainly sand, silt, and clay, which were deposited in shallow water along the coast. At many sites, layers and lenses rich in organic detritus are found. This detritus consists of washed-together remains of plants that grew in the catchment area. The detritus also contains remains of insects and other invertebrates. These remains give a regional picture of the biotas.

The flora includes remains of many species of other woody plants (Figure 4). Most noteworthy are common remains of the tree birch *Betula pubescens* (downy birch), which probably formed copses or low forest on sheltered sites. At present, tree birch is confined to southernmost Greenland, whereas *Betula nana*, common today in Jameson Land, has not been found in the interglacial deposits. On moist localities, alder was growing, probably *Alnus crispa* (green alder), which grows in southwestern Greenland and North America at present. Another woody plant found in interglacial deposits that still grows in southernmost Greenland is *Cornus canadensis* (creeping dogwood) (Figure 5). Dwarf shrubs are represented by no less than 11 species, of which the majority are ericaceous. Fruits and leaves of *Empetrum nigrum* (crowberry) are especially common. Leaves of the snow-bed species *Salix herbacea* (herb-like willow) are also common.

The dwarf shrubs are characterized by Arctic species such as *Dryas octopetala*, *Cassiope tetragona* (white Arctic bell-heather), and *Arctostaphylos alpina* (Alpine bearberry). The fossil assemblages are also rich in remains of herbs and bryophytes. The bryophyte flora with more than 60 identified taxa also includes some southern extralimital species.

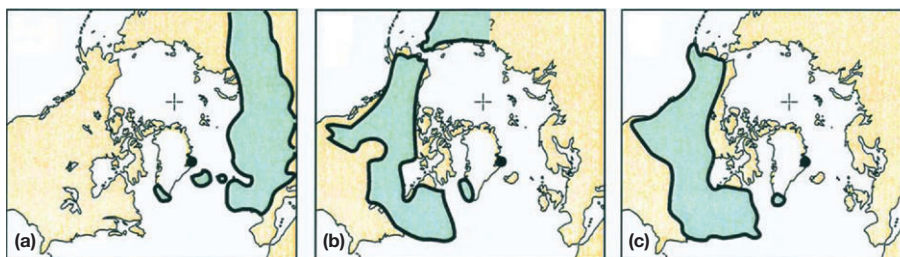


Figure 5 Map over the northern parts of the Earth, showing the present day geographical ranges of (a) *Betula pubescens* (downy birch); (b) *Alnus crispa* (green alder); and (c) *Cornus canadensis* (creeping dogwood). Remains of these species occur in layers from the last interglacial stage in Jameson Land (black dots).

One of the most warmth demanding plant species from the interglacial layers is *Betula pubescens*, which indicates a mean temperature for the warmest month of at least 10–11 °C, corresponding to 5 °C higher than at present. The deposits from Jameson Land have been dated by numerous luminescence dates, which place the deposits in the last interglacial.

Marine deposits from the last interglacial are also fairly widespread in the Thule area in northwest Greenland. Remains of nonmarine plants and animals have been located at a single coastal cliff section. Some of the taxa represent southern extralimital occurrences. These comprise the bryophytes *Pleurozium schreberi* and *Conardia compacta* and the vascular plants *Selaginella selaginoides* (lesser clubmoss), *Menyanthes trifoliata* (buckbean), *Betula* sect. *Nanae* (dwarf birch), and *Gentiana detonsa* (serrate gentian) (Bennike and Böcher, 1992; Hedenäs and Bennike, 2003). The mean temperature for the warmest month of the year was at least 4 °C higher than at present. The layers at Thule are also primarily dated by means of luminescence dating.

In northern Greenland, two sites with plant remains have yielded nonfinite radiocarbon dates. The flora indicates summer temperatures similar to those found in the region at present, which may imply that the deposits belong to an interglacial rather than to an interstadial.

The Last Ice Age

Hugin Sø Interstade

In Jameson Land, the Hugin Sø Interstade, which is correlated with marine isotope stage 5c, represents an ice-free period. The vegetation appears to have consisted of herbs and bryophytes only. Remains of *Saxifraga oppositifolia* (purple saxifrage) are common, and seeds of *Papaver* sect. *Scapiflora* (arctic poppy) also occur. These species are found all over Greenland at present, but they play an increasing role in the vegetation toward the north, and none of them have been found in deposits from the last interglacial. It has been proposed that the mean temperature for the warmest month of the year was 3–4 °C lower than at present (Bennike and Böcher, 1994).

The Last Glacial Maximum

In central Greenland, the temperature minimum during the last Ice Age is dated to about 22 ka BP, and the mean annual temperature was about 25 °C lower than at present. If these

results can be transferred to the marginal parts of Greenland, only the hardest plants could have survived the last Ice Age. The extent of the Inland Ice was far larger than at present during the Last Glacial Maximum and large parts of the extensive shelf areas around Greenland were also covered by the ice sheet.

The Late Glacial

Late Glacial sediments from two localities in southern Greenland have been analyzed for pollen and plant macrofossils (Björck et al., 2002). The concentration of pollen is extremely low, and a large part must be regarded as long-distance transported. Pollen of Poaceae (grasses), *Sagina* type (pearlwort), and *Saxifraga* (saxifrage) presumably come from the local vegetation, and the same applies to a few seeds of *Minuartia* sp. (sandwort), *Saxifraga* cf. *oppositifolia* and Poaceae (grasses). The unstable soils were characterized by a pioneer vegetation with mosses and scattered herbs.

Holocene Vegetation Development

The vegetation development after the last deglaciation in the different parts of Greenland has obviously been different, but nevertheless it can be divided into three main phases: a pioneer phase, a phase of optimum productivity, and a recessional phase. The development reflects direct and indirect climatic forcing, and problems in colonizing the country. The vegetation development has mainly been constructed from regional pollen diagrams from lake deposits. The most important pollen diagrams from South and West Greenland have been published by Fredskild (1973, 1983, 1985a, b). From East Greenland, pollen diagrams have been published by Funder (1978). In the northern parts of Greenland it has proved difficult to find lakes with continuous sequences, and the extremely low pollen production also leads to problems. From these parts of Greenland only a few well-dated pollen diagrams have been published. In West Greenland, some diagrams extend back to the earliest Holocene, whereas none of the diagrams from East and North Greenland covers the first one or two millennia of the Holocene. In addition to pollen analyses, macroscopic plant remains have also been analyzed at many sites, partly to investigate early Holocene floras (Bennike et al., 1999).

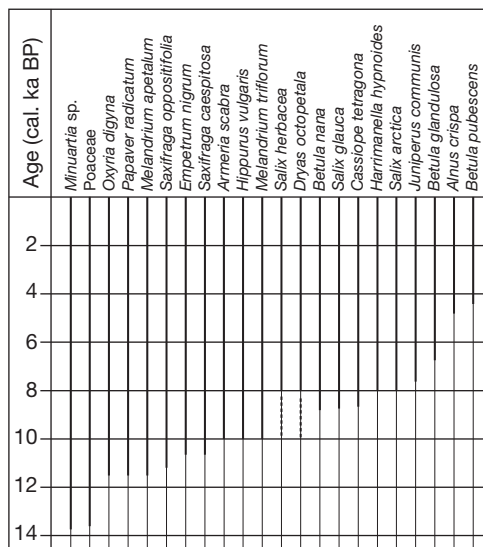


Figure 6 Summary diagram showing the known temporal range of different taxa in Greenland after the last deglaciation (Holocene and Late Glacial).

The pioneer phase is characterized by bryophytes, herbs and grasses, by species that can tolerate low summer temperatures and unstable soils (Fredskild, 1985a; Bennike, 1999; Bennike et al., 1999). In southern Greenland, pollen of *Oxyria digyna* can account for up to 75% of the pollen sum. In eastern Greenland, macrofossils of *Papaver radiculatum* (arctic poppy), *Oxyria digyna*, and *Melandrium apetalum* (nodding lychnis) are present in earliest Holocene deposits (Figure 6). The pioneer phase may have extended over several millennia in the peripheral parts of Greenland, which were deglaciated first. In the inland, the pioneer phase was of shorter duration, because most of Greenland's plant species had immigrated when these parts became ice free, so that a more diverse flora could rapidly colonize these areas.

The first woody plant that appears to have immigrated to Greenland is *Empetrum nigrum*, which was present from ca. 11 ka BP. It was soon followed by *Salix herbacea*, which is known from 10.8 ka BP, and *Vaccinium uliginosum* (Arctic blueberry), which is known from 10.5 ka BP. Other ericaceous dwarf shrub species were also early immigrants.

After a transitional stage a productivity optimum is seen. Although regional differences are evident, large parts of Greenland's lowlands became covered by dense dwarf shrub heaths. This development reflects soil stabilization as species immigrated and summer temperatures increased. In the far south, birch copses expanded, but in the north, fell-field vegetation prevailed. *Betula nana*, *Salix* spp. (willow), and a number of ericaceous plants such as *Empetrum nigrum*, *Vaccinium uliginosum*, and *Cassiope tetragona* came to dominate the heaths.

Gradually decreasing summer temperatures in the mid- and late Holocene led to decreasing pollen production and more open heaths. The northern range limit of some species, such as *Empetrum nigrum*, *Armeria scabra* (thrift), and *Potamogeton filiformis* (slender-leaved pondweed) moved south.

Figure 7 shows a summary diagram from southernmost Greenland as an example of a Greenland pollen succession

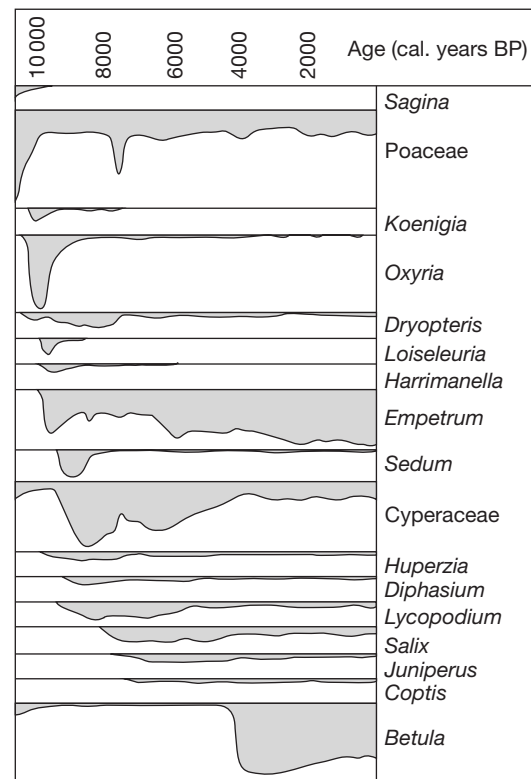


Figure 7 Summary pollen percentage diagram for the Holocene, based on several lakes in the Kap Farvel area in southern Greenland.

(Fredskild, 1973). The pioneer vegetation was dominated by grasses, *Saxifraga*, *Cerastium* (chickweed), and *Sagina*, followed by peaks of *Koenigia islandica* (common koenigia) and *Oxyria digyna*. Different dwarf shrubs, such as *Loiseleuria decumbens* (Alpine azalea), *Harimanella hypnoides* (moss heather), and *Empetrum* were early immigrants. *Angelica archangelica* (angelica) was also an early immigrant, whereas *Salix*, *Juniperus communis*, and *Betula* arrived later. The short-lived marked peak of grass pollen may be synchronous with the 8200 event (Fredskild, 1973). It has not been detected in other Greenland pollen diagrams, but the resolution of most diagrams is too low to allow short events to be detected. Declining *Juniperus* and *Betula* values mark declining summer temperatures in the late Holocene.

Another example of a Greenland Holocene pollen succession is provided by a simplified diagram from the inner part of the Godthåbsfjord in SW Greenland (Figure 8; Fredskild (1983)). The pioneer vegetation was rich in grasses, sedges, and *Oxyria digyna*. *Empetrum* is also here present at an early stage, whereas *Salix glauca* (northern willow) did not immigrate until about 8.7 ka BP. Around 7.4 ka BP *Betula nana* arrived. At about the same time *Juniperus communis* immigrated; the latter species shows a marked fall in the values for pollen percentage and pollen influx at around 4.4 ka. It is assumed that these declines are due to colder and moister summers. The last immigrant is *Alnus crispa*, which expanded rapidly, but shows a gradual decline shortly after its immigration, caused by declining summer temperatures.

In the inner parts of the Scoresby Sund region, a marked rise is seen in the *Betula nana* curve just after the species arrived at about 8.8 ka BP (Funder, 1978). It appears that dwarf birch

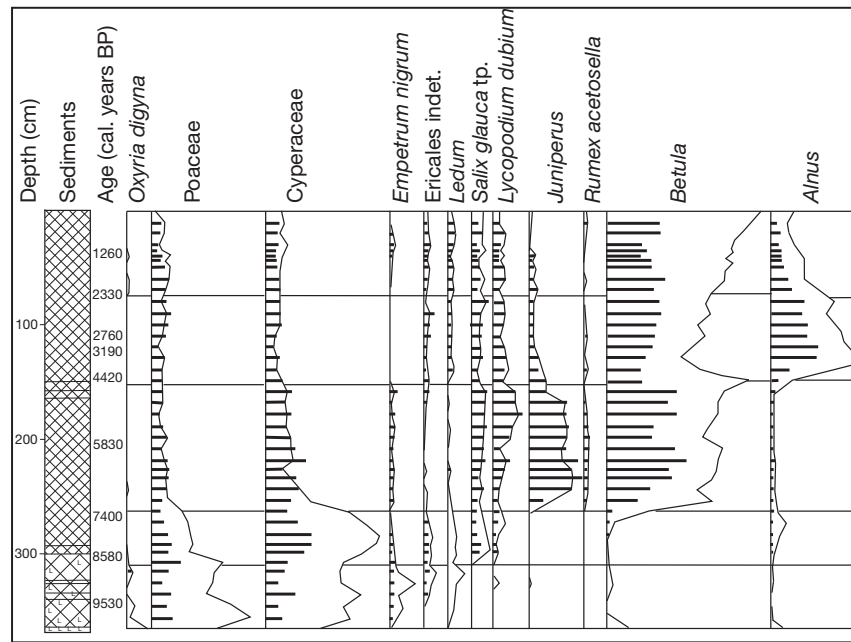


Figure 8 Simplified pollen diagram from Johs. Iversen Sø in the inner part of the Godthåbsfjord in southwestern Greenland. The curves show percentage values, and the bars show influx values. The sediments consist of clay in the bottom, followed by clay-gyttja, and then by ca. 3 m gyttja. Modified from Fredskild (1983).

met with little competition after it arrived. Rich dwarf shrub heaths gave way to more open heaths or fell-field vegetation during the mid- to late Holocene.

The first humans arrived in Greenland 4.5 ka BP, but these Paleo-Eskimos hardly influenced the regional vegetation. However, technological development allowed the first Europeans to colonize Greenland about 1 ka BP. The Norse settled in south and southwest Greenland, where they founded the Eastern and the Western Settlements, with around 300 farms and 20 churches. The economy was mainly based on husbandry, especially cattle, sheep and goats, but also horses and pigs. Distinct changes of the vegetation took place following the landnam (land clearance) (Fredskild, 1988). The Norsemen introduced a range of species, and *Rumex acetosella* (sheep sorrel) and *Rumex acetosa* (green sorrel) spread. *Betula* (especially tree birch) and *Salix* declined, whereas herbs expanded, reflecting widespread cutting of forest and shrubs, followed by grazing. In peat sections in the Western Settlement a layer of charcoal has been observed. This was deposited when the settlers burned the copses after the landnam. Farming and grazing led to increased soil erosion, and sand and clay were washed out into lakes. Overgrazing may have been one of the factors that led to the demise of the Norse between AD 1350 and 1400.

Development of Lake Vegetation

There are numerous lakes in Greenland, with large variation in extent, depth, and water chemistry. Most lakes are ultra-oligotrophic, but subsaline lakes occur locally (Bennike, 2000). In addition to high conductivity, the subsaline lakes are characterized by high alkalinity. These lakes are closed basins without outlets, so that salts that are transported into the lakes from the catchment become concentrated when the water evaporates.

Many lakes have become increasingly oligotrophic and acidic after they formed (Fredskild, 1992). This is presumably most pronounced in areas with acid rocks, such as gneiss and granite. Remains of water plants that prefer high conductivity are found in abundance in the first phase of the lake ontogeny, and the sediments can also contain large amounts of colonies of the green alga *Pediastrum boryanum*, which indicates high primary productivity (Figure 9). This first phase is followed by lower primary productivity and many of the macrophyte species disappear or become rare and sterile. This development reflects decreased outwash of nutrients from the drainage area.

Colonization of Greenland

It has been much debated whether the plant species in Greenland survived the last Ice Age in ice-free areas in Greenland, or if they immigrated after the last deglaciation. It has been suggested, for example, that *Betula pubescens* survived in South Greenland. Tree birch is one of the most warmth-demanding plant species in Greenland, and if it survived the last Ice Age, this would not only mean that ice-free areas were present, but also that summer temperatures were no lower than at present. Also, *Betula nana* has been proposed as an Ice Age survivor. Even though dwarf birch is far less warmth-demanding than tree birch, it also has a climatically determined northern range limit in Greenland, and from a Greenlandic point of view it cannot be considered a cold-adapted plant. Pollen analysis has shown that both species arrived during the Holocene. Iversen proposed that only the hardiest plants survived, a viewpoint followed by Fredskild.

Fieldwork in Jameson Land has resulted in finds of macroscopic remains of plants that grew there during the first

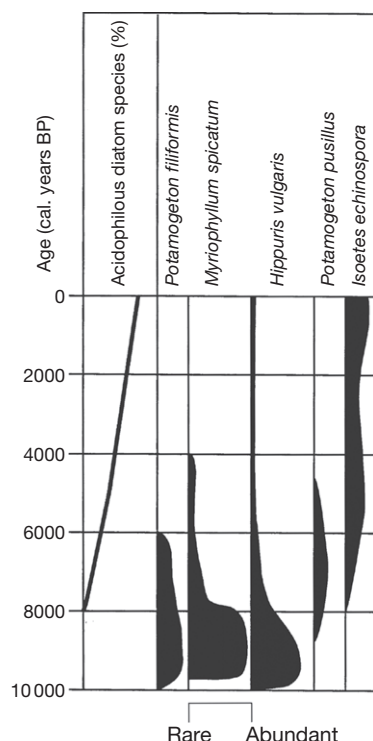


Figure 9 Strongly simplified model showing long-term changes of lakes in the area around the Godthåbsfjord in southwestern Greenland based on pollen, seeds, fruits, and leaf fragments. The plant species that prefer the highest conductivity and the most eutrophic conditions are placed to the left in the diagram.

millennium of the Holocene (Bennike et al., 1999). The oldest deposits only contain remains of a few hardy herbs and bryophytes. Most of the recorded species are presently widely distributed in Greenland, and play an increasingly larger role toward the north, but some species only found at present in the northern parts of Greenland are represented. The first woody plant that seems to have appeared is *Empetrum nigrum*, which is not especially hardy, but which has berries that are adapted to dispersal by birds. All warmth-demanding species are now believed to have immigrated after the last deglaciation. It is proposed that storms and birds, maybe often in combination, played a crucial role in long-distance chance dispersal.

With respect to macrolimnophytes, it has been concluded that the first species to arrive were species that are presently widely distributed in Greenland. More warmth-demanding species followed, and the most warmth-demanding species were the last to arrive (Fredskild, 1992).

The arrival of those plants that produce much pollen can be pinpointed rather precisely. In lake or peat sequences, the pollen curve for such a species may rise drastic, reflecting rapid expansion just after immigration. Birch species are prolific pollen producers, and from pollen analyses it has been possible to establish that *Betula nana* arrived in East Greenland at ca. 8.8 ka BP, and in West Greenland at ca. 7 ka BP. *Betula pubescens* arrived at ca. 4.2 ka BP in southern Greenland.

Possible immigration routes to Greenland can be suggested for those species that have restricted modern day geographical ranges (Figure 10). *Draba sibirica* (siberian whitlow-grass) is,

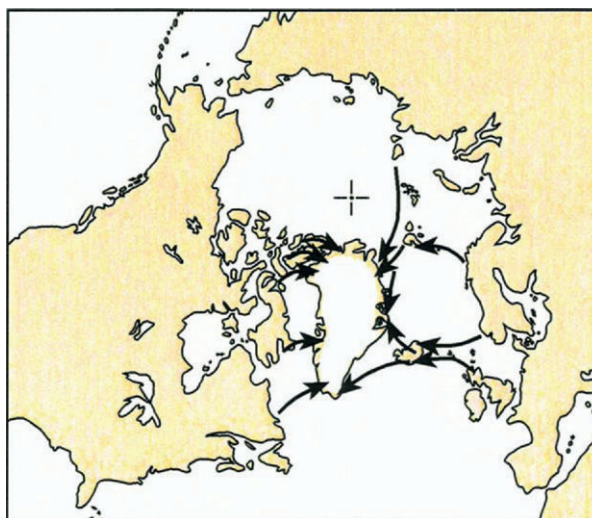


Figure 10 Map of the Earth's northern parts with arrows that show possible immigration routes to Greenland. The routes are suggested from the modern day geographical ranges of the species. The route from Labrador has played a small role, even though the distance to Greenland is fairly short. However, *Alnus crispa* and some other species probably colonized Greenland from Labrador. Even fewer plants appear to have arrived from Baffin Island. Surprisingly, many species have colonized Greenland from northwestern Europe. The distance is long, but the Faeroe Islands and Iceland form stepping stones along the route. If dispersal by birds is important, that may explain this paradox, because a large part of the Greenland migrating birds winter in Europe and Africa. A few species are considered immigrants from Siberia to East Greenland.

for example, considered an immigrant from Siberia to East Greenland. It has previously been suggested that this species immigrated to Greenland via western Siberia and northern North America. However, it is hard to understand why the populations along the migration route should die out.

Molecular data show that populations of plant species from opposite sides of the Atlantic Ocean often have identical or very similar chloroplast haplotypes, which indicates that migration rates across the Atlantic are much higher than previously envisaged by many botanists (Brochmann et al., 2003). On the basis of the genetic results it is concluded that the glacial survival hypothesis is not needed to explain the modern day distributions of plants in the Arctic. Hence, these results support the fossil evidence for postglacial colonization by the majority of the vascular plants in Greenland.

See also: **Beetle Records:** Late Tertiary and Early Quaternary Records. **Ice Core Records:** Greenland Stable Isotopes. **Plant Macrofossil Records:** Arctic Eurasia; Arctic North America.

References

- Bennike O (1990) The Kap København formation: Stratigraphy and palaeobotany of a Plio-Pleistocene sequence in Peary Land, North Greenland. *Meddelelser om Grønland-Geoscience* 23: 85.
- Bennike O (1999) Colonisation of Greenland by plants and animals after the last ice age: A review. *Polar Record* 35: 323–336.

- Bennike O (2000) Palaeoecological studies of Holocene lake sediments from West Greenland. *Palaeogeography, Palaeoclimatology, Palaeoecology* 155: 285–304.
- Bennike O, Abrahamsen N, Matgorzata B, et al. (2002) A multi-proxy study of Pliocene sediments from Île de France, northeast Greenland. *Palaeogeography, Palaeoclimatology, Palaeoecology* 186: 1–23.
- Bennike O, Björck S, Böcher J, Hansen L, Heinemeier J, and Wohlfarth B (1999) Early Holocene plant and animal remains from northeast Greenland. *Journal of Biogeography* 26: 667–677.
- Bennike O and Björck S (2002) Chronology of the last recession of the Greenland ice sheet. *Journal of Quaternary Science* 17: 211–217.
- Bennike O and Böcher J (1992) Early Weichselian interstadial land biotas at Thule, northwest Greenland. *Boreas* 21: 111–117.
- Bennike O and Böcher J (1994) Land biotas of the last interglacial/glacial cycle on Jameson Land, East Greenland. *Boreas* 23: 479–487.
- Bennike O and Weidick A (2001) Late Quaternary history around Nioghalvfjærdstjorden and Jøkelbugten, northeast Greenland. *Boreas* 30: 205–227.
- Björck S, Bennike O, Rosén P, et al. (2002) Anomalously mild Younger Dryas summer conditions in South Greenland. *Geology* 30: 427–430.
- Böcher J (1995) Palaeontomology of the Kap København formation, a Plio-Pleistocene sequence in Peary Land, North Greenland. *Meddelelser om Grønland-Geoscience* 33: 82.
- Brochmann C, Gabrielsen TM, Nordal I, Landvik JY, and Elven R (2003) Glacial survival or tabula rasa? The history of North Atlantic biota revisited. *Taxon* 52: 417–450.
- Fredskild B (1967) Palaeobotanical investigations at Sermermiut, Jakobshavn, West Greenland. *Meddelelser om Grønland* 178(4): 54.
- Fredskild B (1973) Studies on the vegetational history of Greenland. *Meddelelser om Grønland* 198(4): 245.
- Fredskild B (1978) Palaeobotanical investigations of some peat deposits of Norse age at Qagssiarssuk, South Greenland. *Meddelelser om Grønland* 204(5): 41.
- Fredskild B (1983) The Holocene vegetational development of the Godthåbsfjord area, West Greenland. *Meddelelser om Grønland-Geoscience* 10: 28.
- Fredskild B (1985a) Holocene pollen records from West Greenland. In: Andrews (ed.) *Quaternary environments. Eastern Canadian Arctic*, pp. 643–681. Boston: Allen and Unwin.
- Fredskild B (1985b) The Holocene vegetational development of Tugtuliqssuaq and Qeqertat, northwest Greenland. *Meddelelser om Grønland-Geoscience* 14: 20.
- Fredskild B (1988) Agriculture in a marginal area – South Greenland from the Norse Landman (985 A.D.) to the present (1985 A.D.). In: Birks HH, Birks HJB, Kaland PE, and Moe D (eds.) *The Cultural Landscape – Past, Present and Future*, pp. 381–393. Cambridge: Cambridge University Press.
- Fredskild B (1992) The Greenland limnophytes – Their present distribution and Holocene history. *Acta Botanica Fennica* 144: 93–113.
- Fredskild B (1995) Palynology and sediment slumping in a high arctic Greenland lake. *Boreas* 24: 345–354.
- Fredskild B (2000) The Holocene vegetational changes on Qeqertarsuatsiaq, a West Greenland island. *Danish Journal of Geography* 100: 7–14.
- Funder S (1978) Holocene stratigraphy and vegetation history in the Scoresby Sund area, East Greenland. *Bulletin Grønlands Geologiske Undersøgelse* 129: 66.
- Funder S (1989) Quaternary geology of the ice-free areas and adjacent shelves of Greenland. In: Fulton RJ (ed.) *Quaternary Geology of Canada and Greenland*, pp. 741–792. Ottawa: Geological Survey of Canada.
- Funder S, Bennike O, Böcher J, Israelsson C, Petersen KS, and Símonarson L (2001) Late Pliocene Greenland – The Kap København Formation in Peary Land. *Bulletin of the Geological Society of Denmark* 48: 117–134.
- Hedenäs L and Bennike O (2003) Moss remains from the last interglacial at Thule, northwest Greenland. *Lindbergia* 28: 52–58.

Holocene North America

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Glossary

$\delta^{13}\text{C}$: The ratio of two stable isotopes of carbon, ^{13}C and ^{12}C , reported as departures in parts per thousand from a laboratory standard. They can be used to determine the relative abundance of warm-season prairie grasses (C_4 ; $\delta^{13}\text{C}$ around -4° / in speleothems)

to other plants (mainly C_3 ; $\delta^{13}\text{C}$ around -9° / in speleothems).

Speleothem: Calcium carbonate deposits in a cave, in this case, stalagmites (columns growing upward from the cave floor).

Isotopes: Alternate forms of the same element having different numbers of neutrons in their atoms.

Introduction

The purpose of this article is to summarize research on Holocene plant macrofossils from a new or at least a little-used source, alluvial sediments in the upper reaches of small streams, in reconstructing the vegetational history of the central Midwestern United States. The focus is on the spatial and temporal establishment of upland vegetation, being principally deciduous forest, savannah, and prairie. The two most complete sites along the prairie-forest border, one in northeastern Iowa and the other in southwestern Minnesota, will be presented first and in the greatest detail. Then results are briefly described from sites in southeastern Nebraska, well into the presettlement prairie, southwestern Indiana within the presettlement deciduous forest, and along the prairie-forest border in southwestern Missouri.

Plant macrofossils were one of the most under-used, valuable paleoecological tools in North American Quaternary paleoecology. Although they were used first in the early 1900s to establish the foundations of the vegetational history of northwest Europe, their use was almost universally supplanted by pollen analysis. Arguably, the University of Minnesota has played an important role as a center for plant macrofossil research. Although some macrofossil work in Minnesota was done prior to 1950 (Rosendahl, 1948), modern work with a stratigraphic context began with the establishment of a laboratory and reference collection by HE Wright Jr. at the Limnological Research Center, University of Minnesota. Visiting scientist WA Watts, working on lake and marsh sediments, established the importance of macrofossils in paleoecology beginning in the 1960s. This tradition was carried on by John and Hilary Birks (Birks, 1980), Krystyna Wasylikowa, and Johanna and Eberhard Gröger in the late 1960s and 1970s, and by later visitors and students. The small field of macrofossil workers grew as John Matthews, Norton Miller, Stephen Jackson, Rolf Mathewes, Paul and Hazel Delcourt, Dorothy Peteet, and others established their own laboratories, trained students, and contributed substantially to macrofossil research. Recent summaries indicate that the field has matured (Birks and Birks, 2003; Jackson, et al., 1997).

Despite these additional researchers, our understanding of the Holocene vegetation of the central Midwestern United States south of the Wisconsin glacial border has lagged behind that

of glaciated terrain because typical fossiliferous sites are rare. My colleagues and I discovered that sediments in many cutbanks, in first- to third-order streams beyond the glacial border, contain a rich record of pollen and plant macrofossils (Figure 1). These remains (Tables 1 and 2) are present where the ground-water table has been consistently high, preventing oxidation of the sediments and their organic remains.

To study these sequences, at least 2–4 liters of cutbank samples are dug from, at, or near water level (or rarely cored), stored in plastic bags, and brought to the laboratory. A minimum of 200 ml is washed through screens with 0.5 and 0.1 mm mesh, and plant remains are picked from the residue using small tweezers and brushes. The various parts of plants are identified using a modern seed, fruit, and bud collection of over 3100 accessions housed in The Repository, Department of Geoscience, University of Iowa.

By choosing sample sites relatively close to the headwaters of the stream, the chances of long-distance transport by running water are eliminated. Each site represents a short period of deposition, so each is a snapshot of vegetation in the area at one time. Every site is radiocarbon-dated using well-preserved wood, or occasionally leaves or disseminated organic materials. Sampling multiple sites up and down the stream has provided a good representation of Holocene vegetation. Ages reported here have been converted to calibrated radiocarbon dates unless otherwise specified. Conversion of ages to calibrated years was based on 50% median probability and 2-sigma range and was referenced to 1950 AD using CALIB 5.0 (Stuiver et al., 2005).

The richness of the alluvial records (compare Kirchner Marsh with other sites in Tables 1 and 2) and the ability to identify many of the plant remains to species, allow much greater resolution of Holocene plant communities including aquatic, wetland, disturbed ground, forest trees, forest understory herbs, shrubs and vines, and prairie, than is possible from lake or wetland samples. Because of space limitations, only prairie and forest habitats are presented in this article. Other habitats are described in Baker et al. (1996, 1998, 2002) and Baker (2000).

Macrofossils most commonly represented are seeds and fruits, but a wide variety of other plant parts can be identified in many cases. The types of remains identified in this study are listed at the bottom of Tables 1 and 2. The units plotted on the diagrams are numbers of specimens/liter.

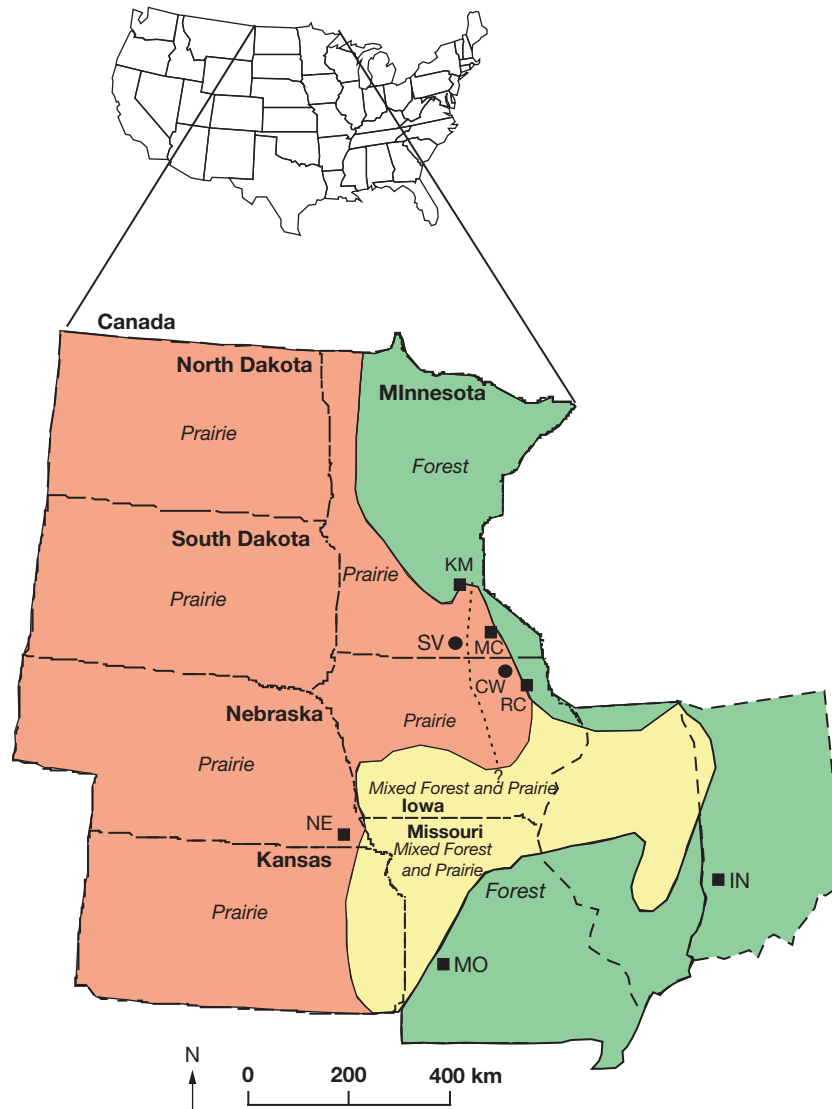


Figure 1 Map of Midwestern sites mentioned in the text and generalized presettlement vegetation. Short-dashed line is mid-Holocene prairie-forest ecotone.

The use of alluvial samples in reconstructing past environments has been validated in several ways:

1. Two studies compared plant remains in modern stream alluvium with the vegetation in the surrounding area. [Baker and Drake \(1994\)](#) investigated alluvial samples from a prairie preserve in northwestern Iowa, and [Strickland \(1998\)](#) examined samples from a forest stream in southeastern Minnesota. In both, the plant remains contained numerous indicators of several habitats in the surrounding vegetation.
2. Interpretation of pollen, macrofossils, and other types of fossils in several Holocene alluvial samples are compatible ([Baker et al., 1996, 2000, 2002; Work et al., 2005](#)).
3. Speleothem isotopes ([Figure 2](#)) provide a high-resolution (annual to decadal) record of events along the prairie-forest border ([Baker, et al., 1996, 2002; Denniston et al., 1999;](#)
4. [Dorale et al., 1992](#)), and $\delta^{13}\text{C}$ fluctuations in these sites indicate the same directions and timing of Holocene vegetational change. The $\delta^{18}\text{O}$ record shows changes concomitant with those in the $\delta^{13}\text{C}$ record and indicates a response to both temperature and evaporation ([Denniston et al., 1999](#)).
4. The $\delta^{13}\text{C}$ in fine-grained alluvial samples correlates well with both associated plant fossils and with a nearby speleothem isotopic record ([Baker et al., 1998](#)).

The diagrams in this article are color-coded to represent the presettlement vegetation on the map. The same colors are used on the diagrams to indicate the close correspondence between Holocene and modern communities. Blue = Late Glacial conifer forest (not shown on map), green = deciduous forest, orange = prairie, and yellow = savannah or savannah-like vegetation.

Table 1 Scientific and common names of trees and forest herbs, plant organs found as macrofossils, and the sites where their fossils have been found during the forested intervals

Scientific name	Common name	Fossil type	KM	RC	MC	NE	IN	MO
Trees								
<i>Acer negundo</i>	boxelder	s		x		x		x
<i>Acer rubrum</i>	red maple	s					x	
<i>Acer saccharinum</i>	silver maple	l				x		cf.
<i>Acer saccharum</i>	sugar maple	b					x	cf.
<i>Acer</i> sp.	maple	s		x		x		x
<i>Alnus serrulata</i>	hazel alder	c						x
<i>Betula nigra</i>	river birch	s					x	x
<i>Betula lutea</i>	yellow birch	s,c			x			
<i>Betula papyrifera</i>	paper birch	s,c	x		x			
<i>Betula</i> sp.	birch	s,c		x				x
<i>Carpinus caroliniana</i>	hornbeam	s		x	x		x	
<i>Carya cordiformis</i>	bitternut hickory	n				x	x	x
<i>Carya laciniosa</i>	shellbark hickory	n					x	
<i>Carya ovata</i>	shagbark hickory	n, w				x	cf.	cf.
<i>Carya</i>	hickory	n			cf.		x	
<i>Fagus grandifolia</i>	beech	s					x	
<i>Fraxinus americana</i>	white ash	s				x		
<i>Fraxinus pennsylvanica</i>	green ash	s				type		type
<i>Fraxinus nigra</i>	black ash	s		x	cf.			
<i>Fraxinus</i> sp.	ash	s			cf.		x	x
<i>Gleditsia triacanthos</i>	honey locust	l				x		type
<i>Juglans cinerea</i>	butternut	n, ls		x		x	x	
<i>Juglans nigra</i>	black walnut	n		x	cf.	x	x	x
<i>Juglans</i> sp.	walnut/butternut	t				x		
<i>Juniperus virginiana</i> type	eastern red cedar	t						x
<i>Larix laricina</i>	tamarack	l		x	x			
<i>Liquidambar styraciflua</i>	sweetgum	b					cf.	
<i>Liriodendron tulipifera</i>	tuliptree	s					x	
<i>Morus rubra</i>	red mulberry	s				x	x	x
<i>Nyssa</i>	blackgum	s					cf.	
<i>Ostrya virginiana</i>	hophornbeam	s		x		x	x	
<i>Picea glauca</i>	white spruce	l,co		x	x			
<i>Picea mariana</i>	black spruce	l,co						
<i>Platanus occidentalis</i>	sycamore	s,ba				x	x	x
<i>Populus deltoides</i>	cottonwood	s		cf.		x	x	
<i>Populus tremuloides</i>	quaking aspen	b			x		type	type
<i>Populus</i> sp.	aspen	b	x	x			x	
<i>Populus</i> type	aspen	b			x			
<i>Prunus serotina</i>	black cherry	s		x				
<i>Prunus</i> sp.	cherry, plum	s						
<i>Quercus macrocarpa</i>	bur oak	s,b				x		x
<i>Quercus marilandica/stellata</i>	blackjack/post oak	s				x	x	
<i>Quercus palustris</i>	pin oak	s					cf.	
<i>Quercus rubra</i>	northern red oak	s						x
<i>Quercus velutinus</i> type	black oak	s						x
<i>Quercus</i> (white oak group)	white oak group	l				x	x	x
<i>Quercus</i>	oak	l		x	x	x	x	x
<i>Salix amygdaloides</i>	peachleaf willow	l						cf.
<i>Salix exigua</i> ssp. interior type	sandbar willow	l				x		x
<i>Salix</i>	willow	s	x	x	x			
<i>Staphylea trifolia</i>	bladdernut	s					x	
<i>Tilia americana</i>	basswood	s		x		x	x	x
<i>Ulmus americana</i>	American elm	b		x		x	x	x
<i>Ulmus rubra</i>	red elm	b		x		x		x
Forest-Floor Herbs								
<i>Agastache foeniculum</i>	giant blue hyssop	s			x			
<i>Aralia nudicaulis</i>	wild sarsaparilla	s			x			
<i>Aralia racemosa</i>	American spikenard	s		x	x		x	
<i>Campanula americana</i>	tall bellflower	s		x	x	x	x	x

(Continued)

Table 1 (Continued)

Scientific name	Common name	Fossil type	KM	RC	MC	NE	IN	MO
<i>Carex backii</i>	Back's sedge	s					type	
<i>Carex jamesii</i>	James' sedge	s						type
<i>Carex squarrosa</i>	squarrose sedge	s					x	
<i>Cerastium nutans</i>	nodding chickweed	s		x				
<i>Claytonia virginica</i>	spring beauty	s		x	x		x	x
<i>Cryptotaenia canadensis</i>	honewort	s		x	x		cf.	x
<i>Eupatorium altissimum</i>	tall thoroughwort	s		x	x			
<i>Eupatorium purpureum</i>	sweet-scented joepeyweed	s		x				
<i>Eupatorium rugosum</i>	white snakeroot	s			type		cf.	type
<i>Galium triflorum</i>	fragrant bedstraw	s						x
<i>Iodanthus pinnatifidus</i>	purplerocket	s		x				
<i>Laportea canadensis</i>	wood nettle	s		x	x	x	x	x
<i>Linum striatum</i>	ridged yellow flax	s						cf.
<i>Polygonum scandens</i>	climbing false buckwheat	s		x				
<i>Ranunculus abortivus</i>	littleleaf buttercup	s		x	x		x	
<i>Ranunculus hispidus</i>	bristly buttercup	s		x	x			
<i>Rudbeckia triloba</i>	browneyed susan	s			x			x
<i>Salvia lyrata</i>	lyreleaf sage	s				x		
<i>Sanicula gregaria</i>	clustered blacksnakeroot	s				x		
<i>Thalictrum dioicum</i>	early meadowrue	s			x			
<i>Taenidia integerrima</i>	yellow pimpernel	s		x				
<i>Triodanis perfoliata</i>	Venus' looking-glass	s						x
<i>Triosteum perfoliatum/auranticum</i>	horse-gentian	s			x			

fossil abbreviations: s = seed or fruit;

l = leaf or leaf fragment; b = bud or bud scale;

c = catkin scale; co = cone;

n = nut or nut fragment; ba = bark

ls = leaf scar, t-twig, w = wood.

Table 2 Scientific and common names of prairie plants, plant organs found as macrofossils, and the sites where their fossils have been found during the prairie and savannah intervals

Taxon	Common name	Fossil type	KM	RC	MC	NE	IN	MO
<i>Agalinus aspera</i>	tall false foxglove	s		x				
<i>Ambrosia psilostachya</i> type	western ragweed	s						x
<i>Amorpha canescens</i>	lead plant	s, l	x	x	x	x		x
<i>Amorpha nana</i>	dwarf false indigo	s, l				x		
<i>Amphicarpea bracteata</i>	American hogpeanut	s		x				
<i>Andropogon gerardii</i>	big bluestem	f		x	x	x		x
<i>Androsace occidentalis</i>	western rockjasmine	s				x		x
<i>Anemone canadensis</i>	Canada anemone	s		x				
<i>Carex cf. bicknellii</i>	Bicknells' sedge	s				x		
<i>Carex meadii</i> type	Meads' sedge	s				x		
<i>Ceanothus americanus</i>	New Jersey tea	s		x		type		
<i>Chamaechrista fasciculata</i>	partridge pea	l						x
<i>Coreopsis nudata</i> type	Georgia tickseed	s						x
<i>Dalea candida</i>	white prairieclover	c	x	x	x	x		x
<i>Dalea purpurea</i>	purple prairieclover	c				cf.		cf.
<i>Dalea</i>	prairieclover	c						x
<i>Desmanthus illinoensis</i>	Illinois bundleflower	l				cf.		
<i>Desmodium canadense</i>	showy ticktrefoil	s		x	x	x		x
<i>Desmodium canescens</i>	hoary ticktrefoil	s						x
<i>Desmodium marilandicum</i> type	smooth small-leaved ticktrefoil	s						x
<i>Desmodium</i> sp.	ticktrefoil	s		x		x		x
<i>Dicranthelium oligosanthes</i> type	Scribner's dicranthelium	f		x				
<i>Eupatorium altissimum</i>	tall thoroughwort	s						x

(Continued)

Table 2 (Continued)

<i>Taxon</i>	<i>Common name</i>	<i>Fossil type</i>	<i>KM</i>	<i>RC</i>	<i>MC</i>	<i>NE</i>	<i>IN</i>	<i>MO</i>
<i>Euphorbia corollata</i>	flowering spurge	s		x				
<i>Helianthus annuus</i> type	common sunflower	s		x		x		
<i>Helianthus grosseserratus</i> type	sawtooth sunflower	s		x	x	x		
<i>Helianthus x laetiflorus</i> type	cheerful sunflower	s			x	x		x
<i>Helianthus</i> sp.	sunflower	s						x
<i>Heliopsis helianthoides</i>	ox-eye	s			x	x	x	
<i>Hypoxis hirsuta</i>	yellow stargrass	s		x	x			x
<i>Lactuca</i> cf. <i>ludoviciana</i>	western wild lettuce	s			x			
<i>Lespedeza capitata</i>	round-headed bushclover	s				x		
<i>Lespedeza</i> (not <i>capitata</i>)	bushclover	s				x		
<i>Liatris aspera</i>	tall blazing star	s		x				
cf. <i>Liatris pycnostachya</i>	prairie blazing star	s		x				
cf. <i>Liatris spicata</i>	dense blazing star	s		x				
<i>Liatris</i>	blazing star	s				x		
<i>Linum sulcatum</i>	grooved flax	s		x	x	x		
<i>Lobelia spicata</i>	palespike lobelia	s		x	x			
<i>Monarda fistulosa</i>	wild bergamot	s			cf.	x	x	x
<i>Monarda pectinata</i>	pony beebalm	s				x		
<i>Monarda</i> cf. <i>punctata</i>	spotted beebalm	s		x				
<i>Oenothera biennis</i>	common evening-primrose	s						x
<i>Oxalis violacea</i>	violet woodsorrel	s			x			cf.
<i>Oxypolis rigidior</i>	stiff cowbane	s			x			
<i>Panicum capillare</i> type	witchgrass	f			x	x		
<i>Panicum</i> cf. <i>virgatum</i>	switchgrass	f		x				
<i>Phlox pilosa</i>	prairie phlox	s		x	x			cf.
<i>Polygala alba</i>	white milkwort	s			cf.	x		
<i>Polygala senega</i>	Seneca snakeroot	s			x			
<i>Polygala verticillata</i>	whorled milkwort	s		x	x	x		x
<i>Potentilla arguta</i>	tall cinquefoil	s		x	x			
<i>Pycnanthemum</i> cf. <i>pilosum</i>	hairy mountainmint	s			x			
<i>Pycnanthemum virginianum</i>	Virginian mountainmint	s		x				
<i>Pycnanthemum</i>	mountainmint	s			x	x		x
<i>Ratibida pinnata</i>	gray-headed coneflower	s		x	x	x		x
<i>Rudbeckia hirta</i>	blackeyed susan	s		x	x			x
<i>Schizachyrium scoparium</i>	little bluestem	f		x	x	x		x
<i>Schrankia nuttallii</i>	sensitive brier	l				x		
<i>Silphium laciniatum</i>	compass plant	s		x				
cf. <i>Silphium perfoliatum</i>	cup plant	s		x				x
<i>Sorghastrum nutans</i>	indiangrass	f		x	x	x		x
<i>Sporobolus asper</i>	rough dropseed	s				x		
cf. <i>Sporobolus</i>	dropseed	s				x		
<i>Stipa spartea</i>	porcupine grass	a		x		x		
<i>Talinum parviflorum</i>	prairie fameflower	s						x
<i>Thalictrum dasycarpum</i>	purple meadow-rue	s		x			x	
<i>Tradescantia</i>	spiderwort	s				x		
<i>Verbena stricta</i> type	prairie vervain	s				x		
<i>Vernonia baldwini</i> type	western ironweed	s				x		
cf. <i>Vernonia</i>	ironweed	s				x		
<i>Veronicastrum virginicum</i>	Culver's root	s		x				
<i>Zizia aptera</i>	heart-leaved golden alexender	s			x			x
<i>Zizia aurea</i>	golden alexander	s		x	x			cf.
Other large grass florets	grass	f						x

s = seed or fruit, l = leaf or leaflet;

f = floret; c = calyx; a = awns.

Northeastern Iowa

Roberts Creek is located in the forest, just east of the prairie-forest border in Iowa (Figure 1). Although this article focuses on the Holocene record, two sites contain Late Glacial

records. The Late Glacial at Roberts Creek (Figure 3), lasting until ~11 000 years BP, is dominated by *Picea* and *Larix* pollen, needles, and cones confirming regional trends, showing that these trees arrived at least by 13 000 years BP and were the dominant Late Glacial taxa. Although the numerous *Picea*

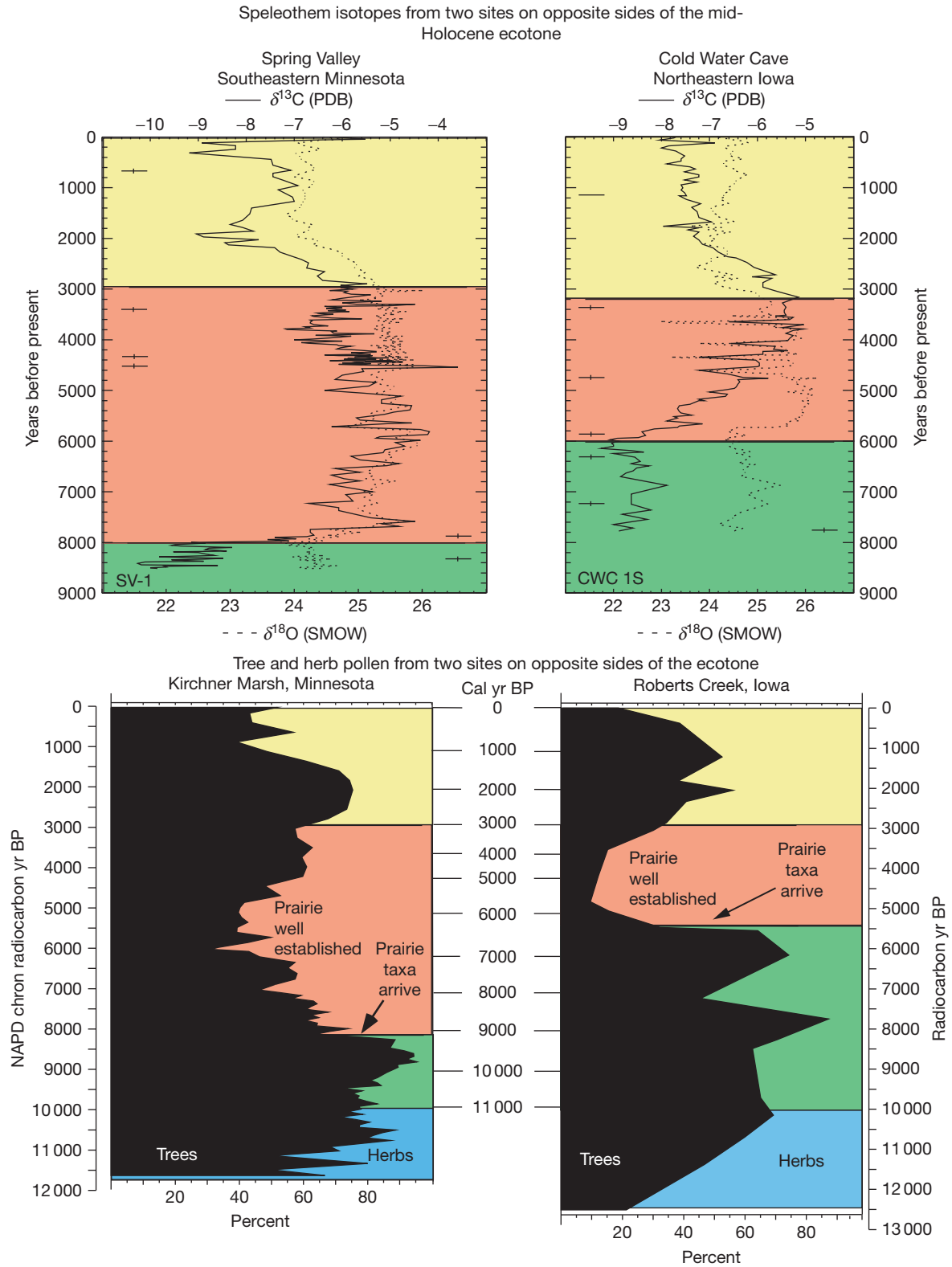


Figure 2 Comparison of arboreal pollen curve and speleothem records. Spring Valley cave and Kirchner Marsh are north and west, and Cold Water Cave and Roberts Creek are south and east, of a mid-Holocene forest/prairie ecotone.

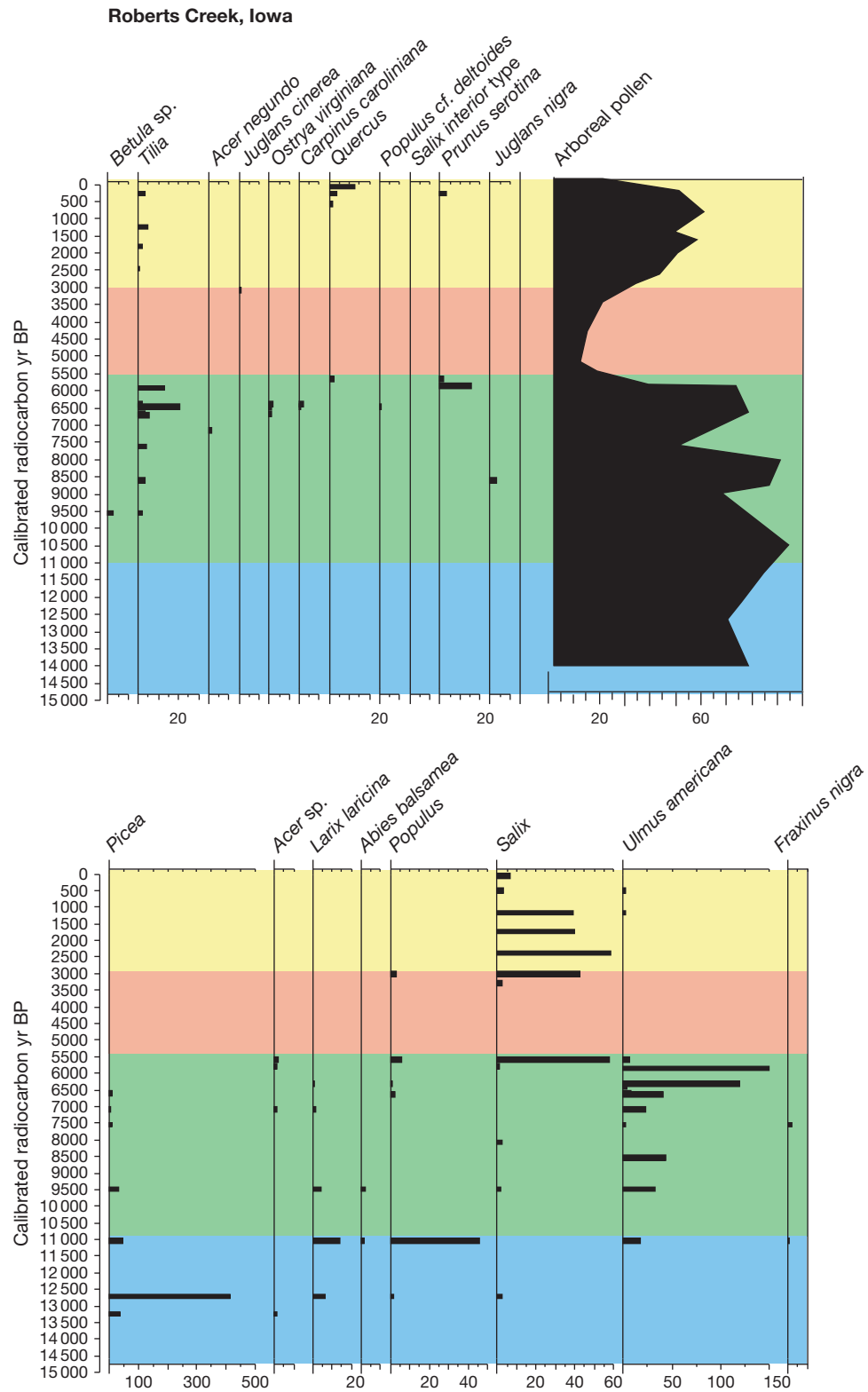


Figure 3 Late Glacial and Holocene macrofossils from Roberts Creek, northeast Iowa; trees and arboreal pollen percentages of the total pollen sum.

needles and pollen were not identified to species, several cones of *P. glauca* and one of *P. mariana* were present. Macrofossils of *Salix*, *Populus*, a single *Acer* sp. fruit fragment and slightly later, *Fraxinus nigra*, indicate that some deciduous trees were present in the spruce-larch forest during the Late Glacial.

The deciduous forest phase contains macrofossils of *Ulmus americana*, *Fraxinus nigra*, *Tilia americana*, *Ostrya virginiana*, *Juglans nigra*, *Carpinus caroliniana*, and *Quercus* sp. (Figure 3). Early in this phase, the presence of *Abies balsamea* needles suggests that an early period of conifer-hardwood forest (not differentiated in the figures) may have been present. Macrofossils of shady forest-floor taxa habitats are also present (Table 2), including *Aralia racemosa*, *Ranunculus abortivus*, *Cryptotaenia canadensis*, *Claytonia* cf. *virginica*, and *Laportea canadensis*, which have seldom, if ever, been found as fossils. The combined macrofossils from trees and forest herbs indicate that closed deciduous forest was present during much of this interval. Near the top of this forested phase, macrofossils of the pioneer prairie taxa, *Monarda fistulosa*, *Rudbeckia hirta*, and *Zizia aurea* appear, while forest taxa are still abundant. These taxa are among the early dominants in modern prairie reconstructions, suggesting that they also were early migrants into forest openings that began to appear. The occurrence of *Picea* and *Larix* needle fragments is a common feature of several midwestern macrofossil sites (Watts and Winter, 1967). These are degraded tips and bases of needles that are resistant to both chemical and mechanical breakdown. Although these conifers could have persisted in the early Holocene in cool, north-facing habitats, their occurrence, their persistence well into the mid-Holocene, and their corroded appearance suggests that they are reworked.

Arboreal pollen, together with arboreal and forest-floor macrofossils, decreased sharply about 5500 years BP, as Poaceae and *Ambrosia* pollen percentages rose rapidly. A rich mix of over 20 taxa of plant macrofossils of prairie plants occurs in the orange zone and the lower part of the prairie phase (Plate 1, Figure 4). Many of these represent presently common tall-grass prairie plants in the three most important families, Poaceae

(*Sorghastrum nutans*, *Andropogon gerardii*, *Schizachyrium scoparium*, *Stipa spartea*, *Dicanthelium oligosanthos* type), Fabaceae (*Amorpha canescens*, *Dalea candida*, *D. purpurea*, *Desmodium canadense*), and Asteraceae (*Helianthus grosseserratus* type, *Rudbeckia hirta*, *Ratibida pinnata*, *Liatris aspera*). *Monarda fistulosa*, *Zizia aurea*, and *Hypoxis hirsuta* are some of the other common prairie taxa represented (Figure 4, Table 2). This diverse assemblage represents species that presently grow in dry, mesic, and wet prairies, and they give a much more complete picture of the prairie than pollen alone. (See Validation of Pollen Studies)

Two particular cutbanks are especially rich in macrofossils (Figure 4), and abundant charcoal and popped seeds indicate that fires removed the myriad of dense prairie stems, which increased runoff and facilitated transport of plant remains to these sites.

In the subsequent savannah phase, the increase in arboreal pollen percentages (Figure 2) is largely *Quercus*, which is generally poorly represented in the macrofossil record. Few macrofossils of other arboreal taxa occur in this interval, and only *Tilia* and the riparian tree *Salix* are consistently present. Only two forest-floor herbs are present near the top of the sequence, suggesting that heavily shaded habitats were rare. Prairie plants continue to be well represented during this last 3500-year period. This assemblage is consistent with an oak savannah.

The post-settlement period, the widely recognized *Ambrosia* pollen zone in North America, is represented by the youngest three samples. It shows a marked increase in numbers and diversity (at least 18 taxa) of weedy-plant macrofossils, with peaks of many alien species as well as native weeds (Baker, et al., 1993, 1996). Most of the macrofossils representing other habitats declined or disappeared in the historical period. Non-native insects also appear in the post-settlement sediments, and the sediments indicate rapid runoff and soil erosion (Baker et al., 1993). These changes mesh well with the historical record of deforestation and cultivation of crops, indicating that the stream cutbanks and their macrofossil assemblages also record changes in human land-use practices.

The high-resolution carbon isotopic record from Coldwater Cave (Figures 1 and 2) fills in the gaps in the alluvial record and supports the conclusions. The carbon isotopic record shows $\delta^{13}\text{C}$ values between -8 and -9 , indicating a strong C_3 -dominated vegetation for the forest from 9000 to 6000 years BP. Beginning about 6000 years BP, the $\delta^{13}\text{C}$ values increased gradually to -5.5 by 3000 years BP. This gradual buildup marks the slow accumulation of organic carbon in the soil over the cave of $\delta^{13}\text{C}$ from C_4 grasses that are dominant in the prairie and present at Roberts Creek. The large fluctuations may indicate periodic droughts similar to those recorded in the ratios of aquatic plants to weeds in the macrofossil record at Kirchner Marsh (Watts and Winter, 1967). The re-establishment of some trees in savannah is indicated by the return of $\delta^{13}\text{C}$ values to -7 to -8 after 3000 years BP.

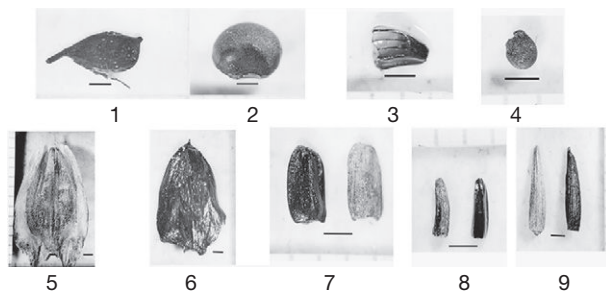


Plate 1 Prairie-plant macrofossils and modern reference specimens. Fossils are all from Roberts Creek, northeast Iowa. 1-Fossil *Amorpha canescens* pod, 2-Fossil *Amphicarpaea bracteata* pod, 3-Fossil *Dalea candida* calyx, 4-Fossil *Hypoxis hirsuta* seed, 5-Modern *Silphium laciniatum* fruit, 6-Fossil *Silphium laciniatum* fruit, 7-Fossil (left) and modern (right) *Ratibida pinnata* fruits, 8-Modern (left) and fossils (right) of *Rudbeckia hirta* fruit, and 9-Modern (left) and fossil (right) of *Liatris aspera* fruit. Bar is 1 mm long.

Southeastern Minnesota

A Late Glacial record is also present at Money Creek and Pine Creek (hereafter simply referred to as Money Creek), which are also in the deciduous forest at present. It is similar to that at

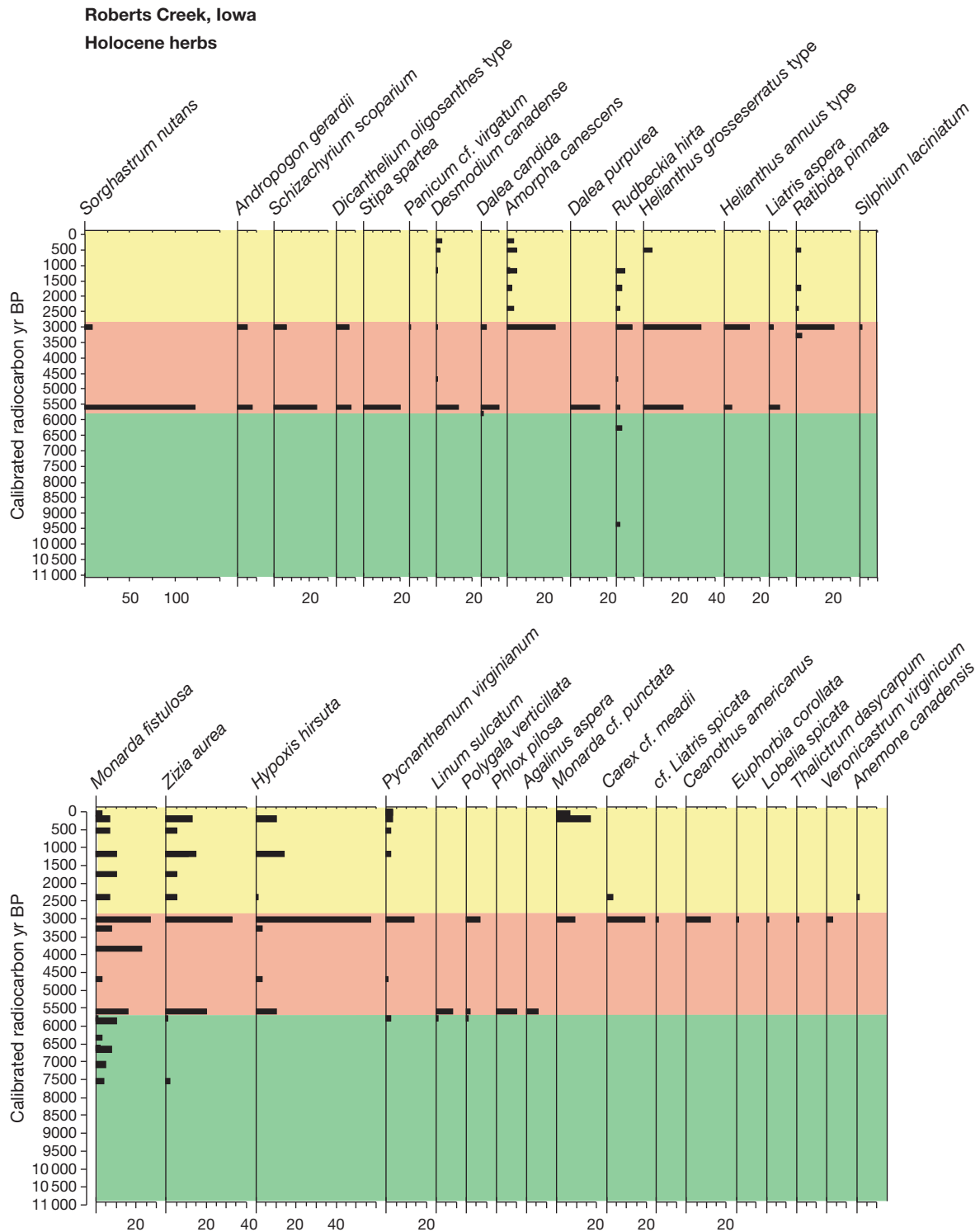


Figure 4 Holocene macrofossils from Roberts Creek, northeast Iowa; prairie plants.

Roberts Creek, except for the occurrence of the sub-Arctic species *Selaginella selaginoides* and *Betula nana* (= *B. glandulosa*). *Picea mariana*, *Larix laricina*, *Populus* sp., *Salix* sp., and *Tilia* were present at Money Creek about 13 000 years BP (Figure 5). The deciduous elements are not abundant, but their occurrence indicates at least a scattered presence in Late Glacial forests.

Although there are significant gaps in the Late Glacial record from about 13 000 to 9700 years BP and in the early Holocene record from 9500 to about 7400 years BP (Baker et al., 2001, 2002), there are close similarities to the Roberts Creek sequence. The early Holocene samples contain *Tilia americana*, *Ulmus americana*, *Carpinus caroliniana*, *Ostrya*

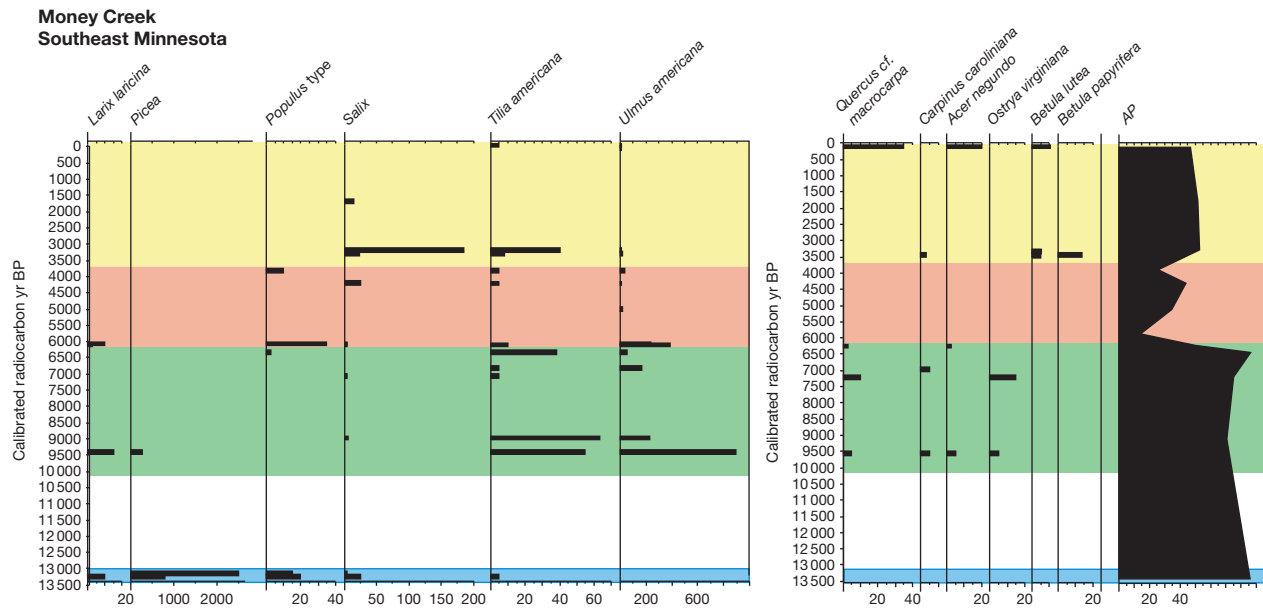


Figure 5 Late Glacial and Holocene macrofossils from Money Creek, southeast Minnesota; trees and arboreal pollen percentages of the total pollen sum.

virginiana, and *Quercus cf. macrocarpa* (Figure 5). In addition, a similar suite of shady forest-floor understory herbs occurs in the deciduous-forest phase, including *Aralia racemosa*, *Laportea canadensis*, and *Campanula americana*. *Monarda fistulosa*, *Heliopsis helianthoides*, and *Helianthus grosseserratus* type, which are pioneer prairie taxa, are present as early as 8300 years BP, suggesting that, as at Roberts Creek, these plants became established in forest openings before the arrival of prairie.

All deciduous forest taxa disappear around 5500 years BP, and arboreal pollen percentages drop precipitously (Figures 5 and 6). Macrofossils of the important tall-grass prairie grasses, *Schizachyrium scoparium* and *Sorghastrum nutans* appear, followed by *Helianthus laetiflorus* type, *Amorpha canescens*, *Andropogon gerardii*, *Desmodium canadense*, *Pycnanthemum* sp., *Rudbeckia hirta*, *Dalea candida*, and *Zizia aptera* (Figure 6). As at Roberts Creek, these taxa and the disappearance of tree taxa indicate that a diverse mesic tall-grass prairie was well established. The dominant prairie-plant families, Poaceae, Fabaceae, and Asteraceae, are also well represented here by plant macrofossils.

After about 3500 year BP, the abundance and diversity of prairie-plant macrofossils decreases, arboreal pollen percentages rebound, especially *Quercus*, and macrofossils of several deciduous trees reappear (Figure 5). Several prairie-plant macrofossils continue to be present, suggesting that oak savannah was established, with prairie growing in open areas between the widely spaced oaks (Figure 6). The tall grasses (*Sorghastrum*, *Schizachyrium*, and *Andropogon*) were not found in the savannah phase. These grasses provide substantial biomass to fuel fires; if they were poorly represented, it could be that fires may have been less frequent and/or not as hot, perhaps allowing the development of the savannah oaks, at least in some areas. Post-settlement samples record a sharp increase of 16 taxa of weedy disturbance plants (Baker et al., 1993). Most of the remains of native vegetation of all habitats disappear from the record.

In contrast to the above sequence, Kirchner Marsh, also along the presettlement prairie-forest ecotone, remained in prairie throughout the mid-Holocene (Watts and Wright, 1967; Wright et al., 1963). There, tree pollen declined sharply about 9000 years BP, as prairie taxa begin to arrive (Figure 2). Prairie was well established by 8000 years BP, and was dominant until about 3000 years BP. Although this marsh/lake sequence provided little macrofossil evidence of upland vegetation, the wetland and aquatic seeds combined with the generally upland pollen assemblage still provided a reasonable reconstruction of Holocene climatic and vegetational change. A stalagmite from Spring Valley Cave (Denniston et al., 1999), west of the ecotone (Figure 1), shows a sharp change in $\delta^{13}\text{C}$ (solid line), indicating a change from C_3 to C_4 grasses about 8000 years BP, reverting to C_3 plants about 3000 years BP. Baker et al. (1992) suggest that dominant presence of dry air from the Pacific moved eastward across northern and central Minnesota between 9000 and 8000 radiocarbon years BP, but was blocked by warm, moist air from the Gulf of Mexico in eastern Iowa and southeastern Minnesota. These sequences indicate that the mid-Holocene advance of prairie across southern Minnesota, as in Iowa, stalled along a sharp prairie-forest ecotone for at least 2500 years BP (Figure 1).

Southeastern Nebraska

Sites along the South Fork of the Nemaha River in Nebraska, well within the presettlement tall-grass prairie, have yielded a relatively complete Holocene record for about the last 9000 years BP (Baker, 2000; Baker et al., 2000). The early Holocene phase with forest elements is present from about 9000 to 8500 radiocarbon years BP. It contains macrofossils of deciduous trees including *Ulmus americana*, *U. rubra*, *Juglans cinerea* type, *Juglans nigra*, *Carya cordiformis*, *Carya ovata*,

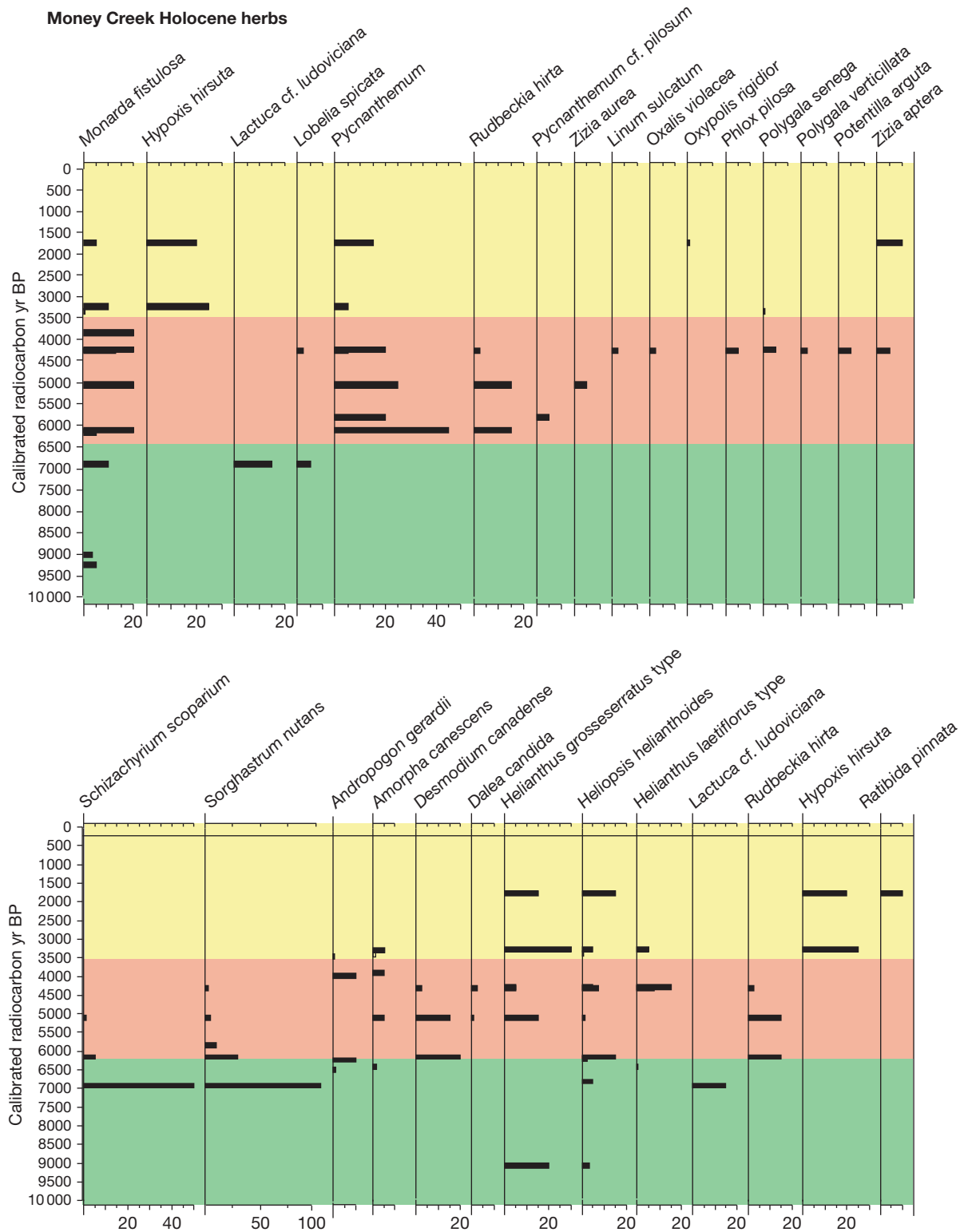


Figure 6 Holocene macrofossils from Money Creek, southeast Minnesota; prairie plants.

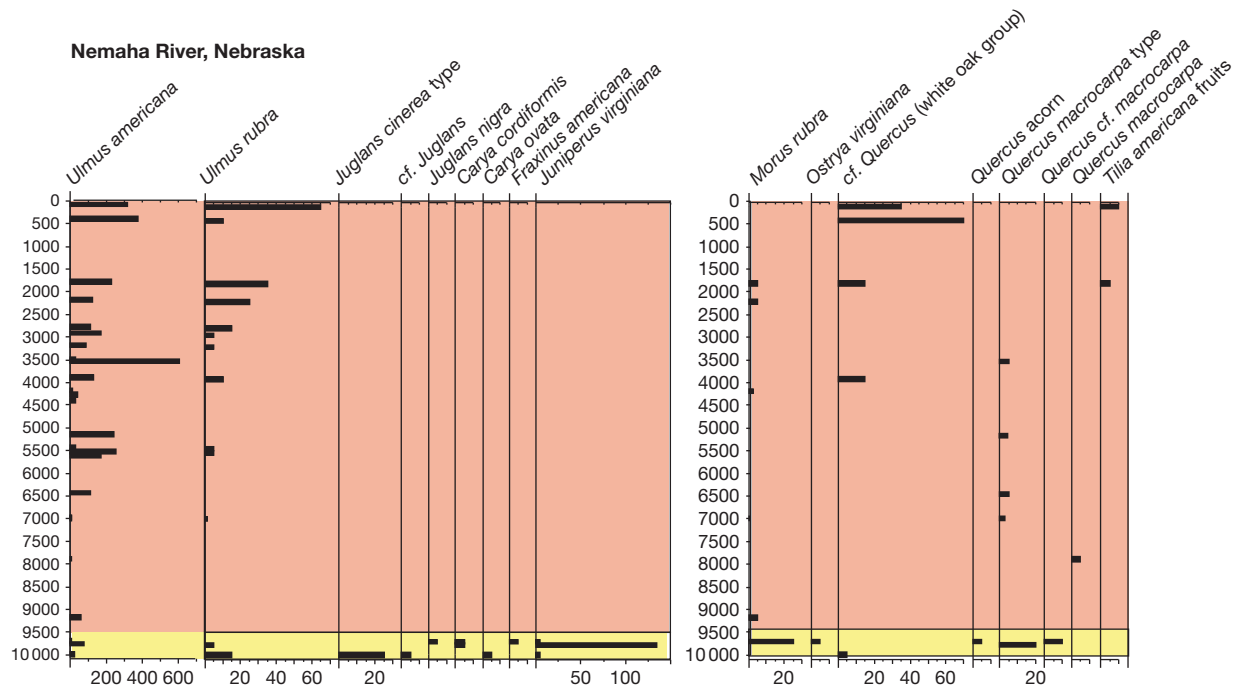


Figure 7 Holocene macrofossils from Nemaha River, southeast Nebraska; upland trees.

Fraxinus americana, *Ostrya virginiana*, *Quercus macrocarpa* type, and *Tilia americana* (Figure 7). These suggest that an upland deciduous forest was present during this part of the early Holocene. Several riparian trees, including *Acer negundo*, *Morus rubra*, and *Salix exigua* ssp. *interior*, indicate that riparian forest was also present (Baker, 2000). A limited forest understory assemblage of *Campanula americana*, *Laportea canadensis*, *Salvia lyrata*, and *Sanicula gregaria* indicates that at least some closed forest was present. The low numbers and diversity suggest that the shady understory habitat was less extensive than at the sites in Minnesota and Iowa. Also, the prairie species *Andropogon gerardii*, *Schizachyrium scoparium*, *Sorghastrum nutans*, *Sporobolus asper*, *Stipa spartea*, *Amorpha canescens*, and *Helianthus x laetiflorus* type (Figure 8), along with low arboreal pollen percentages and prairie phytoliths (Baker et al., 2000), indicate that prairie openings were extensive during this interval. It is likely that the upland deciduous trees grew in copses on mesic sites along the valley sides, and prairie grew on the uplands and in dry sandy soils, constituting a savannah-like habitat.

During the rest of the Holocene in Nebraska (Figures 7 and 8), except for the drought tolerant oaks, the upland trees disappeared, but riparian trees, including *Ulmus* spp., persisted throughout most of the last 9500 years BP. Prairie elements are abundant and diverse throughout. The prairie taxa mentioned above generally increase in numbers and other taxa are sporadically present. A particular short interval occurred from 3300–2900 years BP, when the grasses *Andropogon*, *Schizachyrium*, and *Sorghastrum* along with *Helianthus x laetiflorus* type are particularly abundant, other species like *Dalea candida*, *Lespedeza capitata*, *Desmodium canadense*, *Helianthus grosseserratus* type, and *Heliopsis helianthoides* make a showing, and even riparian-tree macrofossils disappear.

This interval also has a sharp peak in $\delta^{13}\text{C}$ values, indicating greater input of C_4 grasses (Baker et al., 2000). The importance of *Andropogon*, *Schizachyrium*, and *Sorghastrum* along with *Amorpha canescens*, suggests that this may represent a dry period. They have deep root systems and were hardly affected by drought in the 1930s (Weaver, 1968). In any event, it appears that the middle and late Holocene were dominated by prairie.

The record of post-settlement time is not rich, and the abundance of bud scales of *Ulmus americana* and *U. rubra*, and leaf-lobes of either *Quercus alba* or *Q. macrocarpa* suggest that a savannah-like environment may have become established, possibly caused by the cessation of fire after settlement.

Southwestern Indiana

Limited work in southwestern Indiana, well within deciduous forest, has resulted in only nine dated intervals (Figure 9). There are large gaps in time, but critical times are covered sufficiently to give an outline of the Holocene vegetational history.

The early Holocene has the most diverse macrofossil record of trees. *Ulmus americana*, *Carpinus caroliniana*, *Morus rubra*, and *Platanus occidentalis* have multiple occurrences and are abundant in at least some localities. Trees of southeastern or eastern distribution that occur only in the Indiana sites are *Liquidambar styraciflua*, *Fagus grandifolia*, and *Liriodendron tulipifera*. These appear in the early Holocene, but only *Liriodendron* occurs in the middle to late Holocene (where samples are few and widely spaced). Forest understory herbs that occur sporadically are *Aralia racemosa*, *Campanula americana*, *Carex backii* type, *Claytonia virginica* type, cf. *Cryptotaenia canadensis*,

Nemaha River, Nebraska
Holocene herbs

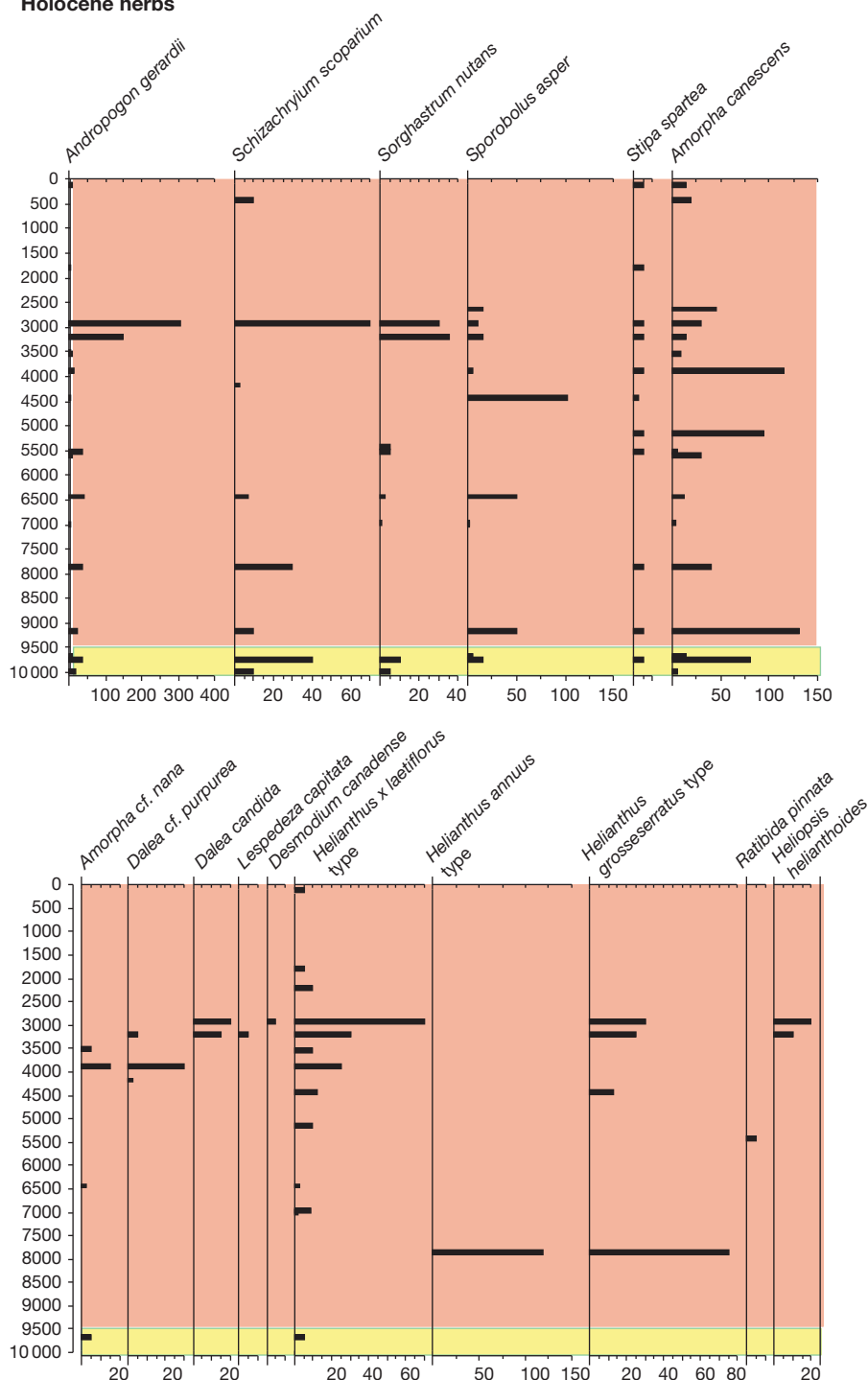


Figure 8 Holocene macrofossils from Nemaha River, Nebraska; prairie plants.

Eupatorium cf. rugosum, *Laportea canadensis*, and *Ranunculus abortivus*. Three species that could represent prairies, *Helianthus helianthoides*, *Monarda fistulosa*, and *Thalictrum dasycarpum* (not shown), all occur at 3830 years BP. From this imperfect record, the vegetation in southwestern Indiana appears to have been forested throughout the Holocene. The only evidence of some

open areas is the occurrence of the three facultative prairie species, but the absence of any of the prairie grasses or other characteristic prairie taxa, and the fact that these three species can also occur in open woodlands, suggests that prairie was not extensive, and probably was not present in this part of Indiana.

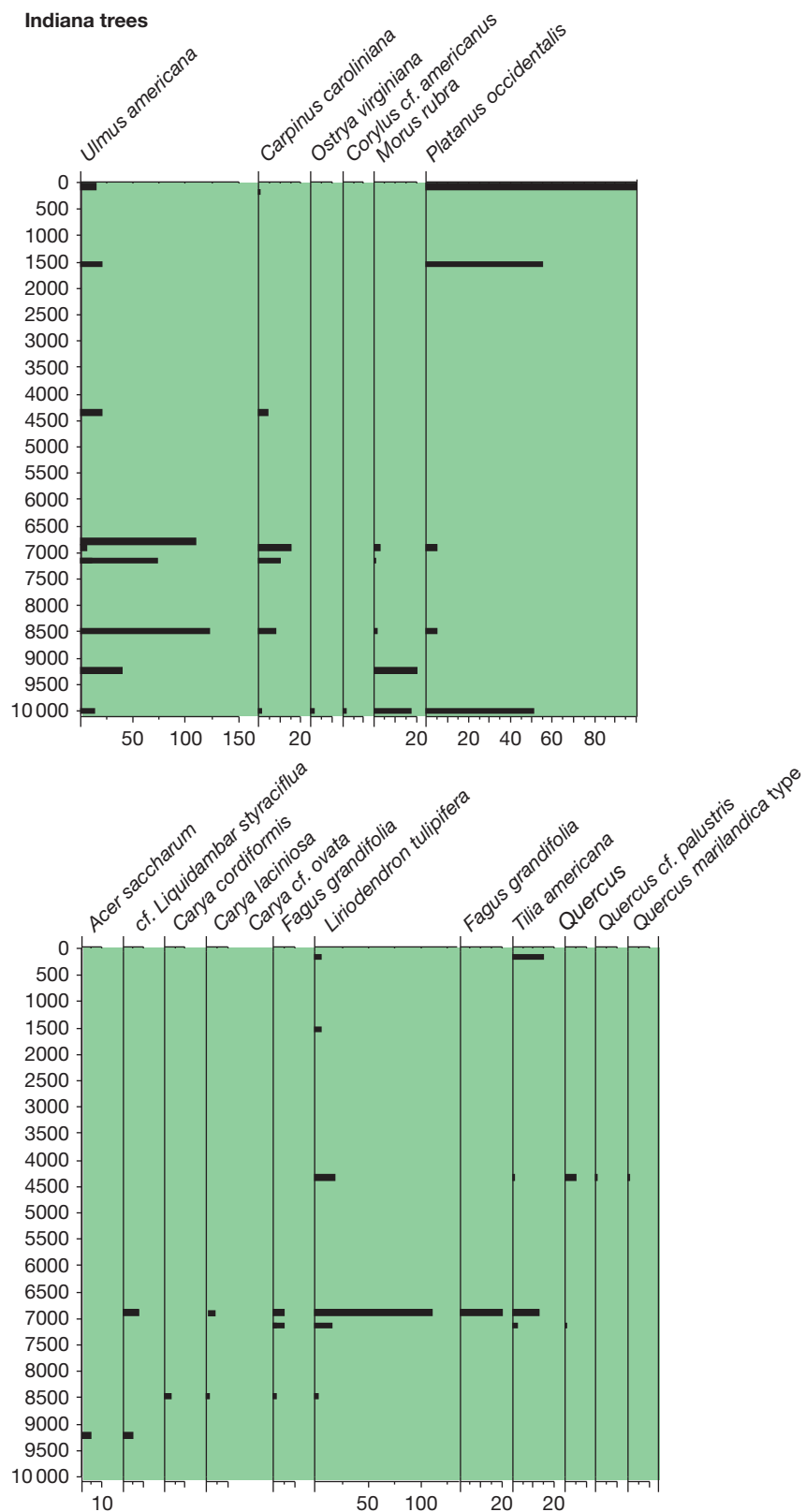


Figure 9 Holocene macrofossils from Indiana; trees.

Sac River, Missouri

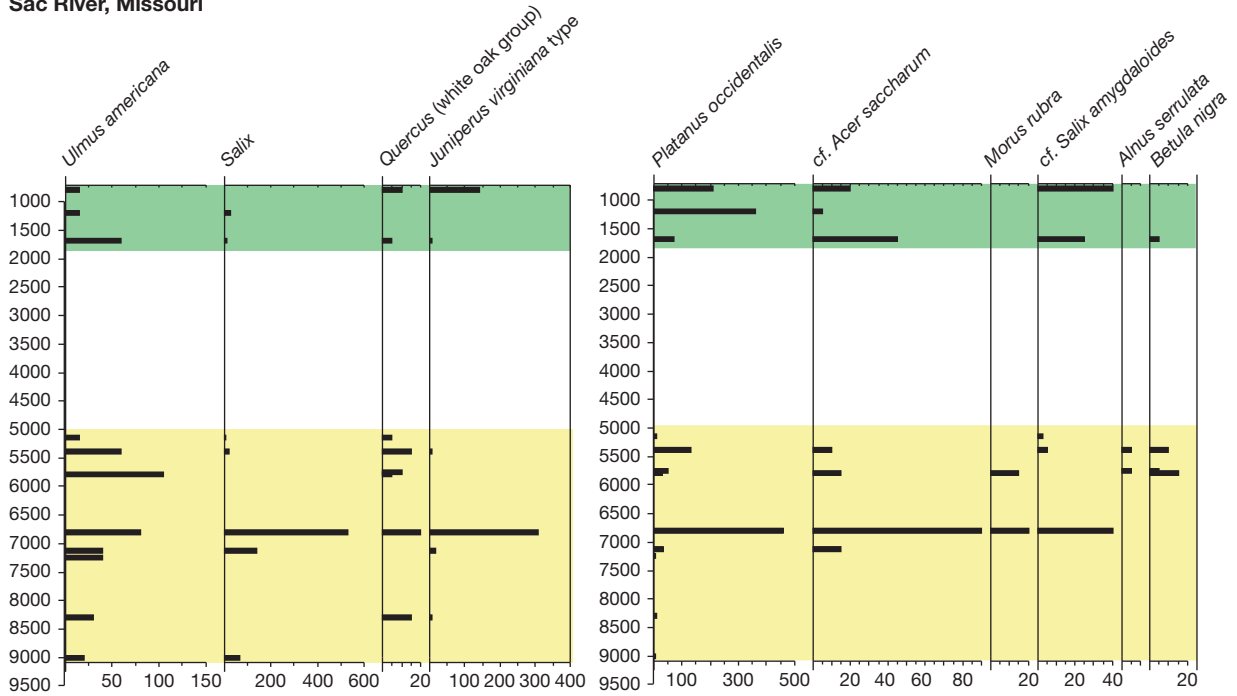


Figure 10 Holocene macrofossils from Sac River, Missouri; trees.

Sac River herbs

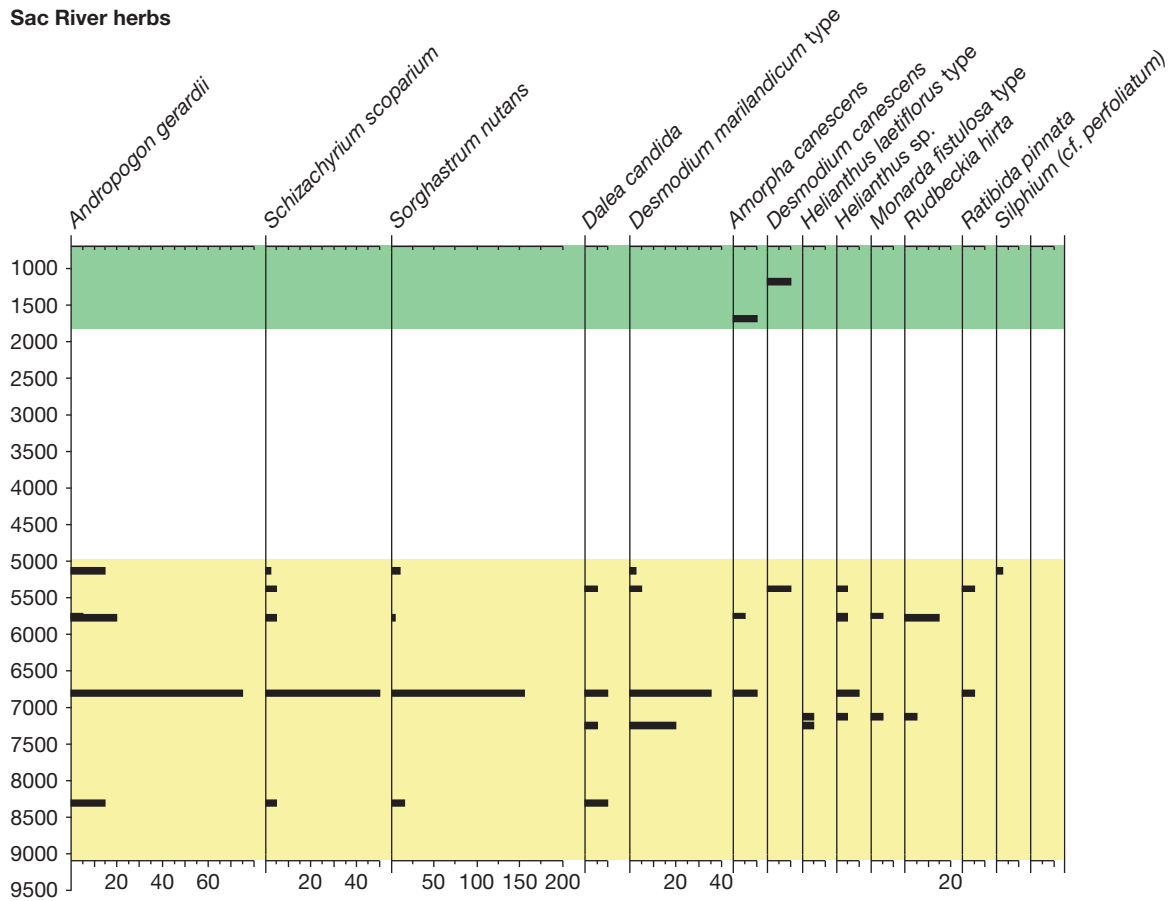


Figure 11 Holocene macrofossils from Sac River Missouri; prairie plants.

Southwestern Missouri

Preliminary work along the Sac River, southwestern Missouri (Baker et al., 2005), has yielded a complete early Holocene record (~9000 to ~5000 years BP), but a very large gap (~5000 to ~1800 years BP) exists in the late Holocene (Figures 10 and 11). The sites are located in the deciduous forest close to the southern limb of the prairie-forest border (Figure 1) and thus should be as sensitive as the Minnesota and Iowa sites, where dramatic shifts of the prairie-forest border occurred.

Only *Ulmus americana*, *Salix* sp., and *Platanus occidentalis* are present in the oldest sample (9000 years BP). By about 7000 years BP, the probable upland trees, *Quercus* (subgenus *leucobalanus*-white oak group), *Juniperus virginiana* type, and cf. *Acer saccharum* become abundant. By about 5500 years BP, the full suite of trees is present including the bottomland tree, *Betula nigra* and shrub, *Alnus serrulata*. Most of these species are also present in the latest Holocene above the gap (Figure 10).

No prairie plants were present at 9000 years BP, but between 8000 and 8500, the tall prairie grasses *Andropogon gerardii*, *Schizachyrium scoparium*, and *Sorghastrum nutans* appear along with the prairie legume, *Dalea candida* (Figure 11). By about 7000 years BP, a full suite of prairie species are present, but by the late Holocene, most of the prairie species have disappeared. It appears that the prairie-forest border was close to its present position throughout the early Holocene. There appear to be no large swings toward either prairie or forest during the period of coverage. The pollen record by Moss also suggests no major shift in the prairie-forest border (Baker et al., 2005). The late Holocene looks to have been relatively forested, perhaps with a few openings. It is possible that major shifts occurred during the late Holocene gap.

Reconstruction of the Prairie-Forest Ecotone

1. The sequences in northeastern Iowa and southeastern Minnesota are similar, with a deciduous forest in the early Holocene, rapid change to diverse dry to mesic prairie about 5500 years BP, and a gradual change to oak savannah about 3000 years BP. The timing of these changes contrasts with sites in southeastern Minnesota a short distance north and west, where prairie arrived about 2500–3000 years earlier. A stable mid-Holocene ecotone between prairie on the west and forest on the east is postulated from about 8000–5500 years BP.
2. In sites within the prairie in eastern Nebraska, prairie replaced mixed open and closed forest about 9000 years BP, with relatively minor shifts in prairie types thereafter. Riparian forests were present throughout, except for a short interval from 3300–2900 years BP, which could represent a drier period.
3. Southwestern Indiana was apparently forested throughout the Holocene, including trees that are mainly eastern or southeastern in distribution. A slightly drier period with more forest openings about 3800 years BP is possible.
4. Southwestern Missouri supported both prairie and forest species in an open forest or savannah environment

throughout the early Holocene, apparently with no significant movement of the prairie forest border.

5. These records demonstrate that a strong contrast existed in the Holocene history of the southern and northern limb of the prairie-forest border. Major shifts from forest to prairie occurred in northern sites. Although there is presently a gap in coverage of middle to late Holocene, it appears that the prairie-forest border in the south was relatively stable.
6. The apparently sudden shift to prairie in southeastern Minnesota and northeastern Iowa was preceded by the establishment of a few pioneer prairie species in forest openings. The rapid disappearance of trees and establishment of many prairie species represents the demise of the forest and the establishment of the tall-grass prairie, heralded by several species in the dominant families Asteraceae, Fabaceae, and Poaceae.
7. The use of macrofossils from rarely studied river-alluvium, supplemented by pollen and stable isotope analyses, have demonstrated previously undetected phytogeographic changes in forest-prairie distribution that varied considerably in relative position throughout the Holocene in the American Midwest. The identification of macrofossils to species level has allowed an impressively detailed reconstruction of vegetation and habitats in areas where a pollen record would be extremely general and much less informative.

See also: Plant Macrofossil Introduction. Carbonate Stable Isotopes: Speleothems. Plant Macrofossil Methods and Studies: Validation of Pollen Studies.

References

- Baker RG (2000) Holocene environments reconstructed from plant macrofossils in stream deposits from southeastern Nebraska. *The Holocene* 10: 357–365.
- Baker RG and Drake P (1994) Holocene history of prairie in Midwestern United States: Pollen versus plant macrofossils. *Ecoscience* 1: 333–339.
- Baker RG, Bettis EA III, and Moss P (2005) Holocene pollen and plant macrofossils in the Sac River Valley. In: Lopinot NH, Ray JH, and Conner MD (eds.) *Regional Research and the Archaic Record at the Big Eddy Site (23CE426)*, pp. 32–137 Southwest Missouri.
- Baker RG, Bettis EA III, Denniston RF, and Gonzalez LA (2001) Plant remains, alluvial chronology, and cave speleothem isotopes indicate abrupt Holocene climatic change at 6 ka in Midwestern USA. *Global and Planetary Change* 28: 285–291.
- Baker RG, Bettis EA III, Gonzalez LA, Denniston RF, Strickland LE, and Krieg JR (2002) Holocene paleoenvironments in southeastern Minnesota – chasing the prairie-forest ecotone. *Palaeogeography, Palaeoclimatology, Palaeoecology* 177: 103–122.
- Baker RG, Bettis EA III, Schwert DP, et al. (1996) Holocene Paleoenvironments of northeast Iowa. *Ecological Monographs* 66: 203–234.
- Baker RG, Fredlund GG, Mandel RD, and Bettis EA III (2000) Holocene Environments of the Central Great Plains. *Quaternary International* 67: 75–88.
- Baker RG, Gonzalez LA, Raymo M, Bettis EA III, Reagan MK, and Dorale JA (1998) Comparison of multiple proxy records of Holocene environments in Midwestern USA. *Geology* 26: 1131–1134.
- Baker RG, Schwert DP, and Bettis EA III (1993) The impact of EuroAmerican settlement on landscapes, vegetation and water quality in northeastern Iowa. *Holocene* 3: 314–323.
- Birks HH (1980) Plant Macrofossils. In: Birks HJB (ed.) *Quaternary Palaeoecology ch.5*, p. 66. London: Edward Arnold (Publishers) Limited.
- Birks HH and Birks HJB (2003) The importance of plant macrofossils in the reconstruction of Late-glacial vegetation and climate: Examples from Scotland, western Norway, and Minnesota, USA. *Quaternary Science Reviews* 22: 453–473.

- Denniston RF, Gonzalez LA, Baker RG, et al. (1999) Speleothem evidence for Holocene fluctuations of the prairie-forest ecotone, north-central USA. *The Holocene* 9: 671–676.
- Dorale JA, Gonzalez LA, Reagan MK, Pickett DA, Murrell MT, and Baker RG (1992) A high-resolution record of Holocene climate change in speleothem calcite from Cold Water Cave, Northeast Iowa. *Science* 258: 1626–1630.
- Jackson ST, Overpeck JT, Webb T III, Keatts SE, and Anderson KH (1997) Mapped plant macrofossil and pollen records of Late Quaternary vegetation change in eastern North America. *Quaternary Science Reviews* 16: 1–70.
- Rosendahl CO (1948) A Contribution to the Knowledge of the Pleistocene Flora of Minnesota. *Ecology* 29: 284–315.
- Strickland LE (1998) *Taphonomy of Plant Remains in Prairie Creek, Nerstrand Big Woods State Park, Rice County, Minnesota: A Modern Analog for Paleoenvironmental Interpretation*. M.S. Thesis, Department of Geoscience, University of Iowa, p. 111.
- Stuiver M, Reimer PJ, and Reimer RW (2005) CALIB. <http://calib.qub.ac.uk/>.
- Watts WA and Winter TC (1966) Plant macrofossils from Kirchner Marsh, Minnesota – a paleoecological study. *Geological Society of America Bulletin* 77: 1339–1359.
- Weaver JE (1968) *Prairie Plants and Their Environment*. pp. 59 and 148–149. Lincoln and London: University of Nebraska Press.
- Work PT, Semken HA, and Baker RG (2005) Pollen, plant macrofossils, and microvertebrates from mid-Holocene alluvium in east-central Iowa, USA: Comparative taphonomy and paleoecology. *Palaeogeography, Palaeoclimatology, Palaeoecology* 223: 204–221.
- Wright HE Jr., Winter TC, and Patten HL (1963) Two pollen diagrams from southeastern Minnesota: Problems in the regional late-glacial and postglacial vegetational history. *Geological Society of America Bulletin* 74: 1371–1396.

Late Glacial Multidisciplinary Studies

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Abiotic and biotic variables can reflect ecological processes and their changes on several temporal and spatial scales. The reconstruction of environmental changes in the Late Glacial period is of special interest because the changes in climate and hence in climate-dependent responses of these variables were both rapid and of large amplitude.

Organisms and environmental gradients show interesting relationships. Therefore, biostratigraphies providing qualitative and quantitative changes of taxa are time series of the behavior of various types of organisms and also of environmental change. Generally, species show responses to ecological gradients by exhibiting optima and tolerances that reflect population size or productivity or some other measure of fitness. They are often unimodal. We use three examples to illustrate how these relationships can be reconstructed in three different ways: qualitative reconstruction of past environmental changes; quantitative reconstruction of past climatic changes; and estimation of the biotic response to climate change, which is possible only if an independent line of evidence for the climate change is available.

Time control is essential for all estimates of rates of change. However, in the Late Glacial period (ca. 16 000–11 500 cal yr BP) time control can be hampered by (1) the absence of organic material suitable for radiocarbon dating, (2) the presence of reworked organic material, (3) plateaus of constant radiocarbon ages, and (4) problems of calibrating the radiocarbon time-scale. Dating techniques other than radiocarbon (e.g., several geophysical techniques) may have large errors or are applicable only at special sites (e.g., annual sediment layers or varves).

Among terrestrial ecosystems, the vegetation (i.e., the primary producers) may provide the best reconstruction of the past climate. Aquatic ecosystems may be of special interest because water plants and aquatic invertebrates (e.g., insects or Cladocera) can respond rapidly to climatic warming or cooling for at least three reasons: Their dispersal of propagules (seeds, fruits, resting stages, or adults) may be fast because of the high mobility of the main vector (e.g., water fowl); their life cycles are relatively short; and they have no special requirements, such as mature soils.

The complementarity of pollen and plant macrofossil analyses may be especially useful for the reconstruction of past environments. Identification of plant macrofossils can often go to the species level (instead of only the genus or family level; see the example of dwarf birch and tree birches), and local presence can be assumed if heavy macroremains are found, whereas pollen may be more easily transported from a long distance. Both aspects may apply to terrestrial and aquatic plants.

Qualitative Reconstruction of Past Environments

Multidisciplinary studies are most precise if all analyses can be made on a single core. Some studies, however, need relatively large volumes of sediment per sample (e.g., beetle analysis); in such cases, several cores and their correlation may be needed. Alternatively, open exposures may provide large sample volumes. Beetles may offer additional information, such as habitat structure or openness of vegetation.

We present two examples of the qualitative reconstruction of past Late Glacial environments, one indicating primarily changes in temperatures and the other indicating changes in both effective moisture (lake level) and temperature.

The Late Glacial in Switzerland

The biostratigraphic changes during the Late Glacial on the Swiss Plateau (the lowland between the Jura Mountains and the Alps) are summarized in [Figure 1](#) and [Table 1](#). Two phases of strong and rapid warming occur at approximately 14 500 and 11 500 cal yr BP (i.e., at the beginning of the Bølling and beginning of the Holocene). They reflect global or hemispheric pulses of climatic warming. The Oldest Dryas on the Swiss Plateau can usually be subdivided into three phases. At the base, the pollen concentration is low, plant macroremains are absent, and many reworked pollen grains are present, indicating the dominance of open ground and scarce vegetation after deglaciation. The second phase reflects a species-rich steppe–tundra (sometimes subdivided into two subphases) ([Caillard, 1984](#)). Then a marked increase in the birch pollen curve can be attributed to a higher abundance of dwarf birch (*Betula nana*); the species identification is based primarily on plant macrofossils (a few remains of tree birches are also found). The sharpest transition for the entire Late Glacial occurs at approximately 14 500 cal yr BP at the beginning of the Late Glacial interstadial (Bølling). The plant macrofossils show a change from dominant dwarf birch to dominant tree birch (the last remains of dwarf birch are found within the *Juniperus*–*Hippophaë* regional pollen zone; the Linnean species *Betula 'alba'* includes several species of tree birches). At the same time as this ecologically important period of afforestation, the assemblages of all groups of aquatic organisms change ([Figure 1](#)). Most inferences of biotic changes involve steps in climate warming (and cooling at the beginning of the Younger Dryas). However, related, climate-induced conditions are important, such as pedogenesis, trophic state, and seasonal circulation patterns in lakes (affecting oxygen conditions). The forested Late Glacial is first dominated by tree birches (Bølling) and then by pine and tree birches (Allerød). Minor fluctuations within this Late Glacial interstadial can only be traced if the sites

Multidisciplinary studies in the Late Glacial of the Swiss plateau

Several sites on the Swiss plateau						Lobsigensee on the Swiss plateau					
Cal years BP	Bio zones	Regional pollen zone	Plant macro fossils	Inferred vegetation	$\delta^{18}\text{O}$ lit. 4	Sediment	Chironomids and chaoborus	Cladocera	Coleoptera and trichoptera	Mollusca	Ostracods and pigments: O_2
	Preboreal	Pinus-corylus, QM		Birch-pine forest, hazel							Strictly anoxic
11 500					T↑					T↑	
	Younger Dryas	Pinus-gramineae-artemisia (some Ephedra)	Less Pinus, Betula dominant	Open pine forest with birch, larch, poplar		Lake marl	Chironomid assemblage dominated by littoral taxa, no taxa indicative of oligotrophic conditions	Bosmina longispina↓, Bosmina longirostris	No faunal changes but lower abundances	Number of individuals↑	meromixis and/or temporarily anoxic
12 000										Minimum number of individuals	
12 500										T↓	
12 650					T↓						
13 000	Allerød	Pinus-betula	Pinus sylvestris, Betula "alba"	Pine birch forests					Temperate fauna 14–16°C mean July temperature	Number of individuals↓	
13 500											
13 600											
14 000	Bølling	B. "alba" Salixartemisia B. "alba"	Betula pubescens, B. pendula Populus tremula	Birch forests			Cold-stenotherm species absent	Sub-arctic species absent	Temperate fauna	Maximum number of individuals	
14 500		Junip.-Hippophaë	Afforestation		T↑	Transit	Faunal change	Faunal change	Faunal change	T↑	
15 000	Oldest Dryas 3 rd phase	Artemisia-betula nana	Betula nana, Salix sp.	Shrub-tundra Salix, B. nana		Silt and clay	Many cold-stenotherm species present	Sub-arctic species present	Boreal and boreo-montan fauna, 10–12°C mean July	Valvata piscinalis↑, pisidium↓	No meromixis
15 500?											
16 000	Oldest Dryas 2 nd phase	Artemisia-helianthemum cyperaceae	Selaginella selaginoides, Gypsophila repens, Potentilla, Onobrychis...	Steppe-tundra, species-rich							
16 500?											
17 000?	Oldest Dryas 1 st phase	Artemisia-pinus and reworked pollen		Open ground, some pioneers		Transit Sand, till					

Figure 1 Overview of bio- and isotope stratigraphies of the Late Glacial on the Swiss Plateau. Ecologically, the most remarkable change from the shrub-tundra in the Oldest Dryas to a juniper-birch forest in the early Bølling (i.e., a shift in dominance from dwarf birches to tree birches) coincides with the rapid shift in oxygen isotopes of at least 3 around 14 500 cal yr BP. It is accompanied by changes in the assemblages of insects (chironomids, Chaoborus, Coleoptera, and Trichoptera) and Cladocera as well as a change in abundances among mollusks. The subdivision of the Oldest Dryas based on pollen and plant macrofossils (see text for details) is also reflected in the biostratigraphies of the invertebrates: First occurrences coincide with the stepwise development of the vegetation. The rapid cooling at the beginning and the rapid warming at the end of the Younger Dryas are recorded in the oxygen isotope ratios and all biostratigraphies. Data from Ammann (1989), Gaillard (1984, 1994), Lotter et al. (1992), and Richoz (1997); the references for Lobsigensee are given in Table 1.

are at sensitive altitudes and if the sampling resolution is sufficiently high. The Younger Dryas cool period is reflected in the lowland by an opening of the forests, when heliophilous taxa such as *Larix* and *Populus* are recorded as plant macrofossils, as well as pollen of *Artemisia* and *Ephedra*. At higher altitudes, a lowering of the timberline was registered.

The Pleniglacial and Late Glacial in Lake Zeribar, Zagros Mountains, Iran

Paleoecological changes between ca. 18 000 and 6000 cal yr BP in the Mediterranean climate area of western Iran are illustrated by different proxy data obtained from a single sediment core from Lake Zeribar, Zagros Mountains, Iran (Figure 2). A pollen diagram (van Zeist and Bottema, 1977) illustrates the dominance of *Artemisia*-Chenopodiaceae steppe-like vegetation in the Pleniglacial and Late Glacial until ca. 12 000 cal yr BP. Trees were absent or limited to a few sheltered places until ca. 15 400 cal yr BP (Heinrich 1 cool oscillation). Their number increased slightly in the Bølling/Allerød interstadial, but it was reduced again during the Younger Dryas stadial. A slight increase in humidity approximately 12 000 cal yr BP favored the expansion of Poaceae-

dominated steppe-like vegetation, but the climate was still too dry for an extensive spread of trees. Until approximately 7000 cal yr BP, the drought-resistant *Pistacia* was the major tree; between ca. 7000 and 6000 cal yr BP the climate became sufficiently humid for oak woodland to become established.

Evidence of fluctuations in water levels and in salinity is obtained from various proxies in the lake (Figure 2). In the macrofossil record (Wasylikowa, 2005), peaks of *Chenopodium rubrum*, which typically expands on seasonally exposed lake-shores, indicate episodes of fluctuating and/or low lake water levels. Two cold episodes (Heinrich 1 and Younger Dryas) were also characterized by increased salinity, as indicated by the presence of halophytic macrophytes, *Ruppia maritima*, *Salicornia europaea*, and *Suaeda* sp. (Wasylikowa 2005); by peaks of brackish water and marine diatoms (Figure 2; Wasylikowa et al., 2006); and by high conductivity values (Snyder et al., 2001). Among the submerged aquatics, *Potamogeton pectinatus* expanded. It tolerates brackish water. It can also withstand short-lasting emergence and it became abundant at the same time as *C. rubrum* (Wasylikowa, 2005). The cause of the lower lake levels was probably dry climate, as indicated by elevated $\delta^{18}\text{O}$ values (Stevens et al., 2001). Higher lake water levels

Table 1 Multidisciplinary studies at Lobsigensee included in [Figure 1](#) for the environmental reconstruction of the Late Glacial periods

- Ammann, B. (1985) Introduction and Palynology: vegetation history and core correlation at Lobsigensee (Swiss Plateau). *Dissertationes Botanicae* 87: 127–135.
- Ammann, B. (1989) Late-Quaternary palynology at Lobsigensee – Regional vegetation history and local lake development. *Dissertationes Botanicae* 137: 1–157.
- Ammann, B., Chaix, L., Eicher, U., Elias, S.A., Gaillard, M.-J., Hofmann, W., Siegenthaler, U., Tobolski, K., Wilkinson, B. (1983) Vegetation, insects, molluscs, and stable isotopes from Late Würm deposits at Lobsigensee (Swiss Plateau). *Revue de Paléobiologie* 2: 221–227.
- Ammann, B., Tobolski, K. (1985) Vegetational development during the Late-Würm at Lobsigensee (Swiss Plateau). *Revue de Paléobiologie* 2: 163–180.
- Chaix, L. (1985) Malacofauna at Lobsigensee. *Dissertationes Botanicae* 87: 148–150.
- Elias, S.A., Wilkinson, B. (1983) Late Glacial fossil insect assemblages from Lobsigensee (Swiss Plateau). *Revue de Paléobiologie* 2: 189–204.
- Elias, S.A., Wilkinson, B. (1985) Fossil assemblages of Coleoptera and Trichoptera at Lobsigensee. *Dissertationes Botanicae* 87: 157–162.
- Gaillard, M.-J. (1983) On the occurrence of *Betula nana* L. pollen grains in the Late-glacial deposits of Lobsigensee (Swiss Plateau). *Revue de Paléobiologie* 2: 181–188.
- Hofmann, W. (1985) Developmental history of Lobsigensee: subfossil Cladocera (Crustacea). *Dissertationes Botanicae* 87: 150–153.
- Hofmann, W. (1985) Developmental history of Lobsigensee: subfossil Chironomidae (Diptera). *Dissertationes Botanicae* 87: 154–156.
- Löffler, H. (1986) An early meromictic stage in Lobsigensee (Switzerland) as evidenced by ostracods and Chaoborus. *Hydrobiologia* 143: 309–314.
- Siegenthaler, U., Eicher, U. (1985) Stable isotopes of carbon and oxygen in carbonate sediments of Lobsigensee. *Dissertationes Botanicae* 87: 162–164.
- Tobolski, K. (1985) Plant macrofossils from Lobsigensee. *Dissertationes Botanicae* 87: 140–143.
- Züllig, H. (1985) Carotinoids from plankton and phototrophic bacteria in sediments as indicators of trophic changes: evidence from the Late-Glacial and the early Holocene of Lobsigensee. *Dissertationes Botanicae* 87: 143–147.

during the Bølling/Allerød interstadial are indicated by the disappearance of *C. rubrum* and the spread of *Najas marina*, which tolerates brackish conditions but, unlike *P. pectinatus*, is sensitive to desiccation. A slight increase in temperature is suggested by the appearance of *Ceratophyllum demersum*, a relatively thermophilous aquatic species. In the Holocene, a series of low lake level episodes occurred between ca. 10 000 and 6300 cal yr BP, but unlike the earlier episodes, the water was fresh and salinity indicators were absent. The curve of the noncombustible mineral fraction in the loss-on-ignition diagram indicates that this was the time of intensified catchment erosion ([Stevens et al., 2001](#)). The replacement of the dominant *Najas marina* by *N. minor* approximately 11 000 cal yr BP could reflect the increase in temperature and decline of salinity.

Quantitative Reconstruction of Climate Change in the Late Glacial: The Role of Plant Macrofossils

Analog or Indicator Species Approaches

If the ecological and environmental tolerances of a taxon are well-known, either through observation or experiment, the

past ecological and environmental conditions can be inferred if the taxon is found as a fossil or subfossil; it can thus be used as an indicator species. Similarly, the environmental tolerances of an assemblage can be inferred from a modern analogue; the analog assemblage approach applies particularly well to vegetation reconstruction. If types of vegetation can be reconstructed from fossil plant assemblages, past habitats and past temperature and humidity can be reconstructed. Plant macrofossils are well suited to the analogue approach to environmental and climatic reconstruction. Past vegetation can be reconstructed in some detail, mainly because of the rather local deposition of plant macrofossils and the possibility of identifying many plant macrofossils to a detailed taxonomic level, often to species.

When pollen data were used to reconstruct the spatial and temporal distribution of vegetation types and the ecotones between them in the Late Glacial of western Norway ([Birks, 1994](#)), a southern and western distribution of birch trees was inferred during the interstadial. The ability of macrofossils to refine pollen data was then used to distinguish between tree birch (e.g., *Betula pubescens*) and dwarf birch (*B. nana*). The macrofossil record indicated that the Late Glacial interstadial of western Norway lacked birch trees. All the *Betula* pollen was derived from trees to the south by long-distance transport, or it came from *B. nana*, which was shown by macrofossil records to be present in some abundance inland but absent along the Norwegian coast ([Birks, 2003](#)). Therefore, temperature and ecotone reconstructions made on the basis of the pollen-inferred occurrence of tree birch had to be revised. Late Glacial assemblages of plant macrofossils in western Norway closely resemble the modern sub-Alpine and Alpine vegetation. The mean July temperature ranges of these modern vegetation zones are known and can thus be inferred for the past. The new interpretations are supported by the entire macrofossil assemblages, which indicate a treeless low-Alpine or mid-Alpine vegetation ([Birks, 1993, 2003](#)).

A special case of the analog approach is the mutual climate range method, used especially with Coleoptera data. By overlapping the modern climate ranges of species within a fossil assemblage, the most likely climate envelope for that assemblage is narrowed down. This approach has been applied with greater mathematical rigor to plant assemblages, particularly macrofossil assemblages with species determinations (probability density functions) ([Kühl et al., 2002](#)). It has not been applied to Late Glacial plant assemblages, but it shows promise as an alternative approach to modern analogs and transfer functions.

Multivariate Approach to Climate Reconstruction

This approach to climate reconstruction involves the use of large modern data sets of organisms that are related to their present environment by a transfer function. The transfer function is applied to fossil data to reconstruct past environments and climates. In the Late Glacial, climate transfer functions are most often used to reconstruct mean July (T_{jul}) or summer temperatures. They have been applied in Late Glacial studies in North America ([Levesque et al., 1994](#)) and Europe ([Lotter, 2000](#); [Brooks and Birks, 2000](#)).

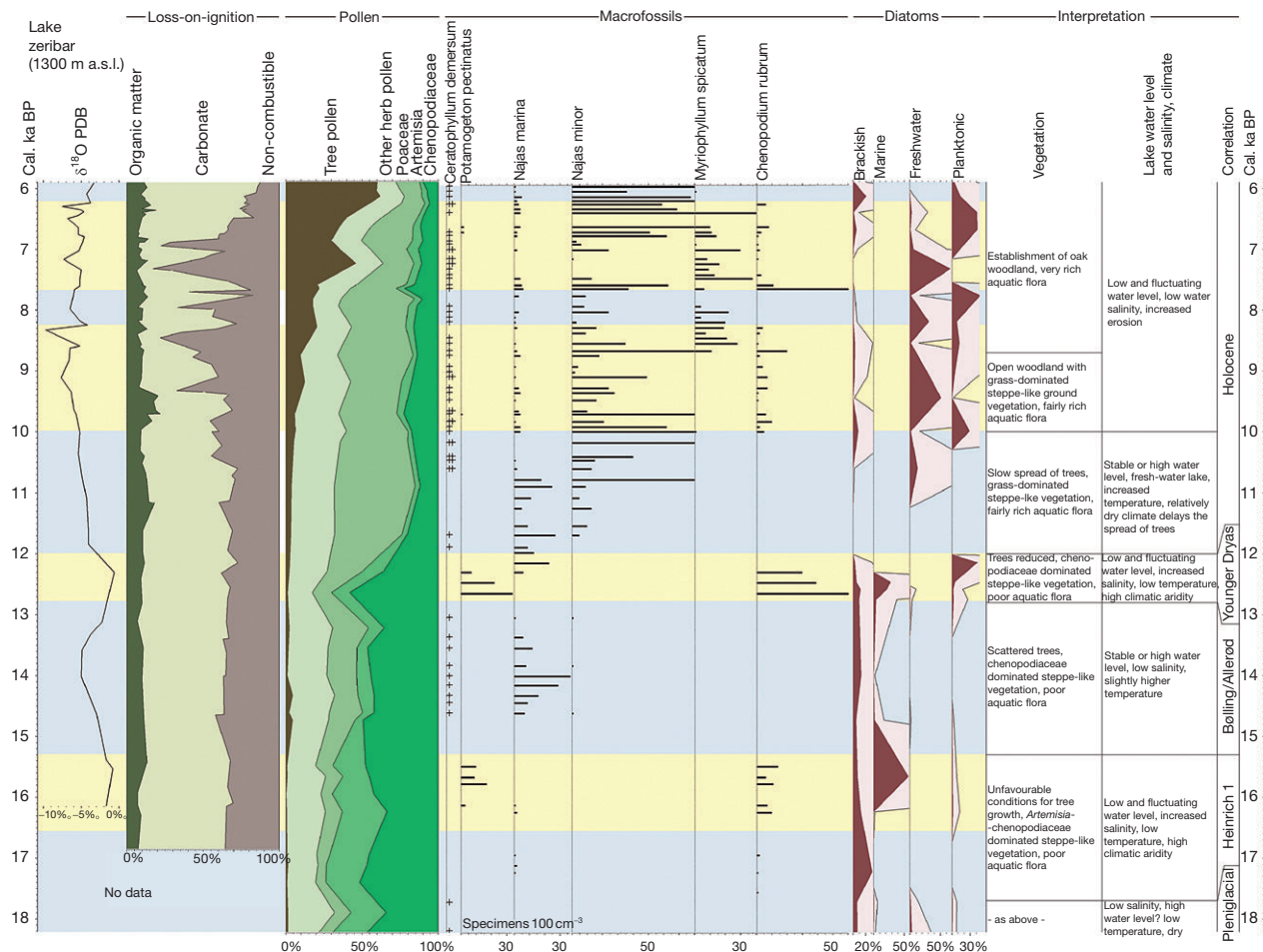


Figure 2 Multiproxy evidence of environmental change at Lake Zeribar, Iran. Yellow zones indicate the episodes of low and fluctuating lake levels.

Temperature Reconstructions in the Late Glacial at Kråkenes, Western Norway: An Example of a Multiproxy Climate Reconstruction

At Kråkenes, western Norway, all of the previously discussed approaches were applied to several data sets from the same Late Glacial sediment core to reconstruct mean July temperature (T_{jul}). The transfer function approach was used on pollen, chironomid, and Cladocera data; the MCR method was applied to the Coleoptera data (but the data were too sparse to make a continuous curve); and the analog approach was used on the plant macrofossil data. The results are compared in **Figure 3** (Birks and Ammann, 2000).

The shapes of the temperature curves are all similar. The Allerød interstadial temperatures were cool, ca. 3–6 °C below the present T_{jul} . All the proxies show a decrease into the Younger Dryas of 1 or 2 °C. This small decrease, probably combined with increased precipitation and snow accumulation in the cirque behind the lake, was sufficient to lead to the formation and advance of a cirque glacier that drained directly into the lake, forming glaciolacustrine sediments. The proxies all show a rapid temperature rise into the early Holocene, reaching modern temperatures after approximately 1000 years (**Figure 3**).

A more detailed comparison of the results in **Figure 3** shows that the absolute temperature values differ between

proxies. The transfer function reconstructions (pollen, chironomids, and Cladocera) differ largely because of features of the modern data sets. Modern pollen assemblages from arctic or alpine situations contain long-distance transported pollen of trees with high pollen productivity and dispersal (*Betula*, *Pinus*, and *Alnus*) that was largely absent in the Late Glacial. Modern chironomid assemblages have lower taxonomic diversity below approximately 9 °C T_{jul} (e.g., on Svalbard) and therefore cannot easily differentiate cold temperatures. The cladoceran modern data set was compiled for Switzerland, where the reconstructions from Cladocera showed overall higher temperature (Birks and Ammann, 2000). The analog reconstructions from the plant macrofossil data are probably the most reliable, although they are ranges rather than single values for each sample, representing the ranges of the low Alpine (10–7.5 °C), the mid-Alpine (7.5–5 °C), and the high Alpine (<5 °C) vegetation zones of today.

The vegetation reconstructions from the macrofossil data are clear; all taxa conform to the proposed modern analog, and none disagrees. Within each vegetation zone, communities can be differentiated—for example, those related to snow beds or winter exposure, snow meltwater flushes, unstable stony ground, grassland, and dwarf-shrub heath. These mosaics also conform to present-day vegetation community mosaics

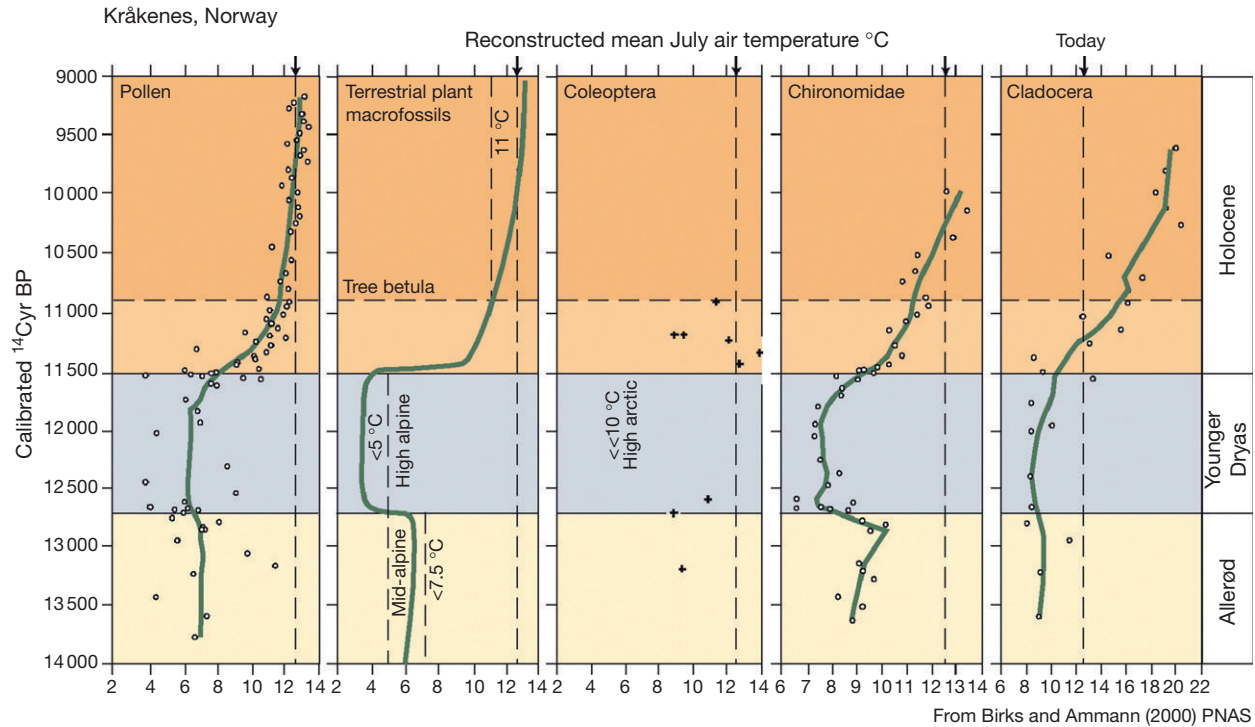


Figure 3 Comparison of mean July air temperature reconstructions at Kråkenes, western Norway. Pollen, chironomids, and Cladocera reconstructions were made with transfer functions. Coleoptera reconstructions were made with the MCR method. The plant macrofossil reconstructions were made by comparison with modern analog assemblages. From Birks HH, and Ammann B (2000) Two terrestrial records of rapid climatic change during the glacial–Holocene transition (14 000–9 000 calendar years B.P.) from Europe. *Proceedings of the National Academy of Sciences, USA* 97: 1390–1394.

Table 2 Comparison of Late Glacial mean July temperature (T_{jul}) reconstructions using different proxies and methods, the temperature changes between the Late Glacial periods, and the rates of these temperature changes at Kråkenes, western Norway

	Mean T_{jul} , AL (°C)	AI–YD drop (°C)	Rate per 25 years (°C)	Mean T_{jul} , YD (°C)	Holocene T_{jul} after 500 years (°C)	Rate per 25 years (°C)
Pollen	7	1	0.7	6	6	0.25
Chironomids	9–10	1.5–2.5	0.7	7.5	3.5	0.2
Cladocera	9	1		8	6	0.3
Macrofossils	~6	1	~ 1	<5	6	0.25
Coleoptera	9.6	~ 2		≪10	~ 5	

AL, Allerød; YD, Younger Dryas.

in the Alpine zones of western Norway. Mosses contributed to these reconstructions (Jonsgard and Birks, 1995), and terrestrial oribatid mites were also valuable habitat indicators (Solhøy and Solhøy, 2000).

Timing and rates of climate changes at Kråkenes

Terrestrial plant macrofossils, mostly *Salix herbacea* leaves, were used to obtain a high-resolution series of AMS ^{14}C dates between 12 700 and 10 000 ^{14}C BP. A calendar-year chronology for the upper Late Glacial and early Holocene was reconstructed by wiggle matching against the Hohenheim tree-ring data that dated the Younger Dryas–Holocene transition to 11 530 $\pm$ 40/–60 cal yr BP (Gulliksen et al., 1998). In the older Late Glacial, a calendar chronology was constructed by comparison with

^{14}C -dated laminated lake sediment series. Rates of change could then be calculated for the various groups of organisms (Birks et al., 2000). The rates of T_{jul} change shown in Figure 3 are summarized in Table 2. All the proxies show a rate of decrease from Allerød to Younger Dryas of 0.7–1 °C per 25 years. Although the actual temperature drop was small, it occurred very rapidly. The rate of temperature change during the first 500 years of the Holocene was 0.2–0.3 °C per 25 years. After the initial very rapid responses to the Holocene temperature rise, rates of change decreased as communities and soils developed and internal successional processes started, thus reducing the sensitivity of the ecosystem to large temperature changes (Birks et al., 2000). These results all hinge on the use of terrestrial plant macrofossils to obtain a high-resolution series of radiocarbon dates.

Biotic Response to Climate Change

The Concept

Two conditions need to be fulfilled if the estimation of biotic responses to climate change is the goal: Some independent line of evidence for climatic change is needed to avoid circular reasoning, and the temporal resolution must be high enough to capture the relevant biological processes.

Various groups of organisms have different life cycles, including generation times and rates of response to environmental changes (e.g., speeds of migration). If we combine several groups of organisms that provide different life histories, we may be able to separate biological processes on several time-scales. The precondition is that our time control is good enough for recording these processes.

For example, the transition from the Late Glacial to the Holocene as one of the most rapid periods of warming was studied on the Swiss Plateau and in the western Alps at two sites with a record of oxygen isotopes (Gerzensee and Leysin) and three sites without carbonates (one of them presented here; Ammann, 2000; Ammann et al., 2000; Brooks, 2000; Schwander et al., 2000; Tobolski and Ammann, 2000; von Grafenstein et al., 2000; Wick, 2000).

The Responses of Flora, Fauna, and Vegetation to Rapid Climatic Changes

Figure 4 presents a comparison between the pollen and plant macrofossil records at the transition from the Late Glacial to the Holocene at three sites at different altitudes in the western Swiss Alps: Gerzensee at 603 m a.s.l., Leysin at 1230 m a.s.l., and Zeneggen at 1530 m a.s.l. At Gerzensee and Leysin, the oxygen isotopes of the bulk carbonates reflect the changes in the summer temperatures at this transition from the Younger Dryas to the Preboreal (independent from biostratigraphies). These isotope changes were correlated with the ones in the GRIP ice core, from which the timescale was then transferred (Schwander et al., 2000). At Zeneggen, however, no autochthonous carbonates occurred in the sediment, so the organic matter of the sediment (expressed as loss on ignition at 550 °C) was used as a proxy for productivity and thus indirectly for summer temperatures. During the cooling of the Younger Dryas, the lowest site remained forested (some opening of the forest is indicated by somewhat higher percentages of *Ephedra*, *Artemisia*, *Selaginella selaginoides*, and other heliophilous herbs, along with lower concentrations of macrofossils from trees). At the two sites at higher altitudes, forests were virtually absent and only a few trees survived during the Younger Dryas. Although Zeneggen is situated nearly 300 m higher, plant macrofossils from trees are richer than at Leysin. This may be explained by the special climate in the central Alps at Zeneggen/Valais, which has a very high insolation and high continentality. Zeneggen is only 30 km from Zermatt and 40 km from Gouillé Rion, where the modern timberline reaches a maximum elevation of 2400 m, with single *Pinus cembra* specimens currently up to 2500 m a.s.l. (i.e., 500–700 m higher than the timberline made by spruce on the northern slope of the Alps, including Leysin).

Placing the results of pollen and macrofossil analyses into a broader context, Figure 5 demonstrates the biotic responses of

groups of organisms other than plants. Changes in species composition and abundances (i.e., qualitative and quantitative changes) are summarized as the scores on the first axes of either PCA or CA (depending on the gradient length). At Zeneggen, the highest site (without a GRIP timescale), there seems to be the following sequence of responses: Cladocera first, pollen second, and plant macrofossils third. The rise in organic matter seems to last longer than the overall changes in the biota recorded in the PCA (axis 1). This is an indication that the changes in species composition in vegetation and cladoceran fauna as recorded in the PCA/CA lasted shorter than the increase in productivity as recorded in the LOI. At the two sites where the transition from the Late Glacial to the Holocene is recorded by oxygen isotopes, the biotic responses are quite complex. First, several changes had already occurred in the last third of the Younger Dryas, especially among the Cladocera. Second, several groups of organisms show two or three steps of overall changes. Some of these changes are associated with both the beginning and the end of the transition as recorded by oxygen isotopes (e.g., plant macrofossils at Gerzensee). Other biotic changes may then be controlled by succession, which was triggered by climatic change.

Conclusions on Biotic Responses

On the basis of not only pollen and plant macrofossils but also groups of aquatic invertebrates, the general conclusions about biotic responses to rapid climatic warming in southern central Europe from the Swiss study are the following (Ammann et al., 2000):

- (1) Biotic lags to the isotope record of temperature increase of organisms with various life cycles were less prominent than expected for both aquatic and terrestrial taxa. Responses occur within approximately 4–16 years, often within the sampling resolution.
- (2) The duration of the biotic response can last longer (or shorter) than the period of the climatic shift recorded in the isotopes. This depends on species-specific environmental thresholds and on biogeographical features such as range limits and the process of succession.
- (3) Altitudinal differences in sites are useful to capture fine-scale changes. Large climatic shifts such as the Younger Dryas are recorded at all altitudes but small-scale changes only at sensitive altitudes (i.e., near the ecotones of the taxa present).
- (4) The nature of the biotic response to climatic change involves at least four groups of processes acting on different timescales:
 - i. Biogeographical changes include local (or global) extinction as well as immigration. Species migration is the slowest of the three processes; migration velocity depends, among other things, on generation times and the maximum distances of dispersal jumps. Migration (i.e., the shift of range limits) may occur in less than 1 year (e.g., algae, microorganisms, and some insects) or it may take millennia (e.g., trees). Species migration also depends on the distances involved and on barriers (e.g., mountains or oceans).
 - ii. Population dynamics after arrival of a species in an area involve the building up of populations. The time needed for this depends, among other things, on age

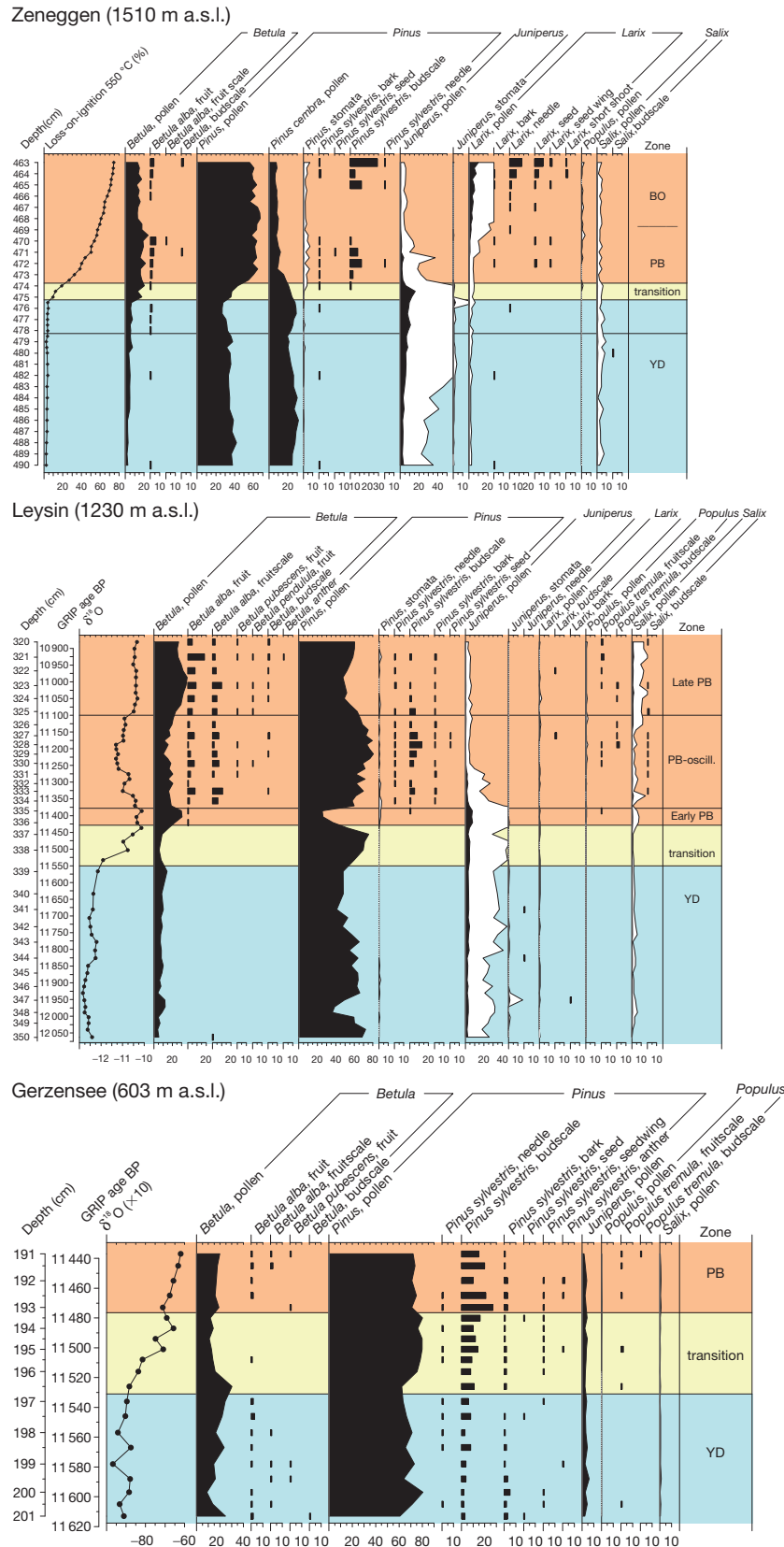


Figure 4 Altitudinal transect of pollen (curves) and plant macrofossils (histograms) as stratigraphies covering the transition from the Younger Dryas to the early Holocene (preboreal); Gerzensee and Leysin presented on a GRIP timescale. At Gerzensee, the forest did not disappear during the Younger Dryas but was more open than before and after. At Leysin and Zeneggen, only a few trees survived the cool period of the Younger Dryas. Timescale after Schwander et al. (2000), and botanical results combined from Wick (2000) and Tobolski and Ammann (2000).

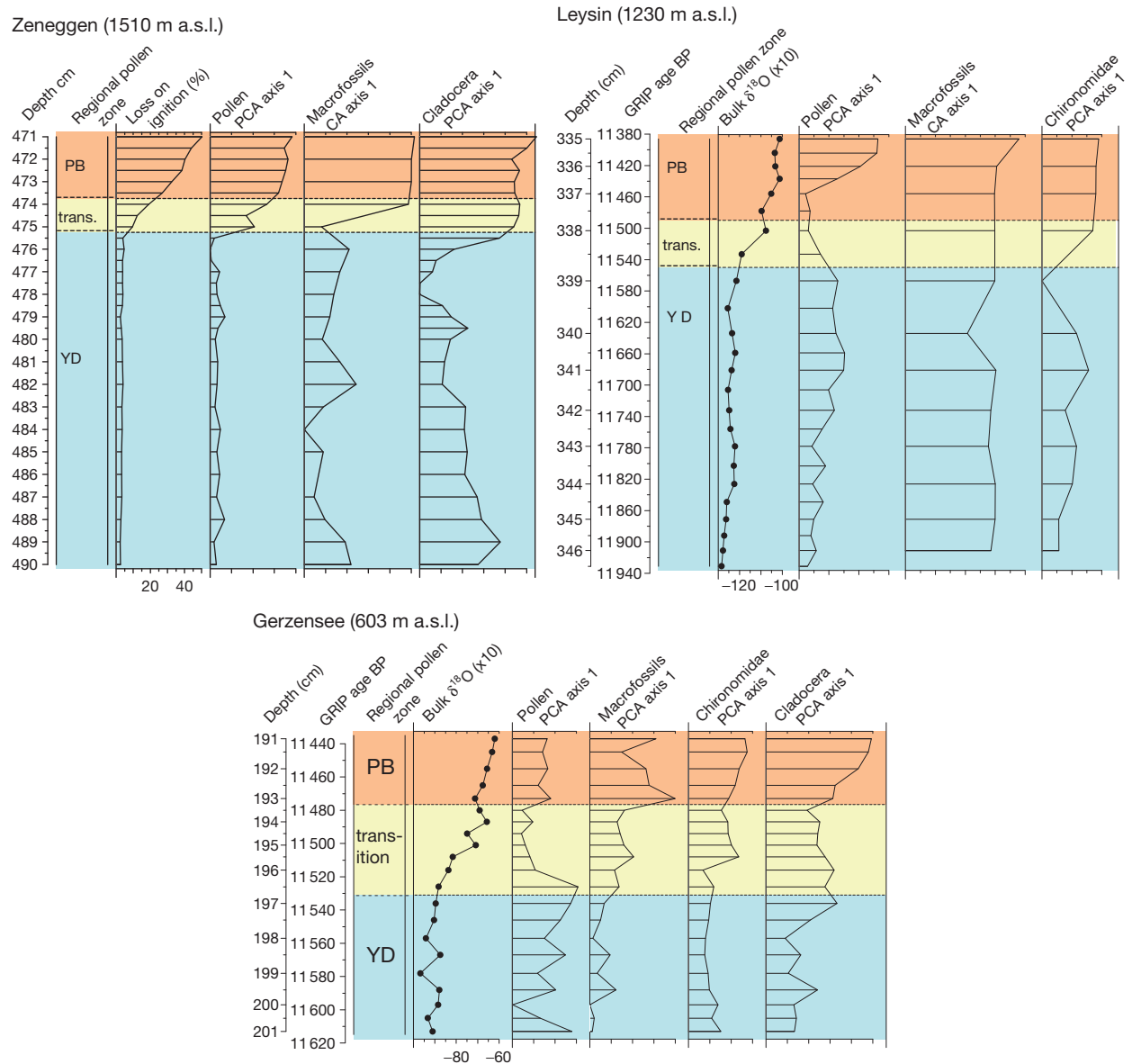


Figure 5 Biotic responses to the rapid warming at the end of the Younger Dryas in the same sites as in [Figure 4](#). The qualitative and quantitative changes in species composition and abundances are summarized as scores of the first axis on a PCA of one or two groups of invertebrates and as a CA for plant macrofossils. Cladocera seem to respond first; pollen, plant macrofossils, and chironomids respond in complex ways and usually in two steps. Responses often occur within the sampling resolution. Modified from Ammann B, Birks HJB, Brooks SJ, et al. (2000) Quantification of biotic responses to rapid climatic changes around the Younger Dryas—A synthesis. *Palaeogeography Palaeoclimatology, Palaeoecology* 159: 313–347.

of first reproduction and generation time. Population changes occur at an intermediate rate.

- iii. Productivity (e.g., pollen production per year or tree ring width): This is the fastest type of response and can occur within 1 year.
- iv. Genetic variability within populations may lead to adaptation. Processes of population genetics (including phenomena such as bottle necks, founder effects, and drift) occur on all timescales and are related to the groups of processes mentioned under (i) and (ii). Both, paleo-records (e.g., for spruce [Latalowa and van der Knaap, 2006](#)) and phylogeography (e.g., for spruce [Scotti et al., 2000](#); [Sperisen et al., 2001](#)) provide

complementary information on tree migration (as an integrated example see for beech [Magri et al., 2006](#)). Again, the results of plant macrofossils may be important here for the taxonomic and spatial distinctions (i.e., the distinction between tree species within a genus and the distinction between presence and absence in the past). An important bridge between paleoecology and phylogeography may develop if the analysis of ancient DNA becomes less problematic, provided that DNA degradation is not too serious. In this context, the analysis of plant macrofossils will become even more important by providing species-specific samples.

See also: Oribatid Mites; Paleoclimate Relevance to Global Warming; Plant Macrofossil Introduction. **Chironomid Records:** Chironomid Overview. **Paleoclimate Reconstruction:** Approaches. **Paleolimnology:** Cladocera. **Plant Macrofossil Methods and Studies:** Treeline Studies; Validation of Pollen Studies. **Radiocarbon Dating:** Conventional Method.

References

- Ammann B (2000) Biotic responses to rapid climatic changes: Introduction to a multidisciplinary study of the Younger Dryas and minor oscillations on an altitudinal transect in the Swiss Alps. *Palaeogeography, Palaeoclimatology, Palaeoecology* 159: 191–201.
- Ammann B, Birks HJB, Brooks SJ, et al. (2000) Quantification of biotic responses to rapid climatic changes around the Younger Dryas—A synthesis. *Palaeogeography, Palaeoclimatology, Palaeoecology* 159: 313–347.
- Birks HH (1993) The importance of plant macrofossils in late-glacial climatic reconstructions: An example from western Norway. *Quaternary Science Reviews* 12: 719–726.
- Birks HH (1994) Late-glacial vegetational ecotones and climatic patterns in western Norway. *Vegetation History and Archaeobotany* 3: 107–119.
- Birks HH (2003) The importance of plant macrofossils in the reconstruction of Lateglacial vegetation and climate: Examples from Scotland, western Norway, and Minnesota, USA. *Quaternary Science Reviews* 22: 453–473.
- Birks HH and Ammann B (2000) Two terrestrial records of rapid climatic change during the glacial–Holocene transition (14,000–9,000 calendar years B.P.) from Europe. *Proceedings of the National Academy of Sciences, USA* 97: 1390–1394.
- Birks HH, Battarbee RW, and Birks HJB (2000) The development of the aquatic ecosystem at Kråkenes Lake, western Norway, during the late-glacial and early-Holocene—A synthesis. *Journal of Paleolimnology* 23: 91–114.
- Brooks SJ (2000) Late-glacial fossil midge stratigraphies (Insecta: Diptera: Chironomidae) from the Swiss Alps. *Palaeogeography, Palaeoclimatology, Palaeoecology* 159: 261–279.
- Brooks SJ and Birks HJB (2000) Chironomid-inferred late-glacial air temperatures at Whitrig Bog, south-east Scotland. *Journal of Quaternary Science* 15: 759–764.
- Gaillard M-J (1984) Etude palynologique de l'Evolution Tardi—et Postglaciaire de la Végétation du Moyen-Pays Romand (Suisse). *Dissertationes Botanicae* 77: 1–322.
- Gaillard M-J and Lemdahl G (1994) Lateglacial insect assemblages from Grand-Maraais, South-Western Switzerland—Climatic implications and comparison with pollen and macrofossil data. In: Cramer J, Lotter A, and Ammann B (eds.) *Festschrift Gerhard Lang Beiträge zur systematic und Evolution, Floristic und Geobotanic, Vegetationsgeschichte und Paläoökologie*. Dissertationes Botanica 234 pp. 287–308.
- Gulliksen S, Birks HH, Possnert G, and Mangerud J (1998) A calendar age estimate of the Younger Dryas–Holocene boundary at Kråkenes, western Norway. *The Holocene* 8: 249–259.
- Jonsgard B and Birks HH (1995) Late-glacial mosses and environmental reconstructions at Kråkenes, western Norway. *Lindbergia* 20: 64–82.
- Kühl N, Gebhardt C, Litt T, and Hense A (2002) Probability density functions as botanical-climatological transfer functions for climate reconstruction. *Quaternary Research* 58: 381–392.
- Latalowa L and van der Knaap, WO (2006) Late Quaternary expansion of Norway spruce *Picea abies* (L.) Karst. in print. In Europe according to pollen data. *Quaternary Science Reviews*.
- Levesque A, Cwynar LC, and Walker IC (1994) A multiproxy investigation of late-glacial climate and vegetation changes at Pine Ridge Pond, southwest New Brunswick, Canada. *Quaternary Research* 42: 316–327.
- Lotter AF, Birks HJB, Eicher U, and Siegenthaler U (1992) Late-glacial climatic oscillations as recorded in Swiss lake sediments. *Journal of Quaternary Science* 7: 187–204.
- Lotter AF, Birks HJB, Eicher U, et al. (2000) Younger Dryas and Allerød summer temperatures at Gerzensee Switzerland, inferred from fossil pollen and cladoceran assemblages. *Palaeogeography, Palaeoclimatology, Palaeoecology* 159: 349–361.
- Magri D, Vendramin GG, Comps B, et al. (2006) A new scenario for the Quaternary history of European beech populations: palaeobotanical evidence and genetic consequences. *New Phytologist* 171: 199–221.
- Schwander J, Eicher U, and Ammann B (2000) Oxygen isotopes of lake marl at Gerzensee and Leysin (Switzerland), covering the Younger Dryas and two minor oscillations, and their correlation to the GRIP ice core. *Palaeogeography, Palaeoclimatology, Palaeoecology* 159: 203–214.
- Scotti I, Vendramin GG, Matteotti LS, et al. (2000) Postglacial recolonization routes for *Picea abies* (L.) Karsten. In Italy as suggested by the analysis of sequence-characterized amplified region (SCAR) markers. *Molecular Ecology* 9: 699–708.
- Snyder JA, Wasyluk K, Fritz SC, and Wright HE (2001) Diatom-based conductivity reconstruction and palaeoclimatic interpretation of a 40-ka record from Lake Zeribar, Iran. *The Holocene* 11: 737–745.
- Solhøy IW and Solhøy T (2000) The fossil oribatid mite fauna (Acari: Oribatida) in late-glacial and early-Holocene sediments in Kråkenes Lake, western Norway. *Journal of Paleolimnology* 23: 35–47.
- Sperisen C, Büchler U, Gugerli F, et al. (2001) Tandem repeats in plant mitochondrial genomes: Application to the analysis of population differentiation in the conifer Norway spruce. *Molecular Ecology* 10: 257–263.
- Stevens LR, Wright HE, and Ito E (2001) Proposed changes in seasonality of climate during the Lateglacial and Holocene of Lake Zeribar, Iran. *The Holocene* 11: 747–755.
- Tobolski K and Ammann B (2000) Macrofossils as records of plant responses to rapid Late Glacial climatic changes at three sites in the Swiss Alps. *Palaeogeography, Palaeoclimatology, Palaeoecology* 159: 251–259.
- van Zeist W and Bottema S (1977) Palynological investigations in western Iran. *Palaeohistoria* 19: 19–85.
- von Grafenstein U, Eicher U, Erlenkeuser H, et al. (2000) Isotope signature of the Younger Dryas and two minor oscillations at Gerzensee (Switzerland): Palaeoclimatic and palaeolimnologic interpretations based on bulk and biogenic carbonates. *Palaeogeography, Palaeoclimatology, Palaeoecology* 159: 215–229.
- Wasylukowa K (2005) Palaeoecology of Lake Zeribar, Iran, in the Pleniglacial, Lateglacial and Holocene, reconstructed from plant macrofossils. *The Holocene* 15: 720–735.
- Wasylukowa K, Witkowski A, Walanus A, et al. (2006) *Palaeolimnology of Lake Zeribar and its climatic implications*. Quaternary Research in press.
- Wick L (2000) Vegetational response to climatic changes recorded in Swiss Late Glacial lake sediments. *Palaeogeography, Palaeoclimatology, Palaeoecology* 159: 231–250.

Pollen Analysis, Principles

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Introduction

Pollen analysis, the study of fossil pollen and spores, is the principal technique for long-term (10^2 – 10^5 years) vegetation reconstructions. Pollen-stratigraphic results have greatly shaped the development of ideas regarding the nature of vegetation and climate during the Quaternary glacial–interglacial cycles. They have provided unique insights into long-term ecological patterns and processes such as the dispersal dynamics and distribution changes of tree species; the ancestry, development, and continuity of modern plant communities and biomes; and the sensitivity and response mechanisms of plants and vegetation to climate changes. They have also shed light on the origin and spread of early agriculture and the role of humans in influencing vegetation.

Many flowering plants liberate enormous quantities of pollen into the air, the production of pollen grains in one forest hectare can be billions (Pohl, 1937). In the atmosphere, pollen is efficiently transported and mixed by wind. Pollen is therefore widely and evenly spread, and fossil pollen records are available from various sedimentary environments such as seas, lakes, ponds, peatlands, soils, and humus layers. The pollen concentration of sediments is usually high; in lake sediment, pollen concentration may often reach 10^3 – 10^5 grains cm^{-3} (Allison et al., 1986), and pollen analysis can often be carried out on sediment samples smaller than 1.0 cm^3 in volume. Where sediment deposition is continuous, sedimentary pollen records may provide an unbroken sequence of useful data. This is one of the major advantages of pollen analysis, namely, a methodological flexibility permitting adjustment of the technique according to the time dimensions relevant for the study. Consequently, the applications of pollen analysis can vary from investigations of the vegetation dynamics of the continents during the glacial–interglacial cycles to studies of annual to subdecadal patterns of the secondary plant succession after a local forest disturbance.

The spread of airborne pollen by wind also results in methodological complexities, some of which constitute the most important limitations and pitfalls of the method. The crux of the problem is that many different patterns of plant abundance can yield the same flux densities of pollen types (Prentice, 1988). The reason for this is that the pattern of pollen dispersal from a singular source, such as a single tree, is typically

leptokurtic: most of the pollen is deposited within 10^1 – 10^3 m from the source, but a small proportion of pollen is transported in the atmosphere over longer, theoretically indefinite, distances (Hjelmroos and Franzén, 1994). It is therefore difficult for a paleoecologist to know whether small proportions of a given taxon in sediment reflect a few individuals of the plant species that formerly grew in the vicinity of the coring site, or whether they came from a larger population that grew far away from the site (Davis, 2000). The problem of precisely defining the pollen source area and quantitatively reconstructing past plant assemblages is crucial, not only because of the risk of erroneous interpretations of the nature of past vegetation and environments, but also because it has been an obstacle for utilizing the full potential of pollen analysis as a central, routinely used tool in ecological research.

This article reviews the general features of pollen analysis as a method in Quaternary science. The focus is particularly on the theoretical basis of pollen analysis as a technique for vegetation reconstruction. Some examples of the application of the technique are also included. These are selected to reflect the variable time and space scales that characterize the use of pollen analysis and also to highlight the problems and pitfalls that limit the use of the technique.

Origin of Pollen and Spores

Pollen grains are unicellular, microscopic (mostly between 15 and $100 \mu\text{m}$) organs that contain the male gametophyte generation of plants. Spores are, in principal, the corresponding organs of the pteridophytes (ferns, mosses, and allies) but differ morphologically and functionally from pollen (Figure 1). In practical terms, pollen analysis usually comprises the identification and tallying of both pollen and spores in an assemblage. Spores are mostly similar in size to pollen grains and have similar preservational characteristics. Hence, any sample that contains pollen is likely to yield spores as well. Pollen grains are produced in the anthers of the male flowers of plants and released during flowering. The purpose of the pollen grain is to germinate on a stigma and provide the sperm nucleus to the ovule. For this purpose, pollen is carried by wind, water, insects, or other animals from the anthers to the stamen of the female flower. Wind-pollinated plants, such as most of the conifers, many other

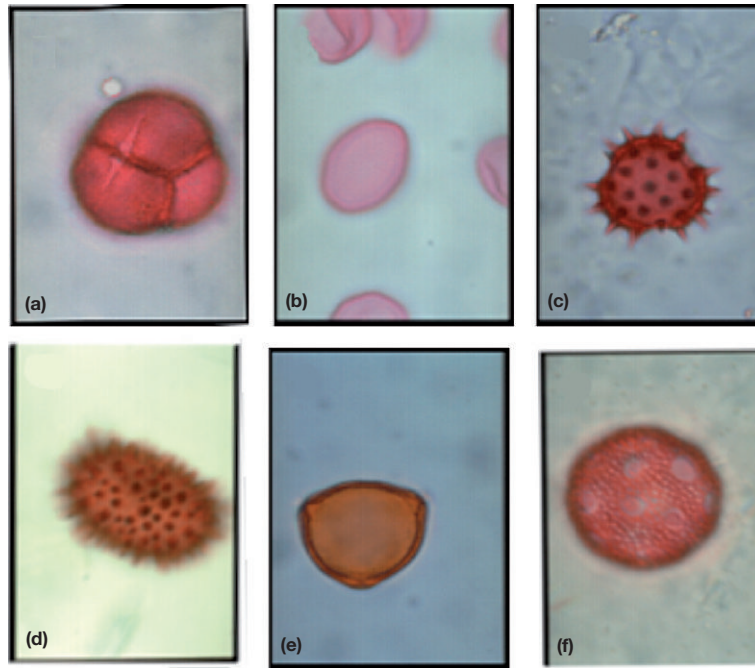


Figure 1 Some examples of different kinds of pollen and spore types: (a) *Vaccinium vitis-idaea* (crowberry), a tetrad pollen type; (b) *Potamogeton natans* (broad-leaved pondweed), an inaperturate pollen grain; (c) *Tussilago farfara* (coltsfoot), an echinate trizonocolporate pollen type; (d) *Cystopteris fragilis* (fragile fern), a monolete spore; (e) *Corylus avellana* (hazel), a trizonoporate pollen type; (f) *Silene dioica* (red campion), a polyaperturate pollen grain. Images from Uppsala University Paleobiology Website: <http://www.palaeontology.geo.uu.se>.

angiosperm trees, grasses, and sedges, usually produce more pollen than animal-pollinated plants. The estimated annual pollen production rates of common wind-pollinated plants reflect the enormous amounts of pollen production. One shoot of hemp can produce over 500 million pollen grains, and a shoot of lodgepole pine (*Pinus contorta*) can produce 8–9 million pollen grains per year; a male plant of common sorrel (*Rumex acetosa*) produces 400 million pollen grains; one 10-year-old branch system of birch, spruce, or oak can produce more than 100 million grains, and pine, over 350 million pollen grains in a year (Pohl, 1937). The pollen productivity of the animal-pollinated plants is, in general, much less, but the production rates of some animal-pollinated plants, such as lime, willow, and heather, are comparable to those of wind-pollinated plants.

Sedimentation and Preservation

Only a fraction of the produced pollen grains falls on the stamen of the maternal part of the flower and fertilizes the ovule. A great majority of pollen falls on land or water and decays, unless pollen is deposited in an anoxic sedimentary environment where it will be preserved. During sedimentation, the cytoplasm (cell contents) of pollen grains is lost, and the fossil pollen grains consist of the hard and chemically resistant outer layers or exines. In addition to the sedimentary environment, preservation of pollen grains depends on the thickness and sporopollenin content of the exine, so some pollen grains preserve better in sediments than others.

Various types of corers and samplers are used to recover sediments from different sedimentary contexts, such as lakes and peatlands (Glew et al., 2001). Pollen is extracted from

sediments in the laboratory. Sediment subsamples are processed in the laboratory by a series of physical and chemical treatments. The standard procedure includes treatment of the sample with dilute hydrochloric acid, potassium hydroxide solution, hydrofluoric acid, and a mixture of sulfuric acid and acetic anhydride (Bennett and Willis, 2003; Fægri and Iversen, 1991; Moore et al., 1991). Hydrofluoric acid is a strong acid that dissolves the silicates; sulfuric acid and acetic anhydride form a chemical compound for acetolysis that removes cellulose and most of the other organic remains from the subsample. The grains may be stained with safranin, which can help make them more distinctive under the microscope. The grains are then suspended, usually in silicone oil or glycerine jelly. Pollen grains are normally counted with 500× magnification.

Plant–Pollen Representation

The results of the pollen counts are usually presented in pollen diagrams showing the percentage values of all or selected pollen types. It is important to note that most pollen grains cannot be identified to the species level. Most are identified either to the genus or family level only. Palynologists working in a given region may come to recognize two or more types of pollen within a genus or family that are specific to that region and can identify these on the basis of identification keys. Pollen of all grass species are identified at the family level (Poaceae), pollen of different pine species at the generic level (*Pinus*), and the three common North-European birch species cannot be distinguished from each other.

Since the first comparisons of pollen diagrams and modern plant abundances, it has been known that pollen records do

not directly reflect plant abundances because some taxa are overrepresented and others underrepresented in pollen samples. A potential way to transform the fossil pollen percentage values to correspond more precisely with real relative plant abundances is to use various types of statistical correction factors developed individually for each pollen type. In order to obtain such factors, it is necessary to study the pollen–plant relationships by comparing modern plant abundances and modern pollen values within forests. Foundations for this work were laid in the temperate woodlands in North America and Europe in the 1960s. Davis (1963) defined an index, termed the *R*-value, for describing the ratio between pollen and vegetation percentages and used the *R*-value to convert the original pollen sums to reflect the vegetation percentages. In contrast, Andersen (1970) compared absolute pollen accumulation values analyzed from moss polsters with the vegetation percentages of corresponding tree species in the same woodland. Significantly, he showed that models describing pollen input cannot consist only of multiplying terms that reflect the relative productivity differences but must also include additive terms to account for the background pollen component originating outside the vegetation area used for calibration purposes. The relationship between pollen percentage values and the vegetation percentage can therefore be depicted as a linear regression curve in which the slope indicates the pollen productivity of the given species and the *y*-axis intercept shows the amount of background component of the same species transported from outside the study area.

An improvement combining features of both the *R*-value method and Andersen's linear regression method is an extended *R*-value (ERV) method, which is based on the percentage values but takes into account the background pollen input (Prentice, 1988; Prentice and Parsons, 1981). In addition, the ERV method includes a site factor that ensures that the corrected percentages at each site total 100%. The ERV model presents the relationship between the pollen percentage and the surrounding vegetation as:

$$p_{ik} = \alpha_i v_{ik} f_k + z_i$$

where p_{ik} is the pollen percentage of the taxon *i* at site *k* and v_{ik} is the abundance of the taxon in the vegetation. α is a coefficient describing the pollen production of taxon *i*, and z_i is the background component of pollen coming from outside the study area. f_k is the site factor ensuring that the right-hand side of the equation will total 100%. It is important to note that these taxa have to be identified at the same taxonomic level. Thus, if the pollen taxon is *Pinus*, then the vegetation taxon must also be *Pinus* (genus level) and would include all the pine species in the area, subsumed at that taxonomic level. There is often confusion on this point, whereby people talk about *Pinus* (pollen taxon, genus level) and then start discussing pine species individually.

The site factor can be expressed as:

$$f_k = \left(1 / \sum_j \alpha_j v_{jk} \right) \left(1 - \sum_j x_j \right)$$

Due to the inclusion of the site factor, ERVs also account for the Fagerlind effect (Prentice and Parsons, 1981), a non-linearity typical of pollen percentage data. The linear regression methods assume a linear relationship between the pollen

values and plant abundance. However, the percentage pollen data are not linear, because they must total 100%. Therefore, an increase in the pollen type of one taxon must be accompanied by a decrease of another, even if the actual abundances of the other taxa remain constant. In general, the ERVs give a markedly better fit to pollen percentage–plant abundance data than do the classical linear regression models.

Although valuable and rational in principle, the generation of reliable correction factors is a complicated process, and their use is never simple or straightforward. Nevertheless, when the representation differences between the main plant taxa are significant, the corrected values undoubtedly often reflect the original plant abundances more closely than the uncorrected values. Yet, it is important to remember that the linear models for pollen input assume constant background pollen and constant pollen productivity for any given taxon, whenever it appears, implying that the use of correction factors is constrained in space and time to the overall vegetation type for which they were originally generated. Vegetation changes during the Holocene have been of such magnitude that the background pollen input has undoubtedly changed, and the linear relationship obtained from modern samples is therefore not a reliable estimate for correcting the fossil pollen values on millennial or longer timescales.

Pollen Dispersal and Model-Based Correction Approaches

To fully account for the complexities of the dispersal of pollen in the atmosphere, a more comprehensive model linking the regional and local vegetation patterns and defining the source area of pollen is needed. Intuitively, it is obvious that small hollows (small openings within forests) receive great quantities of pollen from the vegetation around the sampling site, whereas centers of larger basins, for example, lakes, receive predominantly wind-transported and well-mixed pollen from the air. Consequently, the source area of pollen records from lakes is the regional vegetation, whereas small hollows have a smaller source area and reflect more the local vegetation around the site (Jacobson and Bradshaw, 1981). This simple model has been used as a general rule of interpretation of pollen-stratigraphic records from sampling sites of different sizes. Studies designed to investigate stand-scale forest disturbance or local prehistoric land use can use small hollows or lakes with a radius of less than 50 m, whereas palynologists interested in regional-scale pollen signals (e.g., in climate reconstructions) need to focus on larger lakes, for example, with a radius of 250 m or more. Such a qualitative and intuitive definition does not, however, permit specification of the spatial scale of the study with the precision that would be comparable to that of modern ecological studies. A challenge for palynologists has been to define the pollen source areas more quantitatively and precisely, and, hence, to generate more spatially precise vegetation reconstructions (Figure 2 and Table 1).

Comprehensive efforts to model pollen dispersal and define the source areas have been carried out since the 1980s, and the resulting model is usually termed the 'Prentice–Sugita model' after its key developers (Prentice, 1988; Sugita, 1994).

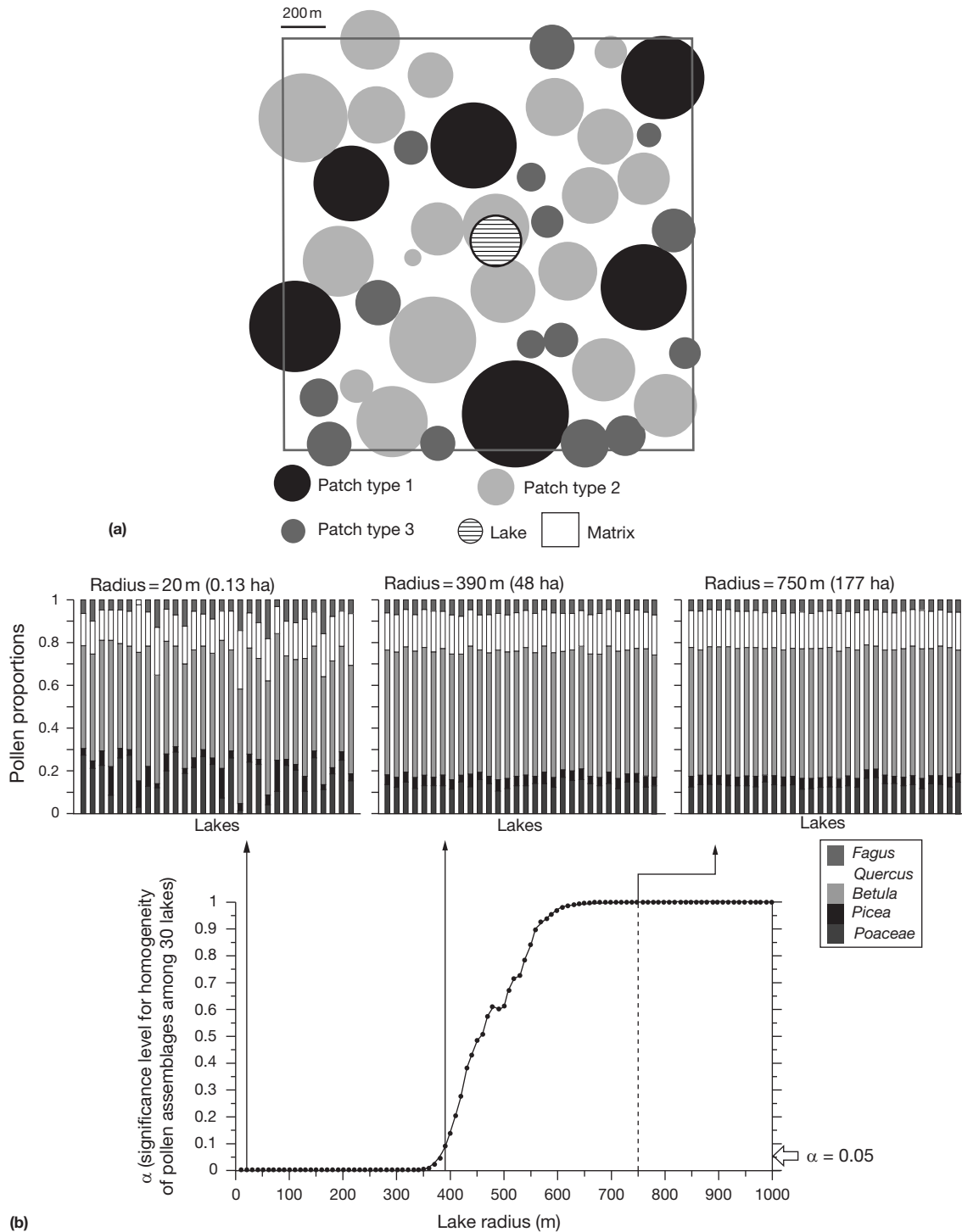


Figure 2 The influence of the sample size on the pollen assemblages obtained from lakes of different radii in a simulated patchy landscape. (a) An example of the simulated landscape, with forest stands in a matrix of grass-dominated vegetation. See Table 1 for stand compositions. (b) Simulated pollen assemblages among 30 lakes with different sizes. The percentage values stabilize when the lake size is 48 ha or larger, indicating that the lakes of this size sense the vegetation as homogenous in this particular landscape at the significance level $\alpha = 0.05$, using the likelihood ratio statistics G^2 . Source: Sugita S (2007a) Theory of quantitative reconstruction of vegetation I: Pollen from large sites REVEALS regional vegetation composition. *The Holocene* 17: 229–241.

The principal purpose of the model is to quantify the pollen input from the vegetation surrounding the sampling site, either to the center of a basin or on the entire surface of a basin. It is important to assess the pollen inputs to the center of a basin

because this is the place from which sediment cores are very often taken for pollen analysis purposes. According to the model, the input depends on the pollen productivity of the plant species contributing to the pollen rain at the site,

Table 1 Stand characteristics in the simulation experiment in Figure 2. The landscape has four stand types. The distribution of taxa in each stand is assumed to be homogeneous. Source: Sugita S (2007a) Theory of quantitative reconstruction of vegetation I: Pollen from large sites REVEALS regional vegetation composition. *The Holocene* 17: 229–241

Taxa	Matrix	Patch type 1	Patch type 2	Patch type 3	Overall composition
Proportion in area	0.40	0.30	0.20	0.10	
Mean size $\pm$ SD (ha)		30 $\pm$ 10	15 $\pm$ 5	2 $\pm$ 1	
Poaceae	0.95	0.01	0.05	0.30	0.423
<i>Picea</i>	0.01	0.75	0.01	0.01	0.232
<i>Betula</i>	0.01	0.20	0.25	0.30	0.144
<i>Quercus</i>	0.01	0.03	0.14	0.38	0.079
<i>Fagus</i>	0.02	0.01	0.55	0.01	0.122

the mean plant abundance at a given distance from the center of the basin, and a pollen deposition function. The pollen deposition function, based on Sutton's physical equation for a ground source (Prentice, 1988), defines the proportion of pollen remaining airborne at a given distance from the plants and is individual to all pollen types because of their individual settling velocities. A characteristic feature common to the shapes of all the deposition functions is their leptokurtic nature: a majority of pollen will be deposited near the source, but a proportion will be transported over theoretically indefinite distances (Prentice, 1988). The distributions may also be highly skewed, due to prevailing wind directions or terrain. This pattern of pollen dispersal from a source explains why plants near a lake are more strongly represented in the pollen input of a basin than those growing progressively further from the basin and implies that some form of distance weighting must be applied to plant abundances while modeling or comparing modern pollen assemblages with modern plant abundances (Prentice, 1988; Sugita, 1994).

A more comprehensive theoretical modeling approach to understand and correct for the differences in pollen productivity, pollen dispersal, and sampling size is called the Landscape Reconstruction Algorithm (LRA; Sugita, 2007a,b). LRA consists of two models. Regional Estimates of VEgetation Abundance from Large Sites (REVEALS) is a model to reconstruct quantitatively the regional composition of vegetation through time. The REVEALS model defines the pollen loading, the average input of pollen on area unit, in the center of a large lake on the basis of mean plant abundance of a given taxon, pollen productivity of each taxon, radius of the lake, maximum distance of the plant from the lake, and a dispersal function. Once the pollen loading is known, it is possible to use the basic equation of the REVEALS model to estimate the past mean plant abundances and hence to reconstruct the past regional composition of vegetation. Simulation experiments and empirical tests have shown that the lakes of size ≥ 100 –500 ha are generally large enough so that the pollen data can provide accurate estimates of regional vegetation (e.g., Hellman et al., 2008).

Once the regional vegetation change through time has been reconstructed with the REVEALS model, it is possible to use the LRA to reconstruct local vegetation features from small sites. This requires the incorporation of the Local Vegetation Estimates (LOVE) model to quantify and subtract background pollen to produce quantitative vegetation reconstructions

within the relevant source area of smaller sites (Sugita, 2007a). The central concept here is the 'relevant source area of pollen' (RSAP). RSAP refers to the size of the area around a given study site, which controls the individualistic features in the pollen assemblages of the site in relation to the uniform regional background pollen. The vegetation patterns within the RSAPs are therefore reflected in the lake-to-lake differences in pollen input in areas of the same regional vegetation (Davis, 2000; Sugita, 1994). The RSAP for a site can be estimated by calculating the r^2 , or maximum likelihood function score, as a measure of goodness-of-fit between the pollen assemblage and the distance-weighted plant abundance around each study site (Sugita, 2007b). The distance beyond which r^2 , or the likelihood score, no longer improves defines the radius of the RSAP for that particular site.

LOVE is theoretically similar to the inverse form of the ERV, but the important difference is that, unlike the ERV, LOVE does not assume a constant background pollen but incorporates it as a parameter that can vary through time and can be estimated using the REVEALS model. Thus, the ideal LRA approach to estimate past vegetation composition in given specifically defined areas in a patchy landscape is to combine the pollen records from a large lake and the REVEALS model to reconstruct past regional vegetation and to combine the pollen records from a number of smaller sites and the LOVE model to reconstruct the past vegetation within the RSAPs of the smaller sites, corresponding with the study areas of interest, for each time window. This provides vegetation reconstructions at both local and regional scales (Figure 3).

A key requirement for the use of the LRA is the existence of pollen productivity and pollen dispersal estimates, reflected by pollen settling velocities, at least for the most common and important plant taxa, especially in northwest Europe where it has been studied most intensively. Obtaining these estimates is a non-trivial task, but the tests have shown that once the information is available, REVEALS and LOVE models have the potential to produce more accurate, quantitative vegetation reconstructions. As with all models, the LRA procedure contains many assumptions regarding pollen transportation and deposition mechanisms, local and regional vegetation patterns, and basin configuration (Sugita, 2007b). It does not, for example, account for stream-borne pollen, a major source of pollen in many lakes. It is thus unclear how reliably the approach can be applied to produce corrected vegetation



Figure 3 (a) A lake of about 20 ha, a typical sampling site for pollen analysis focusing on reconstruction of postglacial regional vegetation patterns in the boreal zone of Europe. The dominant trees in the surrounding forest are species of pine and spruce (Photo: Charlotte Sweeney). (b) Small forest hollow in Suserup Skov, Denmark. Photograph by Gina Hannon.

reconstructions, particularly outside the lake-rich landscapes of the temperate and boreal biomes.

A potential way to avoid the problems related to percentage data and to estimate plant abundance in the surrounding landscape independently for each species is to calculate the direct pollen accumulation rate (PAR) values individually for each taxon (keeping in mind the caveat that this does not equal plant species in most cases). Instead of relative percentage data, PAR is a calculation of the accumulation of pollen at the sediment surface (per cm^2 or other space unit) per year (Davis and Deevey, 1964). This value is individual and independent for each pollen and spore type and therefore avoids the problems arising from the interdependent nature of the relative records. PAR values are usually calculated for the most important tree species, the aim being for PAR values to provide precise records of past tree population sizes, measured either as canopy coverage or biomass. However, methodological investigations have shown that the PARs can vary by factors from 3 up to 10, not only within lakes, but also between lakes located in the same area and surrounded by similar vegetation. The predominant reason for the within-basin variability of PAR is the spatially and temporally uneven pattern of matrix sedimentation and pollen deposition in the lake basin (Seppä et al., 2009b).

Despite these caveats, in some regions, the PAR technique has been successfully applied to estimate past tree population density changes around the study lakes, because they can help to circumvent the problems arising from the nonlinearity of pollen percentage data. Especially valuable have been the applications of PAR values to investigate past competitive interactions, for example, the population growth of an invading tree taxa and the associated population declines of other, less competitive taxa (Figure 4; Bennett and Lamb, 1988; Seppä et al., 2009a). PAR values can be also valuable in the ecotones of the forested and treeless biomes, such as boreal forest and the arctic tundra, where the advances or retreats of the main tree taxa are recorded as major changes in PAR values, while less change is apparent in percentage values. In suitable circumstances, it is possible to use modern surface samples or pollen trap results for generating

threshold PAR values that define the absence, presence, or dominance of key tree species and apply these values to the fossil PAR records for providing semiquantitative estimates of past population densities in the study region. To go further in the direction of quantification will require a survey of modern population sizes of key tree species in the RSAPs of modern sites used for calibration of the PAR values (Figure 5; Seppä et al., 2009b).

Applications of Pollen Analysis

Local-Scale Studies

As indicated by the pollen representation models, the RSAPs for pollen records obtained from small forest hollows or moss polsters are about 50–200 m. Such records reflect forest history on a stand scale, the same spatial scale on which most modern forest ecologists operate, and the scale for which the classical forest succession models describing the predictable sequence of plant communities on annual to decadal time-scales were developed. Consequently, paleoecologists working on local-scale pollen studies can share the concepts, research problems, techniques, and perspectives of forest ecology. This provides opportunities for close collaboration with modern ecologists, whose understanding of local dispersal, establishment, and other time-related ecological processes emerges from observations of contemporary processes and may thus underestimate the underlying critical historical factors (Bradshaw, 1988).

Another major advantage of focusing on local-scale studies is that it is often possible to relate reliably past changes in species composition and structure to external factors causing the changes. Independent evidence for the external factors can be obtained, for example, from analyses of sediment charcoal, tree rings, historical archives, and climate reconstructions based on techniques independent of vegetation dynamics. Comparisons of these records with local-scale pollen records often demonstrate the importance of such local processes as fire, human disturbance, and storms in vegetation dynamics. Immigration or expansion of a species that is simultaneous with a local disturbance

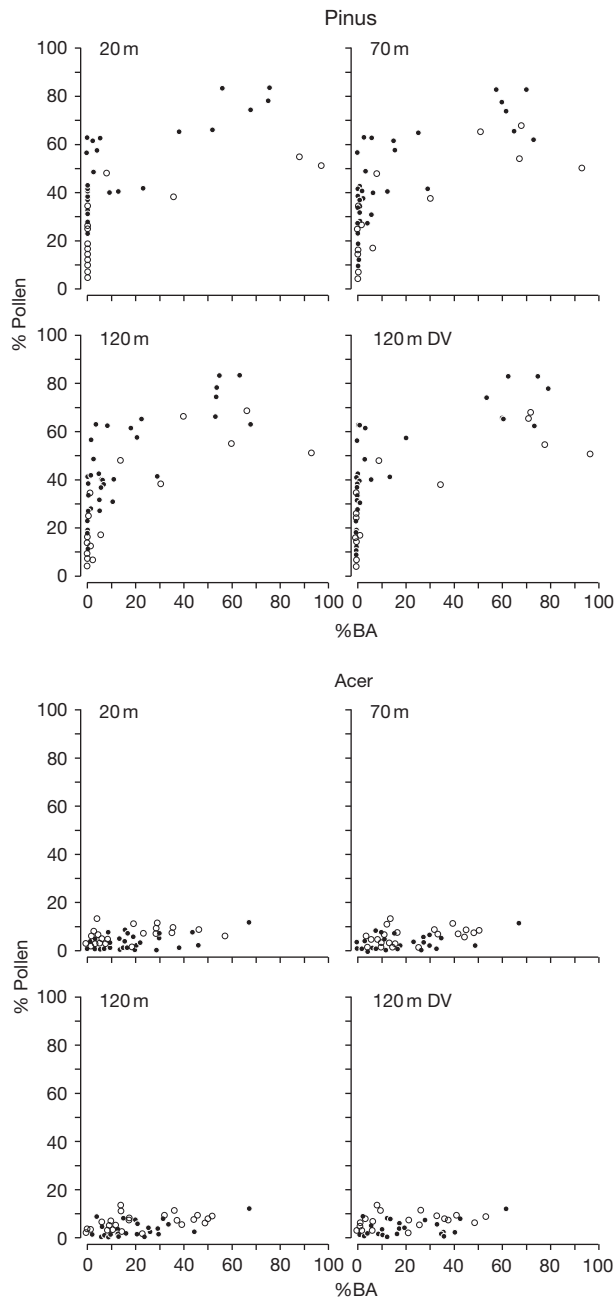


Figure 4 Scatter diagrams showing the relationship between pine (*Pinus*) and maple (*Acer*) pollen percentages and their basal area (BA) percentages in the surrounding vegetation. Pollen percentages for individual sites are constant, but three different vegetation sampling areas have been surveyed: 20 m = 20-m vegetation sampling radius, 70 m = 70-m vegetation sampling radius, and 120 = 120-m vegetation sampling radius. Results from two different data sets are shown, from Fish Creek (open circles) and the eastern Adirondacks (closed circles) in northern New York, USA. Scatter plots such as these provide the basis for estimating the pollen productivity of plant taxa; taxa with high productivity have a steep slope, and those with efficient dispersal have a high y-axis intercept. Source: Jackson ST and Kearsley JB (1998) Quantitative representation of local forest composition in forest-floor pollen assemblages. *Journal of Ecology* 86: 474–490.

such as a fire may have taken place as a response to the altered growth conditions and interspecific competition. For example, the local establishment of beech at its distribution limit in northern Europe seems to have been greatly facilitated by human activities, especially disturbance by fire, and this association with historical disturbances is reflected in the species' modern patchy distribution pattern (Figure 6; Bradshaw and Lindbladh, 2005). On the other hand, stand-scale invasions during intervals with no evidence for disturbance, such as that of hemlock in its western range limit in North America, may be related to rapid or gradual climate changes and associated changes in local growing conditions (Figure 7; Parshall, 2002). These examples reflect the individualistic response patterns and invasion histories of tree species and demonstrate the great relevance of local-scale, pollen-analytical investigations for basic and applied forest ecology.

Regional-Scale Studies

Conventionally, pollen records have been produced from small- to medium-sized (roughly 10–50 ha) lakes and mires and reflect the vegetation of the surrounding few hundreds of square kilometers. Pollen records on this scale have had a major influence on our understanding of vegetation history and concepts of population, community, and biome dynamics during the Quaternary Period. Regional-scale pollen records have shown dramatic vegetation changes during glacial–interglacial cycles in regions adjacent to ice sheets and have provided unique insights into the nature of vegetation during the previous interglacials. They have also shed light on significant vegetation zone shifts during the interglacials, such as the early- to mid-Holocene expansion of savannah vegetation into the modern Saharan desert in Africa, the northward advance of the boreal conifer forest zone of the Northern Hemisphere, and the major deforestations caused by humans in many ecosystems during the late Holocene.

By synthesizing data from regional records over large areas, it has been possible in some instances to reconstruct the location of refugia of the main tree species during the last glaciation, and pollen data have helped map their subsequent spread during the interglacial (Bennett, 1997; Huntley and Birks, 1983). According to conventional interpretation, a post-glacial immigration of a new plant species into a region is reflected in a distinct rise of the percentage values of the corresponding pollen type. Many recent studies, however, have challenged this view by showing that the regional-scale pollen records may regularly provide underestimates for the date of immigration of the new species into a given area. The occurrence of macrofossils and stomata of tree species have provided firm evidence for the presence of tree species hundreds or thousands of years before any significant rise in the taxon's pollen curve (Välranta et al., 2011). This suggests that species may have occurred sparsely, probably under conditions limiting both population growth and pollen productivity, until a population expansion occurred in response to a change of external factors. In many cases, it may be this expansion that is reflected in the visible rise of the pollen curve, not the immigration. Motivated by these observations, some palynologists have also adopted a fresh approach to interpreting

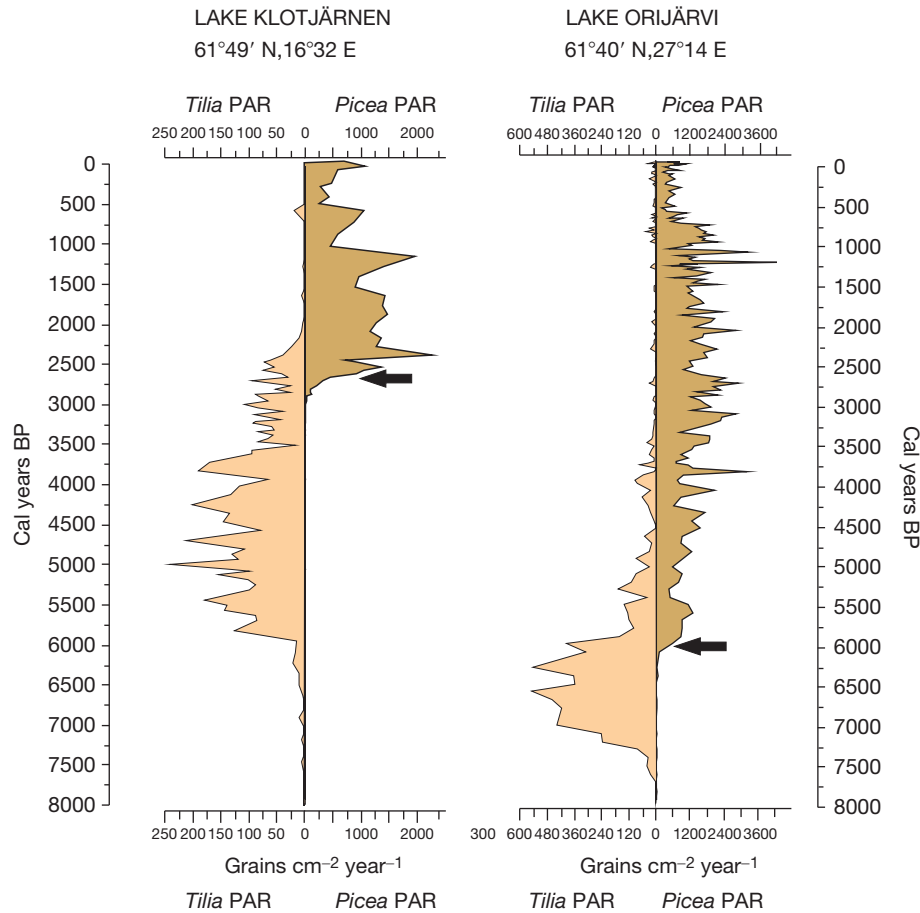


Figure 5 Pollen accumulation rate (PAR) values provide a means of estimating past plant abundances independent of pollen percentage values. Here, the PAR records of Norway spruce (*Picea abies*) and small-leaved lime (*Tilia cordata*) are presented from two lakes in Fennoscandia. The black arrow represents the time of the immigration of spruce, and increasing PAR values show its rise to dominance. The concurrent decline of lime PAR values indicate that the spruce population expansion happened at the expense of the lime population, mostly as a consequence of the competitive replacement. Source: [Seppä H, Alenius T, Giesecke T, Bradshaw RHW, Heikkilä M, and Muukkonen P \(2009b\)](#) Invasion of Norway spruce (*Picea abies*) and the rise of the boreal ecosystem in Fennoscandia. *Journal of Ecology* 97: 629–640.

pollen data. A comparative examination of combined and mapped pollen records from Beringia, a region that had been earlier interpreted as being tundra with no forests or even trees present, showed consistent patterns indicating a presence of tree species during the Last Glacial Maximum and subsequent expansions of the populations ([Brubaker et al., 2004](#)). Overall, the detection of a presence of taxon from pollen data, especially in a sparsely vegetated region, remains difficult, and the most reliable results can be achieved by an integration of pollen analysis with analyses of stomatal and plant macrofossil evidence from the study region ([Birks and Birks, 2001](#)).

Continental-Scale Reconstructions

The continental-scale applications of pollen analysis usually involve pollen-stratigraphic analyses of long sediment cores obtained from ocean bottoms ([Margari et al., 2010](#); [Sanchez-Goni et al., 2008](#)) or from continental sites where sediment has

accumulated undisturbed over tens of thousands of years, for example, in crater lakes or lakes located in tectonic depressions beyond regional glacial ice limits ([Lozhkin et al., 2007](#); [Tzedakis et al., 2006](#)). These sediment cores span tens or even hundreds of thousands of years and can cover full glacial–interglacial cycles. In addition to providing evidence of the nature of vegetation during the previous interglacials and interstadials, the records reflect the pace of vegetation dynamics in response to the Milankovitch-scale climate dynamics, demonstrating alternations of forests and treeless vegetation types consistent with variations in solar insolation and global ice volume. The long records obtained from terrestrial sites in particular allow integration of the land-based and marine records (e.g., the marine isotope and other records) and hence a synthesis and correlation of marine and terrestrial ecosystems.

A difficulty with long pollen records is that their chronological control is, in general, weakly constrained, often being obtained by correlating distinct events and intervals of the pollen diagrams with precisely dated, long records such as ice cores.

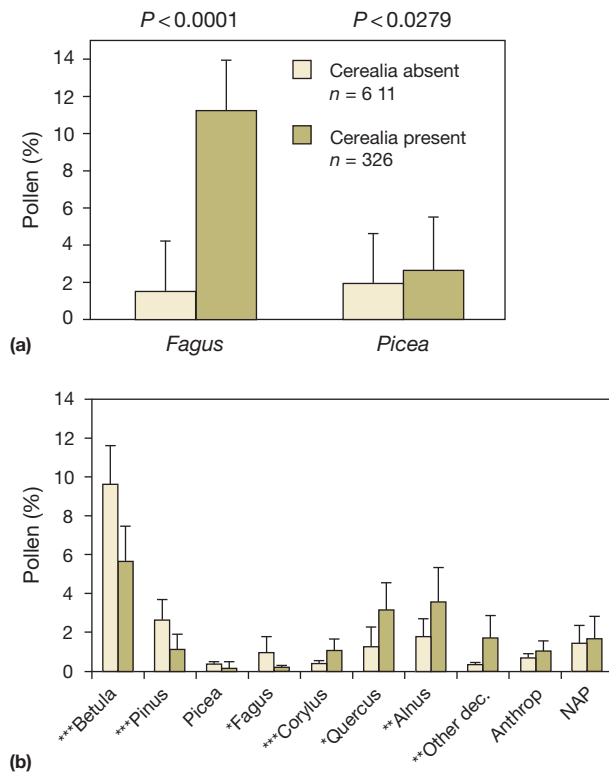


Figure 6 Stand-scale pollen records reflect the contrasting invasion patterns of Norway spruce (*Picea abies*) and European beech (*Fagus sylvatica*) in northern Europe. (a) Pollen percentage values (mean + SD) with a total terrestrial pollen sum for Norway spruce and European beech in small hollow samples with and without cereal grain pollen. The percentages of beech are higher when cereal pollen is recorded in the same samples, while spruce does not indicate similar relationship. (b) The mean forest stand composition in southern Scandinavia during the invasion of spruce and beech. The percentage values (mean + SD) were adjusted by stand-scale correction factors that allow for differential pollen representation. The light brown bars indicate sites invaded by Norway spruce and the brownish green bars sites invaded by European beech. Norway spruce favored forest stands where birch, pine, or beech was abundant, whereas European beech was established in sites with oak, alder, hazel, and other deciduous trees. All sites are totally or mostly 'closed canopy' sites with RSAP roughly comprising the surrounding 20–50 m. Source: Bradshaw RHW and Lindbladh M (2005) Regional spread and stand-scale establishment of *Fagus sylvatica* and *Picea abies* in Scandinavia. *Ecology* 86: 1679–1686.

More reliable correlations can be obtained by analyzing continental and marine proxies from the same deep-sea sequences or by applying tephrochronological techniques to generate more precise chronologies for the continental cores (Figure 8). Improved chronological precision has prompted palynologists to investigate the timing and duration of the interglacials in pollen records and to examine whether the recently realized abrupt climatic events and cycles typical of the glacial intervals influenced the vegetation of the continents adjacent to the Atlantic. Long continental pollen records can also be used for quantitative climate reconstructions because, on the timescale

represented by these records, vegetation patterns were predominantly controlled by the climate regimes. Possibilities for precise reconstructions are limited by the often low level of analog between modern pollen assemblages and those representing the glacial intervals. In general, however, the results indicate trends that are comparable to those obtained by independent techniques such as the stable isotope records of ice cores.

Conclusions and Future Directions

Historically, the main function of pollen analysis in Quaternary-stratigraphic investigations has been to identify chronostratigraphic periods, to correlate sediment sequences, and to provide general reconstructions of vegetation and climate. Conventional pollen diagrams are still valuable when produced in new regions or collected from key sites where long, uninterrupted pollen records can be obtained, but in regions where the palynological studies have a long history, for example, in eastern North America and Europe, conventional pollen analysis is increasingly being replaced by carefully designed studies that address specific questions and concentrate on specific time intervals. This is particularly so in the application of pollen analysis in Quaternary paleoclimatology, where the recent discoveries of major Quaternary climate changes, such as Heinrich events during the last glacial, the abrupt warmings and coolings during the last glacial termination, and the short-lived cold interval about 8.2 ka ago, highlight the need to focus on high-resolution pollen-stratigraphic data from optimally located sites in order to investigate the occurrence of these climate events in the continental regions. A related emerging application of pollen analysis is its use as a method in Earth system science, for example, to reconstruct past continental-scale land cover changes and assess their importance in the global carbon cycle and the climate system (Gaillard et al., 2010).

There is another major trend in pollen analysis, where the aim is to reconstruct past vegetation in more precise terms. The fundamental objective is to provide quantitative vegetation reconstructions that incorporate central ecological attributes such as floristic composition, species abundance, and the spatial distribution of plants in an explicitly defined context of time and space and assess the role of internal and external factors as drivers of change. For this purpose, we must to improve our understanding of modern relationships between vegetation patterns and pollen input, including the critical processes of pollen productivity and dispersal from plants to the deposition site. We must also improve our ability to identify pollen to lower taxonomic resolution (i.e., to the species level), so that plant records and pollen records can be more directly compared. Pollen analysis will have only limited utility (and not reach its full potential) as long as we are able to identify the majority of types to just genus level. Formidable (and perhaps intractable) taxonomic and identification issues remain.

Despite having made significant progress, pollen analysts are still far from reaching these ambitious goals but remain motivated by the breakthrough potential of the method in truly quantitative vegetation reconstructions.

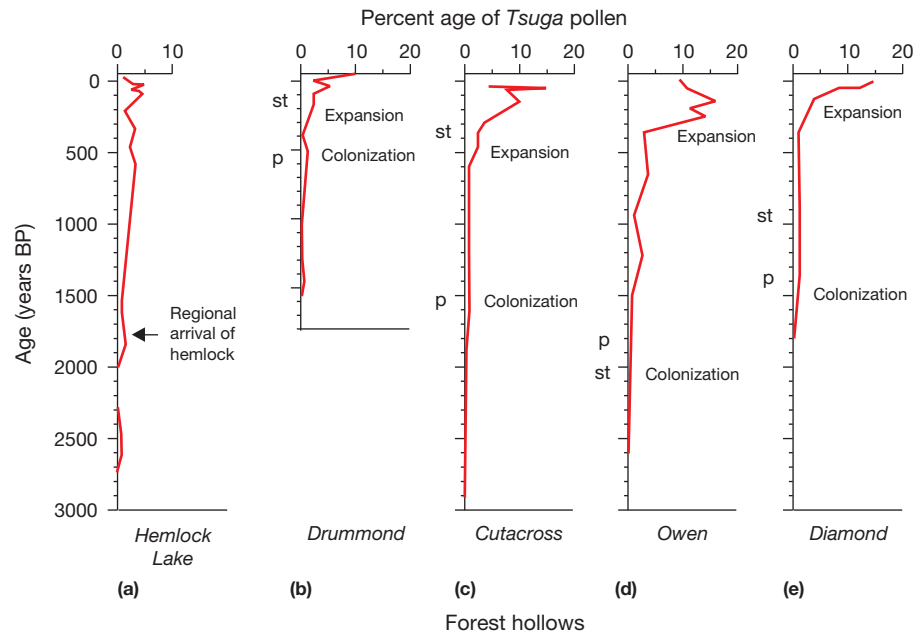


Figure 7 A combined use of stand-scale and regional-scale records can provide a detailed picture of past population dynamics, as exemplified by the investigation of hemlock population history in Wisconsin, USA. Hemlock pollen percentage values from a lake ((a) Hemlock Lake) show the regional arrival of Canadian hemlock (*Tsuga canadensis*) at about 2000 cal years BP. Four stand-scale records (b–e) from small hollows indicate that after its spread, hemlock occurred in the region as scattered populations until a rapid expansion at about 500 cal years BP. The expansion has been a regionally synchronous event, suggesting a climatic change toward moister and cooler conditions. Records of stomata (st) analyzed from the pollen samples corroborate the pollen evidence. Source: Parshall T (2002) Late Holocene stand-scale invasion by hemlock (*Tsuga canadensis*) at its western range limit. *Ecology* 83: 1386–1398.

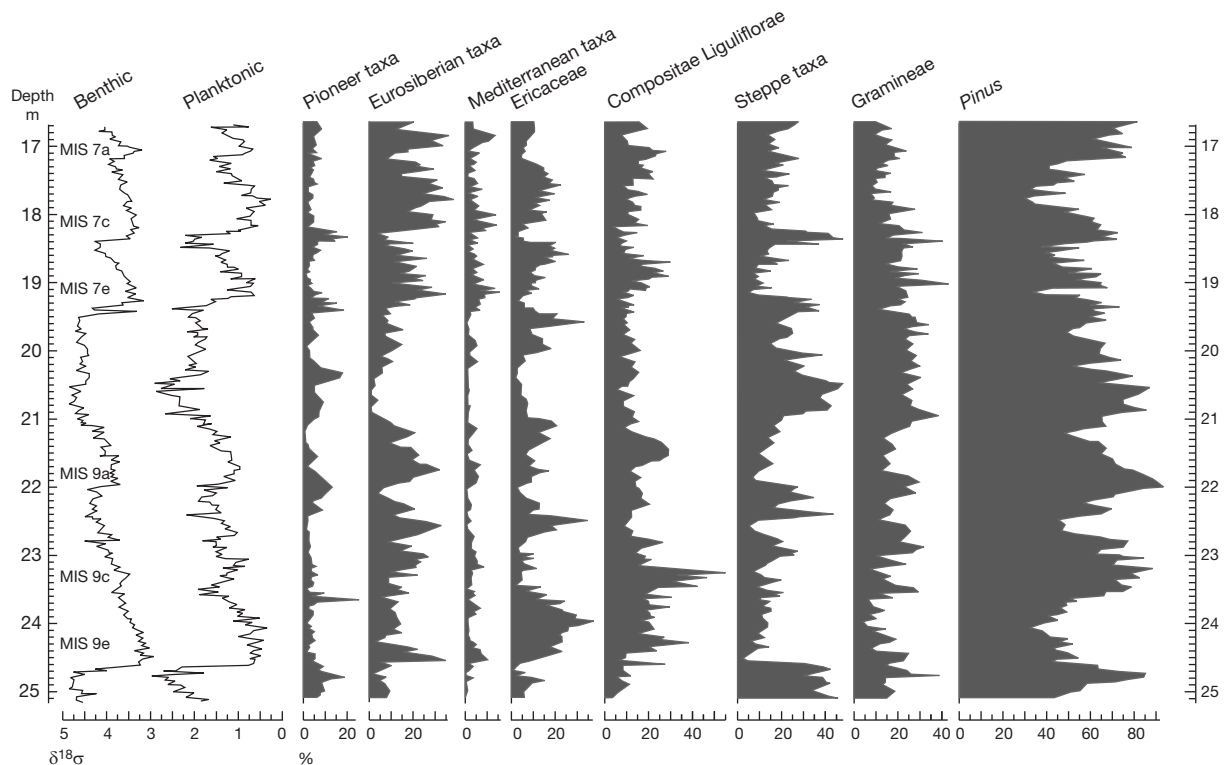


Figure 8 A long, summary pollen diagram from marine sediment core MD01-2443, off the Portuguese coast. Source: Tzedakis PC, Roucoux KH, de Abreu L and Shackleton NJ (2004) The duration of forest stages in southern Europe and interglacial climate variability. *Science* 306: 2231–2235. The corresponding marine isotope stages MIS 9e to MIS 7a and the benthic and planktonic Foraminifera isotope records are indicated on the left. The estimated interval covered by the sequence is 190–350 ka BP.

See also: **Paleobotany:** Overview of Terrestrial Pollen Data. **Pollen Methods and Studies:** Numerical Analysis Methods; POLLSCAPE Model: Simulation Approach for Pollen Representation of Vegetation and Land Cover; Stand-Scale Palynology; Surface Samples and Trapping; The Biome Approach to Reconstructing Past Vegetation; Use of Pollen as Climate Proxies.

References

- Allison TD, Moeller RE, and Davis MB (1986) Pollen in laminated sediments provides evidence for a mid-Holocene forest pathogen outbreak. *Ecology* 67: 1101–1105.
- Andersen ST (1970) The relative pollen productivity and pollen representation of North European trees, and correction factors for tree pollen spectra. *Danmarks Geologiske Undersøgelse, II* 96: 1–99.
- Bennett KD (1997) *Evolution and Ecology: The Pace of Life*. Cambridge: Cambridge University Press p. 241.
- Bennett KD and Lamb HH (1988) Holocene pollen sequences are records of competitive interactions among tree populations. *Trends in Ecology & Evolution* 3: 141–144.
- Bennett KD and Willis K (2003) Pollen analysis. In: Smol JP, Birks HJB, and Last WM (eds.) *Tracking Environmental Change Using Lake Sediments*, pp. 5–32. Dordrecht: Kluwer Academic Publishers.
- Birks HH and Birks HJB (2001) Future uses of pollen analysis must include plant macrofossil analysis. *Journal of Biogeography* 27: 31–35.
- Bradshaw RHW (1988) Spatially precise studies of forest dynamics. In: Huntley B and Webb T III (eds.) *Vegetation History*, pp. 725–751. Dordrecht: Kluwer Academic Publishers.
- Bradshaw RHW and Lindbladh M (2005) Regional spread and stand-scale establishment of *Fagus sylvatica* and *Picea abies* in Scandinavia. *Ecology* 86: 1679–1686.
- Brubaker LB, Anderson PM, Edwards ME, and Lozhkin AV (2004) Beringia as a glacial refugium for boreal trees and shrubs: New perspectives from mapped data. *Journal of Biogeography* 32: 833–848.
- Davis MB (1963) On the theory of pollen analysis. *American Journal of Science* 261: 897–912.
- Davis MB (2000) Palynology after Y2K – Understanding the source area of pollen in sediments. *Annual Review of Earth and Planetary Science* 28: 1–18.
- Davis MB and Deevey E (1964) Pollen accumulation rates: Estimates from late-glacial sediment of Rogers Lake. *Science* 145: 1293–1295.
- Fægri K and Iversen J (1991) *Textbook of Pollen Analysis*. Chichester: Wiley.
- Gaillard M-J, Sugita S, Mazier F, et al. (2010) Holocene land-cover reconstructions for studies on land cover-climate feedbacks. *Climate of the Past* 6: 483–499.
- Glew J, Smol JP, and Last WM (2001) Sediment core collection and extrusion. In: Last WM and Smol JP (eds.) *Tracking Environmental Change Using Lake Sediments: Basin Analysis, Coring, and Chronological Techniques*, vol. 1, pp. 73–105. Dordrecht: Kluwer Academic Publishers.
- Hellman S, Gaillard M-J, Broström A, and Sugita S (2008) The REVEALS model, a new tool to estimate past regional plant abundance from pollen data in large lakes: Validation in Southern Sweden. *Journal of Quaternary Science* 23: 21–42.
- Hjelmroos M and Franzén LG (1994) Implications of recent long-distance pollen transport events for the interpretation of fossil pollen records in Fennoscandia. *Review of Palaeobotany and Palynology* 82: 175–189.
- Huntley B and Birks HJB (1983) *An Atlas of the Past and Present Pollen Maps for Europe: 0–13000 Years Ago*. Cambridge: Cambridge University Press.
- Jacobson G and Bradshaw RHW (1981) The selection of sites for paleovegetational studies. *Quaternary Research* 16: 80–96.
- Lozhkin AV, Anderson PM, Matrosova TV, and Minyuk PS (2007) The pollen records from El'gygytgyn Lake: Implications for vegetation and climate histories of Northern Chukotka since the late middle Pleistocene. *Journal of Paleolimnology* 37: 135–153.
- Margari V, Skinner LC, Tzedakis PC, Ganopolski A, Vautravers M, and Shackleton NJ (2010) The nature of millennial-scale climate variability during the past two glacial periods. *Nature Geoscience* 3: 127–131.
- Moore PD, Webb JA, and Collinson ME (1991) *Pollen Analysis*. Oxford: Blackwell Scientific Publications.
- Parshall T (2002) Late Holocene stand-scale invasion by hemlock (*Tsuga canadensis*) at its Western range limit. *Ecology* 83: 1386–1398.
- Pohl F (1937) Die Pollenerzeugung der Windblütler. *Beihefte zum Botanische Centralblatt* 56: 365–470.
- Prentice IC (1988) Records of vegetation in time and space: The principles of pollen analysis. In: Huntley B and Webb T III (eds.) *Vegetation History*, pp. 17–42. Dordrecht: Kluwer Academic Publishers.
- Prentice IC and Parsons RW (1981) Maximum likelihood linear calibration of pollen spectra in terms of forest composition. *Biometrics* 39: 1051–1057.
- Sanchez-Goni MF, Landais A, Fletcher WJ, Naughton F, Desprat S, and Duprat J (2008) Contrasting impacts of Dansgaard-Oeschger events over a Western European latitudinal transect modulated by orbital parameters. *Quaternary Science Reviews* 27: 1136–1151.
- Seppä H, Alenius T, Giesecke T, Bradshaw RHW, Heikkilä M, and Muukkonen P (2009a) Invasion of Norway spruce (*Picea abies*) and the rise of the boreal ecosystem in Fennoscandia. *Journal of Ecology* 97: 629–640.
- Seppä H, Alenius T, Muukkonen P, Giesecke T, Miller PA, and Ojala AEK (2009b) Pollen accumulation rates as a basis for quantitative tree biomass reconstructions. *The Holocene* 19: 209–220.
- Sugita S (1994) Pollen representation of vegetation in Quaternary sediments: Theory and method in patchy vegetation. *Journal of Ecology* 82: 881–897.
- Sugita S (2007a) Theory of quantitative reconstruction of vegetation I: Pollen from large sites REVEALS regional vegetation composition. *The Holocene* 17: 229–241.
- Sugita S (2007b) Theory of quantitative reconstruction of vegetation II: All you need is LOVE. *The Holocene* 17: 243–257.
- Tzedakis PC, Hooghiemstra H, and Pälike H (2006) The last 1.35 million years at Tenaghi Philippon: Revised chronostratigraphy and long-term vegetation trends. *Quaternary Science Reviews* 25: 3416–3430.
- Väliranta M, Kaakinen A, Kuhry P, Kultti S, Salonen JS, and Seppä H (2011) Scattered late-glacial and early Holocene tree populations as dispersal nuclei for forest development in north-eastern European Russia. *Journal of Biogeography* 38: 922–932.

POLLEN METHODS AND STUDIES

Contents

Use of Pollen as Climate Proxies

Reconstructing Past Biodiversity Development

Numerical Analysis Methods

Databases and Their Application

Surface Samples and Trapping

Stand-Scale Palynology

Changing Plant Distributions and Abundances

The Biome Approach to Reconstructing Past Vegetation

POLLSCAPE Model: Simulation Approach for Pollen Representation of Vegetation and Land Cover

Archaeological Applications

Use of Pollen as Climate Proxies

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Introduction

Climate strongly influences the geographical distribution of plants and of vegetation types because not all plants have evolved the same tolerances for extreme temperatures and the same temperature and moisture requirements for optimal photosynthesis (Woodward, 1987). Past variations of the climate will therefore have influenced the distribution of terrestrial vegetation, and the changes in these distributions may be reconstructed by studying the record of fossil pollen recorded in lake sediments and peat bogs (see [Pollen Analysis, Principles](#)). However, the relation between the pollen composition of lake sediments (pollen diversity and relative abundance) and the species composition of the surrounding vegetation is not straightforward because pollen production, dispersion, and preservation vary from one species to the other. In order to reconstruct past climates from pollen data, it is therefore necessary to directly relate modern pollen data to climatic values, and then apply this relation to fossil pollen samples.

The increasing concern over the current and future changes in climate has led to greater interest in the use of pollen studies to quantify past climate changes, notably for intervals of large variations in climate such as the transition between glacial and interglacial intervals. The results obtained by the transformation of pollen data into climate equally form an ideal source of data for testing the output of global circulation models during the past (paleo-GCMs), a process that allows an assessment of these models for future predictions under different conditions (see [Data–Model Comparisons](#)).

Past variations of the climate have influenced the distribution of terrestrial vegetation, and the changes in these distributions

may be reconstructed by studying the record of fossil pollen recorded in lake sediments and peat bogs (see [Pollen Analysis, Principles](#)). The underlying principle of all attempts to reconstruct climate from pollen data is that of uniformitarianism, that is, the information required for understanding past conditions may be obtained from the relationship between the modern distribution of the taxa concerned and the climate (Birks and Birks, 1980). The distribution of each taxon is controlled by a set of climatic limits, and combinations of taxa indicate a particular set of conditions, with greater or lesser accuracy. By relating the modern distribution of taxa and climate, and subsequently applying this relationship to the observed assemblage of fossil pollen taxa, it is then possible to reconstruct the climate that shaped past plant communities, by using the presence/absence or the relative abundance of the various pollen taxa identified ([Figure 1](#)).

Theoretically, pollen-based climate reconstructions are derived from an inversion of functions that describe the way in which biological organisms respond to their environment. In practice, non-causal statistical equations linking climate and modern proxy data are calculated and then applied to fossil assemblages to reconstruct past climates.

In any form of climate reconstruction based on a transfer function, it is assumed that (1) modern pollen assemblages relate to their present-day environment, (2) the climatic parameter to be reconstructed is either a controlling factor on the abundance of the taxa or is linearly related to another controlling factor, (3) the ecological responses of the taxa have not changed with time, (4) variations in the modern data set are unaffected, or only negligibly, by environmental factors other than the one to be reconstructed (Birks, 1995).

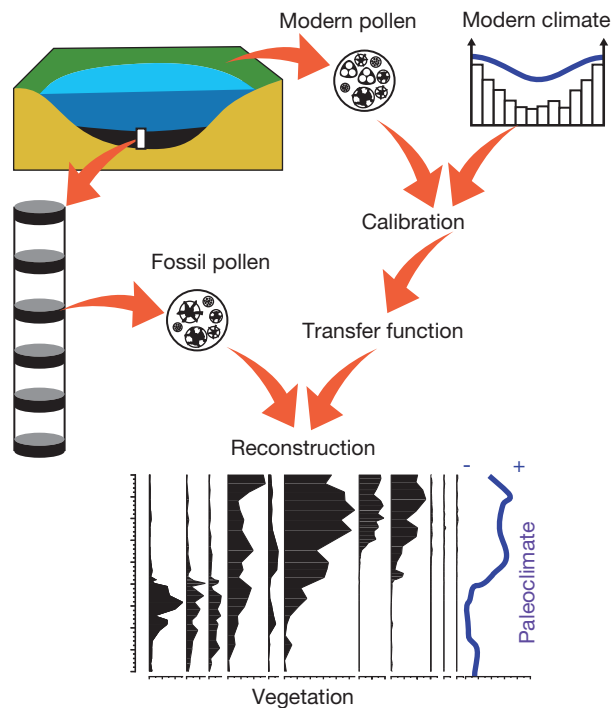


Figure 1 Summary diagram illustrating the steps necessary for the construction of a transfer function. (Modified from an unpublished diagram by S. Juggins.) Modern samples are taken from environments corresponding to those of the fossil samples. A transfer function is formed between the modern samples and the corresponding modern climatic parameter. The transfer function is then applied to the series of fossil samples, to provide an estimation of paleoclimatic changes.

The most intuitive approach is to compare the present-day distributions of selected species with the corresponding distribution of climate variables thought to be critical in determining the distributions of each species. This approach, called indicator species, is univariate because the species are analyzed separately and with respect to one climatic variable at a given time. However, this approach is somewhat limited, because in reality species respond to combinations of climatic variables and are limited by different climatic factors in different parts of their ranges. Different climatic factors are intercorrelated and these correlations are a function of the climatic regimes in which the species live. The relationship of a species to climate may change within its ecological range due to differentiation of physiological ecotypes. The climatic influence on species is made more complex by the competitive effects of other species present.

Multivariate indicator species approaches are preferable, because they avoid these problems to some extent by simultaneously considering the relationship of several species' distributions to more than one climatic variable (Iversen, 1944; Fauquette et al., 1998; Pross et al., 2000). However, these methods still ignore climatically relevant information in the proxy data, notably in terms of species assemblages and species abundances.

Multivariate statistical approaches developed during the last three decades provide the most powerful approach. Transfer function equations are derived by multivariate techniques from an adequate calibration sample of modern distributional data, and applied to paleo-samples to estimate environmental

parameters (Birks, 1995). Although the basis of these methods remains the same, several different methods have been developed over the last few decades for establishing and then inverting the pollen–climate relation.

Estimates of past climates may be obtained for a time series of pollen samples obtained from a single site, or several sites from the same region, giving a continuous time series of climate values (Guiot et al., 1989; Seppä and Birks, 2001). With the advent of continental-scale databases of pollen sequences, various studies have examined the spatial extent of climate changes over a wide area (Cheddadi et al., 1997; Sawada et al., 2004). These studies, which require careful correlation between sites (Dating Techniques), have tended to focus on single time periods, classically the mid-Holocene. Given the low spatial resolution of most climate models, this time-slice approach is appropriate for comparing paleo-GCM output with proxy data climate reconstructions.

In order to transform the information contained in pollen assemblages into climatic values, different approaches and different methods have been developed since the 1940s. In this article, we first review the various different types of methods and approaches that exist, followed by a comparison of the strengths and weaknesses of these methods. Finally, we present three case studies showing climate reconstructions at different temporal and spatial scales.

Advantages and Disadvantages of Pollen Data for Climate Reconstructions

There are now a large number of different types of fossil data that can be used in climate reconstructions, including fossil Coleoptera (see Overview), Chironomids (see Chironomid Overview), plant macrofossils (see Dendroarchaeology; Dendroclimatology), and each type has its own strengths and weaknesses.

Among the advantages of pollen data (Moore et al., 1991), some are as follows:

- (1) Information about the presence of taxa, pollen samples provide a quantified idea of the composition of the vegetation around the site.
- (2) The relatively small amount of sediment required for pollen analysis allows studies to be made at a relatively high temporal resolution.
- (3) Although the temporal resolution is lower than most annual proxy archives (e.g., see Dendrochronology; Varved Lake Sediments), most pollen sequences will cover a longer period of time, and hopefully include a greater variety of climatic changes.
- (4) Sites containing a good pollen record are relatively easy to find in humid regions, although this is more problematic in drier regions where pollen preservation is worse.
- (5) Pollen analysis is a well-established method, having been used for well over 100 years, and there is an extensive catalog of existing literature.

The main disadvantages are as follows:

- (1) Due to differences in pollen productivity, transport, and preservation between different taxa, the relationship between the composition of the pollen assemblage and that of the vegetation is extremely complex and nonlinear (see

POLLSCAPE Model: Simulation Approach for Pollen Representation of Vegetation and Land Cover

- (2) The identification of pollen grains is frequently limited by morphological similarities between grains from different species, and often only the genus can be established. This lack of taxonomic information limits the precision available in the reconstructed climate because many genera include species with contrasting climatic limits.
- (3) Different pollen sites record a different sample of the surrounding vegetation due to topographic differences and basin size. Small forest hollows record an extremely local vegetation signal, usually the vegetation within 10–100 m of the site, and a reconstructed climate will be affected by any local microclimate (see [Stand-Scale Palynology](#)). Large lake basins record a regional or even extra-regional pollen signal ([Jacobson and Bradshaw, 1981](#)), and the climate reconstructed will therefore be more similar to a mean regional climate.

Methods Based on Similarity

Mutual Climate Range Methods

The mutual climate range (MCR) methods use climatic limits (e.g., temperature extremes) for presence of pollen taxa, established from the modern distribution of one or several plant species. A possible range of climate values for a given fossil sample is estimated by combining the modern limits of all taxa found (see [Figure 2](#)).

In the simplest approach, a single taxon with a well-defined climatic tolerance is used to estimate a change in climate by its presence or absence. For example, the dwarf birch, *Betula nana*, which does not occur currently in lowland British Isles, is observed in this region during the Late Glacial interval. By establishing that the modern distribution of *B. nana* is

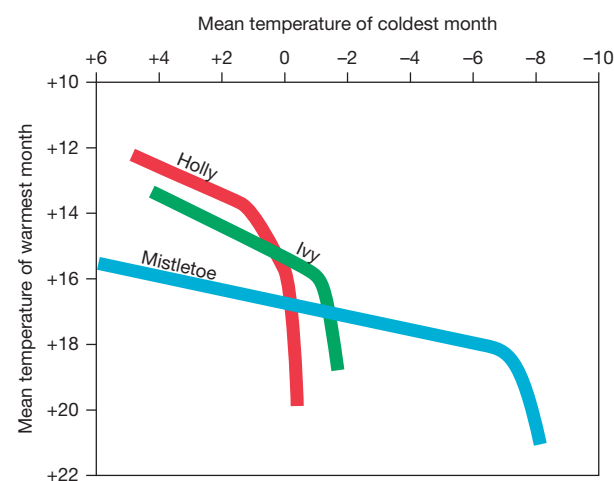


Figure 2 The climatic limits of holly (*Ilex aquifolium*), ivy (*Hedera helix*), and mistletoe (*Viscum album*). Above these limits, the species are not present. Using this diagram, it is therefore possible to estimate past temperatures within a certain range, according to the presence or absence of these three species in a pollen sample. Modified from Iversen J (1944) *Viscum, Hedera and Ilex* as climatic indicators. *Geologiska Föreningen Förhandlingar* 66: 463–483.

limited by maximum summer temperatures over 22 °C, [Conolly and Dahl \(1970\)](#) estimated that during the Late Glacial, the maximum summer temperature in lowland Britain did not exceed 22 °C, compared to a modern value of 30 °C.

A better constraint can be obtained by using two environmental variables and/or several species in parallel. Using the present-day distribution of ivy (*Hedera helix*), holly (*Ilex aquifolium*), and mistletoe (*Viscum album*), constrained by specific winter and summer temperatures, [Iversen \(1944\)](#) estimated the climatic changes of the mid-Holocene in Denmark to within a defined range, showing an increase of 2–3 °C in summer temperature and an increase of 1–2 °C in winter temperatures ([Figure 2](#)).

The complexity can be increased, for example, using climatic envelopes defined with many environmental variables and for larger sets of taxa. In order to reconstruct the climate for a fossil sample, the envelopes for all taxa present are superimposed and the resulting overlap is considered to be the possible paleoclimatic range ([Figure 3](#)). The method has been applied to pollen data from northeastern Australia by [Kershaw and Nix \(1988\)](#), showing warmer and wetter conditions during the mid-Holocene.

In order to improve the methods, [Fauquette et al. \(1998\)](#) have made use of the taxa abundances to refine the climatic range identified by the presence of a given pollen taxon, by defining a set of abundance classes. For example the elm, *Ulmus*, is considered to be present at 0.5%, with an annual temperature range between -3 and 22 °C. Once the percentage passes a defined threshold, the taxon is considered to be abundant, and the climatic range is reduced (*Ulmus* is considered abundant over 8%, and the annual range shrinks to between 4 and 17 °C).

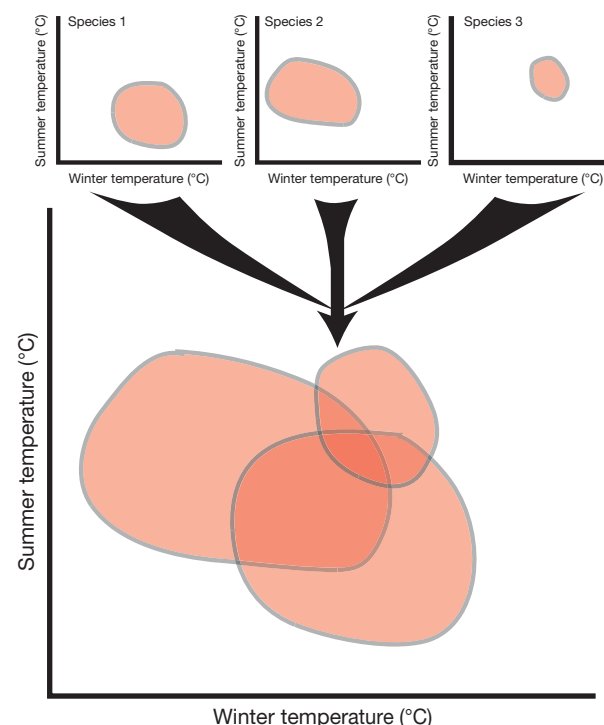


Figure 3 Synthetic mutual climatic range diagram. Three species' climatic ranges are shown. A sample containing all three of these species is assumed to have climatic envelope defined by the overlap of the three ranges.

Finally, Pross et al. (2000) have improved on the method in order to further narrow the climatic ranges obtained. Climatic ranges were reconstructed for a set of synthetic but possible pollen assemblages and compared against the actual modern climate for that synthetic assemblage. Because the actual modern climate range for a given assemblage is narrower than the potential climate range predicted by pure indicator principle, they conclude that the MCR technique exaggerates the ranges and they take into account actual ranges to obtain the most likely climatic intervals.

Because MCR methods do not depend on a relation between modern taxa assemblages and climate values, they work well when the vegetation composition has no modern analog. However, the distributions of one or more taxa in climate space may be discrete and no overlap occurs. This is usually due to a lack of modern distributional data, and will bias the estimates obtained.

Modern Analog Technique

The modern analog technique (MAT) (Guiot, 1990; Jackson and Williams, 2004) uses a large data set of modern pollen samples, and consists of selecting a set of modern samples (or analogs) that most closely resemble each fossil pollen sample. The resemblance is assessed by comparing modern and fossil pollen assemblages in terms of taxa composition and abundance. In most studies, the dissimilarity between assemblages is quantified using squared-chord distances, which consist of a prior transformation of pollen percentages by taking the square root in order to reduce the influence of the few relatively abundant taxa in the samples (Figure 4).

Once a set of modern analogs has been selected, climatic values can be attributed to the fossil sample as a weighted average of the climate of the modern samples. Climatic

estimates at the exact location of each modern pollen-sampling site are obtained by interpolation from a set of meteorological stations. The weights used are the inverse of the chord distance, which allows the most similar analogs to have the greater influence on the climate estimates obtained. The number of analogs chosen is usually fixed at the start of the analysis, but Waelbroeck et al. (1998) suggested using the analog distances: when a sharp increase in distance occurs, no further analogs are selected. Using a modern data set of 227 samples and this method, Guiot et al. (1989) reconstructed annual temperature and precipitation from two long cores from central France, and provided climatic time series that cover the last 140 ka, including both the start and end of the last glacial interval.

The MAT suffers from a number of problems, including assemblages containing taxa that respond to different climatic parameters, the estimation of errors, and dealing with intervals in which the past vegetation has no modern analog. As with the MCR methods, a number of adaptations and improvements have been made to deal with these problems.

In order to deal with the problem of taxa responding to different climatic parameters, Guiot (1990) have used 'bioclimatic' loadings to preferentially weight the distance calculated between the modern and fossil samples. An ordination is made of the modern samples, and the distribution of taxa are examined in order to identify the axis that is most strongly related to the climatic parameter under investigation. The taxa scores along this axis are then used as weights.

The error involved in assigning climatic values is usually estimated from the dispersion of the chosen analogs around the weighted average. This method has the advantage of considering the higher and lower errors separately, resulting in nonsymmetric and more realistic error bars; however, in doing so the total error is underestimated. Nakagawa et al. (2002) have proposed an improved method that uses a linear relationship calculated between observed and estimated climate values at modern pollen sites. The confidence intervals obtained for the linear relation are then applied to past estimates. This requires a dense network of meteorological stations and may not be practical in all regions.

One significant problem for the MAT technique is situations where no good modern analog can be found for a fossil assemblage. These no-analog situations result from vegetation dynamics, anthropogenic influence and lack of adequate modern samples. Davis et al. (2003) proposed a modification to the technique to help avoid this problem, whereby the modern and fossil assemblages were regrouped into plant functional types (PFTs), and analog matching is carried out on the basis of the PFT abundances (scores). Because each PFT represents a set of ecologically similar plant taxa (Prentice et al., 1996), the assumption can be made that a similar climate will always contain a similar composition of PFTs, even if the species composition may vary for nonclimatic reasons.

To further improve the method, information derived from a second climate proxy may be used to constrain the climatic estimates. Cheddadi et al. (1997) estimated paleoclimate from pollen data using MAT constrained with lake-level data. The constraint consists of restricting the set of modern pollen samples considered as analogs of the fossil samples to those locations where the implied change in annual precipitation minus evapotranspiration ($P - E$) is consistent with the regional change in

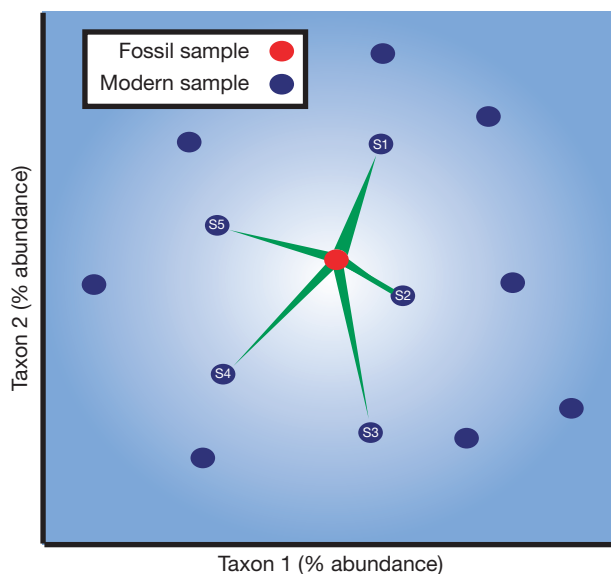


Figure 4 Synthetic diagram of the modern analog technique. The samples are compared according to the percentage abundance of two taxa. In this example, the fossil sample (red) is closest to five samples, and the climate to be assigned will be a weighted average of the climate associated with these five samples. Sample S2, being the most similar, will have the greatest weight in the estimation.

moisture balance as indicated by lake-level data (see [Overview](#)). The constraint led to an improvement in the spatial coherency of the reconstructed paleoclimate anomalies, especially for P – E. The use of a second proxy as a constraint has the disadvantage that it can no longer be used to independently verify the results; however, if the climatic response of the proxy is well understood, this can offer a more consistent set of results.

One of the principal advantages of the MAT method is that it is well adapted to large data sets, and has been extensively used for large-scale reconstructions ([Cheddadi et al., 1997](#); [Sawada et al., 2004](#)), in part because it does not rely on a response function dominated by a single climate variable, and thus allows reconstructions in areas where the composition of plant communities is limited by different parameters. Recent work by Telford and co-workers, however, has shown that the apparent better performance of MAT, when compared to the calibration methods listed below, may be overexaggerated by lack of full statistical independence and spatial autocorrelation of the climate variable.

Response Surface

The analogs described above do not always provide an adequate set of climatic indicators. To emphasize the signal to noise ratio, [Bartlein et al. \(1986\)](#) proposed a method in which the pollen assemblages are projected onto an environmental space formed by the climatic parameters of interest. Surfaces are then fitted to the abundance values of a set of selected taxa by locally weighted regression ([Cleveland and Devlin, 1988](#)). Having fitted response surfaces for the suite of taxa that will be used to make the reconstructions, the values for the taxa at each grid point can be combined to provide a series of assemblages that may be considered as potential analogs for each fossil assemblage. Thereafter the procedure is very similar to that used in direct quantitative analog methods, with the five to ten closest matching assemblages for each fossil assemblage being identified and used to provide a reconstruction and associated confidence intervals. Due to limited interpolation and extrapolation, response surfaces are less affected by the problem of no-analog vegetation. However, the smoothing of the relation between taxa abundances and climate can lead to an underestimation of true paleoclimate values ([Figure 5](#)).

Webb et al. (1993) have used this technique to estimate seasonal temperatures and annual precipitation across eastern North America for several time intervals since the Last Glacial Maximum (LGM), and have shown that, given an adequate data set, large-scale climatic gradients, and the changes in those gradients, can be reconstructed from pollen data.

Methods Based on Calibration

In contrast to the methods based on similarity, those based on calibration establish a mathematical function (a transfer function, Webb and Bryson (1972)), which relates taxa abundances to climate parameters, and so allows estimates of past climatic conditions. Reconstruction using a transfer function takes place in two stages. In the first, a response function is established ($X = R(C,D)$), which links the abundance of the pollen taxa (X) to a set of climatic (C) and nonclimatic (D) controlling factors. If the climate is considered to be the major

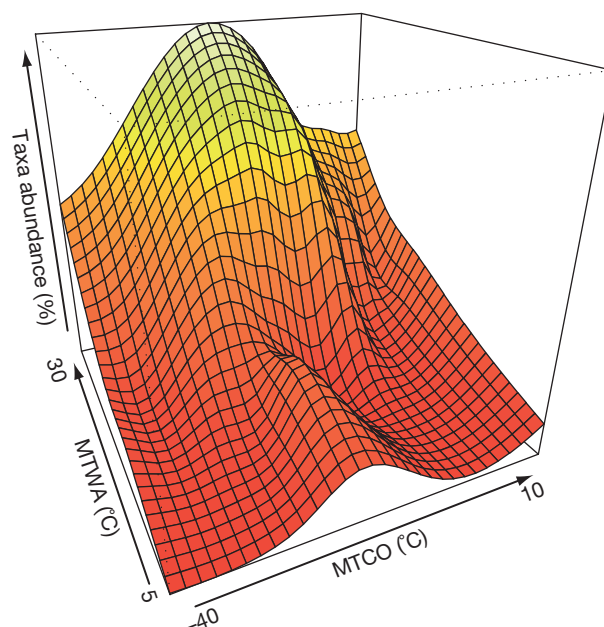


Figure 5 Example response surface. The response surface is fitted to two climate parameters: winter temperatures (MTCO) and summer temperatures (MTWA). The example taxon has its highest abundances at high summer and intermediate winter temperatures.

controlling factor, then the full response function can be approximated to ($X = R(C)$). Once the response function is established, it may be used for past reconstructions by inverting the function to give the transfer function ($C = T(X)$). As before, a number of techniques exist, which are characterized by the method in which the response function is constructed, and below we describe how the response functions for three of these methods are established, followed by a comparative study ([Lotter et al., 2000](#)). This is followed by a description of the use of non-linear neural networks. For further detail, the reader is referred to [Birks \(1995\)](#) and the chapter on reconstruction methods (see [Approaches](#)).

Partial Least Squares Analysis

Partial least squares (PLS) analysis is a modified principal components analysis (PCA) that forces the first axis or component to maximize the covariance between one environmental variable and a linear combination of taxa abundances, thus forming a direct link between the taxa and a single climatic parameter ([Figure 6a](#)). Subsequent components may also be included to further maximize the covariance, but these are constrained to be orthogonal to the preceding component and therefore uncorrelated. If this technique is used with several environmental parameters, it is known as canonical correspondence analysis, and has been used in pollen-based climate reconstructions by Webb and Bryson (1972).

Weighted Averaging and Weighted Average – Partial Least Squares

The preceding method assumes a linear relationship (ascending or descending) between taxa abundances and climatic parameters ([Figure 6a](#)). Although a linear response may be assumed over small ranges of climatic parameters (i.e., short

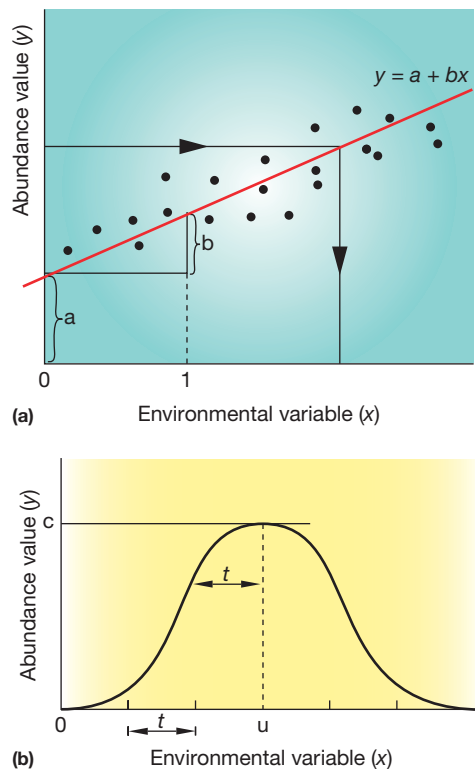


Figure 6 Two possible response functions. (a) Linear; (b) unimodal.

environmental gradients), a more realistic ecological response function is the unimodal relationship (Figure 6b), in which taxon abundance varies across the environmental gradient, with an optimum at a particular value. The technique of weighted averaging (WA) (ter Braak and Looman, 1986) allows these unimodal response functions to be used in climatic reconstructions, and is considered as the unimodal equivalent of linear regression and calibration (Birks, 1995).

The regression step of WA consists of estimating the response function for a group of selected taxa for a given climatic or environmental parameter. For this, all the modern samples where that taxon is present are used, and the optimum is estimated as the average of the climate of these samples, weighted by the abundance of the taxon. Once established, the past climate parameter is estimated as the weighted average of the optimum of all taxa present in a fossil sample, where the weights are the abundances. The estimates obtained suffer from edge effects, which cause underestimates of extreme values. A correction for this has been proposed by ter Braak and Juggins (1993). As before, errors are calculated by bootstrapping (Efron and Tibshirani, 1993). The main advantages of the WA method are that it uses a theoretically robust response model and copes well in no-analog situations. However, the method may be affected by an uneven distribution of the modern sites along the environmental gradient. Further, any extra indirect covariance between the taxa and the environment is ignored, once the climate variable is taken into account.

Weighted average–partial least squares (WA-PLS) avoids this latter problem by allowing further components to be chosen, as in PLS above, which (1) improve the covariance between the weighted averages and the climate; and (2) must

be orthogonal to the preceding component. These secondary components represent other environmental factors that constrain the composition of a given pollen assemblage (ter Braak and Juggins, 1993).

In a high-resolution study of lake sediments, Lotter et al. (2000) have tested and compared the use of PLS, WA, and WA-PLS for the reconstruction of summer temperatures at Gerzensee, Switzerland. A modern pollen data set of 126 samples was used to establish the transfer functions, which were subsequently applied to 118 fossil samples covering the Late Glacial interval. Although root mean squared error of prediction (RMSEP) tests on the modern data set showed WA-PLS to outperform PLS and WA, the differences were slight. Paleotemperatures reconstructed using WA-PLS temperatures covered a larger range than the other techniques, but all methods showed comparable values, particularly when the errors were taken into account. Lotter et al. (2000) suggest that, although the gradient of temperatures covered by the modern data set was large, the set included both taxa which had an optimum outside of the gradient, indicating a linear approach, and taxa whose optima were included in the gradient, indicating a unimodal approach. Finally, a consensus curve was calculated by using all the predictions for the study of Late Glacial climate change. The temperature curve obtained corresponded well to an oxygen isotope curve from the same site, but showed lags in the response of vegetation to the climatic changes.

Artificial Neural Networks

Artificial neural networks (ANNs) are used for a large variety of nonlinear problems and so lend themselves well to the problem of calibrating pollen assemblages and climatic variables. The classic forward-feed backpropagation network has been used in reconstructions for several proxy–environmental data sets (e.g., pollen, tree rings, diatoms). Generally the networks consist of a set of input nodes, which correspond to each taxon, a hidden layer of nodes and a set of output nodes, which correspond to the climatic parameters (Figure 7). The number of nodes in the hidden layer may vary, depending on the level of complexity required. Each hidden node is, in fact, a coefficient, which is adjusted during a period of training using modern samples and their corresponding climate, until the assemblages can be best transformed into climate. As with the MAT method, ANNs require a large modern data set so that they may be accurately trained. Once the network is established, it can be used to estimate past climates, by entering the taxa abundances found in fossil samples. The method has been used in a climatic reconstruction for a number of European pollen sites at the LGM by Peyron et al. (1998) to give a spatially coherent picture of the conditions of this interval, with a marked northwest–southeast gradient between an extremely cold and dry northwest to milder, although still cold and dry, southeast.

Examples of Paleoclimate Reconstruction from Pollen Data

The methods described above have been applied in a wide range of different regions of the world and in applications at different scales of time and space. Here, we review three recent studies that cover different scales, albeit limited to the

Holocene Epoch: a high temporal resolution study from a single site, mapped climate changes for a continent, and a continual regional reconstruction for the last 12 ka.

Single Site/Time Series

Seppä and Birks (2001) used WA-PLS to reconstruct changes in summer temperature and annual precipitation for a core taken from Lake Tsuolbmajavri in northern Finland. The transfer function was set up using a data set of 304 modern pollen samples from Norway, Finland, and northern Sweden, covering a gradient of approximately 10 °C summer temperature and 2500 mm annual precipitation. A two-component WA-

PLS model was chosen, which, after cross-validation, showed RMSEP errors of about 1 °C and 340 mm and a coefficient of determination (r^2) over 0.7.

Summer temperature and annual precipitation were reconstructed along the core at a resolution of ca. 70 years, covering the last 10 ka (Figure 8). The reconstruction, which began with a cool and moist early Holocene, showed clearly the Holocene thermal maximum between ca. 8.2 and 5.7 ka BP, with temperatures that were up to 2 °C warmer than the present day. The latter part of the Holocene shows a return to cooler and more humid conditions. Seppä and Birks (2001) have linked these changes to large-scale changes in the atmospheric circulation across Fennoscandia, with the warm mid-Holocene

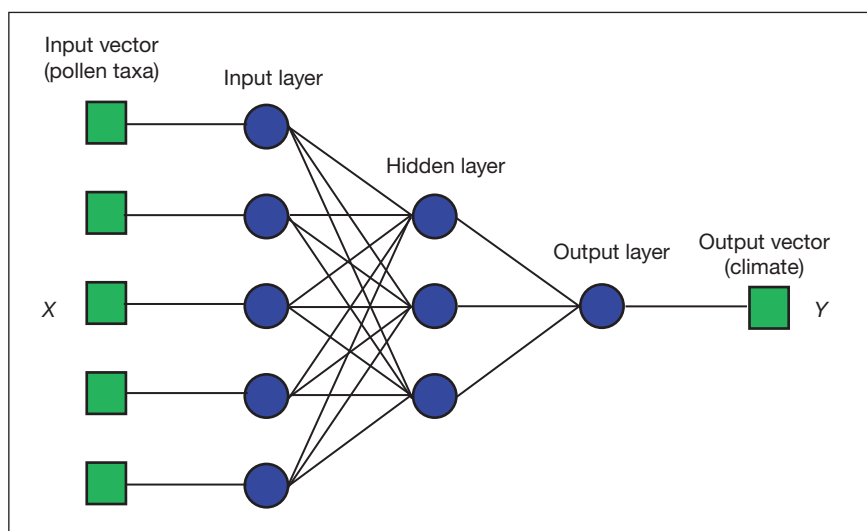


Figure 7 A backpropagation forward-feed neural network with one hidden layer. At the left of the diagram is the input vector. Each input vector represents one pollen sample with, here, five taxa. These abundances are then iteratively related to the output vector, the climate parameter of interest.

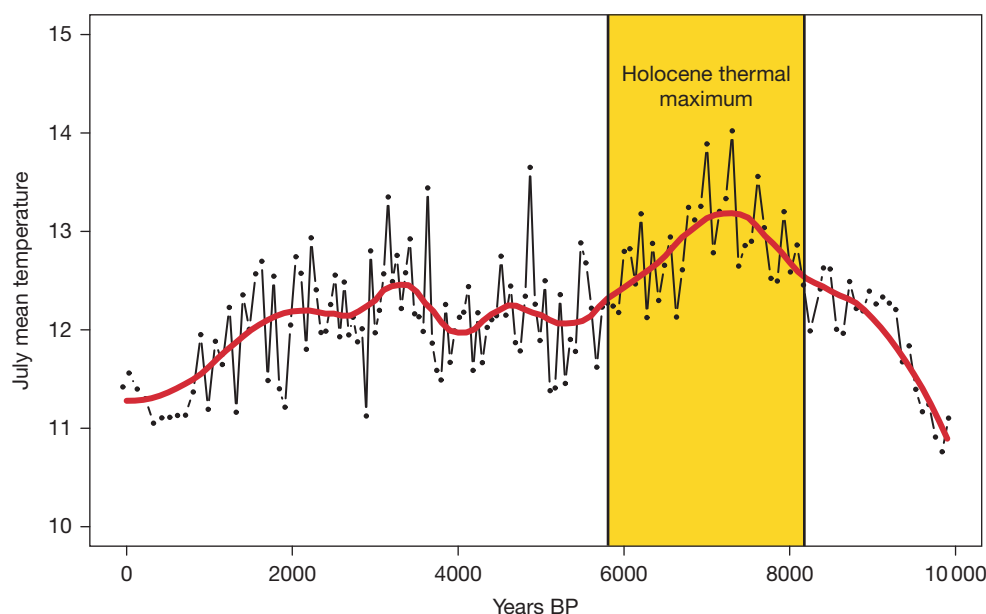


Figure 8 Reconstructed summer temperature from Tsuolbmajavri in northern Finland. Modified from Seppä H and Birks HJB (2001) July mean temperature and annual precipitation trends during the Holocene in the Fennoscandian tree-line area: Pollen-based climate reconstructions. *The Holocene* 11: 527–539.

conditions resulting from the establishment of anticyclonic conditions, in turn caused by the loss of the Fennoscandinavian ice sheet and higher summer insolation. The later cooling follows the reduction in insolation in the later Holocene.

Time-Slice/Spatial Study

To provide data for comparison with the output of an atmospheric GCM, Sawada et al. (2004) estimated the change in summer temperatures for the North American continent at 6 ka BP. Such 'time-slice' reconstructions at large spatial scales have become increasingly feasible, due to the establishment of continental-scale databases of modern and fossil pollen samples (see [Databases and Their Application](#)). The training data set consisted of 4590 pollen samples, and associated temperature values, distributed across the continent, and samples from 545 sites were available for the mid-Holocene. A modified MAT technique was

used to estimate temperature at each of the fossil sites. The modifications consisted of an increase in the number of taxa used to match samples and attempts to more objectively identify both poor analogs and the number of analogs to be used.

The results for the 545 sites were interpolated onto a gridded surface, and then the amount of change in temperature values (i.e., the anomaly) between 6 ka and the present was calculated by subtracting the mid-Holocene surface from an interpolated modern climate. The results ([Figure 9](#)) showed a general trend from a warmer north to a cooler south, albeit with a number of local differences such as a slight cooling in central and north-western Canada. The warming in the north has been linked directly to the increased insolation during the mid-Holocene. However, Sawada et al. (2004) suggest that the southern cooling may result from an inadequate set of modern samples. Mid-Holocene cooling at mid-latitudes has also been observed in previous continental scale studies from both North America

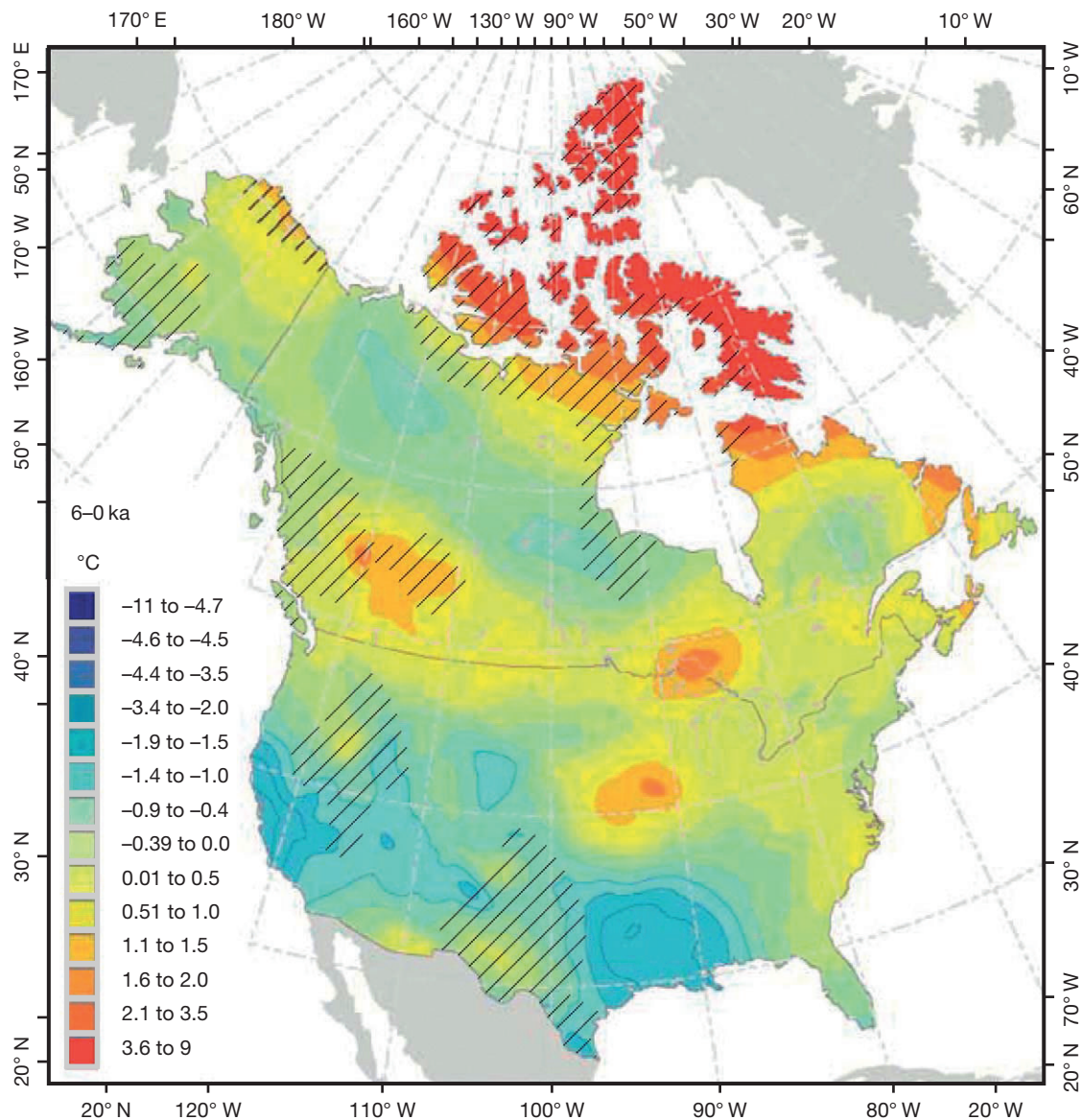


Figure 9 Map of summer temperature anomalies for the North America continent at 6 ka BP. Reproduced from Sawada M, Viau A, Vettoretti G, Peltier WR, and Gajewski K (2004) Comparison of North American pollen-based temperature and global lake-status with CCMA AGCM2 output at 6 ka. *Quaternary Science Reviews* 23: 225–244, Copyright 2004, with permission from Elsevier.

(Gajewski et al., 2000) and Europe (Cheddadi et al., 1997), and appears to be a consistent feature of Holocene climate change.

Continuous Holocene Study

Time-slice reconstructions are static, in the sense that each time period is considered separately. By combining the site-based approach with the time-slice approach in order to improve the signal-to-noise ratio, Davis et al. (2003) introduced an approach in which the regional large-scale changes are considered in a continuous way, allowing an investigation of the evolution of large-scale climate. Again, this study relies on the existence of continental-scale databases.

The study used a modern data set of 2363 samples distributed across Europe and North Africa. These were used to estimate temperature parameters (summer, winter, and annual) for each sample of Holocene age from a set of 510 sites, giving approximately 68 000 climate estimates distributed across Europe and throughout the last 12 ka. Estimates were made using a modified MAT method, in which pollen taxa were classed into vegetation groups to limit the no-analog problem. These were interpolated onto a regular grid in both time and space and then area-averaged into six regions.

The results (Figure 10) showed quite different dynamics of temperature change in the various regions of Europe. Notably, the Holocene thermal maximum appears confined to Northern

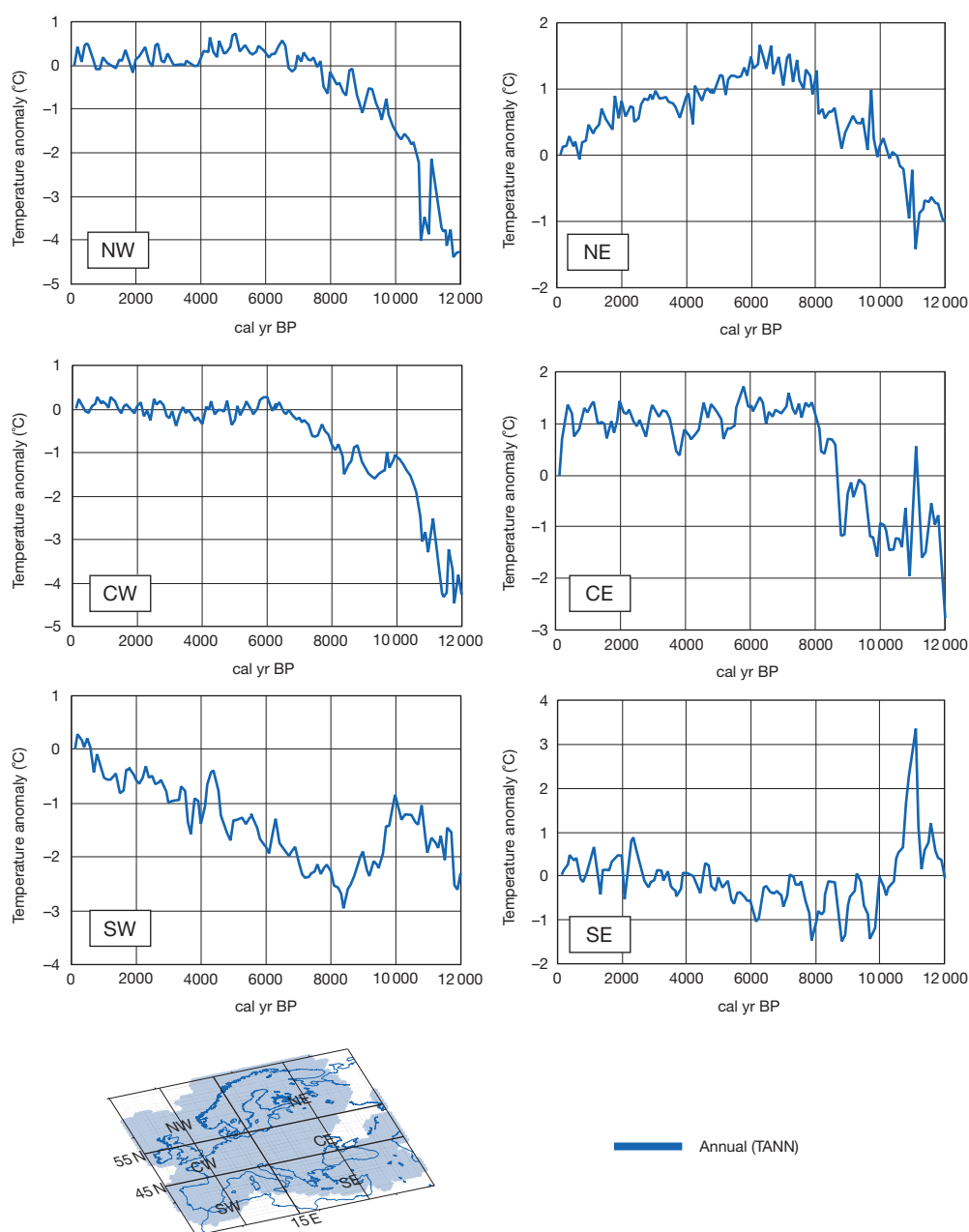


Figure 10 Annual temperature anomalies for six regions of Europe. The inset shows the geographical limits of the regions. Reproduced from Davis BAS, Brewer S, Stevenson AC, Guiot J, and Data Contributors (2003) The temperature of Europe during the Holocene reconstructed from pollen data. *Quaternary Science Reviews* 22: 1701–1716, Copyright 2004, with permission from Elsevier.

Europe, while in the south, temperatures decreased during the early Holocene and remained cooler than present for much of the period. This suggests that the cooler mid-latitudes observed in the spatial studies of mid-Holocene climate (see above) are not anomalous, but instead form part of a general opposing trend of temperature changes between the north and south.

Perspectives

Perhaps the greatest problem facing paleoclimate reconstructions today is that the distribution of the vegetation is assumed to be uniquely controlled by the climate, and nonclimatic parameters, for example, CO₂ concentration or soil depth, are not taken into account. During intervals of substantial changes in these parameters, for example, the last glacial interval, it may be expected that their effects on the vegetation composition may bias any paleoclimatic reconstruction. Vegetation models are the only existing tool that allow the CO₂ effect to be taken into account, and there is therefore interest in using the models for climatic reconstruction, by inverse modeling of pollen data. This approach uses a vegetation model to explore a large number of possible climatic scenarios in order to find the scenario that corresponds to the vegetation observed in a fossil pollen sample (Guiot et al., 2000). Past changes in CO₂ may be set in the model, allowing this bias to be reduced. Work on this approach, together with improvements in our knowledge of the variations of nonclimatic factors, will help to provide more robust future reconstructions.

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See also: **Ice Core Methods:** Overview. **Paleoceanography, Physical and Chemical Proxies:** Oxygen-Isotope Stratigraphy of the Oceans. **Paleoclimate Reconstruction:** Sub-Milankovitch (DO/Heinrich) Events. **Quaternary Stratigraphy:** Chronostratigraphy; Continental Biostratigraphy; Lithostratigraphy; Pseudostratigraphy.

References

- Bartlein PJ, Prentice IC, and Webb T III (1986) Climatic response surfaces from pollen data for some eastern North American taxa. *Journal of Biogeography* 13: 35–57.
- Birks HJB (1995) Quantitative palaeoenvironmental reconstructions. In: Maddy D and Brew JS (eds.) *Statistical Modelling of Quaternary Science Data*, pp. 161–254. Cambridge: Quaternary Research Association.
- Birks HJB and Birks HH (1980) *Quaternary Palaeoecology*. London: Edward Arnold.
- Cheddadi R, Yu G, Guiot J, Harrison SP, and Prentice IC (1997) The climate of Europe 6000 years ago. *Climate Dynamics* 13: 1–9.
- Cleveland WS and Devlin SJ (1988) Locally-weighted regression: An approach to regression analysis by local fitting. *Journal of the American Statistical Association* 83: 596–610.
- Conolly AP and Dahl E (1970) Maximum summer temperature in relation to modern and quaternary distributions of certain arctic-montane species in the British Isles. In: Walker D and West RG (eds.) *Studies in the Vegetation History of the British Isles*, pp. 159–223. Cambridge: Cambridge University Press.
- Davis BAS, Brewer S, Stevenson AC, and Guiot J (2003) Data Contributors. The temperature of Europe during the Holocene reconstructed from pollen data. *Quaternary Science Reviews* 22: 1701–1716.
- Efron B and Tibshirani RJ (1993) *An Introduction to the Bootstrap*, p. 450. London: Chapman and Hall.
- Fauquette S, Guiot J, and Suc J-P (1998) A method for climatic reconstruction of the Mediterranean Pliocene using pollen data. *Palaeogeography, Palaeoclimatology, Palaeoecology* 144: 183–201.
- Gajewski K, Vance R, Sawada M, et al. (2000) The climate of North America and adjacent waters at 6 ka. *Canadian Journal of Earth Sciences* 37: 661–681.
- Guiot J (1990) Methodology of the last climatic cycle reconstruction in France from pollen data. *Palaeogeography, Palaeoclimatology, Palaeoecology* 80: 49–69.
- Guiot J, Pons A, De Beaulieu J-L, and Reille M (1989) A 140000-year climatic reconstruction from two European pollen records. *Nature* 338: 309–313.
- Guiot J, Torre F, Jolly D, Peyron O, Boreux JJ, and Cheddadi R (2000) Inverse vegetation modelling by Monte Carlo sampling to reconstruct palaeoclimates under changed precipitation seasonality and CO₂ conditions: Application to glacial climate in Mediterranean region. *Ecological Modelling* 127: 119–140.
- Iversen J (1944) *Viscum, Hedera and Ilex* as climatic indicators. *Geologiska Föreningen Förhandlingar* 66: 463–483.
- Jackson ST and Williams JW (2004) Modern analogs in Quaternary paleoecology: Here today, gone tomorrow? *Annual Review of Earth and Planetary Sciences* 32: 495–537.
- Jacobson GL and Bradshaw RH (1981) The selection of sites for palaeoenvironmental studies. *Quaternary Research* 16: 80–96.
- Kershaw AP and Nix HA (1988) Quantitative paleoclimatic estimates from pollen data using bioclimatic profiles of extant taxa. *Journal of Biogeography* 15: 589–602.
- Lotter AF, Birks HJB, Eicher U, Hofmann W, Schwander J, and Wick L (2000) Younger Dryas and Allerød summer temperatures at Gerzensee (Switzerland) inferred from fossil pollen and cladoceran assemblages. *Palaeogeography, Palaeoclimatology, Palaeoecology* 159: 349–361.
- Moore PD, Webb JA, and Collinson ME (1991) *Pollen Analysis*. Oxford: Blackwell Scientific Publications.
- Nakagawa T, Tarasov PE, Nishidac K, Gotanda K, and Yasuda Y (2002) Quantitative pollen-based climate reconstruction in central Japan: Application to surface and late quaternary spectra. *Quaternary Science Reviews* 21: 2099–2113.
- Peyron O, Guiot J, Cheddadi R, et al. (1998) Climatic reconstruction in Europe for 18,000 year BP from pollen data. *Quaternary Research* 49: 183–196.
- Prentice IC, Guiot J, Huntley B, Jolly D, and Cheddadi R (1996) Reconstructing biomes from palaeoecological data: A general method and its application to European pollen data at 0 and 6 ka. *Climate Dynamics* 12: 185–194.
- Pross J, Klotz S, and Mosbrugger V (2000) Reconstructing palaeotemperatures for the early and middle Pleistocene using the mutual climatic range method based on plant fossils. *Quaternary Science Reviews* 19: 1785–1799.
- Sawada M, Vau A, Vettoretti G, Peltier WR, and Gajewski K (2004) Comparison of North American pollen-based temperature and global lake-status with CCMA AGCM2 output at 6 ka. *Quaternary Science Reviews* 23: 225–244.
- Seppä H and Birks HJB (2001) July mean temperature and annual precipitation trends during the Holocene in the Fennoscandian tree-line area: Pollen-based climate reconstructions. *The Holocene* 11: 527–539.
- ter Braak CJF and Juggins S (1993) Weighted averaging partial least squares regression (WA-PLS): An improved method for reconstructing environmental parameters from species assemblages. *Hydrobiologia* 269/270: 485–502.
- ter Braak CJF and Looman CWN (1986) Weighted averaging, logistic regression and the Gaussian response model. *Vegetatio* 65: 3–11.
- Waelbroeck C, Labeyrie L, Duplessy JC, et al. (1998) Improving past sea surface temperature estimates based on planktonic fossil faunas. *Paleoceanography* 13: 272–283.

Webb T III and Bryson RA (1972) Late and postglacial climatic changes in the Northern Midwest, USA: Quantitative estimates derived from fossil pollen spectra by multivariate statistical analysis. *Quaternary Research* 2(1): 70–115.

Webb T III, Bartlein PJ, Harrison SP, and Anderson KH (1993) Vegetation, lake-level and climate change in eastern North America. In: Wright HE Jr., Kutzbach JE,

Webb T III, Ruddiman WF, Street-Perrott FA, and Bartlein PJ (eds.) *Global Climates Since the Last Glacial Maximum*, pp. 415–467. Minneapolis: University of Minnesota Press.

Woodward FI (1987) *Climate and Plant Distribution*. London, England: Cambridge University Press.

Reconstructing Past Biodiversity Development

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Introduction

Concerns for the preservation of modern biodiversity in light of climate change and increasingly intensive land use, erosion, and human interference with ecological systems have led to stronger interest in the geological record of changes in past biodiversity. The concept of biodiversity includes genetic variation and variation in species and ecosystem composition. Although exciting progress has taken place recently in our understanding of past genetic diversity (see [Ancient Plant DNA](#)), most paleoecological data record species associations or higher levels of organization of individuals. The most widely available type of paleoecological data is pollen, which, in this discussion, also includes tracheophyte and bryophyte spores that are usually counted during pollen analysis.

Measuring Biodiversity

Following [Whittaker \(1977\)](#), vegetation diversity may relate to communities (alpha diversity), landscapes (gamma diversity), or regions (epsilon diversity). Similarly, changes between communities (beta diversity) or between landscapes (delta diversity) can be defined. Such scales are useful also in a paleoecological context where pollen data – depending on the size of basin – broadly reflect the alpha to epsilon diversities ([Figure 1](#)). In contrast, because of their shorter distance of transport, plant macrofossils mostly reflect the alpha and beta levels of diversity of past vegetation.

The most obvious aspect of vegetation diversity is species richness, which is simply the number of species. In a community of limited extent, the number of plant species may be rather accurately enumerated, but in larger areas, usually only a sample of the total species richness can be investigated. However, the relation between species richness of the sample and that of the entire area is not simple. This is because not only the species richness of the entire unit of vegetation but also the relative frequency of each species is important for how many species are recorded in the sample. In other words, sample species richness is also dependent on evenness. Evenness is a measure of the distribution of individuals for each species (the relative frequency of species). A community in which all species have the same number of individuals (are equally common) is said to have maximum evenness, while any other combination of frequencies results in lower evenness ([Figure 2](#)). As more individuals are registered in a sample, the number of species or taxa grows, a relationship that is depicted in so-called species growth curves. Evenness determines the shape of such species growth curves: it is steep when evenness is high and more gradual when evenness is low ([Figure 3](#)). This means that if evenness is at a maximum, then all species are likely be represented in even a small sample comprising only a few percent of the individuals of the community. In contrast,

if evenness is very low, then only a fraction of the species present in the entire population is likely to be found in even a large sample comprising almost half the individuals of the community.

Many diversity indices have been developed; most of them are composite measures of species richness and evenness. Several of these indices have been applied to fossil records. As explained above, even species richness is influenced by evenness, unless the total species richness of the entire population is recorded. A useful index of evenness is

$$E_{1/D} = \frac{1 / \sum_{i=1}^S p_i^2}{S}$$

where p_i is the proportion of taxon i and S is the total number of taxa in sample. $E_{1/D}$ is independent of richness and theoretically ranges from almost zero (when one taxon is very dominant) to one at maximum evenness ([Smith and Wilson, 1996](#)).

Biases in Fossil Records of Biodiversity

There are three obvious biases when using samples of fossil assemblages as reflections of past biodiversity: taxonomic precision, sample size, and pollen productivity.

Taxonomic Precision

Plant macrofossils can often be identified to specific or at least generic level, but pollen can usually only be identified to genus or often only to family level. This lack of taxonomic precision with pollen data causes a nonlinear relationship between the plant species number and palynological richness (number of pollen taxa). This means that the number of pollen taxa in an assemblage cannot be taken as a direct measure of the number of plant species in the vegetation from which that assemblage was derived. With a low number of species, the relationship is almost 1:1, but in rich communities, the relationship may be 2:1 or even higher. [Odgaard \(1994\)](#) found the relationship to be almost identical in three modern communities with different disturbance regimes and used this finding to correct fossil palynological richness for bias in taxonomic precision.

Sample Size and Pollen Productivity

When pollen is analyzed in a sample of a geological deposit, the microfossil count (for instance, 1000 specimens) represents an infinitesimal fraction of the number of pollen grains once produced in the area of interest and during the period of interest ([Figure 4](#)). Because of the small sample-to-population ratio, the evenness of the population (i.e., all the pollen grains from the area and period of interest) is theoretically important for how many microfossil types (taxa) are found in the sample.

The fact that the observed number of taxa increases with sample size (Figure 3) has led to the common notion that if taxon numbers are to be compared between samples, then the number of individuals in each sample should be identical. Because it is often impractical to tally exactly the same number

of individuals in each sample, especially during pollen counting, some technique is needed to estimate the number of taxa that would have been found in a sample if a lower count had been made. Rarefaction is such a technique. However, it is very important to note that rarefaction does not allow extrapolation to taxon numbers in a larger sample, only estimation of taxon numbers at lower counts than the actual.

Developed in the late 1960s and early 1970s, rarefaction was introduced to paleoecology in the 1970s (Tipper, 1979). It was first used in Quaternary pollen analysis in the 1980s (Birks and Line, 1992; Birks et al., 1988). Intuitively speaking, rarefaction uses the information of frequency of each taxon in an assemblage to estimate at what lower count than the actual count a taxon would be represented by its first individual. For example, a taxon with a frequency of 5% (one out of 20) would be expected to be found with one individual out of the first 20 individuals counted plus or minus an error term. The rarefaction estimate is defined by

$$E(T_n) = \sum_{i=1}^T 1 - \left[\frac{(N - N_i)!(N - n)!}{(N - N_i - n)!N!} \right]$$

where $E(T_n)$ is the expected number of taxa at sample size n ; N is the population size; T is the number of taxa in population, and N_i is the number of individuals of taxon i . In paleoecology, the population size is not known, and here, the actual number of individuals in the samples is taken as the population size,

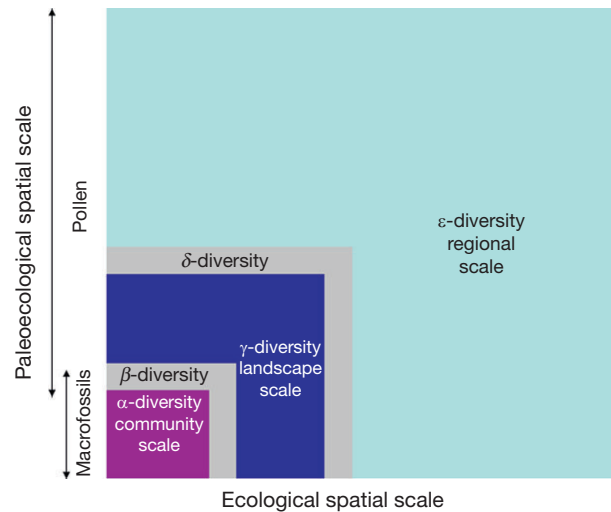


Figure 1 Ecological spatial scales of diversity (*sensu* Whittaker, 1977) compared to the spatial scales reflected by paleoecological studies using pollen and plant macrofossils.

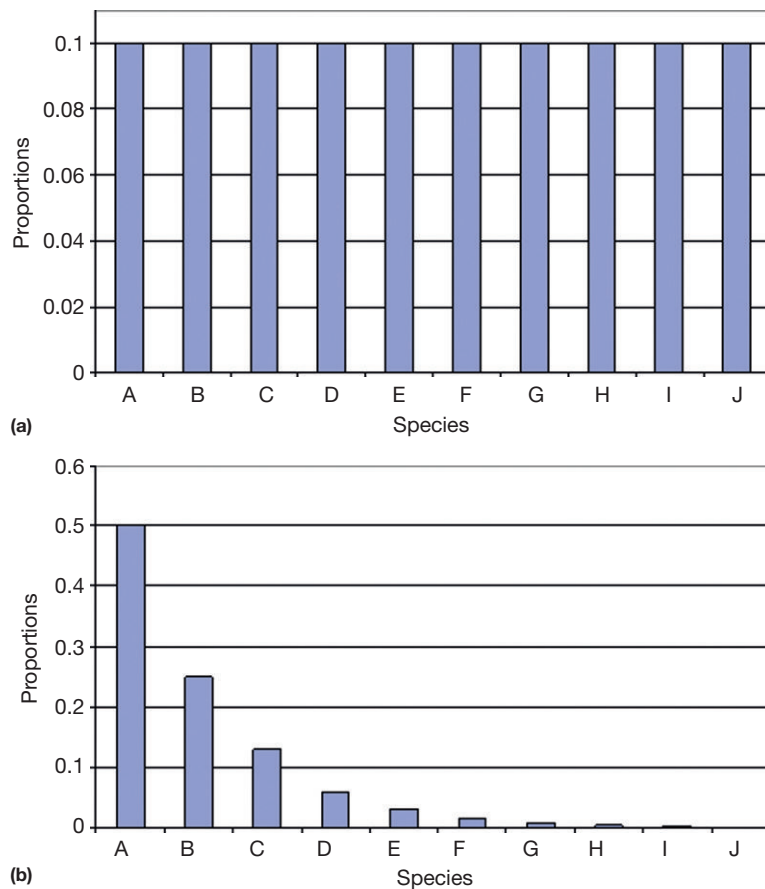


Figure 2 Frequency distributions of assemblages of 10 species (A–J) showing maximum (a) and intermediate evenness (b).

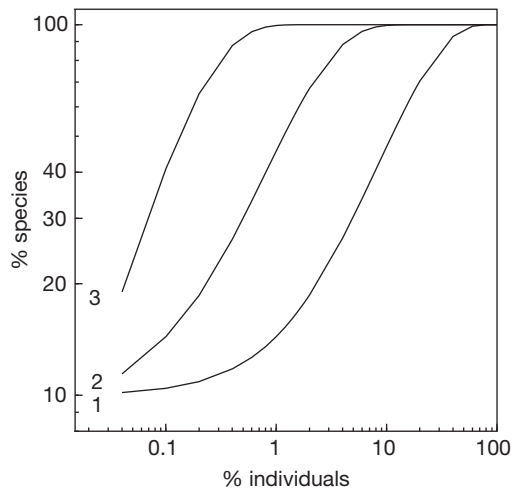
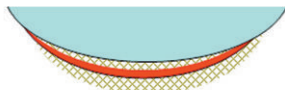


Figure 3 Species accumulation curves showing number of species versus number of individuals for three different assemblages with low (1), intermediate (2), and high (3) evenness.



(a) : 10^{18}



(b) : 10^{12}



(c) : 10^6



(d) : 10^3

Figure 4 Orders of magnitude of pollen grains: (a) produced in the pollen source area of a medium-sized lake (10 ha) during a few decades, (b) fossilized in the lake sediment of interest, (c) sampled by the paleoecologist, and (d) counted on the microscope slide.

while n is the lower number of individuals for which taxon richness is to be estimated.

Rarefaction estimates are often inappropriately interpreted as an index of past species richness. For example, stratigraphic plots of rarefaction estimates based on pollen data tend to be generally parallel to plots of nontree pollen (Figure 5), and this may superficially be interpreted as a reflection of generally higher species richness in nonwoodland communities. It is important to note, however, that such estimates may theoretically be strongly biased by the evenness of the assemblages and by variable population sizes, as well as varying taxonomic precision. The study of Räsänen et al. (2004) provides an

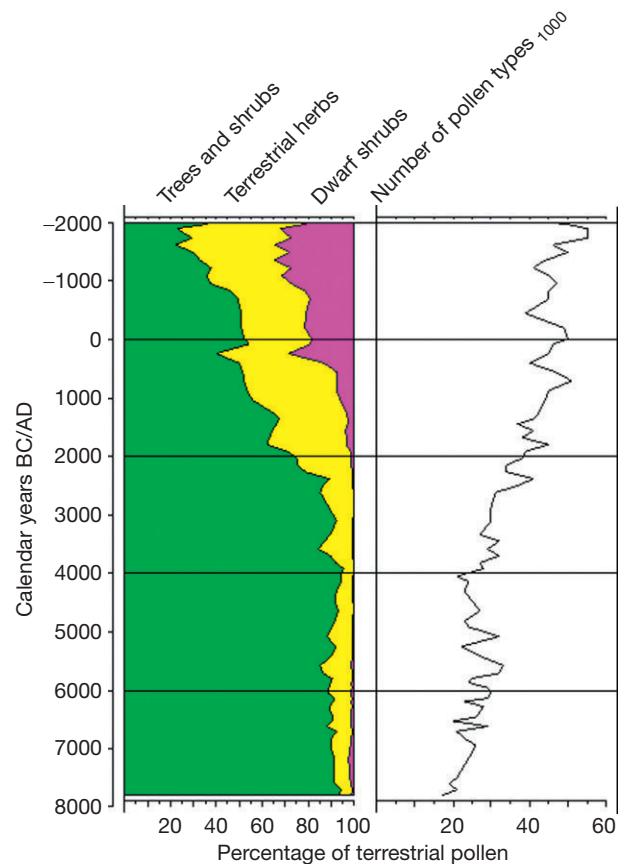


Figure 5 Survey pollen diagram from Holocene sediments from lake Navn Sø, Denmark, showing percentages of pollen from trees (green), grasses and herbs (yellow), and heather (violet), as well as the number of terrestrial pollen types at constant count of 1000 pollen grains. Adapted from Odgaard BV (1999) Fossil pollen as a record of past biodiversity. *Journal of Biogeography* 26: 7–17.

example where rarefaction estimates of palynological richness and evenness of the pollen assemblages correlate significantly, indicating that observed richness is biased by evenness. Furthermore, the underlying assumption in using a constant sample size is that population size is constant throughout the time series, and this is probably only rarely the case. If population sizes vary while sample size is kept constant, the counted fraction of the population varies, and taxon richness may accordingly covary. Odgaard (1999) and van der Knaap (2009) suggested workarounds for the dependence of observed microfossil type richness on population size. Using estimates of relative (Odgaard, 1999) or absolute (van der Knaap, 2009) pollen accumulation rates, these procedures attempt to estimate the number of pollen taxa that would have been found, provided the same fraction of the pollen population (the pollen production of the area) had been tallied.

Interpreting Records of Sample Diversity

This section gives a few examples of pollen-based biodiversity studies illustrating the use of various approaches and numerical methods.

The Dynamics of Landscape Biodiversity

Smith and Cloutman (1988) studied the development of a 15 ha bog in Wales using pollen records from 13 open sections or cores. The chronology was based on radiocarbon dates. Because of the high number of well-dated pollen stratigraphies in a small area, Smith and Cloutman (1988) were able to reconstruct the Holocene landscape variation of the local vegetation. The pollen records reflected the on-site hydrosereal vegetation as well as the vegetation surrounding the site. The large number of sites allowed an interpretation of the past landscapes into units, such as mixed woodland, hazel woodland, blanket peat, grassland, and dry heath. The time series of reconstructions clearly reflect the decrease of biodiversity at landscape level (gamma diversity) as blanket peat expanded over the area (Table 1).

Table 1 Main vegetational units identified by Smith and Cloutman (1988) in six different time slices of the Waun-Fignen-Felen peatland of South Wales. Ages are in radiocarbon years BP

	8000	7500	6500	5700	4700	3700
Grassland	●				●	
Birch/hazel scrub	●					
Hazel woodland	●					
Mixed woodland	●	●	●		●	●
Hazel scrub		●	●			
Reed swamp	●	●	●	●	●	
Carr	●	●	●	●	●	
Blanket peat		●	●	●	●	●
Acid peat			●	●		
Dry heath		●	●	●		
Damp heath			●			

Controls of Palynological Richness

Peros and Gajewski (2008) studied the number of pollen taxa from arctic nonarctic plants in surface lake sediments from northern Canada and concluded that the latitudinal trend of observed palynological richness, rarefacted to counts of 300 or 100, was opposite to the trend documented for the species richness of the vegetation. This finding suggested that palynological richness in that area is a very poor indicator of landscape scale species richness. Using pollen concentration as a proxy of landscape scale pollen productivity, Peros and Gajewski (2008) also found that palynological richness was negatively correlated to pollen productivity in the same dataset. In contrast, they found a positive correlation between pollen sample evenness and palynological richness (Figure 6). The same relations between palynological richness, concentration, and evenness were found in Holocene samples from the area. Peros and Gajewski (2008) were unable to efficiently separate the effects of pollen productivity and evenness, but their data suggested that pollen productivity was especially effective in controlling palynological richness at low productivity gradients.

The importance of the pollen productivity control of palynological richness was strongly underlined by a recent study by Meltsov et al. (2011). Comparing floristic and palynological richness along an openness gradient in Estonia, Meltsov et al. (2011) found that sediment samples from open vegetation (within 250 m from sediment samples) consistently had more pollen taxa than samples from more wooded surroundings. However, the additional pollen taxa in samples from sites with more open vegetation were widespread pollen taxa, virtually unrelated to local vegetation communities. Also, no relation between pollen sample evenness and palynological richness was found in this study. Furthermore, the majority

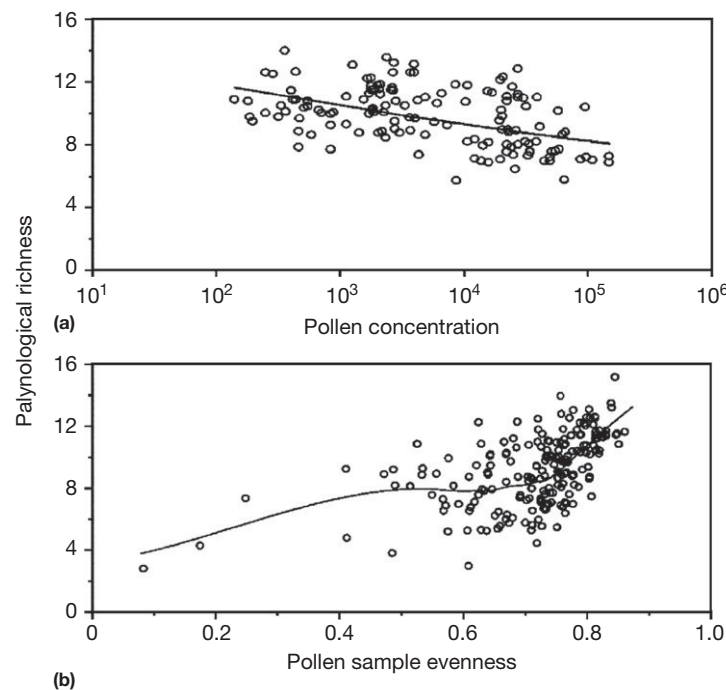


Figure 6 Plot of rarefacted palynological richness (base count 100) against pollen concentration and pollen sample evenness (Hurlburt's probability of interspecific encounter) in North American arctic lake surface samples. Reproduced from Peros MC and Gajewski K (2008) Testing the reliability of pollen-based diversity estimates. *Journal of Paleolimnology* 40: 357–368, with permission from Springer.

of species in the vegetation (two-thirds of the local pollen types) were conspicuously absent from the pollen counts ($n > 1600$) of the sedimentary samples.

These results strongly underline the control exerted by overall pollen productivity on the observed pollen sample richness. Because pollen productivity is lower in open landscapes than in densely wooded areas, widespread but less frequent pollen types are more effectively recorded in samples from open landscapes than from closed ones. Rarefaction estimates of pollen sample richness are, accordingly, a complex reflection of many processes such as pollen population evenness, pollen dispersal, landscape pollen productivity, and possibly floristic richness. Although more work is needed to resolve these complexities, pollen productivity seems a much more important control of palynological richness than does floristic richness.

See also: [Pollen Analysis, Principles](#); [Paleobotany: Ancient Plant DNA](#). **Pollen Methods and Studies:** [Numerical Analysis Methods](#); [Use of Pollen as Climate Proxies](#).

References

- Birks HJB and Line JM (1992) The use of rarefaction analysis for estimating palynological richness from Quaternary pollen-analytical data. *The Holocene* 2: 1–10.
- Birks HJB, Line JM, and Persson T (1988) Quantitative estimation of human impact on cultural landscape development. In: Birks HH, Birks HJB, Kaland PE, and Moe D (eds.) *The Cultural Landscape – Past, Present and Future*, pp. 229–240. Cambridge: Cambridge University Press.
- Meltsov V, Poska A, Odgaard BV, et al. (2011) Palynological richness and pollen sample evenness in relation to local floristic diversity in Southern Estonia. *Review of Palaeobotany and Palynology* 166: 344–351.
- Odgaard BV (1994) Holocene vegetation history of Northwestern Jutland. *Opera Botanica* 123: 1–171.
- Odgaard BV (1999) Fossil pollen as a record of past biodiversity. *Journal of Biogeography* 26: 7–17.
- Peros MC and Gajewski K (2008) Testing the reliability of pollen-based diversity estimates. *Journal of Paleolimnology* 40: 357–368.
- Räsänen S, Hicks S, and Odgaard BV (2004) Pollen deposition in mosses and in a modified Tauber trap from Hailuoto, Finland: What exactly do the mosses record? *Review of Palaeobotany and Palynology* 129: 103–116.
- Smith AG and Cloutman EW (1988) Reconstruction of Holocene vegetation history in three dimensions at Waun-Fignen-Felen, an upland site in South Wales. *Philosophical Transactions of the Royal Society of London, B* 322(1209): 159–219.
- Smith B and Wilson JB (1996) A consumer's guide to evenness indices. *Oikos* 76: 70–82.
- Tipper JC (1979) Rarefaction and rarefaction – The use and abuse of a method in paleoecology. *Paleobiology* 5: 423–434.
- van der Knaap WO (2009) Estimating pollen diversity from pollen accumulation rates: A method to assess taxonomic richness in the landscape. *The Holocene* 19: 159–163.
- Whittaker RH (1977) Evolution of species diversity in land communities. *Evolutionary Biology* 10: 1–67.

Numerical Analysis Methods

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Introduction

Pollen-analytical (palynological) data are counts of different pollen and spore types present in one or more sediment samples. The samples or objects can be from one sediment core collected from the study site (temporal, stratigraphical data) or from one age or depth at several sites (spatial, geographical data). The most common type of spatial data consists of pollen counts from the uppermost, surficial sediments, so-called surface samples. Stratigraphical and spatial data commonly contain counts of many (50–250) different pollen and spore types or taxa in a large number of samples (50–500). Such data are highly multivariate, containing many variables (pollen and spore types) and many objects (sediment samples). In pollen analysis, the number of objects is usually much larger than the number of variables, in contrast to, for example, diatom analysis where the number of taxa usually exceeds the number of samples. With the development of national or continental databases of palynological data, data-sets can also consist of pollen counts from many different cores from a large number of sites over a broad geographical area, thereby combining both temporal and spatial data (Brewer et al., 2012; Mitchell, 2011).

Pollen counts are almost always expressed as percentages or proportions of some calculation or pollen sum (e.g., total terrestrial pollen and spores), thereby transforming the counts into relative proportional or percentage compositional data. The sample sum itself is relatively uninformative in many numerical analyses. It is largely determined by the initial research and sampling design, namely, the number of individual pollen and spores to be counted per sample. Numerical properties of relative palynological data are that the data are ‘closed,’ there may be many zero values (absences) in the data, and there may be many taxa. Closed data require special methods for correct statistical analysis. Reyment and Savazzi (1999) discuss statistical approaches to analyzing closed data involving log-ratios. The large number of zero values can, however, cause problems in the use of log-ratios. Pollen counts are more rarely expressed as pollen concentrations (grains per unit volume (grains cm^{-3}) or dry weight of sediment (grains gm^{-1})). If the sediment matrix accumulation rate (net thickness of sediment accumulation per unit time; cm year^{-1}) can be determined by, for example, age-depth modeling, pollen accumulation rates (PAR) (‘pollen influx’; $\text{grains cm}^{-2} \text{year}^{-1}$) can be estimated. Such ‘absolute’ data are harder to obtain than relative counts. Although their statistical properties are less complex than relative data, absolute data have a high inherent variance and some variance-stabilizing transformation is usually essential prior to numerical analysis (Birks and Gordon, 1985).

A pollen count is a statistical *sample* count of the total amount of pollen present in the sediment level of interest. This count is an estimate of the underlying statistical *population* of interest. Despite the technique of Quaternary pollen analysis being nearly 100 years old, the relevant statistical methods for

assessing the inherent statistical uncertainties attached to pollen counts were only relatively recently developed by J.E. Mosimann, L.J. Maher, and K.D. Bennett. The application of multivariate data-analytical techniques such as cluster analysis and principal components analysis (PCA) to pollen-analytical data began in 1970 through the pioneering work of D.P. Adam, M.B. Dale, D. Walker, H.S. Cole, T. Webb III, E.J. Cushing, A.D. Gordon, and others. Since then, the use of statistical procedures and multivariate numerical techniques has increased greatly (Birks, 1992; Birks and Gordon, 1985) for data collection and assessment, data summarization, data analysis, and data interpretation.

Numerical methods are useful tools at many stages in a palynological investigation. During data collection, they can be invaluable in identifying certain morphologically similar pollen types. In data assessment, statistical techniques are essential in estimating the inherent errors associated with the resulting pollen counts. A range of numerical techniques can be used to detect and summarize major underlying patterns in palynological data. For single stratigraphical sequences, the numerical delimitation of pollen-assemblage zones can be an important step in data summarization. Other numerical techniques such as ordination procedures can summarize temporal or spatial trends in one or several data-sets. Numerical methods are essential for quantitative data analysis in terms of estimating palynological richness, detecting population changes, modeling population responses, quantifying rates of change, detecting temporal patterns such as periodicities and power spectra in time series, and reconstructing quantitatively past climate. The final stage of data interpretation can be greatly aided by numerical techniques in the reconstruction of past vegetation and in the testing of competing hypotheses about underlying causative factors such as climate change and human activity in determining pollen-stratigraphical changes. As broader and more complex questions are asked of palynological data, numerical methods are being constantly updated and modified in an attempt to help answer new research questions (Birks, 2012a).

The ability to interpret output from computer-based numerical analyses (e.g., PCA) in a rigorous and critical way both numerically and palynologically requires that the investigator has a good understanding of the underlying mathematical theory and the biological and geological assumptions of the numerical methods being used. Table 1 lists valuable texts about many of the relevant numerical methods used with palynological data. Table 2 lists texts about numerical techniques in pre-Quaternary deep-time geology, paleoecology, and paleobiology, many of which are relevant in the quantitative analysis of Quaternary palynological data. The texts listed in Tables 1 and 2 are valuable introductions to many of the numerical methods that form part of the tool kit for Quaternary pollen analysis in the twenty-first century.

Because of space limitations, references to primary sources are mainly restricted to publications not covered in the reviews

Table 1 Important reference texts about numerical methods used with Quaternary pollen-analytical data

- Birks HJB and Gordon AD (1985) *Numerical Methods in Quaternary Pollen Analysis*. London: Academic Press.
- Borcard D, Gillet F, and Legendre P (2011) *Numerical Ecology with R*. New York: Springer.
- Fielding AH (2007) *Cluster and Classification Techniques for the Biosciences*. Cambridge: Cambridge University Press.
- Gordon AD (1999) *Classification*. London: Chapman & Hall.
- Gotelli NJ and Ellison AM (2004) *A Primer of Ecological Statistics*. Sunderland, MA: Sinauer Associates.
- Hanrahan G (2009) *Environmental Chemometrics*. Boca Raton, FL: CRC.
- Jongman RHG, ter Braak CJF, and van Tongeren OFR (1987) *Data Analysis in Community and Landscape Ecology*. Wageningen: Pudoc.
- Legendre P and Legendre L (1998) *Numerical Ecology*, 2nd English edn. Amsterdam: Elsevier.
- Lepš J and Šmilauer P (2003) *Multivariate Analysis of Ecological Data using CANOCO*. Cambridge: Cambridge University Press.
- Maddy D and Brew JS (eds.) (1995) *Statistical Modeling of Quaternary Science Data*. Quaternary Research Association Technical Guide 5, Cambridge.
- Manly BFJ (2007) *Randomization, Bootstrap, and Monte Carlo Methods in Biology*, 3rd edn. Boca Raton, FL: Chapman & Hall.
- McCune B, Grace JB, and Urban DL (2002) *Analysis of Ecological Communities*. Gleneden Beach, OR: MjM Software Design.
- Quinn GP and Keough MJ (2002) *Experimental Design and Data Analysis for Biologists*. Cambridge: Cambridge University Press.
- Sokal RR and Rohlf FL (2012) *Biometry*, 3rd edn. New York: W.H. Freeman.
- Zuur AF, Ieno EN, and Smith GM (2007) *Analysing Ecological Data*. New York: Springer.

Table 2 Important reference texts about numerical methods used with pre-Quaternary deep-time, geological, paleontological, and paleoecological data

- Carr JR (1995) *Numerical Analysis for the Geological Sciences*. Englewood Cliffs, NJ: Prentice Hall.
- Davis JC (2002) *Statistics and Data Analysis in Geology*, 3rd edn. New York: Wiley.
- Hammer Ø and Harper DAT (2006) *Paleontological Data Analysis*. Oxford: Blackwell.
- Reyment RA (1971). *Introduction to Quantitative Paleocology*. Amsterdam: Elsevier.
- Reyment RA (1980) *Morphometric Methods in Biostratigraphy*. London and New York: Academic Press.
- Reyment RA (1991) *Multidimensional Palaeobiology*. Oxford: Pergamon.
- Reyment RA and Jöreskog KG (1993) *Applied Factor Analysis in the Natural Sciences*. Cambridge: Cambridge University Press.
- Reyment RA and Savazzi E (1999) *Aspects of Multivariate Statistical Analysis in Geology*. Amsterdam: Elsevier.
- Reyment RA, Blackith RE, and Campbell NA (1984) *Multivariate Morphometrics*. London: Academic.
- Rock NMS (1988) *Numerical Geology*. New York: Springer.

by Birks and Gordon (1985), Birks (1987, 1992, 1995, 1998), Birks and Seppä (2004), and Birks et al. (2010). Many of the numerical methods discussed here are also widely used in paleolimnology (Birks, 1998, 2010). The recent book edited by Birks et al. (2012) provides up-to-date reviews of the numerical and statistical techniques used in paleolimnology, many of which are directly relevant to the analysis of Quaternary palynological data.

Data Collection and Data Assessment

Numerical tools can help in the identification of fossil pollen types and hence in basic data collection. Statistical techniques are essential in the assessment of the statistical uncertainties associated with pollen counting, namely, error estimation.

Identification

Pollen identification depends on careful comparison of fossil material with modern reference material prepared and mounted in the same way as the fossil material. Pollen identification involves assigning fossil grains to preexisting categories of modern taxa. With experience, most grains in a temperate flora can be assigned by memory to modern categories. However, some numerically dominant and ecologically important pollen types may be so similar in overall morphology and appearance that the pollen analyst cannot consistently assign a particular fossil to a modern species. This problem is particularly acute with pollen from several important tree or shrub genera, such as *Pinus* (pine), *Picea* (spruce), and *Betula* (birch). Pollen from these taxa is often a substantial component in pollen assemblages from northern temperate and boreal regions. In such cases, numerical techniques can be invaluable in distinguishing the pollen to lower (species or subgenus) taxonomic levels, thereby enabling subsequent data analysis to capture more of the ecological signal encoded in the assemblage. Pollen-morphological characters are often of different data types (e.g., quantitative measurements, ordinal or multi-state qualitative (e.g., smooth, rough, and very rough), binary or dichotomous (e.g., presence or absence of granules)). If possible, all the available characters should be used in an attempt to achieve reliable identifications. Numerical techniques such as the deviant index for mixed data (Hansen and Cushing, 1973) and classification and regression trees (CART) (Lindbladh et al., 2002) have proved to be powerful means of identifying fossil pine and spruce grains to species level.

Pollen counting and identification are very time-consuming and efforts are continually being made to automate them (Flenley, 2003) using texture measures, artificial neural networks, Gabor transforms, digital moments, and other pattern description and recognition techniques. Major advances are currently being made at Massey University, New Zealand, in automated pollen identification and counting using Classifynder (Holt et al., 2011; <http://www.classifynder.com>). Within a few years, automated pollen analysis will be frequent in many palynological laboratories.

Error Estimation

A pollen count is a sample, hopefully an unbiased sample count, of the total amount of pollen preserved in the sediment sample of interest. As in all sampling, there are statistical uncertainties associated with any sample count. The larger the pollen count, the smaller the uncertainties become. As larger counts require more time to obtain, there is a trade-off between time and level of uncertainty. It is therefore important to estimate the uncertainty associated with all pollen counts (Bennett and Willis, 2001; Maher et al., 2012).

Methods for estimating counting errors and confidence intervals associated with relative pollen data are well developed and are based on the multinomial and negative multinomial distributions (Mosimann, 1965). Errors depend on the size of the count and on the proportion in the count sum of the pollen type of interest. The degree of uncertainty decreases with the size of the count. Methods for estimating errors and confidence intervals associated with pollen concentrations are also available (Maher et al., 2012). Estimating the total error requires careful propagation of errors associated with the various steps involved in deriving concentration values. The total error depends not only on estimating and combining these errors but also on the number of fossil grains counted. Estimating the errors for PAR is similar to estimating concentration errors but with the additional complexity of incorporating the uncertainties associated with estimating sediment accumulation rates based on radiocarbon dates and calibrating radiocarbon ages into calibrated ages (Maher et al., 2012). The total error associated with PAR may be 25–40% or more of the estimated values. It is therefore essential to derive and display reliable estimates of the total errors before attempting interpretation of observed changes in PAR (Bennett and Willis, 2001).

Data Summarization

This involves detecting the major patterns of variation within a single data-set (stratigraphical or geographical data) and the patterns of similarity and difference between two or more stratigraphical sequences or geographical data-sets.

Single Data-Sets

A useful first step in the numerical analysis of a multivariate palynological data-set is to summarize the data as a few groups or in a few dimensions. These are selected to fulfill predefined mathematical criteria, such that the groups maximize the between-group variance or the selected dimensions and axes capture as much of the total variance in the data with the constraint that all the axes are uncorrelated to each other. However, standard techniques for clustering or partitioning (e.g., hierarchical cluster analysis, *K*-means clustering, two-way indicator species analysis) and ordination or dimension reduction (e.g., PCA, correspondence analysis (CA), detrended CA (DCA), principal coordinates analysis, nonmetric multidimensional scaling (NMDS)) do not take account of the stratigraphical or geographical nature of the data. Hence, one may discard potentially important information when these methods are applied to palynological data, with the result that objects with similar pollen composition may be grouped together even though the objects are far apart stratigraphically or geographically. Numerical methods that take account of stratigraphical or geographical information include constrained clustering and partitioning (one-dimensional constraints for stratigraphical data; two-dimensional constraints for spatial data) and constrained ordination (redundancy analysis (RDA)=constrained PCA, canonical CA (CCA)=constrained CA, detrended CCA (DCCA)=constrained DCA). Detailed accounts of these methods are given in Gordon

(1999), ter Braak (1996), Legendre and Legendre (1998), and Legendre and Birks (2012a,b).

Stratigraphical sequences are most commonly summarized as a series of local or site pollen-assemblage zones. Numerical zonation is implemented by stratigraphically constrained clustering or partitioning using an agglomerative or divisive algorithm (Birks, 2012b; Birks and Gordon, 1985) followed by a comparison of the variance of the zones with the variance expected under the broken-stick model to provide a basis for deciding how many zones should be defined (Bennett, 1996). Of the many constrained clustering and partitioning techniques currently available, optimal partitioning using the sum-of-squares criterion and implemented by a dynamic programming algorithm (Birks, 2012b; Birks and Gordon, 1985) is consistently the most robust and hence most useful numerical zonation procedure.

Ordination of stratigraphical data, constrained by object depth or age, by RDA, CCA, or DCCA can be valuable in identifying the major patterns of variation or latent structure, the nature of change, and any trends in the sequence (Birks, 1987). For some research questions, it may be useful not to impose the stratigraphical constraint and to use PCA, CA, DCA, etc., simply to detect the major patterns of variation irrespective of the ordering of the objects. In other research problems, it may be useful to partial out, as covariables, the effects of 'nuisance' variables that are not of primary interest (e.g., Bradshaw et al., 2005; Odgaard, 1994) in partial PCA, CA, DCA, RDA, CCA, or DCCA. Methods based on an assumed linear response of the variables to the underlying latent gradients of variation implicit in PCA and its relatives (RDA, partial PCA, partial RDA) are appropriate with palynological data (stratigraphical or spatial) with a total compositional turnover of 2 or less standard deviations as estimated in DCA or DCCA (ter Braak, 1996). Methods based on an assumed unimodal response model implicit in CA and its relatives (DCA, CCA, DCCA, and their partial relatives) are most appropriate with data showing a high total compositional turnover (>2 standard deviations). Plotting the ordination results (object scores) in a stratigraphical context can provide a useful summary of the major patterns of variation and the nature of changes in the sequence and allows the detection of trends and abrupt changes within the sequence which may be obscured in zonation with its primary concern on partitioning the data sequence into 'homogeneous' units or zones. Careful examination of the taxon (variable) scores can also reveal which pollen taxa are most influential to the object scores for a given ordination axis, thereby providing an ecological interpretation of the observed patterns of variation (Birks, 2012b).

Since the pioneering uses of NMDS in Quaternary palynology (Birks and Gordon, 1985), NMDS is being increasingly used by Quaternary pollen analysts. NMDS produces a low-dimensional representation or map of the objects (pollen samples) (like PCA and CA) but where the rank orders of the similarities between all pairs of objects in the original full-dimensional data are represented monotonically in the low-dimensional ordination plot. There is no unique solution and no underlying response model in NMDS in contrast to PCA or CA. NMDS simply produces a low-dimensional representation or map of the objects. Interpretation in terms of the pollen taxa is often difficult and can involve plotting taxon values on the

resulting ordination plots. There is no guarantee that the NMDS axes are capturing the major patterns of variation or latent structure in the data, in contrast to PCA and CA. Overall, it has few advantages over PCA or CA as a tool in data summarization (see Legendre and Birks, 2012b; Prentice, 1980).

In the analysis of geographical palynological data, spatial constraints can be incorporated in clusterings or partitionings based on cluster analysis (Legendre and Birks, 2012a; Legendre and Legendre, 1998) and in ordinations (Legendre and Birks, 2012b; ter Braak, 1996). The question of whether to apply such constraints depends very much on the research problem of interest – do similar modern pollen assemblages all occur together geographically, or how can modern pollen assemblages be partitioned into coherent geographical areas of similar pollen composition? In the first case, unconstrained analysis would be appropriate; in the second case, constrained analysis would be required. The results of clusterings and ordinations should be displayed as maps of the group membership of objects or object scores on the first few ordination axes (Grimm, 1988). The appropriate number of ordination axes to retain can be assessed by comparison with the broken-stick model (Legendre and Birks, 2012b; Legendre and Legendre, 1998).

CART are useful tools in the analysis of sets of modern pollen assemblages. They describe the dependence of response variable values on the values of predictor variables. They are defined by a recursive binary partitioning of the data-set into subgroups that are successively more and more homogenous in the values of the response variable. At each partition, one predictor is used to define the binary split, and the split that maximizes the homogeneity and difference between the two subgroups is selected. Regression trees can be extended to consider multivariate responses (MRT) (Legendre and Birks, 2012a; Simpson and Birks, 2012). For palynological examples of their use, see Pelánková et al. (2008: CART) and Herzschuh and Birks (2010: MRT). In the first case using CART, modern pollen assemblages were used as predictors to evaluate how modern pollen assemblages reflect modern vegetation types, landscape types, and different climate variables. In the latter case, a multivariate regression tree (MRT) was used to identify the major modern climate thresholds for the composition of modern pollen assemblages. Other useful numerical tools in the analysis of modern or stratigraphical pollen data include (1) indicator species analysis (Legendre and Birks, 2012a) to detect indicator pollen taxa for particular groups of samples (e.g., Herzschuh and Birks, 2010), (2) multiresponse permutation procedures (MRPP) to assess statistical differences between groups of modern pollen samples collected from the same vegetation type and from different vegetation types (e.g., Herzschuh and Birks, 2010), and (3) analysis of similarity (ANOSIM) to assess the statistical significance of overall differences in modern spectra and pairwise ANOSIMs to evaluate differences in modern spectra from different vegetation types. By looking at the overall percentage contribution that each pollen type makes to the average dissimilarity between pairs of groups, one can determine the pollen types most influential in discriminating between the two groups of samples (SIMPER, similarity percentage tests). See Bhattacharya et al. (2011) for a use of ANOSIM and SIMPER with modern pollen data and Birks (2012c) for introductions to these and related techniques.

Two or More Stratigraphical Sequences

When two or more paleoecological variables (e.g., pollen, diatoms) have been studied in the same stratigraphical sequence (multiproxy studies – Birks and Birks, 2006), numerical zonations based on each set of variables and a comparison of the resulting partitions can help to identify common and unique changes in different variables, for example, pollen reflecting catchment vegetation and diatoms reflecting within-lake conditions, or pollen reflecting regional-scale vegetation and plant macrofossils reflecting local vegetation (Birks and Gordon, 1985). Separate ordinations of the different data-sets help to summarize the major patterns within each data-set, and these patterns can be compared by, for example, oscillation logs (Birks, 1987). In cases where one data-set can be regarded as representing ‘response’ variables (e.g., pollen) and another data-set can be regarded as reflecting potential ‘predictor’ or explanatory variables (e.g., stable isotope ratios; independent climate estimates), constrained ordinations (RDA, CCA, DCCA) and associated Monte Carlo permutation tests (ter Braak, 1996) can be used to assess the statistical relationship between the two data-sets. This approach is discussed further under Data Interpretation.

Comparison of two or more pollen-stratigraphical sequences can be undertaken by ordinating the data from all sequences together by unconstrained PCA, CA, or DCA or by constrained RDA, CCA, or DCCA using site identity coded as nominal 1/0 (‘dummy’ variable) constraints. Sequence-slotting with or without external constraints (e.g., tephra layers) can provide a direct comparison of sequences and can, if an absolute chronology is available, assist in the construction of time–space correlation diagrams (Birks and Gordon, 1985; Thompson et al., 2012).

Differences between sequences are more difficult to characterize and display. Difference diagrams (Birks, 2012b; Birks and Gordon, 1985) where the differences in pollen composition at the same time between two sites, each with a reliable chronology, are calculated and displayed, can be useful tools for summarizing differences between sites at, for example, different altitudes or on different soil types.

Two or More Geographical Data-Sets

Patterns in two or more spatial data-sets (e.g., modern pollen assemblages and pollen assemblages prior to European settlement in North America) can be compared by separate ordinations or clusterings of the two data-sets and the mapping of the sample scores or group memberships for different times. Difference maps, where the values of a pollen type for time $t - 1$ are subtracted from the values of the same taxon for time t , can be constructed to highlight geographical patterns of increase or decrease in particular pollen taxa (Birks and Gordon, 1985).

Through a careful and critical use of appropriate numerical procedures, it is possible to derive a consistent and repeatable description and summary of the major patterns of variation within and between stratigraphical and geographical palynological data-sets. Numerical methods are particularly useful for these purposes because they highlight and identify the major ‘signals’ within the data-set(s) and relegate the inherent ‘noise’ within the data to a subordinate role.

Data Analysis

Data analysis is used here to include specialized techniques that estimate particular numerical characteristics from pollen-stratigraphical data. Examples include palynological richness, modeling of population changes and growth rates, rates of change, periodicities and power spectra, and inferred past environment.

Palynological Richness

Rarefaction analysis (Birks and Line, 1992) estimates the palynological richness within and between sequences. It estimates how many taxa would have been found if all the pollen counts had been the same size. The actual minimum count among the sequence(s) of interest is usually used as the base value. The interpretation of palynological richness as a record of past biodiversity is complex and currently unresolved (Odgaard, 1994, 1999; van der Knaap, 2009). In an integrated palynological, socioeconomical, and land use historical study, Hanley et al. (2008) showed that recent changes in palynological richness in several sites in the Scottish uplands were best predicted, in a statistical sense, by livestock prices, a proxy for grazing pressure.

Population Analysis

Stratigraphical PAR can be viewed as temporal records of past plant populations within the pollen source area of the study site. Sequence-splitting (Birks, 2012b; Birks and Gordon, 1985) divides the PAR of individual taxa into units of presence or absence, and when the taxon is present, into units of uniform mean and variance. It focuses on the individualistic behavior of taxa and on similarities and differences in the patterns of change of taxa within one or more sequences (Birks and Gordon, 1985). Sequence-splitting is one type of classification tree where depth is the sole external predictor (Birks, 2012b). PAR of individual taxa can also be modeled using logistic or exponential population growth models and population doubling times can be estimated for individual taxa as a means of comparing population rates of change between taxa, sites, and geographical regions (Birks and Gordon, 1985).

These types of models are *a posteriori* modeling (*sensu* Flenley, 2003) where pollen-analytical data are used and mathematical population models are then fitted to aid data analysis. The models fitted can be relatively complex, such as nonlinear population models (e.g., Jeffers et al., 2011, 2012) that allow testing of alternative hypotheses about potential drivers of change. Their use thus lies between data analysis and data interpretation.

In the alternative *a priori* approach to modeling (Flenley, 2003), a landscape vegetation dynamics or forest succession model is used (e.g., Heiri et al., 2006). These models require parameters for each taxon (e.g., longevity, seed productivity, growth rate, etc.) and independently derived climate estimates to 'drive' the model. The model attempts to simulate what may have happened in terms of vegetation composition and taxon abundance at a given spatial scale, and the simulation results

can be compared with actual pollen-analytical data. These models are used to test ecological hypotheses about, for example, drivers of long-term treeline or forest limit changes (Heiri et al., 2006).

Rate-of-Change Analysis

Rate-of-change analysis (Grimm, 1988; Grimm and Jacobson, 1992) estimates the amount of palynological compositional change per unit time in stratigraphical data. It is estimated by calculating the multivariate dissimilarity (e.g., chord distance) between stratigraphically adjacent samples and by dividing the dissimilarity by the estimated age interval between the sample pairs. An alternative approach is to interpolate the data to constant time intervals, calculate the dissimilarity, and divide by a constant time interval. The data can, if required, be smoothed prior to interpolation (Birks, 1998, 2012b). Rate-of-change analysis is critically dependent on a reliable chronology for the sequence. Because radiocarbon years do not always equal calendar years (Blaauw and Heegaard, 2012) (see *Calibration of the ¹⁴C Record*), a carefully calibrated timescale or an independent absolute chronology (e.g., from laminated sediments) are essential for reliable rate-of-change estimation (Birks, 1998).

Time Series Analysis

Pollen-stratigraphical data represent time series of the changing frequencies or accumulation rates of different pollen types through time. There are two main approaches to time series analysis. There is the time domain approach that is based on the concept of autocorrelation, namely, the correlation between samples in the same sequence that are k time intervals apart. The autocorrelation coefficient is a measure of similarity between samples separated by different time intervals, and it can be plotted as a correlogram to assess the behavior (e.g., periodicities) of the pollen type of interest over time. Time series of two different variables can be compared by the cross correlation coefficient to detect patterns of temporal variation and relationships between variables (e.g., Tinner et al., 1999).

The second approach involves the frequency domain and focuses on bands of frequency or wavelength over which the variance of a time series is concentrated. It estimates the proportion of the variance that can be attributed to each of a continuous range of frequencies. The power spectrum of a time series can help detect periodicities within the data and the main tools are spectral density functions, cross-spectral functions, and the periodogram. Recent palynological examples include Willis et al. (1999) and paleolimnological examples are reviewed by Dutilleul et al. (2012) and Cumming et al. (2012).

Conventional time series analysis makes stringent assumptions of the data, namely, that the intersample intervals are constant and that the data are stationary, and thus, there are no trends in mean or variance in the time series (Dutilleul et al., 2012). In the absence of equally spaced samples, the usual procedure is to interpolate samples to equal time intervals. This is equivalent to a low-pass filter and may result in an underestimation of the high-frequency component in the spectrum. Techniques have recently been developed for spectral

and cross-spectral analysis for unevenly spaced time series, so-called temporal series (Dutilleul et al., 2012), but they do not appear to have been applied to pollen-stratigraphical data, although they have proved useful in analyzing paleolimnological temporal series (Cumming et al., 2012; Dutilleul et al., 2012).

A limitation of all power-spectral analyses is that they provide an integrated estimate of variance for the entire time series. This can be overcome by wavelet power-spectral analysis that identifies the dominant frequencies in different variables and displays how these frequencies have varied through the time series. Brown et al. (2005) show the value of wavelet analysis in the analysis of palynological data and Birks (2012d) outlines paleolimnological and paleoclimatological uses of wavelet analysis for unevenly sampled time series.

Pollen-Based Climate Reconstructions

Since the pioneering work by H.S. Cole and T. Webb III in the quantitative reconstruction of past climate from pollen-stratigraphical data, several approaches have been developed to derive so-called modern 'transfer' or calibration functions that model the relationship between modern pollen assemblages and modern climate. These calibration functions are then used to transform fossil pollen assemblages into estimates of past climate. The various approaches are reviewed by Birks (1995) and Birks et al. (2010) and the problems and pitfalls in pollen-based climate reconstructions in Europe are discussed by Birks and Seppä (2004).

The approaches to quantitative reconstruction of past climate from palynological data fall into three main types (Birks et al., 2010): the indicator species approach, the assemblage approach (e.g., modern analog-based techniques including response surface analysis), and the calibration function or multivariate indicator species approach involving some form of linear or nonlinear regression and calibration (e.g., segmented inverse linear regression, principal components regression, two-way weighted averaging, weighted averaging partial least squares regression). Because each numerical procedure has its own biases, careful model selection using split-sampling cross validation is essential (Birks, 1995; Juggins and Birks, 2012). When several models appear equally appropriate, a consensus reconstruction can be derived by fitting a robust smoother (e.g., a LOWESS (locally weighted least squares) smoother) through the reconstructed values derived from different models (Juggins and Birks, 2012).

A major development is the development of a robust statistical procedure to test if a quantitative reconstruction inferred from fossil pollen assemblages explains more of the variation in the fossil assemblages than reconstructions based on randomly derived climate variables (Telford and Birks, 2011). Reconstructions that fail this basic test of statistical significance thus have limited credibility and should only be interpreted with extreme caution.

A hidden problem in the use of modern pollen-climate calibration functions is spatial autocorrelation, an inherent property of all spatially distributed data. The estimation of the predictive power of any calibration function assumes that the test sets used in cross validation are independent of the sites used in the calibration function (Telford and Birks, 2005).

Cross validation in the presence of spatial autocorrelation violates this assumption. Telford and Birks (2009) have developed methods for detecting and describing spatial autocorrelation, quantifying its effects, testing the significance of a calibration function in autocorrelated environments, and implementing cross validation that is robust to autocorrelation. Some pollen-climate calibration functions have no predictive power because of the strong impact of spatial autocorrelation (Telford and Birks, 2009).

Given these recent statistical developments and an increasing concern about the ecological and paleoecological assumptions in the use of calibration functions (Birks et al., 2010; Juggins and Birks, 2012), the use of such calibration functions for quantitative environmental reconstructions requires care and critical awareness and understanding of the assumptions, strengths and weaknesses, and numerical problems in calibration functions (see Juggins and Birks, 2012).

Similar approaches using calibration functions or modern analog techniques (MATs) and modern pollen and satellite-based images can be used to reconstruct conifer or deciduous forest cover, total vegetation cover, or leaf area index from fossil pollen assemblages over broad areas (e.g., Tarasov et al., 2007; Williams, 2002; Williams et al., 2011).

Data Interpretation

Pollen-stratigraphical data are a complex reflection of the vegetational composition around the study site. Interpretation of these data can be in terms of past vegetation or in terms of possible factors that have influenced changes in vegetation and hence in pollen stratigraphy (Bennett and Willis, 2001).

Vegetation Reconstruction

The reconstruction of past vegetation can be attempted using modern pollen representation factors and models of pollen deposition (e.g., Soepboer et al., 2010), finding pollen of indicator species characteristic of particular vegetation types today (e.g., Herzschuh and Birks, 2010), or comparing quantitatively fossil pollen assemblages with modern pollen assemblages from known vegetation types today, the so-called MAT. The basic idea of the MAT is simple – compare the fossil assemblage with all modern pollen assemblages using an appropriate dissimilarity measure (e.g., chord distance), find the modern assemblage(s) most similar to the fossil assemblage, assign the modern vegetation type(s) of the closest analog(s) to the fossil assemblage, and repeat for all fossil assemblages (Simpson, 2007, 2012). Problems arise in defining a critical threshold to decide if the modern and fossil assemblages are sufficiently similar to indicate that they could be derived from the same vegetation type. Fossil assemblages may, for a variety of reasons, have no close modern analog (Jackson and Williams, 2004). In such cases, the MAT will fail as a tool for interpreting fossil assemblages in terms of past vegetation. Attempts have been made to put the MAT on a more rigorous numerical basis (Simpson, 2007, 2012) using receiver operating characteristic analysis, logistic regression, and Monte Carlo simulation. The robustness of these approaches remains to be

assessed with different data-sets and in different geographical areas (Simpson, 2007, 2012).

An alternative but less formal approach to analog analysis is to ordinate or cluster together the modern and stratigraphical data (Birks and Gordon, 1985) and to examine what modern and fossil samples are positioned near each other or are clustered together. Alternatively, the modern pollen data can be ordinated (e.g., using PCA, CA, or DCA) and the modern vegetation type displayed on the ordination plots of the first few ordination axes. The fossil samples can be positioned as ‘passive’ or ‘supplementary’ objects into the plane formed by the modern ordination axes. Alternatively, fossil samples can be the basis of the primary ordination and modern samples added as ‘supplementary’ objects. A more formal approach involves multiple discriminant analysis (canonical variate analysis). It can be used to display the maximum between-group variance relative to the within-group variance of modern samples from known vegetation types. Fossil samples from unknown vegetation types are then assigned (‘identified’) to one of these modern vegetation types (Birks and Gordon, 1985).

Causative Factors

Many factors can influence the composition of vegetation and hence the composition of pollen assemblages – soil changes, climatic changes, human impact, pathogens, species interactions, disturbance regimes, lags in migration and population expansion, and complex interactions between these factors (Bennett and Willis, 2001). It is a major challenge to test competing hypotheses about underlying causative factors. Statistical techniques such as constrained ordination (e.g., RDA=reduced-rank multivariate regression, CCA, DCCA) and associated Monte Carlo permutation tests can be used to assess the statistical relationships between pollen data (‘responses’) and external forcing factors (‘predictors’), for example, to assess the relative impacts of climatic change and volcanic tephra deposition on pollen assemblages and hence on past vegetation (Lotter and Birks, 1993) or the role of fire on vegetation dynamics (Feurdean et al., 2009). Although statistical techniques exist with permutation tests that can take specific account of the time-ordered nature of pollen-stratigraphical data, the major problem is deriving, for statistical analysis, data that reflect potential forcing factors (Lotter and

Birks, 2003). This remains a major challenge in the interpretation of all Quaternary pollen-analytical data – see Lotter and Birks (1993, 2003), Mitchell (2004), and Jeffers et al. (2011, 2012) for examples of different approaches to hypothesis testing in Quaternary pollen analysis.

Software and Computing

Some numerical methods require the use of software to implement them, and new and important numerical methods relevant to Quaternary palynological data analysis are constantly being developed by applied statisticians and quantitative paleoecologists. The rate of development of new methods greatly exceeds the production of user-friendly dedicated software such as TILIA, C2, or PSIMPOLL. Virtually all the new methods are available as packages and scripts for public use free of charge within the R software development package and environment (www.r-project.org) (Table 3). Given the wide array of freely available R packages (e.g., vegan, rioja, analogue, paleoSig) and scripts relevant to pollen analysts and the ever-increasing power of personal computers, there is no excuse for palynologists to use outdated, suboptimal numerical techniques in the analysis of their hard-earned data. There is inevitably a steep learning curve in the initial stages of using R, just as there is in any other programming environment (e.g., GenStat, MATLAB), but the final rewards are well worth the initial effort. Using R and new specially developed techniques is a challenge for the Quaternary palynological community now and in the immediate future.

There are also many books available now that provide excellent introductions to the use of R in the numerical and statistical analysis of data – see Table 4.

Future Developments

Considerable advances have been made in the numerical analysis of Quaternary palynological data since the pioneering attempts in the late 1960s and early 1970s, particularly in the descriptive and narrative phases of pollen-analytical studies. In the descriptive phase, basic patterns are detected, assessed, described, and classified, and the relevant numerical

Table 3 Software for paleoenvironmental analysis

Package	Web address	Use
TILIA	http://www.neotomadb.org/data/category/tilia	Plotting of palynological data
PSIMPOLL	http://www.chrono.qub.ac.uk/psimpoll/psimpoll.html/	Plotting and analysis of palynological data
rioja	www.staff.ncl.ac.uk/staff/stephen.juggins	Quaternary science data
analogue	http://analogue.r-forge.r-project.org	Analog and weighted averaging
vegan	http://cran.r-project.org	Community ecology analysis
	http://vegan.r-forge.r-project.org	
paleoMAS	http://cran.r-project.org	Paleoecological analysis
fossil	http://cran.r-project.org/web/packages/fossil/index.html	Paleoecological and paleogeographical analysis
simba	www.r-project.org	Similarity coefficients
palaeoSig	http://cran.r-project.org/web/packages/palaeoSig/index.html	Significance tests of reconstructions

Birks HJB, Lotter AF, Juggins S, and Smol JP (eds.) (2012) *Tracking Environmental Change Using Lake Sediments, vol. 5: Data Handling and Numerical Techniques*. Dordrecht: Springer.

Table 4 Books on the use of R for numerical and statistical analyses*General texts*

- Crawley MJ (2007) *The R Book*. Chichester: Wiley.
- Dalgaard P (2008) *Introductory Statistics with R*. New York: Springer.
- Fox J and Weisberg S (2011) *An R Companion to Applied Regression*. Sage, London.
- Logan M (2010) *Biostatistical Design and Analysis Using R*. Chichester: Wiley-Blackwell.
- Torgo L (2011) *Data Mining and R. Learning with Case Studies*. Boca Raton, FL: CRC Press.
- Wright DB and London K (2009) *Modern Regression Techniques Using R. A Practical Guide for Students and Researchers*. London: Sage.
- Zuur AF, Ieno EN, and Meesters EHWG (2009) *A Beginner's Guide to R*. New York: Springer.

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- Bivand RS, Pebesma EJ, and Gómez-Rubio V (2008) *Applied Spatial Data with R*. New York: Springer.
- Borcard D, Gillet F, and Legendre P (2011) *Numerical Ecology with R*. New York: Springer.
- Cowpertwait PSP and Metcalfe AV (2009) *Introductory Time Series with R*. New York: Springer.
- Everitt BS and Hothorn T (2011) *An Introduction to Applied Multivariate Analysis with R*. New York: Springer.
- Wehrens R (2011) *Chemometrics with R*. New York: Springer.
- Williams G (2011) *Data Mining with Rattle and R*. New York: Springer.

Birks HJB, Lotter AF, Juggins S, and Smol JP (eds.) (2012) *Tracking Environmental Change Using Lake Sediments, vol. 5: Data Handling and Numerical Techniques*. Dordrecht: Springer.

techniques are all part of this phase. The various approaches to data analysis discussed above contribute to the narrative phase where plausible, inductively based explanations, generalizations, and reconstructions are derived from the observed patterns. Hypothesis-testing techniques such as RDA, CCA, DCCA, and associated permutation tests are one approach to the analytical phase in science where falsifiable or testable hypotheses are proposed, evaluated, tested, and rejected.

More work is needed to improve and to quantify comparisons and interpretations of stratigraphical data and to test competing hypotheses about possible causative factors in influencing past vegetation and hence fossil pollen assemblages at different spatial and temporal scales. Multidisciplinary paleoecological investigations are resulting in major advances because a multiproxy approach can help to test competing hypotheses (Birks and Birks, 2006; Bradshaw et al., 2005; Lotter and Birks, 2003). However, the rigorous numerical analysis of multiproxy data is still in its infancy.

A potentially important new approach to the interpretation of pollen-stratigraphical data is to use life history and other ecological traits and to link these traits to past environmental conditions through the fossil pollen record (Lacourse, 2009). Special numerical techniques that can simultaneously link paleoenvironmental data (*R*) and plant species traits (*Q*) by means of fossil pollen data (*L*), (RLQ analysis: Dolédec et al., 1996) are required for this type of analysis. Fourth-corner analysis (Dray and Legendre, 2008) is also useful for quantifying and testing the relationships between *R* and *L*. See Lacourse

(2009) for an example of these analyses to demonstrate that climate is the ultimate control on Holocene forest composition and species abundance but long-term community assembly is also constrained through interspecific differences in plant traits.

The numerical analysis of huge data-sets now available from several publicly available databases (Brewer et al., 2012) represents a major challenge because such data-sets have important spatial and stratigraphical properties (Mitchell, 2011) (see [Databases and Their Application](#)). New numerical data-mining techniques that can take account of the geographical and temporal properties of the data are required to help analyze the data (Hastie et al., 2011). Such techniques can be used to answer questions concerning common and unique trends in pollen data at a range of spatial scales, the importance of spatial and temporal autocorrelation at a range of spatial and temporal scales, and the role of causative factors at these different scales. Statistical-learning techniques such as CART, boosted trees, and random forests (Simpson and Birks, 2012) are potentially powerful techniques for analyzing huge pollen-stratigraphical data-sets. A major problem inherent in all databases is data quality (Brewer et al., 2012). The recognition of this problem is leading to new initiatives in data screening (e.g., Gonzales et al., 2009; Whitmore et al., 2005) and in assessing and reducing temporal uncertainties (Blois et al., 2011).

It is now 40 years since the pioneering work on the numerical analysis of pollen-stratigraphical data. The next 40 promises to be equally exciting, rewarding, and challenging.

See also: Pollen Analysis, Principles. **Pollen Methods and Studies:** Databases and Their Application; Reconstructing Past Biodiversity Development; The Biome Approach to Reconstructing Past Vegetation; Use of Pollen as Climate Proxies. **Radiocarbon Dating:** Calibration of the ¹⁴C Record.

References

- Bennett KD (1996) Determination of the number of zones in a biostratigraphical sequence. *New Phytologist* 132: 155–170.
- Bennett KD and Willis KJ (2001) Pollen. In: Smol JP, Birks HJB, and Last WM (eds.) *Tracking Environmental Change using Lake Sediments. Terrestrial, Algal, and Siliceous Indicators*, vol. 3, pp. 5–32. Dordrecht: Kluwer.
- Bhattacharya T, Beach T, and Wahl D (2011) An analysis of modern pollen rain from the Maya lowlands of northern Belize. *Review of Palaeobotany and Palynology* 164: 109–120.
- Birks HJB (1987) Multivariate analysis of stratigraphic data in geology: A review. *Chemometrics and Intelligent Laboratory Systems* 2: 109–126.
- Birks HJB (1992) Some reflections on the application of numerical methods in Quaternary palaeoecology. *Publications of the Karelian Institute, University of Joensuu* 102: 7–20.
- Birks HJB (1995) Quantitative palaeoenvironmental reconstructions. In: Maddy D and Brew JS (eds.) *Statistical Modeling of Quaternary Science Data*, pp. 161–254. Cambridge: Quaternary Research Association. Technical Guide 5.
- Birks HJB (1998) Numerical tools in palaeolimnology – Progress, potentialities, and problems. *Journal of Paleolimnology* 20: 307–332.
- Birks HJB (2010) Numerical methods for the analysis of diatom assemblage data. In: Smol JP and Stoermer EF (eds.) *The Diatoms: Applications for the Environmental and Earth Sciences*, 2nd edn., pp. 23–54. Cambridge: Cambridge University Press.
- Birks HJB (2012a) Conclusions and future challenges. In: Birks HJB, Lotter AF, Juggins S, and Smol JP (eds.) *Tracking Environmental Change Using Lake*

- Sediments. Data Handling and Numerical Techniques*, vol. 5, pp. 653–674. Dordrecht: Springer.
- Birks HJB (2012b) Analysis of stratigraphical data. In: Birks HJB, Lotter AF, Juggins S, and Smol JP (eds.) *Tracking Environmental Change Using Lake Sediments. Data Handling and Numerical Techniques*, vol. 5, pp. 355–378. Dordrecht: Springer.
- Birks HJB (2012c) Overview of numerical methods in palaeolimnology. In: Birks HJB, Lotter AF, Juggins S, and Smol JP (eds.) *Tracking Environmental Change Using Lake Sediments. Data Handling and Numerical Techniques*, vol. 5, pp. 19–92. Dordrecht: Springer.
- Birks HJB (2012d) Introduction and overview of Part III. In: Birks HJB, Lotter AF, Juggins S, and Smol JP (eds.) *Tracking Environmental Change Using Lake Sediments. Data Handling and Numerical Techniques*, vol. 5, pp. 331–354. Dordrecht: Springer.
- Birks HH and Birks HJB (2006) Multi-proxy studies in palaeolimnology. *Vegetation History and Archaeobotany* 15: 235–251.
- Birks HJB and Gordon AD (1985) *Numerical Methods in Quaternary Pollen Analysis*. London: Academic Press.
- Birks HJB, Heiri O, Seppä H, and Björne AE (2010) Strengths and weaknesses of quantitative climate reconstructions based on late-quaternary biological proxies. *The Open Ecology Journal* 3: 68–110. <http://dx.doi.org/10.2174/1874213001003020068>.
- Birks HJB and Line JM (1992) The use of rarefaction analysis for estimating palynological richness from Quaternary pollen-analytical data. *The Holocene* 2: 1–10.
- Birks HJB, Lotter AF, Juggins S, and Smol JP (eds.) (2012) *Tracking Environmental Change Using Lake Sediments. Data Handling and Numerical Techniques*, vol. 5. Dordrecht: Springer.
- Birks HJB and Seppä H (2004) Pollen-based reconstructions of late-Quaternary climate in Europe – Progress, problems, and pitfalls. *Acta Palaeobotanica* 44: 317–334.
- Blaauw M and Heegaard E (2012) Estimation of age-depth relationships. In: Birks HJB, Lotter AF, Juggins S, and Smol JP (eds.) *Tracking Environmental Change Using Lake Sediments. Data Handling and Numerical Techniques*, vol. 5, pp. 379–414. Dordrecht: Springer.
- Blois JL, Williams JW, Grimm EC, Jackson ST, and Graham RW (2011) A methodological framework for assessing and reducing temporal uncertainty in paleovegetation mapping from late-Quaternary pollen records. *Quaternary Science Reviews* 30: 1926–1939.
- Bradshaw EG, Rasmussen P, and Odgaard BV (2005) Mid- to late-Holocene land-use change and lake development at Dallund Sø, Denmark: Synthesis of multiproxy data, linking land and lake. *The Holocene* 15: 1152–1162.
- Brewer S, Jackson ST, and Williams JW (2012) Paleoeoinformatics: Applying geohistorical data to ecological questions. *Trends in Ecology and Evolution* 27: 104–112.
- Brown KJ, Clark JS, Grimm EC, et al. (2005) Fire cycles in North American interior grasslands and their relation to prairie drought. *Proceedings of the National Academy of Sciences of the United States of America* 102: 8865–8870.
- Cumming BF, Laird KR, Fritz SC, and Verschuren D (2012) Tracking Holocene climatic change with aquatic biota from lake sediments: Case studies of commonly used numerical techniques. In: Birks HJB, Lotter AF, Juggins S, and Smol JP (eds.) *Tracking Environmental Change Using Lake Sediments. Data Handling and Numerical Techniques*, vol. 5, pp. 615–642. Dordrecht: Springer.
- Dolédéc S, Chessel D, ter Braak CJF, and Champely S (1996) Matching species traits to environmental variables: A new three-table ordination method. *Environmental and Ecological Statistics* 3: 143–166.
- Dray S and Legendre P (2008) Testing the species traits-environment relationships: The fourth-corner problem revisited. *Ecology* 89: 3400–3412.
- Duttilleul P, Cumming BF, and Lontoc-Roy M (2012) Autocorrelogram and periodogram analyses of palaeolimnological temporal-series from lakes in central and western North America to assess shifts in drought conditions. In: Birks HJB, Lotter AF, Juggins S, and Smol JP (eds.) *Tracking Environmental Change Using Lake Sediments. Data Handling and Numerical Techniques*, vol. 5, pp. 523–548. Dordrecht: Springer.
- Feurdean A, Willis KJ, and Astalos C (2009) Legacy of the past land-use changes and management on the 'natural' upland forest composition in the Apuseni National Park, Romania. *The Holocene* 19: 967–981.
- Flenley J (2003) Some prospects for lake sediment analysis in the 21st century. *Quaternary International* 105: 77–80.
- Gonzales LM, Grimm EC, Williams JW, and Nordheim EV (2009) A modern plant-climate research data-set for modeling eastern North American plant taxa. *Grana* 48: 1–18.
- Gordon AD (1999) *Classification*, 2nd edn. Boca Raton, FL: Chapman & Hall/CRC.
- Grimm EC (1988) Data analysis and display. In: Huntley B and Webb T III (eds.) *Vegetation History*, pp. 43–76. Dordrecht: Kluwer.
- Grimm EC and Jacobson GL Jr. (1992) Fossil-pollen evidence for abrupt climate change during the past 18000 years in eastern North America. *Climate Dynamics* 6: 179–184.
- Hanley N, Davies A, Angelopoulos K, et al. (2008) Economic determinants of biodiversity change over a 400-year period in the Scottish uplands. *Journal of Applied Ecology* 45: 1557–1565.
- Hansen BS and Cushing EJ (1973) Identification of pine pollen of late Quaternary age from Chuska Mountains, New Mexico. *Geological Society of America Bulletin* 84: 1181–1199.
- Hastie TJ, Tibshirani RJ, and Friedman J (2011) *The Elements of Statistical Learning*, 2nd edn. New York: Springer.
- Heiri C, Bugmann H, Tinner W, Heiri O, and Lischke H (2006) A model-based reconstruction of Holocene treeline dynamics in the Central Swiss Alps. *Journal of Ecology* 94: 206–216.
- Herzschuh U and Birks HJB (2010) Evaluating the indicator value of Tibetan pollen taxa for modern vegetation and climate. *Review of Palaeobotany and Palynology* 160: 197–208.
- Holt K, Allen G, Hodgson R, Marsland S, and Flenley J (2011) Progress towards an automated trainable pollen location and classifier system for use in the palynology laboratory. *Review of Palaeobotany and Palynology* 167: 173–175.
- Jackson ST and Williams JW (2004) Modern analogs in Quaternary palaeoecology: Here today, gone yesterday, gone tomorrow? *Annual Review of Earth and Planetary Sciences* 32: 495–537.
- Jeffers ES, Bonsall MB, Watson JE, and Willis KJ (2012) Climate change impacts on ecosystem functioning: Evidence from an *Empetrum* heathland. *New Phytologist* 193: 150–164.
- Jeffers ES, Bonsall MB, and Willis KJ (2011) Stability in ecosystem functioning across a climatic threshold and contrasting forest regimes. *PLoS One* 6: e16134. <http://dx.doi.org/10.1371/journal.pone.0016134>.
- Juggins S and Birks HJB (2012) Quantitative environmental reconstructions from biological data. In: Birks HJB, Lotter AF, Juggins S, and Smol JP (eds.) *Tracking Environmental Change Using Lake Sediments. Data Handling and Numerical Techniques*, vol. 5, pp. 431–498. Dordrecht: Springer.
- Lacourse T (2009) Environmental change controls post-glacial forest dynamics through interspecific differences in life-history traits. *Ecology* 90: 2149–2160.
- Legendre P and Birks HJB (2012a) Clustering and partitioning. In: Birks HJB, Lotter AF, Juggins S, and Smol JP (eds.) *Tracking Environmental Change Using Lake Sediments. Data Handling and Numerical Techniques*, vol. 5, pp. 167–200. Dordrecht: Springer.
- Legendre P and Birks HJB (2012b) From classical to canonical ordination. In: Birks HJB, Lotter AF, Juggins S, and Smol JP (eds.) *Tracking Environmental Change Using Lake Sediments. Data Handling and Numerical Techniques*, vol. 5, pp. 201–248. Dordrecht: Springer.
- Legendre P and Legendre L (1998) *Numerical Ecology*, 2nd English edn. Amsterdam: Elsevier.
- Lindbladh M, O'Connor R, and Jacobson GL (2002) Morphometric analysis of pollen grains for palaeoecological studies: Classification of *Picea* from eastern North America. *American Journal of Botany* 89: 1459–1467.
- Lotter AF and Birks HJB (1993) The impact of the Laacher See Tephra on terrestrial and aquatic ecosystems in the Black Forest, southern Germany. *Journal of Quaternary Science* 8: 263–276.
- Lotter AF and Birks HJB (2003) The Holocene palaeolimnology of Sägistalsee and its environmental history – A synthesis. *Journal of Paleolimnology* 30: 333–342.
- Maher LJ, Heiri O, and Lotter AF (2012) Assessment of uncertainties associated with palaeolimnological laboratory methods and microfossil analysis. In: Birks HJB, Lotter AF, Juggins S, and Smol JP (eds.) *Tracking Environmental Change Using Lake Sediments. Data Handling and Numerical Techniques*, vol. 5, pp. 143–166. Dordrecht: Springer.
- Mitchell FJG (2004) How open were European primeval forests? Hypothesis testing using palaeoecological data. *Journal of Ecology* 93: 168–177.
- Mitchell FJG (2011) Exploring vegetation in the fourth dimension. *Trends in Ecology and Evolution* 26: 45–52.
- Mosimann JE (1965) Statistical methods for the pollen analyst: Multinomial and negative multinomial techniques. In: Kummel B and Raup D (eds.) *Handbook of Paleontological Techniques*, pp. 636–673. San Francisco, CA: W.H. Freeman & Co.
- Odgaard BV (1994) Holocene vegetation history of northern west Jutland, Denmark. *Opera Botanica* 123: 1–171.
- Odgaard BV (1999) Fossil pollen as a record of past biodiversity. *Journal of Biogeography* 26: 7–17.

- Pelánková B, Kuneš P, Chytrý M, Jankovská V, Ermakov N, and Svobodová-Svitavská H (2008) The relationships of modern pollen spectra to vegetation and climate along a steppe-forest-tundra transition in southern Siberia, explored by decision trees. *The Holocene* 18: 1259–1271.
- Prentice IC (1980) Multidimensional-scaling as a research tool in Quaternary palynology – A review of theory and methods. *Review of Palaeobotany and Palynology* 31: 71–104.
- Reyment RA and Savazzi E (1999) *Aspects of Multivariate Statistical Analyses in Geology*. Amsterdam: Elsevier.
- Simpson GL (2007) Analogue methods in palaeoecology: Using the analogue package. *Journal of Statistical Software* 22: 1–29.
- Simpson GL (2012) Analogue methods in palaeolimnology. In: Birks HJB, Lotter AF, Juggins S, and Smol JP (eds.) *Tracking Environmental Change Using Lake Sediments. Data Handling and Numerical Techniques*, vol. 5, pp. 495–522. Dordrecht: Springer.
- Simpson GL and Birks HJB (2012) Statistical learning in palaeolimnology. In: Birks HJB, Lotter AF, Juggins S, and Smol JP (eds.) *Tracking Environmental Change Using Lake Sediments. Data Handling and Numerical Techniques*, vol. 5, pp. 249–327. Dordrecht: Springer.
- Soepboer W, Sugita S, and Lotter AF (2010) Regional vegetation-cover changes on the Swiss Plateau during the past two millennia: A pollen-based reconstruction using the REVEALS model. *Quaternary Science Reviews* 29: 472–483.
- Tarasov P, Williams JW, Andreev A, et al. (2007) Satellite- and pollen-based quantitative woody cover reconstructions for northern Asia: Verification and application to late-Quaternary pollen data. *Earth and Planetary Science Letters* 264: 284–298.
- Telford RJ and Birks HJB (2005) The secret assumption of transfer functions: Problems with spatial autocorrelation in evaluating model performance. *Quaternary Science Reviews* 24: 2173–2179.
- Telford RJ and Birks HJB (2009) Design and evaluation of transfer functions in spatially structured environments. *Quaternary Science Reviews* 28: 1309–1316.
- Telford RJ and Birks HJB (2011) A novel method for assessing the statistical significance of quantitative reconstructions inferred from biotic assemblages. *Quaternary Science Reviews* 30: 1272–1278.
- ter Braak CJF (1996) *Unimodal Models to Relate Species to Environment*. Wageningen: DLO-Agricultural Mathematics Group.
- Thompson R, Clark RM, and Boulton GS (2012) Core correlation. In: Birks HJB, Lotter AF, Juggins S, and Smol JP (eds.) *Tracking Environmental Change Using Lake Sediments. Data Handling and Numerical Techniques*, vol. 5, pp. 415–430. Dordrecht: Springer.
- Tinner W, Hubschmid P, Wehrli M, Ammann B, and Conedera M (1999) Long-term forest fire ecology and dynamics in southern Switzerland. *Journal of Ecology* 87: 273–289.
- van der Knaap WO (2009) Estimating pollen diversity from pollen accumulation rates: A method to assess taxonomic richness in the landscape. *The Holocene* 19: 159–163.
- Whitmore J, Gajewski K, Sawada M, et al. (2005) Modern pollen data from North America and Greenland for multi-scale paleoenvironmental applications. *Quaternary Science Reviews* 24: 1828–1848.
- Williams JW (2002) Variations in tree cover in North America since the last glacial maximum. *Global and Planetary Change* 35: 1–23.
- Williams JW, Tarasov P, Brewer S, and Notaro M (2011) Late Quaternary variations in tree cover at the northern forest-tundra ecotone. *Journal of Geophysical Research* 116: G01017. <http://dx.doi.org/10.1029/2010JG001458>.
- Willis KJ, Kleczkowski A, and Crowhurst SJ (1999) 124,000-year periodicity in terrestrial vegetation during the late Pliocene epoch. *Nature* 397: 685–688.

Databases and Their Application

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Introduction

Palynological data from the recent geological past have been invaluable for understanding ecological dynamics at time-scales inaccessible to direct observation, including the reorganization of communities and ecosystems as species responded individually to climate change, historical influences on contemporary patterns of biodiversity, and biotic responses to climatic change, both gradual and abrupt. Understanding the dynamics of ecological systems requires ecological time series, but many ecological processes operate too slowly to be amenable to experimentation or direct observation. In addition to having ecological significance, fossil-pollen data are crucial for climate-model verification and are essential for elucidating climate–vegetation feedbacks that can amplify or mitigate climate variations.

Basic palynological research is site based, and palynologists have devoted innumerable hours to identifying and counting pollen grains from cores, sections, and excavations. These data are typically published in papers describing one or a few sites. Usually, the data are published graphically, as in a pollen diagram, often only showing a few of the identified taxa, and the actual raw data reside on the investigator's computer or in a file cabinet. These basic data are similar to museum collections, in that they are costly to replace, often irreplaceable, and their value does not diminish with time. Moreover, the raw data are the primary resource for new interpretations and analytical approaches. Also, similar to museum collections, they require cataloging and curation. Whereas physical specimens of large fossils, such as animal bones, are typically accessioned into museums, microfossils, such as pollen, are often not accessioned, and the digital data are the primary objects. Thus, the loss of pollen data is equivalent to losing valuable museum specimens.

Several pollen database cooperatives have assembled pollen data from various parts of the world and have archived these data in relational databases. These databases fulfill two critical needs: (1) they are a secure archive of invaluable paleobiological data and (2) they are essential infrastructure for paleoenvironmental research. Although data from individual sites are valuable in their own right, emergent properties can be investigated by marshalling data from geographic arrays of sites for synoptic, broad-scale ecosystem studies. Examples of such properties are continent-scale climate and vegetation reconstructions, which provide the historical context for

understanding biodiversity dynamics, including genetic diversity. Synoptic climate reconstructions are crucial for climate-model validation.

For the past 20 years, the managers of the various regional pollen databases have collaborated to deploy their data in identical database structures. Some regional pollen databases have had their data merged from the outset, as did, for example, the North American Pollen Database (NAPD), the Latin American Pollen Database (LAPD), and the Pollen Database for Siberia and the Russian Far East (PDSRFE). In recent years, additional efforts have been made to merge other regional databases into a Global Pollen Database (GPD). The logical next step has been to merge the pollen databases into an even broader paleo database containing a wide variety of fossil and related data, which will facilitate even broader studies of community and ecosystem dynamics. This objective is now being achieved with the development of the Neotoma Paleocology Database.

Database History

Ever since Lennart von Post first introduced the science of paleopalynology and the first pollen diagram in 1916 ([von Post, 1967](#)), pollen data have accumulated as slides in desk drawers, count sheets in file cabinets, and files now stored on various electronic media. With the advent of radiocarbon dating in the 1950s, it became possible to compare temporally controlled pollen diagrams across vast areas to assess species migrations and expansions and synchronous and time-transgressive events. Such comparisons require a 'database,' whether it is a pile of pollen diagrams or digital data in a relational database.

Efforts to develop pollen databases began in North America and Europe in the 1970s. H.J.B. Birks and B. Huntley assembled data from sites throughout Europe to produce their massive atlas of isopoll maps at 500-year intervals beginning 13 000 ¹⁴C year BP for 55 taxa ([Huntley and Birks, 1983](#)). For the most part, they extracted data from published pollen diagrams. However, this procedure had limitations, and Huntley was impelled to become one of the original organizers of the European Pollen Database (EPD; [Fyfe et al., 2009](#)). In North America, efforts of T. Webb III and colleagues at Brown University in the 1970s and 1980s led to the first organized pollen database as part of the Cooperative Holocene Mapping

Project (COHMAP Members, 1988; Wright et al., 1993). Although climate data-model comparison was the principal objective of the COHMAP project, the synoptic analyses of the pollen data, particularly maps showing the constantly shifting ranges of species in response to climate change, were revelatory and led to much ecological insight.

The COHMAP pollen 'database' consisted of a multiplicity of flat files with prescribed formats for data and chronologies. FORTRAN programs were written to read these files and to assemble data for particular analyses. T. Webb III managed the COHMAP pollen database at Brown University, but as the quantity of data increased, data management became increasingly cumbersome. Clearly, the data needed to be transferred to a relational database management system (RDBMS). Discussions with E.C. Grimm, who was based in a museum with a mission of data preservation, led to the initiation of the NAPD at the Illinois State Museum in 1990. At the same time in Europe, the International Geological Correlation Project (IGCP) 158b was conducting a major collaborative synthesis of paleoecological data, primarily of pollen, and the need for a pollen database became painfully obvious (Fyfe et al., 2009). A workshop to develop an EPD was held in Sweden in 1989, and the organizers of NAPD and EPD commenced a long-standing collaboration to develop completely compatible databases. NAPD and EPD held several joint workshops and developed the same data structure. Nevertheless, the two databases were independently established, partly because Internet and software capabilities were not yet sufficient to easily manage a merged database. The pollen databases were developed in Paradox®, which at the time was the most powerful RDBMS software readily available for the personal computer (PC) platform. NAPD and EPD established two important protocols: (1) the databases were relational and queryable and (2) they were publicly available. All data in NAPD are in the public domain. EPD contains data that are public domain as well as smaller set of restricted data that require permission from the contributor for use. In an effort to encourage public access and to diminish the number of restricted sites, the EPD Advisory Board has set a three-year limitation, after which restricted status may continue but must be reaffirmed by the contributor.

As the success of the NAPD-EPD partnership grew in terms of contributed datasets and publications utilizing the databases, working groups initiated pollen databases for other regions, including the LAPD in 1994 (Marchant et al., 2002), the PDSRFE in 1995, and the African Pollen Database (APD) in 1996. At its initial organizational workshop, LAPD opted to merge with NAPD, rather than develop a standalone database. PDSRFE also followed this model, and the combined database was called the GPD. APD developed independently, but uses the exact table structure of NAPD and EPD. Pollen database projects have also been initiated in other regions, and the GPD contains some of these data, including the Indo-Pacific Pollen Database (IPPD).

The growth of the Quaternary pollen databases has been paralleled by the development of databases for other fossil types, including the North American Plant Macrofossil Database, FAUNMAP (a database of Quaternary-Pliocene mammals from North America), the North American Non-Marine Ostracode Database, the Diatom Paleolimnology Data Cooperative, and the Strategic Environmental Archaeology

Database/Bugs Database. These databases are similar in that they essentially store lists of taxa with quantitative measurements from stratigraphic contexts, with similar metadata. All of these databases face long-term sustainability issues associated with the costs of database maintenance and development and continued data acquisition. Thus, to consolidate maintenance costs and to build a multiproxy paleo database even more useful for addressing emergent global-scale ecosystem questions than the individual databases alone, the Neotoma Paleocology Database (www.neotomadb.org) was established in 2007.

Named for the packrat, genus *Neotoma*, prodigious collectors of anything in their territories, the Neotoma database realizes the ultimate goal of a multiproxy paleodatabase with global coverage. Neotoma includes fossil data for the past 5 million years (the Pliocene, Pleistocene, and Holocene Epochs) and provides the underlying cyberinfrastructure for a variety of disciplinary database projects. Although Neotoma is a new structure, it was built on the foundation laid particularly by the GPD and FAUNMAP. Neotoma can accommodate virtually any type of fossil data. The data are accessible to anyone with an Internet connection. Neotoma database management is centralized but scientific oversight is distributed, thus providing domain scientists with quality control over their portions of the data. To this end, all data in Neotoma are stored in a single centralized database, but are conceptually organized into virtual *constituent* databases, for example, NAPD and EPD, each of which has its own governance charged with overseeing scientific issues such as taxonomy, data quality standards, data acquisition, and prioritizing tool development. Constituent database cooperatives may develop individualized and branded websites to display and distribute their portions of the database if they so desire. Constituent database projects can appoint 'data stewards' to remotely upload and update data. The primary reason for the centralized design is lower requisite IT cost for development and long-term maintenance – only one database is being maintained rather than a plethora of autonomous databases all requiring managers. Furthermore, the centralized database structure simplifies cross-disciplinary multiproxy analyses and common tool development.

The Neotoma project was initially funded by a 2-year grant from the U.S. National Science Foundation (NSF) Geoinformatics Program and was launched with a workshop at Pennsylvania State University in February 2007. The initial task was to develop a data model suitable for a wide range of paleo data, which was accomplished by merging the GPD and FAUNMAP, which represent a broad spectrum of paleo fossil data and depositional contexts, the former primarily from wet-sediment cores, the latter mainly from terrestrial excavations. Neotoma received a 5-year renewal from the NSF Geoinformatics program, which began in September 2010, when a second workshop was held in Madison, Wisconsin. The two workshops involved 50–60 participants, representing a large spectrum of paleo data, including representatives from several of the pollen database cooperatives. Other research projects utilizing Neotoma are contributing data as well.

The data from the GPD and other paleoecological databases are now available from the Neotoma website. Users can search for individual sites or datasets on a map or by several different

search criteria. Datasets can be downloaded or copied and pasted into a spreadsheet for use with other applications. In addition, the entire database can be downloaded in Microsoft SQL Server or Access format. The database and website are hosted at the Center for Environmental Informatics at Pennsylvania State University.

Neotoma currently contains data from the NAPD, EPD, LAPD, PDSRE, and IPPD. The APD is currently independent, but is slated for incorporation into Neotoma. The LAPD and Japanese Pollen Database (JPD) have also inventoried data for future incorporation into Neotoma (Figure 1).

APD and LAPD contend with much larger and less well-known floras than do the pollen databases for temperate regions. APD has made a major effort at synonymization of pollen taxa and has produced a taxonomy list, which is available from its website. Images of pollen grains are also available from the APD website. Marchant et al. (2002) have published a major compendium of the ecology and distribution of sporophyte parent-taxa in LAPD.

The pollen databases make available data published in many languages, from many countries, and from literature not always widely available. Of particular note, PDSRFE assembled data from more than 175 sites, most of which were either unpublished or published in Russian in publications not widely available outside of Russia; and they published a compendium of these sites in Russian and English, with the original reference, a short description of each site, radiocarbon dates, short interpretation, and pollen diagram (Anderson and Lozhkin, 2002).

Database Structure

Neotoma consists of a large number of tables and fields for data and metadata. Central tables to the database are (Figure 2)

- **Sites:** Name and location of site.
- **Collection units:** Typically cores, sections, or excavation units. Sites may have multiple collection units, which have spatial coordinates.
- **Analysis units:** Individual units analyzed, typically identified by depth and thickness within a collection unit. Collection units may have multiple analysis units.
- **Samples:** A data type tabulated from an analysis unit, for example, a pollen sample, a plant macrofossil sample. An analysis unit may have multiple samples.
- **Datasets:** Samples of a particular data type from a collection unit comprise a dataset, for example, a pollen dataset. A collection unit may have multiple datasets (e.g., pollen, plant macrofossils, ostracodes), and a dataset may have multiple samples of the same data type.
- **Variables:** A variable consists of a taxon, element, and measurement unit. For plants, the element is the plant organ, for example, pollen, spore, seed, needle. Thus, *Picea* pollen and *Picea* needles are different variables. For pollen, possible measurement units are number of identified specimens, percent, or present/absent. Lookup tables exist for taxon names, element types, and measurement units.

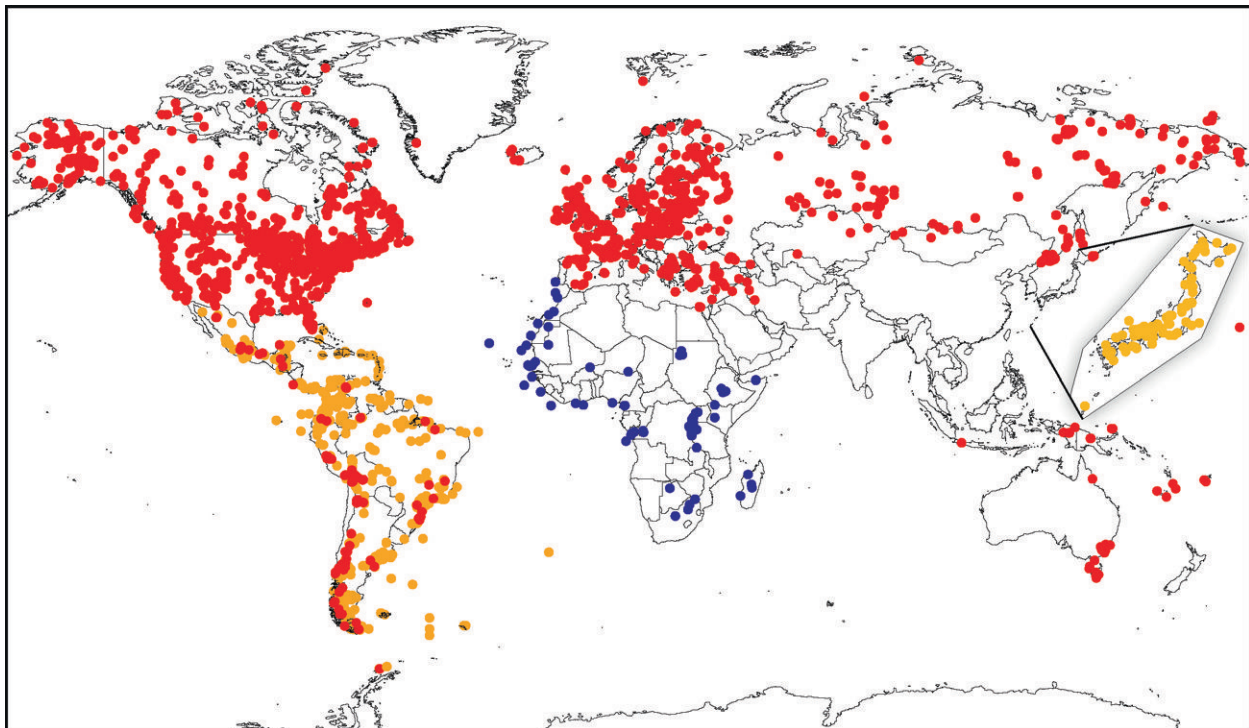


Figure 1 Pollen sites (cores and sections). Red dots – Neotoma Paleocology Database, blue dots – African Pollen Database, yellow dots – inventoried sites for inclusion.

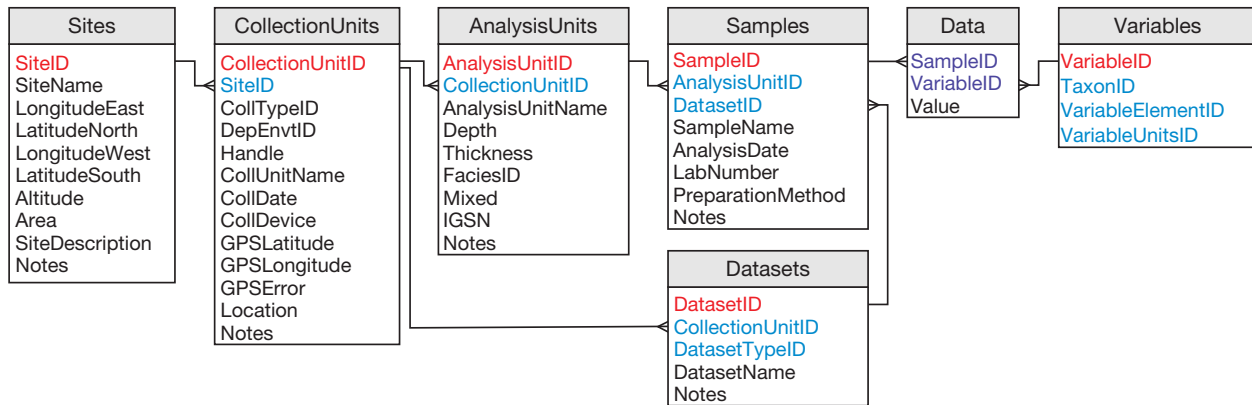


Figure 2 Central tables in the Neotoma Paleoecology Database. Table names are in the gray boxes. Red fields are primary keys; blue fields are foreign keys; violet fields are compound primary keys. Relationships shown by connection lines. IGSN, International geo sample number.

In addition, tables exist for:

- *Contacts*: principal investigators, collectors, analysts, data contributors;
- *Geochronology*: radiocarbon and other radiometric dates;
- *Chronologies*: chronology controls (e.g., calibrated ^{14}C ages, tephras, biostratigraphic ages), age models (algorithms for assigning sample ages based on chronology controls), and ages assigned to individual samples;
- *Lithology*: core or section descriptions;
- *Publications*: bibliographic references for data and metadata;
- *Synonymy*: accepted names and synonyms, original identifications.

Quality Control

The database cooperatives make every effort to ensure data accuracy. Quality control is maintained by having data stewards with palynological training enter data and by a software interface, currently under development, between the data input files and the database files. This interface checks for many kinds of common errors and inconsistencies. Probably, the thorniest quality control issues have been nomenclature and taxonomic synonymization. On a basic level, misspellings are corrected, but, in addition, the database does synonymize some names. The database coordinators must be cognizant of taxonomic resolution. Although some datasets may not contain the taxonomic resolution of others, they may nevertheless be suitable for synoptic studies. Whereas the quality of the data is ultimately the responsibility of the contributor, checking the appropriateness of any dataset is the responsibility of the user.

Taxonomy and Nomenclature

A protocol of the pollen databases is that contributors' names for pollen types are not changed in their meaning. Thus, a taxonomic identification is not questioned; although, of course, the database coordinators may choose not to include datasets with obvious misidentifications. However, some correction and synonymization is desirable and necessary to make the database more functional. Plant taxonomy of higher plants

is largely based on the diploid sporophyte generation, and taxonomic synonymization for pollen involves complications in addition to the usual issues of nomenclatural synonymy. An important problem arises from the fact that sporophyte names are used for pollen types, but pollen types may derive from more than one sporophyte taxon. As a consequence, names are combined, many pollen types have the suffix-type appended to them, and the usage of *cf.* to indicate uncertainty is common.

Taxonomic Hierarchy

The pollen database contains a flexible taxonomic hierarchy, which stores the maximum taxonomic precision made by the analyst but allows users to aggregate taxa into higher level categories. Whereas high taxonomic precision might be desirable for certain ecological studies, lumping is necessary for synoptic studies. The essence of the hierarchy is that each pollen variable has a field that points to the next higher variable in the hierarchy. For example, *Fraxinus nigra* points to *Fraxinus nigra*-type, which points to *Fraxinus*. The hierarchy then facilitates the retrieval of all instances of *Fraxinus* pollen. Changing botanical definitions and taxonomic synonymies are a persistent problem for databases of species occurrences.

Synonymy and Naming Conventions

Three general types of synonymy exist: nomenclatural, syntactic, and pollen morphological. The database synonymizes for nomenclature and syntax, but not for morphology. Nomenclatural synonymy refers to botanical nomenclature. The regional pollen databases follow major works relevant for their area, such as *Flora of North America* (Flora of North America Editorial Committee, 1993 et seq.), *Flora Europae* (Tutin et al., 1964–1993, 1993), *Vascular Plants of Russia and Adjacent States* (Czerepanov, 1995), *Énumération des plantes à fleurs d'Afrique tropicale* (Lebrun and Stork, 1991–1997), and the *Australian Plant Name Index* (Chapman, 1991). Inevitably, with a global database such as Neotoma, nomenclatural differences occasionally occur. In general, the newer authority will be followed, or if one name varies from recommendations of the *International Code of Botanical Nomenclature* (2006), then the name

following the Code will be used. Thus, for example, families with the 'aceae' endings such as Poaceae and Asteraceae are preferred over Gramineae and Compositae.

In Neotoma, Angiosperm family names and higher ranks follow the phylogenetic classification of the Angiosperm Phylogeny Group (APG; [Angiosperm Phylogeny Group, 2003, 2009; Stevens, 2001 et seq.](#)). Some of the APG revisions based on phylogenetics change long-standing taxonomic classifications, but are consistent with pollen morphology. For example, the genus *Celtis* has been moved from the Ulmaceae to the Cannabaceae. The scabrate, triporate pollen grains of *Celtis* are more consistent with the Cannabaceae than with the rugulate, stephanoporate pollen grains of *Ulmus* and other genera remaining in the Ulmaceae. Pollen from the traditional families Chenopodiaceae and Amaranthaceae are commonly grouped together as Chenopodiaceae/Amaranthaceae (or often informally as 'ChenoAms'). Conveniently from the

palynological standpoint, APG has synonymized Chenopodiaceae with Amaranthaceae, thus the Chenopodiaceae/Amaranthaceae simply becomes Amaranthaceae (as does the former superfamily Chenopodiineae).

Nomenclatural synonymization should not change the range of pollen morphotypes intended by the original analyst. For example, APG has synonymized the Agavaceae with the Asparagaceae, a family that includes many more taxa than the former Agavaceae. However, the genera of the former Agavaceae now comprise the subfamily Agavoideae, and the name is thus changed to Asparagaceae subf. Agavoideae. Neotoma has tables for synonyms and for original identifications. Thus, an original identification of Agavaceae can be retained. In this example, the circumscription of potential taxa has not changed. However, synonymization that involves splitting or lumping potentially does change the circumscription of morphotypes included in the analyst's original identification. For example, an

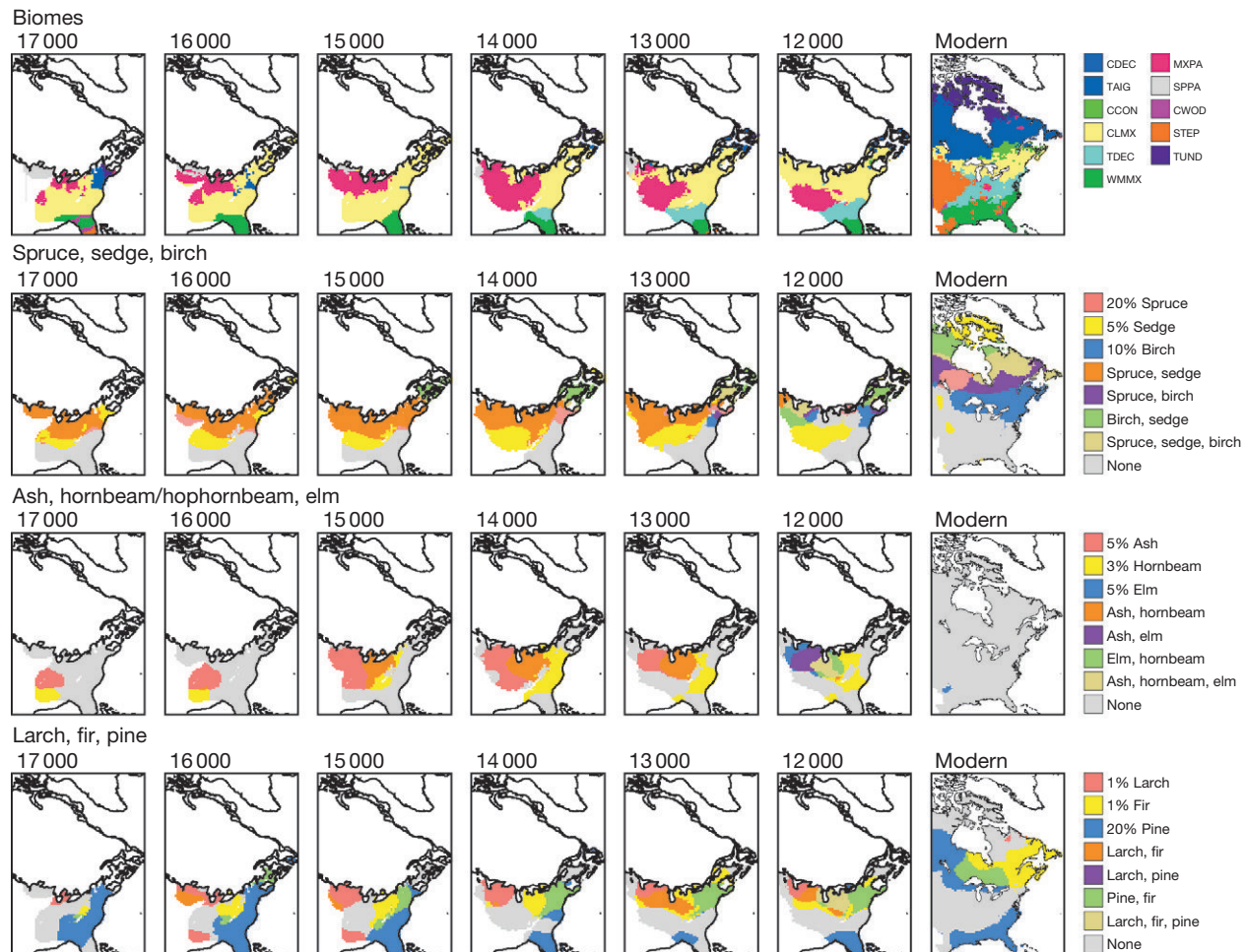


Figure 3 Representations of late Pleistocene and modern plant associations as biomes (top row) and as multitaxon isopoll maps (rows 2–4). Numbers indicate cal year BP. Biomes are: CDEC, cold deciduous forest; TAIG, Taiga; CCON, cool conifer forest; CLMX, cool mixed forest; TDEC, temperate deciduous forest; WMMX, warm mixed forest; MXPA, mixed parkland; SPPA, spruce parkland; CWOD, conifer woodland; STEP, steppe; TUND, tundra. Three taxa are mapped in each multitaxon map, with a single isopoll shown for each plant taxon. Primary colors (red, blue, cyan) indicate regions where only one of the three taxa is present in appreciable numbers. Secondary colors (orange, purple, green) denote associations between pairs of taxa. Beige indicates where all three taxa overlap. Scientific names: spruce, *Picea*; sedge, *Cyperaceae*; birch, *Betula*; ash, *Fraxinus*; hornbeam, *Ostrya/Carpinus*; elm, *Ulmus*; larch, *Larix*; fir, *Abies*; pine, *Pinus*. Reproduced from [Williams JW, Shuman BN, and Webb III T \(2001\)](#) Dissimilarity analyses of late-Quaternary vegetation and climate in eastern North America. *Ecology* 82: 3346–3362, with permission.

identification of Cannabaceae prior to APG would have included the genera *Cannabis* and *Humulus*, but probably not *Celtis*. Thus, in Neotoma these earlier identifications of Cannabaceae have been changed to Cannabaceae *sensu stricto*.

Syntax refers to elements of pollen-type names that are not parts of the Linnean system (i.e., species, genus, family...). Neotoma has a set of rules for pollen syntax; for example, for types the convention is name-type with a hyphen, for example, *Ambrosia*-type, not *Ambrosia* type. For a global database, types are a problem because they often are not precisely defined, or they are defined differently in different regions. Researchers need to be aware of these issues of taxonomic precision when utilizing the database.

Pollen types with different names may be morphologically identical, but the GPD does not attempt to synonymize these for several reasons. The principal reason is that, unlike floras for the sporophyte life-history stage, complete palynological floras and monographs do not exist. The database coordinators, although experienced palynologists, have no published authority on which to base their morphological naming conventions, nor do they have the expertise to determine the full realm of morphological synonyms. They could possibly reject fine morphological distinctions that, in fact, a given palynologist has made with confidence. Although, the GPD does not synonymize based on morphology, some regional pollen databases that contribute to the GPD may synonymize based on intimate knowledge of the palynology of their region. Another important argument against morphological synonymization is that the pollen-type name may indicate the most probable derivative sporophyte taxon based on biogeographic or ecological considerations.

One other database protocol that produces name changes is that, for any dataset, pollen-type names must be mutually exclusive, which is necessary for lumping algorithms to work properly. For example, if a dataset includes *Potentilla* and Rosaceae, the latter is changed to Rosaceae undiff., because Rosaceae includes *Potentilla*. Some of these names can be somewhat cumbersome. For example, many contributed datasets include *Artemisia* and Asteroideae (or the obsolete Tubuliflorae). The latter is changed to Asteraceae subf. Asteroideae undiff.

Database Applications

The pollen databases have been widely used. Applications range from seeking to place a new record into a regional context (e.g., Nelson and Hu, 2008; Schauffler and Jacobson, 2002) to synoptic mapping at continental to global scale (e.g., Allen et al., 2010) and paleoclimatic reconstructions for evaluation of global climate models (e.g., Bartlein et al., 2011). Applications range from evolutionary biology and macroecology (e.g., DiMichele et al., 2004; Giesecke and Bennett, 2004; Petit et al., 2004) to natural resource management (e.g., Cole et al., 1998; Heard and Valente, 2009). Given the diversity of studies utilizing the pollen database, selection of a few representative examples is difficult. Nonetheless, the following two illustrate some of the more powerful synoptic applications.

Biomization is a particularly powerful technique for deriving biomes from pollen assemblages. Rather than using actual species, biomization utilizes plant functional types (PFTs),

which are based on life form, physiology, phenology, leaf form, and climatic tolerances. The method entails assigning PFTs to biomes globally, assigning pollen taxa to PFTs regionally, and using these assignments to assign regional pollen taxa to biomes. For fossil samples, 'affinity scores' are calculated between pollen samples and biomes. The biome with the highest affinity score is assigned to the fossil-pollen sample, subject to a tie-breaking rule (Prentice et al., 1992, 1996; Williams et al., 1998, 2000, 2001). Advantages of biomization are that nonanalog biomes are definable and that no reliance is placed on comparison with modern samples, which may be biased by anthropogenic activities. Williams et al. (2001) mapped biomes and major taxa for the Late Glacial interval (17 000–12 000 cal year BP) in eastern North America (Figure 3). Mapped are two nonanalog biomes – mixed parkland and spruce parkland – that no longer exist. The maps clearly show the changing biomes and taxon distributions through time. This study also showed that the peak interval of dissimilarity from the present preceded the time of maximum climate change. Thus, nonanalog vegetation does not appear to be a nonequilibrium transient state, but, in fact, is in equilibrium with nonanalog climate conditions.

Maps of fossil-pollen data show species migrations from glacial ranges or 'refugia' to their modern ranges. Modern genetic data also contain clues to the past migration history of taxa. In Europe, collaboration between palynologists and geneticists has produced a detailed history of the migration of deciduous *Quercus* from glacial refugia (Brewer et al., 2002; Petit et al., 2002). Because of frequent introgression, the various *Quercus* species share chloroplast DNA (cpDNA) haplotypes regionally, and haplotype lineages cross species lines. Consequently, the taxonomic resolution of the cpDNA lineages is comparable to the identification of pollen to the genus level. Palynological data (Figure 4) and the geographic distribution of haplotypes (Figure 5) show the existence of refugia

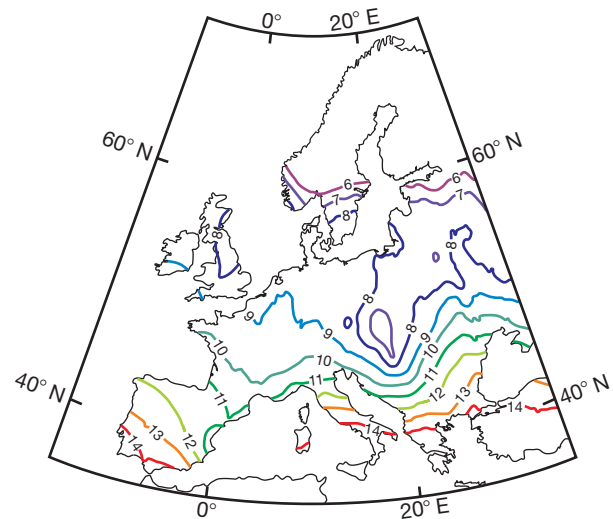


Figure 4 Isopoll map showing time of first arrival of deciduous *Quercus*. Adapted from Brewer S, Cheddadi R, de Beaulieu J-L, Reille M, and Data contributors (2002) The spread of deciduous *Quercus* throughout Europe since the last glacial period. *Forest Ecology and Management* 156: 27–48.

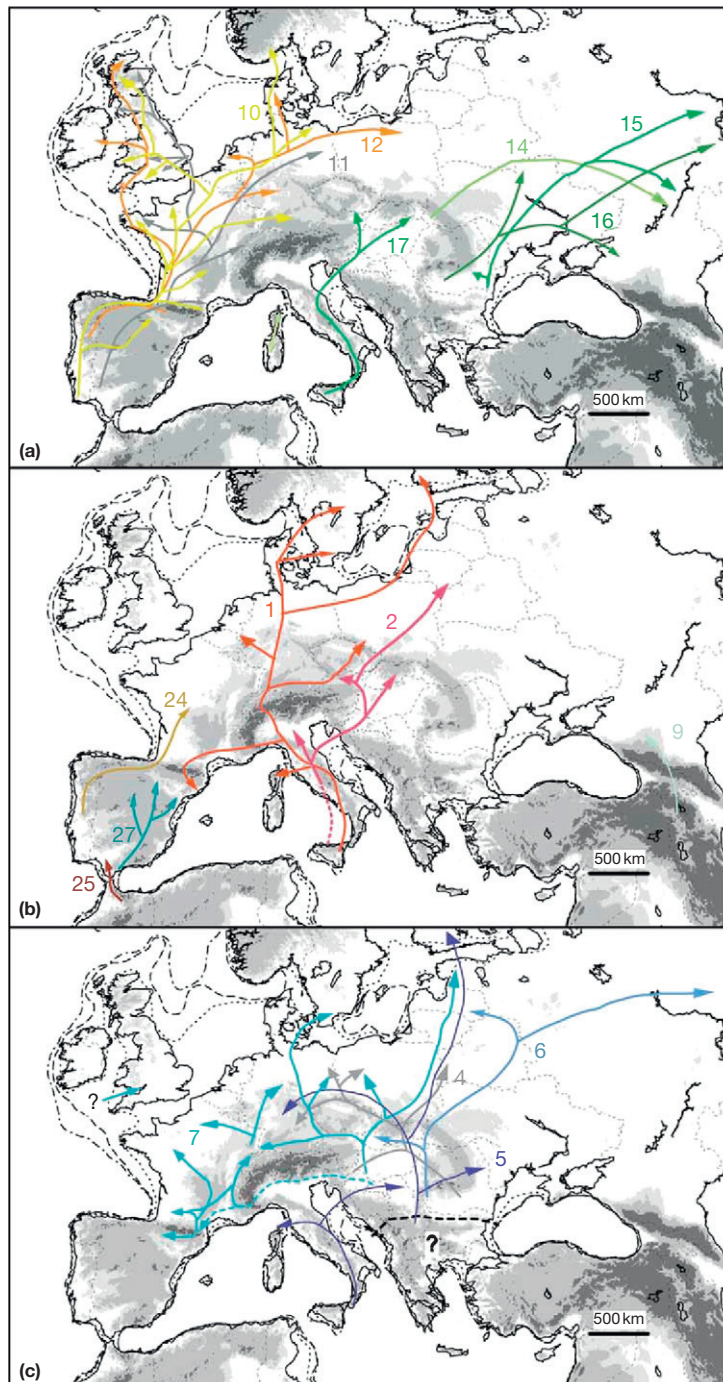


Figure 5 Maps showing inferred postglacial colonization routes for frequent cpDNA haplotypes of deciduous *Quercus*. Colors indicate different lineages, and numbers indicate haplotypes. Maps (a), (b), and (c) show different haplotypes moving from the same refugia. Gray shading indicates altitude. Reproduced from Petit RJ, Brewer S, Bordács S, et al. (2002) Identification of refugia and post-glacial colonisation routes of European white oaks based on chloroplast DNA and fossil pollen evidence. *Forest Ecology and Management* 156: 49–74.

in the Iberian Peninsula, Apennine Peninsula, and Balkan Peninsula; and isopoll maps and the distributions of haplotypes show migration routes north from these southern peninsulas to northern Europe and into central Asia. Pollen and genetic data are in agreement about general migration trends. Synoptic mapping of *Quercus* pollen shows where refugia were and when migrations occurred. Genetic survey adds detail to

the migration history; for example, DNA data show that *Quercus* in the British Isles is related to the Iberian Peninsula populations, rather than to those from the Apennine refugium. Genetic data can also suggest where more palynological work is warranted. For example, a unique haplotype in Corsica and Sardinia suggests a refugium there, although no pollen data are available to confirm it.

Conclusions

The quantity of data in the pollen databases and the hundreds of papers that have utilized these databases demonstrate their great success. The databases are public and access is free. The databases have fostered collaboration between scientists in different disciplines and from different countries, and they are critical to global change research.

The pollen data were some of the first paleo data to be organized into a relational database structure. Many lessons concerning scientific database organization have been learned. To be successful, the database must satisfy needs and wishes of both data contributors and data consumers. Some scientists had initial reluctance to contribute data. However, willingness to participate has increased with the maturation of the database and the increasing evidence of its value as a scientific resource, combined with the broader informatics revolution. A large database effort must involve a number of data cooperatives. Unfunded data contributors will send data 'as is,' but typically they will not spend much time entering or reformatting data to fit database standards.

The incorporation of the pollen databases into the multi-proxy Neotoma Paleoeology Database attends to several long-standing needs. The common cyberinfrastructure should reduce long-term maintenance costs, enhance data preservation, and guard against data loss, while the distributed governance permits expert review and oversight for particular data types. The development of a software interface allowing data stewards from the various pollen database cooperatives to upload and manage data remotely should permit these cooperatives to focus on data entry rather than IT needs. Neotoma will facilitate new multi-proxy studies of Quaternary environments.

See also: Pollen Records, Last Interglacial of Europe. **Plant Macrofossil Methods and Studies:** Validation of Pollen Studies. **Pollen Methods and Studies:** Archaeological Applications. **Pollen Records, Late Pleistocene:** Africa; Australasia; Middle and Late Pleistocene in Southern Europe; Northern Asia; Northern North America; South America; Western North America. **Pollen Records, Postglacial:** Africa; Australia and New Zealand; Northeastern North America; Northern Asia; Northern Europe; Northwestern North America; South America; Southeastern North America; Southern Europe; Southwestern North America.

References

- Allen JRM, Hickler T, Singarayer JS, Sykes MT, Valdes PJ, and Huntley B (2010) Late glacial vegetation of northern Eurasia. *Quaternary Science Reviews* 29: 2604–2618.
- Anderson PM and Lozhkin AV (2002) *Late Quaternary Vegetation and Climate of Siberia and the Russian Far East (Palynological and Radiocarbon Database)*. Magadan: North East Science Center, Far East Branch, Russian Academy of Sciences.
- Angiosperm Phylogeny Group (2003) An update of the Angiosperm Phylogeny Group classification for the orders and families of flowering plants: APG II. *Botanical Journal of the Linnean Society* 141: 399–436.
- Angiosperm Phylogeny Group (2009) An update of the Angiosperm Phylogeny Group classification for the orders and families of flowering plants: APG III. *Botanical Journal of the Linnean Society* 161: 105–121.
- Bartlein PJ, Harrison SP, Brewer S, et al. (2011) Pollen-based continental climate reconstructions at 6 and 21 ka: A global synthesis. *Climate Dynamics* 37: 775–802.
- Brewer S, Cheddadi R, de Beaulieu J-L, and Reille M (2002) Data contributors: The spread of deciduous *Quercus* throughout Europe since the last glacial period. *Forest Ecology and Management* 156: 27–48.
- Chapman AD (1991) *Australian Plant Name Index. Australian Flora and Fauna Series 12*, 4 vols. Canberra: Australian Government Publishing Service.
- COHMAP Members (1988) Climatic changes of the last 18,000 years: Observations and model simulations. *Science* 241: 1043–1052.
- Cole KL, Davis MB, Stearns F, Walker K, and Guntenspergen G (1998) Historical landcover changes in the Great Lakes region. In: Sisk TD (ed.) *Perspectives on Land Use History of North America: A Context for Understanding Our Changing Environment*. United States Geological Survey, Biological Resources Division, Biological Science Report USGS/BRD/BSR 1998-0003, pp. 43–50.
- Czerepanov SK (1995) *Vascular Plants of Russia and Adjacent States (the Former USSR)*. Cambridge: Cambridge University Press.
- DiMichele WA, Behrensmeyer AK, Olszewski TD, et al. (2004) Long-term stasis in ecological assemblages: Evidence from the fossil record. *Annual Review of Ecology, Evolution, and Systematics* 35: 285–322.
- Flora of North America Editorial Committee (1993 et seq.) *Flora of North America North of Mexico*, 14+ vols. New York: Oxford University Press.
- Fyfe RM, de Beaulieu J-L, Binney H, et al. (2009) The European Pollen Database: Past efforts and current activities. *Vegetation History and Archaeobotany* 18: 417–424.
- Giesecke T and Bennett KD (2004) The Holocene spread of *Picea abies* (L.) Karst. in Fennoscandia and adjacent areas. *Journal of Biogeography* 31: 1523–1548.
- Heard MJ and Valente MJ (2009) Fossil pollen records forecast response of forests to hemlock woolly adelgid invasion. *Ecography* 32: 881–887.
- Huntley B and Birks HJB (1983) *An Atlas of Past and Present Pollen Maps for Europe: 0–13000 Years Ago*. Cambridge: Cambridge University Press.
- International Code of Botanical Nomenclature (Vienna Code) (2006) *Regnum Vegetabile*, vol. 146. Ruggell: A.R.G. Gantner Verlag KG.
- Lebrun J-P and Stork AL (1991–1997) *Énumération des plantes à fleurs d'Afrique Tropicale*, 4 vols. Geneva: Conservatoire et Jardins Botaniques de la Ville de Genève.
- Marchant R, Almeida L, Behling H, et al. (2002) Distribution and ecology of parent taxa of pollen lodged within the Latin American Pollen Database. *Review of Palaeobotany and Palynology* 121: 1–75.
- Nelson DM and Hu FS (2008) Patterns and drivers of Holocene vegetational change near the prairie-forest ecotone in Minnesota: Revisiting McAndrews' transect. *New Phytologist* 179: 449–459.
- Petit RJ, Bialozyt R, Garnier-Géré P, and Hampe A (2004) Ecology and genetics of tree invasions: From recent introductions to Quaternary migrations. *Forest Ecology and Management* 197: 117–137.
- Petit RJ, Brewer S, Bordács S, et al. (2002) Identification of refugia and post-glacial colonisation routes of European white oaks based on chloroplast DNA and fossil pollen evidence. *Forest Ecology and Management* 156: 49–74.
- Prentice IC, Cramer W, Harrison SP, Leemans R, Monserud RA, and Solomon AM (1992) A global biome model based on plant physiology and dominance, soil properties and climate. *Journal of Biogeography* 19: 117–134.
- Prentice IC, Guiot J, Huntley B, Jolly D, and Cheddadi R (1996) Reconstructing biomes from palaeoecological data: A general method and its application to European pollen data at 0 and 6 ka. *Climate Dynamics* 12: 185–194.
- Schauffler M and Jacobson GL Jr. (2002) Persistence of coastal spruce refugia during the Holocene in northern New England, USA, detected by stand-scale pollen stratigraphies. *Journal of Ecology* 90: 235–250.
- Stevens PF (2001 et seq.) Angiosperm Phylogeny Website. Version 9, June 2008. <http://www.mobot.org/MOBOT/research/APweb/>.
- Tutin TG, Heywood VH, and Burges NA, et al. (eds.) (1964–1993) In: *Flora Europaea*, 5 vols. Cambridge: Cambridge University Press.
- Tutin TG, Burges NA, Chater AO, et al. (1993) *Psilotaceae to Platanaceae, Flora Europaea*, 2nd edn., vol. 1. Cambridge: Cambridge University Press.
- von Post L (1967) Forest tree pollen in south Swedish peat bog sediments. *Pollen et Spores* 9: 375–401.
- Williams JW, Shuman BN, and Webb T III (2001) Dissimilarity analyses of late-Quaternary vegetation and climate in eastern North America. *Ecology* 82: 3346–3362.
- Williams JW, Summers RL, and Webb T III (1998) Applying plant functional types to construct biome maps from eastern North American pollen data: Comparisons with model results. *Quaternary Science Reviews* 17: 607–627.
- Williams JW, Webb T III, Richard PH, and Newby P (2000) Late Quaternary biomes of Canada and the eastern United States. *Journal of Biogeography* 27: 585–607.
- Wright HE Jr., Kutzbach JE, Webb T III, Ruddiman WF, Street-Perrott FA, and Bartlein PJ (eds.) (1993) *Global Climates Since the Last Glacial Maximum*. Minneapolis: University of Minnesota Press.

Surface Samples and Trapping

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Pollen Surface Samples and Pollen Trapping

The Need for Surface Samples and Pollen Trapping

The pollen assemblages preserved in peat and lake sediments reflect the vegetation that existed at the time when that particular sediment was accumulating. The translation of these assemblages into actual vegetation communities, however, is complicated by the differential pollen productivity, differential pollen dispersal, and differential pollen preservation properties between taxa (see [Pollen Analysis, Principles](#)). Therefore, in order to better understand and quantify fossil pollen assemblages, it is necessary to collect present-day information about the relationship between the pollen being deposited on a lake bottom or peat surface and the vegetation that produces it. Such present-day reference material falls into two broad categories: pollen extracted from surface samples and pollen collected by means of pollen traps. The major difference between the two is that for surface samples, the time period represented by the pollen assemblage is unknown, whereas in the case of pollen traps, the time interval over which pollen is being collected can be controlled. For traps, this knowledge of the time interval allows the rate at which pollen accumulates on a known surface area to be calculated, and in consequence, each pollen taxon can be considered individually rather than in terms of its percentage presence in the whole (or a predefined) assemblage.

Modern reference pollen material has been collected for almost as long as the technique of pollen analysis has been in existence, and some of the earliest work in this field ([Aario, 1943](#)) is still useful and valid today. Collection of moss polsters or lake-surface sediment is the more common procedure and the first to be employed, whereas pollen traps have been used to collect reference material for paleoecological investigations since the 1970s ([Cundill, 1986; Tauber, 1974](#)).

Surface Samples

Surface samples can be obtained from a range of material (see [Surface Samples, Taphonomy, Representation](#)). The most common source materials are moss polsters and lake-surface sediments, because these most closely resemble the characteristics of the sites recording past environments with which their pollen assemblages will be matched. In areas where such materials are missing, however, lichens, soils, and leaf litter have been used. The latter are less desirable because of the uncertainty of pollen capture and preservation, which could lead to a selectively biased pollen assemblage. Surface samples have also been obtained from snow, but in this situation, the quantity of pollen is often too small for any reliable statistical treatment ([Hicks and Isaksson, 2005](#)) or represents an abnormal and rarely occurring event ([Hjelmroos and Franzen, 1994](#)).

Two main strategies have been employed for moss polster surface sample collection. One involves taking single samples of known surface area and predetermined depth, one for each vegetation situation ([Figure 1](#)), the other of taking ten random samples of varying size from within a designated area of the studied vegetation community and amalgamating them ([Hicks et al., 1998](#)). The strategy for sampling should be related to the specific research approach in which the samples will be used (see also the section '[Sample Collection Strategy, Accompanying Data, and Archiving](#)'). The multiple-sample record may be more appropriate in areas with high plant biodiversity (meadows or Arctic tundra) and where the data set will be used for the modern analog approach (see section '[Past Land Use and Anthropogenic Disturbance to the Vegetation](#)') whereas the single-sample method is useful in species-poor communities and where comparisons with data from pollen traps are to be made, because the same surface area can be used for both.

The length of the sampling period covered by a moss polster has been much debated ([Cundill, 1991](#)), although very few experiments have been made to actually assess this. The exceptions are the studies by [Räsänen et al. \(2004\)](#) and [Pardoe et al. \(2010\)](#), who conclude that the time period is probably around 2 years. This is important because not only does pollen production vary enormously from year to year but some species (particularly temperate trees) exhibit regionally coherent cyclicity in their flowering and seed production ([Kelly, 1994; Koenig and Knops, 1998; Ranta and Satri, 2007](#)). In vegetation situations where the dominant species are big pollen producers but one of them shows cyclicity in flowering, if a sample represents only a single year, or even 2 years, the percentage presence of each pollen type can vary enormously depending upon the year in which it is sampled ([Figure 2](#)). In such situations, therefore, the percentage records have very little meaning.

There are several types of samplers available for sampling surface sediments of lakes. The most commonly used are the Kajak and Limnos samplers ([Figure 3](#)), although the uppermost layer of a frigid finger core can also be used ([Glew et al., 2001; Kansanen et al., 1991](#)). The length of time incorporated in a lake-sediment surface sample varies with the rate of sediment accumulation in the lake in question. This may be only a few years in highly productive lakes but can be tens of years in Arctic lakes ([Prentice, 1978](#)). The surface sediment of lakes with inflow streams will contain pollen coming from the catchment area of the lake, whereas the surface sediment of a lake with no inflow will contain predominantly pollen from the air and from plants growing in the lake itself and along its shores. It is potentially possible to obtain an annual sample from varved lakes (see [Varved Lake Sediments](#)), if they can be sampled varve by varve, but even so the pollen within a single varve could be a mixture of pollen produced in different years



Figure 1 Sampling a moss polster as a single (nonamalgamated) surface sample of known surface area. The moss is cut off at the ground surface, but all mineral material is excluded.

because of the mechanisms of sedimentation (Davis, 1973; Davis et al., 1971).

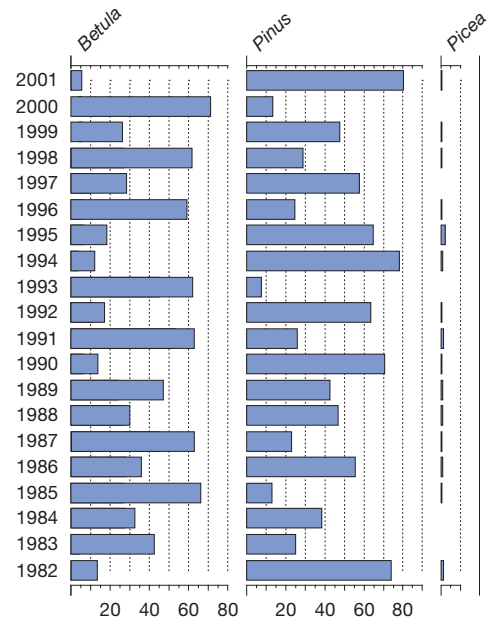
As a general rule, moss polster samples should be used as reference material for interpreting pollen assemblages extracted from peat, whereas samples of surface sediments of lakes are best as a reference for pollen assemblages from lake-sediment profiles. For some specific research questions, however (see Table 1), results from one type of source material are applied to the other.

One distinct advantage that surface samples have over pollen traps is that a large number of samples with a good spatial coverage can be collected during one field season, whereas, for usable data, pollen monitoring with pollen traps has to be continued for at least 5 years, and this often sets a limit on the number of pollen traps that is feasible to maintain.

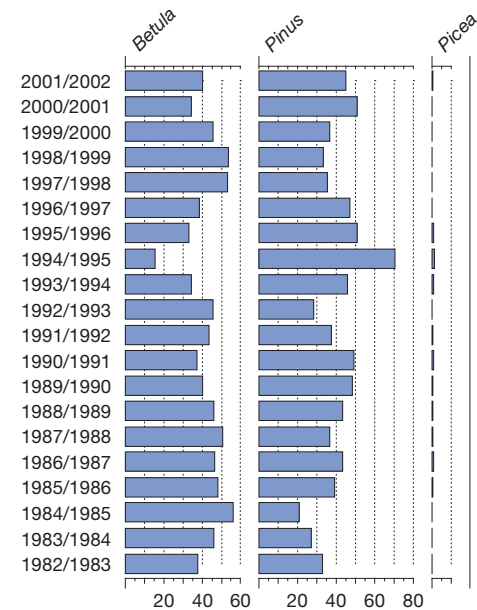
Pollen Trapping

The most commonly used pollen trap in Quaternary studies is the Tauber-type trap (Hicks and Hyvärinen, 1986; Tauber, 1974). This trap, named after the Danish physicist who developed it in order to investigate pollen dispersal, is (in a slightly modified form) now the standard trap used by the pollen monitoring program (PMP; Hicks et al., 2001; Giesecke et al., 2010), a network uniting scientists interested in pollen monitoring. Other major trap types include the Cundill trap (Cundill, 1986), the Burkard sampler (Hirst, 1952), and Cour filters (Cour, 1974). The latter two are used primarily for aerobiological studies, which monitor the pollen content of the air and so will not be considered here. The Cundill trap is applicable for Quaternary studies and is preferred to the Tauber-type trap in situations where the yearly rainfall is extremely high and the vegetation is very dense, that is, in the tropics (Gosling et al., 2003).

The Tauber-type trap is usually sunk into the ground surface so that its sloping collar top with the 5-cm-diameter central opening is at ground level (Figure 4). However, depending on the research question, the Tauber-type traps can be installed at many other positions, such as on a floating raft, on a pole, or on a roof-top (Giesecke and Fontana, 2008; Levetin et al., 2000; Solomon and Silkworth, 1986).



(a)



(b)

Figure 2 Annual pollen quantities that have accumulated on a known area of land surface (the same results as in Figure 5), expressed as a percentage of the terrestrial pollen sum: (a) year-by-year and (b) averages of two consecutive years. Note the wide range in percentage values over the sampling period for the two major tree types, even though the monitoring site has always been in the same place and the surrounding vegetation has remained constant. In this example, *Picea* is not present in the local vegetation.

Ideally, monitoring locations are in the center of small (ca. 200 m diameter) mires so that the situation resembles, as closely as possible, the fossil one with which results will be compared. In forested situations, it is desirable that, if a mire is not available, the trap is placed within an opening in the forest, unless the results are to be used to interpret fossil assemblages

from small forest hollows. In temperate areas, the trap remains in position throughout the year, from the end of one growing season to the end of next one, the contents being collected at the end of the growing season. In tropical areas, without the dormant season, the span of monitoring time may be specific to other annual cycles. In this way, the temporal resolution of the results is one full year (the best resolution that is feasible

to obtain from peat or lake sediments) and includes all pollen coming to the ground surface, both directly from that summer's local pollen emission and also through resuspension and wind transport from sources at even considerable distances (Hicks, 2001).

The lower part of the trap, which is buried in the ground and in which the pollen is collected, can be of any size but should be large enough to contain the precipitation of the whole monitoring year. At the beginning of the monitoring season, enough glycerol is poured into the container part of the trap to cover the entire base. This is to make sure that any pollen entering the trap remains moist and is preserved. Small quantities of formalin



Figure 3 Sampling lake-surface sediment with a Limnos sampler. With this sampler, the loose uppermost sediments can be removed centimeter by centimeter.

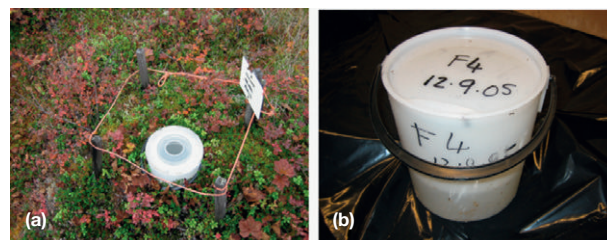


Figure 4 A modified Tauber trap used for monitoring pollen deposition in accordance with the PMP protocol. (a) The trap in position in a *Sphagnum* (Peat moss) tussock. (b) The year's collection ready for transportation to the laboratory. Field procedure is simple and designed to prevent sample loss. The special monitoring lid is removed and placed on a new container while the container that has been in the field for the whole year is closed with a normal entire lid.

Table 1 Sampling strategy in relation to the research question for which the modern pollen data set will be used

Approach	Number and location of samples	Source material				Required vegetation data
		Moss polster	Lake-sediment surface sediment	Soil sample	Pollen trap	
Modern analogs: indicator species approach	Samples from within those vegetation communities for which an analog is required	X	X	X	X	Species list and abundance
Modern analogs: whole assemblage comparative approach	As above but with at least 30 sampling points	X	X	X		Species cover/frequency
Transfer functions: a. local/regional b. subcontinental	A comprehensive evenly spaced network of points covering the area/ regional/subcontinent in question	X	X	X		Major vegetation units. The sites should not be highly anthropogenic
Pollen productivity estimates	At least 30 sampling points randomly through the landscape	X	X	X	X	Detailed vegetation analyses centered on each sample plus regional vegetation out to tens of km (field data + air photos)
Background pollen estimates	Center of large lakes		X			Major vegetation units from satellite images
PARs, thresholds of abundance, measurements of pollen dispersal	A transect of monitoring points across the limit of the species in question or radiating at increasing distances from the species				X	Both detailed local- and coarser-scale regional analyses (field data + air photos)

Note: Although all the source materials are possible, they should be regarded as alternatives. That is, the data set should, in most instances, be all derived from one type of source material (either moss polsters or surface sediments), not a mixture between the different types of collecting sites.

and thymol are also added both to kill any animals and insects that fall into the trap and to prevent bacterial and fungal growth. To prevent contamination of the sample by small rodents, amphibians, or insects attracted to the traps by the glycerol, water or decaying animals, a piece of mesh or few metal bars can be placed somewhat below the opening of the trap (Koff, 2001). There are often practical difficulties with traps, such as vandalism by people, especially when the traps are placed in densely populated areas, or damage by large animals (e.g., bears and bovines) attracted to the glycerol, sometimes making it impossible to conduct pollen trapping at certain locations (Hall, 1990; Tonkov et al., 2001; van der Knaap et al., 2001).

Pollen traps are most easily placed on land and, therefore, produce data applicable for a terrestrial environment, although there is a growing body of evidence to suggest that the results can also be applied to interpret pollen records from lake cores provided that the lakes have no inflow streams and are of the same size as the opening in the forest in which the pollen trap was placed (Hicks and Hyvärinen, 1999). Experiments made by Giesecke and Fontana (2008) with pollen traps floating on the lake-surface to evaluate the relationship between pollen accumulating in this situation and that accumulating at the lake floor and in the terrestrial traps found no substantial differences in observed pollen deposition.

During the laboratory preparation of the pollen trap contents, marker grains are added (Stockmarr, 1971) which enable the total pollen concentration of the trap contents to be calculated. Because the pollen is collected for 1 year and the surface area of the collection opening is known, the pollen accumulation rate (PAR) in grains $\text{cm}^{-2} \text{year}^{-1}$ can then be calculated. The term 'influx' has frequently erroneously (Thompson, 1980) been used in the literature to refer to the amount of pollen $\text{cm}^{-2} \text{year}^{-1}$, but correctly this should be referred to as PAR. The precision of the PAR calculation depends on the relationship between the number of marker grains and the number of pollen grains counted in the preparation, so ideally the results should be expressed together with their confidence intervals (Maher, 1981).

Results of pollen trapping in northern Finland that have continued for more than 40 years demonstrate great annual

variation in the PAR (Figure 5; for the confidence limits on these calculations, see Hicks, 2001), to the extent that the difference in the quantity of pollen from 1 year to the next is frequently bigger than the difference in the quantity of pollen of one taxon (e.g., *Pinus* (Pine)) between different vegetation zones. This annual variation reflects regional pollen production, which is determined by weather conditions (in the north, primarily temperature) of the year before pollen emission. When the long-term average PAR is considered, however, the value reflects the density and abundance of the different taxa in the region surrounding the trap (Table 2). The results from pollen traps, therefore, differ from those from surface samples in that, through the PARs, two signals can be extracted, one of climate and one of plant abundance, depending upon the temporal resolution at which they are considered.

For both surface samples and pollen traps, all microfossils present are identified and counted. Thus, in addition to pollen and spores, coniferous stomata, charcoal (see Charred Particle Analyses), spheroidal carbonaceous particles, testate amoebae, fungal spores, and non-pollen palynomorphs are recorded. This range of microfossils enables the results to be applied to a wider range of research questions than if only pollen and spores were recorded.

Sample Collection Strategy, Accompanying Data, and Archiving

The number of surface samples/pollen traps and their location and the detail and extent of any associated vegetation analyses varies in accordance with the intended use of the data set. Although the prime aim of the results of both types of analysis is a more objective or even quantified interpretation of fossil pollen assemblages, there are numerous ways in which this can be achieved. These include:

- modern analogs, either using the indicator species approach or the whole assemblage comparative approach;
- transfer functions, either at the local/regional or subcontinental scale; and

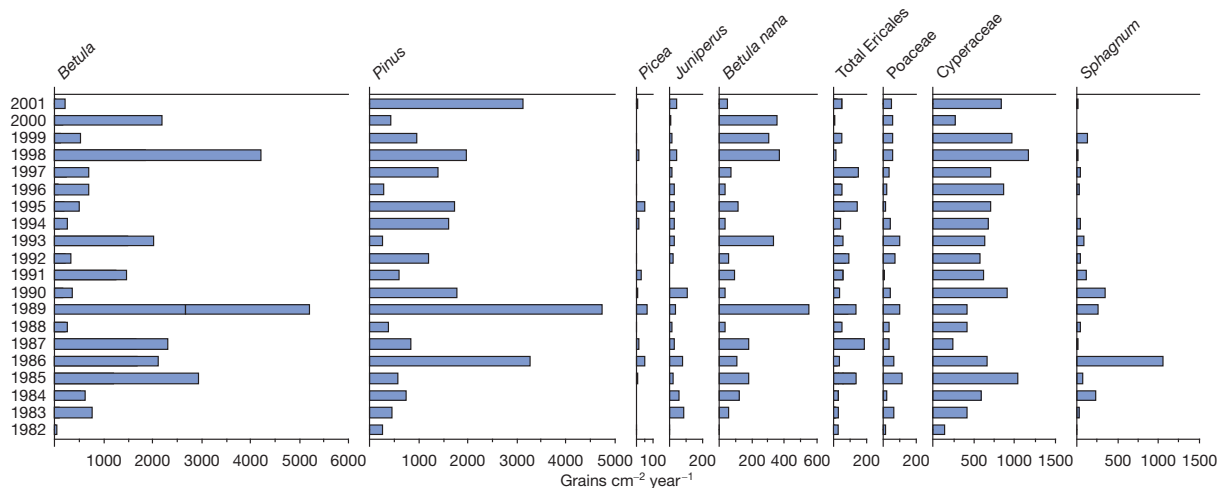


Figure 5 Yearly PARs for the period 1982–2001 monitored at a site in northern Finland located close to the northern forest limit of Scots pine (*Pinus sylvestris*).

Table 2 Threshold PAR values for the presence and abundance of three tree species common in northernmost Fennoscandia based on the long-term average results of pollen monitoring

	Betula (<i>Birch</i>)	Pinus (<i>Pine</i>)	Picea (<i>Spruce</i>)
Not present within 10 km		<300	<15
Not present within 1 km	<600	300–500	15–25
Present but only sparsely	600–1000	500–1300	25–140
Open forest	1000–1600	1300–2500	140–200
Dense forest	>1600	>2500	>200

Note: The numerical value is affected by the size of opening in the forest in which the pollen trap is located. Such threshold values should, therefore, only be applied to fossil situations where the opening was of similar size.

Source: Reproduced from Table 1 of Hicks S and Sunnari A (2005) Adding precision to the spatial factor of vegetation reconstructed from pollen assemblages. *Plant Biosystems* 139(2): 127–134, with permission.

- using parameters derived from the modern reference samples as input for models of pollen dispersal.

The requirements and restrictions on the number and location of surface samples or pollen traps vary in accordance with these three major research methods and are summarized in Table 1.

Unlike surface samples, the whole set of which can be collected during one field season, pollen traps are a long-term commitment, and so their location has to be most carefully planned. Of particular significance is the size of the opening in which the trap is located, because this directly affects the numerical value of the PARs calculated for the individual pollen taxa. Forest openings (in effect the area which is not covered by trees) should be of a similar size if PARs are to be compared between traps. Traps placed within forest will record predominantly local pollen deposition which can be extremely abundant, while those placed in large mires, or even on the water surface of large lakes (>1 km in diameter), will collect relatively smaller quantities of pollen, but the pollen composition will reflect an average for a large regional area.

In the majority of cases, the surface sample or pollen trap data set will be of only limited use unless it is accompanied by information on the vegetation in the area where it is collected. The desirable degree of detail in the accompanying vegetation data, however, varies with the research question being addressed (Table 1). In situations where the sample covers a long time interval, it is conceivable that during the time that the pollen assemblage has accumulated the surrounding vegetation has changed, for example, a 10-year lake-surface sediment sample in an area formerly cultivated and later abandoned, or a pollen trap located in a forest area, which during the time in question is partially felled. In such situations, the surface sample will give a mismatch for the vegetation it is collected to represent. In the case of the pollen trap, however, where data are collected every year, it is possible to follow the changes in PAR brought about by the partial felling of forest, forest fire, or year-to-year variability in climate.

It is possible to archive the results of surface sample and pollen trap analyses in the form of databases. The European Pollen Database and Global Pollen Database (see Databases and Their Application) hold records of fossil pollen sequences and modern (surface) samples, although not all parts of the world are represented. The Neotoma Database includes records of fossil pollen sequences and modern (surface) samples, as well as other paleoecological data. The PMP maintains a

database of the year-by-year records from pollen traps, with data supplied by its members. The completeness of accompanying metadata (latitude, longitude, altitude, sampling medium, date of collection, name of collector, local and regional vegetation, slope and aspect) varies considerably, and so samples should be carefully selected from those in the database so as to fulfill the requirements of the research question to be addressed. Nevertheless, these databases form a valuable publically available source of information for a wide range of different vegetation situations in diverse geographical areas.

Data Applications

Vegetation History

Past vegetation can be reconstructed from surface samples in an objective manner using the modern analog whole assemblage comparative approach. The data set for this should contain samples from the whole range of vegetation communities that are thought to have existed in the past at the location of the fossil profile. A reconstruction will not be possible if the fossil pollen assemblage does not have a modern analog. Numerical ordination or classification techniques (see Numerical Analysis Methods) applied to the surface sample data set are used to separate and characterize different plant communities, which can then be identified in the fossil record (Gaillard et al., 1992). The surface and fossil pollen assemblages can also be combined and analyzed numerically and the fossil samples interpreted according to their position (in, e.g., a principal components analysis plot) relative to the surface samples (Gaillard et al., 1994; Hjelle, 1999a).

In an alternative approach, PARs calculated from pollen trapping can be used to provide threshold values for the closeness and abundance of different species relative to the monitoring site (Table 2). When applied to fossil situations, these give a much more objective picture of past treelines (see Tree-line Studies) and the migration of species (see Changing Plant Distributions and Abundances) than can be obtained from methods which have to rely on pollen percentage values (Hicks, 2001; Seppä and Hicks, 2006).

Past Land Use and Anthropogenic Disturbance to the Vegetation

A simple and frequently used method of identifying and following human interference with the vegetation

(see [Archaeological Applications](#)) is the modern analog indicator species approach. Surface samples for this method are deliberately collected from sites exhibiting the sort of land-use situation that is expected to have existed in the past and visually inspected to pinpoint those species that are uniquely associated with each type of land use. These species then become markers or indicators of that land-use type in the fossil record (Behre, 1981, 1986; Hjelle, 1997, 1999b). As with all modern analog approaches, the modern data set may be lacking in representatives of all past land-use categories. Moreover, the method assumes that the tolerance of the indicator species to the relevant environmental variables has not changed over time. The whole assemblage comparative approach mentioned above can also be applied to reconstruct past land use.

Climate Reconstructions

One common use of surface samples is in climate reconstructions, either at the broad regional scale or at the coarser global scale. A common practice is to use transfer functions in which the pollen data are translated into a climate parameter (see [Approaches, Use of Pollen as Climate Proxies](#), and [Mire and Peat Macros](#)). For this approach, the modern training set should cover the full range of variability in the climate parameter being reconstructed, which usually means obtaining a data set with a wide geographical spread and a dense spatial network (Birks and Seppä, 2004; Birks et al., 2010).

Results from pollen trapping in temperate areas are beginning to reveal that the annual variation in pollen production calibrates with temperature of the summer in the year before pollen emission (Autio and Hicks, 2004). There is the possibility, therefore, that if continuous PAR records at an annual resolution can be obtained from sediment profiles, these would provide a proxy record of summer temperature variation through time.

Models of Pollen Dispersal and Deposition

Models of pollen dispersal rely on surface sample results in order to calculate pollen productivity estimates and establish the composition of the background pollen within a given region (Sugita, 1994; Sugita et al., 1999). Moss polsters or samples of surface sediments of lakes, collected from random locations of investigation area, are most commonly used for pollen productivity estimates (Broström et al., 2008). For estimating background pollen composition, samples must be taken from the centers of large lakes (of a size larger than the relative source area of pollen for the region), because it is the consistent regional component of the pollen assemblage that is required (Hellman et al., 2008). Surface samples or data collected by means of pollen traps can be used to evaluate vegetation reconstruction model performance (Sugita et al., 2010).

See also: Pollen Analysis, Principles; Varved Lake Sediments. **Paleobotany:** Charred Particle Analyses. **Paleoclimate Reconstruction:** Approaches. **Plant Macrofossil Methods and Studies:** Mire and Peat Macros; Surface Samples, Taphonomy, Representation; Treeline Studies. **Pollen Methods and Studies:**

Archaeological Applications; Changing Plant Distributions and Abundances; Databases and Their Application; Numerical Analysis Methods; POLLSCAPE Model: Simulation Approach for Pollen Representation of Vegetation and Land Cover; Use of Pollen as Climate Proxies.

References

- Aario L (1943) Über die wald- und klimaentwicklung an der Lappländischen Eismeerküste in Petsamo. *Annales Botanici Societatis Zoologica Botanicae Fennicae Vanamo* 19: 1–158.
- Autio J and Hicks S (2004) Annual variations in pollen deposition and meteorological conditions on the fell Aakenustunturi in northern Finland: Potential for using fossil pollen as a climate proxy. *Grana* 43: 31–47.
- Behre K-E (1981) The interpretation of anthropogenic indicators in pollen diagrams. *Pollen et Spores* 23: 225–245.
- Behre K-E (1986) *Anthropogenic Indicators in Pollen Diagrams*. Rotterdam: A.A. Balkema.
- Birks HJB, Heiri O, Seppä H, and Bjune AE (2010) Strengths and weaknesses of quantitative climate reconstructions based on late-quaternary biological proxies. *The Open Ecology Journal* 3: 68–110.
- Birks HJB and Seppä H (2004) Pollen-based reconstructions of late-quaternary climate in Europe – Progress, problems, and pitfalls. *Acta Palaeobotanica* 44: 317–334.
- Broström A, Nielsen A, Gaillard M-J, et al. (2008) Pollen productivity estimates of key European plant taxa for quantitative reconstruction of past vegetation: A review. *Vegetation History and Archaeobotany* 17: 461–478.
- Cour P (1974) Nouvelle technique de détection des fluxes et retombées polliniques: étude de la sédimentation des pollens et des spores à la surface du sol. *Pollen et Spores* 16: 103–141.
- Cundill PR (1986) A new design of pollen trap for modern pollen studies. *Journal of Biogeography* 13: 83–98.
- Cundill PR (1991) Comparisons of moss polster and pollen trap data: A pilot study. *Grana* 30: 301–308.
- Davis MB (1973) Redeposition of pollen grains in lake sediment. *Limnology and Oceanography* 18: 44–52.
- Davis MB, Brubaker LB, and Beiswenger JM (1971) Pollen grains in lake sediments: Pollen percentages in surface sediments from Southern Michigan. *Quaternary Research* 1: 450–467.
- Gaillard M-J, Birks HJB, Emanuelsson U, and Berglund BE (1992) Modern pollen/land-use relationships as an aid in the reconstruction of past landuses. *Vegetation History and Archaeobotany* 1: 3–17.
- Gaillard M-J, Birks HJB, Emanuelsson U, Karlsson S, and Lagerås P (1994) Application of modern pollen/land-use relationships to the interpretation of pollen. *Review of Palaeobotany and Palynology* 82: 47–73.
- Giesecke T and Fontana S (2008) Revisiting pollen accumulation rate from Swedish lake sediments. *The Holocene* 18: 293–305.
- Giesecke T, Fontana S, van der Knaap WO, Pidek IA, and Pardoe H (2010) From early pollen trapping experiments to the Pollen Monitoring Programme. *Vegetation History and Archaeobotany* 19: 247–258.
- Glew JR, Smol JP, and Last WM (2001) Sediment core collection and extrusion. In: Last WM and Smol JP (eds.) *Tracking Environmental Change Using Lake Sediments*, vol. 1, pp. 73–105. Dordrecht: Kluwer Academic Publishers.
- Gosling WD, Mayle FE, Killeen TJ, Siles M, Sanchez L, and Boreham S (2003) A simple and effective methodology for sampling modern pollen rain in tropical environments. *The Holocene* 13: 613–618.
- Hall SA (1990) Pollen deposition and vegetation in the southern Rocky Mountains and southwest Plains, USA. *Grana* 29(1): 47–61.
- Hellman S, Gaillard M-J, Broström A, and Sugita S (2008) The REVEALS model, a new tool to estimate past regional plant abundance from pollen data in large lakes: Validation in southern Sweden. *Journal of Quaternary Science* 23(1): 21–42.
- Hicks S (2001) The use of annual arboreal pollen deposition values for delimiting tree-lines in the landscape and exploring models of pollen dispersal. *Review of Palaeobotany and Palynology* 117: 1–29.
- Hicks S, Bozilova E, Dambach K, Drescher-Schneider R, and Latalowa M (1998) Sampling methodologies for the collection of modern pollen data and related vegetation and environment. *Paläoklimaforschung* 27: 141–144.
- Hicks S and Hyvärinen VP (1986) Sampling modern pollen deposition by means of 'Tauber traps': Some considerations. *Pollen et Spores* 28: 219–242.
- Hicks S and Hyvärinen H (1999) Pollen influx values measured in different sedimentary environments and their palaeoecological implications. *Grana* 38: 228–242.

- Hicks S and Isaksson E (2005) Assessing source areas of pollutants from studies of fly-ash, charcoal and pollen from Svalbard snow and ice. *Journal of Geophysical Research* 111: 1–9.
- Hicks S, Tinsley H, and Huusko A (2001) Some comments on spatial variation in arboreal pollen deposition: First records from the Pollen Monitoring Programme (PMP). *Review of Palaeobotany and Palynology* 117: 183–194.
- Hirst JM (1952) An automatic volumetric spore trap. *Annals of Applied Biology* 39: 257–265.
- Hjelle KL (1997) Relationships between pollen and plants in human-influenced vegetation types. *Review of Palaeobotany and Palynology* 99: 1–16.
- Hjelle KL (1999a) Use of modern pollen samples and estimated pollen representation factors as aids in the interpretation of cultural activity in local pollen diagrams. *Norwegian Archaeological Review* 32: 19–39.
- Hjelle KL (1999b) Modern pollen assemblages from mown and grazed vegetation types in western Norway. *Review of Palaeobotany and Palynology* 107: 55–81.
- Hjelmroos M and Franzen L (1994) Implications of recent long-distance pollen transport events on the interpretation of fossil pollen records in Fennoscandia. *Review of Palaeobotany and Palynology* 82: 175–189.
- Kansanen PH, Jaakkola T, Kulmala S, and Suutarinen R (1991) Sedimentation and distribution of gamma-emitting radionuclides in bottom sediments of southern Lake Päijänne, Finland, after the Chernobyl accident. *Hydrobiologia* 222: 121–140.
- Kelly D (1994) The evolutionary ecology of mast seeding. *Trends in Ecology & Evolution* 9: 465–470.
- Koenig WD and Knops JMH (1998) Scale of mast-seeding and tree-ring growth. *Nature* 396: 225–226.
- Koff T (2001) Pollen influx into Tauber traps in Estonia in 1997–1998. *Review of Palaeobotany and Palynology* 117(1–3): 53–62.
- Levetin E, Rogers CA, and Hall SA (2000) Comparison of pollen sampling with a Burkard Spore Trap and a Tauber Trap in a warm temperate climate. *Grana* 39: 294–302.
- Maier LJ (1981) Statistics for microfossil concentration measurements employing samples spiked with marker grains. *Review of Palaeobotany and Palynology* 32: 153–191.
- Pardoe HS, Giesecke T, van der Knaap WO, et al. (2010) Comparing pollen spectra from modified Tauber traps and moss samples: Examples from a selection of woodlands across Europe. *Vegetation History and Archaeobotany* 19: 271–283.
- Prentice IC (1978) Modern pollen spectra from lake sediments in Finland and Finnmark, north Norway. *Boreas* 7: 131–153.
- Ranta H and Satri P (2007) Synchronized inter-annual fluctuation of flowering intensity affects the exposure to allergenic tree pollen in North Europe. *Grana* 46: 274–284.
- Räsänen S, Hicks S, and Odgaard BV (2004) Pollen deposition in mosses and in a modified 'Tauber trap' from Hailuoto, Finland: What exactly do the mosses record?. *Review of Palaeobotany and Palynology* 129: 103–116.
- Seppä H and Hicks S (2006) Integration of modern and past pollen accumulation rate (PAR) records across the arctic tree-line: A method for more precise vegetation reconstructions. *Quaternary Science Reviews* 25: 1501–1516.
- Solomon AM and Silkworth AB (1986) Spatial patterns of atmospheric pollen transport in a montane region. *Quaternary Research* 25(2): 150–162.
- Stockmarr J (1971) Tablets with spores used in absolute pollen analysis. *Pollen et Spores* 13: 616–622.
- Sugita S (1994) Pollen representation of vegetation in Quaternary sediments: Theory and method in patchy vegetation. *Journal of Ecology* 82: 881–897.
- Sugita S, Gaillard MJ, and Broström A (1999) Landscape openness and pollen records: A simulation approach. *The Holocene* 9: 409–421.
- Sugita S, Tarshall T, Calcote R, and Walker K (2010) Testing the Landscape Reconstruction Algorithm for spatially explicit reconstruction of vegetation in northern Michigan and Wisconsin. *Quaternary Research* 74: 289–300.
- Tauber H (1974) A static non-overload pollen collector. *New Phytologist* 73: 359–369.
- Thompson R (1980) Use of the word influx in paleolimnological studies. *Quaternary Research* 14: 269–270.
- Tonkov S, Hicks S, Bozilova E, and Atanassova J (2001) Pollen monitoring in the central Rila Mountains, Southwestern Bulgaria: Comparisons between pollen traps and surface samples for the period 1993–1999. *Review of Palaeobotany and Palynology* 117(1–3): 167–182.
- van der Knaap WO, van Leeuwen JFN, and Ammann B (2001) Seven years of annual pollen influx at the forest limit in the Swiss Alps studied by pollen traps: Relations to vegetation and climate. *Review of Palaeobotany and Palynology* 117(1–3): 31–52.

Relevant Website

<http://www.pollentrapping.net/> – Pollen Monitoring Program.

Stand-Scale Palynology

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Introduction

Conventional pollen analysis records vegetational change at spatial and temporal scales that are largely unfamiliar to the plant ecologist. Forest succession, or the predictable replacement of one group of plants by another, is usually observed at the scale of the woodland stand (10^2 – 10^3 m²) during the lifetime of the ecologist, but pollen analysis of lakes and bogs reconstructs vegetation cover from far larger areas (10^4 – 10^6 m²), typically over time periods of thousands of years. Forest hollows and other ‘closed-canopy’ sites, by contrast, are dominated by pollen and spores that have traveled only a few meters from source plants and thus resolve vegetation dynamics at the scale of the woodland stand. Vegetation can be reconstructed with very high spatial resolution that can be easily compared with plot-based surveys of modern vegetation. Thus, stand-scale palynology provides a key link between paleobotanical and contemporary ecological studies. Ecological processes and properties that can be more easily studied using stand-scale rather than conventional palynology include the following: (1) the effects of disturbance on vegetation, either events such as fire, disease, or storm damage, or chronic disturbance such as animal browsing; (2) species invasions, succession and its relationship to climatic change, and anthropogenic impact; and (3) vegetation structure and openness. Stand-scale palynology is well suited to complement archaeological investigations where the study sites are often small and spatially well defined.

History

Stand-scale studies began in Scandinavia, as did many methodological developments in palynology. The approach was adopted and developed in the British Isles and former colonies and then the initiative passed to North American scientists where many of the recent developments have taken place. Iversen (1964) in Denmark was the first to fully appreciate the benefits of local-scale palynological studies even though isolated earlier studies had taken place in Norway and Sweden. Iversen realized the potential of stand-scale studies to yield specific insights into dynamic ecological processes and in 1949 established a field laboratory within the 200 ha Draved Forest, Denmark, with the aim of developing the subject. Pollen–vegetation relationships, vegetation monitoring, and pollen analysis from 30 stand-scale sites were among the ongoing studies before these unique observations and experiments were terminated due to lack of funding in 2003.

Special mention must be given to the related topic of soil palynology, which was largely developed by Dumbleby (1957, 1985). Soil palynology has developed more within archaeological than Quaternary science, yet shares some of the special features of stand-scale palynology of wetland and mor humus sites. Andersen (1970) was a landmark paper in development

of the relationship between source vegetation and stand-scale pollen deposition and the detailed history of Løvenholm Forest was a showcase for the power of the technique (Andersen, 1984). Iversen used mor humus profiles to describe the development of Draved Forest over recent millennia, while Andersen sampled forest hollows from Løvenholm Forest that covered most of the Holocene. These forests today are important reference areas for the study of undisturbed forest processes, yet the pollen studies showed how most of the forest area had been significantly altered by human activity beginning thousands of years ago. Such studies have improved our detailed understanding of the past relationships between people and vegetation in Northwest Europe.

Andersen also demonstrated a strong empirical relationship between pollen deposition at a single point on the forest floor and the present vegetation within a 20–30 m radius. This finding was supported by Bradshaw (1981) but was later modified by thorough field investigations in the 8500 ha Sylvania Wilderness, Upper Michigan, USA (Calcote, 1995; Sugita, 1994). A review of the subject was made by Bradshaw (1988), and later, Mitchell (1998) covered some further developments and showed how stand-scale studies contributed to discussion about the openness of Northwest European forest structure (Mitchell, 2005). Syntheses of case studies involving suites of stand-scale sites are the best illustration of how stand-scale palynology provides an important interface to issues within dynamic ecology (Bradshaw and Lindbladh, 2005; Bradshaw et al., 2005; Davis et al., 1998). Although forest peats are found in the tropics and subtropics, almost all of the published stand-scale sites are from glaciated regions within boreal and temperate forest areas.

Pollen Representation Theory

Andersen (1970) adopted a largely empirical approach to pollen representation at the stand-scale and developed correction factors to relate pollen counts more closely to source vegetations that have been widely adopted. He observed that within a forest, pollen seemed to land closer to its source than was predicted by physical laws describing the dispersal of small particles. He ascribed this observation to special properties of turbulent air movement under the forest canopy. Prentice (1985) developed a pollen representation theory that accounted for the special properties of stand-scale sites, and his approach laid the foundation for later models that consider pollen production and dispersal issues and allow quantitative landscape reconstruction. The POLLSCAPE model assumes that all pollen disperses through the air as single grains and this assumption is often violated at the stand scale. POLLSCAPE predicts that stand-scale sites have a somewhat larger source area for pollen (>50 m radius) than that observed by Andersen. The differences are not great, however, and there is a general consensus that stand-scale sites sample vegetation at a different scale compared

to conventional sites, which are usually chosen so as to yield a regional record. Stand-scale sites are much smaller than conventional sites and ideally are located under a closed tree canopy. Plant macrofossils, insect remains, and charcoal can also be preserved in stand-scale sites, and the basis for their interpretation is increasing as more sites are investigated. Stand-scale sites are best suited for detecting severe, infrequent fires using charcoal fragments larger than 150 μm (Higuera et al., 2005).

Types of Sites

Pollen is preserved in a variety of settings beneath forest canopies. If the forest canopy is temporarily or permanently removed, a considerably increased proportion of the pollen preserved at any one point derives from vegetation so far from that point that stand-scale resolution is effectively lost (Jacobson and Bradshaw, 1981). Suitable sample sites fall into three broad categories: small, wet, forest hollows (Figure 1); accumulations of mor humus; and soils of low biological activity (Table 1). The sediment in forest hollows may be mud or peat, and the pollen is preserved by waterlogging. Sedimentation can be continuous over thousands of years, but hiatuses are rather common, and the sites are sensitive to fluctuating water tables. Mor humus is a dry, acidic accumulation of partially decomposed plant litter of very low biological activity. This humus type is typically associated with acidic, free-draining soils such as podzols and is found under trees producing a leaf litter of high polyphenolic content. Little or



Figure 1 Small forest hollow in Suserup Skov, Denmark. Photograph by Gina Hannon.

no mixing occurs, and pollen is prevented from downwashing by becoming trapped in aggregates of humic material. Downwashing and mixing does tend to occur in soils, but where pollen concentrations are high, information about past vegetation can be extracted, for example, in soil horizons buried under archaeological monuments (Andersen, 1997).

Each type of site has its own assets and problems for the palynologist. The major asset of all these sites is their small pollen source area and hence their good spatial resolution. They can provide information about taxa that produce sparse amounts of poorly dispersed pollen such as *Arbutus unedo* (strawberry tree), *Ilex aquifolium* (European holly), and *Tilia cordata* (small-leaved lime), whose past history is hard to determine from conventional pollen sites. Frequently, high temporal resolution can be attained to match the spatial resolution, especially in mor humus.

Pollen grains from closed-canopy sites are more exposed to oxidation and microbial action than in conventional sites, posing potential preservation problems. Soil pollen samples tend to be the worst preserved, and large numbers of deteriorated grains or high values of resistant types, such as *Polypodium* (polypody fern) spores, indicate a probable preservation bias. Preservation can be excellent, and several forest hollows with plant macrofossil records have been reported (Figure 2;



Figure 2 Acorn cup of *Quercus petraea* (sessile oak) about 6000 years old recovered from a forest hollow in Suserup Skov, Denmark. Photograph by Peter Wærn-Moors.

Table 1 Properties of stand-scale, closed-canopy sites

	Small hollows	Mor humus	Soils
Length of record (years)	10 ³	10 ² –10 ³	10 ¹
Continuity of record	May be patchy	Good	Poor
Temporal resolution (years/sample)	10–100	1–10	–
Mixing or downwash of pollen	Minimal	Minimal	Yes
Pollen preservation	Usually good	Good	Poor
Macrofossils	Occasionally present	No	No
Charcoal	Present	Usually present	Usually present
Site availability	Restricted	Restricted	Widespread

Hannon et al., 2000). Macrofossils are very rare in humus or soil samples, although charcoal fragments can be found in all site types.

Insights from Case Studies

Sylvania, Upper Michigan, USA

Davis et al. (1998) used pollen analysis of forest hollows to investigate the late Holocene invasion of mixed deciduous stands by *Tsuga canadensis* (eastern hemlock) in Sylvania, Upper Michigan, USA. They asked whether the timing of the invasion was simultaneous throughout the study area and whether the invasibility of the area was affected by species composition. These are questions of importance to biologists that cannot be addressed using traditional palynological techniques. They compiled a dataset comprising ten sediment cores collected along a 10 km transect. Sediment accumulation began at different times from the Late Glacial until only 1200 years ago, but once initiated, apparently continued until the present day. They believed this was attributable to a continuously rising water table, but it does illustrate the special sensitivity of forest hollow sites to fluctuating water tables. Sediment accumulation rates were typically 1–3 cm per century with high pollen concentrations and generally good preservation (Davis et al., 1998).

Windstorms were recorded in the sediments by wood layers and sharp increases in pollen abundance of *Betula* (birch), but the invasion of eastern hemlock was not facilitated by disturbance. The vegetation has retained a patchy structure throughout the Holocene, and *Tsuga* preferentially invaded stands previously dominated by *Pinus strobus* (eastern white pine) around 3000 years ago (Figure 3 H1–H4). The invasion process took a few hundred years. Once *Tsuga* had invaded a stand, it maintained its dominance until today with a gradually increasing abundance, probably reflecting the regional increase in moisture.

Stands that today are dominated by *Acer saccharum* (sugar maple) and *Tilia grandifolia* (large-leaved lime) had multiple origins (Figure 3 M1–M4). Some were converted from *Tsuga* stands following disturbance, while others gradually lost *Quercus* (oak) during long periods of time. The spatial and numerical precision of these vegetation reconstructions is striking and quite distinct from interpretations that can be made from conventional, regional palynological studies. The 3000-year stability of most of the *Tsuga* stands is in contrast to the greater dynamics observed in the mixed hardwood stands and gives new insight into the development of patchiness in vegetation.

Southern Scandinavia

Several forest hollows have been investigated in southern Sweden and Denmark that permit analyses and insights into vegetation dynamics and drivers at a range of spatial scales. Maps of southern Sweden have been prepared that compare stand-scale composition with regional forest composition interpreted from conventional lake and peat sites (Figure 4; Lindbladh and Foster, 2010; Lindbladh et al., 2000). Deciduous forest types dominated the central and western landscapes at 1250 BC (calendar years), while *Pinus* (pine) was abundant

toward the east. Deciduous taxa such as *Alnus* (alder), birch, *Corylus* (hazel), oak, and lime were all common, but no single taxon was a clear regional dominant. *Fraxinus excelsior* (European ash) and *Ulmus* (elm) were less abundant than the other major tree taxa. The stand-scale records from 1250 BC confirm the dominant role of deciduous trees with only a single site recording abundant *Pinus*. Single taxa could dominate individual stands, but although one taxon comprised more than 50% of the basal area in four stands, it was a different taxon (alder, birch, pine, and oak) in each case. A kaleidoscope of forest stands was probably found in a diverse, deciduous forest landscape.

By AD 500 European beech had become established at the landscape scale, particularly in the southeast and western regions (Figure 4). *Pinus sylvestris* (Scots pine) maintained its importance in the east. Norway spruce was becoming a component of the northern forests, and increased importance of *Betula* and *Juniperus communis* (common juniper) in the central part of the region indicated widespread disturbance of forest cover. At the stand scale, the pie diagrams show an increasing importance of birch and pine at the expense of alder and lime. Stand-scale dominance patterns could differ from the regional abundance. For example, beech dominated a stand in the northwest in a region that only supported 2% beech, although beech was more abundant in the regional vegetation to the north.

The maps from the present forest landscape show the widespread dominance of conifers, chiefly spruce but with abundant pine, particularly in the east and southern parts of the region (Figure 4). The present-day map shows a good general agreement with recent forest inventory data, although there is rather little spruce recorded from southern parts of the boreal-temperate forest ecotone. Spruce is close to its natural southern limits here, and there has been much planting during the present century. Young, dense plantations produce little pollen in relation to forest volume, and there is also a possible temporal mismatch between the pollen and tree data. The pollen data are from bioturbated sediments and can integrate pollen input spanning 50–100 years, especially in the soft, unconsolidated surface sediments that are difficult to sample with accuracy. Birch and beech are the commonest deciduous taxa, and this pattern is also observed in the stand-scale records. Spruce, birch, and pine, the three major boreal taxa, comprise over 70% of the forest volume in 9 out of 13 stands. Three of the remaining stands are northern beech outliers, and one record is from a recently clear-felled area. Both the regional and local records show the same general tendency of declining abundance of deciduous taxa (except birch and beech) and an increasing tendency for one or two taxa to dominate both stands and landscapes. The northern beech-dominated stands, however, differ from the surrounding regional composition showing that forest hollows can detect stands that are unrepresentative of the landscape. Nevertheless, most stands broadly reflect the landscape-scale shifts in composition during the last 3250 years, although individual taxa show greater ranges of abundance in the stand-scale records.

Stand-scale compositional trends

The detrended correspondence analysis (DCA) of the calibrated pollen data from the individual forest stands for 1250 BC, AD 500, and the present, portrays the similarities

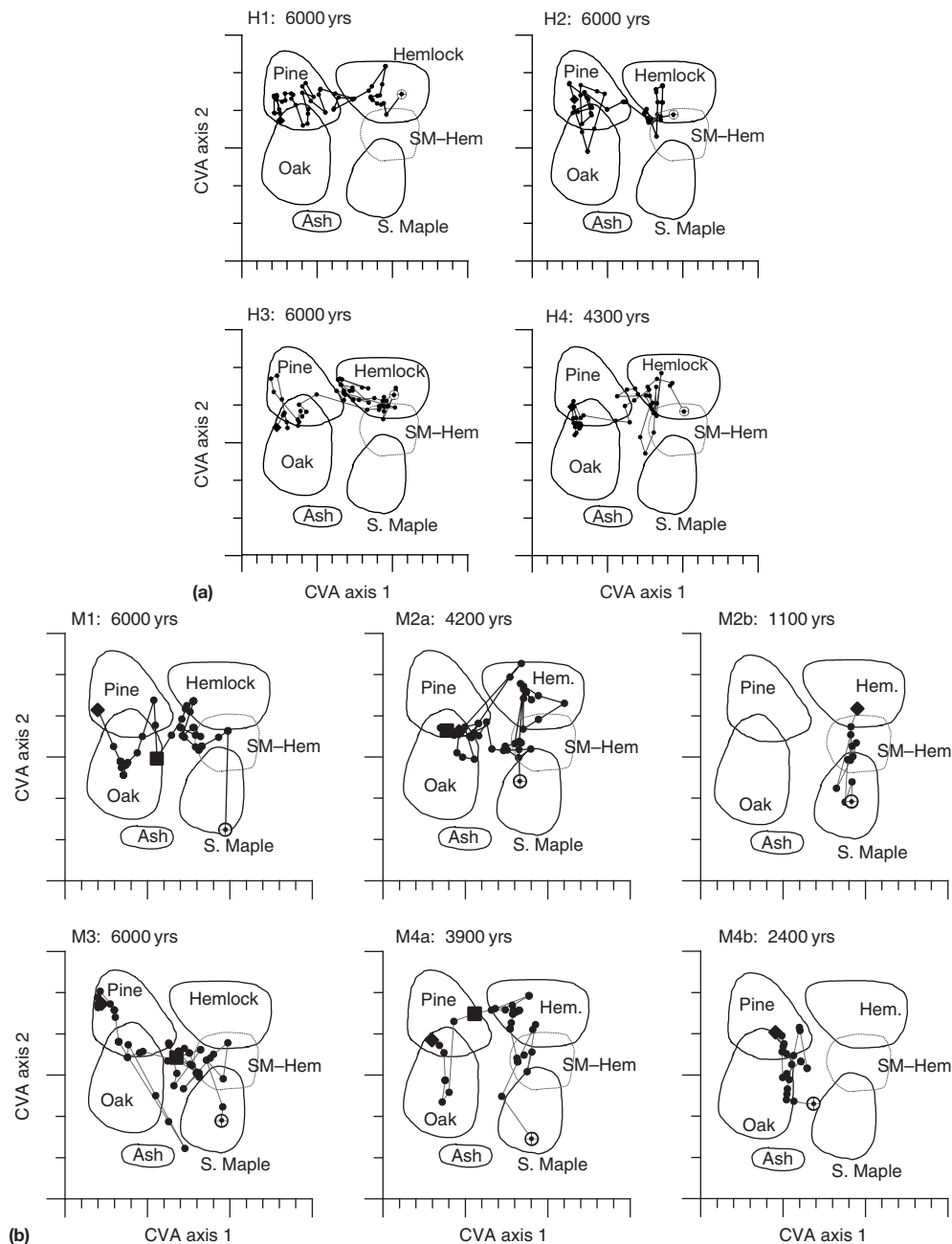


Figure 3 Trajectories through time of pollen assemblages (a) in *Tsuga* hollows and (b) in hardwood hollows projected into canonical variate analysis (CVA) space defined by 66 surface samples. Dots show the position of the trajectories interpolated to 100-year intervals. The oldest sample in each series is shown by a large diamond and the youngest by a circle with a crosshair. Large squares denote 3000 calendar years ago. The outlined areas include surface pollen assemblages from six major forest communities. Reproduced from Davis MB, Calcote RR, Sugita S, and Takahara H (1998) Patchy invasion and the origin of a hemlock-hardwoods forest mosaic. *Ecology* 79: 2641–2659.

and differences between the 16 individual stand histories (Figure 5). All stands move away from the rich deciduous forest indicated by pollen from alder, hazel, oak, and lime with some ash and elm. The commonest successional pathway is from birch and *Carpinus betulus* (European hornbeam) to forest comprising spruce and pine, although two stands became dominated by beech. The role of birch as a prominent taxon in forests undergoing transition from deciduous to coniferous was emphasized in the DCA, and many stands were at

this intermediate stage at AD 500. The same tendency was observed at the regional scale especially in the central region of the study area (Figure 4).

The DCA shows that unique successional pathways can occur and be resolved by stand-scale palynology, but the general pattern in southern Sweden during the last 3250 years has been along the pathways that lead to spruce – and to a lesser extent to beech-dominated stands. The spreading rate of spruce and beech into southern Scandinavia can be estimated from

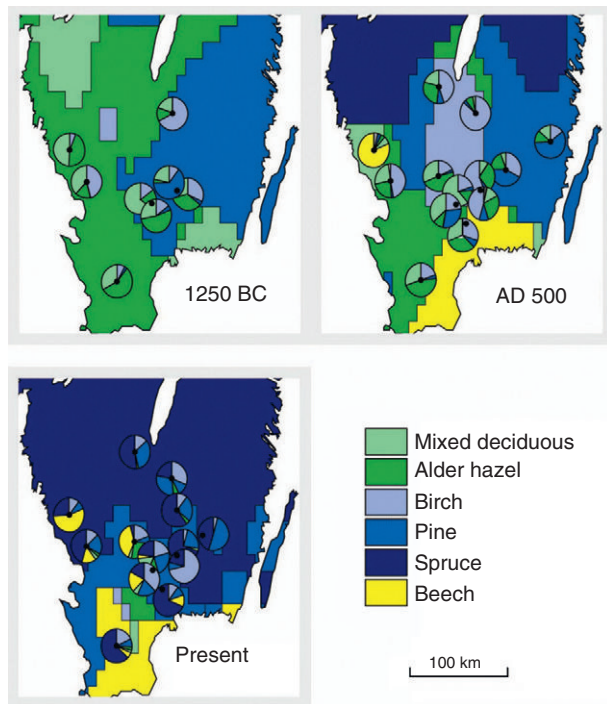


Figure 4 Forest maps of southern Sweden from 1250 BC, AD 500, and the present based on regional pollen sites. The pie diagrams show data from stand-scale sites. Reproduced from Lindbladh M, Bradshaw RHW, and Holmqvist BH (2000) Pattern and process in south Swedish forests during the last 3000 years, sensed at stand and regional scales. *Journal of Ecology* 88: 113–128.

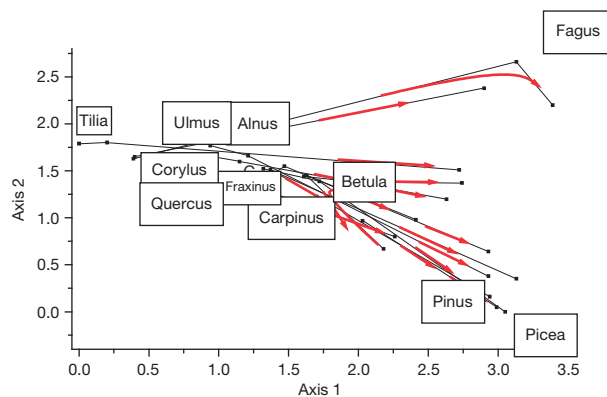


Figure 5 Detrended correspondence analysis (DCA) of 13 stand-scale sites from southern Scandinavia. Nine of the sites are represented by three time points (1250 BC, AD 500, and the present). Four sites cover just AD 500 to the present. The arrows show the direction in time. Reproduced from Lindbladh M, Bradshaw RHW, and Holmqvist BH (2000) Pattern and process in south Swedish forests during the last 3000 years, sensed at stand and regional scales. *Journal of Ecology* 88: 113–128.

the arrival times at the individual stands sampled by small hollows. Data–model comparison of species spread showed that the late Holocene spread of spruce was rather rapid (250 m year^{-1}) throughout the last 4000 years and was only limited by biological processes of dispersal, despite significant

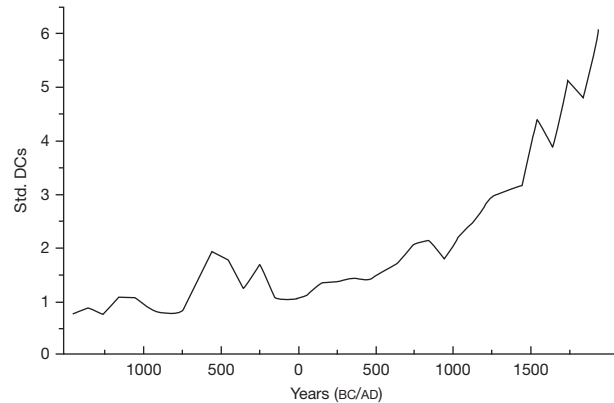


Figure 6 The mean values for rate-of-change analysis for 13 stand-scale sites in southern Scandinavia. The analysis included the calibrated tree pollen data, shrub, and herbaceous taxa that exceeded 2% of the pollen sum in at least one sample. Std. DCs – chord distance per 100 years. Reproduced from Lindbladh M, Bradshaw RHW, and Holmqvist BH (2000) Pattern and process in south Swedish forests during the last 3000 years, sensed at stand and regional scales. *Journal of Ecology* 88: 113–128.

changes in climate. In contrast, the spread of beech was most likely controlled by site properties (Bialozyt et al., 2012).

Rate of vegetational change

Analysis of the rate of vegetational change indicates those time periods when rapid reorganization of forest composition occurred as well as periods of relative stability (Jacobson et al., 1987). Each stand-scale site in the study of records from Scandinavia has its own unique pattern, but certain common features are apparent when all stands are combined into a mean value (Figure 6; Lindbladh et al., 2000). A major increase in rate of change began around AD 1000 and has continued until the present day, with the last 1000 years showing an increase four times greater than that recorded during the preceding 2500 years. These recent changes thus completely overshadow the earlier record. The changes recorded from the last 150 years were the most rapid but represent the culmination of a transformation that was initiated 850 years earlier. Lindbladh and Foster (2010) combined data from 25 small hollows with six regional sites and confirmed the recent increased rate of change in south Scandinavian vegetation. They showed that while south Scandinavian oak populations had been in continuous decline for at least 2000 years, the majority of this decline took place during the last 300 years, corroborating historical records.

Anthropogenic or climatic control of recent dynamics of *T. cordata* and *Fagus sylvatica* in Denmark and southern Sweden?

Much debate focuses on the origins of beech forests in Northern Europe, primarily owing to their current dominance of large areas and their often presumed ‘natural’ status. At the continental scale, European beech is arguably in equilibrium with climate and thus is an expected component of modern southern Scandinavian forests. Paleoeological investigation of sediments in forest hollows in southern Sweden, however, indicates that present-day beech stands often show signs of

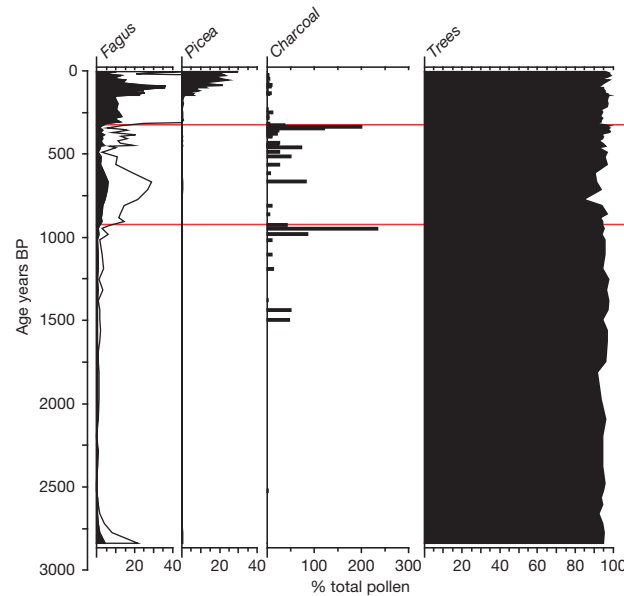


Figure 7 History of beech establishment and its relationship to large charcoal fragments recorded from a forest hollow at Siggaboda, southeast Sweden. Reproduced from Björkman L and Bradshaw RHW (1996). The immigration of *Fagus sylvatica* L. and *Picea abies* (L.) Karst into a natural forest stand in southern Sweden during the last 2000 years. *Journal of Biogeography* 23: 235–244.

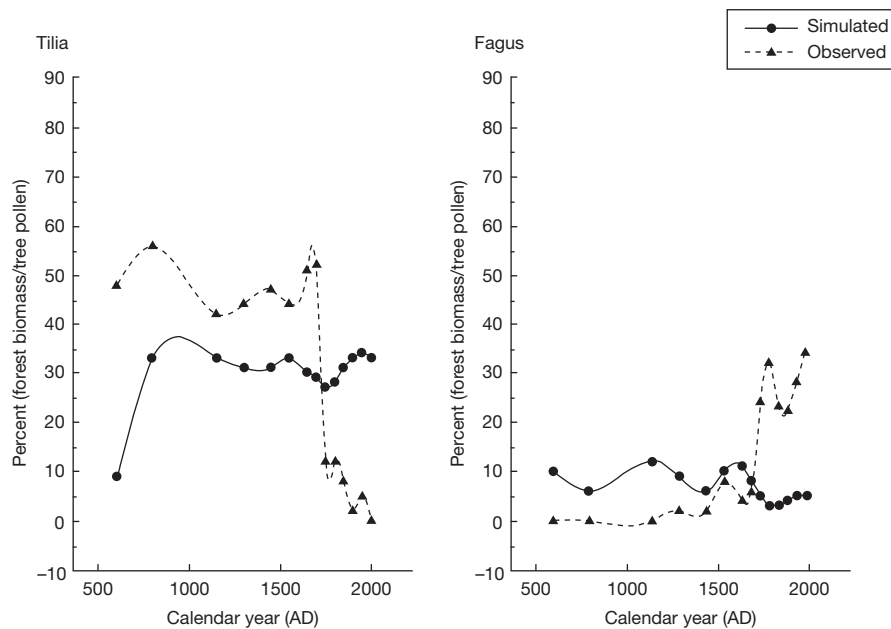


Figure 8 Data model comparisons for lime (*Tilia*) and beech (*Fagus*) in Draved Forest, Denmark. Simulated data are percentage of total forest biomass, and observed data are calibrated pollen percentages from a mor humus profile. Reproduced from Cowling SA, Sykes MT, and Bradshaw RHW (2001) Palaeovegetation-model comparisons, climate change and tree succession in Scandinavia over the past 1500 years. *Journal of Ecology* 89: 227–236.

anthropogenic disturbance immediately prior to the replacement of *T. cordata* (small-leaved lime) by beech. Significant population expansions of beech were closely associated with fire episodes linked to anthropogenic activities at a forest hollow in Siggaboda, southeast Sweden, (Björkman and Bradshaw, 1996; Hannon et al., 2010; Figure 7) and the Siggaboda record is repeated at other sites in the region (Bradshaw and Lindbladh, 2005). Controversy, therefore,

concerns not so much whether beech should be present on a continental scale, but rather whether or not its dominance in southern Sweden and Denmark is a ‘natural’ feature of the landscape (i.e., having nonanthropogenic origins). Do climate-driven simulations of forest composition predict beech dominance for southern Scandinavia?

Simulations with a climate-driven forest stand simulation model in Draved Skov, Jutland, did not support the hypothesis

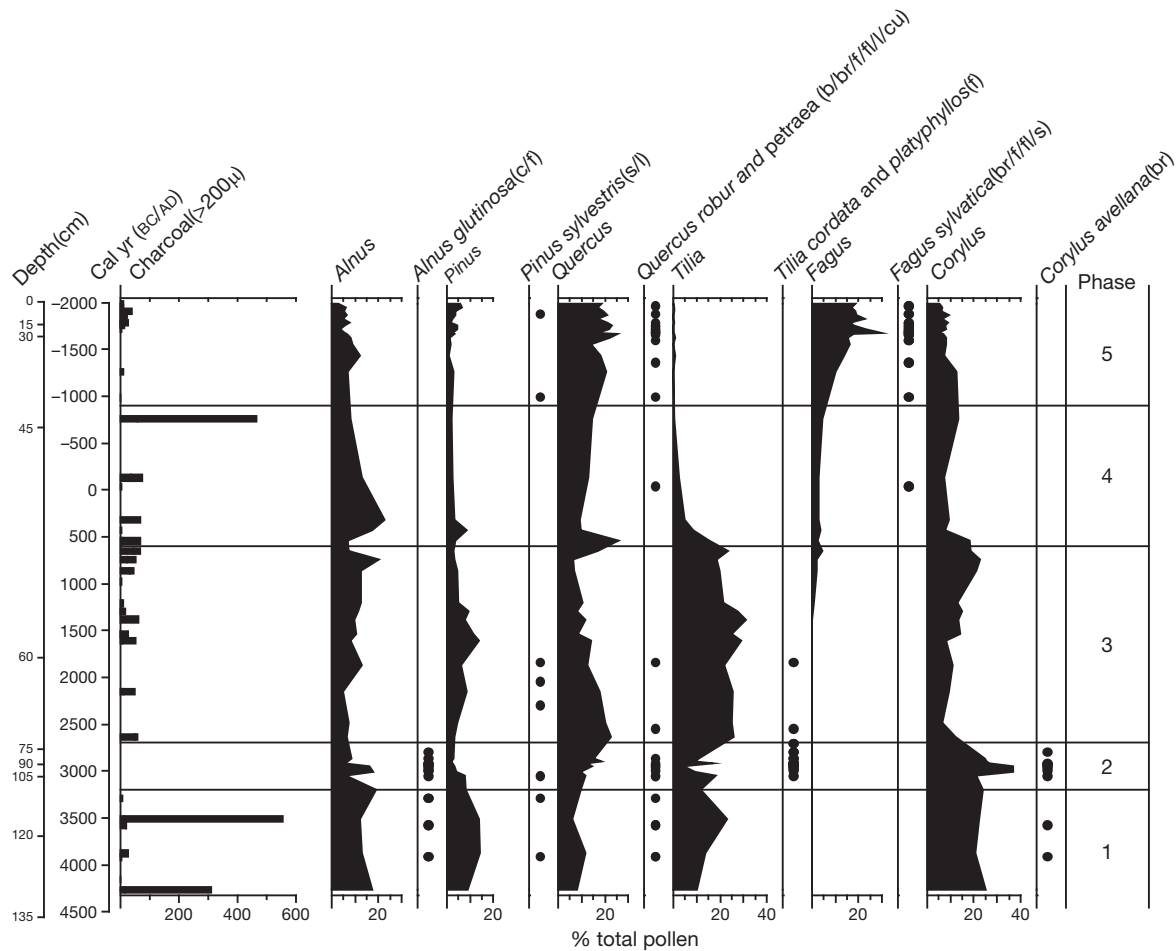


Figure 9 Suserup forest hollow, Denmark. Summary diagram showing selected trees recorded as pollen (filled curves) and macrofossils (●). The charcoal fragments are expressed as number per cm³. Reproduced from Hannon GE, Bradshaw RHW, and Emborg J (2000) 6000 years of forest dynamics in Suserup Skov, a seminatural Danish woodland. *Global Ecology and Biogeography* 9: 101–114.

that historical changes in climate were favorable for stand-scale dominance of beech (Figure 8; Cowling et al., 2001). The results indicated that, following cooling during the Little Ice Age, lime tree populations should have rapidly reestablished their dominance in deciduous forests. Observation, based on a stand-scale pollen analysis from mor humus in Draved Forest, recorded a replacement of lime by beech (Figure 9; Aaby, 1983).

Based on the results of climate-driven simulations, beech should not dominate modern-day northern temperate woodlands in Europe, whereas lime should be present in significantly greater quantities than are observed today (Figure 8). The minor presence of lime has been a long-standing feature of forests in Northwest Europe and can best be explained by anthropogenic interference. Forest clearance for agriculture followed by subsequent abandonment of fields favors the replacement of lime by beech, as does the introduction of domestic livestock into the forest (Björse and Bradshaw, 1998). Thus, the widespread late Holocene replacement of lime by beech in Northwest Europe was driven by anthropogenic forces even though forest changes of comparable scale earlier in the Holocene were climate driven.

Charcoal and macrofossils

Stand-scale forest structure and the importance of burning and forest grazing in the past are issues of conservation and management interest. Plant macrofossil and charcoal studies from southern Scandinavia have helped inform the debate about the degree of forest openness in the past, and relevant data have come from small forest hollows (Bradshaw and Hannon, 2004; Bradshaw et al., 2010; Hannon et al., 2010). The record from Suserup Skov, Denmark, covered the last 6000 years and contained abundant large fragments of charcoal from 5000 years ago until 1000 years ago (Hannon et al., 2000; Figure 9). The charcoal gave evidence of local burning, but the fires were almost certainly ground fires that maintained rather open conditions in the forest and kept out particularly fire-sensitive species. European beech, a very fire-sensitive species, only became abundant when the burning ceased while Scots pine, which tolerates moderate fire, was a natural component of the vegetation only during the periods of most frequent burning.

The plant macrofossil record at Suserup considerably broadens our knowledge of mid- to late Holocene forest composition in Denmark. It shows that both small-leaved lime and *Tilia platyphyllos* (large-leaved lime) grew on site about

5000 years ago as did *Quercus robur* (pedunculate oak) and *Q. petraea* (sessile oak). At this time, there is stand-scale evidence for at least 16 tree and shrub taxa, which is a unique record of diversity for Northwestern Europe that is unmatched in present woodlands. Suserup is a fertile, base-rich site, but its mid-Holocene record was probably characteristic of other similar as yet uninvestigated sites. Understanding the loss of this diversity is a research challenge that has relevance for the protection of European forest biodiversity at present.

Conclusions

Stand-scale palynology permits spatial resolution that provides detailed insight into dynamic vegetation processes that cannot be resolved using conventional pollen analysis. The case studies presented in this article give information about invasion mechanisms, forest structure, local diversity, succession, fire, and disturbance processes that link directly to topical issues in long-term ecological research and conservation biology. The recognition of the former importance of lime in northern and central European forests is one important conclusion from stand-scale studies. Stand-scale sites also link to conventional regional sites, where the research focus is on reconstruction of past climate and land cover. Thus, stand-scale sites provide a pivotal link between Quaternary science and ecology. The understanding of stand-scale sites has benefited from recent developments in pollen representation theory. A promising future application is their increased use in the validation and development of stand-scale forest simulation models.

See also: **Pollen Methods and Studies: Numerical Analysis Methods; POLLSCAPE Model: Simulation Approach for Pollen Representation of Vegetation and Land Cover.**

References

- Aaby B (1983) Forest development, soil genesis and human activity illustrated by pollen and hypha analysis of two neighbouring podzols in Draved Forest, Denmark. *Danmarks Geologiske Undersøgelse II* 114: 1–114.
- Andersen ST (1970) The relative pollen productivity and pollen representation of North European trees, and correction factors for tree pollen spectra. *Danmarks Geologiske Undersøgelse II* 96: 1–99.
- Andersen ST (1984) Forests at Løvenholm, Djursland, Denmark, at present and in the past. *Biologiske Skrifter Kongelige Danske Videnskabernes Selskab* 24: 1–208.
- Andersen ST (1997) Pollen analyses from early Bronze age barrows in Thy. *Journal of Danish Archaeology* 13: 7–17.
- Bialozyt R, Bradley L, and Bradshaw RHW (2012) Modelling the spread of *Fagus sylvatica* and *Picea abies* in southern Scandinavia during the late Holocene. *Journal of Biogeography* 39: 665–675. <http://dx.doi.org/10.1111/j.1365-2699.2011.02665.x>.
- Björkman L and Bradshaw RHW (1996) The immigration of *Fagus sylvatica* L. and *Picea abies* (L.) Karst into a natural forest stand in southern Sweden during the last 2000 years. *Journal of Biogeography* 23: 235–244.
- Björse G and Bradshaw R (1998) 2000 years of forest dynamics in southern Sweden: Suggestions for forest management. *Forest Ecology and Management* 104: 15–26.
- Bradshaw RHW (1981) Modern pollen-representation factors for woods in Southeast England. *Journal of Ecology* 69: 45–70.
- Bradshaw RHW (1988) Spatially-precise studies of forest dynamics. In: Huntley B and Webb T III (eds.) *Vegetation History*, pp. 725–751. Dordrecht: Kluwer Academic Publishers.
- Bradshaw RHW and Hannon GE (2004) The Holocene structure of North-West European temperate forest induced from palaeoecological data. In: Honnay O, Verheyen K, Bossuyt B, and Hermy M (eds.) *Forest Biodiversity: Lessons from History for Conservation*, pp. 11–25. Wallingford: CAB International.
- Bradshaw RHW and Lindbladh M (2005) Regional spread and stand-scale establishment of *Fagus sylvatica* and *Picea abies* in Scandinavia. *Ecology* 86: 1679–1686.
- Bradshaw RHW, Lindbladh M, and Hannon GE (2010) The role of fire in Southern Scandinavia during the late Holocene. *International Journal of Wildland Fire* 19: 1040–1049.
- Bradshaw RHW, Wolf A, and Møller PF (2005) Long-term succession in a Danish temperate deciduous forest. *Ecography* 28: 157–164.
- Calcote R (1995) Pollen source area and pollen productivity: Evidence from forest hollows. *Journal of Ecology* 83: 591–602.
- Cowling SA, Sykes MT, and Bradshaw RHW (2001) Palaeovegetation-model comparisons, climate change and tree succession in Scandinavia over the past 1500 years. *Journal of Ecology* 89: 227–236.
- Davis MB, Calcote RR, Sugita S, and Takahara H (1998) Patchy invasion and the origin of a hemlock-hardwoods forest mosaic. *Ecology* 79: 2641–2659.
- Dimbleby GW (1957) Pollen analysis of terrestrial soils. *New Phytologist* 56: 12–28.
- Dimbleby GW (1985) *The Palynology of Archaeological Sites*. Orlando, FL: Academic Press.
- Hannon GE, Bradshaw RHW, and Emborg J (2000) 6000 years of forest dynamics in Suserup Skov, a seminatural Danish woodland. *Global Ecology and Biogeography* 9: 101–114.
- Hannon GE, Niklasson M, Brunet J, Eliasson P, and Lindbladh M (2010) How long has the 'hotspot' been 'hot'? Past stand-scale structures at Siggaboda nature reserve in southern Sweden. *Biodiversity and Conservation* 19: 2167–2187.
- Higuera PE, Sprugel DG, and Brubaker LB (2005) Reconstructing fire regimes with charcoal from small-hollow sediments: A calibration with tree-ring records of fire. *The Holocene* 15: 238–251.
- Iversen J (1964) Retrogressive vegetational succession in the Postglacial. *Journal of Ecology* 52(supplement): 59–70.
- Jacobson GL and Bradshaw RHW (1981) The selection of sites for paleovegetational studies. *Quaternary Research* 16: 80–96.
- Jacobson GL, Webb T III, and Grimm EC (1987) Patterns and rates of vegetation change during the deglaciation of Eastern America. In: Ruddiman WF and Wright HE Jr. (eds.) *North America and Adjacent Oceans During the Last Deglaciation*, pp. 277–288. Boulder, CO: Geological Society of America.
- Lindbladh M, Bradshaw RHW, and Holmqvist BH (2000) Pattern and process in South Swedish forests during the last 3000 years, sensed at stand and regional scales. *Journal of Ecology* 88: 113–128.
- Lindbladh M and Foster DR (2010) Dynamics of long-lived foundation species: The history of *Quercus* in Southern Scandinavia. *Journal of Ecology* 98: 1330–1345.
- Mitchell FJG (1998) The investigation of long-term successions in temperate woodland using fine spatial resolution pollen analysis. In: Kirby KJ and Watkins C (eds.) *The Ecological History of European Forests*, pp. 213–223. Oxford: CAB International.
- Mitchell FJG (2005) How open were European primeval forests? Hypothesis testing using palaeoecological data. *Journal of Ecology* 93: 168–177.
- Prentice IC (1985) Pollen representation, source area, and basin size – Toward a unified theory of pollen analysis. *Quaternary Research* 23: 76–86.
- Sugita S (1994) Pollen representation of vegetation in Quaternary sediments – Theory and method in patchy vegetation. *Journal of Ecology* 82: 881–897.

Changing Plant Distributions and Abundances

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Introduction

The spread of plants is a continuous process and yet difficult to observe directly because plants are sessile organisms that spread through the dispersal of propagules (organs used for propagation). Natural plant distribution limits are generally set by species-specific climatic parameters that determine their survival and regeneration. However, these areas of potential distribution determined by climate are often more extensive than the actual species distributions (Svenning and Skov, 2004). These discrepancies might indicate that the distributions of many species are dispersal limited (Svenning and Skov, 2007), especially in high and mid-latitudes, where glacial cycles have repeatedly altered plant distributions. Recent concerns regarding the impacts of global change have raised the question of whether plant distributions can shift fast enough to keep up with the changing climatic conditions.

Plants may also spread when previously insurmountable barriers to dispersal are overcome. Over the course of millions of years, this can occur through the emergence of land bridges, with lowered sea levels, or the collision of continents and the closure of straits. Especially during the past 500 years, humans have both deliberately and unwittingly moved innumerable plants over dispersal barriers and thus altered many ecosystems.

The spread of plants can be studied on many different timescales: from the effect of alien species on ecosystems within a few years to subcontinental invasions of major trees lasting for several thousand years and the spread of plant families and genera between continents on the scale of millions of years. Knowledge of the dynamics and patterns of plant invasions is therefore gathered from different disciplines including paleontology and paleoecology, as well as ecology and genetics (Petit et al., 2004). Results of this research are important for evaluation of possible consequences of global warming for species distributions and abundance (Pitelka et al., 1997), as well as assessment of threats from introduced and genetically modified plants to ecosystem structure and biodiversity.

This article focuses on the reconstruction of changes in plant distribution and abundance based on pollen diagrams. The focus is therefore on the distribution of wind-pollinated trees, in subcontinental space, during thousands of years. New insights from DNA studies and numerical modeling contribute greatly to our current understanding, and these aspects are therefore included in this overview.

How to Gain Knowledge About the Spreading of Plants

Information on changes in plant distributions on short timescales can be obtained from a variety of sources such as air photos, satellite images, historical records, herbarium collections, or the age structure of invasive trees within an

area. Paleoecological techniques have to be used to reconstruct the temporal pattern in changes of plant distributions on longer timescales. Plant remains that preserve in favorable environments at or near their place of growth testify to the presence of that plant at the time when the material was deposited. By far, the most abundant plant remains are pollen grains, which preserve well in the sediments of lakes and bogs or in very dry environments. These environments also preserve larger plant remains like needles and seeds (macrofossils) or branches and stumps (megafossils). Both microscopic (pollen) and macroscopic (from bud scales to tree trunks) plant remains are used in reconstructing past plant distributions, yielding different perspectives.

Although the identification of fossil pollen grains is generally only possible to generic or even family level, as for many herbs, macroscopic plant remains are often identifiable to species level. Macrofossils are rarely transported over large distances, and their presence at a location therefore clearly indicates that the plant was present near the site. Macrofossils are often readily identifiable even in the field, and thus, it is not surprising that the earliest information on past vegetation came from macroscopic plant remains (Birks and Seppä, 2010). Andersson (1902) was able to compile the first map depicting past distribution limits of *Corylus avellana* (hazel) in Sweden with the help of many interested laymen who reported finds of hazel nuts outside the distribution of the shrub (Figure 1). At the time when Andersson compiled his map, it was impossible to directly obtain the age of the individual finds, and thus, the age of this northernmost occurrence of *C. avellana* could not be obtained. As in this example, macroscopic plant remains are often conspicuous and are thus discovered during peat harvesting, road building, and archaeological excavations or simply by chance. Tree stumps and trunks above or beyond the treeline are another example where evidence of past species' occurrence is easily found. Since the development of radiocarbon dating in the 1950s, such finds can be directly dated and thus represent a point in space and time, which together with other finds can yield a thorough assessment of past changes in species distribution.

Although macrofossils allow for spatially precise reconstructions, they often yield little information on the average upland vegetation composition, which is more readily obtained from pollen analysis. Moreover, pollen analysis often yields continuous information and inherently delivers some crude age control. The latter aspect was potentially the most important advantage at the time of the technique's development in the 1920s, in the research era before radiocarbon dating. As a consequence, macrofossil analysis received little attention in the decades following the 1920s, especially in Scandinavia, the birthplace of pollen analysis (Birks and Seppä, 2010).

Pollen analytical results, on the other hand, quickly became available, and only 8 years after von Post's 1916 lecture that established pollen analysis as a standard method, von Post

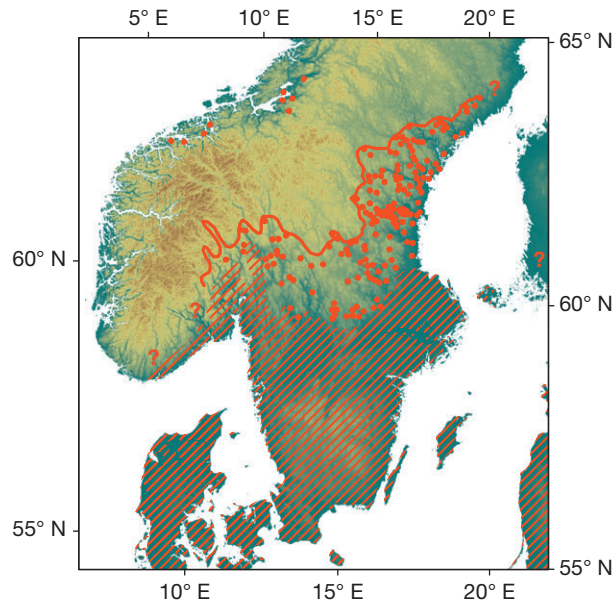


Figure 1 The present (hatched) and past distribution limit (line) of hazel (*Corylus avellana*) in Scandinavia redrawn from Andersson G (1902) Hasseln i Sverige fördom och nu. *Sveriges Geologiska Undersökning Series C 3*: 1–168. Dots mark the location of fossil hazel nut finds reported to Andersson.

(1924) published maps depicting the relative abundance of particular taxa during different time periods in southern Sweden. In these maps, the abundance of a particular taxon was depicted by circles with different radii, representing the pollen abundance of the taxon at the site. The number of available pollen diagrams also quickly increased outside Sweden. Thus, Szafer (1935) treated pollen percentages from 152 sites across Poland as if they were elevation data and produced the first isopollen maps, where the lines represent equal pollen percentages of a specific taxon at a given time. In his maps, he also decided on a threshold value to represent presence of the taxon in the area. The need for such threshold values represents the major handicap in using pollen analytical results to reconstruct past species limits. Wind-pollinated plants produce large amounts of pollen that is transported over vast distances, which makes it difficult to decide whether small amounts originated from locally occurring, but rare specimens or large, distant populations.

Since the time of von Post and Szafer, numerous pollen diagrams covering the last 10–15 ka have been produced from all continents and are an excellent source for studying changes in plant distribution and abundance on long timescales and over vast areas. To make better use of these invaluable datasets, they are collected into databases and made available through the internet. Reported macrofossils finds are also compiled into databases. In the new paleoecological database Neotoma, these datasets can be extracted in conjunction with pollen data.

The application of genetic techniques to the reconstruction of plant spreading has greatly enhanced our insights into patterns and processes, and the application of ancient DNA to these questions is poised to further change our perspectives (Gugerli et al., 2005). Geographical distributions of genetic

markers in plants reveal their kinship and make it possible to reconstruct the colonization routes of different lineages, which have changed established ideas about plant migration (e.g., Magri et al., 2006). The composition and diversity of genetic markers in extant populations gives insight into some of the processes in plant populations that were at work when the species were spreading 10 ka earlier (Petit et al., 2002b). Because plant spreading occurs over long time periods, it is difficult to gain knowledge from experimental, field observations or monitoring studies designed to test ideas. Dispersal theory has evolved models that offer robust methods to test hypotheses, to help understand past plant spreading, and to permit predictions for future scenarios (Clark et al., 1998).

Examples of Late Quaternary Plant Spreading

As a suite of research techniques, paleoecology developed in the mid- to high latitudes of the Northern Hemisphere, regions that changed dramatically between cold and warm stages of the Quaternary. Many plants that are widespread today were restricted to small areas under the climate that prevailed during the glacial intervals. Seen from today's perspective, these ranges are termed refugia (Bennett and Provan, 2008). The potential growth area for many of these plants increased rapidly when temperatures rose during the Late Glacial and the onset of the Holocene. In consequence, plants spread into these newly-available areas and competed for new habitats. However, the plant communities we know today did not move as belts, but species spread independently from each other and new combinations of different plants formed in time and space (Bennett and Provan, 2008; Davis, 1976; Huntley, 1988).

The Holocene changes in the distribution and abundance of deciduous *Quercus* (oak) and *Tilia* (lime) in Europe offer good examples of how the composition of forests changed through time (Figure 2). The maps in Figure 2 are taken from the atlas of past and present pollen maps for Europe (Huntley and Birks, 1983), which shows isopollen maps for European taxa in 500 or 1000 radiocarbon-year time steps. When comparing isopollen maps for different taxa, it has to be kept in mind that different trees produce different amounts of pollen, and it is therefore not possible to compare the numerical values directly. In this example, oak species are large pollen producers, and lime species disperse little pollen. Therefore, their respective pollen percentages were reclassified to improve visual comparison.

For the beginning of the Holocene (Figure 2(d)), lime pollen was frequently found only in sediments from southeastern Europe, while deciduous oak pollen was abundant in records from the Iberian, Apennine, and Balkan peninsulas. At about 10 ka, many east European forests contained lime but not oak, whereas the opposite was the case in western Europe (Figure 2(c)). In the following millennia, both lime and oak species were abundant in central European forests (Figure 2(b)), and it was mainly the abundance of lime that declined toward the present (Figure 2(a)).

Isopollen maps are a useful tool to present changes in plant abundance through time, but because each time slice is presented on a different map, it is difficult to extract

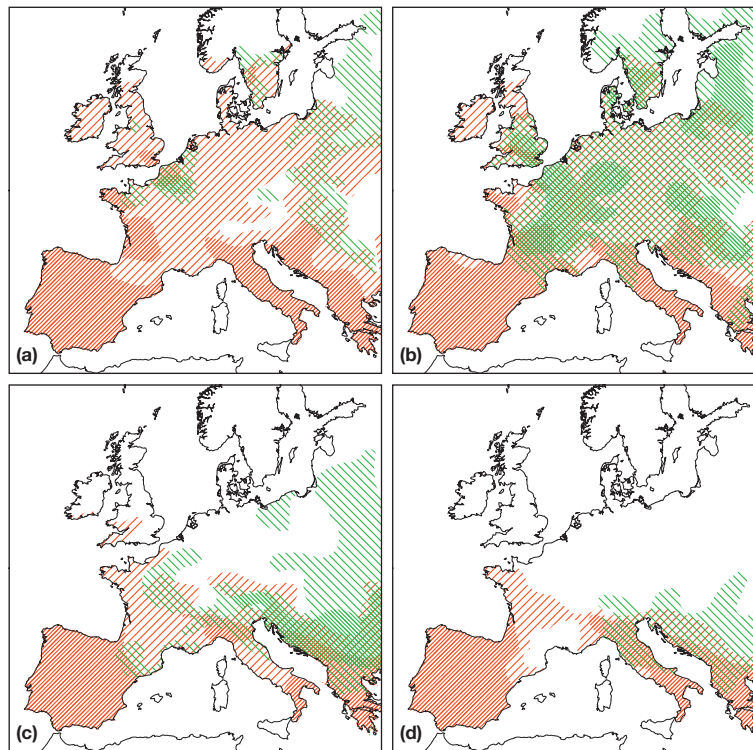


Figure 2 Map of changing distribution and abundance of deciduous oak (red) and lime (green) pollen at about (a) 2000 years ago, (b) 7000 years ago, (c) 10000 years ago, and (d) the beginning of the Holocene about 11500 years ago. The wide-hatched areas mark the distribution of frequent occurrences of deciduous oak and lime as indicated by more than 5 and 1% of their pollen, respectively. The narrow-hatched areas indicate regions where the tree comprised an important proportion of the forest indicated by 25 and 5% of their pollen, respectively. Percentages are based on the sum of tree pollen. Redrawn and simplified from Huntley B, and Birks HJB (1983) *An Atlas of Past and Present Pollen Maps for Europe 0–13000 Years Ago*, pp. 667. Cambridge: Cambridge University Press.

information on the dynamics of plant spreading or continental-scale changes in abundance patterns. An alternative approach is to choose a specific point on the pollen curve (e.g., a threshold value) and map the time of its occurrence in space. These maps depict lines of equal time for the occurrence of a specific event (isochrone maps) (Figures 3 and 4). Often, isochrone maps depict features in pollen diagrams that are interpreted to represent the moment when the species concerned has arrived in the immediate area surrounding a site (Birks, 1989; Davis, 1976). Such maps give valuable insight into the dynamics of a taxon's spread, but they simplify the problem of identifying the arrival of species from pollen data and often give the impression of a frontlike spread of a dense population.

The spread of *Picea abies* (Norway spruce) in Scandinavia provides a good example of this problem. When plotting the time that different threshold values were reached in the abundant pollen diagrams from Scandinavia and adjacent areas, a clear frontlike pattern emerges (Giesecke and Bennett, 2004) (Figure 3). On the basis of this analysis, the spread of Norway spruce occurred in the area of present-day Sweden <5000 years ago. However, several large Norway spruce remains (e.g., cones) extracted from peats in central Sweden date to the early Holocene (Kullman, 2001). Moreover, a few pollen records indicate the presence of the tree several thousand years before it became abundant (Figure 3). Some Finnish pollen diagrams even indicate an early Holocene presence and a later decline of spruce pollen. Furthermore, the spatial patterns of

DNA markers in mitochondrial and nucleus DNA from current Norway spruce populations suggest that the tree spread into Scandinavia from several different areas (Tollefsrud et al., 2008, 2009). The direction of spread reconstructed from pollen cannot explain the pattern of DNA markers and may be only one of the realized routes. The combined evidence suggests that Norway spruce spread into Scandinavia long before it became abundant and that this spread occurred without leaving a clear palynological trace. It is conceivable that a first spread, with only a low population density, followed the retreat of the ice sheet. Winged spruce seeds could have been transported over many kilometers in late winter on ice-crusted snow (Giesecke and Bennett, 2004). This could have led to small scattered populations, some of which subsequently declined while others grew only slowly and were too small to be detectable by conventional pollen analyses for thousands of years.

These contrasting patterns of palynological, genetic and megafossil evidence raise the question of what triggered the wave of expanding populations of Norway spruce during the late Holocene (Giesecke, 2004). It also shows that none of the three sources of information alone can accurately describe the Holocene history of Norway spruce in Scandinavia. Studies on ancient DNA from several sites are needed to obtain more evidence for enhancing our understanding of the postglacial history of this important tree species in Scandinavia.

The spread of deciduous oak species in Europe is a good example of how results from pollen analysis and a survey of

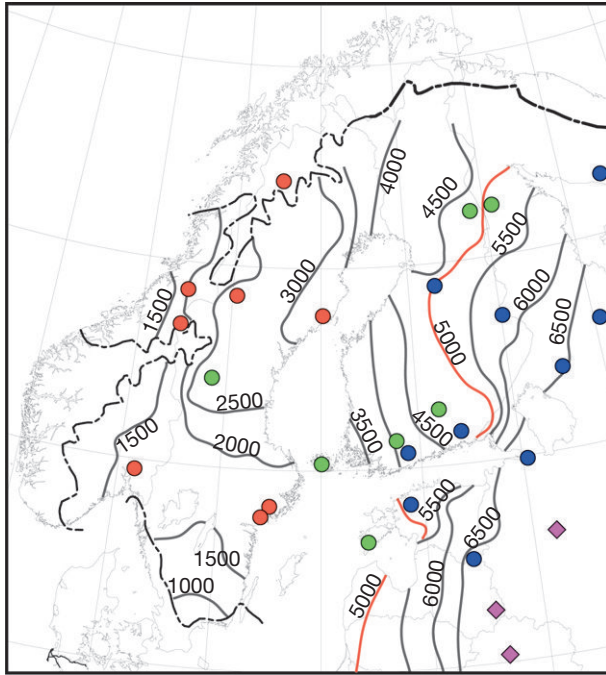


Figure 3 The spread of Norway spruce (*Picea abies*) in Fennoscandia and adjacent areas. Interpolation of the time the 5% threshold was reached in pollen diagrams. Circles show the position of pollen diagrams indicating small spruce populations deduced from the continuous curve or the 1% threshold at approximately 9000 = blue, 7000 = green, and 5000 years ago = red. Purple diamonds mark the position of diagrams with more than 3% spruce pollen during the Late Glacial.

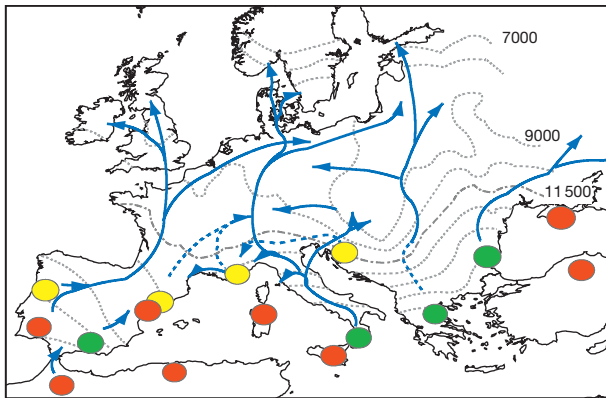


Figure 4 Major spreading routes and dynamics for the spread of deciduous oak in Europe. The time of the beginning of the continuous pollen curve (rational limit) is represented by gray broken lines of equal time. Blue arrows mark the spreading routes reconstructed from extant haplotypes, and circles indicate proposed refugia: green = primary refugia deduced from pollen data; red = primary refugia additionally proposed from genetic data; yellow = secondary refugia. Adapted from Brewer S, Cheddadi R, de Beaulieu JL, and Reille M (2002) The spread of deciduous *Quercus* throughout Europe since the last glacial period. *Forest Ecology and Management* 156: 27–48; Petit RJ, Brewer S, Bordacs S, et al. (2002) Identification of refugia and post-glacial colonisation routes of European white oaks based on chloroplast DNA and fossil pollen evidence. *Forest Ecology and Management* 156: 49–74.

genetic markers in present-day populations complement each other to reconstruct spreading history. As mentioned previously, it is often not possible to detect small populations using pollen analysis, which is a hindrance when trying to gain information on where species were growing during the Last Glacial Maximum (LGM). The survey of genetic markers in living populations permits construction of the kinship between populations throughout Europe. Populations with a specific genetic marker (haplotype) in the south persisted during the northward spread of that population so that the route along which the population was spreading is marked today by individuals sharing the same haplotype. Not all populations with unique haplotypes succeeded in the race for new habitats. This effect is called a genetic bottleneck and results in detectable greater genetic diversity in the refuge areas, with lower genetic diversity toward the periphery of the taxon's range. In this way, genetic markers have been used to verify *Quercus* refugia that were identified from pollen analysis and to identify additional oak refugia (Figure 4). The combination of an isochrone map with proposed spreading routes of major haplotypes of oak (Figure 4) allows new insight into spreading history. The portrayal of the spreading routes takes account of obstacles like mountain ranges, while the isochrones represent an interpretation on a plane. It is interesting to note that the spreading routes are not always perpendicular to the isochrones, supporting the concept that the initial spread occurred at low population densities in a chaotic, rather than directional manner.

The geographic distribution of genetic markers in living *Fagus sylvatica* (beech) populations is another example of how the combination of pollen and DNA analysis reveals new insights into the postglacial spread of species (Magri et al., 2006). Pollen analysis suggested that beech populations occurred around the Mediterranean basin during the LGM and initial summaries suggested colonization from the Balkans toward the west (Giesecke et al., 2007). Genetic results support the location of these palynologically identified LGM populations but show that only a few LGM populations were the source of the northward and westward spread of the tree, while other populations in the south showed little expansion during the Holocene and did not contribute to the gene pool of northern beech populations (Figure 5). Svenning and Skov (2004, 2007) argue that different tree species have expanded their ranges from their refugia to a different extent. Here, we see an example where different populations of the same species behave differently. The data also demonstrate that the shortest distance between a population and the nearest glacial refuge area is not necessarily an indication of the direction of colonization.

Speed and Pattern: How Did It Work?

The Late Glacial and early Holocene spread of temperate and boreal trees for new habitats in treeless areas reached dispersal rates of over 1000 m year⁻¹. This rate incorporates the time that a tree needs to grow from seed to maturity (generation time), as well as the distance over which the seed has to be dispersed. By multiplying the spreading rate with generation time, one obtains the dispersal distance. For example, the early Holocene spread of oak in Europe shows spreading rates of 500 m year⁻¹, which, multiplied by a generation time of about

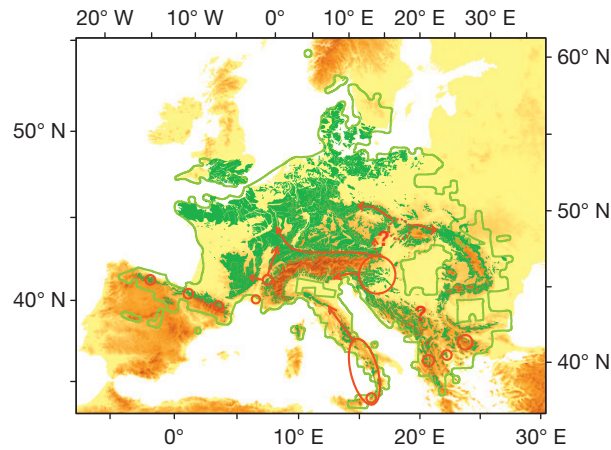


Figure 5 Distribution and Holocene spread of *Fagus sylvatica* (beech) in Europe. The green line marks the natural distribution according to *Atlas Flora Europaea* (Jalas J and Suominen J (eds.) (1972–99) *Atlas Florae Europaeae: Distribution of Vascular Plants in Europe*. Helsinki: The Committee for Mapping the Flora of Europe and Societas Biologica Fennica Vanamo) and the filled green areas the potential distribution of beech (Bohn U, Gollub G, and Hettwer C (2000) *Karte der natürlichen Vegetation Europas*. Massstab 1:2.500.000 Karten und Legende. Bonn: Bundesamt für Naturschutz). Red circles mark tentative location of refuge areas and arrows indicate main colonization routes (Magri D, Vendramin GG, Comps B, et al. (2006) A new scenario for the Quaternary history of European beech populations: Palaeobotanical evidence and genetic consequences. *New Phytologist* 171: 199–221).

30 years, amounts to a seed dispersal distance of 15 km. But acorns are heavy seeds and usually only found several meters from their parent tree. This mismatch between the seed dispersal distances that can be observed today and the distances necessary to explain early Holocene tree spreading is known as Reid's paradox of rapid plant migration (Clark et al., 1998). It is hard to imagine that many acorns are dispersed over 15 km, but it is possible that a single seed out of a population of many thousands is transported over long distances. Dispersal theory describes this concept with seed dispersal functions that have a high kurtosis: they predict that the greatest amounts of seeds will be dispersed close to the source but allow for a small number to be dispersed over long distances.

When applying this concept to the early Holocene spread of oak, it may be assumed that a single seed was deposited many km away from the nearest oak tree. If this seed germinated and grew into a tree, which in turn dispersed seeds that mostly fell close to the tree, this single tree would become the founder of a new population (Figure 6(c)). This mechanism seems to be the most likely explanation for the actual distribution of haplotypes in oaks of western France (Petit et al., 1997). Here, oak trees were sampled within a region with high resolution, and the distribution of chloroplast DNA markers was analyzed. In oaks, chloroplasts are inherited only from the mother (maternally) and thus through the heavy seed. The analysis revealed a mosaic with different areas dominated by distinct haplotypes. Thus, the population of oak trees in each of these areas may descend from one single founding tree, through the maternal line.

Genetic evidence has also yielded another possible mechanism for the high spreading rates that are reconstructed from pollen data. The haplotype distribution of *Fagus grandifolia*

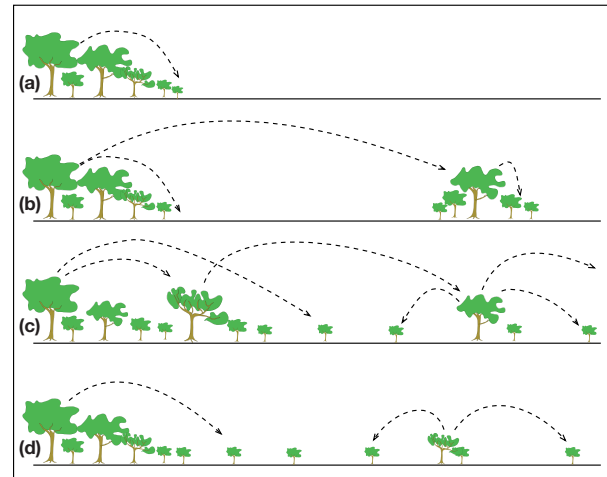


Figure 6 Models of tree spread; broken lines mark dispersal: (a) Continuous moving front: dispersal takes place from a mature dense population to just beyond the distribution limit. (b) Advance as a discontinuous population: few discrete long-distance dispersal events lead to founder populations, which in turn expand and finally merge with each other. (c) Spread at low population density: frequent long-distance dispersal leads to small scattered founder populations, which are new sources for long-distance dispersal events. (d) Merging of disjunct populations: small scattered refuge populations expand and are a source for dispersal events as well as large distant populations. Both populations merge eventually. (a) and (b) Redrawn from Davis MB (1987) *Invasion of forest communities during the Holocene: Beech and hemlock in the Great Lakes region*. In: Gray AJ, Crawley MJ, Edwards PJ (eds.) *Colonization, Succession, and Stability*. Oxford: Blackwell Scientific.

(American beech) and *Acer rubrum* (red maple) in North America suggests that the glacial refugia of both trees were much further north than indicated by pollen data (McLachlan et al., 2005). These populations, which persisted as close as 500 km south of the ice margin, may have consisted of groups of trees in habitats with favorable microclimate (cryptic refugia) too small to be detected by pollen analysis (Figure 6(d)). These northern populations, as well as larger southern populations, expanded after the last glaciation. Without the recognition of these northern refugia, the expansion of these populations would be easily interpreted as a rapid spread of the more southern populations, resulting in apparent high spreading rates that never actually occurred.

However, many regions (such as those covered by ice sheets) were most certainly bare of any trees during the LGM and yet rapid spreading rates have been reconstructed for these areas (McLeod and MacDonald, 1997). Investigations of carefully selected and well-dated lake sediments allow the reconstruction of pollen accumulation rates (PAR), which is a better measure of changes in tree abundance than the interdependent percentage. The increase in PAR estimates of a pollen taxon can be used as a proxy for the population increase of the parent tree and compared to population growth models (Bennett, 1983, 1988). Such studies interpret the often exponential increase in PAR to represent the expansion of small populations that would have been present near the site even before the pollen type is encountered in standard counts.

The above interpretations on the mechanisms of plant spread are in contrast to traditional models of the frontlike

movement of a dense population (Figure 6(a); Davis, 1987). These traditional models find support in the observation of sharp distribution limits of present species ranges (Davis et al., 1991). The model of a moving front may be inferred from the presentation of species spreading in the form of isochrone maps and the common observation that regional population expansions often follow a wavelike pattern. However, the moving front model cannot explain exponential growth curves of PAR estimates (Giesecke, 2005) and regional pattern of haplotypes (Petit et al., 1997), and it also does not comply with theoretical models (Clark et al., 1998). The moving front model may therefore be inadequate to describe tree spreads on the continental or regional scale, although it may still be applicable on the local scale.

Distribution Versus Abundance

In a simple thought experiment, Bennett (1985) showed that, using pollen analysis, it should be impossible to detect a single tree of a given species in a forest of different species. What can be reconstructed from a network of pollen diagrams is the range of common occurrence of a species. A good example to illustrate the difference between the species distribution limit and the area of abundance can be seen for European beech (*F. sylvatica*) in Figure 5. The *Flora Europaea* (Jalas and Suominen, 1972–99) marks the simple occurrence of the species within a 50 by 50 km square, while the map of the potential natural vegetation (Bohn et al., 2000) shows areas where beech would be the dominant or characteristic tree. The latter is the area that can be reconstructed from pollen diagrams, while the former is generally invisible in reconstructions from pollen.

For some species, colonization is directly followed by population expansion and, assuming the rate of expansion is equal at all sites, it is possible to deduce the underlying spreading rates (Bennett, 1986). However, the rate of population expansion can vary between sites (Giesecke, 2005; Giesecke et al., 2007) or the species may spread at low population densities and expand thousands of years later. The possible occurrence of cryptic refugia (see Stewart and Lister, 2001) complicates matters even more, so that published rates of spread based on paleoecological data are potentially erroneous and need careful revision.

On the other hand, once a population has exceeded its detectable size, then pollen analysis is well suited to reconstruct changes in abundance. Many of the factors that might influence the spread of a species will also be important for changes in its abundance. Moreover, for some species, like Norway spruce, the questions have changed from analyzing its postglacial spread to understanding its late Holocene population expansion or the lack thereof during the early Holocene.

Factors Influencing the Spread of Trees and Changes in Abundance

The vegetation history of temperate and boreal trees in the Northern Hemisphere shows that some trees spread rapidly, following the warming at the end of the last glaciation, while others lagged behind, and some spread much later during the

Holocene. Many factors have been discussed to explain these patterns, which can be summarized as follows:

- *Seed dispersal.* The small winged seeds of *Betula* (birch), for example, can easily be picked up by storms and carried over large distances. The dispersal organs of plants may have a different likelihood of becoming dispersed over long distances. This factor, combined with different generation times, results in different theoretical rates of spread. Thus, it can be argued that the time at which a species arrives in a certain place is largely determined by the average spreading rate and the distance from its LGM refugia, taking into account obstacles like mountain ranges.
- *Competition between species.* Many light-demanding species are also good colonizers and thus dominated early phases of forest establishment during the Late Glacial and early Holocene. The trees that followed after had to compete with those that were already established. This interpretation finds support in the observation that many shade-tolerant species often need many generations at one site before they became abundant.
- *Soil development.* Young soils that developed from glacial deposits were rich in nutrients (especially phosphorous), which leached out over time. The growth of peat mosses changed nutrient-rich mires into nutrient-poor raised bogs, and blanket peats developed in formerly forested areas. Thus, changing soil properties may change the competitive balance between species. However, a change in vegetation cover can also cause a change in soil type (Willis et al., 1997), and it is therefore questionable if soil leaching could have been the driver for late Holocene tree spreads or changes in abundance.
- *Changes in forest disturbance.* Natural disturbance regimes, such as fire and storms, have changed in frequency which may influence the balance between species with short versus long generation times. There is clear evidence in Europe that human activity has significantly changed the vegetation over the last 3000–8000 years. By opening the forest, human activity at least accelerated the population expansion of late-spreading trees, and it may even have facilitated it. The utilization of the forest by humans and their domesticated animals also altered the competitive interaction between trees, which may also have allowed species to spread into areas from which they would have been otherwise excluded by competition.
- *Climate change.* The hypothesis that regional Holocene climate change has the potential to explain most patterns of tree spreading is difficult to test because pollen diagrams themselves are the most abundant proxy record of past climate, which leads to circular arguments. Arguments for climate-controlled late Holocene tree spreads are therefore often made by rejecting all alternative hypotheses (Tinner and Lotter, 2006). Advances in modeling Holocene climate, combined with vegetation modeling, will make it possible to further test this hypothesis.
- *Adaptation and gene flow.* Widespread tree species grow under different climate parameters in different parts of their range and often show specific adaptations to those climate parameters. If the acquisition of specific adaptations is slow, then this could explain the delayed spreading of some species (Davis and Shaw, 2001).

Conclusion

Results of paleoecological investigations have yielded abundant information that allows the reconstruction of continental-scale changes in plant abundance. Recent results from surveys of genetic markers in present-day populations reveal colonization pathways, which often differ from the interpreted direction of spread obtained from paleoecological data. This leads to the interpretation that the initial spread of species after the LGM occurred at low abundances and that populations may have expanded only several thousand years later. Inferences about spreading rates are therefore only possible for species where the spread was immediately followed by population expansion. Many of the rates published so far are potentially erroneous and should therefore not be used in studies predicting the impact of climate change on species distributions.

Paleoecological data are well suited to demonstrate past changes in species abundance and have already revealed contrasting patterns of population expansion. Therefore, further research should aim at uncovering the mechanisms and drivers behind the different continental-scale patterns of population dynamics. Genetic studies may yield further surprises, and the utilization of fossil DNA may open a new chapter of research into the processes behind past changes in species distributions and abundances.

See also: Pollen Analysis, Principles. **Paleobotany:** Ancient Plant DNA; Charred Particle Analyses. **Plant Macrofossil Methods and Studies:** Validation of Pollen Studies. **Pollen Methods and Studies:** Archaeological Applications; Databases and Their Application; Surface Samples and Trapping.

References

- Andersson G (1902) Hasseln i Sverige fördom och nu. *Sveriges Geologiska Undersökning Series C* 3: 1–168.
- Bennett KD (1983) Postglacial population expansion of forest trees in Norfolk, UK. *Nature* 303: 164–167.
- Bennett KD (1985) The spread of *Fagus grandifolia* across eastern North America during the last 18000 years. *Journal of Biogeography* 12: 147–164.
- Bennett KD (1986) The rate of spread and population increase of forest trees during the postglacial. *Philosophical Transactions of the Royal Society of London Series B* 314: 523–531.
- Bennett KD (1988) Holocene geographic spread and population expansion of *Fagus grandifolia* in Ontario, Canada. *Journal of Ecology* 76: 547–557.
- Bennett KD and Provan J (2008) What do we mean by 'refugia'? *Quaternary Science Reviews* 27: 2449–2455.
- Birks HJB (1989) Holocene isochrone maps and patterns of tree-spreading in the British-Isles. *Journal of Biogeography* 16: 503–540.
- Birks HJB and Seppä H (2010) Late-Quaternary palaeoclimatic research in Fennoscandia – A historical review. *Boreas* 39: 655–673.
- Bohn U, Gollub G, and Hetwer C (2000) *Karte der natürlichen Vegetation Europas. Massstab 1:2.500.000 Karten und Legende*. Bonn: Bundesamt für Naturschutz.
- Brewer S, Cheddadi R, de Beaulieu JL, and Reille M (2002) The spread of deciduous *Quercus* throughout Europe since the last glacial period. *Forest Ecology and Management* 156: 27–48.
- Clark JS, Fastie C, Hurr G, et al. (1998) Reid's paradox of rapid plant migration – Dispersal theory and interpretation of paleoecological records. *Bioscience* 48: 13–24.
- Davis MB (1976) Pleistocene biogeography of temperate deciduous forests. *Geoscience and Man* 13: 13–26.
- Davis MB (1987) Invasion of forest communities during the Holocene: Beech and hemlock in the Great Lakes region. In: Gray AJ, Crawley MJ, and Edwards PJ (eds.) *Colonization, Succession, and Stability*. Oxford: Blackwell Scientific.
- Davis MB and Shaw RG (2001) Range shifts and adaptive responses to Quaternary climate change. *Science* 292: 673–679.
- Davis MB, Schwartz MW, and Woods K (1991) Detecting a species limit from pollen in sediments. *Journal of Biogeography* 18: 653–668.
- Giesecke T (2004) The Holocene spread of spruce in Scandinavia. *Comprehensive summaries of Uppsala Dissertations from the Faculty of Science and Technology* 1127: 1–46.
- Giesecke T (2005) Moving front or population expansion: How did *Picea abies* (L.) Karst. become frequent in central Sweden? *Quaternary Science Reviews* 24: 2495–2509.
- Giesecke T and Bennett KD (2004) The Holocene spread of *Picea abies* (L.) Karst. in Fennoscandia and adjacent areas. *Journal of Biogeography* 31: 1523–1548.
- Giesecke T, Hickler T, Kunkel T, Sykes MT, and Bradshaw RHW (2007) Towards an understanding of the Holocene distribution of *Fagus sylvatica* L. *Journal of Biogeography* 34: 118–131.
- Gugerli F, Parducci L, and Petit RJ (2005) Ancient plant DNA: Review and prospects. *New Phytologist* 166: 409–418.
- Huntley B (1988) Europe. In: Huntley B and Webb T III (eds.) *Vegetation History*, pp. 345–383. Dordrecht: Kluwer Academic Publishers.
- Huntley B and Birks HJB (1983) *An Atlas of Past and Present Pollen Maps for Europe 0–13000 Years Ago*, pp. 667. Cambridge: Cambridge University Press.
- Jalas J and Suominen J (eds.) (1972–99) *Atlas Florae Europaeae: Distribution of Vascular Plants in Europe*. Helsinki: The Committee for Mapping the Flora of Europe and Societas Biologica Fennica Vanamo.
- Kullman L (2001) Immigration of *Picea abies* into north-central Sweden. New evidence of regional expansion and tree-limit evolution. *Nordic Journal of Botany* 21: 39–54.
- Magri D, Vendramin GG, Comps B, et al. (2006) A new scenario for the Quaternary history of European beech populations: Palaeobotanical evidence and genetic consequences. *New Phytologist* 171: 199–221.
- McLachlan JS, Clark JS, and Manos PS (2005) Molecular indicators of tree migration capacity under rapid climate change. *Ecology* 86: 2088–2098.
- McLeod TK and MacDonald GM (1997) Postglacial range expansion and population growth of *Picea mariana*, *Picea glauca* and *Pinus banksiana* in the western interior of Canada. *Journal of Biogeography* 24: 865–881.
- Petit RJ, Pineau E, Demesure B, Bacilieri R, Ducousso A, and Kremer A (1997) Chloroplast DNA footprints of postglacial recolonization by oaks. *Proceedings of the National Academy of Sciences of the United States of America* 94: 9996–10001.
- Petit RJ, Brewer S, Bordacs S, et al. (2002a) Identification of refugia and post-glacial colonisation routes of European white oaks based on chloroplast DNA and fossil pollen evidence. *Forest Ecology and Management* 156: 49–74.
- Petit RJ, Csaikl UM, Bordacs S, et al. (2002b) Chloroplast DNA variation in European white oaks – Phylogeography and patterns of diversity based on data from over 2600 populations. *Forest Ecology and Management* 156: 5–26.
- Petit RJ, Bialozyt R, Garnier-Gere P, and Hampe A (2004) Ecology and genetics of tree invasions: From recent introductions to Quaternary migrations. *Forest Ecology and Management* 197: 117–137.
- Pitelka LF, Gardner RH, Ash J, et al. (1997) Plant migration and climate change. *American Scientist* 85: 464–473.
- Stewart JR and Lister AM (2001) Cryptic northern refugia and the origins of the modern biota. *Trends in Ecology & Evolution* 16: 608–613.
- Svenning JC and Skov F (2004) Limited filling of the potential range in European tree species. *Ecology Letters* 7: 565–573.
- Svenning JC and Skov F (2007) Could the tree diversity pattern in Europe be generated by postglacial dispersal limitation? *Ecology Letters* 10: 453–460.
- Szafer W (1935) The significance of isopollen lines for the investigation of the geographical distribution of trees in the post-Glacial period. *Bulletin de l'Académie Polonaise des Sciences et des Lettres B* 1: 235–239.
- Tinner W and Lotter AF (2006) Holocene expansions of *Fagus sylvatica* and *Abies alba* in Central Europe: Where are we after eight decades of debate? *Quaternary Science Reviews* 25: 526–549.
- Tollefsrud MM, Kissling R, Gugerli F, et al. (2008) Genetic consequences of glacial survival and postglacial colonization in Norway spruce: Combined analysis of mitochondrial DNA and fossil pollen. *Molecular Ecology* 17: 4134–4150.
- Tollefsrud MM, Sonstebo JH, Brochmann C, Johnsen O, Skroppa T, and Vendramin GG (2009) Combined analysis of nuclear and mitochondrial markers provide new insight into the genetic structure of North European *Picea abies*. *Heredity* 102: 549–562.
- von Post L (1924) Ur de sydsvenska skogarnas regionala historia under postarktisk tid. *Geologiska Föreningen Förhandlingar* 46: 83–128.
- Willis KJ, Braun M, Sumegi P, and Toth A (1997) Does soil change cause vegetation change or vice versa? A temporal perspective from Hungary. *Ecology* 78: 740–750.

The Biome Approach to Reconstructing Past Vegetation

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Introduction: Reconstructing Late Quaternary Vegetation at Large Spatial Scales

When vegetation change is viewed at continental scales, dynamic biogeographic patterns become apparent. Individual taxon ranges shift markedly and whole vegetation formations appear, fragment, or disappear through time. If we assume that large-scale distribution patterns of vegetation are, on millennial timescales, mainly controlled by climate, such changes are likely to reflect variations in key atmospheric circulation features, such as the spatial extent of a monsoon system or the average position of the polar jet and associated storm tracks. These in turn are affected, for example, by orbital variations or the presence of a large ice sheet. Features at this scale are those most effectively simulated by general circulation models (GCMs). Given the present need to model climate reliably in relation to increasing levels of atmospheric CO₂ (Randall et al., 2007), it is important to know how well climate models simulate climates different from those of the present day. Climate model performance can be assessed through the comparison of vegetation predicted from a paleoclimate simulation with actual past vegetation as recorded by paleodata. The approach was first developed for the COHMAP (1988) project and has evolved into the biome approach to vegetation reconstruction (Prentice and Webb, 1998).

This article explains the basis of 'biomization' (Prentice et al., 1996), which uses a globally applicable algorithm to generate maps of vegetation biomes from large arrays of pollen data; it then summarizes the results of the BIOME 6000 project (Prentice and Webb, 1998), which mapped vegetation at 0, 6000, and 18000 ¹⁴C years BP. Further sections discuss insights gained into Quaternary plant biogeography, illustrate model-data comparisons and briefly discuss new related approaches to vegetation reconstruction.

Objective Vegetation Mapping

It is difficult to make a map of modern global vegetation using a consistent and repeatable procedure, and mapping past vegetation is even more challenging. Disparate interpretations of past vegetation due to differences in regional approaches and/or paleoecological methodology make inter-regional compilations of interpreted data difficult, and attempts make global maps of paleovegetation have been subjective and hard to replicate. The biome method provides some distinct advantages over other approaches. It uses an objective and repeatable method to group pollen assemblages (the primary data) directly into globally consistent vegetation types (biomes). These reflect the structure and function of constituent taxa rather than their floristic identity. Maps for a given time slice (e.g., 6000 years BP) are created by assigning a dot, color coded for a particular biome, to each site that has a record. Such maps have relatively coarse spatial detail, but

they are generally preferable to those created by subjective interpolation among data points, especially if the latter mask areas of poor data coverage.

Paleoclimate Reconstructions

The biome method also provides a robust approach to investigate past climate. Reconstructions of past climate from paleodata usually follow the taxon-based transfer-function approach in which numerical relationships are developed between (for example) modern pollen abundances and observed climate variables. The numerical function is applied to fossil pollen assemblages and past climate variables are calculated. This method is constrained by observable modern relationships. In the past, certain taxa occurred in abundances observed nowhere today, and fossil pollen spectra may have no recognizable modern counterpart – so-called 'no-analogue' assemblages (Williams and Jackson, 2007). Once outside the envelope of modern observations, fossil pollen spectra, described in terms of taxon abundance, cannot be reliably related to past climate. The problem becomes acute for times prior to the Holocene, when the climate system and global vegetation were markedly different from today. The biome approach circumvents this problem: plant-climate relationships are described with reference to plant structure and function and are not dependent upon the fossil pollen assemblages having modern counterparts. The vegetation distributions thus obtained can be checked for concordance with those derived from climate-model simulations using the same bioclimatic rules.

The Biome Approach

PFTs and Climate

The distributions of plant species are partly controlled by bioclimatic factors, for example, their susceptibility to heating, cooling, or frost, and a minimum requirement for growing-degree-days. Various classification systems for global vegetation are based on this principle, for example, Holdridge's (1967) life zones. Associated with these bioclimatic responses are morphological and physiological plant traits that equip plants to succeed in certain environments but usually mean they are less suited to others. For example, in the cold conditions of the far North growing-degree-days are below the minimum for the growth of trees. All Mediterranean floras are characterized by species that are drought- and fire-tolerant, although there is little overlap in floristic composition among the Mediterranean regions of the world (Archibald, 1995). Thus global vegetation can be described according to the plant functional types (PFTs) that dominate in various types of climatic conditions, and these PFT assemblages can be classed as formations or biomes, such as arctic tundra or warm temperate evergreen sclerophyll forest. In this type of

classification, species identity is not paramount. Furthermore, the known bioclimatic tolerances of PFTs theoretically allow, and do indeed predict, novel admixtures – no-analogue biomes – under certain climatic conditions.

Climate is, of course, not the only factor affecting plant distribution. Disturbance regimes, differing growth strategies, and interspecific competition also play a role in determining which PFTs dominate a region. For example, drought, fire, and regeneration potentially all affect the balance of grassland and woodland at the moister borders of Savanna and temperate grassland regions (e.g., Scott, 1995). Soil and topography also influence vegetation, occasionally at regional scales, but more often on a local basis that is not resolved in continental-scale reconstructions.

Modeling Vegetation

If bioclimatic requirements and growth rates, regeneration, and competition are taken into account, it is possible to model modern global vegetation effectively. Prentice et al. (1992) developed a global vegetation model (BIOME) that incorporated these principles. BIOME can be fed data from modern climate observations or a climate-model simulation and will produce a vegetation map for those conditions (see Figure 1). In the case of simulations, these may also be paleo-simulations or simulations based on future projections of climate. Later, improved versions of BIOME account for variations in atmospheric CO₂ and other parameters that have a major effect on plant growth and distribution (see, e.g., Kaplan et al., 2003). It is an equilibrium model; for a prescribed climate it generates an optimal vegetation map based on a global grid of points, but it does not incorporate any subsequent interactions of climate and vegetation. More recently dynamic vegetation models such as the Lund–Potsdam–Jena model (Sitch et al., 2003) have been developed and interactive vegetation packages are now incorporated into GCMs. In this way, feedbacks related to vegetation–atmosphere interactions (such as albedo and moisture fluxes) can be modeled. As they are modified and improved, climate models require continual

evaluation. Mapped paleodata play a central role because they can be directly compared with simulated vegetation maps that are in turn driven by climate-model output in so-called data-model comparisons (see below).

Classifying and Mapping Pollen Data as Biomes – Biomization and the BIOME 6000 Project

In a global vegetation reconstruction, biomes should have distinct and realistic ecological qualities, and they must also be distinguishable from other biomes by pollen assemblage composition. In order to compare observed past vegetation with simulated past vegetation, biomes should be consistent with those defined in a vegetation model (such as BIOME). Although these requirements constrain the number of biomes that can be defined, this is not a problem – too many, and the biomization and resultant maps become unnecessarily complicated. Conversely, information may be lost if there are too few biomes. The ideal number depends upon the questions being asked of the data. For example, 18 biomes were used in the BIOME 6000 overview reconstructions for the northern continents and Africa (Prentice et al., 2000), and these included one tundra biome. In contrast, for a more detailed investigation of the Arctic, Bigelow et al. (2003) defined five separate tundra biomes.

Pollen-based biomes are defined using the biomization approach (Prentice and Webb, 1998; Prentice et al., 1996). Because pollen assemblages are conventionally reported as frequency data, the percentage values of pollen taxa are used as the basis of the biomization procedure. An algorithm that relates the taxon abundances in a pollen spectrum to PFTs and PFTs to biomes is then applied to each fossil pollen spectrum from a region of interest. The biome that has the highest resultant score is selected in each case (see Figure 2 for further details).

The biomization of modern pollen samples is validated by comparison with descriptions of modern vegetation. This is not as straightforward as it may seem, because regional vegetation maps are not easily amalgamated, nor are different

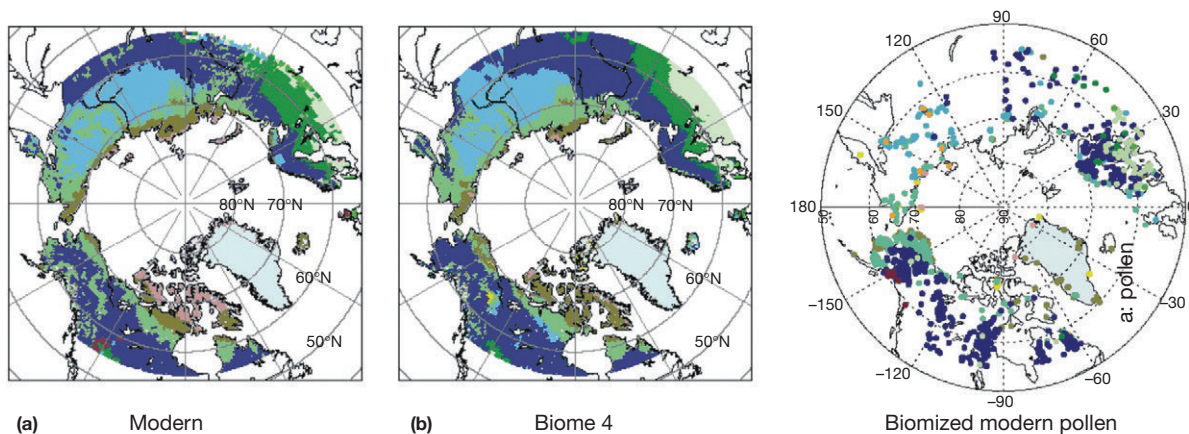


Figure 1 A comparison of mapped modern vegetation for the Arctic and sub-Arctic, the vegetation simulated by the BIOME 4 using a modern climate dataset as input and modern biomes based on biomized pollen spectra. Reproduced from Kaplan JO, Bigelow NH, Bartlein PJ, et al. (2003) Climate change and arctic ecosystems II: Modeling paleodata-model comparisons, and future projections. *Journal of Geophysical Research* 108(D19): <http://dx.doi.org/10.1029/2002JD002559>; Bigelow NH, Brubaker LB, Edwards ME, et al. (2003) Climate change and arctic ecosystems I: Biome reconstructions of tundra vegetation types at 0, 6, and 18 radiocarbon ka in the Arctic. *Journal of Geophysical Research* 108(D19): <http://dx.doi.org/10.1029/2002JD002558>.

Steps in biomization

1. Assemble and check original pollen data, then convert to %
2. Define a set of biomes (regional or global coverage)
3. Define a set of plant functional types (PFTs)
4. Define the biomes by their constituent PFTs
5. Assign all recorded pollen taxa to one or more PFTs

A simple example with 2 taxa, 3 PFTs, and 2 biomes is shown here

Taxon		PFT	
<i>Betula</i>			
<i>Picea</i>			
		Boreal-arctic deciduous broadleaved shrub: PFT contributes to the shrub-tundra biome	Boreal summergreen deciduous broadleaved tree: PFT contributes to the taiga (boreal evergreen forest) biome
		Boreal evergreen needle-leaved tree: PFT contributes to the taiga biome	

6. Assign taxa to biomes in a taxon-biome matrix (presence/absence)

	Taiga biome	Tundra biome
<i>Betula</i>	1	1
<i>Picea</i>	1	0

Example pollen values:
Betula 70%
Picea 30%

7. Calculate affinity score of each taxon for each biome (the following is taken from Prentice and Webb 1998,

$$A_{ik} = \sum_j \delta_{ij} \sqrt{\{ \max[0, (p_{jk} - \theta)] \}}$$

Where A_{ik} is the affinity of pollen sample k for biome i ; summation is over all taxa j ; δ_{ij} is the entry (0 or 1) in the taxon-biome matrix; p_{jk} are pollen percentage values; θ is a threshold pollen percentage.

Underlying this equation is the selection of taxa (taxon-biome matrix entries), the summing of the square roots of the percent values of taxa in a given biome, and the chance to set the threshold percent >0 , that is, to exclude very small pollen values that may be due to long-distance transport of grains from beyond the region of interest.

Assuming the threshold is zero for convenience, $\sqrt{70} = 8.36$ and $\sqrt{30} = 5.47$; therefore the total affinity score for shrub tundra is **8.36** (from *Betula* via the shrub PFT) and the total affinity score for taiga is $8.36 + 5.47 = \mathbf{13.84}$ (from *Betula* via the deciduous tree PFT and *Picea* from the evergreen tree PFT). The assigned biome in this simple example is therefore taiga (evergreen boreal forest)

Figure 2 Steps in biomizing a simple pollen spectrum.

global vegetation maps mutually consistent (Prentice and Webb, 1998). Biomizations have been validated for several continents, using regional maps and/or vegetation data reported with the modern pollen samples (e.g., Bigelow et al., 2003; Figure 1). Visual inspection of the maps shows generally good agreement, although there is some level of error in biome assignments. In some regions, modern vegetation reflects considerable alteration by humans, but even in these areas modern biomizations appear 'realistic' in relation to our understanding of potential natural vegetation (Prentice et al., 1996).

The BIOME 6000 Project

The BIOME 6000 project has produced pollen-based biome maps for much of the globe for 0 and 6000 ^{14}C years BP, and

there are also maps for 18 000 ^{14}C years BP for many regions. The exercise involved hundreds of participants worldwide. Regional experts screened radiocarbon-dated pollen data for taxonomic and chronological reliability, related pollen taxa to one or more PFTs, and assigned each PFT to one or more biomes. Fossil pollen spectra dated to within 500 radiocarbon years of 6000 ^{14}C years BP and within 1000 years of 18 000 ^{14}C years BP were then biomized and the results presented as maps of past biome distributions. Table 1 summarizes the publications in which biome reconstructions can be found; for a useful overview of the climatological and biogeographic implications of the past biome distributions see Prentice et al. (2000). Figure 3 presents biome maps for 0, 6000, and 18 000 ^{14}C years BP. The next section describes some of the prominent biome shifts and mentions probable climatic causes.

Table 1 Published biome reconstructions for 0, 6, and 21 ka

<i>Region/Topic</i>	<i>Publication</i>
Overview of the BIOME 6000 project and methods	Prentice and Webb (1998)
Overview of the northern continents and Africa 0, 6, and 18 ka	Prentice et al. (2000)
Africa and Arabia at 0 and 6 ka	Jolly et al. (1998)
Australia, SE Asia and the Pacific at 0, 6, and 18 ka	Pickett et al. (2004)
The Arctic 0,6, and 18 ka	Bigelow et al. (2003)
Beringia at 0, 6, and 18 ka	Edwards et al. (2000)
Black Sea Corridor	Cordova et al. (2009)
China at 0 and 6 ka	Yu et al. (1998)
China at 6 ka, 18 ka	Yu et al. (2000)
Europe (Southern) and Africa at 18 ka	Elenga et al. (2000)
Former Soviet Union and Mongolia at 0 and 6 ka	Tarasov et al. (1998)
Northern Eurasia at 18 ka	Tarasov et al. (2000)
Japan at 0, 6, and 18 ka	Takahara et al. (2000)
North America (West) at 0, 6, and 21 ka	Thompson and Anderson (2000)
North America (East) late Quaternary	Williams et al. (2000)

Two issues of the *Journal of Biogeography* contain most of the BIOME 6000 vegetation reconstructions for different regions at 0, 6, and 21 ka: Volume 25, No. 6 (November 1998) and Volume 27, No. 3 (May 2000).

Global Vegetation Patterns at 18 000 and 6000 ¹⁴C years BP (18 and 6 ka)

The Last Glacial Maximum

The Last Glacial Maximum (LGM; 18 000 ¹⁴C years BP or 18 ¹⁴C ka) represents an interval when insolation patterns (as controlled by orbital cycles) were similar to present, but in almost every other aspect conditions were markedly different. Ice sheets were extensive across northern land regions, oceans were cooler, atmospheric dust loadings were higher, and CO₂ levels were lower (COHMAP Members, 1988; Jansen et al., 2007). All these features contributed to global climate and growth conditions that contrast greatly with those of today. Paleocological data for the LGM are relatively sparse compared with the data available for the Holocene, but their synthesis nevertheless reveals useful patterns.

The northern continents

Much of the Arctic region was under ice at the LGM. To the South of the ice and across unglaciated Beringia (North-West Canada, Northern Alaska, and North-East Siberia) tundra vegetation dominated (Edwards et al., 2000). Boreal forest was displaced to more southerly latitudes, and its overall extent contracted. In Eurasia particularly, steppe vegetation was far more extensive than today (Tarasov et al., 2000). In the mid-latitudes of Europe temperate deciduous forests and Mediterranean forests and woodlands were also greatly reduced and in many areas replaced by steppe (Elenga et al., 2000). In China, total site coverage is sparse, but a marked reduction of forest over much of Central China is evident in the data (Yu et al., 1998). The predominance of treeless biomes reflects heightened aridity as well as lower temperatures.

The lower latitudes and Southern Hemisphere

In the tropics and subtropics the data are sparser than for northern temperate regions, but they nevertheless clearly indicate major changes in vegetation cover, particularly the fragmentation and reduction in area of tropical moist forest in Africa (Prentice et al., 2000). The desert regions of South-West North America were characterized by conifer woodland; this has been shown to be the consequence of winter storms and rain linked to a southward displacement of the jet stream that was in turn due to intense high pressure over the Laurentide Ice Sheet (COHMAP Members, 1988; Thompson and Anderson, 2000). Globally, the extents of most other desert regions were similar to or greater than today's, reflecting generally greater aridity under cooler global climates. In Australia, the LGM vegetation was characterized by increased dominance of xerophytic PFTs over today, both in the southeast and the northern tropical region, indicating greater aridity (Pickett et al., 2004).

The mid-Holocene

During the middle Holocene (conventionally represented by 6000 ¹⁴C year BP or 6 ka in model simulations) Northern Hemisphere summer insolation was considerably greater than modern. By this time the waning northern ice sheets approximated their modern extent, so there was virtually no influence of residual Pleistocene ice on regional climates (eastern Canada is the one exception – see below). As Prentice and Webb (1998) point out, 6 ka does not represent a global climatic 'optimum' or even 'maximum.' The term 'climatic optimum' is ecologically meaningless because different plant species have different climatic requirements, nor was the warmest period of the Holocene experienced at the same time in all regions. Nevertheless, global average temperatures were most likely 1–2 °C warmer than present, and this makes 6 ka a particularly interesting period to study. The site coverage is generally good, and biomized data allow quite detailed comparisons of modern and 6 ka vegetation distributions.

The northern continents

The arctic treeline advanced northward by up to a few hundred kilometers over its current position in some regions, such as Central Siberia, reflecting warming due to heightened summer insolation. In contrast, forest in Eastern Canada was displaced to the south of its present limits as a result of residual Laurentide ice over Labrador (Bigelow et al., 2003; Dyke, 2004). In temperate Europe, thermophilous temperate forest expanded north of its current position and also southward into Mediterranean regions (Elenga et al., 2000). Steppe vegetation expanded slightly in North America, compared with current limits, but in China this did not happen; instead an enhanced monsoon due to higher summer insolation likely favored the broad distribution of temperate deciduous forest (Yu et al., 2000). In Japan, temperate coniferous forest was prominent, possibly favored over today's broadleaved evergreen/mixed forests by increased seasonality driven by warmer summers (Takahara et al., 2000).

The lower latitudes and Southern Hemisphere

Striking differences in vegetation cover between 6 ka and present are observed in the subtropics, notably in Africa,

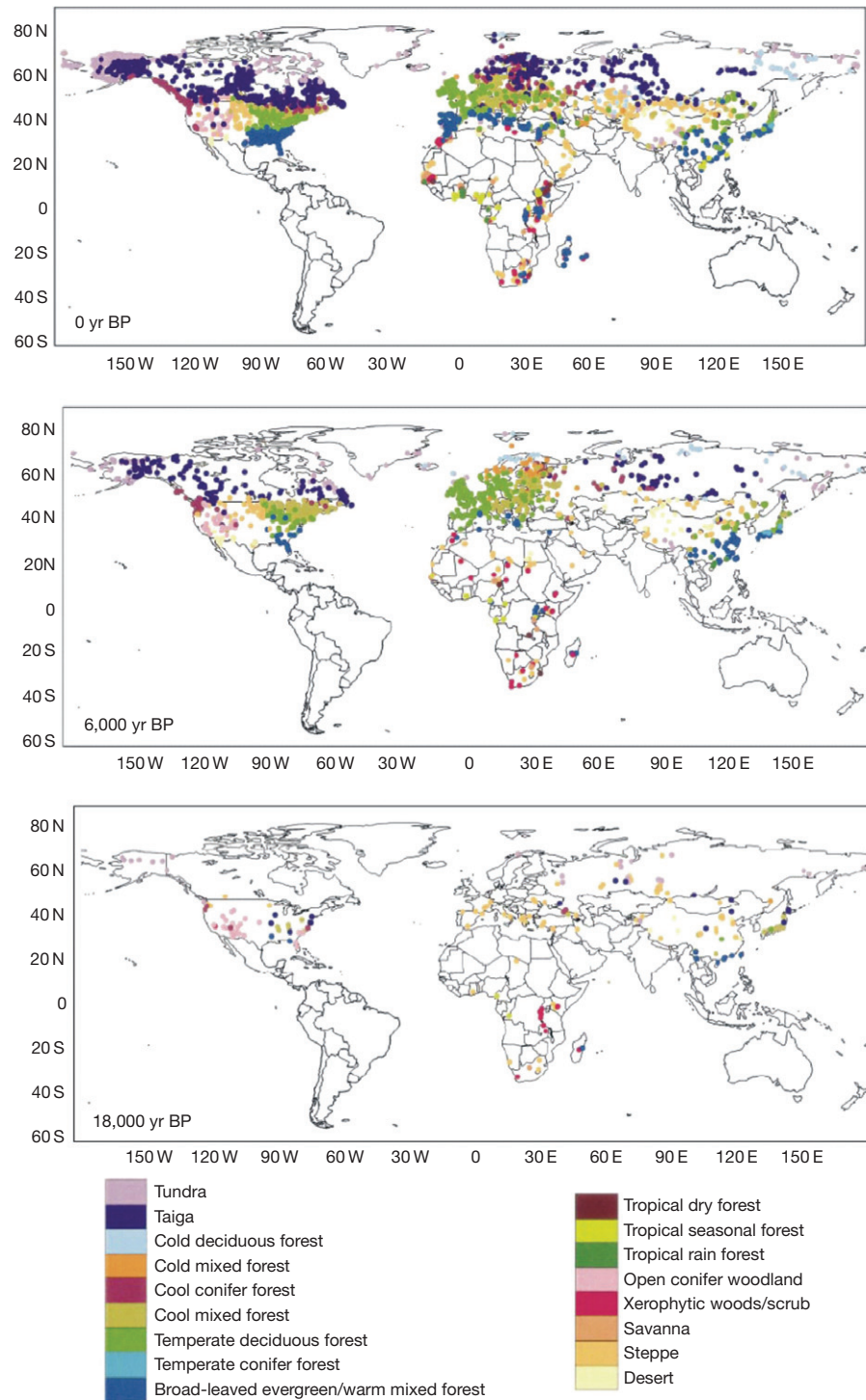


Figure 3 Biome maps for the northern continents and Africa: LGM and mid-Holocene. From Prentice IC, Jolly D, and BIOME 6000 participants (2000) Mid-Holocene and glacial-maximum vegetation geography of the Northern continents and Africa. *Journal of Biogeography* 27: 507–519. Biomes for Australia not shown. Pickett EJ, Harrison SP, Hope G, et al. (2004) Pollen-based reconstructions of biome distributions for Australia, Southeast Asia and the Pacific (SEAPAC region) at 0, 6000 and 18,000 ^{14}C yr BP. *Journal of Biogeography* 31: 1381–1444.

where the biomization suggests that much of the modern Sahara was vegetated by grasslands and scrub-woodland (Jolly et al., 1998; Prentice et al., 2000). This feature appears to be related to a far stronger monsoon driven by summer insolation that was greater than present. However, tropical moist forest extent may have been less than

present; the enhanced monsoon further North may have been linked to less consistent rains near the Equator (Prentice et al., 2000). Relatively small differences between modern and 6 ka vegetation in Australia appear largely to reflect regional variations in moisture availability (Pickett et al., 2004).

Historical Biogeography: Migrations, Refugia, and Novel Biomes

BIOME 6000's global paleovegetation maps illustrate the dynamic nature of late-Quaternary biomes. The synthesized datasets can be used to investigate whether biome distributions changed in response to climate change, or whether biomes were fragmented or even disappeared entirely. Furthermore, in-depth examination of the taxon contributions to various biomes reveals considerable variability in biome composition through time.

Migration Versus Refugia

Temperate deciduous forest and tropical moist forest were both greatly fragmented during the LGM. Nevertheless, their constituent taxa survived and later assembled to form widespread forest biomes in the Holocene. For example, the deciduous forest was virtually eliminated from mainland China (Yu et al., 1998). Prentice et al. (2000) argue that while it is possible that the biome was displaced onto continental shelf areas that were subsequently inundated, the pollen data show that deciduous tree taxa also survived scattered within other biomes that were more widespread in the LGM, such as broadleaved evergreen forest. Thus *in situ* survival may have been as important as migration. Understanding how and where taxa survived major environmental changes and what ecological and evolutionary processes affected them is of relevance to evolutionary biology and conservation biology (see Harrison et al., 2001; McMahon et al., 2011; Qian and Ricklefs, 2000). Species distributions are once again being radically altered, this time by direct human action (such as transformation of ecosystems for agriculture and forestry), and contemporary biomes are threatened with disruption by climate change associated with global warming.

Temporal Changes in the Composition of Biomes

Steppe and tundra in the LGM

In certain cases, the biome assignment is sensitive to how a few 'indicator' taxa are assigned to PFTs (Prentice et al., 2000). This is the case for steppe and tundra biomes across Eurasia and Beringia in the LGM (Edwards et al., 2000; Tarasov et al., 2000). In Alaska and North-West Canada, the LGM pollen assemblages are assigned to tundra, but those further West and South in Eurasia are assigned to steppe, whereas sites in between are sensitive to the details of the biomization. Today, boreal forest tends to separate steppe and tundra geographically (see, e.g., Scott, 1995) and, for the most part, steppe and tundra are well differentiated functionally and floristically. However, if we explore the potential climate space (seasonal conditions of temperature and moisture) occupied by these biomes, we find that under conditions that are too cold and/or dry for forest (as in the LGM), steppe and drier forms of tundra are adjacent biomes and share similar dominant PFTs (grasses and forbs; Kaplan et al., 2003; Figure 4). Thus, underlying the instability in the biomization of the steppe-tundra transition in Eurasia is the fact that tundra and steppe are closely related (in terms of PFTs and geographic distribution) under certain configurations of climate.

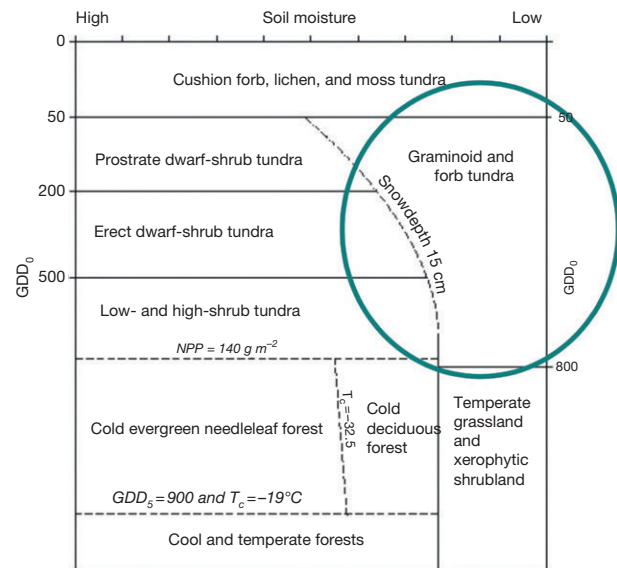


Figure 4 Diagram after Kaplan et al. (2003) showing the climate space for northern biomes. Grass-forb tundra occurs in dry climates that are too cool for true steppe.

Paleoecologists have long questioned the nature of the treeless LGM vegetation of Beringia. It is associated with the presence of diverse Pleistocene megafauna including both herbivores and carnivores and must, therefore, have been productive enough to sustain this extensive faunal ecosystem (see Hopkins et al., 1982). The dominant pollen taxa, comprising grasses, sedges, and *Artemisia* (sagebrush, mugwort), have been interpreted as indicating either tundra or steppe (see above). Neither interpretation is convincing, because debate has focused on the modern properties of tundra and steppe ecosystems. Neither appears a suitable analogue for past conditions. Most modern tundra types are moist and dominated by woody plants that would not support large populations of specialist grazers, and modern steppe grows under conditions that are far warmer than those of the LGM in Beringia.

The refocusing of attention on which PFTs occur in cool, dry climate space, and the possibility that in the LGM Beringian 'tundra' was quite unlike contemporary tundra in both PFT composition and ecological properties, offers a promising solution to a long-standing biogeographic conundrum. In their biomization of northern tundra and boreal ecosystems, Bigelow et al. (2003) defined a grass- and forb-dominated tundra biome (GFT) associated with cool, dry climate space. Such vegetation is rare today in the Arctic, but where it does occur it is associated with locally xeric conditions (Yurtsev, 1982). In the biomization, many LGM pollen spectra across Beringia were assigned to this biome. In addition, the BIOME model output based on LGM paleoclimate simulations showed a high level of productivity for GFT, suggesting that it may have been able to support LGM herbivores (Kaplan et al., 2003). The lesson to be learned here is that our conceptions of biomes rely almost entirely on those that exist currently, whereas we should expect both variability in biomes through time and

different regional sets of biomes during the highly variable climatic conditions of the Quaternary.

Late Quaternary biomes of Eastern North America

Although the BIOME 6000 project focused on just two past time slices, in part because these times were the foci for coordinated climate-model experiments, more highly resolved histories of biome distribution during the late Quaternary are feasible. Williams et al. (2004) produced maps of biome distributions from the LGM to present at 1000-year time slices for Eastern North America (Figure 5). The maps' fine detail is based on a conservative, quantitative interpolation technique and thus the actual data points are obscured. This format has the advantage of illustrating clearly the continual shifting of biome distributions through time and the marked change in dominant biomes from the Late Glacial to the Holocene, but the details are likely to be inaccurate. Williams et al. (2004) also analyzed the changing composition of given biomes through time. For example, cool mixed forest, characterized by conifers, broadleaved trees and shrubs, and herbs, shows changing frequencies of these types through time. In particular, conifers dominate LGM and Late Glacial records, whereas after ca. 6 ka, broadleaved taxa are more abundant than conifers. Among the conifers, *Pinus* (Pine) and *Picea* (Spruce) dominate LGM and Late Glacial records, whereas *Tsuga* (Hemlock) dominates in the later Holocene (Figure 6). Biomizing, by definition, masks changes in taxonomic composition because the PFT–biome relationship drives the method. However, 'deconstruction' of biome composition provides interesting ecological insights and raises questions about the degree to which climate and non-climatic factors each determine these intra-biome changes.

Climate-Model Evaluations Using Biomized Data

As mentioned above, a key aim of the BIOME 6000 project is to produce mapped past vegetation at a global scale in order to evaluate the performance of climate models. It is assumed that the biome configurations of 6000 and 18000 ^{14}C year BP reflect the climate prevailing at those times, that is, large-scale patterns of vegetation are in equilibrium with the observed climate – an assumption that is not unreasonable for these two key times (Prentice and Webb, 1998). Reproduction of those configurations by a vegetation model such as BIOME driven by climate-model simulations is most parsimoniously interpreted as reflecting a successful simulation. If simulation and paleodata disagree, both must be revisited to assess where errors lie.

The Arctic Treeline

This exercise is usefully illustrated by the status of arctic treeline (i.e., the northern limit of the boreal forest) at 6 ka. As mentioned above, the biomized 6 ka treeline is strongly asymmetric latitudinally. A mid-Holocene northward advance might be expected with greater summer insolation, particularly in areas of continental climate, such as Siberia. The model simulations adopted in the data-model comparison by

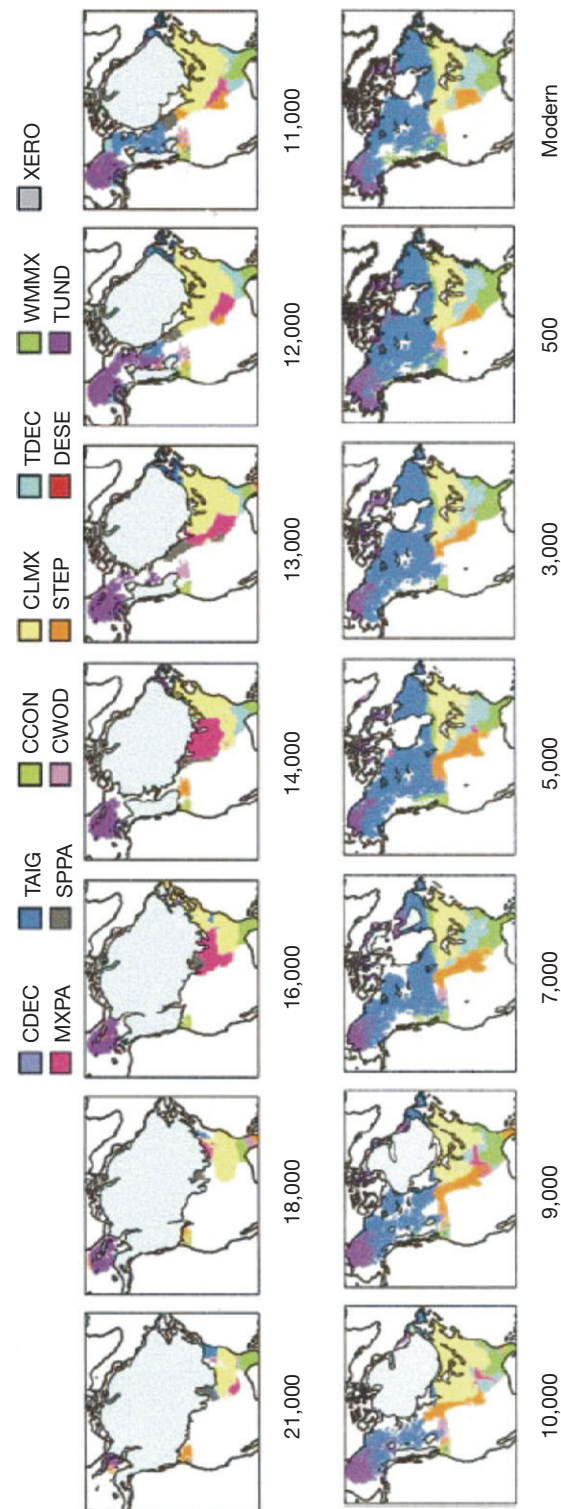


Figure 5 Biome maps for North America at 1000-year time slices, from Williams JW, Shuman BN, Webb T, Bartlein PJ, and Leduc PL (2004) Late-Quaternary vegetation dynamics in North America: Scaling from taxa to biomes. *Ecological Monographs* 74: 309–334. Note that dates are calibrated to calendar years BP. Key to biomes: CCON, cool conifer forest; CDEC, cold deciduous forest; CLMX, cool mixed forest; CWOD, conifer woodland; DESE, desert; MXPA, mixed parkland; SPPA, spruce parkland; STEP, steppe; TAIG, taiga; TDEC, temperate deciduous forest; TUND, tundra; WMMX, warm mixed forest; XERO, xerophytic scrub.

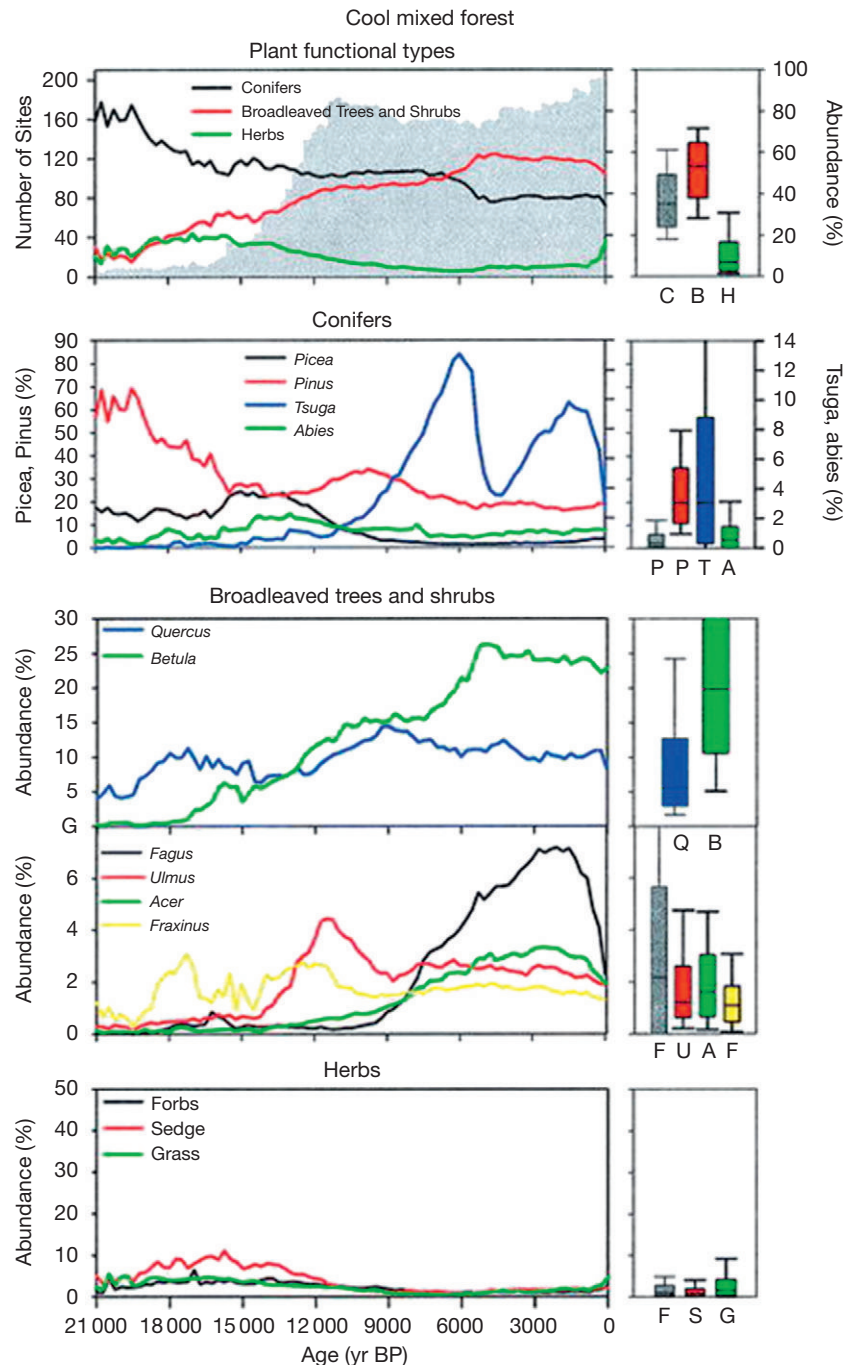


Figure 6 Changes in biome composition through time for North American biomes, from Williams JW, Shuman BN, Webb T, Bartlein PJ, and Leduc PL (2004) Late-Quaternary vegetation dynamics in North America: Scaling from taxa to biomes. *Ecological Monographs* 74: 309–334. The graphs show long-term means of pollen percentages (for PFTs or taxa) for all sites assigned to the cool mixed forest biome (total number of sites shown by the gray histogram in the uppermost graph).

Kaplan et al. (2003) drive such a response in the BIOME model; they also show the treeline displaced northward in Canada, in contrast to the data. This data–model discrepancy can be explained by the last remaining Laurentide ice, which physically prevented the northward advance of vegetation as well as creating regionally cool conditions (Dyke, 2005). This feature was not included in the modeling, and the southward treeline displacement was not simulated.

The West African Monsoon at 6000 ^{14}C years BP

The ‘greening’ of the Sahara is one of the most striking features of vegetation in the Mid Holocene (see above). For some time it has been clear that the primary driver of this feature was the enhanced monsoon related to higher summer insolation in the Northern Hemisphere (e.g., COHMAP Members, 1988). Data on the biomes of the Sahel and Sahara regions were used as benchmarks against which to test model simulations. Could

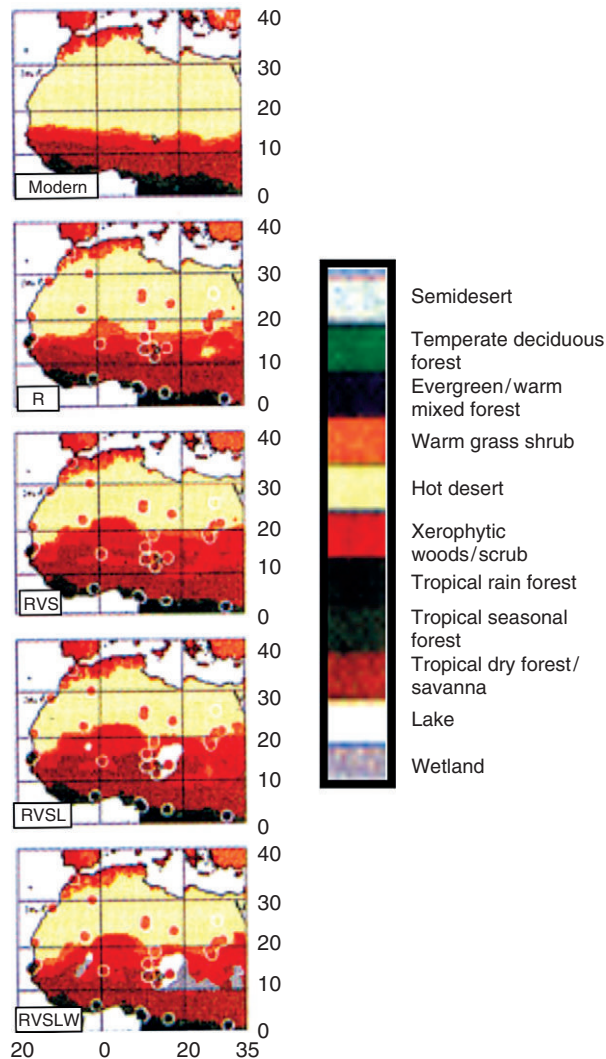


Figure 7 Data-model comparison for West Africa at 6 ka from Broström A, Coe M, Harrison SP, et al. (1998) Land surface feedbacks and paleomonsoons in Northern Africa. *Geophysical Research Letters* 25: 3615–3618. The shading shows simulated vegetation according to the biomes described in the color key. The top panel shows the modern control. All other panels show a different 6-ka simulation. In these panels the paleodata data points are indicated by small circles (same color code), and the simulated vegetation is shaded. R = orbital forcing only; RVS adds vegetation feedback; RVSL adds prescribed lakes (including mega-lake Chad, see [Africa during the Late Quaternary](#)); and RVSLW adds prescribed wetland areas. The simulated climate becomes progressively wetter further North with the addition of more feedbacks into the model, and simulated vegetation better matches (but not perfectly) the observed vegetation.

models simulate enough precipitation far enough north to support the observed biomes of xerophytic woodland and grassland? Until recently paleoclimate models incorporating the insolation anomaly but no surface feedbacks consistently failed to simulate the full magnitude of the feature. Models that encompass feedbacks, for example, the changed vegetation itself and other properties of the land surface such as soil moisture and wetland extent, performed better than those without the feedbacks (Broström et al., 1998; Kohfeld and Harrison, 2000; [Figure 7](#)). A new generation of models

that include both land feedbacks and, importantly, ocean feedbacks, perform far better at simulating moisture Northward into the Sahara ([Braconnot et al., 2007](#)).

Northern Hemisphere Continental Aridity at 6000 ¹⁴C years BP

Wohlfahrt et al. (2004, 2008) analyzed feedbacks due to ocean and vegetation changes for the mid-Holocene as depicted in coupled ocean-atmosphere model simulations with dynamic vegetation. Their studies, which used a large biomized dataset, demonstrate how data-model comparisons using biomized data can be used to diagnose model performance. Although several large-scale features of Northern Hemisphere biome patterns corresponded well with simulations, the increase in simulated mid-continental aridity in Eurasia was not supported by the data. In this case, the atmospheric simulations alone generated xerophytic vegetation in the Asian interior, and this was enhanced by ocean feedback. The vegetation drove positive feedback towards aridity, apparently by the over-simulation of evapotranspiration. The mismatch suggests inconsistencies within those parts of models dealing with interactions between land-surface and atmosphere.

Current Developments and Future Directions

The biome approach was developed to examine continental and global vegetation patterns and as such has limitations as a tool to describe detailed vegetation history at a regional scale or smaller. It is used most successfully at these larger scales. To ensure global compatibility, taxon-PFT and PFT-biome relationships must be universally agreed, and although the method has proved remarkably successful in general, further refinements are possible. One is the incorporation of macrofossil data, which is currently an under-used paleodata resource. In Western North America, for example, where arid conditions often preclude pollen preservation, data from pack-rat middens have been used to good effect in macrofossil-based biomizations ([Thompson and Anderson, 2000](#)). Another approach, particularly promising when the focus is on paleoclimate, is to use plant traits to map vegetation. Traits can be directly selected for their bioclimatic significance, without the need to decide upon one or more PFTs for each taxon. See [Barboni et al. \(2004\)](#) for a demonstration of the utility of this approach and [Harrison et al. \(2010\)](#) for progress towards a global traits classification. Biomization has recently been combined with pollen-source theory ([Sugita, 2007](#)) to improve pollen-based estimations of treeline position ([Binney et al., 2011](#)). This approach may also be applicable to the broad-scale identification of anthropogenically modified vegetation, which is not reliably detected by the original biomization procedure.

See also: Plant Macrofossil Introduction; Pollen Analysis, Principles. Paleobotany: Paleophytogeography. Paleoclimate: Introduction. Paleoclimate Modeling: Last Glacial Maximum GCMs. Plant Macrofossil Methods and Studies: Treeline Studies. Pollen Methods and Studies: Changing Plant Distributions and Abundances; Reconstructing Past Biodiversity Development.

References

- Archibold OW (1995) *Ecology of World Vegetation*. Heidelberg: Springer 528 pp.
- Barboni D, Harrison SP, Bartlein PJ, et al. (2004) Relationships between plant traits and climate in the Mediterranean region: A pollen data analysis. *Journal of Vegetation Science* 15: 635–646.
- Bigelow NH, Brubaker LB, Edwards ME, et al. (2003) Climate change and arctic ecosystems I: Biome reconstructions of tundra vegetation types at 0, 6, and 18 radiocarbon kyr in the Arctic. *Journal of Geophysical Research* 108(D19) <http://dx.doi.org/10.1029/2002JD002558>.
- Binney HA, Gething PW, Nield JM, Sugita S, and Edwards ME (2011) Treeline identification from pollen data: Beyond the limit? *Journal of Biogeography* 38: 1792–1806.
- Braconnot P, Otto-Bliesner B, Harrison S, et al. (2007) Results of PMIP2 coupled simulations of the Mid-Holocene and Last Glacial Maximum – Part 1: Experiments and large-scale features. *Climate of the Past* 3: 261–277.
- Broström A, Coe M, Harrison SP, et al. (1998) Land surface feedbacks and paleomonsoons in Northern Africa. *Geophysical Research Letters* 25: 3615–3618.
- COHMAP Members (1988) Climatic changes of the last 18 000 years: Observations and model simulations. *Science* 241: 1043–1052.
- Cordova CE, Harrison SP, Mudie PJ, Riehl S, Leroy SAG, and Ortiz N (2009) Pollen, plant macrofossil and charcoal records for palaeovegetation reconstruction in the Mediterranean – Black Sea Corridor since the Last Glacial Maximum. *Quaternary International* 197: 12–26.
- Dyke AS (2004) An outline of North American deglaciation with emphasis on Central and Northern Canada. In: Ehlers J and Gibbard PL (eds.) *Quaternary Glaciation – Extent and Chronology, Part II*, pp. 373–424. The Netherlands: Elsevier.
- Dyke AS (2005) Late Quaternary vegetation history of Northern North America based on pollen, macrofossil, and faunal remains. *Géographie physique et Quaternaire* 59: 211–262.
- Edwards ME, Anderson PM, Brubaker LB, et al. (2000) Pollen-based biomes for Beringia 18 000, 6000 and 0 ¹⁴C yr BP. *Journal of Biogeography* 27: 521–554.
- Elenga H, Peyron O, Bonnefille R, et al. (2000) Pollen-based biome reconstructions for Southern Europe and Africa 18 000 yr BP. *Journal of Biogeography* 27: 621–634.
- Harrison SP, Prentice IC, Barboni D, Kohfeld KE, Ni J, and Sutra JP (2010) Ecophysiological and bioclimatic foundations for a global plant functional classification. *Journal of Vegetation Science* 21: 300–317.
- Harrison SP, Yu G, Takahara H, and Prentice IC (2001) Diversity of temperate plants in East Asia. *Nature* 413: 129–130.
- Holdridge LR (1967) *Life Zone Ecology*. San Jose: Tropical Science Center.
- Hopkins DM, Matthews JV Jr., Schweger CE, and Young SB (eds.) (1982) *Paleoecology of Beringia*, pp. 3–28. New York, NY: Academic Press.
- Jansen E, Overpeck J, Briffa KR, et al. (2007) Palaeoclimate. In: Solomon S, Qin D, and Manning M, et al. (eds.) *Climate Change 2007: The Physical Science Basis*, Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change, Cambridge: Cambridge University Press.
- Jolly D, Prentice IC, Bonnefille R, et al. (1998) Biome reconstruction from pollen and plant macrofossil data for Africa and the Arabian peninsula at 0 and 6 ka. *Journal of Biogeography* 25: 1007–1028.
- Kaplan JO, Bigelow NH, Bartlein PJ, et al. (2003) Climate change and arctic ecosystems II: Modeling paleodata-model comparisons, and future projections. *Journal of Geophysical Research* 108(D19) <http://dx.doi.org/10.1029/2002JD002559>.
- Kohfeld KE and Harrison SP (2000) How well can we simulate past climates? Evaluating the models using global palaeoenvironmental datasets. *Quaternary Science Reviews* 19: 321–346.
- McMahon SM, Harrison SP, Armbruster WS, et al. (2011) Improving assessment and modelling of climate change impacts on global terrestrial biodiversity. *Trends in Ecology & Evolution* 26: 249–259.
- Pickett EJ, Harrison SP, Hope G, et al. (2004) Pollen-based reconstructions of biome distributions for Australia, Southeast Asia and the Pacific (SEAPAC region) at 0, 6000 and 18 000 ¹⁴C yr BP. *Journal of Biogeography* 31: 1381–1444.
- Prentice IC, Cramer W, Harrison SP, Leemans R, Monserud RA, and Solomon AM (1992) A global biome model based on plant physiology and dominance, soil properties and climate. *Journal of Biogeography* 19: 117–134.
- Prentice IC, Guiot J, Huntley B, Jolly D, and Cheddadi R (1996) Reconstructing biomes from palaeoecological data: A general method and its application to European pollen data at 0 and 6 ka. *Climate Dynamics* 12: 185–194.
- Prentice IC, Jolly D, and BIOME 6000 participants (2000) Mid-Holocene and glacial-maximum vegetation geography of the Northern continents and Africa. *Journal of Biogeography* 27: 507–519.
- Prentice IC and Webb T III (1998) BIOME 6000: Reconstructing global mid-Holocene vegetation patterns from palaeoecological records. *Journal of Biogeography* 25: 997–1005.
- Qian H and Ricklefs RE (2000) Large-scale processes and the Asian bias in species diversity of temperate plants. *Nature* 407: 180–182.
- Randall DA, Wood RA, Bony S, et al. (2007) Climate models and their evaluation. In: Solomon S, Qin D, and Manning M, et al. (eds.) *Climate Change 2007: The Physical Science Basis*. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change, Cambridge: Cambridge University Press.
- Scott GAJ (1995) *Canada's Vegetation: A World Perspective*. Montreal: McGill-Queen's University Press 361 pp.
- Sitch S, Smith B, Prentice IC, et al. (2003) Evaluation of ecosystem dynamics, plant geography and terrestrial carbon cycling in the LPJ dynamic global vegetation model. *Global Change Biology* 9: 161–185.
- Sugita S (2007) Theory of quantitative reconstruction of vegetation I: Pollen from large sites REVEALS regional vegetation composition. *The Holocene* 17: 229–241.
- Takahara H, Sugita S, Harrison SP, Miyoshi N, Morita Y, and Uchiyama T (2000) Pollen-based reconstructions of Japanese biomes at 0, 6000 and 18 000 ¹⁴C yr BP. *Journal of Biogeography* 27: 665–683.
- Tarasov PE, Volkova VS, Webb T, et al. (2000) Last glacial maximum biomes reconstructed from pollen and plant macrofossil data from northern Eurasia. *Journal of Biogeography* 27: 609–620.
- Tarasov PE, Webb T III, Andreev AA, et al. (1998) Present-day and mid-Holocene biomes reconstructed from pollen and plant macrofossil data from the former Soviet Union and Mongolia. *Journal of Biogeography* 25: 1029–1053.
- Thompson RS and Anderson KH (2000) Biomes of Western North America at 18 000, 6000 and 0 ¹⁴C yr BP reconstructed from pollen and packrat midden data. *Journal of Biogeography* 27: 555–584.
- Williams JW and Jackson ST (2007) Novel climates, no-analog communities, and ecological surprises. *Frontiers in Ecology and the Environment* 5: 448–475.
- Williams JW, Schuman BN, Webb T, Bartlein PJ, and Leduc PL (2000) Late Quaternary biomes of Canada and the Eastern United States. *Journal of Biogeography* 27: 585–607.
- Williams JW, Shuman BN, Webb T, Bartlein PJ, and Leduc PL (2004) Late-Quaternary vegetation dynamics in North America: Scaling from taxa to biomes. *Ecological Monographs* 74: 309–334.
- Wohlfahrt J, Harrison SP, and Braconnot P (2004) Synergistic feedbacks between ocean and vegetation on mid- and high-latitude climates during the mid-Holocene. *Climate Dynamics* 22: 223–238.
- Wohlfahrt J, Harrison SP, Braconnot P, et al. (2008) Evaluation of coupled ocean–atmosphere simulations of the mid-Holocene using palaeovegetation data from the Northern hemisphere extratropics. *Climate Dynamics* 31: 871–890. <http://dx.doi.org/10.1007/s00382-008-0415-508>.
- Yu G, Chen X, Ni J, et al. (2000) Palaeovegetation of China: a pollen data-based synthesis for the mid-Holocene and last glacial maximum. *Journal of Biogeography* 27: 635–664.
- Yu G, Prentice IC, Harrison SP, and Sun X (1998) Pollen-based biome reconstructions for China at 0 ka and 6 ka. *Journal of Biogeography* 25: 1055–1069.
- Yurtsev BA (1982) Relics of the xerophyte vegetation of Beringia in Northeastern Asia. In: Hopkins DM, Matthews JV, Schweger CE, and Young SB (eds.) *Paleoecology of Beringia*, pp. 157–177. New York: Academic Press.

POLLSCAPE Model: Simulation Approach for Pollen Representation of Vegetation and Land Cover

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Introduction

POLLSCAPE (Sugita, 1994, 1998) is a general simulation framework to explore factors and mechanisms affecting pollen deposition on mires and lakes in heterogeneous landscapes. Inter-taxonomic differences in pollen productivity and dispersal characteristics, basin size, spatial patterns of vegetation, and atmospheric conditions such as wind speed and turbulence simultaneously influence pollen dispersal and deposition (Prentice, 1985; Sugita, 1994). Simulations are a useful tool to understand functional relationships between a combination of these factors and pollen assemblages, providing us with plausible scenarios for interpreting fossil pollen data (Bunting and Middleton, 2009; Caseldine et al., 2008; Fyfe, 2006; Sugita et al., 1997). This article describes contributions of POLLSCAPE to the advances in the theory of pollen analysis and palynology, factors and mechanisms considered in POLLSCAPE, and future directions of this simulation approach.

Specifying the spatial scale of vegetation and land cover is an important step for quantitative reconstruction of past environment using fossil pollen (Davis, 2000). Traditionally, paleoecologists and palynologists have been interested in reconstructing past vegetation at a regional to subcontinental scale in order to reconstruct past climate. Pollen records from large- and medium-sized sites are appropriate for this purpose (Jacobson and Grimm, 1986; Prentice et al., 1992; Webb, 1981; Wright et al., 1994). Other ecological questions that require much finer spatiotemporal resolution, such as species invasion, stand-level vegetation dynamics, and natural and anthropogenic impacts on landscape openness, require fossil pollen records from small sites such as forest hollows (Andersen, 1970; Bradshaw and Lindbladh, 2005; Davis et al., 1998; Mitchell, 2005). Although these studies emphasized the importance of the spatial scale of vegetation represented by pollen, the source area of pollen is still ambiguous and difficult to quantify. Use of simulations in Quaternary palynology is still limited, but POLLSCAPE and related approaches have been useful in developing new concepts and methods to define the source area of pollen in complex and realistic landscapes (Sugita, 1994) and thus to reconstruct past vegetation and land cover quantitatively (Sugita, 2007a,b).

POLLSCAPE: A Simulation Framework for a Better Understanding of the Pollen–Vegetation Relationships in Vegetation with Complex Spatial Structure

POLLSCAPE was designed in such a way that pollen–vegetation relationships can be evaluated in realistic landscapes with patchy and heterogeneous vegetation (Davis and Sugita, 1997; Sugita, 1994, 1998; Sugita et al., 1997, 1999). We can gain some insights into the pollen–vegetation relationships

and pollen source area by assuming homogeneous vegetation (Davis, 1963; Prentice, 1985, 1988; Sugita, 1993). However, the spatial structure of vegetation is often complex, and pollen assemblages are expected to vary from site to site even in the same vegetation zone. There are many factors simultaneously influencing pollen dispersal and deposition at pollen sampling sites. POLLSCAPE simulations have been useful in teasing apart factors and mechanisms that most affect the pollen–vegetation relationships in this complex system.

Pollen Dispersal and Deposition Models in POLLSCAPE

POLLSCAPE simulations have used pollen dispersal and deposition models proposed by Prentice (1985, 1988), Sugita (1993), and Sugita et al. (1999). These models describe taxon-specific quantitative relationships between plant abundances surrounding pollen sites and pollen loading. Hereafter, pollen loading is defined as the number of pollen grains deposited onto the surface of bogs and lake water before sedimentation. Prentice's model estimates pollen deposition at a point at the center of a sedimentary basin and thus is appropriate for describing the relationships between plant abundance and pollen assemblages on mires and bogs, where pollen grains rarely move horizontally after deposition. Sugita's model (Sugita, 1993), modified after Prentice (1985, 1988), estimates pollen deposition on the entire surface of a basin and thus is more suitable for describing pollen loading and assemblages in lake sediment. This is because pollen originally deposited over the entire lake surface is redistributed by mixing and focusing before permanent sedimentation (Davis and Brubaker, 1973; Davis et al., 1984). Another model proposed by Sugita et al. (1999) – the Ring-Source model – provides more accurate description of pollen dispersal and deposition over the entire surface of a lake than the model proposed in Sugita (1993).

Major factors and parameters included in these three models are fall speed of pollen and pollen productivity for individual plant taxa, average wind speed, basin size and type (e.g., mire or lake), and the spatial pattern of source plants in the landscape (Figure 1). Basic assumptions of the models (Prentice, 1985; Sugita, 1993) are:

1. The sampling basin is a circular opening in the forest canopy.
2. Wind direction is even in all directions.
3. Wind above canopy is the dominant agent of pollen transport.
4. Each taxon has a constant pollen productivity in space and time.
5. Spatial distribution of source plants for each taxon is described as a function of distance from a point at the center of the basin.

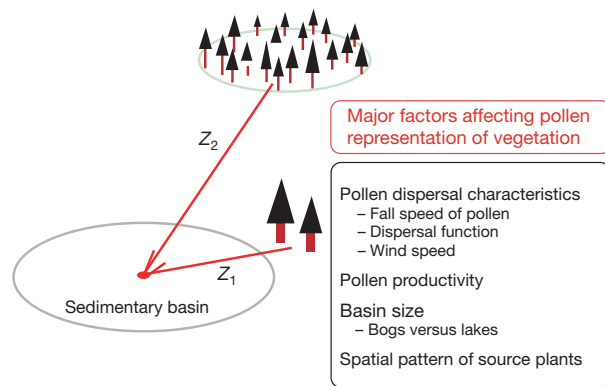


Figure 1 Factors affecting pollen representation of vegetation. Modified from Sugita S and Takahara H (2001) Fourth dimension in ecology: Lessons from the past on vegetation dynamics. *Kagaku* 71: 77–85.

6. Pollen dispersal is approximated by a diffusion model of small particles from a ground-level source under neutral atmospheric condition (Sutton, 1953).

The effects on the models of departures from some of those assumptions have been examined elsewhere (Bunting and Middleton, 2005; Bunting et al., 2004; Gaillard et al., 2008; Hellman et al., 2008b; Nielsen, 2004; Sugita, 2007a,b).

These models of pollen dispersal and deposition affect how plant abundance is registered in pollen assemblages. Plants a few meters away from the basin provide more pollen grains to the sites than plants of the same abundance a few hundred meters away. At the same time, the number of plants which provide pollen to the basin increases as the distance from the basin increases. Thus, 'pollen from lakes and bogs records vegetation as a summation of plant abundance expressed in a distance-weighted fashion' (Sugita, 1998). The Prentice–Sugita model (Prentice, 1985; Sugita, 1993) and Ring-Source model (Sugita et al., 1999) describe this concept in a simple mathematical fashion (Sugita et al., 2010a). Further modification, improvement, and testing of the pollen dispersal/deposition models may be necessary, particularly because pollen dispersal distance from a point source can be sensitive to parameter selection of atmospheric conditions such as wind speed, turbulence, and heights of source plants (Bunting and Middleton, 2005; Jackson and Lyford, 1999; Sjögren et al., 2010). However, their simplicity is also a key to clarify important factors and mechanisms to approximate pollen dispersal and deposition.

On the Source Area of Pollen

Characteristic Radius

Assuming homogeneous vegetation, Prentice's (1985) model provides estimates of the area within which a given proportion of total pollen originates (e.g., 50%, 70%, or 90%) for mires and bogs of different sizes. Prentice (1988) termed the distance for the area as 'the characteristic radius' for each taxon. Sugita's models (Sugita, 1993; Sugita et al., 1999) can also be used in the same way. Figure 2 shows the characteristic radius for three different pollen types using Prentice's and Sugita's models;

(A) spruce (*Picea* spp.), (B) sugar maple (*Acer saccharum*), and (C) oak (*Quercus* spp.) (Sugita, 1993). Spruce, sugar maple, and oak produce heavy, relatively heavy, and relatively light pollen types, respectively (Prentice, 1985; Sugita, 1993). For each taxon, Figure 2 shows the changes in the characteristic radius, depending on the basin size and plant taxa. The top graph for each taxon represents predictions from Prentice's model (1985) for mires, and the bottom graph from Sugita's model (1993) for lakes. For example, when the basin radius is 50 m, 70% of spruce pollen comes from within ca. 450 m for mires and from within ca. 320 m for lakes (Figure 2a). When oak pollen accumulates, 70% comes from within ca. 12 000 m for mires and from within ca. 10 000 m for lakes (Figure 2c). The characteristic radius for sugar maple is intermediate between the two (Figure 2b). Thus, the characteristic radius varies depending on plant taxa, basin size, and basin types, and is model dependent.

The characteristic radius is a conceptually simple and useful measure to approximate the source area of pollen. However, it assumes homogeneous vegetation on a plane land surface and is not able to specify which of these source areas (e.g., 50% or 70%) is appropriate for more spatially complex vegetation patterning.

Relevant Source Area of Pollen

One of the important contributions of POLLSCAPE simulations to palynology is the concept of the 'Relevant Source Area of Pollen' (RSAP) (Sugita 1994). Based on the simulation results, the RSAP is defined as the area beyond which correlations between pollen loading and plant abundance for all taxa do not continue to improve in a given region or vegetation type (Sugita, 1994). In other words, vegetation heterogeneity within the RSAP is recorded as the variation in pollen loading among sites, superimposed on a constant background pollen loading coming from the regional vegetation. Simulation results using a hypothetical landscape with three plant taxa (Figure 3) show that pollen loading coming from beyond the RSAP (i.e., 300 m for 50-m-radius lakes in this scenario) is nearly constant for each taxon at all sites (Figure 4). Several simulations have also shown that only 30–45% of the total pollen arriving at a given site comes from within the RSAP (Sugita, 1994, 1998), which has been supported by empirical studies (e.g., Calcote, 1995). Thus, the RSAP can differ significantly from the 70% characteristic radius that is often used as a measure of the pollen source area (Prentice, 1988).

Simulations also show that the RSAP for sedimentary basins of a given size can differ between different landscapes in various regions. Several POLLSCAPE simulations (Broström et al., 2005; Bunting et al., 2004; Hellman et al., 2009a,b; Nielsen and Sugita, 2005; Sugita et al., 1999) have demonstrated that, besides the basin size, the spatial pattern of vegetation is the most dominant factor influencing the size of the RSAP. Although wind speed and atmospheric turbulence conditions strongly affect the distance of pollen dispersal from a point source (Jackson and Lyford, 1999; Sjögren et al., 2010), stronger winds and unstable atmospheric conditions would not affect the RSAP significantly (Gaillard et al., 2008; Nielsen and Sugita, 2005; Sugita, 2007a,b). Those simulation results are informative and important. More POLLSCAPE simulations

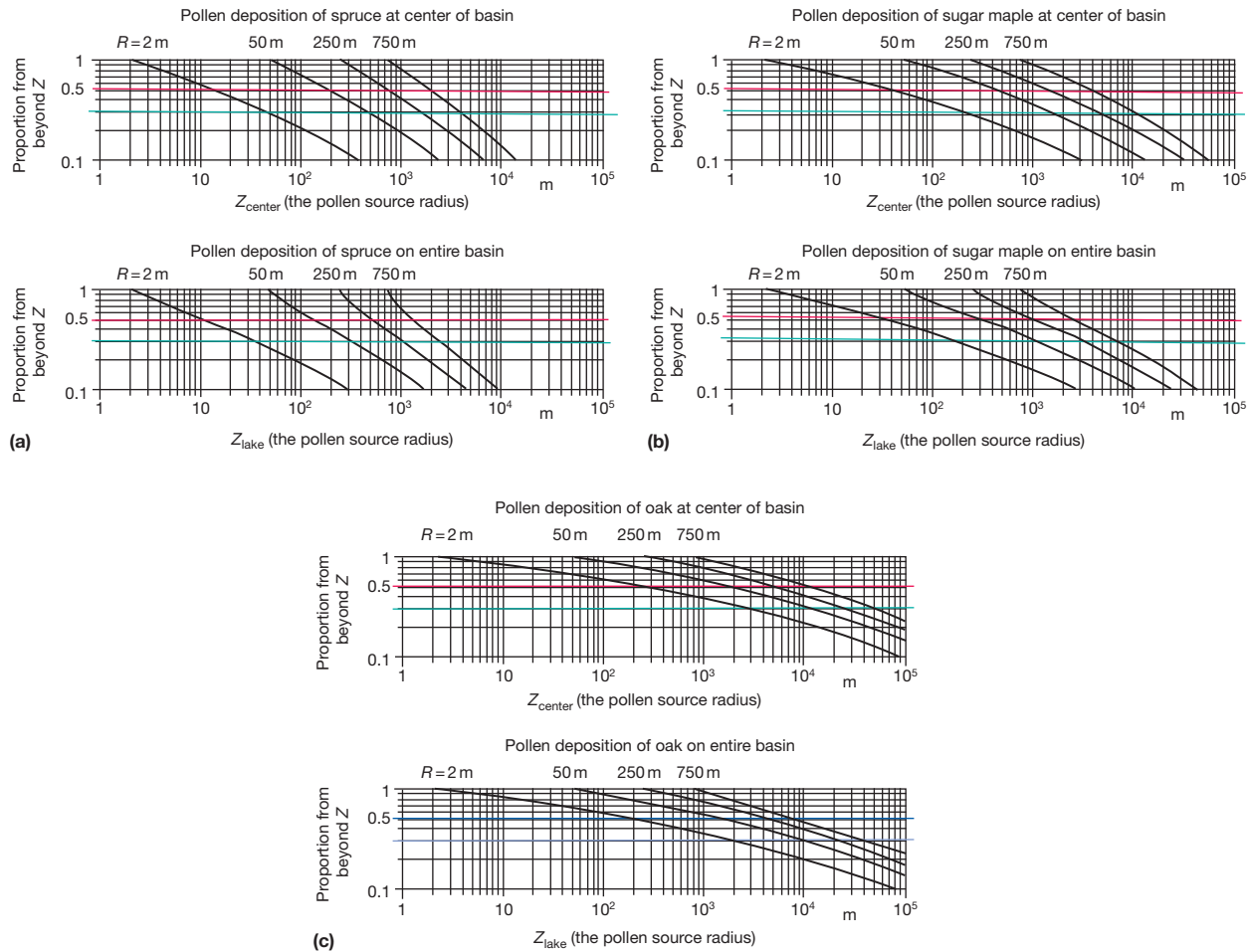


Figure 2 The characteristic radius for spruce (*Picea* spp.), sugar maple (*Acer saccharum*), and oak (*Quercus* spp.) estimated using Prentice's model (1985) for a point at the center of a basin and Sugita's model (1993) for the entire surface of a basin. Radius of a basin varies from 2 to 750 m. Red lines represent the 50% characteristic radii and green lines the 70% characteristic radii. Modified from Sugita S (1993) A model of pollen source area for an entire lake surface. *Quaternary Research* 39: 239–244.

will be necessary to elucidate important properties of the RSAP in the modern and past landscapes.

Empirically, the RSAP can be estimated using data sets of pollen percentages and vegetation abundance surrounding similarly sized sites with the Extended R-value models (Calcote, 1995; Parsons et al., 1983; Sugita, 1994). During the past decade, the number of RSAP estimates has increased in northwestern Europe, the United States, and elsewhere (Table 1). When pollen assemblages from moss polsters and forest hollows located in closed forests and dense woodlands are used, the RSAP estimates are within a 120–130-m radius in the midwest of the United States (Sugita et al., 2010b) and a 50–150-m radius in Norfolk, United Kingdom (Bunting et al., 2005). When moss polsters from semi-open forests and open landscapes are used, the estimates vary from 300 to 1000 m in radius (Broström et al., 2005; Mazier et al., 2008; Räsänen et al., 2007; von Stedingk et al., 2008). The RSAP estimates vary from a few hundred meters to 1700 m in radius when pollen assemblages from medium-sized lakes, mostly < 20–30 ha, in semi-open forests and open landscapes are used (Duffin and Bunting, 2008; Hjelle and Sugita, 2012; Nielsen and Odgaard, 2004; Nielsen and Sugita, 2005; Soepboer et al., 2007).

Other Insights from POLLSCAPE Simulations

Effects of Vegetation Patterning and Basin Size on Pollen Assemblages

POLLSCAPE simulations (Sugita, 1994) demonstrate relatively large site-to-site variation in pollen loading and pollen proportions for forest hollows and small ponds (2 and 50 m in radius) and smaller variations for larger lake sizes (250 and 750 m in radius) (Figure 5). This large variation in pollen loading for hollows and small ponds reflects the differences in plant abundance surrounding the sites in the simulated landscape (Figure 3). On the other hand, small variation in pollen loading and proportion for larger lakes, especially for lakes 750 m in radius, indicates that pollen from these lakes records the surrounding vegetation as homogeneous even when the actual vegetation is heterogeneous. 'Grain size' of vegetation heterogeneity and lake size interact, determining the efficiency with which pollen can detect local abundances of plants. Those results support the empirical observations that pollen data from smaller lakes are especially appropriate for fine-scale vegetation reconstruction, while pollen data from larger lakes are suitable for reconstruction of regional

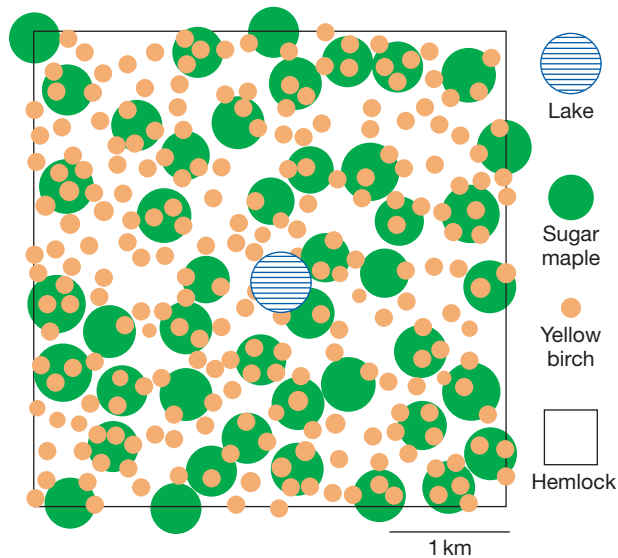


Figure 3 Hypothetical patchy landscape simulated and used for calculating pollen deposition on a circular lake of a given radius. Some POLLSCAPE simulations use this type of patchy and heterogeneous landscapes within 60 km × 60 km or 100 km × 100 km (Sugita, 1994; Sugita et al., 1999) as well as homogeneous regional vegetation beyond the plots out to 100 to 400 km from the sedimentary basin to estimate pollen loading coming from the entire region. Three common tree taxa in the eastern United States – sugar maple (*Acer saccharum*), hemlock (*Tsuga* sp.), and yellow birch (*Betula alleghaniensis*) – are considered in this landscape. Modified from Sugita S (1994) Pollen representation of vegetation in Quaternary sediments: Theory and method in patchy vegetation. *Journal of Ecology* 82: 881–897.

vegetation and climate (Berglund, 1973; Bradshaw and Webb, 1985; Jacobson and Bradshaw, 1981; Janssen, 1972).

Factors and Mechanisms Affecting the Pollen Percentage: Vegetation Relationships

Pollen percentages are difficult to use for quantitative reconstruction of vegetation. The pollen–vegetation relationships for the same species vary from region to region because of the confounding effects of pollen percentage calculation and the differences in background pollen loading (Broström et al., 1998; Prentice and Webb, 1986; Sugita et al., 1999). For example, non-arboreal pollen percentages (NAP%) have been used to approximate landscape openness in past landscapes (Berglund, 1991; Frenzel et al., 1994). However, Broström et al. (1998) showed that the relationship between NAP% and landscape openness within 1000 m from lakes can differ significantly between southern Skåne (i.e., an open agricultural region) and northern Skåne and Småland (i.e., semi-open region) in southern Sweden (Figure 6). POLLSCAPE simulations using a similar set of open and semi-open landscape patterns (Figure 7; Sugita et al., 1999) successfully depicted the relationships that are similar to those observed in Broström et al. (1998) (Figure 6). Sugita et al. (1999) and others clearly show that the differences in both the spatial patterns of vegetation and the regional abundance of vegetation, thus background pollen loading, affect the NAP%–landscape openness relationships. It would be unrealistic to apply the modern

relationship of NAP% and landscape openness obtained from one region for reconstructing the past openness of the landscape elsewhere.

POLLSCAPE simulations have also helped understand potential and limits for the interpretation of pollen records. For example, Sugita et al. (1997) showed that pollen records would not register large-scale disturbance events (e.g., 2500 ha) even if these events occur only a few hundred meters away from study sites. This conclusion was supported in northern Finland using pollen trap data (Hicks, 1994). Data-model comparison of the pollen representation of disturbance in Estonia also supports a similar conclusion. (Kangur, 2002; Koff et al., 2000). Davis and Sugita (1997), Giesecke (2005), Poska et al. (2008), and von Stedingk and Fyfe (2009) used POLLSCAPE to demonstrate how plant migration, vegetation dynamics, and land cover changes would be recorded in fossil pollen data and to discuss implications for the interpretation of the long-term changes in spatiotemporal patterns of vegetation.

POLLSCAPE and Quantitative Reconstruction of Past Vegetation

Quantitative reconstruction of vegetation is hampered by the lack of theoretically sound methods that incorporate inter-taxonomic differences in pollen dispersal and productivity, temporal changes in background pollen, and the complex nature of pollen percentage calculations. Prentice and Parsons (1983) proposed an elegant method for quantitative reconstruction of vegetation using the inverse form of the Extended R-value models. However, unless the changes in background pollen through time are known, their method would hardly be applicable for reconstruction of vegetation and land cover.

Using POLLSCAPE, Sugita (1994) showed that pollen loading coming from beyond the RSAP in a given landscape can be estimated, once the overall regional composition of vegetation is calculated. In theory, the regional composition of vegetation and land cover can be estimated from pollen assemblages from large lakes >100–500 ha (Sugita, 1994, 1998, 2007a), because variation of pollen assemblages among those sites can be small even if vegetation is highly heterogeneous (Figure 5). This represents the first step for the Landscape Reconstruction Algorithm (LRA), a two-step framework for quantitative reconstruction of vegetation proposed in Sugita (2007a, b) (Figure 8). In the first step, the REVEALS (Regional Estimates of VEgetation Abundance from Large Sites) model, which is a generalized form of the R-value model (Davis, 1963; Parsons et al., 1983), estimates the regional vegetation composition using quantitative estimates of pollen productivity as input parameters and pollen assemblages, or counts, from large sites (Sugita, 2007a). The second step uses the LOVE (LOCAL VEgetation Estimates) model to estimate vegetation composition within the RSAP for smaller sites, subtracting the amount of pollen coming from beyond the RSAP in concept (Sugita, 2007b). Although propagation of errors in this model-based approach could potentially be problematic, LRA can overcome the problems of nonlinear relationships between pollen percentages and vegetation composition as well as the difficulties in evaluating background pollen coming from the regional vegetation (Figure 6; Sugita et al., 1999).

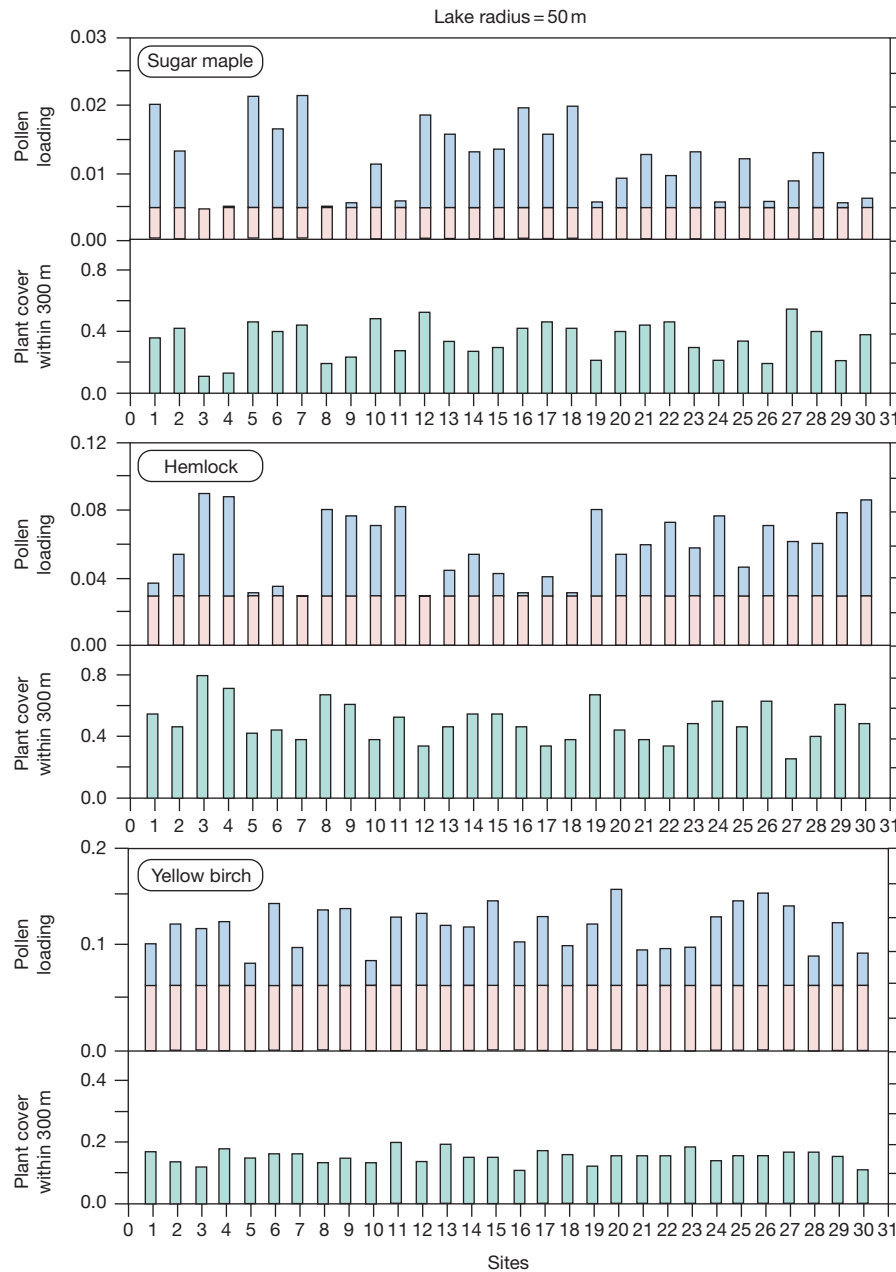


Figure 4 Simulated pollen loading on small ponds (radius = 50 m) using the hypothetical landscape in Figure 3. The relevant source area of pollen based on the Extended R-value model is the area within a 300-m radius from lake shore (Sugita, 1994). Pollen loading coming from within 300 m (light orange) and beyond 300 m (blue) from the shore is compared with plant cover (green) of these taxa. Modified from Sugita S (1998) Modelling pollen representation of vegetation. In: Gaillard M-J and Berglund BE (eds.) *Quantification of Land Surfaces Cleared of Forests During the Holocene – Modern Pollen/Vegetation/Landscape Relationships as an Aid to the Interpretation of Fossil Pollen Data*. Palaeoclimate Research, pp. 1–18. Stuttgart: Gustav Fischer Verlag.

Every step of the LRA approach has been evaluated and validated by POLLSCAPE simulations (Sugita, 2007a,b). In addition, Hellman et al. (2008a), Soepboer et al. (2010), Nielsen and Odgaard (2010), and Sugita et al. (2010b) have tested the LRA approach empirically.

Pollen productivity is one of the most important parameters for POLLSCAPE and the LRA applications (Figure 8). At the same time, POLLSCAPE simulations have provided important clues and insights into how to set up sampling designs to

obtain reliable estimates of pollen productivity (Broström et al., 2005; Gaillard et al., 2008). During the past decade, the number of taxa for which pollen productivity estimates are available has increased dramatically in many parts of Europe and elsewhere (Broström et al., 2008), including those in southern Sweden (Broström et al., 2004), western Norway (Hjelle, 1998; Hjelle and Sugita, 2012), Denmark (Nielsen, 2004), Great Britain (Bunting et al., 2005), northern Michigan and Wisconsin in the United States (Calcote, 1995; Sugita

Table 1 Empirical estimates of the relevant source area of pollen (RSAP) in radius currently available

Vegetation/landscape types	Pollen samples obtained from	
	Moss polsters/hollows	Lakes/reservoirs (mostly < 20–30 ha)
Closed forests/dense woodlands	Midwest, USA: 120–130 m; Sugita et al. (2010b) Norfolk, UK: 50–150 m; Bunting et al. (2005)	
Semi-open forests/open landscapes	S. Sweden: 400 m; Broström et al. (2005) C. Sweden: 500 m; von Stedingk et al. (2008) W. Switzerland: 300 m; Mazier et al. (2008) N. Finland: 1000 m; Räsänen et al. (2007)	Denmark: 1700 m; Nielsen and Sugita (2005) South Africa: 600–900 m; Duffin and Bunting (2008) Swiss Plateau: 800 m; Soepboer et al. (2007) W. Norway: 900–1000 m; Hjelle and Sugita (2012) S. Estonia: 1500–2000 m; Poska et al. (2011)

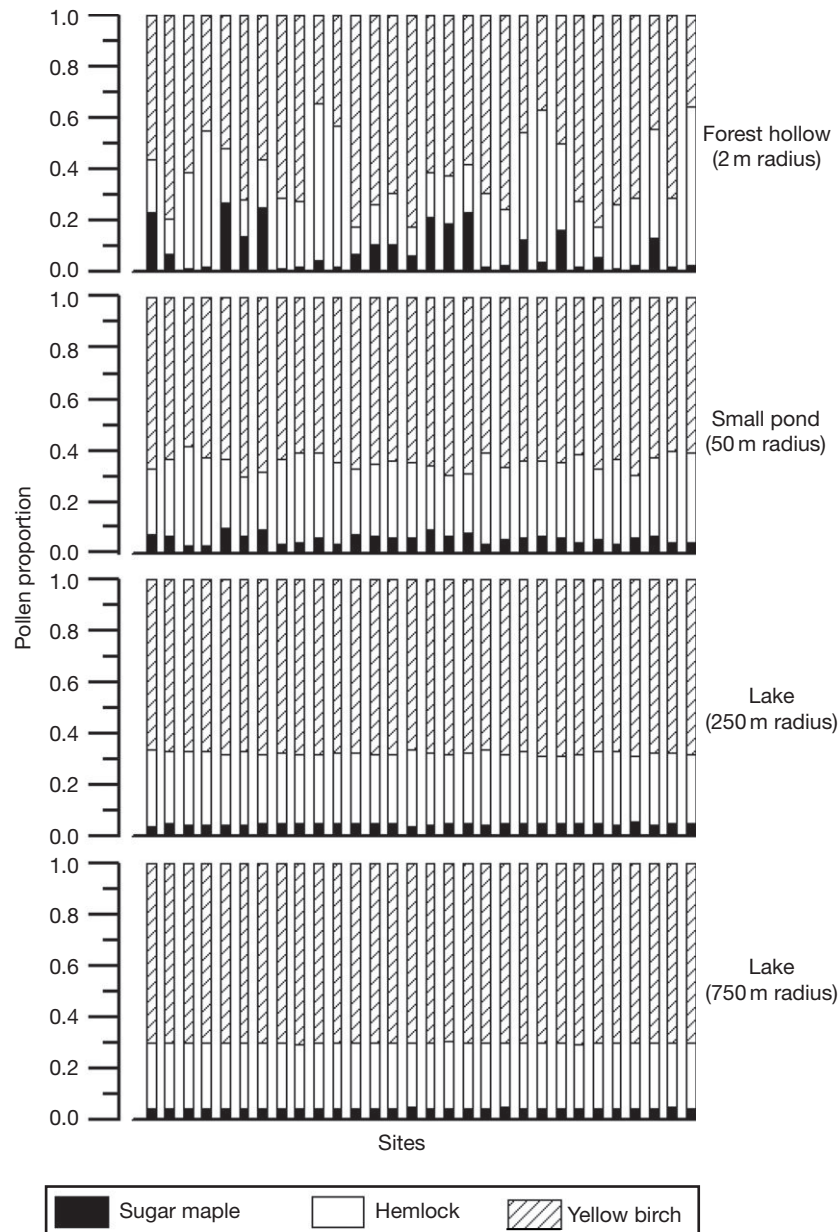


Figure 5 Simulated pollen assemblages at 30 sites using a landscape similar to that shown in [Figure 3](#). Lake radius varies from 2 to 750 m. Modified from Sugita S (1998) Modelling pollen representation of vegetation. In: Gaillard M-J and Berglund BE (eds.) *Quantification of Land Surfaces Cleared of Forests During the Holocene – Modern Pollen/Vegetation/Landscape Relationships as an Aid to the Interpretation of Fossil Pollen Data*. Palaeoclimate Research, pp. 1–18. Stuttgart: Gustav Fischer Verlag.

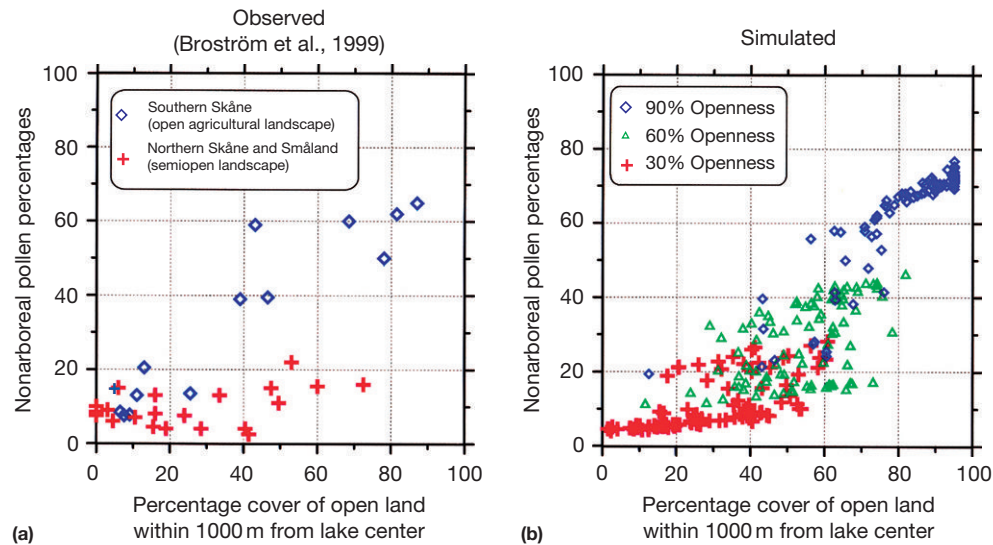


Figure 6 Observed and simulated relationships between non-arboreal pollen percentages and percentage cover of open land within 1000 m from Lake Center in southern Sweden. POLLSCAPE was used to simulate non-arboreal pollen percentages and landscape openness. Modified from Sugita S, Gaillard M-J, and Broström A (1999) Landscape openness and pollen records: A simulation approach. *The Holocene* 9: 409–421.

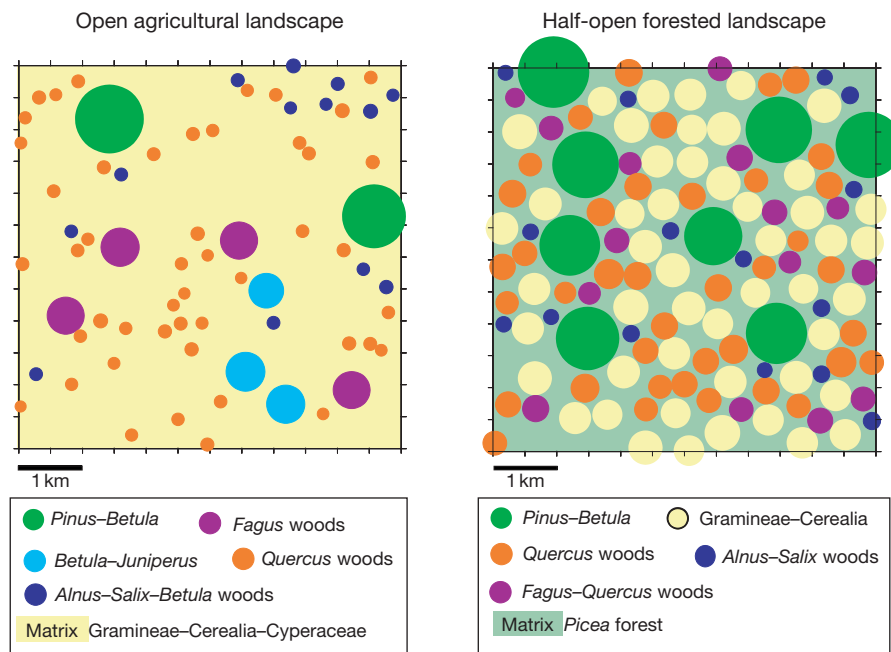


Figure 7 Simulated landscapes for open agricultural region in Skåne, and for half-open forested region in northern Skåne and Småland in southern Sweden. Modified from Sugita S, Gaillard M-J, and Broström A (1999) Landscape openness and pollen records: A simulation approach. *The Holocene* 9: 409–421.

et al., 2006; Sugita et al., 2010b), the Swiss Plateau (Soepboer et al., 2007), the Swiss Jura Mountains (Mazier et al., 2008), Estonia (Poska et al., 2011), northern Finland (Räsänen et al., 2007), central Sweden (von Stedingk et al., 2008), and central South Africa (Duffin and Bunting, 2008). Currently, more studies for pollen productivity estimates are underway in other parts of Europe, central Africa, China, India, Japan, the Great Plains of the United States, northern Canada, and Australia.

Both forward modeling of POLLSCAPE and related approaches (Broström et al., 2005; Bunting and Middleton, 2005, 2009; Bunting et al., 2004; Hellman et al., 2009a,b; Nielsen and Sugita 2005; Sugita et al., 1997, 1999) and inverse modeling of the LRA (Sugita, 2007b; Sugita et al., 2010b) are important and informative for estimating the RSAP in the past landscapes, another important parameter necessary for the LOVE model (Figure 8). Although challenging, all the necessary tools are now at hand, and further POLLSCAPE and LRA

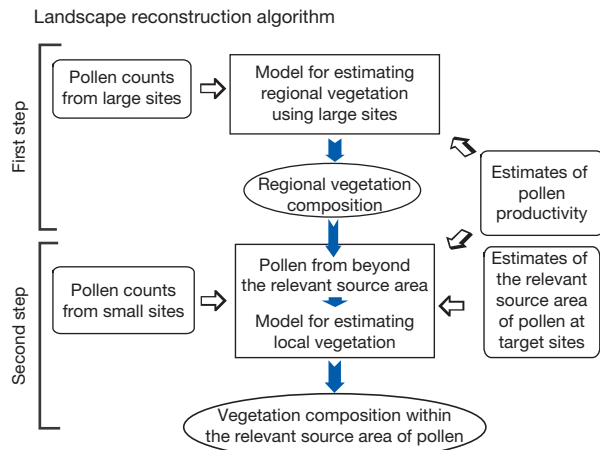


Figure 8 Landscape Reconstruction Algorithm, a two-step framework for quantitative reconstruction of vegetation and land cover. First step is to estimate regional composition of vegetation using the REVEALS model with pollen data from large lakes > 100–500 ha and estimates of pollen productivity. Pollen loading coming from beyond the relevant source area of pollen is then estimated from the REVEALS results. Second step is to reconstruct the local vegetation composition within the relevant source area of pollen using the LOVE model. POLLSCAPE simulations play an important role for estimating the relevant source area of pollen in the past landscapes. Modified from Sugita S (2007b) Theory of quantitative reconstruction of vegetation II: All you need is LOVE. *The Holocene* 17: 243–257.

experiments will be useful for robust estimates of the RSAP in various landscapes through time.

Other POLLSCAPE applications are also found at the NordForsk (Nordic Research Council) POLLANDCAL (POLLEN-LANDscape CALibration) network website (<http://www.geog.ucl.ac.uk/ecrc/pollandcal/>), coordinated by Marie-Jose Gaillard, Linnaeus University, Sweden. Members of the POLLANDCAL network have developed user-friendly packages of POLLSCAPE-related computer programs, such as OPENLAND3 (Eklöf et al., 2004), MOSAIC (Middleton and Bunting, 2004), and HUMPOL (Bunting and Middleton, 2005). These programs enable users to handle both simulated landscapes and empirical vegetation/land cover data, such as aerial photographs and digitized maps, for POLLSCAPE simulations. POLLSCAPE has become not only a useful analytical tool but also a tool to set up hypotheses and design appropriate sampling schemes specific to research objectives during project planning stages. Palynologists and paleoecologists can take full advantage of this research tool to set up a hypothesis-testing approach for a further development of the fields (Graumlich et al., 2005).

References

- Andersen ST (1970) The relative pollen productivity and pollen representation of north European trees, and correction factors for tree pollen spectra. Geological Survey of Denmark. II. Series. No. 96, 99.
- Berglund BE (1973) Pollen dispersal and deposition in an area of Southeastern Sweden – Some preliminary results. In: Birks HJB and West RG (eds.) *Quaternary Plant Ecology*, pp. 117–129. New York: Wiley.
- Berglund BE (1991) The Krageholm Area: The cultural landscape during 6,000 years in Southern Sweden-The Ystad Project. *Ecological Bulletins* 41: 221–244.
- Bradshaw RHW and Lindbladh M (2005) Regional spread and stand-scale establishment of *Fagus sylvatica* and *Picea abies* in Scandinavia. *Ecology* 86: 1679–1686.
- Bradshaw RHW and Webb T III (1985) Relationships between contemporary pollen and vegetation data from Wisconsin and Michigan, USA. *Ecology* 66: 721–737.
- Broström A, Gaillard M-J, Ihse M, and Odgaard B (1998) Pollen-landscape relationships in modern analogues of ancient cultural landscapes in southern Sweden – A first step towards quantification of vegetation openness in the past. *Vegetation History and Archaeobotany* 7: 189–201.
- Broström A, Nielsen AB, Gaillard M-J, et al. (2008) Pollen productive estimates of key European plant taxa for quantitative reconstruction of vegetation: A review. *Vegetation History and Archaeobotany* 17: 461–478.
- Broström A, Sugita S, and Gaillard M-J (2004) Pollen productivity estimates for the reconstruction of past vegetation cover in the cultural landscape of southern Sweden. *The Holocene* 14: 368–381.
- Broström A, Sugita S, and Gaillard M-J (2005) Estimating spatial scale of pollen dispersal in the cultural landscape of southern Sweden. *The Holocene* 15: 252–262.
- Bunting MJ, Armitage R, Binney HA, and Waller M (2005) Estimates of relative pollen productivity and relevant source area of pollen for major tree taxa in two Norfolk (UK) woodlands. *The Holocene* 15: 459–465.
- Bunting MJ, Gaillard MJ, Sugita S, Middleton R, and Broström A (2004) Vegetation structure and pollen source area. *The Holocene* 14: 651–660.
- Bunting MJ and Middleton R (2005) Modelling pollen dispersal and deposition using HUMPOL software, including simulating windroses and irregular lakes. *Review of Palaeobotany and Palynology* 134: 185–196.
- Bunting MJ and Middleton R (2009) Equifinality and uncertainty in the interpretation of pollen data: The Multiple Scenario Approach to reconstruction of past vegetation mosaics. *The Holocene* 19: 799–803.
- Calcote R (1995) Pollen source area and pollen productivity: Evidence from forest hollows. *Journal of Ecology* 83: 591–602.
- Caseldine C, Fyfe R, and Hjelle K (2008) Pollen modelling, paleoecology and archaeology: Virtualization and/or visualization of the past? *Vegetation History and Archaeobotany* 17: 543–549.
- Davis MB (1963) On the theory of pollen analysis. *American Journal of Science* 261: 897–912.
- Davis MB (2000) Palynology after Y2K – Understanding the source area of pollen in sediments. *Annual Reviews of Earth and Planetary Sciences* 28: 1–18.
- Davis MB and Brubaker LB (1973) Differential sedimentation of pollen grains in lakes. *Limnology and Oceanography* 18: 635–646.
- Davis MB, Calcote RR, Sugita S, and Takahara H (1998) Patchy invasion and the origin of a hemlock-hardwoods forest mosaic. *Ecology* 79: 2641–2659.
- Davis MB, Moeller RE, and Ford J (1984) Sediment focusing and pollen influx. In: Haworth EY and Lund JWG (eds.) *Lake Sediments and Environmental History*, pp. 261–293. Leicester: University of Leicester Press.
- Davis MB and Sugita S (1997) Reinterpreting the fossil pollen record of Holocene tree migration. In: Huntley B, Cramer W, Morgan AV, Prentice HC, and Allen JRM (eds.) *Past and Future Rapid Environmental Changes: The Spatial and Evolutionary Responses of Terrestrial Biota. NATO ASI Series*, pp. 181–193. Berlin: Springer.
- Duffin KI and Bunting MJ (2008) Relative pollen productivity and fall speed estimates for southern African savanna taxa. *Vegetation History and Archaeobotany* 17: 507–525.
- Eklöf M, Broström A, Gaillard M-J, and Pilesjö P (2004) OPENLAND 3: A computer program to estimate plant abundance around pollen sampling sites from vegetation maps: A necessary step for calculation of pollen productivity estimates. *Review of Palaeobotany and Palynology* 132: 67–77.
- Frenzel B, Andersen ST, Berglund BE, and Glaser B (1994) Evaluation of land surfaces cleared from forests in Roman Iron Age and the time of migrating Germanic tribes based on regional pollen diagrams. *Palaeoklimaforschung*, p. 134. Mainz: Gustav Fischer Verlag.
- Fyfe R (2006) GIS and the application of a model pollen deposition and dispersal: A new approach to testing landscape hypotheses using the POLLANDCAL models. *Journal of Archaeological Science* 33: 483–493.
- Gaillard M-J, Sugita S, Bunting MJ, et al. (2008) The use of modeling and simulation approach in reconstructing past landscapes from fossil pollen data: A review and results from the POLLANDCAL network. *Vegetation History and Archaeobotany* 17: 419–444.
- Giesecke T (2005) Moving front or population expansion: How did *Picea abies* (L.) Karst. become frequent in central Sweden? *Quaternary Science Reviews* 24: 2495–2509.
- Graumlich LJ, Sugita S, Brubaker LB, and Card VM (2005) Paleoperspectives in ecology. *Ecology* 86: 1667–1668.
- Hellman S, Bunting MJ, and Gaillard M-J (2009) Relevant Source Area of Pollen in patchy cultural landscapes and signals of anthropogenic landscape disturbance in

- the pollen record: A simulation approach. *Review of Palaeobotany and Palynology* 153: 245–258.
- Hellman S, Gaillard M-J, Broström A, and Sugita S (2008a) The REVEALS model, a new tool to estimate past regional plant abundance from pollen data in large lakes: Validation in southern Sweden. *Journal of Quaternary Science* 23: 21–42.
- Hellman SEV, Gaillard M-J, Broström A, and Sugita S (2008b) Effects of the sampling design and selection of parameter values on pollen-based quantitative reconstructions of regional vegetation: A case study in southern Sweden using the REVEALS model. *Vegetation History and Archaeobotany* 17: 445–459.
- Hellman S, Gaillard M-J, Bunting MJ, and Mazier F (2009) Estimating the Relevant Source Area of Pollen in the past cultural landscapes of southern Sweden – A forward modelling approach. *Review of Palaeobotany and Palynology* 153: 259–271.
- Hicks S (1994) Present and past pollen records of Lapland forests. *Review of Palaeobotany and Palynology* 82: 17–35.
- Hjelle KL (1998) Herb pollen representation in surface moss samples from mown meadows and pastures in western Norway. *Vegetation History and Archaeobotany* 7: 79–96.
- Hjelle KL and Sugita S (2012) Estimating pollen productivity and relevant source area of pollen using lake sediments in Norway: How does lake size variation affect the estimates? *The Holocene* 22: 312–323.
- Jackson ST and Lyford ME (1999) Pollen dispersal models in Quaternary plant ecology: Assumptions, parameters, and prescriptions. *The Botanical Review* 65: 39–75.
- Jacobson GL Jr. and Bradshaw RHW (1981) The selection of sites for paleovegetational studies. *Quaternary Research* 16: 80–96.
- Jacobson GL Jr. and Grimm EC (1986) A numerical analysis of Holocene forest and prairie vegetation in central Minnesota. *Ecology* 67: 958–966.
- Janssen CR (1972) The paleoecology of plant communities in the Dommel Valley, North Brabant, Netherlands. *Journal of Ecology* 60: 411–437.
- Kangur M (2002) Methodological and practical aspects of the presentation and interpretation of microscopic charcoal data from lake sediments. *Vegetation History and Archaeobotany* 11: 289–294.
- Koff T, Punning J-M, and Kangur M (2000) Impact of forest disturbance on the pollen influx in lake sediments during the last century. *Review of Palaeobotany and Palynology* 111: 19–29.
- Mazier F, Broström A, Gaillard M-J, Sugita S, Vittoz P, and Buttler A (2008) Pollen productivity estimates and relevant source area of pollen for selected plant taxa in a pasture woodland landscape of the Jura Mountains (Switzerland). *Vegetation History and Archaeobotany* 17: 479–495.
- Middleton R and Bunting MJ (2004) Mosaic v1.1: Landscape scenario creation software for simulation of pollen dispersal and deposition. *Review of Palaeobotany and Palynology* 132: 61–66.
- Mitchell FJG (2005) How open were European primeval forests? Hypothesis testing using paleoecological data. *Journal of Ecology* 93: 168–177.
- Nielsen AB (2004) Modelling pollen sedimentation in Danish lakes around AD 1800 – An attempt to validate the POLLSCAPE model. *Journal of Biogeography* 31: 1693–1709.
- Nielsen AB and Odgaard B (2004) The use of historical analogues for interpreting fossil pollen records. *Vegetation History and Archaeobotany* 13: 33–43.
- Nielsen AB and Odgaard B (2010) Quantitative landscape dynamics in Denmark through the last three millennia based on the Landscape Reconstruction Algorithm approach. *Vegetation History and Archaeobotany* 19: 375–387.
- Nielsen AB and Sugita S (2005) Estimating relevant source area of pollen for small Danish lakes around AD 1800. *The Holocene* 15: 1006–1020.
- Parsons RW, Gordon AD, and Prentice IC (1983) Statistical Uncertainty in Forest Composition Estimates Obtained from Fossil Pollen Spectra via the R-Value Model. *Review of Palaeobotany and Palynology* 40: 177–189.
- Poska A, Meltsov V, Sugita S, and Vassiljev J (2011) Relative pollen productivity estimates of major anemophilous taxa and relevant source area of pollen in a cultural landscape of the hemi-boreal forest zone (Estonia). *Review of Palaeobotany and Palynology* 167: 30–39.
- Poska A, Sepp E, Veski S, and Koppel K (2008) Using quantitative pollen-based land-cover estimations and spatial CA-Markov model to reconstruct the development of cultural landscape at Rouge, South Estonia. *Vegetation History and Archaeobotany* 17: 527–541.
- Prentice IC (1985) Pollen representation, source area, and basin size: Toward a unified theory of pollen analysis. *Quaternary Research* 23: 76–86.
- Prentice IC (1988) Records of vegetation in time and space: The principles of pollen analysis. In: Huntley B and Webb T III (eds.) *Vegetation History*, pp. 17–42. Dordrecht: Kluwer Academic.
- Prentice IC, Cramer W, Harrison SP, Leemans R, Monserud RA, and Solomon AM (1992) A global biome model based on plant physiology and dominance, soil properties and climate. *Journal of Biogeography* 19: 117–134.
- Prentice IC and Parsons RW (1983) Maximum likelihood linear calibration of pollen spectra in terms of forest composition. *Biometrics* 39: 1051–1057.
- Prentice IC and Webb T III (1986) Pollen percentages, tree abundances and the Fagerlind effect. *Journal of Quaternary Science* 1: 35–43.
- Räsänen S, Suutari H, and Nielsen AB (2007) A step further towards quantitative reconstruction of past vegetation in Fennoscandian boreal forests: Pollen productivity estimates for six dominant taxa. *Review of Palaeobotany and Palynology* 146: 208–220.
- Sjögren P, Connor S, and van der Knaap WO (2010) The development of composite dispersal functions for estimating absolute pollen productivity in the Swiss Alps. *Vegetation History and Archaeobotany* 19: 341–349.
- Soepboer W, Sugita S, and Lotter AF (2010) Regional vegetation-cover changes on the Swiss Plateau during the past two millennia: A pollen-based reconstruction using the REVEALS model. *Quaternary Science Reviews* 29: 472–483.
- Soepboer W, Sugita S, Lotter AF, van Leeuwen JFN, and van der Knaap WO (2007) Pollen productivity estimates for quantitative reconstruction of vegetation cover on the Swiss Plateau. *The Holocene* 17: 65–77.
- Sugita S (1993) A model of pollen source area for an entire lake surface. *Quaternary Research* 39: 239–244.
- Sugita S (1994) Pollen representation of vegetation in Quaternary sediments: Theory and method in patchy vegetation. *Journal of Ecology* 82: 881–897.
- Sugita S (1998) Modelling pollen representation of vegetation. In: Gaillard M-J and Berglund BE (eds.) *Quantification of Land Surfaces Cleared of Forests During the Holocene – Modern Pollen/Vegetation/Landscape Relationships as an Aid to the Interpretation of Fossil Pollen Data*. *Palaeoclimate Research*, pp. 1–18. Stuttgart: Gustav Fischer.
- Sugita S (2007a) Theory of quantitative reconstruction of vegetation I: Pollen from large sites REVEALS regional vegetation composition. *The Holocene* 17: 229–241.
- Sugita S (2007b) Theory of quantitative reconstruction of vegetation II: All you need is LOVE. *The Holocene* 17: 243–257.
- Sugita S, Gaillard M-J, and Broström A (1999) Landscape openness and pollen records: A simulation approach. *The Holocene* 9: 409–421.
- Sugita S, Hicks S, and Sormunen H (2010a) Absolute pollen productivity and pollen-vegetation relationships in northern Finland. *Journal of Quaternary Science* 25: 724–736.
- Sugita S, MacDonald GM, and Larsen CPS (1997) Reconstruction of fire disturbance and forest succession from fossil pollen in lake sediments: Potential and limitations. In: Clark JS, Cachier H, Goldammer JG, and Stocks B (eds.) *Sediment Records of Biomass Burning and Global Change. NATO ASI Series I: Global Environmental Change*, pp. 387–412. Berlin: Springer.
- Sugita S, Parshall T, and Calcote R (2006) Detecting differences in vegetation among paired sites using pollen records. *The Holocene* 16: 1123–1135.
- Sugita S, Parshall T, Calcote R, and Walker K (2010b) Testing the Landscape Reconstruction Algorithm for spatially explicit reconstruction of vegetation in northern Michigan and Wisconsin. *Quaternary Research* 74: 289–300.
- Sutton OG (1953) *Micrometeorology*. London: McGraw-Hill.
- von Stedingk H and Fyfe RM (2009) The use of pollen analysis to reveal Holocene treeline dynamics: A modeling approach. *The Holocene* 19: 273–283.
- von Stedingk H, Fyfe RM, and Allard A (2008) Pollen productivity estimates from the forest-tundra ecotone in west-central Sweden: Implications for vegetation reconstruction at the limits of boreal forest. *The Holocene* 18: 323–332.
- Webb T III (1981) The past 11,000 years of vegetational change in eastern North America. *BioScience* 31: 501–506.
- Wright HE Jr., Kutzbach JE, Webb T III, Ruddiman WF, Street-Perrott FA, and Bartlein PJ (1994) *Global Climates Since the Last Glacial Maximum*. Minneapolis, MN: University of Minnesota Press.

Archaeological Applications

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Introduction

The long-term history of human impact on vegetation – or cultural landscape history – has been a major research field within paleobotany/paleoethnobotany/archaeobotany and palynology/paleoecology in Europe since the 1960s. It is a vast topic covering a wide range of questions, approaches, and methodologies and has produced a large body of data, interpretations, and hypotheses.

Reconstruction of past cultural landscapes has traditionally been based upon the study of plant macrofossils (such as fruits, seeds, and leaves; see [Use in Environmental Archaeology](#)) and pollen grains preserved in peat and lake sediments or in cultural layers at archaeological sites. Pollen analysis has been a part of both archaeobotanical and paleoecological research. It was first applied to trace forest history and infer information on past climate. Thanks to Firbas's work on the identification of cereal pollen grains ([Firbas, 1937](#)), [Iversen \(1941\)](#) could, for the first time, identify periods of cereal cultivation in his pollen diagrams. He also could describe a typical succession of pollen assemblages ascribed to human-induced deforestation for the purpose of agriculture, known in the literature as Iversen's 'landnam' ([Figure 1](#)).

Since the publication of the 'landnam' theory by Iversen, the interest in 'human-impact' history, as inferred from 'offsite' pollen stratigraphic data (i.e., data from lake or peat stratigraphies at sites located outside the areas of archaeological findings), has increased rapidly in central and northern Europe. The anthropogenic influence on vegetation has been studied by numerous palynologists from the 1950s to the 1970s (e.g., in Sweden, [Berglund, 1969](#); [Figure 2](#)). Moreover, pollen analysis became gradually more sophisticated during the 1970s and 1980s, thanks to significant progress in the technology of microscopy and in pollen-morphological studies. Publication of increasingly detailed pollen identification keys (e.g., [Punt et al., 1976–2009](#)) greatly improved the value of pollen data for reconstruction of herb-rich, human-made vegetation, including cultivated fields, pastures, and hay meadows.

Since the 1960s, archaeobotanical research has generally included both 'onsite' (i.e., within the archaeological excavation and its findings) and 'offsite' (see above) pollen analytical investigations with the aim of reconstructing the landscape and plant environment in which humans were living and from which humans extracted food and material for building and other purposes. Pollen analytical studies within archaeobotany have mostly developed independently from 'offsite' palynological and paleoecological research on past cultural landscape (see [Use in Environmental Archaeology](#)).

In America, pollen analysis flourished as a paleoecological method, thanks to the pioneer work of Davis (see [Stand-Scale Palynology](#)), and since the 1970s, and especially the 1980s, human impact has also been studied (see review

in [McAndrews, 1988](#)). There are also very good examples of pollen analytical records of past human impact in Japan, China, the rest of Asia, and Australia (see [Huntley and Webb, 1988](#)).

This paper reviews pollen analytical methodology as a tool to reconstruct past human impact on vegetation, past land use, and past cultural landscapes with a focus on western Europe. It presents and discusses the different interpretation approaches that are used to translate pollen data in terms of human-impact and cultural landscape history.

What Questions?

Past Human Impact and Archaeology

Cultural landscape history has been a central research area within archaeology/archaeobotany. Paleobotanical questions raised by archaeologists are primarily related to the use of plants by man and to the local and regional vegetation/landscape/environment surrounding the settlement or monument. These questions might include the possible impact and consequences of different cultures and their activities on the vegetation/landscape/environment and similarly on the possible influence of the vegetation/landscape/environment on humans and their activities, and on how humans used landscapes for collection or production of food and for material to build houses, domestic utensils, etc. The types of land use (e.g., coppicing, slash and burn, pollarding, hay making, grazing, cultivation), and the spatial extent of managed areas, have also been recurrent questions. Both onsite and offsite pollen and plant macrofossil studies have been widely used in an attempt to answer these questions (see [Use in Environmental Archaeology](#)), of which many have been real challenges for pollen analysts and paleoecologists.

Past Human Impact and Climate

Attempts to separate human impact from the effects of climate change on past terrestrial ecosystems have frustrated paleoecologists. Some vegetation changes, such as shifts in forest composition, might be very difficult to ascribe to either climate change or human impact. The slow migration of *Fagus* (beech) through Europe and its late establishment in central and northwestern Europe, is a classic example of such debates (see [Stand-Scale Palynology](#)). The mid-Holocene 'elm decline' is another such example (see [Northern Europe](#)).

The possible feedback effect of human-induced deforestation on past climate changes (e.g., [Ruddiman, 2003, 2005](#)) is also an important and challenging area within climate research that has not yet been satisfactorily solved (e.g., [Gaillard et al., 2000](#)). Such questions require reliable reconstructions of the spatial distribution of human-induced open land in the past,

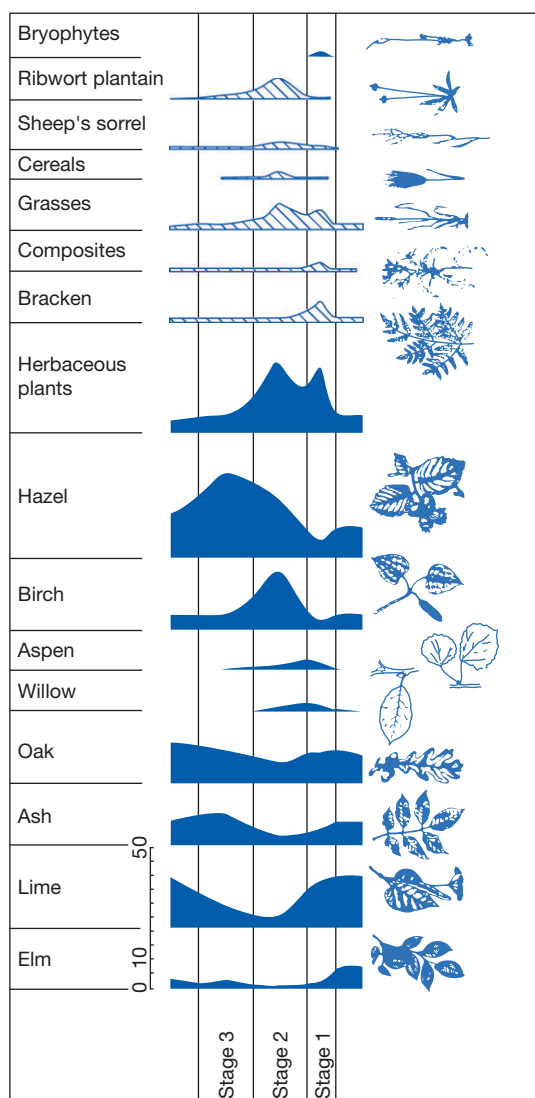


Figure 1 Generalized pollen diagram for a Stone Age 'landnam' with its three phases: (1) first clearance phase, decrease of tree taxa (especially elm), short-lived increase of herbaceous flora; (2) clearing by stone ax and fire, strong decrease in linden, decrease of ash and oak, and increase of anthropogenic indicators such as narrowleaf plantain, first cereal pollen; willow (*Salix*) and aspen (*Populus*) reach low maxima in the beginning, followed by peak of birch; and (3) decrease of birch and human-impact indicators, peak of hazel (*Corylus*), regeneration of elm, linden, oak, and ash. Reproduced from Iversen J (1973) *The development of Denmark's nature since the last glacial*. *Danmarks Geologiske Undersøgelse II V/7-C*: 1–26.

which is still a challenge (e.g., Gaillard et al., 2010). A major achievement of the large ESF (European Science Foundation) project 'European Paleoclimate and Man' was the evaluation of land surfaces cleared of forests during various archaeological periods (e.g., Frenzel et al., 1994). However, many of the reconstructions were based on a simple interpretation of nonarboreal pollen (NAP) percentages (Figure 3(a) and 3(b)). We now know that NAP percentage is not a reliable quantitative index of open land cover (Sugita et al., 1999, see below).

Past Human Impact and Nature Conservation

The reconstruction of cultural landscape history may also help solve questions related to nature conservation. There is an increased awareness among conservation biologists and decision-makers of the value of long-term historical information. Ecological processes operating at longer timescales must be known and understood if an ecosystem is to be maintained or restored for the future (e.g., Bradshaw and Hannon, 2004). Questions concern the need of management continuity or the necessity of a dynamic management for the maintenance of particular species or biodiversity as a whole, for instance, in herb-dominated ecosystems (e.g., Gaillard et al., 2009). The factors necessary for the maintenance of sustainable forest dynamics (such as fire and grazing) are also a subject of constant debate (e.g., Greisman and Gaillard, 2009; Olsson et al., 2010).

Approaches to Reconstruction of Past Human Impact from Pollen Data

The Indicator-Species Approach

The indicator-species approach relies on the modern ecology of species, that is, on the indicator value of species in terms of environmental characteristics such as soil properties, climate (temperature, humidity), and human-induced factors such as deforestation, cultivation, grazing, and mowing. Therefore, interpretation is based on the 'modern analog' concept, that is, the hypothesis that modern and past species' ecological requirements are comparable. In the indicator-species approach *sensu stricto*, the amount of pollen found is not critical. The presence of few pollen grains from an indicator type is the basis for interpretation. However, whether presence is scattered in time or continuous is taken into account and might lead to different conclusions on land use and human impact. Behre (1981) published a very influential paper on the use of various pollen taxa as indicators of anthropogenic vegetation types and land use (Figure 4).

The indicator-species approach and Behre's scheme of human-impact pollen indicators were used successfully by numerous European pollen analysts during the 1980s and 1990s, and were also adapted for northern Europe, for example, northern Norway (Vorren, 1986) and northern Finland (Hicks, 1992). The approach provided a way to identify variations in land-use/landscape management through time. It also made it possible to describe relative changes in the extent of total human impact and different management types. The latter may be described as a development of the indicator-species approach toward a semiquantitative reconstruction method. It was achieved by grouping and summing up the 'anthropogenic indicators' in order to obtain percentage curves of contrasting land uses (e.g., Berglund, 1991; Berglund and Ralska-Jasiewiczowa, 1986).

Anthropogenic pollen indicators

The most widely used anthropogenic indicators in pollen diagrams of northern Europe are pollen grains of cultivated crops (mainly cereals but also buckwheat (*Fagopyrum*), common flax (*Linum usitatissimum*)), narrowleaf plantain

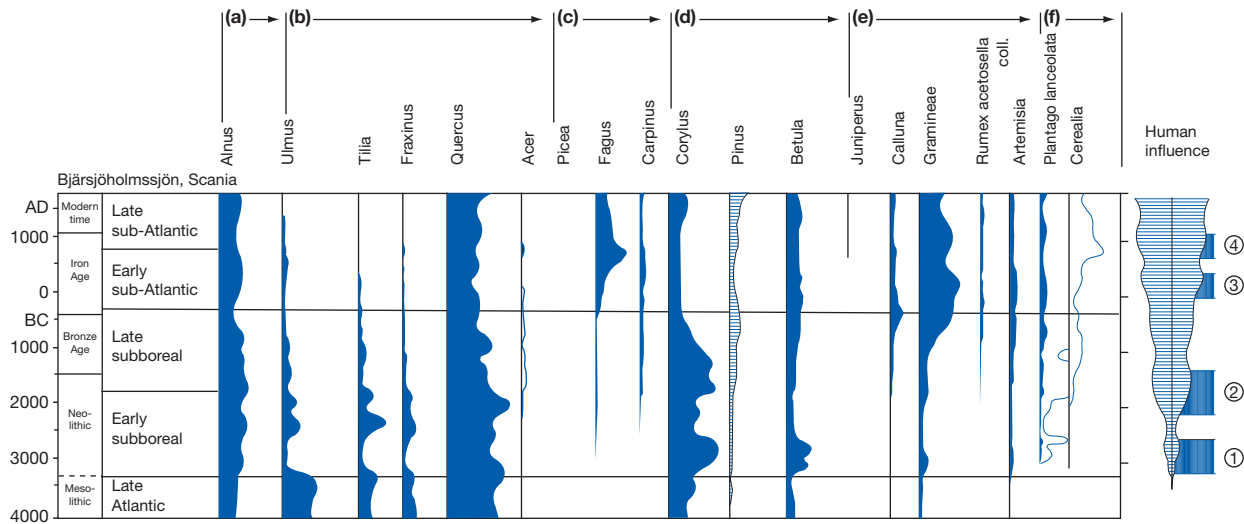


Figure 2 Generalized human-impact diagram for the province of Skåne, southern Sweden, based on the pollen record from Lake Bjärsjöholmssjön. Major indicator pollen for (a) alder carr, (b) broad-leaved forest, (c) late migrating trees, (d) light-demanding trees, pine (*Pinus*), assumed by the author to be long-distance pollen, (e) general human-impact indicators, (f) pastures (*Plantago lanceolata* = narrowleaf plantain), and arable fields (*Cerealia* = cereals). Phases of human impact increase (expansion) 1–4 were later correlated to the following archaeological periods: (1) Early Neolithic, (2) Middle Neolithic, (3) Late Bronze Age, (4) Late Iron Age. Reproduced from Berglund BE (1969) Vegetation and human influence in South Scandinavia during Prehistoric time. In: Berglund BE (ed.) *Impact of Man on the Scandinavian Landscape During the Last Post-Glacial*, pp. 9–28 (Oikos Supplement 12).

(*Plantago lanceolata*), sage (*Artemisia*), common and garden sorrel (*Rumex acetosa/acetosella*), and heather (*Calluna*), together with pollen of the grass, sedge, and daisy families (Gramineae, Cyperaceae, and Compositae, respectively), three groups that may include species from a wide range of vegetation types, including natural, undisturbed communities. The possibility of the development of reed (*Phragmites communis*) or other shore vegetation with both grasses and sedges, concomitant to the assumed increase in human impact, should be ruled out before using these two taxa as anthropogenic indicators. Most of these anthropogenic indicators are also the most commonly found NAP in lake sediments and peat deposits of northwestern Europe, and they are often recorded in relatively high quantities. In North America, the equivalent indicators are the cultivated maize (*Zea mays*) and sage, ragweed (*Ambrosia*), sorrel, plantain and the grass, sedge, daisy, and goosefoot (Chenopodiaceae) families (see pollen diagrams and more details on the following examples in McAndrews, 1988). Pollen diagrams from ~2000-m elevation in Mexico show that human disturbance becomes evident from ca. 3.7 ka BP with the first occurrence of *Zea* together with the increase of goosefoot (*Chenopodium*) type, and the daisy and sedge families. In Guatemala, pollen diagrams are characterized from 8 uncal ka BP by low percentages of savanna trees, appearance of shrubs (Melastomataceae), increase of charcoal particles, and first occurrence of maize pollen, which may indicate that the practice of swidden (slash and burn agriculture) already occurred prior to the Maya culture. The latter displays less than 30% tree pollen. The late Preclassic, Early Classic, Late Classic, and Postclassic phases of the Maya period can often be separated by means of variations in the NAP sum. In Ontario, evidence for maize cultivation from AD 1360 consists of maize pollen together with grasses and purslane (*Portulaca*). It has been

suggested that Aboriginal farmers probably contributed to the forest succession by providing oak (*Quercus robur*) and white pine (*Pinus strobus*) with disturbed habitats necessary for seedling establishment. The peak in aspen from AD 1440 to 1570 indicates disturbance. In Connecticut, European farming started around AD 1700, which is indicated by the appearance of ragweed and sorrel pollen. Chestnut (*Castanea*) blight in AD 1913 caused a decline of chestnut pollen and a secondary succession, shown by a rise in birch followed by oak.

There are a large number of additional pollen indicators that might be identified and provide invaluable information on land use. The grouping of anthropogenic pollen taxa was used in a semiquantitative approach to draw percentage curves of human-induced vegetation types through time in several studies on past cultural landscapes. An outstanding example of this approach is found in the multidisciplinary study of the cultural landscape from 6000 BC until modern times in southern Sweden (Berglund, 1991; Figure 5). The detailed grouping used in the project (Table 1) was fully published for the first time in the first edition of the *Encyclopedia of Quaternary Sciences* (2008).

The pollen indicators are grouped into categories of human-induced vegetation as a basis for percentage calculation of groups of pollen types that can be displayed in a so-called human-impact pollen diagram (e.g., Berglund, 1991; Berglund and Ralska-Jasiewiczowa, 1986). This grouping was used within the Ystad project, southern Sweden, and is based primarily on the results of studies on modern pollen–vegetation relationships with statistical analysis and tests of those relationships (Gaillard et al., 1994; see text for more explanations). In Table 1, undifferentiated (undiff.) pollen is not identified further to group of genera (genera type) or to a single genera. It has to be stressed that many pollen types may be related to

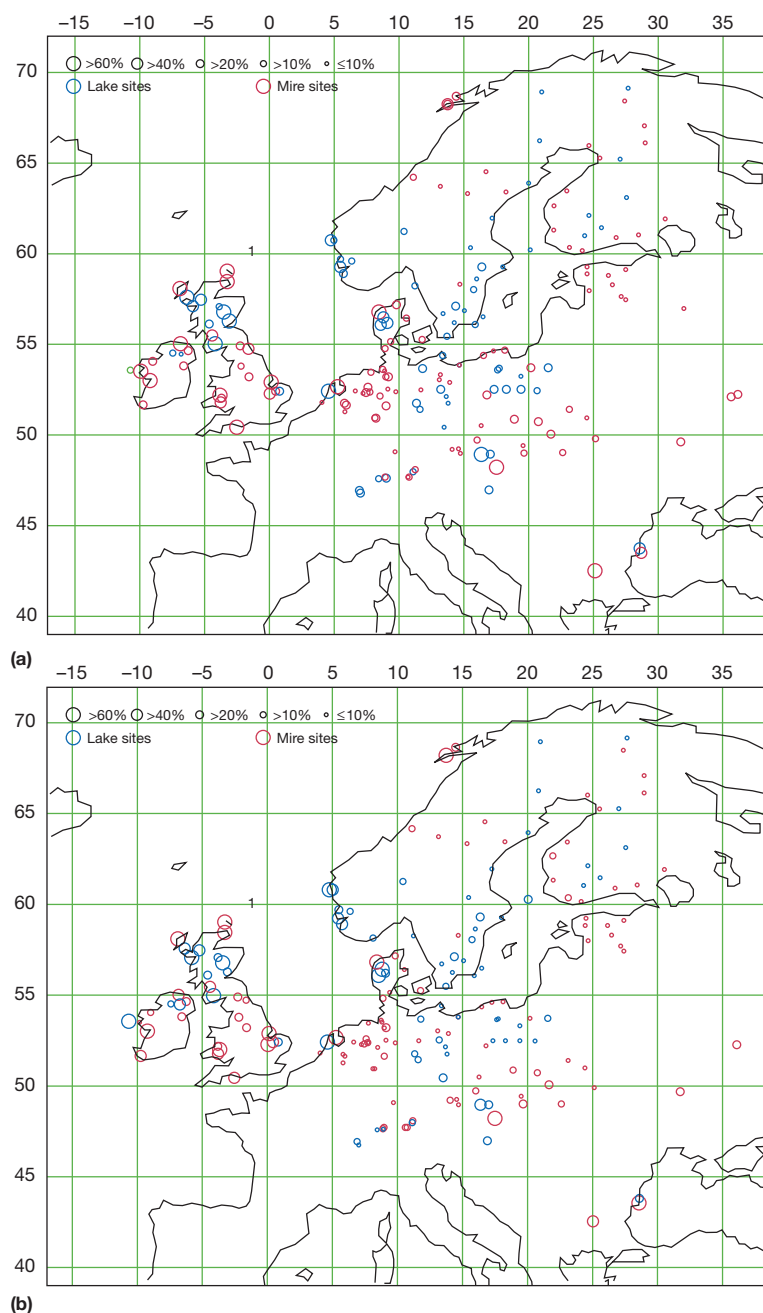


Figure 3 NAP pollen percentages for 195 lowland sites (below 800-m altitude) at (a) 1.8 ka BP and (b) 1.45 ka BP. Lake sites are indicated by blue, mire sites by red circles. The NAP percentages were expected by the authors to ‘express the occurrence of treeless areas in some way’ and to ‘indicate regional differences.’ Sites with high NAP percentages are common in the British Isles and around the North Sea and the Atlantic seaboard for both time windows (1.8 and 1.45 ka BP). Low percentages prevail in northernmost Europe. Intermediate NAP percentages occur in North Central Europe at 1.8 ka BP while the values are low in that area at 1.45 ka BP. The authors also comment that ‘actual degree of deforestation is difficult to estimate from these maps.’ See text for more explanations. Reproduced from Frenzel B, Andersen ST, Berglund BE, and Gläser B (eds.) (1994) *Evaluation of Land Surfaces Cleared from Forests in the Roman Iron Age and the Time of Migrating Germanic Tribes Based on Regional Pollen Diagrams. Palaeoklimaforschung/Palaeoclimate Research*, vol. 12. Stuttgart: Gustav Fischer Verlag.

several of the vegetation categories and that a choice has to be made for the purpose of drawing pollen diagrams. The grouping should, therefore, be understood as one among many possible suggestions. It is the grouping that was most adequate in the south-Swedish context. Examples of pollen taxa that are relatively easy to identify with modern pollen keys (Moore

et al., 1991; Punt et al., 1976–2009) are *Ranunculus acris* type including a large number of *Ranunculus* species, several *Anemone* species (e.g., *Anemone nemorosa* and *Pulsatilla* species), *R. acris* group (*R. acris*, *R. bulbosus*, *R. repens*, and 12 other upland species), *R. ficaria* group (*R. ficaria*), *Anemone nemorosa* group (*A. apennina*, *A. nemorosa*, *A. ranunculoides*, and *A. sylvestris*),

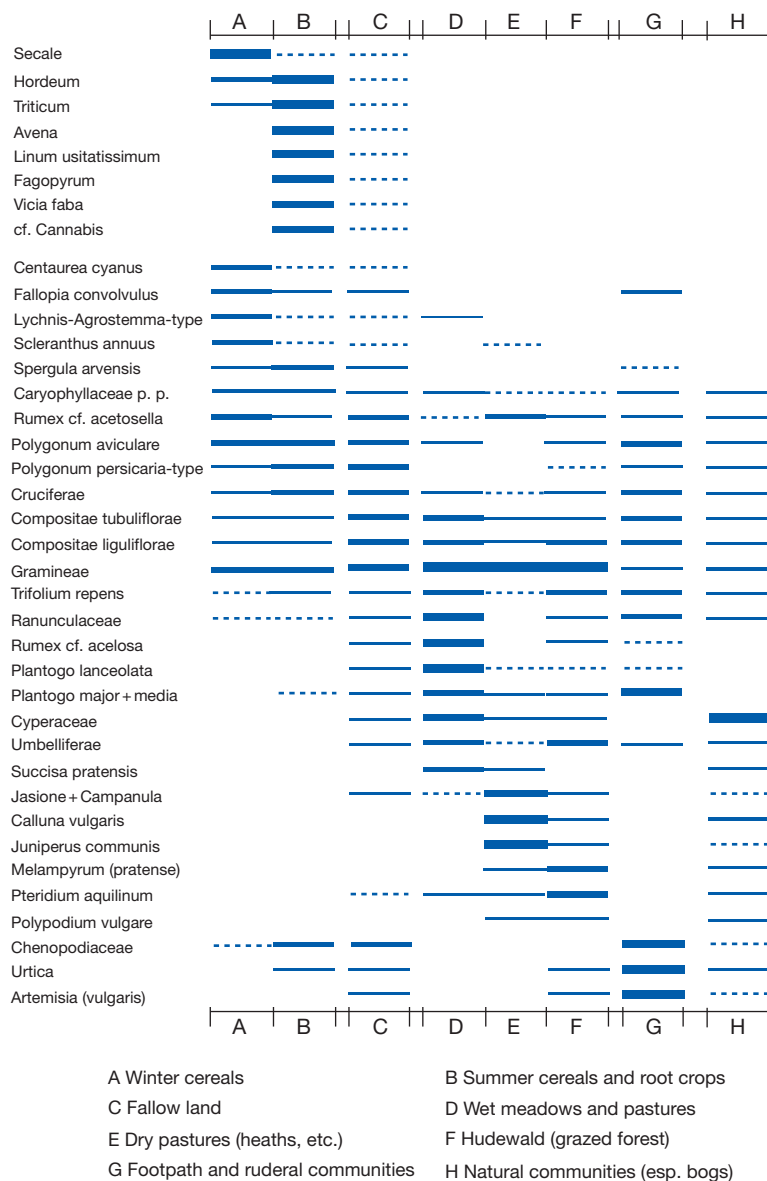


Figure 4 The principal anthropogenic indicators in pollen diagrams and their occurrence in various farming contexts for northwestern Europe (north of the Alps). The frequency of occurrence is shown for the plants themselves and without reference to variation in pollen production and dispersal. Reproduced from Behre K-E (1981) The interpretation of anthropogenic indicators in pollen diagrams. *Pollen et Spores* 23: 225–245.

Anthriscus sylvestris type, *Chaerophyllum hirsutum* type, *Daucus carota* type, *Heracleum sphondylium* type, *Peucedanum palustre* type, *Pimpinella major* type, *Ononis* type, *Trifolium* type (in which *T. pratense*, *T. repens*, and *Medicago* spp. are not too difficult to separate), *Lotus corniculatus*, *Ulex* type (*Genista*, *Ulex*, *Cytisus*), *Vicia* type (*Vicia* spp. and *Lathyrus* spp. with possible separation of several species such as *Lathyrus pratensis*, *Vicia sylvatica*, and *Vicia cracca*), *Cichorium intybus* type (*C. intybus*, *Hieracium* spp., *Hypochaeris* spp., *Lactuca* spp., *Taraxacum* spp., and some other smaller genera), *Scorzonera humilis*, *Centaurea cyanus*, *Centaurea montana*, *Centaurea scabiosa*, *Centaurea nigra* type, *Cirsium* type, *Serratula* type, *Anthemis* type, *Cerastium fontanum* type (mainly *Cerastium* spp., *Stellaria* spp.), *Dianthus superbus* type (*Dianthus* spp., *Petrorhagia* spp.), *Sagina procumbens* type, *Scleranthus annuus* type, *Spergula arvensis* type, *Hypericum perforatum* type, and *Linum*

catharticum (see also Figures 6–14). In Figures 6–14, ‘personal observations of M.-J. Gaillard’ refer to pollen-morphological types that were not described earlier in the literature and that are based on morphological characters observed by M.-J. Gaillard.

Reliable and accurate interpretation of the occurrence of anthropogenic indicators in pollen records requires very careful identification of the pollen grains by means of detailed examination of pollen-morphological characters and a good knowledge of the ecology of the taxa. Straightforward and very precise interpretation is not always possible and may be the exception rather than the rule. The indicator value of species or groups of species may also be evaluated by means of studies of the pollen–vegetation relationship in modern analogs of past human-made vegetation types, for example, in cultivated fields, open or wooded pastures, open or wooded hay

Bjäresjösjön Pollen diagram

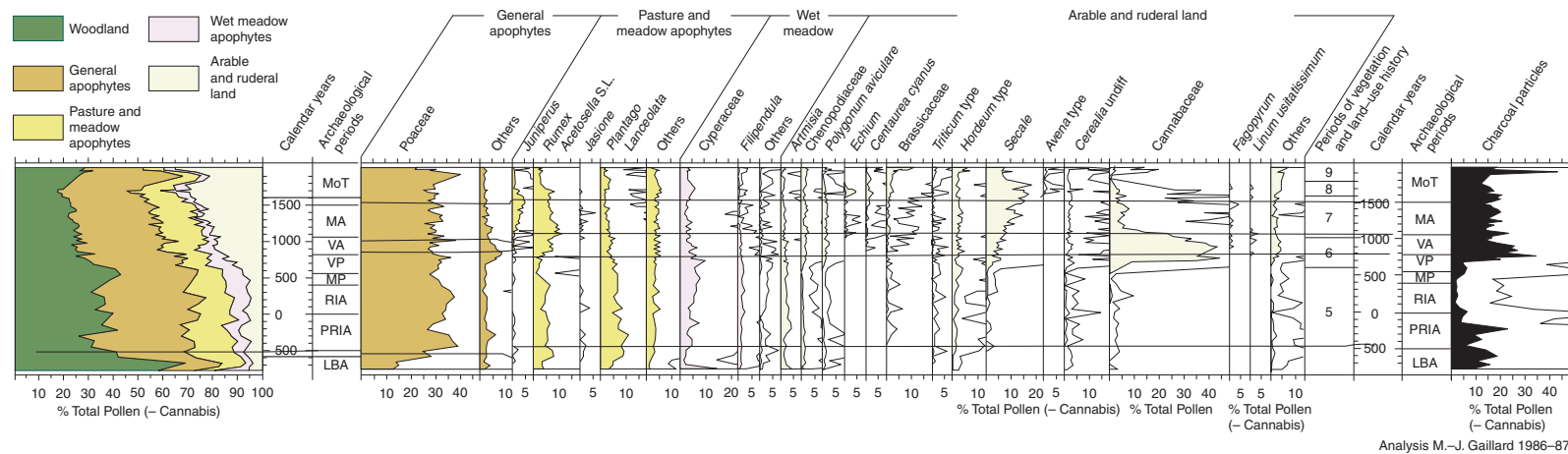


Figure 5 Simplified pollen diagram from Lake Bjäresjösjön (2 ha), southern Sweden. The diagram shows the groups of pollen indicators (see [Table 1](#)) used in a multidisciplinary research project on cultural landscape of the area during the past 6 ka ([Berglund, 1991](#)). The most abundant pollen taxa are presented for herbaceous plants only. The herb pollen types are dominant at Bjäresjösjön with nonarboreal pollen percentage (NAP%) of over 50% from late Bronze Age onward. General apophytes: herbaceous, terrestrial plants that are generally related to human-induced vegetation communities but cannot be ascribed to a particular land use. The record from Lake Bjäresjösjön represents the landscape around the small lake in an area of ~1000–1500-m radius. In a pollen diagram from a larger lake, Lake Krageholmssjön (220 ha), ~2 km north of Bjäresjö, NAP does not reach more than 25% until the ‘modern times’ (MoT). The contrast between these two pollen diagrams from the same area (province of Skåne, southern Sweden) is due to the size of the lakes and not to differences in the local vegetation. The record from Lake Krageholmssjön represents the vegetation of Skåne at the very large spatial scale (100 × 100 km) (see text on pollen calibration for more explanations). Reproduced from [Berglund BE \(ed.\) \(1991\) The Cultural Landscape During 6000 Years in Southern Sweden](#), vol. 41. Copenhagen: Ecological Bulletin.

Table 1 List of types of pollen indicators of human impact**(a) Trees and shrubs of damp soils**

Alnus
Salix
Myrica gale

(b) Shade-tolerant trees and shrubs

Fraxinus
Ulmus
Tilia
Quercus
Acer
Hedera helix
Viburnum
Sambucus nigra
Sambucus racemosa
Cornus sanguinea
Crataegus
Lonicera
Fagus
Carpinus
Picea abies

(c) Light-demanding trees and shrubs

Corylus
Betula
Populus
Pinus
Prunus spp. *Prunus avium/spinosa*
Sorbus aucuparia
Evonymus
Rhamnus cathartica
Cotoneaster spp.
Rosa
 Cf. *Malus sylvestris*
 Cf. *Sorbus intermedia*
Rubus spp.

(d) Cultivated trees

Aesculus hippocastaneum
Juglans regia
Ligustrum vulgaris/Syringa

(e) Forest herbs and ferns

Aegopodium podagraria
Anemone nemorosa group/*Hepatica nobilis*
Humulus
Chrysosplenium
Fragaria
Lamium type
Lysimachia nemorum
Mercurialis
Melampyrum
 Rosaceae undiff.
Sanicula europaea
Scilla type
Stachys sylvatica type
Stellaria holostea
Stellaria nemorum
Symphytum
Trientalis europaea
Vicia type cf. *Lathyrus sylvestris*
Vicia sylvatica type
Viola canina
 Polypodiaceae undiff.
Athyrium filix-femina

(Continued)

Table 1 (Continued)

Dryopteris filix-mas
Dryopteris cristata type
Dryopteris dilatata
Gymnocarpium dryopteris
Lycopodium annotinum
Matteuccia struthiopteris
Polypodium vulgare
Pteridium aquilinum

(f) Dry pastures and heath and 'alvar' vegetation

Juniperus
Calluna
Empetrum
Erica tetralix
 Ericaceae undiff.
Vaccinium spp.
V. cf. V. uliginosum
V. cf. V. vitis-idaea
V. cf. V. myrtillus
Allium spp.
Allium vineale type
Anthyllis vulneraria
Centaurea nigra type
Dianthus type
Genista type
Gentianella campestris type
Globularia
Gypsophila fastigiata
Helianthemum
Jasione montana
Ononis type cf. *O. repens*
Pimpinella saxifraga
Plantago media
Potentilla cf. *P. fruticosa*
Sagina procumbens type
Sedum
Veronica cf. *V. officinalis*

(g) Fresh meadows and pastures

Alchemilla
 Apiaceae undiff.
Anthriscus sylvestris
 Apiaceae cf. *Carum carvi*
Aster type
Bidens type
Campanula type
Cerastium type
Cirsium
Filipendula
Galium type
Geranium type
Heracleum sphondylium
Hypericum perforatum type
 Juncaceae
Knautia
Lathyrus pratensis
Linum catharticum
Lotus type
Medicago sativa
Peucedanum carvifolium group
Pimpinella major type
Plantago lanceolata
 Poaceae undiff.
Polygala vulgaris

(Continued)

Table 1 (Continued)

Potentilla
Prunella type
Primula veris type
Ranunculaceae undiff.
Ranunculus acris type
Ranunculus acris group
Rhinanthus type
Rumex acetosa/acetosella
Saxifraga granulata
Scabiosa
Scorzonera humilis
Serratula
Sinapis type
Succisa pratensis
Trifolium type
Trifolium type cf. *T. medium*
T. pratense
T. repens
Veronica cf. *V. chamaedrys*
Veronica spp.
Vicia type
Vicia cracca type
Viola spp.

(h) Wet meadows and salt meadows

Cyperaceae
Anagallis tenella type
Angelica sylvestris group
Armeria maritima
Caltha palustris
Comarum palustre
Epilobium
Eriophorum
Geum urbanum/G. rivale
Glaux maritima
Hydrocotyle vulgaris
Impatiens
Ligusticum scoticum
Lychnis flos-cuculi
Lysimachia vulgaris group
Mentha type
Menyanthes trifoliata
Oenanthe fistulosa type
Plantago coronopus
Ranunculus ficaria group
Ranunculus flammula group
Schoenoplectus type
Thalictrum
Triglochin
Trollius
Equisetum
Sphagnum

(i) Ruderal communities

Apiaceae cf. *Aethusa cynapium*
Anthriscus caucalis type
Artemisia
Anthemis type
Chaerophyllum hirsutum type
Chenopodiaceae
Cichorium intybus type
Conium maculatum
Daucus carota

(Continued)

Table 1 (Continued)

Euphorbia
Hornungia type
Linaria
Myosotis discolor
Plantago major
Reseda lutea type
Rumex obtusifolius type
Scleranthus perennis
Scrophularia type
Silene dioica type
Sonchus oleraceus type
Solanum
Trifolium campestre
Urtica
(j) Cultivated land
Cerealia undiff.
Avena type
Hordeum type
Secale cereale
Triticum type
Zea mays
Cannabis type
Cannabis type cf. *C. sativa*
Fagopyrum esculentum
Linum usitatissimum type
Agrostemma githago
Anagallis arvensis type
Centaurea cyanus
Faloppia convolvulus type
Polygonum aviculare type
P. persicaria group
Ranunculus arvensis
Scleranthus annuus
Trifolium type cf. *T. arvense*
Valeriana locusta type
Viola arvensis type

meadows. This has been achieved in southern Sweden by selecting modern sites characterized by old-fashioned, traditional land use and management, collecting moss polsters in homogenous vegetation types for pollen analysis, and performing vegetation inventories (e.g., Gaillard et al., 1994). Similar studies have been performed in many parts of Europe, most recently in the Pyrenees and the Southern Alps (Court-Picon et al., 2006; Mazier et al., 2006; see Section 'The Comparative Approach').

Crop cultivation

The classical indicators for cultivation in Europe are pollen of cereals and related weeds (e.g., cornflower (*Centaurea cyanus*) and corn cockle (*Agrostemma githago*); Behre, 1981). Of these, it is only a few weeds that can be identified to species and that are restricted to cultivated fields. The cereal pollen type may include all known cultivated species within the genera wheat (*Triticum*), barley (*Hordeum*), oat (*Avena*), and rye (*Secale*), as well as wild grasses from genera such as *Glyceria*, *Elymus*, and *Bromus*. Moreover, among cereals, only the pollen of rye can be identified to species, but part of the pollen grains from rye may belong to the pollen-morphological type *Hordeum* (e.g., Andersen, 1979; Beug, 2004). Therefore, most percentage

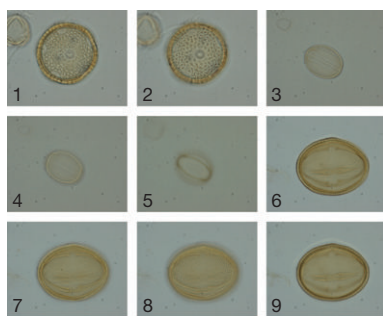


Figure 6 Pictures of pollen grains from plant taxa indicative of human impact. All pictures were taken at magnification 1000 $\times$, except *Succisa pratensis* (400 $\times$). All pollen specimens are from the reference collection at the Paleocology Laboratory, School of Natural Sciences, Linnaeus University, Kalmar, Sweden. They were prepared from dried material collected in field from fresh plants during spring and summer, 1995 and 1996. They were embedded in glycerine jelly after acetolysis. Pictures taken by M.-J. Gaillard. Names in brackets are the pollen-morphological types according to Punt et al. (1976–2009) and Moore et al. (1991). Attribution to man-made vegetation types according to Table 1. Dry pastures, heaths, and alvar vegetation – 1, 2: *Dianthus deltoides* (*D. superbus* type) (pink family); 3–5: *Genista pilosa* (hairy greenweed) (*Genista* type) (characteristic of *Calluna* (heather) heaths, burned); 6–9: *Gentianella uliginosa* (*G. campestris* type) (gentian family).

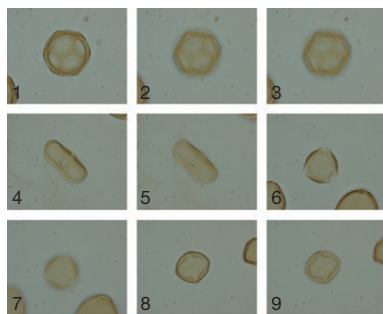


Figure 7 Dry pastures, heaths, and alvar vegetation – 1–3: *Gypsophila fastigiata* (*G. repens* type) (southeastern European species, also occurring in 'alvar' vegetation, a particular environment on the islands of Öland and Gotland in the Baltic Sea, and in Estonia); 4–5: *Pimpinella saxifraga* (*P. major* type); 6–9: *Potentilla fruticosa* (*Potentilla* cf. *P. fruticosa*, personal observations, M.-J. Gaillard; a boreal-subalpine species in northwestern Europe are generally rare, however, common in 'alvar' vegetation) (refer to caption of Figure 6).

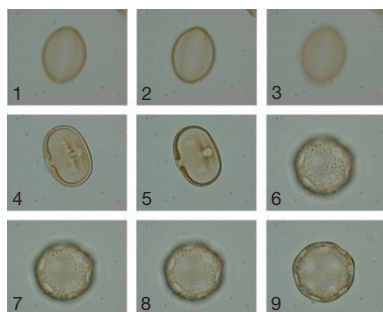


Figure 8 Fresh meadows and pastures – 1–3: *Veronica chamaedrys* germander speedwell (*Rhinanthus* type, *Veronica* cf. *V. chamaedrys*, personal observations, M.-J. Gaillard); 4, 5: *Vicia cracca* (tufted vetch) (*V. cracca* type); 6–9: *Cerastium arvense* (chickweed) (*C. fontanum* type) (refer to caption of Figure 6).

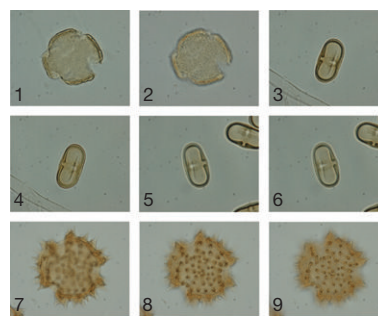


Figure 9 Fresh meadows and pastures – 1, 2: *Ranunculus repens* (creeping buttercup) (*R. acris* type) (*R. acris* group); 3–6: *Anthriscus sylvestris* (*A. sylvestris* type) (umbellifer family); 7–9: *Scorzonera humilis* (Viper's grass) (*S. humilis*) (characteristic of hay meadows in southern Sweden) (refer to caption of Figure 6).

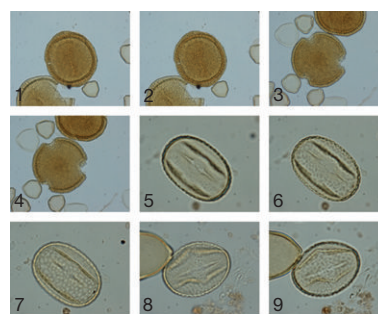


Figure 10 Fresh meadows and pastures – 1–4: *Succisa pratensis* (devil's-bit scabious) (*Succisa*); 5–9: *Trifolium pratense* (meadow clover) (*Trifolium* type, cf. *T. pratense*, personal observations, M.-J. Gaillard) (refer to caption of Figure 6).

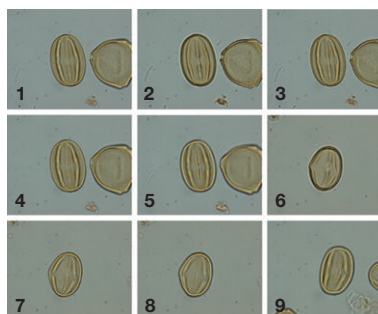


Figure 11 Fresh meadows and pastures: 1–9: *T. repens* (creeping clover) (*Trifolium* type, cf. *T. repens*, personal observations, M.-J. Gaillard). Note: the grain in the right part of pictures 1 to 5 is pollen of *Corylus avellana*. (refer to caption of Figure 6).

curves of rye in the literature might well be an underrepresentation of the species. The pollen of *Hordeum* type, *Avena* type, and *Triticum* type might include different genera and species of both wild grasses and cultivated cereals (Andersen, 1979; Beug, 2004): barley (*Hordeum*) type – *Glyceria* species, some species of *Bromus* and *Agropyron*, *Elymus arenarius* – oat (*Avena*) type – wild oat (*A. fatua*), lopsided oat (*A. strigosa*), some species of *Setaria* – wheat (*Triticum*) type – some species of *Bromus* and European beach grass (*Ammophila arenaria*).

The difficulties in matching cereal pollen-morphological types with true cultivated cereals imply that records of cereal

Viking-age landscape
in Bjäresjö, ca. AD 900

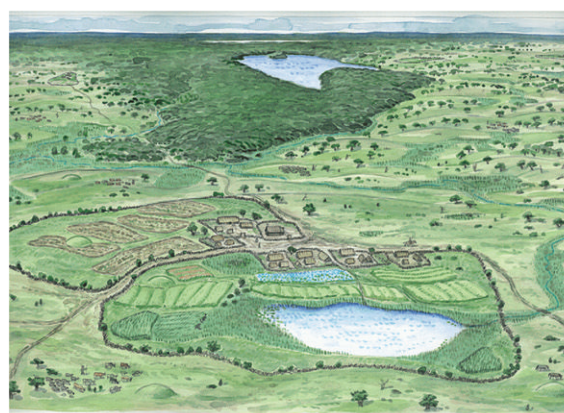
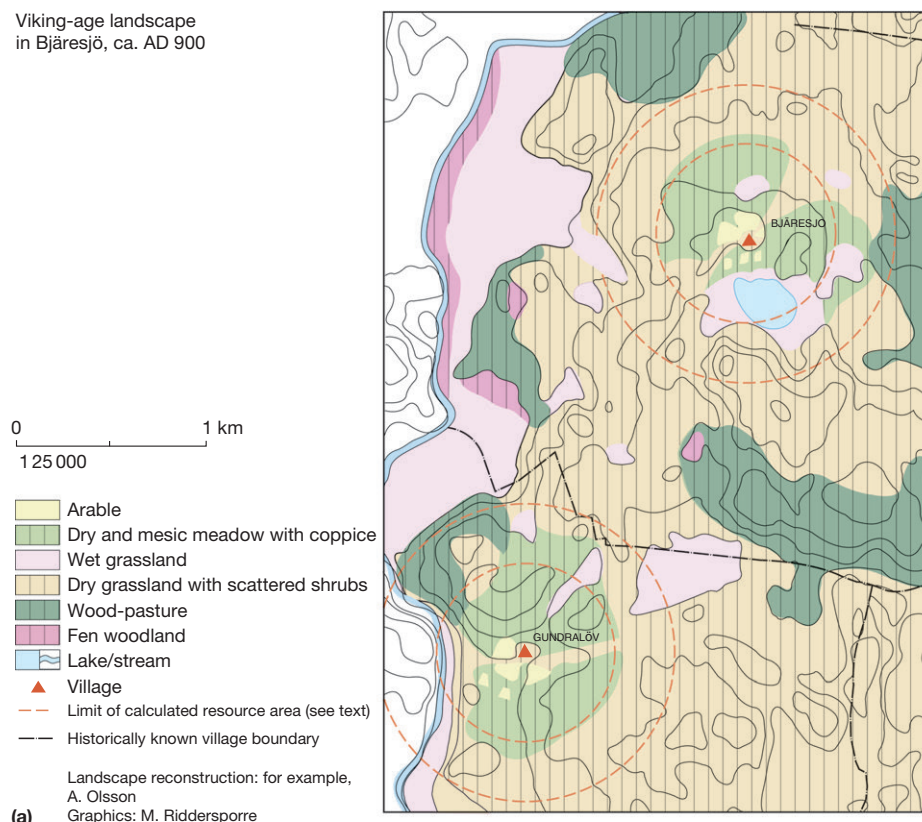


Figure 12 Reconstruction of the Viking Age landscape at Bjäresjö, southern Sweden, based upon studies in archaeology, pollen and plant macrofossil records, and agroecosystem modeling (nutrient flow systems and resource area analysis). (a) Map of the Bjäresjö area showing the possible distribution of the different land uses. (b) Artist's picture of the landscape based on the pollen diagrams and maps produced for the areas around the lakes of Bjäresjösjön (see (a) above, [Figure 5](#)) and Krageholmssjön (large lake in the background). (a) Reproduced from [Berglund BE \(ed.\) \(1991\) The Cultural Landscape During 6000 Years in Southern Sweden](#), vol. 41. Copenhagen: Ecological Bulletin. (b) Reproduced by permission of the artist Nils Forshed, Skövde, Sweden.

pollen have to be interpreted with caution and preferably in conjunction with other pollen data (findings of pollen from weeds for instance) and related paleoecological data, especially plant macrofossil records.

Slash and burn

The practice referred to as slash and burn represents a form of woodland use for agricultural purposes that are strongly associated with Nordic countries. It involved felling of trees and, after drying, setting the wood on fire and then sowing cereals

(often rye) in the clearings; in this way, the poor soils of forested areas were enriched with ash resulting from burning. After the crop was harvested, the area usually reverted to grassland and eventually woodland regenerated. In the eighteenth and nineteenth centuries, slash and burn was intensively practiced in Scandinavia, with the result that the cycle was reduced to 20 years or less, so woodlands failed to fully regenerate. In Finland, the practice coincides with the introduction of arable farming in the Neolithic period and lasted until the early twentieth century ([Vuorela, 1986](#)). In Sweden, the

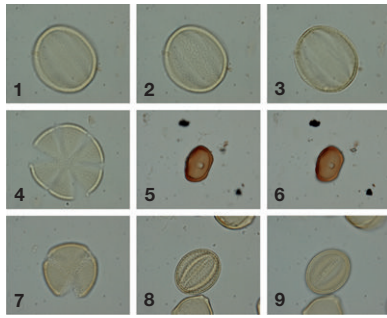


Figure 13 Fresh meadows and pastures – 1–4: *Prunella vulgaris* (self-heal) (*Prunella*); 5–6: *Daucus carota* (wild carrot) (*D. carota* type) (also in dry pastures and ruderal vegetation); 7–9: *Anemone nemorosa* (*R. acris* type *A. nemorosa* group) (characteristic of wooded hay meadows; also in broad-leaved woods) (refer to caption of [Figure 6](#)).

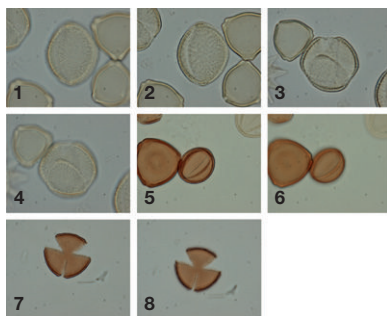


Figure 14 Wet meadows – 1–4: *R. ficaria* (lesser celandine) (*R. acris* type *R. ficaria* group, *R. ficaria*); 5–8: *R. flammula* (lesser spearwort) (*R. acris* type *R. flammula* group) (refer to caption of [Figure 6](#)).

practice was mainly associated with spruce forests ([Lagerås, 1996, 2000](#)). It was common from the time of spruce expansion (time transgressive from northeastern to eastern and southern Scandinavia, from ca. 4 uncal. ka BP in the northeast to ca. 1 ka BP or later in the south) but ceased in the early twentieth century when it fell into disfavor because of associated fire hazards.

There are no proper pollen indicators of slash and burn, but findings of pollen from cereals (rye in particular, or oat (*Avena*) type), combined with high amounts of microscopic and macroscopic charcoal fragments in the peat or lake sediment profiles, are good indications of the occurrence of this practice in the neighborhood of the investigated site (e.g., [Lagerås, 1996, 2000](#)).

Pasture lands/grazing

Common and garden sorrel, narrowleaf and common plantain, white clover (*Trifolium repens*), prostrate knotweed (*Polygonum aviculare*), devil's-bit scabious (*Succisa*), sheep's bit/bellflower (*Jasione/Campanula*), heather, juniper, and the buttercup (*Ranunculaceae*), umbellifer (*Umbelliferae*), daisy, grass, and sedge families are all indicators of grazed areas according to the scheme of [Behre \(1981\)](#). However, they are also common in hay meadows and in ruderal (waste ground) communities around cultivated fields, on fallows, and along tracks and natural herb-rich communities (in particular,

common sorrel, narrowleaf and common plantain, white clover, prostrate knotweed, and the buttercup, umbellifer, daisy, grass, and sedge families). The specific identification of pollen from sorrels or clovers is not simple. It is not uncommon that pollen analysts do not separate species within these genera and simply present curves for the genus *Rumex* (sorrel) or the group *Rumex acetosa/acetosella* (common and garden sorrels), and for the group *Trifolium* type (*Trifolium* species and other genera and species within the clover family) or for the clover family (*Fabaceae* undiff.). Therefore, these pollen types are generally used as indicators of pasture lands and meadows, without any attempt to separate the two types of land use. Moreover, part of the pollen from these taxa may originate from contemporaneous arable fields and other human-modified environments around settlements and along tracks. Some palynologists, though, do attempt to further identify genera and species groups within the buttercup, umbellifer, clover, pink (*Caryophyllaceae*), and daisy families, and within the genera *Rumex* (sorrel), and *Trifolium* (clover), etc.; thanks to the identification keys published in [Punt et al. \(1976–2009\)](#) ([Table 1](#); [Figures 6–14](#)).

Wood pastures and forest grazing

It is not possible to confidently reconstruct the past existence of wood pastures and forest grazing (denser woods where cattle might graze and/or browse) from pollen data. This is because the representation of herbs will be overwhelmed by the dominance of tree pollen, even though they might be more common in grazed woods. Further, herbs and shrubs flower much less (produce fewer pollen grains) in wooded vegetation than in open landscapes.

However, careful examination of pollen assemblages might sometimes be interpreted in terms of wood pastures. In southern Sweden, it was proposed that a shift in pollen percentages (and pollen influx) from a dominance of broad-leaved trees such as oak (*Quercus*), ash (*Fraxinus*), elm (*Ulmus*), and linden (*Tilia*) (in mixture) to a clear dominance of birch (*Betula*) and/or oak, sometimes in connection with very slight increases in herb and/or shrub pollen, might indicate transformation of the mixed broad-leaved forests into wood pastures. Birch and/or oak trees might have been favored/preserved for the purpose of shadowing the grazing land, increasing its fodder production, and (in the case of oak) producing food (acorns) for pigs (e.g., [Lagerås, 2000](#)).

Juniper-rich pastures (the south-Swedish 'fälad') may be indicated by continuous and relatively abundant occurrences of juniper, together with significant representation of grasses, sedges, and sorrel ([Gaillard et al., 1991](#); [Figure 5](#)).

Hay meadows/mowing and the infield-outland/outfield system

As mentioned above, common sorrel, narrow-leaved and common plantain, white clover, prostrate knotweed, and the buttercup, umbellifer, daisy, grass, and sedge families are all indicators of both pastures and hay meadows. None of these taxa can be ascribed exclusively to the practice of mowing. True indicators of mowing probably do not exist, but the occurrence of certain genera or species in combination, associated with high percentages of Gramineae and narrowleaf plantain, might be a strong indication of the occurrence of hay meadows.

Examples of such pollen taxa are, for southern Scandinavia, Viper's grass (*Scorzonera humilis*) (Figure 9), yellow rattle (*Rhinanthus*), fairy flax (*Linum catharticum*), milkwort (*Polygala*), grass of Parnassus (*Parnassia*), devil's-bit scabious (*Succisa pratensis*) (Figure 10), primrose (*Primula*), and eyebright (*Euphrasia* type), which all have very specific pollen morphologies (Gaillard et al., 1994; Punt et al., 1976–2009; Table 1).

The concept of 'infield-outland system' is known in particular from the historical agrarian systems in southern Scandinavia but has many analogs in agrarian systems of Europe (Gaillard et al., 2009). It implies that arable land, together with hay meadows, are confined to the 'infields' that were separated by a fence from the surrounding 'outland' mainly used for grazing, collecting fuel, and other extensive activities. In southern Sweden, the past occurrence of the infield-outland system has been tentatively inferred from pollen data by assuming that pollen assemblages, including indicator taxa of hay meadows and cultivated cereals (often rye), also indicate the existence of infields (e.g., Gaillard et al., 1991; see Section 'The Comparative Approach'; Figure 12(a) and 12(b)).

Pollarding

Tree pollarding (i.e., pruning or cutting back upper branches of certain species of deciduous trees) has been an important element of agrarian systems in many parts of Europe. However, tracing its history on the basis of pollen data is not possible,

either using the indicator-species or the comparative approach (see also below; Gaillard et al., 1994; Figures 15 and 16).

The Comparative Approach

Parallel to the increasingly more sophisticated use of the 'indicator-species' approach, the 1980s were marked by a renewed interest in the 'comparative approach.' This involves the characterization of a range of modern vegetation types by means of contemporary pollen spectra (usually from surface samples) and then the comparison of these spectra with fossil pollen assemblages (Birks and Gordon, 1985). Until the 1990s, the use of modern pollen spectra as an aid in the reconstruction of past vegetation was rarely attempted in northwestern Europe. It was argued that the present-day flora, vegetation, and landscapes of that region were so heavily influenced by millennia of human influence that most modern pollen spectra could be expected to have little or no similarity with past fossil assemblages. Moreover, other disadvantages and limitations of the comparative approach, such as the different sources of modern pollen spectra (moss polsters, surface litter, soils, surface sediments, and pollen traps), also discouraged palynologists from applying that approach. Even though human-influenced landscapes have dominated most of the late Holocene, very few reconstructions of past vegetation/landscape units involving modern pollen spectra were attempted for vegetation types such as arable land, pasture land, or meadows.

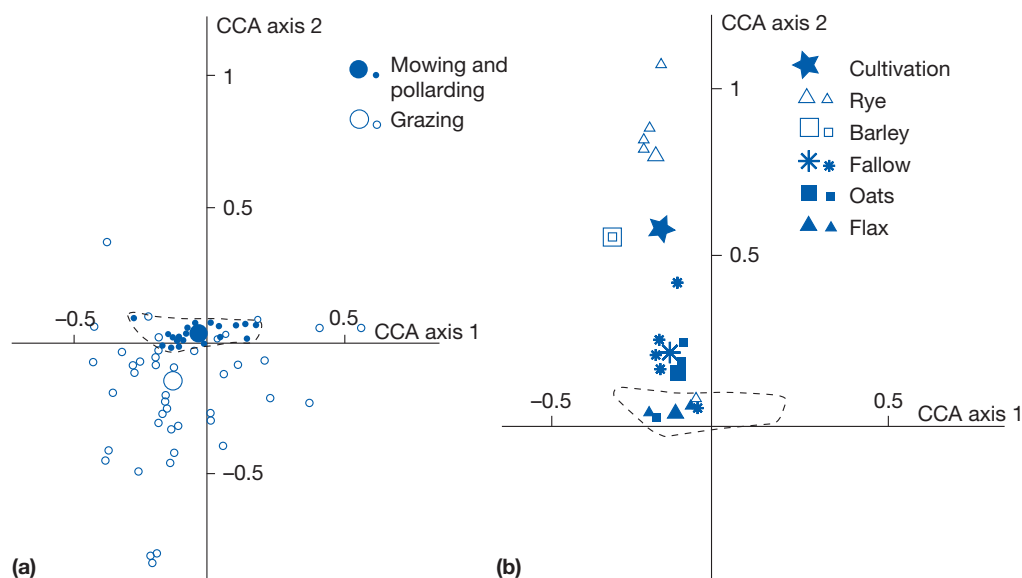


Figure 15 Study of the pollen/vegetation/land-use relationships in southern Sweden. (a) Canonical correspondence analysis (CCA) plots (axes 1 and 2) of 124 surface pollen samples, 60 pollen and spore types, and 16 land-use and landscape variables. The position of all samples, pollen types, and several land-use/landscape variables on axes 1 and 2 is not shown. Only selected samples are shown by their scores (small signatures), and land-use variables are shown by large signatures (dichotomous variables). The plots (a) and (b) show that the land uses – grazing, mowing, and cultivation (rye or barley) – can be separated on the basis of pollen assemblages. Note that pollarding cannot be separated from mowing by means of pollen assemblages. Abbreviations: Cult, cultivation; FAL, fallow; O, oats; FL, flax; M, mowing; P, pollarding; GR, grazing. (a) CCA plot (axes 1 and 2) showing the position of modern pollen samples from 44 grazed and 24 mowed or mowed and pollarded sites in relation to the scores for the grazing, mowing, and pollarding variables. (b) CCA plot (axes 1 and 2) showing the position of modern pollen samples from 11 cultivated sites (16 pollen spectra) with different crops in relation to the scores for the variables crop cultivation, rye, barley, oat, and flax cultivation, and fallow. The scores for the land-use variables (large signatures) are represented by their centroid or weighted average of the site scores (small signatures) where the variable is present. The dashed line delimits the distribution of individual mowed or mowed and pollarded sites. Reproduced from Gaillard MJ, Birks HJB, Emanuelsson U, Karlsson S, Lagerås P, and Olausson D (1994) Application of modern pollen/land-use relationships to the interpretation of pollen diagrams – Reconstructions of land-use history in south Sweden, 3000–0 BP. *Review of Palaeobotany and Palynology* 82: 47–73.

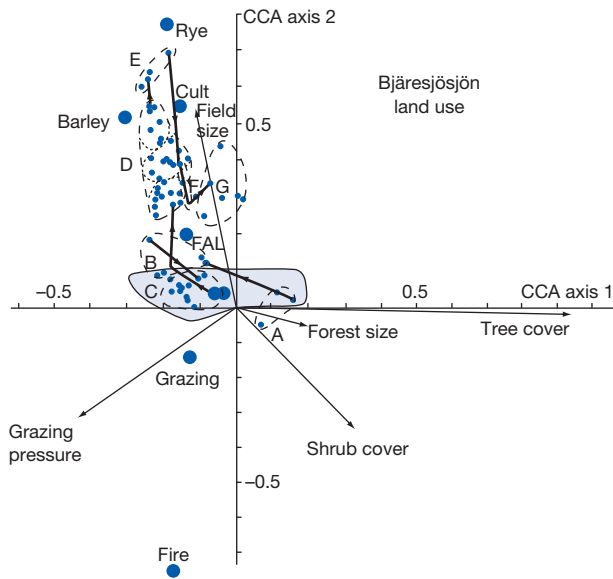


Figure 16 Plot of the 66 fossil pollen samples (black dots) (Lake Bjäresjösjön) positioned as 'passive' samples on CCA axes 1 and 2 of **Figure 15**. Biplot arrows and centroids for 12 selected land-use/landscape variables are shown. Clusters of sample scores (a–g) have been delimited by eye and are joined up in stratigraphical order. Cluster D is subdivided into three subclusters (dotted lines). The dotted area represents the distribution of the scores for mowed or mowed and pollarded sites (see **Figure 15**). Abbreviations: M, mowing; P, pollarding; Fal., fallow; Cult., cultivation. Reproduced from Gaillard MJ, Birks HJB, Emanuelsson U, Karlsson S, Lagerås P, and Olausson D (1994) Application of modern pollen/land-use relationships to the interpretation of pollen diagrams – Reconstructions of land-use history in south Sweden, 3000–0 BP. *Review of Palaeobotany and Palynology* 82: 47–73.

The first attempt at using the comparative approach for human-influenced vegetation in Europe was published by Berglund and Ralska-Jasiewiczowa (1986). Pollen in moss polsters from modern analogs of ancient, nonfertilized grazed and mowed plant communities was used to explore whether these land uses could be separated by means of pollen assemblages. This pioneer work was followed by a series of studies on the pollen/vegetation/land-use relationships in modern analogs of ancient cultural landscapes in southern Sweden (Gaillard et al., 1994, 1998). This resulted in a large database of 120 pollen assemblages and related vegetation, land-use, and other environmental variables such as soil chemistry, from human-influenced sites. One of the most interesting results of this research was the ability to separate hay meadows from grazed pastures by means of pollen assemblages (**Figure 15(a)** and **15(b)**). The use of the comparative approach on a fossil profile from southern Sweden revealed that pollen assemblages dating from the pre-Roman Age up to the Vendel Period (500 BC–AD 800) were significantly similar to the modern pollen assemblages from traditional, nonfertilized hay meadows (**Figure 16**). This was the first convincing independent demonstration of the early occurrence of species-rich meadows comparable to traditionally managed hay meadows. The modern analog and comparative approaches were later adopted by several scientists in Europe, first in Finland, Norway, and Denmark during the 1990s (e.g., Hicks and Birks, 1996) and recently in the French

Pyrenees (Mazier et al., 2006, 2009; **Figures 17** and **18**), and in southern France (Court-Picon et al., 2006).

Calibration of Pollen Data onto Land Cover

Quantification of the human-induced landscape openness in absolute plant abundance values, such as percentage cover, has been an important challenge for palynologists. It has been a research focus in southern Sweden for more than 10 years (e.g., Broström et al., 2004; Gaillard et al., 1998, 2000, 2010; Hellman et al., 2008; Sugita et al., 1999). Pollen assemblages from lake-surface sediments were related to the percentage cover of major landscape units for 40 sites within traditional cultural landscapes in the hope of establishing calibration functions that would allow reconstruction of human-induced landscapes in terms of absolute coverage of vegetation/landscape units such as wooded versus open land, pastureland, cultivated land, etc. Broström et al. (1998) presented empirical data demonstrating that the sum of NAP was not linearly related to open land cover within a 1000-m radius around the pollen sampling point, if pollen samples were from contrasting vegetation regions. Sugita et al. (1999) used a simulation approach involving models of pollen dispersal and deposition (see **POLLSCAPE Model: Simulation Approach for Pollen Representation of Vegetation and Land Cover**) to test the ability of NAP to represent open land. The results showed that NAP percentages cannot be used as a straightforward measure of landscape openness because the share of NAP depends on the proportion and taxa composition of the background pollen component in the pollen assemblage (**Figure 19(A)** and **19(B)**). A different approach than the traditional 'transfer function' methodology needed to be applied in order to tackle this problem. Sugita (2007a,b) proposed the Landscape Reconstruction Algorithm (LRA) approach. It implies estimation of the background term in the equation of the pollen–vegetation relationship according to extended *R*-value (ERV) models (see **POLLSCAPE Model: Simulation Approach for Pollen Representation of Vegetation and Land Cover**). The LRA includes two models, Regional Estimates of Vegetation Abundance from Large Sites (REVEALS) that calculates regional vegetation abundance (RVA, spatial scale 100 × 100 km) using pollen data from large sites (lakes or bogs, ≥100 ha) and the Local Vegetation Estimates (LOVE) model that infers local vegetation abundance within the relevant source area of pollen (RSAP) of the study site (*sensu* Sugita, 1994, ≤100 km² area) using pollen records from small sites (≤10 ha). LOVE uses the REVEALS estimates to calculate the background pollen for small sites and also estimates the RSAP; it then can calculate the local plant abundance within the RSAP (see **POLLSCAPE Model: Simulation Approach for Pollen Representation of Vegetation and Land Cover**). Reliable estimates of pollen productivity estimates (PPEs) are required for application of the LRA (e.g., Broström et al., 2004, 2008). The number of plant taxa for which PPEs are available has rapidly increased in many parts of northern Europe (Broström et al., 2008).

The REVEALS model is now validated in southern Sweden (**Figures 20** and **21**; Hellman et al., 2008) and the Swiss Plateau (Soepboer et al., 2010), and the LOVE model in Michigan and Wisconsin (Sugita et al., 2010) and southern Sweden (Fredh et al. and Cui et al. in progress). The excellent performance of REVEALS to predict the regional coverage of

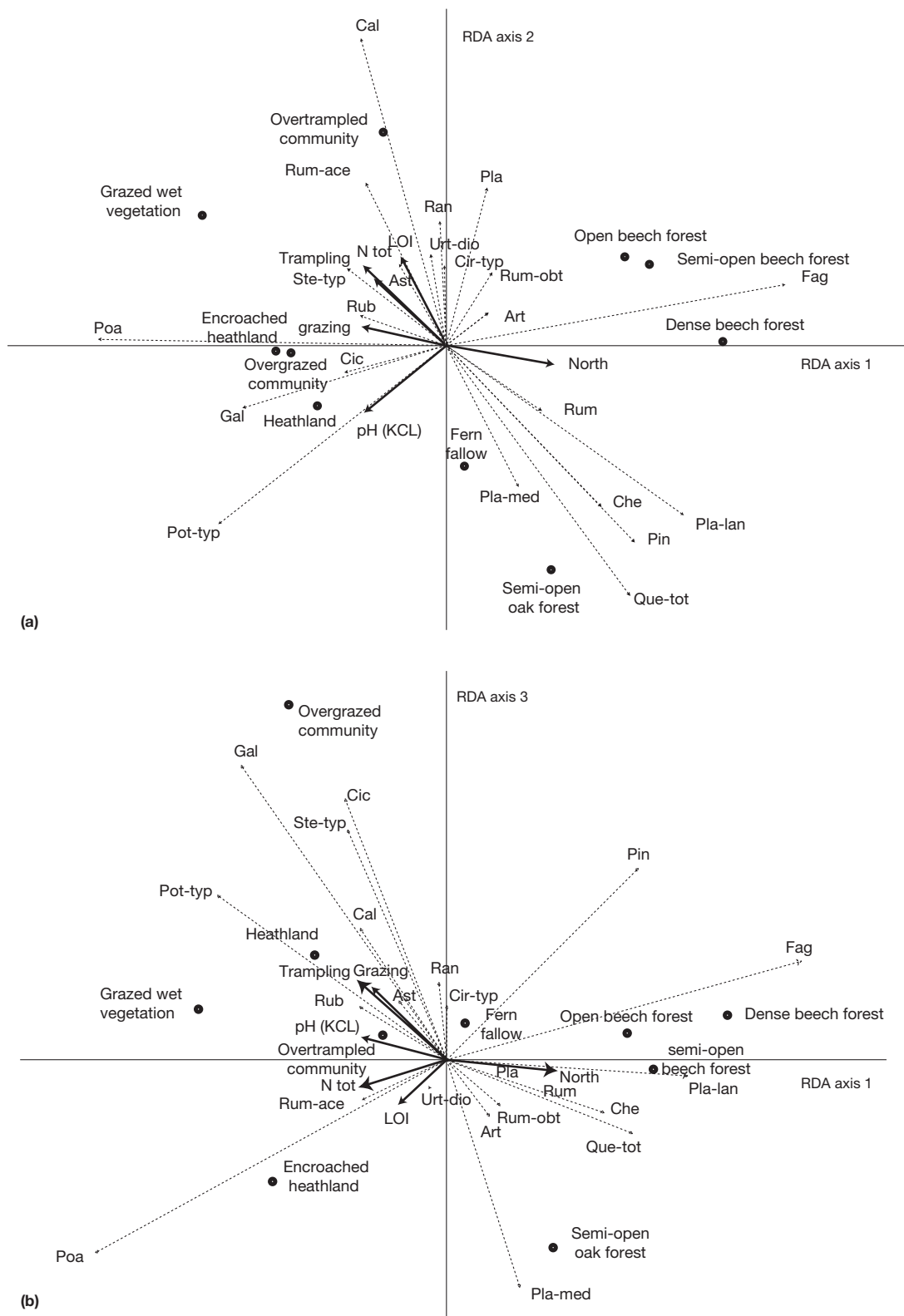


Figure 17 Study of modern pollen assemblages from grazed vegetation in the Western Pyrenean Mountains (France). Redundancy analysis (RDA) of the modern pollen assemblages using 13 explanatory variables (nominal variables are given with their centroids (solid circles) and quantitative variables with solid arrows), four passive variables (italic), and selected taxa (dotted arrows). (a) Taxa scores and environmental variables on RDA axes 1 and 2. (b) Taxa scores and environmental variables on RDA axes 1 and 3. The three first axes explain, respectively, 17.9, 6.2, and 5.4% of the total variation in the modern pollen data and are all significant ($P < 0.001$ after 999 permutations). LOI, loss on ignition. Abbreviations of selected taxa: Art, *Artemisia* (sage); Ast, *Asteraceae* (daisy family); Cal, *Calluna* (heather); Che, *Chenopodiaceae* (goosefoot family); Cic, *Cichoriaceae* (daisy family); Cir-typ, *Cirsium* (thistle) type; Fag, *Fagus* (beech); Gal, *Galium* (bedstraw) type; Poa, *Poaceae* (grass family); Pla, *Plantago* (plantain); Pla-lan, *Plantago lanceolata* (narrowleaf plantain); Pla-med, *Plantago major/media* (common plantain); Pin, *Pinus* (pine); Que-tot: *Quercus* (oak) total, Pot-typ: *Potentilla* (cinquefoil) type; Ran, *Ranunculaceae* (buttercup family); Rub, *Rubiaceae* (madder family); Rum, *Rumex* (sorrel); Rum-ace, *Rumex acetosa/acetosella* (garden sorrel/common sorrel); Rum-obt, *Rumex obtusifolius*; Ste-typ, *Stellaria* type; Urt-dio, *Urtica dioica* (nettle). Reproduced from Mazier F, Galop D, Gaillard MJ, et al. (2009) Multidisciplinary approach to reconstructing local pastoral activities: An example from the Pyrenean Mountains (Pays Basque). *The Holocene* 19: 171–188.

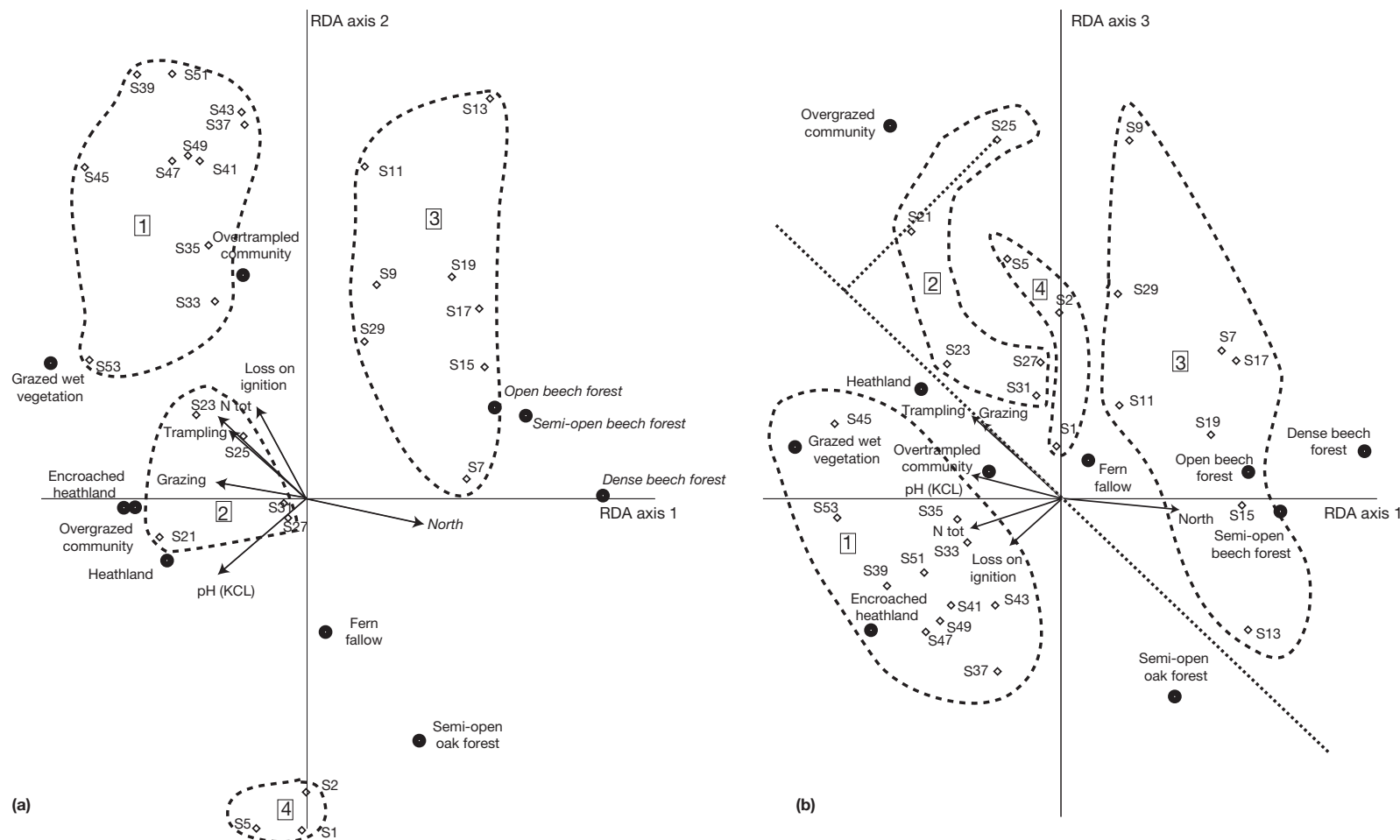


Figure 18 Application of the comparative approach in the Western Pyrenean Mountains (France), as an aid to the interpretation of Holocene pollen assemblages in terms of land use. Plot of 27 fossil pollen samples from the Holocene pollen record of Sourzay (open diamonds) positioned as 'passive' samples on the RDA axes (see Figure 17) with explanatory variables (nominal variables are given with their centroids (solid circles) and quantitative variables with solid arrows) and four passive variables (italic). Numbers (e.g., S53) correspond to the depth of the samples. (a) Fossil samples and environmental variables on RDA axes 1 and 2. (b) Fossil samples and environmental variables on RDA axes 1 and 3. Fossil samples are projected onto the arrow of the grazing pressure variable to provide a relative degree of grazing pressure. Clusters of sample scores (1–4) have been delimited by eye and are numbered in stratigraphical order. Tentative interpretation: Cluster 1: open wet pastureland on rich soils; relatively low pastoral activities. Two major vegetation types: grazed wet vegetation and encroached heathland. Cluster 2: increase in grazing pressure and colonization by trees. From sample 21, there is an indication of a mixture of more or less wooded vegetation with grazing and trampling activities and variable grazing pressure. Note the shift S27–S29–S31 between clusters 2 and 3. Cluster 3: forested phase with little pastoral activity rather rich soils and varying grazing and trampling activities. Cluster 4: open pasture, mosaic landscape with heathland communities, fern fallow, overgrown lawn, and semi-open beech forest; poor soils, and grazing and trampling activities. These different land-use phases were compared with archaeological studies in the area. Reproduced from Mazier F, Galop D, Gaillard MJ, et al. (2009) Multidisciplinary approach to reconstructing local pastoral activities: an example from the Pyrenean Mountains (Pays Basque). *The Holocene* 19: 171–188.

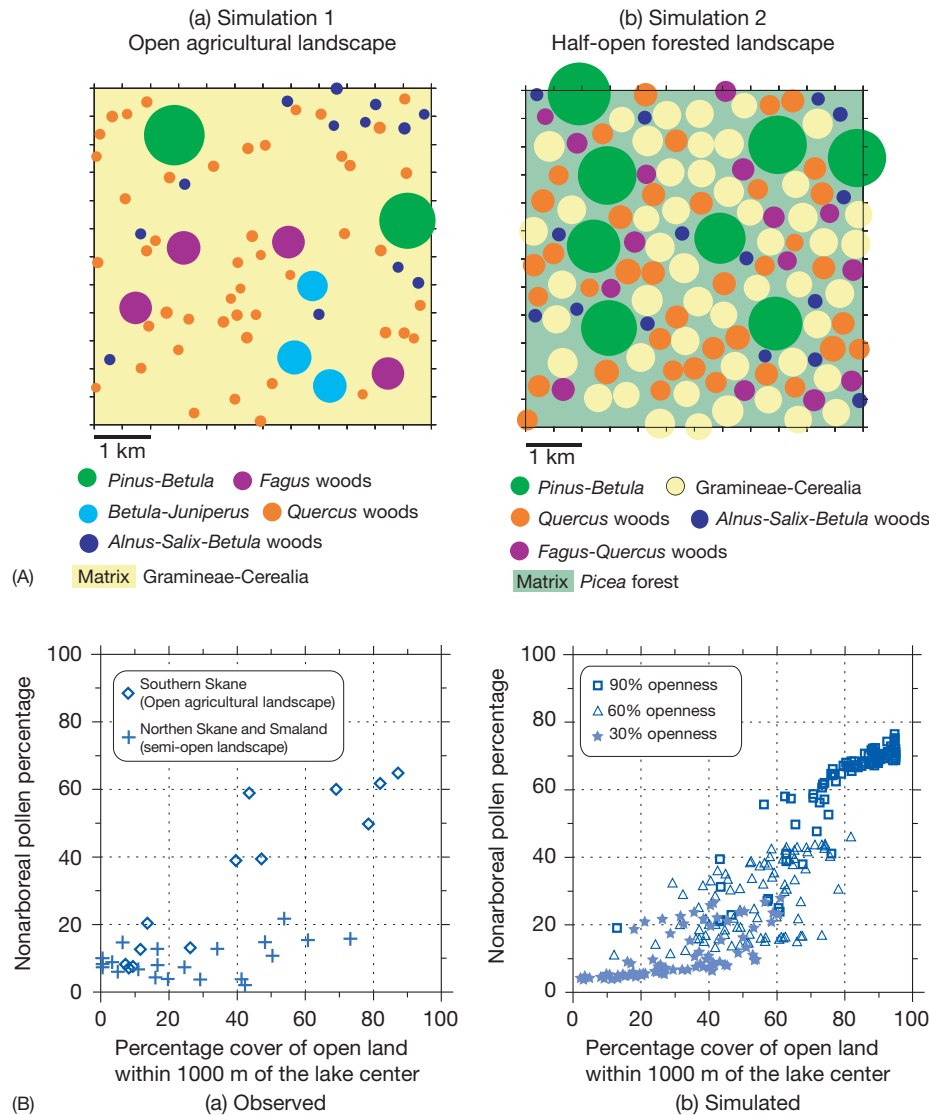


Figure 19 Comparison of the nonarboreal pollen (NAP)/open land relationship in two areas of southern Sweden (the Skåne and Småland provinces) based on empirical and simulated data (see text for explanations). (A) Hypothetical landscapes used for simulations. (a) Open landscape mimicking the open, agricultural landscape in southern Skåne, with a matrix of open land (cultivated fields, meadows, and pastures) and scattered patches of woodland types. (b) Semi-open landscape mimicking the semi-open, forested landscape of northern Skåne and Småland with a matrix of spruce and scattered patches of open land (cultivated fields, meadows, and pastures) and different woodland types. (B) Scatter plots of NAP percentage in surface sediments from small lakes versus percentage cover of open land within 1000 m of the lakes. (a) Empirical relationships between NAP percentages versus percentage cover of open land in southern Skåne (diamonds, Gaillard et al., 1998) and in northern Skåne and Småland (crosses, Broström et al., 1998). (b) Simulated relationships between NAP percentage versus percentage cover of open land within 1000 m of 3.14-ha lakes using the hypothetical landscapes of (A) and models of pollen dispersal and deposition. Three landscape designs with different degrees of openness, that is, 30% (stars), 60% (triangles), and 90% (squares), were used for simulation. The empirical data for Skåne are very similar to the simulation data for a landscape for a regional openness of 90%, while the empirical data for Småland are comparable to the simulation data for a regional openness of 30%. Reproduced from Sugita S, Gaillard MJ, and Broström A (1999) Landscape openness and pollen records: A simulation approach. *The Holocene* 9: 409–421.

vegetation/landscape units such as open versus wooded land, conifer forests, broad-leaved woods, cereal fields, and even individual tree, shrub, and herb taxa (Figure 21) represents a new and exciting potential in the use of pollen analysis for the reconstruction of past landscapes in quantitative terms, that is, past coverage of various plant groups or individual taxa and species, for example, human-induced vegetation types or forested versus open vegetation. Similarly, the LOVE model offers the potential to estimate plant cover at the spatial scale of small sites RSAP, which may help to answer specific questions that traditional interpretation of pollen data cannot solve, for

example, questions on the size of human-induced vegetation/landscape units such as cultivated and grazing land or the share of tree taxa in forested vegetation.

Applications of the LRA, REVEALS, and LOVE Models

The first REVEALS-based reconstructions of Holocene vegetation in southern Sweden indicate that human impact on land cover over the last 3000 years, as well as landscape openness during early Holocene (11500–10000 cal years BP), was much more profound than changes in pollen percentages

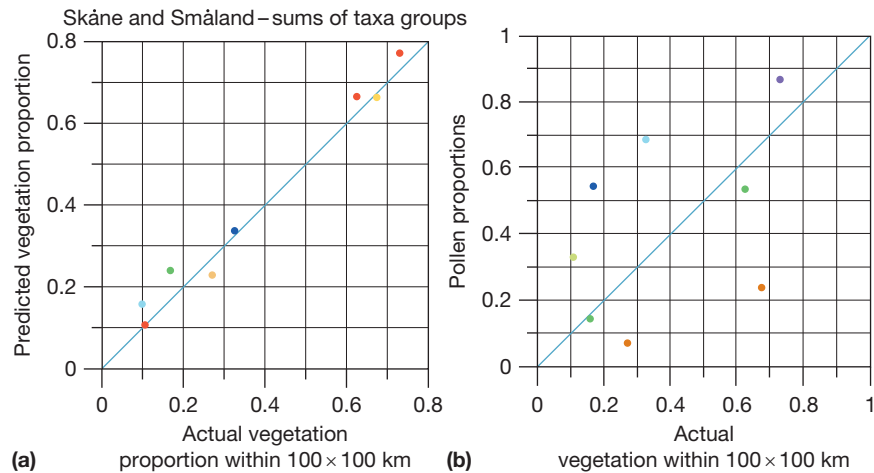


Figure 20 Validation of Sugita's REVEALS model in southern Sweden (see text for explanations). REVEALS estimates of regional vegetation abundance compared (a) with percentage coverage in the present vegetation and (b) with pollen percentages, for the following taxa groups: total trees (vegetation: Skåne – dark blue, Småland – red; pollen: Skåne – light blue, Småland – purple), total herbs (vegetation: Skåne – yellow, Småland – sand; pollen: Skåne – red brown, Småland – orange), sum conifers (vegetation: Skåne – light blue, Småland – brown; pollen: Skåne – green, Småland – dark green), and sum broad-leaved trees (vegetation: Skåne – light green, Småland – pink; pollen: Skåne – dark blue, Småland – light green). Reproduced from Hellman S, Gaillard M-J, Broström A, and Sugita S (2008) Pollen in large REVEALS regional plant abundance – Model validation in southern Sweden and potential for reconstruction of past vegetation abundance and land cover, *Journal of Quaternary Science* 22: 1–22.

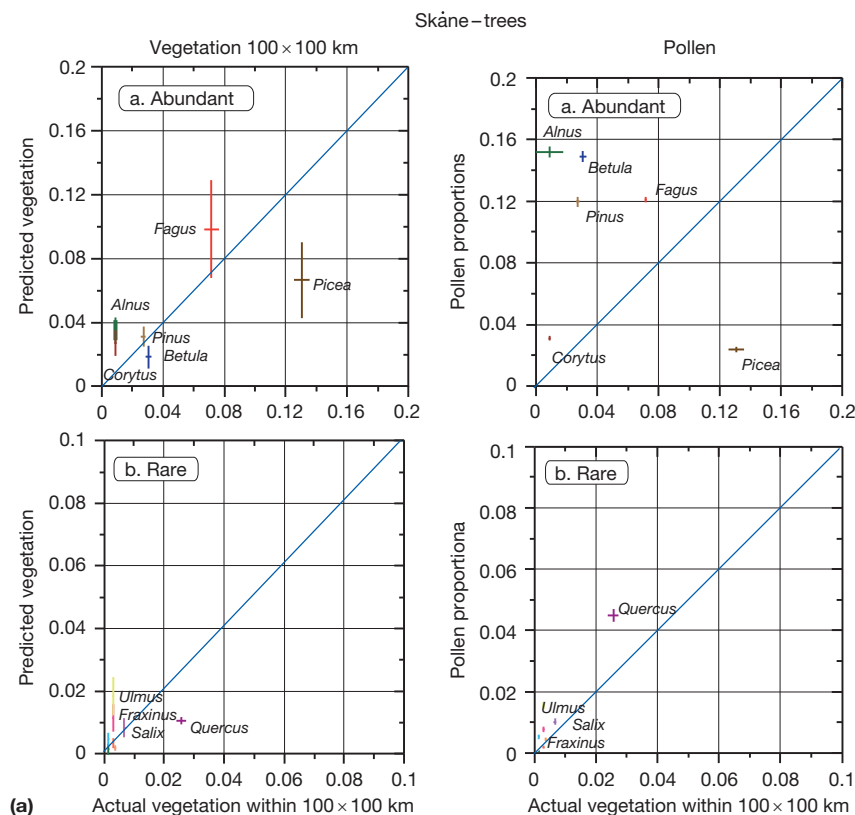


Figure 21 Vegetation proportions predicted by REVEALS (left) and pollen proportions from lake-surface pollen assemblages (right) plotted against vegetation proportions compiled from various sources of vegetation data with their respective error estimates, for (a) abundant taxa (>2%) and (b) rare taxa (<=2%) in Skåne (A, Trees; B, Herbs) and in Småland (C, Trees; D, Herbs) (see caption of Figure 20). Tree taxa: alder (light green), birch (dark blue), hazel (dark red), beech (light red), spruce (dark brown), and pine (light brown), maple (dark green), hornbeam (light blue), ash (dark pink), juniper (light orange), oak (dark purple), willow ssp. (light purple), linden (orange), and elm (light yellow). Abundant herb taxa (Ba, Da): cereals (dark red), daisy family (light red), grasses (light blue), and rye (orange). Rare herb taxa (Bb, Db): heather (dark purple), sedges (blue), *Filipendula ulmaria* (meadowsweet) (pink), narrowleaf plantain (dark green), cinquefoil (yellow), *Ranunculus acris* (meadow buttercup) (light green), madder family (dark brown), and garden sorrel type (light brown). Reproduced from Hellman S, Gaillard M-J, Broström A, and Sugita S (2008) Pollen in large lakes REVEALS regional plant abundance – Model validation in southern Sweden and potential for reconstruction of past vegetation abundance and land cover. *Journal of Quaternary Science* 22: 1–22.

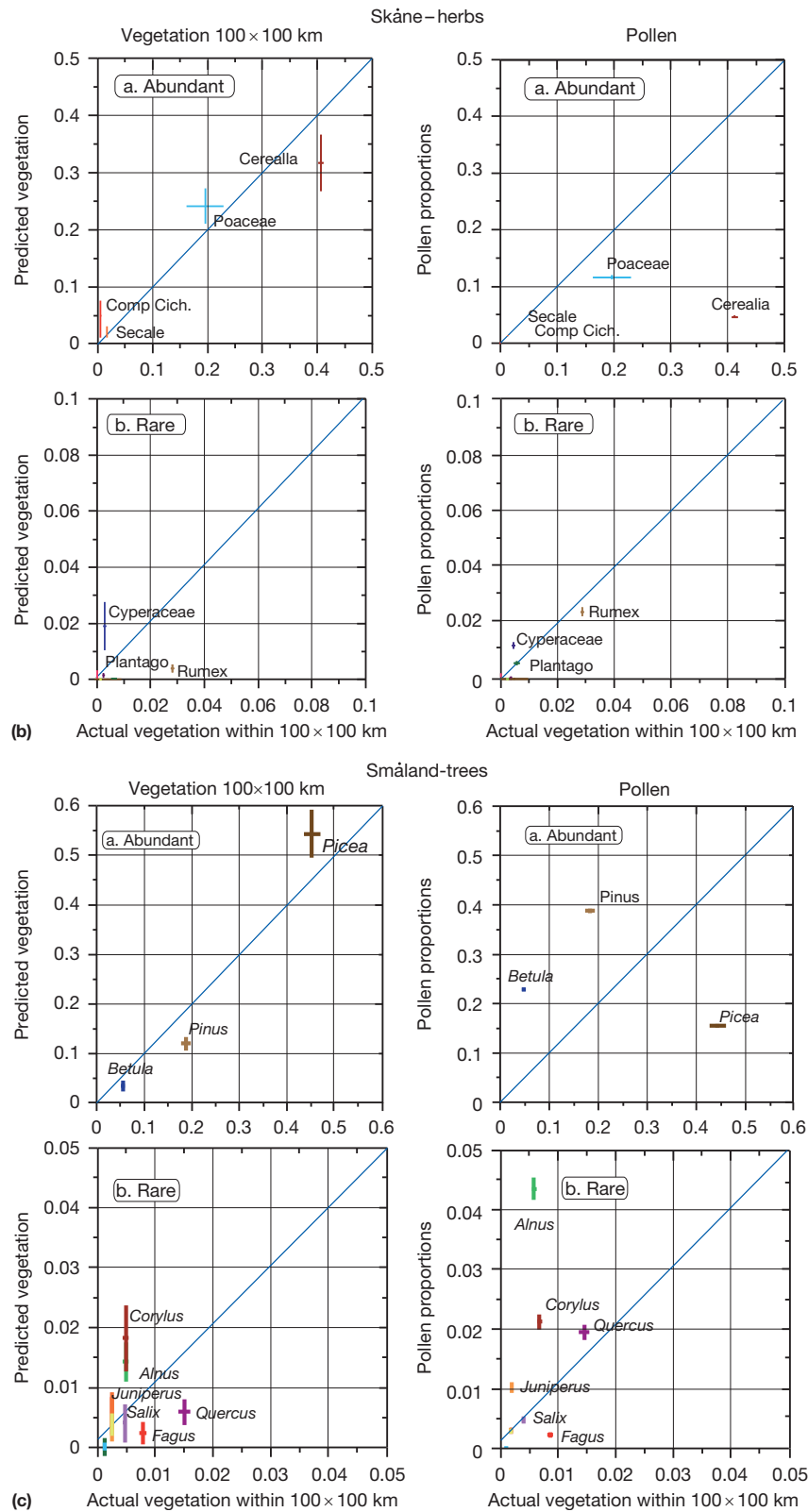


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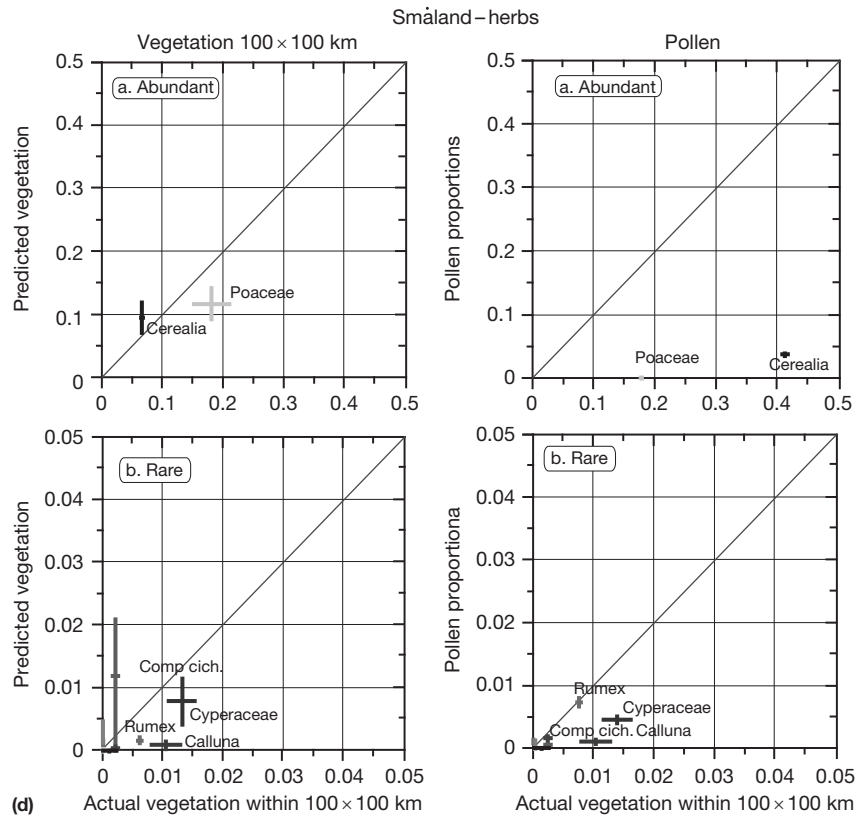


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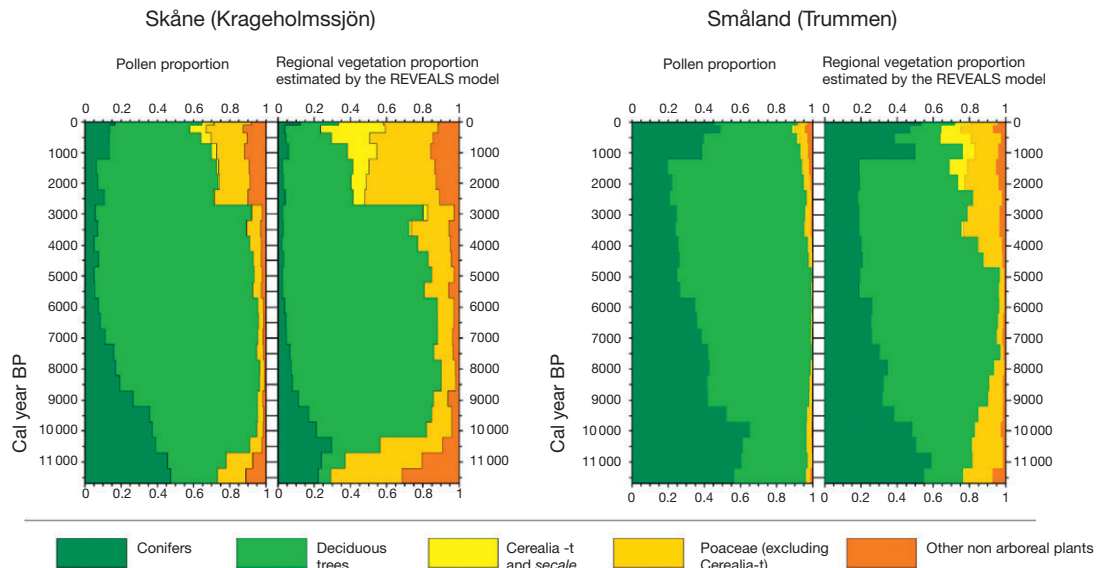


Figure 22 REVEALS estimates of plant groups through the Holocene in southern Sweden. REVEALS reconstructions of Holocene vegetation changes (right in each panel) in southern Sweden based on the pollen records (left in each panel) from Krageholmssjön (province of Skåne, left) and Lake Trummen (province of Småland, right). REVEALS was run with the PPEs from southern Sweden (Broström et al., 2004). Conifers = *Pinus* (pine), *Picea* (spruce), *Juniperus* (juniper); Deciduous trees = *Alnus* (alder), *Betula* (birch), *Corylus* (hazel), *Quercus* (oak), *Ulmus* (elm), *Tilia* (linden), *Fraxinus* (ash), *Fagus* (beech), *Carpinus* (hornbeam), *Salix* (willows); Cerealia-t = cereals; rye excluded; other nonarboreal plants = *Calluna* (heather), Compositae subfamily Cichorioideae (lettuce, dandelions, and others), Cyperaceae (sedges), *Rumex acetosa*-t (sorrels, in particular common sorrel and sheep's sorrel), *Filipendula* (meadowsweet), *Ranunculus acris* (buttercup), *Plantago lanceolata* (narrowleaf plantain); *Secale* = rye; Poaceae = grasses. Reproduced from Gaillard MJ, Sugita S, Mazier F, et al. (2010) Holocene land-cover reconstructions for studies on land cover-climate feedbacks. *Climate of the Past* 6: 483–499.

alone would suggest (Sugita et al., 2008). The cover of open land through the Holocene is shown to be strongly underestimated by percentages of NAP. For instance, at the regional spatial scale, openness estimated by REVEALS for the last 3000 years reached 50–80% in Skåne and 20–40% in Småland (compared to 20–40% and 3–10% NAP, respectively; Figure 22). A REVEALS reconstruction of the regional vegetation of the Swiss Plateau for the past 2000 years also showed that anthropogenic impact is underestimated by NAP (Soepboer et al., 2010). Therefore, spatial structure and openness of vegetation caused by both natural and anthropogenic disturbances over the last 10 000 years may have been more substantial than earlier assumed. Within an ongoing Swedish project, LANDCLIM, REVEALS is applied to estimate land cover during the Holocene in northwest Europe and western Europe north of the Alps for the purpose of climate and dynamic vegetation model evaluation (Figure 23; Gaillard et al., 2010). These REVEALS vegetation reconstructions confirm that landscape openness in the study region was already

significant from 3000 cal years BP or earlier and that the cover of treeless vegetation was larger than NAP percentages suggested. The latter has important implications for a number of paleoecological issues, such as the land cover-climate feedbacks through time. For instance, these REVEALS estimates of open land cover in Europe support the picture proposed by Kaplan et al. (2009). His scenarios of anthropogenic land cover change based on a different modeling approach also suggest a much more open landscape in Europe during the late Holocene than earlier assumed (Gaillard et al., 2010).

The first published LRA application was performed using nine late Holocene (the last 3000 years) pollen records from the major landscapes characteristic of Denmark (Nielsen and Odgaard, 2010; Figure 24). The LOVE estimates of plant cover within the RSAP of the nine sites showed that the Danish landscapes were heavily influenced by pastoral activities throughout the studied period, resulting in a dominance of grassland (40–90%) on fertile soils and a mixture of grass (40–80%) and heather (*Calluna*) (10–60%) heathlands on

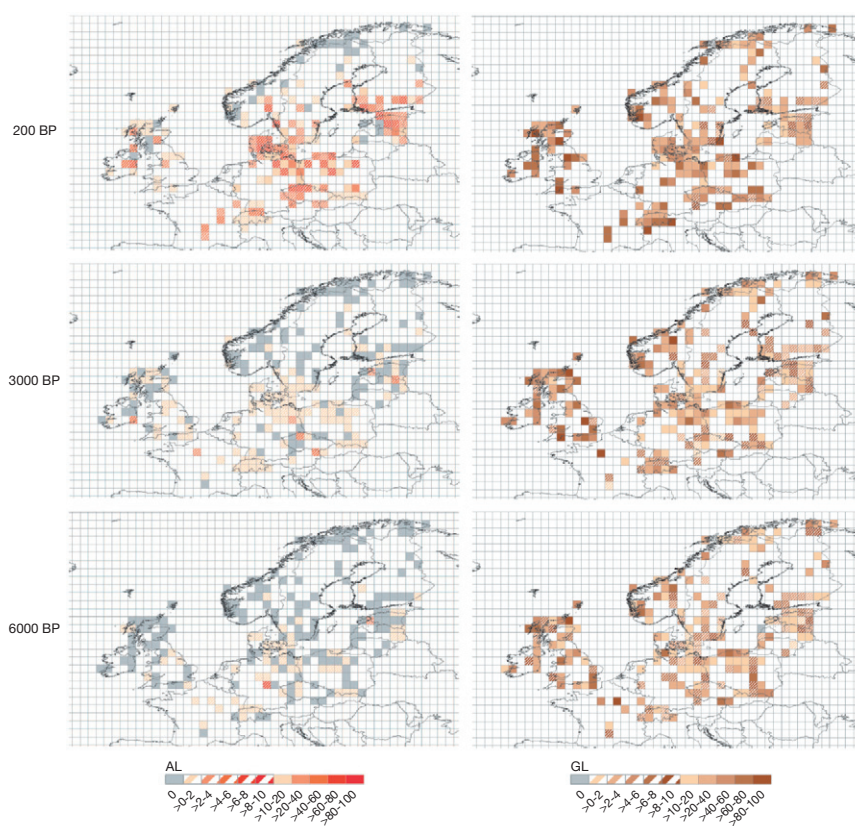


Figure 23 Example of REVEALS estimates produced by the LANDCLIM project for the time windows 6000, 3000, and 200 cal years BP (present = 1950), first generation LANDCLIM maps (produced by Anna-Kari Trondman, Florence Mazier, Anne Birgitte Nielsen, Ralph Fyfe and LANDCLIM members for the purpose of this article; see Gaillard et al. (2010) for a description of the LANDCLIM project): PFT agricultural land (AL = cereals total (*Cerealia* undiff., *Triticum* (wheat) type, *Avena* (oats) type, *Hordeum* (barley) type, *Secale cereale* (rye)) and PFT grassland (GL = *Cyperaceae*, *Filipendula*, *Plantago lanceolata*, *Plantago montana*, *Plantago media*, *Poaceae*, *Rumex* p.p. (mainly *R. acetosa* and *R. acetosella*, i.e., *Rumex acetosa* pollen-morphological type). The first generation of LANDCLIM REVEALS estimates is based on all available Holocene pollen records ($n=420$) from small and large sites (lakes and bogs) including all or part of the time windows 6000, 3000, 600, 200, and 0 cal years BP (1950), and with ≥ 3 dates for the chronological control. These pollen records were collected from the European (EPD), Alpine (ALPADABA), and Czech (PALYCZ) pollen databases (Fyfe et al., 2007; van der Knaap et al., 2000; Kunes et al., 2009, respectively). The PPEs are, for each taxon, the mean of all PPEs obtained within the project area (northwestern Europe and western Europe north of the Alps, Broström et al., 2008). These maps clearly show the strong increase in size of the land surfaces covered by agricultural land (here exclusively cultivated with cereals) and by grassland (here mainly grasses) between 6000 BP (Neolithic) and 3000 BP (Late Bronze Age), and between 3000 and 200 BP (AD 1750–1800).

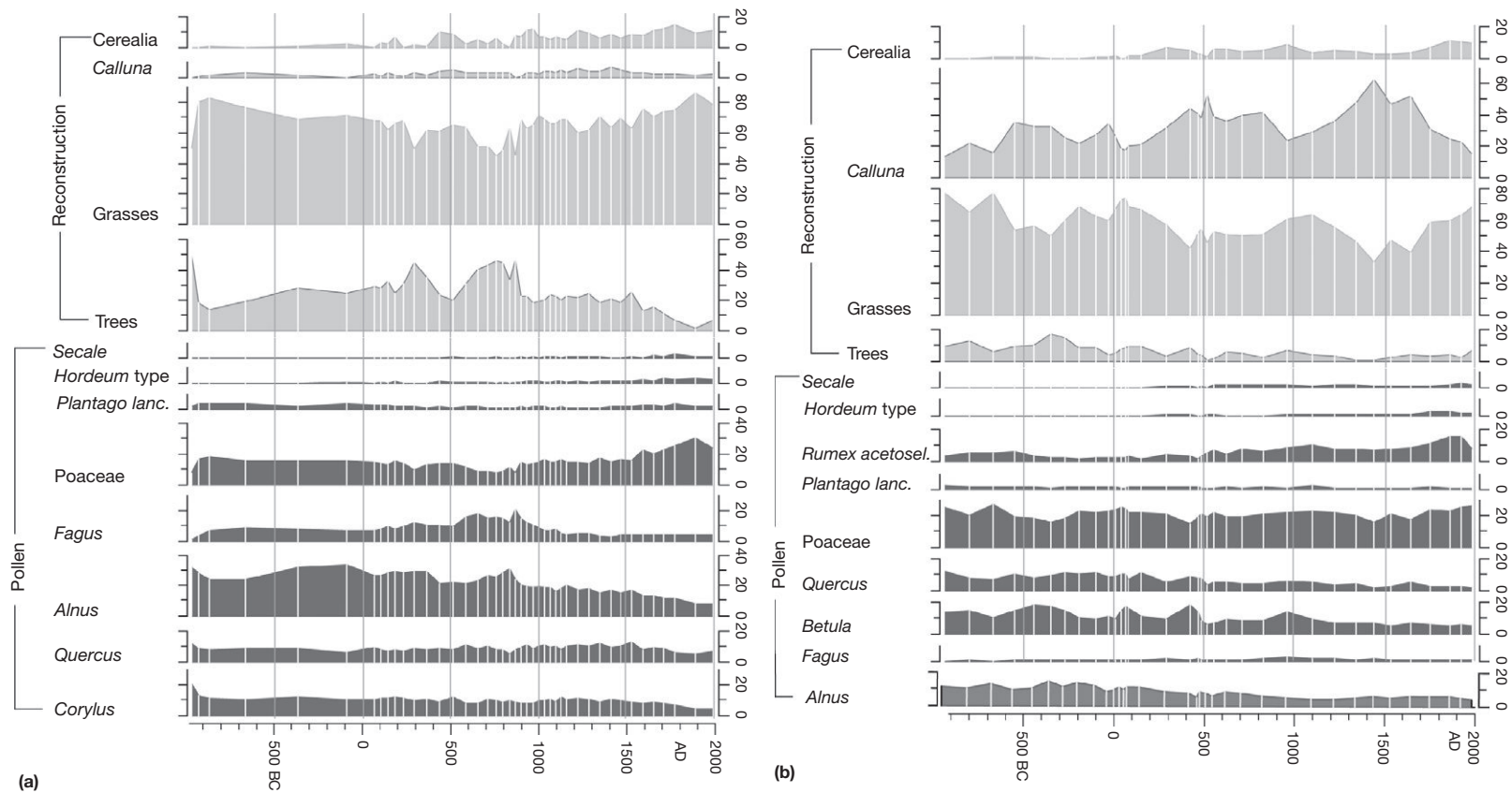


Figure 24 LRA reconstructions of the openland cover during the last 3000 years at two contrasting sites of Denmark, (a) Gudme sø (eastern Denmark) and (b) Skaansø (western Denmark). Reproduced from [Nielsen AB and Odgaard BV \(2010\)](#) Quantitative landscape dynamics in Denmark through the last three millennia based on the Landscape Reconstruction Algorithm approach. *Vegetation History and Archaeobotany* 19: 375–387 (see text for more explanations on LRA). Simplified pollen percentage diagram for major trees and herbs and LRA–LOVE reconstruction for total trees, grasses, *Calluna* (heather), and *Cerealia*. The LOVE reconstructions clearly show that pollen percentages of cereals and grasses underrepresent the cover of these plants in the vegetation. At Gudme sø, grasses cover ~50–80% of the landscape through the studied period, and in Skaansø, ~40–80%. The cover of agricultural land increases at both sites from ca. AD 100 and maintains values of ~10% from AD 1000 at Gudme sø, while they are slightly lower at Skaansø where heathlands (15–60% cover of *Calluna*) represent a large part of the landscape.

sandy soils. It is clear from this study that the degree of openness of the landscape during the last 3000 years was much higher than suggested by the uncorrected pollen percentages (10–<40% grasses and ≤30% *Calluna*). A LOVE vegetation reconstruction at two localities in southern Sweden (partly published in Marlon et al. (2010) was the only means to show that pine was more abundant at one site than at the other, which explained a significant difference between the two sites in terms of fire history, fire activity being highest at the pine-rich site during middle and late Holocene (Marlon et al., 2010; Figure 25).

Conclusions and Prospects

Pollen analysts and paleoecologists have made enormous progress in the interpretation of pollen in terms of precise vegetation abundance and landscape character. The

development of models of the pollen–vegetation relationship including pollen dispersal and deposition models (e.g., Prentice and Parsons, 1983; Sugita, 1994) led to a better understanding of the vegetation quantitative composition and spatial scale represented by pollen percentage assemblages. The computer simulation program POLLSCAPE (Sugita, 1994) and its successor HUMPOL (Bunting and Middleton, 2005) allow us to simulate pollen percentages in basins (bogs or lakes) of different sizes in hypothetical or real landscapes. It provides a means of testing the effect of changing various parameters on pollen percentages, that is, insights on the pollen representation of vegetation (see POLLSCAPE Model: Simulation Approach for Pollen Representation of Vegetation and Land Cover). The LRA (Sugita, 2007a; Sugita, 2007b), with its two models REVEALS and LOVE, is very useful complements to the traditional indicator and comparative approaches, and they are the most sensible methods so far to estimate vegetation abundance in absolute terms (for instance,

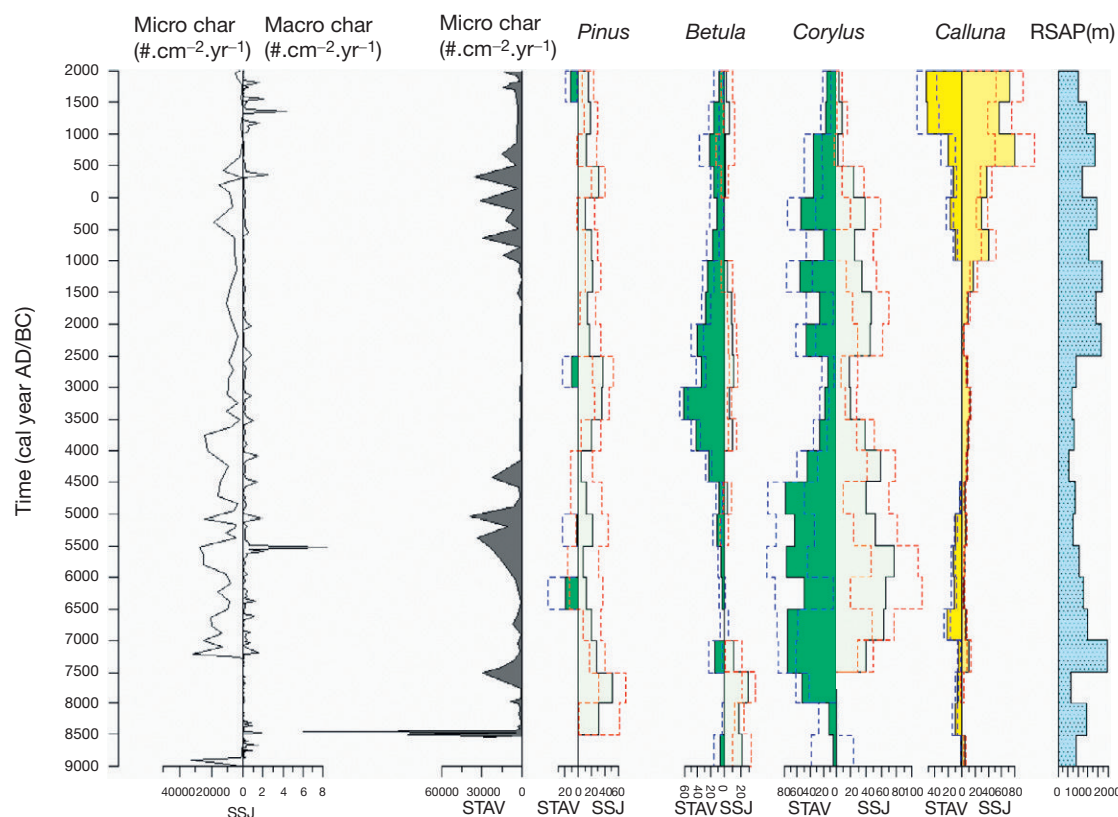


Figure 25 LRA reconstruction of 20 tree and herb taxa at two sites of southern Sweden, Stavsåkra (STAV) and Storasjö (SSJ) (see Olsson et al., 2010 for details on these sites and their fire history; modified from Marlon JF, Cui Q, Gaillard MJ, McWethy D and Walsh M (2010) Humans and fire: Consequences of anthropogenic burning during the past 2 ka. *Pages News* 18: 80–82); only four taxa are shown here, *Pinus* (pine), *Betula* (birch), *Corylus* (hazel), and *Calluna* (heather). The LOVE estimates are shown with their standard errors (blue dashed lines for Stavsåkra, red dashed lines for Storasjö). RSAP = Relevant source area of pollen, i.e., smallest radius of the area for which the reconstruction of plant abundance in % cover is valid (see Sugita, 1994). In this case it varies between 500 and 2500 m radius throughout the Holocene. The RSAP is mostly influenced by the structure of the vegetation (Hellman et al., 2009a,b). The LOVE estimates of *Pinus* and *Betula* show clear differences in species composition of the forests at the two sites, that is, up to 40% cover of *Pinus* and very low values of *Betula* except at the beginning of the Holocene at Storasjö and up to 60% cover of *Betula* and very low values of *Pinus* at Stavsåkra. The charcoal records are presented for comparison. The relatively high fire activity at Storasjö compared to Stavsåkra during the mid-Holocene might be due to the continuous presence of pine in high abundance at Storasjö during the entire Holocene, while Stavsåkra was characterized by a dominance of broad-leaved trees in the close vicinity of the site. This between-site difference in forest composition was not clearly shown by the pollen percentages or the pollen accumulation rates (e.g., Greisman and Gaillard, 2009), which illustrates the importance of using pollen–vegetation modeling for questions of this kind.

in percentage cover or in ha within a defined area). The challenges are now to (1) further evaluate the performance of REVEALS and LOVE in other landscapes than southern Sweden, Denmark, and North America; (2) gather all data necessary for the acquisition of the model parameters, PPEs, and FSP (fall speed of pollen), in other regions of the world; and (3) develop adequate methods to improve the production of landscape maps at regional and local spatial scales. The latter might be found in the use of GISs (geographical information systems) and related modeling, and in spatial modeling statistics.

See also: [Plant Macrofossil Methods and Studies: Use in Environmental Archaeology](#); [Pollen Methods and Studies: Numerical Analysis Methods](#); [POLLSCAPE Model: Simulation Approach for Pollen Representation of Vegetation and Land Cover](#); [Stand-Scale Palynology](#).

References

- Andersen STh (1979) Identification of wild grass and cereal pollen. *Danmarks Geologiske Undersøgelse. Årsbok 1972*: 69–92 (Geological Survey of Denmark. Fearbook 1978).
- Behre K-E (1981) The interpretation of anthropogenic indicators in pollen diagrams. *Pollen et Spores* 23: 225–245.
- Berglund BE (1969) Vegetation and human influence in south Scandinavia during Prehistoric time. In: Berglund BE (ed.) *Impact of Man on the Scandinavian Landscape During the Last Post-Glacial*. Oikos Suppl. 12, 9–28.
- Berglund BE (ed.) (1991) *The Cultural Landscape During 6000 Years in Southern Sweden*, vol. 41. Copenhagen: Ecological Bulletins.
- Berglund BE and Ralska-Jasiewiczowa M (1986) Pollen analysis. In: Berglund BE (ed.) *Handbook of Holocene Palaeoecology and Palaeohydrology*, pp. 455–484. Chichester: Wiley.
- Beug HJ (2004) *Leitfaden der Pollenbestimmung für Mitteleuropa und angrenzende Gebiete*. München: Pfeil-Verlag.
- Birks HJB and Gordon AD (1985) *Numerical Methods in Quaternary Pollen Analysis*. London: Academic Press.
- Bradshaw RHW and Hannon GE (2004) The Holocene structure of North-West European temperate forest induced from palaeoecological data. In: Honnay O, Verheyen K, Bossuyt B, and Hermy M (eds.) *Forest Biodiversity: Lessons from History for Conservation*, pp. 11–25. Cambridge, MA: CAB Publishing.
- Broström A, Gaillard M-J, Ihse M, and Odgaard B (1998) Pollen-landscape relationships in modern analogues of ancient cultural landscapes in southern Sweden – A first step towards quantification of vegetation openness in the past. *Vegetation History and Archaeobotany* 7: 189–201.
- Broström A, Nielsen AB, Gaillard MJ, et al. (2008) Pollen productivity estimates – The key to landscape reconstructions. *Vegetation History and Archaeobotany* 17: 461–478.
- Broström A, Sugita S, and Gaillard MJ (2004) Pollen productivity estimates for the reconstruction of past vegetation cover in the cultural landscape of southern Sweden. *The Holocene* 14: 368–381.
- Bunting J and Middleton R (2005) Modelling pollen dispersal and deposition using HUMPOL software, including simulating windrows and irregular lakes. *Review of Palaeobotany and Palynology* 134: 185–196.
- Court-Picon M, Buttler MA, and de Beaulieu JL (2006) Modern pollen/vegetation/land-use relationships in mountain environments: An example from the Champsaur valley (French Alps). *Vegetation History and Archaeobotany* 15: 151–168.
- Firbas F (1937) Der pollenanalytische Nachweis des Getreidebaus. *Zeitschrift für Botany* 31: 447–478.
- Frenzel B, Andersen ST, Berglund BE, and Gläser B (eds.) (1994) *Evaluation of Land Surfaces Cleared from Forests in the Roman Iron Age and the Time of Migrating Germanic Tribes Based on Regional Pollen Diagrams*. *Palaeoklimaforschung/Palaeoclimate Research*, vol. 12. Stuttgart: Gustav Fischer Verlag.
- Fyfe RM, de Beaulieu JL, Binney H, et al. (2007) The European Pollen Database: Past efforts and current activities. *Vegetation History and Archaeobotany* 18: 417–424.
- Gaillard MJ, Birks HJB, Emanuelsson U, Karlsson S, Lagerås P, and Olausson D (1994) Application of modern pollen/land-use relationships to the interpretation of pollen diagrams – Reconstructions of land-use history in south Sweden, 3000–0 BP. *Review of Palaeobotany and Palynology* 82: 47–73.
- Gaillard MJ, Birks HJB, Ihse M, and Runborg S (1998) Pollen/landscape calibrations based on modern pollen assemblages from surface-sediment samples and landscape mapping – A pilot study in south Sweden. In: Gaillard M-J and Berglund BE (eds.) *Quantification of Land Surfaces Cleared of Forests During the Holocene – Modern Pollen/Vegetation/Landscape Relationships as an Aid to the Interpretation of Fossil Pollen Data*. *Palaeoklimaforschung/Palaeoclimate Research*, vol. 27, pp. 31–52. Stuttgart: Gustav Fischer Verlag.
- Gaillard M-J, Dearing JA, El-Daoushy F, Enell M, and Håkansson H (1991) A late Holocene record of land-use history, soil erosion, lake trophy and lake-level fluctuations at Bjäresjösjön (South Sweden). *Paleolimnology* 6: 51–81.
- Gaillard MJ, Dutoit T, Hjelle K, Koff T, and O'Connell M (2009) European cultural landscapes: Insights into origins and development. In: Krzywinski K, O'Connell M, and Küster H (eds.) *Cultural Landscapes of Europe*, pp. 35–46. Bremen: Aschenbeck Media.
- Gaillard M-J, Sugita S, Broström A, Eklöf M, and Pilesjö P (2000) Long term land-cover changes on term regional to global scales inferred from fossil pollen – How to meet the challenges of climate research? *Pages Newsletter* 8: 30–32.
- Gaillard MJ, Sugita S, Bunting MJ, et al. (2008) The use of modelling and simulation approach in reconstructing past landscapes from fossil pollen data: A review and results from the POLLANDCAL network. *Vegetation History and Archaeobotany* 17: 419–443.
- Gaillard MJ, Sugita S, Mazier F, et al. (2010) Holocene land-cover reconstructions for studies on land cover-climate feedbacks. *Climate of the Past* 6: 483–499.
- Greisman A and Gaillard MJ (2009) The role of climate variability and fire in early and mid-Holocene forest dynamics of Southern Sweden. *Journal of Quaternary Science* 24: 593–611.
- Hellman S, Gaillard M-J, Broström A, and Sugita S (2008) Quantitative reconstruction of regional plant cover using pollen records from large lakes – Validation of the reveals model in southern Sweden. *Journal of Quaternary Science* 22: 1–22.
- Hicks S (1992) Modern pollen deposition and its use in interpreting the occupation history of the island Hailuoto, Finland. *Vegetation History and Archaeobotany* 1: 75–86.
- Hicks S and Birks HJB (1996) Numerical analysis of modern and fossil pollen spectra as a tool for elucidating the nature of fine-scale human activity in boreal areas. *Vegetation History and Archaeobotany* 5: 257–272.
- Huntley B and Webb T III (1988) *Vegetation History*. Dordrecht: Kluwer.
- Iversen J (1941) Landnam in Danmarks Stenalder (land occupation in Denmark's stone age). *Danmarks Geologiske Undersøgelse II* 66: 7–68.
- Iversen J (1973) The development of Denmark's nature since the last glacial. *Danmarks Geologiske Undersøgelse II* V/7-C: 1–26.
- Kaplan J, Krumhardt K, and Zimmermann N (2009) The prehistoric and preindustrial deforestation of Europe. *Quaternary Science Reviews* 28: 3016–3034.
- Kunes P, Abraham V, Kovaik O, and Kopecky M (2009) PALYCZ contributors: Czech Quaternary Palynological Database (PALYCZ): Review and basic statistics of the data. *Preslia* 8: 209–238.
- Lagerås P (1996) Vegetation and land-use in the Småland Uplands, Southern Sweden, during the last 6000 years. *Lundqua Thesis* 36, Lund University. ISSN 0293-3146.
- Lagerås P (2000) Järnålderns odlingsystem och landskapets långsiktiga förändring – Hamnedas röjningsröseområden i ett paleoekologiskt perspektiv (Iron Age farming and long-term landscape change in Hamneda), Southern Sweden – Clearance cairn areas in a palaeoecological perspective. In: Lagerås P (ed.) *Arkeologi och Paleoekologi i Sydvästra Småland*, pp. 167–230. Sweden: Riksantikvarieämbetet.
- Marlon JF, Cui Q, Gaillard MJ, McWethy D, and Walsh M (2010) Humans and fire: Consequences of anthropogenic burning during the past 2 ka. *PAGES News* 18: 80–82.
- Mazier F, Galop D, Brun C, and Buttler A (2006) Modern pollen assemblages from grazed vegetation in the western Pyrenean Mountains (France) – A numerical tool for more precise reconstruction of past cultural landscapes. *The Holocene* 16: 91–103.
- Mazier F, Galop D, Gaillard MJ, et al. (2009) Multidisciplinary approach to reconstructing local pastoral activities – An example from the Pyrenean mountains (Pays Basque). *The Holocene* 19: 171–188.
- McAndrews JH (1988) Human disturbance of North America forests and grasslands: The fossil pollen record. In: Huntley B and Webb T III (eds.) *Vegetation History*, pp. 673–697. Dordrecht: Kluwer.
- Moore PD, Webb JA, and Collinson ME (1991) *Pollen Analysis*, 2nd edn. Oxford: Blackwell.

- Nielsen AB and Odgaard BV (2010) Quantitative landscape dynamics in Denmark through the last three millennia based on the Landscape Reconstruction Algorithm approach. *Vegetation History and Archaeobotany* 19: 375–387.
- Olsson F, Gaillard MJ, Lemdahl G, et al. (2010) A continuous record of fire covering the last 10500 calendar years from southern Sweden: The role of climate and human activities. *Palaeogeography, Palaeoclimatology, Palaeoecology* 291: 128–141.
- Prentice IC and Parsons RW (1983) Maximum likelihood linear calibration of pollen spectra in terms of forest composition. *Biometrics* 39: 1051–1057.
- Punt W, Blackmore S, Clarke GCS, Hoen PP, and Stafford PJ (eds.) (1976–2009) *The Northwest European Pollen Flora I–VIII*. Amsterdam/Boston/London: Elsevier.
- Ruddiman WF (2003) The anthropogenic greenhouse era began thousands of years ago. *Climatic Change* 61: 261–293.
- Ruddiman WF (2005) The early anthropogenic hypothesis a year later. *Climatic Change* 69: 427–434.
- Soepboer W, Sugita S, and Lotter A (2010) Regional vegetation-cover changes on the Swiss Plateau during the past two millennia: A pollen-based reconstruction using the REVEALS model. *Quaternary Science Reviews* 29: 472–483.
- Sugita S (1994) Pollen representation of vegetation in Quaternary sediments: Theory and method in patchy vegetation. *Journal of Ecology* 82: 881–897.
- Sugita S (2007a) Theory of quantitative reconstruction of vegetation. I. Pollen from large lakes reveals regional vegetation. *The Holocene* 17: 229–241.
- Sugita S (2007b) Theory of quantitative reconstruction of vegetation. II. All you need is LOVE. *The Holocene* 17: 243–257.
- Sugita S, Gaillard MJ, and Broström A (1999) Landscape openness and pollen records: A simulation approach. *The Holocene* 9: 409–421.
- Sugita S, Gaillard MJ, Hellman S, and Broström A (2008) Model-based reconstruction of vegetation and landscape using fossil pollen. In: Posluschny A, Lambers K, and Herzog I (eds.) *Layers of Perception*. Proceedings of the 35th Conference on Computer Applications and Quantitative Methods in Archaeology (CAA), Berlin, Germany, 2–6 April 2007, pp. 385–391. Bonn: Dr. Rudolf Habelt GmbH.
- Sugita S, Parshall T, Calcote R, and Walker K (2010) Testing the Landscape Reconstruction Algorithm for spatially explicit reconstruction of vegetation in northern Michigan and Wisconsin. *Quaternary Research* 74: 289–300.
- Van der Knaap WO, van Leeuwen JFN, Fonkhouse A, and Ammann B (2000) Palynostratigraphy of the last centuries in Switzerland based on 23 lake and mire deposits: Chronostratigraphic pollen markers, regional patterns, and local histories. *Review of Palaeobotany and Palynology* 108: 85–142.
- Vorren KD (1986) The impact of early agriculture on the vegetation of northern Norway: A discussion of anthropogenic indicators in biostratigraphical data. In: Behre K-E (ed.) *Anthropogenic Indicators in Pollen Diagrams*, pp. 1–18. Rotterdam/Boston, MA: A.A. Balkema.
- Vuorela I (1986) Palynological and historical evidence of slash-and-burn cultivation in south Finland. In: Behre K-E (ed.) *Anthropogenic Indicators in Pollen Diagrams*, pp. 53–64. Rotterdam/Boston, MA: A.A. Balkema.

Further Reading

- Hellman S, Bunting JM, Gaillard MJ (2009a) Relevant Source Area of Pollen in patchy cultural landscapes and signals of anthropogenic landscape disturbance in the pollen record: A simulation approach. *Review of Palaeobotany and Palynology* 153: 245–258.
- Hellman S, Gaillard MJ, Bunting JM, Florence Mazier F (2009b) Estimating the Relevant Source Area of Pollen in the past cultural landscapes of southern Sweden — A forward modelling approach. *Review of Palaeobotany and Palynology* 153: 259–271.





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DEDICATIONS

The second edition of the Encyclopedia of Quaternary Science is dedicated to the memory of three outstanding scientists who have recently passed away: Russell Coope, Bill Watts, and Aleksis Dreimanis.

Russell Coope

G. Russell Coope (1930–2011) was essentially the founding father of Quaternary Entomology, having initiated the modern era of Quaternary insect fossil research in the late 1950s. Prior to that time, and going back to the nineteenth century, there had been a number of false starts and stops in this field. Previous researchers had assumed that any ice-age insect fossils they found would necessarily represent extinct taxa. This, as it turns out, was a completely false assumption. Beginning with the Late Pleistocene Upton Warren and Chelford sites, Russell was able to determine that fossil insects, especially beetles (Coleoptera) from Pleistocene deposits, represent extant species, albeit species that may live in dramatically different regions today than those where they lived in the past. For many years at the start of this endeavor, Russell was ‘a voice in the wilderness,’ having to convince colleagues in both the Quaternary sciences and modern Entomology that it is possible to identify ‘beetle bits’ (individual sclerites of beetle exoskeleton) to the species level, and that fossil beetle assemblages can yield reliable reconstructions of past environments. One of his greatest battles was to convince skeptical palynologists that the interstadial interval in what is now known as MIS 3, and another one toward the end of the last glaciation, were almost as warm as today, in spite of the fact that the vegetation cover represented in pollen records failed to register this. Likewise, Russell championed the idea that climate oscillations, particularly in the late glacial interval and early Holocene, were extremely rapid and dramatic: switching from full glacial to full interglacial conditions in a matter of decades, at most. Russell’s paleoclimate reconstructions were finally given full credence by the scientific community at large when the oxygen isotope records from Greenland ice cores began to be published, showing nearly the exact same amplitude and timing of climate change he had been publishing since the 1960s.

Russell’s research has had important implications for the field of zoogeography. On the basis of his work, we now know that insects respond to climate change chiefly by shifting their distributional ranges, so that they



Figure 1 Russell Coope at the Chelford quarry fossil site, August 1977. Photograph by Alan V. Morgan, used with permission.

track these changes through space. Through his tenacity and hard work, Russell was able to track down the modern ranges of beetle species he found in Pleistocene deposits to such remote locations as northeastern Siberia and the Tibetan Plateau (cold-adapted species), as well as North Africa and the Mediterranean islands (warm-adapted species).

Russell brought a level of curiosity and enthusiasm for natural history and all things scientific that was infectious. He was a brilliant lecturer and storyteller, holding his audiences practically spellbound. He trained many of the paleoentomologists who took the field forward in recent decades. His seemingly boundless energy kept him working and publishing papers, right up to the end of his life. He published more than 200 papers in his long career. He collaborated with dozens of other Quaternary scientists and was one of the first to appreciate and promote multidisciplinary studies of Pleistocene fossil sites. There is no doubt that his research and publishing legacies will continue to inspire future generations of paleoentomologists.

Bill Watts

William (Bill) A. Watts (1930–2010) was a towering figure in Quaternary paleoecology. He was a skilled botanist and ecologist and this was a secret to his success in unraveling the interglacial, late-glacial, and Holocene vegetation history not only in his native Ireland but also in the midwestern, the southeastern, and eastern USA, and in Italy. A major contribution was his reintroduction and development of quantitative plant macrofossil analysis. Macrofossils can often be identified to lower taxonomic levels than pollen, and Bill took full advantage of the complementarity of both methods. Thus, he was able to determine important features of vegetation history by identifying trees to species level and to illuminate the history of aquatic plants and local vegetational developments.



Figure 2 Photograph of Bill Watts, taken at Laguna de la Roya, Spain. Photograph by Fraser Mitchell, used with permission.

Bill always enjoyed fieldwork and microscope work, carrying out the latter in spare time left over from his job as Provost of Trinity College Dublin and his participation on committees dealing with hospitals and conservation in Ireland. Bill's quiet but intense enthusiasm and his great sense of humor added much to discussions, macrofossil identification sessions, and excursions with him. He and his work are inspirations to those who follow him.



Figure 3 Aleksis Dreimanis, photograph by Stephen Hicock.

Aleksis Dreimanis

Aleksis Dreimanis (1914–2011) was widely considered the world's foremost Quaternary glacial geologist in the 1970s and 80s. Aleksis was born in Latvia. While still an undergraduate at the University of Latvia, he published his first paper on the structural geology of deformed glacial deposits along the banks of the River Daugava – a study that he would return to and revise 60 years later. In 1944, Aleksis was conscripted as a military geologist by the Germans who occupied Latvia. For the next two years, he was separated from his family and forced to do strategic geological mapping in eastern Germany and northern Italy until the end of WWII, followed by internment in prisoner of war and displaced persons camps. In 1948, he escaped to London, England, where he met with the President of the University of Western Ontario who gave him a job in London, Canada (his family was able to join him the following year).

Over the next six decades at the University of Western Ontario, Aleksis ascended to the top of his discipline while gaining international recognition through landmark publications and inspired leadership on committees. His 1970s synthesis and definition of the Wisconsin Stage with Paul Karrow, Jan Terasmae, Dick Goldthwait, and Jane Forsyth became a world standard to which other records were compared. In 1989, he expanded this synthesis and introduced the conceptual till tetrahedron that showed the till continuum with end members occupying the corners of an inverted pyramid. He was among the first geologists to use radiocarbon dating, and he provided samples for inter-lab calibration. He was among the first to quantitatively study the relationship between till lithology and till formation, and to investigate how glacial transport and dynamics determine till texture, deformation, and rheologic states. Aleksis produced over 200 publications spanning eight decades and over a wide range of topics. Largely because of his influence, the study of till genesis and its relation to till properties and past glacial behavior has become a major focus of glacial geologic research in recent decades. The contribution and legacy that forms the body of work of Dreimanis will continue as a major benchmark within the diverse fields of scientific enquiry that are sometimes loosely referred to as 'glacial geology.'

By

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FOREWORD

As with the publication of the first edition in 2007, the publication of the second edition of the *Encyclopedia of Quaternary Science* represents a landmark in the history of publishing in the field of Quaternary Science. Quaternary Science is a multidisciplinary endeavor which seeks to establish as detailed a picture as possible of the manifold environmental changes that have occurred during the most recent geological period, the Quaternary – an interval of time that spans the past 2.59 million years or the past 0.056% of geological time. It is a period of significant climate and environmental change and witnessed the widespread dispersal of our species, *Homo sapiens*, across the planet.

Since Louis Agassiz and Reverend William Buckland traipsed over parts of the Scottish landscape in 1840 in search of evidence of glaciation, a huge literature has emerged on the science of long-term climate and environmental change. Rapid technological advances in the late twentieth century and the proliferation of scientific journals, particularly in an era of electronic publishing, have resulted in an exponential growth and documentation of knowledge on the climate and environmental changes that have occurred during the Quaternary period. More recently, there has been an increasing public interest in applied Quaternary research as a framework for understanding the basis for recent climate changes and for understanding the nature and frequency of geological hazards and vexing issues such as soil erosion and land degradation, and the adverse effects of ocean water temperature increases and acidification on coral reef environments. In a similar manner, the likely magnitude of future sea-level rise and the associated impacts on coastal landscapes in the twenty-first century have attracted wide public interest. In this sense, Quaternary Science is very much on the political agenda and is a critically important subject to address issues of public concern.

The *Encyclopedia of Quaternary Science* edited by Scott Elias of Royal Holloway, University of London, UK, accordingly represents a particularly welcome addition to the literature. The encyclopedia presents an up-to-date and authoritative overview of Quaternary Science. The encyclopedia should enjoy a wide readership as the entries are presented in a very clear and easily readable style. The text of the articles is written at a level that allows undergraduate students to understand the material, while providing active researchers with a ready reference resource for information in the field. Each entry of up to 4000 words covers the salient points of each topic with very clear illustrations. A central theme that pervades the work is the importance of Quaternary Science in providing an historical context for assessing present environmental changes and as basis for modeling potential future changes.

The encyclopedia consists of four volumes in print form and is available electronically. All the entries have been updated and the text totals around 3,500,000 words. As a major reference work, the encyclopedia has a very wide coverage of topics within the Quaternary sciences reflecting the complex and interdisciplinary nature of the science. Each of the major sections begins with a general overview of the topic prepared by a leading expert in the field. The major sections, for example, examine the analytical methods commonly used in paleoenvironmental reconstructions to unravel in a forensic-like manner the nature of former environments and the tempo of environmental change. Accordingly, a great range of topics such as the former extent of ice cover and nature of Quaternary glaciations, the biological responses and resultant fossil records to fluctuating climate, the expansion and contraction of desert environments, and global and local changes in relative sea levels are examined. Other topics covered include dating techniques, Quaternary stratigraphy, fluvial environments, lake level studies, paleosols, paleobotany, ancient DNA, paleolimnology, vertebrate studies, insect fossil studies, paleoceanography, stable isotope studies, ice core records, and human evolution in the Quaternary. One of the new sections in the encyclopedia examines the application of Quaternary proxy evidence in forensic science. All sections provide a clear summary of the latest advances in the fields of research.

This is an outstanding work and the editors and the publisher are to be congratulated for producing an encyclopedia that cogently summarizes the current state of the science.

A handwritten signature in black ink, appearing to read 'C. Murray-Wallace', with a stylized, flowing script.

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INTRODUCTION

It has been my privilege to work with more than 420 leading Quaternary scientists in developing the second edition of the *Encyclopedia of Quaternary Science*. This team of writers and editors represents 28 countries in Europe, Asia, Africa, the Americas, Australia, and New Zealand. Starting with the first edition in 2006, I have had my finger on the pulse of Quaternary science, and this branch of science is truly pulsating! Information now comes from an incredible variety of disciplines: geochemistry, numerical modeling, history, vulcanology, paleobiology, nuclear physics, stratigraphy, sedimentology, climatology, anthropology, archaeology, glacial geology, soil science, ice-stream modeling – the list is staggering. This highly disparate group of people are bound together by one common thread: the desire to know the history of the planet during the last 2.6 million years – the time of the ice ages. For Quaternary scientists, this is a pressing need, not an idle curiosity. Any doubts about this statement can be easily dispelled by a consideration of the lengths to which many of them go to gather the necessary data. Some of them have worked for months in sub-zero temperatures on top of very high mountains, or near the center of polar ice sheets, collecting ice cores. Others have spent many weeks on some of the roughest seas in the world, drilling deep-sea sediment cores. Often the work is more mundane. An oxygen isotope curve for a lengthy marine sediment core represents thousands of hours of patiently picking tiny fossils from layer after layer of sediments, in order to obtain sufficient numbers of calcium carbonate shells to yield samples for isotopic analysis. A map showing proposed ice limits from the last glaciation represents thousands of hours of field mapping of glacial features by dozens of people. Why do all of these people devote their lives to this pursuit of knowledge? Does it really matter so much? The answer becomes clear when you step back and examine the topic of Quaternary science in its proper context.

The world we inhabit has largely been shaped by the events of the Quaternary. All the biological communities that exist today are the end product of a long series of species associations that came together in the past, largely driven by climatic change during the Pleistocene. We cannot properly understand the functioning of modern ecosystems without a solid knowledge of their history, any more than we can understand the plot of a long novel by reading just the final page. We are also living in a time of alarming climate changes. Even though the pace and intensity of some of these changes have not been seen in historical times, there were many rapid, large-scale climatic shifts in the Pleistocene. The best way to predict the effects of global warming on the planet's climate and ecosystems is to look at the history of similarly intense, rapid changes in the prehistoric past. The interval that is most relevant today is the most recent geologic period: the Quaternary. As human populations rise exponentially, increasing numbers of people are exposed to geologic hazards, such as earthquakes (and attendant tsunamis), and volcanic eruptions. These are short-lived events that take place only rarely in any one region. The interval between major events, such as volcanic eruptions, may be centuries or millennia. How do we come to grips with predicting the future likelihood of such erratic phenomena? Again, the answers come from piecing together the ancient history of such events, over many thousands of years.

The Quaternary has been the time when our own species came of age. The beginning of the Quaternary, roughly 2.6 million years ago, was about the time when the earliest member of our genus (*Homo*) first appeared in Africa. Pleistocene environments shaped the course of human evolution, culminating in anatomically modern *Homo sapiens* spreading from Africa throughout most of the world during the last glaciation. Even though human beings largely shape their own environments today, for the vast majority of our species' history, it has been the environment that has shaped us. Our direct ancestors' adaptations to environmental change are deeply ingrained in our genes. Thus, an understanding of the environmental conditions that shaped our species is critical to our understanding of humanity.

Quaternary science is a rapidly changing field, and the articles that appear in this encyclopedia reflect this. New dating techniques, such as cosmogenic nuclide dating, are revolutionizing our understanding of many earth surface processes. The ability to analyze increasingly smaller samples for radiocarbon and stable isotopes of oxygen and hydrogen means that we are gaining a level of precision in the reconstruction of past events that was unheard of just a few years ago. Stable isotope studies of air bubbles trapped in ice cores from Greenland and Antarctica have given Quaternary scientists an entirely new perspective on the rapidity and intensity of climatic change during the last glacial cycle and beyond. Likewise, the discovery of long sequences of annually laminated sediments in both marine and freshwater environments has provided a great leap forward in our ability to resolve the timing of environmental changes in nonpolar regions. The ability to extract and analyze ancient DNA sequences from Pleistocene fossils (both plants and animals) is revolutionizing the field of paleobiology. We are beginning to be able to trace the genetic lineages of a number of different organisms, from beetles to bison. In short, these are very exciting times to be a Quaternary scientist! While it is virtually impossible for any Quaternary researcher or student to keep abreast of all the new discoveries in this multifaceted science, this encyclopedia can be of great help. The articles contained here represent the state of the art in a huge variety of topics, and they offer the opportunity to dig deeper into their respective subjects by providing full citations of the most pertinent literature available. I invite you to come and explore the Quaternary Period in the pages that follow. It is a fascinating story.

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HOW TO USE THE ENCYCLOPEDIA

The material in this encyclopedia is organized into two broad sections. At the start of Volume 1 are four *Introductory Articles* that serve as a broad introduction to the field. It is intended that these articles should serve as the first port of call for readers, to enable them to embark on any of the main entries without the need to consult other material beforehand.

Following the Introductory Articles, the main body of The Encyclopedia is arranged as a series of entries in alphabetical order. These entries fill the remainder of the volumes.

To help realize the full potential of the material in the Encyclopedia we have provided four features to help you find the topic of your choice: a contents list by subject; an alphabetical contents list; cross-references to other articles; and a full subject index.

1. Contents list by subject

Your first point of reference will probably be the contents list by subject. This list appears at the front of each volume, and groups the entries under subject headings describing the broad themes of Quaternary science. This will enable the reader to make quick connections between entries and to locate the entry of interest. Under each main section heading, you will find several subject areas and under each subject area is a list of those entries that covers aspects of that subject, together with the volume and page numbers on which these entries may be found.

2. Alphabetical contents list

The alphabetical contents list, which also appears at the front of each volume, lists the entries in the order in which they appear in the Encyclopedia. This list provides both the volume number and the page number of each entry.

You will find ‘dummy’ entries where obvious synonyms exist for entries or where we have grouped together related topics. Dummy entries appear in both the contents list and the body of the text.

Example

If you were attempting to locate material on *peat studies* via the alphabetical contents list:

PEAT STUDIES *see* PLANT MACROFOSSIL METHODS AND STUDIES: Mire and Peat Macros

The dummy entry directs you to the entry in which *peat* is covered.

If you were trying to locate the material by browsing through the text and you had looked up *Peat Studies*, then the following information would be provided in the dummy entry:

Peat Studies *see* PLANT MACROFOSSIL METHODS AND STUDIES: Mire and Peat Macros

3. Cross-references

All of the entries in the Encyclopedia have been extensively cross-referenced. The cross-references, which appear at the end of the entry, serve three different functions:

- i. To indicate if a topic is discussed in greater detail elsewhere.
- ii. To draw the reader’s attention to parallel discussions in other entries.
- iii. To indicate material that broadens the discussion.

Example

The following list of cross-references appear at the end of the entry GLACIAL LANDFORMS, ICE SHEETS | Growth and Decay

See also: **Introduction:** History of Quaternary Science; **Paleoclimate Reconstruction:** Sub-Milankovitch (DO/Heinrich) Events; **Glacial Climates:** Thermohaline Circulation; **Glaciation, Causes:** Astronomical Theory of Paleoclimates; **Paleoclimate Modeling:** Last Glacial Maximum GCMs; **Glaciations:** Overview; **Sea Level Studies:** Overview; Paleocceanography; **Ice Core Records:** Correlations Between Greenland and Antarctica.

Here you will find examples of all three functions of the cross-reference list: a topic discussed in greater detail elsewhere (e.g., **Paleoclimate Modeling:** Last Glacial Maximum GCMs), parallel discussion in other entries (e.g., **Glacial Climates:** Thermohaline Circulation), and reference to entries that broaden the discussion (e.g., **Glaciations:** Overview).

4. Index

The index provides you with the page number where the material is located. The index entries differentiate between material that is a whole entry, is part of an entry, or is data presented in a figure or a table. Detailed notes are provided on the opening page of the index.

5. Contributors

A full list of contributors is listed at the beginning of each volume.

CONTENTS LIST BY SUBJECT

Location references refer to the volume number and page number (separated by a colon)

Introduction and History of Science

INTRODUCTION | History of Quaternary Science, 1:1
INTRODUCTION | History of Dating Methods, 1:9
INTRODUCTION | Societal Relevance of Quaternary Research, 1:17
INTRODUCTION | Understanding Quaternary Climatic Change, 1:26

Paleoclimatology

Overview and Approaches

PALEOCLIMATE | Introduction, 3:87
PALEOCLIMATE RECONSTRUCTION | Approaches, 3:179
PALEOCLIMATE MODELING | Data-Model Comparisons, 3:135
PALEOCLIMATE | Timescales of Climate Change, 3:93
PALEOCLIMATE | Modern Analog Approaches in Paleoclimatology, 3:102
PALEOCLIMATE | Paleoclimate History of the Arctic, 3:113
PALEOCLIMATE | The Younger Dryas Climate Event, 3:126

Paleoclimate Reconstructions

PALEOCLIMATE RECONSTRUCTION | Pliocene Environments, 3:185
GLACIATION, CAUSES | Tectonic Uplift and Continental Configurations, 2:127
GLACIATION, CAUSES | Astronomical Theory of Paleoclimates, 2:136
PALEOCLIMATE RECONSTRUCTION | Paleodroughts and Society, 3:194
PALEOCLIMATE RECONSTRUCTION | Paleotempestology, 3:209

PALEOCLIMATE RECONSTRUCTION | Sub-Milankovitch (DO/Heinrich) Events, 3:200
PALEOCLIMATE RECONSTRUCTION | Younger Dryas Oscillation, Global Evidence, 3:222
PALEOCLIMATE RECONSTRUCTION | Historical Climatology, 3:237
PALEOCLIMATE RECONSTRUCTION | The Last Millennium, 3:229
PALEOCLIMATE RELEVANCE TO GLOBAL WARMING, 3:244

Climate Forcing Mechanisms

GLACIAL CLIMATES | Thermohaline Circulation, 1:737
GLACIAL CLIMATES | Biosphere Feedbacks, 1:721
GLACIAL CLIMATES | Volcanic and Solar Forcing, 1:748
GLACIAL CLIMATES | Effects of Atmospheric Dust, 1:729

Climate Modelling

PALEOCLIMATE MODELING | Quaternary Environments, 3:147
PALEOCLIMATE MODELING | The Last Interglacial, 3:155
PALEOCLIMATE MODELING | Last Glacial Maximum GCMs, 3:165
PALEOCLIMATE MODELING | Paleo-ENSO, 3:171

Dating Techniques

Radiometric Methods

DATING TECHNIQUES, 1:447
RADIOCARBON DATING | Conventional Method, 4:305
RADIOCARBON DATING | AMS Radiocarbon Dating, 4:316

RADIOCARBON DATING | Sources of Error, 4:324
RADIOCARBON DATING | Variations in Atmospheric ^{14}C , 4:329
RADIOCARBON DATING | Causes of Temporal ^{14}C Variations, 4:336
RADIOCARBON DATING | Calibration of the ^{14}C Record, 4:345
RADIOCARBON DATING | Charcoal, 4:353
RADIOCARBON DATING | ^{14}C of Plant Macrofossils, 4:361
K/Ar AND $^{40}\text{Ar}/^{39}\text{Ar}$ DATING, 2:477
U-SERIES DATING, 4:567
FISSION-TRACK DATING, 1:643
LUMINESCENCE DATING | Thermoluminescence, 2:643
LUMINESCENCE DATING | Optical Dating, 2:653
LUMINESCENCE DATING | Electron Spin Resonance Dating, 2:667
COSMOGENIC NUCLIDE DATING | Overview, 1:407
COSMOGENIC NUCLIDE DATING | Cosmic Ray Interactions in Minerals, 1:418
COSMOGENIC NUCLIDE DATING | Methods, 1:410
COSMOGENIC NUCLIDE DATING | Exposure Geochronology, 1:432
COSMOGENIC NUCLIDE DATING | Landscape Evolution, 1:440

Geomagnetic Excursions and Secular Variations in the Magnetic Field

Geomagnetic Excursions and Secular Variations, 1:705

Biological Methods

LICHENOMETRY, 2:565

Annual Layer Methods

VARVED MARINE SEDIMENTS, 4:582
VARVED LAKE SEDIMENTS, 4:573

Chemical Methods

AMINO-ACID DATING, 1:37

Quaternary Stratigraphy

QUATERNARY STRATIGRAPHY | Tephrochronology, 4:277
QUATERNARY STRATIGRAPHY | Overview, 4:189
QUATERNARY STRATIGRAPHY | Lithostratigraphy, 4:227

QUATERNARY STRATIGRAPHY | Continental Biostratigraphy, 4:206
QUATERNARY STRATIGRAPHY | Pedostratigraphy, 4:250
QUATERNARY STRATIGRAPHY | Morphostratigraphy/Allostratigraphy, 4:243
QUATERNARY STRATIGRAPHY | Climatostratigraphy, 4:222
QUATERNARY STRATIGRAPHY | Sequence Stratigraphy, 4:260
QUATERNARY STRATIGRAPHY | Chronostratigraphy, 4:215

Quaternary Glaciation

Glacial Landforms

GLACIAL LANDFORMS | Introduction, 1:755
GLACIAL LANDFORMS, ICE SHEETS | Growth and Decay, 1:877
GLACIAL LANDFORMS, ICE SHEETS | Evidence of Glacier and Ice Sheet Extent, 1:884
GLACIAL LANDFORMS | Evidence of Glacier Recession, 1:803
GLACIAL LANDFORMS, ICE SHEETS | PaleoeLAS, 1:909
GLACIAL LANDFORMS | Glacitectonic Structures and Landforms, 1:839
GLACIAL LANDFORMS, SEDIMENTS | Glaciogenic Lithofacies, 2:18
GLACIAL LANDFORMS | Moraine Forms and Genesis, 1:769
GLACIAL LANDFORMS, SEDIMENTS | Glaciofluvial Landforms of Deposition, 2:6
GLACIAL LANDFORMS | Glaciofluvial Landforms of Erosion, 1:825
GLACIAL LANDFORMS, SEDIMENTS | Glacial Erratics and Till Dispersal Indicators, 2:81
GLACIAL LANDFORMS, SEDIMENTS | Glaciomarine Sediments and Ice-Rafted Debris, 2:30
GLACIAL LANDFORMS, SEDIMENTS | Tills, 2:62
GLACIAL LANDFORMS, SEDIMENTS | Till Fabric Analysis, 2:76
GLACIAL LANDFORMS | Glacial Land Systems, 1:813
GLACIAL LANDFORMS, EROSIONAL FEATURES | Micro- to Macroscale Forms, 1:865
GLACIAL LANDFORMS, EROSIONAL FEATURES | Major Scale Forms, 1:847
GLACIAL LANDFORMS, ICE SHEETS | Trimlines and Palaeonunataks, 1:918
GLACIAL LANDFORMS | Quaternary Vulcanism: Subglacial Landforms, 1:780
GLACIAL LANDFORMS, SEDIMENTS | Micromorphology of Glacial Sediments, 2:52

GLACIAL LANDFORMS, SEDIMENTS | Clast Form Analysis, 2:1
 GLACIAL LANDFORMS, ICE SHEETS | Evidence of Glacier Flow Directions, 1:895
 GLACIAL LANDFORMS, SEDIMENTS | Glaciolacustrine, 2:43
 GLACIAL LANDFORMS, SEDIMENTS | Glacial Sequence Stratigraphy, 2:85

Permafrost and Periglacial Features

PERMAFROST AND PERIGLACIAL FEATURES | Permafrost, 3:464
 PERMAFROST AND PERIGLACIAL FEATURES | Active Layer Processes, 3:421
 PERMAFROST AND PERIGLACIAL FEATURES | Paraglacial Geomorphology, 3:553
 PERMAFROST AND PERIGLACIAL FEATURES | Slope Deposits and Forms, 3:481
 PERMAFROST AND PERIGLACIAL FEATURES | Frost Mounds: Active and Relict Forms, 3:472
 PERMAFROST AND PERIGLACIAL FEATURES | Patterned Ground, 3:452
 PERMAFROST AND PERIGLACIAL FEATURES | Talus Slopes, 3:566
 PERMAFROST AND PERIGLACIAL FEATURES | Thermokarst Topography, 3:574
 PERMAFROST AND PERIGLACIAL FEATURES | Ice Wedges and Ice-Wedge Casts, 3:436
 PERMAFROST AND PERIGLACIAL FEATURES | Cryoturbation Structures, 3:430
 PERMAFROST AND PERIGLACIAL FEATURES | Blockfields (Felsenmeer), 3:523
 PERMAFROST AND PERIGLACIAL FEATURES | Block/Rock Streams, 3:514
 PERMAFROST AND PERIGLACIAL FEATURES | Rock Weathering, 3:500
 PERMAFROST AND PERIGLACIAL FEATURES | Rock Glaciers and Protalus Forms, 3:535
 PERMAFROST AND PERIGLACIAL FEATURES | Periglacial Fluvial Sediments and Forms, 3:490
 PERMAFROST AND PERIGLACIAL FEATURES | Permafrost and Glacier Interactions, 3:507
 PERMAFROST AND PERIGLACIAL FEATURES | Yedoma: Late Pleistocene Ice-Rich Syngenetic Permafrost of Beringia, 3:542

Tree-ring section

DENDROCHRONOLOGY, 1:453
 DENDROCLIMATOLOGY, 1:459
 GLACIAL LANDFORMS, TREE RINGS | Dendrogeomorphology, 2:91
 GLACIAL LANDFORMS, TREE RINGS | Dendroglaciology, 2:104

Fluvial Environments

FLUVIAL ENVIRONMENTS | Sediments, 1:663
 FLUVIAL ENVIRONMENTS | Terrace Sequences, 1:684
 Glacial–Interglacial Scale Fluvial Responses, 2:112
 FLUVIAL ENVIRONMENTS | Responses to Rapid Environmental Change, 1:676
 FLUVIAL ENVIRONMENTS | Deltaic Environments, 1:693
 PALEOHYDROLOGY, 3:253

History of Quaternary Glaciations

GLACIATIONS | Overview, 2:143
 GLACIATIONS | Transition from Late Neogene to Early Quaternary Environments, 2:151
 GLACIATIONS | Early Quaternary (Pleistocene) and Precursors, 2:167
 GLACIATIONS | Middle Pleistocene in Eurasia, 2:172
 GLACIATIONS | Mid-Quaternary in North America, 2:180
 GLACIATIONS | Middle Pleistocene Glaciations in the Southern Hemisphere, 2:187
 GLACIATIONS | Late Quaternary of Antarctica, 2:216
 GLACIATIONS | Late Quaternary in North America, 2:245
 GLACIATIONS | Late Quaternary in Highland Asia, 2:236
 GLACIATIONS | Late Quaternary of the Southwest Pacific Region, 2:202
 GLACIATIONS | Late Pleistocene in Eurasia, 2:224
 GLACIATIONS | Late Pleistocene in South America, 2:250
 GLACIATIONS | Late Pleistocene Glacial Events in Beringia, 2:191
 GLACIATIONS | Neoglaciation in Europe, 2:257
 GLACIATIONS | Neoglaciation in the American Cordilleras, 2:269

Sea Level Studies

SEA LEVEL STUDIES | Overview, 4:369
 SEA LEVEL STUDIES | Geomorphological Indicators, 4:377
 SEA LEVEL STUDIES | Coral Records of Relative Sea-Level Changes, 4:409
 SEA LEVEL STUDIES | Use of Cave Data in Sea-Level Reconstructions, 4:460
 SEA LEVEL STUDIES | Eustatic Sea-Level Changes – Glacial–Interglacial Cycles, 4:429
 SEA LEVEL STUDIES | Eustatic Sea-Level Changes Since the Last Glacial Maximum, 4:439

SEA LEVEL STUDIES | Isostasy: Glaciation-Induced Sea-Level Change, 4:452
SEA LEVEL STUDIES | Sedimentary Indicators of Relative Sea-Level Changes – High Energy, 4:385
SEA LEVEL STUDIES | Sedimentary Indicators of Relative Sea-Level Changes – Low Energy, 4:396
SEA LEVEL STUDIES | Microfossil-Based Reconstructions of Holocene Relative Sea-Level Change, 4:419
SEA-LEVELS, LATE QUATERNARY | Late Quaternary Relative Sea-Level Changes in High Latitudes, 4:467
SEA-LEVELS, LATE QUATERNARY | Late Quaternary Sea-Level Changes in Greenland, 4:481
SEA-LEVELS, LATE QUATERNARY | Late Quaternary Relative Sea-Level Changes at Mid-Latitudes, 4:489
SEA-LEVELS, LATE QUATERNARY | Late Quaternary Relative Sea-Level Changes in the Tropics, 4:495
SEA-LEVELS, LATE QUATERNARY | Tectonics and Relative Sea-Level Change, 4:503

Paleosols and Wind-Blown Sediments

Paleosol Research

PALEOSOLS AND WIND-BLOWN SEDIMENTS | Overview, 3:357
PALEOSOLS AND WIND-BLOWN SEDIMENTS | Nature of Paleosols, 3:367
LOESS DEPOSITS: ORIGINS AND PROPERTIES, 2:573
PALEOSOLS AND WIND-BLOWN SEDIMENTS | Weathering Profiles, 3:392
PALEOSOLS AND WIND-BLOWN SEDIMENTS | Soil Micromorphology, 3:381
PALEOSOLS AND WIND-BLOWN SEDIMENTS | Soil Morphology in Quaternary Studies, 3:402
PALEOSOLS AND WIND-BLOWN SEDIMENTS | Mineral Magnetic Analysis, 3:375
PALEOSOLS AND WIND-BLOWN SEDIMENTS | Biogeochemical Role of Dust in Quaternary Climate Cycles, 3:412

Dune Studies

DUNE FIELDS | High Latitudes, 1:597
DUNE FIELDS | Mid-Latitudes, 1:606
DUNE FIELDS | Low Latitudes, 1:623

Loess Studies

LOESS RECORDS | Central Asia, 2:585
LOESS RECORDS | China, 2:595
LOESS RECORDS | Europe, 2:606
LOESS RECORDS | North America, 2:620
LOESS RECORDS | South America, 2:629
EOLIAN RECORDS, DEEP-SEA SEDIMENTS, 1:637

Lake Level Studies

LAKE LEVEL STUDIES | Overview, 2:483
LAKE LEVEL STUDIES | Modeling, 2:558
LAKE LEVEL STUDIES | Africa during the Late Quaternary, 2:499
LAKE LEVEL STUDIES | Asia, 2:506
LAKE LEVEL STUDIES | Australia, 2:524
LAKE LEVEL STUDIES | Latin America, 2:531
LAKE LEVEL STUDIES | North America, 2:537
LAKE LEVEL STUDIES | West-Central-Europe, 2:549

Paleobotany

Overview

PALEOBOTANY | Overview of Terrestrial Pollen Data, 2:699

Pollen Studies

POLLEN ANALYSIS, PRINCIPLES, 3:794
POLLEN METHODS AND STUDIES | Use of Pollen as Climate Proxies, 3:805
POLLEN METHODS AND STUDIES | Surface Samples and Trapping, 3:839
POLLEN METHODS AND STUDIES | The Biome Approach to Reconstructing Past Vegetation, 3:861
POLLEN METHODS AND STUDIES | POLLSCAPE Model, 3:871
POLLEN METHODS AND STUDIES | Reconstructing Past Biodiversity Development, 3:816
POLLEN METHODS AND STUDIES | Numerical Analysis Methods, 3:821
POLLEN METHODS AND STUDIES | Changing Plant Distributions and Abundances, 3:854
POLLEN METHODS AND STUDIES | Stand-Scale Palynology, 3:846
PALEOBOTANY | Paleophytogeography, 2:730
POLLEN METHODS AND STUDIES | Databases and Their Application, 3:831

Charred Particle Analysis

PALEOBOTANY | Charred Particle Analyses, 2:716

Quaternary Pollen Records

POLLEN RECORDS, LAST INTERGLACIAL OF EUROPE, 4:1
POLLEN RECORDS, LATE PLEISTOCENE | Africa, 4:9
POLLEN RECORDS, LATE PLEISTOCENE | Australasia, 4:18
POLLEN RECORDS, LATE PLEISTOCENE | Northern Asia, 4:27

POLLEN RECORDS, LATE PLEISTOCENE | Northern North America, 4:39
 POLLEN RECORDS, LATE PLEISTOCENE | Western North America, 4:72
 POLLEN RECORDS, LATE PLEISTOCENE | South America, 4:52
 POLLEN RECORDS, LATE PLEISTOCENE | Middle and Late Pleistocene in Southern Europe, 4:63
 POLLEN RECORDS, POSTGLACIAL | Africa, 4:85
 POLLEN RECORDS, POSTGLACIAL | Australia and New Zealand, 4:104
 POLLEN RECORDS, POSTGLACIAL | Northeastern North America, 4:115
 POLLEN RECORDS, POSTGLACIAL | Northwestern North America, 4:124
 POLLEN RECORDS, POSTGLACIAL | Southeastern North America, 4:133
 POLLEN RECORDS, POSTGLACIAL | Southwestern North America, 4:142
 POLLEN RECORDS, POSTGLACIAL | South America, 4:156
 POLLEN RECORDS, POSTGLACIAL | Northern Asia, 4:164
 POLLEN RECORDS, POSTGLACIAL | Northern Europe, 4:173
 POLLEN RECORDS, POSTGLACIAL | Southern Europe, 4:179
 POLLEN METHODS AND STUDIES | Archaeological Applications, 3:880

Phytoliths

PHYTOLITHS, 3:582

Ancient Plant DNA

PALEOBOTANY | Ancient Plant DNA, 2:705

Plant Macrofossil Studies

PLANT MACROFOSSIL INTRODUCTION, 3:593
 PLANT MACROFOSSIL METHODS AND STUDIES | Surface Samples, Taphonomy, Representation, 3:684
 PLANT MACROFOSSIL METHODS AND STUDIES | Treeline Studies, 3:690
 PLANT MACROFOSSIL METHODS AND STUDIES | Megafossils, 3:621
 PLANT MACROFOSSIL METHODS AND STUDIES | Paleolimnological Applications, 3:657
 PLANT MACROFOSSIL METHODS AND STUDIES | CO₂ Reconstruction from Fossil Leaves, 3:613
 PLANT MACROFOSSIL METHODS AND STUDIES | Rodent Middens, 3:674

PLANT MACROFOSSIL METHODS AND STUDIES | Mire and Peat Macros, 3:637
 PLANT MACROFOSSIL METHODS AND STUDIES | Use in Environmental Archaeology, 3:699
 PLANT MACROFOSSIL METHODS AND STUDIES | Dendroarchaeology, 3:630
 PLANT MACROFOSSIL METHODS AND STUDIES | Validation of Pollen Studies, 3:725

Plant Macrofossil Records

PLANT MACROFOSSIL RECORDS | Arctic Eurasia, 3:733
 PLANT MACROFOSSIL RECORDS | Greenland, 3:760
 PLANT MACROFOSSIL RECORDS | Arctic North America, 3:746
 PLANT MACROFOSSIL RECORDS | Holocene North America, 3:768
 PLANT MACROFOSSIL RECORDS | Late Glacial Multidisciplinary Studies, 3:785

Diatom Studies

DIATOM INTRODUCTION, 1:471
 DIATOM METHODS | Data Interpretation, 1:489
 DIATOM RECORDS | North Atlantic and Arctic, 1:562
 DIATOM RECORDS | Pacific, 1:571
 DIATOM RECORDS | Antarctic Waters, 1:527
 DIATOM RECORDS | Diatom Fossil Records from Marine Laminated Sediments, 1:554
 DIATOM RECORDS | Freshwater Laminated Sequences, 1:540
 DIATOM RECORDS | Large Lakes, 1:546
 DIATOM METHODS | Salinity and Climate Reconstructions from Continental Lakes, 1:507
 DIATOM METHODS | $\delta^{18}\text{O}$ Records, 1:481
 PALEOBOTANY | Silicon Isotopes in Diatoms, 2:734
 DIATOM METHODS | Use in Archaeology, 1:516
 DIATOM METHODS | Diatomites: Their Formation, Distribution, and Uses, 1:501
 Structures and Applications of Biomarkers from Arctic Sea Ice Diatoms, 1:588

Paleolimnology

PALEOLIMNOLOGY | Overview of Paleolimnology, 3:259
 PALEOLIMNOLOGY | Lake Chemistry, 3:292
 PALEOLIMNOLOGY | Physical Properties of Lake Sediments, 3:300
 PALEOLIMNOLOGY | Cladocera, 3:271
 PALEOLIMNOLOGY | Freshwater Mollusks, 3:281

PALEOLIMNOLOGY | Contributions of
Paleolimnological Research to
Biogeography, 3:313
PALEOLIMNOLOGY | Pigment Studies, 3:326
PALEOLIMNOLOGY | Visible and Infrared
Spectroscopical Applications, 3:349
PALEOLIMNOLOGY | Multiproxy Approaches, 3:339

Vertebrate Studies

VERTEBRATE OVERVIEW, 4:590
VERTEBRATE RECORDS | Early Pleistocene, 4:599
VERTEBRATE RECORDS | Early and Middle
Pleistocene of Northern Eurasia, 4:605
VERTEBRATE RECORDS | Mid-Pleistocene of
Southern Asia, 4:651
VERTEBRATE RECORDS | Mid-Pleistocene of
Africa, 4:615
VERTEBRATE RECORDS | Mid-Pleistocene of
Australia, 4:621
VERTEBRATE RECORDS | Mid-Pleistocene of
Europe, 4:639
VERTEBRATE RECORDS | Mid-Pleistocene of
North America, 4:646
VERTEBRATE RECORDS | Late Pleistocene of
Africa, 4:664
VERTEBRATE RECORDS | Late Pleistocene of North
America, 4:673
VERTEBRATE RECORDS | Late Pleistocene of South
America, 4:680
VERTEBRATE RECORDS | Late Pleistocene of
Southeast Asia, 4:693
VERTEBRATE RECORDS | Late Pleistocene
Megafaunal Extinctions, 4:700
VERTEBRATE RECORDS | Late Pleistocene
Mummified Mammals, 4:713
VERTEBRATE STUDIES | Ancient DNA, 4:719
VERTEBRATE STUDIES | Speciation and
Evolutionary Trends in Quaternary
Vertebrates, 4:723
VERTEBRATE STUDIES | Dwarfing and Gigantism in
Quaternary Vertebrates, 4:733
VERTEBRATE STUDIES | Interactions with
Hominids, 4:748

Insect Fossil Studies

Beetle Studies

BEETLE RECORDS | Overview, 1:161
BEETLE RECORDS | Late Tertiary and Early
Quaternary Records, 1:173
BEETLE RECORDS | Middle Pleistocene of Western
Europe, 1:184

BEETLE RECORDS | Late Pleistocene of
Europe, 1:200
BEETLE RECORDS | Late Pleistocene of North
America, 1:221
BEETLE RECORDS | Late Pleistocene of South
America, 1:235
BEETLE RECORDS | Late Pleistocene of
Australia, 1:191
BEETLE RECORDS | Late Pleistocene of Northern
Asia, 1:255
BEETLE RECORDS | Late Pleistocene of Japan, 1:207
BEETLE RECORDS | Late Pleistocene of New
Zealand, 1:244
BEETLE RECORDS | Postglacial Europe, 1:274
BEETLE RECORDS | Postglacial North America, 1:282

Chironomid Studies

CHIRONOMID RECORDS | Chironomid
Overview, 1:355
CHIRONOMID RECORDS | Late Pleistocene of
Europe, 1:373
CHIRONOMID RECORDS | Africa, 1:361
CHIRONOMID RECORDS | Late Pleistocene of
Europe, 1:373
CHIRONOMID RECORDS | Postglacial
Europe, 1:386
CHIRONOMID RECORDS | Postglacial Southern
Hemisphere, 1:398

Oribatid Mite Studies

ORIBATID MITES, 2:680

Paleoceanography

Paleoceanographic Studies

PALEOCEANOGRAPHY | Paleoceanography An
Overview, 2:745
PALEOCEANOGRAPHY, BIOLOGICAL PROXIES |
Corals, Sclerosponges and Mollusks, 2:795
PALEOCEANOGRAPHY, BIOLOGICAL PROXIES |
Benthic Foraminifera, 2:765
PALEOCEANOGRAPHY, BIOLOGICAL PROXIES |
Planktic Foraminifera, 2:825
PALEOCEANOGRAPHY, BIOLOGICAL PROXIES |
Radiolarians and Silicoflagellates, 2:830
PALEOCEANOGRAPHY, BIOLOGICAL PROXIES |
Coccolithophores, 2:783
PALEOCEANOGRAPHY, BIOLOGICAL PROXIES |
Alkenone Paleothermometry Based on the
Haptophyte Algae, 2:755
PALEOCEANOGRAPHY, BIOLOGICAL PROXIES |
Marine Diatoms, 2:816

PALEOCEANOGRAPHY, BIOLOGICAL PROXIES |
Dinoflagellates, 2:800

PALEOCEANOGRAPHY, PHYSICAL AND
CHEMICAL PROXIES | Terrigenous
Sediments, 2:941

PALEOCEANOGRAPHY, PHYSICAL AND
CHEMICAL PROXIES | Oxygen Isotopic
Composition of Seawater, 2:915

PALEOCEANOGRAPHY, PHYSICAL AND
CHEMICAL PROXIES | Oxygen-Isotope
Stratigraphy of the Oceans, 2:907

PALEOCEANOGRAPHY, PHYSICAL AND
CHEMICAL PROXIES | Dissolution of Deep-Sea
Carbonates, 2:859

PALEOCEANOGRAPHY, PHYSICAL AND
CHEMICAL PROXIES | Mg/Ca and Sr/Ca
Paleothermometry from Calcareous Marine
Fossils, 2:871

PALEOCEANOGRAPHY, BIOLOGICAL PROXIES |
Temperature Proxies Census Counts, 2:841

PALEOCEANOGRAPHY, PHYSICAL AND
CHEMICAL PROXIES | Salinity Proxies
 $\delta^{18}\text{O}$, 2:932

PALEOCEANOGRAPHY, PHYSICAL AND
CHEMICAL PROXIES | Nutrient Proxies, 2:899

PALEOCEANOGRAPHY, PHYSICAL AND
CHEMICAL PROXIES | Carbon Cycle
Proxies, 2:849

PALEOCEANOGRAPHY, PHYSICAL AND
CHEMICAL PROXIES | Radioisotope
Proxies, 2:923

PALEOCEANOGRAPHY, BIOLOGICAL PROXIES |
Biomarkers Indicators of Past Climate, 2:775

PALEOCEANOGRAPHY, PHYSICAL AND
CHEMICAL PROXIES | Magnetic Proxies and
Susceptibility, 2:884

Paleoceanographic Records

PALEOCEANOGRAPHY, RECORDS | Early
Pleistocene, 3:1

PALEOCEANOGRAPHY, RECORDS | Late
Pleistocene North Atlantic, 3:9

PALEOCEANOGRAPHY, RECORDS | Late
Pleistocene South Atlantic, 3:18

PALEOCEANOGRAPHY, RECORDS | Late
Pleistocene North Pacific, 3:33

PALEOCEANOGRAPHY, RECORDS | Postglacial
North Atlantic, 3:55

PALEOCEANOGRAPHY, RECORDS | Postglacial
North Pacific, 3:62

PALEOCEANOGRAPHY, RECORDS | Postglacial
South Pacific, 3:73

PALEOCEANOGRAPHY, RECORDS | Postglacial
Indian Ocean, 3:46

Ice Core Studies

Ice Core Studies

ICE CORE METHODS | Overview, 2:278

ICE CORES | History of Research, Greenland and
Antarctica, 2:435

ICE CORE METHODS | Chronologies, 2:303

ICE CORE METHODS | Stable Isotopes, 2:347

ICE CORE METHODS | Conductivity Studies, 2:319

ICE CORE METHODS | Glaciochemistry, 2:326

ICE CORE METHODS | Biological Material, 2:288

ICE CORE METHODS | Microparticle and Trace
Element Studies, 2:342

ICE CORE METHODS | CO₂ Studies, 2:311

ICE CORE METHODS | Methane Studies, 2:334

ICE CORE METHODS | ¹⁰Be and Cosmogenic
Radionuclides in Ice Cores, 2:353

ICE CORE METHODS | Studies of Firn Air, 2:361

Ice Core Records

ICE CORES | History of Carbon Monoxide and
Ultra-Trace Gases from Ice Cores, 2:463

ICE CORES | History of Nitrous Oxide from Ice
Cores, 2:471

ICE CORES | Dynamics of the Greenland Ice
Sheet, 2:439

ICE CORES | Dynamics of the East Antarctic Ice
Sheet, 2:456

ICE CORES | Dynamics of the West Antarctic Ice
Sheet, 2:448

ICE CORE METHODS | Borehole Temperature
Records, 2:298

ICE CORE RECORDS | Greenland Stable
Isotopes, 2:403

ICE CORE RECORDS | Antarctic Stable Isotopes, 2:395

ICE CORE RECORDS | Ice Margin Sites, 2:416

ICE CORE RECORDS | Correlations Between
Greenland and Antarctica, 2:410

ICE CORE RECORDS | Africa, 2:373

ICE CORE RECORDS | Chinese, Tibetan
Mountains, 2:379

ICE CORE RECORDS | South America, 2:387

ICE CORE RECORDS | Thermal Diffusion
Paleotemperature Records, 2:431

Stable Isotope Research – Carbonates

CARBONATE STABLE ISOTOPES | Overview, 1:291

CARBONATE STABLE ISOTOPES |
Speleothems, 1:294

CARBONATE STABLE ISOTOPES | Terrestrial Teeth
and Bones, 1:304

CARBONATE STABLE ISOTOPES | Terrestrial
Organic Materials, 1:315
CARBONATE STABLE ISOTOPES | Non-Lacustrine
Terrestrial Studies, 1:322
CARBONATE STABLE ISOTOPES | Lake
Sediments, 1:333
CARBONATE STABLE ISOTOPES | Nonmarine
Biogenic Carbonates, 1:341

Humans in the Quaternary

ARCHAEOLOGICAL RECORDS | Overview, 1:49
ARCHAEOLOGICAL RECORDS | 2.7 Myr–300 000
Years Ago in Africa, 1:59
ARCHAEOLOGICAL RECORDS | 2.7 Myr–300 000
Years Ago in Asia, 1:67
ARCHAEOLOGICAL RECORDS | 1.9 Myr–300 000
Years Ago in Europe, 1:83
ARCHAEOLOGICAL RECORDS | Global Expansion
300 000–8000 Years Ago, Africa, 1:91
ARCHAEOLOGICAL RECORDS | Global Expansion
300 000–8000 Years Ago, Asia, 1:98
ARCHAEOLOGICAL RECORDS | Global
Expansion 300 000–8000 Years Ago,
Australia, 1:108

ARCHAEOLOGICAL RECORDS | Global Expansion
300 000–8000 Years Ago, Americas, 1:119
ARCHAEOLOGICAL RECORDS | Neanderthal
Demise, 1:146
ARCHAEOLOGICAL RECORDS | Human Evolution
in the Quaternary, 1:135
ARCHAEOLOGICAL RECORDS | Postglacial
Adaptations, 1:154

Use of Quaternary proxies in forensic science

USE OF QUATERNARY PROXIES IN FORENSIC
SCIENCE | Use of Geology in Forensic Science:
Search to Locate Burials, 4:522
USE OF QUATERNARY PROXIES IN FORENSIC
SCIENCE | Analysis Techniques in Forensic
Palynology, 4:556
USE OF QUATERNARY PROXIES IN FORENSIC
SCIENCE | Soils and Sediment, 4:535
USE OF QUATERNARY PROXIES IN FORENSIC
SCIENCE | The Use of Macroscopic Plant Remains
in Forensic Science, 4:542
USE OF QUATERNARY PROXIES IN FORENSIC
SCIENCE | Insects, 4:548
Diatoms, 1:522

CONTENTS

<i>Dedications</i>	<i>v</i>
<i>Foreword</i>	<i>ix</i>
<i>Introduction</i>	<i>xi</i>
<i>Editorial Advisory Board</i>	<i>xiii</i>
<i>Contributors</i>	<i>xv</i>
<i>How to Use the Encyclopedia</i>	<i>xxvii</i>
<i>Contents List by Subject</i>	<i>xxix</i>

VOLUME 1

INTRODUCTORY ARTICLES 1

History of Quaternary Science <i>S A Elias</i>	1
History of Dating Methods <i>A G Wintle</i>	9
Societal Relevance of Quaternary Research <i>S A Elias</i>	17
Understanding Quaternary Climatic Change <i>J J Lowe, M J C Walker, and S C Porter</i>	26

A

ALKENONE STUDIES *see* PALEOCEANOGRAPHY, BIOLOGICAL PROXIES: Alkenone Paleothermometry Based on the Haptophyte Algae

ALLOSTRATIGRAPHY *see* QUATERNARY STRATIGRAPHY: Morphostratigraphy/Allostratigraphy

Amino Acid Dating <i>G H Miller, D S Kaufman, and S J Clarke</i>	37
---	----

ANATOMICALLY MODERN HUMANS *see* ARCHAEOLOGICAL RECORDS: Global Expansion 300 000–8000 Years Ago, Africa; Global Expansion 300 000–8000 Years Ago, Asia; Global Expansion 300 000–8000 Years Ago, Australia; Global Expansion 300 000–8000 Years Ago, Americas

ARCHAEOLOGICAL RECORDS 49

Overview <i>C Gamble</i>	49
-----------------------------	----

2.7 Myr–300 000 Years Ago in Africa <i>J W K Harris, D R Braun, and M Pante</i>	59
2.7 Myr–300 000 Years Ago in Asia <i>R Dennell</i>	67
1.9 Myr–300 000 Years Ago in Europe <i>J McNabb</i>	83
Global Expansion 300 000–8000 Years Ago, Africa <i>A E Close and T Minichillo</i>	91
Global Expansion 300 000–8000 Years Ago, Asia <i>M D Petraglia and R Dennell</i>	98
Global Expansion 300 000–8000 Years Ago, Australia <i>R Cosgrove</i>	108
Global Expansion 300 000–8000 Years Ago, Americas <i>T Goebel</i>	119
Human Evolution in the Quaternary <i>L Cashmore</i>	135
Neanderthal Demise <i>W Davies</i>	146
Postglacial Adaptations <i>G Bailey</i>	154
B	
BEETLE RECORDS	161
Overview <i>S A Elias</i>	161
Late Tertiary and Early Quaternary Records <i>S A Elias and S Kuzmina</i>	173
Middle Pleistocene of Western Europe <i>G R Coope</i>	184
Late Pleistocene of Australia <i>N Porch</i>	191
Late Pleistocene of Europe <i>G Lemdahl and G R Coope</i>	200
Late Pleistocene of Japan <i>M Hayashi</i>	207
Late Pleistocene of North America <i>S A Elias</i>	221
Late Pleistocene of South America <i>A C Ashworth</i>	235
Late Pleistocene of New Zealand <i>M Marra</i>	244
Late Pleistocene of Northern Asia <i>A Sher and S Kuzmina</i>	255

Postglacial Europe <i>P Ponel</i>	274
Postglacial North America <i>S A Elias</i>	282

BERINGIA *see* ARCHAEOLOGICAL RECORDS: Global Expansion 300 000–8000 Years Ago, Americas;
 BEETLE RECORDS: Late Pleistocene of North America; DUNE FIELDS: High Latitudes; GLACIATIONS:
 Late Pleistocene Glacial Events in Beringia; PALEOCEANOGRAPHY, RECORDS: Postglacial North
 Pacific; PLANT MACROFOSSIL RECORDS: Arctic North America; POLLEN RECORDS, LATE
 PLEISTOCENE: Northern North America; VERTEBRATE RECORDS: Late Pleistocene of North America

BIOGENIC CARBONATE STUDIES *see* CARBONATE STABLE ISOTOPES: Nonmarine Biogenic Carbonates

BOND CYCLES *see* PALEOCLIMATE RECONSTRUCTION: Sub-Milankovitch (DO/Heinrich) Events

C

CARBONATE STABLE ISOTOPES 291

Overview <i>H Schwarcz</i>	291
Speleothems <i>H Schwarcz</i>	294
Terrestrial Teeth and Bones <i>H Bocherens and D G Drucker</i>	304
Terrestrial Organic Materials <i>D McCarroll and N Loader</i>	315
Non-Lacustrine Terrestrial Studies <i>J Quade and T Cerling</i>	322
Lake Sediments <i>S M Bernasconi and J A McKenzie</i>	333
Nonmarine Biogenic Carbonates <i>S J Carpenter</i>	341

CARBON DIOXIDE, ATMOSPHERIC CONCENTRATIONS *see* CARBONATE STABLE ISOTOPES:
 Overview; ICE CORE METHODS: CO₂ Studies; PALEOCEANOGRAPHY, PHYSICAL AND CHEMICAL
 PROXIES: Carbon Cycle Proxies ($\delta^{11}\text{B}$, $\delta^{13}\text{C}_{\text{calcite}}$, $\delta^{13}\text{C}_{\text{organic}}$, Shell Weights, B/Ca, U/Ca, Zn/Ca, Ba/Ca);
 PLANT MACROFOSSIL METHODS AND STUDIES: CO₂ Reconstruction from Fossil Leaves

CAVE ART *see* VERTEBRATE STUDIES: Interactions with Hominids

CHARCOAL STUDIES *see* PALEOBOTANY: Charred Particle Analyses

CHIRONOMID RECORDS 355

Chironomid Overview <i>I R Walker</i>	355
Africa <i>H Eggermont and D Verschuren</i>	361
Late Pleistocene of Europe <i>S J Brooks</i>	373
Postglacial Europe <i>G Velle and O Heiri</i>	386

Postglacial Southern Hemisphere <i>J Massferro and M Vandergoes</i>	398
--	-----

CLADOCERA STUDIES *see* PALEOLIMNOLOGY: Cladocera

CLIMATE CHANGE *see* PALEOCLIMATE: Introduction; Timescales of Climate Change; PALEOCLIMATE MODELING: Data–Model Comparisons; Quaternary Environments; The Last Interglacial; Last Glacial Maximum GCMs; Paleo-ENSO; PALEOCLIMATE RECONSTRUCTION: Approaches; Pliocene Environments; Paleodroughts and Society; Sub-Milankovitch (DO/Heinrich) Events; Paleotempestology; Younger Dryas Oscillation, Global Evidence; The Last Millennium; Historical Climatology; Paleoclimate Relevance to Global Warming

CLIMATE MODELING, QUATERNARY *see* PALEOCLIMATE MODELING: Quaternary Environments

COCCOLITHOPHORE STUDIES *see* PALEOCEANOGRAPHY, BIOLOGICAL PROXIES: Alkenone Paleothermometry Based on the Haptophyte Algae; Coccolithophores

COLEOPTERA FOSSIL RECORDS *see* BEETLE RECORDS: Overview; Late Tertiary and Early Quaternary Records; Middle Pleistocene of Western Europe; Late Pleistocene of Australia; Late Pleistocene of Europe; Late Pleistocene of Japan; Late Pleistocene of North America; Late Pleistocene of South America; Late Pleistocene of New Zealand; Late Pleistocene of Northern Asia; Postglacial Europe; Postglacial North America

CORAL STUDIES *see* PALEOCEANOGRAPHY, BIOLOGICAL PROXIES: Corals, Sclerosponges and Mollusks; SEA LEVEL STUDIES: Coral Records of Relative Sea-Level Changes

COSMOGENIC NUCLIDE DATING 407

Overview <i>J C Gosse</i>	407
------------------------------	-----

Methods <i>J M Schaefer and N Lifton</i>	410
---	-----

Cosmic Ray Interactions in Minerals <i>D Lal</i>	418
---	-----

Exposure Geochronology <i>S Ivy-Ochs and F Kober</i>	432
---	-----

Landscape Evolution <i>D E Granger</i>	440
---	-----

CUT-MARKED BONE *see* VERTEBRATE STUDIES: Interactions with Hominids

D

DANSGAARD-OESCHGER EVENTS *see* PALEOCLIMATE RECONSTRUCTION: Paleodroughts and Society

Dating Techniques <i>A J T Jull</i>	447
--	-----

DENDROARCHAEOLOGY *see* PLANT MACROFOSSIL METHODS AND STUDIES: Dendroarchaeology

Dendrochronology <i>B L Coulthard and D J Smith</i>	453
--	-----

Dendroclimatology <i>B H Luckman</i>	459
---	-----

Diatom Introduction	471
<i>V J Jones</i>	
DIATOM METHODS	481
$\delta^{18}\text{O}$ Records	481
<i>M J Leng, P A Barker, G E A Swann, and A M Snelling</i>	
Data Interpretation	489
<i>A Korhola</i>	
Diatomites: Their Formation, Distribution, and Uses	501
<i>R J Flower</i>	
Salinity and Climate Reconstructions from Continental Lakes	507
<i>S C Fritz</i>	
Use in Archaeology	516
<i>N G Cameron</i>	
Diatoms	522
<i>N G Cameron</i>	
DIATOM RECORDS	527
Antarctic Waters	527
<i>C E Stickley, J Pike, and V J Jones</i>	
Freshwater Laminated Sequences	540
<i>H Simola</i>	
Large Lakes	546
<i>A W Mackay</i>	
Diatom Fossil Records from Marine Laminated Sediments	554
<i>J Pike and C E Stickley</i>	
North Atlantic and Arctic	562
<i>N Koç, A Miettinen, and C E Stickley</i>	
Pacific	571
<i>I Koizumi</i>	
Structures and Applications of Biomarkers from Arctic Sea Ice Diatoms	588
<i>S T Belt, G Massé, and M Poulin</i>	
DIATOMS, MARINE <i>see</i> PALEOCEANOGRAPHY, BIOLOGICAL PROXIES: Marine Diatoms	
DINOFLAGELLATES <i>see</i> PALEOCEANOGRAPHY, BIOLOGICAL PROXIES: Dinoflagellates	
DNA, FOSSIL ANIMAL <i>see</i> VERTEBRATE STUDIES: Ancient DNA	
DNA, FOSSIL PLANTS <i>see</i> PALEOBOTANY: Ancient Plant DNA	
DROUGHT HISTORY RECONSTRUCTION <i>see</i> PALEOCLIMATE RECONSTRUCTION: Paleodroughts and Society	
DUNE FIELDS	597
High Latitudes	597
<i>S A Wolfe</i>	
Mid-Latitudes	606
<i>J Sun and D R Muhs</i>	

Low Latitudes <i>N Lancaster</i>	623
-------------------------------------	-----

E

EL NIÑO SOUTHERN OSCILLATION <i>see</i> PALEOCEANOGRAPHY, RECORDS: Postglacial South Pacific; PALEOCLIMATE MODELING: Paleo-ENSO	
ELECTRON SPIN RESONANCE DATING <i>see</i> LUMINESCENCE DATING: Electron Spin Resonance Dating	
EOLIAN SEDIMENTS <i>see</i> PALEOSOLS AND WIND-BLOWN SEDIMENTS: Overview; Nature of Paleosols; Mineral Magnetic Analysis; Soil Micromorphology; Weathering Profiles; Soil Morphology in Quaternary Studies; Biogeochemical Role of Dust in Quaternary Climate Cycles	
Eolian Records, Deep-Sea Sediments <i>D K Rea</i>	637
EQUILIBRIUM LINE ALTITUDE (ELA) RECONSTRUCTION <i>see</i> GLACIAL LANDFORMS, ICE SHEETS: Paleo-ELAs	
ERRATICS, GLACIAL <i>see</i> GLACIAL LANDFORMS, SEDIMENTS: Glacial Erratics and Till Dispersal Indicators	
EXTINCTIONS, QUATERNARY VERTEBRATES <i>see</i> VERTEBRATE RECORDS: Late Pleistocene Megafaunal Extinctions	

F

FELSENMEER (BLOCKFIELDS) <i>see</i> PERMAFROST AND PERIGLACIAL FEATURES: Block/Rock Streams	
Fission-Track Dating <i>J A Westgate, N D Naeser, and B V Alloway</i>	643

FLUVIAL ENVIRONMENTS

Sediments <i>A Aslan</i>	663
Responses to Rapid Environmental Change <i>T E Törnqvist</i>	676
Terrace Sequences <i>D J Merritts</i>	684
Deltaic Environments <i>L Giosan and S L Goodbred</i>	693

FORAMINIFERA STUDIES *see* PALEOCEANOGRAPHY: Paleocyanography An Overview; PALEOCEANOGRAPHY, BIOLOGICAL PROXIES: Temperature Proxies, Census Counts; Benthic Foraminifera; Planktic Foraminifera; PALEOCEANOGRAPHY, PHYSICAL AND CHEMICAL PROXIES: Oxygen-Isotope Stratigraphy of the Oceans

FOSSIL MITES *see* Oribatid Mites

G

Geomagnetic Excursions and Secular Variations <i>A P Roberts and G M Turner</i>	705
--	-----

GLACIAL CLIMATES	721
Biosphere Feedbacks <i>V Brovkin</i>	721
Effects of Atmospheric Dust <i>I Tegen</i>	729
Thermohaline Circulation <i>S Rahmstorf</i>	737
Volcanic and Solar Forcing <i>D T Shindell</i>	748
GLACIAL LANDFORMS	755
Introduction <i>D J A Evans and D I Benn</i>	755
Moraine Forms and Genesis <i>D J A Evans</i>	769
Quaternary Vulcanism: Subglacial Landforms <i>J L Smellie</i>	780
Evidence of Glacier Recession <i>P M Colgan</i>	803
Glacial Landsystems <i>D J A Evans</i>	813
Glaciofluvial Landforms of Erosion <i>A E Kehew, M L Lord, and A L Kozlowski</i>	825
Glacitectonic Structures and Landforms <i>D J A Evans</i>	839
GLACIAL LANDFORMS, EROSIONAL FEATURES	847
Major Scale Forms <i>I S Evans</i>	847
Micro- to Macroscale Forms <i>B R Rea</i>	865
GLACIAL LANDFORMS, ICE SHEETS	877
Growth and Decay <i>M J Siegert</i>	877
Evidence of Glacier and Ice Sheet Extent <i>D M Mickelson and C Winguth</i>	884
Evidence of Glacier Flow Directions <i>C R Stokes</i>	895
Paleo-ELAs <i>A Nesje</i>	909
Trimlines and Paleonunataks <i>C K Ballantyne</i>	918

VOLUME 2**GLACIAL LANDFORMS, SEDIMENTS 1**

Clast Form Analysis 1
D I Benn

Glaciofluvial Landforms of Deposition 6
J L Carrivick and A J Russell

Glaciogenic Lithofacies 18
N Eyles and M Lazorek

Glaciomarine Sediments and Ice-Rafted Debris 30
C Ó Cofaigh

Glaciolacustrine 43
D J A Evans

Micromorphology of Glacial Sediments 52
J F Hiemstra

Tills 62
D J A Evans

Till Fabric Analysis 76
D I Benn

Glacial Erratics and Till Dispersal Indicators 81
D J A Evans

Glacial Sequence Stratigraphy 85
D J A Evans

GLACIAL LANDFORMS, TREE RINGS 91

Dendrogeomorphology 91
H Gärtner and I Heinrich

Dendroglaciology 104
B L Coulthard and D J Smith

Glacial–Interglacial Scale Fluvial Responses 112
M D Blum

GLACIATION, CAUSES 127

Tectonic Uplift and Continental Configurations 127
L A Owen

Astronomical Theory of Paleoclimates 136
A Berger, M-F Loutre, and Q Z Yin

GLACIATIONS 143

Overview 143
J Ehlers and P L Gibbard

Transition from Late Neogene to Early Quaternary Environments 151
M Sarnthein

Early Quaternary (Pleistocene) and Precursors 167
J Ehlers, V Astakhov, P L Gibbard, Ó Ingólfsson, J Mangerud, and J I Svendsson

Middle Pleistocene in Eurasia <i>J Ehlers, V Astakhov, P L Gibbard, J Mangerud, and J I Svendsen</i>	172
Mid-Quaternary in North America <i>C E Jennings, J S Aber, G Balco, R Barendregt, P R Bierman, C W Rovey II, M Roy, L H Thorleifson, and J A Mason</i>	180
Middle Pleistocene Glaciations in the Southern Hemisphere <i>A Coronato and J Rabassa</i>	187
Late Pleistocene Glacial Events in Beringia <i>S A Elias and J Brigham-Grette</i>	191
Late Quaternary of the Southwest Pacific Region <i>D J A Barrell</i>	202
Late Quaternary of Antarctica <i>Ó Ingólfsson</i>	216
Late Pleistocene in Eurasia <i>J Ehlers, V Astakhov, P L Gibbard, J Mangerud, and J I Svendsen</i>	224
Late Quaternary in Highland Asia <i>L A Owen</i>	236
Late Quaternary in North America <i>J T Andrews and A S Dyke</i>	245
Late Pleistocene in South America <i>A Coronato and J Rabassa</i>	250
Neoglaciation in Europe <i>J A Matthews</i>	257
Neoglaciation in the American Cordilleras <i>S C Porter</i>	269

GLOBAL WARMING *see* Paleoclimate Relevance to Global Warming

H

HEINRICH EVENTS *see* PALEOCLIMATE RECONSTRUCTION: Paleodroughts and Society

HISTORICAL CLIMATE RECORDS *see* PALEOCLIMATE RECONSTRUCTION: Historical Climatology; The Last Millennium

HOLOCENE ENVIRONMENTS *see* Dendroclimatology; ARCHAEOLOGICAL RECORDS: Postglacial Adaptations; BEETLE RECORDS: Postglacial Europe; Postglacial North America; CHIRONOMID RECORDS: Postglacial Europe; Postglacial Southern Hemisphere; DIATOM METHODS: Use in Archaeology; GLACIATIONS: Neoglaciation in Europe; Neoglaciation in the American Cordilleras; ICE CORES: Dynamics of the Greenland Ice Sheet; Dynamics of the West Antarctic Ice Sheet; Dynamics of the East Antarctic Ice Sheet; PALEOCEANOGRAPHY, RECORDS: Postglacial Indian Ocean; Postglacial North Atlantic; Postglacial North Pacific; Postglacial South Pacific; PALEOCLIMATE: Timescales of Climate Change; PALEOCLIMATE MODELING: Paleo-ENSO; PALEOCLIMATE RECONSTRUCTION: Historical Climatology; The Last Millennium; PLANT MACROFOSSIL METHODS AND STUDIES: Treeline Studies; PLANT MACROFOSSIL RECORDS: Holocene North America; POLLEN RECORDS, POSTGLACIAL: Africa; Australia and New Zealand; Northeastern North America; Northwestern North America; Southeastern North America; Southwestern North America; South America; Northern Asia; Northern Europe; Southern Europe

HUMAN EVOLUTION *see* ARCHAEOLOGICAL RECORDS: Overview; 2.7 Myr–300 000 Years Ago in Africa; 2.7 Myr–300 000 Years Ago in Asia; 1.9 Myr–300 000 Years Ago in Europe; Human Evolution in the Quaternary

I

ICE CORE METHODS	277
Overview	278
<i>E J Brook</i>	
Biological Material	288
<i>J C Priscu, B C Christner, C M Foreman, and G Royston-Bishop</i>	
Borehole Temperature Records	298
<i>K M Cuffey</i>	
Chronologies	303
<i>J Schwander</i>	
CO ₂ Studies	311
<i>T Blunier and T M Jenk</i>	
Conductivity Studies	319
<i>R Mulvaney</i>	
Glaciochemistry	326
<i>K J Kreutz and B G Koffman</i>	
Methane Studies	334
<i>J Chappellaz</i>	
Microparticle and Trace Element Studies	342
<i>J R McConnell</i>	
Stable Isotopes	347
<i>E J Brook</i>	
¹⁰ Be and Cosmogenic Radionuclides in Ice Cores	353
<i>R Muscheler</i>	
Studies of Firn Air	361
<i>C Buizert</i>	
ICE CORE RECORDS	373
Africa	373
<i>L G Thompson and M E Davis</i>	
Chinese, Tibetan Mountains	379
<i>C P Wake</i>	
South America	387
<i>L G Thompson and M E Davis</i>	
Antarctic Stable Isotopes	395
<i>E J Brook</i>	
Greenland Stable Isotopes	403
<i>B M Vinther and S J Johnsen</i>	
Correlations Between Greenland and Antarctica	410
<i>E J Brook</i>	

Ice Margin Sites <i>V V Petrenko</i>	416
Thermal Diffusion Paleotemperature Records <i>A M Grachev</i>	431
ICE CORES	435
History of Research, Greenland and Antarctica <i>M Aydin</i>	435
Dynamics of the Greenland Ice Sheet <i>C S Hvidberg, A Svensson, and S L Buchardt</i>	439
Dynamics of the West Antarctic Ice Sheet <i>R Bindshadler</i>	448
Dynamics of the East Antarctic Ice Sheet <i>E D Waddington and C S Lingle</i>	456
History of Carbon Monoxide and Ultra-Trace Gases from Ice Cores <i>M Aydin</i>	463
History of Nitrous Oxide from Ice Cores <i>A Schilt</i>	471
ICE SHEETS, PLEISTOCENE <i>see</i> GLACIAL LANDFORMS, ICE SHEETS: Growth and Decay; Evidence of Glacier and Ice Sheet Extent; Trimlines and Paleonunataks; GLACIATIONS: Early Quaternary (Pleistocene) and Precursors; Middle Pleistocene in Eurasia; Mid-Quaternary in North America; Middle Pleistocene Glaciations in the Southern Hemisphere; Late Pleistocene Glacial Events in Beringia; Late Quaternary of the Southwest Pacific Region; Late Quaternary of Antarctica; Late Pleistocene in Eurasia; Late Quaternary in Highland Asia; Late Quaternary in North America; Late Pleistocene in South America; ICE CORES: Dynamics of the Greenland Ice Sheet; Dynamics of the West Antarctic Ice Sheet; Dynamics of the East Antarctic Ice Sheet	
ICE WEDGES, ICE WEDGE CASTS <i>see</i> PERMAFROST AND PERIGLACIAL FEATURES: Ice Wedges and Ice-Wedge Casts	
K	
K/Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ Dating <i>J R Wijbrans and K F Kuiper</i>	477
L	
LAKE CHEMISTRY RECONSTRUCTION <i>see</i> PALEOLIMNOLOGY: Lake Chemistry	
LAKE LEVEL STUDIES	483
Overview <i>R T Jones and J T Jordan</i>	483
Africa during the Late Quaternary <i>M E Edwards</i>	499
Asia <i>G Yu, B Xue, and Y Li</i>	506
Australia <i>J Magee</i>	524

Latin America <i>S E Metcalfe</i>	531
North America <i>J R Stone and S C Fritz</i>	537
West-Central Europe <i>M Magny</i>	549
Modeling <i>J Vassiljev</i>	558
Lichenometry <i>D P McCarthy</i>	565

LITHOSTRATIGRAPHY *see* QUATERNARY STRATIGRAPHY: Lithostratigraphy

Loess Deposits: Origins and Properties <i>D R Muhs</i>	573
---	-----

LOESS RECORDS 585

Central Asia <i>A E Dodonov</i>	585
China <i>S C Porter</i>	595
Europe <i>D-D Rousseau, E Derbyshire, P Antoine, and C Hatté</i>	606
North America <i>H M Roberts, D R Muhs, and E A Bettis III</i>	620
South America <i>M A Zárate</i>	629

LUMINESCENCE DATING 643

Thermoluminescence <i>O B Lian</i>	643
Optical Dating <i>O B Lian</i>	653
Electron Spin Resonance Dating <i>A J T Jull</i>	667

M

MAGNETIC POLARITY STUDIES *see* Geomagnetic Excursions and Secular Variations

MAMMALIAN EVOLUTION *see* Vertebrate Overview; VERTEBRATE RECORDS: Early Pleistocene; Mid-Pleistocene of Africa; Mid-Pleistocene of Southern Asia; VERTEBRATE STUDIES: Ancient DNA; Speciation and Evolutionary Trends in Quaternary Vertebrates

MARINE ISOTOPE STAGES *see* PALEOCEANOGRAPHY, PHYSICAL AND CHEMICAL PROXIES: Oxygen-Isotope Stratigraphy of the Oceans

MEGAFAUNA, PLEISTOCENE *see* Vertebrate Overview; VERTEBRATE RECORDS: Mid-Pleistocene of Australia; Mid-Pleistocene of North America; Early Pleistocene; Late Pleistocene of Africa;

Late Pleistocene of North America; Late Pleistocene of South America; Late Pleistocene of Southeast Asia; Late Pleistocene Megafaunal Extinctions; Mid-Pleistocene of Africa; Mid-Pleistocene of Europe; Mid-Pleistocene of Southern Asia; VERTEBRATE STUDIES: Interactions with Hominids

MEGAFAUNAL EXTINCTION *see* VERTEBRATE RECORDS: Late Pleistocene Megafaunal Extinctions; VERTEBRATE STUDIES: Interactions with Hominids

METHANE STUDIES, ICE CORES *see* ICE CORE METHODS: Methane Studies

Mg/Ca AND Sr/Ca STUDIES *see* PALEOCEANOGRAPHY, PHYSICAL AND CHEMICAL PROXIES: Mg/Ca and Sr/Ca Paleothermometry from Calcareous Marine Fossils

MICROMORPHOLOGY OF SEDIMENTS *see* GLACIAL LANDFORMS, SEDIMENTS: Micromorphology of Glacial Sediments; PALEOSOLS AND WIND-BLOWN SEDIMENTS: Soil Micromorphology

MIDGES *see* CHIRONOMID RECORDS: Africa; Chironomid Overview; Late Pleistocene of Europe; Postglacial Europe; Postglacial Southern Hemisphere

MILANKOVITCH THEORY *see* GLACIATION, CAUSES: Astronomical Theory of Paleoclimates

MOLLUSKS, MARINE *see* PALEOCEANOGRAPHY, BIOLOGICAL PROXIES: Corals, Sclerosponges and Mollusks

MORAINES, GLACIAL *see* GLACIAL LANDFORMS: Moraine Forms and Genesis; Evidence of Glacier Recession; Glacial Landsystems; Glacitectonic Structures and Landforms; GLACIAL LANDFORMS, SEDIMENTS: Glacial Erratics and Till Dispersal Indicators

MORPHOSTRATIGRAPHY *see* QUATERNARY STRATIGRAPHY: Morphostratigraphy/Allostratigraphy

N

NEANDERTHAL DEMISE *see* ARCHAEOLOGICAL RECORDS: Neanderthal Demise

NEOGLACIATION *see* GLACIATIONS: Neoglaciation in Europe; Neoglaciation in the American Cordilleras

O

OPTICALLY-STIMULATED LUMINESCENCE DATING *see* LUMINESCENCE DATING: Optical Dating

Oribatid Mites

J M Erickson and R B Platt Jr.

680

OXYGEN ISOTOPE STAGES *see* PALEOCEANOGRAPHY, PHYSICAL AND CHEMICAL PROXIES: Oxygen-Isotope Stratigraphy of the Oceans

P

PACKRAT MIDDENS *see* PLANT MACROFOSSIL METHODS AND STUDIES: Rodent Middens

PALEOANTHROPOLOGY *see* ARCHAEOLOGICAL RECORDS: Overview; 2.7 Myr–300 000 Years Ago in Africa; 2.7 Myr–300 000 Years Ago in Asia; 1.9 Myr–300 000 Years Ago in Europe; Global Expansion 300 000–8000 Years Ago, Africa; Global Expansion 300 000–8000 Years Ago, Asia; Global Expansion 300 000–8000 Years Ago, Australia; Global Expansion 300 000–8000 Years Ago, Americas; Human Evolution in the Quaternary; Neanderthal Demise; Postglacial Adaptations

PALEOBOTANY

699

Overview of Terrestrial Pollen Data

699

R H W Bradshaw

Ancient Plant DNA

705

N Wales, R Allaby, E Willerslev, and M T P Gilbert

Charred Particle Analyses <i>K J Brown and M J Power</i>	716
Paleophytogeography <i>A E Bjune</i>	730
Silicon Isotopes in Diatoms <i>J J Tyler</i>	734
PALEOCEANOGRAPHY	745
Paleoceanography An Overview <i>D M Anderson and K E Lee</i>	745
PALEOCEANOGRAPHY, BIOLOGICAL PROXIES	755
Alkenone Paleothermometry Based on the Haptophyte Algae <i>S L Ho, B D A Naafs, and F Lamy</i>	755
Benthic Foraminifera <i>R Saraswat and R Nigam</i>	765
Biomarker Indicators of Past Climate <i>J P Sachs, K Pahnke, R Smittenberg, and Z Zhang</i>	775
Coccolithophores <i>J-A Flores and F J Sierro</i>	783
Corals, Sclerosponges and Mollusks <i>T M Quinn and B R Schöne</i>	795
Dinoflagellates <i>A de Vernal, A Rochon, and T Radi</i>	800
Marine Diatoms <i>F Abrantes and I M Gil</i>	816
Planktic Foraminifera <i>H J Dowsett and M M Robinson</i>	825
Radiolarians and Silicoflagellates <i>D Lazarus</i>	830
Temperature Proxies, Census Counts <i>J D Ortiz</i>	841
PALEOCEANOGRAPHY, PHYSICAL AND CHEMICAL PROXIES	849
Carbon Cycle Proxies ($\delta^{11}\text{B}$, $\delta^{13}\text{C}_{\text{calcite}}$, $\delta^{13}\text{C}_{\text{organic}}$, Shell Weights, B/Ca, U/Ca, Zn/Ca, Ba/Ca) <i>B Hönisch and K A Allen</i>	849
Dissolution of Deep-Sea Carbonates <i>S Barker</i>	859
Mg/Ca and Sr/Ca Paleothermometry from Calcareous Marine Fossils <i>Y Rosenthal and B Linsley</i>	871
Magnetic Proxies and Susceptibility <i>R G Hatfield and J S Stoner</i>	884
Nutrient Proxies <i>T M Marchitto</i>	899

Oxygen-Isotope Stratigraphy of the Oceans <i>F C Bassinot</i>	907
Oxygen Isotope Composition of Seawater <i>E J Rohling</i>	915
Radioisotope Proxies <i>R Francois</i>	923
Salinity Proxies $\delta^{18}\text{O}$ <i>P D Naidu</i>	932
Terrigenous Sediments <i>A M Franzese and S R Hemming</i>	941

VOLUME 3

PALEOCEANOGRAPHY, RECORDS	1
Early Pleistocene <i>T de Garidel-Thoron</i>	1
Late Pleistocene North Atlantic <i>D M Anderson</i>	9
Late Pleistocene South Atlantic <i>S Mulitza, A Paul, and G Wefer</i>	18
Late Pleistocene North Pacific <i>T Kiefer</i>	33
Postglacial Indian Ocean <i>P D Naidu</i>	46
Postglacial North Atlantic <i>M W Kerwin and K A Huguen</i>	55
Postglacial North Pacific <i>J I Martínez</i>	62
Postglacial South Pacific <i>F Lamy and R de Pol-Holz</i>	73
PALEOCLIMATE	87
Introduction <i>C J Mock</i>	87
Timescales of Climate Change <i>P J Bartlein</i>	93
Modern Analog Approaches in Paleoclimatology <i>C J Mock and J J Shinker</i>	102
Paleoclimate History of the Arctic <i>G H Miller, J Brigham-Grette, R B Alley, L Anderson, H A Bauch, M S V Douglas, M E Edwards, S A Elias, B P Finney, J J Fitzpatrick, S V Funder, A Geirsdóttir, T D Herbert, L D Hinzman, D S Kaufman, G M MacDonald, L Polyak, A Robock, M C Serreze, J P Smol, R Spielhagen, J W C White, A P Wolfe, and E W Wolff</i>	113
The Younger Dryas Climate Event <i>A E Carlson</i>	126

PALEOCLIMATE MODELING	135
Data–Model Comparisons <i>S P Harrison</i>	135
Quaternary Environments <i>C S Jackson</i>	147
The Last Interglacial <i>M Montoya</i>	155
Last Glacial Maximum GCMs <i>A J Broccoli</i>	165
Paleo-ENSO <i>C J Mock</i>	171
 PALEOCLIMATE RECONSTRUCTION	 179
Approaches <i>B N Shuman</i>	179
Pliocene Environments <i>R Z Poore</i>	185
Paleodroughts and Society <i>C J Mock</i>	194
Sub-Milankovitch (DO/Heinrich) Events <i>L Labeyrie, L Skinner, and E Cortijo</i>	200
Paleotempestology <i>K-b Liu</i>	209
Younger Dryas Oscillation, Global Evidence <i>S Björck</i>	222
The Last Millennium <i>M E Mann</i>	229
Historical Climatology <i>M Chenoweth</i>	237
 Paleoclimate Relevance to Global Warming <i>S H Schneider and M D Mastrandrea</i>	 244
Paleohydrology <i>V R Thorndycraft</i>	253
 PALEOLIMNOLOGY	 259
Overview of Paleolimnology <i>M S V Douglas</i>	259
Cladocera <i>M Rautio and L Nevalainen</i>	271
Freshwater Mollusks <i>C G De Francesco</i>	281
Lake Chemistry <i>D Antoniadis</i>	292

Physical Properties of Lake Sediments <i>K R Hodder and R Gilbert</i>	300
Contributions of Paleolimnological Research to Biogeography <i>K A Moser</i>	313
Pigment Studies <i>S McGowan</i>	326
Multiproxy Approaches <i>N Michelutti and J P Smol</i>	339
Visible and Infrared Spectroscopical Applications <i>P Rosén and H Vogel</i>	349
PALEOLITHIC <i>see</i> ARCHAEOLOGICAL RECORDS: Overview; 2.7 Myr–300 000 Years Ago in Africa; 2.7 Myr–300 000 Years Ago in Asia; 1.9 Myr–300 000 Years Ago in Europe; Global Expansion 300 000–8000 Years Ago, Africa; Global Expansion 300 000–8000 Years Ago, Asia; Global Expansion 300 000–8000 Years Ago, Australia; Global Expansion 300 000–8000 Years Ago, Americas; Neanderthal Demise	
PALEOSOLS AND WIND-BLOWN SEDIMENTS	357
Overview <i>D R Muhs</i>	357
Nature of Paleosols <i>J A Mason and P M Jacobs</i>	367
Mineral Magnetic Analysis <i>M J Singer and K L Verosub</i>	375
Soil Micromorphology <i>R A Kemp</i>	381
Weathering Profiles <i>E A Bettis III</i>	392
Soil Morphology in Quaternary Studies <i>L McFadden</i>	402
Biogeochemical Role of Dust in Quaternary Climate Cycles <i>K E Kohfeld</i>	412
PATTERNED GROUND <i>see</i> PERMAFROST AND PERIGLACIAL FEATURES: Patterned Ground	
PEAT STUDIES <i>see</i> PLANT MACROFOSSIL METHODS AND STUDIES: Mire and Peat Macros	
PEDOSTRATIGRAPHY <i>see</i> QUATERNARY STRATIGRAPHY: Pedostratigraphy	
PERMAFROST AND PERIGLACIAL FEATURES	421
Active Layer Processes <i>N I Shiklomanov and F E Nelson</i>	421
Cryoturbation Structures <i>J Vandenberghe</i>	430
Ice Wedges and Ice-Wedge Casts <i>J Murton</i>	436
Patterned Ground <i>C K Ballantyne</i>	452

Permafrost <i>C R Burn</i>	464
Frost Mounds: Active and Relict Forms <i>N Ross</i>	472
Slope Deposits and Forms <i>C Harris</i>	481
Periglacial Fluvial Sediments and Forms <i>J van Huissteden, J Vandenberghe, P L Gibbard, and J Lewin</i>	490
Rock Weathering <i>J Murton</i>	500
Permafrost and Glacier Interactions <i>R I Waller</i>	507
Block/Rock Streams <i>P Wilson</i>	514
Blockfields (Felsenmeer) <i>B R Rea</i>	523
Rock Glaciers and Protalus Forms <i>A Kääb</i>	535
Yedoma: Late Pleistocene Ice-Rich Syngenetic Permafrost of Beringia <i>L Schirrmeister, D Froese, V Tumskey, G Grosse, and S Wetterich</i>	542
Paraglacial Geomorphology <i>C K Ballantyne</i>	553
Talus Slopes <i>B H Luckman</i>	566
Thermokarst Topography <i>C R Burn</i>	574
PERMAFROST HISTORY <i>see</i> PERMAFROST AND PERIGLACIAL FEATURES: Active Layer Processes; Cryoturbation Structures; Patterned Ground; Permafrost	
Phytoliths <i>M S Blinnikov</i>	582
PIGMENTS, FOSSIL <i>see</i> PALEOLIMNOLOGY: Pigment Studies	
PINGOS <i>see</i> PERMAFROST AND PERIGLACIAL FEATURES: Frost Mounds: Active and Relict Forms	
Plant Macrofossil Introduction <i>H H Birks</i>	593
PLANT MACROFOSSIL METHODS AND STUDIES	613
CO ₂ Reconstruction from Fossil Leaves <i>M Rundgren</i>	613
Megafossils <i>G M MacDonald</i>	621
Dendroarchaeology <i>R H Towner</i>	630

Mire and Peat Macros <i>D Mauquoy and B van Geel</i>	637
Paleolimnological Applications <i>M-J Gaillard and H H Birks</i>	657
Rodent Middens <i>S A Elias</i>	674
Surface Samples, Taphonomy, Representation <i>A C Dieffenbacher-Krall</i>	684
Treeline Studies <i>W Tinner</i>	690
Use in Environmental Archaeology <i>S Jacomet</i>	699
Validation of Pollen Studies <i>S T Jackson and R K Booth</i>	725
PLANT MACROFOSSIL RECORDS	733
Arctic Eurasia <i>F Kienast</i>	733
Arctic North America <i>N H Bigelow, G D Zazula, and D E Atkinson</i>	746
Greenland <i>O Bennike</i>	760
Holocene North America <i>R G Baker</i>	768
Late Glacial Multidisciplinary Studies <i>B Ammann, H H Birks, A Walanus, and K Wasylkowa</i>	785
PLATE TECTONICS <i>see</i> GLACIAL LANDFORMS: Glacitectonic Structures and Landforms; GLACIATION, CAUSES: Tectonic Uplift and Continental Configurations; GLACIATIONS: Late Quaternary in North America	
PLIOCENE ENVIRONMENTS <i>see</i> PALEOCLIMATE RECONSTRUCTION: Pliocene Environments	
Pollen Analysis, Principles <i>H Seppä</i>	794
POLLEN METHODS AND STUDIES	805
Use of Pollen as Climate Proxies <i>S Brewer, J Guiot, and D Barboni</i>	805
Reconstructing Past Biodiversity Development <i>B V Odgaard</i>	816
Numerical Analysis Methods <i>H J B Birks</i>	821
Databases and Their Application <i>E C Grimm, R H W Bradshaw, S Brewer, S Flantua, T Giesecke, A-M Lézine, H Takahara, and J W Williams</i>	831
Surface Samples and Trapping <i>A Poska</i>	839

Stand-Scale Palynology <i>R H W Bradshaw</i>	846
Changing Plant Distributions and Abundances <i>T Giesecke</i>	854
The Biome Approach to Reconstructing Past Vegetation <i>M E Edwards</i>	861
POLLSCAPE Model: Simulation Approach for Pollen Representation of Vegetation and Land Cover <i>S Sugita</i>	871
Archaeological Applications <i>M-J Gaillard</i>	880

VOLUME 4

Pollen Records, Last Interglacial of Europe <i>C Tzedakis</i>	1
--	---

POLLEN RECORDS, LATE PLEISTOCENE 9

Africa <i>M E Meadows and B M Chase</i>	9
Australasia <i>P Kershaw and S van der Kaars</i>	18
Northern Asia <i>A V Lozhkin and P M Anderson</i>	27
Northern North America <i>N H Bigelow</i>	39
South America <i>H Hooghiemstra and J C Berrio</i>	52
Middle and Late Pleistocene in Southern Europe <i>J-L de Beaulieu, P C Tzedakis, V Andrieu-Ponel, and F Guiter</i>	63
Western North America <i>R S Thompson</i>	72

POLLEN RECORDS, POSTGLACIAL 85

Africa <i>A-M Lézine</i>	85
Australia and New Zealand <i>J R Dodson</i>	104
Northeastern North America <i>J W Williams and B N Shuman</i>	115
Northwestern North America <i>D G Gavin and F S Hu</i>	124
Southeastern North America <i>D A Willard</i>	133
Southwestern North America <i>P E Wigand</i>	142

South America <i>H Behling</i>	156
Northern Asia <i>A A Andreev and P E Tarasov</i>	164
Northern Europe <i>M J Bunting</i>	173
Southern Europe <i>L Sadori</i>	179

POTASSIUM-ARGON DATING *see* K/Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ Dating

Q

QUATERNARY STRATIGRAPHY 189

Overview <i>B Pillans</i>	189
Continental Biostratigraphy <i>T van Kolfschoten</i>	206
Chronostratigraphy <i>B Pillans</i>	215
Climatostratigraphy <i>P L Gibbard</i>	222
Lithostratigraphy <i>W E Westerhoff and H J T Weerts</i>	227
Morphostratigraphy/Allostratigraphy <i>P D Hughes</i>	243
Pedostratigraphy <i>A Palmer</i>	250
Sequence Stratigraphy <i>T R Naish, S T Abbott, and R M Carter</i>	260
Tephrochronology <i>B V Alloway, D J Lowe, G Larsen, P A R Shane, and J A Westgate</i>	277

R

RADIOLARIAN STUDIES *see* PALEOCEANOGRAPHY, BIOLOGICAL PROXIES: Radiolarians and Silicoflagellates

RADIOCARBON DATING 305

Conventional Method <i>G T Cook and J van der Plicht</i>	305
AMS Radiocarbon Dating <i>A J T Jull</i>	316
Sources of Error <i>E M Scott</i>	324

Variations in Atmospheric ^{14}C <i>J van der Plicht</i>	329
Causes of Temporal ^{14}C Variations <i>G S Burr</i>	336
Calibration of the ^{14}C Record <i>P J Reimer, R W Reimer, and M Blaauw</i>	345
Charcoal <i>M I Bird</i>	353
^{14}C of Plant Macrofossils <i>C Hatté and A J T Jull</i>	361

ROCK GLACIERS *see* PERMAFROST AND PERIGLACIAL FEATURES: Rock Glaciers and Protalus Forms

RODENT MIDDENS *see* PLANT MACROFOSSIL METHODS AND STUDIES: Rodent Middens

S

SEA LEVEL STUDIES 369

Overview <i>I Shennan</i>	369
Geomorphological Indicators <i>P A Pirazzoli</i>	377
Sedimentary Indicators of Relative Sea-Level Changes – High Energy <i>J A G Cooper</i>	385
Sedimentary Indicators of Relative Sea-Level Changes – Low Energy <i>R J Edwards</i>	396
Coral Records of Relative Sea-Level Changes <i>C D Woodroffe</i>	409
Microfossil-Based Reconstructions of Holocene Relative Sea-Level Change <i>W R Gehrels</i>	419
Eustatic Sea-Level Changes – Glacial–Interglacial Cycles <i>C V Murray-Wallace</i>	429
Eustatic Sea-Level Changes Since the Last Glacial Maximum <i>P L Whitehouse and S L Bradley</i>	439
Isostasy: Glaciation-Induced Sea-Level Change <i>G Milne and I Shennan</i>	452
Use of Cave Data in Sea-Level Reconstructions <i>A Dutton</i>	460

SEA-LEVELS, LATE QUATERNARY 467

Late Quaternary Relative Sea-Level Changes in High Latitudes <i>C Ó Cofaigh and M J Bentley</i>	467
Late Quaternary Sea-Level Changes in Greenland <i>S A Woodroffe and A J Long</i>	481
Late Quaternary Relative Sea-Level Changes at Mid-Latitudes <i>A C Kemp, B P Horton, and S E Engelhart</i>	489

Late Quaternary Relative Sea-Level Changes in the Tropics <i>Y Zong</i>	495
Tectonics and Relative Sea-Level Change <i>A R Nelson</i>	503

SEA SURFACE TEMPERATURE RECONSTRUCTION *see* PALEOCEANOGRAPHY: Paleocceanography An Overview; PALEOCEANOGRAPHY, BIOLOGICAL PROXIES: Alkenone Paleothermometry Based on the Haptophyte Algae; Biomarker Indicators of Past Climate; Coccolithophores; Dinoflagellates; Marine Diatoms; Planktic Foraminifera; Temperature Proxies, Census Counts; PALEOCEANOGRAPHY, PHYSICAL AND CHEMICAL PROXIES: Carbon Cycle Proxies ($\delta^{11}\text{B}$, $\delta^{13}\text{C}_{\text{calcite}}$, $\delta^{13}\text{C}_{\text{organic}}$, Shell Weights, B/Ca, U/Ca, Zn/Ca, Ba/Ca); Mg/Ca and Sr/Ca Paleothermometry from Calcareous Marine Fossils; Oxygen-Isotope Stratigraphy of the Oceans; PALEOCEANOGRAPHY, RECORDS: Late Pleistocene North Atlantic; Late Pleistocene North Pacific; Late Pleistocene South Atlantic; Postglacial Indian Ocean; Postglacial North Atlantic; Postglacial North Pacific; Postglacial South Pacific

SILICOFLAGELLATES *see* PALEOCEANOGRAPHY, BIOLOGICAL PROXIES: Radiolarians and Silicoflagellates

SPELEOTHEMS *see* CARBONATE STABLE ISOTOPES: Speleothems

STABLE ISOTOPE STUDIES, CARBONATES *see* CARBONATE STABLE ISOTOPES: Overview; Speleothems; Terrestrial Teeth and Bones; Nonmarine Biogenic Carbonates; Terrestrial Organic Materials; Non-Lacustrine Terrestrial Studies; Lake Sediments

STABLE ISOTOPE STUDIES, DEEP SEA RECORDS *see* PALEOCEANOGRAPHY, PHYSICAL AND CHEMICAL PROXIES: Oxygen-Isotope Stratigraphy of the Oceans; Oxygen Isotope Composition of Seawater; Salinity Proxies $\delta^{18}\text{O}$; PALEOCEANOGRAPHY, RECORDS: Early Pleistocene; Late Pleistocene North Atlantic; Late Pleistocene South Atlantic; Postglacial Indian Ocean; Postglacial North Atlantic; Postglacial North Pacific; Postglacial South Pacific; ICE CORE METHODS: Stable Isotopes; ICE CORE RECORDS: Antarctic Stable Isotopes; Greenland Stable Isotopes; Correlations Between Greenland and Antarctica

STABLE ISOTOPE STUDIES, ICE CORES *see* ICE CORE METHODS: Stable Isotopes; ICE CORE RECORDS: Antarctic Stable Isotopes; Greenland Stable Isotopes; Correlations Between Greenland and Antarctica

SUB-MILANKOVITCH EVENTS *see* PALEOCLIMATE RECONSTRUCTION: Sub-Milankovitch (DO/Heinrich) Events

T

TALUS SLOPES *see* PERMAFROST AND PERIGLACIAL FEATURES: Talus Slopes

TEETH AND BONES, STABLE ISOTOPE STUDIES *see* CARBONATE STABLE ISOTOPES: Terrestrial Teeth and Bones

TEPHROCHRONOLOGY *see* QUATERNARY STRATIGRAPHY: Tephrochronology

THERMOHALINE CIRCULATION OF THE OCEANS *see* GLACIAL CLIMATES: Thermohaline Circulation

THERMOLUMINESCENCE DATING *see* LUMINESCENCE DATING: Thermoluminescence

TILL, GLACIAL *see* GLACIAL LANDFORMS, SEDIMENTS: Till Fabric Analysis; Tills; Glacial Erratics and Till Dispersal Indicators

TREE RINGS *see* Dendrochronology; Dendroclimatology; GLACIAL LANDFORMS, TREE RINGS: Dendrogeomorphology; Dendroglaciology; PLANT MACROFOSSIL METHODS AND STUDIES: Dendroarchaeology

TREELINE RECONSTRUCTION *see* POLLEN METHODS AND STUDIES: Archaeological Applications

U**USE OF QUATERNARY PROXIES IN FORENSIC SCIENCE 522**

Use of Geology in Forensic Science: Search to Locate Burials 522
L Donnelly

Soils and Sediment 535
R C Murray

The Use of Macroscopic Plant Remains in Forensic Science 542
J H Bock

Insects 548
D E Gennard

Analytical Techniques in Forensic Palynology 556
V M Bryant

U-Series Dating 567
W G Thompson

V

Varved Lake Sediments 573
B Zolitschka

Varved Marine Sediments 582
K A Hughen and B Zolitschka

Vertebrate Overview 590
D C Schreve

VERTEBRATE RECORDS 599

Early Pleistocene 599
L Rook, M Delfino, M P Ferretti, and L Abbazzi

Early and Middle Pleistocene of Northern Eurasia 605
I Vislobokova and A Tesakov

Mid-Pleistocene of Africa 615
L C Bishop and A Turner

Mid-Pleistocene of Australia 621
G J Prideaux

Mid-Pleistocene of Europe 639
R Sardella

Mid-Pleistocene of North America 646
C J Bell

Mid-Pleistocene of Southern Asia 651
J de Vos

Late Pleistocene of Africa 664
T E Steele

Late Pleistocene of North America 673
J I Mead

Late Pleistocene of South America <i>M Ubilla</i>	680
Late Pleistocene of Southeast Asia <i>Y Chaimanee</i>	693
Late Pleistocene Megafaunal Extinctions <i>S A Elias and D C Schreve</i>	700
Late Pleistocene Mummified Mammals <i>C R Harington</i>	713
VERTEBRATE STUDIES	719
Ancient DNA <i>I Barnes, R Barnett, and B Shapiro</i>	719
Speciation and Evolutionary Trends in Quaternary Vertebrates <i>A M Lister</i>	723
Dwarfing and Gigantism in Quaternary Vertebrates <i>M R Palombo and R Rozzi</i>	733
Interactions with Hominids <i>K V Boyle</i>	748
VOLCANIC ASH <i>see</i> QUATERNARY STRATIGRAPHY: Tephrochronology	
VULCANISM <i>see</i> GLACIAL CLIMATES: Volcanic and Solar Forcing; GLACIAL LANDFORMS: Quaternary Vulcanism: Subglacial Landforms; QUATERNARY STRATIGRAPHY: Tephrochronology	
W	
WIND-BLOWN SEDIMENTS (LOESS, SAND DUNES) <i>see</i> Eolian Records, Deep-Sea Sediments; Loess Deposits: Origins and Properties; DUNE FIELDS: High Latitudes; Mid-Latitudes; Low Latitudes; LOESS RECORDS: Central Asia; China; Europe; North America; South America	
Y	
YOUNGER DRYAS OSCILLATION <i>see</i> PALEOCLIMATE RECONSTRUCTION: Younger Dryas Oscillation, Global Evidence	
<i>Index</i>	757

Pollen Records, Last Interglacial of Europe

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Introduction

The Last Interglacial represents the most recent geological interval during which conditions were similar to the present interglacial, but with negligible anthropogenic effects. The most widely used stratigraphic unit of the Last Interglacial is marine isotope stage (MIS) 5e of the deep-sea stratigraphy (Shackleton, 1969), reflecting an interval of minimum global ice volume with sea level as high, or indeed higher, than present. The oldest assignment of a terrestrial deposit to the Last Interglacial is the Eemian, the term introduced by Harting (1874) to describe a subsoil characterized by warm marine mollusks in the Eem valley near Amersfoort in The Netherlands. Deposits of similar age were also found in Germany and Denmark, and a pollen zonation scheme was described by Jessen and Milthers (1928). The Amersfoort site has been used as the type locality for the Eemian Stage (Zagwijn, 1961) for northwest Europe, but because the sequence was condensed, a more complete sedimentary record at Amsterdam terminal has been chosen as the lectostratotype (Cleveringa et al., 2000). The succession of pollen zones described in The Netherlands can be observed in numerous deposits across northwest Europe, with some vegetation events considered sufficiently time-parallel (Zagwijn, 1996). This has led to the use of the term 'Eemian' being applied across northwest Europe, often extended well beyond this region over the whole of Europe, and also used interchangeably with 'Last Interglacial' in more general contexts (e.g., the 'Eemian' in Greenland ice cores). Kukla et al. (2002) have pointed out that although the term Eemian has been used as a chronostratigraphic stage (i.e., with time-parallel boundaries everywhere), in local contexts it is defined using biostratigraphic or climatostratigraphic units whose boundaries may be diachronous; they have therefore recommended that the use of the term 'Eemian' be restricted to northwest Europe. Thus, the generic term 'Last Interglacial' is here used to describe the forest interval in different parts of Europe, which is broadly equivalent with MIS 5e, although their upper and lower boundaries may diverge in detail, depending on location.

Depositional Environments and Vegetation Development

Given its stratigraphical position, the Last Interglacial has been the most extensively studied pre-Holocene stage. South of the last (Weichselian) glacial limit, hundreds of former lake basins overlie sediments of the penultimate (late Saalian: MIS 6) glaciation. These represent kettle hole features formed by buried ice during the retreat of the Saalian ice and filled by Last Interglacial organic sediments, overlain by minerogenic sediments of the ensuing cold interval. In coastal lowlands, Last

Interglacial marine sediments are often interlayered with lacustrine beds, as in the Eemian stratotype at Amersfoort. North of the Weichselian ice limit, several Last Interglacial deposits have also been identified, covered by Weichselian till or deformed by ice (Mangerud, 1991). A prominent feature is the existence of several terrestrial sites with Last Interglacial marine sediments, indicating an extensive marine transgression. The so-called 'Eemian Sea' (Figure 1) flooded main river valleys along the North Sea coast and linked the Baltic Sea and Arctic Ocean over Lake Ladoga, Lake Onega, and the White Sea, turning Fennoscandia into an island (see review by Donner (1995)). This resulted from a combination of greater crustal downwarping and a more easterly located center of the late Saalian ice sheet relative to the Weichselian, rapid deglaciation and somewhat higher Last Interglacial sea levels. South of the late Saalian glacial limit, a concentration of deposits is found in the Alpine foreland, formed in depressions left by the penultimate Alpine glaciation. Additional deposits have been found in similar geomorphological settings related to local ice caps, such as the Grande Pile basin in the Vosges mountains (Woillard, 1978). These often contain longer polleniferous sequences than sites further north, with the entire last climatic cycle represented at Grande Pile. In southern Europe, the overall density of pollen sites of Last Interglacial age is low. However, they usually form part of long sequences that have accumulated relatively undisturbed in volcanic crater lakes or tectonic basins, providing complete records with clear chronostratigraphical relations.

Analyses of these deposits have produced a substantial body of palynological evidence, providing an insight into the vegetational character of large parts of Europe (see Figure 2 for location of some of the sites discussed). Within the classic Eemian area of western and central Europe, pollen diagrams show a characteristic succession of overlapping peaks, suggesting a uniform vegetation development across large regions (e.g., Grüger (1991) and Zagwijn (1996)). Despite varying nomenclatures, pollen zonation schemes from this area are therefore easily correlated (Table 1) and reveal the following broad patterns: *Betula* (birch) formed the first forests followed closely by *Pinus* (pine). These were replaced by *Ulmus* (elm) and *Quercus* (oak) and then by *Corylus* (hazel). Compared to the Holocene, hazel appears to have spread more slowly during the Last Interglacial (i.e., its peak is recorded after, rather than before or during, the oak peak). A *Corylus*–*Taxus*–*Tilia* phase followed, noted by the relatively high values of *Taxus* (yew) achieved in certain places and by the presence of *Tilia tomentosa* (linden), today confined to the Balkans and the Near East (Grüger, 1991). A characteristic feature of the Last Interglacial is the ensuing expansion of *Carpinus* (hornbeam), which completely dominated large parts of European deciduous forests. Following that, *Abies* (fir) and/or *Picea* (spruce) populations expanded. The closing stages of the interglacial were

marked by the re-expansion of pine and birch. Perhaps equally characteristic of the Eemian as the hornbeam phase is the absence of *Fagus* (beech) across Europe, except in Italy where low but continuous presence has been recorded (Follieri et al., 1988; Allen et al., 2005). It remains unclear whether climatic

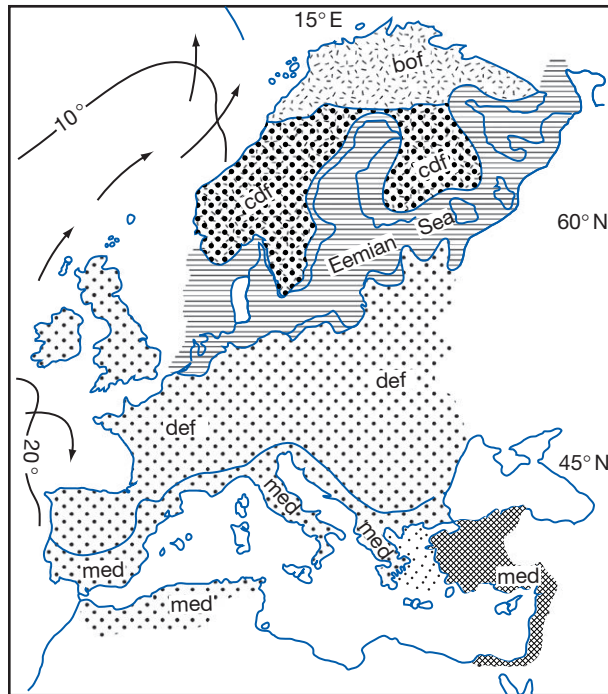


Figure 1 Reconstruction of European paleoenvironments during the Last Interglacial optimum ~125 ka. Vegetation units: boreal forest (bof), mixed coniferous-deciduous forest (cdf), temperate deciduous forest (def), Mediterranean evergreen forest/scrub (med). From van Andel TH and Tzedakis PC (1995) Paleolithic landscapes in Europe and environs, 150 000–25 000 years ago: An overview. *Quaternary Science Reviews* 15: 481–500.

conditions, competition, pathogenic attack, or a combination of these factors during the Last Interglacial (or perhaps during the preceding cold stage) inhibited the expansion of beech. Despite the overall similarities, Zagwijn (1996) has also drawn attention to some regional peculiarities. For example, the yew peak is not strongly expressed in eastern Poland, while the linden peak is not apparent in the southernmost and westernmost sites of the classic Eemian area. The linden peak is most strongly expressed in the north German lowland and can be traced eastward into Poland, Belarus, and western Russia, where yew is absent, thus forming a valuable east-west correlation tool.

Beyond the classic Eemian area, Last Interglacial vegetation successions become, unsurprisingly, modified. Mangerud (1991) considered a transect of sites from northern Germany, mid-Jutland, and western Norway and pointed out that linden and fir dropped out of the succession in Denmark, while in western Norway two additional taxa (yew and hornbeam) were removed. This produces a floristically rather impoverished interglacial succession at Fjøsanger, near Bergen, characterized by pine and birch, an early immigration of oak and some elm, followed by the expansion of hazel and then spruce. Further east, oak and particularly hazel are more strongly represented in diagrams from Ostrobothnia than in the Holocene, while linden is absent. Hornbeam is represented in records from southern Sweden and is also encountered in diagrams from Finland as far north as Ostrobothnia, where it is usually absent in the Holocene. Further north, in Swedish and Finnish Lapland diagrams are dominated by birch or pine with traces of hazel followed by a peak in spruce (see review by Donner (1995)).

In southern Europe, examination of the available data reveals specific regional patterns in the character of the interglacial succession. Thus, during the Last Interglacial, Greek (Tzedakis et al., 2003) and Italian sites (Follieri et al., 1988; Allen et al., 2005) show deciduous oak and elm expanding first, followed by an increase in Mediterranean sclerophylls,

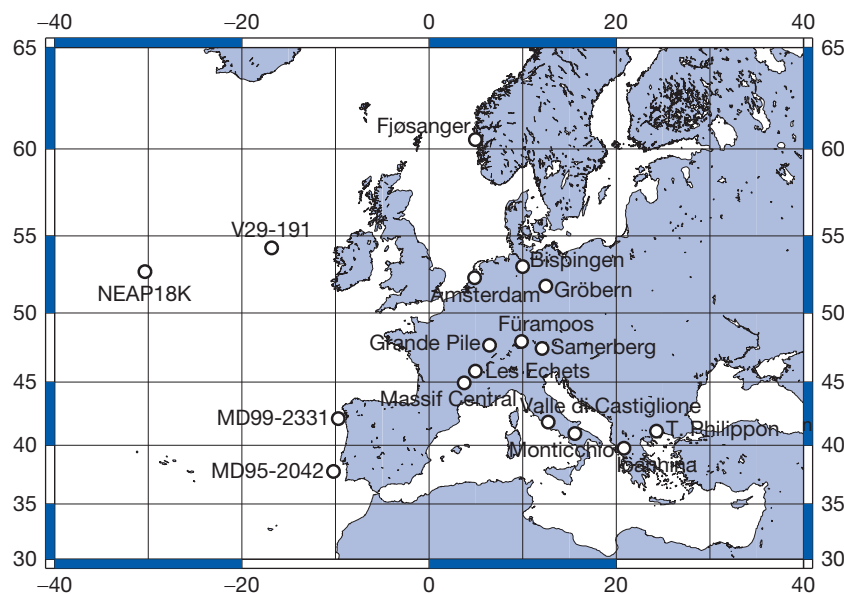


Figure 2 Location of some Last Interglacial sites.

Table 1 Pollen zonation schemes from the classic Eemian area (Zagwijn, 1996; Grüger, 1991). Duration from annually laminated sediments at Bispingen (Müller 1974).

Pollen zones Amsterdam, Amersfoort		Pollen zones Northern Germany		Duration (yr) Bispingen
E7	<i>Pinus</i>	VII	<i>Pinus, Betula</i>	2000?
E6	<i>Picea</i>	VI	<i>Pinus, Picea, Abies</i>	2000?
E5	<i>Carpinus</i>	V	<i>Carpinus, Picea</i>	4000?
E4b	<i>Taxus</i>	IVb	<i>Corylus, Taxus, Tilia</i>	1200?
E4a	<i>Corylus</i>	IVa	<i>Corylus, Quercus</i>	1150
E3b	<i>Quercus–Corylus</i>	III	<i>Pinus, Quercus, Ulmus</i>	450
E3a	<i>Quercus</i>			
E2b	<i>Pinus–Quercus</i>	II	<i>Pinus, Betula, Ulmus</i>	200
E2a	<i>Pinus–Ulmus</i>			
E1	<i>Betula–Pinus</i>	I	<i>Betula, Juniperus, Artemisia</i>	100
Zagwijn (1961)		Grüger (1991)		Müller (1974)

Question marks refer to durations extrapolated on the basis of sediment accumulation rates.

dominated by the largest preanthropogenic peak in *Olea* (olive) in the last 430 ka. About 3 ka into the interglacial, hornbeam and *Ostrya* (hophornbeam) populations expanded; fir then increased and remained dominant until the end of the interglacial at Lago Grande di Monticchio (Allen et al., 2005), while at Ioannina, the final phase of the interglacial shows a re-expansion of deciduous oak (Tzedakis et al., 2003). In contrast to the floristically diverse interglacial successions at wetter sites, records from southernmost Europe and from generally drier locations show a vegetational development that is mostly dominated by deciduous and evergreen oak, other mediterranean sclerophylls, and pine and *Juniperus* (juniper) (e.g., Sánchez Goñi et al. (2005)). In the Massif Central, high-altitude sites (Figure 3) show both central European and Mediterranean affinities (e.g., de Beaulieu and Reille, 1992), thus conveniently linking the stratigraphies north and south of the Alps. Of particular interest are traces of Mediterranean taxa (olive, *Pistacia* (pistachio)) during the hazel zone, a reflection of their large population expansion at lower altitudes.

Duration and Timing of Events

The first important contribution toward an understanding of the duration of the Last Interglacial on land was made by Shackleton (1969) who proposed that the Eemian was not equivalent to the entire MIS 5 of the deep-sea stratigraphy, but rather to an interval within it (MIS 5e). This meant that the Last Interglacial did not have a duration of half an eccentricity cycle (~50 ka), but rather half a precessional cycle

(~10 ka). Astronomical calibration of the marine timescale provided an age of 128 ka for the mid-point of the glacial-to-interglacial transition and 116 ka for the interglacial-to-stadial transition. This appeared to agree with the first Eemian floating chronology from Bispingen, Germany, which indicated a duration of ~11 ka (Müller, 1974). This figure was based on counts of annual laminations spanning part of the Eemian and extrapolation based on sedimentation rates for the remainder. Although the emergent chronology was not anchored by any absolute dates, the good correspondence between the marine and terrestrial estimates for the duration of the Last Interglacial led to the assumption that the lower and upper boundaries of the MIS 5e and the Eemian must have been broadly synchronous.

This view was challenged by Kukla et al. (1997) who tuned the Grande Pile pollen sequence (Woillard, 1978) to the marine stratigraphy of North Atlantic record V29-191 (McManus et al., 1994). This indicated that the terrestrial interglacial lasted 23 ka (130–107 ka) and extended well into MIS 5d until the onset of significant ice rafting in the North Atlantic (event C24 of McManus et al. (1994)). To avoid any circularity arising from aligning terrestrial records to the marine stratigraphy, Shackleton et al. (2002) combined high-resolution foraminiferal isotopic and pollen analyses in a deep-sea core (MD95-2042) off southwest Portugal. The chronology of the sequence was based on inferred sea-level stillstands correlated with radiometrically dated marine coral terraces. This meant that pollen-stratigraphical changes could be directly compared with changes in global ice volume and could also be placed within an absolute chronological framework (Figure 4). The length of the forested interval was 16 ka, with the onset at

Ribains maar (1075 a.s.l.)

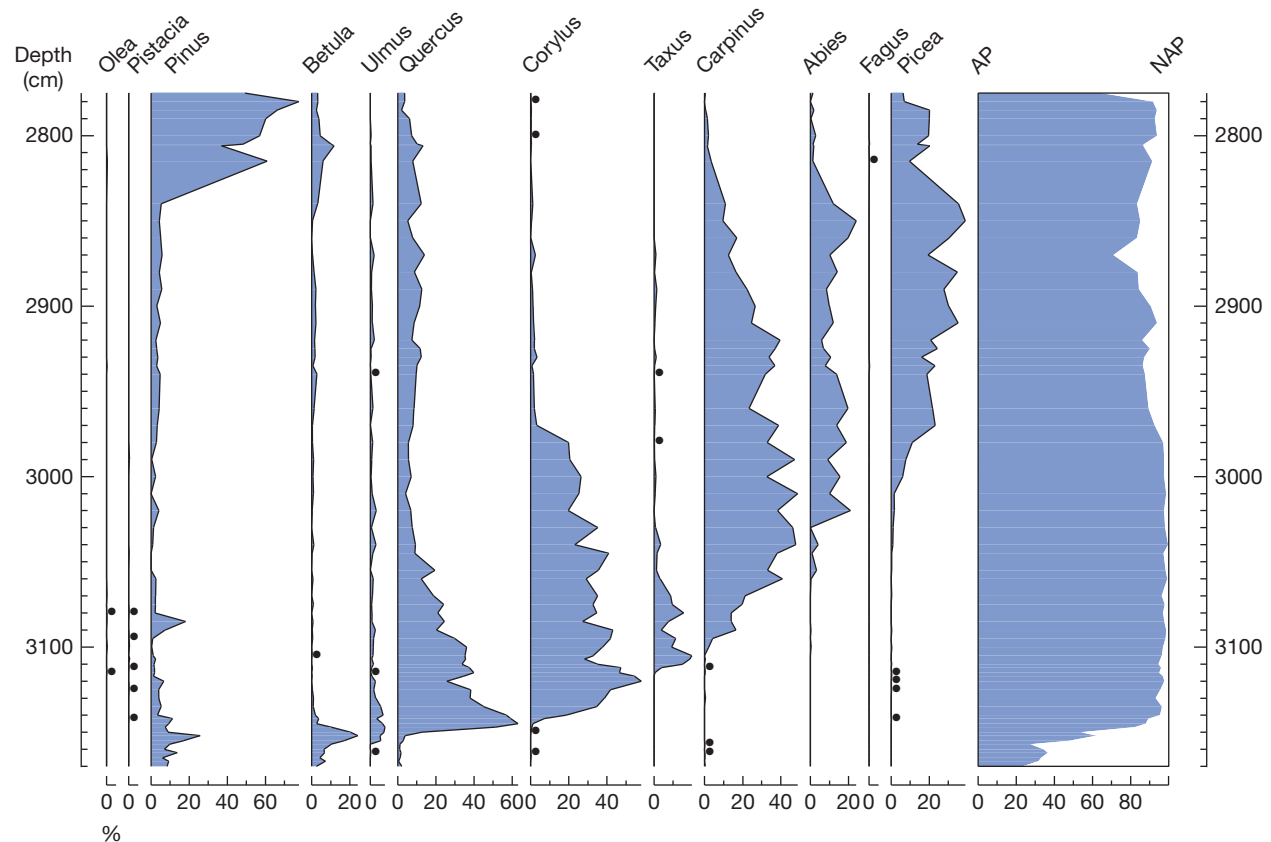


Figure 3 Last Interglacial pollen diagram of selected taxa from Ribains maar, Massif Central (de Beaulieu and Reille, 1992), displaying northern and southern European characteristics. Dots indicate pollen values below 2.5%.

126 ka and the end at 110 ka, while the lower and upper age boundaries for MIS 5e were 132 ka and 115 ka, respectively. Although the actual date for the MIS 6/5e boundary (the mid-point of the deglaciation) may be a matter of some debate, the phase relationship between pollen and benthic $\delta^{18}\text{O}$ records derived from the same sediment sequence is independent of the precise chronology and clearly reveals that neither the lower nor upper boundaries of terrestrial and marine stages are coincident. The onset of interglacial conditions on land (expressed by the full expansion of tree populations) occurred within MIS 5e, well after deglaciation was complete, and it was coincident with a rise to peak sea-surface temperatures (SSTs). This conflicts with the Dutch scheme, where sea level continued to rise during the first 3–4 ka (into the hazel zone) of the terrestrial interglacial (e.g., Zagwijn (1983)). However, the Scandinavian ice sheet reached The Netherlands during MIS 6 and therefore local sea-level history is complicated by glacio-isostatic movements (e.g., Lambeck and Chappell (2001)). At the end of MIS 5e (115 ka), SSTs began a gradual decline, but interglacial conditions persisted on land with tree populations surviving into MIS 5d, albeit at reduced densities. The onset of significant ice rafting in the North Atlantic after 110 ka led to the disruption of the thermohaline circulation and an abrupt change in temperature and precipitation regimes, leading to the elimination of tree populations in southwest Iberia.

On this basis, the timing of events in Europe can be broadly summarized as follows. Southern European sites show evidence of a step-like transition to full interglacial conditions, with an early small increase mainly in oak populations around 129 ka, before a second large expansion of tree populations, defining the onset of the interglacial at ca. 127–126 ka (Figure 5). This is preceded by a Late Glacial vegetation oscillation, coincident with Heinrich Event 11 (HE 11) (Shackleton et al., 2002). However, the magnitude of this oscillation appears to have been smaller than the Younger Dryas during the most recent deglaciation. This means that there was no significant setback to the northward spread of tree populations and can partly explain the rapid vegetation succession that characterizes the early Eemian (as determined in annually laminated sediments). Thus, the degree of diachroneity between southern and northern Europe in the early part of the Last Interglacial appears to have been small (on the order of a few hundred years). However, Müller and Kukla (2004) proposed that north of 50° N, pine and birch continued to dominate landscapes for the first 3–4 ka of the interglacial, and only after the Nordic Seas had warmed sufficiently by 124–123 ka, thermophilous taxa did expand in these areas. The problem with this scheme is that during the early Holocene, radiocarbon dates show that the pine/birch phase did not last more than 1 ka and, if anything, the succession of different zones in the early Eemian is thought to have been even faster. The most parsimonious

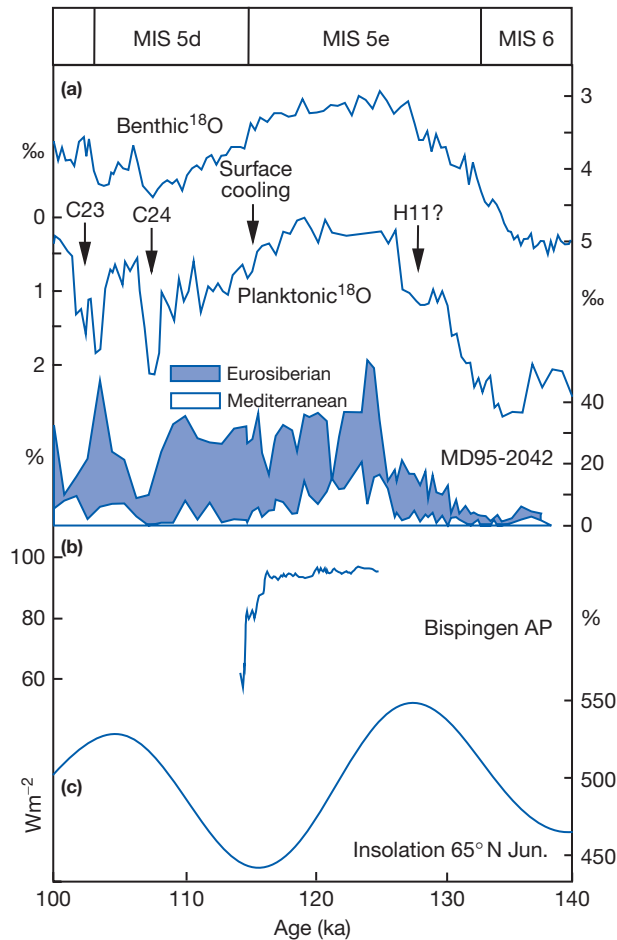


Figure 4 (a) Benthic and planktonic $\delta^{18}O$ data and summary pollen curves from MD95-2042, off southwest Portugal (Shackleton et al., 2002); (b) arboreal pollen (AP) percentages from Bispingen, Germany (Müller, 1974); the onset of the Eemian is here aligned with the terrestrial interglacial in MD95-2042; and (c) June insolation 65° N. (Data from Berger (1978)).

interpretation would be that at ~125 ka southern Europe was characterized by a significant expansion of Mediterranean vegetation (with a high proportion of olive), a reflection of hot summers associated with the insolation maximum, while north of the Alps deciduous forest (mainly hazel) dominated (Figure 1). The hornbeam expansion occurred around 123 ka in the south (Tzedakis et al., 2003; Sánchez-Góñi et al., 2005; Allen et al. 2005) with some degree of S–N diachroneity as shown in the timing of its appearance between Ribains and Grande Pile (de Beaulieu and Reille, 1992). After 120 ka, a reduction in boreal summer insolation led to a decrease in temperatures and an expansion of spruce/fir in northern Europe, followed by taiga after 118 ka. An episode involving a brief reduction in forest cover during the second part of the interglacial ~118 ka is recorded in some southern European sequences (e.g., de Beaulieu and Reille (1992), Shackleton et al. (2002), Tzedakis et al. (2003)). This may be coeval with changes in North Atlantic circulation, but lack of detailed chronological control makes correlations difficult. Studies from annually laminated sites from northern Germany suggest that the elimination of tree populations occurred around

115 ka, near the MIS 5e/5d transition (e.g., Müller (1974)). However, independent chronologies from Portugal (Shackleton et al., 2002), Italy (Allen et al., 2005), and Greece (Tzedakis et al., 2003) suggest that southern European tree populations persisted well into the interval of ice growth until ca. 110–111 ka. Tzedakis (2003) has suggested that this N–S diachroneity may be a reflection of the effects of different bioclimatic parameters, limiting tree growth in the two areas. In the North, tree-population decline is associated with a decrease in the amount of accumulated growing-season warmth as a result of orbital changes after 120 ka. Model simulations suggest that after 120 ka, sea ice and tundra vegetation expanded between 60–90° N (Crucifix and Loutre, 2002) and thus by 115 ka the tundra/taiga boundary may have migrated to ~50° N. Although the orbital changes occurred gradually, the abrupt decline in tree populations at 115 ka suggests a threshold response. Cooling of the Nordic seas may have also contributed to the elimination of forest in northern Germany and Scandinavia by 115 ka (Müller and Kukla, 2004). In southern Europe, however, forest is not limited by the degree of summer warmth, but largely by moisture availability. During the early stages of ice growth (115–110 ka), oceanic heat transport continued to operate over most of the subpolar North Atlantic (e.g., Chapman and Shackleton, 1999), providing a moisture source for tree populations in southern Europe. However, once sufficient ice volume had accumulated to initiate ice rafting in the North Atlantic and disruption of the thermohaline circulation, it led to a reduction in moisture availability and the demise of southern European tree populations. During the closing stages of the interglacial, southern European pollen records show a brief open vegetation event, followed by a re-expansion of tree populations before a larger increase in open vegetation cover indicates the onset of stadial conditions, associated with the occurrence of ice-rafting event C24 (Figure 5).

Some uncertainty still surrounds the length of the forest interval in the 40–50° N band. Did the interglacial there end at ~115 ka, ~110 ka, or somewhere in between? The answer may lie in the stratigraphical thickness of the stadial interval following the end of the Eemian. The Melisey I stadial appears to be stratigraphically very short and is immediately followed by a forest expansion associated with the St. Germain Ia interstadial. Although decreases in sediment accumulation rates may have contributed to that, the same pattern of stratigraphical length replicated across different depositional environments suggests real differences in time span. An argument may, therefore, be made that the end of the Last Interglacial in this region could not have happened at 115 ka, because that would imply that the St. Germain Ia interstadial tree expansion would have occurred before 110 ka, which creates an impossible asymmetry with southern European sequences. Decreases in growing-season warmth and/or cooling of Nordic seas would have a limited impact at this latitude, and therefore taiga could persist until 110 ka. If correct, this would imply a steep vegetation gradient centered around 50° N during the first-half of MIS 5d.

While this view of the timing of changes represents an advance over earlier schemes, some questions remain: (1) Does the onset of the Last Interglacial show a lagged tree population response relative to deglaciation everywhere, or is it geographically dependent? It is interesting to note, for example, that in a combined foraminiferal isotope and pollen study in a deep-sea

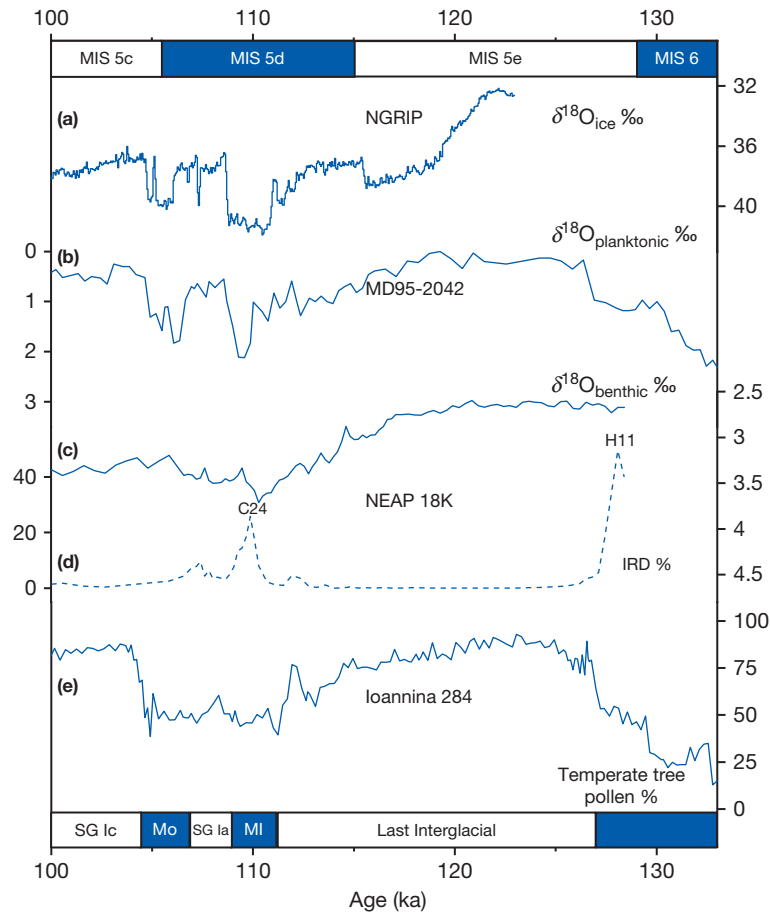


Figure 5 (a) $\delta^{18}\text{O}$ composition of ice in the NGRIP record (North Greenland Ice Core Members (2004). High-resolution record of Northern Hemisphere climate extending into the last interglacial period. *Nature* 431, 147-151). (b) Planktonic $\delta^{18}\text{O}$ data from MD95-2042, off southwest Portugal (Shackleton et al., 2002). (c) $\delta^{18}\text{O}$ composition of benthic Foraminifera (solid line); and (d) relative abundance of ice-rafted detritus (IRD) grains in $>150\ \mu\text{m}$ sediment fraction (dashed line) in core NEAP 18 K in the North Atlantic (Chapman and Shackleton, 1999). Age model used here is based on alignment to the $\delta^{18}\text{O}_{\text{benthic}}$ record of MD95-2042 (Shackleton et al., 2002). (e) Temperate tree pollen percentage from Ioannina I-284; age model is based on a pollen-orbital tuning procedure (Tzedakis et al., 2003). Position of marine isotope and terrestrial stage boundaries shown. 'MI,' 'SG Ia,' 'Mo,' and 'SG Ic' refer to Melisay I, St. Germain Ia, Montaigu, and St. Germain Ic intervals, respectively, as recognized in the Grande Pile pollen record (Woillard, 1978).

core of southeastern USA, the marine and terrestrial stage boundaries coincide at the onset of the Last Interglacial (Heusser and Oppo, 2003). (2) If deglaciation had reached mid-point at $\sim 132\ \text{ka}$ (or earlier) and was complete by ca. 128–127 ka, what could have caused the postulated ice-rafting event (HE 11) at $\sim 128\ \text{ka}$? New evidence from the South China Sea suggests that the demise of ice sheets did not precede boreal summer insolation increase and that the timing of deglaciation is in agreement with Milankovitch theory (Oppo and Sun, 2005). This means that the MIS 6/5e boundary should be placed somewhere around 129 ka. If correct, then the MD95-2042 chronology of Shackleton et al. (2002) may be in need of revision. However, this would shift the onset of the terrestrial interglacial and timing of vegetation events by 3 ka toward younger ages, because the phase lag between the marine and terrestrial boundary is fixed in the same sequence and estimated as $\sim 5\ \text{ka}$. This would imply that the expansion of mediterranean taxa and the olive event took place ca. 124–123 ka, much later than the summer insolation maximum at 127 ka that is thought to have caused it. However, Skinner and Shackleton (2005) showed that changes

in deep-water hydrography in the Portuguese margin during Termination I may lead to apparent differences in the timing of changes in the benthic isotope curve of up to 3–4 ka compared to the equatorial Pacific. By extension, the phase lag of 5 ka for Termination II between marine and terrestrial stages may be overestimated, with most of the early decrease in $\delta^{18}\text{O}_{\text{benthic}}$ values in MD95-2042 attributed to a rise in deep-water temperature and water-mass changes, while the glacio-eustatic sea-level change occurred later. This would reconcile any conflicting dates, producing the following possible timing of events: onset of MIS 5e at $\sim 129\ \text{ka}$, HE 11 at $\sim 128\ \text{ka}$, SST increase and onset of terrestrial interglacial at $\sim 127\ \text{ka}$, expansion of mediterranean taxa at $\sim 126\ \text{ka}$.

Vegetation Character and Comparisons with the Holocene

A striking pattern of the Last Interglacial is the uniform vegetation development across large regions of western and central

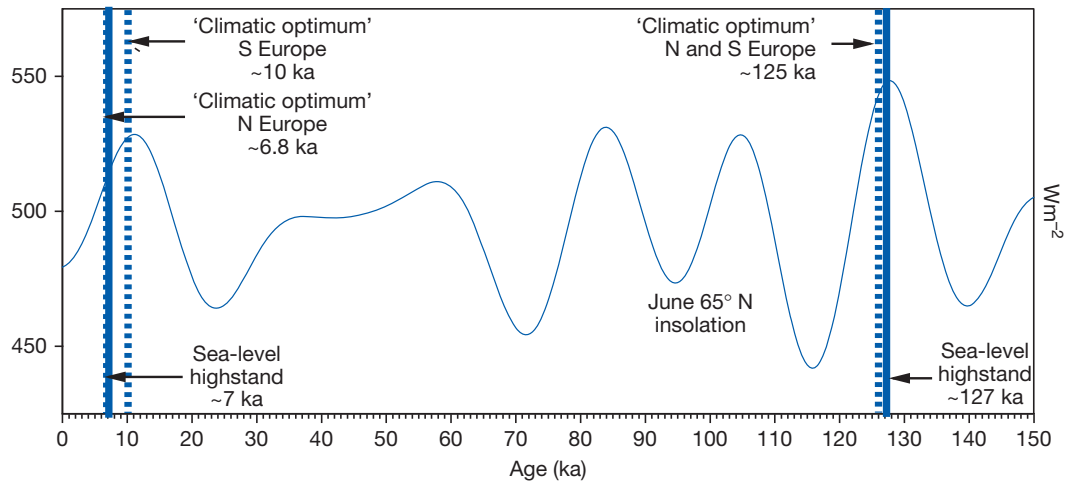


Figure 6 Timing of Holocene and Last Interglacial sea-level highstands and climatic optima in northern and southern Europe relative to the June 65° N insolation curve. From Berger A (1978).

Europe. This similarity becomes even more intriguing when compared to the large degree of variability of the Holocene vegetation succession across the same geographical areas, which in turn raises questions regarding the effects of small-scale anthropogenic disturbance on early Holocene landscapes. In addition, the different distribution of land and sea may have played a role in determining the vegetation character. The presence of the Eemian Sea must have led to milder and more equable climate conditions in the surrounding landmasses. This would have contributed to the establishment of temperate forests and might perhaps go some way in explaining the uniformity of Eemian vegetation over such a large area.

Considering the overall vegetational character of the Last Interglacial across Europe, what emerges is that the climate at the peak of the interglacial was warmer than at any time during the Holocene. This is based on the presence of pollen and macroremains of thermophilous terrestrial and water plants north of their Holocene ranges (e.g., Phillips (1974), Donner (1995)). The question, of course, is when exactly was the interglacial optimum? It has often been assumed that this occurred in the hornbeam zone, mainly on the basis of Holocene analogies. The millennium around 6¹⁴C ka BP (~6.85 cal ka BP) has often been considered as the time of maximum summer warmth in mid- to high latitudes of the Northern Hemisphere (although this applies mainly to Scandinavia, Britain, and the Alps, where it was first defined, and not to southern Europe, where it occurred ~9¹⁴C ka BP (~10 ka)). If the Holocene peak for northern Europe were considered to lie somewhere around 6.85 ka (i.e., ~4.65 ka into the interglacial), then by analogy the same point would lie somewhere in the hornbeam zone of the Last Interglacial. The problem is that the assumption of symmetry in the timing of events between the two interglacials may not be tenable (Figure 6). Sea-level reconstructions based on coral dates (Lambeck and Chappell, 2001) suggest that the Holocene highstand was reached ~7 ka (i.e., 4 ka after the boreal summer insolation peak at 11 ka), while that of the Last Interglacial was reached by ~127 ka

(coeval with the insolation peak). Thus, whereas during the Holocene the presence of ice still continued to influence high-latitude climate and vegetation well into the interglacial, the same did not occur during the Last Interglacial. This means that the optimum of the Last Interglacial occurred earlier (~125 ka) compared to that of the Holocene, closely following the timing of the insolation peak. This would coincide with the olive event in southern Europe and the hazel phase north of the Alps, rather than the hornbeam phase. What emerges is that the timing of the interglacial optimum in southern Europe appears to be more closely associated with the insolation peak and less influenced by the timing of deglaciation. On the other hand, the timing of the optimum in northern Europe varies between interglacials depending on the amount of residual ice volume.

See also: [Paleoclimate Modeling: The Last Interglacial](#). [Plant Macrofossil Methods and Studies: Validation of Pollen Studies](#). [Pollen Methods and Studies: Archaeological Applications](#). [Databases and Their Application](#). [Pollen Records, Late Pleistocene](#): Africa; Australasia; Middle and Late Pleistocene in Southern Europe; Northern Asia; Northern North America; South America; Western North America. [Pollen Records, Postglacial](#): Africa; Australia and New Zealand; Northeastern North America; Northern Asia; Northern Europe; Northwestern North America; South America; Southeastern North America; Southern Europe; Southwestern North America.

References

- Allen JRM and Huntley B and the Monticchio Working Group (2005) The Eemian vegetation and climate record from Lago Grande di Monticchio, southern Italy. In: Sirocko F, Claussen M, McManus J, and Kull C (eds.) *The Climate of the Next Millennium in the Perspective of Abrupt Climate Change During the Late Pleistocene*, DEKLIM/PAGES Conference. Book of Abstracts, pp. 86–87. Germany: Mainz.
- Berger A (1978) Long-term variations of caloric insolation resulting from the earth's orbital elements. *Quaternary Research* 9: 139–167.

- Chapman MR and Shackleton NJ (1999) Global ice-volume fluctuations, North Atlantic ice-rafting events, and deep-ocean circulation changes between 130 and 70ka. *Geology* 27: 795–798.
- Cleveringa P, Meijer T, van Leeuwen RJW, et al. (2000) The Eemian stratotype locality at Amersfoort in the central Netherlands: A re-evaluation of old and new data. *Geologie en Mijnbouw* 79: 197–216.
- Crucifix M and Loutre M-F (2002) Transient simulations over the last interglacial period (126–115kyr BP): Feedback and forcing analysis. *Climate Dynamics* 19: 417–433.
- de Beaulieu J-L and Reille M (1992) Long Pleistocene pollen sequences from the Velay Plateau (Massif Central, France). I. Ribains maar. *Vegetation History and Archaeobotany* 1: 233–242.
- Donner J (1995) *The Quaternary History of Scandinavia*. Cambridge: Cambridge University Press.
- Follieri M, Magri D, and Sadori L (1988) 250,000-year pollen record from Valle di Castiglione (Roma). *Pollen et Spores* 30: 329–356.
- Grüger E (1991) Late Quaternary biostratigraphy in northern Germany 150,000–15,000 years B.P. *Striae* 34: 7–14.
- Harting P (1874) De bodem van het Eemdal. *Verslagen en Mededelingen van de Koninklijke Academie van Wetenschappen afdeling Natuurkunde II* 8: 282–290.
- Heusser L and Oppo D (2003) Millennial- and orbital-scale climate variability in southeastern United States and in the subtropical Atlantic during marine isotope stage 5: Evidence from pollen and isotopes in ODP Site 1059. *Earth and Planetary Science Letters* 214: 483–490.
- Jessen K and Milthers V (1928) Stratigraphical and palaeontological studies of interglacial fresh-water deposits in Jutland and northwest Germany. *Danmarks Geologiske Undersøgelse Raekke II* 48: 1–379.
- Kukla G, McManus JF, Rousseau D-D, and Chuine I (1997) How long and how stable was the last interglacial? *Quaternary Science Reviews* 16: 605–612.
- Kukla GJ, Bender MD, de Beaulieu J-L, et al. (2002) Last Interglacial climates. *Quaternary Research* 58: 2–13.
- Lambeck K and Chappell J (2001) Sea level change through the last glacial cycle. *Science* 292: 679–686.
- Mangerud J (1991) The last interglacial/glacial cycle in northern Europe. In: Shane LCK and Cushing EJ (eds.) *Quaternary Landscapes*, pp. 38–75. Minneapolis: University of Minnesota Press.
- McManus JF, Bond GC, Broecker WS, Johnsen S, Labeyrie L, and Higgins S (1994) High-resolution climate records from the North Atlantic during the last interglacial. *Nature* 371: 326–329.
- Müller H (1974) Pollenanalytische Untersuchungen und Jahresschichtenzählungen an der eem-zeitlichen Kieselgur von Binsingen/Luhe. *Geologisches Jahrbuch A21*: 149–169.
- Müller UC and Kukla GJ (2004) North Atlantic Current and European environments during the declining stage of the last interglacial. *Geology* 32: 1009–1012.
- Oppo DW and Sun Y (2005) Amplitude and timing of sea-surface temperature change in the northern South China Sea: Dynamic link to the East Asian monsoon. *Geology* 33: 785–788.
- Phillips L (1974) Vegetational history of the Ipswichian/Eemian interglacial in Britain and continental Europe. *New Phytologist* 73: 589–604.
- Sánchez Goñi MF, Loutre MF, Crucifix M, et al. (2005) Increasing vegetation and climate gradient in Western Europe over the Last Glacial Inception (122–110ka): Data-model comparison. *Earth and Planetary Science Letters* 231: 111–130.
- Shackleton NJ (1969) The last interglacial in the marine and terrestrial records. *Proceedings of the Royal Society, London B* 174: 135–154.
- Shackleton NJ, Chapman M, Sánchez-Goñi MF, Pailler D, and Lancelot Y (2002) The classic marine isotope substage 5e. *Quaternary Research* 58: 14–16.
- Skinner LC and Shackleton NJ (2005) An Atlantic lead over Pacific deep-water change across termination I: Implications for the application of the marine isotope stage stratigraphy. *Quaternary Science Reviews* 24: 571–580.
- Tzedakis PC (2003) Timing and duration of Last Interglacial conditions in Europe. A chronicle of a changing chronology. *Quaternary Science Reviews* 22: 763–768.
- Tzedakis PC, Frogley MR, and Heaton THE (2003) Last interglacial conditions in southern Europe: Evidence from Ioannina, northwest Greece. *Global and Planetary Change* 36: 157–170.
- van Andel TH and Tzedakis PC (1995) Paleolithic landscapes in Europe and environs, 150,000–25,000 years ago: An overview. *Quaternary Science Reviews* 15: 481–500.
- Woillard GM (1978) Grande Pile peat bog: A continuous pollen record for the last 140,000 years. *Quaternary Research* 9: 1–21.
- Zagwijn W (1961) Vegetation, climate and radiocarbon datings in the Late Pleistocene of the Netherlands, Part I: Eemian and Early Weichselian. *Mededelingen van de Geologische Stichting Nieuwe Series* 14: 15–45.
- Zagwijn WH (1983) Sea-level changes in the Netherlands during the Eemian. *Geologie en Mijnbouw* 62: 437–450.
- Zagwijn WH (1996) An analysis of Eemian climate in western and central Europe. *Quaternary Science Reviews* 15: 451–469.

POLLEN RECORDS, LATE PLEISTOCENE

Contents

Africa

Australasia

Northern Asia

Northern North America

South America

Middle and Late Pleistocene in Southern Europe

Western North America

Africa

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Introduction

Africa is a unique continent with a complex and, indeed, fascinating array of contemporary physical environments associated with prolific biological diversity. Its fauna and flora have been shaped by the combined forces of evolutionary, geological, and climatic change along a trajectory that has become modified, especially so in the relatively recent past, by human activities. Quaternary environmental changes in Africa have been complex, frequent, and of high amplitude, although this is a view of environmental dynamics on the continent that has emerged only in the last few decades. It is important, then, that as accurate a picture of these changes as possible be developed so that we may more fully understand the contemporary status of African biogeography and be able to explore appropriate scenarios of future climate change and its implications. Pollen records, so valuable in reconstructing terrestrial environmental conditions in the Quaternary, have facilitated the compilation of a late Pleistocene vegetation history for Africa, although the temporal and spatial resolution is vastly inferior to that of temperate Europe or North America. As of 2011, the European Pollen Data Base, for example, has more than 1000 pollen records for the late Quaternary, representative of a geographical area of approximately 10 million km². The African Pollen Database (<http://medias.obs-mip.fr/apd/>), on the other hand, has somewhat less than 270 records even though it is the second largest continent on earth and spans an area more than 30 million km².

African pollen records for the late Pleistocene are restricted in number due to a combination of the relatively few studies that have been attempted and environmental conditions that are not always conducive to pollen preservation. This is especially the case in arid and semiarid environments where

organic sediments, traditionally the main source of subfossil pollen, are relatively unusual. Nevertheless, the picture that has emerged of African vegetation at, for example, the Last Glacial Maximum (LGM, ca. 26.5–19 ka; cf. [Clark et al., 2009](#)), is becoming clearer, based on pollen accumulations across a range of different sedimentary environments, including upland bogs and mires, lowland wetlands, lakes, pans, caves, middens, and coastal and marine sediments ([Elena et al., 2000](#)). However, the temporal and spatial coverage of the record remains very incomplete, and there are very large areas of the continent for which remarkably little is known about the development of late Quaternary vegetation patterns. As a result, it becomes necessary, as is the case with the biomization procedure followed by [Elena et al. \(2000\)](#) to interpolate the late Pleistocene paleoecology of much of Africa from the relatively few reliable records that do exist. One significant methodological advance that has emerged recently has been the quantification of the relationships between pollen, plants, and climate at the continental scale employing the African Pollen Database (e.g., [Watrin et al., 2007](#)). The modeling of plant distribution data, pollen percentages, and a climate response surface is used to identify plant taxa that illustrate a high degree of correspondence with climate at the regional scale and, in so doing, take a step toward more reliable vegetation and climate reconstructions for the late Quaternary.

Spanning latitudes 30 degrees north and south of the equator, the continent of Africa classically illustrates how the interactions between underlying geology, soils and geomorphology, and climate have combined over time to produce the contemporary distribution of its major biomes. Broadly speaking, these are identified as several major biomes ([Figure 1](#)), namely, tropical rainforest, savanna woodlands and grasslands, arid and semiarid formations, shrublands under a Mediterranean-type

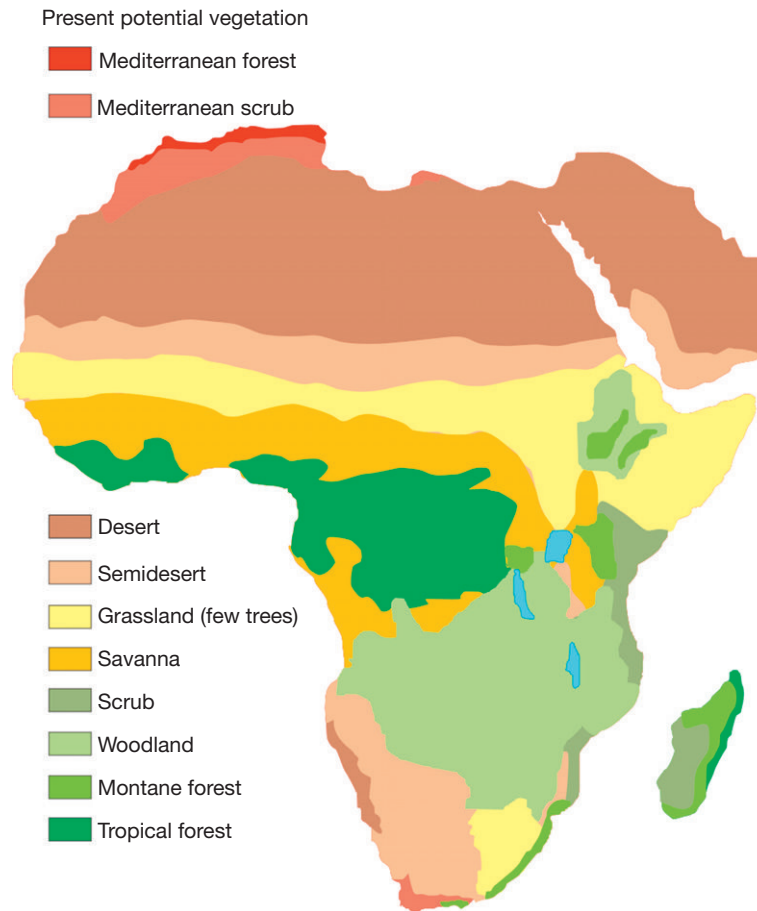


Figure 1 The present potential vegetation of Africa. Adapted from Adams JM and Faure H (1997) (eds.) *Review and Atlas of Palaeovegetation: Preliminary Land Ecosystem Maps of the World since the Last Glacial Maximum*. Oakridge, TN: Oak Ridge National Laboratory (<http://www.esd.ornl.gov/projects/qen/adams1.html>).

climate, and upland or montane regions (Meadows, 1996). This general biogeographical framework is used here as a basis for the description and explanation of late Pleistocene pollen records of Africa. Rather than attempting to describe the entire late Pleistocene, the focus is on the LGM, taken to be coincident with the time when global ice volumes were at their maximum values and sea levels at minima, that is, between 26.5 and 19 ka (Clark et al., 2009), and representative of a time during which global temperatures were significantly lower than today. A summary of the LGM vegetation distribution is shown in Figure 2, revised and adapted from Adams and Faure (1997).

Tropical Rainforests

Given the importance of tropical rainforests in the context of biodiversity and the function of major global biogeochemical cycles, it is frustrating that there are so few pollen records for the late Pleistocene in Africa. Indeed, the first published pollen records for the equatorial forests of Africa were produced little more than two decades ago for two lakes in West Africa (Maley, 1991). There remain rather few terrestrial sites for which there are vegetation histories spanning the LGM from tropical

lowland forests of Africa, although the Holocene is better represented (see Lezine, this volume). Nevertheless, the pollen sequence from Lake Barombi Mbo, Cameroon, has now been more precisely interpreted (Lebamba et al., 2010) and provides an opportunity to glimpse this crucial period of vegetation change from the perspective of the largest area of tropical rainforest outside the Amazon basin.

The mechanisms underlying tropical forest diversity are not definitively known, but the original idea that long-term environmental stability was a principal factor has long since been laid to rest (Meadows, 1999). The replacement of this idea came with the realization, spawned by paleoecological studies, that the tropics – far from being immutable and static over extended periods of geological time – had been subject to major climatic fluctuations during the Quaternary. An alternative viewpoint arose that interpreted areas (habitat islands) of elevated diversity and levels of endemism within the tropics as the Pleistocene refugia of rainforest taxa (Connor, 1986). According to this view, these areas represented places of refuge at moderate altitude, where plant and animal species adapted to the hot and humid climates of lowland equatorial regions survived during the emerging cooler and drier climates of the glacials, when the savannas became much more widespread.

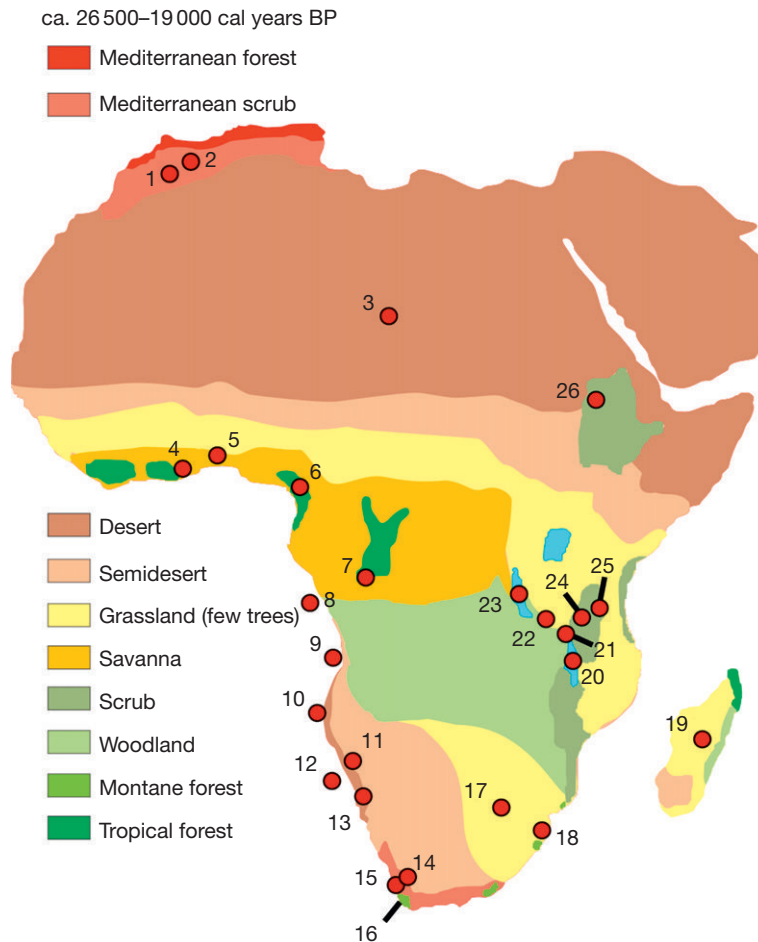


Figure 2 LGM vegetation of Africa. Localities mentioned in the text: (1) Tigalmamine, (2) Lake Ifrah, (3) Tibesti, (4) Lake Bosumtwi, (5) Dahomey Gap, (6) Lake Barombi Mbo, (7) Ngamakala, (8) Congo River fan, (9) ODP 1078C, (10) GeoB 1023–5, (11) Brandberg, (12) GeoB 1711–4, (13) Sossusvlei, (14) Pakhuis (15) Elands Bay, (16) Rietvlei, (17) Tswaing Crater, (18) Mfabeni Peatland, (19) Lake Tritrivakely, (20) Lake Malawi, (21) Lake Masoko, (22) Lake Rukwa, (23) Lake Tanganyika, (24) Dama Swamp, Iringa, (25) Deva-Deva, and (26) Lake Tana. Adapted from Adams JM and Faure H (1997) (eds.) *Review and Atlas of Palaeovegetation: Preliminary Land Ecosystem Maps of the World since the Last Glacial Maximum*. Oakridge, TN: Oak Ridge National Laboratory (<http://www.esd.ornl.gov/projects/gen/adams1.html>).

This view was not without its critics (Connor, 1986), and interpretation of paleoecological evidence from the equatorial region, most especially pollen analysis, led to a reassessment of the elegantly simple refugium hypothesis. The situation is far more complex than earlier viewpoints would have suggested and, certainly for South America, a simplistic view of the refugium hypothesis cannot be upheld (Bush, 2005).

A comprehensive biogeographical understanding of the Quaternary history of tropical lowland forests of Africa has proved elusive. Cooler temperatures, on the order of 4–5 °C lower than present, are uniformly indicated in the pollen records for the LGM. There appears to be considerable expansion of grassland and tropical seasonal forest at the expense of tropical rainforest in the cold phases, indicative of drier conditions. The so-called Dahomey Gap in West Africa was at its most extensive during these times (Dupont and Weinelt, 1996), although rainforest does appear to have survived in certain localities (Dupont et al., 2000; Jahns et al., 1998). Receiving annual rainfall today in excess of 2300 mm, the savannas in the vicinity of Lake Barombi Mbo are themselves

regarded as relict, left over from a time (the LGM) when cooler temperatures and greater aridity were thought to have forced the forests into large remnant patches (Figure 3; Maley and Brénac, 1998). Maley and Brénac (1998) also identify ‘sylvigene cycles’ in the vegetation history, defined as repeated successional changes on an approximately 2000-year recurrence interval which may coincide with the Dansgaard–Oeschger cycles observed in the ice sheet record. The record at Barombi Mbo has recently been subject to considerable further analysis employing a numerical biomization method that reveals more detail about the vegetation and climate changes (Lebamba et al., 2010). This method indicates that open savanna formations were never extensive in the basin and that patches of grassland and deciduous woodland, at their most developed during Heinrich event 1, were more likely enclosed within the forest in the form of a mosaic.

Meanwhile, at Lake Bosumtwi in Ghana, a prolonged lake low stand does appear to coincide with the disappearance of rainforest taxa and their replacement with vegetation with greater montane affinities (Maley, 1991) around the time of

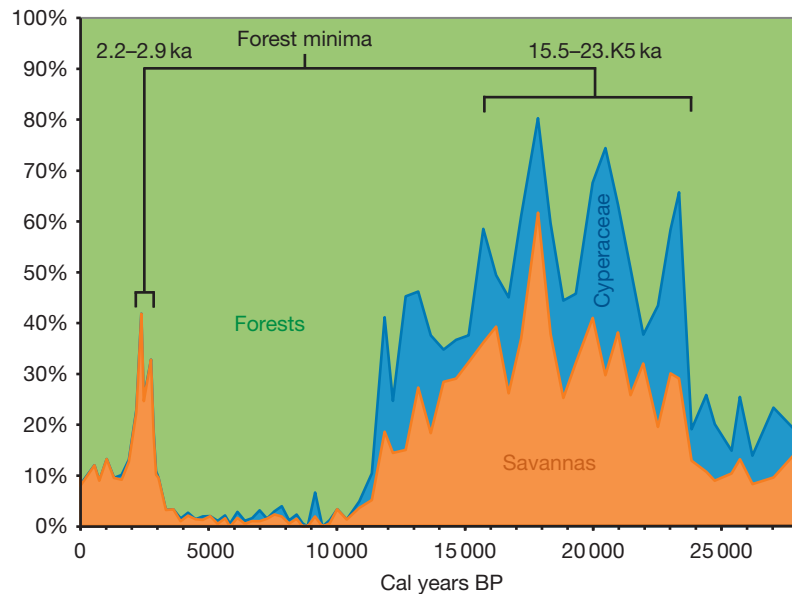


Figure 3 Vegetation changes in the African tropical forest biome as indicated from pollen analysis at Lake Barombi Mbo, Cameroon (courtesy of J Maley).

the LGM. This is supplemented by geochemical and geomorphological evidence that further suggests a return to drier conditions at the time of the Northern Hemisphere Younger Dryas oscillation (Shanahan et al., 2006, 2008). Only marginally drier LGM conditions are recorded at Ngamakala in the Congo basin (Elenga et al., 1994), and here, rainforest taxa remain very prominent in the sequence. Marret et al. (2001) note that these drier conditions may have ended abruptly by 13 000 cal years BP because pollen and other evidence from the offshore sediments of the Congo River fan suggest a dramatic increase in discharge at this time. The importance of these various observations is twofold. First, vegetation changes, as indicated by the pollen record, are more complex than can be explained by the simple retreat of rainforest in response to a drier and cooler LGM. Second, on the basis of the available paleoenvironmental proxies, there is conflicting evidence as to the status of rainforest around the time of the LGM; it may have been entirely replaced by savanna in some localities but appears to have survived, albeit in variable forms, in others. There is evidence from ocean sediments off Gabon, western equatorial Africa, of a weakening of the monsoon system brought about by lower sea surface temperatures in the tropical Atlantic (Kim et al., 2010). Anhuf et al. (2006) argue that tropical forest in Africa was reduced in extent by an estimated 84% in response to increased aridity. Accordingly, unlike in South America, the refugium hypothesis still cannot be rejected as a model of vegetation change in the late Quaternary for the African lowland tropics, even if it does not offer a comprehensive explanation of patterns of diversity and endemism.

The Afromontane Region

The cooler conditions at higher altitudes have proved more conducive to the accumulation of pollen-rich organic sediments and, indeed, provided the first Quaternary paleoecological

reconstructions in tropical Africa (Coetzee, 1967). Nevertheless, the dynamics of the afromontane flora and, in particular, the occurrence of glacial aridity and the existence of Quaternary lowland forest refugia in the mountains have been the subject of considerable debate. Significant evidence in this regard has recently emerged from sites in Central East Africa (Jolly et al., 1997; Marchant and Taylor, 1998; Marchant et al., 1997; Rucina et al., 2009; Street-Perrott et al., 2007) and also from Lake Tana in Ethiopia (Lamb et al., 2007).

The review by Jolly et al. (1997) remains significant because it synthesizes the late Quaternary paleoenvironmental implications of several sites within a relatively small geographical area and, in so doing, offers a more reliable reconstruction than the usual single-site pollen analyses. Although the number of published paleoecological studies in the afromontane region as a whole has increased, many biogeographical questions remain unanswered. For example, did the region act as a glacial refugium for lowland taxa and did the forests of tropical Africa respond to change as homogeneous units, or were there differences in response times resulting in the development of unique and non-analog communities? The so-called inter-lacustrine highlands of Central Africa offer the possibility of addressing such issues, and Jolly and coworkers have successfully explored these issues using six radiocarbon-dated pollen sequences drawn from five sites in Uganda, Burundi, and Rwanda, ranging in age from around 40 000 cal years BP to 'modern.' The LGM emerges as a period characterized by relatively xerophytic ericaceous scrub and grassland that, in at least one instance, is suggestive of a treeless landscape. The authors make no direct paleoclimatic reconstructions, but clearly, such conditions are consistent with a cooler and drier climate at the time. As for a refuge for lowland taxa, the pollen data neither support nor refute the existence of the major East-Central Africa refugium proposed by Colyn et al. (1991).

Two sites in Tanzania – Dama Swamp, Iringa (Mumbi et al., 2008), and Deva-Deva (Figure 4; Finch et al., 2009) from



Figure 4 Deva-Deva swamp in the Uluguru Mountains of Tanzania, a late Pleistocene record from within a biodiversity hotspot (photo courtesy of J Finch).

within the biodiversity hotspot of the Eastern Arc mountains – also help cast light on the validity of the refugium hypothesis. The data suggest relative environmental stability in this region over the LGM interval, although there are indications of some changes in forest composition, perhaps in response to marginally drier climates.

Based on evidence from a site on Mount Kenya, [Rucina et al. \(2009\)](#) suggest that the LGM may have actually been characterized by somewhat greater moisture availability than today but that the vegetation assemblage was a hybrid of higher and lower altitude elements. The indication is that, rather than a simplistic shift of vegetation belts to higher and lower elevations in response to late Quaternary climate change, communities dynamically adjusted, assembling and reassembling as individual species migrated. It seems that vegetation belts did not shift elevations as intact units in response to changes in conditions but rather new associations formed as species reacted individually to variations in climate. Other wetlands on Mount Kenya have provided valuable pollen and associated multiproxy evidence of change in the afro-montane region ([Street-Perrott et al., 2007](#)) in response to precipitation and temperature changes. At Lake Tana in north-west Ethiopia, paleoecological and other evidence suggest that the lake dried out around the time of Heinrich event 1 (post-18 700 cal years BP to ca. 15 100 cal years BP; [Lamb et al., 2007](#)). Certainly, changes in montane forest composition reflecting adjustments to climate variation (and, no doubt, atmospheric chemistry) are also apparent throughout the later Pleistocene in many of the afro-montane sequences considered.

Savanna Woodlands and Grasslands

The savanna biome of Africa, the most spatially extensive complex of vegetation formations on the modern continent, is also significantly understudied from the perspective of the late Pleistocene pollen record. While climate, in concert with fire, is a major driver of vegetation in this biome, the involvement of humans in shaping the vegetation pattern over extended periods also needs to be considered ([Meadows, 1996](#)). Longer sedimentary sequences have been retrieved

from wetland sites in Zambia ([Livingstone, 1971](#)) and South Africa (Wonderkrater, Equus Cave, the Eastern Free State, and Pretoria Salt Pan: [Scott, 1982, 1987, 1989, 1999; Scott et al., 2008](#), respectively) and also from Madagascar ([Gasse and van Campo, 1998](#)). These are supplemented by long lacustrine sequences from Lake Tanganyika ([Tierney et al., 2010; Vincens, 1991, 1993](#)), Lake Albert ([Beuning et al., 1997](#)), Lake Masoko ([Vincens et al., 2007](#)), Lake Rukwa ([Vincens et al., 2005](#)), and Lake Malawi ([Beuning et al., 2010; DeBusk, 1998](#)). The simplistic view of a cooler and drier LGM becomes even more difficult to uphold in light of these records, although [Gasse et al. \(2008\)](#) have recently comprehensively reviewed palaeoenvironmental evidence across the African tropics and concluded that the LGM was indeed ‘generally’ drier. Certainly, and uncontroversially, cooler LGM conditions are widely apparent in the records, but the moisture signal appears to be complex, both spatially and temporally. Evidence of greater aridity at the time of the LGM is definitively indicated at several localities, including at least some of the Rift Valley lakes. Pollen evidence from two cores from the southern part of Lake Tanganyika ([Vincens, 1991, 1993](#)) reveals changes in vegetation patterns related to climatic fluctuations from around 27 000 cal years BP onward. The LGM here is characterized by greater abundance of afro-montane vegetation and a correspondingly lower percent of pollen, indicative of humid conditions ([Vincens, 1991, 1993](#)). Around this time, open and poorly diversified savanna woodland and C_4 grassland prevailed at low and mid-altitudes, with local patches of montane forest including *Podocarpus* spp. indicating cooler and drier climatic conditions than today, with frost during the night at low altitude. Pollen data are supplemented by the molecular analyses of [Tierney et al. \(2010\)](#) who note that greater LGM aridity may further have been associated with a dual (i.e., two short rainy seasons) annual precipitation regime but that the role of lowered CO_2 concentrations in favoring C_4 grasslands also needs to be considered. For the catchments of Lake Malawi and Lake Masoko, on the other hand, the pollen and sedimentological records suggest that, following warmer and more arid conditions from around 26 000–34 000 cal years BP, the LGM was characterized by increased pollen frequencies of montane elements and the aridity that typified this period elsewhere in the tropics did not prevail here ([DeBusk, 1998; Garcin et al., 2006a,b](#)). Indeed, DeBusk goes as far as to suggest that the relatively cool and moist late Pleistocene conditions may have facilitated the interchange of biota between the highlands of east and west Africa along the uplands of the Zaire–Zambezi watershed. These conclusions are to some extent supported by those based on more recently cored sediments in Lake Malawi, where [Beuning et al. \(2010\)](#) suggest relatively minor LGM impacts on vegetation by comparison with other periods earlier in the latest glacial that arose out of what they term ‘megadroughts.’

The savannas of southern Africa do appear to have been subject to lower moisture availability around the time of the LGM. [Scott et al. \(2008\)](#) review late Quaternary environments in this region. For example, around the Tswaing Crater ([Figure 5](#)), [Scott \(1999\)](#) documents cooler and drier conditions associated with the lowering of vegetation belts by up to 1000 m, as well as changes in the relative abundance of C_3 and C_4 grasses ([Scott, 2002](#)), again likely in response to lower carbon dioxide concentrations. [Finch and Hill \(2008\)](#) infer



Figure 5 The Tswaing Crater, northeast of Pretoria, South Africa, and the locality of one of the longest Quaternary terrestrial sediment sequences in Africa. Photo courtesy of the Hartbeespoort Radio Observatory (www.hartrao.ac.za/other/tswaing/tswaing.html).

greater aridity around 21 ka at the Mfabeni Peatland, and coastal forests are thought to have retreated along the Maputaland plain around that time. The first real indications of Angolan vegetation response to late Quaternary climate dynamics have recently become available from a marine core off the west coast of Southern Africa (ODP 1078, Dupont and Behling, 2006; Dupont et al., 2008) suggesting that much more open vegetation prevailed during much of the glacial in response to cooler and drier conditions.

Off the east coast of the continent in Madagascar at Lake Tritrivalakely, Gasse, and van Campo (1998) outline a cool and dry LGM, although these conditions appear to have terminated abruptly at around 17 000 cal years BP with significantly increased moisture availability, a change that may have preceded the equivalent in the Northern Hemisphere by as much as 2000 years.

In summary, late Pleistocene temperature signals in the savannas of Africa are relatively wellresolved, but the precipitation and evaporation trends have proved much more difficult to reconstruct. The problem of reconstructing rainfall amounts and seasonality is no doubt exacerbated by the effect of reduced atmospheric carbon dioxide concentrations, especially in this biome, since woodland and grassland dynamics appear to be particularly sensitive to such changes (Bond et al., 2003) and there are clear advantages for C_4 grasses under lower partial pressures of CO_2 (Ehleringer et al., 1997).

Arid and Semiarid Biomes

Quaternary pollen analysis was developed initially in temperate latitudes, and the sedimentary environments in which pollen typically preserves over time are associated with reducing conditions in permanent or at least seasonal wetlands or in lacustrine sediments. Obviously, such habitats are rare in the arid and semiarid biomes of Africa and palynologists have been forced to seek alternative archives. Even so, the number of records, particularly those extending into the late Pleistocene, is very limited and Elenga et al. (2000) cite only Sossusvlei, in Namibia, and Tibesti, Chad, as sites for which there are LGM records. Both have been interpreted as indicating environments somewhat cooler than the present day with even greater aridity. Subsequent sedimentological studies, however,

from both Namibia (Stuut et al., 2002) and Tibesti (Maley, 2000), indicate that even considering the effect of reduced potential evaporation, these regions may have experienced increased precipitation during the LGM. In Namibia, additional insight can be obtained from pollen assemblages derived from marine records that have been obtained off the southwest African coast (Shi et al., 2000, 2001). Marine core GeoB 1023–5 is from a site off the Cunene River mouth at around latitude 17° S and yields information on the dynamics of northern Namib Desert vegetation from around 21 000 cal years BP onward (Shi et al., 2000). The pollen spectra from this core not only indicate the prominence of desert vegetation but also show increases in semiarid scrub, exemplified by higher *Stoebe*-type pollen frequencies as well as the presence of certain elements of the fynbos flora (e.g., *Restionaceae*, *Ericaceae*) that are associated with wetter conditions and increased winter rainfall which presently prevails further south. The presence of *Stoebe*-type pollen in the region, rather than being the result of long distance transport by stronger winds (Stuut et al., 2002), is corroborated by pollen recovered from hyrax middens in the Brandberg Massif (21° S), which show high levels of this pollen in samples dated to 21 100 and 35 000 cal years BP (Scott et al., 2004). A longer temporal perspective, up to 135 000 years, is provided by another core in the region, GeoB-1711-4 off the central Namibian coast (23° S) (Shi et al., 2001). This record indicates several periods in the late Pleistocene during which vegetation changes along the western margin of Southern Africa conform to a general pattern of increased glacial humidity and interglacial aridification; this pattern has been attributed to equatorward migrations of the westerly storm tracks as a response to increased Antarctic sea-ice extents (cf. Chase and Meadows, 2007; Stuut and Lamy, 2004; Stuut et al., 2004; van Zinderen Bakker, 1976). Dupont and Behling (2006) provide supplementary evidence for terrestrial vegetation change in semiarid southwestern Africa from ODP 1078 that reveals a cold and dry LGM was succeeded by cool and wet conditions during Heinrich event 1.

Mediterranean-Climate Regions

The distinctive climate of regions with warm, dry summers alternating with cool, wet winters is characteristic of the Mediterranean regions of North Africa and of the southwestern part of the continent and is associated with species-rich vegetation formations dominated by evergreen sclerophyllous shrubs. Virtually, nothing is known about the late Pleistocene development of this vegetation type in North Africa. Elenga et al. (2000) document no pollen records for the LGM in this region, but more recent work has cast some light on the situation. In Morocco, Lamb et al. (1989) do provide a pollen record that extends back to around the time of, or at least shortly after, the LGM from Lake Tigalmamine in the Middle Atlas Mountains, and there is a more recent and longer core from Lake Ifrah (Rhouijati et al., 2010). The pollen spectra in basal samples of the 21-m Tigalmamine core are very strongly dominated by grasses, chenopods, and *Artemisia*, resulting in what is referred to as steppe vegetation under climatic conditions that, on this basis, appear to have been too cold and arid to support tree growth at the time. At Lake Ifrah, glacial climates

(29 ka onward) appear to have been colder and generally drier than today with the most intense aridity around 19.5–12 ka (Rhoujjati et al., 2010). Nevertheless, there is evidence for periods of greater moisture availability in the Lake Ifrah record and, indeed, some tree taxa, specifically *Cedrus*, appear to have survived in scattered refugia in the region during glacial times (Cheddadi et al., 2009). The paleoecological evidence thus favors the existence of drier as well as cooler climates for the glacial. Tzedakis (2007) suggests that the contradictory pollen and lake level records for the region during the coldest parts of the last glacial may be reconcilable. Either there was indeed less rainfall associated with increased cloudiness and reduced evapotranspiration or there were increased seasonal contrasts, with colder and wetter winters (more snow) and a more pronounced summer drought. In regard to the Late Glacial, the invasion of evergreen oak (*Quercus* spp.) onto the steppes around Lake Tigalmamine signifies that climate warming was underway by 14 ka and coincided with lower lake levels, and the record at Lake Ifrah is consistent with this reconstruction (Rhoujjati et al., 2010). However, the open, treeless steppe vegetation again dominated around the time of the Younger Dryas (Lamb et al., 1989), and conditions were never sufficiently warm and moist for woodland to prevail around this site until well into the Holocene.

At the other end of the continent, in the southwestern Cape, the late Pleistocene pollen record has been supplemented with records obtained from hyrax (*Procapra capensis*) middens. These unusual archives are revealing themselves to be remarkably sensitive paleoenvironmental archives (Chase et al., 2011; Meadows et al., 2010). The potential of these deposits was first recognized in the 1990s when Scott (1994) – and subsequently Scott and Vogel (2000) – explored pollen sequences in hyrax middens, revealing vegetation changes in terms of both species composition and the relative importance of C_4 , C_3 , and CAM plants from around 20 000 cal years BP onward at Pakhuis in the Cederberg mountains at around 32° S. That work has since been supplemented by several more detailed palynological and stable isotope studies on middens (Chase et al., 2011; Meadows et al., 2010; Scott and Woodborne, 2007a,b) that reveal late Pleistocene environmental dynamics in this region. Changes in plant species composition suggest a lowering of vegetation belts around the time of the LGM, with increased proportions of upland taxa in the record in response to lower temperatures and what is interpreted to be somewhat greater aridity. The stable carbon isotope signals for this interval show no enrichment, as would be expected during a period of lower atmospheric carbon dioxide, and indicate that winter rainfall persisted in the region throughout the interval in question and in fact may have been more effective than in the present day. That such winter rainfall may have been more effective, through a combination of increased annual amounts and a reduction in evaporation rates, is supported by pollen and charcoal data from Elands Bay Cave, a coastal site less than 100 km west of the Cederberg, where there are substantially higher frequencies of afro-montane vegetation indicators than occur today during at least some intervals of the last glacial (Chase and Meadows, 2007; Parkington et al., 2000). The highest resolution record of last glacial–interglacial transition environments obtained thus far comes from a large hyrax midden in the Cederberg (Chase et al., 2011) where a

stable isotope signal of available moisture provides unequivocal evidence of the occurrence of a well-defined, cold, and dry period coincident with the Younger Dryas and pointing to the influence of Northern Hemisphere drivers of climate even as far south as 32° S after Heinrich event 1.

Further south, in the Cape Peninsula region, several pollen cores have been recovered from Rietvlei and the Cape Flats (Schalke, 1973). Pollen from these cores shows significant differences between what has been interpreted as Holocene deposits and those dating to the last glacial. Although many of the ages from older deposits are, arguably, infinite, ranging from $40\,100 \pm 3000$ – $47\,960 \pm 2500$ ^{14}C years BP, adjacent sediments with similar pollen spectra date to $32\,770 \pm 1000$ – $34\,660 \pm 900$ ^{14}C years BP ($\sim 37\,500$ – $39\,500$ cal years BP). Pollen spectra from these deposits are very similar to spectra from modern wetlands adjacent to the humid forests of the south coast, suggesting significantly more effective and less seasonal precipitation during periods of the last glacial cycle.

Higher rainfall for the winter rainfall zone of South Africa appears to be out of phase with vegetation indicators elsewhere in southern Africa at this time and is especially interesting given the implications of lower atmospheric carbon dioxide levels which should reduce plant available moisture.

Conclusions

The paucity of the pollen record for the late Pleistocene of Africa certainly hampers the reliability, if not the validity, of any summary conclusions regarding the nature of vegetation change and its causes. As Hessler et al. (2010) state, there is clearly a need for more records, especially from the northern equatorial tropics of Africa where there are “...no records at all” (p. 2895). Not surprisingly, the strongest paleoenvironmental signal emanating from pollen sequences of the late Pleistocene is that of lower temperatures, and coupled with strong evidence for increased aridity in many biomes, this had the effect of limiting biomass through reduction in forest and woodland taxa. Thus, the tropical rainforests of West and Central Africa were reduced in extent through encroachment of montane vegetation shifting in altitude and of savanna and grassland vegetation shifting in latitude. The expansion of the deserts is less obvious in the pollen record, but this is perhaps a consequence of gaps in the data rather than any resilience on the part of the vegetation to climate change. The extent to which the vegetation changes were a result of climate change, that is, temperature and precipitation reduction, remains a key and, as yet, largely unresolved issue. The complicating issue of atmospheric chemistry changes remains unresolved, for example, Jolly and Haxeltine (1997) use a process-based vegetation model to argue that substantial variation in the vegetation of the East African mountains is possible as a consequence of lowering atmospheric carbon dioxide concentrations and that estimates of LGM cooling (by 4 °C) and aridity (30% lower precipitation) in this biome may both have been overestimated.

Nevertheless, the vegetation changes inferred from pollen analysis are both substantial and in line with major global changes in glaciation and insolation regimes as shown by Jolly et al. (1998). It appears that the same forcing mechanisms that have defined global climate change throughout the

Quaternary (e.g., Milankovitch cycles of orbital variation, polar ice development, oceanic circulation patterns) have been the key drivers behind the described patterns of vegetation change. Across the continent, however, these forcing mechanisms have played a varying role and have had various levels of impact as a function of their particular dynamics and sphere of influence. It is, therefore, too simplistic to argue that glacial and interglacial cycles correspond to cycles of aridity and humidity, respectively. In tropical monsoon-influenced regions, decreased temperatures and a less-active global hydrological system during glacial periods have apparently resulted generally in increased aridity. In temperate regions, however, some evidence does exist that increased hemispheric temperature gradients and polar ice extents have resulted in invigorated westerly systems, and increases in precipitation in the lower mid-latitudes and perhaps even in the subtropics. Each of these cycles operates at a different frequency, with the interplay set upon the backdrop of orbital and ice-forced 100-ka cycles of depressed temperatures and reduced atmospheric CO₂. The amalgam is not easy to decipher.

The studies presented here define a stage in an ongoing process to better understand the relationship between changing African climates and ecology. This knowledge is especially critical when considering our attempts to understand the future effects of potential anthropogenic impacts on climate and the global system as a whole. If management and mitigation of human influence are to be best directed, it is vital that we develop an understanding of the natural variation of the systems being considered, and a sense of causality, so that the direction of change can be best predicted.

See also: Pollen Analysis, Principles. **Lake Level Studies:** Africa during the Late Quaternary. **Paleoclimate Modeling:** Last Glacial Maximum GCMs. **Pollen Methods and Studies:** Databases and Their Application; Reconstructing Past Biodiversity Development; The Biome Approach to Reconstructing Past Vegetation; Use of Pollen as Climate Proxies. **Pollen Records, Postglacial:** Africa. **Vertebrate Studies:** Speciation and Evolutionary Trends in Quaternary Vertebrates.

References

- Adams JM and Faure H (eds.) (1997) *Review and Atlas of Palaeovegetation: Preliminary Land Ecosystem Maps of the World since the Last Glacial Maximum*. Oakridge, TN: Oak Ridge National Laboratory <http://www.esd.ornl.gov/projects/qen/adams1.html>.
- Anhuf D, Ledru M-P, Behling H, et al. (2006) Paleo-environmental change in Amazonian and African rainforest during the LGM. *Palaeogeography, Palaeoclimatology, Palaeoecology* 239: 510–527.
- Beuning KRM, Talbot MR, and Kelts K (1997) A revised 30,000-year paleoclimatic and paleohydrologic history of Lake Albert, East Africa. *Palaeogeography, Palaeoclimatology, Palaeoecology* 136: 259–279.
- Beuning KRM, Zimmerman KR, Ivory SJ, and Cohen AS (2010) Vegetation response to glacial–interglacial climate variability near Lake Malawi in the southern African tropics. *Palaeogeography, Palaeoclimatology, Palaeoecology* 303: 81–92.
- Bond WJ, Midgley GF, and Woodward FI (2003) The importance of low atmospheric CO₂ and fire in promoting the spread of grasslands and savannas. *Global Change Biology* 9: 973–982.
- Bush MB (2005) Of orogeny, precipitation, precession and parrots. *Journal of Biogeography* 32: 1301–1302.
- Chase BM and Meadows ME (2007) Late Quaternary dynamics of southern Africa's winter rainfall zone. *Earth-Science Reviews* 84: 103–138.
- Chase BM, Quick LJ, Meadows ME, Scott L, Thomas DSG, and Reimer PL (2011) Late glacial interhemispheric dynamics revealed in South African hyrax middens. *Geology* 39: 19–22.
- Cheddadi R, Fady B, Francois L, et al. (2009) Putative glacial refugia of *Cedrus atlantica* deduced from Quaternary pollen records and modern genetic diversity. *Journal of Biogeography* 36: 1361–1371.
- Clark PU, Dyke AS, Shakun JD, et al. (2009) The last glacial maximum. *Science* 325: 710–714.
- Coetzee JA (1967) Pollen analytical studies in East and southern Africa. *Palaeoecology of Africa* 3: 1–146.
- Colyn M, Gautier-Hion A, and Verheyen W (1991) A re-appraisal of the palaeoenvironmental history in Central Africa: Evidence for a major fluvial refuge in the Zaire Basin. *Journal of Biogeography* 18: 403–407.
- Connor EF (1986) The role of Pleistocene forest refugia in the evolution and biogeography of tropical biotas. *Trends in Ecology & Evolution* 1: 165–168.
- DeBusk GH (1998) A 37,500 year pollen record from Lake Malawi and implications for the biogeography of Afrotropical forests. *Journal of Biogeography* 25: 479–500.
- Dupont LM and Behling H (2006) Land-sea linkages during deglaciation: High-resolution records from the eastern Atlantic off the coast of Namibia and Angola (ODP site 1078). *Quaternary International* 148: 19–28.
- Dupont LM, Behling H, and Kim J-H (2008) Thirty thousand years of vegetation development and climate change in Angola (Ocean Drilling Program Site 1078). *Climate of the Past* 4: 107–124.
- Dupont LM, Jahns S, Marret F, and Shi N (2000) Vegetation change in equatorial West Africa: Time-slices for the last 150 ka. *Palaeogeography, Palaeoclimatology, Palaeoecology* 155: 95–122.
- Dupont LM and Weinelt M (1996) Vegetation history of the savanna corridor between the Guinean and the Congolian rain forest during the last 150,000 years. *Vegetation History and Archaeobotany* 5: 273–292.
- Ehrlinger JR, Cerling TF, and Helliker BR (1997) C₄ photosynthesis, atmospheric CO₂, and climate. *Oecologia* 112: 285–299.
- Elenga H, Peyron O, Bonnefille R, et al. (2000) Pollen-based biome reconstruction for southern Europe and Africa 18,000 yr BP. *Journal of Biogeography* 27: 621–634.
- Elenga H, Schwartz D, and Vincens A (1994) Pollen evidence of late Quaternary vegetation and inferred climate changes in Congo. *Palaeogeography, Palaeoclimatology, Palaeoecology* 109: 345–356.
- Finch JM and Hill TR (2008) A late Quaternary pollen sequence from Mfabeni Peatland, South Africa. Reconstructing forest history in Maputaland. *Quaternary Research* 70: 442–450.
- Finch J, Leng ML, and Marchant R (2009) Late Quaternary vegetation dynamics in a biodiversity hotspot, the Uluguru Mountains of Tanzania. *Quaternary Research* 72: 111–122.
- Garcin Y, Vincens A, Williamson D, and Guiot J (2006a) Wet phases in tropical southern Africa during the last glacial period. *Geophysical Research Letters* 33: L07703. <http://dx.doi.org/10.1029/2005GL025531>.
- Garcin Y, Williamson D, Taieb M, Vincens A, Mathé, and Majule A (2006b) Centennial to millennial changes in maar-lake deposition during the last 45,000 years in tropical southern Africa (Lake Masoko, Tanzania). *Palaeogeography, Palaeoclimatology, Palaeoecology* 239: 334–354.
- Gasse F, Chalief F, Vincens A, Williams MAJ, and Williamson D (2008) Climatic patterns in equatorial and southern Africa from 30,000 to 10,000 years ago reconstructed from terrestrial and near-shore proxy data. *Quaternary Science Reviews* 27: 2316–2340.
- Gasse F and Van Campo E (1998) A 40,000-yr pollen and diatom record from Lake Tritrivakely, Madagascar, in the southern tropics. *Quaternary Research* 49: 299–311.
- Hessler I, Dupont L, Bonnefille R, et al. (2010) Millennial-scale changes in vegetation records from tropical Africa and South America during the last glacial. *Quaternary Science Reviews* 29: 2882–2899.
- Jahns S, Huls M, and Sarntin M (1998) Vegetation and climate history of west equatorial Africa based on a marine pollen record off Liberia (site GIK 16776) covering the last 400,000 years. *Review of Palaeobotany and Palynology* 102: 277–288.
- Jolly D, Harrison SP, Damnati B, and Bonnefille R (1998) Simulated climate and biomes of Africa during the late Quaternary: Comparison with pollen and lake status data. *Quaternary Science Reviews* 17: 629–657.
- Jolly D and Haxelline A (1997) Effect of low glacial atmospheric CO₂ on tropical African montane vegetation. *Science* 276: 786–788.
- Jolly D, Taylor D, Marchant R, et al. (1997) Vegetation dynamics in central Africa since 18,000 yr BP: Pollen records from the interlacustrine highlands of Burundi, Rwanda and western Uganda. *Journal of Biogeography* 24: 495–512.

- Kim So-Young, Scourse J, Marret F, and Dhong-II Lim (2010) A 26,000 year integrated record of marine and terrestrial environmental change off Gabon, west equatorial Africa. *Palaeogeography, Palaeoclimatology, Palaeoecology* 297: 426–438.
- Lamb HF, Bates CR, Coombes PV, et al. (2007) Late Pleistocene desiccation of Lake Tana, source of the Blue Nile. *Quaternary Science Reviews* 26: 287–299.
- Lamb HF, Eicher U, and Switsur VR (1989) An 18,000 year record of vegetation, lake-level, and climatic change from Tigalmamine, Middle Atlas, Morocco. *Journal of Biogeography* 16: 65–74.
- Lebamba J, Vincens A, and Maley J (2010) Pollen, biomes, forest successions and climate at Lake Barombi Mbo (Cameroon) during the last ca. 33000 cal yr BP: A numerical approach. *Climate of the Past Discussions* 6: 2703–2740.
- Livingstone DA (1971) A 22,000 year old pollen record from the plateau of Zambia. *Limnology and Oceanography* 16: 349–356.
- Maley J (1991) The African rain forest vegetation and palaeoenvironments during late Quaternary. In: Myers N (ed.) *Tropical Forests and Climate*, pp. 79–98. Dordrecht-Boston: D Reidel Publishing Company.
- Maley J (2000) Last Glacial Maximum lacustrine and fluvial formations in the Tibesti and other Saharan mountains, and large-scale climatic teleconnections linked to the activity of the subtropical jet stream. *Global and Planetary Change* 26: 121–136.
- Maley J and Brénac P (1998) Vegetation dynamics, palaeoenvironments and climatic changes in the forests of western Cameroon during the last 28,000 years BP. *Review of Palaeobotany and Palynology* 99: 157–187.
- Marchant R and Taylor D (1998) Dynamics of montane forest in Central Africa during the late Holocene: A pollen-based record from western Uganda. *The Holocene* 8: 375–381.
- Marchant R, Taylor D, and Hamilton AC (1997) Late Pleistocene and Holocene history at Mubwindi Swamp, Southwest Uganda. *Quaternary Research* 47: 316–328.
- Marret F, Scourse JD, Versteegh G, Jansen JHF, and Schneider R (2001) Integrated marine and terrestrial evidence for abrupt Congo River palaeodischarge fluctuations during the last deglaciation. *Journal of Quaternary Science* 16: 761–766.
- Meadows ME (1996) Biogeography. In: Adams WM, Goudie AS, and Orme AR (eds.) *The Physical Geography of Africa*, pp. 161–173. Oxford: Oxford University Press.
- Meadows ME (1999) Biogeography: Changing places, changing times. *Progress in Physical Geography* 23: 257–270.
- Meadows ME, Selanie M, and Chase BM (2010) Holocene palaeoenvironments of the Cederberg and Swartkops mountains, Western Cape, South Africa: Pollen and stable isotope evidence from hyrax dung middens. *Journal of Arid Environments* 74: 786–793.
- Mumbi CT, Marchant R, Hooghiemstra H, and Wooller MJ (2008) Late Quaternary vegetation reconstruction from the Eastern Arc Mountains, Tanzania. *Quaternary Research* 69: 326–341.
- Parkington J, Cartwright C, Cowling RM, Baxter A, and Meadows ME (2000) Palaeovegetation at the last glacial maximum in the Western Cape, South Africa: Wood charcoal and pollen evidence from Elands Bay Cave. *South African Journal of Science* 96: 543–546.
- Rhouijati A, Cheddadi R, Taieb M, Baali A, and Ortu E (2010) Environmental changes over the past c.29,000 years in the Middle Atlas (Morocco): A record from Lake Ifrah. *Journal of Arid Environments* 74: 737–745.
- Rucina SM, Muiruri VM, Kinyanjui RM, McGuinness K, and Marchant R (2009) Late Quaternary vegetation and fire dynamics on Mt. Kenya. *Palaeogeography, Palaeoclimatology, Palaeoecology* 283: 1–14.
- Schalke HJWG (1973) The Upper Quaternary of the Cape Flats area. *Scripta Geologica* 15: 1–57.
- Scott L (1982) A late Quaternary pollen record from the Transvaal bushveld, South Africa. *Quaternary Research* 17: 339–340.
- Scott L (1987) Pollen analysis of hyena coprolites and sediments from Equus Cave, Taung, southern Kalahari (South Africa). *Quaternary Research* 28: 144–156.
- Scott L (1989) Late Quaternary vegetation history and climatic change in the eastern Orange Free State, South Africa. *South African Journal of Botany* 55: 107–116.
- Scott L (1994) Palynology of Late Pleistocene hyrax middens, southwestern Cape Province, South Africa: A preliminary report. *Historical Biology* 9: 71–81.
- Scott L (1999) Vegetation history and climate in the Savanna biome South Africa since 190,000 ka: A comparison of pollen data from the Tswaing Crater (the Pretoria Saltpan) and Wonderkrater. *Quaternary International* 57–58: 215–223.
- Scott L (2002) Grassland development under glacial and interglacial conditions in southern Africa: Review of pollen, phytolith and isotope evidence. *Palaeogeography, Palaeoclimatology, Palaeoecology* 177: 47–57.
- Scott L, Holmgren K, and Partridge TC (2008) Reconciliation of vegetation and climatic interpretations of pollen profiles and other regional records from the last 60 thousand years in the Savanna Biome of southern Africa. *Palaeogeography, Palaeoclimatology, Palaeoecology* 257: 198–206.
- Scott L, Marais E, and Brook GA (2004) Fossil hyrax dung and evidence of Late Pleistocene and Holocene vegetation types in the Namib Desert. *Journal of Quaternary Science* 19: 829–832.
- Scott L and Vogel JC (2000) Evidence for environmental conditions during the last 20,000 years in Southern Africa from ¹³C in fossil hyrax dung. *Global and Planetary Change* 26: 207–215.
- Scott L and Woodborne S (2007a) Pollen analysis and dating of late Quaternary faecal deposits (hyraceum) in the Cederberg, Western Cape, South Africa. *Review of Palaeobotany and Palynology* 144: 123–134.
- Scott L and Woodborne S (2007b) Vegetation history inferred from pollen in late Quaternary faecal deposits (hyraceum) in the Cape winter-rain region and its bearing on past climates. *Quaternary Science Reviews* 26: 941–953.
- Shanahan TM, Overpeck JT, Beck JW, et al. (2008) The formation of biogeochemical laminations in Lake Bosumtwi, Ghana, and their usefulness as indicators of past environmental changes. *Journal of Paleolimnology* 40: 339–355.
- Shanahan TM, Overpeck JT, Wheeler CW, et al. (2006) Paleoclimatic variations in West Africa from a record of late Pleistocene and Holocene lake level stands of Lake Bosumtwi, Ghana. *Palaeogeography, Palaeoclimatology, Palaeoecology* 242: 287–302.
- Shi N, Beug H-J, and Schneider R (2000) Correlation between vegetation in southwestern Africa and oceanic upwelling in the past 21,000 years. *Quaternary Research* 54: 72–80.
- Shi N, Schneider R, Beug H-J, and Dupont LM (2001) Southeast trade wind variations during the last 135 kyr: Evidence from pollen spectra in eastern South Atlantic sediments. *Earth and Planetary Science Letters* 187: 311–321.
- Street-Perrott FA, Barker PA, Swain DL, et al. (2007) Late Quaternary changes in ecosystems and carbon cycling on Mt. Kenya, East Africa: A landscape-ecological perspective based on multi-proxy lake sediment influxes. *Quaternary Science Reviews* 26: 1838–1860.
- Stuut J-BW, Crosta X, Van der Borg K, and Schneider RR (2004) On the relationship between Antarctic sea ice and southwestern African climate during the late Quaternary. *Geology* 32: 909–912.
- Stuut J-BW and Lamy F (2004) Climate variability at the southern boundaries of the Namib (southwestern Africa) and Atacama (northern Chile) coastal deserts during the last 120,000 yr. *Quaternary Research* 62: 301–309.
- Stuut J-BW, Prins MA, Schneider RR, Weltje GJ, Jansen JHF, and Postma G (2002) A 300 kyr record of aridity and wind strength in southwestern Africa: Inferences from grain-size distributions of sediments on Walvis Ridge, SE Atlantic. *Marine Geology* 180: 221–233.
- Tierney JE, Russell JM, and Huang Y (2010) A molecular perspective on late Quaternary climate and vegetation in the Lake Tanganyika basin, East Africa. *Quaternary Science Reviews* 29: 787–800.
- Tzedakis C (2007) Seven ambiguities in the Mediterranean palaeoenvironmental narrative. *Quaternary Science Reviews* 26: 2042–2066.
- van Zinderen Bakker EM (1976) The evolution of late Quaternary paleoclimates of Southern Africa. *Palaeoecology of Africa* 9: 160–202.
- Vincens A (1991) Late Quaternary vegetation history of the South Tanganyika basin: Climatic implications in south central Africa. *Palaeogeography, Palaeoclimatology, Palaeoecology* 86: 207–226.
- Vincens A (1993) Nouvelle séquence pollinique du lac Tanganyika. 30,000 ans d'histoire botanique et climatique du bassin Nord. *Review of Palaeobotany and Palynology* 78: 381–394.
- Vincens A, Buchet G, Williamson D, and Taieb M (2005) A 23,000 year pollen record from Lake Rukwa (8° S, SW Tanzania): New data on vegetation dynamics and climate in Central East Africa. *Review of Palaeobotany and Palynology* 137: 147–162.
- Vincens A, Garcin Y, and Buchet G (2007) Influence of rainfall seasonality on African lowland vegetation during the late Quaternary: Pollen evidence from Lake Masoko, Tanzania. *Journal of Biogeography* 34: 1274–1288.
- Watrin J, Lezine AM, Gajewski K, and Vincens A (2007) Pollen-plant-climate relationships in sub-Saharan Africa. *Journal of Biogeography* 34: 489–499.

Australasia

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Introduction

The vegetation of the Australasian region reflects the major changes that have taken place in its geographical position, and hence climate, through the Cenozoic Era, including a long period of isolation from other parts of the world. Prior to this time, the whole of Australasia was part of the southern land mass of Gondwana, much of which was dominated by a wet temperate to subtropical flora that today is largely restricted to cool to warm temperate rainforests of New Zealand and south-eastern Australia, extending, in small patches, along the eastern Australian coastal ranges to tropical north-eastern Australia and New Guinea (Hill, 1994). After separation from Gondwanaland, New Zealand remained largely within temperate latitudes but Australia has progressively moved northward into warmer and drier latitudes. This period of isolation resulted in the development of the unique nature of Australian vegetation, characterized by a high degree of sclerophyll (the development of thickened, hardened foliage that resists loss of moisture). There are extensive xerophytic woodlands and shrublands, with *Acacia* as a major dominant, within the heart of the continent and mainly *Eucalyptus*-dominated sclerophyll forest, woodland, and shrubland on the wetter periphery (Kershaw et al., 2000) (Figure 1).

The continued northward movement of Australia into tropical latitudes has brought it into close proximity to the tropical rainforest environments of Southeast Asia. Although there has been little faunal mixing between the two regions, as evidenced by the well-established Wallace's Line, there has been significant immigration of rainforest taxa into New Guinea within the last 15 Ma or so as indicated by the characteristically Southeast Asian composition of the lowland rainforest. However, the extent of invasion of rainforest taxa into Australia is debated (Truswell et al., 1987; Whiffin, 2002; Sniderman and Jordan, 2011). The situation is complicated by a possible recycling of Gondwanan taxa into Australia via India and Southeast Asia and by the nature of present-day small drier rainforest patches (tropical deciduous broadleaf forest or monsoon forest) that show affinities with more tropical dry forests on other Gondwanan continents (Morley, 2002).

The significance of this history to late Pleistocene vegetation, as constructed from pollen records, is that, unlike other parts of the world that have not experienced this degree of isolation, there has been minimal input of plants that have evolved elsewhere, and this may have influenced specific responses of the indigenous vegetation to Quaternary climate fluctuations. In the case of New Zealand, this isolation continued through the late Pleistocene. Isolation was broken in Australia within the period of interest with the arrival of people by at least 45 ka (O'Connell and Allen, 2004; here and elsewhere 'ka' refers to 'kiloyears' or thousands of years before present) but people were absent from New Zealand until about the last 1 ka (Wilmshurst et al., 2008). A focus on the

late Pleistocene within the Australasian region as a whole, therefore, allows some assessment of the relative influences of climate and people on the regional vegetation, as evidenced in the pollen record.

Nature of the Data

The vast majority of pollen records within the region are restricted to the Holocene interval with few extending far into the Pleistocene. This is partly a result of a lack of suitable sites for sediment accumulation within a continent dominated by tectonically old land surfaces combined with dry or climatically very variable conditions that are not conducive to pollen preservation. The major exceptions to this pattern are found along the eastern seaboard of Australia, in addition to New Guinea and New Zealand, where all terrestrial records covering at least the whole of the late Pleistocene are located. Both New Guinea and New Zealand are volcanically and tectonically active and have experienced some glaciation while recent volcanic activity has occurred within limited parts of north-eastern and south-eastern Australia. Minor tectonism and glaciation is evident in the highlands of south-eastern Australia. Even in these areas, though, the majority of pollen records are discontinuous.

Complementing and extending the terrestrial pollen data are records from marine cores. These are generally more continuous, have the major advantage of being supported by chronologies derived from associated oxygen isotope records, and can record vegetation from drier areas where no terrestrial sedimentary records exist. They generally lack the resolution of terrestrial sedimentary records due to slower sedimentation rates and some surface sediment mixing and the source of the pollen and its mechanisms of transport to a site are more difficult to determine without extensive modern pollen sampling.

Additional limitations to palynology in Australia are the relative representation and taxonomic resolution of critical taxa. The dominant of more arid vegetation, *Acacia*, is notoriously underrepresented due to relatively poor pollen production and dispersal. Other major taxa such as *Eucalyptus*, Poaceae, Asteraceae, and Chenopodiaceae can seldom be identified beyond the genus or family level despite containing a large number of species. However, the restriction of late Pleistocene records largely to wetter areas that can support a variety of rainforest as well as sclerophyll communities associated with steep rainfall and/or temperature gradients, provides good pollen sensitivity to environmental forcing.

Major features of records, covering most or all of the late Pleistocene for different regions, are selected here to provide a framework of vegetation change within the Australasian region (Figure 1). Charcoal values, where available, are presented with pollen data to provide an indication of the biomass

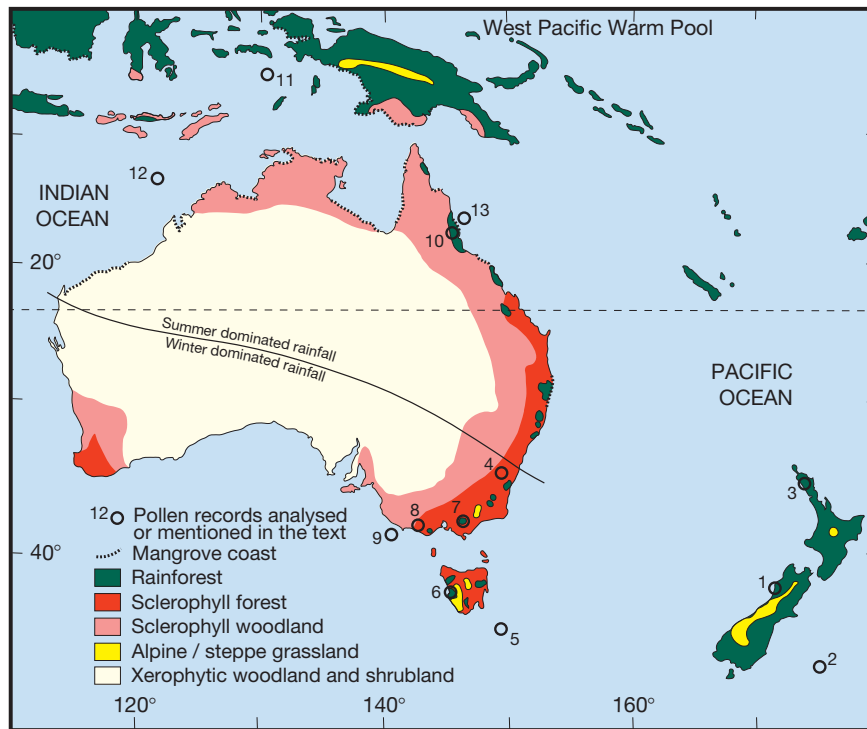


Figure 1 Generalized present day vegetation cover of the Australasian region (Kershaw et al., 2000) and location of selected late Pleistocene pollen records: 1 (Okarito Pakahi), 2 (DSDP Site 594), 3 (Lake Tangonge), 4 (Lake George), 5 (Core GC-3), 6 (Lake Selina), 7 (Caledonia Fen), 8 (Lake Wangoom), 9 (core E-55-6), 10 (Lynch's Crater), 11 (Aru Sea core MD982175), 12 (North Australian Basin core MD982167), 13 (ODP Site 820).

burning that is an important characteristic of Australian environments. The Holocene component of each record is included to provide a recent reference for assessment of degree and direction of past changes while some indication of the extent of variation in vegetation within the last glacial cycle can be visualized from comparison of reconstructed Holocene vegetation (Figure 1) and the estimated vegetation cover at the extreme of the glacial interval, the Last Glacial Maximum (LGM) (Figure 2). The previous glacial cycle in these records, where available, is also included to allow assessment of the degree to which the last glacial cycle is representative of glacial-interglacial cyclicity and to allow a better indication of climate versus human influences in those records where human activity within the last cycle is likely. Marine isotope records that provide largely a measure of Northern Hemisphere ice volume, and regional insolation curves (Berger and Loutre, 1991), are included with pollen diagrams, where appropriate, to allow assessment of Northern and Southern Hemisphere climate forcing influences on the vegetation.

New Zealand

Long marine pollen records from off the eastern coast of New Zealand provide a broad picture of the New Zealand vegetation and, from comparison with associated marine oxygen isotope records, suggest major Northern Hemisphere insolation and ice volume forcing with expansion of forest vegetation during warm and wet interglacials and replacement, to varying degrees, by shrubland and steppe-grassland during

cooler and drier glacial intervals (Heusser and van de Geer, 1994; Mildenhall, 2003).

Indicative patterns of vegetation and climate change for much of the country during the last glacial cycle are provided by the glacial moraine-impounded peat bog of Okarito Pakahi, the only terrestrial record to continuously cover the late Pleistocene interval (Vandergoes et al., 2005) (Figure 3). This record is complemented by the generalized DSDP site 594 marine pollen record that also provides corroborative evidence for the chronology derived from radiocarbon, thermoluminescence, and tephra dates on Okarito Pakahi (Figure 3). High temperatures during the interglacial intervals of MIS 5e and 1, as well as interstadials MIS 5c and 5a, are indicated by high representation of pollen from lowland podocarp-hardwood rainforest, dominated by the podocarp *Dacrydium cupressinum*. This pattern supports the global temperature picture except that the last interglacial does not appear to achieve the degree of climatic amelioration experienced in the Holocene, and the vegetation response was as great in the interstadials of MIS 5c and 5a as it was in MIS 5e. The glacial intervals of MIS 6 and MIS 4–2 are characterized by the significant presence of pollen from herbs (largely Poaceae with notable Asteraceae) apart from the early part of MIS 3, indicating more open vegetation. The generally very high values of pollen from montane-submontane shrubs in the Okarito Pakahi record indicate temperatures substantially cooler than today. The reduced percentages of shrubs during MIS 6 are the result of high values of pollen for the cool temperature rainforest tree/shrub *Nothofagus* subgenus *Fuscospora*. The current absence of this taxon within the vegetation of this area is not a

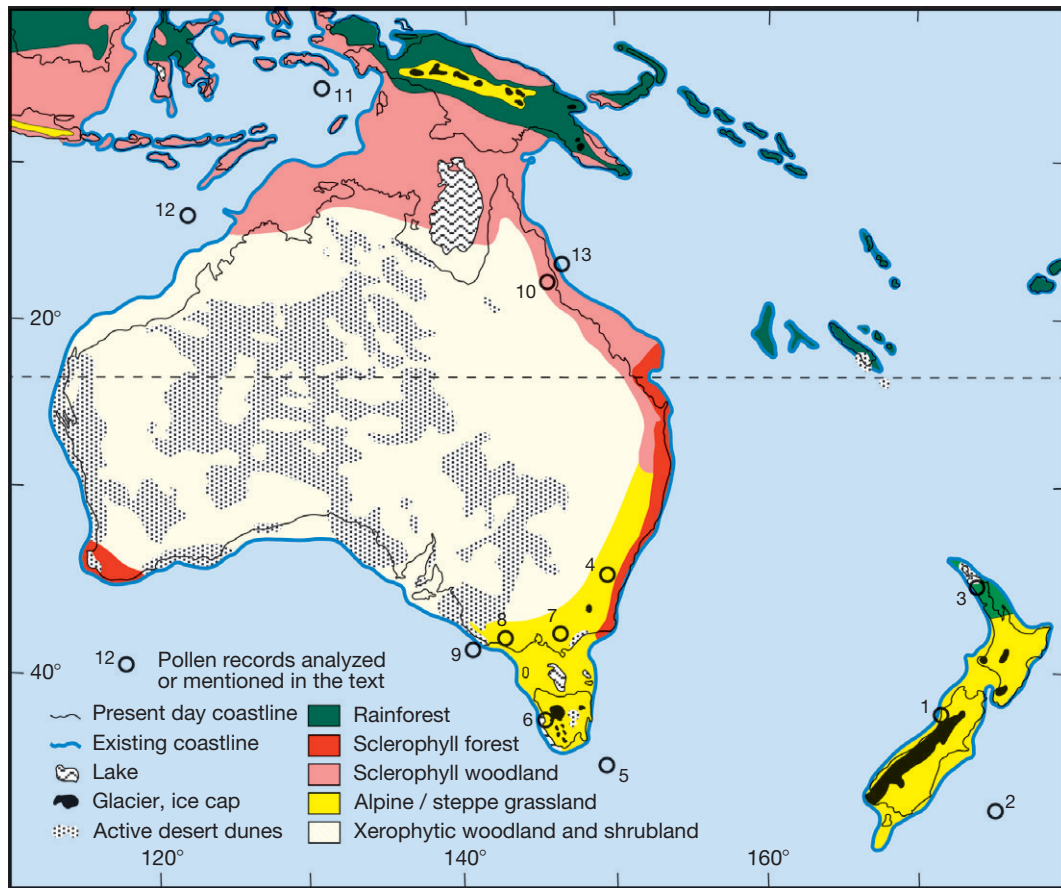


Figure 2 Inferred vegetation cover of the Australasian region at the Last Glacial Maximum (Kershaw et al., 2000). Pollen sites are the same as those in Figure 1.

feature of the island as a whole and has not been satisfactorily explained (e.g., Leschen et al., 2008).

A marked deviation from the Northern Hemisphere pattern is the early entry into maximum glacial conditions as indicated by a sharp rise in herb pollen values, perhaps about 30 ka as opposed to about 25 ka globally. This is suggested to be the result of forcing by regional Southern Hemisphere insolation preceding that of the Northern Hemisphere (Vandergoes et al., 2005). The subsequent maximum in Southern Hemisphere insolation, around 20 ka, may also have been responsible for early deglaciation in the region.

The northernmost part of New Zealand, Northland, is considered to have differed significantly in the nature of vegetation changes to the remainder of New Zealand in that, away from the direct influence of glaciation, precipitation, and also fire have been major controlling factors (Newnham, 1999). The difficulty of obtaining a continuous pollen sequence, though, led to debate over the degree to which forest persisted in this region during the last glacial interval. Some resolution of this question was provided by the construction of a lowland record from the dune-dammed Lake Tangonge (Elliot et al., 2005) (Figure 3). This record is considered to extend back to MIS 4 although the chronology beyond the limit of radiocarbon dating is uncertain. As through much of New Zealand, conifer-hardwood forest dominated by *Dacrydium cupressinum* is

currently dominant but is distinguished by the addition of the subtropical emergent araucarian conifer, *Agathis*, while *Nothofagus* subg. *Fuscospora* cool temperate forest is almost absent. The record suggests the maintenance of a complete forest cover through the last glacial interval with conifer-hardwood forest as a major component. This component was greatest during what is considered to be MIS 4. Conditions were wet but cool, as indicated by low *Agathis* values and the absence of the other major warm taxon, *Ascarina*. The significant representation of both these taxa in the early part of MIS 3 indicates at least as wet but significantly warmer conditions while the subsequent demise of the latter genus and increase in the former genus between about 43 and 30 ka suggests the development of greater seasonality with drier summers and cooler winters (Elliot et al., 2005). Full glacial conditions are indicated from the early date of around 30 ka by the expansion of *Nothofagus* that, combined with evidence for frequent burning, suggests a cool, drier and probably windier climate. The peaks in burning and Poaceae suggest that driest conditions occurred during termination of the Pleistocene. However, the limited dating control opens a possibility that there is a gap in the record here and that much of MIS 2 experienced dry conditions with some disruption of the forest cover. The vegetation of the Holocene is remarkably similar to that of early MIS 3.

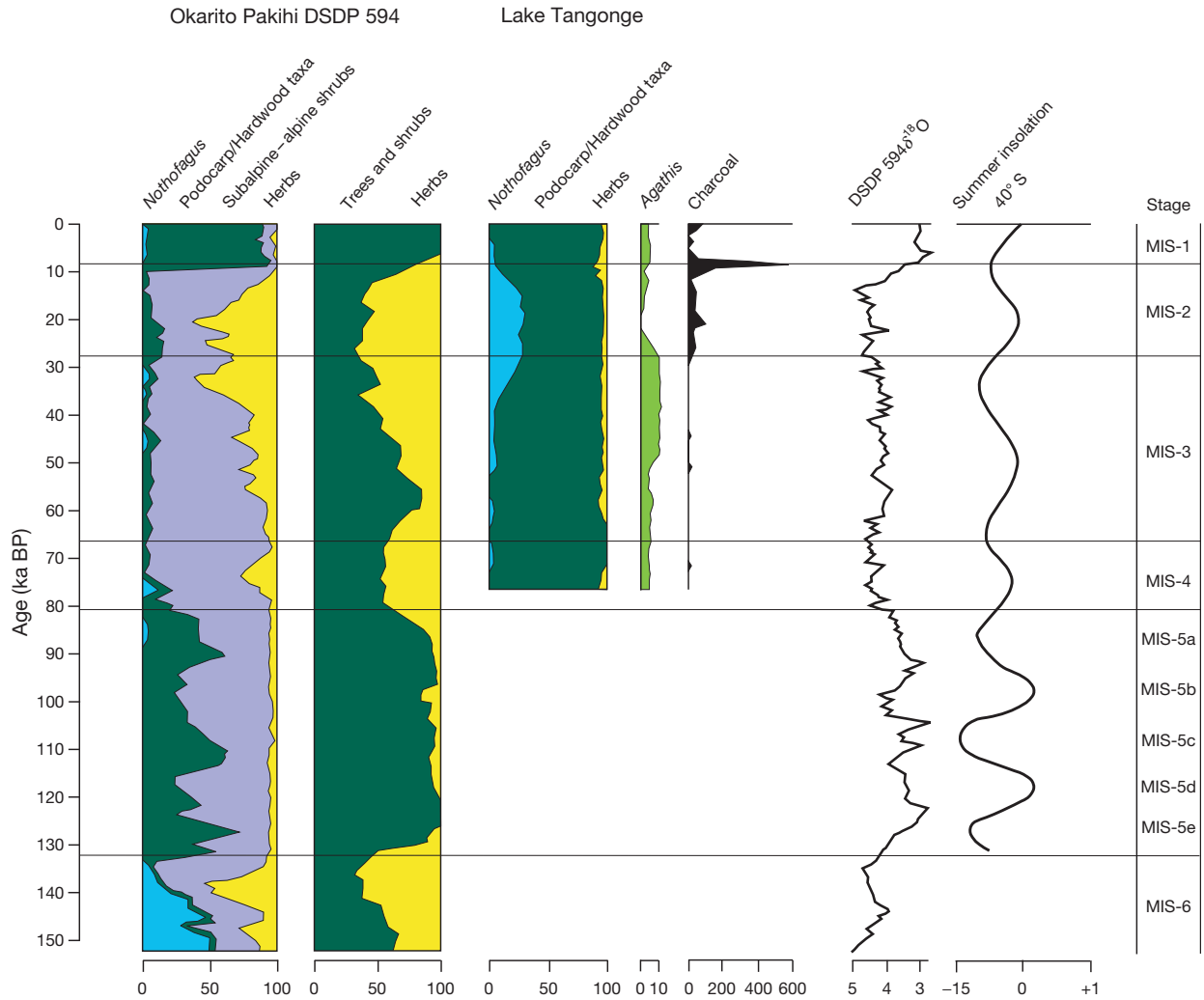


Figure 3 Variation in relative abundance of major or indicator taxa and taxon groups in relation to a dry land pollen sum for selected New Zealand pollen records compared to the DSDP 594 benthic $\delta^{18}\text{O}$ record, the summer insolation curve for 40°S and the marine isotope stratigraphy of [Martinson et al. \(1987\)](#). Charcoal values, expressed as particles/cc $\times 1000$, are also included for one site, Lake Tangonge, as a measure of past biomass burning.

The Northland region records uncertainties, both in terms of continuity and chronology as well as paleoenvironmental reconstruction. These issues are being addressed by multiproxy studies on volcanic maars that are known to cover the last 30 ka but have the potential to extend back continuously to at least 245 ka ([Molloy et al., 2009](#)). A combination of radiocarbon and tephrochronological ages on sediments from Onepoto maar, south of Lake Tangonge, demonstrate the continued occurrence of a forest cover dominated by *Nothofagus* from at least 30 ka to ~ 14 ka with some grassland and herb-field present during the dry, windy height of the last glacial interval, dated here between about 28.5 and 18 ka ([Augustinus et al., 2011](#)). Climatic amelioration from 18 to 14 ka is marked by an increase to dominance of lowland podocarp forest, especially *Dacrydium cupressinum*, and, after some Late Glacial perturbations, the presence of *Ascarina* indicates maximum Holocene temperatures by about 10.5 ka.

South-Eastern Australia

The controversially dated pollen record from Lake George ([Singh and Geissler, 1985](#)) suggested marked changes in the vegetation of south-eastern Australia through the middle and late Pleistocene that could be explained largely by Northern Hemisphere orbital and ice volume forcing and this pattern of change has been confirmed recently by a pollen record from core GC-3 on the Tasman Rise (Sander van der Kaars, unpublished). For the last glacial cycle, details of vegetation change, covering much of the landscape variation in more humid areas, are provided by three records: Lake Selina from the formerly glaciated landscape of western Tasmania ([Colhoun et al., 1999](#)), Caledonia Fen from the south-eastern highlands of Victoria ([Kershaw et al., 2007, 2010](#)), and Lake Wangoom from the western plains of Victoria ([Harle et al., 2002](#)) ([Figure 4](#)).

Lake Selina shows many similarities with Okarito Pakahi in its formation as a result of Pleistocene glaciation, and its setting

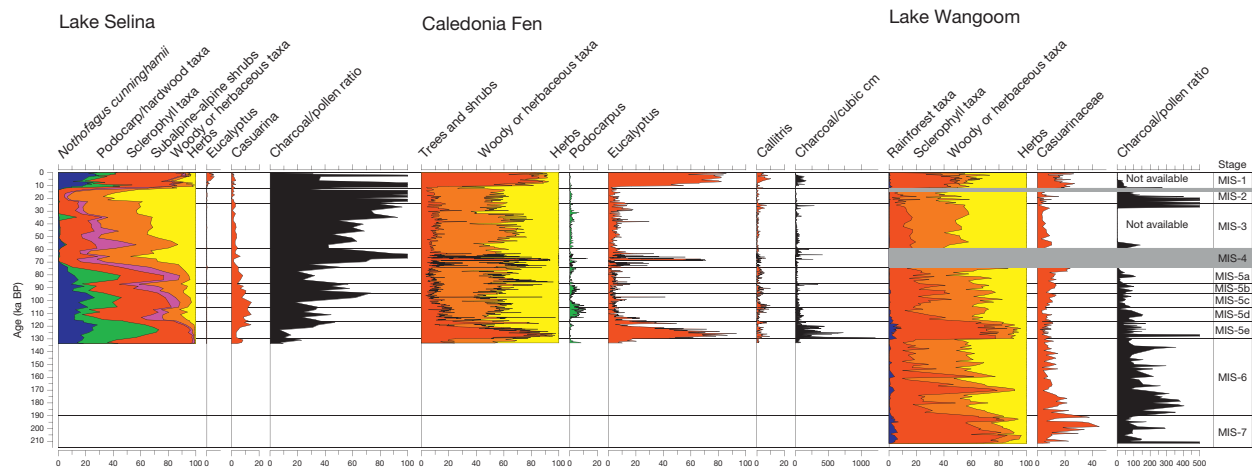


Figure 4 Variation in relative abundance of major or indicator taxa and taxon groups in relation to a dry land pollen sum for selected pollen records from south-eastern Australia compared with charcoal records from individual sites expressed as percentages of dry land pollen (Lake Selina and Lake Wangoom) and particles per cm^3 (Caledonia Fen), and the marine isotope stratigraphy of [Martinson et al. \(1987\)](#).

within Gondwanan temperate rainforest that experiences high rainfall, in excess of 3000 mm per annum. It differs in that it is at a higher altitude, ~ 500 m.a.s.l., that *Nothofagus* subgenus *Menziesii* (specifically *N. cunninghamii*) is the dominant rainforest element although conifers (largely *Phyllocladus*) are well represented, and in the intrusion of sclerophyll taxa into the vegetation. Although radiocarbon dates suggest that the record probably extends to the beginning of MIS 3, U/Th dates, together with comparison with the marine isotope stratigraphy, led to the conclusion that it covered the whole of the last glacial–interglacial cycle ([Colhoun et al., 1999](#)). Certainly, the pattern of change that is so similar to that at Okarito Pakahi would support this view, although the sclerophyll components of the two rainforest-dominated phases at Selina are very different, with Casuarinaceae pollen prominent in the early stage and *Eucalyptus* pollen in the latter stage. In common with Okarito Pakahi, dominance of rainforest during MIS 5e and the Holocene and the prominence of herbs, together with significant representation of alpine/subalpine shrubs, indicate strong temperature control over the sequence. In contrast to Okarito Pakahi, climatic amelioration during MIS 5e appears to have been much greater than in substages 5c and 5a. Despite the high rainfall, and dominance of rainforest during the interglacial intervals, fire has been a major feature of the environment and fire activity appears to have increased through the record.

Caledonia Fen probably formed within a blocked depression of a tributary of the Caledonia River. It lies at 1300 m.a.s.l. within eucalypt-dominated upper montane forest. It provides a rare glimpse into full last glacial conditions at high altitude ([Kershaw et al., 2007](#)). Age extrapolation to the base of the 20 m sequence on the basis of radiocarbon and optically stimulated luminescence ages higher in the sequence that fall within the last 60 ka suggests a late MIS 6 date for site formation. Using this chronology, the interglacials of MIS 5e and 1 stand out due to their dominance of eucalypt pollen and organic sediment. By contrast, almost the whole of the glacial intervals are dominated by steppe-grassland composed predominantly of Poaceae and Asteraceae, suggesting substantially

reduced temperatures. Reduced precipitation is also indicated by high values of salt-tolerant Chenopodiaceae that are excluded from the pollen sum. Low productivity in the shallow lake during glacial times is demonstrated by the lack of pollen from aquatic taxa, presumably because of ice coverage of the lake through much of each year. However, the record is also characterized by a high level of millennial variability that has yet to be fully analyzed. One substantial deviation from the glacial pattern, where forest development and aquatic productivity almost matched those of the interglacials, was subject to higher resolution study ([Kershaw et al., 2010](#)). The event was placed in early MIS 3 by [Kershaw et al. \(2007\)](#) but subsequent OSL ages suggested an MIS 4 age estimate of 66.8–65.5 ka. The abrupt nature of the event is suggestive of a Dansgaard–Oeschger (D–O) event but the fact that it is singular, appears not to date to a D–O event elsewhere, and is so remote from the presumed North Atlantic trigger, makes such a correlation unlikely and, hence, an interpretation difficult. It is also unusual that this degree of climatic amelioration is not recorded during the interstadials of MIS 5, which may only be recognized by fluctuations within the one remaining rainforest-related taxon in the region, *Podocarpus*, and the sclerophyll conifer, *Callitris*.

The low altitude volcanic crater site of Lake Wangoom, situated within open eucalypt forest or woodland, reflects mainly variation in effective rainfall rather than temperature. The published chronology, based on U/Th dates as well as by comparisons with a shorter marine palynological record ([Harle, 1997](#)) and the marine isotope stratigraphy, suggests that it covers the last two glacial cycles ([Harle et al., 2002](#)). In general terms, interglacials are clearly identified by marked expansion of forest vegetation, while glacials are characterized by steppe-grassland dominated by Poaceae and Asteraceae, but the chronological control is insufficient to identify interstadial and stadial phases with any certainty. An alternative chronology, based on radiocarbon and OSL dates within the upper part of record, suggests only one cycle with a major interstadial in MIS 3 achieving forest development at least equivalent to those of the Holocene and what would be the latter part

of MIS 5 at the base. Support for this chronology may be the lack of response of Casuarinaceae in the middle forest phase, perhaps suggesting a different climate scenario. Under either chronology, forest development in the Holocene may be less than expected by Milankovitch forcing in that grass pollen is highly represented and there is a much smaller proportion of pollen from wetter rainforest (predominantly *Nothofagus*) and wet sclerophyll forest (*Pomaderris*) elements. This pattern may be the result of either an overall trend to drier conditions or human impact as indicated by generally highest charcoal values for the record within MIS 3/2, using either timescale.

Northern Australia

For many years, the pollen record of Lynch's Crater from the Atherton volcanic province within the humid tropics of north-east Queensland provided the only basis for reconstruction of late Pleistocene vegetation and environments in northern Australia (Kershaw, 1985). However, in recent years, some indication of variation within the region has been provided by the reconstruction of several pollen records from marine sediments taken from the seas between Australia and Indonesia (Kershaw and van der Kaars, 2012). Two such records (MD982175 and MD982167), which relate to different northern Australian landscapes, are presented in Figure 5. In addition, a third marine record (ODP Site 820) from off the coast of

the humid tropics of north-eastern Australia is included to provide a comparison with Lynch's Crater (Figure 6).

The MD982175 Aru Sea core lies close to both New Guinea and the Sahul continental shelf, an extensive region that connected northern Australia to New Guinea during low sea-level phases. The pollen record covers about the last 155 ka, based on comparison with the associated oxygen isotope record (Beaufort et al., 2010). This record shows a dominance of rainforest pollen during the interglacial periods of MIS 5e and 1, as well as in the interstadials of MIS 5c and 5a (Kershaw and van der Kaars, 2012). Sclerophyll taxa increase slightly during the MIS 5 stadials and substantially during the glacial intervals of MIS 6 and 4–2 although the responses of sclerophyll trees (predominantly *Eucalyptus*) and herbs (almost entirely Poaceae) are different, based on their fossil pollen records. The pattern suggests that glacial climates were generally drier than those of the interglacials, probably the result of a general reduction in the influence of the Asian–Australian monsoon that is considered to be largely in phase with Northern Hemisphere radiation and ice volume forcing (Miller et al., 2005). However, regional influences such as low latitude southern insolation and the drying of the extensive continental shelves have had some importance. The degree of apparent rainforest reduction during glacials may also be exaggerated as a result of increased proximity of Australian eucalypt woodland to the site with the exposure and consequent occupation by vegetation of the Sahul shelf. Charcoal levels are variable and difficult to decipher in relation to changes in vegetation.

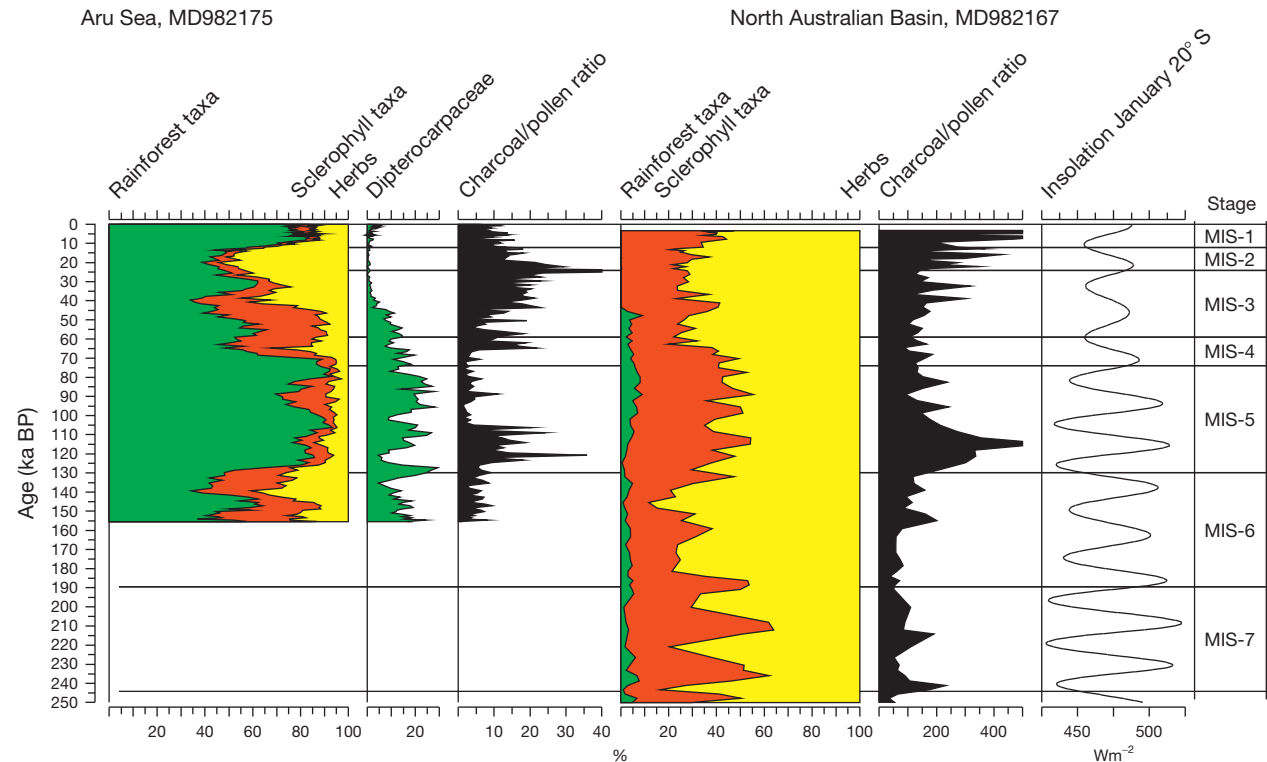


Figure 5 Variation in relative abundance of major or indicator taxa and taxon groups in relation to a dry land pollen sum for selected pollen records from northern Australasia (Aru Sea Core MD982175 and North Australia Basin Core MD982167) compared with charcoal/pollen ratios from each site, the marine isotope stratigraphy of Martinson et al. (1987), and the summer insolation curve for 20° S.

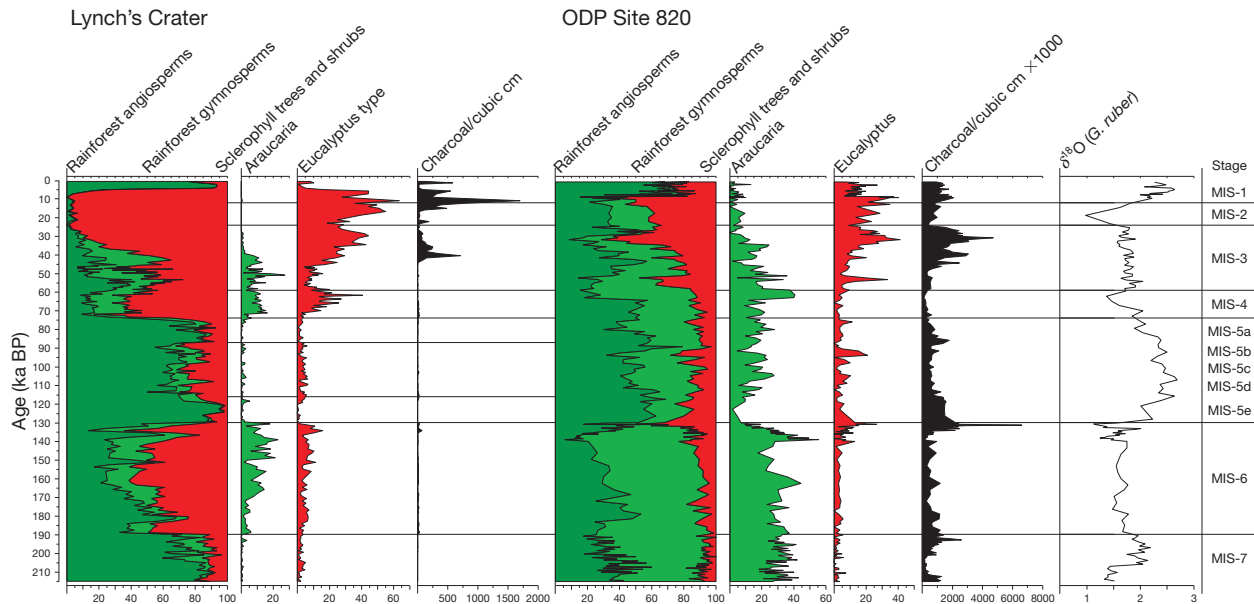


Figure 6 Variation in relative abundance of major or indicator taxa and taxon groups in relation to a dry land pollen sum for selected pollen records from north-eastern Australia (Lynch's Crater and ODP Site 820) compared with charcoal values for each site expressed as particles/cm³, the planktonic $\delta^{18}O$ curve from ODP Site 820, and the marine isotope stratigraphy of [Martinson et al. \(1987\)](#).

However, the beginning of a sustained increase in burning around 40 ka may have been ecologically significant in bringing about the decline in the regional dominant plant family of Southeast Asian lowland rainforest, Dipterocarpaceae.

Core MD982167, from the North Australian Basin, lies adjacent to a region of extensive eucalypt savanna vegetation with small patches of dry (monsoonal) rainforest in the Kimberley region. A summary diagram ([Kershaw and van der Kaars, 2012](#)), shown in relation to the chronology derived from an associated oxygen isotope record (Bradley Opdyke, personal communication), shows the predominance of this savanna woodland throughout the last 250 ka. Glacial-interglacial cyclicity is evident in the proportion of tree and shrub pollen (largely *Eucalyptus*) relative to herb pollen (essentially Poaceae). However, superimposed on this cyclicity is a strong 20 ka Southern Hemisphere precessional frequency. It is hypothesized that this is a monsoon signal emanating from the Indian Ocean, rather than a push from the Asian monsoon winds, with precessionally induced sea surface temperatures controlling much of the precipitation variation. Divorced from these patterns is a sharp increase in charcoal from the beginning of MIS 5 and a sharp reduction in rainforest pollen around 45 ka, just prior to a sustained increase in charcoal values.

In north-eastern Australia, the ODP site 820 record is accompanied by an isotope curve from the foraminiferan *Globigerinoides ruber* produced by [Peerdeman et al. \(1993\)](#). Although there is some debate over the degree of continuity of this core, major features such as the MIS 6/5 boundary are clear and have been used to assist in establishing a chronology for the Lynch's Crater record beyond the limit of radiocarbon dating ([Moss and Kershaw, 2000](#)). This chronology has recently been substantiated by direct OSL dating of the Lynch's Crater record ([Rieser and Wust, 2010](#)). Both records cover about the last

215 ka and reflect variation in the nature and extent of the most extensive patch of tropical rainforest in Australia. Herbs have been omitted from the pollen sums due to very local overrepresentation of Poaceae in the upper part of the Lynch's Crater record.

The record from Lynch's Crater exhibits clear glacial-interglacial cyclicity with pollen from angiosperm-dominated rainforest, which makes up almost entirely the humid tropics rainforest today, peaking during wetter interglacials. However, the Northern Hemisphere signal appears distorted in places, especially during MIS 5 where there is evidence of only one interstadial. Intervening glacial intervals show high values of sclerophyll or savanna woodland pollen. These plant communities surround the modern rainforest, although the pollen signal shows that *Eucalyptus* only became a dominant of this vegetation type in the last glacial period. High values of pollen from rainforest conifers, indicating extensive drier *Araucaria*-dominated rainforest, demonstrate partial replacement of the angiosperm rainforest until about 45 ka ([Turney et al., 2001](#)) when these were totally replaced by sclerophyll vegetation. The sharp and largely sustained increase in charcoal at this time has led to the long-held hypothesis that an increase in burning with the arrival of Aboriginal people resulted in the demise of fire-sensitive *Araucaria* forest and promotion of fire-resistant *Eucalyptus* ([Singh et al., 1981](#)). A similar relationship between *Araucaria* forest, *Eucalyptus*, and charcoal is noted in the ODP Site 820 record at a similar time, but the picture is complicated by an earlier partial reduction in *Araucaria* at the MIS 6/5 boundary that also shows a marked peak in charcoal and some *Eucalyptus* response, at a time before there is any evidence for people on the continent. It has been suggested that the early event was the result of a combination of dry conditions at the very end of MIS 6 and high El Niño/La Niña-Southern Oscillation (ENSO) activity, while the latter was caused by

human impact in combination with ENSO climatic variability (Kershaw et al., 2003). The effect of continued human burning may have retarded final rainforest expansion until about 7 ka, well after the Pleistocene/Holocene boundary.

General Discussion

The illustrated pollen records, covering a wide range of Australasian environments, demonstrate marked and varied responses of vegetation to past environmental forcing. Vegetation variation is regional in character with little evidence of long distance migration, a result, no doubt, of long-term adaptation to particular niches that are maintained in the absence of wholesale disruption that occurs with the development of continental ice sheets. The global influence of Northern Hemisphere insolation and ice volume, so evident in marine isotope records, is most clearly expressed in terrestrial records by temporal vegetation patterns in the most southerly latitudes of the continent where temperature is the dominant variable and in the northern part of Australia where some influence of the Asia monsoon system on precipitation variation is evident.

There is evidence of the superimposition of Southern Hemisphere insolation forcing on the orbital scale in the majority of records, and most particularly in northwestern Australia where the source of monsoon rainfall is predominantly from the surface of the Indian Ocean that has no direct connection to high northern latitudes. A subtle relationship between temperature and precipitation, including seasonal variations, as well as latitudinal movements in the major pressure systems, may have influenced patterns of vegetation change in the mid-latitudes of mainland south-eastern Australia. The suggestion of a distinct warm event at Caledonia Fen at 66 ka may be the result of age uncertainty though a similar event is evident from OSL ages in shorelines of old megalakes in more arid south-eastern Australia (Cohen et al., 2012). Within the latest Pleistocene, well dated records have indicated a Southern Hemisphere influence on the timing of the onset and termination of the coldest part of the last glacial interval, especially in New Zealand.

Some millennial scale variability is evident within the last glacial–interglacial transition but is regionally variable and shows little relationship with features like the Younger Dryas (e.g., Augustinus et al., 2011; Turney et al., 2006) which characterize much of the globe. Before this time-consistent millennial scale variability, as evident elsewhere mainly as D–O events within the last glacial period, is difficult to detect in pollen records, but may be a feature of the Caledonia Fen record (Kershaw et al., 2010) and has been demonstrated to exist within the local pollen and sediments of the Lynch’s Crater record where a teleconnection between Pacific ENSO and North Atlantic D–O cyclicity has been proposed (Turney et al., 2004).

There appears to be little relationship between biomass burning as indicated by charcoal, vegetation, and inferred climate change from the records presented. However, the recent analysis of some 223 sedimentary charcoal records from Australia, many extending into the Pleistocene, strongly suggests that burning has been higher during warmer periods,

including interglacials and D–O interstadials (Mooney et al., 2011).

A feature of several Australian records is a trend toward more sclerophyllous vegetation, generally associated with an increase in burning. A major event is identified around 40–45 ka that, in the absence of a major change in climate at this time and correspondence with earliest archaeological evidence (e.g., O’Connell and Allen, 2004), can, at least in part, be attributed to human burning. However, as indicated most clearly in the ODP Site 820 record, this trend was initiated prior to evidence for the presence of people. Despite the proposal by Singh et al. (1981) that vegetation changes associated with increased burning around 125 000 years ago in the Lake George pollen record were due to Aboriginal activities, and support for increased burning around that time in the northern Australian records presented here, a climatic cause is considered most likely. It has been suggested that the trigger may have been the crossing of a critical threshold in Australia’s movement toward Southeast Asia that constricted water movement through the Indonesian throughflow from the Pacific to Indian Ocean, leading to a build up of warm water forming the West Pacific Warm Pool (Figure 1), and the consequent establishment or intensification of ENSO variability (Isern et al., 1996; Kershaw et al., 2003). This whole scenario may be questioned on the basis of the charcoal analysis by Mooney et al. (2011), who see no systematic change in fire regime through time, including the period of human arrival. However, their analysis extends back only 70 ka and fails to take into account pollen data that allow assessment of the impact of fire, not just its pattern of occurrence.

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See also: Pollen Analysis, Principles. **Glaciation, Causes:** Tectonic Uplift and Continental Configurations. **Lake Level Studies:** Australia. **Pollen Records, Late Pleistocene:** Africa; South America. **Pollen Records, Postglacial:** Australia and New Zealand. **Quaternary Stratigraphy:** Climatostratigraphy.

References

- Augustinus P, D’Costa D, Deng Y, Hagg J, and Shane P (2011) A multi-proxy record of changing environments from ca. 30,000 to 9000 cal. A BP: Onepoto maar palaeolake, Auckland, New Zealand. *Journal of Quaternary Science* 26(4): 389–401.
- Beaufort L, van der Kaars S, Bassinot F, and Moron V (2010) Past dynamics of the Australian monsoon: Precession, phase and links to the global monsoon. *Climate of the Past* 6: 695–706.
- Berger A and Loutre MF (1991) Insolation values for the climate of the last 10 million years. *Quaternary Science Reviews* 10: 297–317.
- Cohen TJ, Nanson GC, Jansen JD, et al. (2012) Late Quaternary mega-lakes fed by the northern and southern river systems of central Australia: Varying moisture sources and increased continental aridity. *Palaeogeography, Palaeoclimatology, Palaeoecology* 356–357: 89–108.

- Colhoun EA, Pola JS, Barton CE, and Heijnis H (1999) Late Pleistocene vegetation and climate history of Lake Selina, western Tasmania. *Quaternary International* 57(58): 5–23.
- Elliot M, Neall V, and Wallace C (2005) A late Quaternary pollen record from Lake Tangonge, far northern New Zealand. *Review of Palaeobotany and Palynology* 136: 143–158.
- Harle KJ (1997) Late Quaternary vegetation change in southeastern Australia: Palynological evidence from marine core E55-6. *Palaeogeography, Palaeoclimatology, Palaeoecology* 131: 465–483.
- Harle KJ, Heijnis H, Chisari R, Kershaw AP, Zoppi U, and Jacobsen G (2002) A chronology for the long pollen record from Lake Wangoom, western Victoria as derived from uranium/thorium disequilibrium dating. *Journal of Quaternary Science* 17: 707–720.
- Heusser LE and van de Geer G (1994) Direct correlation of terrestrial and marine palaeoclimatic records from four glacial–interglacial cycles – DSDP Site 594 Southwest Pacific. *Quaternary Science Reviews* 13: 273–282.
- Hill RS (ed.) (1994) *History of the Australian Vegetation: Cretaceous to Recent*. Cambridge: Cambridge University Press.
- Isern AR, McKenzie JA, and Feary DA (1996) The role of sea-surface temperature as a control on carbonate platform development in the western Coral Sea. *Palaeogeography, Palaeoclimatology, Palaeoecology* 124: 247–272.
- Kershaw AP (1985) An extended late Quaternary vegetation record from north-eastern Queensland and its implications for the seasonal tropics of Australia. *Proceedings of the Ecological Society of Australia* 13: 179–189.
- Kershaw AP, McKenzie GM, Brown J, Roberts RG, and van der Kaars S (2010) Beneath the peat: A refined pollen record from an interstadial at Caledonia Fen, highland eastern Victoria, Australia. In: Haberle SG, Stephenson J, and Prebble M (eds.) *Altered Ecologies: Fire, Climate and Human Influence on Terrestrial Landscapes*, pp. 33–48. Canberra, Australia: ANU E-Press, The Australian National University. Available online at: http://eprint.anu.edu.au/terra_australis/ta32/pdf_instructions.html.
- Kershaw AP, McKenzie GM, Porph N, et al. (2007) A high resolution record of vegetation and climate through the last glacial cycle from Caledonia Fen, south-eastern highlands of Australia. *Journal of Quaternary Science* 22: 481–500.
- Kershaw AP, Quilty PG, van Huet S, David B, and McMinn A (2000) The Quaternary. *Australasian Association of Palaeontologists Memoirs* 23: 461–506.
- Kershaw AP and van der Kaars S (2012) Australia and the southwest Pacific. In: Metcalfe SE and Nash DJ (eds.) *Quaternary Environmental Change in the Tropics*, pp. 236–262. Oxford: John Wiley and Sons.
- Kershaw AP, van der Kaars S, and Moss PT (2003) Late Quaternary Milankovitch-scale climate change and variability and its impact on monsoonal Australia. *Marine Geology* 201: 81–95.
- Leschen RA, Buckley TR, Harman HM, and Shulmeister J (2008) Determining the origin and age of the Westland beech (*Nothofagus*) gap, New Zealand, using fungus beetle genetics. *Molecular Ecology* 17: 1256–1278.
- Martinson DG, Pisias NG, Hays JD, Imbrie J, Moore TC Jr., and Shackleton NJ (1987) Age dating and the orbital theory of the ice ages: Development of a high resolution 0–300,000-year chronostratigraphy. *Quaternary Research* 27: 1–29.
- Mildenhall DC (2003) Deep-sea record of Pliocene and Pleistocene terrestrial palynomorphs from offshore eastern New Zealand (ODP Site 1123, Leg 181). *New Zealand Journal of Geology and Geophysics* 46: 343–361.
- Miller GH, Mangan J, Pollard D, Thompson SL, Felzer BS, and Magee JW (2005) Sensitivity of the Australian monsoon to insolation and vegetation: Implications for human impact on continental moisture balance. *Geology* 32: 885–888.
- Molloy C, Shane PA, and Augustinus PC (2009) Eruption recurrence rates in a basaltic volcanic field based on tephra layers in maar sediments: Implications for hazards in the Auckland volcanic field. *Geological Society of America Bulletin* 121: 1666–1677.
- Mooney SD, Harrison SP, Bartlein PJ, et al. (2011) Late Quaternary fire regimes of Australasia. *Quaternary Science Reviews* 30: 28–46.
- Morley RJ (2002) Tertiary vegetational history of Southeast Asia with emphasis on the biogeographical relationship with Australia. In: Kershaw P, David B, Tapper N, Penny D, and Brown J (eds.) *Bridging Wallace's Line: The Environmental and Cultural History and Dynamics of the SE-Asian–Australian Region*, pp. 49–60. Reiskirchen: Catena Verlag.
- Moss PT and Kershaw AP (2000) The last glacial cycle from the humid tropics of northeastern Australia: Comparison of a terrestrial and a marine record. *Palaeogeography, Palaeoclimatology, Palaeoecology* 155: 155–176.
- Newnham R (1999) Environmental change in Northland, New Zealand during the last glacial and Holocene. *Quaternary International* 57(58): 61–70.
- O'Connell JF and Allen J (2004) Dating the colonisation of Sahul (Pleistocene Australia–New Guinea): A review of recent research. *Journal of Archaeological Science* 31: 835–853.
- Peerdeman FM, Davies PJ, and Chivas AR (1993) The stable oxygen isotope signal in shallow-water, upper-slope sediments off the Great Barrier Reef (Hole 820A). *Proceedings of the Ocean Drilling Program, Scientific Results* 133: 163–173.
- Rieser U and Wust RAJ (2010) OSL chronology of Lynch's Crater, the longest terrestrial record in NE-Australia. *Quaternary Geochronology* 5: 233–236.
- Singh G and Geissler EA (1985) Late Cainozoic history of fire, lake levels and climate at Lake George, New South Wales, Australia. *Philosophical Transactions of the Royal Society of London* 311: 379–447.
- Singh G, Kershaw AP, and Clark R (1981) Quaternary vegetation and fire history in Australia. In: Gill AM, Groves RA, and Noble IR (eds.) *Fire and the Australian Biota*, pp. 23–54. Canberra: Australian Academy of Science.
- Sniderman JMK and Jordan GJ (2011) Extent and timing of floristic exchange between Australian and Asian rain forests. *Journal of Biogeography* 38: 1445–1455.
- Truswell EM, Kershaw AP, and Sluiter IR (1987) The Australian/Malaysian connection: Evidence from the palaeobotanical record. In: Whitmore TC (ed.) *Biogeographic Evolution of the Malay Archipelago*, pp. 32–49. Oxford: Oxford University Press.
- Turney CSM, Kershaw AP, Clemens S, Branch N, Moss PT, and Fifield LK (2004) Millennial and orbital variations in El Niño/Southern Oscillation and high latitude climate in the last glacial period. *Nature* 428: 306–310.
- Turney CSM, Kershaw AP, Lowe JJ, et al. (2006) Climatic variability in the southwest Pacific during the last termination (20–10ka BP). *Quaternary Science Reviews* 25: 886–903.
- Turney CSM, Kershaw AP, Moss P, et al. (2001) Redating the onset of burning at Lynch's Crater (North Queensland): Implications for human settlement in Australia. *Journal of Quaternary Science* 16: 767–771.
- Vandergoes MJ, Newnham RM, Preusser F, et al. (2005) Regional insolation forcing of late Quaternary climate change in the Southern Hemisphere. *Nature* 436: 242–245.
- Whiffin T (2002) Plant biogeography of the SE Asian–Australian region. In: Kershaw P, David B, Tapper N, Penny D, and Brown J (eds.) *Bridging Wallace's Line: The Environmental and Cultural History and Dynamics of the SE-Asian–Australian Region*, pp. 61–82. Reiskirchen: Catena Verlag.
- Wilmschurst JM, Anderson AJ, Higham TFG, and Worthy TH (2008) Dating the late prehistoric dispersal of Polynesians to New Zealand using the commensal Pacific rat. *Proceedings of the National Academy of Sciences* 105: 7676–7680.

Relevant Websites

<http://users.monash.edu.au/~pkershaw/database.html> – Quaternary long marine and terrestrial records of the southern hemisphere.

Northern Asia

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Introduction

Siberia, which spans all of northern Asia, often invokes timeless images of frozen barrens and vast stretches of snow-covered taiga. However, the modern landscape is relatively young, shaped by late Pleistocene conditions that were much warmer, much colder, and much drier than the present. Soviet scientists, who first described the environmental history of Siberia, mounted multi-disciplinary expeditions focused on exposure of ancient sediments found along major river valleys. Even though the reconstructions themselves were detailed, often combining data from palynological, plant macrofossil, faunal, glacial geological, and geomorphological analyses, the discontinuous depositional nature of fluvial and floodplain deposits limited the paleoenvironmental reconstructions to 'snapshots' in time. Local reconstructions were then correlated to form regional stratigraphic schemes that still shape the descriptions of Quaternary Siberia. More recently, scientists have collected ancient sediments deposited in extant lake basins, which provide continuous records of change that have helped clarify and modify the original interpretations of late Pleistocene landscapes.

This article concentrates on palynological records of northern Siberia, from the arctic islands to the southern taiga ($\sim 80^{\circ}$ – 60° N) and from Bering Strait to the polar Urals ($\sim 170^{\circ}$ W– 65° E; [Figure 1](#)). Today, tundra and forest-tundra are found to the north of $\sim 68^{\circ}$ N, except in areas bordering the Bering Strait where tundra and forest-tundra occur as far south as 60° N. Modern conifer-dominated taiga of Siberia occurs between $\sim 68^{\circ}$ and 50° N. The region is divided into three geographic areas: northeast Siberia (Bering Strait to the Verkhoyansk Range), eastern Siberia (Verkhoyansk Range to the Yenisei valley), and western Siberia (Yenisei valley to the Urals). Although investigations into the late Pleistocene have occurred in many areas of northern Siberia, palynological research has focused on the following: (1) the northern Priokhot'ye, upper Kolyma drainage, and Yana-Indigirka-Kolyma lowland of northeast Siberia; (2) the Laptev Sea region of eastern Siberia; and (3) the Ob and Yenisei drainages of the Western Siberian Plain and the Jamal Peninsula of western Siberia.

The original late Pleistocene stratigraphy was proposed by [Saks \(1953\)](#) based on sections in western Siberia, and his scheme was subsequently applied throughout Siberia ([Table 1](#)). This stratigraphy included two cool or glacial intervals: the Sartanskaya and Zyrianskaya stades, which were later considered approximate equivalents to marine isotope stage (MIS) 2 and MIS 4, respectively. Two warm intervals were also defined: the Karginski interstade ($\sim$ MIS 3) and Kazantsevskoye interglaciation ($\sim$ MIS 5). The largest alteration of Saks' scheme was the elimination of the Zyrianskaya stage in western and eastern Siberia and its replacement by the Ermakovskaya stage, which encompasses MIS 4 and MIS 5a

through MIS 5d (see [Laukhin, 2011](#)). Continuing research has shown that key stratotypes used in defining the early stratigraphic schemes are no longer valid and questions persist concerning the timing and/or characteristics of the Siberian stratigraphy (see [Astakhov and Nazarov, 2010](#); [Laukhin, 2011](#) for western Siberia and [Sher et al., 2005](#) for eastern and northeastern Siberia). Despite proposed changes to the late Pleistocene nomenclature (e.g., [Astakhov and Nazarov, 2010](#)), the newest terminology has not as yet been recognized by the Russian International Stratigraphic Commission of the Quaternary. Thus, we will continue to refer to the four-part stratigraphy, which remains the interpretive framework in much of the Russian literature. Because of discrepancies in terminology and timing between the Siberian regions, we organize the chapter subheadings by isotope stages but with reference to the regional nomenclature. Although the late glaciation (~ 12500 – 10000 ^{14}C years BP) technically incorporates the latest Pleistocene, we do not include it here because postglacial warming throughout much of northern Siberia began ~ 12500 BP. The Zyrianskaya or late Ermakovskaya stage also is not discussed, because the vegetation throughout much of northern Siberia is inferred to be the same as in the better-studied Sartanskaya stade ([Figure 2](#)). We begin with the most recent cool interval, working back through time to the last interglaciation.

MIS 2 (Sartanskaya Stade)

Northern Asia experienced an incredible change in its geography during the Last Glacial Maximum (LGM; MIS 2). With lowered global sea levels, the continent was connected to North America via the Bering Land Bridge, and Siberia extended 400–700 km farther north, connecting the arctic islands to the mainland. This glacial interval is known for its abundance of large mammal remains and a curious pollen assemblage. The juxtaposition of faunal and floral paleodata, particularly in Beringia (from the Lena River to northwestern Canada) gave rise to a single question that has motivated research for decades: was the LGM vegetation a rich steppe that supported moderate to large populations of Pleistocene megafauna, or was it a depauperate tundra or polar desert where only a small number of these herd animals could survive?

The unusual pollen spectra that characterize the LGM records are dominated by graminoids (members of the grass (Poaceae) and sedge (Cyperaceae) families) and wormwood (*Artemisia*; MIS 2 zones in [Figures 3–6](#)). This assemblage has been reported in sites from the western Siberian Plain (see [Volkova, 2002](#)), Taymyr Peninsula (see [Andreev et al., 2011](#)), Laptev coast (see [Andreev et al., 2011](#)), Lena delta (see [Andreev et al., 2011](#); [Sher et al., 2005](#)), middle Lena drainage ([Müller et al., 2010](#)), New Siberian Islands (see [Andreev et al., 2011](#)), Yana-Indigirka-Kolyma lowlands (see [Lozhkin et al., 2002a](#)),

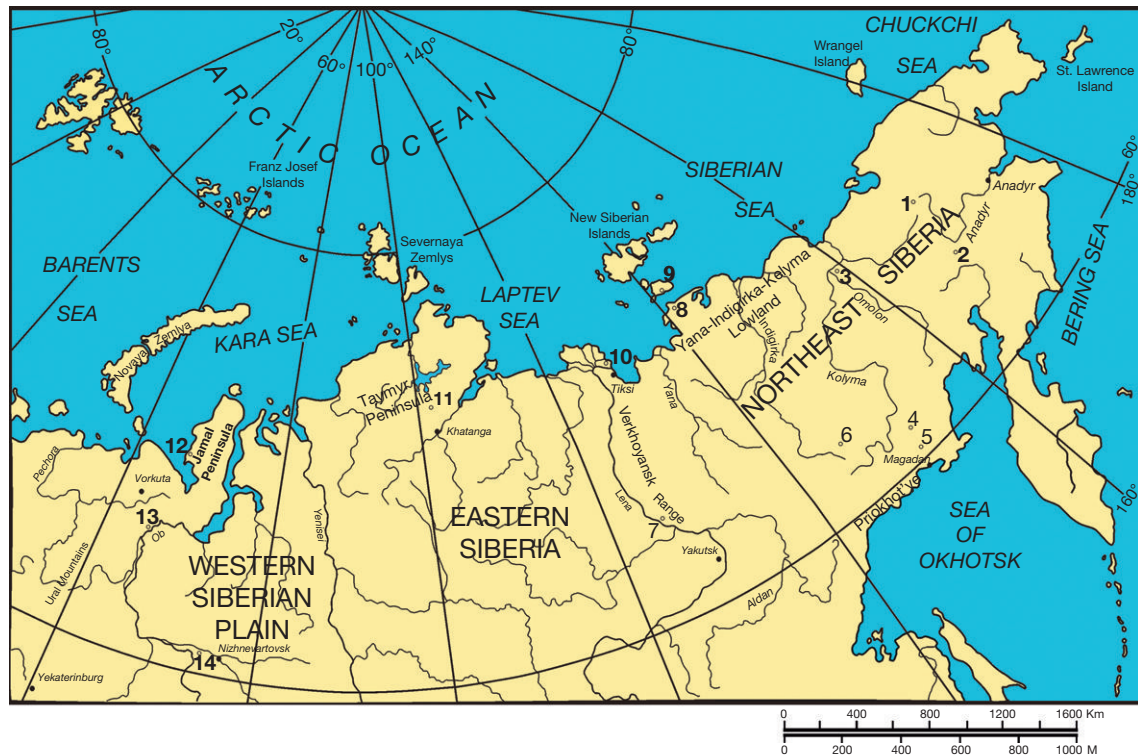


Figure 1 Map of northern Siberia showing locations named in the text. The site key is as follows: (1) Lake El'gygytyn; (2) Main river sites; (3) Duvanny Yar; (4) Lake Elikhan; (5) Lake Alut; (6) Kirgylakh mammoth site; (7) Lake Billyakh; (8) Oyogos Yar; (9) Bol'shoy Lyakhovsky Island; (10) Lena Delta sites, Mamontovy Khayata; (11) Yamal Peninsula sites; Lake Levinson-Lessing; (12) Yamal section sites; (13) Suryshkary and Zolotoy Mys sections.

Table 1 Late Pleistocene stratigraphic nomenclature, ages, and approximate marine isotope stage equivalents

Original Siberian scheme	Western Siberia	Eastern Siberia	Northeast Siberia
Sartanskaya MIS 2	Sartanskaya ~10 000–22/23 000 MIS 2	Sartanskaya ~10 300–30 000 MIS 2	Sartanskaya 12 500–26 500 MIS 2
Karginski MIS 3	Karginski ~22/23 000–50 000 MIS 3	Karginski ~30 000–50 000 MIS 3	Karginski ~26 500–60 000 MIS 3
Zyrianskaya MIS 4	Ermakovskaya ~50 000–110 000 MIS 4, MIS 5a, MIS 5b, MIS 5c, MIS 5d	Ermakovskaya or Zyrianskaya ~50 000–110 000 MIS 4, MIS 5a, MIS 5b, MIS 5c, MIS 5d	Zyrianskaya MIS 4 ~60 000–75 000
Kazantsevskoye MIS 5	Kazantsevskoye ~115 000–130 000 MIS 5e	Kazantsevskoye MIS ~115 000–130 000 MIS 5e	Kazantsevskoye ~75 000–128 000 MIS 5

Source: Saks VN (1953) *Quaternary Period in the Soviet Arctic*. Leningrad, Moscow: Vodtransizdat (in Russian); Volkova VS, Arkhipov SA, Babushkin AE, et al. (2003) *Cenozoic of Western Siberia*. Novosibirsk: GEO (in Russian); Andreev AA, Schirmermeister L, Tarasov PE, et al. (2011) Vegetation and climate history of the Laptev Sea region (Arctic Siberia) during Late Quaternary inferred from pollen records. *Quaternary Science Reviews* 30: 2182–2199; Anderson PM and Lozhkin AV (2001) The stage 3 interstadial complex (Karginskii/middle Wisconsinan interval) of Beringia: Variations in paleoenvironments and implications for paleoclimatic interpretations. *Quaternary Science Reviews* 20: 93–125; Lozhkin AV, Anderson PM, Matrosova TV, and Minyuk PS (2007) The pollen record from El'gygytyn Lake: Implications for vegetation and climate histories of northern Chukotka since the late Middle Pleistocene. *Journal of Paleolimnology* 37: 135–153.

upper Kolyma and upper Indigirka drainages (see Lozhkin et al., 2002a,b), northern Priokhot'ye (see Lozhkin et al., 2002a), and Chukotka (see Anderson et al., 2002a). Soviet scientists were the first to classify the glacial vegetation as tundra–steppe, suggesting that vast, relatively homogeneous and productive grassland covered northern Eurasia and far northwestern North America (see Yurtsev, 1982). This vegetation was also termed steppe–tundra and arctic steppe.

The LGM pollen spectra typically are associated with a faunal assemblage that is best known for the woolly mammoth (*Mammuthus primigenius*) and other large mammals, such as horse (*Equus*) and bison (*Bison*). The combination of faunal and paleobotanical data eventually gave rise to the concept of mammoth–steppe biome (see Guthrie, 1982). In the early 1980s, paleoecologists working with samples from lakes rather than sections noted that extremely low pollen accumulation

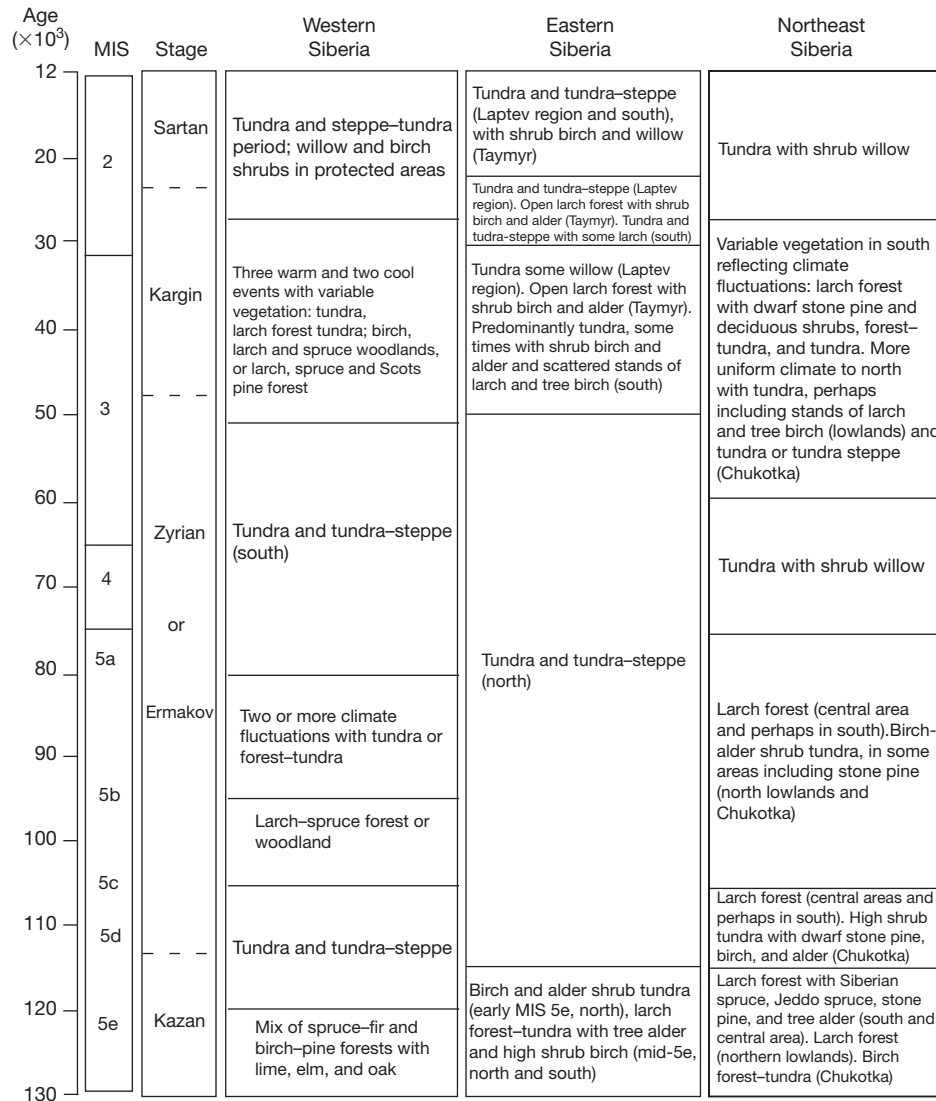


Figure 2 Vegetation of western, eastern, and northeast Siberia during the late Pleistocene. Age in the figure is in years before present $\times 10^3$.

rates and increased sedimentation in lake basins suggested unproductive landscapes, such as polar desert (see Hopkins et al., 1982). Furthermore, minor pollen taxa indicated comparatively depauperate arctic tundra rather than northern grassland. These conflicting interpretations gave rise to the 'productivity paradox': how could the late Pleistocene megafauna, which required seasonally rich vegetation growing in relatively fertile soils, survive in landscapes that were either barren or provided only limited food sources? Each side of this controversy found 'fatal flaws' in the other's data. For example, the vertebrate paleontologists noted that the major pollen taxa and many minor types represented ecologically diverse families or genera, making precise assignment to steppe or tundra impossible. The palynologists commented that poorly understood taphonomy and dating of many fossil-bone horizons prevented reasonable estimates of animal numbers or unquestionable co-occurrence of species at a site.

During the height of the productivity debate, a third line of thought developed that suggested neither extreme view was

correct. This thinking was led by Yurtsev (1982), who after further study of relict steppic communities found today in northeast and eastern Siberia, proposed that the LGM vegetation was more likely a tundra-steppe mosaic rather than a relatively uniform biome. He suggested that plant communities varied in composition and distribution based primarily on soil moisture and temperature. The most arid communities in this gradient experienced the greatest growing season warmth (thereby providing the most extensive summer thaw and soil aeration) and were most similar but not identical to zonal dry steppe. Plant associations more like those in arctic tundra occupied the other end of the gradient, occupying cool, poorly drained soils. Various grass-dominated communities (e.g., dry steppe-like meadows, mesic meadows) occurred between these extremes. Other paleoecologists, based on palynological data and/or feeding habits of the fauna, also proposed a landscape characterized by a mosaic of vegetation types (see Hopkins et al., 1982).

As further debate reached an impasse, scientists began pursuing other approaches for reconstructing the Sartanski

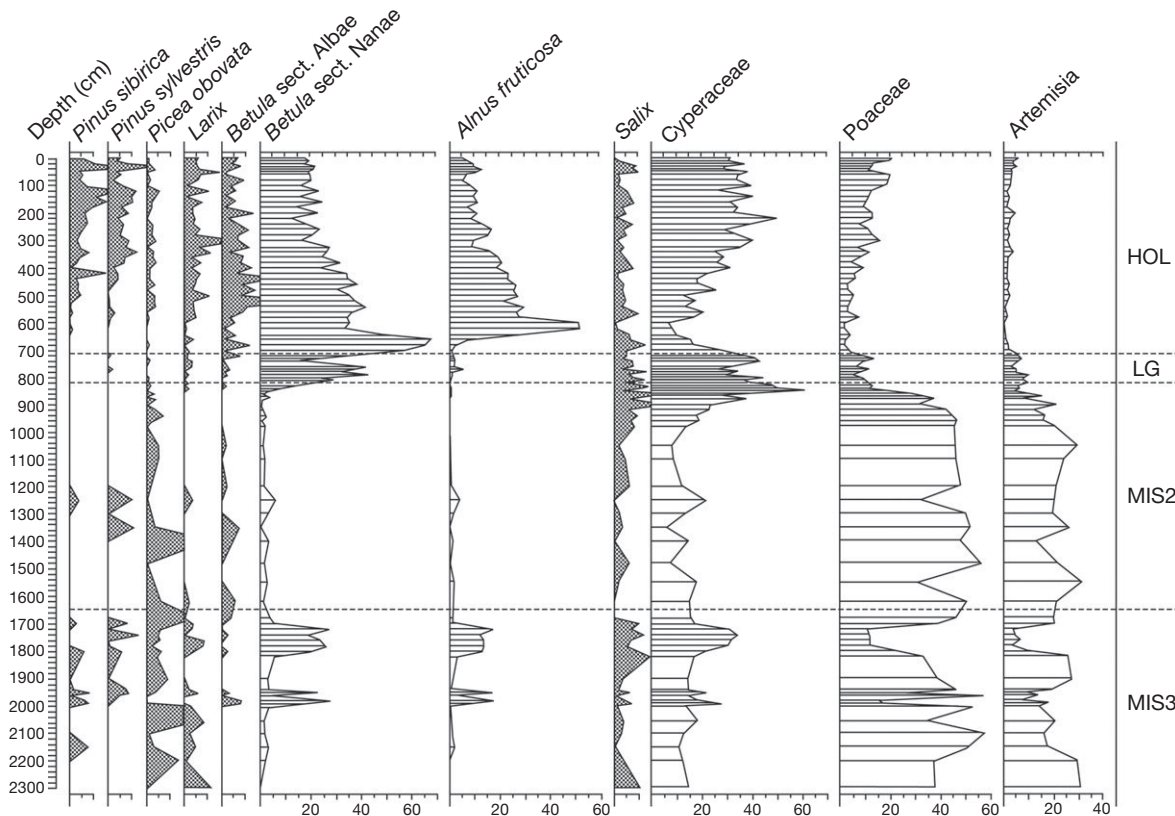


Figure 3 Pollen percentage diagram from Lake Levinson-Lessing, eastern Siberia. Data provided by A.A. Andreev (see Andreev et al., 2011). Figure shows pollen zones associated with the Holocene (HOL), late glaciation (LG), marine isotope stage 2 (MIS 2), and marine isotope stage 3 (MIS 3). High pollen percentages of shrubs (*B. sect. Nanae*, *Alnus fruticosa*) indicate shrub tundra and warmer conditions. High pollen percentages of *Cyperaceae* (sedge) and *Poaceae* (grass) indicate herb-dominated tundra or steppe-tundra and cool climate.

landscapes. Palynologists began looking for more rigorous methods that would improve upon the previous qualitative paleovegetation interpretations. For example, the collection of modern pollen samples from lakes and bogs throughout the arctic and subarctic greatly increased the ability to statistically define possible modern analogs for the MIS 2 vegetation. Comparisons using dissimilarity measurements have been done for high-latitude lakes in North America, but analogs have been calculated for only a single northern Siberian site (Lozhkin et al., 2007). Rather than indicating a no-analog landscape, as is often assumed for LGM spectra, statistics indicated that possible LGM analogs for Sartanski-age samples existed on southern and central Wrangel Island. Today, the vegetation on Wrangel Island is classified as discontinuous, herb-dominated arctic tundra, although Yurtsev (1982) noted that steppic plants are important components of some of the island's plant communities. He further suggested that these landscapes "may be the closest living approximations of the northern variants of late Pleistocene tundra-steppe" (Yurtsev, 1982, p. 172).

The development of computer models that merge an understanding of modern composition, structure, and function of arctic ecosystems with pollen-based vegetation reconstructions provide a second means for looking at LGM landscapes. These biome models use pollen data from a given time to assign plant functional types (e.g., tree vs. shrub, deciduous vs. evergreen). These types are then combined to determine a biome, the latter

being floristically distinguishable by plant and pollen composition and occupying a unique bioclimatic space (Kaplan et al., 2003). LGM biomes have been mapped for Beringia (Bigelow et al., 2003), the former Soviet Union (Tarasov et al., 2000), and the circum-arctic (Kaplan et al., 2003). Each of these experiments, including one in which the vegetation was determined using four different climate models (Kaplan et al., 2003), indicates an absence of forest and presence of various tundra types during the LGM. The mosaic nature of that tundra, including some communities with shrubby species, is particularly clear in Beringia (Bigelow et al., 2003). The indication from these analyses is that the vegetation of northern Siberia was more similar to tundra than steppe and that modern zonal steppe is not a regional analog for the Sartanskaya vegetation. In contrast, LGM biome reconstructions for Lake Billyakh (site 7, Figure 1) suggest a significant presence of both steppe and tundra communities in the southwestern foothills of the Verkhoyansk Range (Müller et al., 2010).

Another approach returned to the interdisciplinary investigations that characterized early Soviet research, with teams of Quaternary specialists gathering a broad spectrum of data (e.g., insects, plant and animal macrofossils, ostracods, cryolithology) to be used in paleoenvironmental reconstructions. Much of this research centered on the Laptev Sea region (see Andreev et al., 2011), with the Mamontovy Khayata site (site 10, Figure 1) being one of the most intensely studied (see Kienast et al.,

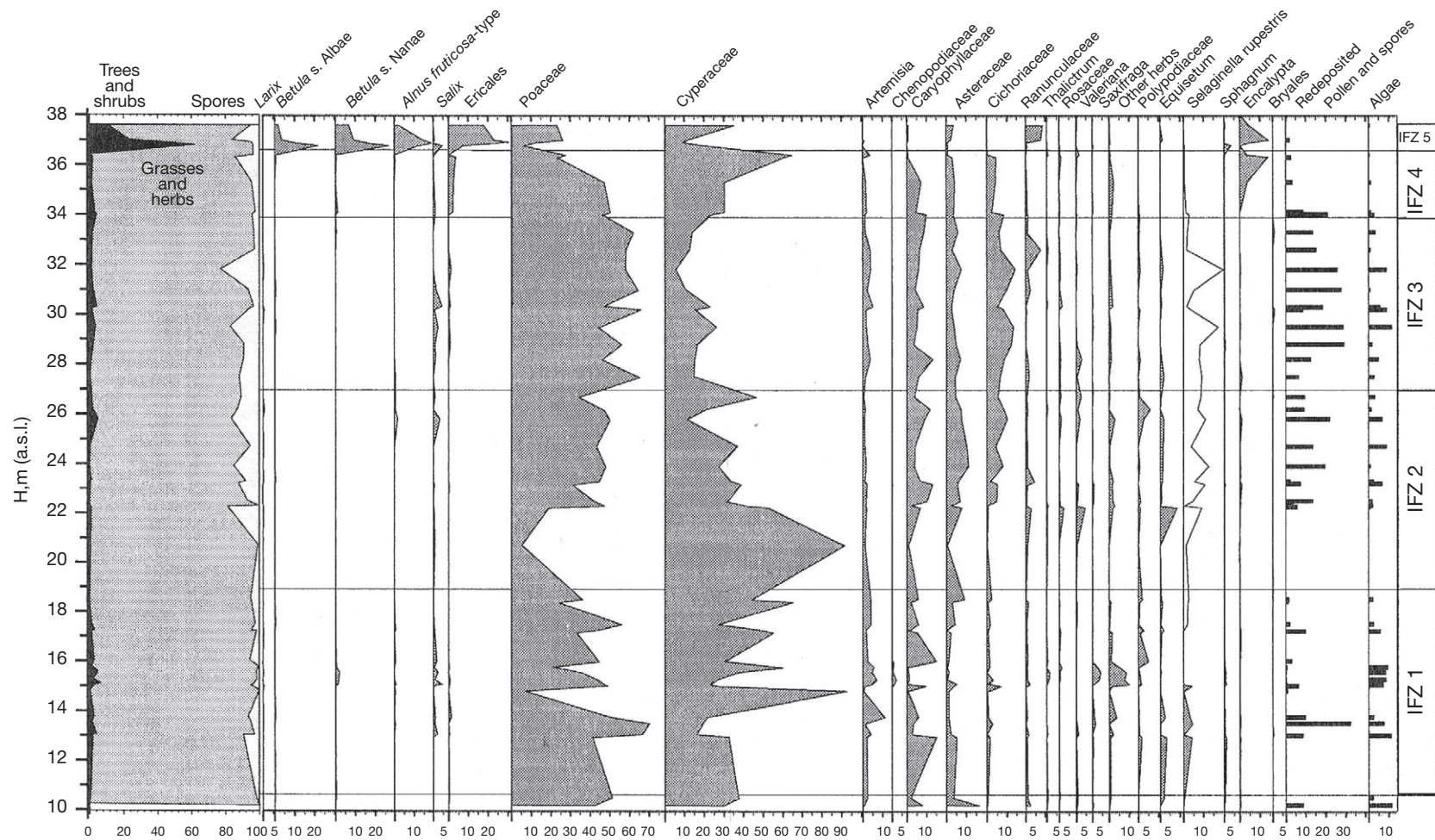


Figure 4 Summary pollen percentage diagram from Mamontovy Khayata, eastern Siberia. Data provided by A.A. Andreev (see [Andreev et al., 2011](#)). Zones correspond to marine isotope stages 1–4. High pollen percentages of grass (Poaceae) and sedge (Cyperaceae) indicate tundra or tundra steppe and cold climate. High percentages of shrubs (*B. sect. Nanae*, *Alnus fruticosa*) indicate shrub tundra and warmer conditions.

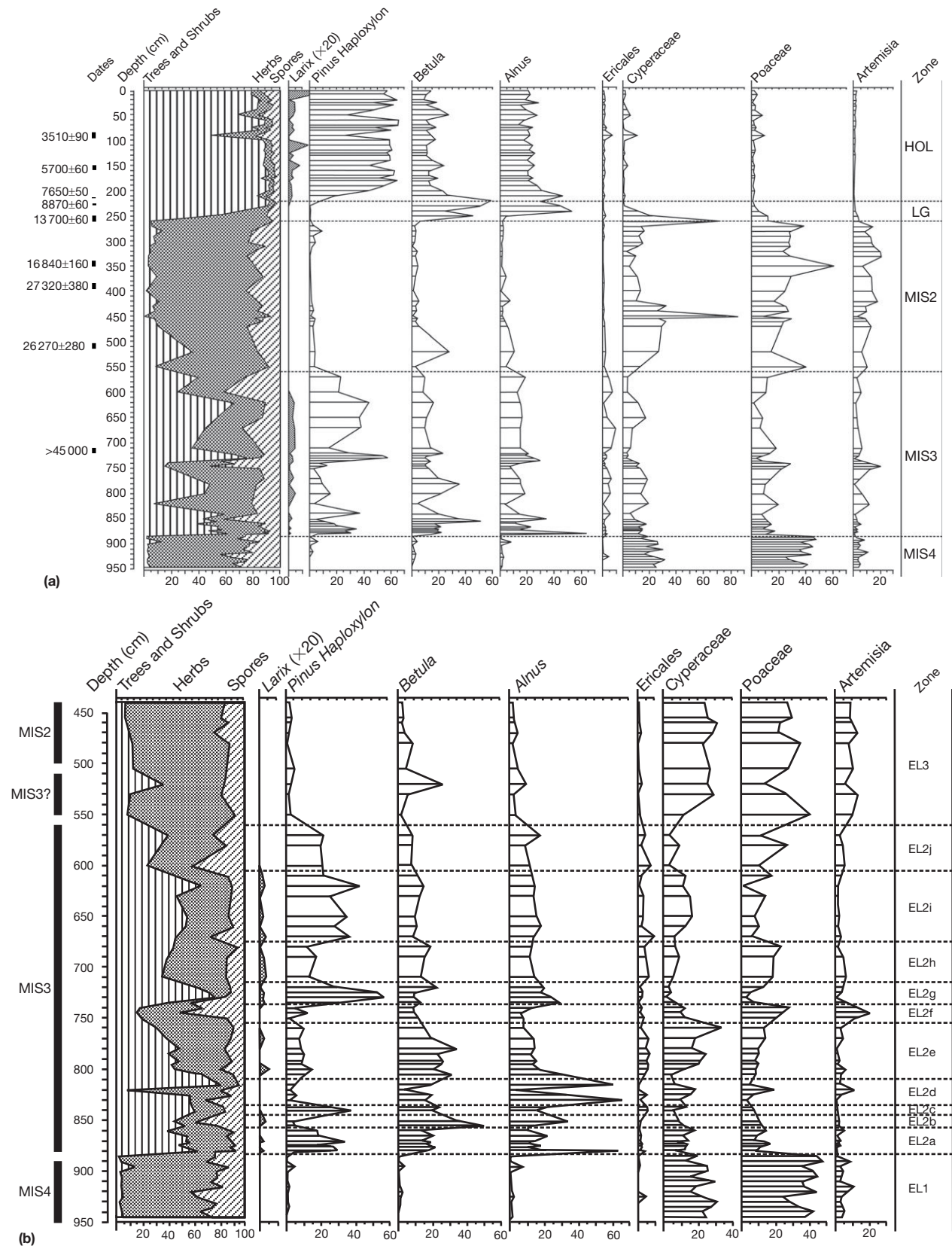


Figure 5 Pollen percentage diagram from Lake Elikchan, northeast Siberia. Modified with permission from Lozhkin and Anderson (2011) Forest or no forest: Implications of the vegetation record for climatic stability in Western Beringia during Oxygen Isotope Stage 3. *Quaternary Science Reviews* 30: 2160–2181. Figure 5(a) shows pollen zones associated with the Holocene (HOL), late glaciation (LG), marine isotope stage 2 (MIS 2), marine isotope stage 3 (MIS 3), and marine isotope stage 4 (MIS 4). Figure 5(b) illustrates fluctuations in pollen during MIS 3. Peaks in tree and shrub pollen reflect forest establishment, whereas high percentages of herb pollen indicate tundra.

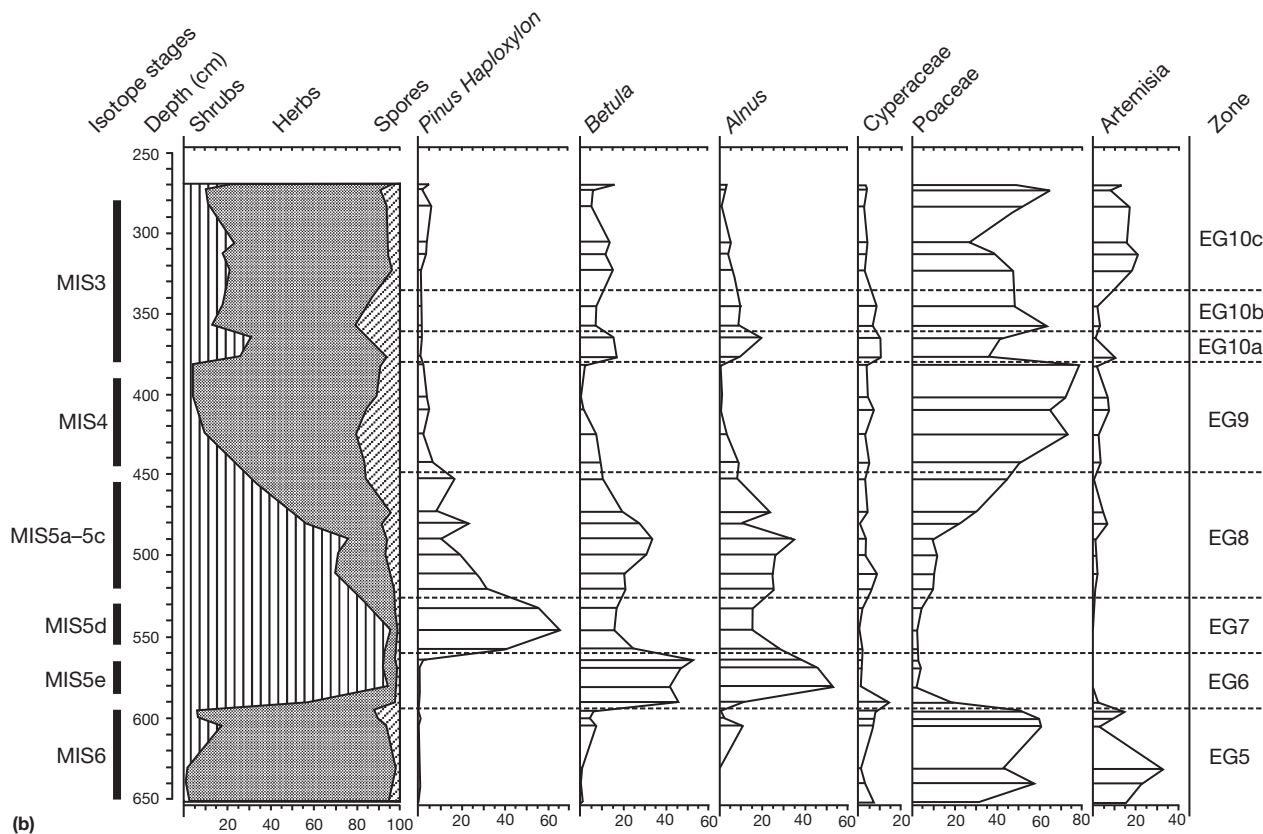
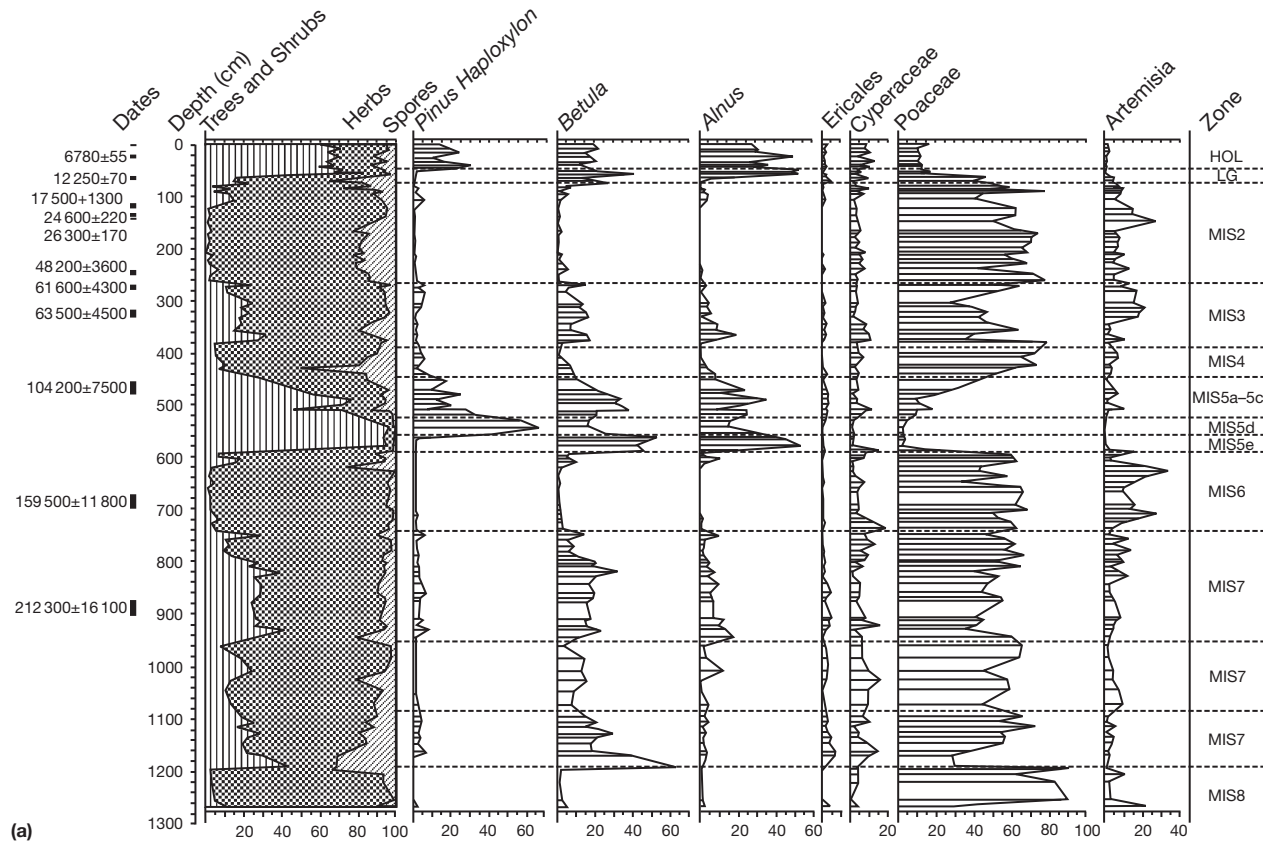


Figure 6 Pollen percentage diagram from Lake El'gygytyn, northeast Siberia. Modified with permission from Lozhkin and Anderson (2011) Forest or no forest: Implications of the vegetation record for climatic stability in Western Beringia during Oxygen Isotope Stage 3. *Quaternary Science Reviews* 30: 2160–2181. Pollen zones in Figure 6(a) have been classified as Holocene (HOL), late glaciation (LG), and marine isotope stages 2–8. Figure 6(b) shows MIS 3–MIS 6 in more detail. Herb pollen is low during warm intervals and high during cold intervals. Highest percentages of shrub pollen indicate the warmest climate.

2005; Sher et al., 2005). Pollen data from this region are typically dominated by graminoids (Figure 4; zones III and IV), with minor pollen taxa suggesting variations in microhabitats. Plant macrofossil and insect data reveal a mosaic of arctic meadow and steppe communities, with habitats varying from extremely dry to aquatic. Kienast et al. (2005) noted many similarities to relict steppes of present-day eastern Siberia. The great taxonomic diversity at Mamontovy Khayata contrasts to the floristically depauperate polar desert, suggesting high arctic vegetation is an unlikely analog for the Sartanskaya stade (Sher et al., 2005).

In yet another approach, Andreev et al. (2011) noted variations in percentages of key pollen taxa in the Laptev Sea region. Particularly, wormwood pollen, an indicator of xeric and/or disturbed habitats, increases from east to west, suggesting more arid conditions in Taymyr Peninsula as compared, for example, to Bykovsky Peninsula (Figures 3 and 4). The Taymyr plant macrofossil data, with the predominance of xerophytic species, and the more mixed-habitat plant remains from the Bykovsky sites also support this conclusion (Kienast et al., 2001, 2005). Extending this comparison eastward, pollen data from northern Chukotka (e.g., site 1, Figures 1 and 6), and southern northeast Siberia (e.g., Elikchan (site 4, Figures 1 and 5)) show wormwood pollen that generally varies between ~10% and 20%. These values are greater as compared to Bykovsky Peninsula sites (typically <10%; e.g., site 10, Figures 1 and 4) but less than on Taymyr Peninsula (≥20%; e.g., Levinson-Lessing Lake, site 11, Figures 1 and 3), suggesting that the moistest areas in the far north occurred in eastern areas of the Laptev region. However, reliance on a single pollen taxon to infer dryness can be oversimplified. For example, Kuzmina et al. (2011), working in the Main River region (site 2, Figure 1), noted that the insect assemblage, which has few thermophilous steppe species, indicates relatively moist conditions in southern Chukotka, in contrast to other areas of northern Siberia. These results support previous conclusions that central Beringia was characterized by relatively mesic environments, which possibly acted as a 'barrier' preventing the dispersal of more xeric species between the two continents (see Elias and Crocker, 2008).

In addition to the proposed east-west gradients, a third spatial pattern involving tree presence is hinted at in the paleobotanical records. Tree pollen is rare in northern Siberian records. However, the low but consistent appearance of larch (*Larix*) pollen in the Billyakh Lake diagram indicates the local establishment of the tree (Müller et al., 2010). As suggested by the biome reconstructions, tundra-steppe or tundra was the dominant vegetation, but stands of Dahurian larch (*Larix dahurica*) likely were distributed in protected valley sites, possibly throughout southeastern areas of eastern Siberia. Pollen data from western Siberia indicate forest-tundra, formed by Siberian spruce (*Picea obovata*) and Siberian larch (*Larix sibirica*) established in areas to the south of 56° N (see Volkova, 2002). These data suggest a zonal distribution of tundra/tundra-steppe to the north and forest-tundra to the south during the LGM. However, plant macrofossil studies indicate that larch and spruce were present at least in small populations on southern Taymyr Peninsula ~18 000–19 000 years ago (Binney et al., 2009), raising the possible existence of refugial forest communities in the far north. These cryptic refugia in the Siberian lowlands possibly acted as seed sources for the repopulation of larch which established in the northern interior during the Holocene (Anderson et al., 2002b).

The above approaches bring much needed new information and analyses to the 'productivity' debate, even though they resolve the past environments on differing spatial scales. That is, models relying on palynological data emphasize broad vegetation types, whereas multidisciplinary studies provide insights into relatively local landscapes. Nonetheless, points of agreement do exist. The Sartanskaya vegetation was a mosaic of herb-dominated communities that contained elements of arctic tundra and steppe. In the most severe areas, vegetation was likely discontinuous. Shrubs, particularly willow (*Salix*) and birch (*Betula* sect. *Nanae*), were present but restricted to more mesic and protected sites. Small populations of larch may have survived the LGM in southern areas of northern Siberia or in cryptic refugia in the northern lowlands. The mosaic quality of the vegetation was apparently independent of topography, because it characterized both the Siberian lowlands and more mountainous areas to the south. As Schweger et al. (1982) noted, the survival of the late Pleistocene megafauna was perhaps more dependent on a mosaic of grazing habitats rather than the occurrence of a single, highly productive, uniform habitat.

MIS 3 (Karginski Interstade)

Although the Karginski interstade was defined by Saks in the 1950s, it was only some 20 years later that Kind (1974) formalized the sequences of interstadial climatic changes into three warm and two cool events (Table 2). While many

Table 2 Examples of Karginski interstadial divisions

Age (years BP × 10 ³)	Subhorizon name	Relative climate within Karginski interstade
Kind Siberian Scheme		
22–30	Lipovskoy–Novoselovskii	Warm
30–33	Konotzelski	Cool
33–43	Malokhetskii	Maximum warmth
43–45	unnamed	Cool
45–50	unnamed	Warm
Western Siberia		
24–29	Verkhnelobanovskii	Warm
30–34	Lokhpodgortskii	Cool
35–41	Zolotomysski	Warm
42–43	Kiriasski	Cool
44–50	Shuryshkarski	Warm
Northeast Siberia		
27–31	Kuranakh–Salinski	Warm
31–34	Konotzelski	Cool
34–40	Malokhetskii	Warm
40–44	Kirgilyakhskii	Cool
44–60	Elikchan 4	Warm

Source: Kind NV (1974) *Late Quaternary Geochronology According to Isotope Data*. Moscow: Nauka (in Russian); Arkhipov SA and Volkova VS (1994) *Geologic History, Landscapes, and Climates of the Pleistocene in West Siberia*. Novosibirsk: RAS Siberian Branch (in Russian), including original subhorizons; see also Volkova VS, Arkhipov SA, Babushkin AE, Kulkova IA, Guskov SA, Kuzmina OB, et al. (2003) *Cenozoic of Western Siberia*. Novosibirsk: GEO (in Russian); Anderson PM and Lozhkin AV (2001) The stage 3 interstadial complex (Karginskii/middle Wisconsinan interval) of Beringia: Variations in paleoenvironments and implications for paleoclimatic interpretations. *Quaternary Science Reviews* 20: 93–125; Lozhkin and Anderson (2011) Forest or no forest: Implications of the vegetation record for climatic stability in Western Beringia during Oxygen Isotope Stage 3. *Quaternary Science Reviews* 30: 2160–2181.

Siberian paleoecologists accepted this stratigraphic framework, others have long doubted its validity. V.I. Astakov and A.V. Sher were two of the most vocal opponents, with arguments centering on the following: (1) the accuracy of age assignments of the Karginski deposits and (2) whether interstadial environments were significantly warmer than during MIS 2. Astakov and colleagues returned to key western Siberian sections to redate the basic sediment stratigraphy, using ^{14}C , luminescence, and Th/U techniques (see Astakhov and Nazarov, 2010). They found that the original stratotype for the Karginski interstade was invalid, because the 'interstadial' units actually dated to $\sim 130\,000$ years ago (i.e., MIS 5e). This finding supported conclusions drawn by Sher and colleagues, who, based on paleoecological data from well-dated sites in the northern lowlands, inferred that conditions varied little between MIS2 and MIS3, with sites previously assigned to the Karginskaya stage being more likely of last interglacial age (see Sher et al., 2005). Additional investigations into the timing and nature of the late Pleistocene in western Siberia, headed by Laukhin (2011), agreed that the original Karginski stratotype was invalid, but other dated sections indicated Kind's scheme to be basically accurate. Lozhkin and Anderson (2011) in a summary of MIS 3 pollen data from western Beringia proposed that aspects of both Kind's and Sher's arguments were correct with interstadial variability characterizing southern areas and with more stable, stadial-like environments in the north. As suggested by these discussions and outlined in the following vegetation history (Figure 2), the Karginski interstade appears to be distinguished by complex spatial and temporal patterns that set it apart from stadial environments.

From the earliest studies, palynological research in northeast Siberia documented spatial and temporal variability in MIS 3 vegetation and climate (see Anderson and Lozhkin, 2001). Although the age assignments of some sites to the Karginski interstade have been questioned (see Sher et al., 2005), the Elikchan (site 4 Figures 1 and 5) and Alut (site 5, Figure 1) lake records confirm the instability of interstadial vegetation with forests likely expanding and contracting rather rapidly, at least in southwestern areas of the region (Lozhkin and Anderson, 2011). The Elikchan record suggests that MIS 3 experienced two warm (zones EL2a–EL2c; EL2g–EL2i), two cool (EL2d, EL2f), and two moderate (EL2e, EL2j) intervals with vegetation shifting from near modern (i.e., Dahurian larch with dwarf stone pine, *Pinus pumila*) to shrub tundra. The timing of the paleoclimatic changes is still problematic (Lozhkin and Anderson, 2011) with contradictions occurring between lakes and with some of the better-dated sections. For example, the Kirgilyakhski interval, defined from the Kirgilyakh mammoth site (site 6, Figure 1), indicates cool conditions from $\sim 40\,000$ to $44\,000$ BP (Anderson and Lozhkin, 2001). Such an interpretation is in agreement with the Alut record but corresponds to a warm interval at Elikchan Lake. The southern lake and section data do concur that the interstadial climatic optimum happened in mid-Karginski times, as was suggested by Kind (Table 2). During this interval, vegetation and climate are inferred to be like modern (cf. zone EL2g, Figure 5(b) to zone HOL, Figure 5(a)).

In Chukotka, El'gygytyn Lake (site 1, Figures 1 and 6) provides the most continuous vegetation history for MIS 3, although the level of analysis is not comparable to that from the southern lakes (Lozhkin and Anderson, 2011).

Shrub tundra persisted in the El'gygytyn region throughout the interstade. However, relatively high percentages of alder (*Alnus*) pollen suggest that summers were somewhat warmer $\sim 54\,000$ – $60\,000$ BP (zone EG10a, Figure 6(b)) as compared to $\sim 46\,000$ – $27\,000$ BP (zone EG10c). A similar shift from relatively warm to cooler conditions is recorded in several Main River sites in southern Chukotka (site 2, Figure 1), with indications of scattered larch stands during warmer intervals (Anderson et al., 2002a). The northern establishment of Dahurian larch–birch woodlands or forest–tundra in the Yana–Indigirka–Kolyma lowlands was inferred from pollen data (Anderson and Lozhkin, 2001), although site assignment to the Karginski interval subsequently was questioned (see Sher et al., 2005). However, the recent discovery of larch trunks in a peat horizon dated to $\sim 40\,000$ – $43\,000$ BP at the Duvanny Yar site (site 3, Figure 1; Zanina et al., 2011) suggests that the initial age assignments and paleovegetation interpretations are correct. Additional paleobotanical data from Duvanny Yar support the presence of open Dahurian larch forest with alder, birch, and willow shrubs, implying a climate similar to modern at this time. Between $\sim 35\,000$ and $38\,000$ BP, moist tundra was dominant in the lowlands, suggesting cooler and/or drier conditions.

Eastern portions of the eastern Siberia lowlands were dominated by treeless vegetation during MIS 3 (see Andreev et al., 2011; Figure 4), although low percentages (2–9%) of forest–tundra insects indicate the existence of small pockets of Dahurian larch (Sher et al., 2005). Numerous ^{14}C -dated sites record the dominance of tundra–steppe or tundra with a mosaic of graminoid meadows, steppic and disturbed habitats, and willow thickets in protected settings. The monolithic nature of the MIS 3 and MIS 2 environments, particularly as documented from Bykovsky Peninsula (site 10, Figure 1), confirmed to Sher and colleagues that interstadial conditions were essentially stable across northern Siberia and that the environmental fluctuations proposed by Kind were in error (Sher et al., 2005). However, they also noted that several bioindicators suggested somewhat warmer summer climates between $32\,000$ and $42\,000$ BP, giving some small degree of variability to the interstade (Andreev et al., 2011; Kienast et al., 2005; Sher et al., 2005).

In contrast to the Lena delta region, open Dahurian larch forests with shrub birch and alder established in the Taymyr lowland (site 11, Figures 1 and 3) from $\sim 48\,000$ to $25\,000$ BP, suggesting a climate that was warmer and wetter than modern (see Andreev et al., 2011). However, discontinuous steppic vegetation characterized northern areas of the Peninsula from $\sim 32\,000$ BP throughout MIS 2. Herb-dominated communities also prevailed during the interstade in southeastern eastern Siberia. The Lake Billyakh record (site 7; Figure 1) suggests modest climatic differentiation, as increases in Dahurian larch and shrub birch hint at somewhat ameliorated conditions prior to $\sim 31\,000$ BP (Müller et al., 2010). Evidence for sustained warm periods is absent, although brief and relatively warm events occurred $\sim 47\,500$, $\sim 38\,000$, and $\sim 33\,000$ BP.

Despite reassigning the original Karginski stratotype to MIS 5e, additional research in the Western Siberian Plain suggests much of the original paleoenvironmental history is valid (see Arkhipov and Volkova, 1994). For example, the Shuryshkary sections (site 13; Figure 1) contain 'warm' deposits occurring stratigraphically between stadial sediments and above a last

interglacial horizon (see [Laukhin, 2011](#)). Paleobotanical data revealed variable interstadial conditions as represented by fluctuating vegetation types: (1) forest–tundra; (2) north taiga open woodland (tree birch (*Betula* sect. *Albae*), Siberian spruce, Scots pine (*Pinus sylvestris*), and Siberian stone pine (*Pinus sibirica*)); (3) southern north taiga; (4) south taiga (climatic optimum; Scots pine, Siberian spruce, Siberian fir (*Abies sibirica*), tree birch); (5) dark coniferous woodland; (6) forest–tundra; (7) tundra; and (8) forest–tundra ([Volkova et al., 2003](#)). This scheme is more complex than Kind’s original stratigraphy and is more in line with the numerous vegetation fluctuations in southern northeast Siberia. In contrast, data from Yamal Peninsula (site 12, [Figure 1](#)) record relatively stable climate as indicated by persistent graminoid tundra between ~33 000 and 25 000 BP ([Andreev et al., 2006](#)). This stable pattern has a greater affinity to northern eastern Siberia.

MIS 5 (Mid- and Lower Ermakovski and Kazantsevskoye Stages, Western and Eastern Siberia; Kazantsevskoye Stage, Northeast Siberia)

As compared to other late Pleistocene stages, little controversy exists concerning the paleoenvironments of the last interglaciation. Although stratigraphic issues are not completely

resolved (e.g., [Astakhov and Nazarov, 2010](#); [Laukhin, 2011](#)), there is a general consensus that modern vegetation zones moved hundreds of km northward ([Figure 2](#)) and that climate was significantly warmer than the present during the climatic optimum.

Most of northeast Siberia was dominated by forest during MIS 5e ([Lozhkin and Anderson, 1995](#)). As compared to the modern vegetation, the interglacial forests of the Magadan region contained greater numbers of tree birch and temperate trees (Jeddo spruce (*Picea ajanensis*), fir (*Abies*), hazel (*Corylus*), tree alder), representing range extensions of up to 1000 km. Like today, the interior supported Dahurian larch forest, but this forest included Siberian spruce and tree alder (*Alnus hirsuta*). The presence of Siberian spruce in the upper Kolyma and Indigirka drainages suggest range extensions of ~700–800 km. Dahurian larch migrated ~600 km northward, eliminating coastal tundra and a similar distance eastward into areas of southern Chukotka. Birch forests also established in parts of Chukotka, representing northward range extensions of ~1000 (*Betula ermanii*) and ~600 km (*Betula cajanderi*). El’gygytyn Lake (site 1, [Figure 1](#)) is the only continuous record of paleoenvironmental change for all of northern Siberia ([Lozhkin et al., 2007](#)) and provides some of the best documentation of vegetation and climate change within MIS 5. For example, El’gygytyn pollen data indicate an increase in dwarf stone

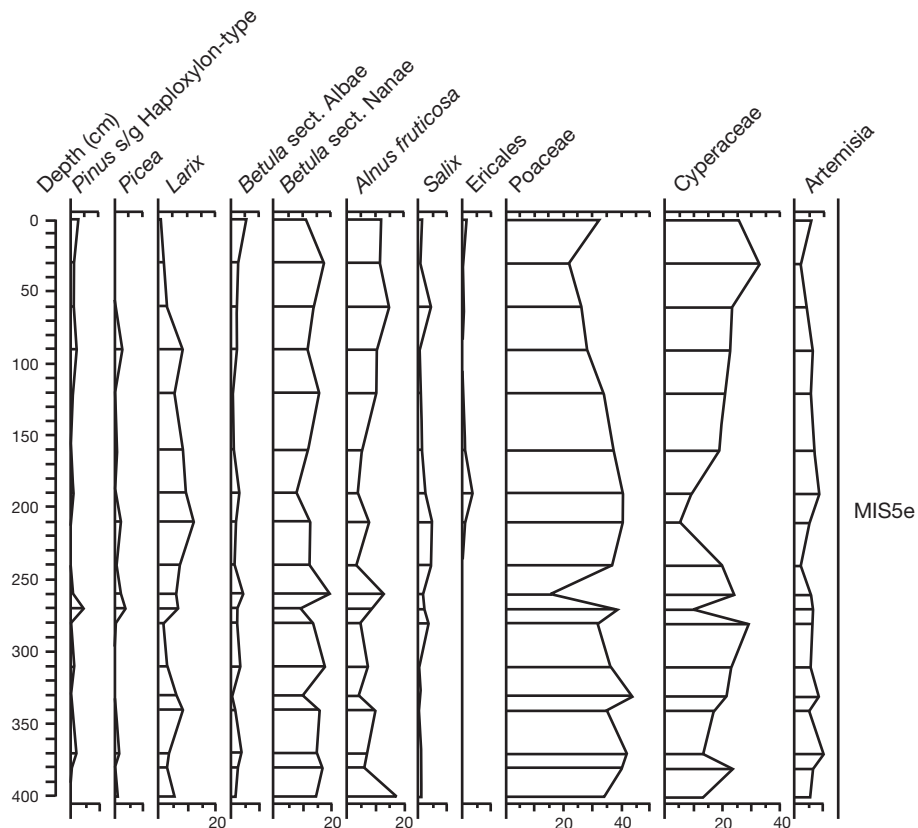


Figure 7 Summary pollen percentage diagram from Oyogos Yar, eastern Siberia, showing the climatic optimum of the last interglaciation (marine isotope stage 5e). Data provided by A.A. Andreev (see [Andreev et al., 2011](#)). The high percentages of larch (*Larix*), which is an underrepresented pollen type, suggest the establishment of larch forest or forest–tundra and warmer than present climate.

pine during MIS 5d (Figure 6; zone EG 7), suggesting greater winter precipitation as compared to the climatic optimum (zone EG 6). Substages 5a–5c (zone EG 8) are indistinguishable, but plant communities and thus climate appear similar to modern.

In eastern Siberia, shrub tundra with shrub birch and alder established on Bol'shoy Lyakhovsky Island (site 9, Figure 1) during the early and middle portions of the climatic optimum (see Andreev et al., 2011). Plant macrofossils for this interval consisted mainly of subarctic shrub taxa (Kienast et al., 2008). At the same time, paleobotanical and insect data indicate the presence of larch forest or forest–tundra with shrub alder and birch, representing a ~270-km northward displacement (see Andreev et al., 2011). Pollen data from the Oyogos Yar site (site 8, Figures 1 and 7) confirm the presence of larch during MIS 5e. Scattered stands of tree birch may also have been present. Additionally, forests established in the western Laptev Sea region and in the middle Lena drainage (Andreev et al., 2011).

Although Astakhov and Nazarov (2010) suggested that most western Siberian sites previously assigned to the Kazantsevskoye interglaciation are more likely of middle Pleistocene age, recent studies have confirmed the occurrence of interglacial sites in the region (Laukhin, 2011; e.g., Shuryshkary section, lower Zolotoy Mys, site 13, Figure 1). In the western Siberian Plain, MIS 5e is marked by the northward displacement of all vegetation zones by as much as 700 km (Arkipov and Volkova, 1994; Volkova et al., 2003; Figure 2). Dark coniferous forest (e.g., Siberian spruce, Siberian fir, Siberian stone pine) occupied the northern plain, extending to the current coastline and eliminating arctic tundra and forest–tundra landscapes. Southern areas of this conifer-dominated taiga included broadleaf species (e.g., linden (*Tilia*), elm (*Ulmus*), oak (*Quercus*), hazel). Today, oak and hazel are not found east of the Urals, indicating a ~2000-km range extension (Volkova et al., 2003). Some variability may have occurred during MIS 5e (Laukhin, 2011), but additional data are required to confirm these trends. MIS 5d is marked by a dramatic vegetation change to tundra or tundra–steppe, whereas MIS 5c is characterized by Siberian larch–Siberian spruce forest or open woodland. The latest vegetation of late MIS 5 was birch forest–tundra.

See also: Plant Macrofossil Introduction; Pollen Analysis, Principles. **Beetle Records:** Late Pleistocene of Northern Asia. **Glacial Climates:** Biosphere Feedbacks. **Glaciations:** Late Pleistocene Glacial Events in Beringia. **Introduction:** Understanding Quaternary Climatic Change. **Paleoclimate:** Introduction; Paleoclimate History of the Arctic; Timescales of Climate Change. **Paleoclimate Modeling:** Data–Model Comparisons; Last Glacial Maximum GCMs; Quaternary Environments; The Last Interglacial. **Paleoclimate Reconstruction:** Approaches. **Plant Macrofossil Records:** Arctic Eurasia. **Pollen Methods and Studies:** Changing Plant Distributions and Abundances; Databases and Their Application; The Biome Approach to Reconstructing Past Vegetation; Use of Pollen as Climate Proxies. **Pollen Records, Late Pleistocene:** Northern North America. **Pollen Records, Postglacial:** Northern Asia. **Sea-Levels, Late Quaternary:** Late Quaternary Relative Sea-Level Changes in High Latitudes. **Vertebrate Records:** Late Pleistocene Megafaunal Extinctions.

References

- Anderson PM, Kotov AN, Lozhkin AV, and Trumpe MA (2002a) Palynological and radiocarbon data from late Quaternary deposits of Chukotka. In: Anderson PM and Lozhkin AV (eds.) *Late Quaternary Vegetation and Climate of Siberia and the Russian Far East (Palynological and Radiocarbon Database)*, pp. 35–79. Washington, DC: NOAA; and Magadan: NESF FEB RAS (in English and Russian).
- Anderson PM and Lozhkin AV (2001) The stage 3 interstadial complex (Karginskii/middle Wisconsinan interval) of Beringia: Variations in paleoenvironments and implications for paleoclimatic interpretations. *Quaternary Science Reviews* 20: 93–125.
- Anderson PM, Lozhkin AV, and Brubaker LB (2002b) Implications of a 24,000-yr palynological record for a Younger Dryas cooling and for boreal forest development in Northeastern Siberia. *Quaternary Research* 57: 325–333.
- Andreev AA, Forman SL, Ingolfsson O, and Manley WF (2006) Middle Weichselian environments on western Yamal Peninsula, Kara Sea based on pollen records. *Quaternary Research* 65: 275–281.
- Andreev AA, Schirmermeister L, Tarasov PE, et al. (2011) Vegetation and climate history of the Laptev Sea region (Arctic Siberia) during Late Quaternary inferred from pollen records. *Quaternary Science Reviews* 30: 2182–2199.
- Arkipov SA and Volkova VS (1994) *Geologic History, Landscapes, and Climates of the Pleistocene in West Siberia*. Novosibirsk: RAS Siberian Branch (in Russian).
- Astakhov V and Nazarov D (2010) Correlation of upper pleistocene sediments in northern West Siberia. *Quaternary Science Reviews* 29: 3615–3629.
- Bigelow NH, Brubaker LB, Edwards ME, et al. (2003) Climate change and Arctic ecosystems: 1 Vegetation changes north of 55° N between the last glacial maximum, mid-Holocene, and present. *Journal of Geophysical Research* 108(D19) Article No. 1870, Alt11–1–Alt11–25.
- Binney HA, Willis KJ, Edwards ME, et al. (2009) The distribution of late-quaternary woody taxa in northern Eurasia: Evidence from a new macrofossil database. *Quaternary Science Reviews* 28(23–24): 2445–2464.
- Elias SA and Crocker B (2008) The Bering Land Bridge: A moisture barrier to the dispersal of steppe–tundra biota? *Quaternary Science Reviews* 27: 2473–2483.
- Guthrie RD (1982) Mammals of the mammoth steppe as paleoenvironmental indicators. In: Hopkins DM, Matthews JV Jr., Schweger CE, and Young SB (eds.) *Paleoecology of Beringia*, pp. 307–326. New York: Academic Press.
- Hopkins DM, Matthews JV Jr., Schweger CE, and Young SB (eds.) (1982) *Paleoecology of Beringia*. New York: Academic Press.
- Kaplan JO, Bigelow NH, Prentice IC, et al. (2003) Climate change and arctic ecosystems: 2 Modeling, paleodata-model comparisons, and future projections. *Journal of Geophysical Research* 108(D19) 8171 Alt 12–1–Alt 12–17.
- Kienast F, Schirmermeister L, Siegart C, and Tarasov P (2005) Palaeobotanical evidence for warm summers in the East Siberian Arctic during the last cold stage. *Quaternary Research* 63: 283–300.
- Kienast F, Schirmermeister L, Tarasov P, and Andreev AA (2008) Continental climate in the East Siberian Arctic during the Last Interglacial: Implications from palaeobotanical records. *Global and Planetary Change* 60: 535–562.
- Kienast F, Siegart C, Dereviagin A, and Mai DH (2001) Climatic implications of late Quaternary plant macrofossil assemblages from the Taymyr Peninsula, Siberia. *Global and Planetary Change* 31: 265–281.
- Kind NV (1974) *Late Quaternary Geochronology According to Isotope Data*. Moscow: Nauka (in Russian).
- Kuzmina SA, Sher AV, Edwards ME, et al. (2011) The late Pleistocene environment of the Eastern West Beringia based on the principal section at the Main River, Chukotka. *Quaternary Science Reviews* 30: 2091–2123.
- Laukhin SA (2011) 'Warm' stages in the West Siberian Late Pleistocene. *Quaternary International* 241: 51–67.
- Lozhkin AV and Anderson PM (1995) The last interglaciation in northeast Siberia. *Quaternary Research* 43: 147–158.
- Lozhkin AV and Anderson PM (2011) Forest or no forest: Implications of the vegetation record for climatic stability in Western Beringia during Oxygen Isotope Stage 3. *Quaternary Science Reviews* 30: 2160–2181.
- Lozhkin AV, Anderson PM, Matrosova TV, and Minyuk PS (2007) The pollen record from El'gygytyn Lake: Implications for vegetation and climate histories of northern Chukotka since the late Middle Pleistocene. *Journal of Paleolimnology* 37: 135–153.
- Lozhkin AV, Anderson PM, and Trumpe MA (2002a) Palynological records and radiocarbon data from late Quaternary deposits of the Yana-Kolyma lowland, the Kolyma basin, and Northern Priokhot'ye. In: Anderson PM and Lozhkin AV (eds.) *Late Quaternary Vegetation and Climate of Siberia and the Russian Far East*

- (*Palynological and Radiocarbon Database*), pp. 80–163. Washington, DC: NOAA; and Magadan: NESF FEB RAS (in English and Russian).
- Lozhkin AV, Grinenko OV, Anderson PM, and Trumpe MA (2002b) Paleobotanical characteristics and radiocarbon data from late Quaternary deposits of the Indigirka and Yana River basins. In: Anderson PM and Lozhkin AV (eds.) *Late Quaternary Vegetation and Climate of Siberia and the Russian Far East (Palynological and Radiocarbon Database)*, pp. 164–195. Washington, DC: NOAA; and Magadan: NESF FEB RAS (in English and Russian).
- Müller S, Tarasov PE, Andreev AA, Tütken T, Gartz S, and Diekmann B (2010) Late Quaternary vegetation and environments in the Verkhoyansk Mountains region (NE Asia) reconstructed from a 50-kyr fossil pollen record from Lake Billyakh. *Quaternary Science Reviews* 29: 2071–2086.
- Saks VN (1953) *Quaternary Period in the Soviet Arctic*. Leningrad, Moscow: Vodtransizdat (in Russian).
- Schweger CE, Matthews JV Jr., Hopkins DM, and Young SB (1982) Paleoeecology of Beringia – A synthesis. In: Hopkins DM, Matthews JV Jr., Schweger CE, and Young SB (eds.) *Paleoeecology of Beringia*, pp. 425–444. New York: Academic Press.
- Sher AV, Kuzmina SA, Kuznetsova TV, and Sulerzhitsky LD (2005) New insights into the Weichselian environment and climate of the East Siberian Arctic, derived from fossil insects, plants, and mammals. *Quaternary Science Reviews* 24: 533–569.
- Tarasov PE, Volkova VS, Webb T, et al. (2000) Last glacial maximum biomes reconstructed from pollen and plant macrofossil data from northern Eurasia. *Journal of Biogeography* 27(3): 609–620.
- Volkova VS (2002) Palynological characteristics and radiocarbon data of Late Quaternary deposits of the Western Siberian Plain. In: Anderson PM and Lozhkin AV (eds.) *Late Quaternary Vegetation and Climate of Siberia and the Russian Far East (Palynological and Radiocarbon Database)*, pp. 196–256. Washington, DC: NOAA; and Magadan: NESF FEB RAS (in English and Russian).
- Volkova VS, Arkhipov SA, Babushkin AE, et al. (2003) *Cenozoic of Western Siberia*. Novosibirsk: GEO (in Russian).
- Yurtsev BA (1982) Relicts of the xerophyte vegetation of Beringia in northeastern Asia. In: Hopkins DM, Matthews JV Jr., Schweger CE, and Young SB (eds.) *Paleoeecology of Beringia*, pp. 157–178. New York: Academic Press.
- Zanina OG, Gubin SV, Kuzmina SA, Maximovich SV, and Lopatina DA (2011) Late-Pleistocene (MIS 3-2) palaeoenvironments as recorded by sediments, palaeosols, and ground-squirrel nests at Duvanny Yar, Kolyma lowland, northeast Siberia. *Quaternary Science Reviews* 30: 2107–2123.

Northern North America

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Introduction

This entry summarizes our current knowledge of late Quaternary pollen-based vegetation reconstructions in northern North America, north of about 60°N.

The vegetation of northern North America is primarily boreal forest and tundra. In most areas, the boreal forest is dominated by conifers (depending on locality, spruce, pine, larch, or toward the south, fir), although in some areas, birch or aspen may be important (Elliott-Fisk, 2000). The tundra vegetation varies depending on climate and substrate. Graminoid meadows or tussock tundra are common on water-saturated landscapes, while shrubs are found in better-drained settings. Shrubs become progressively shorter and less frequent toward the north. In the northernmost Arctic, shrubs are excluded entirely, and the vegetation is a sparse covering of grasses, mosses, lichens, and liverworts (CAVM Team, 2003).

Regionally, the northern limit of trees is controlled by climate, but on a local scale, factors such as aspect and drainage exert a strong influence. Spruce and pine are the dominant treeline species across the continent, though stands of cottonwood north of the treeline are present in Alaska (Viereck and Little, 1975). Treeline in central Canada is a gradational boundary, with a wide swath of lichen woodland and forest-tundra separating the closed forest from the tundra (Figure 1) (Elliott-Fisk, 2000).

This article covers the time period from the last interglaciation (ca. 125 ka) to the late Holocene. All ages are written in kiloyears and abbreviated as 'ka' (thousands of years ago). All radiocarbon dates younger than about 21 ¹⁴C ka have been calibrated by Calib ver. 5.0.1 (Stuiver and Reimer, 1993) and are written as 'ka.' Ages from other dating methods (e.g., luminescence) are assumed to correlate with the calibrated radiocarbon timescale and are also written as 'ka.'

Summary of Pollen Records in Northern North America

Alaska and Yukon

Marine Isotope Stage 5 (MIS 5 (ca. 128–73 ka)) is a complex period of warm (MIS 5e, 5c, 5a) and cool (5d and 5b) episodes. The entire interval is colloquially known as the last interglaciation, although in the Arctic, probably only the climate at onset (MIS 5e) was similar or warmer than today (CAPE Last Interglacial Project Members, 2006). MIS 5-aged sites are primarily located in areas that were unglaciated during the last Ice Age (MIS 2), such as the northern half of Alaska, the northern Yukon, and unglaciated uplands of Baffin Island (Figure 1). In western Alaska, lake cores (e.g., Squirrel Lake (Figure 2)) and peat deposits (Edwards et al., 2003; Matthews, 1974a) indicate the westward and northward expansion of

spruce. Buried near-shore marine deposits on Togiak Bay (Figure 2) in southwest Alaska (CAPE Last Interglacial Project Members, 2006; Kaufman et al., 2001) also probably document spruce expansion. Spruce distribution in the north is a function of summer climate, indicating that in the north and west, at least, summers were warmer than they are today. In interior Alaska, pollen records from several sites are not significantly different from today's, except that ferns (Polypodiaceae) were apparently more abundant, suggesting wetter and possibly warmer climate than today (CAPE Last Interglacial Project Members, 2006). Near treeline in the northern Yukon today, pollen analyses of MIS 5-aged peat at Ch'ijee's Bluff indicate northward expansion of spruce and *Typha* (cattail) (Figure 2). Many of the peat sites mentioned above also have detailed plant macrofossil records, which provide additional strong evidence for the local presence of spruce and other taxa far north of their current limits. MIS 5-aged pollen has also been recovered from Baffin Island; see 'Canada East of the Yukon' below.

Pollen records dating to MIS 4 (ca. 72–60 ka) and MIS 3 (ca. 60–30 ka) indicate widespread tundra across Alaska, although some spruce were present in the Alaskan interior, the Yukon, and probably on the Bering Land Bridge during MIS 3 (Matthews, 1974b; Wooller et al., 2011; Wetterich et al., 2012). In western Alaska, the record from Squirrel Lake is dominated by grass, sedge, and *Artemisia* pollen with minor amounts of willow and birch (Anderson and Lozhkin, 2001), although a pollen study from the western Seward Peninsula consistently records about 20% spruce during MIS 3, indicating the trees were present, probably on the land bridge itself (Wetterich et al., 2012). Sites in the Yukon are all terrestrial peat deposits; at most localities, spruce pollen frequencies are less than 20%, suggesting only scattered trees within a birch or graminoid tundra (Schweger and Matthews, 1991). However, abundant spruce needles and pollen were recovered in sediments dated to about 38 ka, indicating that spruce may have been locally abundant (Matthews et al., 1990a).

The most recent glaciation (MIS 2 (ca. 29–18 ka)) was a period of severe climate, with maximum cooling during the Last Glacial Maximum (LGM) about 21 ka. Pollen records from unglaciated Alaska and the Yukon (Figure 3) indicate the vegetation was herb and forb dominated, with grass, sedge, and *Artemisia* being the most common taxa. As a rule, sites in western Alaska (i.e., Zagoskin, Kaiyak, and Burial lakes) contain more *Thalictrum* (tundra and woodland plant that favors wet meadows and bogs) than sites to the east (i.e., Bluefish Exposure) (Abbott et al., 2010; Ager, 2003; Anderson, 1985; Zazula et al., 2006a). In addition, the well-preserved Kitluk soil on the Seward Peninsula (Goetcheus and Birks, 2001; Wolf, 2001) and some peat cores from the Bering Land Bridge (Elias et al., 1996) (Figure 1) contain higher frequencies of shrub pollen than sites further inland. Taken together, these data suggest a moisture gradient (as there is today) from a



Figure 1 Map showing localities that are mentioned in the text. Alaskan treeline is based on *Alaska Trees and Shrubs* (Viereck and Little, 1972) and the Canadian treeline is based on *The National Atlas of Canada*, 5th edn. (Energy Mines and Resources Canada, 1993).

relatively mesic west to a drier east. However, macrofossils from the Kitluk soil also indicate a dry-adapted vegetation on the uplands of the Seward Peninsula and a peat core from Norton Sound, just south of the Seward Peninsula (Ager and Phillips, 2008), contains little shrub pollen, suggesting the shrubs were patchy and the vegetation was a mosaic.

The Late Glacial (ca. 18–12 ka) marks the transition from full glacial conditions to the onset of postglacial warming. As climate warmed during this transition, effective moisture (precipitation minus evaporation) increased, so that many lakes came into existence, especially after about 16–14 ka. From the LGM to about 16 ka, pollen records remain dominantly unchanging (mostly graminoids and forbs), but after about 16 ka, birch pollen becomes significantly more important across the region ('the birch rise'), from western Alaska to the Mackenzie River (Figures 1 and 4) (Anderson et al., 2004; Ritchie, 1984). This change in the pollen records marks a significant shift in the vegetation from a dominantly herbaceous tundra to a dominantly shrub tundra, probably reflecting both increasing summer warmth and increased moisture (Finney et al., 2004). The apparent absence of a pattern in the birch expansion, along with its astonishingly rapid rise at some sites, suggests that it survived the LGM *in situ* in protected sites, perhaps reproducing vegetatively (Brubaker et al., 2005).

In the early Holocene (ca. 11–8 ka), many sites across interior Alaska and the Yukon (and western Northwest Territory) contain a short episode of elevated *Populus* (probably cottonwood (*Populus balsamifera*)) and *Salix* (willow) pollen frequencies. *Populus* pollen does not preserve well or travel far from its source (Edwards and Dunwiddie, 1985); frequencies preserved in the pollen records (up to 20%) are far higher than

modern frequencies, indicating widespread *Populus*, possibly as gallery forests along streams (Anderson et al., 2004). Due to a combination of bulk radiocarbon samples and changes in ^{14}C production in the atmosphere, precisely dating the *Populus* zone is problematic (Bartlein et al., 1995), but the period dates roughly between about 12 ka and 10 ka BP (Edwards et al., 2005). Significant *Populus/Salix* woodland development occurred in interior Alaska, the north slope of the Brooks Range (along with abundant cottonwood macrofossils (Mann et al., 2010)), interior Yukon, and western Northwest Territory (Anderson et al., 2004; Ritchie, 1987). *Populus* probably survived in the region during the LGM (Brubaker et al., 2005), although genetic analyses cannot yet identify the refugial population (Breen et al., 2012). Today, *P. balsamifera* forms clonal stands far north of the spruce limit, easily propagates vegetatively, and could have survived the LGM as isolated gallery stands in protected river floodplains.

Spruce pollen and macrofossils appeared in interior Alaska after the *Populus/Salix* expansion, suggesting it was growing locally at least by 10 ka and possibly slightly earlier (Figure 4). However, many pollen records contain low spruce frequencies (<5%) prior to the main spruce rise. This may suggest its survival as refugial populations during the LGM (Brubaker et al., 2005). While the northern limit of spruce in Alaska did not expand beyond its current limits during the early Holocene, a single spruce twig dated to about 9.2 ka was recently recovered from north-coastal Seward Peninsula (Figure 1) (Wetterich et al., 2012). Pollen morphological analyses elsewhere indicate the earliest Holocene-aged spruce pollen in Alaska and the Yukon was from white spruce, later shifting to black spruce about 5 ka (Anderson et al., 2004).

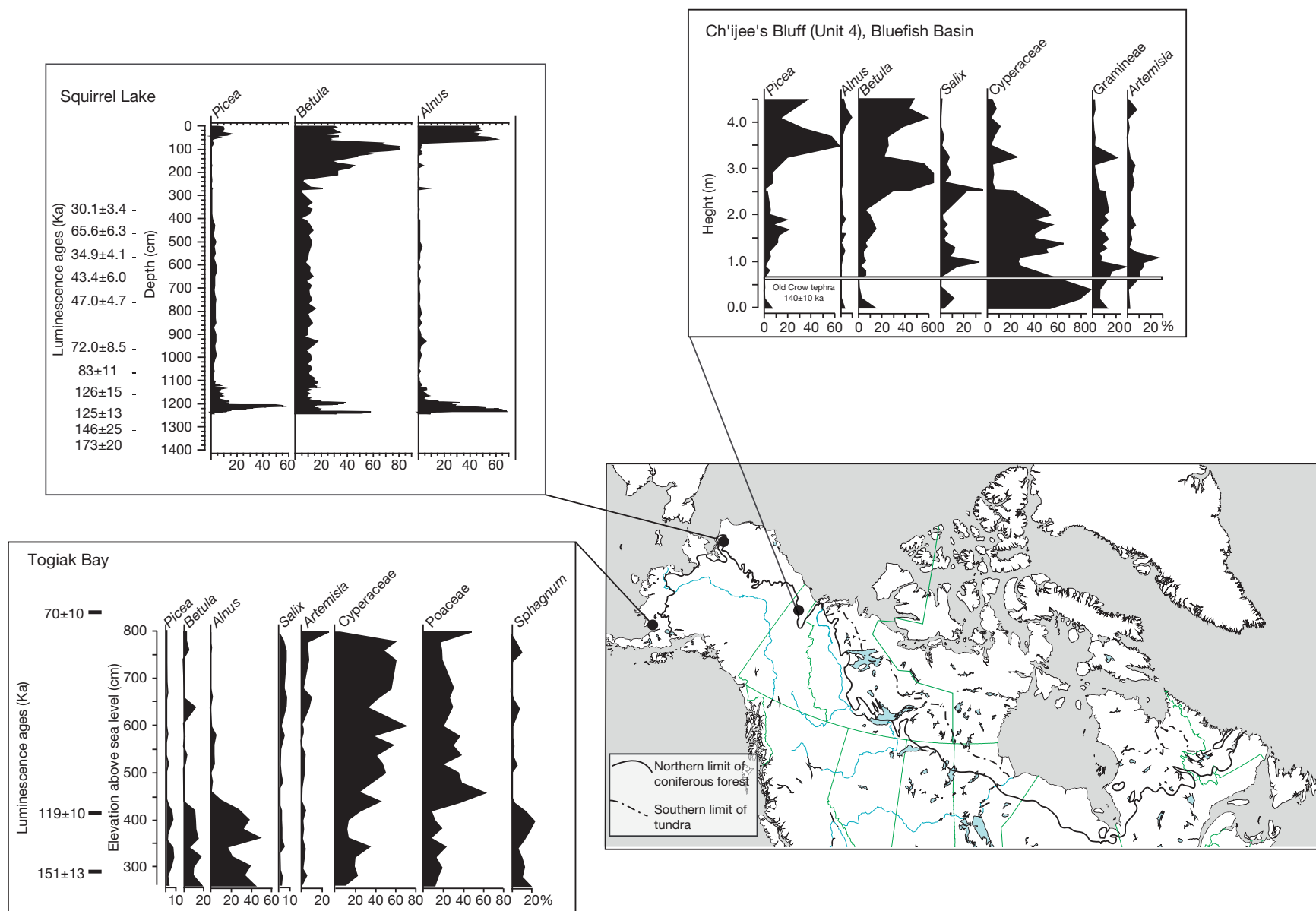


Figure 2 MIS 5-aged pollen records in Alaska and the Yukon. The data are redrawn from publications (Ch'ijee's Bluff (Matthews et al., 1990b) and Togiak Bay (Kaufman et al., 2001)) or kindly provided by the analyst (Squirrel Lake (P. Anderson)). In all records for this and other figures, terrestrial pollen percents are based on the sum of all terrestrial pollen, while spore percentages are based on the sum of terrestrial pollen and spores.

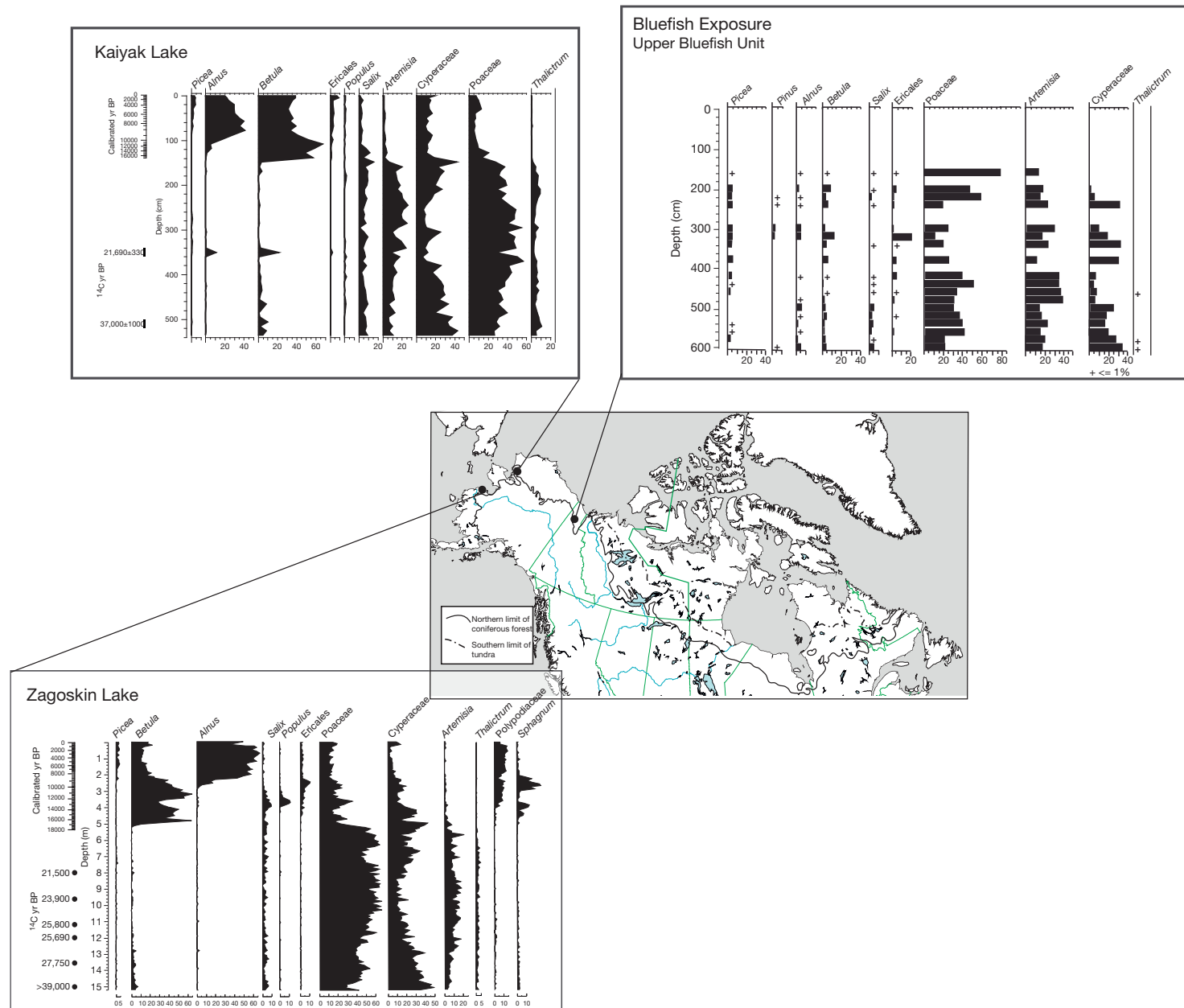


Figure 3 MIS 3 to MIS 2-aged pollen records in Alaska and the Yukon. The data are either redrawn from publications (Bluefish Exposure (Zazula et al., 2006a)) or are calculated from raw pollen data housed in the Neotoma Paleocology Database (see ‘Relevant Websites’ for the URL). Age models in this and subsequent figures are either the author’s published chronology, or in the case where calibrated time scales needed to be built, the dates thought accurate by the investigator with the calibrated age being the median of a one standard deviation probability curve for that date have been used (Reimer et al., 2004). The chronology was then modeled using linear interpolation between the calibrated dates.

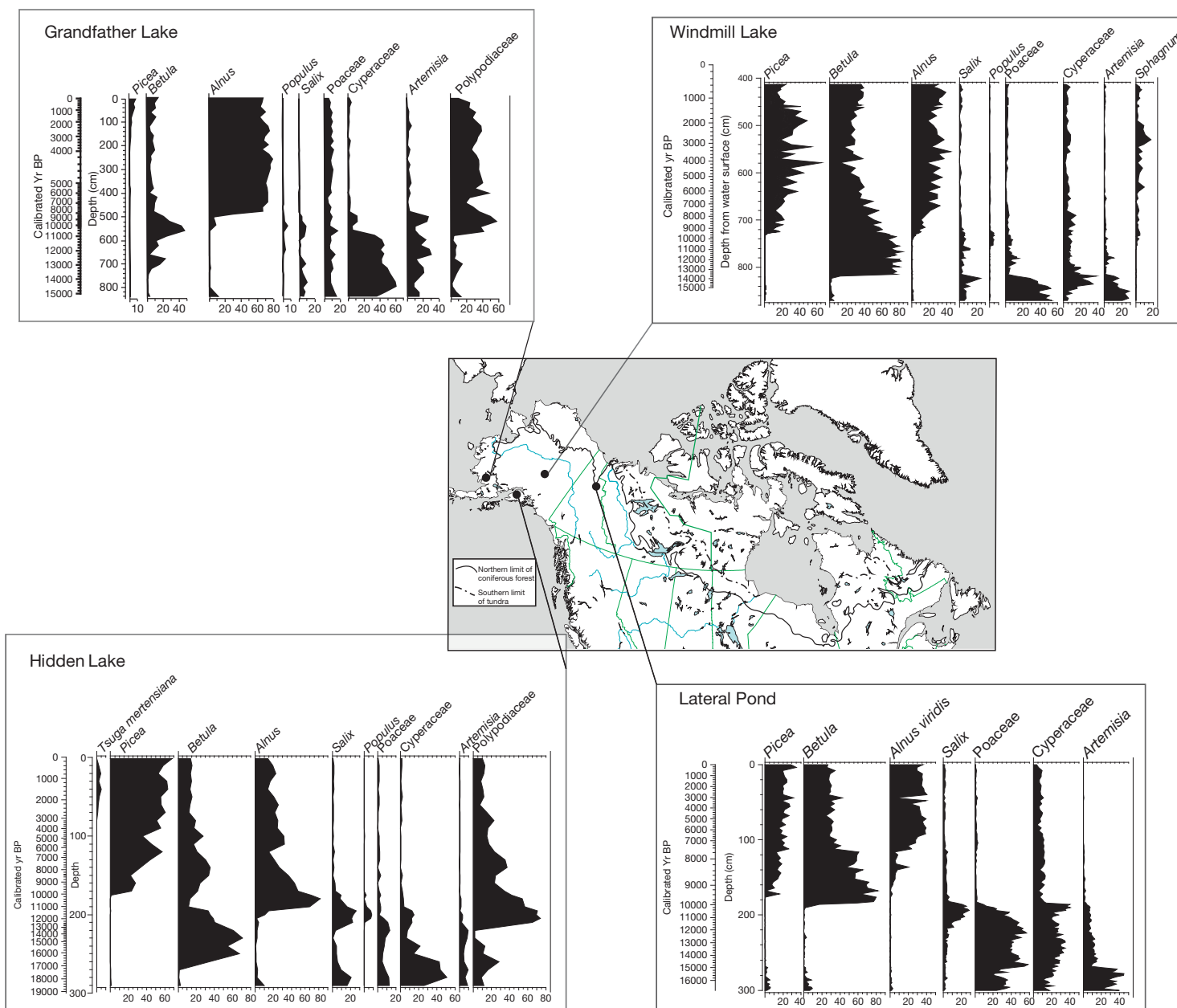


Figure 4 Late Glacial and Holocene-aged pollen records from Alaska and the Yukon. All the pollen data are available at the Neotoma Paleocology Database. Age models are as in **Figure 3**.

By about 10 ka, the alder pollen frequency increased rapidly in northwestern and southwestern Alaskan pollen cores (Figure 4). In the Alaskan interior and Yukon, alder frequencies crossed the 20% threshold (indicating local presence) by about 7–8 ka, and slightly later immediately north of the Brooks Range (Anderson et al., 2004). The very rapid alder rise in the western cores, combined with an apparent delay to the east, suggests alder survived the LGM, either in Alaska or on the now-submerged Bering Land Bridge (Brubaker et al., 2005).

Because it was ice-covered during the last glaciation, southern Alaska has a different vegetation history than interior and northern Alaska. Deglaciation is dated to shortly before 14 ka around Cook Inlet (Figure 1) (Reger et al., 2007). Unlike sites in the interior, only a few records indicate a brief period of herbaceous tundra prior to the development of a birch-dominated tundra (Ager and Brubaker, 1985). Peat cores from the northern Kenai Peninsula (Figure 1) indicate the development of birch shrub tundra by about 14 ka (Jones et al., 2009). While just northeast of Anchorage (Figure 1), this transition is dated to about 13 ka (Yu et al., 2008). At roughly 9–11 ka, sites on the Anchorage Lowland and Kenai Peninsula record transient increases in *Populus* and willow frequencies (Ager and Brubaker, 1985; Anderson et al., 2006b; Jones et al., 2009) similar to that seen in the interior. Shortly afterward, alder frequencies increase, indicating a shift to an alder-dominated vegetation, at least in some sites (Ager, 1983; Ager and Brubaker, 1985; Anderson et al., 2006b; Jones et al., 2009). This alder rise is earlier on the Kenai Peninsula than in the interior (roughly 9–10 ka on the Kenai Peninsula vs. 7–8 ka in the interior), which is consistent with the early alder to the south. Spruce was probably locally present as early as 10 ka at Hidden Lake (Figure 4) and by about 5 ka in the southern Kenai Peninsula. Spruce pollen is not differentiated between interior and coastal species, but it may initially document the arrival of interior taxa, followed by the coastal Sitka spruce in the late Holocene (Ager, 2000). Lastly, mountain hemlock was present locally within the past few thousand years.

Some sites around the Gulf of Alaska clearly record a brief period (ca. 13–11 ka) of tundra expansion at the expense of boreal shrubs and trees. The timing of this shift suggests it is correlated with the well-known Younger Dryas event (Kokorowski et al., 2008). Pleasant Island in southeast Alaska documents the expansion of herbs at the expense of pine and alder, while a pollen and macrofossil record from southwestern Kodiak Island indicates the expansion of a heath vegetation and loss of ferns (Kokorowski et al., 2008). Hidden Lake (Figure 4) also documents a reduction of ferns but slightly earlier. Taken together, these records indicate a brief episode of cooler and/or drier climate. Although the evidence is most compelling at the Gulf of Alaska sites, there is some evidence that similar vegetation shifts are documented in southwest Alaska, in the Alaska Range, and north of the Brooks Range (Kokorowski et al., 2008).

Canada East of the Yukon

During MIS 5, several lakes on Baffin Island (Figure 1) document MIS 5 vegetation (Frechette et al., 2006; Miller et al., 1999). In all cases, the lakes record increased shrubs (birch, alder, and willow) and higher pollen concentration relative to

today. These and other proxies (Francis et al., 2006) suggest that summer air temperatures were significantly warmer than modern, which is consistent with the contemporaneous sites in the Yukon and Alaska.

The Laurentide and Innuitian ice sheets covered much of northern Canada during the last glaciation. However, northwestern Northwest Territories and some of the Arctic islands were unglaciated during MIS 2 (Dyke et al., 2002). Laurentide deglaciation began in the west, with the Mackenzie Delta and Tuktoyaktuk Peninsula (Figure 1) being deglaciated by about 14 ka, although the Hudson Bay basin was still ice-covered around 8.5 ka, and the last remnant of ice did not disappear from northern Québec until after about 6 ka (Dyke, 2004). The Innuitian ice sheet covered the Canadian Arctic Archipelago until about 10 ka (or perhaps earlier – see Dyke, 2004), with subsequent rapid retreat over the next millennium, so that by about 8.5 ka much of the ice sheet had reached its contemporary ice limits (England et al., 2006). As a result of this asymmetrical melting of the two ice sheets, much of the Arctic Archipelago was deglaciated several millennia before northern Québec and Labrador.

On the Tuktoyaktuk Peninsula (Sleet Lake, Figures 1 and 5), a birch shrub tundra from about 14 ka to about 12 ka is interrupted by the spread of spruce and a boreal shrub, sweet gale (*Myrica*), ca. 12–8 ka, indicating a 30–40 km extension of tree-line (spruce needles were also recovered), which retreated southward by about 8 ka (Spear, 1993). After 8 ka, the pollen record is dominated by birch and alder, indicating the establishment of the modern tundra at about this time. At a higher elevation on the Brock Plateau to the east (Figure 1), a pollen record beginning by about 13.9 ka (possibly as early as 16 ka) indicates the vegetation was initially an *Artemisia*-willow tundra, suggesting very dry conditions. A birch shrub tundra grew in the area by about 11 ka (with no evidence of a spruce expansion) and the regional expansion of alder occurred by about 7 ka, although it is unclear whether it grew locally or not (Ruhland et al., 2009).

Lakes in the Mackenzie drainage to the south (MacDonald, 1995) and in the Mackenzie Mountains to the west (i.e., Keele Lake (Szeicz and MacDonald, 2001); Figure 5) record the presence of a birch shrub tundra by about 13 ka. Spruce (initially white spruce, with black spruce shortly afterward) was probably locally present shortly before about 11 ka along the Mackenzie River but was later in the mountains (about 9 ka), probably due to cooler temperatures at higher elevations. Increased spruce pollen influx during the early to middle Holocene in the alpine sites also indicates that spruce may have been more abundant or treeline slightly higher than it is today.

So far, the oldest reliably dated postglacial pollen record from the Canadian Arctic Archipelago is from Melville Island (Figures 1 and 6), where the record begins about 13 ka, suggesting that deglaciation was earlier here than elsewhere. Here, the initial vegetation was grass-dominated, with sedges and *Oxyria digyna* (an Arctic/alpine snowbed plant) as secondary taxa (Peros et al., 2010). Other Arctic Archipelago pollen records (Banks, Victoria, and Ellesmere islands) begin about 10 or 11 ka (Figures 1 and 6), shortly after deglaciation (with the exception of Banks Island which was mostly unglaciated during MIS 2). Graminoids dominate all records at their outset, though *Artemisia* and willows are abundant initially on Victoria Island (Peros and Gajewski, 2008) and *Oxyria digyna* is common at

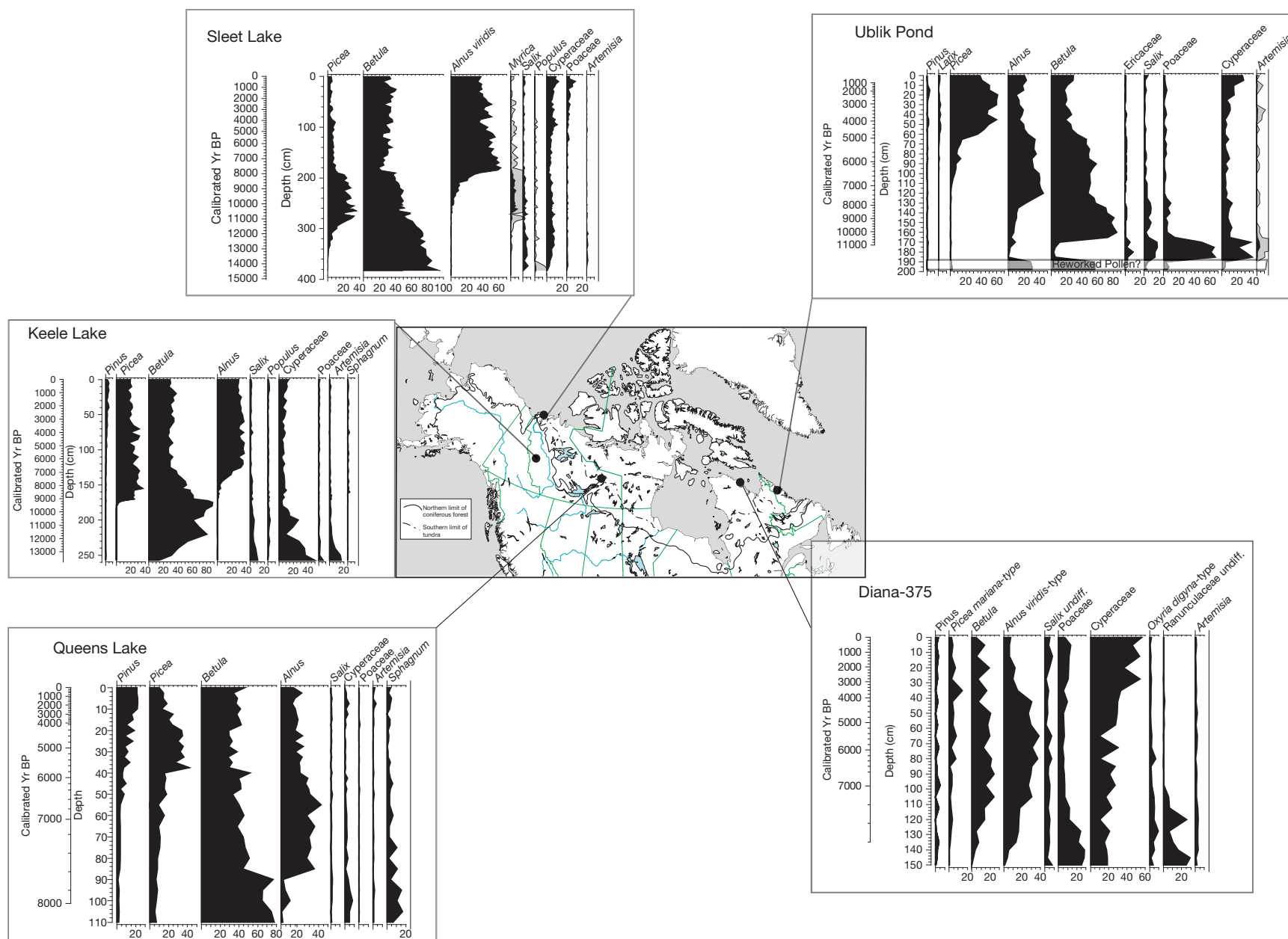


Figure 5 Late Glacial and Holocene-aged pollen records from Northwest Territories, Nunavut, Québec, and Labrador. All the data are available in the Neotoma Paleocology Database. The Sleet Lake chronology follows the depth-age model recommended by the analyst (Spear, 1993); other age models are as in Figure 3.

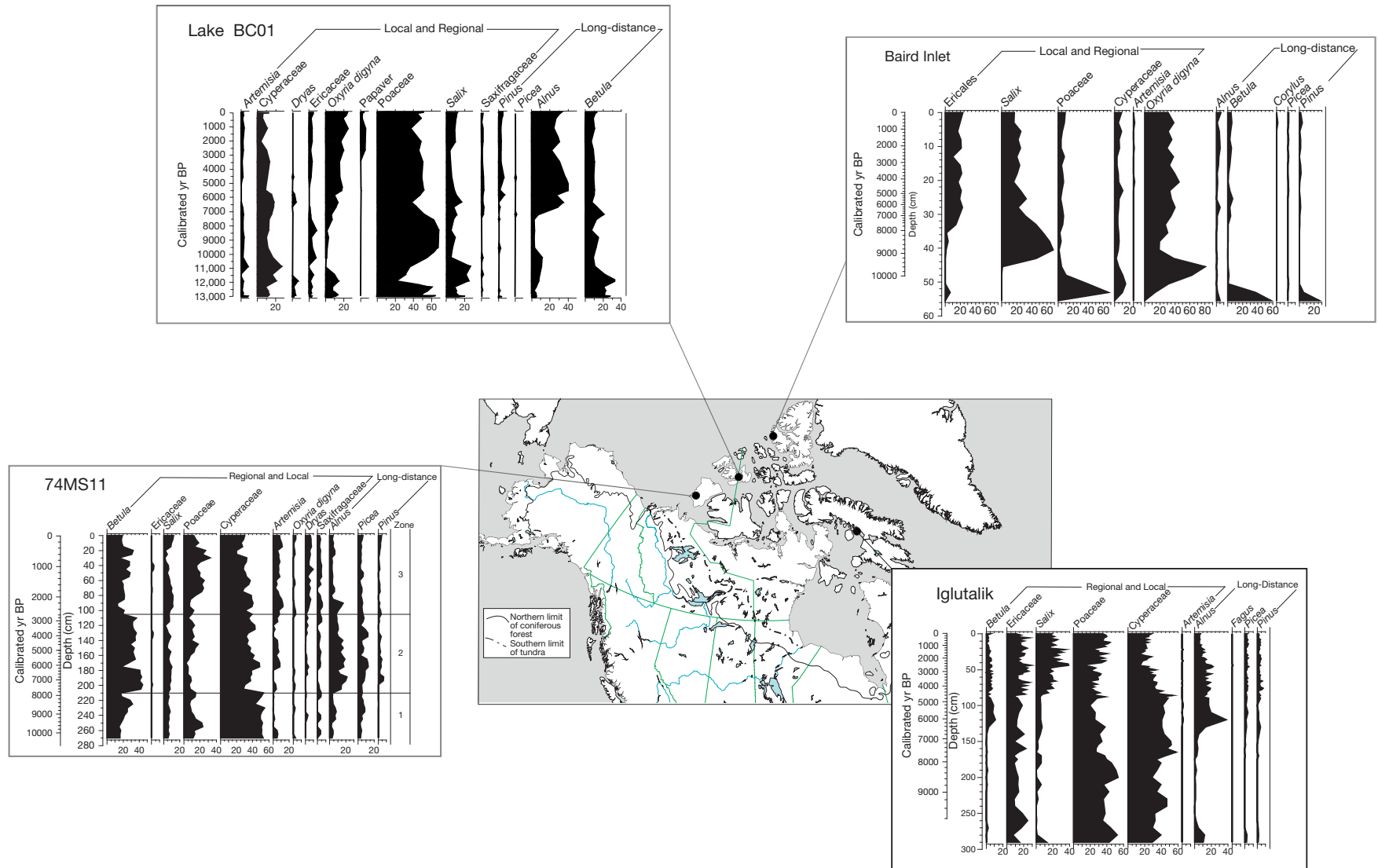


Figure 6 Holocene-aged pollen records from Arctic Canada. With the exception of Lake BC01 (Peros et al., 2010), all the data are available in the Neotoma Paleoecology Database and age models are as in Figure 3. Local/regional terrestrial pollen percentages are based on the sum of the local and regional terrestrial taxa. Long-distance percentages are based on the sum of the local/regional taxa plus the long-distance taxa. Note that *Betula* is considered local/regional in the southern to middle Arctic, while it is considered long-distance in the northern Arctic. In addition, the basal sample from Baird Inlet contains 53 long-distance pollen grains, but only six local pollen grains (four are Ericales). As a result, the local pollen frequencies are highly skewed and have been omitted from the diagram, even though the long-distance taxa are still shown.

Baird Inlet on Ellesmere Island (Hyvärinen, 1985). Willows are more abundant at several of these sites (i.e., Victoria, Melville, Somerset, and Ellesmere islands) during the early Holocene than they are today (Figure 6). In addition, pollen concentrations (and influx when it is calculated) are often higher during the early Holocene than today, suggesting that plant abundance and productivity was greater than today.

On Baffin Island (Figures 1 and 6), despite some local variation, grasses and/or sedges dominate the early Holocene vegetation. In the south, some shrubs (heaths) were present, while in the north, forbs and, in one case, ferns were relatively important (Gajewski and MacDonald, 2004; Short et al., 1985). Off the southern tip of Baffin Island, the Robinson Lake pollen record is initially dominated by grass, followed by a transition to sedge and *O. digyna*-dominated tundra (Miller et al., 1999).

By the middle Holocene (ca. 7 ka), Canada was mostly deglaciated. Lakes that were already in existence prior to this time generally document the development of the modern vegetation by about 7 ka. However, several records also document vegetation succession following deglaciation, while others suggest mid-Holocene treeline fluctuations.

Several pollen cores (cf. Queen's Lake (Figure 5)) from the forest-tundra ecotone indicate that a birch shrub tundra was well established before ca. 8 ka and alder became important by about 7 ka. Spruce (probably black spruce) pollen frequencies reached maximum amounts (up to 40%) about 7–4.5 ka, decreasing gradually afterward, indicating a treeline advance that is asynchronous with the early Holocene treeline changes documented on the Tuktoyaktuk Peninsula. Pine (*Pinus*) pollen frequencies increase to modern levels by about 2 ka, probably reflecting population growth south of the site (Gajewski and MacDonald, 2004).

A transect of pollen records crossing the treeline of interior Québec indicates that shortly after deglaciation (ca. 7 ka), alder and spruce (probably black spruce) pollen frequencies rapidly increased, reaching their northern limits by about 6 ka. In addition, one site also suggests the expansion of cottonwood (*Populus*) and larch (*Larix*), indicating that the earliest forest was somewhat different from today's (Gajewski and MacDonald, 2004). Alder frequencies are at their highest in the middle Holocene, decreasing about 4.5 ka. For the past 3 ka, within the forest-tundra, spruce pollen frequencies decreased gradually, while birch and nonarboreal pollen increased, suggesting slow deforestation (Gajewski and MacDonald, 2004), although the overall forest limit in this region is notable for its apparent stability throughout the middle and late Holocene (Payette et al., 2002).

Coastal regions of northern Québec were probably ice-free by about 8 ka (Dyke, 2004). The pollen record from Diana-375 (Figure 5) documents an initial postglacial vegetation of herbaceous tundra, which was replaced by a birch and alder shrub tundra by about 6.8 ka. Shrub pollen frequencies decrease by about 3.7 ka, concurrent with increases in graminoid pollen, suggesting an expansion of herbaceous (sedge-dominated) tundra in the late Holocene (Richard, 1981).

In northern Labrador, like Québec, pollen records document the rapid colonization of the landscape after deglaciation. Coastal regions may have been ice-free as early as 9 ka, while the interior was still ice-covered until about 8 ka (Dyke, 2004). Despite a basal date that is possibly too old

(ca. 12 ka), the coastal site of Ubluk Pond (Figure 5) (Short and Nichols, 1977) documents a brief period of graminoid and willow tundra preceding the transition to a birch shrub tundra, which was followed by alder expansion about 7.5 ka. Spruce frequencies begin increasing about 5 ka, reaching maximum amounts (ca. 50%) by about 2.5 ka, after which frequencies gradually decrease to modern levels. This spruce pollen increase is seen at a number of sites and suggests a transient northward and altitudinal treeline shift (Lamb, 1985; Short and Nichols, 1977).

Lastly, in the Arctic Archipelago, the mid- to late Holocene pollen records from Banks Island (Figures 1 and 6) document slight changes in vegetation. About 7 ka, the cores indicate higher proportion of local (birch) and long-distance (alder) shrub pollen, suggesting vegetation was probably shrubbier than today at least in the southern part of the island. After about 2–3 ka, shrub pollen frequencies generally decreased, and the vegetation reverted to a sedge-dominated tundra (Gajewski and MacDonald, 2004).

On Baffin Island (Figures 1 and 6), the middle to late Holocene vegetation began with relatively shrubby vegetation, with willows, birches, and ericads being the dominant shrubs. Sites in the southern part of the island also record significant increases in alder pollen (Iglutalik; Figure 6), but it is unlikely that it ever reached the island. Later, by about 4–3 ka, shrubs may have become less common, as some sites record reduced birch and willow pollen during the late Holocene (Short et al., 1985).

Discussion and Conclusions: LGM and Holocene Thermal Maximum Vegetation

Last Glacial Maximum

The LGM was the coldest part of the last Ice Age, centered about 21 ka, which generally corresponds with the period of greatest ice advance (Dyke, 2004; Dyke et al., 2002). Sea level was depressed by about 125 m, exposing the Bering Land Bridge and the continental shelf. One consequence of this change is sites that are near the coast today would have had a much more continental climate during the LGM. General circulation models indicate that both summer and winter temperatures in unglaciated Alaska and Yukon were 2–5 °C cooler than modern, sharply decreasing toward the east and the Laurentide Ice Sheet (Braconnot et al., 2007). Insect and pollen-based reconstructions are roughly similar to the model simulations. Chironomids (an aquatic insect) suggest that summer air temperatures were 2–4 °C cooler than modern in western Alaska (Kurek et al., 2009). Beetle-based reconstructions are somewhat variable, with summer temperatures about as warm as today at some sites (Seward Peninsula) and others significantly colder (up to 7 °C colder in the Yukon). A pollen-based reconstruction using the modern analog technique suggests summer temperatures were about 4 °C cooler and winter temperatures about 2 °C cooler (Viau et al., 2008).

The climate models also indicate increased aridity during the LGM with greatest aridity toward the margins of the ice sheet (Braconnot et al., 2007). Extensive dune fields, increased loess production, and near absence of lakes in unglaciated

Alaska and Yukon support the model results (Hopkins et al., 1982; Muhs et al., 2003).

Vegetation reconstructions during the LGM are at the center of a long-standing controversy (Hopkins et al., 1982). Pollen-based reconstructions tend to focus on the very low pollen influx (almost two orders of magnitude less than the Holocene) (Ritchie, 1984) and the abundance of herb and forb pollen. This suggested to researchers that there had been a very sparse graminoid vegetation, and in places, a polar desert (Anderson et al., 2004). The difficulty is that the quantity of bones from large grazers (mammoth, bison, and horse) from this interval implies a vegetation that was richer and more diverse than indicated by the pollen records (Guthrie, 1990). This apparent contradiction is somewhat resolved with high-resolution dating studies, indicating that mammoths, anyway, were more abundant before and after the LGM than during the LGM (Guthrie, 2003), and by vegetation modeling efforts, suggesting that the graminoid vegetation could have been more productive during the LGM than it is today, especially in central and western Alaska (Allen et al., 2010).

Central to the controversy is how to interpret the grass, sedge, and *Artemisia* pollen; this combination of taxa is not seen in today's vegetation. Pollen analysis cannot determine grass and sedge pollen to anything finer than the family level. As a result, those taxa are difficult to interpret in terms of vegetation or climate. *Artemisia* pollen is also somewhat problematic because in the north there are *Artemisia* species with steppic affinities and others with tundra affinities. Macrofossils recovered in the Yukon indicate the steppic taxon (along with other steppic plants) grew there during the LGM, but there is no indication that it grew in sites farther west. Other indicator taxa (species with specific environmental preferences that can be identified in the pollen record) can help with interpreting the pollen record. For example, many records contain weedy taxa which are typical of disturbed soils, suggesting that the LGM vegetation was not a continuous plant-covered surface but was probably pocked by areas of sediment, as is typical for high-latitude localities today (CAVM Team, 2003). Or, as noted above, *Thalictrum*, a relatively mesic taxon, is more common in the west than in the east, suggesting a moisture gradient during the LGM. While no exact modern analog of LGM vegetation exists today, analyses using plant functional types (vegetation classifications based on plant phenology (deciduous vs. evergreen), leaf morphology (broad leaf vs. needle leaf), life form (tree, shrub, or herbaceous), and climatic tolerances) indicate the closest modern biome is today found in disjunct locations across the northern tundra where shrubs are highly reduced and graminoid and forb taxa dominate (Bigelow et al., 2003).

Despite the widespread presence of herbaceous taxa during the LGM, trees may have been present in refugia. Evidence is accumulating that conifers were present in the Yukon just prior to the onset of the LGM (Zazula et al., 2006b) and in eastern Siberia (albeit a shrubby pine) during the LGM (Anderson et al., 2010). Although LGM-aged conifer macrofossils are yet to be found in Alaska or the Yukon, low (5% or less) but consistent spruce pollen (Figure 3) as well as the pattern of expansion during the Holocene suggest it may have survived the LGM in cryptic refugia (Brubaker et al., 2005). In addition, DNA analyses of white spruce in North America suggest that

populations growing today in unglaciated Alaska and the Yukon are significantly different from southern populations, suggesting they have been isolated from each other over glacial intervals (Anderson et al., 2006a).

Holocene Thermal Maximum

Many of the records discussed above document vegetation shifts which may be the result of warmer than modern climate during the Holocene. This includes the *Populus* expansion in Alaska, spruce treeline changes in the Northwest Territories, western Alaska, and eastern Canada, as well as shifts in shrub abundance in Arctic Canada. However, these shifts are not contemporaneous; they occur earlier in the west (early Holocene) and later in the east (middle to late Holocene) (Kaufman et al., 2004).

During the first half of the Holocene, due to variations in the Earth's orbit around the Sun, the amount of solar radiation reaching the high latitudes was greater than modern during the summer but less than modern during the winter. In the early Holocene, climate models suggest that because of this shift, temperatures were as much as 4 °C warmer than modern in summer and 2–4 °C cooler than modern in winter (Kutzbach et al., 1998), although by the middle Holocene, these anomalies had relaxed to about 1 °C warmer in summer and 1 °C cooler in winter (see PMIP2 website below). Pollen-based climatic reconstructions for the middle Holocene (6 ka) suggest a fair amount of variability across the north, although summer heat (measured as growing degree days) was generally higher than today, especially in northeastern Canada (Bartlein et al., 2011).

Spruce expansion on the Tuktoyaktuk Peninsula in northwest Canada (Ritchie et al., 1983) and north coastal Seward Peninsula in western Alaska (Wetterich et al., 2012) suggests ameliorating climate. The treeline shift on the Tuktoyaktuk Peninsula is about 30–40 km, whereas the Seward Peninsula spruce twig is about 100 km beyond the modern treeline to the east. However, it seems most likely that the Seward Peninsula spruce derived from a small population on the Bering Land Bridge. A pollen record from St. Paul Island on the southern land bridge records elevated spruce pollen percentages about 12 ka (Colinvaux, 1981) and no spruce wood has yet been recovered elsewhere on the Seward Peninsula (Hopkins et al., 1981) that would indicate a range expansion from the modern spruce limits to the east. The presence of spruce at both the Tuktoyaktuk Peninsula and the Seward Peninsula suggests about a 2–3 °C increase in summer temperature (Wetterich et al., 2012; Kaufman et al., 2004). As both sites are near the coast today, they would have been affected by lower sea levels as well as changes in insolation during the Holocene Thermal Maximum (HTM), both of which would have led to increased continentality and warmer summers.

The early Holocene *Populus* woodland of Alaska and the Yukon is a unique vegetation not found in the north today. This deciduous vegetation was mirrored in eastern Siberia, although the Siberian woodland included larch (a deciduous conifer) instead of *Populus* (Edwards et al., 2005). The expansion of these deciduous trees during the HTM may reflect not only warm springs as noted below but also the relatively cold

winters, because both these taxa tolerate very cold temperatures. Today, in Alaska, *Populus* can be found north of the spruce limit, and in eastern Siberia, larch forms treeline.

Macrofossil evidence from the north slope of the Brooks Range suggests it was cottonwood (*P. balsamifera*) that expanded during the HTM, at least in that region. Cottonwood reacts favorably to warm springs, and this partially explains its expansion (Kaufman et al., 2004). However, other factors may have been equally important. *Populus* may have been limited earlier by a lack of available substrates. Cottonwood is an early successional species found today mainly on aggrading floodplains. A careful study of drainage histories in the central foothills of the Brooks Range indicates widespread alluviation contemporaneous with the *Populus* expansion as well as downcutting and reduced *Populus* presence (Mann et al., 2002, 2010).

Treeline expansion in the eastern Northwest Territories, which was glaciated until about 8 ka, dates to the mid-Holocene (Queen's Lake and others), at the same time that lake productivity increased (Gajewski and MacDonald, 2004). Radiocarbon-dated buried forest soils west of Hudson Bay (deglaciated about 6 ka) indicate a 250 km northward extension of treeline about 4 ka (Sorenson et al., 1971), although pollen records to the north and west do not register this shift (Gajewski and MacDonald, 2004; Seppä et al., 2003).

Increased shrub abundance in the eastern Canadian Arctic (especially southern regions) dates to the middle and late Holocene, and sites on southern Baffin Island probably indicate local expansion of birch, although stronger or more frequent southerly winds blowing pollen from the south cannot be excluded. Pollen-based climate reconstructions (alder and boreal taxa excluded from the analysis) indicate summers were 1–2 °C warmer than modern (Kerwin et al., 2004).

Treeline sites in northern Québec and Labrador document modest HTM vegetation shifts. In Québec, treeline was apparently no farther north than it is today, but the forest-tundra ecotone of the middle Holocene may have more trees than today (Gajewski and MacDonald, 2004); temperature reconstructions suggest summers were less than 1 °C warmer than today (Kerwin et al., 2004).

The time-transgressive nature of the HTM is due in part to the presence of the Laurentide Ice Sheet over much of northern Canada (Kaufman et al., 2004). Ice Sheets affect climate both regionally (large-scale circulation effects) and locally (cold wind and/or water draining the ice sheet) (Bartlein et al., 1991). As the Laurentide Ice Sheet thinned, atmospheric circulation would have taken on its modern configuration, and local effects would have been time-transgressive across the region as the remaining ice melted. This may explain the evidence for cooling within an overall warming trend. For example, one of the most dramatic local effects was the draining of the large Laurentide proglacial lakes and resulting cold water flush into the North Atlantic at about 8.2 ka (Barber et al., 1999). This short (less than 1000 yr), sharp climatic reversal is seen only in one pollen diagram (Seppä et al., 2003) probably because sites that would have been closest to the cold flush would have either been covered by ice or by the proglacial lakes themselves. The vegetation recovery was rapid, however, and evidence of HTM warming returned with the continued expansion of shrubs and trees.

In sum, the HTM is a time-transgressive episode controlled by a combination of insolation changes and the timing of deglaciation. Pollen records document these changes by transient increases of tree and shrub pollen, which occur during the early Holocene in the west and the middle to late Holocene to the east.

Conclusions

Pollen, because it is widespread and generally identifiable to the genus level, is a robust means of reconstructing past vegetation, and the pollen records summarized above provide vegetation reconstructions spanning the past ca. 125 ka. During that interval, climate has been both colder (MIS 4, MIS 3, and MIS 2) and warmer (MIS 5 and HTM) than modern. During warm periods, trees and/or shrubs grew farther north than they do today. During cold periods, trees and most shrubs virtually disappear, leaving a herbaceous vegetation. Pollen data also record changes in plant productivity, with low pollen abundance (e.g., during the LGM) reflecting relatively sparse vegetation cover.

Pollen data are most robust when compared with other proxy records, especially plant macrofossils. Pollen, because it travels widely and because of variations in pollen production and preservation, sometimes results in a biased vegetation reconstruction. This is particularly true at treeline and in the tundra where both long-distance pollen and low local pollen production can be a problem. For example, the presence of spruce needles at Sleet Lake (Figure 5) during the HTM definitively documents that the spruce limit had expanded northward and it was not just a wind shift that brought increased spruce pollen to the site. In addition, in cases where plants are near their ecophysiological limits and are reproducing only vegetatively, the pollen record may underestimate the abundance of a taxon. This is probably the case during the Late Glacial of Alaska and the Yukon when birch pollen is so rare. In this case, even though there are no birch macrofossils, the rapid birch expansion with climatic amelioration suggests it was already present but not pollinating.

See also: **Beetle Records:** Late Pleistocene of Australia. **Glaciations:** Late Quaternary in North America. **Lake Level Studies:** North America. **Paleoceanography, Physical and Chemical Proxies:** Oxygen-Isotope Stratigraphy of the Oceans. **Plant Macrofossil Records:** Arctic North America. **Pollen Records, Late Pleistocene:** Western North America. **Sea Level Studies:** Eustatic Sea-Level Changes Since the Last Glacial Maximum.

References

- Abbott MB, Edwards ME, and Finney BP (2010) A 40,000-yr record of environmental change from Burial Lake in Northwest Alaska. *Quaternary Research* 74: 156–165.
- Ager TA (1983) Holocene vegetational history of Alaska. In: Wright HE (ed.) *The Holocene. Late Quaternary Environments of the United States* pp. 128–141. Minneapolis, MN: University of Minnesota Press.
- Ager TA (1998) Ostglacial vegetation history of the Kachemak Bay area, Cook Inlet, south-central Alaska. In: Kelley KD and Gough LP (eds.) *Geologic Studies in Alaska*

- by the U.S. Geological Survey, pp. 147–165. Washington, DC: U.S. Geological Survey U.S. Geological Survey Paper 1615.
- Ager TA (2003) Late Quaternary vegetation and climate history of the central Bering land bridge from St. Michael Island, western Alaska. *Quaternary Research* 60: 19–32.
- Ager TA and Brubaker LB (1985) Quaternary palynology and vegetational history of Alaska. In: Bryant V and Holloway F (eds.) *Pollen Records of Late Quaternary North American Sediments*, pp. 353–384. Dallas, TX: American Association of Stratigraphic Palynologists.
- Ager TA and Phillips RL (2008) Pollen evidence for late Pleistocene Bering Land Bridge environments for Norton Sound, northeastern Bering Sea, Alaska. *Arctic, Antarctic, and Alpine Research* 40: 451–461.
- Allen JRM, Hickler T, Singarayer JS, Sykes MT, Valdes PJ, and Huntley B (2010) Last glacial vegetation of northern Eurasia. *Quaternary Science Reviews* 29: 2604–2618. <http://dx.doi.org/10.1016/j.quascirev.2010.05.031>.
- Anderson PM (1985) Late Quaternary vegetational change in the Kotzebue Sound northwestern Alaska. *Quaternary Research* 24: 307–321.
- Anderson PM, Edwards ME, and Brubaker LB (2004) Results and paleoclimate implications of 35 years of paleoecological research in Alaska. In: Gillespie AR and Porter SC (eds.) *The Quaternary Period in the United States*, pp. 427–440. Amsterdam: Elsevier.
- Anderson RS, Hallett DJ, Berg E, et al. (2006a) Holocene development of Boreal forests and fire regimes on the Kenai lowlands of Alaska. *The Holocene* 16: 791–803.
- Anderson LL, Hu FS, Nelson DM, Petit RJ, and Paige KN (2006b) Ice-age endurance: DNA evidence of a white spruce refugium in Alaska. *Proceedings of the National Academy of Sciences of the United States of America* 103: 12447–12450.
- Anderson PM and Lozhkin AV (2001) The stage 3 interstadial complex (Karginskii/middle Wisconsinan interval) of Beringia: Variations in paleoenvironments and implications for paleoclimatic interpretations. *Quaternary Science Reviews* 20: 93–126.
- Anderson PM, Lozhkin AV, Solomatkina TB, and Brown TA (2010) Paleoclimatic implications of glacial and postglacial refugia for *Pinus pumila* in western Beringia. *Quaternary Research* 73: 269–276. <http://dx.doi.org/10.1016/j.yqres.2009.09.008>.
- Barber DC, Dyke AS, Hillaire-Marcel C, et al. (1999) Forcing of the cold event of 8200 years ago by catastrophic drainage of Laurentide lakes. *Nature* 400: 344–348.
- Bartlein PJ, Anderson PM, Edwards ME, and McDowell PF (1991) A framework for interpreting paleoclimatic variations in eastern Beringia. *Quaternary International* 10–12: 73–83.
- Bartlein PJ, Edwards ME, Shafer SL, and Barker ED Jr. (1995) Calibration of radiocarbon ages and the interpretation of paleoenvironmental records. *Quaternary Research* 44: 417–424.
- Bartlein PJ, Harrison SP, Brewer S, et al. (2011) Pollen-based continental climate reconstructions at 6 and 21 ka: A global synthesis. *Climate Dynamics* 37: 775–802. <http://dx.doi.org/10.1007/s00382-010-0904-1>.
- Bigelow NH, Brubaker LB, Edwards ME, et al. (2003) Climate change and Arctic ecosystems: 1 Vegetation changes north of 55°N between the last glacial maximum, mid-Holocene, and present. *Journal of Geophysical Research* 108(D19) <http://dx.doi.org/10.1029/2002JD002558> 11–11–25.
- Braconnot P, Otto-Bliesner B, Harrison S, et al. (2007) Results of PMIP2 coupled simulations of the Mid-Holocene and Last Glacial Maximum – Part 1: Experiments and large-scale features. *Climate of the Past* 3: 261–277.
- Breen AL, Murray DF, and Olson MS (2012) Genetic consequences of glacial survival: The late Quaternary history of balsam poplar (*Populus balsamifera* L.) in North America. *Journal of Biogeography* 39: 918–928. <http://dx.doi.org/10.1111/j.1365-2699.2011.02657.x>.
- Brubaker LB, Anderson PM, Edwards ME, and Lozhkin AV (2005) Beringia as a glacial refugium for boreal trees and shrubs: New perspectives from mapped pollen data. *Journal of Biogeography* 32: 833–843.
- CAVM Team (2003) Circumpolar arctic vegetation map. In: Walker DA (ed.) *Conservation of Arctic Flora and Fauna Map No. 1*. Anchorage: U.S. Fish and Wildlife Service.
- CAPE Last Interglacial Project Members (2006) Last interglacial arctic warmth confirms polar amplification of climate change. *Quaternary Science Reviews* 25: 1383–1400.
- Colinvaux PA (1981) Historical ecology in Beringia: The south land bridge coast at St Paul Island. *Quaternary Research* 16: 18–36.
- Dyke AS (2004) An outline of North American deglaciation with emphasis on central and northern Canada. In: Ehlers J and Gibbard PL (eds.) *Quaternary Glaciations – Extent and Chronology, Part II*, pp. 373–424. Amsterdam: Elsevier.
- Dyke AS, Andrews JT, Clark PU, et al. (2002) The Laurentide and Innuitian ice sheets during the Last Glacial Maximum. *Quaternary Science Reviews* 21: 9–31.
- Edwards ME, Brubaker LB, Lozhkin AV, and Anderson PM (2005) Structurally novel biomes: A response to past warming in Beringia. *Ecology* 86: 1696–1703.
- Edwards ME and Dunwiddie PW (1985) Dendrochronological and palynological observations on *Populus balsamifera* in northern Alaska, U.S.A. *Arctic and Alpine Research* 17: 271–278.
- Edwards ME, Hamilton TD, Elias SA, Bigelow NH, and Krumhardt AP (2003) Interglacial extension of the boreal forest limit in the Noatak valley, northwest Alaska: Evidence from an exhumed river-cut bluff and debris apron. *Arctic, Antarctic, and Alpine Research* 35: 460–468.
- Elias SA, Short SK, Nelson CH, and Birks HH (1996) Life and times of the Bering land bridge. *Nature* 382: 60–63.
- Elliott-Fisk DL (2000) The taiga and boreal forest. In: Barbour MG and Billings WD (eds.) *North American Terrestrial Vegetation*, 2nd edn., pp. 41–73. Cambridge: Cambridge University Press.
- Energy Mines and Resources Canada (1993) *National Atlas of Canada*, 5th edn. Ottawa: Energy, Mines, and Resources Canada.
- England J, Atkinson N, Bednarski J, Dyke AS, Hodgson DA, and Ó Cofaigh C (2006) The Innuitian ice sheet: Configuration, dynamics and chronology. *Quaternary Science Reviews* 26: 689–703.
- Finney BP, Rühland K, Smol JP, and Fallu M-A (2004) Paleolimnology of the North American subarctic. In: Pienitz R, Douglas MSV, and Smol JP (eds.) *Long-Term Environmental Change in Arctic and Antarctic Lakes*, pp. 269–317. Dordrecht: Springer.
- Francis DL, Wolfe AP, Walker IR, and Miller GH (2006) Interglacial and Holocene temperature reconstructions based on midge remains in sediments of two lakes from Barin Island, Nunavut, Arctic Canada. *Palaeogeography, Palaeoclimatology, Palaeoecology* 236: 107–125.
- Frechette B, Wolfe AP, Miller GH, Richard PJH, and de Vernal A (2006) Vegetation and climate of the last interglacial on Baffin Island, Arctic Canada. *Palaeogeography, Palaeoclimatology, Palaeoecology* 236: 91–106. <http://dx.doi.org/10.1016/j.palaeo.2005.11.034>.
- Gajewski K and MacDonald GM (2004) Palynology of North American arctic lakes. In: Pienitz R, Douglas MSV, and Smol JP (eds.) *Long-term Environmental Change in Arctic and Antarctic Lakes*, pp. 89–116. Dordrecht: Springer.
- Goetcheus VG and Birks HH (2001) Full-glacial upland tundra vegetation preserved under tephra in the Beringia National Park, Seward Peninsula, Alaska. *Quaternary Science Reviews* 20: 135–147.
- Guthrie RD (1990) *Frozen Fauna of the Mammoth Steppe*. Chicago, IL: University of Chicago Press.
- Guthrie RD (2003) Rapid body size decline in Alaskan Pleistocene horses before extinction. *Nature* 426: 169–171.
- Hopkins DM, Matthews JV, Schweger CE, and Young SB (1982) *Paleoecology of Beringia*. New York: Academic Press pp. 489.
- Hopkins DM, Smith PA, and Matthews JV Jr. (1981) Dated wood from Alaska and the Yukon: Implications for forest refugia in Beringia. *Quaternary Research* 15: 217–249.
- Hyvärinen H (1985) Holocene pollen stratigraphy of Baird Inlet, east-central Ellesmere Island, Arctic Canada. *Boreas* 14: 19–32.
- Jones MC, Peteet DM, Kurdyla D, and Guilderson TP (2009) Climate and vegetation history from a 14 000-year peatland record, Kenai Peninsula, Alaska. *Quaternary Research* 72: 207–217.
- Kaufman DS, Ager TA, Anderson NJ, et al. (2004) Holocene thermal maximum in the western Arctic (0–180°W). *Quaternary Science Reviews* 23: 529–560.
- Kaufman DS, Manley WF, Wolfe AP, et al. (2001) The last interglacial to glacial transition, Togiak Bay, southwestern Alaska. *Quaternary Research* 55: 190–202.
- Kerwin MW, Overpeck JT, Webb RS, and Anderson KH (2004) Pollen-based summer temperature reconstructions for the eastern Canadian boreal forest, subarctic, and Arctic. *Quaternary Science Reviews* 23: 1901–1924.
- Kokorowski HD, Anderson PM, Mock CJ, and Lozhkin AV (2008) A re-evaluation and spatial analysis of evidence for a Younger Dryas climatic reversal in Beringia. *Quaternary Science Reviews* 27: 1710–1722.
- Kurek J, Cwynar LC, Ager TA, Abbott MB, and Edwards ME (2009) Late Quaternary paleoclimate of western Alaska inferred from fossil chironomids and its relation to vegetation histories. *Quaternary Science Reviews* 28: 799–811. <http://dx.doi.org/10.1016/j.quascirev.2008.12.001>.
- Kutzbach JE, Gallimore R, Harrison SP, Behling PJ, Selin R, and Laarif R (1998) Climate and biome simulations for the past 21,000 years. *Quaternary Science Reviews* 17: 473–506.
- Lamb HF (1985) Palynological evidence for postglacial change in the position of tree limit in Labrador. *Ecological Monographs* 55: 241–258.
- MacDonald GM (1995) Vegetation of the continental Northwest Territories at 6 ka BP. *Géographie Physique et Quaternaire* 49: 37–43.
- Mann DH, Groves P, Reanier RE, and Kunz ML (2010) Floodplains, permafrost, cottonwood trees, and peat: What happened the last time climate warmed suddenly in Arctic Alaska? *Quaternary Science Reviews* 29: 3812–3830.

- Mann DH, Peteet DM, Reanier RE, and Kunz ML (2002) Responses of an arctic landscape to Lateglacial and early Holocene climatic changes: The importance of moisture. *Quaternary Science Reviews* 21: 997–1021.
- Matthews JV (1974a) Quaternary environments at Cape Deceit (Seward Peninsula, Alaska): Evolution of a tundra ecosystem. *Geological Society of America Bulletin* 85: 1353–1384.
- Matthews JV (1974b) Wisconsin environment of interior Alaska: Pollen and macrofossil analysis of a 27 meter core from the Isabella Basin (Fairbanks, Alaska). *Canadian Journal of Earth Sciences* 11: 828–841.
- Matthews JV, Schweger CE, and Hughes OL (1990a) Plant and insect fossils from the Mayo Indian Village section (central Yukon): New data on middle Wisconsin environments and glaciation. *Géographie Physique et Quaternaire* 44: 15–26.
- Matthews JV, Schweger CE, and Janssens JA (1990b) The last (Koy-Yukon) interglaciation in the northern Yukon evidence from unit 4 at Ch'ijee's Bluff, Bluefish Basin. *Géographie Physique et Quaternaire* 44: 341–362.
- Miller GH, Mode WN, Wolfe AP, et al. (1999) Stratified interglacial lacustrine sediments from Baffin Island, Arctic Canada: Chronology and paleoenvironmental implications. *Quaternary Science Reviews* 18: 789–810.
- Muhs DR, Ager TA, Bettis EA, et al. (2003) Stratigraphy and palaeoclimatic significance of Late Quaternary loess-palaeosol sequences of the Last Interglacial-Glacial cycle in central Alaska. *Quaternary Science Reviews* 22: 1947–1986.
- Payette S, Eronene M, and Jasinski JJP (2002) The circumboreal tundra-taiga interface: Late Pleistocene and Holocene changes. *Ambio Special Report; Tundra-Taiga Treeline Research* 12: 15–22.
- Peros MC and Gajewski K (2008) Holocene climate and vegetation change on Victoria Island, western Canadian Arctic. *Quaternary Science Reviews* 27: 235–249.
- Peros M, Gajewski K, Paull T, Ravindra R, and Podritske B (2010) Multi-proxy record of postglacial environmental change, south-central Melville Island, Northwest Territories, Canada. *Quaternary Research* 73: 247–258.
- Reger RD, Sturmman AG, Berg EE, and Burns PAC (2007) *A Guide to the Late Quaternary History of Northern and Western Kenai Peninsula, Alaska. Division of Geological and Geophysical Surveys, Guidebook 8*. Anchorage: State of Alaska Division of Geological and Geophysical Surveys.
- Reimer PJ, Baillie MGL, Bard E, et al. (2004) IntCal04 terrestrial radiocarbon age calibration, 0–26 kyr BP. *Radiocarbon* 46: 1029–1058.
- Richard PJH (1981) *Paléophytogéographie Postglacaire en Ungava. Collection Paleo-Quebec, No. 13*. Montréal: Université du Québec à Montréal.
- Ritchie JC (1984) *Past and Present Vegetation of the far Northwest of Canada*. Toronto: University of Toronto Press.
- Ritchie JC (1987) *Postglacial Vegetation of Canada*. Cambridge: Cambridge University Press.
- Ritchie JC, Cwynar LC, and Spear RW (1983) Evidence from north-west Canada for an early Holocene Milankovitch thermal maximum. *Nature* 305: 126–128.
- Ruhland K, St Jacques JM, Beierle BD, Lamoureux SF, Dyke AS, and Smol JP (2009) Lateglacial and Holocene paleoenvironmental changes recorded in lake sediments, Brock Plateau (Melville Hills), Northwest Territories, Canada. *The Holocene* 19: 1005–1016. <http://dx.doi.org/10.1177/0959683609340999>.
- Schweger CE and Matthews JV (1991) The last (Koy-Yukon) interglaciation in the Yukon: Comparisons with Holocene and interstadial pollen records. *Quaternary International* 10–12: 85–94.
- Seppä H, Cwynar LC, and MacDonald GM (2003) Post-glacial vegetation reconstruction and a possible 8200 cal. yr BP event from the low arctic of continental Nunavut, Canada. *Journal of Quaternary Science* 18: 621–629.
- Short SK, Mode WN, and Davis PT (1985) The Holocene record from Baffin Island: Modern and fossil pollen studies. In: Andrews JT (ed.) *Quaternary Environments*, pp. 608–642. Boston, MA: Allen & Unwin.
- Short SK and Nichols H (1977) Holocene pollen diagrams from subarctic Labrador-Ungava: Vegetational history and climate change. *Arctic and Alpine Research* 9: 265–290.
- Sorenson CJ, Knox JC, Larsen JA, and Bryson RA (1971) Paleosols and the forest border in Keewatin, N.W.T. *Quaternary Research* 1: 468–473.
- Spear RW (1993) The palynological record of Late-Quaternary Arctic tree-line in northwest Canada. *Review of Palaeobotany and Palynology* 79: 99–111.
- Stuiver M and Reimer PJ (1993) Extended ¹⁴C database and revised CALIB radiocarbon calibration program. *Radiocarbon* 35: 215–230.
- Szeicz JM and MacDonald GM (2001) Montane climate and vegetation dynamics in easternmost Beringia during the Late Quaternary. *Quaternary Science Reviews* 20: 247–257.
- Viau AE, Gajewski K, Sawada MC, and Bunbury J (2008) Low- and high-frequency climate variability in eastern Beringia during the past 25 000 years. *Canadian Journal of Earth Sciences — Revue Canadienne des Sciences de la Terre* 45: 1435–1453.
- Viereck LA and Little EL (1972) *Alaska Trees and Shrubs. Agriculture Handbook No. 410*. Washington, DC: U.S. Department of Agriculture.
- Viereck LA and Little EL (1975) Volume 2. *Alaska Trees and Common Shrubs Atlas of United States Trees*. Washington, DC: United States Department of Agriculture, Miscellaneous, Publication 1293.
- Wetterich S, Grosse G, Schirmeister L, et al. (2012) Late Quaternary environmental and landscape dynamics revealed by a pingo sequence on the northern Seward Peninsula, Alaska. *Quaternary Science Reviews* 39: 26–44.
- Wolf VG (2001) *A Window to the Past: Macrofossil Remains from an 18,000 Year-Old Buried Surface, Seward Peninsula, Alaska*. Unpublished MS thesis, University of Alaska Fairbanks.
- Wooler MJ, Zazula GD, Blinnikov M, et al. (2011) The detailed palaeoecology of a mid-Wisconsinan interstadial (ca. 32 000 14C a BP) vegetation surface from interior Alaska. *Journal of Quaternary Science* 26: 746–756. <http://dx.doi.org/10.1002/jqs.1497>.
- Yu Z, Walker KN, Evenson EB, and Hajdas I (2008) Lateglacial and early Holocene climate oscillations in the Matanuska Valley, south-central Alaska. *Quaternary Science Reviews* 27: 148–161.
- Zazula GD, Schweger CE, Beaudoin AB, and McCourt GH (2006a) Macrofossil and pollen evidence for full-glacial steppe within an ecological mosaic along the Bluefish River, eastern Beringia. *Quaternary International* 142–143: 2–16.
- Zazula GD, Telka AM, Harington CR, Schweger CE, and Mathewes RW (2006b) New spruce (*Picea* spp.) macrofossils from Yukon Territory: Implications for late Pleistocene refugia in Eastern Beringia. *Arctic* 59: 391–400.

Relevant Websites

<http://www.neotomadb.org/index.php/>. – Neotoma Paleocology Database.
<http://pmp2.lsce.ipsl.fr/>. – Website.

South America

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General Setting of South America for Paleoecological Studies

The late Pleistocene is taken here to cover the last ~130 000 years (130 ka) and includes the last and the present interglacial, separated by the last glacial. Many pollen records showing vegetation dynamics reflect the Holocene and the Late Glacial interval. Several records also reflect older parts of the last glacial, and only in a few sedimentary basins in South America, mainly in the Colombian Andes, can pollen records that represent the full late Pleistocene be retrieved. On a continental scale, multisite paleoecological results were summarized by Flenley (1979), Markgraf (1993), and Morley (2000). Multisite studies of more regional importance include Marchant et al. (2001), Van der Hammen and Hooghiemstra (2000), and Heusser (2003).

The South American continent includes nearly all biomes: from the wettest rainforest in Pacific Colombia (Chocó) with over 12 000 mm annual precipitation to extremely dry conditions in the Atacama Desert with less than 100 mm annual precipitation; from Amazonian forest types, ranging from evergreen to raingreen (semideciduous) and dry forest, to cold and wet páramo in the northern Andes, dry puna in the Central Andes, and wet Magellanic moorlands in Patagonia. Aspects of biogeography and late Quaternary history have been covered by Prance (1982), Prance and Lovejoy (1985), Vuilleumier and Monasterio (1986), Whitmore and Prance (1987), and Markgraf (2001).

Long Pollen Records, Biome Evolution, and Nonanalog Vegetation Communities

The Eastern Cordillera of Colombia includes remarkably deep basins, where, during long intervals of the Pleistocene, subsidence maintained the equilibrium with sediment infill. The altitudinal position of the surface of these basins (~2550 m) is half the amplitude of the altitudinally shifting upper forest line (UFL) during a glacial–interglacial cycle, viz. from around 3200 m at the present day to about 2000 m during a glacial (Figure 1). Along the altitudinal gradient, distinct vegetation belts can be recognized (Van der Hammen, 1974): tropical lowland forest from 0 to 1000 m, sub-Andean forest (=lower montane forest) from 1000 to 2300 m, Andean forest (=upper montane forest) from 2300 to 3200 m, dwarf forest and shrub in the subpáramo from 3200 to 3500 m, grass-páramo from 3500 to 4200 m, and superpáramo from 4200 to 4800 m (Figure 2). Each altitudinal belt is characterized by a distinct flora that has left identifiable pollen grains in the sediments. Downcore changes in the proportion of pollen grains originating from different vegetation belts show altitudinal shifts of those belts.

Most important are the basins of Fúquene (5°27'N, 73°46'W, 2580 m) and Bogotá (4°50'N, 74°12'W, 2550 m); which are located some 100 km apart. Constructing a robust timeframe for these mainly lacustrine sediments is difficult, and interpretations have changed over time. The pollen record from the almost 600-m-deep Bogotá basin starts some 3 Ma (e.g., Andriessen et al., 1993; Torres et al., 2005) but ends around 30 ka when the paleolake Bogotá drained. The depth of the Fúquene basin, separated from the Bogotá basin by a watershed of 3200 m altitude, is unknown, and so is the age of the oldest sediments. But sediments have continued to be deposited in Lake Fúquene up to the present time. Therefore, the last 30 ka of the pollen-based history of vegetation and climate change, not present in the Bogotá basin, can be found in the sediments of Lake Fúquene.

From these long Pleistocene records, we know that during the late Pleistocene, montane forests and páramo vegetation of the northern Andes have had a long and dynamic history, characterized by many repetitive climatic cycles and immigration events. The north Andean flora received trees migrating from the south (e.g., *Polylepis*, an Austral-Antarctic element) and from the Northern Hemisphere (e.g., *Alnus* and *Quercus*, both Holarctic elements). On timescales of several millions of years, the pollen record shows a continuum of change, driven by geological changes (e.g., tectonic changes in the basin and closure of the Panamanian Isthmus), evolution of ecosystems, and global climate change. As a consequence, late Pleistocene ecosystems are the product of a long history of change. The prelude of the late Pleistocene is documented in the pollen records of the 357-m-deep core Funza-1 and the 586-m-deep core Funza-2, plotted after the 2005 chronology (Torres et al., 2005 and unpublished data). The precise interpretation of pollen records is based on calibration sets of recent pollen rain (Figure 3). Moss samples collected along an altitudinal transect are prepared in the same way as the samples making up a pollen diagram. The results (Figure 4) show a clear relationship between the modern dominant pollen taxa and altitude. This relationship is applied to intervals of the past, assuming this relationship remains valid. This assumption should be treated with caution: the long pollen record of Funza has shown that many north Andean forest communities of the past do not have a modern analog. Also, the páramo vegetation composition during an interglacial setting (during full Quaternary time, this setting prevailed over ~10% of the time only) seems different from the vegetation composition during glacial conditions, when both low temperatures and low atmospheric carbon dioxide (CO₂) concentrations (low *p*CO₂) prevailed (Boom et al., 2001).

A molecular paleobotanical study of sediments from the Bogotá basin led to the hypothesis that the altitudinal distribution of C₃ and C₄ plants during modern time differs from that during glacial times (the hypothesis was formulated for

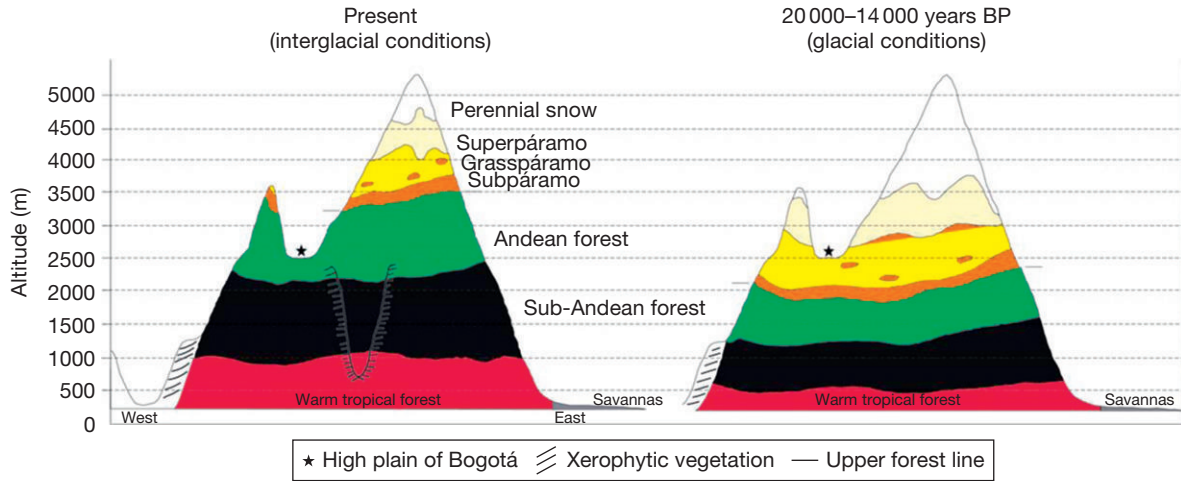


Figure 1 Altitudinal distribution of main vegetation belts in the Colombian Andes for the present day and the Last Glacial Maximum (LGM). Modified from Van der Hammen T (1974) The Pleistocene changes of vegetation and climate in tropical South America. *Journal of Biogeography* 1: 3–26; Wille M, Hooghiemstra H, Behling H, Van der Borg K, and Negret AJ (2001) Environmental change in the Colombian subandean forest belt from 8 pollen records: The last 50 ka. *Vegetation History and Archaeobotany* 10: 61–77.



Figure 2 Photographs of distinct vegetation belts: (a) superpáramo, (b) grasspáramo, (c) subpáramo, and (d) Andean forest.

the Last Glacial Maximum (LGM; **Figure 5**)). Under modern conditions, C_4 plants (herbs resistant to dryness and, potentially, photosynthesis, do not suffer from low pCO_2) occur at low elevations (below ~ 2000 m) in the inter-Andean valleys (dry because of a strong effect of the rain shadow) and the savannas. However, during glacial times, C_4 plants had a

competitive advantage over C₃ plants (including all trees and many herbs) and migrated upslope. As a consequence, it is hypothesized that the present-day páramo vegetation, dominated by C₃ herbs, was dominated by C₄ herbs, including *Sporobolus lasiophyllus*, which is rare in the modern páramo, during the LGM (Boom et al., 2001; 2002). Although results have to be

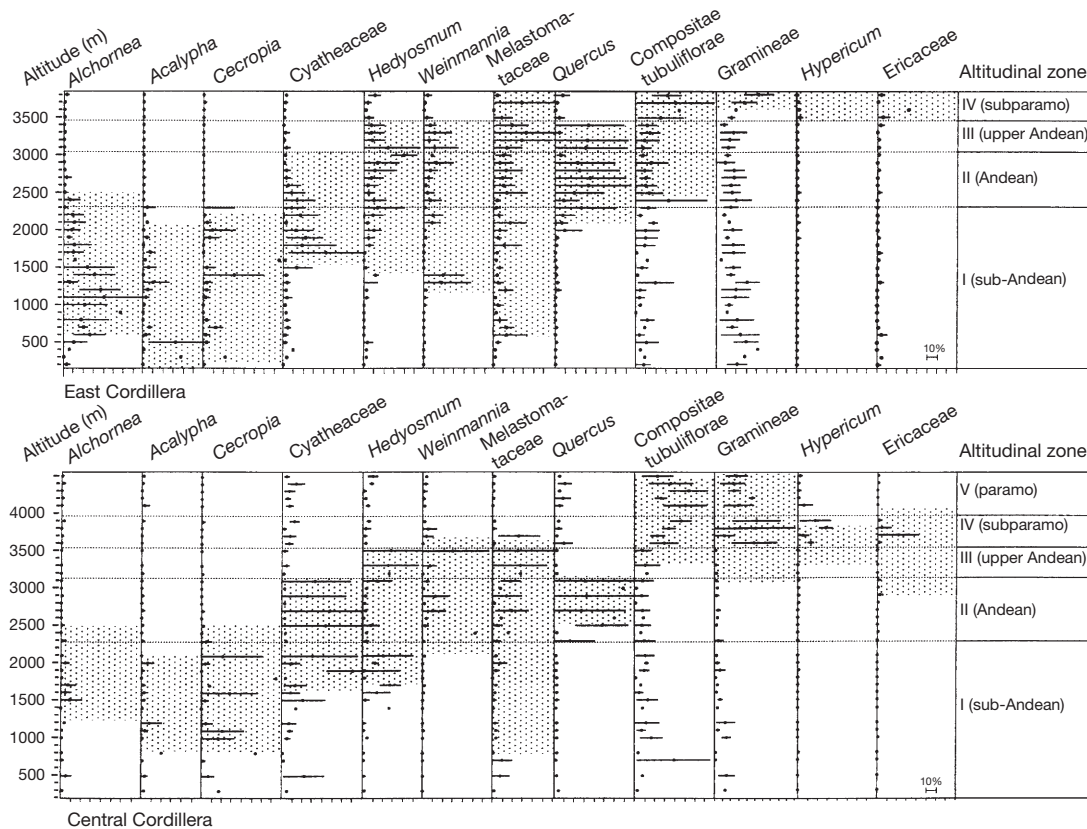


Figure 3 Modern pollen rain. Changes in pollen representation of surface samples in altitudinal intervals of 100 m. The present-day altitudinal range of the taxa is shaded. Reproduced from Van't Veer R and Hooghiemstra H (2000) Montane forest evolution during the last 650,000 years in Colombia: A multivariate approach based on pollen record Funza-I. *Journal of Quaternary Science* 15: 329–346.

confirmed, here is another example that past vegetation communities may reflect nonanalogs of present-day vegetation.

The geologically recent immigration of oak (*Quercus*) into South America is most relevant for the late Pleistocene of northern South America (Hooghiemstra, 2006). This Northern Hemisphere tree migrated into South America when the Panamanian Isthmus formed, around 4 Ma. Oak migrated slowly and arrived in the area of the Bogotá basin around 475 ka (Van't Veer and Hooghiemstra, 2000). A zonal belt with oak forest developed after 330 ka. During the period 220–135 ka, zonal *Quercus* and *Weinmannia* forests partly replaced *Podocarpus*-dominated forest. *Quercus* possibly spread southward in the sub-Andean forest belt, using the north–south oriented valleys of the Magdalena and Cauca rivers. At several places, *Quercus* could easily have expanded upslope and reached inter-Andean plains, such as the high plains of Bogotá and Fúquene. Today, the southernmost extension of *Quercus* is at the Colombian–Ecuadorian border, indicating a relatively low migration speed.

Late Pleistocene Dynamics of the Northern Andes

From the northern Andes, pollen records are available that show the dynamic history of vegetation change with great detail and the inferred sequence of climate change. Pollen

record Fúquene-2 (Van Geel and Van der Hammen, 1973) shows the interval from ca. 32 ka to the present with a temporal resolution of 300 years (Figure 6). The partly overlapping pollen record Fúquene-7C (Mommersteeg and Hooghiemstra, 2007) shows the interval from ca. 88 ka to the present with an average temporal resolution of 450 years. Seven carbon-14 (^{14}C) ages provide chronological control for the upper 40 100 calendar years before present (cal years BP), and the relationship between orbital forcing and lake-level stands was used to provide the chronology for the lower part of the record (Mommersteeg and Hooghiemstra, 2007).

Ten main intervals were identified and were related to the chronostratigraphy of marine isotope stages (MIS): 88.5–84.0 ka (MIS 5b), 84.0–70.5 ka (MIS 5a), 70.5–64.0 ka (MIS 4), 64.0–58.0 ka (MIS 4), 58.0–47.0 ka (MIS 3.3), 47.0–38.0 ka (MIS 3), 38.0–26.8 ka (MIS 3.1), 26.8–16.4 ka (MIS 2), 16.4–8.0 ka, and 8.0–6.3 ka. The UFL shifted between 2000 and 3300 m, and the inferred temperature depression varied from 1.5 to 2.0 °C at MIS 5.1, 3.0 to 4.0 °C at MIS 4, 1.0 to 2.0 °C at MIS 3.3, 2.5 to 4.0 °C at MIS 3.1, and 4.5 to 8.0 °C at MIS 2 (LGM); temperatures were similar to present-day temperatures during 8.0–6.3 ka. Precipitation change is inferred from forest composition and lake-level changes. This pollen record shows that forest near the UFL was dominated by *Quercus* (e.g., from ca. 60 to 47 cal years BP), and *Polylepis* (e.g., from ca. 47 to 36 cal years BP) in an alternating manner.

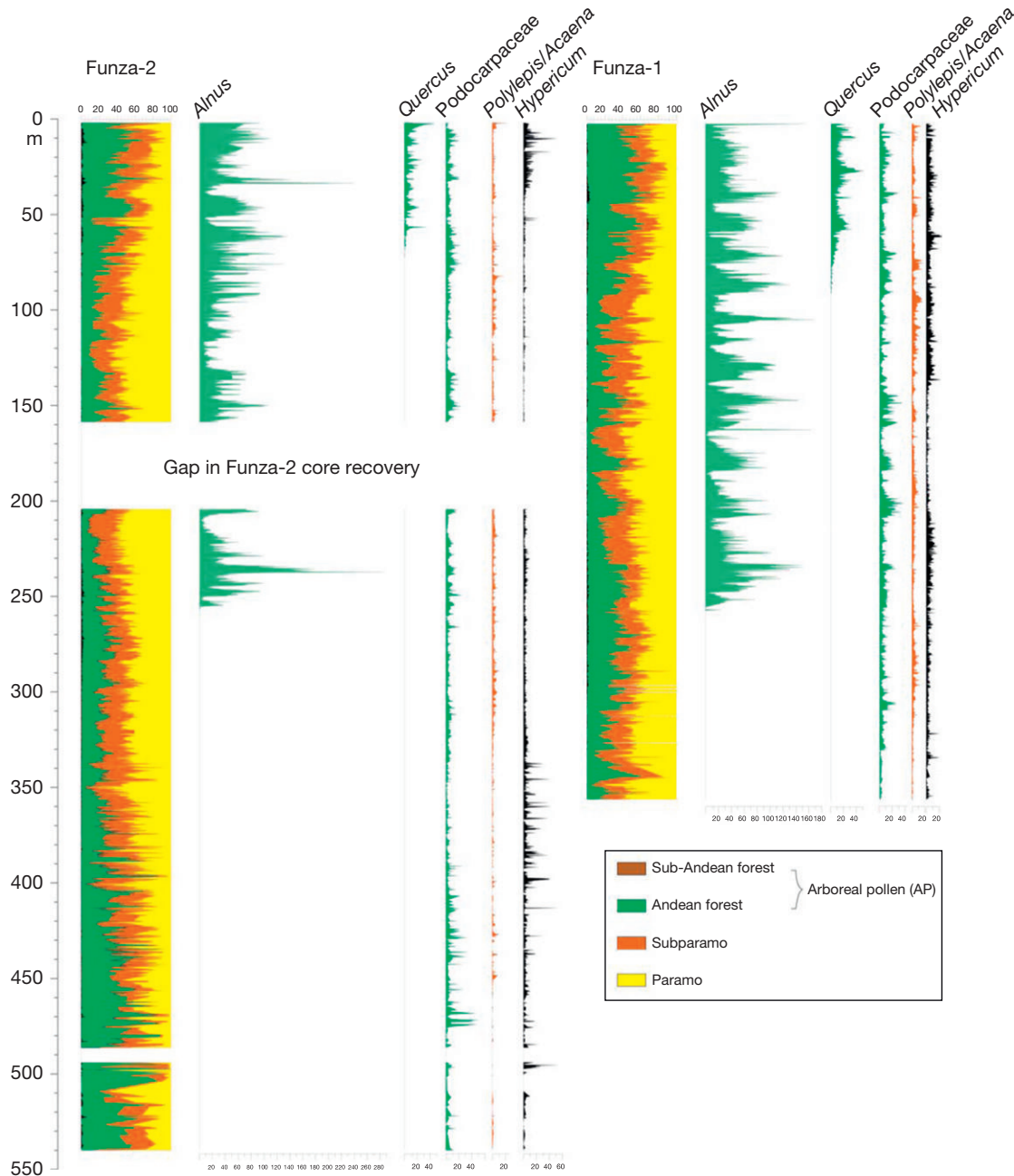


Figure 4 Main pollen diagrams of cores Funza-1 and Funza-2 from the sedimentary basin of Bogotá (Colombia) showing how late Pleistocene vegetation and climate change was preceded by a long series of Pleistocene ice ages. Reproduced from Torres and Hooghiemstra, unpublished data.

In the case where climatic conditions are centrally in the ecological envelope of a taxon, a given change in climate conditions may have less effect than in the case (for another taxon) where climatic conditions are marginal and close to the limits of the ecological envelope (Thompson et al., 2000). In other words, a central position in the ecological envelope may cause a low sensitivity to climate change while the same amount of change may cause significant vegetation change in a marginal position or even force a taxon to disappear. Therefore, individual taxa respond to environmental change, but as vegetation belts have many taxa in common that respond

almost similarly, vegetation belts too seem to respond as a quasi-unit. The significant difference between *Polylepis*-dominated and *Quercus*-dominated forests in pollen records of Fúquene-7C suggests that floristic changes in the composition of ecotone forest may also occur (partly) independently of climate change. This pollen record suggests a sensitive balance between altitudinally organized competing ecosystems. It shows that minor changes in temperature conditions at millennial timescales are clearly reflected when the examined natural archive lies at a strategic position and temporal resolution is adequate. Main periodicities of climate-forced

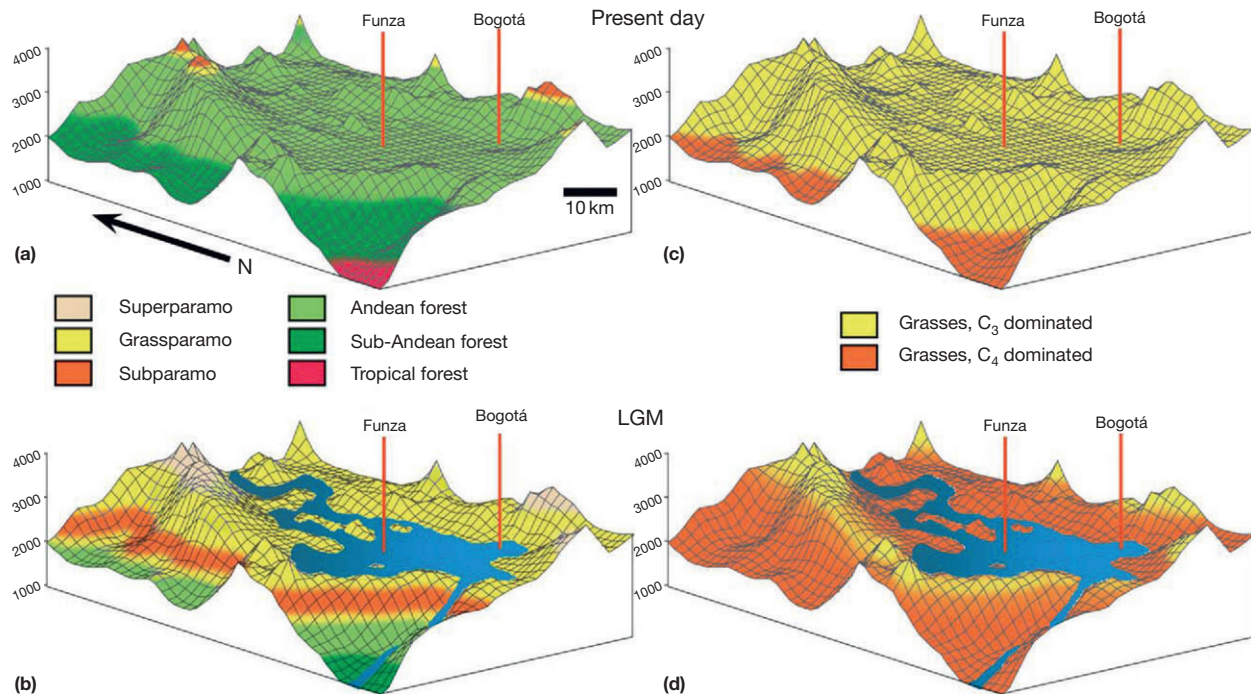


Figure 5 Altitudinal vegetation distribution in the area of the basin of Bogotá, Colombia. (a) Present-day vegetation belts; (b) multisite-based reconstruction of vegetation distribution during the LGM; (c) present-day altitudinal distribution of plants with a C₃ and C₄ photosynthetic metabolism: páramo vegetation is dominated by C₃ herbs; (d) tentative reconstruction for the LGM of maximum possible extent of C₄ grass distribution when $p\text{CO}_2$ is reduced to 180 parts per million V: páramo is tentatively dominated by C₄ herbs. Reproduced from Boom A, Marchant R, Hooghiemstra H, and Sinneghe Damsté JS (2002) CO₂- and temperature-controlled altitudinal shifts of C₄- and C₃-dominated grasslands allow reconstruction of paleo-atmospheric $p\text{CO}_2$. *Palaeogeography, Palaeoclimatology, Palaeoecology* 177: 151–168.

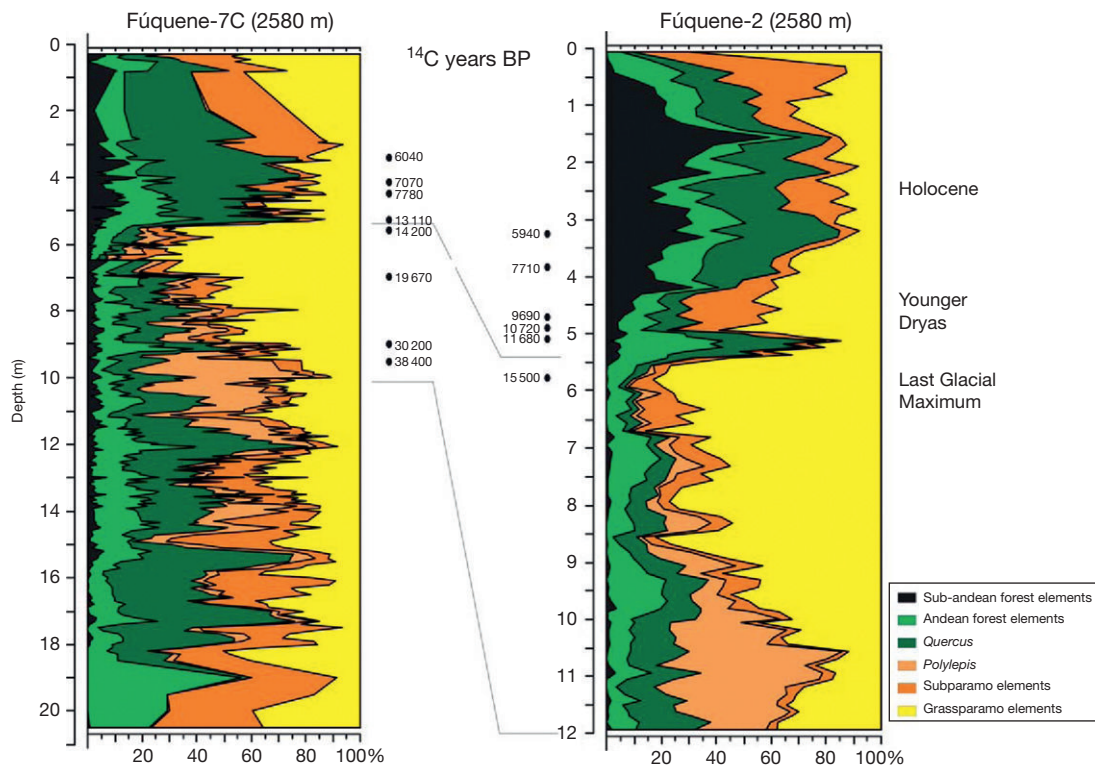


Figure 6 Main pollen diagram of core Fúquene-2 and core Fúquene-7C showing vegetation changes during the last 88000 cal years. During late glacial time, core Fúquene-7C has a hiatus and the Younger Dryas climate oscillation is reflected only in core Fúquene-2. Reproduced from Mommersteeg H and Hooghiemstra H (2007) Millennium-scale climate variability in the tropics from 88,000-yr pollen record Fúquene-7C (Colombia). *Journal of Quaternary Science*.

vegetation change, shown by the UFL record in particular, are concentrated in the 1000–4250-year frequency band.

An evaluation of the forest dynamics from pollen sites below 2000 m altitude indicated the degree to which modern vegetation belts were displaced and compressed during the LGM in the southern Colombian Andes (Table 1; Wille et al., 2001).

Considering the full altitudinal range of montane forests, the lower montane forest belt experienced the greatest compression during the last interglacial–glacial cycle; during the LGM, its altitudinal span was only 600 m, compared to 1300 m today. Although the lower limit of the lower montane forest belt is less clearly defined than its upper limit, the vertical interval is estimated at 55% of the present day; the difference in surface area is highly dependent on the locality under consideration. The salient ecophysical factor at the transition from the sub-Andean to the Andean forest belt is the occurrence of night frost; a few frosty nights may exclude frost-sensitive species and may have a decisive impact on the composition of the

vegetation. As a consequence, an explanation of the vertical vegetation distribution should be based more heavily on the ‘average temperature of the coldest month’ than on the frequently used ‘average annual temperature.’ This is a recommendation for the use of the ‘biome method’ (Prentice et al., 1992; elaborated for Colombia by Marchant et al., 2001) rather than the ‘Holdridge system’ to classify tropical ecosystems on ecophysiological grounds.

From phytogeographical studies, it became clear that the sub-Andean forest belt received many taxa from northern and southern latitudes by migration and from high and low elevations by adaptation (Van der Hammen and Cleef, 1986) and that it is floristically richer than the Andean forest belt. This high diversity coincides with a highly dynamic history on Quaternary timescales (Table 1). In the northern Andes, there are more coincidences between high diversity and a highly dynamic ecosystem. For example, the geologically youngest part of the Colombian Andes, namely, the Cocuy area in the Eastern Cordillera, shows the highest diversity of plants in the Espeletiinae group (Hooghiemstra et al., 2006; Van der Hammen and Cleef, 1986). Also, the páramo ecosystem is a good example of a biome with a very dynamic Pleistocene history coinciding with high species diversity: Figure 7 shows the archipelago of isolated páramo ‘islands’ on the highest mountain tops during modern (interglacial) conditions and extended areas with páramo vegetation during glacial intervals. Thus, periods of isolation (during ~20% of Pleistocene time), permitting allopatric speciation processes, are interrupted by periods with extended areas of páramo vegetation permitting

Table 1

Vegetation belt	Modern range (m)	LGM range (m)	Compression (%)
Tropical lowland forest	0–1000	0–800	~20
Lower montane forest	1000–2300	800–1400	~55
Upper montane forest	2300–3200	1400–2000	~33
Subpáramo, grasspáramo	3200–4200	2000–3000	~0

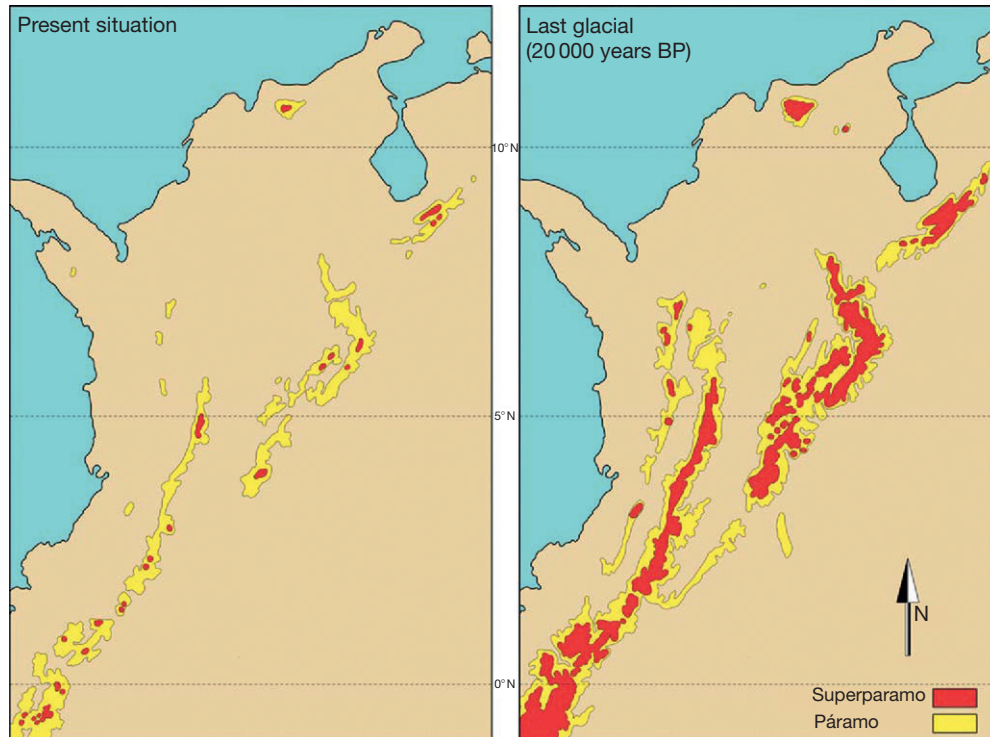


Figure 7 Extension and distribution of páramo during present-day and LGM conditions. The present setting reflects maximum isolation whereas the glacial setting offers opportunities for genetic material to be distributed over large areas. Reproduced from Van der Hammen T and Cleef AM (1986) Development of high Andean páramo flora and vegetation. In: Vuilleumier F and Monasterio M (eds.) *High Altitude Tropical Biogeography*, pp. 153–201. New York: Oxford University Press.

gene flow (during ~80% of Pleistocene time). The glacial-interglacial rhythm of expansion and contraction of the páramo area is considered an important mechanism to explain the high diversity of the high-Andean flora.

Pollen Record from the Colombian Savannas

The history of the savanna biome is best documented in the savannas of the Llanos Orientales in northeastern Colombia. This savanna area extends into Venezuela and represents the largest savanna ecosystem in northern South America (Huber, 1987; Huntley and Walker, 1982; Mistry, 2000; Sarmiento and Monasterio, 1975). Precipitation regimes are distinctly seasonal, related to the annually migrating Intertropical Convergence Zone (ITCZ), the main mechanism that brings seasonal variability in this ecosystem (Haug et al., 2001). Fires are frequent in the dry season and many trees have developed corky bark as a protection for fire (Mistry, 2000). Forest is mainly located in narrow corridors along the drainage system, forming gallery forest, and the higher elevations are covered by shrubby grasslands or pure grassland (Figure 8). Characteristic savanna trees are *Curatella americana* (Dilleniaceae), *Byrsonima crassifolia* (Malpighiaceae), and *Palicourea rigida* (Rubiaceae). Gallery forest shares many arboreal taxa with adjacent rainforests, but the palm *Mauritia flexuosa* is often abundant along streams and lake borders and forms characteristic palm forest (morichal). The open vegetation of the savannas is characterized by grasses. Many poaceous genera are represented in the savanna, but in paleoecological studies, pollen morphology does not distinguish between poaceous genera, and the proportion of grassy vegetation is reflected as one record.

Neotropical savannas occur north and south of the equator, separated by the Amazonian rainforest area where annual precipitation is minimally 1500–1800 mm and maximally a short (2 months), dry interval (Behling and Hooghiemstra, 2001; Van der Hammen and Hooghiemstra, 2000). Apart from climatic savannas, edaphic savannas also occur: these savannas lie on porous sandstone tablelands where the local vegetation is hardly able to make use of the abundant precipitation. A pollen diagram from such edaphic savanna is available

from the sandstone plateau of Araracuara, near the Caquetá River (Colombian Amazon). This pollen record shows that *Bonnetia* shrub with moraceous and melastomataceous trees alternates with open herb vegetation dominated by Rapatea-ceae, Poaceae, and *Xyris* (Berrio et al., 2003).

Figure 9 shows five pollen records from savanna lakes located along a 500-km east–west transect through the Llanos Orientales. They include Laguna Sardinas (4°58' N, 69°28' W, 180-m elevation), Laguna Angel (4°28' N, 70°34' W, 200-m elevation; Behling and Hooghiemstra, 1998), El Piñal (4°04' N, 70°14' W, 180-m elevation; Behling and Hooghiemstra, 2001), Laguna Loma Linda (3°18' N, 73°23' W, 310-m elevation), and Laguna Las Margaritas (3°23' N, 73°26' W, 290-m elevation; Wille et al., 2003). The openness of a savanna, indicating the degree of climatic humidity and the length of the dry period, is reflected in these pollen diagrams by the proportions of pollen of Poaceae, savanna shrub (*Byrsonima* and *Curatella*), and the group of elements of the gallery forest (from dry to more humid), respectively. Radiocarbon dates provide a robust time control and show that the full interval since the LGM is included. The general conclusions of these five pollen records and two additional pollen records from Laguna Carimagua (4°04' N, 70°13' W, 180-m elevation; Berrio et al., 2000) and Laguna Chenevo (4°05' N, 70°21' W, 150-m elevation; Berrio et al., 2002) are shown in Figure 10.

At the LGM, climatic conditions were very dry and these conditions continued up to the Late Glacial. Between ca. 10700 and 10070 ¹⁴C years BP (at El Piñal site up to 9200 ¹⁴C years BP), gallery forest had expanded, indicating less dry conditions. In all pollen records, the Holocene started between 9700 and 9200 ¹⁴C years BP with dry to very dry conditions that continued for several millennia. Climatic conditions became wetter between 7100 and 6100 ¹⁴C years BP in the western section of the savanna area close to the foothills of the Cordillera, and between 6400 and 5300 ¹⁴C years BP in the eastern section.

All pollen records show another increase in climatic humidity in the short interval between 4000 and 3600 ¹⁴C years BP. A southward migration of the average position of the ITCZ possibly supplied more moisture from the Atlantic Ocean to these latitudes (Brenner et al., 2001; Haug et al., 2001; Marchant and Hooghiemstra, 2004), causing drier climates in the

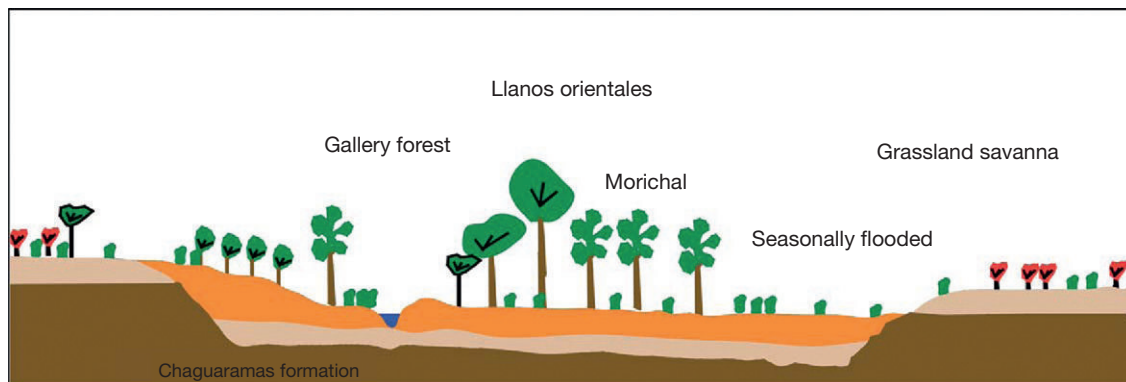


Figure 8 Schematic cross-section through the Colombian savanna area showing the relationship between landscape topography, drainage system, and vegetation. Reproduced from Berrio JC, Hooghiemstra H, Behling H, Botero P, and Van der Borg K (2002) Late-Quaternary savanna history of the Colombian Llanos Orientales from Lagunas Chenevo and Mozambique: A transect synthesis. *The Holocene* 12: 35–48.

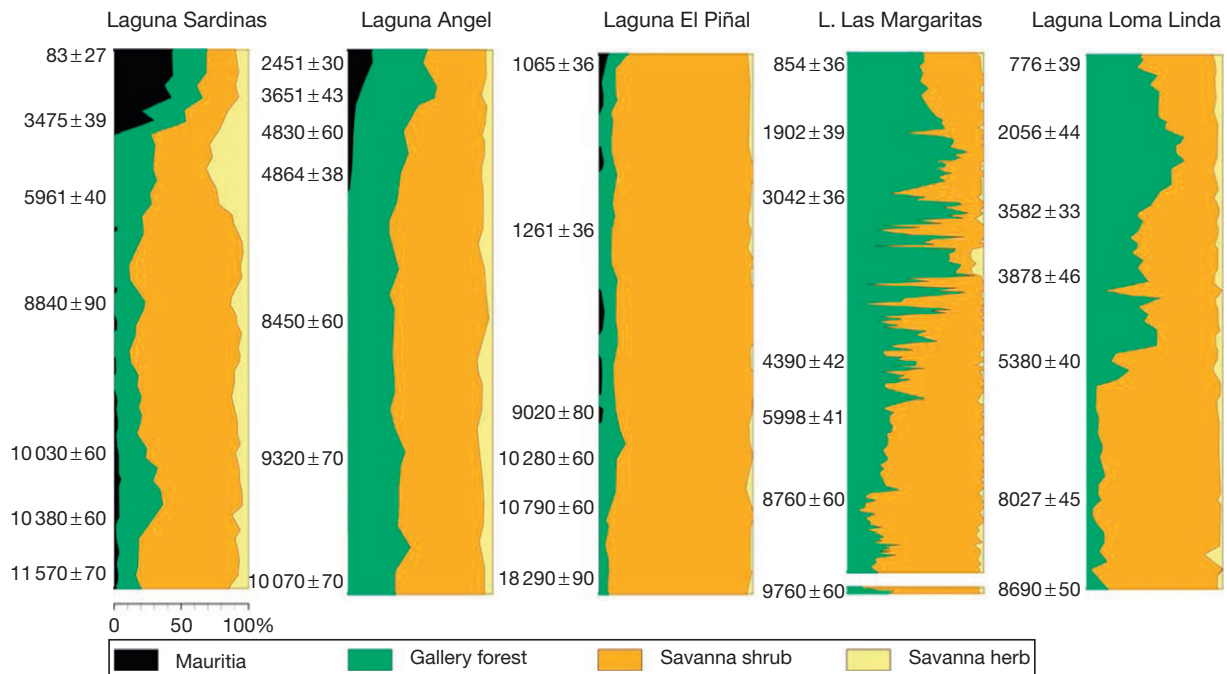


Figure 9 Main pollen diagrams of five sites located on a transect through the savanna of the Llanos Orientales in Colombia. From left (east) to right (west) are shown Lake Sardinas, Lake Angel, Lake El Piñal, Lake Las Margaritas and Lake Loma Linda. Reproduced from Berrio JC, Hooghiemstra H, Behling H, Botero P, and Van der Borg K (2002) Late-Quaternary savanna history of the Colombian Llanos Orientales from Lagunas Chenevo and Mozambique: A transect synthesis. *The Holocene* 12: 35–48; Behling H and Hooghiemstra H (2000) Holocene Amazon rain forest-savanna dynamics and climatic implications: High resolution pollen record Laguna Loma Linda in eastern Colombia. *Journal of Quaternary Science* 15: 687–695; Wille M, Hooghiemstra H, Van Geel B, Behling H, De Jong A, and Van der Borg K (2003) Submillennium-scale migrations of rainforest-savanna boundary in Colombia; ^{14}C wiggle-matching and pollen analysis of core Las Margaritas. *Palaeogeography, Palaeoclimatology, Palaeoecology* 193: 201–223, with permission.

Caribbean and moister conditions at more equatorial latitudes. Increased precipitation levels allowed *Mauritia* and *Mauritiella* palms, which were rare during the previous interval, to become abundant in the Llanos Orientales between 4000 and 3700 ^{14}C years BP. These palms formed extensive stands on soils with stagnating water. This climate-driven vegetation change may have coincided with the first human settlements in this savanna area. Expansion of settlements in the savannas was possibly also related to more favorable (wetter) climate conditions. Natural climate change forced vegetation change but may also have triggered expansion and migration of human populations, leaving mixed signals in pollen records which are often difficult to disentangle.

How Stable/Dynamic is the Amazonian Rainforest? The Pollen Evidence

The nature of the Ice Age vegetation of the Amazon remains highly controversial. Explanations of the huge biodiversity in the Amazonian rainforest are based on a variety of assumptions. The scarce data from diverse proxies (pollen, plant macrofossils, lake levels, paleodune fields, noble gases, stable isotopes, sediment sections, etc.) are interpreted in directions that best conform to these assumptions, hindering a quasi-objective, evidence-based interpretation of the available data. The papers by Colinvaux et al. (2000) and Van der Hammen

and Hooghiemstra (2000) are good reflections of the state of the art and convincingly indicate that more research at strategic sites is needed before a conclusive decision on Haffer's forest refugia hypothesis can be made. Logistically, the Amazonian rainforest biome is difficult to access for researchers. But it is even more important to find adequate sites that contain well-preserved pollen records (and other proxies) to document the dynamics of the Amazonian rainforest biome. The *terra firma* area, of relatively high elevation and never inundated by the Amazonian drainage system, has thus far offered few sites with pollen records. The low areas of Amazonia include the presently active drainage system and the wide river valleys that were filled with riverine sediments supplied by these drainage systems in the past, when these rivers had different courses. The latter explains the limitations of paleoecological research in the abandoned and now active parts of the river valleys: rivers were changing their courses continuously, preventing sites from accumulating sediments during long periods of time. Therefore, the few available pollen records from rainforest sites mostly come from abandoned meanders and rarely cover an interval longer than the Holocene. Once a pollen record from a rainforest area has been produced, there are practical problems: a significant proportion of the trees are palynologically 'silent,' that is, they do not produce large numbers of pollen grains because they are insect pollinated. Moreover, a large number of trees recognized by their pollen grains have undefined 'ecological envelopes,' hindering the ecological categorization of many

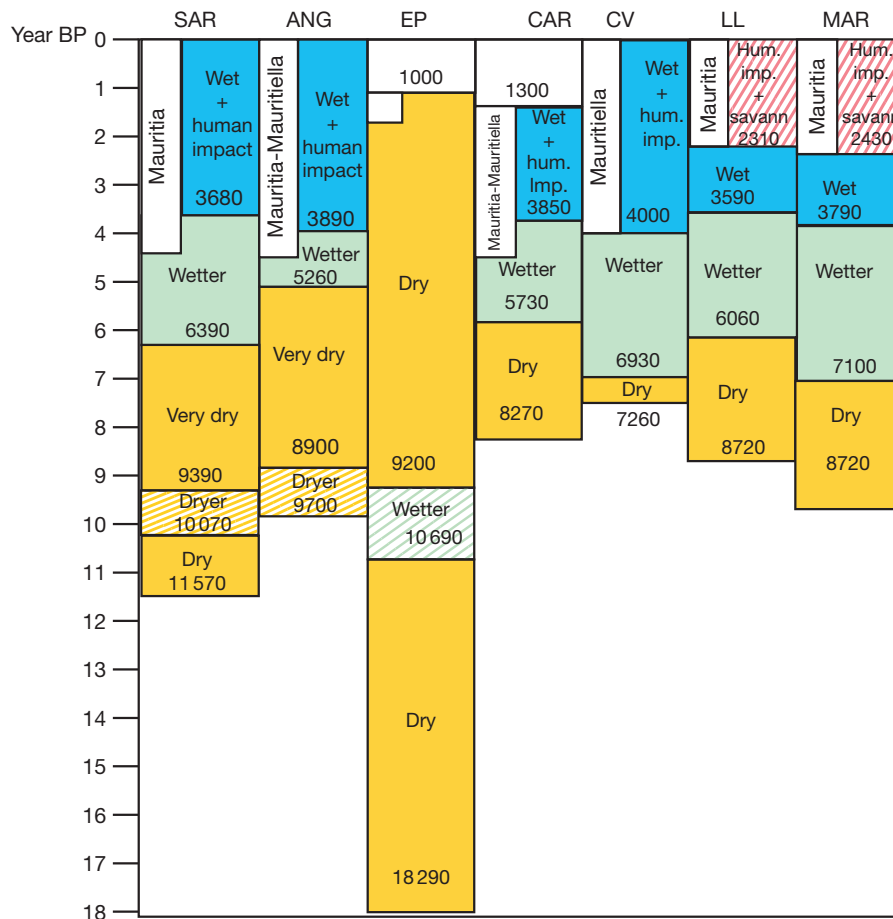


Figure 10 Synthesis of environmental change in the savannas of the Llanos Orientales in Colombia from seven sites reflecting the last 18,000 years. From left (east) to right (west) are shown Laguna Sardinas, Laguna Angel, Laguna El Piñal, Laguna Carimagua, Laguna Chenevo, Laguna Loma Linda, and Laguna Las Margaritas.

records (each with low representation) that are needed to reconstruct details of environmental change.

Pollen records spanning longer intervals are located at the outer borders of the Amazonian rainforest: the Carajás site in southeast Brazil (Absy et al., 1991), and Lagunas Bella Vista and Chaplin in the Bolivian Amazon (Mayle et al., 2000). Site La Pata (Colinvaux et al., 1996) is the only pollen record that shows vegetation dynamics from the 'interior' of the rainforest during the Holocene and a significant part of the last glacial. Figure 11 shows the locations of the most important pollen sites. The centrally located site, La Pata, shows the continuous presence of forest, but that forest changed into a drier type during the LGM (Ledru et al., 1998). The marginally located sites, Carajás and Chalpin, show alternations between periods of forest (wet climatic conditions) and periods of dry vegetation (varying from dry forest to savanna). Interestingly enough, these alternations in humidity in the marginal areas seem related to precession forcing of climate change.

The refugia hypothesis of Haffer (1969, 1982) postulates that during the LGM, the Amazon basin was covered with a significant surface of dry, savanna-like vegetation and rainforest had retreated to a number of isolated refugia. This theory is neither validated nor adequately rejected by fossil data. There is

progress in understanding speciation processes. Apart from the classical mechanism of allopatric speciation, sympatric speciation is a process that stimulates thinking in terms of forest refugia. This kind of speciation is now widely accepted as a relevant mechanism to understand how the Amazonian rainforest could have reached such high levels of biodiversity. In general, precipitation change has a more severe effect on rainforest composition and distribution than low-scale temperature change (estimates for the LGM temperature cooling in Amazonia vary at 3–6 °C; Farrera et al., 1999). The main forcing factor for moisture change, brought about by orbitally forced migrations of the ITCZ, is a stable factor of the climate system. Contrary to this, however, temperature change in Amazonia possibly increased greatly at the onset of Pleistocene ice ages. Therefore, it is doubtful whether Pleistocene climates of the Amazonian rainforest should be characterized as 'dynamic,' as is the case for the páramo biome.

Relatively basic problems in Amazonian paleoecology remain unsolved: there is much discussion on how the presence of pollen grains of cool montane trees (*Podocarpus* in particular) should be interpreted, in terms of LGM temperature. *Podocarpus* has important representatives in the cool Andes, but several species also inhabit Amazonian lowlands. As the

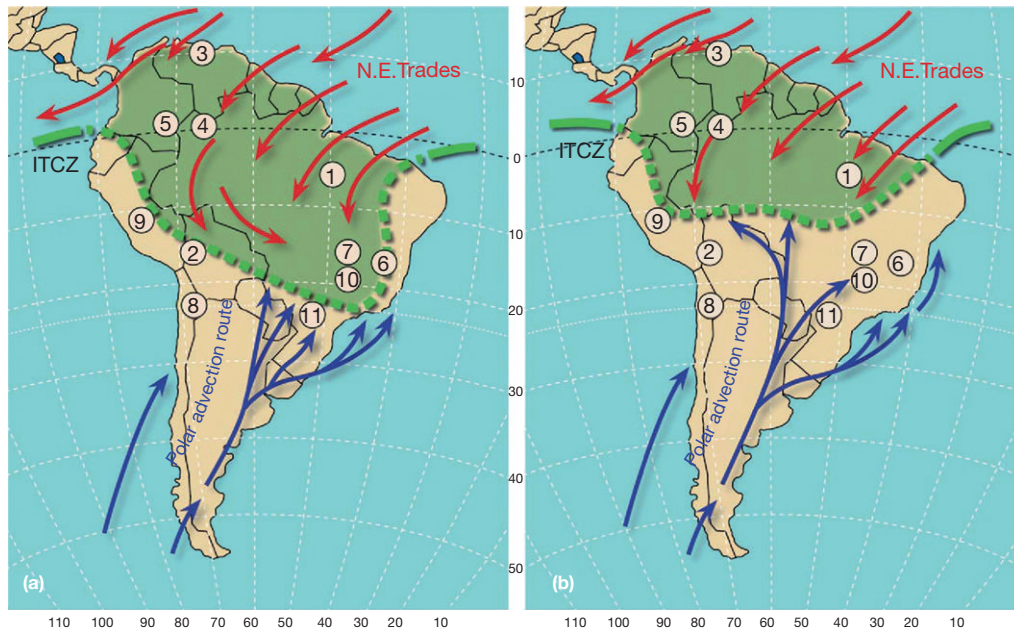


Figure 11 Map showing main pollen sites in and around the Amazonian rainforest area (sites 1–11) and the mean position of the ITCZ at the present day (a) and between 12 400 and 8800 cal years BP (b). Sites are (1) Serra dos Carajas, Brazil; (2) Lake Titicaca, Bolivia; (3) Lake Valencia, Venezuela; (4) São Gabriel da Cachoeira, Brazil; (5) Rio Caquetá, Colombia; (6) upper Rio Doce, Brazil; (7) Brasília, Brazil; (8) Atacama Desert, Chile; (9) Junin Plain, Peru; (10) Salitre, Brazil; (11) Araucaria forest area, Brazil. The shaded continental area delimits the zone of influence of the moist equatorial air masses (northeast trades) east of the Andes; warming of the continent during the austral summer induces a significant southward migration of the ITCZ. Reproduced from Martin L, Bertaux J, Corrège T, et al. (1997) Astronomical forcing of contrasting rainfall changes in tropical South America between 12,400 and 8800 cal yr BP. *Quaternary Research* 27: 117–122.

relation between climatic parameters and distribution of *Podocarpus* species is not well established, debates on the interpretation of *Podocarpus* peaks in pollen records continue (Behling and Lichte, 1997; Bush et al., 1990; Van der Hammen and Hooghiemstra, 2000). The cooperative initiatives between modelers and paleoecologists of the 1990s (Thompson et al., 2000) need to be followed up.

Understanding how the climate system operates is crucial to improve our interpretation of proxy records of environmental change. Martin et al. (1997) were able to explain the contrasting rainfall changes in tropical South America between 12 400 and 8800 cal years BP, clarifying a mechanism of more general importance for the tropical regions of South America. They compared Lake Titicaca (16°S, 69°W) in Bolivia with Lake Valencia (10°N, 67°W) in Venezuela. Between 12 400 and 8800 cal years BP, the water level of Lake Titicaca dropped gradually to 60 m lower than that today, indicating a large deficit in rainfall over the Altiplano. In contrast, during the same interval, Lake Valencia changed from an ephemeral lake to an overflowing lake. The present-day mean position of the ITCZ during the austral summer has a southern position, but between 12 400 and 8800 cal years BP the conclusive evidence from some ten sites and from insolation data shows that the mean position of the ITCZ was ~10° north of its modern position. This mechanism explains the antiphase rainfall relationship between Bolivian and Venezuelan latitudes, reflecting a migration of the ITCZ at the precession rhythm of orbital climate forcing. Precessional forcing also explains the presence of savanna with ca. 20-ka intervals in the pollen record of

Carajas (viz. around 65, 45, 25, and 6 ka BP) with intervals of wet forest in between (Absy et al., 1991).

See also: Pollen Records, Last Interglacial of Europe. **Lake Level Studies:** Latin America. **Pollen Methods and Studies:** Databases and Their Application. **Pollen Records, Late Pleistocene:** Africa; Australasia; Middle and Late Pleistocene in Southern Europe; Northern Asia; Northern North America; Western North America. **Vertebrate Records:** Late Pleistocene of South America.

References

- Absy ML, Cleef A, Fournier M, et al. (1991) Mise en évidence de quatre phases d'ouverture de la forêt dense dans le sud-est de l'Amazonie au cours des 60,000 dernières années. Première comparaison avec d'autres régions tropicales. *Comptes Rendus Académie des Sciences, Paris* 313: 673–678.
- Andriessen PMA, Helmens K, Hooghiemstra H, Riezebos PA, and Van der Hammen T (1993) Absolute chronology of the Pliocene–Quaternary sediment sequence of the Bogotá area, Colombia. *Quaternary Science Reviews* 12: 483–501.
- Behling H and Hooghiemstra H (1998) Late Quaternary paleoecology and paleoclimatology from pollen records of the savannas of Llanos Orientales in Colombia. *Palaeogeography, Palaeoclimatology, Palaeoecology* 139: 251–267.
- Behling H and Hooghiemstra H (2001) Neotropical savanna environments in space and time: Late Quaternary inter-hemispheric comparisons. In: Margraf V (ed.) *Inter-Hemispheric Climatic Linkages*, pp. 307–323. San Diego: Academic Press.
- Behling H and Lichte M (1997) Evidence of dry and cold climatic conditions at glacial times in tropical Southeastern Brazil. *Quaternary Research* 48: 348–358.
- Berrio JC, Arbeláez MV, Duivenvoorden JF, Cleef AM, and Hooghiemstra H (2003) Pollen representation and successional vegetation change on the sandstone plateau

- of Araracuara, Colombian Amazonia. *Review of Palaeobotany and Palynology* 126: 163–181.
- Berrio JC, Hooghiemstra H, Behling H, Botero P, and Van der Borg K (2002) Late-Quaternary savanna history of the Colombian Llanos Orientales from Lagunas Chenevo and Mozambique: A transect synthesis. *The Holocene* 12: 35–48.
- Berrio JC, Hooghiemstra H, Behling H, and Van der Borg K (2000) Late Holocene history of savanna gallery forest from Carimagua area, Colombia. *Review of Palaeobotany and Palynology* 111: 295–308.
- Boom A, Marchant R, Hooghiemstra H, and Sinnenghe Damsté JS (2002) CO₂ and temperature controlled altitudinal shifts of C₄- and C₃-dominated grasslands allow reconstruction of paleo-atmospheric pCO₂. *Palaeogeography, Palaeoclimatology, Palaeoecology* 177: 151–168.
- Boom A, Mora G, Cleef AM, and Hooghiemstra H (2001) High altitude C₄ grasslands in the northern Andes: Relicts from glacial conditions? *Review of Palaeobotany and Palynology* 115: 147–160.
- Brenner M, Hodell DA, Curtis JH, Rosenmeier ME, Binford MW, and Abbott MB (2001) Abrupt climate change and pre-Columbian cultural collapse. In: Markgraf V (ed.) *Interhemispheric Climate Linkages*, pp. 87–103. San Diego: Academic Press.
- Bush MB, Colinvaux PA, Wiemann MC, Piperno DR, and Liu K-b (1990) Late Pleistocene temperature depression and vegetation change in Ecuadorian Amazonia. *Quaternary Research* 34: 330–345.
- Colinvaux PA, De Oliveira PE, and Bush MB (2000) Amazonian and neotropical plant communities on glacial time-scales: The failure of the aridity and refuge hypotheses. *Quaternary Science Reviews* 19: 141–169.
- Colinvaux PA, De Oliveira PE, Moreno JE, Miller MC, and Bush MB (1996) A long pollen record from lowland Amazonia: Forest and cooling in glacial times. *Science* 247: 85–88.
- Farrera I, Harrison SP, Prentice IC, et al. (1999) Tropical climates at the Last Glacial Maximum: A new synthesis of terrestrial palaeoclimate data. I. Vegetation, lake-levels and geochemistry. *Climate Dynamics* 15: 823–856.
- Flenley J (1979) *The Equatorial Rain Forest: A Geological History*, p. 162. London: Butterworths.
- Haffer J (1969) Speciation in Amazonian forest birds. *Science* 165: 131–137.
- Haffer J (1982) General aspects of the refuge theory. In: Prance GT (ed.) *Biological Diversification in the Tropics*, pp. 6–24. New York: Columbia University Press.
- Haug G, Hughen KA, Sigman DM, Peterson LC, and Röhl U (2001) Southward migration of the Intertropical Convergence Zone through the Holocene. *Science* 293: 1304–1308.
- Heusser CJ (2003) Ice age southern Andes: A chronicle of paleoecological events. *Developments in Quaternary Science*, vol. 3, p. 240. Amsterdam: Elsevier.
- Hooghiemstra H (2006) Immigration of oak into northern South America: A paleo-ecological document. In: Kappelle M (ed.) *Ecology and Conservation of Neotropical Montane Oak Forests*, vol. 185, pp. 18–28. Berlin: Springer. Ecological Studies.
- Hooghiemstra H, Wijninga VM, and Cleef AM (2006) The paleobotanical record of Colombia: Implications for biogeography and biodiversity. *Annals of the Missouri Botanical Garden* 93: 297–325.
- Huber O (1987) Neotropical savannas: Their flora and vegetation. *Trends in Ecology and Evolution* 2: 67–71.
- Huntley BJ and Walker BH (eds.) (1982) *Ecology of Tropical Savannas*, vol. 42. Berlin: Springer. Ecological Studies.
- Ledru MP, Bertaux J, Sifeddine A, and Suguio K (1998) Absence of last glacial maximum records on lowland tropical forests. *Quaternary Research* 49: 233–237.
- Marchant R, Behling H, Berrio JC, et al. (2001) Mid- to Late-Holocene pollen-based biome reconstructions for Colombia. *Quaternary Science Reviews* 20: 1289–1308.
- Marchant R and Hooghiemstra H (2004) Rapid environmental change in African and South American tropics around 4000 years before present: A review. *Earth-Science Reviews* 66: 217–260.
- Markgraf V (1993) Climatic history of Central and South America since 18,000 yr BP: Comparison of pollen records and model simulations. In: Wright HE Jr., Kutzbach JE, Webb T III, Ruddiman WF, Street-Perrott FA, and Bartlein PJ (eds.) *Global Climates Since the Last Glacial Maximum*. Minneapolis: University of Minnesota Press.
- Markgraf V (ed.) (2001) *Interhemispheric Climate Linkages*, p. 454. San Diego: Academic Press.
- Martin L, Bertaux J, Corrège T, et al. (1997) Astronomical forcing of contrasting rainfall changes in tropical South America between 12,400 and 8800 cal yr BP. *Quaternary Research* 27: 117–122.
- Mayle FE, Burbridge R, and Killeen TJ (2000) Millennial-scale dynamics of southern Amazonian rainforest. *Science* 290: 2291–2294.
- Mistry J (2000) *World Savannas. Ecology and Human Use*, p. 344. Harlow, UK: Prentice Hall, Pearson Education.
- Mommersteeg H and Hooghiemstra H (2007) Millennium-scale climate variability in the tropics from 88,000-yr pollen record Fuquene-7C (Colombia). *Journal of Quaternary Science*.
- Morley RJ (2000) *Origin and Evolution of Tropical Rain Forests*, p. 362. Chichester, UK: John Wiley & Sons.
- Prance GT (ed.) (1982) *Biological Diversification in the Tropics*, p. 714. New York: Columbia University Press.
- Prance GT and Lovejoy TE (1985) *Key Environments: Amazonia*, p. 442. Oxford: Pergamon Press.
- Prentice IC, Cramer W, Harrison SP, Leemans R, Monserud RA, and Solomon AM (1992) A global biome model based on plant physiology and dominance, soil properties and climate. *Journal of Biogeography* 19: 117–134.
- Sarmiento G and Monasterio M (1975) A critical consideration of the environmental conditions associated with the occurrence of savanna ecosystems in tropical America. In: Golley FB and Medina E (eds.) *Tropical Ecosystems: Trends in Terrestrial and Aquatic Research*, pp. 223–250. Berlin: Springer.
- Thompson RS, Anderson KH, and Bartlein PJ (2000) Atlas of relations between climatic parameters and distributions of important trees and shrubs in North America. Denver, CO: US Geological Survey. Professional Paper 1650, vol. 1 (p. 269), vol. 2 (p. 423), vol. 3 (p. 386).
- Torres V, Vandenberghe J, and Hooghiemstra H (2005) An environmental reconstruction of the sediment infill of the Bogotá basin (Colombia) during the last 3 million years from abiotic and biotic proxies. *Palaeogeography, Palaeoclimatology, Palaeoecology* 226: 127–148.
- Van der Hammen T (1974) The Pleistocene changes of vegetation and climate in tropical South America. *Journal of Biogeography* 1: 3–26.
- Van der Hammen T and Cleef AM (1986) Development of high Andean páramo flora and vegetation. In: Vuilleumier F and Monasterio M (eds.) *High Altitude Tropical Biogeography*, pp. 153–201. New York: Oxford University Press.
- Van der Hammen T and Hooghiemstra H (2000) Neogene and Quaternary history of vegetation, climate and plant diversity in Amazonia. *Quaternary Science Reviews* 19: 725–742.
- Van Geel B and Van der Hammen T (1973) Upper Quaternary vegetational and climatic sequence of the Fúquene area (Eastern Cordillera, Colombia). *Palaeogeography, Palaeoclimatology, Palaeoecology* 14: 9–92.
- Van't Veer R and Hooghiemstra H (2000) Montane forest evolution during the last 650,000 years in Colombia: A multivariate approach based on pollen record Funza-I. *Journal of Quaternary Science* 15: 329–346.
- Vuilleumier F and Monasterio M (eds.) (1986) *High Altitude Tropical Biogeography*. New York: Oxford University Press.
- Whitmore TC and Prance GT (eds.) (1987) *Biogeography and Quaternary History in Tropical America*, p. 214. Oxford: Clarendon Press.
- Wille M, Hooghiemstra H, Behling H, Van der Borg K, and Negret AJ (2001) Environmental change in the Colombian subandean forest belt from 8 pollen records: The last 50 kyr. *Vegetation History and Archaeobotany* 10: 61–77.
- Wille M, Hooghiemstra H, Van Geel B, Behling H, De Jong A, and Van der Borg K (2003) Submillennium-scale migrations of rainforest-savanna boundary in Colombia: ¹⁴C wiggle-matching and pollen analysis of core Las Margaritas. *Palaeogeography, Palaeoclimatology, Palaeoecology* 193: 201–223.

Middle and Late Pleistocene in Southern Europe

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Introduction

Pollen analysis is one of the most efficient tools for the reconstruction of past environmental changes. Southern Europe has yielded numerous key pollen sequences, some continuous over long time intervals. After discussing the history of paleoecological research and identifying major reference sections, we use these records to trace spatial and temporal vegetation variability through southern Europe during glacial–interglacial cycles, interpreting the records in terms of climate forcing, and discussing the major trends in Mediterranean biodiversity during the Pleistocene, with special attention to late Pleistocene.

History of Research

For northern and central Europe, the first pollen records that provided information on Pleistocene vegetation history were recovered from The Netherlands, Britain, Denmark, and Germany. These records provided a framework for understanding the vegetation history of Pleistocene interglacials. However, in northern Europe, repeated advances and retreats of ice sheets caused repeated sediment accumulation and erosion. As a consequence, the records are usually discontinuous in space and time and are often difficult to place in a stratigraphic context. In contrast, southern Europe contains some sites, such as subsiding basins or crater lakes (maars) in tectonically active regions that have the potential for continuous sediment accumulation over long intervals. The investigation of such sites (Figure 1) during recent decades provided information about vegetation and climate changes during several glacial–interglacial cycles and has opened the door to detailed correlations with other types of long records (loess sequences, deep-sea sediment cores, and ice cores).

Greece

The first long pollen sequence (and still the longest) described in southern Europe comes from a subsiding basin in Macedonia, northeast Greece: Tenaghi Philippon (Wijmstra, 1969). A 120-m core yielded sediments spanning more than 1 Ma. Wijmstra and his collaborators have devoted more than 25 years to the study of this record, which was one of the first in Europe showing many climatic cycles during the Pleistocene (van der Hammen et al., 1971; van der Wiel and Wijmstra, 1987a, 1987b; Wijmstra and Smit, 1976) in agreement with the Milankovitch (1930) theory and contrary to the ‘classical’ four-part glacial–interglacial scheme proposed by Penck and Brückner (1909). Tzedakis et al. (2003c) and Tzedakis et al. (2006) proposed a new chronology for this sequence, based on an initial alignment

with the marine $\delta^{18}\text{O}$ record, which was further refined by a direct pollen–orbital tuning procedure; according to this, the base of the sequence extends back to 1.35 Ma.

By correlating with the northern European pollen record, Wijmstra and colleagues were the first to demonstrate a latitudinal gradient in upper Pleistocene European vegetation dynamics. Oak forest played a major role during most of the interglacial, without the well-marked succession of forest trees characteristic of the mid-European interglacial sequences. Oak forests also succeeded in expanding during the post-Eemian (marine isotopic stage (MIS) 5e) interstadials when boreal forests were growing in northern Europe; the Macedonian interstadials Drama and Elevteroupolis were convincingly correlated with the Dutch interstadials Brörup and Odderade (MIS 5c and 5a, respectively). A new core recovered in 2005 will be subjected to a higher resolution multidisciplinary study (Pross et al., 2007).

The discovery of the Tenaghi Philippon record led other researchers to explore records from other subsiding basins, such as Ioannina (Bottema, 1974; Tzedakis, 1993, 1994), Xinias (Bottema, 1979), and Kopais (Okuda et al., 2001; Tzedakis, 1999). A new core from Ioannina, in northwest Greece, is being analyzed at much higher resolution (ca. 200 years). Results so far cover the last 130 ka (Tzedakis et al., 2002, 2003b), the penultimate glacial (Roucoux et al., 2011) and the penultimate interglacial complex (Roucoux et al., 2008). The forest phases are also characterized by early expansion of deciduous and evergreen oak (*Quercus* spp.) and Mediterranean sclerophylls but are followed by a more diverse vegetation succession compared to that inferred from the Tenaghi Philippon record. This diverse vegetation includes elm (*Ulmus*), hazel (*Corylus*), hornbeam (*Carpinus betulus*), fir (*Abies*), and beech (*Fagus*). The penultimate interglacial complex, corresponding to MIS 7, includes three major temperate intervals (Zitza I, Zitza II, and Zitza III, respectively, corresponding to substages 7e, 7c, and 7a). Of these, the longest and the warmest interval is Zitza II (Roucoux et al., 2008). At the end of Zitza I, an abrupt shift to colder and drier climatic conditions is observed, in agreement with the French Lac du Bouchet record (Reille et al., 2000).

In the northwest Balkan Peninsula (former Yugoslavia), there are regions, such as the Ljubljana plain, with a geological setting suitable for Pleistocene long sequences (Sercelj, 1966). However, the recent political situation has not allowed these areas to be investigated for their potential Pleistocene records. In Albania, after research initiated on Lake Malik (Denèfle et al., 2000), the multidisciplinary study of Lake Ohrid (704 m a.s.l.) (Lezine et al., 2010) provides a record of the last glacial–interglacial cycle in a mountain zone and complements the Greek sites.

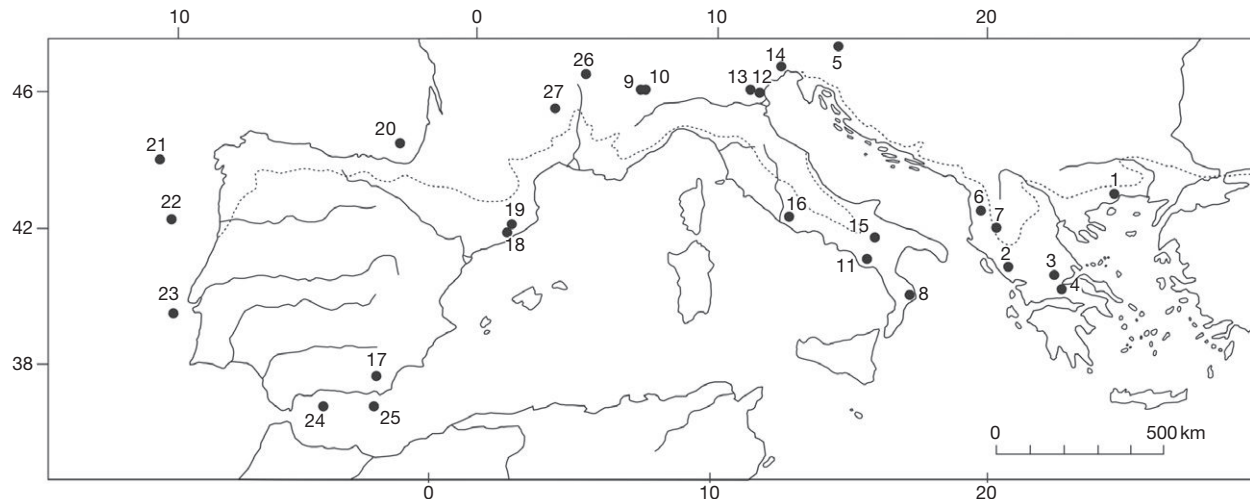


Figure 1 Location map of the cited major Pleistocene sites. (1) T. Philippon, (2) Ioannina, (3) Xinias, (4) Kopais, (5) Ljubljana plain, (6) Lake Malik, (7) Lake Orhid, (8) Vrica, (9) Leffe, (10) Pianico-Sèllere, (11) Vallo di Diano, (12) Venice Lagoon, (13) Fimon, (14) Azzaro D., (15) Monticchio, (16) Valle di Castiglione, (17) Padul, (18) Banõlas, (19) Olot, (20) MD04-2845, (21) MD01-2447, (22) MD95-2039, (23) MD95-2042, (24) ODP 976, (25) MD95-2043, (26) Les Echets, and (27) Velay.

Italy

Due to active neotectonic activity, several uplifted Italian marine sequences were used to define the Plio/Pleistocene boundary. The official stratotype for this boundary has been defined at Monte San Nicola, Sicily (Gibbard and Head, 2010), replacing the previous type site at Vrica (Aguirre and Pasani, 1985). A review of the other sections is available in Bertini (2003). Further north in Europe, the pre-Alpine site of Leffe (Figure 1), north of Bergamo, was first studied by Lona (1950) and Lona and Follieri (1957) and later reinvestigated by Ravazzi and Rossignol-Strick (1995) and Muttoni et al. (2007). This site provides an example of early Pleistocene climatic cycles spanning the interval 0.87–1.9 Ma in a region wetter than Tenaghi Philippon. These pollen spectra were dominated by tree taxa that today are extinct in Europe. These ‘Tertiary relicts’ include cedar (*Cedrus*), hickory (*Carya*), wingnut (*Pterocarya*), and *Eucommia*. The glacial intervals were short and not sharply defined, and the interglacials were rather wet. Close to Leffe, a thick annually laminated lacustrine sequence outcrops at Pianico-Sèllere. This interglacial complex, initially correlated with the last interglacial (MoscarIELLO et al., 2000), has been dated by the occurrence of the Brunhes/Matuyama boundary (marking a paleomagnetic reversal) at ca. 800 ka (Pinti et al., 2001). A controversy arose when Brauer et al. (2007a) attributed an age of ca. 400 ka to a tephra (volcanic ash) layer included in the section. Nevertheless, in terms of both palynostratigraphy and magnetostratigraphy (Roulleau et al., 2009; Scardia and Muttoni, 2009), this sequence cannot be contemporaneous with MIS 11, and the correlation with MIS 19 is more likely. Even by that time, the pollen data show that almost all the exotic (or Tertiary relic) trees had already disappeared. The timing of the vegetation succession was established by varve counting and high-resolution pollen analyses. The vegetation sequence is comparable with that of the last interglacial. The floristic differences between Leffe and Pianico are consistent with the vegetational changes characterizing the transition from the early to the middle Pleistocene.

A long core in the Venice lagoon (Massari et al., 2004) yielded a full Pleistocene record, but it provided results that are difficult to interpret due to phases of fluvial erosion by the Po River. Not far to the east along the Adriatic coast, a 95-m long core at Azzaro Decimo (Pini et al., 2009) provided a pollen record for the last 215 ka, unfortunately disturbed by several layers of sterile gravels resulting from fluvial activity. Lake Fimon (Pini et al., 2010), in the Berici Hills, west of Padova, provides also a detailed record of the last climatic cycle. The interval contemporaneous with MIS 5 is registered in 5 m of sediments, allowing the acquisition of a high-resolution pollen diagram. Vegetation dynamics during the last interglacial (corresponding to MIS 5e and the lower part of 5d Shackleton et al., 2003) is marked by the late arrival and relatively small contribution of European hornbeam (*C. betulus*) pollen and the relative abundance of beech (*Fagus*) during the later phase.

In southern Italy, south of Naples, the middle Pleistocene sequence of Vallo di Diano (Russo Ermolli and Cheddadi, 1997) includes two interglacials correlated with MIS 15 and 13.

During the 1980s, the intensive exploration of crater lakes and paleolakes from central and southern Italy yielded several long pollen records. Intensive studies at Lago Grande di Monticchio, well dated by tephra layers and by annual laminations, gave a detailed description of the vegetation and climate changes during the last glacial–interglacial cycle and suggested a duration of ca. 17 700 years for MIS 5e (the last interglacial interval) (Allen et al., 1999, 2009; Brauer et al., 2007b), in close agreement with the estimates from the Ioannina sequence based on tuning of pollen sequences to orbital forcing curves (pollen-orbital tuning) (Tzedakis et al., 2003b). Most of these sediment sequences do not extend beyond the last interglacial (see Pollen Records, Last Interglacial of Europe), except at Valle di Castiglione, where Follieri et al. (1988) obtained a pollen sequence for the last 270 ka. At this site, researchers first identified a phase of major expansion of Mediterranean

tree taxa such as olive (*Olea*), pistachio (*Pistacia*), and evergreen oak (*Quercus ilex*) at the beginning of the last interglacial. Here, pollen from mesophilous trees, such as hazel (*Corylus*), European hornbeam, beech, and fir (*Abies*), a major component of the pollen spectra. This is also the case for the two temperate episodes which follow MIS 5c and 5a and the three major temperate phases of the penultimate interglacial complex (Roma I, II, and III at Valle di Castiglione, which are equivalent to MIS 7e, 7c, and 7a). *Zelkova*, which had disappeared from continental western Europe during the middle Pleistocene, was still present on the Italian peninsula until the last Pleniglacial (MIS 2).

Spain

Unlike Greece, Spain, at the same latitude, is poor in Pleistocene records. The promising site of Padul ([Figure 1](#)) lies in a large tectonic depression in the foothills of Sierra Nevada, which contains a thick succession of peat and silt layers. The three pollen diagrams published by [Florschütz et al. \(1971\)](#) and [Pons and Reille \(1988\)](#) are difficult to interpret, partly due to poor pollen preservation in most of the peat layers ([Pons and Reille, 1988](#)), probably due to changes in the water table and desiccation phases. The pollen zone IVb (70–24 m) ([Florschütz et al., 1971](#)) shows two major temperate episodes (j and m, considered to predate the last interglacial) with abundant oaks (undifferentiated) alternating with dry phases characterized by dominant steppe with stands of pine, but the record's resolution is too low to allow a consistent correlation with the other Mediterranean long sequences. The interval between 24 m and 12 m in core IV, with an abundant warm flora alternating with dry episodes, is convincingly correlated by [Pons and Reille \(1988\)](#) with the interval 14.8 m–6 m in their core Padul 2, but they diverge in their interpretation which correlates this section with the early Würm (MIS 5c to 5a) contrary to [Florschütz et al. \(1971\)](#) who attribute the early temperate episode to the last interglacial (MIS 5a). Further studies and absolute dates are needed to refine the chronology at this site.

The few other Spanish sites providing information on the Pleistocene vegetation are in Catalonia. These include Banòles ([Pérez-Obiol and Julià, 1994](#)), with a detailed record of the last glacial, and Olot ([Burjachs, 1994](#)), with a partial record of the last interglacial (characterized by phases with abundant European hornbeam and fir).

Fortunately, pollen analyses of marine cores close to the Iberian coast provide an integrated perspective of long-term vegetation changes on the adjacent continent and, moreover, provide direct land–sea correlations. The first direct palynological correlation of MIS 5e with the Eemian interglacial was identified in a core from the Iberian margin ([Turon, 1984](#)), and provided a starting point for more intensive research. After a thorough study of the interglacial complex offshore the Tajo river catchments ([Sanchez-Goni et al., 1999](#)), a description of the last four glacial–interglacial cycles is now available from cores MD95-2039, MD01-2443, MD01-2444, MD01-2447, and MD04-2845 ([Desprat, 2005](#); [Desprat et al., 2006](#); [Margari et al., 2010](#); [Roucoux et al., 2011](#); [Sanchez-Goni et al., 2008](#); [Tzedakis et al., 2004b](#)). A characteristic of the cold phases on the Iberian coast is the abundance of Ericaceae. During MIS 5,

oaks are the dominant forest elements during the temperate intervals, and a marked occurrence of evergreen trees is inferred during an early phase of MIS 5e. The last glacial cycle is also registered in marine cores from the Alboran sea (ODP 076, MD95-2043) ([Combourieu-Nebout et al., 1993](#); [Combourieu-Nebout et al., 2002](#); [Sanchez-Goni et al., 2002](#)).

France

In contrast to the peninsulas of Spain and Italy, which are isolated from middle Europe by mountain ranges, the French Mediterranean lowlands are connected with continental Europe by the Rhône/Saone corridor. The Pleistocene vegetation was poorly documented in France till the 1970s. The discovery of la Grande Pile by Seret and [Woillard \(Woillard, 1978\)](#), the first west European site with a continuous record of the last glacial–interglacial cycle, formed an impetus for new investigations, leading to the discovery of its more southern counterpart of Les Echets ([de Beaulieu and Reille, 1984](#)). The two sites show forest dynamics during the last interglacial very similar to those of the classical Eemian (the same succession of forest types and same absence of beech). The early Würm interstadials (St. Germain I and II corresponding to MIS 5c and 5a) show complex forest successions including episodes of dominant mesophilous trees at a time when boreal forests expanded in the Netherlands and northern Germany (the Brörup and Odderade interstadials). This provides evidence of a steeper climatic gradient in Europe during this interval, compared with the Eemian and the Holocene. However, these results were so unexpected in 1975 that it was suggested that the St. Germain I and II layers at Grande Pile represented a massive redeposition of the last interglacial sediments. At Les Echets ([Figure 1](#)), two new cores recovered in 2001 provided additional information for the upper part of the sequence (last glacial) showing the local consequences of Heinrich events and Dansgaard–Oeschger events ([Verès et al., 2007](#); [Verès et al., 2009](#); [Wohlfarth et al., 2008](#)) and of the transition from the penultimate glacial to the last interglacial ([Gandouin et al., 2007](#)).

However, another major Pleistocene key sequence soon confirmed the records from Grande Pile and Les Echets. This came from the crater lakes and paleolakes of the Velay Plateau (Massif Central), a highland region close to the Mediterranean Cevennes, built up by volcanic activity from the Oligocene to the Late Glacial. The exploration of three maars, Lac du Bouchet, Ribains, and Praclaux ([de Beaulieu and Reille, 1992](#); [Reille and de Beaulieu, 1995](#); [Reille et al., 1998](#)), allowed the development of a composite pollen sequence which covers the four last glacial–interglacial cycles (i.e., the last ca. 450 ka) ([Reille et al., 2000](#)). Due to its location close to the 45th parallel, this record may be important as a link between the classical Mediterranean long sequences and the north European record. Below the last interglacial (see [Pollen Records, Last Interglacial of Europe](#)), four cold intervals (correlated with MIS 6, 8, 10, and 12) alternate with three temperate ‘complexes’ correlated with MIS 7, 9, and 11. As during MIS 5, each of the latter starts abruptly with a strong temperate interglacial (Bouchet 1 = MIS 7e, Landos = MIS 9e, Praclaux = beginning of MIS 11) marked by the classical succession of European forest trees. This vegetation sequence starts with pioneer trees, then warm mixed oak forests

expand and are replaced by less thermophilous trees (yew, hornbeam, and firs) and finally, boreal spruce and pine forests mark the end of the interglacial. The three main interglacials are followed by a succession of two or three short stadials (similar to Melisey 1 and 2, and contemporaneous with MIS 5d and 5b) alternating with two or three temperate or mild intervals as shown by the expansion of the less temperate deciduous trees and/or of boreal conifers. These intervals are reminiscent of the Saint Germain interstadials in MIS 5. In most cases, the initial interglacial is marked by an important development of warmth-demanding trees, which are increasingly less abundant during the subsequent interstadials. This gives a decreasing sawtooth structure to regional pollen curves which shows a striking similarity with the climate record inferred from Antarctic ice cores (Petit et al., 1999), illustrating a global interhemispheric signature (Figure 2).

Trends in Vegetation Dynamics

Mediterranean Climate Setting and Beginning of Ice Age

The gradual cooling which has occurred since the end of the Eocene is well documented in the pollen records from southern Europe. It began around 3.6 Ma with the initiation of climatic phases marked by summer aridity. This environmental stress promoted different adaptations among regional plant species, including a differentiation among the already existing sclerophyllous taxa, which evolved into the present-day Mediterranean flora (Comboureu-Nebout, 1993; de Beaulieu et al., 2005; Suc et al., 1995). However, Tzedakis (2007)

suggested that the Mediterranean-type climate is not exclusively a post-3.6 Ma phenomenon but may have appeared intermittently during the course of the Tertiary.

The early Pleistocene represents a transition between climate governed by subtropical environments toward regimes marked by an intensification of glacial cycles between 1 Ma and 800 ka. During this interval, several taxa that had persisted from the Tertiary flora were still abundant, although their role progressively decreased. The Leffe sequence (Ravazzi and Rossignol-Strick, 1995) presents an example of early Pleistocene 40-ka glacial cycles. The Jaramillo paleomagnetic event dates the upper part of this sequence to ca. 1 Ma. The pollen sequence shows seven glacial-interglacial cycles characterized by the same recurrent successions of vegetation types, with forests almost always dominant. Each cycle (Figure 3(a)) starts with a short phase of moderately open vegetation, with a maximum abundance of xerophilous herbs (sage (*Artemisia*) and other steppe elements), which is associated with a cold and dry episode associated with a glacial interval. Then deciduous trees expand, first oak, hazel, and hornbeam, followed by a maximum of Juglandaceae (hickory and wingnut), associated with hemlock (*Tsuga*), which corresponds to a warm and wet climatic optimum, finally the expansion of mountain conifers (pine, spruce, and cedar) correspond to a climate cooling. This is then followed by second broadly similar succession, with *Quercus*, *Ulmus*, *Carpinus*, and *Carya*, terminating in a prolonged coniferous phase, before the onset of a relatively short steppe vegetation phase. This suggests the persistent influence of precession in southern Europe even during the interval of obliquity-dominated glacial cycles (Tzedakis,

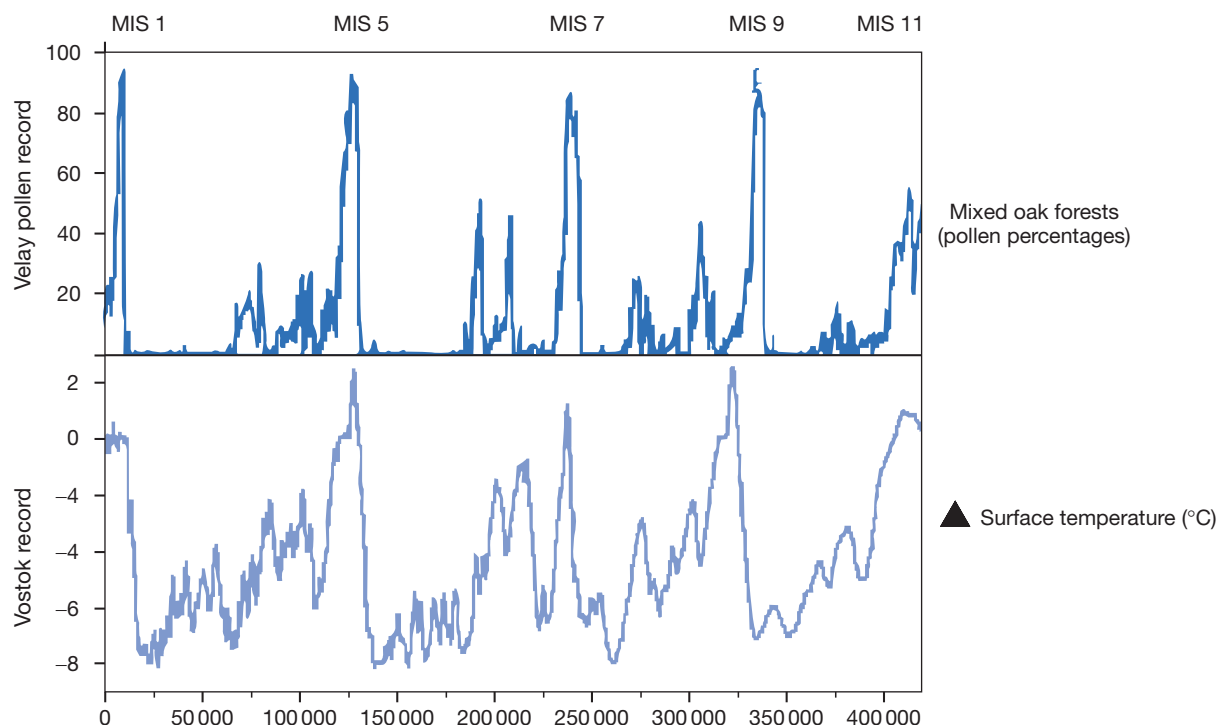


Figure 2 Comparison between the air surface temperature curve during the last 450 000 years at Vostok, Antarctica (based on the isotopic composition of the ice: from Petit et al., 1999) and the pollen curves of the sum of the most thermophilous trees in the Velay sequence (redrafted from Tzedakis et al., 1997). They show impressively similar saw-tooth structures. The slight differences in timing are, in great part, due to different age models.

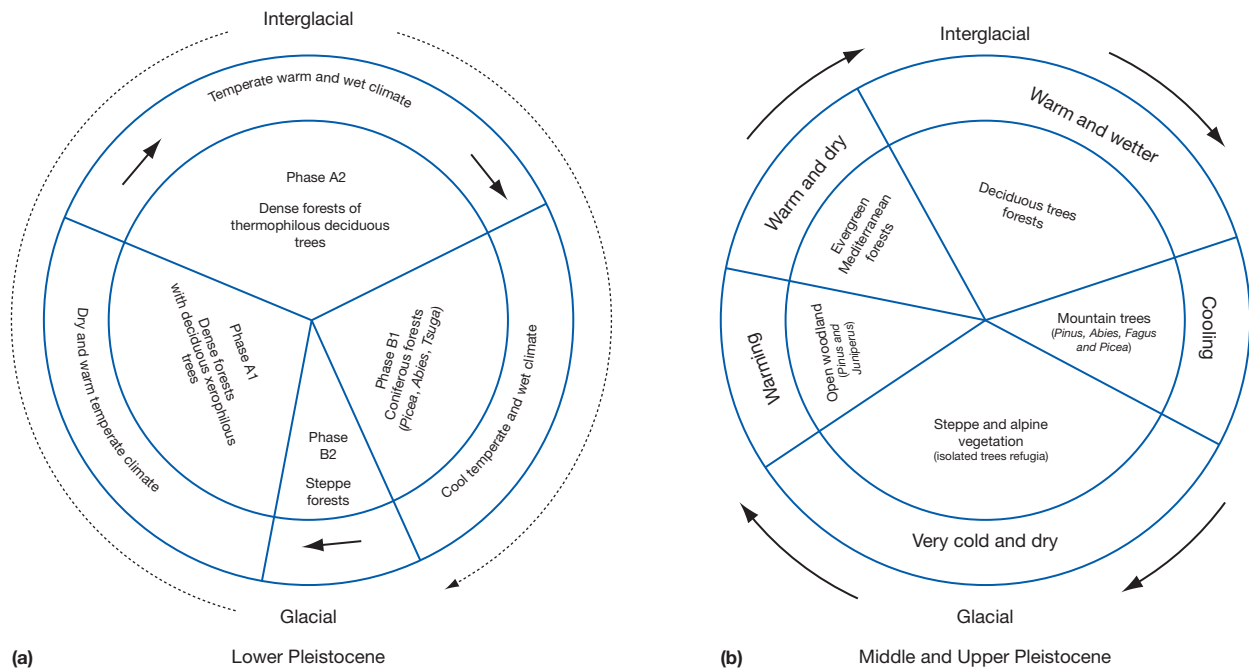


Figure 3 Differences between vegetation cycles during the lower Pleistocene (a) and during middle/upper Pleistocene (b). From de Beaulieu J-L, Andrieu-Ponel V, Cheddadi R, et al. (2006) Apport des longues séquences lacustres à la connaissance des climats et des paysages pléistocènes. *C.R. Palévol* 5: 65–72.

2005). The pollen flora is rich in 'Tertiary' taxa such as Taxodiaceae, hemlock (*Tsuga*), cedar, and the deciduous hickory, wingnut, *Eucommia*, Persian ironwood (*Parrotia*), horse chestnut (*Aesculus*), and sweetgum (*Liquidambar*).

The lower part of Tenaghi Philippon record from ca. 1.35 Ma is contemporaneous with Leffe (Tzedakis et al., 2006). The main difference between the two sequences is the major role played by oak and Mediterranean pines at Tenaghi Philippon due to a drier climate.

Pleistocene Extinctions and the South European Refuges

The causes of the change from 40-ka cycles to 100-ka cycles (governed by orbital eccentricity), which became dominant during the first part of middle Pleistocene, are still unclear. The $\delta^{18}\text{O}$ curves from benthic Foraminifera in deep-sea cores clearly show that the MIS 22, at ca. 900 ka, is the first glacial episode marked by major ice accumulation and very low temperatures on the continents, as became the rule in the middle and late Pleistocene. As shown by the Italian sequences and, to a lesser degree, by Tenaghi Philippon, this glaciation was certainly an interval of major extinction for the most of the Tertiary temperate flora of Europe. After this episode, the Juglandaceae never played a major role in the vegetation, and this group declined in importance in the European vegetation as the more medio-European taxa (European hornbeam, beech, and spruce) became dominant.

The lower part of the middle Pleistocene (between the Brunhes/Matuyama boundary and ca. 450 ka) is poorly documented except at Tenaghi Philippon. This remains a unique European key site that provides a nearly continuous record of the past 1.35 Ma. At this site, Tzedakis et al. (2006) demonstrated that the cold period corresponding to MIS 16 represents

a second major period of extinction in Greece. Cedar, hickory, and *Eucommia* disappeared or became restricted to very small refuges and never again played a significant role in forest ecosystems. But at Pianico-Sèllere, in Northern Italy, these tree taxa were already absent by 800 ka, pointing to regional lags in the timing of local extinctions.

The history of the extirpation of the few remaining 'Tertiary relicts,' such as wingnut and *Zelkova*, from the European flora is well known. *Zelkova* was able to persist in central Italy until MIS 2 and two small, isolated populations still survive in Sicily and Crete (Quézel and Médail, 2003). This case illustrates the role of Mediterranean peninsulas in the conservation of temperate trees during glacial episodes. Recent studies linking paleoecology and phylogeography have shown that the location of refuges during the last glaciation constrained the distribution of forest ecosystems during the Holocene (Hewitt, 2000; Petit et al., 2002, 2003). This evidence must likewise be valid for the previous climatic cycles and may explain part of the variability in forest dynamics between the different interglacials.

Climatic Cycles and Orbital Forcing

On the basis of the Tenaghi Philippon record, Tzedakis et al. (2006) consider that the forest dynamics during all the interglacials following MIS 16 present common trends (of a 'modern' character) marked by episodes with abundant hornbeam in addition to oak (four hornbeam peaks between MIS 16 and MIS 12) and more frequent occurrences of fir. The two interglacials correlated with MIS 15 and 13 at Vallo di Diano are true to this model, also showing a great abundance of fir.

Following MIS 12, the spatial pattern of the glacial-interglacial cycles across southern Europe rests on the correlation between the five longest pollen sequences (Tenaghi Philippon, Velay, and valle di Castiglione and the deep-sea sediment cores off the Iberian Peninsula and from Alboran sea). In a first attempt at correlation, Tzedakis et al. (1997) developed a common chronology based on a tuning of the interglacial inceptions with the marine isotope timescale (Figure 4), which is mostly governed by orbital eccentricity and axial tilt. Later, a more detailed investigation led Magri and Tzedakis (2000) to identify direct relations between vegetation phases and precession cycles (Berger, 1977). An expansion of the Mediterranean taxa, pistachio (*Pistacia*), green olive (*Phillyrea*), and olive (*Olea*), generally at the beginning of most of the warm intervals, was correlated by Magri and Tzedakis (2000) with summer occurring at perihelion leading to arid summers. Later, the shift toward September perihelion was correlated with the maximum biomass of mesophilous forests. March perihelion is correlated with maximum extension of herbs. In light of this concept, Tzedakis et al. (2003a,b, 2006) reexamined the Greek sequences and adopted a timescale for Tenaghi Philippon that is slightly different from the one proposed in 1997. In any case, there is a striking correlation between the SPECMAP curves (Figure 4) and the succession of forested and nonforested episodes in the pollen sequences. This clearly demonstrates that each of the temperate intervals attributed to MIS 11, 9, and 7 is marked by a succession of climatic oscillations including cold stadials. Not less than 13 major warm phases are documented between MIS 12 and the

present (Figure 4). However, a gradient in vegetation sequences exists between France, Italy, and Greece, which illustrates the persistence of the present temperature and precipitation gradients between the three regions. In France and Italy, the vegetation history of temperate stages shares features in common with central European vegetation sequences (Iversen, 1958; Figure 3(b)). This includes a 'late-temperate' phase with abundant 'mountain' taxa (beech, fir, yew, spruce). In Greece, oak is always the dominant tree. There, even during glacial intervals, this tree remained present, sometimes abundant, showing that wood stands never totally disappeared. Tzedakis et al. (2002) showed a gradient between Ioannina and Tenaghi Philippon corresponding to different mesic situations.

Suborbital Forcing

During the 1990s, evidence accumulated which indicates short-term climatic changes during the last glacial interval, both from marine (Heinrich events, Bond events: Bond et al., 1997) and ice (Dansgaard-Oeschger events: Dansgaard et al., 1993) records. Cores from the Alboran sea show that these events are reflected in the western Mediterranean (Baudin et al., 2007; Combourieu-Nebout et al., 2002) and their consequences on the continent are now inferred from pollen records from France (Andrieu-Ponel et al., 2006; Wohlfarth et al., 2008), Italy (Allen et al., 2009), and Greece (Margari et al., 2009; Müller et al., 2011; Tzedakis et al., 2004a). On the continent, studies on lake-level changes also show that cyclic fluctuations during the

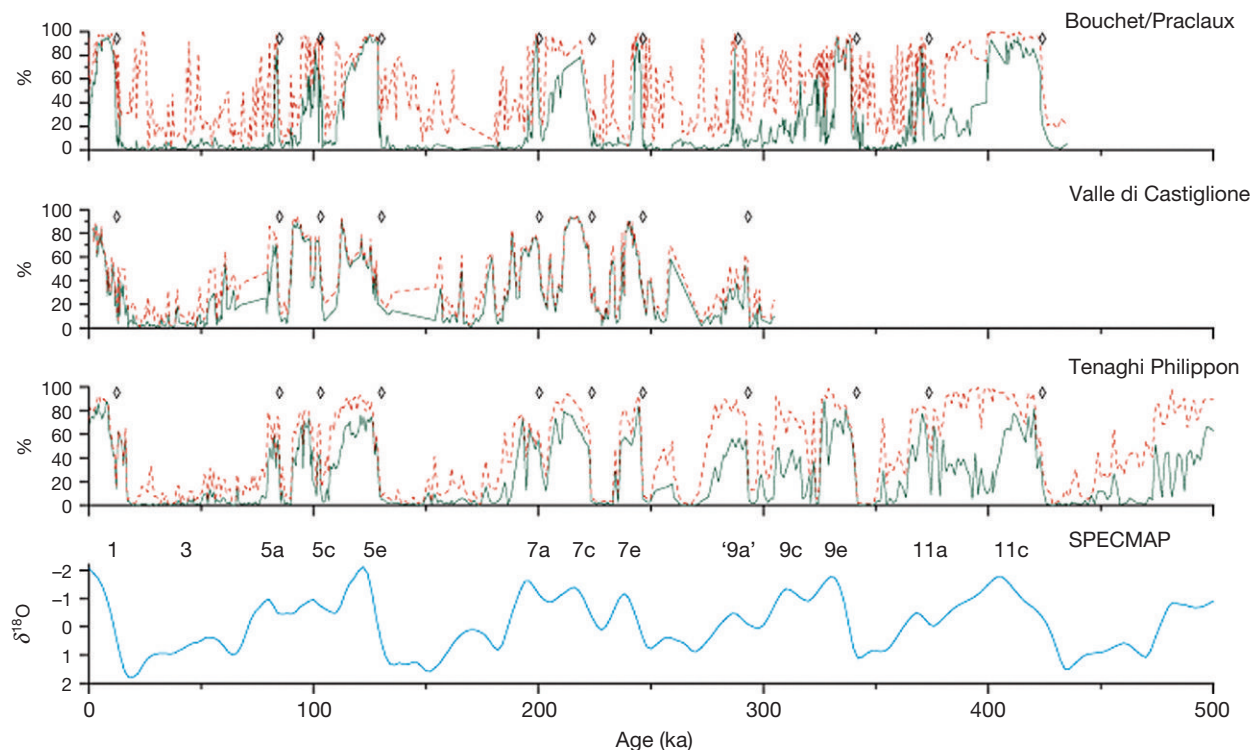


Figure 4 Comparison of south European terrestrial records and the marine oxygen isotope sequence. The solid lines represent the arboreal pollen (AP) curves minus *Pinus*; dotted lines the AP curves (including *Pinus*); and diamonds represent the control points used to align the chronologies. Modified from Tzedakis PC, Andrieu V, de Beaulieu J-L, et al. (1997) Comparison of terrestrial and marine records of changing climate of the last 500 000 years. *Earth and Planetary Science Letters* 150: 171–176.

Holocene correlated with changes in solar activity. Magny et al. (2007) have suggested similar (or independent) changes in the Holocene hydrological regime in the Mediterranean regions. However, comparable information for intervals prior to the last glacial remains lacking. An exception is an analysis of the changes in diatom populations during the last interglacial at Ribains, Velay, which suggests successive changes in seasonality during the last interglacial (Rioual et al., 2007). The challenge for the next decade may be to apply very high-resolution multi-disciplinary studies to the rare long Pleistocene series described here allowing better identification and understanding of sub-orbital climate variability in the Mediterranean regions.

Conclusion

The Mediterranean regions (mostly Greece and Italy) are unique in the world in terms of density of key Pleistocene pollen sequences. Moreover, the Velay sites provide a rare opportunity to connect middle Europe with southern Europe. The Mediterranean regions also constitute modern hot spots of biodiversity (Quézel and Médail, 2003). This conjunction provides a rare opportunity to put a long-term, paleoecological perspective on present-day ecology (Rull and Vegas-Vilarrubia, 2011), in order to address questions of biodiversity conservation (Samways et al., 2006; cf. Hampe and Pétit, 2005) in the context of global warming and increased human pressure.

See also: Pollen Records, Last Interglacial of Europe. **Pollen Records, Late Pleistocene:** Africa; Australasia; Northern Asia; Northern North America; South America; Western North America.

References

- Aguirre E and Pasani G (1985) The Pliocene–Pleistocene boundary. *Episodes* 8(2): 116–120.
- Allen JRM, Brandt U, Brauer A, et al. (1999) Rapid environmental changes in southern Europe during the last glacial period. *Nature* 400: 740–742.
- Allen JRM, Brandt U, and Huntley B (2009) Last interglacial palaeovegetation, palaeoenvironments and chronology: A new record from Lago Grande di Monticchio, Southern Italy. *Quaternary Science Reviews* 28(15–16): 1521–1538.
- Andrieu-Ponel V, Beaulieu JL (de), Cheddadi R, Guiter F, et al. (2006) *Analyse multidisciplinaire de la nouvelle séquence climatique (EC1) des Echets (alt., 267 m, Ain, France)*. Châtenay-Malabry: Andra, 173 pp.
- Baudin F, Combouirieu-Nebout N, and Zahn R (2007) Signatures of rapid climatic changes in organic matter records in the Western Mediterranean sea during the last glacial period. *Bulletin de la Société Géologique de France* 178(1): 3–13.
- Berger A (1977) Support for the astronomical theory of climate changes. *Nature* 268: 44–45.
- Bertini A (2003) Pliocene climatic cycles and altitudinal forest development from 2.7 Ma in the Northern Apennines (Italy): Evidences from the pollen record of the Stirone section (5.1 to 2.2 Ma). *Geobios* 34(3): 253–265.
- Bond G, Showers W, Cheseby M, et al. (1997) A pervasive millennial-scale cycle in North Atlantic Holocene and glacial climates. *Science* 278: 1257–1266.
- Bottema S (1974) *Late Quaternary Vegetation History of Northwestern Greece*. PhD Thesis, Rijksuniversiteit te Groningen, Groningen.
- Bottema S (1979) Pollen analytical investigations in Thessaly (Greece). *Palaeohistoria* 21: 19–40.
- Brauer A, Allen JRM, Mingram J, Dulski P, Wulf S, and Huntley B (2007a) Evidence for last interglacial chronology and environmental change from southern Europe. *Proceedings of the National Academy of Sciences of the United States of America* 104(2): 450–455.
- Brauer A, Wulf S, Mangili C, and Moscarillo A (2007b) Tephrochronological dating of varved interglacial lake deposits from Piànico-Sellere (Southern Alps, Italy) to about 400 ka. *Journal of Quaternary Science* 22(1): 85–96.
- Burjachs F (1994) The palynology of the upper Pleistocene and Holocene of the North-East Iberian Peninsula: Pla de l'Estany (Catalonia). *Historical Biology* 9: 17–33.
- Combouirieu-Nebout N, Turon JL, Zahn R, Capotondi L, Londeix L, and Pahnke K (2002) Enhanced aridity and atmospheric high-pressure stability over the Western Mediterranean during the North Atlantic cold events of the past 50 ky. *Geology* 30: 863–866.
- Combouirieu-Nebout N (1993) Vegetation response to Upper Pliocene glacial/interglacial cyclicity in the Central Mediterranean. *Quaternary Research* 40: 228–236.
- Dansgaard W, Johnsen SJ, Clausen HB, et al. (1993) Evidence for general instability of past climate from a 250-kyr ice-core record. *Nature* 364: 218–220.
- de Beaulieu J-L, Andrieu-Ponel V, Cheddadi R, et al. (2006) Apport des longues séquences lacustres à la connaissance des climats et des paysages pléistocènes. *C.R. Palévol* 5: 65–72.
- de Beaulieu J-L, Miras Y, Andrieu-Ponel V, and Guiter F (2005) Vegetation dynamics in the North-West Mediterranean regions: Instability of the Mediterranean bioclimate. *Plant Biosystems* 13: 114–126.
- de Beaulieu J-L and Reille M (1984) A long Upper Pleistocene pollen record from les Echets, near Lyon, France. *Boreas* 13: 111–132.
- de Beaulieu J-L and Reille M (1992) Long Pleistocene pollen sequences from the Velay Plateau (Massif Central, France). I. Ribains maar. *Vegetation and Historical Archaeobotany* 1: 233–242.
- Denéfle M, Lezine AM, Fouache E, et al. (2000) A 12 000-year pollen record from Lake Maliq, Albania. *Quaternary Research* 54(3): 423–432.
- Desprat S (2005) *Réponses climatiques marines et continentales du Sud-Ouest de l'Europe lors des derniers interglaciaires et des entrées en glaciations*. Thesis, Bordeaux University.
- Desprat S, Sanchez-Goni MF, Turon JL, et al. (2006) Climatic variability of Marine Isotope Stage 7: Direct land-sea-ice correlation from a multiproxy analysis of a northwestern Iberian margin deep-sea core. *Quaternary Science Reviews* 25: 1010–1026.
- Florschütz F, Menendez Amor J, and Wijmstra TA (1971) Palynology of a thick Quaternary succession in southern Spain. *Palaeogeography, Palaeoclimatology, Palaeoecology* 10: 233–264.
- Follieri M, Magri D, and Sadori L (1988) 250 000-year pollen record from Valle di Castiglione (Roma). *Pollen et Spores* 30: 329–356.
- Gandouin E, Ponel P, Andrieu-Ponel V, et al. (2007) Past environment and climate changes at the last Interglacial/Glacial transition (Les Echets, France) inferred from subfossil chironomids (Insecta). *Comptes Rendus Geoscience* 399(5): 337–346.
- Gibbard PL and Head MJ (2010) The newly-ratified definition of the Quaternary system/period and redefinition of the Pleistocene series/epoch, and comparison of proposals advanced prior to formal ratification. *Episodes* 33: 152–158.
- Hampe A and Pétit RJ (2005) Conserving biodiversity under climate change: The rear edge matters. *Ecology Letters* 8: 461–467.
- Hewitt GM (2000) The genetic legacy of the Quaternary ice ages. *Nature* 405: 907–913.
- Iversen J (1958) The bearing of glacial and interglacial epochs on the formation and extinction of plant taxa. *Uppsala Universitets Arsskrift* 6: 210–215.
- Lezine A-M, von Grafenstein U, Andersen N, et al. (2010) Lake Ohrid, Albania, provides an exceptional multi-proxy record of environmental changes during the last glacial–interglacial cycle. *Palaeogeography, Palaeoclimatology, Palaeoecology* 287(1–4): 116–127.
- Lona F (1950) Contributi alla storia della vegetazione e del Clima nella Val Padana. Analisi pollinica del giacimento villafranchiano di Ieffe (Bergamo). *Atti della Società Italiana di Scienze Naturali* 89: 123–178.
- Lona F and Follieri M (1957) Successione pollinica della serie superiore (Günz-Mindel) di Ieffe (Bergamo). *Veröffentlichungen des geobotanischen Institutes Rübel in Zürich* 34: 86–98.
- Magny M, de Beaulieu J-L, Drescher-Schneider R, et al. (2007) Holocene climate changes in the central Mediterranean as recorded by lake-level fluctuations at Lake Accesa (Tuscany, Italy). *Quaternary Science Reviews* 26: 1736–1758.
- Magri D and Tzedakis PC (2000) Orbital signatures and long-term vegetation patterns in the Mediterranean. *Quaternary International* 73(74): 69–78.
- Margari V, Gibbard PL, Bryant CL, and Tzedakis PC (2009) Character of vegetational and environmental changes in southern Europe during the last glacial period: Evidence from Lesbos Island, Greece. *Quaternary Science Reviews* 28: 1317–1339.
- Margari V, Skinner LC, Tzedakis PC, Ganopolski A, Vautravers M, and Shackleton NJ (2010) The nature of millennial-scale climate variability during the past two glacial periods. *Nature Geoscience* 3: 127–133. <http://dx.doi.org/10.1038/NGEO740>.

- Massari F, Rio D, Serandrei Barbero R, et al. (2004) The environment of Venice area in the past two million years. *Palaeogeography, Palaeoclimatology, Palaeoecology* 202: 273–308.
- Milankovitch M (1930) Mathematische Klimalehre und astronomische Theorie der Klimaschwankungen. In: Köppen W and Geiger R (eds.) *Handbuch der Klimatologie* I (A): pp. 1–176. Berlin: Gebrüder Borntraeger.
- Moscariello A, Ravazzi C, Brauer A, et al. (2000) A long lacustrine record from the Pianico-Sèllere Basin (Middle–Late Pleistocene, Northern Italy). *Quaternary International* 73/74: 47–68.
- Müller UC, Pross J, Tzedakis PC, et al. (2011) The role of climate in the spread of modern humans into Europe. *Quaternary Science Reviews* 30(3–4): 273–279.
- Muttoni G, Ravazzi C, Breda M, et al. (2007) Magnetostratigraphic dating of an intensification of glacial activity in the southern Italian Alps during Marine Isotope Stage 22. *Quaternary Research* 67: 161–173.
- Okuda M, Yasuda Y, and Setoguchi T (2001) Middle to Late Pleistocene vegetation history and climate changes at Lake Kopais, Southeast Greece. *Boreas* 30: 73–82.
- Penck A and Brückner E (1901–1909) *Die Alpen im Eiszeitalter*, 1042 pp. Leipzig: Chr. Herm. Tauchnitz.
- Pérez-Obiol R and Julià R (1994) Climatic change on the Iberian Peninsula recorded in a 30 000-yr pollen record from Lake Banyoles. *Quaternary Research* 41: 91–98.
- Petit R, Aguinagale I, de Beaulieu J-L, et al. (2003) Glacial refugia: Hotspots but not melting pots of genetic diversity. *Science* 300: 1563–1565.
- Petit R, Brewer S, Bordacs S, et al. (2002) Identification of refugia and post-glacial colonisation routes of European white oaks based on chloroplast DNA and fossil evidence. *Forest Ecology and Management* 156: 49–74.
- Petit JR, Jouzel J, Raynaud D, et al. (1999) Climate and atmospheric history of the past 420 000 years from the Vostok ice core, Antarctica. *Nature* 399: 429–436.
- Pini R, Ravazzi C, and Donegana M (2009) Pollen stratigraphy, vegetation and climate history of the last 215 ka in the Azzano Decimo core (plain of Friuli, North-Eastern Italy). *Quaternary Science Reviews* 28(13–14): 1268–1290.
- Pini R, Ravazzi C, and Reimer PJ (2010) The vegetation and climate history of the last glacial cycle in a new pollen record from Lake Fimon (Southern Alpine foreland, N-Italy). *Quaternary Science Reviews* 29(23–24): 3115–3137.
- Pinti DL, Quidelleur X, Chiesa S, Ravazzi C, and Gillot P-Y (2001) K–Ar dating of an early Middle Pleistocene age distal tephra in the interglacial varved succession of Pianico (Southern Alps, Italy). *Earth and Planetary Science Letters* 5819: 1–7.
- Pons A and Reille M (1988) The Holocene- and Upper Pleistocene pollen record from Padul (Granada, Spain): A new study. *Palaeogeography, Palaeoclimatology, Palaeoecology* 66: 243–263.
- Pross J, Tzedakis PC, Schmiedl G, et al. (2007) Tenaghi Philippon (Greece) revisited: Drilling a continuous lower-latitude terrestrial climate archive of the last 250 000 years. *Scientific Drilling* 5: 44–46.
- Quézel P and Médail F (2003) *Ecologie et biogéographie des forêts du bassin méditerranéen*. Paris: Elsevier.
- Ravazzi C and Rossignol-Strick M (1995) Vegetation change in a climatic cycle of Early Pleistocene age in the Lefke Basin (northern Italy). *Palaeogeography, Palaeoclimatology, Palaeoecology* 117: 105–122.
- Reille M, Andrieu V, de Beaulieu J-L, Guenet P, and Goeury C (1998) A long pollen record from Lac du Bouchet, Massif Central, France for the period 325 to 100 ka (OIS 9c to OIS 5e). *Quaternary Science Reviews* 17: 1107–1123.
- Reille M and de Beaulieu J-L (1995) Long Pleistocene pollen records from the Praclaux Crater, South-Central France. *Quaternary Research* 44: 205–215.
- Reille M, de Beaulieu J-L, Svobodova H, Andrieu-Ponel V, and Goeury C (2000) Pollen biostratigraphy of the last five climatic cycles from a long continental sequence from the Velay region (Massif Central, France). *Journal of Quaternary Science* 15(7): 665–685.
- Rioul P, Andrieu-Ponel V, de Beaulieu J-L, Reille M, Svobodova H, and Battarbee RW (2007) Diatom responses to limnological and climatic changes at Ribains maar (French Massif Central) during the Eemian and Early Würm. *Quaternary Science Reviews* 26(11–12): 1557–1609.
- Roucoux KH, Tzedakis PC, Frogley MR, Lawson IT, and Preece RC (2008) Vegetation history of the marine isotope stage 7 interglacial complex at Ioannina, NW Greece. *Quaternary Science Reviews* 27: 1378–1395.
- Roucoux KH, Tzedakis PC, Lawson IT, et al. (2011) Vegetation history of the penultimate glacial period (Marine Isotope Stage 6) at Ioannina, north-west Greece. *Journal of Quaternary Sciences* 26(6): 616–626.
- Rouleau E, Pinti DL, Rouchon V, Quidelleur X, and Gillot P-Y (2009) Tephro-chronostratigraphy of the lacustrine interglacial record of Pianico, Italian Southern Alps: Identifying the volcanic sources using radiogenic isotopes and trace elements. *Quaternary International* 204: 31–43.
- Rull V and Vegas-Vilarrubia T (2011) What is long term in ecology? *Trends in Ecology and Evolution* 26(1): 3–4.
- Russo Ermolli E and Cheddadi R (1997) Climatic reconstruction during the middle Pleistocene: A pollen record from Vallo di Diano (Southern Italy). *Geobios* 30(6): 735–744.
- Samways MJ, Ponel P, and Andrieu-Ponel P (2006) Palaeobiodiversity emphasizes the importance of conserving landscape heterogeneity and connectivity. *Journal of Insect Conservation* 10: 215–218.
- Sanchez-Goni MF, Cacho I, Turon JL, et al. (2002) Synchronicity between marine and terrestrial responses to millennial scale climatic variability during the last glacial period in the Mediterranean region. *Climate Dynamics* 19: 95–105.
- Sanchez-Goni MF, Eynaud F, Turon JL, and Shackleton NJ (1999) High resolution palynological record off the Iberian margin: Direct land–sea correlation for the last interglacial complex. *Earth and Planetary Science Letters* 171: 123–137.
- Sanchez-Goni MF, Landais A, Fletcher WJ, et al. (2008) Contrasting impacts of Dansgaard-Oeschger events over a western European latitudinal transect modulated by orbital parameters. *Quaternary Science Reviews* 27: 1136–1151.
- Scardia G and Muttoni G (2009) Paleomagnetic investigations on the Pleistocene lacustrine sequence of Pianico-Sèllere (northern Italy). *Quaternary International* 204: 44–53.
- Serčelj A (1966) Pelodne analize Pleistocenskih in Holocenskih sedimentov Ljubljanskega barga. *Razprave IX, Slovenska Akademija Znanosti Umetnosti* 9: 431–472.
- Shackleton NJ, Sanchez-Goni MF, Paillet D, and Lancelot Y (2003) Marine Isotope Substage 5e and the Eemian Interglacial. *Global and Planetary Changes* 36: 151–155.
- Suc J-P, Bertini A, Combourieu-Nebout N, et al. (1995) Structure of west Mediterranean vegetation and climate since 5.3 Ma. *Acta Zoologica Cracoviense* 38: 3–16.
- Turon JL (1984) Direct land/sea correlations in the last interglacial complex. *Nature* 309(5970): 673–676.
- Tzedakis PC (1993) Long-term tree populations in northwest Greece through multiple Quaternary climatic cycles. *Nature* 364: 437–440.
- Tzedakis PC (1994) Vegetation change through glacial–interglacial cycles: A long pollen sequence perspective. *Philosophical Transactions of the Royal Society of London, Series B* 345: 403–432.
- Tzedakis PC (1999) The last climatic cycle at Kopais, central Greece. *Journal of the Geological Society of London* 155: 425–434.
- Tzedakis PC (2005) Towards an understanding of the response of Southern European vegetation to orbital and suborbital climate variability. *Quaternary Science Reviews* 25: 1585–1599.
- Tzedakis PC (2007) Seven ambiguities in the Mediterranean palaeoenvironmental narrative. *Quaternary Science Reviews* 26: 2042–2066.
- Tzedakis PC, Andrieu V, de Beaulieu J-L, et al. (1997) Comparison of terrestrial and marine records of changing climate of the last 500 000 years. *Earth and Planetary Science Letters* 150: 171–176.
- Tzedakis PC, Frogley MR, and Heaton THE (2003a) Last interglacial conditions in Southern Europe: Evidence from Ioannina, Northwest Greece. *Global and Planetary Change* 36: 157–170.
- Tzedakis PC, Frogley MR, Lawson IT, Preece RC, Cacho I, and de Abreu L (2004a) Ecological thresholds and patterns of millennial-scale climate variability: The response of vegetation in Greece during the last glacial period. *Geology* 32: 109–112.
- Tzedakis PC, Hooghiemstra H, and Pälike H (2006) The last 1.35 million years at Tenaghi Philippon: Revised chronostratigraphy and long-term vegetation trends. *Quaternary Science Reviews* 25: 3416–3430.
- Tzedakis PC, Lawson IT, Frogley MR, Hewitt GM, and Preece RC (2002) Buffered tree population changes in a Quaternary refugium: Evolutionary implications. *Science* 297: 2044–2047.
- Tzedakis PC, McManus G, Hooghiemstra H, Oppo DW, and Wijmstra TA (2003b) Comparison of changes in vegetation in northeast Greece with records of climate variability on orbital and suborbital frequencies over the last 450 000 years. *Earth and Planetary Science Letters* 21: 197–212.
- Tzedakis PC, Roucoux KH, de Abreu L, and Shackleton NJ (2004b) The duration of forest stages in southern Europe and interglacial climate variability. *Science* 306: 2231–2235. <http://dx.doi.org/10.1126/science.1102398>.
- van der Hammen T, Wijmstra TA, and Zagwijn H (1971) The floral record of the late Cenozoic of Europe. In: Turekian KK (ed.) *The Late Cenozoic Glacial Ages*, pp. 391–424. New Haven, CT: Yale University Press.
- van der Wiel AM and Wijmstra TA (1987a) Palynology of the lower part (78–120) of the core Tenaghi Philippon II, Middle Pleistocene of Macedonia, Greece. *Review of Palaeobotany and Palynology* 52: 73–88.

- van der Wiel AM and Wijmstra TA (1987b) Palynology of 112.8–197.8 m interval of the core Tenaghi Philippon III, Middle Pleistocene of Macedonia. *Review of Palaeobotany and Palynology* 52: 89–117.
- Verès D, Lallier-Verges E, Wohlfarth B, et al. (2009) Climate-driven changes in lake conditions during late MIS 3 and MIS 2: A high-resolution geochemical record from Les Echets, France. *Boreas* 38(2): 230–243.
- Wijmstra TA (1969) Palynology of the first 30 metres of a 120 m deep section in northern Greece. *Acta Botanica Neerlandica* 18: 511–527.
- Wijmstra TA and Smit A (1976) Palynology of the middle part (30–78 metres) of the 120 m deep section in northern Greece (Macedonia). *Acta Botanica Neerlandica* 25: 297–312.
- Wohlfarth B, Veres D, Ampel L, et al. (2008) Rapid ecosystem response to abrupt climate changes during the last glacial period in Western Europe, 40–16 ka. *Geology* 36(5): 407–410.
- Woillard G (1978) Grande Pile peat bog: A continuous pollen record for the last 140 000 years. *Quaternary Research* 9: 1–21.

Western North America

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Introduction

Pollen records from a variety of settings provide evidence of large-scale changes in vegetation and climate during the late Pleistocene (ca. 130 to 10 ka) in Western North America (WNA). This is the interval from the beginning of the last interglacial (MIS 5e) to the beginning of the current interglacial (the Holocene (MIS 1)). The late Pleistocene ended with the demise of the last of the megafauna (mammoths, mastodons, camels, llamas, horses, etc.) and was the last major episode of Earth history before humans transformed the globe. Our knowledge of the late Pleistocene in this region is uneven, and many more records are available for the last 20 ka of this interval than for the previous 100 kyr.

Pollen data from WNA illustrate several modes of vegetation change through the late Pleistocene, including shifts in the geographic and elevational ranges of plant taxa, and changes in the structure and composition of vegetation communities. The vegetation changes recorded in pollen records were caused primarily by global changes related to the radiation balance of the Earth and resulting shifts in climate zones. However, changes in the concentration of carbon dioxide in the atmosphere affected the water requirements of plants and may have caused them to be limited to wetter-than-present sites during portions of the late Pleistocene. Changes in the seasonal and geographic distributions of incoming solar radiation alone, independent of their effects on climatic circulation, may have also played a role in vegetation change, as may have changes in fire and other disturbance regimes (such as insect infestations). Ecological processes (e.g., succession), substrate requirements, and dispersal capabilities of individual taxa would have also influenced the pattern of vegetation change.

Environments of Western North America

WNA has a wide range of physiographic settings (Figure 1, Table 1) from the Great Plains on the eastern margin, through a broad region of high mountains and plateaus interspersed with valleys and plains, to the shores of the Pacific Ocean on the west. Winter temperatures are mild in the southwest and Mexico and along the Pacific Coast (Figure 2, upper left) while cold winters occur on the northern Great Plains and in the northern Rocky Mountains. Summers are hot in the southwest, in the Central Valley of California, and on the Great Plains (Figure 2, upper right). Although much of interior WNA is arid (due in large part to the rain shadows of the Sierra Nevada, Cascade, and other major mountain ranges), there are moist environments in the Pacific Northwest, northern Rocky Mountains, the high mountains of the central Rockies, and in the high Sierra Nevada mountains of eastern California (Figure 2, lower right; these are the regions with the most late Quaternary pollen records). Winter precipitation is dominant along the Pacific Coast, northern Idaho, and western Montana.

Summer rainfall prevails in northern Mexico, the eastern part of the southwest, parts of the Colorado Plateau, the southern Rocky Mountains, and the Great Plains. The regions between the pronounced areas of winter and summer precipitation have mixed regimes with varying ratios of warm-season to cool-season precipitation. Extremely dry summers occur today in central and southern California.

The present-day pattern of vegetation types follows the climatic gradients described above, and examples of the range of western vegetation are shown in Figures 3 and 4 (see Table 2 for a list of common and scientific names). Dense conifer-dominated forests occur in the wet environments of the Pacific Northwest and in northern California (Figure 3a). More open conifer-dominated forests occur in the high mountains of the interior (Figure 4a), frequently grading into a woodland where scattered trees are embedded in a matrix of shrubs (Figures 3b, 4c, and 4d). Alder and other hardwoods are present in the forests of the Pacific Northwest, while aspen, another hardwood, is a common plant in high mountains across WNA. Oak woodlands or scrub occur in California west of the crest of the Sierra Nevada (Figure 3c), in the southwest, and in the southern portions of the Colorado Plateau and Rocky Mountains. The intermountain valleys of the Great Basin and Colorado Plateau host shrublands dominated by sagebrush and other small-stature plants that are tolerant of aridity and heat (Figure 3d), and treeless tundra occurs in some of the highest mountains (Figure 4c). Grasslands occur on the Great Plains, in portions of the southwest and coastal California, in the central valley of California, and in eastern Washington.

Aspects of Quaternary Palynology in WNA

Fossil pollen records have been obtained in WNA and the surrounding region from a variety of water bodies, including the Pacific Ocean, large lakes in tectonic basins, small lakes and ponds (frequently in glaciated terrain), and wetland and mire deposits. The depositional setting influences the size of the area that is being recorded in the pollen data, with marine records representing broad areas whose rivers empty in the ocean near the pollen site. Large lakes with inflowing rivers may also receive pollen from a broad swath of vegetation spanning a large range of elevational and geographic settings. Pollen records from small lakes and mires usually reflect a much more local vegetation signal.

A relatively small number of wind-pollinated pollen taxa dominate most WNA pollen records. Unfortunately, some of the most prolific pollen producers represent a large number of species that grow under a broad range of environmental conditions. For example, many species of pine occur in woodlands and forests across the region under a variety of combinations of temperature and moisture conditions (Figure 5), but these different species generally cannot be differentiated based

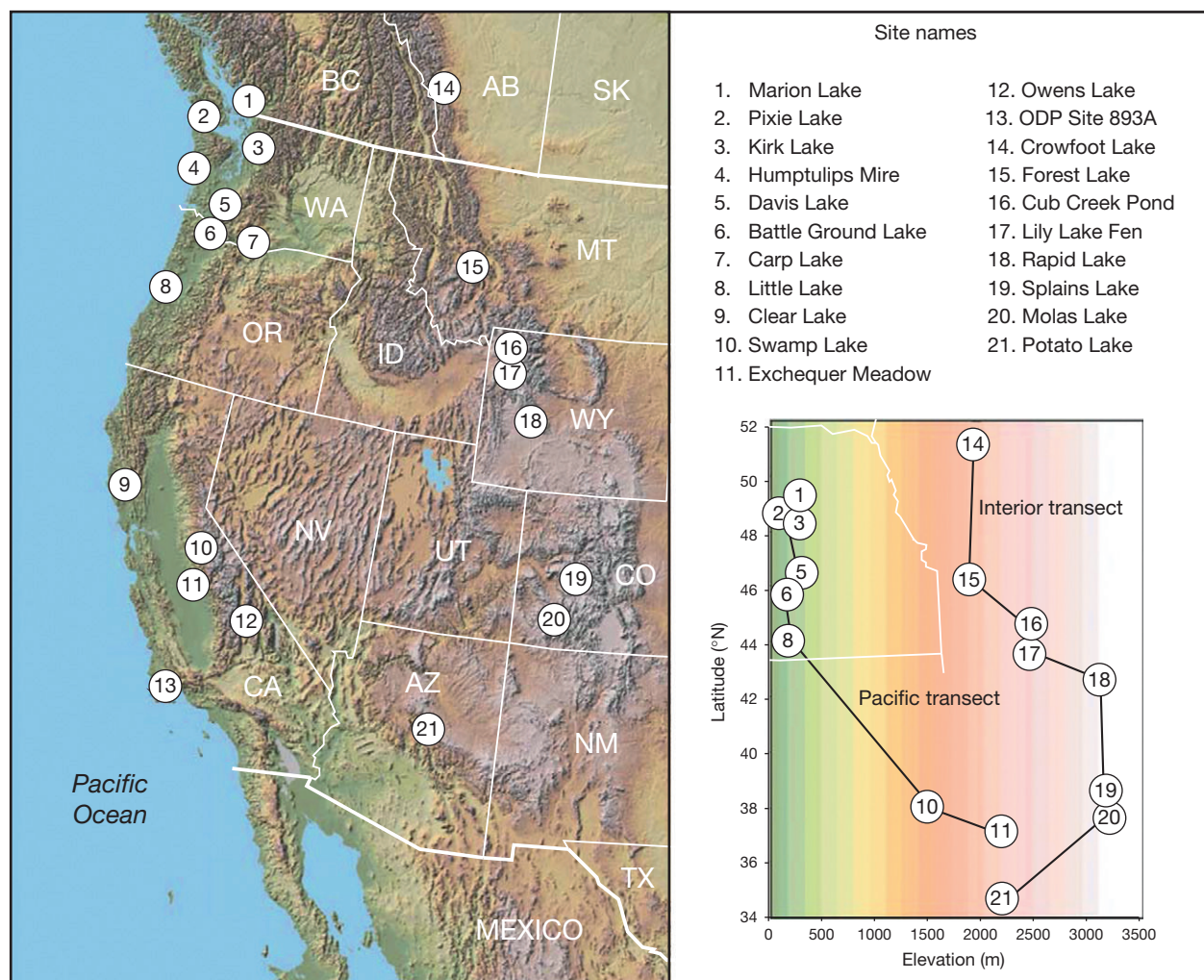


Figure 1 The left panel is a shaded relief map of the physiography of western North America. The numbered circles on this map show the location of sites discussed in the text (the site names are shown in the upper right panel). The lower right panel shows the latitudinal and elevational placement of the sites used in the Pacific and Interior Late Glacial transects. The vertical color bands in this panel provide a key to the elevations shown on the shaded relief map.

on their pollen. Juniper pollen also represents a large number of species from different environments, while spruce (Figure 5), fir, Douglas fir, and hemlock are limited to cool and/or moist climates. Oak lives under warm summers and mild winters, whereas sagebrush represents hot summers and dry conditions (Figure 5). The pollen of saltbush and related taxa provide evidence of hot and dry conditions. Many of the plants in tundra, grassland, and desert environments are insect-pollinated and release relatively small amounts of pollen. Such plants are frequently invisible in pollen records, even if they are abundant on the landscape. The pollen of pine and other windborne 'exotic' pollen taxa blown in from nearby communities frequently overwhelms the native pollen of these communities.

Late Pleistocene Pollen Records

There are many pollen records in WNA and it is not feasible to cover all of them in detail in this article. Instead, the emphasis

here is on some of the key modes of change, their apparent rapidity, and the interpreted causes of these changes. The pollen records discussed have differing spatial and temporal scales of resolution and differing levels of age control between and within records. Pollen records from Ocean Drilling Program (ODP) cores, such as that from ODP 893a from the Santa Barbara Basin (discussed below), can be dated through reference to the marine oxygen isotope stratigraphy of the host sediments. For terrestrial records, radiocarbon dating provides age control for the past 40 ka and volcanic ash stratigraphy (tephrochronology) provides additional chronological controls, especially in the Pacific Northwest and northern Rocky Mountains. Unfortunately, it is extremely difficult to obtain reliable age controls on lacustrine sediments that are older than the reach of radiocarbon dating (all ages discussed below are given in calendric time). Consequently, the following text will focus on the overall patterns of change and variability through the late Pleistocene, rather than on detailed correlations among these records or between these records

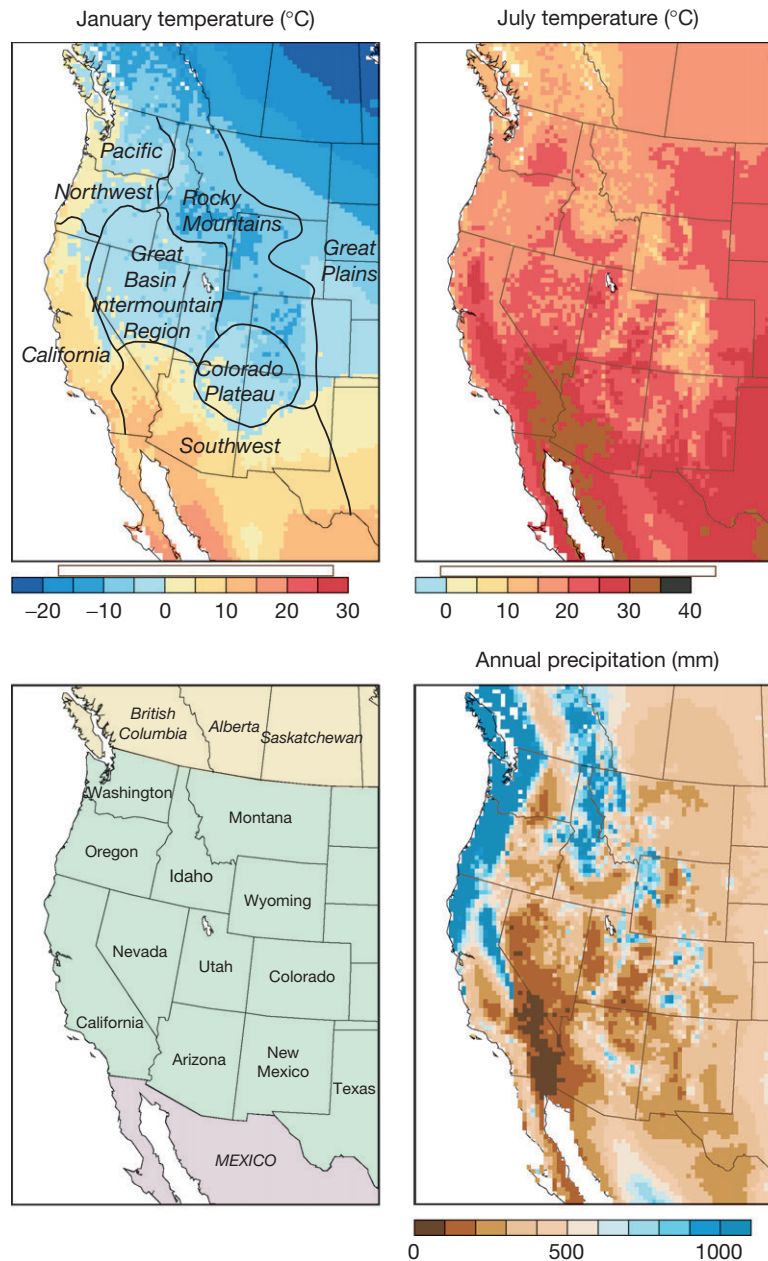


Figure 2 Aspects of the modern climate of western North America are shown in the right panels and the upper left panel. A sketch map of major geographic regions overlies the climate data in the upper left panel and the names of states and provinces are shown in the lower left panel.

and other chronologies of climate change. The following discussion is organized by time following the terminology of the marine isotope stages (see [Table 3](#)).

Marine Isotope Stage 5

Few WNA pollen records extend back through Marine Isotope Stage (MIS) 5 and here we examine the pattern and general timing of variability during this time in four of the most continuous pollen profiles. The record from ODP Site 893A ([Heusser, 2000](#)) in the Santa Barbara Basin provides a well-dated suite of some of the most dramatic changes seen in a broad source area in southern California. The transition from

the penultimate glaciation (MIS 6) into MIS 5 is registered by the rapid decline of juniper approximately 130 ka ([Figure 6](#)) followed by a succession of maxima of pine, then oak, and then sagebrush over the course of the last interglacial. Oak reached its highest levels in the record between 130 and 120 ka, presumably reflecting the warmest conditions. Two subsequent high periods of oak (MIS 5c and 5a) alternate with periods of juniper dominance (MIS 5d and 5b), which, coupled with the ages provided by the marine oxygen isotope stratigraphy, reflect the warm-cold-warm-cold-warm sequence seen in the marine record of MIS 5.

Owens Lake, now drained to feed the water needs of southern California, was located at the southern end of Owens



Figure 3 Examples of present-day vegetation in the states adjacent to the Pacific Ocean: (a) dense old-growth conifer forest in western Oregon; (b) juniper woodland in a sagebrush matrix in eastern Oregon; (c) oak woodland in southern California; (d) sagebrush shrubland in eastern Oregon. Photographs by R. Thompson.

Valley, on the east side of the Sierra Nevada (**Figure 6**). The interval of the pollen record from this site that is thought to cover the last interglacial (**Woolfenden, 2003**) is shown in **Figure 6**. Although the assigned ages are inexact, the Owens Lake record shows an apparently rapid transition from MIS 6 to MIS 5 environmental conditions approximately 130 ka, similar to the ODP 893A record from the other side of the Sierra Nevada. At Owens Lake, juniper declined rapidly at the end of the penultimate glacial interval (MIS 6), followed by the beginning of a steady rise in pine (an increasing trend that continued through MIS 5e) and then increases in mountain mahogany, sagebrush, and saltbush. Oak (not shown in **Figure 6** at Owens Lake) achieved higher pollen percentages during the last interglacial here than during the current interglacial, perhaps indicating wetter conditions.

At Clear Lake in northern California west of the Sierra Nevada, oak provides a strong signal in the pollen record (**Adam et al., 1988**) that is interpreted to be indicative of warm conditions, as in southern California (**Figure 7**). Here oak pollen is at its highest levels of the late Pleistocene immediately prior to ~120 ka, suggesting conditions warmer than those of the Holocene during the last interglacial. As with the record from ODP 893A, fluctuations in the pollen data appear to register the sequence of changes seen in the marine record through MIS 5 (although the age controls for the Clear Lake record are not sufficient to prove this point).

The pollen record from Carp Lake, Washington (**Whitlock et al., 2000**), located on the eastern flank of the Cascade Range,

begins with what appears to be the last interglacial (**Figure 7**). Some of the pollen taxa (including juniper and oak) represented in MIS 1 (the current interglacial) are present prior to ~120 ka in this record, and spruce is nearly absent. Spruce, an indicator of colder conditions at this site, is present after that and the relatively high levels of sagebrush and grass pollen suggest open and somewhat dry environments. There are no indicators of pronounced warmth until what appears to be MIS 5a at ca. 80 to 75 ka. The vegetation of this interval was like nothing else during the late Pleistocene — juniper, western hemlock, and oak (along with fir, western hemlock, and saltbush, which are not shown in **Figure 7**) were at high levels, and, unlike the MIS 1 portion of the record, pine and red alder were at low levels.

Marine Isotope Stages 4 and 3

Juniper and sagebrush increase ca. 70 ka in both the ODP 893A and Carp Lake records (**Figures 6 and 7**), indicating the arrival of colder and drier conditions. Oak increases after ~60 ka in the Clear Lake record, but remains well below the levels seen in MIS 5 and MIS 1, suggesting sustained cool conditions through MIS 3 at this site. Oak is absent outside of MIS 5 and MIS 1 in the ODP893A pollen record.

At Carp Lake the onset of cooler and drier conditions nearly 70 ka is marked by decline in oak and juniper and increase in pine and sagebrush (**Figure 7**). Although conditions at this site were more 'glacial' during MIS 4 than during MIS 5 and MIS 3,



Figure 4 Examples of present-day vegetation in Rocky Mountain/Colorado Plateau region: (a) moist sub-alpine spruce-fir forest in Utah; (b) a dry bristlecone pine stand at treeline in Colorado; (c) alpine tundra in Colorado; (d) lower treeline of a ponderosa pine-Douglas fir forest in Colorado. Photographs by R. Thompson.

Table 1 Locations and citations for sites mentioned in the text

No.	Site name	State or province	Latitude ($^{\circ}$ N)	Longitude ($^{\circ}$ W)	Elevation (m)	Reference
1	Marion Lake	British Columbia	49.308	122.547	305	Mathewes, 1973
2	Pixie Lake	British Columbia	48.596	124.197	70	Brown and Hebda, 2002
3	Kirk Lake	Washington	48.233	121.617	190	Cwynar, 1987
4	Humptulips Mire	Washington	47.283	123.911	110	Heusser, C.J. et al., 1999
5	Davis Lake	Washington	46.592	122.250	282	Barnosky, 1981
6	Battle Ground Lake	Washington	45.800	122.492	155	Barnosky, 1985
7	Carp Lake	Washington	45.918	120.881	714	Whitlock et al., 2000
8	Little Lake	Oregon	44.168	123.582	217	Worona and Whitlock, 1995
9	Clear Lake	California	39.000	122.450	404	Adam et al., 1988
10	Swamp Lake	California	37.950	119.817	1554	Smith and Anderson, 1992
11	Exchequer Meadow	California	37.000	119.833	2219	Davis and Moratto, 1988
12	Owens Lake	California	36.381	117.966	1067	Woolfenden, 2003
13	ODP Site 893a	California	34.300	120.000	-576	Heusser, L.E., 2000
14	Crowfoot Lake	Alberta	51.650	116.417	1940	Reasoner, 1996
15	Forest Lake	Montana	46.452	112.166	1895	Brant, 1980
16	Cub Creek Pond	Wyoming	44.490	110.284	2500	Waddington and Wright, 1974
17	Lily Lake Fen	Wyoming	43.767	110.317	2469	Whitlock, 1993
18	Rapid Lake	Wyoming	42.729	109.194	3134	Fall et al., 1994
19	Splains Lake	Colorado	38.833	107.078	3165	Fall, 1997
20	Molas Lake	Colorado	37.746	107.684	3200	Maier, 1961
21	Potato Lake	Arizona	34.462	111.345	2205	Anderson, 1993

it was not as cold and dry as during MIS 2, when sagebrush and grass dominate the Carp Lake record. Pine declined after 60 ka and sagebrush and grass increased. This was the beginning of a long and oscillatory slide toward the glacial conditions

Table 2 Common and scientific names for pollen taxa and plants mentioned in the text

<i>Conifers</i>	
Douglas fir	<i>Pseudotsuga menziesii</i>
Fir	<i>Abies</i>
Hemlock	<i>Tsuga</i> (mountain hemlock is <i>T. mertensiana</i> , western hemlock is <i>T. heterophylla</i>)
Juniper	<i>Juniperus</i> (the pollen type as used here may include other genera in the Cupressaceae and related families)
Pine	<i>Pinus</i> (pinyon pine, as used here, includes <i>P. edulis</i> and <i>P. monophylla</i> ; ponderosa pine is <i>P. ponderosa</i>)
Spruce	<i>Picea</i> (sitka spruce is <i>P. sitchensis</i>)
<i>Hardwood trees</i>	
Alder	<i>Alnus</i> (red alder is <i>A. rubra</i>)
Aspen	<i>Populus tremuloides</i>
Birch	<i>Betula</i>
Mountain mahogany	<i>Cercocarpus</i>
Oak	<i>Quercus</i>
<i>Shrubs and herbs</i>	
Grass	Poaceae
Sagebrush	<i>Artemisia</i> section <i>tridentatae</i> (this includes <i>A. tridentata</i> (Figure 5) and related dry-adapted species)
Saltbush	Chenopodiaceae and <i>Amaranthus</i>

of MIS 2, although a marked short-lived reversal of the trend is seen in the increase in pine and decrease in sagebrush and grass ca. 40 ka.

The pollen record from Humptulips Mire in western Washington (Heusser et al., 1999) apparently began during MIS 4 (Figure 7). The levels of spruce and grass pollen during this interval suggest cold and dry conditions, perhaps exceeding those of MIS 2 at this site. The subsequent increase in pine and western hemlock after ~60 ka suggests warmer conditions, although the high levels of pine and the low levels of alder suggest cooler and drier conditions than during MIS 1.

Marine Isotope Stage 2

There are substantially more data available for MIS 2 than for earlier times, although many montane sites were occupied by ice at the LGM (upper left of Figure 8). For the California and Washington State records discussed above, the LGM is generally the interval of greatest dominance of pollen taxa indicative of cold and/or dry conditions. In the Clear Lake record, oak is at its lowest levels (Figure 7), and the Carp Lake record has its highest values for spruce, sagebrush, and grass (coupled with the lowest values for pine). At Carp Lake, the end of MIS 2 is marked by a sharp transition as spruce and sagebrush essentially disappear and pine, western hemlock, oak, and red alder rise. The Humptulips Mire record also shows a marked change at this time, with decreases in pine, grass, and mountain hemlock and increases in western hemlock, alder, and spruce (presumably here the maritime Sitka spruce rather than the cold-adapted species of the interior).

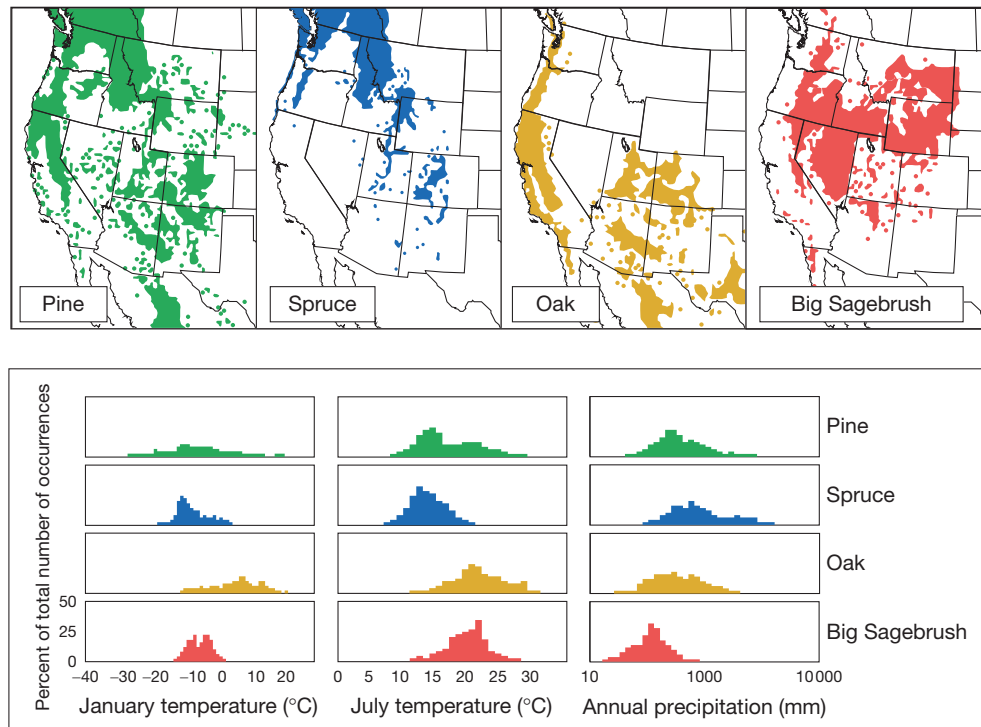


Figure 5 The present-day geographic and climatic distributions in western North America of four taxa that are well-represented in late Pleistocene pollen in the region (Little, 1971, 1976). The heights of the bars in the histograms indicate the percentage of the total number of occurrences of the taxon that falls within the climatic range indicated on the horizontal axis (Thompson et al., 1999a,b).

Table 3 The approximate ages (from [Martinson et al., 1987](#)) and global environments for the marine isotope stages (MIS)

MIS 5:	130 000 to 75 000 yr BP, the marine isotope stratigraphy has large amplitude swings within this interval with warm substages (from oldest to youngest) 5e, 5c, and 5a separated by relatively cool conditions during substages 5d and 5b. MIS 5e is often viewed as the 'last interglacial.'
MIS 4:	75 000 to 60 000 yr BP, the initial period of continental-scale glaciation, which was of a smaller magnitude than the Last Glacial Maximum (LGM).
MIS 3:	60 000 to 25 000 yr BP, a long interval of conditions intermediate between the warmth of the interglacials and the cold of the LGM.
MIS 2:	25 000 to 10 000 yr BP, the interval encompassing the LGM (~21 000 yr BP) and the transition to the current interglacial.
MIS 1:	10 000 yr BP to the present: The current interglacial, the Holocene.

Biomes of the LGM

The greater number of sites available for the LGM provides the basis for synoptic-scale reconstruction of past biomes and comparisons with their modern distributions. The lower portion of [Figure 8](#) shows the modern geographic distributions (compiled from the data presented by [Küchler \(1964\)](#)) and climatic tolerances of three biomes that cover most of WNA, and the upper right panel shows the distribution of these biomes as estimated from modern pollen data ([Thompson and Anderson, 2000](#)). The biomes presented here are forest (where trees dominate and there is relatively little open vegetation; examples shown in [Figures 3a](#) and [4a](#)), woodland (trees scattered with a matrix of shrubs and herbs; examples – [Figures 3b, 3c, 4b, and 4d](#)), and shrubland–grassland (trees are largely absent; example – [Figure 3d](#)).

In general, there is a good match between the two modern maps, although the pollen-based maps do have woodland

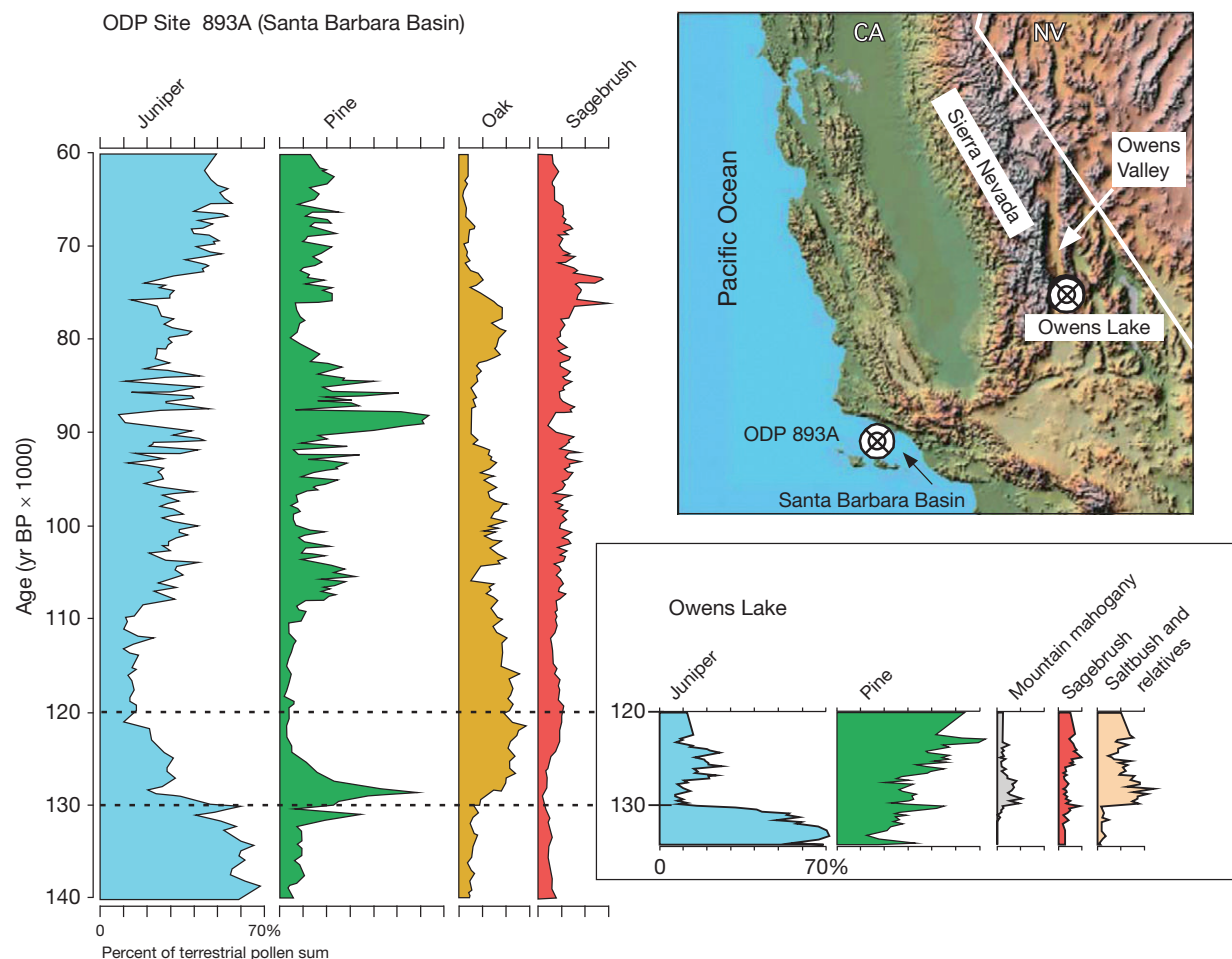


Figure 6 Marine isotope stage (MIS) 5 in the pollen record from ODP Site 893A and MIS 5e in the pollen record from Owens Lake, California ([Table 1](#)). These data were digitized from the published diagrams. For ODP Site 893A the age model was taken from the published diagram on pollen groups and applied here to the individual pollen taxa.

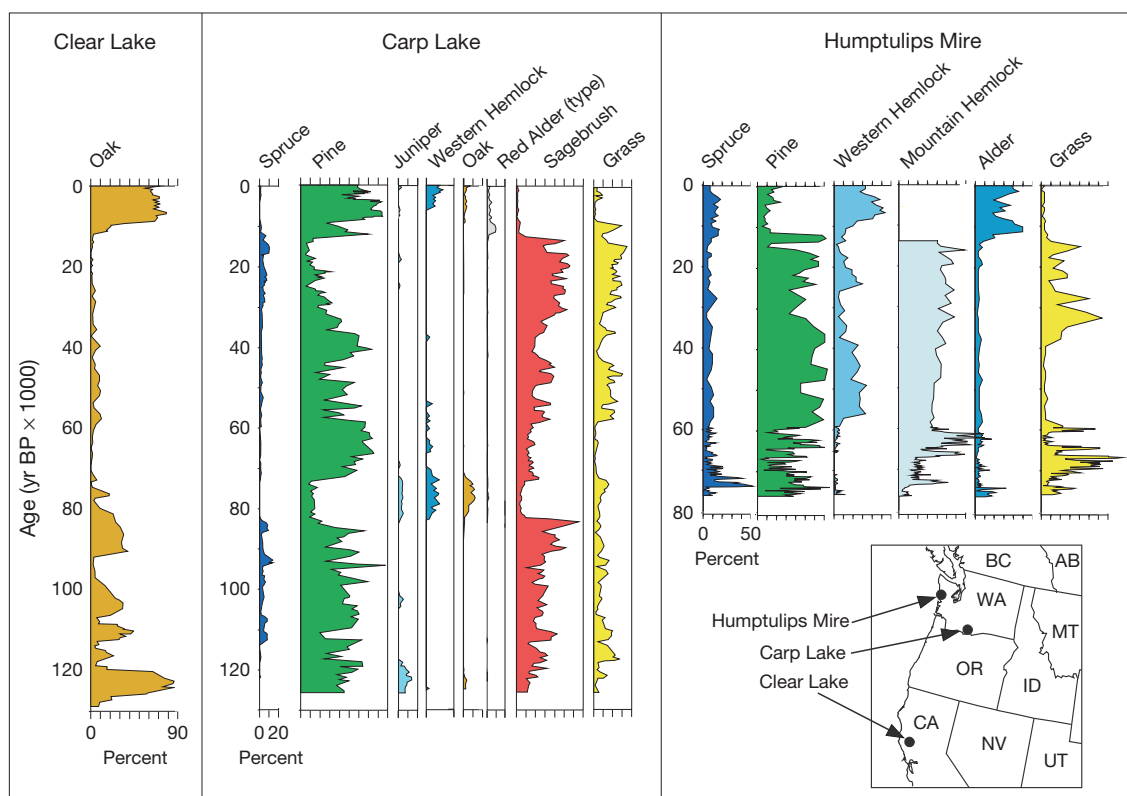


Figure 7 Records spanning all or much of the late Pleistocene from Clear Lake, California, Carp Lake, Washington, and Humptulips Mire, Washington (Table 1). These data were digitized from the published diagrams. For Clear Lake the ages were determined by taking the MIS boundaries plotted by Adam et al. (1988) and interpolating the ages of the samples between these boundaries.

farther north (this may be a mismatch between the two approaches, because pollen spectra receiving input from forest and shrubland–grassland would integrate them into a woodland). Conifer-dominated forests prevail in the Pacific Northwest and in the high elevations of the Sierra Nevada and Rocky Mountains. A few lower-elevational pollen records (yellow in Figure 8) show the presence of sagebrush- or grassland-dominated vegetation. Woodland (usually with pinyon pine, juniper, and/or oak short-stature trees) occurs in dry environments on the lower mountain slopes and plateaus of the interior, and shrubland–grassland occurs on the Great Plains, in the intermountain valleys, and in the central valley of California. Although the geographic ranges of the biomes intermingle and their climatic ranges overlap (Figure 8), in general, forest has the highest moisture levels combined with the lowest temperatures. Shrubland–grassland has lower moisture availability coupled with warm summers and winters as cold as those in forests. Woodland has intermediate moisture availability, with summers as warm as shrubland–grassland, but winters warmer than the other two biomes.

The biome reconstructions for the LGM (Figure 8, upper left panel) indicate that forest was restricted to the westernmost maritime portion of the Pacific Northwest. Across most of the rest of the landscape the vegetation was an open-conifer woodland, where subalpine conifers (in the north) or pygmy woodland conifers (to the south) lived in an open matrix with sagebrush, grass, and other steppe plants. Cold and dry climates are inferred for these latter areas, although the

physiological effects of low levels of atmospheric carbon dioxide may have restricted trees to wetter sites than today (see Van de Water et al. (1994) for evidence that trees in WNA had a higher density of stomata at the LGM than today, due to the lower levels of atmospheric carbon dioxide. The larger number of stomata would have released more water vapor from the trees, requiring them to use more water than they do under present-day conditions). Plant macrofossil records in the southwestern deserts (at lower latitudes and lower elevations than the pollen records) indicate conditions wetter and cooler than at present. The contrast between drier conditions to the north and wetter conditions to the south suggests that the westerlies were displaced southward at the LGM (Thompson et al., 1993).

Transition from glacial to interglacial conditions

Most of the Pleistocene pollen records in western North America begin after the LGM, due in part to the presence of glaciers in the high mountain basins where many of the records are located (Figure 8). These glaciers began to retreat by 17 ka, and there are many pollen profiles that record the transition from glacial to interglacial conditions in WNA, a transition that is expressed differently in different places and at different elevations. Figure 9 illustrates some of the changes seen across WNA during the transition from glacial to interglacial (Holocene) conditions between 17 and 9 ka along two north–south transects of pollen sites (Figure 1), one along the Pacific margin and the other in the more dry and continental interior. As discussed below, there is considerable variability imposed

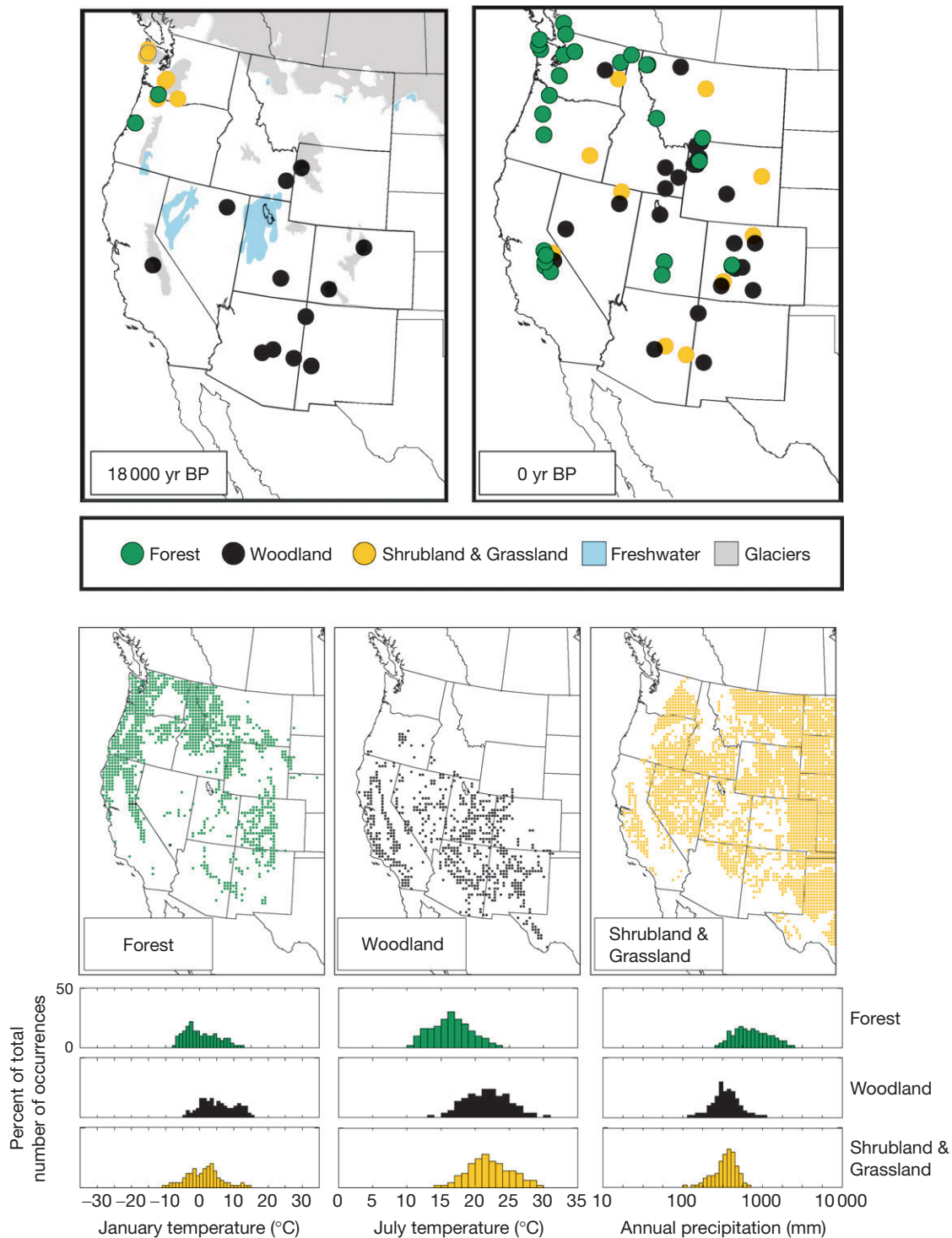


Figure 8 The upper panels illustrate the biomes of western North America at the Last Glacial Maximum (LGM) inferred from pollen data compared with those inferred for today (Thompson and Anderson, 2000). Glaciers and lakes on the LGM panel from Dyke et al. (2003). The lower panels illustrate the present geographic (Küchler, 1964) and climatic distributions of modern vegetation assemblages that approximate the biomes identified in the pollen data.

on the long-term trend at individual sites. Some of this variability may be tied to the Younger Dryas (e.g., Reasoner and Jodry (2000)) or other North Atlantic fluctuations, although the age controls available for most sites are not sufficient to make definitive correlations with distant events.

Pacific transect. This transect includes the wettest environments in present-day WNA (Figure 2), and elevations along it range from near-sea level in the north to high elevations in the Sierra Nevada to the south (Figure 1). At the three northern sites glacial-age pine forests are replaced by forests dominated

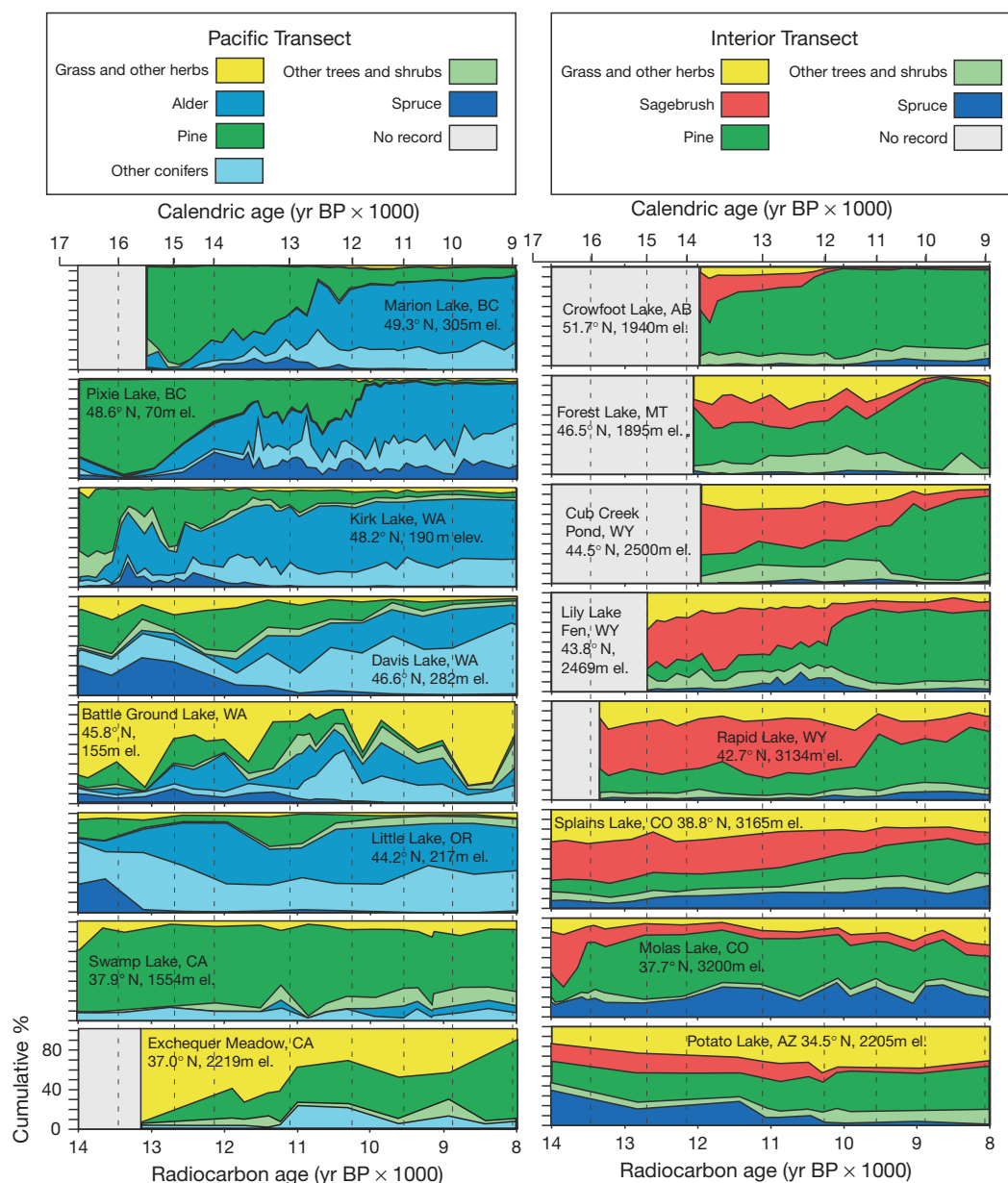


Figure 9 Two latitudinal transects of pollen sites illustrating the nature and timing of the transition from Pleistocene to Holocene vegetation in western North America (Table 1). The data are from the North American Pollen Database (NAPD, see Table 1 for citations and site information) and the age assignments are based on the NAPD age models. Calendric ages are shown on top, and radiocarbon years are plotted on the bottom (scaled by radiocarbon years indicated in NAPD files). Vertical dashed lines indicate 1000 year intervals in calendric time. The pollen taxa shown in this figure include pine, spruce, and sagebrush (illustrated in Figure 5), and alder. For the Pacific transect, the 'other conifers' category includes fir, Douglas fir, juniper, and hemlock. The 'other conifers' group is less important for the interior transect and is included in the 'other trees and shrubs' category. Sagebrush is relatively unimportant for the Pacific transect and is included in the 'grass and other herbs' category. Taxa shown in yellow and orange indicate open environments, those in green and blue are forest types.

by other conifers (including initially spruce, and then Douglas fir and hemlock) and alder by 16 to 13 ka. At the fourth site, Davis Lake (WA), spruce was more important than at the sites to the north at the close of the glacial interval, and there were more pollen indicators of open vegetation. These glacial taxa declined between 17 and 14 ka, and by 13 ka were largely replaced by other conifers (fir, Douglas fir, juniper) and alder. The transition at the next site to the south, Battle Ground

Lake, is more complex than the previous sites, with more open vegetation throughout and more variability. The early part of the interval had sagebrush and grass as indicators of open conditions; by the end of this interval sagebrush was largely gone, grass diminished, and there was a new suite of taxa indicating open vegetation. A peak in 'other conifers' prior to 12 ka is a brief interval of overlap between a short-lived maximum in fir pollen and the initial rise of Douglas fir. At this site

some of the important trees of the Holocene, including hemlock, were not established until 9 ka. Forests were present during the glacial interval at the final site of this transect in the Pacific Northwest (Little Lake (OR)), although sagebrush and other indicators of open vegetation were present. Here spruce departed earlier than 15 ka and forests with alder, fir, and Douglas fir became established.

The two southerly sites on the Pacific transect are located in the high Sierra Nevada Range of eastern California. At Swamp Lake, the basal glacial-age pollen spectra indicate pine-dominated forests with some indicators of open vegetation (including sagebrush and grass). Pine dominance continues through the glacial–interglacial transition, although there are significant increases in alder, other conifers (especially juniper), and other trees and shrubs about 13 ka. The most southerly site (Exchequer Meadow) is a high-elevation meadow today and the early part of its record is dominated by sagebrush, grass, and other plants indicative of open vegetation. The representation of trees increases through the interval of interest, with juniper at high levels between 13 and 12 ka, followed by pine dominance with fir and oak.

Interior transect. This transect covers environments that are generally colder, drier, and higher than those along the Pacific transect. The modern forests along the transect have pine, spruce, Douglas fir, and fir as common elements. The widespread occurrence of sagebrush (Figure 5) reflects the dryness and open vegetation of the valleys in the Rocky Mountain region. The basal spectra of the three northern sites date to slightly younger than 14 ka, and the base of Lily Lake Fen (WY) dates to ~15 ka. The earliest recorded vegetation at these four sites is significantly more open than that of today, and the sagebrush, grass, juniper, and birch pollen characteristic of this early interval are replaced by pine-dominated forest by 12 to 10 ka. The next site to the south, Rapid Lake (WY), shows a similar transition at ca. 11 ka when sagebrush and grass decline as pine, spruce, fir, and saltbush rise.

The transition at Splains Lake, farther south in Colorado, was similar in composition and scope, but apparently occurred more gradually as pollen taxa indicative of open environments decline from 70% at 16 ka to less than 30% by 11 to 10 ka. Pine, spruce, and fir increased at this site after 12 ka. At Molas Lake in southwestern Colorado, only the pollen spectra older than 16 ka have significant representations of open vegetation indicators. A forest with pine and spruce is dominant after that time, although grass and herbs increase after 12 ka. The southernmost site, Potato Lake (AZ), has a glacial-age spruce–pine forest with sagebrush that was gradually replaced by a pine–grass complex (with spruce disappearing around 12 ka). The trend at this site is similar to that seen in the macrofossil records from the southwest deserts, and, unlike the pollen records discussed above, forest here gave way to increasingly open and dry environments through the glacial–interglacial transition.

Summary

During the late Pleistocene in WNA, forests were at their greatest extents during the warm climate intervals in MIS 5 and MIS 1, perhaps due to both the climate and the relative abundance of

carbon dioxide in the atmosphere. During these times, oak and other warmth-requiring plants spread northward and to higher elevations. During the coldest climates of the late Pleistocene (MIS 4 and MIS 2) these plants retracted downslope and to the south, forests diminished, and plants adapted to cold and dry conditions spread into areas that today support moist forest environments. Vegetation apparently changed rapidly to many of the climatic variations described above, and the amplitude and rate varied among geographic and elevational settings.

See also: Pollen Analysis, Principles. **Paleobotany:** Overview of Terrestrial Pollen Data. **Plant Macrofossil Methods and Studies:** CO₂ Reconstruction from Fossil Leaves; Rodent Middens. **Plant Macrofossil Records:** Holocene North America. **Pollen Methods and Studies:** Databases and Their Application; The Biome Approach to Reconstructing Past Vegetation; Use of Pollen as Climate Proxies. **Pollen Records, Late Pleistocene:** Northern North America.

References

- Adam DP, Sims JD, and Throckmorton CK (1988) 130,000-yr continuous pollen record from Clear Lake, Lake County, California. *Geology* 9: 373–377.
- Anderson RS (1993) A 35,000 year vegetation and climate history from Potato Lake, Mogollon Rim, Arizona. *Quaternary Research* 40: 351–359.
- Barnosky CW (1981) A record of late Quaternary vegetation from Davis Lake, southern Puget lowland, Washington. *Quaternary Research* 16: 221–239.
- Barnosky CW (1985) Late Quaternary vegetation near Battle Ground Lake, southern Puget Trough, Washington. *Geological Society of America Bulletin* 96: 263–271.
- Brant LA (1980) *A Palynological Investigation of Postglacial Sediments at Two Locations Along the Continental Divide near Helena, Montana*. State College, Pennsylvania: Pennsylvania State University PhD Dissertation.
- Brown KJ and Hebda RJ (2002) Origin, development, and dynamics of coastal temperate conifer rainforests of southern Vancouver Island, Canada. *Canadian Journal of Forest Research* 32: 353–372.
- Cwynar LC (1987) Fire and the forest history of the North Cascade Range. *Ecology* 68: 791–802.
- Davis OK and Moratto MJ (1988) Evidence for a warm dry early Holocene in the western Sierra Nevada of California: Pollen and plant macrofossil analysis of Dinkey and Exchequer Meadows. *Madroño* 35: 132–149.
- Dyke AS, Moore A, and Robertson L (2003) Deglaciation of North America. *Geological Survey of Canada Open File* 1574.
- Fall PL (1997) Timberline fluctuations and late Quaternary paleoclimates in the southern Rocky Mountains, Colorado. *Geological Society of America Bulletin* 109: 1306–1320.
- Fall PL, Davis PT, and Zielinski GA (1994) Late Quaternary vegetation and climate of the Wind River Range, Wyoming. *Quaternary Research* 43: 393–404.
- Grimm, E. North American Pollen Database. www.ncdc.noaa.gov.
- Heusser CJ, Heusser LE, and Peteet DM (1999) Humptulips revisited: A revised interpretation of Quaternary vegetation and climate of western Washington, USA. *Palaeogeography, Palaeoclimatology, Palaeoecology* 150: 191–221.
- Heusser LE (2000) Rapid oscillations in western North America vegetation and climate during oxygen isotope stage 5 inferred from pollen data from Santa Barbara Basin (Hole 893A). *Palaeogeography, Palaeoclimatology, Palaeoecology* 161: 407–421.
- Küchler AW (1964) Potential natural vegetation of the conterminous United States. *American Geographical Society Special Publication* 36: 1–16, (with separate map at scale 1:3,168,000).
- Little EL Jr. (1971) Atlas of United States trees, volume 1: Conifers and important hardwoods. *Forest Service Miscellaneous Publication* 1146. Washington, DC: U.S. Department of Agriculture.
- Little EL Jr. (1976) Atlas of United States trees, volume 3: Minor western hardwoods. *Forest Service Miscellaneous Publication* 1314. Washington, DC: U.S. Department of Agriculture.

- Maher LJ Jr. (1961) *Pollen Analysis and Postglacial Vegetation History in the Animas Valley Region, Southern San Juan Mountains, Colorado*. Minneapolis: University of Minnesota, PhD Dissertation.
- Martinson DG, Pisias NG, Hays JD, Imbrie J, Moore TC Jr., and Shackleton NJ (1987) Age dating and the orbital theory of ice ages: Development of a high resolution 0 to 300,000-year chronostratigraphy. *Quaternary Research* 27: 1–29.
- Mathewes RW (1973) A palynological study of postglacial vegetation changes in the University Research Forest, southeastern British Columbia. *Canadian Journal of Botany* 51: 2085–2103.
- Reasoner MA (1996) *Late Quaternary Alpine and Sub-Alpine Lacustrine Records: Canadian and Colorado Rocky Mountains*. Edmonton: University of Alberta, PhD Dissertation.
- Reasoner MA and Jodry MA (2000) Rapid response of alpine timberline vegetation to the Younger Dryas climate oscillation in the Colorado Rocky Mountains, USA. *Geology* 28: 51–54.
- Smith SJ and Anderson RS (1992) Late Wisconsin paleoecologic record from Swamp Lake, Yosemite National Park, California. *Quaternary Research* 38: 91–102.
- Thompson RS, Anderson KH, and Bartlein PJ (1999a) Atlas of relations between climatic parameters and distributions of important trees and shrubs in North America – Introduction and conifers. *U.S. Geological Survey Professional Paper* 1650a: p. 269.
- Thompson RS, Anderson KH, and Bartlein PJ (1999b) Atlas of relations between climatic parameters and distributions of important trees and shrubs in North America – Hardwoods. *U.S. Geological Survey Professional Paper* 1650b: p. 423.
- Thompson RS and Anderson KH (2000) Biomes of western North America at 18,000, 6000 and 0 ¹⁴C yr BP reconstructed from pollen and packrat midden data. *Journal of Biogeography* 27: 555–584.
- Thompson RS, Shafer SL, Strickland LE, Van de Water PK, and Anderson KH (2004) Quaternary vegetation and climate change in the western United States: Developments, perspectives, and prospects. In: Gillespie AR, Porter SC, and Atwater BF (eds.) *Developments in Quaternary Science, 1: The Quaternary Period in the United States*, pp. 403–426. The Netherlands: Elsevier Amsterdam.
- Thompson RS, Whitlock C, Bartlein PJ, Harrison SP, and Spaulding WG (1993) Climatic changes in the western United States since 18,000yr b.p. In: Wright HE Jr., Kutzbach JE, Webb T III, Ruddiman WF, Street-Perrott FA, and Bartlein PJ (eds.) *Global Climates Since the Last Glacial Maximum*, pp. 468–513. Minneapolis: University of Minnesota Press.
- Van de Water PK, Leavitt SW, and Betancourt JL (1994) Trends in stomatal density and ¹³C/¹²C ratios of *Pinus flexilis* needles during the last glacial–interglacial cycle. *Science* 264: 239–243.
- Waddington JCB and Wright HE Jr. (1974) Late Quaternary changes on the east side of Yellowstone Park, Wyoming. *Quaternary Research* 4: 175–184.
- Whitlock C (1993) Postglacial vegetation and climate of Grand Teton and southern Yellowstone National Parks. *Ecological Monographs* 63: 173–198.
- Whitlock C, Sarna-Wojcicki AM, Bartlein PJ, and Nickmann RJ (2000) Environmental history and tephrostratigraphy at Carp Lake, southwestern Columbia Basin, Washington, USA. *Palaeogeography, Palaeoclimatology, Palaeoecology* 155: 7–29.
- Woolfenden WB (2003) A 180,000-year pollen record from Owens Lake, CA: Terrestrial vegetation change on orbital scales. *Quaternary Research* 59: 430–444.
- Worona MA and Whitlock C (1995) Late Quaternary vegetation and climate history near Little Lake, central Coast Range, Oregon. *Geological Society of America Bulletin* 107: 867–876.

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POLLEN RECORDS, POSTGLACIAL

Contents

Africa

Australia and New Zealand

Northeastern North America

Northwestern North America

Southeastern North America

Southwestern North America

South America

Northern Asia

Northern Europe

Southern Europe

Africa

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Introduction

During the last glacial–interglacial transition, the orbitally induced increase in solar radiation during summer had pronounced effects on tropical monsoon activity. Enhanced thermal contrast between land and sea and the related land–sea pressure gradient were responsible for strong inflow of moist maritime surface winds toward the African continent. Enhanced monsoon fluxes over tropical Africa led to the establishment of numerous lakes and the wide development of tropical plant communities throughout the African tropics during the Holocene (e.g., [Hoelzmann et al., 1998](#); [Lézine et al., 2011](#); [Watrin et al., 2009](#)). The greening of the Saharan desert supported abundant faunas and led to the early development of grazing and agriculture ([Petit-Maire, 1999](#)).

The shift from dry and cool to humid and warm climate conditions in tropical Africa has been studied by numerous authors (e.g., [de Menocal et al., 2000](#); [Gasse, 2000](#); [Kröpelin et al., 2008](#); [Lézine et al., 2005](#)) with the aim of investigating the timing and structure of the last glacial–interglacial transition in the African tropics. The role of the North Atlantic Ocean circulation has also been proposed to explain the millennial-scale climate oscillations that punctuated this interval. In addition, paleoclimatic models connecting environmental changes with variations in the Earth's orbit ([Kutzbach and Street-Perrott, 1985](#)) or investigating atmosphere–vegetation feedback in the climate system ([Broström et al., 1998](#)) have been used to study the mechanisms of climate variability.

Environmental reconstruction of tropical Africa from pollen data is severely limited by the incompleteness of the record, owing primarily to major discontinuities in lacustrine sediments

preserved in a predominantly arid climate. Long and complete pollen records from the interval lasting from the last glaciation to the present are extremely rare (e.g., Sacred Lake ([Coetzee, 1967](#)); Barombi M'Bo ([Maley, 1991](#)); Lake Bambili ([Assi-Kaudhis et al., 2008](#))). In this context, marine sedimentary sequences offer the possibility of providing additional evidence of environmental and climate changes on the nearby continent. In this review, I use pollen data from both continental and marine sedimentary sequences ([Figure 1](#) and [Table 1](#)) to discuss short- and long-term climate changes that occurred in tropical Africa during the last glacial–interglacial transition and the Holocene and the related variations in plant distribution.

Evidence of Past Atmospheric Circulation: Influx Values in Marine Pollen Sequences off West Africa

Core KS84068 (4°28'N, 4°11'W) recovered in the Gulf of Guinea off the Ivory Coast and core A180-48 (15°19'N, 18°06'W) recovered off Senegal, in zones out of any direct influence of river discharges, provide detailed evidence for important changes in the strength of the two main air fluxes that drive the West African climate ([Lézine, 1998](#)). These are the continental trade wind of NE–SW direction, which crosses the dry and hot Saharan desert, and the Atlantic monsoon, which blows in the opposite direction and carries humid air masses to the continent in summer. The confluence of these two air masses (the Inter-Tropical Convergence Zone) moves from roughly 4°N in January to 20–25°N in July ([Figure 2](#)). Continent–ocean exchanges are deduced from the variations in pollen influx (number of grains per cm² per year; [Figure 3](#)). Core KS8403, with a sediment accumulation rate

Table 1 List of African pollen sites cited in the text and figures

Site name	Country	References
<i>Continental pollen sites</i>		
Acacus	Lybia	Schulz (1994)
Abiyata	Ethiopia	Lézine and Bonnefille (1982)
Ahakagyezi	Uganda	Taylor (1990)
Albert	RD Congo	Beuning et al. (1997) Ssemmanda and Vincens (1993)
Badda	Ethiopia	Hamilton (1982)
Bafounda	Cameroon	Tamura (1990)
Baie-Saint-Jean	Mauritania	Lézine (unpublished)
Bal	Nigeria	Salzmann and Waller (1998)
Barombi-Mbo	Cameroon	Maley and Brenac (1998)
Bilanko	Congo	Elenga and Vincens (1990)
Bilma	Niger	Schulz et al. (1990)
Bir Atrun	Sudan	Ritchie and Haynes (1987)
Bogoria	Kenya	Vincens (1987)
Bosumtwi	Ghana	Maley and Livingstone (1983)
Bujuku	Uganda	Livingstone (1967)
Bwamba Pass	Uganda	Nakimera (2001)
Chemchane	Mauritania	Lézine et al. (1990)
Coraf	Congo	Elenga et al. (2001)
Danka	Ethiopia	Hamilton (1982)
Dega Sala	Ethiopia	Bonnefille and Mohammed (1994)
Diogo	Senegal	Lézine (1988)
Elmenteita	Kenya	Mworia-Maitima (1999)
Ijenda	Burundi	Roche and Bikwemu (1989)
Iringa	Tanzania	Mumbi et al. (2008)
Kaigama	Nigeria	Salzmann and Waller (1998)
Kaisungor	Kenya	Van Zinderen Bakker (1964)
Kajemarum	Nigeria	Salzmann and Waller (1998)
Kaméléte	Gabon	Ngomanda et al. (2005) Ngomanda et al. (2007)
Kamiranzovu	Rwanda	Hamilton (1982)
Karisoke	Rwanda	Ntaganda (1991)
Kashiru	Burundi	Bonnefille and Chalié (2000)
Kiguhu	Rwanda	Roche and Ntaganda (1999)
Kilimanjaro	Tanzania	Coetzee (1967)
Kimili	Kenya	Hamilton (1982)
Kitandara	Uganda	Livingstone (1967)
Kitina	Congo	Elenga et al. (1996)
Koitobos	Kenya	Hamilton (1982)
Kuluwu	Nigeria	Salzmann and Waller (1998)
Kuruyange	Burundi	Bonnefille et al. (1991)
Kuwasenkoko	Rwanda	Hamilton (1982)
Laboot	Kenya	Hamilton (1982)
Lompoul	Senegal	Lézine (1988)
Mahoma	Uganda	Livingstone (1967)
Maridor	Gabon	Ngomanda (2005)
Masoko	Tanzania	Vincens et al. (2003) Vincens et al. (2007)
Mbalang	Cameroon	Vincens et al. (2010)
Mboandong	Cameroon	Richards (1986)
Mubwinbi	Uganda	Marchant et al. (1997)
Muchoya	Uganda	Taylor (1990)
Naivasha	Kenya	Mworia-Maitima (1991)
Ndurumu	Burundi	Jolly and Bonnefille (1992)
Ngamakala	Congo	Elenga et al. (1994)
Nguène	Gabon	Ngomanda et al. (2007)
Nyabessan	Cameroon	Ngomanda et al. (2009)
Njupi	Cameroon	Zogning et al. (1997)
Orgoba	Ethiopia	Mohammed (1992)

(Continued)

Table 1 (Continued)

Site name	Country	References
Ossa	Cameroon	Reynaud-Farrera et al. (1996)
Oyo	Sudan	Ritchie (1994)
Potou	Senegal	Lézine (1988)
Rusaka	Burundi	Bonnefille et al. (1995)
Rutundu	Kenya	Coetzee (1967)
Sacred	Kenya	Coetzee (1967)
Seguedine	Mali	Baumhauer and Schulz (1984)
Sélé	Benin	Salzmann and Hoelzmann (2005)
Selima	Sudan	Ritchie and Haynes (1987)
Shum Laka	Cameroon	Kadomura and Kiyonaga (1994)
Simbiro	Ethiopia	Mohammed (1992)
Sinnda	Congo	Vincens et al. (1998)
Songolo	Congo	Elenga et al. (2001)
Tamsaa	Ethiopia	Mohammed and Bonnefille (1998)
Tanganyika	Tanzania	Vincens (1991)
Taoudenni	Mali	Schulz (1991)
Termit	Niger	Schulz et al. (1995)
Tibesti Mts	Chad	Schulz (1980)
Tilla	Nigeria	Salzmann (2000)
Tilo	Ethiopia	Lamb (2001)
Tjéri	Chad	Maley (1981)
Touba N'Diaye	Senegal	Lézine (1988)
Victoria	Uganda	Kendall (1969) Beuning (1999) Nakimera (2001)
Yoa	Chad	Kröpelin et al. (2008) Lézine (2009)
<i>Marine cores</i>		
A180-48		Lézine et al. (1995)
KS-84063		Lézine et al. (1994)
KW31		Lézine and Cazet (2005)

ranging from 4 to 10 cm per 1000 years, provides an exceptional record for the time interval of 19 000–3000 cal yr BP. Two well-marked peaks of pollen influx are recorded for the onset of deglaciation (2000 grains per cm² per year) and the Younger Dryas interval (3000 grains per cm² per year) at levels characterized by the occurrence of pollen grains from the Saharan desert (such as *Artemisia* (Sage) and *Ephedra* (Mormon tea)) and high concentrations of eolian dust. These testify to the enhanced trade wind circulation during dry events. Conversely, early Holocene levels centred ca. 9500 cal yr BP are characterized by minimum values of pollen influx (770 grains per cm² per year) as a response of SW–NE dominant wind fluxes during the early Holocene humid maximum. Core A180-48 confirms the importance of pollen influx values during the Younger Dryas event with a maximum of 9600 grains per cm² per year. The base of the core is poorly dated, which precludes any reliable record before 15 000 cal yr BP.

Reinforced eolian activity of NE direction over the eastern tropical Atlantic corresponds to phases of severe dryness on the nearby continent, leading to the lowering of lake and water table levels (Figure 3; Gasse, 2000). The low values of pollen influx recorded before the Younger Dryas (at a time interval corresponding to the Bölling/Alleröd pollen zones of Northern Hemisphere) and the early Holocene correspond to the reversal of dominant wind fluxes and enhanced monsoon circulation responsible for high lake levels throughout North Africa (Gasse, 2000).

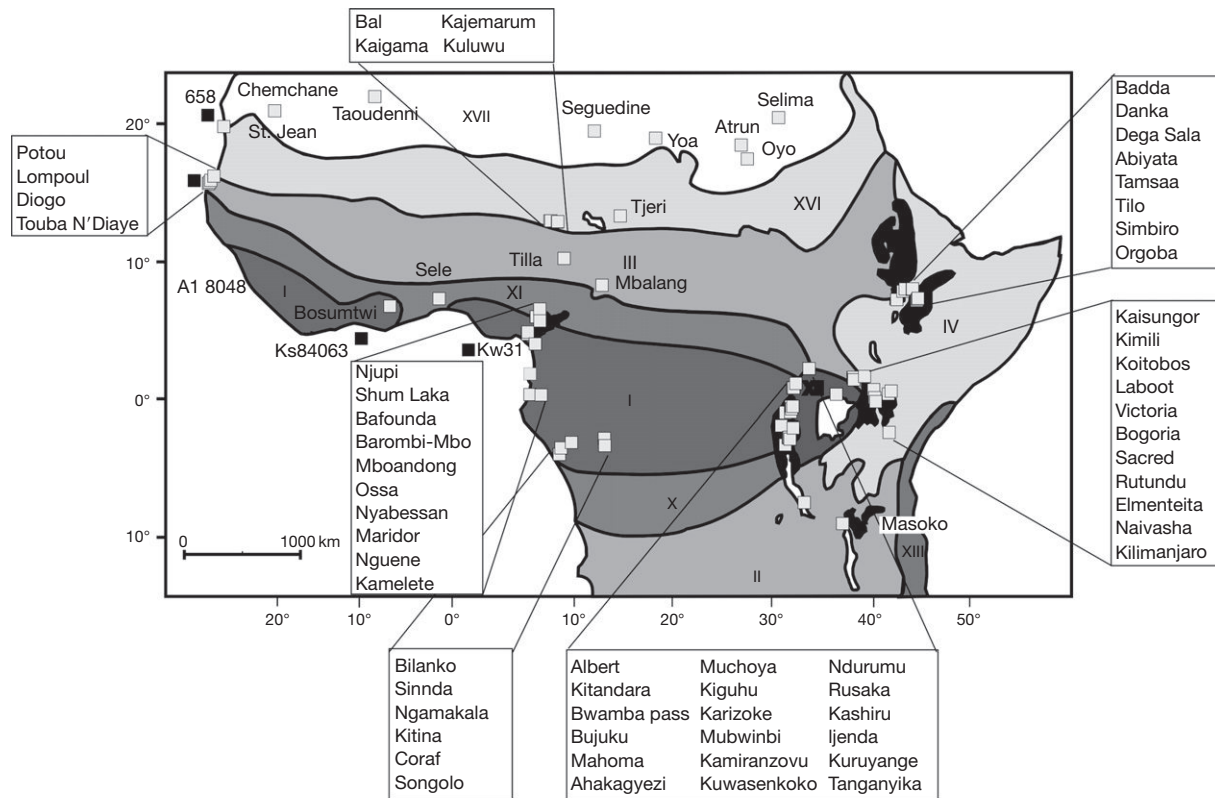


Figure 1 Location map of pollen sites mentioned in the text. Modern definitions of vegetation according to [White \(1983\)](#) with: I, Guineo–Congolian forests; II, Zambezian dry forests and woodlands (including Miombo); III, Sudanian dry forests and woodlands; IV, Somalia–Masaï wooded grasslands; X, Guineo–Congolian/Zambezian transition zone; XI, Guineo–Congolian/Sudanian transition zone; XII, Lake Victoria mosaic; XIII, Zanzibar–Inhambane mosaic; XVI, Sahelian wooded grasslands; XVII, Saharan desert. In black: Afro-montane forests and grasslands.

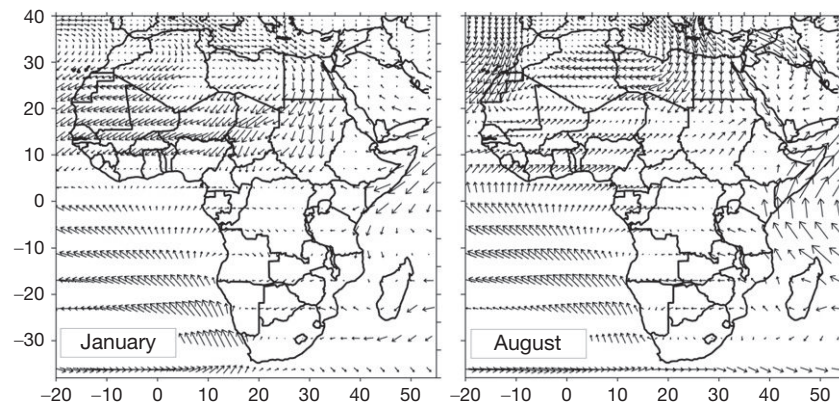


Figure 2 Modern surface wind fields for Africa for boreal winter (January) and summer (August). During boreal summer, heating over land surface allows moist maritime air from the equatorial Atlantic to penetrate into western and central subtropical Africa. Strong southwesterly winds associated with the summer Asian monsoon develop in East Africa, while westerly airstreams of the Atlantic monsoon penetrate eastward and reach the Eastern highlands. Regional atmospheric circulation reverses during boreal winter, and dry northeast winds blow over subtropical Africa.

Pollen records of the Younger Dryas dry event on the African continent are extremely rare because most of the lakes dried up at this time. It is, however, probable that the interruption of the forest advance at Lake Bosumtwi ($6^{\circ}32'N$, $1^{\circ}25'W$) between 15 000 and 10 600 cal yr ([Maley, 1991](#)) and Lake Albert ($1^{\circ}50'N$, $31^{\circ}10'E$) between 13 300 and 11 200 cal yr BP ([Beuning et al., 1997](#)), and the well-identified phase of extension of halophytic herbaceous vegetation at Lake Magadi

around 12 900 cal yr BP ([El Moutaki, 1994](#)) correspond to this event. In southeast Africa, however, rather different climate conditions prevailed as shown, for instance, at Lake Masoko ($9^{\circ}20'S$, $33^{\circ}45'E$) ([Vincens et al., 2007](#)) with high rainfall prior to 11 700 cal yr BP (including the Younger Dryas event) allowing a semideciduous forest (forest with some deciduous and some evergreen trees) to expand. A dramatic change in vegetation composition at 11 700 cal yr BP led to the wide expansion

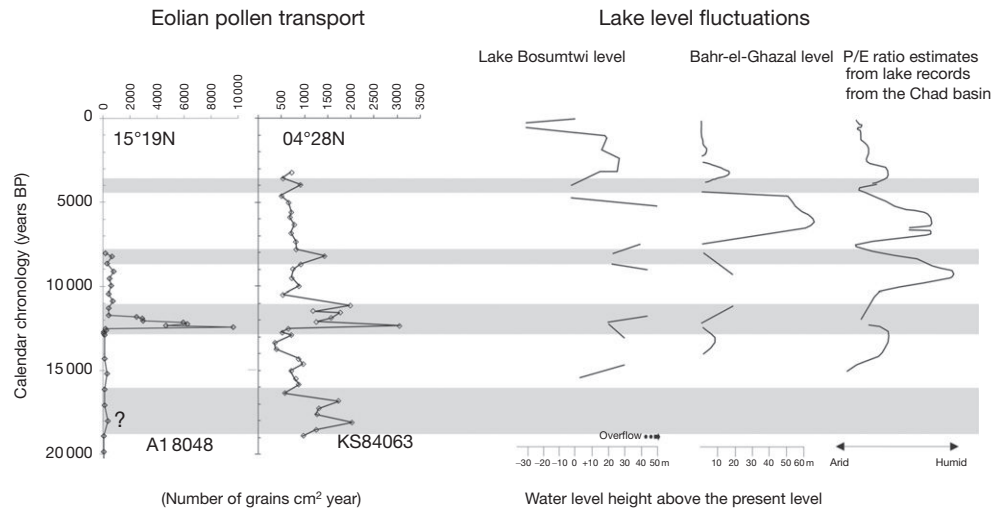


Figure 3 Pollen evidence for dry climate phases in western tropical Africa. Two curves at the left show the variations in trade wind intensity inferred from two pollen influx records (A18048 and KS84063) from the eastern Atlantic Ocean at 15° N and 4° N (Lézine, 1998). Peaks in pollen influx correspond to dry phases recorded by episodes of low lake level (in records from Lake Bosumtwi, Bahr-el-Ghasal, and the Chad basin) on the nearby continent. Two main dry phases are recorded at the onset of the deglaciation and the 'Younger Dryas' event, whereas two more discrete phases of trade wind intensity are dated from ca. 8000 to 4000 cal yr BP. Reproduced from Gasse F (2000) Hydrological changes in the African tropics since the Last Glacial Maximum. *Quaternary Science Reviews* 19: 189–211.

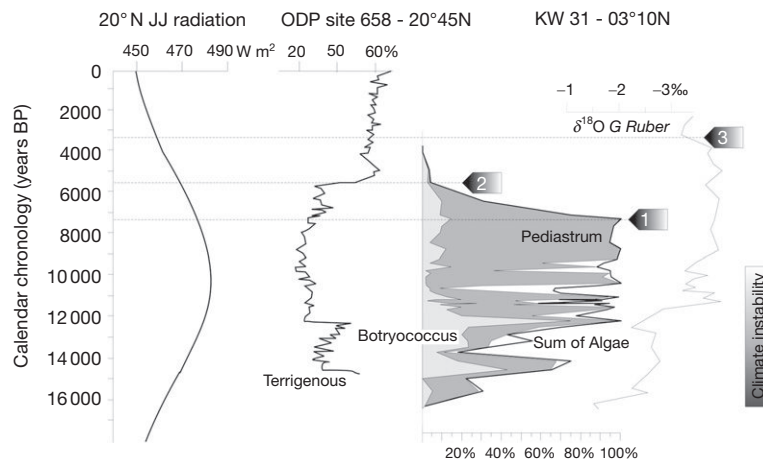


Figure 4 The African Humid Period as recorded around 21° N (ODP site 568 terrigenous record) from de Menocal et al. (2000) (left) and around 3° N (Core KW31 algae record) from Lézine et al. (2005) (right). The figure shows that the onset of deglaciation was very unstable, characterized by phases of intense flowing (peaks in *Pediastrum* curve) and interruption of river discharges to the ocean. Peaks of *Botryococcus* record phases of dominant surficial runoff in a poorly vegetated environment. At the end of the African Humid Period, the record of the onset of modern conditions was abrupt near the northern tropic and more progressive near the equator, occurring in three steps at 7000, 5500, and 3500 cal yr BP. (*G. Ruber* = *Globigerinoides ruber*, a foraminiferan sensitive to salinity changes).

of open and dry plant communities of Zambezi affinity. This change is interpreted as reflecting not only a decrease in annual rainfall but also a change in rainfall distribution with a dry season changing from 3–4 months prior to 11 700 cal yr BP to 5–6 months after.

Evidence of Past Rainfall Variations: Fresh-Water Algae Record in Marine Pollen Sequences off West Africa

Owing to its location off the mouth of the Niger River, the palynological content of core KW31 (3°31'1"N, 05°34'1"E;

Lézine et al., 2005) is thought primarily to reflect variations in fresh water flowing to the ocean. In this core, pollen grains are associated with an abundant and highly diverse fresh-water algal flora, including *Pediastrum* (a colonial green alga), which has been used to characterize the successive hydrological phases of the African Humid Period. As depicted by the *Pediastrum* curve, strong fresh-water discharges started ca. 14 500 cal yr BP, at the time of the Bölling/Allerød interval, with a peak dated from 13 650 cal yr BP (KW31) (Figure 4) and culminated during the Holocene. The whole interval, which corresponds to the increase in summer radiation at the last glacial–interglacial transition, was climatically very unstable, characterized by an

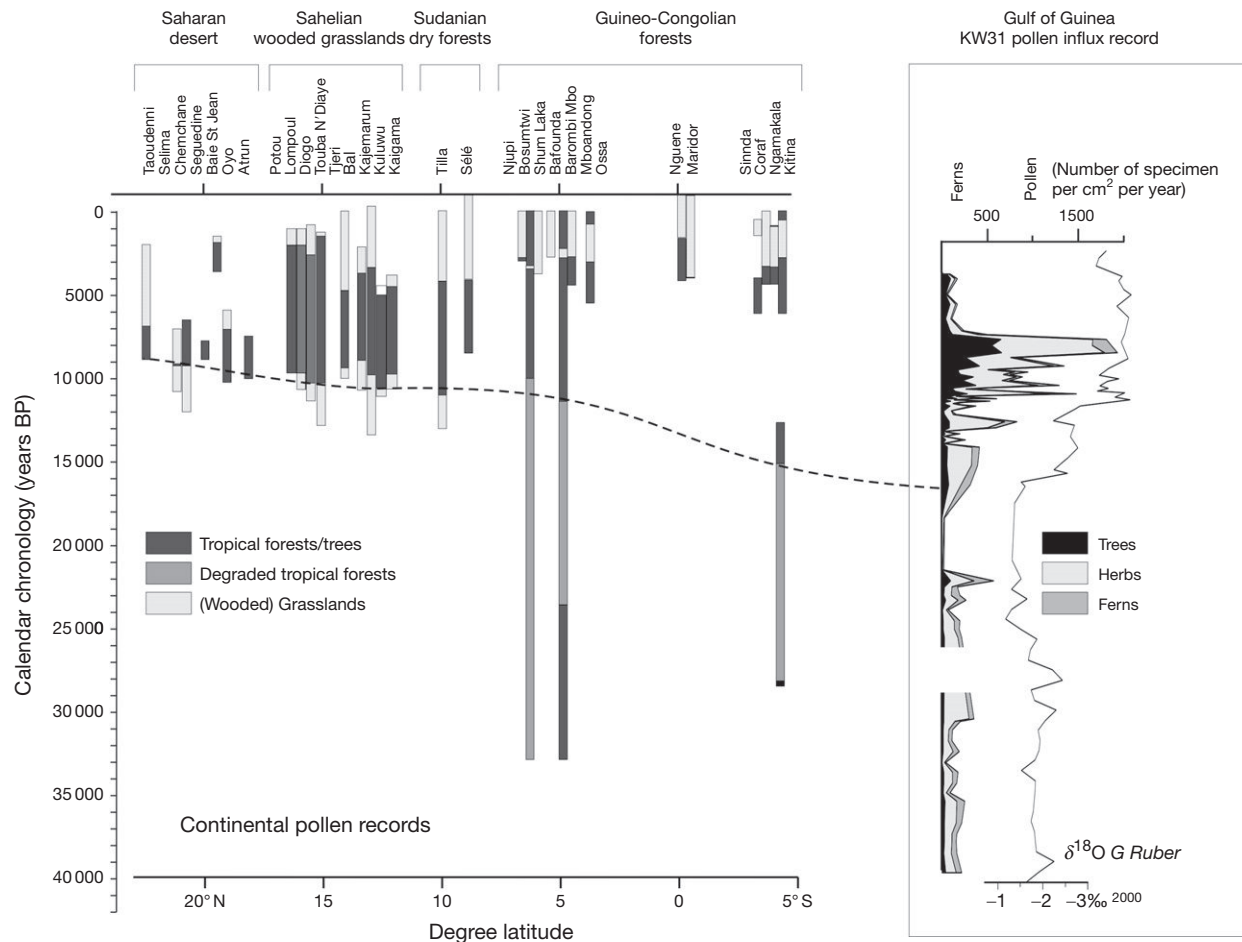


Figure 5 West African vegetation changes at the last glacial-interglacial transition. This figure shows the variations in pollen influx recorded in core KW31 recovered off the mouth of the Niger River in the Gulf of Guinea and the occurrence of tropical trees in the nearby continent from about 5° S to 25° N. Dashed line indicates the northward progression of tropical trees, which started about 14 500 cal yr BP near the equator and reached the northern tropic about 9 000–8 500 cal yr BP. Reproduced from [Lézine A-M and Cazet J-P \(2005\)](#) High resolution pollen record from core KW31, Gulf of Guinea, documents the history of the lowland forests of West Equatorial Africa since 40,000 years. *Quaternary Research* 64: 432–443.

alternation of phases of intense flowing and abrupt interruption of the fresh water supply to the ocean. Two main short 'dry' events are successively recorded between 13 400 and 12 100 cal yr BP, then around 11 400–11 200 cal yr BP, that closely correlate with events recorded in the GRIP ice record ([Blunier et al., 1997](#)) or more widely throughout the globe (the Younger Dryas interval). During these intervals, the low values of *Pediastrum* and the correlative increasing values of *Botryococcus* or *Staurastrum* (both green algae) indicate that river flow was considerably reduced and replaced by surficial runoff. At the end of the African Humid Period, a shift to drier conditions developed progressively from 7 100 cal yr BP, and the modern Niger River outflow was definitively established at 3 500 cal yr BP. The terrigenous (eolian) record at ODP site 658 ([de Menocal et al., 2000](#)) strongly contrasts with the KW31 algae record by showing that, near the Tropic of Cancer, the onset and the end of the African Humid Period occurred very abruptly, within decades to centuries around 5 500 cal yr BP.

Vegetation Response to Enhanced Monsoon Circulation and Global Warming at the Onset of the Last Deglaciation

Extension of Tropical Forest Ecosystems in West Africa and the Greening of the Saharan Desert

KW31 pollen content confirms previous studies carried out at Bosumtwi (Ghana) and, to a lesser extent, at Barombi M'Bo (Cameroon), according to which forests were strongly reduced in area during the last glacial stage. Increased monsoon rains since 15 000 cal yr BP led to the wide expansion of forests throughout north tropical Africa. However, comparisons between ocean and continent pollen records ([Figure 5](#)) point to a noticeable time lag between the strengthening of monsoon fluxes in the Gulf of Guinea and the vegetation response on the nearby continent. The continental pollen records suggest that forests expanded in the Cameroon midlands at 11 600 cal

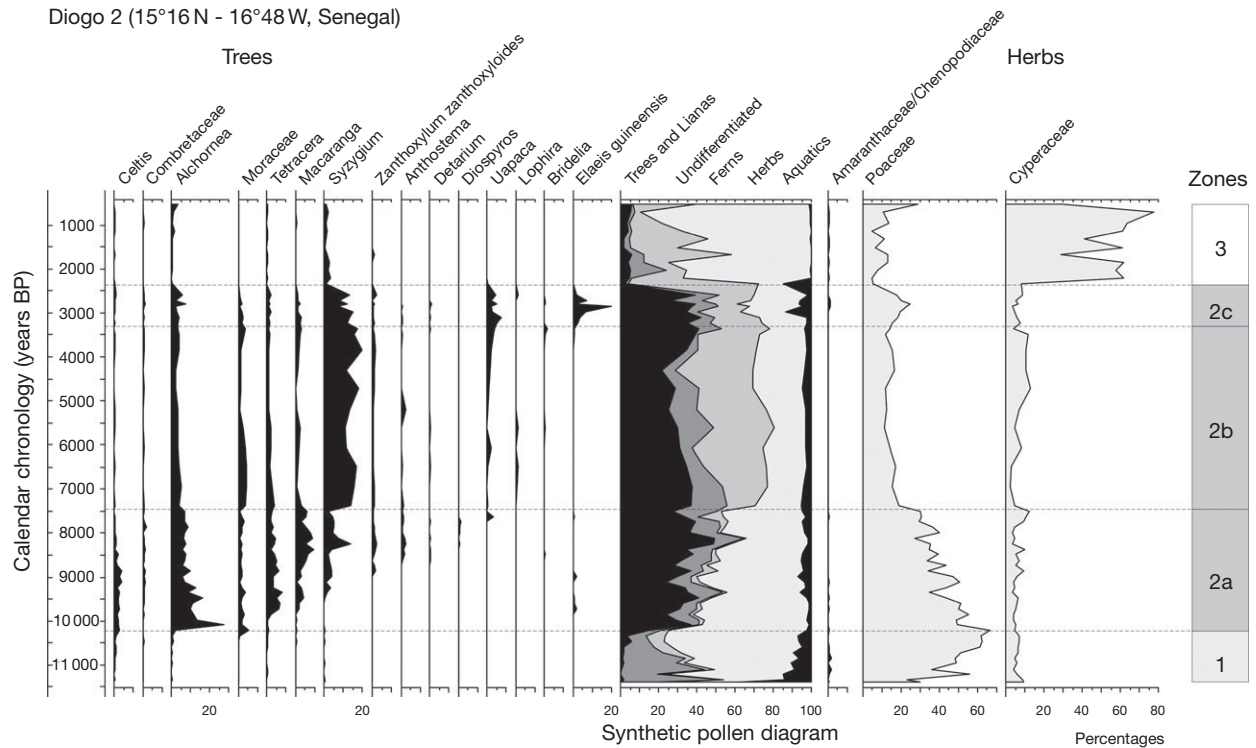


Figure 6 Diogo 2 (Senegal) summary pollen diagram showing percentages of the main pollen taxa. To the left, the main trees, to the right, the main herbaceous plants, and, in the middle, a synthetic diagram showing the variations of the main groups of taxa: trees, undifferentiated (trees, shrubs, herbs, or lianas), ferns, and upland and aquatic herbs. Percentages are calculated using a sum including all the determinable pollen and fern spores. (Data from the African Pollen Database). Reproduced from [Lézine A-M \(1998\)](#) Pollen records of past climate changes in West Africa since the Last Glacial Maximum. In: Issar A and Brown N (eds.) *Water, Environment and Society in Time of Climatic Change*, pp. 295–317. Dordrecht: Kluwer Academic Publishers.

yr BP, that is, more than 3 millennia after the start of the Niger River flow to the ocean and the extension of moist forests in southern Congo. Moist forests were established in the Sudanian zone at 11000 cal yr BP, then the western Guineo-Congolian forest domain, and the Sahel one millennium later, between 10500 and 9500 cal yr BP ([Figure 6](#)). Lastly, tropical trees entered the Saharan desert around 9500–9300 cal yr BP ([Figure 7](#)).

The Diogo interdunal depression ([Figure 6](#)) lies inland from the Atlantic coastal strand of Senegal north of the Cap Vert Peninsula at about 15°N ([Lézine, 1989](#)). It yielded a continuous pollen record for the interval lasting from 11500 cal yr BP to the present. From the beginning of the sequence to 10300 cal yr BP, the regional vegetation was dominated by herbaceous plant types, mainly Poaceae (grass), suggesting the prevalence of scrub- and grasslands (zone 1). The abrupt increase in tree pollen percentages by 10300 cal yr BP records the sudden expansion of tropical gallery forests since most of the taxa are of Guineo-Congolian and Sudanian phytogeographical affinity (*Celtis* (hackberry), *Tetracera alnifolia*, *Elaeis guineensis* (African oil palm), *Alchornea*, *Macaranga*, Moraceae (fig family), etc.) (zone 2a). Around 7500 cal yr BP, the most characteristic Guineo-Congolian forest trees disappeared and swamp forests dominated by *Syzygium* developed suggesting climate deterioration (zone 2b). From 3500 to 2500 cal yr BP, Guineo-Congolian trees (*Elaeis guineensis*, *Uapaca*, *Anthostema*

senegalense) went through a second phase of expansion recording increasing moisture conditions in the Niayes area (zone 2c). This wet phase abruptly ended at 2500 cal yr BP, leading to a rapid degradation of the forests and the setting of modern environmental conditions (zone 3).

The Oyo depression lies in one of the most arid area of NW Sudan at roughly 19°N. The pollen diagram ([Figure 7](#)) shows that tropical tree species of Sudano-Sahelian affinity (*Piliostigma*, *Grewia tenax*, *Lannea*, *Celtis*) occurred in the immediate surroundings of the lake from 9500 cal yr BP to the mid-Holocene and did not disappear before 5000 cal yr BP ([Ritchie, 1994](#)). At this date, the regional climate became drier and the vegetation was dominated by Sahelian *Acacia* (thorn tree) scrub- or grasslands. Desert conditions were established definitively around 5000 yr BP and the Oyo basin completely dried out.

From west to east, all the pollen diagrams from the southern fringe of the modern Saharan desert record a similar presence of pollen types from tropical-Sudanese (Sudano-Guinean) and Sahelian phytogeographical zones. More than 50 tropical tree pollen types have been enumerated from 11 pollen sites from 16°15' to 25°45'N ([Table 2](#)), which testify to the presence of the parent trees in the immediate surroundings of the lakes. However, their very low percentages indicate that tropical trees were scarce and most probably primarily restricted to gallery forests along freshwater bodies. Pollen

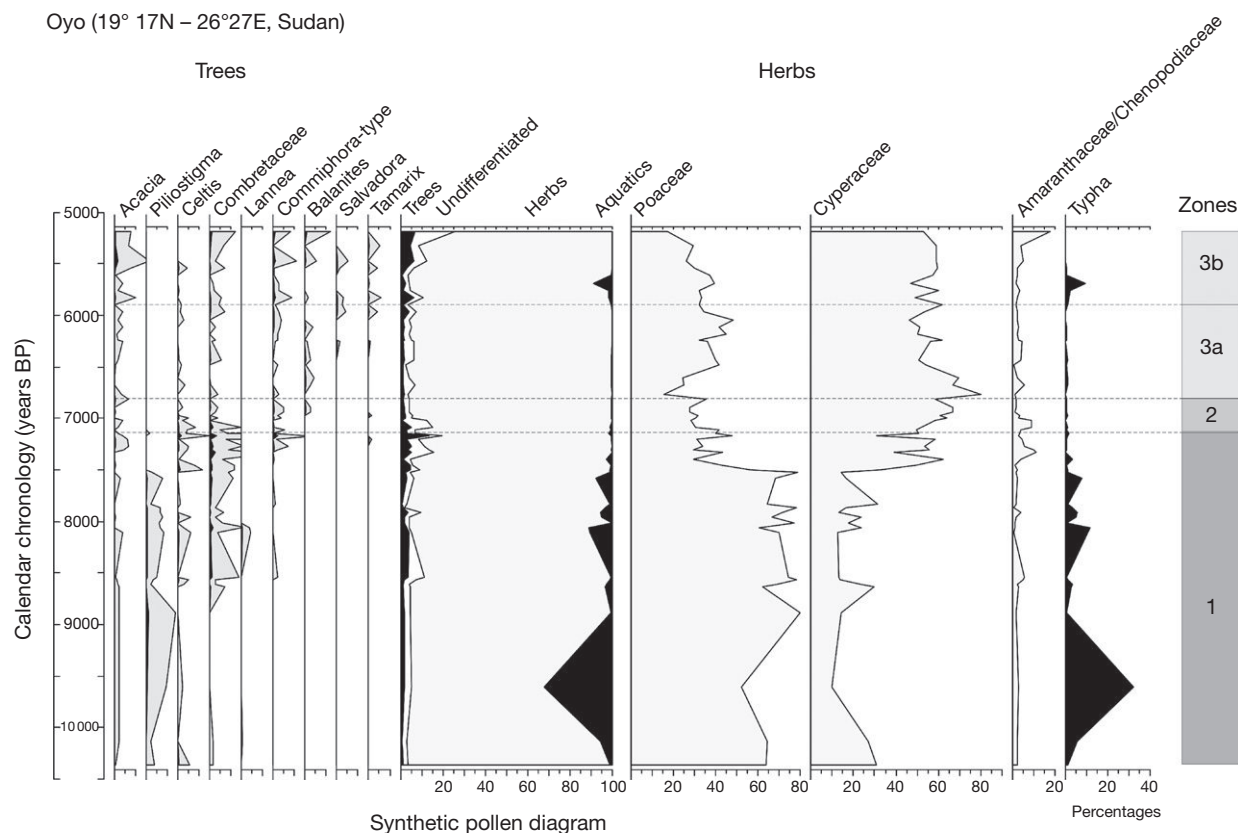


Figure 7 Oyo (Sudan) summary pollen diagram. See [Figure 6](#) for details. (Data from the African Pollen Database). Reproduced from [Ritchie JC \(1994\)](#) Holocene pollen spectra from Oyo, northwestern Sudan: Problems of interpretation in a hyperarid environment. *The Holocene* 4(1): 9–15.

spectra remained dominated by pollen from herbaceous plant types (Poaceae, Amaranthaceae/Chenopodiaceae (amaranth/goosefoot families)), while pollen grains from xerophytic plant communities that characterize the modern desert were also present (e.g., *Ziziphus mauritiana*-type, *Salvadora persica*-type, *Tamarix* (tamarisk), *Ephedra*, *Artemisia*). This indicates that plants that nowadays occupy distinct vegetation (eco-climatic) zones in North Tropical Africa (tropical-Sudanian, tropical-Saharan, Saharan desert) were found together, representing a complex, no-analogue situation ([Figure 8](#); [Watrin et al., 2009](#)). During the early to mid-Holocene, the vegetation of the Saharan desert considerably increased its diversity while accommodating humid adapted species from tropical forests and wooded grasslands.

The Retreat of ‘Cool’ Species from the Lowland Guineo–Congolian Forests

Pollen records of the last glacial stage document the general descent of upper treeline by 1000–1700 m on the tropical mountains. Ericaceous (heath) scrubs and grasslands widely expanded in the surroundings of most of the lakes or peat bogs of East Africa, in response to cool and dry climate conditions ([Farrera et al., 1999](#)) and/or substantial lowering of CO₂ concentration ([Figure 9](#); [Street-Perrott et al., 1997](#); [Jolly and Haxeltine, 1997](#); [Wu et al., 2007](#)). Forests were considerably

reduced in area in the highlands, whereas some of their components entered the lowland Guineo–Congolian forest as recorded in pollen records at lakes Bosumtwi (Ghana) and Barombi M’Bo (Cameroon).

The pollen diagram from Bois de Bilanko, Congo (3° 31’S, 15° 1’E; 600 m alt; [Figure 10](#)), shows that montane plant species (*Olea capensis* (black ironwood), *Podocarpus* spp. and *Ilex mitis* (African holly)) persisted in the lowlands of southern Congo until slightly after 12 000 cal yr BP. Related pollen percentages reach significant values of 5%, 35%, and 10%, respectively, attesting to the presence of the corresponding trees near the core site in association with elements of lowland tropical forests. Then the progressive climate warming led to the disappearance of these heat-intolerant plant species from the lowlands. The latest occurrence of *Podocarpus* and *Olea* (olive) pollen grains in the Atlantic central African pollen sites is recorded at Njupi, Cameroon, around 2500 cal yr BP.

The northern boundary of the Guineo–Congolian forest domain reveals a surprising phase of *Olea capensis* (East African olive) expansion at the beginning of the Holocene, ca. 10 500 cal yr BP. Noticeable pollen percentages of *Olea* are recorded at several lowland sites from 3° to 10°N: up to about 20% at Bosumtwi at 10 000 cal yr BP, that is, more than twice as much as the percentages reached during the glacial times; 14% at Tilla where the *Olea* phase spans from 10 630 to 9800 cal yr BP; 3% at Tjeri from 9500 to 8100 cal yr BP. Because pollen

Table 2

[illegible]

Table 2 (Continued)

<i>Phytogeographical affinity</i>	<i>Pollen types</i>		<i>Termit</i>	<i>El Atrun</i>	<i>Bilma</i>	<i>Oyo</i>	<i>Baie St Jean</i>	<i>Seguedine</i>	<i>Chemchane</i>	<i>Selima</i>	<i>Taoudenni</i>	<i>Acacus</i>	<i>Tibesti Mt</i>
	<i>Family</i>	<i>Genus and species</i>	<i>16° 15' N, 10° 57' E</i>	<i>18° 10' N, 26° 39' E</i>	<i>18° 41' N, 12° 55' E</i>	<i>19° 16' N, 26° 11' E</i>	<i>19° 28' N, 16° 18' W</i>	<i>20° 10' N, 12° 57' E</i>	<i>21° N, 12° W</i>	<i>21° 22' N, 29° 19' E</i>	<i>22° 39' N, 4° W</i>	<i>25° 45' N, 10° 35' E</i>	<i>21° 22' N, 4° E</i>
Tropical-Sudanian	Palmae	<i>Borassus/</i> <i>Hyphaene</i>	*	*	*	*		*			*	*	
Tropical-Sudanian	Palmae	<i>Phoenix</i>											*
Tropical-Sudanian	Rubiaceae	<i>Rubiaceae undiff.</i>							*				
Tropical-Sudanian	Rubiaceae	<i>Mitragyna</i> -type <i>inermis</i>	*	*		*					*		
Tropical-Sudanian	Rubiaceae	<i>Nauclea</i> -type			*								
Tropical-Sudanian	Rutaceae	<i>Zanthoxylum</i> -type							*				
Tropical-Sudanian	Sapindaceae	Sapindaceae <i>undiff.</i>	*		*						*		
Tropical-Sudanian	Sapotaceae	Sapotaceae <i>undiff.</i>									*		
Tropical-Sudanian	Sterculiaceae	<i>Sterculia</i> -type							*		*		
Tropical-Sudanian	Ulmaceae	<i>Celtis integrifolia</i> - type	*	*	*	*	*		*	*	*		

Gray shading indicates pollen types occurring in at least two pollen sites.

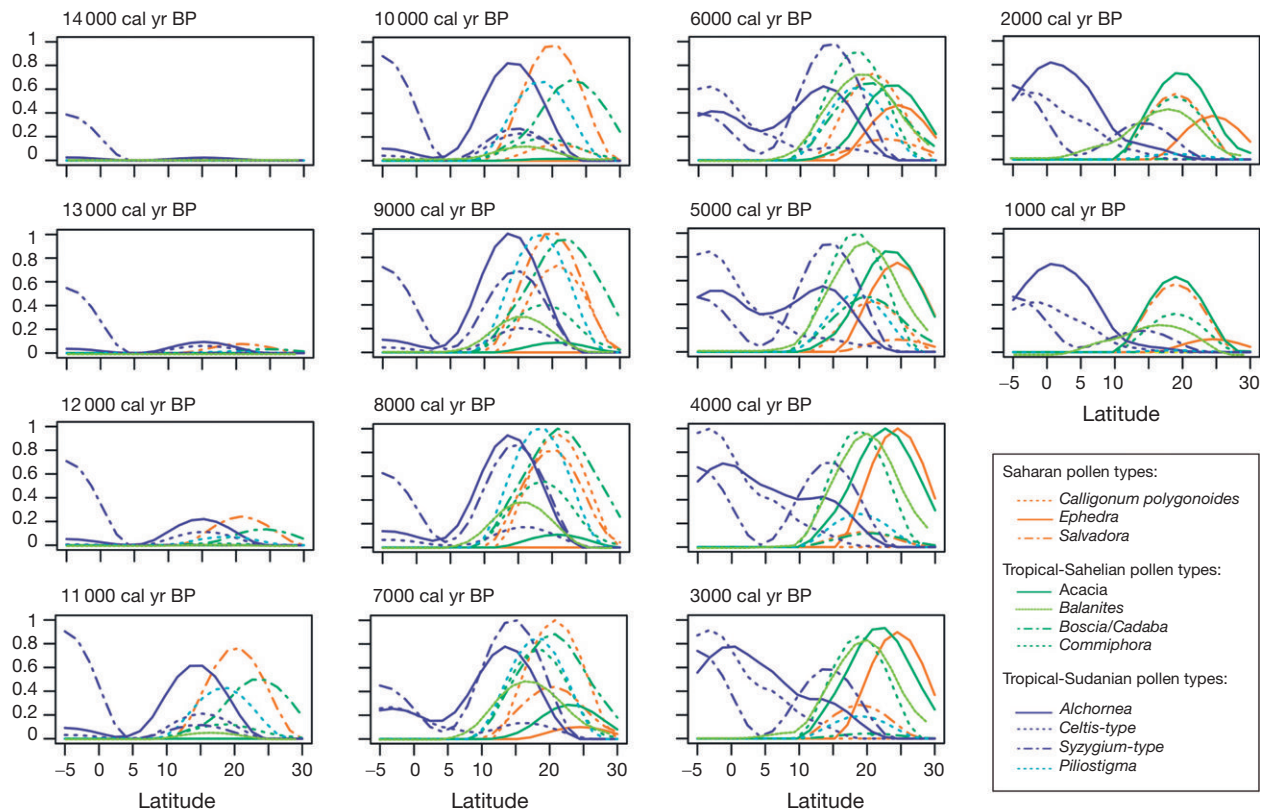


Figure 8 Latitudinal distribution of selected pollen taxa from 14 000 to 1 000 yr BP in western tropical Africa. Curves show probability densities calculated from pollen percentages (normalized for each taxon and varying from 0 to 1). This figure shows that, during the Holocene, plants that belong to distinct vegetation zones today occupied a common area at the time of the ‘green Sahara,’ about 9000 years BP. Reproduced from [Watrin J, Lézine A-M, and Hély C \(2009\) Plant migration and ecosystems at the time of the ‘green Sahara’. *CR Geosciences* 341: 656–670.](#)

percentages more than 10% are thought to indicate the presence of the parent plants (Coetzee, 1967), it is clear that *Olea* expanded in the immediate surroundings of Bosumtwi and Tilla and was also probably present in the southern Lake Chad basin. The *Olea* phase predates the expansion of the other pioneer forest types (such as *Alchornea*, *Macaranga*). Its significance (both climatic and biologic) is not yet fully understood. It is probable that *Olea capensis*, which is known to be a forest edge colonizing species (Kendall, 1969), benefited from the absence of competition from other trees and was able to extend its range during early postglacial times. It disappeared when the increase in temperature and rainfall was sufficient to permit the expansion of more heat- and moisture-tolerant pioneer trees.

The Upward Forests Expansion in Eastern Africa

Numerous pollen diagrams document the upward forest expansion in East Africa (Figure 9). However, most of the data comes from sedimentary sequences, which are discontinuous or poorly dated, thereby obviating any reliable reconstruction of vegetation dynamics at a regional scale. In addition, pollen sites have been obtained from a wide altitudinal gradient in various types of environments from lowland wooded grasslands to upland forests or alpine scrublands, from the edge of

the Guineo–Congolian forest massif at the west to the eastern side of the Rift valley. I present here three examples from continuous pollen records from the lowland dry forests surrounding Lake Tanganyika, the montane forest belt in Mount Kenya, and the upper limit of the forest in the Ruwenzori massif, in order to illustrate the postglacial forest history of East Africa.

Lake Tanganyika lies within the Zambezian Miombo woodland at 773 m altitude (Figure 11). The catchment basin extends over 231 000 km² to more than 3000 m altitude and is occupied by a wide range of vegetation associations including lowland dry woodlands and wooded grasslands, the so-called ‘Miombo,’ mainly composed of Caesalpiniaceae plant types (*Brachystegia*, *Julbernardia*, and *Isobertinia*), upland rainforests, and alpine scrub- and grasslands. Several cores from the northern (Sd14) and southern (MPU12) sub-basins were studied by Vincens (1989, 1991) and yielded detailed records of the last glacial–interglacial transition and the Holocene. Ericaceous plant communities, which widely expanded in the surroundings of the lake during the last glacial within degraded Miombo formations (zone 1), started to decrease by 17 600 cal yr BP and nearly disappeared from the lowlands by 14 000 cal yr BP in the southern basin and 12 000 cal yr BP in the northern basin (zone 2). During the interval from 17 600 to 12 000 cal yr BP, montane forest elements (*Olea* and *Myrica* (bayberry))

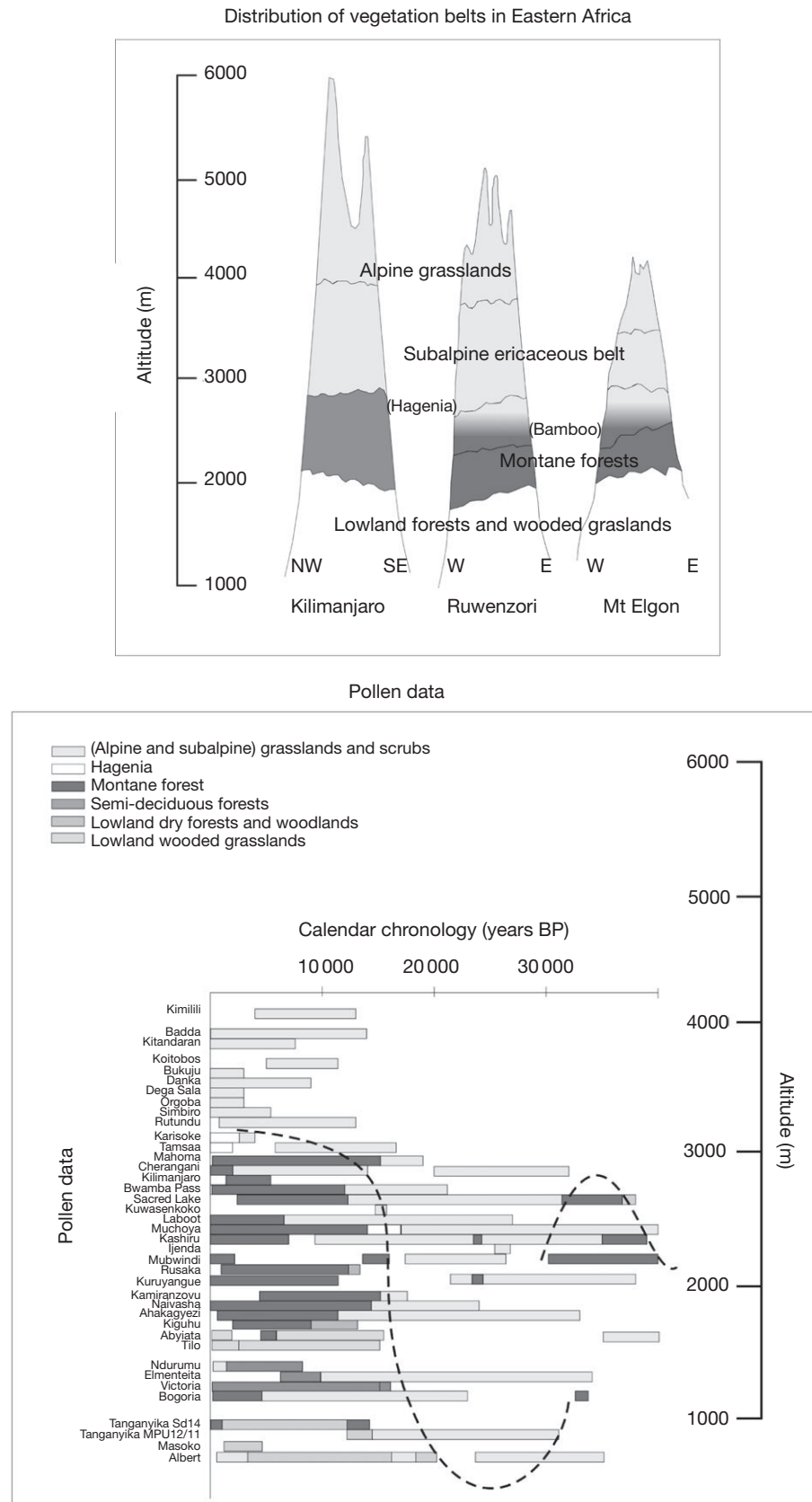


Figure 9 East African vegetation changes at the last glacial–interglacial transition. Top: the modern distribution of the main vegetation belts in the East African mountains (reproduced from [Hedberg 1951](#)) Vegetation belts of the East African Mountains. *Svensk Botanisk Tidskrift* 45(1): 141–202). Bottom: pollen sites plotted according to elevation from 750 (Lake Albert) to 4040 m (Kimilili). Forests disappeared from the African uplands during the last glacial interval except for a short phase centered around 35 000 cal yr BP. Trees started to expand from 15 000 cal yr BP.

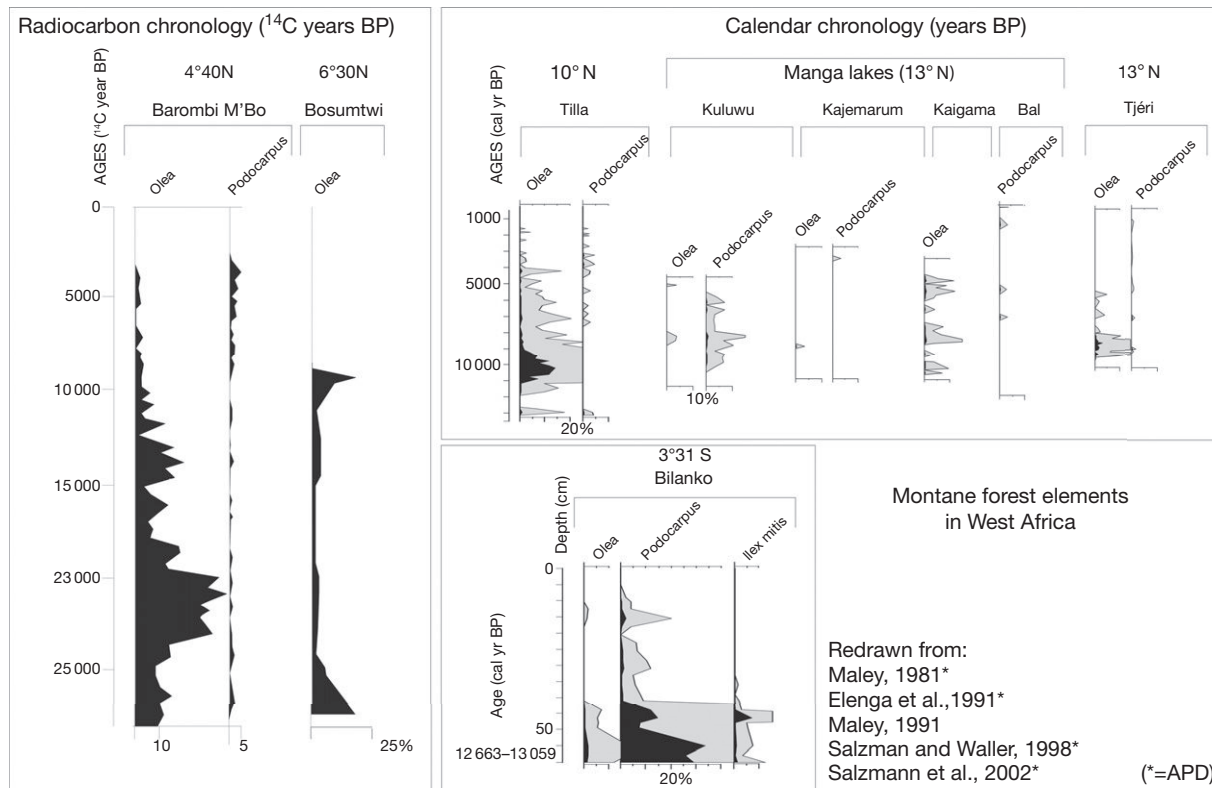


Figure 10 Occurrence of tree pollen types from 'cold' plant species in West African diagrams. Barombi M'Bo and Bosumtwi data are redrawn from Maley (1991). The other pollen data are from the African Pollen Database.

expanded in the nearby mountainous areas in response to the progressive climate amelioration. Miombo plant types, mainly *Brachystegia* and *Uapaca kirkiana*-type, started to spread during the same interval, while the Miombo flora greatly diversified after 12 000 cal yr BP and included more wet tolerant plant types with Guineo–Congolian affinity, such as *Tetrorchidium*, *Holoptelea grandis*, and *Pycnanthus angolensis* (zone 3). This clearly indicates the onset of moist climate conditions near Lake Tanganyika. The late Holocene interval is poorly documented at Lake Tanganyika. However, the uppermost levels of core Sd14 suggest the degradation of lowland vegetation types during the last two millennia and the establishment of *Podocarpus* populations in the uplands.

Sacred Lake is a crater lake situated at 2400 m altitude (Figure 12). It yielded one of the most famous and longer pollen sequences from East Africa, spanning from about 38 500 cal yr BP to the present (Coetzee, 1967). The lake is situated in the montane rainforest zone, in the humid forest composed of *Pygeum africanum*, *Polyscias kikuyensis*, *Macaranga kilimandscharika*, *Neoboutonia macrocalyx*, *Apodytes dimidiata*, *Cassipourea malosana*, *Afrocrania volkensii*, *Syzygium guineense*, and so on. From the base to 12 200 cal yr BP, the pollen assemblages are dominated by herbaceous types, mainly Poaceae and *Artemisia*, which attest to dry and cold conditions during the glacial interval. It is probable, that, except for a short episode spanning from 36 000 to 30 000 cal yr BP during which forest types occur, the landscape around Sacred Lake remained open and similar to that of the modern ericaceous belt. From 16 500 cal yr BP (zone 2), the progressive increase

of *Hagenia abyssinica* suggests that the treeline started to move upward. It reached the core site around 12 200 cal yr BP when *Hagenia* was associated with *Myrica*. *Hagenia* and *Myrica* are heliophilous colonizing taxa nowadays found in the ericaceous belt or at the upper limit of the forest. Their occurrence could account for an early stage of forest development at the core site. Then the disappearance of elements from the ericaceous belt and the correlative increase of *Afrocrania*, *Olea*, *Rapanea*, *Macaranga*, *Dombeya*, and *Pygeum* record the development of the humid montane rainforest in response to warmer and moister climate conditions (zone 3). Finally, toward the top of the sequence after about 3500 cal yr BP, the dramatic increase in *Podocarpus* pollen and the decrease of pollen from humid plant types indicates the replacement of the humid forest by more dry forests around the lake (zone 4).

Mahoma is a glacial lake situated at 2960 m altitude in the Ruwenzori massif (Figure 13). It yielded a 9.5 m long core studied by Livingstone in 1967. The lake lies at the boundary between the montane forest belt and the ericaceous zone. Three main phases of vegetation development can be distinguished. At the base of the sequence, from 22 000 to 15 000 cal yr BP, the vegetation surrounding Mahoma Lake was dominated by plants from alpine grass- or scrublands mainly composed of Poaceae, *Dendrosenecio*, Ericaceae, and *Artemisia*. Of interest is the large representation of small trees and shrubs, including *Myrica*, *Hagenia abyssinica* and *Anthospermum*, which attest, as at Sacred Lake at approximately the same interval, to an early stage of plant

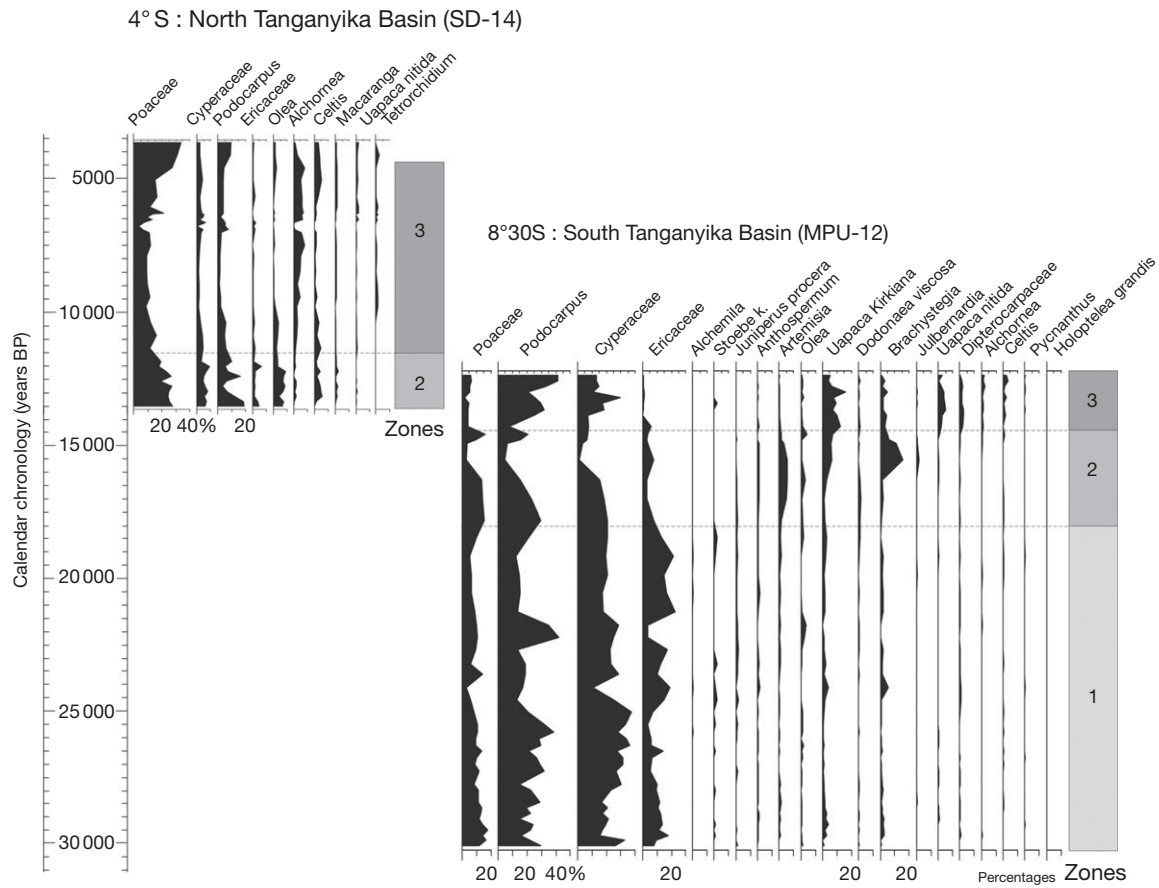


Figure 11 Summary pollen diagrams from Lake Tanganyika showing the last glacial–interglacial transition (core MPU-12 from the Southern Basin) and the Holocene (core Sd14 from the Northern Basin) (Data from the African Pollen Database). Reproduced from Vincens A (1989) *Paléoenvironnements du bassin North-Tanganyika (Zaire, Burundi, Tanzanie) au cours des 13 derniers mille ans: apport de la palynologie. Review of Palaeobotany and Palynology* 61: 69–88; Vincens A (1991) *Late Quaternary vegetation history of the South-Tanganyika basin. Climatic implications in South Central Africa. Palaeogeography, Palaeoclimatology, Palaeoecology* 86: 207–226.

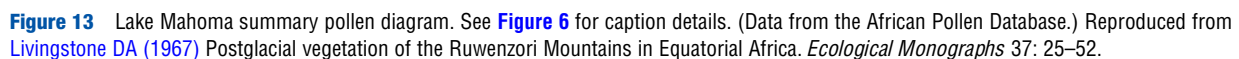
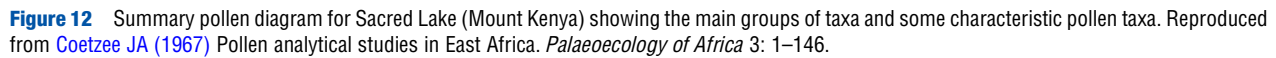
colonization after the retreat of glaciers. The pollen from ‘alpine’ taxa markedly decreased after 15000 cal yr BP and definitively disappeared from this record around 10000 cal yr BP. They were replaced by pollen from forest plant types such as *Olea*, *Celtis*, Myrsinaceae (mainly *Rapanea*), and *Acalypha*. This pollen assemblage does not correspond to any modern analogue. Most of the taxa are elements of the present day montane forest or are found at the boundary between the forest and the ericaceous belt, whereas *Celtis* and *Acalypha* belong to vegetation associations nowadays found downslope from the montane forest. Lastly, the last four millennia record the decrease of *Celtis* and the increase of *Podocarpus* in response to probable climate deterioration. The high proportions of Poaceae pollen in the upper levels of the Mahoma pollen sequence, moreover, account for increased anthropogenic pressure and correlative forest clearance in the Ruwenzori Mountains during the last millennium.

The End of the African Humid Period

The Collapse of the Tropical and Equatorial Forests

Considerable changes in vegetation have occurred in tropical Africa during the last few thousands of years at the end of the

African Humid Period (Vincens et al., 1999). In the lowlands of the equatorial forest domain (Figure 14), the individual response of the environment at each pollen site was clearly dependent upon local hydrogeological conditions. The replacement of forests by more open landscapes, wooded grasslands, grasslands, or woodlands occurred within an interval lasting from 4500 to 1300 cal yr BP. Some lakes dried out, such as Sinnda, while other sites remained forested (Bosumtwi, Barombi M’Bo) or were only partially affected, with rain or swamp forests being replaced by more open formations with increasing importance of light-dependent trees. To the north, the development of the modern ‘Dry Dahomey Gap,’ which interrupts the Guineo–Congolian forest domain, dates from 4000 cal yr BP (Lake Sélé; Salzmann and Hoelzmann, 2005), whereas the disappearance of the gallery forests from the Manga area in northern Nigeria took place at a date varying from roughly 4500 and 3500 cal yr BP and from the western Sahel only at around 2500 cal BP (Diogo 2, Figure 6). In this latter area, specific hydrological conditions linked to the proximity of the sea partly explain the persistence of dense forest communities more than 2000 years after the onset of dry climate condition over the Sahel. New high-resolution pollen data from Cameroon (Lake Mbalang (Vincens et al., 2010); Lake Bambili (Assi-Kaudjhis et al., 2008; Lézine et al., in preparation)) confirm



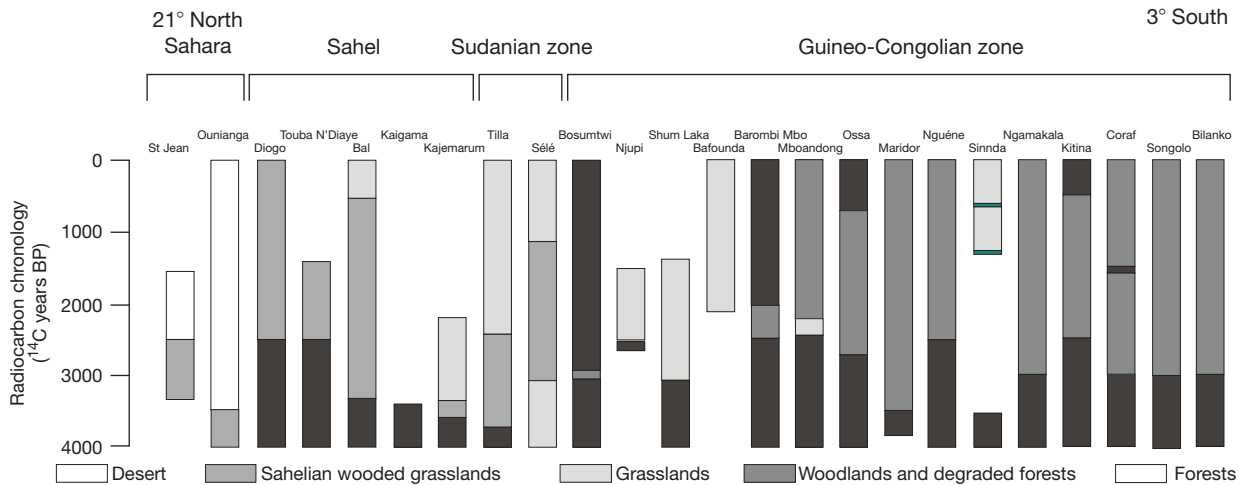


Figure 14 The end of the African Humid Period in western Africa. The vegetation response to the 4000 yr BP dry event varied considerably according to the local hydrological settings and occurred within a span of 2500 years. Reproduced from Vincens A, Schwartz D, Elenga H, et al. (1999) Forest response to climate changes in Atlantic Equatorial Africa during the last 4000 yr BP and inheritance on the modern landscapes. *Journal of Biogeography* 26: 879–885.

the ‘threshold’ effect in most of the forest response to the climate deterioration at the end of the African Humid Period. The dramatic switch from forest to grassland communities occurred ca. 3000 cal yr BP at both sites after a long period of progressive forest destabilization starting in the mid-Holocene in response to superimposed climate fluctuations. These included a long-term gradual decrease of summer insolation and the short-term variability likely linked to El Niño/Southern Oscillation and sea surface temperature changes in the tropical Atlantic (Jury et al., 2002; Moy et al., 2002).

The Retreat of Tropical Trees from the Saharan Desert

The discovery by Kröpelin et al. (2008) of a groundwater-fed lake in the hyperarid desert of northern Chad was important because the lake yielded a unique continuous sedimentary sequence of late Holocene age for the entire Saharan desert. It significantly improves our knowledge of Saharan paleoecology, particularly with respect to the adaptation of plants to the shift from moist to arid conditions and the setting of the modern Saharan desert. The Lake Yoa pollen record (19°02'N, 20°19'E; Figure 15) shows the progressive retreat of tropical plant communities, which were widely spread throughout North Africa (Watrén et al., 2009) during the African Humid Period, and their replacement by herbaceous-dominated desert ecosystems. The occurrence between 6000 and 4750 cal yr BP of pollen from numerous tropical plants together with ferns and other indicators of surface runoff indicates a dominant monsoonal climatic regime. Their low percentages, however, show that dense tropical forest formations never existed at the time. Trees likely occurred as gallery forest in an open landscape, characterized by extended herbaceous plant communities dominated by grasses and containing numerous Saharan and Sahelian elements.

From 4750 to 2700 cal yr BP, a transition zone, characterized by the disappearance of most of the tropical humid indicators and their replacement by Sahelian (Sudano-Sahelian, then

Sahelo-Saharan) elements including *Tribulus*, *Boerhavia*, and *Blepharis*, which are therophytes (annual plants) with a very short life cycle (Gillet, 1968) and thus well adapted to long rainless periods, confirms the increasing intensity of dry environmental conditions. Then, after 2700 cal yr BP, the increase in psammophilous (sand-loving) plant types (mainly *Cornulaca*, *Ephedra*; White, 1983) confirms the local expansion sand dunes around Lake Yoa and the establishment of modern climate conditions characterized by the complete reversal of prevailing winds, which are predominantly in the NE–SW direction.

Concluding Remarks

This review of postglacial pollen records from the African tropics highlights two major observations:

- (1) Pollen data reported here show that the African ecosystems were highly sensitive to climate change. The impact of climate changes was not only the upslope and down slope displacements of existing vegetation belts in the eastern highlands or the northward and southward displacements of existing phytogeographical zones in the western lowlands. As observed by Livingstone (1967, p. 50), “...taxa appear to have behaved individually rather than as members of a migrating community.” As a consequence, the biodiversity of the African ecosystems was considerably more diverse during the last few thousands years, during the last glacial stages, and during the Holocene than at present.
- (2) The presence of available fresh water in interdunal depressions and the reactivation of extensive fluvial systems during the early Holocene, which was linked to enhanced monsoonal rainfall, allowed tropical trees to migrate north and to persist for several thousand years in the Saharan desert. The role of local hydrological conditions has to be carefully taken into account when interpreting pollen data in terms of climate parameters.

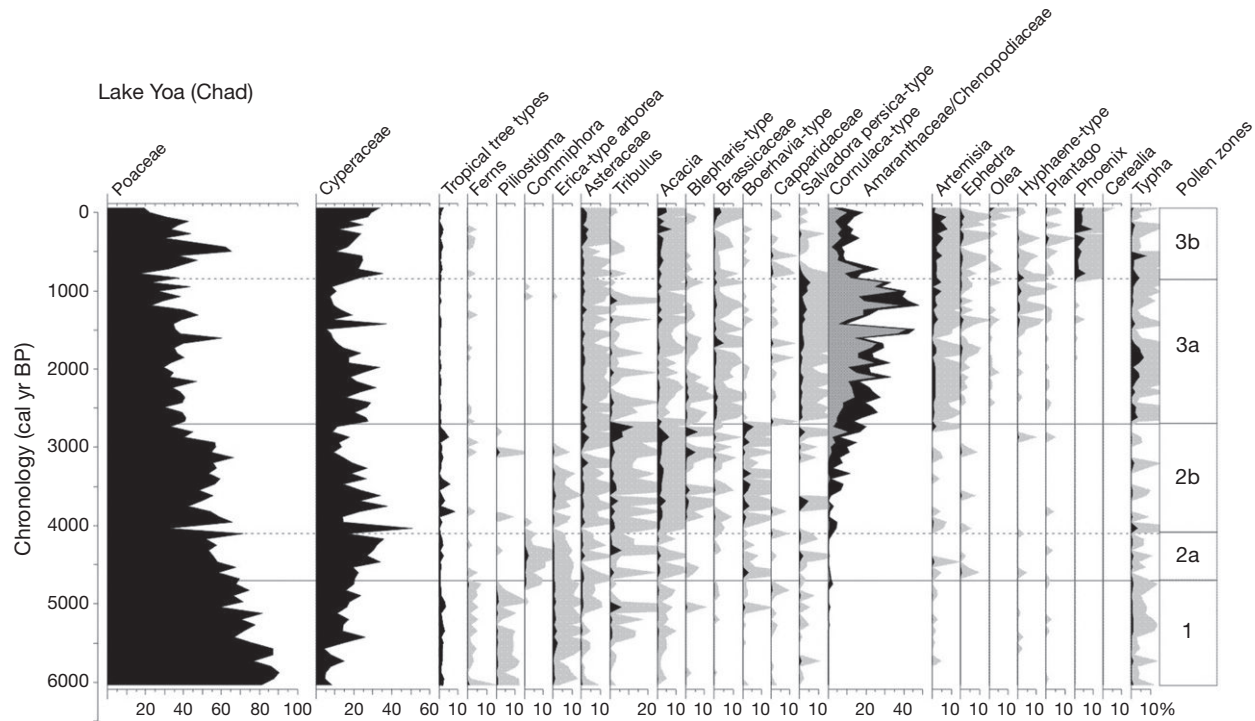


Figure 15 Lake Yoa summary pollen diagram (Ounianga lakes, Northern Chad) showing percentages of main pollen types. Grey shading indicates 10 × magnification. Reproduced from Lézine A-M (2009) Timing of vegetation changes at the end of the Holocene Humid Period at the northern edge of the Atlantic and Indian monsoon systems. *CR Geosciences* 341: 750–759.

Acknowledgments

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See also: **Pollen Methods and Studies: Stand-Scale Palynology. Pollen Records, Late Pleistocene: Africa. Glacial Climates: Biosphere Feedbacks.**

References

- Assi-Kaudjhis C, Lézine A-M, and Roche E (2008) Contribution à l'étude de la végétation d'altitude (Afrique centrale atlantique) à partir des analyses préliminaires de la carotte de Bambili (Nord-ouest du Cameroun), depuis 17 000 ans BP. *ECO-GEO-TROP* 32: 131–143.
- Baumhauer R and Schulz D (1984) The Holocene lake of Seguedine, Kaouar, NE Niger. *Palaeoecology of Africa* 16: 283–290.
- Beuning KRM (1999) A re-evaluation of the Late Glacial and early-Holocene vegetation history of the Lake Victoria region. *Palaeoecology of Africa* 26: 115–136.
- Beuning KRM, Talbot MR, and Kelts K (1997) A revised 30,000-year paleoclimatic and paleohydrologic history of Lake Albert, East Africa. *Palaeogeography, Palaeoclimatology, Palaeoecology* 136: 259–279.
- Blunier T and Brook E (2001) Timing of millennial-scale climate change in Antarctica and Greenland during the last glacial period. *Science* 291: 109–112.
- Blunier T, Schwander J, Stauffer B, et al. (1997) Timing of the Antarctic Cold Reversal and the atmospheric CO₂ increase with respect to the Younger Dryas event. *Geophysical Research Letters* 24: 2683–2686.
- Bonnefille R and Chalié F (2000) Pollen-inferred precipitation time-series from equatorial mountains, Africa, the last 40 kyr BP. *Global and Planetary Change* 26: 25–50.
- Bonnefille R and Mohammed U (1994) Pollen-inferred climatic fluctuations in Ethiopia during the last 3000 years. *Palaeogeography, Palaeoclimatology, Palaeoecology* 109: 331–343.
- Bonnefille R, Rioulet G, and Buchet G (1991) Nouvelle séquence pollinique d'une tourbière de la crête Zaïre-Nil (Burundi). *Review of Palaeobotany and Palynology* 67: 315–330.
- Bonnefille R, Rioulet G, Buchet G, et al. (1995) Glacial/Interglacial record from Intertropical Africa, high resolution pollen and carbon data at Rusaka, Burundi. *Quaternary Science Reviews* 14: 917–936.
- Broström A, Coe M, Harrison SP, et al. (1998) Land surface feedbacks and paleomonsoons in Northern Africa. *Geophysical Research Letters* 25: 3615–3618.
- Coetzee JA (1967) Pollen analytical studies in East Africa. *Palaeoecology of Africa* 3: 1–146.
- De Menocal P, Ortiz J, Guilderson T, et al. (2000) Abrupt onset and termination of the African Humid Period: Rapid climate responses to gradual insolation forcing. *Quaternary Science Reviews* 19: 347–361.
- El Moutaki S (1994) *Transition Glaciaire-Interglaciaire et Younger Dryas dans l'hémisphère Sud (1°–20° Sud): analyse palynologique à haute résolution de sondages marin et continentaux (lac et marécage)*. PhD Thesis, University of Aix-Marseille III.
- Elenga H, Schwartz D, and Vincens A (1994) Pollen evidence of late Quaternary vegetation and inferred climate changes in Congo. *Palaeogeography, Palaeoclimatology, Palaeoecology* 109: 345–356.
- Elenga H, Schwartz D, Vincens A, et al. (1996) Diagramme pollinique Holocène du lac Kitina (Congo): mise en évidence de changements paléobotaniques et paléoclimatiques dans le massif forestier du Mayombe. *Comptes-rendus de l'Académie des Sciences Paris* 323: 403–410.
- Elenga H and Vincens A (1990) Paléoenvironnements quaternaires récents des plateaux Batéké (Congo): étude palynologique des dépôts de la dépression du bois de Bilanko. In: Lanfranchi R and Schwartz D (eds.) *Paysages quaternaires de l'Afrique Centrale Atlantique*, pp. 271–282. Paris: Orstom.

- Elenga H, Vincens A, and Schwartz D (1991) Présence d'éléments forestiers montagnards sur les Plateaux Batéké (Congo) au Pléistocène supérieur: nouvelles données palynologiques. *Palaeoecology of Africa* 21: 239–252.
- Elenga H, Vincens A, Schwartz D, et al. (2001) Le marais estuarien de la Songolo (Sud Congo) à l'Holocène moyen et récent. *Bulletin de la Société Géologique de France* 172: 359–366.
- Farrera I, Harrison SP, Prentice IC, et al. (1999) Tropical climates at the last glacial maximum: A new synthesis of terrestrial palaeoclimate data. I. Vegetation, lake-levels and geochemistry. *Climate Dynamics* 15: 823–856.
- Gasse F (2000) Hydrological changes in the African tropics since the Last Glacial Maximum. *Quaternary Science Reviews* 19: 189–211.
- Gillet H (1968) *Le peuplement végétal du massif de l'Ennedi (Tchad)*. University of Paris Thesis.
- Hamilton A (1982) *Environmental History of East Africa – A study of the Quaternary*. London: Academic Press.
- Hedberg O (1951) Vegetation belts of the East African Mountains. *Svensk Botanisk Tidskrift* 45(1): 141–202.
- Hoelzmann P, Jolly D, Harrison SP, Laarif F, Bonnefille R, and Pachur H-J (1998) Mid-Holocene land-surface conditions in northern Africa and the Arabian peninsula: A data set for analysis of biogeophysical feedbacks in the climate system. *Global Biogeochemical Cycles* 12(1): 35–51.
- Jolly D and Bonnefille R (1992) Histoire et dynamique du marécage tropical de Ndurumu (Burundi), données polliniques. *Review of Palaeobotany and Palynology* 75: 133–151.
- Jolly D and Haxeltine A (1997) Effect of low glacial atmospheric CO₂ on tropical African montane vegetation. *Science* 276: 786–788.
- Jury MR, Enfield DB, and Mélice JL (2002) Tropical monsoons around Africa: Stability of El Niño–Southern Oscillation associations and links with continental climate. *Journal of Geophysical Research* 107(C10): 3151. <http://dx.doi.org/10.1029/2000JC000507>.
- Kadamura H and Kiyonaga J (1994) Origin of grassfields landscape in the West Cameroon Highlands. In: Kadamura H (ed.) *Savannization Processes in Tropical Africa II*, pp. 47–85. Japan: Tokyo Metropolitan University.
- Kendall R (1969) An ecological history of the Lake Victoria basin. *Ecological Monographs* 39(2): 121–176.
- Kröpelin S, Verschuren D, Lézine A-M, et al. (2008) Climate-driven ecosystem succession in the Sahara: The last 6000 years. *Science* 320: 765–768.
- Kutzbach JE and Street-Perrott FA (1985) Milankovitch forcing of fluctuations in the level of tropical lakes from 18 to 0 kyr BP. *Nature* 317: 130–134.
- Lamb HF (2001) Multi-proxy records of Holocene climate and vegetation change from Ethiopian crater lakes. *Proceedings of the Royal Irish Academy* 101B: 35–46.
- Lézine A-M (1988) Les variations de la couverture forestière mésophile d'Afrique occidentale au cours de l'Holocène. *Comptes-rendus de l'Académie des Sciences Paris* 307: 439–445.
- Lézine A-M (1989) Late Quaternary vegetation and climate of the Sahel. *Quaternary Research* 32: 317–334.
- Lézine A-M (1998) Pollen records of past climate changes in West Africa since the Last Glacial Maximum. In: Issar A and Brown N (eds.) *Water, Environment and Society in Time of Climatic Change*, pp. 295–317. Dordrecht: Kluwer Academic.
- Lézine A-M (2009) Timing of vegetation changes at the end of the Holocene Humid Period at the Northern edge of the Atlantic and Indian monsoon systems. *CR Geosciences* 341: 750–759.
- Lézine A-M, Assi-Kaudjhis C, Roche E, Vincens A, and Achoundong G (in press) Toward an understanding of the response of west African montane forests to climate change.
- Lézine A-M and Bonnefille R (1982) Diagramme pollinique Holocène d'un sondage du Lac Abiyata (Ethiopie, 7°42' Nord). *Pollen et Spores* 24: 463–480.
- Lézine A-M, Casanova J, and Hillaire-Marce CI (1990) Across an early Holocene humid phase in western Sahara: Pollen and isotope stratigraphy. *Geology* 18: 264–267.
- Lézine A-M and Cazet J-P (2005) High resolution pollen record from core KW31, Gulf of Guinea, documents the history of the lowland forests of West Equatorial Africa since 40,000 years. *Quaternary Research* 64: 432–443.
- Lézine A-M, Duplessy J-C, and Cazet J-P (2005) West African monsoon variability during the last deglaciation and the Holocene: Evidence from Fresh Water Algae, Pollen and isotope data from Core KW31, Gulf of Guinea. *Palaeogeography, Palaeoclimatology, Palaeoecology* 219(3–4): 225–237.
- Lézine A-M, Hély C, Grenier C, Braconnot P, and Krinner G (2011) Sahara and Sahel vulnerability to climate changes, lessons from Holocene hydrological data. *Quaternary Science Reviews* 30(21–22): 3001–3012.
- Lézine A-M, Tastet JP, and Leroux M (1994) Evidence of atmospheric paleocirculation over the Gulf of Guinea since the Last Glacial Maximum. *Quaternary Research* 41: 390–395.
- Lézine AM, Turon JL, and Buchet G (1995) Pollen analyses off Senegal: Evolution of the coastal palaeoenvironment during the last deglaciation. *Journal of Quaternary Science* 10: 95–105.
- Livingstone DA (1967) Postglacial vegetation of the Ruwenzori Mountains in Equatorial Africa. *Ecological Monographs* 37: 25–52.
- Maley J (1981) *Etudes palynologiques dans le bassin du Tchad et paléoclimatologie de l'Afrique nord-tropicale de 30 000 ans à l'époque actuelle*. Paris: Orstom.
- Maley J (1991) The African rain forest vegetation and palaeoenvironments during Late Quaternary. *Climatic Change* 19: 79–98.
- Maley J and Brenac P (1998) Vegetation dynamics, palaeoenvironments and climatic change in the forests of western Cameroon during the last 28,000 years BP. *Review of Palaeobotany and Palynology* 99: 157–187.
- Maley J and Livingstone DA (1983) Extension d'un élément montagnard dans le sud du Ghana (Afrique de l'Ouest) au Pléistocène supérieur et à l'Holocène inférieur: premières données polliniques. *Comptes-rendus de l'Académie des Sciences Paris* 296: 1287–1292.
- Marchant R, Taylor D, and Hamilton A (1997) Late Pleistocene and Holocene history at Mubwindi Swamp, Southwest Uganda. *Quaternary Research* 47: 316–328.
- Mohammed MU (1992) *Paléoenvironnement et paléoclimatologie des derniers millénaires en Ethiopie. Contribution palynologique*. PhD Thesis, Marseille, France: University of Aix-Marseille III.
- Mohammed MU and Bonnefille R (1998) A late Glacial / late Holocene pollen record from a highland peat at Tamsaa, Bale Mountains, south Ethiopia. *Global and Planetary Change* 16–17: 121–129.
- Moy CM, Seltzer GO, Rodbel DT, and Anderson DM (2002) Variability of El Niño/Southern Oscillation activity at millennial timescales during the Holocene epoch. *Nature* 420: 162–165.
- Mumbi CT, Marchant R, Hooghiemstra H, and Wooller MJ (2008) Late Quaternary vegetation reconstruction from the Eastern Arc Mountains, Tanzania. *Quaternary Research* 69(2): 326–341.
- Mworia-Maitima J (1991) Vegetation response to climatic change in Central Rift Valley, Kenya. *Quaternary Research* 35: 234–245.
- Mworia-Maitima J (1999) Late Quaternary pollen record from lake Elmenteita. *Palaeoecology of Africa* 26: 103–114.
- Nakimera I (2001) *The Impact of Human Activities and Climate on the Vegetation in the Lake Victoria Region and on the Rwenzori Mountain and Its Neighbourhood*. PhD Thesis, Kampala, Uganda: Makerere University.
- Ngomanda A (2005) *Dynamique des écosystèmes forestiers du Gabon au cours des cinq derniers millénaires*. PhD Thesis, France: University of Montpellier.
- Ngomanda A, Chepstow-Lusty A, Makaya M, et al. (2005) Vegetation changes during the past 1300 years in western equatorial Africa: A high resolution pollen record from Lake Kamalété, Lopé Reserve, Central Gabon. *The Holocene* 15(7): 1021–1031.
- Ngomanda A, Jolly D, Bentaleb I, et al. (2007) Lowland rainforest response to hydrological changes during the last 1500 years in Gabon, Western Equatorial Africa. *Quaternary Research* 67: 3,411–425.
- Ngomanda A, Neumann K, Schweizer A, and Maley J (2009) Seasonality change and the third millennium BP rainforest crisis in southern Cameroon (Central Africa). *Quaternary Research* 71(3): 307–318.
- Ntaganda C (1991) *Paléoenvironnements et paléoclimats du Quaternaire supérieur au Rwanda par l'analyse palynologique des dépôts superficiels*. PhD Thesis, Belgium: University of Liège.
- Petit-Maire N and Bouysse P (coordinators) (1999) *Maps of the World Environments during the Last two Climatic Extremes*. Commission for the Geological Map of the World and Agence nationale pour la Gestion des Déchets radioactifs, Paris. Two maps and notes.
- Reynaud-Farrera I, Maley J, and Wirrmann D (1996) Végétation et climat dans les forêts du Sud-Ouest Cameroun depuis 4 770 ans BP: Analyse pollinique des sédiments du Lac Ossa. *Comptes-rendus de l'Académie des Sciences Paris* 322: 749–755.
- Richards K (1986) Preliminary results of pollen analysis of a 6000 year core from Mboandong, a crater lake in Cameroon. In: Baker RGE, Richards K, and Rimes CA (eds.) *The Hull University, Cameroon Expedition 1981–1982*, pp. 14–28. Department of Geography, University of Hull Unpublished Final Report.
- Ritchie JC (1994) Holocene pollen spectra from Oyo, northwestern Sudan: Problems of interpretation in a hyperarid environment. *The Holocene* 4(1): 9–15.
- Ritchie JC and Haynes CV (1987) Holocene vegetation zonation in the eastern Sahara. *Nature* 330: 645–647.
- Roche E and Bikwemu G (1989) Paleoenvironmental change on the Zaire-Nile ridge in Burundi; the last 20,000 years: An interpretation of palynological data from the

- Kashiru Core, Ijenda, Burundi. In: Mahaney WC (ed.) *Quaternary and Environmental Research on East African Mountains*, pp. 231–242. Rotterdam: Balkema.
- Roche E and Ntaganda C (1999) Analyse palynologique de la séquence sédimentaire Kiguhu II (région des Birunga, Rwanda). *Evolution du paléoenvironnement et du paléoclimat dans le domaine afro-montagnard du Rwanda au cours de l'Holocène. Geo-Eco-Trop* 22: 71–82.
- Salzmänn U (2000) Are modern savannas degraded forests? A Holocene pollen record from the Sudanian vegetation zone of NE Nigeria. *Vegetation History and Archaeobotany* 9: 1–15.
- Salzmänn U and Hoelzmann P (2005) The Dahomey Gap: An abrupt climatically induced rain forest fragmentation in West Africa during the late Holocene. *The Holocene* 15: 190–199.
- Salzmänn U, Hoelzmann P, and Morczinek I (2002) Late Quaternary climate and vegetation of the Sudanian zone of Northeast Nigeria. *Quaternary Research* 58: 73–83.
- Salzmänn U and Waller M (1998) The Holocene vegetational history of the Nigerian Sahel based on multiple pollen profiles. *Review of Palaeobotany and Palynology* 100: 39–72.
- Schulz E (1980) Zur Vegetation der östlichen zentralen Sahara und zu ihrer Entwicklung im Holozän. *Würzburger Geographische Arbeit* 51: 1–194.
- Schulz E (1991) The Taoudenni-Agorgott pollen record and the holocene vegetation history of the Central Sahara. In: Petit-Maire N (ed.) *Paléoenvironnements du Sahara. Lacs Holocènes à Taoudenni (Mali)*, pp. 143–160. Paris: CNRS.
- Schulz E (1994) The southern limit of the Mediterranean vegetation in the Sahara during the Holocene. *Historical Biology* 9: 137–156.
- Schulz E, Joseph A, Baumhauer R, et al. (1990) Upper Pleistocene and Holocene history of the Bilma Region (Kawar, NE-Niger). *Publications Occasionnelles du Centre International pour la Formation et les Echanges en Géosciences* 22: 281–284.
- Schulz E, Pomel S, Abichou A, and Salzmänn U (1995) Climate and man. *Questions and answers from both sides of the Sahara. Publications Occasionnelles du Centre International pour la Formation et les Echanges en Géosciences* 31: 35–47.
- Ssemmanda I and Vincens A (1993) Végétation et climat dans la Bassin du lac Albert (Ouganda, Zaïre) depuis 13 000 ans B.P.: Apport de la palynologie. *Comptes-rendus de l'Académie des Sciences Paris* 316: 561–567.
- Street-Perrott FA, Huang Y, Perrott RA, et al. (1997) Impact of lower atmospheric CO₂ on tropical mountain ecosystems: Carbon-isotope evidence. *Science* 278(5342): 1422–1426.
- Tamura T (1990) Late Quaternary landscape evolution in the West Cameroon Highlands and the Adamaoua Plateau. In: Lanfranchi R and Schwartz D (eds.) *Paysages quaternaires de l'Afrique centrale atlantique*, pp. 298–318. Paris: ORSTOM.
- Taylor DM (1990) Late Quaternary pollen records from two Ugandan mires: Evidence for environmental change in the Rukiga Highlands of southwest Uganda. *Palaeogeography, Palaeoclimatology, Palaeoecology* 80: 283–300.
- Van Zinderen Bakker EM (1964) A pollen diagram from Equatorial Africa, Cherangani, Kenya. *Geologie en Mijnbouw* 43: 123–128.
- Vincens A (1987) La sédimentation pléistocène supérieur-holocène. 4.1.6. Pollens et spores: indices de l'évolution climatique. *Bulletin des Centres de Recherches Exploration-Production Elf-Aquitaine* 11: 437–451.
- Vincens A (1989) Paléoenvironnements du bassin North-Tanganyika (Zaire, Burundi, Tanzanie) au cours des 13 derniers mille ans: Apport de la palynologie. *Review of Palaeobotany and Palynology* 61: 69–88.
- Vincens A (1991) Late Quaternary vegetation history of the South-Tanganyika basin. Climatic implications in South Central Africa. *Palaeogeography, Palaeoclimatology, Palaeoecology* 86: 207–226.
- Vincens A, Buchet G, and Servant M, ECOFIT Mbalang collaborators (2010) Vegetation response to the 'African Humid Period' termination in Central Cameroon (7°N) – New pollen insight from Lake Mbalang. *Climate of the Past* 6: 281–294.
- Vincens A, Garcin Y, and Buchet G (2007) Influence of rainfall seasonality on African lowland vegetation during the Late Quaternary: pollen evidence from Lake Masoko, Tanzania. *Journal of Biogeography* 34: 1274–1288.
- Vincens A, Schwartz D, Bertaux J, et al. (1998) Late Holocene climatic changes in western Equatorial Africa inferred from pollen lake Sinnda, Southern Congo. *Quaternary Research* 50: 34–45.
- Vincens A, Schwartz D, Elenga H, et al. (1999) Forest response to climate changes in Atlantic Equatorial Africa during the last 4000 yr BP and inheritance on the modern landscapes. *Journal of Biogeography* 26: 879–885.
- Vincens A, Williamson D, Thevenon F, et al. (2003) Pollen-based vegetation changes in southern Tanzania during the last 4200 years: Climate change and/or human impact. *Palaeogeography, Palaeoclimatology, Palaeoecology* 198: 321–334.
- Watrin J, Lézine A-M, and Hély C (2009) Plant migration and ecosystems at the time of the 'green Sahara'. *CR Geosciences* 341: 656–670.
- White F (1983) *La végétation de l'Afrique*. Paris: Orstom – UNESCO.
- Wu H, Guiot J, Brewer S, Guo Z, and Peng C (2007) Dominant factors controlling glacial and interglacial variations in the treeline elevation in tropical Africa. *PNAS* 104: 9720–9724.
- Zogning A, Giresse P, Maley J, and Gadel F (1997) The Late Holocene palaeoenvironment in the Lake Njupi area, west Cameroon: Implications regarding the history of Lake Nyos. *Journal of African Earth Sciences* 24: 285–300.

Australia and New Zealand

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Introduction

Geological, Climate, and Biogeographic Setting

Australia and New Zealand had similar positions in terms of latitude and proximity to Antarctica until the late Cretaceous, with all these regions located in high southern latitudes under the influence of westerly air flow. From the occurrence of forests of various conifers and with *Nothofagus* prominent among the angiosperms, it is likely that they had moist and mild climates. New Zealand separated from this connection about 80 Ma and Australia from Antarctica about 35 Ma, and from these times the two regions had different floristic histories.

The Australian region of the Indian–Australian Plate drifted northward into the mid-latitudes and tropics. By the Pliocene, only the south and Tasmania remained under the influence of the westerlies, and the climate of a large part of the continent was dominated by the influence of a subtropical high, causing a large area to become arid, and the far north to come under a monsoon climate. The flora diversified rapidly under these conditions, rainforests and particularly the conifers decreased in extent, while *Eucalyptus*, *Acacia*, and *Banksia* woodlands and heaths developed and arid shrublands and grasslands gained prominence. During this interval, there was very little landscape renewal and soils became depleted in nutrients through leaching.

In New Zealand, the climate remained firmly controlled by its position in relation to the westerlies, but the central spine of mountains allowed a relatively drier eastern South Island environment to develop. The high mountains, mid-southern latitudes, maritime environment, and active volcanism and with regular regionally extensive glaciation from the late Tertiary meant that New Zealand had relatively regular landscape renewal compared to Australia. This resulted in much more nutrient-rich soils, no strongly seasonal drying cycle, and hence a better environment for forest.

A consequence of Australia's drift northward was that it also came in contact with Asian floras and faunas. New Zealand was much more isolated so there was some limited exchange between Australia and Asia but very little between New Zealand and anywhere else. By the late Tertiary, New Zealand's flora had retained a strong similarity with its early Tertiary and Australian counterparts, while for Australia many new environments and adapted floras were in place. The long period of isolation meant that both regions developed highly endemic floras. Several glacial–interglacial cycles took place from the late Pliocene, and in the very latest part of the Quaternary human impacts caused natural systems to be subjected to prolonged environmental perturbations. It is the last phase of these events that the present contribution considers.

Lead-up to the Last Glacial Maximum

The last great glacial cycle began about 125 ka, and in MIS 5e the global climate was relatively warm. From about 100 ka, there was a general trend of cooling toward the Last Glacial Maximum (LGM), centered around 21 calendar ka. The climate changes toward this pivotal time were not straightforward. Every study shows a great deal of climate variability, with abrupt events, relatively warm intervals, and relatively cool intervals, of various time spans and severity along the way (e.g., [Alverson et al. \(1999\)](#), [Bradley \(1999\)](#), and [Alley \(2000\)](#)). The big changes are known to be controlled by changes in Earth's orbital parameters ([Hayes et al., 1977](#)). The scale of changes that occurred on submillennial timescales are still questions of debate. There remains some uncertainty of the actual timing of the peak of LGM. Some glaciers in New Zealand, for instance, seemed to have reached their maximum size several thousands of years before 21 ka BP.

The LGM pattern in Australia is one of general aridity rather than of ice cover as occurred for much of Europe and northern North America. The extent of ice in Australia was small, being restricted to the mountainous southeast and in Tasmania. Similarly, the direct impact of ice on vegetation was small; however, periglacial conditions and increased continentality had strong regional impacts in some areas. Inland Australia and indeed many now-humid areas were so dry at the LGM that few sites of pollen deposition have been found. Thus, the database for reconstruction of vegetation at the LGM is small. Broad land bridges from the mainland to New Guinea and Tasmania created a land surface about 30% greater than today; opportunities for migration of plants, animals, and humans were enhanced.

The vegetation patterns, especially in the south, were greatly modified by these environments. A large part of southeast Australia was converted from forest and woodland to a semi-arid steppe, dominated by grasses, Asteraceae, sometimes chenopods, and only small numbers of forest or heathland species were present (see [Figure 1 \(a\)](#)). It appears that for the south-west, heath and woodland taxa survived right through the LGM. There was no ice in that region, and no great increase in continentality. It is possible that the very high biodiversity led to some inbuilt resilience to conditions at the time so that although species composition changed, no great shift in vegetation structure occurred. For latitudes less than about 30°S, along the eastern coast, there is less evidence of grassland and steppe, and woodland and forest formed the main natural vegetation structure as it does today. In most cases, the vegetation had a drier aspect to it; thus, rainforest areas seemed to be replaced by *Eucalyptus* woodland. There are no reliable records from the arid interior for the LGM – it may be that these areas

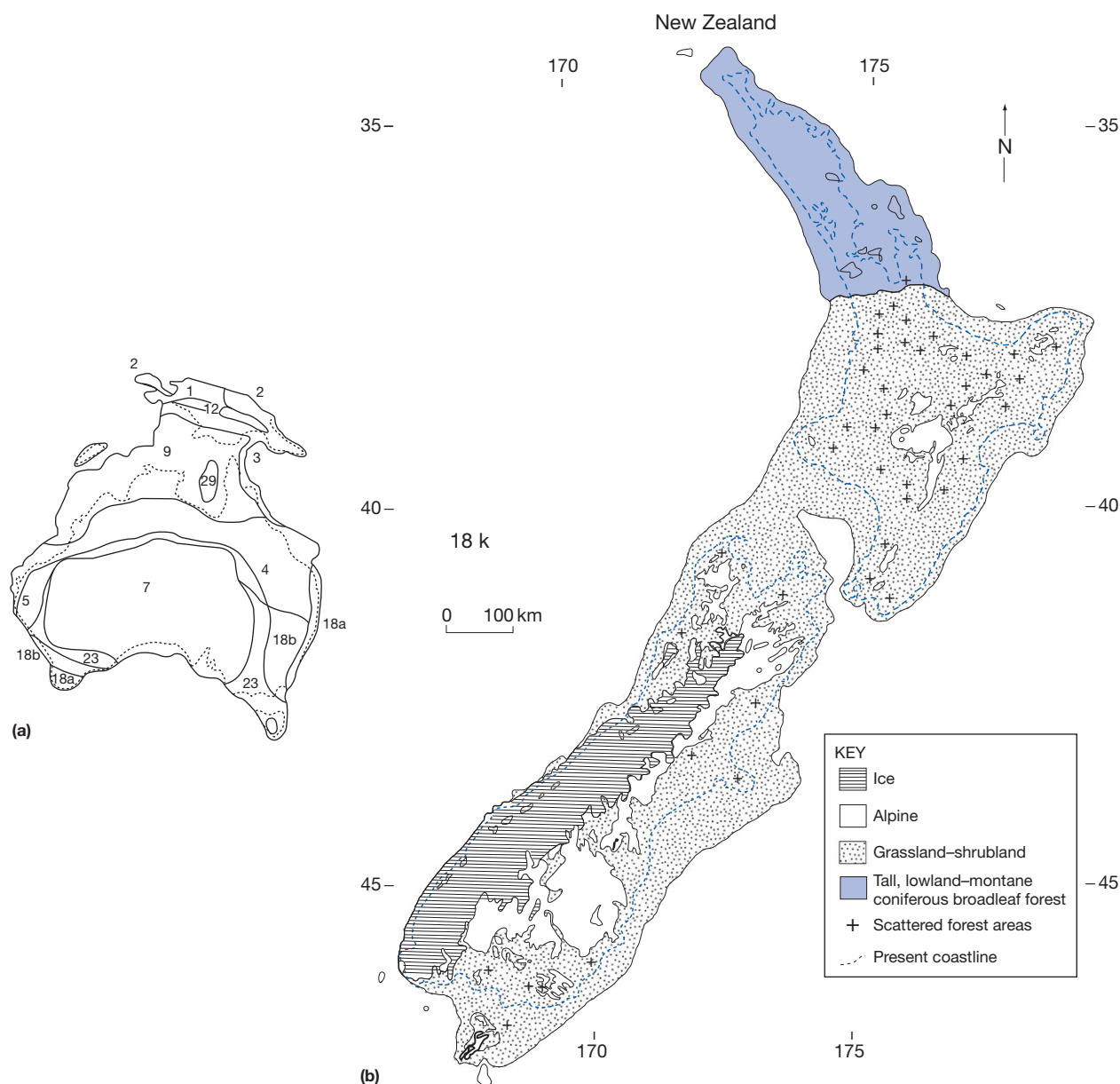


Figure 1 Estimate of vegetation cover for (a) Australia and (b) New Zealand at Last Glacial Maximum. (a) 1. Tropical rainforest (evergreen or semievergreen forest of humid tropics, usually tall) (corresponds to Olson seasonal tropical forest and broadleaved humid forest. Also includes swamp forest). Leaf cover above a level 8 m off ground, >50%. No more than 50% loss of canopy leaf cover at any one time during average year. 2. Monsoon or dry forest (medium height, deciduous or mainly deciduous forest of warm climates) (not corresponding to an Olson category; approximately a subdivision of seasonal tropical forest). Leaf /branch cover above a level 8 m off the ground, >50% during peak month of leafiness. >50% loss of canopy leaf cover at some stage in average year, except for Australian *Eucalyptus* dry forests, where limit of rainforest is defined by where *Eucalyptus* becomes >50% of canopy. 3. Tropical woodland. 4. Tropical scrub. 5. Tropical semidesert (sparse scrub or sparse grassland) (Corresponds to subdivision of Olson desert and semi-desert). Less than 2% vegetation cover above 80 cm off the ground. 25–4% vegetation cover between 0 and 80 cm off the ground, during an average year. 9. Savanna (dense grassland with a scattering of trees and/or bushes) (subdivision of Olson tropical savanna and woodland). 12. Montane tropical forest (evergreen, adapted to cool temperatures) (not corresponding to an Olson category). 18. Semiarid temperate woodland or scrub (various open woody vegetation types) (corresponds to Olson warm conifer and other woodlands). (18a = temperate woodland, 18b = temperate scrub.) Cover above 8 m, less than 20%. Cover 0.8–8 m, above 20%. 23. Temperate semidesert (sparse shrubland or grassland) (subdivision of Olson desert and semidesert). 4–25% total above ground cover. 29. Lakes and open water. (b) Reproduced from McGlone MS, Salinger MJ, and Moar NT (1993) Paleovegetation studies of New Zealand's climate since the Last Glacial Maximum. In: Wright HE Jr., Kutzbach JE, Webb T III, et al. (eds.) *Global Climates since the Last Glacial Maximum*, p. 301. Minneapolis: University of Minnesota Press, with permission from University of Minnesota Press.

also gained a drier aspect, perhaps becoming more open in structure. However, these regions today see great variability in rainfall and it may be that very little change took place. In short, the overall picture suggests a reduction in biomass, and, although climate seems to have driven this, it is not clear if reduced CO₂ was also a factor.

For New Zealand, lowered sea level joined the main islands into an area the size of a small continent (Figure 1(b)). A substantial ice cap covered South Island and high mountain areas in North Island at places like Mt. Egmont and the central volcanic plateau. New Zealand, then, does have a genuine postglacial history as opposed to most of Australia. Periglacial conditions were widespread, and large parts of the South and central North Island were covered by grasslands. Small patches of scrub and forest probably survived in protected valleys. True forest was restricted to the north, perhaps in latitudes north above 38°S (Figure 1(b)).

Recovery from the LGM

Recovery from the LGM grasslands and steppe, as viewed in terms of vegetation change, was most likely a process that had several phases. Initially the changes were small: increasing numbers of shrub and tree pollen in the south, gradual conversion of grasslands into shrublands, then shrublands with scattered trees. Examples of this from Tasmania, South Island, and southeast mainland Australia are shown in Figure 2. Around the present rainforest areas of both Australia and New Zealand, small numbers of rainforest taxa appeared or even survived in LGM refugia. About 10 radiocarbon ka, a significant change took place. Tree numbers increased rapidly in all areas where woodland and forest occur today. This coincides with many lake systems becoming permanent water bodies after being largely dry basins, sometimes associated with active dune fields, throughout the LGM. This is a clear indication that precipitation values were approaching modern (Harrison and Dodson, 1993).

Some effort has been spent in trying to identify a Younger Dryas signal in records from both Australia and New Zealand (e.g., McGlone (1995)). No clear indication of this has been observed; indeed, no major or significant changes have been identified in the interval, such as Heinrich Events, causing change in vegetation cover or vegetation recovery after the LGM. Perhaps these phenomena were not significant in the region; the vegetation histories examined so far have suffered from lack of sensitivity or the sites have insufficient time resolution to resolve the matter. Examples of high resolution records for the interval 15–10 ka are shown in Figure 3. These generally indicate small degrees of change that were little different from the degree of change observed for much of the Holocene.

Patterns of Change in the Holocene in Australia

The Holocene opens with the main patterns of vegetation close to modern for large parts of Australia. This means that since the LGM, significant adjustments involving migration and

re-establishment had occurred during about 8 radiocarbon ka, an interval almost as long as the Holocene itself.

In southeast Australia, the majority of detailed records are from the Western Basalt Plains of Victoria and adjacent south-east South Australia. This largely reflects the abundance of sites in close proximity to several research centers. The Basalt Plains is an area dominated by cracking clays. This excludes many plant species that might otherwise be suited to the Mediterranean-type climate of mild moist winters and hot dry summers, and probably gives a distorted view of vegetation change when it is applied widely across the region. From the early Holocene, woodland was already established (e.g., Dodson (1979)). The understorey of the woodlands had a relatively open character and that was maintained through the Holocene. The greatest shifts in vegetation structure occurred after the time of European settlement. In mountainous regions of northeast Victoria and southern New South Wales, the early Holocene opens with woodland and shrubland and a mid-Holocene interval with evidence of expanded moist and relatively warm taxa such as *Pomaderris* spp. (native hazel tree) (e.g., Martin (1999)). However, there is no clear evidence of expansion of alpine *Eucalyptus* species above the present snowline (at about 1800 m a.s.l.), so this change is regarded as having had small dimensions. In the Barrington Tops region of central-eastern New South Wales, this same mid-Holocene period was marked by a small expansion of temperate rainforest taxa. From about 4 ka BP, the vegetation opens up again, with snow grasses being more prominent and this has been regarded as associated with a slight cooling and drying. This pattern was reversed after 1.5 ka BP, and is associated with renewed peat growth. It is hypothesized that this is due to decreased evaporation (Dodson et al., 1986; Dodson, 1987). There may be a role for fire in the dynamics of these patterns but the extent of this is not clear.

Tasmania is treated separately here because its vegetation contains a significant number of endemic conifer and podocarp species. The early Holocene opens with mixtures of woodland and shrubland (e.g., Macphail (1979)). Then, within a thousand years, there is an expansion of more mesic vegetation, and from the mid-Holocene a cooling trend is evident as grassland taxa expand and *Pomaderris* declines. Much of the vegetation is very sensitive to the frequency, intensity, and probably seasonality of the fire regime, and many records do not include fire histories to see the impact of these. Cooling in the late Holocene and the expansion of wetland may have expanded the area of fire shadow habitats and allowed for greater development of *Athrotaxis* (pencil pine) woodlands and forest, as seen at Solomons Jewel Lakes where expansion of *Athrotaxis* follows *Sphagnum* hummock and hollow formation (Dodson, 2000).

The Atherton Tableland of northeast Australia has rich soils and it is moist all year due to the Australian monsoon and winter trade winds. The natural vegetation is dominated by tropical rainforest. The area was probably covered by sclerophyll forest (woody evergreen plants having leaves that are hard and tough, so reducing the rate of water loss) and remnant rainforest patches at LGM. In the early Holocene, the dominance of rainforest over sclerophyll forest took about 2–3 ka. This trend is assumed to be a lag effect due to the relatively slow migration of rainforest taxa from LGM refugia; in other words, vegetation change significantly lagged the trend

in climate (e.g., Kershaw (1983)). After about 3 ka BP, there is a reduction in rainforest taxa which may be related to increased burning by Aboriginal people, a slight drying of the climate, or both (Kershaw, 1986).

At the Top End, in the Northern Territory, there are few records and these all come from the late Holocene. Most of these come from coastal plains and mainly reflect marine transgressions and sedimentation regimes. Rapid sedimentation

began about 7–6 ka BP, and the records are dominated by mangrove pollen (e.g., Crowley et al. (1990)). Continued sedimentation gradually changed the environments to freshwater swamps, especially after about 5.3 ka BP, with occasional marine incursions. The regional terrestrial vegetation signals are generally swamped but where present suggest *Eucalyptus* woodlands and rainforest taxa, as is the case today (Woodroffe et al., 1986; Clark and Guppy, 1988).

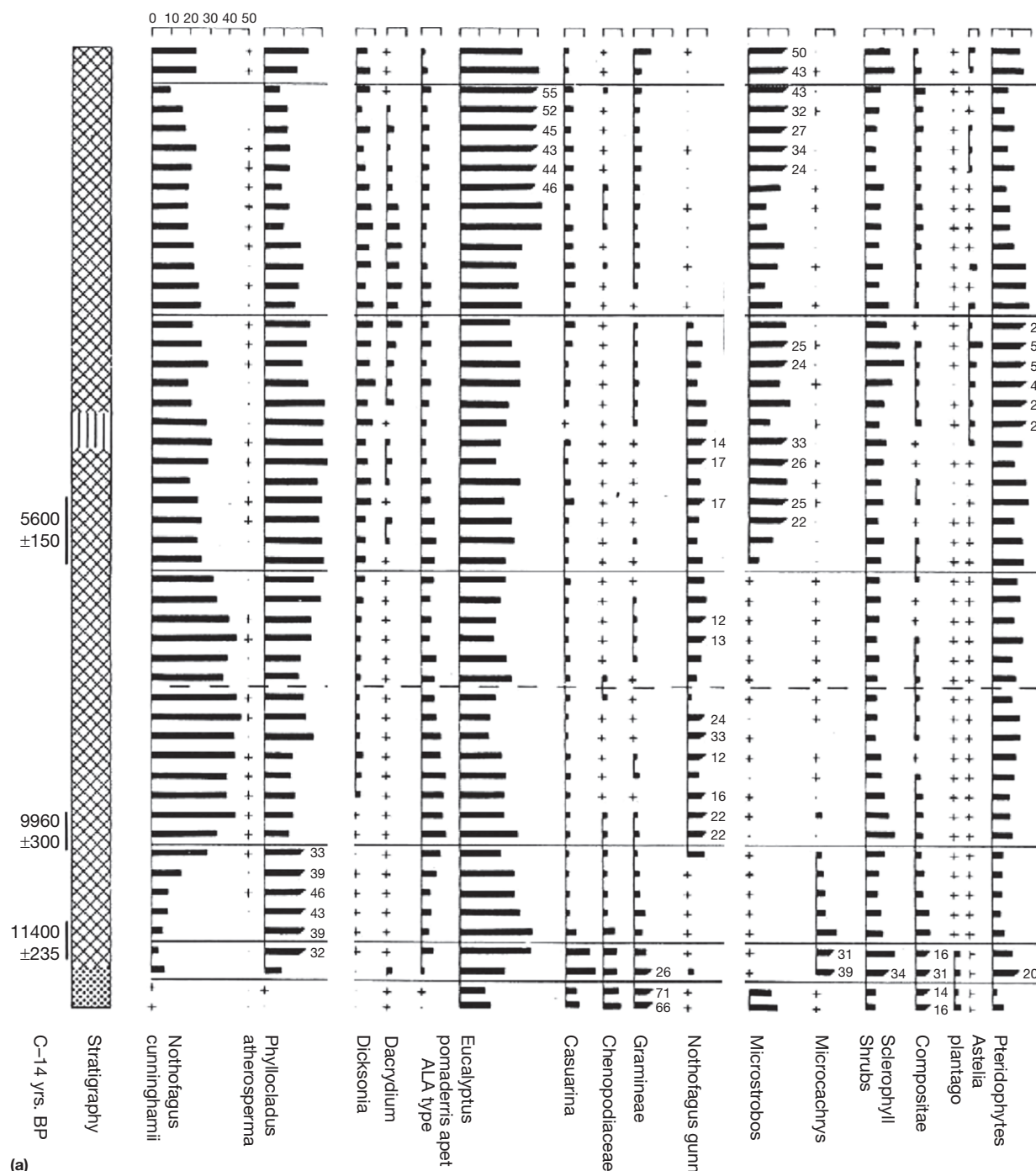


Figure 2 (a) Pollen diagram from Eagle Tarn, Upland Tasmania (redrawn from Macphail MK (1979) Vegetation and climates in southern Tasmania since the last glaciation. *Quaternary Research* 11: 306–341).

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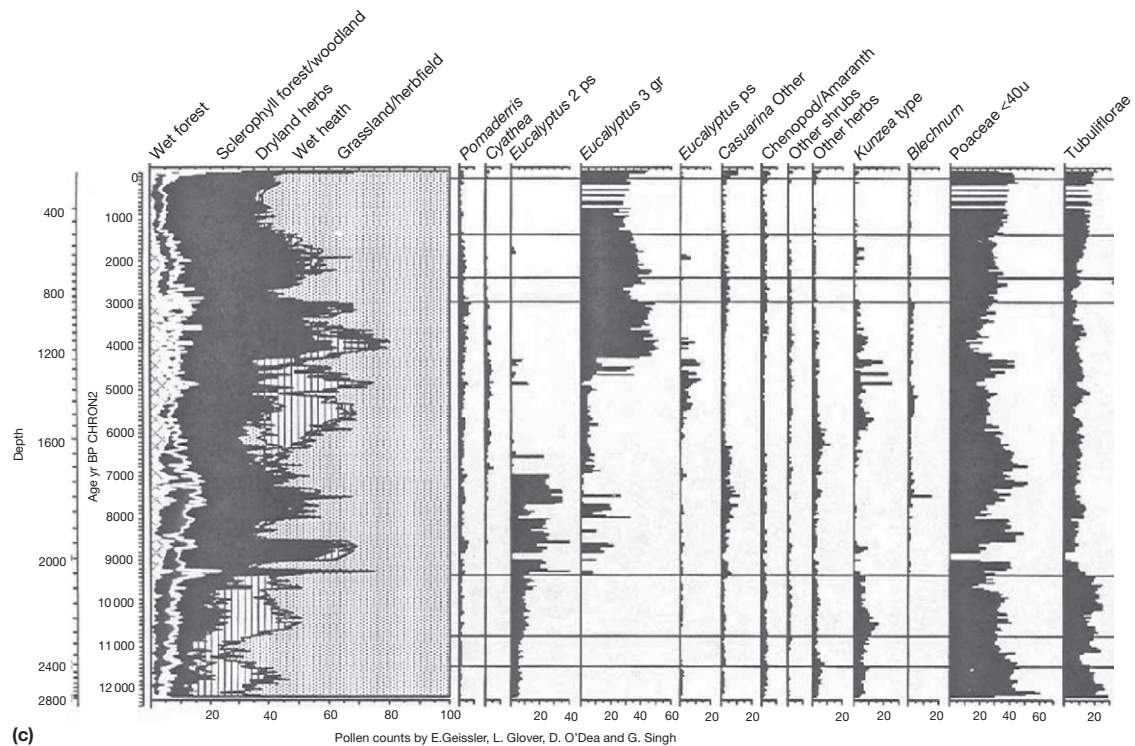
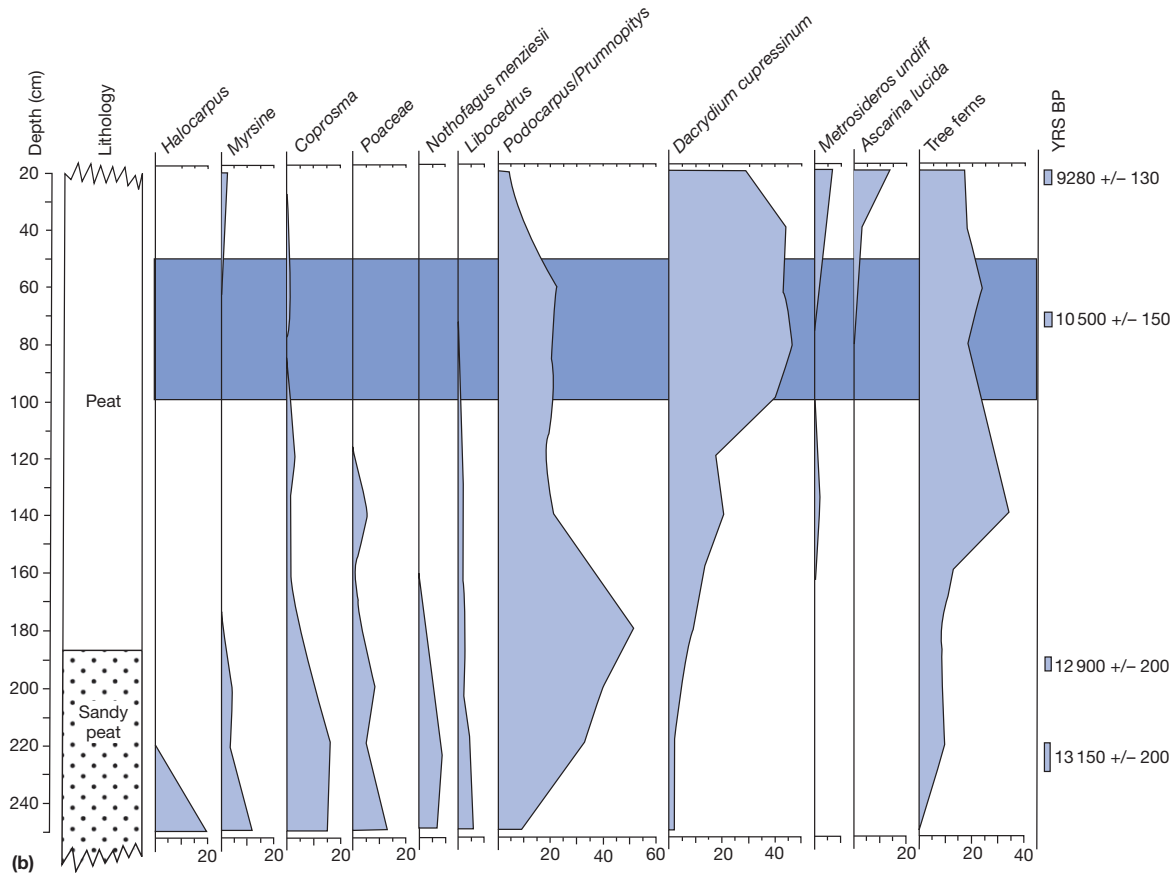


Figure 2 (b) Pollen diagram from Durham Rd, North Island, New Zealand (redrawn from [McGlone MS \(1995\)](#) Lateglacial landscape and vegetation change and the Younger Dryas climatic oscillation in New Zealand. *Quaternary Science Reviews* 14: 867–881). (c) Examples of Holocene-age pollen diagrams from Australia Bega Swamp, New South Wales (redrawn from [Hope GS, Kershaw AP, et al. \(2004\)](#) History of vegetation and habitat change in the Austral-Asian region. *Quaternary International* 118–119: 103–126).

There are few Holocene records of vegetation change in the semiarid and arid regions of Australia. In some cases, they come from peats associated with groundwater discharge and are therefore associated with very local wetlands. Until the records are better developed, the degree of impact on vegetation as recorded for southeast Australia, and accordingly the recovery from LGM conditions was more subdued. The main change seems to be that *Casuarina* became less important during the Holocene while *Eucalyptus* became more important. The reasons for this are unclear. It is unlikely to be a trend in soil leaching because most of the landscape has ancient and highly leached soils. Because of the high biodiversity of the region, and the low resolving power of pollen analysis in identifying potential *Casuarina* or *Eucalyptus* species in this transition, it is difficult to target the possible causes of the change beyond possible successional change. *Casuarina* is a nitrogen fixer and the later expansion of *Eucalyptus* may depend on subtle changes in soil nitrogen balance. No clear relationship between possible fire regime change and vegetation change has been identified for the region.

In summary, the records from Australia show a series of subtle vegetation changes across the Holocene. Apart from the southwest, there is a general pattern of an increased mesic interval in the early to mid-Holocene, which appears to show differences in timing from place to place. The differences in

timing for the beginning of this phase may depend on real regional differences in the establishment of climate patterns, but could also be related to lag times for key taxa to migrate into those regions after the LGM. The differences in timing at the end of the mesic phase may be related to a combination of climate shifts and innate resilience and subsequent reaction of vegetation to new conditions. It is possible that fire, perhaps in combination with climate and anthropogenic impacts, may have been important at some sites in setting the actual timing of changes, although it is a debated topic (e.g., Hallam (1975) and Horton (1982, 2000)). The southwest-southeast differences may relate to the fact that inbuilt resilience in the former could be due to the high biodiversity, which allowed many related species to fill gaps in times of change. In this case, some of the change goes unnoticed because it is difficult to observe with the taxonomic resolving power of routine palynology. It is known that even in the Tertiary, fire regimes in the absence of humans were only a little different from the historical period (Atahan et al., 2004; Dodson et al., 2005)

Patterns of Change in the Holocene in New Zealand

For most of New Zealand, the Holocene opens with many vegetation types about to undergo further significant change

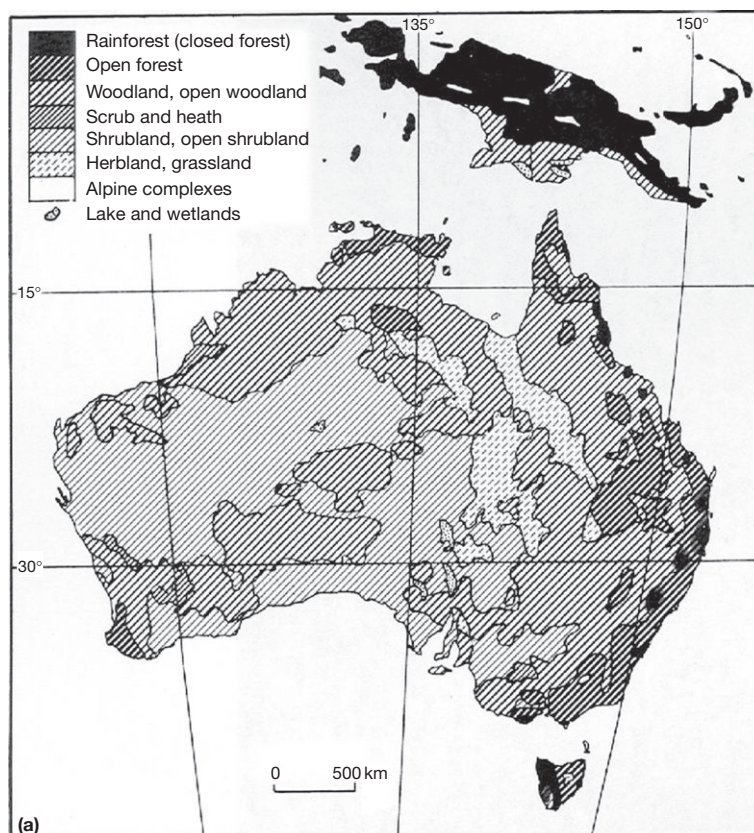


Figure 3 Estimated vegetation cover during the Holocene optimum for (a) Australia and (b) New Zealand. (b) Reproduced from McGlone MS, Salinger MJ, and Moar NT (1993) Paleovegetation studies of New Zealand's climate since the Last Glacial Maximum. In: Wright HE Jr., Kutzbach JE, Webb T III, et al. (eds.) *Global Climates since the Last Glacial Maximum*, p. 298. Minneapolis: University of Minnesota Press, with permission from University of Minnesota Press.

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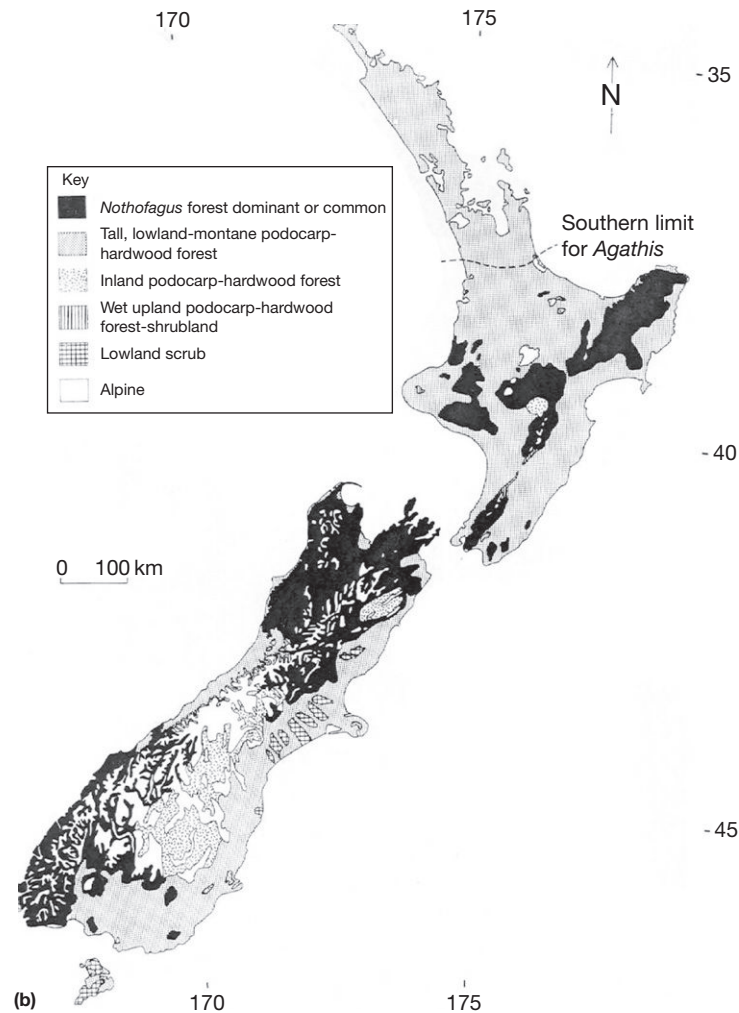


Figure 3 (Continued)

before they assume their modern character. Central North Island and South Island have a Holocene history initially dominated by recovery from LGM conditions, and with possibly periodic interruptions from volcanism in the North Island. Forests survived LGM in the far north, above about 38° S, and Holocene vegetation changes were more subtly related to climate rather than dominated by a recovery mode.

In the early Holocene, some southern sites open their records with shrubby taxa such as *Halocarpus*, *Myrsine* (matipo), *Coprosma*, or *Phyllocladus* (celery pine) species, and especially near subalpine areas there is a strong grassland element (Figure 4). These then become dominated by one or more tree Podocarpaceae taxa, such as *Podocarpus-Prumnopitys*, and *Dacrydium cupressinum* (rimu). Some sites have early Holocene records with forest already established and presumably these were near forest refugia at the LGM. Then, depending on location, *Nothofagus* spp. (southern beech) begin to appear, before the early Holocene. In some places afforestation was rapid, especially in central North Island and northern South Island, and a few thousand years later in southern and eastern South Island. There is a pattern of establishment that probably reflects migration from refugia, mainly in the mountains

around southern North Island, Nelson, and possibly southland. The migration rates were slow but when *Nothofagus* appeared it grew to dominate the forest canopy and its dense nature created low light conditions at ground level, ultimately leading to the elimination of podocarps. Subsequently, large areas of forests were dominated by one or more *Nothofagus* species. On the lowlands of central western South Island, *Nothofagus* was not present in the Holocene (it was about 30 ka BP). The region is still dominated by tall podocarp forests and sometimes the region is referred to as the beech (*Nothofagus*) gap (see Figure 4 for examples). Other small regions remained occupied by podocarp forests, for example, some wetlands occupied by *Dacrycarpus dacrydioides*, or shrublands in areas adjacent to large active braided streams.

In addition to these changes, there was one other significant change noted for some central and northern sites. This was an increase in *Ascarina lucida* (hutu), a shrub to small tree that is frost sensitive. In many pollen diagrams, there is an early Holocene interval of greater abundance of *Ascarina* (see Figure 4). The decline after about 6 ka BP (McGlone and Moar, 1977) is thought to represent a cooling or more frost-prone environments after this time.

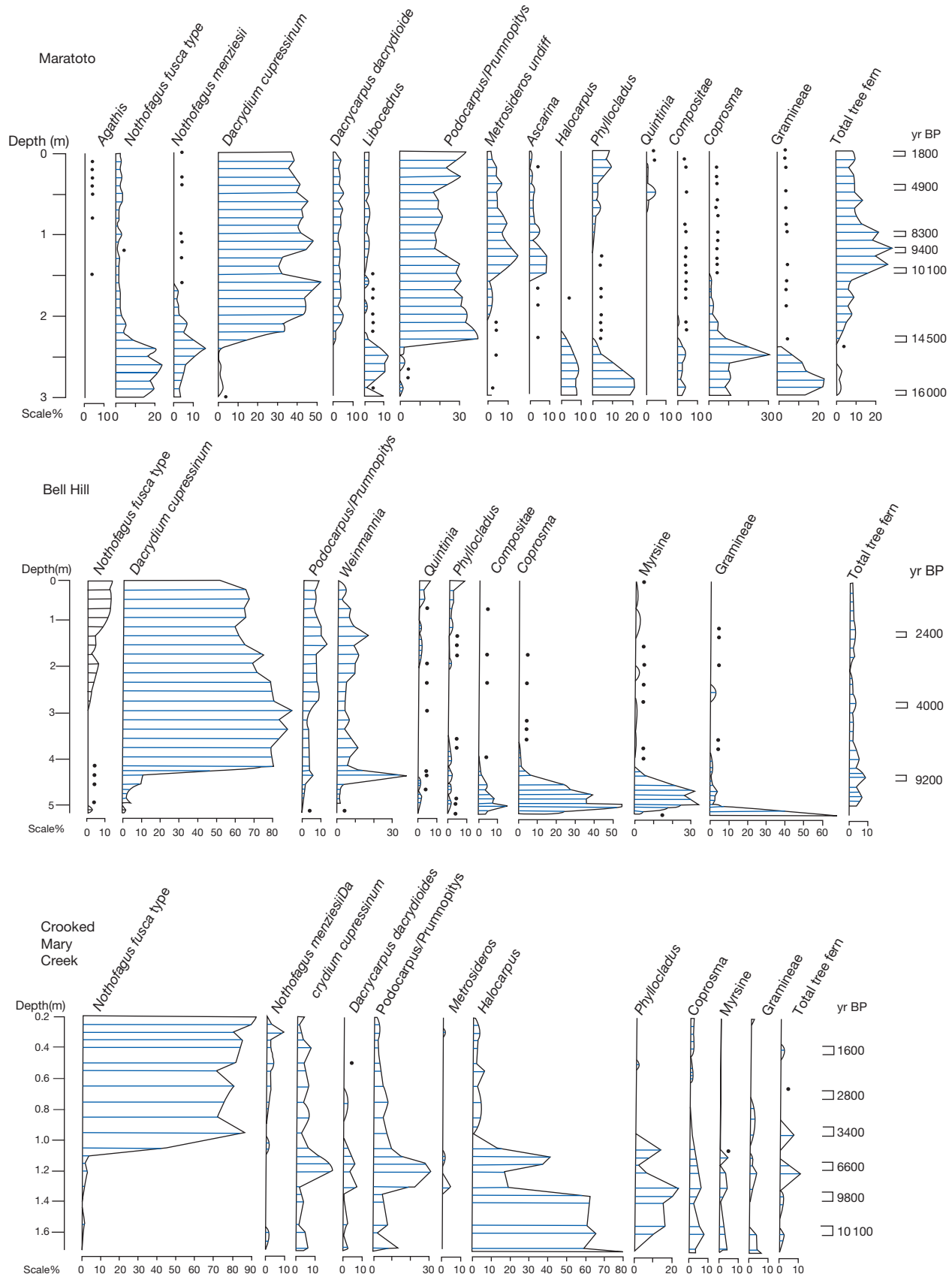


Figure 4 Examples of Holocene-age pollen diagrams from New Zealand: Maratoto (North Island), and Bell Hill and Crooked Mary Creek (South Island). Reproduced with permission, University of Minnesota Press.

Northern New Zealand retained forest cover during LGM, probably as a result of the maritime environment and warmer climate in the region. The vegetation adjustments in the post-glacial were therefore much less than further south. Pollen diagrams show that *Agathis australis* (kauri) was present and that *Nothofagus menziesii* (silver beech) was prominent (e.g., Dodson et al. (1988), Kershaw and Strickland (1988), and Newnham (1999)). During the Holocene, only small vegetation adjustments appear to occur for these regions.

Anthropogenic Impacts

Australia and New Zealand had vastly contrasting human histories and hence human impacts on vegetation. The evidence suggests that humans arrived in Australia from Asia about 55–60 ka BP. People occupied most of the continent before LGM and had thus adapted to the full range of environments, and had coped with full glacial conditions that were severe in the south, and most especially in the southeast. In stark contrast, humans arrived in New Zealand within the last thousand years. In this case, the people were of Polynesian origin and were spreading to their most southerly location, well beyond their original homelands in the tropics. They brought their crops and fishing methods with them, and although the environment was different from the tropical and subtropical Pacific, they did not need to weather any great climate or natural habitat shifts, that were not of their own making, after settlement. European settlement began in both countries about 200 years ago. Native people were displaced, and new forms of agricultural practices and urbanization took place, which initiated a series of major and probably irreversible changes in the environments.

The long period of Aboriginal occupation of Australia before the Holocene appears to have led to long-standing patterns of land use, burning, and hunting. Accordingly, it is difficult to identify any vegetation adjustments that relate specifically to changing patterns of human impact in the Holocene. A long-standing debate is the timing of the demise of a large browsing and grazing fauna termed the Australian megafauna (see [Late Pleistocene Megafaunal Extinctions](#)). Although there are some Holocene dates, these are possibly unreliable and it is likely that the megafauna were extinct by the late Pleistocene. Interpretation of possible adjustments in vegetation composition and structure related to these is fraught with the difficulty of matching the timing of the extinctions with climate change and human impacts such as hunting. Present knowledge suggests this as a late Pleistocene question rather than a Holocene one. One thing we do know is that recent work suggests that burning patterns were already well established before the Holocene.

Polynesian settlement had a very strong impact on vegetation systems in New Zealand. People arrived into a well-forested landscape (below tree-line) and began clearing for agriculture and hunting of the large and abundant bird fauna. The birds were naïve and apparently easily hunted to extinction ([Anderson, 1989](#)). Consequently, major changes in vegetation cover took place, and most of this occurred within a few hundred years. The scale of human impact was more severe in South Island, where forest recovery rates may have been

lower in the cooler conditions compared to the north. Wood charcoal is very abundant in the soils and it is clear that fire was an important tool used by people to clear forests for crops and an aid in hunting. The megapode birds became extinct in a short period due to hunting and loss of habitat.

European settlement began in Australia and New Zealand in the late eighteenth century, and from about AD 1840, much of the good soils were being utilized for agriculture and human settlements. In pollen diagrams, the first signs of European settlement are associated with clearing of woodland and forest, the establishment of grassland, and the introduction of many new species, including weeds. The latter include *Pinus*, *Plantago*, and *Echium*, plus sometimes an increase in native species such as *Pteridium*, which were favored by clearing ([Figure 5\(b\)](#)).

Some Outstanding Questions

Several late Pleistocene and Holocene abrupt and relatively short-lived climate events are widely recognized in many records from the Northern Hemisphere. These include Dansgaard–Oeschger events, the Younger Dryas cold interval, and the 8.2-ka and 4.2-ka dry and cool events, the Medieval Warm Period and the Little Ice Age. It is important to establish whether these can also be identified in Australia and New Zealand and thereby distinguish whether these were events caused by regional- or global- scale mechanisms. For those that are not represented globally, more regional explanations need to be identified for their causes. In all cases, there is no unequivocal representation of these events in either New Zealand or Australia. The reason for this is not clear; possible explanations include that they were of such minor magnitude that their effects were minimal, or perhaps the time resolution of the records studied so far has been too coarse to reveal them, or perhaps the long evolution of the floras of the region in isolation has predisposed the floras that have been studied to show little reaction to these events. If this is the case, then proxy data series unrelated to vegetation change may be necessary to identify them.

A major question attracting attention is the significance of fire in the environment, and the role of humans in changing its frequency, intensity, and patterns. It is generally considered that this was important in shaping vegetation and environments in Australia and New Zealand, but probably at different times. New Zealand is a wonderful laboratory for testing human impacts. We know that the changes wrought since European settlement have created environments that have proved to be hostile to most native plant and animal species. Since Polynesian settlement, perhaps 700 years ago, significant amounts of clearance, at least accompanied by fire, have been the norm. Natural fire was associated with volcanism and lightning, but because there were no anthropogenic effects before 700 yr BP, it becomes possible to investigate records for their dynamics and trends in comparison with the modern day. Research in this area will undoubtedly open up insights in to how systems work, and enable environmental managers to make better informed decisions on biodiversity, conservation, and on the dynamics of natural systems.

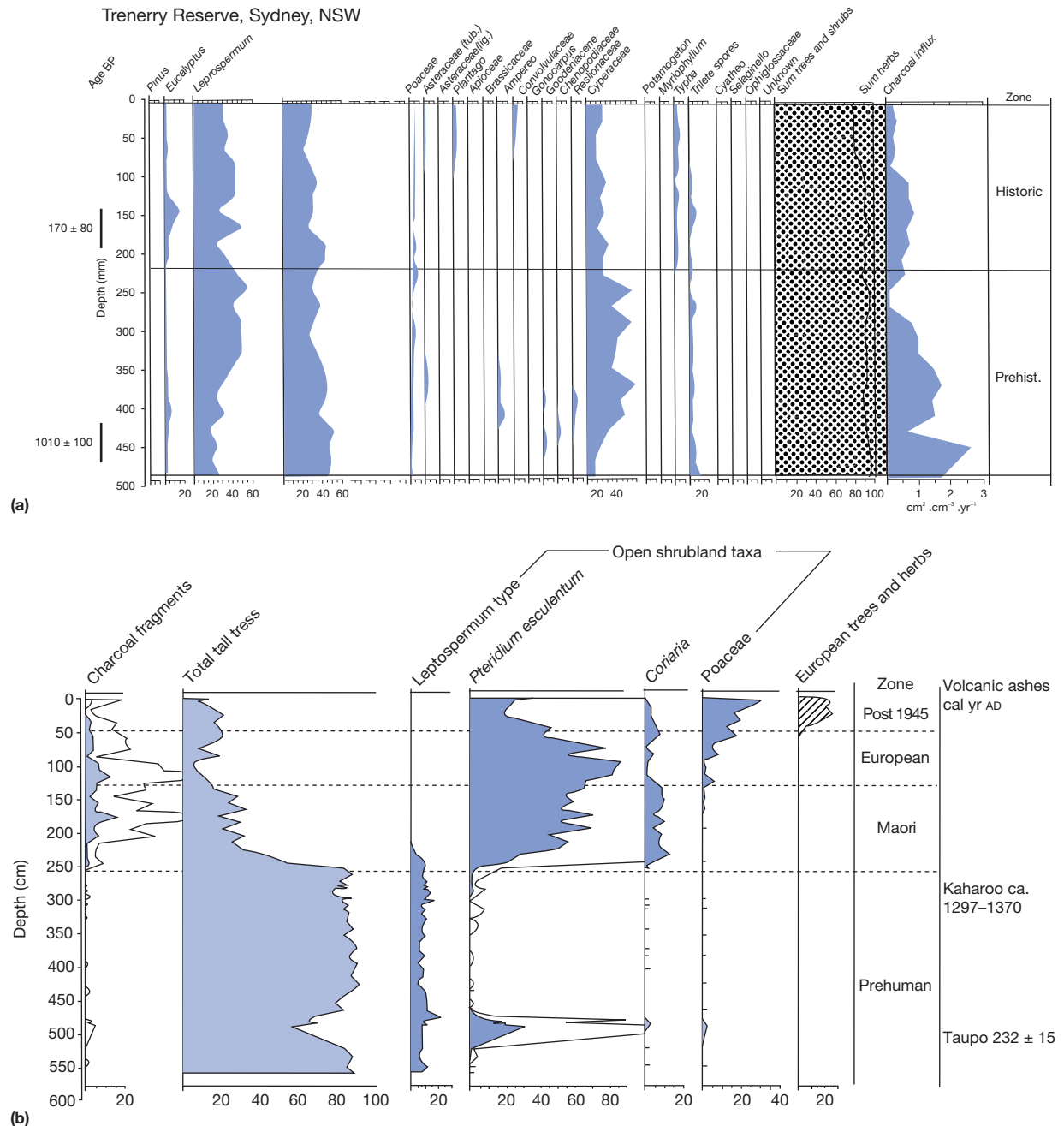


Figure 5 Selected pollen diagrams showing human impacts on vegetation in Australia and New Zealand.

See also: **Plant Macrofossil Methods and Studies:** Rodent Middens. **Vertebrate Records:** Late Pleistocene Megafaunal Extinctions.

References

- Alley RB (2000) *The Two-Mile Time Machine*, p. 290. New Jersey: Princeton University Press.
- Alverson K, Oldfield F, and Bradley RS (eds.) (1999) *Past Global Changes and their Significance for the Future*, p. 479. Amsterdam: Elsevier.

- Anderson A (1989) *Prodigious Birds: Moas and Moa Hunting in Prehistoric New Zealand*. Cambridge: Cambridge University Press.
- Atahan P, Dodson JR, and Itzstein-Davey F (2004) A fine-resolution Pliocene pollen and charcoal record from Yallalie, south-western Australia. *Journal of Biogeography* 31: 199–205.
- Bradley RS (1999) *Paleoclimatology: Reconstructing Climates of the Quaternary*, p. 610. San Diego: Academic Press.
- Clark RL and Guppy J (1988) A transition from mangrove forest to freshwater wetland in the monsoon tropics of Australia. *Journal of Biogeography* 15: 665–684.
- Crowley GM, Anderson P, Kershaw AP, and Grindrod J (1990) Palynology of a Holocene sequence from the Mulgrave river, humid tropical north-east Queensland. *Australian Journal of Ecology* 15: 231–240.
- Dodson JR (1979) Late Pleistocene vegetation and environments near Lake Bullenmerri, western Victoria. *Australian Journal of Ecology* 4: 419–427.

- Dodson JR (1987) Mire development and environmental change, Barrington Tops, New South Wales, Australia. *Quaternary Research* 27: 73–81.
- Dodson JR (2000) A vegetation and fire history in a subalpine woodland and rainforest region, Solomons Jewel Lakes, Tasmania. *The Holocene* 11: 111–117.
- Dodson JR, Enright NJ, and McLean RF (1988) A late Quaternary vegetation history for far northern New Zealand. *Journal of Biogeography* 15: 647–656.
- Dodson JR, Greenwood PG, and Jones RL (1986) Holocene forest and wetland dynamics at Barrington Tops, New South Wales. *Journal of Biogeography* 13: 538–563.
- Dodson JR and Lu J (2000) A late Holocene vegetation and environment record from Byenup Lagoon, south-western Australia. *Australian Geographer* 31: 41–54.
- Dodson JR, Robinson M, and Tardy C (2005) Two fine resolution Pliocene charcoal records and their bearing on prehuman fire frequency in southwestern Australia. *Austral Ecology* 30: 592–599.
- Hallam SJ (1975) *Fire and Hearth: A Study of Aboriginal Usage and European Usurpation in South-West Australia*. Canberra: Australian Institute of Aboriginal Studies.
- Harrison SP and Dodson JR (1993) Climates of Australia and New Guinea since 18,000 B.P. In: Wright HE Jr., Kutzbach JE, and Webb T, III et al. (eds.) *Global Climates since the Last Glacial Maximum*, p. 569. Minneapolis: University of Minnesota Press.
- Hays JD, Imbrie J, and Shackleton NJ (1977) Variations in the Earth's orbit: Pacemaker of the ice ages? *Science* 198: 529–530.
- Hope GS, Kershaw AP, et al. (2004) History of vegetation and habitat change in the Austral-Asian region. *Quaternary International* 118–119: 103–126.
- Horton D (1982) The burning question: Aborigines, fire and Australian ecosystems. *Mankind* 13: 237–251.
- Horton D (2000) *The Pure State of Nature*. Sydney: Allen and Unwin.
- Kershaw AP (1983) A Holocene pollen diagram from Lynch's Crater, north-eastern Queensland, Australia. *New Phytologist* 75: 173–191.
- Kershaw AP (1986) The last two glacial-interglacial cycles from north-eastern Australia: Implications for climate change and Aboriginal burning. *Nature* 322: 47–49.
- Kershaw AP and Strickland K (1988) A Holocene pollen diagram from Northland, New Zealand. *New Zealand Journal of Botany* 26: 145–152.
- Luly JG (1991) *A Pollen Analytical Investigation of Holocene palaeoenvironments at Lake Tyrrell, Semi-arid Northwestern Victoria, Australia*. PhD Thesis, p. 223. Canberra: Australian National University Unpublished.
- Luly JG (2001) On the equivocal fate of late Pleistocene *Callitris* Vent. (Cupressaceae) woodlands in arid South Australia. *Quaternary International* 83–85: 155–168.
- Martin ARH (1999) Pollen analysis of Digger's Creek Bog, Kosciuszko National Park: Vegetation history and tree-line change. *Australian Journal of Botany* 47: 725–744.
- McGlone MS (1995) Lateglacial landscape and vegetation change and the Younger Dryas climatic oscillation in New Zealand. *Quaternary Science Reviews* 14: 867–881.
- McGlone MS and Moar NT (1977) The *Ascarina* decline and post-glacial climate change in New Zealand. *New Zealand Journal of Botany* 15: 485–489.
- Macphail MK (1979) Vegetation and climates in southern Tasmania since the last glaciation. *Quaternary Research* 11: 306–341.
- Newnham RN (1999) Environmental change in Northland, New Zealand during the last glacial and Holocene. *Quaternary International* 57–58: 61–70.
- Pearson S (1999) Late Holocene biological records from the middens of Stick-nest rats in the central Australian arid zone. *Quaternary International* 59: 39–46.
- Pearson S and Dodson JR (1993) Stick-nest rats as palaeoenvironmental indicators in the Australian arid zone. *Quaternary Research* 39: 347–354.
- Pearson S, Lawson E, Head L, McCarthy L, and Dodson J (1999) The spatial and temporal patterns of Stick-nest Rat middens in Australia. *Radiocarbon* 49: 295–308.
- Pickett EJ, Harrison SP, Hope G, et al. (2004) Pollen-based reconstructions of biome distributions for Australia, Southeast Asia and the Pacific (SEAPAC region) at 0, 6000 and 18,000 ¹⁴C yr BP. *Journal of Biogeography* 31: 1381–1444.
- Singh G (1981) Late Quaternary pollen records and seasonal palaeoclimates of Lake Frome, south Australia. *Hydrobiologia* 82: 419–430.
- Woodroffe CD, Thom BG, and Chappell JMA (1986) Development of widespread mangrove swamps in mid-Holocene times in northern Australia. *Nature* 317: 711–713.

Northeastern North America

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Introduction

Northeastern North America (NENA) is here defined as all US states and Canadian provinces east of 100° W that were wholly or partially covered by the Laurentide Ice Sheet at the Last Glacial Maximum, ca. 21 000 cal yr BP (21 ka). The many lakes, mires, and hollows in this deglaciated landscape have made NENA a center of palynological research for six decades, with over 500 NENA records archived in the Neotoma Paleoecology Database (Figure 1) (Grimm et al., 2013). This dense network permits detailed reconstructions of the spatial and temporal patterns of postglacial vegetation history. Plant taxa exhibit highly varied postglacial vegetation histories, characterized by large shifts in range and abundance, and relatively rapid rates of change between 16 and 8 ka (Jacobson et al., 1987; Shuman et al., 2005; Williams et al., 2004). Higher-order vegetation characteristics, such as biomes, plant functional type distributions, and associations among plant taxa, emerge from the individualistic behavior of plant taxa. Comparison of pollen records with independent paleoclimatic proxies indicates that NENA vegetation history has been primarily shaped by climatic variability at centennial to orbital timescales and by the postglacial opening up of new lands for colonization caused by the retreat of the Laurentide Ice Sheet. Other contributing factors include variations in atmospheric CO₂ concentrations (Cowling, 1999), the late Pleistocene megafaunal extinctions (Gill et al., 2009), and human land use (Denevan, 1992; Foster, 2002).

This article first reviews the history of NENA palynology, with respect to methodological advances and insights gained into the processes governing long-term vegetation dynamics. Pollen abundance maps for selected taxa illustrate postglacial vegetation history, and community-scale vegetation dynamics are represented by rate-of-change maps, maps of the minimum dissimilarities between fossil and modern pollen spectra (which indicate past vegetation formations with no modern analog), and reconstructed maps of woody cover. Finally, this article focuses on New England as a case study of the linkages between late Quaternary vegetation dynamics and climatic variability at centennial to millennial timescales. Space constraints and the rich history of palynological research prohibit a full review of all facets of NENA vegetation history, so interested readers are directed to other syntheses and reviews (Foster, 2002; Gaudreau and Webb, 1985; Grimm and Jacobson, 2004; Grimm et al., 2001; Jackson et al., 1997; Jacobson et al., 1987; Richard, 1993; Ritchie, 1987; Shuman et al., 2009; Webb, 1988; Webb et al., 1993b; Williams et al., 2004), as well as to the primary data stored in the Neotoma Paleoecology Database.

Climate and Physiography

NENA encompasses significant north–south and east–west gradients in climate and a diversity of bedrock types and glacial

deposits. Indices of air temperature, radiation flux, and winter severity (e.g., mean temperature of the warmest and coldest months of the year, growing-degree days, number of frost-free days, potential evapotranspiration, day length) trend from north to south. Indices of moisture availability (e.g., annual precipitation, relative humidity, rainfall variability, and actual evapotranspiration) generally increase from west to east. Gradients of moisture availability are particularly steep between 100° and 95° W, with mesic conditions to the east.

Topographic relief in NENA is generally low (Bostock, 1970; Fenneman and Johnson, 1946). The Precambrian Laurentian Shield underlies most of eastern Canada and portions of the northern United States and is generally characterized by low relief and an average elevation of 500 m, although areas of locally high relief occur in the Adirondacks and the Tomgat Mountains of Labrador. In the Appalachian Mountains, ranging from Newfoundland to Alabama, elevations exceed 1500 m. West of the Appalachians, the folded sedimentary rocks of the Valley and Ridge structural province give way to the generally low-relief landscape of central North America. Although relief is generally low to moderate in formerly glaciated areas, the irregular depth and texture of glacial deposits produces variable edaphic conditions at local to regional scales.

The subcontinental-scale distribution of vegetation in NENA prior to European settlement was strongly shaped by the intersection between the north–south temperature and east–west moisture gradients, with finer-scale ecological variation due in part to the patchwork of glaciogenic and periglacial sedimentary deposits and exposed bedrock. In the semiarid upper Midwest, grasslands predominated, with the prairie–forest ecotone trending NW–SE from Manitoba to Indiana. Treeless conditions also prevail in northern Québec and Labrador, where cold winter temperatures and short growing seasons limit tree growth. Forests in NENA are characterized by a strong south-to-north zonation in species composition and physiognomy, from the deciduous forests of the east-central United States to the mixed hardwood–conifer forests in New England and Great Lakes region, to the coniferous forests of eastern Canada (Küchler, 1975; Rowe, 1972). North of 50° N, tree density decreases, as the vegetation gradually changes from closed boreal forests to lichen woodlands to forest–tundra to open tundra. Within each of these broad vegetation formations, there is substantial variation in plant community composition and structure.

History of Palynological Research in NENA

Prior to the advent and widespread application of pollen analysis in the mid-twentieth century, inferences about postglacial vegetation history were primarily based upon the distribution of extant vegetation formations, in particular the

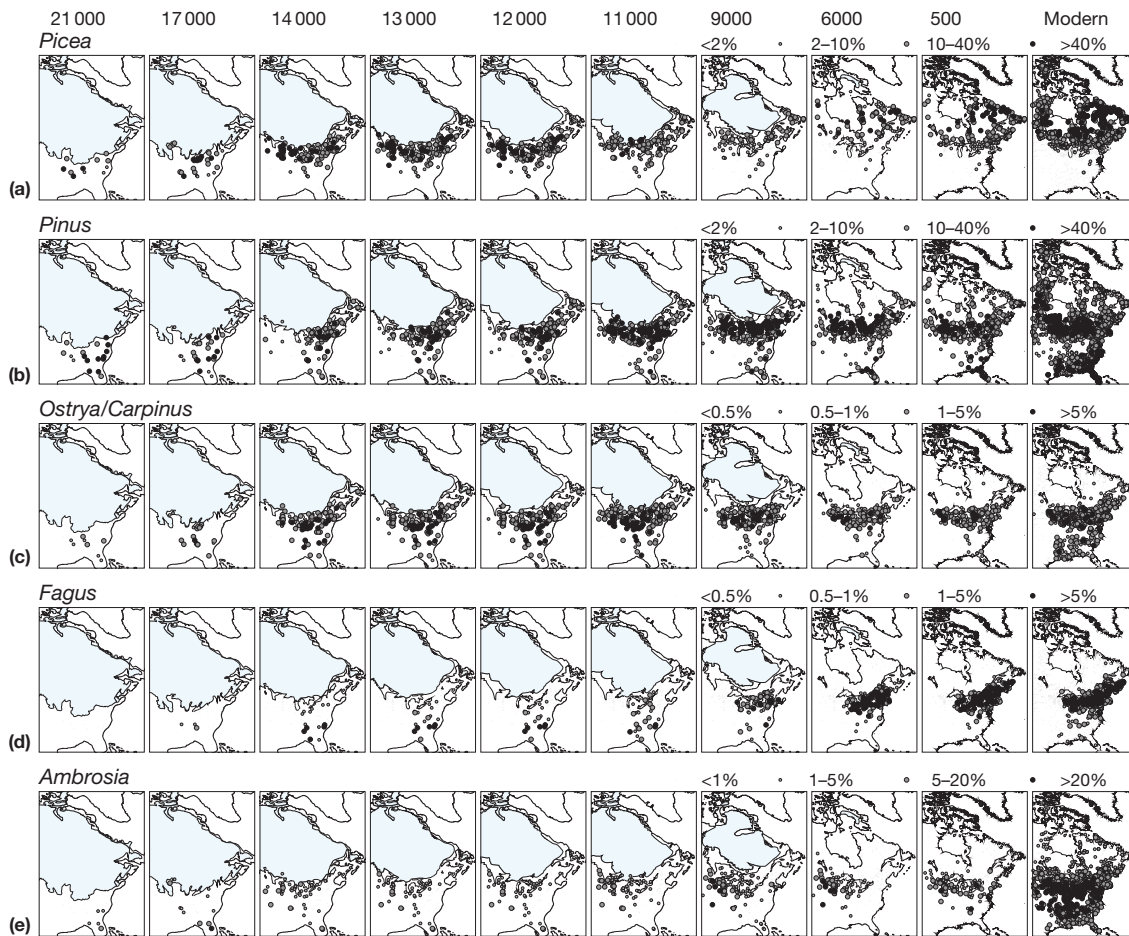


Figure 1 Mapped percent pollen abundances for (a) *Picea* (spruce), (b) *Pinus* (pine), (c) *Ostrya/Carpinus* (hophornbeam/hornbeam), (d) *Fagus* (beech), and (e) *Ambrosia* (ragweed). Data from Williams et al. (2004).

distribution of outlier vegetation types, which were interpreted as relicts of previously continuous distributions or as pioneer communities. Before radiometric dating was available, palynologists concentrated on testing the prevailing theories of plant biogeography and plant succession, honing pollen taxonomy, and developing regional pollen stratigraphies for the purposes of stratigraphic correlation. Initial investigations were focused in New England, where Deevey (1939) established a still-cited regional pollen stratigraphic zonation, and in the upper Midwest, where Potzger (1946) collected 25 pollen records to test theories about climax vegetation type and post-glacial plant migration. Other early workers included Cushing, Davis, Leopold, Ogden, Sears, Terasmae, Whitehead, and Wright.

The application of radiocarbon dating to organic sediments soon freed pollen data from their role as stratigraphic correlators and allowed new hypotheses to be tested about the synchronicity or time transgression of vegetational events. The number of pollen analysts and pollen records rapidly proliferated during the 1960s and 1970s, providing the foundation for regional- to subcontinental-scale vegetation reconstructions. In the upper Midwest, transects of radiocarbon-dated pollen diagrams demonstrated the eastward expansion of the prairie during the mid-Holocene (McAndrews, 1966). Davis (1976, 1981b) was the first

to map pollen abundances and show time-transgressive changes consistent with a general pattern of postglacial migration and expansion. Inferred migration rates were surprisingly rapid – on the order of 100–500 m year⁻¹ (Davis, 1981b) – although these estimates may be too high (McLachlan et al., 2005). Moreover, species exhibited individualistic responses to climate change (Cushing, 1965; Davis, 1981b), overturning the previous Clementsian paradigm that vegetation formations responded cohesively to climate change (Braun, 1950). In a classic example, *Tsuga canadensis* expanded into New England several thousand years before *Fagus grandifolia* (Figure 1(d)), despite their close modern association (Davis, 1981b).

Advances during the 1970s and 1980s expanded these mapping efforts and developed statistical techniques to extract quantitative paleoenvironmental information from pollen data (Bartlein et al., 1984). These advances were enabled by the consolidation of fossil pollen records into centralized databases and the collection of thousands of surface pollen samples into calibration datasets that documented pollen-climate and pollen-vegetation relationships (Whitmore et al., 2005; Williams et al., 2006). These pollen databases are still actively supported and used and are now incorporated into the Neotoma database (Blois et al., 2011a; Grimm et al., 2013). Early work showed that the principal axes of variation in the

composition of surface pollen samples corresponded to those for the surrounding vegetation (Webb, 1974) and that pollen abundances positively correlated to plant population density (Delcourt et al., 1984). Primary emphasis soon shifted to pollen-based paleoclimatic reconstructions (Bartlein et al., 1984) and the use of such reconstructions as an independent benchmark for paleoclimatic simulations produced by general circulation models (COHMAP Members, 1988; Webb et al., 1993a). Pollen-based paleoclimatic reconstructions continue (Bartlein et al., 2011) but must now take into account the effects of low glacial CO₂ concentrations on plant physiology (Wu et al., 2007) and the possibility that the realized climatic niche for species may shift somewhat over time (Pearman et al., 2008; Veloz et al., 2012). There has been a renewed interest in using pollen data as a paleovegetation proxy, for example, for reconstructing the past distributions of biomes, plant functional types, and leaf area indices (Gonzales et al., 2008; Prentice et al., 2011; Williams et al., 2011) and using these reconstructed vegetation maps for model validation.

In recent years, an increasing number and variety of paleoclimatic records have provided independent evidence of postglacial climate history, both globally and for NENA. This independent climatic framework is freeing pollen data from its role as a paleoclimatic indicator, analogous to radiocarbon dating having freed pollen from its stratigraphic role (Webb et al., 2004), and allows instead the testing of hypotheses regarding the climatic forcing of vegetational dynamics. Events interpreted from NENA pollen diagrams can be correlated to the abrupt climatic changes recorded in Greenland ice cores and elsewhere (Gonzales and Grimm, 2009; Peteet, 1995; Shuman et al., 2002; Yu, 2007). Unfortunately, intersite correlations are hampered by imprecise age constraints: most NENA records were collected prior to 1980 and were dated using conventional radiocarbon methods on bulk sediment samples, which are prone to incorporation of 'old' radiocarbon-depleted carbon from groundwater sources and can be inaccurate by thousands of years (Grimm et al., 2009). Improvements in radiocarbon dating have come from AMS radiocarbon dating techniques (with a 1 σ error typically <200 years) and careful selection of materials for dating. Cross validation analyses of pollen stratigraphic events among well-dated 'benchmark' pollen records indicate that spatial interpolations of these events carried an uncertainty of about 500 years (Blois et al., 2011b). Multiproxy analyses at single sites permit direct assessment of vegetational responses to local climatic change, with minimal stratigraphic correlation uncertainty (Booth et al., 2004; Shuman et al., 2004).

In addition to the new opportunities opened up by multiproxy studies, exciting new directions involve the intersection of traditional pollen-based interpretations of vegetation history with new lines of evidence and new theories. Contemporary patterns of genetic diversity now provide a complementary line of evidence about postglacial migrational rates and pathways and the locations of full-glacial refugia (Jaramillo-Correa et al., 2009; Soltis et al., 2006). New evidence for accelerated rates of evolutionary response to twenty-first-century climate change, evolutionary adaption in invasive species, and within-species adaptation of populations to local environments is forcing a reappraisal of the standard paradigm that evolutionary responses to late Quaternary climate change were minimal

(Davis et al., 2005; Soltis et al., 2006). The long records of ecological dynamics provided by subfossil pollen records offer the opportunity to test theories based on observation of contemporary ecosystems and their variation along spatial environmental gradients. For example, Clark and McLachlan (2003) used eight pollen records in eastern North America to argue that Holocene forest composition and plant abundance were more stable than predicted by the Hubbell theory of neutral ecological drift (Hubbell, 2001).

NENA Vegetational History

The timing and pattern of response to late Quaternary climate change and the retreat of the Laurentide Ice Sheet varied widely among plant taxa (Figure 1), although most arboreal taxa expanded in range and abundance. Cold-tolerant taxa such as spruce, pine, birch, and alder (*Picea*, *Pinus*, *Betula*, and *Alnus*) colonized areas vacated by the Laurentide Ice Sheet, and temperate taxa such as oak, hemlock, maple, chestnut, and beech (*Quercus*, *Tsuga*, *Acer*, *Castanea*, and *Fagus*) expanded from glacial refugia in the southern Appalachians, southern Coastal Plain, and Florida. However, postglacial vegetation history was not a simple northward expansion, nor was change limited to the end Pleistocene/early Holocene transition. Taxa such as pine, sage, and ragweed (*Pinus*, *Artemisia*, and *Ambrosia*) include significant east–west shifts, whereas other taxa, such as ash, hophornbeam/hornbeam, and elm (*Fraxinus*, *Ostrya/Carpinus*, and *Ulmus*) apparently peaked in abundance during the Late Glacial and early Holocene intervals and were relatively uncommon thereafter (Williams and Jackson, 2007). During the Holocene, variations in moisture availability apparently resulted in widespread impacts upon the position and character of the prairie–forest ecotone and the composition of eastern forests (Shuman et al., 2004; Williams et al., 2009). Over the last 1000 years, spruce has re-expanded into northern New England and the northern Great Lakes, and beech and hemlock abundances have declined, presumably in response to regional cooling.

Mapped sequences of pollen percentages for five taxa – spruce, pine, hophornbeam/hornbeam, and ragweed (Figure 1) – illustrate the individualistic responses of plant taxa to postglacial environmental change. The scale of these maps emphasizes vegetation change at subcontinental scales, although finer-scale phenomena are apparent as well. At this scale, pollen abundances usually are positively correlated with the regional abundance of the parent plant taxon (Solomon and Webb, 1985), so here, we use the maps of pollen abundance to infer past changes in plant distributions, while recognizing that the underlying pollen–vegetation relationships are nonlinear and regionally variable (Prentice and Webb, 1986). A second set of maps (Figure 2) shows rates of palynological change, emergence and disappearance of novel plant communities during the Late Glacial interval, and changes in vegetation physiognomy. Additional map series may be viewed in animated form at <http://www.ncdc.noaa.gov/paleo/pollen/viewer/webviewer.html> (Williams et al., 2004).

Spruce (*Picea*)

In eastern North America, spruce (*Picea*) pollen represents three species: white spruce (*Picea glauca*), black spruce (*Picea*

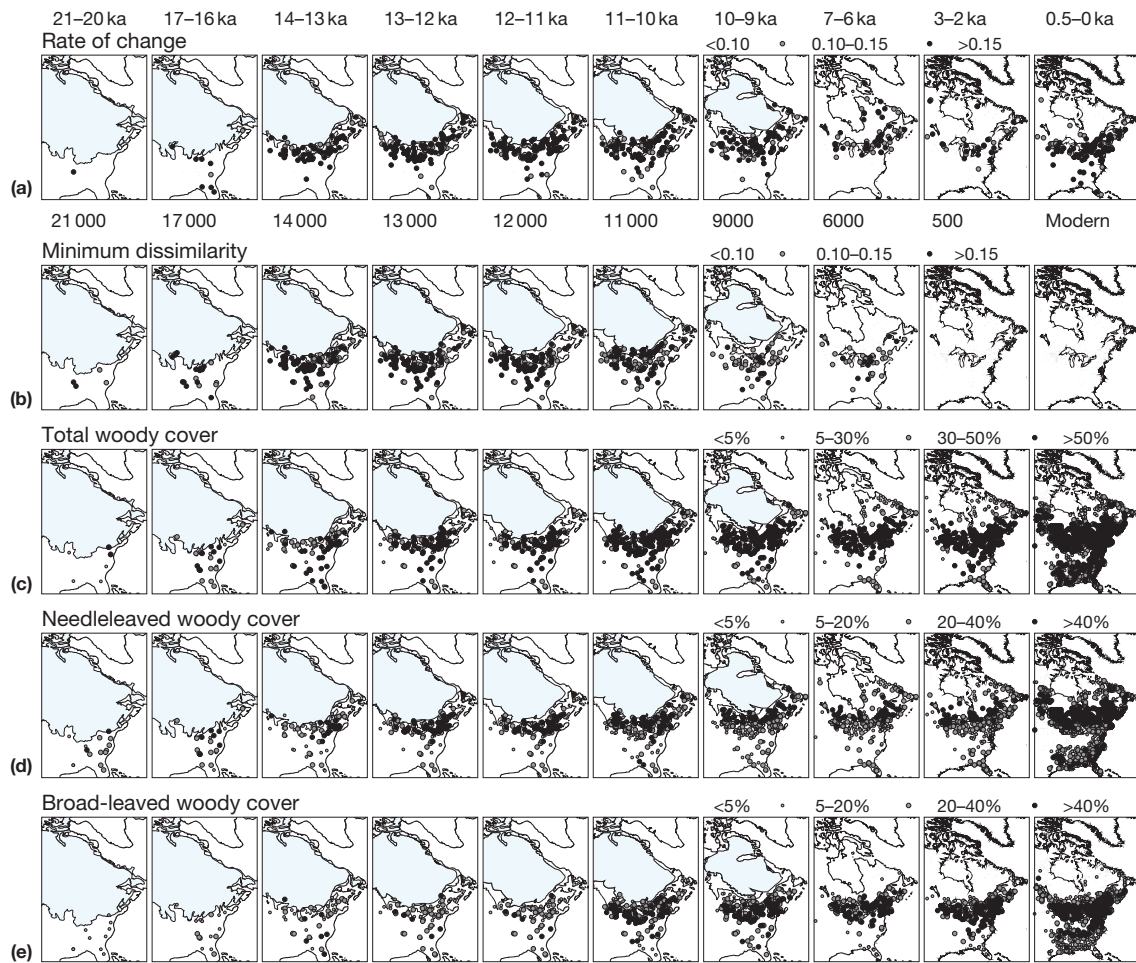


Figure 2 Maps of (a) palynological rate of change, (b) minimum dissimilarity from present, and the fractional areal cover for (c) woody plants, (d) needle-leaved woody plants, and (e) broad-leaved woody plants. Rates of change and minimum dissimilarities are computed using the squared-chord dissimilarity metric. Rates of change represent the within-site dissimilarities between pollen assemblages from adjacent time intervals and hence are an indicator of local rates of community turnover. The minimum dissimilarity analyses show the dissimilarity between a fossil pollen sample and its closest analog found from an all-North American dataset of surface pollen samples (Williams et al., 2001), and hence, large values indicate fossil samples with no close analog in modern pollen samples. Estimates of woody cover are calculated by application of modern-analog approaches to pollen data and remotely sensed indices of vegetation cover (Williams, 2003; Williams et al., 2011).

mariana), and red spruce (*Picea rubens*). A fourth species (*Picea critchfieldii*) apparently went extinct during the last deglaciation (Jackson and Weng, 1999). Spruce pollen is usually identified to the level of genus, although careful morphological measurements do permit species-level identifications (Lindbladh et al., 2003).

Spruce pollen (Figure 1(a)) documents the northward expansion of cold-tolerant arboreal taxa into formerly glaciated regions. At the Last Glacial Maximum, spruce pollen was most abundant in the continental interior but is common to all full-glacial NENA records. There is minimal overlap between its present and full-glacial range, indicating that spruce dynamics have been governed by both colonization into deglaciated regions and extirpation of trailing-edge populations. Some spruce populations appear to have persisted continuously in refugia (Anderson et al., 2006; Schauffler and Jacobson, 2002). Spruce pollen percentages were moderate to high across NENA from 21 to 12 ka, but declined between 12 and 9 ka as pine expanded westward and high spruce abundances became

limited to a narrow zone between pine and the Laurentide Ice Sheet. During this period, spruce rapidly migrated northwest into the corridor between the Laurentide and Cordilleran Ice Sheets (Williams et al., 2004). Early estimates of rates of *Picea* spread were as high as 2 km year^{-1} (Ritchie and MacDonald, 1986) but have been recently revised downward to 0.3 km year^{-1} (Yansa, 2006). After the collapse of the Laurentide Ice Sheet ca. 8.2 ka, spruce expanded across eastern Canada. In the late Holocene, spruce has re-expanded into interior New England, due to an increase in red spruce and a decrease in white spruce (Lindbladh et al., 2003).

Pine (*Pinus*)

Pine pollen, like spruce, was widely abundant at the Last Glacial Maximum and documents the expansion of pine species northward and westward inland into deglaciated areas (Figure 1(b)). Exactly which species is unclear; there are roughly a dozen species of pine in eastern North America, all

indistinguishable palynologically except for eastern white pine (*Pinus strobus*) because it is the sole eastern representative of *Pinus* subgenus *Strobus*. Pine pollen was present throughout NENA between 18 and 21 ka but was most abundant in the east. Between 17 and 13 ka, pine pollen abundances dropped to trace levels in the upper Midwest, and pine is absent from macrofossil records (Jackson et al., 1997), indicating that pine was regionally rare or absent. Pine then rapidly expanded westward across the Great Lakes region and upper Midwest into central Canada. Note that the apparent westward expansion of pine into the Midwest at 13 ka in Figure 1b is probably too early, due to inaccurate radiocarbon dates for some pollen records (Grimm et al., 2009). New AMS chronologies from Crystal Lake, IL, and Appleman Lake, IN, place the pine rise around 11.7 ka (Gill et al., 2009; Gonzales and Grimm, 2009). During the early Holocene, the northern limit of pine closely tracked the northward retreat of the Laurentide Ice Sheet, reaching its current position by 8 ka. Pine pollen abundances continued to increase in northern Canada during the Holocene.

Hophornbeam/Hornbeam (*Ostrya/Carpinus*)

Several NENA pollen types from temperate deciduous hardwood trees, notably hophornbeam/hornbeam (Figure 1(c)), ash, and elm, are characterized by peak pollen percentages during the Late Glacial interval and early Holocene and relatively low abundances during the middle and late Holocene. (*Ostrya* and *Carpinus* are genera in Betulaceae that are similar palynologically and are usually mapped as a single pollen type.) These taxa are important constituents of the Late Glacial and early Holocene 'no-analog' pollen assemblages (Williams et al., 2007) and during the Late Glacial interval had closely overlapping distributions with northern conifers such as spruce and larch (*Larix*). The Late Glacial/early Holocene peak abundances correspond to greater-than-present seasonality of insolation and temperature (Delcourt and Delcourt, 1994; Williams et al., 2007), suggesting that hophornbeam/hornbeam and ash flourished under warm but short summers and were able to tolerate cold winters that excluded other thermophilous taxa. Despite these variations in abundance, the northern limits for hophornbeam/hornbeam and ash were stable during the Late Glacial interval and expanded north only slightly following the retreat of the Laurentide Ice Sheet. *Ulmus*, conversely, apparently experienced both a range expansion from the southern United States into the upper Midwest, arriving by 16 ka, and a peak in pollen abundances between 12 and 6 ka.

Beech (*Fagus*)

Many temperate arboreal taxa (e.g., beech (Figure 1(d)), hemlock, maple, lime (*Tilia*), eastern white pine, oak, chestnut, and hickory (*Carya*)) that are now common in the temperate deciduous and cool mixed forests of NENA (Küchler, 1975) were regionally rare during the Last Glacial Maximum. These taxa migrated into NENA from refugia in the southern Appalachians, southern Coastal Plain, and lower Mississippi Valley (Davis, 1983; McLachlan et al., 2005), with significant differences among taxa in the migrational path and timing of arrival. Beech is a useful example because its pollen type is associated with a single species, American beech (*F. grandifolia*). Previous pollen

data syntheses have suggested that beech full-glacial refugia were located on the southern Coastal Plain or lower Mississippi Valley and that beech migrated north into New England along a coastal route east of the Appalachians (Davis, 1981b). Low beech pollen abundances suggest arrival in New England as early as 13 ka (Davis, 1969). Beech abundances begin to rise around 9 ka, several thousand years after the hemlock rise (Davis, 1969; Oswald et al., 2007), despite the currently close association between beech and hemlock. Beech pollen is found in eastern Wisconsin by 6 ka, about 2000 years after its occurrence in western Michigan, suggesting that Lake Michigan acted as a barrier to migration (Davis et al., 1986; Webb, 1987).

These pollen-based interpretations are being reexamined in light of phylogeographic analyses of extant tree populations in eastern North America, and there are some unresolved differences between the palynological and molecular data. The spatial pattern of genetic variation in beech suggests that (1) additional beech refugia existed in eastern North Carolina and perhaps as far north as northern Kentucky and (2) migration into the Midwest was primarily from the south, not the east (McLachlan et al., 2005). The phylogeographic evidence for such cryptic refugia thus suggests that fossil-based estimates of migration rate (e.g., 200 m year⁻¹ for *Fagus*) (Davis, 1983) are too high, although the current molecular-based migration rates (<100 m year⁻¹) (McLachlan et al., 2005) are based on highly conservative assumptions about starting position and duration of migration.

Ragweed (*Ambrosia*)

The mapped series of ragweed (Figure 1(e)) illustrates shifts in herbaceous abundances in the Midwest during the Holocene, which were driven by changes in moisture availability and, in the last 500 years, the effects of European land use. *Ambrosia* has many species and is a common indicator of unforested to semi-forested vegetation, whether due to aridity or anthropogenic land clearance (Grimm, 2001; McAndrews, 1988). Ragweed was present in the northern Great Plains since the Last Glacial Maximum but did not become regionally abundant until 11 ka. Ragweed abundances peak between 9 and 6 ka. Ecophysiological evidence suggests that the ragweed peak may have been caused by drier-than-present winters that amplified the ecological sensitivity to variable levels of summer precipitation (Grimm, 2001). A high-resolution record from Kettle Lake, ND, shows that ragweed abundances repeatedly fluctuated during the middle Holocene with a periodicity of 130 years (Clark et al., 2002). In New England, a peak in ragweed pollen abundances between 10.1 and 7.7 ka indicates more arid climates and increased forest openness (Faison et al., 2006).

Most modern agricultural plant species are insect-pollinated or have limited dispersal distances (e.g., maize (*Zea mays*)) and so are not well represented in lake and mire pollen records. Primary evidence of recent human land use, therefore, comes from changes in the abundance of weedy species. The dramatic increase in ragweed after 0.5 ka (Figure 1(e)) provides a particularly clear signal of European settlement and land clearance today, and the *Ambrosia* rise is widely used as an age control in the upper portion of pollen records (McAndrews, 1988). Other plant taxa displaying post-European increases in pollen abundance include the grass family (Poaceae), sorrel (*Rumex*),

and plantain (*Plantago*). Pollen records with clear evidence of Native American land clearance and burning are scarce and suggest that Native Americans had significant but localized impacts upon NENA vegetation (Ekdahl et al., 2004; Muñoz and Gajewski, 2010; Parshall and Foster, 2002). Some areas of known agricultural activity by Native Americans, such as the Mississippi River Valley, have relatively few palynological studies (Ollendorf, 1993).

Rates of Change

Rates of palynological change in NENA were high between 17 and 8 ka, with the fastest rates of change between 13 and 11 ka (Figure 2(a)). Late Glacial rates of change were high in the entire region. The rate-of-change maps reflect widespread local turnover in the identity of dominant species and the reorganization of late Pleistocene plant communities into their Holocene counterparts (Jacobson et al., 1987; Shuman et al., 2005; Williams et al., 2001). The timing and identity of the taxa involved in the community reorganizations varied regionally (Williams et al., 2004). After 9 ka, rates of palynological change decreased, although plant community composition continued to change regionally. Between 7 and 6 ka, for example, rates of change were high for Midwestern sites as the prairie–forest border shifted eastward, in New England and New York as pine declined and hemlock and deciduous forest taxa expanded, and in northern Québec and Labrador as spruce pollen abundances increased and sedge (Cyperaceae) and willow (*Salix*) abundances decreased, indicating a northward shift in treeline.

No-Analog Plant Associations and Biomes

Many Late Glacial pollen assemblages are compositionally unlike all modern pollen assemblages in North America (Figure 2(b)). These ‘no-analog’ pollen assemblages emerge from the individualistic behavior of plant taxa and indicate plant associations not present on the modern landscape (Cushing, 1965; Williams and Jackson, 2007). These assemblages are characterized by a mixture of pollen from cold-tolerant conifers (spruce and larch), deciduous taxa (ash, hophornbeam/hornbeam, hazel, cottonwood, and willow), and herbaceous taxa (sedges and sage) and are unusual for the mixture of boreal and temperate arboreal taxa, higher-than-present pollen abundances of ash and/or hophornbeam/hornbeam (Figure 1(c)), and the scarcity of pine, alder, and birch pollen. The palynological dissimilarity scores are a conservative index of the distribution of no-analog species associations because some species are not well represented in pollen records and pollen morphology typically restricts taxonomic identifications to the genus level, so that some novel species-level associations likely are obscured (Jackson and Williams, 2004).

Equilibrium niche theory predicts that these plant communities should have formed in response to environmental conditions also not found at present (Jackson and Overpeck, 2000; Williams et al., 2001), and paleoclimatic simulations suggest that the no-analog communities grew in areas with higher-than-present seasonality of insolation and temperature (Williams and Jackson, 2007). However, pollen-based reconstructions using the expanded response surface (ERS) technique, which attempts to extend pollen–climate relationships beyond the limits of

the modern climate space, do not indicate higher-than-present seasonality (Gonzales et al., 2009). This discrepancy between paleoclimatic simulations and pollen-based paleoclimatic reconstructions is not yet resolved; one possibility is that the ERS method may not fully account for shifts in species’ realized niches during periods of no-analog climates (Veloz et al., 2012). Recent work suggests that the no-analog plant communities also may be linked to the late Pleistocene megafaunal extinctions, perhaps because the megafaunal population declines released some hardwood taxa from herbivore suppression and shifted ecosystems from an herbivore-dominated to fire-dominated disturbance regime (Gill et al., 2009; Robinson et al., 2005).

Tree Cover

Pollen-based reconstructions of woody cover, based on calibrations of modern pollen data with remotely sensed vegetation datasets, show a widespread expansion in forest cover in NENA since the Last Glacial Maximum (Figures 2(c–e)) (Williams, 2003). At the Last Glacial Maximum, open forests and woodlands prevailed to the east and steppelike conditions to the west. Closed forests developed first along the eastern seaboard and expanded westward between 14 and 11 ka, then northward until 9 ka. Late Glacial forests were initially dominated by needle-leaved trees, with mixed deciduous–conifer forests established by 12 ka and broad-leaved deciduous forests established between 11 and 9 ka. The current gradient in decreasing tree cover from the boreal forest to tundra in Québec appears to have been in place by 6 ka and has been stable since then. The vegetational impacts of European land use are manifested as a widespread decrease in woody cover after 0.5 ka, particularly for broad-leaved woody taxa.

Climatic Controls on New England Vegetation History

New England offers a particularly detailed case study of the climatic drivers of postglacial vegetation history. In Deevey’s (1939) original zonation, New England pollen stratigraphies were divided into three zones: a lowermost A zone, characterized by high abundances of spruce pollen and other coniferous taxa, indicating the development of boreal forests soon after deglaciation; an intermediate B zone, characterized by a maximum in pine pollen abundances, and a C zone, characterized by high abundances of oak pollen and other temperate tree taxa. The C zone has three subzones: C-1, in which oak and hemlock abundances are high; C-2, characterized by low abundances of hemlock and rising hickory abundances; and C-3, in which hemlock abundances recover and chestnut abundances also rise (Deevey, 1939). This sequence of events is replicated in pollen records throughout southern New England.

Independent age controls and paleoclimatic records now make it possible to match Deevey’s vegetation zones to regional and hemispheric climatic events. The period of high spruce abundances is now known to date to between 13 and 11.6 ka in southern New England (Figure 3) and is linked to a return to cooler conditions in the North Atlantic and adjacent

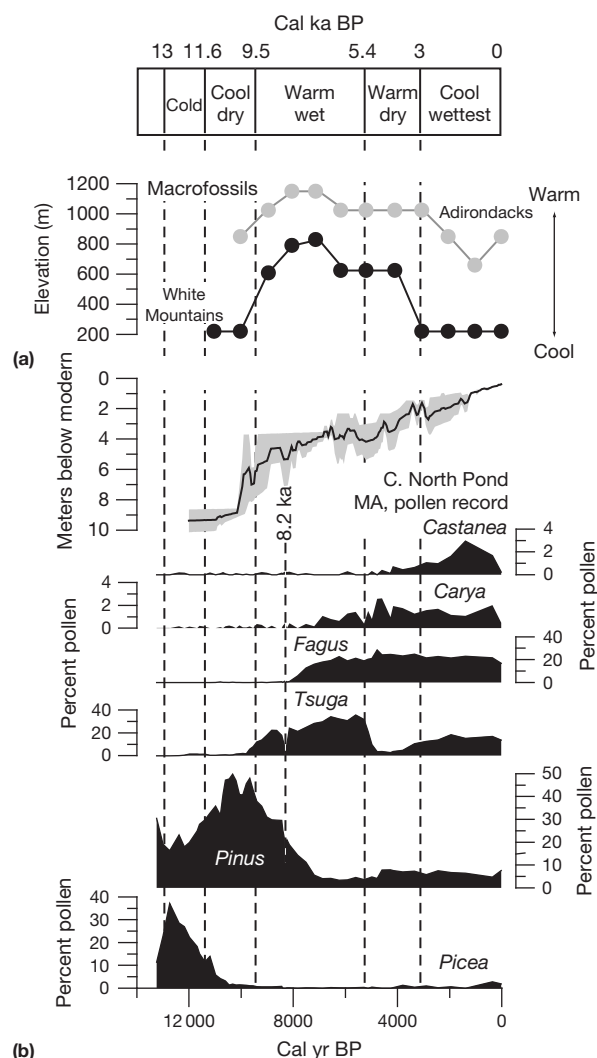


Figure 3 Evidence for the climatic control of postglacial vegetation history in New England at orbital to centennial timescales, based on fossil pollen records and nearby independent paleoclimatic records. (a) Maximum elevation of *Pinus strobus*. (b) Davis Pond, MA, water level. In panel a, plant macrofossil records of the upper elevational limit of Eastern white pine (*Pinus strobus*) provide a paleotemperature indicator (Jackson and Whitehead, 1991; Spear et al., 1994), while in panel b, a lake level reconstruction from Davis Pond, MA, provides a paleohydrological proxy (Newby et al., 2011). Changes in these paleoclimatic proxies correspond closely to major vegetation events in the pollen record from North Pond, MA (Whitehead and Crisman, 1978).

areas during the Younger Dryas Chronozone (Oswald et al., 2007; Peteet et al., 1993; Shuman et al., 2009). Lake level records indicate that the period of maximum pine abundances occurred during an interval of cool and dry conditions, dated between 11.6 and 8.2 ka (Figure 3). The pine maximum was time-transgressive, occurring first in southern New England and propagating north (Shuman et al., 2009) and at many sites was followed by a peak in ragweed abundances (Faison et al., 2006). Modern vegetation gradients were established between 9.5 and 8 ka (Oswald et al., 2007).

During the Holocene, centennial-scale climate variability profoundly affected forest composition. The most striking event is the mid-Holocene collapse in hemlock populations, marked by a regionwide and abrupt decline in pollen abundances. At some sites, this collapse occurred in less than 10 years (Allison et al., 1986). The apparent synchronicity and rapidity of the hemlock decline, and apparent lack of an effect on other tree taxa initially led paleoecologists to suspect that the mortality was caused by a pest or pathogen outbreak (Davis, 1981a). This hypothesis was reinforced by finds of hemlock needles with damage consistent with hemlock looper and head capsules from hemlock looper and spruce budworm, all in paludified dune field sediments dating to this period (Bhiry and Filion, 1996). Although a pest outbreak remains a viable mechanism for hemlock mortality, recent evidence has suggested that hydrological variability and drought may have been the ultimate causal agents (Foster et al., 2006; Shuman et al., 2001; Yu et al., 1997). Key lines of evidence include (1) the demonstration that oak abundances in southern coastal New England collapsed at the same time as the regionwide hemlock decline, suggesting a common climatic driver (Foster et al., 2006) and (2) evidence for lowered lake levels during the interval of low hemlock abundances (Figure 3) (Shuman et al., 2001). Recent analyses of lake sediments at New Long Pond, MA, suggest a series of repeated centennial-scale droughts (Newby et al., 2009), whose onset coincides closely with the hemlock decline (Shuman et al., 2009). Transitions between New England archaeological phases also appear to correspond to these periods of vegetation and climate change (Muñoz et al., 2010). Thus, the detailed evidence from New England strongly indicates that vegetation change is directly linked to climate change, both with respect to millennial-scale shifts in vegetation composition and more abrupt, decadal- to centennial-scale vegetation events (Shuman et al., 2009).

Summary

NENA has been a major locus of paleovegetational research during the past six decades, and the resultant dense network of fossil pollen records has enabled detailed maps and other syntheses of the spatial and temporal patterns of vegetation history. Such maps reveal the now-classic observation that plant taxon responses to climate change were highly individualistic, with each taxon displaying a unique trajectory in its shifting distribution over time. Rates of change were fastest between 13 and 8 ka and plant associations with no modern analog were common until 12 ka. Vegetation changes were primarily driven by climate change but influenced also by other factors such as rising CO₂ concentrations, the late Pleistocene megafaunal extinctions, and human arrival. In New England, local- to regional-scale syntheses of paleovegetational and paleoclimatic time series offer strong evidence of climatic controls on vegetation history. Major research directions at present include the generation of new high-resolution and well-dated paleovegetational records, the establishment of independent paleoclimatic records to provide an independent climatic framework for interpreting paleovegetational data, and the synthesis of paleoecological data with new biogeographic and evolutionary theories and data.

See also: **Archaeological Records:** Global Expansion 300 000–8000 Years Ago, Americas. **Lake Level Studies:** North America. **Paleoclimate:** The Younger Dryas Climate Event. **Plant Macrofossil Records:** Holocene North America.

References

- Allison TD, Moeller RE, and Davis MB (1986) Pollen in laminated sediments provides evidence of mid-Holocene forest pathogen outbreak. *Ecology* 67: 1101–1105.
- Anderson LL, Hu FS, Nelson DM, Petit RJ, and Paige KN (2006) Ice-age endurance: DNA evidence of a white spruce refugium in Alaska. *Proceedings of the National Academy of Sciences* 103: 12447–12450.
- Bartlein PJ, Harrison SP, Brewer S, et al. (2011) Pollen-based continental climate reconstructions at 6 and 21 ka: A global synthesis. *Climate Dynamics* 37: 775–802.
- Bartlein PJ, Webb T III, and Fleri E (1984) Holocene climate change in the northern Midwest: Pollen-derived estimates. *Quaternary Research* 22: 361–374.
- Bhury N and Filion L (1996) Mid-Holocene hemlock decline in eastern North America linked with phytophagous insect activity. *Quaternary Research* 45: 312–320.
- Blois J, Goring S, and Smith A (2011a) Integrating paleoecological databases. *Eos* 92: 48.
- Blois JL, Williams JW, Grimm EC, Jackson ST, and Graham RW (2011b) A methodological framework for improved paleovegetation mapping from late-Quaternary pollen records. *Quaternary Science Reviews* 30: 1926–1939.
- Booth RK, Jackson ST, and Gray CED (2004) Paleoecology and high-resolution paleohydrology of a kettle peatland in upper Michigan. *Quaternary Research* 61: 1–13.
- Bostock H (1970) Physiographic regions of Canada. In: Douglas RJW (ed.) *Geology and Economic Minerals of Canada*, pp. 10–30. Ottawa, ON: Geological Survey of Canada Economic Geology Report Number 1.
- Braun EL (1950) *Deciduous Forests of Eastern North America*. New York: Blakiston.
- Clark JS, Grimm EC, Donovan JJ, Fritz SC, Engstrom DR, and Almendinger JE (2002) Drought cycles and landscape responses to past aridity on prairies of the Northern Great Plains, USA. *Ecology* 83: 595–601.
- Clark JS and McLachlan JS (2003) Stability of forest biodiversity. *Nature* 423: 635–638.
- COHMAP Members (1988) Climatic changes of the last 18,000 years: Observations and model simulations. *Science* 24: 1043–1052.
- Cowling SA (1999) Simulated effects of low atmospheric CO₂ on structure and composition of North American vegetation at the Last Glacial Maximum. *Global Ecology and Biogeography Letters* 8: 81–93.
- Cushing EJ (1965) Problems in the Quaternary phytogeography of the Great Lakes region. In: Wright HE Jr. and Frey DG (eds.) *The Quaternary of the United States*, pp. 403–416. Princeton: Princeton University Press.
- Davis MB (1969) Climatic changes in southern Connecticut recorded by pollen deposition at Rogers Lake. *Ecology* 50: 409–422.
- Davis MB (1976) Pleistocene biogeography of temperate deciduous forests. *Geoscience and Man* XIII: 13–26.
- Davis MB (1981a) Outbreaks of forest pathogens in Quaternary history. *Proceedings of the Fourth International Palynological Conference*, pp. 216–227. Lucknow, India: Birbal Sahni Institute of Palaeobotany.
- Davis MB (1981b) Quaternary history and the stability of forest communities. In: West DC, Shugart HH, and Botkin DB (eds.) *Forest Succession*, pp. 132–177. New York: Springer-Verlag.
- Davis MB (1983) Quaternary history of deciduous forests of eastern North America and Europe. *Annals of the Missouri Botanical Garden* 70: 550–563.
- Davis MB, Shaw RG, and Etterson JR (2005) Evolutionary responses to changing climate. *Ecology* 86: 1704–1714.
- Davis MB, Woods KD, Webb SL, and Futyma RP (1986) Dispersal versus climate: Expansion of *Fagus* and *Tsuga* into the Upper Great Lakes region. *Plant Ecology* 67: 93–103.
- Deevey ES Jr. (1939) Studies on Connecticut lake sediments. I. A postglacial climatic chronology for southern New England. *American Journal of Science* 237: 691–724.
- Delcourt HR and Delcourt PA (1994) Postglacial rise and decline of *Ostrya virginiana* (Mill.) K Koch and *Carpinus caroliniana* Walt in eastern North America: Predictable responses of forest species to cyclic changes in seasonality of climates. *Journal of Biogeography* 21: 137–150.
- Delcourt PA, Delcourt HR, and Webb T III (1984) *Atlas of Mapped Distributions of Dominance and Modern Pollen Percentages for Important Tree Taxa of Eastern North America*. Dallas, TX: American Association of Stratigraphic Palynologists Foundation.
- Denevan WM (1992) The pristine myth: The landscape of the Americas in 1492. *Annals of the Association of American Geographers* 82: 369–385.
- Ekdhall EJ, Teranes JL, Guilderson TP, et al. (2004) Prehistorical record of cultural eutrophication from Crawford Lake, Canada. *Geology* 32: 745–748.
- Faison EK, Foster DR, Oswald WW, Hansen BCS, and Doughty E (2006) Early Holocene openlands in southern New England. *Ecology* 87: 2537–2547.
- Fenneman NM and Johnson DW (1946) *Physiographic Divisions of the Conterminous US*. Washington, DC: US Geological Survey.
- Foster DR (2002) Insights from historical geography to ecology and conservation: Lessons from the New England landscape. *Journal of Biogeography* 29: 1269–1275.
- Foster DR, Oswald WW, Faison EK, Doughty ED, and Hansen BCS (2006) A climatic driver for abrupt mid-Holocene vegetation dynamics and the hemlock decline in New England. *Ecology* 87: 2959–2966.
- Gaudreau DC and Webb T III (1985) Late-Quaternary pollen records and isochrone maps for the northeastern United States. In: Bryant VM Jr. and Holloway RG (eds.) *Pollen Records of Late-Quaternary North American Sediments*, pp. 245–280. Dallas, TX: American Association of Stratigraphic Palynologists Foundation.
- Gill JL, Williams JW, Jackson ST, Lininger K, and Robinson GS (2009) Pleistocene megafaunal collapse preceded novel plant communities and enhanced fire regimes. *Science* 326: 1100–1103.
- Gonzales LM and Grimm EC (2009) Synchronization of late-glacial vegetation changes at Crystal Lake, Illinois, USA with the North Atlantic Event Stratigraphy. *Quaternary Research* 72: 234–245. <http://dx.doi.org/10.1016/j.yqres.2009.05.001>.
- Gonzales LM, Williams JW, and Grimm EC (2009) Expanded response-surfaces: A new method to reconstruct paleoclimates from fossil pollen assemblages that lack modern analogues. *Quaternary Science Reviews* 28: 3315–3332.
- Gonzales LM, Williams JW, and Kaplan JO (2008) Variations in leaf area index in northern and eastern North America over the past 21,000 years: A data-model comparison. *Quaternary Science Reviews* 27: 1453–1466.
- Grimm EC (2001) Trends and palaeoecological problems in the vegetation and climate history of the northern Great Plains, USA. *Biology and Environment: Proceedings of the Royal Irish Academy* 101B: 47–64.
- Grimm EC and Jacobson GL Jr. (2004) Late-Quaternary vegetation history of the eastern United States. In: Gillespie AR, Porter SC, and Atwater BF (eds.) *The Quaternary Period in the United States*, pp. 381–402. Amsterdam: Elsevier.
- Grimm EC, Keltner J, Cheddadi R, et al. (2013) Pollen databases and their application. In: Elias SA (ed.) *Encyclopedia of Quaternary Science*, 2nd edn. Amsterdam: Elsevier.
- Grimm EC, Lozano-Garcia S, Behling H, and Markgraf V (2001) Holocene vegetation changes in the Americas. In: Markgraf V (ed.) *Interhemispheric Climate Linkages*, pp. 325–370. San Diego: Academic Press.
- Grimm EC, Maher LJ Jr., and Nelson DM (2009) The magnitude of error in bulk-sediment radiocarbon dates from central North America. *Quaternary Research* 72: 301–308.
- Hubbell SP (2001) *The Unified Neutral Theory of Biodiversity and Biogeography*. Princeton, NJ: Princeton University Press.
- Jackson ST and Overpeck JT (2000) Responses of plant populations and communities to environmental changes of the late Quaternary. *Paleobiology* 26(supplement): 194–220.
- Jackson ST, Overpeck JT, Webb T III, Keatts SE, and Anderson KH (1997) Mapped plant-macrofossil and pollen records of late Quaternary vegetation change in eastern North America. *Quaternary Science Reviews* 16: 1–70.
- Jackson S and Weng C (1999) Late Quaternary extinction of a tree species in eastern North America. *Proceedings of the National Academy of Sciences* 96: 13847–13852.
- Jackson ST and Whitehead DR (1991) Holocene vegetation patterns in the Adirondack Mountains. *Ecology* 72: 641–654.
- Jackson ST and Williams JW (2004) Modern analogs in Quaternary paleoecology: Here today, gone yesterday, gone tomorrow? *Annual Review of Earth and Planetary Sciences* 32: 495–537.
- Jacobson GL Jr., Webb T III, and Grimm EC (1987) Patterns and rates of vegetation change during the deglaciation of eastern North America. In: Ruddiman WF and Wright HE Jr. (eds.) *North America and Adjacent Oceans During the Last Deglaciation*, pp. 277–288. Boulder: Geological Society of America.
- Jaramillo-Correa JP, Beaulieu J, Khaza DP, and Bousquet J (2009) Inferring the past from the present phylogeographic structure of North American forest trees: Seeing the forest for the genes. *Canadian Journal of Forest Research* 39: 286–307.
- Küchler AW (1975) *Potential Natural Vegetation of the Conterminous United States*. New York: American Geographical Society.
- Lindbladh M, Jacobson GL Jr., and Schauffler M (2003) The postglacial history of three *Picea* species in New England, USA. *Quaternary Research* 59: 61–69.

- McAndrews JH (1966) *Postglacial History of Prairie, Savanna, and Forest in Northwestern Minnesota*. Durham, NC: Torrey Botanical Club.
- McAndrews JH (1988) Human disturbances of North American forests and grasslands: The fossil pollen record. In: Huntley B and Webb T III (eds.) *Vegetation History*. Dordrecht: Kluwer Academic Publishers.
- McLachlan JS, Clark JS, and Manos PS (2005) Molecular indicators of tree migration capacity under rapid climate change. *Ecology* 86: 2088–2098.
- Muñoz SE and Gajewski K (2010) Distinguishing prehistoric human influence on late-Holocene forests in southern Ontario, Canada. *The Holocene* 20: 967–981.
- Muñoz SE, Gajewski K, and Peros MC (2010) Synchronous environmental and cultural change in the prehistory of the northeastern United States. *Proceedings of the National Academy of Sciences* 107: 22008–22013.
- Newby PE, Donnelly JP, Shuman BN, and MacDonald D (2009) Evidence of centennial-scale drought from southeastern Massachusetts during the Pleistocene/Holocene transition. *Quaternary Science Reviews* 28: 1675–1692.
- Newby PE, Shuman BN, Donnelly JP, and MacDonald D (2011) Repeated century-scale droughts over the past 13,000 yr near the Hudson River watershed, USA. *Quaternary Research* 75: 523–530.
- Ollendorf AL (1993) *Changing landscapes in the American Bottom (USA): An interdisciplinary investigation with an emphasis on the late-prehistoric and early-historic periods*. PhD Dissertation. University of Minnesota.
- Oswald WW, Faison EK, Foster DR, Doughty ED, Hall BR, and Hansen BCS (2007) Post-glacial changes in spatial patterns of vegetation across southern New England. *Journal of Biogeography* 34: 900–913.
- Parshall T and Foster DR (2002) Fire on the New England landscape: Regional and temporal variation, cultural and environmental controls. *Journal of Biogeography* 29: 1305–1317.
- Pearman PB, Guisan A, Broennimann O, and Randin CF (2008) Niche dynamics in space and time. *Trends in Ecology & Evolution* 23: 149–158.
- Petee D (1995) Global Younger Dryas? *Quaternary International* 28: 93–104.
- Petee DM, Daniels RA, Heusser LE, Vogel JS, Southon JR, and Nelson DE (1993) Late-glacial pollen, macrofossils and fish remains in northeastern U.S.A. – The Younger Dryas oscillation. *Quaternary Science Reviews* 12: 597–612.
- Potzger JE (1946) Phytosociology of the primeval forest in central-northern Wisconsin and upper Michigan, and a brief post-glacial history of the Lake Forest Formation. *Ecological Monographs* 16: 211–250.
- Prentice IC, Harrison SP, and Bartlein PJ (2011) Global vegetation and terrestrial carbon cycle changes after the last ice age. *New Phytologist* 189: 988–998.
- Prentice IC and Webb T III (1986) Pollen percentages, tree abundances and the Fagerlind effect. *Journal of Quaternary Science* 1: 35–43.
- Richard PJH (1993) The origin and postglacial dynamics of the mixed forest in Quebec. *Review of Palaeobotany and Palynology* 79: 31–68.
- Ritchie JC (1987) *Postglacial Vegetation of Canada*. Cambridge: Cambridge University Press.
- Ritchie JC and MacDonald GM (1986) The patterns of post-glacial spread of white spruce. *Journal of Biogeography* 13: 527–540.
- Robinson GS, Burney LP, and Burney DA (2005) Landscape paleoecology and megafaunal extinction in southeastern New York State. *Ecological Monographs* 75: 295–315.
- Rowe JS (1972) Forest Regions of Canada. Pub. no. 1300, Canadian Forestry Service, Department of Fisheries and the Environment, Ottawa.
- Schauffler M and Jacobson GL Jr. (2002) Persistence of coastal spruce refugia during the Holocene in northern New England, USA, detected by stand-scale pollen stratigraphies. *Journal of Ecology* 90: 235–250.
- Shuman BN, Bartlein PJ, Logar N, Newby P, and Webb T III (2002) Parallel climate and vegetation responses to the early Holocene collapse of the Laurentide Ice Sheet. *Quaternary Science Reviews* 21: 1793–1805.
- Shuman B, Bartlein PJ, and Webb T III (2005) The magnitudes of millennial and orbital-scale climatic change in eastern North America during the Late Quaternary. *Quaternary Science Reviews* 24: 2194–2206.
- Shuman BN, Bravo J, Kaye J, Lynch JA, Newby P, and Webb T III (2001) Late Quaternary water-level variations and vegetation history at Crooked Pond, southeastern Massachusetts. *Quaternary Research* 56: 401–410.
- Shuman BN, Newby P, and Donnelly JP (2009) Abrupt climate change as an important agent of ecological change in the northeast U.S. throughout the past 15,000 years. *Quaternary Science Reviews* 28: 1693–1709.
- Shuman BN, Newby PC, Huang Y, and Webb T III (2004) Evidence for the close climatic control of New England vegetation history. *Ecology* 85: 1297–1310.
- Solomon A and Webb T III (1985) Computer-aided reconstruction of late-Quaternary landscape dynamics. *Annual Review of Ecology and Systematics* 16: 63–84.
- Soltis DE, Morris AB, McLachlan JS, Manos PS, and Soltis PS (2006) Comparative phylogeography of unglaciated eastern North America. *Molecular Ecology* 15: 4261–4293.
- Spear RW, Davis MB, and Shane LCK (1994) Late Quaternary history of low- and mid-elevation vegetation in the White Mountains of New Hampshire. *Ecological Monographs* 64: 85–109.
- Velloz S, Williams JW, He F, Liu Z, and Otto-Bliesner B (2012) No-analog climates and shifting realized niches during the late Quaternary: Implications for 21st-century predictions by species distribution models. *Global Change Biology* 18: 1698–1713.
- Webb T III (1974) Corresponding patterns of pollen and vegetation in lower Michigan: A comparison of quantitative data. *Ecology* 55: 17–28.
- Webb SL (1987) Beech range extension and vegetation history – Pollen stratigraphy of two Wisconsin lakes. *Ecology* 68: 1993–2005.
- Webb T III (1988) Glacial and Holocene vegetation history: Eastern North America. In: Huntley B and Webb T III (eds.) *Vegetation History*, pp. 385–414. Dordrecht: Kluwer Academic.
- Webb RS, Anderson KH, and Webb T III (1993a) Pollen response-surface estimates of late-Quaternary changes in the moisture balance of the northeastern United States. *Quaternary Research* 40: 213–227.
- Webb T III, Bartlein PJ, Harrison SP, and Anderson KH (1993b) Vegetation, lake levels, and climate in eastern North America for the past 18,000 years. In: Wright HE Jr., Kutzbach JE, Webb T III, Ruddiman WF, Street-Perrott FA, and Bartlein PJ (eds.) *Global Climates Since the Last Glacial Maximum*, pp. 415–467. Minneapolis, MN: University of Minnesota Press.
- Webb T III, Shuman BN, and Williams JW (2004) Climatically forced vegetation dynamics in North America during the late Quaternary period. In: Gillespie AR, Porter SC, and Atwater BF (eds.) *The Quaternary Period in the United States*, pp. 459–478. Amsterdam: Elsevier.
- Whitehead DR and Crisman TL (1978) Paleolimnological studies of small New England (U.S.A.) ponds. Part I. Late-glacial and postglacial trophic oscillations. *Polskie Archiwum Hydrobiologii* 25: 471–481.
- Whitmore J, Gajewski K, Sawada M, et al. (2005) North American and Greenland modern pollen data for multi-scale paleoecological and paleoclimatic applications. *Quaternary Science Reviews* 24: 1828–1848.
- Williams JW (2003) Variations in tree cover in North America since the Last Glacial Maximum. *Global and Planetary Change* 35: 1–23.
- Williams JW and Jackson ST (2007) Novel climates, no-analog communities, and ecological surprises. *Frontiers in Ecology and the Environment* 5: 475–482.
- Williams JW, Jackson ST, and Kutzbach JE (2007) Projected distributions of novel and disappearing climates by 2100 AD. *Proceedings of the National Academy of Sciences* 104: 5738–5742.
- Williams JW, Shuman B, and Bartlein PJ (2009) Rapid responses of the Midwestern prairie-forest ecotone to early Holocene aridity. *Global and Planetary Change* 66: 195–207.
- Williams JW, Shuman B, Bartlein PJ, et al. (2006) *An Atlas of Pollen-Vegetation-Climate Relationships for the United States and Canada*. Dallas, TX: American Association of Stratigraphic Palynologists Foundation.
- Williams JW, Shuman BN, and Webb T III (2001) Dissimilarity analyses of late-Quaternary vegetation and climate in eastern North America. *Ecology* 82: 3346–3362.
- Williams JW, Shuman BN, Webb T III, Bartlein PJ, and Leduc PL (2004) Late Quaternary vegetation dynamics in North America: Scaling from taxa to biomes. *Ecological Monographs* 74: 309–334.
- Williams JW, Tarasov PA, Brewer S, and Notaro M (2011) Late-Quaternary variations in tree cover at the northern forest-tundra ecotone. *Journal of Geophysical Research – Biogeosciences* 116: G01017. <http://dx.doi.org/10.1029/2010JG001458>.
- Wu H, Guiot J, Brewer S, Guo Z, and Peng C (2007) Dominant factors controlling glacial and interglacial variations in the treeline elevation in tropical Africa. *Proceedings of the National Academy of Sciences* 104: 9720–9724.
- Yansa CH (2006) The timing and nature of late Quaternary vegetation changes in the northern Great Plains, USA and Canada: A re-assessment of the spruce phase. *Quaternary Science Reviews* 25: 263–281.
- Yu Z (2007) Rapid response of forested vegetation to multiple climatic oscillations during the last deglaciation in the northeastern United States. *Quaternary Research* 67: 297–303.
- Yu Z, Andrews JH, and Eicher U (1997) Middle Holocene dry climate caused by change in atmospheric circulation patterns: Evidence from lake levels and stable isotopes. *Geology* 25: 251–254.

Relevant Websites

<http://www.ncdc.noaa.gov/paleo/paleo.html> – NOAA Paleoclimatology Data Portal.
<http://www.ncdc.noaa.gov/paleo/pollen/viewer/webviewer.html> – Pollen Viewer.
www.neotomadb.org – Neotoma Paleocology Database.

Northwestern North America

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Introduction

The vegetation history of northwestern North America (NWN) following the last glaciation is marked by several major transitions that are consistent with a rich and complex climate history. The region today supports very heterogeneous vegetation corresponding to broad-scale gradients in moisture availability and temperature and fine-scale topography in mountainous areas (Figure 1). This heterogeneity presents a major challenge for understanding the regional vegetation history because local geomorphic and climatic conditions can modify species responses to climate change at any particular site (Whitlock, 1992). Nevertheless, some common patterns have emerged from more than five decades of paleoecological studies in this region.

Our aim in this article is not to present a comprehensive summary of the wide range of findings from the hundreds of pollen records from NWN, much of which would reiterate previous articles (Table 1). Rather, we aim to highlight robust findings from a subset of records representative of broad vegetation zones. This discussion focuses on the area north of 42° N and west of the continental divide. We first describe the broad climatic gradients and the corresponding ecosystems across NWN. We then summarize the major vegetational changes based on key pollen profiles from several subregions. The history of each subregion is discussed with respect to three major climatic periods: the Last Glacial Maximum (LGM, broadly defined as the interval 25–18 ka; ka, thousands of years before present), the last glacial–interglacial transition (14.5–11.6 ka), and the Holocene (11.6–0 ka). All ages are based on calibrated radiocarbon dates.

Broad Patterns of Climate and Vegetation

The dominant westerly flow of moisture results in a major longitudinal moisture gradient due to coastal mountain ranges that produce heavy orographic precipitation to the west and a rain shadow to the east. Thus, the primary vegetation gradient is characterized by a maritime moist conifer forest west of the crest of the coastal ranges and a mixture of sagebrush steppe, mixed conifer, and pine forests east of the coastal ranges. In the southern half of NWN, little rainfall occurs from July to September, due to the dominance of a subtropical high-pressure system centered over the Pacific. This results in a seasonally dry maritime forest along the coast in Oregon and Washington, and very dry shrub steppe sagebrush in the interior. Further to the east, beyond the continental divide, summer moisture increases from monsoon-like precipitation. In the northern half of NWN and generally north of the subtropical high pressure, increasing onshore flow results in a hypermaritime coastal climate dominated by dense forests of

shade-tolerant conifers up to 60° N. In the high-snow areas along the coast, treeline declines from 1500 m in western Washington to 500 m in south-central Alaska. East of the coastal ranges, summer moisture increases with latitude due to increased summer convective precipitation, especially in mountainous areas. From south-central British Columbia northward, the inland vegetation transitions from sagebrush steppe to pine forests, then to subboreal and boreal spruce forest (i.e., the Cold Desert to Western Cordillera to Boreal Cordillera transition on Figure 1.). An exception to this overall pattern occurs to the west of the continental divide from northern Idaho to central British Columbia, where high precipitation results in a moist interior forest, often called the interior wet belt, dominated by *Tsuga heterophylla* (western hemlock) and *Thuja plicata* (western redcedar) that are disjunct from their main coastal distribution (Gavin, 2009). In central Alaska, the boreal forest is mainly composed of black and white spruce, with aspen and birch on warm and disturbed sites. In general, treeline in the interior forests is hundreds of meters higher and declines at a slower rate with latitude compared to the maritime forests. The forest limit in Alaska abuts the southern flanks of the Brooks Range and is limited westward toward the cold-maritime climates near the Bering Sea. For detailed accounts of vegetation–climate relationships in this region, we refer readers to detailed monographs focused on Oregon and Washington (Franklin and Dyness, 1988), British Columbia (Meidinger and Pojar, 1991), or Alaska (Gallant et al., 1995; Viereck et al., 1992).

Alaska

This section should be read in conjunction with [Northern North America](#).

Last Glacial Maximum

Despite the fact that much of the region was unglaciated during the last glaciation, most of the Alaskan pollen records do not extend beyond the past 14 000 years. At all of the sites where full-glacial sediments have been recovered, the pollen spectra of the LGM are dominated by Poaceae, Cyperaceae, and *Artemisia* (Eisner and Colinvaux, 1990; Livingstone, 1955, 1957; Oswald et al., 1999). These pollen assemblages are difficult to interpret because of their broad ecological tolerances (e.g., Cwynar, 1982). However, minor taxa such as *Encalypta rhabdocarpa* and *Selaginella siberica* suggest that sparse, xeric tundra prevailed under the cold, dry climate of the late Pleistocene.

An issue that remained controversial for several decades was whether some of the boreal-forest tree species existed in eastern Beringia (Alaska and adjacent Canada) during the LGM (Hopkins et al., 1981; Hultén, 1937). Based on the spatial

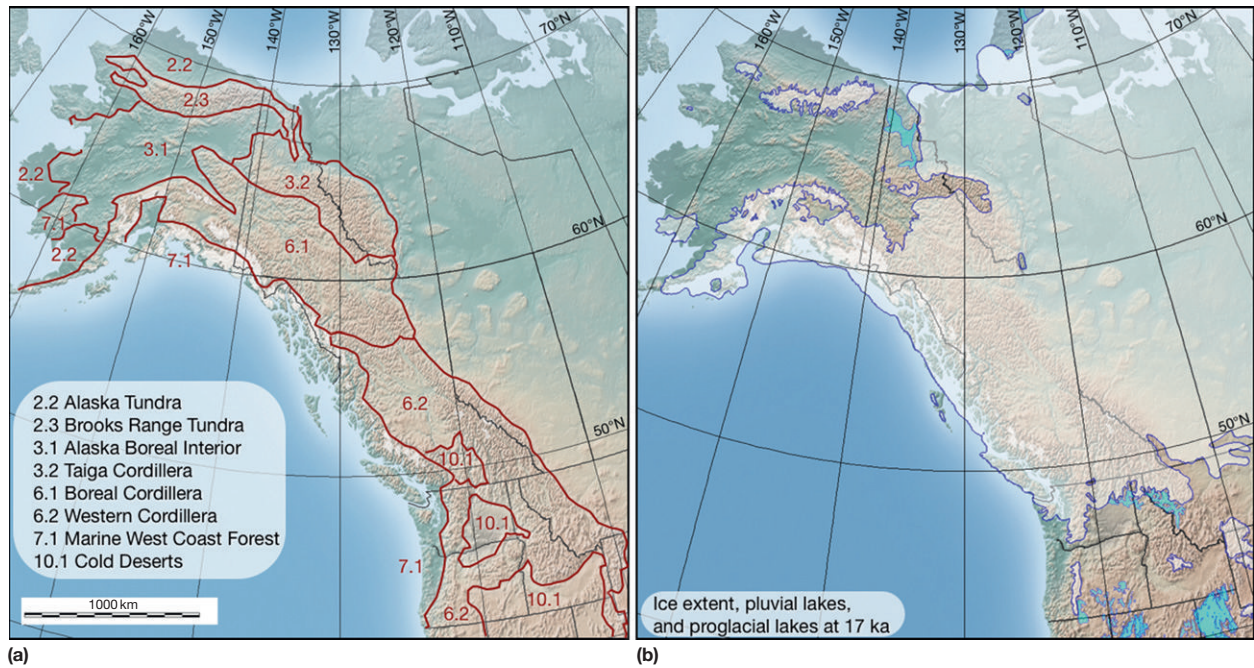


Figure 1 (a) Shaded-relief map of northwest North America using a color scheme that differentiates humid (green) and xeric (brown) regions. Ecoregions differentiate the major vegetation types and climatic regions. The map shows all eight ecoregions occurring within our area of focus from the ecoregions of North America (Commission for Environmental Cooperation, 1997). (b) Extent of the Cordilleran ice sheet, other ice fields, pluvial lakes, and proglacial lakes at 18 000 cal. years before present (Dyke et al., 2003).

Table 1 Previously published regional syntheses of paleovegetational change from northwest North America. This table is limited to papers published since 1990

Region	Citations
Alaska	Anderson and Brubaker (1994), Bigelow et al. (2003), Brubaker et al. (2005), Edwards et al. (2005)
Yukon	Cwynar and Spear (1995), Vermaire and Cwynar (2010)
Washington, Oregon, Idaho, and Western Montana	Thompson and Anderson (2000), Thompson et al. (2003), Whitlock (1992), Whitlock et al. (2008)
Columbia River Basin (Interior Washington, Idaho, and British Columbia)	Mehring (1996), Walker and Pellatt (2008)
British Columbia	Hebda (1995), Mathewes (1991)
Southern Vancouver Island	Brown and Hebda (2003)
Coastal forests	Hebda and Whitlock (1997)

patterns of individual pollen types, Brubaker et al. (2005) concluded that the common tree and shrub taxa in the Alaskan pollen records survived the last LGM in eastern Beringia. This includes *Picea* (spruce), a dominant taxon in the modern boreal forests, as evidenced by the fact that trace amounts of spruce pollen are common in the LGM sediments from a number of sites. This hypothesis cannot be tested rigorously using fossil data; spruce macrofossils dated to the LGM have not been found. However, a recent analysis of chloroplast DNA in extant trees revealed unique haplotypes and a higher diversity of haplotypes in Alaska than areas farther south (Anderson

et al., 2006), offering unambiguous evidence that spruce did survive the LGM in Alaska. Together, the DNA and pollen data suggest that tundra prevailed on the regional landscape during the LGM, but small populations of trees existed in isolated sheltered areas.

Transition from the Pleistocene to the Holocene

Between 15 and 12 ka, pollen records indicate that herb tundra transitioned to shrub tundra dominated by *Betula* (birch) in response to increases in temperature and effective moisture. *Betula*, *Salix* (willow), and *Cyperaceae* became the dominant taxa, and tundra vegetation cover became more continuous than previously. Associated with these vegetational changes are the stabilization of eolian dunes, aggradation of streams, and formation of organic peat deposits (Mann et al., 2002). A number of pollen diagrams from Alaska display prominent peaks of *Populus balsamifera*, suggesting that this species formed woodlands in at least some local habitats (Figure 2). A recent synthesis of the regional pollen diagrams shows a cluster of *Populus* peaks between 13 and 10 ka (Edwards et al., 2005). These are thought to be evidence of warmer-than-present climatic conditions during the early Holocene, presumably in response to peak values of summer insolation (Kaufman et al., 2004). This age range likely reflects that many of the pollen diagrams are based on bulk-sediment ¹⁴C dating, which tends to yield anomalously old dates. Several recent pollen records with AMS-¹⁴C ages on terrestrial plant macrofossils suggest that the *Populus* interval was post-Younger Dryas. Furthermore, *Populus* peaks occurred at different times at different sites, and midge-based temperature reconstructions from three sites in interior Alaska provide compelling evidence

that the early Holocene was not warmer, probably as a result of ocean–atmosphere–land interactions associated with the states of the Arctic Oscillation and El Niño Southern Oscillation (Clegg et al., 2011).

Pollen records from Alaska indicate that climate cooling during the Younger Dryas Chronozone (YDC; 12.9–11.6 ka) altered the species composition and decreased the cover of terrestrial vegetation in at least some areas of Alaska. Evidence includes decreased tree cover in southeastern Alaska (Engstrom et al., 1990) and the expansion of herb taxa at the expense of *Betula* shrubs in south-central and southwestern Alaska (Hu et al., 1995, 2002; Peteet and Mann, 1994). In northern Alaska, Mann et al. (2002) report a contraction in the range of *Populus balsamifera*, presumably in response to a return to cold and dry conditions. Other evidence supporting YDC-related climatic fluctuations in Alaska includes high-resolution analysis of biogenic silica (diatom remains) in lake sediments (Hu et al., 2002, 2006; Kaufman et al., 2010), geomorphic and paleoecological records of stream down-cutting, eolian dune activity, and reduced peat formation (Mann et al., 2002), and glacier fluctuations (Briner et al., 2002).

The Holocene

Pollen records from several sites with high-quality chronological control indicate the early Holocene expansion of *Populus balsamifera*. This expansion may have resulted from climatic warming and/or from the presence of extensive disturbed mineral soils during the early postglacial interval (Hu et al., 1993). In interior Alaska, *Betula* tree species such as *Betula neoalaskana*, a common component of early-successional stands on the modern landscape, may have also gained importance at this time. Although studies of modern pollen samples indicate that separation of *B. neoalaskana* pollen from shrub-*Betula* species is possible (Clegg et al., 2005; Edwards et al., 1991), few fossil-pollen studies have attempted to distinguish these species in Alaska (Brubaker et al., 1983; Tinner et al., 2008). However, existing pollen and macrofossil records of tree-*Betula* seeds reveal local presence no later than 9 ka (Hu et al., 1993; Tinner et al., 2006).

Pollen assemblages with a substantial amount of *Picea* began to occur as early as 10 ka in the eastern and southern Alaska (Figure 2). The first spruce species to expand in the early Holocene pollen assemblages is *Picea glauca* (white spruce), which probably formed woodland with tree birch between 10 and 9 ka in interior Alaska. Pollen profiles that separate *Picea* pollen to the species level indicate that both *P. glauca* and *P. mariana*, the two common species of today's Alaskan boreal forests, were both present on the landscape by 9 ka in eastern Alaska (Tinner et al., 2006). The expansion of white spruce appears to have occurred from the east to west (Anderson and Brubaker, 1994), and the rates of population increases exhibit major spatiotemporal patterns. In eastern Alaska, *Picea* pollen percentages reached near-modern values within 600 years of initial appearance. Several pollen records from central Alaska display a rise in *Picea* pollen percentages at about the same time as the eastern sites, but the abundance of *Picea* pollen remained substantially below modern abundance until 5 to 6 ka. In central Alaska, boreal forests dominated by *P. mariana*

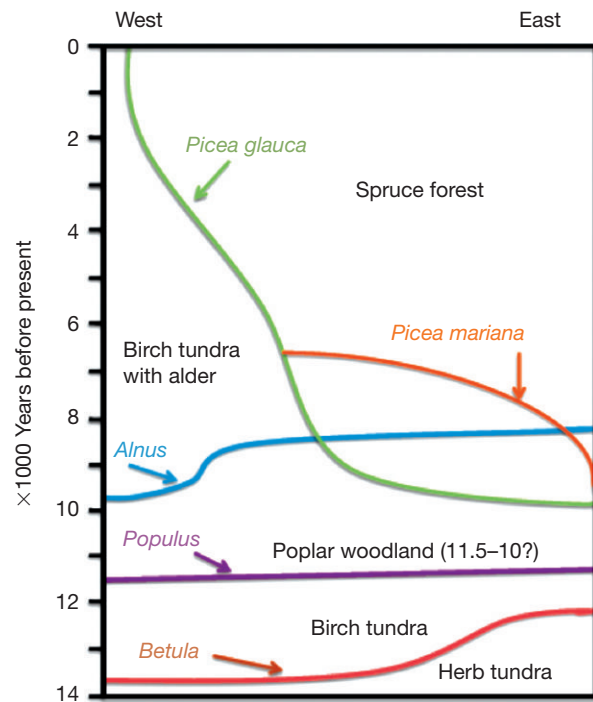


Figure 2 Schematic diagram showing the spatiotemporal dynamics of postglacial vegetation in Alaska, based on fossil-pollen analysis. Lines indicate the approximate timing of population expansions of select major taxa in the Alaskan pollen records.

similar to those we see today became established around 6.5 ka. Associated with this vegetational change was a pronounced increase in forest fire frequency, which was probably driven by the high flammability of *P. mariana* forests (Higuera et al., 2009; Lynch et al., 2003). Pollen percentages and accumulation rates in lake sediments suggest that the westward migration of sparse *Picea* stands halted around 8 ka, presumably because of an episode of climatic cooling (Hu et al., 1993). *Picea* then resumed its westward expansion to reach its modern western range limit during the late Holocene (Ager and Sims, 1981; Brubaker et al., 1983). A striking spatial pattern based on the dense network of Holocene pollen records from interior Alaska is that in contrast to the rapid spread across the eastern sites, spruce expanded extremely slowly from central to southwestern Alaska, and there is not a clear climatic or ecological explanation for this contrast (Kaltenrieder et al., 2011). As a result of the late expansion of spruce in southwestern Alaska, birch tundra was the prevalent vegetation type for $\geq 10,000$ years, until 3–4 ka when *P. glauca* invaded the region. A recent study showed that the abundance of birch shrubs fluctuated with climatic changes possibly associated with solar-irradiance fluctuations and ocean–atmosphere interactions at hemispheric scales (Hu et al., 2003).

The establishment of spruce forests similar to those of today occurred between 6.5 and 5 ka in interior Alaska (Anderson and Brubaker, 1994). *Picea* pollen percentages began to increase between 6 and 4 ka (Anderson and Brubaker, 1994) near the modern treelines in northwestern Alaska and around 4 ka in southwestern Alaska (Brubaker et al., 2001). Pollen abundance

reached modern values ~2–4 ka, suggesting that the modern forest-tundra ecotone became established during this interval (Anderson and Brubaker, 1994; Brubaker et al., 2001; Hu et al., 1995; Kaltenrieder et al., 2011). Spruce pollen never exceeded 10% of the terrestrial pollen sum during the late Holocene in these regions, suggesting that population densities have never been higher after the development of the modern tundra-forest ecotone in western Alaska. However, this assessment may be a result of the lack of high-temporal-resolution pollen records from western Alaska. The first detailed palynological study of lake sediments of the past 700 years revealed pronounced fluctuations of altitudinal treeline in the Copper River Basin of southeastern Alaska (Tinner et al., 2008). Tinner et al. (2008) conducted multiproxy analyses of lake sediments at decadal-to-centennial timescales to infer ecosystem and climatic fluctuation associated with the Little Ice Age (LIA; ~500–150 years ago). Their pollen data revealed diminished abundance of major tree taxa, including *P. glauca*, *P. mariana*, and *Betula neoalaskana*, as a result of decreases in temperature and effective moisture during the LIA. These changes were accompanied by the expansions of shrub and herbaceous species, including *Alnus viridis*, *Betula glandulosa/nana*, and *Epilobium*.

Alnus (alder) is a major taxon in the many Holocene pollen diagrams from Alaska, reaching 70–80% of the pollen spectra at some sites. The earliest expansion of alder occurred at sites in the northwestern Brooks Range around 9500 years ago, and alder appears to have expanded southward and eastward in interior Alaska. It became a major component of the pollen diagrams at many sites around 8000 cal BP, suggesting the rapid expansion of this taxon across of the landscape around this time. Because alder is a nitrogen fixer, its Holocene expansion probably markedly increased nitrogen availability and accelerated the rate of nitrogen cycling in Alaska, especially where alder is a prevalent component of the terrestrial plant communities such as southwestern Alaska. The N effects of the Holocene alder expansion have been documented with the analyses of lake sediments for biogenic silica, C/N ratio, and nitrogen and carbon isotopes (Hu et al., 2001), although it remains unknown how the increased N availability affected other species that are associated with alder.

In the pollen records from the North Slope of Alaska, *Picea* and *Alnus* are present in the Holocene sediments, but the abundance is low for both taxa. Spruce trees have never been locally present in this region based on the late Quaternary fossil records, and the presence of spruce in the pollen diagrams from the North Slope represents long-distance transport from areas south of the Brooks Range. In contrast, *Alnus* expanded in the Brooks Range and southern Arctic Foothills between 8 and 7 ka (Bergstrom, 1984; Eisner, 1991; Livingstone, 1955; Oswald et al., 1999, 2003). Tundra communities similar to those of today became established during the early–middle Holocene (Oswald et al., 1999, 2003). A detailed palynological study (Oswald et al., 2003) demonstrated that the trajectories of Holocene vegetation development differed between two major geomorphic substrates related to the Quaternary glacial history of the region. On the Sagavanirktok surface (>500,000 years old), relatively dry and sparse tundra dominated by prostrate shrubs transitioned to tussock tundra dominated by *Betula nana*,

Eriophorum vaginatum, Ericaceae species, and *Sphagnum* mosses (Walker et al., 1994) around 7500 years ago. In contrast, at a nearby site on the Itkillik II surface (~14,000 years old), pollen analysis revealed that prostrate-shrub tundra communities prevailed on the landscape since the early Holocene, although vegetation cover may have increased around 7500 years ago (Oswald et al., 2003). Oswald et al. attributed this vegetation-history contrast to differences in soil texture and geomorphology between the two geomorphic surfaces. Specifically, as effective moisture increased from early to middle Holocene, the fine-texture soils and gentle topography of the older Sagavanirktok surface promoted buildup of soil organic matter, shallowing of permafrost table, and development of water-logged soils, which would have favored the development of tussock tundra. These changes probably did not occur on the younger Itkillik II surface because of better drainage associated with coarser-texture and less-acidic soils.

Marine West Coast Forest

Last Glacial Maximum

A 7.7-m peat core from coastal western Washington, located south of the maximum extent of the Juan de Fuca lobe of the Cordilleran ice sheet, revealed a coarse vegetation history that may extend to the previous interglacial (Heusser et al., 1999). The full-glacial pollen assemblages are comprised mainly of *Pinus* (pine, mainly the shore-pine variant of *P. contorta*), *Tsuga mertensiana* (mountain hemlock), and Poaceae, suggesting cold open parkland, in contrast to dense *Picea sitchensis* (Sitka spruce)/*T. heterophylla* (western hemlock) at the site today. Similar pollen assemblages, but with more warm-adapted *T. heterophylla* pollen, were found in the low elevations of the coastal ranges of western Oregon and southwest Washington (Grigg and Whitlock, 2002; Grigg et al., 2001). In the unglaciated western Cascade Range of central Oregon, a debris flow deposit dated to >35 ka contained abundant macrofossils of tree species common today 1000 m higher in elevation, including *Picea engelmannii* (Engelmann spruce) and *Abies lasiocarpa* (subalpine fir). Further to the east, the High Cascades were covered by an ice sheet at the time (Figure 1). Taken together, these records indicate that forest cover occurred within 10s to 100s of kilometers from the ice sheet maximum, that subalpine forest types extended to low elevations, but also that warm-adapted conifer tree species persisted at some sites in Oregon. The most abundant warm-adapted species in the region today, *Pseudotsuga menziesii* (Douglas-fir), was very rare but likely present in Oregon during the last glacial (Gugger and Sugita, 2010).

Ice-free areas existed in certain locations north of the southern ice sheet margin during the LGM. In combination with lower sea level, many headlands along the British Columbia and Alaska coast remained ice-free allowing for persistence of plant and animal populations in refugia. Refugia have been identified on the basis of pollen and macrofossils on the Queen Charlotte Islands, Alexander Archipelago, and Vancouver Island (Carrara et al., 2007; Hebda, 1997; Warner et al., 1982). These findings are supported by phylogeographic studies (reviewed in Shafer et al., 2010) and findings of tree

macrofossils appearing in offshore postglacial sediments near Vancouver Island (Lacourse et al., 2003).

Transition from Pleistocene to Holocene

At 14.5 ka, a rapid warming initiated the retreat of many alpine glaciers, while at the same time, the Cordilleran ice sheet limit was moving northward. In the Oregon Coast Range, this transition involved a change from subalpine parkland to lowland tree species (*Alnus rubra* (red alder) and *Pseudotsuga menziesii*). The abrupt increase of *Pseudotsuga menziesii* is notable, given its rarity or absence during much of the glacial maximum (Grigg and Whitlock, 1998; Worona and Whitlock, 1995). Further north, from Puget Sound northward, the initial colonizer of deglaciated landscapes was almost entirely *Pinus contorta*, followed by a mix of subalpine species (*Abies*, *Picea*, and *Tsuga mertensiana*) between 14 and 13 ka. Here, the transition out of this parkland vegetation occurred later (at 12 to 11 ka) than at sites further south, typically close in time to the first major increase of *Pseudotsuga menziesii* (Cwynar, 1987).

The YDC (12.9–11.6 ka), as recorded by several nonvegetation climate proxies in this region, was a distinct cold and dry interval (Barron et al., 2003; Heine, 1998; Vacco et al., 2005), but its expression in the vegetation record is variable. Pollen records show little change in forests in western Oregon directly associated with the YDC, possibly reflecting the fact that warm-adapted vegetation was well established before the YDC and that many species (e.g., *Pseudotsuga*) have broad climatic tolerances (Grigg and Whitlock, 1998). From Vancouver Island northward, pollen assemblages from coastal sites show subalpine-adapted species (*Tsuga mertensiana*) and low pollen influxes coeval with the YDC (Brown and Hebda, 2002; Lacourse, 2009; Mathewes, 1993).

The Holocene

The early Holocene is marked by increased summer insolation (8% more relative to modern at 50°N) that strengthened the Pacific Subtropical high-pressure system, creating warm stable and dry air in coastal regions. Throughout the Pacific

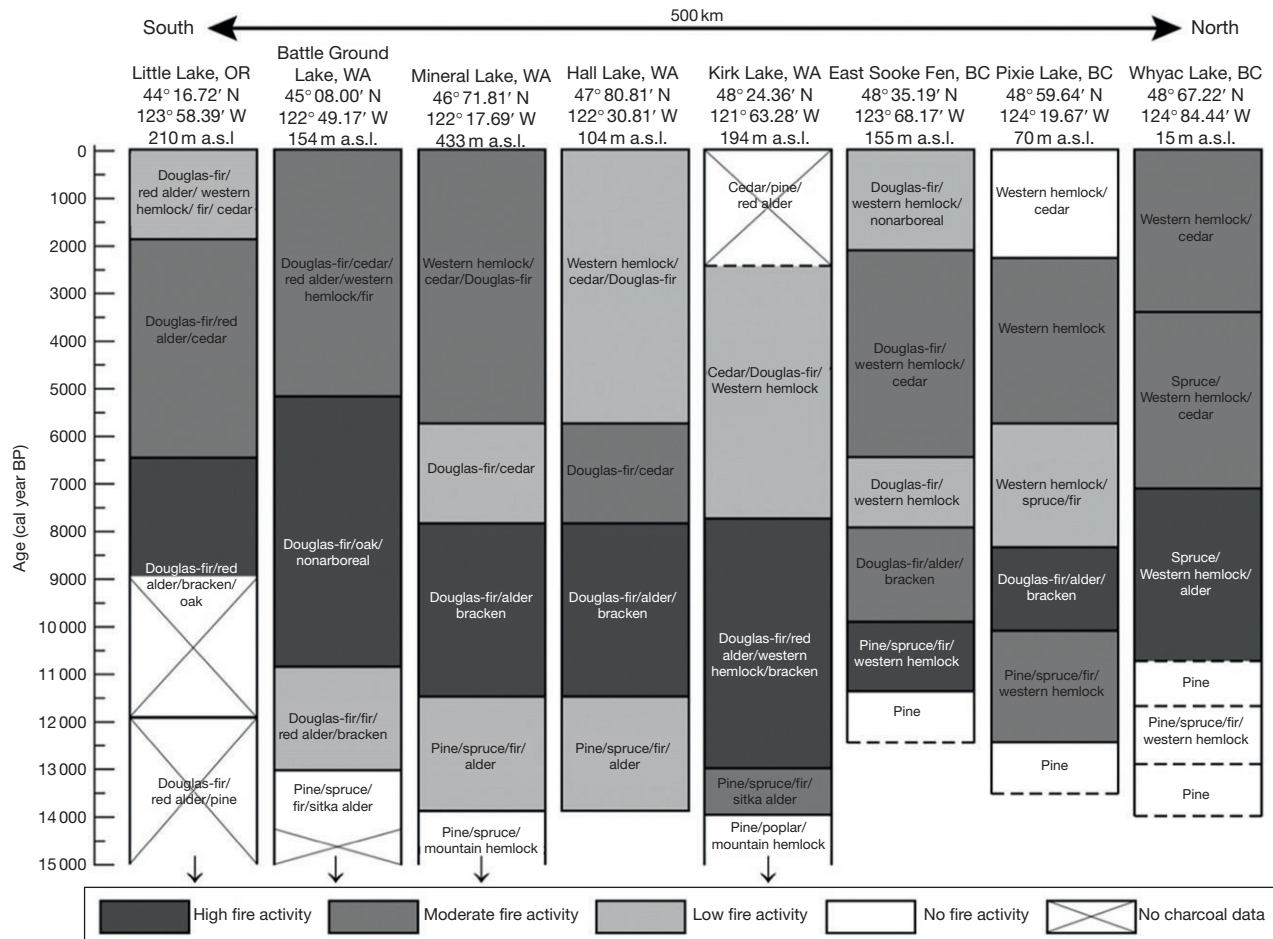


Figure 3 Pollen zones and inferred fire activity for eight low-elevation sites in the Marine West Coast Forest ecoregion from south (left) to north (right): Little Lake, OR (Long et al., 1998; Worona and Whitlock, 1995), Battle Ground Lake, OR (Barnosky, 1985; Walsh et al., 2008), Mineral Lake and Hall Lake, WA (Sugita and Tsukada, 1982), Kirk Lake, WA (Cwynar, 1987), East Sooke Fen, Pixie Lake, and Whyac Lake, BC (Brown and Hebda, 2002). Reproduced from Walsh MK, Whitlock C, Bartlein PJ (2008) A 14,300-year-long record of fire–vegetation–climate linkages at Battle Ground Lake, southwestern Washington. *Quaternary Research* 70: 251–264.

Northwest, the early Holocene is marked by hotter summers and increased drought and forest fire (Figure 3). Records from insect remain in lake sediments in southern British Columbia quantified July temperatures at 3 °C warmer than present (Gavin et al., 2011), in agreement with a temperature reconstruction based on pollen data from southeast Oregon (Minckley et al., 2007). In western Oregon and Washington, increased fire and summer warmth fostered widespread early-successional *Pseudotsuga*–*Alnus rubra* forests with abundant *Pteridium* spores indicating full-sun conditions after severe forest fires (Brown and Hebda, 2002; Sugimura et al., 2008). There is mounting evidence, however, that the early Holocene was not uniformly warm and dry, but was marked by distinct century-scale intervals of increased moisture. The remnants of the waning ice sheet to the north (which had its last major collapse at 8.5 ka) may have still been influencing the path of the jet stream across western North America (Gavin et al., 2011).

The transition out of the early Holocene was marked by increasing representation of shade-tolerant late-successional species (*Tsuga* and *Abies*), the decline of *Pseudotsuga* and *Alnus rubra*, and a decrease in fire. This transition began as early as 8 ka, but modern pollen assemblages were not reached until some time between 6 and 3 ka. The wide range in timing of the establishment of modern assemblages may be due to progressive development of ecosystem properties that favor acid-tolerating species such as *T. plicata*, a species that arrived comparatively recently north of Vancouver Island (Hebda and Mathewes, 1984). In southeast Alaska, the cooler and wetter climate was marked by several vegetation changes, including the expansion of *Tsuga* and development of paludified soils (Hansen and Engstrom, 1996). One reversal (warming) in the Neoglacial cooling trend was a period of increased fire found at several sites in British Columbia and Washington, beginning at 2.5 ka and possibly attributable to human-set fires (Hallett et al., 2003; Prichard et al., 2009; Sugimura et al., 2008).

Western Cordillera

Last Glacial Maximum

Few sediment records extend to the glacial maximum in the unglaciated terrain of eastern Washington and Oregon (Bartlein et al., 2010). In the rain shadow of the Cascade Range, near the modern forest–steppe boundary, the LGM vegetation was dominated by *Artemisia*, Poaceae, and herb taxa (Whitlock et al., 2000). Modeling of paleoclimate and species distributions over western North America for the glacial maximum is consistent with the inferences from the few existing geologic records: contraction of warm-adapted conifers (e.g., *Pseudotsuga*) and expansion of cold-adapted *Picea* and *Artemisia* (Bartlein et al., 1998).

Transition from Pleistocene to Holocene

Most of the Cordilleran ice sheet persisted in valley bottoms until the Holocene; therefore, few pollen records in inland British Columbia encompass the late Pleistocene. East of the Cascade Range, *Artemisia* and *Pinus* dominated the pollen assemblages, indicating a more open and drier vegetation type than occurs today. At Waitts Lake in northeastern Washington,

increased *Artemisia* coincided with the YDC, though *Artemisia* was already increasing prior to a tephra marker at 13.1 ka (Mack et al., 1978; Walker and Pellatt, 2008). Peak *Artemisia* during the YDC was followed by increased tree (*Pinus* and *Alnus*) pollen until 10.5 ka. Further south in Idaho and western Montana, a *Picea* parkland predominated at higher elevations before 13 ka, which changed to a closed pine forest during the YDC (Brunelle et al., 2005).

The Holocene

Increased warm-adapted taxa occur throughout the Western Cordillera during the early Holocene. At lower elevations, increased Poaceae and decreased tree pollen suggest an uphill retreat of the forest-grassland ecotone (Mack et al., 1978). In the mountains of central Idaho and western Montana, cool-adapted conifers (*Picea* and *Abies*) declined with increased *Pseudotsuga*/*Larix* pollen from 11.5 to 7.5 ka (Brunelle et al., 2005). In British Columbia, recently deglaciated land was usually colonized by *Pinus* cf. *contorta*, though pollen records indicate a fairly rapid primary succession to a more diverse forest (Gavin et al., 2011). Modern forest conditions generally developed by 4 ka with an increase in mesic taxa, such as *Picea* and *Abies* in subalpine forest sites.

An event during the early Holocene that deserves special mention is the eruption of Mount Mazama that created Crater Lake 7627 ± 150 years ago (Zdanowicz et al., 1999). The ash deposits from this eruption blanketed the northwest, especially east of the Cascade Range. This ash layer created the seed bed for today's forests and altered the trajectories of forest change. In some areas, extremely thick ash deposits led to 'tephra plains' that remain very dry and sparsely vegetated today. In other areas, ash deposits thickened soils and may have allowed for greater water retention. Colonization of forest on steep coarse rock talus in the mountains may have required the addition of tephra to function as a mulch and thus changed the character of the steeper parts of the landscapes (Mehring et al., 1977).

In wetter subzones of the Western Cordillera, forests contain many species disjunct from their main coastal distribution. Originally hypothesized to be the result of a Miocene vicariance (following Cascadian orogeny), this biogeographic pattern and its potential root in a northern Idaho refugium have received much focus from geneticists and paleoecologists (Carstens et al., 2005; Daubenmire, 1968). The emerging picture is that each species has a unique history, that many plant species dispersed during the Holocene from the coast, and that the modern mesic-forest composition in northern Idaho may be a very recent feature of the landscape (Gavin, 2009; Gavin and Hu, 2006; O'Connell et al., 2008; Shafer et al., 2010). Further north within the belt of mesic forests, the earliest detection of the modern forest dominant (*T. heterophylla*) occurred at 4.5 ka, coincident with the decline of *Betula* and transition from a continental to a cool and wet maritime climate (Gavin et al., 2011).

In the Western Cordillera, forests dominated by pine and mixed conifers support a complex mixed-severity fire regime. Vegetation and fire histories are of great importance for land managers who are setting restoration targets based on the historical range of variation in vegetation and disturbance

rates. Pollen and charcoal records from the Western Cordillera are summarized in several papers (Gavin et al., 2007; Marlon et al., 2009; Whitlock et al., 2008). Fires are a long-term presence from wet coastal forests to dry interior basins, and the Holocene vegetation history is more precisely interpreted alongside independent fire-history data. Some especially high-resolution studies have demonstrated a highly episodic history of fire, showing that any reference to disturbance regimes from the past 300 years alone does not describe the degree of variability that occurs over longer periods, including events of high severity and low frequency that may be considered highly undesirable today (Colombaroli and Gavin, 2010). Over the entire Holocene, the regional coherence of paleorecords indicates that increased fire during the early Holocene thermal maximum promoted disturbance-adapted vegetation. There is some evidence that east of the continental divide increased insolation resulted in increased summer moisture, more mesic vegetation, and reduced fire (Whitlock et al., 2008).

Cold Deserts

In general, the pollen assemblages from this region are dominated by broadly dispersed and highly productive pollen taxa. Thus, pollen records from these regions reflect broad areas. In addition, these drought-adapted taxa have wide ecological amplitudes, resulting in more stable populations through short-term climate change. However, the vegetation changes that are detected in these regions are concordant with the paleoclimatic inferences from the coastal region. For example, a well-studied core from Bear Lake confirms the increased extent of cold- and dry-adapted steppe taxa during glacial maxima (Jimenez-Moreno et al., 2007).

Pollen data from eastern Oregon and Washington indicate less dense vegetation than present at 14 ka (Mehring, 1996; Minckley et al., 2007). The pollen assemblages from the cold deserts show little sensitivity to the YDC, though all sites show increased herb taxa after 11 ka, indicating warmer and drier conditions. Decreased fire at this time likely reflects reduced biomass continuity (Minckley et al., 2007). After 7 ka, *Pinus* forest increased at high elevation sites and modern pollen assemblages were reached at most sites shortly thereafter.

Conclusions

The vegetation history of NWA is becoming more fully understood due to a large number of recent studies. The region is poised to provide information on postglacial vegetation dynamics similar to that available from eastern North America. Maps of taxa and community types through the late Quaternary may soon be possible in this region. This rich data set will further our understanding of the millennial-scale controls of vegetation change, while high-resolution pollen analysis has begun to reveal coherent decadal-to-century-scale changes. These latter efforts provide important information on ecological responses to rapid climate change.

The paleoecological record from the Cordillera provides an example of how populations persist on the landscape through periods of rapid climate change and how combinations of climate change and natural disturbances may rapidly

transform entire ecosystems. The existence of pockets of endemism today suggests a long-term history of isolation and persistence. Applying both genetic and fossil investigations into the history of patterns of disjunction and endemism is a powerful biogeographic approach to understand the origin of modern patterns of diversity. There remains much potential by merging these lines of evidence in the study of vegetation in the Pacific Northwest.

The paleoecological record can also inform managers in setting restoration target points. Large areas of dry forest in the Pacific Northwest have 'missed' several cycles of fire since the period of effective fire suppression. Disagreement on the character of the historical range of variability of fire frequency and fire severity exists despite decades of research on the topic. The sediment record, where it is of sufficiently high resolution, is a powerful tool to describe the historical range of variability in natural disturbances and its link to climate change.

See also: Pollen Analysis, Principles. **Chironomid Records:** Chironomid Overview. **Glaciations:** Late Quaternary in North America. **Paleobotany:** Paleophytogeography. **Plant Macrofossil Records:** Holocene North America. **Pollen Methods and Studies:** Changing Plant Distributions and Abundances. **Pollen Records, Late Pleistocene:** Northern North America.

References

- Ager TA and Sims JD (1981) Holocene pollen and sediment record from the Tangle Lake area, central Alaska. *Journal of Palynology* 5: 85–98.
- Anderson PM and Brubaker LB (1994) Vegetation history of northcentral Alaska – A mapped summary of late-Quaternary pollen data. *Quaternary Science Reviews* 13: 71–92.
- Anderson LL, Hu FS, Nelson DM, et al. (2006) Ice-age endurance: DNA evidence of a white spruce refugium in Alaska. *Proceedings of the National Academy of Sciences of the United States of America* 103: 12447–12450.
- Barnosky CW (1985) Late Quaternary vegetation near Battle Ground Lake, southern Puget Trough, Washington. *Geological Society of America Bulletin* 96: 263–271.
- Barron JA, Heusser L, Herbert T, and Lyle M (2003) High-resolution climatic evolution of coastal northern California during the past 16,000 years. *Paleoceanography* 18: 1020. <http://dx.doi.org/10.1029/2002PA000768>.
- Bartlein P, Anderson K, Anderson P, et al. (1998) Paleoclimate simulations for North America over the past 21,000 years: Features of the simulated climate and comparisons with paleoenvironmental data. *Quaternary Science Reviews* 17: 549–585.
- Bartlein PJ, Harrison SP, Brewer S, et al. (2010) Pollen-based continental climate reconstructions at 6 and 21 ka: A global synthesis. *Climate Dynamics* 37: 775–802.
- Bergstrom MF (1984) *Late Wisconsin and Holocene History of a Deep Arctic Lake, North-Central Brooks Range, Alaska*. M.S. Thesis, Ohio State University, Columbus, OH.
- Bigelow NH, Brubaker LB, Edwards ME, et al. (2003) Climate change and Arctic ecosystems: 1. Vegetation changes north of 55 degrees N between the last glacial maximum, mid-Holocene, and present. *Journal of Geophysical Research-Atmospheres* 108(D19): 8170. <http://dx.doi.org/10.1029/2002JD002558>.
- Briner J, Kaufman D, Werner A, et al. (2002) Glacier readvance during the late glacial (Younger Dryas?) in the Ahklun Mountains, southwestern Alaska. *Geology* 30: 679–682.
- Brown KJ and Hebda RJ (2002) Origin, development, and dynamics of coastal temperate conifer rainforests of southern Vancouver Island, Canada. *Canadian Journal of Forest Research* 32: 353–372.
- Brown KJ and Hebda RJ (2003) Coastal rainforest connections disclosed through a Late Quaternary vegetation, climate, and fire history investigation from the Mountain Hemlock Zone on southern Vancouver Island, British Columbia, Canada. *Review of Palaeobotany and Palynology* 123: 247–269.
- Brubaker LB, Anderson PM, Edwards ME, et al. (2005) Beringia as a glacial refugium for boreal trees and shrubs: New perspectives from mapped pollen data. *Journal of Biogeography* 32: 833–848.

- Brubaker LB, Anderson PM, and Hu FS (2001) Vegetation ecotone dynamics in Southwest Alaska during the Late Quaternary. *Quaternary Science Reviews* 20: 175–188.
- Brubaker LB, Garfinkel H, and Edwards M (1983) A late Wisconsin and Holocene vegetation history from the central Brooks Range – Implications for Alaskan paleoecology. *Quaternary Research* 20: 194–214.
- Brunelle A, Whitlock C, Bartlein P, et al. (2005) Holocene fire and vegetation along environmental gradients in the Northern Rocky Mountains. *Quaternary Science Reviews* 24: 2281–2300.
- Carrara PE, Ager TA, and Baichtal JF (2007) Possible refugia in the Alexander Archipelago of southeastern Alaska during the late Wisconsin glaciation. *Canadian Journal of Earth Sciences* 44: 229–244.
- Carstens BC, Brunfeldt SJ, Demboski JR, et al. (2005) Investigating the evolutionary history of the Pacific Northwest mesic forest ecosystem: Hypothesis testing within a comparative phylogeographic framework. *Evolution* 59: 1639–1652.
- Clegg BF, Kelly R, Clarke GH, et al. (2011) Nonlinear response of summer temperature to Holocene insolation forcing in Alaska. *Proceedings of the National Academy of Sciences of the USA* 108(48): 19299–19304.
- Clegg BF, Tinner W, Gavin DG, et al. (2005) Morphological differentiation of *Betula* (birch) pollen in northwest North America and its palaeoecological application. *The Holocene* 15: 229–237.
- Colombarelli D and Gavin DG (2010) Highly episodic fire and erosion regime over the past 2,000 y in the Siskiyou Mountains, Oregon. *Proceedings of the National Academy of Sciences of the United States of America* 107: 18909–18914.
- Commission for Environmental Cooperation (1997) *Ecological Regions of North America*. Montreal, Québec: Commission for Environmental Cooperation.
- Cwynar LC (1982) A late-Quaternary vegetation history from Hanging Lake, northern Yukon. *Ecological Monographs* 52: 1–24.
- Cwynar LC (1987) Fire and forest history of the North Cascade Range. *Ecology* 68: 791–802.
- Cwynar LC and Spear RW (1995) Paleovegetation and paleoclimatic changes in the Yukon at 6 ka bp. *Géographie physique et Quaternaire* 49: 29–35.
- Daubenmire R (1968) Soil moisture in relation to vegetation distribution in mountains of northern Idaho. *Ecology* 49: 431–438.
- Dyke AS, Moore A, and Robertson L (2003) Deglaciation of North America (Open File 1574). Geological Survey of Canada.
- Edwards ME, Brubaker LB, Lohzkin AV, et al. (2005) Structurally novel biomes: A response to past warming in Beringia. *Ecology* 86: 1696–1703.
- Edwards ME, Dawe JC, and Armbruster WS (1991) Pollen size of *Betula* in northern Alaska and the interpretation of late Quaternary vegetation records. *Canadian Journal of Botany* 69: 1666–1672.
- Eisner WR (1991) Palynological analysis of a peat core from Imnavait Creek, the North Slope, Alaska. *Arctic* 44: 279–282.
- Eisner WR and Colinvaux PA (1990) A long pollen record from Ahlhorak Lake, arctic Alaska. *Review of Palaeobotany and Palynology* 63: 35–52.
- Engstrom DR, Hansen BCS, and Wright HE Jr. (1990) A possible Younger Dryas record in southeastern Alaska. *Science* 250: 1383–1385.
- Franklin JE and Dyrness CT (1988) *Natural Vegetation of Oregon and Washington*. Oregon: Oregon State University Press.
- Gallant AL, Binnian EF, Omerik JM, et al. (1995) *Ecoregions of Alaska*. Washington, DC: United States Government Printing Office.
- Gavin DG (2009) The coastal-disjunct mesic flora in the inland Pacific Northwest of USA and Canada: Refugia, dispersal and disequilibrium. *Diversity and Distributions* 15: 972–982.
- Gavin DG, Hallett DJ, Hu FS, et al. (2007) Forest fire and climate change in western North America: Insights from sediment charcoal records. *Frontiers in Ecology and the Environment* 5: 499–506.
- Gavin DG, Henderson ACG, Westover K, et al. (2011) Abrupt Holocene climate change and potential response to solar forcing in western Canada. *Quaternary Science Reviews* 30: 1243–1255.
- Gavin D and Hu F (2006) Spatial variation of climatic and non-climatic controls on species distribution: The range limit of *Tsuga heterophylla*. *Journal of Biogeography* 33: 1384–1396.
- Grigg L and Whitlock C (1998) Late-glacial vegetation and climate change in western Oregon. *Quaternary Research* 49: 287–298.
- Grigg LD and Whitlock C (2002) Patterns and causes of millennial-scale climate change in the Pacific Northwest during Marine Isotope Stages 2 and 3. *Quaternary Science Reviews* 21: 2067–2083.
- Grigg LD, Whitlock C, and Dean WE (2001) Evidence for millennial-scale climate change during marine isotope stages 2 and 3 at Little Lake, western Oregon, USA. *Quaternary Research* 56: 10–22.
- Gugger PF and Sugita S (2010) Glacial populations and postglacial migration of Douglas-fir based on fossil pollen and macrofossil evidence. *Quaternary Science Reviews* 29: 2052–2070.
- Hallett DJ, Lepofsky DS, Mathewes RW, et al. (2003) 11 000 years of fire history and climate in the mountain hemlock rain forests of southwestern British Columbia based on sedimentary charcoal. *Canadian Journal of Forest Research* 33: 292–312.
- Hansen B and Engstrom D (1996) Vegetation history of Pleasant Island, southeastern Alaska, since 13,000 yr BP. *Quaternary Research* 46: 161–175.
- Hebda RJ (1995) British Columbia vegetation and climate history with focus on 6 ka BP. *Géographie physique et Quaternaire* 49: 55–79.
- Hebda RJ (1997) Late Quaternary paleoecology of Brooks Peninsula. *Brooks Peninsula: An Ice Age Refugium on Vancouver Island*, pp. 1–48. Victoria, BC: Ministry of Environment, Lands and Parks and BC Parks.
- Hebda RJ and Mathewes RW (1984) Holocene history of cedar and native Indian cultures of the North American Pacific Coast. *Science* 225: 711–713.
- Hebda RJ and Whitlock C (1997) Environmental history. In: Schoonmaker PK (ed.) *The Rain Forests of Home: Profile of a North American Bioregion*, pp. 227–254. Washington, DC: Island Press.
- Heine JT (1998) Extent, timing, and climatic implications of glacier advances Mount Rainier, Washington, U.S.A., at the Pleistocene/Holocene transition. *Quaternary Science Reviews* 17: 1139–1148.
- Heusser CJ, Heusser LE, and Peteet DM (1999) Humptulips revisited: A revised interpretation of Quaternary vegetation and climate of western Washington, USA. *Palaeogeography, Palaeoclimatology, Palaeoecology* 150: 191–221.
- Higuera PE, Brubaker LB, Anderson PM, et al. (2009) Vegetation mediated the impacts of postglacial climate change on fire regimes in the south-central Brooks Range, Alaska. *Ecological Monographs* 79: 201–219.
- Hopkins DM, Smith PA, and Matthews JV (1981) Dated wood from Alaska and the Yukon – Implications for forest refugia in Beringia. *Quaternary Research* 15: 217–249.
- Hu FS, Brubaker LB, and Anderson PM (1993) A 12000 year record of vegetation change and soil development from Wien Lake, Central Alaska. *Canadian Journal of Botany* 71: 1133–1142.
- Hu FS, Brubaker LB, and Anderson PM (1995) Postglacial vegetation and climate-change in the northern Bristol Bay Region, southwestern Alaska. *Quaternary Research* 43: 382–392.
- Hu FS, Finney BP, and Brubaker LB (2001) Effects of Holocene *Alnus* expansion on aquatic productivity, nitrogen cycling, and soil development in southwestern Alaska. *Ecosystems* 4: 358–368.
- Hu FS, Kaufman D, Yoneji S, et al. (2003) Cyclic variation and solar forcing of Holocene climate in the Alaskan subarctic. *Science* 301: 1890–1893.
- Hu FS, Lee BY, Kaufman DS, et al. (2002) Response of tundra ecosystem in southwestern Alaska to Younger-Dryas climatic oscillation. *Global Change Biology* 8: 1156–1163.
- Hu FS, Nelson DM, Clarke GH, et al. (2006) Abrupt climatic events during the last glacial-interglacial transition in Alaska. *Geophysical Research Letters* 33.
- Hultén E (1937) *Outline of the History of Arctic and Boreal Biota During the Quaternary Period: Their Evolution During and After the Glacial Period as Indicated by the Equiformal Progressive Areas of Present Plant Species*. PhD Dissertation, Lund University, Stockholm, Sweden.
- Jimenez-Moreno G, Anderson RS, and Fawcett PJ (2007) Orbital- and millennial-scale vegetation and climate changes of the past 225 ka from Bear Lake, Utah-Idaho (USA). *Quaternary Science Reviews* 26: 1713–1724.
- Kaltenrieder P, Tinner W, Lee B, et al. (2011) A 16000-year record of vegetational change in south-western Alaska as inferred from plant macrofossils and pollen. *Journal of Quaternary Science* 26: 276–285.
- Kaufman D, Ager T, Anderson N, et al. (2004) Holocene thermal maximum in the western Arctic (0–180 degrees W). *Quaternary Science Reviews* 23: 529–560.
- Kaufman DS, Anderson RS, Hu FS, et al. (2010) Evidence for a variable and wet Younger Dryas in southern Alaska. *Quaternary Science Reviews* 29: 1445–1452.
- Lacourse T (2009) Environmental change controls postglacial forest dynamics through interspecific differences in life-history traits. *Ecology* 90: 2149–2160.
- Lacourse T, Mathewes RW, and Fedje DW (2003) Paleoeecology of late-glacial terrestrial deposits with in situ conifers from the submerged continental shelf of western Canada. *Quaternary Research* 60: 180–188.
- Livingstone DA (1955) Some pollen profiles from Arctic Alaska. *Ecology* 36: 587–600.
- Livingstone DA (1957) Pollen analysis of a valley fill near Umiat, Alaska. *American Journal of Science* 255: 254–260.
- Long C, Whitlock C, Bartlein P, et al. (1998) A 9000-year fire history from the Oregon Coast Range, based on a high-resolution charcoal study. *Canadian Journal of Forest Research* 28: 774–787.
- Lynch A, Rivers A, and Bartlein P (2003) An assessment of the influence of land cover uncertainties on the simulation of global climate in the early Holocene. *Climate Dynamics* 21: 243–256.
- Mack RN, Rutter NW, Valastro S, et al. (1978) Late Quaternary vegetation history at Waits Lake, Colville River Valley, Washington. *Botanical Gazette* 139: 499–506.
- Mann DH, Heiser PA, and Finney BP (2002) Holocene history of the Great Kobuk Sand Dunes, northwestern Alaska. *Quaternary Science Reviews* 21: 709–731.

- Marlon JR, Bartlein PJ, Walsh MK, et al. (2009) Wildfire responses to abrupt climate change in North America. *Proceedings of the National Academy of Sciences of the United States of America* 106: 2519–2524.
- Mathewes R (1991) Climatic conditions in the western and northern cordillera during the last glaciation – Paleoeological evidence. *Géographie physique et Quaternaire* 45: 333–339.
- Mathewes RW (1993) Evidence for Younger Dryas-age cooling on the North Pacific coast of America. *Quaternary Science Reviews* 12: 321–331.
- Mehring PJ (1996) *Columbia River Basin Ecosystems: Late Quaternary Environments*. (Interior Columbia Basin Ecosystem Management Project). Walla Walla WA, USA: US Forest Service and Bureau of Land Management.
- Mehring PJ, Blinman E, and Petersen KL (1977) Pollen influx and volcanic ash. *Science* 198: 257–261.
- Meidinger D and Pojar J (1991) *Ecosystems of British Columbia, Special Report Series 06*. BC Ministry of Forests.
- Minckley T, Whitlock C, and Bartlein P (2007) Vegetation, fire, and climate history of the northwestern Great Basin during the last 14,000 years. *Quaternary Science Reviews* 26: 2167–2184.
- O'Connell LM, Ritland K, and Thompson SL (2008) Patterns of post-glacial colonization by western redcedar (*Thuja plicata*, Cupressaceae) as revealed by microsatellite markers. *Botany* 86: 194–203.
- Oswald WW, Brubaker LB, and Anderson PM (1999) Late Quaternary vegetational history of the Howard Pass area, northwestern Alaska. *Canadian Journal of Botany* 77: 570–581.
- Oswald WW, Brubaker LB, Hu FS, et al. (2003) Holocene pollen records from the central Arctic Foothills, northern Alaska: Testing the role of substrate in the response of tundra to climate change. *Journal of Ecology* 91: 1034–1048.
- Peteet DM and Mann DH (1994) Late-glacial vegetational, tephra, and climatic history of southwestern Kodiak Island, Alaska. *Ecoscience* 1: 255–267.
- Prichard SJ, Gedalof Z, Oswald WW, et al. (2009) Holocene fire and vegetation dynamics in a montane forest, North Cascade Range, Washington, USA. *Quaternary Research* 72: 57–67.
- Shafer ABA, Cullingham CI, Cote SD, et al. (2010) Of glaciers and refugia: A decade of study sheds new light on the phylogeography of northwestern North America. *Molecular Ecology* 19: 4589–4621.
- Sugimura WY, Sprugel DG, Brubaker LB, et al. (2008) Millennial-scale changes in local vegetation and fire regimes on Mount Constitution, Orcas Island, Washington, USA, using small hollow sediments. *Canadian Journal of Forest Research* 38: 539–552.
- Sugita S and Tsukada M (1982) The vegetation history in western North America 1. Mineral and Hall lakes. *Japanese Journal of Ecology* 32: 499–516.
- Thompson RS and Anderson KH (2000) Biomes of western North America at 18,000, 6000 and 0 C-14 yr BP reconstructed from pollen and packrat midden data. *Journal of Biogeography* 27: 555–584.
- Thompson RS, Shafer SL, Strickland LE, et al. (2003) Quaternary vegetation and climate change in the western United States: Developments, perspectives, and prospects. *The Quaternary Period in the United States*, pp. 403–426. Amsterdam: Elsevier.
- Tinner W, Bigler C, Gedy S, et al. (2008) A 700-year paleoecological record of boreal ecosystem responses to climatic variation from Alaska. *Ecology* 89: 729–743.
- Tinner W, Hu FS, Beer R, et al. (2006) Postglacial vegetational and fire history: Pollen, plant macrofossil and charcoal records from two Alaskan lakes. *Vegetation History and Archaeobotany* 15: 279–293.
- Vacco D, Clark P, Mix A, et al. (2005) A speleothem record of Younger Dryas cooling, Klamath Mountains, Oregon, USA. *Quaternary Research* 64: 249–256.
- Vermaire JC and Cwynar LC (2010) A revised late-Quaternary vegetation history of the unglaciated southwestern Yukon Territory, Canada, from Antifreeze and Eikland ponds. *Canadian Journal of Earth Sciences* 47: 75–88.
- Viereck LA, Dyrness CT, Batten AR, and Wenzlick KJ (1992) *The Alaska Vegetation Classification (No. GTR-PNW-286)*. Department of Agriculture, Forest Service, Pacific Northwest Research Station.
- Walker IR and Pellatt MG (2008) Climate change and ecosystem response in the northern Columbia River basin – A paleoenvironmental perspective. *Environmental Reviews* 16: 113–140.
- Walker MD, Walker DA, and Auerbach AN (1994) Plant communities of a tussock tundra landscape in the Brooks Range Foothills, Alaska. *Journal of Vegetation Science* 5: 843–866.
- Walsh MK, Whitlock C, and Bartlein PJ (2008) A 14,300-year-long record of fire–vegetation–climate linkages at Battle Ground Lake, southwestern Washington. *Quaternary Research* 70: 251–264.
- Warner B, Mathewes R, and Clague J (1982) Ice-free conditions on the Queen Charlotte Islands, British Columbia, at the height of Late Wisconsin Glaciation. *Science* 218: 675–677.
- Whitlock C (1992) Vegetational and climatic history of the Pacific Northwest during the last 20,000 years – implications for understanding present-day biodiversity. *Northwest Environmental Journal* 8: 5–28.
- Whitlock C, Marlon J, Briles C, et al. (2008) Long-term relations among fire, fuel, and climate in the north-western US based on lake-sediment studies. *International Journal of Wildland Fire* 17: 72–83.
- Whitlock C, Sarna-Wojcicki A, Bartlein P, et al. (2000) Environmental history and tephrostratigraphy at Carp Lake, southwestern Columbia Basin, Washington, USA. *Palaeogeography, Palaeoclimatology, Palaeoecology* 155: 7–29.
- Worona MA and Whitlock C (1995) Late Quaternary vegetation and climate history near Little Lake, central Coast Range, Oregon. *Geological Society of America Bulletin* 107: 867–876.
- Zdanowicz CM, Zielinski GA, and Germani MS (1999) Mount Mazama eruption: Calendrical age verified and atmospheric impact assessed. *Geology* 27: 621.

Southeastern North America

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Introduction

The southeastern United States has a diverse flora and fauna that reflects the topographic and latitudinal gradients characterizing the region. This region ranges from Virginia and Tennessee in the north to the Gulf of Mexico in the south and is bounded on the west by the Mississippi River and the east by the Atlantic Ocean. It was south of the maximum extent of continental ice during the Last Glacial Maximum (LGM) (21.5 ± 1.5 cal ka BP) (Jackson et al., 2000; Figure 1). As such, the southeastern United States served as a refugium for plant and animal species that migrated south during the last glaciation. Postglacial sedimentary records provide details on migration rates and environmental tolerances of taxa during the interval ranging from the LGM to the present day. This article summarizes more than 75 years of research on the vegetational history of the southeastern United States and highlights current and potential future directions for paleoecological research in the region.

Initial research on southeastern United States floras began in the 1930s, with studies of pollen records from peat bogs by Sears (1935). Although chronologic control was poor before the advent of radiocarbon dating in 1950, these researchers developed a fundamental understanding of the history and development of the southeastern flora through studies of pollen from lake and bog sediments. After 1950, radiocarbon dating facilitated analysis of the timing of millennial-scale events associated with the last glaciation and the response of southeastern United States vegetation to climatic amelioration of the deglacial and Holocene intervals. Basic biogeographic questions were researched throughout the region by Braun (1950), Whitehead (1965), Watts (1979), Delcourt and Delcourt (1984), and others in a series of site studies of lakes and wetlands. Since the 1980s, the ability to date very small samples with ^{14}C accelerator mass spectroscopy, to undertake quantitative paleoecological analyses, and to integrate paleoecological results with climate models has enabled more sophisticated analyses of broad climatic impacts and patterns of late Quaternary vegetational change (Jackson et al., 2000; Webb et al., 1998). Recent attention has focused on the following: (1) development of pollen records with greater temporal resolution to evaluate the timing of forest response to decadal- to -centennial-scale climate fluctuations and (2) improved calibration of pollen assemblages with environmental parameters including temperature and hydrology (Sugita, 2007; Whitmore et al., 2005; Willard et al., 2011). Improved geochronology is allowing more direct comparison of pollen records with ice core and marine sediment records, improving regional and extraregional reconstructions of vegetation and changes in air temperature and precipitation. In the following sections, the modern distribution of vegetation and pollen is discussed, followed by a discussion of patterns of change in plant communities from the LGM into the present interglacial interval.

Modern Forest Distribution and Composition in the Southeastern United States

The diverse vegetation of the southeastern United States reflects the complex topography and physiography of the region (Figure 2(a)). Both the Atlantic and Gulf coastal plains are dominant elements of southeastern United States, characterized by low topographic relief. The Atlantic coastal plain extends southward from New Jersey to Florida, grading westward into the Gulf coastal plain and extending as far north as southern Illinois at the head of the Mississippi embayment (Figure 2(a)). The western margin of the Atlantic coastal plain is clearly defined by the fall line, the boundary between the upland Piedmont province and Atlantic coastal plain that marks a mid-Pliocene highstand of sea level. The Piedmont, Blue Ridge, Ridge and Valley, and Appalachian plateau provinces are more topographically diverse, including fold belts with significant relief and relatively flat plateaus with minimal relief. Coupling the extreme local variability in topography and substrate with climatic gradients related both to latitude and altitude, these regions encompass great diversity in forest composition (Figure 2(b)).

Vegetational associations in the southeastern United States include the wetlands and tropical vegetation of south Florida, southeastern evergreen forests, oak-pine forests, and a series of deciduous forests including the Appalachian oak/oak-chestnut forests, mixed mesophytic forests, and western mesophytic forests (Figure 2(b); Braun, 1950; Kuchler, 1964; Greller, 1988). The southern one-third of the Florida peninsula consists of extensive wetlands, including freshwater marshes and forested wetlands of the Florida Everglades and saline-influenced coastal marshes and mangrove forests along Florida Bay and the Gulf of Mexico. The freshwater marsh communities of the Everglades ecosystem consist primarily of saw grass marshes and water lily sloughs with tree islands interspersed throughout the system (Willard et al., 2001). These communities occur primarily on a peat substrate, and the community composition is determined by hydroperiod (annual period of inundation), water depth, and fire regime.

Forested wetlands of the Atlantic and Gulf coastal plains include Everglades tree islands, coastal mangrove swamps, and floodplains of coastal plain rivers. The latter have acted as sediment sinks during the Holocene, and lateral accretion of sediment during flood events has influenced the structure of floodplain environments, with coarse-grained sediments deposited on levees and finer-grained sediments deposited in backswamps distal to the stream. The hydrologic, topographic, and geomorphic diversity (substrate elevation, distance from river mouth, soil texture, and organic matter content) inherent to coastal plain floodplains determines the structure of floodplain vegetation (Rice and Peet, 1997; Townsend, 2001). Floodplain forests include cypress and tupelo (*Nyssa*) in backswamps that are flooded for prolonged periods of time

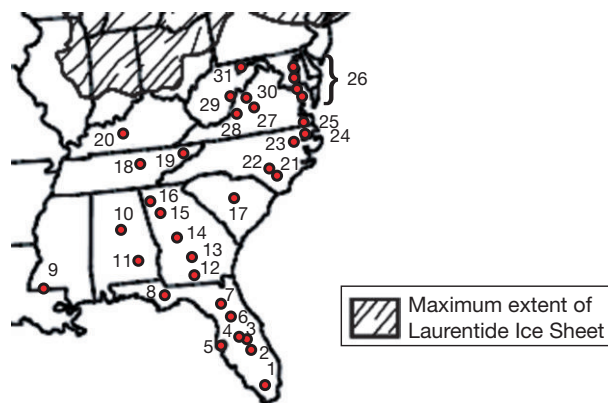


Figure 1 Map of southeastern United States with location of sites discussed within the text. Site numbers are keyed to localities in [Table 1](#). Shaded area shows maximum extent of continental ice sheet during Last Glacial Maximum.

(pocosin vegetation of [Figure 2\(b\)](#)) and hardwood assemblages (oak, sweet gum, ash (*Fraxinus*)) on the better-drained levees ([Hupp, 2000](#)).

The remainder of the Florida peninsula and southern Mississippi, Alabama, Georgia, and the eastern Carolinas are occupied by southeastern evergreen forests. This vegetation association occupies the Atlantic and Gulf coastal plain provinces and is dominated by longleaf pine (*Pinus palustris*) and broad-leaved evergreens such as live oak (*Quercus virginiana*). Longleaf pine forests occupy xeric sites and consist of an overstory of longleaf pine, an understory of oaks, and an herbaceous layer consisting primarily of grasses ([Braun, 1950](#)). Deciduous taxa are more common in the northern reaches of the association, where loblolly pine (*P. taeda*) forests include a number of hardwoods (oak, *Quercus*; hickory, *Carya*; sweet gum, *Liquidambar*; and maple, *Acer*).

The oak–pine association lies north of the southeastern evergreen forests, extending southward from southern New Jersey to Georgia and westward to the Mississippi River ([Figure 2\(b\)](#)). It encompasses part of the Piedmont plateau along the Atlantic coast and the Piedmont, Valley and Ridge,

Table 1 Sites included in summary of paleoecological data from the southeastern United States. Site numbers are keyed to sites illustrated in [Figure 1](#)

Site no.	Site name	State	Age	Citation
1	Everglades	FL	Late Holocene	Willard et al. (2001)
2	Lake Annie	FL	Early Wisconsin, Late Glacial to late Holocene	Watts (1975b)
3	Lake Tulane	FL	Wisconsin, full glacial to late Holocene	Grimm et al. (2006)
4	Scott Lake	FL	Middle to late Holocene	Watts (1971)
5	Tampa Bay	FL	Full glacial through deglacial; late Holocene	Willard et al. (2007)
6	Mud Lake	FL	Early Wisconsin, middle to late Holocene	Watts (1969)
7	Sheelar Lake	FL	Middle Wisconsin to late Holocene	Watts and Hansen (1994)
8	Camel Lake	FL	Late Wisconsin to early Holocene	Watts et al. (1992)
9	Tunica Hills	LA	Late Glacial to late Holocene	Delcourt and Delcourt (1977)
10	Cahaba Pond	AL	Early to late Holocene	Delcourt et al. (1983)
11	Goshen Springs	AL	Early Wisconsin to late Holocene	Delcourt (1980)
12	Lake Louise	GA	Early Wisconsin, middle to late Holocene	Watts (1971)
13	McClelland Sandpit Site	GA	Early Holocene	Zayac et al. (2001)
14	Sandy Run Creek	GA	Late deglacial to late Holocene	Lamoreaux et al. (2009)
15	Pigeon Marsh	GA	Full glacial to late Holocene	Watts (1975a)
16	Quicksand Pond	GA	Early Wisconsin to late Holocene	Watts (1970)
17	White Pond	SC	Full glacial to late Holocene	Watts (1980)
18	Anderson Pond	TN	Middle Wisconsin to late Holocene	Delcourt (1979)
19	Shady Valley Bog	TN	Early to late Holocene	Sears (1935)
20	Jackson Pond	KY	Full glacial to late Holocene	Wilkins et al. (1991)
21	Singletary Lake	NC	Early Wisconsin to late Holocene	Frey (1951)
22	Fort Bragg	NC	Early to late Holocene	Goman and Leigh (2004)
23	Roanoke River	NC	Late Holocene	Willard et al. (2011)
24	Rockyhock Bay	NC	Early Wisconsin to late Holocene	Whitehead (1973)
25	Dismal Swamp	VA	Early to late Holocene	Whitehead (1972)
26	Chesapeake Bay–Potomac River mouth	MD	Early to late Holocene	Willard et al. (2003, 2005)
26	Chesapeake Bay–Patuxent River mouth	MD	Late deglacial to late Holocene	Willard et al. (2003, 2010)
26	Chesapeake Bay–Kent Island	MD	Early to late Holocene	Willard et al. (2003, 2005)
26	Chesapeake Bay and tributaries	MD	Early to late Holocene	Brush (1984); Brush et al. (1982)
27	Hack Pond	VA	Late Glacial to late Holocene	Craig (1969)
28	Browns Pond	WR	Deglacial to late Holocene	Kneller and Peteet (1999)
29	Cranberry Glades	WV	Late Glacial to late Holocene	Watts (1979)
30	Potts Mountain Pond	VA	Early to late Holocene	Watts (1979)
31	Buckles Bog	MD	Full glacial to late Holocene	Maxwell and Davis (1972)

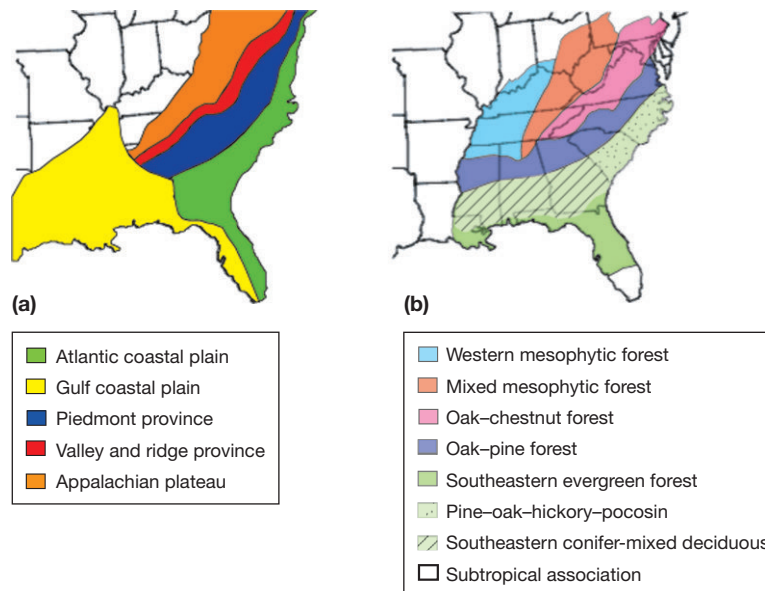


Figure 2 (a) Physiographic regions of southeastern United States, (b) Forest associations of the southeastern United States. Adapted from Braun LE (1950) *Deciduous Forests of Eastern North America*. Philadelphia: The Blakiston Company; K  chler AW (1964) *Potential Natural Vegetation of the Conterminous United States*. New York: American Geographical Society. Special Publication No.36, pp.1–116; Greller A (1988) Deciduous forest. In: Barbour MG and Billings WD (eds.) *North American Terrestrial Vegetation*, pp. 287–316. Cambridge: Cambridge University Press.

and Appalachian plateau along the Gulf coast. Although deciduous hardwoods are the dominant elements of these forests, pines are a consistent member of the association, thus distinguishing it from the oak–hickory region to the north (Braun, 1950). Oak dominates forests of the oak–pine association, and pine, hickory, and various hardwoods are subdominant (Braun, 1950). North of the Gulf coastal plain, the oak–pine association is a transitional zone with considerable overlap in ranges of trees from the mixed mesophytic forests to the north and evergreens to the south. Longleaf pine is a common component in southern reaches of this association, along with *Magnolia*, oak, hickory, and *Castanea* (chestnut). In higher elevations, beech (*Fagus*), sweet gum, maple, and other mesophytic taxa are common. The James River in Virginia represents the northern boundary of this association; north of the river, longleaf pine, and live oak are minor forest components.

A series of deciduous forests lie north of the oak–pine association and include the oak–chestnut, mixed mesophytic, and western mesophytic associations described by Braun (1950). These forests cover much of the unglaciated Appalachian plateau, the southern Appalachian Mountains, and part of the Piedmont plateau (Figure 2(a) and 2(b)). These associations are dominated by hardwoods (oak, hickory, maple, tuliptree (*Liriodendron*), hemlock (*Tsuga*), basswood (*Tilia*)), and species dominance is determined primarily by site characteristics. On the Appalachian plateau, oak, hickory, maple, and beech are the primary taxa. The Piedmont plateau is characterized by a rich mix of oaks and hickories, and loblolly pine (*Pinus taeda*) and shortleaf pine (*P. echinata*) are common in secondary forests. In the southern Appalachian Mountains, there is a gradient from diverse stands of hardwoods at low elevations to forests with spruce (*Picea*), fir (*Abies*), and eastern white pine (*Pinus strobus*) at high elevation (Braun, 1950).

Pollen Signature of Modern Vegetation

Calibration of the pollen record with modern vegetation in the southeastern United States has been undertaken in a comprehensive fashion (Delcourt et al., 1984; Webb et al., 1993). These synoptic views, focused primarily on lake records, provide a regional synthesis of surrounding vegetation. More recently, a number of studies examined pollen–vegetation relationships in wetlands. These studies documented more local-scale variation in order to examine impacts of local sedimentation, hydrology, and land cover change (Willard et al., 2001, 2005, 2011). Below, the general patterns of pollen distribution in the southeastern United States are summarized for key taxa within the region.

Pines of the southeastern United States include species whose ranges are limited by soil characteristics, moisture availability, and winter temperature (Iverson et al., 1999; Thompson et al., 2000). Southeastern pines are most common on sandy coastal plain sediments and decrease in abundance northward to a minimum in oak–chestnut forests of the mid-Atlantic. Although pine tends to be overrepresented in pollen records, this is paralleled by its abundance in forests (Delcourt and Delcourt, 1984). Where pines are rare in the watershed, their pollen is similarly uncommon in sediments. Recent analysis of pine pollen abundance in estuarine and shallow marine sediments clearly documents the positive relationship between pine pollen abundance and abundance in adjacent forests, as well as the correlation between abundance of southeastern pine pollen with both latitude and winter temperature (Willard et al., 2005).

Oak and hickory encompass numerous species with broad ecological ranges. Oak is most abundant on the Cumberland plateau of Tennessee and Alabama, in the Ridge and Valley and Blue Ridge provinces, decreasing southward as the association grades into more conifer-rich forests (Delcourt and Delcourt,

1984). Hickory is most abundant in the interior lowlands of Tennessee and Kentucky and is rare on the coastal plain (Delcourt and Delcourt, 1984; Iverson et al., 1999). Its pollen abundance closely parallels its distribution in forests, providing a reliable proxy for its presence and abundance in the pollen record. Within both genera, it is difficult to distinguish between species based on pollen morphology because size and subtle morphological differences between species may vary among reference collections. In a few cases, however, size can be used to differentiate species. For example, *Carya floridana*, a species restricted to Florida, is significantly smaller than other *Carya* species that have a more temperate/northern distribution (Whitehead, 1965). Such difficulties in distinguishing hickory or oak species in the pollen record make identification of particular edaphic or topographic characteristics based on species identification of pollen problematic.

Spruce, fir, and eastern white pine are common elements in the northern hardwoods and boreal forest regions and also occur rarely at high elevations in the central Appalachian Mountains of the southeastern United States. Hemlock and beech also are restricted to mid- to high elevations of the Appalachians (Delcourt and Delcourt, 1984). Other species with restricted distribution include those characteristic of backswamps of coastal plain rivers and other swamp complexes, including tupelo, cypress, and *Salix* (willow). Tupelo and willow pollen are found most commonly in backswamp sediments, with poor representation in adjacent levee deposits (Willard et al., 2011). Cypress pollen is restricted to sediments deposited in cypress swamps (Willard et al., 2001, 2011). Although its pollen is difficult to distinguish from that of the Cupressaceae (including the cedars *Juniperus* and *Chamaecyparis*), the latter occur on drier, shallow, upland soils, and the accompanying taxa in the upland associations differ completely from those of alluvial backswamps.

Quaternary Palynological Records from Lakes, Estuaries, and Wetlands of the Southeastern United States

The unglaciated southeastern United States is an excellent region to examine vegetational response to retreat of the Laurentide Ice Sheet and associated climatic fluctuations during the last ~22,500 years. The combined analysis of lacustrine, estuarine, and wetland deposits over the past few decades has enriched our interpretation of regional response to climate fluctuations on differing scales. Lacustrine sediments have been used extensively to interpret the vegetational history of the region because they contain long-term, relatively continuous records of sediment deposition. Lake deposits have been studied intensively throughout the region, including karst lakes of the Florida peninsula, sinkhole lakes throughout the Appalachians, and 'Carolina bays' along the Atlantic coast. Many of these lakes contain records covering at least the last 20,000 years with varying degrees of temporal resolution due to both varying sedimentation rates and density of analysis. Recent analyses have included estuarine records with high sedimentation rates that provide decadal- to -centennial-scale temporal resolution, as well as wetland records from marshes and forested wetlands (Willard et al., 2001, 2011; Willard et al., 2007).

Estuarine sediments preserve a record of regional forest composition and have the potential for extremely high pre-Colonial sedimentation rates (up to 1 cm/decade at some sites in Chesapeake Bay; Willard et al., 2005). Estuarine pollen records may be integrated with interpretation of past water quality conditions (temperature, oxygen content, turbidity) from floral, faunal, and geochemical proxies (Cronin et al., 2003; Willard et al., 2003), providing a crucial link between terrestrial and estuarine processes and a means to evaluate the impact of post-Colonial land use activities on the watershed, tributaries, and estuary.

Because of their sensitivity to relatively subtle changes in climate and hydrology and the generally excellent preservation of pollen, wetland records provide a unique perspective on wetland distribution and extent over long time intervals. Marsh records may provide sufficient temporal resolution to document the response of wetland communities to late Holocene climate fluctuations. These records provide particularly good evidence for impacts of land use change since European colonization, including altered sedimentation rate and type, plant community shifts, and interpretations of relative differences in pre- and post-Colonial hydrology.

LGM (22.5–19.5 cal ka BP)

Sedimentary records of the LGM are preserved in lakes, wetlands, and ponds throughout the southeastern United States as well as in outcrops. Well-developed chronologies for these sections are difficult to establish because of low deposition rates, sedimentary hiatuses, and radiocarbon dating issues related to carbon source, root intrusion, and carbon content of sediments (Jackson et al., 2000). To evaluate the paleoenvironmental history of the LGM in the southeastern United States, Jackson et al. (2000) incorporated 9 of the 18 sites used in the Cooperative Holocene Mapping Project (COHMAP) pollen database (Webb et al., 1993, 1998) with more recently reported sites to maximize the chronological accuracy of their LGM reconstruction. Sites with low sedimentation rates during the LGM period were excluded; for sites with variable sedimentation rates, only samples closely associated with LGM age dates were included to provide a snapshot of vegetation during the interval. All radiocarbon dates presented here have been calibrated to calendar years and are presented as calendar kiloyears before present (cal ka BP).

During the LGM, taiga, steppe, and boreal forest occupied sites near the ice margin, and the southern limit of the boreal forest region was up to 1200 km south of its present limit in Canada (Delcourt and Delcourt, 1985). Alpine tundra has been documented in LGM sediments near the ice margin in Pennsylvania and mid-elevations in the central Appalachians of West Virginia (Watts, 1979), whereas taiga appears to have been limited to western localities near the ice margin (Jackson et al., 2000). Boreal forests, characterized by abundant spruce throughout most of their range, extended southward from near the ice margin into Georgia and Alabama (Jackson et al., 2000). Fir was concentrated in a narrow band along the southern limit of the forest where moisture is thought to have been more abundant during the growing season (Delcourt and Delcourt, 1985). Mixed conifer/deciduous forests dominated the Atlantic coastal plain, with northern hardwoods present in

the Carolinas and southern pines more abundant in Georgia and northern Florida (Jackson et al., 2000). Hickory and sweetgum were common elements of LGM forests as far south as Tampa Bay (Jackson et al., 2000; Willard et al., 2007).

Deglacial Interval (19.5–10 cal ka BP)

Sites throughout the southeastern United States exhibited significant changes in forest composition around 17 cal ka BP, although sites near the northern limit of the full-glacial boreal forest responded as early as 19 cal ka BP (Delcourt and Delcourt, 1984). Significant decreases in abundance of northern pines and spruce from Florida to Virginia and Tennessee are accompanied by greater abundance of pollen of more mesic and cool-temperate deciduous taxa after ~17 cal ka BP (Figure 3). The decreased abundance of spruce pollen after ~17 cal ka BP indicates a northward shift of their populations after the glacial maximum, but spruce and fir pollen remained common throughout the deglacial interval in both Virginia and Tennessee. The incorporation of rare mesic species with boreal species into a mixed conifer deciduous forest is interpreted to represent the continuation of relatively cool climatic conditions during an interval with greater summer precipitation than the full glacial. In southern sites (Florida, South Carolina), the common occurrence of pollen of temperate hickory species indicates a southward persistence of its population center during the early deglacial.

A second vegetational shift occurred ~15 cal ka BP at most southeastern sites, with the intensity and timing of this change varying latitudinally. In northern and central Florida, hickory abundance decreased and pine abundance increased after 15 cal ka BP, most likely reflecting greater abundance of southern pines such as shortleaf and loblolly pine. At White Pond in South Carolina, hickory, oak, beech, and hop hornbeam pollen abundance increased after 15 cal ka BP. These patterns are mirrored throughout the region (Figure 3) and represent the expansion of mesic trees into the mid-latitudes of the southeastern United States as climatic conditions ameliorated (Delcourt and Delcourt, 1985). These deglacial forests differ distinctly from midlatitude forests of the Holocene in both species composition and dominance. Hemlock and eastern white pine (*P. strobus*) occurred south of their modern distribution, and cypress occurred much farther inland than its present coastal limits. Although few sites have sufficient temporal resolution to evaluate vegetational response to centennial-scale events such as the Younger Dryas, there are indications of significant forest changes at Cahaba Pond, AL, where beech disappeared and pine/oak increased (Figure 4). Likewise, variability in abundance of pine, oak, and Amaranthaceae (pigweed) pollen in lake sediments from Florida indicates forest response to altered hydrology throughout this interval, although clarification of the timing of specific vegetational events is needed (Grimm et al., 2006; Willard et al., 2007). Although records from the high-altitude Browns Pond in the Virginia Appalachians indicate warmer conditions during the Younger Dryas (Kneller and Peteet, 1999), Chesapeake Bay pollen records indicate cooler drier conditions on the coast (Willard et al., 2010). Additional records with high temporal resolution and robust chronologies are needed to improve our understanding of the response of

regional vegetation to deglacial climatic events including the Bölling-Allerød and Younger Dryas.

Early Holocene (10–5 cal ka BP)

Early Holocene forests of the southeastern United States typically were oak–pine forests with decreased representation of mesic trees relative to the late deglacial interval. In Florida, high percentage of oak, Asteraceae (aster) and Poaceae (grass) pollen in early Holocene sediments indicates the presence of closed oak scrub forest with widespread areas of prairielike vegetation (Watts and Hansen, 1994). Pollen assemblages from Cahaba Pond in Alabama were dominated by oak whereas common nonarboreal pollen, mesic, and hydric taxa were relatively rare, and species richness was low, prompting interpretations of increased evapotranspiration stress and decreased precipitation (Delcourt et al., 1983). Hydrologic variability also was documented in assemblages from Dismal Swamp, North Carolina, where oak and hickory dominate early Holocene assemblages (Figure 4). Fluctuating abundance of herbaceous and aquatic taxa in this site are interpreted as recording local fluctuations in precipitation and the water table (Whitehead, 1972). Oak–chestnut forests dominated early Holocene forests of the Virginia mountains and Chesapeake Bay region, and periodic cooling, including the 8 cal ka BP event, is documented by increased montane conifers and decreased pine pollen abundance, respectively (Grimm et al., 2006; Willard et al., 2005). Dominance of oak, decreased abundance of hop hornbeam, spruce, and mesic species and increased abundance of trees around the pond all indicate drier conditions with lower, seasonally fluctuating water levels at Jackson Pond, Kentucky (Wilkins et al., 1991). The record from Anderson Pond in central Tennessee stands in contrast to other sites, demonstrating the continued common occurrence of mesic species throughout the early Holocene (Delcourt, 1979). The earliest documented occurrence of tupelo–cypress backswamps along Atlantic coastal plain rivers ranges from ~9 to 6 cal ka BP in Georgia and North Carolina (Goman and Leigh, 2004; LaMoreaux et al., 2009), near abandoned channel deposits. Although the high abundance of tupelo pollen indicates locally long hydroperiods related to fluvial processes, these complete pollen assemblages are characterized by greater early Holocene abundance of oak pollen relative to pine, indicating a regionally dry climate.

Late Holocene (5 cal ka BP to Present)

After ~5 cal ka BP, insolation changes caused cooler summers and warmer winters of the neoglaciation, resulting in substantial changes in forest composition throughout the southeastern United States. The most striking change was a widespread increase in pine abundance throughout the entire Atlantic coastal plain from Florida peninsula to New Jersey (Watts, 1979). Based on the modern distribution of pine species, this represents a northward expansion of the southern pines (shortleaf, slash, longleaf, and loblolly pine), which require relatively warm winters ($\geq 0.6^\circ\text{C}$), high moisture availability ($>885\text{-mm}$ annual precipitation), and unconsolidated soils (Iverson et al., 1999; Thompson et al., 2000). The shift from oak to pine dominance in these coastal plain sites suggests a

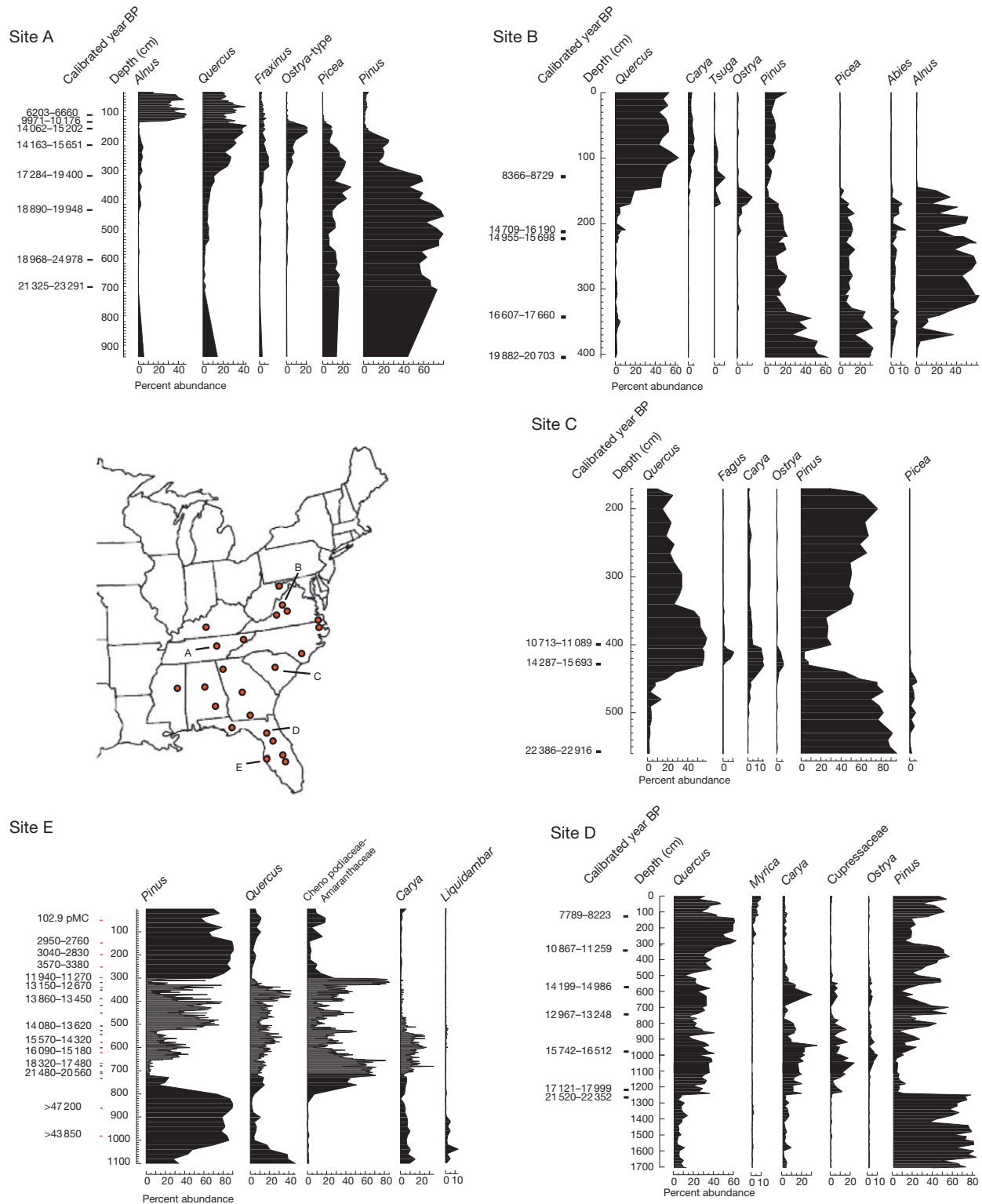


Figure 3 Pollen records of full glacial through deglacial age in the southeastern United States. Points on map indicate sites with information relevant to reconstructions discussed in text. Percent abundance data for key taxa are presented. Data for generation of pollen diagrams was obtained from the North American Pollen Database (<http://www.ncdc.noaa.gov/paleo/napd.html>). (a) Anderson Pond, Tennessee (Site 1, [Figure 1](#)); (b) Browns Pond, Virginia (Site 30, [Figure 1](#)); (c) White Pond, South Carolina (Site 17, [Figure 1](#)); (d) Sheelar Lake, Florida (Site 7, [Figure 1](#)); (e) Tampa Bay, Florida (Site 5, [Figure 1](#)). More complete information for sites is provided in [Table 1](#).

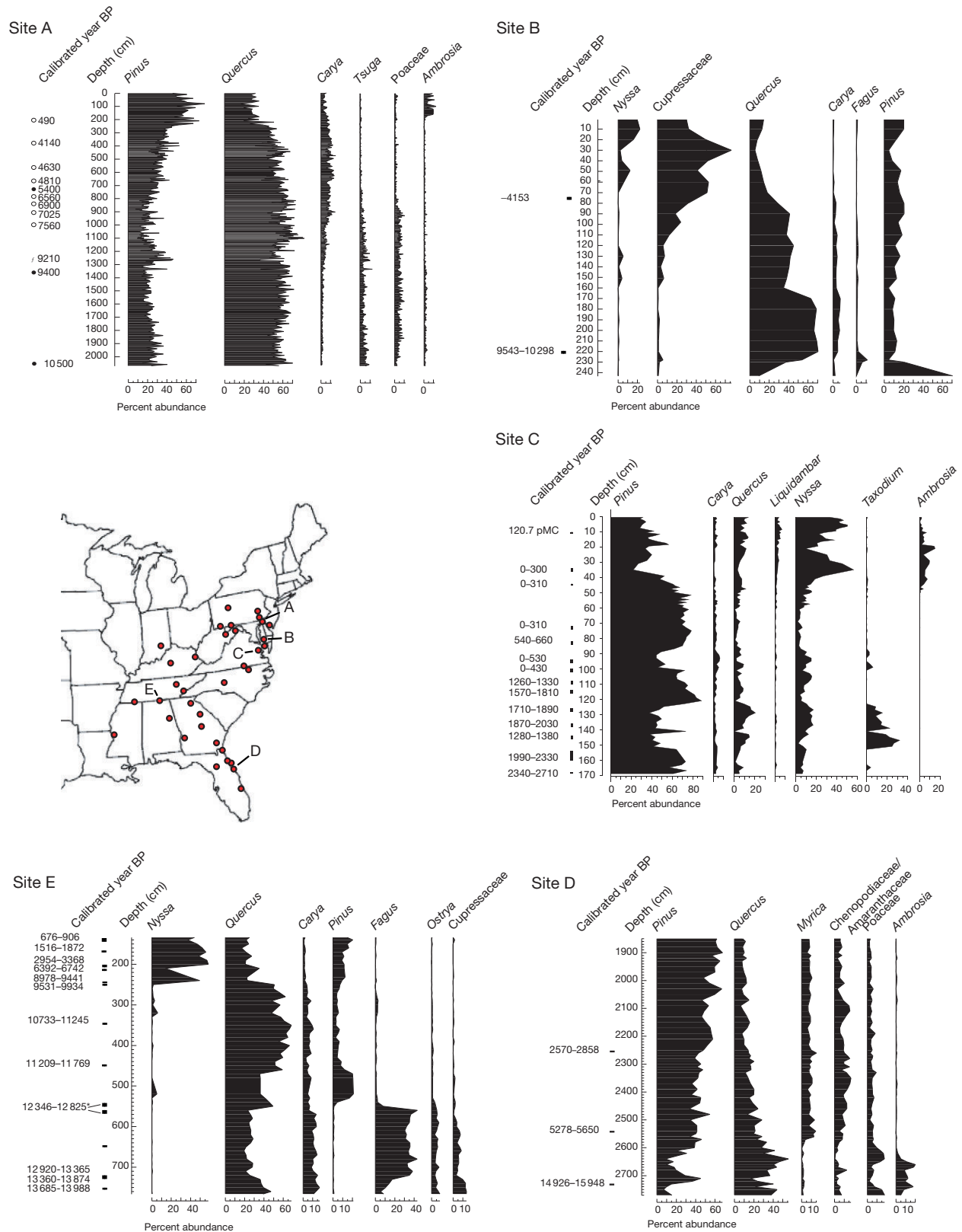


Figure 4 Holocene pollen records from the southeastern United States. Points on map indicate sites with information relevant to Holocene reconstructions discussed in text. Percent abundance data for key taxa are presented. Data for generation of pollen diagrams was obtained from the North American Pollen Database (<http://www.ncdc.noaa.gov/paleo/napd.html>). (a) Chesapeake Bay Core MD99-2207 (Site 26, **Figure 1**); (b) Dismal Swamp, Virginia (Site 25, **Figure 1**); (c) Roanoke River, North Carolina (Site 23, **Figure 1**); (d) Lake Annie, Florida (Site 2, **Figure 1**); (e) Cahaba Pond, Alabama (Site 10, **Figure 1**). More complete information for sites is provided in **Table 1**.

replacement of early Holocene oak–hickory-dominated forests by northerly expanding southeastern evergreen forests up to the southern Chesapeake Bay watershed. Spruce and fir locally expanded their populations into mid- and high elevations in the central and southern Appalachian Mountains (Delcourt and Delcourt, 1985), and chestnut expanded northward to help form extensive oak–chestnut forests. Local changes in hydrology and sedimentation resulted in development of swamp forests in the coastal plain (such as Dismal Swamp, North Carolina), around the Carolina Bays, and around coastal plain ponds (e.g., Cahaba Pond, Alabama). Although isolated tupelo–cypress swamps have been identified from early Holocene sediments (Goman and Leigh, 2004; LaMoreaux et al., 2009), most of the existing swamp forests originated in the late Holocene (Whitehead, 1972). Recent evidence from a 2500-year record from Roanoke River backswamp documents substantial hydrologic variability over centennial scales (Willard et al., 2011), but an improved chronology is necessary to correlate it with specific climate events.

High-resolution records from Chesapeake Bay document a forest response to quasiperiodic cool intervals superimposed upon the entire Holocene record (Willard et al., 2005). These consist of periodic decreases in pine abundance every ~1500 years that correlate with intervals of increased ice rafting in the North Atlantic Ocean and solar minima interpreted from ice core records. Although this remains to be substantiated in other sites with high temporal resolution, it suggests that forests of the southeastern United States were capable of responding to centennial-scale climate events in a manner recognizable in the sedimentary pollen record.

Implications for Future Research

Postglacial records of southeastern United States vegetation continue to provide valuable reconstructions of biogeographic trends and paleoclimatic interpretations of terrestrial records. As additional records with high temporal resolution are generated from the region, our understanding of the sensitivity of forests and other plant communities to relatively short-term climate shifts will be improved. Calibration of assemblage changes from terrestrial, estuarine, marine, and ice core records will expand our ability to model past climate events using coupled atmospheric and oceanic circulation models.

These long-term records also provide a context to evaluate more recent changes in plant communities due to land use changes since European colonization of southeastern North America. Land clearance has been shown to increase sedimentation rates in tributaries, estuaries, and bottomland forests (Brush, 1984; Willard et al., 2003, 2011), and secondary and tertiary forests now mantle much of the southeastern United States. Restoration efforts for degraded ecosystems such as the Florida Everglades and Chesapeake Bay are relying upon floral and faunal-based reconstructions of the natural variability of the ecosystem to develop sustainable restoration goals in the context of modern conditions and predicted future warming. The increasing use of multiple proxies (i.e., pollen, plant macrofossils, organic biomarkers) provides a promising means to improve reconstructions of different climatic and

environmental parameters. Further improvements in dating techniques and our understanding of radiocarbon reservoir corrections in different depositional settings should allow interpretation of vegetation changes over decadal- to centennial timescales. Such records are critical to understand the magnitude and rates of vegetation change under natural climate variability. The resulting understanding of vegetation and terrestrial response to different modes of climate variability provides a context to evaluate anthropogenic impacts on southeastern United States forests and their likely response to a range of future climate and land use scenarios.

See also: [Pollen Methods and Studies: Databases and Their Application](#). [Pollen Records, Postglacial: Northeastern North America](#); [Northwestern North America](#); [South America](#); [Southwestern North America](#).

References

- Braun LE (1950) *Deciduous Forests of Eastern North America*. Philadelphia: The Blakiston Company.
- Brush GS (1984) Patterns of recent sediment accumulation in Chesapeake Bay (Virginia–Maryland, U.S.A.) tributaries. *Chemical Geology* 44: 227–242.
- Brush GS, Martin EA, DeFries RS, et al. (1982) Comparisons of ^{210}Pb and pollen methods for determining rates of estuarine sediment accumulation. *Quaternary Research* 18: 196–217.
- Craig AJ (1969) Vegetational history of the Shenandoah Valley, Virginia. *Geological Society of America Special Paper* 123: 283–296.
- Cronin TM, Dwyer GS, Kamiya T, et al. (2003) Medieval Warm Period, Little Ice Age and 20th century temperature variability from Chesapeake Bay. *Global and Planetary Change* 36: 17–29.
- Delcourt HR (1979) Late Quaternary vegetation history of the eastern highland rim and adjacent Cumberland Plateau of Tennessee. *Ecological Monographs* 49: 255–280.
- Delcourt PA (1980) Goshen Springs: Late-Quaternary vegetation record for southern Alabama. *Ecology* 61: 371–386.
- Delcourt PA and Delcourt HR (1977) The Tunica Hills, Louisiana-Mississippi: Late glacial locality for spruce and deciduous forest species. *Quaternary Research* 7: 218–237.
- Delcourt PA and Delcourt HR (1984) Late-Quaternary paleoclimates and biotic responses in eastern North America and the western North Atlantic Ocean. *Palaeogeography, Palaeoclimatology, Palaeoecology* 48: 263–284.
- Delcourt HR and Delcourt PA (1985) Quaternary palynology and vegetational history of the southeastern United States. In: Bryant VM and Holloway RG (eds.) *Pollen Records of Late-Quaternary North American Sediments*, pp. 1–37. Austin, TX: American Association of Stratigraphic Palynologists Foundation.
- Delcourt HR, Delcourt PA, and Spiker EC (1983) A 12,000-year record of forest history from Cahaba Pond, St. Clair County, Alabama. *Ecology* 64: 874–887.
- Delcourt PA, Delcourt HR, and Webb T III (1984) *Atlas of Mapped Distributions of Dominance and Modern Pollen Percentages for Important Tree Taxa of Eastern North America*. AASP Contributions Series, vol. 14, pp. 1–131. College Station, TX: American Association of Stratigraphic Palynologists.
- Frey DG (1951) Pollen succession in the sediments of Singletary Lake, North Carolina. *Ecology* 32: 518–533.
- Goman M and Leigh DS (2004) Wet early to middle Holocene conditions on the upper Coastal Plain of North Carolina, USA. *Quaternary Research* 61: 256–264.
- Greller AM (1988) Deciduous forest. In: Barbour MG and Billings WD (eds.) *North American Terrestrial Vegetation*, pp. 287–316. Cambridge: Cambridge University Press.
- Grimm EC, Watts WA, Jacobson GL Jr., et al. (2006) Evidence for warm wet Heinrich events in Florida. *Quaternary Science Reviews* 25: 2197–2211.
- Hupp CR (2000) Hydrology, geomorphology, and vegetation of Coastal Plain rivers in the south-eastern USA. *Hydrological Processes* 14: 2991–3010.
- Iverson LR, Prasad AM, Hale BJ, and Sutherland EK (1999) Atlas of current and potential future distributions of common trees of the eastern United States. USDA Forest Service General *Technical Report*, NE-265 USDA Forest Service, Radnor, Pennsylvania.

- Jackson ST, Webb RS, Anderson KH, et al. (2000) Vegetation and environment in eastern North America during the Last Glacial Maximum. *Quaternary Science Reviews* 19: 489–508.
- Kneller M and Peteet D (1999) Late-glacial to early Holocene climate changes from a central Appalachian pollen and macrofossil record. *Quaternary Research* 51: 133–147.
- Küchler AW (1964) *Potential Natural Vegetation of the Conterminous United States*, pp. 1–116. New York: American Geographical Society Special Publication 36.
- LaMoreaux HK, Brook GA, and Knox JA (2009) Late Pleistocene and Holocene environments of the Southeastern United States from the stratigraphy and pollen content of a peat deposit on the Georgia Coastal Plain. *Palaeogeography, Palaeoclimatology, Palaeoecology* 280: 300–312.
- Marshall CH, Pielke RA Sr., Steyaert LT, et al. (2004) The impact of anthropogenic land-cover change on the Florida peninsula sea breezes and warm season sensible weather. *Monthly Weather Review* 132: 28–52.
- Maxwell JA and Davis MB (1972) Pollen evidence of Pleistocene and Holocene vegetation on the Allegheny Plateau. *Quaternary Research* 2: 506–530.
- Rice SK and Peet RK (1997) *Vegetation of the Lower Roanoke River Floodplain*. Durham, NC: The Nature Conservancy.
- Sears PB (1935) Types of North American pollen profiles. *Ecology* 16: 488–499.
- Sugita S (2007) Theory of quantitative reconstruction of vegetation II: All you need is LOVE. *The Holocene* 17: 243–257.
- Thompson RS, Anderson KH, and Bartlein PJ (2000) Atlas of relations between climatic parameters and distributions of important trees and shrubs in North America – Introduction and conifers. *United States Geological Survey Professional Paper* 1650-A: 1–269.
- Townsend PA (2001) Relationships between vegetation patterns and hydroperiod on the Roanoke River floodplain, North Carolina. *Plant Ecology* 156: 43–58.
- Watts WA (1969) A pollen diagram from Mud Lake, Marion County, north-central Florida. *Geological Society of America Bulletin* 80: 631–642.
- Watts WA (1970) The full-glacial vegetation of northwestern Georgia. *Ecology* 51: 17–33.
- Watts WA (1971) Postglacial and interglacial vegetation history of southern Georgia and central Florida. *Ecology* 52: 676–690.
- Watts WA (1975a) Vegetation record for the last 20,000 years from a small marsh on Lookout Mountain, northwestern Georgia. *Geological Society of America Bulletin* 86: 287–291.
- Watts WA (1975b) A late-Quaternary record of vegetation from Lake Annie, south-central Florida. *Geology* 3: 344–346.
- Watts WA (1979) Late Quaternary vegetation of central Appalachia and the New Jersey coastal plain. *Ecological Monographs* 49: 427–469.
- Watts WA (1980) Late-Quaternary vegetation history at White Pond on the Inner Coastal Plain of South Carolina. *Quaternary Research* 13: 187–199.
- Watts WA and Hansen BCS (1994) Pre-Holocene and Holocene pollen records of vegetation history from the Florida peninsula and their climatic implications. *Palaeogeography, Palaeoclimatology, Palaeoecology* 109: 163–176.
- Watts WA, Hansen BCS, and Grimm EC (1992) Camel Lake: A 40000-yr record of vegetational and forest history from northwest Florida. *Ecology* 73: 1056–1066.
- Webb T III, Anderson KH, Bartlein PJ, et al. (1998) Late Quaternary climate change in eastern North America: A comparison of pollen-derived estimates with climate model results. *Quaternary Science Reviews* 17: 587–606.
- Webb T III, Bartlein PJ, Harrison S, et al. (1993) Vegetation, lake levels, and climate in eastern North America for the past 18,000 years. In: Wright HE Jr., Kutzbach JE, and Webb T, III et al. (eds.) *Global Climates Since the Last Glacial Maximum*, pp. 415–467. Minneapolis, MN: University of Minnesota Press.
- Whitehead DR (1965) Palynology and Pleistocene phytogeography of unglaciated eastern North America. In: Wright HE Jr. and Frey DG (eds.) *The Quaternary of the United States*, pp. 417–432. Princeton, NJ: Princeton University Press.
- Whitehead DR (1972) Developmental and environmental history of the Dismal Swamp. *Ecological Monographs* 42: 301–315.
- Whitehead DR (1973) Late Wisconsin vegetational changes in unglaciated eastern North America. *Quaternary Research* 3: 621–631.
- Whitmore J, Gajewski K, Sawada M, et al. (2005) Modern pollen data from North America and Greenland for multi-scale paleoenvironmental applications. *Quaternary Science Reviews* 24: 1828–1848.
- Wilkins GR, Delcourt PA, Delcourt HR, et al. (1991) Paleoeecology of central Kentucky since the last glacial maximum. *Quaternary Research* 36: 224–239.
- Willard D, Bernhardt C, Brown R, et al. (2011) Development and application of a pollen-based paleohydrologic reconstruction from the Lower Roanoke River Basin, North Carolina, USA. *The Holocene* 21: 305–317.
- Willard DA, Bernhardt CE, Brooks GR, et al. (2007) Deglacial climate variability in central Florida, USA. *Palaeogeography, Palaeoclimatology, Palaeoecology* 251: 366–382.
- Willard DA, Bernhardt CE, Korejwo DA, et al. (2005) Impact of millennial-scale Holocene climate variability on eastern North American terrestrial ecosystems: Pollen-based climatic reconstruction. *Global and Planetary Change* 47: 17–35.
- Willard DA, Bernhardt CE, Newell WL, et al. (2010) Deglacial to Holocene sedimentary records from Chesapeake Bay. *Geological Society of America Abstracts with Programs* 42: 176.
- Willard DA, Cronin TM, and Verardo S (2003) Late-Holocene climate and ecosystem history from Chesapeake Bay sediment cores, USA. *The Holocene* 13: 201–214.
- Willard DA, Weimer LM, and Holmes CW (2001) The Florida Everglades ecosystem: Climatic and anthropogenic impacts over the last two millennia. *Bulletins of American Paleontology* 361: 41–55.
- Zayac T, Rich F, and Newson L (2001) The paleoecology and depositional environments of the McClelland Sandpit Site, Douglas, Georgia. *Southeastern Geology* 40: 259–272.

Southwestern North America

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Landscape and Climate

Dynamic change and great vertical and spatial diversity have characterized the Holocene vegetation of the western and southwestern portions of the North American continent. In the United States, this region encompasses the modern states of California, Nevada, Utah, Arizona, New Mexico, and adjacent regions of Oregon, Idaho, and Texas and, in northern Mexico, the states of Baja California, Baja California Sur, Sonora, Sinaloa, Chihuahua, Durango, and western Coahuila. Driven by the interaction of climate, topography, and distance from the Pacific Ocean for some areas and from the Gulf of Mexico in other areas, vegetation communities as diverse as desert shrub, semiarid woodlands, montane forests, and tundra can be found within relatively short distances of each other in the region. Topographically, the region varies from sea level coastal plains, deep interior depressions over 200 ft below sea level, to extensive interior plateaus sometimes lying above 1300 m with flanking mountain ranges with peaks well above 3000 m (Hunt, 1967). The north-south to northwest-southeast trending mountain ranges which dominate the region serve both to impede the flow of moisture-laden storms from the Pacific to the west and to channel monsoonal storms northward from the Gulf of California and the Gulf of Mexico.

For discussion purposes, the area can be divided into three large regions based upon the dominant climate. These include (1) most of California west of the crest of the Sierra Nevada Mountains and north of the Transverse Ranges at the southern end of the Central Valley, (2) the high-elevation interior plateau known as the Great Basin which lies east of the crest of the Sierra Nevada Mountains, and stretches to the Wasatch Range to the east and to the Ochoco Mountains in the north, and (3) the Desert Southwest which lies south and southeast of the Great Basin and south of the Transverse Ranges in southern California and extending into northern Mexico in both areas.

Coastal vegetation ranges from rainforest in the north to chaparral in the south. Before Euro-American settlement, the interior Central Valley was characterized by mixtures of shrubs, grasses, and other herbaceous plants in the north and graded southward into shrubby steppe. Oak-dominated deciduous woodlands flanked (and to a lesser degree still flank) the Central Valley on all sides. Lying above the deciduous forests, the intermediate-elevation slopes of the Sierra Nevada Mountains still support pine-dominated montane forests. These forests transition into pine-dominated subalpine woodlands at higher elevations and, at even higher elevations, to subalpine grasslands and tundra.

East of the crest of the Sierra Nevada Mountains, the lowest elevation basins in the Great Basin are characterized by shadscale (saltbush)-dominated desert shrub communities (Billings, 1951). On the surrounding slopes, these transition into sagebrush-dominated steppe. Piñon pine and juniper dominate semiarid woodlands that cover the intermediate mountain slopes of the region. In the case of the lower

mountain ranges in this region, semiarid woodland comprises the highest-elevation vegetation community. Limber pine-dominated subalpine woodlands are found above the semiarid woodland in a few of the higher ranges. Only the highest mountain ranges of the region support narrow zones of ponderosa pine- and white fir-dominated montane forests between the semiarid and subalpine woodlands. Subalpine grasslands and/or alpine tundra occur rarely above the subalpine woodland and only on the highest mountain ranges.

Similar to the Great Basin, the Desert Southwest is a region of dramatic contrasts in topographically governed microclimate and vegetation. Mojave Desert vegetation characterizes most of southern Nevada, eastern California, and northwestern Arizona. The most common valley floor vegetation community within the Mojave Desert is desert shrub dominated by creosote bush and white bursage. Joshua tree is characteristic of many of the alluvial fans surrounding the broad basins of this area. However, the vegetation of the region also includes a great diversity of other shrubs and cacti. In fact, desert vegetation diversity increases southeastward from the Great Basin into southern Arizona. In central Arizona, the Mojave Desert vegetation gives way to the richly diverse Sonoran Desert. A great variety of cacti characterize the vegetation communities of the Sonoran Desert. Saguaro cactus, paloverde, and brittlebush are common in these communities. In extreme southeastern Arizona and southwestern New Mexico, the Sonoran Desert gives way to the Chihuahuan Desert. Primarily a shrub desert lying above 1100 m, most of the Chihuahuan Desert lies further south in the Mexican states of Chihuahua, Coahuila, and parts of Durango, Zacatecas, Nuevo Leon, and San Luis Potosi (Lowe, 1977). Much of the altiplano is dominated by Chihuahuan Desert shrub vegetation. Tarbush, creosote bush, and whitethorn are among the most abundant species represented in these communities, though grasses, yuccas, and cacti are also common (Lowe, 1977). Changes in soils often result in dramatic shifts in vegetation community composition over very short distances in the Chihuahuan Desert (Lowe, 1977, p. 21).

Piñon pine and juniper codominate semiarid woodlands that cover many of the lower-elevation mountain ranges and the intermediate-elevation slopes of the higher ranges of the northern portion of the Desert Southwest (Lowe, 1977). Above this on the higher ranges in the western portion of this region, limber pine and, occasionally, white fir occur. With increasing elevation in the northeastern part of this region, semiarid woodland transitions into montane woodland dominated by ponderosa pine. Above this, fir becomes more common; on the highest mountains, spruce may be found mixed with fir.

In southern Arizona, however, oak and oak pine woodlands and chaparral replace piñon-juniper semiarid woodlands on intermediate-elevation slopes. Higher mountain ranges still support ponderosa pine at lower elevations in the montane forest and fir or spruce-fir forests at highest elevations. Climatologically, the three regions can be characterized by the seasonal distribution of rainfall and temperature. California north of

the Transverse Ranges can be divided into a northern area that is characterized by cool wet winters and warm dry summers. The southern portion of this area is dominated by warm moist winters and hot, dry to moist summers. In the Great Basin, cold, moist winters and hot, dry summers are the norm. The Desert Southwest is characterized by warm, dry winters and hot, moist summers. The climate of these regions reflects the effects of both the heterogeneous topography and the varying impact of three major air mass systems that intersect over it (Houghton et al., 1975). The predominance of these air masses varies both seasonally and annually to influence the weather of the region today. In the past, changes in their average position have resulted in significant changes in the distribution of vegetation communities within the region and upon the animals and peoples that have depended upon them. These systems converge upon the central Great Basin from the west, southeast, and northeast and can be described as follows:

- *Pacific*: a regime dominated by maritime polar air masses. These moist, cool air masses sweeping in from the Pacific Ocean to the west produce cool, wet winters, and their absence in summer makes the growing season hot and dry.
- *Gulf*: a regime dominated by maritime and continental tropical air masses. These warm, moist air masses entering the southwest from either the Gulf of Mexico or the Gulf of California to the south produce hot, moist summers with midsummer torrential rainfall (originating during convective storms) and warm dry winters.
- *Polar*: a regime dominated by continental polar and Arctic air masses. These cold dry continental air masses drop down from the North American interior and extend into the Great Basin from the northeast during the winter.

Movements of these air systems, and their impact on local and regional climate, are reflected in changes in hydrology, erosion and deposition processes, and vegetation. Displacement of winter and summer storm tracks and variations in the penetration of the summer monsoon are all affected by the realignment of these pressure systems through time.

Pollen Records

Climate not only affects the spatial distribution of vegetation in the region, but it also determines where well-preserved pollen records are found. The distribution of pollen (palynological) evidence in the region is primarily constrained by moisture differences. The major sources of late Quaternary pollen records in the greater southwest include lacustrine and marsh deposits, dry cave sediments, alluvium and cienega deposits, and ancient woodrat middens (a source unique for the North American record).

In general, the number of well-preserved pollen records from aquatic environments at all elevations decreases southward in this region, reflecting the greater rarity of lakes and marshes due to greater aridity. Most lacustrine records span only a few thousand years, though some span the full Holocene and, in rare cases, extend well into the Pleistocene. Springs and associated ephemeral ponds are the primary source for well-preserved Holocene pollen records in the southern portion of

this region. These often provide records spanning the Holocene; however, the destruction by drought of pollen during the middle Holocene drought is often a problem. Also, they are often complex and difficult to interpret because of dramatic changes in deposition rates and occasional hiatuses. However, high organic production at some of these sites resulting in rapid rates of sediment accumulation makes them amenable to the formation of high-frequency vegetation histories. Such records can be compared with tree-ring data to generate detailed records of regional climate dynamics and vegetation response.

Pollen from ancient woodrat middens usually is excellently preserved by the urine that encases the nest materials. Thus, they can provide very detailed records of past vegetation. However, their interpretation must be constrained by the fact that they primarily represent the collecting activities of woodrats. These activities actually provide some pollen types which may be underrepresented or absent from the airfall pollen record. Because they are usually found in sheltered localities, woodrat middens have much less airfall pollen than open locations. Woodrat midden pollen samples are obtained from the radiocarbon-dated layers of crystallized urine-encased nest materials that comprise an ancient nest (Betancourt et al., 1990). Each pollen spectrum obtained from individual layers represents a single 'snapshot' of a particular point in time at a specific place on the landscape.

Alluvial, cienega, and cave pollen samples from the North American west and southwest suffer from stratigraphic or, more specifically, chronometric problems. Alluvial and cienega deposits can represent highly variable depositional environments that contain scant organic materials for radiocarbon dating. Deposits in the dry caves of the interior often contain abundant well-preserved pollen, as well as material for radiocarbon dating, but because rodents nest in the loose deposits, they are often heavily mixed. In comparison with mechanically abraded alluvial pollen, and cienega pollen that has been subject to episodes of degradation, cave deposits are often extremely well preserved.

The Data

Pollen records from the west and southwest are not as abundant as those from the eastern or northwestern portions of North America. In addition, many of the analyzed records have either not been published or are only partially published. Many are only recorded in doctoral dissertations, master's thesis, or project reports or 'gray literature.' More recently, some have been published on the Internet. The coverage in this article is primarily restricted to those sites that have recently been published in peer-reviewed literature and provide more detailed, regionally representative vegetation and climate histories. In a few cases, however, some 'gray literature' will be included where it provides needed continuity in the chronometric and spatial pollen record. Older palynological research is summarized in several review articles by Mehringer (1985), Hall (1985), Adam (1985), Mehringer (1986), and Martin and Mehringer (1965), and more recent research in the Great Basin by Wigand and Rhode (2002).

The Holocene vegetation record from the west and southwest is as diverse as its landscape. However, it can be

characterized by several general trends as well as by crosscutting climatic events that are encountered in most of the region. These events can be tracked by tracing the response of climate indicator plant species that are characterized by wide spatial distributions and relatively rapid response to changes in climate. Differences in the magnitude of these events from one area to the next reflect the varying predominance of the major climate influencing air masses discussed above. Some of the salient trends and crosscutting events from a regionwide perspective are summarized below. Select sites will be briefly discussed later. Note that dates have been given in calibrated years before present (cal BP) and radiocarbon years BP (rcyr BP).

Vegetation and Climate History of the North American West and Southwest

General Summary

The Pleistocene/Holocene transition has been variously placed at about 13 900 cal BP (12 000 rcyr BP) or 11 540 cal BP (10 000 rcyr BP) based upon geologic, glacial, botanical, or other criteria. We will simply indicate that the major break in vegetation assemblages between the Pleistocene and Holocene in the west is based upon the use of dendrograms generated by dissimilarity coefficient analysis of pollen spectra. These analyses place the break between the Pleistocene and the Holocene usually about 13 900 cal BP (12 000 rcyr BP) to 13 470 cal BP (11 500 rcyr BP). The interval following this break, which corresponds to the Younger Dryas in Europe, is usually characterized by initial drought followed by effectively moister conditions between 13 120 cal BP (11 200 rcyr BP) and 11 920 cal BP (10 200 rcyr BP), probably driven by colder temperatures and reduced evaporation rates rather than by an increase in precipitation. The vegetation of that interval is purely Holocene in nature.

Warm, moist conditions seem to have predominated much of the interior of the region between 11 920 cal BP (10 200 rcyr BP) and 9530 cal BP (8500 rcyr BP). These conditions may have resulted from increased late spring through mid-summer precipitation as shown by the pollen record, woodrat midden plant macrofossils (Wigand and Rhode, 2002), and spring discharge in the northern Mohave Desert (Quade et al., 1998).

During the middle Holocene, much warmer and drier conditions seem to have prevailed in much of the northern portion of the interior west between 8910 cal BP (8000 rcyr BP) and 6290 cal BP (5500 rcyr BP). However, between 9660 cal BP (8600 rcyr BP) and 6290 cal BP (5500 rcyr BP) the southern part of the region apparently enjoyed a significant increase in late spring or mid-summer rainfall. These monsoonal climates gave way to conditions that were relatively drier after 3030 cal BP (2800 rcyr BP). Between 6440 cal BP (5600 rcyr BP) and 6290 cal BP (5500 rcyr BP) a major episode of increased winter precipitation recorded from pollen localities as far north as the Plateau of eastern Washington to well into the northern Mojave Desert, and southern California brought the middle Holocene drought to a sudden halt. Although there was a brief return to slightly drier conditions after this event, climate was never as severe as before 6290 cal BP (5500 rcyr BP).

Increasing winter precipitation after 5820 cal BP (5000 rcyr BP) climaxed in the northern interior between 4570 cal BP (4000 rcyr BP) and 2670 cal BP (2600 rcyr BP) during the

Neopluvial. Renewed drought cycles characterized the interval between 2670 cal BP (2600 rcyr BP) and 1700 cal BP (1700 rcyr BP), but were interrupted by a dramatic winter wet event centered between about 2140 cal BP (2100 rcyr BP) and 1850 cal BP (1900 rcyr BP). Between about 1700 cal BP (1700 rcyr BP) and 910 cal BP (1000 rcyr BP) increased grass abundance, piñon pine expansion, and increased water depth in previously marginal ponds during the summer confirm a period of highly variable, near-decadal variations in precipitation with a shift from winter to late spring to mid-summer dominance.

A return to winter-dominated precipitation in the northern portion of the region about 910 cal BP (1000 rcyr BP) signaled the beginning of generally much drier conditions, punctuated by a significant warm, moist event about 690 cal BP (730 rcyr BP). A return to drier conditions followed until about 240 cal BP (250 rcyr BP), when a return to cooler, moister conditions may signal the last phase of the Little Ice Age in the region. These conditions ended about 150 years ago. Vegetation since then probably reflects the impact of both the Euro-American settlement as well as climate. A few key pollen records whose locations can be found on [Figure 1](#) are discussed below.

The Details

Coastal California The Clear Lake microfossil record from the north central coastal range of California reveals the dramatic shift from pine and cypress (*Pinus* and *Cupressaceae*) family dominated conifer forest to oak (*Quercus*)-dominated



Figure 1 Location of published, radiocarbon-dated Holocene pollen records from the western and southwestern United States mentioned in the text. Most of these are not in the NOAA National Climatic Data Center's North American Pollen Database. The sites are 1, Diamond Pond; 2, Fish Lake; 3, Wildhorse Lake; 4, Clear Lake; 5, Little Valley; 6, Lead Lake; 7, Hidden Cave; 8, Ruby Marshes; 9, Crescent Spring; 10, Exchequer Meadow; 11, Lower Pahranaagat Lake; 12, Tulare Lake; 13, Ballona Estuary; 14, San Joaquin Marsh; 15, Montezuma Well; 16, Twin Lakes; 17, Beef Pasture; 18, Double Adobe; and 19, Summit Lake.

deciduous forest during the Pleistocene/Holocene transition (Adam, 1988; see (Northwestern North America)). With a sample roughly every 500 years, this record reveals a maximum extent of oak forest about 8340 cal BP (7500 rcyr BP). Thereafter, various shrubs and low-growing trees in the buckthorn family (*Ceanothus* and *Rhamnus*) became more abundant, remaining so (except for the period between about 4570 cal BP (4000 rcyr BP) to 2670 cal BP (2600 rcyr BP)) until the present (Adam, 1988). Chinquapin (*Chrysolepis*), alder (*Alnus*), willow (*Salix*), and less commonly sagebrush (*Artemisia*) and grasses comprise the major part of the Holocene plant community in the area.

In east central California the record from Exchequer Meadow located in the montane forest on the west slope of the Sierra Nevada Mountains indicates that the transition from a sagebrush-dominated alpine tundra and wet sedge meadow community to a pine (*Pinus*)-fir (*Abies*)-dominated montane forest had begun as early as 13 910 cal BP (12 000 rcyr BP) (Davis and Moratto, 1988). However, it was not fully established until about 9530 cal BP (8500 rcyr BP). Lower elevation deciduous oak forest expansion is reflected in the rise of oak (*Quercus*) pollen about 13 470 cal BP (11 500 rcyr BP), but it is not until about 10 130 cal BP (9000 rcyr BP) that it is clearly

well established at lower elevations in the region. After a middle Holocene lull, renewed abundance of sedge (Cyperaceae) pollen about 6290 cal BP (5500 rcyr BP) reflects increased regional moisture evidenced in pollen records elsewhere in the region (Davis et al., 1985).

The Holocene portion of a pollen record from the Tulare Lake Basin near the southern end of the Central Valley records a history of lake and marsh fluctuation in its littoral vegetation, primarily sedge and cattail (*Typha*) pollen, and pelagic algae (primarily *Pediastrum* sp. and *Botryococcus* sp.) (Davis, 1999) (Figure 2). The absence of a pollen or algae record for the interval about 11 500 BP may indicate that the basin was totally dry during the so-called Clovis drought (Davis, 1999, Figure 3; Haynes, 1991). Pollen from the base of the Holocene section of the Tulare Lake core dating to about 11 540 cal BP (10 000 rcyr BP) suggests that by about 11 920 cal BP (10 200 rcyr BP), marsh vegetation (as indicated by the presence of littoral vegetation pollen) is restricted, but the abundance of pelagic algae at about 10 020 cal BP (8900 rcyr BP) and 8860 cal BP (7900 rcyr BP) indicates at least two episodes of deep lake (Davis, 1999, Figure 3). The terrestrial pollen indicates the nearby presence of an extensive juniper woodland with a

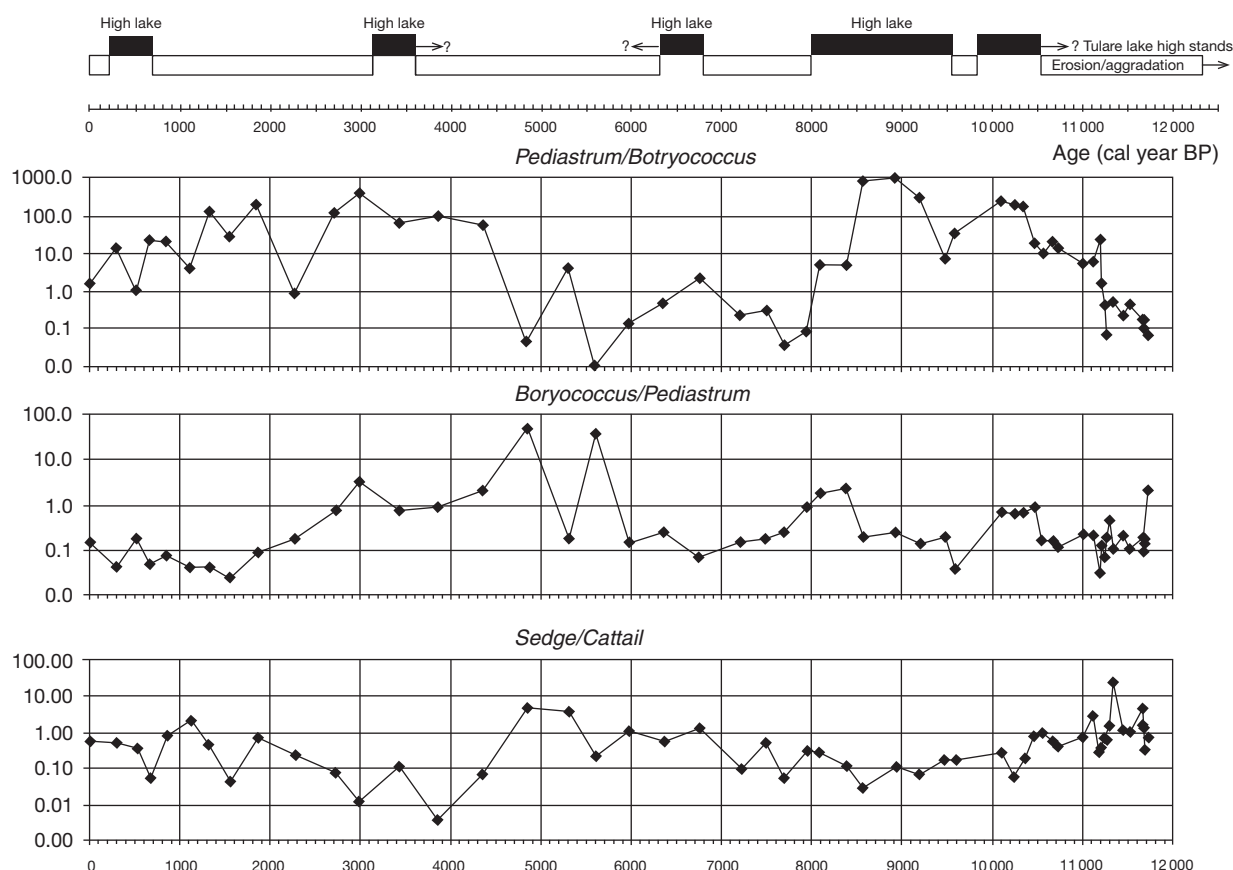


Figure 2 Comparison of reconstructed Holocene high lake stands of Tulare Lake with ratios of littoral marsh species and pelagic algae. High stands were mapped in shoreline trenches by Dr. R. Negrini and this author between 2003 and 2007. The ratio of *Pediastrum* to *Botryococcus* algae indicates slightly more oligotrophic (fresher) water conditions versus more eutrophic (stagnant) water conditions. *Pediastrum* abundance suggests recent recharge of the lake. The sedge-to-cattail ratio reflects marsh extent and stability. Cattails appear when there is renewed influx of fresher water into the Tulare Lake Basin. Sedges predominate when there are extended periods of low water. Note that sedge is usually abundant between high lake stands and lower during high lake stands. The algae and pollen used in these ratios have been standardized with respect to pollen abundance per sample so that they reflect actual increases and decreases in the pollen of these species in the record.

sagebrush understory that disappeared by 8910 cal BP (8000 rcyr BP) to be replaced by a saltbush (*Chenopodiaceae-Amaranthus* pollen)-dominated desert shrub community. Copious sunflower family (primarily *Aster*-type) pollen from the Holocene record suggests the abundance of both shrubs and forbs of the Asteraceae around Tulare Lake (Davis, 1999, Figure 3). The greatest abundance of aster-type pollen occurs between 7880 cal BP (7000 rcyr BP) and 6290 cal BP (5500 rcyr BP), suggesting increased late-spring to middle-summer precipitation. Oak pollen from the Tulare Lake record suggests that oak was regionally common in the hills surrounding the basin between 10 130 cal BP (9000 rcyr BP) and 3830 cal BP (3500 rcyr BP) (Davis, 1999, Figure 3).

After episodic middle Holocene drought, increasingly wetter conditions are marked by the expansion of littoral vegetation about 5220 cal BP (4500 rcyr BP) (Davis, 1999, Figure 3). The Holocene climax of marsh expansion about 4660 cal BP (4100 rcyr BP) is followed by an abrupt decline of littoral vegetation and its replacement by pelagic algae,

indicating that water depth in the Tulare Lake Basin rose rapidly to drown the marsh, resulting in a dramatic expansion of the lake just after 4570 cal BP (4000 rcyr BP) (Davis, 1999, Figure 3). The climax of pine pollen (a reflection of increased winter rainfall) in the Tulare Lake record at the same time as the pelagic algae (about 2880 cal BP (2700 rcyr BP) to 3030 cal BP (2800 rcyr BP)) suggests a regional increase in precipitation at that time (Davis, 1999, Figure 3). Extremely low charcoal abundance during this interval also suggests wetter conditions. Increasing saltbush pollen coupled with decreasing pine, oak, and littoral vegetation pollen, as well as declining pelagic algae abundance, suggests significantly drier conditions during the last 1980 cal BP (2000 rcyr BP) (Davis, 1999, Figure 3).

A 7-ka-long pollen record from the San Joaquin Marsh provides additional evidence that coastal southern California had a moister climate during the middle Holocene (Davis, 1992). Prior to 3200 cal BP (3000 rcyr BP), pollen from *Aster*-type plants dominated the pollen record (Davis, 1992, Figure 3). After that, pollen from plants from the saltbush family

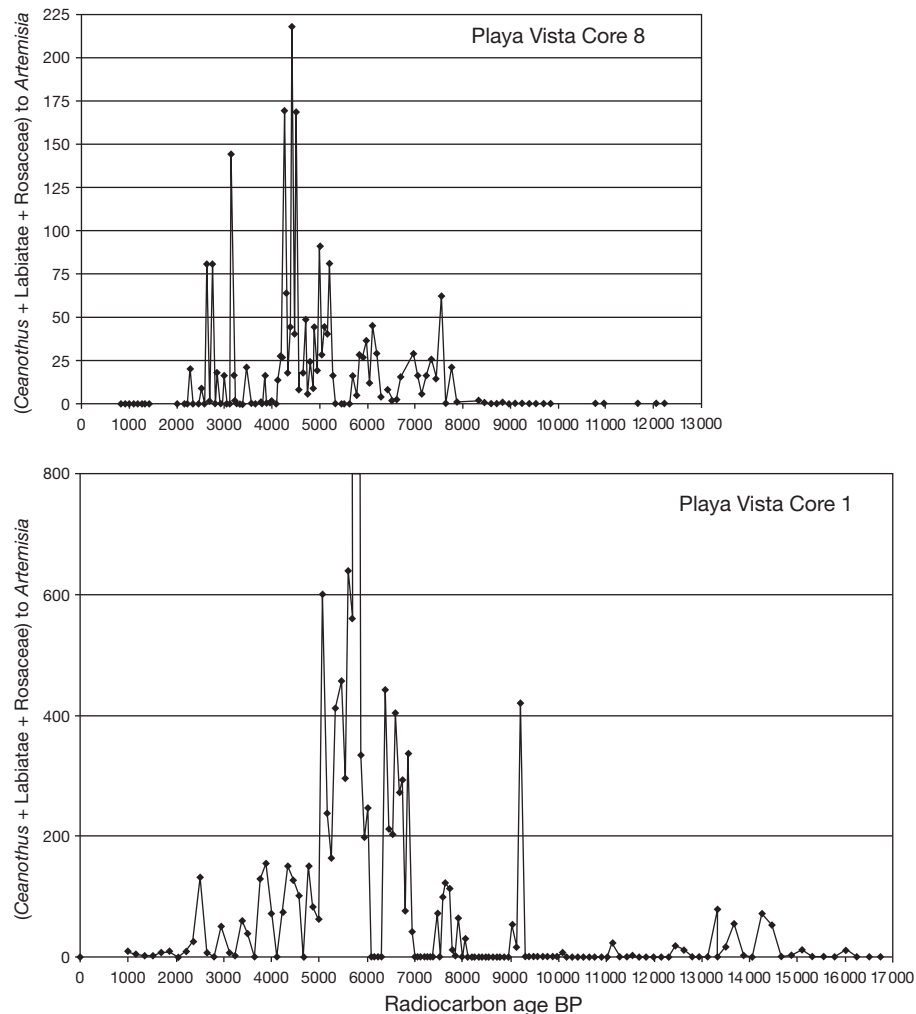


Figure 3 Comparison of the ratio of *Ceanothus* + *Labiatae* + *Rosaceae* to *Artemisia* pollen for both Playa Vista Cores 1 and 8. The primary source of *Ceanothus*, *Labiatae*, and *Rosaceae* pollen is the moister climate coastal chaparral. *Artemisia* (primarily California sagebrush) is characteristic of the drier coastal chaparral. The period of greatest chaparral expansion between 8000 and 5000 rcyr BP corresponds to events throughout the southwestern United States.

predominates. In addition, the greater abundance of pine, oak, buckthorn (*Ceanothus*), and sagebrush pollen provides further evidence of moister climate. Higher values of Liguliflorae-type pollen prior to 5820 cal BP (5000 rcyr BP) suggest that this moisture may have been in the form of late-spring through midsummer precipitation. Ongoing reanalysis by this author of pollen from 12- to 16-ka-long cores from the Ballona Estuary in the southwest corner of the Los Angeles basin confirms this conclusion. There, pollen from shrubs in the mint (*Lamiaceae*), buckthorn (*Ceanothus*), and rose (*Rosaceae*) families increases in abundance, indicating a major chaparral expansion between 8 and 5 ka ([Figure 3](#)).

A pollen record from Laguna Canyon south of Los Angeles indicates that lake level rose as a result of increased winter precipitation along the southern coast of California, resulting in extensive flooding of the surrounding valley floor. Sustained flooding established a sedge-rich marsh, though there was some tall coastal cattail as well. For brief intervals, the lake levels may have been slightly deeper, allowing the establishment of emergent aquatic plants such as water smartweed. The lake(s) may have ended not as a result of drier climate but probably as a result of a breach in the sill which maintained the lake in Laguna Canyon.

Summer paleoclimatic reconstructions using tree-ring records have been utilized in analyzing the salt variations in Soda Lake, a saline lake in the Carrizo Plain just east of the Temblor Range in the southeastern portion of the Central Valley of California. Limber pine tree-ring cores from the Transverse Ranges just south-southeast of Soda Lake indicate that there is a relatively strong correspondence between mesic and xeric conditions and the percentage of salt in the lake and possibly in the kind of salt being precipitated. The analysis also documented a major expansion of Jeffrey pines onto the upper mountain slopes, beginning about 225 years ago. Stand establishment dates indicate that limber pines dominated the upper mountain slopes during the last millennium until 225 years ago, when Jeffrey pine became dominant. This shift indicates both a lengthening of the growing season in the area and a post-Little Ice Age increase in precipitation. This research is ongoing but holds tremendous promise for tracking the climate of the last few hundred years in the region.

Great Basin. Three pollen records from an elevational transect up the west slope of Steens Mountain in south central Oregon provide a record of Holocene vegetation change ([Mehring, 1985, 1986](#); [Wigand, 1987](#); [Johnson et al., 1994](#)). The longest record from Fish Lake (2250-m elevation) provides a 13-ka record of high-elevation sagebrush community dynamics. A ratio of sagebrush to grass pollen indicates that drier conditions predominated between 8910 cal BP (8000 rcyr BP) and 6290 cal BP (5500 rcyr BP) ([Mehring, 1985](#)). During the same interval, sagebrush became more common than grass around Wildhorse Lake at 2565-m elevation. The 9.4-ka-long record from Wildhorse Lake indicates that grasses did not regain their early Holocene dominance until after 4570 cal BP (4000 rcyr BP) ([Mehring, 1985](#)). Diamond Pond, at 1265 m at the base of the northwest slope of Steens Mountain, records the shifting boundary between the lower sagebrush and desert shrub communities during the last 6000 radiocarbon years. A brief but major wet episode about 6290 cal BP (5500 rcyr BP) resulted in a shift from greasewood

(*Sarcobatus*) to sagebrush predominance in the shrub communities around Diamond Pond. Between 4570 cal BP (4000 rcyr BP) and 1980 cal BP (2000 rcyr BP) a further shift to wetter conditions resulted in the expansion of western juniper (*Juniperus occidentalis*) in the volcanic complex around and east of Diamond Pond ([Mehring and Wigand, 1990](#)). Close correspondence between the pollen record from Diamond Pond and the macrofossil record from ancient woodrat middens in the same area documents the actual arrival and subsequent departure of western juniper. Juniper macrofossils from the ancient woodrat middens also permitted direct identification of western juniper in the pollen record from Diamond Pond ([Mehring and Wigand, 1990](#)).

The charcoal record reconstructed from microscopic charcoal counted on the pollen slides reveals a dynamic relationship between fuel accumulation during wet climate episodes and fire during subsequent drought in the Diamond Craters complex ([Wigand, 1987](#)). It also indicates that when climates became sufficiently dry after 1980 cal BP (2000 rcyr BP), catastrophic fires resulted in the ultimate demise of juniper woodland at lower elevation sites, such as those around Diamond Craters ([Figure 4\(a\)-4\(c\)](#)). Such fires characterized other intervals as well, when forests that expanded during episodes of wetter climate were destroyed at their lower elevational distribution by fire. Increased grass pollen relative to juniper pollen between 1630 cal BP (1600 rcyr BP) and 910 cal BP (1000 rcyr BP) indicates a period when the precipitation

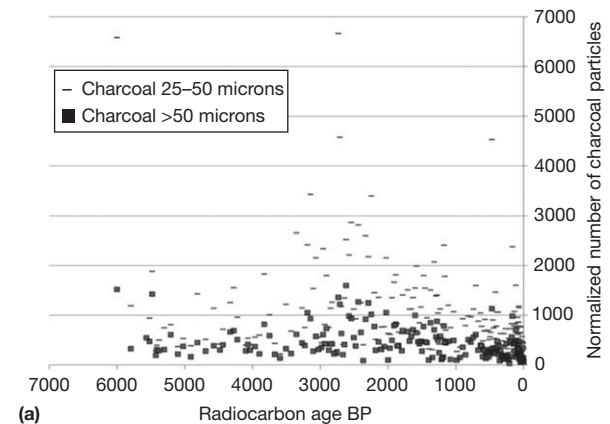


Figure 4 (a) Influx rates of two sizes of charcoal at Diamond Pond, Oregon, for the ~6000 years. Charcoal abundance (as a measure of fire frequency and magnitude) shifts on a centennial to millennial scale. (b) In general, charcoal increases in abundance and size during extended episodes of wetter climate. On the other hand, it decreases in abundance and size during extended episodes of drier climate. During the transition from a period of drier climate to one of wetter climate, charcoal abundance (fire occurrence) is lower than during the later part of a mesic episode when periodic droughts begin to punctuate the wet episode. (c) Comparison of the charcoal-to-pollen ratio from Diamond Pond and Summit Lake in the northern Great Basin. Diamond Pond is located on the ecotone between the desert shrub and lower sagebrush communities. Summit Lake lies within the upper sagebrush community in the Black Rock Range of north central Nevada. There are other charcoal records in the region that are similar in the timing of charcoal abundance suggesting periods of increased fire frequency due to climate.

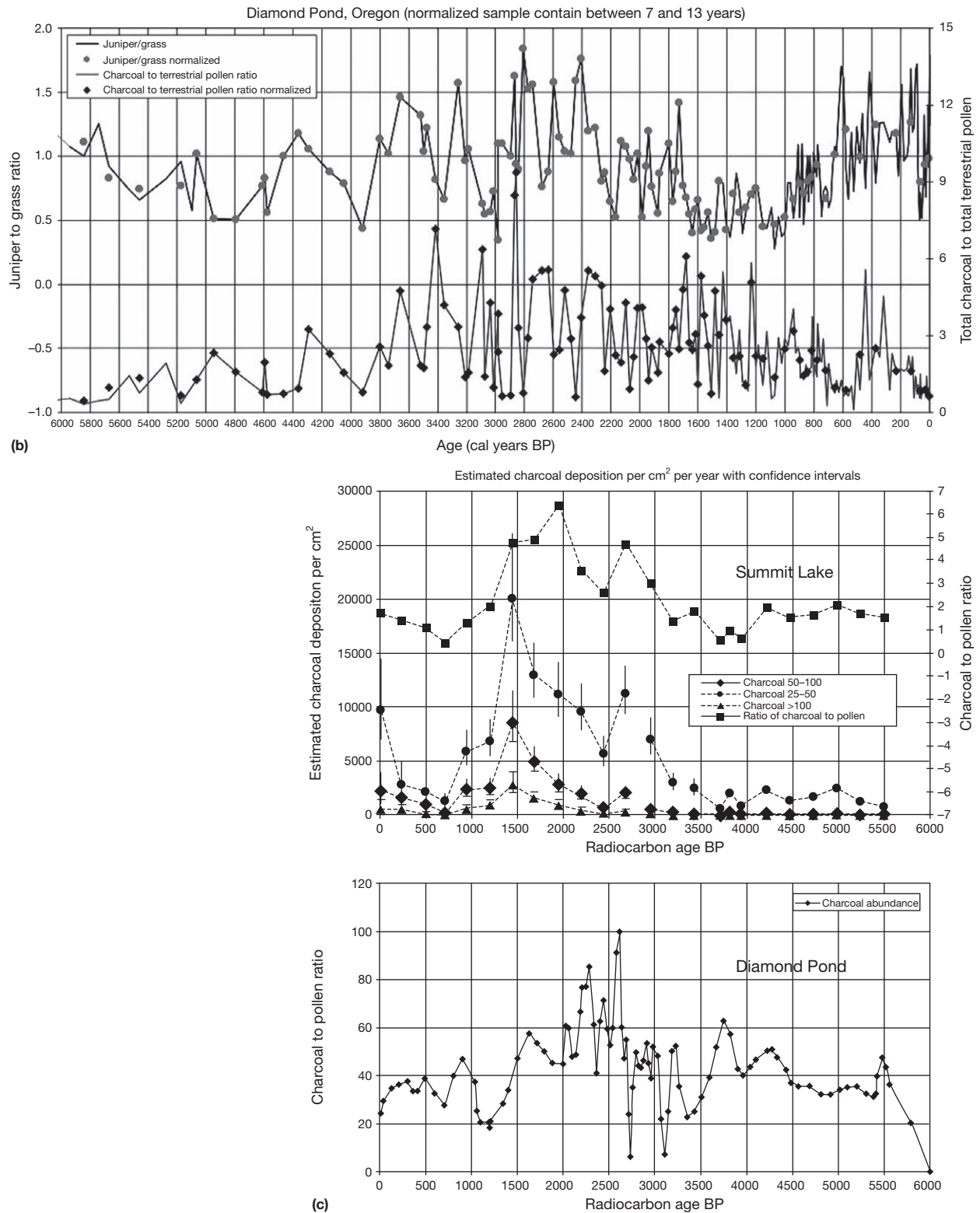


Figure 4 (Continued)

pattern shifted from winter-dominated to a late-spring through midsummer pattern (Wigand, 1987).

A regionally restricted and very brief climate event is found in a transect of pollen records stretching from Sardine Meadow (south of the Sierra Valley) on the western end, through Summit Lake in the Black Rock Range, to Grays Lake in southeastern Idaho. These records have revealed a striking warm, wet event ~730 rcyr BP that does not appear in records north or south of this swath (Figure 5). Today, this stretch of the northern Great Basin is unique in having high May and June precipitation relative to winter precipitation. Each of the pollen records mentioned above shows a dramatic increase in aquatic plant and algae production. In the Sierra Valley northwest of Reno, NV, this coincides with an archaeological site comprised of over 49 hearths that may have been used to process camas root (Wigand, 2003; Wigand, 2004a,b).

In the western Great Basin south of Reno, Nevada, a 5.5-ka-long record from Little Valley provides additional evidence of a period of wetter climate between 4580 cal BP (4000 rcyr BP) and 1980 cal BP (2000 rcyr BP) ago (Wigand and Rhode, 2002). Lying in the montane forest just east of Lake Tahoe,

the pollen from the Little Valley core reveals increasing pine abundance relative to sagebrush after 6290 cal BP (5500 rcyr BP). Increased alder, birch (*Betula*), and willow pollen between 4580 cal BP (4000 rcyr BP) and 1980 cal BP (2000 rcyr BP) indicates cooler, wetter conditions in the Little Valley meadows (Wigand and Rhode, 2002, Figure 11).

Expansion of marshes in the Carson Sink west of the Stillwater Mountains, as a result of increased precipitation and runoff from the Sierra Nevada Mountains during this interval, is confirmed by the appearance of abundant cattail pollen in the deposits of Hidden Cave (Wigand and Mehringer, 1985, Figure 36). Cattail pollen may have arrived naturally but most probably reflects human activity in the cave. However, today, the nearest marsh is kilometers distant from the cave, and its presence in the cave between 4 and 3 ka suggests that marshes were probably much closer to the site. Wetter conditions between 4580 cal BP (4000 rcyr BP) and 3200 cal BP (3000 rcyr BP) are confirmed by the relative increase in sagebrush pollen relative to saltbush (*Chenopodiaceae*) pollen at Crescent Spring, Utah, in the northwestern Bonneville Basin (Mehringer, 1985, Figure 11). Late Holocene precipitation

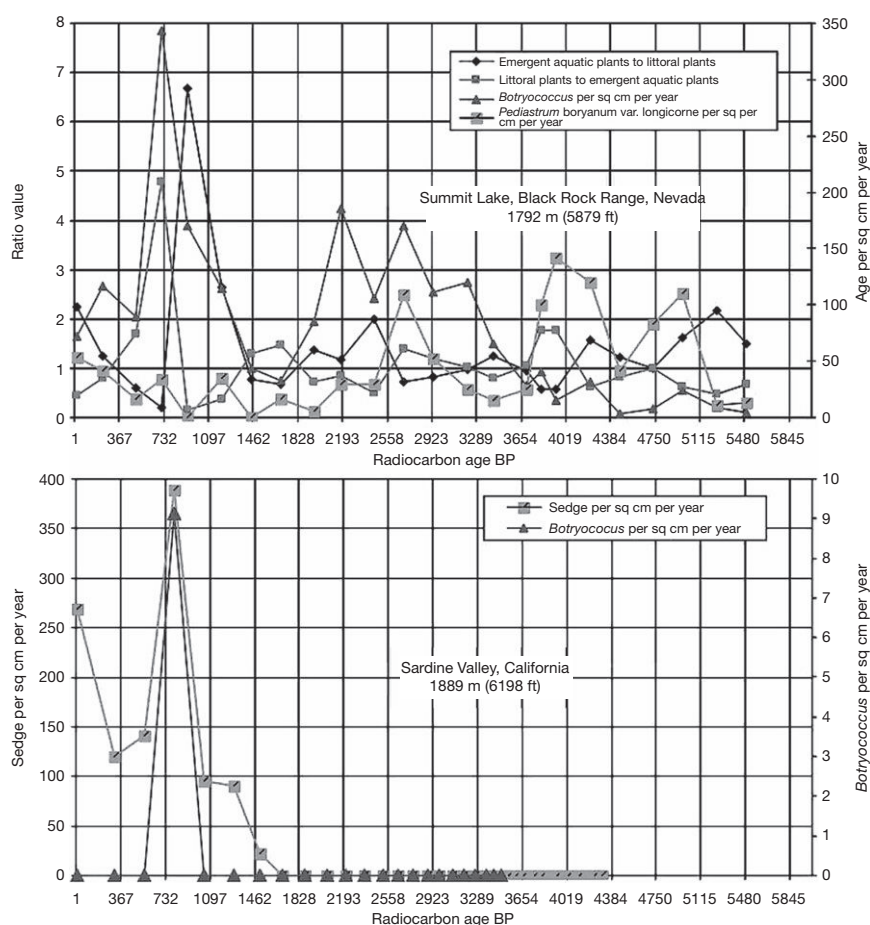


Figure 5 Pollen records in a band across the northern portion of the Great Basin are not revealing an episode of warm, moist climate centered about 730 rcyr BP. Key pollen and acid resistant algae types in records stretching from the Modoc Plateau to Gray's Lake in southeastern Idaho reveal a sudden increase in the discharge of springs and an increase in marsh depth. They also indicate a dramatic increase in the productivity of aquatic plant pollen and algae reflecting fresher water conditions and warmer climate. Here, the key pollen and algae in a record from northern California and north central Nevada are compared.

increase stands in stark contrast to the severe middle Holocene drought recorded in pollen from the sediments of Hidden Cave (Wigand and Mehringer, 1985) and from lake and marsh sediments in the Ruby Valley of east central Nevada by Thompson (1992).

Additional analysis of the Hidden Cave pollen record and of grain-size analyses run on many of the same strata from which the pollen was taken reveal a very close relationship between vegetation dynamics and surficial processes in the Carson Sink (Figure 6). During intervals of dramatic climatic transition ~8910 (8000) and 4580 (4000) years ago (when climates were transitioning into and out of the middle Holocene drought), vegetation assemblages fluctuate wildly. During the same interval, sediment analysis indicates that numerous colluvial sediment units are deposited in Hidden Cave. On the other hand, when vegetation was transitioning from late Pleistocene juniper woodlands and marshes bordering the Carson Sink about 10 130 cal BP (9000 rcyr BP), sediment analysis indicates an increase in units rich in aeolian sediments. This indicates that as the lake and marshes were drying, aeolian erosion and deposition began to increase (Wigand, 1997).

We have also found that comparison of ratios of several of the dominant pollen types found in the Hidden Cave record differentiates periods of summer/spring precipitation from periods of winter-dominated precipitation. The ratio of

'saltbush/sagebrush' pollen reveals the sequence of dry and wet cycles for the last 11540 cal BP (10 000 rcyr BP) (Figure 7). On the other hand, the ratio of 'grass/greasewood' pollen reveals the pattern of intervals when spring and summer precipitation becomes more common versus a winter-dominated precipitation pattern. This is also indicated by a rise in single-leaf piñon (*Pinus monophylla*) pollen.

A 2.3-ka-long core from the Carson Sink east of Fallon, Nevada, provides additional evidence of the late-spring through midsummer shift in annual precipitation, documented in the Diamond Pond core. A significant increase in pine pollen relative to juniper pollen between 1630 cal BP (1600 rcyr BP) and 910 cal BP (1000 rcyr BP) corresponds to an expansion of single-leaf piñon documented in the ancient woodrat nests of the region (Wigand and Rhode, 2002, Figures 13 and 14).

A recent pollen record from Pyramid Lake (Mensing et al., 2004) suggests that the early middle Holocene (7520 cal BP (6600 rcyr BP) to 6290 cal BP (5500 rcyr BP)) was extremely dry, although it included one short but intense wet phase. Evidence of a greatly reduced volume and depth of Pyramid Lake suggests that Lake Tahoe probably did not overflow its rim during this interval. The climate became much more mesic after 6180 cal BP (5400 rcyr BP) with episodes of moister climate corresponding with the Neopluvial between 4580 cal

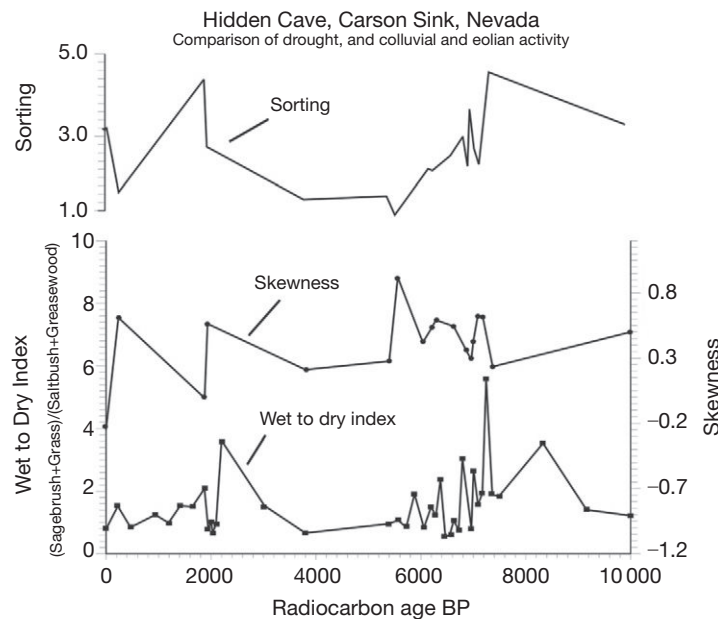


Figure 6 Ratios of major pollen types in samples from Hidden Cave compared with grain-size analysis of sediments of the same samples indicate a strong correlation between climate, vegetation response to climate change, and the dynamics of dominant surface erosion processes. Colluvial processes are responsible for most of the postglacial sediments that occur in the cave, but their nature varies considerably through the Holocene. Occasionally, the sediments are characterized by coarse grain size and poor sorting. This indicates that there was significant, high-energy slope wash, probably due to high precipitation events. At other times, the sediments are characterized by finer particle size and contain abundant aeolian sands that were deflated from the drying lake bottoms and marshes that lie near the cave. These sediments were deposited on the slopes above the cave and then swept into the cave during heavy rainfall events. At Hidden Cave, the early Holocene shift toward drier conditions (as the vegetation cover became less dense and pluvial Lake Lahontan dried) was punctuated by highly variable erosion rates dominated by high-energy colluvial events reflecting torrential rains. During the middle Holocene, aeolian sediments become more abundant and form a large portion of the cave's sediments though they are still being brought into the cave as colluvium. During the transition between the middle and late Holocene, colluvial sediments in the cave become coarser again and highly variable in their nature reflecting greater variability in climate as well as an increase in precipitation. Vegetation lagging behind the increase in precipitation cannot serve as a competent armor against soil erosion until its biomass has increased significantly.

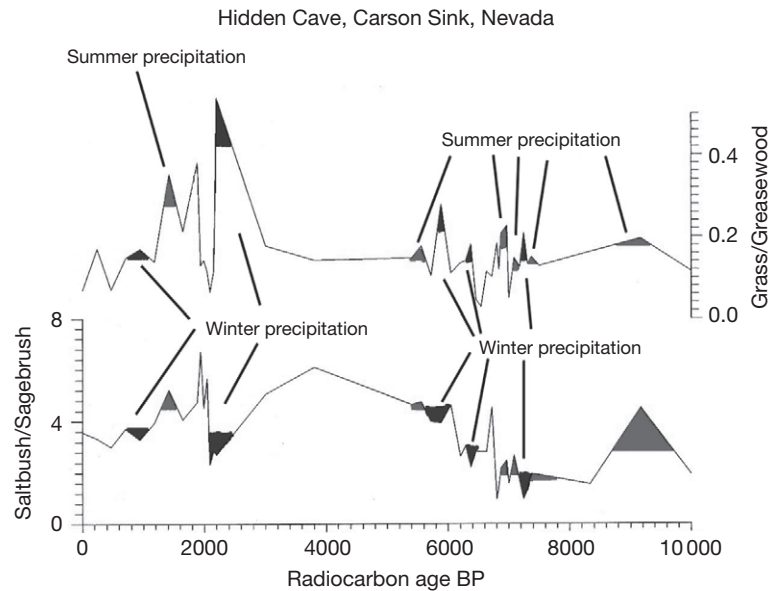


Figure 7 Ratios generated from some of the climate indicator pollen species in the pollen record at Hidden Cave offer a potential opportunity to monitor shifts in the seasonality of precipitation for the entire Holocene. In particular, the shift in predominance of winter versus summer precipitation is extremely important when the dynamics of vegetation around the cave and erosion processes are studied. The ratios of saltbush to sagebrush and grass to greasewood are used to reflect seasonality. An increase in both sagebrush and grass reflects increased winter precipitation. An increase in the abundance of both grass and greasewood without an increase in sagebrush pollen abundance indicates summer-shifted precipitation. That is, grass abundance increases in response to increases in both summer or winter precipitation. Sagebrush only increases with increased winter precipitation. However, when both saltbush and grass abundance increase (without an increase in sagebrush), a shift to summer-dominated precipitation is indicated. This diagram reveals several significant shifts between periods dominated by summer rainfall and others dominated by winter precipitation. In general, neither winter nor summer precipitation characterizes the area around Hidden Cave. However, there are some episodes where the shift is pronounced and persists for some time.

BP (4000 rcyr BP) and 1980 cal BP (2000 rcyr BP) (Mensing et al., 2004, Figure 5). The past 2.4 ka appear to have had three cycles of wetter climate (Mensing et al., 2004, Figure 5). One was centered around 1630 cal BP (1600 rcyr BP) and corresponds to the period of late-spring to summer-shifted precipitation postulated elsewhere (Wigand and Rhode, 2002). Wetter climate between 1132 cal BP (1100 rcyr BP) and 780 cal BP (800 rcyr BP) corresponds with a return to increased winter precipitation in the northern Great Basin (Wigand, 1987) and between 390 cal BP (350 rcyr BP) and 170 cal BP (150 rcyr BP) with the 'Little Ice Age' (Wigand and Rhode, 2002). The timing and magnitude of the Pyramid Lake wet/dry cycles compares favorably with those found in other pollen records (Wigand, 1987; Wigand and Rhode, 2002), in stable isotope data (Benson et al., 2002), in ancient woodrat midden records (Tausch et al., 2004; Mehringer and Wigand, 1990), and with ages of submerged rooted stumps along the eastern Sierra Nevada (Stine, 1994; Lindstrom, 1990). These drought episodes appear to correspond with the timing of ice drift minima (solar maxima) identified from North Atlantic marine sediments (Mensing, 2004).

Mensing et al. (2008) review the pollen evidence from four sites revealing evidence of drought within the Great Basin during the last 2000 years. Using pollen ratios of climate indicator species indicating drier versus moister climate, they reveal periods of drought conditions. They then compared those records with the ages of tree stumps submerged beneath lakes in the western Great Basin, tree-ring chronologies, pack rat middens, and $\delta^{18}\text{O}$ data from sediments. Mensing and his

colleague's pollen records provide evidence of the magnitude of some of the droughts of the last two millennia. Their pollen records indicate that the climate between 2000 and 1800 cal BP was dry, with the driest sites being in the western Great Basin. The century ending ~1200 cal BP may have been dry, but the pollen record does not support severe drought. Their high-resolution pollen records from Pyramid Lake and Mission Cross Bog clearly identify a drought ending ~800 cal BP (350 rcyr BP) and 170 cal BP (150 rcyr BP) cal BP, whereas only the Pyramid Lake record indicates a drought ending at 550 cal BP. Evidence for more mesic climate at this time in northeastern Nevada suggests that this drought may have been fairly localized.

At the eastern edge of the central Great Basin, a record from Blue Lake, UT, analyzed by Louderback reveals a latest Pleistocene through Holocene record of vegetation change (Louderback and Rhode, 2009). Data from the upper 4 m of the core reveal that when Lake Bonneville occupied the Blue Lake site, pine and sagebrush communities characterized the nearby shorelines following a decline in lake level after ~12 000 cal BP. Wetlands developed by ~11 900 cal BP. Bulrush-dominated marshes were established no later than 10 800 cal BP. The Blue Lake wetlands desiccated during a warm, dry early middle Holocene ~8300–6500 cal BP. More mesic conditions beginning ~6500 cal BP are documented by the return of marshes at the expense of playa and grass (probably salt grass) meadow communities.

Regionally, sagebrush abundance increased relative to desert shrub species. Single-leaf piñon pine migrated into the

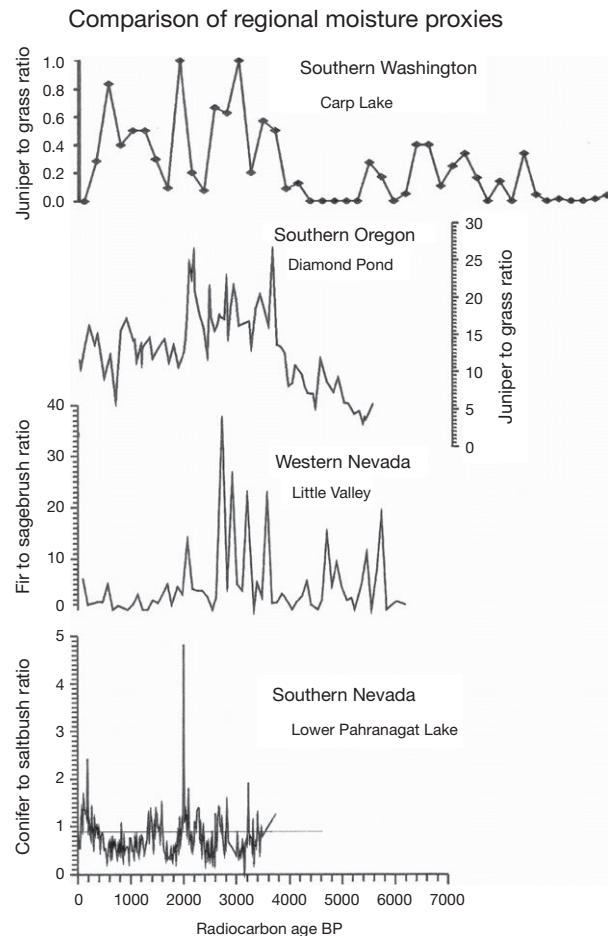


Figure 8 Comparison of drought indices generated from the pollen of climate indicator plant species at four western US pollen localities in the Intermountain West. They range from Carp Lake in the north to Lower Pahranaagat Lake in the south. Allowing for slight differences in the position of the climatic events due to the vagaries of radiocarbon dating, there is significant similarity between the records. Differences usually reflect the origin point of the event. If the source of moisture driving a wet episode or period is the northern Pacific, that event is strongest in the northern portion of the west and at higher elevations. If the event is derived from monsoonal rains, the event is stronger in the southern portion of the west. The plant species used in these ratios vary only in order to select the most representative wet versus dry climate indicator plant species from each site.

region (probably the nearby Goshute Mountains) after ~8000 cal BP. Late Holocene fluctuations include cooler and relatively moister intervals from ~4400 to 3400 and ~2700 to 1500 cal BP (these coincide with those in the Diamond Pond record). Warmer conditions between ~3400 and 2700 cal BP and after 1500 cal BP are also revealed in the pollen record from Blue Lake.

Desert West A high-resolution pollen record from Lower Pahranaagat Lake, about 90 km north-northeast of Las Vegas, Nevada, records about 4090 cal (3700 rcyr) of low-elevation vegetation dynamics in the northern Mojave Desert (Wigand and Rhode, 2002, Figure 17). With a sample about every 14 years for the span of the record, very high-frequency climate cycles could be identified and compared with long tree-ring

records in the region (Wigand and Rhode, 2002, Figure 17). Although the magnitude is not as great as in the northern Great Basin records, the three major wet phases of the Neopluvial between 4000 and 2000 rcyr BP are also evident (Figure 8; Wigand and Rhode, 2002, Figure 18). These were first recorded at Diamond Pond. In addition, the period of late-spring to midsummer-shifted precipitation recorded in northern and western Great Basin pollen records is also prominent in the pollen sequence from Lower Pahranaagat Lake (Wigand and Rhode, 2002, Figure 18).

Radiocarbon dates on peats and spring deposits at Ash Meadows, a huge spring complex with great time depth, have revealed at least three intervals of early to middle Holocene spring resurgence. These events took place about 9200 cal BP, about 8100 cal BP, and finally about 6800 cal BP. The earliest episode appears to be associated with the appearance of springs in areas where they do not currently exist, suggesting a much greater precipitation input than later during the Holocene. The other two episodes appear to be similar to other middle to late Holocene spring events.

Stable isotope data extracted from single-leaf piñon pine and Utah juniper remains from ancient woodrat middens sampled near the Yucca Mountain high-level nuclear waste storage facility study area reveal a pattern of variable physiological stress on these two species (Figure 9). As might be expected, piñon is less resistant to drought and temperature stress than is Utah juniper. When the stable isotope record is compared with the record of aeolian activity produced by Lancaster (1993), a clear relationship is demonstrated between physiological stress on both juniper and piñon resulting from climate variability. This is also reflected in the initiation and termination of periods of aeolian activity in southern Nevada (Figure 10).

Palynological investigations during the middle and late 1950s led Martin (1963) to propose a period of middle Holocene increased monsoonal precipitation for central and southern Arizona. He worked with pollen records obtained from cienega (marsh) deposits, where pollen preservation is often poor, deposition rates are highly variable, and some climatic episodes are missing from the stratigraphic record. Martin proposed that the climate of the middle Holocene of the Desert Southwest was characterized by heavy summer rains resulting in higher values of saltbush pollen. Martin associated increased values of *Aster*-type pollen with higher water tables, floodplain alluviation, fewer and lighter summer storms, and perhaps increased winter precipitation. If this is the case, then the record from Double Adobe IV suggests a Neopluvial moist episode similar to those seen between 4580 cal BP (4000 rcyr BP) and 1980 cal BP (2000 rcyr BP) in the northern Mojave Desert and the Great Basin (Martin, 1963, Figure 19). An extended period of increased *Aster*-type (Compositae) pollen is centered around 4250 cal BP (3800 rcyr BP) at Double Adobe. He also recorded that an early Holocene transition from *Aster*-type pollen dominance to saltbush-dominated pollen spectra occurred about 8860 cal BP (7900 rcyr BP) (Martin, 1963, Figure 20).

The Montezuma Well pollen sequence from central Arizona records over 10 ka of vegetation dynamics (Davis and Shafer, 1992). Declining values of pollen from the cypress family (probably *Juniperus*), ash (*Fraxinus*), and pine occurred between about 11540 cal BP (10 000 rcyr BP) and 8340 cal BP (7500 rcyr BP) (Davis and Shafer, 1992). Middle Holocene

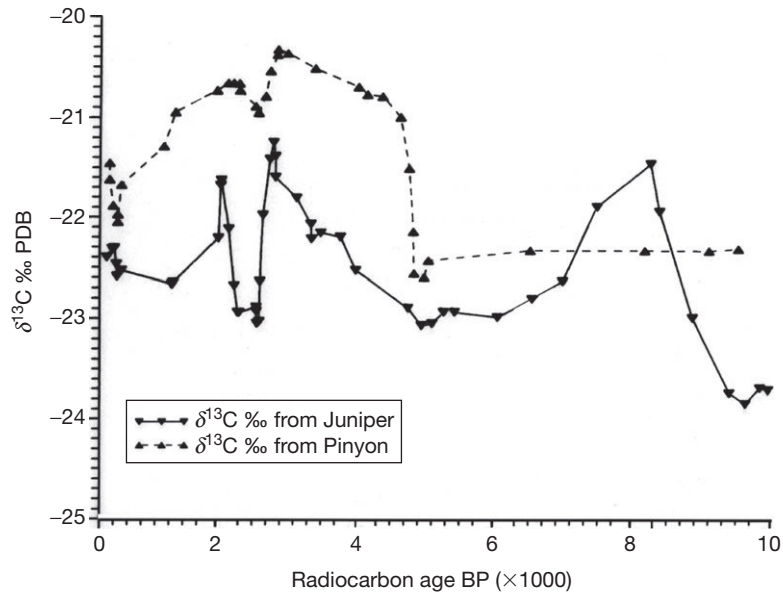


Figure 9 Carbon isotope ($\delta^{13}\text{C}$) measurements were obtained on single-leaf piñon and Utah juniper macrofossils recovered from ancient woodrat middens spanning the Holocene in the northern Mojave Desert. The samples were obtained from middens within a hundred-kilometer radius of the Nevada Test Site. Each midden was represented by a paired (piñon and juniper) sample – in a few cases there is a juniper but no piñon sample. Elevationally, the middens were within a couple of hundred meters of each other. Carbon 13 is a measure of plant stress. As plants experience stress (either as a result of drought or frost during the growing season), the abundance of carbon 13 increases. The data plotted here indicate that (1) both piñon and juniper were more stressed by the middle to late Holocene shift to cooler annual temperatures than they had been by middle Holocene drier climates when summer precipitation characterized the area and (2) the piñon sample of each sample pair indicated greater stress in response to the change in climate than did juniper at the same locality.

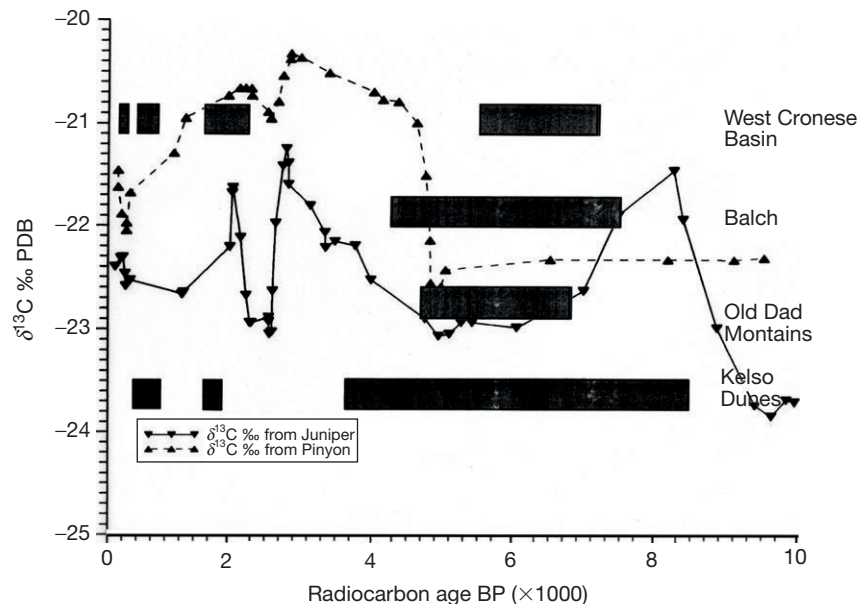


Figure 10 The carbon 13 measurements plotted above are compared with dated aeolian dune activity in the northern Mojave Desert. In general, when carbon 13 values indicate higher stress on piñon and juniper, dune movement declines and ceases. On the other hand, dune movement characterizes intervals when plant stress decreases, but follows an episode of higher precipitation so that there is abundant alluvium available for aeolian transport.

drought, characterized by significantly increased values of saltbush pollen after 7880 cal BP (7000 rcyr BP) occurred at the same time that sedge pollen suffered a dramatic decline (Davis and Shafer, 1992). Saltbush pollen declined after 4570 cal BP (4000 rcyr BP), and cypress family pollen values increased,

suggesting an increase in regional precipitation contemporaneous with the Neopluvial event seen further north and west. Resurgence of sedge pollen values after 4570 cal BP (4000 rcyr BP) also substantiates the regional increase in precipitation (Davis and Shafer, 1992; see also Figure 8).

Two well-dated pollen records from extreme southwestern Colorado, Beef Pasture and Twin Lakes, reveal more detail of Holocene vegetation in the northeastern corner of the Desert Southwest. Beef Pasture lies near the base of the spruce zone in the La Plata Mountains, while Twin Lakes lies in the middle (Petersen, 1988). Spruce (*Picea*) pollen from the Twin Lakes record suggests that prior to about 8600 rcyr BP, spruce was not as abundant around the site. Two episodes between 9660 cal BP (8600 rcyr BP) to 6720 cal BP (5800 rcyr BP) imply that spruce reached its greatest Holocene density in the La Plata Mountains (Petersen, 1988, Figure 26; Petersen and Mehringer, 1976). This may have been driven by warm, moist, perhaps monsoonal climates in the early middle Holocene. Declining spruce pollen values from the Beef Pasture record indicate that climatic conditions became relatively drier after 3030 cal BP (2800 rcyr BP) (Petersen, 1988, Figure 27). This may reflect the same climatic transition that occurred near the end of the Neopluvial at Diamond Pond in the northern Great Basin and in other sites between the two areas.

Petersen (1988, 1994) attempted a reconstruction of the relative positions of the lower and upper spruce tree lines of the La Plata Mountains (Petersen, 1988, Figure 51). He believed that lower spruce treeline reflected the relative proportion of summer to winter precipitation. Upper treeline was influenced by temperature. He suggested that, except for a period between 1300 and 800 calendar years ago, upper treeline retreated to lower elevations for much of the last 2 ka (Petersen, 1988). Several episodes of lower spruce treeline retreat to higher slopes occurred during that same interval. The result was a dramatic shrinking of the spruce zone in the La Plata Mountains. Counts

of piñon-type pine pollen coinciding with the upslope retreat of spruce indicate a relatively increased summer to winter precipitation (Petersen, 1988, Figure 51). Petersen used this information to explain the success of Anasazi farmers in the Four Corners area during that interval.

In the desert regions of northern Mexico, the potential for reconstructing ancient vegetation and climate histories has been explored using the combination of plant macrofossils and pollen from ancient woodrat middens (Anderson and Van Devender, 1991). This technique fills a crucial proxy data void in a region that lacks pollen records from aquatic, dry cave, or alluvial contexts. Currently, only cursory vegetation histories and their possible paleoclimates have been reconstructed for the coastal lowlands of Sonora, Mexico (Anderson and Van Devender, 1995). In particular, the spatial movement of indicator plant species (e.g., piñon and creosote bush) has been reconstructed for very localized areas. However, as the spatial and temporal distribution of woodrat midden strata increases, the possibility of reconstructing more detailed vegetation and climate histories for the region will increase.

A final example of the value of utilizing ancient woodrat midden records to enhance the pollen records of the same region is the dated expansion of single-leaf piñon in the western Great Basin from the northern Mojave Desert. Directly dated piñon needles reveal the expansions and retreats of piñon during the last 22 000 years along the eastern front of the Sierra Nevada Mountains (Figure 11). This movement was controlled by changes in moisture and temperature during the late Quaternary.

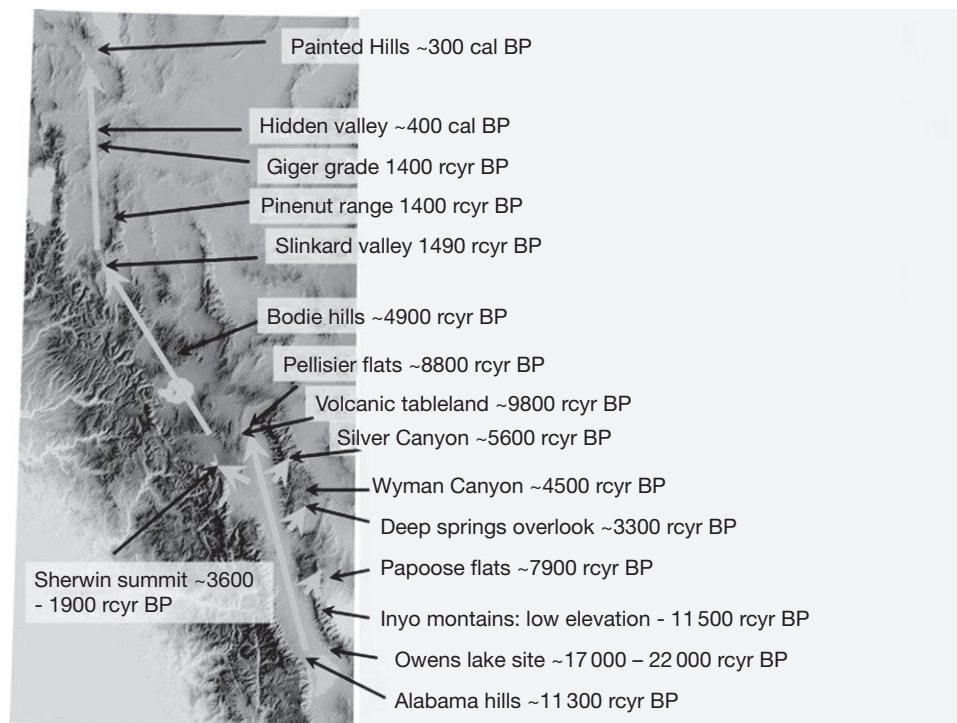


Figure 11 The radiocarbon-dated movement of piñon pine along the eastern front of the Sierra Nevada Mountains. It seems to have begun its expansion from the extreme northwestern Mojave Desert and then used the Owens River and the foothills lying on either side as a corridor. When conditions improved (warmer and moister), piñon would expand. As conditions became cooler and/or drier, piñon would retreat.

See also: **Pollen Methods and Studies:** Databases and Their Application. **Pollen Records, Postglacial:** Northeastern North America; Northwestern North America; Southeastern North America.

References

- Adam DP (1985) Quaternary pollen records from California. In: Bryan VM Jr. and Holloway RG (eds.) *Pollen Records of Late-Quaternary North American Sediments*, pp. 125–140. Dallas, TX: American Association of Stratigraphic Palynologists.
- Adam DP (1988) *Professional Paper 1363: Palynology of Two Upper Quaternary Cores from Clear Lake*, Lake County, California, pp. 86. US Geological Survey, Washington, DC.
- Anderson RS and Van Devender TR (1991) Comparison of pollen and macrofossils in packrat (*Neotoma*) middens: A chronological sequence from the Waterman Mountains, southern Arizona, USA. *Review of Paleobotany and Palynology* 68: 1–28.
- Anderson RS and Van Devender TR (1995) Vegetation history and paleoclimates of the coastal lowlands of Sonora, Mexico – Pollen records from packrat middens. *Journal of Arid Environments* 30: 295–306.
- Benson L, Kashgarian M, Rye R, et al. (2002) Holocene multidecadal and multicentennial droughts affecting Northern California and Nevada. *Quaternary Science Reviews* 21: 659–682.
- Betancourt JL, Van Devender TR, and Martin PS (eds.) (1990) *Packrat Middens: The Last 40,000 years of Biotic Change*. Tucson, AZ: University of Arizona Press.
- Billings WG (1951) Vegetational zonation in the Great Basin of western North America. In: International Union Societe Biologique Series 8, 9: Les Bases Ecologiques de la Regeneration de la Vegetation des Zones Arides, 101–122. Paris: UISB.
- Davis OK (1992) Rapid climatic change in coastal southern California inferred from pollen analysis of San Joaquin Marsh. *Quaternary Research* 37: 89–100.
- Davis OK (1999) Pollen analysis of Tulare Lake, California: Great Basin-like vegetation in Central California during the full glacial and early Holocene. *Review of Palaeobotany and Palynology* 107: 249–257.
- Davis OK, Anderson RS, Fall PL, et al. (1985) Palynological evidence for early Holocene aridity in the southern Sierra Nevada, California. *Quaternary Research* 24: 322–332.
- Davis OK and Moratto MJ (1988) Evidence for a warm dry early Holocene in the western Sierra Nevada of California: Pollen and plant macrofossil analysis of Dinkey and Exchequer Meadows. *Madrono* 35: 132–149.
- Davis OK and Shafer DS (1992) A Holocene climatic record for the Sonoran Desert from pollen analysis of Montezuma Well, Arizona, USA. *Palaeogeography, Palaeoclimatology, Palaeoecology* 92: 107–119.
- Hall SA (1985) Quaternary pollen analysis and vegetational history of the Southwest. In: Bryant VM Jr. and Holloway RG (eds.) *Pollen Records of Late-Quaternary North American Sediments*, pp. 95–123. Dallas, TX: American Association of Stratigraphic Palynologists.
- Haynes CV Jr. (1991) Geoarchaeological and paleohydrological evidence for a Clovis-age drought in North America and its bearing on extinction. *Quaternary Research* 35(3): 438–450.
- Houghton JG, Sakamoto CM, and Gifford RO (1975) Nevada's Weather and Climate. In: *Special Publication 2. Nevada Bureau of Mines and Geology, Mackay School of Mines*, p. 78. Reno: University of Nevada.
- Hunt CB (1967) *Natural Regions of the United States and Canada*. San Francisco, CA: W.H. Freeman p. 725.
- Johnson CG Jr., Clausnitzer RR, Mehringer PJ Jr., and Oliver CD (1994) Biotic and abiotic processes of eastside ecosystems: The effects of management on plant and community ecology, and on stand and landscape vegetation dynamics. *General Technical Report PNW-GTR-322*, p. 66. US Department of Agriculture, US Forest Service, Pacific Northwest Research Station, Portland.
- Lancaster N (1993) Development of Kelso Dunes, Mojave Desert, California. *National Geographic Research and Exploration* 9: 444–459.
- Lindstrom S (1990) Submerged tree stumps as indicators of middle Holocene aridity in the Lake Tahoe Basin. *Journal of California and Great Basin Anthropology* 12: 146–157.
- Louderback LA and Rhode DE (2009) 15,000 Years of vegetation change in the Bonneville basin: The Blue Lake pollen record. *Quaternary Science Reviews* 28(2009): 308–326.
- Lowe CH (1977) *Arizona's Natural Environment: Landscapes and Habitats*. Tucson, AZ: University of Arizona Press p. 136.
- Martin PS (1963) *The Last 10 000 Years: A fossil Pollen Record of the American Southwest*. Tucson, AZ: University of Arizona Press p. 87.
- Martin PS and Mehringer PJ Jr (1965) Pleistocene Pollen Analysis and Biogeography of the Southwest. In: Wright HE Jr. and Frey DG (eds.) *The Quaternary of the United States*, pp. 433–451. Princeton, NJ: Princeton University Press.
- Mehringer PJ Jr. (1985) Late-Quaternary pollen records from the interior Pacific Northwest and northern Great Basin of the United States. In: Bryant VM Jr. and Holloway RG (eds.) *Pollen Records of Late-Quaternary North American Sediments*, pp. 167–189. Dallas, TX: American Association of Stratigraphic Palynologists.
- Mehringer PJ Jr. (1986) Prehistoric: Environments. In: D'Azevedo WL (ed.) *Handbook of North American Indians. Great Basin*, vol. II, pp. 31–50. Washington, DC: Smithsonian Institution.
- Mehringer PJ Jr. and Wigand PE (1990) Comparison of late Holocene environments from woodrat middens and pollen. In: Betancourt JL, Van Devender TR, and Martin PS (eds.) *Packrat Middens: The Last 40 000 years of Biotic Change*, pp. 294–325. Tucson, AZ: University of Arizona Press.
- Mensing SA, Benson LV, Kashgarian M, and Lund S (2004) A Holocene pollen record of persistent droughts from Pyramid Lake, Nevada, USA. *Quaternary Research* 62: 29–38.
- Mensing S, Smith J, Burkle Norman K, and Allan M (2008) Extended drought in the Great Basin of western North America in the last two millennia reconstructed from pollen records. *Quaternary International* 188(2008): 79–89.
- Petersen KL (1988) Climate and the Dolores River Anasazi: A paleoenvironmental reconstruction from a 10,000-year pollen record, La Plata Mountains, Southwestern Colorado. University of Utah, Anthropological Papers Number 113, p. 66. University of Utah Press, Salt Lake City.
- Petersen KL (1994) A warm and wet Little Climatic Optimum and a cold and dry Little Ice Age in the southern Rocky Mountains, USA. *Climatic Change* 26: 243–269.
- Petersen KL and Mehringer PJ Jr. (1976) Postglacial timber-line fluctuations, La Plata Mountains, southwestern Colorado. *Arctic and Alpine Research* 8: 275–288.
- Quade J, Forester RM, Pratt WL, and Caner C (1998) Black mats, spring-fed streams, and late-glacial-age recharge in the southern Great Basin. *Quaternary Research* 49(2): 129–148.
- Stine S (1994) Extreme and persistent drought in California and Patagonia during mediaeval time. *Nature* 369: 546–549.
- Tausch R, Nowak C, and Mensing S (2004) Climate change and associated vegetation dynamics during the Holocene: The paleoecological record. In: Chambers UC and Miller JR (eds.) *Great Basin Riparian Ecosystems: Ecology, Management and Restoration*, pp. 24–48. Covelo, CA: Island Press.
- Thompson RS (1992) Late Quaternary environments in Ruby Valley, Nevada. *Quaternary Research* 37: 1–15.
- Wigand PE and Mehringer PJ Jr (1985) Pollen and seed analyses. In: Thomas DH (ed.) pp. 108–124. Nevada: The Archaeology of Hidden Cave. New York: American Museum of Natural History.
- Wigand PE (1987) Diamond pond, Harney County, Oregon: Vegetation history and water table in the eastern Oregon desert. *Great Basin Naturalist* 47: 427–458.
- Wigand PE (1997) Native American diet and environmental contexts of the Holocene revealed in the pollen of human fecal material. *Nevada Historical Society Quarterly* 40(1): 105–116.
- Wigand PE (2003) Southeastern Idaho: Sensitive Climatic Transition Region in the Northern Rockies. Report on Grays Lake Pollen Analysis submitted to Richard Holmer, Department of Anthropology, Idaho State University. 81 pp, 68 Figures, 10 Tables.
- Wigand PE (2004) Local Marsh and Regional Vegetation and Fire History and Its Environmental Context as Reconstructed from Sediment Cores at Summit Lake, Nevada. Report submitted to the Bureau of Land Management, Winnemucca Office. 42 pp, 16 Figures, 7 Tables.
- Wigand PE (2004) Modern Ecology and Paleoecology of the Sierra Valley of Northeastern California. Final report submitted to Far Western Anthropological Group, Inc. 63 pp, 32 Figures, 5 Tables.
- Wigand PE and Palacios-Fest M (2008) The SR133 Realignment Project (Laguna Canyon, Southern California): Pollen and Other Macrobotanical Analyses. Caltrans Laguna Canyon Project. Statistical Research, Inc.
- Wigand PE and Rhode D (2002) Great Basin vegetation history and aquatic systems: The last 150,000 years. In: Hershler R, Madsen DB, and Currey DR (eds.) *Great Basin Aquatic Systems History. Smithsonian Contributions to Earth Sciences*, vol. 33, pp. 309–367. Washington, DC: Smithsonian Institution Press.

South America

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Introduction

The South American continent is mostly covered by different tropical, subtropical, and temperate vegetation types forming a high number of very diverse ecosystems (Seibert, 1996). Important tropical rainforest ecosystems include the Amazon rainforests, especially in northern Brazil, the Atlantic rainforest, forming a small strip along the eastern Brazilian coast, and the Chocó rainforest, mostly distributed along the western Colombian Pacific coastal lowland.

The major tropical savanna ecosystems north of the equator are the Llanos Orientales in Colombia and the Orinoco Llanos in Venezuela. Other large savanna areas include the Gran Sabana in southeastern Venezuela, the Rupununi–Rio Branco savanna in Suriname, and the Roraima–Rupununi savanna in Brazil. South of the equator is the Cerrado, the major savanna region occurring mostly in central Brazil.

In northeastern Brazil, the semiarid Caatinga is the dominant vegetation type. In southeastern Brazil, there is a gradient from the coast to the inland, going from the Atlantic rainforest to the semideciduous forest, deciduous forest, and Cerrado. *Araucaria* forests and patches of grassland (Campos) cover the southern Brazilian highlands. Further south, in the southeastern South American lowlands Pampa, grasslands (Uruguay, Argentina), Chaco dry forests (Argentina), and the Patagonian steppe (Argentina, Chile) are the most important ecosystems.

The Andes are the major mountain system in South America, stretching along the western side of the continent from 5° N to 55° S. The variety of ecosystems in the Andes is very high, including altitudinal gradients from mountain rainforest, cloud forests, dry forests, deserts, high altitude Paramos, and Puna tundra grasslands, as well as latitudinal gradients from tropical, subtropical, and temperate vegetation types from rainforest to deserts. The Atacama desert is found on the coastal side of the central Andes and the Valdivian rainforest, and the *Araucaria* forests occur to the west of the southern Andes.

All those different vegetation types are related to different climatic conditions. In tropical South America, the annual precipitation and its temporal distribution are controlled by the seasonal migration of the intertropical convergence zone (ITCZ), which shifts from 7–9° N in January to 10–20° S in July, and by the topographic features of the continent. As a result, two different main climate types are represented in tropical South America: a climate without a dry season in the equatorial latitudes and climates with marked dry seasons north and south of the equatorial latitudes. North of the equator, the dry season lasts from October to March, and south of the equator, it occurs from April to September. For most of the savanna regions with seasonal climates, the annual precipitation ranges from 1200 to 2000 mm, and the average annual temperature is 20–26 °C. In northeastern Brazil, the ITCZ remains north of the equator and causes semiarid conditions with a long dry season of 6–11 months. The mean annual

rainfall is less than 250–750 mm. South America north of the equator is influenced by the northeasterly trade winds blowing from the Caribbean, whereas the region south of the equator is influenced by the Atlantic Southeasterlies, which originate from the subtropical high-pressure cell over the South Atlantic Ocean. The climate of the northern Andes is also influenced by the easterly trade winds. Advections of Antarctic cold fronts produce heavy rainfall when they meet tropical air masses. These climate patterns occur mainly in the regions of southern and southeastern Brazil. One consequence of these climate patterns is that southern Brazil has a humid climate with little or no marked dry season. The average annual precipitation is between 1400 and 2200 mm. The climate in southern South America is under influence of the Westerlies (Nimer, 1989; Weischet, 1996).

The main questions for this article are: What can postglacial pollen records tell us about the environmental history of South America? What has happened to the different South American ecosystems during postglacial times? How stable was vegetation and climate during the Holocene? What was the human impact on vegetation and environment?

Postglacial pollen records provide insights on past vegetation and climate changes from different regions of South America. Compared with Europe or North America, the paleovegetation and climate of South America is poorly studied. So far, probably not more than 200 postglacial pollen records are available for sites in South America (see the Global Pollen Database website). Most of them have been published during the last 20 years. This article provides an introduction to a limited number of pollen records. Several major ecosystems or regions from South America with detailed and well-dated representative pollen records have been selected for this overview. The radiocarbon ages cited here are in uncalibrated ¹⁴C ages (years BP).

Amazon Rainforest and Coastal Mangroves

The Amazon rainforest is the largest rainforest ecosystem in South America and covers an area of about 5 600 000 km². The Amazon Basin is low lying, with more than 1 000 000 km² below 100 m elevation. For example, the water level at Manaus, which lies 1200 km inland, is only 14 m a.s.l. As a consequence, the central and eastern part of the Amazon Basin and its vegetation as well as the coastal region with its mangroves and salt marshes must have been strongly influenced by the postglacial sea-level rise and dynamics.

The influence of Holocene sea-level changes on the Amazon Basin is documented by the Rio Curuá pollen record from the Caxiuanã forest reserve in eastern Amazonia (Figure 1; Behling and Costa, 2000). Four different Holocene paleoenvironmental intervals have been identified. First, there was a transition from an active to a passive fluvial system and a well-drained and diverse *terra firme* (unflooded upland) Amazon rainforest with very limited development of inundated

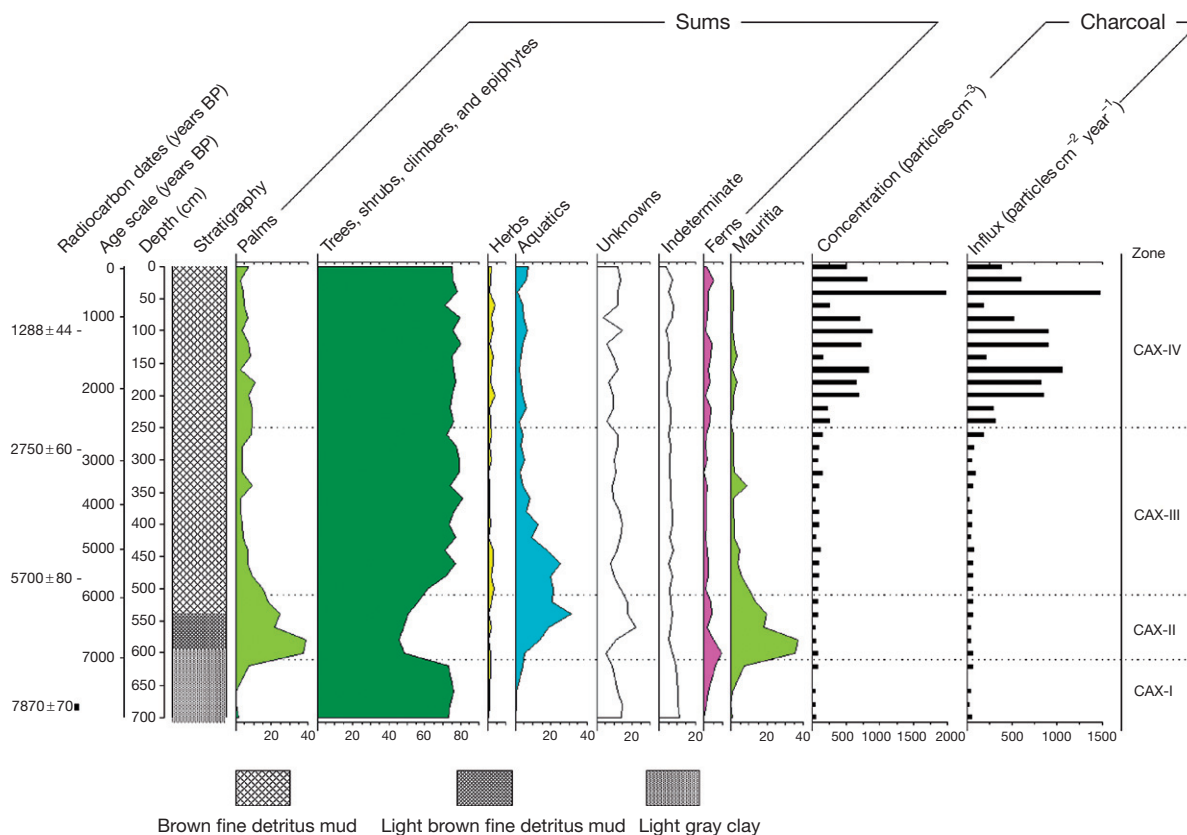


Figure 1 Summary pollen diagram from Rio Curuá. Modified from Behling H and Costa ML (2000) Holocene environmental changes from the Rio Curuá record in the Caxiuanã region, eastern Amazon Basin. *Quaternary Research* 53: 369–377.

forests (Várzea and Igapó forests) existed between >7990 and 7030 years BP. Figure 2 shows an inundated Várzea forest in Amazonia. Second, a sluggish river with a local *Mauritia* palm swamp and similar regional vegetation as before occurred between 7030 and 5970 years BP. Third, a passive river, forming shallow-lake conditions and with still-abundant *terra firme* forest, developed in the study region between 5970 and 2470 years BP, and finally, a blocked or impeded river with high water levels and marked increase of inundated forests occurred during the last 2470 years BP. Increased charcoal during this last interval suggests the first strong presence of humans in this region. Similar environmental changes, including the *Mauritia* palm swamp phase, have also been found in the Lago Calado record in the central Amazonia, suggesting that this environment was a regional phenomenon for low-lying areas in the Amazon Basin (Behling et al., 2001).

Based on the Lago do Crispim pollen record from the Amazon coastal area, mangroves first developed along the river close to the core site between 7550 and 6620 years BP (Figure 3; Behling and Costa, 2001). There is evidence that areas, originally covered by dense, tall coastal Amazon rainforest, were partly replaced by mangrove and some Restinga (coastal shrub and herbaceous vegetation) during the early Holocene. Decreasing *Rhizophora* (mangrove) pollen abundances document a retreat of mangroves, reflecting sea-level regression starting at around 7000 years BP. The marked reduction of mangroves near the lake indicates a lower relative sea level between around 6620 and 3630 years BP. During this

interval, a local *Mauritia/Mauritiella* palm swamp formed. Marked coastal environmental changes occurred at around 3630 years BP, driven by sea-level transgression. Mangroves expanded again and were close to the site. The local palm swamp was replaced by a Cyperaceae (sedge) swamp. Rainforest and coastal Restinga vegetation adjacent to the swamp were replaced by salt marshes as sea level rose. The Atlantic Ocean was close to the core site, but the site, which is only 1–2 m above modern sea level, was apparently never affected by marine incursions during the Holocene. Reduced mangrove vegetation, since ca. 1840 years BP, may be due to a slightly lower relative sea level or to human impact.

In western Amazonia, the Atlantic sea-level rise played a minor role. The pollen record from the lower terrace of Rio Caquetá in Colombia indicates that the terrace was poorly drained after 4000 years BP, as a result of higher river levels, probably due to higher precipitation rates in western Amazonia (Behling et al., 1999).

Amazon Rainforest and the Neighboring Savannas

Pollen records from the transition zone between Amazon rainforests and neighboring savanna ecosystems reflect the dynamic of the Amazon forest area (Behling et al., 2010). The pollen record from Laguna Loma Linda from the Llanos Orientales in Colombia is found in such a transition zone,



Figure 2 Seasonally inundated Várzea forest of the Amazon lowland in Brazil (Photo H. Behling).

which today is covered with Amazon rainforest (Figure 4; Behling and Hooghiemstra, 2000). Figure 5 shows a typical appearance of the Llanos Orientales in Colombia with grass savanna and gallery forest in the background. During the interval from 8700 to 6000 years BP, the vegetation of the region was dominated by grass savanna with only a few woody savanna taxa, such as *Curatella* (sandpaper tree) and *Byrsonima* (nance). Gallery forest along the drainage system was poorly developed. This is also evident from other pollen records from the central part of the Llanos Orientales. Compared to today, precipitation must have been significantly lower and seasonality stronger. During the interval from 6000 to 3600 years BP, rainforest taxa increased markedly, reflecting an increase in precipitation. Amazonian rainforest and gallery forest taxa were abundant, such as Moraceae/Urticaceae, Melastomataceae, *Alchornea*, *Cecropia* (trumpet tree), and *Acalypha* (copperleaf), while Poaceae (grasses) were reduced in frequency. From 3600 to 2300 years BP, rainforest taxa continued to increase; Moraceae/Urticaceae became very frequent, and Myrtaceae and *Myrsine* became common. Savanna vegetation decreased continuously throughout this interval. Pollen data from the central region in the Llanos Orientales indicate a marked increase of *Mauritia* palms since about that time. This indicates that precipitation was still increasing and that the length of the annual dry season was possibly shortened. From 2300 years BP onwards, grass savanna (mainly represented by Poaceae) expanded, and *Mauritia* palms became frequent. This reflects an increased human impact on the vegetation.

There is also evidence of late Holocene Amazon rainforest expansion in other regions, both north and south of the equator, reflecting a change to wetter climatic conditions with higher precipitation rates and longer wet periods since the mid-Holocene and especially during the late Holocene (Behling and Hooghiemstra, 2001). In southwestern Amazonia, there is evidence of Amazon rainforest expansion

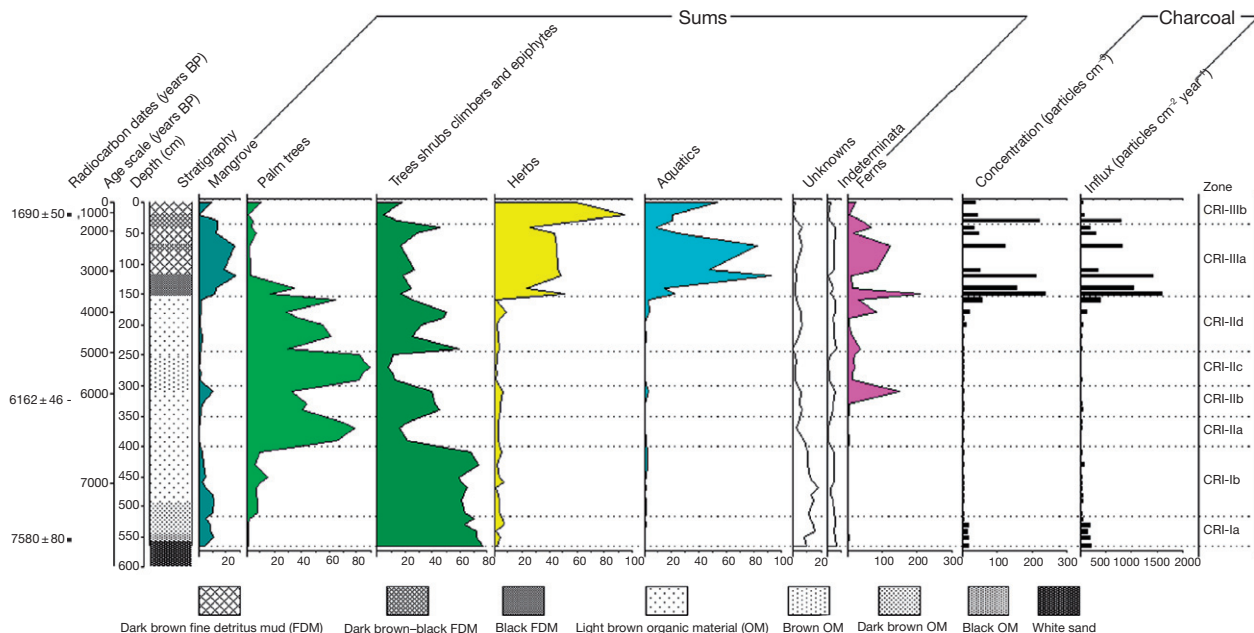


Figure 3 Summary pollen diagram from Lago Crispim. Modified from Behling H and Costa ML (2001) Holocene vegetational and coastal environmental changes from the Lago Crispim record in northeastern Pará State, eastern Amazonia. *Review of Palaeobotany and Palynology* 114: 145–155.

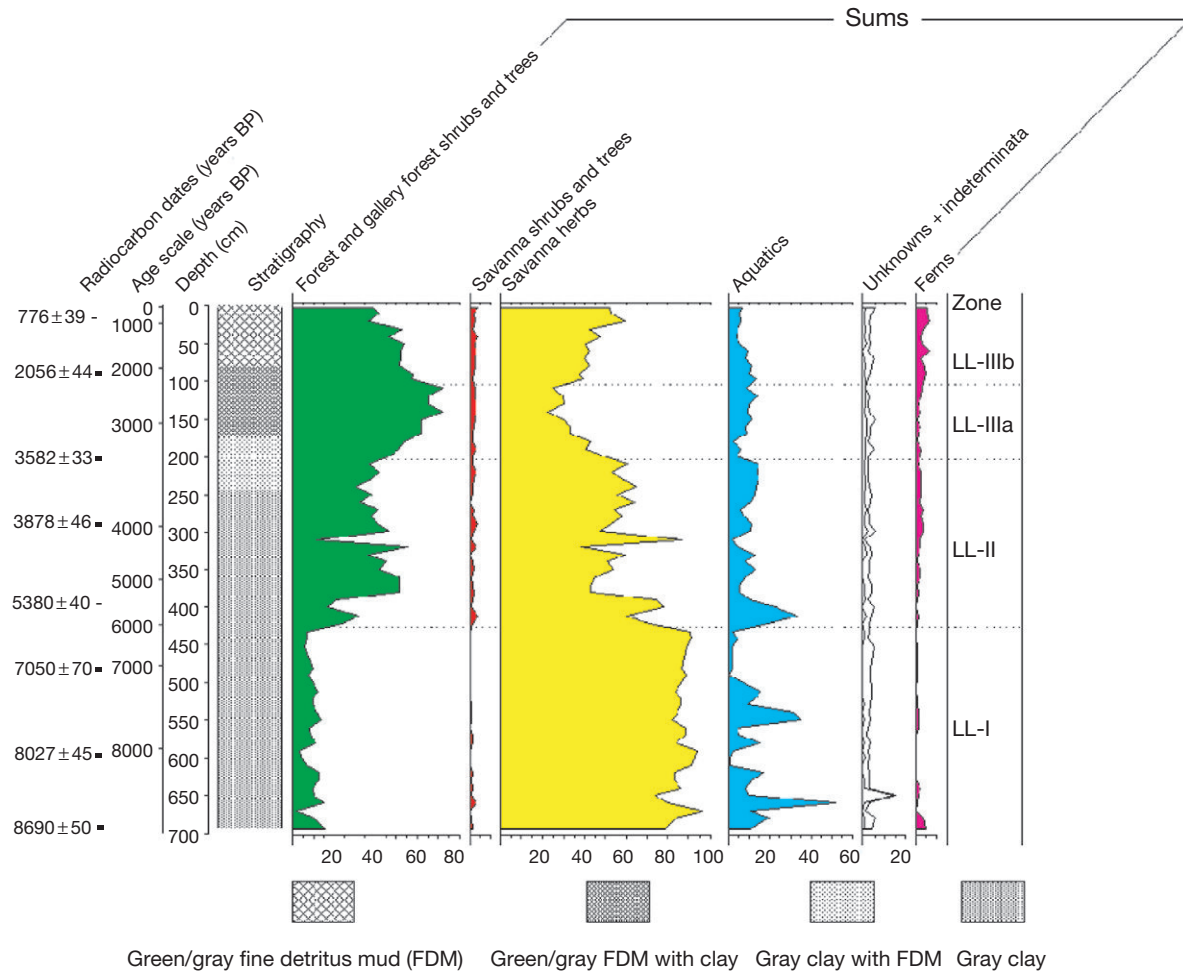


Figure 4 Summary pollen diagram from Laguna Loma Linda. Modified from Behling H and Hooghiemstra H (2000) Holocene Amazon rain forest – Savanna dynamics and climatic implications: High-resolution pollen record Laguna Loma Linda in eastern Colombia. *Journal of Quaternary Sciences* 15: 687–695.



Figure 5 Savanna and gallery forest of the Llanos Orientales in Colombia (Photo H. Behling).

(Laguna Bella Vista and Chaplin, Bolivia) during the late Holocene, at least since 3000 years BP (Mayle et al., 2000). The expansion of the Amazon rainforest after 5460 years BP is also reported for the southeastern Amazon region (Behling, 2002a).

Caatinga in Northeastern Brazil

The modern vegetation in northeastern Brazil is primary Caatinga vegetation, which is subdivided into several physiognomic vegetation types, including grassland, xerophytic thorn shrub savanna, and shrub woodland. Pollen analytical studies have been carried out on a peat bog core located in the Icatu River Valley, within a system of stabilized sand dunes at the middle São Francisco River (10°24'S, 43°13'W) in northeastern Brazil (De Oliveira et al., 1999). The study region was dominated by the *Mauritia* palm during the early Holocene until about 4240 years BP, suggesting wetter climatic conditions than during the late Holocene. Between 8910 and 6790 years BP, there was a decrease in forest taxa and a gradual increase of Caatinga and Cerrado taxa. Later, the absence of pollen and spores in the core indicate semiarid conditions. Between about 6230 and 4540 years BP, there was a return of mosaic vegetation including gallery forest, Cerrado, and Caatinga, indicating the return of moister climatic conditions. After 4240 years BP, the increase in Caatinga and Cerrado taxa and the reduction of gallery forest taxa in the landscape reflect the establishment of the present vegetation and climatic conditions.

Cerrado, Semideciduous Forest, and Atlantic Rainforest of Southeastern Brazil

A pollen record has been studied from a 16 m long core from Lago do Pires, which today is located in semideciduous forest area, in a transition zone between the Atlantic rainforest to the east and the Cerrado to the west (Figure 6; Behling, 1995a). During the early Holocene interval from 9720 to 5530 years BP, the presence of Cerrado vegetation with some gallery forest indicates low precipitation and a dry season of about 6 months. Within this period, an interval from 8810 to 7500 years BP characterized by expanding gallery forest, suggests greater precipitation and a shorter dry season (6–5 months). From 5530 to 2780 years BP, the vegetation was characterized by extensive forest in the valleys, and Cerrado on hills, indicating higher precipitation and a dry season of about 5 months. From 2780 to 970 years BP, denser Cerrado vegetation on the hills and slopes suggests a dry season of 4–5 months. During the latest Holocene, the presence of closed semideciduous forest is indicative of modern climatic conditions. Markedly extended high-elevation grassland in the Serra do Mar coastal mountain range north of Rio de Janeiro indicates dry climatic conditions during the early Holocene until about 4910 years BP (Behling and Safford, 2010). Later, wetter conditions are indicated by reduced high-elevation grassland and the extension of Atlantic rainforest into higher elevation areas.

Araucaria Forest and Campos of Southern Brazil

Subtropical *Araucaria* forest is found mostly on the southern Brazilian highlands between latitudes 24° and 30° S at 1000–1400 m elevation. Several paleoecological studies from the *Araucaria* forest and Campos regions of the southern Brazilian highlands have been carried out in the last decade (Behling, 2002b). Figure 7 shows the *Araucaria* forest and Campos on the southern Brazilian highland. Data from the states of Paraná (Serra Campos Gerais: Behling, 1997), Santa Catarina (Serra do Rio Rastro, Morro da Igreja, Serra da Boa Vista: Behling, 1995b), and Rio Grande do Sul (Aparados da Serra: Roth and Lorscheitter, 1993; Cambará do Sul: Behling et al., 2004) have proved that extensive areas of Campos vegetation existed on the highlands through glacial, early, and mid-Holocene times. The dominance of Campos vegetation was attributed to cold and dry glacial conditions followed by warm and dry early Holocene climates. A dry season lasting probably about 3 months each year was characteristic for the early and mid-Holocene interval (Behling, 1997, 2002b). Initial expansion of *Araucaria* forests started by migration from the gallery forests along the rivers about 3000 years BP, which indicates a return to somewhat wetter climates. A marked expansion of *Araucaria* forests started on the highlands, replacing Campos vegetation in Santa Catarina State about 1000 years BP and in Paraná State (Serra Campos Gerais) about 1500 years BP, reflecting a very

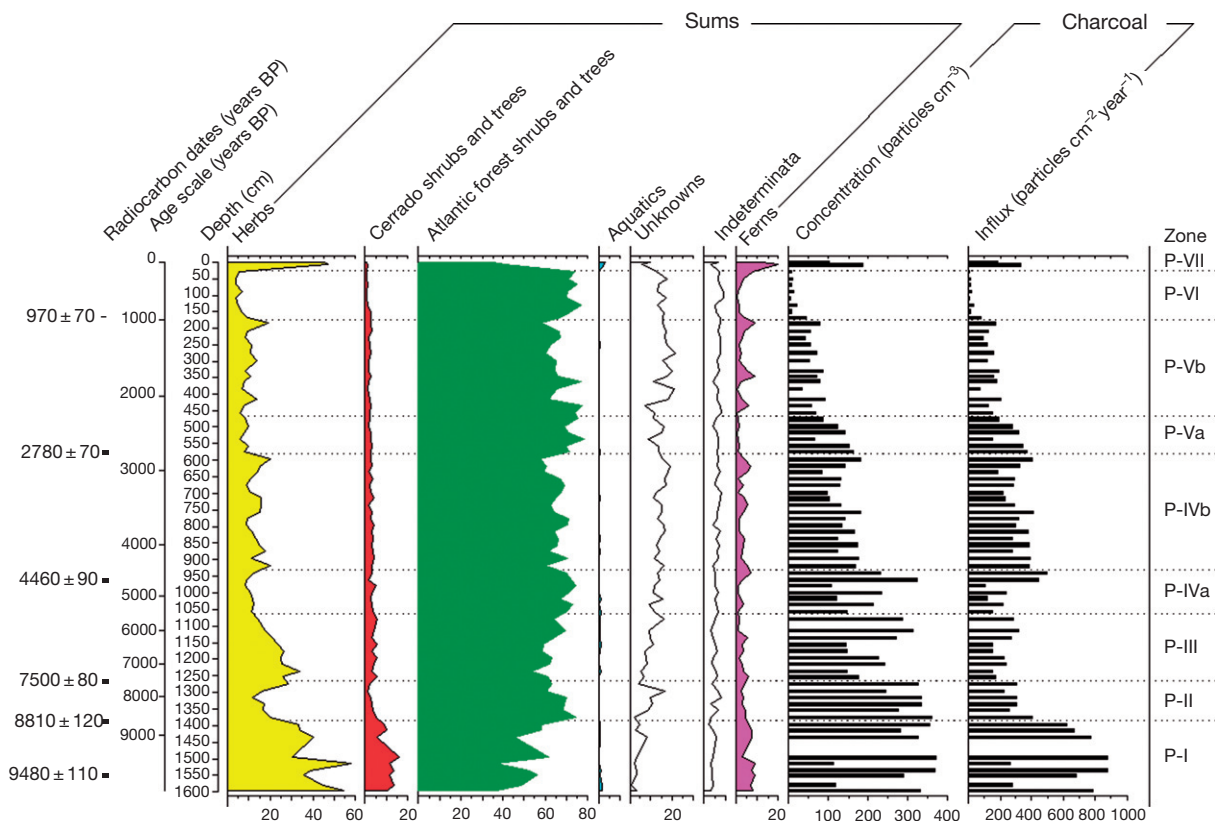


Figure 6 Summary pollen diagram from Lago do Pires. Modified from Behling H (1995) A high-resolution Holocene pollen record from Lago do Pires, SE Brazil: Vegetation, climate and fire history. *Journal of Paleolimnology* 14: 253–268.

humid climate without a marked seasonal dry season. That burning is a strong factor controlling the dynamic between *Araucaria* forest and grassland during the late Holocene is shown by the record from the São José dos Ausentes region in Rio Grande do Sul State (Jeske-Pieruschka et al., 2010).

New data from the high-resolution Cambará do Sul record, which extends to 42 000 years BP, indicate that Campos vegetation prevailed on the southernmost highland during early and mid-Holocene times, but first small *Araucaria* forest populations migrated along river valleys to the study region (Figure 8; Behling et al., 2004). At the beginning of the Holocene, the species composition of the grassland changed, probably related to the Holocene warming. During the late Holocene interval at 3950 years BP, *Araucaria* forests expanded

and created a net of gallery forests along the streams. At that time, Campos was still the predominant vegetation, until 1140 years BP. Then, *Araucaria* forest trees, especially *Araucaria angustifolia* (Paraná pine) itself, replaced the Campos vegetation of the study region. For a certain interval between AD 1520 and 1780, *Weinmannia* trees became frequent in the *Araucaria* forest. Some years later, Poaceae and then Cyperaceae pollen increased, indicating that cattle had invaded the forest, an indication of increasing human impact. The introduction of exotic pines into southern Brazil about 180 years ago, the collection of tree fern trunks, the intensive logging of the *Araucaria* trees during the last 50–60 years, and the transformation of the original forest to a secondary forest are also registered in the pollen record.



Figure 7 *Araucaria* forest with Campos (grassland) on the southern Brazilian highland (Photo H. Behling).

Pampa of Argentina

The Holocene history of vegetation and climate of five Pampa sites are summarized by Prieto (1996). During the late early Holocene, grassland communities existed until 7000 years BP in the central part of the Pampa and until about 5000 years BP in the southwestern part of the Pampa. During the late Holocene, pollen data indicate vegetation that is similar to modern psammophytic (sand-adapted) and halophytic (salt-tolerant) communities of the southern Pampa, associated with communities characteristic of moister edaphic conditions. The drier climatic conditions changed to a subhumid-dry climate after 5000 years BP (Arroyo Sauce Chico, Cerro La China), about 4000 years BP (Fortín Necochea) and before 3000 years BP (Empalme Querandíes), with a trend to toward more humid conditions.

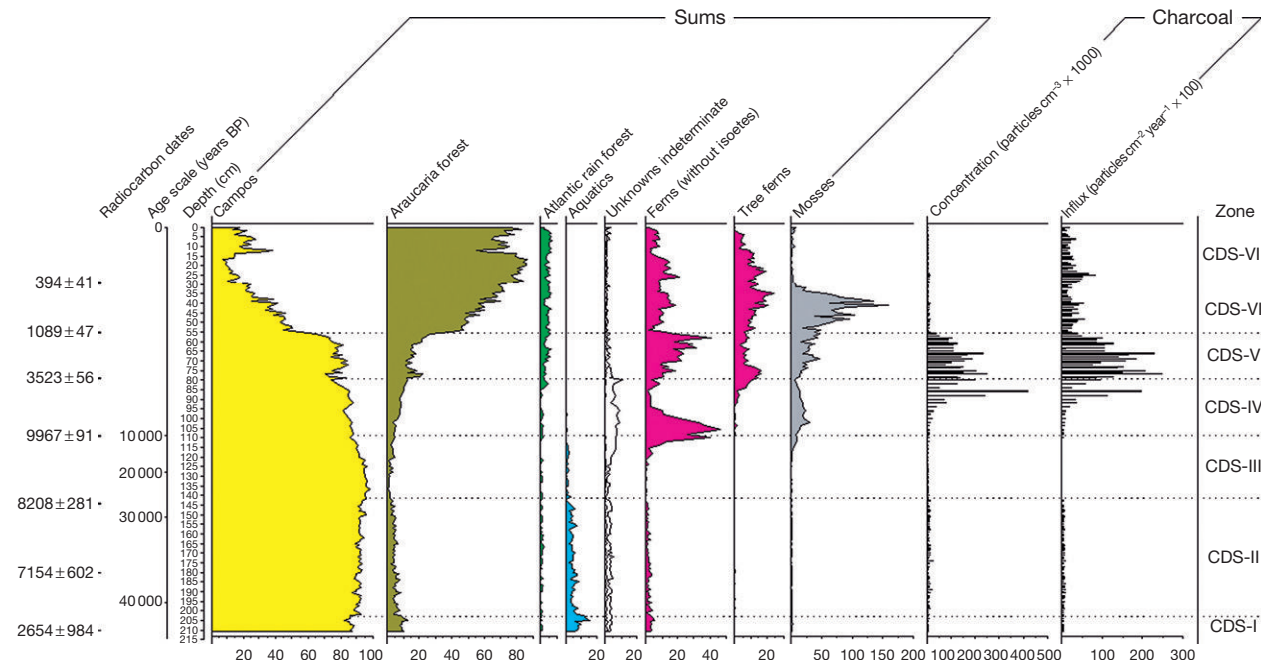


Figure 8 Summary pollen diagram from Cambará do Sul. Modified from Behling H, Pillar V, Orlóci L, and Bauermann SG (2004) Late Quaternary *Araucaria* forest, grassland (Campos), fire and climate dynamics, studied by high-resolution pollen, charcoal, and multivariate analysis of the Cambará do Sul core in southern Brazil. *Palaeogeography, Palaeoclimatology, Palaeoecology* 203: 277–297.

Andes in South America

The number of published pollen records from the Andes, especially from Colombia and Chile, is much greater than from the South American lowlands. Due to the great latitudinal extent of the Andes from 5° N to 55° S and the range's varied topographic features and altitudinal gradients, crossing different climatic zones, the recorded postglacial vegetation history is rather complex and cannot be generalized. Vegetation and climate patterns vary depending on location. Great vegetation changes occur within very short distances, for instance, from cloud forest with perhumid climatic conditions to a dry forest with long annual dry seasons and low rainfall rates.

The pollen records from the Colombian Andes, such as from Lake Fúquene (2580 m a.s.l.), indicate that glacial paramo (low vegetation growing above altitudinal treeline) and subparamo vegetation, indicative of cold and dry conditions, was replaced by Andean forest reflecting warm and humid conditions during the early Holocene. The onset of the drier climate prevailing today began in the middle Holocene (Vélez et al., 2003). Investigations on lake and peat sediments in the Cajas National Park (2°45'S, 79°10'W) of the western Andes in Ecuador indicate the occurrence of 11 drier and wetter phases during the Holocene (Hansen et al., 2003). Studies in the central Peruvian Andes from Laguna Baja (7°42'S, 77°32'W, 3575 m a.s.l.) document a decrease of pollen from Podocarpaceae (podocarp) and Urticales, reflecting an increase of warmer and wetter climatic conditions during the early Holocene. The first evidence for anthropogenic activities can be seen after 4000 years BP (Hansen and Rodbell, 1995).

Studies in the temperate forest region of the southern Andes at 40° S indicate that west of the Andes, a more extensive southern beech (*Nothofagus*) parkland existed during the early Holocene, and east of the Andes, there were a high number of steppe taxa, together with *Nothofagus dombeyi* type. On both sides of the Andes, a warmer and seasonally drier climate existed than is found there today (summarized by Markgraf in Grimm et al., 2001). Paleoecological studies on vegetations dynamics of the northern Patagonian rainforest in the Chonos Archipelago in southern Chile (44–46° S) document *Tepualia* and *Weinmannia* as important forest trees, reflecting a warmer climate between 10 600 and 6000 years BP than present. Later *Pilgerodendron* recovered and *Nothofagus–Pilgerodendron–Tepualia* forests occurred in that area until the present (Haberle and Bennett, 2004).

For further reading on post glacial pollen records in South American, see Behling (1998, 2002c), Behling and Hooghiemstra (2001), Grimm et al. (2001), and Markgraf (1993).

See also: **Pollen Records, Late Pleistocene: South America.**

References

- Behling H (1995a) A high resolution Holocene pollen record from Lago do Pires, SE Brazil: Vegetation, climate and fire history. *Journal of Paleolimnology* 14: 253–268.
- Behling H (1995b) Investigations into the Late Pleistocene and Holocene history of vegetation and climate in Santa Catarina (S Brazil). *Vegetation History and Archaeobotany* 4: 127–152.
- Behling H (1997) Late Quaternary vegetation, climate and fire history in the *Araucaria* forest and campos region from Serra Campos Gerais (Paraná), S Brazil. *Review of Palaeobotany and Palynology* 97: 109–121.
- Behling H (1998) Late Quaternary vegetational and climatic changes in Brazil. *Review of Palaeobotany and Palynology* 99: 143–156.
- Behling H (2002a) Late Quaternary vegetation and climate dynamics in southeastern Amazonia interfered from Lagoa da Confusão in Tocantins State, northern Brazil. *Amazoniana* 17: 27–39.
- Behling H (2002b) South and Southeast Brazilian grasslands during Late Quaternary times: A synthesis. *Palaeogeography, Palaeoclimatology, Palaeoecology* 177: 19–27.
- Behling H (2002c) Carbon storage increases by major forest ecosystems in tropical South America since the Last Glacial Maximum and the early Holocene. *Global and Planetary Change* 33: 107–116.
- Behling H, Berrio JC, and Hooghiemstra H (1999) Late Quaternary pollen records from the middle Caquetá river basin in central Colombian Amazon. *Palaeogeography, Palaeoclimatology, Palaeoecology* 145: 193–213.
- Behling H, Bush MB, and Hooghiemstra H (2010) Biotic development of Quaternary Amazonia: A palynological perspective. In: Hoorn C and Wesselingh FP (eds.) *Amazonia, Landscape and Species Evolution*, 1st edn., pp. 335–345. Oxford: Blackwell.
- Behling H and Costa ML (2000) Holocene environmental changes from the Rio Curuá record in the Caxiuanã region, eastern Amazon Basin. *Quaternary Research* 53: 369–377.
- Behling H and Costa ML (2001) Holocene vegetational and coastal environmental changes from the Lago Crispim record in northeastern Pará State, eastern Amazonia. *Review of Palaeobotany and Palynology* 114: 145–155.
- Behling H and Hooghiemstra H (2000) Holocene Amazon rain forest – Savanna dynamics and climatic implications: High resolution pollen record Laguna Loma Linda in eastern Colombia. *Journal of Quaternary Sciences* 15: 687–695.
- Behling H and Hooghiemstra H (2001) Neotropical savanna environments in space and time: Late Quaternary interhemispheric comparisons. In: Markgraf V (ed.) *Interhemispheric Climate Linkages*, pp. 307–323. New York: Academic Press.
- Behling H, Keim G, Irion G, Junk W, and Nunes de Mello J (2001) Holocene environmental changes inferred from Lago Calado in the Central Amazon Basin (Brazil). *Palaeogeography, Palaeoclimatology, Palaeoecology* 173: 87–101.
- Behling H, Pillar V, Orlóci L, and Bauermann SG (2004) Late Quaternary *Araucaria* forest, grassland (Campos), fire and climate dynamics, studied by high resolution pollen, charcoal and multivariate analysis of the Cambará do Sul core in southern Brazil. *Palaeogeography, Palaeoclimatology, Palaeoecology* 203: 277–297.
- Behling H and Safford HD (2010) Late-glacial and Holocene vegetation, climate and fire dynamics in the Serra dos Órgãos Mountains of Rio de Janeiro State, southeastern Brazil. *Global Change Biology* 16: 1661–1671.
- Brunschön C and Behling H (2010) Reconstruction and visualization of upper forest line and vegetation changes in the Andean depression region of southeastern Ecuador since the last glacial maximum – A multi-site synthesis. *Review of Palaeobotany and Palynology* 163: 139–152.
- De Oliveira PE, Barreto AMF, and Suguio K (1999) Late Pleistocene/Holocene climatic and vegetational history of the Brazilian caatinga: The fossil dunes of the middle São Francisco River. *Palaeogeography, Palaeoclimatology, Palaeoecology* 152: 319–337.
- Grimm EC, Lozano-García S, Behling H, and Markgraf V (2001) Holocene vegetation and climate variability in the Americas. In: Markgraf V (ed.) *Interhemispheric Climate Linkages*, pp. 325–370. San Diego: Academic Press.
- Haberle SG and Bennett KD (2004) Postglacial formation and dynamics of northern Patagonian rainforest in the Chonos Archipelago, southern Chile. *Quaternary Science Reviews* 23: 2433–2452.
- Hansen BCS and Rodbell DT (1995) A late-glacial/Holocene pollen record from the eastern Andes of Northern Peru. *Quaternary Research* 44: 216–227.
- Hansen BCS, Rodbell DT, Seltzer GO, Leon B, Young KR, and Abbott M (2003) Late-glacial and Holocene vegetational history from two sides in the western Cordillera of southwestern Ecuador. *Palaeogeography, Palaeoclimatology, Palaeoecology* 194: 79–108.
- Jeske-Pieruschka V, Fidelis A, Bergamin RS, Vélez E, and Behling H (2010) *Araucaria* forest dynamics in relation to fire frequency in southern Brazil based on fossil and modern pollen data. *Review of Palaeobotany and Palynology* 160: 53–60.
- Markgraf V (1993) Climatic history of Central and South America since 18,000 yr BP: Comparison of pollen records and model simulations. In: Wright HE Jr., Kutzbach JE, Webb T III, Ruddiman WF, Street-Perrott FA, and Bartlein PJ (eds.) *Global Climates since the Last Glacial Maximum*, pp. 357–385. Minneapolis and London: University of Minnesota Press.

- Mayle F, Burbridge R, and Killeen TJ (2000) Millennial-scale dynamics of southern Amazonian rain forests. *Science* 290: 2291–2294.
- Nimer E (1989) *Climatologia do Brasil*. Rio de Janeiro: Brazilian Geography and Statistics Institute 421 pp.
- Prieto AR (1996) Late Quaternary vegetational and climatic changes in the Pampa grassland of Argentina. *Quaternary Research* 45: 73–88.
- Roth L and Lorscheitter ML (1993) Palynology of a bog in Parque Nacional de Aparados da Serra, East Plateau of Rio Grande do Sul, Brazil. *Quaternary of South America and Antarctic Peninsula* 8: 39–69.
- Seibert P (1996) *Farbatlas Südamerika: Landschaft und Vegetation*. Stuttgart: Ulmer 288 pp.
- Vélez MI, Hooghiemstra H, Metcalfe S, Martinez I, and Mommersteeg H (2003) Pollen- and diatom based environmental history since the Last Glacial Maximum from the Andean core Fúquene-7, Colombia. *Journal of Quaternary Science* 18: 17–30.
- Weischet W (1996) *Regionale Klimatologie, 1 Die Neue Welt*. Stuttgart: Teubner 468 pp.

Relevant Website

<http://www.ncdc.noaa.gov/paleo/gpd.html> – Global Pollen Database (GPD).

Northern Asia

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Introduction

The territory of northern Asia represents approximately 15% of the Earth's land surface and includes a wide variety of climatic and vegetation zones. It is situated between the Ural Mountains (ca. 60° E) in the west and the Pacific Ocean in the east (ca. 180° E), and between the Arctic Ocean in the north and ca. 47° N latitude in the south (**Figure 1**). The latter boundary corresponds to the southern limit of boreal coniferous forests in Kazakhstan and Mongolia. West of 90° E, the topography is relatively flat. Elevations of the West Siberian Lowland do not exceed 200 m, and hilly plains and low-elevation (up to 1500 m) mountains occur in northern Kazakhstan. East of 90° E, the topography is more complex. Lowlands occupy a 50–600-km-wide band along the Arctic coast, while central and southern regions consist of high plateaus and mountains above 1000 m. The highest elevations occur in the Altai (4506 m) and Hangai (3905 m) mountains and on the Kamchatka Peninsula (4750 m).

The climate of northern Asia is cold and continental. A strong Siberian anticyclone controls winter weather. The mean January temperatures vary from -16°C in the south to -48°C in the interiors of east Siberia. The absolute temperature minimum registered in the Verkhoyansk Mountains is -68°C . July temperatures (T_{VII}) decrease northward from ca. 20°C to less than 4°C . More than 65% of the annual precipitation (P) falls during the warm season, when the weather is controlled by the westerlies and the Pacific monsoon. P reaches 600 mm in the Ural Mountains and 1000 mm in the Far East. In Kazakhstan, central Yakutia, and Mongolia, situated far from oceanic influence, P is less than 300 mm.

The spatial distribution of vegetation shows the clear influence of climate. In the mountains, it is complicated by altitude and slope orientation. Various moss, grass, dwarf shrub, and shrub tundra types (tundra zone) occupy arctic lowlands and the upper belt of the mountains and high plateaus. Southward, this is gradually replaced with boreal cold deciduous and evergreen conifer forests (taiga) dominated by larch (*Larix*), Siberian spruce (*Picea obovata*), Scots pine (*Pinus sylvestris*), Siberian pine (*P. sibirica*), and fir (*Abies sibirica*). Broad-leaved trees are represented mostly by cold- and drought-resistant birch (*Betula pubescens* and *B. pendula*) and aspen (*Populus tremula*). Temperate deciduous taxa, such as elm (*Ulmus*) and lime (*Tilia*), play minor roles in the modern vegetation. Wych elm (*U. glabra*) and Siberian small-leaved lime (*T. sibirica*) trees occur sporadically in the West Siberian Lowland and Altai Mountains, and Siberian elm (*U. pumila*) may grow in the floodplain forests south of Lake Baikal.

Since **Dokturovskii and Kudryashov (1923)** provided pollen identification keys for the main arboreal taxa in Russia, scientists from the former Soviet Union published numerous articles and monographs reconstructing postglacial changes in northern Eurasian vegetation by means of pollen analysis; for a comprehensive synthesis, see **Neishtadt (1957)**, **Khotinskii**

(1977), **Grichuk (1984)**, **Peterson (1993)**, and **Velichko et al. (1997)**. The collapse of the Soviet Union intensified the level of international cooperation, resulting in the appearance of extensive compilations of existing northern Eurasian pollen data (e.g., **Edwards et al., 2000**; **Gunin et al., 1999**; **Tarasov et al., 1998**; **Williams et al., 2011**; and references therein) and in the publication of new high-resolution and better dated pollen records. Since the late 1970s, there has been a rapid increase in the number of published pollen diagrams from northern Asia. **Khotinskii (1977)** had used 26 pollen sequences to reconstruct the Holocene vegetation history of northern Asia, but now, there are ca. 200 records archived in the global and European pollen databases.

In this article, the authors review the postglacial vegetational and climate changes in northern Asia based on pollen records with reliable chronologies, rather than attempting to synthesize a complete northern Asia-wide picture of postglacial environmental changes, which requires further refinement of site spatial coverage and chronologies. Unless otherwise indicated, dates in this discussion are in calendrical-equivalent (calibrated) years before present.

The Ural Mountains

The pollen record from Lake Lyadhej-To ($68^{\circ}15'\text{N}$, $65^{\circ}45'\text{E}$, 150 m a.s.l., site 1 in **Figure 1**), situated in the shrub–herb tundra zone at the NW rim of the polar Ural Mountains, provides information on the Holocene environmental history of this area (**Andreev et al., 2005**). Pollen of sedges (Cyperaceae), grasses (Poaceae), and dwarf birches (*Betula* sect. *Nanae*) (**Figure 2**) from the lowermost sediments suggests sparse tundra-like vegetation around the lake ca. 10.7–10.55 cal ka BP. Approximately 10.55–8.8 cal ka BP birch forest with some shrub alder grew around the lake, reflecting the warmest climate of the Holocene with T_{VII} 11–13 °C. Pollen records from the adjacent areas also reflect a major spread of birch at that time in what is now treeless tundra (**Andreev et al., 2005** and references therein). Higher values of dwarf birch and sedge pollen and significantly lower pollen concentrations point to climatic deterioration during 8.8–5.5 cal ka BP. Birch forests completely disappeared from the lake's vicinity at about 6 cal ka BP, and shrub and herb tundra communities became dominant in the regional vegetation after about 5.5 cal ka BP. Pollen-based reconstruction suggests significant cooling between 5.5 and 3.5 cal ka BP with T_{VII} up to 8 °C, leading to establishment of a shrub–herb tundra similar to the modern vegetation.

West Siberia

Pollen diagrams from the West Siberian Lowland are quite numerous (e.g., **Volkova and Mikhailova, 2002** and references

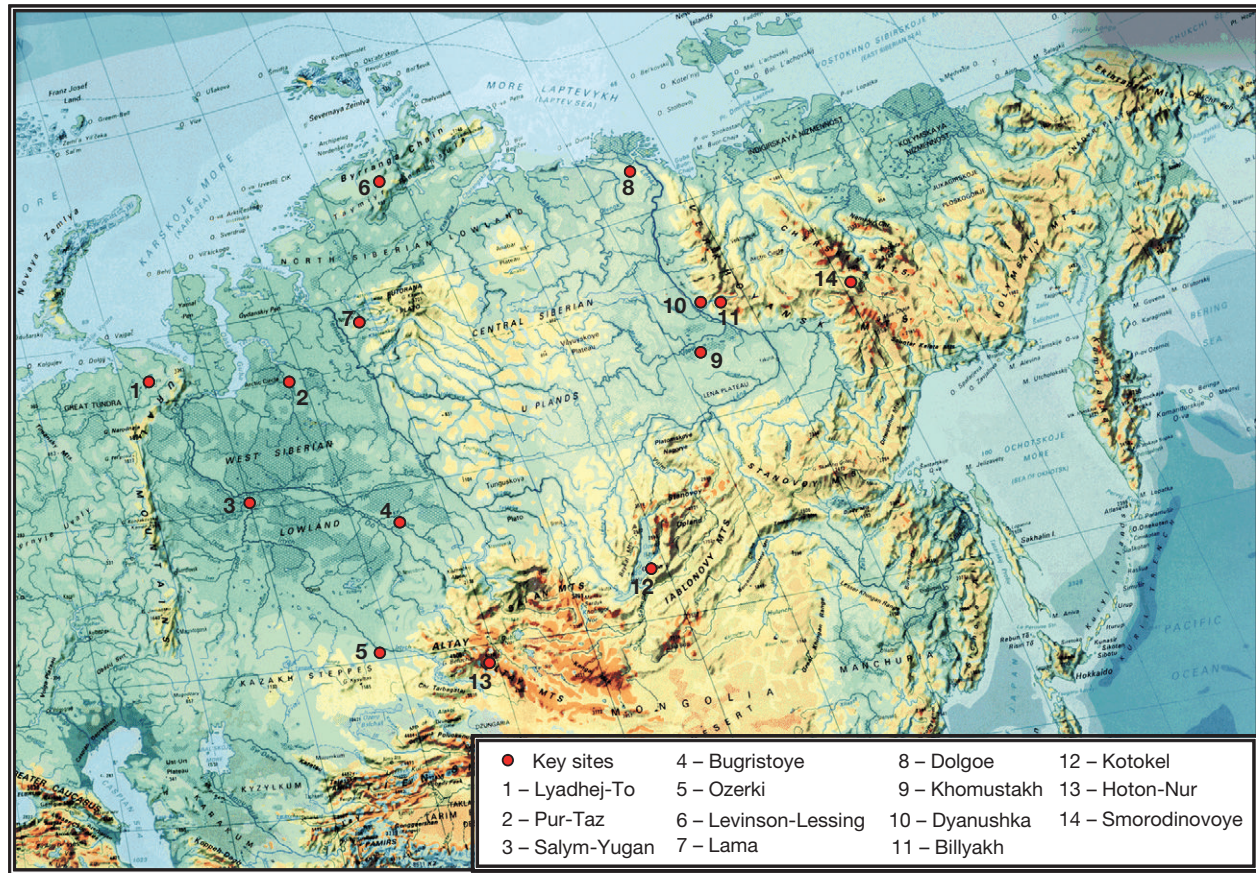


Figure 1 Map of northern Asia showing the location of cited sites.

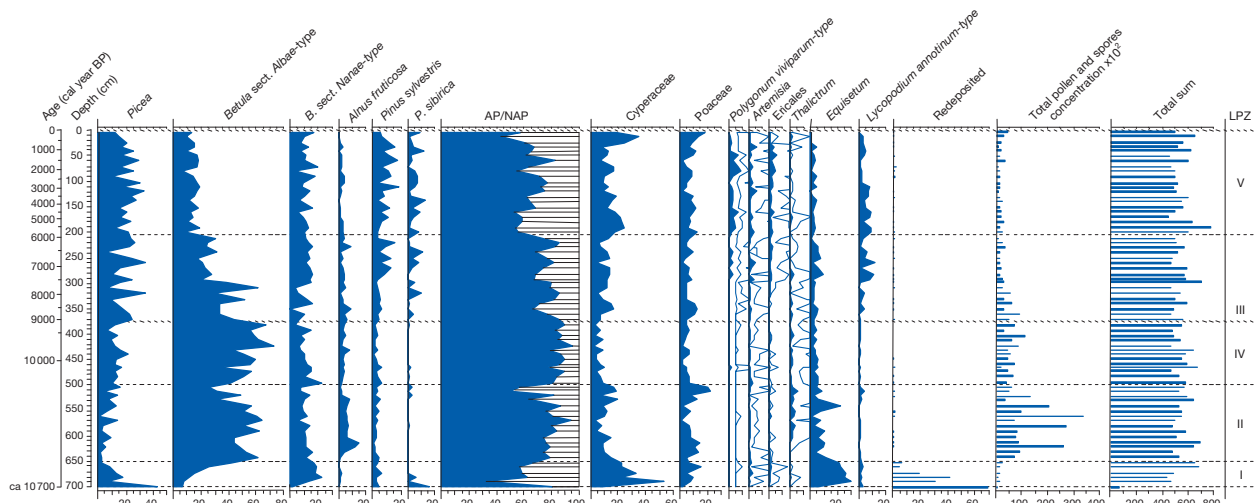


Figure 2 Percentage palynological diagram of selected taxa from Lake Lyadhej-To record. Pollen zonation is according to Andreev et al. (2005).

therein). The Pur-Taz peat section from the flat watershed between the Pur and Taz Rivers ($66^{\circ}42'N$, $79^{\circ}44'E$, site 2 on Figure 1) is one of the better-studied sequences from the forest-tundra transitional zone with Siberian larch (*L. sibirica*) as a dominant tree. The section yielded pollen and macrofossil records of vegetation and climate history at the Arctic treeline ca. 10.4–5.1 cal ka BP (Peteet et al., 1998). The presence of

larch and tree birch pollen in the Late Glacial/early Holocene indicates that regional summer temperatures were warm enough to support the growth of trees. An increase in spruce pollen between 10 and 5.1 cal ka BP suggests a movement of spruce into the north as the result of warmer-than-present early Holocene climate. Quantitative pollen-based climate reconstruction (Andreev and Klimanov, 2000) shows that the

warmest interval (T_{VII} 2.5 °C higher than today) occurred in the area ca. 6.8–5.6 cal ka BP. This is approximately the same time as the dominance of spruce (*Picea*) macrofossils at the Pur-Taz site. The decrease in evergreen conifer tree pollen and macrofossils registered in the uppermost peat layer suggests a late Holocene shift toward colder environments similar to today (Peteet et al., 1998).

Pollen data from the Salym–Yugan area (60°10'N, 72°50'E, site 3 in Figure 1) help to trace the Holocene vegetation history in the southern part of West Siberia (Pitkänen et al., 2002). The modern vegetation of the area consists of forests (mainly Scots pine) and open mire communities. Pollen data suggest that open sedge associations dominated the vegetation between ca. 10.9 and 9.9 cal ka BP and trees, such as larch, spruce, and birch, were poorly represented in the regional vegetation. On the other hand, the pollen sequence from Bugristoye (58°15'N, 85°20'E, site 4 in Figure 1) demonstrates that larch (up to 40%), spruce (up to 25%), and birch (up to 60%) absolutely dominated the pollen spectra as early as 11.4–10.8 cal ka BP, suggesting significant afforestation of the Ket'–Chulym watershed at that time (Blyakharchuk and Sulerzhitsky, 1999). This conclusion agrees with other West Siberian pollen data (Volkova and Mikhailova, 2002). In the Salym–Yugan area, birch and pine forests became established shortly after 9.9 cal ka BP, but spruce was rather abundant until about 4.3 cal ka BP, while the Bugristoye record demonstrates a decrease in spruce abundance shortly after 6.3 cal ka BP, reflecting climate deterioration. Peat accumulation at the Bugristoye site stopped shortly after ca. 4 cal ka BP. After 4 cal ka BP, modern birch–pine forest dominated the vegetation in southern West Siberia, although large regions were covered with open mire vegetation.

Kazakhstan

Reliable records of the Holocene vegetation and climate dynamics in the vast area of northern and central Kazakhstan

were not available until the 1990s (Kremenetski et al., 1997 and references therein). Ozerki swamp (50°25'N, 80°28'E, 210 m; site 5 in Figure 1) located in the Irtysh River valley provides the most complete postglacial pollen record from the forest–steppe transition zone in northern Kazakhstan. In this area, modern steppe vegetation is dominated by grasses and *Artemisia* (sage) species and Chenopodiaceae (goosefoot family) and *Ephedra* (Mormon tea) grow on compacted soils (Tarasov et al., 1997). High summer temperatures and low precipitation limit the spread of patchy birch and pine forests to the river valleys and low mountains. The Ozerki record suggests that an oxbow depression currently occupied by the swamp was filled with water at about 15 cal ka BP, suggesting amelioration of the very dry and cold 'glacial' climate. The lowermost pollen zone (Figure 3) reveals the highest percentages of nonarboreal pollen (NAP) taxa (up to 85%), suggesting rather dry environments and steppe vegetation. However, the occurrence of *Hippophae* (sea-buckthorn), spruce, and larch pollen in this pollen zone can be interpreted as the spread of sea-buckthorn shrubs and trees along the Irtysh River between 15 and 12 cal ka BP. These taxa can be found today in the Altai Mountains, southeast of the study site. Since 12 cal ka BP, the vegetation composition has been similar to that found today in the Kazakhstan steppe. Around 8.5 cal ka BP, birch pollen exceeded 70%, suggesting expansion of birch forest in the area and probably wetter-than-present climate conditions. After 6.5 cal ka BP, the pollen spectra reflect the spread of pine from the south Ural and West Siberia to northern and central Kazakhstan. Regional pollen data (Kremenetski et al., 1997; Tarasov et al., 1997) suggest that Scots pine was already established in northern Kazakhstan by 7 cal ka BP, but did not reach its modern limit in central Kazakhstan until after 2 cal ka BP. Thus, the afforestation of the steppe zone reached a maximum only during the last millennium (Tarasov et al., 2005). Because none of these changes in vegetation can be attributed to human activities, a climatic explanation should be invoked.

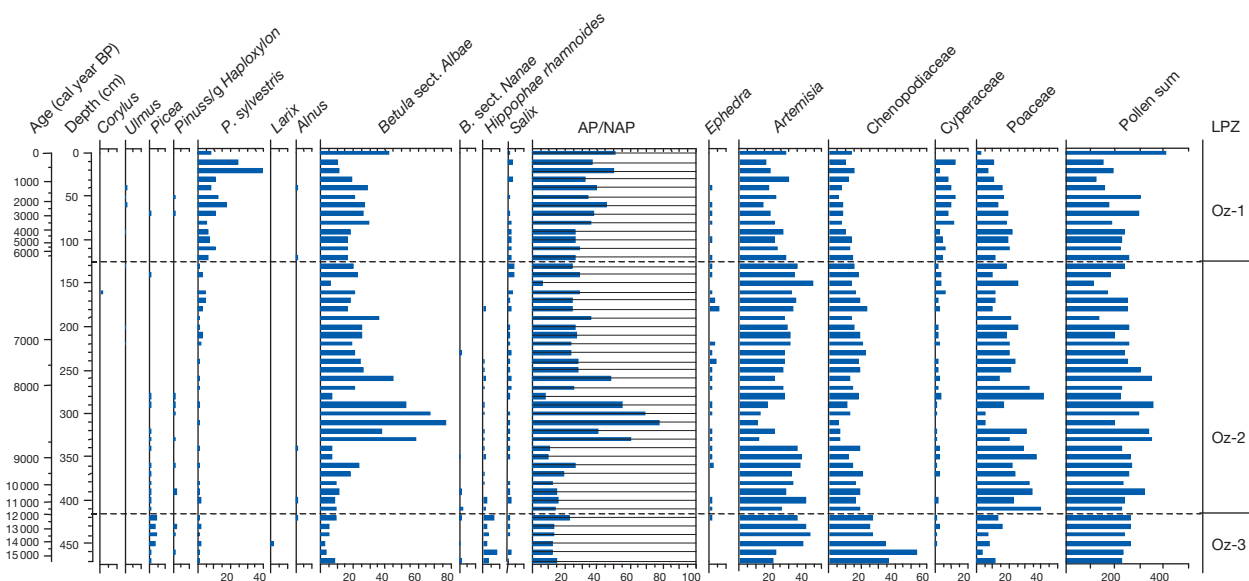


Figure 3 Percentage palynological diagram of selected taxa from Ozerki Swamp record. Pollen zonation is according to Tarasov et al. (1997).

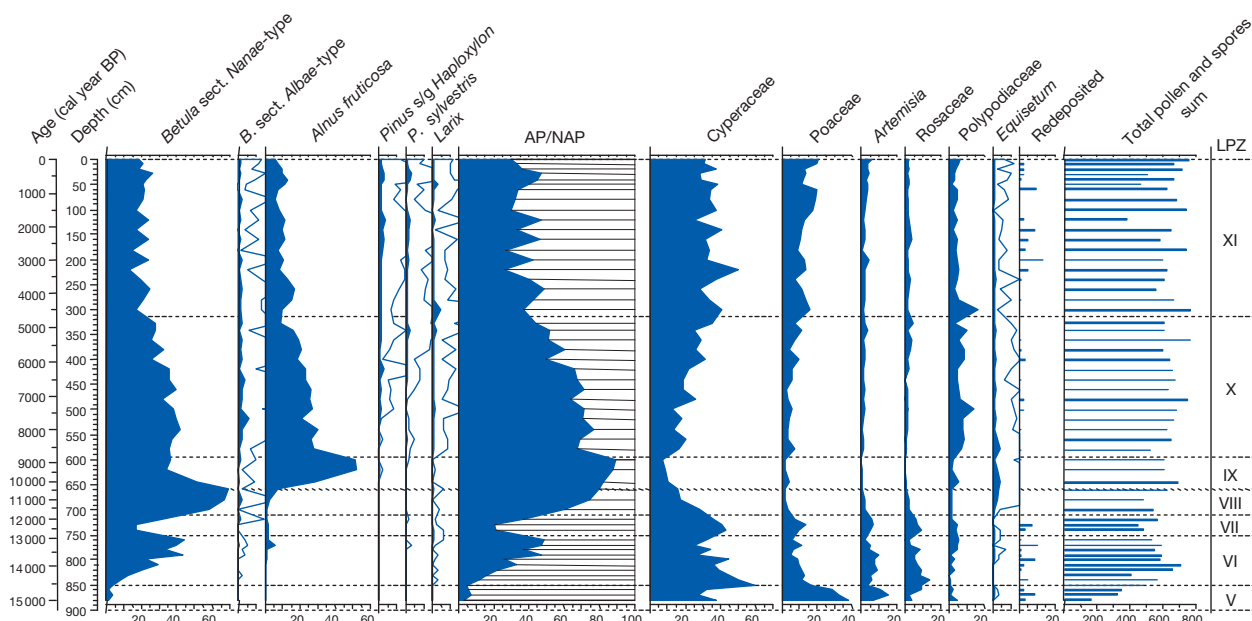


Figure 4 Percentage palynological diagram of selected taxa from Lake Levinson-Lessing record. Pollen zonation is according to Andreev et al. (2003).

Central Siberia

Environmental changes since about 32 ^{14}C ka BP from northern central Siberia are recorded in the sediments of Levinson-Lessing Lake (74°28'N, 98°38'E, site 6 in Figure 1) in the northeastern Taymyr Peninsula (Andreev et al., 2003). The lake is situated in the tundra zone. Pollen spectra (Figure 4) contain large amounts of herbaceous taxa in the lower part of the record, including grasses and sage, reflecting severe continental climate and scarce, steppe-like vegetation until ca. 13.7 cal ka BP. Dwarf shrub and sedge tundra vegetation had limited coverage and survived only in wet habitats. Dramatic increases in shrub birch and willow pollen and in total pollen concentration, associated with a significant decrease in sage, grasses, and other herb taxa pollen, reflect a climatic amelioration after 13.7 cal ka BP corresponding with the Allerød in Europe. This was followed by an increase in herbaceous taxa pollen, likely indicating a cooling associated with the Younger Dryas (YD). Pollen spectra from the Pleistocene/Holocene transition dated to ca. 11.5–11.2 cal ka BP reflect a shift in vegetation from herb-dominated to shrubby birch–willow tundra. Shrub alder occurred in the area approximately 10 cal ka BP and disappeared after 3.8 cal ka BP. Dwarf birches, which were broadly distributed in the region during the early and middle Holocene, also almost disappeared after 3.2 cal ka BP, when the vegetation became similar to the modern herb tundra. Quantitative paleoclimatic interpretation of the pollen spectra suggests that T_{VII} was 2–4 °C higher than the present during the early Holocene and became similar to the present after 6.3 cal ka BP.

A continuous pollen record from Lama Lake, from western Taymyr (69°32'N, 90°12'E, site 7 in Figure 1), provides detailed environmental information for the Late Glacial and Holocene in central Siberia (Andreev et al., 2004). At present, the vegetation cover in the region varies, depending on

altitude: dense spruce–larch–birch taiga dominates at the lower elevations, while shrub and herb tundra cover the higher mountains. The pollen data suggest that scarce, steppe-like plant communities dominated the vegetation around the lake during the Late Glacial (Figure 5). Tundra-like communities with arctic dwarf shrubs, sedges, and Brassicaceae (cabbage family) species grew in wetter habitats. Reconstructed climate fluctuations may be correlated with the Bølling/Allerød warming and YD cooling. The Late Glacial/Preboreal transition occurred at about 11.5 cal ka BP. It is characterized by a significant increase in dwarf birch and willow pollen accompanied by a relatively high NAP content, suggesting a broad distribution of shrubby and meadow associations. Abundant spores of *Sphagnum* (peat moss) and Polypodiaceae (polypody fern family) indicate wet habitats around the lake. High contents of larch and alder pollen in the early Holocene sediments indicate that larch occurred in the area as early as 11 cal ka BP, while shrub alder came to the area 200 years later. Spruce did not reach the area before about 10.3 cal ka BP. The spruce pollen content and the total pollen concentration increased dramatically in the deposits dated to ca. 10.1–8.8 cal ka BP, indicating the broad distribution of spruce under warm summer conditions. The paleoclimatic reconstruction from the Lake Lama pollen record suggests that T_{VII} during the early Holocene was ca. 1.5–3.5 °C above modern values. Other paleobotanical records (Andreev et al., 2004 and references therein) also confirm the broad distribution of spruce in the western Taymyr during this interval. A gradual decrease in spruce pollen in the sediments younger than 5.2 cal ka BP reflects a gradual deterioration of the regional climate. A significant increase in birch pollen percentages in the upper part of the Lake Lama core likely mirrors the increased role of birches in the local forests after 2.5 cal ka BP. The sharp decrease in the arboreal pollen (AP) content and the increase in sage pollen content at 30–15 cm depth may be correlated with the Little Ice Age.

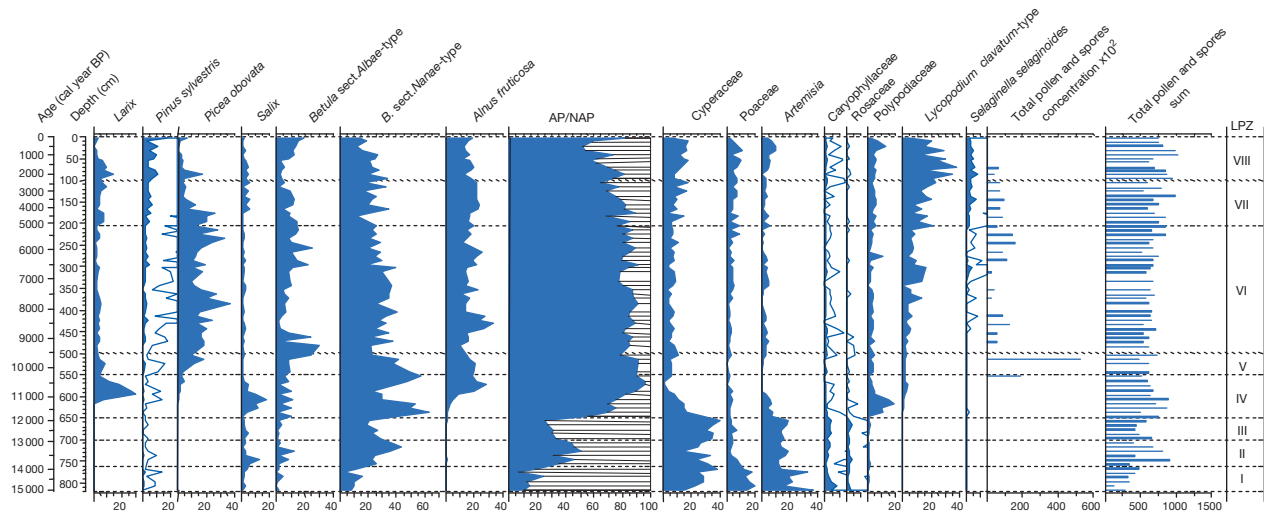


Figure 5 Percentage palynological diagram of selected taxa from Lake Lama record. Pollen zonation is according to Andreev et al. (2004).

East Siberia

The pollen record from Dolgoe Lake, situated in the lower Lena River (71°52'N, 127°04'E, site 8 in Figure 1), is valuable for the reconstruction of paleoenvironments in the present-day larch forest-tundra, dominated by dwarf birch and shrub alder (Pisaric et al., 2001). Sedges, grasses, and sages dominate pollen spectra deposited shortly before 14.5 cal ka BP, reflecting open vegetation with only a sparse cover of herbs and grasses around the lake. An increase in shrub birch pollen higher in the sequence points to the establishment of shrub birch tundra between 14.5 and 13 cal ka BP. A sharp decrease in shrub birch pollen and rise in herbaceous pollen percentages between 13 and 11.5 cal ka BP suggest a return to a more open vegetation cover. The timing of this vegetation shift is synchronous with YD oscillation. After 11.5 cal ka BP, shrub birch pollen percentages rise, associated with the onset of the Holocene. Around 11 cal ka BP, birch was replaced as the dominant taxon by alder, likely by shrub alder (*A. fruticosa*). The increase in larch pollen percentages after 9.5 cal ka BP, coupled with the presence of larch in the stomata and macrofossil records, suggests that it grew north of Dolgoe Lake between 9.5 and 3.8 cal ka BP. At about 7.6 cal ka BP, spruce increased in abundance in the pollen record. Today, spruce grows 350 km to the south. However, the presence of its pollen and stomata in the lake record indicates that spruce occurred in the local vegetation between 7.6 and 3.8 cal ka BP, providing evidence of a considerable northward range extension in the mid-Holocene. Low AP percentages and AP concentrations in the sediments dated younger than 3.8 cal ka BP suggest a change in the environment and a shift from predominantly woodland vegetation to the modern shrub tundra with isolated larch stands.

Khomustakh Lake (63°43'N, 129°22'E, site 9 in Figure 1) has been selected as a key site for the study of postglacial vegetation in the central part of eastern Siberia (Velichko et al., 1997). At present, pine forests occupy sandy soils in the area, whereas clayey soils are occupied by larch forests. Sedimentation and pollen accumulation in the lake started about 12 cal ka BP (Figure 6), probably as a consequence of the Allerød climate

amelioration. Rather high levels of dwarf birch, sage, grasses, and other herbaceous pollen taxa point to their important role in the local vegetation, codominant with open herb and shrub tundra-like vegetation. A subsequent increase in sage and other herb pollen contents likely reflects the YD cooling. The transition to the Holocene is dated to about 11.5 cal ka BP (see International Stratigraphic Chart website) and is marked by a dramatic increase in tree birch pollen. During the Preboreal interval, birch dominated the vegetation; however, the permanent presence of larch in the pollen spectra indicates that it also was an important component in the local forests until about 6.8 cal ka BP. After 7 cal ka BP, Scots pine migrated to the area and quickly occupied sandy habitats in central and southern Yakutia. Quantitative climate reconstruction based on the Khomustakh pollen record suggests that T_{VII} was up to 1.5 °C warmer than the present in the middle Holocene, causing degradation of the permafrost and facilitating the spread of pine. Since 6.8 cal ka BP, forests composed of pine started to dominate in these regions. Recent paleoecological studies (Müller et al. (2009) site 11 and Werner et al. (2010) site 10 in Figure 1) reveal similar vegetation changes.

South Siberia

This area is extremely important from the paleobotanical viewpoint, because it appears likely that, during the last glaciation, the hills and mountains of southern Siberia provided refuge for the tree species that make up the present boreal forest belt (Grichuk, 1984). Postglacial pollen records from this region mainly come from Lake Baikal and its vicinity (Demske et al., 2005; Tarasov et al., 2002; and references therein). The pollen diagram (Figure 7) from Lake Kotokel (52°46'N, 108°46'E, 458 m, site 12 in Figure 1), situated near the eastern coast of Lake Baikal, illustrates changes in vegetation since the last glacial (Bezrukova et al., 2010; Tarasov et al., 2002, 2009). At present, the lake is surrounded by boreal forest dominated by Scots and Siberian pine, larch, and birch trees. Stands of shrub pine (*P. pumila*), birch, and alder are abundant in the

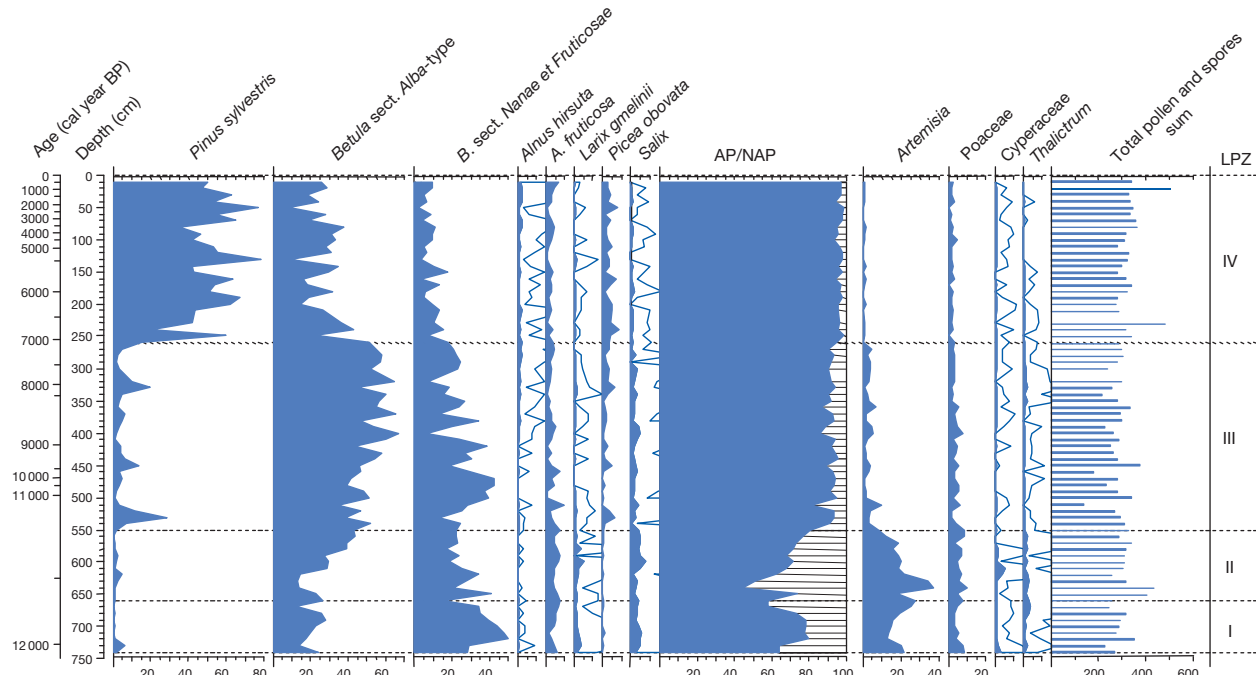


Figure 6 Percentage palynological diagram of selected taxa from Lake Khomustakh record. Pollen zonation is according to Velichko et al. (1997).

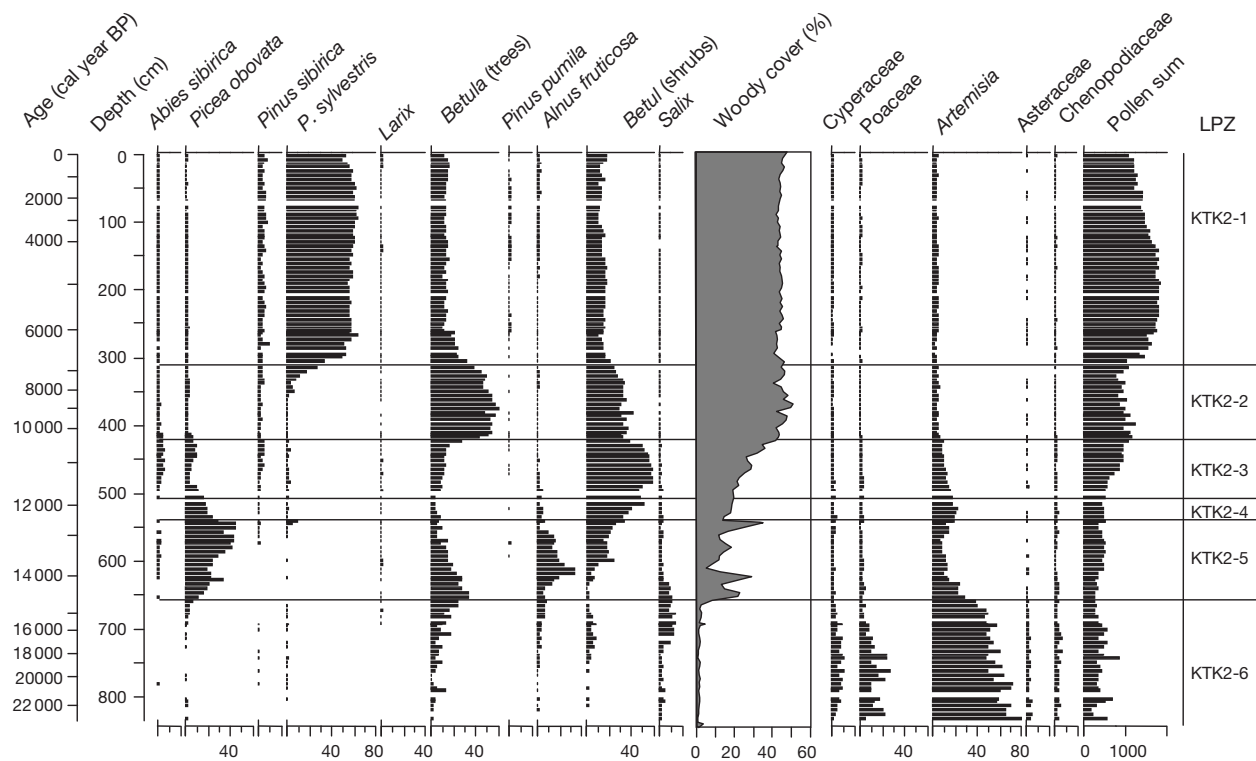


Figure 7 Percentage palynological diagram of selected taxa from Lake Kotokel record. Pollen zonation and quantitative changes in woody cover percentages within a 21×21 km window are according to Bezrukova et al. (2010).

mountain tundra belt, beyond the continuous treeline. The interval prior to about 14.7 cal ka BP was very dry and cold, as indicated by high percentages for steppe herbs and pollen from woody plants below 5% (Figure 7). Since that time,

climate became warmer and wetter, as marked by a gradual increase in tree and shrub pollen percentages and in woody coverage to 20–30% ca. 14.5–14 cal ka and ca. 13.3–12.8 cal ka BP. These two intervals correspond to the Meiendorf and

Allerød interstadials, which were interpreted as part of the undivided Bølling/Allerød interstadial in the Baikal region (Bezrukova et al., 2005). The increase in tundra shrubs and sedge pollen percentages indicating the YD in the region between 12.7 and 11.65 cal ka BP is in agreement with the formal definition and dating of this cold oscillation based on the Greenland NGRIP ice core records. The maximal spread of the taiga communities in the region is associated with a warmer and wetter climate than the present prior to ca. 7 cal ka BP. This was followed by a wide spread of Scots pine, indicating the onset of modern environments. Since that time, pollen spectra show very little change, likely indicating that vegetation composition and distribution patterns became similar to those of today.

Mongolia

A comprehensive synthesis of pollen and plant macrofossil records from Mongolia is presented by Gunin et al. (1999). That monograph presents reconstructions of Late Glacial and Holocene vegetation dynamics. Several other papers discuss vegetation and environmental changes at the regional scale (Fowell et al., 2003; Rudaya et al., 2009; Tarasov et al., 2000, 2004; and references therein). A ^{14}C -dated pollen record from Hoton–Nur (48°40'N, 88°18'E, 2083 m), a large freshwater lake in the northern Mongolian Altai (site 13 in Figure 1), spans the whole Holocene (Figure 8). Pollen concentration and preservation in the lower part of the 9.2-m record from Hoton–Nur was extremely poor, suggesting severe 'glacial' conditions and scarce vegetation around the lake. Pollen spectra of the Ho-3 pollen zone showed relatively high frequencies of pollen of Arctic–alpine taxa, including shrub alder and birch, sedges, and grasses (Gunin et al., 1999). This suggests a slight amelioration of the Late Glacial climate and the spread of tundra vegetation at higher elevations and river valleys. However, high percentages of sage and Chenopodiaceae pollen in the Ho-3 zone suggest that before about 10.1 cal ka BP, the

dominant vegetation close to the lake was dry steppe (Tarasov et al., 2000). In the Ho-2 pollen zone dated to ca. 10.1–4.5 cal ka BP, the dominance of spruce, pine (most probably Siberian pine), and larch pollen likely indicates that patches of boreal conifers played a more important role in the local vegetation than they do today. Expansion of boreal evergreen conifers in the region would require noticeably wetter conditions than those prevailing today. However, the presence of relatively high percentages of herbaceous pollen suggests an open mosaic of forest–steppe-like vegetation. Pollen from the uppermost zone demonstrates that vegetation around the lake reflected drier conditions and became similar to the modern steppe with small forest patches of larch and Siberian pine, taxa that are less sensitive to water stress than spruce. Quantitative reconstruction of the Holocene vegetation and climate dynamics in the semiarid Mongolian Altai (Rudaya et al., 2009) suggests that boreal woodland replaced the primarily open landscape of northwestern Mongolia at about 10 cal ka BP in response to a noticeable increase in P from 200–250 to 450–550 mm. A decline of the forest vegetation and a return to a predominance of open vegetation types occurred after 5 cal ka BP when P decreased to 250–300 mm.

Northeastern Asia

A sediment core from the Smorodinovoye Lake in the Upper Kolyma region (64°46'E, 141°06'E, site 14 in Figure 1) provides a ca. 27 cal ka record of vegetation changes from northeastern Siberia (Anderson et al., 2002). High percentages of grass and sage pollen and the variety of xeric taxa such as rock spike moss (*Selaginella rupestris*) (Figure 9) suggest that open grass–sage vegetation dominated the landscape during the Late Glacial. A dramatic rise in birch percentages around 12.8 cal ka BP implies rapid establishment of shrub tundra in response to postglacial warming. Birch–willow tundra dominated the vegetation between ca. 12.8 and 11.8 cal ka BP. A decline in shrub birch and an increase

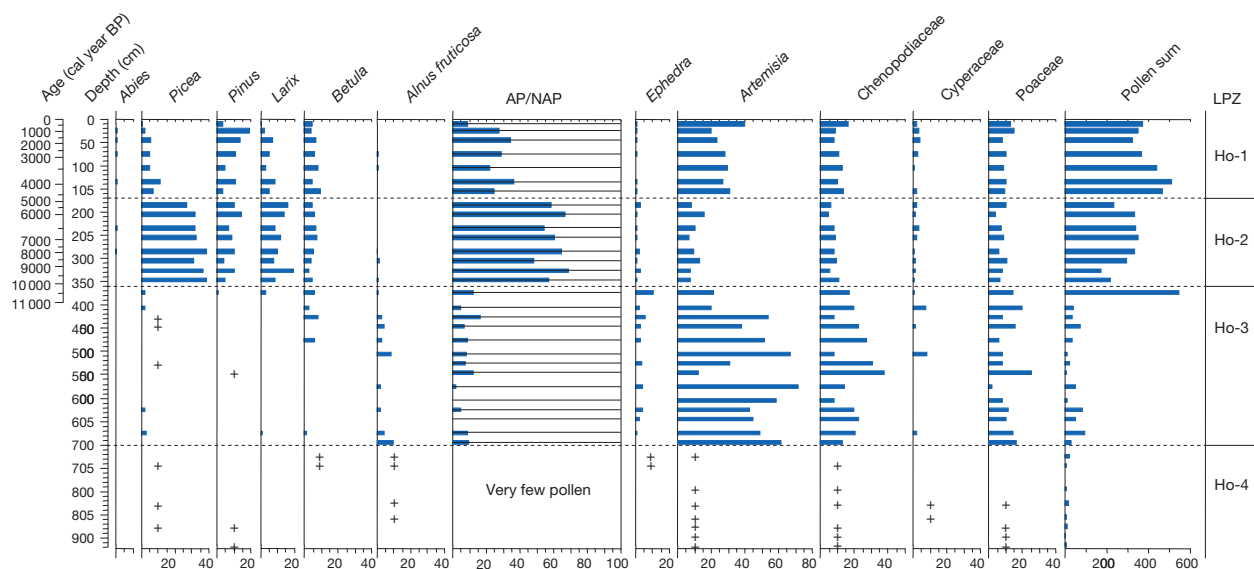


Figure 8 Percentage palynological diagram of selected taxa from Lake Hoton–Nur record. Pollen zonation is according Tarasov et al. (2000).

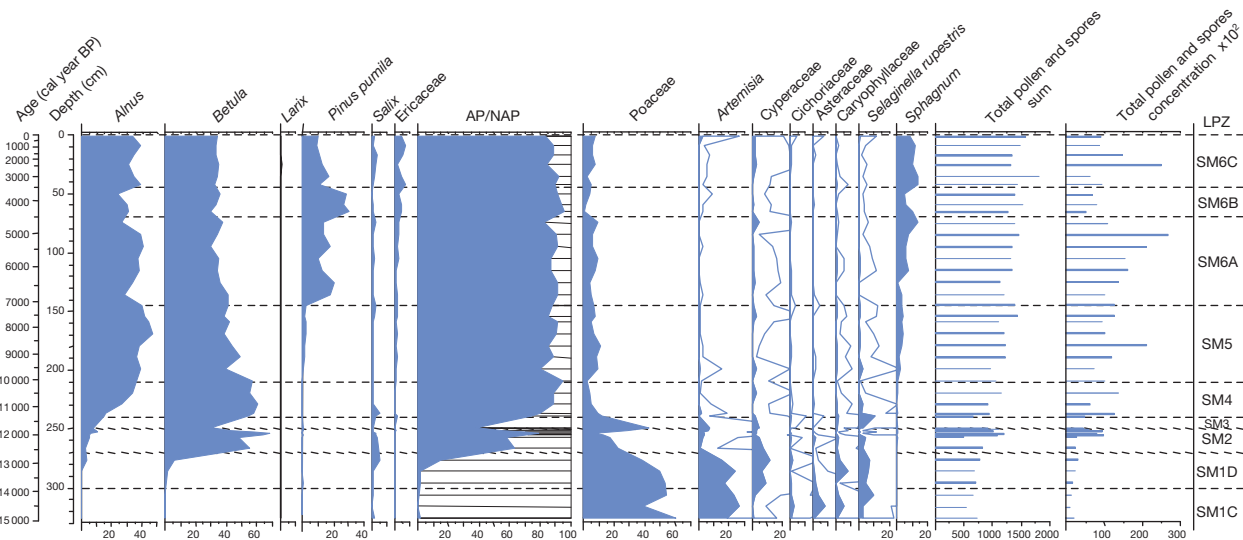


Figure 9 Percentage palynological diagram of selected taxa from Lake Smorodinovoye record. Pollen zonation is according to Anderson et al. (2002).

in grass and sage pollen percentages indicate that the vegetation reverted to herb-dominated tundra shortly after 11.8 cal ka BP. The vegetation change suggests that summer temperatures and *P* became lower than during the previous interval (12.8–11.8 cal ka BP), but not as extreme as during the Late Glacial. This short-term episode most likely represents the YD cooling. An increase in birch and willow pollen from ca. 11.3–9.5 cal ka BP suggests that shrub tundra again became dominant with the onset of the Holocene. Alder percentages increased at ca. 11 cal ka BP, but alder was not widespread until about 10 cal ka BP. The presence of a larch needle and single larch pollen grains indicates that larch was present in the area as early as about 10.9 cal ka BP. During 9.5–7 cal ka BP, birch and alder continued to dominate the pollen assemblage, indicating that shrub tundra with isolated larch trees dominated the vegetation. Shortly after 7 cal ka BP, shrub pine occurred in the area, as suggested by the composition of the pollen assemblages. Pollen spectra suggest that since that time, birch, alder, and heath shrubs were common on the landscape and, along with the dwarf Siberian pine, were abundant in the understory of the larch forest. High shrub tundra, dominated by dwarf Siberian pine, also became established on the mountain slopes above the altitudinal tree limit.

Conclusions

Generally, the pollen records discussed above demonstrate spatially and temporally coherent patterns of environmental changes. Open sage- and grass-dominated communities were widespread in northern Asia during the Late Glacial. A significant increase of shrub pollen registered in several of the records may be correlated with the Bølling/Allerød warming, ca. 13.7 cal ka BP. The subsequent increase in NAP reflects the cooling/drying of the YD interval. Many records show that the Late Glacial/Preboreal transition occurred at about 11.5 cal ka BP and was characterized

by a significant increase in birch, shrub alder, and willow pollen and in woody cover percentages (Tarasov et al., 2007; Williams et al., 2011). Records from geographically different sites demonstrate different patterns of vegetation changes connected with local environmental conditions and migration history of the arboreal species. However, the interval from ca. 10.5 to ca. 8.8 cal ka BP was the warmest postglacial episode (the Holocene climatic optimum) in the arctic regions of northern Asia (e.g., Andreev et al., 2005; Velichko et al., 1997). In contrast, records from the more southerly sites, within the boreal forest and forest-steppe zone, show an early Holocene climate amelioration that is less pronounced than that of the middle Holocene. The explanation of these phenomena can be found in the unequal influence of the lower-than-present sea level and higher-than-present summer isolation during the Late Glacial and early Holocene.

See also: **Pollen Methods and Studies: Databases and Their Application.** **Pollen Records, Late Pleistocene: Northern Asia.**

References

- Anderson PM, Lozhkin AV, and Brubaker LB (2002) Implications of a 24,000-yr palynological record for a Younger Dryas cooling and for boreal forest development in northeastern Siberia. *Quaternary Research* 57: 325–333.
- Andreev AA and Klimanov VA (2000) Quantitative Holocene climatic reconstruction from Arctic Russia. *Journal of Paleolimnology* 24: 81–91.
- Andreev AA, Tarasov PE, Klimanov VA, et al. (2004) Vegetation and climate changes around the Lama Lake, Taymyr Peninsula during the late Pleistocene and Holocene reconstructed from pollen records. *Quaternary International* 122: 69–84.
- Andreev AA, Tarasov PE, Ilyashuk BP, et al. (2005) Holocene environmental history recorded in the Lake Lyadhej-To sediments, Polar Urals, Russia. *Palaeogeography, Palaeoclimatology, Palaeoecology* 223: 181–203.
- Andreev AA, Tarasov PE, Siegert Ch, et al. (2003) Vegetation and climate changes on the northern Taymyr, Russia during the Upper Pleistocene and Holocene reconstructed from pollen records. *Boreas* 32: 484–505.
- Bezrukova EV, Abzaeva AA, Letunova PP, et al. (2005) Postglacial history of Siberian spruce (*Picea obovata*) in the Lake Baikal area and the significance of this species as a paleo-environmental indicator. *Quaternary International* 136: 47–57.

- Bezrukova EV, Tarasov PE, Solovieva N, et al. (2010) Vegetation and environmental dynamics in southern Siberia during the last 47 kyr derived from pollen and diatom records from Lake Kotokel. *Palaeogeography, Palaeoclimatology, Palaeoecology* 296: 185–198.
- Blyakharchuk TA and Sulerzhitsky LD (1999) Holocene vegetational and climatic changes in the forest zone of Western Siberia according to pollen records from the extrazonal palsa bog Bugristoye. *The Holocene* 9: 621–628.
- Demske D, Heumann G, Granoszewski W, et al. (2005) Late glacial and Holocene vegetation and regional climate variability evidenced in high-resolution pollen records from Lake Baikal. *Global and Planetary Change* 46: 255–279.
- Dokturovskii VS and Kudryashov VV (1923) Pollen in peat. *Izvestiya Nauchno-eksperimental'nogo Torfyanogo Instituta* 5: 33–44 (in Russian).
- Edwards ME, Anderson PM, Brubaker LB, et al. (2000) Pollen-based biomes for Beringia 18,000, 6000 and 0 ¹⁴C yr BP. *Journal of Biogeography* 27: 521–554.
- Fowell SJ, Hansen BCS, Peck JA, et al. (2003) Mid to late Holocene climate evolution of the Lake Telmen basin, north central Mongolia, based on palynological data. *Quaternary Research* 59: 353–363.
- Grichuk VP (1984) Late Pleistocene vegetation history. In: Velichko AA (ed.) *Late Quaternary Environments of the Soviet Union*, pp. 155–178. Minneapolis, MN: University of Minnesota Press.
- Gunin PD, Vostokova EA, Dorofeyuk NI, et al. (eds.) (1999) *Geobotany 26: Vegetation Dynamics of Mongolia*. Dordrecht: Kluwer Academic.
- Khotinskii NA (1977) *Golotsen Severnoi Evrazii*. Moscow: Nauka (in Russian).
- Kremenetski CV, Tarasov PE, and Cherkinsky AE (1997) Postglacial development of Kazakhstan pine forests. *Géographie physique et Quaternaire* 51: 391–404.
- Müller S, Tarasov PE, Andreev AA, and Diekmann B (2009) Late Glacial to Holocene environments in the present-day coldest region of the Northern Hemisphere inferred from a pollen record of Lake Billyakh, Verkhoyansk Mts., NE Siberia. *Climate of the Past* 5: 74–94.
- Neishtadt MI (1957) *Istoriya Lesov SSSR v Golotsene*. Moscow: Publishing House of Academy of the Sciences of the USSR (in Russian).
- Peteet D, Andreev A, Bardeen W, and Mistretta F (1998) Long-term arctic peatland dynamics, vegetation and climate history of the Pur-Taz region, Western Siberia. *Boreas* 27: 115–126.
- Peterson GM (1993) Vegetational and climatic history of the western former Soviet Union. In: Wright HE Jr., Kutzbach JE, and Webb T, III et al. (eds.) *Global Climates since the Last Glacial Maximum*, pp. 169–193. Minneapolis, MN: University of Minnesota Press.
- Pisarcic MFJ, MacDonald GM, Velichko AA, and Cwynar LC (2001) The Lateglacial and Postglacial vegetation history of the northwestern limits of Beringia based on pollen, stomate and tree stump evidence. *Quaternary Science Reviews* 20: 235–245.
- Pitkänen A, Turunen J, Tahvanainen T, and Tolonen K (2002) Holocene vegetation history from the Salym-Yugan mire area, West Siberia. *The Holocene* 12: 353–362.
- Rudaya N, Tarasov P, Dorofeyuk N, et al. (2009) Holocene environments and climate in the Mongolian Altai reconstructed from the Hoton-Nur pollen and diatom records: A step towards better understanding climate dynamics in Central Asia. *Quaternary Science Reviews* 28: 540–554.
- Tarasov PE, Bezrukova EV, and Krivonogov SK (2009) Late glacial and Holocene changes in vegetation cover and climate in southern Siberia derived from a 15 kyr long pollen record from Lake Kotokel. *Climate of the Past* 5: 73–84.
- Tarasov P, Brovkin V, and Wagner M (2005) What drives the climate? *PAGES News* 13: 24–25.
- Tarasov P, Dorofeyuk N, and Metel'tseva E (2000) Holocene vegetation and climate changes in Hoton-Nur basin, northwest Mongolia. *Boreas* 29: 117–126.
- Tarasov PE, Dorofeyuk NI, Sokolovskaya VT, et al. (2004) Late Glacial and Holocene vegetation changes recorded in the pollen data from the Hangai mountains, Central Mongolia. In: Yasuda Y and Shinde V (eds.) *Monsoon and Civilization*, pp. 23–50. New Delhi: Roli Books.
- Tarasov PE, Dorofeyuk NI, and Vipper PB (2002) The Holocene dynamics of vegetation in Buryatia. *Stratigraphy and Geological Correlation* 10: 88–96.
- Tarasov PE, Jolly D, and Kaplan JO (1997) A continuous Late Glacial and Holocene record of vegetation changes in Kazakhstan. *Palaeogeography, Palaeoclimatology, Palaeoecology* 136: 281–292.
- Tarasov PE, Webb T III, Andreev AA, et al. (1998) Present-day and mid-Holocene biomes reconstructed from pollen and plant macrofossil data from the former Soviet Union and Mongolia. *Journal of Biogeography* 25: 1029–1053.
- Tarasov P, Williams JW, Andreev A, et al. (2007) Satellite- and pollen-based quantitative woody cover reconstructions for northern Asia: Verification and application to late-Quaternary pollen data. *Earth and Planetary Science Letters* 264: 284–298.
- Velichko AA, Andreev AA, and Klimanov VA (1997) Climate and vegetation dynamics in the tundra and forest zone during the Late Glacial and Holocene. *Quaternary International* 41/42: 71–96.
- Volkova VS and Mikhailova IV (2002) Evolution of geological processes, environment and climate of the Holocene in Siberia. In: Vaganov EA, Derevyanko AP, and Grachev MA, et al. (eds.) *Osnovnye Zakonomernosti Global'nykh i Regional'nykh Izmenenii Klimata i Prirodnoi Sredy v Pozdnem Kainozoe Sibiri*, pp. 58–70. Novosibirsk: Izdatel'stvo Instituta Archaeologies i Etnografii.
- Werner K, Tarasov P, Andreev A, et al. (2010) A 12.5-ka history of vegetation dynamics and mire development with evidence of the Younger Dryas forest presence in the Verkhoyansk Mountains, East Siberia, Russia. *Boreas* 39: 56–68.
- Williams JW, Tarasov P, Brewer S, and Notaro M (2011) Late Quaternary variations in tree cover at the northern forest-tundra ecotone. *Journal of Geophysical Research* 116: G01017. <http://dx.doi.org/10.1029/2010JG001458>.

Relevant Websites

- <http://www.europeanpollendatabase.net/> – European Pollen Database.
- <http://www.ncdc.noaa.gov> – National Climatic Data Center, NOAA Satellite and Information Service, for access to pollen databases.
- <http://www.ncdc.noaa.gov/paleo/gpd.html> – Global Pollen Database.
- <http://www.stratigraphy.org/column.php?id=Chart/Time%20Scale> – International Stratigraphic Chart.

Northern Europe

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Introduction

Pollen records retrieved from sediment accumulations provide insights into past vegetation and thus past habitat and environment. Not surprisingly, pollen records from the current interglacial (the Holocene) occur more abundantly and have been subjected to more detailed study than records from any other interval. This article reviews some of the key issues raised by pollen records from the postglacial interval in northern Europe. For the purposes of this work, the postglacial interval is considered to be the time from the onset of full interglacial conditions at the start of the Holocene until the present, and northern Europe includes the Scandinavian countries.

Terrestrial pollen records are usually retrieved from waterlogged sedimentary environments such as peat bogs and lakes; records from such environments are both likely to be well preserved and to have the best stratigraphic integrity. These sedimentary systems form in low lying, wet parts of the landscape and thus are most common in areas with variable topography and/or impermeable substrates. One of the legacies of glaciation in northern Europe is a landscape with many suitable places for initiation of sediment accumulation, with a wide range of depressions formed by processes such as scour and erosion of bedrock, or deposition of moraine dams and of undulating till fields which include features such as kettle holes. Fluvial environments also provide preservation opportunities, from paleochannel infilling and deposition in oxbow lakes to the development of floodplain-wide wetlands such as alder carrs. Other suitable preservation locales are formed by coastal processes (e.g., sea-level isolation basins, salt marsh sediments). Many important records have also been retrieved from peat deposits such as raised mires and blanket peat; the formation of these sedimentary archives is controlled primarily by climate factors and the presence of impermeable substrates allowing waterlogging rather than underlying topography, but initiation of peat accumulation often postdates the end of the glacial interval by several thousand years giving only partial records of the postglacial (Barber, 1993).

Despite the range of sediment-forming contexts available, site distribution is patchy. Sites are not distributed randomly across the region, and so pollen records are not available from all landscapes or from all site types within one landscape area. For example, unglaciated parts of southern England have produced far fewer complete postglacial pollen records than the formerly glaciated uplands of the Lake District in northwest England. Lack of data for some parts of the landscape is therefore still an important limitation, although less so in northern Europe than in most of the rest of the world. The concentration of population in northern Europe has led to a concentration of universities and other research institutes that can support palynological research and thus ensure that the local area is well studied; this too can lead to patchy coverage. For example, in Scandinavia, more records are available for the relatively

densely populated south than for the geographically larger but less settled central and northern areas. Wetlands and lakes are also under pressure from human population, through draining and peat extraction, exposing sites that may contain good pollen records and adding urgency to their collection. In places this population density leads to a serious loss of pollen records, especially in lowland agriculturally productive areas where extensive drainage of wetlands and conversion to agricultural uses has substantially reduced the availability of sites that have the potential to yield pollen records.

A selection of European pollen records is curated in the European Pollen Database (EPD), and some of the data can be accessed through the NOAA website or directly through the EPD website. The overview of pollen records for different parts of Europe (mainly northwest Europe) presented in the report on International Geoscience Program project 158B (Berglund et al., 1996) does not cover the whole area, yet at least some of these 'pollen gaps' are due at least in part to an absence of sites rather than a lack of research activity. There are approximately 300 pollen records from northwest Europe archived in the EPD.

Broadly, the postglacial pollen record from northern Europe is a record of the reestablishment of forest biomes in areas occupied during the glacial interval by arctic and subarctic tundra and steppe communities, or by ice. The other key feature of the postglacial interval is the unprecedented spread of human activity both in space and in intensity, which forms part of the argument over nomenclature for the present interglacial. The older British term Flandrian (and other regional terminologies) treats the present postglacial interval as just the latest in a sequence of Pleistocene interglacials (Hyyvärinen, 1978), whereas strictly speaking the term 'Holocene' treats the postglacial as the beginning of a new geological epoch of equivalent status to the Pleistocene (see International Stratigraphic Chart website).

Chronologies

Before the development and application of radiocarbon dating in the 1950s, postglacial pollen records were correlated with each other largely on the basis of the arrival of particular tree species or changes in the pollen assemblage or peat structure which were believed to represent synchronous responses across the region to Holocene climatic events. The Blytt-Sernander scheme divided the Holocene into four broad climatic intervals primarily on the basis of variations in peat stratigraphy and associated plant macrofossils (Von Post, 1916). It was widely applied to pollen records and used as a tool to correlate records across northern Europe (see *History of Quaternary Science; Table 1*). The terminology associated with this scheme (e.g., 'Atlantic period') is still encountered widely in allied literatures and textbooks. In peatlands, periods of changed decay conditions and accumulation rates marked by different colors and

Table 1 The Blytt–Sernander scheme and the commonly used European pollen zone numbering system based on it

<i>Blytt–Sernander zone name</i>	<i>Inferred climate</i>	<i>European pollen zone number</i>	<i>Pollen record characteristics</i>
Sub-Atlantic (most recent)	Cold and wet, oceanic	IX	Climatic deterioration, spread of bogs
Sub-Boreal	Warm and dry, continental	VIII	Mixed-oak forest; some 'late arrival' species occur, for example, beech, hornbeam
Atlantic	Warm and wet, oceanic	VII	Mixed-oak forest: lime and alder arrive at start
Late Boreal	Warm and dry	VI	Oak, elm arrive
Early Boreal	Warm and dry	V	Large amounts of hazel Pine dominance

preservation conditions of peats were also attributed to a response to these major climate changes.

Radiocarbon dating showed the flaws in this simple assumption of synchronicity, but also opened up new areas of investigation. One of the attractions of researching postglacial pollen records is the ready applicability of radiocarbon dating of the organic components to establish a chronology, which enables correlation between sites within regions, with archaeological records, and with other evidence from the wider landscape. However, radiocarbon dating is not without its own challenges, and each radiocarbon-based chronology has associated errors and areas of uncertainty. Tephrochronology offers the possibility of more confident age correlation between sites at some points in the past, but the patchy distribution of distal tephra from distant (mainly Icelandic) sources means that tephrochronology cannot be a replacement for radiocarbon dating in northwest Europe.

Dating shows that the basal age of pollen records across northern Europe varies. This is due in part to the time-transgressive nature of glacial retreat, but the establishment of full interglacial conditions does not always correlate with

the beginning of an organic sedimentary record. For example, the onset of peat formation in upland and exposed coastal locations appears to have been highly time transgressive, depending as it does on a combination of climatic conditions, soil type and permeability, and catchment hydrologic regime. In some locations, the timing of peat development appears to be strongly related to human activity and changes in the tree cover and erosion regimes, and in others, the timing of the creation and abandonment of paleochannel features suggests similar controls.

The Postglacial Pollen Record

Figure 1 is a schematic composite pollen diagram for the postglacial interval in northwest Europe showing the main features characteristic of such diagrams. In this section, some of the key issues raised by the study of diagrams from across the region are discussed, which show strong geographical variation in both the timing and the relative importance of events.

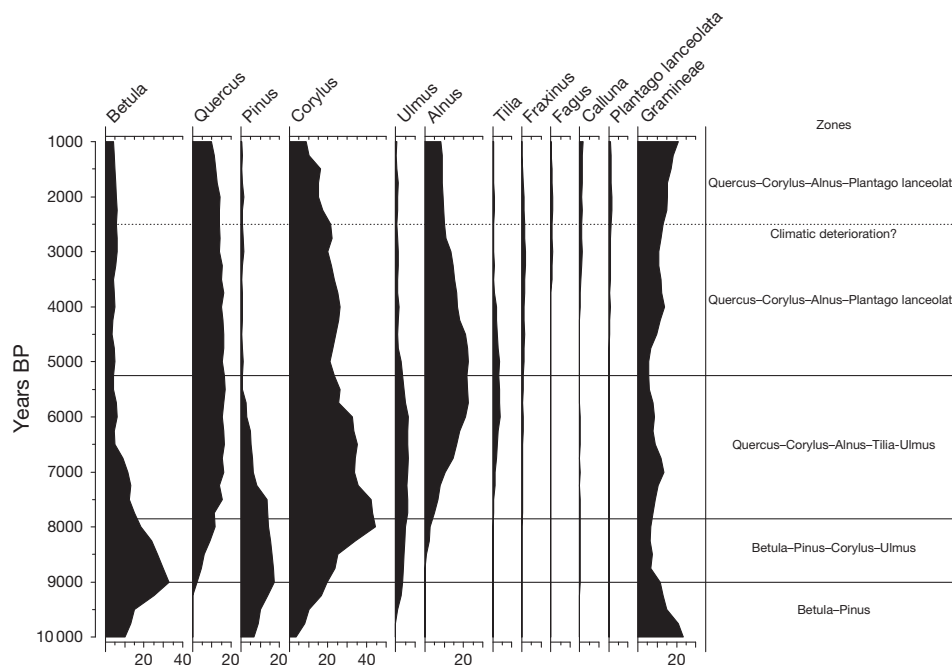


Figure 1 Schematic pollen record from the postglacial of northern Europe. Zone boundaries indicate the approximate limits of the Blytt–Sernander climate zones (see Table 1) in this particular case.

Recolonization by Forest

One area of research opened up by the radiocarbon chronology was the investigation of the rates and routes of the reinvasion of northern Europe by temperate tree species in the first part of the postglacial interval. Tree species appear to have survived glacial periods in refugia in southern regions and on the exposed continental shelf. Recent research has also begun to search for 'cryptic refugia,' microclimatic 'pockets' in the north European plain where some temperate species survived or were able to colonize during Late Glacial times (Stewart and Lister, 2001), allowing rapid expansion as soon as local conditions were suitable.

By using sites with radiocarbon chronologies, it is possible to map the distribution of pollen of particular taxa across Europe at specific points in time (creating isopoll maps) and thus reconstruct their changing distributions. Huntley and Birks (1983) compiled an atlas documenting the spread of tree taxa across Europe during the Late Glacial and postglacial. More recent research has tended to either develop detailed isopoll maps for smaller parts of Europe (Prentice, 1983), or to focus on reconstruction of a single point in the past (e.g., the BIOME 6000 project).

Macrofossil evidence provides an important check and correction on the maps produced, especially at range limits or for taxa that are less well represented in the pollen record. Phytogeography, studying the present-day distribution of genetic markers across a species range, offers additional evidence for migration routes, and the extraction of ancient DNA from subfossil plant remains including pollen grains, seeds, and other plant parts offers opportunities to test reconstructed patterns of population dynamics.

Isopoll mapping shows that taxon response to climate amelioration and the opening up of new potential habitat in northern Europe was essentially individualistic (Huntley and Prentice, 1993). The arrival of the various tree taxa was not synchronous and cannot be correlated across regions (although the assumption of synchronicity is still widely applied within defined areas). In formerly glaciated areas, forest biomes reestablished at the end of a primary succession sequence, in which pioneering mosses and herbs gave way to pioneering shrubs and trees. Different plant species played the dominant roles in this process, across different parts of Europe (Huntley, 2009). The process of invasion of a new species into already established woodland is less clearly understood than initial invasion. Studying the process probably requires stand-scale resolution data, which are not always available. Most postglacial pollen records come from relatively large basins (diameter measured in 10s of meters or more) and therefore reflect a larger area of landscape than the stand (see Jacobson and Bradshaw, 1981). Comparison with other interglacials opens up questions about the reasons for the different timing and pattern of tree invasion and expansion in each interglacial, suggesting that chance factors may be as important as climatic and biological variations. Isopoll maps also provide information for the study of changes in range limits of taxa during the postglacial and changes in treeline composition and position, which can elucidate a climatic signal.

The Wildwood: What Is the Natural Vegetation of Northern Europe?

According to bioclimatic models, northern Europe should largely be tree covered today. The relative openness of the

landscape actually observed is generally attributed to millennia of human activity, and one research focus has been reconstructing and mapping the latest stage of the 'WildWood' (Rackham, 2000), the hypothesized natural forests of northern Europe, before the impact of agricultural activity (Kaplan et al., 2009). These maps are of particular interest as a guide for conservation and landscape restoration activities. This approach assumes, however, that the main anthropogenic impact on vegetation occurred after the widespread adoption of agriculture.

A date of around 6000 cal years BP is often used for these studies. This just predates most records of arable and pastoral agriculture in northwest Europe (Bonsall et al., 2002; Gkiasta et al., 2003). However, it is toward the end of the 'Atlantic' period, which is argued to represent 'better' climatic conditions than today (Bonsall et al., 2002). Certainly over the following millennia, extensive spread of blanket mire (peatland) has been seen in the oceanic climate zone of coastal northwest Europe, and there has been a general southern retraction of northern treelines and downslope movement of upper tree-lines in response to cooler and wetter conditions. These observations suggest that the 5000 ¹⁴C years BP wildwood may not be an attainable target for ecological restoration projects. However, the extent to which these vegetation changes are a pure climate response is debatable. Both human activity and soil processes leading to leaching, gleying, and, from an arboreal plant's perspective, general deterioration of soil conditions in exposed environments as the present interglacial progressed will also have played a part in the changing extent and character of woodland.

Maps showing inferred vegetation for the mid-Holocene (e.g., Bennett, 1989) show something of the great diversity of woodland type that must have prevailed across Europe. However, interpretation of pollen records in terms of vegetation associations and dominance is not simple, and calibration of the pollen record is a key research agenda in contemporary European palynology.

The degree of openness within this wildwood is currently an important and controversial theme for palynologists and landscape managers alike, following Vera's (2000) book, which argued that the wildwood would have had a large proportion of clearings, maintained by large native herbivores. This has led to some reevaluation of the pollen record in terms of openness (e.g., Svenning, 2002), and a lively debate into the interpretation of a range of proxies of past habitat structure is ongoing.

The Elm Decline

When radiocarbon dating was applied to postglacial pollen diagrams, most events (e.g., the arrival of oak and alder in northwest Europe) that had been used by palynologists as biostratigraphic markers were found to be asynchronous. One event stood out; the decline in elm (*Ulmus*), originally identified by Iversen (1941), appeared to be broadly synchronous across northern Europe (Figure 1). A recent review by Parker et al. (2002) shows that statistically the timing of the elm decline varies by about 1000 years across the region, although how much of this is true regional and catchment variation in the event (e.g., real variation in the actual timing of the event), and how much due to the uncertainties of the

dating process (e.g., a synchronous event being 'blurred' by dating error) is hard to establish.

The causes of the elm decline have been extensively debated, and three main possibilities have emerged. These have been reviewed in Pegler and Birks (1993). The first possibility is that the decline is in response to climatic deterioration at the end of the 'Atlantic period' or to soil changes (Parker et al., 2002), although, because elm is apparently the only broad-leaved deciduous forest tree taxon affected, this is not a widely supported hypothesis. A second alternative is direct human agency, partly through the collection of leaf hay for animal feed, which would suppress flowering, and thus, pollen production, and also weaken the tree by creating wounds through which disease could enter. Again, the question of why only elm would be selected for this practice makes this hypothesis less likely. Another possible explanation for a human-caused decline is that elm was preferentially cleared due to particular soil conditions, or that elm was peculiarly vulnerable to competition from other trees in recolonizing areas following human disturbance of woodland during slash and burn type agriculture. The third alternative is that the decline was caused by a severe outbreak of disease affecting elm populations across the region (Pegler and Birks, 1993). This theory is often seen as attractive, especially following the 1980s outbreak of Dutch elm disease in the area. It is supported by a few high-resolution studies of the elm decline at specific sites, for example, at Diss Mere in Norfolk, England, which found a very rapid decline over less than a decade at a time when human activity was not apparent in the record (Pegler and Birks, 1993). The elm bark beetle (*Scolytus scolytus*) is one known vector for the Dutch elm disease-causing agent, a fungus, *Ceratocystis ulmi*. Occasional remains have been recovered from sediments at around the time of the decline (Girling and Greig, 1985). Other apparent pathogen related single-taxon declines occur in other geographical regions, most notably the hemlock (*Tsuga*) decline in eastern North America around 4800 ¹⁴C years BP.

The preferred explanation seems to be that several factors were involved in the elm decline across northern Europe. Human activity was increasing in intensity and changing in type during this mid-Holocene interval, as Neolithic agriculture spread in northern Europe. Woodland management activities such as the collection of leaf hay and increased browsing pressure from domesticated livestock could well have provided increased opportunities for invasion of a disease vector into otherwise healthy trees. Even if human activity were neither the primary cause nor an exacerbating factor in the elm decline, humans are likely to have taken advantage of the opening up of woodland caused by the death of elm trees. It is easier to clear diseased or dead trees than it is to remove living ones, and the natural glades created by the death of trees will have been attractive to grazing animals. Therefore, human activity may have been sufficient to prevent extensive elm regeneration.

Human Activity

The timing and nature of early agricultural activity is a challenge for pollen research. Vegetation disturbance caused by human activity in the landscape before the widespread adoption of agriculture can be detected in pollen records. Especially in the northern and western fringes of Europe, the detection of

Mesolithic period human-mediated modification of landscape is an increasingly active research field (e.g., Whitehouse and Smith, 2004). One way in which people often modify vegetation is by the use of fire. Changes in fire regime can be investigated by analysis of charred particles on pollen slides as well as the investigation of macroscopic charcoal recovered from sediments. Such investigations are an important adjunct to the pollen data and provide more detailed information about landscape change. Whether the first farming in a region occurred as 'slash and burn' style shifting cultivation or the establishment of permanent settlements with gardens, initial openings of the forest canopy will have been small. Detection of small openings in the forest is not easy using pollen signals. Charred particle analyses are often used as evidence of increased fire incidence associated with landscape use by human populations.

Pollen markers for early, that is, Neolithic, agricultural activity include increases in the presence of herbs that have a strong association with human activity, such as ribwort plantain (*Plantago lanceolata*). However, because these taxa also occur naturally in the landscape and do not produce large amounts of pollen compared to trees, their detection is not always easy. Cereal crops are generally poorly represented in pollen records. Among the cereals, only the pollen of rye (*Secale*) can be identified to species. Plant macrofossil evidence indicates that early farmers mostly grew cultivars of wheat (*Triticum*) and barley (*Hordeum*). Comparison of archaeological and paleoecological evidence of cereal agriculture and of human presence often shows that the pollen record evidence 'lags' behind the archaeological evidence.

Evidence of human activity and of agriculture is time transgressive across the European pollen record, and further complicated by the patchy and nonrandom distribution of pollen sites in the landscape. Early agricultural settlement and landscape modifying activity seem to focus largely on areas with light, well-drained soils, which were probably relatively thinly forested and were also most suitable for Neolithic agricultural practices. Archaeological evidence is clustered on areas such as the chalk and limestone Downs and Wolds of southern Britain. This may partly be an artifact of discovery because later peatland and wetland spread may obscure much of the Neolithic landscape in lowland areas. In areas such as the Rhine valley, where deep alluvium deposits have built up during the last few thousand years, the archaeological record of the first farmers is likely to be buried under several meters of sediment. Areas with chalk or sand substrates, apparently attractive to Neolithic agriculturalists, also tend to lack sites suitable for the preservation of pollen records (Brown, 1997).

The main phase of clear human activity-related changes in the landscape recorded in postglacial pollen records from northern Europe is therefore the first broad opening up of the woodland canopy. This likely occurred due to population increase leading to more land being cleared for use or as a result of pastoral activities. The grazing of domesticated livestock is usually managed to varying degrees, generally leading to a concentration of grazers in particular areas such as those within easy reach of the base settlement. Use of land by livestock, including cattle, sheep, goats, and pigs, can substantially suppress the regeneration of woodland tree and shrub species either because the animals deliberately seek out and

preferentially eat seedlings or because seedlings are destroyed as a by-effect of grazing (Davis, 1987). With time, this can lead to greater openness of the woodland canopy as older trees gradually die without replacement, leading to woodland clearance by default rather than by active removal of timber or deliberate clearance. The detection of landscape openness, and its implications for the density of settlement and nature of agricultural practices, is a key driver behind the PollandCal project (<http://www.ecrc.ucl.ac.uk/pollandcal/>).

Another major impact of human activity has been its role in changing the distribution of species (see Crosby, 2003). Examples of this include the introduction into northern Europe of nonnative herbivores (including sheep, goats, and rabbits) and of crop species (such as maize, potatoes, and flax) and associated weeds. The introduction of plants for ornamental and practical purposes has accelerated in the last millennium, as has the translocation of plants beyond their range limits. For example, to the best of our knowledge, larch (*Larix*) and spruce (*Picea*) are not native to the United Kingdom in the present interglacial (Bennett, 1989), yet both are planted widely in the uplands and are recorded in more recent pollen records. In some cases, introductions have had a profound ecological effect on the landscape and on ecosystems. Introduced species may become highly invasive and may displace or outcompete native species; in Britain, *Rhododendron ponticum* (Dehnen-Schmutz and Williamson, 2006) and the gray squirrel (Manchester and Bullock, 2000) are two notable examples that have affected recent vegetation. Other plant species have been moved beyond their apparent natural range limits, for example, with the planting of beech (*Fagus*) and lime (*Tilia*) across northern areas (Bennett, 1989). The macrofossil record is a far more sensitive detector of arriving plant species than the pollen record, but their presence does show up in some pollen diagrams. Relatively recent arrivals and ornamental plantings such as spruce (*Picea*) plantations can actually serve as useful age marker horizons in the pollen record of the last few hundred years.

Current Research Challenges

The postglacial interval in northern Europe is richly represented in pollen records, far more so than any other part of the Quaternary, but this richness of records simply opens up more questions. Reconstructing the postglacial vegetation dynamics of the region is still challenging. Site positioning is not random, and some areas with distinctive geologies, soils, and therefore, presumably ecologies are still 'palynological deserts,' poorly represented or entirely absent from the available database of pollen records. Site size and the dynamics of wetland sedimentary systems are also a challenge for interpretation, because changes in basin size and surface type are known to affect the spatial resolution of the pollen signal preserved. Therefore, one site, undergoing hydrosere succession, records different spatial samples of the surrounding landscape in different seral stages than nearby sites. Comparison between sites of different character must also be adjusted to allow for this effect. Our understanding of the complex relationship between vegetation mosaics and the pollen assemblage forming at a sample point within that mosaic is improving, but there are still many uncertainties in this vital area, particularly around

questions of openness of past forest canopies. Future developments should also progress toward a more integrated paleoecology, drawing on the full range of biotic records available in interpretation and reconstruction.

Intersite comparison and combination of sites for reconstruction of broader patterns of vegetation dynamics also depends on the establishment of reliable chronologies. So far, this has mostly been done by radiocarbon dating, with all the associated uncertainties. The discovery of tephra layers of various ages in some parts of the region holds out the promise of greater confidence in intersite correlation at some specific points in the postglacial. Correlation with the marine record presents a further challenge, which may yield valuable insights into the complex connections between climate and human activity as drivers of postglacial vegetation history in northern Europe.

Summary

Northern Europe has an extensive archive of pollen records of the postglacial interval, but there are still gaps in the spatial coverage. The pollen record reveals the patterns of reinvasion of tree taxa following the last glaciation, as well as the increasing intensity and diversity of human impact on the landscape. However, interpretation of the pollen signal is a developing field, with much still to be done before we can reconstruct past landscapes with confidence and get full value from the large body of data available for the region.

See also: Pollen Analysis, Principles. **Pollen Methods and Studies:** Changing Plant Distributions and Abundances; POLLSCAPE Model: Simulation Approach for Pollen Representation of Vegetation and Land Cover. **Pollen Records, Late Pleistocene:** Middle and Late Pleistocene in Southern Europe. **Quaternary Stratigraphy:** Tephrochronology. **Radiocarbon Dating:** Conventional Method; Sources of Error.

References

- Barber KE (1993) Peatlands as scientific archives of past biodiversity. *Biodiversity and Conservation* 2: 474–489.
- Bennett KD (1989) A provisional map of forest types for the British Isles 5000 years ago. *Journal of Quaternary Science* 4: 141–144.
- Berglund BE, Birks HJB, Ralska-Jasiewiczowa M, and Wright HE Jr. (eds.) (1996) *Palaeohydrological Events During the Last 15,000 Years: Regional Syntheses of Palaeoecological Studies of Lakes and Mires in Europe*. Chichester: Wiley 764 pp.
- Bonsall C, Macklin MG, Anderson DE, and Payton RW (2002) Climate change and the adoption of agriculture in north-west Europe. *European Journal of Archaeology* 5: 9–23.
- Brown AG (1997) *Alluvial Geoarchaeology: Floodplain Archaeology and Environmental Change*. Cambridge: Cambridge University Press ch. 6, pp. 192–218.
- Crosby AW (2003) *The Columbian Exchange: Biological and Cultural Consequences of 1492*, 2nd edn New York: Praeger 320 pp.
- Davis SJM (1987) *The Archaeology of Animals*. New Haven, CT: Yale University Press.
- Dehnen-Schmutz K and Williamson M (2006) *Rhododendron ponticum* in Britain and Ireland: Social, economic and ecological factors in its successful invasion. *Environment and History* 12: 325–350.
- Girling MA and Greig J (1985) A first fossil record for *Scolytus scolytus* (F.) (elm bark beetle): Its occurrence in elm decline deposits from London and the implications for neolithic elm disease. *Journal of Archaeological Science* 12: 347–351.

- Gkiasta M, Russell T, Shennan S, and Steele J (2003) Neolithic transition in Europe: The radiocarbon record revisited. *Antiquity* 77: 45–62.
- Huntley B (2009) European post-glacial forests: Composition changes in response to climate change. *Journal of Vegetation Science* 1: 507–518.
- Huntley B and Birks HJB (1983) *An Atlas of Past and Present Pollen Maps for Europe: 0–13,000 Years Ago*. Cambridge: Cambridge University Press.
- Huntley B and Prentice IC (1993) Holocene vegetation and climates of Europe. In: Wright HE Jr., Kutzbach JE, Webb T III, Ruddiman WF, Street-Perrott FA, and Bartlein PJ (eds.) *Global Climates Since the Last Glacial Maximum*. Minneapolis, MN: University of Minnesota Press.
- Hyvärinen H (1978) Use and definition of the term Flandrian. *Boreas* 7: 182.
- Iversen J (1941) Landnam i Danmarks Stenalder (land occupation in Denmark's Stone Age). *Danmarks Geologiske Undersøgelse II, Række* 66: 1–68.
- Jacobson GL and Bradshaw RHW (1981) The selection of sites for paleovegetational studies. *Quaternary Research* 16: 80–96.
- Kaplan JO, Krumhardt KM, and Zimmerman N (2009) The prehistoric and preindustrial deforestation of Europe. *Quaternary Science Reviews* 28: 3016–3034.
- Manchester SJ and Bullock JM (2000) The impacts of non-native species on UK biodiversity and the effectiveness of control. *Journal of Applied Ecology* 37: 845–864.
- Parker AG, Goudie AS, Anderson DE, Robinson MA, and Bonsall C (2002) A review of the mid-Holocene elm decline in the British Isles. *Progress in Physical Geography* 26(1): 1–45.
- Pegler SM and Birks HJB (1993) The mid-Holocene *Ulmus* fall at Diss Mere, south-east England – Disease and human impact? *Vegetation History and Archaeobotany* 2: 61–68.
- Prentice IC (1983) Pollen mapping of regional vegetation patterns in south and central Sweden. *Journal of Biogeography* 10: 441–454.
- Rackham O (2000) *The History of the Countryside*, 2nd edn. London: Phoenix Press 448 pp.
- Stewart JR and Lister AM (2001) Cryptic northern refugia and the origins of the modern biota. *Trends in Ecology and Evolution* 16: 608–613.
- Svenning JC (2002) A review of natural vegetation openness in north-western Europe. *Biological Conservation* 104: 133–148.
- Vera FWM (2000) *Grazing Ecology and Forest History*. Wallingford: CABI Publishing.
- Von Post L (1916) Forest tree pollen in south Swedish peat bog deposits (translated into English by Davis MB and Faegri K, 1967). *Pollen et Spores* 9: 375–401.
- Whitehouse NJ and Smith DN (2004) 'Islands' in Holocene forests: Implications for forest openness, landscape clearance and 'culture-steppe' species. *Environmental Archaeology* 9: 203–212.

Relevant Websites

- <http://www.europeanpollendatabase.net/> – European Pollen Database.
- <http://www.ncdc.noaa.gov/paleo/pollen.html> – NOAA.
- <http://www.stratigraphy.org/column.php?id=Chart/Time%20Scale> – International Stratigraphic Chart.

Southern Europe

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The Current State of Knowledge of Postglacial Pollen Records of Southern Europe

Pollen analysis has proved to be a powerful tool not only to demonstrate changes in plant communities through time but also to investigate changes in past climate and human exploitation of the landscape. The southernmost regions of Europe, together with the Carpathian region, have been found to be important biodiversity reservoirs for the whole continent during glacial intervals. Southern Europe extends south of the 45° parallel. From a biogeographic standpoint, it consists mainly of Mediterranean countries and the islands of the Mediterranean Sea. For the sake of brevity, this article discusses only the most recent papers in which detailed references of previous regional works are provided. A list of sites is provided in [Tables 1–4](#), grouped according to their geographic distribution, and the list of sites used for this synthesis is shown in [Figure 1](#). (In [Figure 1](#) and [Tables 1 to 4](#) a reference to most southern European pollen sites (place, author(s), publication year) is made. A complete list of references can be found in regional syntheses and in recent papers quoted in the text, as it was not possible to include all references.) Dates in this discussion are in calendrical-equivalent (calibrated) years before present (see [Calibration of the ¹⁴C Record](#)). The classical pollen chronozones used for northern and central Europe are deliberately not used in this article because this model does not generally fit the dynamics and timing of changes in the vegetation of southern Europe. Knowledge of the Late Glacial (see [Middle and Late Pleistocene in Southern Europe](#)) vegetation trends is also needed for a better understanding of postglacial vegetation dynamics in southern Europe.

The postglacial reforestation of the various southern European regions shows differences not only in timing but also in vegetation types and dynamics. Many postglacial pollen records of southern Europe show substantial differences between early and late Holocene vegetation development, suggesting a general change from wetter to drier climatic conditions (e.g., [Sadori and Narcisi, 2001](#)). These changes have often been interpreted as reflecting aridification phases leading to the establishment of the present-day Mediterranean climate ([Carrión et al., 2010b](#); [Jalut et al., 2000](#); [Pérez-Obiol et al., 2011](#); [Sadori et al., 2011](#)). However, according to an alternative interpretation, the opening of the forests since the mid-Holocene was primarily linked to human impact, increasing from the early Neolithic onwards ([Kaplan et al., 2009](#)). Palynological research on postglacial environmental changes has, therefore, led to divergent interpretations in terms of natural, as opposed to anthropogenic, factors. One way to shed greater light on the causes of mid-Holocene landscape changes is to use those natural archive records that can be unambiguously attributed to climatic forcing. In many parts of southern Europe, lake level, lake isotope, cave speleothem, and deep-sea sedimentary records (e.g., [Bar-Matthews and Ayalon, 2011](#); [Magny and Miramont Sivan, 2002](#); [Roberts et al., 2011](#)) independently suggest a distinction

between an earlier part of the Holocene characterized by conditions moister than present and favorable to temperate deciduous trees, and a later interval with a drier climate, particularly in summer, favorable to sclerophyllous trees.

Iberian Peninsula and Balearic Islands

Recently, [Carrión et al. \(2010a\)](#) published an updated synthesis of Late Glacial and postglacial records of this region. They succeeded not only in documenting but also in explaining, with selected and meaningful examples, the complex late Quaternary vegetation history of the Iberian Peninsula and Balearic Islands. [Carrión et al. \(2010a\)](#) include a complete list of the currently available regional, terrestrial, and marine pollen records as a supplementary file (see also [Table 1](#)).

At the beginning of the postglacial in Portugal and continental Spain, an increase in moisture and temperature is suggested by the development of deciduous broadleaf forests at low elevations close to the coast and mainly pine (*Pinus*) forest inland and in most mountain regions. Evidence for Mediterranean vegetation with evergreen oaks, probably both kermes oak (*Quercus coccifera*) and holm oak (*Q. ilex*), mixed with steppe and grassland plants, was found in pollen spectra that predate the Holocene in southeastern Spain from the subarid region of Almería. This early reforestation suggests that mesophilous and thermophilous tree refugia were present in southeastern Spain in the late Pleistocene. Evergreen vegetation dominated by the *Quercus ilex* type, with deciduous oak, sparse cork oak (*Q. suber*) and considerable amounts of pine, probably reflecting altitudinal changes of vegetation in the Sierra Nevada mountains, was found in southeastern Spain at Padul during the Late Glacial, since at least 13 ka BP. A similar age for this early reforestation, in which *Quercus* (both evergreen and deciduous) and other angiosperm taxa played an important role, is also found in the lower Guadiana valley, Portugal ([Fletcher et al., 2007](#)). Increases in *Quercus* pollen since the Late Glacial interval show the earliest occurrences in most pollen sites of southernmost and coastal regions of Iberia. After the Late Glacial reforestation, the marked recurrence of glacial conditions typical of northern countries and preceding the postglacial is found in the mountain area of Lago de Sanabria in northeastern Spain. The expansion of deciduous oaks occurred rather late, after 11 ka BP, at Estany de Banyoles in Cataluña.

For the Late Glacial/postglacial transition in Portugal, a relatively humid climate can be deduced from the early paludification of Lagoa do Golfo where deciduous, riparian, and coniferous forests dominated, replacing sclerophyllous maquis shrubland vegetation dominated by evergreen shrubs. Along the coast, at Orla and in the basins of the Tejo and Sado rivers and probably along the whole coastline between Porto and Sines, considerable areas of maritime pine (*Pinus pinaster*) forests were recorded in areas with sandy, unconsolidated soils.

Table 1 Iberian peninsula and Balearic Islands. List of the main Postglacial records.

<i>Iberian Peninsula and Balearic Islands</i>	
<i>Site(s)</i>	<i>Author(s)</i>
Albufera de Alcudia	Burjachs et al., 1994
Algendar, Cala'n Porter, Son Bou, Hort Timoner	Yll et al., 1997
Antas, San Rafael, Roquetas de Mar	Pantaléon-Cano, 1998; Pantaléon-Cano et al., 2003
Banyoles	Pérez-Obiol and Julià, 1994
Beliche-Guadiana	Fletcher et al., 2007
Besós, Cubelles, Ullastret	Riera-Mora and Esteban-Amat, 1994
Burg	Pèlachs et al., 2009, 2011; Bal et al., 2011; Pérez-Obiol et al., 2011
Cala Galdana	Yll et al., 1994
Canal de Navarrés	Carrión and van Geel, 1999
Charco da Candieira	Van der Knaap and Van Leeuwen, 1995; 1997
Cañada de la Cruz	Carrión et al., 2001b
El Carrizal	Franco Múgica et al., 2005
Estany del Calgalell	Riera Mora, 1990
Gádor	Carrión et al., 2003
Hoyos de Iregua	Gil García et al., 2002
Lagoa Comprida	Janssen and Woldring, 1981
Lagoa de Albufeira	Queiroz, 1989
Lagoa de Lucenza, Fraga, As Aguiladas, Castelo Cerveira	Santos et al., 2000
Lagoa de Marinho	Ramil-Rego et al., 1998
Lagoa Santo André	Santos and Sánchez Goñi, 2003
Lagoa Travessa	Mateus, 1985, 1989
Lago de Ajo	Watts, 1986
Lago de Sanabria	Hannon, 1985; Muñoz-Sobrino et al., 2004; Turner and Hannon, 1988
Lago Las Pardillas	Sanchez Goñi and Hannon, 1999
Laguna Medina	Reed et al., 2001
Laguna de la Roya	Allen et al., 1996
Laguna de las Lamas	Ruiz, 1994
Laguna de las Sanguijuelas, Lleguna	Muñoz Sobrino et al., 2004
Laguna del Hormillo	Gómez-Lobo, 1993
Laguna Quintanar de la Sierra	Peñalba, 1994; Peñalba et al., 1997
La Mancha plain	Dorado Valiño et al., 2002
La Piedra	Muñoz-Sobrino et al., 1996
Les Palanques, Sidera	Pérez Obiol, 1988
Ojos del Tremendal	Stevenson, 2000
Portalet	González-Sampériz et al., 2005
Pozo do Carballal	Ramil Rego et al., 1998
Atxuri, Belate, Lago de Arreo, Los Tornos, Quintanar de la Sierra, Saldropo	Peñalba, 1989; Peñalba et al., 1997
Rascafría	Franco-Múgica et al., 1998
Salines	Burjachs et al., 1997
Siles	Carrión, 2002
Turbera de Padúl	Florschütz et al., 1971; Pons and Reille, 1988
Urtiaga	Sanchez Goñi, 1992
Villaverde	Carrión et al., 2001a

In the Iberian Peninsula, the postglacial vegetation appears even more diversified than the Late Glacial. In fact, clear differences of vegetation development and composition between the areas with continental and Mediterranean climates are

Table 2 Southern France and Corsica. List of the main Postglacial records.

<i>Southern France and Corse</i>	
<i>Site(s)</i>	<i>Author(s)</i>
Bocca Antilio, Prato di Caldane, Oriente, Nino, Vaccaja, Marmano, Les Pozzi, Plateau d'Ese, Cavallara, La Camera, Piano di Renuccio, Punta Tozzarella, Ovace	Reille, 1975
Barcaggio, Bastani, Cannuta, Capitello, Crovani, Le Fango, Ostriconi, Saleccia	Reille, 1992
Belvis	Jalut et al., 1975
Biot	Nicol-Pichard and Dubard, 1998
Bious, Castet, Estarres	Jalut et al., 1988
Biscaye	Mardones and Jalut, 1983; Reille and Andrieu, 1995
Capestang	Jalut, 1995; Jalut et al., 2009
Étang de Mauguio	Planchais, 1982
Fos-sur-Mer	Triat, 1975
Freychinède	Jalut et al., 1982; Reille, 1993
Lac du Creno	Reille, 1992; Reille et al., 1997
Lagune de Canet	Planchais, 1985
La Pouretère	Aubert et al., 2004
Lourdes, Le Monge	Reille and Andrieu, 1995
Marais des Baux	Andrieu-Ponel et al., 2000
Marsillargues	Planchais and Duzer, 1978; Planchais, 1982
Palavas	Planchais, 1987
Perafita	Miras et al., 2010
Pinarello, Étangs de Terrenzana, del Sale, de Palo, de Palombaggia	Reille, 1984
Tourves	Nicol-Pichard, 1987

found. The scientific debate about the role of climatic change has been mostly restricted to vegetation dominated by oak and pine. Many sites show high values of pine persisting since the early Holocene, suggesting that mountain pine forests partly covered the inner and Atlantic regions of the Iberian Peninsula. However, during intervals with open woodlands, the effects of over-representation of pine pollen in percentage diagrams should be taken into account. Modern pollen-vegetation relationship studies carried out in northwestern Spain (the Las Pardillas region) provide evidence for the local presence of pine at the beginning of postglacial reforestation (Sánchez Goñi and Hannon, 1999). As suggested by Carrión et al. (2010a), different causes, such as continentality, altitude, and aridity, may have favored pines as opposed to oaks. When analyzing this evidence, the fact that several pine species with quite different ecological features are often not distinguished in the pollen diagrams, has to be taken into account. Thus, the existence of pine woodlands during the Late Glacial and early postglacial can be considered a consequence of the orographic influence in several of the Iberian mountain pollen records. Carrión et al. (2010a), using both palynological and macrofossil data, proposed the following distribution in the Iberian Peninsula: European black pine (*Pinus nigra*) and Scots pine (*P. sylvestris*), representing the inheritance of the glacial pine

Table 3 Peninsular Italy and Sicily. List of the main Postglacial records.

<i>Peninsular Italy and Sicily</i>	
<i>Site(s)</i>	<i>Author(s)</i>
Biviere di Gela	Noti <i>et al.</i> , 2009
Canolo Nuovo	Schneider, 1985; Gröger, 1977
Fociomboli	Braggio Morucchio <i>et al.</i> , 1980
Gorgo Basso	Tinner <i>et al.</i> , 2009
Lagaccione	Magri, 1999
Lagdei	Bertoldi, 1980
Lago Agoria di Fondo, Lago Casanova	Cruise, 1990a, 1990b
Lago Agoria di Mezzo	Braggio Morucchio and Guido, 1975
Lago Albano, Lago di Nemi	Lowe <i>et al.</i> , 1996; Mercuri <i>et al.</i> , 2002
Lago Alimini Piccolo	Magri, 1999; Magri and Di Rita, 2005; Di Rita and Magri, 2009
Lago Baccio, Lago del Greppo, Lago Nero	Mori Secci, 1996
Lago Battaglia	Caroli, 2005; Caroli and Caldara, 2006
Lago d'Averno	Gröger and Thulin, 1998; Gröger <i>et al.</i> , 2002
Lago dell'Accesa	Drescher Schneider <i>et al.</i> , 2006
Lago delle Fate, Laghi dell'Orgials	Ortu <i>et al.</i> 2005, 2006
Lago delle Lame, Lago Nero, Prato Mollo	Cruise, 1990a
Lago di Bargone	Cruise <i>et al.</i> , 1996
Lago di Martignano	Kelly and Huntley, 1991
Lago di Masacciucoli / Lago di Massacciucoli	Menzio <i>et al.</i> , 2001
Lago di Mezzano	Sadori <i>et al.</i> , 2004; Sadori <i>et al.</i> , 2011
Lago di Monterosi	Bonatti, 1966
Lago di Pergusa	Sadori and Narcisi, 2001
Lago di Ripa Sottile	Ricci Lucchi <i>et al.</i> , 2000
Lago di Vico	Frank, 1969; Magri and Sadori, 1999
Lago Grande di Monticchio	Watts <i>et al.</i> , 1996; Allen <i>et al.</i> , 2002
Lago Lagastro, Lago Riane, Lago Rotondo Rovegno	Branch, 2004
Lago Lungo	Calderoni <i>et al.</i> , 1994
Lago Padule, Prato Spilla	Lowe and Watson, 1993
Lajone	Braggio Morucchio <i>et al.</i> , 1978; Guido <i>et al.</i> , 2004
Naples port	Allevato <i>et al.</i> , 2010; Di Pasquale <i>et al.</i> , 2010
Piana del Fucino	Magri and Follieri, 1991
Pisa port, Arno river	Benvenuti <i>et al.</i> , 2006; Mariotti Lippi <i>et al.</i> , 2006
Pisa, Rapallo, Sestri, Versilia coastal plains	Bellini <i>et al.</i> , 2009
Rome port, Tiber delta	Bellotti <i>et al.</i> , 2007; Sadori <i>et al.</i> , 2010
Stagno di Maccarese	Di Rita <i>et al.</i> , 2010
Stracciaccia	Giardini, 2006
Valle di Baccano	Ciuffarella, 1996
Valle di Castiglione	Follieri <i>et al.</i> , 1988; Alessio <i>et al.</i> , 1986
Urgo di Pietra Giordano	Bertolani Marchetti <i>et al.</i> , 1984

Table 4 Southern Balkan Peninsula, Dalmatian Islands and Crete. List of the main Postglacial records.

<i>Southern Balkan Peninsula, Dalmatian Islands and Crete</i>	
<i>Site(s)</i>	<i>Author(s)</i>
Agia Gallini	Bottema, 1980
Begbunar	Lazarova <i>et al.</i> , 2009
Bokanjčko Blato	Gröger, 1996
Delphinos, Kournas	Bottema and Sarpaki, 2003
Edessa, Giannitsa, Khimaditis	Bottema, 1974; Bottema and Woldring, 1990
Goce Delchev, Visokata Ela	Stefanova, 1997
Gramousti, Rezina	Willis, 1992a and 1992b
Ioannina	Bottema, 1974; Tzedakis, 1993; Lawson <i>et al.</i> , 2004
Kopais	Allen 1986, 1990; Tzedakis, 1999
Kupena mire	Božilova <i>et al.</i> , 1989; Huttunen <i>et al.</i> , 1992
Lailias, Paiko, Voras, Pieiras	Gerasimidis, 1985; Gerasimidis and Athanasiadis, 1995
Lake Arkutino	Božilova and Beug, 1992
Lake Bolata	Tonkov <i>et al.</i> , 2011
Lake Dalgoto	Stefanova <i>et al.</i> , 2003; Stefanova and Ammann, 2003
Lake Doirani	Athanasiadis <i>et al.</i> , 2000
Lake Durankulak	Božilova and Tonkov, 1985, 1998; Božilova and Filipova, 1986
Lake Ohrid	Lézine <i>et al.</i> , 2010; Wagner <i>et al.</i> , 2009
Lake Ribno	Tonkov <i>et al.</i> , 2002
Banderishko	
Lake Sedmo Rilsko	Božilova and Tonkov, 2000
Lake Shabla-Ezeretz	Filipova, 1985; Božilova and Filipova, 1986
Lake Shkodra	Sadori <i>et al.</i> , in preparation
Lake Srebarna	Lazarova and Božilova, 2001
Lake Trilistnika	Tonkov <i>et al.</i> , 2009
Lake Varna, lake Beloslav	Božilova and Beug, 1994
Lake Voulkaria	Jahns, 2005
Lake Vrana	Schmidt <i>et al.</i> , 2000
Lerna	Jahns, 1993
Maliq	Denfle <i>et al.</i> , 2000; Fouache <i>et al.</i> , 2001; Bordon <i>et al.</i> , 2009; Fouache <i>et al.</i> , 2010
Malo Jezero	Beug, 1967, Jahns and van der Bogaard, 1998; Jahns, 2002; Sadori <i>et al.</i> , 2011
Matnitsa	Filipovitch, 1985
Mire Straldza	Tonkov <i>et al.</i> 2009
Moor Praso	Stefanova and Oeggli, 1993
Rhodopes	Athanasiadis <i>et al.</i> , 1993; Gerasimidis and Athanasiadis, 1995
Suho Ezero	Božilova and Smith, 1979; Božilova <i>et al.</i> , 1990
Tenaghi Philippon	Wijmstra, 1969
Tschokljovo	Tonkov and Božilova, 1992
Veliko Jezero	Jahns and van der Bogaard, 1998; Jahns, 2002
Vidl	Brande, 1973
Voras	Athanasiadis and Gerasimidis, 1986; Gerasimidis and Athanasiadis, 1995
Xinias	Bottema, 1979

woodlands, were probably abundant in the mountains, whereas stone pine (*P. pinea*), occurring since the glacial, was limited to the warmer southern and southwestern regions. Aleppo pine (*P. halepensis*) was widespread in the east and Ebro Valley and favored by the opening of the environment and the onset of

matorral (shrubland and woodland ecoregion of the Mediterranean). Mountain pine (*P. uncinata*) and Scots pine would have formed the timberline in the Eurosiberian thermic (i.e., the boundary between the Mediterranean and Eurosiberian vegetation regions in the northern Iberian Peninsula – see Moreno

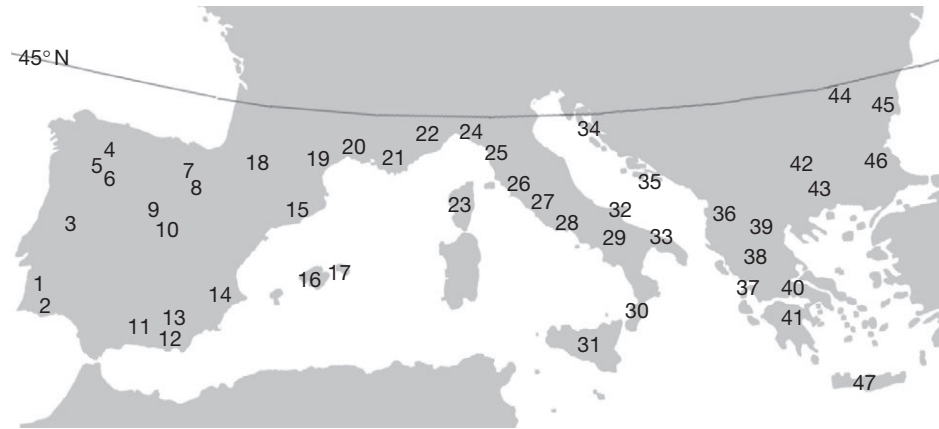


Figure 1 List of sites used for this synthesis. 1. Lagoa Travessa – Mateus, 1989; Lagoa de Albufeira – Queiroz, 1989. 2. Beliche-Guadiana – Fletcher et al., 2007. 3. Serra de Estrela: Charco da Candieira – Van der Knaap and Van Leeuwen, 1995, 1997; Lagoa Comprida – Janssen and Woldring, 1981. 4. Lago de Ajo – Watts, 1986. 5. Lagoa de Lucenza, Fraga, As Aguilladas, Castelo Cerveira – Santos et al., 2000. 6. Laguna de la Roya – Allen et al., 1996; Laguna de las Sanguijuelas, Lleguna – Muñoz-Sobrino et al., 2004; Lago de Sanabria – Turner and Hannon, 1988. 7. Las Pardillas – Sánchez-Goñi and Hannon, 1999. 8. Quintanar de La Sierra – Peñalba, 1989; Peñalba et al., 1997. 9. El Carrizal – Franco Múgica et al., 2005. 10. Rascafría – Franco Múgica et al., 1998. 11. Turbera de Padúl – Florschütz et al., 1971; Pons and Reille, 1988. 12. Almería: Antas, San Rafael, Roquetas de Mar – Pantaléon-Cano, 1998; Pantaléon-Cano et al., 2003; Gádor – Carrión et al., 2003. 13. Sierra de Segura: Siles – Carrión, 1998; Villaverde – Carrión et al., 2001a; Cañada de la Cruz – Carrión et al., 2001b. 14. Canal de Navarres – Carrión and van Geel, 1999. 15. Banyoles – Pérez-Obiol and Julià, 1994. 16. Majorca: Cala Galdana – Yll et al., 1997; Albufera de Alcudia – Burjachs et al., 1994. 17. Minorca: Algendar, Cala'n Porter, Son Bou, Hort Timoner – Yll et al., 1997. 18. Pyrenees: Bious, Castet, Estarres – Jalut et al., 1988; Biscaye – Mardones and Jalut, 1983; Reille and Andrieu, 1995; Lourdes, Le Monge – Reille and Andrieu, 1995; Belvis – Jalut et al., 1975; La Pourrière – [Citation missing here?]. 19. Capetang – Jalut, 1995; Jalut et al., 2009. 20. Palavas – Planchais, 1987; Marais des Baux – Andrieu-Ponel et al., 2000. 21. Biot – Nicol-Pichard and Dubard, 1998. 22. Lago delle Fate, Laghi dell'Orgials – Ortu et al., 2005, 2006. 23. Corsica: 27 pollen sites (see Table 2) Reille, 1975; Reille, 1992; Reille et al., 1997. 24. Ligurian Apennines: Lago Agoriaie di Fondo, Lago Casanova – Cruise, 1990a, 1990b; Lago Agoriaie di Mezzo – Braggio Morucchio and Guido, 1975. Tosco-Emilian Apennines: Lagdei – Bertoldi, 1980; Lago Padule, Prato Spilla – Lowe and Watson, 1994. 25. Pisa, Rapallo, Sestri, Versilia coastal plains – Bellini et al., 2009. Pisa port, Arno river – Benvenuti et al., 2006; Mariotti – Lippi et al., 2006. 26. Lago dell'Accesa – Drescher-Schneider et al., 2006. 27. Lagaccione – Magri, 1999; Lago di Mezzano – Sadori et al., 2004; Lago di Vico – Magri and Sadori, 1999; Lago di Martignano – Kelly and Huntley, 1991; Stagno di Maccarese – Di Rita et al., 2010; Rome Port – Bellotti et al., 2007; Sadori et al., 2010; Valle di Castiglione – Follieri et al., 1988; Lago Albano, Lago di Nemi – Lowe et al., 1996; Mercuri et al., 2002. 28. Lago d'Averno – Gröger and Thulin, 1998; Gröger et al., 2002; Naples Port – Di Pasquale et al., 2010. 29. Lago Grande di Monticchio – Watts et al., 1996; Allen et al., 2002. 30. Canolo Nuovo – Schneider, 1985; Gröger, 1977. 31. Lago di Pergusa – Sadori and Narcisi, 2001; Gorgo Basso – Tinner et al., 2009; Biviere di Gela – Noti et al., 2009. 32. Lago Battaglia – Caroli, 2005; Caroli and Caldara, 2006. 33. Lago Alimini Piccolo – Magri, 1999; Magri and Di Rita, 2005; Di Rita and Magri, 2009. 34. Cres Island: Lake Vrana – Schmidt et al., 2000. Vidl – Brande, 1973. 35. Mijet Island: Malo Jezero – Beug, 1967; Jahns and van der Bogaard, 1998; Jahns, 2002; Sadori et al., 2011; Veliko Jezero – Jahns and van der Bogaard, 1998; Jahns, 2002. 36. Lake Shkodra – Sadori et al., in preparation; Zanchetta et al., submitted; Maliq – Denèfle et al., 2000; Fouache et al., 2001; Bordon et al., 2009; Fouache et al., 2010. Lake Ohrid – Wagnr et al., 2009;. 37. Lake Voulkaria – Jahns, 2005; Sadori et al., 2011. 38. Gramousti and Rezina – Willis, 1992a and 1992b; Ioannina – Bottema, 1974; Tzedakis, 1993; Lawson et al., 2004. 39. Edessa, Giannitsa, Khimaditis – Bottema, 1974; Bottema and Woldring, 1990. 40. Xinias – Bottema, 1979; Kopais – Allen 1986, 1990; Tzedakis, 1999. 41. Lerna – Jahns, 1993. 42. Southwestern mountain region of Bulgaria. Rila mountains: Lake Suho Ezero – Božilova et al., 1990; Lake Sedmo Rilsko – Božilova and Tonkov, 2000; Lake Trilistnika – Tonkov et al., 2008. Rhodope mountains: Kupena mire – Božilova et al., 1989; Huttunen et al., 1992. Pirin mountains: Lake Ribno Banderishko – Tonkov et al., 2002; Mozgovitsa – Tonkov, 2003; Moor Praso – Stefanova and Oegg, 1993; Goce Delchev, Visokata Ela – Stefanova, 1997. 43. Tenaghi Philippon; Wijmstra, 1969. 44. Lake Srebarna – Lazarova and Božilova, 2001. 45. Lake Varna – Božilova and Beug, 1994; Lake Durankulak – Božilova and Tonkov, 1985, 1998, Božilova and Filipova, 1986; Lake Shabla-Ezeretz – Filipova, 1985, Božilova and Filipova, 1986. 46. Lake Arkutino – Božilova and Beug, 1992. 47. Delphinos, Kournas – Bottema and Sarpaki, 2003.

et al., 1990), moving to higher elevations during the early Holocene.

However, around 11.5 ka BP, in many mountain Atlantic sites of the Iberian Peninsula (Cantabrian Cordillera and Serra de Estrela), birch (*Betula*) played an important role, preceding and accompanying the expansion of deciduous oaks from around 10.8 ka BP. The reduced amount of steppe vegetation (herbs are mainly represented by grasses) suggests that oceanic conditions were already prevailing in this area by the early Holocene (Carrión et al. 2010a). At the time of the North Atlantic cooling event (ca. 8.2 ka BP, e.g. Rohling and Pälike, 2005), a decrease of pine and an increase of evergreen and deciduous angiosperm trees is recorded from mountain sites in

central and southeastern Spain (Rascafría, Villaverde). A decrease of steppe and an increase of deciduous oaks and thermophilous maquis took place in Almería, indicating the Holocene peak in available moisture. An expansion of fir (*Abies*) and beech (*Fagus*) is recorded in northern Catalonia, while the climate signal is unexpectedly inconsistent either in the Balearic Islands or in the northwestern Iberian regions under Atlantic influence (Muñoz-Sobrino et al., 2004; Yll et al., 1997). In the northwestern Iberian Peninsula, a climate change towards cooler conditions may have brought about a decrease in deciduous oak and an increase in birch. This cooling is radiocarbon dated at around 9.2 cal ka BP, about one millennium before the North Atlantic cooling event.

In southwestern Portugal (Lagoa Travessa), a decrease in pine occurred soon after 7.4 ka BP, and it was replaced mainly by evergreen vegetation after 5.5 ka BP. During the entire mid-Holocene, in some Iberian continental and mountain zones, pine forests seem to have persisted where conifers were not replaced by broadleaf trees as recorded in other areas of southwestern Europe. The substitution of pine with evergreen oak at Navarrés and with deciduous oak at Villaverde occurred at 6.8 ka BP, chronologically matching an apparent change toward wetter (and cooler?) conditions found at Banyoles (decreasing deciduous oak, increasing fir), and at Lago de Ajo (decreasing pine and birch, increasing hazel (*Corylus*)). In contrast, the sudden and sharp decrease in box (*Buxus*) in the mid-Holocene, which led to its local extinction at Minorca in about one millennium, can be interpreted as a sign of increased xeric conditions (Yll et al., 1997) and has been recorded in all the Balearic Islands.

Could the change detected in many peninsular sites be only due to improved edaphic (soil) conditions, which allowed some mesophilous and thermophilous taxa to spread? Although the mid-Holocene was an interval with very low fire incidence and minima in xerophilous taxa, optimal conditions for mesophilous plants were hardly attained in the Mediterranean part of the peninsula until 5.8 ka BP. According to de Beaulieu et al. (2005), the trend toward aridification found in the Balearic Islands also became evident in the peninsula by about 6.5 ka BP. This was due to significant changes in the Mediterranean atmospheric forcing, which progressed from 6.5 to 6 ka BP with the establishment of a drier climate and an increase in seasonality. A different interpretation is advanced and well substantiated for southeastern Spain by Carrión et al. (2010a). In the mountainous areas, in fact, in the period from 7.5 to 5 ka BP, several species of pines, evergreen, deciduous oaks, and other broadleaf trees were widespread. This represented the mesophytic optimum, the xerophytic minimum, and the period of lowest fire activity.

Pérez-Obiol et al. (2011) offer a useful summary of the mid-Holocene climate changes registered in palynological records of southwestern Spain. They propose a rough temporal framing of Holocene climatic changes as follows: a humid phase (12–7 ka BP), a transition phase (7–5.5 ka BP), and an aridification phase (5.5 ka BP–). The timing, extent, and progress of the establishment of the Mediterranean climate had a degree of variability depending on the biogeographic region considered. Close examination of several pollen sequences carried out by Pérez-Obiol et al. (2011) revealed climatic transformations in the flora and vegetation between 7 and 4 ka BP. Three main spatial and taxonomic responses have been identified by them: (1) in littoral regions, deciduous broadleaf trees were frequently dominant and then replaced by sclerophyllous and evergreen forests; (2) in continental regions and sub-Mediterranean mountains, the dominance of pine throughout the whole Holocene signals a change of less magnitude; and (3) in southeastern semiarid Mediterranean regions, the main changes are reflected by alternation between steppe and shrub communities.

The varied responses of the diverse Mediterranean vegetation ecosystems to the gradual mid-Holocene climatic change probably accounts for the time-transgressive nature of changes registered in regional pollen spectra. Many authors, however, attributed these changes to human impact, leading to the

dominance of evergreen sclerophyllous vegetation in the Mediterranean in recent millennia.

Jalut et al. (2000) identified another episode of increased xerophytes and almost complete disappearance of mesophilous trees in the peninsula around 5.1–4.7 ka BP. The 400-year interval is an artifact of imprecise dating. However, the pollen records show sudden change. This climate change is likely to have contributed to the extinction of box in the Balearic Islands.

Recently, Carrión and colleagues (Carrión et al., 2010b) carried out an analysis of the aridification trends and processes that have occurred in the Murcian-Almerian bioprovince, southeast Spain, during the Holocene. The area has experienced a trend toward desertification since at least the Miocene. The human impact is highly variable in this region, and this complicates the distinction between natural and anthropogenic causes. Environmental change was highly variable across southeastern Spain, starting earlier in low-elevation areas and river basins, after 5 ka BP. Forest degradation in the mountains started later (4.3–3.6 ka BP) and reached its maximum during the Roman occupation, around 2 ka BP. The high degree of present-day xerophytization (i.e., plant adaptation to dry conditions) has multiple causes, resulting from a combination of anthropogenic disturbance and resource exhaustion under enhanced aridity.

Pollen records from southern Portugal identify significant vegetation changes between 5 and 3 ka BP, consisting of the expansion of seminatural heathlands and shrublands at the expense of forest vegetation. A wet phase occurred in the mid-Holocene in northwestern and north-central Iberia with increasing hazel, Ericaceae (heath family), and beech, the last being recorded after 4.5 ka BP. Around 3.9 ka BP, the xerophytic component increased in the rest of the peninsula, along with anthropogenic indicators and fires.

Strong evidence of cereal cultivation occurred in some sites soon after 3.2 ka but mainly since Roman times or later, first in lowland sites and then in sites at higher altitudes. The beginning of the extensive cultivation of olive (*Olea*) and grape (*Vitis*) seems to date from the Middle Ages. An arid period preceded and accompanied the Romanization of Iberia (2.7–1.9 ka BP, Jalut et al., 2000), so the effects of human landscape modification must be taken into account from this time onwards.

Southern France, Andorra, and Corsica

The Holocene vegetation history and climate change in Mediterranean France and in the northern slopes of the Pyrenees has been partially summarized in recent papers (Andrieu-Ponel et al., 2000; Jalut et al., 2000, 2009), while the postglacial palynological record of Corsica is discussed in Reille et al. (1997) (see also Tables 2 and 5). Most studies in the coastal regions of Languedoc, Camargue, and Provence (Figure 1) were carried out in flood plains and marshy areas, where pollen was often scarce and poorly preserved, pollen from local vegetation masked the regional pollen rain, and problems with radiocarbon dates often arose (e.g., Andrieu-Ponel et al., 2000). In this regard, a new radiocarbon-dated diagram from Capestang, Languedoc (Jalut et al., 2009), is particularly interesting.

Table 5 Timing and features of the main postglacial vegetational events in southern Europe

Event	Age	Features
<i>Iberian Peninsula and Balearic Islands</i>		
Spread of Postglacial woodlands	Since 10.2 ka BP onward	Spread of deciduous oak forests in coastal and southern mountain sites, of pine and/or birch in most of the northern mountain and continental sites
Start of an aridification phase	7.5–7 ka BP	A slow and prolonged climate forcing does probably account for its time-transgressive character
Climate change toward cooler and/or wetter conditions?	Since 6 ka BP	Found mainly at mountain sites, it could also be due to improved edaphic conditions which allowed the spread of more demanding water taxa
Strong climate change with increase in aridity and seasonality	6–4.5 ka BP	First detected in Balearic Islands (reduction and extinction of <i>Buxus</i> at Minorca), then in the rest of the Mediterranean part of the peninsula
Human presence	3.5–3 ka BP	Forest opening through cutting, burning, and grazing (possible synergy of human and climate forcing) and begin of cultivation
Arid phase	2.6–1.9 ka BP	Detected through pollen ratios, it preluded and accompanied the impact of the Romans
Human impact	2 ka BP onward	Strong forest clearance and extensive land exploitation. Cultivation of <i>Juglans</i> since ca. 2 ka BP
<i>Southern France and Corsica</i>		
Spread of Postglacial woodlands	Since around 1 ka BP	Conifer trees on the Pyrenees mountains, thermophilous forests at middle and low elevations
Arid phase	7.500–7 ka BP	Detected through pollen ratios
Increase of aridity	6 ka BP	Expansion of Mediterranean taxa. It was interpreted as the effect of human impact, but at least a synergy of climate and humans should be taken into account
Climate change with increase in aridity and seasonality and human presence	4.500–4 ka BP	Human populations probably just strengthened a climate forcing which led to a well-developed Mediterranean vegetation. Evidence of grazing from the Pyrenees
Human impact	Since around 2.6 ka BP	Cultivation both in plain and in mountain regions. Cultivation of <i>Juglans</i> since ca. 2.2 ka BP
<i>Peninsular Italy and Sicily</i>		
Spread of Postglacial woodlands	Since around 10.500 ka BP	Spread of thermophilous forests (deciduous <i>Quercus</i> dominant) in central and southern Italy and with <i>Abies</i> in northern Italy
Beginning of a climate trend toward aridity	(8) 7.5 ka BP	Beginning of <i>Q. ilex</i> continuous curve in central Italy (8 ka BP) and increase (7.2 ka BP) of <i>Olea</i> in Sicily
Arid climate change	ca. 3.7. ka BP	Forest clearance without fires and changes in wood composition, detected in Latium. Does it correspond to the northern Italy decrease of <i>Abies</i> (and partly <i>Quercus</i>) of 3.8 ka BP?
Human presence	3.6–3.5 ka BP	Human forest clearance, cultivation of cereals, legumes, olive (probable in Sicily), presence of secondary anthropic indicators and use of fire
Human impact	Since around 2.7 ka BP	Extensive cultivations, strong land exploitation. since around 2.7 ka BP <i>Juglans</i> in Sicily, <i>Castanea</i> widespread near Rome. Disappearance of <i>Abies</i> at Monticchio?
<i>Southern Balkan Peninsula, Dalmatian Islands, and Crete</i>		
Spread of Postglacial woodlands	Since around 10.600 ka BP	Spread of thermophilous forests (deciduous <i>Quercus</i> dominant) and conifer woods at some mountain sites
Arid phase at low-elevation sites?	9–8 ka BP	Maxima of <i>Pistacia</i> pollen interpreted as an indication of open environments
Climate trend toward wetness	7.5–6 ka BP	Reduction of deciduous <i>Quercus</i> and increase of one or more of the following taxa: <i>Corylus</i> , <i>Carpinus orientalis/Ostrya</i> , <i>Ulmus</i> , <i>Tilia</i> , <i>Abies</i> , and <i>Pinus</i> in upland sites of northern Greece and western Bulgaria
Arid climate event	7.2 ka BP	Reduction of deciduous <i>Quercus</i> , increase of <i>Pistacia</i> and <i>Juniperus</i> detected in Mljet Island
Human presence	5–4.5 ka BP	Probable man-induced forest clearance, increasing pasture, and secondary anthropic indicators
Human impact	Since around 4–3.5 ka BP	Probable extensive cultivations of <i>Olea</i> since the Bronze Age; <i>Juglans</i> (with <i>Ceratonia</i> in Crete) since around 3.3 ka BP

In the mountains of Corsica, Late Glacial pollen assemblages included alder (*Alnus*), birch, deciduous oak, juniper (*Juniperus*), and low amounts of pine pollen (<20%). Such a low percentage of pine pollen suggests its absence in the island from 13.9 to 11.9 ka BP, but pine dominated forests developed in the following centuries (Reille et al., 1997). At high elevations in Corsica, two

types of mesophilous forests were dominant from 8.9 to 2.7 ka BP: deciduous oak forest and mixed forest with yew (*Taxus*). At lower elevations, tree heath (*Erica arborea*) spread, while *Quercus ilex* type did not play a significant role until 5.1 ka BP.

At the onset of the Late Glacial reforestation, open landscapes with conifer trees and shrubs were common both in the

Pyrenees and on the plains of southern France. The Pyrenees acted as a Late Glacial refugium for mesophilous and thermophilous trees. Since ca. 12 ka BP, temperate forests, including mainly deciduous oaks with birch, hazel, and hornbeam, dominated the low and medium elevations in southern France and also along the Mediterranean coast. After 9.5 ka BP, fir pollen increased in plains records (Biot, Rhone valley, and Palavas), reflecting its expansion in the southern French Alps, in the northern Rhone valley, and in the southern Massif Central. Until recently, the slight increases of *Quercus ilex* and other sclerophyllous taxa from ca. 8 to 7 ka BP were interpreted as exclusively anthropogenic. It was thought that Neolithic populations heavily utilized this tree. Actually, beech and fir also increased during this interval together with Mediterranean taxa, thus leading to a different climatic reconstruction.

The new palynological record (Jalut et al., 2009) from the former pond of Capeatang (Languedoc) spans the entire Holocene, providing information for the last 12 000 years. After a phase dominated by pine, a deciduous forest is established about 10.2 ka BP. Hazel is more abundant than deciduous oak until 8.5 ka BP. Beech shows an expansion together with *Quercus ilex* type from 4.5 to 3 ka BP. The contemporary presence of such different vegetation types, a beech forest and a Mediterranean forest, can be explained by their differing altitudinal distributions, as the latter type of vegetation is the more thermophilous. Around 3–2.8 ka BP, *Fagus* and deciduous *Quercus* decrease and the forest becomes more open, with increased *Pinus*, grasses, and evergreen thermophilous or Mediterranean taxa.

Jalut et al. (2000) characterized climate changes along the Mediterranean coast of Spain and southwestern France, identifying an arid phase from 8.3 to 7.8 ka BP, followed by four other similar changes, the last of which was recorded in the Middle Ages. Human impact probably strengthened the impact of climate forcing, leading to well-developed Mediterranean vegetation during the late Holocene. Could migration delay and establishment time (e.g., edaphic requirements) have influenced the timing and selection of tree establishment? If so, then climatic forcing could be ruled out as the cause of the spread of beech and fir. On the basis of malacofauna (mollusk) analyses, de Beaulieu et al. (2005) concluded that during the Holocene, the Mediterranean ecosystems were patchy and frequently open. Such a structure may partly explain the ability of mountain taxa such as fir and beech to colonize some empty niches in the Mediterranean hills during their major expansion phases. In addition, long distance transport of exotic pollen cannot be excluded.

The recent study of the Planells de Perafità fen (Andorra) led Miras et al. (2010) to hypothesize a quite early human impact, dating back at ca. 8.4–8.1 ka BP. Because of uncertainties in grass pollen identification, the claimed cultivation of cereals in ca. 5.1 ka BP in Languedoc is not certain. Around 5.1–4.5 ka BP, at medium and high altitudes in the Pyrenees, pasture maintenance was a common practice (Jalut et al., 2000). Clear signs of human landscape modification are found in the Pyrenees since ca. 2.7 ka BP and in the plain close to Montpellier (Capeatang) since ca. 2.6 ka BP. A similar date marks the start of presumed forest clearance in the vegetation of Corsica. There, Jalut et al. (2000) detected an arid period, beginning around 2.7 ka BP. As in Spain, so also in France, sparse walnut (*Juglans*) pollen grains are found in

pollen spectra from ca. 7.8 ka BP, while the cultivation of this tree began around 2.2 ka BP. The exploitation of olive and chestnut (*Castanea*) in southern France dates back to the Middle Ages.

Peninsular Italy and Sicily

From the sector of northern Italy south of 45°, information on past vegetation comes from a few Alpine and some north Apennine mountain lakes. Most Holocene pollen records of central and southern Italy originate from extant or drained crater lakes, while the only insular record spanning the whole postglacial is from the center of Sicily (see Tables 3 and 5). An overview of the palynological research in Italy, including both Pleistocene and Holocene records, was recently published by Magri (2007) and is recommended for a complete reference list.

Throughout Italy, Late Glacial pollen spectra are characterized by significant levels of mesophilous tree pollen, especially deciduous oaks, birch, hazel, linden (*Tilia*), and elm (*Ulmus*). The only exceptions are the southernmost alpine lacustrine records (Laghi dell'Orgials and Lago delle Fate, Maritime Alps). The pollen diagrams from the Maritime Alps show the typical features of central European countries and the classical pollen chronozones, with pines dominant both in the Late Glacial and in the postglacial and broadleaf trees almost absent. In central Sicily and in most central Italian sites, sporadic evidence of thermophilous trees and shrubs occurs after the end of the glacial. Refugia areas for most trees were not far from the investigated sites. The expansion of forests was temporarily interrupted during the Younger Dryas interval but actually started at most sites by 12.5 ka BP. Closed forest conditions developed at different times in central and southern Italy ranging from 11.9 ka BP in the north to 10.2 ka BP in the south. At Gorgo Basso (Tinner et al., 2009), a coastal lake of Sicily, true forest conditions were never reached, but a *Pistacia* shrubland became established around 10 ka BP.

Postglacial pollen spectra are characterized by high, though fluctuating, percentages of deciduous oaks from the mountain areas of Sicily (together with evergreen oaks at Lago di Pergusa) to northern Italy (northern Apennines). Deciduous oak forests are generally dominant in the plains and hilly sites of central Italy. Throughout the postglacial in Latium (e.g., Lagaccione, Lago di Vico, Valle di Castiglione, Lago Albano, Lago Lungo), they are accompanied by beech (*Fagus*) during the first part of the postglacial. Other main taxa in the first half of the postglacial are described in the following paragraphs.

Fir (*Abies*) was very important in the western sector of the peninsula and was found from the early postglacial in the northern Apennines (Lowe and Watson, 1994). It played an important role, with expansion in areas close to the sea from 9.8 to 7 ka BP (Bellini et al., 2009) in eastern Liguria and northern Tuscany. It was recorded from the beginning of the pollen diagram (7 ka BP) to 6 ka BP at Massaciuccoli, but it probably occurred earlier as suggested from the more southern site of Lago dell'Accesa, where it was found in the interval 10.2–6 ka BP. Fir pollen occurs sporadically in northern Latium Holocene records, suggesting a long distance transport.

Hazel pollen has been found from the early postglacial, followed by *Quercus ilex* type with beech, after ca. 9.8 ka BP in northern Latium. Hazel and fir pollen have been found (the

latter since around 6.8 ka BP) in postglacial spectra from Basilicata (Lago Grande di Monticchio). *Quercus ilex* type pollen with olive pollen (the latter increasing since around 8 ka BP) has been found in Sicily. In Sicily, oak forests persisted, even if progressively decreasing, from about 8.3–8.1 ka BP to about 4.2 ka BP. In Latium, mixed deciduous oak woods were widespread from 10 to 4.1 ka BP. Favorable climatic conditions for fir and deciduous oak forests persisted in the northern Apennines until ca. 4.2 ka BP, when beech woods increased. Fires were almost absent regionally from ca. 8.3 to 4.5 ka BP, but fires of anthropogenic origin began around 4 ka BP. Human disturbance also coincided with increased aridity in Sicily as early as ca. 8.9 ka BP. At about the same time, the expansion of Mediterranean vegetation in central Italy began. It is surprising that at ca. 6.8 ka BP, upland sites of the northern Apennines registered the first appearance of sparse grains of olive and a late expansion of beech, possibly for edaphic or topographic reasons.

In the last few years, research has been concentrated mainly in coastal areas such as lagoons, marshy zones, coastal plains, and ancient harbor areas. The studied records depict a mosaic of different vegetation types for the last 10 ka with a high regional variability. From 9 to 6 ka BP, the records from the Ligurian and Tyrrhenian coast are characterized by high percentages of fir, probably growing close to the sea. At the time, Mediterranean vegetation, characterized by *Pistacia* and then by *Olea*, was already established in Sicily at both Gorgo Basso and Biviere di Gela (Tinner et al., 2009). The development of open mesothermophilous woods and Mediterranean maquis in the coastal plains of southern Liguria and Tuscany dates back to 7 ka BP after the local disappearance of fir. More to the south, just north of the Tiber delta, at Stagno di Maccarese, a quite similar situation is found, with riparian fresh water trees playing an important role (Di Rita et al., 2010). From the Adriatic side of the peninsula in Apulia, the available data come from Lago Alimini Piccolo (Di Rita and Magri, 2009) and the former Lago Battaglia, with evergreen Mediterranean vegetation dominant from the beginning of the records at ca. 5.5 ka BP and at ca. 6 ka BP, respectively. The last couple of millennia are well represented in harbor basins, of riverine (Pisa port, Mariotti-Lippi et al., 2006), deltaic (Rome port, Sadori et al., 2010) and marine (Naples port, Allevato et al., 2010; and Lago d'Averno, Gröger and Thulin, 1998) origin.

Recently, Sadori et al. (2011) carried out a comparison of selected pollen records from the central and eastern Mediterranean to detect climate changes in the mid-Holocene. The role of changing seasonality appears to be crucial for this region (Magny et al., 2012). Precipitation seasonality increased during the early to mid-Holocene, with winter precipitation attaining a maximum and summer precipitation, a minimum. At least three rapid climate events with changes in plant biomass are recorded: an abrupt and short change around 8.2 ka BP, one centered around 6 ka BP, and one soon after 3 ka BP. From the beginning of the Bronze Age (ca. 4.4 ka BP in this region), human impact overlapped with a climate change, probably bipartite, toward dryness.

In fact, anthropogenic pollen types have been found in many sites of central and southern Italy since about 3.9–3.8 ka BP. Forest opening occurred in Latium at about 4.2–4.1 ka BP,

associated with aridity. This sudden decrease in regional arboreal pollen concentration followed a previous decline at ca. 4.9–4.7 ka BP. This twofold climate change was followed rapidly by strong human impact. Since Roman times, humans were affecting the vegetation throughout Italy. This probably overlapped with climate change, and the two factors cannot be disentangled.

Southern Balkan Peninsula, Dalmatian Islands and Crete

Since the review by Willis (1994), many new data from the region have become available (see Tables 4 and 5). Among these, several studies from the western mountains of Bulgaria, shedding new light on glacial refugia, are available (Tonkov, 2003; Tonkov et al., 2008). New pollen records have been published from Crete (Bottema and Sarpaki, 2003 and references therein) and from the Dalmatian islands of Mljet and Cres and from western Greece (Jahns, 2005). A particular increase in information about eastern Mediterranean palynology comes from Albania, where big lacustrine basins are present. Two of them are wide and transboundary lakes, namely Lake Ohrid and Lake Shkodra. There are current (Lake Shkodra; Zanchetta et al., 2012) and already well documented (ancient Lake Maliq, Denéfle et al., 2000; and Lake Ohrid, Wagner et al., 2009) research about them and the postglacial vegetation and climate history of the region. A review article (Sadori et al., 2011) presented new pollen curves from Mljet, Voulkaria and Maliq.

Regionally, coniferous trees, with patches of mesophilous and thermophilous taxa, characterized the Late Glacial interstadial. The early expansion of oak woodland (ca. 13.9 ka BP) at Kopais, Greece, mirrors a similar spread in southeastern Spain and central Italy. Due to sampling problems at the Lake Ohrid (284 m max depth), the 40 ka core investigated by Wagner et al. (2009), taken at the considerable water depth of 105 m, shows a hiatus during the Late Glacial. Last glacial vegetation included deciduous oaks, which accompanied the pine spread of the first half of the Holocene. Postglacial forest expansion was, in fact, mainly by deciduous oak and accompanying taxa, but it varied widely across the Balkan Peninsula, suggesting different possible locations of single taxa refugia. The spread of forest was not synchronous. A dense canopy forest was established in northwestern Greece by 12.5 ka BP, while in northern Greece and in southern Bulgaria, woodlands remained fairly open until ca. 10.2 ka BP. Around 11.9 ka BP, an intermediate forest cover was established on the island of Cres (Croatia). In the northern Pirin Mountains of Bulgaria, deciduous woods with pines and birch upslope began to spread at this time. Along the southeastern Black Sea coast (Filipova-Marinova et al., 2004), there is evidence for early postglacial migration of deciduous trees (oak, elm, linden, European ash (*Fraxinus excelsior*) and maple (*Acer*)) from their Late Glacial refugia in the Strandzha Mountains of southeast Bulgaria.

In the mountains of northwest Greece, dense forests were established by 11.9 ka BP or a few centuries thereafter. A slight but noteworthy expansion of *Pistacia* (probably *P. terebinthus* in most sites) occurred during the early Holocene throughout much of Greece (Lawson et al., 2004). Its peak between 10.2

and 8.9–8.3 ka BP suggests an aridity phase. This spread was also recorded in Crete, with maximum values at 8.9 ka BP. Here, open deciduous and evergreen oak woodland dominated. There is no evidence of this dry phase in either Bulgaria or in the northern Adriatic island of Cres. In western Greece, at Lake Voulkaria (Jahns, 2005), pollen of pistachio (*Pistacia*) is present since the beginning of the Holocene. A change was found in most upland sites from Greece, Bulgaria, and Albania (Denèfle et al., 2000) between 8.3 and 6.8 ka BP. This was marked by decreasing levels of deciduous oak pollen and by increasing levels of one or more of the following taxa: hazel, hornbeam or ironwood (*Carpinus orientalis/Ostrya*), elm, linden, beech, fir, and pine. As strikingly similar vegetation dynamics were found during previous interglacials, many possible explanations have been advanced, but the pattern rules out anthropogenic intervention (see Pollen Records, Last Interglacial of Europe). Among the explanations, a shift toward wetter/cooler conditions and the development of soils with humic horizons (Tonkov, 2003 and references therein) seem probable. On the other hand, the expansion of hornbeam (*Carpinus betulus/Ostrya*) and beech found at Lake Maliq may point to conditions of lower availability of moisture and warmer winters because these tree taxa are tolerant of dryness but intolerant of winter cold (Denèfle et al., 2000).

An important change was found in a quite different landscape at the Adriatic island of Mljet around 8.1 ka BP. Here, a partial replacement of deciduous oak by juniper and *Phillyrea* resembles the start of the slow spread of many sclerophyllous trees found in Sicily and in central Italy. This change is synchronous with the aridity phase in the southwestern Mediterranean from 8.3 to 7.8 ka BP.

Unequivocal human intervention in the Balkans is detected only very late in the pollen record. The first sign of anthropogenic disturbance consists in a reduction of both forest density (increasing pasture and secondary anthropogenic indicators) and/or diversity (reduction of deciduous oaks and appearance or increase of fruit trees and/or Ericaceae). Such signals are found in most pollen diagrams only after 5.7–5.1 ka BP. It is worthwhile to note again that a ‘Mediterraneanization’ of the climate could likewise have produced such a vegetation change. Fruit tree pollen appeared very early, although in low amounts. Most of these trees were present in the region since early Holocene. In some sites, walnut pollen grains appeared or increased at ca. 3.5 ka BP (together with *Ceratonia* in Crete). Some centuries later, the pollen of these cultivars was spread throughout Greece and Dalmatia. The onset of olive cultivation is difficult to define because it is not clear if the expansion found at least since the Bronze Age (as in Sicily) could be ascribed to a natural spread of this native plant. From Greek times (i.e., late Bronze Age) onwards, human impact produced strong landscape changes culminating in the present conditions.

Conclusions

The postglacial vegetation history of southern Europe is characterized by dynamic processes in which no additional plant taxa entered the region. All the pollen taxa found in the region have been here since the end of the last glacial. It is clear that a

complex pattern of climatic change prevailed across the region throughout the postglacial interval (Table 5).

Some common trends appeared in southern Europe (Spain, France, Italy, and western Balkans), in addition to a number of rapid climate events with changes in plant biomass (Pérez-Obiol et al., 2011; Sadori et al., 2011). Moreover, strong regional differences in vegetation history occurred, likely linked to regional differences in the climate and to problems in chronological assessments of individual records. Pollen data can in fact be the product of either climate changes or human activity or both. The dilemma can be resolved using a multiproxy approach, comparing pollen data to information about lake level changes and data from lake and cave isotopes, the records of which are less ambiguous with respect to an anthropogenic signal. In addition, when synthesized, primary paleoclimate data from isotope records and speleothems are compared with regional-scale climate modeling simulations, both model output and proxy data suggest an east-west division in Mediterranean climate history (Roberts et al., 2011).

The onset of reforestation began around 12.4 ka BP, with different vegetation dynamics and woodland densities in the various regions. In recent years, it has become clear that establishing the timing of these events requires improved chronologies for regional pollen records. The so-called climate optimum for deciduous oaks (the dominant taxon of the early Holocene) began around 10.2 ka BP, but its duration and areal extent varied from region to region. The 8.2 ka BP abrupt event identified in North Atlantic cores has only been detected in some Mediterranean sites. Many southern European records show substantial differences between early and late Holocene vegetation. They suggest a general evolution from wetter to drier climatic conditions with clearly arid phases (see Paleoclimate Relevance to Global Warming) in southwestern Europe, leading to the establishment of the modern Mediterranean climate. This trend began around 8.3 ka BP, marked by slight increases in pollen from sclerophyllous taxa and culminating around 5.1–4.5 ka BP. In contrast, southeastern Europe apparently experienced increased humidity from ca. 8.3 to 5.7 ka BP. Around 5.7–5.1 ka BP, marked changes in woodland range, density, and composition occurred, and typical Mediterranean vegetation spread throughout the Mediterranean basin. However, anthropogenic effects cannot be ruled out even if there is no doubt that humans alone could not have caused major synchronous changes in vegetation throughout the Mediterranean region. In recent millennia, humans modified the vegetation through livestock grazing, cultivation, burning, and associated living activities. Such activities either caused or exacerbated the decline in forest cover, disguising the climate signal. Human impact on the southern Mediterranean landscapes is hard to detect in pollen diagrams before the Bronze Age (Sadori et al., 2011), probably because prior to this, human impacts were predominantly local in scale and did not affect regional vegetation patterns to any great extent. The best way to study these phenomena is by combining continuous records of proxy vegetation and climate change such as lake cores located as close as possible to archaeological sites, which are the direct link to the region’s cultural history. Throughout the postglacial, human populations modified the plant landscape more or less strongly, but the climatic changes were the determining factors on the development of southern European environments.

Human populations were obliged to adapt to natural environmental variations, increasing their impact and the environmental consequences of global changes.

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See also: Paleoclimate Relevance to Global Warming; Pollen Records, Last Interglacial of Europe. **Pollen Records, Late Pleistocene:** Middle and Late Pleistocene in Southern Europe. **Radiocarbon Dating:** Calibration of the ^{14}C Record; Conventional Method.

References

- Allevato E, Russo Ermolli E, Boetto G, and Di Pasquale G (2010) Pollen-wood analysis at the Neapolis harbour site (1st-3rd century AD, southern Italy) and its archaeobotanical implications. *Journal of Archaeological Science* 37: 2365–2375.
- Andrieu-Ponel VP, Ponel JT, Jull AJT, de Beaulieu J-L, Bruneton H, and Leveau P (2000) Toward the reconstruction of the Holocene vegetation history of Lower Provence: Two new pollen profiles from Marais des Baux. *Vegetation History and Archaeobotany* 9: 71–84.
- Bar-Matthews M and Ayalon A (2011) Mid-Holocene climate variations revealed by high-resolution speleothem records from Soreq Cave, Israel and their correlation with cultural changes. *The Holocene* 21: 163–171.
- Bellini C, Mariotti-Lippi M, and Montanari C (2009) The Holocene landscape history of the NW Italian coasts. *The Holocene* 19: 1161–1172.
- Bottema S and Sarpaki A (2003) Environmental change in Crete: A 9000-year record of Holocene vegetation and the effect of the Santorini eruption. *The Holocene* 13: 733–749.
- Carrión JS, Fernández S, González-Sampériz P, et al. (2010a) Expected trends and surprises in the Lateglacial and Holocene vegetation history of the Iberian Peninsula and Balearic Islands. *Review of Palaeobotany and Palynology* 162: 458–475.
- Carrión JS, Fernández S, Jiménez-Moreno G, et al. (2010b) The historical origins of aridity and vegetation degradation in southeastern Spain. *Journal of Arid Environments* 74: 731–736.
- de Beaulieu J-L, Miras Y, Andrieu-Ponel V, and Guiter F (2005) Vegetation dynamics in the North-West Mediterranean regions: Instability of the Mediterranean bioclimate. *Plant Biosystems* 139: 114–126.
- Denèfle M, Lezine AM, Fouache E, and Dufaure JJ (2000) A 12,000-year pollen record from Lake Maliq, Albania. *Quaternary Research* 54: 423–432.
- Di Rita F, Celant A, and Magri D (2010) Holocene environmental instability in the wetland north of the Tiber delta (Rome, Italy): Sea-lake-man interactions. *Journal of Paleolimnology* 44: 51–67.
- Di Rita F and Magri D (2009) Holocene drought, deforestation and evergreen vegetation development in the central Mediterranean: A 5500 year record from Lago Alimini Piccolo, Apulia, southeast Italy. *The Holocene* 19: 295–306.
- Filipova-Marinova M, Christova R, and Božilova E (2004) Palaeoecological conditions in the Bulgarian Black sea area during the Quaternary. *Environmental Micropaleontology, Microbiology and Meiobenthology* 1: 136–154.
- Fletcher WJ, Boski T, and Moura D (2007) Palynological evidence for environmental and climatic change in the lower Guadiana valley, Portugal, during the last 13000 years. *The Holocene* 17: 481–494.
- Grüger E and Thulin B (1998) First results of biostratigraphical investigations of Lago d'Averno near Naples relating to the period 800 BC–800 AD. *Quaternary International* 47–48: 35–40.
- Jahns S (2005) The Holocene history of vegetation and settlement at the coastal site of Lake Voulkaria in Acarnania, western Greece. *Vegetation History and Archaeobotany* 14: 55–66.
- Jalut G, Dedoubat JJ, Fontugne M, and Otto T (2009) Holocene circum-Mediterranean vegetation changes: Climate forcing and human impact. *Quaternary International* 200: 4–18.
- Jalut G, Esteban Amat A, Bonnet L, Gauquelin T, and Fontugne M (2000) Holocene climatic changes in the Western Mediterranean, from south-east France to south-east Spain. *Palaeogeography, Palaeoclimatology, Palaeoecology* 160: 255–290.
- Kaplan JO, Krumhardt KM, and Zimmerman N (2009) The prehistoric and preindustrial deforestation of Europe. *Quaternary Science Reviews* 28: 3016–3034.
- Lawson I, Frogley M, Bryant Ch, Preece R, and Tzedakis PC (2004) The Lateglacial and Holocene environmental history of the Ioannina basin, north-west Greece. *Quaternary Science Reviews* 23: 1599–1625.
- Lowe JJ and Watson C (1994) Lateglacial and early Holocene pollen stratigraphy of the northern Apennines, Italy. *Quaternary Science Reviews* 12: 727–738.
- Magny M and Miramont Sivan O (2002) Assessment of the impact of climate and anthropogenic factors on Holocene Mediterranean vegetation in Europe on the basis of palaeohydrological records. *Palaeogeography, Palaeoclimatology, Palaeoecology* 186: 47–59.
- Magny M, Peyron O, Sadori L, Ortu E, Zanchetta G, Vannièr B, and Tinner W (2012) Contrasting patterns of precipitation seasonality during the Holocene in the south- and north-central Mediterranean. *Journal of Quaternary Science* 27: 290–296.
- Magri M (2007) Advances in Italian palynological studies: late Pleistocene and Holocene records. *Geologica Föreningens i Stockholm Förhandlingar* 129: 337–344.
- Mariotti-Lippi M, Bellini C, Trinci C, Benvenuti M, Pallecchi P, and Sagri M (2006) Pollen analysis of the ship site of Pisa San Rossore, Tuscany, Italy: The implications for catastrophic hydrological events and climatic change during the late Holocene. *Vegetation History and Archaeobotany* 16: 453–465.
- Miras Y, Ejarque A, Orengo H, Riera Mora S, Palet JM, and Poiraud A (2010) Prehistoric impact on landscape and vegetation at high altitudes: An integrated palaeoecological and archaeological approach in the eastern Pyrenees (perafita valley, Andorra). *Plant Biosystems* 144: 924–939.
- Moreno JM, Pineda FD, and Rivas-Martínez S (1990) Climate and vegetation at the Eurosiberian-Mediterranean boundary in the Iberian Peninsula. *Journal of Vegetation Science* 1: 233–244.
- Muñoz-Sobrino C, Ramil-Rego P, and Gómez-Orellana L (2004) Vegetation of the Lago de Sanabria area (NW Iberia) since the end of the Pleistocene: A palaeoecological reconstruction on the basis of two new pollen sequences. *Vegetation History and Archaeobotany* 13: 1–22.
- Pérez-Obiol R, Jalut G, Julià R, et al. (2011) Mid-Holocene vegetation and climatic history of the Iberian Peninsula. *The Holocene* 21: 75–93.
- Reille M, Gamisans J, de Beaulieu J-L, and Andrieu V (1997) The late-glacial at Lac de Creno (Corsica, France): A key site in the western Mediterranean basin. *New Phytologist* 135: 547–559.
- Roberts N, Brayshaw D, Kuzucuoglu C, Pérez-Obiol R, and Sadori L (2011) The mid-Holocene climatic transition in the Mediterranean: Causes and consequences. *The Holocene* 21: 3–13.
- Rohling EJ and Palike H (2005) Centennial-scale climate cooling with a sudden cold event around 8,200 years ago. *Nature* 434: 975–979.
- Sadori L, Giardini M, Giraudi C, and Mazzini I (2010) The plant landscape of the imperial harbour of Rome. *Journal of Archaeological Science* 37: 3294–3305.
- Sadori L, Jahns S, and Peyron O (2011) Mid-Holocene vegetation history of the central Mediterranean In: Roberts N, Sadori L, and Pérez Obiol R (eds.) Special Issue: Mid-Holocene Climate Change in the Mediterranean Region and Its Consequences. *The Holocene* 21: 117–129.
- Sadori L and Narcisi B (2001) The postglacial record of environmental history from Lago di Pergusa (Sicily). *The Holocene* 11: 655–671.
- Sánchez Goñi MF and Hannon GE (1999) High-altitude vegetational pattern on the Iberian Mountain Chain (north-central Spain) during the Holocene. *The Holocene* 9: 39–57.
- Tinner W, van Leeuwen JFN, Colombaroli D, et al. (2009) Holocene environmental changes at Gorgo Basso, a Mediterranean coastal lake in southern Sicily, Italy. *Quaternary Science Reviews* 1–13.
- Tonkov S (2003) Holocene palaeovegetation of the northwestern Pirin mountains (Bulgaria) as reconstructed from pollen analysis. *Review of Palaeobotany and Palynology* 124: 51–61.
- Tonkov S, Bozilova E, Possnert G, and Velčev A (2008) A contribution to the postglacial vegetation history of the Rila Mountains, Bulgaria: The pollen record of Lake Trilistnika. *Quaternary International* 190: 58–70.
- Wagner B, Lotter AF, Nowaczyk N, et al. (2009) A 40,000-year record of environmental change from ancient Lake Ohrid (Albania and Macedonia). *Journal of Paleolimnology* 41: 407–430.
- Willis KJ (1994) The vegetational history of the Balkans. *Quaternary Science Reviews* 13: 769–788.
- Yll EI, Pérez-Obiol R, Pantaléon-Cano J, and Roure JM (1997) Palynological evidence for climatic change and human activity during the Holocene on Minorca (Balearic Islands). *Quaternary Research* 48: 339–347.
- Zanchetta G, Van Welden A, Baneschi I, et al. (2012) Multiproxy record for the last 4500 years from Lake Shkodra (Albania/Montenegro). *Journal of Quaternary Science* 27: 780–789.

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QUATERNARY STRATIGRAPHY

Contents

Overview

Continental Biostratigraphy

Chronostratigraphy

Climatostratigraphy

Lithostratigraphy

Morphostratigraphy/Allostratigraphy

Pedostratigraphy

Sequence Stratigraphy

Tephrochronology

Overview

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Introduction

Stratigraphy, derived from the Latin 'stratum' and the Greek 'graphia,' was traditionally considered to be the science of rock strata. However, the definition has now been broadened to include nonlayered sedimentary rocks (e.g., diamictite), igneous rocks, and metamorphic rocks. Indeed, the general scope of stratigraphy includes all rocks forming the crust of the Earth (and other planets), whether consolidated or unconsolidated, and their organization into distinctive stratigraphic units based on their properties and attributes. Stratigraphic procedures include the description, classification, naming, and correlation of stratigraphic units for the purposes of establishing their relationships in space and time. Thus, stratigraphy is concerned not only with the stratigraphic order and age relationships of rock bodies, but also with their distribution, lithology, fossil content, and all other properties and attributes. Stratigraphy also includes the interpretation of environment and mode of origin of rocks and their geological history (Jackson, 1997; Salvador, 1994).

The use of the term 'rock' to describe Quaternary strata that are unconsolidated or semiconsolidated may seem inappropriate

at first sight. However, such usage is consistent with that adopted by the International Stratigraphic Guide (Salvador, 1994), and therefore, the term 'rock' is used here in its broadest sense.

Stratigraphic classification, which underpins almost all Quaternary research, is the arbitrary but systematic organization of rocks into units based on any of their attributes and properties. The three most common formal units of stratigraphic classification are lithostratigraphic (based on lithological properties), biostratigraphic (based on fossil content) and chronostratigraphic (based on time of formation), but other kinds of stratigraphic units are also defined, including magnetostratigraphic, sequence stratigraphic, climatostratigraphic, and so on. While each kind of stratigraphic unit may be useful in stratigraphic classification in certain areas, or for particular purposes, chronostratigraphic units are recognized worldwide and offer the best means of international communication among stratigraphers. Formally defined global chronostratigraphic units comprise the international Geological Time Scale (GTS), and include such major units as the Holocene, Pleistocene, Pliocene, and Miocene Series, which together constitute the Quaternary and Neogene Systems.

It is essential that named stratigraphic units have their characteristics and boundaries clearly described so that they may be recognized away from the place (type locality or type section) where they were originally defined and named. The nomination of a type locality or section also allows other people the opportunity to visit and study the site and to ensure that everyone has the same understanding of the character and definition of the unit.

Comprehensive rules for defining stratigraphic units are given in the International Stratigraphic Guide (Salvador, 1994).

Stratigraphic Correlation: Principles and Methods

Stratigraphic correlation is the process by which stratigraphic units in different places are shown to be similar in character and/or stratigraphic position. Correlation may be on the basis of lithology, fossil content, age, or any other property (Jackson, 1997).

Some Basic Principles

A fundamental principle in stratigraphic correlation is the Law of Superposition, first articulated by Nicholas Steno in the seventeenth century, which states that in any stratigraphic sequence that is not overturned, the youngest stratum is at the top and the oldest is at the base. The Law of Superposition is important in establishing the relative age of units in a stratigraphic sequence, and also for local, regional, and global correlation. For example, the correlation of oxygen isotope stages between two or more deep-sea cores can usually be achieved by matching the pattern of isotope variations from the sea floor downward provided that the sediment cores lack unconformities and that no significant sediment is missing at the core top.

Another fundamental principle is that older landforms and associated deposits are commonly found higher in the landscape than younger ones. This is due to the general tendency of erosional processes to lower the landscape, often assisted by tectonic uplift. Thus, a flight of river terraces will increase in age with elevation, as a consequence of progressive fluvial down-cutting. Up until the early 1970s, marine terraces were often correlated from region to region on the basis of elevation. However, with the recognition of the significant role of tectonism and the advent of modern dating methods, especially U/Th dating of corals, the shortcomings of altimetric correlation became apparent.

Commonly Used Methods of Correlation

1. *Lithostratigraphy*: Lithostratigraphy is the basis for most geological mapping and for description of stratigraphic sections, both in outcrop and in drill holes.
2. *Biostratigraphy*: Whether it be microfossils in deep-sea sediments or the distinctive floras and faunas (including human culture) of continental deposits, biostratigraphy is one of the principal methods of correlation. Some of the best known large mammal fossils of the Quaternary, such as woolly mammoths (Figure 1), are most important for continental biostratigraphy.



Figure 1 The author, holding a late Pleistocene woolly mammoth tusk, recovered by gold miners in Bonanza Creek, near Dawson City in the Yukon, northwest Canada. Such vertebrate fossils are well preserved in the permafrost zones of the Northern Hemisphere.

3. *Tephrostratigraphy*: Tephrae represent almost instantaneous events and enable very precise correlation both within and between marine and terrestrial sequences. Once a tephra is dated at one site, the age can be 'transferred' to all other sites where the tephra occurs.
4. *Magnetostratigraphy*: The major magnetic reversals provide important chronostratigraphic markers in the Quaternary, including those associated with the proposed Lower/middle Pleistocene boundary (Matuyama/Brunhes reversal), the Plio/Pleistocene boundary (top of Olduvai Subchron), and the base of the Gelasian Stage (Gauss/Matuyama reversal). Magnetostratigraphy also plays an important role in marine-terrestrial correlation.
5. *Geochronology*: Geological dating methods yield numerical age estimates that are being increasingly used for correlation. However, all age estimates have field and laboratory uncertainties that must be assessed.
6. *Climatostratigraphy*: Since the recognition of widespread glacial deposits of Quaternary age in the early nineteenth century, global climatic changes, driven by astronomical variations, have provided a climatostratigraphic framework for correlation.
7. *Sea level*: Sea level is recognized as a potential global reference for correlation of shoreline deposits. However, tectonic and isostatic factors, a lack of precise numerical

ages, and problems in identification of paleo-sea-level datums have hindered its use. The Last Interglacial shore-line (~120–130 ka), with the paleo-sea-level a few meters higher than at present, is one of the most widespread datums for measuring tectonic deformation.

8. *Aminostratigraphy*: Fossil proteins contain amino acids that undergo racemization depending on elapsed time, the nature of the fossil material, and a range of environmental factors, including temperature. The extent of racemization is used for correlation.
9. *Oxygen isotope stratigraphy*: Oxygen isotope variations in Foraminifera from deep-sea cores have become the global reference for Quaternary stratigraphic correlation. Astronomical tuning provides a precise chronology.
10. *Sequence stratigraphy*: Originally developed by the oil exploration industry for interpreting seismic section, sequence stratigraphy works best in continental shelf sediments where facies architecture is controlled by relative sea-level changes. In Wanganui Basin, New Zealand, sequence stratigraphic models can be tested against the known history of Quaternary sea-level changes.
11. *Pedostratigraphy*: Paleosols are used for correlation, particularly in loess sequences, and they have also been used for paleoclimate reconstructions.
12. *Morphostratigraphy*: Glacial moraines and outwash terraces are often correlated using morphostratigraphy.
13. *Magnetic susceptibility*: Magnetic susceptibility variations in Chinese loess sequences show a pattern very similar to that of oxygen isotope records in deep-sea cores, and allow

correlation on a cycle-by-cycle basis between the two records, supplemented by magnetostratigraphy.

14. *Chemostratigraphy*: Chemostratigraphy is particularly valuable in ice cores and deep-sea cores. For example, the sulfate concentration in ice cores is used to identify volcanic eruptions.
15. *Geophysical measurements*: Geophysical measurements are used for correlation in deep-sea cores and in downhole logging of boreholes.

A Global Stratotype Section and Point for the Pleistocene

After a century of uncertainty, the 18th International Geological Congress (London, 1948) resolved to select a stratotype for the Pliocene–Pleistocene boundary and to ‘place the boundary at the horizon of the first indication of climatic deterioration in the Italian Neogene succession’ (King and Oakley, 1949). After consideration of the suitability of several sections over the next three decades (see discussion in Aguirre and Pasini, 1985), the Vrica section 4 km south of Crotona in Calabria, southern Italy (Figure 2, inset), was finally selected as the Global Stratotype Section and Point (GSSP), culminating in the ratification of the Plio–Pleistocene boundary, at the base of the marine claystones conformably overlying bed *e* in the Vrica section (Aguirre and Pasini, 1985). Subsequently, the GSSP for the Pleistocene (and Quaternary) was changed to the base of the Gelasian Stage, defined in the Monte San Nicola section in Sicily (Gibbard et al., 2010; Rio et al., 1998),

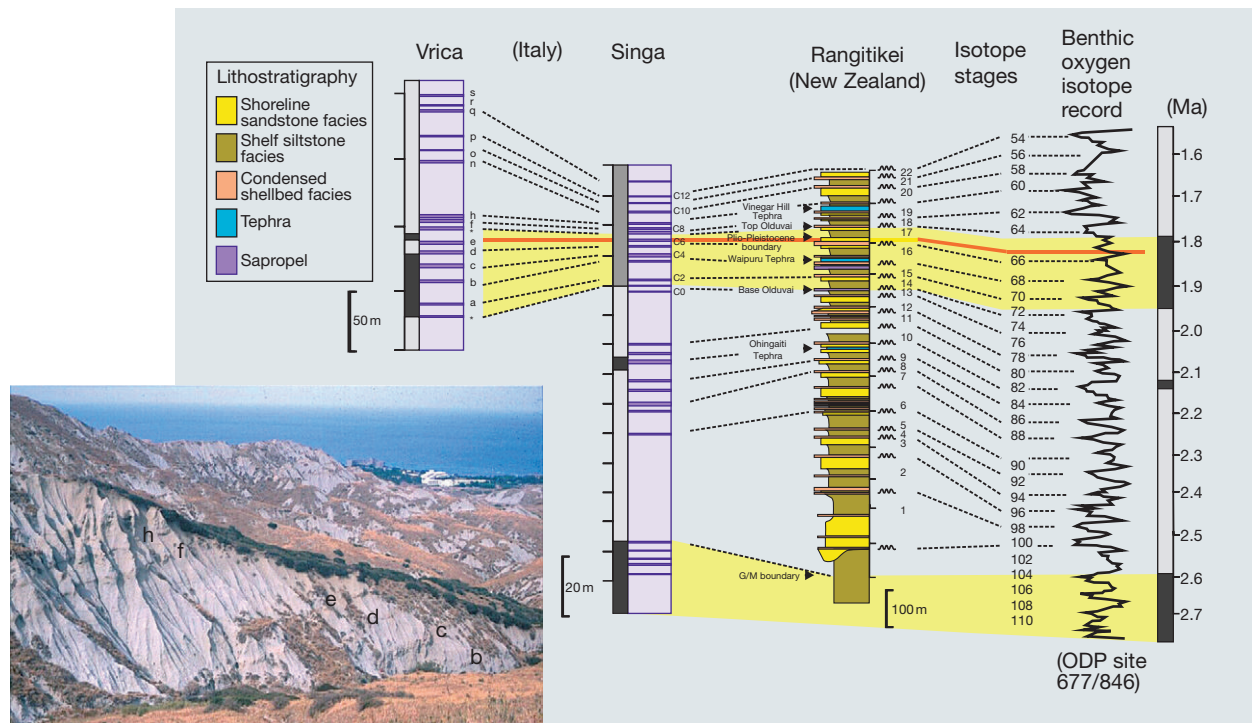


Figure 2 Correlation of Vrica GSSP to Wanganui Basin, New Zealand. Reproduced from Naish TR, Abbott ST, Alloway BV, et al. (1998) Astronomical calibration of a Southern Hemisphere Plio–Pleistocene reference section, Wanganui Basin, New Zealand. *Quaternary Science Reviews* 17: 695–710. Inset shows Vrica section with sapropel layers labeled.

with the Vrica GSSP to be used to define the base of the Calabrian Stage.

A primary requirement of any GSSP is that it provides multiple criteria for correlation of the boundary. In the case of the Vrica and Monte San Nicola GSSPs, the main methods of correlation are magnetostratigraphy and several marine microfossil datums, including Foraminifera, coccoliths, and ostracods (Aguirre and Pasini, 1985; Rio et al., 1998). In addition, regional lithostratigraphic correlation of sapropels has enabled the sections to be astronomically tuned, yielding an age of 1.806 Ma for the Vrica GSSP and 2.588 Ma for the Monte San Nicola GSSP (Lourens et al., 1996).

Prior to detailed study of the Italian sections, the principal means of correlation was climatostratigraphy. Thus, the base of the Pleistocene Series was equated with a putative major global cooling event that was variously identified on the basis of glacial till, pollen analysis, and faunal evidence. In New Zealand, for example, the incoming of the Sub-Antarctic scallop, *Zygochlamys delicatula*, to Wanganui Basin was estimated to represent the base of the Pleistocene, at a stratigraphic level that we now recognize to be at about 2.5 Ma, just above the Gauss/Matuyama boundary. At the time, it was widely accepted that the Pleistocene was characterized by just four major glaciations (named Gunz, Mindel, Riss, and Wurm in the European Alps), but later, when it became apparent that there were many more glacial episodes, the difficulties of using only climatostratigraphy became readily apparent.

The first precise correlation of the Vrica GSSP to New Zealand was achieved by Beu and Edwards (1984), and refined by Naish et al. (1998), using a combination of biostratigraphy, magnetostratigraphy, and cyclostratigraphy (Figure 2). It is noteworthy that this correlation can be achieved without numerical ages – it is not necessary to know the precise age of a chronostratigraphic boundary to successfully correlate it to other areas.

Stratigraphy by Numbers?

Modern dating methods provide numerical age estimates for a spectacular range of geological materials. Indeed, there are many situations in which age is the only means of correlation. However, correlation on the basis of age is dependent on a full evaluation of dating uncertainties and limitations, some examples of which are given in the following section.

Radiocarbon Dating at the Limit

The type locality of the Salmon Springs Glaciation near Sumner in Washington, northwestern USA, contains two units of interbedded gravel and till (drift), separated by 1.5 m of nonglacial sediments that include tephra, silt, and peat. The peat was dated at $71\,500 \pm 1700$ –1400 years BP (Stuiver et al., 1978), and both drift units were considered to represent a glacial episode during the early to the middle part of the last (Wisconsin) glaciation. The Salmon Springs Drift was widely correlated throughout the northwest United States, largely on the basis of ordinal sequence (glacial deposits that preceded

the last advance of the Cordilleran ice sheet were generally correlated with the Salmon Springs).

However, the age of the Salmon Springs Glaciation was considerably revised when Easterbrook et al. (1981) reported a zircon fission track age of 0.84 ± 0.21 Ma for the tephra at the type locality. Furthermore, the immediately overlying silts were of reversed polarity magnetization, consistent with an age of >0.78 Ma (i.e., older than the Mauyama/Brunhes polarity reversal). Presumably, the ~ 70 ka radiocarbon age was grossly in error as a result of minor, but significant, contamination by modern carbon. It is noteworthy that the validity of the Salmon Springs Drift as a stratigraphic unit is not compromised by the dramatic change in age because it is defined at its type locality, which remains unchanged. Indeed, the identification of a tephra at the type locality enhances the correlation potential of the Salmon Springs Drift with other sites.

A strikingly similar dating situation was encountered in the glacial deposits of Tasmania, where *Phyllocladus* wood in varved clays associated with the Linda Moraine gave a radiocarbon age of $26\,480 \pm 800$ years BP (Gill, 1956). This age was for a long time accepted as dating the Last Glacial Maximum in Tasmania. Subsequently, redating of the site yielded a radiocarbon age of $>40\,000$ years BP (Colhoun, 1985), and sediments within the Linda Moraines were shown to have reversed polarity (Barbetti and Colhoun, 1988), and therefore yielded an age of >0.78 Ma. Furthermore, pollen assemblages from an underlying paleosol contain abundant *Nothofagus brassii*, a taxa only known from pre-Pleistocene deposits in Tasmania. Thus, the Linda Moraine can be assigned an early Pleistocene age on the basis of magnetostratigraphy, biostratigraphy, and geochronology.

These two examples are a timely reminder that multiple correlation criteria should be employed wherever possible.

Aminostratigraphy and Amino Acid Dating of Bone

In the absence of independently dated calibration samples, the extent of amino acid racemization (usually given as a D/L ratio) in fossil shell material has proved a powerful tool for stratigraphic correlation (referred to as aminostratigraphy). However, in many cases, reliable calibration allows conversion of D/L ratios into numerical age estimates. Examples include marine bivalves in shoreline deposits (Hearty et al., 1986; Murray-Wallace, 1995), nonmarine mollusks (Bowen et al., 1989), and eggshell (Miller et al., 1999).

Dating the first human colonizations of North America is a controversial topic, with many archaeologists favoring a 'late' arrival, in the range 12–13 ka, although significantly older ages have been postulated. Some of the critical human fossils were found in the 1920s and 1930s in California and like many other human fossils, their stratigraphic context was not well documented. Thus, any age information had to come from direct dating of the bones rather than dating of associated materials.

The development of amino acid racemization (AAR) dating in the late 1960s offered the possibility of directly dating fossil bones that were previously unable to be dated by radiocarbon. The main advantage of AAR, at the time, was the small sample size (a few grams of bone) and an extended age range (~ 100 ka) relative to radiocarbon. Based on the extent of

aspartic acid racemization in bone, Bada et al. (1974) reported ages up to 48 ka for several skeletons from California, generating an intense controversy about the validity of the ages and the antiquity of humans in America. As AAR dating requires knowledge of the rate of racemization, calibration was obtained using one bone that had previously yielded a radiocarbon age of $17\,150 \pm 1470$ years BP on its collagen fraction.

Subsequently, the same skeletons were dated by U/Th and AMS radiocarbon, yielding ages <10 ka (Bada et al., 1984). In particular, the amino acid extracts from the calibration sample, collagen from which had previously been dated at ~ 17 ka, gave an AMS radiocarbon age of 5100 ± 500 years. Naturally, using the revised, younger age of the calibration sample considerably reduced the AAR age estimates. The combined results of the U/Th, AAR, and AMS radiocarbon measurements indicate Holocene ages for the skeletal material and reinforce the necessity for cross-checking of ages using multiple techniques. Furthermore, had the field stratigraphy been documented at the outset, the erroneous ages may well have been identified earlier and much of the controversy avoided.

The Timing of the Last Interglacial from Shoreline Deposits

The Last (Eemian) Interglacial was the last time, prior to the Holocene that the eustatic sea level was at or slightly higher than the modern level, and the last time when ice volumes and global climate were similar to the present. Thus, shoreline deposits of the last interglacial age are widely preserved, close to

the present sea level on tectonically stable coasts, and valuable as a global stratigraphic marker. On tectonically active coasts, the height of the Last Interglacial shoreline is an excellent datum for calculating uplift rates (Figure 3).

Uranium-series ages on fossil corals from the last interglacial reefs are generally in the range of 135–115 ka. However, the timing of the onset and the duration of the last interglacial varies widely from place to place, and a detailed reconstruction of sea-level changes during this time interval remains elusive. Age variations might be a consequence of several factors, including dating uncertainties and isostatic and tectonic effects, as well as uncertainties associated with paleo-sea-level indicators.

In a careful study, utilizing TIMS U-series dating of corals from Western Australia, Stirling et al. (1998) concluded that coral reef growth started at 128 ± 1 ka along the entire Western Australian coastline, when the relative sea level was at least 3 m above the present level. From a regressive reef sequence at Mangrove Bay (Figure 4), the termination of the last interglacial period occurred at 116 ± 1 ka, but the main period of reef building was between 128 and 121 ka. All samples came from sites that were considered to be tectonically stable and that were not significantly affected by glacio-isostatic adjustments, being remote from former ice sheets. To eliminate the possibility of reworking, coral samples were collected in growth position. Samples were also checked for possible diagenesis using a stringent set of criteria, including lack of evidence for recrystallization ($<1\%$ calcite), low ^{232}Th concentration,

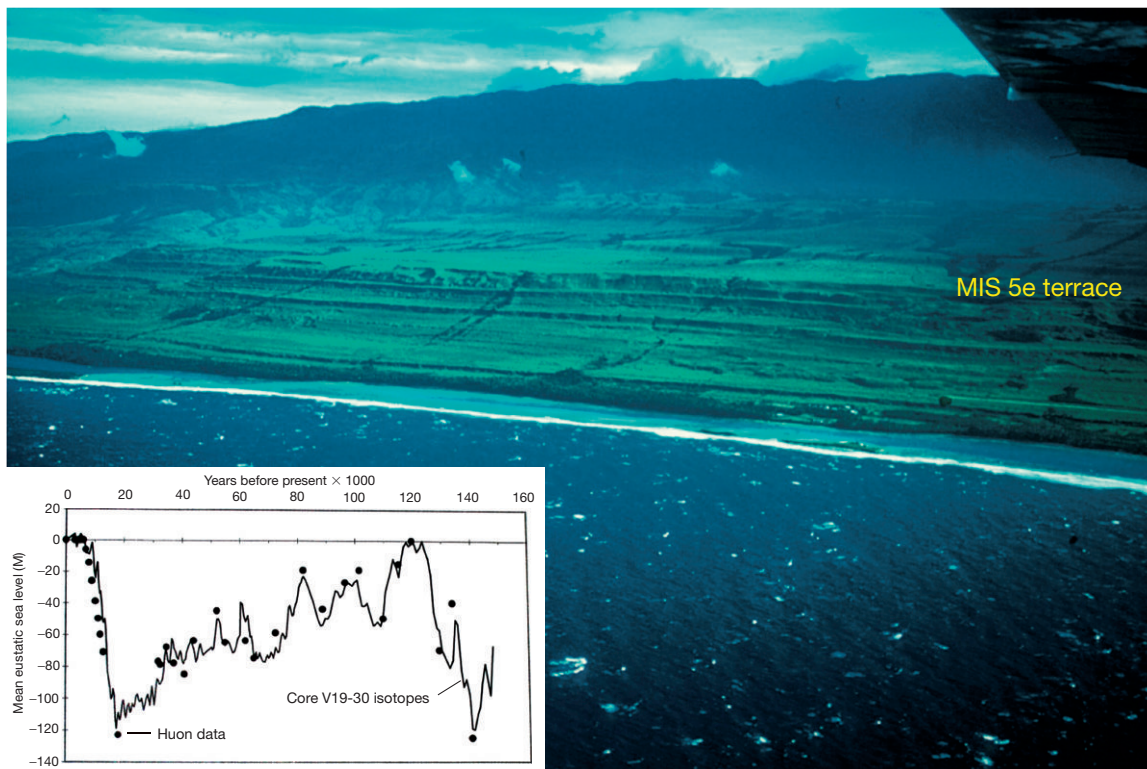


Figure 3 Uplifted coral terraces at Sialum on the Huon Peninsula, Papua New Guinea. The Last Interglacial reef complex is at an elevation of ~ 180 m, indicating a mean uplift rate of ~ 1.5 m/ka. Inset shows sea-level record of the last 140 ka as reconstructed from Huon Peninsula and deep-sea core data. Reproduced from Pillans B, Chappell J, and Naish TR (1998) A review of the Milankovitch climatic beat: Template for Plio-Pleistocene sea-level changes and sequence stratigraphy. *Sedimentary Geology* 122: 5–21. Photo by B.G. Thom.

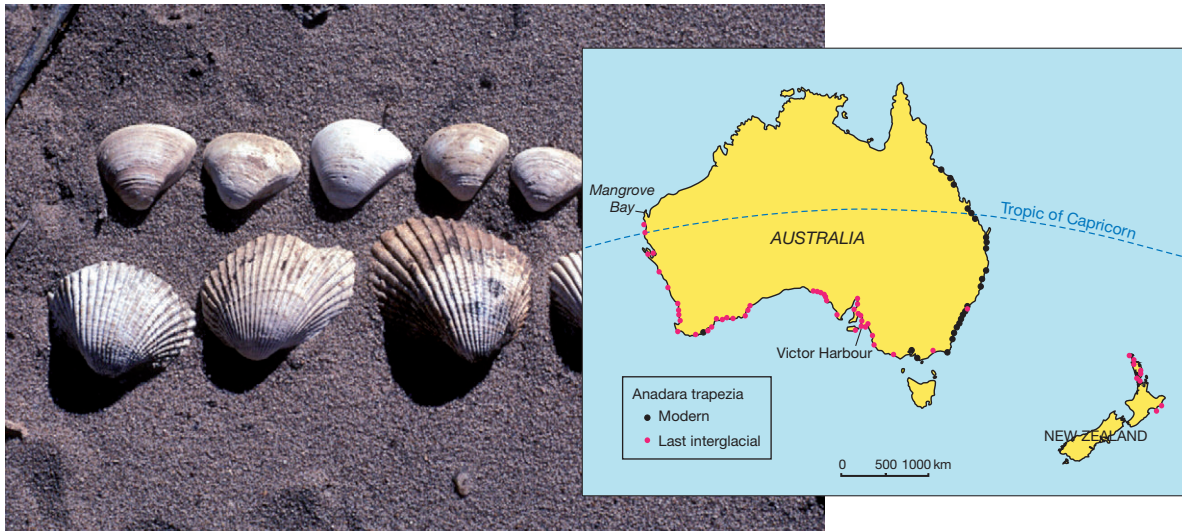


Figure 4 Distribution of *Anadara trapezia* in Australia and New Zealand. Photo shows specimens of *A. trapezia* (ribbed shells) and *Anapella cycladea* (smooth shells) from Last Interglacial shoreline deposits (Glanville Formation) at Victor Harbour, South Australia (see Murray-Wallace et al., 2000).

typical U concentration (~ 3 ppm), and initial $^{234}\text{U}/^{238}\text{U}$ ratio in the modern range of $149 \pm 4\%$.

Identification and dating of last interglacial shorelines in extratropical regions of Australia and elsewhere are hampered by a lack of corals, which have proved to be the most reliable dating medium. Uranium-series dating of fossil mollusk shells has been widely tried, but it is considered to be unreliable because of its open system behavior (e.g., Kaufman et al., 1971). Living mollusk shells contain very low uranium concentrations, typically a few ppb. However, fossil mollusks contain significantly higher U, typically in the ppm range, indicative of postmortem U uptake and leading to age underestimates. Eggins et al. (2005) used laser-ablation multi-collector ICPMS measurements on a robust mollusk shell (*Anadara trapezia*) of the last interglacial age to investigate the nature of U uptake. A profile of measurements 5 mm long from the interior to the exterior of the shell showed increasing U concentrations and $^{230}\text{Th}/^{234}\text{U}$ ratios toward the exterior, consistent with progressive U uptake. The oldest apparent ages (~ 120 ka), obtained at the exterior surface, are consistent with the last interglacial age, but significantly younger apparent ages (~ 60 ka) were obtained from the interior of the shell.

On the east coast of Australia, the last interglacial coastal sand barriers contain abundant fossil kauri (*Agathis* sp.) logs, which have yielded radiocarbon ages on different extracts from the wood that range from background to as young as 11 ka (Chappell et al., 1996). The true age of the barriers (~ 120 – 130 ka) is established from U/Th dating of fossil corals at one site, so the radiocarbon ages are erroneously young, reflecting variable amounts of contamination. In the 1960s and early 1970s, radiocarbon dating of shoreline deposits from many parts of the world indicated that sea level was within 10 m of the present level in the age range of 20–40 ka. However, Thom (1973) concluded that almost all these studies were in error either because the radiocarbon ages were wrong or because they had been incorrectly associated with sea-level events.

Biostratigraphy and aminostratigraphy are important for identifying the Last Interglacial shoreline in the Australasian region, with many sites in southern Australia and New Zealand containing the warm-water estuarine bivalve, *A. trapezia* (Murray-Wallace et al., 2000) – Figure 4. Similarly, in the Mediterranean, the Last Interglacial is recognized by the presence of the warm water ‘Senegalese’ fauna, including the distinctive gastropod, *Strombus bubonius*, coupled with aminostratigraphy, which is calibrated by U/Th dating of corals (Hearty et al., 1986).

The Demise of the Australian Megafauna

Most of Australia’s largest mammals, reptiles, and birds (megafauna) became extinct in the late Pleistocene, but the causes are hotly debated. Possible causes include direct human hunting, alteration of vegetation by human burning, and climate change, particularly increased aridity leading into the Last Glacial Maximum. Roberts et al. (2001) reported OSL and U/Th burial ages for megafauna from 28 sites across Australia and concluded that the extinction occurred $\sim 46,400$ years ago, soon after the arrival of people in Australia. Thus, humans rather than climate change were implicated as the cause of the extinction. Their dating concentrated mainly on sites that contained articulated bones, to avoid problems with possible reworking of older material. Radiocarbon ages were not directly used because of known contamination problems in materials older than ~ 35 ka.

In a parallel study, Miller et al. (1999) reported more than 700 AAR ages on *Genyornis* eggshell from three different climatic regions of Australia and concluded that *Genyornis*, a large (ostrich-sized) flightless bird, became extinct at 50 ± 5 ka. Subsequently, Miller et al. (2005) concluded, based on $\delta^{13}\text{C}$ measurements of eggshell, that the extinction of *Genyornis* was the result of a dramatic change in available food sources, possibly caused by human burning of landscapes, which rapidly changed the vegetation from a drought-adapted mosaic of

trees, shrubs, and grasses to the modern fire-adapted desert scrub. Unlike *Genyornis*, the modern Australian emu (*Dromaius*) survived, probably because it was smaller and faster (and hence less attractive as a hunting target to humans), and also because it had a less specialized grazing diet that made it more adaptable to the vegetation changes. A notable feature of these studies by Miller et al. (1999, 2005) is that the majority of fossil eggshell fragments were collected from modern deflation hollows in sand dunes, where the eggshell occurred as a surface lag deposit (Figure 5). In other words, little stratigraphic context was available, nor was it necessary, to document the demise of *Genyornis* because they were able to date the fossils directly.

The study by Roberts et al. (2001) has been criticized because they excluded evidence for megafauna survival well after 46 ka, including many archaeological sites, which, inevitably, contain disarticulated remains of animals that were hunted (Johnson, 2005). However, if the abundance of a megafauna species was very low and the species was being actively sought out by hunters, then archaeological sites would likely have a much greater chance of recording their presence than nonarchaeological sites such as caves and swamps that are passive, a chance accumulators of bones. Johnson (2005) listed 13 sites with evidence indicating the possible survival of megafauna after 46 ka, all of which contained disarticulated remains, and the majority of which contained stratigraphic evidence of overlapping human occupation. Furthermore, there was a significant correlation between body mass and age, with smaller species generally surviving longer than larger species. Johnson (2005) therefore concluded that while the abundances of all elements of the megafauna were reduced to drastically low levels around 50 ka, some species may have persisted in very low numbers over restricted geographic ranges, and the final extinction of remnant populations may have been spread over a comparatively long period (up to 30 ka) after this event.

The demise of the Australian megafauna highlights some limitations that are common to all biostratigraphic correlations which use first and last appearance datums for correlation. First, when a species or a group of species are present in very low abundance, appearance datums will be very dependent on geographic range, and second, the preservation potential of fossils will be very low. It also highlights the challenge of dating by association which is common to much of Quaternary

stratigraphy. How do we know that the material being dated is the same age as the material we want to know the age of? Miller et al. (1999, 2003) circumvented this problem by directly dating the *Genyornis* eggshells, while Roberts et al. (2001) employed stringent criteria to ensure that their OSL ages represented the time of burial of the megafauna fossils they wanted to know the age of. Reworking of older fossil material into younger deposits is the curse of biostratigraphy, but by only dating articulated skeletons, Roberts et al. (2001) minimized its impact in their study. Recent direct dating of fossil megafauna teeth by combined ESR and U-series methods also provides a way of identifying reworked materials (e.g., Grün et al., 2010).

Marine–Terrestrial Correlation

During the nineteenth and early twentieth centuries, Quaternary stratigraphy was largely founded on the study of continental deposits, particularly glacial deposits. At the same time, shoreline deposits provided links to the oceanic realm. However, in the second half of the twentieth century, and since then, the retrieval of deep ocean sediment cores has resulted in a major shift in focus and a revolution in Quaternary stratigraphy.

In particular, the oxygen isotope stratigraphy of deep-sea cores provided a climatostratigraphic template against which almost all other stratigraphies could be compared. This came about largely because the deep-sea sediment record is more continuous than most long continental records, with the exception of the Chinese loess record. Equally important was the development of the astronomically calibrated timescale for oxygen isotope stages (e.g., Lourens et al., 1996). Thus, marine–terrestrial correlation has become a significant part of Quaternary stratigraphy, some examples of which are discussed in the following paragraphs.

Wanganui Basin

Wanganui Basin in the North Island of New Zealand contains the most complete, on-land, shallow-marine record of Quaternary climatic and sea-level change yet described (Figure 6). Sedimentation generally kept pace with basin subsidence, resulting in a 3 km thick basin fill of dominantly coastal and shallow marine sediments over the last 3.6 Ma. The basin fill comprises cyclic, unconformity-bound strata, mostly representing environments located landward of the contemporaneous shelf edge. Gentle neotectonic upwarping of the eastern margin has produced on-land exposures through as many as 47 superposed cyclothem deposited since Marine Isotope Stage 104 (~2.6 Ma). Nearly all of the 41 and 100 ka Pleistocene sea-level cycles are represented by individual depositional sequences comprising transgressive (TST), highstand (HST), and regressive system tracts (RST). In general, major sedimentary facies within the basin fill were deposited in a range of coastal plain, shoreface, and shelf marine environments during the rise, highstand, and falling part of each glacioeustatic cycle. Thus, the Wanganui stratigraphic record represents deposition dominantly during odd-numbered (interglacial) isotope stages. Glacial stages are represented by regressive shoreline facies that are truncated above by surfaces of marine planation and bioerosion, and that mark both cyclothem and sequence boundaries.



Figure 5 Recently exposed *Genyornis* eggshell in late Pleistocene dune sand, Port Augusta area, South Australia. Photo by G.H. Miller.

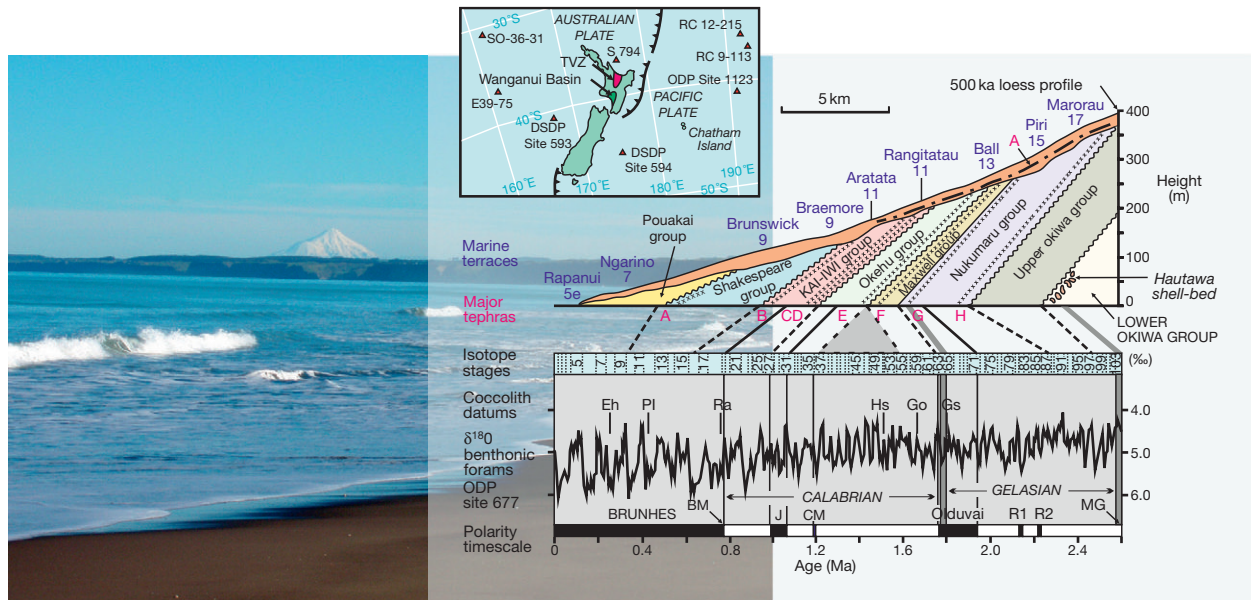


Figure 6 Left: Castlecliff Beach, Wanganui Basin, looking west to Mt Taranaki andesite stratovolcano. Right: Generalized cross-section (modified from Pillans, 1994a,b) showing stratigraphic relationships between marine terraces, and major lithostratigraphic units of the shallow marine basin fill, including correlation with the marine isotope record of ODP Site 677. *Major tephras*: A. Rangitawa (0.345 Ma), B. Kupe (0.63 Ma), C. Kaukatea (0.86 Ma), D. Potaka (1.00 Ma), E. Rewa (1.20 Ma), F. Pakihikura (1.58 Ma), G. Waipuru (1.79 Ma), H. Ohingaiti. Reproduced from Pillans B, Alloway B, Naish T, Westgate J, Abbott S and Palmer A (2005) Silicic tephras in Pleistocene shallow-marine sediments of Wanganui Basin, New Zealand. *Journal of the Royal Society of New Zealand* 35: 43–90. *Major biostratigraphic datums* (coccoliths): Eh. FAD *Emiliana huxleyi*, Pl. LAD *Pseudoemiliana lacunosa*, Ra. LAD *Reticulofenestra asanoi*, Hs. LAD *Helicosphaera sellii*, Go. FAD *Geophyrocapsa oceanica*, Gs. *Geophyrocapsa sinuosa*. Reproduced from Naish TR, Abbott ST, Alloway BV, et al. (1998) Astronomical calibration of a Southern Hemisphere Plio-Pleistocene reference section, Wanganui Basin, New Zealand. *Quaternary Science Reviews* 17: 695–710.

Southward migration of the basin depocenter and gentle marginal upwarping have not only resulted in excellent exposures of the basin fill, but have also led to the development of an extensive flight of marine terraces around the basin margin. The terraces were formed during each interglacial from ~680 ka (MIS17) to ~60 ka (MIS3), with overlying loess beds representing each glacial stage from MIS12 to MIS2 (Naish et al., 1998; Palmer and Pillans, 1996; Pillans, 1983). This direct stratigraphic association between loess and marine terrace sequences is unusual and allows unequivocal marine–terrestrial correlation (Pillans, 1994a,b) – **Figure 7**. Pollen analyses of interbedded peats, as well as plant phytoliths recovered from the loess, allow the vegetation history to be established and contribute to paleoclimate reconstructions.

An integrated chronology for the basin has been developed from radiometric ages of tephras, magnetostratigraphy, sequence stratigraphy, biostratigraphy, and cyclostratigraphic correlation to the oxygen isotope stratigraphy of deep-sea cores east of New Zealand (Pillans et al., 2005). However, none of these techniques would have been successful without careful field mapping and lithological description of stratigraphic sections.

Rangitawa Tephra: A Widespread Marker Bed in the South West Pacific Region

The Taupo Volcanic Zone (TVZ) in New Zealand is the source of some of the largest Quaternary volcanic eruptions in the world – Machida (2002) lists 12 tephras with volumes >100 km³ that

were erupted from the TVZ during the last 1.6 Ma, the last being the Taupo Tephra only 1800 years ago.

The Rangitawa Tephra (345 ka) was perhaps the largest eruption from the TVZ in the last 500 ka, probably associated with the eruption of the Whakamaru group ignimbrites (Kohn et al., 1992). First identified in the early 1950s in Wanganui Basin, Rangitawa Tephra is found throughout the central and southern North Island, in the South Island, and in deep-sea cores more than 1000 km from its source. Thus, the tephra is a valuable widespread mid-Pleistocene marker bed that enables marine–terrestrial correlations throughout the New Zealand region. Pillans et al. (1996) reported 51 independent numerical age estimates for Rangitawa Tephra, in the range 176–456 ka, only 15 of which were considered reliable. The weighted mean age of nine reliable fission track age determinations was 345 ± 12 ka, in excellent agreement with a weighted mean age of 340 ± 7 ka for two reliable astronomically calibrated age estimates from deep-sea cores.

At the well-documented Rangitatau East section (**Figure 7**) in Wanganui Basin, the tephra occurs in the upper part of Loess L10 with a well-developed paleosol immediately above. At the Mt Curl section (**Figure 8**), further east, the tephra is underlain by loess and overlain by interglacial coastal dune sand. Thus, the tephra is interpreted to have fallen at the end of a glacial period, consistent with pollen analyses of associated sediments at other sites (Kohn et al., 1992), and consistent with its position in two deep-sea cores close to the MIS 10/9 boundary (**Figure 8**).

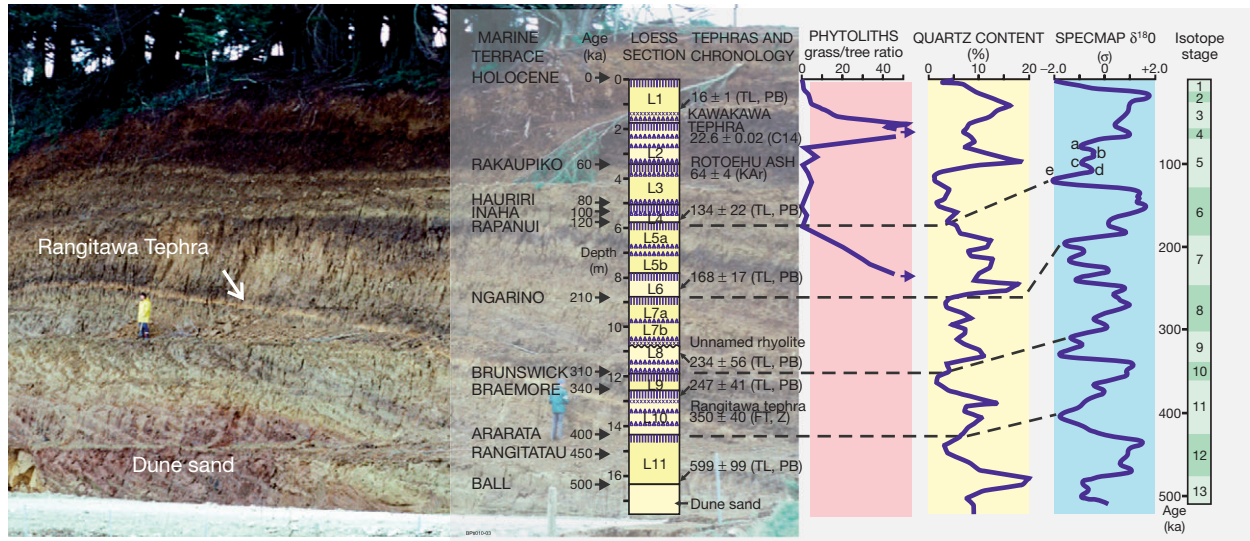


Figure 7 Stratigraphy and chronology of loess and tephras at Rangitatau East section, Wanganui Basin, and relationships to marine terraces (Palmer and Pillans, 1996).

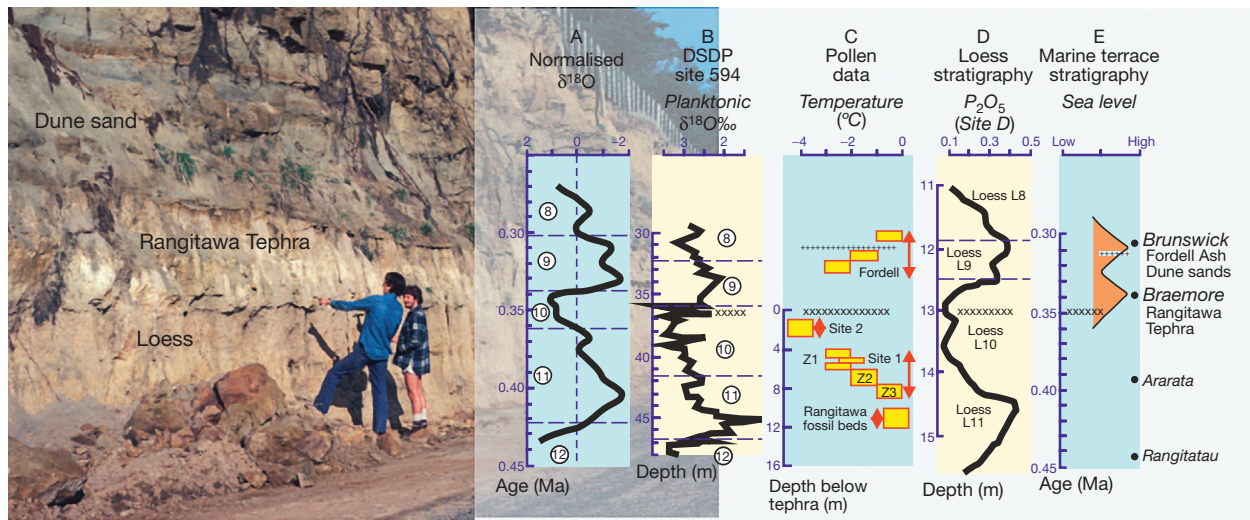


Figure 8 Stratigraphy and correlation of Rangitatau Tephra in marine and terrestrial sequences, New Zealand. Reproduced from Kohn BP, Pillans B, and McGlone MS (1992) Zircon fission track age for middle Pleistocene Rangitatau Tephra, New Zealand: Stratigraphic and paleoclimatic significance. *Palaeogeography, Palaeoclimatology, Palaeoecology* 95: 73–94. Photo shows Rangitatau Tephra underlying MIS 9 coastal sand dune at Mt Curl section, Wanganui Basin.

In the late 1970s and early 1980s, the tephras identified in Rangitatau East and Mt Curl sections in Wanganui Basin were considered to represent separate eruptions because their fission track ages were significantly different (~ 370 ka vs. ~ 240 ka, respectively). However, the two tephras had very similar field character, mineralogy, and glass shard chemistry and were nowhere identified in the same section. Were they the same tephra or were they two very similar tephras of different ages? This question, which is often asked in tephrostratigraphy, is not always easily answered for the very reason that tephras erupted from the same source can indeed be very similar in character. The question was answered when it was found that the early (pre-1981) glass ages were too young because of uncorrected track annealing and uncertainties in estimation

of neutron dose measurement, while the zircon ages were too young because of underetching (Kohn et al., 1992; Pillans et al., 1996). Subsequent glass fission track ages were corrected for track annealing using the isothermal plateau technique (Westgate, 1989), and zircons were redated with standardization of etching conditions and use of the zeta calibration technique to overcome uncertainties in decay constants and neutron fluences (Pillans et al., 1996).

The Matuyama/Brunhes Boundary

The last major polarity reversal of the Earth's magnetic field, known as the Matuyama/Brunhes polarity transition or reversal, has long been used as a major chronostratigraphic marker

in Quaternary studies. The Matuyama/Brunhes boundary (MBB) is widely identified in both marine and continental sequences and is also a key time marker for the chronology of human evolution (Pillans, 2003).

As summarized by Pillans (2003), the MBB is widely used to define the Lower–middle Pleistocene boundary ever since the Burg Wartenstein Symposium ‘Stratigraphy and Patterns of Cultural Change in the middle Pleistocene’ held in Austria in 1973 recommended that “The beginning of the middle Pleistocene should be so defined as to either coincide with or be linked to the boundary between the Matuyama Reversed Epoch and the Brunhes Normal Epoch of paleomagnetic chronology” (Butzer and Isaac, 1975).

The age of the MBB is independently established at ~780 ka from astronomical calibration of deep-sea isotope records (Tauxe et al., 1996) and $^{40}\text{Ar}/^{39}\text{Ar}$ dating of basaltic to andesitic lavas that were erupted during the polarity transition (Singer and Pringle, 1996). However, some age estimates are up to ~790 ka (e.g., Berger et al., 1995; Hou et al., 2000). While the 10 ka spread in age estimates might be partly explained by differences in intercalibration standards for $^{40}\text{Ar}/^{39}\text{Ar}$ dating (Renne et al., 1998), it should be noted that the MBB transition was not an instantaneous change in magnetic polarity but may have lasted several thousand years (Singer and Pringle, 1996; Tauxe et al., 1996). Thus, a single, precise age for the MBB is somewhat unrealistic and, indeed, is not necessary for many stratigraphic applications.

Well-defined paleomagnetic records in deep-sea sediments indicate that the MBB is located in interglacial Marine Isotope Stage 19 (Tauxe et al., 1996) – Figure 9. However, its inferred position in continental sediments is somewhat more variable, ranging from glacial MIS 20 to interglacial MIS 19 in Asian and European loess sequences. This discrepancy was analyzed by Zhou and Shackleton (1999), who concluded that the measured position of the MBB in loess sequences was misleading. Their conclusion was based on the stratigraphic position of a layer of

microtektites from the Australasian tektite field (see below), which occurred above the MBB in loess, but below the MBB in marine cores. They argued that there was a variable delay in remanence acquisition in the loess as a result of postdepositional weathering processes, which resulted in a downward displacement of the position of the MBB. More recently, Liu et al. (2008) have questioned the identification of Australasian microtektites in the Chinese loess and argued that the true position of the MBB in Chinese loess is in the upper part of soil S8, which in turn should be correlated with the upper part of MIS 19.

In strongly weathered, generally unfossiliferous Quaternary continental sequences in Australia, the MBB is one of the most important chronostratigraphic markers in the Lower–middle Pleistocene (Figure 9). However, the primary component of magnetic remanence is variably affected by chemical weathering, resulting in a secondary weathering-induced remanence that may completely or partially mask the primary component. Thus, there may be an apparent downward displacement in the position of the MBB as a result of Brunhes normal overprinting. In such settings, where strata are strongly weathered and secondary magnetic overprints are suspected, the presence of reversed polarity remanence provides a minimum age of 780 ka (Pillans, 2003).

Tektites

Tektites (from the Greek word ‘tektos,’ meaning molten) are glassy objects produced from melted crustal rocks that are caused by large hypervelocity meteorite impacts on sediments (Jackson, 1997). On impact, molten blobs of rock are ‘splashed’ into the atmosphere and cooled during flight to form shapes that include spheroid, elliptical, and dumbbell shapes, depending on the degree of rotation. Most tektites also have secondary surface features that are consistent with aerodynamic ablation during hypersonic flight.

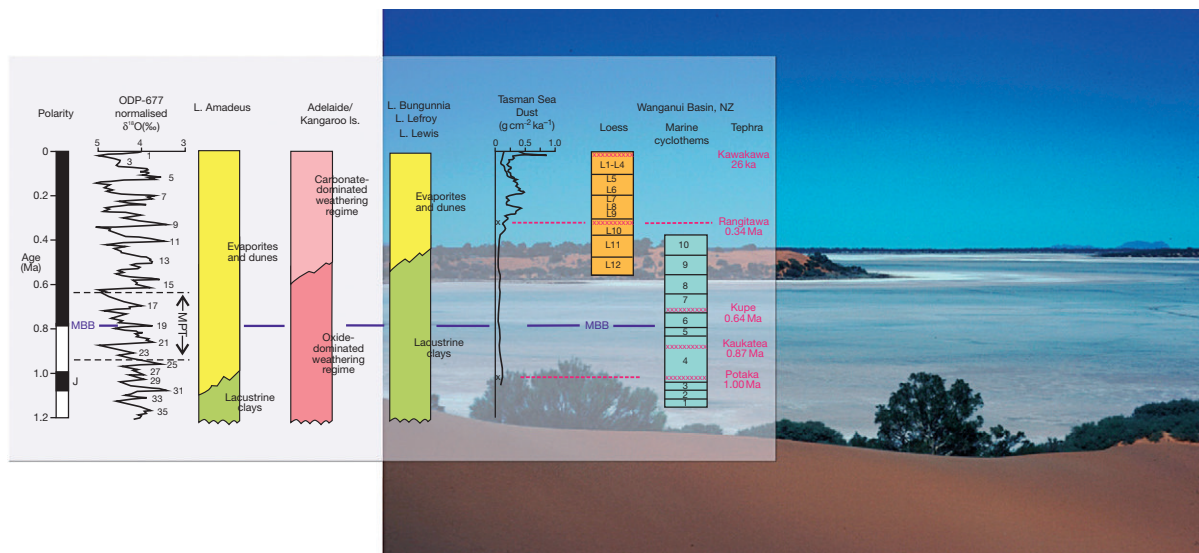


Figure 9 The Matuyama/Brunhes boundary (MBB) links marine and terrestrial sequences in Australasia, demonstrating that major climatic changes across the Mid-Pleistocene transition were broadly synchronous (redrawn from Pillans, 2003). However, the major arid shift was significantly earlier at Lake Amadeus (photo: J. Magee).

About 800 000 years ago, a large extraterrestrial impact occurred in southeast Asia, producing tektites, microtektites (<1 mm size), and impact debris, which are found over more than 10% of Earth's surface (Schnetzler and McHone, 1996), including much of Australia and surrounding oceans (Figure 10). The distribution, size, and concentration of tektites and microtektites indicate a likely impact site somewhere in southern Laos or Thailand, but the location of the impact site has not yet been discovered.

Like tephra, tektites are important chronostratigraphic markers, because they represent almost instantaneous events, can be directly dated by methods such as fission-track, K/Ar, and $^{40}\text{Ar}/^{39}\text{Ar}$, and can be correlated over large distances, including both marine and terrestrial deposits.

Although there has been considerable debate concerning the age of the Australasian tektites, their age is now firmly established through magnetostratigraphy of deep-sea cores, in which microtektites occur just prior to the Matuyama/Brunhes polarity transition (Schneider et al., 1992), as well as direct laser fusion $^{40}\text{Ar}/^{39}\text{Ar}$ dating of tektite glass (Izett, 1992; Hou et al., 2000). The most accurate and precise age for the tektites is the weighted mean age of 803 ± 3 ka reported by Hou et al. (2000). Previously, a number of field studies supported a late Pleistocene age in the range 5–25 ka (e.g., Lovering et al., 1972), but these occurrences are now considered to represent reworking into younger sediments (e.g., Fudali, 1993; Shoemaker and Uhlherr, 1999). There has long been speculation that the impact

may have triggered the M/B reversal. However, the estimated time (12–15 ka) between the impact and the reversal appears to be too long for a causal link between the two events. Impact glass (Darwin Glass) in western Tasmania, which yields an $^{40}\text{Ar}/^{39}\text{Ar}$ age of 816 ± 7 ka (Lo et al., 2002), appears to be from a separate impact during the same time period.

Even in deposits where reworking is suspected, the presence of tektites provides a maximum age of 800 ka for the enclosing sediment. For example, waterworn tektites have been found in diamond-bearing alluvial gravel terraces some 25 km downstream of the Argyle diamond mine in northwestern Australia (Fudali, 1993), indicating a maximum age of 800 ka for the terraces. Australasian microtektites are also present in Chinese loess sequences, where their presence as a discrete layer was used to correct the misleading position of the Matuyama/Brunhes polarity transition (Zhou and Shackleton, 1999) – Figures 10 and 11. Australasian tektites associated with Acheulean-like stone tools in South China yield an $^{40}\text{Ar}/^{39}\text{Ar}$ age of 803 ± 3 ka, thus making them the oldest known large cutting tools in East Asia (Hou et al., 2000).

Chinese Loess

The classic loess sequences of China (Figure 11) are arguably the most complete continental record of the Quaternary. Initially, magnetostratigraphy provided the overall chronology,

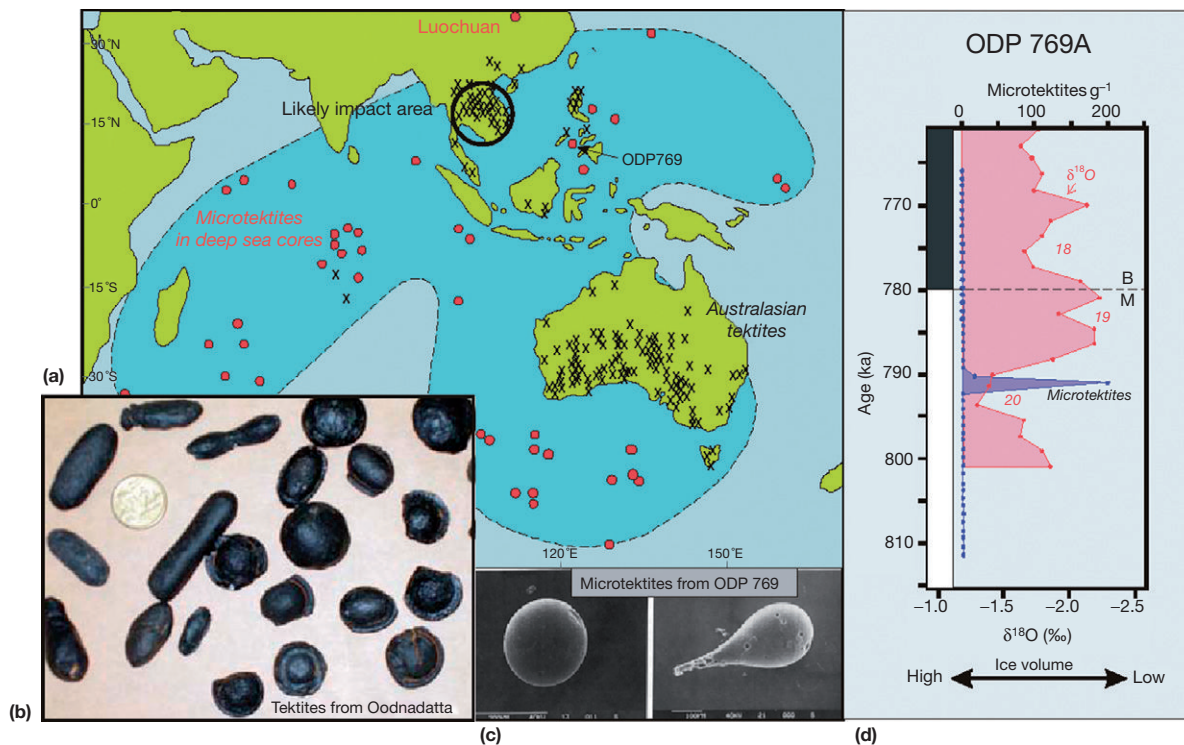


Figure 10 (a) The Australasian tektite field. Reproduced from Prasad MS, Gupta SM, and Kodagali VN (2003) Two layers of Australasian impact ejecta in the Indian Ocean? *Meteoritics and Planetary Science* 38: 1373–1381. (b) Tektites from the Oodnadatta area, central Australia. (c) Microtektites in deep-sea cores. (d) Chronology of microtektites in ODP core 690. (c,d) Reproduced from Schneider DA, Kent DV, and Mello, GA (1992) A detailed chronology of the Australasian impact event, the Brunhes-Matuyama geomagnetic polarity reversal, and global climatic change. *Earth and Planetary Science Letters* 111: 395–405.

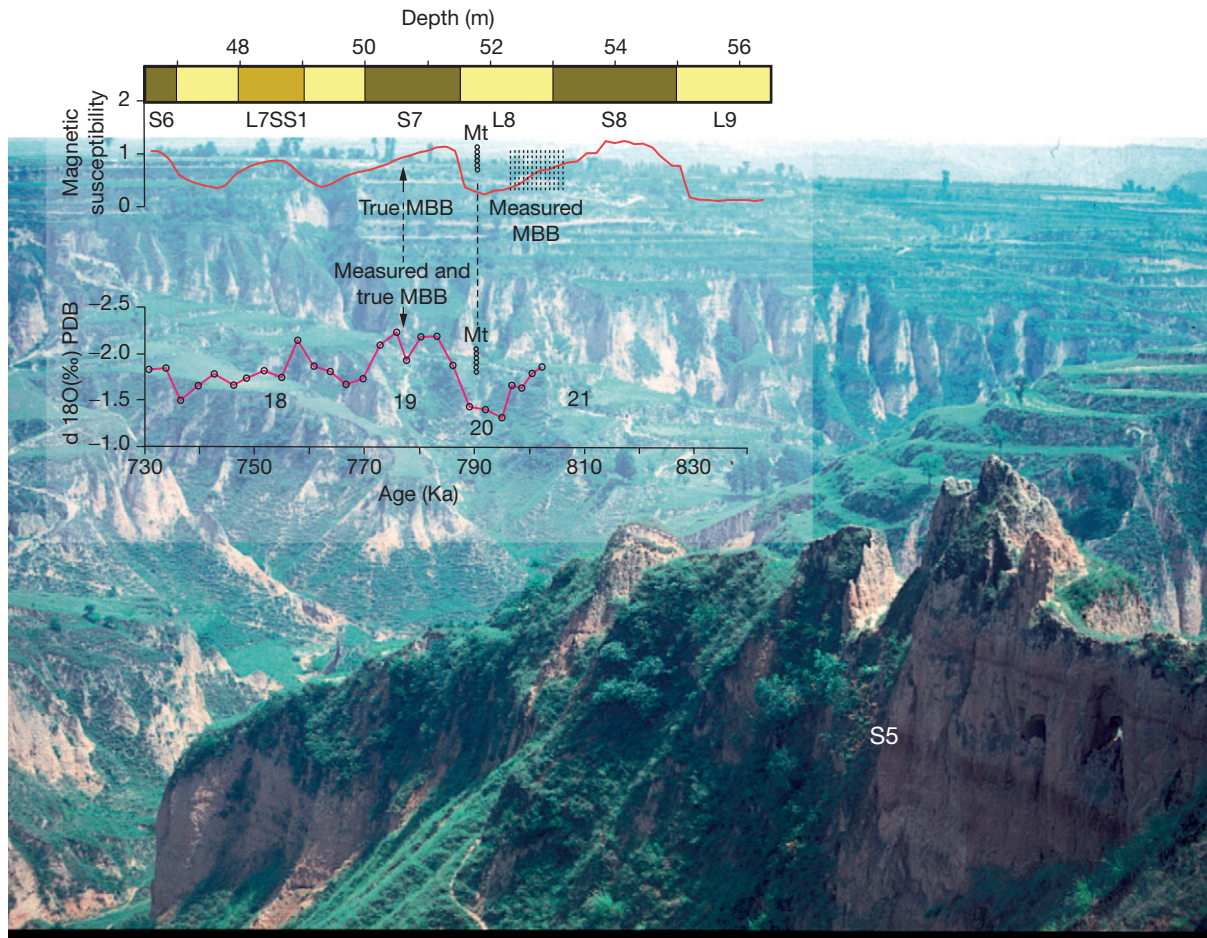


Figure 11 Loess/paleosol sequence at Luochuan section, China, and contrasting positions of the Matuyama/Brunhes boundary in Luochuan loess versus deep-sea cores. Reproduced from Zhou LP and Shackleton NJ (1999) Misleading positions of geomagnetic reversal boundaries in Eurasian loess and implications for correlation between continental and marine sedimentary sequences. *Earth and Planetary Science Letters* 168: 117–130.

which was later supplemented by a higher-resolution time scale based on magnetic susceptibility variations (see Kukla (1987) and references therein). A cycle-by-cycle correlation of the Chinese loess/paleosol sequence, down to the level of the Red Clay (~2.6 Ma), with the marine isotope record was first accomplished by Kukla (1987). Hovan et al. (1989) established direct links between Chinese loess and the marine isotope record by measuring the dust content of a marine core in the North Pacific. Most recently, Sun et al. (2006) have tuned the grain size records of two Chinese loess sections to orbital parameters to provide an independent astronomical timescale for the loess records, allowing improved correlation to the marine isotope record.

Pollen Analysis of Marine Sediments

Reconstruction of vegetation changes based on pollen analysis has long played a central role in reconstructing Quaternary climates. However, while detailed ice-core and marine isotope records, in particular, have provided new insights into global climate changes, there is a dearth of well-dated, long, continuous terrestrial pollen records that can be used to examine the detailed phase relationships between these and other

paleoclimate time series. Commonly, the chronology for terrestrial pollen records has been established by transferring the marine isotope timescale, on the assumption that there is synchronicity of certain events between the two records (e.g., Tzedakis et al., 1997), but clearly the best approach is to directly analyze both pollen and oxygen isotopes in the same marine sediment core (e.g., Heusser and Shackleton, 1979; Mildenhall et al., 2004).

A high-resolution pollen and oxygen isotope record of the Last Interglacial from deep-sea core MD95-2042, located off the coast of Portugal, indicates that the full interglacial conditions on land (as evidenced by expansion of forest vegetation) lasted from 126 to 110 ka, whereas full interglacial marine conditions (MIS 5e) lasted from 132 to 115 ka (Shackleton et al., 2003). In other words, there is an implied lag of 5–6 ka between the terrestrial and marine records at this site. Tzedakis (2005) has noted that during the last two interglacials, major tree population expansion in southern Europe was closely associated with the peak in Northern Hemisphere June insolation at 65°N, and less influenced by the timing of deglaciation. Given that different vegetation communities are limited by different bioclimatic parameters, this association is unlikely to be universally applicable.

A 350 ka high-resolution pollen record from DSDP Site 594 (Figure 6), ~250 km east of the South Island of New Zealand, was obtained by Heusser and van de Geer (1994). Their study was significant because it demonstrated that, within the limits of sampling resolution (~2.4 ka), New Zealand terrestrial vegetation changes covaried with the global oxygen isotope record from the same core. In other words, Northern and Southern Hemisphere climate records were essentially synchronous. Nelson et al. (1985) had previously reached the same general conclusion from their study of lithological variations in the core from DSDP Site 594 – increased terrigenous silt inputs from expanded glaciers in the South Island corresponded with times of enriched $\delta^{18}\text{O}$, indicating that alpine glaciations in New Zealand were in phase (± 3 ka) with global climate changes over the past 750 ka.

Conclusion

The stratigraphy and correlation of continental glacial deposits dominated Quaternary science for many years, culminating in the widespread application of the so-called four glaciation model, which was based on classical glacial sequences in the European Alps. When oxygen isotope stratigraphy of deep-ocean cores provided evidence for many more glacial intervals in the Quaternary, the incompleteness of the continental record was revealed. A lack of reliable ages beyond the range of radiocarbon dating also imposed limitations on glacial

stratigraphy on-land, whereas the glacial cycles in marine records were more readily dated.

The rise to primacy of the marine isotope stages (MIS) as the global template for Quaternary stratigraphy can be traced to two seminal papers. The first was the study of oxygen isotope variations in core V28-238 from the western equatorial Pacific (Figure 12) by Shackleton and Opdyke (1973), who famously predicted that ‘it is highly unlikely that any superior stratigraphic subdivision of the Pleistocene will ever emerge.’ The second was the paper by Hays et al. (1976), which demonstrated the control of orbital perturbations and paved the way for astronomical tuning.

Another significant paleoclimatic archive that has become available in recent years is the highly detailed ice cores from Antarctica, Greenland, and various smaller ice caps. The longest and oldest ice cores are from Antarctica and extend back over 800 000 years (EPICA Community Members, 2004; Luthi et al., 2008). A detailed study of such cores is revolutionizing our understanding of global climate change in the Quaternary, in much the same way that deep-sea cores did previously.

Ice cores and deep-sea cores notwithstanding, Quaternary stratigraphy of continental sequences remains central to the discipline of Quaternary research, including the history of human evolution (cf. deMenocal, 2004). One challenge remains, and that is, to establish precise correlations between marine, ice, and continental archives using all methods of stratigraphic correlation. The global correlation chart (Figure 13) and recent results from precisely dated speleothems demonstrate that this challenge is being met (e.g., Cheng et al., 2009; Wang et al., 2001).

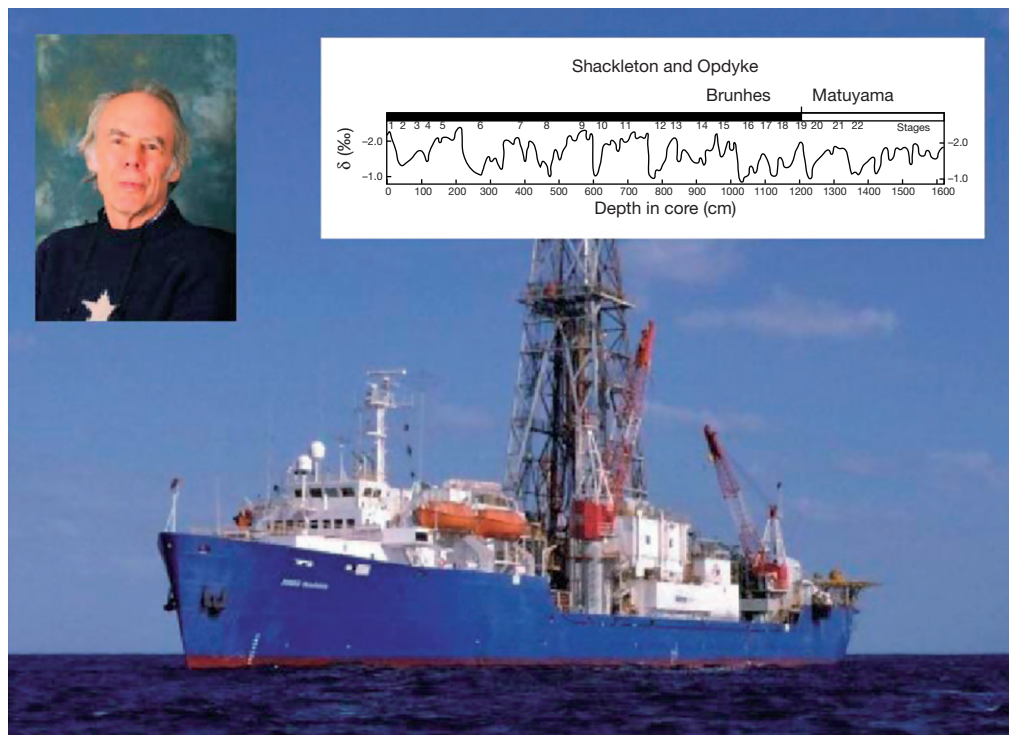


Figure 12 Analysis of deep-sea sediment cores, such as core V28-238 (Shackleton and Opdyke, 1973) revolutionized Quaternary stratigraphy in the 1960s and 1970s. Sir Nicholas Shackleton (1937–2006) was a leader in this research. Photo of Joides Resolution drilling ship courtesy of International Ocean Drilling Program, Texas A&M University.

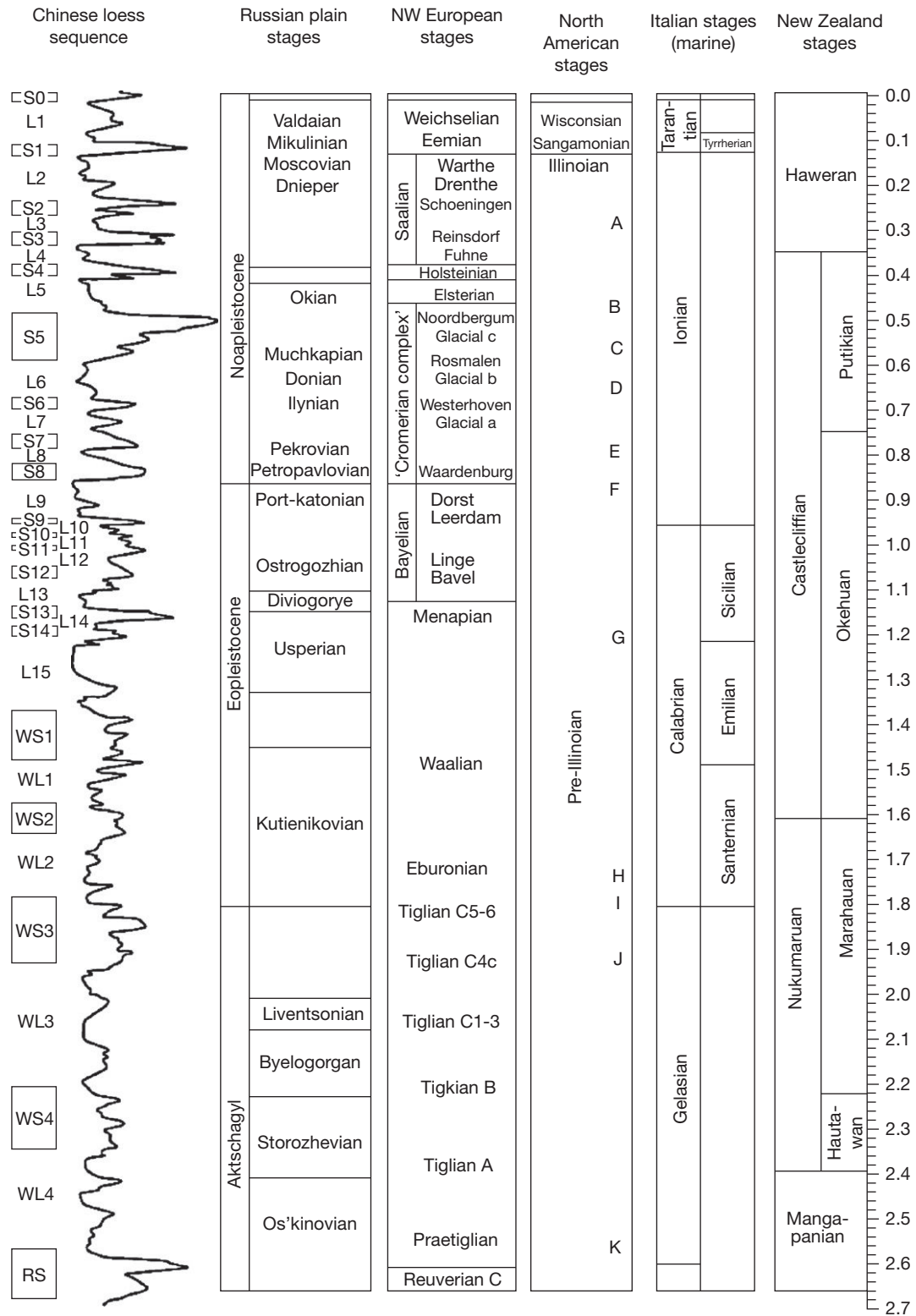


Figure 13 (Continued)

See also: Amino Acid Dating; U-Series Dating. **Glacial Landforms:** Glacitectonic Structures and Landforms; Introduction; Quaternary Vulcanism: Subglacial Landforms. **Glacial Landforms, Erosional Features:** Major Scale Forms; Micro- to Macroscale Forms. **Glacial Landforms, Ice Sheets:** Evidence of Glacier and Ice Sheet Extent; Growth and Decay. **Paleoceanography, Physical and Chemical Proxies:** Oxygen-Isotope Stratigraphy of the Oceans. **Quaternary Stratigraphy:** Climatostratigraphy; Continental Biostratigraphy; Lithostratigraphy; Morphostratigraphy/Allostratigraphy; Pedostratigraphy; Sequence Stratigraphy; Tephrochronology. **Radiocarbon Dating:** Variations in Atmospheric ^{14}C .

References

- Aguirre E and Pasini G (1985) The Pliocene-Pleistocene boundary. *Episodes* 8: 116–120.
- Bada JL, Gillespie R, Gowlett JAJ, and Hedges REM (1984) Accelerator mass spectrometry radiocarbon ages of amino acid extracts from Californian palaeoindian skeletons. *Nature* 312: 442–444.
- Bada JL, Schroeder RA, and Carter GF (1974) New evidence for the antiquity of man in North America deduced from aspartic acid racemization. *Science* 184: 791–793.
- Barbetti M and Colhoun EA (1988) Reversed magnetisation of glaciolacustrine sediments from western Tasmania. *Search* 19: 151–153.
- Berger WH, Bickert T, Wefer G, and Yasuda MK (1995) Brunhes-Matuyama boundary: 790 ky date consistent with ODP Leg 130 oxygen isotope records based on fit to Milankovitch template. *Geophysical Research Letters* 22: 1525–1528.
- Beu AG and Edwards AR (1984) New Zealand Pleistocene and Late Pliocene glacio-eustatic cycles. *Palaeogeography, Palaeoclimatology, Palaeoecology* 46: 119–142.
- Bowen DQ, Hughes S, Sykes GA, and Miller GH (1989) Land-sea correlations in the Pleistocene based on isoleucine epimerization in non-marine molluscs. *Nature* 340: 49–51.
- Buizer KW and Isaac GL (eds.) (1975) *After the Australopithecines* p. 911. The Hague: Mouton.
- Chappell J, Head J, and Magee J (1996) Beyond the radiocarbon limit in Australian archaeology and Quaternary research. *Antiquity* 70: 543–552.
- Cheng H, Edwards RL, Broecker WS, et al. (2009) Ice age terminations. *Science* 326: 248–252.
- Colhoun EA (1985) Glaciations of the West Coast Range, Tasmania. *Quaternary Research* 24: 39–59.
- deMenocal PB (2004) African climate change and faunal evolution during the Pliocene-Pleistocene. *Earth and Planetary Science Letters* 220: 3–24.
- Easterbrook DJ, Briggs ND, Westgate JA, and Gorton MP (1981) Age of the Salmon Springs Glaciation in Washington. *Geology* 9: 87–93.
- Eggins SM, Grun R, McCulloch MT, et al. (2005) In situ U-series dating by laser-ablation multi-collector ICPMS: New prospects for Quaternary geochronology. *Quaternary Science Reviews* 24: 2523–2538.
- EPICA Community Members (2004) Eight glacial cycles from an Antarctic ice core. *Nature* 429: 623–628.
- Fudali RF (1993) The stratigraphic age of australites revisited. *Meteoritics* 28: 114–119.
- Gibbard P, Head MJ, Walker M, et al. (2010) Formal ratification of the Quaternary System/Period and the Pleistocene Series/Epoch with a base at 2.588 Ma. *Journal of Quaternary Science* 25: 96–102.
- Gill ED (1956) Radiocarbon dating for glacial varves in Tasmania. *The Australian Journal of Science* 19: 80.
- Grün R, Eggins S, Aubert M, Spooner N, Pike AWG, and Müller W (2010) ESR and U-series analyses of faunal material from Cuddie Springs, NSW, Australia: Implications for the timing of the extinction of the Australian megafauna. *Quaternary Science Reviews* 29: 596–610.
- Hays JD, Imbrie J, and Shackleton NJ (1976) Variations in the Earth's orbit: Pacemaker of the ice ages. *Science* 194: 1121–1132.
- Hearty PJ, Miller GH, Stearns CE, and Szabo BJ (1986) Aminostratigraphy of Quaternary shorelines in the Mediterranean basin. *Geological Society of America Bulletin* 97: 850–858.
- Heusser L and Shackleton NJ (1979) Direct marine-continental correlation: 150,000-year oxygen isotope-pollen record from the North Pacific. *Science* 204: 837–839.
- Heusser LE and van de Geer G (1994) Direct correlation of terrestrial and marine paleoclimate records from four glacial-interglacial cycles – DSDP Site 594 Southwest Pacific. *Quaternary Science Reviews* 13: 273–282.
- Hou Y, Potts R, Yuan B, et al. (2000) Mid-Pleistocene Acheulean-like stone technology of the Bose Basin, South China. *Science* 287: 1622–1626.
- Hovan SA, Rea DK, Pisias NG, and Shackleton NJ (1989) A direct link between the China loess and marine $\delta^{18}\text{O}$ records: Aeolian flux to the north Pacific. *Nature* 340: 296–298.
- Izett GA and Obradovich JD (1992) Laser-fusion $^{40}\text{Ar}/^{39}\text{Ar}$ ages of Australasian tektites (Abstract). *Lunar and Planetary Science* 23: 593–594.
- Jackson JA (ed.) (1997) *Glossary of Geology*, p. 769. Alexandria: American Geological Institute.
- Johnson CN (2005) What can the data on late survival of Australian megafauna tell us about the cause of their extinction? *Quaternary Science Reviews* 24: 2167–2172.
- Kaufman A, Broecker WS, Ku T-L, and Thurber DL (1971) The status of U-series methods of mollusk dating. *Geochimica et Cosmochimica Acta* 35: 1155–1189.
- King WBR and Oakley KP (1949) Definition of the Pliocene-Pleistocene boundary. *Nature* 163: 186–187.
- Kohn BP, Pillans B, and McGlone MS (1992) Zircon fission track age for middle Pleistocene Rangitawa Tephra, New Zealand: Stratigraphic and paleoclimatic significance. *Palaeogeography, Palaeoclimatology, Palaeoecology* 95: 73–94.
- Kukla G (1987) Loess stratigraphy in central China. *Quaternary Science Reviews* 6: 191–219.
- Liu Q, Roberts AP, Rohling EJ, Zhu R, and Sun Y (2008) Post-depositional magnetization lock-in and the location of the Matuyama-Brunhes geomagnetic reversal boundary in marine and Chinese loess sequences. *Earth and Planetary Science Letters* 275: 102–110.
- Lo CH, Howard KT, Chung SL, and Meffre S (2002) Laser fusion argon-40/argon-39 ages of Darwin impact glass. *Meteoritics and Planetary Science* 37: 1555–1562.
- Lourens LJ, Antonarakou A, Hilgen FJ, Van Hoof AAM, Vergnaud-Grazzini C, and Zachariasse WJ (1996) Evaluation of the Plio-Pleistocene astronomical timescale. *Palaeogeography* 11(4): 391–413.
- Lovering JF, Mason B, Williams GE, and McColl DH (1972) Stratigraphical evidence for the terrestrial age of australites. *Journal of the Geological Society of Australia* 18: 409–418.
- Luthi D, Le Floch M, Bereiter B, et al. (2008) High-resolution carbon dioxide concentration record 650,000–800,000 years before present. *Nature* 453: 379–382.
- Machida H (2002) Quaternary volcanoes and widespread tephra of the world. *Global Environmental Research* 6: 3–17.
- Mildenhall DC, Hollis CJ, and Naish TR (2004) Orbitally-influenced vegetation record of the Mid-Pleistocene Climate Transition, offshore eastern New Zealand (ODP Leg 181, Site 1123). *Marine Geology* 205: 87–111.
- Miller GH, Fogel ML, Magee JW, Gagan MK, Clarke SJ, and Johnson BJ (2005) Ecosystem collapse in Pleistocene Australia and a human role in megafaunal extinction. *Science* 309: 287–290.
- Miller GH, Magee JW, Johnson BJ, et al. (1999) Pleistocene extinction of *Genyornis newtoni*: Human impact on Australian megafauna. *Science* 283: 205–208.
- Murray-Wallace CV (1995) Aminostratigraphy of Quaternary coastal sequences in southern Australia – an overview. *Quaternary International* 26: 69–86.
- Murray-Wallace CV, Beu AG, Kendrick GW, Brown LJ, Belperio AP, and Sherwood JE (2000) Palaeoclimatic implications of the occurrence of the arcoid bivalve *Anadara trapezia* (Deshayes) in the Quaternary of Australasia. *Quaternary Science Reviews* 19: 559–590.
- Naish TR, Abbott ST, Alloway BV, et al. (1998) Astronomical calibration of a Southern Hemisphere Plio-Pleistocene reference section, Wanganui Basin, New Zealand. *Quaternary Science Reviews* 17: 695–710.
- Nelson CS, Hendy CH, Jarrett GR, and Cuthbertson AM (1985) Near-synchronicity of New Zealand alpine glaciations and Northern Hemisphere continental glaciations during the past 750 kyr. *Nature* 318: 361–363.
- Palmer AS and Pillans B (1996) Record of climatic fluctuations from ca. 500 ka loess deposits and paleosols near Wanganui, New Zealand. *Quaternary International* 34–36: 155–162.
- Pillans B (1983) Upper Quaternary marine terrace chronology and deformation, South Taranaki, New Zealand. *Geology* 11: 292–297.
- Pillans B (1994a) Direct marine-terrestrial correlations, Wanganui Basin, New Zealand: The last 1 million years. *Quaternary Science Reviews* 13: 189–200.
- Pillans B (1994b) New Zealand Quaternary stratigraphy: Integrating marine and terrestrial records of environmental change. *Journal of Geography (Japan)* 103: 760–769.
- Pillans B (2003) Subdividing the Pleistocene using the Matuyama-Brunhes boundary (MBB): An Australasian perspective. *Quaternary Science Reviews* 22: 1569–1577.

- Pillans B, Alloway B, Naish T, Westgate J, Abbott S, and Palmer A (2005) Silicic tephra in Pleistocene shallow-marine sediments of Wanganui Basin, New Zealand. *Journal of the Royal Society of New Zealand* 35: 43–90.
- Pillans B, Kohn BP, Berger G, et al. (1996) Multi-method dating comparison for Mid-Pleistocene Rangitawa Tephra, New Zealand. *Quaternary Science Reviews* 15: 641–653.
- Renne PR, Swisher CC, Deino AL, Karner DB, Owens TL, and DePaolo DJ (1998) Intercalibration of standards, absolute ages and uncertainties in $^{40}\text{Ar}/^{39}\text{Ar}$ dating. *Chemical Geology* 145: 117–152.
- Rio D, Sprovieri R, Castradori D, and Di Stephano E (1998) The Gelasian Stage (Upper Pliocene): A new unit of the global standard chronostratigraphic scale. *Episodes* 21: 82–87.
- Roberts RG, Flannery TF, Ayliffe LK, et al. (2001) New ages for the last Australian megafauna: Continent-wide extinction about 46,000 years ago. *Science* 292: 1888–1892.
- Salvador A (ed.) (1994) *International Stratigraphic Guide* p. 214. Boulder: Geological Society of America.
- Schneider DA, Kent DV, and Mello GA (1992) A detailed chronology of the Australasian impact event, the Brunhes–Matuyama geomagnetic polarity reversal, and global climatic change. *Earth and Planetary Science Letters* 111: 395–405.
- Schnetzler CC and McHone JF (1996) Source of Australasian tektites: Investigating possible impact sites in Laos. *Meteoritics and Planetary Science* 31: 73–76.
- Shackleton NJ and Opdyke ND (1973) Oxygen isotope and paleomagnetic stratigraphy of equatorial Pacific core V28-238: Oxygen isotope temperatures and ice volumes on a 10^5 to 10^6 scale. *Quaternary Research* 3: 39–55.
- Shackleton NJ, Sanchez-Goni MF, Pailler D, and Lancelot Y (2003) Marine Isotope Substage 5e and the Eemian Interglacial. *Global and Planetary Change* 36: 151–155.
- Shoemaker EM and Uhlherr HR (1999) Stratigraphic significance of australites in the Port Campbell Embayment, Victoria. *Meteoritics and Planetary Science* 34: 369–384.
- Singer BS and Pringle MS (1996) Age and duration of the Matuyama–Brunhes geomagnetic polarity reversal from $^{40}\text{Ar}/^{39}\text{Ar}$ incremental heating analyses of lavas. *Earth and Planetary Science Letters* 139: 47–61.
- Stirling CH, Esat TM, Lambeck K, and McCulloch MT (1998) Timing and duration of the Last Interglacial: Evidence for a restricted interval of widespread coral reef growth. *Earth and Planetary Science Letters* 160: 745–762.
- Stuiver M, Heusser CT, and Yang IC (1978) North American glacial history back to 75,000 years BP. *Science* 200: 16–21.
- Sun Y, Clemens SC, An Z, and Yu Z (2006) Astronomical timescale and paleoclimatic implications of stacked 3.6-Myr monsoon records from the Chinese Loess Plateau. *Quaternary Science Reviews* 25: 33–48.
- Tauxe L, Herbert T, Shackleton NJ, and Kok YS (1996) Astronomical calibration of the Matuyama–Brunhes boundary: Consequences for magnetic remanence acquisition in marine carbonates and the Asian loess sequences. *Earth and Planetary Science Letters* 140: 133–146.
- Thom BG (1973) The dilemma of high interglacial sea-levels during the Last Glaciation. *Progress in Physical Geography* 5: 170–246.
- Tzedakis PC (2005) Towards an understanding of the response of southern European vegetation to orbital and suborbital climate variability. *Quaternary Science Reviews* 24: 1585–1599.
- Tzedakis PC, Andrieu V, de Beaulieu J-L, et al. (1997) Comparison of terrestrial and marine climate records of changing climate of the last 500,000 years. *Earth and Planetary Science Letters* 150: 171–176.
- Wang YJ, Cheng H, Edwards RL, et al. (2001) A high-resolution absolute-dated late Pleistocene monsoon record from Hulu Cave, China. *Science* 294: 2345–2348.
- Westgate JA (1989) Isothermal plateau fission track ages of hydrated glass shards from silicic tephra beds. *Earth and Planetary Science Letters* 95: 226–234.
- Zhou LP and Shackleton NJ (1999) Misleading positions of geomagnetic reversal boundaries in Eurasian loess and implications for correlation between continental and marine sedimentary sequences. *Earth and Planetary Science Letters* 168: 117–130.

Continental Biostratigraphy

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Introduction

One of the most debated questions in geological sciences is what is the age. The age of a particular event, the age of a remarkable discovery, the age of a glaciations, the age of the hominines dispersal into Europe, or the extinction of, for instance, the woolly mammoth. There are different ways to solve questions about the age: to date the fossil record. A classical, widely used method is biostratigraphy, that is, the science of dating rocks and sediments using animal and plant fossils. The fact that during the Quaternary the continental flora and fauna almost constantly changed means that biostratigraphy for dating provides a good basis for correlation of stratigraphic sequences or events at one locality to those at another locality, for example, for correlation between Europe and Asia. However, biostratigraphy is a relative dating method and must be independently calibrated with numerical dating methods before numerical ages can be assigned to biostratigraphic zones or events.

The Quaternary Changes in Flora and Fauna

The rapid and continuous climatic changes during the Quaternary induced important changes in environmental conditions and available habitats. This is particularly the case in higher latitudes or in areas such as the western part of the North European plain, where the conditions are less diverse, in comparison to mountainous areas such as central Italy or the Balkan Peninsula. The forest vegetation, which dominates the flora of northwestern Europe during interglacial phases, disappeared during the cold stages and was replaced by the so-called Mammoth Steppe, treeless vegetation types which are 'analogous but not necessarily identical in plant composition to present-day's tundra, steppe, and alpine meadows.' The climatic changes that characterize the past 2.5 Ma of our geological history had a huge impact on the composition of the flora and fauna and on the distribution of species in particular at the higher latitudes. The composition of the Quaternary flora and fauna had undergone changes due to (a) the evolution of species, (b) the extinction of species, and (c) the dispersal or migration of species.

The Evolution of Species

The evolution of species is a natural phenomenon that can be observed in the entire geological history of the earth. The idea of evolution has been put forward by, for example, Darwin (1809–82) and Wallace (1823–1913) and are extensively described by Darwin in *The Origin of Species* (full title *On the Origin of Species by Means of Natural Selection, or the Preservation of Favoured Races in the Struggle for Life*) published on

24 November 1859. As stated by both scientists, natural selection is the base for evolution. Natural selection can take place if:

- (1) there is more offspring that can grow to adulthood;
- (2) there will be a struggle for survival among individuals;
- (3) in sexually reproducing species, no two individuals are identical. Variation is rampant;
- (4) much of this variation is heritable.

From this, one can infer that in a world of stable populations where each individual must struggle to survive, those with the 'best' characteristics will be more likely to survive, and those desirable traits will be passed to their offspring; and that these advantageous characteristics are inherited by following generations, becoming dominant among the population through time. This is natural selection that leads to the evolution of species that can be observed in the Quaternary fossil record.

Many different species show evolutionary changes during the past 2.5 Ma. However, there are also species with a long stratigraphical range that hardly evolve or do not evolve at all. This applies to nearly all plant species but also to, for example, mollusks, insects, amphibians, and, reptiles. Among the Quaternary mammals there are species that do not show clear evolutionary changes such as the Chiroptera or bats. Another example is the pygmy shrew *Sorex minutus*, a species that only shows some fluctuations in size during the past 2.5 Ma. The pygmy shrew is widespread in Europe and lives under different climatic conditions. It inhabits areas with good ground cover although it is uncommon in woodland. The Pleistocene climatic and inherent environmental changes influenced the geographical distribution of the species but hardly affected the morphology of the dentition and the skeleton of the species. The fossil record of the species does not show a clear evolution. This applies also to a number of forest dwellers such as the wood mouse *Apodemus sylvaticus*, squirrels (e.g., the red squirrel *Sciurus vulgaris*) and glirids (e.g., the garden dormouse *Eliomys quercinus* and the common dormouse *Muscardinus avellanarius*).

Fortunately, there are a large number of species that clearly evolve even in a relative short period as the Quaternary. The evolution is in some cases rather rapid, in others rather slow and not all the evolutionary changes occur in the same period of time. Some species mainly evolve during the early Pleistocene whereas others show clear changes during a later phase of the Quaternary. Remarkable is the fact that many species show comparable evolutionary changes. A general feature in the evolution is:

- (1) the increase in the height of the crown of the premolars; and
- (2) the increase of complexity of the enamel pattern of the premolars.

Both features are related to the adaptation to more abrasive nutrition. The deterioration of the climate led to the reduction

of woodland vegetation and an expansion of a more open habitat with a dominance of grasses. For many herbivores, this implies a change to more abrasive food.

Change in size is another feature that can be observed in a number of Quaternary species. Pleistocene wolverines show a remarkable increase in size during the middle Pleistocene. The same, however, less extensive, can be seen in wolves. Horses, on the other hand, increase in size during the early part of the middle Pleistocene whereas they reduce in size from the later part of the middle Pleistocene onwards. Changes in size are generally not unidirectional and are, therefore, not a good feature for biostratigraphical purposes. It can only be used in combination with other data.

Among the species that show a clear evolution during the Quaternary and are, hence, biostratigraphically relevant are the mammoths belonging to the lineage *Mammuthus rumanus*–*Mammuthus meridionalis*–*Mammuthus trogontherii*–*Mammuthus primigenius* and a large number of voles representing different vole-lineages.

Mammoth Evolution

Many paleontologists investigated the patterns and processes in the evolution of the Eurasian elephantids or mammoths and is hence, one of the best known lineages. Mammoth evolution began in Africa and during the late Pliocene mammoths migrated to Eurasia. The late Pliocene and Pleistocene fossil record indicates that the Eurasian mammoth underwent very significant changes: a shortening and heightening of the cranium and mandible (Figure 1), an increase in the relative height of the molars, an increase in the number of plates and thinning of the dental enamel (Lister et al., 2005). The earliest representatives of the European mammoth *Mammuthus rumanus* from the interval 3.5–2.5 Ma is characterized by upper M³ molars with 8–10 plates, a hypsodonty index (i.e., the height of the crown divided by the width × 100) of about 120 and an average thickness of the enamel of about 3.8 mm. The upper M³ of the most advanced representative of the mammoth lineage *Mammuthus primigenius* has 20–28 plates, an average hypsodonty index of about 200 and enamel with a thickness that ranges between 1.0 and 2.5 mm with an mean of about 1.6 mm.

Previously it was assumed that the transformation within the *Mammuthus rumanus*–*Mammuthus meridionalis*–*Mammuthus trogontherii*–*Mammuthus primigenius* lineage was gradual and more or less simultaneous across the species' range. However, recent investigations of the geographical variation across the whole of northern Eurasia indicated a more complex model (Lister et al., 2005). The transition between *Mammuthus rumanus* and *Mammuthus meridionalis* is still poorly known; the number of late Pliocene fossil remains is too limited. The early- and early middle Pleistocene *Mammuthus meridionalis* is better known. The species was widely dispersed in Eurasia and migrated to America during the early Pleistocene. The European fossil record indicated that the species hardly evolved despite the fact that the species occurred in the region for almost 2 Ma. In eastern Asia, probably in China, however, *Mammuthus meridionalis* evolved into *Mammuthus trogontherii* in the interval 2.0–2.5 Ma. *Mammuthus trogontherii* expanded its original range and spread to NE Siberia by 1.2 Ma and toward Europe around 1 Ma. In NE Siberia *Mammuthus trogontherii* began a transformation into *Mammuthus primigenius* as early as 700 ka. The European population of *Mammuthus trogontherii* also hardly evolved and only shows a reduction of size in the interval between 1 Ma and 200 ka. *Mammuthus primigenius* with its origin in NE Siberia invaded Europe during the late middle Pleistocene about 200 ka ago.

The Evolution of Voles

Rodents are well represented in the present-day fauna; nearly 40% of all living mammalian species are rodents. They represent a huge variety of animals including, for example, beavers, squirrels, marmots, hamsters, rats, mice, voles, lemmings, dormice, and porcupines. Voles are a rather recent branch of diminutive grazers in the rodent evolution. They arose from the hamster lineage during the late Pliocene and are with the about 100 extant species very successful since, in particular during the Quaternary. The evolution of the voles shows the occurrence of radiations leading to a number of different lineages. All these lineages show in principle the same type of evolution.

The first changes that can be observed are an increase in height of the crown. The brachiodont molars develop into hypsodont molars with high enamel-free areas or dentine

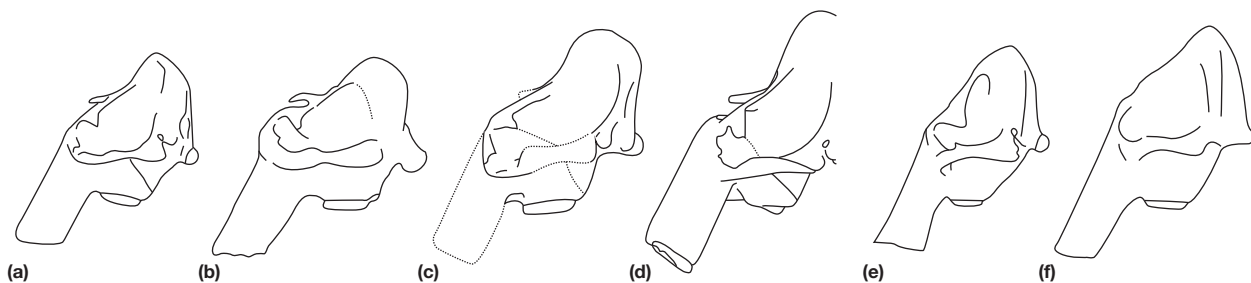


Figure 1 *Mammuthus* crania from (a) Liventsovka, Russia, ~2.3 Ma, type locality of '*M. gromovi*' (after Azzaroli, 1977); (b) Chilhac, France (*M. meridionalis*, ~1.9 Ma, after Boeuf, 1990); (c) Upper Valdarno, Italy (*M. meridionalis* type locality, ~1.8 Ma, after Azzaroli, 1966); (d) Scoppito, Italy ('*M. meridionalis vestinus*', ~1.2 Ma, after Maccagno, 1962); (e) Cherny Yar, Russia, type of '*M. trogontherii chosaricus*,' after Dubrovo, 1966); (f) Debica, Poland (typical *M. primigenius*, after Kubiak, 1980). Scale bar 50 cm. Note the apparent increase in cranium height between (b) and (c). Reproduced from Lister AM, Sher A, van Essen H, and Wei G (2005) The pattern and process of mammoth evolution in Eurasia. *Quaternary International* 126–128: 49–64.

tracts. The increase of height of the crowns leads to a delay in the formation of roots, an evolution that finally results in voles with molars without roots. A second feature that can be observed in the evolution of voles is the development of crown cement in a number of the lineages. Crown cement occurs in the reentrant angles of the hypsodont molars. In addition, different lineages show an increase in the length as well as an increase in complexity of the posterior upper molar (M^3) and the anterior lower molar (M_1).

Another general feature that can be observed in different lineages are changes in the differentiation of the thickness of the enamel. The occlusal surface of the molars of voles shows a number of dentine field covered with enamel at the outer side. The anterior enamel edge of the salient angles of the lower molars is concave, the posterior convex. The concave edges occlude first during the longitudinal movement of the mastication; they are the so-called leading edges. The convex edges are the trailing edges. In a large number of living species, the enamel of the trailing edges is much thinner than that of the leading edges. The opposite can be observed in many more primitive species. The transition of trailing edges that are thicker to a situation in which the leading edges are relatively thicker is called the evolution in the differentiation in the thickness of the enamel.

The evolutionary changes described above do not occur simultaneously in the different lineages or within the lineages in the same region. The evolution of voles is a complicated story, studied by a large number of paleontologists. Because of the geographically wide range of the species, the often dominant occurrence in the Quaternary fossil record and the extremely quick evolution that can be observed in a number of lineages, voles are extremely important for Quaternary continental stratigraphy. In particular for Eurasia, the *Mimomys savini*, *Arvicola terrestris* lineage is important for the subdivision of the middle and late Pleistocene. *Mimomys savini*, a vole with rooted molars, is the ancestor of the living Water Vole *A. terrestris*. The species occurred in the European faunas during the late early and the early middle Pleistocene. The transition of populations of water voles with rooted molars to populations with more hypsodont, unrooted molars referred to the genus *Arvicola*, took place during the first half of the middle Pleistocene. This transition seems to be well established since populations with a small percentage of rooted molars are known from several localities in Germany, Italy, Czechia, and Russia. The genus *Arvicola* shows clearly a gradual evolution in the differentiation in the thickness of the enamel since its appearance during the early middle Pleistocene. The enamel differentiation appeared to be an important marker to indicate the evolutionary stage of *Arvicola* and hence the relative age of the fossil record.

The *Microtus (Allophaiomys)*–*Microtus (Microtus)* lineage is important for the biozonation of the later part of the early Pleistocene. The molars of that well-known lineage are already without roots. They first of all show changes in the differentiation of the thickness of the enamel and later, near to the end of the early Pleistocene, there is a huge radiation that leads to a number of different lineages such as the *Microtus (Stenocranius) hintoni*–*M. (St.) gregaloides*–*Microtus (St.) gregalis* lineage. For the early part of the Quaternary, one of the important lineages is the *Mimomys occitanus*–*Mimomys hajnachensis*–*Mimomys*

polonicus–*Mimomys pliocaenicus*–*Mimomys ostramosensis* lineage (Neraudeau et al., 1995) (Figure 2).

Extinction

Biostratigraphically important is also the extinction of species. Several examples demonstrate the global or local extinction of species. Well known are the late Pleistocene–early Holocene extinctions of mainly larger mammals that can be observed in almost every continent. In Eurasia, species such as *Ursus spelaeus*, *Mammuthus primigenius*, *Coelodonta antiquitatis*, *Equus hydruntinus*, and *Megaloceros giganteus* disappeared at the end of the late Pleistocene or the beginning of the Holocene. Extinction in the smaller mammal fauna of Eurasia can be observed in middle Pleistocene faunas. Species such as *Drepanosorex savini*, *Talpa minor*, and *Trogotherium cuvieri* are relicts from the early Pleistocene that became extinct during the middle Pleistocene.

Extinction is often preceded by a strong reduction of the geographical range of a species and a retreat to a refugial area. The late Pleistocene European rhinoceros species *Stephanorhinus kirchbergensis* and *Stephanorhinus hemitoechus* show a progressive southward contraction of their geographical range before getting extinct before the Late Glacial Maximum (Stuart, 1993, 2005; Stuart et al., 2004).

Migration and Dispersal of Species

The migration of mammal species is certainly the major factor in the changes of composition of the Quaternary faunas in a specific region. These migrations are first of all caused by the alternation of the available habitats due to the changes in climate and environment (van Kolfschoten, 1995). Eurasian cold stage faunas from the last and penultimate glacial period are characteristic and rather well known; species such as *Dicrostonyx gulielmi*, *Lemmus lemmus*, *Mammuthus primigenius*, *Coelodonta antiquitatis*, and *Rangifer tarandus* expanded their range southwards and occur together with species which prefer a more steppic environment such as ground squirrels (*Spermophilus undulates*) and hamsters (*Cricetulus migratorius* and *Cricetus cricetus*) that expanded their geographical range north and westwards.

A rise in temperature led to more distinct steppic conditions in the lower latitudes. These steppic conditions resulted in the increase of the relative number of steppe elements. Lemmings and other cold stage indicators withdrew northwards and species such as the steppe lemming *Lagurus lagurus* migrated westward and invaded Northwestern Europe. A rise in temperature is followed in certain areas such as the North European plain by an increase of oceanic influences. That results in a climate, which induces a re-establishment of forests with thermophilous broad-leaved and coniferous trees and the return of forest dwellers such as glirids (*Eliomys quercinus*, *Muscardinus avellanarius*), wild boar (*Sus scrofa*), and cervids (*Cervus (Dama) dama* and *Capreolus capreolus*) (van Kolfschoten, 1992).

This general picture of alternating species, more or less the same species that ‘comes and goes’, is applicable to the late middle and late Pleistocene and may be also to the earlier cold stages. However, one can distinguish a migration of species that is more or less independent of, and not only the result of

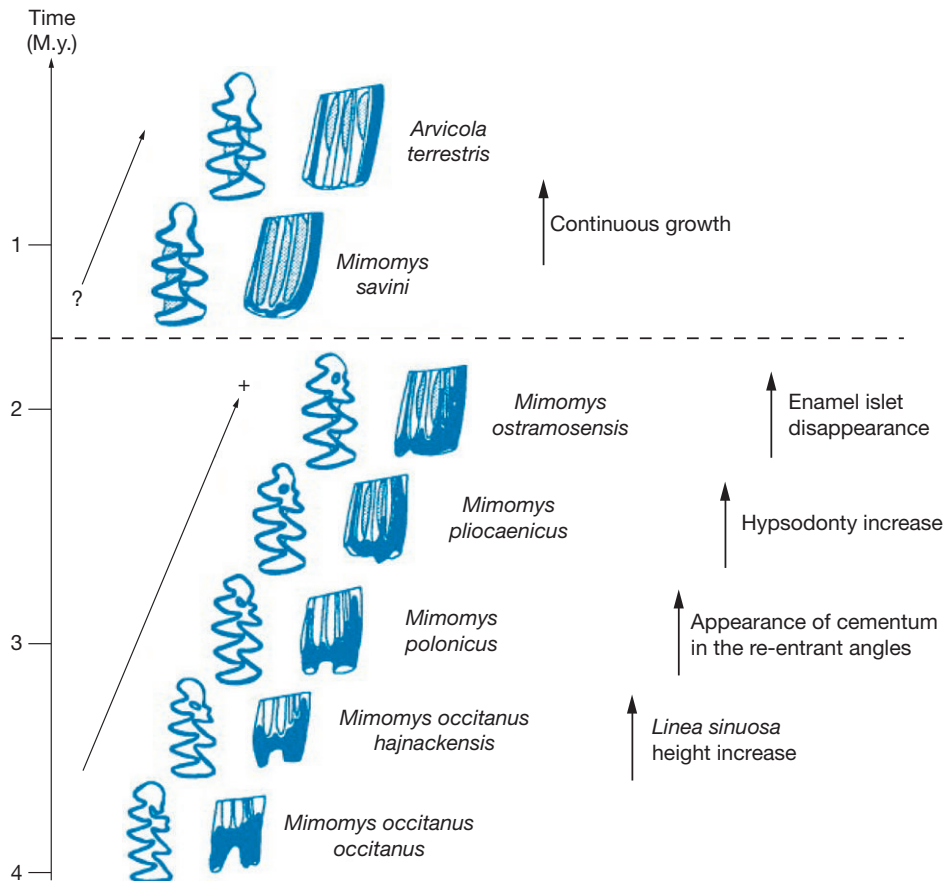


Figure 2 Phyletic gradualism in the *Mimomys occitanus*–*M. ostromosensis* lineage, showing the major evolutionary trends: increased hypsodonty related to a higher linea sinuosa; appearance of cementum in the re-entrant angles, and disappearance of the enamel islet. Continuous growth only occurs between *M. savini* and *A. terrestris* cantiana (after Chaline, 1987, modified). Reproduced from Neraudeau D, et al. (1995) Plio-Pleistocene water voles. *Palaeontology* 38(1): 77–85.

climatic change. The contemporary migration of a number of species characterizes the so-called major dispersal events. In addition, there are also incidental immigrations of single species. The woolly rhinoceros *Coelodonta antiquitatis*, for example, has its origin in Asia and invaded Europe during the middle Pleistocene, together with other cold stage faunal elements. The species is, since its appearance, a permanent element of the European cold stage faunas. The straight-tusk elephant *Elephas (P.) antiquus* invaded North Western Europe for the first time during the early middle Pleistocene and occurs during every interglacial period ever since. *Hippopotamus amphibius* also invaded Europe during the early middle Pleistocene, most probably during the same warm phase in which *Elephas (P.) antiquus* appeared. However, *Hippopotamus* did not return until the late Pleistocene (Eemian), after its withdrawal at the end of the early middle Pleistocene.

Another aspect is the migration and dispersal at subspecies level; the migration and dispersal of populations to areas which were inhabited by the same species. Tracing these migrations demands significant morphological differences between the 'new' population and the 'old' one, as is the case between the Late Saalian and Eemian *A. terrestris* populations of Northwestern Europe. The gradual change in differentiation of the enamel is interrupted and shows an important

fluctuation at the Saalian–Eemian transition. An explanation of this feature is the invasion of less derived populations. *A. terrestris* disappeared from the northern areas which were covered by inland ice and along the edge of the inland ice shield during the Late Saalian. The re-immigration of that area took place through the invasion of less-advanced populations which survived in areas located more to the south-(east). The presence of more primitive features in southern populations during the Pleistocene is very plausible since it has been demonstrated that the living *Arvicola* populations also show a strong morphological cline in the development of the enamel differentiation. The southern populations are in this aspect much more primitive than the northern populations (van Kolfschoten, 1992).

Biostratigraphical Subdivision of the Quaternary

Changes in the composition of the Quaternary mammalian fauna form the base for a biostratigraphical subdivision. In particular, as these changes are rapid and can be traced in a geographically extensive area the presence of a certain species, its evolutionary stage, or (although restricted) the absence of a species, can be used for dating deposits. No absolute

age-estimation but the indication of a relative age; dating in the sense of older than, or younger than, or about the same age.

The extensive and well-investigated Quaternary mammalian fossil record offers the possibility to divide the record into clusters and to establish a biostratigraphical subdivision or zonation of the Quaternary period (Figure 3). Each biostratigraphical unit or biozone is characterized by a distinct faunal assemblage. A number of mammal biozonations have been established by different authors. However, many of these zonations are not defined according to the official guide to stratigraphic nomenclature, and the terminology used by some authors is confusing. Furthermore, the application of several biozonations is geographically restricted. Two different biozonations are widely used in Quaternary biostratigraphy: the first one is based on the changes in the Eurasian larger mammal faunal community, the second one on the evolution in the smaller mammal assemblages. The coexistence of two different biozonations, one for the larger and one for the smaller mammalian record, is historically grown and is, for example, the result of specialization within the community of mammalian paleontologists.

Larger Mammal Biozonation

The widely adapted larger mammal biozonation is mainly based on the Italian fossil record and is established by Azzaroli and coworkers (Azzaroli et al., 1988). The Quaternary larger mammal faunas are divided into three assemblages: Villafranchian, Galerian, and Aurelian faunas (Figure 3). The Villafranchian is divided into an early, middle, and late Villafranchian; the transitions are marked by major events (Figure 4). The *Leptobos* event, indicating the appearance of species of the genus *Leptobos* in the Italian mammal faunas, marks the beginning of the Villafranchian. The first occurrence of *Mammuthus meridionalis* and *Equus stenonis*, the so-called elephant–*Equus* event, marks the transition of the early to middle Villafranchian. However, Rook and Martinez-Navarro (2010) have recommended against using the term ‘elephant–*Equus* event’ because true modern elephants (genus *Mammuthus*) and modern one-toed horses are recorded in some early Villafranchian faunas. The appearance of *Canis etruscus*, the ‘Wolf’ event, traditionally marked the middle to late Villafranchian transition, but it is now known that the expansion of wolf-like dogs across Eurasia was a diachronous event. A new term, ‘*Pachycrocuta brevirostris* event’ has recently been proposed because this hyaenid carnivore had a major continent-wide impact on faunal assemblages (Rook and Martinez-Navarro, 2010). The ‘end-Villafranchian’ dispersal event was originally described as a total faunal turnover, and massive extinctions and replacements mark the transition to the Galerian. However, it appeared that the original definition of the boundary between the Villafranchian and the Galerian is not applicable since the duration of the total faunal turnover took altogether about 0.5 Ma.

The Villafranchian starts long before the Quaternary, the oldest Villafranchian faunas have an age of about 3.5–3.6 Ma (Rook and Martinez-Navarro, 2010). The elephant–*Equus* event has an age of 2.5–2.6 Ma, the ‘*Pachycrocuta brevirostris* event’ age of 1.8 Ma. The stratigraphical position of the Villafranchian–Galerian transition is still problematic. Some authors put the transition at an age of about 1.1–1.2 Ma,

whereas others put the beginning of the Galerian close to the Brunhes/Matuyama boundary with an age of 0.78 Ma.

The Galerian–Aurelian transition is marked by the renewal of the fauna that coincides with the extinction of species such as those of the megacerine cervids of the *Megaceroides verticornis* group and *Megaloceros savini*, while modern taxa appear for the first time (Glozzi et al., 1997).

Smaller Mammal Biozonation

The smaller mammal biozonation that is widely used as a standard for the early and middle Pleistocene of Eurasia, is mainly a modified version of the Hungarian biozonation established by Kretzoi (1965). In this subdivision, three biozones are recognized for the Pleistocene period in the literature often erroneously referred to as ‘stages’: Villányian, Biharian, and Toringian. The Villányian faunas can be recognized by the dominance of voles of the genus *Mimomys* and the absence (or the occurrence in only a very low percentage) of voles of the genus *Microtus*. The Biharian faunas are characterized by a dominance of *Microtus* co-occurring with *Mimomys*. The Biharian biozone is divided into two substages: Lower and Upper Biharian. The disappearance of the subgenus *Microtus* (*Allophaiomys*) marks the transition from the Lower to the Upper Biharian. The Toringian biozone can be recognized by the *Arvicola*–*Microtus* assemblages. *Mimomys* is missing in Toringian faunas.

The chronostratigraphical age of boundaries of the biozones and hence, the chronostratigraphical ranges of the biozones are still a matter of debate. The Villányian started long before the beginning of the Quaternary at about 3.6–3.7 Ma. One of the most important abrupt changes in the smaller mammalian fauna took place during the transition of the Villányian to the Biharian biozone. The major faunal ‘turnover,’ in particular the reconstruction of the vole community, and the expansion of rootless voles characterized by the appearance of *Microtus* (*Allophaiomys*) (a group of voles, easily recognizable and well represented in the fossil record of the northern hemisphere), was dated until recently just after the Olduvai Event. This transition formed one of the strong ‘continental’ arguments to put the Plio–Pleistocene boundary near the top of the Olduvai. New data, however, indicate that the age of this faunal ‘turnover’ is not indisputable. Tesakov argues that this faunal ‘turnover’ took place long before the Olduvai Event. He dates the transition between 2.1 and 2.2 Ma. The transition between the Lower and the Upper Biharian is rather well dated. Most authors agree that *Microtus* (*Allophaiomys*) disappeared during or just before the Jaramillo paleomagnetic Subchron. The Biharian–Toringian transition is studied in great detail because the age of this transition is very important for dating the earliest Paleolithic sites in Europe. The first occurrence of the genus *Arvicola*, marking the beginning of the Toringian, dates around 0.5 Ma (van Kolfschoten, 1992; von Koenigswald and van Kolfschoten, 1996).

Additional Biozonations

The biozonations described above are mainly used by Eurasian continental biostratigraphers often next to regional zonations applicable in the area of investigation. There is no elaborated



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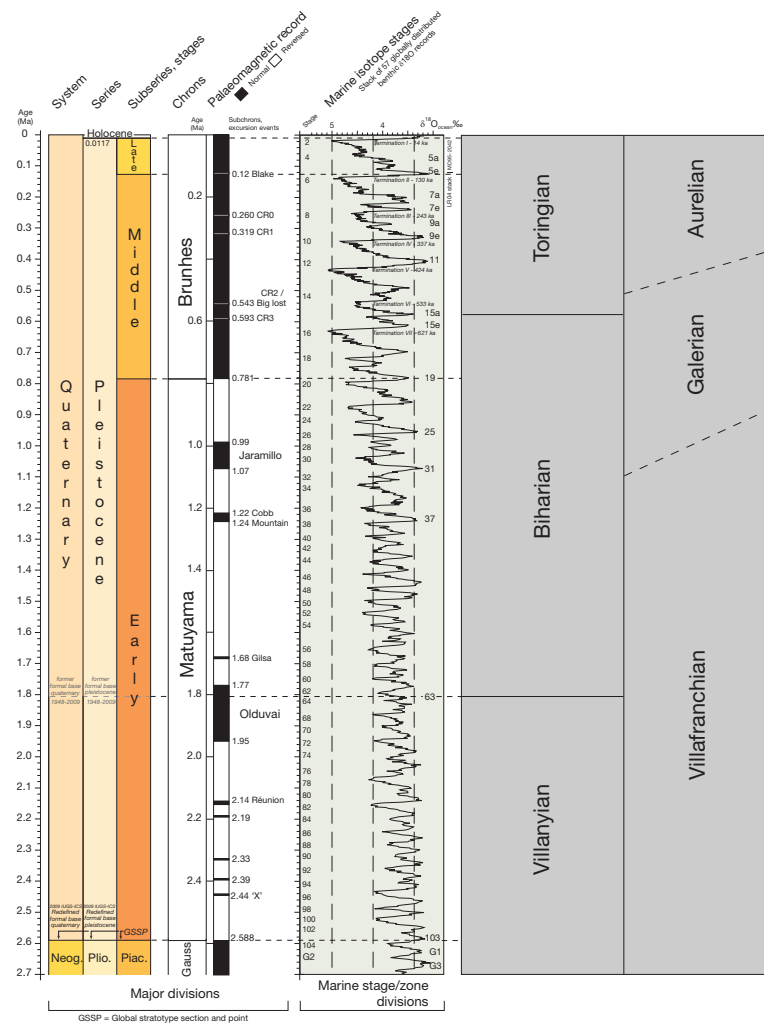


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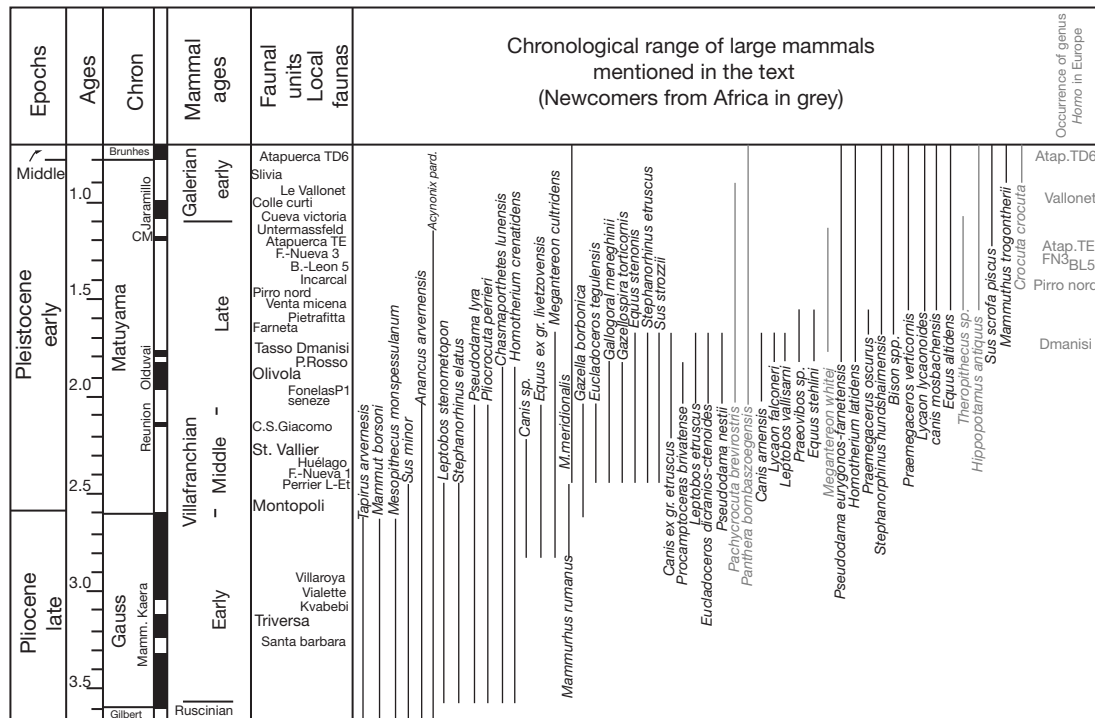


Figure 4 Villafranchian chronology with range chart for selected European large mammals (from Rook and Martinez-Navarro, 2010). Newcomers from Africa shown in grey.

widely applicable biozonation of the continental floral record despite the fact that the impact of climate change during the Quaternary on the floral composition in most regions is huge and the stratigraphical range of many species is very fragmented and/or restricted. Floral species hardly show any evolution during the Quaternary and also the number of extinctions is restricted.

In the marine realm, there are a number of micropalaeontological zonations (Gibbard and van Kolfschoten, 2004; Lourens et al., 2004). Biozonations have been established for the fossil record of different microfossils: planktonic Foraminifera, calcareous nannofossils, diatoms, Radiolaria, and dinoflagellates. The planktonic Foraminifera biozonation and the calcareous nannoplankton biozonation are the major tools for subdividing marine sediments. The application of the other fossils is restricted to areas with specific conditions.

Concluding Remarks

Detailed studies of Quaternary deposits from the continental as well as from the marine environment resulted in a number of biozonations that play a role in the debate on dating and correlating strata. This is in particular the case for early Quaternary deposits. The use of biostratigraphical arguments is less relevant in younger deposits because the biozonation is too coarse for the detailed information one is looking for. The stratigraphical setting of middle and late Pleistocene strata in many cases offers the possibility to correlate them to regional stages such as the widely adapted northwest European Stages. Furthermore, there is a tendency to use the Marine Isotope

Stages as a reference in the debate on the age of continental finds, in spite of the very restricted number of direct correlations between continental and marine zonations.

See also: Diatom Introduction; Vertebrate Overview.

Paleoceanography, Biological Proxies: Planktic Foraminifera. **Quaternary Stratigraphy:** Overview. **Vertebrate Records:** Early and Middle Pleistocene of Northern Eurasia; Early Pleistocene; Late Pleistocene Megafaunal Extinctions; Late Pleistocene Mummified Mammals; Late Pleistocene of Africa; Late Pleistocene of North America; Late Pleistocene of South America; Late Pleistocene of Southeast Asia; Mid-Pleistocene of Africa; Mid-Pleistocene of Australia; Mid-Pleistocene of Europe; Mid-Pleistocene of North America; Mid-Pleistocene of Southern Asia. **Vertebrate Studies:** Ancient DNA.

References

- Azzaroli A (1977) *Evolutionary patterns of Villafranchian elephants in Central Italy*. Memorie. Classe di Scienze fisiche, matematiche e naturali, serie VIII, volume XIV, sezione II a, 4: 149–168. Rome: Atti della Accademia Nazionale dei Lincei.
- Azzaroli A, de Giulio C, Ficcarelli G, and Torre D (1988) Late Pliocene to early mid-Pleistocene mammals in Eurasia faunal succession and dispersal events. *Palaeogeography Palaeoclimatology Palaeoecology* 66: 77–100.
- Boeuf O (1990) Originalité et importance de la faune Plio-Pléistocène de Chilhac (Haute-Loire), France. *Quartärpaläontologie* 8: 13–28.
- Chaline J (1987) Arvicolid Data (Arvicolidae, Rodentia) and Evolutionary Concepts. *Evolutionary Biology* 21: 237–310.
- Dubrov IA (1966) On systematic position of fossil elephant from the Khozarsky faunal complex. *Bull. Comissii po izuch, chtvert perioda* 32: 63–74.
- Gibbard PL and van Kolfschoten T (2004) The Pleistocene and Holocene Epochs. In: Gradstein FM, Ogg JG, and Smith A (eds.) *A Geologic Time Scale*, pp. 441–452. Cambridge: Cambridge University Press.

- Gliozzi E, Abbazzi L, Argenti P, et al. (1997) Biochronology of selected mammals, molluscs and ostracods from the Middle Pliocene to the Late Pleistocene in Italy. The state of the art. *Rivista Italiana di Paleontologia e Stratigrafia* 103(3): 369–388.
- Kretzoi M (1965) Die Nager und Lagomorphen von Voigtstedt in Thüringen und ihre chronologische Aussage. *Palaontologische Abhandlungen* 2(3): 587–660.
- Kubiak H (1980) The skulls of *Mammuthus primigenius* (Blumenbach) from Dębica and Bzianka near Rzeszów, South Poland. *Folia Quaternaria* 51: 31–45.
- Lister AM, Sher A, van Essen H, and Wei G (2005) The pattern and process of mammoth evolution in Eurasia. *Quaternary International* 126–128: 49–64.
- Lourens L, Hilgen F, Shackleton NJ, Laskar J, and Wilson D (2004) The Neogene Period. In: Gradstein FM, Ogg JG, and Smith A (eds.) *A Geologic Time Scale*, pp. 409–440. Cambridge: Cambridge University Press.
- Maccagno AM (1962) *l'Elephas meridionalis* Nesti di contrada "Madonna della Strada" Scoppito (l'Aquila). Guglielmo Genovese: Napoli.
- Neraudeau D, Viriot L, Chaline J, Laurin B, and van Kolfschoten T (1995) Discontinuity in the Eurasian Water Vole Lineage (Arvicolidae, Rodentia). *Palaeontology* 38(1): 77–85.
- Rook L and Martínez-Navarro B (2010) Villafranchian: The long story of a Plio-Pleistocene European large mammal biochronological unit. *Quaternary International* 219: 134–144.
- Stuart AJ (1993) The Failure of Evolution: Late Quaternary Mammalian Extinctions in the Holarctic. *Quaternary International* 19: 101–109.
- Stuart AJ (2005) The extinction of woolly mammoth (*Mammuthus primigenius*) and straight-tusked elephant (*Palaeoloxodon antiquus*) in Europe. *Quaternary International* 126–128: 171–177.
- Stuart AJ, Kosintsev PA, Higham TFG, and Lister AM (2004) Pleistocene to Holocene extinction dynamics in giant deer and woolly mammoth. *Nature* 431: 684–689.
- van Kolfschoten T (1992) Aspects of the migration of mammals to northwestern Europe during the Pleistocene, in particular the re-immigration of *Arvicola terrestris*. *Courrier Forschung Senckenberg* 153: 213–220.
- van Kolfschoten T (1995) On the application of fossil mammals to the reconstruction of the palaeoenvironment of northwestern Europe. *Acta Zoologica Cracoviensia* 38: 73–84.
- von Koenigswald W and van Kolfschoten T (1996) The *Mimomys*–*Arvicola* boundary and the enamel thickness quotient (SDQ) of *Arvicola* as stratigraphic markers in the Middle Pleistocene. In: Turner C (ed.) *The early Middle Pleistocene of Europe*, pp. 211–226. Rotterdam: Balkema.

Chronostratigraphy

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Introduction

Chronostratigraphy is the part of stratigraphy that deals with the relative time relations of rock bodies, and a chronostratigraphic unit is a body of rock that was formed during a specified interval of geological time (Salvador, 1994). (The use of the term 'rock' to describe Quaternary strata that are unconsolidated or semiconsolidated may seem inappropriate at first sight. However, such usage is consistent with that adopted by the International Stratigraphic Guide (Salvador, 1994), and thus the term 'rock' is used here in its broadest sense.)

To facilitate international communication, the rock record of the earth is divided into standardized global chronostratigraphic units that make up the international geological time scale (GTS), managed by the International Commission on Stratigraphy (ICS). No GTS can be final, and the GTS is continually being refined as new data become available. The latest iteration is GTS2012 (Gradstein et al., 2012), which follows earlier iterations such as GTS2004 (Gradstein et al., 2004) GTS1989 (Harland et al., 1990) and GTS1982 (Harland et al., 1982). Chronostratigraphic subdivisions of the Cenozoic Erathem are shown in Figure 1.

Each chronostratigraphic (time-rock) unit of the GTS has an equivalent geochronologic (time) unit (Table 1). This dual hierarchy means that geologists refer to the rocks of the Pleistocene Series, but events in the Pleistocene Epoch. Similarly the terms 'early,' 'middle,' and 'late' are used as qualifiers for time units, while 'lower' and 'upper' are used as qualifiers for time-rock units. Zalasiewicz et al. (2004) have proposed to end the distinction between the two parallel classifications as a way of simplifying stratigraphic terminology. If their proposal were to be adopted, the terms Eonothem, Erathem, System, Series, and Stage would become redundant, in favor of Eon, Era, Period, Epoch, and (disputably) Age. And the qualifiers Upper and Lower would be made redundant in favor of Early and Late.

Global Boundary Stratotype Sections and Points

International subdivisions of the Phanerozoic part of the GTS are defined by their lower boundaries at Global Stratotype Sections and Points (GSSPs), often referred to as golden spikes. Note that a GSSP automatically defines the upper boundary of the underlying unit, thereby eliminating the possibility of overlap between adjacent units.

The principal requirements of a GSSP are

1. Location in a stratigraphically continuous section.
2. Good exposure and easy access for further study.
3. Precise definition at a lithological marker horizon.
4. The presence of multiple criteria for global correlation.

Typically, GSSPs have been defined in marine sedimentary sequences to make use of marine biostratigraphic markers for global correlation. However, this is not a necessary requirement. Similarly, for reasons of ease of access, until 2008, no GSSP had

been defined in drill core material. Both of these considerations were fully assessed when the Holocene GSSP was defined in a Greenland ice core (see Section 'Holocene Series').

Defining the Quaternary

The term Quaternary was introduced by the Italian mining engineer, Giovanni Arduino (1714–95), when he distinguished four orders of geological strata – Primary, Secondary, Tertiary, and Quaternary (Schneer, 1969). However, its definition and usage in the modern sense was not really established until the XVIIIth International Geological Congress in London in 1948. At this conference, it was resolved to select a stratotype for the Pliocene–Pleistocene boundary (stated as equivalent to the Tertiary–Quaternary boundary) and to “place the boundary at the first indication of climatic deterioration in the Italian Neogene succession (King and Oakley, 1949).” After consideration of several sections over the next three decades, the Vrica section in Calabria, southern Italy, was chosen and ratified as the Pliocene–Pleistocene stratotype (Aguirre and Pasini, 1985).

Unfortunately, the formal definition of the Quaternary was left unresolved when Aguirre and Pasini (1985: 116) stated that “The subject of defining the boundary between the Pliocene and Pleistocene was isolated from ... the status of the Quaternary within the chronostratigraphic scale.” The Quaternary was subsequently shown variously as a System above the Tertiary System (Salvador, 1994), or as a System above the Neogene System (Harland et al., 1990), but was never formally ratified as such. In each case, the base of the Quaternary (=base Pleistocene) was assigned an age of ~1.8 Ma.

When the Quaternary was shown as an informal climatostratigraphic unit in GTS2004, the international Quaternary community was incensed, and much discussion ensued (e.g., Aubry et al., 2005; Gibbard et al., 2005; Pillans and Naish, 2004). A significant outcome was that a majority of Quaternary scientists favored a definition of the Quaternary with its base at ~2.6 Ma to encompass the time interval during which (1) the Earth's climate has been strongly influenced by bipolar glaciation and (2) the genus *Homo* first appeared and evolved (Pillans and Naish, 2004). Figure 2 summarizes the major temporal and latitudinal relations between orbital forcing and the Earth's late Pliocene–early Pleistocene climate (3.0–1.5 Ma) as accounted in various records, including high-resolution deep ocean $\delta^{18}\text{O}$ ice volume record from southwest Pacific ODP Site 1123, glacioeustatic cyclothems of Wanganui Basin New Zealand, ice rafting as recorded by magnetic susceptibility at North Pacific ODP Site 882, a median grain size profile from Jingchuan loess section in China, and a pollen summary diagram from ODP 658 (offshore West Africa) showing progressive aridification. A step-like increase in African aridity at about 2.8 Ma is linked to a significant event in hominid evolution in East Africa as the genera *Paranthropus* and *Homo* emerge from a single lineage (Pillans and Naish, 2004).

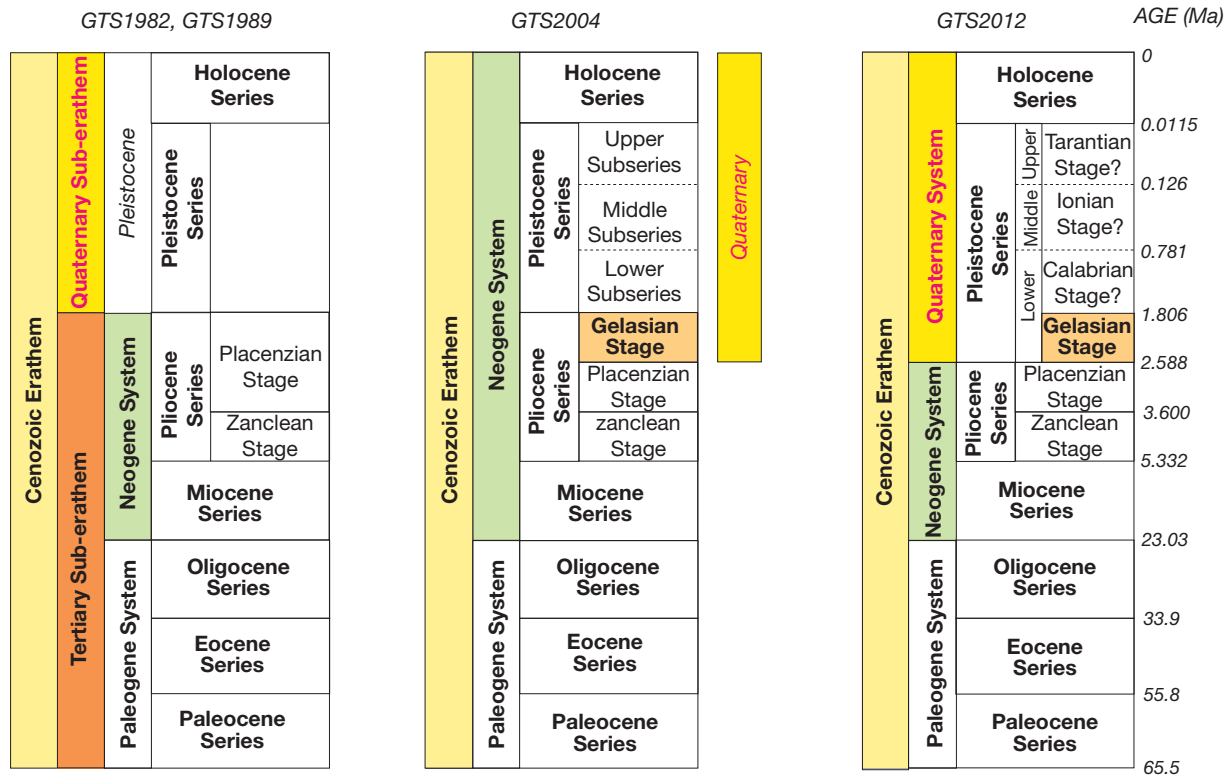


Figure 1 The evolving chronostratigraphic subdivision of the Cenozoic according to recent and proposed iterations of the international geological time scale (GTS). Sources: GTS1982 (Harland et al., 1982), GTS1989 (Harland et al., 1990), GTS2004 (Gradstein et al., 2004) and GTS2012 (Gradstein et al., 2012). For simplicity, stages of pre-Pliocene Series are not shown.

Table 1 Chronostratigraphic (time-rock) units of the GTS and their geochronologic (time) equivalents

Chronostratigraphic units	Geochronologic units
Eonothem	Eon
Erathem	Era
System	Period
Series	Epoch
Stage	Age

Source: Salvador A (ed.) (1994) *International Stratigraphic Guide*, p. 214. Boulder: Geological Society of America.

In August 2005, a joint ICS-INQUA Task Group, which had been established to resolve the issue of how best to define the Quaternary, made the following recommendations:

1. The Quaternary should be recognized as a formal chronostratigraphic/geochronologic unit.
2. The lower boundary of the Quaternary should coincide with the base of the Gelasian Stage (~2.6 Ma) and thus be defined by the Gelasian GSSP.
3. The Quaternary should have the rank of either System/Period above a shortened Neogene, or Sub-erathem/Sub-era within the Cenozoic and correlative with the upper part of an extended Neogene.

In September 2005, a meeting of the ICS recommended that the Quaternary be formally defined as a Sub-Era of the Cenozoic Era and that its base be defined by the GSSP for the Gelasian

Stage of the Upper Pliocene (see below). However, this proposal was rejected by INQUA, in favor of a Quaternary System/Period.

Finally, in June 2009, the Executive Committee of the International Union of Geological Sciences (IUGS) formally ratified a proposal by ICS to define the base of the Quaternary System at the GSSP for the Gelasian Stage at Monte San Nicola in Italy (Gibbard et al., 2010). At the same time, the base of the Pleistocene was also lowered to coincide with the Gelasian GSSP. Thus, the Gelasian Stage was transferred from the Pliocene Series to the Pleistocene Series (Figure 1).

The Gelasian Stage

The Gelasian Stage was originally defined as the youngest of three stages that make up the Pliocene Series. The GSSP for the Gelasian Stage is located in the Monte San Nicola section (Figure 3) near the town of Gela in Sicily (Rio et al., 1998). It is defined at the base of the marly layer overlying sapropel MPRS 250, an organic-rich horizon with an astronomically calibrated age of 2.588 Ma (Lourens et al., 1996, 2004). The Gauss/Matuyama paleomagnetic boundary occurs about 1 m (20 ka) below MPRS 250 and provides a close approximation for global correlation of the base of the Gelasian (=base of Quaternary System).

The Pleistocene Series

The GSSP for the Pliocene–Pleistocene boundary (=base Pleistocene) was originally located in the Vrica section, 4 km south

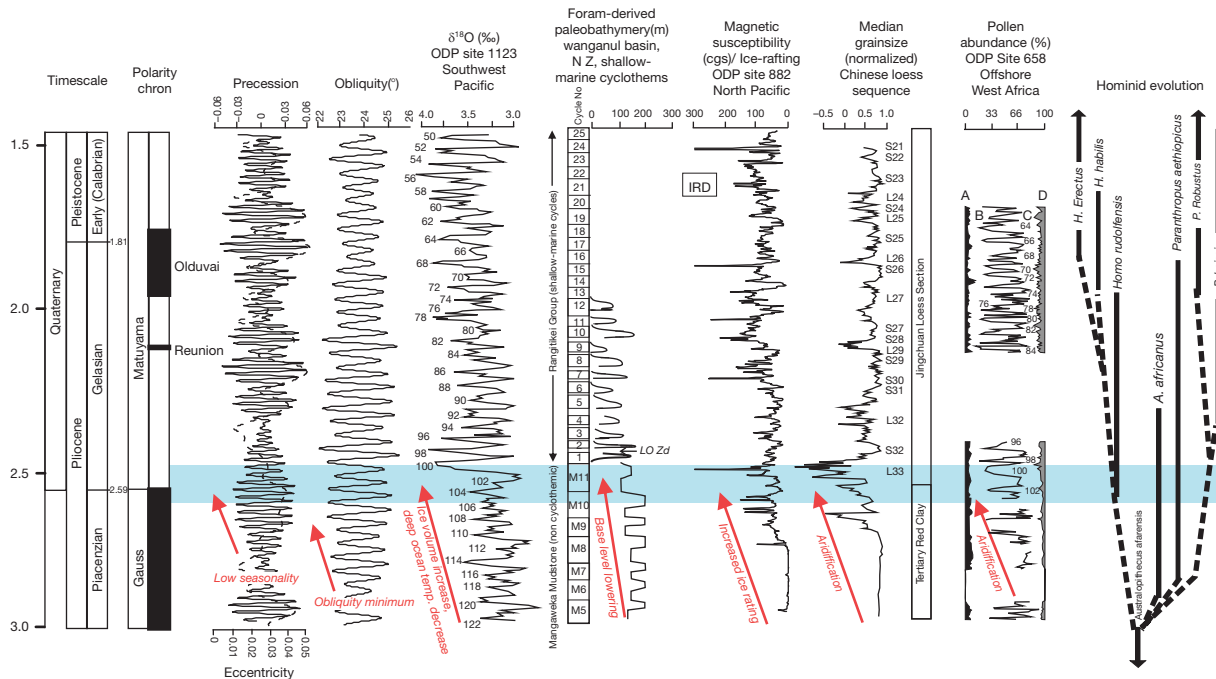


Figure 2 Summary of links between orbital forcing and paleoclimatic changes around 2.6 Ma that define the base of the Quaternary. Reproduced from Pillans B and Naish T (2004) Defining the Quaternary. *Quaternary Science Reviews* 23: 2271–2282.



Figure 3 GSSP (arrowed) for the Gelasian Stage (Rio et al., 1998), GSSP for the Pleistocene Series and Quaternary System. Source: Web site of the International Commission on Stratigraphy, <http://www.stratigraphy.org/>.



Figure 4 Former GSSP (top of marker bed 'e') for the Pleistocene Series, Vrica section, southern Italy (Aguirre and Pasini, 1985), now the GSSP for the Calabrian Stage. Source: Web site of the International Commission on Stratigraphy, <http://www.stratigraphy.org/>.

of Crotona in Calabria, southern Italy (Figure 4). The boundary was defined at the top of a sapropel layer 'e' (Aguirre and Pasini, 1985) with an astronomically calibrated age of 1.806 Ma (Lourens et al., 1996, 2004). The boundary lies just above the top of the Olduvai Subchron, which facilitates global correlation. This GSSP has now been proposed as the base of the Calabrian Stage (Cita et al., 2008).

Subdivision of the Pleistocene Series

Three working groups of the ICS are currently seeking suitable GSSPs to define the Upper, Middle, and Lower Pleistocene.

Lower Pleistocene

By definition, the GSSP for the Lower Pleistocene should be the base Pleistocene GSSP at Monte San Nicola section in Sicily. It is proposed that the Lower Pleistocene Subseries will comprise the Gelasian and Calabrian Stages.

Middle Pleistocene

Participants in the Burg Wartenstein Symposium 'Stratigraphy and Patterns of Cultural Change in the middle Pleistocene,' held in Austria in 1973, recommended that the beginning of the middle Pleistocene be defined to either coincide with, or be linked to, the Matuyama–Brunhes paleomagnetic reversal boundary (Butzer and Isaac, 1975). A similar recommendation

was made by the INQUA Working Group on Major Subdivision of the Pleistocene at the XIIth INQUA Congress in Ottawa in 1987 (Richmond, 1996), and although potential GSSPs in Japan, Italy, and New Zealand were discussed, no decision was reached. Pillans (2003) also advocated the Matuyama–Brunhes boundary (MBB), because it constitutes the most recognizable chronostratigraphic marker in weathered continental deposits.

At present, the SQS Working Group on the Lower–middle Pleistocene Boundary is considering candidate sections in Italy and Japan, having resolved at the 32nd International Geological Congress in Florence in 2004 to place the boundary as close as possible to the MBB.

Upper Pleistocene

The Middle–Upper Pleistocene boundary is generally placed at the beginning of the Last (Eemian) Interglacial or at the beginning of Marine Isotope Stage 5 (MIS 5), with an age of ~130 000 years (Gibbard, 2003). However, detailed pollen analyses of marine cores to the west of Portugal have shown that the base of MIS 5 is significantly older (~60 000 years) than the base of the Eemian (Shackleton et al., 2003; Figure 5).

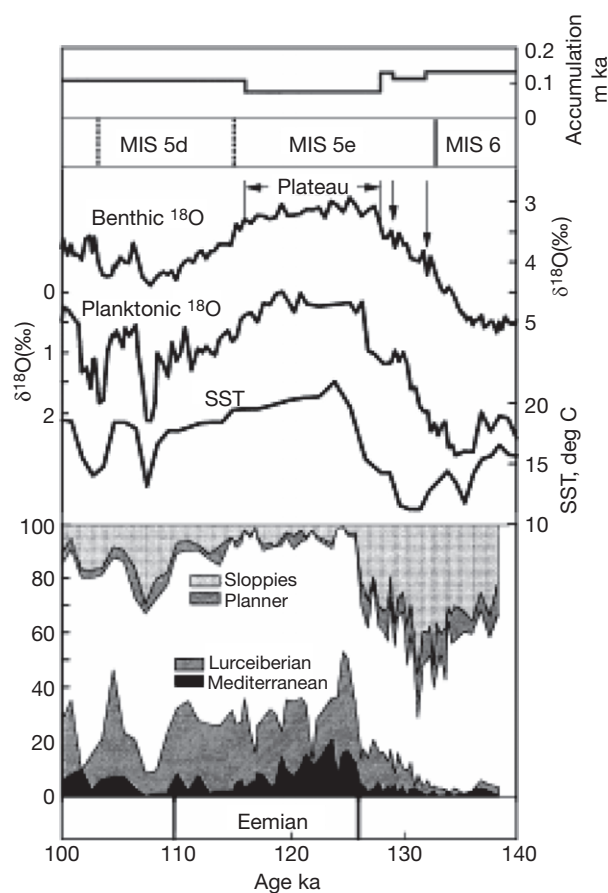


Figure 5 Marine and continental records of the last (Eemian) interglacial compared with MIS 5e in marine core MD95-2024. Reproduced from Shackleton NJ, Sanchez-Goni MF, Paillet D, and Lancelot Y (2003) Marine Isotope Substage 5e and the Eemian Interglacial. *Global and Planetary Change* 36: 151–155.

Gibbard (2003) has argued that the Upper Pleistocene GSSP should be chosen in a high-resolution terrestrial sequence at the base of the Eemian, and he nominates a depth of 63.5 m in the Amsterdam Terminal borehole as a candidate GSSP.

Holocene Series

For many years, the base of the Holocene Series (= top of Upper Pleistocene) was broadly accepted as the end of the last glacial period, with an age of 10 000 radiocarbon years. However, this definition was never ratified by ICS. In 2008, the GSSP was formally defined at a depth of 1492.3 m in the NGRIP ice core from Greenland, reflecting the first signs of climatic warming at the end of the Younger Dryas (Walker et al., 2008, 2009). The age of the boundary, based on multi-parameter annual layer counting is 11 734 cal ¹⁴C years BP. This GSSP is the first to be defined in an ice core.

Local Chronostratigraphic Units and Time Scales

In most regions of the world, fossils remain the principal means of establishing the age and sequence of Phanerozoic rocks, including many Quaternary deposits. However, while it is generally possible, using biostratigraphy to correlate deposits with the major chronostratigraphic units of the GTS at Series level and above, regional endemism of biota often makes correlation at finer scales (e.g., Stage level) problematic. Establishment of local time scales is a solution to this problem.

Numerous local 'time scales' have been erected for regional subdivision of the Quaternary, but the majority of these are, strictly speaking, climatostratigraphic units, not chronostratigraphic units. Furthermore, few of the regional stages have been formally defined using Local boundary Stratotype Sections and Points (LSSPs). A notable exception is in New Zealand, where most Cenozoic local stages are defined (or are in the process of being defined) by LSSPs in marine sequences. These include the New Zealand Haweran, Castlecliffian, and Nukumaruan Stages with LSSPs in Wanganui Basin (Cooper, 2004).

The need for local time scales is increasingly debated. However, in relatively isolated regions such as New Zealand, so long as fossils remain the principal means of correlation, it seems likely that local time scales will be required.

Chronostratigraphic Markers

A chronostratigraphic horizon or marker is a stratigraphic surface that is everywhere of the same age (Salvador, 1994). Although they are theoretically without thickness, in practice, the term chronostratigraphic horizon (or chronohorizon) has been applied to very thin and distinctive markers that are essentially the same age over their whole geographic extent. Examples of chronostratigraphic markers in the Quaternary include tephras, magnetic polarity reversal horizons, and tektites. The latter are glassy objects produced from melted crustal rocks that are caused by large hypervelocity meteorite impacts. About 800 000 years ago, a large extraterrestrial impact occurred in southeast Asia, producing tektites, which are

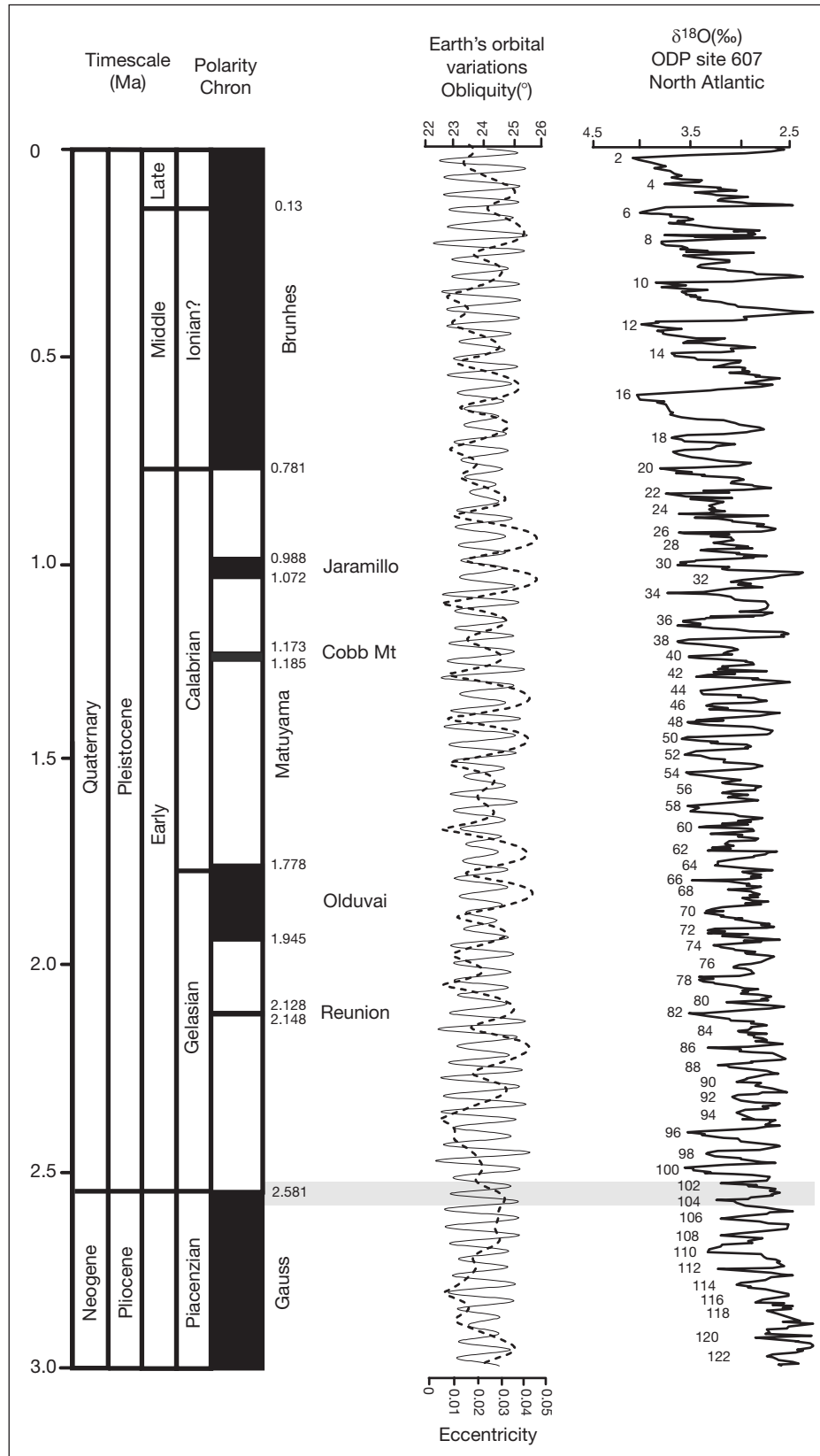


Figure 6 Chronostratigraphic units, paleomagnetic datums and ages (from Lourens et al., 2004) for the Quaternary. Orbital variations in obliquity and eccentricity from Berger and Loutre (1991). Oxygen isotope record from ODP Site 607 from Raymo et al. (1989).

found over a wide area, including much of Australia and surrounding oceans – these are usually referred to as Australites or Australasian tektites. Zhou and Shackleton (1999) used the Australasian tektite layer to demonstrate the misleading position of the MBB in Chinese loess, concluding that there was a time delay in remanence acquisition in the loess and a downward displacement of the apparent position of the MBB. Such a study is a pertinent reminder that magnetic reversal boundaries are not everywhere isochronous.

Finer Subdivision of the Quaternary Using Marine Isotope Stages

Based on analysis of Foraminifera in deep-sea cores, oxygen isotope records, have been divided into marine isotope stages (MIS), numbered from the top downward (Figure 6). Because the $\delta^{18}\text{O}$ signal dominantly reflects changes in the oxygen isotopic composition of the global ocean, the stages have been used to provide a framework for global correlation of marine cores, limited only by the mixing time of the ocean and deep-sea sediment bioturbation rates. Indeed, as Shackleton and Opdyke (1973: 48) famously predicted: “Thus it is highly unlikely that any superior stratigraphic subdivision of the Pleistocene will ever emerge.”

MIS are numbered so that odd numbers correspond with $\delta^{18}\text{O}$ minima (interglacial stages) and even numbers correspond with $\delta^{18}\text{O}$ maxima (glacial stages). MIS 1 corresponds to the present interglacial period and stages are numbered sequentially down to MIS 103 at the base of the Quaternary (Figure 6), and beyond.

MIS boundaries are placed at points of maximum change between maxima and minima. However, additional events (maxima and minima) have been identified and numbered within many stages. Prell et al. (1986) introduced a numbering system for these events whereby, for individual isotopic events within any stage, the integer portion of the code represents the MIS number and the decimal number refers to the event. As in MIS nomenclature, even number decimals refer to $\delta^{18}\text{O}$ maxima and odd number decimals refer to $\delta^{18}\text{O}$ minima. Thus, event 5.5 is the lower interglacial event in MIS 5 corresponding to Stage 5e of Shackleton and Opdyke (1973). In the same way, upper MIS boundaries are assigned a zero decimal, so event 5.0 corresponds to the MIS 4/5 boundary.

Although not representing true chronostratigraphic horizons, MIS and isotopic events are widely used as such, particularly for correlation between deep-sea cores. Furthermore, they are frequently used for correlation between marine and terrestrial sequences, a practice that ignores the likely leads and lags between marine and terrestrial indicators of past environmental change (cf. the ~6000 year estimated offset between the pollen and isotopic records of MIS 5.5 recorded in cores off the coast of Portugal by Shackleton et al. (2003)).

Astronomical Timescale

Prior to 1950 and the development of modern radiometric dating techniques such as radiocarbon, Quaternary chronology was rather speculative. Early attempts to infer ages for various

glacial deposits using the chronology of orbital variations and resultant solar insolation curves (calculated by Milankovitch and others) were greeted with some skepticism. However, in the late 1960s and early 1970s, the astronomical chronology received strong support from dated coral reef sequences in New Guinea and Barbados (e.g., Broecker et al., 1968). However, it was Hays et al. (1976) who provided the first convincing evidence that the independent contributions of all the three orbital elements (obliquity, precession, and eccentricity) could be recognized in the geological record.

Today, the astronomical chronology of deep-sea cores is highly developed, and based on astronomical calibrations, the ages of all isotope stages and magnetic reversals in the Quaternary (and much of the Neogene) are well established (e.g., Lisiecki and Raymo, 2005; Lourens et al., 1996; Shackleton et al., 1990; see Figure 6). Astronomical calibration of marine sequences in the Mediterranean provides highly precise numerical ages for individual sedimentary cycles, including the Pleistocene and Gelasian (=Quaternary) GSSPs (Lourens et al., 2004). Astronomical calibration of the shallow marine sedimentary record from the Wanganui Basin in New Zealand has also been completed for the entire Quaternary (Naish et al., 1998, and updated by Pillans et al., 2005) and provides the primary chronology for LSSPs of the New Zealand Quaternary stages (Cooper, 2004).

See also: **Glacial Landforms, Ice Sheets:** Evidence of Glacier and Ice Sheet Extent. **Glacial Landforms, Erosional Features:** Major Scale Forms. **Glaciation, Causes:** Astronomical Theory of Paleoclimates. **Paleoceanography, Physical and Chemical Proxies:** Oxygen-Isotope Stratigraphy of the Oceans. **Paleoclimate Reconstruction:** Pliocene Environments. **Quaternary Stratigraphy:** Overview; Tephrochronology.

References

- Aguirre E and Pasini G (1985) The Pliocene-Pleistocene boundary. *Episodes* 8: 116–120.
- Aubry MP, Berggren WA, van Couvering J, McGowan B, Pillans B, and Hilgen F (2005) Quaternary: Status, rank, definition, survival. *Episodes* 28: 118–120.
- Berger A and Loutre MF (1991) Insolation values for the climate of the last 10 million years. *Quaternary Science Reviews* 10: 297–317.
- Broecker WS, Thurber DL, Goddard J, Ku T-L, Matthews RK, and Mesolella KJ (1968) Milankovitch hypothesis supported by precise dating of coral reefs and deep-sea sediments. *Science* 159: 297–300.
- Butzer KW and Isaac GL (eds.) (1975) *After the Australopithecines*, p. 911. The Hague: Mouton.
- Cita MB, Capraro L, Ciaranfi N, et al. (2008) The Calabrian stage redefined. *Episodes* 31: 408–419.
- Cooper RA (ed.) (2004) The New Zealand geological timescale. In: *Institute of Geological and Nuclear Sciences Monograph*, vol. 22, p. 284. Lower Hutt, New Zealand: Institute of Geological and Nuclear Sciences.
- Gibbard PL (2003) Definition of the middle-upper Pleistocene boundary. *Global and Planetary Change* 36: 201–208.
- Gibbard PL, Smith AG, Zalasiewicz JA, et al. (2005) What status for the Quaternary? *Boreas* 34: 1–6.
- Gibbard P, Head MJ, Walker M and The Subcommission on Quaternary Stratigraphy (2010) Formal ratification of the Quaternary System/Period and the Pleistocene Series/Epoch with a base at 2.58 Ma. *Journal of Quaternary Science* 25: 96–102.

- Gradstein FM, Ogg JG, and Smith AG (eds.) (2004) *A Geological Time Scale 2004*, p. 589. Cambridge: Cambridge University Press.
- Gradstein FM, Ogg JG, Schmitz M, and Ogg G (eds.) (2012) *The Geological Time Scale*, p. 1142. Oxford: Elsevier.
- Harland WB, Armstrong RL, Cox AV, Craig LE, Smith AG, and Smith DG (eds.) (1990) *A Geological Time Scale 1989*, p. 263. Cambridge: Cambridge University Press.
- Harland WB, Cox AV, Llewellyn PG, Pickton CAG, Smith AG, and Walters R (1982) *A Geological Time Scale*, p.131. Cambridge: Cambridge University Press.
- Hays JD, Imbrie J, and Shackleton NJ (1976) Variations in the Earth's orbit: Pacemaker of the ice ages. *Science* 194: 1121–1132.
- King WBR and Oakley KP (1949) Definition of the Pliocene-Pleistocene boundary. *Nature* 163: 186–187.
- Lisiecki LE and Raymo ME (2005) A Pliocene-Pleistocene stack of 57 globally distributed benthic $\delta^{18}\text{O}$ records. *Paleoceanography* 20: PA1003.
- Lourens LJ, Antonarakou A, Hilgen FJ, Van Hoof AAM, Vergnaud-Grazzini C, and Zachariasse WJ (1996) Evaluation of the Plio-Pleistocene astronomical timescale. *Paleoceanography* 11(4): 391–413.
- Lourens LJ, Hilgen FJ, Laskar J, Shackleton NJ, and Wilson D (2004) The Neogene period. In: Gradstein FM, Ogg JG, and Smith AG (eds.) *A Geological Time Scale 2004*, pp. 409–440. Cambridge: Cambridge University Press.
- Naish TR, Abbott ST, Alloway BV, et al. (1998) Astronomical calibration of a Southern Hemisphere Plio-Pleistocene reference section, Wanganui Basin, New Zealand. *Quaternary Science Reviews* 17: 695–710.
- Pillans B (2003) Subdividing the Pleistocene using the Matuyama-Brunhes boundary (MBB): An Australasian perspective. *Quaternary Science Reviews* 22: 1569–1577.
- Pillans B, Alloway B, Naish T, Westgate J, Abbott S, and Palmer A (2005) Silicic tephra in Pleistocene shallow-marine sediments of Wanganui Basin, New Zealand. *Journal of the Royal Society of New Zealand* 35: 43–90.
- Pillans B and Naish T (2004) Defining the Quaternary. *Quaternary Science Reviews* 23: 2271–2282.
- Prell WL, Imbrie J, Martinson DG, Morley JJ, Pisias NG, Shackleton NJ, and Streeter HF (1986) Graphic correlation of oxygen isotope stratigraphy: Application to the Late Quaternary. *Paleoceanography* 1: 137–162.
- Raymo ME, Ruddiman WF, Backman J, Clement BM, and Martinson DG (1989) Late Pliocene variations in the Northern Hemisphere ice sheet and North Atlantic deep water circulation. *Paleoceanography* 4: 413–446.
- Richmond GM (1996) The INQUA-approved provisional lower-middle Pleistocene boundary. In: Turner C (ed.) *The Early Middle Pleistocene in Europe*, pp. 319–327. Rotterdam: Balkema.
- Rio D, Sprovieri R, Castradori D, and Di Stephano E (1998) The Gelasian stage (upper Pliocene): A new unit of the global standard chronostratigraphic scale. *Episodes* 21: 82–87.
- Salvador A (ed.) (1994) *International Stratigraphic Guide*, p. 214. Boulder: Geological Society of America.
- Schneer CJ (ed.) (1969) *Toward a History of Geology*, p. 469. Cambridge, MA: MIT Press.
- Shackleton NJ, Berger A, and Peltier WR (1990) An alternative astronomical calibration of the lower Pleistocene timescale based on ODP Site 677. *Transactions of the Royal Society of Edinburgh* 81: 251–261.
- Shackleton NJ and Opdyke ND (1973) Oxygen isotope and paleomagnetic stratigraphy of equatorial Pacific core V28-238: Oxygen isotope temperatures and ice volumes on a 10^5 to 10^6 scale. *Quaternary Research* 3: 39–55.
- Shackleton NJ, Sanchez-Goni MF, Pailler D, and Lancelot Y (2003) Marine isotope substage 5e and the Eemian interglacial. *Global and Planetary Change* 36: 151–155.
- Walker M, Johnsen S, Rasmussen SO, et al. (2008) The Global Stratotype Section and Point (GSSP) for the base of the Holocene Series/Epoch (Quaternary System/Period) in the NGRIP ice core. *Episodes* 31: 264–267.
- Walker M, Johnsen S, Rasmussen SO, et al. (2009) Formal definition and dating of the GSSP (Global Stratotype Section and Point) for the base of the Holocene using the Greenland NGRIP ice core, and selected auxiliary records. *Journal of Quaternary Science* 24: 3–17.
- Zalasiewicz J, Smith A, Brenchley P, et al. (2004) Simplifying the stratigraphy of time. *Geology* 32: 1–4.
- Zhou LP and Shackleton NJ (1999) Misleading positions of geomagnetic reversal boundaries in Eurasian loess and implications for correlation between continental and marine sedimentary sequences. *Earth and Planetary Science Letters* 168: 117–130.

Climatostratigraphy

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In contrast to the rest of the Phanerozoic, the Quaternary has a long-established tradition of sediment sequences being divided on the basis of represented climatic changes, particularly sequences based on glacial deposits in central Europe and mid-latitude North America. This approach was adopted by early workers for terrestrial sequences because it seemed logical to divide till (glacial diamicton) sheets and nonglacial deposits or in stratigraphical sequences into *glacial* (glaciation) and *interglacial* periods, respectively (cf. West, 1977; Bowen, 1978; West, 1984). In other words, the divisions were fundamentally lithological (Figure 1). The overriding influence of climatic change on sedimentation and erosion in the Quaternary has meant that, despite the enormous advances in knowledge during the last century and a half, climate-based classification has remained central to the subdivision of the succession. Indeed, the subdivision of the modern ocean-sediment isotope stage sequence is itself based on the same basic concept. It is this approach that has brought Quaternary geology so far, but at the same time has caused considerable confusion to workers attempting to correlate sequences from enormously differing geographical and thus environmental settings. This is because of the great complexity of climatic change and the very variable effects of the changes on natural systems.

The recognition of climatic events from sediments is an inferential method and by no means straightforward. Sediments are not unambiguous indicators of contemporaneous climate, so that other evidence such as fossil assemblages, characteristic sedimentary structures (including periglacial structures) or textures, soil development, and so on must be relied upon wherever possible to illuminate the origin and climatic affinities of a particular unit. Local and regional variability of climate complicates this approach in that sequences are the result of local climatic conditions, yet there remains the need to equate them to a global scale. For at least the first half of the twentieth century, the preferred scale was that developed for the Alpine region at the turn of the century by Penck and Brückner (1909) (Figure 2).

Since the mid-twentieth century, a comparable scheme developed in northern Europe has been dominant, at least in Europe, but similar schemes were established elsewhere, such as North America, or the former USSR (Table 1). More recently, these schemes have tended to be replaced by the marine or ice-core oxygen isotope records (see *Oxygen-Isotope Stratigraphy of the Oceans; Correlations Between Greenland and Antarctica*). Today the burden of correlation lies in equating local, highly fragmentary, yet high-resolution terrestrial and shallow marine sediments on the one hand, with the potentially continuous, yet comparatively lower resolution ocean-isotope sequence on the other (cf. Gibbard and van Kolfschoten, 2005).

Terminology

Before the impact of the ocean-core isotope sequences, an attempt was made to formalize the climate-based stratigraphical

terminology in the American Code of Stratigraphic Nomenclature (1961) where so-called geologic-climate units were proposed. Here a geologic-climate unit is based on an inferred widespread climatic episode defined from a subdivision of Quaternary rocks (American Code, 1961). Several synonyms for this category of units have been suggested, the most recent being climatostratigraphical units (Mangerud et al., 1974) in which a hierarchy of terms is proposed. In subsequent, stratigraphic codes, however (Hedberg, 1976; North American Commission on Stratigraphic Nomenclature, 1983; Salvador, 1994), the climatostratigraphic approach has been discontinued since it was considered that for most of the geological column 'inferences regarding climate are subjective and too tenuous a basis for the definition of formal geologic units' (North American Commission, 1983: p 849). This view does not find favor with Quaternary scientists, however, since it is difficult to envisage a scheme of stratigraphic subdivision for recent earth history that does not specifically acknowledge the climate-change factor (Lowe and Walker, 1997). Accordingly, Quaternary stratigraphical sequences continue to be divided into geologic-climatic units based on proxy climatic indicators, and hence following with this approach, the Pleistocene–Holocene boundary (the base of the Holocene Series/Epoch) for example, is defined on the basis of the inferred climatic record. Boundaries between geologic-climate units were to be placed at those of the stratigraphic units on which they were based.

The American Code (1961) defines the fundamental units of the geologic-climate classification as follows:

A *glaciation* is a climatic episode during which extensive glaciers developed, attained a maximum extent, and receded. A *stadial* ('stade') is a climatic episode, representing a subdivision of a glaciation, during which a secondary advance of glaciers took place. An *interstadial* ('interstade') is a climatic episode within a glaciation during which a secondary recession or standstill of glaciers took place.

An *interglacial* ('interglaciation') is an episode during which the climate was incompatible with the wide extent of glaciers that characterize a glaciation.

In Europe, following the work of Jessen and Milthers (1928), it is customary to use the terms *interglacial* and *interstadial* to define characteristic types of nonglacial climatic conditions indicated by vegetational changes; *interglacial* to describe a temperate period with a climatic optimum at least as warm as the present interglacial (Holocene, Flandrian: see below) in the same region, and *interstadial* to describe a period that was either too short or too cold to allow the development of temperate deciduous forest or the equivalent of interglacial-type in the same region (West, 1977).

In North America, mainly in the United States, the term *interglaciation* is occasionally used for interglacial (cf. American Code, 1961). Likewise, the terms *stade* and *interstade* may be used instead of *stadial* and *interstadial*, respectively (cf. American Code, 1961). The origin of these terms is not certain but the latter almost certainly derive from the French language word *stade* (m), which is unfortunate since in French *stade*

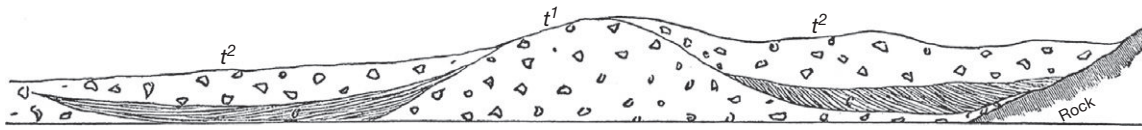


Figure 1 Fossiliferous sand deposits between tills exposed in the Cowden Burn railway cutting at Neilston in Renfrewshire, Scotland. Reproduced from Geikie J (1874) *The great Ice Age and Its Relation to the Antiquity of Man*. London: W. Isbister, Figure 27.

means (chronostratigraphical) stage (cf. Michel et al., 1997), for example, *stade isotopique marin* = marine isotope stage.

It will be readily apparent that, although in longstanding usage, the glacially based terms are very difficult to apply outside glaciated regions. Moreover, Suggate and West (1969) recognized, the term Glaciation or Glacial is particularly inappropriate since modern knowledge indicates that cold rather than glacial climates have tended to characterize the periods intervening between interglacial events over most of the earth. They therefore proposed that the term 'cold' stage (chronostratigraphy) be adopted for 'glacial' or 'glaciation.' Likewise, they proposed the use of the term 'warm' or 'temperate' stage for interglacial, both being based on regional stratotypes. Lüttig (1965) also recognized this problem and attempted to avoid the glacial connotations by proposing the terms *cryomer* and *thermomer* for cold and warm periods, respectively. These terms have found little acceptance, however. The local nature of these definitions indicates that they cannot necessarily be used across great distances or between different climatic provinces (Suggate, 1974; Suggate and West, 1969; West, 1977) or indeed across the terrestrial/marine facies boundary. In addition, it is worth noting that the subdivision into glacial and interglacial is mainly applied to the Middle and late Pleistocene.

Boundaries

Perhaps the biggest problem with climate-based nomenclature is where the boundaries should be drawn. Ideally they should be placed at the climate change, but since the events are only recognized through the responses they initiate in depositional or biological systems, a compromise must be agreed. As Bowen (1978) emphasizes, there are many places where boundaries could be drawn, but in principle they are generally placed at mid-points between temperature maxima and minima, for example, in ocean-sediment sequences. This positioning is arbitrary but is necessary because of the complexity of climatic changes. However, problems may arise when attempts are made to determine the chronological relationship of boundaries drawn in sequences of differing temporal resolution or sediment facies, and indeed determined using differing proxies. By contrast in temperate Northwest Europe the base of an interglacial or interstadial (see Overview) is very precisely defined. It is placed at the point where herb-dominated (cold-climate) vegetation is replaced by forest. The top (i.e., the base of the subsequent glaciation or cold stage) is drawn where the reverse occurs (Jessen and Milthers, 1928; Turner and West, 1968). It is unclear, however, how this relates to the timing of the actual climate change recorded or how the change is recorded by other proxies.

Climatostratigraphic Units are not Chronostratigraphic Units

In the second half of the twentieth century, it was a recognized that Quaternary time should be subdivided as far as possible in keeping with the rest of the geological column using time, or chronostratigraphy, as the basic criterion (e.g., Gibbard and West, 2000; van der Vlerk 1959). Because stages are the fundamental working units in chronostratigraphy (see Chapter Chronostratigraphy), they are considered appropriate in scope and rank for practical intraregional classification (Hedberg, 1976). However, the definition of stage-status chronostratigraphical units, with their time-parallel boundaries placed in continuous successions wherever possible, is a serious challenge especially in terrestrial Quaternary climate-dominated sequences. In these situations, boundaries in a region may be time-parallel but over greater distances problems may arise as a result of diachroneity. It is probably correct to say that only in continuous sequences that span entire interglacial–glacial–interglacial climatic cycles can an unequivocal basis for the establishment of stage events using climatic criteria be truly successfully achieved. There are the additional problems that accompany such a definition of a stage, including the question of diachroneity of climate changes and the detectable responses to those changes. For example, it is well known that there are various 'lag'-times of physical or biological responses to climatic stimuli. Thus, in short, climate-based units cannot be the direct equivalents of chronostratigraphical units because of the time-transgressive nature of former. This distinction of a stage in a terrestrial sequence from that in a marine sequence should be remembered.

In general practice today, the climatic subdivisions have been used interchangeably with chronostratigraphical stages by the majority of workers. While this approach, which gives rise to alternating 'cold' and 'warm' or 'temperate' event-based stages, has been advocated for 40 years, there remains considerable confusion about the precise distinction between the schemes, particularly among the non-geological community, that is working on Quaternary sequences. In Europe, many of the terms in current use, perhaps surprisingly, do not have defined boundary- or unit-stratotypes. This problem has been recognized and steps are now being taken to define units formally through the work of the INQUA Subcommission (Section) on European Quaternary Stratigraphy. However, many fail to see the need for this, especially those who rely on geochronology, particularly radiocarbon (see Conventional Method), for correlation. For example, despite repeated attempts to propose a Global Stratotype Section and Point boundary stratotype for the base of the Holocene Series, Pleistocene–Holocene (Weichselian–Flandrian) boundary (Olausson, 1982) in the past, only recently has a universally

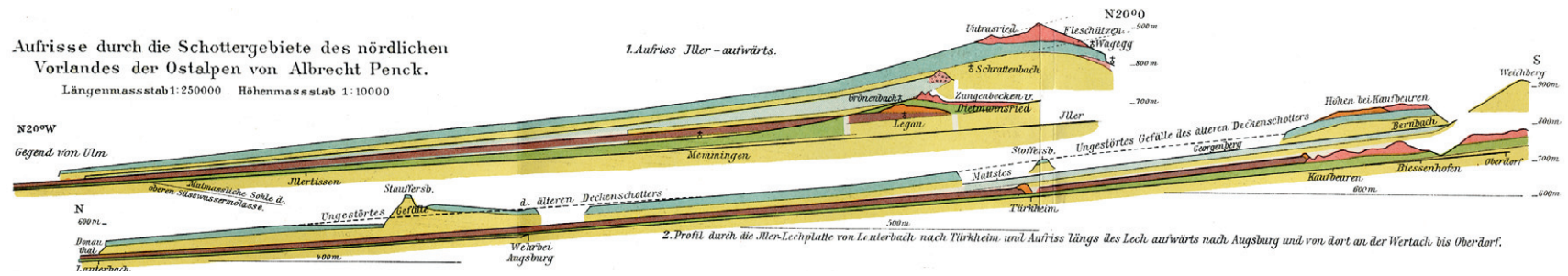


Figure 2 Sketch longitudinal sections through the gravel district of the northern Alpine Foreland of southern Germany showing the distribution of glaciofluvial sediments and their relationship to end-moraines of alpine glaciations by Albrecht Penck. These units are those upon which the original Alpine glacial classification was based. Reproduced from Penck A and Brückner E (1909) *Die Alpen im Eiszeitalter. Erster Band*, 393 pp. Leipzig: Taunitz, see [Table 1](#).

Table 1 Penck and Brückner's (1909–11) scheme of glaciations (and interglacials) in the Alpine region, compared to those proposed for other regions of the Northern Hemisphere

	<i>Alps</i>	<i>Northern Europe</i>	<i>North America</i>	<i>European USSR</i>
Authors	Penck and Brückner (1909-11)	Woldstedt (1958)	Flint (1957)	Flint (1957)
Glacial	Würm	Weichsel	Wisconsin	Valdai
Interglacial	Riss/Würm	Eem	Sangamon	Mikulino
Glacial	Riss	Saale	Illinoian	Moscow/Dneipr
Interglacial	Mindel/Riss	Holstein	Yarmouth	Lichvin
Glacial	Mindel	Elster	Kansan	Oka
Interglacial	Günz/Mindel	Cromer	Aftonian	Muchkap
Glacial	Günz		Nebraskan	?

acceptable boundary been defined (Walker et al., 2008). Subdivision of the Holocene is likely to rely on geochronology for the foreseeable future.

In addition to the formal definition of local sequences, the Pleistocene is now being subdivided into a series of 'standard stages,' comparable in scale to those defined for the preceding Neogene. The scheme, developed for marine sequences in Italy and summarized by Cita (2008), relies on the division of the sequence based on a range of parameters independent of climate. Advances in micropaleontology, paleomagnetic chronology and stable-isotope cyclostratigraphy in the upper Cenozoic marine strata of southern Italy have been used as a basis for the recognition of proposed Pleistocene global chronostratigraphical units in the same situation as the Quaternary/Pleistocene boundary stratotype, independent of climatic oscillations. In this scheme the Pleistocene is divided into the: Gelasian, Calabrian, 'Ionian,' and 'Tarentian' stages. This scheme is in the process of being defined at the time of writing (Cita, 2008).

Nomenclatural Complexities

In languages other than English, the situation is more confused. For example, in German, the terms *Glazial* and *Interglazial* (also '*Kaltzeit*' and '*Warmzeit*,' respectively) are used as equivalents of the English-language stage. Such an approach, on the face of it, seems expedient until one considers certain stages that have been correctly, formally defined in the Netherlands' Middle and early Pleistocene, which are commonly used throughout Europe. Here, for example, the Bavelian Stage includes two interglacial and two glacial events, likewise the Tiglian Stage comprises at least three interglacials and two glacials, and the Cromerian Complex Stage includes four interglacial and three glacial events (de Jong, 1988; Zagwijn, 1992). Each of these interglacials is comparable in their characteristics to the last interglacial or Eemian which is a discrete stage, also defined in the Netherlands. This problem of conflicting status has led workers to adopt the noncommittal term *complex* to group events. One example is the Saalian of Germany – originally defined as a glaciation, this chronostratigraphical stage includes at least one interglacial, as currently defined (Litt and Turner, 1993). This attempt to circumvent the nomenclatural problem by defining a 'Saalian Complex' is a

fudge at best but one that is occasioned by linguistic and long-term historical precedent, as much as by geological needs.

Global Correlation

The original intention was that 'cold' or 'warm' or 'temperate' stages should represent the first-rank climate oscillations recognized, an approach that originates from the earliest classification of ice-age events (e.g., Geikie, 1874). However, it has since been realized that some, if not all the events, are themselves internally complex. Subdivision of these stages into substages or zones was to be based, in the case of temperate stages, on biostratigraphy, and in the case of cold stages principally on lithostratigraphy and or pedostratigraphy. Within the range of radiocarbon dating (ca. 30 ka), the most satisfactory form of subdivision is frequently that based on radiocarbon years (cf. Shotton and West, 1969). However, high-resolution investigations, such as the ice-core investigations (see *Stable Isotopes, Correlations Between Greenland and Antarctica*) have allowed the recognition of ever more climatic oscillations of decreasing intensity or wavelength within the first-rank time divisions. These events are stretching the ability of the stratigraphical terminology to cope with the escalating numbers of names they generate. Terms such as 'event,' 'oscillation,' or 'phase' are currently in use to refer to short or small-scale climatic events (often referred to as 'sub-Milankovitch oscillations': see *Sub-Milankovitch (DO/Heinrich) Events*). Clear hierarchical patterns are becoming blurred but perhaps this should be seen as a positive development since the system must reflect the need to classify events that are recognized. Moreover, as our ability to resolve smaller and smaller scale oscillations increases, a more detailed nomenclature will inevitably emerge. One possible approach is to avoid attempted chronostratigraphical classification in favor of an event-scheme based on the recognition of 'diachronic units' (e.g., Curry et al., 2011) but in reality this differs little from that based on climate.

Therefore, for many Quaternary workers, chrono- and climatostratigraphical terminology remain interchangeable. Although this situation is clearly unsatisfactory, because of the imprecision that it may bring to interregional and ultimately to global correlation, realistically it is likely to continue for the foreseeable future. The long-term goal should be to clarify the situation by continuing to develop a formally

defined, chronostratigraphically based system that is fully compatible with the rest of the geological column, supported by reliable geochronology.

References

- American Commission on Stratigraphic Nomenclature (1961) Code of stratigraphic nomenclature. *Bulletin of the American Association of Petroleum Geologists* 45: 645–660.
- Bowen DQ (1978) *Quaternary Geology*. Oxford: Pergamon Press.
- Cita MB (2008) Summary of Italian marine stages. *Episodes* 31/2: 251–254.
- Curry BB, Grimley DA, and McKay ED III (2011) Quaternary glaciations in Illinois. In: Ehlers J, Gibbard PL, and Hughes PD (eds.) *Quaternary Glaciations – Extent and Chronology – A Closer Look. Developments in Quaternary Science* 15: 467–487. Elsevier: Amsterdam.
- De Jong J (1988) Climatic variability during the past three million years, as indicated by vegetational evolution in northwest Europe and with emphasis on data from The Netherlands. *Philosophical Transactions of the Royal Society of London B* 318: 603–617.
- Flint RF (1957) *Glacial and Pleistocene Geology*. New York: Wiley Interscience.
- Geikie J (1874) *The Great Ice Age and its Relation to the Antiquity of Man*. London: W. Isbister.
- Gibbard PL and van Kolfschoten Th (2005) Pleistocene and Holocene series. In: Gradstein F, Ogg J, and Smith A (eds.) *A Geologic Time Scale 2004*, p. 589. Cambridge: Cambridge University Press.
- Gibbard PL and West RG (2000) Quaternary chronostratigraphy: The nomenclature of terrestrial sequences. *Boreas* 29: 329–336.
- Hedberg HD (1976) *International Stratigraphic Guide*. New York: J. Wiley & Sons.
- Jessen K and Milthers V (1928) Stratigraphical and palaeontological studies of interglacial freshwater deposits in Jutland and north-west Germany. *Danmarks Geologisk Undersøgelse, II Række* 48.
- Litt Th and Turner C (1993) Arbeitsergebnisse der Subkommission für Europäische Quartärstratigraphie: Die Saalesequenz in der Typusregion (Berichte der SEQS 10). *Eiszeitalter und Gegenwart* 43: 125–128.
- Lowe JJ and Walker MJC (1997) *Reconstructing Quaternary Environments*, p. 446. London: Longmans.
- Lüttig G (1965) Interglacial and interstadial periods. *Journal of Geology* 73: 579–591.
- Mangerud J, Andersen STh, Berglund BE, and Donner JJ (1974) Quaternary stratigraphy of Norden, a proposal for terminology and classification. *Boreas* 3: 109–128.
- Michel J-P, Fairbridge RW, and Carpenter MSN (1997) *Dictionnaire des Sciences de la Terre*, 3rd edn., p. 346. Chichester: Masson, Paris, J. Wiley & Sons.
- Mitchell GF, Penny LF, Shotton FW, and West RG (1973) A correlation of the Quaternary deposits of the British Isles. *Geological Society of London, Special Report* 4, 99 p.
- North American Commission on Stratigraphic Nomenclature (1983) North American stratigraphic code. *American Association of Petroleum Geologists Bulletin* 67: 841–875.
- Nystuen JP (ed.) (1986) Regler og råd for navnsetting av geologiske enheter i Norge. *Norsk Geologisk Tidsskrift* 66(supplement 1) 96 pp.
- Olausson E (ed.) (1982) The Pleistocene/Holocene boundary in south-western Sweden. *Sveriges Geologiska Undersökning* C794.
- Penck A and Brückner E (1909) *Die Alpen im Eiszeitalter. Erster Band*, p. 393. Leipzig: Taunitz.
- Salvador A (1994) *International Stratigraphic Guide*, 2nd edn., p. 214. International Union of Geological Sciences and Geological Society of America.
- Shotton FW and West RG (1969) Stratigraphical table of the British Quaternary. Appendix B1. In: George TN, Harland WB, Ager DV, Ball HW, Blow HW, Casey R, Holland CH, Hughes NF, Kellaway GA, Kent PE, Ramsbottom WHC, Stubblefield J, and Woodland AW (eds.) *Recommendations on Stratigraphical Usage. Proceedings of the Geological Society of London* 165: 155–157.
- Suggate RP (1974) When did the Last Interglacial end? *Quaternary Research* 4: 246–252.
- Suggate RP and West RG (1969) Stratigraphic nomenclature and subdivision in the Quaternary. Working group for Stratigraphic Nomenclature, INQUA Commission for Stratigraphy (unpublished discussion document).
- Turner C and West RG (1968) The subdivision and zonation of interglacial periods. *Eiszeitalter und Gegenwart* 19: 93–101.
- van der Vlerk IM (1959) Problems and principles of Tertiary and Quaternary stratigraphy. *Quarterly Journal of the Geological Society of London* 105: 49–64.
- Walker M, Johnsen S, Rasmussen SO, et al. (2008) The Global Stratotype Section and Point (GSSP) for the base of the Holocene Series/Epoch (Quaternary System/Period) in the NGRIP ice core. *Episodes* 31(2): 264–267.
- West RG (1977) *Pleistocene Geology and Biology*, 2nd edn. London: Longmans.
- West RG (1984) Interglacial, interstadial and oxygen isotope stages. *Dissertationes Botanicae* 72: 345–357.
- Woldstedt P (1958) *Das Eiszeitalter. Band II*. Stuttgart: Ecke.
- Zagwijn WH (1992) The beginning of the Ice Age in Europe and its major subdivisions. *Quaternary Science Reviews* 11: 583–591.

Lithostratigraphy

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Introduction

Ever since Earth scientists have started to study the Earth they have all attempted to describe and classify what they saw and communicate their observations. Since observations in Earth science deal with the properties of strata on top of each other (in outcrops or in drillholes), these communications dealt with stratigraphy; that is, the description of the layers that were observed. Layers of rock can be described in many ways, and so it happened, and still happens. Depending on the background of the Earth scientist, description and classification was at first based on fossil content (biostratigraphy) and/or lithology (lithostratigraphy). This was especially so in the eighteenth and nineteenth century. Advances in chemistry and physics in the twentieth century enabled other methods of classification of rocks (e.g., chemo-, chrono-, magneto-, and seismostratigraphic). Although this was a clear advance in geosciences, it soon appeared that it also obscured clear communication. Fortunately, this was broadly understood and in 1976, the first edition of the International Stratigraphic Guide was published (Hedberg, 1976). This publication showed the way to clarification of concepts. Advancing knowledge in stratigraphy led to a second edition in 1994 (Salvador, 1994). This edition incorporated developments and concepts that were not included in the first edition. Recent publications by Walsh (2005a,b,c) point out the need for further clarification. This article is based on the principles of lithostratigraphic classification as defined by Hedberg (1976) and Salvador (1994) and the modifications proposed by Walsh (2005a). It is further based on the application of these principles on Quaternary deposits and on recent experiences in Northwestern Europe (Bowen, 1999a,b; Ebbing et al., 2003; Gullentops et al., 2001; Kemna and Westerhoff, 2007; McMillan, 2002, 2005; Rijdsdijk et al., 2005; Weerts et al., 2005; Westerhoff, 2009; Westerhoff et al., 2003).

Lithostratigraphic Classification

What Is Lithostratigraphy?

There are as many definitions of lithostratigraphy as there are textbooks. In short, lithostratigraphy is the classification of bodies of rock based on the observable lithologic properties of the strata and their relative stratigraphic positions (after Hedberg, 1976; Nichols, 1999; Salvador, 1994). This implies that observable lithologic properties and stratigraphic position are the only criteria to be used when defining lithostratigraphic units. In the Quaternary record, the 'rock' is often not lithified. In many definitions of lithostratigraphy, the concept of the study of outcrops is explicitly or implicitly present. Outcrops of Quaternary deposits are scarce and often short-lived. Furthermore, Quaternary deposits are absent or very thin in large

parts of the world. Thickness, however, is no criterion in defining lithostratigraphic units (Hedberg, 1976; Salvador, 1994). Apart from that, the description of rock units or deposits encountered in drillholes can replace observations at outcrops. So, the general lithostratigraphic concepts can be applied to the Quaternary sedimentary record for surficial as well as subsurface reconnaissance just as they do to older deposits.

There is common agreement on mapability as the key criterion for the validity of a lithostratigraphic unit (Bowen, 1983; Brookfield, 2004; Cox and Sumbler, 1998; Hedberg, 1976; Krumbein and Sloss, 1963; Nichols, 1999; Salvador, 1994). Typically, lithostratigraphic units are defined for and used in detailed geologic mapping (scale of about 1:25 000). In fact a new formation needs to be tested on its mapability. Therefore, a lithostratigraphic unit that cannot be mapped makes no sense. Before proceeding into definitions and considerations, we should remember that, just like any other stratigraphic classification, lithostratigraphy is only a tool! It is not a goal in itself. Its purpose is communication, interpretation, and correlation. To be able to do so we need (i) clear definitions, and (ii) clear rules of application.

There are of course several stratigraphic methods (lithostratigraphy, biostratigraphy, chronostratigraphy, sequence stratigraphy, and so on). Figure 1 shows a simplified outcrop that is classified in various ways. Note that the boundaries between the several stratigraphic units sometimes coincide but just as often they do not. One should always make clear which type of stratigraphic method and unit one uses and one should apply the definitions and rules of application of that particular type properly. This may seem tedious (and in fact it often is) but it is absolutely necessary to avoid confusion.

Following Hedberg (1976), Salvador (1994), and Walsh (2005a), a lithostratigraphic scheme: (i) has a hierarchic structure with the formation as the central unit, (ii) has a clear nomenclature, (iii) describes each unit properly, and (iiii) contains mapable units only.

Hierarchic Structure

Every rock unit has to be assigned to a formation. A formation may be divided into several members and beds, but this is not compulsory. A bed can be placed directly within a formation or within a member. Formations may be grouped into subgroups, groups, and supergroups but this is also not compulsory. At the subgroup level and higher, the lithostratigraphic units are based on interpretation rather than on observation. These higher hierarchy units consist of lower hierarchy units identifiable by observations (in core descriptions or in outcrops) that are grouped together by interpretation that usually is based on some kind of conceptual model. In our approach, lithostratigraphic units typically range from architectural element assemblages to basin fills (Figure 2).

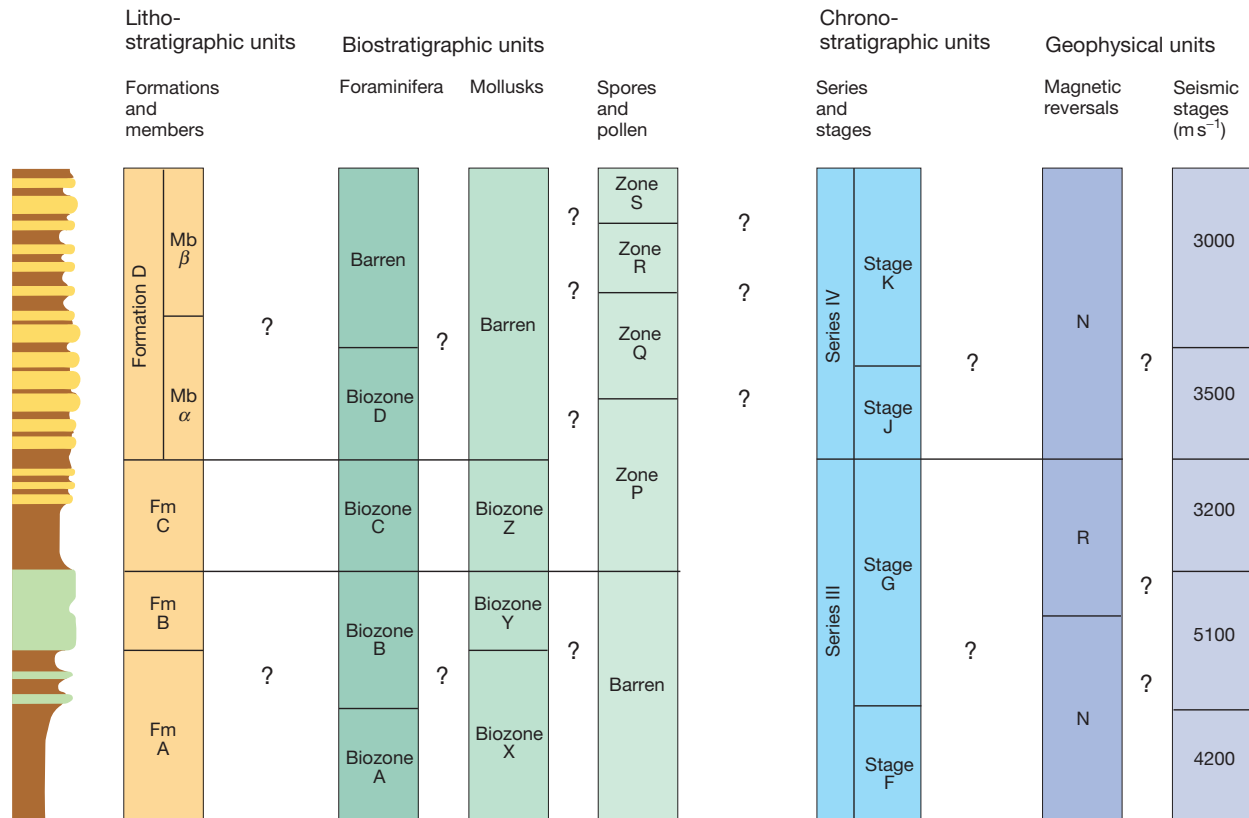


Figure 1 A hypothetical outcrop classified by several stratigraphic methods. Modified from Salvador A (ed.) (1994) *International Stratigraphic Guide: A Guide to Stratigraphic Classification, Terminology, and Procedure*. *International Subcommission on Stratigraphic Classification of IUGS International Commission on Stratigraphy*, 2nd edn., 214 p. Trondheim: International Union of Geological Sciences; Boulder, CO: Geological Society of America.

A formation in our sense is regarded as an architectural element assemblage (cf. Miall, 1985) and is therefore nearly always lithologic heterogeneous. Although this may sound strange, it is completely in line with the international guidelines. “The degree of change in lithology required to justify the establishment of distinct formations [...] is not amenable to strict and uniform rules. It may vary with the complexity of the geology of a region and the detail needed to portray satisfactorily its rock framework and to work out its geologic history” (Salvador, 1994, p. 33). Formations usually consist of a network of architectural elements (Miall, 1985, 1996) that are confined to a certain stratigraphic level. Because of the mapability criterion, we strongly discourage the widespread habit of assigning formal member or bed status to individual architectural elements or facies units. It is perfectly sound to identify these units within a formation without that formal status (Figure 3). Assigning formal lithostratigraphic status to them will in many cases lead to unjustifiable correlations. Indeed, there are numerous examples where this has happened. With respect to this, one should bear in mind that lithostratigraphy is not synonymous with lithology.

The definition of a member in our sense slightly differs from that of Salvador (1994). We propose to define a member as a named entity within the formation when it possesses lithologic properties distinguishing it from adjacent parts of the formation and/or it is present at a distinct stratigraphic level within the formation.

A bed in our view is used only to identify a well recognizable stratum (strata) that acts as a regional marker. Consequently, we advise limited use of the bed. A bed is typically less than 2 m thick. It should be noted that we deviate from the guidelines of Hedberg (1976) and Salvador (1994) here.

Rules of Application and Nomenclature

Defining a lithostratigraphic unit

Reading the nomenclatural regulations of any branch of science usually makes scientists yawn. Yet, it is an aspect of science that deserves our attention. In that light the publications of Walsh (2005a,b,c) provide interesting reading for Earth scientists dealing with stratigraphy. Although his papers mainly deal with bio- and chronostratigraphy, his observations are valid in lithostratigraphy as well. We use many of his views here.

Every lithostratigraphic unit should be properly defined and named. Hedberg (1976) and Salvador (1994) provide nomenclatural rules. Each unit should have a name that consists of two parts: the name of the unit (*nominal* part) and the rank of the unit (*hierarchical* part). The name should be derived from the location where the unit is defined, based on a drill-hole or an outcrop at that location. The drillhole or outcrop serves as its *stratotype*. A stratotype should be accessible for other researchers. In the Quaternary record, drillholes usually serve as stratotypes. Because the deposits themselves cannot be

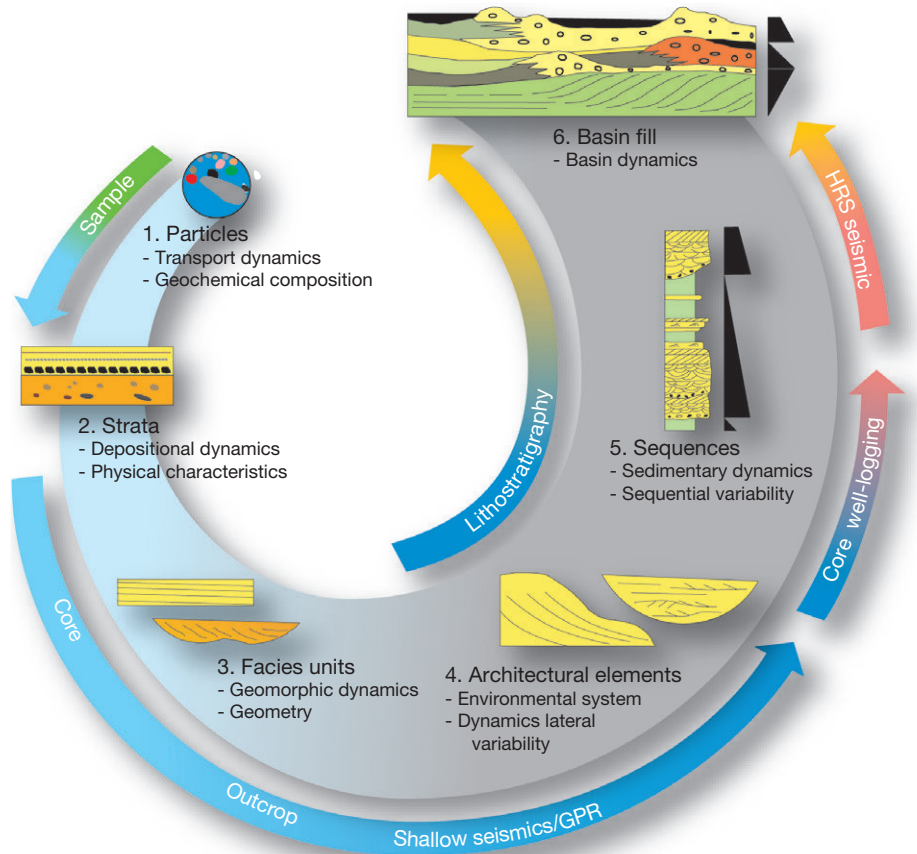


Figure 2 Lithostratigraphy and scale. Lithostratigraphic units typically range from architectural elements and sequences to basin fills. Adapted from Heinz J and Aigner Th (2003) Hierarchical dynamic stratigraphy in various Quaternary gravel deposits, Rhine glacier area (SW Germany): Implications for hydrostratigraphy. *International Journal of Earth Sciences* 92: 923–938.

observed at the stratotype location, authors should provide an adequate description of the deposits in the drillhole along with the exact location of the drillhole.

Let us now assume that we have an excellent drillhole near the Eem river in the Netherlands (Figure 4(a) and 4(b)). At a certain level in that drillhole, marine deposits are present. These deposits can be recognized in the descriptions of the drillholes that are present in the database of the Geological Survey of the Netherlands. The deposits can be mapped over a large area and Lorié (1906) already described them. Zonneveld (1959, p. 54) presents a map of the deposits, compiled from work by Bennema and Pons (1952) and Van der Heide (1957). He calls the deposits 'Eem beds' (Dutch: Eemlagen; Zonneveld, 1959, pp. 56–58) but does not introduce them as a lithostratigraphic unit. This was done by Zagwijn (1961). Because he uses a series of drillholes near the Eem river to define the unit, the name of the unit is 'Eem.' Assigning a rank to a unit is always more or less arbitrary. We know that the Eem unit consists of siliciclastic marine deposits that can be easily recognized from core descriptions (based on their marine mollusks) and that the unit has a widespread occurrence at a

certain stratigraphic level. The lithologic properties of the unit strongly differ from the bounding units. Thus, when the unit was first introduced by Zagwijn (1961, p. 20) it was defined as a formation; the Eem Formation. Unfortunately, his definition was based on mixture of litho- and biostratigraphic criteria. He includes organic deposits 'related' to the marine deposits in the Eem Formation because of their (palynological inferred!) age. It also lacks a clear stratotype. A formal stratotype was defined in the formal definition of the Eem Formation by Doppert et al. (1975, pp. 48–49; summarized in Van Staalduinen et al., 1979, pp. 42–43). The excellent drillhole 32B0119 of the Geological Survey of the Netherlands serves as the original (holo)stratotype in their definition (Figure 4(a)). Apart from the marine deposits described above they also assign related organic deposits to the Eem Formation.

Nomenclatural rules

Once a lithostratigraphic unit has been defined, changes in its status should be documented properly. From mapping a given unit, it might, for instance, appear that the original stratotype only describes a part of the deposits of that unit. In that case,

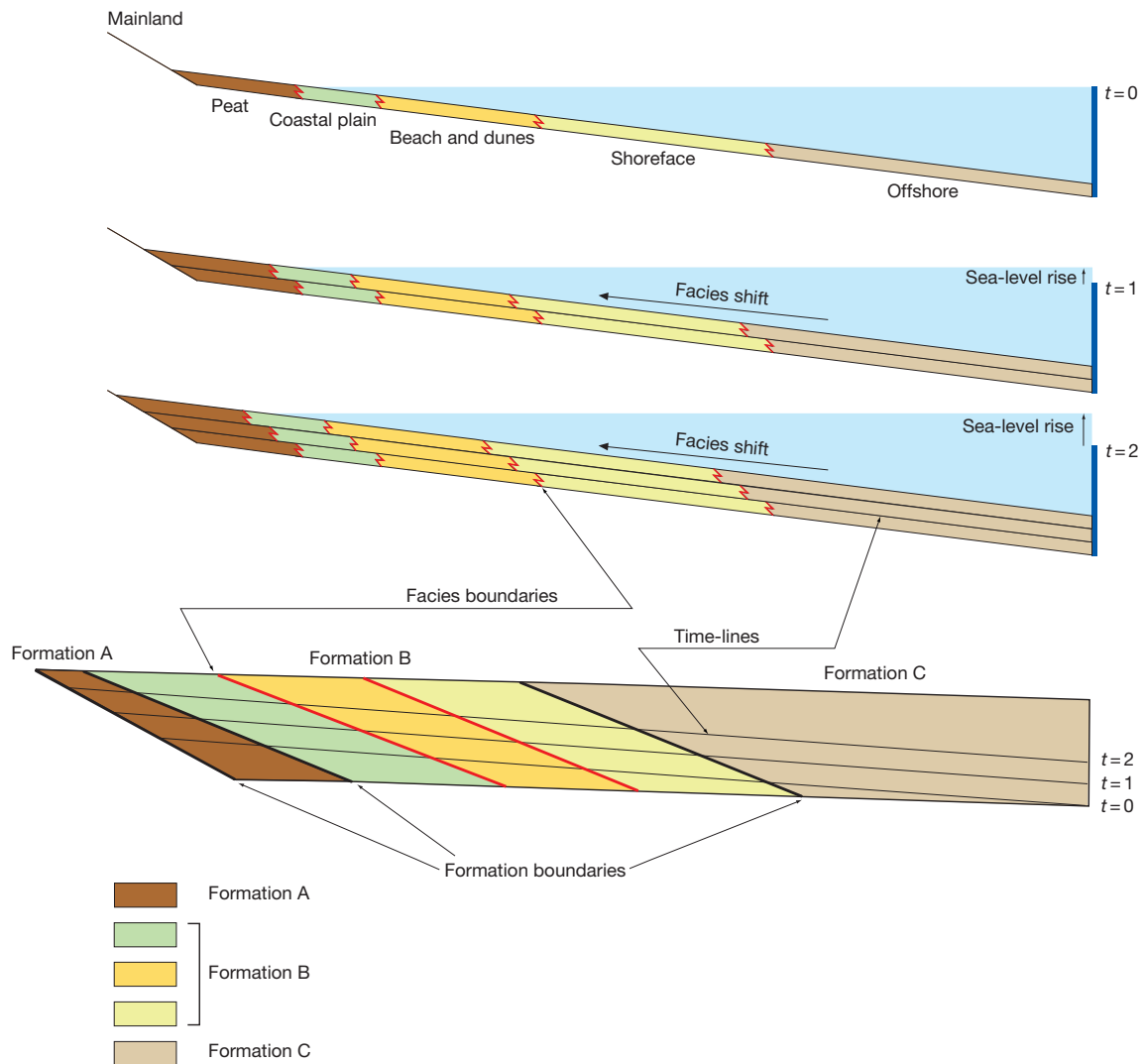


Figure 3 Relationships between the boundaries of lithostratigraphic units (defined by lithological characteristics resulting from the depositional environment) and timelines in a succession of strata formed during gradual sea-level rise (modified from Nichols G (1999) *Sedimentology and Stratigraphy*, p. 238. Oxford: Blackwell Science). Note that the facies units cross timelines; age is no criterion in a lithostratigraphic classification. Also note the displacement of the depositional environment when sea level rises. The continental deposits (peat) form formation A; the coastal plain (predominantly clay), beach and dune (sand), and the shoreface deposits (sand) form formation B; the offshore deposits (sand) form formation C. The three sandy environments (i.e., beach and dunes, shoreface, and offshore) can be distinguished by their different lithologic properties (sand median, sedimentary structure, specific mollusk content).

one or more stratotypes should be added. Sometimes, the original stratotype even has to be replaced by a new one. Hedberg (1976) and Salvador (1994) give nomenclatural rules, which are summarized in Table 1. The original stratotype of a lithostratigraphic unit is by definition its holostatotype. The original author might add one or more parastratotypes if (s)he deems this necessary. Modifications defined afterwards lead to the definition of lecto-, neo-, or hypostatotypes. These rules must be obeyed to be able to always exactly know the status of a given stratotype. When looked at in detail, it appears that Hedberg's and Salvador's nomenclature only reveals the historic status of the stratotype. Walsh (2005a, pp. 320–323) therefore proposes another nomenclatural classification for stratigraphic units. He introduces three formal classes of stratotypes: (i) nominal, (ii) exemplary, and (iii) boundary-defining

stratotypes (Table 2). Walsh's (2005a) classification scheme defines the function of the stratotypes. We propose to use both Hedberg's/Salvador's nomenclatural rules and Walsh's rules for stratotype definition. This means that every stratotype has to be given both a status and a function. Walsh's (2005a,b,c) papers are mainly concerned with bio- and chronostratigraphy. In those fields, boundary-defining stratotypes are important. The nature of the unit boundaries changes over space. Therefore, the practical utility of boundary-defining stratotypes in lithostratigraphy is very limited. In lithostratigraphy, the nominal and exemplary functions of the stratotypes are of main importance. Caution should be taken in defining exemplary stratotypes too strictly as this leads to very rigid interpretations. We prefer to use loose exemplary stratotypes to describe the deposits of a given unit. If perception of a certain lithostratigraphic unit changes, its

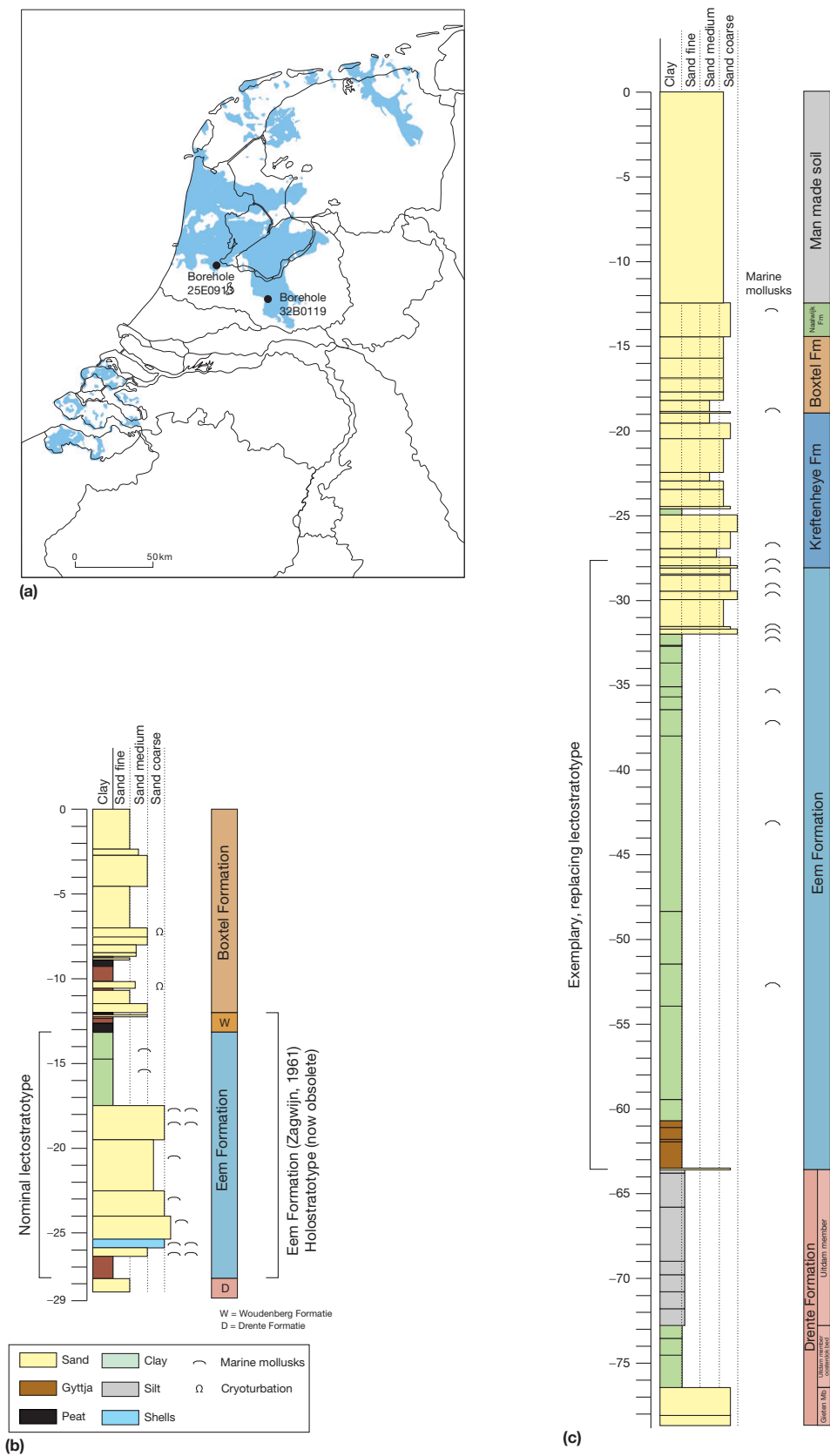


Figure 4 Application of the nomenclatural rules to the Eem Formation, the Netherlands. (a) Simplified map of the occurrence of the Eem Formation including the location of the stratotypes. (b) Original stratotype of the Eem Formation (drillhole 32B0119), now obsolete. (c) New stratotype of the Eem Formation (drillhole 25E0913). See text for further explanation.

Table 1 Stratotype nomenclature, status

<i>Stratotype name</i>	<i>Stratotype status</i>
Holostratotype	The original stratotype designated by the original author at the time of proposing the stratigraphic unit
Lectostratotype	A stratotype for a previously defined stratigraphic unit selected later in the absence of an adequately defined original stratotype (defining lectostratotype) or A changed interpretation of the original holostratotype (changing lectostratotype) or A new stratotype that gives a more adequate description of the stratigraphic unit than its obsolete holostratotype (replacing lectostratotype)
Neostratotype	A new stratotype selected to replace an older one which has been destroyed or otherwise made inaccessible
Parastratotype	An additional stratotype used in the original definition by the original author to illustrate the diversity and/or heterogeneity of the stratigraphic unit or some critical feature that is not evident or exposed in the holostratotype
Hypostratotype	A stratotype proposed after the original definition of the holostratotype (and parastratotypes) in order to extend the knowledge of the unit to other geographic areas where it was not recognized earlier (extending hypostratotype) or A stratotype that expresses regional differences of the unit that were not defined by the original author (regional hypostratotype)

Note that we extended the original meaning of the lecto- and the parastratotype.

When using those status stratotypes, authors should indicate the *exact* status.

Source: Modified from Hedberg HD (ed.) (1976) *International Stratigraphic Guide. A Guide to Stratigraphic Classification, Terminology, and Procedure*, 200 p. New York: John Wiley; Salvador A (ed.) (1994) *International Stratigraphic Guide: A Guide to Stratigraphic Classification, Terminology, and Procedure. International Subcommission on Stratigraphic Classification of IUGS International Commission on Stratigraphy*, 2nd edn., 214 p. Trondheim: International Union of Geological Sciences; Boulder, CO: Geological Society of America.

hierarchical rank might also change. It is perfectly sound to 'demote' or 'promote' a formation to a member or (sub)group, as long as the arguments are published and the previous rank of the newly defined unit is described in its history. In the 'demotion' case, the original holostratotype of the demoted unit will probably serve as an exemplary regional hypostratotype of the new 'mother' unit. Finally, it might appear that a certain lithostratigraphic unit is not valid. In that case, the unit should be abandoned. 'Its' deposits should be incorporated in one or more existing or newly defined units. In both cases, this should be included in the definition of these units.

Let us now return to our Eem Formation example. We have to assign one or more functions to the holostratotype (Table 3). As the name of the Eem Formation is derived from the river Eem that flows near the location of drillhole 32B0119, it is a nominal stratotype. The drillhole gives an example of Eem Formation deposits, so it is also an exemplary stratotype.

Table 2 Functional stratotype nomenclature. See text for further details

Exemplary stratotype	A specific stratigraphic section used to serve as a concrete example of an author's concept of one or more attributes of a particular stratigraphic unit
Nominal stratotype	A specific stratigraphic section that serves as the name bearer of a particular stratigraphic unit
Strict stratotype	All of the stratotype must always be included in the concept denoted by the name of the stratigraphic unit which it bears
Loose stratotype	At least some part of the stratotype must always be included in the concept denoted by the name of the stratigraphic unit which it bears
Boundary-defining stratotype	A specific stratigraphic section in which one or two specified horizons are used to define one or both boundaries of a particular stratigraphic unit. Such boundary definitions may be permanent or subject to modification with new data
Unit stratotype	A specific, physically superposed stratigraphic section whose designated base and top are used to define both the base and the top of a particular stratigraphic unit, both being defined by permanent <i>golden spikes</i>
Boundary stratotype	A specific, physically superposed stratigraphic section in which one specified horizon is used to define the base or the top of a particular stratigraphic unit, being defined by a permanent <i>golden spike</i>

Source: Modified from Walsh SL (2005a) The role of stratotypes in stratigraphy.

Part 1. Stratotype functions. *Earth-Science Reviews* 69: 307–332.

In the original definition also, the boundaries of the unit were defined, so it has a third function as well as it is a unit stratotype. Finally, drillhole 32B0119 also served as the stratotype of the Eemian stage in the Netherlands (Zagwijn, 1961). Walsh (2005c) discusses at length the 'age-based-on-formation' principle, based on considerations by Wood et al. (1941). Although scientifically correct as long as the correct nomenclature is used (in our case 'Eem Formation' for the lithostratigraphic unit and 'Eemian stage' for the chronostratigraphic unit), we consider this practice unfortunate. It easily leads to confusion and it should be discouraged when introducing new units. For nomenclatural stability's sake, it is hardly possible – and indeed undesirable – to change widely used existing 'age-based-on-formation' nomenclature. However, every stratigrapher should be aware of the possible confusion (we refer the readers to Lawson (1979) and Walsh (2005a,b,c) for further reading on this topic).

With the ongoing 1:50 000 mapping of the Netherlands, it became clear that the deposits in the 32B0119 drillhole only represented a small part of the facies present in the Eem Formation. Furthermore, the deposits in the stratotype 'represented' only a small part of the Eemian stage (e.g., Jelgersma and Breeuwer, 1975; Zagwijn, 1983; several 1:50 000 Geological Survey of the Netherlands map sheets and their explanatory annotations). This called for a new stratotype both for the Eem Formation and for the Eemian stage. In 1997, the

Table 3 Changing stratotypes of the Eem Formation

Author	Stratotype	Status	Function
Zagwijn (1961)	Not given. Three interpreted sections near the Eem river in his paper probably are meant to serve as stratotype	Not defined	Not defined, probably nominal, exemplary and unit
Doppert et al. (1975)	Drillhole 32B0119 near the Eem river, interval between 12.00 and 27.71 m below the surface	Holostratotype	Nominal, exemplary and unit
Bosch et al. (2003)	Drillhole 25E0913 at Amsterdam Terminal, interval between 28.10 and 63.50 m below the surface	Replacing lectostratotype	Loose exemplary
	Drillhole 32B0119 near the Eem river, interval between 13.21 and 27.71 m below the surface	Changing lectostratotype	Nominal
Bosch et al. (2006)	Drillhole 25E0913 at Amsterdam Terminal, interval between 28.10 and 63.50 m below the surface	Replacing lectostratotype	Loose exemplary
	Drillhole 32B0119 near the Eem river, interval between 13.21 and 27.71 m below the surface	Changing lectostratotype	Nominal
	Drillhole P5-4 (P5-2 in Oele, 1971), interval between 2.75 and 6.0 m below the surface	Regional hypostratotype for the Eem Formation, holostratotype for the Brown Bank Member	Loose exemplary for the Eem Formation, nominal for the Brown Bank Member

Geological Survey of the Netherlands carried out a new drill-hole in Amsterdam from which undisturbed cores were taken in a former glacial basin that was formed during the Saalian glacial. The location was chosen because it was suitable for both chrono- and lithostratigraphic redefinition of the stratotype (De Gans et al., 2000). The 'Eemian' sediment units (sic!) in the drillhole were described in detail by Van Leeuwen et al. (2000, pp. 166–173). When the Eem Formation was redefined, Bosch et al. (2003) replaced the original holostratotype of the formation with this new drillhole (25E0913; Figure 3). Bosch et al. (2003) returned to the original concept of the Eem Formation. In their view, it only comprises the recognizable marine deposits at that particular stratigraphic level. The 'related' organic deposits are excluded from the formation. It had further appeared that a part of the deposits were fluvial reworked. The reworked marine deposits can be recognized based on their coarser lithology, often containing gravel (e.g., see Busschers et al., 2005; De Gans and De Groot, 1995; Törnqvist et al., 2000). They were also excluded from the formation. Both changes concern only a minor part of the original Eem Formation deposits. Finally, Bosch et al. (2003) included the former Brown Bank Formation as a member into the Eem Formation. The Brown Bank unit was recognized as a mapable unit of predominantly marine origin of early Weichselian age by Oele (1971) and Cameron et al. (1984, 1989). It occurs at the same stratigraphic level as the Eem Formation. Age, however, is no reason for assigning the deposits to another formation (Figure 4). (Note that this is a fine example of the confusion that often arises when the 'age-based-on-formation' principle is misunderstood!)

What are the nomenclatural consequences of all these changes? Should we rename the Eem Formation to Amsterdam Formation? No! Clearly, the original holostratotype is no longer valid because it only describes a small part of the formation. The *concept* of the formation, however, has not changed. So, the name Eem Formation is maintained. The original holostratotype is now a *changing lectostratotype*. It retains its function as *nominal* stratotype but has lost its other functions. Drillhole 25E0913 is the new *loose exemplary, replacing lectostratotype* (erroneously

Table 4a Requirements of a lithostratigraphic unit description

Unit property	Contents
Name	Name and rank of the unit Name and rank of next higher hierarchic unit <i>Derivatio nominis</i>
Description of the lithologic properties	Lithology (general, dominant, subordinate, sporadic) Nature of the boundaries (lower and upper) Other characteristic features Regional differential lithologic properties, including short descriptions of potential units of lower rank that form part of the unit Thickness (minimal, maximal, average)
Stratotypes	Status and function stratotypes cf. Tables 2 and 3, including exact intervals in drillholes and/or sections Geographic coordinates, including a small detailed map of the stratotype area Description of the drillhole(s) or section(s), exactly indicating the unit interval Indicative map of the areal extent of the unit
Genesis	Description of the depositional environment(s), as far as relevant for the unit's lithologic properties
Relation to other (litho)stratigraphic units	Indication of possible confusion with other units due to lithologic similarities and/or intercalation
Relation to earlier described and defined units	Name of the older unit(s) that is (are) replaced by this one, if relevant Original publication where the present unit was defined, if relevant
Age	As precisely as possible, either chronostratigraphic or absolute
References	References deemed necessary for understanding the nature and properties of the unit

classified as parastratotype by Bosch et al., 2003). Although it may have boundary-defining use in chronostratigraphy, it is not defined as such in lithostratigraphy. Drillhole P5-4 (P5-2 in

Oele, 1971) serves as the *nominal* and *loose exemplary* stratotype for the Brown Bank Member and as a *regional hypostatotype* for the Eem Formation (Bosch et al., 2006).

Description of a Unit

A formal lithostratigraphic unit should be properly described. The description should contain its lithologic properties,

stratigraphic position, and (historic) nomenclature. It should also contain relevant information on its genesis, its (often presumed) age, its relation with other lithostratigraphic units. The latter is of extreme importance for mapping and modeling purposes. Table 4a summarizes the requirements of an adequate description (derived from the British Geological Survey and the Geological Survey of the Netherlands). Table 4b summarizes the description of the Eem Formation.

Table 4b Short description of the Eem Formation as example

<i>Eem formation</i>	
Name	<i>Unit:</i> Eem Formation <i>Mother unit:</i> Upper North Sea Group cf. Van Adrichem Boogaert & Kouwe (eds., 1993) <i>Derivation nominis:</i> Harting (1874, 1875) used the term 'Eem' to describe marine deposits in the subsurface near the Eem river. Zagwijn (1961) and Doppert et al. (1975) adopt the name for formal the lithostratigraphic unit
Description of the lithologic properties	<i>General lithology:</i> Gray calcareous sand (median 150–420 µm) bearing remains of marine mollusks, and gray calcareous clay bearing marine mollusks <i>Dominant lithology:</i> Gray calcareous sand (median 150–420 µm) bearing marine mollusks <i>Subordinate lithology:</i> Gray calcareous clay bearing remains of marine mollusks; marine shell beds <i>Sporadic lithology:</i> Diatomite (gyttja) <i>Lower boundary:</i> The Eem Formation usually rests conformably on glacial deposits of the Drente Formation from which it can easily be distinguished. Its base usually consists of sand, except in the glacial basins where thick beds of clay form the base of the unit. To the south of the maximum Saalian ice extent, the unit rests nonconformably on older sandy to clayey units <i>Upper boundary:</i> The nature of the upper boundary of the Eem Formation varies from sharp to diffuse and erosive to nonerosive over short distances. A general description of the upper boundary cannot be given <i>Other characteristic features:</i> Typical for the unit is the occurrence of marine mollusks, with diagnostic species (e.g., <i>Bittium reticulatum</i>, <i>Venerupis aurea</i> var. <i>senescens</i>, <i>Echinocyamus pusillus</i>, <i>Ostrea</i>) <i>Regional differential lithologic properties:</i> In the Saalian glacial basins, clay beds of >10 m thickness are present (>25 m in the replacing lectostratotype). Within the formation one unit of lower rank is discerned: ● Brown Bank Member (after Cameron et al., 1984; 1989; Rijdsdijk et al., 2005; Oele, 1971). This unit consists of (regressive) marine to lacustrine silty clay and fine sand of early Weichselian age. It occurs only in the present North Sea where it forms the top of the formation <i>Thickness:</i> The unit is typically around 10–15 m thick. In the glacial basins, thickness can be as much as 70 m. To the south of the Saalian glaciation, the unit is thin (1–10 m)
Stratotypes	Drillhole 25E0913 at Amsterdam Terminal, interval between 28.10 and 63.50 m below the surface; loose exemplary, replacing lectostratotype (Van Leeuwen et al., 2000) Drillhole 32B0119 near the Eem river, interval between 13.21 and 27.71 m below the surface; nominal, changing lectostratotype (Cleveringa et al., 2000; Doppert et al., 1975; Zagwijn, 1961) Drillhole P5-4 (P5-2 in Oele, 1971), interval between 2.75 and 6.0 m below the surface; loose exemplary regional hypostatotype for the Eem Formation, nominal holostratotype for the Brown Bank Member
Genesis	Marine, from transgressive at its base to regressive near the top. In the deeper parts of the southern North Sea, the unit grades into (glacial) lacustrine deposits
Relation to other (litho)stratigraphic units	Because the original 'Eem' stratotype also serves as the nominal stratotype for the Eemian stage, the lithostratigraphic Eem Formation is often interpreted chronostratigraphical. Also, the unit is often wrongly correlated to nonmarine units of Eemian age
Relation to earlier described and defined units	The unit comprises the Eem Formation as defined by Zagwijn (1961) and Doppert et al. (1975) and the former Brown Bank Formation (Cameron et al., 1984, 1989; Oele 1971) with the exception of: ● 'related' peat beds that now form part of the Woudenberg Formation ● predominantly coarse-grained fluvial deposits containing reworked marine shells, now part of the Kreftenheye Formation The formation was earlier defined by: ● Zagwijn WH (1961) Vegetation, climate and radiocarbon datings in the Late Pleistocene of the Netherlands. Part I. Eemian and Early Weichselian. <i>Mededelingen Geologische Stichting N.S.</i> 14: 15–45 ● Doppert JW Chr., Ruegg GHJ, van Staaldin CJ, Zagwijn WH, and Zandstra JG (1975) Formaties van het Kwartair en Boven-Tertiair in Nederland. In: Zagwijn WH and van Staaldin CJ (eds.), <i>Toelichting bij geologische overzichtskaarten van Nederland</i>, pp. 11–56. Haarlem: Rijks Geologische Dienst
Age	Upper Pleistocene (Eemian to Middle Weichselian)
References	We refer the reader to Bosch et al. (2006), who give a list of relevant publications. Printing of the list here is beyond the scope of this paper

Examples from the Quaternary Record

Splitting or Lumping? Scale and Hierarchy of Lithostratigraphic Units

When defining lithostratigraphic units from outcrops or from a single drillhole, some workers are inclined to identify as many units as possible. They like to split up the sedimentary record in many formal lithostratigraphic units. Their goal is to have a detailed local framework which often seems to serve a detailed scheme of climate-induced events (Van Huissteden et al., 1986; Zagwijn and Paepe, 1968). Others argue that those detailed local frameworks hamper communication and correlation on a higher level. These workers prefer to identify few formal lithostratigraphic units. They describe the internal lithologic variation within the lithostratigraphic units in terms of facies variation. In other words, they lump facies associations into lithostratigraphic units. Splitting or lumping is a matter of taste. Based on our mapping and modeling experiences, we prefer the latter. If one clearly identifies and describes the facies units within a given lithostratigraphic unit, communication and correlation is as easy at the local level as it is at higher scale levels. Further, a proliferation of names is inhibited.

As stated above, the formation is the central unit in lithostratigraphic classification. One of the main problems in applying lithostratigraphy in mapping is scale. In the past, many Quaternary geologists defined formations from outcrops or drillholes when working in a given area. Workers in other areas did the same. When mapping proceeded, it often appeared that they had defined the same unit, or parts of a unit that was defined elsewhere also. Figure 5 gives an example from glacial units in northern America (after Karrow, 1989). Karrow's figure suffices for communication and correlation in the Ohio–Ontario region. Things worsen, however, if the glacial units are correlated further away. This soon leads to a proliferation of names. The same happened in Great Britain where workers in the continental record defined many 'glaciogenic' and 'fluvial' formations. Bowen (1999a,b) attempted to clarify the nomenclatural jungle that had developed over the years, despite the efforts of the authors their classification does not always lead to clarification. In particular, lithostratigraphic

units of fluvial and glacial origin were 'grouped' together without an attempt to regroup them in order to get rid of the proliferation of names at the formation level. Bowen (1999b, p. 7) recognized the problem, but apparently he and his co-authors were not able to solve it at the time. McMillan (2005) clearly reduces the confusion in defining three glaciogenic groups (with many regional subgroups) and the Britannia Catchment Group with 24 regional subgroups. All the previously defined fluvial and glaciogenic lithostratigraphic units are maintained and placed into the subgroups. Although admirable and clearly a major clarification, this places the subgroup in the center of the lithostratigraphic classification. This does not really matter as long as one realizes that it deviates from the concepts of Hedberg (1976) and Salvador (1994).

In the Netherlands, a completely revised lithostratigraphic classification of the Tertiary and Quaternary was published in 2000 (available on the internet via www.dinoloket.nl), 2003 (Westerhoff et al., 2003), and 2005 (Rijsdijk et al., 2005). Here, all lithostratigraphic units were placed in four genetic categories (i.e., marine, fluvial, glacial, and local deposits). To maintain a clear link with the existing lithostratigraphic classification of the complete (predominantly sedimentary) record by Van Adrichem Boogaert and Kouwe (1993), the Upper Tertiary (Neogene) and Quaternary deposits were all placed in the Upper North Sea Group. The number of Quaternary formations defined by Doppert et al. (1975) slightly decreased. Many new formations contain several members.

Let us now consider the fluvial deposits of the Meuse in the new classification scheme. They are present in the southeastern part of the Netherlands (Figure 6). South of the Feldbiss Fault, Meuse deposits have been preserved in a well-documented flight of river terraces dating from the Pliocene to the Holocene (e.g., Felder and Bosch, 1989; Van den Berg, 1996; Van den Berg and van Hoof, 2001). In the subsiding area north of the fault, the Meuse deposits form a stacked unit with a terrace expression for the youngest units only (Saalian and younger). The Meuse deposits mainly consist of coarse sand and gravel. Locally, clay and loam beds are present. In South Limburg, the quartz content of the deposits increase with age (= height of the terraces). In McMillan's (2005) concept, the Meuse deposits would be placed together in a 'Meuse' Catchment Subgroup that would contain several formations. Westerhoff and Weerts (2003) and Westerhoff et al. (2003) chose to place the deposits into one formation, the Beegden Formation (named after its newly defined nominal, loose exemplary holotype; drillhole 58D0203 near the town of Beegden, Westerhoff and Weerts, 2003). No members are discerned in the thick stacked Beegden Formation deposits north of the Feldbiss fault, simply because they cannot be properly mapped. Consequently, the Meuse terrace deposits in South Limburg are included into the Beegden Formation as Members. River terraces are really morphostratigraphic units. To be valid as lithostratigraphic unit, each 'terrace' member needs to be described lithostratigraphically and it needs to have one or more stratotypes. Note that the Dutch approach maintains the formation as the central unit of a lithostratigraphic classification.

Another example from the revised lithostratigraphic scheme concerns the definition of Upper Pliocene and Lower Pleistocene fluvial deposits in the southern part of the Netherlands. This fluvial record results from the interplay of Rhine,

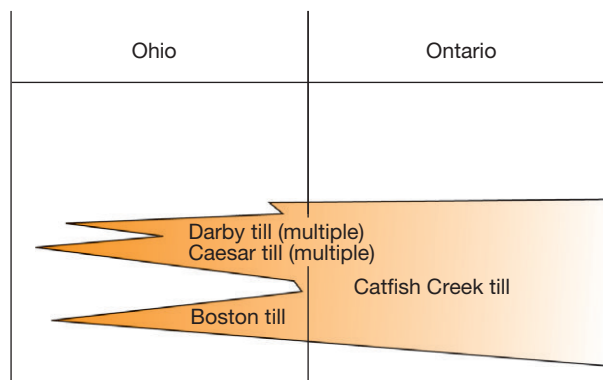


Figure 5 Correlation of independently defined glacial units in the Ohio–Ontario region (modified from Karrow PF (1989) Quaternary continental stratigraphy and neocatastrophism. *Quaternary Science Reviews* 8: 277–282). It is clear that the Boston, Caesar and Darby tills are a part of the Catfish Creek till unit. See text for further explanation.

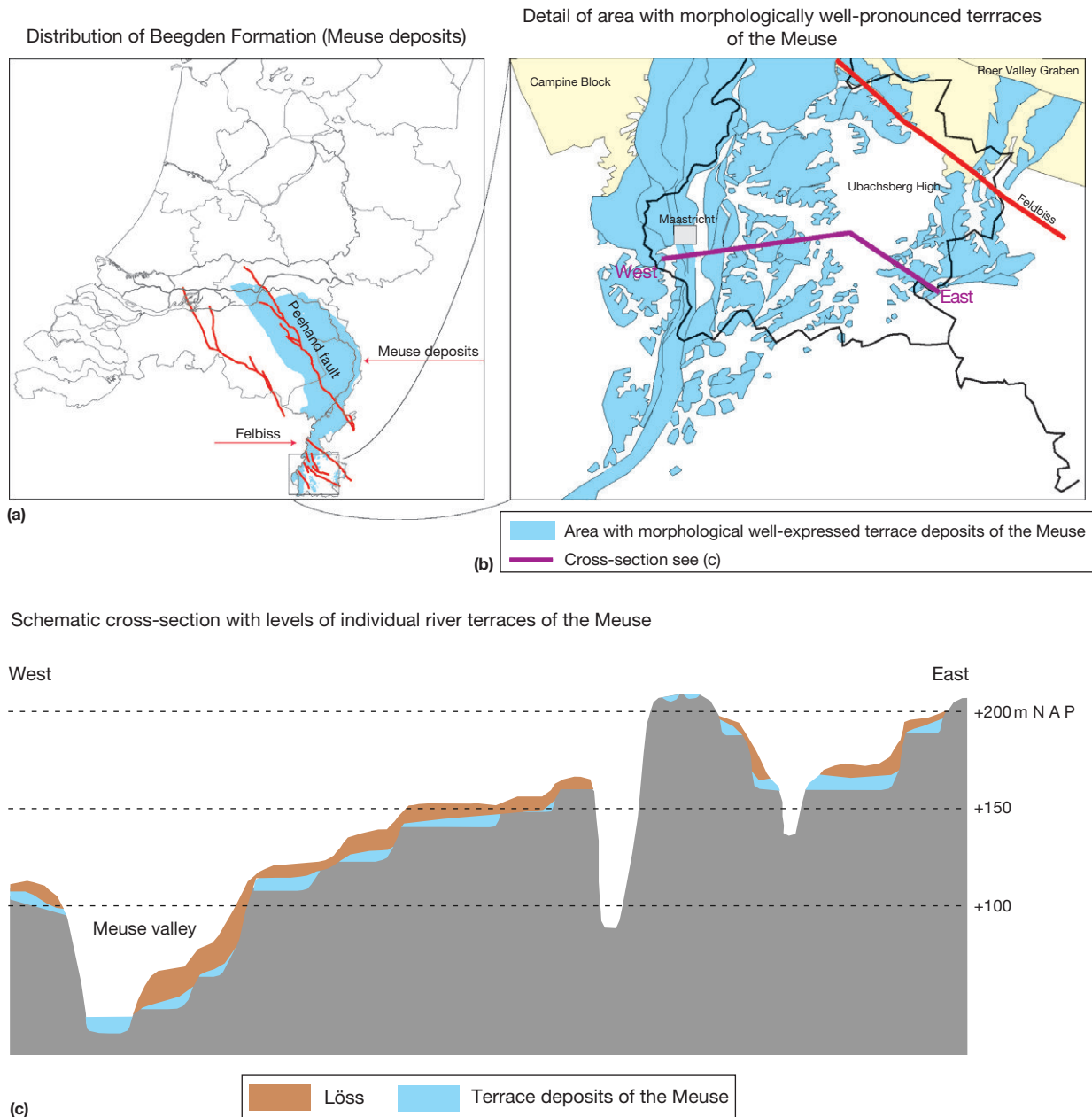


Figure 6 Meuse deposits in the southeastern part of the Netherlands (a) and in South Limburg (b and c). See text for further explanation. Modified from Van den Berg MW (1996) *Fluvial Sequences of the Maas. A 10 Ma Record of Neotectonics and Climate Change at Various Time-Scales*, 181 pp. PhD Thesis, University Wageningen.

Meuse, and smaller rivers draining the central and northern part of Belgium. Lithostratigraphically, their deposits were defined by a mixture of litho-, chrono-, and biostratigraphical criteria. Pollen-defined (sub)stages, however, had a major control on the definition of lithostratigraphical units.

This lithostratigraphic subdivision was developed during the first half of the twentieth century. Based on differences in macro plant remains, which showed a climatic deterioration, a distinction was made between the Pliocene Reuver Clay and the Pleistocene Tegelen Clay (Reid and Reid, 1915). It is noteworthy that this (litho)stratigraphic assignment concerned only the clay deposits. The associated fluvial sand

deposits were not encountered in this subdivision. Subsequent research on mammal remains and pollen analysis confirmed that the Plio–Pleistocene transition could be placed in between the two clay deposits (Meijer, 1998; Zagwijn, 1998). Additional research on the sediment-petrographical composition showed a predominant stable heavy-mineral content (a.o. tourmaline, metamorphic minerals, staurolite, zircon) for the sand deposits below the Reuver Clay. The fluvial deposits found in between the Reuver and Tegelen Clay, however, showed a predominance of unstable heavy minerals (a.o. garnet, epidote, alterite/saussurite, hornblende). For a long time, it was thought that this major shift in heavy-mineral

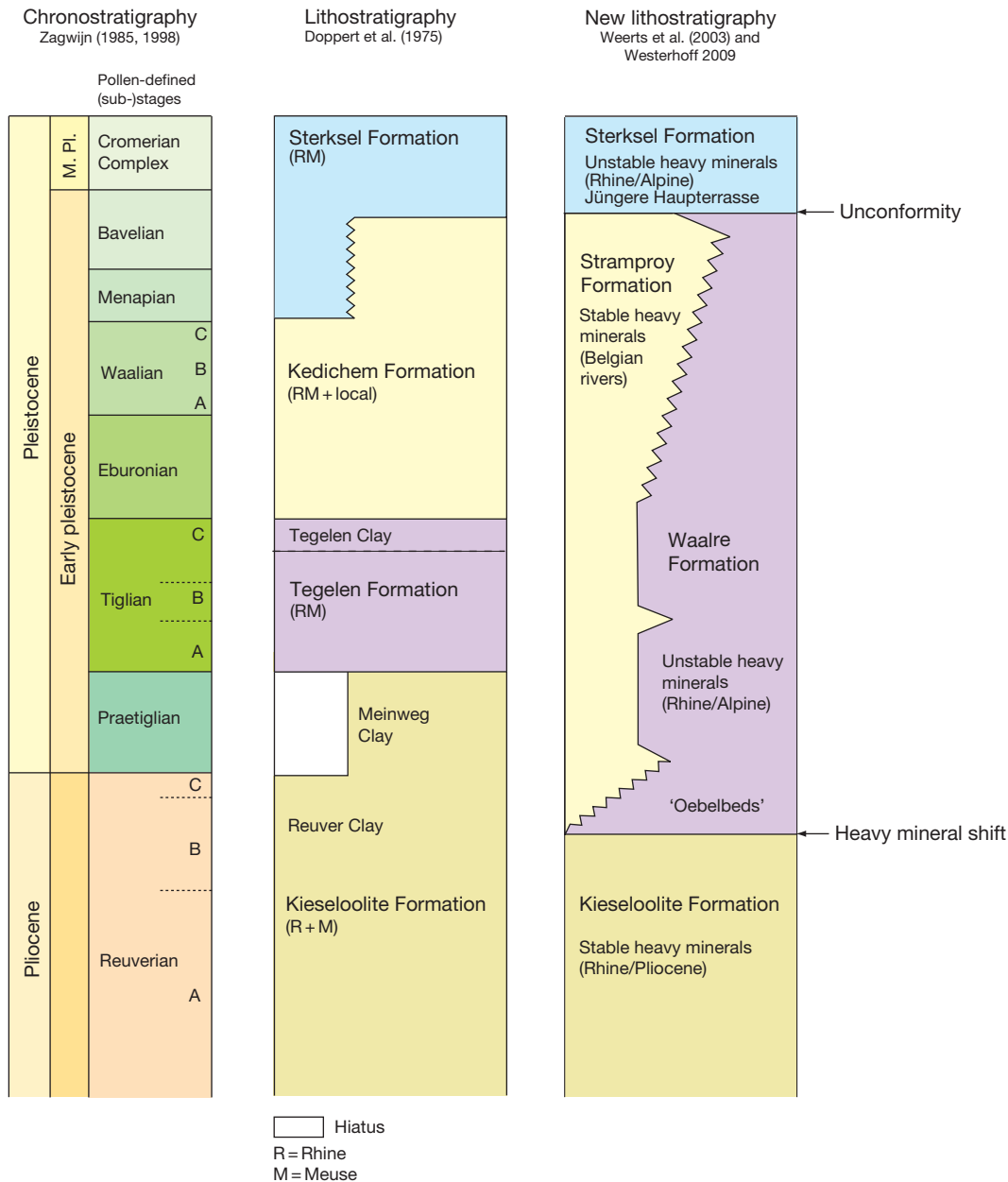


Figure 7 Lithostratigraphy of the Upper Pliocene and early Pleistocene fluvial deposits in the SE Netherlands (Lower Rhine area). The chronostratigraphy is based on the work published by Zagwijn (1985, 1998). The now abandoned lithostratigraphy compared with Doppert et al. (1975) is shown in the middle column. At the right site, the revised lithostratigraphy (Weerts et al., 2003; Westerhoff, 2009) is given.

composition marked the biologically typified Plio–Pleistocene transition (Zonneveld, 1947a,b). The combined evidence was summarized in the lithostratigraphical scheme of Doppert et al. (1975). In this scheme, the Upper Pliocene and Lower Pleistocene fluvial deposits are assigned to three superimposed lithostratigraphic formations (Figure 7). (i) The lower unit comprises all Pliocene fluvial deposits and was assigned as Kieseloolite Formation. (ii) The middle unit is known as Tegelen Formation, a mixed Rhine–Meuse deposit with an assumed (pre)Tiglian to Early Eburonian age. (iii) The third unit is termed the Kedichem Formation. It overlies the Tegelen Formation and comprises all fluvial (mixed Rhine–Meuse and

Belgian rivers) and local deposits that postdate the Tiglian Stage.

New mapping results based on a large number of borehole and well-logging data obtained since 1975, have demonstrated that the interrelationship of the lithostratigraphic units discerned is much more complex than hitherto thought. Sediments with a lithological and petrographical composition similar to the Tegelen Formation were already deposited before the end of the Pliocene but also during post-Tiglian time (Boenigk, 1970, 1978, 2002; Boenigk and Frechen, 2006; Kemna, 2005; Kemna and Westerhoff, 2007). Deposits belonging to the Kedichem Formation comprise material derived

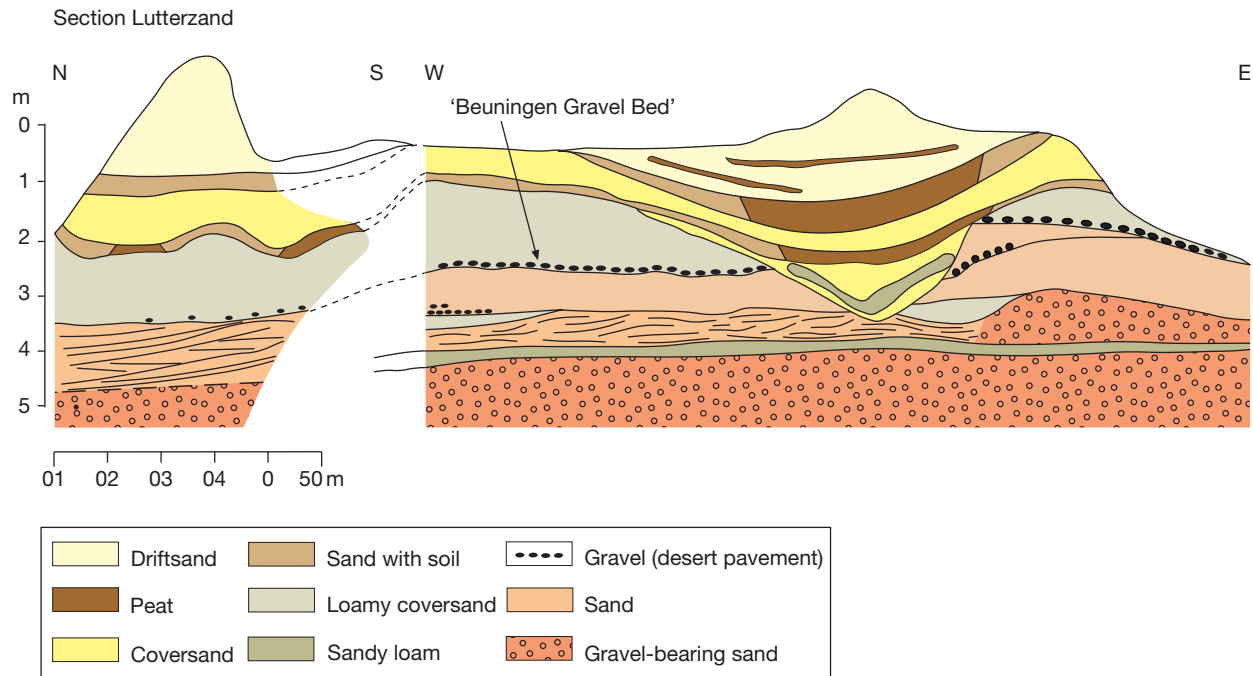


Figure 8 General section of a part of the Lutterzand exposures with the 'Beuningen Gravel Bed' and other desert pavements. Note that only in a part of the section there is one gravel bed present. In other parts, two or three beds are present while gravel beds are lacking in the northern part of the section. Modified from Wijmstra TA and Schreve-Brinkman EJ (1971) *The Lutterzand section, VI*. In: Van der Hammen T and Wijmstra TA (eds.) *Upper Quaternary of the Dinkel Valley. Mededelingen Rijks Geologische Dienst, Nieuwe Serie 22*: 87–99.

from two main and different source areas (i.e., the Rhine–Meuse river system and rivers that drained the central and northern part of Belgium) and interdigitate clearly with deposits of the Tegelen Formation. It is worthwhile mentioning that, according to the definition (cf. Doppert et al., 1975) and mainly determined by their supposed ages, all three of the formerly defined formations encompass river terrace deposits of the Meuse that occur in the southern part of Limburg.

Because of the reasons mentioned above, we have carried out a redefinition of the Upper Pliocene and Lower Pleistocene fluvial deposits in the southern part of the Netherlands (Ebbing et al., 2003; Weerts et al., 2003; Westerhoff, 2009; Westerhoff et al., 2003). The content as well as the concept behind the two previously defined Tegelen and Kedichem Formations have changed considerably and therefore we also have replaced the original names of these formations (Salvador, 1994; Walsh, 2005a). Instead, two new formations are introduced (Westerhoff, 2009):

- The Waalre Formation describes mixed Rhine–Meuse deposits formed during the late Pliocene and early Pleistocene. The deposits are dominated by unstable ('Rhine') heavy-mineral associations.
- The Stramproy Formation encompasses deposits that are mainly formed by rivers draining the central and northern part of Belgium. The heavy-mineral composition comprises dominantly a stable assemblage.

Both new formations occur next to each other and interdigitate repeatedly. The distinction between the two formations is primarily based on the differences in provenance and is macroscopically well observable in the field and in borehole

samples. Furthermore, both formations have a clear stratigraphical position as they are bounded by two easily mapable marker horizons. The lower marker horizon reflects the marked change in petrography between the pre-Rhine deposits of the Kieseloolite Formation (typified by stable heavy-mineral associations) and the (Alpine) Rhine deposits (typified by unstable heavy-mineral associations). The upper horizon is marked by the coarse-grained gravel-bearing base of the Sterksel Formation which represents a major unconformity that occurs in large parts of the depositional area.

Compared to the former Tegelen and Kedichem Formations, the lower and upper boundary of the Waalre and Stramproy Formations are not determined by pollen-defined substages (i.e., Reuverian, Early Eburonian) but by a clear lithostratigraphical position that is distinguished using easily mapped major changes in lithology.

The 'Bed' in a Lithostratigraphic Sense

As stated earlier, a bed in our view is used only to identify a well recognizable stratum (strata) that acts as a regional marker. Consequently, we advise limited use of the bed in a lithostratigraphic sense. A bed is typically less than 2 m thick. Traditionally, beds are recognized from outcrops. A widely used bed in the Quaternary of the Netherlands is the 'Beuningen Gravel bed' that was defined by Van der Hammen et al. (1967) and Van der Hammen (1971) in the Dinkel area (Figure 8). This bed represents a deflation level that by the authors was directly linked to a distinct chronostratigraphic level in the Upper Pleniglacial of the Weichselian (note the direct link of inferred age and lithostratigraphic unit). It consists of a few

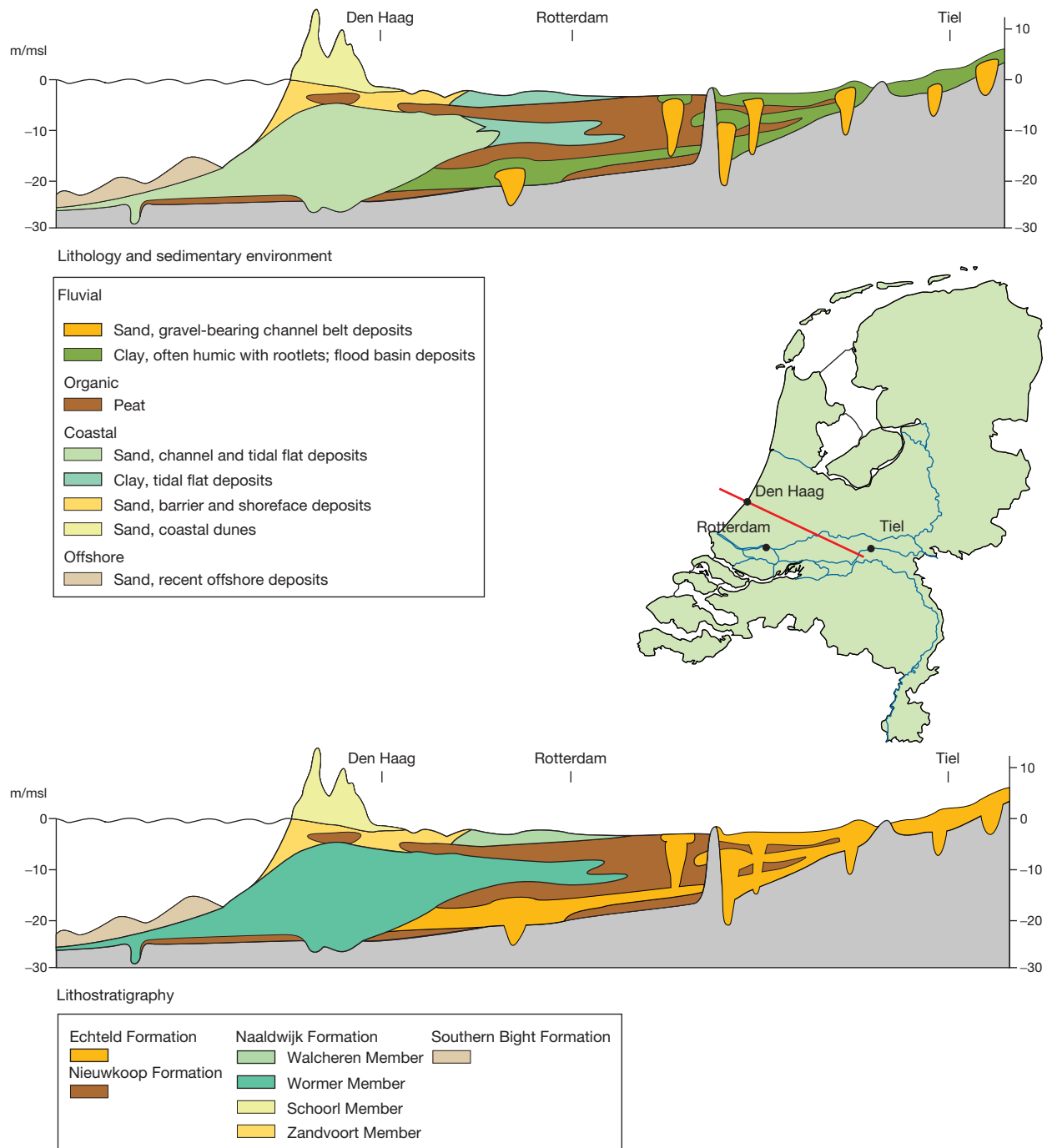


Figure 9 Holocene fluvial and coastal plain deposits in an across section from the western Netherlands. In the upper part, details on lithology and sedimentary facies are shown. The lower section shows the lithostratigraphy at the formation level. Note the stratigraphic position of the Basal Peat Bed between older underlying deposits and the base of the Naaldwijk Formation. Reproduced from Weerts HJT, Westerhoff WE, Cleveringa P, Bierkens MFP, Veldkamp JG, and Rijdsdijk KF (2005) Quaternary geological mapping of the lowlands of the Netherlands, a 21st century perspective. *Quaternary International* 133–134: 159–178; modified from Van der Spek (1994).

centimeters-thin bed of coarse sand and fine gravel that is often seen in outcrops and encountered in drillholes in the eastern part of the Netherlands. Now if one studies schematic sections based on outcrops in the type area by [Wijmstra and Schreve-Brinkman \(1971\)](#) and [Wijmstra and Schalke \(1971\)](#), it is obvious that there are several deflation levels present in the area,

all represented by centimeter-thin coarse sand and gravel. In some sections three or four levels are present, while at other locations, none are found due to subsequent erosion ([Figure 8](#)). Should we now use this 'bed' as a formal lithostratigraphic unit? The Beuningen Gravel Bed is a facies unit; a centimeter-thin coarse-grained desert pavement. Given the

paleoclimatic conditions during the Weichselian pleniglacial in the Netherlands, those desert pavements can be expected at several levels. This was also recognized by Van Huissteden et al. (1986, p. 57), who represented the Beuningen Gravel Bed as diachronous in their schemes, but failed to dismiss it as a lithostratigraphic unit. When one does not have outcrops at his hand, it is impossible to say if one encounters the 'true' Beuningen Gravel Bed or just another local deflation level. So, (i) the unit does not represent a clear regional marker and (ii) it is impossible to map. Conclusion: The Beuningen Gravel Bed should be abolished as lithostratigraphic unit.

Another widely used bed in Dutch stratigraphy is de basal peat bed ('Basisveen' or basal peat bed). It represents the peat that underlies the clastic wedge of Holocene deposits in the coastal plain that formed as a result of the Holocene rising sea level. It has been described and used by many authors under slightly different names (e.g., see De Jong and Hageman, 1960, p. 649, 'Basisveen'; De Mulder and Bosch, 1982, p. 120, 'Lower Peat'; Hageman, 1969, p. 375, 'Lower Peat'; Vos and Van Heeringen, 1997, p. 21 'Basal Peat'; several 1:50 000 Geological Survey of the Netherlands map sheets and their explanatory annotations). It was used both as a formal and an informal unit, sometimes described as a member and sometimes as a bed. Weerts et al. (2003) defined and formally described the 'Basisveen Laag' (English: Basal Peat Bed) as a lithostratigraphic bed (see also Westerhoff et al., 2003, p. 351, 'Basisveen Laag'; Weerts et al., 2005, p. 166, 'Basal Peat Bed'). The Basal Peat Bed always consists of a few centimeters to 2–3 m thick peat bed that rests on top of pretransgressive deposits and it is always covered by Holocene transgressive clastic marine or tidal deposits of the Naaldwijk Formation (Figure 9). As such, the organic formation of Figure 3 might represent the bed. Note that its stratigraphic position is always clear, but that the unit is highly diachronous! Its occurrence is widespread. In our view, this unit clearly represents a lithostratigraphic bed because it (i) occurs at an unambiguous, important stratigraphic level and (ii) it can easily regionally be mapped.

Conclusion

In this article, we gave a brief outline of how to apply a lithostratigraphic approach on Quaternary deposits. A main starting point is the adage that lithostratigraphy is a tool for communication among Earth scientists themselves and users of geological information. It, therefore, needs to be applied with clear rules and definitions. Around the world there are many examples of confusing stratigraphic schemes which hamper the clarification of the Earth's history and moreover the practical use of geological information in applied geoscientific studies. Especially, the latter will benefit from clearly defined lithostratigraphic units that are capable to serve as carrier for sediment of layer properties in for example numerical modeling.

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See also: **Fluvial Environments: Sediments, Glacial Landforms, Sediments: Glaciomarine Sediments and Ice-Rafted Debris. Quaternary Stratigraphy: Chronostratigraphy; Climastratigraphy; Continental Biostratigraphy; Morphostratigraphy/Allostratigraphy; Overview; Pedostratigraphy; Sequence Stratigraphy; Tephrochronology.**

References

- Benneba J and Pons LJ (1952) Donken, fluviaal Laagterras en Eemzee-afzettingen in het westelijk gebied van de grote rivieren. *Boor en Spade* 5: 126–137.
- Boenigk W (1970) *Zur Kenntnis des Altquartärs bei Brüggem (westlichen Niederrhein)*. Germany: Sonderveröffentlichungen Geologisches Institut Universität Köln 17: 1–141.
- Boenigk W (1978) Gliederung der altquartären Ablagerungen in der Niederrheinischen Bucht. *Fortschritte Geologie Rheinland und Westfalen* 28: 135–212.
- Boenigk W (2002) The Pleistocene drainage pattern in the Lower Rhine Basin. *Netherlands Journal of Geosciences/Geologie en Mijnbouw* 81: 201–209.
- Boenigk W and Frechen M (2006) The Pliocene and Quaternary fluvial archives of the Rhine system. *Quaternary Science Reviews* 25: 550–574.
- Bosch JHA, Busschers FS, and Weerts HJT (2003) Eem Formatie. Beschrijving lithostratigrafische eenheid. Description of the unit, to be found at www.dinoloket.nl.
- Bosch JHA, Busschers FS, Weerts HJT, and Passchier S (2006) Eem Formatie. Beschrijving lithostratigrafische eenheid. Description of the unit, to be found at www.dinoloket.nl.
- Bowen DQ (1983) *Quaternary Geology: A Stratigraphic Framework for Multidisciplinary Work*, 237 pp. Oxford: Pergamon Press.
- Bowen DQ (ed.) (1999a) *A Revised Correlation of Quaternary Deposits in the British Isles*, 174 pp. London: Geological Society. Special Report No. 23.
- Bowen DQ (1999b) On the correlation and classification of Quaternary deposits and land-sea correlations. In: Bowen DQ (ed.) *A Revised Correlation of Quaternary Deposits in the British Isles*, pp. 1–9. London: Geological Society. Special Report No. 23.
- Brookfield ME (2004) *Principles of Stratigraphy*, 340 pp. Oxford: Blackwell Publishing.
- Busschers FS, Weerts HJT, Wallinga J, et al. (2005) Sedimentary architecture and optical dating of Middle and Late Pleistocene Rhine-Meuse deposits – Fluvial response to climate change, sea-level fluctuation and glaciation. *Netherlands Journal of Geosciences/Geologie en Mijnbouw* 84: 25–41.
- Cameron TDJ, Laban C, and Schüttenhelm RTE (1984) Flemish Bight. Sheet 52° N/02° E. Quaternary Geology, 1:250,000 series British Geological Survey and Geological Survey of the Netherlands.
- Cameron TDJ, Laban C, and Schüttenhelm RTE (1989) Middle and Upper Pleistocene and Holocene stratigraphy in the Southern North Sea between 52° and 54° N, 2° to 4° E. In: Henriët JP and de Moor G (eds.) *The Quaternary and Tertiary Geology of the Southern Bight, North Sea*, pp. 119–135. Belgium: University of Ghent.
- Cleveringa P, Meijer T, van Leeuwen RJW, et al. (2000) The Eemian stratotype locality at Amersfoort in the central Netherlands: A re-evaluation of old and new data. In: Van Kolfschoten Th and Gibbard PL (eds.) *The Eemian – Local Sequences, Global Perspectives. Geologie en Mijnbouw/Netherlands Journal of Geosciences* 79: 197–216.
- Cox BM and Sumbler MG (1998) Lithostratigraphy: Principles and practice. In: Doyle P and Bennett MR (eds.) *Unlocking the Stratigraphical Record. Advances in Modern Stratigraphy*, pp. 11–27. Chichester: Wiley.
- De Gans W, Beets DJ, and Centineo MC (2000) Late Saalian and Eemian deposits in the Amsterdam glacial basin. In: Van Kolfschoten Th and Gibbard PL (eds.) *The Eemian – Local Sequences. Global Perspectives. Geologie en Mijnbouw/Netherlands Journal of Geosciences* 79: 147–160.
- De Gans W and De Groot TAM (1995) The lower Rhine delta. In: Schirmer W (ed.) *Quaternary Field Trips in Central Europe*, pp. 550–557. München: Friedr. Pfeil.

- De Jong JD and Hageman BP (1960) De legenda voor de Holocene afzettingen op de nieuwe geologische kaart van Nederland, schaal 1:50.000. *Geologie en Mijnbouw* 39: 644–653.
- De Mulder EFJ and Bosch JHA (1982) Holocene stratigraphy, radiocarbon datings and palaeogeography of central and northern North-Holland (the Netherlands). *Mededelingen rijks geologische dienst* 36: 111–160.
- Doppert JW Chr, Ruegg GHJ, van Staalduinen CJ, Zagwijn WH, and Zandstra JG (1975) Formaties van het Kwartair en Boven-Tertiair in Nederland. In: Zagwijn WH and van Staalduinen CJ (eds.) *Toelichting bij geologische overzichtskaarten van Nederland*, pp. 11–56. Haarlem: Rijks Geologische Dienst.
- Ebbing JHJ, Weerts HJT, and Westerhoff WE (2003) Towards an integrated land-sea stratigraphy of the Netherlands. *Quaternary Science Reviews* 22: 1579–1587.
- Felder WM and Bosch PW (1989) *Geologische kaart van Zuid-Limburg en omgeving. Afzettingen van de Maas*. Haarlem: Rijks Geologische Dienst.
- Gullentops F, Bogemans F, de Moor G, Paulissen E, and Pissart A (2001) Quaternary lithostratigraphic units (Belgium). In: Bultynck P and Dejonghe L (eds.) *Guide to a Revised Lithostratigraphic Scale of Belgium*. *Geologica Belgica* 4(1–2): 153–164.
- Hageman BP (1969) Development of the western part of The Netherlands during the Holocene. *Geologie en Mijnbouw* 48: 373–388.
- Harting P (1874) De bodem van het Eemland. *Verslagen Koninklijke Akademie van Wetenschappen, Afdeling Natuurkunde II* (deel VIII): 282–290.
- Harting P (1875) Le Système Émien. *Archives Néerlandaises Sciences Exactes et Naturelles de la Société Hollandaises des Sciences (Haarlem)* 10: 443–454.
- Hedberg HD (ed.) (1976) *International Stratigraphic Guide. A Guide to Stratigraphic Classification, Terminology, and Procedure*, p. 200. New York: Wiley.
- Jelgersma S and Breeuwer JB (1975) Toelichting bij de kaart glaciële verschijnselen gedurende het Saalien, 1: 600.000. In: Zagwijn WH and van Staalduinen CJ (eds.) *Toelichting bij geologische overzichtskaarten van Nederland*, pp. 93–103. Haarlem: Rijks Geologische Dienst.
- Karrow PF (1989) Quaternary continental stratigraphy and neocatastrophism. *Quaternary Science Reviews* 8: 277–282.
- Kemna HA (2005) Pliocene and lower Pleistocene Stratigraphy in the Lower Rhine Embayment, Germany. *Kölner Forum für geologie und Paläontologie* 14: 1–121.
- Kemna HA and Westerhoff WE (2007) Remarks on the palynology-based chronostratigraphical subdivision of Pliocene terrestrial deposits in NW-Europe. *Quaternary International* 164–165: 185–196.
- Krumbein WC and Sloss LL (1963) *Stratigraphy and Sedimentation*, 2nd edn., p. 660. San Francisco: W.H. Freeman.
- Lawson JD (1979) Stability of stratigraphical nomenclature. *Newsletter on Stratigraphy* 7: 159–165.
- Lorié J (1906) De geologische bouw der Geldersche Vallei, benevens beschrijving van eenige nieuwe grondboringen. VII. *Verhandelingen Koninklijke Akademie van Wetenschappen*, Amsterdam, Sectie 2, Deel 13: 100 pp.
- McMillan AA (2002) Onshore Quaternary geological surveys in the 21st century – A perspective from the British Geological Survey. *Quaternary Science Reviews* 21: 889–899.
- McMillan AA (2005) A provisional Quaternary and Neogene lithostratigraphical framework for Great Britain. *Netherlands Journal of Geosciences/Geologie en Mijnbouw* 84: 87–107.
- Meijer T (1998) References of relevant publications about Pliocene and Early Pleistocene deposits in the Netherlands. *Mededelingen Nederlands Instituut voor Toegepaste Geowetenschappen TNO* 60: 579–602.
- Miall AD (1985) Architectural-element analysis: A new method of facies analysis applied to fluvial deposits. *Earth-Science Reviews* 22: 261–308.
- Miall AD (1996) *The Geology of Fluvial Deposits. Sedimentary Facies, Basin Analysis, and Petroleum Geology*, 582 pp. Berlin: Springer.
- Nichols G (1999) *Sedimentology and Stratigraphy*, 355 pp. Oxford: Blackwell Science.
- Oele E (1971) The Quaternary geology of the Dutch part of the North Sea, North of the Frisian Isles. *Geologie en Mijnbouw* 48: 467–480.
- Reid CI and Reid EM (1915) The Pliocene Floras of the Dutch–Prussian border. *Mededelingen Rijksopsporingsdienst voor Delfstoffen* 6: 1–178.
- Rijsdijk KF, Passchier S, Weerts HJT, Laban C, van Leeuwen RJW, and Ebbing JHJ (2005) Revised Upper Cenozoic stratigraphy of the Dutch sector of the North Sea basin: Towards an integrated lithostratigraphic and allostratigraphic approach. *Netherlands Journal of Geosciences/Geologie en Mijnbouw* 84: 129–146.
- Salvador A (ed.) (1994) *International Stratigraphic Guide. A Guide to Stratigraphic Classification, Terminology, and Procedure*. International Subcommission on Stratigraphic Classification of IUGS International Commission on Stratigraphy, 2nd edn., p. 214. Trondheim: International Union of Geological Sciences. Boulder, CO: Geological Society of America.
- Törnqvist TE, Wallinga J, Murray AS, de Wolf H, Cleveringa P, and de Gans W (2000) Response of the Rhine-Meuse system (west-central Netherlands) to the last Quaternary glacio-eustatic cycles: A first assessment. *Global and Planetary Change* 27: 89–111.
- Van Adrichem Boogaert HA and Kouwe WFP (eds.) (1993) Stratigraphic Nomenclature of the Netherlands, Revision and Update by RGD and NOGEP. *Mededelingen Rijks Geologische Dienst* 50: 1–39 (page numbering renewed in each chapter).
- Van den Berg MW (1996) *Fluvial Sequences of the Maas. A 10 Ma Record of Neotectonics and Climate Change at Various Time-Scales*, 181 pp. PhD Thesis, University Wageningen.
- Van den Berg MW and van Hoof T (2001) The Maas terrace sequence at Maastricht, SE Netherlands: Evidence for 200 m of late Neogene and Quaternary uplift. In: Maddy D, Macklin MG, and Woodward JC (eds.) *River Basin Sediment Systems, Archives of Environmental Change*, pp. 45–86. Rotterdam: Balkema.
- Van der Hammen T (1971) The Upper Quaternary stratigraphy of the Dinkel Valley, II. In: Van der Hammen T and Wijmstra TA (eds.) Upper Quaternary of the Dinkel Valley. *Mededelingen Rijks Geologische Dienst, Nieuwe Serie* 22: 59–72.
- Van der Hammen T, Maarleveld GC, Vogel JC, and Zagwijn WH (1967) Stratigraphy, climatic succession and radiocarbon dating of the last glacial in the Netherlands. *Geologie en Mijnbouw* 46: 79–95.
- Van der Heide S (1957) Correlations of marine horizons in the Middle and Upper Pleistocene of the Netherlands. *Geologie en Mijnbouw Nieuwe Serie* 19: 272.
- Van der Spek AJF (1994) Large-scale evolution of Holocene tidal basins in the Netherlands. PhD Thesis, Universiteit Utrecht, Utrecht, The Netherlands.
- Van Huissteden K, Vandenberghe J, and van Geel B (1986) Late Pleistocene Stratigraphy and Fluvial History of the Dinkel Basin (Twente, Eastern Netherlands). *Eiszeitalter und Gegenwart* 36: 43–59.
- Van Leeuwen RJW, Beets DJ, Bosch JHA, et al. (2000) Stratigraphy and integrated facies analysis of the Eemian in Amsterdam-Terminal. In: Van Kolfschoten Th and Gibbard PL (eds.) *The Eemian – Local Sequences, Global Perspectives*. *Geologie en Mijnbouw/Netherlands Journal of Geosciences* 79: 161–196.
- Van Staalduinen CJ, van Adrichem Boogaert HA, Bles MJM, et al. (1979) The Geology of the Netherlands. *Mededelingen Rijks Geologische Dienst* 31(2): 9–49.
- Vos PC and Van Heeringen RM (1997) Holocene geology and occupation history of the Province of Zeeland (SW Netherlands). *Mededelingen Nederlands Instituut voor Toegepaste Geowetenschappen TNO* 59: 5–109.
- Walsh SL (2005a) The role of stratotypes in stratigraphy. Part 1. Stratotype functions. *Earth-Science Reviews* 69: 307–332.
- Walsh SL (2005b) The role of stratotypes in stratigraphy. Part 2. The debate between Kleinpell and Hedberg, and a proposal for the codification of biochronologic units. *Earth-Science Reviews* 70: 47–73.
- Walsh SL (2005c) The role of stratotypes in stratigraphy. Part 3. The Wood Committee, the Berkely school of North American mammalian stratigraphic paleontology, and the status of provincial golden spikes. *Earth-Science Reviews* 70: 75–101.
- Weerts H, Cleveringa P, Ebbing JHJ, de Lang FD, and Westerhoff WE (2003) De lithostratigrafische indeling van Nederland. Formaties uit het Tertiair en Kwartair. *Rapport NITG-TNO 03-051-A* (Internal report).
- Weerts HJT, Westerhoff WE, Cleveringa P, Bierkens MFP, Veldkamp JG, and Rijsdijk KF (2005) Quaternary geological mapping of the lowlands of the Netherlands, a 21st century perspective. *Quaternary International* 133–134: 159–178.
- Westerhoff WE (2009) Stratigraphy and sedimentary evolution. *The lower Rhine-Meuse System During the Late Pliocene and Early Pleistocene (southern North Sea Basin). Geology of the Netherlands 2, TNO-Utrecht*. PhD These, Vrije Universiteit, Amsterdam.
- Westerhoff WE and Weerts HJT (2003) Formatie van Beegden. Beschrijving lithostratigrafische eenheid. Description of the unit, to be found at www.dinoloket.nl.
- Westerhoff WE, Wong TE, and Geluk MC (2003) De opbouw van de ondergrond. In: De Mulder EFJ, Geluk MC, Ritsema I, Westerhoff WE, and Wong TE (eds.) *De ondergrond van Nederland. Geologie van Nederland* 7: 247–352. Utrecht: Nederlands Instituut voor Toegepaste Geowetenschappen TNO.
- Wijmstra TA and Schalk HWJG (1971) Other sections in the Dinkel valley, VIII. In: Van der Hammen T and Wijmstra TA (eds.) *Upper Quaternary of the Dinkel Valley. Mededelingen Rijks Geologische Dienst, Nieuwe Serie* 22: 131–140.
- Wijmstra TA and Schreve-Brinkman EJ (1971) The Lutterzand section, VI. In: Van der Hammen T and Wijmstra TA (eds.) *Upper Quaternary of the Dinkel Valley. Mededelingen Rijks Geologische Dienst, Nieuwe Serie* 22: 87–99.
- Wood HE II, Chaney RW, Clark J, et al. (1941) Nomenclature and correlation of the North American continental Tertiary. *Geological Society of America Bulletin* 52: 1–48.
- Zagwijn WH (1961) Vegetation, climate and radiocarbon datings in the Late Pleistocene of the Netherlands. Part I. Eemian and Early Weichselian. *Mededelingen Geologische Stichting NS* 14: 15–45.
- Zagwijn WH (1983) Sea-level changes in the Netherlands during the Eemian. *Geologie en Mijnbouw* 62: 437–450.

- Zagwijn WH (1985) An outline of the Quaternary stratigraphy of the Netherlands. *Geologie en Mijnbouw* 64: 17–24.
- Zagwijn WH (1998) Borders and boundaries: A century of stratigraphical research in the Tegelen-Reuver area of Limburg (the Netherlands). *Mededelingen Nederlands Instituut Toegepaste Geowetenschappen TNO* 60: 19–34.
- Zagwijn WH and Paepe R (1968) Die Stratigraphie der Weichselzeitlichen Ablagerungen der Niederlande und Belgiens. *Eiszeitalter und Gegenwart* 19: 129–146.
- Zonneveld JIS (1947a) De grens Plio-Pleistoceen in Z.O. Nederland. *Geologie en Mijnbouw N.S.* 9: 180–190.
- Zonneveld JIS (1947b) Het Kwartair van het peelgebied en de naaste omgeving. Een sedimentpetrologische studie. *Mededelingen van de Geologische Stichting, Serie C* VI(3): 1–233.
- Zonneveld JIS (1959) Litho-stratigrafische eenheden in het Nederlandse Pleistoceen. *Mededelingen van de Geologische Stichting, Nieuwe Serie* 12: 31–64.

Morphostratigraphy/Allostratigraphy

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Morphostratigraphy

Morphostratigraphy can be defined as the subdivision of sedimentary units based on surface form. [Frye and Willman \(1962\)](#) defined a morphostratigraphical unit as 'a body of rock that is identified primarily from the surface form it displays.' However, morphostratigraphical units often incorporate some element of lithological criteria ([Bowen, 1978](#), p. 94). As a result, the spatial distribution of morphostratigraphical units is frequently presented using geomorphological maps, incorporating both morphological and lithological criteria. Morphostratigraphy is widely employed in Quaternary Science to subdivide a range of sedimentary successions and is a key component of geomorphology – the study of Earth surface processes and landforms ([Hughes, 2010](#)). Several examples of its application are outlined in the following sections.

Glacial Sediments

Interpretation of glacial landforms and associated sediments, and the mapping of their relative position in the landscape often provides the basis for stratigraphical subdivision – especially, though not exclusively, in upland settings ([Hughes, 2010](#); [Figures 1 and 2](#)). However, Quaternary deposits are often not morphologically expressed at the surface because of erosion or burial by more recent sediments. In these circumstances, morphostratigraphy alone is limited in its power to subdivide the glacial record, and support is required from lithostratigraphical criteria ([Pitkäranta, 2009](#)). This is highlighted in papers by [Owen et al. \(1998\)](#) and [Hughes et al. \(2005, 2006\)](#) who recognized the importance of a combined morphological and sedimentological approach to subdivide glacial deposits in mountain regions. The linkage of landforms and subsurface sediments and relating them genetically to process landform studies underpins the 'landsystems' approach in glacial geomorphology ([Evans, 2003](#)).

Fluvial Sediments

Terraced fluvial successions represent a classic example where morphostratigraphy has been widely applied (e.g., [Antoine et al., 2000](#); [Sancho et al., 2008](#)). Interpretation of terrace successions on the basis of altitude is commonly the reverse of that in glaciated areas, as the terrace age usually decreases with lower altitude. This is because phases of fluvial aggradation interrupted by phases of incision often lead to a step-like profile in valley-side fluvial sediments and a terraced morphology ([Figure 3](#)). While the relative height of terrace surfaces reflects the relative ages of these surfaces, it conceals the complexities of the internal structure of terrace units. Thus, based on surface form alone, only a simplified history of fluvial activity can be inferred. A more comprehensive understanding of fluvial history can be achieved using a three-dimensional

alluvial stratigraphical approach utilizing spatial variability in surface form and lithological characteristics ([Mackey and Bridge, 1995](#)). Thus, most modern studies of alluvial stratigraphy combine morphological mapping with lithological evidence from sediment cores, test pits, or section exposures (e.g., [Howard et al., 2004](#); [Kasse et al., 2005](#); [Woodward et al., 2008](#); [Figure 3](#)).

Lake-Level Fluctuations

Changes in lake levels through time are often recorded by paleoshorelines, sometimes termed paleostrandlines. Former shorelines record lake-level stability during high stands and provide important information regarding changing moisture conditions during the Quaternary. For example, [Schuster et al. \(2005\)](#) identified paleoshorelines of Holocene Lake Mega-Chad, in Africa, using satellite imagery to identify morphodynamic features of sedimentary systems. Most studies, however, utilize both morphological and lithological criteria identified in the field. [Briggs et al. \(2005\)](#) reconstructed late Pleistocene and late Holocene lake highstands in the Pyramid Lake subbasin of Lake Lahontan, Nevada, USA, on the basis of morphological evidence of paleoshorelines and trenches cut through beach deposits. Similarly, in Mongolia, [Komatsu et al. \(2001\)](#) identified paleoshorelines using both satellite/field-based morphological mapping and near-surface lithology.

However, because shorelines tend to record lake-level stability, rapid fluctuations in lake levels may not be clearly recorded by former lake shorelines. Given the rapid millennial to centennial global climate fluctuations that occurred during the last cold stage, the degree to which these changes are recorded in lake shorelines is questionable. Lake-level histories are therefore better recorded in lithostratigraphical and biostratigraphical evidence from subsurface sediments across a lake basin (cf. [Digerfeldt et al., 2000](#); [Frogley et al., 2001](#)). As with other evidence partly recorded by surface form, while morphostratigraphy is an important component in understanding environmental change in lacustrine settings, surface forms should be integrated with other records.

Sea-Level Change

Changes in sea level are often recorded by landforms such as raised beaches. For example, morphological mapping has been a prerequisite of studying the altitudinal variation of raised shoreline deposits in northwest Europe ([Bates et al., 2000](#); [Cullingford, 1977](#)). However, these and other studies utilized natural field sections and boreholes to understand the morpho- and lithostratigraphy of former shoreline areas. Furthermore, reconstruction of sea-level history on the basis of morphostratigraphy alone fails to fully take into account sea-level transgressions and regressions through time. For example, in the Firth of Forth area of eastern Scotland, only seven

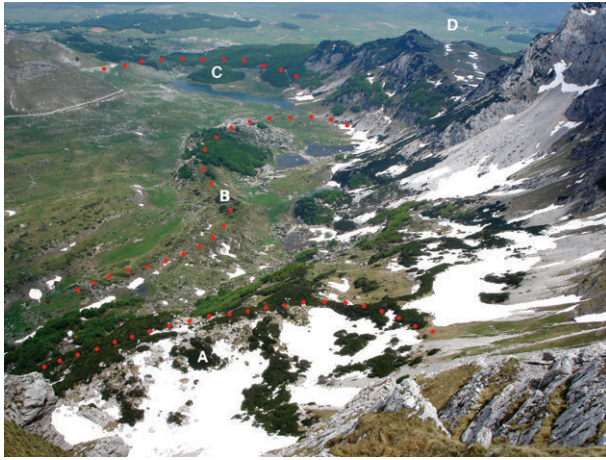


Figure 1 Surface landforms are especially important in subdividing glacial deposits in mountainous regions. In this photograph, four sets of stratigraphically separate suites of moraines (labeled A–D) can be identified in a cirque → valley → piedmont system in the Durmitor massif, Montenegro. Photograph by Philip Hughes, May 2005.

shorelines, including the present, are recorded by surface morphology, while 11 shorelines are recorded by both surface and buried shorelines (Sissons, 1983; Figure 4). Thus, when used alone, morphostratigraphical approaches are limited in their capacity to understand the order of sedimentary changes through time in former coastal settings and should ideally be combined with subsurface lithostratigraphical investigation.

The examples discussed above represent only some of the types of sedimentary successions to which morphostratigraphy can be applied. Morphostratigraphy can be used to subdivide any sedimentary succession where surface landform is an important defining criterion. Some studies also utilize landform morphology to differentiate between erosional surfaces of different ages, such as shore platforms or former cliff lines (e.g., Bates et al., 2000; Dawson et al., 1999), although when strictly defined, morphostratigraphy concerns only the ordering and subdivision of sediments and associated landforms. Thus, while erosional forms may be important toward understanding geomorphological relationships, they cannot be included within morphostratigraphical units because they are not sediment units. However, in the example of shore platforms, associated sediments such as beach deposits can be defined as morphostratigraphical units. This rule should apply to all Quaternary environmental settings and, for example, in glacial environments, neither cirques nor troughs can represent morphostratigraphical units – these can only be defined stratigraphically by the associated sediments that these landforms contain.

Some authors have argued that morphostratigraphical units should have no formal stratigraphical status and that they should not be applied with any formal stratigraphical implication (e.g., Bowen, 1978; Rawson et al., 2002, pp. 11–12; Richmond, 1959). However, by integrating morpho- and lithostratigraphy, there is a strong case for accepting morphostratigraphical units as formal stratigraphical units. In many cases, surface form defines the boundaries between sedimentary units, yet in practice, it is rare that morphostratigraphical units themselves are not defined using at least some

lithological criteria (Bowen, 1978, p. 94). For example, glacial units may be recognized by lithological criteria incorporating a range of sedimentological characteristics such as striated subrounded boulders and diamicton exposures, but separated by morphological criteria, such as moraine position. The fact that morphological and lithological criteria are often used to describe and subdivide glacial sediments led Hughes et al. (2005) to argue for the integration of morphostratigraphy within lithostratigraphy and the development of a formalized morpho-lithostratigraphical approach. In this scheme, morphostratigraphical units can be subdivided using formal lithostratigraphical terms, such as bed, member, formation, and group. Soil development may be an important characteristic of certain deposits and landforms. Pedostratigraphical units are separate and of equal rank to lithostratigraphical units, but both share overlapping textural and compositional attributes, and may be useful as a marker horizon when applied in conjunction with the morphostratigraphy (Figure 5). Hughes et al. (2005) argued that the formalization of morpho-lithostratigraphical approaches was useful for the development of robust terrestrial stratigraphical frameworks and for the systematic comparison of sediment units across wide areas in not just glacial, but in a range of Quaternary sediments. Unfortunately, however, the past status of morphostratigraphy has inhibited the integration of geomorphological data within the formal stratigraphical method.

Allostratigraphy

An allostratigraphical unit is a body of sedimentary rock that is defined and identified by its bounding discontinuities (North American Commission on Stratigraphic Nomenclature, 1983, pp. 865–867). The fundamental units are the same as for lithostratigraphy, but with the prefix allo-, that is, allomember/alloformation. Allostratigraphical units may be defined to distinguish between (a) successions of lithologically similar deposits bounded by discontinuities, (b) contiguous, genetically related heterogeneous deposits bounded by discontinuities, and (c) geographically separated units separated by discontinuities. In the *North American Stratigraphic Code* (NACSN, 1983), the bounding discontinuities are described as unconformities, disconformities, or the present-day land surface, although fault boundaries are excluded (Rawson et al., 2002, pp. 10–11). In some instances, prolonged surface exposure may sometimes be recorded by soil development resulting in allostratigraphical boundaries being marked by soil horizons or paleosols (Autin, 1992).

Allostratigraphical units are essentially the same as unconformity-bounded units and are analogous to syntems as defined in the *International Stratigraphic Guide* (Salvador, 1994, p. 50). Allostratigraphy also represents a form of sequence stratigraphy. However, sequence stratigraphy tends to be applied at a larger scale and, unlike allostratigraphy, does not cover the special case of sediment or rock bodies of similar lithologies separated by discontinuities (Rawson et al., 2002, pp. 10–11). The latter is an important advantage of allostratigraphy because using a lithostratigraphical approach, lithologically similar units separated by discontinuities units would be classified as a single unit. Conversely, allostratigraphy also

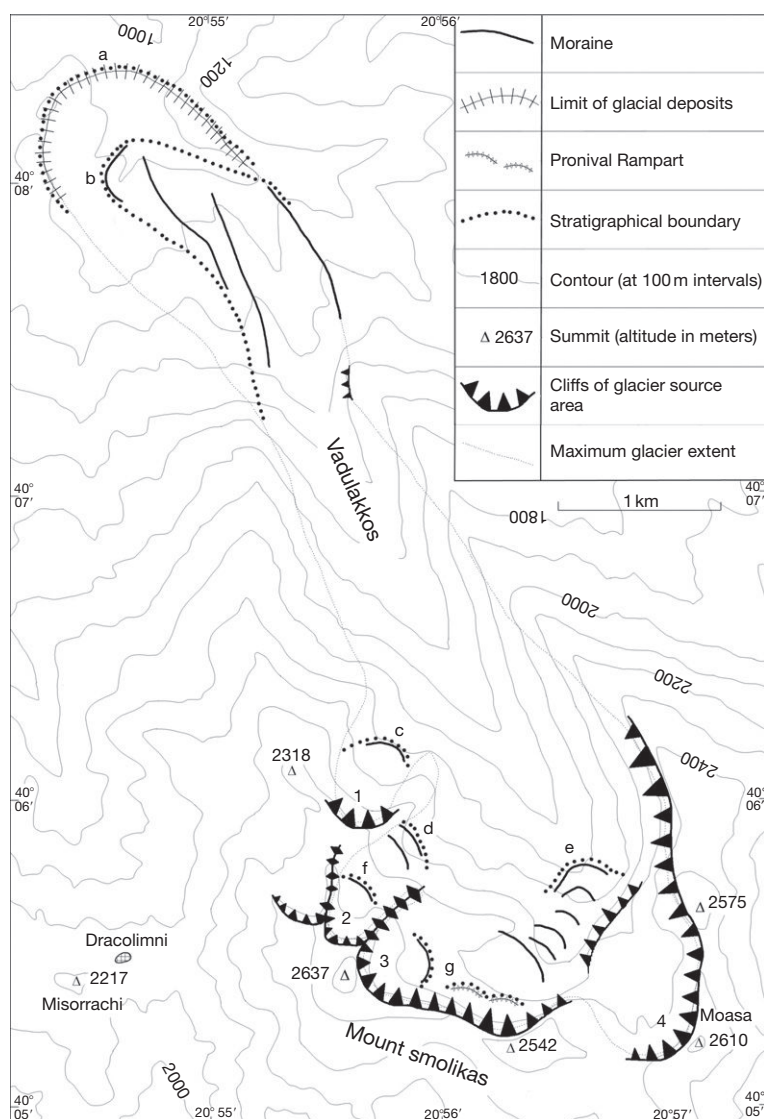


Figure 2 A glacial geomorphological map from Mount Smolikas, Greece. Here, the glacial sequence can be subdivided on the basis of moraine position. Glacial units are identified by the letters a–g and correspond to the following subdivision, in order of age (oldest first), Unit 1: a, Unit 2: b, Unit 3: c/d/e, and Unit 4: f/g. Reproduced from Hughes PD, Gibbard PL, and Woodward JC (2005) Quaternary glacial records in mountain regions: A formal stratigraphical approach. *Episodes* 28: 85–92.

provides an objective method for the definition of sediment units that contain genetically related but heterogeneous sedimentary deposits (Figure 6). For example, Miall (1997, p. 12) argued that “allostratigraphic methods enable the erection of a sequence framework that avoids the cumbersome nature of lithostratigraphy, whereby lateral changes in facies within a unit of comparable age require a change in name.” A common goal of allostratigraphy is that, for both homogeneous and heterogeneous sediment units, the defining discontinuities have time-stratigraphical significance and it is in this respect that the approach is similar to morphostratigraphy, which assumes time-stratigraphical significance for surface forms. Consequently, most Quaternary sedimentary successions that can be subdivided primarily on the basis of surface form, that is, using morphostratigraphy, can be subdivided using an allostratigraphical approach.

The application of allostratigraphy is discussed in the following paragraphs with reference to alluvial, lacustrine, glacial, and fluvial deposits – examples used in the original definition in the *North American Stratigraphic Code* (NACSN, 1983). However, the approach is not limited to these settings and can be applied to a range of Quaternary sedimentary successions where discontinuities are important criteria in differentiating between sediment units, including marine shoreline successions (Figure 4).

Lake-Level Fluctuations

Sedimentation in lakes is largely continuous although variable in terms of rate and composition depending on the depth of the water column and environmental conditions within the lake and also at the surface. When erosion occurs, it is often

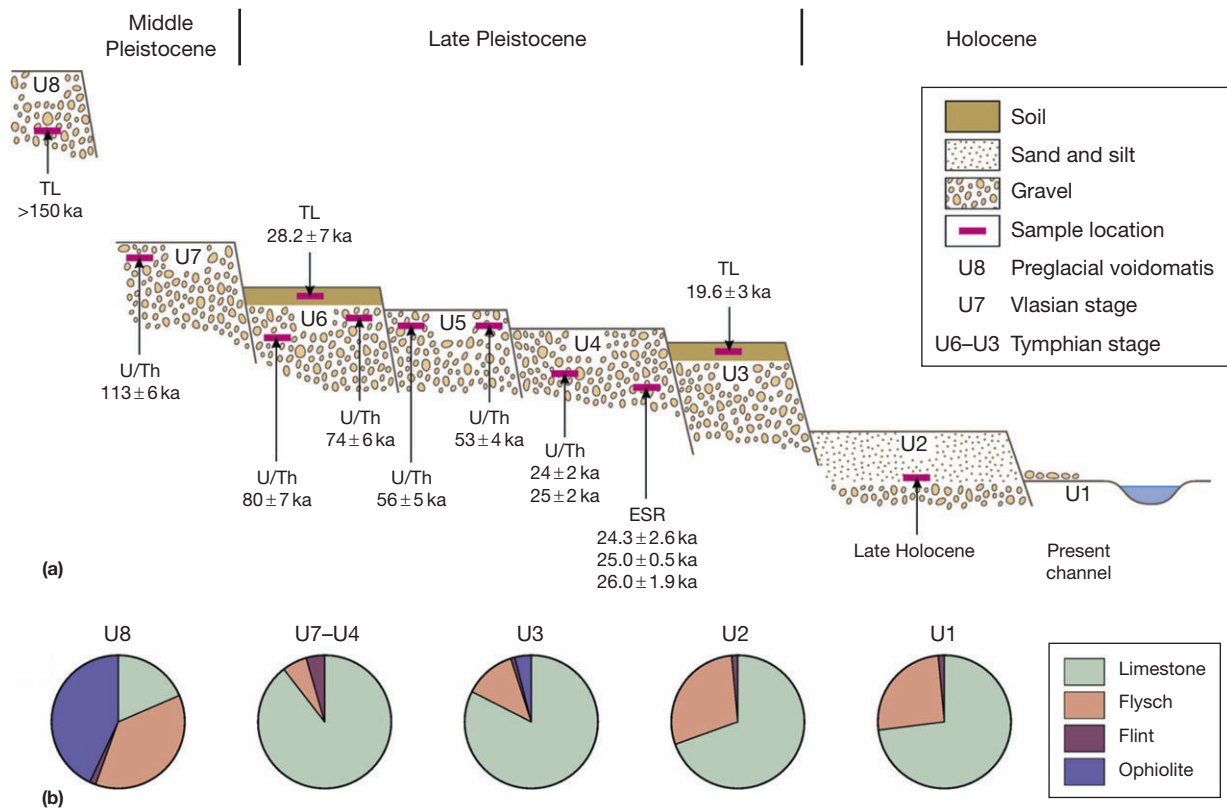


Figure 3 (a) Schematic diagram showing the Pleistocene and Holocene fluvial record in the Voidomatis River basin in northwest Greece (based on Hamlin, 2000; Lewin et al., 1991). (b) The lithological composition of the gravels in the coarse-grained alluvial units (after Lewin et al., 1991). Reproduced from Woodward JC, Hamlin RHB, Macklin MG, Hughes PD, and Lewin J (2008) Glacial activity and catchment dynamics in northwest Greece: Long-term river behaviour and the slackwater sediment record for the last glacial to interglacial transition. *Geomorphology* 101: 44–67.

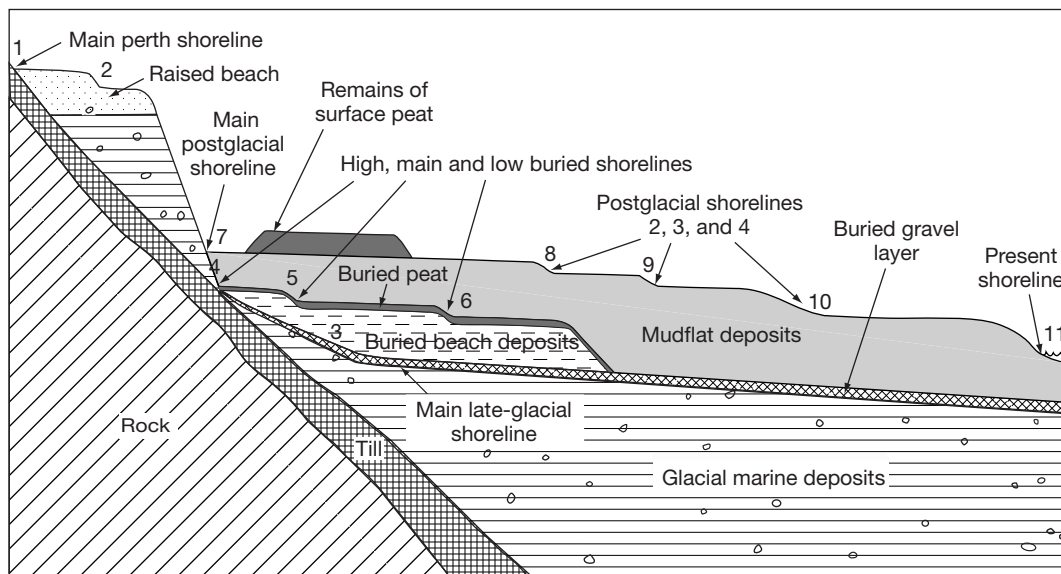


Figure 4 Paleoshorelines in the Firth of Forth, Scotland, based on a combination of morpho- and lithostratigraphical evidence. Shorelines are numbered in order of age from 1, the oldest shoreline, to 11, the present shoreline. Reproduced from Sissons JB (1983) Shorelines and isostasy in Scotland. In: Smith DE and Dawson AG (eds.) *Shorelines and Isostasy*, pp. 210–225. London: Academic Press.

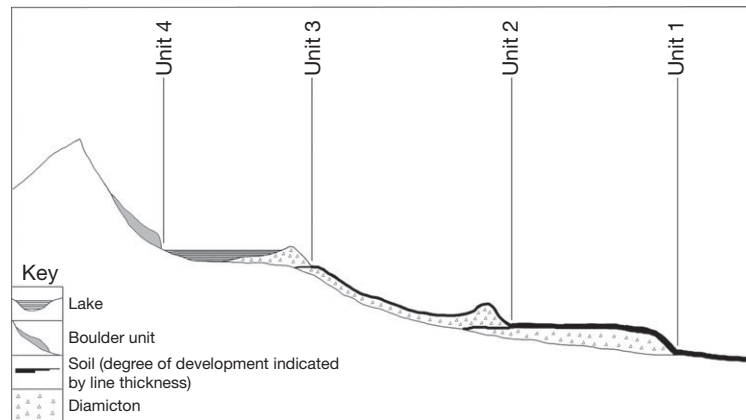


Figure 5 Down-valley cross-section of a hypothetical glaciated cirque-valley system. The diamicton units are glacial in origin and the boulder unit represents a periglacial rock glacier. Sediment landforms are subdivided on the basis of morphostratigraphical position in conjunction with lithological and pedological attributes. Reproduced from Hughes PD, Gibbard PL, and Woodward JC (2005) Quaternary glacial records in mountain regions: A formal stratigraphical approach. *Episodes* 28: 85–92.

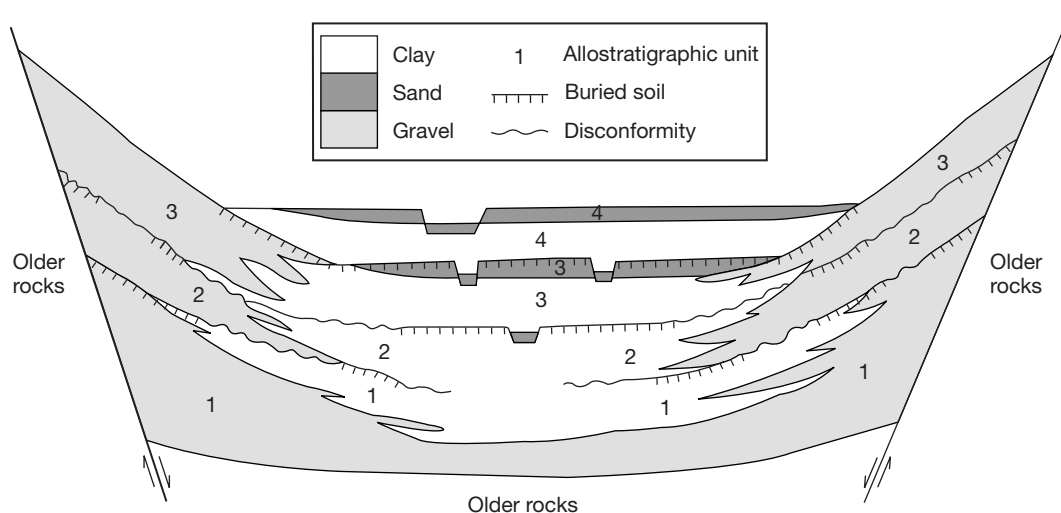


Figure 6 An example of allostratigraphical classification in alluvial and lacustrine deposits in tectonic sedimentary basin (graben). Allostratigraphy provides a useful method for the subdivision of sediment units that contain genetically related but heterogeneous sedimentary deposits. Rather than dividing units on the basis of lithology, they are separated on the basis of discontinuities (or disconformities). Reproduced from [North American Commission on Stratigraphic Nomenclature \(1983\)](#) North American stratigraphic code. *American Association of Petroleum Geologists Bulletin* 67: 841–875. AAPG ©1983. Redrawn with permission of the American Association of Petroleum Geologists.

indicative of major changes in the geomorphological state of the lake basin, such as major lake-level regressions, or subsequent debris flows, or landslides. Lateral variation in lithology between gravel and clay deposits will reflect position in the basin and vertical variation in lithology may reflect changes in lake level. In a lithostratigraphical approach, the lacustrine clays and the alluvial gravels would be divided into separate stratigraphical units and the presence of multiple units separated by discontinuities would not be recognized. Furthermore, an allostratigraphical approach enables genetically related heterogeneous deposits to be grouped within a single stratigraphical unit, as the subdivision of sediment units relies on the presence of a discontinuity (Figure 6).

Allostratigraphy has been employed by numerous researchers to understand the stratigraphy of lacustrine deposits, including at Lake Bonneville ([Oviatt et al., 1994](#)) and Lake Canyon ([Pederson, 2000](#)), both in the United States, and also at Lake Rotorua in New Zealand. Here, [Marx et al. \(2009\)](#) used allostratigraphy to subdivide and order lacustrine terraces (paleoshorelines). They argued that “flights of terraces do not yield the stacking relationships useful in erecting a standard lithostratigraphy, but allostratigraphy provides an objective means of grouping heterogeneous lacustrine deposits that are nevertheless genetically related” (p. 352). At Lake Rotorua, bounding discontinuities are represented by pyroclastic deposits such as tephra marker horizons.

Glacial Sediments

Glacial units are usually separated by erosional boundaries, reflecting each subsequent glacier advance. Moreover, glacial units may appear indistinguishable on the basis of lithological criteria alone. They can often be separated on the basis of their lateral extent and on the basis of moraine position. Consequently, glacial deposits appear to have all the characteristics suited to the allostratigraphical approach in that they represent successions of lithologically similar, contiguous, and geographically separated units – all separated by discontinuities. Thus, glacial units have been recognized as allostratigraphical units by workers such as [Borgert et al. \(1999\)](#), [Dahms \(2002\)](#), and [James et al. \(2002\)](#).

Allostratigraphy is sometimes thought especially suitable for glacial successions because sediment units may appear indistinguishable on the basis of lithological criteria alone “due to the complexity and small-scale variation of the lithologic units in these deposits” ([Räsänen et al., 2009](#), p. 4). [Räsänen et al. \(2009\)](#) argued for the combined use of allo- and lithostratigraphy in a similar manner to [Hughes et al. \(2005\)](#) who argued for the integrated use of morpho- and lithostratigraphy. However, the two approaches vary in an important respect: in [Räsänen et al. \(2009, p. 9\)](#) the ‘allounits are given preference, and the lithostratigraphic units are subordinate to the allostratigraphy’; while in [Hughes et al. \(2005\)](#), morphological and lithological units are considered of equal importance. A detailed discussion is provided in [Hughes \(2010\)](#).

Fluvial Sediments

Fluvial successions are another classic case where allostratigraphy can be applied. In many river basins, alternating phases of fluvial deposition and incision produce terraced sedimentary units bounded by discontinuities. As noted earlier, most workers apply both morpho- and lithostratigraphical approaches (e.g., [Kasse et al., 2005](#); [Lewin et al., 1991](#)) and, outside of North America, there are few examples where allostratigraphy has been applied. However, because fluvial successions often represent both contiguous discontinuity-bounded and geographically separated discontinuity-bounded deposits of similar lithology, an allostratigraphical approach can be applied.

In a study of the Amite River (Louisiana, United States), [Autin \(1992\)](#) argued that allostratigraphy provides several benefits for studying floodplain stratigraphy: (i) the criteria for the definition of alloformations conform with the formally defined stratigraphical procedures ([NACSN, 1983](#)); (ii) alloformations are three-dimensional bodies of sediment whose deposits are genetically associated with a given meander belt created during a discrete time interval; (iii) each alloformation contains a set of heterogeneous sedimentary deposits and their evaluation can be related to a set of fluvial processes; and (iv) this approach to geologic mapping establishes a relative chronology for the units identified.

[Autin \(1992\)](#) concluded that if alternatives to the allostratigraphical approach are adopted, then interpretations of some floodplain fluvial sequences are limited. For example, [Autin \(1992\)](#) argued that while a morphostratigraphical approach may enable the association between constructional landscapes

and sedimentary deposits to be inferred, a three-dimensional sediment body is not delineated. In addition, if sediment units are treated as lithostratigraphical units, the formations would reflect depositional processes and environments but would provide little or no information regarding valley chronology and evolution. Subsequent work in Mississippi Valley of Louisiana has utilized allostratigraphy, lithostratigraphy, and pedostratigraphy depending on the specific problem being addressed (e.g., [Autin, 1996](#); [Heinrich, 2006](#); [McCulloh et al., 2003](#)).

Morphostratigraphy and Allostratigraphy in Practice

Morphostratigraphy and allostratigraphy play a fundamental role in geomorphology and Quaternary stratigraphy ([Hughes, 2010](#)). Several geological surveys in the United States, such as Louisiana ([Heinrich, 2006](#); [McCulloh et al., 2003](#)), Idaho ([Breckenridge et al., 2005](#)), and other state Geological Surveys utilize allostratigraphy to subdivide Quaternary deposits. However, elsewhere, morpho- and, especially lithostratigraphy, has dominated. This is probably because allostratigraphy is a relatively newer concept, first appearing in 1983 in the *North American Stratigraphic Code* ([NACSN, 1983](#)). Allostratigraphy is rarely applied in some countries, such as the United Kingdom, for example, where the British Geological Survey employs lithostratigraphy to subdivide the Quaternary stratigraphical record (e.g., [McMillan et al., 2005](#)).

Some have argued that morphostratigraphical units should have no formal stratigraphical status and that they should not be applied with any formal stratigraphical implication (e.g., [Bowen, 1978](#); [Rawson et al., 2002](#), pp. 11–12; [Richmond, 1959](#)). However, in practice, morphostratigraphical units are rarely defined without using at least some lithological criteria ([Bowen, 1978](#), p. 94). The fact that morphological and lithological criteria are often used to describe and subdivide glacial sediments led [Hughes et al. \(2005\)](#) to argue for the integration of morphostratigraphy within lithostratigraphy. [Räsänen et al. \(2009\)](#) made a similar argument for the integration of allostratigraphy and lithostratigraphy when dealing with glacial deposits. In fact, there is a clear consensus that whether morphostratigraphy or allostratigraphy is adopted to subdivide and order any type of Quaternary deposits, each is always best applied in conjunction with lithostratigraphy ([Hughes, 2010](#); [Räsänen et al., 2009](#)).

See also: **Fluvial Environments:** Terrace Sequences. **Glacial Landforms:** Glacial Landsystems. **Lake Level Studies:** Overview. **Quaternary Stratigraphy:** Lithostratigraphy; Pedostratigraphy. **Sea Level Studies:** Geomorphological Indicators.

References

- Antoine P, Lautridou JP, and Laurent M (2000) Long-term fluvial archives in NW France: Response of the Seine and Somme Rivers to tectonic movements, climatic variations and sea-level changes. *Geomorphology* 33: 183–207.
- Autin WJ (1992) Use of alloformations for definition of Holocene meander belts in the middle Amite River, southeastern Louisiana. *Geological Society of America Bulletin* 104: 233–241.

- Autin WJ (1996) Pleistocene stratigraphy in the southern Lower Mississippi Valley. *Engineering Geology* 45: 87–112.
- Bates MR, Bates CR, Gibbard PL, et al. (2000) Late Middle Pleistocene deposits at Norton Farm on the West Sussex coastal plain, southern England. *Journal of Quaternary Science* 15: 61–89.
- Borgert JA, Lundeen KA, and Thackray GD (1999) Glacial geology of the Southeastern Sawtooth Mountains. In: Hughes SS and Thackray GD (eds.) *Guidebook to the Geology of Eastern Idaho*, pp. 205–217. Pocatello: Idaho Museum of Natural History.
- Bowen DQ (1978) *Quaternary Geology*. Oxford: Pergamon Press.
- Breckenridge RM, Othberg KL, and Kauffman JD (2005) *Surficial Geologic Map of the Mica Bay Quadrangle, Kootenai County, Idaho*. Moscow, ID: Idaho Geological Survey.
- Briggs RW, Wesnousky SG, and Adams KD (2005) Late Pleistocene and late Holocene lake highstands in the Pyramid Lake subbasin of Lake Lahontan, Nevada, USA. *Quaternary Research* 64: 257–263.
- Cullingford RA (1977) Lateglacial raised shorelines and deglaciation in the Earn–Tay area. In: Gray JM and Lowe JJ (eds.) *Studies in the Scottish Lateglacial Environment*, pp. 15–32. Oxford: Pergamon.
- Dahms DE (2002) Glacial stratigraphy of Stough Creek Basin, Wind River Range, Wyoming. *Geomorphology* 42: 59–83.
- Dawson AG, Smith DE, Dawson S, Brooks CL, Foster IDL, and Tooley MJ (1999) Lateglacial climate change and coastal evolution in western Jura, Scottish Inner Hebrides. *Geologie en Mijnbouw* 77: 225–232.
- Digerfeldt G, Olsson S, and Sandgren P (2000) Reconstruction of lake-level changes in lake Xínias, central Greece, during the last 40 000 years. *Palaeogeography, Palaeoclimatology, Palaeoecology* 158: 65–82.
- Evans DJA (ed.) (2003) *Glacial Landscapes*, p. 532. London: Arnold.
- Frogley MR, Griffiths HI, and Heaton THE (2001) Historical biogeography and late Quaternary environmental change of Lake Pamvotis, Ioannina (north-western Greece): Evidence from ostracods. *Journal of Biogeography* 28: 745–756.
- Frye JC and Willman HB (1962) Morphostratigraphic units in Pleistocene stratigraphy. *American Association of Petroleum Geologists Bulletin* 46: 112–113.
- Hamlin RHB, Woodward JC, Black S, and Macklin MG (2000) Sediment fingerprinting as a tool for interpreting long-term river activity: the Voidomatis basin, NW Greece. In: Foster IDL (ed.) *Tracers in Geomorphology*, pp. 473–501. Chichester: John Wiley and Sons.
- Heinrich PV (2006) Pleistocene and holocene fluvial systems of the lower pearl river, Mississippi and Louisiana, USA. *Gulf Coast Association of Geological Societies Transactions* 56: 267–278.
- Howard AJ, Macklin MG, Bailey DW, Mills S, and Andreescu R (2004) Late-glacial and Holocene river development in the Teleorman Valley on the southern Romanian Plain. *Journal of Quaternary Science* 19: 271–280.
- Hughes PD (2010) The role of geomorphology in Quaternary stratigraphy: Morphostratigraphy, lithostratigraphy and allostratigraphy. *Geomorphology* 123: 189–199.
- Hughes PD, Gibbard PL, and Woodward JC (2005) Quaternary glacial records in mountain regions: A formal stratigraphical approach. *Episodes* 28: 85–92.
- Hughes PD, Gibbard PL, and Woodward JC (2006) Middle Pleistocene glacier behaviour in the Mediterranean: Sedimentological evidence from the Pindus Mountains, Greece. *Journal of the Geological Society of London* 163: 857–867.
- James LA, Harbor J, Fabel D, Dahms D, and Elmore D (2002) Late Pleistocene glaciations in the northwestern Sierra Nevada, California. *Quaternary Research* 57: 409–419.
- Kasse C, Hoek WZ, Bohncke JP, et al. (2005) Late Glacial fluvial response of the Niers-Rhine (western Germany) to climate and vegetation change. *Journal of Quaternary Science* 20: 377–394.
- Komatsu G, Brantingham PJ, Olsen JW, and Baker VR (2001) Paleoshoreline geomorphology of Böön Tsagaan Nuur, Tsagaan Nuur and Örog Nuur: The valley of the Lakes, Mongolia. *Geomorphology* 39: 83–98.
- Lewin J, Macklin MG, and Woodward JC (1991) Late Quaternary fluvial sedimentation in the Voidomatis Basin, Epirus, northwest Greece. *Quaternary Research* 35: 103–115.
- Mackey SD and Bridge JS (1995) Three-dimensional model of alluvial stratigraphy: Theory and application. *Journal of Sedimentary Research* B65: 7–31.
- Marx R, White JDL, and Manville V (2009) Sedimentology and allostratigraphy of post-240 ka to pre-26.5 ka lacustrine terraces at intracaldera Lake Rotorua, Taupo Volcanic Zone, New Zealand. *Sedimentary Geology* 220: 349–362.
- McCulloh RP, Heinrich PV, and Snead JI (2003) Geology of the Ville Platte Quadrangle Louisiana. To Accompany the Ville Platte 30 × 60 Minute Geologic Quadrangle. In: *Louisiana Geological Survey, Geological Pamphlet No. 14*, p. 11. Baton Rouge: Louisiana State University.
- McMillan AA, Hamblin RJO, and Merritt JW (2005) An overview of the lithostratigraphical framework for the Quaternary and Neogene deposits of Great Britain (onshore). British Geological Survey Research Report, RR/04/04, 38 pp. <http://nora.nerc.ac.uk/3241/1/RR04004.pdf>.
- Miall AD (1997) *The Geology of Stratigraphic Sequences*. Berlin: Springer-Verlag.
- North American Commission on Stratigraphic Nomenclature (1983) North American stratigraphic code. *American Association of Petroleum Geologists Bulletin* 67: 841–875.
- Oviatt CG, McCoy WD, and Nash WP (1994) Sequence stratigraphy of lacustrine deposits: A Quaternary example from the Bonneville basin, Utah. *Geological Society of America Bulletin* 106: 133–144.
- Owen LA, Derbyshire E, and Fort M (1998) The Quaternary glacial history of the Himalaya. *Quaternary Proceedings* 6: 91–120.
- Pederson JL (2000) Holocene paleolakes of Lake Canyon, Colorado Plateau: Paleoclimate and landscape response from sedimentology and allostratigraphy. *Geological Society of America Bulletin* 112: 147–158.
- Pitkäranta R (2009) Lithostratigraphy and age estimations of the Pleistocene erosional remnants near the centre of the Scandinavian glaciations in western Finland. *Quaternary Science Reviews* 28: 166–180.
- Räsänen ME, Auir JM, Huittu JV, Klap AK, and Virtasalo JJ (2009) A shift from lithostratigraphic to allostratigraphic classification of Quaternary glacial deposits. *GSA Today* 19: 4–11.
- Rawson PF, Allen PM, Brencley PJ, et al. (2002) *Stratigraphical Procedure*. London: The Geological Society.
- Richmond GM (1959) Report of the Pleistocene committee, American Commission on stratigraphic nomenclature. *American Association of Petroleum Geologists Bulletin* 43: 633–675.
- Salvador A (1994) *International Stratigraphic Guide: A Guide to Stratigraphic Classification, Terminology and Procedure*. 2nd edn. Boulder, CO: International Union of Geological Sciences and The Geological Society of America.
- Sancho C, Peña JL, Muñoz A, et al. (2008) Holocene alluvial morphopedosedimentary record and environmental changes in the Bardenas Reales Natural Park (NE Spain). *Catena* 73: 225–238.
- Schuster M, Roquin C, Durringer P, et al. (2005) Holocene lake Mega-Chad palaeoshorelines from space. *Quaternary Science Reviews* 24: 1821–1827.
- Sissons JB (1983) Shorelines and isostasy in Scotland. In: Smith DE and Dawson AG (eds.) *Shorelines and Isostasy*, pp. 210–225. London: Academic Press.
- Woodward JC, Hamlin RHB, Macklin MG, Hughes PD, and Lewin J (2008) Glacial activity and catchment dynamics in northwest Greece: Long-term river behaviour and the slackwater sediment record for the last glacial to interglacial transition. *Geomorphology* 101: 44–67.

Pedostratigraphy

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Introduction

Landscapes comprise diverse landforms, parent materials of different composition, and geomorphic surfaces formed by erosional and depositional events, all of various ages. Landscapes can evolve rapidly at various times in response to both long-period climatic cycles and tectonics, and relatively instantaneous events such as storms and earthquakes. At other times and locations, landscape evolution is slow, and driven by more benign processes of background erosion rates, dust accretion, and weathering. Consequently, soils can form continuously through long periods of landscape evolution, and/or they can form rapidly following catastrophic events.

Pedostratigraphy can be defined as the study of the stratigraphical and spatial relationships and implications of surface and buried soils (adapted from Wysocki et al., 2005).

In the current landscape, soils form in a variety of environmental conditions, which can be analyzed by the classic equation formulated by Dokuchaev and Jenny: $S=f$ (climate, vegetation and organisms, relief, parent material, and time).

Past landscapes and their soils, whether buried or still at the surface, can be analyzed in exactly the same way, except that here the effects of burial or exhumation must also be considered. Therefore, landscapes contain soils of different morphology reflecting the environment in which they formed. Inevitably, these soils will continue to form, aggrade through slow sediment accretion, be truncated by erosion, or be buried. However, more complicated, and perhaps more sophisticated, models have also been put forward (Follmer, 1998), because in the Dokuchaev analysis it is difficult to integrate the effects of changing climates and vegetation, let alone the hydrological changes that take place as a soil is variously weathered, buried, or exhumed.

Soil scientists have begun to realize that soil genesis can involve many specific and complex processes (Follmer, 1998) under the imprint of environmental conditions that prevail at each moment in time. The induced weathering can then set up pedogenic pathways where the properties of adjacent soils may diverge, and then continue to diverge, run a parallel course, or converge once more. Once a soil is buried, these relationships become more complicated.

Definitions

A paleosol is a soil formed in a landscape of the past, and may be buried, nonburied or exhumed. Buried soils represent periods of slow or nil deposition and slow or nil erosion reacting to the environmental conditions of the time and through subsequent centuries. Because landscapes consist of different surfaces on many kinds of deposits, a soil may be independent of the deposits into which it forms (Catt, 1986). Thus, soils can form 'soil stratigraphic' (or 'pedostratigraphic') units in exactly the same way that 'lithostratigraphic,' 'chronostratigraphic,' and other stratigraphic units are recognized and defined. The

North American Stratigraphic Code (1983) defines a pedostratigraphic unit as 'a buried, traceable, three-dimensional (3D) body of rock that consists of one or more differentiated pedologic horizons.'

A pedostratigraphic unit is a body of sediment or rock that consists of one or more pedological horizons that developed in one or more lithostratigraphic, allostratigraphic, or lithodemic units. These units are distinguished by the presence of one or more pedologic horizons that display clear signs of soil formation (pedogenesis). All the morphological features that are used to recognize surface soils can be applied to buried soils as well (Table 1), but the apparent absence of a feature is not an indication that no buried soil is present. For example, soil organic matter can disappear in buried oxidized sediments, and soil structure is often destroyed postburial.

The upper boundary of a pedostratigraphic unit is defined in the code as being the top of the 'uppermost pedologic horizon formed by pedogenesis in a buried soil horizon,' and the lower boundary is the lower boundary of a horizon with definite pedogenic features. These requirements become problematical in entirely weathered sequences, and in upbuilding soil profiles, for example, soil formation continuing through slow loess accretion. Pedostratigraphic units are different from surface soils in two important ways. A soil forming at the surface is taken to include overlying plant litter layers or peat and carbonaceous deposits. These are specifically excluded from pedostratigraphic units, but they may be defined as lithostratigraphic or biostratigraphic units. Also, the parent material C horizon, where present, is usually included in the profile of a surface soil whereas the code, requiring the recognition of definite morphological features for inclusion in a pedostratigraphic unit, often excludes C horizons.

Although the definition of a pedostratigraphic unit in the code requires 'traceability,' few buried soils are continuously traceable for long distances. Because they represent a ground surface, they just as often undergo erosion, or changes in character, as the surface soil mantle does.

Pedostratigraphic units are unique in that they are a product of alteration at or near the surface of the earth. Their lithology and other physical, chemical, and mineralogical properties can, because of this alteration, differ markedly from those of the parent material. Like magneto- and chronostratigraphic units, a single pedostratigraphic unit may be defined in many different parent materials of diverse composition and age. However, unlike the former two stratigraphic units, pedostratigraphic units are independent of time concepts and may be diachronous, yet they may carry significant evidence of age.

Selected Milestones in Pedostratigraphy

Buried soils have been recognized in Quaternary deposits since 1873 when A. H. Worthen of the Illinois Geological Survey

Table 1 Selected morphological, physical and, chemical pedogenic features, their implications, and potential for preservation in paleosols

<i>Pedogenic feature</i>	<i>Implication</i>	<i>Weathering</i>	<i>Preservation potential</i>
<i>Morphological</i>			
Color	Liberation of iron oxides	Weak–strong	High
Organic matter	Prolonged plant growth	Weak–strong	Low
Soil structure	Prolonged plant growth	Weak–strong	Moderate
Clay formation	Moderate weathering; wetting and drying	Moderate–strong	High
Clay coatings	Clay movement	Moderate–strong	High
Carbonate concretions	Leaching	Weak–strong	Moderate
Fragipan	Wetting and drying	Moderate	Low
Root traces	Plant growth and root exploration	Weak–strong	High
<i>Micromorphological</i>			
Argillans	Clay movement	Moderate–strong	High
Fabric development	Pedological ‘organization’	Moderate–strong	Moderate
<i>Chemical</i>			
Lower pH and BS%	Leaching	Weak–strong	Low
Depletion in K ₂ O	Leaching, plant uptake, weathering	Moderate–strong	High
Forms and amounts of P	Plant uptake and return, clay movement, leaching	Moderate–strong	High
<i>Physical</i>			
Lower dry bulk density	Formation of soil structure, burrowing by organisms	Moderate–strong	High
Increased macroporosity	Formation of soil structure, plant root channels, burrowing by earthworms	Moderate–strong	Moderate
Increased plant available water	Formation of soil structure, plant root channels, weathering, organic matter	Moderate–strong	Moderate

showed that a humus-rich soil had developed on one till before being buried by another. In doing so, he established not only the stratigraphic value of buried soils, but also their paleoclimatological value. Thus, the recognition of buried soils was instrumental in developing the glacial stratigraphy of North America, and later, central Europe.

Modern concepts of pedostratigraphy were introduced in pioneering works by [Butler \(1959\)](#), [Ruhe \(1956\)](#), [Ruhe \(1969\)](#), [Morrison \(1964a\)](#), [Morrison \(1964b\)](#), and [Morrison \(1978\)](#).

[Ruhe \(1956\)](#) defined a geomorphic surface as ‘a portion of the land surface that is specifically defined in space and time. It may occupy many parts of the landscape and may include many landforms.’ A geomorphic surface is a spatial time plane linking erosional, soil forming, and depositional parts of the land surface. Soil-forming parts of the landscape need to be metastable long enough to form at least an A horizon, although the transition to erosional areas can be marked by exposed B (subsoil) horizons, while transition areas to deposition are marked by aggrading A (topsoil) horizons. Buried geomorphic surfaces are common in tectonically and volcanically active areas where landforms are episodically mantled by tephra, loess, or dune sands ([Tonkin, 1994](#)), and in areas of growing overlapping fans and valley fills.

Referring to geomorphic surfaces as ‘time planes’ can be misleading because they are rarely as instantaneous as a tephra ([Figure 1](#)), a lahar, or a flood deposit, for example. Usually, the metastable phase is exactly as the name suggests, with an array of soils of slightly different age on different elements of the surface, and different rates and spatial location of erosion and deposition elsewhere. Thus, geomorphic surfaces, both at the surface and buried, are often somewhat diachronous.

[Butler \(1959\)](#) recognized that erosion and deposition cycles can be episodic, for instance, driven by climate change. Periods of metastability, during which recognizable soils develop, are

followed by instability that creates erosional and depositional surfaces. Together, the metastable and unstable phases constitute a ‘K-cycle.’ K-cycles are best recognized in hill country, rather than continuously eroding steep land, where the soil-residence time is long enough for clearly defined soils to develop before being eroded or buried.

A seminal application and development of new concepts in pedostratigraphy was made by [Morrison \(1964a\)](#) while working on the Lahontan and Bonneville pluvial-lake sequences in the Great Basin of the US. He found that paleosols are the best records of metastability in landscape evolution and are, thus, important parts of the stratigraphic record.

Where geomorphic surfaces are buried by varying depths of sediment, a number of possible soil-stratigraphic scenarios arise ([Morrison, 1978](#)) and these have subsequently been recognized as soil profile forms by [Tonkin \(1994\)](#). Simple soil profile forms have no buried soil features to some arbitrarily defined depth, which Tonkin takes as 2 m but which others (e.g., [Hewitt 1998](#)) might take as 1 m. Truncated soil profile forms have lost upper horizons to erosion. An A or E horizon might be found developing in former subsoil horizons. Aggrading soil profile forms have thick A horizons and B horizons that preserve the morphological features of once having been formed at the surface ([Figure 2](#)).

Composite soil profiles develop in thin accessions of sediment where rapid repetitions of A, E, or B horizons are temporarily preserved to form bi- or polysequel soils. The ultimate fate of such soils, if they are preserved, is usually to become welded into a single soil whose complex history might be difficult to unravel. Good examples of such soils can be found in thin accessions of ash surrounding volcanoes ([Figure 3](#)).

Compound soil profile forms are also bi- or polygenetic, but an interval of less weathered parent material is preserved

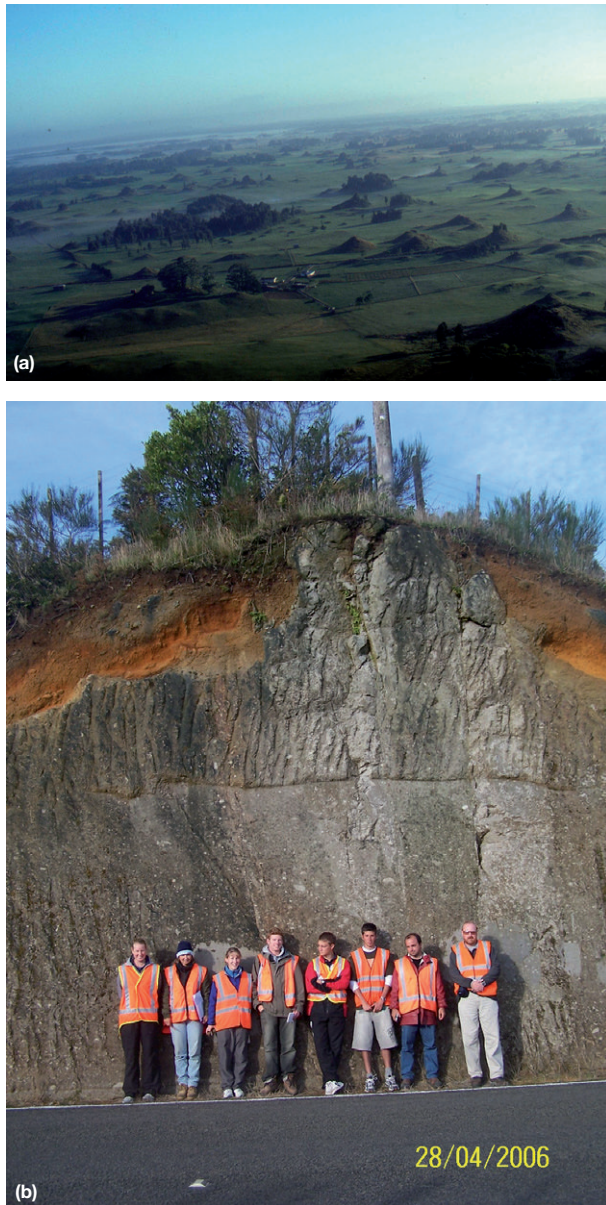


Figure 1 The Mamaku Plateau west of Rotorua in the North Island of New Zealand was formed by the eruption of the Mamaku Ignimbrite 230 ka. The plateau has been episodically partially mantled and partially stripped of loess and tephra since its formation, resulting in an array of buried and geomorphic surfaces, defined by erosion surfaces and/or soils. (a) An oblique aerial view. Note the prominent erosional remnants or tors up to 20 m high (photograph by author). (b) A section through a tor on the central and highest part of the plateau. The ignimbrite here is overlain by a tephra mantle less than 15 000-cal years old. Any older loess, tephra, and soils were stripped by wind, fluvial, and frost action during the last glaciation. Elsewhere on the plateau, fragments of the complete loess and tephra record are preserved between erosion surfaces. Complete sections representing 230 000 years of accumulation occur on the margins of the plateau (photograph by author).

between the surface and buried soils that is less likely to be obscured by weathering, at least in the short term. Episodic accumulation of lapilli-grade tephra from andesite and rhyolite tephra are good examples (Figure 4).



Figure 2 Loess and paleosols near Wanganui, southwestern North Island, New Zealand. The site is a 300 ka marine terrace mapped by Pillans (1990), who appears in the photograph. The prominent ginger paleosol is the Last Interglacial soil complex formed by three soil-forming phases representing marine oxygen isotope stages 5e, 5c and 5a. Loess and tephra continued to accumulate slowly during the formation of this soil, and it is a good example of an aggrading profile. Above the thick ginger soil complex are three more loess units separated by weaker soils (photograph by author).

Persistent profile forms (Tonkin, 1994) are soils that maintain a similar morphology over significant periods of time, and any two soils may be of vastly different age, yet have similar morphological properties. Spodosols and ultisols often persist without apparent change for long periods of time.

The pioneering works of Ruhe (1956) and Butler (1959), and Morrison (1964a) and Morrison (1964b) reawakened interest in paleopedology and pedostratigraphy in the soil science and Quaternary science communities. In 1965, paleopedology became a commission of the International Quaternary Association (INQUA) and a working group of the International Society of Soil Science. The two societies sponsored the 'First International Symposium on Paleopedology' in Amsterdam in 1970. The meeting brought together scientists from a wide range of disciplines from geology and geomorphology to agriculture. From this meeting, a selection of papers was published in a book edited by Yaalon (1971).



Figure 3 The Tirau soil, west of Rotorua, New Zealand, is developed in thin, distal accessions of rhyolite ash, and is an example of a composite soil. The scientist pictured is Alan Pullar, who, with Colin Vucetich (Figure 4) was largely responsible for detailing the tephrostratigraphy of the Central North Island of New Zealand (Vucetich and Pullar, 1969). At shin level is the 26.5 ka (cal) Kawakawa Tephra erupted from Lake Taupo, which rests on a weak paleosol formed 30–50 ka. Above the Kawakawa Tephra, is volcanic loess weakly weathered to form a B horizon, containing two more thin rhyolite tephras. At Pullar's hand level is the Rerewhakaaitu Tephra ca. 17 ka (cal). At this level and above, pedogenic features become more pronounced in response to late Pleistocene warming and warm, humid Holocene conditions. There is no loess above this level, and the soil is developed in a major tephra at ca. 14 ka, followed by several thinner ones on the surface. This is also a good example of an upbuilding profile. Each B horizon was once an A horizon at or near the soil surface (photograph by J. A. Pollok).

Problems with Nomenclature

It is not surprising that scientists trained in different aspects of geology, geomorphology, soil science, and agriculture, all working on aspects of pedostratigraphy and paleosols, should have different views, methods, and nomenclatures. Nonetheless, disagreements over nomenclature have been a persistent problem in pedostratigraphy and paleopedology. A Paleopedology Manual (Catt, 1990) was produced to describe methods used in the study and interpretation of paleosols,



Figure 4 The Taupo soil, north of Lake Taupo, New Zealand, is developed in thick deposits of Taupo Tephra (dates are in radiocarbon years). Beneath are a succession of rhyolite tephra erupted from Lake Taupo (Waimihia and Opepe) and the Rotorua region (Whakatane, Rotoma, Waiohau and Rotorua). Each is thick enough for a thin C horizon to be preserved, and this is an example of a compound profile. The person is Colin Vucetich, mentioned in Figure 3 (photograph by J. A. Pollok).

but it did not address the conflicts in nomenclature. The Second International Symposium on Paleopedology, which was held in 1993, sought to address conflicts regarding terminology, definitions, classification, and genetic interpretations (Follmer, 1998). The submissions were published via four Paleopedology Commission newsletters (Follmer, 1995).

A distinction is often made between buried paleosols and relict paleosols. The latter are paleosols formed on old geomorphic surfaces such as moraines, terraces, and fans, and other metastable landforms, which began forming when soil-forming conditions were different from those of the present day, but were still developing close to the ground surface. The debate (Bronger and Catt, 1998; Morrison, 1998) is over the extent to which all surface soils are in some way relict, especially those that began forming prior to the Holocene. Catt (1998) suggests that the term 'relict' has become obsolete and favors simply 'paleosol.' Nonburied paleosols should show distinct evidence that the environment of soil development was different from the present. Johnson (1998) argues that the term paleosol should be used for

buried soils, not surface soils on the ground, as all surface soils are always being subjected to modification in an ever-changing environment. Holliday (1998) questions the need for the term 'paleosol,' preferring to use simply 'soil' and 'buried soil.' He suggests reserving the term paleosol to soils that are lithified throughout their profile. Others have suggested putting minimum age limits on paleosols (Bronger and Catt, 1998), for example, older than the Holocene. Such proposals have met with protest because there are numerous examples of paleosols in Holocene-aged deposits that have formed on ancient landscapes, for example, the paleosols on alluvium on the Mississippi floodplain (Aslan and Autin, 1998).

The 'geosol' was introduced and defined by Morrison (1964b) and Morrison (1967) as the term for a pedostratigraphic unit, because he found that paleosols are significant elements of the stratigraphic record deserving recognition and classification.

Geosols

A geosol is defined (Catt, 1998) as a 'traceable, mappable 3D body of soil material comprising one or more differentiated pedologic horizons; it has a consistent stratigraphic position and is defined at a type locality where the horizons are buried by younger deposits, but may be traced to sites where it crops out on the present land surface. Geosols may or may not transgress lithostratigraphic, allostratigraphic, or lithodemic units. They change laterally in response to differences in parent material, the original buried topography, drainage, vegetation, and other factors, giving different pedofacies. Geosols may be monogenetic, polygenetic, or compound (multistory). Compound geosols (pedocomplexes) comprise two or more soils (pedomembers), which are separated over large areas by thin unmodified deposits or by unconformities; they often merge laterally into polygenetic geosols'.

Therefore, geosols may be regarded not just as soils or paleosols, but as soilscapes that can be recognized as laterally extensive stratigraphic horizons (Morrison, 1978). The term has been questioned by some (e.g., Holliday (1998)) for not clarifying soil stratigraphic issues, and lacking utility and application. Retallack (1998) also points to situations when the use of geosol is not appropriate. Many alluvial sedimentary sequences, particularly on fans, can have literally hundreds of buried soils, of different character and limited traceability. Each soil may be truncated by a paleochannel or merge with other buried soils. These cannot easily be named and characterized as geosols. In this situation, it is more useful to describe and categorize the different types of buried soil, and he suggests the name 'pedotype' for this eventuality.

Although accepted in the North American Stratigraphic Code (1983) as the fundamental and only pedostratigraphic unit, the International Stratigraphic Guide (1994) has yet to adopt the term, and in fact includes no soil stratigraphic units.

The Application of Pedostratigraphy to Paleoclimatology and Correlation

Eastern European Loess

In the 1960s and 1970s, the loess sequences of central Europe were studied intensively. In what is now the Czech Republic

and Slovakia, Kukla and Koci (1972) and Kukla (1975) described multiple buried soil-loess cycles in deposits on hillsides and river terraces. Their cycle begins with rapid warming from cold and arid climate conditions in which calcareous loess is generated, to temperate moister conditions generating colluvial sediments from hillsides and terrace risers, the boundary they called a 'markline.' During warm interglacial conditions, a decalcified brown forest soil develops through the colluvium into the loess below, although the markline remains decipherable. Deteriorating climate sees cold climate loess deposited once more. Morphological evidence of the multiple glacial cycles was supported by attendant variations in snail faunas (Lozek, 1969). These cycles became known through Europe as 'pedocomplexes.' At favorable sites, the cycles could be subdivided into stadial cycles of relatively warm and humid conditions versus cycles of cold and dry conditions.

Paleomagnetic studies by Koci (reported in Kukla, 1975) showed that sequences, in particular the Cervený Kopec section near Brno, extended well below the Brunhes-Matuyama paleomagnetic boundary (~0.78 Ma).

It became clear that the alpine and continental glacial stratigraphy are only a partial record of climate change during the Quaternary. The oxygen-isotope stratigraphy of deep-ocean cores preserves a much better record of fluctuating Quaternary climate. Recognizing that the marklines representing rapid postglacial warming in the loess sequences were similar to 'terminations' representing similar rapid warming in the oceanic record, Kukla (1975, 1987) was able to provide a convincing continent to deep ocean climostratigraphic correlation for the last million years.

The Loess Plateau of China

The loess-paleosol sequences of the Loess Plateau of China rank as the most complete record of terrestrial climate change during the extended Quaternary period (the past 2.5 Ma). Loess-paleosol sequences over 150-m thick are common in the area enclosed by the big bend of the Huang He River north and west of Xian. The loess is coarse and sandy close to its source in the deserts of the northwestern part of the region, and becomes finer textured and more weathered in the moister southeastern part of the region. It is here, in sections such as Baoji (Ding et al., 1991), that the intervening paleosols become more prominent (Figure 5).

Magnetostratigraphy, and in particular magnetic susceptibility, has proven to be the tool that enabled the deep loess-paleosol sequences, first in Europe and then in China (Kukla, 1987; Liu et al., 1987), to be correlated to the marine oxygen isotope record. Initially only the major glacial-interglacial cycles were correlated, but now more detailed correlations of intervening, minor paleosols are made (Ding et al., 1991). There is still considerable debate as to the origin of the magnetic susceptibility signal. It might be an effect of dilution, where very fine magnetic particles form a constant aerosolic rain, but are relatively diluted by coarse nonmagnetic loess particles during glacial periods. Others have shown that at least part of the signal is due to the formation of fine-grained ferromagnetic particles during pedogenesis.

Soil morphology and micromorphology, used in conjunction with pollen and phytolith studies, are able to distinguish



Figure 5 The Luoquan loess section, Loess Plateau, China. The view shows the upper 50 m of the almost 200 m loess thickness. The prominent paleosol in the middle of the image is S5 (ca. 550 ka) with S4-S1 faintly visible above (photograph by Brad Pillans).

between full forest Alfisols in interglacials, and Mollisols and Inceptisols formed during interstadial weathering.

Beneath the main loess sequence in China is an extensive deposit called the Red Clay, which has now also been shown to be loess (Ding et al., 2000). The Red Clay was deposited between 7 and 2.6 Ma and was transported by westerly winds from dust-source regions in Northwestern China. At 2.6 Ma, a major shift in atmospheric circulation took place, and loess was instead deposited by northerly winds during the winter monsoon.

More detailed paleoclimatic studies are now appearing in which the pedostratigraphy of the Chinese loess deposits is being used to follow the history and variability of the East Asian paleomonsoon climate (e.g., An, 2000). Chinese scientists and their collaborators from the west lead the world in the study of loess–paleosol sequences and their climatic interpretation.

Palouse Loess, Northwestern USA

In the Palouse region of the Pacific Northwest of the USA, Busacca et al. (1992) and McDonald and Busacca (1992) provide a pedostratigraphic framework of loess accumulation, soils, tephra, and paleoclimate. The Palouse loess and paleosols are interesting for several reasons. First, they contain both petrocalcic (cemented by calcite) and duripans (cemented by silica), and their formation complicates the morphology and interpretation of paleosols. Second, their Washtucna paleosol apparently formed at the height of the Last Glacial Maximum, when mean accumulation rates of loess were lower rather than higher (Sweeney et al., 2005) because of the configuration of the then prevailing anticyclones. Glacial outburst floods then provided a ready source of material for more rapid

accumulation in the Late Glacial. Accumulation rates decreased in arid conditions just prior to the Holocene allowing the Sand Hills Coulee soil to form. Accumulation at a slow rate resumed in the Holocene. The Washtucna paleosol is also remarkable for an intense concentration of cicada burrows. O’Geen and Busacca (2001) showed that these cicadas were associated with sagebrush, and that this vegetation must have been dominant at the time. These studies prove the value of pedostratigraphy in challenging the established paradigm of loess accumulation during glacials and soil formation during interglacials.

The pedostratigraphy of the Palouse loess also illustrates the effects of soil formation during sediment accumulation, and that each paleosol has features inherited from upbuilding. Thus A horizons gradually become B horizons as slow loess accretion continues. This effect is probably the rule rather than the exception in many sedimentary environments. Even where loess accumulation has been very slow, in an extreme leaching and weathering environment on the super-humid West Coast of the South Island of New Zealand, Almond and Tonkin (1999) describe the pedogenesis of soils as being by upbuilding rather than from the surface downwards.

Loess in the Southern North Island of New Zealand

In the southern North Island of New Zealand, loess is widespread on Quaternary terraces formed in the last 500 ka. The general model for the deposition of this loess is that during cold climates erosion is maximized, rivers aggrade, and dust is blown from the aggrading flood plain and deposited as loess on higher surfaces (Vella, 1963). During warm climates the land surface is more stable, rivers degrade, soils develop in the loess, and marine terraces are cut by high sea levels (Palmer and Pillans, 1996). Most of the loess is quartzofeldspathic and

is derived from Mesozoic greywacke-argillite rocks, and Plio-Pleistocene mudstones and sandstones. However, there is also a significant input of both tephra and tephric loess from andesite and rhyolite volcanoes in the north of the district (Palmer and Pillans, 1996). Quartzofeldspathic loess dominates in the particularly cold glacials when erosion in the hinterland was at a maximum. The soils between loess units are characterized by upbuilding profiles of slow loess accumulation and tephra accretion (Lowe et al., 2008b).

Each loess can be matched to a river aggradation surface from which it was derived. Each major paleosol formed in a full interglacial is matched by a marine terrace. Less significant soils formed during interstadials when rivers temporarily degraded and the loess supply slowed.

Laboratory Studies

It is not always possible to infer taxonomic and paleoclimatic information from the morphology of paleosols alone. For example, progressive oxidation of organic matter in loess paleosols often confounds their classification and interpretation. Soils, decalcified during formation, are frequently recalcified with carbonate derived from overlying loess (Bronger, 2003). Micromorphology can detect A horizon fabrics, distinguish between primary and secondary carbonate (Bronger, 2003), and provide clear evidence of clay illuviation in the form of argillans (Brewer, 1964; Catt, 1990).

The intensity of weathering that soils have undergone can be assessed using both the amount and ratios of elements and oxides. Runge et al. (1974) used forms and amounts of phosphorus to identify paleosols in loess in the South Island of New Zealand. Their results show that both total and plant available forms of P are reduced in the B horizons of paleosols due to plant uptake and leaching, although a certain amount is returned to topsoil horizons by plant roots and litter. Childs (1975) used element distributions in New Zealand loess to confirm the location of paleosols. For quartzofeldspathic loess, he found that the distribution of Al, Si, and Zr provide evidence for uniformity of composition, while the depletion of P, K, and Ca indicated paleosols. In North Island loess containing tephra, he found that Si also became mobile (due to formation of allophane) indicating paleosols.

Molar ratios are also commonly used. For example, Rutter et al. (1978) used silica:sesquioxide and potassium:silica ratios to differentiate three paleosols in the Central Yukon. In tephric and quartzofeldspathic loess in the North Island of New Zealand, Palmer and Pillans (1996) used dry bulk density, percent quartz, magnetic susceptibility, and potassium distribution to study a loess-paleosol sequence dating back to 0.5 Ma.

Tephra Sequences

Tephra are time markers of events in volcanic eruptions, landscape evolution, paleoclimate, natural hazards and archaeology in many parts of the world, particularly near plate boundaries where volcanoes are more common (Lowe et al., 2008a). However, rhyolitic and dacitic tephras, in particular, can also be found many hundreds of kilometers from source. Soils that form on these tephras record information of relative stability and

paleoclimate. Textural and mineralogical differences between tephras in any sequence can make interpretation and pedostratigraphy difficult and it is usually important to study both proximal and distal sections. The sequence near Taupo, New Zealand (Figure 4), is an example of a proximal section of some of the tephras shown, whereas the section in Figure 3 is distal. The rate of soil formation in each deposit is a function of texture, climate, and mineralogy. It should be noted that because tephras are dispersed in different directions from source due to different wind directions, and are from different sources, the two sites illustrated do not contain the same tephras.

The amount of weathering and the extent of soil formation of each paleosol depend on climate (Figure 3) and the quiescent period between eruptions. In buried Andisols in Japan, Watanabe et al. (1998) found that the most developed Holocene paleosols, characterized by humus accumulation and allophane formation, formed in the Holocene Climatic Optimum. Younger paleosols show less weathering regardless of the time available for formation. The most important influence on the weathering products of tephras and tephra sequences is the silica concentration in the soil solution and the availability of aluminum to co-precipitate (Lowe, 1986). It is the balance between rainfall, drainage, tephra thickness, and depth of burial by further tephra and/or loess accretions that influences the formation of either allophane or halloysite. Conditions that favor allophane and imogolite formation include higher more evenly distributed rainfall (generally 1200 mm or more and few dry periods in summer); rapid drainage facilitating Si leaching; less influence of the organic cycle; shallow burial depth; and andesitic or basaltic (higher Al:Si) rather than rhyolitic composition. The opposite weathering conditions favor halloysite formation. Lowe (1986) found the length of weathering time to be subordinate, although tephric paleosols older than 350,000 years tend to be dominated by halloysite.

Tephra and paleosol sequences have also proven to be favorable repositories for opal phytoliths from plants. Kondo et al. (1994) examined phytoliths from tephra/loess/paleosol sequences in the North Island of New Zealand (one was close, and similar to the section shown in Figure 2). They found that phytoliths characteristic of grass or shrubs dominated parts of the sequences deposited in glacial periods while forest-derived phytoliths were dominant in the interglacial and interstadial periods.

Alluvial Sequences

Paleosols are a common feature of alluvial sequences, particularly in areas of tectonic subsidence or where rapid sedimentation or river avulsion has taken place both in Quaternary and ancient settings (Srivastava et al., 2010). Models of fluvial sedimentation are useful for interpreting paleosols in alluvial sequences (Aslan and Autin, 1998; Kraus, 2002). In each case, the paleosols inform paleoclimate and paleoenvironment although textural and compositional differences between soil parent materials may complicate interpretations. The carbon that is sometimes preserved in Holocene to late Pleistocene paleosols can be dated to provide a chronostratigraphy (e.g., Hall and Periman, 2007).

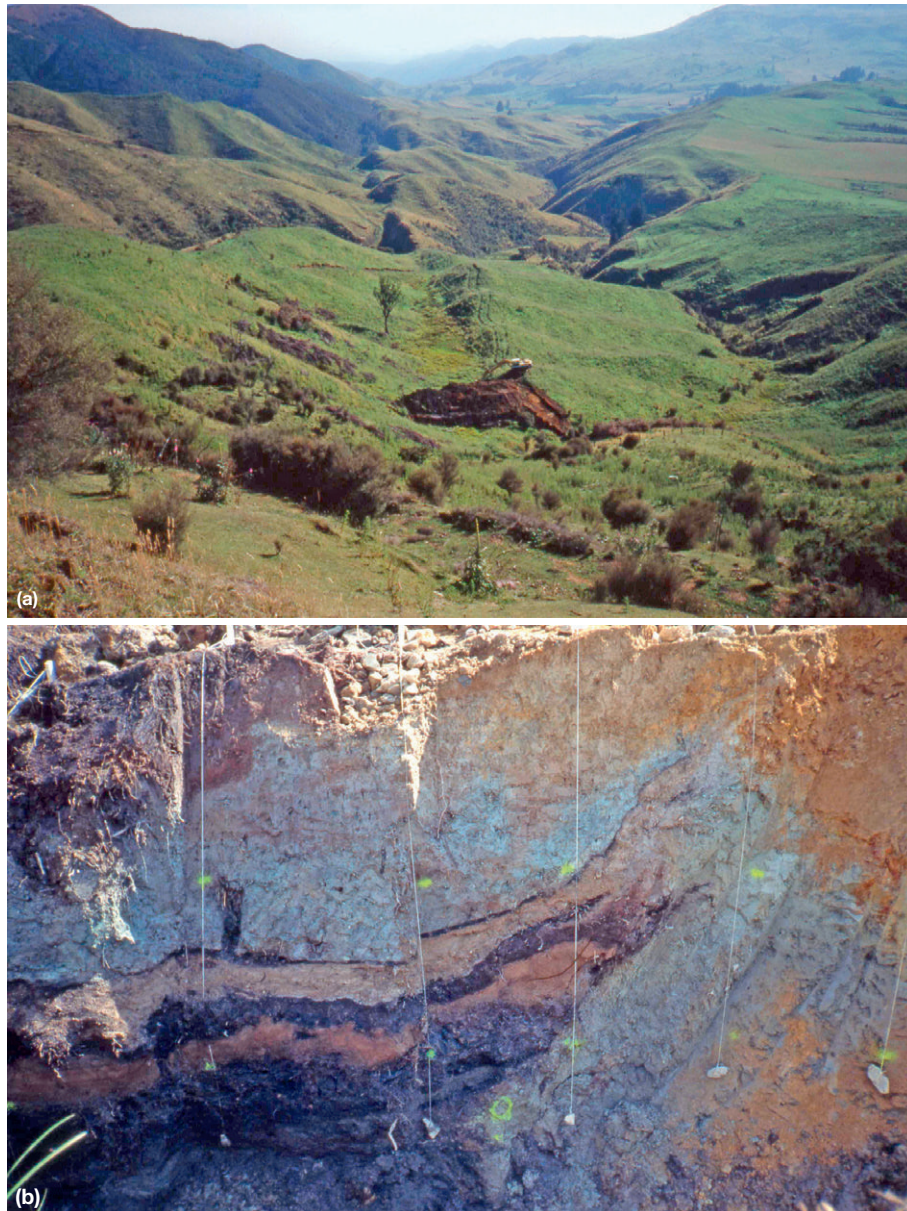


Figure 6 (a) The Ruahine fault, eastern North Island, New Zealand. The fault has dextral movement at $3\text{--}4\text{ mm year}^{-1}$, and is just one of several major strike-slip faults in the region. To the left (west) is Mesozoic greywacke-argillite and to the right (east) is Pliocene marine mudstone and limestone. Long-term uplift has been to the west, but more recent earthquakes in the Holocene have uplifted the eastern side. The fault trench affords excellent opportunities to study the magnitude and offset caused by earthquakes on the fault by excavating a trench with a digger (photograph by Vince Neall). (b) The main fault plane is clearly seen within the fault trench separating peaty paleosols and tephra (left) from blue-gray uplifted Pliocene mudstone (right). The upper black peaty paleosol formed on top of the Taupo Tephra (see [Figure 4](#)) and the second paleosol formed on the Waimihia Tephra. Soil formation on the Taupo Tephra was interrupted by deposition of greywacke and mudstone rubble following an earthquake. The history of fault movement can be pieced together by careful study of the fault trench, using tephrochronology, and radiocarbon dating of organic matter. Vertical strings and yellow dots are 1 m apart (photograph by Vince Neall).

Fans

Soils and paleosols preserved on and within alluvial, colluvial, and debris flow fans provide information on periods of fan building both spatially and temporally ([Carboni et al., 2006](#); [Srivastava et al., 2009](#)). In some environments, their carbon provides the only material dateable by the radiocarbon method. [Bennett et al. \(2006\)](#) found that laterally extensive

paleosols on fans in San Joaquin valley, California, marked sequence boundaries, in keeping with sequence stratigraphic models of fan growth. They found that ground-penetrating radar was able to image the paleosols, which are interrupted by numerous channels and channel deposits. In the Himalayan region, debris flow and alluvial fans preserve paleosols, which enables chronostratigraphic interpretation ([Srivastava](#)

et al., 2009). Loess cover on the fans preserves strongly developed paleosols and fluvial deposits preserve weakly developed paleosols in response to tectonics and climate change.

Fault Trenches

In studies of paleoseismicity, trenches are cut across fault lines and fault scarps where the geomorphic setting of the site favors sedimentation that has buried and preserved the evidence of past earthquakes (Figure 6). Pedostratigraphy is vital for a full understanding of events, rates, and chronology. Soils and peat layers preserved in fault trenches can provide evidence of quiescent periods between earthquakes and may be dated to provide limiting ages for fault movement. Fresh slope, alluvial, or colluvial deposits that bury soils can indicate a seismic event, although care must be taken to eliminate the possibility of storm or normal slope deposits. The stratigraphic relationships and offset of the paleosol and other deposits in the fault trench are then used to reconstruct a seismic history (e.g., Dolan et al., 2000; Langridge et al., 2003; Schermer et al., 2004).

Geoarchaeology

Pedostratigraphy is a vital tool when interpreting geology and archaeology of past civilizations. It may be used in both relatively short and long time periods. Butzer (2004) describes multiple paleosols in coastal eolian sands in the southwest of South Africa, dating back to Marine Oxygen Isotope 9, which preserve records of human occupation, their activities, and the foods they ate. The paleosols range from simple soils with weathered B horizons to calcretes or plinthite depending on the climate in which they formed and the temporal stability of the site. In Greece, red soils and paleosols (mostly Alfisols) developed in alluvium form chronosequences that span more than 200 000 years and are useful in interpreting human history of the region (van Andel, 1998). The degree of soil formation is linked to their age. At Pompeii, the soils buried by products of the cataclysmic eruption of AD 79 were rich volcanic soils that drew people to the area for agriculture. Their artifacts, buildings, and bodies are preserved on that ancient land surface (e.g., Jashemski and Meyer, 2002). Paleosols that developed in older volcanic deposits reveal a history of at least six Plinian eruptions in the last 17 000 years, some preserving evidence of human occupation.

Conclusions

Pedostratigraphy is the study of the stratigraphy, spatial relationships, and implications of surface and buried soils. A paleosol is a soil formed in a landscape of the past, and may be buried, nonburied, or exhumed. Buried soils may have the status of pedostratigraphic units, recognized in the North American Stratigraphic Code.

Pedostratigraphy gained impetus in a series of seminal studies in the 1950s and 1960s where it was used to good effect in landscape models that are widely quoted even today. These studies found that paleosols are the best records of metastability in landscape evolution, and are thus important parts of the stratigraphic record. However, scientists have had difficulty in

agreeing on a common nomenclature for use in pedostratigraphy, and this is probably a reflection of the use of pedostratigraphy by Quaternary scientists from different specializations for a variety of applications.

Pedostratigraphy has been used to best effect in the study of loess–paleosol sequences. It was first demonstrated in Europe that the loess sequences contained multiple soils, and that these soil–loess pairs represented many more fluctuations in climate than suggested by the record of Alpine glaciation. The multiple fluctuations in climate inevitably drew comparison with the oxygen isotope record of the deep sea. The application of magnetostratigraphy and magnetic susceptibility confirmed the correlation, leading to advanced and detailed land–ocean correlations, particularly in China.

Pedostratigraphy is also informative in a number of other settings where paleosols are useful in interpreting the geological and geomorphic setting and paleoclimate. Examples include tephra and alluvial sequences, fans, and fault trenches. Pedostratigraphy is also embraced by geoarchaeologists who need to understand the stratigraphy and environment of past civilizations.

See also: [Paleosols and Wind-Blown Sediments: Soil Micromorphology; Soil Morphology in Quaternary Studies.](#)
Quaternary Stratigraphy: [Chronostratigraphy; Climatostratigraphy; Continental Biostratigraphy; Lithostratigraphy; Sequence Stratigraphy; Tephrochronology.](#)

References

- Almond PC and Tonkin PJ (1999) Pedogenesis by upbuilding in an extreme leaching and weathering environment, and slow loess accretion, South Westland, New Zealand. *Geoderma* 92: 1–36.
- An ZS (2000) The history and variability of the East Asian paleomonsoon climate. *Quaternary Science Reviews* 19: 171–187.
- Aslan A and Autin WJ (1998) Holocene flood-plain soil formation in the southern lower Mississippi Valley: Implications for interpreting alluvial paleosols. *Geological Society of America Bulletin* 110(4): 433–449.
- Bennett GL, Weissmann GS, Baker GS, and Hyndman DW (2006) Regional-scale assessment of a sequence-bounding paleosol on fluvial fans using ground-penetrating radar, eastern San Joaquin Valley, California. *Geological Society of America Bulletin* 118: 724–732.
- Brewer R (1964) *Fabric and Mineral Analysis of Soils*, p. 470. New York: Wiley.
- Bronger A (2003) Correlation of loess–paleosol sequences in East and Central Asia with SE Central Europe: Towards a continental Quaternary pedostratigraphy and paleoclimatic history. Paleopedology: V International Symposium and Field Workshop, Suzdal, Russia. *Quaternary International* 106–107: 11–31.
- Bronger A and Catt JA (1998) Summary outline and recommendations on paleopedological issues: Revisitation of Concepts in Paleopedology. *Quaternary International* 51–52: 5–16.
- Busacca AJ, Nelstead KT, McDonald EV, and Purser MD (1992) Correlation of distal tephra layers in loess in the Channeled Scabland and Palouse of Washington State. *Quaternary Research* 37: 281–303.
- Butler BE (1959) *Periodic Phenomena in Landscapes as a Basis for Soil Studies*. Australia: CSIRO Soil Publication No. 14.
- Butzer KW (2004) Coastal eolian sands, paleosols, and Pleistocene geoarchaeology of the Southwestern Cape, South Africa. *Journal of Archaeological Science* 31: 1743–1781.
- Carboni S, Palomba M, Vacca A, and Carboni G (2006) Paleosols provide sedimentation, relative age, and climatic information about the alluvial fan of the River Tirso (Central-Western Sardinia, Italy). *Quaternary International* 156–157: 79–96.

- Catt JA (1986) *Soils and Quaternary Geology: A Handbook for Field Scientists. Monographs on Soil and Resource Surveys* No. 11: p. 267. Oxford: Clarendon Press.
- Catt JA (ed.) (1990) Paleopedology Manual. *Quaternary International* 6: 95.
- Catt JA (1998) Report from working group on definitions used in paleopedology: Revisitation of concepts in paleopedology. *Quaternary International* 51–52: 84.
- Childs CW (1975) Distribution of elements in two New Zealand Quaternary loess columns. In: Suggate RP and Cresswell MM (eds.) *Quaternary Studies, Selected Papers from IX INQUA Congress, Christchurch, New Zealand Royal Society of New Zealand Bulletin*, vol. 13, pp. 95–100, 321.
- Ding Z, Rutter N, Liu T, Evans ME, and Wang Y (1991) Climatic correlation between Chinese loess and deep sea cores: A structural approach. In: Liu T (ed.) *Loess, Environment and Global Change*, pp. 168–186. Beijing: Science Press.
- Ding ZL, Rutter NW, Sun JH, Yang SL, and Liu TS (2000) Re-arrangement of atmospheric circulation at about 2.6 Ma over northern China: Evidence from grain size records of loess–paleosol and red clay sequences. *Quaternary Science Reviews* 19: 54–558.
- Dolan JF, Sieh K, and Rockwell TK (2000) Late Quaternary activity and seismic potential of the Santa Monica fault system, Los Angeles, California. *Geological Society of America Bulletin* 112: 1559–1581.
- Follmer LR (ed.) (1995) *Paleopedology Commission Newsletter Series* No. 8: May 1993 Symposium Part A, 25 p. No. 9, July 1993 Symposium Part B, 39 p. No. 10, August 1993 Symposium Part C, 36 p. No. 11, Part 1, July 1993 (Addendum to No. 10) Podostratigraphic Issues, 12 p. Part 2, August 1993 (Addendum to No. 10) Symposium Part D, Revised and Reprinted October 1995, 35 p.
- Follmer LR (1998) Preface. In: Follmer LR, Johnson DL, and Catt JA (eds.) *Revisitation of Concepts in Paleopedology*, 1–3. *Quaternary International* 51–52.
- Hall SA and Periman RD (2007) Unusual Holocene alluvial record from Rio Del Oso, Jemez Mountains, New Mexico. *Paleoclimate and Archaeological Significance. New Mexico Geological Society Guidebook, 58th Field Conference* *Geology of the Jemez Mountains Region*, vol. II, pp. 459–468.
- Hewitt AE (1998) New Zealand soil classification. *Landcare Research Science Series* No. 1: p. 133. Lincoln, Canterbury, New Zealand: Manaaki Whenua Press.
- Holliday VT (1998) Soils, paleosols and geosols: A commentary. *Quaternary Perspectives* 2: 4.
- Jashemski WMF and Meyer FG (2002) *The Natural History of Pompeii*, p. 504. Cambridge, UK: Cambridge University Press.
- Johnson DL (1998) Paleosols are buried soils: Revisitation of Concepts in Paleopedology. *Quaternary International* 51–52: 7.
- Kondo R, Childs C, and Atkinson I (1994) *Opal Phytoliths of New Zealand*, p. 85. Lincoln: Manaaki Whenua Press.
- Kraus MJ (2002) Basin-scale changes in floodplain paleosols: Implications for interpreting alluvial architecture. *Journal of Sedimentary Research* 72: 500–509.
- Kukla GJ (1975) Loess stratigraphy of central Europe. In: Butzer K and Isaac GL (eds.) *After the Australopithecines*, pp. 99–188. The Hague: Mouton.
- Kukla GJ (1987) Loess stratigraphy in central China. *Quaternary Science Reviews* 6: 191–219.
- Kukla GJ and Koci A (1972) End of the last interglacial in the loess record. *Quaternary Research* 2: 374–383.
- Langridge R, Campbell J, Hill N, et al. (2003) Paleoseismology and slip rate of the Conway Segment of the Hope Fault at Greenburn Stream, South Island New Zealand. *Annals of Geophysics* 46: 1119–1139.
- Liu XM, Liu TS, Xu TC, Liu C, and Chen MY (1987) A preliminary study on magnetostratigraphy of a loess profile in the Xifeng area, Gansu Province. In: Liu TS (ed.) *Aspects of Loess Research*, pp. 164–174. Beijing: China Ocean Press.
- Lowe D (1986) Controls on the rates of weathering and clay mineral genesis in airflow tephras: A review and New Zealand case study. In: Colman SM and Dethier DP (eds.) *Rates of Chemical Weathering of Rocks and Minerals*, ch. 12, pp. 265–330. Orlando, FL: Academic Press.
- Lowe D, Alloway B, and Shane P (2008a) Far-flung markers. In: Graham I (ed.) *A Continent on the Move: New Zealand Geoscience into the 21st Century*, 170–173 GSNZ Miscellaneous Publication 124, 388 p.
- Lowe D, Tonkin P, Palmer A, and Palmer J (2008b) Dusty horizons. In: Graham I (ed.) *A Continent on the Move: New Zealand Geoscience into the 21st Century*, 270–273 GSNZ Miscellaneous Publication 124, 388 p. Lower Hutt, NZ: Geological Society of New Zealand, Wellington NZ and GNS Science.
- Lozek V (1969) Palaontologische Charakteristik der Loess-Serien. In: Demek J and Kukla J (eds.) *Periglazialzone Loess und Palaolithikum der Tschechoslowakei*, pp. 43–59. Brno: Geographical Institute, Czechoslovakian Academy of Sciences.
- McDonald EV and Busacca AJ (1992) Late Quaternary stratigraphy of loess in the Channeled Scabland and Palouse regions of Washington State. *Quaternary Research* 38: 141–156.
- Morrison RB (1964a) *Lake Lahontan: Geology of the Southern Carson Desert*. vol. 401 p. 156. Nevada: U.S. Geological Survey Professional Paper.
- Morrison RB (1964b) *Soil Stratigraphy: Principles, Applications to Differentiation of Quaternary Deposits and Landforms, and Applications to Soil Science*. PhD Dissertation, University of Nevada, Reno. *Dissertation Abstracts* 28(4), 1967.
- Morrison RB (1967) Principles of quaternary soil stratigraphy. In: Morrison RB and Wright HE Jr. (eds.) *Proceedings of the VII Congress of International Association for Quaternary Research (INQUA)*, 1965, *Quaternary Soils*, vol. 9, pp. 1–69. Reno, NV: Center for Water Resources Research, Desert Research Institute, University of Nevada.
- Morrison RB (1978) Quaternary soil stratigraphy: Concepts, methods, and problems. In: Mahaney R (ed.) *Quaternary Soils*, Geoabstracts, Norwich.
- Morrison RB (1998) Report from the working group on podostratigraphy. In: Follmer LR, Johnson DL, and Catt JA (eds.) *Revisitation of Concepts in Paleopedology*, 81–86. *Quaternary International* 51–52.
- North American Commission on Stratigraphic Nomenclature (1983) North American Stratigraphic Code. *American Association of Petroleum Geologists Bulletin*, vol. 67, 841–875.
- O'Geen AT and Busacca AJ (2001) Faunal burrows as indicators of paleo-vegetation in eastern Washington, USA. *Palaeogeography, Palaeoclimatology, Palaeoecology* 169: 23–37.
- Palmer AS and Pillans BJ (1996) Record of climatic fluctuations from ca. 500 ka loess deposits and paleosols near Wanganui, New Zealand. *Quaternary International* 34–36: 155–162.
- Pillans B (1990) Late Quaternary marine terraces, South Taranaki–Wanganui, New Zealand Geological Survey, Miscellaneous Series Map 18..
- Retallack GJ (1998) Core concepts of paleopedology. In: Follmer LR, Johnson DL, and Catt JA (eds.) *Revisitation of Concepts in Paleopedology*, 203–312. *Quaternary International* 51–52.
- Ruhe RV (1956) Geomorphic surfaces and the nature of soils. *Soil Science* 82: 441–455.
- Ruhe RV (1969) *Quaternary landscapes in Iowa*, p. 255. Ames, IA: Iowa State University Press.
- Runge ECA, Walker TW, and Howarth DT (1974) A study of Late Pleistocene loess deposits, South Canterbury, New Zealand: Part 1. Forms and amounts of phosphorus compared with other techniques for identifying paleosols. *Quaternary Research* 4(1): 76–84.
- Rutter NW, Foscolos AE, and Hughes OL (1978) Climatic trends during the Quaternary in central Yukon based upon pedological and geomorphological evidence. In: Mahaney WC (ed.) *Quaternary Soils*. Norwich: Geo Books.
- Salvador A (ed.) (1994) *International Stratigraphic Guide: A Guide to Stratigraphic Classification, Terminology and Procedure*, 2nd edn., p. 214. Boulder, CO: The International Union of Geological Sciences and The Geological Society of America.
- Schermer ER, Van Dissen R, Berryman KR, Kelsey HM, and Cashman SM (2004) Active faults, paleoseismicity, and historical fault rupture in northern Wairarapa, North Island, New Zealand. *New Zealand Journal of Geology and Geophysics* 47: 101–122.
- Srivastava P, Rajak MK, and Singh LP (2009) Late Quaternary alluvial fans and paleosols of the Kangra basin, NW Himalaya: Tectonic and paleoclimatic interpretations. *Catena* 76: 135–154.
- Srivastava P, Rajak MK, Sinha R, Pal DK, and Bhattacharyya T (2010) A high-resolution micromorphological record of the Late Quaternary paleosols from the Ganga–Yamuna interfluvies: Stratigraphic and paleoclimatic implications. *Quaternary International* 227: 127–142.
- Sweeney MR, Busacca AJ, Richardson CA, Blinnikov M, and McDonald EV (2005) Glacial anticyclone recorded in Palouse loess or northwestern United States. *Geology* 32: 705–708.
- Tonkin PJ (1994) Principles of soil-landscape modelling and their application to the study of soil–landform relationships in New Zealand. In: Webb TH (ed.) *Soil-Landscape Modelling in New Zealand*. Proceedings of a Workshop Held at Aokautere, New Zealand, 8–9 February 1993, p. 173. *Landcare Research Science Series* No. 5: pp. 20–37. Lincoln, Canterbury: Manaaki Whenua Press.
- van Andel TH (1998) Paleosols, red sediments, and the Old Stone Age in Greece. *Gearchaeology* 13: 361–390.
- Vella P (1963) Upper Pleistocene succession in the inland part of Wairarapa Valley, New Zealand. *Transactions of the Royal Society of New Zealand: Geology* 2: 63–78.
- Vucetich CG and Pullar WA (1969) Stratigraphy and chronology of Late Pleistocene volcanic ash beds in Central North Island, New Zealand. *New Zealand Journal of Geology and Geophysics* 12: 784–837.
- Watanabe M, Aoki K, and Sakagami K (1998) Humus accumulation in Holocene paleosols formed in Japanese tephra. *Catena* 34: 35–46.
- Worthen AH (1873) Geology of Sangamon County. In: Worthen AH (ed.) *Geology and Palaeontology: Illinois Geological Survey*, vol. V, pp. 235–319. Champaign, IL: Illinois State Geological Survey.
- Wysocki DA, Schoeneberger PJ, and LaGarry HE (2005) Soil surveys: A window to the subsurface. *Geoderma* 126: 167–180.
- Yaalon DH (ed.) (1971) *Paleopedology: Origin, Nature and Dating of Paleosols*, p. 350. Jerusalem: International Society of Soil Science and the Israel University Press.

Sequence Stratigraphy

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Introduction

Sedimentary cyclicity is an intrinsic feature of marine sedimentary basins, and is controlled by relative sea-level change resulting from the interplay of tectonics, sediment supply, and eustasy (Miall, 1997). Higher frequency cycles of glacio-eustatic origin have periodicities of tens or hundreds of thousands of years. Lower frequency cycles are millions to hundreds of millions of years in length and are driven by sea-level changes tied ultimately to plate tectonic movements.

Seismic reflection images of continental margins depict unconformity-bounded sedimentary cycles with an internal arrangement of reflectors that correspond to stratal surfaces (=depositional surface of shelf margin) (Mitchum et al., 1977). In profile, the stratal architecture of sedimentary cyclicity is revealed on a basin-wide scale, and its development can be tracked across many successive cycles of relative sea-level change. Mitchum et al. (1977) termed individual cycles of this type 'third-order seismic sequences,' where a sequence is defined as "a stratigraphic unit composed of a relatively conformable succession of genetically related strata and bounded at its top and base by unconformities or their correlative conformities."

Historic studies of cyclic Quaternary sediments grouped repetitive faunal and lithological facies into units termed 'cyclothem' (Fleming, 1953; Vella, 1963), and stratal relationships within similarly cyclic sequences were later inferred using outcrop sedimentary facies analysis (Loutit et al., 1988; van Wagoner et al., 1990). Since the late 1980s, a modern sequence stratigraphic approach has been applied successfully to cyclic outcrop successions of Quaternary age (Abbott and Carter, 1994; Ito, 1992; Naish and Kamp, 2007), in which a much higher stratal resolution (down to centimeters) can be differentiated than is the case for conventional seismic analysis. Following the seismic work of Payton et al. (1977) and Vail et al. (1987), seismic sequences have become variously known as stratigraphic sequences, depositional sequences, or (as used herein) sequences.

Clarity of thought is aided by drawing a clear distinction between a sequence stratigraphic model (SSM) – which summarizes the geometric stratal relationships of sediments deposited during a single major cycle of sea-level change – and a global sea-level model (GSM), which hypothesizes a sea-level curve for past times (Carter and Naish, 1998; Carter et al., 1991). A major aspect of classic sequence stratigraphy (Payton, 1977) was the development of techniques whereby a sea-level curve could be reconstructed from geometric analysis of sequences on seismic profiles, which led to the widely used sea-level curves of Vail et al. (1977) and Haq et al. (1987). Such studies contain an inescapable element of circular reasoning, and it is the single greatest strength of Quaternary sequence

stratigraphic studies that they are made against the independent, and now well-tested, model of sea-level variation that is provided by oceanic oxygen isotope measurements (e.g., Shackleton et al., 1990; Lisiecki and Raymo, 2006).

Sequence Stratigraphic Models

An SSM is a summary of the stratal architecture and facies composition that occur within a sequence (Carter et al., 1991). The stratal patterns and sedimentary environments that define the architecture of the classic SSM of Payton et al. (1977) and Vail et al. (1987) are depicted in Figure 1. Between bounding disconformities formed by subaerial exposure during lowered sea level, the model depicts an arrangement of stratal geometries (onlap, downlap, toplap, backlap) that track the development of the sequence as it built firstly landward and then seaward during a relative sea-level cycle (Figures 1(a) and 2). Stratal geometries allow sequences to be subdivided into component packages of strata called 'systems tracts,' each associated with a segment of the controlling sea-level curve. Lowstand, transgressive, highstand, and shelf-margin systems tracts are recognized. Systems tracts are defined as a linkage of contemporaneous depositional systems, where a depositional system is in turn defined as a three-dimensional assemblage of lithofacies (Brown and Fisher, 1977; Van Wagoner et al., 1988). Thus, the SSM predicts a systematic distribution of environmentally controlled lithofacies (Figure 1(b)).

During the initial falling limb of the sea-level curve, the shoreline moves basinward toward and ultimately below the shelf edge. The former continental shelf is exposed to subaerial erosion, forming a lower bounding unconformity referred to as the sequence boundary (SB1). Subsequently, sediment is delivered to the slope where it onlaps the SB to form a basin-floor turbidite fan followed above by a wedge of terrestrial, shoreline, and shelf facies, the whole termed as a lowstand systems tract (LST). Subaerial exposure and consequent slow rates of lignite or soil formation at SBs are confirmed by their marked ¹⁰Be-isotope signature (Carter et al., 2002).

With rising sea level, the shoreline moves landward and inundates the lowstand coastal plain. Commonly, a transgressive surface of erosion (TSE; = ravinement surface, RS) is superimposed upon the former subaerial SB, accompanied by a shell lag containing reworked shoreface mollusks and remane fossils. Any fluvial channels that crossed the shelf to the former lowstand shoreline are delimited by the SB below, but have their LST fill beveled above by the TSE. Sediment bodies within such fluvial channels are referred to as incised valley fill. Deposition of a transgressive systems tract (TST) follows, with shallow marine sediments onlapping the TSE/SB in a landward

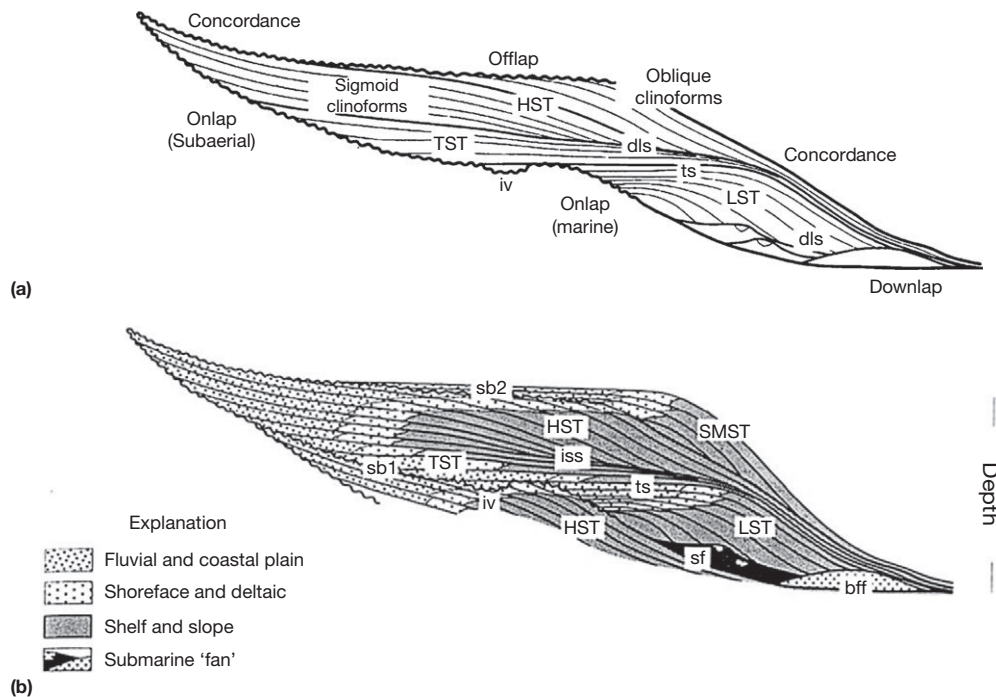


Figure 1 The third-order sequence model. (a) Systems tracts and their internal stratal geometries (offlap, onlap, downlap, backlap, toplap). Reproduced from Miall AD (1997) *The Geology of Stratigraphic Sequences*, p. 433. Berlin and Heidelberg: Springer-Verlag. (b) Systems tracts, surfaces, and the distribution of lithofacies (Christie-Blick, 1991). LST, lowstand systems tract; TST, transgressive systems tract; HST, highstand systems tract; SMST, shelf-margin systems tract; sb, sequence boundary; ts, transgressive surface; dls, downlap surface; iv, incised valley; bff, basin-floor fan; sf, slope fan; iss, interval of sediment starvation.

direction. Above the TSE and any shell lag, the vertical succession of lithofacies deepens upward from inner to offshore shelf facies. The distal TST tapers to a feather edge as the decreasing rate of terrigenous sediment supply causes stratal surfaces to converge seaward. The sediment-starved offshore shelf seafloor that results is marked in outcrop by a local flooding surface (LFS). ¹⁰Be-isotope studies show that the outcrop LFS corresponds with a significant hiatus (Graham et al., 1998), and the surface is often followed by a condensed, *in situ* shellbed with offshore fauna termed the mid-cycle shellbed (MCS) or “backlap” shellbed (Kidwell, 1991; Kondo et al., 1998). Such mid-cycle condensed successions are often associated with the presence of authigenic minerals such as glauconite and phosphate (Loutit et al., 1988).

As the rate of sea-level rise slows toward highstand and subsequent fall, clinoform strata advance basinward by offlap and downlap to form the highstand systems tract (HST). An abrupt boundary termed the downlap surface (DLS) occurs between the base of the HST and the underlying MCS or, in its absence, the top of the TST. The vertical succession of highstand facies shallows upward from offshore sediments at the base through shoreline and then into nonmarine facies at the top. As for the distal TST, distal highstand deposition is characterized by stratal convergence associated with low rates of terrigenous sediment supply.

At the approach of the next sea-level low, which ends each cycle, the shoreline and locus of deposition again shift basinward. Should the lowstand shoreline not fall below the shelf edge, Payton et al. (1977) distinguishes a SMST as forming at the start of the next new sequence, instead of a succeeding LST. As for the LST and HST, a SMST is characterized by an off-lapping, shallowing-upward facies succession.

Though it was part of the initial formulation of the SSM based on third-order cyclicity (Payton et al., 1977), the SMST has found little subsequent use. Rather, later authors have preferred to recognize two other types of regressive sediment bodies as characteristic of the falling sea-level phase. Regressive packages of strata form when the supply of sediment is large enough to form an offlapping body of sediment that builds into the basin as sea-level falls and the shoreline moves seaward (Plint, 1988), as is characteristic of Quaternary sequences from active-margin basins where high sediment fluxes are derived from nearby orogens. When they are bounded below by a regressive surface of erosion (RSE), generated by wave-base scour of the seafloor during regression, such deposits are assigned to a forced regressive systems tract (FRST) (Hart and Long, 1996; Hunt and Tucker, 1992). However, in many cases, similar sand-dominated regressive packages are gradationally based, presumably reflecting subsidence rates high enough to prevent a sea-level lowered wave base impinging on the seafloor. Such gradationally based sediment packages are common in the Wanganui Basin and are termed regressive systems tracts (RST) (Naish and Kamp (1997a). Sea-level lowstand is associated with the formation of the upper sequence-bounding unconformity (SB1).

Five Fundamental Sequence-Bounding and Intrasequence Surfaces

Following the theoretical analysis of Embry (1990), the study of Quaternary sequences has played a key role in developing our understanding of the outcrop characteristics of the

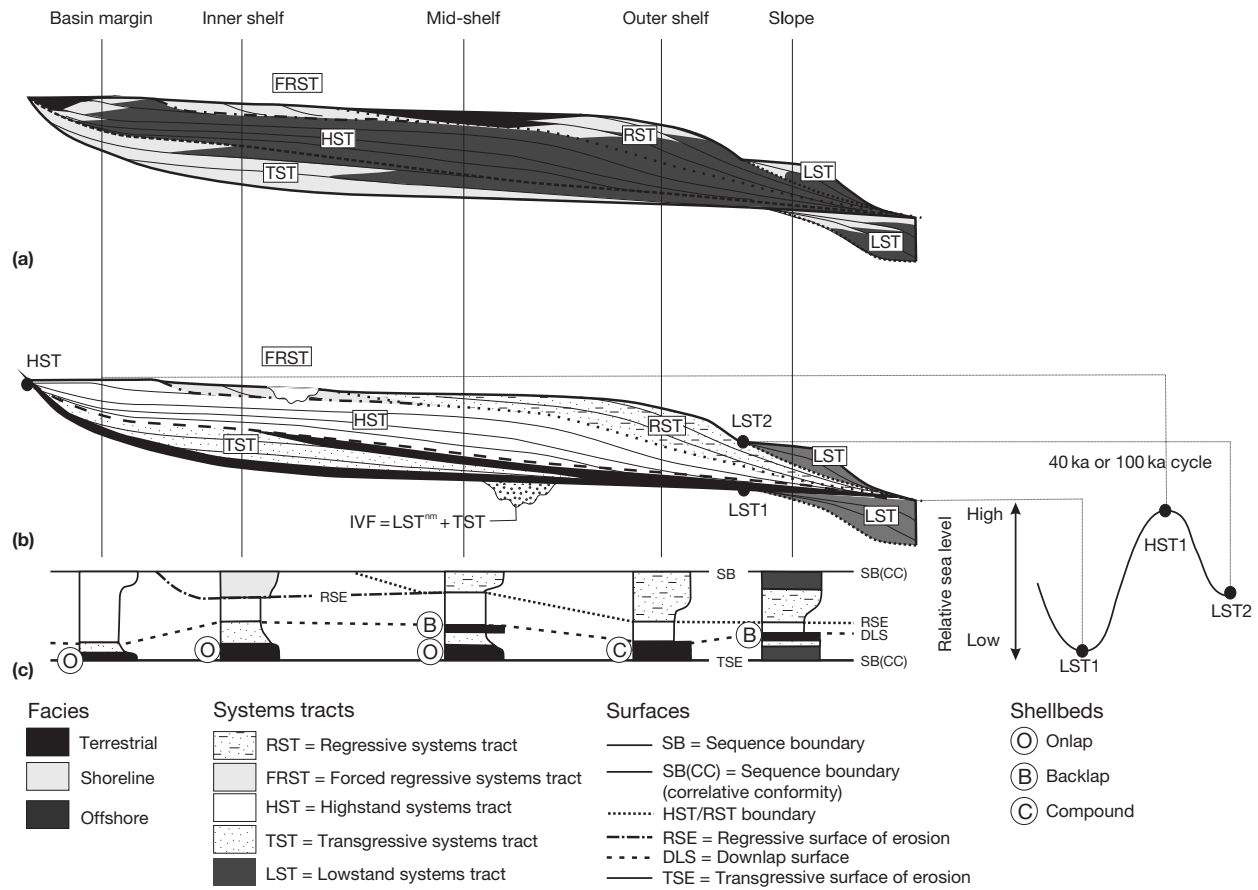


Figure 2 Quaternary sequence model based on outcrop studies of active-margin basins. (a) Quaternary sequence model emphasizing systems tracts, stratal architecture, and generalized lithofacies distribution. (b) Quaternary sequence model emphasizing systems tracts, and the distribution of onlap and backlap shellbeds. (c) Indicative sequence motifs showing across-shelf variation in the vertical succession of facies, surfaces and systems tracts. Adapted from Naish TR and Kamp PJJ (1997) Sequence stratigraphy of sixth-order (41 ka) Pliocene-Pleistocene cyclothems, Wanganui Basin, New Zealand: A case for the regressive systems tract. *Geological Society of America Bulletin* 109(8): 978–999; Saul G, Naish T, Abbott ST, and Carter RM (1999) Sedimentary cyclicity in the marine Plio–Pleistocene: Sequence stratigraphic motifs characteristic of the last 2.5 my. *Geological Society of America Bulletin* 111: 524–537.

stratigraphic discontinuities that punctuate a typical stratigraphic sequence. However, note that a widely used term in older literature – the maximum flooding surface – is a conceptual horizon that has no necessary physical manifestation in outcrop (Carter et al., 1998). Recognition of the key physical surfaces, and the abrupt facies shifts that occur across them, are primary tools for sequence stratigraphic interpretation. The characteristics of sequence surfaces are summarized in Figure 2 and below.

Sequence Boundary

A SB is a surface formed by subaerial exposure that occupies a sequence-bounding position between an underlying RST, or HST, in the case where erosion has removed the RST and an overlying TST. The SB may be marked by incised fluvial channels, by a paleosol, or by terrestrial facies such as lignite, all of which formed during the sea-level lowstand associated with the formation of the SB. An SB-correlative conformity (conceptual horizon projected from the SB into adjacent conformable strata) may occur inland if aggradation of coastal plain facies

continues uninterrupted across the contact between two sequences. Offshore, an SB-correlative conformity extends basinward from beneath the LST.

Ravinement Surface or Transgressive Surface of Erosion

The RS and TSE are alternative, widely used synonyms for the surface that is wave-cut by the landward passage of a transgressing shoreface. The RS often erodes any underlying LST strata, its lower bounding SB and sometimes the upper part of the underlying sequence. The RS may be penetrated by marine crustacean burrows (*Thalassinoides*, *Ophiomorpha*) and pholad borings, and is often overlain by an onlapping shell lag or conglomerate.

Local Flooding Surface

The LFS marks the development of a sediment-starved seabed seaward of the shore-connected prism of the TST, and results from water deepening associated with shoreline transgression. The LFS is usually marked by a burrowed diastemic surface

between the top of the TST and an overlying MCS containing an *in situ* offshore shelf fauna.

Downlap Surface

The DLS is an abrupt but conformable boundary formed by downlap of the HST on to the underlying MCS or (in its absence) TST. The DLS may be burrowed and often coincides with a sharp lithological contrast between an underlying *in situ* shellbed and an overlying HST shelf siltstone.

Regressive Surface of Erosion

The RSE is a surface cut by a lowering wave base during regression and sea-level fall (Plint, 1988), and, where present, comprises the boundary between the HST and FRST. As for the DLS, a sharp lithological contrast may occur across the RSE in shelf settings, where it separates HST mudstone from FRST sand. In rapidly subsiding basins, where a balance exists between the rates of subsidence and sea-level fall, the HST passes gradationally upward into the RST without an intervening RSE.

Parasequences

In many depictions of the SSM, marine flooding surfaces correspond to the boundaries between high order, shallowing-upward, sedimentary cycles referred to as parasequences. These are regarded as the fundamental building blocks of sequences and systems tracts by Van Wagoner et al. (1988), who defined them as “a relatively conformable succession of genetically related beds . . . , bounded by flooding surfaces and their correlative surfaces.”

Depositional sequences are ultimately the product of the interplay between sedimentation, subsidence, and eustasy. Changes in the episodicity, magnitude, and rates of these fundamental parameters interact to vary the architecture and facies composition of depositional sequences according to the order of cyclicity and the particular basin setting. These considerations led Carter et al. (1991) to conclude that higher order sequences (e.g., parasequences) within lower order sequences are likely to be topologically distinct only at their particular level, rather than stacked as in a nested set of Russian dolls. These authors suggested that, rather than having a primary significance, parasequences are simply a convenient descriptive label, the scale of which varies according to sequence order.

Seven Empirical Orders of Sea-Level Cyclicity

Among the voluminous literature on sequence stratigraphy, examples exist of sequence architecture, which is interpreted in terms of about seven different orders of sea-level cyclicity (Fulthorpe, 1991). The first and second orders are ultimately tied to plate tectonic processes (Carter et al., 1991). Third- and fourth-order cycles are variously attributed to regional tectonic change (Saul et al., 1998) or to eustasy of unknown origin (Payton et al., 1977). The fifth, sixth, and seventh orders of sea-level cyclicity correspond to glacio-eustasy at orbitally forced Milankovitch scales (Miall, 1997).

The product of these various orders of sea-level variation can be integrated as a GSM (Haq et al., 1987), but such curves are bedeviled by their low resolution (sequences 1–10 My long) and uncertain accuracy. In reality, we only possess an acceptably accurate model of sea-level change for the last few million years, as underpinned by high resolution oxygen isotope time-series and related studies (Lisiecki and Raymo, 2005). It is the accuracy of the oceanic oxygen isotope sea-level curve as a proxy against which to assess stratigraphic change that makes studies of Quaternary sequence stratigraphy so exceptionally powerful.

Sequence Stratigraphy in Quaternary Sediments

Extending back to 2.6 Ma and beyond, orbitally forced glacio-eustasy has operated throughout the Quaternary with dominant frequencies of 20 ka (precession), 41 ka (obliquity), and 100 ka (eccentricity), and magnitudes between 70–130 m (see review by Pillans et al., 1998). Eustatic variation at these Milankovitch frequencies has driven numerous transgressions and regressions of the shorelines across the world's continental shelves, and imposed a distinct cyclicity on the Quaternary sedimentary successions that reside beneath modern continental shelves. A number of such basins located adjacent to active plate margins have been uplifted and exposed during the late Quaternary (Table 1), thereby allowing their detailed sequence stratigraphy to be described against the background of a known sea-level history. The discussion of Quaternary sequences below is based on the published literature on these active-margin successions.

Eustasy as the Dominant Control on Sediment Deposition

Tuning (or calibration) of the isotopic proxy sea-level curve with astronomically based master curves (e.g., seasonal insolation)

Table 1 Well documented outcropping Quaternary sequences from active-margin plate margins

Basin	Region and country	Key references
Wanganui	North Island, New Zealand	Abbott and Carter (1994), Naish and Kamp (1997a), Saul et al. (1999)
Esmeraldas-Caraquez, Canoa	Central Ecuador	Cantalamesa and Di Celma (2004), Di Celma et al. (2005)
Merced Group	San Francisco, USA	Carter et al. (2002), Clifton et al. (1988)
Kazusa Group, Omma Formation	Central Japan	Ito (1992, 1998), Kitamura et al. (1994)
Peradriatic	Italy	Cantalamesa et al. (2005), Rio et al. (1996)
Corinth	Greece	McMurray and Gawthorpe (2000)

has established that between 2.6 and 0.8 Ma, sea level varied systematically at the 41 ka obliquity wavelength, but that since 0.8 Ma, the 100 ka eccentricity signal has dominated. Magnitudes of Quaternary glacioeustasy can be calibrated between the oxygen isotope record and the sea-level changes inferred from uplifted coral terraces younger than the last interglacial (Chappell, 1974). This calibration has been used to estimate eustatic magnitudes of ~70 m for obliquity with cycles and up to 130 m for eccentricity cycles (Chappell et al., 1986). The oxygen isotope record from deep sea cores thus furnishes an independently established proxy eustatic curve against which to interpret Quaternary sequence stratigraphy.

Generally, rates of Quaternary glacio-eustasy outpace both subsidence and sediment supply as the dominant control on the sedimentary architecture of continental margin basins. Exceptions include the supply-driven Mississippi delta complex (Boyd et al., 1988) and the Canterbury Plains of New Zealand (Leckie, 1994), where different segments of the same coast in each case display concurrent offlapping and onlapping behavior during the Holocene highstand. However, more generally, the sequence architecture of some basin successions with good chronological control can be correlated to the oxygen isotope proxy sea-level curve. Following the conventional SSM, sediments deposited across a Quaternary highstand shelf can be matched to interglacial sea-level highs, while their contained sequence boundaries correspond to the intervening glacial lowstands (Abbott and Carter, 1994; Ito, 1992; Pillans et al., 1994, 2005).

Sedimentary Facies Analysis

The architecture and depositional history of sequences are rendered using standard sedimentary facies analysis. This approach uses sediment attributes such as grain size, physical sedimentary structures, biogenic sedimentary structures, and macrofossil content to interpret environments of deposition (Walker, 1984). Lithofacies analysis is enhanced by identification of the significant surfaces that punctuate sequence successions and by focused sedimentological and paleoecological analysis. The taxonomic composition of macrofossil assemblages is indicative of environmentally diagnostic paleocommunities (Abbott and Carter, 1997; Kidwell, 1991; Kondo et al., 1998). Thus, estuarine faunas are distinct from open shoreface faunas, and both are distinct from offshore subtidal communities. Taphonomic attributes of macrofossil material indicate sedimentary processes (Kidwell, 1991b). For example, broken and abraded assemblages indicate reworking in an energetic environment (death assemblages), whereas pristine assemblages characterized by the presence of paired valves (life assemblages) indicate deposition in low energy settings. Foraminiferal paleoecology is also a powerful tool with which to interpret facies and track bathymetric changes in Quaternary marine strata (Abbott, 1997; Haywick and Henderson, 1992; Naish and Kamp, 1997b). Overall, analysis of the sedimentology and paleoecology of Quaternary sedimentary successions is highly robust because close contemporary analogues often occur in adjacent modern sedimentary environments.

The distribution of facies within a sequence is the basis from which stratal geometries and sequence architecture can

be inferred. For example, a deepening-upward facies succession that fines upward from physically stratified shoreline sand into pervasively bioturbated shelf mud, has inferred onlapping stratal geometry such as characterizes the TST. Conversely, a shallowing- and coarsening-upward succession represents offlapping geometry such as that found in the HST–RST. Thus a conformable succession of gradually changing facies may often correspond to a systems tract (Walker, 1990).

The Importance of Shellbeds

Shell-rich facies are a hallmark of eustatically influenced Quaternary sequences and their interpretation is crucially important for resolving sequence architecture. Shell material becomes concentrated under conditions of low net terrigenous sediment accumulation created by either reworking and bypass in shallow energetic sandy environments (type A shellbeds; Abbott and Carter, 1994), nondeposition in nearshore deeper channel or offshore deeper water muddy environments (type B shellbeds, *op. cit.*), or by infaunal penetration of specialist boring taxa into a current-swept, hard substrate (type C shellbeds, *op. cit.*).

All such shelly deposits represent stratigraphic condensation, and type A and B shellbeds both imply convergence of stratal surfaces toward sequence and systems tract bounding surfaces (Kidwell, 1991a; Kondo et al., 1998; Naish and Kamp, 1997a). Individual shellbeds can be assigned an inferred mode of origin with respect to sequence architecture (Figures 2 and 3), as follows. Basal onlap shellbeds represent transgressive reworking of contemporaneous shelly shoreline and estuarine sediments at TSE, with an admixture of underlying remanie fossils. Fossil content comprises mostly mixed, rounded, and transported assemblages dominated by sandy shoreface and estuarine mollusks. Inferred backlap shellbeds correspond to MCS, and are characterized by offshore shelf faunas, which include *in situ* articulated bivalves. Downlap shellbeds that mark the contact of MCS and HST are characterized by similar offshore shelf faunas but display a more dispersed biofabric than do the immediately underlying backlap shellbeds. Finally, inferred toplap shellbeds occur within the upper parts of RST or FRST, where stratal convergence results from sand bypass in high-energy shoreface environments. Skeletal material is preserved in the sediments in which the organisms lived, and though assemblages are reworked, disarticulated, and fragmented, they are neither significantly rounded nor display the other typical lag features that characterize basal onlap shellbeds. In some sequences, onlap and backlap shellbeds are superimposed to form a compound shellbed.

Quaternary Sequences

How Are They Special?

It is a prime feature of Quaternary sediments from active-margin basins that they exhibit a much greater resolution of sequence architecture than that depicted by the generic third-order SSM. For example:

1. Figure 2 depicts an SSM for the Quaternary, based mainly on outcrop studies of uplifted basins adjacent to active plate margins; and

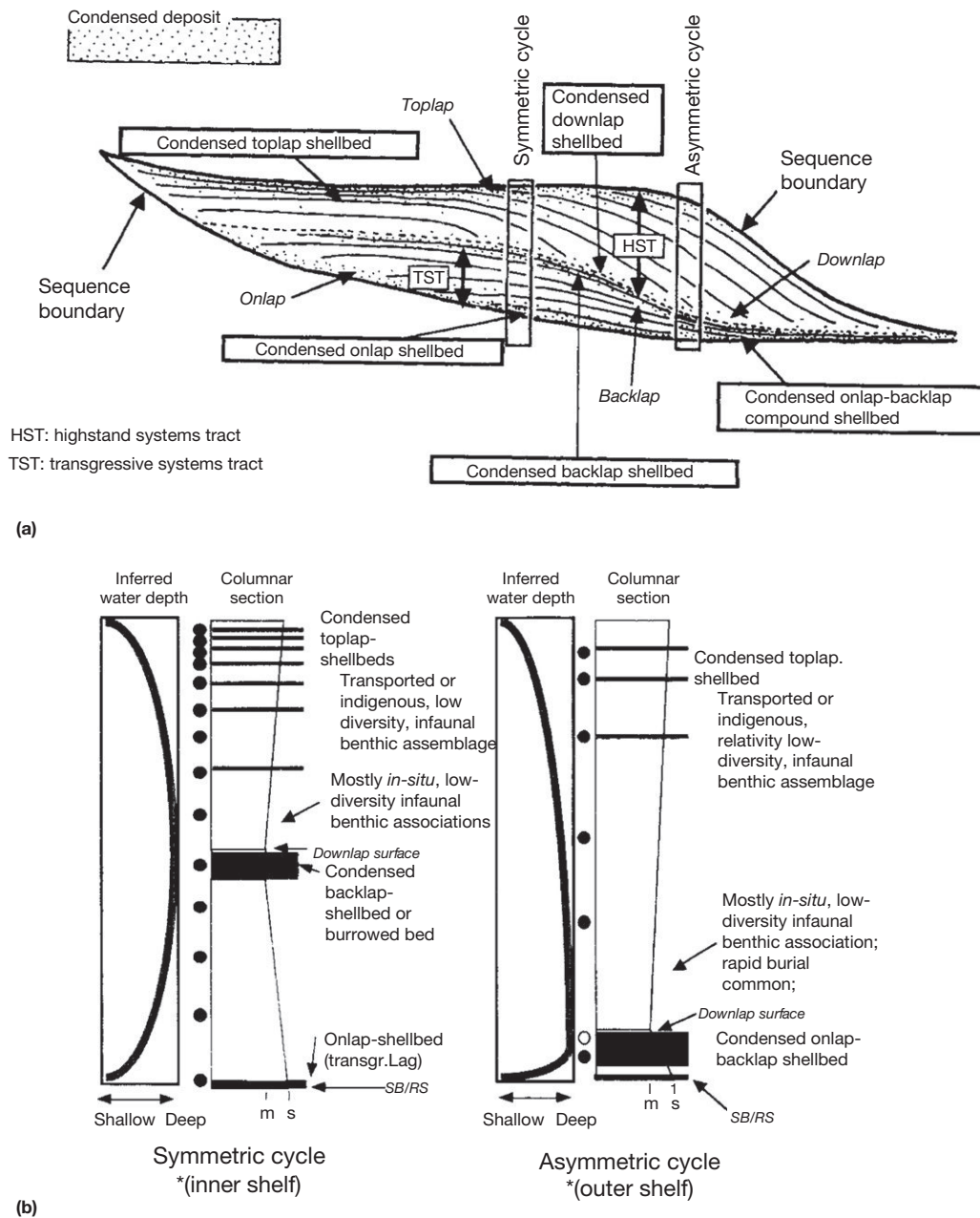


Figure 3 The association between shellbed type and sequence architecture in Plio-Pleistocene sequences. Reproduced from Kondo Y, et al. (1998) *The relationship between shellbed type and sequence architecture: Examples from Japan and New Zealand. Sedimentary Geology* 122: 109–127.

(a) Distribution of shellbed types within an idealized sequence, showing their relationship to stratal convergence and systems tract-bounding surfaces.

(b) Vertical facies succession within a sequence showing the stratigraphic context of shellbed types.

2. **Figure 4** is an example of a measured section from the Canoa Basin, Ecuador, labeled to show the component facies, sequences, systems tracts, and surfaces.

A second important feature of Quaternary sequences is that subaerial sequence boundaries are not usually preserved but are instead superposed by the TSE cut by the passage of a transgressing marine shoreface. A TSE may cause erosion of up to several tens of meters into the top of the underlying sequence, which removes all traces of original sequence boundaries almost basin wide (Abbott, 1992; Abbott and

Carter, 1994) and results in the preservation of top-truncated sequences (Castlecliff motif of Saul et al., 1998). Deep erosion during transgression also leads Quaternary sequences deposited in inner-shelf and basin margin settings to possess proportionally thick TSTs.

A third difference between the Quaternary SSM and the generic third-order SSM is the common presence of strata deposited during the falling limb of the relative sea-level curve. Regressive packages of strata deposited during falling sea level form when the supply of sediment is large enough

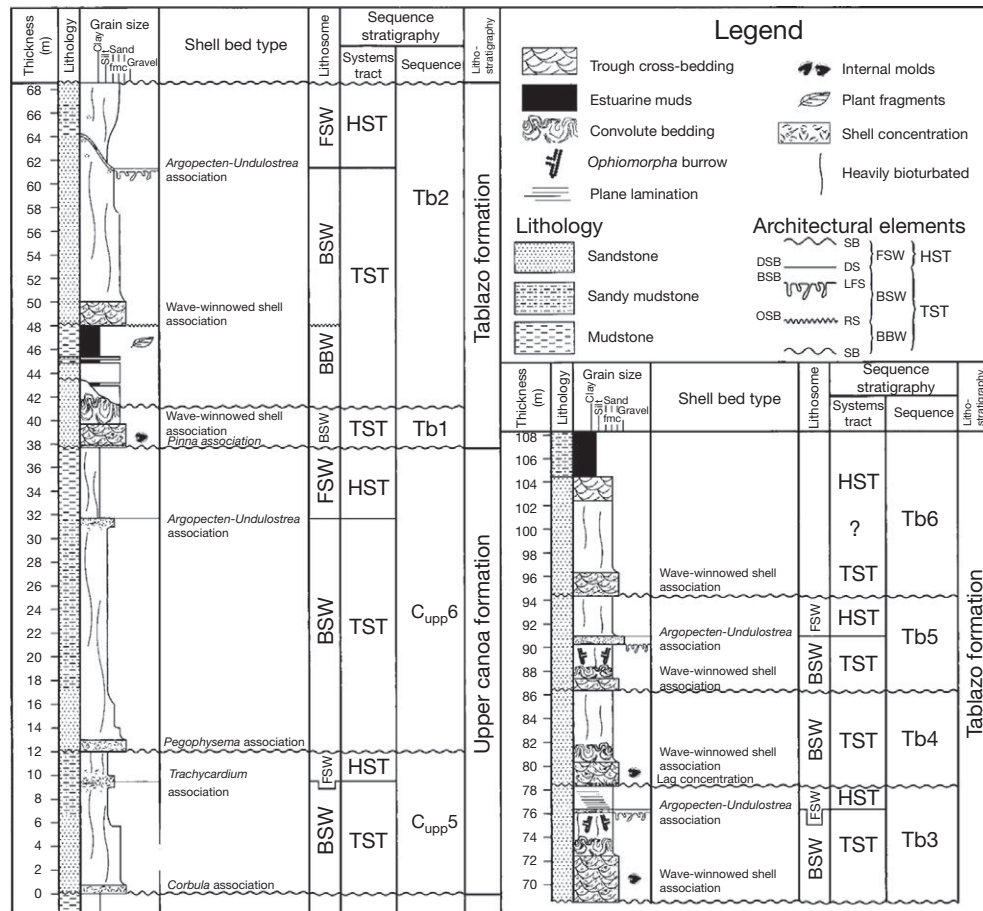


Figure 4 Vertical succession of facies (including shellbeds), sequences, and systems tracts within the Canoa formation, Canoa Basin, central Ecuador. Reproduced from Di Celma C, Ragaini L, Cantalamessa G, and Landini W (2005) Basin physiography and tectonic influence on sequence architecture and stacking pattern: Pleistocene succession of the Canoa Basin (central Ecuador). *Geological Society of America Bulletin* 117: 1226–1241.

to form an offlapping body of sediment that builds into the basin as the shoreline regresses. As described earlier, such sediment bodies are either bounded below by an RSE and termed FRST, or else grade up from underlying HST sediments to comprise a simple RST.

Fourth, very high sedimentation rates and rapid shellbed formation accompanied active-margin basin-filling during Quaternary eustatic cycles. Individual 40 ka-long late Pliocene sequences attain 120 m or more in thickness, with HST–RST alone over 100 m thick (Haywick et al., 1992; Naish and Kamp, 1997a) (Figure 5). Therefore, average sedimentation rates at times reached 5 m ka^{-1} or more, and MCS up to $\sim 1 \text{ m}$ in thickness, despite being ‘condensed’ with respect to their enclosing TST–HST sediment, nonetheless took only a few thousand years to accumulate (Abbott and Carter, 1994).

Finally, and fifth, it is only in Quaternary successions that the complete three-dimensional architecture of continental margin sequences can be studied within a known sea-level context. Outcrop studies can be accomplished not only on uplifted Plio-Pleistocene shelf (Abbott and Carter, 1994) and slope (Ito, 1992) successions, but also on the important shore-face-coastal cliff-coastal plain interface that is preserved in the treads of flights of uplifted marine terraces back to mid-Pleistocene in age (Chappell, 1974; Pillans, 1983).

Sequence Motifs

The vertical stratigraphic succession of facies, faunas, and surfaces within a sequence varies according to its location along a hypothetical shore-normal profile. Particular types of stratigraphic succession can be differentiated as sequence motifs. As discussed by Saul et al. (1999), sequence motifs are arbitrary end-members in a continuous spectrum of possible sequence architectures. Furthermore, the interplay of subsidence and sediment supply, and therefore shore-normal variation in sequence motifs, varies in detail both within and between basins. The generic motifs depicted in Figure 2(c) are drawn from earlier studies of geographically named motifs, generalized here in order to illustrate typical cross-basin variation in Quaternary sequence architecture and facies composition (Figure 6).

Basin margin motif

The edges of active-margin basins often record a history of progressive Quaternary uplift. Wave-cut terraces are formed at interglacial highstands, and flights of terraces formed at successive highstands extend from the modern highstand shoreline into the uplifting hinterland (Pillans, 1983). Terrace treads comprise sediments that represent the landward margins

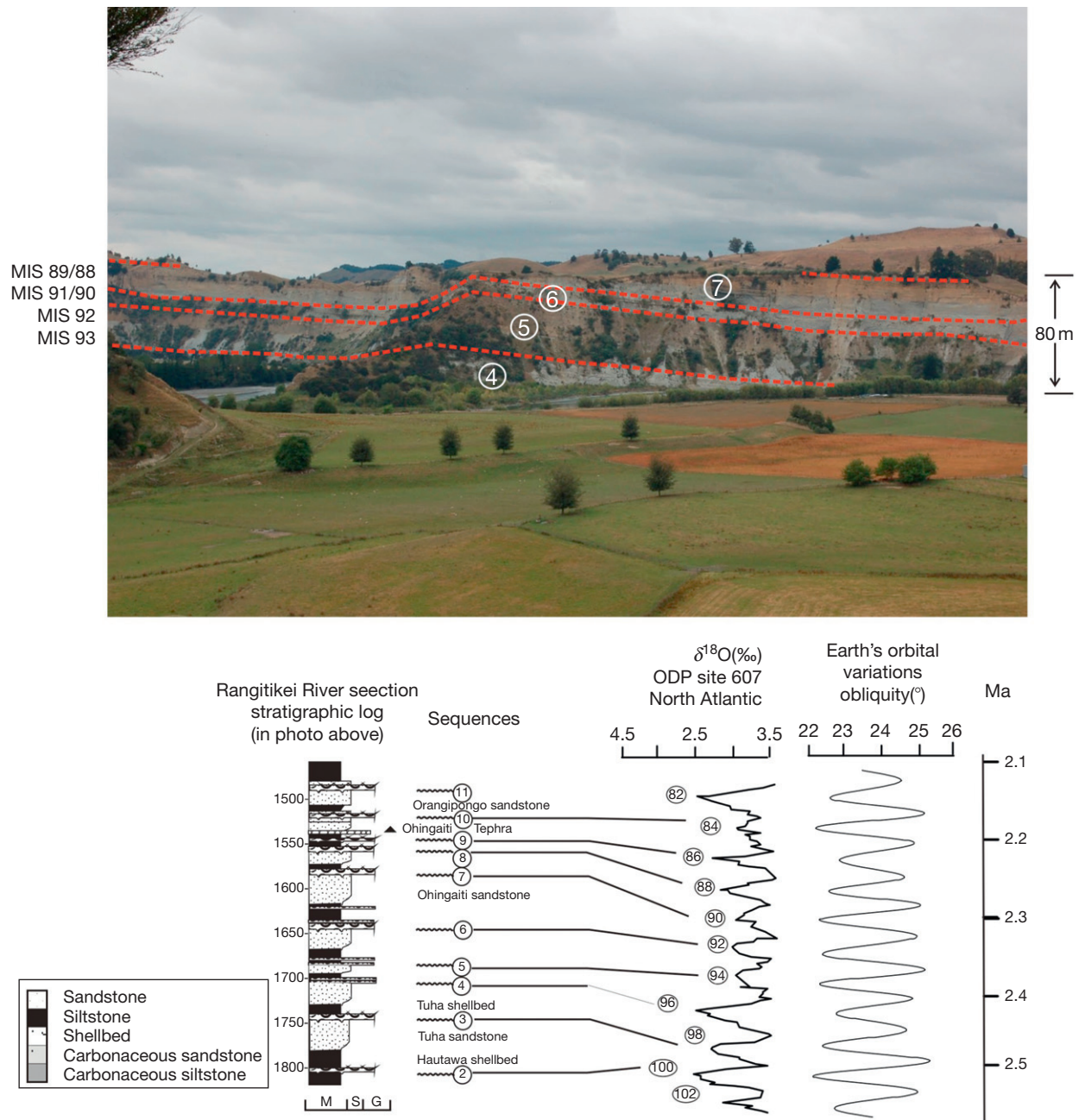


Figure 5 Photo of outcrop exposure of Early Quaternary (2.5–2.0 Ma), 40-ka-duration, shallow-marine sequences exposed in cliffs of the Rangitikei Valley, Wanganui Basin. Sequence boundaries (TSEs) are marked by red dashed line. Individual 40-ka sequences are up to 45 m thick. The stratigraphic column shows these sequences correlated with deep sea oxygen isotope record. Modified from Naish TR and Kamp PJJ (1997) Sequence stratigraphy of sixth-order (41 ka) Pliocene–Pleistocene cyclothem, Wanganui Basin, New Zealand: A case for the regressive systems tract. *Geological Society of America Bulletin* 109(8): 978–999.

of depositional sequences, and sequences formed in this paleogeographic setting are intimately related to the geomorphic expression of the terraces (see [Morphostratigraphy/Allostratigraphy](#)).

For most of the sea-level cycle, the basin margin is subaerially exposed. Sedimentation in this setting consists of nonmarine facies, including substantial paleosol development, together with shallow marine facies deposited at and near each interglacial sea-level peak. The sediments can be partitioned into

transgressive and HSTs. In this setting, sequence-bounding RS may terminate at paleocliffs that represent the position of the highstand paleoshoreline.

Inner-shelf sequence motif

The inner-shelf setting is inundated from mid- to late-rise through to highstand. In contrast to the basin margin motif, paleodepths in this setting may reach several tens of meters, so a proportion of sequences in this paleogeographic setting

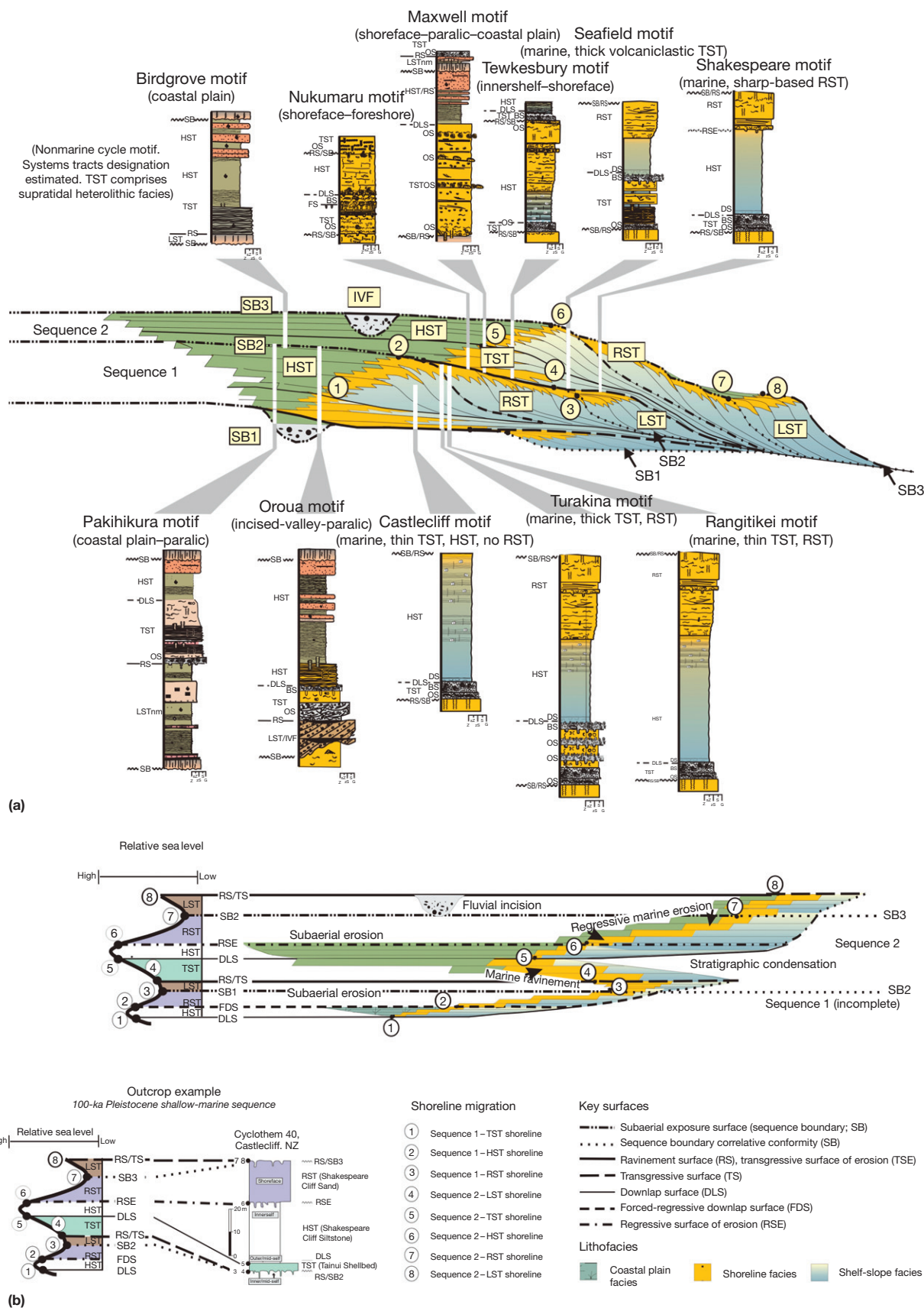


Figure 6 (a) Conceptual sequence stratigraphic and (b) chronostratigraphic models for orbital scale Quaternary sequences based on Wanganui Basin. The stratigraphic architecture timing of development of systems tracts and key surfaces are shown with respect to a relative cycle of sea level. Outcrop motifs of Wanganui sequences are illustrated in the context of a 2D coastal plain and outer shelf progradational setting. Reproduced from [Naish T, Field B, Zhu H, et al. \(2005\)](#) Integrated outcrop, drill core, borehole and seismic stratigraphic architecture of a cyclothemic, shallow-marine depositional system, Wanganui Basin, New Zealand, *Journal of the Royal Society of New Zealand* 35: 91–122.

consists of facies deposited at shelf depths. Thick TSTs begin with an onlap shellbed that deepens upward. Because this motif is proximal to the basin margin, transgressive sedimentation switches rapidly to highstand deposition as relative sea level approaches its peak. Therefore, backlap shellbeds may not be well developed at the transgressive-HST boundary. Highstand deposits are mud-rich, but may become sand-rich toward the top. During the ensuing sea-level fall, there is an accompanying abrupt basinward shift in deposition and wave base is lowered to the point where it scours the shelf. This basinward passage of wave base may generate a RSE, over which lies a body of shoreface sand assigned to the forced RST. As for the basin margin, the inner-shelf paleogeographic position may also be affected by tectonic uplift. Thus, inner-shelf sequences from the Peradriatic basin (Cantalamesa and Di Celma, 2004) are stacked in a basinward stepwise manner, describing a low-order tectonically forced regression.

Mid-shelf sequence motif

As for the proximal motifs, sequences in the mid-shelf paleogeographic position also commence with a relatively thick TST with a basal onlap shellbed. In contrast to the inner-shelf motif, the uppermost TST in the mid-shelf position is more likely to be isolated from terrigenous sediment, and the time until downlap by the highstand is likely to be greater. Thus, here and also in the more distal motifs, shell material accumulates *in situ* to form well-developed backlap shellbeds.

Highstand deposition on the mid-shelf is heralded by downlapping shelf mud that 'quenches' shell accumulation, resulting in an abrupt yet conformable DLS between shell-rich and relatively shell-poor facies. Shelly clumps and bands at the base of some highstand successions represent downlap shellbeds, but these are not as well developed as the underlying backlap shellbeds. As for the inner-shelf motif, highstand shelf mud passes into shoreline sand deposited during falling sea level. However, on the mid-shelf, the rate of subsidence is higher, so wave base is not lowered to the extent where it cuts a RSE. Instead, the transition is conformable and the overlying sands are attributed to the RST rather than the forced RST.

Outer-shelf sequence motif

This sequence motif results from deposition on the outer shelf, between early relative sea-level rise through to the late part of sea-level fall. The outer-shelf motif is highly asymmetrical in the sense that sand-rich transgressive facies are not well developed compared to the thick, sand-rich RST. The poor development of the TST may result from rapid transgression over a low gradient coastal plain, which then becomes isolated from the supply of terrigenous sediment. The resulting TSTs are thin and particularly rich in skeletal material. The basal onlap shellbed may be directly overlain by a backlap shellbed to form a compound shellbed. As for the mid-shelf motif, the DLS forming the boundary between the transgressive and HSTs is an abrupt shell-rich to relatively shell-poor boundary. The upper part of the motif is a gradationally based RST of shoreline sand.

Slope sequence motif

Quaternary sequences of this motif are less well known from outcrop, so this motif is based on the Plio-Pleistocene of the Hawke's Bay basin described by Haywick et al. (1992). This setting lies basinward of the lowstand shoreline, so is

inundated through the entire relative sea-level cycle. Erosional sequence-bounding surfaces associated with subaerial exposure or marine ravinement are often not present. Instead, sequences in this paleogeographic position are bounded by correlative conformities. Another feature of this motif is that it comprises exclusively offshore facies because the slope paleogeographic setting is submerged to depths of over 100 m during most of the relative sea-level cycle. Only the basinward-tapering edges of the transgressive and HSTs are represented and, because of the pronounced convergence of stratal surfaces in the distal TST, backlap shellbeds may be well developed. Facies successions shallow upward as the regressing shoreline reaches lowstand. In the case of the Hawke's Bay basin, this coincided with a reorganization of terrigenous sediment transport paths, such that LSTs comprise coquina limestone.

Case Study 1: Characterizing, 5th- and 6th-Order Quaternary Shallow-Marine Sequences from Outcrop, Borehole, Drillcore, And Seismic Data from Wanganui Basin, New Zealand

Continental margins are relatively rare that afford an opportunity to study the stratigraphic architecture of shallow-marine depositional systems on spatial scales ranging from core and borehole (millimetres/centimetres) to outcrop and seismic (tens of meters), during periods of known sea-level oscillations. Wanganui Basin, New Zealand is one of a small number of Pliocene-Pleistocene basins worldwide (Browne & Naish 2003; Kitamura et al., 2000) in which sedimentation evidently kept pace with subsidence through much of the basin history, resulting in a 5-km-thick record of predominantly shelf and shallow-water sediment. Gentle upwarping of the eastern margin of the basin has produced spectacular exposures, along coastal cliffs and inland (Figure 5) within the heavily dissected and uplifted landscape, through as many as 44 depositional sequences deposited during the last 2.5 Ma (Carter & Naish, 1998; Naish et al., 1998; Saul et al., 1999).

The Wanganui stratigraphic succession comprises a range of siliciclastic and carbonate sediments deposited in shelf, shoreline, and coastal plain depositional environments during orbitally controlled sea-level fluctuations of 41 ka- and 100 ka-duration. Over the last 10 years, this margin has gained a reputation as one of the world's outstanding outcrop examples of a Pliocene-Pleistocene shallow-marine, clastic depositional system (e.g., summary papers: Abbott and Carter 1994, Abbott 1997; Naish & Kamp 1997a; Naish et al., 1998; Saul et al., 1999). Outcrop analogues of this quality are rare, as most Quaternary shelf margins underlie flooded continental shelves (e.g., Gulf of Mexico).

Late Pliocene to mid-Pleistocene (ca. 2.1–0.4 Ma) strata exposed in the coastal cliff sections northwest of Wanganui City, North Island New Zealand, comprise 25, 6th(41 ka)- and 5th(100 ka)-order, shallow-marine to marginal marine stratigraphic sequences, deposited during global glacio-eustatic sea-level cycles corresponding to Marine Isotope Stages (MIS) 78–10. Naish et al. (2005) characterized the sequences using: (1) a series of drill cores sited above and behind the coastal outcrops, which recovered a composite record of ca. 450 m, (2) a new high-resolution multichannel seismic reflection profile acquired along the beach adjacent to the coastal cliffs, and (3) downhole digital logs from the boreholes (Figures 7 and 8). They integrated the outcrop and subsurface data sets to

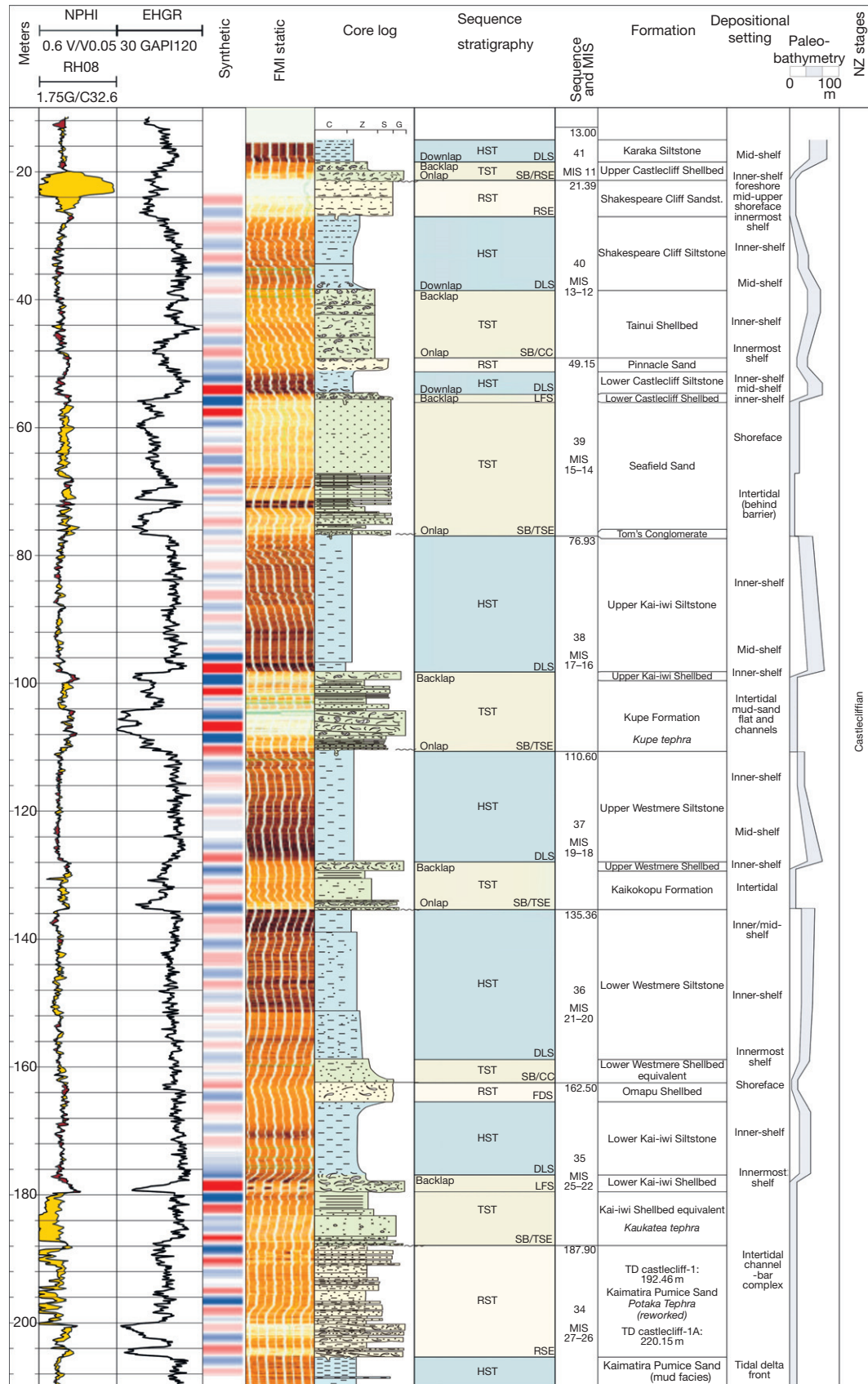


Figure 7 Summary composite log of the Castlecliff-1/1A drillcore, Wanganui Basin, and borehole shows neutron porosity (NPHI), density (RH08), and natural gamma (EHGR) borehole electric logs, synthetic seismic well-tie, lithologic log, sequence stratigraphic interpretation, oxygen isotope stages, lithostratigraphy, and environmental and paleobathymetric interpretations. Reproduced from Naish T, Field B, Zhu H, et al. (2005) Integrated outcrop, drill core, borehole and seismic stratigraphic architecture of a cyclothem, shallow-marine depositional system, Wanganui Basin, New Zealand, *Journal of the Royal Society of New Zealand* 35: 91–122.

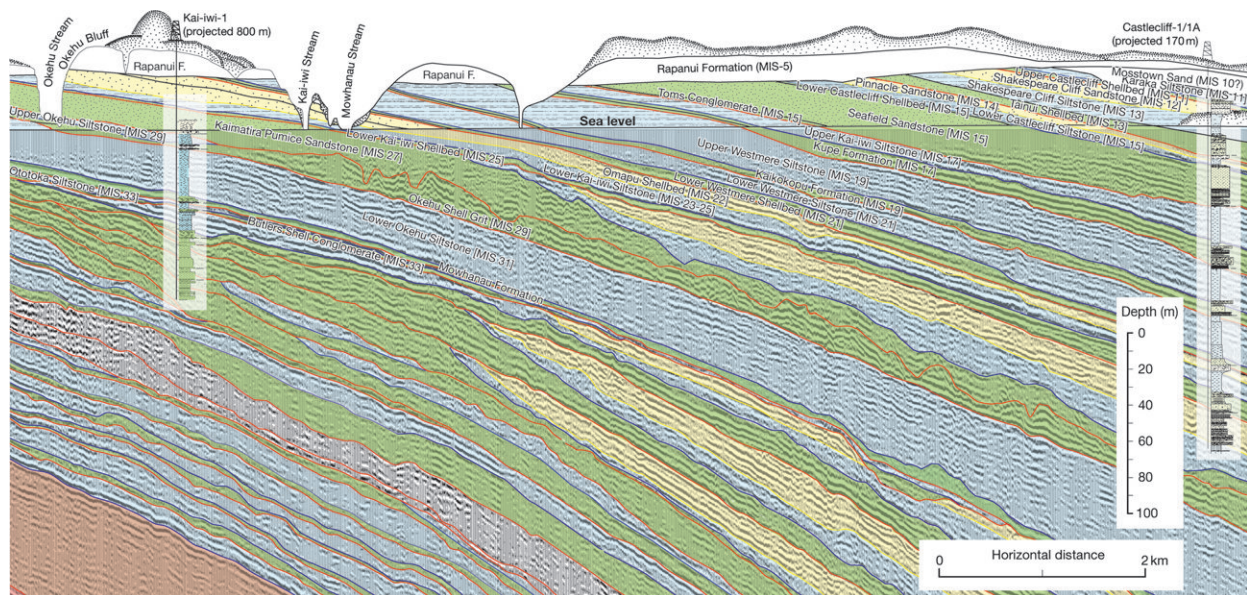


Figure 8 Pleistocene, shallow marine sequences exposed in the coastal Castlecliff section, Wanganui Basin, New Zealand are correlated with drill core logs from ‘behind-outcrop’ boreholes and high-resolution seismic data from along the beach at the base of the coastal cliffs (after Naish et al. (2005); reproduced from Graham (2008)).

produce a high-resolution model of the stratigraphic signatures and 2-D architecture of a cyclical, shallow-marine depositional system.

Case Study 2: Outcrop Examples of Orbitally Influenced Pleistocene Shelf Sequences Preserving Forced Regressive Deposits, Croton Basin, Southern Italy

On a global scale, the Croton Basin preserves one of the best-developed and most complete Pleistocene marine records available in outcrop, as important as those in California, New Zealand, and Japan. The preservation of high-resolution Quaternary sequences occurred through the interaction between high-amplitude global sea-level fluctuations within an extension rift basin. The 550 m-thick composite succession contains 13 orbital-scale sedimentary cycles correlated with Marine Isotope Stages 33 to 7 during a long-term shallowing from upper slope-outer shelf to marginal marine-coastal plan environments between ~1.2–0.25 Ma (Massari et al., 2002). An unconformity of tectonic origin, which is marked by an abrupt shallowing and the first evidence of fluvial incision, punctuates the succession at the base of Sequence 9 removing the record corresponding to MIS 16–12 (~0.65–0.45 Ma) (Figure 9). Correlation of the sedimentary cycles with the deep-sea oxygen isotope record is constrained by an integrated, tephro-, bio-, magneto stratigraphy (Rio et al., 1996; Massari et al., 2002). The cyclothemic nature of the record is characterized by a stack of simple or composite, seaward-prograding, sand-dominated tongues deposited during regressions and intervening aggradational mud-dominated deposits related to transgressive-deepening and highstand episodes. High rates of basal subsidence and high rates of sediment supply have allowed exceptional outcrop examples of forced regressive systems tracts to be preserved in some sequences (Massari et al., 1999) (Figure 10).

Astronomical Calibration

Calibration of orbitally forced oxygen isotope records from the deep sea with astronomical solutions for orbital precession, obliquity, and eccentricity have led to the development of an oxygen isotope numerical timescale extending back to the late Miocene (Lourens et al., 1996; Shackleton et al., 1990). Peaks and troughs in oxygen isotope records are consecutively numbered as isotope stages. For Quaternary shelf successions, the preserved sequences generally correspond to interglacial (odd-numbered) stages, while sequence-bounding unconformities correspond to glacial (even-numbered) stages.

Extended Quaternary successions in the Boso Peninsula, Japan, and the Wanganui Basin, New Zealand, have been correlated in detail with the sea-level history contained in the oceanic oxygen isotope scale. For example, Naish et al. (1999) and Pillans et al. (2005) used geomagnetic reversals and numerous dated tephra horizons to calibrate 43 Wanganui sequences with oxygen isotope stages 1–100 (Figure 11), and Pickering et al. (1999) correlated Boso sequences with isotope stages 15–35. Apart from their intrinsic interest, such calibrations enable absolute dates to be estimated for the boundaries of local Stages, for biostratigraphic events such as first and last appearances, and for many otherwise undated stratigraphic units (cf. Carter 2005).

Summary

1. Glacioeustasy driven by orbital precession (20 ka), obliquity (41 ka), and eccentricity (100 ka) has imposed sedimentary cyclicity on the fill of Quaternary continental margin basins. Such cycles are represented as fifth- to seventh-order sequences by sediments within recently uplifted active-margin basins, each sequence being attributable to deposition during one cycle of sea-level change.

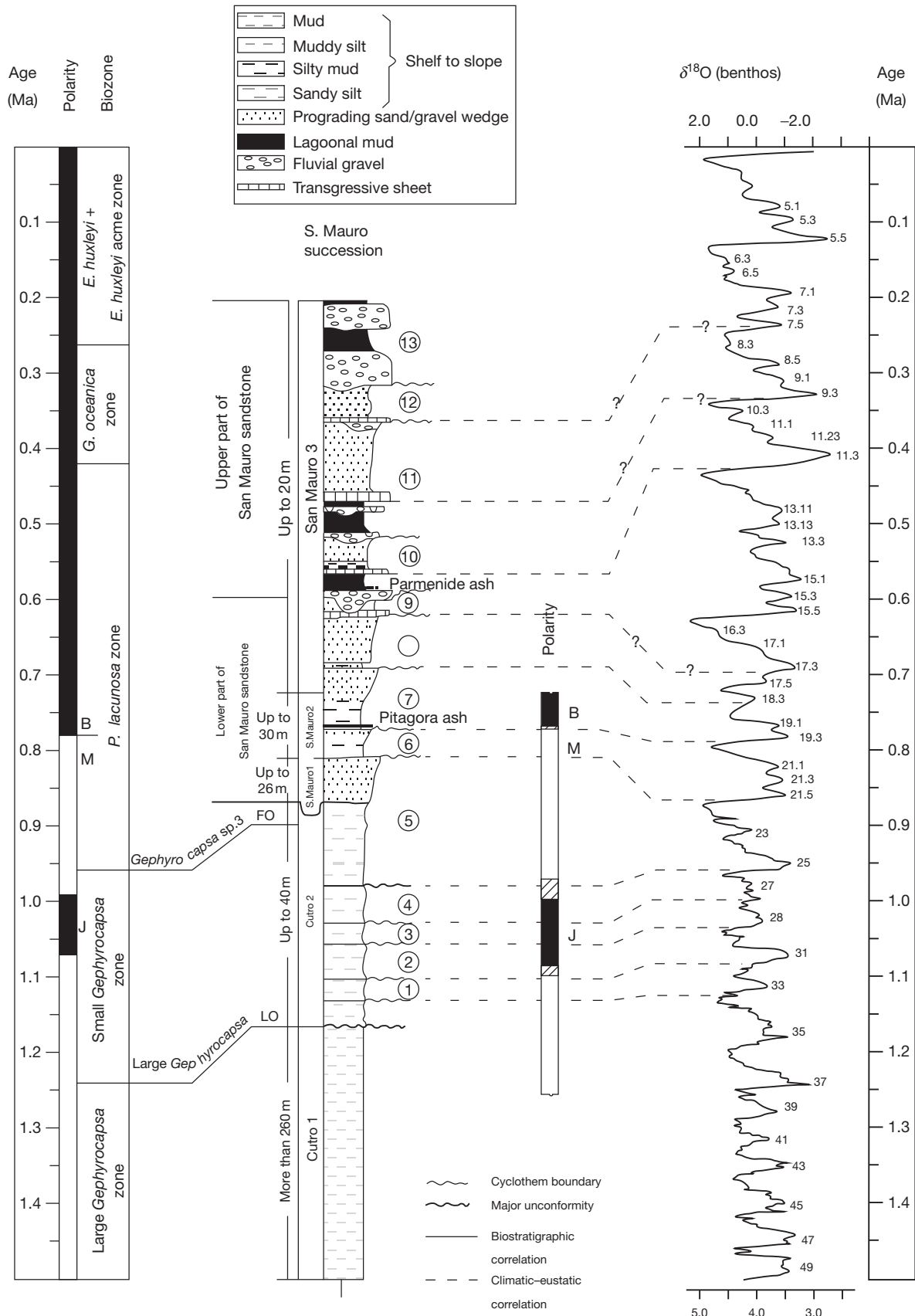


Figure 9 Correlation of the Pleistocene San Mauro shallow-marine cyclothem succession, Croton Basin, with the marine oxygen isotope stratigraphy. Reproduced from [Massari F, Rio D, Sagveti M, et al. \(2002\)](#) Interplay between tectonics and glacio-eustasy: Pleistocene succession of the Croton basin, Calabria (southern Italy). *GSA Bulletin* 114: 1183–1209.

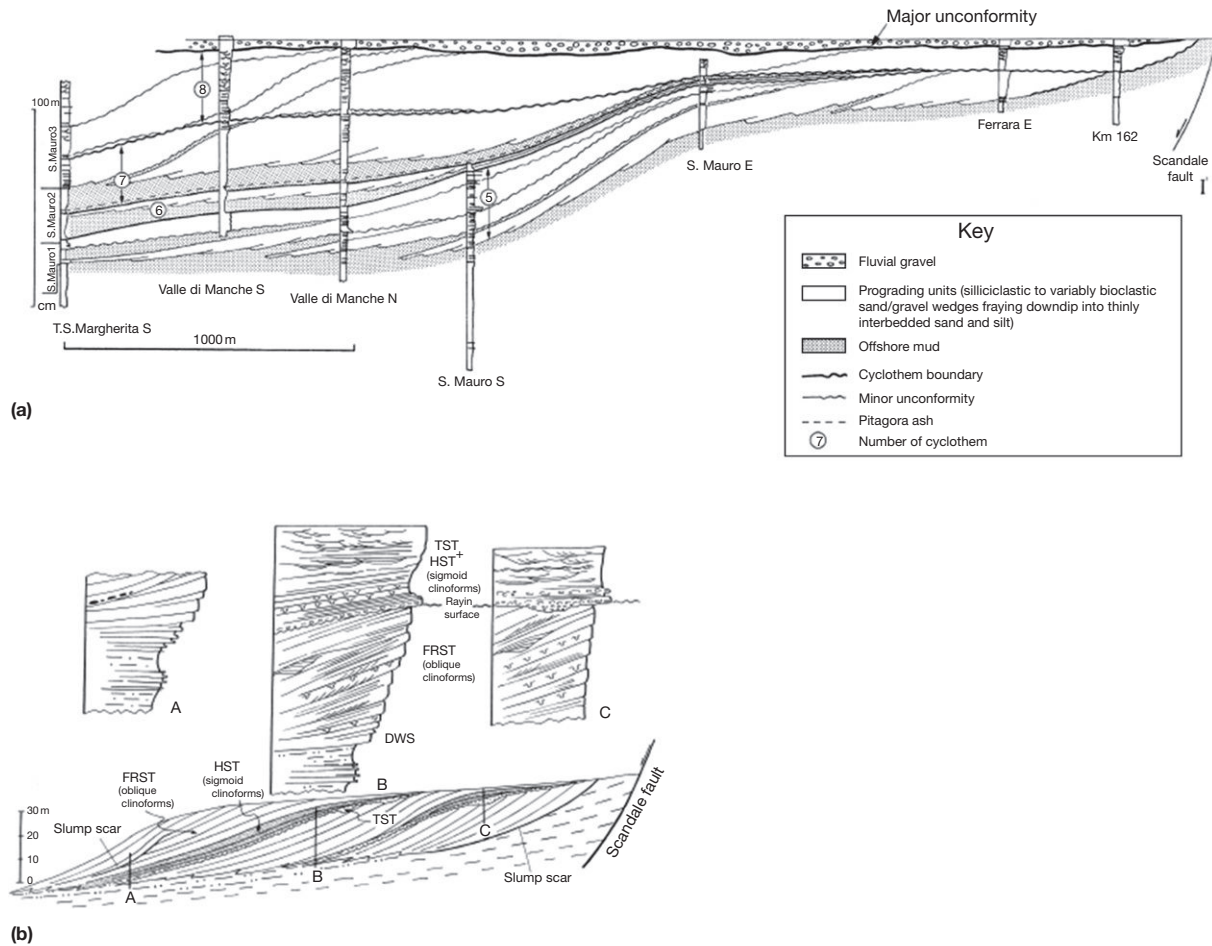
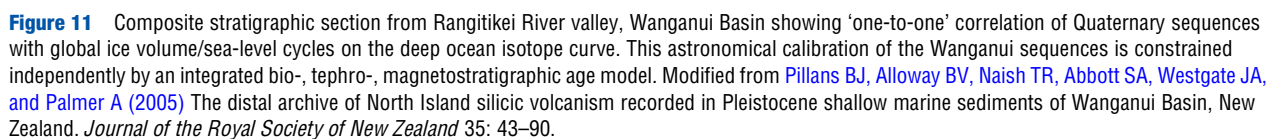


Figure 10 (a) Schematic, dip-oriented cross-section linking cyclothem sections within the Croton Basin shows the 2-D sequence stratigraphic architecture of orbitally influenced Pleistocene shallow-marine sequences. Note the preservation of thick, seaward prograding, shore-connected sediment wedges interpreted as forced regressive deposits. (b) A more detailed example of the sequence architecture of Croton Basin sequences showing the downward shift of facies during forced regression, marked by an RSE or DWS (downward shift surface) beneath the FRST.

- Analysis of sedimentary sequences is undertaken using standard sedimentary facies analysis, supported by paleoecological and taphonomic analysis of fossil faunas. Stratal architecture within sequences is inferred from the succession of facies (grouped as systems tracts) and stratigraphic surfaces that occur within a sequence in the following invariant order: lower SB, LST, TSE, TST, LFS, MCS, DLS, HST, regressive or forced RST, and upper SB.
- The vertical representation of facies and systems tracts within a Quaternary sequence varies along a hypothetical cross-shelf profile. Five sequence motifs are recognized, which characterize the spectrum of sedimentary variation that exists between the inner basin margin and the offshore basin slope.
- Calibration of lengthy Quaternary sequence successions from Japan (Boso Peninsula) and New Zealand (Wanganui Basin) with the orbitally tuned oxygen isotope timescale has been achieved using geomagnetic reversals and dated tephra. Sequences deposited in shelf settings correspond to odd-numbered (interglacial) stages on the oxygen isotope timescale, while the even-numbered (glacial) stages are represented by sequence-bounding unconformities. Such calibration allows numerical ages to be assigned to otherwise undated stratigraphic entities such as local lithostratigraphic units, relative time scales, and fossil first and last appearances.
- Studies of Quaternary sequence stratigraphy have demonstrated:
 - a much greater resolution of sequence architecture than is depicted by generic sequence stratigraphic models;
 - that subaerial sequence boundaries are rarely preserved intact, but instead are superposed by a TSE cut by shoreline passage during the rising sea-level stage of the next eustatic cycle;
 - the common preservation of thick, sandy RST strata during the falling stage of a eustatic cycle;
 - the occurrence of highstand and RST sedimentation rates as high as 5 m/ka, and the formation of shellbeds in as little as a few ka, in sequences from uplifted active-margin basins; and
 - the availability for study of the three-dimensional architecture of complete continental margin sequences in the light of the independent proxy sea-level history that is represented by the oceanic oxygen isotope record.



See also: Glaciation, Causes: Astronomical Theory of Paleoclimates. Quaternary Stratigraphy: Lithostratigraphy.

References

- Abbott ST (1992) The mid-Pleistocene Waiomio Shellbed (c. 500–600 ky), Wanganui Basin, New Zealand. *Alcheringa* 16: 171–180.
- Abbott ST (1997) Foraminiferal paleobathymetry and mid-cycle architecture of mid-Pleistocene depositional sequences, Wanganui Basin, New Zealand. *Palaios* 12: 267–281.
- Abbott ST and Carter RM (1994) The sequence architecture of mid-Pleistocene (0.35–0.95 Ma) cyclothems from New Zealand: Facies development during a period of orbital control on sea-level cyclicity. In: de Boer PL and Smith DG (eds.) *Orbital Forcing and Cyclic Sequences*, vol. 19, pp. 367–394. Oxford: International Association of Sedimentologists Special Publication.
- Abbott ST and Carter RM (1997) Macrofossil associations from mid-Pleistocene cyclothems, Castlecliff section, New Zealand: Implications from sequence stratigraphy. *Palaios* 12(2): 188–210.
- Abbott ST, Naish TR, Carter RM, and Pillans BJ (2005) Sequence stratigraphy of the Nukumaruan stratotype, Nukumarua and Ototoke beaches, Wanganui Basin. *Journal of the Royal Society of New Zealand* 35: 123–150.
- Boyd R, Suter J, and Penland S (1988) Implications of modern sedimentary environments for sequence stratigraphy. In: James DP and Leckie DA (eds.) *Stratigraphy, Sedimentology: Surface and Subsurface*, pp. 33–36. Calgary, Alberta: Canadian Society of Petroleum Geologists Memoir 15.
- Brown LF Jr. and Fisher WL (1977) Seismic-stratigraphic interpretation of depositional systems: Examples from Brazilian rift and pull-apart basins. In: Payton CE (ed.) *Seismic Stratigraphy – Applications to Hydrocarbon Exploration*, pp. 213–248. Tulsa, OK: American Association of Petroleum Geologists Memoir 26.
- Browne GH and Naish TR (2003) Facies development and sequence architecture of a late Quaternary fluvial-marine transition, Canterbury Plains and shelf, New Zealand; implications for forced regressive deposits. *Sedimentary Geology* 158(1–2): 57–86.
- Cantalamesa G and Di Celma C (2004) Sequence response to syndepositional regional uplift: Insights from high-resolution sequence stratigraphy of late Early Pleistocene strata, Periadriatic Basin, central Italy. *Sedimentary Geology* 164(3–4): 283–309.
- Cantalamesa G, Di Celma C, and Ragaini L (2005) Sequence stratigraphy of the Punta Ballena member of the Jama Formation (Early Pleistocene, Ecuador): Insights from integrated sedimentologic, taphonomic and paleoecologic analysis of molluscan shell concentrations. *Palaeogeography Palaeoclimatology Palaeoecology* 216: 1–25.
- Carter RM (2005) A New Zealand climatic template back to c. 3.9 Ma: ODP Site 1119, Canterbury Bight, south-west Pacific Ocean, and its relationship to onland successions. *Journal of the Royal Society of New Zealand* 35: 9–42.
- Carter RM, Abbott ST, Fulthorpe CS, Haywick DW, and Henderson RA (1991) Application of global sea-level and sequence stratigraphic models in southern hemisphere Neogene basins from New Zealand. In: MacDonald DIM (ed.) *Sedimentation, Tectonics and Eustasy – Sea-level Changes at Active Plate Margins*, vol. 12, pp. 41–65. Oxford: International Association of Sedimentologists Special Publication.
- Carter RM, Abbott ST, Graham IJ, Naish TR, and Gammon PR (2002) The Middle Pleistocene Merced-2 and -3 sequences from Ocean Beach, San Francisco. *Sedimentary Geology* 153: 23–41.
- Carter RM, Fulthorpe CS, and Naish TR (1998) Sequence concepts at seismic and outcrop scale: The distinction between physical and conceptual stratigraphic surfaces. *Sedimentary Geology* 122: 165–179.
- Carter RM and Naish TR (1998) A review of Wanganui Basin, New Zealand: A global reference section for shallow marine, Plio-Pleistocene (2.5–0 Ma) cyclostratigraphy. *Sedimentary Geology* 122: 37–52.
- Chappell J (1974) Geology of coral terraces, Huon Peninsula, New Guinea: A study of Quaternary tectonic movements and sea-level changes. *Geological Society of America Bulletin* 85: 553–570.
- Chappell J and Shackleton NJ (1986) Oxygen isotopes and sea level. *Nature* 324: 137–140.
- Christie-Blick N (1991) Onlap, offlap, and the origin of unconformity-bounded depositional sequences. *Marine Geology* 97: 35–56.
- Clifton HE, Hunter RE, and Gardner JV (1988) Analysis of eustatic, tectonic, and sedimentologic influences on transgressive and regressive cycles in the Upper Cenozoic Merced Formation, San Francisco, California. In: Kleinspehn KL and Paola C (eds.) *New Perspectives in Basin Analysis*, pp. 109–128. New York: Springer-Verlag.
- Di Celma C, Ragaini L, Cantalamesa G, and Landini W (2005) Basin physiography and tectonic influence on sequence architecture and stacking pattern: Pleistocene succession of the Canoa Basin (central Ecuador). *Geological Society of America Bulletin* 117: 1226–1241.
- Embry AF (1990) A tectonic origin for third-order depositional sequences in extensional basins – Implications for basin modelling. In: Cross TA (ed.) *Quantitative Dynamic Stratigraphy*, pp. 491–501. Englewood Cliffs, NJ: Prentice Hall.
- Fleming CA (1953) The geology of the Wanganui subdivision. *Bulletin of the NZ Geological Survey* 52: 1–362.
- Fulthorpe CS (1991) Geological controls on seismic sequence resolution. *Geology* 19: 61–65.
- Graham IJ (ed.) (2008) A continent on the move: New Zealand geoscience into the 21st century. *The Geological Society of New Zealand in association with GNS Science*. Manaaki Whenua Press.
- Haq BU, Hardenbol J, and Vail PR (1987) Chronology of fluctuating sea levels since the Triassic. *Science* 235: 1156–1167.
- Hart BS and Long BF (1996) Forced regressions and low-stand deltas: Holocene Canadian examples. *Journal of Sedimentary Research* 66: 820–829.
- Haywick DWH, Carter RM, and Henderson RA (1992) Sedimentology of 40,000 year Milankovitch controlled cyclothems from central Hawke's Bay, New Zealand. *Sedimentology* 39: 675–696.
- Haywick DWH and Henderson RA (1992) Foraminiferal paleobathymetry of Plio-Pleistocene cyclothem sequences, New Zealand. *Palios* 6(6): 586–599.
- Hunt D and Tucker ME (1992) Stranded parasequences and the forced regressive wedge systems tract: Deposition during base-level fall. *Sedimentary Geology* 81: 1–9.
- Ito M (1992) High-frequency depositional sequences of the upper part of the Kazusa Group, a Middle Pleistocene forearc basin fill in Boso Peninsula, Japan. *Sedimentary Geology* 76: 155–175.
- Ito M (1998) Submarine fan sequences of the lower Kazusa Group, a Plio-Pleistocene forearc basin fill in the Boso Peninsula, Japan. *Sedimentary Geology* 122: 69–95.
- Kidwell SM (1991a) Condensed deposits in siliciclastic sequences: Expected and observed features. In: Einsele G, Ricken W, and Seilacher A (eds.) *Cycles and Events in Stratigraphy*, pp. 683–695. Heidelberg: Springer-Verlag.
- Kidwell SM (1991b) The stratigraphy of shell concentrations. In: Briggs DEG and Allison PA (eds.) *Taphonomy – Releasing Information from the Fossil Record*, pp. 211–290. New York: Plenum Press.
- Kitamura A, Kondo Y, Sakai H, and Horii M (1994) 41,000 year orbital obliquity expressed as cyclic changes in lithofacies and molluscan content, Early Pleistocene Omma Formation, central Japan. *Palaeogeography, Palaeoclimatology, Palaeoecology* 112: 345–361.
- Kitamura A, Matsui H, and Oda M (2000) Constraints on the timing of systems tract development with respect to sixth-order (41 ka) sea-level changes; an example from the Pleistocene Omma Formation, Sea of Japan. *Sedimentary Geology* 131(1–2): 67–76.
- Kondo Y, et al. (1998) The relationship between shellbed type and sequence architecture: Examples from Japan and New Zealand. *Sedimentary Geology* 122: 109–127.
- Leckie D (1994) Canterbury Plains, New Zealand: Implications for sequence stratigraphic models. *American Association of Petroleum Geologists Bulletin* 78: 1240–1256.
- Lisiecki LE and Raymo ME (2005) A Pliocene-Pleistocene stack of 57 globally distributed benthic $\delta^{18}\text{O}$ records. *Paleoceanography* 20: <http://dx.doi.org/10.1029/2005PA001153>. PA1003.
- Lourens LJ, et al. (1996) Evaluation of the Plio-Pleistocene astronomical timescale. *Paleoceanography* 11: 391–413.
- Loutit TS, Hardenbol J, and Vail PR (1988) Condensed sections: The key to age determination and correlation of continental margin sequences. In: Wilgus CK, et al. (eds.) *Sea-Level Changes: An Integrated Approach*, vol. 42, pp. 183–213. Tulsa, OK: Society of Economic Paleontologists and Mineralogists Special Publication.
- Massari F, Rio D, Sagvetti M, et al. (2002) Interplay between tectonics and glacio-eustasy: Pleistocene succession of the Crotona basin, Calabria (southern Italy). *GSA Bulletin* 114: 1183–1209.
- Massari F, Sagvetti M, Rio D, D'Alessandro A, and Prosser G (1999) Composite sedimentary record of Pleistocene glacio-eustatic cycles in a shelf setting (Crotona basin, south Italy). *Sedimentary Geology* 127: 85–110.
- McMurray LS and Gawthorpe RL (2000) Along-strike variability of forced regressive deposits: Late Quaternary, northern Peloponnese, Greece. In: Hunt D and Gawthorpe RL (eds.) *Sedimentary Responses to Forced Regression*, vol. 172, pp. 363–377. London: Geological Society Special Publication.
- Miall AD (1997) *The Geology of Stratigraphic Sequences*. Berlin and Heidelberg: Springer-Verlag p. 433.

- Mitchum RM, Vail PR, and Thompson S (1977) Seismic stratigraphy and global changes of sea level, part 2: The depositional sequence as a basic unit for stratigraphic analysis. In: Payton CE (ed.) *Seismic Stratigraphy Applications to Hydrocarbon Exploration*, vol. 26, pp. 53–62. Tulsa, OK: American Association of Petroleum Geologists Memoir.
- Naish T, Field B, Zhu H, et al. (2005) Integrated outcrop, drill core, borehole and seismic stratigraphic architecture of a cyclothem, shallow-marine depositional system, Wanganui Basin, New Zealand. *Journal of the Royal Society of New Zealand* 35: 91–122.
- Naish TR and Kamp PJJ (1997a) Sequence stratigraphy of sixth-order (41 ky) Pliocene-Pleistocene cyclothem, Wanganui Basin, New Zealand: A case for the regressive systems tract. *Geological Society of America Bulletin* 109(8): 978–999.
- Naish TR and Kamp PJJ (1997b) High resolution foraminiferal depth palaeoecology of late Pliocene shelf sequences and systems tracts, Wanganui Basin, New Zealand. *Sedimentary Geology* 110: 237–255.
- Naish TR, et al. (1998) Astronomical calibration of a southern hemisphere Plio-Pleistocene Reference Section, Wanganui Basin, New Zealand. *Quaternary Science Reviews* 17: 695–710.
- Payton CE (ed.) (1977) *Seismic Stratigraphy Applications to Hydrocarbon Exploration*, pp. 53–62. Tulsa, OK: American Association of Petroleum Geologists Memoir.
- Pickering HT, Souter C, Oba T, Taira A, Schaaf M, and Platzman E (1999) Glacio-eustatic control on deep-marine clastic forearc sedimentation, Pliocene-mid-Pleistocene (c. 1180–600 Ka) Kazusa Group, SE Japan. *Journal of the Geological Society of London* 156: 125–136.
- Pillans B (1983) Upper Quaternary marine terrace geochronology and deformation, South Taranaki, New Zealand. *Geology* 11: 292–297.
- Pillans B, Chappell J, and Naish TR (1998) A review of the Milankovitch climate beat: Template for Plio-Pleistocene sea-level changes and sequence stratigraphy. *Sedimentary Geology* 122: 5–21.
- Pillans BJ, Alloway BV, Naish TR, Abbott SA, Westgate JA, and Palmer A (2005) The distal archive of North Island silicic volcanism recorded in Pleistocene shallow marine sediments of Wanganui Basin, New Zealand. *Journal of the Royal Society of New Zealand* 35: 43–90.
- Pillans BJ, Roberts AP, Wilson GS, Abbott ST, and Alloway BV (1994) Magnetostratigraphic, lithostratigraphic and tephrostratigraphic constraints on lower and middle Pleistocene sea-level changes, Wanganui Basin, New Zealand. *Earth and Planetary Science Letters* 121(1–2): 81–98.
- Plint AG (1988) Sharp-based shoreface sequences and 'offshore bars' in the Cardium Formation of Alberta: Their relationship to relative changes in sea level. In: Wilgus CK, et al. (ed.) *Sea-Level Changes: An Integrated Approach*, vol. 42, pp. 357–370. Tulsa, OK: Society of Economic Paleontologists and Mineralogists Special Publication.
- Rio D, et al. (1996) Reading Pleistocene eustasy in a tectonically active shelf setting (Crotone Peninsula, southern Italy). *Geology* 24: 743–746.
- Saul G, Naish T, Abbott ST, and Carter RM (1999) Sedimentary cyclicity in the marine Plio-Pleistocene: Sequence stratigraphic motifs characteristic of the last 2.5 my. *Geological Society of America Bulletin* 111: 524–537.
- Shackleton NJ, Berger A, and Peltier WR (1990) An alternative astronomical calibration of the lower Pleistocene time-scale based on ODP site 677. *Transactions of the Royal Society of Edinburgh, Earth Sciences* 81: 251–261.
- Vail PR, Mitchum RM, and Thompson S III (1977) Seismic stratigraphy and global changes of sea level, part 4: Global cycles of relative changes of sea level. In: Payton CE (ed.) *Seismic Stratigraphy Applications to Hydrocarbon Exploration*, pp. 83–97. Tulsa, OK: American Association of Petroleum Geologists Memoir 26.
- van Wagoner JC, Mitchum RM, Campion KM, and Rahmanian VD (1990) Siliciclastic Sequence Stratigraphy in Well Logs, Cores, and Outcrops. *American Association of Petroleum Geologists Methods in Exploration* 7: 55.
- Van Wagoner JC, et al. (1988) An overview of the fundamentals of sequence stratigraphy and key definitions. In: Wilgus CK, et al. (ed.) *Sea-Level Changes: An Integrated Approach*, vol. 42, pp. 39–45. Tulsa, OK: Society of Economic Paleontologists and Mineralogists Special Publication.
- Vella P (1963) Plio Pleistocene cyclothem, Wairarapa, New Zealand. *Transactions of the Royal Society of New Zealand* 22: 15–50.
- Walker RG (1984) General introduction: Facies, facies sequences, and facies models. In: Walker RG (ed.) *Facies Models*, pp. 1–9. St Johns, Newfoundland: Geological Association of Canada.
- Walker RG (1990) Facies modelling and sequence stratigraphy. *Journal of Sedimentary Petrology* 60: 777–786.

Tephrochronology

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Glossary

Tephra: This word originates from the Greek word *tephra* (τεφρα) meaning 'ashes.' It is an all-embracing term for the explosively erupted, loose fragmental (pyroclastic) products of a volcanic eruption, and includes all grain sizes ranging from fine dust to large blocks that may be several meters across. The plural is either tephra or tephra. (Strictly, 'tephra' is feminine singular and hence in Greek is used as a collective noun with singular verb forms, but appending 's' is quite appropriate in modern usage.) Tephra may include fall deposits (commonly called tephra fall or tephra fallout) and unconsolidated deposits derived from pyroclastic density currents (surges, flows, and transitional gas-water supported flow deposits). The term 'airfall' is no longer used.

Tephrochronology: The use of tephra as isochrons (time-parallel marker beds) to link sequences in different places via precise tie-points and to establish relative or numerical ages.

Tephrostratigraphy: Refers to the study of sequences of tephra layers or strata and associated deposits and their distribution, and their relative and numerical ages. It involves defining, describing, and characterizing or 'fingerprinting' tephra or tephra sequences using their physical, mineralogical, or geochemical properties, and underpins tephrochronology.

Cryptotephra: Refers to distal tephra deposits, preserved as glass shard and/or crystal concentrations in peat bogs, lake, marine, or aeolian sediments, or ice-cores, or soils, that are essentially not visible as a layer in the field. The term and its derivatives (cryptotephrostratigraphy, cryptotephrochronology) originate from the Greek *kryptein*, to hide, to convey the hidden or concealed nature of these typically fine-grained, sparse deposits.

Tephrochronometry: Refers to obtaining a numerical age or date for a tephra layer or cryptotephra.

Introduction

The word 'tephra' originates from a Greek word *tephra* (τεφρα) meaning 'ashes.' It is an all-embracing term for the explosively erupted, loose fragmental (pyroclastic) products of a volcanic eruption, and includes all grain sizes ranging from the finest dust to blocks the size of cars. The plural is either tephra or tephra – the former (as a singular collective noun) is more technically correct but the latter avoids ambiguity and is in common usage. Tephra may include fall deposits (commonly called tephra fall or tephra fallout) and unconsolidated deposits derived from pyroclastic flows or surges. Related terms include tephrostratigraphy, tephrochronology, and tephrochronometry. Note that in these derivatives the letter 'o' rather than 'a' is the appropriate connecting letter. The first known written use of 'tephra' was by Aristotle who described an eruption on the island of Hieria in the Lipari (Aeolian) Islands near Sicily around 350 BC. The first modern usage was by Sigurdur Thorarinsson of Iceland in 1944, who resurrected the term to fit with Greek words lava and magma and to link these with classical volcanology that derives from the Roman name for the island Vulcano, the southernmost of the Lipari Islands.

Tephra deposits have two special features: (1) they are erupted and deposited over very short time periods, geologically speaking, usually a matter of only hours or days to

perhaps weeks or months and (2) they can be spread widely over land and sea to form a thin blanket that (unless reworked) has the same age wherever it occurs. Therefore, once it is identified by its mineralogical and geochemical properties, a tephra layer provides a time-parallel marker bed or isochron for an 'instant' in time, that instant being the date of the eruption that produced the layer. 'Tephrochronology' is thus the use of tephra deposits as isochrons to link sequences in different settings via precise tie-points and to establish and transfer relative or numerical ages (Lowe and Hunt, 2001). In recent times, the term 'tephrochronology' has tended to be used in a broader way as a stand-alone term for all aspects of tephra studies (e.g., Lowe, 2011; Shane, 2000).

'Tephrostratigraphy' is another term that is commonly used and refers to the study of sequences of tephra layers and related deposits and their relative and numerical ages. It involves defining, describing, mapping, and characterizing or 'fingerprinting' tephra or tephra sequences using their physical, mineralogical, or geochemical properties, and fundamentally underpins tephrochronology. Several lines of evidence are needed to ensure confidence in most tephra correlations: (1) stratigraphic position, which provides relative age; (2) intrinsic compositional properties that are diagnostic and persistent; and (3) numerical age, normally based on radiometric, incremental, or other dating techniques.

Early Studies: Pioneers of Tephrochronology

Studies of tephra layers as a tool for research in various disciplines effectively began in the 1920s and 1930s in New Zealand, Japan, Iceland, South America, and USA. In New Zealand and Japan, pioneer tephra studies were carried out in association with soil surveys (Grange, 1931; Uragami et al., 1933). Sigurdur Thorarinsson, an Icelandic volcanologist, is widely regarded as the modern 'father of tephrochronology' (Figure 1). He worked with founder palynologist Lennart von Post in Sweden at the University of Stockholm from 1932 and realized that the numerous Icelandic tephra layers offered great possibilities for correlation and dating. A few years earlier, two Finnish geologists V. Auer and Th. G. Sahlstein had used tephra layers as marker horizons in their studies of vegetation history of Tierra del Fuego. Together with Hakon Bjarnason, Thorarinsson began work on the identification and mapping of some widespread tephra layers (Bjarnason and Thorarinsson, 1940) to obtain dated horizons in Icelandic bogs as an aid to studies of vegetation changes and rates of soil formation and soil erosion. Later, Thorarinsson (1944) published his doctoral thesis (in Swedish) in which he defined the terms 'tephra' and 'tephrochronology.' His immense contributions to the discipline are summarized in two of his last papers prepared for an international tephra meeting held in Iceland in 1980 (Thorarinsson, 1981a,b).



Figure 1 Sigurdur Thorarinsson in the field. Photograph courtesy of D. Reidel Publishing Company, Dordrecht, The Netherlands.

Characteristics of Tephra

Magnitude and Dispersal

Tephra layers are derived from explosive eruptions that inject particles (pyroclasts) to tropospheric and even stratospheric levels. The particles are carried upward by an eruption column that consists of a lower gas-thrust region and an upper convective region. A column will continue to rise convectively until its density is equal to that of the surrounding atmosphere. It will then expand laterally but will also continue upward due to its momentum to form a broad umbrella cloud that plays an important role in the transport of pyroclasts. Temperature of the erupted material and mass eruption rate determines the height of an eruption column which, along with particle size, wind strength, and direction, exerts the principal controls on the long-distance transport of tephra. Three dominant threshold settling velocity values affect the distribution and sorting of airborne tephra: (1) fragments with large settling velocities follow ballistic trajectories that are little affected by wind and only slightly by the expanding eruption cloud, (2) particles suspended by turbulence in the eruption cloud that are too heavy to be suspended by atmospheric winds, and (3) those light enough to be suspended by wind independently of the eruption cloud. Wind may modify the trajectories of all but the largest fragments; fragments held in turbulent suspension within the eruption cloud begin to fall according to their settling velocities as the energy within the cloud dissipates (Figure 2).

Tephra particles range in size from ash (<2 mm) and lapilli (2–64 mm) to blocks (dense, angular) and bombs (vesicular, rounded) (>64 mm) that may reach several meters in diameter. Densities vary from low density, vesicular pumice, and scoria to dense crystals and rock fragments. Materials may be juvenile (formed from magma involved in the eruption) or accidental or xenolithic (derived from preexisting rock or from material beneath the volcano, respectively). Silicic tephra are typically much more widely dispersed than those of basaltic composition.

Tephra are distributed in two-end member patterns with many variations: circular patterns from low-eruption columns during calm conditions (Figure 3(a)) and elliptical to fan-shaped patterns from high-eruption columns that encounter strong unidirectional winds (Figure 3(b)). The most typical pattern is fan shaped with the apex at or near the source with multicomponent sheets radiating out from a common source or overlapping with slightly different distribution trends depending on shifting wind directions. Tephra blankets tend to be gentle wedges that systematically thin along an axis away from source and are best defined by an isopach map constructed from numerous thickness measurements. The layers decrease exponentially in thickness and grain size with increasing distance from source, a reflection of sorting of particles by size and density as they fall through the atmosphere. In some historical eruptions (e.g., Mt. St. Helens, 1980), isolated areas of increased thickness are present in the distal portions of tephra plumes. Several tephra transport models have been developed which relate distance from source to grain size for volcanic eruptions of varying intensity and for varying atmospheric wind profiles and seasonality. Unusually, powerful eruptions give rise to tephra layers easily recognizable

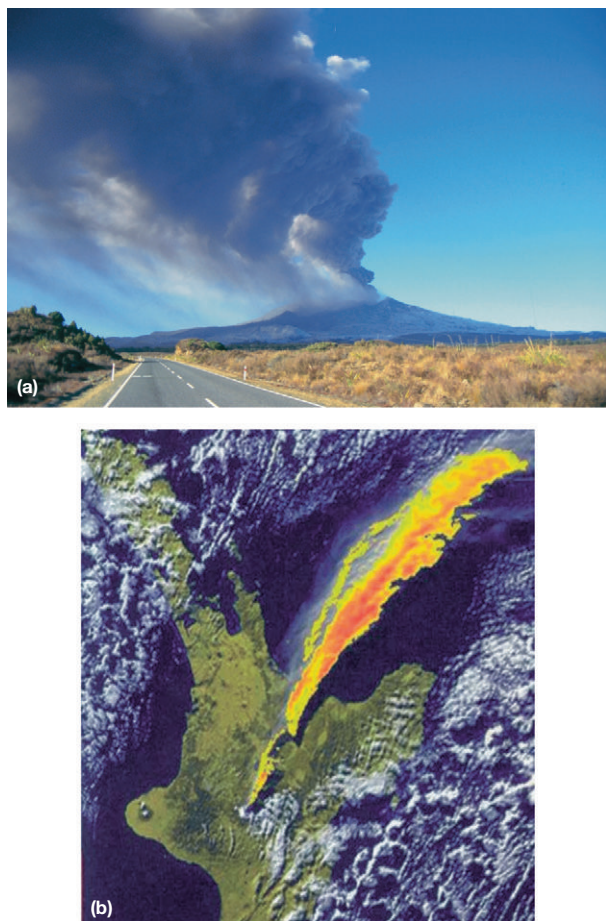


Figure 2 (a) Tephra plume of the 1996 eruption of Ruapehu volcano. Photograph courtesy of Vern Manville, University of Leeds. (b) Eruption plume of Ruapehu volcano, central North Island, New Zealand, as seen from space in 1995. Photograph courtesy of Landcare Research, New Zealand.

1000 km or more from source (Figure 4(a) and 4(b)). The fine ash can be carried over areas of continental size or larger and extend far beyond the recognizable limits of the dispersal fan.

The initial eruption of the Icelandic Eyjafjöll volcano during the Northern Hemispheric spring of 2010 has shown that it is not the most voluminous of eruptions that necessarily result in the widest dispersal of tephra. The AD 2010 Eyjafjöll eruption was relatively small scale (>0.01 to <1 km³) when compared to the super-eruptions of Youngest Toba (ca. 74 ka before present (BP); Figure 4(a)) or Lava Creek B (ca. 0.64 Ma; Figure 4(b)), both of which have explosivity indices at the highest end of the scale (VEI 7–8) with erupted volumes in excess of >100 km³. Yet, this relatively low-magnitude event upwind of mainland Europe and the prevailing wind direction led to widespread and unprecedented disruption to air travel over Europe with significant economic consequences. Following the onset of the eruption, satellite imagery demonstrated the rapid transportation of ash by westerly winds over mainland Europe, eventually expanding to large areas of the North Atlantic Ocean extending toward the eastern seaboard of Canada (UK Met Office, 2010). The AD 2010 Eyjafjöll event

highlights that weather patterns at the time of eruption can not only play a critical role in the transportation and deposition of a relatively small amount of tephra, but also serves as a clear reminder of the unexpected and costly impacts of ash inundation in distal areas downwind of volcanic sources (Davies et al., 2010).

Sparse glass shards from widely dispersed tephra occasionally occur within acid layers preserved in high-latitude ice cores, and geochemical analysis of these can be used to link to other records or to refine calendrical age. For example, the Icelandic-sourced Vatnaöldur tephra layer, deposition of which coincided with the Norse settlement at ca. 874, has Greenland ice-core dates of AD 871 ± 2 and 877 ± 7 (Grönvold et al., 1995; Zielinski et al., 1997). The powerful Mazama eruption in Oregon, USA, has an ice-core age of 7627 ± 150 calendar year BP (Zdanowicz et al., 1999; Figure 5). The ice-core derived calendar ages improve the possibility for correlation of paleoclimate records developed from terrestrial and lacustrine sediments in areas where these tephtras are known. Such records are now able to be better linked to ice-core and marine records via tephrochronology (e.g., Davies et al., 2007, 2008).

Field Characteristics

Macroscopic

Tephtras, depending on the distance from source, can range from millimeter- to meter-thickness, typically mantle preexisting topography (Figure 6) and often display marked lateral variations in thickness and continuity. In general, bedding in near-source (proximal) tephra deposits consists of alternating gradational coarse- to finer-grained layers without sharp bedding, but whether bedding is sharp or gradational depends upon several factors such as the duration and energy fluctuations of an eruption, tephra volume and discharge rate, direction and strength of winds during the eruption, and the duration between eruptions during which erosion, weathering, or winnowing of fines from the tephra may take place. Where the upper part of a tephra layer (in terrestrial settings) is weathered and modified by soil-forming processes, it becomes a paleosol upon burial by subsequent deposits (Figure 7).

Thin (centimeter-thick), distal tephra layers transported by winds over long distances are commonly massive to normal graded (Figure 8) and may be affected by bioturbation, soft-sediment deformation and locally reworked material washing into the site. Lower contacts are generally sharp and planar, whereas the upper contacts can be gradational admixtures of nonvolcanic grains over millimeters to meters up section. Some tephra layers can display tractional sedimentary structures such as trough cross-stratification from reworking in shallow marine and fluvial aqueous environments (Figure 9).

Difficulties can arise relating to the preservation of millimeter-thick distal tephra in marine and other sediments. Although such tephtras may occur as a single, thin continuous horizon, in many cases glass shard concentrations may be discontinuous and dispersed through a band of sediment up to several decimeters in thickness. This dispersal presents the problem of determining which stratigraphic position

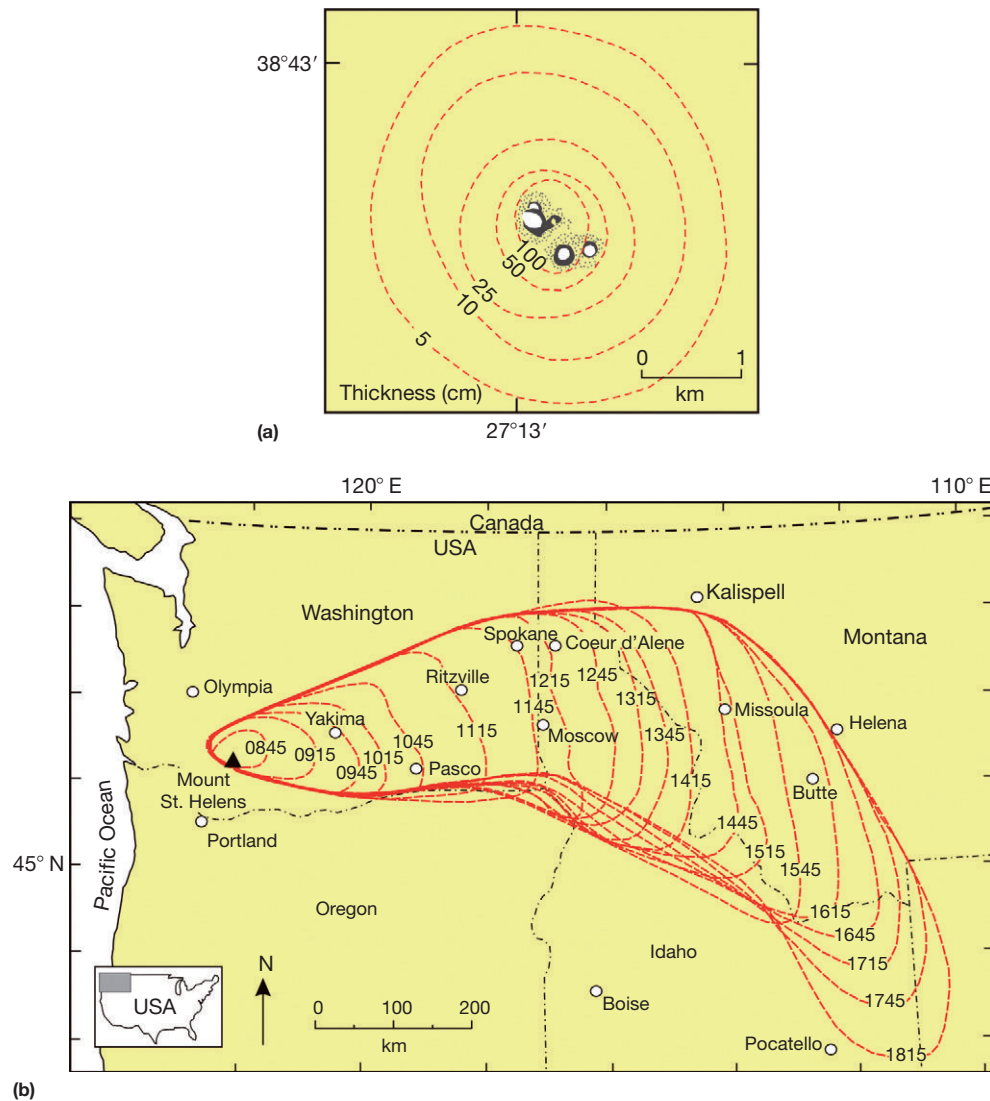


Figure 3 Tephra distribution patterns. (a) Tephra distribution patterns. Circular isopach map of Algar do Carrao I scoria-fall deposit associated with a low-intensity basaltic eruption at Tereira, Azores (reproduced from Self S (1976) The recent volcanology of Terceira, Azores. *Journal of the Geological Society of London* 132: 645–666), and (b) elliptical pattern associated with tephra fallout from the Mt. St. Helens eruption, 1980, that spread as far as 1000 km from the volcanic center in less than 10 h (reproduced from Sarna-Wojcicki AM, Shipley S, Waitt RB Jr, Dzuris D, and Wood SH (1981) Areal distribution, thickness, mass, volume, and grain size of air-fall ash from six major eruptions of 1980. In: Lipman PW and Mullineaux DR (eds.). *The 1980 Eruptions of Mount St. Helens*, pp. 577–600. Washington: US Geological Survey Professional Paper 1250).

represents the point of actual tephra deposition. Usually, the point of maximum shard concentration in a section or core is regarded as marking the tephra-fall datum (chronohorizon) (e.g., Almond, 1996). Typically, there are additional problems associated with bioturbation and postdepositional reworking (e.g., Austin et al., 2004; Gehrels et al., 2006; Newnham et al., 2007a; Payne and Gehrels, 2010). These difficulties also apply to cryptotephra deposits described below.

Microscopic

Distal tephra deposits, preserved as glass-shard and/or crystal concentrations in peat bogs, lake, marine, or aeolian sediments, or ice-cores, or soils, may be invisible to the naked eye in the field (Lowe, 2011). Such deposits, usually of fine ash

grain-size, are referred to as 'cryptotephra,' from the Greek *kryptein*, to hide, to convey the hidden or concealed nature of these diminutive deposits. Until relatively recently, tephra records reported from sites in Europe were predominantly visible layers of ash, prominent in sedimentary sequences because the individual glass shards are of relatively large size, and/or because shard concentrations in the sediment were extremely high. Although visible ash layers had been correlated to sites distant from volcanic sources – such as the Icelandic Vedde Ash found in Norway (Birks et al., 1996) and the Laacher – see tephra from the Eifel district in Germany (van den Bogaard and Schmincke, 1985), traced as far as Turin in north Italy – the areas of Europe in which tephrochronology could be used as an adjunct to stratigraphical investigations were relatively limited. Andrew Dugmore was the first to identify a cryptic

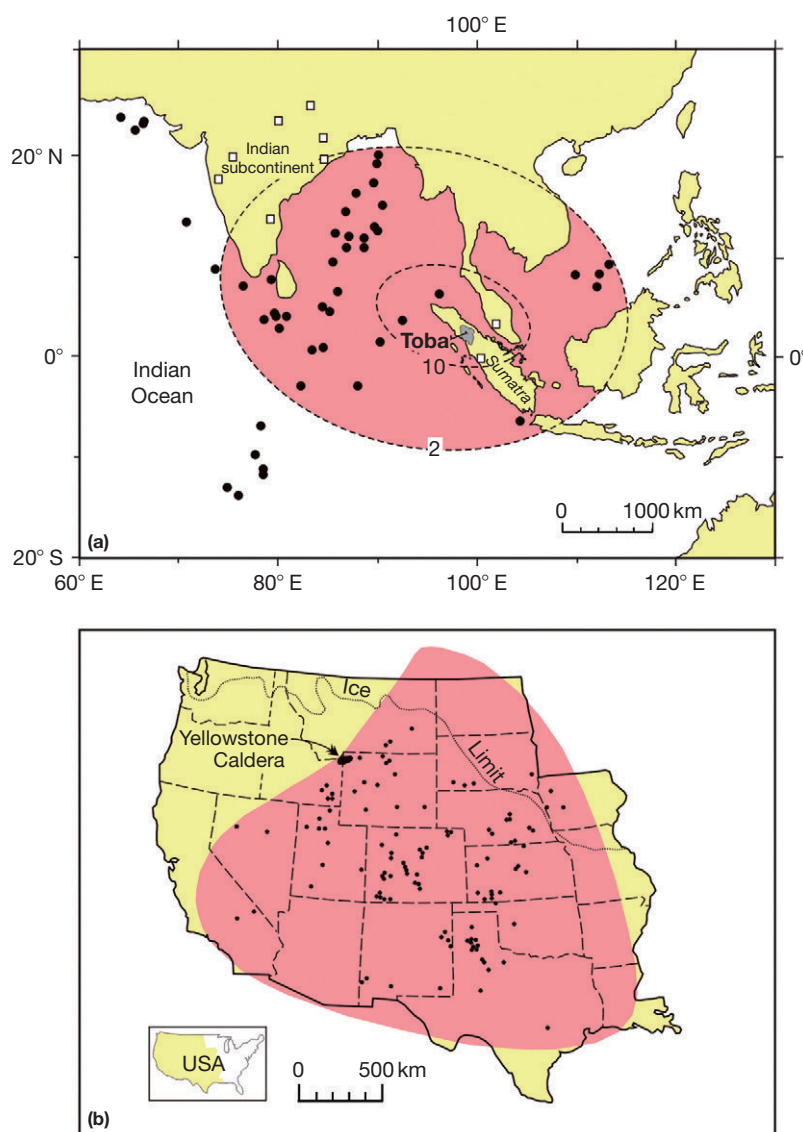


Figure 4 Distributions of two exceptionally large Pleistocene eruptions: (a) Youngest Toba Tephra (ca. 74 ka BP) erupted from Toba Caldera Complex in northern Sumatra. Isopach values for Toba Tephra are in centimeters (reproduced from Lee M-Y, Chen C-H, Wei K-Y, Lizuka Y, and Carey S (2004) First Toba supereruption revival. *Geology* 32: 61–64, see also Matthews et al., 2012); (b) Lava Creek B (ca. 0.64 Ma) from Yellowstone Caldera, west-central United States of America (reproduced from Izett GA and Wilcox RE (1982) Map showing localities and inferred distribution of the Huckleberry Ridge, Mesa Falls and Lava Creek ash beds (Pearlette family ash beds) of Pliocene and Pleistocene age in the Western United States and southern Canada. *US Geological Survey Miscellaneous Investigations Series Map I-325* scale 1:4 000 000).

Icelandic tephra in the United Kingdom (Dugmore, 1989a). The later development of a density separation (flotation) method (Turney, 1998) led to the detection of the 10.3 ^{14}C ka BP Vedde Ash in Scotland, and subsequent studies further demonstrated the presence of microscopic glass shards in sediments well beyond the limits of visible tephra records. Since its discovery in Scotland, the Vedde Ash has been identified in Scandinavia, The Netherlands, and NW Russia (Figure 10; see also Blockley et al., 2007). More than 34 different tephras have been recognized in western Europe and the NE Atlantic region spanning the period between ca. 18.5 and 8 ^{14}C ka BP (Davies et al., 2002). These originated from four main volcanic provinces: Iceland, the Campanian district in Italy, the Eifel district, and the Massif Central in France, and collectively offer

considerable potential for developing precise tie lines between ice, marine, and terrestrial records, and hence a basis for correlation that is independent of other dating methods (Turney et al., 2004; Lane et al., 2012). The observed occurrence of cryptotephra from the Icelandic Eyjafjöll AD 2010 eruption provides an important modern analogue for understanding ash dispersal patterns over Europe and the NE Atlantic region in the past, particularly in relation to the critical role of weather patterns, as well as evaluating the dispersal of different eruptive phases in widely separated localities (Davies et al., 2010).

Chemical traces

A single eruption injects various quantities of insoluble silicate matter (tephra) and gases into various levels of the

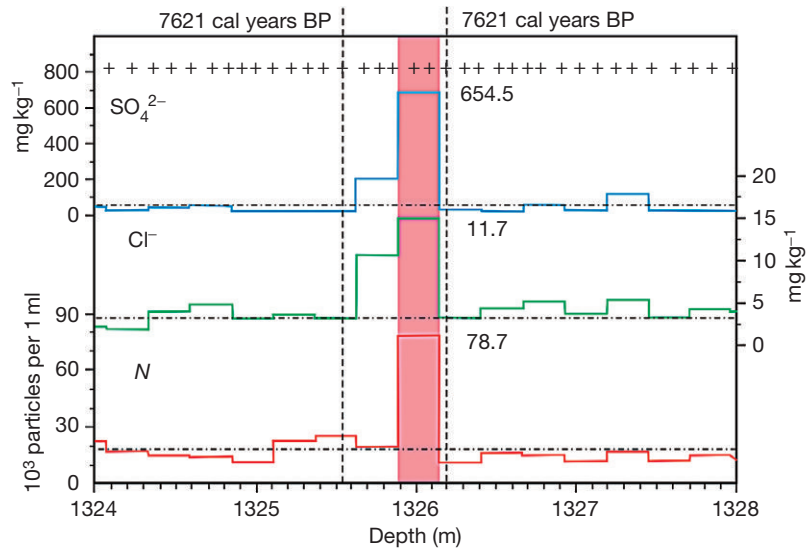


Figure 5 Time series of SO_4^{2-} , Cl^- , and microparticle concentration (N) in GISP2 ice core, showing peaks assigned to Mount Mazama eruption fallout. Reproduced from Zdanowicz CM, Zielinski GA, and Germani MS (1999) Mount Mazama eruption: Calendrical age verified and atmospheric impacts assessed. *Geology* 27: 621–624.



Figure 6 Mesa Falls Tephra sourced from Yellowstone Caldera, USA, mantling preexisting topography. Inset – gradational coarse-fine internal structure of fall deposits. Photograph by Brent Alloway.

atmosphere. Being larger and heavier than the aerosols produced by the oxidation of gases released, the ash components quickly settle out and consequently have limited climate influence except for areas in the immediate vicinity of the eruption. In contrast, aerosols such as H_2SO_4 (oxidized from SO_2 and H_2S) will remain in the stratosphere for several years providing a source for climate disturbances as well as providing a nucleus for various chemical reactions that can lead to ozone loss. The El Chicon (1982) and Mount Pinatubo (1991) eruptions illustrated that volcanic aerosols can potentially have a significant impact on regional to global atmospheric concentrations as well as an influence on global mean temperature. Ice-core parameters that document the presence of volcanic aerosols include SO_4^{2-} and electrical properties of ice such as

conductivity, which provides a measure of total acids and total salts. More recently, volcanic records in ice cores have reached a new level of detail through the measurement of both SO_4^{2-} and electrical properties on the Greenland GISP2 and GRIP ice cores. In some cases, sparse, very fine-grained ($<20\ \mu\text{m}$) glass shards from widely dispersed tephra are found within acid layers (see Figure 5; see also Abbott and Davies, in press). The longest existing continuous record of volcanism from an ice core is the 110 ka record from the GISP2 core which reflects high-sulfur-producing equatorial and Northern Hemisphere eruptions. The ability to obtain volcanic records from both the Greenland and Antarctic ice cores enhances the capabilities of detecting past eruptions that may have had a global impact (e.g., equatorial Tambora 1815 and Krakatau

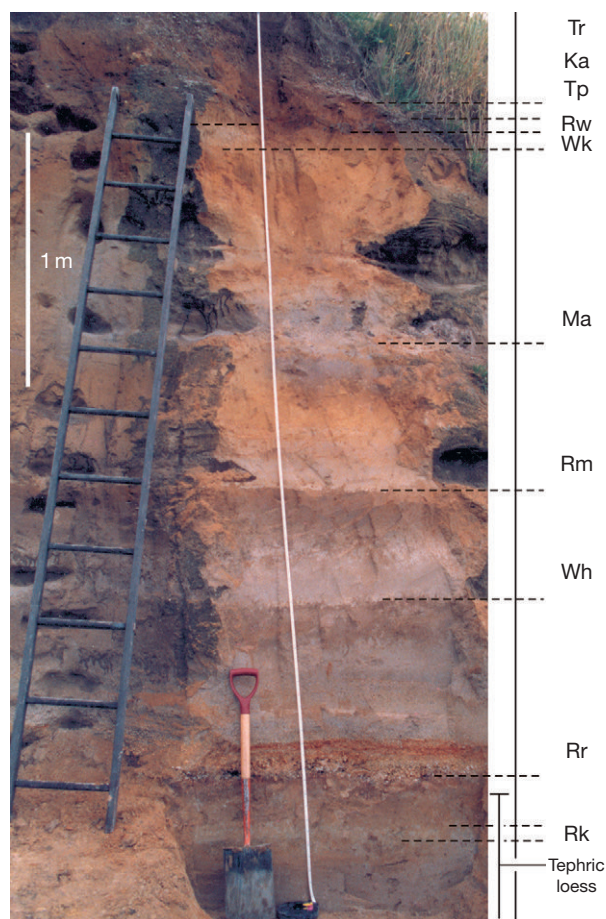


Figure 7 Road cutting near Rotorua, North Island, New Zealand, showing tephra layers and associated buried paleosols (dark brownish to yellowish-brown horizons) dating back ca. 18 cal ka. Deposition of each tephra was followed by a period of quiescence and soil formation; then the soil was buried by tephra from a new eruption. German explorer Dr Ernst Dieffenbach described a site near here in 1839, the first recorded description of tephra in New Zealand. All the tephra apart from Taupo (Tp) were erupted from nearby Okataina caldera. Names and eruption ages are as follows (volcano name in brackets): Tr, Rotomahana Mud AD 1886 (Tarawera); Ka, Kaharoa ca. AD 1314 (Tarawera); Tp, Taupo ca. AD 232 (Taupo); Rw, Rotokawau ca. 3.6 cal ka BP (Rotokawau craters); Wk, Whakatane ca. 5.5 cal ka BP (Haroharo); Mk, Mamaku, ca. 8.0 cal ka BP (Haroharo); Rm, Rotoma ca. 9.5 cal ka BP (Haroharo); Wh, Waiohau ca. 14 cal ka BP (Tarawera); Rr, Rotorua ca. 15.6 cal ka BP (Okareka/Haroharo); Rk, Rerewhakaaitu (pale layer within grayish loess deposits) ca. 17.6 cal ka BP (Tarawera). Photograph by B.E. Green in Hodder et al. (1990).

1883 eruptions). When the signal is found in only one of the polar regions, a source eruption from the hemisphere of the ice core is implied. That is the case for the Icelandic Laki (1783) and Alaskan Katmai (1912) eruptions, signals found in almost all Greenland ice cores, and for the New Zealand Tarawera eruption (1886) signal found in Antarctic cores (Delmas, 1992; Zielinski, 2000; Figure 11).

Mineralogical and Geochemical Characteristics

The use of tephra layers to define isochronous horizons, traceable between different stratigraphic successions, is only

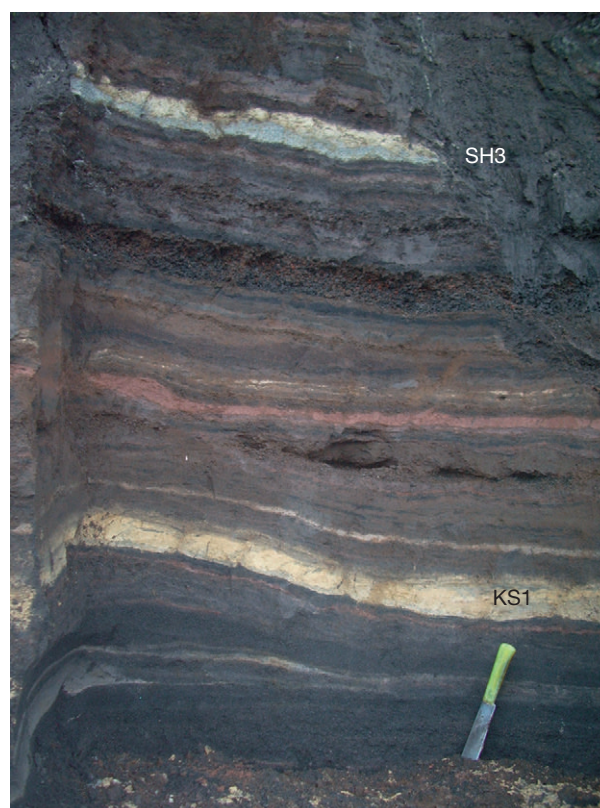


Figure 8 Tephra sequence deposited between 1.3 and 2.0 ^{14}C ka BP, 14 km NNE of Kliuchevskoi volcano summit, Kamchatka, Russia. The thick, upper light-colored tephra layer is sourced from Shiveluch volcano (SH3) with an age of ~ 1.4 ^{14}C ka BP, and the thick, lower pale brown tephra layer is sourced from Ksudach volcano (KS1) with an age of ~ 1.8 ^{14}C ka BP. Both tephra are prominent Holocene marker beds throughout Kamchatka and have been described in Braitseva et al. (1997). Photograph courtesy of Vera Ponomareva, Institute of Volcanology and Seismology, Russia.

possible if each horizon has unique characteristics, which not only aid detection and identification, but also reduce the possibility of spurious correlations between tephra of different ages but with similar physical and geochemical properties (Lowe, 2011).

Ferromagnesian mineral assemblages

Ferromagnesian mineralogical assemblages of a tephra horizon $< \sim 100$ km from the eruption center can aid in the identification of source volcano (Froggatt and Lowe, 1990). Tephra derived from the same center, or from several centers with similar geological characteristics and histories, will tend to have similar mineralogical composition, but changes through time are also known. Tephra from typical calc-alkaline volcanoes contain one or more of the mineral phases orthopyroxene, clinopyroxene, amphiboles (hornblende and cummingtonite), and biotite. Volcanoes that erupt peralkaline magmas may contain sodic phases such as aegirine, riebeckite, ferrohedenbergite, and aenigmatite. The minerals can be easily examined as loose grains (crystals) under a binocular or petrological microscope after the tephra sample has been crushed and wet sieved to remove particles < 63 μm in size. Distal tephra

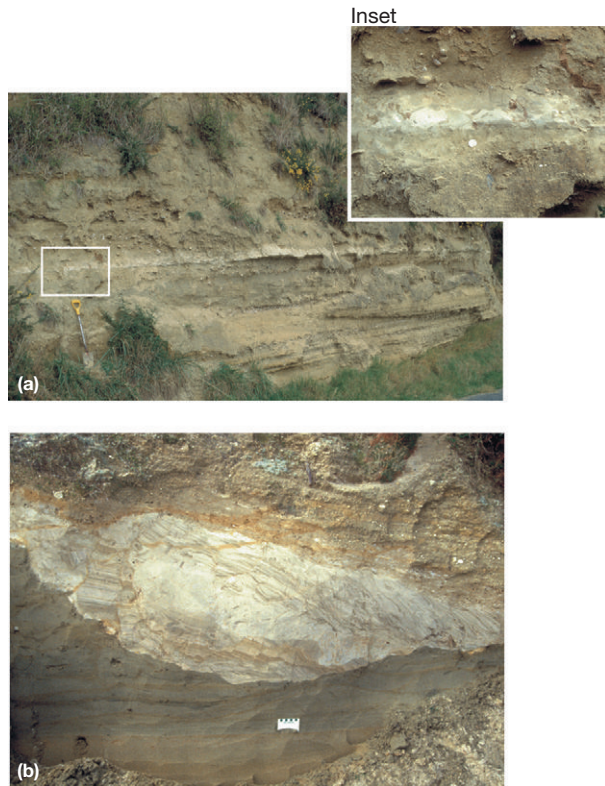


Figure 9 (a) Laterally continuous tephra bed within cross-stratified highly fossiliferous shallow marine sandy siltstones. Inset: Extensive burrowing and sharp but irregular upper and lower contact. (b) Discontinuous lenses of pumiceous and crystal-rich sands, pumiceous granules and gravels within cross-bedded sands, Wanganui Basin, New Zealand. Photographs: Brent Alloway.

beds (>100 km from source) are often mineral-poor because of atmospheric sorting during transport. For such samples, the ferromagnesian minerals can be concentrated using heavy liquids or electromagnetic methods. Problems with the use of mineralogies include the total dissolution of some minerals (e.g., biotite) and preferential dissolution of orthopyroxene relative to hornblende in some acid peat settings, resulting in incorrect ratio estimates (Hodder et al., 1991). Platy minerals such as biotite are easily winnowed from deposits that have been subjected to wind or water action. Therefore, the presence of a particular mineral is more instructive than its absence. Because many tephra beds contain similar mineral assemblages, it is only practical to use mineralogy as a tool when accurate stratigraphic control is available. In the search for a more diagnostic basis for distinguishing between tephra beds of similar mineralogy, researchers now routinely utilize geochemical techniques to fingerprint tephra mineral components, an approach made increasingly more precise and accessible through the development of high-precision analytical instruments with high spatial resolution.

Glass geochemistry

The chemical composition of volcanic glass (e.g., Figure 12) provides a well-established means for the stratigraphic

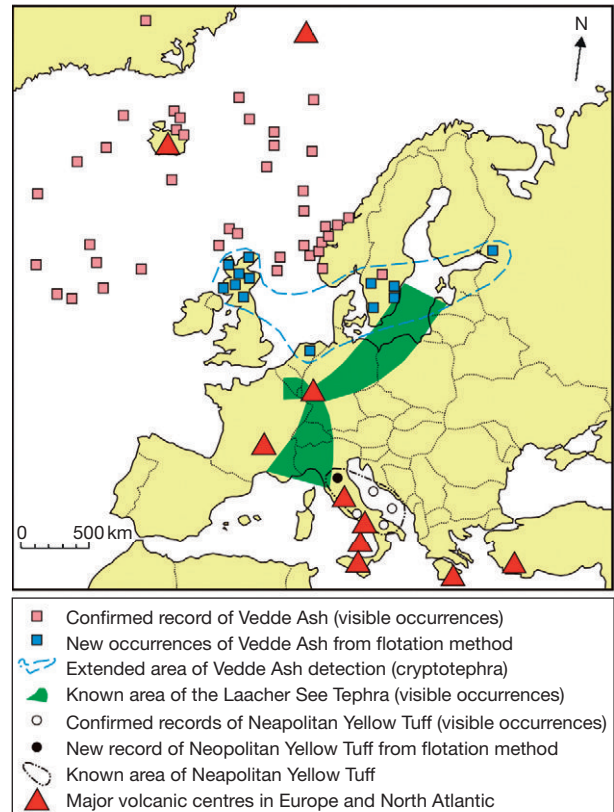


Figure 10 Known distribution of the Vedde Ash (VA), Laacher See (LST), and Neapolitan Yellow Tuff (NYT). The locations of the principal volcanic centers that were active during the period ca. 18.5–8 ^{14}C ka BP are also identified. Reproduced from Davies SM, Branch NP, Lowe JJ, and Turney CSM (2002) Towards a European tephrochronological framework for Termination 1 and the Early Holocene. *Philosophical Transactions Royal Society of London A* 360: 767–802.

correlation of tephra. Numerous studies attest to the application of electron microprobe (EMP) analysis to enable geochemical fingerprinting of tephra by their major element abundances. Nine major elements determined by EMP are routinely analyzed (e.g., Shane, 2000). However, it is imperative to use recognized standards and to check analyses against reference samples, assay of which should be published alongside newly acquired data (Kuehn et al., 2011; Lowe, 2011; Westgate et al., 2008). Many tephra beds can be distinguished on the basis of a few glass major element oxides such as FeO_t and CaO (wt%) (Figure 13(a)). Within-tephra homogeneity and compositional differences between tephra deposits reflect derivation from separate magma batches that lacked pre-eruptive compositional gradients. Homogeneous glass populations, derived from a single eruptive event, are characterized by low standard deviations and tight clustering of data points on element–element plots (Figure 13(b)). Some tephra have major element compositions that comprise several discrete populations (Figure 13(c) and 13(d)) or a range of compositions (e.g., Pyne-O'Donnell et al., 2008; Shane et al., 2008). Once recognized, such compositional variability can enhance correlation by providing additional fingerprinting criteria, but sampling from the full spatial and temporal range of deposits of an eruptive sequence is required (Shane et al., 2008).

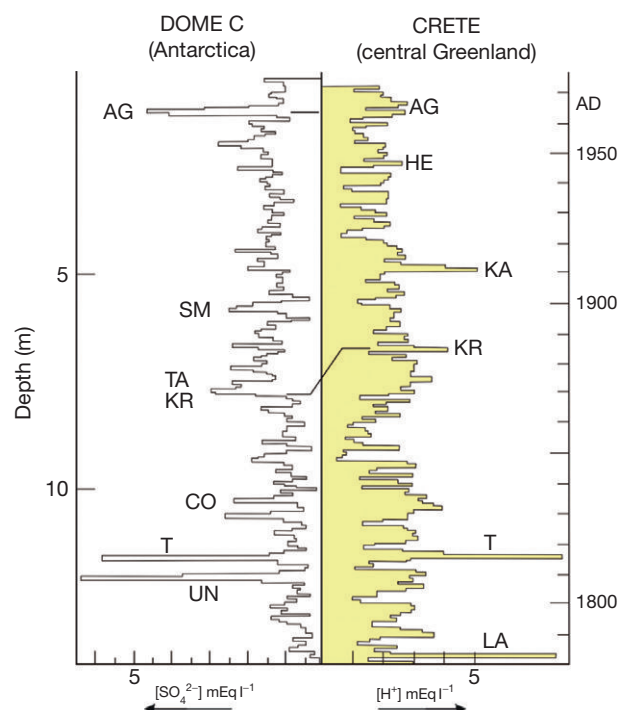


Figure 11 Comparison of eruption-derived acid fallout deposited in Greenland and Antarctic snow over the last 210 years. On the right-hand side, the acidity profile (electrical conductivity measurements (ECM)) obtained at Crete, central Greenland, is shown. On the left-hand side are shown the sulfate measurements obtained at Dome C. The following eruptions are identified: Agung (AG), Santa Maria (SM), Tarawera (TA), Krakatoa (KR), Coseguina (CO), Tambora (T), Unknown (UN), Hekla (HE), Katmai (KA), and Laki (LA). Reproduced from Delmas RJ (1992) Environmental information from ice cores. *Reviews of Geophysics* 30: 1–21. Copyright 1992, American Geophysical Union.

Redeposited and/or mixed tephra beds can also be distinguished by glass major-element compositional analysis, which indicates when a glass population most likely represents the mixed products of separate eruptive events. Increasingly, single shard, laser ablation, inductively coupled, plasma mass spectrometry (LA-ICP-MS) is being used to provide high-quality, grain-specific trace element data that can aid in the correlation of tephtras (Pearce et al., 2004, 2007, 2008, 2011). Furthermore, the high spatial resolution offered by single-shard analysis by LA-ICP-MS provides the potential to shed light on processes operating during fractional crystallization and evolution of magma (e.g., Allan et al., 2008).

Some studies have added a cautionary note regarding the postdepositional chemical instability of some glass, especially basaltic, and hence potential for miscorrelation (Blockley et al., 2005; Pollard et al., 2003).

Fe–Ti oxides and silicate mineral chemistry

Grain-specific analyses of Fe–Ti oxides and estimates of eruption temperature (T) and oxygen fugacity (fO_2) (oxidation state of the magma) have been used to fingerprint tephra deposits (see Shane, 1998; Figure 14). The composition of the titanomagnetite phase is particularly diagnostic of eruptive center and can be used to distinguish many tephra beds from

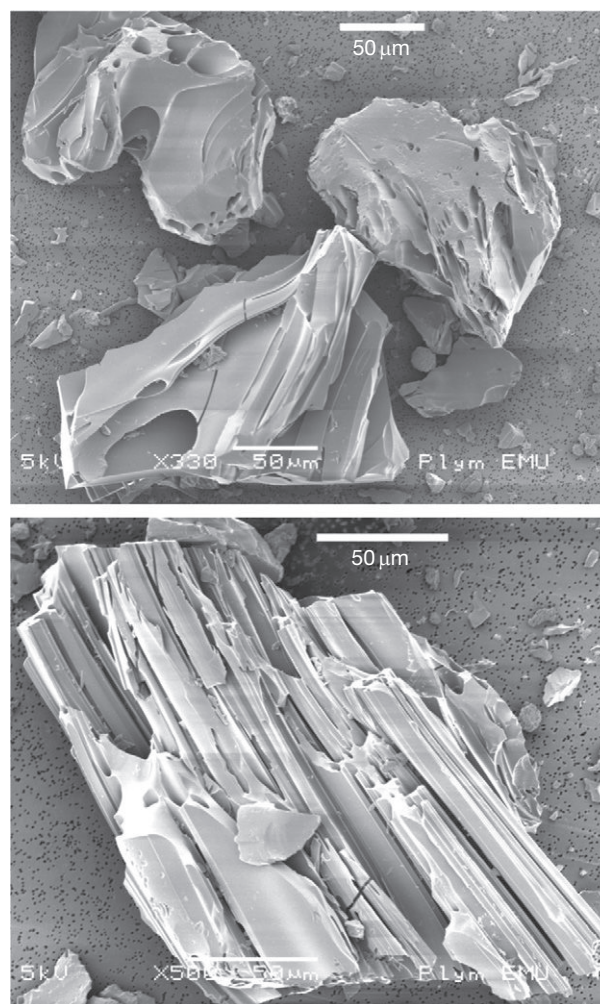


Figure 12 Scanning electron micrograph images of mid-Holocene rhyolitic glass shards extracted from a peat core in North Island, New Zealand. Such shards can be polished and analyzed by EMP to identify their source and likely correlative. Reproduced from Gehrels MJ, Lowe DJ, Hazell ZJ, and Newnham RM (2006) A continuous 5000-year Holocene cryptotephrostratigraphic record from northern New Zealand: Implications for tephrochronology and volcanic hazard assessment. *The Holocene* 16: 173–187.

the same volcano. TiO_2 is the most variable major oxide, ranging between ~6.5 and 16 wt% for different eruptions. MgO , MnO , and Al_2O_3 may vary by a factor >2 in both the spinel and rhombohedral (ilmenite) phases between different tephra beds. The compositional range within individual eruptive units is narrow if the tephra was derived from a thermally homogeneous magma. However, some tephtras are derived from mingled magmas and thus display a wide compositional range in Fe–Ti oxides. Individual eruptive events that display uniform Fe–Ti oxide compositions and thus narrow ranges in T (<~20 °C) and log fO_2 (<~0.5) allow these features to identify individual magma batches. These criteria can help distinguish tephra deposits of similar bulk or glass composition that originated from the same volcano, and can be used to identify the volcanic source.

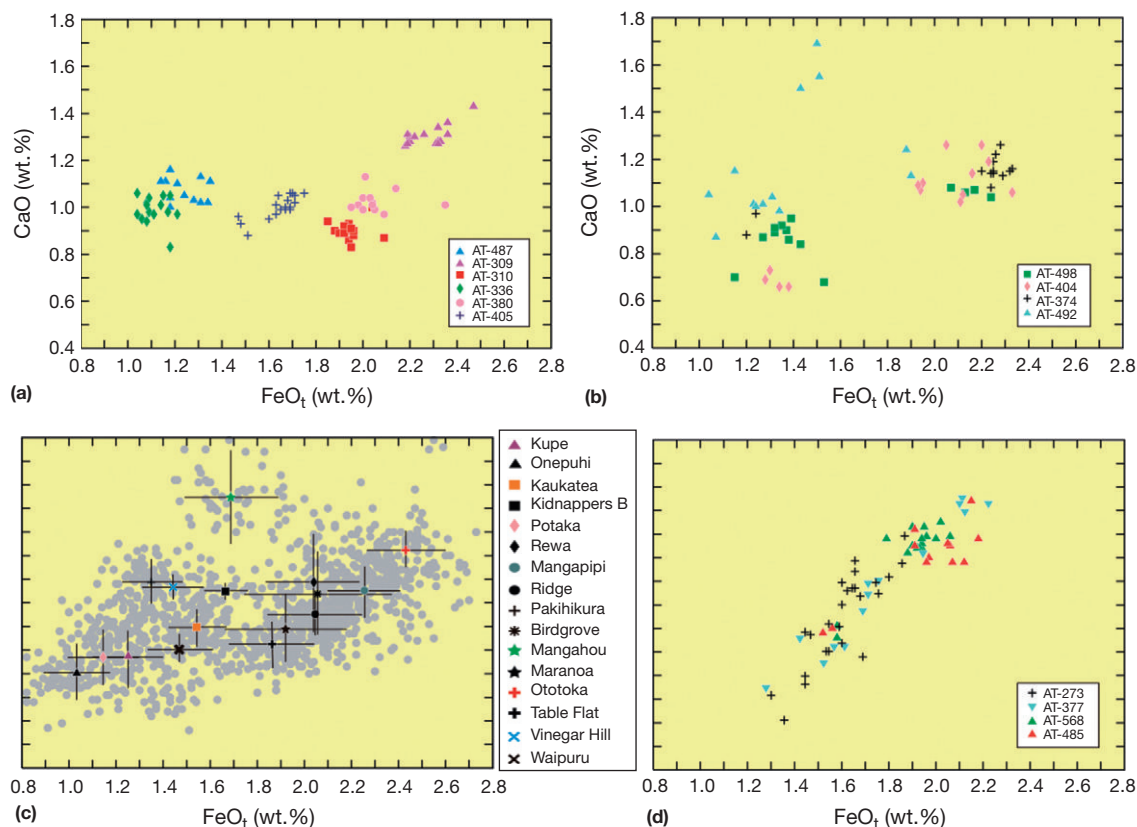


Figure 13 Examples of the glass composition in rhyolitic tephra beds determined by electron microprobe analysis. (a) Plot of CaO versus FeO_t (wt%) compositions of glass shards from six selected tephra layers showing their compositional homogeneity; (b) CaO versus FeO_t (wt%) compositions of glass shards of all named and stratigraphically discrete tephra beds in Wanganui Basin, New Zealand. Error bars represent 1 standard deviation. Gray dots represent all Wanganui analyses (reproduced from Pillans B, Alloway BV, Naish T, Westgate JA, Abbot S, and Palmer AS (2005) Silicic tephra in Pleistocene shallow marine sediments of Wanganui Basin, New Zealand. *Journal of the Royal Society of New Zealand* 35: 43–90; see also Figure 24). Examples of compositional heterogeneity found in individual tephra beds: (c) Two (AT-498, -404, and -374) and three (AT-492) discrete compositional groups. (d) Dominant shard population with a single outlier shard (AT-568 and -485) and near continuum of compositions (AT-273 and -377). All examples (except Figure 13(b)) come from ODP cores retrieved east of New Zealand (e.g., Alloway et al., 2005).

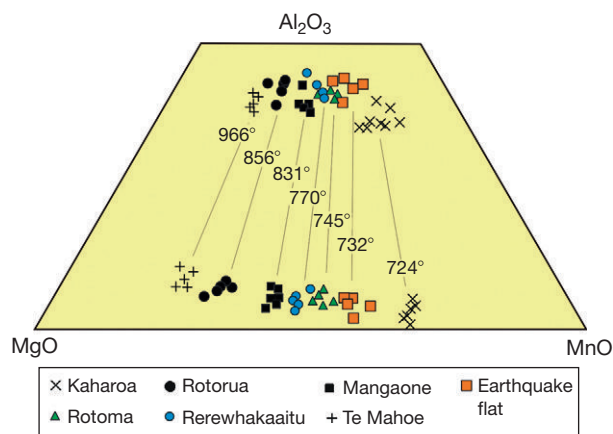


Figure 14 The minor element composition of spinel (Al₂O₃-rich phase) and ilmenites (Al₂O₃-poor phase) in rhyolite tephra from Okataina Volcanic Centre (Shane, 1998), demonstrating their fingerprinting ability. Also shown is estimated eruption temperature based on equilibrium between spinel and ilmenites.

Silicate minerals, on the other hand, have a set crystal structure and only limited chemical substitutions are possible. Thus, in many silicic tephra deposits such minerals cannot be used to distinguish between eruptive events. In addition, the most common phases often display considerable composition range within individual deposits, such as orthopyroxene (En range >10%) and are commonly zoned, for example, plagioclase (An range 10–20%). The compositions of amphiboles and pyroxenes have been used successfully in some studies, however, to support tephra identification (e.g., Cronin et al., 1996). Biotite, which is common in tephra with high-K glass (>4 wt% K₂O) can also be valuable (Shane et al., 2003). The FeO/MgO ratio in biotite does not correlate with FeO/MgO in the coexisting glass phase so that tephra beds with similar glass compositions can be discriminated by their different biotite compositions. Some individual tephra deposits display sequential changes in biotite composition that allow separate phases of the eruption to be identified, greatly increasing the potential precision for correlation. In addition, devitrified lavas that are unsuitable for glass analysis can be correlated to coeval tephra deposits by their biotite compositions.

Chronology

Direct Dating of Tephra

Historical accounts: eye-witness descriptions of tephra falls/eruptions

Historical accounts have provided accurate eruption years for many eruptions in countries where volcanic activity is common, in particular during the past 200–300 years. The most widely known historical account was contained in two letters from Pliny the Younger to the Roman historian Tacitus, describing the death of his uncle, the famous scholar Pliny the Elder, by asphyxiation while observing the AD 79 eruption of Vesuvius. Yet the 1815 eruption of Tambora (Sumbawa, Indonesia), the largest and deadliest eruption in recorded history, had little scientific impact because of the volcano's remote location and slow global communication at the time. No scientific expedition was dispatched to Tambora until 1847, and another 8 years passed before an expedition report was published. In contrast, the Krakatau (Sunda Straits, Indonesia) eruption in 1883 was the first to be quickly and widely reported because the telegraph system had been developed by then.

There are many countries around the world where eruptions have been documented through oral or written accounts. Frequently, the correlation between an eruption described in such accounts and a particular volcanic eruption or resulting lava flow or tephra deposit may be obvious (e.g., Etna, 1669; Krakatau, 1883), and in some cases, prehistoric oral accounts (legends) can be related to independent tephrostratigraphic and archaeological evidence (e.g., Cronin and Neall, 2000). More often, however, correlation between recorded eruption years and tephrostratigraphy is not a straightforward one. This is especially the case where there are numerous active volcanoes in isolated terrains, such as in Iceland. Historical accounts kept by community officials or by annalists (e.g., *Annales Islandici*; Storm, 1888) have provided dates for >100 eruptions in Iceland during the last nine centuries. The accounts vary from a few words to detailed descriptions of eruption sites and areas affected by tephra fall. Written sources have documented the dates of all historical eruptions of Hekla volcano, the majority of the historical eruptions of volcanoes below the ice caps of Vatnajökull and Myrdalsjökull, and the eruptions of other volcanoes in Iceland and off the southwest and northern coasts. The quality of the sources is variable, with contemporary documentation being the most reliable (Larsen and Eiríksson, 2008). Eruption locations are sometimes clearly described, but many are inaccurate, some are obviously wrong or vague or the location is not specified, making correlation between written and tephrostratigraphic records difficult. A detailed, spatially extensive stratigraphic record is therefore a prerequisite to validate eye-witness accounts, helped by documenting field and laboratory characteristics. Pairs or series of tephra layers are favored over individual layers. Thus, the first Icelandic tephra layers securely dated were those of the Hekla (AD 1693), Katla (AD 1721), and Hekla (AD 1766) eruptions (Thorarinsson, 1967).

Fission-track dating

Fission-track (FT) dating of tephra beds using coexisting zircon and glass phases has played a significant role in advancing the

stratigraphy and chronology of Plio-Pleistocene successions in New Zealand, United States of America, Canada, Japan, and East Africa. However, application of the zircon FT technique to the dating of distal tephra was often hindered by low abundance of zircons, fine grain size, and presence of detrital grains. Volcanic glass shards are typically the most abundant component of such sediments and commonly the only datable phase. Up until recently, hydrated glass shards were seldom used to determine the age of Quaternary volcanic rocks because glass has a propensity for partial fading (natural size reduction) of fission tracks even at ambient temperatures and because of the time-consuming nature of establishing correction procedures. During the late 1970s, FT dating of volcanic glass shards provided the first numerical ages for tephra layers in New Zealand and United States of America. However, the glass ages could not be considered reliable unless the glass was checked for FT fading because the latter would lead to lower apparent ages. The isothermal plateau fission track (ITPFT) dating technique applied to hydrated glass shards was developed in the late 1980s (Westgate, 1989). This technique corrected for the effects of partial track fading by heating at 150 °C for 30 days. Accurate and precise ITPFT-ages were obtained consistently on tephra-derived glass provided that it had experienced simple thermal histories. The precision is better than 10% of the mean age for most samples, but can be improved considerably if the sample is counted several times by different operators. A weighted mean can then be derived.

The glass-ITPFT technique is particularly well suited to the dating of distal rhyolitic tephra, permitting a direct and detailed comparison of stratigraphic and chronologic data. Since then, the ITPFT technique has been widely applied to a variety of proximal to distal depositional settings and environments in the United States of America (e.g., Preece et al., 1999), western Canada, New Zealand (e.g., Alloway et al., 1993; Pillans et al., 2005; Shane et al., 1996), Japan, East Africa, and Indonesia (e.g., Alloway et al., 2004; Chesner et al., 1991). The use of this dating technique, especially when combined with other tephrochronological tools, has led to important advances in Quaternary stratigraphy and regional to interregional correlation.

^{40}Ar – ^{39}Ar dating

The $^{40}\text{Ar}/^{39}\text{Ar}$ technique provides an opportunity to obtain high-precision ages on young tephra deposits. Notable applications include the precise dating of tephra within hominid-bearing deposits in north-eastern Africa (e.g., Walter, 1994; Walter et al., 1991). In New Zealand, the $^{40}\text{Ar}/^{39}\text{Ar}$ dating technique has played a pivotal role in determining the chronology and dynamics of caldera-forming ignimbrite eruptions in the central Taupo Volcanic Zone over the last 1.6 Ma (Houghton et al., 1995) and in the older Coromandel Volcanic Zone (Briggs et al., 2005). Initial $^{40}\text{Ar}/^{39}\text{Ar}$ studies performed step-heated Ar-release experiments on bulk feldspar and sanidine separates, and laser fusion analysis on mineral subsamples to obtain age spectra and isochrons. A precision of about ± 30 ka for Quaternary-aged plagioclase and better than 1% on sanidine was obtained. However, the use of single-crystal laser fusion (SCLF) has revealed the common occurrence of contaminants and partially degassed relict crystals with individual pumice blocks sometimes containing crystals

with a range of ages (e.g., Gansecki et al., 1998). Such studies have shown that an accurate age for some tephra deposits can be assessed only by grain-specific techniques. The SCLF $^{40}\text{Ar}/^{39}\text{Ar}$ technique has been applied to tephra deposits as young as the Holocene age (Chen et al., 1996). Because of the small volume of Ar involved in SCLF, K-rich phases are commonly required. This limits its application to calc-alkaline volcanic provenances. Unfortunately, the technique has not been widely applied to distal tephra beds because of their fine grain size and crystal-poor character.

Dating Enveloping or Entrained Materials

Radiocarbon dating

Since its advent around 1950, radiocarbon (^{14}C) dating has been the dominant technique for dating tephra layers deposited within the past 50–60 ka. To provide accurate ages, ^{14}C dating has three key requirements.

1. Samples containing carbon, including wood, leaves, twigs, charcoal, peat, lake sediment, pollen, or shells, must be contemporaneous, or very nearly so, with the eruption that generated the tephra, that is, terrestrial or aquatic plant material metabolizing at the time of the eruption was killed by that event and either (a) entrained in the deposit (as in the case of charcoal or carbonized wood within hot pyroclastic density currents) (Figure 15(a)), or (b) directly overlain and hence buried by the deposit, or (c) directly underlain by the deposit (e.g., tephra fallout on a peat bog that subsequently resumed its growth) (Figure 15(b)).
2. Samples must contain sufficient ^{14}C – in excess of background levels – to enable an age to be derived, otherwise a ‘minimum’ age is the result.
3. Radiocarbon ages obtained must be calibrated, that is, converted to calendar years rather than ^{14}C years which, because of fluctuations in ^{14}C productivity in the atmosphere and other factors, do not exactly match calendar (solar) years. Calibrations on terrestrial samples spanning the past 50 cal ka are made using an internationally agreed set of calibration curves (IntCal09) based on cross-matching ^{14}C ages against tree-ring-based sequences to 12.4 cal ka ago, and laminated marine and lake sediments and U-series-dated corals, speleothems, and foraminiferal deposits (Reimer et al., 2009).

Ages are measured by two methods: (1) *radiometric laboratories* synthesize either benzene (C_6H_6) or CO_2 and count the decay of ^{14}C in these over time (called beta counting) and (2) *accelerator mass spectrometric (AMS) laboratories* synthesize graphite (or CO_2) and count the relative number of ^{14}C atoms directly (the isotope ratio of ^{14}C relative to that of ^{13}C or ^{12}C is measured). Both methods have been widely used for dating tephtras, the AMS system increasingly so where sample sizes are tiny or where separate components, such as pollen grains within sediments, are dated to help acquire better age estimates for associated tephra layers (e.g., Newnham et al., 2007a).

Radiocarbon ages on tephtras are subject potentially to a wide range of errors (see Walker, 2005). However, only laboratory counting statistics (normally increased by a laboratory ‘error multiplier’) are reported as part of the age measurement,

conventionally in years BP and at 1 standard deviation (1σ). The ‘present’ refers to 1950. Reported 1σ errors, representing 68% probability, may range from ± 15 years for high-precision ages on optimal material of late Holocene age to ± 2 ka or more for late Pleistocene samples or for samples containing small amounts of carbon. Although ^{14}C laboratories provide ages routinely with a 1σ error term, there is a trend now for ages on tephtras to be evaluated at 2σ , representing 95% probability (Hajdas et al., 2006; Lowe et al., 2008; Walker, 2005).

Wiggle-match dating using dendrochronology or depositional age modeling using peat/lacustrine stratigraphy

Wiggle-match dating is the process of matching a time-series of ^{14}C samples, usually from wood or peat or lacustrine deposits, to the shape of ^{14}C calibration curves to derive a more precise age estimate for an eruption event than would be otherwise obtainable from ^{14}C ages or dendrochronology. The matching process is called flexible age-depth modeling where it involves sediments (Lowe, 2011). Hogg et al. (2003) fitted five contiguous ^{14}C ages on decadal blocks from a 50-year-old tree killed



Figure 15 (a) Carbonized log entrapped within pyroclastic density current deposits (Taupo Ignimbrite) near Taupo, North Island, New Zealand. Such logs have provided numerous ^{14}C ages for the Taupo eruption that generated the ignimbrite (Froggatt and Lowe, 1990; Wilson, 1993; Hogg et al., 2012). Photograph: Brent Alloway. (b) Distal rhyolitic tephra-fall layer (white) preserved within peat near Hamilton, North Island, New Zealand. The age of the mid-Holocene ash layer can be obtained by dating the encapsulating peat using the ^{14}C method (Hogg et al., 1987). Photograph courtesy of Maria Gehrels, University of Plymouth.

by the Kaharoa eruption of Mt. Tarawera, New Zealand, to the wiggles of the high-precision Southern Hemisphere calibration curve. Once all five ages were 'locked' onto the curves, it enabled the estimation of the death of the tree, and thus the date of the Kaharoa eruption, at AD 1314 $\pm$ 12 from the date calculated for its outermost ring. The eruption of the Greek volcano of Thera (Santorini) in the Aegean Sea, and the associated widespread tephra, was dated to 1627–1600 BC using wiggle-match dating of wood from a chair burnt during the eruption (Galimberti et al., 2004). In a study of the Kaipo bog sequence in New Zealand, improved ages on late-glacial and Holocene tephra layers intercalated with peat deposited since ca. 18 cal ka were derived using a Bayesian-based age-depth model wiggle-matched against the IntCal04 curves (Hajdas et al., 2006; Lowe et al., 2008). Wiggle-match dating/flexible age-depth modeling have also been conducted on tephra-bearing lake sediments in United States of America, central Europe, and Scandinavia (Blockley et al., 2008; Kuehn et al., 2009; Wohlfarth et al., 2006).

Luminescence dating

Thermoluminescence (TL) and optically stimulated luminescence (OSL) dating have become widely used in Quaternary studies. Many applications are still experimental, however, and specialists disagree as to the reliability of these methods with certain materials and over certain time ranges. Direct luminescence dating of tephra beds is uncommon but some ages have been determined and compared with independent numerical ages to show that purified fine-silt-sized glass can be reliably dated over the age range 5–450 ka (Berger, 1992). More often, luminescence dating is applied to fine-grained mineral fractions of bracketing paleosol-loess deposits. In one such example, TL dates from late Pleistocene loess successions in eastern Washington, USA, indicated that the ages of tephra interbeds were significantly older than those implied by earlier ^{14}C dating, suggesting a much longer eruptive history (>50 ka) for Mt. St. Helens than previously accepted (Berger and Busacca, 1995; Figure 16). Tephra ages have also been determined by TL dating of bracketing loess from sites in Alaska (Berger et al., 1996), New Zealand (Berger et al., 1992), and Japan (Watanuki et al., 2005). For example, one of the Sheep Creek group of tephra, SCt-F in the Fairbanks area, was dated at 190 ka (Berger et al., 1996; Westgate et al., 2008). In North Island, New Zealand, the Rotoehu Ash was OSL dated at ca. 45 cal ka BP (Lian and Shane, 2000), in approximate agreement with previous ages deduced from several other methods (Danisik et al., 2012).

Luminescence dating complements ^{14}C dating over its ~60-ka age range, and allows ages to be obtained when no carbon is available. Generally, there is good agreement between the two techniques (e.g., Demuro et al., 2008) but discrepancies in others (e.g., Lowe et al., 2010). Large sets of luminescence ages from the same stratigraphic horizon (from which a measure of reproducibility of luminescence dating can be derived) are not commonly obtained. Beyond the ~60-ka limit to ^{14}C dating, TL and OSL dating provide an extension into the late Quaternary where methods based on radioactive decay (K/Ar, U-series, and fission track) can be applied for comparison. One of the most comprehensive multimethod comparisons was applied to Rangitawa Tephra (Pillans et al.,

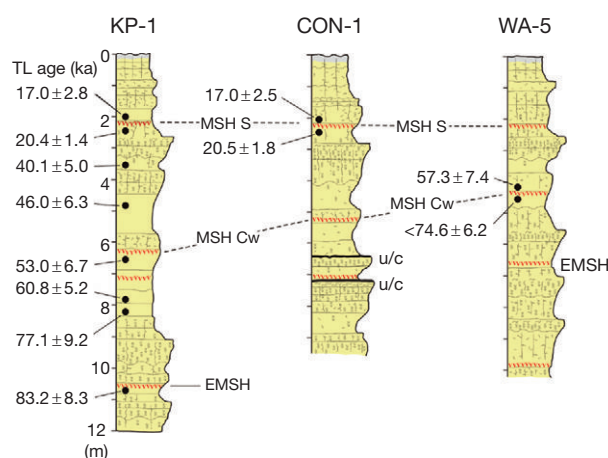


Figure 16 TL ages (ka) for the samples of loess and Mt. St. Helens tephra (Set S, MSH S; Set C, MSH-C; Early Mt. St. Helens, EMSH). Reproduced from Berger GW and Busacca AJ (1995) Thermoluminescence dating of Late Pleistocene loess and tephra from eastern Washington and southern Oregon, and implications for the eruptive history of Mount St. Helens. *Journal of Geophysical Research* 100: 22, 361–22, 374. Copyright 1995, American Geophysical Union.

1996) – a 350-ka marine and terrestrial interregional tephra marker bed in New Zealand dated by luminescence, fission track (glass and zircon), K/Ar, Ar/Ar methods, and orbital tuning (Figure 17).

Paleomagnetism

In many studies, numerical ages derived from glass-ITPFT, SCLF Ar/Ar, and TL methods are used in conjunction with magnetostratigraphic information. This combination is helpful because the age of geomagnetic polarity transition zones is well known for late Cenozoic time and provides an independent check on the reliability of numerical age estimates. The geomagnetic polarity sequence registered in the bracketing sediment is itself a useful signature for correlation purposes. For example, the identification of relatively short-lived polarity subchrons, such as the Jaramillo Normal Subchron (1.07–0.99 Ma), has greatly facilitated the construction of a tephrostratigraphy in the basins of the southern North Island (Pillans et al., 2005; Shane et al., 1996) and loess cover-beds in the Fairbanks region of central Alaska (see Preece et al., 1999; Figure 18). Magnetic measurements on ignimbrites and widespread ash-fall deposits in Japan provided the means both to correlate these units and to link the source volcanic area with marine and other sequences over a distance of 1000 km (Hayashida et al., 1996).

Applications of Tephrochronology

Case Studies

Volcanic hazard assessment and eruption frequency from long tephra records

Tephrochronological studies help to establish the recurrence intervals of major episodes of volcanic activity. Detailed tephrostratigraphic records provide a basis for predicting the likely frequency and magnitude of volcanic activity in the

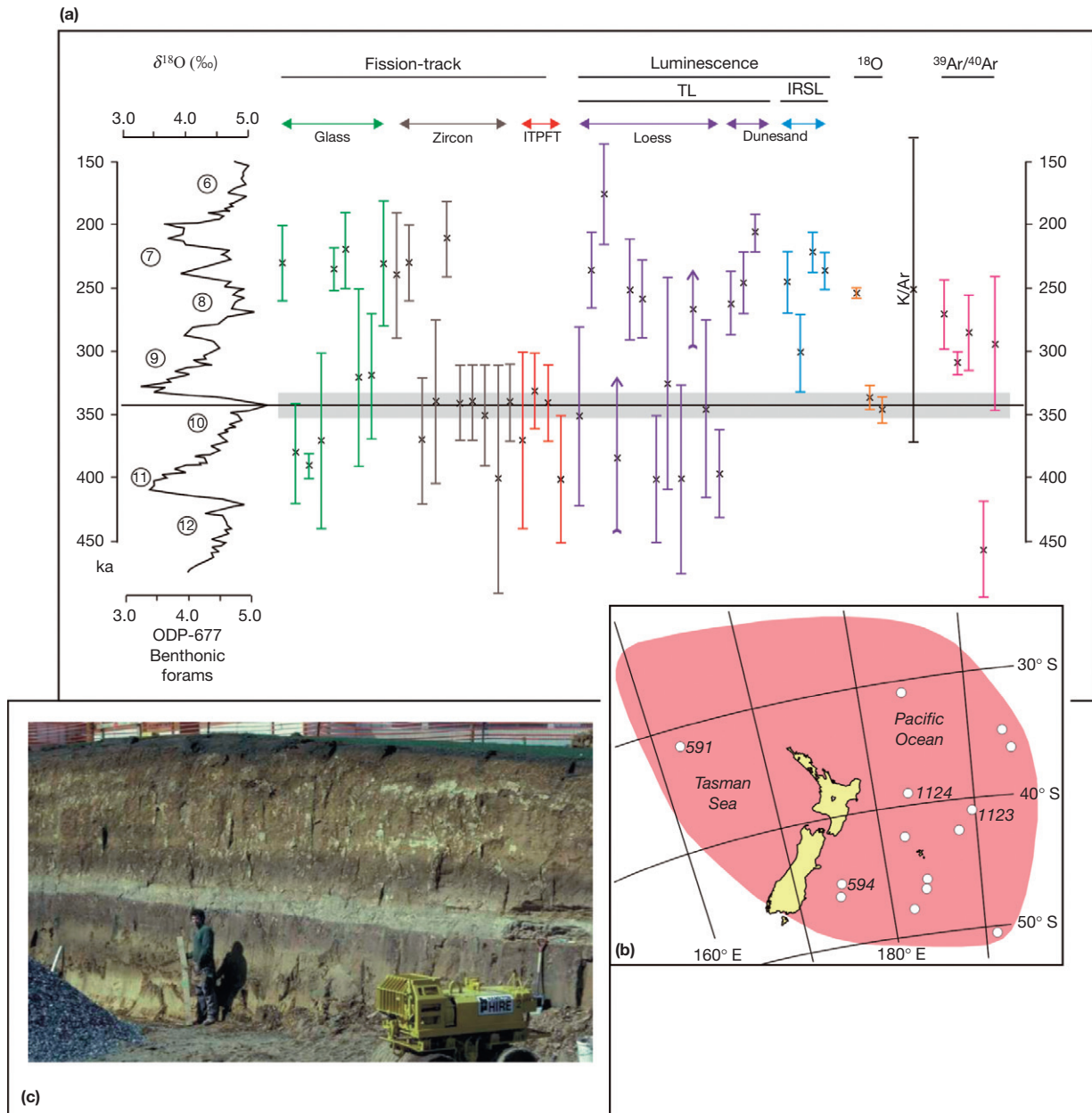


Figure 17 (a) Summary of numerical age determinations on Rangitawa Tephra, with 1 SD error bars plotted (reproduced from Pillans B, Froggatt P, Kohn BP, et al. (1996) Multi-method dating comparison for mid-Pleistocene Rangitawa Tephra, New Zealand. *Quaternary Science Reviews (Quaternary Geochronology)* 15: 641–653). (b) Minimum known extent of Rangitawa Tephra (see also Matthews et al., 2012). (c) Rangitawa Tephra showing up as the prominent white layer (just above person's head) in section of weathered tephra and paleosols (darker horizons) in Hamilton, North Island, New Zealand. Rangitawa Tephra (0.35 Ma) overlies a buried paleosol older than 0.78 Ma (Lowe et al., 2001). Photograph by David Lowe.

future. A notable example is the tephra hazard zonation map for Mt. St. Helens that was produced 2 years before the May 1980 eruption (Crandell and Mullineaux, 1978). The zones reflected the thickness-distance relations of three past tephra-fall episodes representative of eruption magnitudes and dominant wind patterns and speeds in the Pacific north-west. The boundaries of the zones were placed arbitrarily at distances of 30, 100, and 200 km from the volcano and were intended to aid users in locating areas with respect to potential tephra thicknesses. This zonation map should probably have been

extended to cover distal areas >200 km downwind in light of the wide distribution and impacts of thin tephra-layer accumulation that ensued. Nowadays, numerical and probabilistic models have been developed for use in estimating the extent and thickness of potential tephra falls and may aid in delimiting tephra-hazard zones (e.g., Bonadonna et al., 2002; Hurst and Smith, 2004). The inputs to the models include wind data and seasonality, eruption rate and duration, column height, grain-size distribution, and settling velocities.

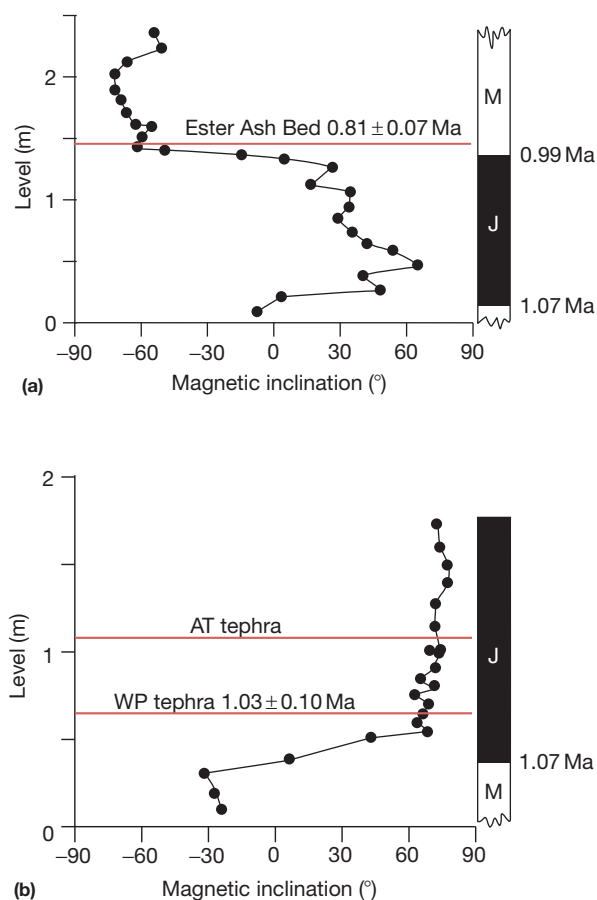


Figure 18 (a) Magnetic inclination changes in the basal Gold Hills Loess which contains the Ester Ash Bed, at Ester Island, 4 km west of Fairbanks, Alaska. (b) Magnetic inclination changes in loess containing the AT and WP tephra beds at Ester II. M, Matuyama reversed chron; J, Jaramillo normal subchron. Reproduced from Preece SJ, Westgate JA, Stemper BA, and Pewe TL (1999) Tephrochronology of late Cenozoic loess at Fairbanks, central Alaska. *Geological Society of America Bulletin* 111: 71–90.

Long tephra records can contribute significantly to an understanding of eruption frequency and tephra dispersal. A notable example is the tephrostratigraphic record from Lago Grande di Monticchio in southern Italy (Wulf et al., 2004). Here, 340 distal tephra layers are preserved within 72.5 m of lacustrine sediments deposited over the last 100 ka (Figure 19). Most tephras ($n=313$) preserved within this long core are derived from volcanic eruptions of the Campanian province, which still represents an area of volcanic risk for the Naples metropolitan area. Other tephras are related to high-explosive events of Roman and Sicilian-Aeolian volcanoes ($n=17$) or cannot be correlated with any distinct volcanic source ($n=10$). In addition to this continuous, high temporal resolution record of climate and environmental change from a maar lake, the intercalated tephra sequence provides new information about the eruptive tempo of explosive Italian volcanoes and dispersal of tephra.

Until recently, assessments of volcanic hazards for the city of Auckland (New Zealand's largest metropolitan area

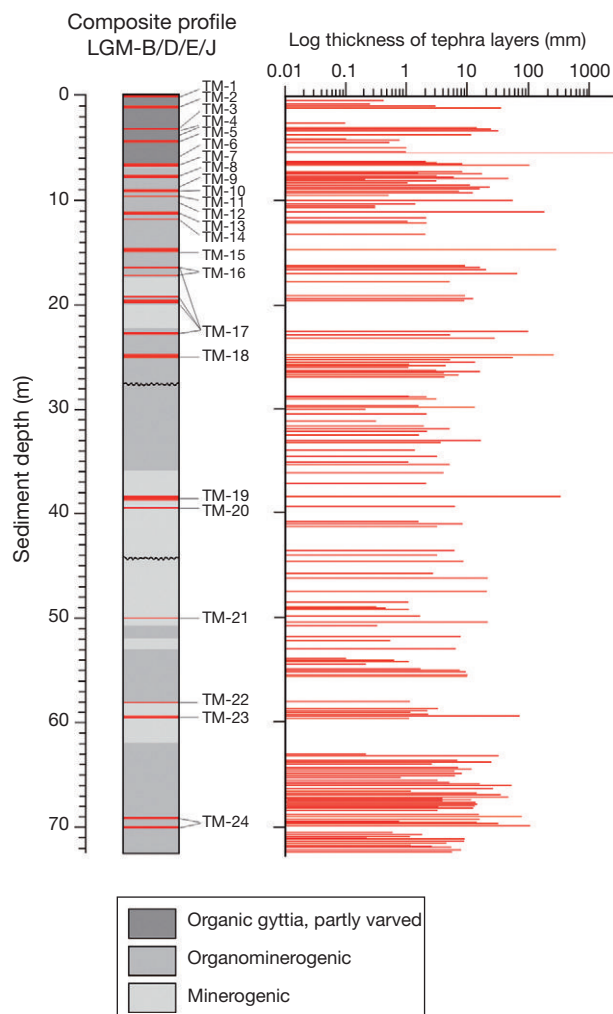
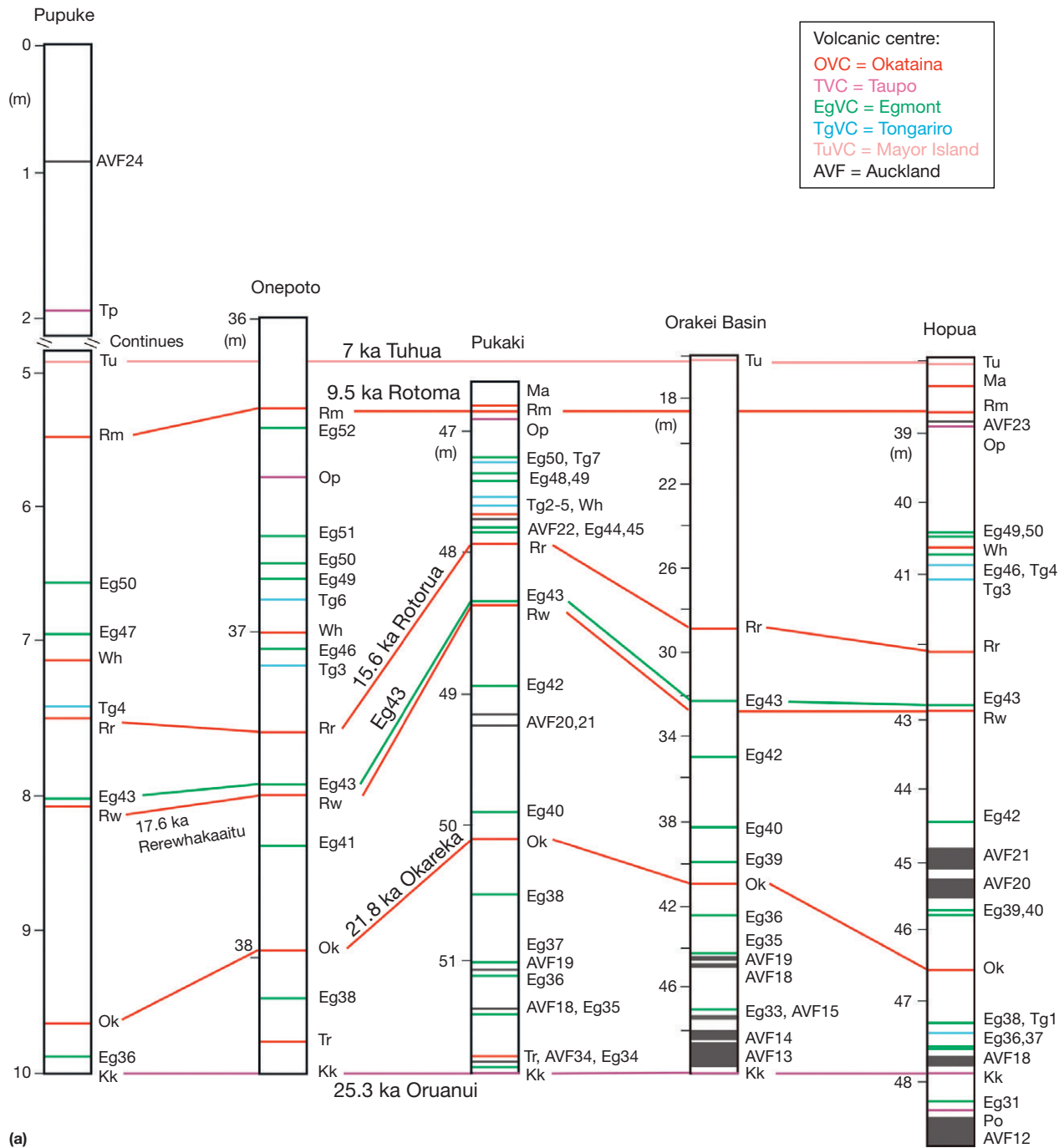


Figure 19 Sedimentology and tephrostratigraphy of composite profile LGM-B/D/E/J from Lago Grande di Monticchio. Reprinted from Wulf S, Krami M, Brauer A, Keller J, and Negendank JFW (2004) Tephrochronology of the 100 ka lacustrine sediment record of Lago Grande di Monticchio (southern Italy). *Quaternary International* 122: 7–30, Copyright 2004, with permission from Elsevier.

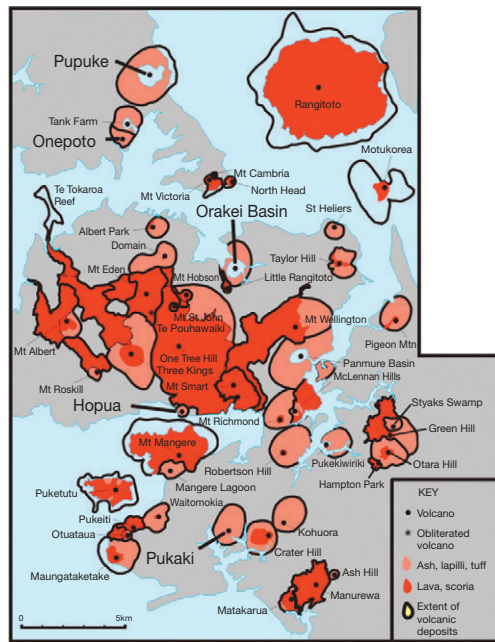
with ~1.5 million people) focused entirely on the threat from the local basaltic Auckland Volcanic Field (AVF), such threats being assessed most recently using a Bayesian-statistical framework (Lindsay et al., 2009). The threat of inundation by tephra fall from distant rhyolitic and andesitic centers in central and western North Island (~200–260 km away) was not considered in any regional or local authority plan until Newnham et al. (1999) demonstrated that such a threat existed. Since then, new long records obtained from Auckland maars have further highlighted the region's vulnerability to frequent impact by ash fall, distal, and local (Sandiford et al., 2001; Shane and Hoverd, 2002; Molloy et al., 2009; Figure 20). Source volcanoes for distal tephra include the rhyolitic Taupo (TVC) and Okataina (OVC) volcanic centers, the peralkaline Tuhua Volcanic Centre (TuVC) of Mayor Island, and the andesitic Egmont (EgVC) and Tongariro (TgVC) volcanic centers. A total of 106 different tephra layers from local and distant



(a)

Figure 20 (a) Post-ca. 25 ka tephra preserved in the Pupuke, Onepoto, Pukaki, Orakei, and Hopua cores, Auckland, New Zealand. Tie lines are shown for key tephra marker beds that provide age control. Sources of distal tephras are color coded. Modified from Molloy C, Shane P, and Augustinus, P (2009) Eruption recurrence rates in a basaltic volcanic field based on tephra layers in maar sediments: Implications for hazards in the Auckland volcanic field. *GSA Bulletin* 121: 1666–1677. (b) Map of the Auckland Volcanic Field (AVF) showing the locations of Pupuke, Onepoto, Pukaki, Orakei, and Hopua craters, where long sediment cores have been retrieved. (c) Pukaki maar crater in Mangere, South Auckland. The crater is nearly circular and has an area of 1.76 km². The crater floor is 3.67 m above mean high sea level. Pukaki was the first maar crater in the Auckland area to be cored. A 52.5-m long sediment core extending back to ca. 25 cal ka BP was retrieved at the indicated site. (d) The location of distant rhyolitic and andesitic centers in central and western North Island, New Zealand. (b, c, and d) Modified from Alloway BV, Lowe DJ, Barrell DJA, et al. (2007) Toward a climate event stratigraphy for New Zealand over the past 30 000 years (NZ-INTIMATE project). *Journal of Quaternary Science* 22: 9–35.

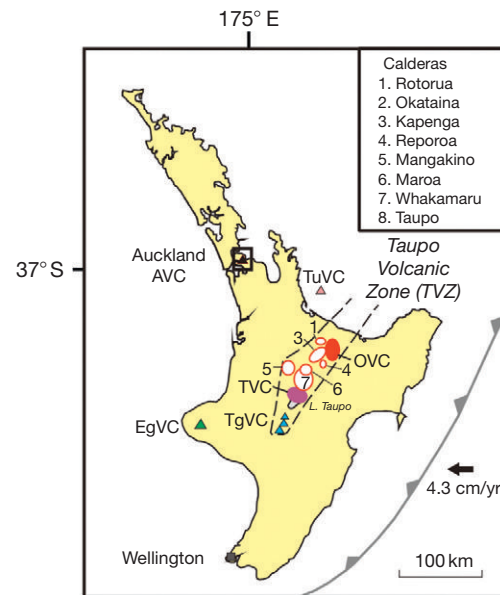
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(b)



(c)



(d)

Figure 20 (Continued)

volcanoes (>0.5 mm thick) in the last 80 ka have been recorded in the Auckland maars, an average recurrence of ca. 755 years (Molloy et al., 2009). These comprise 52 events from EgVC (1 per 1.5 ka), 24 AVF events (1 per 3.5 ka), 21 Taupo Volcanic Zone rhyolite (TVC and OVC) events (1 per 3.8 ka), 7 TgVC events (1 per 11.4 ka), and 2 TuVC events (1 per 40 ka). Unrecognized cryptotephra (<0.5 mm) are also likely to be present in the cores. Such cryptotephra also potentially have significant implications for hazard assessments as exemplified by the 1995–96 eruptions of Ruapehu Volcano. Tephra from these eruptions disrupted aviation at Auckland but did not leave a macroscopic layer. Most recently, Newnham et al. (2010) identified an increase in respiratory-related mortality in Auckland (also Hamilton) as a possible consequence of the effects of diffuse, fine-grained ash fallout from the 1996 Mt. Ruapehu eruptions. The Auckland study indicates that volcanic hazard assessment based on distal tephra records may be applicable to many large cities

elsewhere that are not currently considered at risk from volcanism (Magill and Blong, 2005; Magill et al., 2006; Newnham et al., 1999).

Geomorphic and landscape reconstruction

Well-developed regional and interregional tephrochronological frameworks can contribute to an understanding of many environmental issues. The main advantage that tephrochronology brings to interdisciplinary paleoenvironmental research is the unparalleled precision and confidence with which disparate terrestrial and marine records can be correlated and dated. An excellent example of this occurs in the North Island of New Zealand which has a history of large volcanic eruptions spanning the entire Quaternary. The biggest eruption in the Taupo Volcanic Zone in the past 60 ka was the Oruanui (or Kawakawa) eruption (Wilson 2001; Figure 21(a) and 21(b)) that occurred ca. 25.3 cal ka BP (Vandergoes et al., in press) during the extended Last Glacial Maximum (LGM). With a

bulk volume of $\sim 1200 \text{ km}^3$ (530 km^3 as magma), the eruption formed the modern Taupo caldera and generated an ignimbrite and very widespread tephra-fall deposits. The preservation of this tephra in loess deposits, in alluvial/colluvial gully fills, and on river terraces, and its absence from unstable sites, have permitted a detailed assessment of geomorphic activity and landscape response during the LGM (Pillans et al., 1993; Manville and Wilson, 2004; Figure 21(c)). Pollen data for sediments associated with products of the Oruanui eruption, and those ^{14}C dated in the range ca. 17–23 ^{14}C ka BP, have also assisted in determining LGM vegetation in northern, central, and southern parts of the North Island and the West Coast of the South Island (Newnham et al., 2007b).

A similar integrated paleoenvironmental study has been achieved for another key tephra in New Zealand, Rerewhakaaitu Tephra (Figure 21(a)), erupted from Mt. Tarawera volcano ca. 17.6 cal ka at a time of major environmental change. Terrestrial evidence, based on fluvial aggradation and downcutting relationships, loess accumulation rates, paleovegetation patterns, and buried soil development and mineralogy, show that marked amelioration of climate occurred at around the

time the Rerewhakaaitu Tephra was deposited (Newnham et al., 2003) (Figures 21(c) and 22). Similarly, marine evidence shows major changes in accumulation rates of sediment and aeolian quartz and abundance of various marine organisms, while foraminiferal oxygen and carbon isotope records indicate that the arrival of the meltwater signal occurred close to, or just after, deposition of the Rerewhakaaitu Tephra. This tephra is thus an important onshore–offshore regional marker for the Last Termination and onset of postglacial environmental conditions.

Glacier advance and retreat, dating of ice

Glaciers in volcanic regions, and the landforms they generate, have been subjected to numerous tephra falls in the Holocene. Tephra layers deposited on glaciers or ice caps are buried in the accumulation areas and may be preserved in the ice. Tephra layers of known age can thus be used to date the ice and contribute greatly to the understanding of glacier dynamics and how glaciers respond to climate and environment.

Iceland is a good example because of frequent eruptive activity and numerous glaciers ranging from small cirque

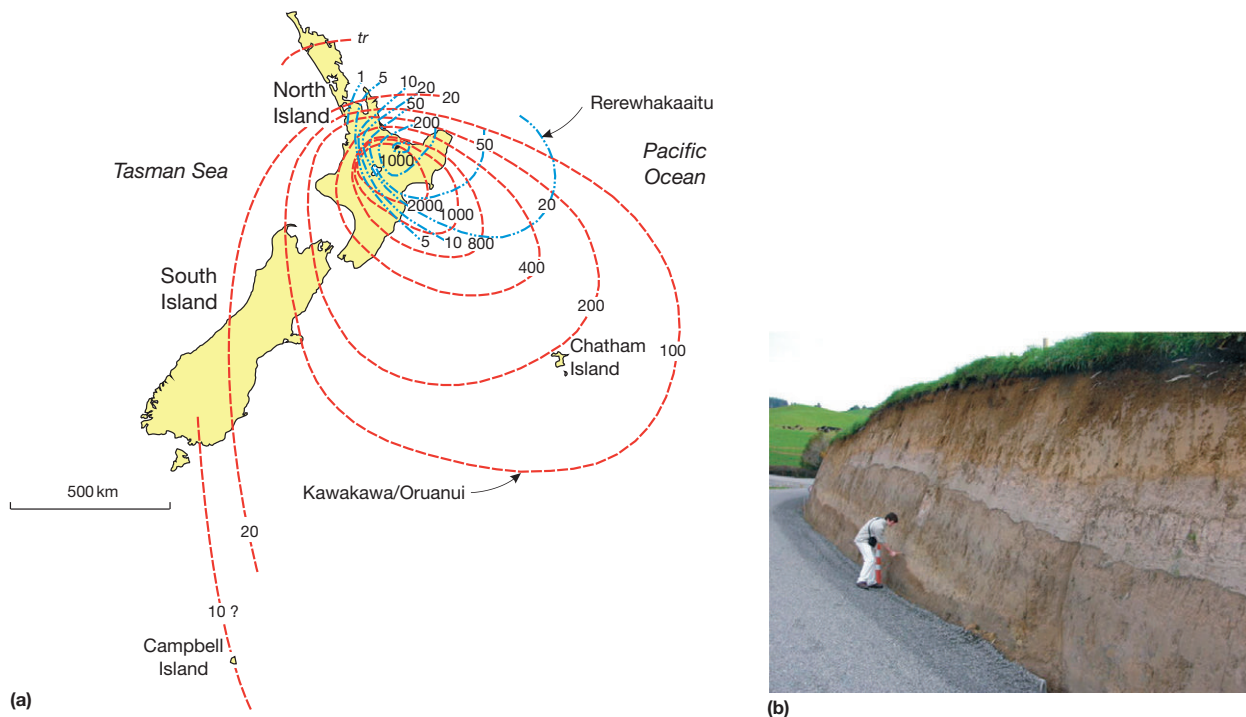


Figure 21 (a) Map of distributions of Kawakawa/Oruanui (red dashed line) and Rerewhakaaitu (blue dot-dash line) tephra. Isopachs in millimeters. Adapted from Campbell IB (1986) New occurrences and distribution of Kawakawa Tephra in South Island, New Zealand. *New Zealand Journal of Geology and Geophysics* 29: 425–435; Wilson CJN (2001) The 26.5 ka Oruanui eruption, New Zealand: An introduction and overview. *Journal of Volcanology and Geothermal Research* 112: 133–174; Newnham RM, Eden DN, Lowe DJ, and Hendy CH (2003) Rerewhakaaitu Tephra, a land-sea marker for the last termination in New Zealand, with implications for global climate change. *Quaternary Science Reviews* 22: 289–308, with permission from Elsevier. (b) Section near Putaruru, North Island, showing Kawakawa/Oruanui tephra as prominent pinkish-white marker bed ~ 0.5 m thick in middle of section overlying tephric loess and paleosol. Pale tephra bed near base (by person) is ca. 50 cal ka old Rotoehu Ash, which overlies a dark paleosol on loess. Photograph: David Lowe. (c) Summary diagram for the period ca. 30–10 ^{14}C ka BP in central and southern North Island, New Zealand note age scale is in ^{14}C yrs. Adapted from Pillans B, McGlone M, Palmer A, Mildenhall D, Alloway BV, and Berger G (1993) The Last Glacial Maxima in central and southern North Island, New Zealand: A paleoenvironmental reconstruction using the Kawakawa Tephra Formation as a chronostratigraphic marker. *Paleogeography, Paleoclimatology, Paleoecology* 101: 283–304; Manville V and Wilson CJN (2004) The 26.5 ka Oruanui eruption, New Zealand: A review of the roles of volcanism and climate in the post-eruptive sedimentary response. *New Zealand Journal of Geology and Geophysics* 47: 525–547.

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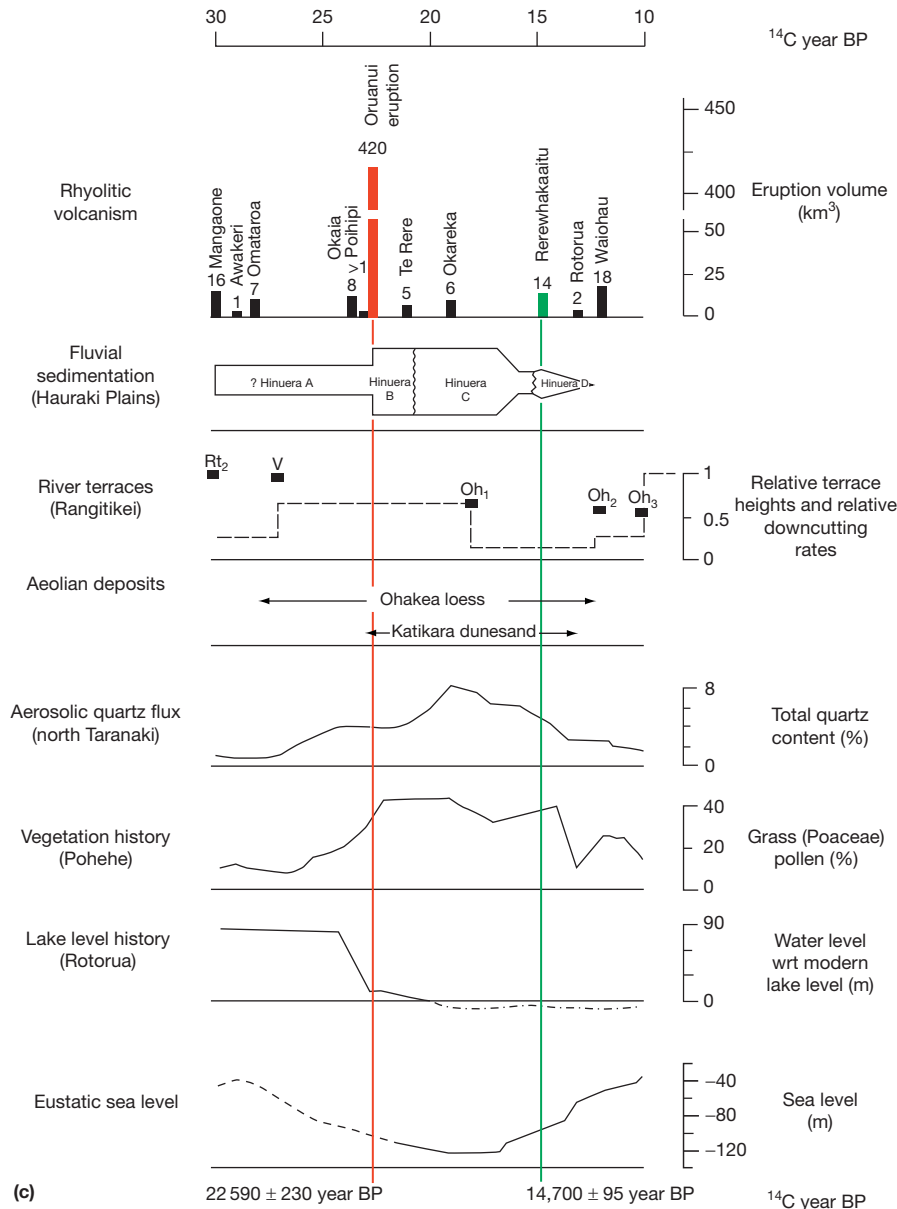


Figure 21 (Continued)

glaciers to ice caps such as 8000 km² Vatnajökull. Early work includes Thorarinsson's studies of an outlet glacier of the Langjökull ice cap where he demonstrated that the most advanced terminal moraine since the retreat of the inland ice in the early Holocene postdated the AD 1693 tephra layer from Hekla volcano (Thorarinsson, 1949) and related to the Little Ice Age (LIA). Dugmore (1989b) used tephrochronology to date the maximum advance of an outlet glacier of Myrdalsjökull ice cap in southern Iceland to the mid-Holocene (>4.5 cal or 4 ^{14}C ka BP) and a gradual retreat to its LIA moraine and the present position. His findings contradicted the opinion that most outlet glaciers in Iceland had their maximum extension during the LIA. Further, tephrochronological dating of moraines in Iceland has confirmed mid-Holocene glacier advances in northern and central Iceland that were locally more extensive than the LIA advances (e.g., Dugmore and Kirkbride, 2001;

Gudmundsson, 1997). Tephrochronology has also been applied to reconstruct the retreat of the inland ice including glacier advances/retreats and the formation of ice-dammed lakes in northern Iceland (Norddahl and Hafliðason, 1992), glacial advances in southern Iceland (Kirkbride and Dugmore, 2008), and ice-cap response to Holocene climate change (Kirkbride and Dugmore, 2006). Tephra layers preserved in the Vatnajökull ice cap have been used to date the ice and to record the eruption history of the volcanoes below the ice cap, revealing that the volcanic activity is intermittent, with a periodicity of 130–140 years (Larsen et al., 1998).

Traces of tephra deposited on the Greenland ice sheet have served as marker horizons in the Greenland ice cores since Palais et al. (1991) were the first to find glass particles in the GISP2 ice core from a known volcanic eruption – that of Öræfajökull AD 1362, erupted from southeast Iceland. Previously, acid fallout

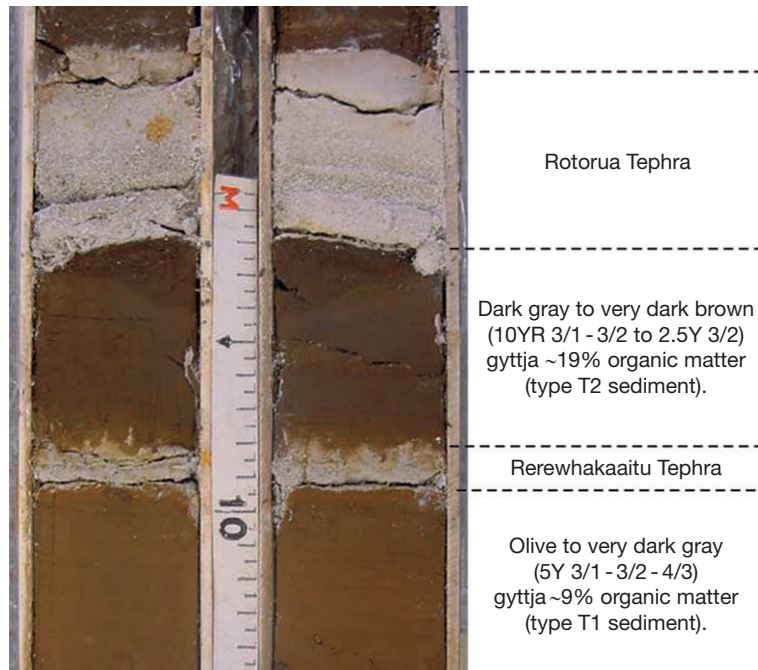


Figure 22 Close-up photo of core from Lake Rotorua (Hamilton, New Zealand) showing darkening in sediment color just above ca. 17.6 cal ka BP Rerewhakaaitu Tephra that reflects an increase in organic content because of reafforestation and amelioration in climate. Note: Colors are for fresh sediment (core oxidized shortly after opening). Reproduced from Newnham RM, Eden DN, Lowe DJ, and Hendy CH (2003) Rerewhakaaitu Tephra, a land-sea marker for the last termination in New Zealand, with implications for global climate change. *Quaternary Science Reviews* 22: 289–308.

from volcanic eruptions had served as time markers in the ice (as noted above), in particular the large acidity peaks relating to the Laki fissure eruptions of 1783–85 (Hammer et al., 1980). Glass shards from several eruptions of known age have now been found in the Greenland ice cores, including glass from the Vatnaöldur tephra (~AD 874), as noted earlier, and the Laki eruptions (Fiaccio et al., 1994). Another example is the detection of the widespread early Holocene Saksunarvatn tephra, dated to ca. 9 ^{14}C ka BP on terrestrial organic material (Mangerud et al., 1986), as a visible ash layer in the GRIP and GISP2 ice cores (Figure 23) where it has been assigned an age of 10.3 ice-core ka (Grönvold et al., 1995; Zielinski et al., 1997). This tephra provides an important tephrochronological linkage between terrestrial, marine, and ice-core records in the North Atlantic region. Tephrae have also been found in Antarctic ice cores (Dunbar et al., 2003; Kyle et al., 1981; Narcisi et al., 2005; Zielinski, 2000).

Sedimentary basin history

The recognition of tephra has been an essential tool for dating lake sediments, shoreline levels, and reconstructing drainage evolution particularly in Plio-Pleistocene pluvial lakes of the Great Basin in south-western United States (e.g., Reheis et al., 2002; Williams, 1994). Precise identification of tephra layers has also proven to be important in understanding the growth and development of sedimentary basins as well as for the correlation of glacio-eustatic sedimentary cycles across marine basins. Wanganui Basin, New Zealand, is one of several Plio-Pleistocene basin margins in the world in which sedimentation matched subsidence through much of its history, resulting in a 5-km thick sequence of predominantly shelf and shallow-water sediment (Pillans et al., 2005). Gentle



Figure 23 A 1-cm thick visible tephra horizon identified within the NGRIP ice core, Greenland. Photograph courtesy of Sigfus Johnsen, The Niels Bohr Institute, Copenhagen, Denmark.

neotectonic upwarping of the eastern margin along the plate boundary has produced on-land exposures through as many as 45 superposed cyclothem since 2.5 Ma. This succession represents the most complete, on-land, shallow-marine record of late Neogene climatic and sea-level change yet described. Twenty-eight tephrae are recognized from glass-shard composition and stratigraphic position, and were dated using magnetostratigraphy, orbitally tuned cyclostratigraphy and fission-track (ITPFT) dating (Figure 24). Many tephra deposits can be reliably correlated to distal sedimentary successions

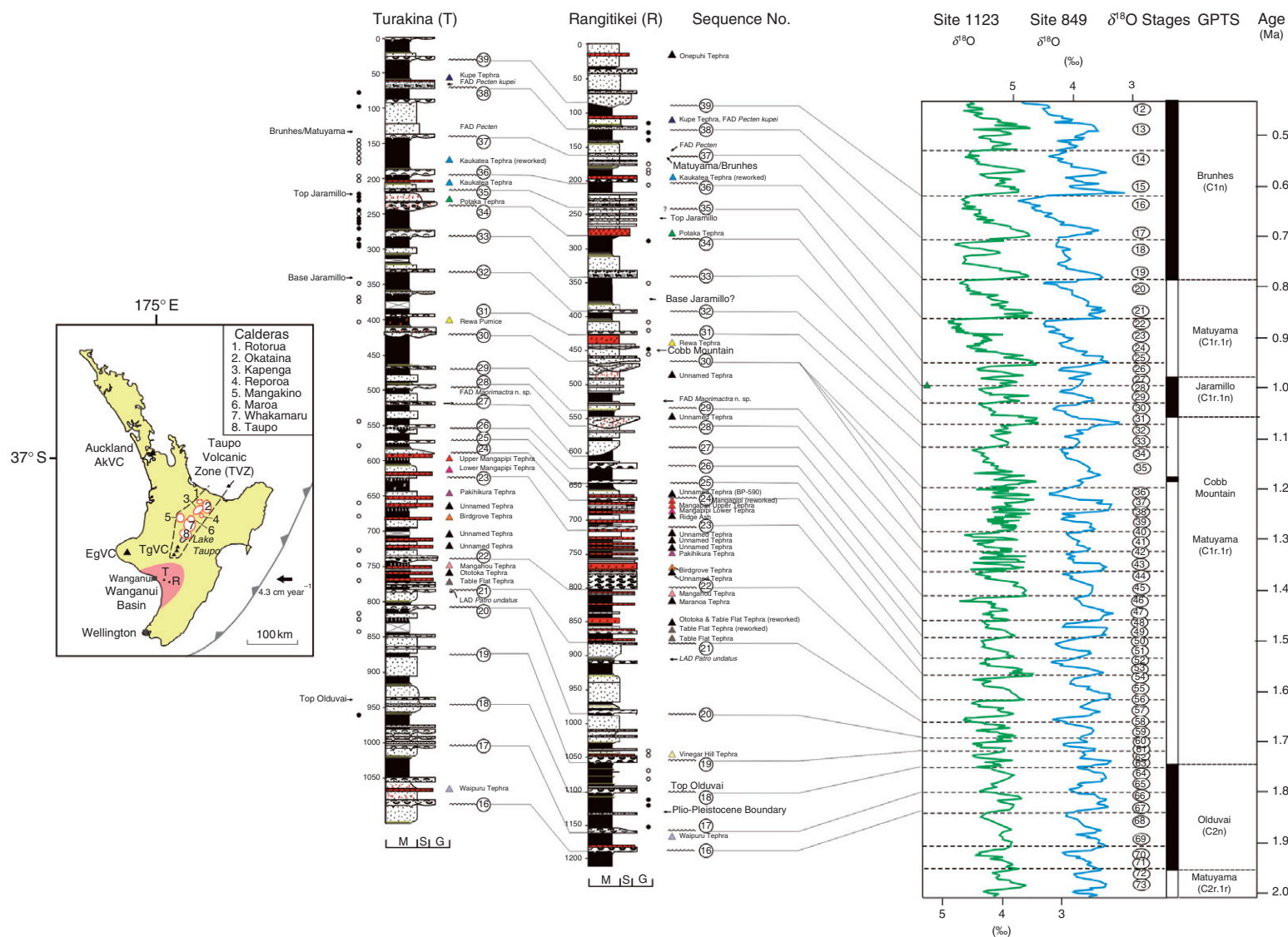


Figure 24 Composite stratigraphic columns for two major sections (Turakina and Rangitikei) in the Wanganui Basin, New Zealand, showing positions of tephra horizons, lithologies, sequence boundaries, and macrofossil LADs and FADs, as well as paleomagnetic samples and boundaries. The onshore cyclostratigraphic records (Sequences 39–16) are compared with the $\delta^{18}\text{O}$ record from Site 1123 retrieved east of New Zealand and the standard equatorial core record obtained from Site 849. LAD, last appearance datum; FAD, first appearance datum. Reproduced from Alloway BV, Pillans BJ, Naish TR, Carter LC, and Westgate JA (2005) Onshore-offshore correlation of Pleistocene rhyolitic eruptions from New Zealand: Implications for TVZ eruptive history and paleoenvironmental construction. *Quaternary Science Reviews* 24: 1601–1622.

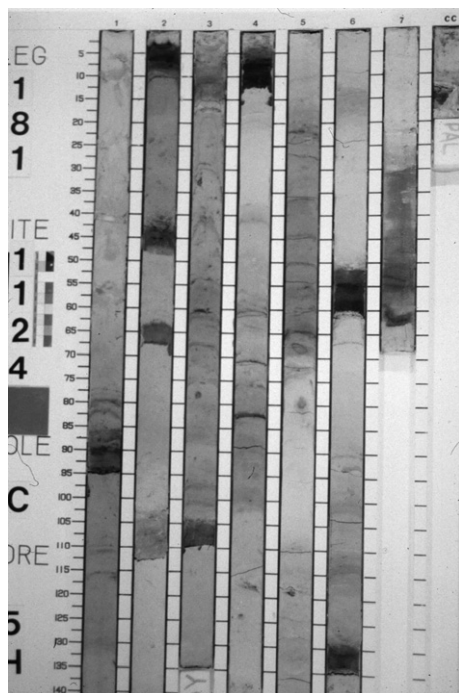


Figure 25 Pinkish-gray tephra beds (dark colored) in light greenish-gray clay to silty clay bearing nannofossil ooze (light-colored) from ODP Site 1124, east of North Island, New Zealand. Photograph courtesy of Lionel Carter, NIWA, Wellington, New Zealand.

throughout the North Island, and in marine cores east of New Zealand obtained by the Ocean Drilling Programme (ODP) (Alloway et al., 2005; Figure 25). The orbitally tuned chronology for the ODP-cores, calibrated by numerical ages on tephra and magnetostratigraphy, enhances inter-regional correlation and provides an important framework for future paleoenvironmental reconstructions.

Hominid evolution

Most of the hominid remains and associated artifacts from the East African rift system have been found in Plio-Pleistocene volcanoclastic sediments. Consequently, the chronology of hominids in East Africa is largely constrained by a combination of regional correlation of tephra layers and isotopic dating (e.g., McDougall, 1985; Walter, 1994; WoldeGabriel et al., 2005). Some eruptions were of sufficient magnitude or duration to generate widespread tephra deposits that occur as a series of dated marker horizons throughout the fossil-bearing deposits of East Africa. Although many of the larger eruptive events have been directly dated, the ages of many other tephra are only constrained by indirect stratigraphic interpolation between dated horizons. The geochemical compositions of volcanic glasses from each eruption are distinctive and provide a definitive means to establish broad tephrostratigraphic correlations. This same tephrostratigraphic approach has been used to extend the East African tephra correlations into the continuous and well-dated marine cores from the western Indian Ocean and Gulf of Aden that contain several macroscopic Plio-Pleistocene volcanic-ash layers, nearly 1000 km east and northeast of hominid localities in Ethiopia and Kenya (Brown et al., 1992; deMenocal and Brown, 1999;

Figure 26). Major-element compositions of glass shards extracted from these marine sediments were used to establish precise tephrostratigraphic correlations into the fossil-bearing East African sedimentary sequences. Moreover, controversy concerning the ages of key tephra marker beds relating to significant temporal junctures in early hominid evolution could be readily tested against correlatives independently dated in orbitally tuned marine records.

Archaeology

Tephrochronology has increasing application in archaeological studies because they form isochronous horizons enabling the correlation of equivalent-aged successions. A good example is the Late Bronze Age explosive eruption of Thera (Santorini) which had a devastating impact on prosperous Cycladic maritime culture and contributed to a major change in the course of Western history (McCoy and Heiken, 2000). The eruption, dated ca. 1627–1600 BC (see the section ‘Chronology’ above), generated a widespread ash found in deep-sea cores in the Aegean, eastern Mediterranean, and Black Sea, as well as in lakes in eastern Turkey (Figure 27). Scattered patches of the ash have also been identified on islands east of Thera and at archaeological sites on Crete – usually reworked into building construction or within Late Minoan destruction debris. Traces have also been identified in Nile delta sediments and in soils of Turkey.

Another example where tephrochronology has advanced archaeological studies is in New Zealand where a key rhyolitic eruptive, Kaharoa Tephra (KT), has helped resolve the controversial timing of initial Polynesian settlement (Lowe et al., 2000; Newnham et al., 1998). Because the settlement occurred so recently in human history – now known to have been within the last 750 years – dating methods including ^{14}C were problematic and two conflicting theories for the timing of settlement emerged, one ‘early’ and the other ‘late.’ KT was erupted in the austral winter of AD 1314 $\pm$ 12 from Mt. Tarawera volcano and dispersed over >30,000 km² of northern and eastern North Island (Hogg et al., 2003). It provides a unique ‘settlement datum’ in three ways. Firstly, the initial deforestation signals, the simultaneous decline in forest trees, and sustained increase in bracken fern (*Pteridium*) and charcoal (as determined from analysis of pollen and spores from sediments containing KT), occurred only a few decades before KT deposition, that is, from late in the thirteenth century to ca. AD 1300 (Figure 28). Secondly, tree-derived seeds nibbled by the Pacific rat, *Rattus exulans*, an introduced predator that accompanied the early eastern Polynesian voyagers to New Zealand, are all <750 calendar years old and, apart from a single exception, have been found only above the KT layer, not below it (Wilmschurst and Higham, 2004; Wilmschurst et al., 2008). Finally, no cultural remains or artifacts have yet been found beneath KT at any archaeological sites (Lowe, 2011; Lowe et al., 2000). This means that such sites must be younger than AD 1314 $\pm$ 12, in close agreement with ages for the earliest archaeological remains in New Zealand including the oldest known at Wairau Bar (South Island) dated using moa eggshell at AD 1288–1300 (Higham et al., 1999). Collectively, the archaeological and paleoenvironmental evidence, linked tephrochronologically by the KT datum, indicates that the earliest Polynesian settlement of New Zealand was ca. AD 1250–1300 (see also

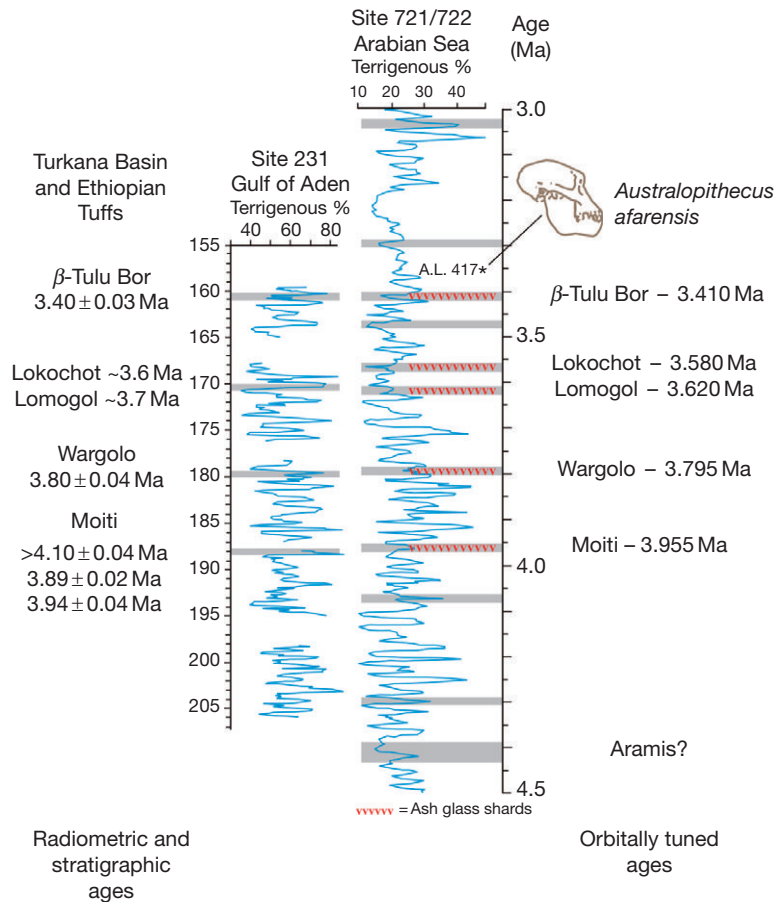


Figure 26 Comparison of radiometric and stratigraphic (interpolated) ages for East African tuffs between 4.0 and 3.4 Ma and their orbitally tuned ages derived from the marine sediment chronostratigraphy at ODP sites 721 and 722 in the Arabian Sea. Reproduced from deMenocal PB and Brown FH (1999) Pliocene tephra correlations between East African hominid localities, the Gulf of Aden, and the Arabian Sea. In: Agusti J, Rook L, and Andrews P (eds.) *Hominoid Evolution and Climatic Change in Europe*, vol. 1. Cambridge: Cambridge University Press.

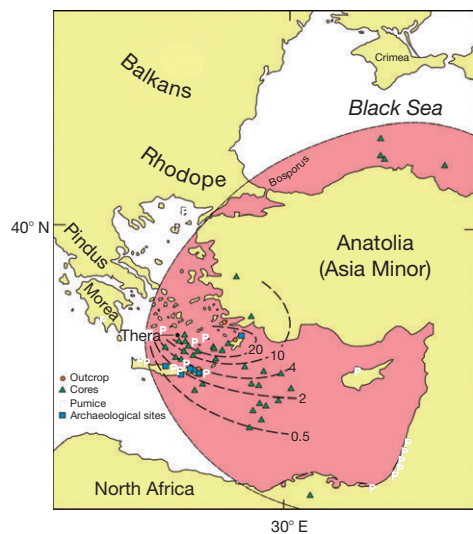


Figure 27 Regional distribution of tephra from the Thera eruption (shaded area) in the Late Bronze Age. Tephra thickness isopachs are in centimeters. Symbols identify sites where tephra has been identified in cores (triangles) and at archaeological sites (squares). Reproduced from McCoy FW and Heiken G (2000) *The Late-Bronze Age Explosive Eruption of Thera (Santorini), Greece: Regional and Local Effects*, pp. 43–70. Boulder, CO: Geological Society of America, Special Paper 345.

Wilmschurst et al., 2008). The association of KT with early human settlement is similar to the Vatnaöldur tephra or *Landnámsslag* (settlement layer) in Iceland of the mid-870s AD that coincides closely with anthropogenic impacts relating to permanent Nordic settlement.

Conclusions

Since the early pioneering work of Thorarinsson and others, the value of tephra in providing time-parallel marker horizons or isochrons is now well understood. Tephra are routinely detected and identified in both visible and cryptotephra forms, and are used in a diverse range of disciplinary fields including stratigraphy, sedimentology, geomorphology, archaeology, and paleoenvironmental reconstruction over a range of time scales. Tephrochronology is also an essential tool for establishing the frequency/periodicity of volcanic activity and for assessing volcanic hazards.

A number of significant recent advances have contributed to the growing usage of tephrochronology. First, the application of high-precision analytical techniques (e.g., EMP analysis and LA-ICP-MS) has made it possible to obtain detailed major- and trace-element compositions from individual glass shards and for 'fingerprinting' individual tephra beds or tephra

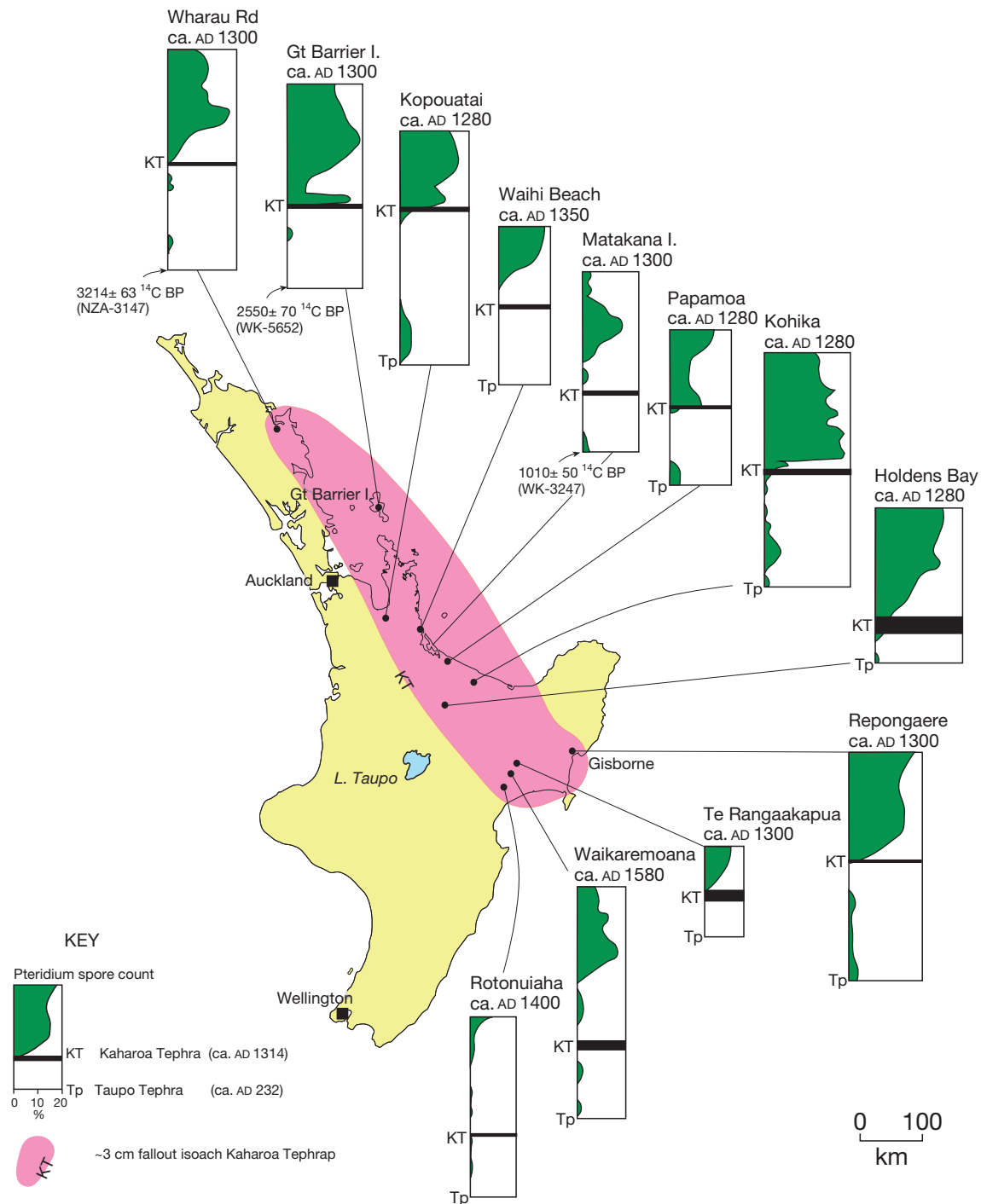


Figure 28 Summary diagram of 12 bracken (*Pteridium*) spore profiles, North Island, New Zealand, containing the Kaharoa Tephra settlement datum (KT). In most profiles, the deforestation signal (increase in *Pteridium* and charcoal, decline of tall trees) occurs at around or after the deposition of KT, but in four (Kopouatai, Papamoa, Kohika, Holden's Bay) it occurs just before, by perhaps a few decades. Ages at the top of each profile are approximate estimates ($\pm$ few decades) for the start of deforestation. Taupo Tephra (Tp) provides a predeforestation datum in most profiles (Hogg et al., 2012); in three other sites, ¹⁴C ages are given. Reproduced from Lowe DJ, Newnham RM, and McCraw JD (2002) Volcanism and early Maori society in New Zealand. In: Torrence R and J. Grattan (eds.) *Natural Disasters and Cultural Change*, pp. 126–161. London: Routledge.

successions of similar mineralogy and/or provenance. The second significant advance has been the detection of cryptotephra well beyond the limits of their previously known visible distribution. Studies of cryptotephra offer considerable

potential for further developing integrated tephrochronological frameworks over much larger areas than previously possible, such as that being developed for Europe and the NE Atlantic region. However, a future challenge of working with

cryptotephra will be quantifying the glass-shard concentration and relating this to ash densities observed and predicted in the atmosphere during an eruptive event (e.g., Eyjafjöll 2010 eruption). Third, the development of the ITPFT dating method has enabled ages to be obtained on many distal tephra that previously were undated because their main component, glass, was regarded as an unreliable medium. Finally, perhaps the most exciting development is the use of tephrochronology, uniquely, to effect more precise correlations between marine, ice-core, and terrestrial records. This application holds the key to testing the reliability of high-precision correlations between sequences and current theories about the degree of synchronicity of climate change at regional to global scales.

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See also: Dendrochronology; Fission-Track Dating; Geomagnetic Excursions and Secular Variations; K/Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ Dating. **Luminescence Dating:** Thermoluminescence. **Quaternary Stratigraphy:** Pedostratigraphy; Sequence Stratigraphy. **Radiocarbon Dating:** Conventional Method.

References

- Abbott PM and Davies SM (in press) Volcanism and the Greenland ice-cores: the tephra record. *Earth-Science Reviews*.
- Allan ASR, Baker JA, Carter L, and Wysoczanski RJ (2008) Reconstructing the Quaternary evolution of the world's most active silicic volcanic system: Insights from a 1.65 Ma deep ocean tephra record sourced from the Taupo Volcanic Zone, New Zealand. *Quaternary Science Reviews* 27: 2341–2360.
- Alloway BV, Pillans BJ, Naish TR, Carter LC, and Westgate JA (2005) Onshore-offshore correlation of Pleistocene rhyolitic eruptions from New Zealand: Implications for TVZ eruptive history and paleoenvironmental construction. *Quaternary Science Reviews* 24: 1601–1622.
- Alloway BV, Pillans BJ, Sandhu AS, and Westgate JA (1993) Revision of the marine chronology in Wanganui Basin, New Zealand, based on the isothermal plateau fission-track dating of tephra horizons. *Sedimentary Geology* 82: 299–310.
- Alloway BV, Pribadi A, Westgate JA, et al. (2004) Correspondence between glass-FT and ^{14}C ages of silicic pyroclastic flow deposits sourced from Maninjau Caldera, West-Central Sumatra. *Earth and Planetary Science Letters* 227: 121–133.
- Almond PC (1996) Loess, soil stratigraphy and Aokautere Ash on Late Pleistocene surfaces in South Westland, New Zealand: Interpretation and correlation with the glacial stratigraphy. *Quaternary International* 34–36: 163–176.
- Annales Islandici 1400–1800*, I–VI. Reykjavik: Hid islenzka bókmenntafélag, 1922–1987.
- Austin WEN, Wilson LJ, and Hunt JB (2004) The age and chronostratigraphical significance of North Atlantic Ash Zone II. *Journal of Quaternary Science* 19: 137–146.
- Berger GW (1992) Dating volcanic ash by use of thermoluminescence. *Geology* 20: 11–14.
- Berger GW and Busacca AJ (1995) Thermoluminescence dating of late-Pleistocene loess and tephra from eastern Washington and Southern Oregon, and implications for the eruptive history of Mount St. Helens. *Journal of Geophysical Research* 100: 22361–22374.
- Berger GW, Pewe TL, Westgate JA, and Preece S (1996) Age of Sheep Creek tephra (Pleistocene) in central Alaska from thermoluminescence dating of bracketing loess. *Quaternary Research* 45: 263–270.
- Berger GW, Pillans BJ, and Palmer AS (1992) Dating loess up to 800 ka by thermoluminescence. *Geology* 20: 403–406.
- Birks HH, Gulliksen S, Halldason H, Mangerud J, and Possnert G (1996) New radiocarbon dates for the Vedde Ash and the Saksunarvatn Ash from western Norway. *Quaternary Research* 45: 119–127.
- Bjarnason H and Thorarinsson S (1940) Datering av vulcaniska asklager i isländsk jordmån. *Geografisk Tidsskrift* 43: 5–30.
- Blockley SPE, Bronk Ramsey C, Lane CS, and Lotter AF (2008) Improved age modelling approaches as exemplified by the revised chronology for the central European varved lake Soppensee. *Quaternary Science Reviews* 27: 61–71.
- Blockley SPE, Lane CS, Lotter AF, and Pollard AM (2007) Evidence for the presence of the Vedde Ash in central Europe. *Quaternary Science Reviews* 26: 3030–3036.
- Blockley SPE, Pyne-O'Donnell SDF, Lowe JJ, et al. (2005) A new and less destructive laboratory procedure for the physical separation of distal tephra glass shards from sediments. *Quaternary Science Reviews* 24: 1952–1960.
- Bonadonna C, Macedonio G, and Sparks RSJ (2002) Numerical modelling of tephra fallout associated with dome collapse and Vulcanian explosions: Application to hazard assessment on Monserrat. In: Druitt TH and Kokelaar BP (eds.) *The Eruption of Soufriere Hills Volcano, Monserrat, from 1995 to 1999*, vol. 21, pp. 517–537. London: Geological Society Memoirs.
- Braitseva OA, Ponomareva VV, Sulerzhitsky LD, Melekestsev IV, and Bailey J (1997) Holocene key-marker tephra layers in Kamchatka, Russia. *Quaternary Research* 47: 125–139.
- Briggs RM, Houghton BF, McWilliams M, and Wilson CJN (2005) $^{40}\text{Ar}/^{39}\text{Ar}$ ages of silicic volcanic rocks in the Tauranga-Kaimai area, New Zealand: Dating the transition between volcanism in the Coromandel Arc and the Taupo Volcanic Zone. *New Zealand Journal of Geology and Geophysics* 48: 459–469.
- Brown FH, Sarna-Wojcicki AM, Meyer CE, and Haileab B (1992) Correlation of Pliocene and Pleistocene tephra layers between the Turkana Basin of East Africa and the Gulf of Aden. *Quaternary International* 13–14: 55–67.
- Chen Y, Smith PE, Evensen NM, York D, and Lajoie KR (1996) The edge of time: Dating young volcanic ash layers with the ^{40}Ar – ^{39}Ar laser probe. *Science* 274: 1176–1178.
- Chesner CA, Pose WI, Deino A, Drake R, and Westgate JA (1991) Eruptive history of Earth's largest Quaternary caldera (Toba, Indonesia) clarified. *Geology* 19: 200–203.
- Crandell DR and Mullineaux DR (1978) Potential hazards from future eruptions of Mt. St. Helens Volcano, Washington. *US Geological Survey Bulletin* 1383-C: 26.
- Cronin SJ and Neall VE (2000) Impacts of volcanism on pre-European inhabitants of Taveuni, Fiji. *Bulletin of Volcanology* 62: 199–213.
- Cronin SJ, Wallace RC, and Neall VE (1996) Sourcing and identifying andesitic tephras using major oxide titanomagnetite and hornblende chemistry, Egmont volcano and Tongariro Volcanic Centre. *Bulletin of Volcanology* 58: 33–40.
- Danišik M, Shane P, Schmitt AK, et al. (2012) Re-anchoring the late Pleistocene tephrochronology of New Zealand based on concordant radiocarbon ages and combined $^{238}\text{U}/^{230}\text{Th}$ disequilibrium and (U-Th)/He zircon ages. *Earth and Planetary Science Letters* 349–350: 240–250.
- Davies SM, Branch NP, Lowe JJ, and Turney CSM (2002) Towards a European tephrochronological framework for termination 1 and the Early Holocene. *Philosophical Transactions Royal Society of London A* 360: 767–802.
- Davies SM, Elmquist M, Bergman J, Wohlfarth B, and Hammarlund D (2007) Cryptotephra sedimentation processes within two lacustrine sequences from West Central Sweden. *The Holocene* 17: 319–330.
- Davies SM, Larsen G, Wastegård S, et al. (2010) Widespread dispersal of Icelandic tephra: How does the Eyjafjöll eruption of 2010 compare to past Icelandic events? *Journal of Quaternary Science* 25: 605–611.
- Davies SM, Wastegård S, Rasmussen TL, et al. (2008) Identification of the Fugloyarbanki tephra in the NGRIP ice core: A key tie-point for marine and ice-core sequences during the last glacial period. *Journal of Quaternary Science* 23: 409–414.
- Delmas RJ (1992) Environmental information from ice cores. *Reviews of Geophysics* 30: 1–21.

- deMenocal PB and Brown FH (1999) Pliocene tephra correlations between East African hominid localities, the Gulf of Aden, and the Arabian Sea. In: Agustí J, Rook L, and Andrews P (eds.) *Hominoid Evolution and Climatic Change in Europe*, vol. 1. Cambridge: Cambridge University Press.
- Demuro M, Roberts RG, Froese DG, Arnold LJ, Brock F, and Bronk Ramsey C (2008) Optically stimulated luminescence dating of single and multiple grains of quartz from perennally frozen loess in western Yukon Territory, Canada: Comparison with radiocarbon chronologies for the late Pleistocene Dawson tephra. *Quaternary Geochronology* 3: 346–364.
- Dugmore AJ (1989a) Icelandic volcanic ash in late-Holocene peats in Scotland. *Scottish Geographical Magazine* 105: 169–172.
- Dugmore AJ (1989b) Tephrochronological studies of Holocene glacier fluctuations in South Iceland. In: Oerlemans J (ed.) *Glacier Fluctuations and Climate Change*, pp. 37–55. Dordrecht: Kluwer.
- Dugmore AJ and Kirkbride M (2001) Timing and significance of mid-Holocene glacier advances in Northern and Central Iceland. *Journal of Quaternary Science* 16: 145–153.
- Dunbar NW, Zielinski GA, and Voisins DT (2003) Tephra layers in the Siple Dome and Taylor ice cores, Antarctica: Sources and correlations. *Journal of Geophysical Research* 108: 2374.
- Facciolo JJ, Thordarson T, Germani MS, et al. (1994) Atmospheric aerosol loading and transport due to the 1783–84 Laki eruption in Iceland, interpreted from ash particles and acidity in the GISP2 ice core. *Quaternary Research* 42: 231–240.
- Froggatt PC and Lowe DJ (1990) A review of late Quaternary silicic and some other tephra formations from New Zealand: Their stratigraphy, nomenclature, distribution, volume and age. *New Zealand Journal of Geology and Geophysics* 33: 89–109.
- Galimberti M, Bronk Ramsey C, and Manning SW (2004) Wiggle-match dating of tree-ring sequences. *Radiocarbon* 46: 917–924.
- Ganseccki CA, Mahood GA, and McWilliams MO (1998) New ages for the climactic eruptions at Yellowstone: Single-crystal $^{40}\text{Ar}/^{39}\text{Ar}$ dating identifies contamination. *Geology* 26: 343–346.
- Gehrels MJ, Lowe DJ, Hazell ZJ, and Newnham RM (2006) A continuous 5000-year Holocene cryptotephrostratigraphic record from Northern New Zealand: Implications for tephrochronology and volcanic hazard assessment. *The Holocene* 16: 173–187.
- Grange LI (1931) Volcanic-ash showers. A geological reconnaissance of volcanic-ash showers of the central part of the North Island. *New Zealand Journal of Science and Technology* 12: 228–240.
- Grönvold K, Oskarsson N, Johnsen SJ, et al. (1995) Ash layers from Iceland in the Greenland GRIP ice core correlated with oceanic and land based sediments. *Earth and Planetary Science Letters* 135: 149–155.
- Gudmundsson HJ (1997) A review of the Holocene environmental history of Iceland. *Quaternary Science Reviews* 16: 81–92.
- Hajdas I, Lowe DJ, Newnham RM, and Bonani G (2006) Timing of the late-glacial climate reversal in the Southern Hemisphere using high-resolution radiocarbon chronology for Kaipo bog, New Zealand. *Quaternary Research* 65: 340–345.
- Hammer CU, Clausen HB, and Dansgaard W (1980) Greenland ice sheet evidence of postglacial volcanism and its climatic impact. *Nature* 288: 230–235.
- Hayashida A, Kamata H, and Danhara T (1996) Correlation of widespread tephra deposits based on paleomagnetic directions: Link between a volcanic field and sedimentary sequences in Japan. *Quaternary International* 34–36: 89–98.
- Higham TFG, Anderson AJ, and Jacomb C (1999) Dating the first New Zealanders: The chronology of Wairau Bar. *Antiquity* 73: 420–427.
- Hodder APW, de Lange PJ, and Lowe DJ (1991) Dissolution and depletion of ferromagnesian minerals from Holocene tephra in an acid bog, New Zealand, and implications for tephra correlation. *Journal of Quaternary Science* 6: 195–208.
- Hodder APW, Green BE, and Lowe DJ (1990) A two-stage model for the formation of clay minerals from tephra-derived volcanic glass. *Clay Minerals* 25: 313–327.
- Hogg AG, Higham TFG, Lowe DJ, Palmer J, Reimer P, and Newnham RM (2003) A wiggle-match date for Polynesian settlement of New Zealand. *Antiquity* 77: 116–125.
- Hogg AG, Lowe DJ, and Hendy CH (1987) University of Waikato radiocarbon dates I. *Radiocarbon* 29: 263–301.
- Hogg AG, Lowe DJ, Palmer JG, Boswijk G, and Bronk Ramsey C (2012) Revised calendar date for the Taupo eruption derived by ^{14}C wiggle-matching using a New Zealand kauri ^{14}C calibration data set. *The Holocene* 22: 439–449.
- Houghton BF, Wilson CJN, McWilliams MO, et al. (1995) Chronology and dynamics of a large silicic magmatic system: Central Taupo Volcanic Zone, New Zealand. *Geology* 23: 13–16.
- Hurst T and Smith W (2004) A Monte Carlo methodology for modelling ashfall hazards. *Journal of Volcanology and Geophysical Research* 138: 393–403.
- Kirkbride MP and Dugmore AJ (2006) Responses of mountain ice caps in Central Iceland to Holocene climate change. *Quaternary Science Reviews* 25: 1692–1707.
- Kirkbride MP and Dugmore AJ (2008) Two millennia of glacier advances from Southern Iceland dated by tephrochronology. *Quaternary Research* 70: 398–411.
- Kuehn SC, Froese DG, Carrara PE, Foit FF Jr., Pearce NJG, and Rotheisler P (2009) Major- and trace-element characterization, expanded distribution, and a new chronology for the latest Pleistocene Glacier Peak tephra in Western North America. *Quaternary Research* 71: 201–216.
- Kuehn SC, Froese DG, Shane PAR, and Intercomparison Participants (INTAV) (2011) The INTAV intercomparison of electron-beam microanalysis of glass by tephrochronology laboratories: results and recommendations. *Quaternary International* 246: 19–47.
- Kyle PR, Jerek PA, Mosley-Thompson E, and Thompson LG (1981) Tephra layers in the Byrd Station ice core, Antarctica, and their climatic importance. *Journal of Volcanological and Geothermal Research* 11: 29–39.
- Lane CS, Blockley SPE, Lotter AF, Finsinger W, Filipi ML, and Matthews IP (2012) A regional tephrostratigraphic framework for central and southern European climate archives during the Last Glacial to Interglacial transition: comparisons north and south of the Alps. *Quaternary Science Reviews* 36: 50–58.
- Larsen G and Eiriksson J (2008) Late Quaternary terrestrial tephrochronology of Iceland – Frequency of explosive eruptions, type and volume of tephra deposits. *Journal of Quaternary Science* 23: 109–120.
- Larsen G, Gudmundsson MT, and Björnsson H (1998) Eight centuries of periodic volcanism at the center of the Iceland hotspot revealed by glacier tephrostratigraphy. *Geology* 26: 943–946.
- Lian OB and Shane PAR (2000) Optical dating of paleosols bracketing the widespread Rotoehu tephra, North Island, New Zealand. *Quaternary Science Reviews* 19: 1649–1662.
- Lindsay J, Marzocchi W, Jolly G, Constantinescu R, Selva J, and Sandri L (2009) Towards real-time eruption forecasting in the Auckland Volcanic Field: Application of BET_EF during the New Zealand National Disaster Exercise 'Ruaumoko'. *Bulletin of Volcanology* 72: 185–204.
- Lowe DJ (2011) Tephrochronology and its application: A review. *Quaternary Geochronology* 6: 107–153.
- Lowe DJ and Hunt JB (2001) A summary of terminology used in tephra-related studies. *Les Dossiers de l'Archéo-Logis* 1: 17–22.
- Lowe DJ, Newnham RM, McFadgen BG, and Higham TFG (2000) Tephra and New Zealand archaeology. *Journal of Archaeological Science* 27: 859–870.
- Lowe DJ, Shane PAR, Alloway BV, and Newnham RM (2008) Fingerprints and age models for widespread New Zealand tephra marker beds erupted since 30000 years ago: A framework for NZ-INTIMATE. *Quaternary Science Reviews* 27: 95–126.
- Lowe DJ, Tippet JM, Kamp PJJ, Liddell IJ, Briggs RM, and Horrocks JL (2001) Ages on weathered Plio-Pleistocene tephra sequences, Western North Island, New Zealand. *Les Dossiers de l'Archéo-Logis* 1: 45–60.
- Lowe DJ, Wilson CJN, Newnham RM, and Hogg AG (2010) Dating the Kawakawa/Oruanui eruption: Comment on 'Optical luminescence dating of a loess section containing a critical tephra marker horizon, SW North Island of New Zealand' by R. Grapes et al.. *Quaternary Geochronology* 5: 493–496.
- Magill CR and Blong RJ (2005) Volcanic risk ranking for Auckland, New Zealand: I. Methodology and hazard information. *Bulletin of Volcanology* 67: 331–339.
- Magill CR, Hurst AW, Hunter LJ, and Blong RJ (2006) Probabilistic tephra fall simulations for the Auckland Region, New Zealand. *Journal of Volcanology and Geothermal Research* 153: 370–386.
- Mangerud J, Furnes H, and Johanssen J (1986) A 9000 year old ash bed on the Faroe Islands. *Quaternary Research* 21: 85–104.
- Manville V and Wilson CJN (2004) The 26.5 ka Oruanui eruption, New Zealand: A review of the roles of volcanism and climate in the post-eruptive sedimentary response. *New Zealand Journal of Geology and Geophysics* 47: 525–547.
- Matthews NE, Smith VC, Costa A, Durant AJ, Pyle DM, and Pearce NJG (2012) Ultra-distal tephra deposits from super-eruptions: examples from Toba, Indonesia, and Taupo Volcanic Zone, New Zealand. *Quaternary International* 258: 54–79.
- McCoy FW and Heiken G (2000) *The Late-Bronze Age Explosive Eruption of Thera (Santorini), Greece: Regional and Local Effects*, pp. 43–90. Boulder, CO: Geological Society of America Special Paper 345.
- McDougall I (1985) K-Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ dating of the hominid-bearing Pliocene-Pleistocene sequence at Koobi Fora, Lake Turkana, Northern Kenya. *Geological Society of America Bulletin* 96: 159–175.
- Molloy C, Shane P, and Augustinus P (2009) Eruption recurrence rates in a basaltic volcanic field based on tephra layers in maar sediments: Implications for hazards in the Auckland volcanic field. *GSA Bulletin* 121: 1666–1677.
- Narcisi B, Petit JR, Delmonte B, Basile-Doelsch I, and Moggi V (2005) Characteristics and sources of tephra layers in EPICA-Dome C ice record (East Antarctica): Implications for past atmospheric circulation and ice core stratigraphic correlations. *Earth and Planetary Science Letters* 239: 253–265.

- Newnham RM, Dirks KM, and Samaranyake D (2010) An investigation into long-distance health impacts of the 1996 eruption of Mt. Ruapehu, New Zealand. *Atmospheric Environment* 44: 1568–1578.
- Newnham RM, Eden DN, Lowe DJ, and Hendy CH (2003) Rerewhakaaitu Tephra, a land-sea marker for the last termination in New Zealand, with implications for global climate change. *Quaternary Science Reviews* 22: 289–308.
- Newnham RM, Lowe DJ, and Alloway BV (1999) Volcanic hazards in Auckland, New Zealand: A preliminary assessment of the threat posed by central North Island silicic volcanism based on the Quaternary tephrostratigraphical record, pp. 27–45. London: Geological Society. Special Publications 161.
- Newnham RM, Lowe DJ, Giles TM, and Alloway BV (2007a) Vegetation and climate of Auckland, New Zealand, since ca. 32 000 cal. yr ago: Support for an extended LGM. *Journal of Quaternary Science* 22: 517–534.
- Newnham RM, Vandergoes MJ, Garnett M, Lowe DJ, Prior C, and Almond PJ (2007b) Test of AMS ^{14}C dating of pollen concentrates using tephrochronology. *Journal of Quaternary Science* 22: 37–51.
- Newnham RM, Lowe DJ, McGlone MS, Wilmshurst JM, and Higham TFG (1998) The Kaharoa Tephra as a critical datum for earliest human impact in Northern New Zealand. *Journal of Archaeological Science* 25: 533–544.
- Norddahl H and Hafliðason H (1992) The Skógar tephra, a Younger Dryas marker in North Iceland. *Boreas* 21: 23–41.
- Palais JM, Taylor K, Mayewski PA, and Grootes P (1991) Volcanic ash from the 1362 AD Öræfajökull eruption (Iceland) in the Greenland ice sheet. *Geophysical Research Letters* 18: 1241–1244.
- Payne RJ and Gehrels MJ (2010) The formation of tephra layers in peatlands: An experimental approach. *Catena* 81: 12–23.
- Pearce NJG, Alloway BV, and Westgate JA (2008) Mid-Pleistocene silicic tephra beds in the Auckland region, New Zealand: Their correlation and origins based on the trace element analyses of single glass shards. *Quaternary International* 178: 16–43.
- Pearce NJG, Denton JS, Perkins WT, Westgate JA, and Alloway BV (2007) Correlation and characterisation of individual glass shards from tephra deposits using trace element laser ablation ICP-MS analyses: Current status and future potential. *Journal of Quaternary Science* 22: 721–736.
- Pearce NJG, Westgate JA, Perkins WT, and Preece SJ (2004) The application of ICP-MS methods to tephrochronological problems. *Applied Geochemistry* 19: 289–322.
- Pearce NJG, Westgate JA, Perkins WT, and Wade SC (2011) Trace-element microanalysis by LA-ICP-MS: The quest for comprehensive chemical characterisation of single, sub-10 μm volcanic glass shards. *Quaternary International* 246: 57–81.
- Pillans B, Alloway BV, Naish T, Westgate JA, Abbot S, and Palmer AS (2005) Silicic tephra in Pleistocene shallow marine sediments of Wanganui Basin, New Zealand. *Journal of the Royal Society of New Zealand* 35: 43–90.
- Pillans B, Froggatt P, Kohn BP, et al. (1996) Multi-method dating comparison for mid-Pleistocene Rangitapu Tephra, New Zealand. *Quaternary Science Reviews (Quaternary Geochronology)* 15: 641–653.
- Pillans B, McGlone M, Palmer A, Mildenhall D, Alloway BV, and Berger G (1993) The Last Glacial Maxima in central and Southern North Island, New Zealand: A paleoenvironmental reconstruction using the Kawakawa Tephra formation as a chronostratigraphic marker. *Paleogeography, Paleoclimatology, Paleocology* 101: 283–304.
- Pollard AM, Blockley SPE, and Ward KR (2003) Chemical alteration of tephra in the depositional environment: Theoretical stability modelling. *Journal of Quaternary Science* 18: 385–394.
- Preece SJ, Westgate JA, Stemper BA, and Pewe TL (1999) Tephrochronology of late Cenozoic loess at Fairbanks, Central Alaska. *Geological Society of America Bulletin* 111: 71–90.
- Pyne-O'Donnell SDF, Blockley SPE, Turney CSM, and Lowe JJ (2008) Distal volcanic ash layers in the Late glacial Interstadial (GI-1): Problems of stratigraphic discrimination. *Quaternary Science Reviews* 27: 72–84.
- Reheis MC, Sarna-Wojcicki AM, Reynolds RL, Repenning CA, and Mifflin MD (2002) Pliocene to middle Pleistocene lakes in the western Great Basin: Ages and connections. *Great Basin Aquatic Systems History: Smithsonian Contributions to the Earth Sciences*, vol. 33, pp. 53–108. Washington, DC: Smithsonian Institution Press.
- Reimer PJ, Baillie MGL, Bard E, et al. (2009) Intcal09 and Marine09 radiocarbon age calibration curves, 0–50 000 years cal BP. *Radiocarbon* 51: 1111–1150.
- Sandiford A, Alloway BV, and Shane P (2001) A 28 000–6600 cal. yr record of local and distal volcanism preserved in a paleolake, Auckland, New Zealand. *New Zealand Journal of Geology and Geophysics* 44: 323–336.
- Shane PAR (1998) Correlation of rhyolitic pyroclastic eruptive units from the Taupo Volcanic Zone, New Zealand by Fe–Ti oxide compositional data. *Bulletin of Volcanology* 60: 224–238.
- Shane PAR (2000) Tephrochronology: A New Zealand case study. *Earth-Science Reviews* 49: 223–259.
- Shane PAR, Black TM, Alloway BV, and Westgate JA (1996) Early to middle Pleistocene tephrochronology of North Island, New Zealand: Implications for volcanism, tectonism and paleoenvironments. *Geological Society of America Bulletin* 108: 915–925.
- Shane P and Hovard J (2002) Distal record of multi-sourced tephra in Onepoto Basin, Auckland, New Zealand: Implications for volcanic chronology, frequency and hazards. *Bulletin of Volcanology* 64: 441–454.
- Shane PAR, Nairn IA, Martin SB, and Smith VC (2008) Compositional heterogeneity in tephra deposits resulting from the eruption of multiple magma bodies: Implications for tephrochronology. *Quaternary International* 178: 44–53.
- Shane PAR, Smith VC, and Nairn IA (2003) Biotite composition as a tool for the identification of Quaternary tephra beds. *Quaternary Research* 59: 262–270.
- Storm G (1888) *Islandske Annaler indtil 1578. Det norske historiske Kildeskriftfond*, p. 667. Norway: Christiania p. 667.
- Thorarinsson S (1944) Tefrokronologiska studier på Island; Thjorsardalur och dess foerøedelse; Tephrochronological studies in Iceland. *Geografiska Annaler* 1–2: 1–217.
- Thorarinsson S (1949) Some tephrochronological contributions to the volcanology and glaciology of Iceland. Glaciers and climate. *Geografiska Annaler* 31: 239–256.
- Thorarinsson S (1967) The eruptions of Hekla in historical times. In: Einarsson T, Kjartansson G, and Thorarinsson S (eds.) *The Eruption of Hekla, 1947–48*, pp. 1–177. Reykjavik: Societas Scientiarum Islandica.
- Thorarinsson S (1981a) Tephra studies and tephrochronology; a historical review with special reference to Iceland. In: Self S and Sparks RSJ (eds.) *Tephra Studies; Proceedings of the NATO Advanced Study Institute: Tephra Studies as a Tool in Quaternary Research, Part of NATO Advanced Study Institutes Series; Series C, Mathematical and Physical Sciences*, pp. 1–12. Dordrecht: D. Reidel.
- Thorarinsson S (1981b) The application of tephrochronology in Iceland. In: Self S and Sparks RSJ (eds.) *Tephra Studies; Proceedings of the NATO Advanced Study Institute: Tephra Studies as a Tool in Quaternary Research, Part of NATO Advanced Study Institutes Series; Series C, Mathematical and Physical Sciences*, pp. 109–134. Dordrecht: D. Reidel.
- Turney CSM (1998) Extraction of rhyolitic component of Vedde microtephra from minerogenic lake sediments. *Journal of Paleolimnology* 19: 199–206.
- Turney CSM, Lowe JJ, Davies SM, et al. (2004) Tephrochronology of Last Termination sequences in Europe: A protocol for improved analytical precision and robust correlation procedures (a joint SCOTAV–INTIMATE proposal). *Journal of Quaternary Science* 19: 111–120.
- UK Met Office (2010) Volcanic Ash Advisory Centre, London, UK. http://www.metoffice.gov.uk/aviation/vaac/vaacuk_vag.html (28 April 2010).
- Uragami K, Yamada S, and Nagamura Y (1933) Studies on the volcanic ashes in Hokkaido. *Bulletin of the Volcanological Society of Japan* 1: 79–90.
- van den Bogaard P and Schmincke H (1985) A widespread isochronous Late Quaternary tephra layer in Central and Northern Europe. *Geological Society of America Bulletin* 96: 1554–1571.
- Vandergoes MJ, Hogg AG, Lowe DJ, et al. (in press) A revised age for the Kawakawa/Oruanui tephra, a key marker for the Last Glacial Maximum in New Zealand. *Quaternary Science Reviews*.
- Walker M (2005) *Quaternary Dating Methods*. Chichester, England: John Wiley & Sons Ltd.
- Walter RC (1994) Age of Lucy and the First family: Single crystal $^{40}\text{Ar}/^{39}\text{Ar}$ dating of the Denen Dora and lower Kada Hadar members of the Hadar Formation, Ethiopia. *Geology* 22: 6–10.
- Walter RC, Manega PC, Hay RL, Drake RE, and Curtis GH (1991) Laser-fusion $^{40}\text{Ar}/^{39}\text{Ar}$ dating of bed I, Olduvai Gorge, Tanzania. *Nature* 354: 145–149.
- Watanuki T, Murray AS, and Tsukamoto S (2005) Quartz and polymineral luminescence dating of Japanese loess over the last 0.6 Ma: Comparison with an independent chronology. *Earth and Planetary Science Letters* 240: 774–789.
- Westgate JA (1989) Isothermal plateau fission-track ages of hydrated glass shards from silicic tephra beds. *Earth and Planetary Science Letters* 95: 226–234.
- Westgate JA, Preece SJ, Froese DG, et al. (2008) Changing ideas on the identity and stratigraphic significance of the Sheep Creek tephra beds in Alaska and the Yukon Territory, northwestern North America. *Quaternary International* 178: 183–209.
- Williams SK (1994) Late Cenozoic tephrostratigraphy of deep sediment cores from the Bonneville Basin. *Geological Society of America Bulletin* 105: 1517–1530.
- Wilmshurst JM, Anderson AJ, Higham TFG, and Worthy TH (2008) Dating the late prehistoric dispersal of Polynesians to New Zealand using the commensal Pacific

- rat. *Proceedings of the National Academy of Sciences of the United States of America* 105: 7676–7680.
- Wilmshurst JM and Higham TFG (2004) Using rat-gnawed seeds to independently date the arrival of Pacific rats and humans to New Zealand. *The Holocene* 14: 801–806.
- Wilson CJN (1993) Stratigraphy, chronology, styles and dynamics of late Quaternary eruptions from Taupo volcano, New Zealand. *Philosophical Transactions of the Royal Society of London A* 343: 205–306.
- Wilson CJN (2001) The 26.5 ka Oruanui eruption, New Zealand: An introduction and overview. *Journal of Volcanology and Geothermal Research* 112: 133–174.
- Wohlfarth B, Blaauw M, Davies SM, et al. (2006) Constraining the age of Late glacial and early Holocene pollen zones and tephra horizons in southern Sweden with Bayesian probability methods. *Journal of Quaternary Science* 21: 321–334.
- WoldeGabriel G, Hart WK, Katoh S, Beyene Y, and Suwa G (2005) Correlation of Plio-Pleistocene tephra in Ethiopian and Kenyan rift basins: Temporal calibration of geological features and hominid fossil records. *Journal of Volcanology and Geothermal Research* 147: 81–108.
- Wulf S, Kraml M, Brauer A, Keller J, and Negendank JFW (2004) Tephrochronology of the 100 ka lacustrine sediment record of Lago Grande di Monticchio (southern Italy). *Quaternary International* 122: 7–30.
- Zdanowicz CM, Zielinski GA, and Germani MS (1999) Mount Mazama eruption: Calendrical age verified and atmospheric impacts assessed. *Geology* 27: 621–624.
- Zielinski GA (2000) Use of paleo-records in determining variability within the volcanism-climate system. *Quaternary Science Reviews* 19: 417–438.
- Zielinski GA, Mayewski PA, Meeker LD, et al. (1997) Volcanic aerosol records and tephrochronology of the Summit, Greenland, ice cores. *Journal of Geophysical Research* 102: 26625–26640.

RADIOCARBON DATING

Contents

Conventional Method

AMS Radiocarbon Dating

Sources of Error

Variations in Atmospheric ^{14}C

Causes of Temporal ^{14}C Variations

Calibration of the ^{14}C Record

Charcoal

^{14}C of Plant Macrofossils

Conventional Method

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Introduction

Radiometric and accelerator mass spectrometric (AMS) analyses for carbon-14 (^{14}C) are fundamentally different as the former is accomplished by measuring the radiation produced by decay of ^{14}C atoms (i.e., β -particles of $E_{\text{max}} = 156 \text{ keV}$) while AMS measures the number of ^{14}C atoms relative to the number of ^{13}C (or ^{12}C). One gram of 'modern' carbon contains approximately 6×10^{10} atoms of ^{14}C and a typical AMS will measure approximately 100 atoms s^{-1} (Fifield, 1999), thereby allowing very small samples to be measured within a short period of time. Radiometric measurement presents a different set of problems:

1. Although the number of atoms per gram is high, the natural activity is extremely low because of the long half-life (5730 years). For modern carbon, the activity is approximately $226 \text{ Bq kg}^{-1} \text{ C}$, that is, on average, only 13.56 atoms in 1 g of modern carbon decay every minute. For old samples with an age of about 50 000 years (the practical limit of the method), the activity is less than $1 \text{ Bq kg}^{-1} \text{ C}$.
2. The ^{14}C β -particles are low energy, therefore, they are not able to penetrate counter materials from the outside and consequently the source of ^{14}C has to be internal.

While AMS undoubtedly has many advantages, radiometric analysis was the only option available until the late 1970s, and it is still used by a large number of laboratories in the ^{14}C -dating community, although the actual number of ^{14}C measurements made will be dominated by the AMS technique.

Irrespective of the method employed, for accurate ^{14}C dating to be achieved, only the ^{14}C that was part of the organism when it died should be measured. Therefore, any extraneous carbon that has subsequently entered the sample (mainly from the burial environment) must be removed. In addition, the fraction that is most suitable for dating must be isolated, for

example, cellulose from wood and collagen from bone. This is achieved by a combination of chemical and physical means, the so-called sample pretreatment. A detailed discussion of pretreatment methods is beyond the scope of this article; however, Mook and Streurman (1983) provide a good starting point. Also, Table 1 lists the approximate weights of samples required for radiometric analysis.

Solid-Source Counting

Modern day radiometric ^{14}C analysis is carried out by either liquid scintillation counting (LSC) or gas proportional counting (GPC); however, the early ^{14}C -dating laboratories adopted

Table 1 Typical percentage carbon contents and weight ranges of raw samples required for radiometric ^{14}C age measurement

Material	Typical carbon content (%)	Weight (g)
Wood	50	4–20
Charcoal	70	3.5–15.0
Dry peat	50	4–25
Wet peat	5	40–200
Humus	0–5	120–1000
Bone	0–30	50–500
Carbonate	12	15–60

Note: The weight ranges are quite wide since the quality of the sample will have a bearing on the amount of carbon remaining after pretreatment and the measurement device will have a bearing on the amount of sample required, for example, small versus large GPC (see Table 3).

solid-source counting techniques, originally developed by Libby (and coworkers), the inventor of the ^{14}C -dating method (Libby, 1952).

In their screen wall counter (Anderson et al., 1951), background reduction was achieved using a 20-cm thick steel plate as passive shielding while a ring of Geiger–Müller (GM) counters arranged around the counter and operated in anticoincidence was used as an active guard. The original system required about 8.5 g carbon per sample, spread over an area of 400 cm². Although the background could be reduced to an acceptable 4.5 counts per minute (cpm), efficiency was low (approximately 5%) because of high self-absorption, and this defined the upper dating range at <25 000 years before present (BP – the time unit for reporting ^{14}C ages; see section ‘Age Calculation’). Therefore, the method was soon abandoned in favor of either LSC or GPC. Figure 1 illustrates the basic sample preparation steps for AMS, LSC, and GPC.

Liquid Scintillation Counting

Instrumentation

It has been known for over half a century that certain aromatic compounds (fluors or scintillants) can convert the absorbed energy from nuclear radiation into photons of light. In LSC, photomultiplier tubes (PMTs) are employed to convert the weak light output into electrical pulses (Figure 2). Light impinging on the photocathode causes the emission of photoelectrons, which are accelerated toward the PMT anode, but before reaching it, they impinge on a series of dynodes, each of

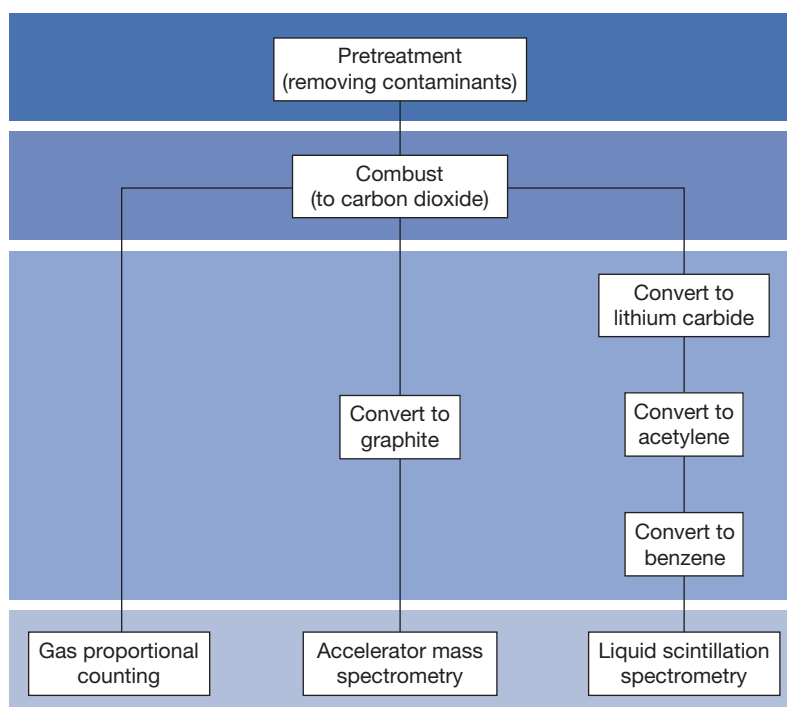


Figure 1 Main steps involved in preparing samples for ^{14}C dating by accelerator mass spectrometric analysis, liquid scintillation counting, and gas proportional counting.

which causes an electron gain such that there is an overall gain of 10^7 – 10^{10} at the anode.

A major problem with early LSC designs employing a single PMT was its thermionic noise (dark current), which constituted the majority of the instrument background. Virtually all modern day scintillation counters are based on the design of Hiebert and Watts (1953) which employs: (i) two diametrically

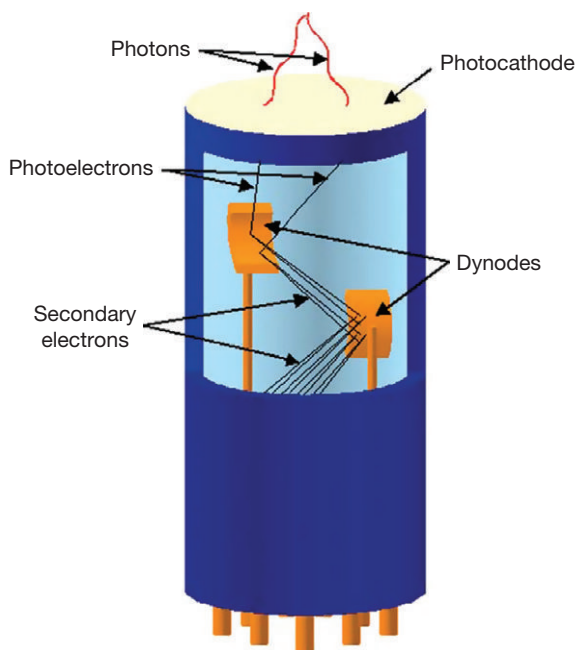


Figure 2 Schematic representation of a photomultiplier tube indicating photons impinging on the photocathode and the electron gain at each dynode.

opposed PMTs with the sample vial situated between them, (ii) fast coincidence circuitry (20 ns resolving time) to differentiate between decay events occurring in the sample vial (mainly coincident) and random thermionic noise, and (iii) summation circuitry to sum the individual PMT pulses (Figure 3).

Between the 1950s and 1980s, LSC advances were mainly modifications to existing instrumentation and included:

- Incorporation of cosmic guard detectors.
- Increasing the passive shielding.
- Reducing the PMT voltage.
- Masking PMTs or vials to minimize PMT crosstalk (light-dark current).

Low-background materials for vial (quartz or Teflon™) and vial holder (e.g., delrin) construction also brought about significant background reductions.

In the 1980s, the instrument manufacturers began to realize the potential versatility of the LSC for modern environmental analysis and combined a range of background reduction features with modern microprocessor and multichannel analyzer (MCA) technology, making the technique truly spectrometric. Thus, many modern LSCs used for ^{14}C dating contain one or more of the following background reduction features.

Enhanced passive shielding

The normal 5 cm thickness of lead was increased substantially (e.g., Figure 4) to a minimum of 10 cm; with the inclusion of cadmium and copper linings, this has the following effects on the background:

- It reduces the background from environmental γ photons associated with building and instrument construction materials.
- It reduces the 'soft' cosmic muon component.

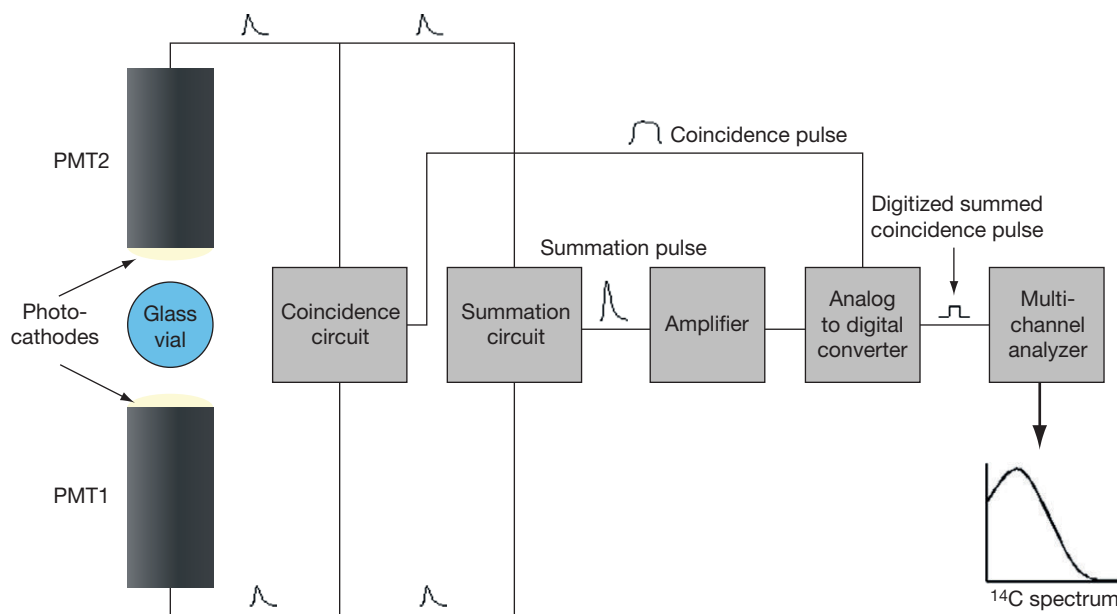


Figure 3 Schematic diagram of a liquid scintillation counter indicating the diametrically opposed photomultiplier tube (PMT), coincidence and summation circuits, and other electronic units.

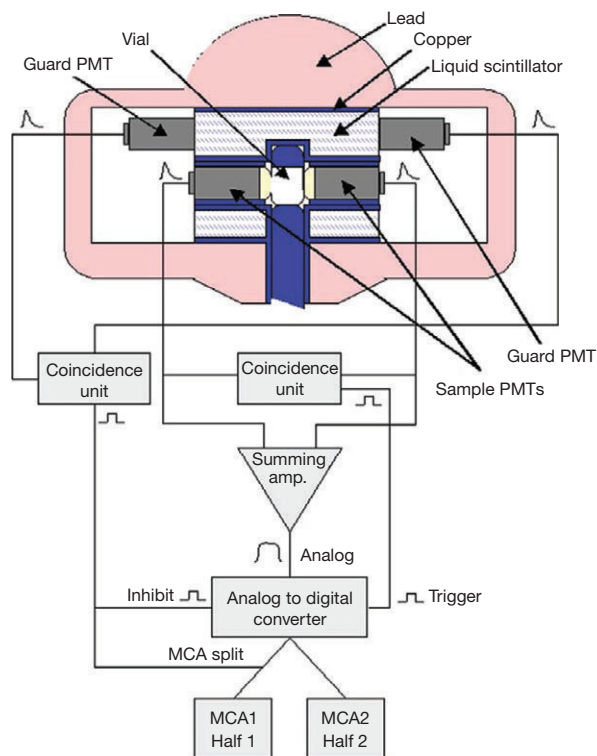


Figure 4 Schematic representation of the Quantulus liquid scintillation counting illustrating the enhanced passive shielding, guard detector (liquid scintillant), and anticoincidence circuitry. MCA, multichannel analyzer; PMT, photomultiplier tube.

- The cadmium and copper linings absorb secondary X-rays and thermal neutrons.

However, an anticoincidence shield (active guard detector) is required to reduce the high-energy γ photon and hard cosmic ray components.

Anticoincidence shielding

An anticoincidence shield typically consists of a volume of scintillating material that surrounds the sample detector assembly and its PMTs (Figure 4). The shield has its own PMTs, operating in anticoincidence with the sample PMTs to reject events detected simultaneously by the guard and sample PMTs, and events detected only by the guard (e.g., Kojola et al., 1984). The guard will reject much of the residual environmental γ radiation, the soft cosmic component, and the harder cosmic muon flux.

Pulse discrimination electronics

As an alternative or in addition to the more traditional approach of enhanced passive shielding and an anticoincidence shield, some manufacturers employ circuitry that distinguishes true β -events from background events on the basis of differences in pulse shapes. The different forms of pulse discrimination electronics include the following:

- Integration of the delayed component of each pulse and comparison with the total pulse integral to produce an

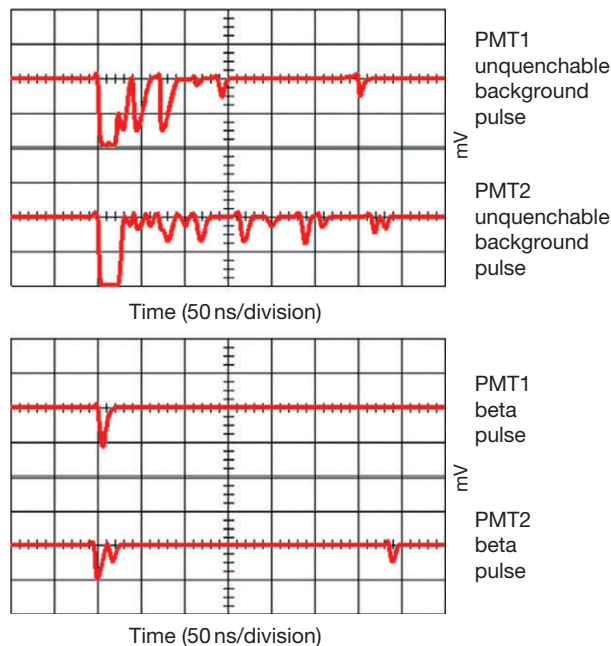


Figure 5 Comparison of pulse shapes from a nonquenchable background event and a true β -event. PMT, photomultiplier tube.

amplitude-independent parameter that relates to the pulse shape (Kaiholo, 1991) or

- A comparison of the number of after-pulses from different events (Figure 5): the unquenchable component of background is the result of low photon yield Cerenkov events consisting of a prompt component followed by a significant delayed component, comprising a burst of after-pulses. In contrast, fewer after-pulses will follow the prompt pulses from ^{14}C β -events. Valenta (1987) designed circuitry to discriminate unquenchable background events from true β -events on the basis of the number of after-pulses that follow the prompt pulse, relative to the amplitude of this prompt pulse.

A further technique compares the ratio of the pulse amplitudes (as opposed to pulse shapes) produced by the two PMTs (Kaiholo, 1991) as a discrimination criterion since photons produced in the cocktail from a radioactive decay event are distributed in a relatively even manner so that the number impinging on the photocathode of each PMT will be similar. Therefore, the pulse amplitude ratios will vary around unity. In contrast, photons produced in the vial wall or in the PMTs will show greater variation and can be discriminated out.

Quasi-active guard detector

The pulse-shape analysis features described above can also be used in conjunction with a quasi-active guard detector, so called because it has no PMTs. It consists of a modified detector assembly, typically constructed from bismuth germanate ($\text{Bi}_4\text{Ge}_3\text{O}_{12}$ or BGO) or a long fluorescence lifetime scintillating plastic. The guard surrounds the sample vial and is also optically coupled to the face of each PMT. The BGO or the slow

Table 2 Effect of various background reduction features on LSC performance

Instrument model	Background reduction features	Vial type	Volume of benzene (ml)	Optimum window parameters count time = 5000 min			
				% Efficiency	Background (cpm)	E^2V^2/B^a ($\pm 2\sigma$)	Maximum measurable age (years BP)
Packard 4530	None	Standard borosilicate glass	5	61.7	5.52	$17\,200 \pm 1000$	43500
PerkinElmer Quantulus ^b	Enhanced passive shield + liquid scintillator anticoincidence shield + PAC	Standard borosilicate glass	5	65.0	0.51	$211\,000 \pm 8000$	53200
PerkinElmer Quantulus ^b	Enhanced passive shield + liquid scintillator anticoincidence shield + PAC in low-background laboratory	Standard borosilicate glass	5	65.0	0.27	$391\,000 \pm 21000$	55700
PerkinElmer Quantulus ^c	Enhanced passive shield + liquid scintillator anticoincidence shield + PAC	Teflon™	5	84.4	0.45	$395\,000 \pm 17000$	55800
PerkinElmer Quantulus ^c	Enhanced passive shield + liquid scintillator anticoincidence shield + PAC	MG Silica	5	84.7	0.58	$310\,000 \pm 12000$	54800
Kvartett ^d	Nal(Tl) anticoincidence shield	Quartz	3.8	71.0	0.31	$235\,000 \pm 12000$	53700
Kvartett ^d	Nal(Tl) anticoincidence shield	Quartz	3.3	71.0	0.56	$98\,000 \pm 4000$	50200
Beijing Instrument ^e	Enhanced passive shield + Nal(Tl) anticoincidence shield	Teflon™	5	76.3	0.40	$364\,000 \pm 16000$	55500
PerkinElmer 2770TR/SL ^f	Pulse-shape analysis + BGO quasi-active guard	Standard borosilicate glass	4.6	70.3	0.49	$213\,000 \pm 9000$	53300
PerkinElmer 2770TR/SL ^f	Pulse-shape analysis + BGO quasi-active guard	Teflon™	4.6	70.0	0.44	$236\,000 \pm 10\,000$	53700

^aWhere E , counting efficiency; B , background count rate; BGO, bismuth germanate ($\text{Bi}_4\text{Ge}_3\text{O}_{12}$); Nal(Tl), thallium-activated sodium iodide; PAC, AC/power adapter/charger; V , sample volume.

^bKaiholu (1991).

^cHogg et al. (1991).

^dEinarsson (1992).

^eSanyuan and Yuanming (1992).

^fCook (1995).

fluor in the plastic modifies background pulses making them much longer in duration than true β -events and this aids background discrimination. With plastic guards, a secondary fluor must be used to avoid overlap of the emission spectrum of the primary fluor with the absorption spectrum of the guard.

Table 2 gives performance data for (i) a standard LSC without background reduction features and (ii) a range of modern LSCs.

Sample Analysis

Benzene synthesis

In addition to instrumentation requirements, successful ^{14}C analysis depends critically on sample conversion into a suitable chemical form for counting. C_6H_6 is now almost exclusively used as the preferred chemical form for LSC (Figure 1). Its synthesis involves the following steps:

1. Sample conversion to CO_2 and subsequent purification. (After CO_2 purification, a small subsample is removed for $\delta^{13}\text{C}$ measurement to enable a fractionation correction to be made.)
2. Conversion of CO_2 to lithium carbide by reaction with molten lithium under vacuum.
3. Addition of water to the cooled carbide to generate C_2H_2 .
4. Cyclotrimerization of the C_2H_2 to C_6H_6 using a chromium- or vanadium-based catalyst.

Counter setup

Optimized LSC performance requires a counting window that maximizes the figure of merit, E^2/B (where E =efficiency and B =background count rate). This should be set up such that the lower discriminator excludes potential hydrogen-3 (^3H) events and the upper discriminator is set close to the sample ^{14}C spectrum endpoint. Using a modern computer-based MCA and a simple Basic program, it is possible to determine such an optimum window. Normally, a window will be selected which is both close to the maximum E^2/B and in which minimal changes in counting efficiency accompany any changes in quenching (a type of balance point counting).

Sample quenching

Small variations in C_6H_6 counting efficiency occur due to both impurities in the synthesized C_6H_6 and dissolved oxygen. These interfere with the conversion of β -particle energy to photons of light, that is, they cause quenching. Reduced light output results in the production of smaller PMT anode pulses and hence a shift in the β -spectrum to lower energies with a consequent loss in efficiency. To compensate for variable efficiency, a quench curve is produced in which a number of high, known-activity ^{14}C standards are counted in an identical counting geometry to samples (same vials, C_6H_6 weight, etc.) but with increasing amounts of a quenching agent added. Counting efficiency in the optimum window is then plotted against a quench indicating parameter (QIP). This QIP is derived from the spectrum of a γ photon-emitting external standard that is housed in a small shield within the LSC and moved (mechanically or pneumatically) to a position in close proximity to the scintillation vial on request. The γ photons from the standard cause Compton electron production in the vial. These electrons are subject to the same quenching as the ^{14}C β -particles and so the changes in the ^{14}C spectrum are mirrored by changes in the external standard spectrum. The QIP is therefore a measure of the degree of quenching of the external standard spectrum. All counting efficiencies can be normalized to the same value as the least quenched standard by applying a quench factor (QF) derived from the graph of efficiency versus QIP. All samples are then counted and the QIP values noted. These QIP values are used as above to determine a QF, which is applied to the net count rate of each sample.

Counting procedure

Most laboratories store the synthesized C_6H_6 for a minimum of 3 weeks to allow any radon-222 (^{222}Rn ; a decay product of uranium-238 (^{238}U) with a half-life of 3.8 days) contained in the sample to decay, otherwise the α -particles from the ^{222}Rn and the α - and β -particles from its short-lived daughters cause

an increased count rate in the ^{14}C counting window. Then, a fixed weight of sample C_6H_6 is added to a suitable vial containing the fluor(s). Most ^{14}C -dating laboratories use only a primary fluor at relatively high concentrations (12–15 mg g^{-1} C_6H_6). A primary (typically butyl-PBD) and a secondary fluor (bis-MSB) are recommended for modern PerkinElmer (Packard) TriCarb instruments that employ nonprogrammable pulse-shape analysis (called TR-LSC). The bis-MSB is used to enhance pulse-shape discrimination by suppressing after-pulsing in ^{14}C events (Cook and Anderson, 1992).

Most laboratories that use borosilicate vials will operate a quasi-simultaneous batch counting approach in which a batch of samples is cycled many times through the instrument, each sample counting for a relatively short time in each cycle to minimize the bias on any single vial caused by background changes. A typical batch would consist of about 20–30 vials and would include:

1. Background standards in the form of scintillation-grade C_6H_6 and C_6H_6 synthesized from infinite age (with respect to ^{14}C) material such as interglacial wood, geological age carbonate, anthracite, etc.
2. Modern reference standards, that is, C_6H_6 synthesized from the internationally recognized standard SRM-4990C (Oxalic acid II) (or SRM 4990 Oxalic acid I). Their activities can be related to the equilibrium living activity of an organism.
3. C_6H_6 synthesized from samples of an unknown age.
4. C_6H_6 synthesized from samples that have an internationally agreed consensus age.
5. A high-activity ^{14}C - C_6H_6 standard to monitor counter stability.
6. A high-activity ^3H - C_6H_6 standard to monitor spill of ^3H counts into the ^{14}C energy region.

Figure 6 illustrates the layout of a typical batch of samples in an LSC, with a background standard followed by a modern reference standard and several unknown age samples. This sequence is repeated several times with other standards at the end of the batch.

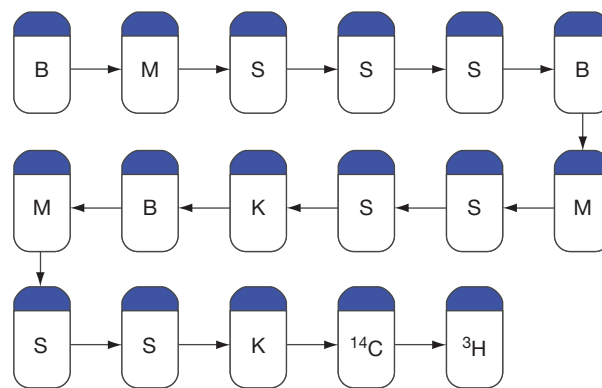


Figure 6 Layout of a typical batch of vials containing benzene for quasi-simultaneous batch counting where B, background (i.e., ^{14}C -free sample); M, modern reference standard (typically Oxalic acid II); S, unknown age sample; K, known age sample; ^{14}C , high-activity ^{14}C standard; and ^3H , high-activity ^3H standard.

After a number of cycles (e.g., 40 of 50 min counts) several checks will be undertaken of LSC stability and then (i) net count rates for modern reference standards, known age standards and unknown age samples will be determined from the gross count rates by subtracting a mean background count rate and (ii) QFs for individual standards and samples will be determined. These data will then be used to calculate ^{14}C ages (see section 'Radiometric ^{14}C Age and Error Calculations').

Modern low-background LSCs have been used to date small samples, for example, Haas and Trigg (1991) dated samples in the range 200–600 mg, however, a major drawback is the extended counting times required. This approach has now largely been abandoned in favor of AMS. In contrast, Pearson (1979) used large samples (approximately 13 g C_6H_6) to produce high-precision measurements (± 20 years BP). He used a standard LSC without background reduction features but optimized all aspects of the method and introduced a number of correction factors for the counting process. Much of his high-precision LSC work, together with the GPC work of Minze Stuiver, was on producing the high-precision dendrochronological calibration curve (e.g., Pearson and Stuiver, 1993; Stuiver and Pearson, 1993).

Gas Proportional Counting

Instrumentation

Gas counters were developed during the beginning of the twentieth century, long before the advent of radiocarbon dating, and consist basically of a cylindrical volume with a thin wire stretched along the axis (Figure 7).

The volume contains a gas, and an electrical field is applied between the cylinder and the wire. The polarity is chosen so that the central wire is positive (anode) with respect to the outer cylinder (cathode). The electric field is at a maximum at the anode wire and decreases rapidly toward the cathode; using thin wires, very high field values are achieved close to the anode. The counting gas (originally argon/alcohol) will be ionized by particles or electromagnetic radiation and the electrons produced will drift toward the anode because of the electric field. Very close to the wire, the field becomes so strong that multiplication starts (typically $>10^6$) and an avalanche of electrons develops. The charge detected at the wire depends

strongly on the applied voltage and in the so-called proportional region this charge (after multiplication) is proportional to the energy of the decay event. This is the regime for GPC. For more details the reader is referred to Friedlander et al. (1981) and Watt and Ramsden (1964).

The practical solution to the problems of low activities and low energies described in the section 'Introduction' was developed in the 1950s, and involved the employment of both GPC counters (for ^{14}C detection) and GM counters (for cosmic ray background reduction), a passive shield, and the use of a gaseous form of carbon (synthesized from the sample material) as the counting gas. In normal operation, the radioactive source is outside the GPC; however, the low-energy β -particles emitted by ^{14}C cannot penetrate the counter walls. Using carbon in the gaseous form as the counting gas solved this problem. Thus, the counting gas is also the sample. CO_2 is the most commonly used counting gas (Mook, 1983). The emphasis on CO_2 is caused by the fact that (i) CO_2 is the primary gas that has to be produced in all cases, after sample pretreatment, via acid hydrolysis, wet oxidation, or combustion; (ii) the cryogenic properties of CO_2 facilitate handling and thus minimize contamination; and (iii) all laboratories involved in high-precision work (like calibration of the ^{14}C timescale) use CO_2 . Some laboratories prefer the use of the other gases in order to avoid the high purity requirements for CO_2 as a counting gas.

GPC counters

Basically, two types of GPC have been developed for CO_2 gas counting: quartz tubes coated with a thin (typically $0.01\ \mu\text{m}$) conductive gold layer acting as the cathode (e.g., de Vries et al., 1959), and copper tubes (e.g., Tans and Mook, 1978). In both cases, great care has to be taken that the construction materials are free from radioactive contamination.

Key parameters affecting low-level $^{14}\text{CO}_2$ counting are gas pressure, anode voltage, and gas multiplication (Kromer and Münnich, 1992).

The gas pressure determines the precision of the ^{14}C measurement. For example, a statistical precision of 0.1% corresponds to 10^6 counts. If this needs to be measured in a 1-day counting period, the counter has to contain 2 mol of carbon, equivalent to 50 l of CO_2 gas at National Toxicology Program (NTP).

In addition, the CO_2 pressure affects background and the demands on gas purity. Because problems with electrical insulation and impurity sensitivity pose limits to the counter diameter, Tans and Mook (1978) developed a system of seven parallel-connected (sub)counters, operating with CO_2 at pressures up to 6 bar. All seven are made of quartz and can operate at the same high voltage, close to 7 kV. The anode wires are stainless steel with a diameter of $25\ \mu\text{m}$. These numbers represent the ideal design parameters and in practice, the counters operate at pressures around 3 bar.

Special counters

Special GPC designs include multiwire counters and mini-counters. The sensitivity in GPC can be improved by increasing the size of the counter and by increasing the gas pressure; however, increasing the diameter and/or the pressure would require higher voltages to be applied across the counter. This

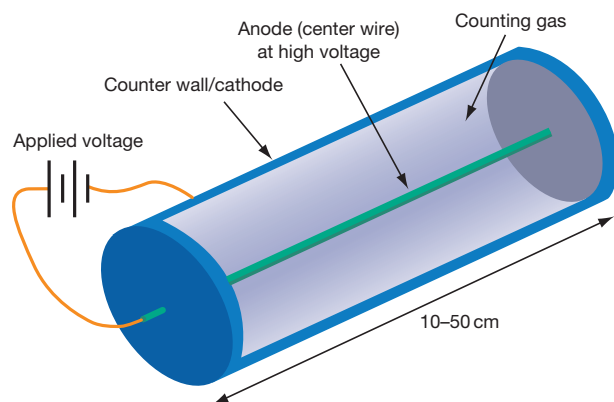


Figure 7 Schematic representation of a gas proportional counter.

can be avoided by dividing the sensitive volume into several wall-less counter elements, the so-called multielement counter (Povinec, 1992); however, in practice their operation is rather complex.

Mini-counters comprise an array of counters designed for small samples (typically 10–100 mg C), inside a single anticoincidence shield (Oulet et al., 1986). Since ^{14}C dating requires a precision of 0.5% or better, counting times are long (many days) and this puts stringent constraints on the stability of the counting system. Again, the introduction of AMS made the mini-counters more or less obsolete for practical dating purposes.

Passive shielding

The purpose is the same as in LSC although shields for GPC can be more complex in design. A typical counter shield is shown in Figure 8. Iron is used to absorb the γ radiation from the

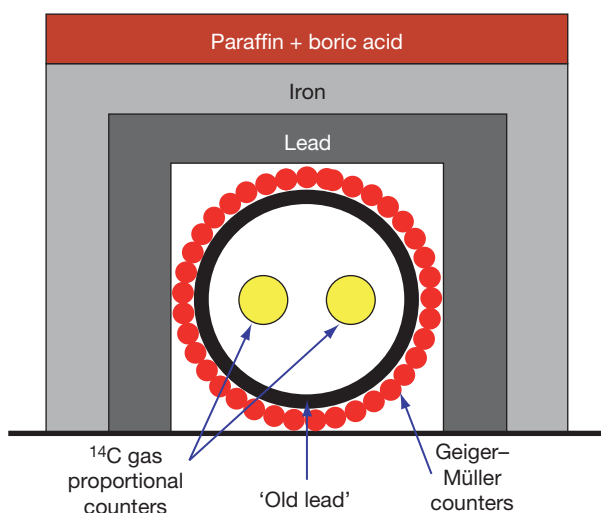


Figure 8 Typical passive and active shield assembly for gas proportional counters.

surroundings (potassium-40 (^{40}K) in concrete, etc.) and the nucleon component in the cosmic radiation. Neutrons are moderated to thermal energies by paraffin wax and subsequently absorbed by boron. The innermost shield, between the GM counters and the GPC, consists of 'old' (=low radio-activity) lead. Passive shielding reduces the cosmic ray background by about 10%.

Anticoincidence shielding

In most setups, a ring of GM counters connected in parallel surrounds the GPC. These operate in anticoincidence with the GPC, thereby acting as an active shield to register the cosmic ray background radiation. This active shield reduces the cosmic ray background by about 90%.

Electronics

A typical schematic overview of a measuring setup is shown in Figure 9. Each set of GPC and GM ring is connected to a controller unit, which includes adjustable discriminators (low, medium, and high), the settings of which depend on whether the detected pulse is selected as an α -particle (from ^{222}Rn), meson (cosmic ray particle), β -particle (from ^{14}C), or purity count. A separate controller collects external parameters such as barometric pressure, and counter gas pressure and temperature of each individual counter.

Sample Analysis

Counting gas

As with LSC, the advantages of the gas counting technique over solid-source counters are the high detection efficiency and 4π geometry, while in addition, the problems of cross-contamination using solid graphite can be avoided. The choice of gas is made with a view to chemistry (sample treatment) and purification. CO_2 , CH_4 , C_2H_6 , and C_2H_2 fulfill these requirements and are commonly used by ^{14}C laboratories. de Vries (1956) developed the CO_2 method, which is the simplest for

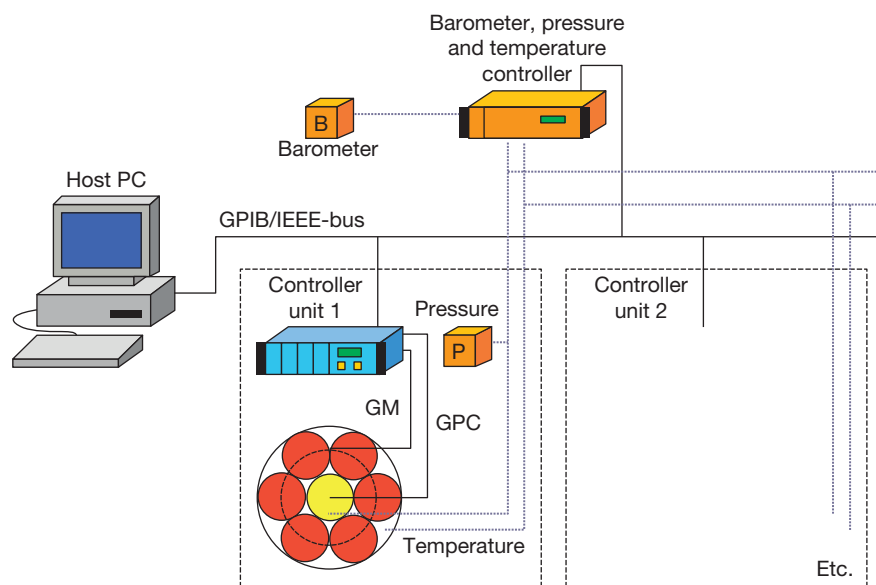


Figure 9 Typical schematic overview of a gas proportional counting setup. GM, Geiger-Müller; GPC, gas proportional counting.

gas production from sample materials; however, it is very sensitive to small amounts of impurities such as oxygen: concentrations greater than $1:10^6$ are sufficient to reduce the gas gain as electrons originating from ^{14}C decay and drifting to the anode of the GPC may become attached to the electronegative impurities in the CO_2 gas.

CO₂ preparation

The sample is combusted to CO_2 in a stream of oxygen (O_2), typically over several hours, to ensure complete combustion (Figure 10). Next, a copper oxide (CuO) furnace at 800°C oxidizes carbon monoxide (CO) to CO_2 while silver (Ag) wool at 450°C removes any halides present, and finally, potassium permanganate (KMnO_4) solution removes sulfides. Water (H_2O) is trapped cryogenically in CO_2 /methanol (CH_3OH) at -78°C and the final product, CO_2 , is frozen in liquid N_2 . Finally, the CO_2 is passed over Cu at 500°C to oxidize oxides of nitrogen. The CO_2 (typically a few liters) is then stored in a steel or glass container, from which it can be transferred

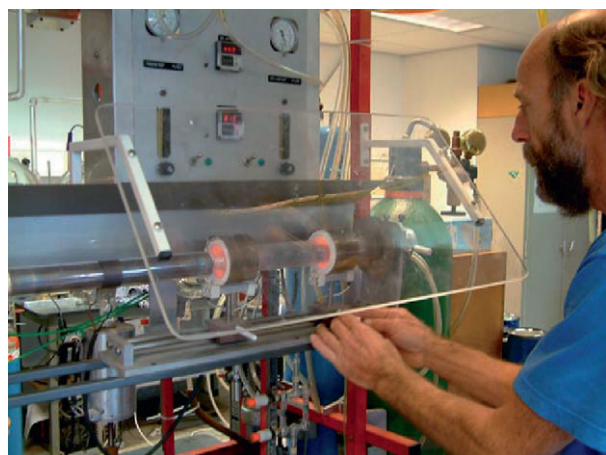


Figure 10 Sample undergoing combustion in a stream of oxygen.

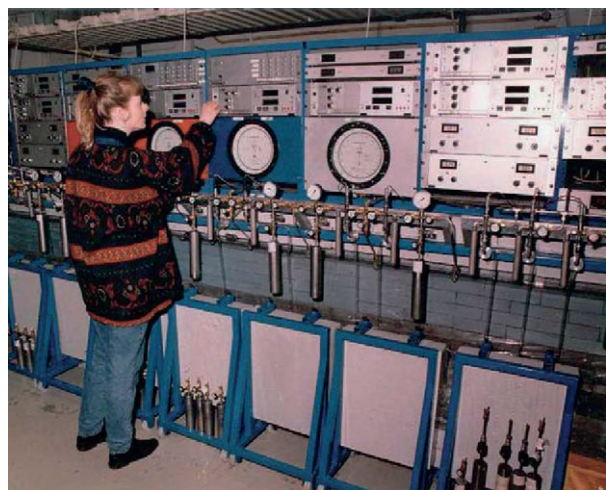


Figure 11 The Groningen gas proportional counter facility.

cryogenically back and forth to the GPC. As with C_6H_6 , the CO_2 is stored for about 1 month to allow any ^{222}Rn to decay.

A few milliliters of CO_2 are removed for $\delta^{13}\text{C}$ measurement, necessary for isotopic fractionation correction. The Groningen GPC facility is shown in Figure 11.

Counting procedure

Counting times depend on parameters like precision required or age of the sample, but are typically 24 h per sample. Backgrounds and standards are measured monthly but for a few days. Table 3 illustrates the limits of detection, etc., of several GPC systems in use in Groningen.

For GPC, high-precision (defined as 0.2% or better) measurements are possible using large counters like Groningen counter number 9. As with LSC, such precision is needed for calibration of the radiocarbon timescale (Pearson and Stuiver, 1993; Stuiver and Pearson, 1993) and for dating projects where high temporal resolution is required (e.g., van der Plicht and Bruins, 2005). The latter may serve as an example project for stringent quality control.

Activity calculations

The total number of counts N during a counting time t_c is:

$$N(AB)t_c$$

where A represents the net ^{14}C count rate, and B the background count rate. However, this simple relationship requires corrections for barometric pressure and CO_2 purity. Higher barometric pressure means a greater thickness and mass of atmosphere, which acts as a shield against the cosmic ray flux. In routine CO_2 counter operation, the counter itself determines the gas purity. At fixed counting conditions (such as voltage and pressure), electronegative impurities result in a decrease in pulse height, which causes a shift of the counter plateau towards higher voltages. A similar shift occurs in the meson characteristic curve, which is measurable by the decrease in the count rate at the steep part of the curve (Figure 12). The decrease in meson count rate can be converted to that in the β count rate through a comparison of both slopes (Brenninkmeijer and Mook, 1979).

Radiometric ^{14}C Age and Error Calculations

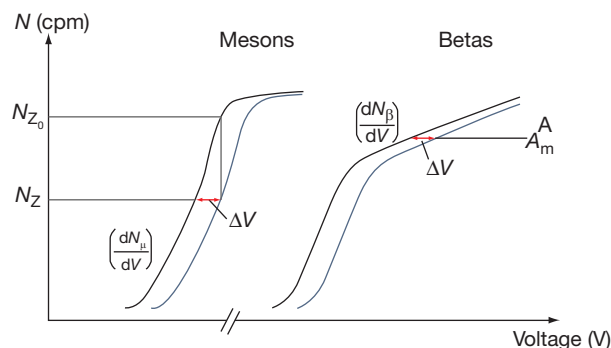
Age Calculation

For LSC: once the net count rates per gram of C_6H_6 have been determined for oxalic acid standards, known age standards and unknown age samples, an activity correction for quenching is applied. Similarly, for GPC: once the net count rates per unit mass of CO_2 have been determined, activity corrections for gas purity and atmospheric pressure are applied. Following these procedures, activities are further corrected for fractionation using a fractionation factor (FF) calculated using the $\delta^{13}\text{C}$ value. This correction is common to LSC and GPC. Fractionation arises from differential uptake of isotopes by physical and/or chemical processes. These processes, both in nature and in the laboratory (during sample preparation), are dependent on mass. Therefore, fractionation changes the ^{14}C content of the sample (relative to the ^{12}C content) and consequently the

Table 3 Performance characteristics of some of the large and small Groningen gas proportional counters

Counter	Number of counters	Volume (ml)	Pressure (mbar)	Minimum weight of carbon (g)	Maximum measurable age (years BP)	A_0 (cpm) ^a	Background (cpm)
9	7	6840	2907	12.5	58000	118.3	13.9
9a	1	977	2907	1.9	47000	17.0	2.7
9b	3	2931	2907	5.3	52000	50.4	5.7
9c	3	2931	2907	5.3	52000	50.9	5.5
8	1	288	3100	0.75	46000	5.85	1.0

^aFrom measurement of Oxalic acid I; BP, before present; cpm, counts per minute.

**Figure 12** Shift in the β -curve and accompanying meson curve caused by the presence of electronegative impurities in the counting gas.

^{14}C age, if a correction factor is not applied. Correction can take place because the ^{13}C isotope is also fractionated (relative to the ^{12}C) and because the change in the $^{13}\text{C}/^{12}\text{C}$ isotopic ratio is proportional to the change in the $^{14}\text{C}/^{12}\text{C}$ ratio. Therefore, the FF is effectively a normalization factor that corrects the different ^{14}C activities of contemporaneous samples using their $\delta^{13}\text{C}$ values. The $\delta^{13}\text{C}$ value of -25% (0.025) in the equations is the theoretical value for wood and the factor of 2 in the equations takes into account the fact that $^{14}\text{C}/^{12}\text{C}$ fractionation is twice that of $^{13}\text{C}/^{12}\text{C}$.

For unknown age samples and Oxalic acid II standards,

$$\text{FF} = 1 - [2(0.025 + ^{13}\text{C})]$$

For Oxalic acid I standards it is:

$$\text{FF} = 1 - [2(0.019 + ^{13}\text{C})]$$

Therefore, the normalized sample activity is:

$$A_{\text{SN}} = A_s [1 - [2(0.025 + ^{13}\text{C})]]$$

where A_s is the measured sample activity uncorrected for fractionation and A_{SN} is the normalized sample activity.

Similarly, the normalized activity for Oxalic acid II is

$$A_{\text{ON}} = 0.7459 \times A_{\text{OXII}} \{1 - [2(0.025 + ^{13}\text{C}_{\text{OXII}})]\}$$

while the normalized activity for Oxalic acid I is

$$A_{\text{ON}} = 0.95 \times A_{\text{OXI}} \{1 - [2(0.019 + ^{13}\text{C}_{\text{OXI}})]\}$$

where A_{OX} is the measured oxalic acid activity and A_{ON} is the normalized activity.

The A_{ON} activity was originally based on the age corrected activity of tree rings growing during AD 1890 and in equilibrium with the AD 1890 atmosphere. AD 1890 is the last time when the ^{14}C record was not significantly perturbed by anthropogenic influences. The first primary reference standard (Oxalic acid I) was produced in 1957 and has an internationally accepted reference value of: $0.95 \times$ the activity of Oxalic acid I, in AD 1950, normalized to a $\delta^{13}\text{C}$ value of -19% (with respect to Vienna Pee Dee Belemnite (VPDB)) = activity of AD 1890 wood. The measurements were not made in AD 1950 but this was deemed to be an appropriate reference year. Therefore, when a sample is dated, its age is always quoted in years BP where 'present' is AD 1950. Since the samples and oxalic acid standards have been decaying at the same rate post-1950, there is no need for decay corrections. By 1978, supplies of Oxalic acid I were practically exhausted and Oxalic acid II, synthesized from sugar beet grown in 1977, was introduced in 1980 (Mann, 1983): $0.7459 \times$ the specific activity of Oxalic acid II normalized to a $\delta^{13}\text{C}$ value of -25% equates to the Oxalic acid I reference value. The 0.95 and 0.7459 factors take into account the fact that the oxalic acid standards were produced from materials that grew during the nuclear era.

From the above, the fraction modern (F) of the samples is calculated using Oxalic acid I as follows:

$$F = \frac{A_s \{1 - [2(0.025 + ^{13}\text{C})]\}}{0.95 \times A_{\text{OXI}} \{1 - [2(0.019 + ^{13}\text{C}_{\text{OXI}})]\}} = \frac{A_{\text{SN}}}{A_{\text{ON}}}$$

Similarly, using Oxalic acid II:

$$F = \frac{A_s \{1 - [2(0.025 + ^{13}\text{C})]\}}{0.7459 \times A_{\text{OXII}} \{1 - [2(0.025 + ^{13}\text{C}_{\text{OXII}})]\}} = \frac{A_{\text{SN}}}{A_{\text{ON}}}$$

This is the activity ratio ^{14}a (also called 'percent modern'; Mook and van der Plicht, 1999).

From this, the radiocarbon age (t) can be calculated as follows:

$$t = \frac{1}{\lambda} \ln \left[\frac{1}{F} \right] = 8033 \ln \left[\frac{1}{F} \right]$$

where t = age in years BP since the death of the organism and λ = decay constant for ^{14}C ($\ln 2/5568$). The half-life of 5568 years is the original half-life used by Libby and is used by convention in age calculations and the definition of 'BP.'

Stuiver and Polach (1977) and Mook and van der Plicht (1999) provide a detailed discussion of radiocarbon conventions and nomenclature.

Error Calculation

Because radioactive decay is a random process, a repeated measurement of the same sample under identical conditions will not give the same result. The uncertainty in a measured activity is given by the standard deviation (σ). The standard deviation in a total number of counts N during a counting time t_c is $\sqrt{N}$ (Poisson error). Radiocarbon laboratories do not have a standardized method of determining errors on age measurements because of the variations in counting procedures but at a minimum, the error should include Poisson errors from the sample, Oxalic acid standard(s) and background count rates, together with errors on the relevant correction factors. Furthermore, the overall error is normally subject to a minimum value based on the standard deviation of a set of replicate analyses of a bulk sample.

Typical measurement precision for ^{14}C laboratories is 0.5% or better and this applies to all three detection techniques, AMS, LSC, and GPC.

An indepth discussion of error calculations and quality assurance issues can be found in this encyclopedia.

See also: Dating Techniques. **Plant Macrofossil Records:** Late Glacial Multidisciplinary Studies. **Introduction:** History of Dating Methods; History of Quaternary Science. **Radiocarbon Dating:** Causes of Temporal ^{14}C Variations; Sources of Error; Variations in Atmospheric ^{14}C .

References

- Anderson EC, Arnold JR, and Libby WF (1951) Measurement of low level radiocarbon. *The Review of Scientific Instruments* 22: 225–230.
- Brenninkmeijer CAM and Mook WG (1979) The effect of electronegative impurities on CO_2 proportional counting: An on-line purity test counter. In: Berger R and Suess HE (eds.) *Proceedings of the Ninth International Radiocarbon Conference*, pp. 185–196. Los Angeles and La Jolla.
- Cook GT (1995) Enhanced low-level LSC performance for carbon-14 dating using a bismuth germinate ($\text{Bi}_4\text{Ge}_3\text{O}_{12}$) quasi-active guard. *Radioactivity and Radiochemistry* 6: 10–15.
- Cook GT and Anderson R (1992) A radiocarbon dating protocol for use with Packard scintillation counters employing burst-counting circuitry. *Radiocarbon* 34: 381–388.
- de Vries HL (1956) Purification of CO_2 for use in a proportional counter for ^{14}C age measurements. *Applied Science Research B5*: 387–400.
- de Vries HL, Stuiver M, and Olsson I (1959) A proportional counter for low level counting with high efficiency. *Nuclear Instruments and Methods* 5: 111–114.
- Einarsson S (1992) Evaluation of a prototype low-level liquid scintillation multi-sample counter. *Radiocarbon* 34: 366–373.
- Fifield LK (1999) Accelerator mass spectrometry and its applications. *Reports on Progress in Physics* 62: 1223–1274.
- Friedlander G, Kennedy JW, Macias ES, and Miller JM (1981) *Nuclear and Radiochemistry*. New York: Wiley.
- Haas H and Trigg V (1991) Low-level scintillation counting with a LKB Quantulus counter establishing optimal parameter settings. In: Ross H, Noakes JE, and Spaulding JD (eds.) *Liquid Scintillation Counting and Organic Scintillators*, p. 669. MI, USA: Lewis Publishers.
- Hiebert RD and Watts RJ (1953) Fast coincidence circuits for H^3 and C^{14} measurements. *Nucleonics* 11: 38–41.
- Hogg A, Polach H, Robertson S, and Noakes JE (1991) Application of high purity synthetic quartz vials to liquid scintillation low-level ^{14}C counting of benzene. In: Ross H, Noakes JE, and Spaulding JD (eds.) *Liquid Scintillation Counting and Organic Scintillators*, p. 123. MI, USA: Lewis Publishers.
- Kaiholia L (1991) Liquid scintillation counting performance using glass vials in the Wallac 1220 Quantulus. In: Ross H, Noakes JE, and Spaulding JD (eds.) *Liquid Scintillation Counting and Organic Scintillators*, p. 495. MI, USA: Lewis Publishers.
- Kojola H, Polach H, Nurmi J, Oikari T, and Soini E (1984) High resolution low-level liquid scintillation β -spectrometer. *Applied Radiation and Isotopes* 35: 949–952.
- Kromer B and Münnich KO (1992) CO_2 gas proportional counting in radiocarbon dating – Review and perspective. In: Taylor RE, Long A, and Kra RS (eds.) *Radiocarbon After Four Decades: An Interdisciplinary Perspective*, pp. 184–197. New York: Springer.
- Libby WF (1952) *Radiocarbon Dating*. Chicago: University of Chicago Press.
- Mann WB (1983) An international reference material for radiocarbon dating. *Radiocarbon* 25: 519–527.
- Mook WG (1983) International comparison of proportional gas counters for ^{14}C activity measurements. *Radiocarbon* 25: 475–484.
- Mook WG and Streunman HJ (1983) Physical and chemical aspects of radiocarbon dating. In: Mook WG and Waterbolk HTJ (eds.) *^{14}C and Archaeology*, vol. 8, pp. 31–55. New York: PACT Publications.
- Mook WG and van der Plicht J (1999) Reporting ^{14}C activities and concentrations. *Radiocarbon* 41: 227–239.
- Ottel RL, Huxtable G, and Sanderson DCW (1986) The development of practical systems for ^{14}C measurement in small samples using miniature counters. *Radiocarbon* 28: 603–614.
- Pearson GW (1979) Precise ^{14}C measurement by liquid scintillation counting. *Radiocarbon* 21: 1–21.
- Pearson GW and Stuiver M (1993) High-precision bidecadal calibration of the radiocarbon timescale, 500–2500 BC. *Radiocarbon* 35: 25–34.
- Povinec P (1992) ^{14}C gas counting: is there still a future? *Radiocarbon* 34: 406–413.
- Sanyuan G and Yuanming X (1992) Instrumentation and software for low-level liquid scintillation counting radiocarbon dating. *Radiocarbon* 34: 374–380.
- Stuiver M and Pearson GW (1993) High-precision bidecadal calibration of the radiocarbon timescale, AD 1950–500 BC and 2500–6000 BC. *Radiocarbon* 35: 1–24.
- Stuiver M and Polach HA (1977) Reporting of ^{14}C data. *Radiocarbon* 19: 355–363.
- Tans PP and Mook WG (1978) Design, construction and calibration of a high accuracy carbon-14 counting set up. *Radiocarbon* 21: 22–40.
- Valenta RJ (1987) Reduced Background Scintillation Counting. US Patent No. 4,651,006.
- van der Plicht J and Bruins HJ (2005) Quality control of Groningen ^{14}C results from Tel Rehov. In: Levy TE and Higham T (eds.) *The Bible and Radiocarbon Dating – Archaeology, Text and Science*, pp. 256–270. London: Equinox.
- Watt DE and Ramsden D (1964) *High Sensitivity Counting Techniques*. New York: Pergamon Press.

AMS Radiocarbon Dating

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Symbols

A Activity of radiocarbon in a sample
F Fraction of modern carbon

F_m Fraction of modern carbon measured before blank correction
 λ Radioactive decay constant

Accelerator Mass Spectrometry Fundamentals

Over the last nearly three decades, accelerator mass spectrometry (AMS) has developed from a curiosity to become by far the most common method of radiocarbon dating. The technique developed from the need for more sensitivity and smaller sample sizes than could be achieved by decay counting (Cook and van der Plicht, this volume). The first reports of the use of an accelerator combined with a mass spectrometer system were reported in 1977 (see Tuniz et al., 1998 for the historical discussion). AMS has become one of the most important tools in the geochronologist's arsenal to try to determine the age of material. For radiocarbon, precise measurements are needed to determine an age. For details of the principles of radiocarbon dating, the reader is referred to the section of van der Plicht (this volume).

A fundamental difference between radionuclide measurement using AMS differs from the decay counting methods (Cook and van der Plicht, this volume) is the number of atoms of the radionuclide in the sample is measured directly, instead of waiting for individual radioactive decay events. For radiocarbon (^{14}C) dating, this makes the technique 1000 to 10 000 times more sensitive than counting of the radioactive decay. AMS also has important applications for other longer lived radionuclides.

It is easy to understand the usefulness of AMS by looking at the radioactive decay equation

$$dN/dt = -\lambda N$$

Here, the number of atoms of ^{14}C , N , is related directly to the rate of radioactive decay (dN/dt) through the decay constant, λ . So, the decay rate is very slow; for 1 g of modern carbon, the decay rate is 13.59 dpm g^{-1} . In AMS, one typically measures a sample of 0.5 mg C, and over 120 000 atoms can be counted from a modern sample. If one takes a counting rate of 120 000 atoms in 1080s, or 111 counts/second – from 0.5 mg of carbon, then this translates to $111 \times 60/0.0005 = 1.33 \times 10^7$ cpm g^{-1} , a difference compared to decay counting of over a million.

The extreme sensitivity is achieved by accelerating sample atoms as ions to high energies using a particle accelerator and using nuclear particle detection techniques (Tuniz et al., 1998). The first purpose built machines for AMS radiocarbon dating relied on the design of Purser et al. (1980) or the redesign of existing Van de Graaff accelerator systems, some of which

continue to provide valuable data. Early machines dedicated to AMS radiocarbon were installed at the Universities of Oxford, Arizona, Toronto, and Nagoya in the early 1980s. Newer designs resulted from competition after the expiry of the original patent (see Purser et al., 1980), and this equipment is currently commercially available from National Electrostatics Corporation (NEC; Middleton, Wisconsin, USA) and High Voltage Engineering Europa (Amersfoort, the Netherlands). The field has grown continuously since that time, with over 70 purpose built AMS machines operating in 2010, and orders for a number of new machines are pending.

A picture of the 3 MV AMS at the University of Arizona is shown in Figure 1 and later a diagram of the design of the 3 MV AMS machine (based on an NEC accelerator) is shown in Figure 2.

Today, high external precision of about $\pm 0.35\%$ in ^{14}C content, or ± 30 years in uncalibrated radiocarbon age is easily achieved on a single 0.5-mg-sized sample target in 20 min of measurement time. Samples as small as 100 μg or less have been successfully dated and even smaller samples (~ 10 – $20 \mu\text{g}$) have been measured for special experiments, with reduced precision due to limited time available to measure the sample. With longer counting times or using multiple target measurements to reduce error, one can achieve better than ± 20 years in radiocarbon age.

The AMS method has also greatly improved the measurements of longer lived radionuclides such as ^{26}Al , ^{10}Be , ^{36}Cl , ^{41}Ca , and ^{129}I , which were very difficult to measure using counting techniques; measurements of small amount of these radionuclides is now routine (Fifield, 1999; Jull and Burr, 2006; Tuniz et al., 1998). Many geochronological applications using AMS radiocarbon and other AMS-based methods are discussed elsewhere in this encyclopedia.

An Example of AMS Operation

In order to demonstrate how the AMS operations, we will look at an example, from an AMS system at the University of Arizona. An AMS system at Arizona consists of the following basic components and sequence of processes. There are many good descriptions of the methodology of AMS (Fifield 1999; Jull and Burr 2006; Jull et al., 2003; Kutschera 2005; Synal and Wacker, 2010; Tuniz et al., 1998), so we refer the reader to these publications for detailed operational information.



Figure 1 Photograph of the University of Arizona 3 MV accelerator mass spectrometer, built by National Electrostatics Corporation.

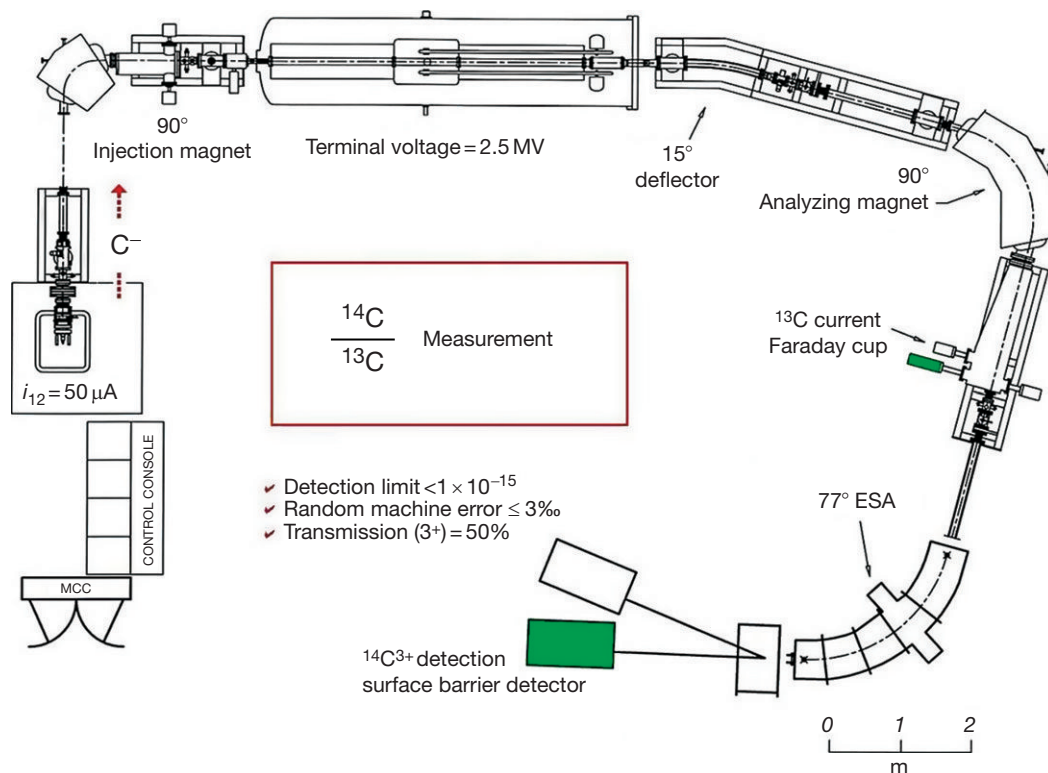


Figure 2 Diagram of the operation of the 3 MV AMS. The major elements of the machine are shown. The ion source generates $50 \mu\text{A}$ or more of C^- , these ions are accelerated and passed through the low-energy magnet, where the first separation by mass occurs. Molecule ions are destroyed, once the ions are further accelerated to 2.5 MV and passed through the stripping canal (**Figure 3(e)**).

The quantity produced by the AMS measurement is usually expressed as the fraction of modern carbon (F) which can be related to the equations shown by van der Plicht (2005) earlier in this volume. Calculations are discussed later in this article.

In this discussion, one refers to **Figure 2**, which shows a diagram of the 3 MV NEC system at Arizona. This figure summarizes the major design elements of the system and also

operational conditions for ^{14}C . One can also look at some pictures of the various components.

1. The ion source (**Figure 3**) generates negative carbon ions ($20\text{--}60 \mu\text{A C}^-$) by Cs sputtering from a graphite target. A diagram of this process is shown in **Figure 4**. The graphite target must be prepared from the original sample material;

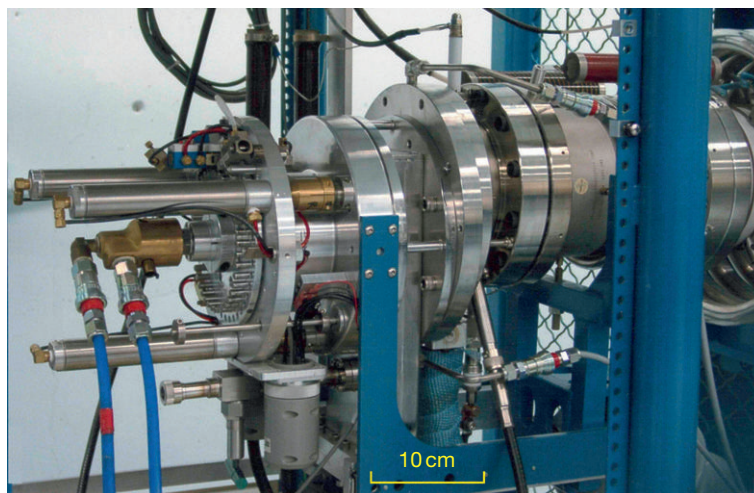


Figure 3 Source for the generation of C^- ions by sputtering of the graphite target.

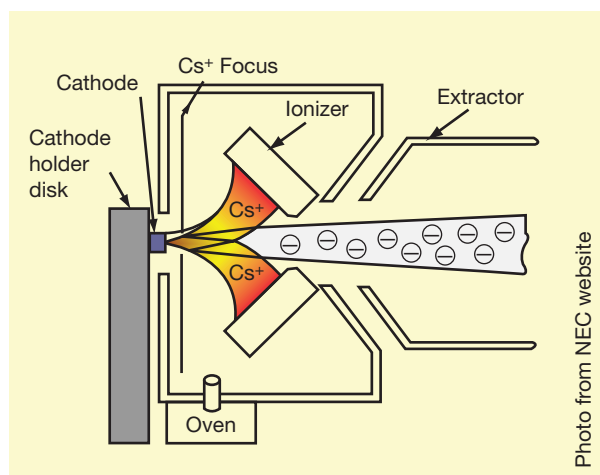


Figure 4 Diagram of the ion-source operation.

the chemistry is discussed later. Good sample currents are 50–70 μA . At Arizona, one measures each target for 200 s, and the complete target wheel of 40 targets is rotated a total of six times, giving a total analysis time on one sample of about 20 min. Usually, one runs one wheel of 40 targets per day. In some laboratories, gas ion sources are under development, but these have not yet reached the reliability of the solid graphite source (Kim et al. 2002).

2. The injection magnet (Figure 5) performs the initial separation of the negative ions by mass. At this point, molecular ions such as hydrides of carbon (CH^-) are also present. N^- is unstable, so an important possible interference is removed. Mass 14 is measured most of the time and in our 3 MV Pelletron machine, masses 13 and 12 are pulsed rapidly through the machine for short intervals.
3. The accelerator (Figure 6), which generates a high voltage of 2.5 million volts (for ^{14}C measurements) and accelerates the C^- ions toward the central part of the machine, which is at high voltage and usually called the 'terminal.'
4. The terminal houses a gas canal known as the 'stripper.' The 2.5 million volt C^- ions enter this gas canal and

interact with a gas. The canal is differentially pumped, so that a high vacuum can be maintained in the accelerator tubes. Because they are moving so fast, they lose several electrons from their electron cloud and as a result become positively charged (see Figure 7 for a diagram of this process). Any molecules, such as CH^- are destroyed in this process. The positively charged ions are accelerated away from the positively charged terminal to the exit of the accelerator.

5. The ions exit the accelerator and are then separated by energy and charge, using the first of two electrostatic deflectors. This device deflects a beam of ions using an electrostatic field and a narrow defining exit slit. The 3^+ ions are selected.
6. The 3^+ ions are now passed through a large 90° magnet. When mass 12 or 13 is injected, these ion beams stop in Faraday cups after the high-energy magnet and the current is measured. If mass 14 is injected, the ^{14}C passes through a 77° electrostatic analyzer, shown in the earlier diagram (Figure 2), and then hits an energy-sensitive solid-state detector (Figure 8). A gas detector (Figure 9) is also used in some AMS machines and for other radionuclides. In the case of the surface-barrier detector, it has the property that it produces a pulse proportional in height to the energy of the ion, for every ion hitting the detector. A thin foil in front of the detector slows down the few $^{14}N^{3+}$ ions that make it through the machine. A similar process occurs in a gas detector, although we can get additional separation with this type of device. The number and energy of the ions are separated by computer, and the ^{14}C can be distinguished from any other ions which are counted. The count rate for a modern sample is around 120 counts per second. An example of the measurement for a counting spectrum is shown in Figure 10; this shows the low number of ^{14}N ions to the left of the ^{14}C .
7. The ratio of ^{14}C – ^{13}C current in the sample is compared to that of the standards. Appropriate corrections are made for the small natural variations of $^{13}C/^{12}C$ ratio and for the background. The radiocarbon age can now be calculated.



Figure 5 Injection magnet, which performs the first mass separation. Molecular ions are still present at this stage.

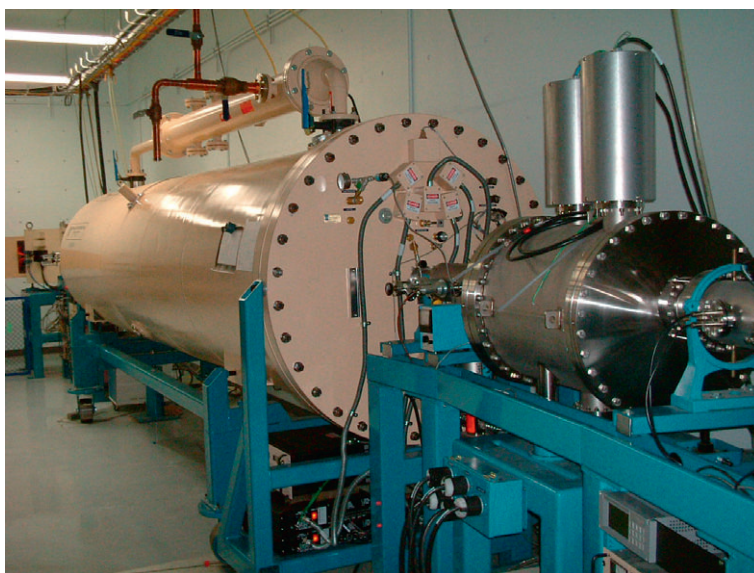


Figure 6 Accelerator tank. Note the electrostatic deflector after the accelerator.

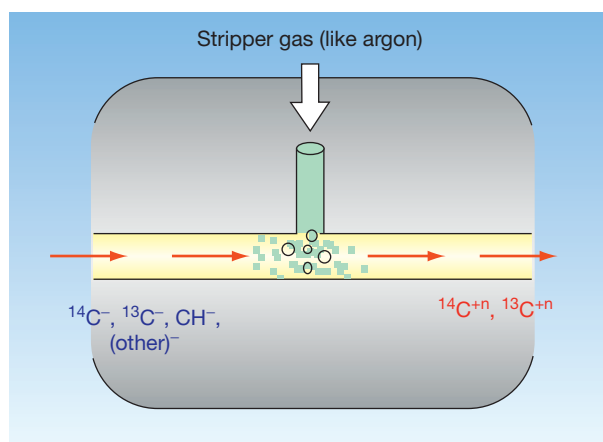


Figure 7 Diagram of the stripping process at the accelerator terminal.

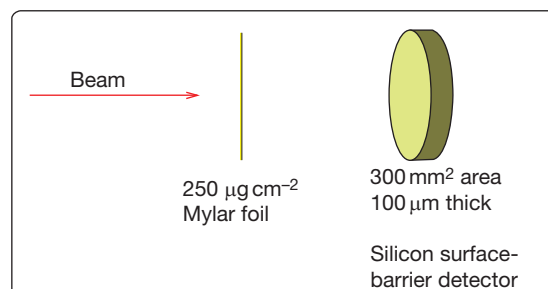


Figure 8 Arrangement of the silicon surface-barrier detector and Mylar foil to slow down $^{14}\text{N}^{3+}$ relative to $^{14}\text{C}^{3+}$.

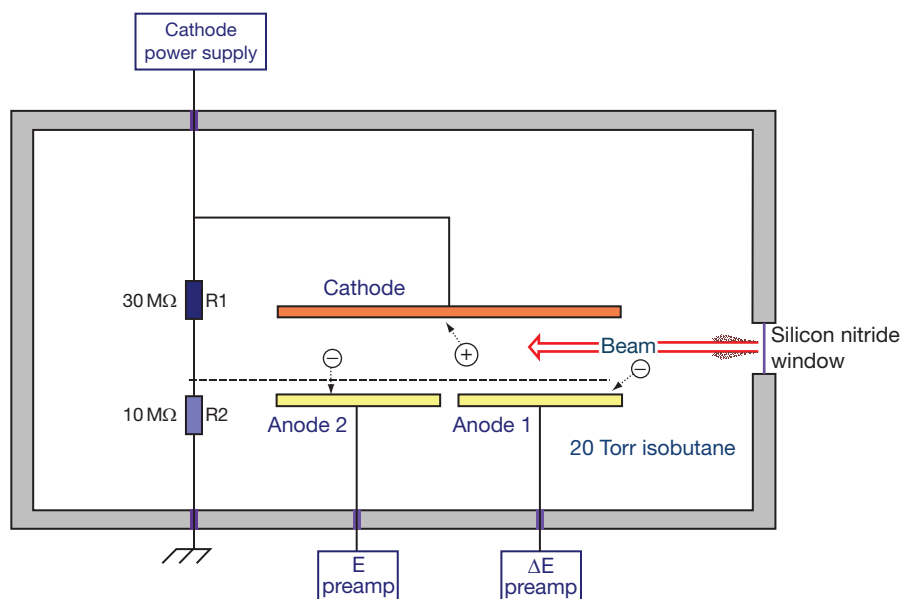


Figure 9 Drawing of a typical layout for a gas detector system. The beam entering the detector from the right slows down giving the first (ΔE) signal indicating the energy loss; the second stage on the left gives the residual energy (E). Courtesy: Dr. George Burr.

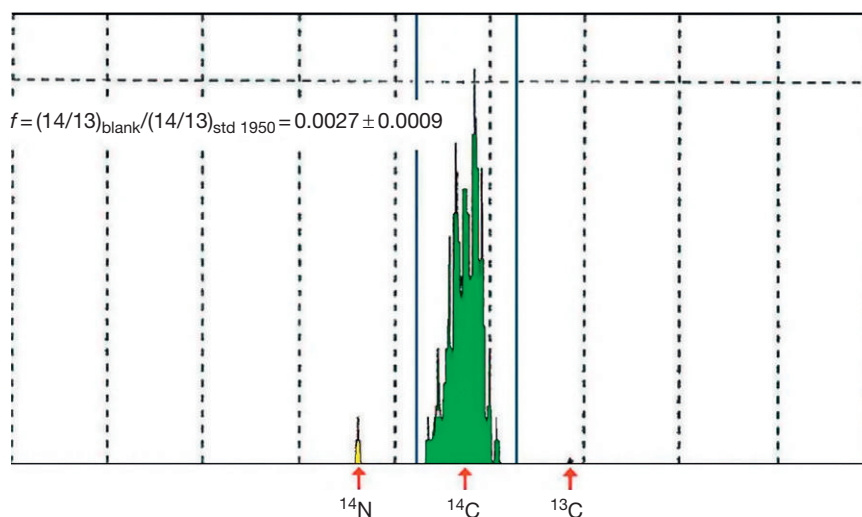


Figure 10 A ^{14}C counting spectrum.

Chemical Preparation of Samples

An important aspect of AMS, like all radiocarbon dating, is to make sure that the samples which are studied have been cleaned of possible contaminations. Because of the small sample size, it is important that the sample is clean and all contaminants have been removed. Some basic procedures for AMS radiocarbon samples at Arizona were summarized by Jull et al. (2004):

1. Acid–base–acid (ABA) method for charcoal, wood, cellulose, plant material, animal tissue: After physical inspection, samples are cleaned with 1 N HCl acid, 0.1% NaOH, and 1 N HCl (ABA pretreatment), washed with distilled water, dried, and combusted at 900 °C with CuO. Hatté et al. (2001) discussed some modifications and potential problems with the ABA method.
2. Carbonates: In general, samples are etched with 100% H_3PO_4 to remove 50–85% of the carbonate, dried, and hydrolyzed with H_3PO_4 as discussed by Burr et al. (1992).
3. Selective combustion for sediments. After cleaning in 1 N HCl and drying, the sample is combusted at 400 °C in ~0.3 atm. oxygen gas (McGeheh et al. 2001).
4. Oxidative acid cleaning for old charcoal. A new method to handle old charcoal samples was developed by Bird et al. (1999). In this procedure, chromic acid is used to remove all except the most resistant charcoal in a sample. This is of particular interest for charcoal samples >20 ka.
5. Textiles, parchment, canvas, art works, and artifacts: The samples are given the ABA pretreatment and after washing and drying, they are Soxhlet extracted with hexane, then ethanol and finally methanol. After washing in distilled

water, and drying, they are combusted at 900 °C with CuO. Bruhn et al. (2001) proposed a more complex, six-stage cleaning procedure for art works.

Of course, for some specialized samples, other methods need to be developed for these applications.

After cleaning and combustion, samples are converted to graphite using a catalytic decomposition reaction, using an Fe catalyst and Zn as the reducing agent for the reaction $2\text{CO} \rightarrow \text{CO}_2 + \text{C}$.

Finally, the graphite powder is pressed into a target holder of Arizona design, which is now widely used in all NEC ion sources. The target material is pressed from both sides in a press-anvil arrangement.

Calculation of Results

Calculation and normalization to standards

The measurement of ^{14}C follows the procedures and calculations described in detail by Donahue et al. (1990) and further discussed by Burr and Jull (2009). McNichol et al. (2001) have reviewed the question of interlaboratory differences in calculations.

As discussed by Cook and van der Plicht (this volume), the result from any radiocarbon measurement needs to be expressed in the form:

$$F = A_{\text{SN}}/A_{1950\text{AD}}$$

where A_{SN} is the activity of ^{14}C in the sample compared to the value for 'modern carbon,' arbitrarily defined as the value of 1950 AD wood. All radiocarbon measurements are normalized to a common $^{13}\text{C}/^{12}\text{C}$ ratio quoted as $\delta^{13}\text{C}$ of -25‰ , to eliminate any effects of isotopic fractionation either in the sample or the measurement.

Because of depression of the ^{14}C signal due to fossil fuel burning, the value of AD 1950 material is taken to be:

$$A_{1950\text{AD}} = 0.95A_{\text{HOxI}}$$

where HOxI is the original Oxalic-I standard. Since that time, a second primary standard known as HOxII was developed by the US National Institute of Standards and Technology (NIST) as is designated as standard reference material SRM-4990C by NIST. This material can be related to HOxI as follows:

$$A_{1950\text{AD}} = 0.95A_{\text{HOxI}} = 0.7459A_{\text{HOxII}}$$

It is important to note that the value of A_{HOxI} was normalized to a $\delta^{13}\text{C}$ value of -19‰ instead of the $0.7459A_{\text{HOxII}}$ which was normalized to -25‰ .

Donahue et al. (1990) discussed the calculation of F from the raw AMS results, and this was also followed up by McNichol et al. (2001). Many AMS laboratories actually measure $^{14}\text{C}/^{12}\text{C}$ and not $^{14}\text{C}/^{13}\text{C}$ in this calculation; the reader is referred to these papers for details of the calculation and, in particular, the importance of the isotopic corrections.

Errors

Errors are calculated from the larger of the Poisson error (determined from $\sqrt{N}/N$) and the standard error of the individual measurements. Various laboratories use different methods to estimate the total error, but the best practice is to

take the larger of these two errors as the minimum error. Donahue et al. (1990) discussed the practice at Arizona, which is used in many laboratories. In this case, an additional error known as the 'machine random error' is added to the Poisson error before comparing this error to the estimate derived from the standard deviation of the individual measurements. Detailed error analysis considerations are also given in the section by Scott (this volume) and also by Burr and Jull (2010).

Blank corrections

During the processing of samples in the chemistry lab, it is impossible to avoid some small amount of contamination introduced either during the chemistry or conversion into graphite. The 'blank' in the machine itself is negligible. One can define a blank component, f , where:

$$f = F_{\text{B}}/F_{\text{std}}$$

where F_{B} is the fraction of modern carbon in the blank material itself; this is often assumed to be recent or modern carbon. However, other studies suggest one component of the blank may be an 'average' contamination from the last samples run on the line. Donahue et al. (1990) showed that the fraction of modern (F) for the sample can be expressed as:

$$F = F_{\text{m}}[1 - f\{1/F_{\text{m}} - 1\}]$$

where F_{m} is the value of the fraction modern measured for the sample, and F is the final corrected value.

New Trends: Smaller and Smaller Machines

Starting in the mid-1990s, there was a developing trend to make small AMS machines. This trend began about 1996 when a compact design (Hughey et al., 1997) was proposed and was followed by the development of an 0.5 MV machine of a different design by NEC in cooperation with ETH-Zürich (Suter et al., 1999).

The 0.5–1 MV 'Small' Tandem

As already mentioned, the first design for a <1 MV AMS was that of Hughey et al. (1997), who developed the instrument for biomedical work. About the same time, scientists at the ETH in Zurich (Suter et al., 1997) discussed the possibilities of a 0.5–1 MV AMS which would operate in the 1+ or 2+ charge state (as opposed to 3+, which the reader will recall, is used in the larger machine example given above). Suter et al. (2000) later reported on the design of a prototype 0.5 MV machine which would use the 1+ charge state. In this case, the design resulted in a compact machine of about 3–4 m on each side. Although the machine could operate in other than the 1+ charge state, due to transmission considerations most of the discussion on these small machines has involved the 1+ charge state.

Several 0.5 MV machines (Figure 11) are operational at eight facilities, located in Europe, the United States, and China. As noted in the next section, there are new designs going to even lower terminal voltages, perhaps as low as 200 kV, which would eliminate the need for a costly accelerator. One prototype design of this machine was constructed in

Zürich (Synal et al., 2005), and there are at least two additional machines of this type planned (see Synal et al., 2010).

However, there are a number of practical limitations to these very small machines. The control of the stripper gas pressure is delicate and essential. In the $1+$ charge state, molecular ions can exist. The assumption used for higher energy AMS operating conditions that the molecular ions are destroyed cannot be taken for granted at low energy. High stripper gas pressure is needed to remove molecular interferences. Southon et al. (2004) have also reported on the technical difficulties of these operating conditions.

200 kV Machines

Synal et al. (2005) and Synal and Wacker (2010) discussed the development of a 200 kV machine which eliminated both the accelerator tank and also the accelerator tubes. The device designed by Synal and coworkers consists of two gap lenses and a stripper tube and is called 'MICADAS' by the developers. The potential is generated by a 200 kV commercial power supply. The device, of course, must be placed in the mass spectrometer (which makes it much larger). Wacker et al.

(2010) report on routine operation of this new machine, with backgrounds similar to other AMS systems for chemical blanks (ca. $4\text{--}6 \times 10^{-15}$). They also estimated the contribution to the observed blanks from molecular interferences in the ion beam to about $2\text{--}3 \times 10^{-15}$ $^{14}\text{C}/^{12}\text{C}$. In 2010, machines are being constructed for a facility in Mannheim, Germany, and another in Debrecen, Hungary.

The 'Single Stage' AMS

Schroeder et al. (2004) of NEC (Middleton, Wisconsin, USA) reported on the design of another machine, which has since been installed at the University of Lund and included a 300 kV accelerator, also without any pressurized gas tank (Figure 6). This is described as 'single stage AMS', although it still actually has several stages. This device generates the 300 kV by floating the high-energy part of the machine at 300 kV with a commercial power supply, a section of accelerator tube in air provides the acceleration (see Figure 12). The stripper is then at ground potential, and the high-energy mass spectrometry is performed following the stripper without any further acceleration (see Figure 13). This interesting design apparently has similar

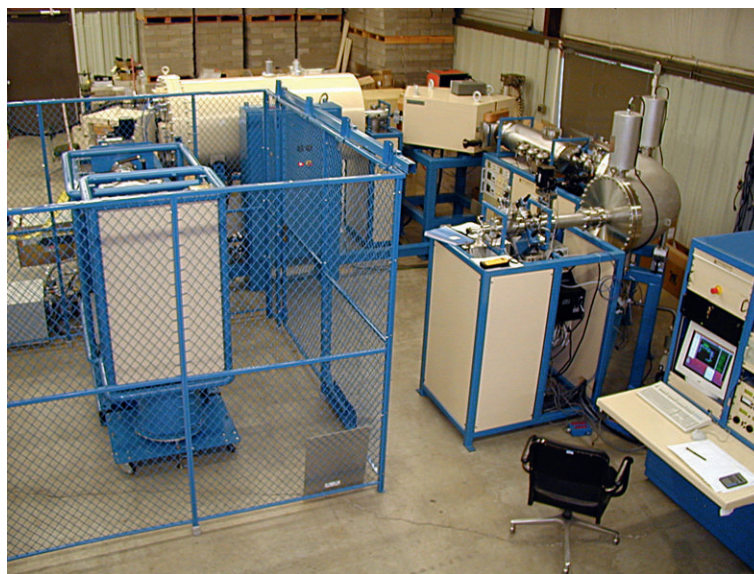


Figure 11 Photographic of the 0.5 MV AMS. Courtesy: National Electrostatics Corporation.

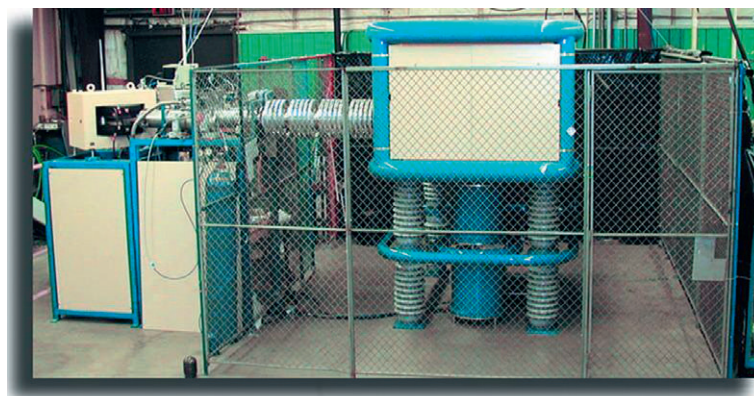


Figure 12 Photograph of the 0.3 MV single stage AMS. Courtesy: National Electrostatics Corporation.

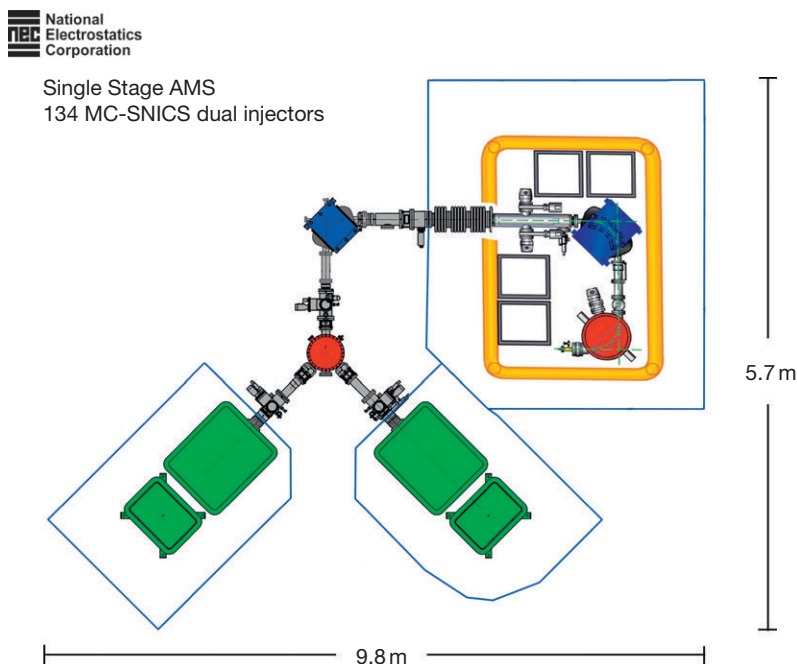


Figure 13 Diagram of the 0.3 MV single stage AMS. Courtesy: National Electrostatics Corporation.

backgrounds to the 0.2–0.5 MV machines and is a new competitor to larger AMS machines for ^{14}C dating (Skog, 2010). These machines are operational at several commercial facilities in the United States for radiocarbon dating and radiocarbon tracer work applied to biomedical applications.

See also: Dating Techniques. **Introduction:** History of Dating Methods. **Radiocarbon Dating:** Causes of Temporal ^{14}C Variations; Conventional Method; Sources of Error.

References

- Bird MI, Ayliffe LK, Fifield LK, et al. (1999) Radiocarbon dating of 'old' charcoal using a wet oxidation, stepped-combustion procedure. *Radiocarbon* 41: 127–140.
- Bruhn F, Duhr A, Grootes PM, Mintrop A, and Nadeau MJ (2001) Chemical removal of conservation substances by 'Soxhlet'-type extraction. *Radiocarbon* 43: 229–237.
- Burr GS, Edwards RL, Donahue DJ, Druffel ERM, and Taylor FW (1992) Mass spectrometric ^{14}C and U-Th measurements in coral. *Radiocarbon* 34: 611–618.
- Burr GS and Jull AJT (2009) Accelerator Mass Spectrometry for radiocarbon research. In: Gross ML and Caprioli R (eds.) *Encyclopedia of Mass Spectrometry*, vol. 5, pp. 656–669. Amsterdam: Elsevier.
- Burr GS and Jull AJT (2010) Accelerator mass spectrometry for radiocarbon research. In: Gross ML and Caprioli R (eds.) *Encyclopedia of Mass Spectrometry*, pp. 656–669. Amsterdam: Elsevier.
- Donahue DJ, Linick TW, and Jull AJT (1990) Isotope-ratio and background corrections for accelerator mass spectrometry radiocarbon measurements. *Radiocarbon* 32: 135–142.
- Fifield LK (1999) Accelerator mass spectrometry and its applications. *Reports on Progress in Physics* 62: 1223–1274.
- Hatté C, Morvan J, Noury C, and Paterne M (2001) Is classical acid-alkali-acid treatment responsible for contamination? An alternative proposition. *Radiocarbon* 43: 177–182.
- Hughey B, Klinkowstein RE, Shefer RE, Skipper PL, Tannenbaum SR, and Wishnok JS (1997) Design of a compact 1 MV AMS system for biomedical research. *Nuclear Instruments and Methods in Physics Research Section B* 123: 153–158.
- Jull AJT and Burr GS (2006) Accelerator mass spectrometry: Is the future bigger or smaller? *Earth and Planetary Science Letters* 243: 305–325.
- Jull AJT, Burr GS, Beck JW, et al. (2003) Using accelerator mass spectrometry for geochronology of the climatic record and connections with the ocean. *Journal of Environmental Radioactivity* 69: 3–19.
- Jull AJT, Burr GS, McHargue LR, et al. (2004) New frontiers in dating of geological, paleoclimatic and anthropological applications using accelerator mass spectrometric measurements of ^{14}C and ^{10}Be in diverse samples. *Global and Planetary Change* 41: 309–323.
- Kim SW, Schneider RJ, von Reden KF, and Hayes JM (2002) Tests of negative ion beams from a microwave ion source with a charge exchange canal for accelerator mass spectrometry applications. *Review of Scientific Instruments* 73: 846–848.
- Kutschera W (2005) Progress in isotope analysis at ultra-trace level by AMS. *International Journal of Mass Spectrometry* 242: 145–160.
- McGehehin J, Burr GS, Jull AJT, et al. (2001) Comparison of sediment dating techniques. *Radiocarbon* 43: 687–690.
- McNichol AP, Jull AJT, and Burr GS (2001) Converting AMS data to radiocarbon values: Considerations and conventions. *Radiocarbon* 43: 313–320.
- Purser KH, Liebert RB, and Russo CJ (1980) MACS – An accelerator radioisotope measuring system. *Radiocarbon* 22: 794–806.
- Schroeder JB, Hauser TM, Klody GM, and Norton GA (2004) Initial results with low energy single stage AMS. *Radiocarbon* 46: 1–4.
- Skog G (2010) Status of the single stage AMS machine at Lund University after 4 years of operation. *Nuclear Instruments and Methods in Physics Research Section B* 268: 895–897.
- Southon J, Santos G, Druffel-Rodriguez K, et al. (2004) The Keck Carbon Cycle AMS Laboratory, University of California, Irvine: Initial operation and a background surprise. *Radiocarbon* 46: 41–49.
- Suter M, Jacob S, and Synal H-A (1997) AMS of ^{14}C at low energies. *Nuclear Instruments and Methods in Physics Research Section B* 123: 148–152.
- Suter M, Jacob SWA, and Synal HA (2000) Tandem AMS at sub-MeV energies. *Nuclear Instruments and Methods in Physics Research Section B* 172: 144–151.
- Synal H-A, Stocker M, and Suter M (2005) MICADAS: A new compact radiocarbon AMS system. In: *Abstracts, 10th International Conference on Accelerator Mass Spectrometry, Berkeley, CA*.
- Synal HA and Wacker L (2010) AMS measurement technique after 30 years: Possibilities and limitations of low energy systems. *Nuclear Instruments and Methods in Physics Research Section B* 268: 701–707.
- Tuniz C, Bird JR, Fink D, and Herzog GF (1998) *Accelerator Mass Spectrometry: Ultrasensitive Analysis for Global Science*. Boca Raton, FL: CRC Press.
- Wacker L, Bonani G, Friedrich M, et al. (2010) MICADAS: Routine and high-precision radiocarbon dating. *Radiocarbon* 52(2): 252–262.

Sources of Error

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Errors and Uncertainties in Radiocarbon Dating

As an estimated carbon-14 (^{14}C) age, X , can be considered as a realization of a random process (due to the nature of radioactive decay), a conceptual model may be considered that states that there is a true but unknown ^{14}C age (often in statistical terminology called a parameter, written as μ), and that our estimated ^{14}C age can be modeled as a realization of a random variable with expected value equal to the true age and variance (or uncertainty) represented by a further unknown parameter, σ^2 . The commonly used probability model for the ^{14}C -dating process is the Gaussian or normal model, defined by the mean which equals the true but unknown ^{14}C age and the standard deviation or measure of uncertainty. This is frequently written as:

$$X \sim N(\mu, \sigma^2)$$

This uncertainty, or σ , is estimated in the error (standard deviation or σ) that is reported with a measured (estimated) ^{14}C age. It is typically based on counting statistics, but may also include components due to other laboratory corrections (e.g., error on fractionation correction). In recent years, there have been several publications where the use of a laboratory error multiplier has been introduced. Such an error multiplier captures sources of variation in the estimated ^{14}C age that are not accounted for in the quoted error.

The interpretation of a single measured ^{14}C age and its associated error is as follows: it is assumed that the measured ^{14}C age (an age estimate) is a realization of a random Gaussian process with a 'true but unknown ^{14}C age' and σ equal to the quoted error. The best estimate of the true ^{14}C age is the measured ^{14}C age, but the true value is highly (95%) likely to lie in the range:

$$\text{Estimated radiocarbon age} \pm 2 \times \text{quoted error}$$

Traceability

It is often important to be able to compare results from different laboratories or from the same laboratory at different times. This is, in many cases, achieved by establishing a chain of calibrations to primary, national, or international standards. Traceability may be demonstrated through comparison of measurement results on a certified reference material, or through using an accepted procedure. Frequently, numerous measurements are needed to define reference materials. This can be achieved as one result of proficiency trials for laboratories or laboratory round-robin/intercomparisons. For ^{14}C dating, the primary standard is 1890 wood, but the reference material most closely associated (and often called the primary standard) is oxalic acid (NBS OxI and NIST OxII). Additional secondary standards have also been developed, including IAEA C1–C6. Such secondary

standards (and further in-house reference materials) are measured on a routine basis, providing in-house quality assurance.

We can now investigate the basis for this conceptual model more carefully and also some of common measurement terms and descriptors that are used.

The Nature of Measurement and Uncertainty

All measurement is subject to uncertainty, which can be considered to have two components, one resulting from the analytical procedure and the other arising from the sample selection process. Every time an analytical measurement is made we get a different result. Typically we cannot analyze the entire sample, it must be selected from the environment it is found in, it may not be fully representative (there may be some heterogeneity, i.e., slightly different ^{14}C composition), so these sources of variation represent the uncertainty due to sampling. Often the emphasis in the laboratory is on the analytical uncertainty, since that is more readily quantified and modeled, but the uncertainty due to sampling is equally important although more difficult to quantify.

An estimate of analytical uncertainty is frequently attached to a measurement (age estimate) and reflects the range of values within which the true value is expected to lie. The uncertainty of a measurement describes the spread of values which might have been observed were the measurement repeated under *identical* conditions. In such a case, the repeat or replicate measurements will be different, and the spread of values is often characterized by a standard deviation, characterizing the precision of measurement but also the natural heterogeneity of the sample material in the field. To estimate the overall uncertainty, it may be necessary to estimate the individual components separately and then combine them using classical error propagation rules.

Measurement Uncertainty and Error

It is important to distinguish between uncertainty and error: error is a single value defined below, uncertainty takes the form of a range.

The error of a measurement, e , is the result of a measurement minus the true value and may include a random component and a systematic component. The error should always be reported with the measurement, and the overall uncertainty is then often expressed as the interval 'measurement $\pm e$ ' or 'measurement $\pm 2e$ '. The actual meaning of such ranges in terms of confidence or 'plausibility' depends on the stochastic model (or probability distribution) chosen to model the random error components. For a Gaussian model, the 1σ and 2σ intervals represent approximately 67% and 95% confidence, respectively.

Accuracy and Precision

Measurements are routinely described as being accurate and/or precise.

Accuracy is the closeness of agreement between a measurement and the true or reference value. If we imagine a series of measurements, each with the same true value, and if the average of the measurements does not equal (within error) the true value, then the measurement is said to be biased, where the bias is the difference between the *expected value* or average of a large series of measurements and the true value. Bias is usually considered to be systematic.

Precision is the closeness of agreement between a series of independent measurements obtained under identical conditions. Precision depends on the distribution of random errors, and is commonly computed as the standard deviation of the results. As the standard deviation increases, the precision decreases.

The archery targets in [Figure 1](#) represent accurate and precise, inaccurate and precise, accurate and imprecise, inaccurate and imprecise measurement from left to right, respectively. These and many other definitions can be found in the Royal Society of Chemistry, [Analytical Methods Committee \(AMC, 2003\)](#).

Uncertainty from Sampling

In the radiocarbon context, we can imagine that the ^{14}C measurement is made on a small component of the whole sample (e.g., a few grains from a silo), or that the archaeological sampling from a midden (which includes the materials to be dated), the security of the sampled materials in connection with the event being dated represent examples of uncertainty due to sampling. These examples also illustrate how difficult it is to quantify such uncertainty.

Stochastic Models

There are many stochastic models, but the most widely used (in ^{14}C dating) is the *Gaussian* or *normal probability model*.

The Gaussian probability model for a ^{14}C measurement X , is written as $X \sim N(\mu, \sigma^2)$, where μ is the expected value of X and σ is the standard deviation.

[Figure 2](#) shows 1000 estimated ages (true age 5000 years before present (BP)), the normal density function, with $\sigma = 30$ years.

Estimating μ and σ

The usual method of estimating μ and σ is based on a series of ^{14}C measurements on the same material made under identical

conditions. These measurements are denoted by $x_1, \dots, x_n$; the estimate of μ is the sample arithmetic average, and that of σ is s , the sample standard deviation. For the example above, the average of all 1000 results is 5001 years (so we could claim the measurement is unbiased) and the sample standard deviation is 31 years.

It is worth noting that the standard error of the mean (s.e.m.) is a measure of precision associated with the estimate of μ and is defined as

$$s/\sqrt{n}$$

thus distinguishing it from the standard deviation.

Estimating Uncertainty

The conventional radiocarbon age is reported as $t \pm \sigma_t$ years BP where σ_t represents the analytical uncertainty on the measurement, as estimated by the laboratory and typically quantified as a standard deviation.

The analytical estimate of σ_t is provided by the laboratory and will typically take into account the analytical uncertainties on the individual components which make up the conventional radiocarbon age.

The most basic form of the radiometric ^{14}C age equation is

$$t = \frac{1}{\lambda} \ln \left[\frac{A_0}{A_t} \right]$$

where t = sample age, which is the time that has elapsed since removal of the carbonaceous material from the carbon cycle (e.g., death of an organism); λ = decay constant, ($\ln 2/t_{1/2}$) where $t_{1/2} = 5568$ years (Libby half-life); A_t = the activity of the carbonaceous material t years after death as measured in the radiocarbon laboratory; and A_0 = 'modern equilibrium living activity' of the sample, again as measured in the radiocarbon laboratory.

A_0 and A_t are measured in the laboratory and both are corrected for fractionation effects. A_0 is measured using standard materials (ideally NIST oxalic acid), while A_t is the activity of the unknown sample. Both will also be adjusted/corrected for background using samples of 'infinite age.'

The analytical uncertainty for t will, therefore, include the uncertainty associated with modern standards, background samples, and fractionation. The actual analytical uncertainty on t (as a function of A_0 and A_t) is again derived through a fairly simple error propagation formula. However, it is also increasingly common that laboratories will take a long series of replicate measurements on a reference material and assess the standard deviation of the set. If the standard deviation is greater than the quoted errors on the individual estimated ages, then this would suggest an unaccounted source of variation and scientists may choose to quote the larger of the standard

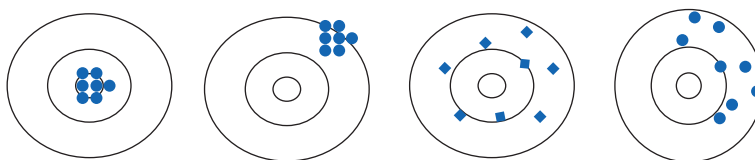


Figure 1 Accuracy and precision.

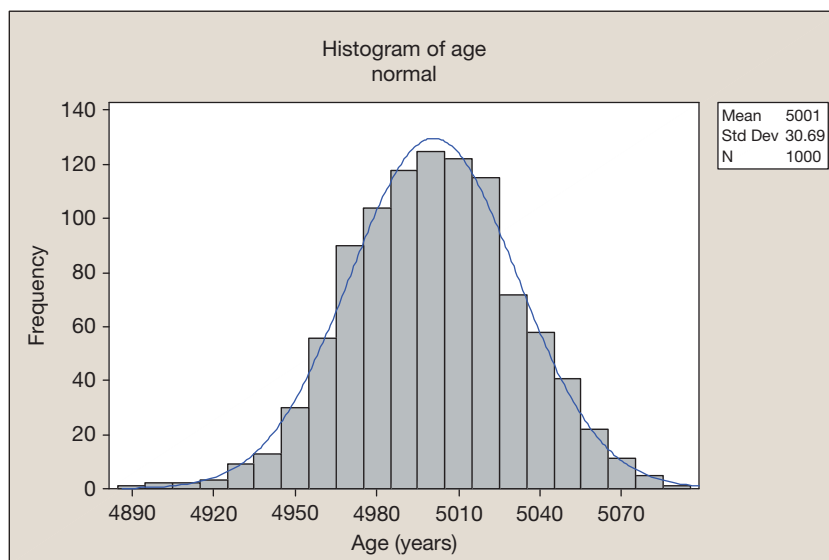


Figure 2 A Gaussian ^{14}C model.

deviation of the set and the individual quoted error, or they may use an error multiplier approach.

How Do We Know Whether the Uncertainty Estimate Is Realistic?

Imagine a situation (known as *repeatability* conditions) in which independent measurements are made on identical samples in the same laboratory, using the same equipment. Such a situation might commonly arise using *reference materials* (typically available in bulk), such as a large quantity of known-age wood. A laboratory will routinely prepare a large number of such samples, each with a measured age and uncertainty. If the uncertainty estimate is realistic, then the between-sample variation should be explained fully by the uncertainties on the individual measurements. If this is not the case, it points to the fact that there are sources of error, contributing to the dispersion of results, which have not been taken into account.

Another measure of whether the uncertainty estimate is realistic can be obtained in a situation (known as *reproducibility* conditions), where independent measurements are obtained by the same method on identical samples in different laboratories. Such a set of conditions is commonly encountered in a *collaborative* or *proficiency trial*.

Interlaboratory Collaborative Trials

Proficiency testing is often used as a method for assessing the accuracy of laboratories in conducting particular measurements. It usually involves distributing 'identical' portions of the test material to each laboratory and analysis of the reported results, which helps each laboratory assess the accuracy of their measurement.

The radiocarbon community has devoted considerable care and effort to establishing and maintaining primary standards and reference materials and in the routine organization of laboratory intercomparisons or collaborative trials to verify

comparability of measurements. From such a series of collaborative trials, secondary standards or reference materials have been developed, including internationally recognized materials such as ANU sucrose (now also known as IAEA-C6), Chinese sucrose, and the IAEA C1–C6 series (Rozanski et al., 1992), augmented by additional oxalic acid samples (now IAEA C7 and C8; Le Clercq et al., 1998). The activity of these materials has been estimated from large numbers of measurements made by many laboratories. Recently, further natural materials from the Third and Fourth International Radiocarbon Intercomparisons (TIRI (Gulliksen and Scott, 1995) and FIRI (Bryant et al., 2000)) have been added to this list. The activities of these standards and reference materials span both the applied ^{14}C age range and the chemical composition range of typical samples. A summary of the intercomparisons organized within the ^{14}C -dating community can be found in Scott (2003). The results from the most recent intercomparisons have indicated that laboratories are generally providing accurate results, but pointed to variation in the results beyond that described by the quoted uncertainties. This provides some evidence that the quoted errors are underestimates of the laboratory precision for some laboratories. Figure 3 shows the results from FIRI (Scott, 2003) for one such reference material; each horizontal line represents the age $\pm 2\sigma$ quoted by a laboratory. Such a plot clearly shows the scatter in the results, but in addition further analysis can reveal evidence of laboratory bias (if present) and also assess the reliability of commonly quoted errors.

Figure 4 shows the difference (offset) between a laboratory's result and the 'true' age (or consensus value) evaluated from the FIRI intercomparison exercise (Scott, 2003). A small number of laboratories showed evidence of small but significant biases.

Calibration

Conventional ^{14}C ages are frequently converted to a calibrated age (on the historical timescale). In radiocarbon terms, this

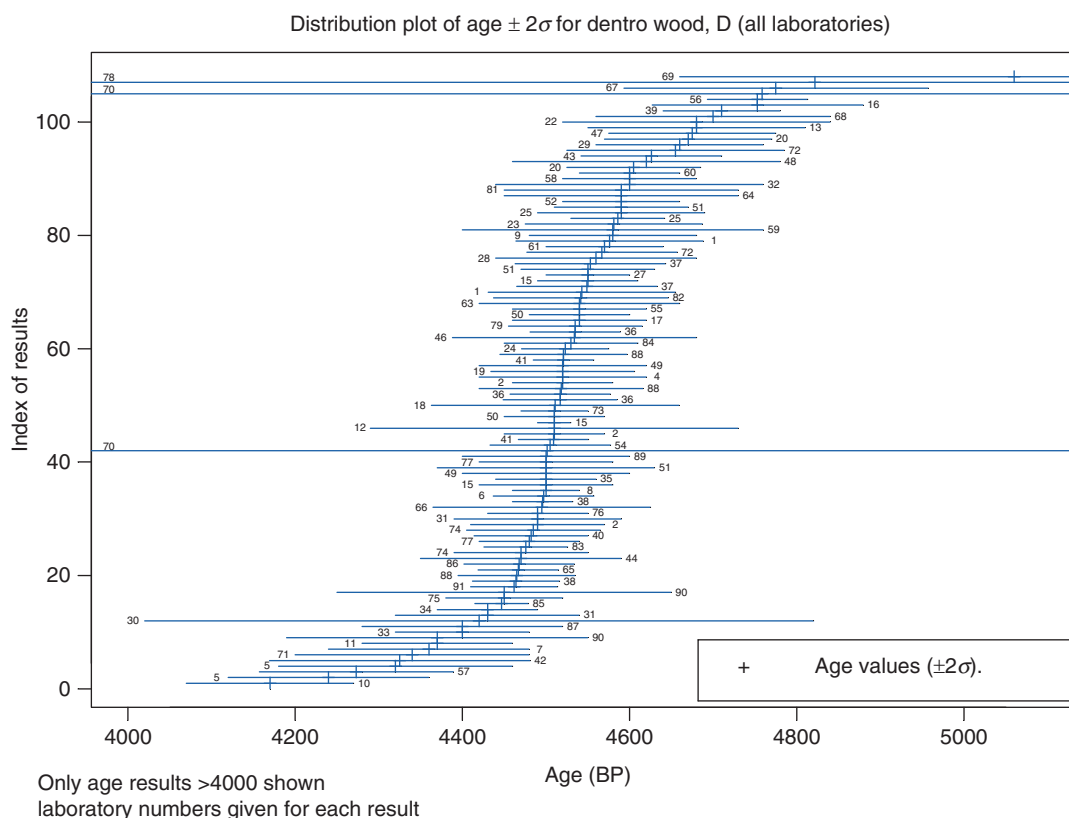


Figure 3 Distribution plot of ages from a collaborative trial. BP, before present.

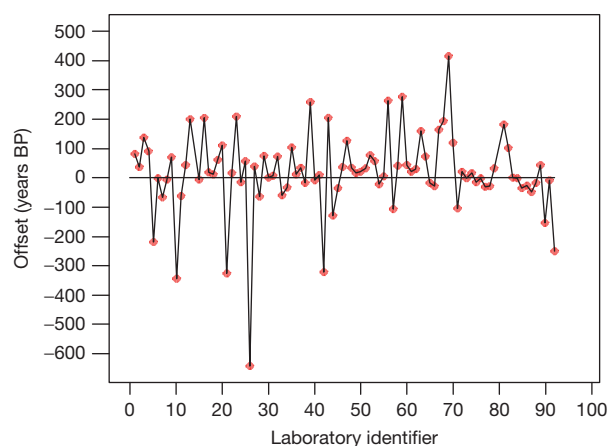


Figure 4 Estimation of bias from a collaborative trial. BP, before present.

means that we have a series of independently known-age samples (such as tree-rings), and in each sample, the ^{14}C age is estimated. The relationship between the estimated age and the true age is modeled and on the basis of the modeled relationship, a ^{14}C age can be calibrated to AD/BC timescale. This relationship has been shown to be of a complex, often nonlinear form, which has significant implications for the transformation of the measured ^{14}C age to a calibrated age

(IntCal04, 2004). The error in years BP on the estimated ^{14}C age must also undergo a transformation to give the corresponding error on the calendar-year scale. However, complexities arise due to the complex pattern of ^{14}C variations. The simplest complication is that although we often model the estimated ^{14}C age using a Gaussian model, once transformed or calibrated, the resulting stochastic model for the calibrated age is no longer Gaussian, it is often multimodal and no longer symmetric. The net effect is that often, the calibrated result has a greater uncertainty. The most commonly used calibration programs adopt a probabilistic approach, resulting in a calibrated-age probability distribution. Figure 5 shows a typical calibration of a single ^{14}C age. Commonly, probabilities are ranked and summed to find the 68.3% (1σ) and 95.4% (2σ) confidence intervals which can result in small, disjoint age ranges. The three most widely used calibration programs are Calib, BCal, and OxCal and an example using OxCal is shown in Figure 5. The main use of the calibration program is for terrestrial samples, although special calibration data sets are available for marine samples. All three programs are based on the same internationally accepted calibration data sets (IntCal04, 2004).

See also: Dating Techniques. **Introduction:** History of Dating Methods. **Radiocarbon Dating:** AMS Radiocarbon Dating; Conventional Method.

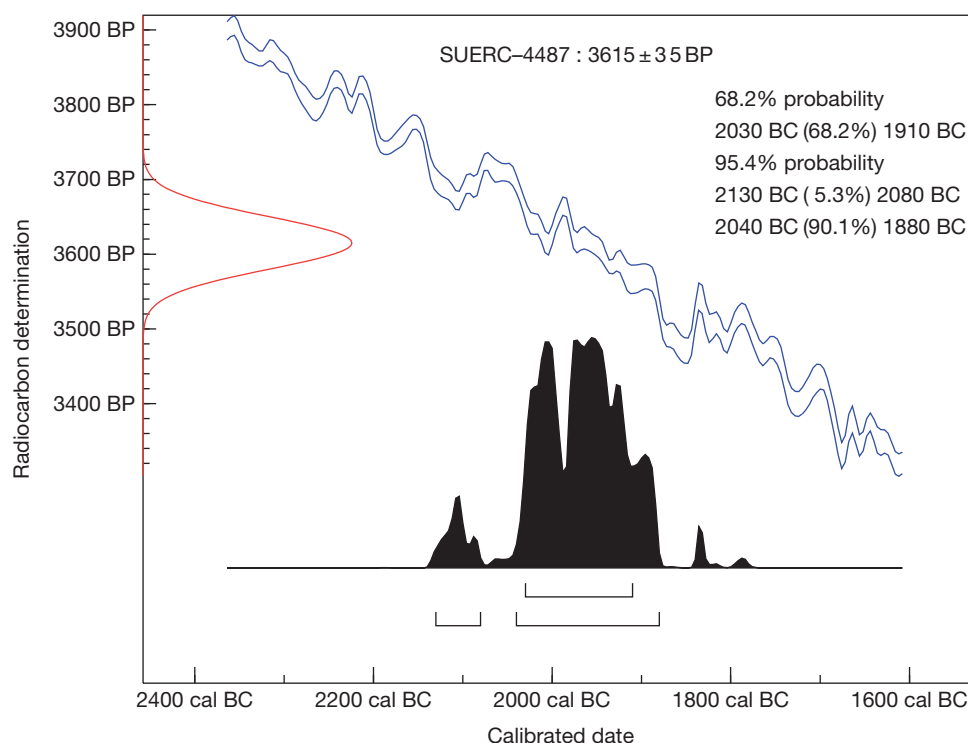


Figure 5 Calibration of a ^{14}C age. BP, before present. Atmospheric data from Stuiver M, Reimer PJ, and Braziunas TF (1998) High-precision radiocarbon age calibration for 1998 terrestrial and marine samples. *Radiocarbon* 40: 1041–1083; OxCal v3.9 Bronk Ramesy (2003); cub r:4 sd:12 prob usp[chron].

References

- Analytical Methods Committee (2003) Part 1: Accuracy, Precision and Uncertainty. Technical Brief No. 13, September 2003. London: Royal Society of Chemistry Available online at: http://www.olinet.com/pdfs/R26_STATISTICS-TERMINOLOGY.pdf.
- Bryant C, Carmi I, Cook G, et al. (2000) Sample requirements and design of an inter-laboratory trial for radiocarbon laboratories. *Nuclear Instruments and Methods in Physics Research B* 172: 355–359.
- Gulliksen S and Scott EM (1995) TIRI report. *Radiocarbon* 37: 820–821.
- IntCal04 (2004) Calibration issue. *Radiocarbon* 46.
- Le Clercq M, van der Plicht J, and Groening M (1998) New ^{14}C reference materials with activities of 15 and 50pMC. *Radiocarbon* 40: 295–297.
- Rozanski K, Stichler W, Gonfiantini R, et al. (1992) The IAEA ^{14}C inter-comparison exercise 1990. *Radiocarbon* 34: 506–519.
- Scott EM (2003) The Third International Radiocarbon Intercomparison (TIRI) and the Fourth International Radiocarbon Intercomparison (FIRI), 1990–2002. Results, analyses, and conclusions. *Radiocarbon* 45: 135–408.

Variations in Atmospheric ^{14}C

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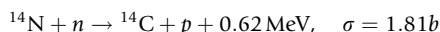
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High-energy cosmic rays, mainly galactic and solar protons, impinge on the Earth's atmosphere. They produce nuclear reactions, with spallation products and neutrons as the main constituents. The high-energy neutrons slow down by scattering and finally reach thermal energies. These thermal neutrons have a large cross-section for a nuclear reaction with nitrogen, the main constituent of the atmosphere (78%): $^{14}\text{N}(n,p)^{14}\text{C}$. Other reactions, such as $^{16}\text{O}(p,3p)^{14}\text{C}$ and $^{16}\text{O}(n,2pn)^{14}\text{C}$, are much less abundant. The ^{14}C nuclei produced are oxidized with atmospheric oxygen to $^{14}\text{CO}_2$ and hence become part of the 'global carbon cycle.' Via photosynthesis in plants, ^{14}C enters the biosphere, which is in equilibrium with the atmosphere. The atmosphere and living terrestrial organisms thus contain ^{14}C in minute concentrations: the natural level is $^{14}\text{C}/^{12}\text{C} \approx 10^{-12}$ corresponding to a level of natural radioactivity of $\sim 225 \text{ Bq kgC}^{-1}$. Upon the death of the organism, the ^{14}C decays as a function of time with a half-life of 5730 years. Thus, by measuring the remaining amount of ^{14}C in fossil remains, the time of death can be established. This is the basis of the ^{14}C (radiocarbon) dating method.

The radiocarbon dating method is complicated by the fact that the natural ^{14}C content is not constant through time. The secular variations in atmospheric CO_2 were first discovered by de Vries (1958). These natural variations cause the ^{14}C timescale to differ from the real (calendar) timescale, resulting in the need for calibration. The ^{14}C dating method and timescale calibration are dealt with elsewhere in this encyclopedia. Here, atmospheric ^{14}C variations, both natural and anthropogenic, and their causes are discussed.

Production

Atmospheric ^{14}C is produced from the interaction of ^{14}N and secondary cosmic ray neutrons by the following exothermic nuclear reaction:



where the cross-section σ is expressed in barn(b): $1b = 10^{-24} \text{ cm}^2$.

The natural production of ^{14}C varies because the flux of primary cosmic ray protons (which produce a cascade of secondary particles, especially neutrons) is not constant. The galactic component of the proton flux is constant, but changes in the screening effect are the major cause of variations in atmospheric ^{14}C production. In the solar system, this screening occurs by the effects of the interplanetary magnetic field and in the vicinity of the Earth, it occurs by the geomagnetic field.

Near the Earth, cosmic ray particles encounter magnetic scattering from the geomagnetic field. This interaction varies greatly with latitude so that the production of ^{14}C is strongly latitude dependent. This is shown in Figure 1, in which the production of ^{14}C (number of nuclei/ cm^2/s) is plotted as a function of geomagnetic latitude. Lingenfelter and Ramaty

(1970) showed that, on average, the global production is $\sim 2.1 \text{ }^{14}\text{C cm}^{-2} \text{ s}^{-1}$.

Variations in the global ^{14}C production as a function of the geomagnetic field can be calculated using

$$\frac{Q(M = M(t))}{Q(M = M(0))} = \left[\frac{M(0)}{M(t)} \right]^{0.5}$$

where $M(t)$ and $M(0)$ are the past and present geomagnetic dipole moment, respectively. The present field is $M(0) = 8 \times 10^{25} \text{ gauss cm}^3$.

In addition, the following empirical relations are found for the production Q (in nuclei/ cm^2/s):

$$Q = 2.310 - 0.024aa$$

and

$$Q = 2.091 - 0.0041 < S >$$

where aa is the so-called geomagnetic aa index, and $<S>$ is the averaged sunspot number, as used by Stuiver and Quay (1980).

A high sunspot activity corresponds to a strong field, resulting in strong shielding, which scatters away more cosmic ray protons from Earth. Hence, less cosmic rays impinge on the Earth's atmosphere, resulting in less production of ^{14}C . In contrast, low sunspot activity corresponds to a weaker field, resulting in increased production of ^{14}C .

The Holocene

By radiocarbon dating of tree-rings of known calendar age, calibration curves are obtained. The most recent calibration curve is INTCAL09, published by Reimer et al. (2009). Using these curves, ^{14}C ages can be transformed into calendar ages, but this procedure is not mathematically straightforward because of the variations of the ^{14}C content in nature. Bronk Ramsey (1998) developed special software to calculate the probability distribution of calibrated ages that includes special procedures for 'wobble-matching' and 'sequencing'.

The dendrochronological part of INTCAL09 is shown in Figure 2. The data are not shown as calibration points on the graph but are plotted as $\Delta^{14}\text{C}$ (expressed in ‰). The $\Delta^{14}\text{C}$ signal is the deviation from the standard ^{14}C activity, corrected for radioactive decay, expressed in ‰. The $\Delta^{14}\text{C}$ signal is calculated as

$$\Delta^{14}\text{C} = \left[\exp \left(\frac{-\text{BP}}{8033} \right) \times \exp \left(\frac{\text{cal BP}}{8267} \right) - 1 \right] \times 1000 (‰)$$

Note that BP is the conventional ^{14}C age (i.e., activity measured relative to the standard), corrected for isotopic fractionation and calculated using the conventional half-life. The standard year corresponds to AD 1950. The historical date relative to AD 1950 is expressed in cal BP. The conventional half-life is $t_{1/2} = 5568$ years, yielding $t_{1/2}/\ln 2 = 8033$; the physical half-life is $t_{1/2} = 5730$ years yielding $t_{1/2}/\ln 2 = 8267$.

Currently, dendrochronology offers calibration back to 12 410 cal BP (Reimer et al., 2009). The absolute German Oak chronology presently extends to 10 429 cal BP, whereas the Preboreal Pine chronology covers the range of 9891–12 410 cal BP. The latter pine chronology was originally a floating chronology; it is now dendrochronologically matched to the absolute oak chronology, hence the pine chronology is absolute. The dendrochronological part of INTCAL09, thus, covers the complete Holocene and reaches back halfway into the Younger Dryas period.

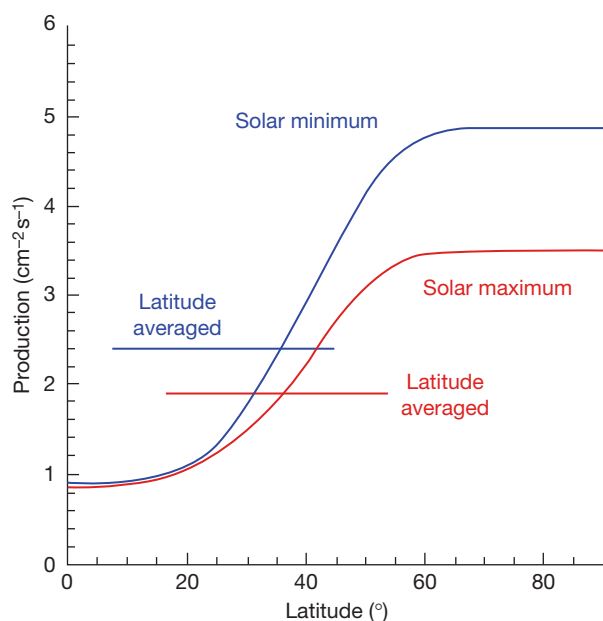


Figure 1 Natural production of ^{14}C (number of nuclei/cm 2 /s) as a function of geomagnetic latitude.

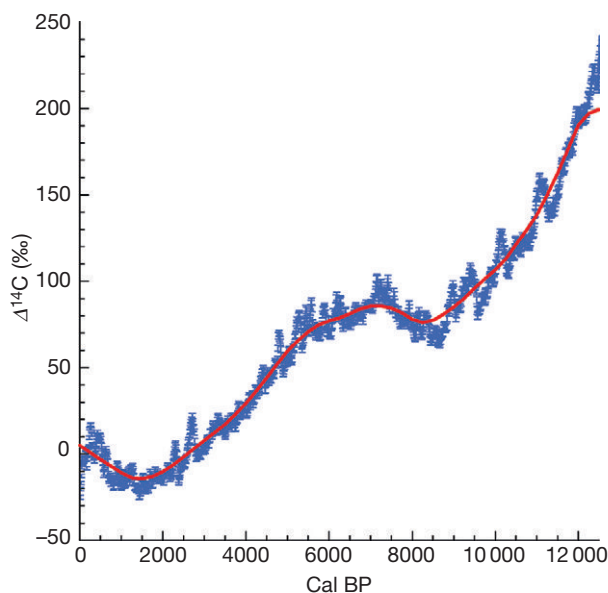


Figure 2 Atmospheric ^{14}C content for the past 12 500 years, expressed in $\Delta^{14}\text{C}$ (‰). The $\Delta^{14}\text{C}$ is calculated from the INTCAL09 calibration dataset. Only the data derived from dendrochronology are shown. A long-term smoothing function is drawn through the data.

Smoothed curves can be drawn through the $\Delta^{14}\text{C}$ data, as is shown in Figure 2. This represents the long-term trend of the ^{14}C dataset. It had already been noted in the early days of radiocarbon that the data show a quasi-sinusoidal variation for the last 7000 years. Damon et al. (1989) showed that the data can be fitted with the following formula:

$$\Delta^{14}\text{C}(\text{‰}) = 35 + 51 \sin(5.826 \times 10^{-4} \text{ cal BP} - 2.401)$$

This function corresponds to a period of 10 780 years, an amplitude of 51‰ and a phase lag of 2.401 radians or 137.6°. The $\Delta^{14}\text{C}$ data fitted with this sine function are shown in Figure 3. Also shown are summarized data obtained by McElhinny and Senanayake (1982) of archaeomagnetic determinations of the Earth's magnetic dipole moment. These magnetic data show quasi-sinusoidal variation for the past ca. 10 000 years, with a maximum of about 40% above present-day value (ca. 6000 cal BP) and a minimum of 80% of present-day value (ca. 6000 cal BP). This shows that the long trend in the $\Delta^{14}\text{C}$ data is consistent with the variations in the geomagnetic values. The inverse relation between $\Delta^{14}\text{C}$ and the geomagnetic field is due to modulation of the ^{14}C production by the Earth's dipole moment.

Also, further back in time throughout the Holocene, the ^{14}C production rate variation induced by change in the geomagnetic dipole intensity appears to cause most of the long-term $\Delta^{14}\text{C}$ trend, as shown by Stuiver and Braziunas (1993).

Figure 4 shows the residual $\Delta^{14}\text{C}$ obtained by subtracting the long-term trend as shown in Figure 2. This means that for natural $\Delta^{14}\text{C}$ variations, the geomagnetic component is removed. The wiggles around this long-term trend, known as 'Suess wiggles,' are referred to as fine structures by Damon and Peristykh (2000). They represent forcing mechanisms other than the geomagnetic influence for cosmogenic isotope production that remain, such as solar, ocean, and possibly climatic forcing.

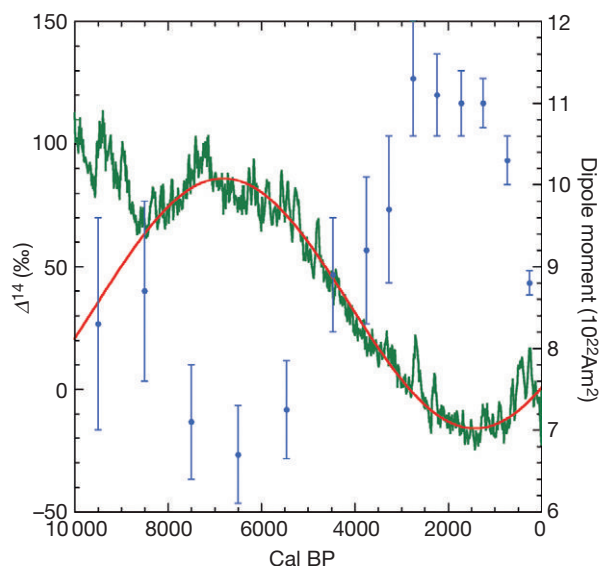


Figure 3 The atmospheric $\Delta^{14}\text{C}$ signal for the last 10 000 years, compared with archaeomagnetic determinations of the strength of the Earth's dipole moment. Data from McElhinny and Senanayake (1982).

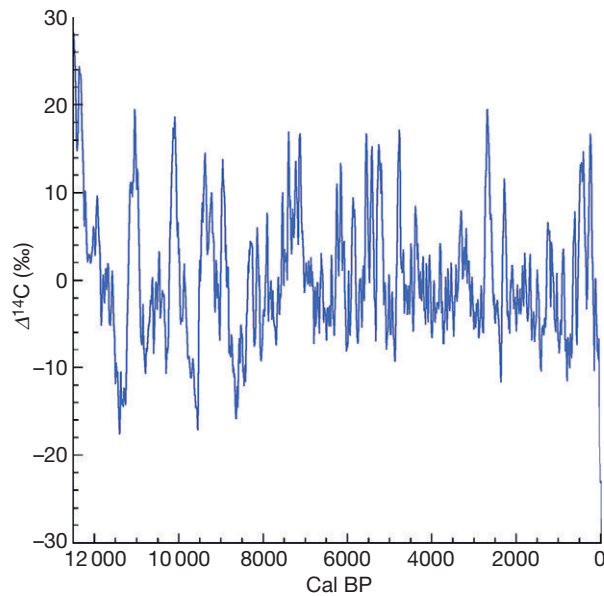


Figure 4 Residual $\Delta^{14}\text{C}$ obtained by deducting the long-term trend as shown in [Figure 2](#).

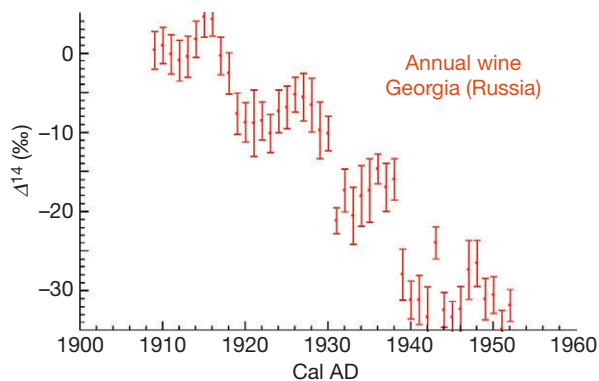


Figure 5 A 50-year record of annual wine samples from the Republic of Georgia showing the 11-year sunspot cycle. Data from [Burchuladze et al. \(1980\)](#).

For the past millennium, this fine structure has been recognized in the known solar minima: the Dalton, Maunder, Spörer, Wolf, and Oort minima around AD 1825, 1700, 1500, 1350, and 1050, respectively. These solar minima correspond with an increase in $\Delta^{14}\text{C}$. Note that observation of sunspots has demonstrated the absence or dearth of sunspots during the Maunder Minimum, as discussed by [Eddy \(1976\)](#). This absence of sunspots is associated with lower solar wind intensity, and an increase in cosmic ray intensity and subsequent increased production of ^{14}C . Frequency analysis of the $\Delta^{14}\text{C}$ record shows dominant periods of approximately 2250, 208, and 89 years, as shown by [Damon and Peristykh \(2000\)](#) and [Damon et al. \(1989\)](#). Very prominent peaks, such as the 89-year cycle (the so-called Gleisberg cycle), are clearly solar in origin. There is evidence for lower frequency cycles (e.g., a 512-year period) that may have an origin in oceanic effects ([Damon and Peristykh, 2000](#)). In addition, the well-known 11-year solar cycle is present albeit with small amplitude. This is referred to as a ‘hyperfine’

structure and has been measured with high precision in annual wine samples from the Republic of Georgia ([Burchuladze et al., 1980](#)). These data are plotted in [Figure 5](#).

The ca. 2000-year oscillation in the $\Delta^{14}\text{C}$ record during the Holocene is synchronous with fluctuations in another cosmogenic isotope record, ^{10}Be , and climatic deterioration as observed in ice-rafted debris from deep sea cores. This observation in variability of the ^{14}C signal, therefore, plays an important role in climate-forcing research and solar effects, as discussed by [Bond et al. \(2001\)](#).

The tree-ring-based record of varying $\Delta^{14}\text{C}$ is used by [Solanki et al. \(2004\)](#) for reconstruction of the sunspot record during the Holocene. It shows that the current episode of high sunspot activity is the most intense for several thousand years. However, without a full understanding of the chain of cause and effect, we cannot be sure exactly how the changing Sun impacts our climate.

In this respect, it is interesting to note that rapid climatic oscillations, known as Dansgaard–Oeschger events, appear during the Late Glacial with a periodicity of ca. 1470 years. Modeling experiments by [Braun et al. \(2005\)](#) suggest that this climate cycle could have been triggered by solar forcing, despite the fact that a 1470-year solar cycle is absent.

The Deglaciation

For the Holocene, ^{14}C variations are well known and observed in the tree-ring record. The ^{14}C oscillations on decadal and century time scales can be explained as solar driven. For the Late Glacial and deglaciation periods, however, atmospheric ^{14}C has been influenced by ocean-related ^{14}C redistribution between the carbon reservoirs ([Stuiver et al., 1991](#)). The postulated mechanism for oceanic $\Delta^{14}\text{C}$ change is via the unstable thermohaline circulation in the North Atlantic, in which lowered salinity results in increased water stratification and reduced deep water formation. The latter, in turn, parallels reduced upwelling of ^{14}C -deficient water elsewhere in the oceans, thus increasing surface ocean and atmospheric $\Delta^{14}\text{C}$ levels ([Stuiver and Braziunas, 1993](#)). Simulations show that ventilation changes can cause fluctuations of 25‰ in atmospheric $\Delta^{14}\text{C}$. The rapid atmospheric $\Delta^{14}\text{C}$ rise at the onset of the Younger Dryas (50‰ in 200 calendar years), however, remains unexplained, as discussed by [Marchal et al. \(2001\)](#).

The tree-ring record is the most powerful tool for observing natural ^{14}C variations, because they are absolutely dated, and they represent an atmospheric ^{14}C signal. The dendrochronological record currently reaches 12 450 cal BP, well into the Younger Dryas interstadial ([Reimer et al., 2009](#)). This record is the backbone of the radiocarbon calibration curve. The latest calibration curve, INTCAL09 ([Reimer et al., 2009](#)), has been extended beyond the tree-ring record back to 50 000 cal BP. This extension is based on marine datasets: Foraminifera and corals dated by ^{14}C and corrected for a site-specific marine reservoir effect. The Foraminifera are from the Cariaco Basin of the coast of Venezuela, collected from laminated sediment which is matched to the absolute time scale ([Hughen et al., 2006](#)); the corals are from various Pacific Ocean sites and dated by U-series isotopes. They are selected to meet criteria established by the INTCAL working group ([Reimer et al., 2009](#)).

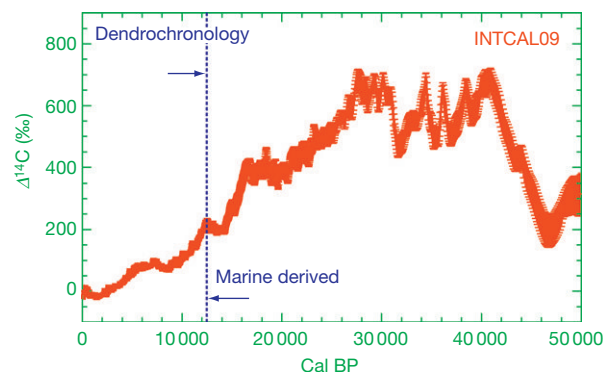


Figure 6 The atmospheric $\Delta^{14}\text{C}$ signal from INTCAL09 for the complete time range of the ^{14}C dating method (Reimer et al., 2009). This record is based on dendrochronology and (beyond ca. 12 400 years ago) on marine records.

Thus, the calibration curve for the time range beyond 12 450 cal BP is marine derived. The $\Delta^{14}\text{C}$ signal, corresponding to the complete INTCAL09 dataset, is shown in Figure 6 (and is an extension of Figure 2).

The Middle/Upper Paleolithic

The time range of the ^{14}C dating method is about 50 000 years; in exceptional cases, 60 000 years can be reached. Variations in the natural ^{14}C content are known from the calibration records that constitute INTCAL09 (Reimer et al., 2009). For the time range beyond the tree-ring record, that is, 12 410 cal BP, this period is currently developed from marine records. In general, a variety of data exist showing natural ^{14}C variations. They are derived from various natural archives, dated by both ^{14}C and another, independent method; for example, terrestrial macroremains (dated by ^{14}C) from laminated sediments (dated by varve counting) as measured by Kitagawa and van der Plicht (1998) and speleothems dated by both ^{14}C and U-series isotopes as measured by Hoffmann et al. (2010). All dating methods for such records have their advantages and disadvantages. The difficulties include such things as nonatmospheric ^{14}C materials (carbonates) that need reservoir corrections, varve counting errors, uncertainties in U-series isotope geochemistry, and uncertainties in ice core chronologies. For a more complete discussion of such records, see van der Plicht (2000). All these datasets show natural ^{14}C variations, some of them with large and thus far unexplained magnitudes.

Until recently, there was too much inconsistency between the datasets, such that calibration was not recommended because records could differ by as much as millennia (van der Plicht et al., 2004). As chronology is an essential ingredient for the Middle/Upper Paleolithic, this spawned debates in archaeology (Bronk Ramsey et al., 2006).

An overview of the most important data available for the complete ^{14}C dating range of 50 000–60 000 years is shown in Figure 7, which can be viewed as an extension of Figure 2 (showing the dendrochronological record). For clarity, the dendrochronological record is not shown again here. The data are as follows: terrestrial macrofossils from laminated

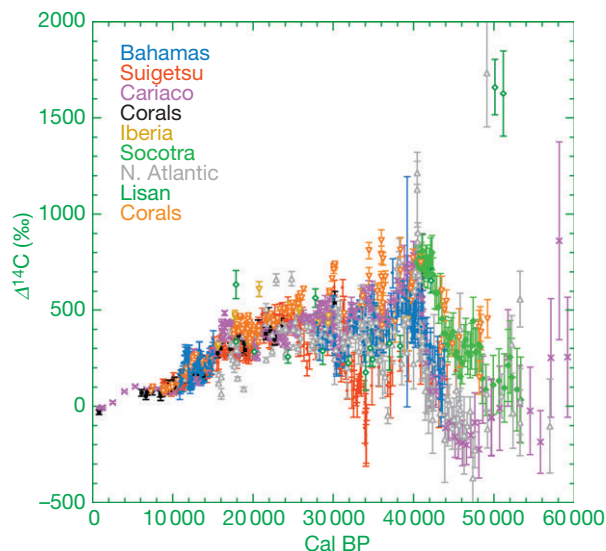


Figure 7 The $\Delta^{14}\text{C}$ data from selected individual 'calibration' records beyond dendrochronology. Note that errors along the horizontal axis are not shown.

sediment from Lake Suigetsu, Japan (Kitagawa and van der Plicht, 1998); a speleothem record from the Bahamas (Hoffmann et al., 2010); Foraminifera from the Cariaco Basin (Hughen et al., 2006); Foraminifera from the Iberian Sea (Bard et al., 2004); Atlantic and Pacific corals (Fairbanks et al. (2005); and furthermore, a speleothem record from Socotra, Arabia; North Atlantic Foraminifera; and calcite from Lake Lisan (Israel) deposits. For full references, see van der Plicht et al. (2004) or Reimer et al. (2009).

The data shown here are plotted as $\Delta^{14}\text{C}$. Note that for calculating $\Delta^{14}\text{C}$, the absolute age must be included, so the plot may be misleading. In addition, the errors along the horizontal axis (the 'absolute timescale') are not shown because they are not known.

Older data with large errors (based on alpha spectrometric U-series measurements, TL/OSL/ESR dates, etc.) are not shown here – see van der Plicht (2000) for these data and references listed.

The extremes are represented by the Japanese varve record (Kitagawa and van der Plicht, 1998) and the Bahamian speleothem (Hoffmann et al., 2010). In the Japanese varve record, the natural ^{14}C content – which is elevated back to 26 000 cal BP – tends to go back to present levels around 30 000 cal BP. Note that this record is the only one based on terrestrial/atmospheric sample material. Its weakness is the possible presence of hiatuses, which cannot be excluded. The other extreme is represented by the speleothem records, for which carbonate material is dated by both ^{14}C and U-series isotopes.

Excursions are shown in the Cariaco Basin record by Hughen et al. (2004), at ca. 35 000 and 41 000 cal BP. These correspond to geomagnetic excursions known as Mono Lake and Laschamps, respectively.

The Cariaco curve is almost an 'average' of the Bahamian speleothem and the Japanese varve datasets (see Figure 7). Note that the absolute timescale for the Cariaco Basin record is the GISP2 ice core timescale of Meese et al. (1997).

The production of cosmogenic isotopes – ^{14}C and also ^{10}Be , ^{26}Al , ^{36}Cl , ^{41}Ca – and variations therein on the century/millennium timescale are determined mainly by the geomagnetic field strength.

Mono Lake and Laschamp-type excursions are observed in cosmogenic isotopes, other than ^{14}C , in various records: for ^{10}Be in Antarctic ice cores, Greenland ice cores and in marine sediments; and for ^{36}Cl in Greenland ice cores and terrestrial packrat mittens. References can be found in the review by van der Plicht (2000).

Paleomagnetic fields can be obtained from remanence measurements, such as data from Tric et al. (1992), as shown in Figure 8. Other measurements show a general similar pattern: high values around 50 ka ago, followed by a distinct minimum around 40 ka.

When comparing Figures 7 and 8, it appears that, in general, the ^{14}C levels in the atmosphere compared with the geomagnetic dipole moment over the past 50 ka reveal a clear inverse relationship. Quantitatively, the ^{14}C variations appear to be of the right order of magnitude.

Differences between various paleomagnetic records can be ascribed to the effect of localized nondipole fields and to variations in the magnetic properties of the minerals in the sediment. It is difficult to derive the global dipole moment from the data.

The cosmogenic isotope ^{14}C is part of the global C cycle and depends on the dynamics of the carbon system on Earth. Most of the exchangeable carbon is dissolved and held in the deep oceans, which have a relatively slow turnover time. The result is that the ^{14}C levels in the atmosphere and terrestrial biosphere lag behind any change in the production rate caused by an alteration in the geomagnetic dipole field intensity. A further

complication is a changing ventilation rate of the oceans during glaciation.

Paleomagnetic fields can also be reconstructed using cosmogenic isotope deposition records. Geomagnetic field intensity variations, however, are not the only factor that has influenced the cosmogenic nuclide records. In the case of ^{14}C , variations of the global carbon cycle (caused by reorganizations of the ocean circulation patterns from the Last Glacial to the present interglacial) and solar variations are superimposed.

The ^{10}Be nuclides are not affected by these variations because they are part of a very different geochemical cycle. The signal depends on precipitation and climate; in addition, its deposition rates in marine sediments vary as a function of lateral sediment redistribution.

Mazaud et al. (1991) constructed the ^{14}C production rate based on the data from Tric et al. (1992) as shown in Figure 8. Muscheler et al. (2005) calculated cosmogenic isotope production based on ^{10}Be deposition rates.

Paleomagnetic field reconstruction based on cosmogenic isotopes has the advantage that it is complementary to methods based on remanence measurements. This production is sensitive mainly to the dipole moment as the shielding effect of the nondipole components decreases faster with increasing distance from Earth. The modulation of cosmic rays takes place in space, far outside the atmosphere, where the effect of nondipole fields is negligible.

In contrast, remanence measurements in volcanic rocks or sediments are based on spot measurements on the Earth's surface that might be influenced by changes in quadrupole (or higher) moments of the field.

The Anthropogenic Era

During the past century, the atmospheric ^{14}C signal has largely been influenced by two different anthropogenic effects: fossil fuel burning and nuclear explosions. Fossil fuels have geological age; hence, all of their ^{14}C has decayed. The combustion of fossil fuels adds ^{14}C -free CO_2 into the atmosphere; this dilution causes the atmospheric ^{14}C content to decrease. This effect (known as the 'Suess effect') is clearly seen in (pre)modern tree-rings (Figure 9(a)). Since the onset of the industrial revolution, the atmospheric ^{14}C content had decreased by 2% in 1950. Since ca. 1950, the Suess effect is no longer directly visible because a large amount of artificial ^{14}C has been injected into the atmosphere by nuclear explosions. As a result, the ^{14}C concentration in the atmosphere increased in 1963 by a factor of 2 in the Northern Hemisphere and with a somewhat smaller factor in the Southern Hemisphere. Since the Nuclear Test Ban Treaty came into effect in 1963, the ^{14}C concentration in the atmosphere has been decreasing due to rapid exchange between atmosphere and other carbon reservoirs, mainly the ocean and the biosphere. A selection of data is shown in Figure 9(b) (see Meijer et al., 1994 and references therein).

The large pulse of artificial ^{14}C injected into the atmosphere (the so-called 'bomb peak') enables the use of ^{14}C as a tracer for studying the global carbon cycle (Levin and Hessheimer, 2000; Levin et al., 2008). The bomb peak is a spike that can be

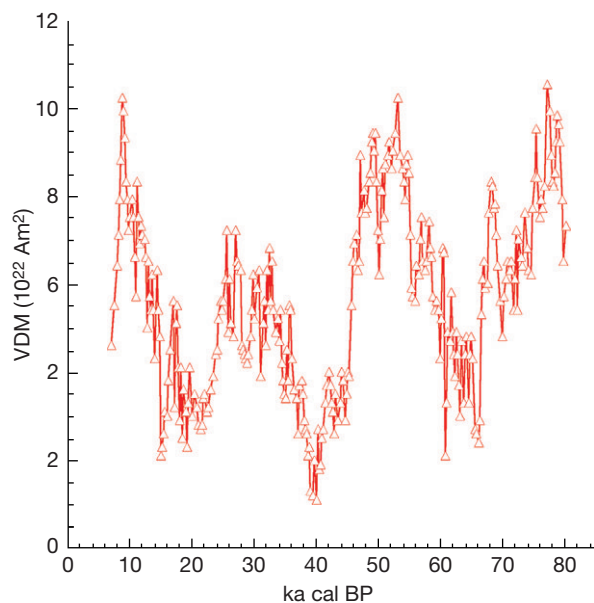


Figure 8 Geomagnetic dipole field (virtual dipole moment, VDM) changes for the past 80 000 years based on paleomagnetic measurements. Reproduced from Tric E, Valet JP, Turcholka P, et al. (1992) Paleointensity of the geomagnetic field during the last 80 000 years. *Journal of Geophysical Research* 97(B6): 9337–9351.

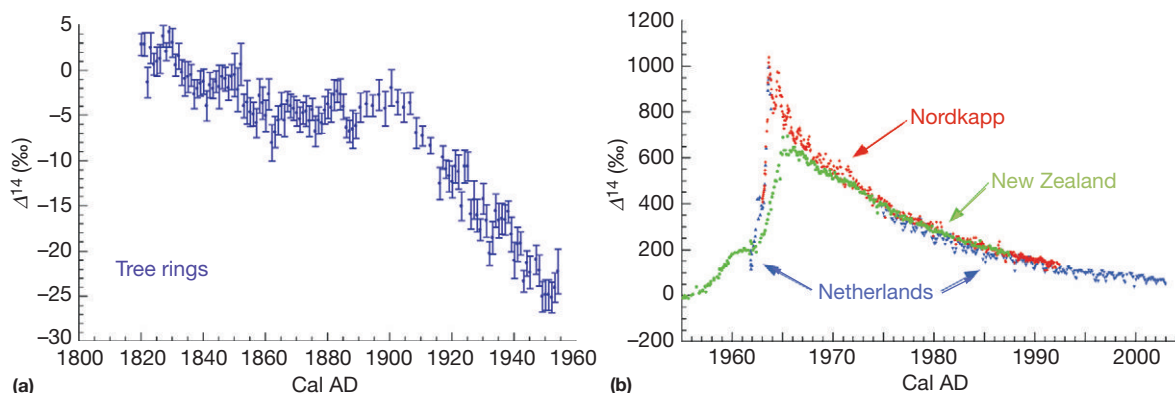


Figure 9 Modern anthropogenic influence on the atmospheric ^{14}C content: (a) the Suess effect – burning of fossil fuels caused the atmospheric $\Delta^{14}\text{C}$ signal to decrease until ca. 1950 (data from Stuiver and Quay, 1980); (b) the bomb effect – until 1963, the atmospheric $\Delta^{14}\text{C}$ signal increased sharply due to atmospheric nuclear explosions.

used for dating modern soils and peat deposits. In addition, the bomb peak can be used for dating in forensic and biomedical applications.

Conclusion

The cosmogenic isotope ^{14}C has many applications in the dating of late quaternary events, particularly the dating of organic remains back to 50 000 years ago. The ^{14}C timescale needs to be calibrated into calendar ages because of variations in the atmospheric ^{14}C content through time. These variations yield scientific information on matters such as forcing of climate change, carbon cycle research, cosmogenic isotope production, and solar fluctuations.

The ^{14}C calibration record, which yields the $\Delta^{14}\text{C}$ signal, is well known for the Holocene based on tree-ring records, which extend back to more than 12 000 years ago. The curve is extended back to 50 000 years ago based on marine records. There has been remarkable progress in obtaining calibration datasets and, thus, a record for past $\Delta^{14}\text{C}$. But the oldest part of the record is still to be considered as ‘work in progress,’ as not all terrestrial records are consistent and there are uncertainties in ^{14}C reservoir corrections for the marine records. What is still needed is a set of well-established anchor points for both timescales – ^{14}C and absolute. A truly terrestrial calibration curve may not be too far away, if chronologies based on Kauri trees can be established (Balter, 2006). This would provide a record of natural $\Delta^{14}\text{C}$ variations without ambiguities, giving a better insight into changes in the Earth’s system in the past 50 000 years.

See also: **Radiocarbon Dating: Calibration of the ^{14}C Record; Causes of Temporal ^{14}C Variations.**

References

- Balter M (2006) Radiocarbon dating’s final frontier. *Science* 313: 1560–1563.
- Bard E, Rostek F, and Ménot-Combes G (2004) Radiocarbon calibration beyond 20,000 yr BP by means of planktonic foraminifera of the Iberian Margin. *Quaternary Research* 61: 204–214.
- Bond G, Kromer B, Beer J, et al. (2001) Persistent solar influence on North Atlantic climate during the Holocene. *Science* 294: 2130–2136.
- Braun H, Christl M, Rahmstorf S, et al. (2005) Possible solar origin of the 1470 year glacial climate cycle demonstrated in a coupled model. *Nature* 438: 208–211.
- Bronk Ramsey C (1998) Probability and dating. *Radiocarbon* 40: 461–474.
- Bronk Ramsey C, Buck CE, Manning SW, Reimer P, and van der Plicht J (2006) Developments in radiocarbon calibration for archaeology. *Antiquity* 80: 783–798.
- Burchuladze AA, Pagava SV, Povinec P, Togonidze GI, and Usacev S (1980) Radiocarbon variations with the 11-year solar cycle during the last century. *Nature* 287: 320–322.
- Damon PE, Cheng S, and Linick TW (1989) Fine and hyperfine structure in the spectrum of secular variations of atmospheric ^{14}C . *Radiocarbon* 31: 704–718.
- Damon PE and Peristykh AN (2000) Radiocarbon calibration and application to geophysics, solar physics, and astrophysics. *Radiocarbon* 42: 137–150.
- De Vries H (1958) Variation in concentration of Radiocarbon with time and location on earth. *Proceedings Koninklijke Nederlandse Akademie Wetenschappen B* 61(2): 1–9.
- Eddy JA (1976) The Maunder minimum. *Science* 192: 1189–1202.
- Fairbanks RG, Mortlock RA, Chiu TC, et al. (2005) Radiocarbon calibration curve spanning 0 to 50,000 years B.P. based on paired $^{230}\text{Th}/^{234}\text{U}/^{238}\text{U}$ and ^{14}C dates on pristine corals. *Quaternary Science Reviews* 24: 1781–1796.
- Hoffmann DL, Beck JW, Richards DA, et al. (2010) Towards radiocarbon calibration beyond 28 ka using speleothems from the Bahamas. *Earth and Planetary Science Letters* 289: 1–10.
- Hughen K, Southon J, Lehman S, Bertrand C, and Turnbull J (2006) Marine-derived ^{14}C calibration and activity record for the past 50,000 years updated from the Cariaco Basin. *Quaternary Science Reviews* 25: 3216–3227.
- Kitagawa H and van der Plicht J (1998) Atmospheric radiocarbon calibration to 45,000 yr BP: Late Glacial fluctuations and cosmogenic isotope production. *Science* 279: 1187–1190.
- Levin I, Hammer S, Kromer B, and Meinhardt F (2008) Radiocarbon observations in atmospheric CO_2 : Determining fossil fuel CO_2 over Europe using Jungfraujoch observations as background. *Science of the Total Environment* 391: 211–216.
- Levin I and Hessheimer V (2000) Radiocarbon – A unique tracer of global carbon cycle dynamics. *Radiocarbon* 42: 69–80.
- Lingenfelter RE and Ramaty R (1970) Astrophysical and geophysical variations in ^{14}C production. In: Olsson IU (ed.) *Radiocarbon Variations and Absolute Chronology*. 12th Nobel Symposium, Proceedings, pp. 513–537. New York: Wiley.
- Marchal O, Stocker TF, and Muscheler R (2001) Atmospheric radiocarbon during the Younger Dryas: Production, ventilation, or both? *Earth and Planetary Science Letters* 185: 383–395.
- Mazaud A, Laj C, Bard E, Arnold M, and Tric E (1991) Geomagnetic field control of ^{14}C production over the last 80 ky: Implications for the radiocarbon time scale. *Geophysical Research Letters* 18: 1885–1888.
- McElhinny MW and Senanayake WE (1982) Variations in the geomagnetic dipole 1: The past 50000 years. *Journal of Geomagnetism and Geoelectricity* 34: 39–51.
- Meese DA, Gow AJ, Alley RB, et al. (1997) The Greenland ice sheet project-2 depth-age scale: methods and results. *Journal of Geophysical Research* 102(C12): 26411–26424.
- Meijer HAJ, van der Plicht J, Gislefoss JS, and Nydal R (1994) Comparing long term atmospheric ^{14}C and ^3H records near Groningen, the Netherlands with Fruholmen, Norway and Izana, Canary Islands ^{14}C stations. *Radiocarbon* 37: 39–50.

- Muscheler R, Beer J, Kubik PW, and Synal HA (2005) Geomagnetic field intensity during the last 60000 years based on ^{10}Be and ^{36}Cl from the Summit ice cores and ^{14}C . *Quaternary Science Reviews* 24: 1849–1860.
- Reimer PJ, Baillie MGL, Bard E, et al. (2009) IntCal09 and Marine09 radiocarbon age calibration curves, 0–50,000 years cal BP. *Radiocarbon* 51: 1111–1150.
- Shackleton NJ, Fairbanks RG, Chiu TC, and Parrenin F (2004) Absolute calibration of the Greenland time scale: Implications for Antarctic time scales and for $\Delta^{14}\text{C}$. *Quaternary Science Reviews* 23: 1513–1522.
- Solanki SK, Usoskin LG, Kromer B, Schüssler M, and Beer J (2004) Unusual activity of the sun during recent decades compared to the previous 11000 years. *Nature* 431: 1084–1087.
- Stuiver M and Braziunas TF (1993) Sun, ocean, climate and atmospheric CO_2 : An evaluation of causal and spectral relationships. *The Holocene* 3: 289–305.
- Stuiver M, Braziunas TF, Becker B, and Kromer B (1991) Climatic, solar, oceanic and geomagnetic influences on Late-Glacial and Holocene atmospheric $^{14}\text{C}/^{12}\text{C}$ change. *Quaternary Research* 35: 1–24.
- Stuiver M and Quay P (1980) Changes in atmospheric carbon-14 attributed to a variable sun. *Science* 207: 11–19.
- Tric E, Valet JP, Tucholka P, et al. (1992) Paleointensity of the geomagnetic field during the last 80,000 years. *Journal of Geophysical Research* 97(B6): 9337–9351.
- van der Plicht J (2000) Varve/comparison issue. *Radiocarbon* 42: 313–452.
- van der Plicht J, Beck JW, Bard E, et al. (2004) NOTCAL04: Comparison/calibration ^{14}C records 26–50 cal kyr BP. *Radiocarbon* 46: 1225–1238.

Causes of Temporal ^{14}C Variations

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Introduction

A principal assumption underlying the radiocarbon dating method is that the amount of radiocarbon (^{14}C) in the atmosphere is constant over the useful range of the technique (Libby, 1955). In the calculation of a radiocarbon age, the ^{14}C content of a sample is measured relative to a standard, and that standard is chosen to reflect a fixed atmospheric ^{14}C value. It is now known that the ^{14}C content of the atmosphere is not invariant; however, radiocarbon age calculations continue to assume constant atmospheric ^{14}C . This is done with the caveat that ages are quoted in radiocarbon years and are distinguished from calendar years. To account for the difference, radiocarbon years are calibrated to calendar years using detailed information about the radiocarbon content of the atmosphere at the time the sample formed.

The key to understanding atmospheric radiocarbon variations through time is the development of time series records that reflect prehistoric atmospheric ^{14}C values. The most precise time series of this type is based on tree rings. Trees absorb carbon dioxide and sample the atmosphere as they grow. Certain trees form distinct annual rings and can be dated very accurately using dendrochronology. Dendrochronological dates are used to fix the calendar age of a tree ring, and ^{14}C measurements on the wood determine the atmospheric radiocarbon at the time the tree grew. The first measurements of this type began in the 1950s when it was observed that the radiocarbon content of the atmosphere was not static (de Vries, 1958; Suess, 1954). Today, an internationally agreed-upon composite tree-ring record (INTCAL09, Reimer et al., 2009) is available to convert radiocarbon dates to calendar dates. Implicit in this conversion is a correction for temporal variability in atmospheric radiocarbon. The current tree-ring record contains thousands of radiocarbon measurements and provides a continuous record with annual to decadal resolution for about the last 12 600 years. The curve is a compilation of results for trees from North America and Europe.

Beyond the limit of the tree-ring calibration, it is necessary to rely on other geological archives to document past atmospheric ^{14}C variations. These include (1) lake and ocean sediments, calibrated by counting varved layers, or with $\delta^{18}\text{O}$ profiles from the same cores; and (2) corals, sediments, or speleothems dated directly with U–Th isotopes. None of these techniques provide time series records with the precision or resolution of the tree-ring record; however, they do allow meaningful inferences to be made about temporal atmospheric ^{14}C variations beyond 12 600 years, to the limit of the radiocarbon method at about 50 000 years.

Natural Radiocarbon Variations

The impetus for developing radiocarbon time series was driven by the need to achieve accurate radiocarbon dates with respect

to calendar ages. These time series also provide fundamental insights into processes that affect the amount and distribution of carbon on Earth. There are three fundamental natural processes that affect atmospheric radiocarbon, including (1) changes in the Earth's geomagnetic field, (2) variable solar activity, and (3) carbon cycle dynamics. Changes in the Earth's geomagnetic field over the past 50 000 years are now well known and the effect of these changes on atmospheric radiocarbon is also understood. This variability occurs over thousands of years. Solar activity produces periodic variations on time scales of years to hundreds of years. Carbon cycle dynamics affect the net amount of carbon in the atmosphere at any given time, and the rate of carbon exchange between different reservoirs on the Earth's surface. The balance established by the carbon cycle reflects the contemporary global climate.

Anthropogenic Radiocarbon Variations

Radiocarbon is a useful tracer because it is carried in carbon dioxide molecules, a key component in Earth's carbon cycle. In addition to the natural influences governing atmospheric ^{14}C , there are three measurable anthropogenic influences. The first is a small (about 2%) decrease in the amount of atmospheric radiocarbon over the past several centuries. This effect was caused by the widespread use of fossil fuels, which are depleted in ^{14}C , sometimes referred to as the industrial effect. The industrial effect began to have a significant impact on atmospheric ^{14}C in the late seventeenth century and continues to influence atmospheric radiocarbon today. A second anthropogenic influence on atmospheric ^{14}C was the production of radiocarbon by nuclear weapons testing. This effect began to increase atmospheric radiocarbon in the mid-1950s. There is a distinct maximum in atmospheric radiocarbon that occurred in 1963, when the Partial Test Ban Treaty, prohibiting atmospheric detonation of nuclear weapons, was signed. A third, and relatively minor source of atmospheric radiocarbon, is produced as a by-product of the nuclear energy industry.

Definitions

The definition of a radiocarbon date includes several conventions that need to be stated explicitly to preface the discussion of temporal radiocarbon variations. These conventions were developed by the radiocarbon community over many years and formalized by Stuiver and Polach (1977). They are as follows:

1. The present year for radiocarbon dating is defined as AD 1950. This date corresponds to the time when the radiocarbon technique was first developed. Radiocarbon dates are expressed in years before present (BP) or years before 1950.
2. The half-life for a radiocarbon measurement is assumed to be 5568 years (the 'Libby' half-life).

- As stated earlier, the original amount of radiocarbon in the atmosphere is assumed to be a constant and known value. This value is quantified using the National Institute of Standards and Technology (NIST – formerly, National Bureau of Standards) Oxalic Acid I standard (NIST-SRM-4990B). Secondary standards used for analysis are adjusted to the original NIST Oxalic Acid I standard value.
- Radiocarbon measurements are normalized for isotopic fractionation by adjusting the measured activities, or carbon isotope ratios, to $\delta^{13}\text{C} = -25\text{‰}$ (see below).

Two of the statements (nos. 2 and 3) listed above are not valid. The actual half-life of radiocarbon is closer to 5730 years, and the amount of radiocarbon in the atmosphere is not invariant. However, the operational definition outlined earlier ensures consistent reporting between different laboratories and makes comparisons of radiocarbon dates unambiguous. The inaccuracies implicit in this definition are removed when the radiocarbon ages are converted to calendar years using a calibration curve.

Correction for Isotopic Fractionation – $\delta^{13}\text{C}$ Correction

The adjustment of a radiocarbon measurement to remove isotopic fractionation is a critical point when discussing temporal ^{14}C variations because some temporal ^{14}C variations have amplitudes that are of the same magnitude as the corrections. Carbon has two stable isotopes – ^{12}C and ^{13}C . The ratio of these two isotopes is essentially the same everywhere in the atmosphere; however, this ratio may be changed when carbon is removed from the atmosphere, for example, through photosynthesis. When plants absorb carbon dioxide from the atmosphere, they discriminate against both ^{13}C and ^{14}C , relative to ^{12}C . This gives plants slightly reduced $^{13}\text{C}/^{12}\text{C}$ ratios, as compared with the contemporary atmosphere in which they grew. For the purposes of radiocarbon measurement, it is necessary to adjust for these small differences by normalizing the measured radiocarbon content to correspond to a uniform $^{13}\text{C}/^{12}\text{C}$ value. To facilitate comparison of $^{13}\text{C}/^{12}\text{C}$ ratios, they are expressed in the δ notation as follows:

$$\delta^{13}\text{C} = \left[\frac{\left(\frac{^{13}\text{C}}{^{12}\text{C}} \right)_s}{\left(\frac{^{13}\text{C}}{^{12}\text{C}} \right)_{\text{STD}}} - 1 \right] \times 1000\text{‰}$$

where $(^{13}\text{C}/^{12}\text{C})_s$ is the ratio of a sample and $(^{13}\text{C}/^{12}\text{C})_{\text{STD}}$ is the ratio in the standard. $\delta^{13}\text{C}$ is a unitless proportion, like percent, and is quoted in parts per thousand, or permil (‰). The standard for $\delta^{13}\text{C}$ is a marine carbonate, so samples from the ocean have $\delta^{13}\text{C}$ values close to zero. Radiocarbon dates are adjusted to $\delta^{13}\text{C} = -25\text{‰}$, which is a typical value for trees.

Calculation of $\Delta^{14}\text{C}$

The quantity $\Delta^{14}\text{C}$ is useful when describing temporal radiocarbon variations (relative radiocarbon isotope ratio, units of permil (‰)). Radiocarbon dates expressed in years before present (BP), present $\equiv$ AD 1950. Half-life of $^{14}\text{C} \equiv 5568$ years (the 'Libby' half-life)). $\Delta^{14}\text{C}$ values express the difference between the sample and the modern atmosphere (AD 1950). There are

two ways to calculate $\Delta^{14}\text{C}$, depending on how one measures the amount of radiocarbon, comprising (1) decay counting or (2) mass spectrometry (measuring isotopic ratios). Counting laboratories measure decay rates (activities) based on the number of beta particles counted per unit time, from the decay of ^{14}C . In a counting laboratory, the activity of ^{14}C is measured in disintegrations per minute (dpm). This is the sample activity A_s . To normalize the sample to $\delta^{13}\text{C} = -25\text{‰}$, A_s may be recalculated to the normalized sample activity A_{SN} using the equation (Donahue et al., 1990):

$$A_{\text{SN}} = A_s \left[\frac{1 - \frac{25}{1000}}{\delta^{13}\text{C} + \frac{1}{1000}} \right]^2$$

In this expression, $\delta^{13}\text{C}$ is measured from the sample. The exponent in the expression is related to the mass difference between ^{14}C and ^{12}C (two atomic mass units). $\Delta^{14}\text{C}$ is then calculated as (Stuiver and Polach, 1977):

$$\Delta^{14}\text{C} = \left[\frac{A_{\text{SN}} e^{\lambda(1950-t)}}{A_{\text{ON}}} - 1 \right] \times 1000\text{‰}$$

where t is the age of the sample in calendar years, λ is the decay constant for ^{14}C appropriate for the 5730 year half-life, and A_{ON} is the activity of the oxalic acid standard at 1950, normalized to -19‰ . Notice that when $\Delta^{14}\text{C}$ is positive, then the sample has a greater ^{14}C activity than the 1950 atmospheric level, and when $\Delta^{14}\text{C}$ is negative, the sample has a smaller ^{14}C activity than the 1950 atmospheric level. In this expression, the age of the sample, t , must be known from independent information.

Accelerator mass spectrometry (AMS) laboratories measure sample isotope ratios instead of sample activities. Either $^{14}\text{C}/^{13}\text{C}$ ratios or $^{14}\text{C}/^{12}\text{C}$ ratios may be utilized for radiocarbon dating purposes. In the case of $^{14}\text{C}/^{13}\text{C}$, the normalized sample ratio, $(^{14}\text{C}/^{13}\text{C})_{\text{SN}}$, is calculated using the formula (Donahue et al., 1990):

$$\left(\frac{^{14}\text{C}}{^{13}\text{C}} \right)_{\text{SN}} = \left(\frac{^{14}\text{C}}{^{13}\text{C}} \right)_s \left[\frac{1 - \frac{25}{1000}}{\delta^{13}\text{C} + \frac{1}{1000}} \right]$$

where $(^{14}\text{C}/^{13}\text{C})_s$ is the measured ratio for the sample and the $\delta^{13}\text{C}$ value is measured from the sample. Notice that when measuring $^{14}\text{C}/^{13}\text{C}$ ratios, the exponent in the conversion factor is reduced to 1 because the difference in mass between the two isotopes is 1 atomic mass unit.

The fraction modern carbon (F) value is calculated as

$$F = \frac{(^{14}\text{C}/^{13}\text{C})_{\text{SN}}}{(^{14}\text{C}/^{13}\text{C})_{\text{ON}}}$$

where $(^{14}\text{C}/^{13}\text{C})_{\text{ON}}$ is the measured ratio for the oxalic acid standard, normalized to its value in 1950, with $\delta^{13}\text{C} = -19\text{‰}$.

For AMS measurements, $\Delta^{14}\text{C}$ may be calculated as follows:

$$\Delta^{14}\text{C} = F e^{\lambda(1950-t)}$$

Counting and mass spectrometry yield equivalent results and the time series discussed in the following sections come from both sources.

Natural Radiocarbon Variations

Radiocarbon is one of many cosmogenic isotopes produced continuously in the upper atmosphere (Lal and Peters, 1967). It is formed predominantly by the reaction $^{14}\text{N}(\text{n},\text{p})^{14}\text{C}$ (Libby, 1955). The global average production of ^{14}C is currently about $2 \text{ atoms cm}^{-2}\text{s}^{-1}$ and the vast bulk of this production is in the atmosphere. Radiocarbon is also produced *in situ* in the lithosphere and hydrosphere, but in negligible amounts. The production of ^{14}C depends on galactic cosmic rays (GCRs), composed primarily of high-energy protons. Although the GCR flux has been constant on a timescale of millions of years (Lal, 1974), it is modulated by the Earth's magnetic field (geomagnetic field) and the Sun's magnetic field (heliomagnetic field), producing temporal variations (Castignoli and Lal, 1980). This temporal variability affects the number of secondary thermal neutrons produced by nuclear reactions between cosmic rays and atmospheric molecules. Once radiocarbon is produced in the atmosphere, it is subject to further temporal fluctuations associated with the transport and exchange of carbon between different carbon reservoirs, through the carbon cycle. These factors are discussed in the next section.

^{14}C Variability Related to Earth's Geomagnetic Field

The Earth's geomagnetic field directly influences the production of radiocarbon in the atmosphere. It serves to deflect incoming galactic cosmic radiation. The geomagnetic field is not homogeneous and the production of ^{14}C is highest at the poles where magnetic field lines are steeply inclined (Figure 1). The geomagnetic influence on radiocarbon production operates on a scale of thousands of years. Figure 2(a) shows the marked decrease in $\Delta^{14}\text{C}$, which has occurred over the past 50 000 years. Figure 2(b) shows a relative paleointensity curve for the Earth's geomagnetic field inferred from globally distributed marine sediment records (Guyodo and Valet, 1999). It can be seen by comparison of the two figures that the radiocarbon content of the atmosphere is inversely related to the geomagnetic field strength. This finding was first shown

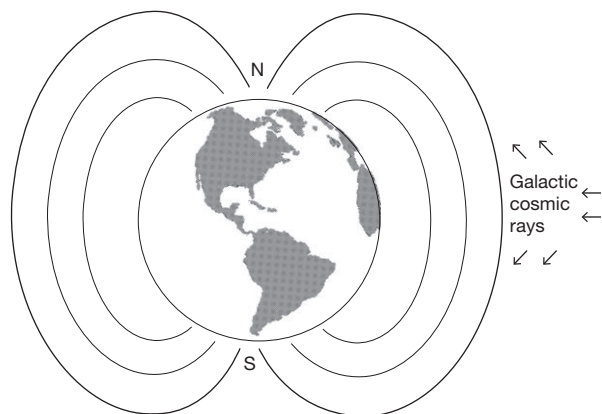


Figure 1 Diagrammatic sketch of the geomagnetic field surrounding the Earth. The geomagnetic field deflects cosmic radiation most strongly at the Equator, where the geomagnetic field is subparallel to the planet's surface.

by Suess (1970) and Bucha (1970). Bard et al. (1990) were the first to document atmospheric $\Delta^{14}\text{C}$ variations beyond the limit of the tree-ring record, where radiocarbon dates diverge from calendar ages by thousands of years. Although a first-order correlation between the geomagnetic field strength and atmospheric ^{14}C is evident, this does not account for all the long-term variability. The details are complicated by multiple ^{14}C pathways in nature within the carbon cycle.

It is worth noting that time series of ^{10}Be , another cosmogenic isotope produced in the atmosphere, are more closely tied to the Earth's geomagnetic field intensity than radiocarbon (McHargue et al., 1995). This is because ^{10}Be has an atmospheric residence time of less than one year. A well-known feature of ^{10}Be studies is a striking maximum in ^{10}Be production between about 35 and 45 ka BP that was observed first in polar ice cores (Raisbeck et al., 1987) and later in marine and lake sediments. Radiocarbon scientists have made an effort to identify a radiocarbon excess from this time period; however, this is more difficult for ^{14}C than for ^{10}Be . As the 35–45 ka BP period is about six radiocarbon half-lives BP, the original ^{14}C signal would have decayed to about 1% or 2% of its original magnitude. This is not true for ^{10}Be , which has a half-life of 1.3 million years.

^{14}C Variability Related to Solar Activity

Atmospheric radiocarbon production is influenced by the Sun's heliomagnetic field in the same sense that it is affected by the Earth's geomagnetic field. On a timescale of years, the heliomagnetic field modulates the GCR flux in a periodic manner. A well-known example of this behavior is observed in the 22-year solar cycle from sunspot time series. A sunspot is a relatively cool region of the Sun's surface characterized by intense magnetic activity. Sunspots were first observed by Galileo and sunspot records extend back in time for centuries. Every 11 years, the heliosphere reverses polarity producing an 11-year periodicity and a 22-year cycle.

As solar variations exhibit periodicities, it is possible to use spectral analyses of $\Delta^{14}\text{C}$ time series to search for changes in past solar activity, predating those known from historical sunspot records. Stuiver and Braziunas (1993) used the Maximum Entropy Method of spectral analysis to study a tree-ring time series with annual resolution, covering the past 450 years. This effort yielded a 10.4-year periodicity, which the authors attributed to the Sun's 11-year cycle. They used the same method to identify a 206-year solar period from a bidecadal time series. In a similar study, Damon and Peristykh (2000) used a discrete Fourier transform method to identify the same 10.4-year period from an annual tree-ring record. The authors also identified a prominent 88-year solar period and a 207-year solar period from a decadal tree-ring record (Figure 3). Such studies require the highest possible precisions for $\Delta^{14}\text{C}$ ($\pm 2\%$), currently obtained only from tree rings.

The Sun also varies on longer time scales and the amplitude of sunspot number variation changes with time as well. Relatively few sunspots occurred between about 1645 and 1715, a period known as the Maunder minimum. This period is plainly evident in decadal tree-ring radiocarbon time series (Figure 4), and was first recognized by Stuiver (1961). Elevated

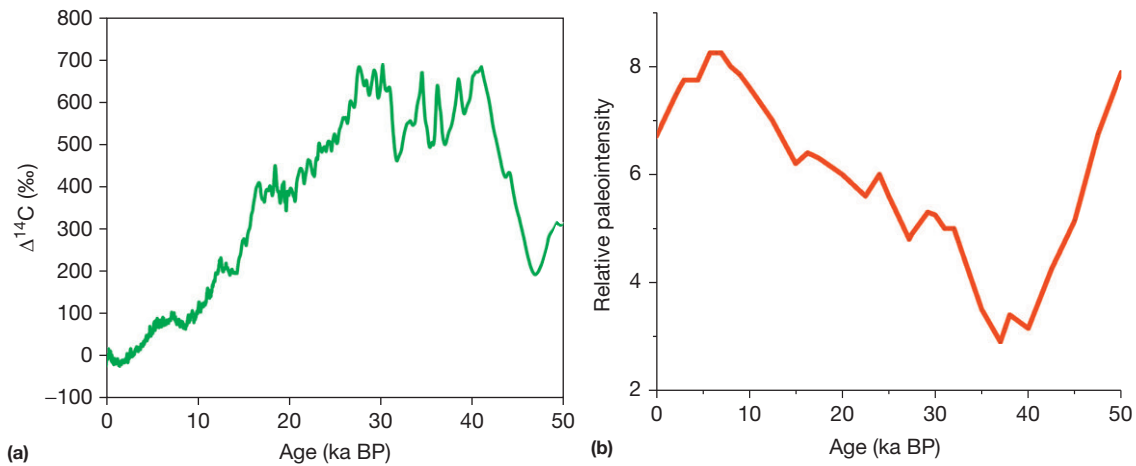


Figure 2 (a) Inferred atmospheric $\Delta^{14}\text{C}$ for the last 50 ka. Data from the INTCAL09 data compilation (Reimer et al., 2009). (b) Relative paleomagnetic field intensity of the Earth for the last 50 ka. Data from the SINT800 record (Guyodo and Valet, 1999). Note the inverse relationship between the relative magnetic intensity and atmospheric $\Delta^{14}\text{C}$.

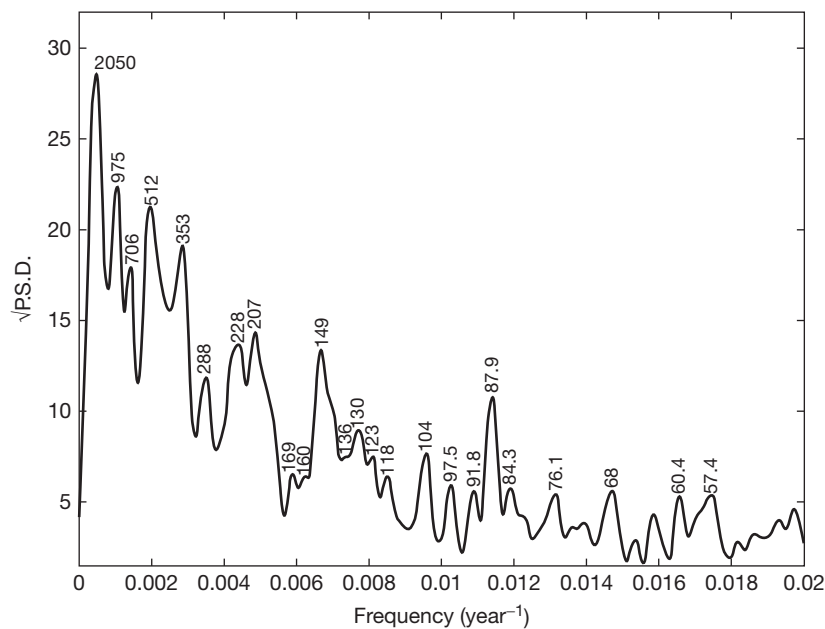


Figure 3 Fourier spectrum of detrended decadal $\Delta^{14}\text{C}$ data. Reproduced from Damon PE and Peristykh AN (2000) Radiocarbon calibration and application to Geophysics, Solar Physics and Astrophysics. *Radiocarbon* 42: 137–150.

$\Delta^{14}\text{C}$ values during this period resulted from reduced heliomagnetic field intensities, which permitted a larger number of GCRs to enter the Earth's atmosphere. There have also been times when the heliomagnetic field was stronger than it is today, and there is evidence of protracted periods of reduced or enhanced heliomagnetic field intensities that lasted thousands of years (Lal et al., 2005). These were detected in ice core samples from Summit, Greenland. Polar ice contains *in situ* ^{14}C whose production is not influenced by the Earth's geomagnetic field or by changes in the carbon cycle. Hence, ice core samples are able to provide a direct record of heliomagnetic field intensity. Lal et al. (2005) found two periods of enhanced radiocarbon production, by a factor of ~ 2 (8.5–9.5 and 27–32 ka BP);

and a period of reduced radiocarbon production, by a factor of ~ 1.5 (12–16 ka BP).

^{14}C production by solar flares

A solar flare is a sudden outburst of particles from the Sun's surface. Radiation from these bursts is emitted across a wide range of the electromagnetic spectrum. Some particles travel to the Earth's atmosphere and produce secondary thermal neutrons in the same manner as GCRs. Lingenfelter and Ramaty (1970) demonstrated that a single solar flare can produce measurable increases in atmospheric $\Delta^{14}\text{C}$. For example, they calculated that a particular solar flare, which occurred on 23 February 1956, is expected to have caused a 7.5‰ increase

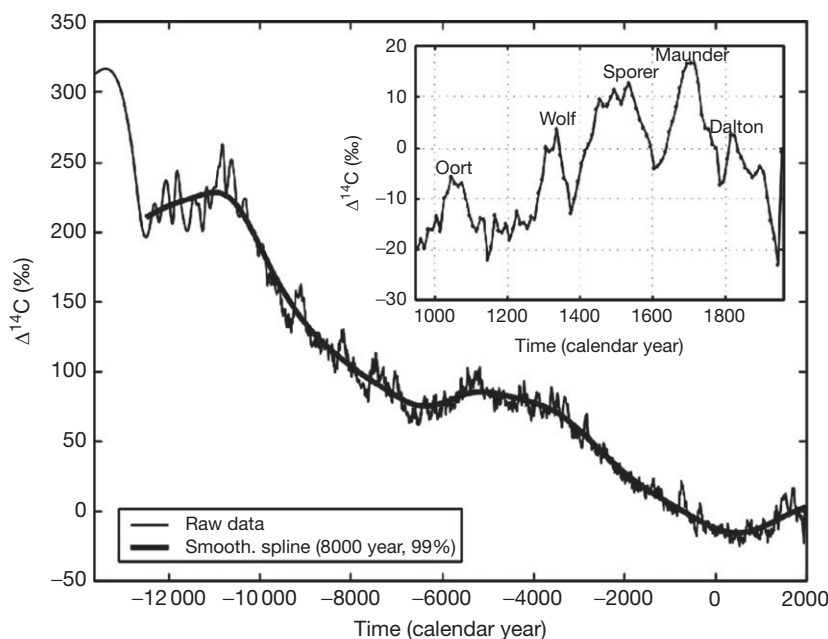


Figure 4 Atmospheric $\Delta^{14}\text{C}$ for the last 16 ka from the INTCAL98 tree-ring compilation. Inset shows solar minima (Oort, Wolf, Spörer, Maunder, and Dalton) which appear as $\Delta^{14}\text{C}$ maxima in the unfiltered data. Reproduced from Damon PE and Peristykh AN (2000) Radiocarbon calibration and application to Geophysics, Solar Physics and Astrophysics. *Radiocarbon* 42: 137–150.

in the contemporary level of atmospheric $\Delta^{14}\text{C}$. This was one of the strongest flares on record (10^{31} – 10^{32} ergs), but not a particularly rare one. Similar flares occurred on 29 September 1989; 28 October 2003; and 2–4 November 2003. Lingenfelter and Ramaty noted that the solar flare contribution to the global atmospheric ^{14}C in 1956 was impossible to resolve due to a much larger concurrent increase in atmospheric ^{14}C from above ground nuclear tests. Scientists have looked to tree-ring time series to identify solar flare components, but the radiocarbon produced by flares is diluted by the relatively large inventory of ^{14}C already in the atmosphere. In addition, the annual resolution of tree rings is too long to identify solar flares that may last hours or days. The effect of solar flares on atmospheric radiocarbon production is counterintuitive. Solar flares enhance ^{14}C production, while sunspot maxima imply reduced ^{14}C production. Both effects tend to occur at the same time. It is clear from Figure 4 that the latter effect has been more significant over the past millennium.

^{14}C production by supernovae

Lingenfelter and Ramaty (1970) also considered the possibility of producing atmospheric ^{14}C from nearby supernovas. They identified two potential pathways of atmospheric radiocarbon production related to supernovas: (1) a short-term increase produced by gamma rays made in the explosion and (2) a much longer-term transient increase produced by direct enhancement of the GCR flux. Damon and others investigated the possibility of elevated atmospheric ^{14}C from a supernova that was observed in AD 1006. Of all supernovas on record, this supernova was the brightest and closest to Earth. Figure 5 shows an annually resolved tree-ring time series from AD 1003 to AD 1030, which features a sharply rising 2-year $\Delta^{14}\text{C}$ pulse with an amplitude of 8‰, possibly related to the 1006 supernova.

^{14}C Variability Related to the Carbon Cycle

Once ^{14}C forms in the upper atmosphere, it rapidly oxidizes to carbon dioxide and becomes efficiently mixed by atmospheric transport and gas exchange. This results in a nearly homogeneous atmospheric ^{14}C content at any given time. The fate of a particular carbon dioxide molecule is subsequently controlled by the carbon cycle, and by extension, the amount of ^{14}C in the atmosphere at any given time is also controlled by carbon cycle dynamics. The carbon cycle connects carbon reservoirs of the atmosphere, hydrosphere, biosphere, and lithosphere in a dynamic balance (Figure 6). More than 90% of the Earth's exchangeable carbon is in the ocean and less than 2% is in the atmosphere. Hence, ocean–atmosphere coupling is strongly leveraged by the mass of carbon in the ocean. Ocean–atmosphere exchange, in particular, exerts dominant temporal control over the amount of CO_2 and ^{14}C in the atmosphere.

The carbon cycle balance can be understood in terms of (1) the number and size of carbon reservoirs on Earth, (2) the exchange mechanisms that transfer carbon between these reservoirs, (3) the rates of exchange between carbon reservoirs, and (4) the dynamics that control the balance of the various exchange pathways. Of particular interest is how these factors might have changed during the past 50 000 years. Studies from the 1950s began to trace the movement of CO_2 between the oceans and the atmosphere, and within the ocean depths using mass balance, or box models to describe the carbon cycle (Craig, 1957; Revelle and Suess, 1957). A key finding in subsequent work was that the ocean is naturally divided into two reservoirs: (1) a thin surface layer (above the thermocline) that exchanges very quickly with the atmosphere (on order 10–20 years), and (2) a much larger deep ocean reservoir that exchanges very slowly (on order thousands of years). This structure was incorporated into the model of

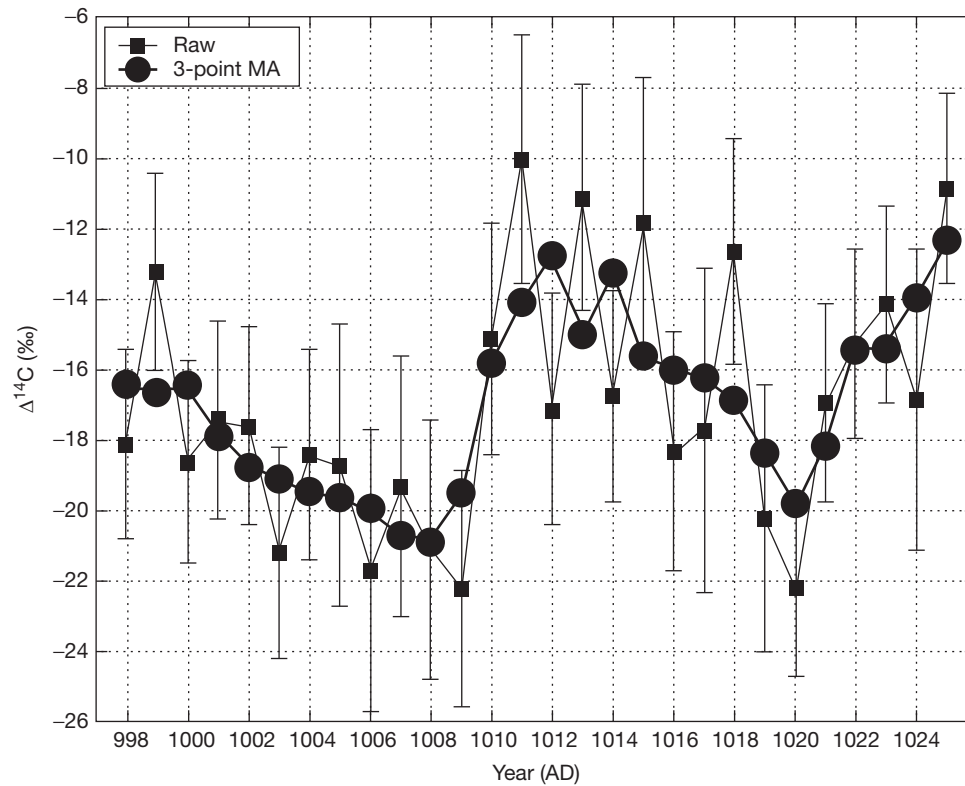


Figure 5 A small but distinct atmospheric $\Delta^{14}\text{C}$ peak that could be related to a supernova observed on 30 April 1006. Reproduced from Damon PE and Peristykh AN (2000) Radiocarbon calibration and application to Geophysics, Solar Physics and Astrophysics. *Radiocarbon* 42: 137–150.

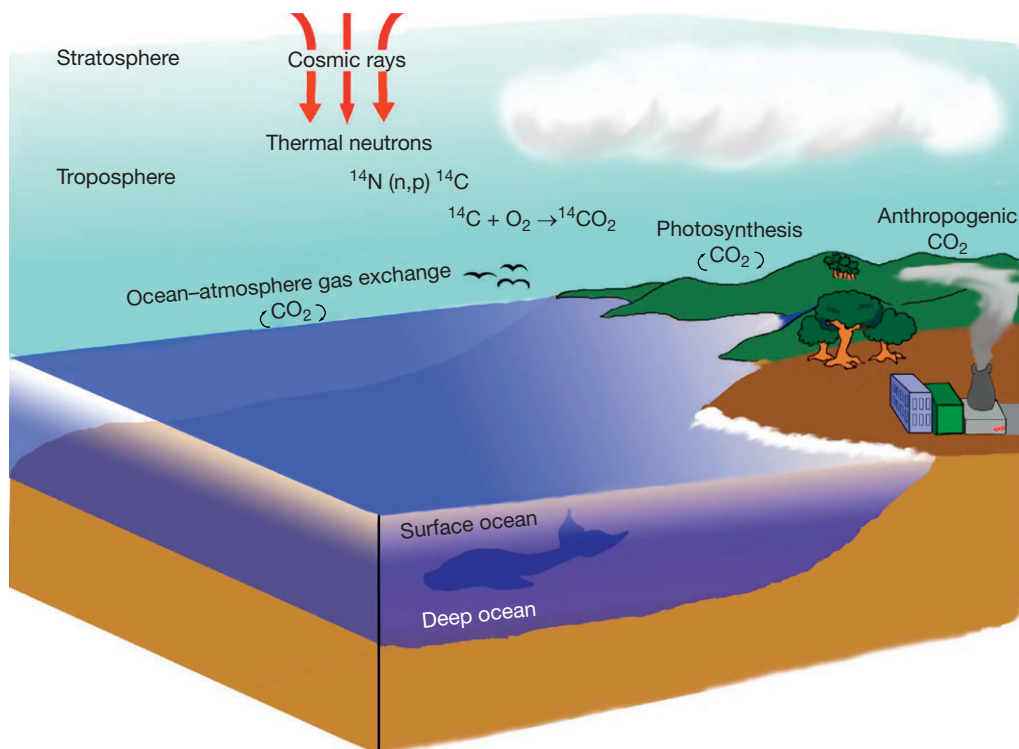


Figure 6 Diagrammatic representation of the carbon cycle. Radiocarbon is produced from secondary thermal neutrons in the atmosphere. The deep ocean is the largest carbon reservoir on Earth. The surface ocean rapidly exchanges with the atmosphere, over a decade or two.

Oeschger et al. (1975) who distinguished a mixed-surface ocean carbon reservoir from a deep ocean carbon reservoir. Their model reliably predicted the major features of atmospheric ^{14}C variability in the latter half of the twentieth century, following massive inputs of bomb-produced ^{14}C .

Mass balance models are more loosely constrained during times of climate change, such as during the last Ice Age. This is because the amount and rate of carbon exchange between reservoirs varied substantially in the past, and the details of these changes are unknown. The Last Glacial Maximum occurred approximately 20 000 years ago, and since then, sea

level has risen over 120 m. This implies large carbon reservoir changes, unseen during the Holocene, where the atmospheric radiocarbon record is well established. The most recent large change in atmospheric radiocarbon occurred 11 000–12 000 calendar years ago. This 10% drop in $\Delta^{14}\text{C}$ was first observed in corals from Papua New Guinea (Edwards et al., 1993). This feature is very likely related to climate change associated with the Younger Dryas climatic event.

Anthropogenic Radiocarbon Variations

Man's activities have left a distinct imprint on the atmospheric radiocarbon record. Three primary anthropogenic inputs can be identified in the atmosphere: (1) ^{14}C -depleted carbon from fossil fuel burning, (2) radiocarbon from nuclear weapons testing, and (3) radiocarbon from nuclear fuels reprocessing.

Fossil Fuels

Anthropogenic effects were first recognized by Suess (1954), who observed a 2% reduction in atmospheric ^{14}C starting in the early eighteenth Century (Figure 7). This reduction was caused by the onset of the industrial revolution. The atmospheric inventory of ^{14}C became diluted when humans began to burn fossil fuels in large amounts. Fossil fuels are typically millions of years old and any radiocarbon in the fuels has long since decayed away. Fossil fuel burning produced a gradual drop in $\Delta^{14}\text{C}$ over time. A side effect of the drop in radiocarbon is a ^{14}C age plateau, which makes it impossible to distinguish radiocarbon ages between the eighteenth and twentieth centuries.

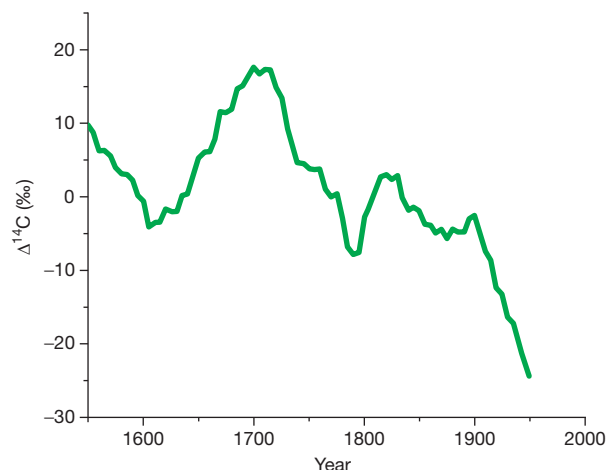


Figure 7 Industrial effect on atmospheric $\Delta^{14}\text{C}$ over the past several centuries. A broad plateau exists because of fossil fuel burning. First identified by Suess (1954). Data is from the INTCAL09 dataset (Reimer et al., 2009).

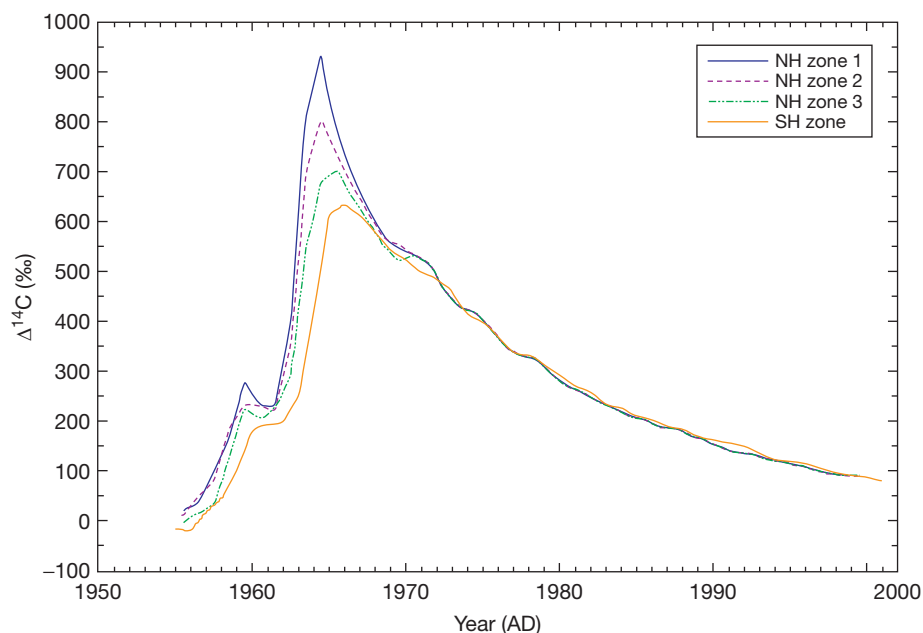


Figure 8 Atmospheric radiocarbon from nuclear weapons testing in the 1950s and 1960s. Note the different amplitude and rate of increase in the Southern Hemisphere, as compared with the Northern Hemisphere. Zones represented in the figure are NH1 – Northern Hemisphere high latitude sites, NH2 and NH3 – Northern Hemisphere mid- to low latitude sites, SH – Southern Hemisphere. Reproduced from Hua Q and Barbetti M (2004) Review of tropospheric bomb ^{14}C data for carbon cycle modeling and age calibration purposes. *Radiocarbon* 46: 1273–1298.

Nuclear Weapons Testing and Nuclear Power

When Libby first published on the radiocarbon method, anthropogenic inputs of radiocarbon were minimal. Within 10 years however, worldwide nuclear weapons testing would nearly double the amount of radiocarbon in the Earth's atmosphere. Anthropogenic radiocarbon was injected into the stratosphere during these tests and it entered the troposphere within a couple of years (Nydal, 1963). Atmospheric ^{14}C increased rapidly in the Northern Hemisphere, with the highest $\Delta^{14}\text{C}$ levels occurring in the mid- to high latitudes (Figure 8). Very few tests were made in the Southern Hemisphere and atmospheric $\Delta^{14}\text{C}$ changed more slowly there, with reduced peak radiocarbon levels. The maximum amplitude of the ^{14}C perturbation in the Northern Hemisphere was reached in 1963, around the time when the Partial Test Ban Treaty was signed. After 1963, the radiocarbon content of the atmosphere dropped off sharply. Within 10 years, atmospheric ^{14}C was reduced to about one half of its peak levels. This reduction was primarily due to the efficient exchange of CO_2 between the atmosphere and the surface ocean. While the atmosphere continued to drop, surface ocean values rose, with peak values occurring around 1970 (Linick, 1980; Nydal et al., 1980). Figure 9

illustrates the dramatic increases which occurred in the surface ocean during the 1960s. Surface ocean ^{14}C is seen to steadily increase from the late 1950s to 1972, when it reached peak $\Delta^{14}\text{C}$ values of approximately 200‰. Figure 9 also illustrates the utility of ^{14}C as a chemical tracer, showing evidence of equatorial upwelling associated with low $\Delta^{14}\text{C}$ values, and enhanced CO_2 exchange in oceanic gyres at mid-latitudes showing relatively high $\Delta^{14}\text{C}$ values. This aspect of the anthropogenic input of ^{14}C made it ideal for tracer studies of the carbon cycle (Damon et al., 1978).

To the benefit of all, radiocarbon is no longer being produced in large amounts by nuclear weapons testing. However, it is still being released into the atmosphere as a by-product of the nuclear power industry. At the LaHague reprocessing plant in France, ^{14}C is introduced directly to the atmosphere through a 100 m chimney (Fontugne et al., 2004), and at the Sellafield nuclear reprocessing plant in England, ^{14}C is released directly into the marine environment (Cook et al., 2004). While these types of releases still occur, they are miniscule in comparison with the quantities of ^{14}C released during the era of nuclear weapons testing. Nuclear fuel reprocessing currently produces only a small fraction of the natural total production of atmospheric ^{14}C .

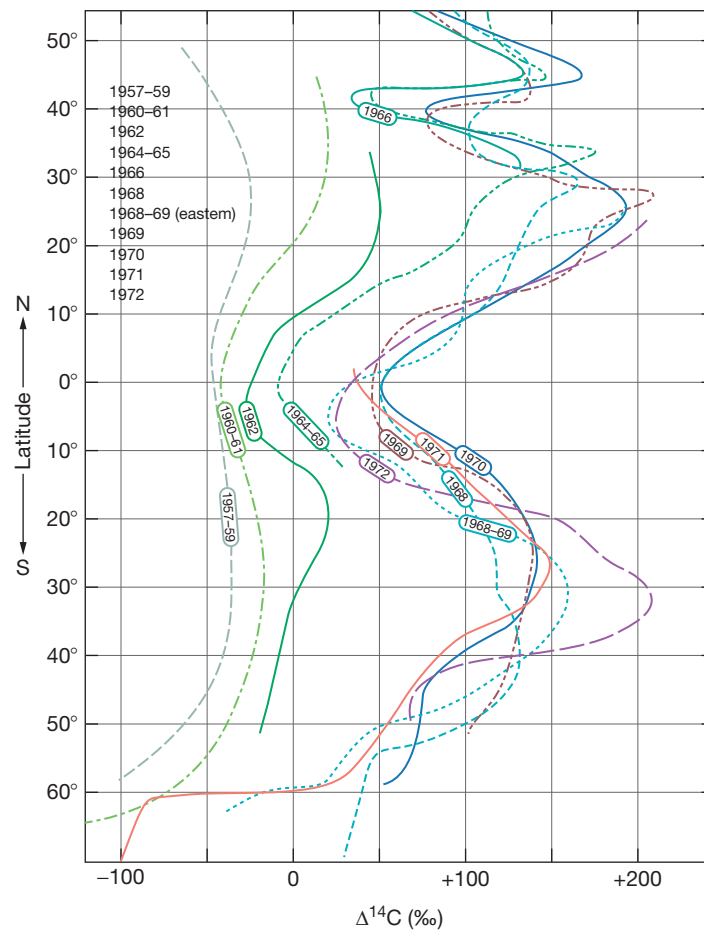


Figure 9 Latitudinal variation of $\Delta^{14}\text{C}$ inorganic carbon dissolved in surface waters of the Central Pacific Ocean, 1957–72. Modified from Linick TW (1980) Bomb-produced carbon-14 in the surface water of the Pacific Ocean. *Radiocarbon* 22: 599–606.

See also: Dating Techniques; Dendrochronology; Geomagnetic Excursions and Secular Variations. **Introduction:** History of Dating Methods. **Paleoceanography:** Paleoceanography An Overview. **Radiocarbon Dating:** AMS Radiocarbon Dating; Calibration of the ^{14}C Record; Variations in Atmospheric ^{14}C .

References

- Bard E, Hamelin B, Fairbanks RG, and Zindler A (1990) Calibration of ^{14}C timescale over the past 30,000 years using mass spectrometric U–Th ages from Barbados corals. *Nature* 345: 405–410.
- Beck JW, Richards DA, Edwards RL, et al. (2001) Extremely large variations of atmospheric ^{14}C concentration during the last glacial period. *Science* 292: 2453–2458.
- Bucha V (1970) Influence of the earth's magnetic field on radiocarbon dating. In: Olsson IU (ed.) *Proceedings XII Nobel Symposium*, pp. 501–511. New York: Wiley.
- Castignoli G and Lal D (1980) Solar modulation effects in terrestrial production of carbon-14. *Radiocarbon* 22: 133–158.
- Cook GT, MacKenzie AB, Muir GPK, Mackie G, and Gulliver P (2004) Sources of anthropogenic ^{14}C to the North Sea. *Radiocarbon* 46: 869–876.
- Craig H (1957) The natural distribution of carbon and the exchange time of carbon between atmosphere and sea. *Tellus* 9: 1–17.
- Damon PE, Lerman JC, and Long A (1978) Temporal fluctuations of atmospheric ^{14}C : Causal factors and implications. *Annual Reviews of Earth and Planetary Science* 6: 457–494.
- Damon PE and Peristykh AN (2000) Radiocarbon calibration and application to Geophysics, Solar Physics and Astrophysics. *Radiocarbon* 42: 137–150.
- de Vries H (1958) Variation in concentration of radiocarbon with time and location on earth. *Koninklijke Nederlandse Akademie van Wetenschappen Proceedings, Series B* 61: 94–102.
- Donahue DJ, Linick TW, and Jull AJT (1990) Isotope-ratio and background corrections for accelerator mass spectrometry radiocarbon measurements. *Radiocarbon* 32: 135–142.
- Edwards RL, Beck JW, Burr GS, et al. (1993) A large drop in atmospheric $^{14}\text{C}/^{12}\text{C}$ and reduced melting in the Younger Dryas, documented with ^{230}Th ages of corals. *Science* 260: 962–968.
- Fontugne M, Maro D, Baron Y, Hatté C, Hebert D, and Douville D (2004) ^{14}C sources and distribution in the vicinity of La Hague Nuclear Reprocessing Plant: Part I – Terrestrial environment. *Radiocarbon* 46: 827–830.
- Guyodo Y and Valet J-P (1999) Global changes in intensity of the Earth's magnetic field during the past 800 kyr. *Nature* 399: 249–252.
- Hua Q and Barbetti M (2004) Review of tropospheric bomb ^{14}C data for carbon cycle modeling and age calibration purposes. *Radiocarbon* 46: 1273–1298.
- Lal D (1974) Long term variations in the cosmic ray flux. *Philosophical Transactions of the Royal Society of London Series A* 277: 395–411.
- Lal D, Jull AJT, and Pollard D (2005) Evidence for large century time-scale changes in solar activity in the past 32 kyr BP, based on in-situ cosmogenic ^{14}C in Greenland ice at Summit. *Earth and Planetary Science Letters* 234: 335–349.
- Lal D and Peters B (1967) Cosmic ray produced radioactivity on the earth. *Handbuch Physik* 46: 551–612.
- Libby WF (1955) *Radiocarbon Dating*, 2nd edn., p. 175. Chicago, IL: University of Chicago Press.
- Lingenfelter RE and Ramaty R (1970) Astrophysical and geophysical variations in C-14 production. In: Olsson IU (ed.) *Proceedings XII Nobel Symposium*, pp. 513–535. New York: Wiley.
- Linick TW (1980) Bomb-produced carbon-14 in the surface water of the Pacific Ocean. *Radiocarbon* 22: 599–606.
- McHargue LR, Damon PE, and Donahue DJ (1995) Enhanced cosmic-ray production of ^{10}Be coincident with the Mono Lake and Laschamp geomagnetic excursions. *Geophysical Research Letters* 22: 659–662.
- Nydal R (1963) Increase in radiocarbon from the most recent series of thermonuclear tests. *Nature* 6: 212–214.
- Nydal R, Lövseth K, and Skogseth FH (1980) Transfer of bomb ^{14}C to the ocean surface. *Radiocarbon* 22: 626–635.
- Oeschger H, Siegenthaler U, Schotterer U, and Gugelmann A (1975) A box diffusion model to study the carbon dioxide exchange in nature. *Tellus* 27: 168–192.
- Reimer PJ, Baillie MGL, Bard E, et al. (2009) INTCAL09 and MARINE09 radiocarbon age calibration curves, 0–50,000 cal BP. *Radiocarbon* 51(4): 1111–1150.
- Revelle R and Suess HE (1957) Carbon dioxide exchange between atmosphere and ocean. *Tellus* 9: 18.
- Stuiver M (1961) Variations in radiocarbon concentration and sunspot activity. *Journal of Geophysical Research* 66: 273–276.
- Stuiver M and Braziunas TF (1993) Sun, ocean, climate and atmospheric $^{14}\text{CO}_2$: An evaluation of causal and spectral relationships. *The Holocene* 3: 289–305.
- Stuiver M and Polach H (1977) Reporting of ^{14}C data. *Radiocarbon* 19: 355–363.
- Suess HE (1954) Natural radiocarbon and the rate of exchange of carbon dioxide between the atmosphere and the sea. *Proceedings, Williams Bay Conference*, p. 52. NAS-NSF Publ September 1953.
- Suess HE (1970) The three causes of the secular C-14 fluctuations, their amplitudes and time constants. In: Olsson IU (ed.) *Proceedings XII Nobel Symposium*, pp. 596–606. New York: Wiley.
- van der Plicht JJW, Beck JW, Bard E, et al. (2004) NOTCAL04 – Comparison/calibration ^{14}C records 26–50 cal kyr BP. *Radiocarbon* 46: 1225–1238.

Calibration of the ^{14}C Record

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Introduction

^{14}C ages, or radiocarbon ages as they are also known, are calculated from the ratio of the radioactive carbon isotope ^{14}C to total carbon or one of the stable isotopes (^{12}C or ^{13}C) in the sample compared to a standard. This calculation assumes that the level of ^{14}C in the atmosphere has been constant since the time the sample grew or was formed. If that were the case, then no calibration would be needed. However, <10 years after the establishment of the radiocarbon dating method, discrepancies between radiocarbon and tree-ring ages were noted and potential mechanisms were discussed (de Vries, 1958). Such discrepancies are especially important when comparing records with a timescale based on radiocarbon ages with those based on other methods, such as layer counting or historical events, or for calculating rates of change.

The concentration of ^{14}C in the atmosphere at any given time depends on a number of factors including the strength of the geomagnetic field and the solar wind intensity, which influence ^{14}C production, and the uptake or release by the earth's carbon reservoirs, particularly the ocean (see [Causes of Temporal \$^{14}\text{C}\$ Variations](#) and [Variations in Atmospheric \$^{14}\text{C}\$](#)). Ocean surface water, which has partially equilibrated with the atmosphere, diffuses or sinks from the surface into the deep ocean where it is isolated from the atmosphere for hundreds or thousands of years (see [Thermohaline Circulation](#) and [Paleoceanography An Overview](#)). The carbon in the water has time to undergo radioactive decay before it is brought to the surface again through upwelling. This lower oceanic radiocarbon content means that samples formed in the ocean surface waters are on an average ~400 years older than samples formed in the atmosphere at the same time. This offset between the atmosphere and ocean surface age is referred to as the marine reservoir effect or marine reservoir age. In addition to natural causes of atmosphere ^{14}C variations, anthropogenic releases of fossil fuel and ^{14}C produced by nuclear weapons testing have altered the ^{14}C concentration in the atmosphere, biosphere, and oceans.

Thus began the quest for known-age samples and precise ^{14}C measurements to correct for the variations in atmospheric, and eventually oceanic, ^{14}C levels through what became known as radiocarbon calibration. The first high-resolution atmospheric calibration based on ring-counted *Sequoia gigantea* was published as both a calibration curve and a table of calendar and radiocarbon ages extending from 1000 to 1850 cal AD (Stuiver and Suess, 1966). Although the resolution and the precision were not as high as the most recent international calibration curve IntCal09, this was undoubtedly a remarkable achievement (Figure 1). Between 1966 and 1979, numerous radiocarbon calibration curves were built using various data sets, criteria for inclusion of data, and methods of combination or smoothing. The lack of standardization

created confusion when comparisons were made between calibrated ages from different curves. In 1979, an international workshop was held to establish a consensus calibration (Klein et al., 1982). Since then, an international working group has produced updates to the calibration curve on ~5-year intervals with the most recent internationally ratified curve, IntCal09, extending to 50 000 calibrated years before AD 1950 (cal BP) (Reimer et al., 2009).

Calibration and Comparison Records

Tree-Rings

Tree-rings are one of the best recorders of atmospheric ^{14}C concentration because many temperate tree species reliably form one ring each year. Trees incorporate carbon into their cell structure through the process of photosynthesis, which uses carbon dioxide from the atmosphere. While some compounds, such as lignin, are added to the cell structure of the living sapwood, the cellulose itself is formed during a single growth season. The wood from one or more tree-rings taken together is usually pretreated with acid and alkali to remove the mobile sugars, resins, tannins, and most of the lignin in a process known as the de Vries method, or, in some cases, either holocellulose or α -cellulose is chemically extracted from the wood (Stuiver et al., 1984). The pretreated samples are then radiocarbon dated. Because there is some small uncertainty in the actual calendar age of the wood used for the ^{14}C measurement, due to different ring widths for each year in a block of wood or to material lost to saw cuts, the term calibrated or 'cal' age is used rather than calendar age.

Tree-rings can be counted from a known felling date to give a calendar age for each ring, which can then be measured for ^{14}C content. However, to extend the calibration beyond the oldest living trees, a sequence of ring width patterns from wood taken from historical buildings and subfossil trees extracted from bogs or river gravels can be matched to construct a dendrochronological record (Figure 2). A number of tree-ring chronologies have been constructed including the Belfast Irish Oak chronology and the Hohenheim oak and pine chronologies. These chronologies are based on tens to hundreds of trees for any given year. For example, the Hohenheim Oak chronology has a mean replication of over a hundred trees per year (Friedrich et al., 2004). The robustness of these chronologies has been demonstrated through cross-checking between independently constructed chronologies.

Corals

Corals have been extremely useful for ^{14}C calibration, both for developing calibration curves specific to marine samples and for extending the atmosphere calibration with the use of a

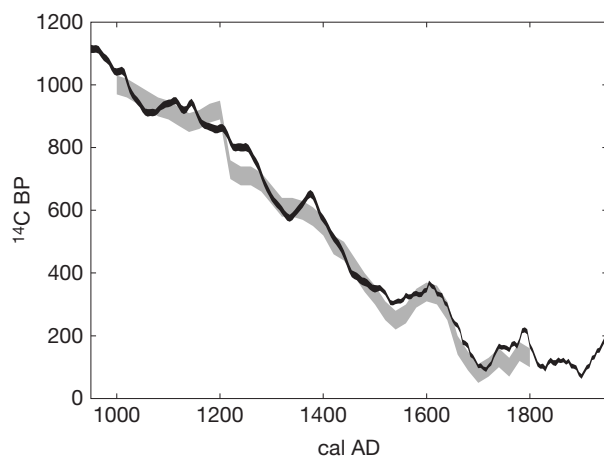


Figure 1 The first atmospheric calibration curve (gray) as measured by Stuiver and Suess (1966) compared to the 2009 international calibration curve IntCal09 (black), both shown as one-standard deviation error envelopes.

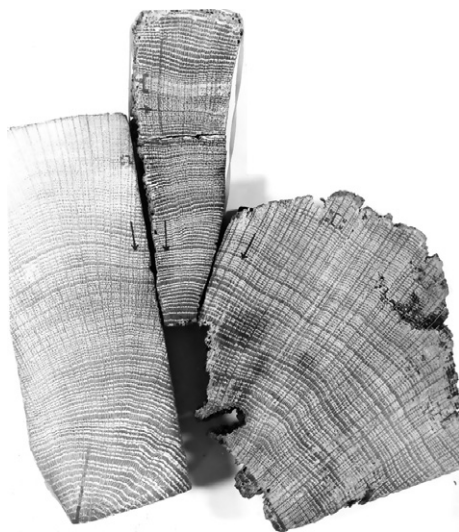


Figure 2 An example of tree-ring width pattern matching for three pieces of Irish oak, which form part of the Belfast Irish Oak chronology. Arrows point to the same year on all three pieces of wood. Photograph courtesy of Prof. M.G.L. Baillie, Queen's University Belfast.

reservoir correction, which will be discussed in more detail in a later section. Reef corals have been shown to build their skeletons of aragonite (calcium carbonate) in isotopic equilibrium with the dissolved organic carbon in the ocean water where they grow. Like tree-rings, reef corals also have annual banding, which can be counted. Unfortunately, individual corals rarely live more than a few hundred years at most, and no long chronologies have been built. However, because uranium from seawater is incorporated in the calcium carbonate, the age of the coral can be measured independently by uranium series dating (^{238}U – ^{230}Th and ^{235}U – ^{231}Pa), and an uncertainty in the calendar age is calculated (see **U-Series Dating**). Potential problems exist with contamination in corals due to dissolution of the coral aragonite and recrystallization as calcite, potentially incorporating older or younger carbon. This is

primarily a problem with corals that have been exposed to freshwater due to changes in sea level or uplift of the coral reef. Checks on the amount of calcite in a sample are done with x-ray diffractions, and samples with quantifiable calcite of $>1\%$ are not recommended for calibration data sets (Reimer et al., 2002).

Varved Sediments

Varved (annually laminated) sediments can potentially supply calibration records through ^{14}C dating of macrofossils, such as Foraminifera or terrestrial material contained in the layers from marine and lake sediments, respectively (see **Varved Lake Sediments**; **Varved Marine Sediments**). The conditions under which the laminations form must be well understood and occur annually with high fidelity. Multiple cores need to be stratigraphically correlated, or dated marker events, such as volcanic tephra (see **Tephrochronology**), need to be present to ensure that there are no hiatuses in the deposition. Sediments are often not continuously varved from the present back through time, so they must be anchored by independent means such as wiggle-matching to the tree-ring curve (see section '**Calibration Methods**'). Uncertainty in the anchoring and in the varve counts needs to be quantified in order to provide reasonable error estimates on the independent age. Ideally, only fragile plant or insect macrofossils, which would not survive extended transport, are used for dating lake sediments for calibration records. At present, only the Cariaco Basin varved sediment record has been incorporated into internationally accepted calibration curves (Reimer et al., 2009).

Aragonite deposits in some varved lake sediments may also be ^{14}C and ^{238}U – ^{230}Th dated. These records could potentially be incorporated into a calibration curve. The ^{238}U – ^{230}Th age must be corrected for initial ^{230}Th , which increases the uncertainty of the age. As with marine samples, the ^{14}C reservoir corrections may vary with time.

Speleothems

Cave deposits known as speleothems (see **Speleothems**) hold the potential for providing near-atmospheric calibration records beyond the end of the tree-ring records as CO_2 (from the atmosphere or soil) is incorporated into the calcite or aragonite when it forms (Hoffmann et al., 2010). The speleothem age can be determined from U–Th dating although usually a growth model is used to provide ages along the axis of the speleothem, where the samples for ^{14}C dating are taken. As with aragonite, the U–Th age must be corrected for initial ^{230}Th . Uncertainties in estimating the fraction of carbon from rock weathering (^{14}C free or 'dead carbon') incorporated into the speleothem require a correction, which is estimated from the overlap with the tree-ring curve or through modeling using stable isotopes. Speleothems were not included in the IntCal09 curve but will be included in the planned update.

Nonvarved Marine Sediments

Nonvarved marine sediments have not been considered as 'calibration' records in the strictest sense because their time-scale is not independently derived. However, if there is a

plausible physical mechanism for a climatic connection, some of these records can be ‘tuned’ or tie-pointed to other timescales (see [Chronologies](#)) through correlation of climate proxies, such as $\delta^{18}\text{O}$, alkenone-derived temperature, varve thickness, or color to the $\delta^{18}\text{O}$ temperature record in ice cores or U-Th-dated speleothems. The uncertainty in the derived timescale must include the uncertainties in both records and in the match between the records (e.g., [Hughen et al., 2006](#)). The ^{14}C ages of planktonic Foraminifera from the Cariaco Basin and Iberian Margin records tie-pointed to the Hulu Cave speleothem timescale ([Bard et al., 2004](#); [Hughen et al., 2006](#)) contributed greatly to the extension of IntCal09 calibration curve to 50 000 cal BP.

Calibration Curves

The combination of ^{14}C data sets to form calibration curves has ranged from simple averages in given windows of time to the implementation of a stochastic model for the underlying radiocarbon calibration curve. The random walk model used in the 2009 IntCal calibration curves incorporates observations with independent errors in the ^{14}C age uncertainty and calendar age, including asymmetry resulting from accumulating counting errors, and allows for the covariance between observations ([Blackwell and Buck, 2008](#)).

Calibration curves have been constructed for the most commonly sampled carbon reservoirs – the atmosphere and the surface-mixed layer of the ocean. The underlying assumption is that local or regional differences from these curves are small compared to the uncertainty of the measurements. When this is not the case, a correction must be applied (see section [‘Problems and Uncertainties in Calibration’](#)). The Northern

Hemisphere atmospheric curve is based on mid-latitude tree-ring measurements as far back as these records are available and then relies on reservoir-corrected marine records. For the Southern Hemisphere, mid-latitude tree-ring ^{14}C measurements have been made, which provide a separate calibration curve for a limited period, which is extended with a random effects model that incorporates the potential variability in the offset. For the marine curve, there have been very few measurements of long calibration records for the Holocene, so the marine curve for this period has been modeled from a simple ocean–atmosphere box-diffusion model using the atmospheric curve as input. The portion of the Marine09 curve older than 10 500 cal BP is based on data from various ocean regions, which has been combined into a ‘global’ ideal curve by applying assumed constant regional reservoir corrections and then subtracting an average 405-year correction ([Reimer et al., 2009](#)).

Calibration Methods

Single Age Calibration

A radiocarbon age is conventionally given as a mean with one standard deviation, for example, 4560 ± 40 BP ([Stuiver and Polach, 1977](#)). Calibration, in the past, was done with a simple table of radiocarbon and calibrated ages for various standard deviations. However, as calibration curves were extended further back in time, it became impractical to publish complete tables. Graphical methods for calibration were also used. The calibration curve was intercepted by the radiocarbon age plus and minus one or two standard deviations to yield a corresponding calibrated age range or ranges ([Figure 3\(a\)](#)). This classical intercept method, while straightforward, does not

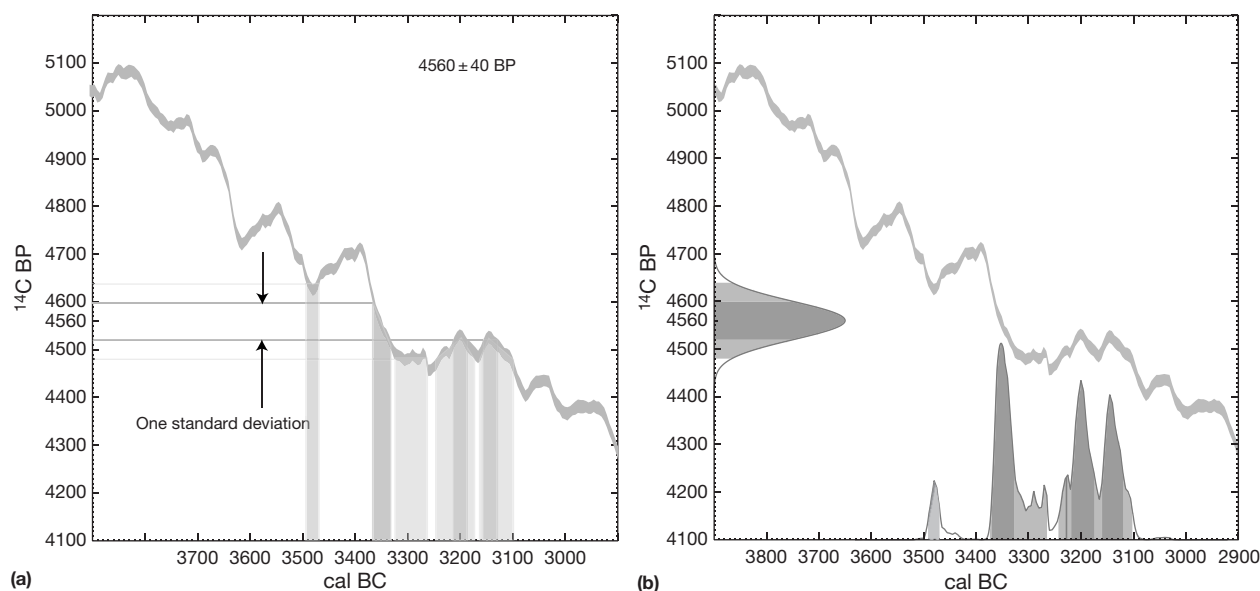


Figure 3 (a) An example calibration for a radiocarbon age of 4560 ± 40 BP using the classical intercept method. The radiocarbon age ± 1 and 2σ errors are drawn as lines intersecting the calibration curve, and the corresponding calibrated age ranges are shown as light gray and dark gray boxes, respectively. The IntCal09 calibration curve is shown as a one-standard deviation error envelope. (b) Calibration for the same radiocarbon age using the probability method. The 68% probability ranges are dark gray and the 95% probability ranges are light gray in both the radiocarbon and the calibration distributions.

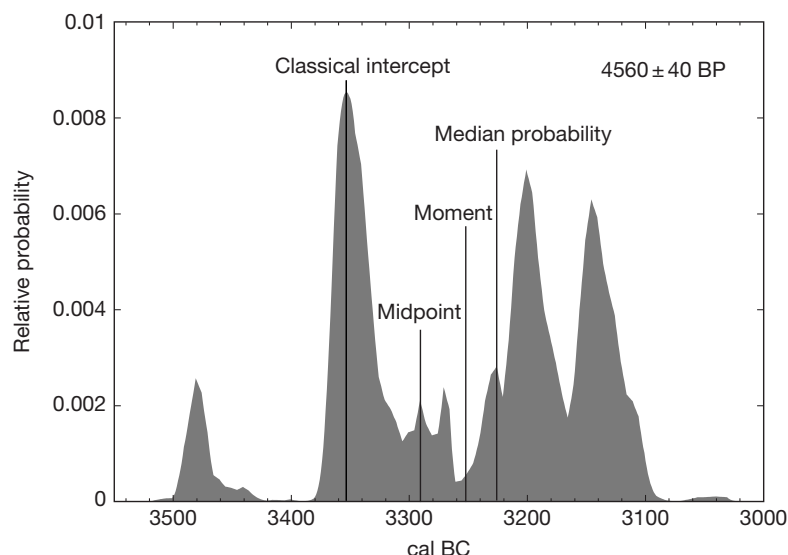


Figure 4 Age estimators for the probability distribution of the radiocarbon age 4560 ± 40 BP calculated with the IntCal09 calibration curve including the classical intercept of the radiocarbon age with the calibration curve, the midpoint of the 95% calibrated age ranges, the median probability, and the moment or weighted average of the probability distribution.

include the full range of the uncertainty on the radiocarbon age and provides no information about the relative probability of multiple calibrated age ranges.

A probability density function (pdf) can be calculated, which retains information about the shape of the distribution. The pdf of the radiocarbon ages R around the mean radiocarbon age U is assumed normal with a standard deviation σ , such that

$$p(R) = \left(\frac{1}{\sigma(2\pi)^{1/2}} \right) \exp - \left[\frac{R - U^2}{2\sigma^2} \right]$$

where

$$\sigma^2 = \sigma_{\text{sample}}^2 + \sigma_{\text{curve}}^2$$

To obtain the probability density along the calendar axis, R can be replaced by the calibration curve function $g(t)$. By determining $g(t)$ for each calendar year t and transferring the corresponding probability of the distribution to the t axis, the probability density is transformed to calendar year dependency (Stuiver and Reimer, 1989). One and two standard deviation confidence intervals may be obtained from the age ranges, which contain 68 and 95% of area under the probability curve (Figure 3(b)).

The shape of the calibration curve at a particular point in time very obviously affects the resulting calibration. Often, the calibration results in broad calibrated age ranges; however, when a region of the calibration curve is very steep, then the resulting calibrated age ranges may be narrower than the radiocarbon age range. The calibrated pdf provides a complete description of the sample calibrated age but is difficult to use in calculations or graphics. Various attempts have been made to define a single number that represents the calibrated age of a sample. While some estimators, such as the moment (also called a weighted average) or the median (the calibrated age at which 50% of the probability falls before and 50% after), are

more robust than others (Telford et al., 2004), there is no good single estimator, and these may fall at relatively low probabilities (Figure 4). The intercept of the radiocarbon age with the calibration curve picks out only one, albeit the highest, of the peaks in probability but could shift significantly with only small changes in the radiocarbon age. Various Bayesian techniques have been developed to deal with the entire pdf in calculations, and these will be discussed later.

Wiggle-Matching

In situations where a time series of radiocarbon ages can be obtained, it is possible to improve the precision of the calibrated age considerably, especially if some of the ages fall on a steep portion or distinctive wiggle in the calibration curve. In its simplest form, a series of radiocarbon ages from tree-rings of known spacing can be moved along the calibration curve until a best match is achieved, classically calculated as the minimum of the weighted sum of squares of the difference. A good example of this is the wiggle-match dating of a series of 'royal' tombs of the Iron Age pastoralists in the High Altai region of Siberia. The dates of the tombs were a long-running controversy among art historians. Initial radiocarbon ages of some of the timbers used in tomb construction gave calibrated age ranges of several hundred years because of a plateau in the calibration curve ~ 800 – 400 BC. A relative chronology for the construction of the tombs was refined with a 'floating' dendrochronology (Marsadolov, 1996), but there was no fixed calendar date. A wiggle-match of ^{14}C ages of a single timber from the Pazyryk 2 tomb with bark attached was used to anchor the dendrochronology and provide a date for the tombs (Figure 5). While easily demonstrated, the uncertainty in the match using the classical statistics method is not rigorously determined, whereas Bayesian methods (see next section) provide a more robust estimate of uncertainty.

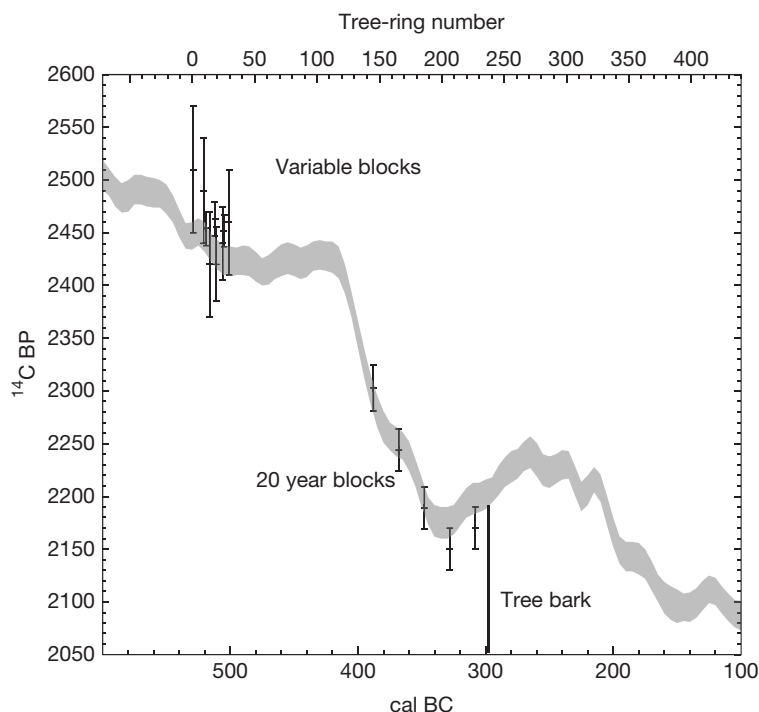


Figure 5 Wiggle-match of the Pazyryk radiocarbon ages versus tree-ring number to the IntCal09 calibration curve. The curve is plotted as a one-standard deviation error envelope, and the radiocarbon ages are given with one-sigma error bars. A bar marks the position of the bark relative to the radiocarbon samples.

The Bayesian wiggle-match gave a felling date for the timber from the Pazyryk 2 tomb of 301–282 BC at two standard deviations (Mallory et al., 2002, see also Kuzmin et al., 2004).

Similar wiggle-matches have been done to anchor floating chronologies for calibration data sets such as the match between the Cariaco Basin varved sediment record and tree-ring ^{14}C . Wiggle-matches, where the spacing in calendar time between ^{14}C -dated samples is not known, such as peat or sediment, can be done by finding the best fit within a physically reasonable range of sedimentation rates. Most of the progress in this area has been made using a Bayesian approach.

Bayesian Modeling

Prior information such as sample stratigraphy, varve counts, or marker events can be used to constrain the probability distributions and thus refine calibrated age ranges. For example, stratigraphic ordering requires that an upper sample be younger than a lower sample in a sequence. This *a priori* information can be used to constrain the original probability distributions so that the modified (or posterior) probability of the upper samples has less probability in the older ranges and more probability in the younger ranges, and vice versa for the lower samples (Figure 6). This is done formally by the application of Bayes's theorem of conditional probability to radiocarbon calibration (Buck et al., 1991). Bayesian calibration programs were initially designed for archaeological applications but can also be used to assess the reliability of stratigraphy or varve counting in paleoenvironmental studies. Recent research has focused on the theoretical framework necessary to apply Bayesian analysis to sedimentation rates, outlier

identification, and calibrated age-depth models (Blaauw and Christen, 2011; Bronk Ramsey, 2008).

Problems and Uncertainties in Calibration

Radiocarbon Error Estimates and Laboratory Offsets

Radiocarbon age errors are often given as the uncertainty resulting from the statistical nature of counting and do not include other possible errors that could result from, for example, sample preparation. An underestimation of the error will of course result in an underestimation of the calibrated age range. Interlaboratory comparison exercises on replicate samples have shown that for a number of laboratories, the errors are underestimated and also that small but significant offsets occur relative to consensus values (see Sources of Error). It is difficult to provide guidelines as to how much, if at all, the error should be increased in the absence of information from the laboratory.

Marine Reservoir Corrections

The actual difference between the atmosphere and surface ocean age (reservoir age) is not a constant but varies with time and location (see Causes of Temporal ^{14}C Variations). Regional upwelling of aged ocean water or input of freshwater necessitates a correction, ΔR , which to a first approximation is considered constant (Stuiver et al., 1986). ΔR can be calculated from the difference between the ^{14}C age of known calendar age marine organisms, such as mollusks, and the marine calibration curve value for that calendar age. Such samples must have

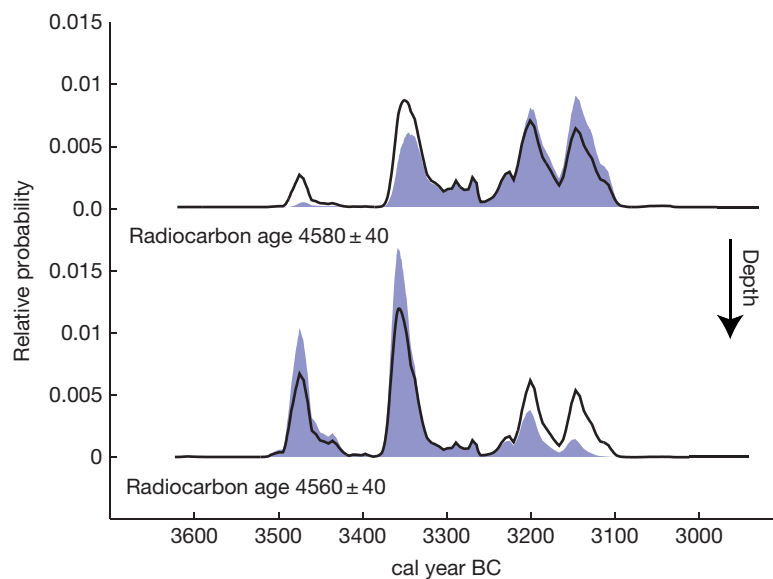


Figure 6 Probability distributions for two ^{14}C samples taken from a stratigraphic sequence with a simple calibration (black line) and with prior information (stratigraphic ordering) included using Bayes's theorem (solid purple). The radiocarbon ages of 4580 ± 40 BP (upper) and 4560 ± 40 BP (lower) are reversed as sometimes occurs in sediment records. The stratigraphic ordering forces the upper sample distribution toward younger calibrated ages and the lower toward older calibrated ages.

been collected before nuclear weapon testing (and ideally prior to the industrial era) because the underlying assumption that the atmosphere and ocean are in equilibrium is not valid during times of rapid atmospheric ^{14}C change. These samples are often difficult to obtain, particularly samples that are known to have been collected live or soon after death. In addition, different species of mollusk occupy different habitats and have different feeding patterns, which, in regions of carbonate rocks or freshwater input, can affect the ΔR value. ΔR is subtracted from the radiocarbon age, and the uncertainty is included in the radiocarbon age error before calibration with a marine calibration curve (Stuiver et al., 1986).

In many locations, ΔR has varied considerably with time because of changes in ocean circulation patterns or ice cover. Changes in ΔR have been documented through ^{14}C dating of marine and terrestrial pairs in archaeological sites or marine cores or by comparison of marine sample ages to well-dated tephra (Tephrochronology) layers. In regions where ΔR does vary with time t , it becomes somewhat of a circular argument to utilize a $\Delta R(t)$ value to correct a radiocarbon age for a sample without first knowing t . Regional marine calibration curves will undoubtedly be developed in the future to incorporate this information.

Freshwater Reservoir Corrections

Freshwater bodies generally reach equilibrium with the atmosphere relatively quickly, so lacustrine and riverine ^{14}C ages are calibrated with the atmospheric calibration curve. However, lakes and rivers, particularly in regions of carbonate bedrock, may have considerable reservoir ages (see ^{14}C of Plant Macrofossils). Moreover, old soil or peat carbon is often eroded into lakes. In volcanic regions, CO_2 released from subglacial eruptions or fissures in the lake bottom may contribute to a large

reservoir offset. ^{14}C measurements of surficial sediment samples or modern lacustrine plant material are used as a correction to the ^{14}C age before calibration, but the reservoir age may not have been constant over time. Lakes in newly deglaciated terrains, even in noncarbonate regions, may have an initial reservoir age due to the input of carbon that has been stored in the glacier since its formation, but the reservoir age tends to decrease as the water is replaced with groundwater and rainfall. ^{14}C analysis of terrestrial macrofossils or the identification of independently dated tephra layers is necessary to establish whether or not the reservoir age has been constant.

Mixed Marine and Terrestrial Samples

Animals that consume a mixed marine and terrestrial diet will contain carbon from both the atmospheric and marine reservoirs. When the tissue is radiocarbon dated, the marine proportion must be estimated and the calibration curve adjusted accordingly, or the results will be erroneous. The percent of marine carbon in the diet can be estimated in some cases from ethnohistorical documents, dental pathology, or stable isotopes, particularly of carbon and nitrogen. Much work has been done to determine the isotopic offset between the diet and various tissues in animals. Isotopic fractionation of carbon during photosynthesis causes an offset of the plant $^{13}\text{C}/^{12}\text{C}$ ratio from that of the carbon dioxide in atmosphere. Most plants follow one of two major photosynthetic pathways (C-3 and C-4), which have isotope fractionations that are distinctive, although CAMS plants may use either pathway. Additional fractionation also occurs when CO_2 dissolves in water, so that the $^{13}\text{C}/^{12}\text{C}$ of algae and phytoplankton falls between these two terrestrial plant values. It is therefore not possible to distinguish how much marine carbon has been consumed using carbon isotopes alone if both C-3 plants

(most temperate plants) and C-4 plants (tropical grasses including millet and maize) have been consumed. Nitrogen isotopes are fractionated with each step from primary production to consumption, so that marine samples, where there are multiple steps in the food chain, have higher $^{15}\text{N}/^{14}\text{N}$. By measuring both carbon and nitrogen isotopes in the sample, it is usually possible to estimate the marine carbon in the diet, although a complete analysis of food resources may be necessary to improve the uncertainty in the estimates. Where there are complicating factors such as freshwater fish in the diet, manuring of crops, or fractionation of nitrogen isotopes due to extreme heat or drought, measurement of other stable isotopes may be useful in separating the dietary sources.

Atmospheric Regional Effects

Despite the assumption, for calibration purposes, that the atmosphere is well mixed within each hemisphere, regional and local differences in atmospheric ^{14}C do occur and are designated as ΔR_a . Due to upwelling of old carbon in the Southern Ocean, the large ocean area and higher wind-driven air-sea mixing, the Southern Hemisphere atmospheric ^{14}C is strongly latitude dependent. At mid-latitudes, the Southern Hemisphere atmosphere is on average about 35 years older than the Northern Hemisphere, although this has varied from 0 to 100 years over the past two millennia (Hogg et al., 2009). In addition, the position of the intertropical convergence zone (ITCZ), the region of low pressure where the Northern and Southern Hemisphere trade winds meet, does not follow the equator and migrates seasonally as well as temporally. Thus, a plant growing near the ITCZ may utilize a mixture of Northern and Southern Hemisphere atmospheric CO_2 , which makes calibration uncertain. Other non-anthropogenic ^{14}C differences may be caused by higher ^{14}C production at high latitude and altitude, air-sea gas exchange near ocean margins in upwelling zones, CO_2 release from degradation of peat or soil, or outgassing of volcanoes or hydrothermal vents (see [Causes of Temporal \$^{14}\text{C}\$ Variations](#)). The first two of these have an effect of 10 or 20 years on the radiocarbon age of a sample at most, whereas the latter two can have a much larger effect in the immediate proximity of the source (e.g., Bruns et al., 1980; Walker et al., 2008).

There are also small offsets due to seasonal input of ^{14}C from the stratosphere into the troposphere. Plants growing early in the year may record a different ^{14}C level than those growing later in the year, which may be accentuated by climate changes. Small but significant differences, which would have an effect on calibration, have been measured between trees growing in Germany and those in Turkey during cold climate intervals (Kromer et al., 2001). An offset of 19 years has also been measured in herbarium specimens from Egypt (Dee et al., 2010). Fossil fuel usage may cause a localized depression in ^{14}C levels in addition to the general atmospheric ^{14}C decline from the beginning of the industrial era until the nuclear era, resulting in large calibrated age ranges.

Nuclear weapons testing resulted in high ^{14}C production in the upper atmosphere in the 1950s–1960s. Most of ‘bomb’ ^{14}C was input into the Northern Hemisphere and has gradually mixed throughout the atmosphere, but the distribution has not been globally uniform (Levin and Kromer, 2004).

Identification of the ‘bomb’ peak in sediments or peats provides a precise and useful time marker. Atmospheric and tree-ring ^{14}C measurements from various regions have been used to calibrate ^{14}C samples from the mid-1950s to the present. ‘Bomb’ calibrations have become increasingly useful in forensics and environmental applications. Mixing of ‘bomb’ ^{14}C into the ocean was even more variable making calibration impossible except in a few regions where records exist.

Future of Calibration

Floating European chronologies have recently been linked to the Hohenheim pre-boreal pine chronology by wiggle-matching both to the Huon pine chronology adjusted for the Southern Hemisphere offset (Hua et al., 2009). However, extending European chronologies through the last glacial period is a slow process due to the scarcity of trees growing in Europe at that time. Subfossil deposits of long-lived kauri (*Agathis australis*) in New Zealand offer the potential for tree-ring records back to the limit of radiocarbon dating at ~55 000–60 000 years ago. Even if these records cannot be firmly anchored to existing kauri chronologies extending from the present, they will provide useful information about variations in atmospheric ^{14}C (Turney et al., 2010).

Significant progress is being made on varved lake records. For instance, ‘missing’ varves in the Lake Suigetsu sequence are now being identified in the Lake Suigetsu 2006 Project by overlapping multiple cores and improved varve-counting techniques. ^{14}C dating of terrestrial macrofossils in the sediments may provide a reliable record of atmospheric ^{14}C levels for much of the radiocarbon age range and help to quantify the extent of reservoir changes in the marine records (Staff et al., 2010).

Other areas of research will have an impact on ^{14}C calibration. The extent of past marine reservoir changes is critical for the reliability of the marine calibration records and much work is being done to quantify these in many parts of the world. ^{14}C dating of organic material trapped in Ar–Ar (see [K/Ar and \$^{40}\text{Ar}/^{39}\text{Ar}\$ Dating](#))-dated volcanic deposits could provide essential atmospheric ^{14}C data as well. Independent dating of the geomagnetic excursions (see [Geomagnetic Excursions and Secular Variations](#)), such as the Laschamps event ~40 000 years ago, may provide a time marker for synchronizing ^{14}C records from different environments. Modeling of ^{14}C production based on paleomagnetic intensity data and records of other cosmogenic isotopes, such as ^{10}Be from ice cores, contributes to the understanding of ^{14}C variations (Laj et al., 2002; Muscheler et al., 2008), which may aid in selection of data sets for calibration curve construction.

Conclusions

Radiocarbon calibration curves have changed from relatively simple tree-ring-only compilations to the incorporation of different types of archives. This has required the development of specialized statistical tools and an improved understanding of ^{14}C distribution in the global carbon reservoirs. Although simple calibrations can result in imprecise ages owing to the

varying ^{14}C content of the atmosphere or ocean, events can be pinpointed more precisely by incorporating additional information such as stratigraphy in a Bayesian framework. As radiocarbon measurements on small samples have become more precise, so too has the need for consideration of regional reservoir effects. Extension and refinement of the calibration curves will require concerted effort by many individuals. While consensus ^{14}C calibration curves are an essential tool for comparisons between records, there is also the need to provide a sufficient means of recognition for the researchers who provide the individual calibration data sets.

See also: Dendroclimatology; Varved Lake Sediments; Varved Marine Sediments. **Introduction:** History of Dating Methods. **Radiocarbon Dating:** Causes of Temporal ^{14}C Variations. **Sea Level Studies:** Coral Records of Relative Sea-Level Changes; Sedimentary Indicators of Relative Sea-Level Changes – Low Energy.

References

- Bard E, Menot-Combes G, and Rostek F (2004) Present status of radiocarbon calibration and comparison records based on Polynesian corals and Iberian Margin sediments. *Radiocarbon* 46(3): 1189–1202.
- Blaauw M and Christen JA (2011) Flexible paleoclimate age-depth models using an autoregressive gamma process. *Bayesian Analysis* 6: 457–474.
- Blackwell PG and Buck CE (2008) Estimating radiocarbon calibration curves. *Bayesian Analysis* 3: 225–248.
- Bronk Ramsey CB (2008) Deposition models for chronological records. *Quaternary Science Reviews* 27: 42–60.
- Bruns M, Levin I, Munnich KO, Hubberten HW, and Fillipakis S (1980) Regional sources of volcanic carbon-dioxide and their influence on C-14 content of present-day plant-material. *Radiocarbon* 22: 532–536.
- Buck CE, Kenworthy JB, Litton CD, and Smith AFM (1991) Combining archaeological and radiocarbon information: A Bayesian approach to calibration. *Antiquity* 65(249): 808–821.
- de Vries H (1958) Variation in concentration of radiocarbon with time and location on earth. *Proceedings of the Koninklijke Nederlandse Akademie Van Wetenschappen Series B: Palaeontology Geology Physics Chemistry Anthropology* B61: 94–102.
- Dee MW, Brock F, Harris SA, et al. (2010) Investigating the likelihood of a reservoir offset in the radiocarbon record for ancient Egypt. *Journal of Archaeological Science* 37: 687–693.
- Friedrich M, Remmele S, Kromer B, et al. (2004) The 12,460-year Hohenheim oak and pine tree-ring chronology from Central Europe – A unique annual record for radiocarbon calibration and palaeo-environment reconstructions. *Radiocarbon* 46(3): 1111–1122.
- Hoffmann DL, Beck JW, Richards DA, et al. (2010) Towards radiocarbon calibration beyond 28 ka using speleothems from the Bahamas. *Earth and Planetary Science Letters* 289: 1–10.
- Hogg A, Palmer J, Boswijk G, Reimer P, and Brown D (2009) Investigating the interhemispheric C-14 offset in the 1st millennium AD and assessment of laboratory bias and calibration errors. *Radiocarbon* 51: 1177–1186.
- Hua Q, Barbetti M, Fink D, et al. (2009) Atmospheric ^{14}C variations derived from tree rings during the early Younger Dryas. *Quaternary Science Reviews* 28: 2982–2990.
- Hughen K, Southon J, Lehman S, Bertrand C, and Turnbull J (2006) Marine-derived C-14 calibration and activity record for the past 50,000 years updated from the Cariaco Basin. *Quaternary Science Reviews* 25: 3216–3227.
- Klein J, Lerman JC, Damon PE, and Ralph EK (1982) Calibration of radiocarbon-dates – Tables based on the consensus data of the workshop on calibrating the radiocarbon time scale. *Radiocarbon* 24: 103–150.
- Kromer B, Manning SW, Kuniholm PI, Newton MW, Spurk M, and Levin I (2001) Regional $^{14}\text{CO}_2$ offsets in the troposphere: Magnitude, mechanisms and consequences. *Science* 294: 2529–2532.
- Kuzmin YV, Slusarenko IY, Hajdas I, Bonani G, and Christen JA (2004) The comparison of ^{14}C wiggle-matching results for the 'floating' tree-ring chronology of the Ulandryk-4 burial ground (Altai Mountains, Siberia). *Radiocarbon* 46: 943–948.
- Laj C, Kissel C, Mazaud A, Michel E, Muscheler R, and Beer J (2002) Geomagnetic field intensity, North Atlantic Deep Water circulation and atmospheric Delta C-14 during the last 50 kyr. *Earth and Planetary Science Letters* 200(1–2): 177–190.
- Levin I and Kromer B (2004) The tropospheric $^{14}\text{CO}_2$ level in mid-latitudes of the Northern Hemisphere (1959–2003). *Radiocarbon* 46(3): 1261–1272.
- Mallory JP, McCormac FG, Reimer PJ, and Marsadolov LS (2002) The date of Pazyryk. In: Boyle K, Renfrew C, and Levine MA (eds.) *Ancient Interactions: East and West in Eurasia*, p. 344. Cambridge: The McDonald Institute for Archaeological Research.
- Marsadolov LS (1996) Problems and prospects of absolute tree-ring dating for the Sayan-Altai archaeological monuments (first millennium BC). In: Dean JS, Meko D, Sweetman T (eds.) *Tree Rings, Environment and Humanity: Proceedings of the International Conference*, Tucson, Arizona, 17–21, Tucson, May 1994. AZ: Radiocarbon, Dept. of Geosciences, University of Arizona p 557–566.
- Muscheler R, Kromer B, Björck S, et al. (2008) Tree rings and ice cores reveal ^{14}C calibration uncertainties during the Younger Dryas. *Nature Geoscience* 1: 263–267.
- Reimer PJ, Baillie MGL, Bard E, et al. (2009) IntCal09 and Marine09 radiocarbon age calibration curves, 0–50,000 years cal BP. *Radiocarbon* 51: 1111–1150.
- Reimer PJ, Hughen KA, Guilderson TP, et al. (2002) Preliminary report of the first workshop of the IntCal04 radiocarbon calibration/comparison working group. *Radiocarbon* 44(3): 653–661.
- Staff RA, Ramsey CB, and Nakagawa T (2010) Suigetsu 2006 Project Members: A re-analysis of the Lake Suigetsu terrestrial radiocarbon calibration dataset. *Nuclear Instruments and Methods in Physics Research Section B: Beam Interactions with Materials and Atoms* 268: 960–965.
- Stuiver M, Burk RL, and Quay PD (1984) $^{13}\text{C}/^{12}\text{C}$ ratios in tree rings and the transfer of biospheric carbon to the atmosphere. *Journal of Geophysical Research* 89: 731–748.
- Stuiver M, Pearson GW, and Braziunas T (1986) Radiocarbon age calibration of marine samples back to 9000 cal yr BP. *Radiocarbon* 28(2B): 980–1021.
- Stuiver M and Polach HA (1977) Discussion: Reporting of ^{14}C data. *Radiocarbon* 19(3): 355–363.
- Stuiver M and Reimer P (1989) Histograms obtained from computerized radiocarbon age calibration. *Radiocarbon* 31(3): 817–823.
- Stuiver M and Suess HE (1966) On the relationship between radiocarbon dates and true sample ages. *Radiocarbon* 8: 534–540.
- Telford RJ, Heegaard E, and Birks HJB (2004) The intercept is a poor estimate of a calibrated radiocarbon age. *The Holocene* 14(2): 296–298.
- Turney CSM, Fifield LK, Hogg AG, et al. (2010) The potential of New Zealand kauri (*Agathis australis*) for testing the synchronicity of abrupt climate change during the last glacial interval (60,000–11,700 years ago). *Quaternary Science Reviews* 29: 3677–3682.
- Walker BD, McCarthy MD, Fisher AT, and Guilderson TP (2008) Dissolved inorganic carbon isotopic composition of low-temperature axial and ridge-flank hydrothermal fluids of the Juan de Fuca Ridge. *Marine Chemistry* 108: 123–136.

Charcoal

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Introduction

Wood from an Egyptian tomb, linen used to wrap the Dead Sea scrolls, and a bread roll were among the first samples of known age used to test the radiocarbon dating technique. The bread roll was pyrolyzed to charcoal by the volcanic eruption of Pompeii in AD 79, and the radiocarbon date obtained on the sample was within error of the known age of the eruption (Arnold and Libby, 1949). This was the first ever radiocarbon date on charcoal, followed by a further dozen analyses of charcoal and charred wood from The Netherlands that suggested that charcoal could routinely produce reliable radiocarbon dates with minimal sample pretreatment (De Vries and Barendsen, 1954). Libby (1955) considered charcoal to be one of the most reliable materials for radiocarbon dating, and since that time, thousands of radiocarbon dates on charcoal have been produced by over 130 laboratories around the world. Radiocarbon dates on charcoal have contributed substantially to the chronological framework that underpins our understanding of human history and prehistory, as well as trajectories of change in the natural environment over the last 50 000–60 000 years.

In the last two decades, the ability to measure the radiocarbon activity of a few milligrams of carbon by accelerator mass spectrometry (AMS) (see [AMS Radiocarbon Dating](#)) has enabled the measurement of samples an order of magnitude smaller in size than was previously possible (see [Conventional Method](#)). This has not only allowed the dating of samples too small to date by conventional techniques, but has also extended the range of carbonized materials that can be dated to aerosol-sized 'black carbon' and individual biomarker compounds produced during combustion/pyrolysis (e.g., Reddy et al., 2002). As a result, radiocarbon dating of 'charcoal' is increasingly being used to support research into the biogeochemical cycling of carbon as well as the more traditional geochronological applications in archaeology and paleoenvironmental reconstruction.

'Charcoal': Definitions, Structure, and Chemistry

Pyrolysis accompanying the incomplete combustion of organic material produces a wide spectrum of particulate, carbon-rich, and generally O–H–S–N-poor compounds. This material is known by a variety of names including charcoal, microcharcoal, carbonaceous spherules, soot, fusain, elemental carbon, oxidation-resistant elemental carbon, microcrystalline graphite, black carbon, and graphitic black carbon. The very range of names used to describe this material reflects not only its compositional complexity but also its importance across a range of disciplines, each of which has evolved not only a terminology but also a set of methodologies for the analysis of this material and interpretation of the results.

In an archaeological and paleoenvironmental context, charcoal generally refers to macroscopic material resulting from the incomplete combustion of woody plant tissues ([Figure 1](#)). The

term 'charcoal' is widely recognized and hence it is used in this review, but an important conceptual development has been the recognition that 'charcoal' and the components that make up charcoal form part of a 'combustion continuum' (e.g., Masiello, 2004). At the low-temperature end of this continuum lies slightly charred biomass of low aromaticity, high reactivity, and comparatively large (mm to cm) size, while at the other end lie submicron-sized black carbon particles and a range of individual pyrogenic molecules, formed from gas phase condensation at high temperature and having high aromaticity and low reactivity ([Figure 2](#)). Macroscopic 'charcoal' lies at the center of this continuum and thus potentially contains a wide variety of components of variable reactivity.

The morphology and chemical composition of charcoal is initially controlled by the material that is pyrolyzed. Woody tissues are composed primarily of lignin and cellulose, cross-linked into a structure that is rigid and rich in polyphenolic compounds, while leaves contain higher proportions of non-structural material such as sugars and waxes. Most macroscopic charcoal preserved in the geologic record is derived from wood, and the open porous structure of woody tissues is often preserved intact after conversion of the wood to charcoal ([Figure 3](#)). The chemistry of charcoal is also dependent on the conditions of pyrolysis, with a general increase in carbon content and aromaticity along with a decrease in H/C and O/C ratios as temperature increases (Masiello, 2004). Black carbon forms from the condensation of polycyclic aromatic hydrocarbons into hydrogen-poor solid spheres generally <1 µm in diameter and as a result retains none of the structures of the pyrolyzed parent material (e.g., Gustafsson et al., 2001; [Figure 4](#)).

Franklin (1951) noted that the x-ray spectra of many pyrolyzed materials suggested the presence of microcrystalline graphitic carbon, and that upon heating to higher temperatures, microcrystalline graphite formed readily from some precursor materials and not others. She proposed the existence of 'organized' and 'nonorganized' components in most pyrolyzed material, with the proportion of each dependent upon the chemical structure of the starting material. Support for this model has come from high-resolution transmission electron microscopy, which has identified the presence of nanometer-scale domains of microcrystalline graphite in an 'amorphous' matrix in natural macroscopic charcoal fragments (Cohen-Ofri et al., 2005; [Figure 5](#)). This complexity and heterogeneity has significant implications for radiocarbon dating.

Charcoal Stability

The presumption of long-term stability of charcoal in the environment is based on several observations including the high degree of resistance of some charcoals to a range of oxidants (Bird and Gröcke, 1997; Skjemstad et al., 1996; Wolbach and Anders, 1989) and its preservation for long periods of time in the geologic record (Cope and Chaloner, 1980). It is

also known that charcoal is a significant component of many soils where it can persist for thousands of years (e.g., Hopkins et al., 1990; Pessenda et al., 2001) and can form a significant component of the 'inert' soil carbon pool (Lehmann et al., 2008).

However, evidence is accumulating that charcoal may not be as stable as previously supposed. A number of mass balance

studies have demonstrated significant loss of PC on decadal to centennial timescales (Bird et al., 1999a,b; Czimczik et al., 2005; Hammes et al., 2008) but not on annual timescales (Eckmeier et al., 2007; Major et al., 2009). Support for the degradation of at least some components of soil charcoal has also come from Glaser et al. (1998) who concluded that a proportion of the highly aromatic fraction of some humic acids may be derived from charcoal, and from evidence for the *in situ* production of humic acids from charcoal (Ascough et al., 2011a).

In addition to microbial degradation, it has also been shown that charcoal is susceptible to photooxidation, albeit at a slower rate than organic carbon (Skjemstad et al., 1996), and experiments have demonstrated that exposure of black carbon to normal atmospheric concentrations of ozone leads to rapid surface oxidation, in turn rendering black carbon particles soluble in distilled water (Chughtai et al., 1991).

Two main factors during production have been shown to influence the susceptibility to chemical oxidation of charcoal and by inference are likely to be of importance in determining the stability of charcoal in the natural environment. The first is the efficiency of the pyrolysis process, with high-temperature and low-oxygen availability favoring the formation of carbon-rich charcoal containing little labile organic matter. Such charcoals have a high carbon content (>75%) and a high resistance to chemical oxidation. The second is the density of the organic matter and hence the surface area of the charcoal produced by



Figure 1 Charcoal embedded in lime. Formed during the eruption of Pompeii in AD 79. Photograph by Jennifer Stephens, Jill Thompson, and Rick Jones, with permission.

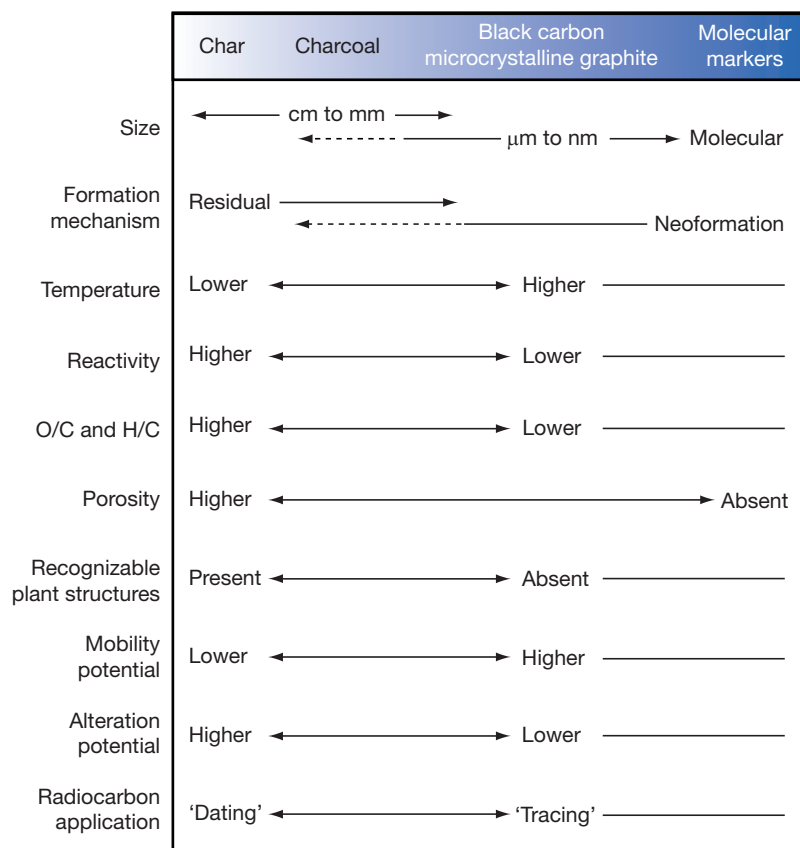


Figure 2 The 'combustion continuum' and variations in the characteristics of pyrolyzed material along the continuum of relevance to radiocarbon dating.

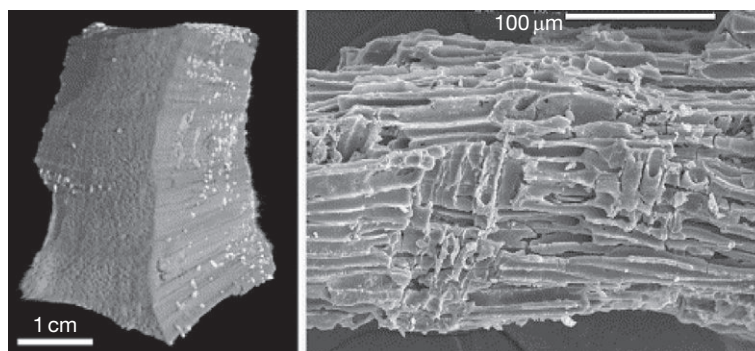


Figure 3 Left, X-ray tomographic image of charcoal formed by a volcanic eruption at 52 000 years BP near Maninjau, Sumatra (photograph by Iain Young, with permission). Right, scanning electron micrograph of a charcoal fragment dated at 48 000 years BP from layer 39 of the Devils Lair archaeological site in South-West Australia. Reproduced from [Turney CSM, Bird MI, Fifield LK, et al. \(2001\)](#) Breaking the radiocarbon barrier and early human occupation at Devil's Lair southwestern Australia. *Quaternary Research* 55: 3–13.

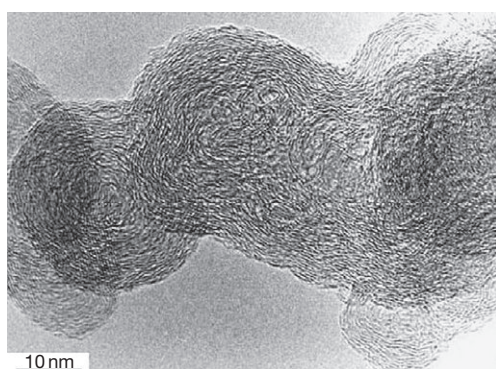


Figure 4 High-resolution transmission electron micrograph of black carbon in SRM-1650 showing spherical structures resulting from gas phase condensation. Reproduced from [Gustafsson Ö, Bucheli TD, Kukulska Z, et al. \(2001\)](#) Evaluation of a protocol for the quantification of black carbon in sediments. *Global Biogeochem. Cycles* 15: 881–890.

pyrolysis. High-density woody material, with comparatively low surface area, yields charcoal that is more resistant to chemical oxidation than charcoal produced from lower density organic matter, such as leaves ([Bird and Gröcke, 1997](#)).

While it is known that charcoal exhibits a wide range of behaviors when subjected to chemical or thermal oxidation, the variations in charcoal chemistry that must underlie these behaviors has received little attention. The emerging evidence suggests that while there may well be some highly refractory carbon present in most charcoal, a significant proportion may be labile and reactive on comparatively short timescales (e.g., [Ascough et al., 2011b](#)).

The significance of charcoal stability for radiocarbon dating has been underscored by an intensive study of the Nauwalabila I archaeological site in Northern Australia. Attempts to date charcoal at depth in the sequence using a range of pretreatment protocols were completely unsuccessful due to pervasive alteration and contamination of the charcoal as a result of oxidative degradation and exposure to a fluctuating groundwater table ([Bird et al., 2002](#)). In addition, comparison of modern and ancient charcoal from archaeological sites in Israel using an extensive suite of analytical techniques has demonstrated that the fossil charcoals have undergone

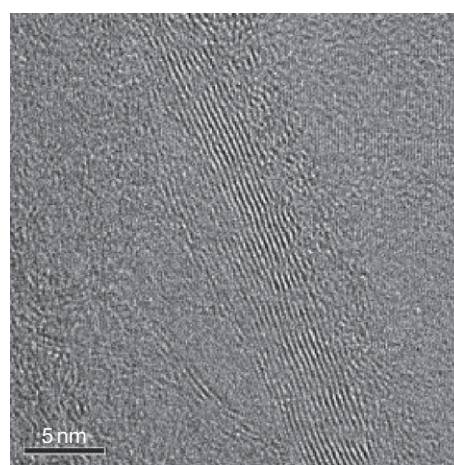


Figure 5 High-resolution transmission electron micrograph of modern charcoal (*Ceratonia*) showing the presence of 'organized' domains of microcrystalline material in an amorphous matrix, as discussed by [Cohen-Ofri et al. \(2005\)](#). The image was taken by Dr. Ronit Popovitz-Biro, Department of Chemical Infrastructure, Weizmann Institute of Science, Israel. Scale bar is 5 nm in length.

significant chemical alteration, in turn raising the possibility that significant contamination could have been introduced during the process of alteration ([Cohen-Ofri et al., 2005](#)).

Charcoal Contamination

The main prerequisite for the reliable measurement of radiocarbon activity is the material being analyzed. If charcoal will degrade, as appears to be the case, this must also mean that there is potential for contamination that is not easily removed, as contaminants may have been involved in irreversible reactions with the charcoal surface. There are several mechanisms by which charcoal may be contaminated to the extent that the reliability of a radiocarbon measurement on the sample may be compromised:

- (i) Charcoal has a large internal and external surface area and acts as a sorbent for a range of organic and inorganic

compounds, with 'activated' carbon being an extreme example (Gustafsson et al., 2001). During what is often millennia between the formation of charcoal and its sampling for analysis, charcoal in the natural environment will have been exposed to soil solutions containing a range of organic compounds and some, including carbohydrates, amino acids, and phenolics, have been shown to adsorb readily to natural charcoal (Pietikäinen et al., 2000). These compounds may be loosely bound but some may undergo irreversible reactions with functional groups on the charcoal surface. pH is likely to exert a significant control on both the degradation and contamination of charcoal (e.g., Braadbaart et al., 2009).

- (ii) The capacity of charcoal to adsorb metabolizable compounds, coupled with its porous nature, makes it a suitable substrate for microbial colonization, and several studies have demonstrated enhanced soil microbial activity associated with charcoal (e.g., Zackrisson et al., 1996). This cycling of carbon between the charcoal and the local environment can continue long after the initial formation of the charcoal and a portion of this material may be irreversibly incorporated in the charcoal.
- (iii) Combustion of organic matter leaves a residue of ash, the main component of which is commonly CaO (lime) along with oxides of other cations. Over time, this material becomes hydrated and is ultimately converted to carbonate (Demeyer et al., 2001). Calcium carbonate is often intimately associated with charcoal in alkaline environments and represents an additional potential source of carbon contamination. In some environments, iron carbonate (siderite) may also be formed during pedogenesis, and as this mineral is relatively insoluble in acid, it may not be readily removed by some pretreatment techniques designed to eliminate contaminants (Hatté et al., 2001).

Sampling Charcoal for Radiocarbon Dating

The recognition that charcoal forms part of a 'combustion continuum' means that there is no single technique for sampling charcoal for radiocarbon measurement. Sampling of macroscopic charcoal, from archaeological sites and soil/sedimentary sequences is relatively straightforward. Fragments can be handpicked from the sedimentary matrix, wrapped in solvent-washed aluminum foil, and sealed in a labeled plastic bag. Provided a modicum of care is exercised in sampling so that no foreign contaminants are introduced, and provided the stratigraphic context of the sample is recorded, the sample can then be dated using radiocarbon. Smaller fragments of charcoal, down to a few hundred microns in diameter, can be concentrated by dry or wet sieving. Charcoal retained on the sieve can be further concentrated by swirling the material in water and decanting off the charcoal and organic particles, which remain in suspension longer than the heavier mineral material. Final purification of a charcoal separate can then be achieved by handpicking a few milligrams of charcoal under a binocular microscope. At some particle sizes, determined by the importance of the sample and patience of the analyst, handpicking of charcoal becomes impractical and chemical

treatments are required to concentrate sufficient charcoal for dating as discussed in the following section.

Pretreatment of Charcoal for Radiocarbon Dating

Modern AMS target production lines can potentially yield finite radiocarbon dates on charcoal of ~55 000 years for a 1 mg target (Bird et al., 1999a,b; Pigati et al., 2007), but this requires that samples be completely free of carbon not indigenous to the charcoal. Chemical pretreatments for charcoal can be classified into two broad categories, the first applied to macroscopic charcoal physically separated from a sedimentary matrix and the second aimed at chemically or thermally isolating fine charcoal from other organic matter in a sedimentary matrix. The aim of all these pretreatments is to remove both younger and older organic as well as inorganic carbon from the charcoal so that the measured radiocarbon activity is directly related to the radiocarbon activity of the sample at the time it was pyrolyzed.

There are generic precautions that should be taken to minimize laboratory contamination, regardless of pretreatment strategy. Thus all pretreatment steps should be carried out in either a laminar flow cabinet, in a HEPA-filtered clean laboratory, or in capped vials with all vessels and implements pre-cleaned either by combustion overnight at 550 °C (Pyrex), 880 °C (quartz) or by washing with a hot solution of 0.1 M $K_2Cr_2O_7$ in a 2 M solution of H_2SO_4 . All water used should be as pure as possible.

Acid-Alkali-Acid

De Vries and Barendsen (1954) first used combinations of acid and alkali extraction as a pretreatment for radiocarbon measurement of 'charred wood' and demonstrated that in some cases acid-soluble material was considerably younger than either the alkali-soluble or insoluble fractions. These results indicated that charcoal could be contaminated by younger carbon, but the contamination could apparently be easily removed. The acid wash is designed to remove 'fulvic acids,' and the alkali wash is designed to remove 'humic acids.' Recognition that alkaline solutions will take up atmospheric carbon dioxide (and hence ^{14}C) that might then be irreversibly incorporated into the sample led to the addition of a final acid wash designed to eliminate this possibility.

This general protocol, which is widely used for many organic sample types, has evolved into the acid-alkali-acid (AAA) pretreatment, although, in practice, the actual protocol employed differs substantially between laboratories. The AAA technique is widely used for charcoal but has also been used with success on cremated bone remains (Nakamura et al., 2010). The acid used is generally HCl but the base may be NaOH, NH_4OH , or KOH, while the strength of these solutions varies between 0.1 and 1 M, the temperature from ambient to boiling, and the time from less than 1 h to 1 day, with the possibility of multiple extractions, particularly at the alkali step. In part, this flexibility in application is required because of widely differing responses of charcoal to the procedure, as the alkali extraction step, in particular, may result in anything from zero to complete sample dissolution.

The addition of an initial HF or HF/HCl wash to remove silicates or aqua regia ($\text{HNO}_3 + \text{HCl}$) to remove iron minerals and thereby expose protected organic contaminants to the reagents and the replacement of the final HCl wash with either H_2SO_4 or an acid oxidation reagent such as nitric acid have been proposed to improve the performance of the AAA pretreatment (e.g., Gillespie et al., 1992; Haesaerts et al., 2010; Hatté et al., 2001). An additional improvement to the AAA pretreatment has been the use of a range of techniques including Raman and Fourier transform infrared spectroscopy to screen charcoal samples for radiocarbon measurement and monitor the removal of humic acid contamination through successive steps of the AAA procedure (Alon et al., 2002; Rebollo et al., 2008).

Acid-Base Oxidation with Stepped Combustion

In order to address the issue that the AAA pretreatment does not always remove contamination from charcoal to a consistently low level, Bird et al. (1999a,b) developed a more rigorous pretreatment for charcoal samples. The acid-base oxidation (ABOX) pretreatment is based on methods developed to quantify 'elemental carbon' in sediments (Bird and Gröcke, 1997; Wolbach and Anders, 1989) and is similar to AAA but replaces the final acid wash with an acid oxidation step designed to chemically oxidize any contaminants tightly bound to the charcoal surface.

Handpicked fragments of charcoal are crushed in a mortar and pestle and a 10–200 mg aliquot is weighed into a 50 ml capped polypropylene centrifuge tube. 20 ml of 6 M HCl is added to the tube, which is then capped and left to stand for an hour, after which it is centrifuged and rinsed twice. If the sample contains appreciable mineral impurities, a 1:1 mix of concentrated HCl and HF can be used in place of a simple acid wash to dissolve any mineral phases. 30 ml of 1 M NaOH is then added to the tube, which is left to stand for 30 min before rinsing.

Thirty milliliters of acid-dichromate oxidant solution (0.1 M $\text{K}_2\text{Cr}_2\text{O}_7$ in a 2 M solution of H_2SO_4) is added, followed by heating to 60 °C for 14–24 h. The oxidation time is a compromise required to ensure that sufficient sample material is left for further analysis. With larger samples or more dense charcoal fragments, the oxidation time can be increased to 72 h or longer. Following completion of the oxidation step, the sample is washed several times with hot water with fine particles that remain in suspension for longer than 5 min also discarded, and then the sample is dried and weighed.

The concept of stepped combustion as developed by Bird et al. (1999a,b), derives from thermal techniques for quantifying fine 'charcoal' in a range of environmental materials including aerosols (Cachier et al., 1989) and sediments (e.g., Gustafsson et al., 2001) and can be applied to raw, AAA, or ABOX pretreated samples. Rather than simply combusting the total sample at high temperature, heating at 330–360 °C in an oxygen atmosphere results in the combustion of any organic contaminants still adhering to the charcoal surface or added during sample handling. A second combustion step at 650 °C initiates the combustion of the charcoal and a final combustion step at 850–900 °C leads to the complete combustion of the uncontaminated 'core' of the sample. A graphite target for

radiocarbon determination can be made from the CO_2 produced during each combustion step.

As thermal oxidation can be expected to remove less resistant phases first and work progressively from the outside to the interior of the particles, any contaminants remaining after the sample pretreatment will be combusted during the low-temperature step, and the 'reliability' of the radiocarbon measurement can be assessed by the coherence of the results from the high-temperature steps. In practice, the low-temperature fraction is usually contaminated to some degree and can simply be discarded.

Tests on raw, AAA, and ABOX pretreated samples of known age and known 'infinite' age have demonstrated that acid-base oxidation with stepped combustion (ABOX-SC) out-performs other protocols by reducing contamination to consistently low levels of ~0.1pMC (percent modern carbon) or better, equivalent to an 'age' of approximately 60 000 years BP and meaning that under favorable conditions, reliable dates up to ~58 000 years BP are achievable (Bird et al., 2003). Several studies have now demonstrated that the ABOX pretreatment provides reliable decontamination and can perform better than AAA pretreatments, particularly for samples older than 40 000 years (e.g., Douka et al., 2010; Higham et al., 2009a,b; Turney et al., 2001).

Other Techniques

The ABOX-SC pretreatment is very harsh and often results in large losses of sample material. Likewise, in many instances, the alkali step of the AAA pretreatment may result in virtually complete dissolution of charred samples (Hedges et al., 1989; Rebollo et al., 2008). A recently developed alternative technique is the use of catalytic addition of hydrogen to reductively separate labile and refractory PC components. This is achieved via pyrolysis at up to 550 °C under high H_2 gas pressures, known as hydrogen pyrolysis (hypy). Hypy results in total reductive conversion and removal of common PC contaminants, such as cellulose and humic substances, isolating a chemically consistent, highly aromatic component from both macroscopic charcoal and fine charcoal from complex matrices such as soils (Ascough et al., 2009, 2010). In this regard, hydrogen pyrolysis offers advantages as a ^{14}C pretreatment, as it is possible to effectively degrade compound classes routinely targeted for removal without unnecessary loss of the resistant indigenous charcoal component of interest for dating, thus minimizing sample loss.

Low-temperature oxygen plasma ashing has also shown promise as a technique for decontaminating charcoal prior to radiocarbon dating. Plasma ashing has been demonstrated to be rapid and controllable, and to perform as well as the AAA technique in decontamination. Because the technique is limited to the sample surface accessible to the plasma, it may not always remove all contamination but provides a viable alternative in cases where more aggressive techniques cannot be used because of small sample size (Bird et al., 1999a,b).

Radiocarbon Dating of 'Fine Charcoal'

Many techniques for the isolation and quantification of products of combustion too small to be separated by hand have been developed over the last decade and radiocarbon measurements of these samples are increasingly being used for the

purposes of dating and source attribution. Source attribution commonly relies on the fact that, unlike charcoal derived from biomass burning, fossil fuel-derived combustion products do not contain radiocarbon. These techniques can be categorized into those aimed at the isolation of particulate material and those that isolate individual compounds produced during combustion. For recent reviews of these methods, see Hammes et al. (2007), Masiello (2004), and Schmidt et al. (2001).

Techniques aimed at particulate isolation fall broadly into two groups, using both an acid treatment to remove carbonates and an HF treatment (used as required) to remove silicates that can protect organic matter. Thermochemical techniques utilize the fact that organic matter oxidizes at a lower temperature than charcoal. Thus, any carbon surviving after heating in an oxygen-containing atmosphere to a temperature of generally 340–375 °C for a predetermined time is considered to have derived from burning (e.g., Gustafsson et al., 2001). The main issue with these techniques, in the context of radiocarbon measurement, is the possibility that the thermal treatment leads to the charring of some organic matter that, in turn, can ‘fix’ contamination into the sample.

Wet chemical techniques use a variety of chemical oxidants including potassium dichromate, potassium permanganate, sodium hypochlorite, nitric acid, or photochemical oxidation using UV-radiation (e.g., Eckmeier et al., 2009; Hammes et al., 2007), at a range of temperatures for a range of times to oxidize organic material. The main issue with chemical oxidation is the possibility of sample loss through the procedure, which is important only if quantification of the material is required.

Compound-specific techniques use either solvent extraction or oxidation to isolate compounds known to derive from combustion, usually polycyclic aromatic hydrocarbons (Gustafsson et al., 2009; Reddy et al., 2002), levoglucosan (Elias et al., 2001), or benzene carboxylic acids (Glaser et al., 1998). These are then separated and quantified by high pressure liquid chromatography or gas chromatography–mass spectrometry. Advances in AMS that now enable on-line, continuous-flow radiocarbon determinations of individual compounds separated by gas chromatography are likely to lead to an increase in the analysis of these compounds in the future for both geochronological and source attribution applications.

In the context of radiocarbon dating, the choice of any of the above techniques is dependent on the application. If geochronology is the prime aim, then a rigorous protocol combining wet chemical and thermochemical techniques, such as ABOX-SC, is most appropriate, whereas if quantification for the purposes of source attribution is the aim, then a ‘gentler’ protocol should be adopted.

Conclusion: Interpretation of Radiocarbon Dates on Charcoal

Discussion in the preceding sections has highlighted that ‘charcoal’ should be thought of as part of a combustion continuum, the components of which range from macroscopic to molecular and from refractory to comparatively labile. Assuming that the fundamental issue of contamination has been dealt with through appropriate pretreatment and analytical protocols,

there remain several potential issues in the interpretation of the radiocarbon measurement of a sample. These relate to the nature of the charcoal isolated for analysis, the context from which the sample was obtained, and the purpose for which the analysis was made.

If the purpose of the measurement is to obtain the age of an event then the radiocarbon ‘date’ must be calibrated to calendar years. Charcoal produced from the combustion of terrestrial biomass produced from photosynthetic fixation of atmospheric CO₂ is not subject to an inherent reservoir effect, and hence calibration to calendar years can be based on the atmospheric calibration curve (see [Calibration of the ¹⁴C Record](#)). However, Gavin (2001) found that radiocarbon dates on charcoal derived from forest fires on Vancouver Island were 180–670 years older than the known age of the fires, and Nakamura et al. (2010) found macroscopic charcoal to be older than cremated bone, in both cases due to the inbuilt age of the wood prior to combustion. Inbuilt ages in charcoal of a few decades are likely to be common as wood takes time to grow, and this needs to be considered when interpreting radiocarbon dates on charcoal ([Figures 6 and 7](#)).

Radiocarbon dates on macroscopic charcoal are likely to provide a reasonably accurate estimate of the time of burning in most cases, but often such dates are used to infer the timing of a separate event by stratigraphic association. In a case where charcoal is clearly associated with a hearth, the association between the age of formation of the charcoal and the timing of human use of the hearth is fairly direct. In many instances, this is not the case, and the age of an event is inferred simply from a stratigraphic association between charcoal and an artifact or change in the composition of a sedimentary sequence. This becomes increasingly difficult to substantiate as the size of the charcoal decreases for two reasons:



Figure 6 Archaeological excavation at Pedra Furada, Capivara National Park, Brazil. A mid-Holocene hearth associated with abundant charcoal fragments is visible in the center bottom of the frame 1.5 m below the modern ground surface. Charcoal fragments are also visible in the walls of the excavation as black specks. Photograph: Michael Bird.



Figure 7 A cleaned eroding fluvial section that revealed a cross-section of a Norse charcoal production pit, in Eyjafjallsveit, Southern Iceland. The pit was cut through the Eldja AD 935 tephra and after its use was covered with eolian material from natural soil accumulation and eventually the tephra fall of Hekla AD 1341. Radiocarbon dating of the basal charcoal fill places the use of the pit into the late twelfth to thirteenth centuries AD. Photograph by Mike Church, with permission.

- (i) Physical bioturbation after deposition can result in the vertical translocation of material, and this becomes more likely as charcoal particle size decreases and sediment particle size increases. The importance of this process is site specific as the good agreement between dates on coarse charcoal and organic carbon from the equivalent levels of soil profiles in the Amazon Basin has indicated no large-scale translocation (Pessenda et al., 2001), while other studies have indicated that millimeter-sized charcoal can be redistributed vertically over several meters in coarse sandy sediments (Bird et al., 2002).
- (ii) It is likely that the more refractory components of charcoal, with a longer lifetime in the environment, are transmuted into finer size fractions as charcoal is degraded. Fine particles are more likely to be transported by wind or water over long distances before deposition in the location from which they are ultimately collected. An extreme example of this has been provided by Dickens et al. (2004) who isolated 'graphitic black carbon' from subrecent marine sediments of the West Coast of North America and demonstrated that the material was almost entirely depleted of radiocarbon, meaning that the 'age' of the material bears no relationship to the age of deposition in the sediments.

Substantial advances have been made in the development of techniques for isolation of charcoal for radiocarbon analysis, most of which are underpinned by the development of AMS measurement techniques and hence the ability to analyze milligram quantities of material. This has enabled a substantial increase in the types of material that can be measured and the natural matrices from which charcoal can be isolated for measurement. However, the interpretation of radiocarbon measurements on charcoal still relies fundamentally on a thorough appreciation of the methodological issues and complications that are specific to the purpose for which the

measurements are being taken and that are also often specific to an individual sample. Radiocarbon dating of charcoal is increasingly going to rely on the application of a suite of additional chemical and spectroscopic techniques to check the suitability and purity of each sample prior to analysis (e.g., Alon et al., 2002; Cohen-Ofri et al., 2005; Rebollo et al., 2008), on cross-comparison with results obtained using other sample types (Haesaerts et al., 2010; Nakamura et al., 2010), and on geochronological techniques (e.g., Turney et al., 2001).

See also: Fission-Track Dating; K/Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ Dating. **Luminescence Dating:** Electron Spin Resonance Dating; Thermoluminescence. **Pollen Records, Late Pleistocene:** Northern North America. **Radiocarbon Dating:** ^{14}C of Plant Macrofossils; AMS Radiocarbon Dating; Calibration of the ^{14}C Record; Conventional Method.

References

- Alon D, Mintz G, Cohen I, Weiner S, and Boaretto E (2002) The use of Raman spectroscopy to monitor the removal of humic substances from charcoal: Quality control for ^{14}C dating of charcoal. *Radiocarbon* 44: 1–11.
- Arnold JR and Libby WF (1949) Age Determinations by radiocarbon content: Checks with samples of known age. *Science* 110: 678–680.
- Ascough PL, Bird MI, Francis SM, and Lebl T (2011a) Alkali extraction of archaeological and geological charcoal: Evidence for diagenetic degradation and formation of humic acids. *Journal of Archaeological Science* 38: 69–78.
- Ascough PL, Bird MI, Francis SM, et al. (2011b) Variability in oxidative degradation of charcoal: Influence of production conditions and environmental exposure. *Geochimica et Cosmochimica Acta* 75(9): 2361–2378.
- Ascough PL, Bird MI, Meredith W, et al. (2009) Hydrolysis as a tool for enhanced ^{14}C age measurement and quantification of Black Carbon in depositional environments. *Quaternary Geochronology* 4: 140–147.
- Ascough PL, Bird MI, Meredith W, et al. (2010) Hydrolysis: Implications for radiocarbon pre-treatment and characterization of Black Carbon. *Radiocarbon* 52: 1336–1350.
- Bird MI, Ayliffe LK, Fifield K, et al. (1999a) Radiocarbon dating of 'old' charcoal using a wet oxidation – Stepped combustion procedure. *Radiocarbon* 41: 127–140.
- Bird MI, Charville-Mort PDJ, Ascough PL, Wood R, Higham T, and Apperley D (2010) Assessment of oxygen plasma ashing as a pre-treatment for radiocarbon dating. *Quaternary Geochronology* 5: 435–442.
- Bird MI, Fifield LK, Santos GM, et al. (2003) Radiocarbon dating from 40–60 ka BP at Border Cave South Africa. *Quaternary Science Reviews (Quaternary Geochronology)* 22: 943–947.
- Bird MI and Gröcke D (1997) Determination of the abundance and carbon-isotope composition of elemental carbon in sediments. *Geochimica et Cosmochimica Acta* 61: 3413–3423.
- Bird MI, Moyo E, Veenendaal E, Lloyd JJ, and Frost P (1999b) Stability of elemental carbon in a Savanna soil. *Global Biogeochemical Cycles* 13: 923–932.
- Bird MI, Turney CSM, Fifield LK, et al. (2002) Radiocarbon analysis of the early archaeological site of Nauwalabila I Arnhemland Australia: Implications for sample suitability and stratigraphic integrity. *Quaternary Science Reviews* 21: 1061–1075.
- Braadbaart F, Poole I, and van Brussel AA (2009) Preservation potential of charcoal in alkaline environments: An experimental approach and implications for the archaeological record. *Journal of Archaeological Science* 36: 1672–1679.
- Cachier HP, Bremond MP, and Buat-Menard P (1989) Determination of atmospheric soot carbon by a simple thermal method. *Tellus* 41B: 379–390.
- Chughtai AR, Jassim JA, Peterson JH, Stedman DH, and Smith DM (1991) Spectroscopic and solubility characteristics of oxidized soots. *Aerosol Science and Technology* 15: 112–126.
- Cohen-Ofri I, Weiner L, Boaretto E, Mintz G, and Weiner S (2005) Modern and fossil charcoal: Aspects of structure and diagenesis. *Journal of Archaeological Science* 33(3): 428–439.
- Cope MJ and Chaloner WG (1980) Fossil charcoal as evidence of past atmospheric composition. *Nature* 283: 647–649.

- Czimczik CI, Preston CM, Schmidt MWI, and Schulze E-D (2003) How surface fire in Siberian Scots pine forests affects soil organic carbon in the forest floor: Stocks molecular structure and conversion to black carbon (charcoal). *Global Biogeochemical Cycles* 17: 1024.
- Czimczik CI, Schmidt MWI, and Schulze E-D (2005) Effects of increasing fire frequency on black carbon and organic matter in Podzols of Siberian Scots pine forests. *European Journal of Soil Science* 56: 417–428.
- De Vries HL and Barendsen GW (1954) Measurements of age by the carbon-14 technique. *Nature* 174: 1138.
- Demeyer A, Voundi Nkana JC, and Verloo MG (2001) Characteristics of wood ash and influence on soil properties and nutrient uptake: An overview. *Bioresource Technology* 77: 287–295.
- Dickens AF, Gelinis Y, Masiello CA, Wakeham S, and Hedges JI (2004) Reburial of fossil organic carbon in marine sediments. *Nature* 427: 336–339.
- Douka K, Higham T, and Sinityn A (2010) The influence of pretreatment chemistry on the radiocarbon dating of Campanian Ignimbrite-aged charcoal from Kostenki 14 (Russia). *Quaternary Research* 73: 583–587.
- Eckmeier E, Gerlach R, Skjemstad JO, Ehrmann O, and Schmidt MWI (2007) Only small changes in soil organic carbon and charcoal found one year after experimental slash-and-burn in a temperate deciduous forest. *Biogeosciences Discussions* 4: 595–614.
- Eckmeier E, van der Borg K, Tegtmeier U, Schmidt MWI, and Gerlach R (2009) Dating charred soil organic matter – Comparison of radiocarbon ages from macrocharcoals and chemically separated charcoal carbon. *Radiocarbon* 51: 437–443.
- Elias VO, Simoneit BRT, Cordeiro RC, and Turcq B (2001) Evaluating levoglucosan as an indicator of biomass burning in Carajas Amazonia: A comparison to the charcoal record. *Geochimica et Cosmochimica Acta* 65: 267–272.
- Franklin RE (1951) Crystallite growth in graphitizing and non-graphitizing carbons. *Proceedings of the Royal Society of London: Series A* 209: 196–218.
- Gavin DG (2001) Estimation of inbuilt age of soil charcoal from fire history studies. *Radiocarbon* 43: 27–44.
- Gillespie R (1997) Burnt and unburnt carbon: Dating charcoal and burnt bone from the Willandra Lakes, Australia. *Radiocarbon* 39: 225–236.
- Gillespie R, Hammond AP, Goh KM, et al. (1992) AMS radiocarbon dating of a Late Quaternary tephra site at Graham's Terrace, New Zealand. *Radiocarbon* 34: 21–28.
- Glaser B, Haumaier L, Guggenberger G, and Zech W (1998) Black carbon in soils: The use of benzenecarboxylic acids as specific markers. *Organic Geochemistry* 29: 811–819.
- Gustafsson Ö, Bucheli TD, Kukulska Z, et al. (2001) Evaluation of a protocol for the quantification of black carbon in sediments. *Global Biogeochemical Cycles* 15: 881–890.
- Gustafsson Ö, Kruså M, Zencak Z, et al. (2009) Brown clouds over South Asia: Biomass or fossil fuel combustion? *Science* 323: 495–498.
- Haesaerts P, Borziac I, Chekha VP, et al. (2010) Charcoal and wood remains for radiocarbon dating Upper Pleistocene loess sequences in Eastern Europe and Central Siberia. *Palaeogeography, Palaeoclimatology, Palaeoecology* 291: 106–127.
- Hammes K, Schmidt MWI, Smernik RJ, Currie LA, Ball WP, and Nguyen TH (2007) Comparison of black carbon quantification methods using reference materials from soil, water, sediment and the atmosphere, and implications for the global carbon cycle. *Global Biogeochemical Cycles* 21: <http://dx.doi.org/10.1029/2006BG002914>.
- Hammes K, Torn MS, Lapenas AG, and Schmidt MW (2008) Centennial black carbon turnover observed in a Russian steppe soil. *Biogeosciences* 5: 339–350.
- Hatté C, Morvan J, Noury C, and Paterne M (2001) Is classical acid–alkali–acid treatment responsible for contamination/an alternative proposition. *Radiocarbon* 43: 177–182.
- Hedges REM, Law IA, Bronk CR, and Housley RA (1989) The Oxford accelerator mass spectrometry facility: Technical developments in routine dating. *Archaeometry* 31: 99–113.
- Higham TFG, Barton H, Turney CSM, Barker G, Bronk Ramsey C, and Brock F (2009a) Radiocarbon dating of charcoal from tropical sequences: Results from the Niah Great Cave, Sarawak, and their broader implications. *Journal of Quaternary Science* 24: 189–197.
- Higham T, Brock F, Peresani M, Broglio A, Wood R, and Douka K (2009b) Problems with radiocarbon dating the Middle to Upper Paleolithic transition in Italy. *Quaternary Science Reviews* 28: 1257–1267.
- Hopkins MS, Graham AW, Hewett R, Ash J, and Head J (1990) Evidence of late Pleistocene fires and eucalypt forest from a North Queensland humid tropical rainforest site. *Australian Journal of Ecology* 15: 345–347.
- Lehmann J, Skjemstad J, Sohi S, et al. (2008) Australian climate–carbon cycle feedback reduced by soil black carbon. *Nature Geoscience* 1: 832–835.
- Libby WF (1955) *Radiocarbon Dating*. Chicago: University of Chicago Press.
- Major J, Lehmann J, Rondon M, and Goodale C (2009) Fate of soil-applied black carbon: Downward migration, leaching and soil respiration. *Global Change Biology* 16: 1366–1379.
- Masiello CA (2004) New directions in black carbon organic geochemistry. *Marine Chemistry* 92: 201–203.
- Masiello CA and Druffel ERM (1998) Black carbon in deep-sea sediments. *Science* 280: 1911–1913.
- Nakamura T, Sagawa S, Yamada T, et al. (2010) Radiocarbon dating of charred human bone remains preserved in urns excavated from medieval Buddhist cemetery in Japan. *Nuclear Instruments and Methods in Physics Research B* 268: 985–989.
- Pessenda LCR, Gouveia SEM, and Aravena R (2001) Radiocarbon dating of total soil organic matter and humin fraction and its comparison with ^{14}C ages of charcoal. *Radiocarbon* 43: 595–601.
- Pietikäinen J, Kiikkilä O, and Fritze H (2000) Charcoal as a habitat for microbes and its effect on the microbial community of the underlying humus. *Oikos* 89: 231–242.
- Pigati JS, Quade J, Wilson J, Jull AJT, and Lifton NA (2007) Development of low-background vacuum extraction and graphitization systems for ^{14}C dating of old (40–60 ka) samples. *Quaternary International* 16: 4–14.
- Rebollo NR, Cohen-Ofri I, Popovitz-Biro R, et al. (2008) Structural characterization of charcoal exposed to high and low pH: Implications for ^{14}C sample preparation and charcoal preservation. *Radiocarbon* 50: 289–307.
- Reddy CM, Pearson A, Xu L, et al. (2002) Radiocarbon as a tool to apportion sources of polycyclic aromatic hydrocarbons and black carbon in environmental samples. *Environmental Science and Technology* 36: 1774–1782.
- Schmidt MWI, Skjemstad JO, Czimczik CI, et al. (2001) Comparative analysis of black carbon in soils. *Global Biogeochemical Cycles* 15: 163–168.
- Skjemstad JO, Clarke P, Taylor JA, Oades JM, and McClure SG (1996) The chemistry and nature of protected carbon in soil Australian. *Journal of Soil Research* 34: 251–271.
- Turney CSM, Bird MI, Fifield LK, et al. (2001) Breaking the radiocarbon barrier and early human occupation at Devil's Lair Southwestern Australia. *Quaternary Research* 55: 3–13.
- Wolbach WS and Anders E (1989) Elemental carbon in sediments: Determination and isotopic analysis in the presence of kerogen. *Geochimica et Cosmochimica Acta* 53: 1637–1647.
- Zackrisson O, Nilsson M-C, and Wardle DA (1996) Key ecological function of charcoal from wildfire in the boreal forest. *Oikos* 77: 10–19.

¹⁴C of Plant Macrofossils

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Glossary

Aging or rejuvenation: Shift in apparent radiocarbon age caused by an anomalously low (aging) or high (rejuvenation) ¹⁴C content of the original carbon. One of the fundamental assumptions underlying radiocarbon dating is that the material being dated was originally in equilibrium with the contemporary atmosphere and drew its carbon (and hence, its ¹⁴C) directly from the atmospheric 'reservoir.' However, if the material was formed in an environment which had a different ¹⁴C content to the atmosphere, then its radiocarbon age will be higher (aging) or lower (rejuvenation) than that of contemporary wood, depending on whether its carbon reservoir contained less (see 'hard water effect') or more (see 'bomb carbon') ¹⁴C than the atmosphere.

Bomb carbon: This expression refers to the significant quantity of ¹⁴C that was injected into the atmosphere by nuclear weapons testing between 1945 and the mid-1970s. The increased activity reached its peak around 1963; atmospheric ¹⁴C reached twice the normal level in the Northern Hemisphere and increased by about 60% in the Southern Hemisphere. It has since decreased to about 10% above 'prebomb' levels, due mainly to the exchange of CO₂ between the atmosphere and the oceans.

Hard water effect: The 'hard water effect' refers to the situation in which the radiocarbon age of fresh water (or organisms living in the water) in contact with limestone (calcium carbonate) can be increased by the dissolved carbonate in the water. As the limestone contains no ¹⁴C, the effect is to dilute the concentration of the ¹⁴C in the incoming CO₂.

Introduction

Until the late 1980s, the majority of dates for lacustrine or palustrine paleoclimatic studies were obtained by beta-counting of samples from organic lake muds or bulk peat sediment. However, since then, the preferred strategy is to date plant macrofossil material by accelerator mass spectrometry (AMS) (see [AMS Radiocarbon Dating](#)). This is partly because AMS offers the possibility of dating a much higher level of stratigraphic resolution (since it is possible to obtain dates on very small samples of material) and partly because of the widely held view that AMS dates on plant macrofossils are inherently more reliable than those obtained from the sediment matrix. In this case, the source of the carbon is known and therefore should not be composed of heterogeneous material that could be of different ages. In general, no account has been taken from plant macrofossils that may be species-dependent or that the plant macrofossils themselves may be redeposited. Consequently, inconsistencies even within the plant macrofossils chronologies are still common. This suggests that the choice of the appropriate material is not trivial and requires serious investigations.

Many studies on terrestrial paleoclimatic records aim to address the Last Glacial–Holocene transition. The characterization of this transition is mostly based on lacustrine or palustrine records. Establishment of a reliable and accurate chronology remains one of the major challenges in such an environment and time-slice. Indeed, because of the variations in the amount of ¹⁴C in atmospheric CO₂, this means that the measured amount of ¹⁴C is not linearly related to age, and the aging is highly critical.

We consider here plant macrofossils accounting for the context from which they are extracted. We then go deeper in the difficulties, from the archaeological context to aquatic

environment. Finally, we give some reminders about sampling, storage, and pretreatments of samples, and inherent possible contaminations.

Archaeological Contexts

In Quaternary studies, archaeological contexts are also an important area where plant macrofossil studies are of great importance. In archaeology, establishment of human evolution chronology is a key issue. Dating support has to be related without ambiguity to human life, industry, or art. Charcoal in a foyer can be linked to a human fire but not directly to the cave paintings drawn just above on the cave walls. A small amount of charcoal on the painting itself would give the date of the work. Besides charcoal, olive pits as food residues, plant leaf as basket component, and wood as timber are dedicated plant support for the ¹⁴C dating of human evolution. Nevertheless, even when directly related to human activity, ¹⁴C geochronology is not without risk.

'Old Wood' Effect

Quaternary scientists have long recognized the problem of the 'old wood' effect in radiocarbon-dating charcoal and wood samples, the age of which may be hundreds of years older than expected. This is particularly true for long-lived tree species such as oak and juniper. Every year, tree growth adds woods to the outside part of the trunk. The date measured on heartwood may be already many centuries old by the time the tree was cut down. An ideal material to date, if available, would be twigs of such trees as hazel or wicker willow that can

integrate 5 years at the worst. Another difficulty is that of possible time-lags between felling and final deposition. The timber may have had an extensive history of use and reuse. The date we are measuring is the death of the tree and not its last use.

Volcanic Area

Volcanic areas may be subjected to atmospheric CO_2 dilution by magmatic CO_2 . Since, the magmatic CO_2 is ^{14}C -free, this results of a ^{14}C dilution in plants of the surrounding areas (Cook et al., 2001; Pasquier-Cardin et al., 1999). A dilution by more than 85% has been observed in Mammoth Mountain, CA, in 1997. ^{14}C activity measurements can then be used as a way to track changes in volcanic activity. However, plant macrofossils inherit this dilution leading to an important artificial radiocarbon aging which can be as small as 200 years or up to 20 000 years. As volcanic activity is highly variable, it is thus really difficult to evaluate the magnitude of the aging. Using coupled $\delta^{13}\text{C}$ – ^{14}C measurements, Pasquier-Cardin et al. (1999) showed the way to evaluate the dilution ratio for volcanoes where clear ^{13}C enrichment in plant has been evidenced in Azores. Garnett and Billett (2007) follow a similar approach in Tyrrhenian Sea.

Tephra records can be also a partial solution to the dating problem facing peat–climate studies. The techniques of wiggle-matched radiocarbon dates and tephrochronology have been used in combination to obtain accurate and precise age estimates for prehistoric tephra and hence for peat layers (Molisso et al., 2010; Pilcher et al., 1996).

Soil and (Dry) Sediment

Scientists face new supplementary complications when drilling or coring for paleoclimate studies into soil or sediment. Indeed at the opposite of archaeological sites where human occupation layers are relatively well defined, no artifacts are present in the soil or sediment to fix the limits of the different units. Major units are often well described by color or major texture changes but within these units, laminated smaller layers are very rare. Few proofs of *in situ* fossilization without any further reworking are present. In nonaqueous and well-oxygenated environments, very rare large macrorests subsist to diagenesis. The tender part of the leaf, twig, and trunk rapidly disappears through bacterial activity. Only resistant parts of the superior plant can be found some thousands years later. This is the case of phytoliths (small silicate bodies produced in abundance in the tissue of most of the grasses), pollen and, on occasion, of seeds.

Bioturbation and Mechanic Disturbance

Whatever the sediment, action of roots, animal burrows, water and fine grain illuviation can distort the original record and even upward and downward movement might displace the target macrorests. The largest uncertainty with the macrofossil sample is ‘is the macrorest in place? or did it move upward or downward?’ In some lucky cases, some fragile seeds might

resist diagenesis and can be used as a reliable support for ^{14}C dating. In a moving environment, the fragile part of the macrorest is destroyed. Its presence proves the *in situ* fossilization. This is the case of birch seeds surrounded by a fragile sheath which presence attests that the macrorest was not subjected to a displacement posterior to its fossilization.

Phytoliths

Phytoliths are increasingly used for isotopic studies performed on organic matter linked to the silicate skeleton building (Hodson et al., 2008; Kelly et al., 1991, 1998; Smith and Anderson, 2001; Smith and White, 2004). Phytoliths trap trace amounts of organic carbon (0.09–1.3 wt%) that is encased in silica (Mulholland and Prior, 1993). The very small amount of carbon now required for a ^{14}C measurement can be used as ^{14}C dating support if it is associated with the highest control of the blank quality (Santos et al., 2010).

Nevertheless, their shape (often needles) and their high resistance toward diagenesis is not an advantage to check the *in situ* position without reworking of the chosen phytoliths in the section. This can be really problematic when a forest has replaced the savanna, the former having an extended and strong rhizosphere.

Pollen

Because many continental paleoclimatic studies are based on pollen assemblages, it appears judicious to directly use pollen grains in support of ^{14}C dating. However, very few ^{14}C datings are performed on pollen grains themselves (Long et al., 1992; Mensing and Southon, 1999), most are performed on pollen concentrate from sediment (Brown et al., 1989, 1992; Kilian et al., 2002; Piotrowska et al., 2004; Zhou et al., 1997), and the results are highly variable.

Based on pollen concentrates, Zhou et al. (1997) succeeded in obtaining a consistent chronology in paleosols and peat from the loess/desert transitional belt in China. However, pollen concentrates from loess or sand samples appear to provide ages that are either too old or too young. Zhou et al. (1997) observe that on the one hand, pollen are badly preserved in dry environments like sands, and on the other hand, reworked pollen are included in pollen concentrates. This reworked pollen can be either older (fossilized) spore pollen that could be transported with dust prior to sediment deposition or younger pollen transported through hydrological or agricultural activities. In Lake Baikal, Piotrowska et al. (2004) succeeded in establishing a good chronological framework based on concentrate pollen carried out using chemical treatment, microsieving, and heavy liquid separation.

At Lake Goszcz (Poland), Kilian et al. (2002) did not succeed in establishing a coherent temporal framework based on concentrate pollen ^{14}C dating, whereas ^{14}C dating on superior plant macrofossils (Goslar et al., 1995) provided reliable chronology. Indeed, all pollen concentrate samples showed an aging that was caused by the presence of the selected fractions of some resistant phytoplankton affected by the ‘hard water effect.’

Even with manually separated pollens, some problems remain. Indeed, this precise and laborious method (5000 of

spruce or 10000 pollen grains of pine to obtain 70 μg of carbon) does not exempt the sample from the possibility that some of the pollen grains can be reworked. Mensing and Southon (1999) demonstrate incorporation of older pollen grains in lacustrine sediment due to erosion from near-shore sediments due to Lake Moran (California) level fluctuation. They arrived at the same conclusion for marine sediments of the Santa Barbara Basin. This example of old pollen incorporation may be extended to most of lacustrine or palustrine environments linked to moraines and continental ice sheet melting (e.g., Great Lakes, North European lakes, Russia), where water inflow drains old sediments that may contain very old pollen grains.

Lacustrine and Palustrine Records

Many studies on terrestrial paleoclimatic records aim to address the Last Glacial-Holocene transition. The characterization of this transition is mostly based on lacustrine or palustrine records. Establishment of a reliable and accurate chronology remains one of the major challenges in such an environment and time-slice. Indeed, as the variations in the amount of ^{14}C in atmospheric CO_2 , means that the measured amount of ^{14}C is not linearly related to age, the aging is highly critical. This article now turns its focus on plant macrofossils available in these humid environments, from algae to superior plant macrofossils.

Algae and Aquatic Moss Macrofossils

Aquatic cells photosynthesize subaquatically and hence build carbon from dissolved inorganic carbon (DIC) into their cellular material, so they reflect the ^{14}C : ^{12}C ratios of the water from which they grow. The DIC is influenced by (1) exchange with the atmospheric CO_2 reservoir, (2) decomposition of organic matter, (3) dissolved carbonate from surrounding

limestone catchments, and (4) residence time of lake or peat-bog water. This means that the ^{14}C activity of DIC does not reflect the ^{14}C activity of atmosphere, but is ^{14}C -depleted. This results in an artificial aging, the so-called 'hard water effect' inherited by algae cells and which can show wide variation (Fontana, 2005; Macdonald et al., 1991). See Figure 1 on bulk organic matter.

Consequently, algal macrofossils should be avoided if possible. If there is no other choice, an evaluation of the reservoir age should be associated with the measurements. This could be done by the collection, in a 'reference' level, of algae macrofossils and subaerial plants or beetles macrofossils. But once again, attention must be brought to the choice of the reference level and to the beetle species. Indeed, the level chosen to determine the reservoir age should correspond to an equivalent climatic context to the one to which the correction will be done. Likewise, the ecological niche of beetle species has to be known in order to avoid beetles that feed with algae or that burrow.

There are some examples of peat bogs that do not seem to be subjected to the 'hard water effect,' such as the case of lacustrine or palustrine environments developed on glacial till (Blaauw et al., 2004; Walker, 2001). Nevertheless, such occurrences are not common.

Terrestrial Plant Macrofossils

Based on the assumption that superior (terrestrial) plants are less susceptible to contamination by inert carbon than their aquatic counter parts, since they obtain their carbon by subaerial photosynthesis, many scientists have advocated the exclusive use of superior plant macrofossils for radiocarbon-dating lacustrine and palustrine sediment sequences. The development of the AMS radiocarbon-dating technique has facilitated the use of these plant macrofossils as a routine material for dating, as a reasonable analytical precision in radiocarbon measurement

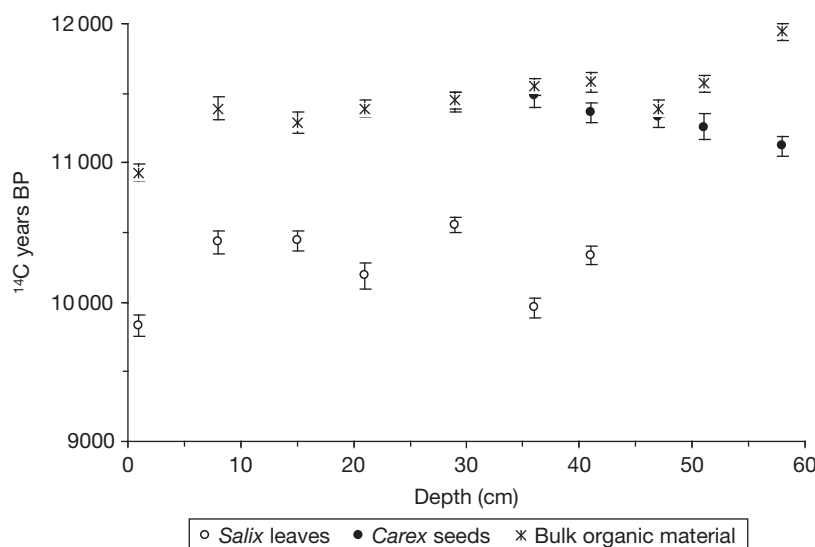


Figure 1 Radiocarbon ages obtained from macrofossil and bulk organic matter samples from Finglas River, southwest Ireland (modified from Turney et al., 2000). Notice the systematic aging of radiocarbon ages obtained on bulk organic material compared to *Salix* leaves and the age difference between *Carex* seeds and *Salix* leaves.

can usually be obtained from as little as 2 mg of organic matter. This amount can typically be extracted from just a few fossil seeds or bracts, or even from a single fossil leaf.

Potamogeton Seeds and Other Floating Superior Plant Macrofossils

Floating superior plants photosynthesize from the CO_2 available at the water surface that mostly results from a mixture of free atmospheric CO_2 and degassing DIC. Indeed, a high content of mineral carbon in water can result from strong limestone catchment erosion or from high organic matter mineralization, such as is commonly observed following a production bloom that may result in DIC degassing. Consequently, floating superior plants inherit part of the DIC ^{12}C : ^{14}C composition and even if diluted, the 'hard water effect' affects the ^{14}C activity of *Potamogeton* seeds. The conversion of this activity into a date would induce significant aging (i.e., the sample appears too old) or rejuvenation (too young) in the case of modern environments with high organic matter mineralization. Indeed, the DIC may thus be imprinted by peak bomb organic carbon and then the floating plants appear ^{14}C -enriched (Olsson and Kaup, 2001).

There are some examples of peat bogs where DIC degassing is not significant, and dating obtained on *Potamogeton* seeds gives a reliable chronological framework (Walker et al., 2001). Nevertheless, since this is linked to water organic production and thus to seasonal cycle, the degassing can fluctuate greatly and therefore, a quantitative evaluation is difficult. It is hazardous to work on this kind of macrofossil. Therefore, it is better to avoid the use of floating plant macrofossils for dating as much as possible.

Sphagnum, Terrestrial Mosses, and Smaller-Sized Superior Plant Macrofossils

As ombotrophic peat deposits are very common and sphagnum-formed, sphagnum macrofossils are often used for ^{14}C dating (Kilian et al., 1995). These superior plants do not grow in the water context and are not subjected to a 'hard water effect' or DIC degassing influence; however, they are not free of aging effects due to the 'hard water effect'! Indeed, palustrine areas may be highly organic, especially in ombotrophic bogs, and then very efficient environments for organic matter mineralization. As this mineralization concerns recent as well as old organic matter, the resulting mineralized CO_2 emitted by the 'soil' shows an old apparent ^{14}C -age. Thus, at the ground surface, CO_2 results in a mixture between free atmospheric CO_2 and mineralized organic matter CO_2 . At a lower scale, this effect can be linked to the well-known canopy effect in a tropical forest. Close to the surface, the smaller-sized superior plants like sphagnum and mosses may thus show ^{14}C aging (Jungner et al., 1995). Some mosses are calciphilous and thus may incorporate limestone as a carbon source, this effect may result in an aging (Zazula et al., 2006). These mosses may not constitute the most reliable support to establish a chronology.

Considering the uncertainty ranges associated with ^{14}C dating, this aging can sometimes not be significant, and chronologies based on ^{14}C on sphagnum can be reliable. But with

the increasing precision of ^{14}C dating, the aging risk linked to organic matter mineralization has to be taken into account.

Larger-Sized Superior Plant Macrofossils

Finally, larger superior plant macrofossils seem to be the most suitable macrofossils for ^{14}C dating. A plethora of such macrofossils can be found in peat and lake sediment. We can list needles of *Pinus*, *Larix*, *Picea*, seeds of *Betula*, *Carex*, *Eleocharis*, leaves of *Salix*, *Dryas*, *Sphagnum*, *Erica*, *Calluna* as the most common (Figure 2).

On account of their high resistance to diagenesis, *Carex* seeds were intensively used in support of chronological frameworks. However, this effective resistance does not permit verification of the state of preservation and then the avoidance of some possibly reworked or redeposited seeds that would be recognized from indices on macrofossil shape and entirety. This was the origin of some inconsistent chronologies based on macrofossils of several centimeter tall superior plants (Turney et al., 2000). We would recommend the collection of 'fragile' macrofossils from which it is easy to check that they are unlikely to have been physically transferred within the sediment matrix and/or otherwise reworked from the surrounding catchment and/or reworked by fauna and/or the rhizosphere. For example, intact *Betula* seeds surrounded by a fragile voile or survival of the fragile *Salix* leaves can be an ideal candidate for ^{14}C dating to establish a chronological framework.

Toward Specific Compounds

The dating of pleniglacial sediments remains problematic due to the very low content of identifiable plant macrofossils. The development of AMS technology now allows ^{14}C measurements on extremely low amounts of carbon (as little as 10 μg , potentially even less) and offers the new prospect of specific-compound dating. In this way, we focus on organism-specific organic compounds called 'biomarkers' or 'specific compounds.' This technique was mainly used in marine environments (from Eglinton et al., 1996) but is more and more common on continental paleoclimatic studies (Hou et al., 2010; Smittenberg et al., 2004; Uchikawa et al., 2008).

Two ways of sample preparation coexist: elimination of all nonselected compounds and extraction of the selected compound from a complex mixture by chromatography. The first one remains time-consuming and fastidious but can easily be done in all laboratories not fully equipped with chromatography whereas the second is more direct but requires to be equipped in preparative gas chromatography (Prep-GC), preparative liquid chromatography (Prep-LC), and/or preparative high pressure liquid chromatography (Prep-HPLC). Chromatographs are either associated with a fraction collector for offline physical measurement (after conversion into CO_2 then graphite) or coupled to a gas source AMS for online measurements.

As has been done for macrorest deposits, the selection of the targeted biomarker has to account for the physiology and the way of life of the organism from which it is extracted. Sillafin (diatom-specific protein) extraction is thus done to establish a chronological framework on paleoclimatic records

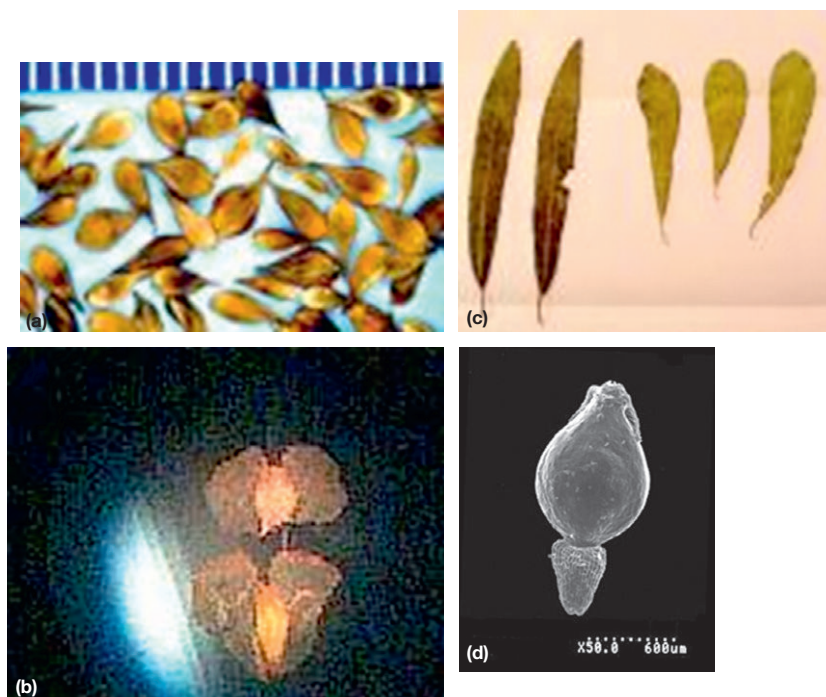


Figure 2 Pictures of some common macrofossils. (a) *Carex* seeds, (b) *Betula* seeds, (c) *Salix* leaves (from <http://tchester.org>), (d) *Eleocharis* seed (from <http://geography.berkeley.edu>).

deriving from diatom assemblages or isotopic signatures (Hatté et al., 2008; Ingalls et al., 2004), but this material does not solve the problem of the hard-water effect. Also, *n*-alkanes from the degradation of leaf wax (Collister et al., 1994) may now be sampled and analyzed by Prep-GC (Uchikawa et al., 2008) to establish lacustrine chronology. Only higher plants have wax on their leaves. This would be one way to avoid any hard-water effect linked to algae, but it does not avoid the reworking risk.

Besides chronological issues, specific-compound radiocarbon dating is used to elucidate the effect of different transport mechanisms, particle association, and chemical reactivity between different carbon reservoir and thus to address key issues of modern carbon cycle understanding (Hou et al., 2010; Huang et al., 1999; Mollenhauer et al., 2009; Pearson and Eglinton, 2000). This is done by comparing radiocarbon ages of lignin phenols and co-occurring plant-wax lipids like long-chain *n*-alkanes, *n*-alcohols, and *n*-fatty acids. Likewise, specific-compound radiocarbon dating allows detection of potential fossil-fuel-derived contamination (Lichtfouse et al., 1995).

Reminders and Suggestions

Storage and Macrofossils Preparation

The core storage and the preparation and identification of small macrofossil samples are usually carried out in a nonsterile environment. Wohlfarth et al. (1998) has shown that the long-term storage of wet macrofossil samples appears to have a significant effect on the radiocarbon age obtained, even when the samples are kept cool. Fungal spores and microorganisms

may easily be incorporated into the sample and can easily contaminate the sample if stored cool and wet for several months. Indeed, fungal metabolism takes up CO_2 from the fossil leaf material but also includes carbon from contaminating nutrients in the ambient water during growth. Wohlfarth et al. (1998) observe that even when stored in slightly acidic water (pH 2) and under cool and dark conditions, *Paecilomyces farinosus* may easily grow and absorb CO_2 from the remaining air in the glass bottle and from the ambient water. This recent carbon, when added to the sample, may account for the large errors of several hundred to several thousand years encountered in dating series. They refer to radiocarbon ages that are $\sim 400\text{--}700$ ^{14}C years too young in samples of around 12000 ^{14}C years BP, depending on whether the macrofossils were dried at 50°C or whether they were stored for 2 years at room temperature in deionized water. Earlier investigations on bulk sediments from marine and lacustrine cores, which had been stored for many years at a temperature of 4°C , also showed unreliable radiocarbon dates due to contamination by recent carbon from modern, terrestrial bacteria (Colman et al., 1996; Geyh et al., 1974). Similar studies designed to constrain the impact of storage and chemical treatment on isotopic composition reveal that inadequate storage conditions might result in a 2‰ $\delta^{13}\text{C}$ distortion on low organic carbon sediment (Gauthier and Hatté, 2008).

Small Samples

Even though the AMS dating method can provide results with acceptable accuracy from small amounts of carbon, a minimum amount of material is still required. If the samples are too small, they may yield very little datable carbon after

pretreatment and can be more easily biased by contaminants, in addition to having larger standard deviations. Processing sample under continuous flow of inert gas (He or N₂) can be considered as an alternative to the common vacuum lines. Working under gas pressure decreases the risk of modern carbon contamination by incorporation of atmospheric CO₂. This method does not require so many changes in the technological suite. Likewise, but harder to implement and to process, chemical treatments on fragile samples might be treated in a gloves box with a clean neutral gas atmosphere.

Chemical Pretreatments

The most common chemical pretreatment used for plant macrofossils consists of three steps, as follows:

1. acid leaching to remove any carbonate and any 'fulvic' acids,
2. alkali treatment to remove 'humic' acids, and
3. acid rinsing to remove modern carbon potentially precipitated during the alkali step.

A cellulose extraction can be performed on large wood samples (Leavitt and Danzer, 1993). Sometimes, but fortunately rarely, contaminants can be incorporated during the chemical pretreatment.

The alkali step may induce atmospheric carbon incorporation into the organic structure or adsorption of volatile organic molecules on cell walls that cannot always be removed by the classical method using a subsequent acid step (Hatté et al., 2001; Kilian et al., 2002). Some laboratory solutions have been proposed that are based on an additional oxidation step (Bird et al., 1999; Hatté et al., 2001) to discard most of the modern carbon incorporated during the alkali step. An alternative would be to perform the chemical pretreatments under an inert atmosphere. However, no chemical pretreatment can correct for fungal and microbial contamination linked to the core and macrofossil storage.

Conclusion

The aim of this article was to give a general view of plant macrofossils founded in Quaternary sediments or archaeological settlements and to discuss their suitability and reliability for ¹⁴C-AMS dating.

Three conclusions can be made:

- Long-term and cool storage of cores and wet macrofossil samples prior to the radiocarbon measurement has significant consequences on the radiocarbon age obtained.
- A chronology based on a type of macrofossil appearing as coherent in one specific environment does not necessarily mean that this macrofossil is the most reliable support for ¹⁴C dating. Nothing is universal!!
- A good choice of a macrofossil for ¹⁴C dating several 10 years ago does not mean that it is still the best choice today. Increasing precision of ¹⁴C dating means that a serious consideration of the meaning of the choice of macrofossil and its relationship with the event to be characterized must be made.

Briefly, ¹⁴C on well-chosen plant macrofossils constitutes a powerful tool to establish chronological frameworks in terrestrial environments.

See also: Dating Techniques; Plant Macrofossil Introduction.
Radiocarbon Dating: AMS Radiocarbon Dating.

References

- Bird MI, Ayliffe LK, Fifield LK, et al. (1999) Radiocarbon dating of "old" charcoal using a wet oxidation, stepped-combustion procedure. *Radiocarbon* 41: 127–140.
- Blaauw M, van der Plicht J, and van Geel B (2004) Radiocarbon dating of bulk peat samples from raised bogs: Non-existence of a previously reported "reservoir effect"? *Quaternary Science Reviews* 23: 1537–1542.
- Brown TA, Farwell GW, Grootes PM, and Schmidt FH (1992) Radiocarbon AMS dating of pollen extracted from peat samples. *Radiocarbon* 34: 550–556.
- Brown TA, Nelson DE, Mathewes RW, Vogel JS, and Southon JR (1989) Radiocarbon dating of pollen by accelerator mass spectrometry. *Quaternary Research* 32: 205–212.
- Collister JW, Rieley G, Stern B, Eglinton G, and Fry B (1994) Compound specific ¹³C analyses of leaf lipids from plants with differing carbon dioxide metabolism. *Organic Chemistry* 21: 619–628.
- Colman SM, Jones GA, Rubin M, King JW, Peck JA, and Orem WH (1996) AMS radiocarbon analyses from Lake Baikal, Siberia: Challenges of dating sediments from a large, oligotrophic lake. *Quaternary Science Reviews* 15: 669–684.
- Cook AC, Hainsworth LJ, Sorey ML, Evans WC, and Southon JR (2001) Radiocarbon studies of plant leaves and tree rings from Mammoth Mountain, CA: A long-term record of magmatic CO₂ release. *Chemical Geology (Isotope Geoscience Section)* 177: 117–131.
- Eglinton TI, Aluwihare LI, Bauer JEE, Druffel ERM, and McNihol AP (1996) Gas chromatographic isolation of individual compounds from complex matrices for radiocarbon dating. *Analytical Chemistry* 68: 904–912.
- Fontana SL (2005) Holocene vegetation history and palaeoenvironmental conditions on the temperate Atlantic coast of Argentina, as inferred from multi-proxy lacustrine records. *Journal of Paleolimnology* 34: 445–469.
- Garnett M and Billett M (2007) Do riparian plant fix CO₂ lost by evasion from surface waters? An investigation using carbon isotopes. *Radiocarbon* 49: 993–1001.
- Gauthier C and Hatté C (2008) Effects of handling, storage, and chemical treatments on delta C-13 values of terrestrial fossil organic matter. *Geochemistry, Geophysics, Geosystems* 9: Q08011. <http://dx.doi.org/10.1029/2008GC001967>.
- Geyh MA, Krumhein WE, and Kudrass HR (1974) Unreliable ¹⁴C dating of long-stored deep-sea sediment due to bacterial activity. *Marine Geology* 17: M45–M50.
- Goslar T, Arnold M, Bard E, et al. (1995) High concentration of atmospheric C-14 during the Younger Dryas cold episode. *Nature* 377: 414–417.
- Hatté C, Hodgins G, Jull AJT, Bishop B, and Tesson B (2008) Marine chronology based on ¹⁴C dating on diatoms proteins. *Marine Chemistry* 109: 143–151.
- Hatté C, Morvan J, Noury C, and Paternie M (2001) Is classical acid-alkali-acid treatment responsible for contamination? An alternative proposition. *Radiocarbon* 43: 177–182.
- Hodson MJ, Parker AG, Leng MJ, and Sloane HJ (2008) Silicon, oxygen and carbon isotope composition of wheat (*Triticum aestivum* L.) phytoliths: Implications for palaeoecology and archaeology. *Journal of Quaternary Science* 23: 331–339.
- Hou J, Huang Y, Brodsky C, et al. (2010) Radiocarbon dating of individual lignin phenols: A new approach for establishing chronology of Late Quaternary Lake Sediments. *Analytical Chemistry* 82: 7119–7126.
- Huang Y, Li B, Bryant C, Bol R, and Eglinton G (1999) Radiocarbon dating of aliphatic hydrocarbons. A new approach for dating passive-fraction carbon in soil horizons. *Soil Science Society of America Journal* 63: 1181–1187.
- Ingalls A, Anderson RF, and Pearson A (2004) Radiocarbon dating of diatom-bound organic compounds. *Marine Chemistry* 92: 91–105.
- Jungner H, Sonninen E, Possnert G, and Tolonen K (1995) Use of bomb-produced ¹⁴C to evaluate the amount of CO₂ emanating from two peat-bogs in Finland. *Radiocarbon* 37: 567–573.
- Kelly EF, Amundson R, Marino BD, and DeNiro MJ (1991) Stable isotope ratios of carbon in phytoliths as a quantitative method of monitoring vegetation and climate change. *Quaternary Research* 35: 222–233.

- Kelly EF, Blecker SW, Yonker CM, Olson CG, Wohl EE, and Todd LC (1998) Stable isotope composition of soil organic matter and phytoliths as paleoenvironmental indicators. *Geoderma* 82: 59–81.
- Kilian MR, Van der Plicht J, and van Geel B (1995) Dating raised bogs: New aspects of AMS ¹⁴C wiggle matching, a reservoir effect and climatic change. *Quaternary Science Reviews* 14: 959–966.
- Kilian MR, Van der Plicht J, van Geel B, and Goslar T (2002) Problematic ¹⁴C-AMS dates of pollen concentrates from Lake Gosciarz (Poland). *Quaternary International* 88: 21–26.
- Leavitt SW and Danzer SR (1993) Method for batch processing small wood samples to holocellulose for stable carbon isotope analysis. *Analytical Chemistry* 65: 87–89.
- Lichtfouse E, Berthier G, Houot S, Barriuso E, Bergheud V, and Vallaeys T (1995) Stable carbon isotope evidence for the microbial origin of C14–C18 n-alkanoic acids in soils. *Organic Geochemistry* 23: 849–852.
- Long A, Davis OK, and Delanois J (1992) Separation and C-14 dating of pure pollen from lake-sediments – nanofossil AMS dating. *Radiocarbon* 34: 557–560.
- MacDonald GM, Beukens RP, and Kieser WE (1991) Radiocarbon dating of limnic sediments: A comparative analysis and discussion. *Ecology* 72: 1150–1155.
- Mensing S and Southon JR (1999) A simple method to separate pollen for AMS radiocarbon dating and its application to lacustrine and marine sediments. *Radiocarbon* 41: 1.
- Molisso F, Insinga D, Marzaioli F, Sacchi M, and Lubritto C (2010) Radiocarbon dating versus volcanic event stratigraphy: Age modelling of Quaternary marine sequence in the coastal region of the Eastern Tyrrhenian Sea. *Nuclear Instruments and Methods in Physics Research Section B: Beam Interactions with Materials and Atoms* 268: 1236–1240.
- Mollenhauer G, Huang Y, Kreutz R, Kusch S, Rethemeyer J, and Schefuss E (2009) Riverine export of terrestrial organic carbon investigated by radiocarbon dating of lignin phenols in river-influenced marine core-top sediments. In: *Union AG (ed.) American Geophysical Union, Fall Meeting 2009*, San Francisco.
- Mulholland SC and Prior CA (1993) AMS radiocarbon dating of phytolith. In: Pearsall DM (ed.) *Current Research in Phytolith Analysis: Applications in Archeology and Paleocology*, MASCA: Research Papers in Science and Archeology University of Pennsylvania Museum.
- Olsson IU and Kaup E (2001) The varying radiocarbon activity of some recent submerged estonian plants grown in the early 1990s. *Radiocarbon* 43: 809–820.
- Pasquier-Cardin A, Allard P, Ferreira T, et al. (1999) Magma-derived CO₂ emissions recorded in ¹⁴C and ¹³C content of plants growing in Furnas Caldera, Azores. *Journal of Volcanology and Geothermal Research* 92: 195–207.
- Pearson A and Eglinton TI (2000) The origin of n-alkanes in Santa Monica Basin surface sediment: A model based on compound-specific $\Delta^{14}\text{C}$ and $\delta^{13}\text{C}$ data. *Organic Geochemistry* 31: 1103–1116.
- Pilcher JR, Hall VA, and McCormac FG (1996) An outline tephrochronology for the Holocen of the north Ireland. *Journal of Quaternary Science* 11: 485–494.
- Piotrowska N, Bluszcz A, Demske D, Granoszewski W, and Heumann G (2004) Extraction and AMS radiocarbon dating of pollen from Lake Baikal sediments. *Radiocarbon* 46: 181–187.
- Santos GM, Alexandre A, Coe HHG, Reyerson PE, Southon JR, and De Carvalho CN (2010) The phytolith ¹⁴C puzzle: A tale of background determinations and accuracy tests. *Radiocarbon* 52: 113–128.
- Smith FA and Anderson KB (2001) Characterization of organic compounds in phytoliths: Improving the resolving power of phytoliths ¹³C as a tool of paleoecological reconstruction of C3 and C4 grasses. In: Meunier J-D (ed.) *Phytoliths: Applications in Earth Science and Human History*. Rotterdam: Balkema.
- Smith FA and White JWC (2004) Modern calibration of phytolith carbon isotope signatures for C3/C4 paleograssland reconstruction. *Palaeogeography, Palaeoclimatology, Palaeoecology* 207: 277–304.
- Smittenberg RH, Hopmans EC, Schouten S, Hayes JM, Eglinton TI, and Sinninghe Damsté JS (2004) Compound specific radiocarbon dating of the varved Holocene sedimentary record of Saanich Inlet, Canada. *Paleoceanography* 19: <http://dx.doi.org/10.1029/2003PA000927>.
- Turney CSM, Coope GR, Harkness DD, Lowe JJ, and Walker DA (2000) Implications for the dating of Wisconsinan (Weichselian) Late-Glacial events of systematic radiocarbon age differences between plant macrofossils from a site in SW Ireland. *Quaternary Research* 53: 114–121.
- Uchikawa J, Popp BN, Schoonmaker JE, and Xu L (2008) Direct application of compound specific radiocarbon analysis of leaf waxes to establish lacustrine sediment chronology. *Journal of Paleolimnology* 39: 43–60.
- Walker MJC (2001) Rapid climate change during the last glacial-interglacial transition; implications for the stratigraphic subdivision, correlation and dating. *Global and Planetary Change* 30: 59–72.
- Walker MJC, Bryant CL, Coope GR, Harkness DD, Lowe JJ, and Scott EM (2001) Towards a radiocarbon chronology of the late-glacial: Sample selection strategies. *Radiocarbon* 43: 1007–1019.
- Wohlfarth B, Skog G, Possnert G, and Holmqvist BH (1998) Pitfalls in the AMS radiocarbon-dating of terrestrial macrofossils. *Journal of Quaternary Science* 13: 137–145.
- Zazula GD, Schweger CE, Beaudouin AB, and McCourt GH (2006) Macrofossil and pollen evidence for full-glacial steppe within an ecological mosaic along the Bluefish River, eastern Beringia. *Quaternary International* 142–143: 2–19.
- Zhou W, Donahue DJ, and Jull AJT (1997) Radiocarbon AMS dating of pollen concentrated from eolian sediments: Implications for monsoon climate change since the late Quaternary. *Radiocarbon* 39: 19–26.

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SEA LEVEL STUDIES

Contents

Overview

Geomorphological Indicators

Sedimentary Indicators of Relative Sea-Level Changes – High Energy

Sedimentary Indicators of Relative Sea-Level Changes – Low Energy

Coral Records of Relative Sea-Level Changes

Microfossil-Based Reconstructions of Holocene Relative Sea-Level Change

Eustatic Sea-Level Changes – Glacial–Interglacial Cycles

Eustatic Sea-Level Changes Since the Last Glacial Maximum

Isostasy: Glaciation-Induced Sea-Level Change

Use of Cave Data in Sea-Level Reconstructions

Overview

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Introduction

Understanding relative sea-level changes helps explain critical interactions in Earth environmental systems throughout the Quaternary. Relative sea-level changes record transfers of mass between oceans and continents during expansion and contraction of great ice sheets driven by climate change. They record extreme events, including storm surges and tsunamis, as well as predictable daily tides. They also record vertical movements in the Earth's crust over a wide range of timescales, from coseismic uplift and subsidence during great, $M_w > 8$, plate-boundary earthquakes, through century- and millennial-scale movements driven by the process of glacial isostatic adjustment (GIA), to timescales beyond the Quaternary, as global ocean volumes change as a result of plate-tectonic movements. Vertical changes

in relative sea level result in major horizontal shifts in coastal environments. The combination of all these factors means that no single location on Earth records all of these changes. No location can be considered stable and therefore a record of 'absolute' sea-level change over different time-scales. Perhaps the most significant advance in studies of Quaternary sea-level change is the recognition that key questions will only be answered by the integration of multiproxy-based reconstructions of sea level with model predictions.

For the timescale of the whole Quaternary this can be simply illustrated by the example of oxygen isotope records from ocean sediments used to reconstruct global eustatic sea-level change through the Quaternary. Oxygen isotope variations, $\delta^{18}\text{O}$, in Foraminifera preserved in deep ocean sediments are a commonly cited proxy for developing continuous reconstructions

of ice volume and sea level over late Quaternary time and earlier. But direct conversion of $\delta^{18}\text{O}$ changes into ice volume and sea level has two critical problems (Clark and Mix, 2002; Duplessy et al., 2002; Lea et al., 2002). First, $\delta^{18}\text{O}$ in fossil oceanic Foraminifera is a function of both the isotopic composition and the temperature of the water in which the Foraminifera grew. Second, $\delta^{18}\text{O}$ of ocean water is a function of both mean ocean $\delta^{18}\text{O}$, itself reflecting continental ice volume and the average $\delta^{18}\text{O}$ of continental ice, and local hydrological effects. Thus, we require information on ice volumes on each of the major ice sheets and their different isotopic compositions (Duplessy et al., 2002) as well as temperature and salinity changes of the water in which the Foraminifera grew. In summary, graphs that show $\delta^{18}\text{O}$ changes in oceanic Foraminifera through multiple glacial cycles, together with a second axis labeled meters sea-level change are not direct observations of the latter but estimates, or reconstructions, from models based on a long series of assumptions and calculations, some of which are not well constrained.

General Principles

Along with many aspects of Quaternary Science, research into sea-level changes advanced significantly with the development of improved methods for dating sediment, especially radiocarbon dating. Perhaps, equally significant were the activities of international groups of scientists under the auspices of International Geological Correlation Programme (IGCP) and INQUA. From the 1970s onward, through a series of IGCP Projects and INQUA Commissions and Subcommissions, national, regional, and international sea-level research benefited from high-profile, interdisciplinary conferences, and field meetings that brought forward many new ideas, researchers, and research topics that ranged in spatial scale from local to global. Early debates noted the regionally different patterns of emerged and submerged shorelines and sediments. Glacio-isostasy was an early hypothesis and part explanation, but the significance of hydro-isostasy and the effects of gravitational attraction on the ocean surface only became accepted with advances in geophysical models and their comparison with reconstructions. Early model-sea-level reconstruction comparisons suffered from highly variable quality control on reconstructions of relative sea level and their incorporation into databases. Methodological debates and international initiatives under IGCP and INQUA brought an improved degree of rigor in data collection and interpretation during the 1980s (Devoy, 1987, chapters 1 and 7; van de Plasche, 1986). These are summarized further in following sections.

From the 1960s onward, the literature contains a recurring debate on the general pattern of eustatic sea-level change since the Last Glacial Maximum; was it a generally smooth function with time or characterized by a series of submillennial oscillations of decimeter to meter scale superimposed on the broader scale pattern of change? To what extent are these climatically induced changes or noise in the sedimentary record, uncertainties in the age and elevation estimates of paleosea-level indicators? Despite better age controls and improved sampling techniques, these complexities continue to confound interpretations. Much of the difference of opinion may be resolved by a more explicit consideration of scales, both spatial and temporal, within a hypothesis testing framework. Not all studies address these issues explicitly, with the danger that

factors important at one scale may not be adequately resolved for a study of processes assumed dominant at another scale. For example, when studying the spatial pattern of isostatic deformation of the crust, it is inherently assumed that local factors that control each sea-level index point used have been quantified, taken into account, or removed from subsequent analyses.

For each geographical location (φ) the change in relative sea level (ΔRSL) at time t , where t is the time relative to present (Shennan et al., 2012):

$$\Delta\text{RSL}(\varphi, t) = \Delta\text{EUS}(t) + \Delta\text{ISO}(\varphi, t) + \Delta\text{TTECT}(\varphi, t) + \Delta\text{LOCAL}(\varphi, t) + \Delta\text{UNSP}(\varphi, t) \quad [1]$$

where:

- $\Delta\text{EUS}(t)$ is the time-dependent eustatic sea level, derived from the model of ice history, that would result by distributing any meltwater evenly across a rigid, nonrotating planet and neglecting self-gravitation in the surface load.
- $\Delta\text{ISO}(\varphi, t)$ is the total isostatic effect of the glacial rebound process including the ice load (glacio-isostatic), water load (hydro-isostatic) and rotational contributions to the redistribution of ocean mass.
- $\Delta\text{TTECT}(\varphi, t)$ is the tectonic effect, which can operate at different timescales.
- $\Delta\text{LOCAL}(\varphi, t)$ is the total effect of local processes within the coastal system. Since we use observations from the sedimentary record to reconstruct relative sea-level elevation we can express these as the sum of $\Delta\text{TIDE}(\varphi, t) + \Delta\text{SED}(\varphi, t)$, where $\Delta\text{TIDE}(\varphi, t)$ is the total effect of tidal regime changes and the elevation of the sediment with reference to tide levels at the time of deposition (e.g., Shennan and Horton, 2002; Uehara et al., 2006). $\Delta\text{SED}(\varphi, t)$ is the total effect of sediment consolidation since the time of deposition. Numerous studies show that this can be both a major process in coastal evolution (Horton and Shennan, 2009; Tornqvist et al., 2008; van Asselen, 2010) and a key variable in reconstructing relative sea-level change from index points taken from sediments.
- $\Delta\text{UNSP}(\varphi, t)$ is the sum of other unspecified factors, either not quantified or not thought of. Implicitly most studies assume their total effect close to zero and random.

When we plot reconstructions of relative sea-level change using geological sea-level index points and find a scatter of data points that do not overlap, this indicates one or more of the following: (1) underestimation of age and/or elevation error terms; (2) plotting data from a too large spatial extent, so incorporating differential GIA; (3) underestimation of $\Delta\text{LOCAL}(\varphi, t)$; (4) evidence of $\Delta\text{UNSP}(\varphi, t)$ (Shennan et al., 2012).

Depending upon the hypothesis of the study in question, one or more of the factors on the right-hand side of eqn [1] are assumed constant, known, or negligible, depending upon the temporal or spatial scale appropriate. If these are not considered, then the study is in danger of becoming just a description of proxy data observed in the field or laboratory. As demonstrated in other disciplines such as geomorphology, as the time span considered becomes smaller, then the spatial scale of dependent variables to be considered also decreases. Figure 1 illustrates global to local scale processes that contribute to sea-level change (Shennan et al., 2012).

For timescales in the order of 10^2 – 10^5 years, we would mostly consider those relationships in blue (Figure 1), for

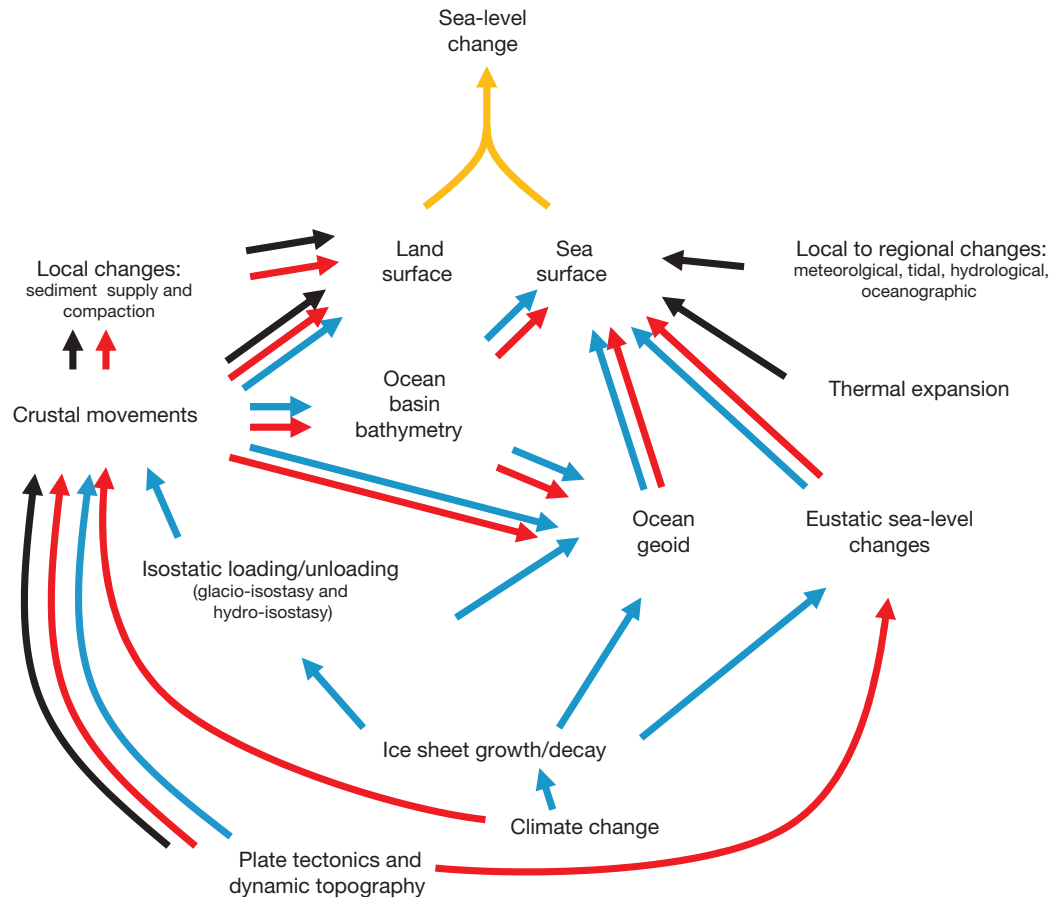


Figure 1 Schematic diagram of global to local scale processes that contribute to sea-level change. Note that it does not include Earth rotation. For timescales in the order of 10^2 – 10^5 years we would mostly consider those relationships in blue; for timescales $\sim 10^0$ – 10^2 years, and the local scale of individual coastal environments we must also consider those in black; and for long timescales, $>10^5$ years, those relationships in red.

example, when comparing geological indicators of sea level with GIA model estimates of sea level. At shorter time periods, 10^0 – 10^2 years, and the local scale of individual coastal environments we must also consider those in black. Factors at these scales could dominate the elevation at which sea-level index points form, yet we may be tempted to assume that they can be accounted for when such indicators are used to reconstruct sea level in the past.

At long timescales, we must also consider those relationships in red. This includes the influence of sediment erosion and deposition on the ocean geoid and solid earth. Across Quaternary timescales, global sediment transfer from the continents to oceans since the mid-Pliocene are of the same order of magnitude as the glacial–interglacial transfer of ice and meltwater, so the effects of sediment unloading and loading on ocean bathymetry and the ocean geoid deserve investigation alongside climate change and GIA (Shennan, 2011).

Note how a factor such as plate tectonics can be considered over different timescales: driving changes in ocean shape, sometimes called tectono-eustasy; crustal deformation adjacent to plate boundaries so influencing Holocene-scale relative sea-level change at sites such as Barbados and Huon Peninsula; and abrupt changes during great earthquakes, with coseismic land movements, displacement of the sea floor and tsunami generation.

Sea-Level Index Points and Reconstructions of Relative Sea Level

IGCP Projects 61 and 200 (van de Plassche, 1986) promoted a universally applicable approach to the reconstruction of relative sea-level histories for different locations and environments. Reliable reconstructions of past changes in sea level relative to present, with quantified uncertainty terms, come from *in situ* sediments and morphological features whose origin was controlled by paleosea level. Where such sediments and morphological features survive they can be used as sea-level index points by defining four attributes: location, age, altitude, and tendency. Although this formal approach and terminology never attained universal application, the principles behind them can be seen in all reconstructions of paleo-sea levels, as evidenced in the separate entries in the sea-level section.

Location Attribute of a Sea-Level Index Point

Once it is established that the sediment or morphological feature has not been transported from its original location, this attribute consists simply of its geographical coordinates.

Age Attribute of a Sea-Level Index Point

The majority of reconstructions used for studies of sea-level changes since the Last Glacial Maximum have their ages measured by radiocarbon techniques and most modern studies use calibrated ages expressed with 95% confidence limits. Other widely used methods for establishing a chronology include various measurements based on sediment characteristics, such as luminescence, isotopic properties, and archaeological evidence. Morphological expressions of past sea level, including beaches, notches, cliffs have the great advantage that they provide a spatial expression of the consistency or variation in relative sea level, but their use is severely limited if they cannot be dated reliably. Organic material incorporated in raised shorelines, for example, is not always *in situ*, so provides only a limiting age of the feature. The most reliable sea-level index points should have at least one type of corroborating evidence to support the radiocarbon age and to demonstrate continuity of sedimentation or formation.

Elevation Attribute of a Sea-Level Index Point

Few sea-level index points formed exactly at paleomean sea level. Many represent environments within the upper part of the tidal range but in total they cover the full tidal range, the shallow subtidal zone and, for limiting dates, beyond. Limiting dates come from either samples from freshwater environments inland of the paleocoastline and at or above the past high-tide level or fully marine environments for which only a minimum water depth can be given. In reconstructing relative sea-level change, the dated freshwater sample must lie, on a plot of elevation against age, at or above data points such as those sea-level index points whose origin was directly controlled by the paleosea level. The same argument applies to dated samples that represent marine environments. They must lie at or below the other index points on the altitude versus age plot.

'Indicative meaning' defines the relationship of the sample to tidal range and allows us to measure relative sea-level change, defined relative to present (van de Plassche, 1986). Indicative meaning comprises two parameters, a reference water level (e.g., mean high water of spring tides – MHWST; or a tide level with an offset, called the indicative difference, e.g., MHWST +0.2 m) and indicative range (the vertical range over which the dated indicator could occur as it is forming). The reference water level allows calculation of sea-level change relative to present, based on the elevation of both the index point and the reference water level to a common datum. This may be the national datum or one defined locally through a series of observations of present-day tide levels. Most analyses assume no change in tidal range since the time of deposition but various modeling studies show this may be an important factor, with changes up to a few meters depending on the amplification of the tidal wave as the coastline shape and bathymetry change through time (Gehrels et al., 1995; Shennan and Horton, 2002; Uehara et al., 2006). Other factors that contribute to the height uncertainty or error term come from all of the measurements that must be made during the sampling process.

The total height uncertainty or error term, e_h is calculated by:

$$e_h = \sqrt{(e_1^2 + e_2^2 + \dots + e_n^2)}$$

where $e_1 \dots e_n$ are the separate sources of uncertainty, expressed as 95% confidence limits.

A final factor with a potentially significant effect is sediment consolidation. There is no widely accepted approach to quantifying this, not least because the necessary physical parameters of the sediments most probably have not been measured. Some studies include an empirical correction for sediment consolidation while others aim to minimize the problem by concentrating on sea-level index points underlain by sediments that undergo minimal or no consolidation. In paleoestuarine sequences, these index points usually come from thin organic tidal marsh sediment or peat that overlies glacial till or other pre-Holocene sediments. Much of the scatter of index points in relative sea-level graphs from estuaries with thick Holocene sequences derives from the net effect of sediment consolidation, where the index points from the basal tidal marsh sediment or peat mostly lie within the upper part of the scatter (Shennan and Horton, 2002). Residuals from the general trend show statistically significant correlations with various sediment characteristics related to consolidation. In general, for paleoestuarine environments best-fit relative sea-level reconstructions or predictions would lie toward the top of the scatter.

Tendency Attribute of a Sea-Level Index Point

The tendency of a sea-level index point describes the increase (positive sea-level tendency) or decrease (negative sea-level tendency) in marine influence recorded by the index point. Most analyses of the pattern of sea-level tendencies consider changes at the local scale, either within or between estuaries, such as relationships between horizontal shifts in the coastal sedimentary environments, sediment accumulation, consolidation and the potential driving mechanism of relatively small, ~decimeters, vertical changes in sea level (Allen, 1997). In the analysis of larger-scale processes, such as testing relative sea-level predictions from isostatic adjustment models, observations of a change in sign of the tendency are useful in determining the turning points in relative sea-level curves, whether from the near-field or far-field. Furthermore, sea-level tendencies show whether relative sea level was slightly above or below present in the last few thousand years, thus identifying the geographic limits of the transition from mid-Holocene relative land uplift to relative submergence.

Sea-Level Changes: Local to Global

Each of the contributions to the Quaternary sea levels section summarizes the range of different processes that control registration of changes in sea level in different locations. Implicitly or explicitly in different degrees, they address the issues of scale, age, location, tendency, elevation, and the associated errors. They reflect the great diversity of issues that studies of Quaternary sea-level change contribute to:

- global ice volumes and meltwater discharge since the Last Glacial Maximum;
- regional patterns of isostatic uplift;

- coseismic and interseismic deformation of the crust close to plate boundaries;
- late Holocene trends as a background to global warming;
- centimeter scale changes caused by short-term climate, oceanographic, or hydrological changes.

To illustrate the inter-relationships within these studies, we can look at the progress in integrating quantitative geophysical models, global meltwater discharge (eustatic sea-level change), and local reconstructions of relative sea-level change around the world. As noted above, perturbations to the ocean geoid and deformation of the solid Earth in response to climate-induced ice and ocean water mass redistribution mean that no single location records only the eustatic component. Far-field locations, those distant from the late Quaternary ice sheets, most closely resemble the eustatic component but they still record relative sea-level change. Indeed, attempts to fit relative sea-level reconstructions from far-field locations with numerical models of GIA and models of global ice distributions reveal significant misfits (Bassett et al., 2005; Lambeck et al., 2002; Milne et al., 2002; Peltier, 1998, 2002a,b). These differences arise from the number of unknown parameters, including Earth model parameters, ice model parameters, as well as uncertainties in relative sea-level reconstructions. While GIA models successfully explain the major differences in relative sea-level changes on a global scale, frequently summarized by figures of six characteristic zones it is infrequently recognized that existence of the zones and the position of the boundaries between them are a function of both the Earth and deglaciation models adopted.

Most approaches use an Earth model that is a spherically symmetric, self-gravitating, compressible Maxwell viscoelastic body. Differences exist between models in how to treat the viscous structure from the surface to the core-mantle boundary. Lambeck et al. (1996) compared three-, four-, and five-layer models and while their optimum solution is for a five-layer model, a three-layer model gives similar results and is adequate for predicting relative sea-level change. In contrast, Peltier (1998, 2004) models the depth-dependence of mantle viscosity in much smaller steps, each ~50–70-km depth, but the major changes occur at the base of the model lithosphere and ~670-km depth. In a three-layer model, the upper layer, to simulate the lithosphere, has a large viscosity so that it behaves as an elastic plate on late Quaternary timescales. The second layer, a model for the upper mantle, extends from the base of the model lithosphere to the 670-km deep seismic velocity discontinuity, below which, the core-mantle boundary simulates the lower mantle.

Differences in model parameters were much greater in analyses during the 1990s, well illustrated by the value used for the depth to the base of the lithosphere, ranging from the ‘thin lithosphere’ models, ~30 km to ‘thick lithosphere’ models, ~200 km. With subsequent analyses using relative sea-level data from more locations, with greater reliability as a result of using the approaches discussed above, and other types of data, including the gravity field, perturbations in Earth rotation and measurements of present rates of crustal motion, differences in model parameters are now less. For example, most models use a lithosphere in the range ~65–95 km, except for modeling areas close to plate boundaries for which a spherically symmetric Earth model is not applicable.

In order to test Earth model parameters, we require relative sea-level data from a range of areas because of the different sensitivities to mass redistributions of ice and water. The glacial isostatic component of predicted relative sea-level change for regions formerly beneath large ice sheets at the Last Glacial Maximum, such as Scandinavia and North America, show a pronounced sensitivity to deeper Earth structure, especially lower mantle viscosity. In contrast, predictions of relative sea-level change for regions beneath or close to smaller ice sheets, such as the British Isles, are highly sensitive to shallow Earth structure, especially lithospheric thickness and the viscosity of the upper mantle. Data from far-field locations help constrain upper and lower mantle parameters but are relatively insensitive to different models for the lithosphere.

Despite more than 150 years of research, estimates of ice sheet dimensions, from local to global scales, at different times during the Quaternary remain contentious. The CLIMAP study of the Earth at the Last Glacial Maximum (CLIMAP, 1981) presented two reconstructions: a ‘minimum model’ with 127 m of ice-equivalent sea-level lowering, or eustatic change; and a ‘maximum model’ with 163 m of eustatic change. More recent assessments narrow the uncertainty of eustatic sea level at the Last Glacial Maximum to a range from ~114 to 135 m (Clark and Mix, 2002; Lambeck et al., 2002; Milne et al., 2002; Peltier, 2002b). The isostatic effects, $\Delta\text{ISO}(\varphi, t)$ in eqn [1], and any tectonic movement, $\Delta\text{TECT}(\varphi, t)$, mean that no site, even in the far-field, exactly records eustatic sea-level change, only relative sea-level change at the site, $\Delta\text{RSL}(\varphi, t)$. As a result, models of eustatic sea-level change, expressed as ice-equivalent sea level, differ as a result of their calibration against different datasets of relative sea-level reconstructions and the Earth models used to predict relative sea-level change. For example, the differences between the ICE-4G and ICE-5G (Figure 2) result primarily from new data becoming available from Bonaparte Gulf and Sunda Shelf together with various lines of evidence that pointed to a larger, multidomed Laurentide ice sheet (Peltier, 2004).

The main far-field records (Figure 3) come from Barbados (Bard et al., 1990; Fairbanks, 1989), Tahiti (Bard et al., 1996), Huon (Chappell and Polach, 1991; Cutler et al., 2003), Bonaparte Gulf (Lambeck et al., 2002; Yokoyama et al., 2000) and Sunda Shelf (Hanebuth et al., 2000). The data from Barbados, Tahiti, and Huon require a correction for long-term tectonic movement, usually corrected assuming a uniform tectonic uplift of 0.25, –0.1, and 1.76 mm year^{–1}, respectively.

A further, highly significant complication in reconciling the far-field relative sea-level records with global ice sheet reconstructions is the high dependence upon where it is assumed the ice was at each point in time, since this influences the effect of gravitational attraction on the water in the ocean. When each ice sheet melted remains contentious and this can be illustrated with the debate on the existence of major changes in the rate of global meltwater discharge, especially the so-called meltwater pulses: MWP1a ~14.0 ka BP and MWP1b ~11.3 ka BP.

The existence, or not, of meltwater pulses and their magnitude has considerable implications for understanding rapid climate change and numerous feedback mechanisms in the climate–ocean–Earth system. Differences remain in the

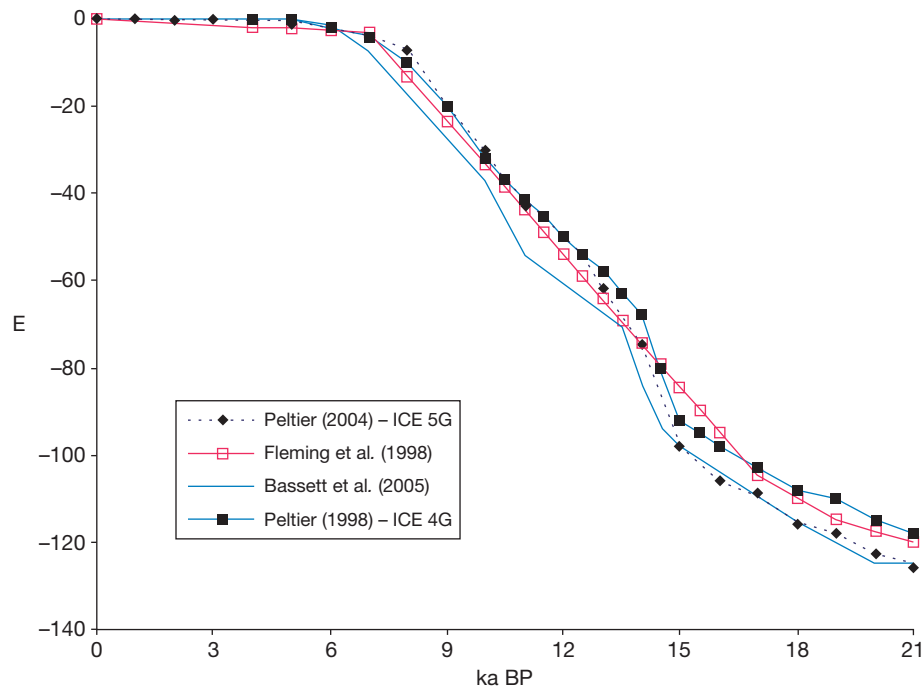


Figure 2 Four recent models of global ice melt since the Last Glacial Maximum shown as ice-equivalent sea-level, or eustatic sea-level change (Bassett et al., 2005; Fleming et al., 1998; Peltier, 1998, 2004). Differences arise from different interpretations of the data used to delimit relative sea levels at the Last Glacial Maximum and different Earth and ice model parameters used to explain relative sea-level changes at sites (see Figure 3).

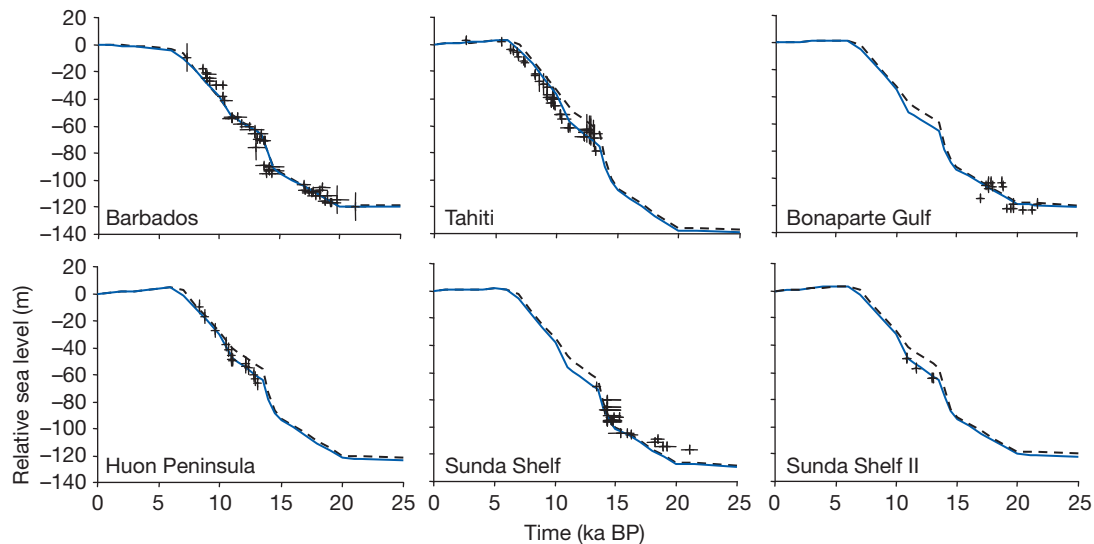


Figure 3 Observed sea-level index points and sea-level predictions at five far-field locations (data from Bassett et al., 2005). Predictions differ on the basis of the assumed contribution to meltwater pulse 1a from the Antarctic ice sheet (in units of equivalent eustatic sea-level change). Dashed line is for no contribution from the Antarctic ice sheet; solid line for a dominant contribution from the Antarctic ice sheet, 15 m out of 24 m equivalent eustatic sea-level rise. Sunda II shows the subset of cores from the northerly sites, sufficiently far away from the other Sunda sites to give different relative sea-level predictions.

interpretation of field observations, with the evidence for MWP-1a generally more accepted than MWP-1b (Bard et al., 1996; Fairbanks, 1989, 1990). Most eustatic models include a distinct MWP-1a but others argue for a nearly uniform rise for the whole of the period ~15–7 ka BP (Figure 2). At Barbados,

Fairbanks (1989) recorded MWP-1a as ~24 m relative sea-level rise in less than 1000 years. The geographical source of the meltwater in MWP-1a impacts upon both the spatially variable relative sea-level change recorded at far-field and near-field sites and our understanding of climate change and the

processes leading to rapid deglaciation. Peltier (2005) argues for a Northern Hemisphere source, primarily but not exclusively, Laurentide, whereas others suggest a significant Antarctic contribution (Bassett et al., 2005; Clark et al., 2002). Figures 2 and 3 demonstrate the significance of these contrasting views in relation to relative sea level at the key far-field locations.

Using the same model of eustatic sea-level change (Figure 2) and an Earth model tuned to fit the Barbados reconstructions of relative sea level, Bassett et al. (2005) show there are significant differences in the predictions of relative sea-level change for the other far-field locations (Figure 3), depending upon whether the dominant contribution for MWP-1a was from Antarctica or not. The dashed line is for no contribution from the Antarctic ice sheet and the solid line for a dominant contribution from the Antarctic ice sheet, 15 m out of 24 m equivalent eustatic sea-level rise. Peltier (2004) offers a different analysis, based on different Earth and ice models, that shows almost the opposite, a predominant Laurentide source for MWP-1a. Peltier (2004) also draws attention to the broad error terms in the far-field data that define MWP-1a, showing that any amplitude in the range of 10–25 m would appear to be allowed. This highlights the more general difficulty of identifying changes in the rate of sea-level change given the error terms of most index points. Nevertheless, it is possible to test explicitly such an hypothesis, for example, the magnitude of MWP-1a, by seeking its signal in a near-field location where the rate of glacio-isostatic uplift is of similar magnitude but of opposite sign.

Figure 4 shows relative sea-level changes at Arisaig, northwest Scotland, since ~16 000 cal years BP (Shennan et al., 2005). The marine limit occurs between ~34 and 38 m above present. Rapid relative sea-level fall continues until at least ~14 500 cal years BP. This fall constrains the acceleration in global meltwater discharge around 14 000 cal years BP, MWP-1a. If meltwater discharge exceeded the rate of relative sea-level fall then sediment

sequences around 15 m above present (Figure 4) would record a change in sea-level tendency, from negative to positive then back to negative. No reversal in relative sea level occurs in any of the sites around this elevation although it remains possible that a reversal falls within the gap of data 13 000–14 000 cal years BP. Negative sea-level tendency for all sea-level reconstructions from ~16 000 to ~12 000 cal years BP limit the acceleration of eustatic sea-level rise during meltwater pulse 1a to ~30 mm year⁻¹ or ~11 km³ year⁻¹ meltwater discharge. This is approximately half the rate predicted by Fairbanks (1989, 1990).

One other continuing debate is the nature of late Holocene deglaciation. The majority of opinion up to the 1980s was that all significant ice melt ceased abruptly at ~7 ka BP. Since then a number of analyses show a gradual end to melting with up to 3 m equivalent sea-level rise since ~7 ka BP, with some debate on when melting ceased, by 4 ka or continuing to preindustrial times (Fleming et al., 1998; Lambeck et al., 1998, 2002; Nakada and Lambeck, 1988; Peltier, 1998, 2004; Peltier et al., 2002; Shennan et al., 2002). The rapid termination of ice melting shown in Figure 3 reflects the poor constraint provided by the long time-series far-field data. Numerous studies show that this produces a Holocene highstand around 6000 BP that is too sharply defined compared to reconstructions at other far-field and near-field sites (Fleming et al., 1998). The predicted relative sea-level curve in Figure 4 shows a flattened peak to the mid-Holocene highstand and numerous sea-level index points from the field area indicate that the highstand persisted more than 1000 years before the onset of any measurable relative fall in sea level (Shennan et al., 2005). These point to gradual cessation of melting of the Laurentide and Antarctic ice sheets rather than an abrupt termination. The model used in Figure 4 has melting ending at 2000 BP and also gives predictions consistent with reconstructions of mid-Holocene relative sea-level above present on Pacific Islands (Peltier et al., 2002; Shennan et al., 2002).

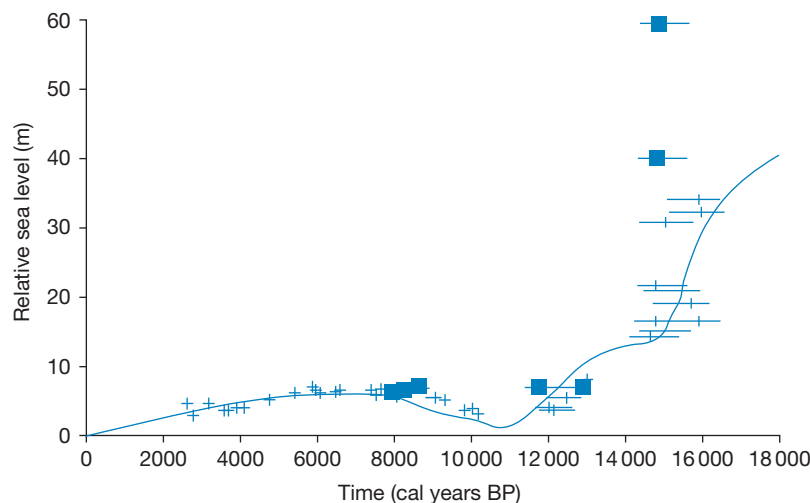


Figure 4 Reconstructions and model predictions of relative sea-level change 18 000 cal years BP to present at Arisaig, northwest Scotland (Shennan et al., 2005). Age error bars for sea-level index points indicate the maximum and minimum calibrated ages; the vertical bar crosses at the median calibrated age and shows the sum of all quantified or estimated elevation errors. Relative sea level must lie at or below limiting dates, shown as solid symbols ■. Model predictions come from the modified version of Peltier et al. (2002), described as model 3 in Shennan et al. (2002).

See also: Radiocarbon Dating: AMS Radiocarbon Dating; Sea Level Studies: Coral Records of Relative Sea-Level Changes; Eustatic Sea-Level Changes – Glacial–Interglacial Cycles; Eustatic Sea-Level Changes Since the Last Glacial Maximum; Geomorphological Indicators; Isostasy: Glaciation-Induced Sea-Level Change; Microfossil-Based Reconstructions of Holocene Relative Sea-Level Change; Sedimentary Indicators of Relative Sea-Level Changes – High Energy; Sedimentary Indicators of Relative Sea-Level Changes – Low Energy; Use of Cave Data in Sea-Level Reconstructions; Sea-Levels, Late Quaternary: Late Quaternary Relative Sea-Level Changes at Mid-Latitudes; Late Quaternary Relative Sea-Level Changes in High Latitudes; Late Quaternary Relative Sea-Level Changes in the Tropics; Late Quaternary Sea-Level Changes in Greenland; Tectonics and Relative Sea-Level Change.

References

- Allen JRL (1997) Simulation models of salt-marsh morphodynamics: Some implications for high-intertidal sediment couplets related to sea-level change. *Sedimentary Geology* 113: 211–233.
- Bard E, Hamelin B, Arnold M, et al. (1996) Deglacial sea-level record from Tahiti corals and the timing of global meltwater discharge. *Nature* 382: 241–244.
- Bard E, Hamelin B, Fairbanks RG, and Zindler A (1990) Calibration of the ^{14}C timescale over the past 30,000 years using mass spectrometric U–Th ages from Barbados corals. *Nature* 345: 405–410.
- Bassett SE, Milne GA, Mitrovica JX, and Clark PU (2005) Ice sheet and solid earth influences on far-field sea-level histories. *Science* 309: 925–928.
- Chappell J and Polach H (1991) Post-glacial sea-level rise from a coral record at Huon Peninsula, Papua New Guinea. *Nature* 349: 147–149.
- Clark PU, Mitrovica JX, Milne GA, and Tamisiea M (2002) Sea-level fingerprinting as a direct test for the source of global meltwater pulse 1A. *Science* 295: 2438–2441.
- Clark PU and Mix AC (2002) Ice sheets and sea level of the Last Glacial Maximum. *Quaternary Science Reviews* 21: 1–7.
- CLIMAP (1981) *Seasonal Reconstruction of the Earth's Surface at the Last Glacial Maximum*. Boulder, CO: Geological Society of America Map and Chart Series C36.
- Cutler KB, Edwards RL, Taylor FW, et al. (2003) Rapid sea-level fall and deep-ocean temperature change since the last interglacial period. *Earth and Planetary Science Letters* 206: 253–271.
- Devoy RJN (1987) *Sea Surface Studies: A Global View*. Beckenham: Croom Helm p. 649.
- Duplessy JC, Labeyrie L, and Waelbroeck C (2002) Constraints on the ocean oxygen isotopic enrichment between the Last Glacial Maximum and the Holocene: Paleocceanographic implications. *Quaternary Science Reviews* 21: 315–330.
- Fairbanks RG (1989) A 17000-year glacio-eustatic sea level record: Influence of glacial melting rates on the Younger Dryas event and deep-ocean circulation. *Nature* 342: 637–642.
- Fairbanks RG (1990) The age and origin of the 'Younger-Dryas climate event' in Greenland ice cores. *Paleoceanography* 5: 937–948.
- Fleming K, Johnston P, Zwart D, Yokoyama Y, Lambeck K, and Chappell J (1998) Refining the eustatic sea-level curve since the Last Glacial Maximum using far- and intermediate-field sites. *Earth and Planetary Science Letters* 163: 327–342.
- Gehrels WR, Belknap DF, Pearce BR, and Gong B (1995) Modelling the contribution of M2 tidal amplification to the Holocene rise of mean high water in the Gulf of Maine and the Bay of Fundy. *Marine Geology* 124: 71–85.
- Hanebuth T, Stattegger K, and Grootes PM (2000) Rapid flooding of the Sunda Shelf: A late-glacial sea-level record. *Science* 288: 1033–1035.
- Horton BP and Shennan I (2009) Compaction of Holocene strata and the implications for relative sea-level change on the east coast of England. *Geology* 37: 1083–1086.
- Lambeck K, Johnston P, Smither C, and Nakada M (1996) Glacial rebound of the British Isles – III. Constraints on mantle viscosity. *Geophysical Journal International* 125: 340–354.
- Lambeck K, Smither C, and Johnston P (1998) Sea-level change, glacial rebound and mantle viscosity for northern Europe. *Geophysical Journal International* 134: 102–144.
- Lambeck K, Yokoyama Y, and Purcell A (2002) Into and out of the Last Glacial Maximum: Sea-level change during Oxygen Isotope Stages 3 and 2. *Quaternary Science Reviews* 21: 343–360.
- Lea DW, Martin PA, Pak K, and Spero HJ (2002) Reconstructing a 350 ky history of sea level using planktonic Mg/Ca and oxygen isotope records from a Cocos Ridge core. *Quaternary Science Reviews* 21: 283–293.
- Milne GA, Mitrovica JX, and Schrag DP (2002) Estimating past continental ice volume from sea-level data. *Quaternary Science Reviews* 21: 361–376.
- Nakada M and Lambeck K (1988) The melting history of the Late Pleistocene Antarctic ice sheet. *Nature* 333: 36–40.
- Peltier WR (1998) Postglacial variations in the level of the sea: Implications for climate dynamics and solid-earth geophysics. *Reviews of Geophysics* 36: 603–689.
- Peltier WR (2002a) Global glacial isostatic adjustment: Palaeogeodetic and space-geodetic tests of the ICE-4G (VM2) model. *Journal of Quaternary Science* 1: 491–510.
- Peltier WR (2002b) On eustatic sea level history: Last Glacial Maximum to Holocene. *Quaternary Science Reviews* 21: 377–396.
- Peltier WR (2004) Global glacial isostasy and the surface of the ice-age Earth: The ICE-5G (VM2) model and GRACE. *Annual Review of Earth and Planetary Sciences* 32: 111–149.
- Peltier WR (2005) On the hemispheric origins of meltwater pulse 1a. *Quaternary Science Reviews* 24: 1655–1672.
- Peltier WR, Shennan I, Drummond R, and Horton BP (2002) On the postglacial isostatic adjustment of the British Isles and the shallow viscoelastic structure of the Earth. *Geophysical Journal International* 148: 443–475.
- Shennan I (2011) Palaeoclimate: Sea level from global to local. *Nature Geoscience* 4: 283–284.
- Shennan I, Hamilton S, Hillier C, and Woodroffe S (2005) A 16,000-year record of near-field relative sea-level changes, northwest Scotland, United Kingdom. *Quaternary International* 133–134: 95–106.
- Shennan I and Horton BP (2002) Holocene land- and sea-level changes in Great Britain. *Journal of Quaternary Science* 1: 511–526.
- Shennan I, Peltier WR, Drummond R, and Horton BP (2002) Global to local scale parameters determining relative sea-level changes and the post-glacial isostatic adjustment of Great Britain. *Quaternary Science Reviews* 21: 397–408.
- Shennan I, Milne GA, and Bradley SL (2012) Late Holocene vertical land motion and relative sea-level changes: Lessons from the British Isles. *Journal of Quaternary Science* 27: 64–70.
- Tornqvist TE, Wallace DJ, Storms JEA, et al. (2008) Mississippi Delta subsidence primarily caused by compaction of Holocene strata. *Nature Geoscience* 1: 173–176.
- Uehara K, Scourse JD, Horsburgh KJ, Lambeck K, and Purcell AP (2006) Tidal evolution of the northwest European shelf seas from the Last Glacial Maximum to the present. *Journal of Geophysical Research* 111: C09025.
- van Asselen S (2010) The contribution of peat compaction to total basin subsidence: Implications for the provision of accommodation space in organic-rich deltas. *Basin Research* 23: 239–255.
- van de Plassche O (1986) *Sea-Level Research: A Manual for the Collection and Evaluation of Data*, p. 618. Norwich: GeoBooks.
- Yokoyama Y, Lambeck K, De Deckker P, Johnston P, and Fifield LK (2000) Timing of the Last Glacial Maximum from observed sea-level minima. *Nature* 406: 713–716.

Geomorphological Indicators

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Introduction

At any time, marine and coastal processes produce many kinds of geomorphologic marks, especially on rocky shores, that are related to the contemporary sea-level position. After a rise in the relative sea level, marks that became submerged are generally eroded or fossilized when capped by sediments. In the case of a fall in the relative sea level, marine marks are rapidly eroded if they remain in the reach of wave action in the mid-littoral zone. Above in the supralittoral zone, which is never submerged but may be kept wet by spray, erosion is limited to microholes produced by lichens or Cyanobacteria. At higher elevations, the marine or maritime action becomes negligible and geomorphologic marks are exposed only to the slow action of weathering and diagenesis processes. A good knowledge of the marine features related to the present sea level is essential to identify and interpret marks related to former sea levels (Pirazzoli, 1996, 2005).

Erosion Marks

Erosion features which may be useful as sea-level indicators can be preserved only in hard rock and occur in a vertical range that depends mainly on site exposure.

Notches (Tidal, Abrasion, and Structural)

A marine notch is an indentation or undercutting, a few centimeters to several meters deep, left by sea erosion in coastal rocks. In coastal regions affected by rapid vertical movements, incipient notches may form sequences of stepped erosional shorelines, like the nine ‘ripple notches’ developed between 4 and 1.7 ka BP in Antikythira Island (Greece; Pirazzoli, 1986, photo 16), or the spectacular ‘staircase’ of 185 successive raised shorelines built from 8.3 sidereal ka BP to the present, during the emergence of the land due to glacio-isostatic uplift on the eastern coast of Hudson Bay (Canada), reported by Fairbridge and Hillaire-Marcel (1977).

Pirazzoli (1986) has classified several kinds of erosion notches. Tidal notches are one of the most precise sea-level indicators. They are typical mid-littoral erosion features, especially in limestone coasts. In a sheltered site with a moderate tidal range, the most common profiles are recumbent V-shaped or U-shaped forms (Figure 1(a) and 1(a’)), with a vertex located near mean sea level (m.s.l.), the base of the notch near the lowest tide level and the top near the highest tide level. If sea level does not change, the undercut will deepen (Figure 1(b)) and undermine the cliff until the roof collapses. After removal of the fallen material by waves, the base of the former notch will look like a small bench, while a new notch will start forming inside the cliff.

Though early studies ascribed this zonal erosion to solution phenomena, it is now recognized that tidal notch development is mainly related to bioerosion (Torunski, 1979; Trudgill, 1976) and zones of rapid tidal notch development correlate well with zones of major eroding organisms, the maximum density of grazing organisms being in the range of the neap tide (semidiurnal tide of small range which occurs near the time of the first and last lunar quarters, between spring tides) or near m.s.l. Rates of undercutting of $0.2\text{--}5.0\text{ mm year}^{-1}$ have been reported in the literature, the most frequent values being in the order of $1.0\text{--}1.5\text{ mm year}^{-1}$ in the tropics and of 1 mm year^{-1} in the Mediterranean (Laborel et al., 1999).

In areas sheltered from wave action, elevated or submerged tidal notches can be used to specify former sea-level positions with up to a decimeter confidence and also to provide qualitative information on the rate of sea-level change and on tectonic movements. A rapid change in the relative sea level, greater than the tidal range (Figure 1(c)), will uplift or submerge the whole notch, preserving it from further mid-littoral erosion, and a new notch will develop in the new intertidal zone. If the rapid vertical displacement is less than the tidal range, or if the movement is slow and gradual, the floor/roof of the notch will be eroded and the height of the notch will increase (Figure 1(e)). Repeated rapid movements will produce successive overlapping indentations, the so-called ripple notches (Figures 1(d) and 2).

When taken away from mid-littoral erosion, tidal notches may be preserved for a long time and late Holocene features (Figure 3) may be hard to distinguish from last Interglacial ones (Figure 4). In Sumba Island (Indonesia), a tidal notch at 350 m in altitude, of an estimated age of 730 ka, is still clearly visible (Pirazzoli et al., 1993; Figure 3).

Where mechanical action such as abrasion by wave-borne sand or gravel predominates, notches cut in hard rock have a rounded and polished appearance. Such abrasion notches may develop at any level reached by waves or by hydrodynamic agents acting in combination with granular deposits. As wave energy lessens in shallow water, abrasion notches are mainly found in the upper part of the mid-littoral zone, often at cliff base. The vertical precision of abrasion notches as sea-level indicator is generally weak (± 1 to several meters), depending on the local tide and site exposure.

In rock formations consisting of horizontal or gently inclined beds, differential erosion may excavate the weakest layers, along with joints and fissures and thus produce linear reentrants with notch-like profiles. Such features are therefore of a structural origin, that is usually easy to recognize in the field: the attitude does not remain horizontal but follows the original line of weakness, the notch profile may vary along its length, the height of the notch corresponds to certain layers in the rock. Structural notches should be recognized as such and not used as sea-level indicators.

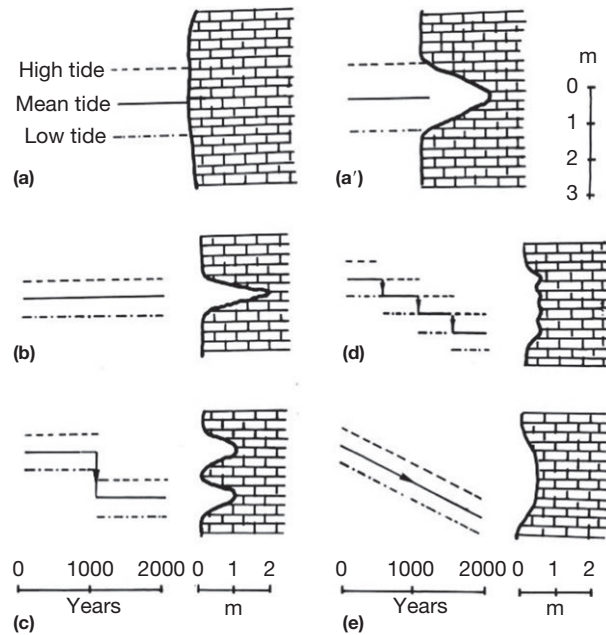


Figure 1 Tidal notch profiles resulting from different compilations of relative sea-level change in sheltered places. The maximum erosion rate is assumed to be 1.0 mm year^{-1} and to occur at m.s.l. Reproduced from Pirazzoli PA (1986) *Marine notches*. In: Van de Plassche O (ed.) *Sea-Level Research: A Manual for the Collection and Evaluation of Data*, pp. 361–400. Norwich: Geo Books.



Figure 2 Marks left by uplifted tidal notches on the limestone cliffs near cape Ladiko, Rhodes Island, Greece. The uppermost notch, at about $+3.7 \text{ m}$ in elevation, developed with a stable relative sea level until ca. $4.9 \text{ }^{14}\text{C ka BP}$, when it was uplifted almost 2 m coseismically. A sequence of six up-and-down coseismic vertical displacements followed, leaving a well-recognizable series of ripple notches (Pirazzoli et al., 1989). Local tidal range is usually less than 0.2 m (photo 4550, May 1978).

Erosion Benches, Platforms, and Strand Flats

When the rock is easily eroded or the slope of a cliff is too gentle to allow an undercut to develop, marine erosion tends to excavate a bench, which, increasing in size, may later become an erosion platform.

The formation of coastal benches and platforms is usually ascribed to the removal by waves of the weathered part of cliffs.



Figure 3 The tidal notch, developed all around a limestone block, has transformed the block into a mushroom rock. South coast of Okinawa Island (Japan), between Mabuni and Gushikami. The retreat point of the notch is at about 2.5 m above present m.s.l. (local spring tidal range: 1.7 m). This 2.5 m uplift is ascribed to a single coseismic vertical displacement dated about $2.35 \text{ }^{14}\text{C ka BP}$ (Kawana and Pirazzoli, 1985; photo 6610, March 1981).



Figure 4 This tidal notch, developed around a middle Pleistocene coral limestone block, projecting from the uplifted surface of the last Interglacial terrace (MIS 5e, minimum age), was formed during the last Interglacial on the northwest coast of Barbados (Schellmann and Radtke, 2004; photo G853, October 2002).

The lowest level of possible weathering probably corresponds to that of constant immersion in seawater, usually in the intertidal zone. Wave abrasion is effective in most rock formations and may occur at any elevation between the highest levels reached by storm waves and the maximum depth at which sediments can be moved by waves. Most often, when a cliff is retreating, it leaves an erosion platform sloping gently seaward in the intertidal zone. The average elevation of shore platforms usually increases with rock hardness. Certain platforms may also develop above high tide level when water supplied by waves can stagnate behind an outer rampart. In high latitudes, strand flats (Klemsdal, 2005) are low shore platforms where disintegrated rock material, resulting from freeze–thaw alternations, has been washed away by waves.



Figure 5 The Kawamine Terrace, developed at about 70–80 m in altitude in the southwestern part of Kikai Island (Ryukyus, Japan), is an erosion surface cut into middle Pleistocene units during an undated high interstadial later than isotope stages 5a (stage 3?; Ota and Omura, 1992). Below, raised Holocene coral-reef terraces (capped by vegetation and villages) can be seen (photo 2221, November 1974).

The accuracy in the reconstruction of past sea-level positions from elevated or submerged benches and platforms will therefore increase when the tidal range is small and will be improved for solid calcareous rocks liable to be attacked not only by physical and chemical processes, but also by bioerosion. In all cases, the reliability of erosion coastal benches, platforms, and strand flats as sea-level indicators will depend mainly on the identification and understanding of the processes which were active in the development of these coastal features.

Purely wave-cut platforms are generally not datable. This can be a problem in areas characterized by very active tectonic movements, like Kikai Island, which is very close to the Ryukyu Trench (about 80 km) and where boundaries of terraces of different ages usually coincide with active faults, some of which also displace a single terrace (Figure 5). Relative sea-level changes may also produce multiple stepped benches. In this case, even ignoring the precise process of bench formation, a general rule to estimate sea-level change is to refer the inactive features, when possible, to the elevation of the present-day active counterpart (Figure 6).

Honeycombs, Tafoni, Pools, and Potholes

Honeycombs are small cavernous features assuming a cell-like structure. These patterns, which may be due to several physical processes, initially develop as many shallow depressions, but continued development produces deep chambers that are separated by thin walls of unweathered rock (Mustoe, 2005). Honeycombs have been reported as occurring in a variety of medium- to coarse-grained rocks, such as sandstones, granites, lavas, conglomerates, gneiss, schists, and limestone. Tafoni are cavernous weathering features of greater size (from a decimeter to several meters) developed on the rock surface. Since honeycombs and tafoni are typically found not only in coastal outcrops but also in inland deserts, their use as approximate sea-level indicators should always be confirmed by other kind of evidence.



Figure 6 A three-stepped bench can be observed at cape Plitri, on the west coast of Corfu Island, Greece. The elevation of the steps is +1.6 m, +0.8 m, and the present sea level (in the absence of wind, the local tidal range is almost negligible). Of the two raised benches, the uppermost one is partly jagged by wave projections and is related to a contemporary Holocene tidal notch that developed between at least 3030–2750 and 790–400 cal years BC, when a coseismic uplift of 0.8 m occurred. The intermediate bench (on which an observer is standing) developed since that time, until it was also uplifted coseismically at a more recent, indeterminate date. The third, lower bench is developing at the present sea level (Pirazzoli et al., 1994; photo D167, June 1991).

Erosion pools are flat-bottomed depressions, frequently found on limestone benches. They may result from solution and/or bioerosion processes in the mid-littoral and supralittoral zones. Elevated erosion pools on sheltered coasts can be used as indicators of past sea levels, with an accuracy of the order of the local tidal range, whereas on exposed coasts they only indicate former supralittoral areas in the reach of waves.

Erosion pools should not be confused with bioconstructed pools (see below).

Shore potholes are more or less rounded depressions, generally with greater depth than width, scoured by abrasion into a rocky shore, where sand, gravel, pebbles, and even boulders are circulated by wave motion. On exposed coasts, potholes may be formed at various levels, ranging from some meters above high tide to well below low tide; they are therefore inaccurate indicators of past sea-level positions.

Sea Caves and Arches

Sea caves are cavities excavated by marine erosion into a cliff in the range of wave action. Sea caves generally occur in the weaker parts of a rock formation, following soft layers, joints, faults, breccias, shale beds, unconformable strata, irregular sedimentation, or internal structures of lava flows. They are generally inaccurate sea level indicators. In limestone formations, however, sea caves are often of karstic origin, enlarged or reworked by the sea; their floor, if regular and flat, may be related to a former low tide position (Figure 7).

A sea cave opening to the sea on both sides of a headland, island or stack, beneath a connecting bridge, is a sea arch (Figure 8).

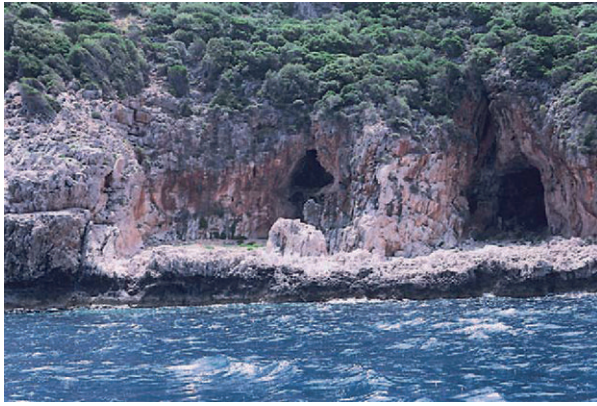


Figure 7 The floor of two abandoned sea caves extends to a bench at about 1 m above present sea level on the Yeditepe coast, south of the Orontes delta, Turkey. This bench was formed close to sea level, until it was uplifted by a coseismic uplift in the sixth century AD. This example shows clearly that the floor of sea caves cut into a hard limestone formation, which tends to the same level as the bench, may have a value as paleosea-level indicator similar to that of the nearby erosion bench (photo A745, May 1988).



Figure 8 Sea arches cut in a Miocene conglomerate at Cape Metsu in Tanabe Bay, Kii Peninsula, Japan. A double arch, called Megane-iwa (rock spectacles) is visible above a partially dissected bench resulting from a late Holocene sea level slightly higher than at present. The base of the two arches is at +3.6 and +2.8 m, respectively, above present m.s.l. (local spring tidal range: 1.46 m), suggesting that the arches have been formed at a slightly higher sea level. Below the higher arch, a third lower arch (not visible in the photo) exists. As the ceiling of its opening is at +0.1 m and the bottom at -0.9 m, it must be in formation at the present sea level (Takahashi, 1973; photo 553, April 1974).

Spurs, Grooves, Furrows, and Surge Channels

The occurrence of spur and groove systems has been reported from the upper part of the outer slope in most coral-reef areas. They consist of spurs a few meters high aligned more or less perpendicular to the reef front, alternating with almost bare grooves (Figure 9). The composite origin of these remarkable geomorphologic features (probably of growth for the spurs, erosion for the grooves) is still a matter of debate. Smaller-size furrows and surge channels into the outer bio-organic ridge of the reef occasionally extend grooves. Such features, when



Figure 9 Spurs and grooves are visible from the air near the southern reef of Takapoto Atoll, Tuamotus, French Polynesia (photo 5455, March 1980).



Figure 10 A series of straight surge channels, almost parallel to each other and perpendicular to the reef front, is partially emerged on the northern coast of Kikai Island (Ryukyus, Japan). Such emergence is not a normal feature and testifies of the recent uplift of the island, that has reached about 10 m in four steps occurred around 6065, 3250, 2700, and 1700 ^{14}C years BP, respectively (Nakata et al., 1979). The local tidal range is about 1.3 m and water level is at -0.3 m (photo 2214, November 1974).

recognized in fossil form, may be useful in the reconstruction of relative sea-level changes, though with a limited accuracy (± 1 m at best; Figure 10).

Bioeroded Holes

Borer organisms often leave morphological marks that can be used as sea-level indicators. As they can usually live at any depth in the sublittoral zone, they will indicate a sea level higher or equal than their marks and the same can be said for erosion features left by boring sponges (*Cliona*) and by sea urchins. However, when the elevated upper limit of burrows (e.g., of *Lithophaga* in the Mediterranean) is clearly defined

and forms a horizontal line, this corresponds to a former biological m.s.l. (Laborel and Laborel-Deguen, 1994), which can even be dated by radiocarbon assay when the articulated shells of the burrowing animals have been preserved inside the burrows.

Sea-urchin niches are small, rounded pools, that have been excavated by sea urchins at any depth in the sublittoral zone. When found emerged, such niches give evidence of relative sea-level fall, but confusion should be avoided with similar small basins that may have a meteoric origin or correspond to an initial phase of erosion pool development.

Depositional Features Associated with Erosion Marks

Though erosion features often make possible the estimation of past sea-level positions, they are usually insufficient to provide age estimates, for which adequate organic datable remains are essential. Fortunately, some depositional processes that contribute to shape certain geomorphic sea-level indicators can leave datable remains.

Marine Terraces

A marine terrace is any relatively flat surface of marine origin, bounded by a steeper ascending slope on one side and by a steeper descending slope on the opposite side. Marine terraces may result from marine abrasion or denudation (marine-cut terraces, or shore platforms, see Figure 5), or consist of shallow water to slightly emerged accumulations of materials removed by shore erosion (marine-built terraces), or also have a polygenic origin, with occurrence of *in situ* deposits after an abrasion phase (Figure 11).

In glacio-isostatically uplifting regions, small glacio-marine deltas situated near the margin of retreating ice sheets may be



Figure 11 A wide, flat marine terrace extends for many kilometers east of Chah Bahar, Iran, especially between Beris (photo) and the Pakistani border. This terrace is bordered landward by a series of higher marine terraces. It is an abrasion Pleistocene terrace at about +40 m, cut into Tertiary marls and capped by a 1–3 m thick layer of organogenic sandy limestone of coastal environment (panchina), containing marine shells and other debris cemented by calcium carbonate. Still undated (except than by radiocarbon estimates ≥ 28 ka), it can be ascribed tentatively to the last Interglacial cycle or to a polycyclic late Quaternary period (photo E123, May 1994).

preserved as flat surfaces related to the sea-level position at the time of their development (Andrews, 1986; Figure 12).

The occurrence of a series of stepped marine terraces usually results from eustatic changes in sea level superimposed on a tectonic trend. Uplifted terraces that act as a continuous tape recorder, each step developing when the rising sea level overtakes the rising land, may be most useful for Quaternary environmental studies, since each terrace can be considered as a fossil counterpart of the present-day terrace, platform, or reef flat. In seismic areas, however, terrace steps may have a tectonic origin (Figure 13).

When the uplift rate is sufficiently rapid (>1 mm year⁻¹), each major marine terrace may correspond to a different interglacial period or stage. In this case, ages increase with elevation, the upper levels also being the less well preserved. For slower



Figure 12 The lower five islands delta (Bay of Fundy, Canada) is a glacio-marine delta, dated 14–13 ka BP, which was subsequently uplifted by glacio-isostatic movements (photo A553, July 1987).



Figure 13 Five Holocene terrace steps can be observed on the northern coast of the Beagle Channel, at Playa Larga near Ushuaia (Argentina). In each of the four lower steps (marked 1–4), shell debris corresponding to beach deposits provided age estimates of about 405 years BP for terrace 1 at +1.6 m, 3095 years BP for terrace 2 at +3.8 m, 4335 years BP for terrace 3 at +5.2 m, and 5615 years BP for terrace 4 at +7.5 m (Gordillo et al., 1992). A clear relative sea-level fall occurred during the Holocene. According to the stepped pattern, coseismic uplifts with a return time of the order of 1.5 ka can be deduced for this area relatively protected from storm surges. Spring tidal range is 1.24 m (photo D77, November 1990).

uplift rates, major marine terraces may have a polycyclic origin, with sea level coming back at the terrace level after a period of exposure to weathering.

The age of a Pleistocene terrace is generally deduced from radiometric dating of marine organisms collected in growth position. U-Series and electron spin resonance (ESR) are the most frequently used dating methods for Pleistocene marine terraces, and radiocarbon for Postglacial and Holocene terraces. From a geomorphologic point of view, the best sea-level indicator of a terrace is the limit between the upper part of the terrace and the adjacent cliff base, especially if it is marked by a notch, because it can be correlated with the peak of the corresponding eustatic transgression. Its accuracy as sea-level indicator depends on the local tidal range and exposure to wave action.

Marine deposits left in uplifting regions by a tsunami (like the tsunami produced 7.1 radiocarbon ka by the Storegga Slide on Northern Europe coasts) may provide accurate stratigraphical indicators, enabling the estimate of the change in the relative sea level since the tsunami event (Smith et al., 2004).

Correlations between marine terraces and sea-level changes have been attempted for the last 300 ka in Papua New Guinea (Chappell, 1983), the last 400 ka in Barbados (Schellmann and Radtke, 2004), the last 680 ka in New Zealand (Pillans, 1983), and even the last million years in Sumba Island, Indonesia (Pirazzoli et al., 1993).

Algal Rims, Microatolls, Trottoirs, Cornices, and Surf Benches

Algal rims are reef-like structures developing on the outer edge of reef flats (mainly Porolithon) or on the rocky substrates on coasts subjected to regular wave action (Laborel, 2005). On reef flats, algal rims generally produce a slightly higher ridge, easy to recognize morphologically. In subtropical areas, algal rims (mainly Neogoniolithon) may contribute, together with the vermetid *Dendropoma*, to the development of trottoirs and even of microreefs growing up to sea-level, where they tend to extend laterally. They are very accurate (up to ± 0.1 m) sea-level indicators and easy to date by radiocarbon. Uplifted fossil forms enable detailed studies of algal-rim evolution in relation to sea level (Figure 14).

Microatolls are circular organic reef structures (diameter 1–6 m) consisting of a slightly raised rim built by coral (usually *Porites*), algae, vermetids, and other organisms surrounding a shallow depression on the shore or on a coral reef (Bird, 2005). When found in elevated position, in areas where moating phenomena are not possible, coral microatolls are excellent indicators of former low tide level (Figure 15).

Bioconstructed pools should not be confused with erosion pools. In the former, the walls of the basins are bioconstructed by vermetids or calcareous algae at the uppermost limit of the sublittoral zone (Figure 16). When preserved, elevated bioconstructed pools are excellent paleosea-level indicators, easy to date with radiocarbon, with an accuracy of up to ± 0.1 m.

The term trottoir, French for sidewalk, designates a simple crust of vermetids or calcareous algae, up to several centimeters thick, covering a rocky substratum, and forming a continuous, almost horizontal passage at sea level, one to several meters wide, between the rocky coast and the sea (e.g., see Figure 14). It is an erosion feature partly shaped by bioconstruction,

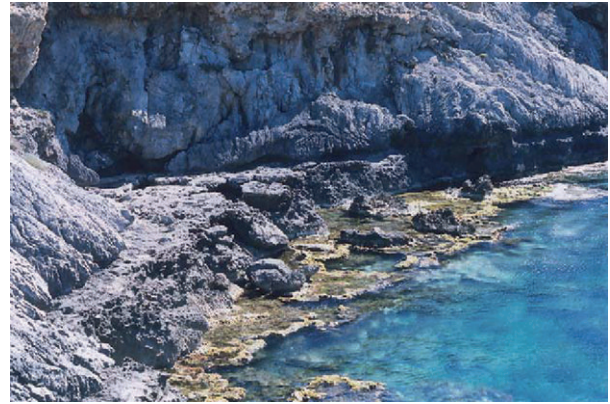


Figure 14 A coseismic uplift affecting all the western part of Crete (Greece) occurred in the fourth century (probably on 21 July 365) AD. At Damnoni, on the south coast of the island, the vertical displacement was of about 1.6 m. Algal reefs and trottoirs were suddenly emerged. Since that time, new algal reefs have developed at the present sea level. A comparison between the elevated and the active reefal and morphological counterparts gives evidence of the excellent accuracy of this kind of sea level indicator (photo G479, June 2000).



Figure 15 These two microatolls of *Porites* in growth position on the barrier reef of Bora Bora (Society Islands, French Polynesia) show, with ± 0.1 m accuracy, a former spring low tide position about 0.8 m higher than at present, that lasted until less than 2 ka (Pirazzoli and Montaggioni, 1988; photo 7666, October 1982).

because vermetids or calcareous algae that protect the substratum against further marine erosion rapidly cover the rocks at sea level. It is a very accurate (up to ± 0.1 m) sea-level indicator easy to date by radiocarbon.

The term trottoir has also been used in the literature to designate a purely bioconstructional overhang up to some decimeters wide (that should more appropriately be called a cornice), that is, a narrow intertidal reef construction made by *Lithophyllum*, *Dendropoma*, coralline algae, and also ostraids occasionally. Trottoirs are relatively frequent in the Mediterranean and in the tropics. Cornices have been reported from even wider areas. They are both excellent sea level indicators, especially in microtidal areas, and have been widely used to reconstruct and date relative sea-level changes or uplift movements during the Holocene, especially in seismic areas.



Figure 16 The outer rim of these active pools with microatoll shape is formed by *Dendropoma* and other vermetids, which are regularly wetted by waves on the front of the fringing reef of Tubuai, Austral Islands, French Polynesia. Local tidal range is 0.6 m and water level is near m.s.l. (photo 7855, October 1983).



Figure 17 Active surf bench cut into Quaternary reef limestone at 0.5 m above the high tide level near Cape Ara, Okinawa, Japan. Local tidal range is 1.7 m and water level is 0.3 m below m.s.l. (photo 6590, February 1981).

A special case is that of surf benches that occur mainly on limestone coasts exposed to persistent surf action. Here organic incrustations may develop on the floor of a tidal notch or near the outer edge of a bench, protecting the substrate rock and thus, locally inhibiting erosion. While erosion will proceed above the accretion level, a surf bench will begin to form, and may become several meters wide (Figure 17). Surf benches as high as 2 m above high tide level have been reported in the literature but are not horizontal; their elevation reaches the highest levels on the most exposed sites of headlands, declining when wave action decreases (Focke, 1978). As with erosional benches, reference to the present-day counterpart in the same place is useful, when possible, to assess relative sea-level changes (e.g., Figure 14).

Reef Flats, Coral Conglomerates, Beachrocks, and Speleothems

A reef flat consists mainly of a pavement cemented by calcareous algae and incorporating various detrital elements (coral



Figure 18 Boundary area between the active reef flat and a coral conglomerate flagstone about 1 m high at Temoe Atoll, Gambier Islands, French Polynesia. In this remote and presently uninhabited island, close to the southernmost limit of coral-reef growth in the South Pacific, a relative sea-level fall of about 0.8 m took place during the last 2 ka. In the case of Temoe Atoll, petrological analysis was not necessary to define the position of the former spring low tide level, because this level was locally marked by an almost horizontal fissure at about 0.5 m above the present m.s.l., separating the ancient reef flat with corals in growth position from the above conglomerate flagstone (Pirazzoli, 1987; photo 7470, October 1982).

fragments, Foraminifera, and *Halimeda*). A reef flat is initially the top of a reef framework, exposed to the atmosphere at low spring tides. In a second episode, skeletal detritus, torn away from the outer reef by storm waves, fills in depressions and irregularities in the reef framework. Lithification of the resulting coral conglomerate usually starts in the subtidal zone, the intergranular spaces being gradually occluded with marine phreatic (water in the zone of saturation) cements. When the deposits reach the intertidal zone, intergranular cements mainly occur in the form of marine vadose (above the zone of saturation, in the zone of aeration) types. When elevated reef flats still contain reef organisms in growth position, a former spring low tide level can be deduced from their surface. On the other hand, when they consist of coral conglomerate (Figure 18), an accurate estimation of the relative sea-level change usually requires petrological analysis. The method of investigation (Montaggioni and Pirazzoli, 1984) consists of sampling vertical sections at several levels and identifying the sections on the former environment (marine phreatic or marine vadose) of the first generations of intergranular cements. Once the boundary between the two former environments of cementation has been determined, the amount of relative sea-level change can be easily estimated from the difference in level between this boundary and the present low tide position.

The term beachrock applies to beach sediments deposited within the intertidal zone and cemented by calcium carbonate, especially in warm waters (Russel and McIntire, 1965; Figure 19). A beachrock is generally a good indicator of sea-level (with a vertical uncertainty depending on the tidal range). However, radiocarbon ages of beachrock samples should be interpreted with caution, because they may be reworked deposits, older than the beach lithification.



Figure 19 Elevated slabs of beachrock partly colonized by vegetation on the southern coast of Tunisia near Lella Meriem, north of Zarzis, testify of an intertidal zone higher than at present. Holocene emergence, probably of hydro-isostatic 5000 ^{14}C years BP, gradually declining afterwards to the present situation (Morhange and Pirazzoli, 2005). Water level is at high tide. Beach deposits near this level have been dated about 2000 years BP (photo G878, November 2002).

Speleothems from karst systems submerged by the sea can also provide former sea-level indications, especially when they are capped by a marine biogenic cover: the U-series dating of the outer part of the speleothem at a certain depth will indicate that the sea-level was still below that depth at that time, whereas the date of the beginning of the marine bioconstruction will approximate the time when the sea rose above the biogenic overgrowth. With this kind of indicator, useful late Pleistocene sea-level reconstructions have been provided from Florida, the Bahamas, and recently also from the Mediterranean (Antonioli et al., 2004).

See also: U-Series Dating. **Glaciation, Causes:** Tectonic Uplift and Continental Configurations. **Paleoceanography, Biological Proxies:** Biomarker Indicators of Past Climate. **Radiocarbon Dating:** AMS Radiocarbon Dating; Conventional Method. **Sea Level Studies:** Use of Cave Data in Sea-Level Reconstructions.

References

- Andrews JT (1986) Elevation and age relationships: Raised marine deposits and landforms in glaciated areas. In: Van de Plassche O (ed.) *Sea-Level Research: A Manual for the Collection and Evaluation of Data*, pp. 67–95. Norwich: Geo Books.
- Antonioli F, Bard E, Potter E-K, Silenzi S, and Improta S (2004) 215-ka history of sea-level oscillations from marine and continental layers in Argentarola cave speleothems (Italy). *Global and Planetary Change* 43: 57–78.
- Bird E (2005) Appendix 5: Glossary of coastal geomorphology. In: Schwartz ML (ed.) *Encyclopedia of Coastal Science*, pp. 1155–1192. Dordrecht: Springer.
- Chappell J (1983) A revised sea-level record for the last 300,000 years from Papua New Guinea. *Search* 14(3–4): 99–101.
- Fairbridge RW and Hillaire-Marcel C (1977) An 8,000-yr palaeoclimatic record of the 'Double-Hale' 45-yr solar cycle. *Nature* 268: 413–416.
- Focke JM (1978) Limestone cliff morphology on Curaçao (Netherlands Antilles), with special attention to the origin of notches and vermetid/coralline algal surf benches. *Zeitschrift für Geomorphologie* 22: 329–349.
- Gordillo S, Bujalesky GG, Pirazzoli PA, Rabassa JO, and Saliège JF (1992) Holocene raised beaches along the northern coast of the Beagle Channel, Tierra del Fuego, Argentina. *Palaeogeography, Palaeoclimatology, Palaeoecology* 99: 41–54.
- Kawana T and Pirazzoli PA (1985) Holocene coastline changes and seismic uplift in Okinawa Island, the Ryukyus, Japan. *Zeitschrift für Geomorphologie Supplementband* 57: 11–31.
- Klemsdal T (2005) Strandflat. In: Schwartz ML (ed.) *Encyclopedia of Coastal Science*, pp. 914–915. Dordrecht: Springer.
- Laborel J (2005) Algal rims. In: Schwartz ML (ed.) *Encyclopedia of Coastal Science*, pp. 24–25. Dordrecht: Springer.
- Laborel J and Laborel-Deguen F (1994) Biological indicators of relative sea-level variations and of co-seismic displacements in the Mediterranean region. *Journal of Coastal Research* 10: 395–415.
- Laborel J, Morhange C, Collina-Girard J, and Laborel-Deguen F (1999) Littoral bioerosion, a tool for the study of sea-level variations during the Holocene. *Bulletin of the Geological Society of Denmark* 45: 164–168.
- Montaggioni LF and Pirazzoli PA (1984) The significance of exposed coral conglomerates from French Polynesia (Pacific Ocean) as indicators of recent relative sea-level changes. *Coral Reefs* 3: 29–42.
- Morhange C and Pirazzoli AP (2005) Mid-Holocene emergence of southern Tunisian coasts. *Marine Geology* 220: 205–213.
- Mustoe G (2005) Honeycomb weathering. In: Schwartz ML (ed.) *Encyclopedia of Coastal Science*, pp. 529–530. Dordrecht: Springer.
- Nakata T, Koba M, Jo W, Imaizumi T, Matsumoto H, and Suganuma T (1979) Holocene marine terraces and seismic crustal movement. *Science Reports of Tohoku University, Geography* 29(2): 195–204.
- Ota Y and Omura A (1992) Contrasting styles and rates of tectonic uplift of coral reef terraces in the Ryukyu and Daito Islands, southwestern Japan. *Quaternary International* 15/16: 17–29.
- Pillans B (1983) Upper Quaternary marine terrace chronology and deformation, South Taranaki, New Zealand. *Geology* 11: 292–297.
- Pirazzoli PA (1986) Marine notches. In: Van de Plassche O (ed.) *Sea-Level Research: A Manual for the Collection and Evaluation of Data*, pp. 361–400. Norwich: Geo Books.
- Pirazzoli PA (1987) A reconnaissance and geomorphological survey of Temoe Atoll, Gambier Islands (South Pacific). *Journal of Coastal Research* 3(3): 307–323.
- Pirazzoli PA (1996) *Sea-Level Changes – The Last 20000 Years*. Chichester: Wiley.
- Pirazzoli PA (2005) Sea-level indicators, geomorphic. In: Schwartz ML (ed.) *Encyclopedia of Coastal Science*, pp. 836–838. Dordrecht: Springer.
- Pirazzoli PA and Montaggioni LF (1988) Holocene sea-level changes in French Polynesia. *Palaeogeography, Palaeoclimatology, Palaeoecology* 68: 153–175.
- Pirazzoli PA, Montaggioni LF, Saliège JF, Segonzac G, Thommeret Y, and Vergnaud-Grazzini C (1989) Crustal block movement from Holocene shorelines: Rhodes Island (Greece). *Tectonophysics* 170: 89–114.
- Pirazzoli PA, Radtke U, Hantoro WS, et al. (1993) A one million-year-long sequence of marine terraces on Sumba Island, Indonesia. *Marine Geology* 109: 221–236.
- Pirazzoli PA, Stiros SC, Laborel J, et al. (1994) Late Holocene shoreline changes related to papaeoseismic events in the Ionian Islands, Greece. *The Holocene* 4(4): 397–405.
- Russell RG and McIntire WG (1965) Southern hemisphere beach rock. *Geographical Review* 55: 17–45.
- Schellmann G and Radtke U (2004) The marine Quaternary of Barbados. *Kölner Geographische Arbeiten* 81: 1–137.
- Smith DE, Shi S, Cullingford RA, et al. (2004) The Holocene Storegga Slide tsunami in the United Kingdom. *Quaternary Science Reviews* 23: 2291–2321.
- Takahashi T (1973) Formation and evolution of shore platforms around southern Kii Peninsula. *Science Reports of Tohoku University, Geography* 23: 63–89.
- Torunski H (1979) Biological erosion and its significance for the morphogenesis of limestone coasts and for nearshore sedimentation (Northern Adriatic). *Senckenbergiana Maritima* 11: 193–265.
- Trudgill ST (1976) The marine erosion of limestones on Aldabra Atoll, Indian Ocean. *Zeitschrift für Geomorphologie Supplementband* 26: 164–200.

Sedimentary Indicators of Relative Sea-Level Changes – High Energy

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Introduction: Definition of High-Energy Coasts

Coasts exposed to high levels of wave energy are routinely termed 'high-energy coasts.' Precise definition of the term is difficult, however, as the distinction between high- and low-energy coasts is entirely relative. For example, the coasts of Washington and Oregon experience higher levels of wave energy than neighboring California, although all of these would be viewed as high-energy environments in comparison with the sheltered coastline of Chesapeake Bay. In common usage, the term is used to distinguish wave-exposed coastlines (typically those facing the open ocean) from wave-sheltered coastlines (those found in embayments, back-barrier lagoons, semi-enclosed seas, and similar settings). Because of the high levels of wave energy being dissipated on such coasts, fine-grained sediments (clay and silt) are rare and deposition is usually limited to sand, gravel, and boulders.

High-energy coasts may contain relict deposits of low-energy settings that have now become exposed through coastal reorientation and/or sea-level changes (e.g., back-barrier sediments exposed on the foreshore). These features will not be considered here. Instead, the discussion will focus on those sedimentary features formed under high-energy conditions that provide an indication of sea level during their formation.

Often the sediments of high-energy coasts are exposed in association with those of contemporary adjacent low-energy sediments (e.g., washovers deposited into lagoonal muds). In such circumstances, the organic-rich sediments of sheltered environments are usually favored for the study of former sea levels because of their close relationship to particular tidal levels, the lower vertical range of dynamic activity (see below), and the abundance of dateable *in situ* organic material.

Nondepositional high-energy coasts may also contain sedimentary indicators of former sea levels in the form of erosional features produced by wave action and occasional coarse-grained sediments in coastline reentrants. Organic deposits occur on some low-latitude, high-energy coasts in the form of coral reefs.

Sedimentary processes and the resulting depositional or erosional features on high-energy coasts are highly variable in time and space. In order to relate Quaternary deposits to the sea level at which they were deposited, studies of contemporary sedimentary environments are necessary in order to identify the past environment of deposition and its relationship to present sea level.

Sea-Level Indicators on High-Energy Coasts

In assessing the record of past sea level, any potential indicator must exist in a known or predictable relationship to sea level at the time of deposition. If a chronological framework is to be

established it must also contain some dateable material incorporated within the sediment. For ancient sedimentary deposits to be correctly interpreted, a lithofacies model is therefore required.

A feature of high-energy coasts is that sedimentary environments may extend from many meters below to many meters above contemporary sea level ([Figure 1](#)), thus potentially reducing the vertical precision of sea-level indicators in this environment. Most sedimentary activity takes place in the zones of wave energy dissipation, particularly in the surf and swash zones. These zones, however, migrate laterally and vertically according to the nature and degree of wave energy being delivered to them. A surf zone typically becomes wider and extends further seaward during a storm compared to more typical conditions. The associated surf zone sediments would also be expected to adjust accordingly. Extremes of sedimentary activity in the vertical dimension are associated with very high-energy events (frontal storms, hurricanes, and tsunami).

The distance below sea level to which coastal deposition extends depends on wavelength (longer waves produce sediment movement at greater depths) while the landward extent of sedimentation is linked to wave run-up, which may be dramatically increased during storms. Thus, care must be taken in interpreting the origin of sedimentary deposits on high-energy coasts as storm or fair-weather-related features. In some instances, it has not been possible to unambiguously assign a storm, tsunami, or fair-weather origin to such deposits (e.g., [Switzer et al., 2009](#)). For example, controversy exists regarding the interpretation of the Hulopoe Gravels on Molokai, Hawaii, as normal shoreline sediments or a tsunami deposit ([Dawson and Shi, 2000](#)).

The large vertical range of activity also means that past indicators of former sea levels may occur within the range of contemporary coastal processes. This renders them vulnerable to modification by erosion or contemporary deposition. In a review of previous studies of supposed raised coastal landforms in eastern Ireland, [Carter \(1983\)](#) concluded that there was no unequivocal evidence of a higher than present sea level: all the features described lay within the range of modern storm activity and alternative explanations (surge, high waves, and tidal range changes) existed for all the cited evidence of higher than present sea-level features in earlier studies.

In contrast, high-level storm-related deposits are formed above the range of most contemporary processes and therefore have an enhanced preservation potential. It is thus important to distinguish them from deposits of more typical energy conditions. While salt marsh deposits are often regarded as good indicators of tidal levels in low-energy settings, [Cooper and Power \(1993\)](#) identified active salt marsh deposits on a high-energy rock coast at levels up to 9 m above contemporary sea level, where they were sustained by a combination of freshwater seepage and salt spray from large waves.

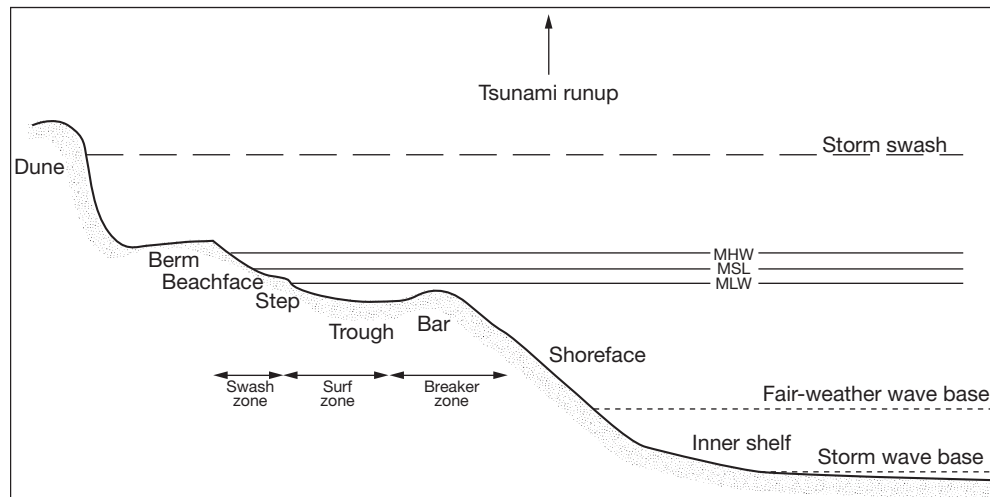


Figure 1 Indicative distribution of coastal depositional environments in relation to tidal reference levels.

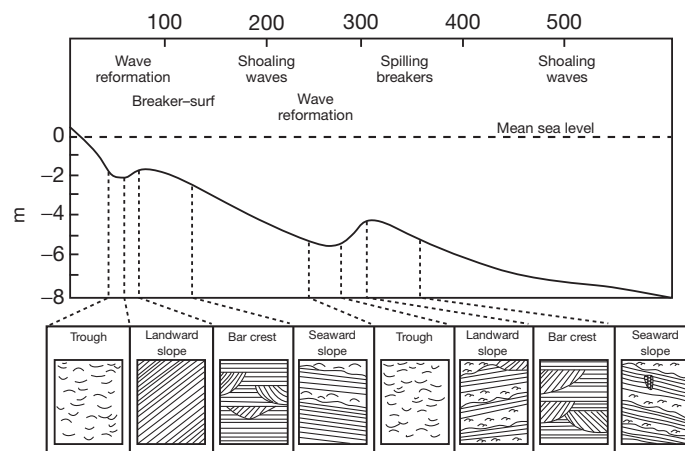


Figure 2 Cross-shore distribution of sedimentary facies in relation to wave energy zones on a barred sandy coast. Reproduced from Davidson-Arnott RGD and Greenwood B (1976) Facies relationships on a barred coast, Kouchibouguac Bay, New Brunswick, Canada. In: Davis RA (ed.) *Beach and Nearshore Sedimentation*, pp. 149–168. Tulsa, OK: SEPM, Special Publication No. 24.

Shoreface-Inner Shelf Facies

Identification of former sea level thus relies on the identification of sediments deposited in a known relationship to sea level at the time of deposition. A number of classic studies of the 1970s and early 1980s (e.g., Clifton et al., 1971; Davidson-Arnott and Greenwood, 1976; Shipp, 1984) described the sedimentary facies of contemporary coastal deposits (primarily barred and nonbarred sand beaches) in different environmental settings at varying scales of resolution. The deposits of these systems typically extend from the subaerial dune and back beach, across the shoreface to the inner shelf. The model of Clifton et al. (1971) related the sedimentary facies of the nearshore environments primarily to the degree of asymmetric flow produced by shoaling waves in the nearshore zone. In deeper water, small ripples dominated the seabed, while in the zone of wave attenuation, large bedforms were produced before waves actually break. High current velocities produced planar sand facies in the breaker zone itself and in the surf zone another zone of large bedforms existed. Swash action at the

intertidal beach was associated with planar, seaward-dipping bedding. Shipp (1984) identified distinctive upper shoreface, longshore trough, longshore bar, landward slope bar crest and seaward slope facies on the shoreface of a high-energy coast. These were followed seaward by a transition to the inner shelf (coincident with fair-weather wave base). Davidson-Arnott and Greenwood (1976) presented a detailed facies description of barred environments, subdividing them into bar crest, seaward slope, landward slope, and trough. Each facies had a distinctive suite of physical and biogenic sedimentary structures that was repeated at each bar (Figure 2).

From these site-based studies, generic models of shoreface-beach sedimentary facies were developed that comprise lower, mid, and upper shoreface, foreshore, overwash, and dune facies (Moslow, 1984). Although not explicit in this regard, these accounts of the nature of sedimentary facies in various depositional subenvironments provide a basis for assessing the relationship between a sedimentary facies and sea level during its deposition. The arrangement of facies can be used to

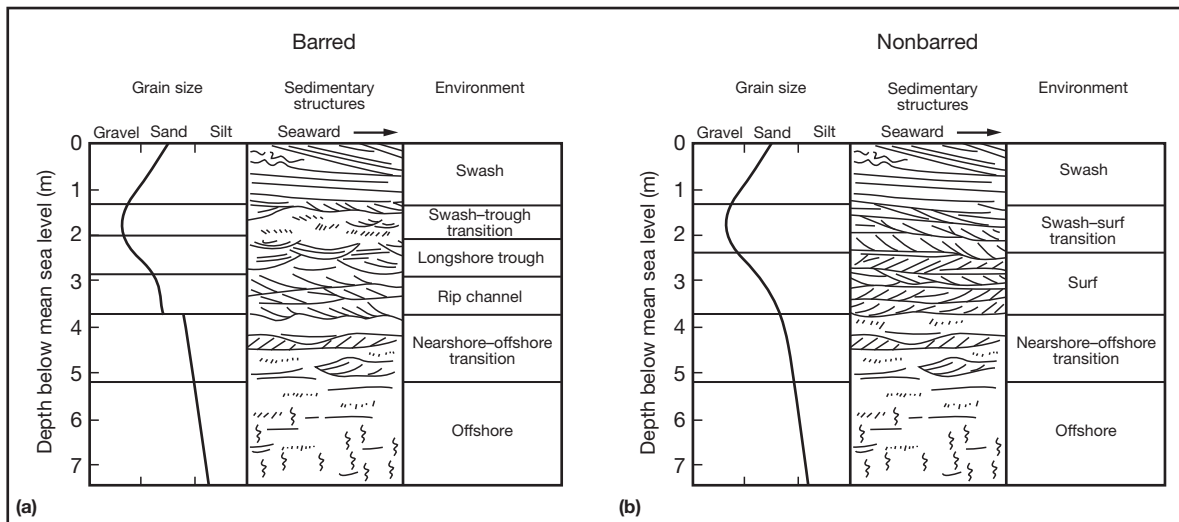


Figure 3 Comparative facies of barred (a) and nonbarred (b) sand beaches in relation to contemporary sea level. Reproduced from Hunter RE, Clifton HE, and Phillips RL (1979) Depositional processes, sedimentary structures and predicted vertical sequences in barred nearshore systems, southern Oregon coast. *Journal of Sedimentary Petrology* 49: 711–726.

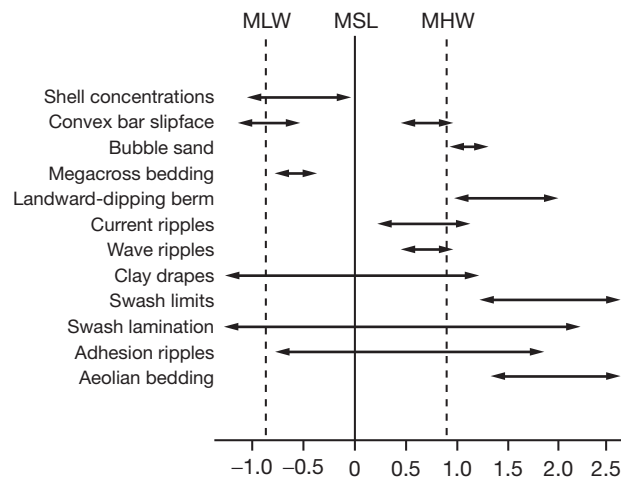


Figure 4 Relationship to sea level of various sedimentary structures on Netherlands beaches (based on information in Roep, 1986).

identify the type of depositional environment (Figure 3) and its relationship to others but it cannot readily identify the precise relationship to former sea level with accuracy better than a few meters.

Roep (1986) explicitly explored the relationship between beach sedimentary characteristics and sea level at three barrier beaches in the Netherlands. This was done using temporal analysis of profile variability on modern beaches coupled with careful assessment of contemporary sedimentary facies and individual sedimentary structures (Figure 4). A number of sedimentary indicators were noted to have a close relationship to contemporary sea level. The high water swash limit, for example, was found to be the best lithological marker in relation to sea level. This separates the seaward-dipping foreshore laminae from the landward-dipping backshore.

The study by Roep (1986) does not necessarily have universal applicability: the extent of specific features is likely to

vary from site to site depending on tidal range, wave climate, beach morphology, etc. The elevation of swash run-up, for example, will vary according to modal wave conditions and beachface gradient and composition. Nevertheless, the approach indicates the potential for refining estimates of former sea-level position from high-energy coasts by studying contemporary facies in equivalent settings.

Inlet-Associated Facies

Inlets on barrier-lagoon coasts contain a distinctive arrangement of sedimentary facies that may be preserved as an inlet migrates (Moslow, 1984). A number of studies have shown that inlet-associated sediments occupy substantial portions (up to 40%) of barrier-associated sedimentary records (Moslow, 1984). Because inlets are dominated by tidal currents which operate within a relatively fixed elevation range relative to the

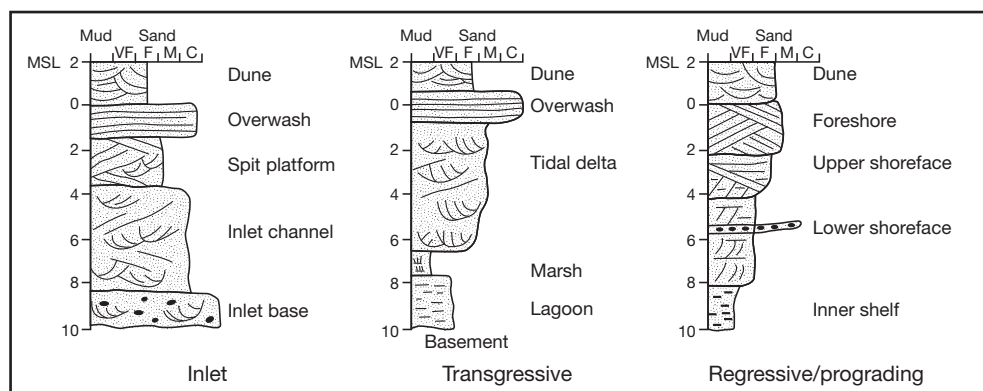


Figure 5 Inlet-associated sedimentary facies compared with transgressive and regressive facies arrangement on a sandy coast (based on Moslow, 1984).

tidal frame, the identification of particular inlet facies can be an aid in assessing past sea level.

Commonly, three sedimentary facies are associated with tidal inlets: the inlet floor, inlet channel, and spit platform (Figure 5). The inlet floor typically comprises a coarse lag deposit, the channel cross-bedded coarse sand and shell, and the spit platform, low angle, planar cross-bedded parallel laminations of sand and shell. The spit platform develops in the intertidal zone and is thus the best indicator of sea level. The channel facies relate to the size of the tidal prism and the erodibility of the underlying substrate and not specifically to sea level.

Gravel and Boulder Beaches

Coarse-grained beaches, which dominate glacial and paraglacial coasts as a result of erosion of gravel and boulders from diamictons, differ fundamentally from sand beaches in that their high permeability and high infiltration rates reduce surface flow, particularly of backwash. This eliminates or reduces the potential for seaward return of sediments and induces a net landward movement of gravel systems. Orford et al. (1991) provide a review of the characteristics of coarse-grained beaches in the context of Quaternary sea-level change. Coarse-grained barriers are therefore more likely to preserve a record of landward movement of sediment under high-energy conditions than of sea-level changes. The sedimentary facies of gravel beaches are less well studied than sand beaches, although Bluck (1967) presented a facies model based on grain shape and packing that enables identification of varying depositional environments on gravel beaches.

Rock Coasts

Erosional processes on high-energy rock coasts produce a range of features that potentially relate to sea level. The essential element of rocky coasts is the cliff-shore platform couplet in which a retreating cliff is associated with a remnant shore platform left behind as the higher part of the cliff is eroded. The relationship of shore platforms to sea level is highly variable and they have been noted to occur intertidally, supratidally, and subtidally, depending on the cliff lithology and the level and nature of wave activity operating on it.

Similarly, caves cut by waves that exploit zones of weakness and remnant sea stacks and arches all form close to sea level but have variable relationships to it. As sea-level indicators, they have a very large vertical error margin (several meters). Marine notches occur on cliffed coasts as a result of wave abrasion, biological processes, and chemical processes. Pirazzoli (1986) identified three types of notch in relation to sea level. Supratidal notches develop in the wave spray zone above high water. Their precision in identifying former sea level is relatively low. Subtidal notches may develop though bioerosion at a particular level and they may, if the life range of the responsible organisms is known, be more closely indicative of sea level. While intertidal notches can be formed by wave-induced erosion, on the more exposed parts of the coast they form at variable heights above sea level. These and several potential biogenic indicators including barnacle lines and corals, enable estimates of former sea level to be made with centimeter-scale accuracy (Pirazzoli, 1986).

Beachrock

Beachrock (Figure 6) is a carbonate-cemented sandstone that forms on the shoreline of tropical and warm temperate beaches. It can form relatively rapidly (within a few hundred years) and therefore provides a good potential indicator of sea-level position (Hopley, 1986). It is widely distributed on high-energy beaches around southern Africa, the Caribbean, and Australia. Its position within the intertidal zone renders beachrock a potentially good indicator of former sea level, particularly, if the tidal range is small. However, the precise levels at which beachrock forms are not well known and this reduces its potential vertical accuracy as a sea-level indicator Kelletat (2006). Hopley (1986) summarized the potential shortcomings of beachrock as a sea-level indicator: misidentification of true beachrock (it is often associated with aeolianite), the need to use the uppermost level of the beachrock and the possibility of derived shell material in whole rock radiocarbon age determinations. To these can be added the potential for slumping of whole blocks after deposition and reworking of old beachrock and incorporation into younger deposits. This underscores the need for additional facies analysis in assigning an indicative sea level. Nonetheless whole rock beachrock



Figure 6 Beachrock is common on warm temperate to tropical coasts. Beachrock at Umhlanga Rocks, KwaZulu-Natal province, South Africa (a) contains diagnostic lower beachface sedimentary structures (trough cross-bedding) and trace fossils. Shallowly submerged beachrock on the Datça Peninsula, Turkey (b) preserves a former shoreline position. Beachrock cementation can incorporate coarse beach clasts as well as sand as in this example (c) from Goana Island, British Virgin Islands.

dating was used to develop a high-resolution late Holocene sea-level curve for South Africa (Ramsay, 1995). Thomas (2009) used OSL dating of beachrock to avoid issues related to diagenesis in India. The early formation of beachrock has the added advantage for sea-level studies of lithifying and thus

preserving shoreline facies that might otherwise be eroded or migrate under a rising sea level (Cooper, 1991).

Sedimentary Response to Sea-Level Change

On depositional coasts, the sedimentary response to changing sea level is mediated by a range of important factors. Sediment supply can determine whether a depositional sequence accumulates. Under falling sea-level conditions this can lead to a distinctive series of laterally extensive shoreline facies. Even under rising sea level, an abundance of sediment can lead to an advancing shoreline (Roy et al., 1994). Sediment type has been noted to vary between terrigenous during lowstands and carbonate during highstands as fluvially incised valleys are drowned and terrigenous input is trapped in drowning valleys during transgression (Stolper et al., 2005).

The accommodation space, as mediated by antecedent topography, is a vital factor in determining the nature of any sedimentary sequence. It controls the availability of space in which sediments can accumulate and the capacity of accumulated sediments to be preserved. A particular aspect of the antecedent topography is the basement slope over which the plane of sea-level migrates. If steep, lateral migration rates will be slower, providing more time for development of shoreline equilibrium forms and slower rates of increase in accommodation space. If gently sloping, migration rates will be faster and shoreline features are more likely to be stranded on the shelf under transgression or on the coastal plain under regression. As the sea-level changes, so too will the accommodation space and the influence of inherited topographic features (Roy et al., 1994).

Because of the variability of these factors, there is, thus, to be expected a variable response to an apparently uniform sea-level change. Added to this, the rate of sea-level change itself is also an important determinant of coastal sedimentary response. Despite this complexity, a number of simplified models of shoreline response to changing sea level have been presented.

Orford et al. (1991) presented a model of gravel barrier platform evolution in response to changing sea level. The responses were mediated by sediment supply, accommodation space and rate of sea-level rise. A series of potential evolutionary paths were identified and sedimentary responses included barrier breaching, sealing, rollover, accretion, erosion, and overstepping.

Carter (1988) identified three simplified models of coastal profile response to sea-level rise (Figure 7). In the erosional response model, sediment eroded from above sea level is envisaged being transported seaward and retained on the shoreface where it enables the nearshore profile to be maintained. Such a model requires an erodible bluff at the rear of a beach that yields sand-sized material suitable for beach building. A similar theoretical construct is involved in the 'Bruun Rule' of shoreline erosion, a now-dated approach that has been shown to be a poor predictor of actual shoreline response (Cooper and Pilkey, 2004). Under the rollover scenario, the coastal barrier sediment volume is maintained, largely unchanged by the process of barrier overwashing of sand from the shoreface. Thus, translation of the barrier takes place as an essentially intact feature in a landward direction

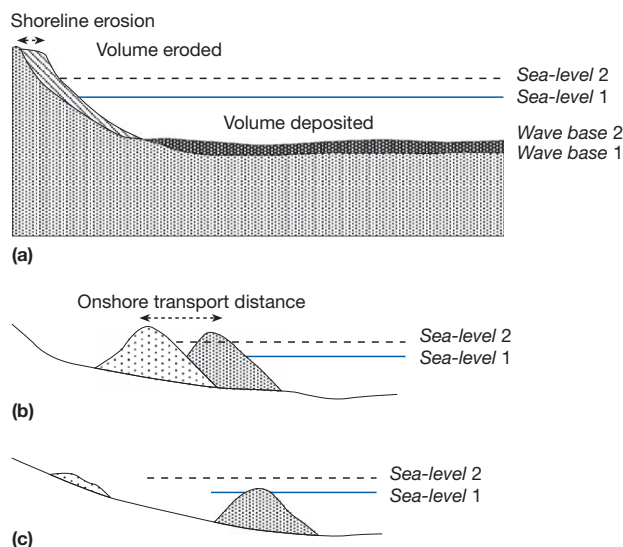


Figure 7 Three simple modes of response to sea-level rise. Reproduced from Carter RWG (1988) *Coastal Environments*. Cambridge: Cambridge University Press.

across the underlying basement slope. Barrier island migration has been documented along much of the US east coast. The third model 'barrier overstepping' is envisaged to occur when a beach is unable to respond to rapidly rising sea level by migration of erosion and is consequently drowned in place and preserved. Because of their slower morphological response times to sea-level rise, the probability of *in situ* drowning is higher for gravel than sand barriers (Carter, 1988).

These three models likely represent end-members or extreme cases. It is likely that combinations of their elements constitute the situation in nature. More sophisticated geomorphic-stratigraphic models of shoreline response to rising sea level have been developed (Roy et al., 1994; Stolper et al., 2005). While each of the models makes simplifying assumptions, they enable shoreline behavior under changing sea level to be simulated numerically and compared with observed sedimentary records in order to investigate the importance of the various controls outlined above.

Van Goor et al. (2003) assessed the potential response of tidal inlets to rising sea level and concluded that an inlet could persist if sedimentation within the inlet-associated environments kept pace with sea level. If not, the inlet could be drowned. Former inlets of the Netherlands Wadden Sea were believed to have been drowned during the Holocene transgression by sea-level rise rates of over 0.8 m per century.

Sedimentary Indicators of Relative Sea-Level Change: Transgression and Regression

There are abundant sedimentary records of trends in former sea-level trend. While these may not provide a precise estimate of sea level at a given time, they indicate the direction of relative sea-level movement. The sequence stratigraphy approach to sea-level change (see article in this volume) is based on the balance between the changing accommodation

space for sedimentation and the rate of sediment supply with which that space may be filled. Thus the vertical stacking of sedimentary facies and the nature of the contacts between them can be interpreted in the context of sea-level change. An account of the principles and application of sequence stratigraphy is provided elsewhere in this volume.

Transgressive situations occur over wide areas when sea-level rises. Local transgressive sequences can also be produced when shortages of sediment promote coastal erosion and consequent landward migration of the shoreline. Identifying regional and local patterns in coastal sequences is thus important in determining the nature of the driving forces of change. Regressive situations can occur where there is a regional fall in sea level that reduces accommodation space and causes a seaward migration of the coast. A similar regression can be caused by a local abundance of sediment. In the latter scenarios, a coastline can continue to accumulate sediment and extend seaward even under conditions of sea-level rise if there is an abundance of sediment. The essential characteristic of transgressive stratigraphic sections is the deposition of progressively deeper water facies over shallower water facies, whereas in regressive stratigraphies the reverse is the case (Figure 8). Investigations of transgressive and regressive stratigraphies in Australia (Roy et al., 1994) show a somewhat more complex range of preserved sequences that reflect differences in accommodation space, sediment supply, and antecedent topography. Transgressive sequences varied from preserved barriers where the substrate slope is gentle and transgressive sandsheets where the slope is steeper. In regressive situations, with adequate sediment supply, strandplains are formed that prograde seawards over shallow marine deposits.

There are, however, a variety of other sedimentological indicators of sea-level change. For example, Marchant and Flemming (1978) described granite boulders on a beach on the wave-exposed Cape Coast of South Africa. The nearest source area was offshore in 20 m water depth. Current velocities

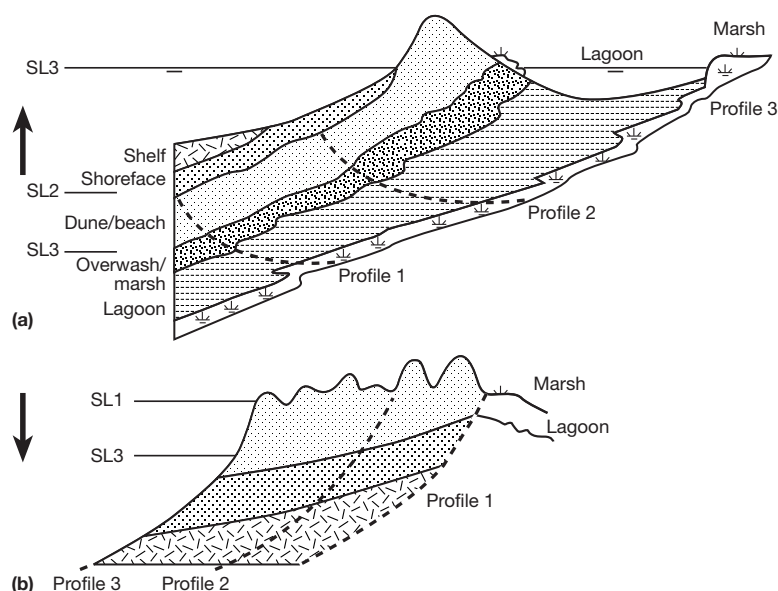


Figure 8 Transgressive and regressive barrier sequences. Under transgressive conditions (a), sea-level changes are associated with landward transport of sediments and stacking of deeper water over shallower water facies. Under regressive conditions (b), falling sea level is associated with seaward advance of the shoreline and vertical stacking of shallow water facies over deeper water facies. Both situations depicted depend on adequate sediment supply. Compare to [Figure 9](#) in which the preservation potential of sequences under different sediment supply rates is considered.

required to move such boulders from the source area were over 9 m s^{-1} , far beyond the range of velocities produced by even extreme waves at contemporary sea level. Current velocities in the surf zone velocities, however, would be capable of transporting such boulders. It was concluded that the boulders had been carried landward during transgression as the surf zone migrated landward during the Holocene sea-level rise.

[Dunbar and Barrett \(2005\)](#) assessed the sediment distribution on modern continental shelves in relation to bed shear stress and water depth. They found consistent trends in high-, medium-, and low-wave energy settings and applied the approach to Pliocene deposits where the paleobathymetry (and hence sea level) could be estimated if assumptions were made regarding the past wave climate.

Preservation Potential

Under regressive conditions the preservation potential of shoreline facies is relatively high since the progressive lowering of energy concentration in the tidal frame gradually leaves the deposits and landforms stranded above all but the most energetic waves. These features may be eroded or reworked under storm action at lower sea levels. Their preservation potential is much enhanced if they are buried by subsequent deposition of, for example, sand dunes, or protected from subsequent modification by the formation of more seaward beachridges.

The preservation potential of transgressive shoreline facies in relation to the rate of sea-level change was assessed by [Davis and Clifton \(1987\)](#) ([Figure 9](#)). Those authors concluded that under slowly rising sea level, the potential existed for almost total destruction of the transgressive coastal sequence and formation of an erosional surface or ravinement surface on which subsequent marine deposition would take place.

Under a rapidly rising sea level, however, the potential for *in situ* preservation of progressively shallower sediments was much higher and a full vertical sequence of deep to shallow water facies might be preserved.

Quaternary High-Energy Shorelines and Sea-Level Indicators

A number of studies of Quaternary sediments in high-energy settings have utilized knowledge of contemporary facies to identify past sea-level positions. Sea level is the plane around which the energy of coastal forces is distributed. Typically, the rates of sea-level change (mm year^{-1}) are much smaller than the interannual, seasonal, or even diurnal variability in the vertical levels at which coastal processes operate (m-vertical scale). Consequently, the identification of the role of sea-level change itself is difficult to ascertain in the sedimentary record of high-energy coasts.

Interpretation of coastal response to sea-level change at decadal to centennial timescales is difficult because the sedimentary response due to sea-level change is often masked by responses to short-term fluctuations in energy, sediment budget adjustments, and morphological evolution ([Cooper and Pilkey, 2004](#)). Working at the centennial scale, however, [Orford et al. \(1995\)](#) established a strong relationship between the retreat rate of a swash-aligned (i.e., with negligible long-shore drift) gravel barrier in western France and historical sea-level rise rates. It is more typically at the millennial timescale on high-energy coasts that sedimentary indicators are used to reconstruct sea-level history. For example, [Murray-Wallace \(2002\)](#) assessed the history of sea level in South Australia over the past 11 interglacials by studying the facies of raised beach features (usually barriers and dunes) in southern

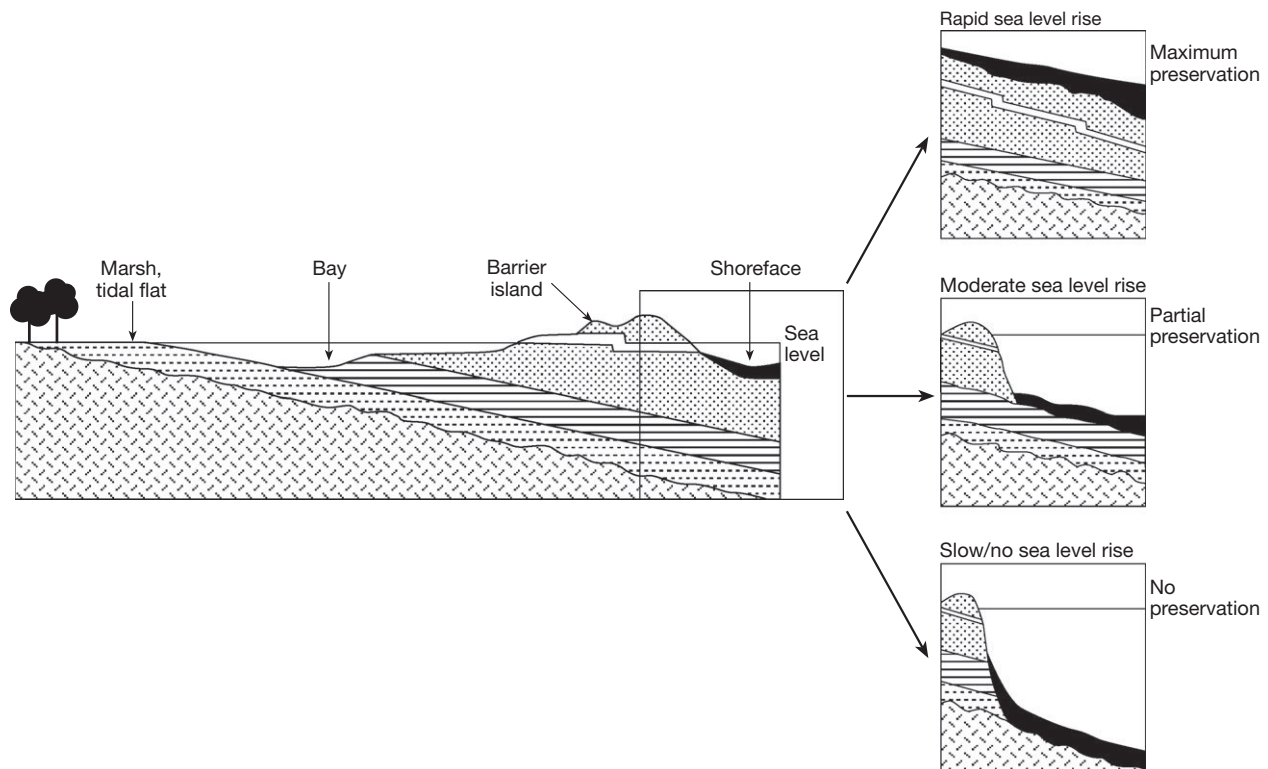


Figure 9 Preserved sequences in a transgressive situation under low, intermediate and rapid sea-level rise. Reproduced from Davis RA and Clifton HE (1987) Sea-level change and the preservation potential of wave-dominated and tide-dominated coastal sequences. In: Nummedal D, Pilkey OH, and Howard JD (eds.) *Sea-Level Fluctuation and Coastal Evolution*, pp. 167–178. Tulsa, OK: SEPM, Special Publication No. 41.

Australia. The levels of vertical resolution were in the order of a few meters. A number of studies have been made of cemented and unconsolidated Quaternary sediments in high-energy coastal settings. Their success in deducing former sea levels depends on clear identification of facies or erosional features associated with depositional environments with a specific relationship to sea level.

Roep and Beets (1988) utilized knowledge of the distribution of diagnostic sand barrier sedimentary structures (Roep, 1986) in relation to sea level (deepest dry eolian sand, highest marine burrow, range of low-angle bars, range of structureless sand, highest high angle bar, highest clay lamina, and highest ripples) in combination with dating of contained articulated shells to trace the mid-late Holocene sea-level history in the western Netherlands. Following the same reasoning, Van Heteren et al. (2000) utilized optical dating of sandy sediments at the beach–dune sedimentary interface to track sea-level changes on a coastal barrier in Massachusetts. The well-defined contact between dune and beach facies was taken as an indicator of former sea level (~2.3–2.7 m above mean spring high tide under contemporary conditions). The sea-level curve derived from dating of the basal dune sands coincided closely with a curve derived from organic deposits in an adjacent marsh, supporting the contention that this particular indicator had remained in a constant relationship to sea level over a few thousand years. The contact between aeolian and foreshore deposits was similarly used to track late Holocene sea-level rise in Alabama (Rodríguez and Meyer, 2006).

Former sea levels associated with raised gravel beach shorelines in western France were assessed by facies mapping and comparison with modern equivalents in the immediate vicinity (Haslett and Curr, 2001). A last interglacial transgressive sequence associated with rapid sea-level rise was preserved as it was overstepped by sea level. This was overlain by a prograding sequence that developed during a prolonged stillstand of sea level accompanied by an abundant sediment supply.

Through facies analysis and interpretation of the diagenetic history of emergent beachrock near Durban, South Africa, Cooper and Flores (1991) identified sediments of the surf zone, breaker zone, swash zone, back beach, and dune that gave an approximation of the sea-level position at which they formed. A rock platform cut into this lithified unit and an associated boulder beach facies recorded a later sea-level position. Both sets of beach facies were interpreted to have been deposited between +4 and +5 m above present mean sea level.

Working on the rocky coast of Banks Peninsula, New Zealand, Bal (1997) assessed a number of features (shore platforms, sea caves, sea stacks, notches, and cliffs). He identified a set of features related to contemporary sea level and a higher set of similar features at 5–6 m above their contemporary equivalents. These were associated with a last interglacial shoreline, whose level could be determined if it was assumed that wave energy conditions were similar during the last and present interglacials.

In now-submerged settings, direct observation of sediments associated with former sea levels is more difficult. Terrestrial

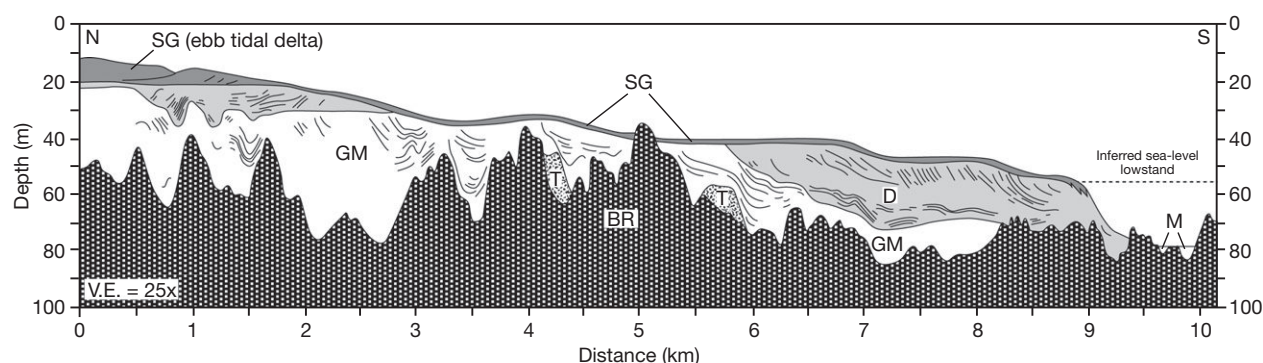


Figure 10 Submerged river delta on the Maine continental shelf. SG, sandy gravel; BR, bedrock; GM, glaciomarine mud; T, till; M, mud; D, sandy deltaic sediments. Reproduced from Barnhardt WA, Belknap DF, and Kelley JT (1997) Stratigraphic evolution of the inner continental shelf in response to late Quaternary relative sea level change, northwestern Gulf of Maine. *Geological Society of America Bulletin* 109: 612–630.

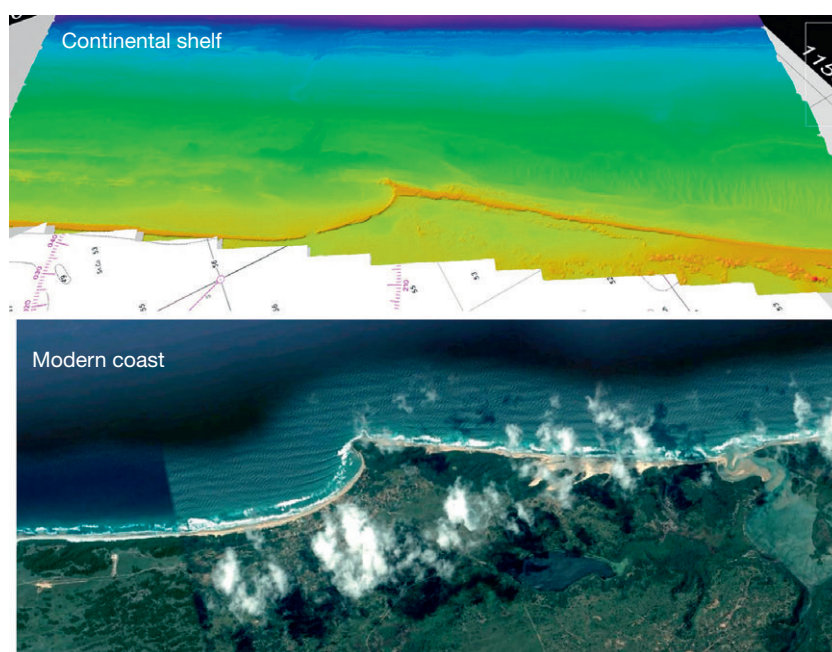


Figure 11 Multibeam bathymetry reveals a submerged zeta bay shoreline on the KwaZulu-Natal continental shelf (image courtesy of Dr Peter Ramsay, Marine Geosolutions). Note the similarity with the contemporary shoreline (lower photograph).

sediments are often dredged from the seabed, indicating that sea level was formerly lower but *in situ* investigations are necessary to identify former indicators of sea level. Consequently, much use has been made of seismic stratigraphic approaches to interpret sea-level history, often accompanied by underwater coring. A number of studies of lower than present sediments on the continental shelf have indicated the presence of discrete sedimentary units that are out of context in their present setting. These include drowned deltas (Barnhardt et al., 1997), spits, barriers, and beaches associated with lowstands of sea level (Stea et al., 1994). Barnhardt et al. (1997) described submerged deltas off large rivers in the Gulf of Maine that developed during lower than present sea-level stands (Figure 10). The deltas were identified on the basis of their internal structure and on their composition. They are composed mainly of sand and exist in a setting where contemporary mud deposition is the dominant process.

Recent advances in high-resolution multibeam echosounding have revealed many examples of drowned coastal geomorphic forms that indicate the positions of former shorelines. These are best preserved when RSL rise is rapid or when the shoreline deposits have been lithified (Figure 11). On the shelf of South Africa and Australia are several lines of submerged beachrock and aeolianite ridges that mark former shorelines (Cooper, 1991; Roy et al., 1994). Gardner et al. (2005, 2007) document and illustrate a variety of such features from the Florida continental shelf including deltas and barrier islands that closely resemble landforms on the contemporary west Florida coast. The dating of these deposits, however, remains problematic.

Combined analysis of emerged and submerged sedimentary deposits can enable a clear understanding of the processes and patterns of deposition and help decipher the relationship between different sedimentary units and their preservation.

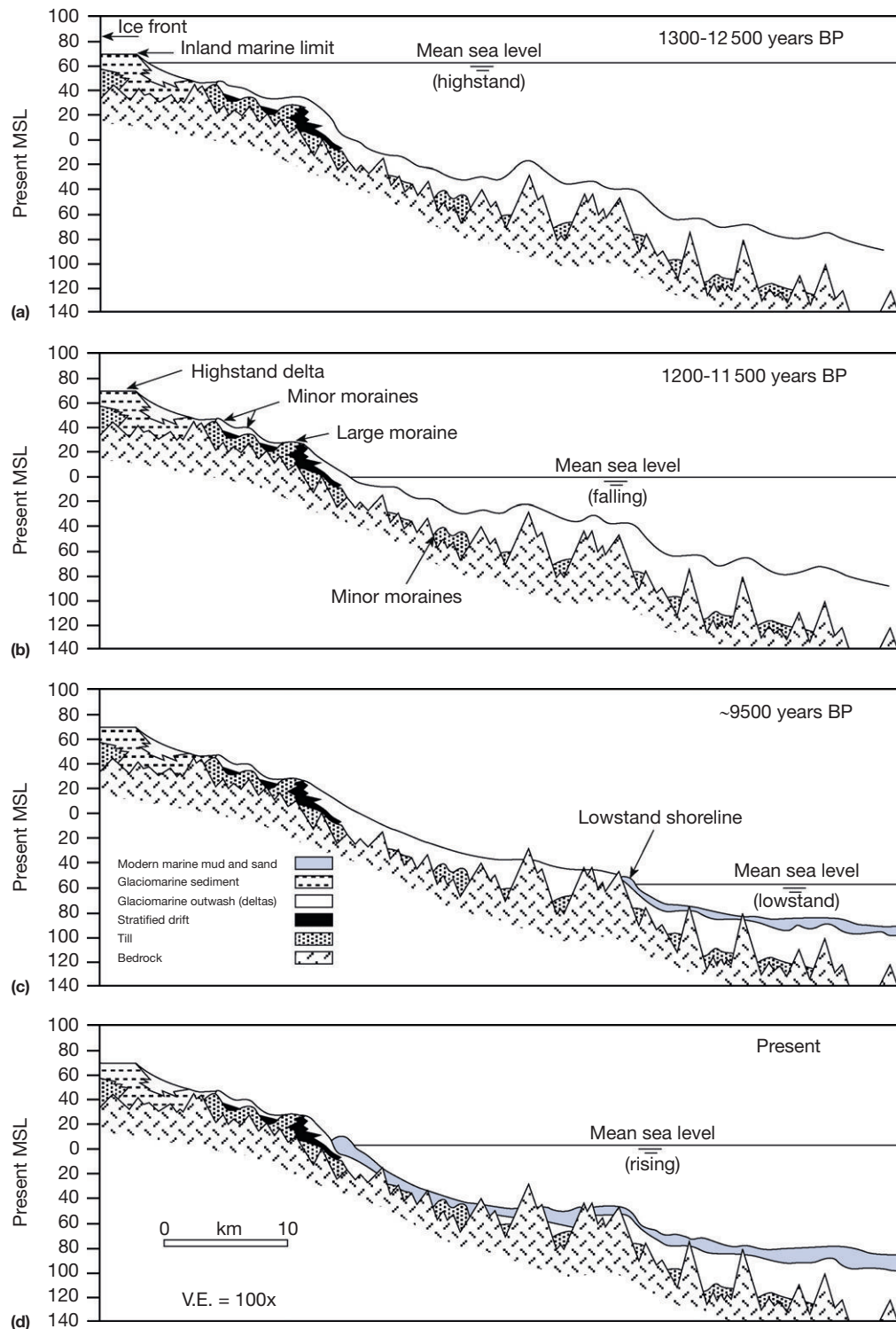


Figure 12 Diagram showing sequence of deposition in relation to higher and lower than present sea levels in the Gulf of Maine. Reproduced from Shipp RC, Belknap DF, and Kelley JT (1991) Seismic stratigraphic and geomorphic evidence for a sea-level lowstand in the Northern Gulf of Maine. *Journal of Coastal Research* 7: 341–364.

Working in the northern Gulf of Maine, an area affected by substantial isostatic uplift following deglaciation, Shipp et al. (1991) undertook a combined study of onshore and offshore coastal deposits that recorded deposition between relative sea

levels between 70 m above and 60 m below the present and spanning glacial to contemporary temperate climatic conditions (Figure 12). Recently, a combination of high-resolution multibeam bathymetry, seismic stratigraphy, seabed coring,

and radiometric dating enabled Kelley et al. (2010) to record submerged shorelines and associated deposits during a period of prolonged sea-level stability in the Gulf of Maine when the shorelines were sites of human occupation.

See also: Paleoclimate Modeling: Last Glacial Maximum GCMs. **Paleosols and Wind-Blown Sediments:** Nature of Paleosols; Overview; Soil Micromorphology. **Quaternary Stratigraphy:** Sequence Stratigraphy. **Sea Level Studies:** Overview. **Sea-Levels, Late Quaternary:** Late Quaternary Relative Sea-Level Changes in High Latitudes.

References

- Bal A (1997) Sea caves, relict shore and rock platforms: Evidence for the tectonic stability of Banks Peninsula, New Zealand. *New Zealand Journal of Geology and Geophysics* 40: 299–305.
- Barnhardt WA, Belknap DF, and Kelley JT (1997) Stratigraphic evolution of the inner continental shelf in response to late Quaternary relative sea level change, northwestern Gulf of Maine. *Geological Society of America Bulletin* 109: 612–630.
- Bluck BJ (1967) Sedimentation in beach gravels: Examples from South Wales. *Journal of Sedimentary Petrology* 37: 128–156.
- Carter RWG (1983) Raised coastal landforms as products of modern process variations, and their relevance in eustatic sea-level studies: Examples from eastern Ireland. *Boreas* 12: 167–182.
- Carter RWG (1988) *Coastal Environments*. Cambridge: Cambridge University Press.
- Clifton HE, Hunter RE, and Phillips RL (1971) Depositional structures and processes in the non-barred high energy nearshore. *Journal of Sedimentary Petrology* 41: 651–670.
- Cooper JAG (1991) Beachrock formation in low latitudes: Implications for coastal evolutionary models. *Marine Geology* 98: 145–154.
- Cooper JAG and Flores RM (1991) Shoreline deposits and diagenesis resulting from two Late Pleistocene stillstands near +4 and +5 m, Durban, South Africa. *Marine Geology* 97: 325–343.
- Cooper JAG and Pilkey OH (2004) Sea level rise and shoreline retreat: Time to abandon the Bruun Rule. *Global and Planetary Change* 43: 157–171.
- Cooper JAG and Power J (2003) Perched salt marshes on a high energy coast: Implications for sea level studies. *Journal of Coastal Research* 19: 357–363.
- Davidson-Arnott RGD and Greenwood B (1976) Facies relationships on a barred coast, Kouchibouguac Bay, New Brunswick, Canada. In: Davis RA (ed.) *Beach and Nearshore Sedimentation*, pp. 149–168. Tulsa, OK: SEPM Special Publication No. 24.
- Davis RA and Clifton HE (1987) Sea-level change and the preservation potential of wave-dominated and tide-dominated coastal sequences. In: Nummedal D, Pilkey OH, and Howard JD (eds.) *Sea-Level Fluctuation and Coastal Evolution*, pp. 167–178. Tulsa, OK: SEPM Special Publication No. 41.
- Dawson AG and Shi S (2000) Tsunami deposits. *Pure and Applied Geophysics* 157: 875–897.
- Dunbar GB and Barrett PJ (2005) Estimating palaeobathymetry of wave-graded continental shelves from sediment texture. *Sedimentology* 52: 253–269.
- Gardner JV, Calder BR, Hughes Clarke JE, Mayer LA, Elston G, and Rzhanova Y (2007) Drowned shelf-edge deltas, barrier islands and related features along the outer continental shelf north of the head of De Soto Canyon, NE Gulf of Mexico. *Geomorphology* 89: 370–390.
- Gardner JV, Dartnell P, Mayer LA, Hughes Clarke JE, Calder BR, and Duffy G (2005) Shelf-edge deltas and drowned barrier-island complexes on the Northwest Florida outer continental shelf. *Geomorphology* 64: 133–166.
- Haslett SK and Curr RHF (2001) Stratigraphy and palaeoenvironmental development of Quaternary coarse clastic beach deposits at Plage de Mezpeurleuch, Brittany, France. *Geological Journal* 36: 171–182.
- Hunter RE, Clifton HE, and Phillips RL (1979) Depositional processes, sedimentary structures and predicted vertical sequences in barred nearshore systems, southern Oregon coast. *Journal of Sedimentary Petrology* 49: 711–726.
- Kellett D (2006) Beachrock as sea-level indicator? Remarks from a geomorphological point of view. *Journal of Coastal Research* 22: 1558–1564.
- Kelley JT, Belknap DF, and Claesson S (2010) Drowned coastal deposits with associated archaeological remains from a sea-level 'slowstand': Northwestern Gulf of Maine, USA. *Geology* 38: 695–698.
- Marchant JW and Flemming BW (1978) Granite erratics in a sandstone boulder-beach in False Bay – Independent evidence for a marine transgression. *Transactions of the Geological Society of South Africa* 81: 219–221.
- Moslow TF (1984) *Depositional Models of Shelf and Shoreface Sandstones*. Tulsa, OK: American Association of Petroleum Geologists Continuing Education Course Note Series #27.
- Murray-Wallace CV (2002) Pleistocene coastal stratigraphy, sea-level highstands and neotectonism of the southern Australian passive continental margin – A review. *Journal of Quaternary Science* 17: 469–489.
- Orford JD, Carter RWG, and Jennings SC (1991) Coarse clastic barrier environments: Evolution and implications for Quaternary sea level interpretations. *Quaternary International* 9: 87–104.
- Orford JD, Carter RWG, McKenna J, and Jennings SC (1995) The relationship between the rate of mesoscale sea-level rise and the rate of retreat of swash-aligned gravel-dominated barriers. *Marine Geology* 124: 177–186.
- Pirazzoli PA (1986) Marine notches. In: Van de Plassche O (ed.) *Sea-Level Research: A Manual for the Collection and Evaluation of Data*, pp. 361–399. Norwich: Geo Books.
- Ramsay PJ (1995) 9000 years of sea-level change along the Southern African coastline. *Quaternary International* 31: 71–75.
- Rodriguez AB and Meyer CT (2006) Sea-level variation during the Holocene deduced from the morphologic and stratigraphic evolution of Morgan Peninsula, Alabama, USA. *Journal of Sedimentary Research* 76: 257–269.
- Roep TB (1986) Sea-level markers in coastal barrier sands: Examples from the North sea coast. In: Van de Plassche O (ed.) *Sea-Level Research: A Manual for the Collection and Evaluation of Data*, pp. 97–128. Norwich: Geo Books.
- Roep TB and Beets DJ (1988) Sea level rise and paleotidal levels from sedimentary structures in the coastal barriers in the western Netherlands since 5600 BP. *Geologie en Mijnbouw* 67: 53–60.
- Roy PS, Cowell PJ, Ferland MA, and Thom BG (1994) Wave-dominated coasts. In: Carter RWG and Woodroffe CD (eds.) *Coastal Evolution*, pp. 121–186. Cambridge: Cambridge University Press.
- Shipp RC (1984) Bedforms and depositional sedimentary structures of a barred nearshore system, eastern Long Island, New York. *Marine Geology* 60: 235–259.
- Shipp RC, Belknap DF, and Kelley JT (1991) Seismic stratigraphic and geomorphic evidence for a sea-level lowstand in the Northern Gulf of Maine. *Journal of Coastal Research* 7: 341–364.
- Stea RR, Boyd R, Fader GBJ, Courtney RC, Scott DB, and Pecore SS (1994) Morphology and seismic stratigraphy of the inner continental shelf off Nova Scotia, Canada: Evidence for a –65 m lowstand between 11,650 and 11,250 ¹⁴C yr B.P. *Marine Geology* 117: 135–154.
- Stolper D, List JH, and Thiel ER (2005) Simulating the evolution of coastal morphology and stratigraphy with a new morphological-behaviour model (GEOMBEST). *Marine Geology* 218: 17–36.
- Switzer AD, Pucillo K, Hareddy RA, Jones BG, and Bryant EA (2005) Sea level, storm, or tsunami: Enigmatic sand sheet deposits in a sheltered coastal embayment from Southeastern New South Wales, Australia. *Journal of Coastal Research* 21: 655–663.
- Thomas PJ (2009) Luminescence dating of beachrock in the southeast coast of India – Potential for Holocene shoreline reconstruction. *Journal of Coastal Research* 25: 1–7.
- Van Goor MA, Zitman TJ, Wang ZB, and Stive MJF (2003) Impact of sea-level rise on the morphological equilibrium of tidal inlets. *Marine Geology* 202: 211–227.
- Van Heteren S, Huntley DJ, Van de Plassche O, and Roberts RK (2000) Optical dating of dune sand for the study of sea-level change. *Geology* 28: 411–414.

Sedimentary Indicators of Relative Sea-Level Changes – Low Energy

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Introduction

Relative sea-level (RSL) change is the result of variations in ocean level (eustatic) and vertical land movements (isostatic). Throughout the Quaternary, both eustatic and isostatic changes have accompanied the waxing and waning of large, high latitude, continental ice sheets. The RSL changes experienced at locations around the globe are predominantly influenced by proximity to these ice sheets, and generalized patterns of RSL change can be predicted on the basis of geographical location and ice-sheet history. The most detailed records of RSL currently available are concerned with changes that have occurred since the end of the last Quaternary glaciation ~11 000 years ago, during the Holocene.

RSL changes are reconstructed from physical features and biological organisms whose distributions are related to sea level in a consistent manner (see [Lambeck et al., 2010](#)). Information concerning RSL change can be extracted from geomorphological evidence (e.g., wave-cut platforms and tidal notches), or from sedimentary environments and their associated flora and fauna. Sea-level data from these diverse sources have produced RSL records from locations around the globe.

This article examines the evidence for RSL change that is contained within low-energy sediments. It begins by outlining the principal, low-energy coastal environments that are sources of RSL data. It then considers the nature of the sea-level information they contain, and the methods employed to generate and analyze these data. It examines the relative strengths and weaknesses of these methodologies, highlighting recent developments and areas of current research interest.

Low-Energy Sedimentary Environments

Low-energy coastlines are common in regions around the globe where waves and tides are small and currents slow. They are typified by accumulations of fine-grained sediment (silty sand to clay) that accrete to form intertidal flats with distinctive floral and faunal communities. The lowest-energy environments are typically found in small, sheltered embayments, lagoons, and estuaries with restricted fetch ([Woodroffe, 2002](#)). Local hydrodynamic conditions may also give rise to low-energy environments in certain parts of an inlet, such as in the lee of a higher-energy gravel spit or sand bar. While most common in these protected areas, lower-energy sediments can also be found along open coasts where large tidal ranges promote the accumulation of extensive intertidal flats that dampen wave action. Some examples of low-energy sedimentary environments are illustrated in [Figure 1](#).

Sources of Sea-Level Data

The lithological and biological characteristics of low-energy intertidal sediments are intimately linked to the tidal frame. As a consequence of their position at the interface between

land and sea, intertidal zones exhibit strong environmental gradients. The highest elevations at the upper limit of tidal influence are associated with terrestrial organisms that can withstand periodic inundation by saline waters, or marine organisms that can withstand prolonged periods of subaerial exposure. As elevation decreases (hydroperiod increases), environmental conditions change, and plant and animal communities alter until almost marine conditions are reached at the lowest tidal levels. This vertical zonation related to tide level means that RSL information can be extracted from intertidal sediments. This is achieved by analyzing the nature of the sediments themselves (e.g., grain size, organic content, elemental/isotopic composition) and the plant and animal remains preserved within them.

Characteristic Intertidal Environments

Intertidal environments can be broadly subdivided into low elevation, unvegetated tidal flats, and higher-elevation vegetated surfaces. Tidal flats are commonly low-relief accumulations of sand, silt, and clay that are covered by most high tides. These flats may be dissected by tidal creeks, the size and density of which are related to the tidal range ([Allen, 2000](#)). While sediment source varies with location, in low-energy environments most material is redistributed across the intertidal flats by gentle wave action and tidal currents. The deposition of finer-grained sediments is facilitated by flocculation and biological packaging (e.g., production of fecal pellets). Similarly, the cohesive nature of clay sediments, coupled with consolidation due to escaping pore water and binding by organisms (e.g., algal mats), reduces the subsequent remobilization of deposited material.

The accumulation of sediment elevates intertidal surfaces, reducing the depth and period of tidal inundation. When sufficiently high relative to the tidal frame, these surfaces become vegetated. At this point, minerogenic sedimentation is augmented by the additional supply of organic material from above and below ground production ([Allen, 2000](#)). The presence of vegetation tends to slow currents and dampen waves, permitting the accumulation of more fine-grained sediments. Vegetation may also serve to bind intertidal sediments, reducing their propensity for erosion and transport. Together, these processes promote the accumulation of characteristic high-elevation organic-rich deposits. The nature of the organic-rich sediments will vary depending on the vegetation communities they support. In high to middle latitudes, saltmarsh plants occupy these high-elevation environments, while toward the tropics, mangrove swamps become widespread ([Bird, 2000](#)).

Saltmarshes

The floral character of a saltmarsh reflects a number of factors such as location, local hydrographic conditions, salinity regime, substrate, and elevation relative to the tidal frame ([Allen and Pye, 1992](#)). In many areas, the general stages of



Figure 1 Some examples of low-energy sedimentary environments. (a) An estuarine fringing saltmarsh in Southampton Water, southern Britain, showing a meandering, dendritic tidal creek network. © Photograph, Long AJ. (b) A back-barrier saltmarsh accumulating behind a large, high-energy gravel barrier in southern Britain. © Photograph, Long AJ. (c) A flat, well-vegetated, mature saltmarsh accumulating in a microtidal lagoon in southern Britain. © Photograph, Edwards RJ. (d) Open, low-energy mudflats fronting a mangrove swamp in Indonesia © Photograph, Horton BP.

saltmarsh evolution outlined in [Figure 2](#) can be recognized, although the precise species involved and their links to tidal levels may vary. Low-elevation sediments are first colonized by pioneer species such as *Spartina* and *Salicornia* ([Figure 2\(a\)](#)). A saltmarsh dominated by these environments is referred to as being youthful or immature ([Frey and Basan, 1985](#)). As the saltmarsh develops, the vegetated platform increases in elevation and reduces in gradient as sediments infill the available accommodation space ([Figure 2\(b\)](#)). A mature or old marsh has a larger proportion of taxa associated with high-elevation environments, such as *Puccinellia* and *Phragmites* species.

[Figure 3](#) shows three examples of contrasting saltmarsh environments. Minerogenic saltmarshes ([Figure 3\(a\)](#)) are typical of low-energy coasts around northwest Europe. The organic content of their sediments and the preservation of plant material are much less than in organogenic saltmarshes such as those encountered on parts of the eastern coast of North America ([Figure 3\(b\)](#)). Finally, [Figure 3\(c\)](#) shows a wide, flat expanse of saltmarsh accumulating behind a mangrove swamp in southeast Australia. This region has experienced a general encroachment of mangrove communities into saltmarsh areas in recent years. The cause of this environmental change may be due to a number of factors, including changes in RSL, tidal prism, sedimentation, and surface elevation ([Rogers et al., 2005](#)).

Mangroves

In tropical regions, saltmarsh vegetation is largely replaced by mangrove communities ([Figure 4\(a\)](#)). Mangrove trees grow best in low-energy environments with muddy substrates and can form extensive fringes where tidal ranges are large ([Bird, 2000](#)). They share a number of features in common with the saltmarsh systems outlined above. The presence of mangroves traps sediment suspended in tidal waters, promoting accretion and elevating the sediment surface. Low-elevation surfaces are characterized by pioneering species such as *Avicennia*, pictured in [Figure 4\(b\)](#) with its pneumatophores protruding from the adjacent unvegetated intertidal flats. As the sediment surface elevation increases, these pioneers are replaced by other mangrove species such as *Rhizophora*.

The Utility of Low-Energy Sedimentary Environments

The use of intertidal environments in sea-level research stems from the fact that subenvironments can be distinguished on the basis of their sedimentary character (lithostratigraphy) and associated flora or fauna (biostratigraphy); and that these subenvironments can be linked to the tidal frame ([van de Plassche, 1986](#)). Traditionally, attention has been focused on the easily recognizable transition between marine and

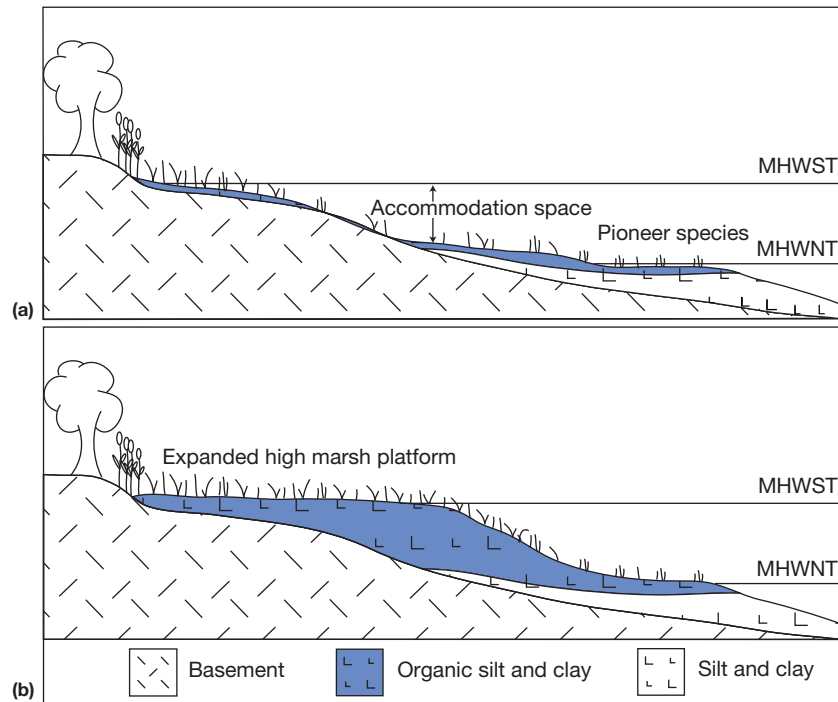


Figure 2 A sketch of the evolution of a typical minerogenic saltmarsh. (a) Pioneer species colonize unvegetated intertidal flats, enhancing sediment accumulation and promoting infilling of available accommodation space. Low-marsh species dominate, and the saltmarsh is referred to as immature. (b) Sediment accumulation results in an elevated saltmarsh surface and an expanded platform colonized by high-marsh species. An increased proportion of high-marsh environments indicates a mature saltmarsh. MHWST = mean high water of spring tides. MHWNT = mean high water of neap tides.

terrestrial conditions which is expressed in the lithostratigraphy by the interchange of minerogenic and organic-rich deposits (e.g., [Tooley, 1978](#)). The age of organic, peaty deposits can be established by radiocarbon dating, permitting a chronology for the sedimentary sequence to be developed (chronostratigraphy). In recent years, advances in the methods used to discriminate between subenvironments (e.g., microfossil analysis) and improvements in dating techniques (e.g., accelerator mass spectrometry (AMS) radiocarbon dating, and short-lived radionuclides such as ^{137}Cs , ^{210}Pb , etc.) have expanded the range of sedimentary contexts from which RSL data can be extracted.

In comparison with their higher-energy counterparts, low-energy sedimentary environments are less susceptible to processes such as erosion and transport. Given sufficient sediment and accommodation space, it is possible for relatively unbroken sequences to accumulate, and these provide ideal sources of RSL data.

Despite their potential utility, the records of RSL change preserved within these sequences are not always straightforward to extract. While sea-level studies are inherently concerned with extrinsic controls on saltmarsh development, marsh morphodynamics will inevitably involve a wider range of phenomena which can also leave imprints in the sedimentary record (e.g., autocyclicity; [Singh Chauhan, 2009](#)). This reinforces the fact that the sequences themselves do not comprise direct measurements of sea level, but rather record changes in a number of proxies for sea level. These proxies are influenced by other factors in addition to RSL change and therefore represent composite signals that require deconvolution. Consequently, the resulting records

will be affected by the particular indicator used, the capacity of that indicator to register change, and the preservation of this signature in the geological record. This will, in turn, depend on a number of factors including the magnitude and rate of RSL change, the sensitivity and precision of the indicator used, and the successive environmental conditions experienced after its incorporation into the sedimentary sequence.

Reconstructing RSL Change From Low-Energy Sedimentary Environments

A variety of strategies and methodologies are employed to distil RSL signals from the composite records contained within low-energy sediments. These are all underpinned by the multiproxy approach, employing a combination of litho-, bio-, and chronostratigraphic data. Lithostratigraphy is usually determined from a series of sediment cores collected at a study site and described in the field ([Figure 5](#)). In some circumstances, lithostratigraphy may also be examined in exposed sections ([Allen and Haslett, 2002](#)). Selected cores or monoliths are then returned to the laboratory for further analysis. This analysis may include quantitative sedimentology (e.g., grain size or loss on ignition), geochemical assessment (e.g., stable isotopes and elemental ratios of organic carbon and total nitrogen), and micropaleontology (e.g., pollen, diatoms, or Foraminifera) to establish the nature of the environment in which the sediment accumulated. The age of sediment samples is then commonly inferred by radiocarbon dating the organic material they contain.



Figure 3 Three examples of vegetated saltmarshes from around the world. (a) A minerogenic saltmarsh in Poole Harbour, southern Britain, showing defined intertidal vegetation zones grading into terrestrial woodland. (b) An organogenic saltmarsh from Connecticut, USA, showing a similar vegetation zonation of intertidal species, backed by woodland. (c) A sparsely vegetated, high-elevation saltmarsh plain backing mangroves in southeastern Australia. © All Photographs, Edwards RJ.

Sea-Level Indicators

An ideal sea-level indicator will be related to sea level in a consistent and quantifiable manner. In this way, changes in the indicator can be used as proxies for changes in RSL (Pirazzoli, 1996). For example, the altitude of past RSL can be inferred from a sediment sample if the current altitude of that sample is known, and if the environment in which it



Figure 4 (a) A tropical mangrove swamp in Indonesia, showing fine-grained intertidal mudflats. © Photograph, Horton BP. (b) An open, temperate mangrove stand showing pneumatophores extending into unvegetated intertidal flats. © Photograph, Edwards RJ.

formed has a quantified vertical relationship to sea level. Quantifying this vertical relationship (termed the indicative meaning) is important to account for the differing vertical distributions of coastal subenvironments and associated sea-level indicators (van de Plassche, 1986). Figure 6 illustrates the components of an indicative meaning which are defined in relation to the tidal frame. Any unquantified change in tidal range through time will introduce vertical error into RSL reconstructions. Samples from low-energy sedimentary environments are used to reconstruct RSL change in two principal ways: (1) by attempting to fix the vertical position (altitude) of sea level relative to a geodetic datum and (2) by examining switches in depositional environment indicating an increase or decrease in marine influence. A selection of approaches to RSL reconstruction that employ one or both of these perspectives are illustrated below. Detailed descriptions of these approaches, along with their associated errors and limitations are discussed in van de Plassche (1986), Shennan (1986), and Tooley and Shennan (1987).

Sea-Level Index Points

An established methodology for reconstructing vertical changes in RSL from sedimentary environments has been developed as part of several International Geosciences Programme (IGCP) Projects (Tooley, 1985). This methodology uses sea-level index



Figure 5 (a) Lithostratigraphic data is commonly collected in the field using hand-corers. Here, a narrow-bore gouge auger is being used to investigate sediments in an Indonesian mangrove. © Photograph, Horton BP. (b) The altitude of sediment samples and boreholes must be measured, and their vertical relationship to tide level established. Here, a total station is being used to level samples from intertidal flats in Indonesia. © Photograph, Horton BP. (c) Saltmarsh sediments recovered in a Russian-type corer are described in the field before being removed for laboratory analysis. In this section, bands of dark peat alternate with pale silty clay. © Photograph, Edwards RJ.

points (SLIPs) that fix the altitude of former RSL in time and space (Shennan, 1986). In order to be established as a SLIP, a sediment sample must possess information concerning its location (latitude and longitude), altitude (relative to a leveling datum), age (commonly inferred from radiocarbon dating), and its indicative meaning (Figure 6).

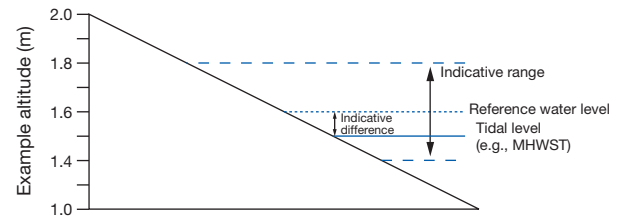


Figure 6 A sketch showing the components of an indicative meaning. The indicative range describes the vertical interval over which a sea-level indicator is found. The indicative difference describes the position of the mid-point of that indicative range (called the reference water level) relative to a defined tidal level. In this example, the reference water level is MHWST + 0.1 m, and the indicative range is ± 0.2 m.

Traditionally, SLIPs have been established from lithostratigraphic contacts between terrestrial and marine sediments, with supporting microfossil data (e.g., pollen, Foraminifera, and diatoms) being used to delimit the onset or removal of brackish/marine conditions and to confirm that the contacts are conformable (Shennan, 1986; Tooley, 1985). This focus on a restricted set of depositional environments arose because of the need to establish an indicative meaning (derived from the nature of the lithostratigraphy above and below a contact) and the requirement for organic material suitable for radiocarbon dating. The lithostratigraphic change from terrestrial peat to marine mud is referred to as a transgressive contact, while the switch from marine mud to terrestrial peat is termed a regressive contact. Transgressive and regressive contacts indicate an increase or decrease in marine influence, respectively. They are used as purely descriptive terms that do not imply an underlying cause or process. For example, a regressive contact (intertidal mud to terrestrial peat) may arise from the accumulation of sediment with or without a change in RSL.

Considerable attention has been directed toward quantifying the magnitude of uncertainty associated with determining the age and altitude of a SLIP (Shennan, 1986). For example, in the case of the age of a SLIP, these may relate to the uncertainties inherent in calibrating radiocarbon dates, or may be associated with contamination by allochthonous carbon (e.g., rootlet penetration). Additional errors can be introduced if the lithological contacts from which dateable material is recovered are associated with breaks in sedimentation, erosion, or reworking. In the traditional lithologically based scheme, the SLIP approach requires contacts to be conformable so that up-core changes reliably reflect the transitions between spatially adjacent subenvironments (Shennan, 1986). This may be difficult to establish from sedimentology alone and can be a particular problem for transgressive contacts where reworking mixes material to produce ‘pseudotransitional’ deposits (Waller et al., 2006). For these reasons, supporting data (e.g., biostratigraphy) are required to demonstrate that the dated sample forms part of a sequence of change extending across the contact but not restricted to it.

While biostratigraphy is the preferred tool for elucidating variations in marine influence across lithological contacts, poor microfossil preservation can result in insufficient data to reliably determine the nature of environmental change. In recent years, there has been growing interest in potential alternative geochemical approaches to supplement the

information provided by biostratigraphy. The isotopic signature of bulk organic carbon ($\delta^{13}\text{C}$) and the elemental ratio of organic carbon and total nitrogen (C/N) are routinely used to examine the source of organic matter in coastal and estuarine environments. In many cases, a good correlation exists between salinity and $\delta^{13}\text{C}$, suggesting that isotopic signatures can be used to distinguish between terrestrial, brackish, and marine environments (e.g., Yu et al., 2010). In regions where C3 and C4 plants occupy different saltmarsh elevation zones, the potential exists to recover an isotopic fingerprint of this biological zonation, although inputs from allochthonous carbon and/or diagenetic changes (e.g., decomposition) can distort these signatures (see Kemp et al., 2010). In the future, compound-specific analyses may provide one means of avoiding the complication of in-washed material (e.g., Johnson et al., 2007; Tanner et al., 2007).

Where C4 plants are absent from saltmarsh environments, the combination of $\delta^{13}\text{C}$ and C/N ratio has been used to detect relative shifts in marine influence on the basis that tidally introduced (allochthonous) particulate organic carbon will decrease with increasing elevation (see Lamb et al., 2006; Wilson et al., 2005). In this case, it is the trends in $\delta^{13}\text{C}$ and C/N ratio rather than the precise values which are considered diagnostic, with particularly marked excursions being associated with the upper limit of marine influence. While further work is needed to better understand the role of complicating factors, geochemical signatures have the potential to provide information on RSL change in Holocene sediments (Lamb et al., 2007). However, at present, the resulting SLIPs will be associated with larger vertical error terms than those produced by microfossil-based reconstructions due to the difficulties of precisely quantifying sample-indicative meaning.

Uncertainties relating to reliably establishing indicative meaning are not the only potential sources of vertical error associated with SLIPs. In addition to altitude errors that may be introduced during fieldwork (e.g., leveling to a datum, measuring depth in core, etc.), perhaps the most significant unknown variable is the extent to which samples may have been lowered through time due to autocompaction of the sediment column (Allen, 2000). In general terms, the effects of compaction are most pronounced in organic-rich sequences and will have greatest impact on the altitudes of SLIPs established from the transgressive contacts of thick, intercalated coastal peat deposits. At present, there is no agreed protocol for dealing with the effects of sediment compaction, although its effects are widely recognized (e.g., Horton and Shennan, 2009; Long et al., 2006a,b; Törnqvist et al., 2008; van Asselen et al., 2009). While a range of approaches have been used to investigate and address its impacts (e.g., Brain et al., 2011; Edwards, 2006; Massey et al., 2006; Paul and Barras, 1998), an initial assessment can simply be made by distinguishing between SLIPs from differing stratigraphic contexts (see Horton and Shennan, 2009).

In light of these collected uncertainties, SLIPs are plotted on age-altitude graphs as individual data points with associated error bars. The magnitude of these error terms are an important limiting factor in determining the maximum resolution at which RSL changes can be measured using this methodology. SLIPs fix the former altitude of RSL in time, but do not provide any information on the nature of sea-level change between

points. As a result, the number, distribution, and quality of data points provide fundamental constraints on the detail with which patterns of RSL change can be reconstructed.

An example of the age-altitude approach

To illustrate the SLIP approach, a hypothetical sedimentary sequence with associated radiocarbon dates is presented in Figure 7. The lithostratigraphic section (Figure 7(a)) is typical of the intercalated low-energy sedimentary sequences found beneath minerogenic saltmarshes in northwest Europe (Allen, 2000; Long et al., 2000). The type of litho- and chronostratigraphic data furnished by a transect of sediment cores is shown in Figure 7(b) and these data are used to produce an age-altitude plot (Figure 7(c)). For simplicity, in this example, all transgressive and regressive contacts are assigned an indicative meaning equivalent to the elevation of MHWS ± 0.5 m.

In addition to the SLIPs, two dated samples without indicative meanings are included on the diagram. While these cannot be directly linked to a former RSL, given knowledge of their depositional environment, they can be used to define the limit of possible RSL altitude (and are thus referred to as limiting dates). In this example, one limiting date is derived from the base of a terrestrial peat (Core E) and must have accumulated above the limit of marine influence (RSL below). Conversely, a second limiting date is taken from subtidal sediments in the same core (RSL above). These dates can be used to constrain the band of sea-level change drawn around the scatter of SLIPs (Figure 7(c)).

The age-altitude plot in Figure 7(c) exhibits a number of features typical of these kinds of record. Data are not evenly distributed through time with the result that there are intervals of 1000 years or more for which no sea-level data are available. Where data coverage is good, vertical scatter is introduced by differential sediment compaction. This is evident by the tendency for intercalated SLIPs (open boxes) to plot below basal SLIPs (filled boxes) of similar age. For example, a large amount of compaction in Core E is responsible for the scatter of points around 1000 BP. Finally, the calibration of radiocarbon dates results in age uncertainties of differing magnitude, and this produces SLIPs with variable age errors. In reality, many of the SLIPs would also be associated with differing altitude errors. The combination of these limitations means that a generalized band of RSL change is plotted around the data. The thickness of this band defines the upper limit of resolution at which the record can be reliably interrogated.

A variant of the age-altitude approach

A modified version of the SLIP methodology has been employed to extract sea-level data from isolation basins. These rocky basins may be located along high-energy coasts, but contain sedimentary sequences that have accumulated in low-energy conditions, protected from the open sea behind a rock sill (Figure 8). Classic examples of isolation basins are found along coastlines experiencing long-term crustal uplift (e.g., Greenland, Scandinavia, Scotland). This uplift results in RSL fall and the isolation of these sedimentary basins from marine influence (Figure 9). This isolation is recorded by microfossil assemblages (e.g., diatoms, Foraminifera) that are sensitive to changes in salinity. These microfossils are preserved within isolation basin sediments and record changes from marine

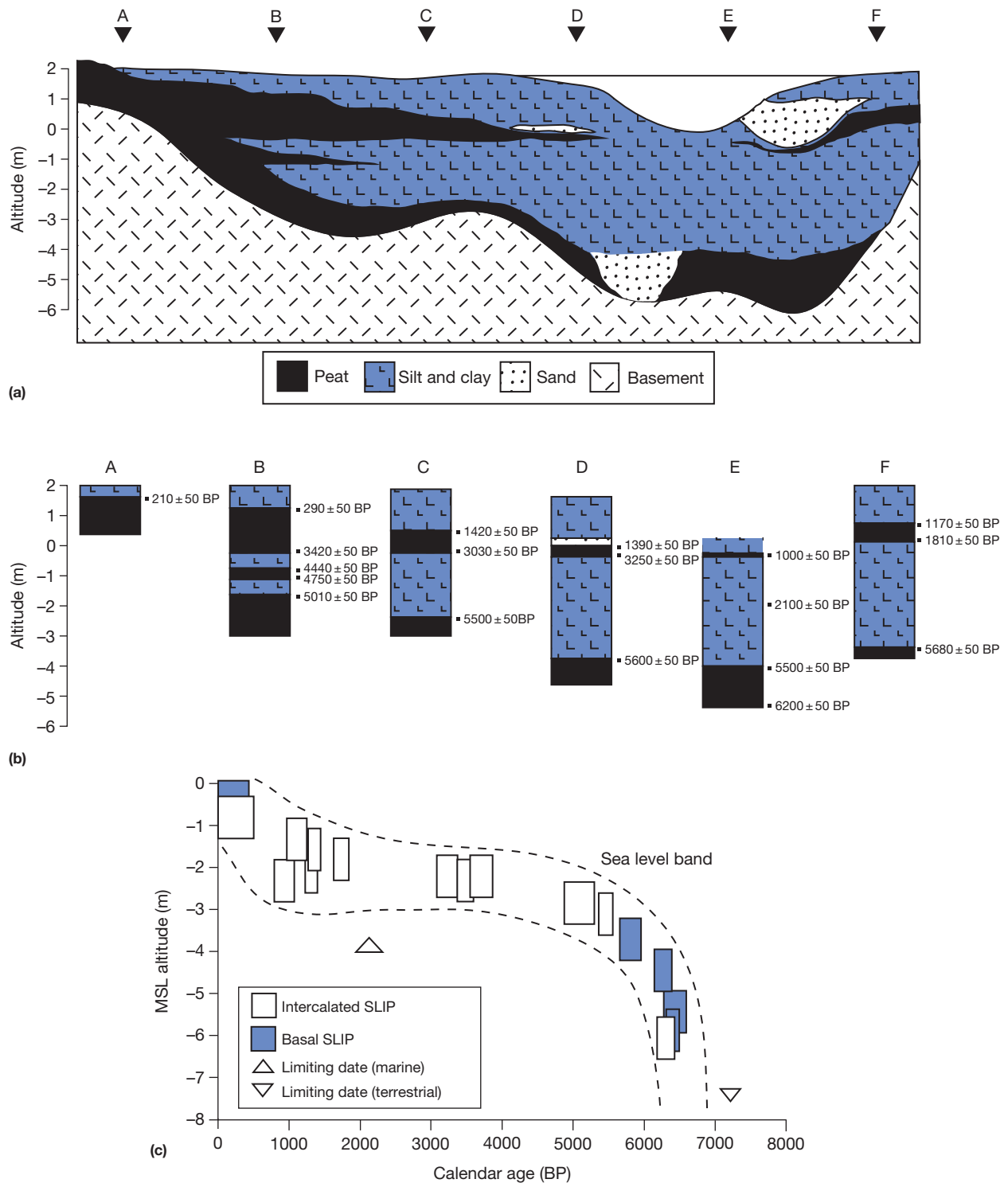


Figure 7 An example stratigraphic sequence used to furnish sea-level index points (SLIPs) for age–altitude analysis. (a) A cross-section of the hypothetical lithostratigraphy indicating the location of a transect of boreholes. (b) The lithostratigraphy as represented by six boreholes, with associated chronological data provided by a series of hypothetical radiocarbon dates. (c) The resulting age–altitude plot of SLIPs plotting changes in mean sea level over time. The height and width of the boxes represents the altitude and age uncertainties associated with each SLIP. Mean high-water spring tides (MHWST) is defined as 2 m above mean sea level. All SLIPs were assigned an indicative meaning of ± 0.5 m. Two limiting dates (without indicative meanings) are shown as triangles.



Figure 8 (a) A staircase of isolation basins in western Greenland. The rock sills separating each basin are clearly visible. © Photograph, Long AJ. (b) Sediments recovered from an isolation basin using a Russian-type corer. This section shows a classic isolation sequence, changing from a marine, gray silty clay at the base, through to a brown, organic gyttja. © Photograph, Long AJ.

through brackish to freshwater depositional environments (Figure 9).

In rapidly uplifting areas (e.g., Greenland), this isolation occurs comparatively quickly and is recorded in the litho- and biostratigraphy by an abrupt isolation contact. This contact can be dated and used as a SLIP, although the altitude of the index point is reconstructed differently from those established in saltmarshes. Instead of using the altitude of the sediment

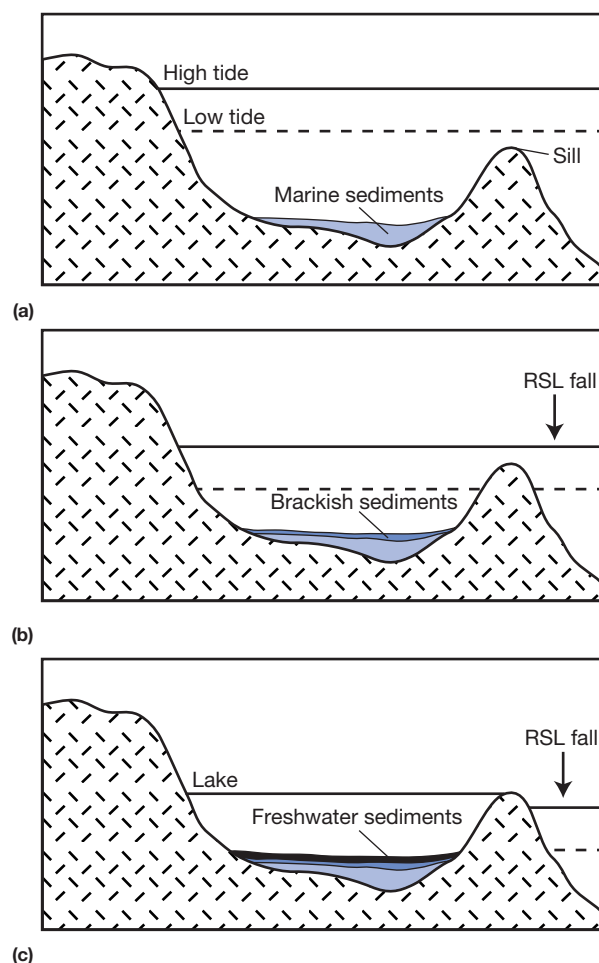


Figure 9 A schematic diagram showing the isolation of a tidal basin by falling relative sea level. (a) The basin is inundated at all stages of the tide, and marine sediments accumulate. (b) The basin is isolated from the sea at low tides and brackish conditions are experienced. (c) High water is now below the height of the sill and the isolation basin becomes a freshwater lake.

sample, SLIPs from isolation basins use the altitude of the lowest part of the rock sill separating the basin from the open sea (Figure 9). In this way, isolation basins have provided important, long records of RSL change from regions close to former centers of ice loading. While these index points are not susceptible to the influence of sediment compaction, their reliable interpretation relies on successfully locating the lowest point of the sill and assumes that the sill itself is impermeable to seawater. Figure 10 shows a simplified example of data collected from an isolation basin study in western Greenland (see Long et al., 2006a,b for details). A transect of boreholes were recovered from an isolation basin with a sill 1.61 m above modern sea level (Figure 10(a) and 10(b)). A series of diatom samples were taken across the transition from a marine gray silt clay into a freshwater gyttja, and an AMS radiocarbon date from this contact returned an age of 4150 ± 50 BP (Figure 10(c)). These data were used to establish a SLIP which, when plotted on an age–altitude diagram with other SLIPs from the area, charts RSL change over the last 8000 years (Figure 10(d)).

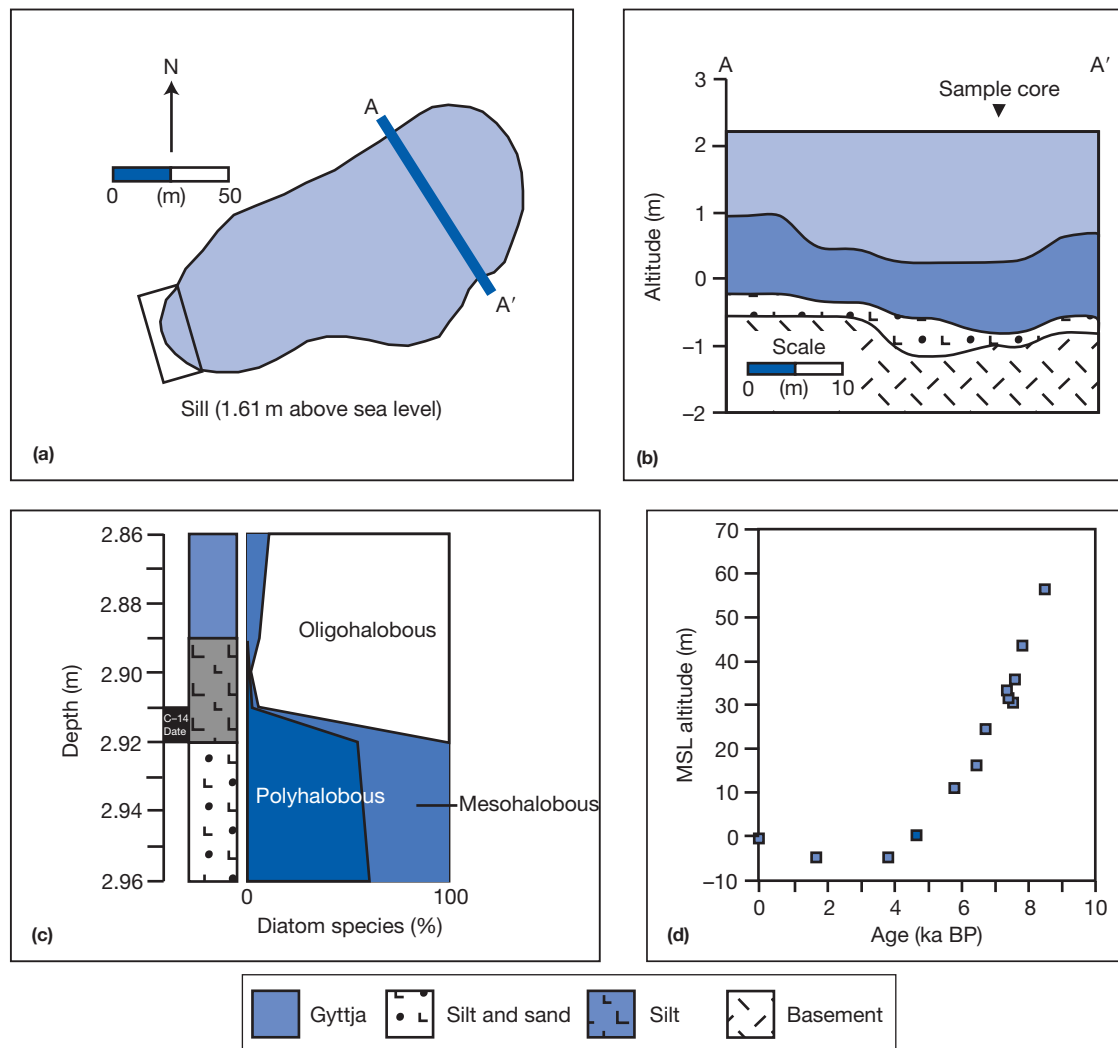


Figure 10 Example of sea-level data from an isolation basin in western Greenland. (a) Plan of the isolation basin showing transect location and position of sill. (b) Lithostratigraphy of the isolation basin showing sample core location. (c) A simplified stratigraphic column plotted with a summary of the diatom analysis, which shows the shift from polyhalobous species (indicative of marine conditions) to oligohalobous species (associated with freshwater conditions). (d) An age–altitude plot derived from isolation basin data collected in western Greenland (example site data plotted as a filled box). Figures compiled and adapted from Long et al. (2006a).

In regions associated with smaller amounts of uplift (e.g., northwest Scotland), complete isolation from the sea takes much longer with the result that brackish water, transitional sediments accumulate in greater thicknesses, and stratigraphic contacts are less abrupt. This complicates the identification and interpretation of the isolation contact and ongoing research is focusing on improving reconstructions from these transitional environments, either by a more sophisticated treatment of microfossil data, or by additional isotopic/elemental analysis (e.g., Mackie et al., 2007).

Sea-Level Tendencies

An alternative and complementary perspective to age–altitude analysis can be supplied by considering the timing and nature of changes in depositional environment. Instead of attempting to quantify vertical RSL movements, attention is focused on

delimiting periods of time when sea-level indicators from a particular area display increases or decreases in marine influence. Increases in marine influence are termed positive sea-level tendencies, while decreases in marine influence are called negative sea-level tendencies (Shennan, 1986).

Sea-level tendencies have been defined, employed, and (mis)interpreted in differing ways in the literature, which has led some authors to suggest that the terms are potentially misleading and no longer helpful (Allen, 2003). If they are to be used with sedimentary archives, it is important to recognize that sea-level tendencies are the product of a balance between sedimentation and local tide-level changes and consequently are not synonymous with rises and falls in RSL. While in mature marshes or settings with well-quantified sedimentation rates, they can provide insights into changes in the rate of RSL rise, even similar tendencies observed across a region need not be related to RSL change. For example, in southern Britain, rapid

colonization by *Spartina anglica*, a hybrid saltmarsh grass capable of withstanding long periods of submergence, produced dramatic increases in sediment accumulation and saltmarsh growth throughout the region during the early part of the twentieth century (see Allen, 2000 and references therein). This gave rise to a widespread negative sea-level tendency during a time in which instrumental measurements indicate rising RSL.

While it should therefore be used with caution, the tendency approach can help tackle certain limitations inherent with age–altitude analysis. For example, in some instances SLIPs from closely spaced transgressive and regressive contacts in the same core are indistinguishable on an age–altitude plot, resulting in a loss of stratigraphic information (e.g., Core B in Figure 7). However, these contrasting SLIPs will be clearly distinguished by tendency analysis. Tendencies can also be useful in identifying inflections in the RSL curve when dealing with SLIPs from isolation basins, since these records are unaffected by variations in sedimentation rate. Another potential strength of the tendency approach is that it expands the range of material that can be used to provide sea-level information. All dateable indicators that show increases or decreases in marine influence can be considered, irrespective of whether their altitude or indicative meaning is quantified. Ultimately, in order to be successful, the tendency approach requires a large number of dated indicators, and this, coupled with its inherent ambiguities, may be one of the reasons why it is not widely applied in sea-level studies.

Marsh Elevation Diagrams

Improvements in microfossil-based reconstructions of RSL have increased the accuracy of indicative meanings and expanded the range of sediments that can be employed in sea-level studies. A new generation of sea-level curves is seeking to combine detailed marsh elevation diagrams with precise SLIPs. These records are of sufficient resolution that they can be compared with proxy climate data, or combined with tidal records (Horton and Edwards, 2006). This type of approach is currently being developed to investigate the relationship between climate and sea level at multidecadal to centennial scales (e.g., Gehrels et al., 2005, 2006, 2008; Kemp et al., 2009, 2011). The use of marsh elevation diagrams (sometimes called paleoenvironmental curves) was developed in the organogenic marshes of the eastern USA (e.g., Varekamp et al., 1992). Multiple sediment samples are collected from individual cores, and the microfossils (commonly Foraminifera) contained within them are used to precisely fix the former elevation of the sediment surface relative to the local tidal frame (Figure 11(a)). The organic-rich nature of the sediments permits large numbers of radiocarbon dates to be collected for each core, and detailed accumulation histories can be developed (Figure 11(b)). By combining these two detailed datasets, the pattern of tide-level change through time can be determined, in this case shown as a curve of MTL rise (Figure 11(c)).

The marsh elevation diagram approach produces, in effect, a series of stratigraphically constrained SLIPs from a single core, with ages determined by its accumulation history. These records can be used to plot general changes in the altitude of RSL but are also capable of capturing more subtle variations in water level. In theory, while marsh elevation diagrams will vary

across a study site, reflecting the local balance between sediment accumulation and RSL (Figure 11(a)), when they are combined with detailed accumulation histories (Figure 11(b)), the same record of RSL change will be extracted (Figure 11(c)). In reality, a record may contain breaks in sedimentation or changes in rate that are not reliably captured by the available age data. In this event, the resulting pattern of RSL change may be distorted (Horton and Edwards, 2006). Consequently, the success of this approach requires a strong age framework (Edwards, 2004) and should incorporate a detailed program of lithostratigraphic investigation to ensure that the sample borehole is located in a position that captures the dominant environmental changes experienced at a site.

Given the importance of accurately determining accumulation history, attention is increasingly being focused on the construction of composite chronologies that employ a range of dating methods to infer the age of intertidal sediments. Approaches have employed various combinations of short-lived radionuclides (e.g., ^{210}Pb , ^{137}Cs , ^{241}Am), chronostratigraphic markers (e.g., tephra, pollen, spheroidal carbonaceous particles, lead isotopic signatures), paleomagnetism, and a range of techniques based around AMS radiocarbon dating (e.g., Gehrels et al., 2006a,b; Marshall et al., 2007, 2009). Increasingly sophisticated statistical manipulations of age data also have the potential to improve the development of age–depth relationships (e.g., Blaauw et al., 2007; Heegaard et al., 2005; Parnell et al., 2008; van de Plassche et al., 2001), although these will ultimately be limited by the fidelity of the sedimentary sequences themselves (Edwards, 2007).

Records produced using marsh elevation diagrams will also be susceptible to the influence of sediment compaction in a similar manner to intercalated SLIPs. Some studies remove the long-term RSL trend in their data, which reflects the overall influence of factors like glacio-isostatic movement and sediment compaction (e.g., Kemp et al., 2011). Other studies endeavor to decompact their records by incorporating age–altitude data from basal SLIPs (e.g., Gehrels et al., 2005).

Figure 12 shows an example record developed from replicate cores recovered from two saltmarshes in North Carolina, USA (Kemp et al., 2009). Marsh elevation reconstructions developed from a foraminiferal transfer function for tide level are combined with detailed chronologies derived from pollen analysis, short-lived radionuclides (^{210}Pb , ^{137}Cs), bomb-spike, and high-precision AMS radiocarbon dating. The resulting records, which exhibit similar patterns of change, show good agreement with data from neighboring tide gauges and extend these instrumental records back to AD 1500. Analysis of the long-term trends in records such as these can be used to examine the timing and magnitude of any recent acceleration in the rate of RSL rise. Ongoing research aims to produce multiple records of this sort with the ultimate aim of identifying common features that may reflect larger-scale, climate-related changes in ocean level (see discussion in Gehrels, 2010).

Concluding Remarks

Low-energy sedimentary environments and their biological components are rich sources of RSL information. Collections of SLIPs, which are predominantly sourced from low-energy

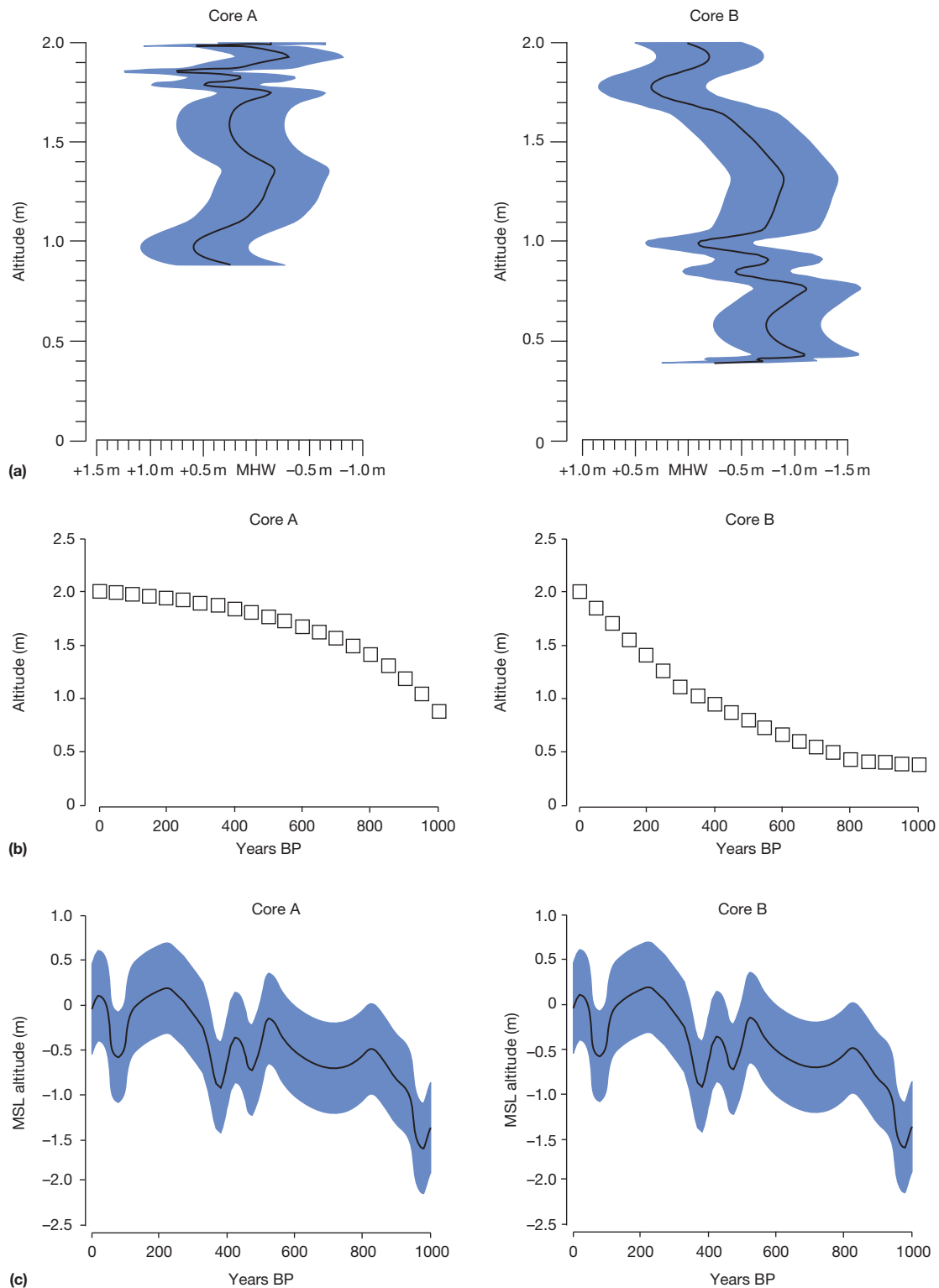


Figure 11 A hypothetical example showing the marsh elevation-diagram approach. (a) Marsh elevation diagrams produced from two cores at the same site. Marsh surface elevation is plotted relative to mean high water, which is 2 m above mean tide level. (b) Accumulation curves for both cores. The contrasting age–depth relationships reflect local differences in sediment accumulation rate. (c) Plots of MTL rise produced by combining the marsh elevation diagrams with the accumulation curves. The example demonstrates that the same RSL record can be extracted from both cores.

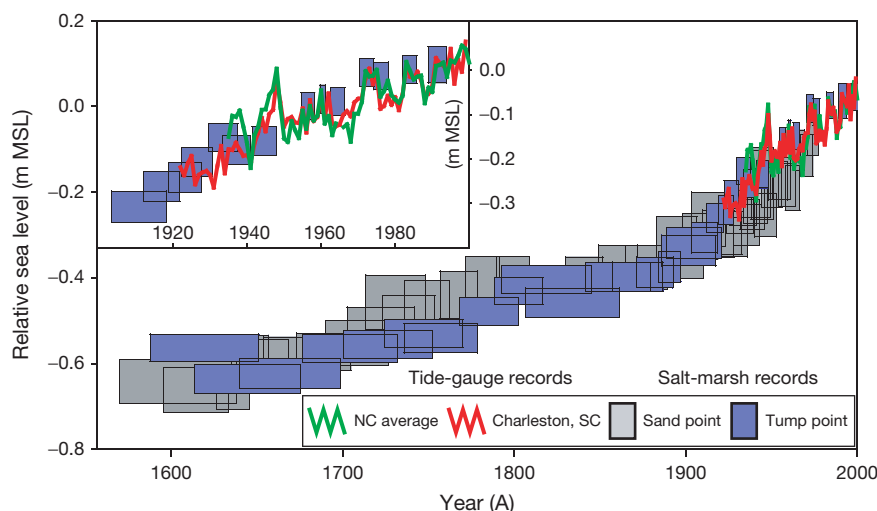


Figure 12 An example of high-resolution sea-level reconstruction developed for two sediment cores from North Carolina, USA. Reproduced from Kemp AC, Horton BP, Culver SJ, et al. (2009) Timing and magnitude of recent accelerated sea-level rise (North Carolina, United States). *Geology* 37(11): 1035–1038. Sample elevation was reconstructed from a foraminiferal transfer function for tide level. Sediment accumulation histories were developed with composite chronologies comprising a pollen chronohorizon, short-lived radionuclides (^{210}Pb , ^{137}Cs), and AMS radiocarbon dating (including the use of ‘bomb-spike’ and high-precision radiocarbon dating). The inset illustrates the good agreement between reconstructed RSL and the instrumental record from two neighboring tide gauges (see Kemp et al., 2009 for details).

sedimentary environments, enable the spatial patterns of RSL change to be established. In turn, these patterns are used to elucidate climate–ice–ocean relationships and the local scale processes that influence their expression in the sedimentary record (e.g., Engelhart et al., 2009; Horton and Shennan, 2009). The provision of reliable RSL data in the form of SLIPs plays a central role in glacial isostatic adjustment modeling and several efforts are underway to compile and disseminate databases of index points to facilitate this process (e.g., Brooks and Edwards, 2006; Brooks et al., 2008). Ongoing research seeks to improve the precision and resolution of existing RSL records and expand the range of environments from which sea-level data can be obtained. Advances in dating techniques, especially those involving minerogenic sediments and improvements in establishing sample indicative meaning, will help to bridge the gap between short duration instrumental records and longer-term geological reconstructions. Understanding sea-level rise and variability is a critical endeavor, both in terms of the range of scientific disciplines that use sea-level data and the wider societal issues associated with the impacts of future sea-level rise on our heavily populated coastlines (Church et al., 2010). While considerable uncertainty currently exists regarding the nature of this rise and its impacts, low-energy sedimentary environments will unquestionably play a critical role in future efforts to improve our understanding of sea-level change.

See also: Geomagnetic Excursions and Secular Variations.

Radiocarbon Dating: AMS Radiocarbon Dating; Calibration of the ^{14}C Record; Conventional Method. **Sea Level Studies:** Eustatic Sea-Level Changes – Glacial–Interglacial Cycles; Eustatic Sea-Level Changes Since the Last Glacial Maximum; Isostasy: Glaciation-Induced Sea-Level Change; Microfossil-Based Reconstructions of Holocene Relative Sea-Level Change; Overview; Sedimentary Indicators of

Relative Sea-Level Changes – High Energy. **Sea-Levels, Late Quaternary:** Late Quaternary Relative Sea-Level Changes at Mid-Latitudes; Late Quaternary Relative Sea-Level Changes in High Latitudes; Late Quaternary Relative Sea-Level Changes in the Tropics.

References

- Allen JRL (2000) Morphodynamics of Holocene salt marshes: A review sketch from the Atlantic and Southern North Sea coasts of Europe. *Quaternary Science Reviews* 19: 1155–1231.
- Allen JRL (2003) An eclectic morphostratigraphic model for the sedimentary response to Holocene sea-level rise in northwest Europe. *Sedimentary Geology* 161: 31–54.
- Allen JRL and Haslett SK (2002) Buried salt-marsh edges and tide-level cycles in the mid Holocene of the Caldicot Level (Gwent), South Wales. *The Holocene* 12: 291–301.
- Allen JRL and Pye K (1992) Saltmarshes. *Morphodynamics, Conservation and Engineering Significance*. Cambridge: Cambridge University Press.
- Bird E (2000) *Coastal Geomorphology: An Introduction*. Chichester: John Wiley & Sons Ltd.
- Blaauw M, Christen J, Mauquoy D, van der Plicht J, and Bennett K (2007) Testing the timing of radiocarbon-dated events between proxy archives. *The Holocene* 17: 283–288.
- Brain MJ, Long AJ, Petley DN, Horton BP, and Allison RJ (2011) Compression behaviour of minerogenic low energy intertidal sediments. *Sedimentary Geology* 233: 28–41.
- Brooks A and Edwards RJ (2006) The development of a sea-level database for Ireland. *Irish Journal of Earth Science* 24: 13–27.
- Brooks AJ, Bradley SL, Edwards RJ, Milne GA, Horton BP, and Shennan I (2008) Post-glacial relative sea-level observations from Ireland and their role in glacial rebound modelling. *Journal of Quaternary Science* 23(2): 175–192.
- Church JA, Woodworth PL, Aarup T, and Wilson S (2010) *Understanding Sea-level Rise and Variability*. Chichester: Wiley-Blackwell.
- Edwards RJ (2004) Constructing chronologies of sea-level change from salt-marsh sediments. In: Buck CE and Millard AR (eds.) *Tools for Constructing Chronologies: Crossing Disciplinary Boundaries*, pp. 191–213. London: Springer Verlag.
- Edwards RJ (2006) Mid to late Holocene relative sea-level change in southwest Britain and the influence of sediment compaction. *The Holocene* 16(4): 575–587.

- Edwards RJ (2007) Sea levels: Resolution and uncertainty. *Progress in Physical Geography* 31(6): 621–632.
- Engelhart SE, Horton BP, Douglas BC, Peltier WR, and Tornqvist TE (2009) Spatial Variability of Late Holocene and 20th Century Sea Level Rise along the US Atlantic Coast. *Geology* 37: 1115–1118.
- Frey RW and Basan PB (1985) Coastal salt marshes. In: Davis RA (ed.) *Coastal Sedimentary Environments*, 2nd edn., pp. 225–301. New York: Springer-Verlag.
- Gehrels WR (2010) Sea-level changes since the Last Glacial Maximum: An appraisal of the IPCC Fourth Assessment Report. *Journal of Quaternary Science* 25: 26–38.
- Gehrels WR, Hayward BW, Newnham RM, and Southall KE (2008) A 20th century sea-level acceleration in New Zealand. *Geophysical Research Letters* 35: L02717.
- Gehrels WR, Kirby JR, Prokoph A, et al. (2005) Onset of recent rapid sea-level rise in the western Atlantic Ocean. *Quaternary Science Reviews* 24: 2083–2100.
- Gehrels WR, Marshall WA, Gehrels MJ, et al. (2006) Rapid sea-level rise in the North Atlantic Ocean since the first half of the 19th century. *The Holocene* 16: 949–965. (Erratum: The Holocene 17: 419–420).
- Heegaard E, Birks H, and Telford R (2005) Relationships between calibrated ages and depth in stratigraphical sequences: An estimation procedure by mixed-effect regression. *The Holocene* 15: 612–618.
- Horton BP and Edwards RJ (2006) *Quantifying Holocene sea-level change using intertidal foraminifera: Lessons from the UK*. Fredericksburg, VA: Cushman Foundation for Foraminiferal Research. Special Publication 40.
- Horton BP and Shennan I (2009) Compaction of Holocene strata and the implications for relative sea-level change on the east coast of England. *Geology* 37(12): 1083–1086.
- Johnson BJ, Moore KA, Lehmann C, Bohlen C, and Brown TA (2007) Middle to late Holocene fluctuations of C3 and C4 vegetation in a Northern New England Salt Marsh, Sprague Marsh, Phippsburg Maine. *Organic Geochemistry* 38: 394–403.
- Kemp AC, Horton BP, Culver SJ, et al. (2009) Timing and magnitude of recent accelerated sea-level rise (North Carolina, United States). *Geology* 37(11): 1035–1038.
- Kemp AC, Horton BP, Donnelly JP, Mann ME, Vermeer M, and Rahmstorf S (2011) Climate related sea-level variations over the past two millennia. *Proceedings of the National Academy of Sciences* 108: 11017–11022.
- Kemp AC, Vane CH, Horton BP, and Culver SJ (2010) Stable carbon isotopes as potential sea-level indicators in salt marshes, North Carolina, USA. *The Holocene* 20: 623–636.
- Lamb AL, Vane CH, Wilson GP, Rees JG, and Moss-Hayes VL (2007) Assessing $\delta^{13}\text{C}$ and C/N ratios from organic material in archived cores as Holocene sea level and palaeoenvironmental indicators in the Humber Estuary, UK. *Marine Geology* 244: 109–128.
- Lamb AL, Wilson GP, and Leng MJ (2006) A review of coastal palaeoclimate and relative sea-level reconstructions using $\delta^{13}\text{C}$ and C/N ratios in organic material. *Earth-Science Reviews* 75: 29–57.
- Lambeck K, Woodroffe CD, Antonioli F, et al. (2010) Palaeoenvironmental records, geophysical modelling, and reconstruction of sea-level trends and variability on centennial and longer timescales. In: Church JA, Woodworth PL, Aarup T, and Stanley Wilson W (eds.) *Understanding Sea-Level Rise and Variability*, pp. 61–121. Chichester: Wiley Blackwell.
- Long AJ, Roberts DH, and Dawson S (2006a) Early Holocene history of the west Greenland Ice Sheet and the GH-8.2 event. *Quaternary Science Reviews* 25: 904–922.
- Long AJ, Waller MP, and Stupples P (2006b) Driving mechanisms of coastal change: Peat compaction and the destruction of late Holocene coastal wetlands. *Marine Geology* 225: 63–84.
- Long AJ, Scaife RG, and Edwards RJ (2000) Stratigraphic architecture, relative sea-level, and models of estuary development in southern England: New data from Southampton Water. In: Pye K and Allen JRL (eds.) *Coastal and Estuarine Environments: Sedimentology, Geomorphology and Geoarchaeology*, vol. 175, pp. 253–279. Geological Society, London: Geological Society. Special Publications.
- Mackie EAV, Lloyd JM, Leng MJ, Bentley MJ, and Arrowsmith C (2007) Assessment of $\delta^{13}\text{C}$ and C/N ratios in bulk organic matter as palaeosalinity indicators in Holocene and Lateglacial isolation basin sediments, northwest Scotland. *Journal of Quaternary Science* 22: 579–591.
- Marshall WA, Clough R, and Gehrels WR (2009) The isotopic record of atmospheric lead fall-out on an Icelandic salt-marsh since AD 50. *Science of the Total Environment* 407: 2734–2748.
- Marshall WA, Gehrels WR, Garnett MH, Freeman SPHT, Maden C, and Xu S (2007) The use of 'bomb spike' calibration and wiggle-matched high-precision AMS ^{14}C analyses to date salt-marsh sediments deposited during the past three centuries. *Quaternary Research* 68: 325–337.
- Massey AC, Paul MA, Gehrels WR, and Charman DJ (2006) Autocompaction in Holocene coastal back-barrier sediments from south Devon, southwest England, UK. *Marine Geology* 226: 225–241.
- Parnell AC, Haslett J, Allen JRM, Buck CE, and Huntley B (2008) A flexible approach to assessing synchronicity of past events using Bayesian reconstructions of sedimentation history. *Quaternary Science Reviews* 27: 1872–1885.
- Paul MA and Barras BF (1998) A geotechnical correction for post-depositional sediment compression: Examples of the Forth Valley, Scotland. *Journal of Quaternary Science* 13: 171–176.
- Pirazzoli PA (1996) *Sea Level Changes: The Last 20 000 Years*. Chichester: John Wiley and Sons Ltd.
- Rogers K, Saintilan N, and Heijnis H (2005) Mangrove encroachment of salt marsh in Western Port Bay, Victoria: The role of sedimentation, subsidence, and sea level rise. *Estuaries* 28: 551–559.
- Shennan I (1986) Flandrian sea-level changes in the Fenland. II. Tendencies of sea-level movement, altitudinal changes, and local and regional factors. *Journal of Quaternary Science* 1: 155–179.
- Singh Chauhan PP (2009) Autocyclic erosion in tidal marshes. *Geomorphology* 110: 45–57.
- Tanner BR, Uhle ME, Kelley JT, and Mora CI (2007) C3/C4 variations in salt-marsh sediments: An application of compound specific isotopic analysis of lipid biomarkers to late Holocene paleoenvironmental research. *Organic Geochemistry* 38: 474–484.
- Tooley MJ (1978) *Sea-Level Changes. North-West England during the Flandrian stage*. Oxford: Clarendon Press.
- Tooley MJ (1985) Sea levels. *Progress in Physical Geography* 9: 113–120.
- Tooley MJ and Shennan I (1987) *Sea-Level Changes* p. 397. Oxford: Blackwell.
- Törnqvist TE, Wallace DJ, Storms JEA, et al. (2008) Mississippi Delta subsidence primarily caused by compaction of Holocene strata. *Nature Geoscience* 1: 173–176.
- van Asselen S, Stouthamer E, and van Asch ThWJ (2009) Effects of peat compaction on delta evolution: A review on processes, responses, measuring and modelling. *Earth-Science Reviews* 92: 35–51.
- van de Plassche O (1986) *Sea-Level Research: A Manual for the Collection and Evaluation of Data*. Norwich: Geobooks.
- van de Plassche O, Edwards RJ, van der Borg K, and de Jong AFM (2001) ^{14}C wiggle-match dating in high-resolution sea-level research. *Radiocarbon* 43: 391–402.
- Varekamp J, Thomas E, and van de Plassche O (1992) Relative sea-level rise and climate change over the last 1500 years. *Terra Nova* 4: 293–304.
- Waller MP, Long AJ, and Schofield JE (2006) Interpretation of radiocarbon dates from the upper surface of late-Holocene peat layers in coastal lowlands. *The Holocene* 16: 51–61.
- Wilson GP, Lamb AL, Leng MJ, Gonzalez S, and Huddart D (2005) Variability of organic $\delta^{13}\text{C}$ and C/N in the Mersey Estuary, U.K. and its implications for sea-level reconstruction studies. *Estuarine, Coastal and Shelf Science* 64: 685–698.
- Woodroffe C (2002) *Coasts: Form, Process and Evolution*. Cambridge: Cambridge University Press.
- Yu F, Zong Y, Lloyd JM, et al. (2010) Bulk organic $\delta^{13}\text{C}$ and C/N as indicators for sediment sources in the Pearl River delta and estuary, southern China. *Estuarine, Coastal and Shelf Science* 87: 618–630.

Coral Records of Relative Sea-Level Changes

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Coral and Sea Level

Coral and coral reefs have played an invaluable role in the determination of past relative sea-level changes throughout the Quaternary at a range of time scales. Individual corals grow by calcification, their polyps secreting carbonate that is added to the exoskeleton of the colony. Coral polyps maintain a symbiotic relationship with species of dinoflagellate algae, called zooxanthellae. These limit the depth within which corals can survive to the photic zone (typically 0–30 m in clear tropical waters). Corals cannot tolerate more than very brief exposure when emerged above the water surface at lowest tides. Coral reefs are thus restricted to a depth range that varies from around mean low water, to depths at which water clarity limits photosynthesis (excepting cold-water corals found in high latitudes).

Coral reefs cover more than 250 000 km² of the earth's surface largely within the tropics where sea surface temperatures exceed 18 °C (Veron, 1995). The greatest diversity of corals occurs in the Indo-Pacific with a second region centered on the western Atlantic; reefs are absent from the Mediterranean and reef development is limited in the eastern Atlantic and eastern Pacific Oceans (Figure 1). The distribution of species across an individual reef, and of particular growth morphologies, varies according to depth, and in relation to several other environmental parameters such as wave energy, light availability, and sediment loading. Individual coral colonies can continue to grow for several hundreds of years, and after their death may be incorporated into the reef structure, together with other calcareous organisms such as Foraminifera, molluscs, and coralline algae. The reef fabric may be interpreted and the position of the surface of the sea during the time of coral growth can sometimes be inferred on the basis of species or growth form. The growth form of corals varies from delicate branching corals, through encrusting forms, to hemispherical colonies of massive corals. Large colonies of coral contain various geochemical records of conditions reflecting ambient water chemistry experienced during their growth. Such proxy reconstructions recovered from modern living corals have been correlated with, and extend, instrumental and historical records, whereas fossil corals provide an archive of water chemistry conditions at previous times during the Quaternary. Those corals that have been particularly constrained by exposure within the intertidal zone, provide the clearest evidence of former sea levels, particularly if it can be demonstrated that the coral is in growth position.

The role that reefs play in reconstructing sea level is comprehensively reviewed by Hopley et al. (2007, chapter 3), Montaggioni and Braithwaite (2009, pp. 405–428), and in numerous entries in the *Encyclopedia of Modern Coral Reefs* (Hopley, 2011). This article examines the contribution that coral records have made to the reconstruction of Quaternary sea level at four different time scales. First, the value of coral for

the recognition and dating of interglacial sea-level highstands reconstructed from terraces of reef limestone is described, particularly on uplifted shorelines throughout the tropics (Figure 2(a)). Second, the pattern of postglacial sea-level rise determined from drilling the foundations of modern reefs is examined, indicating the broad coincidence of reconstructions from disparate sites. Third, and in contrast, the role of coral in clarifying the variability of relative Holocene sea-level curves along tropical shorelines in the far-field (i.e., remote from former ice sheets), recording variability of record at different sites, is outlined. Finally, the potential of intertidal corals, called microatolls, to track interannual sea-level variations is reviewed.

Quaternary Sea-Level Highstands

The significance of fossil reef limestone, emerged and forming terraces elevated above modern shorelines particularly around the Pacific Ocean, was recognized in the early twentieth century by Reginald Daly who realized that sea-level history provided further support for former Ice Ages that were recorded by moraines marking a series of glacial advances in Europe (Daly, 1915). The advent of suitable dating techniques made it possible to place these shorelines into a sequence based on age, discriminating Holocene from Pleistocene, and more recently, differentiating successive interglacial sea-level highstands. Fossil coral reefs have played an important role over recent decades in refining the history of these sea-level variations. A sequence of reef terraces on Barbados, and a more extensive series on the Huon Peninsula on the north coast of Papua New Guinea, contain corals, now elevated by uplift. Dates on these corals constrain the age of past shorelines.

Whereas Holocene corals can be dated using radiocarbon dating techniques (appropriate to determine ages within the past 40 000 years), Pleistocene corals are largely beyond the range of radiocarbon dating, and have been dated primarily by the uranium-series disequilibrium dating methods, supplemented by several other techniques such as electron spin resonance. U-series dating measures the accumulation of ²³⁰Th as a daughter isotope in the uranium decay chain over the past 500 000 years, and U-series and ¹⁴C date comparisons on corals have extended calibration of the latter technique (Bard et al., 1990). With the initial application of U-series dating, Veeh (1966) showed that there was widespread evidence throughout the tropics for a Last Interglacial shoreline around 120 000 years old at heights averaging 6 m above the present sea level. The islands and continental shorelines, such as Bermuda, Bahamas, Seychelles, and Western Australia, were considered stable. However, several of these island sites, such as Oahu in the Hawaiian Islands, are now considered to have undergone flexural adjustments in response to loading of the ocean floor by adjacent volcanic islands. Mangaia in the Cook Islands for example, together with Mitiaro (Figure 2(b)) and

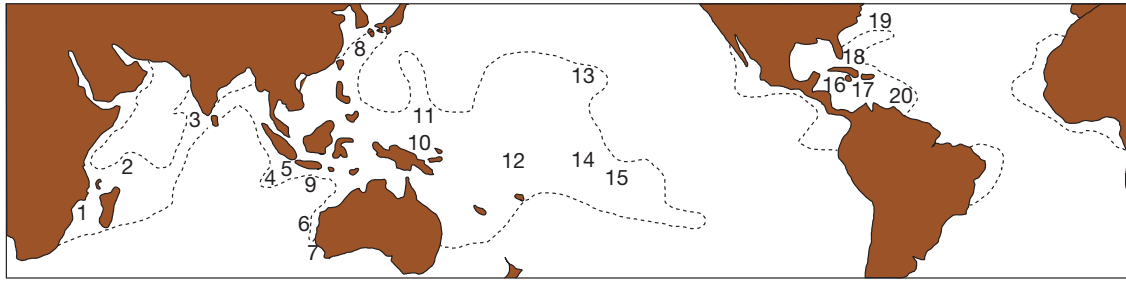


Figure 1 Present-day distribution of coral reefs (based on [Veron, 1995](#)), showing location of key sites discussed in text: 1, Comoros Islands; 2, Seychelles; 3, Maldives; 4, Cocos (Keeling) Islands; 5, Christmas Island; 6, Abrolhos Islands; 7, Rottneat Island; 8, Ryukyu Islands; 9, Sumba, Indonesia; 10, Huon Peninsula, Papua New Guinea; 11, Guam; 12, Tonga; 13, Hawaiian Islands; 14, Tahiti; 15, Cook Islands; 16, Cayman Islands; 17, Haiti; 18, Bahamas; 19, Bermuda; 20, Barbados.



Figure 2 Pleistocene reef terraces indicating former highstands of sea level: (a) the northeastern corner of Christmas Island in the Indian Ocean, uplifted by flexure as it migrates over a bulge in the ocean floor prior to subduction into the Java trench, shows a series of terraces that mark former reefal shorelines, the lowermost of which is Last Interglacial in age; (b) the coast of Mitiaro in the southern Cook Islands which is composed of Pleistocene reef limestones that show a pattern of former grooves, mirroring, though at greater spacing to, the modern spur and groove which can be seen seaward of the fringing reef crest; (c) a sequence of two Pleistocene reef limestones on the neighboring makatea island of Mauke, also in the southern Cook Islands, the lower limestone was deposited during marine oxygen isotope stage 7 and the upper (the base of which is marked by the survey staff) is Last Interglacial in age (marine oxygen isotope substage 5e); (d) an erosional notch cut into Tertiary limestone (foreground) and the Last Interglacial reef limestone (indicated by the person in the background) marks a former highstand of the sea on Cayman Brac in the Cayman Islands – similar evidence is widespread throughout many of the islands in the West Indies, especially the Bahamas; (e) a similar reef limestone terrace characterizes the shoreline in Tonga in the Pacific Ocean (here shown on Lifuka in the Ha'apai group in central Tonga); and (f) the southernmost example of such a reef terrace occurs at Fairbridge Bluff on the island of Rottneat off the coast of Western Australia, where a reef existed during the Last Interglacial beyond the latitudinal limit to which present-day corals construct modern reefs.

Mauke (Figure 2(c)), are makatea islands comprising a volcanic interior but with a rim of limestone (makatea) that contains reef units that were formed during marine oxygen isotope stages 7 and 5. These reef units occur at different heights on the different islands as a result of differential rates of flexure in response to loading by the younger volcanic island of Rarotonga (Woodroffe et al., 1991).

Particularly useful have been those shorelines that have been uplifted on which there is a flight of raised reefs generally with the higher, and more landward, ones being older. Three such terraces occur above sea level on Barbados; more occur on the rapidly uplifted Huon Peninsula in Papua New Guinea. On the basis of dating of the Huon Peninsula terraces from several sites at which there have been different rates of uplift, a history of sea-level positions has been developed, and it has been possible to link this record with that of oxygen isotopes from deep-sea cores (Chappell and Shackleton, 1986). Figure 3 shows the sea-level curve derived from the Huon Peninsula and deep-sea core evidence. More terraces are found on those coasts that are uplifting rapidly, such as at active plate margins, including islands in Indonesia such as Sumba, along the coast of Japan including the Ryukyu Islands, and in the Caribbean, for example on the island of Haiti. With refinements in dating it has been possible to extend the chronology back to identify older interglacial reefs. Reefs associated with oxygen isotope stage 7 are now widely recognized (Figure 2(c)), and reef limestones associated with stages 9 and 11 have also been dated (Siddall et al., 2006). Diagenetic changes, particularly recrystallization of coral mineralogy (aragonite to calcite), impose constraints, but the advent of thermal ionization mass spectrometry (TIMS) U-series dating, and modeling of open system behavior based on isotope ratios, have enabled greater precision for age determinations on increasingly small samples (Thompson and Goldstein, 2005).

The record of sea-level highstands derived from radiometric dating of corals from Pleistocene reef limestones provides chronological support for control on the timing of ice-sheet growth and ocean volume variations through the superimposition of variations in the eccentricity, obliquity, and precession of the earth's orbit. More recently, it offers the prospect of resolving millennial-scale, suborbital oscillations of amplitude

10–15 m that appear overprinted on the longer-term oscillations (Figure 3). Such variations have been observed during glacial periods, 55–30 ka (Yokoyama et al., 2001), and appear within interglacial reef terraces such as those on Huon Peninsula, Oahu, and Barbados during oxygen isotope stage 5 (Thompson and Goldstein, 2005).

Figure 4 illustrates how the reef limestone terraces are superimposed in different ways on shorelines that are experiencing different rates and patterns of vertical movement. Whereas fossil shorelines are stretched out on uplifting coasts, they are superimposed on coasts that are stable, such that the Last Interglacial is generally the most prominent terrace (Figures 2(d), 2(e), and 2(f)). A one-dimensional forward computer spreadsheet model of reef growth based on a sea-level compilation like that shown in Figures 3 and 4 has been developed (Koelling et al., 2009). This enables simulation of the anticipated sequence of reefs in different tectonic settings. It produces stepped reef terraces that resemble those observed on Huon Peninsula and in other settings (Figure 5).

It is also possible to identify sequences of Pleistocene reefs on shorelines that have been subsiding. Coral atolls provide one example of such a setting and here carbonate layers have formed like icing on a cake, at periods when the sea has been high. Dating of successive highstands, separated by solutional discontinuities, constrains the pattern of subsidence (Figure 4). On the one hand, these sequences provide a record of sea-level stands, on the other hand, there is a need to establish the rate of coastal uplift or subsidence, and whether it has been constant.

Last Interglacial reef limestones are prominent from many sites in the tropics and there is an emerging consensus that they formed during a relatively stable period of sea level 2–6 m above present between 128 and 116 ka (Siddall et al., 2006). However, there is less agreement about the details of smaller oscillations, as indicated by reef growth, within that highstand. Several researchers have identified what appear to be multiple reef units within the Last Interglacial. Others, through comparison of constructional reef with erosional notches, have inferred that the interglacial culminated with a rapid rise of several meters such as could occur if the West Antarctic ice sheet collapsed (Blanchon et al., 2009).

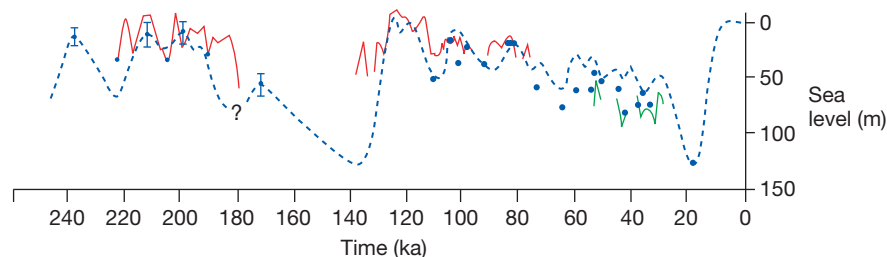


Figure 3 A curve of sea level over the past 240,000 years derived from dating of raised reef terraces on Huon Peninsula correlated with the oxygen isotope record from deep-sea cores. Reproduced from Chappell J and Shackleton NJ (1986) Oxygen isotopes and sea level. *Nature* 324: 137–140. The curve has been derived by correcting the observed elevation of fossil corals for uplift using an uplift rate inferred from the height of the Last Interglacial shoreline. Evidence for a more detailed sequence of oscillations has been proposed on the basis of further dating of corals from Huon (large dots show samples used by Chappell et al., 1996, oscillations during glacial times, green line, is after Yokoyama et al., 2001). Recently, closer scrutiny of U-series ages and their correction for open system behavior has also implied oscillations during interglacial times (red lines). Reproduced from Thompson WG and Goldstein SL (2005) Open-system coral ages reveal persistent suborbital sea-level cycles. *Science* 308: 401–404.

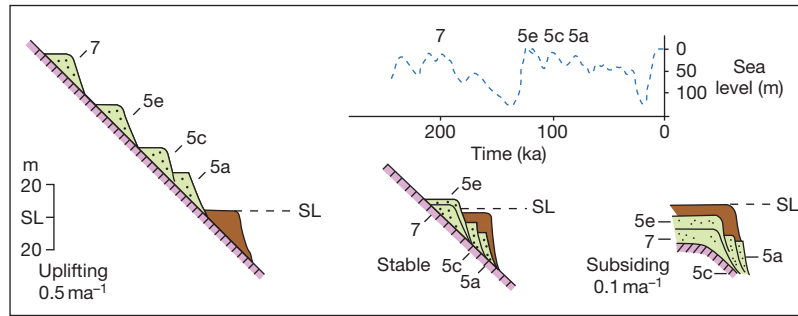


Figure 4 Schematic representation of reef terraces on shorelines that have different histories of vertical movement. The inset shows a simplified version of the sea-level curve shown in [Figure 3](#), highstands identified with marine oxygen isotope stage 7 and substages 5e, 5c, and 5a are marked. The relative location of terraces representing those highstands on shorelines that are uplifting at an average rate of 0.5 m a^{-1} , stable, and subsiding at 0.1 m a^{-1} is shown, and can be compared with the modern reef that has developed during the Holocene.

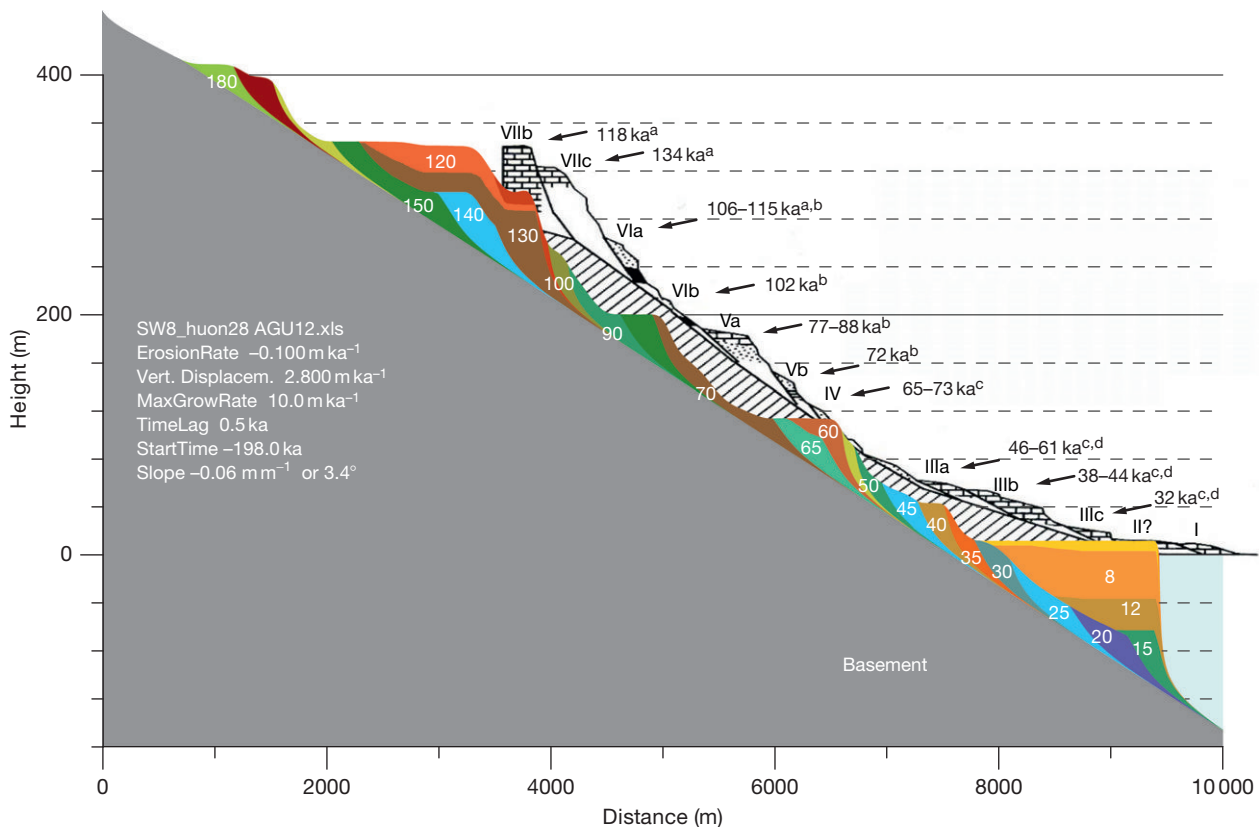


Figure 5 A one-dimensional forward computer spreadsheet model of reef growth based on a sea-level compilation like that shown in [Figures 3](#) and [4](#) has been developed. This enables simulation of the anticipated sequence of reefs in different tectonic settings. It produces stepped reef terraces that resemble those observed on Huon Peninsula (illustrated) and in other settings (based on [Koelling et al., 2009](#)).

Postglacial Sea-Level Rise

At the Last Glacial Maximum, around 20 000 years ago, sea level was of the order of 120–130 m lower than present as a result of the volume of water that was locked up in ice sheets. Reconstruction of the shoreline at this time is largely derived from morphological evidence such as submerged cliffs and notches, and deposits of shallow-water shells. Reefs occurred along what were to become the foreslopes of modern reefs and these deposits are now largely unknown or inaccessible.

The pattern of ice melt and concomitant sea-level rise during the termination, however, has been reconstructed primarily from the vertical growth of reefs that tracked the rising sea ([Figure 6](#)). The initial reconstruction was based on drilling of offshore reefs around Barbados ([Fairbanks, 1989](#)). An extended set of analyses of the deeper corals from this site, combined with a model of continental deglaciation, indicated that the Last Glacial Maximum may have experienced this period of low sea level from as early as 26 000 years ago ([Peltier and Fairbanks, 2006](#)). The transgression is recorded at several

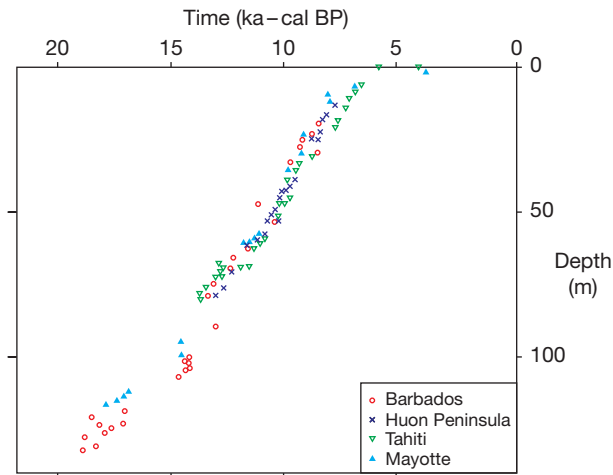


Figure 6 Record of postglacial sea-level rise from the peak of the glacial until the apparent cessation of ice melt around 6000 years ago (data from Barbados, Fairbanks, 1989; Huon Peninsula, Chappell and Polach, 1991; Tahiti, Montaggioni et al., 1997; and Mayotte, Comoro Islands, Camoin et al., 2004).

other sites. The latter part of this sea-level rise was determined simultaneously by drilling through the Holocene reef on the Huon Peninsula in New Guinea on the other side of the world, which showed a similar pattern of rise (Chappell and Polach, 1991). Subsequently, comparable long records of postglacial sea-level rise have been derived from Mayotte in the Comoros Islands and Tahiti in the Society Islands (Bard et al., 2010), with only a slightly shorter record from the Abrolhos Islands off the coast of western Australia (Eisenhauer et al., 1993). Independent support for the reconstruction of this pattern of sea-level rise comes from dating of muddy shoreline deposits such as mangrove wood and peat on the continental shelf off northwestern Australia and on the Sunda Shelf.

The curves derived from these disparate sites are remarkably similar and have been used in attempts to interpret phases of ice melt (meltwater pulses) and rapid sea-level rise (Figure 6). However, there is a limit to the rate at which reefs can accrete vertically. Although individual colonies of coral can grow rapidly, with massive corals averaging an extension of around 10 mm a^{-1} and branching corals up to 100 mm a^{-1} , the reef itself is a porous aggregation of coral containing calcareous algae and loose calcareous sediment, incompletely lithified into a structure and partially eroded by physical and biological erosion processes. Reefs rarely accrete vertically at rates of more than 10 mm a^{-1} , and in most cases at rates considerably less than this. Consequently, reefs have been unable to keep pace with the most rapid rates of sea-level rise that have been driven by pulses of meltwater, except on those coasts that are uplifting (Montaggioni, 2005). The peak of the last glaciation appears to have been terminated by rapid polar melting around 19 ka with little evidence that reefs in existence at that time were able to keep pace with the rising sea. Several phases of reef generation have been suggested separated by meltwater pulses. There is general acceptance of a rapid rise in sea level associated with meltwater pulse 1A (MWP-1A) around 14 ka when a rise of about 20 m occurred in less than 500 years, but there is less consensus in relation to subsequent periods of rapid sea-level

rise (Montaggioni and Braithwaite, 2009). One consequence of phases of rapid sea-level rise may be the drowning of reefs. Recently, extensive relict reefs have been described from water depths of around 30 m, both in the muddy waters of the southern Gulf of Carpentaria, Australia (Harris et al., 2008), and at the site of the southernmost Pacific reefs at Lord Howe Island (Woodroffe et al., 2010).

Holocene Sea-Level Variations

Whereas the early stages of postglacial sea-level rise appear to be generally similar at points around the tropics, or discrepancies can be modeled in terms of mantle viscosity, the pattern of mid- and late Holocene sea-level change varies geographically. As the rate of sea-level rise has slowed, regional disparities in the pattern of relative sea-level history have become increasingly apparent (Lambeck et al., 2010). Corals have played an important role in deciphering these sea-level changes.

A significant difference between the pattern of sea-level change in the Pacific and that in the West Indies was first recognized by Fairbridge (1961) who plotted radiocarbon ages for coral and other shoreline indicators from around the tropics onto one age–depth diagram. He interpreted the evidence to indicate a series of oscillations in sea level over the past few millennia, but the evidence for periods of sea level above its present level came primarily from Australia, whereas that for sea level remaining below present level came from Florida and the Caribbean. Since the 1970s, it has been increasingly accepted that the sea-level envelope which best describes the Pacific region reached a level close to present around 6000 years ago with evidence of a mid-Holocene sea-level highstand, whereas in the Atlantic region, sea level has been continuously rising up until present but at a decelerating rate (Adey, 1978).

It is now recognized that the earth's crust has undergone vertical movements in response to the loading of water in the ocean basins and across continental shelves. This is termed hydroisostasy. Rapid uplift of those areas that were covered by ice sheets has occurred, and intermediate field areas in what is termed the forebulge have undergone submergence as a result of compensation within the mantle (Lambeck et al., 2010). Those parts of the tropics that are in the far-field and therefore not influenced directly by the isostatic rebound that has accompanied the melting of the ice sheets, appear to have experienced a slight lowering of sea level over recent millennia. This process has been termed ocean siphoning; as the volume of water in the ocean is spread throughout the ocean basins that have increased slightly in size as a result of adjustments in the forebulge region (Mitrovica and Milne, 2002).

This is most clearly seen by the contrast in the elevation of sea-level evidence between the western Atlantic and the Indo-Pacific region. In the western Atlantic, there is no evidence for the sea level having been above its present level during the Holocene. Reefs contain fewer coral species in the West Indies than in the Pacific region. One species of staghorn coral, *Acropora palmata*, dominates the reef crest and grows most prolifically within the upper 5 m on the reef front. It has been widely used to reconstruct sea level because it is considered an indicator of shallow water (Lighty et al., 1982;

Toscano and Macintyre, 2003). Although the depth of living *A. palmata* has frequently been observed to be concentrated in this depth range, it is important to note that individual colonies of the coral are also present in deeper water. Use of material of this species recovered in cores has been further criticized because it is rarely possible to determine whether the dated material was actually in its growth position, and there are situations where fragments of this branching coral have been transported either to greater depths or onto land and deposited in storm ridges, implying that uncertainties of sea-level position could range over 10 m (Blanchon, 2005; Gischler, 2006). Figure 7 summarizes radiocarbon dating results, extending an earlier sea-level reconstruction for the West Indies and indicating sea-level rise at a decelerating rate up until present based on ages determined on *A. palmata*.

In contrast, there is widespread evidence from the Indo-Pacific region that the sea reached a level close to present at least 6000 years ago and that it has been slightly higher since then in much of the region. Figure 7 also shows radiocarbon dates from coral, and intertidal indicators such as mangrove sediments, from the northern Great Barrier Reef (Larcombe et al., 1995). The clearest evidence comes from regions where massive corals, particularly microatolls (described below), are found in their position of growth (Chappell, 1983). Whereas there are emergent reef limestones of Last Interglacial age on many islands in the Pacific, as described by Daly, it is also clear that some of the emergent reef limestones are Holocene in age and record a time of higher sea level.

Interpretation of the details of sea-level position requires a careful consideration of the evidence. On many island shorelines, there is a conglomerate of coral boulders, such as seen in Figure 8(a). However, storms are very effective at dislodging corals and transporting them, and depositing an accumulation of rubble. There has been an ongoing controversy as to the

extent to which radiocarbon dating of corals from within a boulder conglomerate provides evidence for a higher sea level. Some boulder conglomerates, containing disoriented fragments of coral, appear to be storm deposits (Figure 8(b)). Clearer evidence for the position of the sea can be discriminated where coral colonies can be unequivocally shown to be in the position in which they grew. For example, if corals and associated organisms, including *Tridacna* clams, occur in their position of growth and at an elevation above where they can now grow, the conglomerate can be inferred to have formed as part of a former reef that grew to a higher sea level (Figure 8(c)). Where fossil coral is found in its position of growth, there remains the need to establish the relationship to former sea level, often the coral is an indicative marker, indicating, as in the case of *A. palmata* that the sea surface was above the coral by anything up to about 5 m. In the Indo-Pacific, there are several corals that are considered to indicate water depths of less than 10 m; these include *A. robusta*, *A. humilis*, and *Pocillopora verrucosa* (Montaggioni and Braithwaite, 2009). Sometimes, fossil coral is found in growth position, but cannot be clearly related to a modern equivalent to verify the elevation to which it would have grown (Figure 8(d)). The vertical errors introduce uncertainties into the interpretation of former sea levels together with error terms associated with the dating techniques. Elsewhere, it may be possible to detect topographical features, such as the algal rim, or spur and groove features that characterize modern reef crests (e.g., the emergent Holocene limestone on Guam, Figure 9(a)), emerged above their modern equivalents. Encrusting vermetid gastropods, oysters, barnacles, and tubeworms have each been used as shallow-water indicators, as has the occurrence of cement types within reef deposits (Montaggioni and Braithwaite, 2009). One of the most convincing lines of evidence comes from regions where massive corals can be shown to have been constrained by former low tide levels.

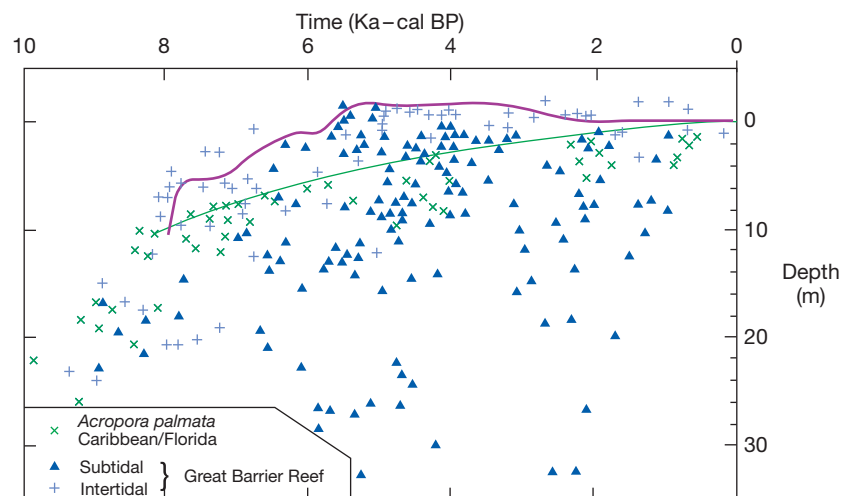


Figure 7 Holocene sea level as indicated by radiocarbon dates on coral in the West Indies and northern Australia. An Atlantic sea-level curve (green line) is shown based on ages determined on *Acropora palmata* from Florida and the West Indies (after Toscano and Macintyre, 2003). Dates on corals from the Great Barrier Reef include fossil samples that grew in shallow water as well as indicators of intertidal elevations (after Larcombe et al., 1995), and imply a sea-level curve for northeastern Australia (purple line) that reached present sea level around 6000 years ago and has been higher than present for several millennia. It is important to note that there are vertical error terms (which are not shown) associated with each sample based on uncertainty as to the extent to which it indicates the elevation of former sea level, and age errors associated with the dating techniques (error terms are generally comparable to the size of the symbol).



Figure 8 Mid-to-late Holocene sea-level evidence at, and above, modern sea level from the Indo-Pacific region. (a) Many reef rims are characterized by a conglomerate, such as the coral conglomerate, that outcrops along the seaward shore of this reef island on the Cocos (Keeling) Islands in the eastern Indian Ocean; (b) a boulder conglomerate on Suvarrow Atoll in the northern Cook Islands in which the coral boulders are disoriented and appear to have been emplaced by storm activity; (c) a conglomerate immediately adjacent to that shown above, on Suvarrow, in which the corals are in growth position, and are accompanied by other biota (such as the *Tridacna* clam immediately behind the machete), which indicates a growing mid-Holocene reef at an elevation above that to which these organisms are presently able to grow; and (d) *in situ* branches of the blue octocoral *Helopora*, exposed on Addu Atoll in the Maldives, which indicate a sea level close to, and possibly slightly above, modern.



Figure 9 Microatolls: (a) the large flat-topped corals in the foreground are microatolls on Guam, they occur in an embayment into Holocene limestone (background) in which a spur and groove pattern (like that seen on the reef front in [Figures 2\(b\)](#) and [8\(a\)](#)) is preserved; (b) microatolls of a massive species of *Porites* within a field of branching *Acropora* on the reef flat in the Cocos (Keeling) Islands; (c) large specimens of *Porites* microatolls on Cocos from which it has been possible to extract a record for most of the twentieth century; and (d) uplifted corals from the southwest coast of Simeulue, off the western Sumatra, Indonesia. The upper surface of the coral has experienced a series of changes in water level related to the Indian Ocean dipole and coseismic uplift, culminating in the entire coral being lifted out of the water in the 2004 Aceh-Andaman earthquake (see [Meltzner et al., 2006](#); photo courtesy of Kerry Sieh).

Such large flat-topped corals are called microatolls and are described in the following section.

Not all corals within a reef, however, reflect the precise level of the sea. Most corals grow in shallow water and radiocarbon dates, even if they are on corals that are in their growth position, are directional, rather than indicative, of sea-level position (Montaggioni and Braithwaite, 2009). For example, it can be seen that most ages of fossil coral recorded from drilling studies on reefs in the Great Barrier Reef, and shown in Figure 7, fall below the curve interpreted to represent sea-level position.

Reef growth can adopt one of several strategies: keep up, catch up, or give up, depending on the reef's relationship to sea level and the rate and direction of sea-level change (Neumann and Macintyre, 1985). Reef growth involves the consolidation of reefal material into a structure; where rates of sea-level rise are too rapid, typically greater than 10 mm a^{-1} , the reef is drowned. At slightly slower rates of rise, the reef is displaced upward and landward, termed backstepping. As rates of sea-level rise decelerate, reef growth may catch up with sea level as appears to have happened on many atolls and on outer reefs on the Great Barrier Reef (Hopley, 1982). Where the rate of rise is similar to, or less than, the rate of reef growth, reefs have the potential to keep up with sea level, as has occurred with the reefs around much of the Caribbean. If the sea falls, as has happened across much of the Indo-Pacific region, then a reef may become emergent, with a relatively bare reef flat replacing formerly diverse coral communities, as shown by the emergent Holocene conglomerate platform that underlies many reef islands on the margins of atolls (Figure 8(a)).

Many Holocene reefs established over Pleistocene substrates as the rise in sea level slowed around 7000 years BP. Initial coring and dating of fringing reefs indicated similarity between reefs in Hanauma Bay in Oahu, Hawaii, and those at Galeta Point in Panama. In each case, the dating implied a reef growth history in which the reef grew up and then prograded out. The interpretation that this supported a similar decelerating sea-level curve in both locations went counter to evidence elsewhere in the Pacific that there had been a mid-Holocene sea level slightly above present. This discrepancy can be resolved if the Hawaiian reef is interpreted as a catch-up reef (Fletcher and Jones, 1996).

If sea level is stable, a reef that has reached sea level generally builds seaward, called progradation. The way in which it builds seaward varies in different settings, in some cases it may involve stepped build-out, and in others it may build uniformly (Kennedy and Woodroffe, 2002). Each of these sea-level and reef-growth scenarios can be illustrated by different reefs from around the world, illustrating the range of responses that Holocene reef morphology can show to variations in the rate of sea-level change. These growth strategies have been comprehensively described by Montaggioni (2005).

Microatolls and Interannual Variations in Sea Level

Individual colonies of massive coral can grow up to a level close to mean low tide and then continue to grow outward, forming flat-topped discoid corals, termed microatolls (Figure 9(a)). Microatolls initially termed miniature atolls, are living on the outer margin but are predominantly dead and

limited by water level on their upper surface (Figure 9(b)). Microatolls in their growth position are fixed biological sea-level indicators and can be useful indicators of the level that previously limited growth, with the larger specimens reaching several meters across and containing a growth record of tens to hundreds of years (Figure 9(c)).

Fossil microatolls have been used extensively to reconstruct Holocene sea-level variations on reef tops. For example, the significance of microatolls was recognized during the 1973 Royal Society expedition to the Great Barrier Reef, northeastern Australia, when it was realized that those found above sea level indicated a higher stand of the sea (Scoffin and Stoddart, 1978). Further study showed that sea level has been decreasing over the past 6000 years with respect to the inner Great Barrier Reef (Chappell, 1983). Much of the Indo-Pacific reef province has experienced relative sea levels in the mid- and late Holocene that have been slightly above present. Microatolls have sometimes been preserved at a height presently above that of their living counterparts on reef flats in the eastern Indian Ocean, Southeast Asia, northern Australia, and across much of the equatorial Pacific Ocean (Smithers and Woodroffe, 2000).

Typically, microatolls comprise a single colony of massive *Porites*, though several other genera can also adopt the microatoll form and grow to several meters in diameter. Microatolls grow upward until constrained by a water level close to mean low tide level, after which they continue lateral growth and become disk-shaped corals (Figure 10(a)), contrasting with dome-shaped colonies that are in water deep enough not to be constrained by sea level (Figure 10(b)). On coasts that experience seismic uplift, a coral previously not limited by sea level may be elevated above the water level, and the exposed upper surface will die, but with continued lateral growth at a lower elevation (Figure 10(c)). A spectacular example of this occurred in association with the 2004 Aceh-Andaman earthquake with uplift on the island of Simeulue, off the western coast of Sumatra (Meltzner et al., 2006; see Figure 9(d)). Where a microatoll, previously limited by water level, experiences an increase in the level of the sea, it can resume vertical growth and begin to overgrow the formerly dead upper surface (Figure 10(d)). Successive falls in water level result in exposure of the living rim, and a series of terracettes develop, recording the fall in water level (Figures 10(e)). Fluctuations of water level with a periodicity of several years are recorded on the upper surface of a microatoll as a series of concentric undulations or annuli (Figure 10(f)). This pattern has been described on microatolls from reef flats in the central Pacific tracking, but lagging behind, sea-level fluctuations that are linked to the atmospheric-oceanic El Niño phenomenon that is a major control on the ocean surface in the central Pacific Ocean (Woodroffe and McLean, 1990).

Banding within the coral provides insight into events in the life history of the coral. It can reveal times at which growth has been lowered by changes in the water surface. In areas where storms are experienced, overturning of the colony can occur during individual storms, or microatolls may have responded to the moating of water that could have occurred behind boulder ramparts formed as a result of storms. In these cases, their upper surface may record elevation of water level within impounded moats above that of regional sea level. Although in

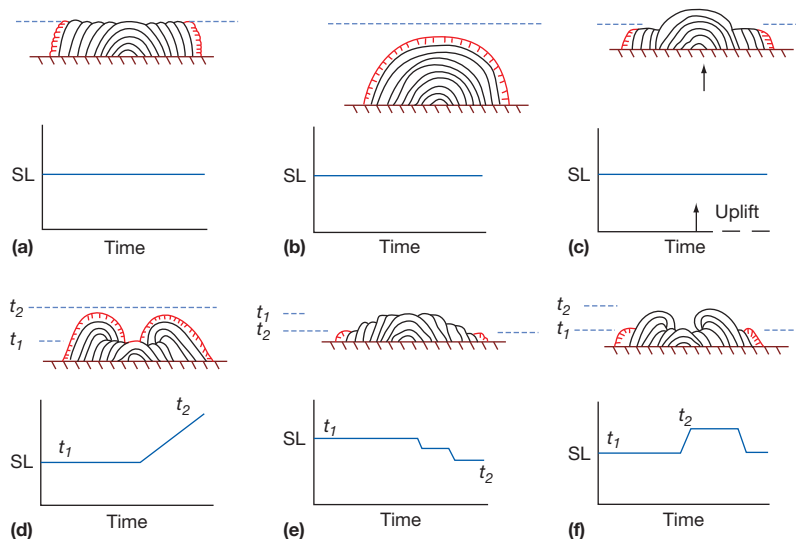


Figure 10 A schematic illustration of the response of the upper surface of microatolls to changes of sea level or uplift of the land, with annual bands (laid down by the living tissue that is shown in red) within the coral skeleton recording the pattern of coral growth; (a) if sea level remains constant from year to year, a massive coral that has grown up to sea level continues to grow outward but with its upper surface constrained at water level; (b) if the coral does not reach water level, it adopts a domed growth form and is not constrained by water level; (c) if the coast undergoes uplift, then a coral previously not limited by sea level may be raised, and the exposed upper surface will die, but with continued lateral growth at a lower elevation (as in [Figure 9\(d\)](#)); (d) if the water level increases, a coral previously constrained by exposure at low tides can resume vertical growth and begin to overgrow the formerly dead upper surface; (e) if water level decreases, then the microatoll adopts a series of terraces; and (f) if there are fluctuations of water level with a periodicity of several years, then the upper surface of the microatoll consists of a series of concentric undulations – such a pattern can be seen on microatolls from reef flats in the central Pacific where El Niño results in interannual variations in sea level (based on [Woodroffe and McLean, 1990](#); [Lambeck et al., 2010](#)).

open-water situations microatolls can enable centimeter-scale reconstructions of former sea level, they are subject to misinterpretation if moating has occurred, and it is therefore important to assess the geomorphological setting within which these corals grew before using fossil specimens to draw conclusions about past sea levels. In addition to tracking sea level, emerged microatolls have been used to record uplift associated with earthquakes, as for example, along the Sumatran coast ([Sieh et al., 2008](#)), and may contain evidence of fluctuations in sea level at decadal to century time scales ([Yu et al., 2009](#)).

See also: U-Series Dating. **Paleoceanography, Biological Proxies:** Corals, Sclerosponges and Mollusks. **Radiocarbon Dating:** AMS Radiocarbon Dating; Conventional Method. **Sea Level Studies:** Eustatic Sea-Level Changes – Glacial–Interglacial Cycles; Eustatic Sea-Level Changes Since the Last Glacial Maximum; Geomorphological Indicators; Isostasy; Glaciation-Induced Sea-Level Change. **Sea-Levels, Late Quaternary:** Late Quaternary Relative Sea-Level Changes in the Tropics.

References

- Adey WH (1978) Coral reef morphogenesis: A multidimensional model. *Science* 202: 831–837.
- Bard E, Hamelin B, and Delanghe-Sabatier D (2010) Deglacial Meltwater Pulse 1B and Younger Dryas sea levels revisited with boreholes at Tahiti. *Science* 327: 1235–1237.
- Bard E, Hamelin B, Fairbanks RG, and Zindler A (1990) Calibration of the ^{14}C timescale over the past 30,000 years using mass spectrometric U–Th ages from Barbados corals. *Nature* 345: 405–410.
- Blanchon P (2005) Comments on "Corrected western Atlantic sea-level curve for the last 11,000 years based on calibrated ^{14}C dates from *Acropora palmata* framework and intertidal mangrove peat" by Toscano and Macintyre. *Coral Reefs* 24: 183–186.
- Coral Reefs* (2003) 22:257–270.
- Blanchon P, Eisenhauer A, Fietzke J, and Liebetrau V (2009) Rapid sea-level rise and reef back-stepping at the close of the last interglacial highstand. *Nature* 458: 881–885.
- Camoin GF, Montaggioni LF, and Braithwaite CJR (2004) Late glacial to post glacial sea levels in the Western Indian Ocean. *Marine Geology* 206: 119–146.
- Chappell J (1983) Evidence for smoothly falling sea levels relative to north Queensland, Australia, during the past 6000 years. *Nature* 302: 406–408.
- Chappell J, Omura A, Esat T, et al. (1996) Reconciliation of late Quaternary sea levels derived from coral terraces at Huon Peninsula with deep sea oxygen isotope records. *Earth and Planetary Science Letters* 141: 227–236.
- Chappell J and Polach H (1991) Post glacial sea level rise from a coral record at Huon Peninsula, Papua New Guinea. *Nature* 349: 147–149.
- Chappell J and Shackleton NJ (1986) Oxygen isotopes and sea level. *Nature* 324: 137–140.
- Daly RA (1915) The glacial-control theory of coral reefs. *Proceedings of the American Academy of Arts and Science* 51: 155–251.
- Eisenhauer A, Wasserburg GJ, Chen JH, et al. (1993) Holocene sea-level determination relative to the Australian continent: U/Th (TIMS) and ^{14}C (AMS) dating of coral cores from the Abrolhos Islands. *Earth and Planetary Science Letters* 114: 529–547.
- Fairbanks RG (1989) A 17,000-year glacio-eustatic sea level record: Influence of glacial melting rates on the Younger Dryas event and deep-ocean circulation. *Nature* 342: 637–642.
- Fairbridge RW (1961) Eustatic changes in sea level. *Physics and Chemistry of the Earth*, pp. 99–185. New York: Pergamon Press.
- Fletcher CH and Jones AT (1996) Sea-level highstand recorded in Holocene shoreline deposits on Oahu, Hawaii. *Journal of Sedimentary Research* 66: 632–641.
- Gischler E (2006) Comment on "Corrected western Atlantic sea-level curve for the last 11,000 years based on calibrated ^{14}C dates from *Acropora palmata* framework and intertidal mangrove peat" by Toscano and Macintyre. *Coral Reefs*

- 22: 257–270 (2003), and their response in *Coral Reefs* 24: 187–190 (2005). *Coral Reefs* 25: 273–279.
- Harris PT, Heap AD, Marshall JF, and McCulloch M (2008) A new coral reef province in the Gulf of Carpentaria, Australia: Colonisation, growth and submergence during the early Holocene. *Marine Geology* 251: 85–97.
- Hopley D (1982) *The Geomorphology of the Great Barrier Reef: Quaternary Development of Coral Reefs*. New York: Wiley.
- Hopley D (ed.) (2011) *Encyclopedia of Modern Coral Reefs*, p. 1236. Berlin: Springer.
- Hopley D, Smithers SG, and Parnell K (2007) *Geomorphology of the Great Barrier Reef: Development, Diversity and Change*, p. 546. Cambridge: Cambridge University Press.
- Kennedy DM and Woodroffe CD (2002) Fringing reef growth and morphology: A review. *Earth-Science Reviews* 57: 257–279.
- Koelling M, Webster JM, Camoin G, Iryu Y, Bard E, and Seard C (2009) SEALEX – Internal reef chronology and virtual drill logs from a spreadsheet-based reef growth model. *Global and Planetary Change* 66: 149–159.
- Lambeck K, Woodroffe CD, Antonioli F, et al. (2010) Paleoenvironmental records, geophysical modeling, and reconstruction of sea-level trends and variability on centennial and longer timescales. In: Church J, Woodworth PL, Aarup T, and Wilson WS (eds.) *Understanding Sea-Level Rise and Variability*, pp. 61–121. Chichester: Wiley Interscience.
- Larcombe P, Carter RM, Dye J, Gagan MK, and Johnson DP (1995) New evidence for episodic post-glacial sea-level rise, central Great Barrier Reef, Australia. *Marine Geology* 127: 1–44.
- Lighty RG, Macintyre IG, and Stuckenrath R (1982) *Acropora palmata* reef framework: A reliable indicator of sea level in the western Atlantic for the past 10,000 years. *Coral Reefs* 1: 125–130.
- Meltzner AJ, Sieh K, Abrams M, et al. (2006) Uplift and subsidence associated with the great Aceh-Andaman earthquake of 2004. *Journal of Geophysical Research* 111: B02407. <http://dx.doi.org/10.1029/2005JB003891>.
- Mitrovica JX and Milne GA (2002) On the origin of late Holocene sea-level highstands within equatorial ocean basins. *Quaternary Science Reviews* 21: 2179–2190.
- Montaggioni LF (2005) History of Indo-Pacific coral reef systems since the last glaciation: Development patterns and controlling factors. *Earth-Science Reviews* 71: 1–75.
- Montaggioni LF and Braithwaite CJR (2009) *Quaternary Coral Reef Systems: History, Development Processes and Controlling Factors*, p. 532. Amsterdam: Elsevier.
- Montaggioni LF, Cabioch G, Camoinau GF, et al. (1997) Continuous record of reef growth over the past 14 k.y. on the mid-Pacific island of Tahiti. *Geology* 25: 555–558.
- Neumann AC and Macintyre I (1985) Reef response to sea level rise: keep-up, catch-up or give-up. *Proceedings of the 5th International Coral Reef Congress* 3: 105–110.
- Peltier WR and Fairbanks RG (2006) Global glacial ice volume and Last Glacial Maximum duration from an extended Barbados sea level record. *Quaternary Science Reviews* 25: 3322–3337.
- Scoffin TP and Stoddart DR (1978) The nature and significance of micro atolls. *Philosophical Transactions of the Royal Society, London B* 284: 99–122.
- Siddall M, Chappell J, and Potter EK (2006) Eustatic sea level during past interglacials. In: Sirocko F, Clausen M, Sanchez-Goni MF, and Litt T (eds.) *The Climate of Past Interglacials, Developments in Quaternary Science*, vol. 7, pp. 75–92. Amsterdam: Elsevier.
- Sieh K, Natawidjaja DH, Meltzner AJ, et al. (2008) Earthquake supercycles inferred from sea-level changes recorded in the corals of West Sumatra. *Science* 322: 1674–1678.
- Smithers SG and Woodroffe CD (2000) Microatolls as sea-level indicators on a mid-ocean atoll. *Marine Geology* 168: 61–78.
- Thompson WG and Goldstein SL (2005) Open-system coral ages reveal persistent suborbital sea-level cycles. *Science* 308: 401–404.
- Toscano MA and Macintyre IG (2003) Corrected western Atlantic sea-level curve for the last 11,000 years based on calibrated ¹⁴C dates from *Acropora palmata* framework and intertidal mangrove peat. *Coral Reefs* 22: 257–270.
- Veeh HH (1966) Th 230/U238 and U234/U238 ages of Pleistocene high sea level stand. *Journal of Geophysical Research* 71: 3379–3386.
- Veron JEN (1995) *Corals in Space and Time: The Biogeography and Evolution of the Scleractinia*. Sydney: University of New South Wales Press.
- Woodroffe CD, Brooke BP, Linklater M, et al. (2010) Response of coral reefs to climate change: Expansion and demise of the southernmost Pacific coral reef. *Geophysical Research Letters* 37: L15602. <http://dx.doi.org/10.1029/2010GL044067>.
- Woodroffe CD and McLean RF (1990) Microatolls and recent sea level change on coral atolls. *Nature* 344: 531–534.
- Woodroffe CD, Short S, Stoddart DR, Spencer T, and Harmon RS (1991) Stratigraphy and chronology of Pleistocene reefs in the southern Cook Islands, South Pacific. *Quaternary Research* 35: 246–263.
- Yokoyama Y, Esat TM, and Lambeck K (2001) Last glacial sea-level change deduced from uplifted coral terraces of Huon Peninsula, Papua New Guinea. *Quaternary International* 83: 275–283.
- Yu K-F, Zhao J-X, Done T, and Chen T-G (2009) Microatoll record for large century-scale sea-level fluctuations in the mid-Holocene. *Quaternary Research* 71: 354–360.

Microfossil-Based Reconstructions of Holocene Relative Sea-Level Change

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Introduction

Microfossils are an important tool for reconstruction of former sea levels, particularly in the Holocene. They are valuable sea-level indicators because their modern counterparts (i.e., microorganisms) are found in narrowly defined niches in the intertidal zone at specific height (or depth) ranges. The vertical living ranges of microorganisms such as Foraminifera, diatoms, and testate amoebae are controlled by limiting ecological parameters, including tidal flooding frequency, subaerial exposure, food availability, substrate composition, and salinity. These parameters play a role in interspecific competition, which results in vertical intertidal zonation of species assemblages (Figure 1). If the upper and lower limits of the vertical ranges of microorganisms are measured in the modern intertidal environment, it is possible to establish, from sediments containing the fossilized microorganisms, the elevation relative to sea level at which the sediments were deposited. This is an example of the principle of uniformitarianism: 'the present is the key to the past.'

There are many published Holocene sea-level studies in which microfossils are employed and this review does not attempt to be comprehensive. Some key references are included, while most illustrated examples in this review are chosen to highlight important concepts. The case studies are primarily from the coastal lowlands of western Europe and the salt marshes of eastern North America, but they are relevant for sea-level studies in comparable environments elsewhere.

This paper covers the application of Foraminifera, diatoms, testate amoebae, ostracods, and pollen in Holocene sea-level reconstructions. Testate amoebae and ostracods are less often used than Foraminifera and diatoms. However, testate amoebae in particular have the potential to deliver very precise Holocene sea-level information from salt-marsh sediments. Pollen analysis is commonly applied in sea-level studies, but pollen grains cannot be regarded as *in situ* sea-level indicators in the same sense as the other microfossil groups. Finally, the recent development of quantitative methods (i.e., the use of transfer functions) and the application of chronological techniques are briefly discussed. The transfer function methodology has become quite popular in the last decade and some advantages and pitfalls associated with the application of transfer functions are highlighted.

Microfossil Groups Used in Sea-Level Studies

Foraminifera

Foraminifera are single-celled organisms that are found in most marine environments, from the intertidal zone to the deep ocean. Shallow water Foraminifera are most useful for sea-level studies as their living range can be most easily related to sea level (Gehrels, 1994). Environmental conditions in the

intertidal zone produce intraspecific competition which results in a foraminiferal vertical zonation, often with narrow vertical ranges of distinct assemblages (Figure 1). The limiting factor that controls foraminiferal zonation is tolerance to subaerial exposure and only the hardest agglutinated Foraminifera (e.g., *Jadammina macrescens* in marshes in eastern North America) are capable of surviving in the uppermost intertidal zone. The potential of salt-marsh Foraminifera as sea-level indicators was first highlighted by Scott and Medioli (1978). Many studies have shown that intertidal foraminiferal zonation occurs in tidal marshes around the world, from high latitude salt marshes to low latitude mangrove environments (Gehrels, 2002). The most precise sea-level reconstructions based on Foraminifera have been produced in microtidal salt marshes on the eastern seaboard of North America (e.g., Gehrels et al., 2005; Kemp et al., 2009) and New Zealand (Gehrels et al., 2008). These reconstructions form a link between geological reconstructions and instrumental observations. They have revealed that ongoing sea-level rise represents a significant departure from slower rates of sea-level rise in the late Holocene. This sea-level acceleration started during the early twentieth century (Woodworth et al., 2009).

A distinct advantage of the use of Foraminifera as sea-level indicators is that in salt marshes abundances are generally high, while species diversity is low. The agglutinated species that are found in the upper parts of salt marshes are also well preserved in fossil sediments. Some agglutinated species have a universal occurrence, most notably *J. macrescens*, *Trochammina inflata*, and *Miliammina fusca*. Foraminifera are less useful in sediments from lower in the intertidal zone. Here calcareous genera are found (e.g., *Elphidium*, *Ammonia*, *Haynesina*, *Quinqueloculina*) and these are not well preserved in sediments due to dissolution, especially when sediments are organic in nature and acidic conditions are present. In deeper waters, the ecological controls that produce intertidal zonation become less influential and reduce the sea-level indicative value of Foraminifera.

Diatoms

Diatoms are single-celled algae that are found in almost every aquatic environment, both freshwater and marine. Diatom valves are built from silica and their preservation potential in fossil sediments is high. Like Foraminifera, they occupy limited environmental ranges in the modern intertidal zone (Nelson and Kashima, 1993). However, diatoms can tolerate both fresh water and marine conditions and their distributions are controlled by a number of environmental variables, including salinity, substrate type, temperature, and nutrient supply. Tidal conditions produce spatially variable conditions in the intertidal zone to which diatoms adapt. As a consequence, distinct diatom assemblages are found within narrow elevational ranges which makes them suitable as sea-level indicators.

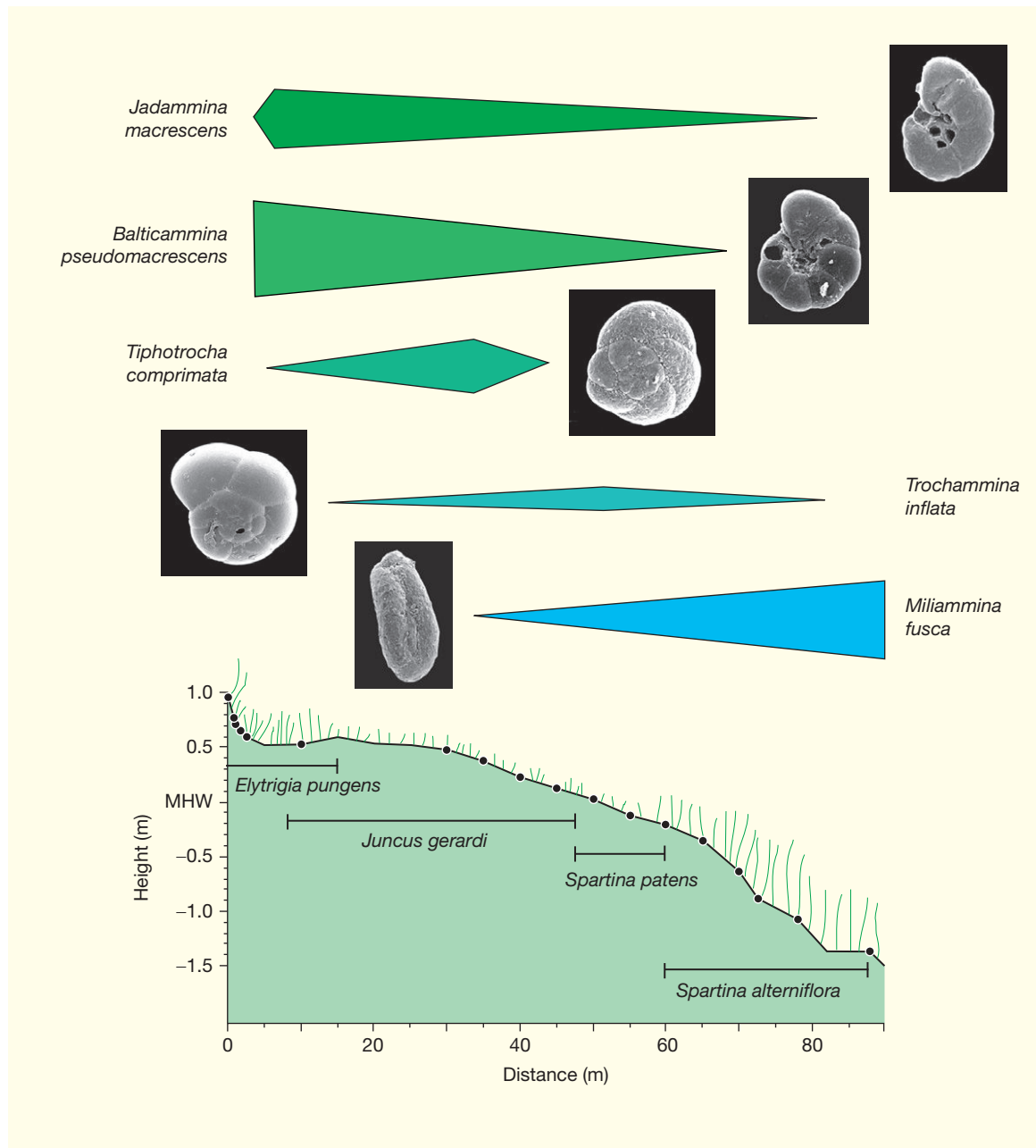


Figure 1 An example of intertidal zonation displayed by Foraminifera in samples collected along the sloping surface of a salt marsh in eastern Maine (USA). Plant zones are also shown. Redrawn and modified from [Gehrels WR \(1994\)](#) Determining relative sea-level change from salt-marsh foraminifera and plant zones on the coast of Maine, U.S.A. *Journal of Coastal Research* 10: 990–1009.

Diatoms have been classified into salinity classes; where halophobes are freshwater diatoms, oligohalobes can tolerate low salinities up to 2‰, mesohalobes live in brackish water between salinities of 2 and 30‰, and polyhalobes are found in marine waters with salinities greater than 30‰ ([Palmer and Abbott, 1986](#)). However, ecological classifications of diatoms are still under debate. For example, more recent classification schemes have been extended to include other ecological parameters in addition to salinity ([Denys, 1991/2; Vos and de Wolf, 1993](#)). Changes in the relative abundances of these groups in sediment are indicative of changing marine

influence, often interpreted as transgressions or regressions. A classic diatom-based study of Holocene sea-level changes is the work by [Devoy \(1979\)](#). More recent studies increasingly use quantitative methods such as transfer functions (e.g., [Long et al., 2010; Sawai et al., 2004](#)).

Testate Amoebae

Testate amoebae are single-celled organisms that live in many terrestrial and aquatic habitats, including peat bogs, soils, lakes, and salt marshes. In the literature, they are also referred

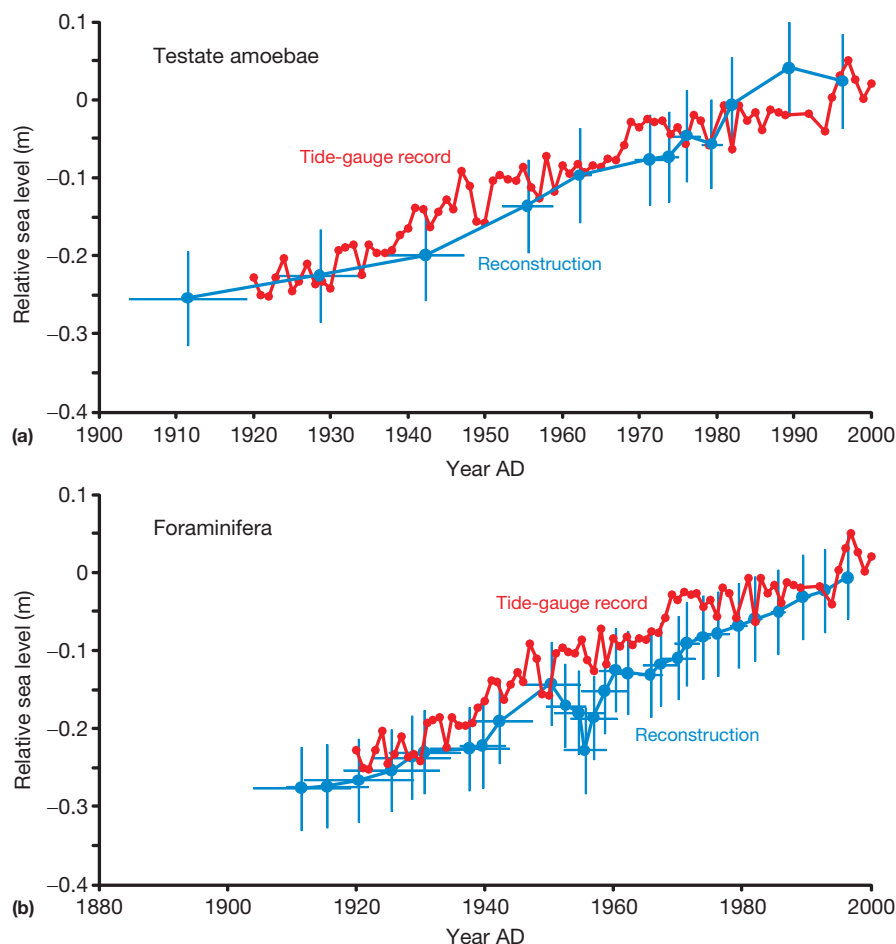


Figure 2 Sea-level reconstructions for Chezzetcook (Nova Scotia, Canada), based on (a) salt-marsh testate amoebae and (b) salt-marsh Foraminifera. The reconstructions are compared with the tide-gauge record from nearby Halifax. Redrawn and modified from Charman DJ, Gehrels WR, Manning C, and Sharma C (2010) Reconstruction of recent sea-level change using testate amoebae. *Quaternary Research* 73: 208–291.

to as rhizopods, thecamoebians, and arcellaceans. Traditionally, testate amoebae have been considered to be exclusively freshwater organisms, although some taxa have occasionally been found in coastal marshes in studies that focused on Foraminifera (e.g., Scott et al., 1991). These testate amoebae were greater than 63 μm in size, as this size fraction is commonly used for foraminiferal analyses. However, studies in salt marshes in the United Kingdom and eastern North America have shown that testate amoebae smaller than 63 μm occur in very high abundances with much greater species diversities (Charman et al., 1998, 2002; Gehrels et al., 2006a; Roe et al., 2002). The preservation potential of testate amoebae is good in recent salt-marsh sediments, coastal raised bogs, and isolation basins, but in other coastal sediments (e.g., coastal reed swamps) preservation may be variable (Roe et al., 2002). An important environmental control on the distribution of testate amoebae is wetness, while salinity appears to limit their occurrence to the uppermost part of salt marshes (Charman et al., 2002). Testate amoebae are promising sea-level indicators. In the few places where they have been studied they occupy narrower intertidal niches than any other microfossil group (Charman et al., 1998; Gehrels et al., 2001, 2006a).

Charman et al. (2010) produced the first sea-level reconstructions based on testate amoebae found in short salt-marsh cores for two sites in eastern North America. The reconstructions are only 100 years long, but compare favorably with tide-gauge records and with reconstructions based on foraminiferal analyses (Figure 2). Importantly, assemblages of testate amoebae show strong similarities on both sides of the Atlantic Ocean, which enhances their applicability across large geographical areas. Important taxa with widespread occurrences in the upper salt marsh include *Tracheleuglypha dentata*, *Trinema lineare*, and *Euglypha rotunda*. *Centropyxis cassis*, *Centropyxis platystoma*, *Diffugia pristis*, and *Cyphoderia ampulla* are common in the high to middle salt marsh. In most salt marshes, testate amoebae are confined to the highest part of the marsh (and above) and are not found below the level of mean high spring tide, although occasionally they can occur as low as the mean high water level (Gehrels et al., 2006a).

Ostracods

Ostracods are small crustaceans, usually between 0.3 and 1.5 mm in size, that produce bivalved calcareous shells. They are found in both fresh and marine waters. Like Foraminifera,

marine ostracods occur in the entire marine realm, from intertidal areas to the deep ocean. Assemblages appear to be restricted to certain depth ranges. However, they are not considered to be precise sea-level indicators, mostly because the shells are easily transported (van Harten, 1986) and because many environmental parameters control their distributions, including salinity, water chemistry, temperature, and substrate (Frenzel and Boomer, 2005). Possible exceptions are assemblages that are found in intertidal environments. Here, species are generally restricted to low-marsh habitats and tidal channels and pools, and their use in sea-level applications is often in combination with Foraminifera. Frenzel and Boomer (2005) list many examples of the use of ostracods in Holocene sea-level studies.

Pollen

Pollen are the first microfossil group used in sea-level reconstructions (e.g., Godwin, 1940), albeit as a chronological tool. Pollen grains are transported, by wind or water, and reconstructing specific tidal levels from sediments based on pollen assemblages is therefore challenging. However, pollen assemblages are valuable in giving a general indication of the position of the sediments within the tidal frame. For example, many salt-marsh sediments contain a distinct pollen assemblage, including Chenopodoaceae and *Plantago maritima* and a range of other pollen types (Roe and van de Plassche, 2005), which allows differentiation from other environments. If the indicative meaning of a sea-level index point is based on pollen alone, the vertical uncertainties can be as large as the entire span of the upper intertidal zone, between mean high water and the highest astronomical tide level. Possible exceptions are mangrove environments, where other microfossils are often poorly preserved. Engelhart et al. (2007) found that pollen distributions in mangrove swamps in Indonesia appear to be zoned and well preserved in cores and could therefore be used to reconstruct sea-level changes.

Seminal pollen-based Holocene sea-level investigations were conducted by Tooley (1978) and Shennan (1982). Before the advent of radiocarbon dating, pollen analyses were used for providing a chronology based on widely recognized vegetation changes (Godwin, 1940). Pollen can be used to improve chronology if vegetational change provides a well preserved marker which has been dated in other locations. In coastal sediments in the United Kingdom, the Alder rise provides an early Holocene chronohorizon, while the Elm decline is a useful marker for middle Holocene sediments. For more recent coastal sediments, European settlement and forest clearance have provided some useful pollen markers along the eastern seaboard of North America (Gehrels et al., 2005; Kemp et al., 2009) and in New Zealand (Gehrels et al., 2008). In eastern North America, the expansion of ragweed (*Ambrosia*), dock (*Rumex*), and the decline of oak (*Quercus*) provide the most notable pollen signals. Figure 3 shows a pollen diagram from a salt marsh at Chezzetcook, Nova Scotia, eastern Canada. The arrival of settlers around AD 1780 left a distinct pollen marker which can be used to date the sediments (Gehrels et al., 2005). In some European coastal sites where pine plantations have been established, a recent rise in pine pollen provides a useful historical marker (Long et al., 1999).

In many sea-level studies from coastal lowlands, pollen analyses are often performed across boundaries between peat and minerogenic sediments (e.g., Shennan, 1982). Sea-level index points may be obtained from such boundaries if they are shown to represent a marine–terrestrial contact and if it can be demonstrated that the boundary represents an *in situ* stratigraphic contact. Pollen analyses are useful for the latter. For example, if the reconstructed vegetational succession contains a clear salt-marsh assemblage, sandwiched in between tidal flat and freshwater marsh deposits, the contact can be assumed to be nonerosional. Even if a stratigraphic contact from which a sea-level index point is derived is shown to be nonerosional, other factors, such as autocompaction, could still produce displacement and should be considered.

Methods of Microfossil-Based Sea-Level Reconstruction

Derivation of Sea-Level Index Points

A sea-level index point may be established when a fossil sea-level indicator is dated, its height is measured, and the height relative to sea level of its modern counterpart (the ‘indicative meaning’) is known. In theory therefore, any biota that is preserved in the fossil record and for which an indicative meaning and an age can be established can form the basis for a sea-level reconstruction. Microfossils are especially valuable for sea-level reconstructions in coastal lowlands where alternating beds of terrestrial and marine deposits provide evidence for transgressions and regressions. In these environments, sea-level index points are obtained from the contacts between peat and silt/clay beds, as these contacts are formed very high in the intertidal zone. The indicative meaning of these stratigraphic contacts is derived from the altitude of the boundary between reed beds and salt marsh in the modern environment which, in the English Fenland for example, is found very near the level of the mean high water of spring tides (Shennan, 1982). Microfossil analyses are useful to corroborate and further quantify the indicative meaning. The distribution of microorganisms in the modern environment forms the basis for achieving this (as discussed above; Gehrels et al., 2001).

A distinct disadvantage of using sea-level index points from peat-silt/clay contacts is that Holocene sequences of coastal deposits are subject to long-term consolidation. This process displaces the original position of sea-level index points downward and leads to overestimation of the rate of relative sea-level rise as sea-level index points from intercalated coastal deposits represent only limiting positions of sea level (Horton and Shennan, 2009). When sea-level index points are derived from a peat unit directly overlying a hard substrate (‘basal peat’), this problem is significantly reduced or even eliminated (Gehrels, 1999). Basal peat is formed at the leading edge of the Holocene transgression and its indicative meaning therefore is close to the highest astronomical tide. However, unless the upper limit of reed beds is known in the modern environment, sea-level index points derived from basal peat may still only represent limiting sea-level positions (i.e., sea level lies below the sea-level index point). This is an issue in the coastal plains of Belgium, the Netherlands, and Germany for example, where coastal fens have been reclaimed and the true indicative

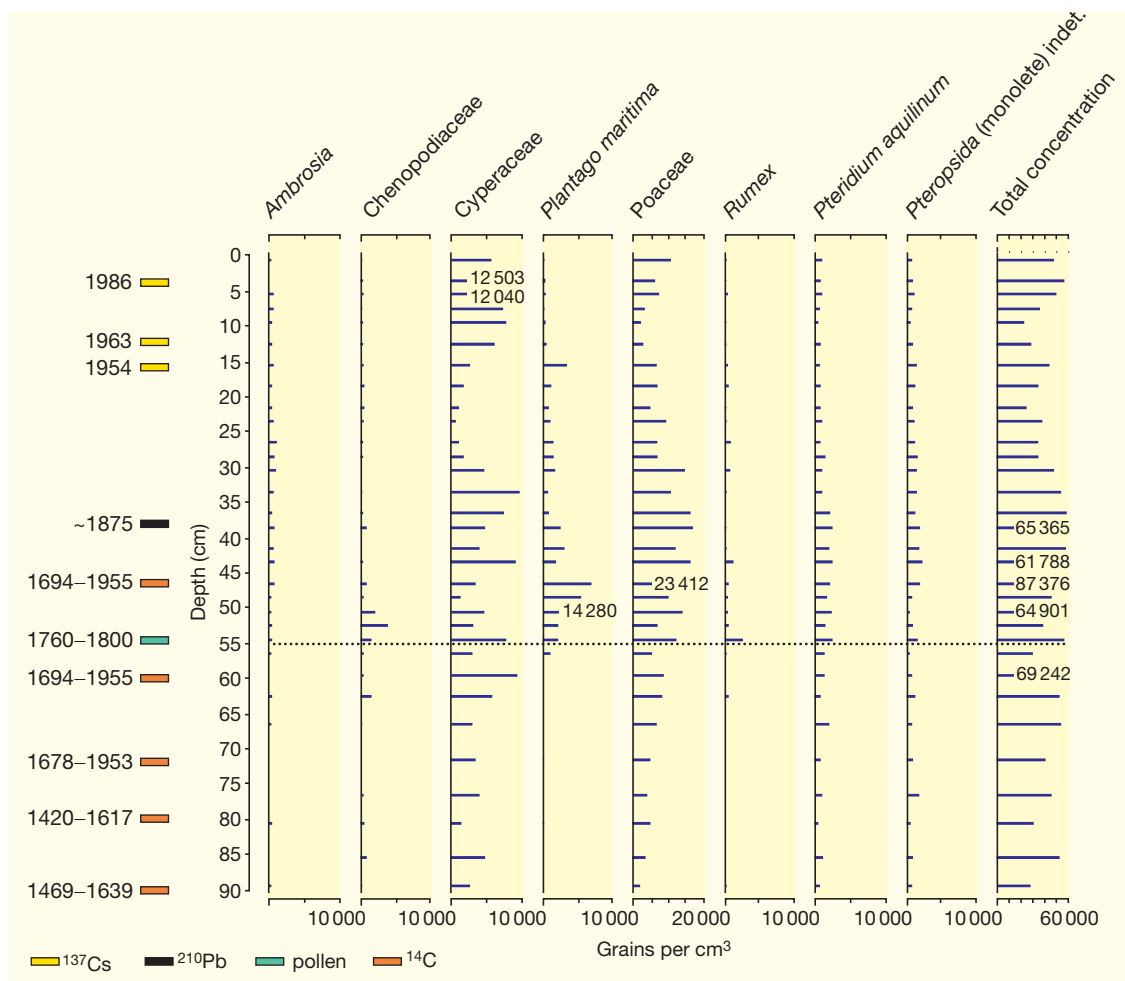


Figure 3 Pollen diagram from a salt marsh in Chezzetcook, Nova Scotia, Canada, used to establish a sea-level history (Gehrels et al., 2005). Deforestation following European settlement around 1780 encouraged the spread of weeds (*Ambrosia*, *Rumex*) and produced a useful pollen marker which ties in with other chronological tools. The cultivation of hay has also increased the concentration of grass pollen (*Poaceae*). Redrawn and modified from Gehrels WR, Kirby JR, Prokoph A, et al. (2005) Onset of recent rapid sea-level rise in the western Atlantic Ocean. *Quaternary Science Reviews* 24: 2083–2100.

meaning of basal peat cannot be accurately determined. The situation is more advantageous if basal peat deposits are found that contain Foraminifera distinctive of salt-marsh environments (e.g., Gehrels et al., 2005) and if pristine environments from which the indicative meaning of microfossils can be established are abundantly present.

Transgressive and Regressive Sequences

Biostratigraphic changes in salt-marsh sediments can provide detailed continuous records of past sea-level changes (Thomas and Varekamp, 1991). Many sea-level records based on biostratigraphical methods have been derived from salt marshes along the Atlantic coast of North America, primarily North Carolina, Connecticut, Maine, and Nova Scotia. The provision of accommodation space by sea-level rise generates the potential for salt-marsh sediment to record sea-level changes, a concept that is well known in seismic stratigraphy over larger geological time scales. Transgressions in salt marshes are recorded when

sea-level rise outpaces accretion. Alternatively, regressions are recorded when accretion rates are faster than sea-level rise. The corollary of this is that, if accretion rates can be measured by dating methods, it is possible to calculate sea-level rise. If auto-compaction is significant, it is even possible to detect small sea-level falls, provided that autocompaction can be determined, for example, by dating basal peat (Gehrels, 1999). Both qualitative and quantitative methods have been used to detect biostratigraphical changes and relate these to changes in sea level.

Whatever the method of reconstruction, quantitative or qualitative, microfossils used in sea-level reconstruction should ideally be derived from sediments that have formed in the uppermost intertidal zone. Here, the relationship between elevation and species assemblages is most strongly defined. In deeper waters, the ecological parameter primarily responsible for the correlation between elevation and foraminiferal assemblages, that is, flooding duration, is no longer the limiting factor for species distribution and other environmental controls (e.g., nutrient levels, substrate type, water temperature,

wave energy, etc.) become more influential (Woodroffe, 2009). When a large number of environmental factors need to be taken into account, the risk that any of these has not remained constant over time increases and the uniformitarian assumptions of microfossil-based sea-level reconstructions may (unknowingly) be violated. Species diversity also tends to increase in deeper water, so that it is often not possible to define the lower limit of indicator species, a problem that makes it difficult to obtain accurate sea-level reconstructions. Other sedimentary factors that can complicate sea-level reconstructions from low intertidal (or even subtidal) sediments include erosion and rapid changes in sedimentation rate. Near the highest tide mark, changes in accommodation space are primarily produced by relative sea-level change, and the rate of sediment accretion approaches that of the rate of sea-level rise (Allen, 1990), although the process of autocompaction should also be considered.

Isolation Basins

In areas of isostatic uplift, sea-level changes can be reconstructed from microfossils present in basins that were isolated from the sea at some stage in their uplift history. The isolation process produces distinct stratigraphic evidence preserved in contemporary lakes or raised bogs, with marine deposits buried beneath freshwater lake deposits or terrestrial peat. The isolation event is marked in the stratigraphy by the contact between the marine and terrestrial deposits. The dating of the contact provides the age of isolation, whereas the height of the sill that separates the basin from the sea represents the height of the former position of sea level during isolation. In some isolation basins, marine deposits can be found above freshwater sediments, which indicates that sea water reentered the basin when the rate of uplift fell behind the rate of sea-level rise. From each of the stratigraphic contacts between marine and freshwater sediments, it is possible to obtain a sea-level index point (Shennan et al., 1994).

Lloyd (2000) worked in an isolation basin at Loch nan Corr in northwest Scotland along a macrotidal coastline. While in many sea-level studies, a large tidal range poses problems, here it creates an opportunity to recognize various stages of transgressions and regressions (Figure 4). The evidence is obtained from analyses of Foraminifera and testate amoebae. The rate of uplift in this area is relatively slow and it took about 7000 years for the isolation process to be completed. Lloyd (2000) was able to obtain five sea-level index points to reconstruct the relative sea-level changes at Loch nan Corr.

Seismic Vertical Motion

Uplift and subsidence events produced by earthquakes occur suddenly and the changes in microfossils are abrupt and often dramatic. The changes are usually also visible in the lithostratigraphy, with soils or marsh deposits directly overlain by tidal flat deposits or vice versa. Foraminifera and diatoms have been used to quantify the amount of vertical displacement produced by earthquakes, especially along the Pacific coast of North America (e.g., Hemphill-Haley, 1995) and in New Zealand (e.g., Hayward et al., 2007). In northeastern Japan, Sawai

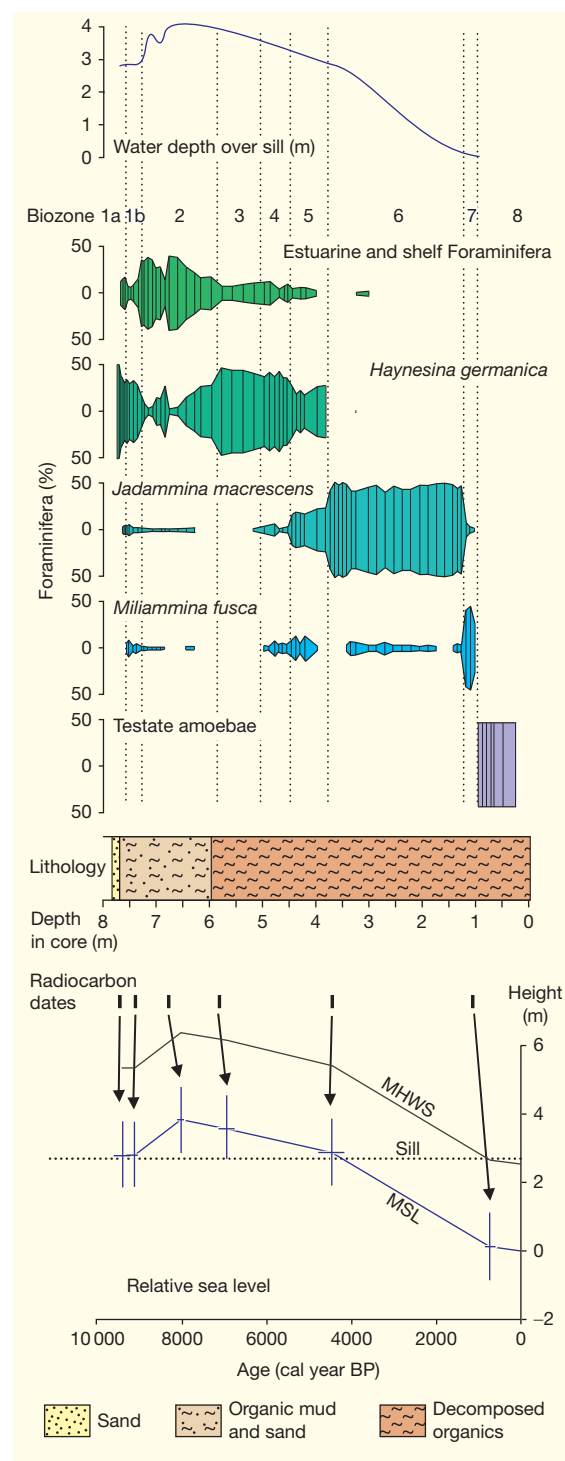


Figure 4 Changes in Foraminifera indicate the slow withdrawal of the sea from an isolation basin in northwest Scotland over the last 9000 years. The isolation process had reached completion when freshwater testate amoebae started living in the basin. From the microfossil changes, six radiocarbon dates and the height of the sill that separates the basin from the sea, a sea-level history has been reconstructed. Redrawn and modified from Lloyd J (2000) Combined foraminiferal and thecamoebian environmental reconstruction from an isolation basin in NW Scotland: Implications for sea-level studies. *Journal of Foraminiferal Research* 30: 294–305.

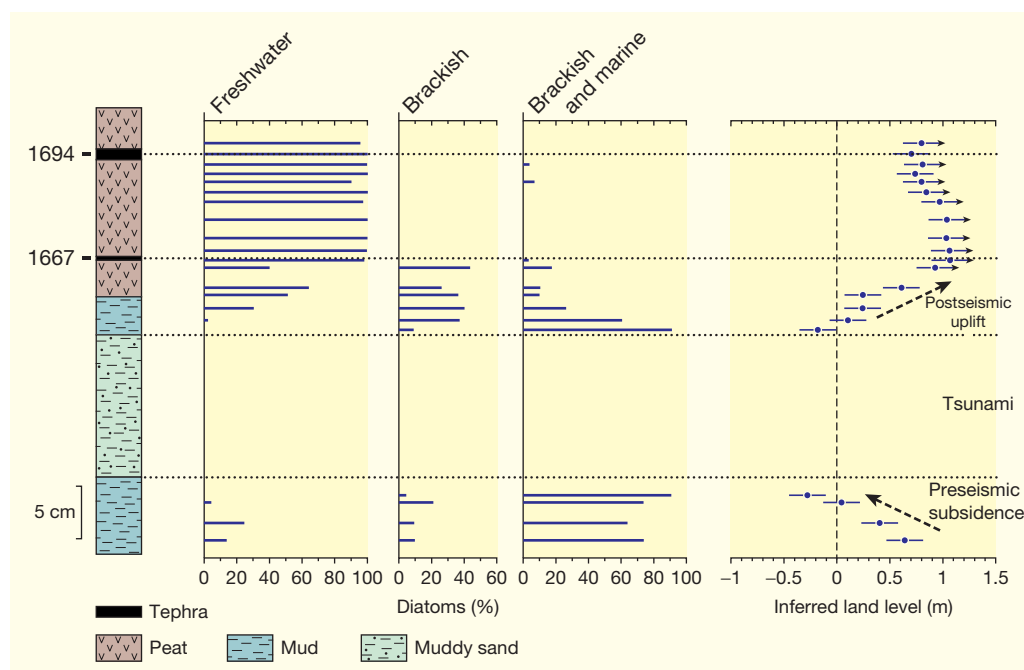


Figure 5 Diatoms in a coastal sequence in northeastern Japan have recorded preseismic subsidence and postseismic uplift of almost a meter. The seismic event produced a tsunami deposit. Redrawn and modified from Sawai Y, Satake K, Kamataki T, et al. (2004) Transient uplift after a 17th-century earthquake along the Kuril subduction zone. *Science* 306: 1918–1920.

et al. (2004) used diatom evidence to quantify preseismic and postseismic vertical land movements caused by large earthquakes (Figure 5).

Transfer Functions

Relative sea-level changes can be reconstructed both qualitatively and quantitatively. Both approaches involve the determination of the indicative meaning of sea-level indicators in the modern environment. With qualitative methods, the water-level relationship is determined in the modern environment by surveying. The observed modern ranges of indicators are then applied to the fossil counterparts by simply assuming that the surveyed range also applies to the fossil assemblage. With quantitative techniques, the heights of sea-level indicator microfossils are also surveyed, but their indicative meaning is established by statistical methods. The regression models used for this purpose can be linear or unimodal, depending on the distribution of the species along the height gradient, and they provide quantitative information on the ranges and optima of modern microfossil taxa with their errors. These values are then used to calculate the height of deposition of a given stratigraphic level based on the microfossil content of that horizon. The calculation of errors is an important feature of this technique and makes it arguably a more objective method than qualitative approaches.

The term ‘transfer function’ is used to describe the set of regression equations that is developed from modern relationships between modern environmental parameters and relative abundances of microorganisms. After its formulation, the

transfer function is applied to fossil sediments, usually in a core, to reconstruct past changes of the same environmental parameter(s) from the changes in microfossil assemblages. Transfer functions have been used in paleoecological studies since the early 1970s (Imbrie and Kipp, 1971), but in sea-level reconstruction, this quantitative approach was not applied until much later (Gehrels, 2000; Guilbault et al., 1995; Horton et al., 1999). These studies were based on Foraminifera, but transfer functions are also used in sea-level reconstructions based on diatoms (e.g., Sawai et al., 2004) and testate amoebae (e.g., Charman et al., 2010).

An important validation of transfer functions is possible when sea-level reconstructions derived from transfer functions can be directly compared with sea-level observations. Gehrels (2000) produced such a comparison for two salt marshes in Maine (Figure 6). One core is from a middle intertidal setting and includes high abundances of *M. fusca*, a middle-to-low marsh foraminifer which can also be found on tidal flats. The other core is from the high marsh and includes mostly typical high marsh Foraminifera, such as *J. macrescens*, *Balticamina pseudomacrescens*, and *Tiphrotrocha comprimata*. The resulting sea-level reconstructions based on a weighted-averaging transfer function reproduce the trends observed in the tide-gauge records well, but the variability in the reconstruction of middle marsh sediments is much greater than that for the high marsh core.

Recent studies have indicated that the application of transfer functions can lead to spurious results, especially when analogs for fossil assemblages are not present in the training set (Woodroffe, 2009). In these cases, it is preferable to combine the use of statistics with qualitative methods based on a visual assessment of the training sets (Long et al., 2010).

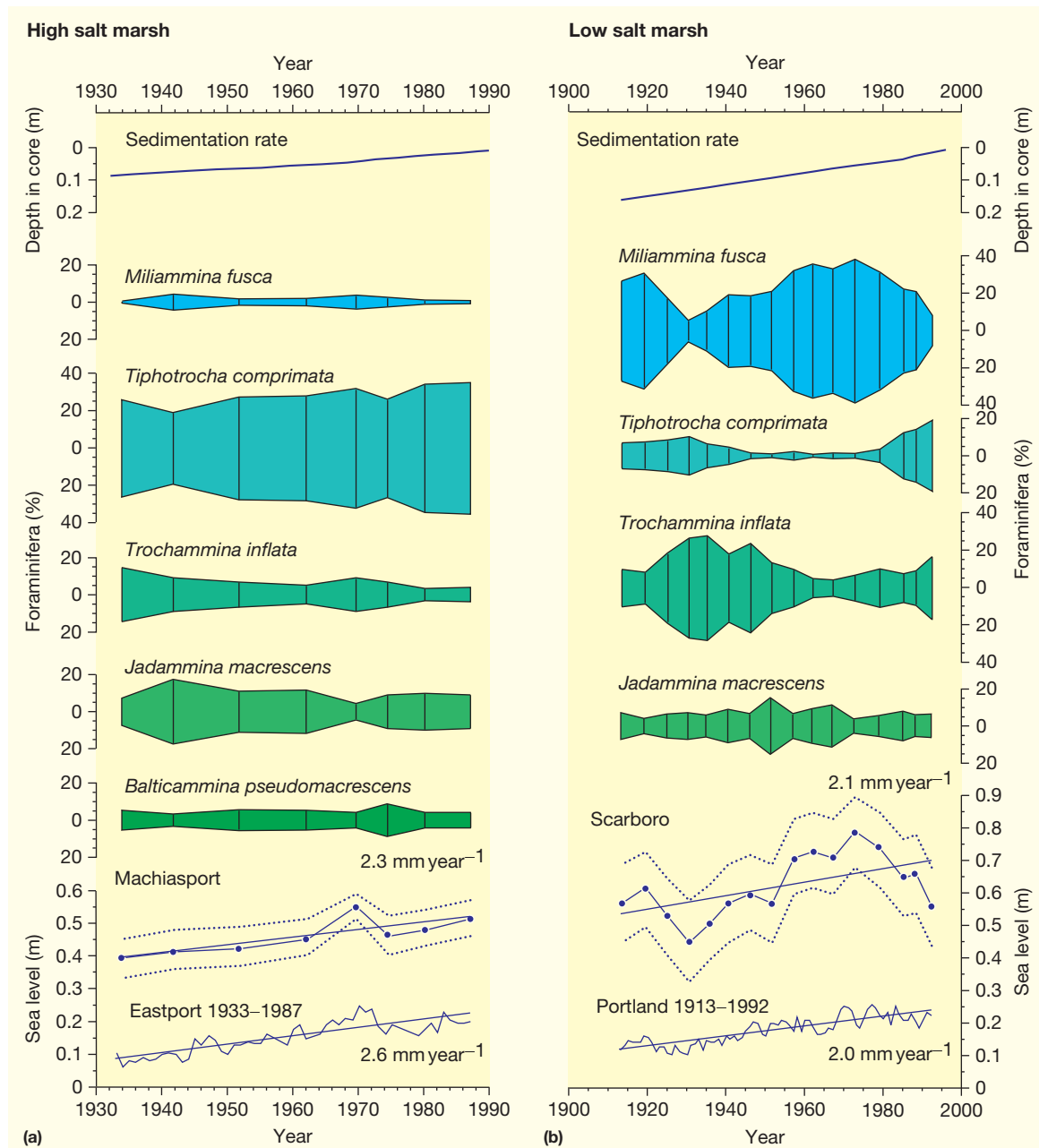


Figure 6 Two recent sea-level reconstructions from (a) high salt-marsh deposits and (b) low salt-marsh deposits from Maine (USA). The high salt-marsh reconstruction from Machiasport is more accurate and less 'noisy,' as is shown by a comparison with a nearby tide-gauge record. Redrawn and modified from [Gehrels WR \(2000\)](#) Using foraminiferal transfer functions to produce high-resolution sea-level records from salt-marsh deposits, Maine, USA. *The Holocene* 10: 367–376.

Chronologies

Radiocarbon dating is the traditional method to provide chronologies for microfossil-based Holocene sea-level reconstructions when sediments are older than ~300 years. Preferred materials for dating are plant fragments, horizontally embedded within the sediment sequence (e.g., [Gehrels et al., 2005](#)) or subsurface rhizomes and stalks with a well-defined relationship to the surface (e.g., van de Plassche, 2000). Macrofossils provide greater chronological precision than bulk dating of

organic sediments ([Törnqvist et al., 1998](#)). In rare cases, optically stimulated luminescence (OSL) dating has been applied, for example, in Denmark ([Szkornik et al., 2008](#)).

More recent sea-level reconstructions commonly rely on the ^{210}Pb method (e.g., [Gehrels, 2000](#)). A reliable ^{210}Pb chronology, however, is dependent on the choice of sedimentation model and ^{137}Cs measurements are therefore helpful to establish additional markers which can be used for calibration. In addition to ^{137}Cs , stable lead isotopic ratios can be used to validate the chosen ^{210}Pb model through detection of important changes in

the isotopic composition of atmospheric Pb produced by industrialization and car emissions (Gehrels et al., 2006b, 2008). More advanced radiocarbon methods, including high-precision and 'bomb-spike' dating, also offer potential for dating recent sediments (Marshall et al., 2007).

Conclusions

Microfossil groups that are used as indicators of sea-level changes in Holocene deposits include Foraminifera, diatoms, testate amoebae, and ostracods. Pollen are useful as indicators of coastal vegetation changes. Along isostatically subsiding coastlines, the most precise and accurate sea-level information is derived from microfossils that occur in the highest part of the intertidal zone. Microfossil analyses provide the indicative meaning for Holocene sea-level index points derived from transgressive and regressive stratigraphic contacts in buried coastal deposits. Detailed examinations of continuous biostratigraphic sequences may provide (sub)centennial scale fluctuations of relative sea-level change. This approach has been successfully used in salt marshes. The application of transfer functions has enabled the quantification of errors associated with sea-level reconstructions. Along isostatically uplifting coasts, microfossils preserved in isolation basins record changes in relative sea level. Abrupt relative sea-level changes produced by earthquakes can also be quantified by microfossils. Chronologies for microfossil-based sea-level reconstructions are based on a variety of methods, including ^{14}C , OSL, ^{210}Pb , ^{137}Cs , pollen, and geochemical markers.

See also: Diatom Introduction; Pollen Analysis, Principles. **Paleoceanography, Biological Proxies:** Benthic Foraminifera. **Sea Level Studies:** Eustatic Sea-Level Changes Since the Last Glacial Maximum; Geomorphological Indicators; Isostasy; Glaciation-Induced Sea-Level Change; Overview; Sedimentary Indicators of Relative Sea-Level Changes – Low Energy. **Sea-Levels, Late Quaternary:** Late Quaternary Relative Sea-Level Changes at Mid-Latitudes; Late Quaternary Relative Sea-Level Changes in High Latitudes; Tectonics and Relative Sea-Level Change.

References

- Allen JRL (1990) Constraints on measurement of sea-level movements from salt-marsh accretion rates. *Journal of the Geological Society* 147: 5–7.
- Charman DJ, Gehrels WR, Manning C, and Sharma C (2010) Reconstruction of recent sea-level change using testate amoebae. *Quaternary Research* 73: 208–291.
- Charman DJ, Roe HM, and Gehrels WR (1998) The use of testate amoebae in studies of sea-level change: A case study from the Taf estuary, South Wales, UK. *The Holocene* 8: 209–218.
- Charman DJ, Roe HM, and Gehrels WR (2002) Modern distribution of saltmarsh testate amoebae: Regional variability of zonation and response to environmental variables. *Journal of Quaternary Science* 17: 387–409.
- Denys L (1991/2) A check-list of the diatoms in the Holocene deposits of the western Belgian coastal plain with a survey of their apparent ecological requirements. *Belgian Geological Survey, Professional Paper* 246, 1–41.
- Devoy RJN (1979) Flandrian sea level changes and vegetational history of the lower Thames Estuary. *Philosophical Transactions of the Royal Society of London B* 285: 355–407.
- Engelhart SE, Horton BP, Roberts DH, Bryant CL, and Corbett DR (2007) Mangrove pollen of Indonesia and its suitability as a sea-level indicator. *Marine Geology* 242: 65–81.
- Frenzel P and Boomer I (2005) The use of ostracods from marginal marine, brackish waters as bioindicators of modern and Quaternary environmental change. *Palaeogeography, Palaeoclimatology, Palaeoecology* 225: 68–92.
- Gehrels WR (1994) Determining relative sea-level change from salt-marsh foraminifera and plant zones on the coast of Maine, U.S.A. *Journal of Coastal Research* 10: 990–1009.
- Gehrels WR (1999) Middle and late Holocene sea-level changes in eastern Maine reconstructed from foraminiferal saltmarsh stratigraphy and AMS ^{14}C dates on basal peat. *Quaternary Research* 52: 350–359.
- Gehrels WR (2000) Using foraminiferal transfer functions to produce high-resolution sea-level records from saltmarsh deposits, Maine, USA. *The Holocene* 10: 367–376.
- Gehrels WR (2002) Intertidal foraminifera as palaeoenvironmental indicators. In: Haslett SK (ed.) *Quaternary Environmental Micropalaeontology*, Chapter 5, pp. 91–114. London: Arnold Publishers.
- Gehrels WR, Hayward BW, Newnham RM, and Southall KE (2008) A 20th century sea-level acceleration in New Zealand. *Geophysical Research Letters* 35: L02717. <http://dx.doi.org/10.1029/2007GL032632>.
- Gehrels WR, Hendon D, and Charman DJ (2006a) Distribution of testate amoebae in salt marshes along the North American East Coast. *Journal of Foraminiferal Research* 36: 201–214.
- Gehrels WR, Marshall WA, Gehrels MJ, et al. (2006b) Rapid sea-level rise in the North Atlantic Ocean since the first half of the 19th century. *The Holocene* 16: 948–964.
- Gehrels WR, Kirby JR, Prokoph A, et al. (2005) Onset of recent rapid sea-level rise in the western Atlantic Ocean. *Quaternary Science Reviews* 24: 2083–2100.
- Gehrels WR, Roe HM, and Charman DJ (2001) Foraminifera, testate amoebae and diatoms as sea-level indicators in UK saltmarshes: A quantitative multiproxy approach. *Journal of Quaternary Science* 16: 201–220.
- Godwin H (1940) Studies in the post-glacial history of British vegetation. III. Fenland pollen diagrams. IV. Postglacial changes of relative land and sea level in the English Fenland. *Philosophical Transactions of the Royal Society of London B* 230: 239–303.
- Guilbault J-P, Clague JJ, and Lapointe M (1995) Amount of subsidence during a late Holocene earthquake – evidence from fossil tidal marsh foraminifera at Vancouver Island, west coast of Canada. *Palaeogeography, Palaeoclimatology, Palaeoecology* 118: 49–71.
- Hayward BW, Grenfell HR, Sabaa AT, Southall KE, and Gehrels WR (2007) Foraminiferal evidence of Holocene subsidence and fault displacements, coastal South Otago, New Zealand. *Journal of Foraminiferal Research* 37: 344–359.
- Hemphill-Haley E (1995) Diatom evidence for earthquake-induced subsidence and tsunami 300 years ago in southern coastal Washington. *Geological Society of America Bulletin* 107(3): 367–378.
- Horton BP, Edwards RJ, and Lloyd JM (1999) A foraminiferal-based transfer function: Implications for sea-level studies. *Journal of Foraminiferal Research* 29: 117–129.
- Horton BP and Shennan I (2009) Compaction of Holocene strata and the implications for relative sea-level change on the east coast of England. *Geology* 37: 1083–1086.
- Imbrie J and Kipp NG (1971) A new micropaleontological method for quantitative paleoclimatology: Application to a late Pleistocene Caribbean core. In: Turekian KK (ed.) *The Late Cenozoic Glacial Ages*, pp. 71–181. New Haven: Yale University Press.
- Kemp AC, Horton BP, Culver SJ, et al. (2009) Timing and magnitude of recent accelerated sea-level rise (North Carolina, United States). *Geology* 37: 1035–1038.
- Lloyd J (2000) Combined foraminiferal and thecamoebian environmental reconstruction from an isolation basin in NW Scotland: Implications for sea-level studies. *Journal of Foraminiferal Research* 30: 294–305.
- Long AJ, Scaife RG, and Edwards RJ (1999) Pine pollen in intertidal sediments from Poole Harbour, UK; implications for late Holocene sediment accretion rates and sea-level rise. *Quaternary International* 55: 3–16.
- Long AJ, Woodroffe SA, Milne GA, Bryant CL, and Wake LM (2010) Relative sea level change in west Greenland during the last millennium. *Quaternary Science Reviews* 29: 367–383.
- Marshall WA, Gehrels WR, Garnett MH, Freeman SPHT, Maden C, and Xu S (2007) The use of 'bomb spike' calibration and high-precision AMS ^{14}C analyses to date salt-marsh sediments deposited during the past three centuries. *Quaternary Research* 68: 325–337.
- Nelson AR and Kashima K (1993) Diatom zonation in southern Oregon tidal marshes relative to vascular plants, foraminifera and sea-level. *Journal of Coastal Research* 9: 673–697.
- Palmer AJM and Abbott WH (1986) Diatoms as indicators of sea-level change. In: van de Plassche O (ed.) *Sea-level Research: A Manual for the Collection and Evaluation of Data*, pp. 457–487. Norwich: GeoBooks.

- Roe HM, Charman DJ, and Gehrels WR (2002) Fossil testate amoebae in coastal deposits in the UK: Implications for studies of sea-level change. *Journal of Quaternary Science* 17: 411–429.
- Roe HM and van de Plassche O (2005) Modern pollen distribution in a Connecticut saltmarsh: Implications for studies of sea-level change. *Quaternary Science Reviews* 24: 2030–2049.
- Sawai Y, Satake K, Kamataki T, et al. (2004) Transient uplift after a 17th-century earthquake along the Kuril subduction zone. *Science* 306: 1918–1920.
- Scott DB and Medioli FS (1978) Vertical zonations of marsh foraminifera as accurate indicators of former sea-levels. *Nature* 272: 528–531.
- Scott DB, Suter JR, and Kisters E (1991) Marsh foraminifera and acellaceans of the lower Mississippi Delta: Controls on spatial distributions. *Micropalaeontology* 37: 373–392.
- Shennan I (1982) Interpretation of Flandrian sea-level data from the Fenland, England. *Proceedings of the Geologists' Association* 93: 53–63.
- Shennan I, Innes JB, Long AJ, and Zong Y (1994) Late Devensian and Holocene relative sea-level changes at Loch nan Eala, near Arisaig, northwest Scotland. *Journal of Quaternary Science* 9: 261–283.
- Szkornik K, Gehrels WR, and Murray AS (2008) Aeolian sand movement and relative sea-level rise in Ho Bugt, western Denmark, during the Little Ice Age. *The Holocene* 18: 951–965.
- Thomas E and Varekamp JC (1991) Paleo-environmental analyses of marsh sequences (Clinton, Connecticut): Evidence for punctuated rise in relative sealevel during the latest Holocene. *Journal of Coastal Research* 11: 125–158. Special Issue.
- Tooley MJ (1978) *Sea-Level Changes – North-West England during the Flandrian Stage*, p. 232. Oxford: Clarendon Press.
- Törnqvist TE, van Ree MHM, van 't Veer R, and van Geel B (1998) Improving methodology for high-resolution reconstruction of sea-level rise and neotectonics by paleoecological analysis and AMS ^{14}C dating of basal peats. *Quaternary Research* 49: 72–85.
- van de Plassche O (2000) North Atlantic climate-ocean variations and sea level in Long Island Sounds, Connecticut, since 500 cal yr AD. *Quaternary Research* 53: 89–97.
- Van Harten D (1986) Ostracode options in sea-level studies. In: van de Plassche O (ed.) *Sea-level Research: A Manual for the Collection and Evaluation of Data*, pp. 489–502. Norwich: GeoBooks.
- Vos PC and de Wolf H (1993) Diatoms as a tool for reconstructing sedimentary environments in coastal wetlands; methodological aspects. *Hydrobiologia* 269/270: 285–296.
- Woodroffe SA (2009) Recognising subtidal foraminiferal assemblages: Implications for quantitative sea-level reconstructions using a foraminifera-based transfer function. *Journal of Quaternary Science* 24: 215–223.
- Woodworth PL, White NJ, Jevrejeva S, Holgate SJ, Church JA, and Gehrels WR (2009) Evidence for the accelerations of sea level on multi-decade and century timescales. *International Journal of Climatology* 29: 777–789. <http://dx.doi.org/10.1002/joc.1771>.

Eustatic Sea-Level Changes – Glacial–Interglacial Cycles

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Introduction

Changes in relative sea-level result from the interplay of several processes operating at different rates and over contrasting spatial and temporal scales. During the Quaternary, the dominant mechanism responsible for sea-level changes has been the progressive buildup and decay of continental-scale ice sheets in response to Milankovitch forcing insolation changes. Relative changes in sea level, expressed at the global scale, were termed eustatic by [Suess \(1906\)](#), and evidence for such changes can be seen along many of the world's coastlines, apart from regions where former evidence has been largely destroyed by erosional processes (e.g., parts of northern Europe where late Pleistocene glaciation has eroded much of the pre-Holocene coastal stratigraphical record or regions that have experienced significant erosion at times of glacial low sea level). Fluctuations in global ice volume account for the majority of relative sea-level change in the Quaternary and accordingly are termed glacioeustatic in nature.

During the Last Glacial Maximum (LGM), excluding Antarctica, an additional 13% of the Earth's surface was covered in ice, corresponding with an ice volume of approximately $52.5 \times 10^6 \text{ km}^3$ ([Yokoyama et al., 2001](#)). Maximum ice sheet thickness exceeded 2500 m over extensive areas of Fennoscandia, particularly the Gulf of Bothnia in northern Europe, and over 3000 m for the Laurentide Ice Sheet centered over Hudson Bay on the North American continent ([Boulton et al., 1985](#); [Lowe and Walker, 1997](#)).

Studies of sedimentary successions on continental shelves and shallow marine platforms such as southeastern Australia (New South Wales), Bonaparte Gulf, northwestern Australia, and Barbados have indicated a maximum sea-level lowering of between <130 and 121 m during the LGM some 20–22 ka ago ([Bard et al., 1990](#); [Ferland et al., 1995](#); [Lambeck and Chappell, 2001](#); [Yokoyama et al., 2001](#)). These findings are consistent with independent observations such as modeling global ice volumes and oxygen isotope evidence that place ice-equivalent sea level between 135 and 120 m below present sea level during the LGM ([Clark and Mix, 2002](#)). These estimates for full glacial sea-level lowering are less than originally predicted on the basis of model calculations of ocean volume accommodation space and estimates of water locked up in continental ice, which had placed LGM ice-equivalent sea level at approximately –154 m ([Williams et al., 1998](#)).

A thorough understanding of the response of the ocean floor and continental shelves to differential water loads is central to accurately quantifying the magnitude of eustatic sea-level change from glacial maxima to interglacials and therefore accounting for the difference between observed and predicted values for eustatic (ice-equivalent) sea level. Larger water volumes will have a greater loading effect on the ocean floors and continental shelves and accordingly will reduce the magnitude of relative sea-level rise as observed along coastlines

during interglacials. In a similar manner, crustal isostatic loading will respond differently to ice or water (liquid state). For every 100 m of ice loading, the continental crust is isostatically depressed by 27 m and it is depressed up to 30 m in response to loading by water due to its higher density.

The geological record reveals major contrasts in the rate of changes in ice volume with time and, as a corollary, glacioeustatic sea-level changes. The last glacial cycle, for example, was characterized by an overall gradual increase in ice volume following the last interglacial maximum over a period of approximately 96 ka (i.e., from the end of the last interglacial maximum ca. 116 to the LGM at 20 ka: [Figure 1](#)). The broad pattern of sea-level changes accompanying the development of the continental ice sheets during the last glacial cycle conforms to a saw-tooth pattern of sequential decline. The record reveals a progressive increase in ice volume punctuated with periodic rises in sea level during interstadials. However, during deglaciation following the LGM, sea-level rise was rapid and occasionally occurred at rates approximately 20 times faster than the rate of sea-level lowering during ice sheet development.

Sea-level curves that depict major changes in the height of the sea surface with respect to the land (i.e., relative sea levels) since the start of the late Pleistocene (ca. 128 ka) are derived from restricted regions. Although such records are site-specific and involve assumptions of a 6 m above present sea-level (APSL) datum for the last interglacial maximum (128 ka) sea surface and a constant rate of uplift since that time, they do show broad similarities globally ([Aharon, 1984](#); [Steinen et al., 1973](#); [Figure 1](#)). Thus, the sea-level curves derived from the emergent coral terraces of Barbados and Huon Peninsula, Papua New Guinea, show a striking similarity in the long-term general trend of sea level, reflecting the dominantly glacioeustatic cause. Local variations in paleo sea level are due to glacio–hydro-isostatic feedback effects on continental shelves of contrasting widths and, in the near field of former ice sheets, progressive mantle adjustments in response to fluctuating ice volumes. These trends are further complicated by steric effects on sea level (e.g., changes in water volume accompanying changes in surface water temperatures).

The Nature of Glacial–Interglacial Cycles

The broad pattern of relative sea-level changes during glacial–interglacial cycles of the past 1 Ma has been inferred in part from evidence for relative sea-level changes of the last glacial cycle (approximately the past 128 ka). This has involved uniformitarian reasoning that similar patterns of eustatic sea-level changes are evident in the longer geological record and is predicated by the hypothesis that long-term insolation changes are primarily responsible for ice volume changes (Milankovitch hypothesis). The validity of this reasoning rests heavily on the spectral signatures of oxygen isotope ($\delta^{18}\text{O}$) data from deep-sea and ice

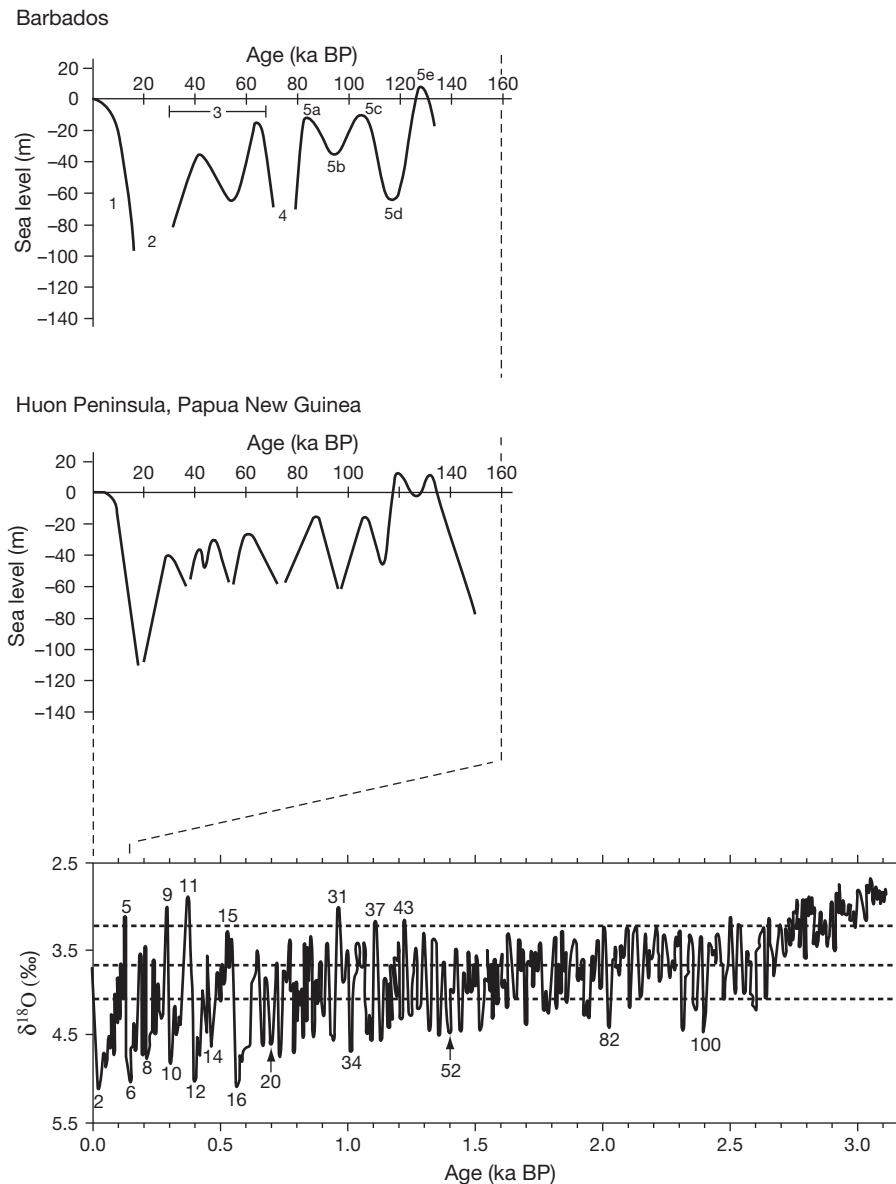


Figure 1 Glacioeustatic sea-level curves for the last glacial cycle (late Pleistocene and Holocene) from Barbados (Steinen et al., 1973) and Huon Peninsula, Papua New Guinea (Aharon, 1984) illustrating the general pattern of sequential decline of sea level with time until the attainment of lowest sea levels during the Last Glacial Maximum (20–22 ka ago). The marine oxygen isotope stages of the last glacial cycle are indicated on the Barbados sea-level record. The records of the last glacial cycle are correlated with the benthic oxygen isotope record from equatorial Atlantic core ODP-607 (Raymo, 1992). Core ODP-607 spans the past 3.2 Ma. Solid horizontal lines indicate the $\delta^{18}\text{O}$ values for the Holocene (top line), substage 5c (middle line), and the stage 2/1 boundary (bottom line) as determined at this core site. Oxygen Isotope Stages (OIS) are indicated for some of the glacials (even numbers) and interglacials (odd numbers). The $\delta^{18}\text{O}$ values for OIS 5 (actually substage 5e), 9, 11, 31, 37, and 43, as determined at this core site, imply sea levels higher than present for these interglacials. The 100 ka cycles appear to commence at about 1 Ma ago (the middle Pleistocene transition) and the previous 2 Ma are dominated by the 40 ka cycles characterized by lower amplitude sea-level variations. The progressive draw-down of sea level in response to the formation of the Antarctic Ice Sheet is evident in the basal 600 ka record in this core.

cores which for the past seven glacial cycles show a very generalized similarity in the timing and amplitude of isotopic changes, and as a corollary, eustatic sea-level changes (Masson-Delmotte et al., 2010; Raymo, 1992). Because a 0.1‰ change in $\delta^{18}\text{O}$ is equivalent to a 10-m change in eustatic sea level, the magnitude of eustatic sea-level fluctuations may be indirectly inferred from the isotopic record of deep-sea cores, with due allowance for local variations in isotopic signals in response to

water temperatures and salinity. The timing of these changes may be inferred from average rates of sedimentation in deep-sea cores, which are generally calibrated either by radiocarbon dating toward the top of cores for the past 50 ka, U-series for the past 400 ka, or by recognition of the Brunhes–Matuyama geomagnetic polarity reversal at 780 ka for longer records (i.e., the change from normal to reversed magnetic polarity, which generally represents the first magnetic reversal to be observed down

from the top of cores). The isotope records have adopted a numbering scheme such that odd numbers refer to warmer intervals, such as interglacials and interstadials, whereas even numbers refer to glacials or stadials (Figure 1).

The most detailed pattern of relative sea-level changes has been derived for the last glacial cycle, which commenced at the end of the last interglacial maximum (marine oxygen isotope substage 5e) approximately 116 ka ago. Whereas knowledge of paleo sea levels for earlier glacial cycles is more reliant on indirect inferences from isotopic records (particularly before interglacial stages 7 at 220 ka or 9 at 322 ka), sea levels of the last glacial cycle have been directly determined from geomorphological and stratigraphical evidence from many emergent coastlines, particularly those located near subduction complexes. Radiocarbon and U-series dating of emergent fossil coral reef successions indicating former sea levels, correlated with inferences derived from the marine oxygen isotope records of ice volume change (and as a corollary, eustatic sea-level changes), have provided the basic framework for understanding the pattern of eustatic sea-level changes for the past 140 ka. Initially, the evidence for interstadial sea-level highstands of Stage 5 (substages 5c and 5a) and Stage 3, as seen in Barbados and Papua New Guinea, were correlated with isotopically inferred sea levels for these intervals (Shackleton and Opdyke, 1973). Numerous subsequent studies have corroborated the correlations of isotopic events with discrete eustatic high stands from many localities throughout the world.

The last glacial cycle commenced with an abrupt fall in sea level at the end of the last interglacial maximum (OIS substage 5e) at approximately 116 ka. The last interglacial maximum occurred from approximately 128 to 116 ka ago. Although field evidence suggests a range of values, sea level was at least 2 m APSL during this period, and estimates from Indian and Pacific Ocean islands suggest values between 2 and 9 m APSL (Hearty et al., 2007). Isotopic records suggest that sea level fell by as much as 20 m within a few thousand years at the end of the last interglacial maximum (Shackleton, 1987).

OIS 5 has five subdivisions. Following the last interglacial maximum, eustatic sea levels fell during the stadial of substage 5d and rose during interstadial 5c to -10 m at approximately 105 ka. This was followed by a subsequent fall in sea level during the stadial of 5b and rose again during the subsequent interstadial 5a at 82 ka (-15 to -20 m). Interstadial 5a was followed by the stadial of Stage 4, which in turn was followed by the interstadial of Stage 3.

Eustatic sea levels of Stage 3 have been subject to several interpretations. Early reports of radiocarbon ages suggested that sea levels during interstadial Stage 3 were close to present approximately 30–40 ka ago. Subsequent studies, however, revealed that the radiocarbon ages underestimated the true age of the fossils and that the deposits correlated with the last interglacial maximum as supported by U-series dating and paleoclimatological reconstructions (Chappell, 1982; Thom, 1973). The radiocarbon ages represented minimum, apparent ages reflecting between 1% and 2% contamination by modern ^{14}C that could not be removed during sample pretreatment. More recent investigations have shown that sea levels during Stage 3 (70–30 ka) were 20–40 m lower than originally determined based on a reinterpretation of the Huon Peninsula emergent coral reef record (Chappell et al., 1996). Interstadial

Stage 3 sea levels for the interval 70–30 ka range between -90 and -45 m, with a modal class of about -65 m. The reinterpreted paleo sea levels for Stage 3 from the Huon Peninsula (Chappell et al., 1996) were previously anticipated based on inferences from marine isotope records based on studies of fossil foraminifer (Shackleton, 1987) and have been corroborated from other far-field environments such as the Lacedpede Shelf in southern Australia (Hill et al., 2009).

The LGM (Stage 2) followed Stage 3 and represents the period of lowest sea level during the last glacial cycle. Although isotopic records bracket the LGM between about 11 and 29 ka, sea level maintained its lowest level between 30 and 19 ka (Lambeck et al., 2002). At this time, continental areas were significantly larger, resulting from the lower sea levels. Australia had a land area up to 25% greater than present and land bridges extended between the mainland and Tasmania as well as Papua New Guinea. Similarly, the Sunda Shelf provided a land connection between Borneo, Java, Sumatra, Malaya, and Vietnam, and Beringia represented a major land bridge between the North American and Asian continents.

The inferred general pattern of eustatic sea-level changes for the past seven glacial cycles is based on the similarity of the spectral signature of oxygen isotope records of earlier glacial cycles with the last glacial cycle, as first noted by Emiliani (1955). In the most general sense, as seen in the benthic oxygen isotope record of equatorial Atlantic core ODP-607 (Raymo, 1992), the past six glacial cycles all show a broadly similar asymmetric pattern reflecting rapid deglaciation and sea-level rise heralding the onset of interglacials, and the slower formation of ice sheets over the course of each glacial cycle. The glacial cycles preceding the most recent cycle all show interglacials as the times of highest sea level. Core ODP-607 suggests that isotope Stages 5e, 9, and 11 were warmer than the current Holocene interglacial and, as a corollary, experienced higher sea levels. Based on a comparison of several deep-sea cores, Shackleton (1987) concluded that continental glaciation is likely to have been greater in Stage 6 than in Stage 2, and that Stages 12 and 16 experienced more extensive ice cover and lower sea levels, a finding corroborated by the northern Red Sea oxygen isotope record which revealed that during OIS 12 global ice volume was approximately 15% greater than during the LGM (Rohling et al., 1998). A similar pattern of glacial and interglacial intensities is evident in the EPICA Dome C ice core record from Antarctica (Masson-Delmotte et al., 2010).

Approximately 1.2 Ma ago, a notable change is evident in marine oxygen isotope records (the middle Pleistocene transition) from the 41-ka dominated cycles of the early Pleistocene to the 100-ka duration glacial cycles that have characterized middle and late Pleistocene time (Raymo et al., 2006; Shackleton et al., 1990). This trend is also evident in ODP-607, where the amplitude in isotopic signal ($\delta^{18}\text{O}$) in the higher frequency 41 ka cycles is less pronounced than in the 100-ka dominated cycles. The 3.2 Ma record of ODP-607 also shows an apparent overall increase in ice volume between 3.2 and 2.6 Ma that heralded the onset of the Quaternary. In the longer geological record, a series of plate tectonic reorganizations contributed to the establishment of the Antarctic Ice Sheet accounting for this draw-down in eustatic sea level and include the closure of the Isthmus of Panama (ca. 3 Ma), the opening of the Drake Passage between the Antarctic Peninsula

and the southern tip of South America (ca. 23 Ma), and the collision of Africa with Europe in the late Miocene.

High-Frequency Sea-Level Changes and Heinrich Events in the Last Glacial Cycle

Studies of ice cores have provided a greater opportunity to resolve short-term climate changes than previously possible with deep-sea sediments. Greenland ice cores reveal numerous abrupt changes in $\delta^{18}\text{O}$ during the last glacial cycle, termed Dansgaard–Oeschger cycles. Dansgaard–Oeschger cycles typically have a duration of approximately 1460 years, of which the last 600 years is represented by a distinct cold phase that ends abruptly with a pronounced change to warmer conditions. North Atlantic deep-sea sediments reveal the presence of discrete layers of lithic fragments (e.g., angular fragments of quartz, limestone, and dolomite) termed ice-rafted debris (IRD), which appears to precede the warming phases of Dansgaard–Oeschger events. Dansgaard–Oeschger cycles have been grouped into Bond Cycles in North Atlantic deep-sea sediments, and they end with well-defined IRD horizons termed Heinrich events. The IRD is associated with mobile sea ice in the North Atlantic Ocean. Oxygen isotope evidence from ice cores suggests that Dansgaard–Oeschger events may potentially be accompanied by sea-level changes of several tens of meters because their isotopic signature is about 50–75% that of the amplitude of glacial–interglacial changes in $\delta^{18}\text{O}$.

The Huon Peninsula, Papua New Guinea, reveals evidence for short-term sea-level changes associated with Heinrich events. The coral reef terraces reveal six discrete phases of sea-level rise and subsequent fall between 9 and 20 m lasting approximately 1–2 ka for the period 65–30 ka ago (Chappell, 2002). These results imply potential multiple sources for the meltwater returned to the world ocean associated with these oscillations, because the complete melting of the Greenland Ice Sheet would raise sea level by only approximately 5 m. Stratigraphic analyses of the geometry of the reef terraces at Huon Peninsula and their dating by U-series reveal that the rates of sea-level rise associated with these interstadials did not exceed the more rapid phases of deglacial sea-level rise following the LGM.

Evidence for Glacioeustatic Sea-Level Changes Prior to the LGM

Evidence for former sea-level highstands is apparent in regions that have been subjected to tectonic uplift, particularly near subduction zones (Chappell, 1974; Pirazzoli et al., 1993), and in regions characterized by a relatively high degree of tectonic stability (e.g., intraplate settings such as Australia) (Murray-Wallace, 2002). In general, the most detailed records of former eustatic sea level have been derived from flights of uplifted coral reef terraces, where there is a clear physical separation of highstand events due to high rates of tectonic uplift, as well as the opportunity to confidently define paleo sea level from well-dated coral reef facies. Uplifted settings also record interstadial events when sea level was significantly lower than present but higher than glacial lowstands (e.g., interstadial OIS 3 at –65 m). Uplifted regions also tend to provide a more

complete record of eustatic sea-level changes than tectonically more stable areas which tend to preserve evidence only for the highest points of glacial cycles (i.e., interglacial sea-level highstands).

In emergent reef settings, interglacial highstand records may extend back 1 Ma (Pirazzoli et al., 1993). In general, however, eustatic sea-level records typically extend back to only about 400 ka, a function of geological preservation rather than a coincidence with the practical limit of U-series dating. Along tectonically more stable coastlines, there is a greater likelihood that stratigraphical evidence for former sea levels has been destroyed by erosion, particularly by fluvial processes during sea-level lowstands. Accordingly, coastlines characterized by higher levels of tectonic stability tend to preserve evidence for the past two glacial cycles, particularly the penultimate (OIS 7 at 220 ka) and last interglacial (OIS substage 5e at 128 ka) records, as well as the current and ongoing Holocene interglacial.

Classical Field Localities for the Reconstruction of Paleo Sea Levels

Many areas of the world's coastlines have been examined for evidence of former eustatic sea levels. The Huon Peninsula in Papua New Guinea and Barbados represent examples of subduction settings in which the interaction of high rates of tectonic uplift (ranging from 3.5 to 0.27 m ka^{–1}, respectively) and relative sea-level changes have resulted in well-defined marine terraces. Coral reef terraces have formed on the Huon Peninsula and Barbados and are represented by *in situ* reef facies and associated lagoonal deposits. The marine terraces in these settings each generally record the culmination of a marine transgression, a highstand, and the onset of a fall in relative sea level at the end of a highstand (forced regression).

In contrast, the Coorong Coastal Plain in southern South Australia is represented by temperate carbonate coastal barriers deposited in response to high wave energy on a low-gradient coastal plain subject to slow rates of uplift (0.07 m ka^{–1}). The latter is an example of trailing edge, a passive continental margin, and is significant for its particularly long record of Quaternary sea-level changes.

Huon Peninsula, Papua New Guinea

Arguably the most detailed picture of eustatic sea-level changes for the later Quaternary has been derived from the Huon Peninsula, Papua New Guinea (Chappell, 1974; Chappell et al., 1996). The Huon Peninsula preserves evidence for numerous interglacial and interstadial sea-level highstands in the form of emergent fringing reef complexes (Figure 2). The peninsula is situated in a particularly complex geotectonic setting on the South Bismark Sea Plate, a microplate between the northward trending Indo-Australian Plate and the southwesterly trending Pacific Plate in a zone of one of the fastest rates of plate convergence in the world (ca. 120 km Ma^{–1}). Uplift of the Huon Peninsula is due to its proximity to the subduction zone at the New Britain Trench and a series of thrust complexes to the southwest of the peninsula.

The marine terraces extend up to 1000 m APSL. Uplift appears to be coseismic in nature, with the rates of long-term uplift being higher closer to the subduction zone. Uplift rates

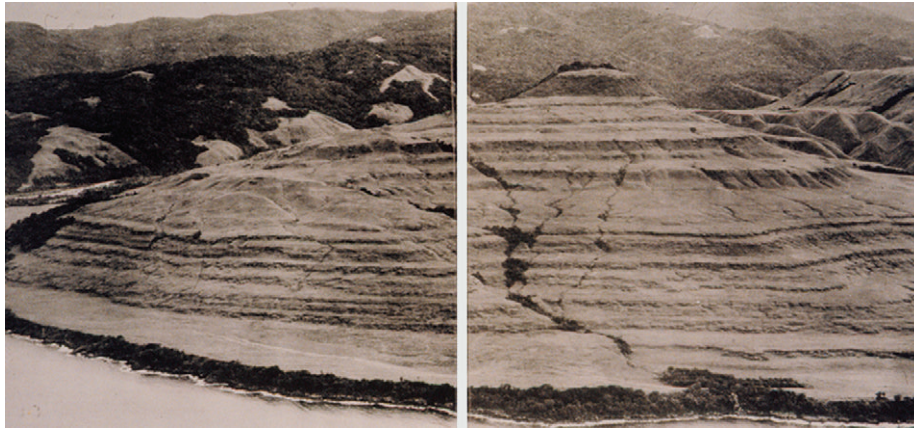


Figure 2 Oblique aerial photograph of the Huon Peninsula, Papua New Guinea showing the well-defined uplifted coral reef terraces. Each terrace has developed during a sea-level highstand during either an interglacial or interstadial. From Figure 15.1, Ollier CD, *Tectonics and Landforms*. Copyright © 1981 Longman, London.



Figure 3 A relict sea cliff, visor, and notch, Barbados. The relict sea cliff, locally termed the Second High Cliff, has a northerly trend on the western side of Barbados and represents the seaward margin of a former reef terrace of one of the island's oldest reefs. A last interglacial highstand (minimum age) gave rise to the visor and notch and modified the coastal cliff formed in middle Pleistocene coralline limestone. Photograph by Murray-Wallace.

averaged since the last interglacial (128 ka) range from 0.7 m ka^{-1} on the northwestern sector of the peninsula to 3.5 m ka^{-1} closest to the subduction zone. A detailed record of eustatic highstand events is preserved for the past 300 ka with the principal interstadials and interglacial maxima represented. The individual reef terraces record evidence for rising sea level and the ensuing highstand, the latter in the form of regressive lithofacies associations. The Huon Peninsula sea-level record also reveals evidence for short-term sea-level events associated with Heinrich events as previously discussed. Accordingly, the region represents one of the world's most detailed archives of eustatic change in the later Quaternary.

Barbados

Barbados is located in the Caribbean Sea approximately 150 km east of the Lesser Antilles. Covering an area of 430 km^2 , the island is a fore-arc ridge, representing the subaerial component of an accretionary prism resulting from the North American

Plate being subducted beneath the Caribbean Plate (Schellmann and Radtke, 2004). Mud diapirism since late Eocene time is responsible for ridge formation. Approximately 85% of the island is covered by Pleistocene coralline limestone that ranges between 15 and 130 m in thickness and extend up to 300 m APSL. The coralline limestone formed from a series of fringing reefs that have developed as the island progressively emerged during the Pleistocene ($0.23\text{--}0.38 \text{ m ka}^{-1}$). The northern part of the island has a glacioeustatic sea-level record extending beyond 600 ka, whereas the island's southern sector has an emergence record younger than 400 ka (Schellmann and Radtke, 2004). Two prominent, high coastal cliffs extend around the island, with the First High Cliff trending east–west on the southern part of the island and the Second High Cliff trending north–south on the northwestern part of the island (Figure 3). The prominent cliffs represent the boundaries between individual reef complexes and have been modified by subsequent sea-level changes and subaerial processes.

The Barbados relative sea-level record is particularly significant because its correlation with the Huon Peninsula record compellingly demonstrates that the changes in sea level are predominantly an expression of global processes (Steinen *et al.*, 1973). A detailed record of glacioeustatic sea-level highstands for the past four glacial cycles has been derived from the Barbados reef complexes based on U-series and electron spin resonance dating (Schellmann and Radtke, 2004). The record indicates, for example, that sub-Milankovitch sea-level oscillations were responsible for the formation of three morphologically distinct reef terraces in oxygen isotope substage 5e with uplift-corrected sea levels of approximately 0 m (i.e., present sea level) between 132 and 128 ka, –2 m at 128 ka, and –13 m between 120 and 118 ka (Schellmann and Radtke, 2004).

Coorong Coastal Plain, South Australia

The Coorong Coastal Plain in southern South Australia represents globally one of the longest Pleistocene highstand records for interglacial sea levels set on a passive, trailing-edge continental margin (Murray-Wallace *et al.*, 1998; Figure 4). Covering an area of some 24 000 km², the coastal plain is a broad, flat landscape with an average seaward gradient of 0.5%. A series of mixed calcareous–siliceous barrier shoreline complexes are preserved across the coastal plain and have formed in response to the high wave energy regime that characterizes this coastline. The barriers comprise eolianite (dune deposits) with interbedded beach and lacustrine facies as well as numerous paleosols (Figure 5). The relict coastal barriers, typically up to 30 m above the general surface of the coastal plain, are between 1 and 2 km in shore normal cross-section, up to 10 km apart, and occur subparallel to the modern coastline. Although composite features, having formed in more than one interglacial highstand, the barriers tend to increase in age landwards.

Preservation of these large-scale coastal landforms is the result of epeirogenic uplift and regional calcrete development. Emergence of the coastal plain has resulted from the combined effects of regional uplift of the Murray Basin in response to plate tectonic interactions between the Indo-Australian and Pacific Plates as part of regional uplift of the southern Australian margin (0.07 m ka^{–1}) and, at a more local scale, crustal doming around the locus of late Quaternary volcanism (0.13 m ka^{–1}; Murray-Wallace *et al.*, 1998; Sandiford, 2003).

The eolianites of the relict coastal barriers have been dated by luminescence and amino acid racemization (whole-rock), and a consistent picture has been derived of barrier formation during interglacial sea-level highstands (Huntley *et al.* 1994; Murray-Wallace *et al.*, 2001). An average rate of uplift of the central portion of the coastal plain based on the altitude of back-barrier lagoon facies, an indicator of former sea level, of 12 successive relict coastal barriers in a transect from Robe to Naracoorte is slow in global terms only, 0.07 mm year^{–1}. Accordingly, intertidal back-barrier lagoon facies of the last interglacial maximum in this region only occur up to 9 m APSL, an order of magnitude lower than for the last interglacial successions on the Huon Peninsula, Papua New Guinea, and imply a eustatic (ice-equivalent) sea level close to present for the last interglacial maximum in this region. The slow rate of tectonic uplift in this region means that mainly evidence for former interglacial highstands is preserved on the coastal plain. The coastal plain succession indicates a general comparability in

the height of successive interglacial highstands. Interglacial sea levels did not deviate by more than 6 m of present sea level for middle and late Pleistocene time, suggesting that the sea returned repeatedly across the continental shelf to a similar regional datum. The data also indicate that significantly higher sea level stands associated with ‘warmer interglacials’ (Burke, 1993) are not expressed in this record. An older barrier succession of presumed early Pleistocene age extends further inland in the Murray Basin, but its age has yet to be reliably determined.

Timing of Quaternary Glacioeustatic Sea-Level Changes

Dating of fossil corals using U-series disequilibrium (U/Th and Pa/Th) represents the foremost method for determining the timing of glacioeustatic sea-level highstands for the latest middle Pleistocene (400–128 ka) and late Pleistocene (128–11.5 ka) records. Radiocarbon dating has been the preeminent method for dating the deglacial sea-level rise following the LGM. Technological developments in the detection of the U-series by thermal ionization mass spectrometry (TIMS) have greatly reduced analytical uncertainties and accordingly refined the resolution with which discrete sea-level highstand events can be dated (Gallup *et al.*, 2002).

The application of laser ablation multicollector inductively coupled plasma mass spectrometry (MC-ICPMS) has significantly enhanced analytical sampling techniques, permitting numerous U-series measurements across individual mollusk and coral specimens (Eggins *et al.*, 2005). This is a significant advance because it permits a more rigorous assessment of the integrity of individual fossils and the degree to which fossils have been contaminated during diagenesis by nonindigenous uranium with the identification of contamination pathways across individual fossils. Collectively, these refinements have led to a clearer definition of the timing of specific sea-level highstands for the last glacial cycle.

Interglacial Sea Levels: Are all Interglacial Highstands of Comparable Magnitude?

One of the interesting and, to some extent, unresolved questions concerning Quaternary glacial cycles is whether sea level during each interglacial highstand attained a similar height. Early studies of former eustatic sea levels in the Mediterranean Basin assumed a progressive lowering of sea level during the Pleistocene, which is now attributed to a tectonic overprint on glacioeustatic sea-level changes.

In the longer geological record, the principal evidence for sea levels during interglacials is inferred from stable isotope records over long time series, particularly for middle and early Pleistocene time. In many respects, stable isotope records from deep-sea and ice cores provide the only means of inferring the magnitude of former high sea-level stands, as the marginal marine stratigraphical record for most of the world’s coastline provides only a fragmentary record of relative sea-level changes for much of early and middle Pleistocene time.

The marine oxygen isotope record shows that in the most general sense, interglacial sea levels were broadly comparable within a band of ±10 m. Stage 11 (423–362 ka) is likely to

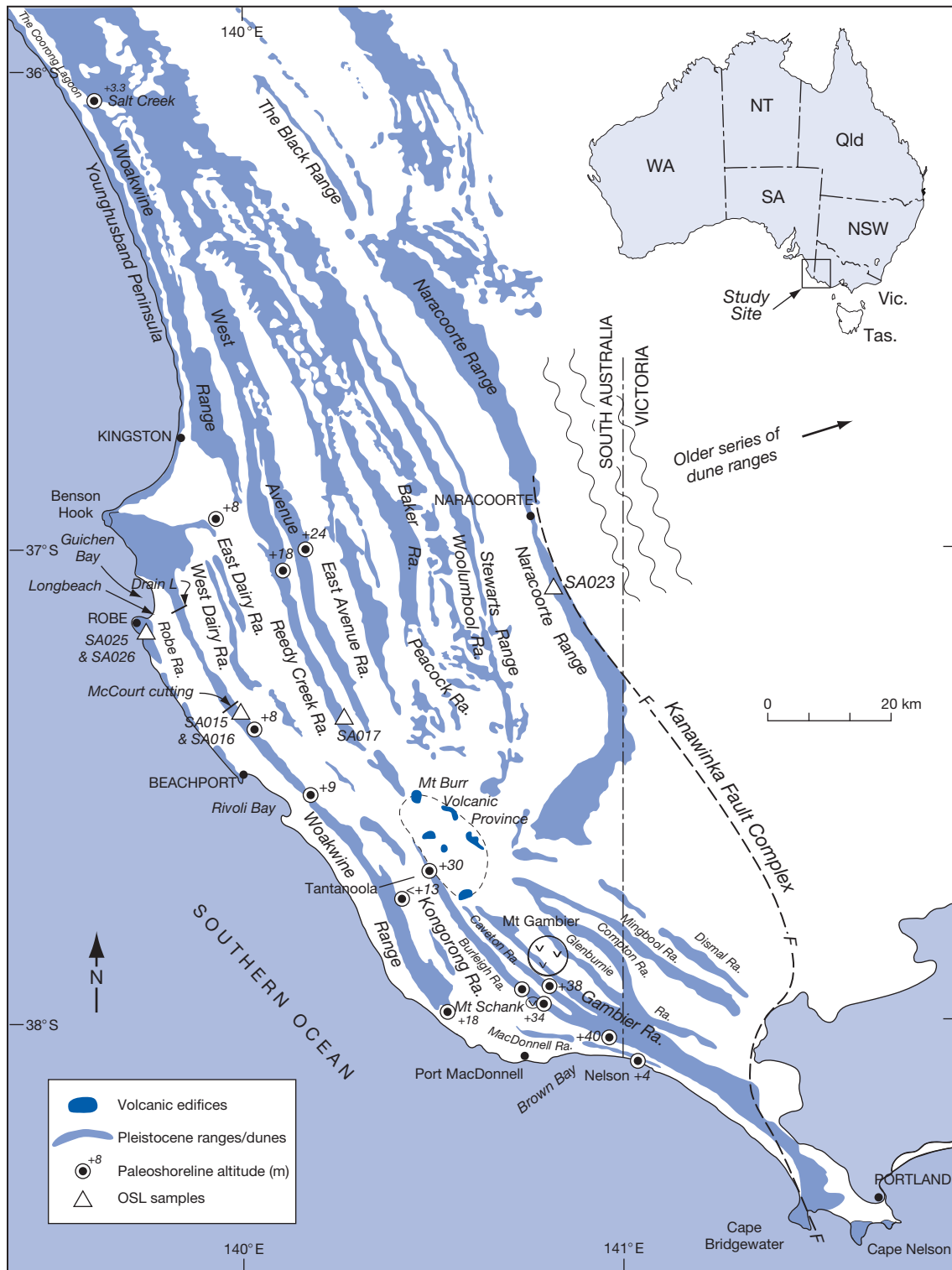


Figure 4 Map of the Coorong Coastal Plain in southern South Australia. The coastal plain preserves a record of interglacial highstand barriers (shaded) and back-barrier lagoon deposits. The coastal barrier deposits represented on this map have all formed since the beginning of the middle Pleistocene with the oldest barrier at Naracoorte. The barriers are progressively young toward the modern coastline.

have been the warmest interglacial in the past 500 ka and, as a corollary, characterized by significantly higher sea levels (Burkley, 1993). The suggestion that Stage 11 was characterized by a sea level 20 m APSL (Hearty et al., 1999) remains controversial

because few coastlines have been found to preserve anomalously high coastal facies of this age (Bowen, 2010) and not all oxygen isotope records show evidence for an inferred sea level higher than the last interglacial maximum for Stage 11.



Figure 5 Cross-section of an erosional remnant of the late Pleistocene (marine oxygen isotope substage 5c; ca. 105 ka) coastal barrier, Robe Range at Cape Northumberland, south of Mount Gambier, South Australia. The cliff face is 25 m high. Landward migrating aeolian tabular and trough cross bedding, as well as erosional truncations, are clearly evident. Paleosols and protosols are associated with the principal erosional truncations. Relative sea level was approximately 15 m below present at the time of deposition of the Aeolian dunes. Photograph by Murray-Wallace.

The Last Interglacial Maximum

Arguably one of the most enduring and significant geomorphological expressions of eustatic sea-level changes in the more recent geological record is the last interglacial maximum (OIS substage 5e; ca. 128–116 ka; [Figure 6](#)). For much of the world's coastlines, sediments and landforms of this age occur APSL and in many regions have been used as a regional datum to quantify geodetic changes that have occurred over the past 120 ka ([Bordoni and Valensise, 1998](#); [Murray-Wallace, 2002](#)). In Pacific and Indian oceans, fossil corals remote from plate boundaries consistently indicate former sea levels of between 2 and 9 m APSL ([Hearty et al., 2007](#); [Veeh, 1966](#)), and a more recent review of last interglacial coastal successions and geophysical modeling suggest a 95% probability that global (ice-equivalent) sea level peaked at 6.6 m APSL ([Kopp et al., 2009](#)). A 6-m reference level has been widely cited as a de facto eustatic sea level for the last interglacial maximum, from which corrections have been made for local tectonic deformation. In southern Australia, the most consistent elevation for the last interglacial shoreline is along the Eyre Peninsula (Gawler Craton) coastline where a value of 2 m APSL is recorded along approximately 500 km of coastline ([Murray-Wallace and Belperio, 1991](#)). This area has been geologically stable for the past 1.45 Ga, suggesting that it is particularly suited to quantifying eustatic sea level in the far field of Pleistocene ice sheets. Because the Holocene highstand (7 ka) along this coastline is up to 1 m APSL ([Belperio et al., 2002](#)), the elevation of the last interglacial shoreline is likely to reflect the combined effects of eustatic (ice-equivalent) sea level and hydro-isostasy, and suggests the eustatic sea level for the last interglacial was similar to the Holocene, a finding in accord with geophysical modeling ([Lambeck and Nakada, 1992](#)).

Uranium-series dating has been instrumental in defining the timing and duration of the last interglacial maximum (OIS 5e; [Edwards et al., 1987](#); [Eisenhauer et al., 1996](#); [Muhs 2002](#); [Stirling et al., 1995](#); [Veeh, 1966](#)). Early studies from tectonically



Figure 6 Holocene supratidal flats near the apex of Gulf St Vincent, South Australia that were formerly intertidal flats during the last interglacial maximum, attesting to the general coincidence of sea levels during the Holocene and last interglacial (OIS substage 5e) highstands. Photograph by Murray-Wallace.

uplifted settings such as the Huon Peninsula, Papua New Guinea, where there is a clear physical separation of reef terraces of last interglacial age, revealed evidence for two highstand events. An early phase at 138 ± 5 ka (Reef VIIa) and a late phase at 118 ± 2 ka (Reef VIIb) corresponded with uplift-corrected paleo sea-level culminations of 5 ± 5 m and 6.5 ± 4 m, respectively ([Aharon and Chappell, 1986](#)). High-precision uranium-series dating of corals from Bermuda, the Bahamas, Hawaii, and Australia suggest that sea level during the last interglacial (OIS 5e) was at least as high as present between 128 and 116 ka ([Muhs, 2002](#)).

Other approaches to quantifying the timing and nature of relative sea-level changes during the last interglacial maximum have involved high-resolution sampling of Foraminifera for isotopic analyses from marine cores. Oxygen isotope analyses on the planktonic foraminifer *Globigerinoides ruber* from two cores from the central Red Sea reveal the presence of at least two sea-level oscillations during MIS 5e with the sea-level culminations centered on 123.5 ka and 119.5 ([Rohling et al., 2008](#)). Oxygen isotope measurements on Foraminifera from rapidly deposited marine sediments provide the opportunity to infer glacioeustatic sea-level changes with a greater age resolving power than for coral reefs, given the slow growth response of the latter to rapid relative sea-level changes. The Red Sea record reveals a rate of relative sea-level rise of 1.6 ± 0.8 m per century at the onset of MIS 5e at approximately 123.5 ka when sea-level rose above present levels ([Rohling et al., 2008](#)). The global review of last interglacial coastal successions found that close to the onset of the last interglacial maximum, when sea level was ≤ -10 m, the rate of sea-level rise was at least 5.6 m ka^{-1} ([Kopp et al., 2009](#)). Collectively, these observations are of interest in the context of future projections of global sea-level rise.

Conclusion

The Quaternary has witnessed repeated major changes in eustatic sea level that accompanied the growth and decay of Pleistocene ice sheets and valley glaciers. The principal driving mechanism was insolation changes registered at the surface of the Earth. Globally, the timing and magnitude of these eustatic

changes have tended to conform to a consistent pattern, as indirectly indicated by the marine oxygen isotope record and corroborated by geomorphological and stratigraphical evidence from tectonically uplifted and more stable coastal areas. At a broad scale since at least the start of the middle Pleistocene, glacioeustatic sea-level variations have occurred as part of glacial cycles. These cycles have seen the slow and progressive buildup of continental ice sheets, and concomitant sea-level lowering by as much as 130–120 m over a time span close to 100 ka. High temporal resolution studies of ice cores for the last glacial cycle have indicated short-term (ca. 1460 years) but significant glacioeustatic sea-level changes (9–20 m) associated with major ice breakouts (Heinrich events) in the North Atlantic and illustrate the complexity of eustatic sea-level changes in the recent geological record. Seven glacial cycles have occurred since the start of the middle Pleistocene approximately 780 ka ago. Glacial cycles of the early Pleistocene involved lower amplitude sea-level fluctuations and occurred over shorter periods (41 ka cycles).

See also: U-Series Dating. **Glaciation, Causes:** Astronomical Theory of Paleoclimates. **Paleoclimate Reconstruction:** Sub-Milankovitch (DO/Heinrich) Events. **Radiocarbon Dating:** Conventional Method. **Sea Level Studies:** Coral Records of Relative Sea-Level Changes; Eustatic Sea-Level Changes Since the Last Glacial Maximum; Isostasy; Glaciation-Induced Sea-Level Change; Overview.

References

- Aharon P (1984) Implications of the coral-reef record from New Guinea concerning the astronomical theory of ice ages. In: Berger A, Imbrie J, Hays J, Kukla G, and Saltzman B (eds.) *Milankovitch and Climate: Understanding the Response to Astronomical Forcing*, pp. 379–389. Dordrecht: Reidel.
- Aharon P and Chappell J (1986) Oxygen isotopes, sea level changes and temperature history of a coral reef environment in New Guinea over the last 10^5 years. *Palaeogeography, Palaeoclimatology, Palaeoecology* 56: 337–379.
- Bard E, Hamelin B, and Fairbanks RG (1990) U-Th ages obtained by mass spectrometry in corals from Barbados: Sea level during the past 130,000 years. *Nature* 346: 456–458.
- Belperio AP, Harvey N, and Bourman RP (2002) Spatial and temporal variability in the Holocene sea-level record of the South Australian coastline. *Sedimentary Geology* 150: 153–169.
- Bordoni P and Valensise G (1998) Deformation of the 125 ka marine terrace in Italy: Tectonic implications. In: Stewart I and Vita-Finzi C (eds.) *Coastal Tectonics*, pp. 71–110. London: Geological Society. Special Publications 146.
- Boulton GS, Smith GD, Jones AS, and Newsome J (1985) Glacial geology and glaciology of the last mid-latitude ice sheets. *Journal of the Geological Society of London* 142: 447–474.
- Bowen DQ (2010) Sea level 400 000 years ago (MIS 11): Analogue for present and future sea level? *Climate of the Past* 6: 19–26.
- Burke LH (1993) Late Quaternary interglacial stages warmer than present. *Quaternary Science Reviews* 12: 825–831.
- Chappell J (1974) Geology of coral terraces, Huon Peninsula, New Guinea: A study of Quaternary tectonic movements and sea-level changes. *Geological Society of America Bulletin* 85: 553–570.
- Chappell J (1982) Radiocarbon dating uncertainties and their effects on studies of the past. In: Ambrose W and Duerden P (eds.) *Archaeometry: An Australasian Perspective*, pp. 322–335. Canberra: Australian National University Press.
- Chappell J (2002) Sea level changes forced ice breakouts in the Last Glacial cycle: New results from coral terraces. *Quaternary Science Reviews* 21: 1229–1240.
- Chappell J, Omura A, Esat T, et al. (1996) Reconciliation of late Quaternary sea levels derived from coral terraces at Huon Peninsula with deep sea oxygen isotope records. *Earth and Planetary Science Letters* 141: 227–236.
- Clark PU and Mix AC (2002) Ice sheets and sea level of the Last Glacial Maximum. *Quaternary Science Reviews* 21: 1–7.
- Edwards RLE, Chen JH, Ku T-L, and Wasserburg GJ (1987) Precise timing of the Last Interglacial period from mass spectrometric determination of Thorium-230 in corals. *Science* 236: 1547–1553.
- Eggins SM, Grün R, McCulloch MT, et al. (2005) In situ U-series dating by laser-ablation multi-collector ICPMS: New prospects for Quaternary geochronology. *Quaternary Science Reviews* 24: 2523–2538.
- Eisenhauer A, Zhu ZR, Collins LB, Wyrwoll K-H, and Eichstätter R (1996) The Last Interglacial sea level change: New evidence from the Abrolhos islands, West Australia. *Geologische Rundschau* 85: 606–614.
- Emiliani C (1955) Pleistocene temperatures. *Journal of Geology* 63: 538–578.
- Ferland MA, Roy PS, and Murray-Wallace CV (1995) Glacial lowstand deposits on the outer continental shelf of southeastern Australia. *Quaternary Research* 44: 294–299.
- Gallup CD, Cheng H, Taylor FW, and Edwards RL (2002) Direct determination of the timing of sea level change during Termination II. *Science* 295: 310–313.
- Hearty PJ, Hollin JT, Neuman AC, O'Leary MJ, and McCulloch M (2007) Global sea-level fluctuations during the Last Interglaciation (MIS 5e). *Quaternary Science Reviews* 26: 2090–2112.
- Hearty PJ, Kindler P, Cheng H, and Edwards RL (1999) A +20 m middle Pleistocene sea-level highstand (Bermuda and the Bahamas) due to partial collapse of Antarctic ice. *Geology* 27: 375–378.
- Hill PJ, De Deckker P, Von der Borch C, and Murray-Wallace CV (2009) Ancestral Murray River on the Lacepede Shelf, southern Australia: Late Quaternary migrations of a major river outlet and strandline development. *Australian Journal of Earth Sciences* 56: 135–157.
- Huntley DJ, Hutton JT, and Prescott JR (1994) Further thermoluminescence dates from the dune sequence in the southeast of South Australia. *Quaternary Science Reviews* 13: 201–207.
- Kopp RE, Simons FJ, Mitrovica JX, Maloof AC, and Oppenheimer M (2009) Probabilistic assessment of sea level during the last interglacial stage. *Nature* 462: 863–868.
- Lambeck K and Chappell J (2001) Sea level change through the last glacial cycle. *Science* 292: 679–686.
- Lambeck K and Nakada M (1992) Constraints on the age and duration of the last interglacial period and on sea-level variations. *Nature* 357: 125–128.
- Lambeck K, Yokoyama Y, and Purcell T (2002) Into and out of the Last Glacial Maximum: Sea-level change during Oxygen Isotope Stages 3 and 2. *Quaternary Science Reviews* 21: 343–360.
- Lowe JJ and Walker MJC (1997) *Reconstructing Quaternary Environments*, 2nd edn. London: Longman.
- Masson-Delmotte V, Stenni B, Pol K, et al. (2010) EPICA Dome C record of glacial and interglacial intensities. *Quaternary Science Reviews* 29: 113–128.
- Muhs DR (2002) Evidence for the timing and duration of the last interglacial period from high-precision uranium-series ages of corals on tectonically stable coastlines. *Quaternary Research* 58: 36–40.
- Murray-Wallace CV (2002) Pleistocene coastal stratigraphy, sea-level highstands and neotectonism of the southern Australian passive continental margin – A review. *Journal of Quaternary Science* 17: 469–489.
- Murray-Wallace CV and Belperio AP (1991) The last interglacial shoreline in Australia – A review. *Quaternary Science Reviews* 10: 441–461.
- Murray-Wallace CV, Belperio AP, and Cann JH (1998) Quaternary neotectonism and intra-plate volcanism: The Coorong to Mount Gambier Coastal Plain, Southeastern Australia: A review. In: Stewart IS and Vita-Finzi C (eds.) *Coastal Tectonics*, pp. 255–267. London: Geological Society. Special Publications 146.
- Murray-Wallace CV, Brooke BP, Cann JH, Belperio AP, and Bourman RP (2001) Whole-rock aminostratigraphy of the Coorong Coastal Plain, South Australia: Towards a 1 million year record of sea-level highstands. *Journal of the Geological Society of London* 158: 111–124.
- Pirazzoli PA, Radtke U, Hantoro WS, et al. (1993) A one million-year-long sequence of marine terraces on Sumba Island, Indonesia. *Marine Geology* 109: 221–236.
- Raymo ME (1992) Global climate change: A three million year perspective. In: Kukla GJ and Went E (eds.) *Start of a Glacial*, pp. 207–223. Berlin: Springer-Verlag.
- Raymo ME, Lisiecki LE, and Nisancioglu KH (2006) Plio-Pleistocene ice volume, Antarctic climate and the global $\delta^{18}\text{O}$ record. *Science* 313: 492–495.
- Rohling EJ, Fenton M, Jorissen FJ, Bertrand P, Ganssen G, and Caulet JP (1998) Magnitudes of sea-level lowstands of the past 500,000 years. *Nature* 394: 162–165.
- Rohling EJ, Grant K, Hemleben CH, et al. (2008) High rates of sea-level rise during the last interglacial period. *Nature Geoscience* 1: 38–42.
- Sandiford M (2003) Neotectonics of southeastern Australia: Linking the Quaternary faulting record with seismicity and in situ stress. In: Hillis RR and Müller RD (eds.) *Evolution and Dynamics of the Australian Plate*, pp. 107–119. Geological

- Society of Australia Special Publication 22, and Geological Society of America Special Paper 372.
- Schellmann G and Radtke U (2004) *The Marine Quaternary of Barbados*, p. 137. Kölner Geographische Arbeiten, Geographisches Institut Der Universität Zu Köln. Heft 81.
- Shackleton NJ (1987) Oxygen isotopes, ice volume and sea level. *Quaternary Science Reviews* 6: 183–190.
- Shackleton NJ, Berger A, and Peltier WR (1990) An alternative astronomical calibration of the lower Pleistocene timescale based on ODP Site 677. *Transactions of the Royal Society of Edinburgh: Earth Sciences* 81: 251–261.
- Shackleton NJ and Opdyke ND (1973) Oxygen isotope and paleomagnetic stratigraphy of equatorial Pacific core V28-238: Oxygen isotope temperatures and ice volumes on a 10^5 year and 10^6 year scale. *Quaternary Research* 3: 39–55.
- Steinen RP, Harrison RS, and Matthews RK (1973) Eustatic low stand of sea level between 125,000 and 105,000 BP: Evidence from the subsurface of Barbados, West Indies. *Geological Society of America Bulletin* 84: 63–70.
- Stirling CH, Esat TM, McCulloch MT, and Lambeck K (1995) High precision U-series dating of corals from Western Australia and implications for the timing and duration of the Last Interglacial. *Earth and Planetary Science Letters* 135: 115–130.
- Suess E (1906) *Das Antlitz der Erde*. Wien: Temstey, 3 volumes.
- Thom BG (1973) The dilemma of high interstadial sea levels during the last glaciation. *Progress in Physical Geography* 5: 170–246.
- Veeh HH (1966) $\text{Th}^{230}/\text{U}^{238}$ and $\text{U}^{234}/\text{U}^{238}$ ages of Pleistocene high sea level stand. *Journal of Geophysical Research* 71: 3379–3386.
- Williams M, Dunkerley D, De Deckker P, Kershaw P, and Chappell J (1998) *Quaternary Environments*, 2nd edn., p. 329. London: Arnold.
- Yokoyama Y, De Deckker P, Lambeck K, Johnston P, and Fifield LK (2001) Sea-level at the Last Glacial Maximum: Evidence from northwestern Australia to constrain ice volumes for oxygen isotope stage 2. *Palaeogeography, Palaeoclimatology, Palaeoecology* 165: 281–297.

Eustatic Sea-Level Changes Since the Last Glacial Maximum

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Introduction

Eustatic, or ice-equivalent, sea-level change is defined to be the uniform shift in the height of the ocean surface that takes place, in the absence of gravitational and rotational effects, ocean dynamics, and solid Earth deformation, due to the input of meltwater to the ocean (Farrell and Clark, 1976). In reality, noneustatic processes cause sea-level change to be spatially variable, and it is this spatially-variable signal that we sample when we infer past sea levels from field observations.

Eustatic sea-level (ESL) change, the global mean of this spatially varying signal, is an important quantity because it is directly equivalent to the changing mass of the global ice sheets, and it therefore provides an independent method for calibrating the marine oxygen isotope record and the total volume of continental ice at the Last Glacial Maximum (LGM). If we can also identify the source and rate of ice mass change over time, for example using sea-level records, and tie these to past variations in climatic conditions, this will improve our understanding of the response of ice sheets to climate change and provide a baseline for future predictions. Since we do not have complete global records of past sea levels, we use models to understand the processes that cause sea-level change to be spatially variable and interpret the records that have been preserved.

The ocean is not bounded by vertical cliffs, and therefore, changes in ocean mass are accompanied by changes in ocean area. The addition of meltwater increases the surface area of the ocean, while the removal of water, for storage in ice sheets, decreases the surface area of the ocean. This change in ocean area is taken into account in the calculation of ESL by integrating over time:

$$S_{\text{eust}}(t) = -\frac{\rho_{\text{ice}}}{\rho_{\text{water}}} \int_{t_0}^t \frac{\dot{V}_{\text{ice}}(t)}{A_{\text{ocean}}(t)} dt \quad [1]$$

$S_{\text{eust}}(t)$ is the eustatic shift in the height of the ocean surface between time t and some initial time (t_0), ρ_{ice} is the density of ice, and ρ_{water} is the density of ocean water. $\dot{V}_{\text{ice}}(t)$ is the rate of change of grounded ice volume at time t , and $A_{\text{ocean}}(t)$ is the area of the ocean at time t . Equation [1] is a statement of the conservation of mass; it makes the assumption that mass which is lost by the ice sheets is instantaneously gained by the oceans.

Since the end of the LGM at ~20 ka BP, sea-level change has been dominated by the flux of meltwater into the oceans following the deglaciation of the major ice sheets, and it has been postulated that sea-level change in tectonically stable regions far from the major ice sheets (referred to as far-field sites) approximates ESL (Milliman and Emery, 1968). Unfortunately, this is an oversimplification. Sea-level change takes place due to the deformation of either of the two bounding

surfaces: the ocean surface and the ocean floor (Figure 1). Ocean floor deformation is spatially variable and arises, in part, due to changes in ocean and ice loading during a glacial cycle. The temporal average of the ocean surface is a gravitationally equipotential surface, and perturbations to its height are also spatially variable, arising due to the redistribution of solid earth or surface masses, as well as changes in the total mass of ocean water (the eustatic component). Due to the temporally- and spatially-varying interplay between the ocean's bounding surfaces, sea-level change deviates from the eustatic value, even in far-field locations, and the eustatic signal must, therefore, be inferred by comparing observations of sea-level change to model predictions.

The Sea-Level Equation

Glacial isostatic adjustment (GIA) models are used to solve the sea-level equation, which considers sea-level changes due to variations in the mass of water in the ocean (the eustatic component) and sea-level changes due to spatially varying perturbations to the bounding surfaces of the ocean (the isostatic component), in order to predict sea-level change at any point on the Earth's surface:

$$SL(x, t) = S_{\text{eust}}(t) + SL_i(x, t) + SL_o(x, t) + SL_r(x, t) \quad [2]$$

$SL(x, t)$ is the sea level at time t in location x . $S_{\text{eust}}(t)$ is the spatially uniform shift in the height of the ocean surface due to eustatic changes in ocean volume, as defined in eqn [1]. The spatially varying isostatic component of sea-level change consists of perturbations to the ocean floor and ocean surface due to ice loading ($SL_i(x, t)$), ocean loading ($SL_o(x, t)$), and rotational feedback within the Earth system ($SL_r(x, t)$). Due to the viscous memory of the Earth's mantle following loading, these isostatic processes continue to cause sea-level change for thousands of years after the cessation of eustatic changes. A more detailed discussion of the processes that influence sea-level change during a glacial cycle is presented elsewhere in this encyclopedia.

GIA model predictions of sea-level change are dependent upon the ice sheet loading history and the rheology of the Earth. If the predictions do not match the observations, then the ice history and/or earth rheology are altered until a good fit is attained at all sites where sea-level observations are being considered. Once this is achieved, the eustatic signal may be inferred from the ice sheet history.

In Figure 2, we show model predictions of the departure from eustasy at 21 and 6 ka BP, as calculated by Milne and Mitrovica (2008). The predictions are based on the Bassett et al. (2005) deglaciation history and an earth model that is tuned to fit the far-field observations. Red areas in Figure 2(a) and 2(b) indicate regions where relative sea level at 21/6 ka BP

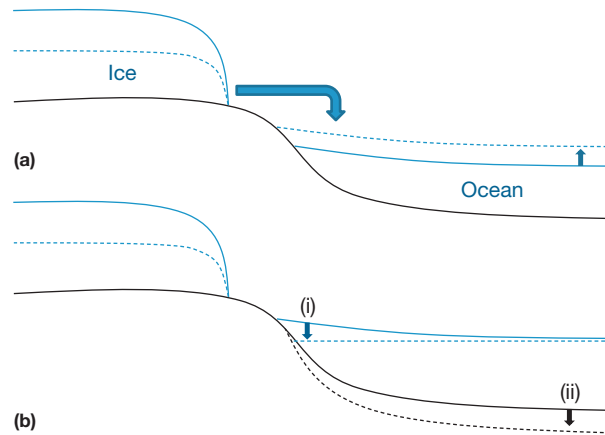


Figure 1 (a) Eustatic sea-level rise due to the influx of meltwater into the ocean. Initial ice and ocean surfaces are shown by solid lines; final ice and ocean surfaces are shown by a dashed line. (b) Isostatic sea-level changes due to (i) ocean surface fall in response to the decrease in the mass of the gravitationally-attracting ice sheet and (ii) ocean floor subsidence due to the increase in ocean loading. Initial surfaces are shown by solid lines; final surfaces shown by dashed lines.

lay above eustatic sea level, and blue areas indicate regions where relative sea level lay below eustatic sea level. The transition from yellow to blue identifies locations where sea level was identical to ESL at 21/6 ka BP. Note that the location of this transition changes between the two times, indicating that there is no location where sea level has tracked eustasy continuously since the LGM (Milne and Mitrovica, 2008). In order to determine ESL changes since the LGM, we seek sea-level records that closely approximate eustasy for each time interval of interest, and interpret each record after accounting for local isostatic effects. The location of such records for the Late Glacial and the Holocene periods are plotted in Figure 2(a) and 2(b), respectively.

A range of sea-level indicators are used when compiling these records. Details of each type are discussed briefly below, but the reader is referred to the relevant Encyclopedia entries for a complete discussion of these data. In all cases, it is important to determine the indicative meaning of any sea-level indicator, that is, the vertical relationship between the altitude of the sea-level indicator and the contemporary tide level (van de Plassche, 1986). A correction may also need to be applied to account for tectonics, subsidence, or compaction. Finally, we account for local isostatic effects before inferring the ESL history. This is carried out by comparing each sea-level record to model predictions, as outlined above.

GIA Modeling

To illustrate this procedure (or at least parts of it), in the following we compare observations of sea-level change to the results from two GIA models. The first, which will be referred to as Bradley et al., combines the results from two recent studies; one which focused on fitting a suite of far-field sea-level data from the LGM to 10 ka BP (Bassett et al., 2005) and another which has been calibrated to Holocene sea-level data from China and the Malay–Thailand Peninsula (Bradley et al., 2008). The Bradley et al. deglacial model is combined with an

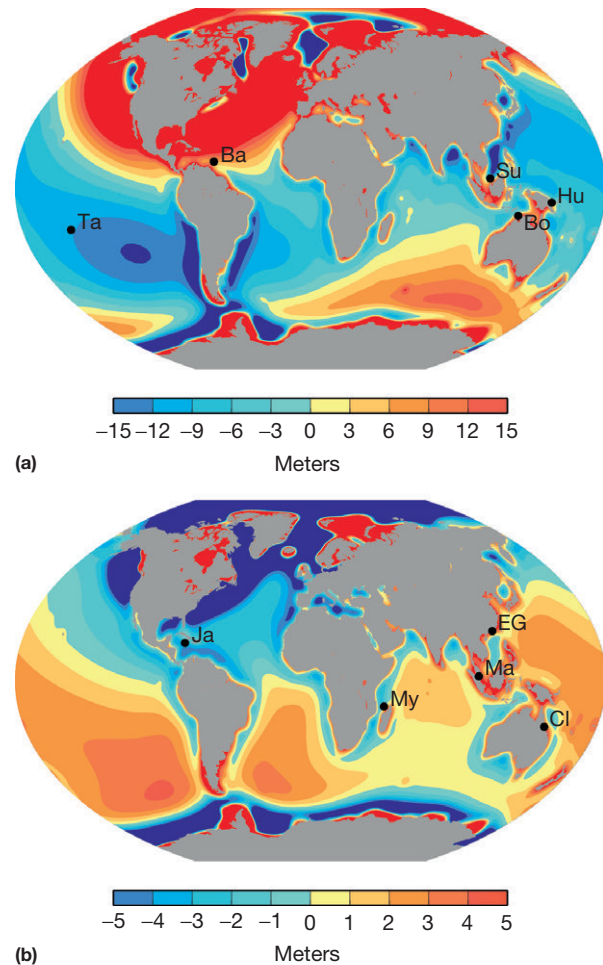


Figure 2 Predicted total sea-level change minus the model eustatic component for a time window extending from either 21 cal ka BP (a) or 6 cal ka BP (b) to the present day. The zero line marks where the total sea-level change is equal to the model eustatic value, with red and blue colors indicating sea levels that are, respectively, above and below the eustatic value. Note that the darkest red and blue colors indicate regions where the departure from eustatic sea level is greater than 15 m (a) and 5 m (b). Locations of sites referred to in the text are: Tahiti (Ta), Barbados (Ba), Sunda Shelf (Su), Bonaparte Gulf (Bo), Huon Peninsula (Hu), Jamaica (Ja), Mayotte (My), Malaysia (Ma), East Guangdong (EG), and Cleveland Bay (Cl). Reproduced from Milne GA and Mitrovica JX (2008) Searching for eustasy in deglacial sea-level histories. *Quaternary Science Reviews* 27: 2292–2302, with permission from Elsevier.

earth model whose parameters were chosen to represent average global earth properties; the model has a lithospheric thickness of 96 km and upper and lower mantle viscosities of 0.5×10^{21} and 1×10^{22} Pa s, respectively. These values are typical of those adopted in other GIA studies, and this will be referred to as the optimal earth model for the Bradley et al. GIA model. The second GIA model that we use is ICE-5G (Peltier, 2004). This is combined with a 90 km-thick lithosphere and the 'VM2' viscosity model (Peltier, 2004), which is characterized by a mean upper mantle viscosity of 0.5×10^{21} Pa s and a mean uppermost lower mantle viscosity of 1.6×10^{21} Pa s.

The eustatic component of these two GIA models is plotted in Figure 3. The magnitude and rate of ESL rise predicted by the

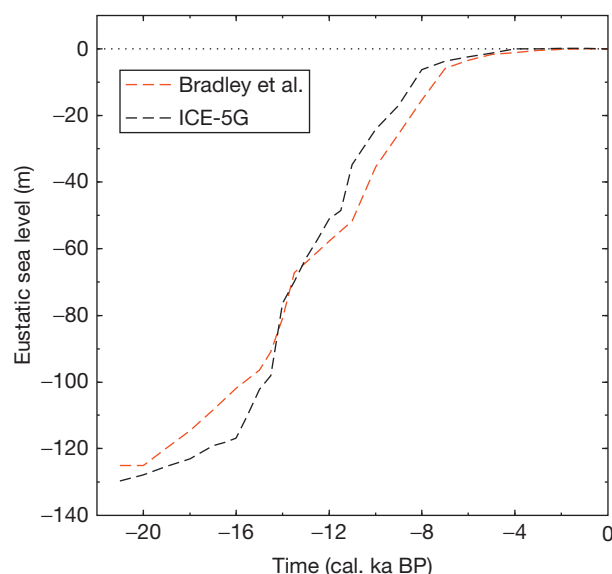


Figure 3 Eustatic sea-level changes derived from the Bradley et al. (red) and ICE-5G (black) GIA models.

two models during the last deglaciation is very similar, with total ESL change since the LGM predicted to be in the range 125–130 m, and rates of ESL rise predicted to peak at 20–30 mm year⁻¹ around 14.5 ka BP before dropping below 2.5 mm year⁻¹ from 8 to 7 ka BP onwards. The two GIA models have been tuned, using similar iterative processes, to fit independent datasets of sea-level observations and geodetic data. The differences in the predictions are indicative of the current level of uncertainty in the reconstruction of ESL change since the LGM.

We proceed by discussing, in detail, records of far-field sea-level change for the Late Glacial, Holocene, and late Holocene periods, and compare these to model predictions in order to infer ESL change since the LGM.

Sea-Level Changes During the Late Glacial

The Late Glacial covers the period between the LGM and the final deglaciation of the major continental ice sheets. During the LGM, global grounded ice volume reached its maximum, resulting in a mean lowering of sea level by $\sim 125 \pm 5$ m (Fleming et al., 1998). The LGM is estimated to have persisted from 26 ka BP until 21 ka BP (Clark et al., 2009; Peltier and Fairbanks, 2006; Yokoyama et al., 2000), and can be identified within sea-level records (see Figure 4(a), 4(b), and 4(e)) as a minimum lowstand.

Several outstanding issues remain unresolved with regard to ESL change during the Late Glacial. First, the timing and magnitude of the LGM lowstand can be used to constrain the total volume of ice within the major ice sheets, but not how that volume was distributed (Milne et al., 2002). Secondly, the inferred rise in ESL during the Late Glacial was not uniform, but included several periods of accelerated sea-level rise, commonly referred to as meltwater pulses (mwps). The source of meltwater for the largest of these, mwp-1A, is still the focus of

heated discussion (Gehrels, 2010), while the precise timing, magnitude, and even existence of some of the other mwps is still widely debated (Bard et al., 2010; Gehrels, 2010). In the two GIA models that we use, the Bradley et al. model assumes a dominant mwp-1A contribution from Antarctica, while in the ICE-5G model mwp-1A is primarily sourced from the Northern Hemisphere, with a dominant contribution from the North American ice sheets.

Data from the five far-field sites which have traditionally been used to infer ESL during the Late Glacial are shown in Figures 4(a)–4(e); the locations are shown as black dots on Figure 2(a). Sea-level observations from these sites provide a record of local sea-level change and are not a direct measure of ESL.

Barbados

The Barbados (13.2° N, 59.5° W) sea-level curve (Bard et al., 1990; Fairbanks, 1989; Peltier and Fairbanks, 2006; Figure 4(e)) is the longest and oldest record of sea-level change from the Late Glacial. Observations are derived from fossilized coral samples which have been dated using uranium–thorium (U–Th) and calibrated ¹⁴C to produce a record which extends from 24 to 7 ka BP. The island is located within the Lesser Antilles Island Arc and is undergoing long-term tectonic uplift due to the subduction of the South American plate beneath the Caribbean plate. To account for this, the data have been corrected assuming a steady uplift rate of 0.35 mm year⁻¹ (Peltier and Fairbanks, 2006). The estimated depth range of each coral type is then used to assign an indicative meaning, and hence paleo-sea-level, for each dated observation. This second correction accounts for the large vertical error bars on some of the data.

The extensive Barbados record provides information on the character of the LGM lowstand and post-LGM ESL rise that is not available from other Late Glacial records. The data suggest there was a sustained LGM lowstand, during which ESL remained stationary at an average depth of approximately –120 m (possible range –138 to –115 m) for ~ 5 ka prior to the onset of deglaciation. However, the large vertical errors on the coral samples (*Montastrea annularis*) do not rule out the possibility of an abrupt sea-level rise ~ 22 ka BP (see Peltier and Fairbanks (2006) for further discussion).

Following the LGM lowstand, the data show a continuous rise for 15 ka with sea level reaching 9.5 m below present by 7 ka BP. The rate of sea-level rise over this deglacial period is not steady, with two apparent periods of accelerated, rapid sea-level rise; a 20–24 m jump between 14.2 and 13.8 ka BP and a second, smaller, 15 m rise between 11.5 and 11.1 ka BP. These were the first observations to be defined as mwps (Fairbanks, 1989), respectively mwp-1A and mwp-1B, although it should be noted that they are constrained by gaps in the observational record associated with breaks in the core (Figure 4(e)).

Huon Peninsula

The Huon Peninsula (6° S, 147.5° E) is located in the Western Pacific and experiences the fastest uplift of the five sites selected due to the subduction of the Australian Plate below the Pacific Plate. The sea-level curve (Figure 4(c)) is derived from a

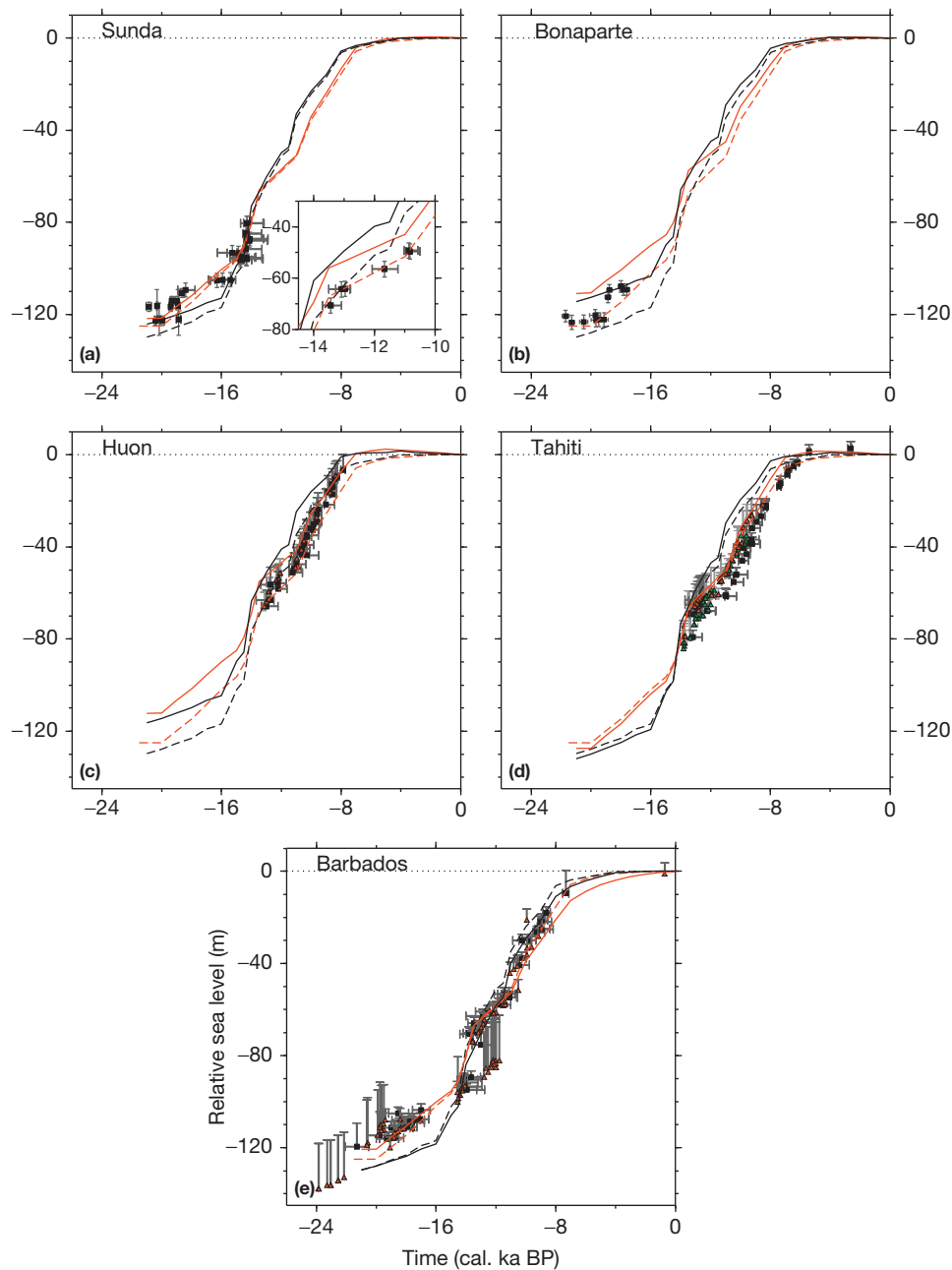


Figure 4 Observed and predicted sea levels at the five Late Glacial far-field sites ((a) to (e)), as shown in [Figure 2\(a\)](#), with [Figure 4](#) (a; inset) displaying data and predictions for the northerly Huon site situated ~ 800 km from the main data set. Black squares and red triangles are sea-level observations, dated using calibrated ^{14}C and U–Th, respectively. Green triangles are recently published, U–Th dated sea-level observations for Tahiti ([Bard et al., 2010](#)). Site-specific sea-level predictions are shown for two published GIA models. Solid black line: ICE-5G with the VM2 earth model ([Peltier, 2004](#)). Solid red line: Bradley et al. ice history with optimal earth model (see main text). The dashed black and red lines represent eustatic sea-level predictions for the ICE-5G and Bradley et al. GIA models, respectively (see [Figure 3](#)).

combination of calibrated ^{14}C ([Chappell and Polach, 1991](#)) and U–Th ([Cutler et al., 2003](#)) dated coral samples. Different tectonic rates have been used to correct the two data sets, $1.76 \text{ mm year}^{-1}$ for the ^{14}C data ([Chappell et al., 1996](#)) and $1.85\text{--}2.8 \text{ mm year}^{-1}$ for the U–Th data ([Cutler et al., 2003](#)). The data infer a steady rise in sea level from -66 m at 13.1 ka BP to -6.5 m at 7.8 ka BP , with no evidence for acceleration during mwp-1B. However, note that there is once again a gap

in the observational record during this period, between 12.1 and 11.2 ka BP .

Tahiti

Tahiti (18°S , 149.5°W) is a ‘hot spot’ volcanic island located within the South Pacific Ocean which undergoes long-term slow subsidence. To account for this, sea-level observations

have been corrected assuming an uplift rate of $-0.1 \text{ mm year}^{-1}$ (Bard et al., 1996) or $-0.25 \text{ mm year}^{-1}$ (Bard et al., 2010). The data are taken from two studies (Bard et al., 1996; Bard et al., 2010) which date a range of coral species from offshore cores using a combination of calibrated ^{14}C and U–Th. The oldest points in the Bard et al. (1996) data set (Figure 4(d)) suggest a rapid 10 m-rise in sea level between 13.8 and 13.4 ka BP, just after the proposed timing of mwp-1A. Although this jump is only constrained using a few data points, preliminary results from a more recent study support these findings (Deschamps et al., 2012). During the proposed time interval of mwp-1B, it is possible to infer a small ($<6 \text{ m}$) jump in the Bard et al. (1996) record; however, as the authors highlighted, when the uncertainty in the depth estimates of the coral samples is taken into consideration, it is equally possible to fit a linear trend to the sea-level rise over this period. Indeed, Bard et al. (2010) present a higher resolution data set and conclude that ‘the Tahiti data define a relatively smooth sea-level rise, with no significant accelerations during the time interval of mwp-1B.’ This highlights the importance of not over-interpreting a sparse data set.

Sunda Shelf

The Sunda Shelf (Hanebuth et al., 2000; Hanebuth et al., 2009) is a geologically stable continental shelf in southeast Asia. The sea-level observations are derived from calibrated ^{14}C -dated organic material (such as mangrove roots, peaty detritus) collected from two offshore drill sites, $\sim 800 \text{ km}$ apart (Hanebuth et al., 2000). The main dataset (shown on the main plot in Figure 4(a)) comes from the southern site (4.75° N , 109.75° E ; Hanebuth et al., 2000; Hanebuth et al., 2009), and a shorter record (shown in the inset on Figure 4(a)) comes from the northern site (9.39° N , 107.99° E ; Hanebuth et al., 2000).

At the southern site, following the LGM lowstand at $\sim 123 \text{ m}$ below present, the rate of sea-level rise is not steady, with two apparent periods of accelerated rise. The first is constrained by the data to be a 10 m jump at 19.6 ka BP. This is of a similar magnitude and timing to the jumps described for the Bonaparte Gulf (see below) and Irish Sea (Clark et al., 2004). However, the global nature of this ‘19 ka event’ is still under debate (Peltier and Fairbanks, 2006). The second jump of $\sim 16 \text{ m}$ between 14.6 and 14.3 ka BP has been associated with mwp-1A, although it is dated to have occurred 500–300 years earlier than the jump in the Barbados record. The shorter sea-level record at the northern site (inset on Figure 4(a)), which covers 13.4 to 10.8 ka BP, shows no evidence for a significant jump or acceleration in the rate of sea-level rise during the mwp-1B event.

Bonaparte Gulf

The Bonaparte Gulf (12° S , 128° E) lies on the broad, tectonically stable continental shelf which surrounds Australia. The sea-level observations (Figure 4(b)) are derived from calibrated ^{14}C dating of organic material (such as marine molluscs) taken from a range of intertidal zones (Yokoyama et al., 2000). Despite the short nature of this sea-level record, it constrains the LGM lowstand between 22 and 19 ka BP to be 122 m below present, and a sharp sea-level rise of $\sim 10 \text{ m}$ is observed at the onset of deglaciation, around 19 ka BP.

Summary

From these five far-field sites, we note several similarities in the pattern of sea-level change during the Late Glacial. First, the magnitude of the LGM lowstand, at $\sim 122 \text{ m}$ below present, is constrained by data at Barbados (Figure 4(e)), Sunda Shelf (Figure 4(a)) and Bonaparte Gulf (Figure 4(b)). At the latter two sites, a rapid (within 1 ka) rise of $\sim 10 \text{ m}$ at 19 ka BP is also observed. Second, there is evidence from Barbados, Tahiti, and the Sunda Shelf to support a brief period of rapid sea-level rise (mwp-1A) during the early deglaciation period, although the precise details are yet to be resolved. Despite evidence for a 15-m sea-level rise between 11.5 and 11.0 ka BP at Barbados (mwp-1B), there is little evidence for an event of this magnitude at other far-field sites. This has led to the suggestion that its magnitude has been overestimated, and that it was not a globally observed event. It has also been postulated that the 8.2 ka BP cooling event was driven by a meltwater pulse, sourced from the drainage of proglacial lakes in North America (Barber et al., 1999). This hypothesis has not yet been confirmed by the sea-level record, and we do not discuss it further here.

Comparisons to Modeling

As mentioned in the Introduction, sea-level observations provide a record of the spatial and temporal pattern of sea-level change at each site and are not a direct indication of ESL. One method to constrain ESL is to compare GIA model predictions of sea-level change with observations of sea-level change at each site. If a good fit is obtained at all sites, then the sea-level equivalent of the ice volume changes that drive the GIA model is taken to be a good approximation of ESL change.

Comparing model predictions of local sea-level change (solid black (ICE-5G) and solid red (Bradley et al.) lines in Figure 4) to the observed data, it is evident that local sea-level change is spatially variable and neither model produces a good fit at all sites. The pre-21 ka BP Barbados data were not available when the GIA models were developed, hence the models are not tuned to fit these data, and predictions are only plotted for 21 ka BP onwards. The magnitude of the local LGM lowstand is captured relatively well at Barbados (Figure 4(e)) and Sunda Shelf (Figure 4(a)) by the Bradley et al. model, but it is over-predicted at Barbados by the ICE-5G model, and both models under-predict the lowstand at Bonaparte Gulf (Figure 4(b)). This misfit could be resolved by adjusting the total ice volume in the input ice model, or, as will be briefly discussed below, by adjusting the choice of earth model. At Huon and Tahiti, the Bradley et al. model provides a closer match to the observed sea level during the Late Glacial than ICE-5G, which under-predicts the observed sea level. However, both models do a good job of predicting the rate of sea-level rise. After $\sim 15 \text{ ka BP}$, the sea-level predictions for Barbados are similar for the two models, and they both fit well, with the predictions bounding the observed data. Prior to 15 ka BP, the Bradley et al. model fits the data better.

The ESL curves derived from the two GIA models are also shown on Figures 4(a)–4(e) and 3 (dashed black and red lines for the ICE-5G and Bradley et al. models, respectively). We reiterate that we do not compare these curves to the sea-level

data, which are representative of local sea level as opposed to ESL, and we refer the reader to [Figure 2](#), where we plot the spatial distribution of the departure from eustasy, to put into context the difference between the local sea level and ESL curves. In particular, note that the observed magnitude of the LGM lowstand or the mwp-1A jump at each site may be higher or lower than the 'eustatic' magnitude. The driving mechanisms behind these differences are discussed in more detail in [Milne and Mitrovica \(2008\)](#).

There are several differences between the ESL histories derived from the two GIA models, including the depth (130 or 125 m) of the minimum ESL at the LGM and the dominant source of meltwater for mwp-1A (either an Antarctic or North American source). The ongoing debate regarding the magnitude (~26 m ESL in both models) and source of mwp-1A has been discussed widely elsewhere ([Gehrels, 2010](#)), and it is now generally believed that the pulse was smaller than initially determined from the Barbados record. The differences between the two ESL histories are primarily due to variations in the volume and timing of melting within the adopted input ice model and to a lesser extent the adopted earth model. At present, there is no optimal global GIA model; the limitations of each must be considered when choosing which to use.

The predictions in [Figure 4](#) have been generated using a single earth model for each GIA model, but, as discussed in greater detail in the Encyclopedia entry on isostasy, sea-level predictions are sensitive to the choice of earth model. To examine this, sea-level predictions at the five far-field sites are generated using the Bradley et al. model ([Figure 5\(a\)–5\(e\)](#)) using four additional earth models. These span a realistic range of upper and lower mantle viscosities, as constrained by previous GIA studies ([Kaufmann and Lambeck, 2002](#); [Mitrovica and Forte, 2004](#)). No single earth model gives the best fit to the data at all sites, and this is to be expected due to the heterogeneous nature of the Earth. The greatest sensitivity is found prior to 15 ka BP, where the local LGM lowstands are predicted to vary over 10–12 m. Some sites are more sensitive to variations in upper mantle viscosity (e.g., Huon and Tahiti) while others are more sensitive to lower mantle viscosity variations (e.g., Sunda Shelf).

Sea-Level Changes During the Holocene

The rate of ESL rise rapidly decreased in the Mid-Holocene following the final deglaciation of the northern hemisphere ice sheets. However, this rate did not immediately reach zero, largely due to the continued input of meltwater from Antarctica throughout the Mid–Late Holocene ([Nakada and Lambeck, 1988](#)), and variations in the volume of the Greenland ice sheet ([Fleming and Lambeck, 2004](#); [Simpson et al., 2009](#); [Tarasov and Peltier, 2002](#)). There exist a large number of observations relating to the local timing of sea-level stabilization or highstand throughout the Indian and Pacific oceans during the Mid-Holocene ([Woodroffe and Horton, 2005](#)), but few continuous records of sea-level change in a single location throughout this period. In [Figure 6](#), we plot data from five sites (shown as black dots in [Figure 2\(b\)](#)) where relatively continuous records

have been obtained, along with predictions of local and ESL change from the two GIA models.

The sea-level indicators used at these sites include shell fragments and beach-rock material from the intertidal zone (China, [Figure 6\(b\)](#)), mangrove and other organic material (Malaysia, [Figure 6\(a\)](#); Australia, [Figure 6\(c\)](#); and Jamaica, [Figure 6\(e\)](#)), fossil microatolls (Australia, [Figure 6\(c\)](#)), fossil oyster beds (Australia, [Figure 6\(c\)](#)), calcareous Foraminifera (Australia, [Figure 6\(c\)](#)), and coral material (Mayotte, [Figure 6\(d\)](#)). ^{14}C dating has been used to establish the age of all the sea-level indicators, with the exception of the coral material, which was dated using U–Th dating. Details of the calibration method used for each dataset can be found in the original data sources, along with details of the indicative meaning of each material and the associated depth errors. We now briefly describe the data and examine how well they are reproduced by the two GIA models.

Malaysia

In [Figure 6\(a\)](#), data from the Straits of Malacca (2.01° N, 102.64° E; [Geyh et al., 1979](#); [Hesp et al., 1998](#)) and Singapore (1.23° N, 103.92° E; [Bird et al., 2007](#)) are combined into one plot because the sites lie within 200 km of each other, and the spatial variation in sea-level predictions across the region is smaller than the errors on the data. No tectonic corrections have been applied to the data as vertical crustal motion in Malaysia is $<0.1 \text{ mm year}^{-1}$ ([Tjia, 1996](#)). A compaction correction is applied to the Singapore data ([Bird et al., 2007](#)), but may be a source of error in the Malacca data set. Both data sets record a sudden decrease in the rate of sea-level rise around 8 ka BP, with the Singapore data recording a hiatus in sea-level rise around 7.7 ka BP ([Bird et al., 2007](#)). The Malacca data then clearly record a ~4 m highstand at ~4.5 ka BP.

East Guangdong, Southeast China

The East Guangdong data (22.64° N, 114.98° E; [Zong, 2004](#)) also record a late Holocene highstand, but it is lower, peaking at ~2 m above present sea level, and occurs later, between ~3.5 and 2.5 ka BP ([Figure 6\(b\)](#)). The rate of early Holocene sea-level rise is much lower at this site than at the Malaysian sites, and the decrease to Mid-Holocene rates is much more gradual. No tectonic corrections have been applied to the data, although the region may be subject to a slight uplift due to the proximity of the plate margin ([Zong, 2004](#)).

Cleveland Bay, Australia

The tectonically stable Australian data ([Figure 6\(c\)](#); 19.89° S, 146.95° E; [Woodroffe, 2009](#)) do not give a clear picture of the rate of sea-level change during the early Holocene due to poor temporal constraints, but they do provide very tight constraints on sea-level changes at this site during the last 5 ka. A highstand of ~2 m is established prior to 6 ka BP, this is maintained beyond 4 ka BP, and then present-day sea level is reached around 2 ka BP.

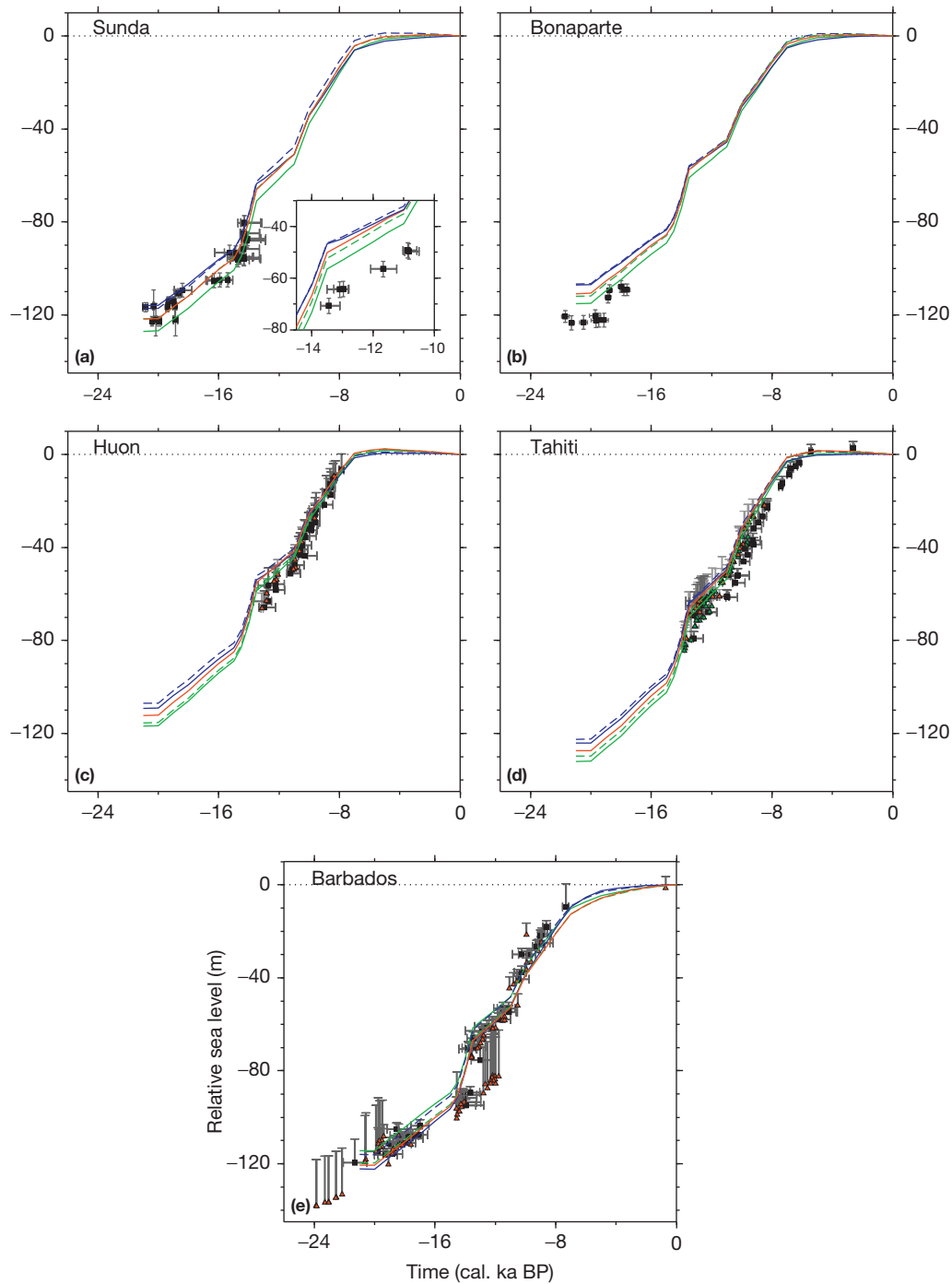


Figure 5 Comparison of sea-level predictions, for a range of earth models, to sea-level observations at the five Late Glacial far-field sites ((a) to (e)). Symbols for the data are as for [Figure 4](#). All predictions are generated using the Bradley et al. GIA model with a lithospheric thickness of 96 km. The solid red line represents the optimal earth model, as in [Figure 4](#). The dashed lines demonstrate the effect of varying the upper mantle viscosity from 1×10^{20} Pa s (blue) to 1×10^{21} Pa s (green) for a lower mantle viscosity of 1×10^{22} Pa s. The solid lines demonstrate the effect of varying the lower mantle viscosity from 1×10^{21} Pa s (blue) to 5×10^{22} Pa s (green) for an upper mantle viscosity of 5×10^{20} Pa s.

Mayotte

The sea-level data from Mayotte ([Figure 6\(d\)](#); 12.83°S, 45.29°E; [Camoïn et al., 2004](#); [Zinke et al., 2003](#)) are taken from a series of cores in the onshore coral reef and the Mayotte Lagoon. Data from the reef slope were rejected

to ensure that our sea-level curve is reconstructed purely from *in situ* material. We emphasize that a degree of caution should be exercised when using reef material to reconstruct past sea levels since different corals have different indicative meanings, and it is difficult to verify whether coral growth

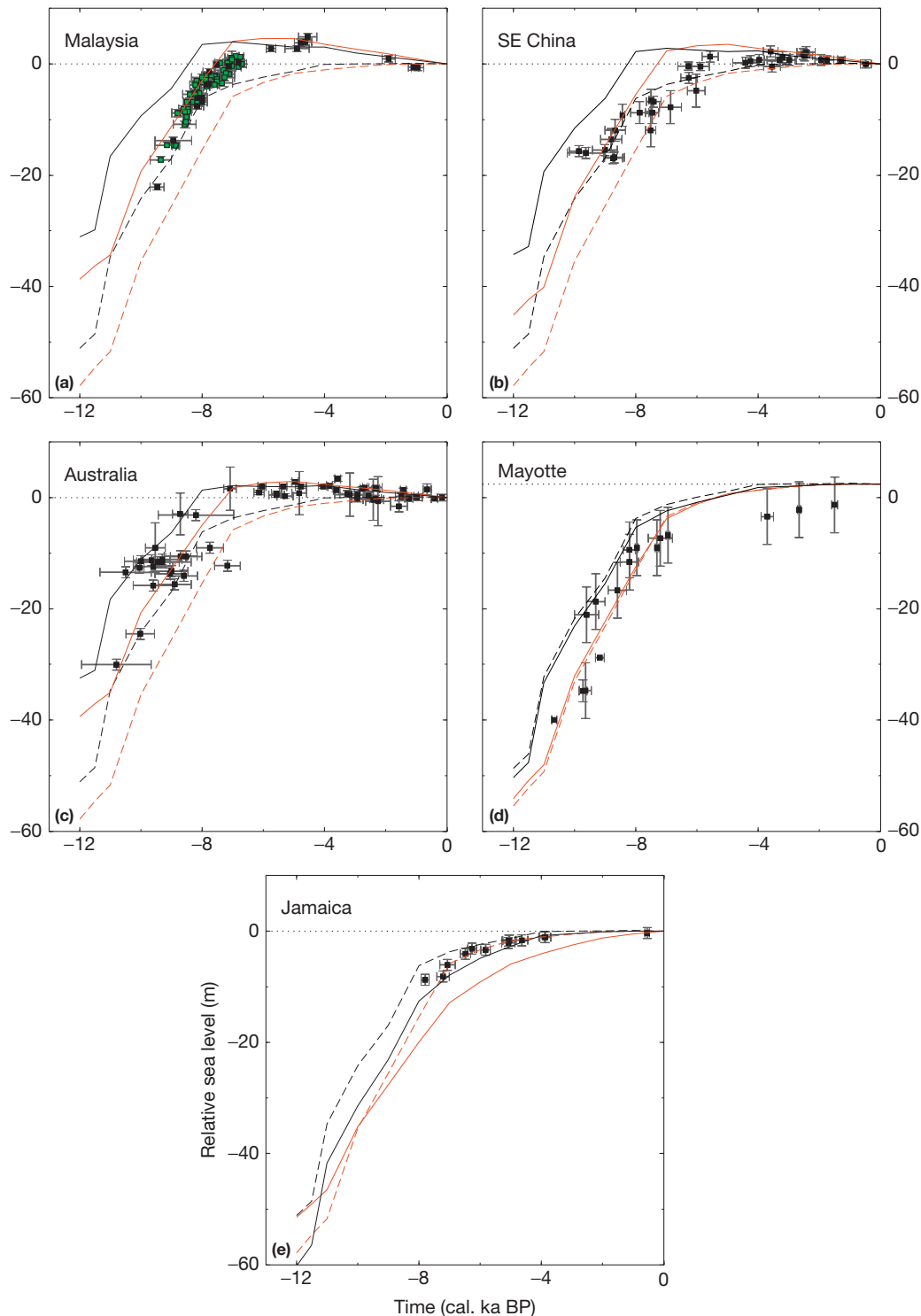


Figure 6 Observed and predicted sea levels at the five Holocene far-field sites ((a) to (e)), as shown in [Figure 2\(b\)](#). Black squares are sea-level observations as described in the main text. Green squares in [Figure 6\(a\)](#) are the data from [Bird et al. \(2007\)](#), black squares in [Figure 6\(a\)](#) are the data from [Geyh et al. \(1979\)](#) and [Hesp et al. \(1998\)](#). Site-specific and eustatic sea-level predictions are as for [Figure 4](#).

depths remained constant during all rates of sea-level rise. Despite the limited number of data points during the Mid-to-Late Holocene, the data do infer a decrease in the rate of sea-level rise around ~ 8 ka BP, as seen at the other sites, but they do not record a Holocene highstand.

Jamaica

The Jamaican data ([Figure 6\(e\)](#); 18.15° N, 78.11° W; [Digerfeldt and Hendry, 1987](#)) do not extend far enough back in time to quantify the transition from rapid sea-level rise

(prior to 8 ka BP) to slower Mid-Holocene sea-level rates. However, they do provide a very tight constraint on sea-level change from 8 ka BP onwards, and, similar to the Mayotte data, they clearly do not record a Holocene highstand, with sea level remaining close to present from 4 ka BP onwards.

Comparisons to Modeling

The Malaysian data are reproduced very well by the Bradley et al. model, which captures both the slow-down in the rate of sea-level rise and the magnitude and timing of the local highstand. The model does not reproduce the hiatus in sea-level rise around 7.7 ka BP that is hinted at by the high-resolution Singapore data. This inflection may be due to non-GIA vertical land motion, or variations in the melt rate of different ice sheets that are not currently replicated in the GIA model.

The Bradley et al. model also provides a reasonable reconstruction of the East Guangdong data, although the timing of the decrease in the rate of sea-level rise is predicted to take place ~ 1 ka earlier than implied by the data. At both this and the Malaysian sites the ICE-5G model predicts an earlier stabilization of sea-level rates than is observed, indicating that the timing of the cessation of melting in this model requires further tuning, assuming the observations are not biased by non-GIA processes.

It is difficult to determine which GIA model most accurately predicts sea-level change at the Australian site during the final stages of rapid sea-level rise due to the large degree of scatter in the data; however, the magnitude and timing of the local highstand is captured well by both models.

The lack of Holocene highstand at the Mayotte and Jamaican sites is replicated by both models, with the best fit to the Mayotte data provided by ICE-5G model and the best fit to the Jamaican data provided by the Bradley et al. model. Both GIA models predict that Mayotte closely tracks ESL change. Reasons for the variation in Holocene sea-level change at these five sites are discussed further below.

Holocene Sea-Level Highstands

ESL, or global ocean mass, is predicted to have increased monotonically during the Holocene, with rates tending to zero as deglaciation slowed. Observed sea levels at Malaysia, East Guangdong, and Cleveland Bay, Australia, exhibit considerable deviation from this pattern. In these three cases, sea levels rose above present 7–6 ka BP, and peaked 1–5 m above present sea level ~ 5 ka BP, before receding to their current level. This Mid-Holocene highstand is predicted by GIA models (Clark et al., 1978), and is related to the increase in ocean basin capacity during the late Holocene, as we discuss below.

During glaciation, ice loading caused mantle material to be forced laterally out from beneath the major ice sheets, resulting in uplift adjacent to the ice sheets. These ‘peripheral bulges’ are a glacio-isostatic feature, and typically form in oceanic regions. As the ice sheets melted, the bulges began to collapse; a process that continues to the present day due to the viscous properties of the mantle. The subsidence of a peripheral bulge creates additional space within the ocean basins resulting in a global fall in the ocean surface. This process is known as ‘ocean syphoning’ (Mitrovica and Peltier, 1991). Sea-level fall due

to peripheral bulge subsidence is augmented by the hydro-isostatic process of ‘continental levering’ (Walcott, 1972). Shallow continental margins were transgressed toward the end of the Late Glacial, and the associated ocean loading initiated a ‘tilting’ of these margins which results in ocean floor subsidence offshore. Late Holocene ocean-surface lowering due to the influx of water into these subsiding regions is approximately the same magnitude as that due to bulge subsidence (Mitrovica and Milne, 2002). Ocean syphoning has been the dominant contributor to sea-level change at locations far from the former ice sheets since the decline in rate of the eustatic signal. It takes place in the absence of a change in ocean mass; it is not a eustatic change.

A secondary consequence of continental levering is onshore uplift, which enhances the magnitude of a Holocene highstand. The magnitude of this enhancement will vary spatially due to coastal geometry: A site that is situated further inland, for example, at the head of an estuary, will experience greater uplift, and hence will record a higher highstand, than a site at the mouth of the estuary. By modeling the effect of ocean loading along complex coastlines, observations of the differential highstand between two sites enable bounds to be placed upon mantle viscosity (Kendall and Mitrovica, 2007).

The sea-level data from Mayotte and Jamaica do not record a Holocene highstand because the effect of ocean syphoning is balanced by sea-level rise in response to ocean loading and ice loading, respectively. In the first case, the volcanic island of Mayotte is tectonically stable, but its small size means that it moves with the ocean floor, which is currently subsiding due to ocean loading at a predicted rate of $0.13\text{--}0.25\text{ mm year}^{-1}$ (Camoin et al., 2004). The rate of sea-level rise due to subsidence balances the rate of sea-level fall due to ocean syphoning, with the result that this site almost perfectly tracks ESL during the Holocene (see Figures 2 and 6(d)). Jamaica is also subsiding, but due to a different process: It lies on a peripheral bulge which was created as a result of loading by the North American ice sheets. Now the ice has melted, the land is subsiding, and this counteracts the fall in sea level due to ocean syphoning.

Holocene ESL Changes

In general, the data in Figure 6 are fit within error by one of the GIA model predictions (solid lines), and therefore we may place some confidence in the ESL history extracted from the GIA models. Recall that we must rely on model reconstructions to determine past eustatic changes since all sea-level observations are contaminated by local and global (i.e., syphoning) isostatic effects.

Both deglacial models predict continued rapid ESL rise, at rates of $\sim 10\text{ mm year}^{-1}$, from the Late Glacial into the early Holocene. The ICE-5G model then predicts a decrease to $\sim 1.5\text{ mm year}^{-1}$ between 8 and 7 ka BP, while the Bradley et al. model adopts a gradual decrease 7–5 ka BP, concurrent with the final deglaciation of the Laurentide ice sheet. Between 7 and 4 ka BP, the ICE-5G model predicts $\sim 3.5\text{ m}$ ESL change due to ongoing melting of the Antarctic ice sheet, whereas the Bradley et al. model predicts a contribution of $\sim 5.5\text{ m}$ ESL from Antarctica since 7 ka BP. In the latter model, ice mass loss continues through to 2 ka BP, after which the eustatic

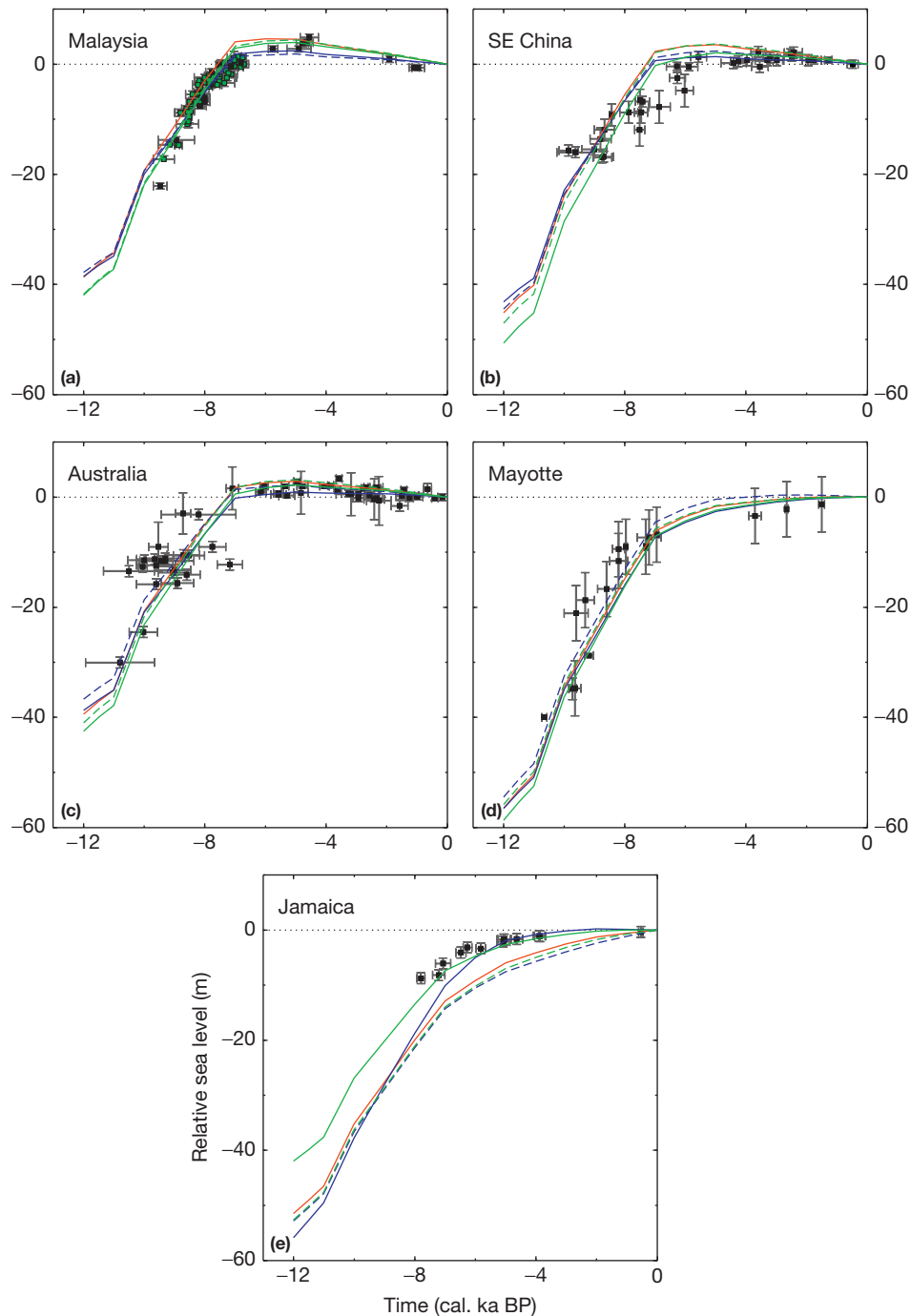


Figure 7 Comparison of sea-level predictions, for a range of earth models, to sea-level observations at the five Holocene far-field sites ((a) to (e)). Symbols for the data are as for [Figure 6](#), model predictions are as for [Figure 5](#).

contribution to sea-level change drops to zero. The Bradley et al. model is similar to [Lambeck's \(2002\)](#) eustatic reconstruction between 7 and 5 ka BP, and all three models are similar from 4 ka BP onwards. In the late Holocene, the ICE-5G model differs slightly from the other models in that it assumes that the mass of the Greenland ice sheet dropped below its present value by ~ 0.1 m ESL around 1 ka BP, before readvancing to its present extent at the beginning of the twentieth century.

Such high-resolution details of the recent ESL history must be resolved using near-field sea-level and ice-sheet data, as the majority of far-field data lack the required precision.

As in the case of the Late Glacial predictions (see [Figure 5](#)), the Holocene sea-level predictions are dependent upon the assumed earth structure as well as the deglaciation history ([Figure 7](#)). This factor is of lesser importance in the Holocene due to the smaller amplitude of mass redistribution over the

surface of the Earth during this period. However, the predictions still exhibit a significant degree of sensitivity in regions which experience the most deformation (e.g., Jamaica), and, as mentioned previously, highstand magnitudes are sensitive to mantle viscosity (e.g., East Guangdong, Malaysia).

Sea-Level Changes During the Late Holocene

For the last ~ 100 years, direct measurements of sea-level change have been recorded by tide gauges, supplemented in the last 15 years by satellite altimetry data, yielding a present-day eustatic rate of $\sim 2 \text{ mm year}^{-1}$ (Cazenave et al., 2009). An active area of sea-level research lies in trying to bridge the gap between the instrumental record and the paleo-record in order to place current rates of ESL rise into a longer-term context. This requires sea-level indicators that (i) cover the last few millennia, (ii) can accurately resolve decimeter-scale changes on 100-year time scales, and (iii) have a chronology that extends into the early twentieth century. Three key data sets have currently been utilized: microatolls, salt marshes, and archaeological sites (see the relevant Encyclopedia entries).

Microatolls are highly sensitive to small sea-level changes at low latitudes, and have been used to reconstruct late Holocene sea-level change in the Southern Cook Islands (Goodwin and Harvey, 2008). At this southern Pacific location, ocean syphoning caused sea-level to fall at $0.4\text{--}0.8 \text{ mm year}^{-1}$ throughout the majority of the last millennium until modern sea-level rise reversed this trend.

Salt marshes provide an excellent record of millennial-scale sea-level change (Long et al., 2010), but they are mostly found outside the tropics, where ESL changes are likely to be overprinted by isostatic effects. However, Gehrels et al. (2006) have successfully used salt marsh records to reconstruct slow rates of ESL fall during the last millennium and demonstrate that contemporary rates of ESL rise are $1\text{--}2 \text{ mm year}^{-1}$ greater than late Holocene rates.

Archaeological data also indicate slow rates ($<0.2 \text{ mm year}^{-1}$) of ESL fall during the last millennium, and can be used to place bounds on the timing of the onset of modern sea-level rise (Lambeck et al., 2004).

However, a degree of caution must be applied to the interpretation of high-resolution observations of sea-level change during the last few hundred years. Aside from GIA, spatially-variable sea-level change takes place due to variations in the temperature and salinity profile of the oceans and perturbations to atmospheric and oceanic circulation patterns, on multiannual to multidecadal time scales. The magnitude of sea-level change induced by these processes can be several times current eustatic rates, as demonstrated by the present-day range of sea surface displacement rates recorded by satellite altimetry (Figure 8). Several decades of observations are therefore required to isolate the GIA signal from the 'dynamic' signal (Kopp et al., 2010). Since low rates of eustatic change have persisted for the majority of the last 2–3 ka, care must also be taken to verify that observations from this period can be attributed to GIA, and not other causes of local sea-level change.

Summary

ESL change is the change in the height of the ocean surface, in the absence of gravitational and rotational effects and solid earth deformation, due to an influx of ocean mass in the form of meltwater. In reality, sea-level change varies continuously in space and time due to the aforementioned gravitational and rotational effects and solid earth deformation, and consequently there is nowhere where the eustatic change can be directly measured. ESL change must be inferred, using GIA models to interpret records of sea-level change. We present a comparison of ten far-field sea-level records with predictions derived from two GIA models. While these records closely approximate ESL changes, they contain additional isostatic and tectonic components, and must be interpreted with this in mind. The close fit between observations and predictions verifies the accuracy of the GIA models, with any misfit

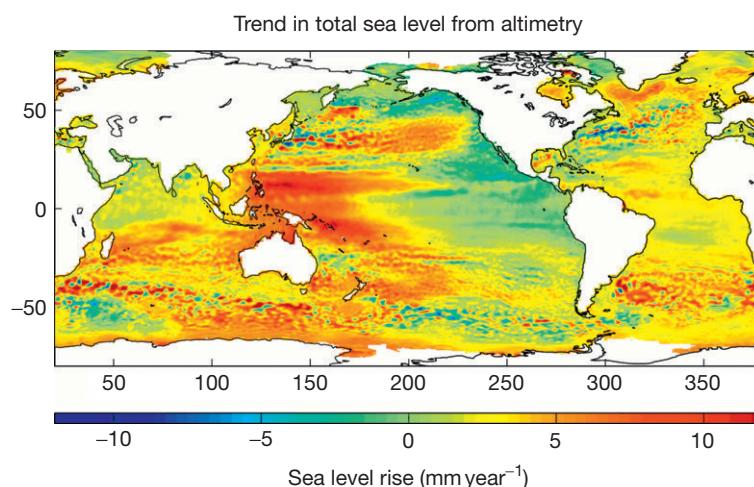


Figure 8 The regional change in sea level based on the 17-year trend from 1993 to 2009 calculated using radar altimeter data from several satellites. Reproduced from Figure 1 of Willis et al. (2010) Global sea level rise: Recent progress and challenges for the decade to come. *Oceanography* 23(4): 26–35, with permission.

attributed to incorrect assumptions of ice history or earth rheology, or the influence of non-GIA processes. The eustatic history inferred from the GIA models (**Figure 3**) is characterized by a lowstand between ~26 and ~21.5 ka BP, followed by rapid ESL rise until ~7 ka BP, with rates peaking at 20–30 mm year⁻¹ around 14.5 ka BP, then a much slower rate of increase (<2.5 mm year⁻¹) which tends to zero by 3–2 ka BP. During each period, rates varied due to dynamic ice sheet processes.

See also: Amino Acid Dating. **Radiocarbon Dating:** Sources of Error. **Sea Level Studies:** Coral Records of Relative Sea-Level Changes; Geomorphological Indicators; Isostasy: Glaciation-Induced Sea-Level Change; Overview; Sedimentary Indicators of Relative Sea-Level Changes – High Energy; Sedimentary Indicators of Relative Sea-Level Changes – Low Energy. **Sea-Levels, Late Quaternary:** Tectonics and Relative Sea-Level Change.

References

- Barber DC, Dyke A, Hillaire-Marcel C, et al. (1999) Forcing of the cold event of 8,200 years ago by catastrophic drainage of Laurentide lakes. *Nature* 400: 344–348.
- Bard E, Hamelin B, Arnold M, et al. (1996) Deglacial sea-level record from Tahiti corals and the timing of global meltwater discharge. *Nature* 382: 241–244.
- Bard E, Hamelin B, and Delanghe-Sabatier D (2010) Deglacial meltwater pulse 1B and Younger Dryas sea levels revisited with boreholes at Tahiti. *Science* 327: 1235–1237.
- Bard E, Hamelin B, Fairbanks RG, and Zindler A (1990) Calibration of the ¹⁴C timescale over the past 30,000 years using mass-spectrometric U–TH ages from Barbados corals. *Nature* 345: 405–410.
- Bassett SE, Milne GA, Mitrovica JX, and Clark PU (2005) Ice sheet and solid earth influences on far-field sea-level histories. *Science* 309: 925–928.
- Bird MI, Fifield LK, Teh TS, Chang CH, Shirlaw N, and Lambeck K (2007) An inflection in the rate of early mid-Holocene eustatic sea-level rise: A new sea-level curve from Singapore. *Estuarine, Coastal and Shelf Science* 71: 523–536.
- Bradley SL, Milne GA, Zong Y, and Horton B (2008) Modelling sea-level data from China and Malay–Thai Peninsula to infer Holocene eustatic sea-level change. American Geophysical Union, Fall Meeting 2008. Abstract #GC33A-0763.
- Camoin GF, Montaggioni LF, and Braithwaite CJR (2004) Late glacial to post glacial sea levels in the Western Indian Ocean. *Marine Geology* 206: 119–146.
- Cazenave A, Dorninh K, Guinehut S, et al. (2009) Sea level budget over 2003–2008: A reevaluation from GRACE space gravimetry, satellite altimetry and Argo. *Global and Planetary Change* 65: 83–88.
- Chappell J, Omura A, Esat T, et al. (1996) Reconciliation of late Quaternary sea levels derived from coral terraces at Huon Peninsula with deep sea oxygen isotope records. *Earth and Planetary Science Letters* 141: 227–236.
- Chappell J and Polach H (1991) Post-glacial sea-level rise from a coral record at Huon Peninsula, Papua New Guinea. *Nature* 349: 147–149.
- Clark PU, Dyke AS, Shakun JD, et al. (2009) The Last Glacial Maximum. *Science* 325: 710–714.
- Clark JA, Farrell WE, and Peltier WR (1978) Global changes in postglacial sea level: A numerical calculation. *Quaternary Research* 9: 265–287.
- Clark PU, McCabe AM, Mix AC, and Weaver AJ (2004) Rapid rise of sea level 19,000 years ago and its global implications. *Science* 304: 1141–1144.
- Cutler KB, Edwards RL, Taylor FW, et al. (2003) Rapid sea-level fall and deep-ocean temperature change since the last interglacial period. *Earth and Planetary Science Letters* 206: 253–271.
- Deschamps P, Durand N, Bard E, et al. (2012) Ice-sheet collapse and sea-level rise at the Bølling warming 14,600 years ago. *Nature* 483: 559–564.
- Digerfeldt G and Hendry MD (1987) An 8000 year Holocene sea-level record from Jamaica: Implications for interpretation of Caribbean reef and coastal history. *Coral Reefs* 5: 165–169.
- Fairbanks RG (1989) A 17,000-year glacio-eustatic sea level record: Influence of glacial melting rates on the Younger Dryas event and deep-ocean circulation. *Nature* 342: 637–642.
- Farrell WE and Clark JA (1976) On postglacial sea level. *Geophysical Journal of the Royal Astronomical Society* 46: 647–667.
- Fleming K, Johnston P, Zwartz D, Yokoyama Y, Lambeck K, and Chappell J (1998) Refining the eustatic sea-level curve since the Last Glacial Maximum using far- and intermediate-field sites. *Earth and Planetary Science Letters* 163: 327–342.
- Fleming K and Lambeck K (2004) Constraints on the Greenland Ice Sheet since the Last Glacial Maximum from sea-level observations and glacial-rebound models. *Quaternary Science Reviews* 23: 1053–1077.
- Gehrels R (2010) Sea-level changes since the Last Glacial Maximum: An appraisal of the IPCC Fourth Assessment Report. *Journal of Quaternary Science* 25: 26–38.
- Gehrels WR, Marshall WA, Gehrels MJ, et al. (2006) Rapid sea-level rise in the North Atlantic Ocean since the first half of the nineteenth century. *The Holocene* 16: 949–965.
- Geyh MA, Kudrass HR, and Streif H (1979) Sea-level changes during the Late Pleistocene and Holocene in the Strait of Malacca. *Nature* 278: 441–443.
- Goodwin ID and Harvey N (2008) Subtropical sea-level history from coral microatolls in the Southern Cook Islands, since 300 AD. *Marine Geology* 253: 14–25.
- Hanebuth TJJ, Stattegger K, and Bojanowski A (2009) Termination of the Last Glacial Maximum sea-level lowstand: The Sunda-Shelf data revisited. *Global and Planetary Change* 66: 76–84.
- Hanebuth T, Stattegger K, and Grootes PM (2000) Rapid flooding of the Sunda Shelf: A late-glacial sea-level record. *Science* 288: 1033–1035.
- Hesp PA, Hung CC, Hilton M, Ming CL, and Turner IM (1998) A first tentative Holocene sea-level curve for Singapore. *Journal of Coastal Research* 14: 308–314.
- Kaufmann G and Lambeck K (2002) Glacial isostatic adjustment and the radial viscosity profile from inverse modeling. *Journal of Geophysical Research-Solid Earth* 107: 15.
- Kendall RA and Mitrovica JX (2007) Radial resolving power of far-field differential sea-level highstands in the inference of mantle viscosity. *Geophysical Journal International* 171: 881–889.
- Kopp RE, Mitrovica JX, Griffies SM, Yin J, Hay CC, and Stouffer RJ (2010) The impact of Greenland melt on local sea levels: A partially coupled analysis of dynamic and static equilibrium effects in idealized water-hosing experiments. *Climatic Change* 103: 619–625.
- Lambeck K (2002) Sea-level change from mid-Holocene to recent time: An Australian example with global implications. Mitrovica JX and Vermeersen LA (eds.) *Ice Sheets, Sea Level and the Dynamic Earth*, pp. 33–50. Washington: American Geophysical Union.
- Lambeck K, Anzidei M, Antonioli F, Benini A, and Esposito A (2004) Sea level in Roman time in the Central Mediterranean and implications for recent change. *Earth and Planetary Science Letters* 224: 563–575.
- Long AJ, Woodroffe SA, Milne GA, Bryant CL, and Wake LM (2010) Relative sea level change in west Greenland during the last millennium. *Quaternary Science Reviews* 29: 367–383.
- Milliman JD and Emery KO (1968) Sea levels during the past 35000 years. *Science* 162: 1121–1123.
- Milne GA and Mitrovica JX (2008) Searching for eustasy in deglacial sea-level histories. *Quaternary Science Reviews* 27: 2292–2302.
- Milne GA, Mitrovica JX, and Schrag DP (2002) Estimating past continental ice volume from sea-level data. *Quaternary Science Reviews* 21: 361–376.
- Mitrovica JX and Forte AM (2004) A new inference of mantle viscosity based upon joint inversion of convection and glacial isostatic adjustment data. *Earth and Planetary Science Letters* 225: 177–189.
- Mitrovica JX and Milne GA (2002) On the origin of late Holocene sea-level highstands within equatorial ocean basins. *Quaternary Science Reviews* 21: 2179–2190.
- Mitrovica JX and Peltier WR (1991) On postglacial geoid subsidence over the equatorial oceans. *Journal of Geophysical Research-Solid Earth* 96: 20053–20071.
- Nakada M and Lambeck K (1988) The melting history of the Late Pleistocene Antarctic ice sheet. *Nature* 333: 36–40.
- Peltier WR (2004) Global glacial isostasy and the surface of the ice-age earth: The ice-5G (VM2) model and grace. *Annual Review of Earth and Planetary Sciences* 32: 111–149.
- Peltier WR and Fairbanks RG (2006) Global glacial ice volume and Last Glacial Maximum duration from an extended Barbados sea level record. *Quaternary Science Reviews* 25: 3322–3337.
- Simpson MJR, Milne GA, Huybrechts P, and Long AJ (2009) Calibrating a glaciological model of the Greenland ice sheet from the Last Glacial Maximum to present-day using field observations of relative sea level and ice extent. *Quaternary Science Reviews* 28: 1631–1657.
- Tarasov L and Peltier WR (2002) Greenland glacial history and local geodynamic consequences. *Geophysical Journal International* 150: 198–229.
- Tjia HD (1996) Sea-level changes in the tectonically stable Malay–Thai Peninsula. *Quaternary International* 31: 95–101.
- van de Plassche O (ed.) (1986) *Sea-Level Research: A Manual for the Collection and Evaluation of Data*. Norwich: Geobooks.

- Walcott RI (1972) Past sea levels, eustasy and deformation of the earth. *Quaternary Research* 2: 1–14.
- Woodroffe SA (2009) Testing models of mid to late Holocene sea-level change, North Queensland, Australia. *Quaternary Science Reviews* 28: 2474–2488.
- Woodroffe SA and Horton BP (2005) Holocene sea-level changes in the Indo-Pacific. *Journal of Asian Earth Science* 25: 29–43.
- Yokoyama Y, Lambeck K, De Deckker P, Johnston P, and Fifield LK (2000) Timing of the Last Glacial Maximum from observed sea-level minima. *Nature* 406: 713–716.
- Zinke J, Reijmer JJG, Thomassin BA, Dullo WC, Grootes PM, and Erlenkeuser H (2003) Postglacial flooding history of Mayotte lagoon (Comoro Archipelago, southwest Indian Ocean). *Marine Geology* 194: 181–196.
- Zong YQ (2004) Mid-Holocene sea-level highstand along the southeast coast of China. *Quaternary International* 117: 55–67.

Isostasy: Glaciation-Induced Sea-Level Change

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Introduction

Relative sea level (RSL) is defined as the height of the ocean surface relative to the solid Earth (or ocean floor) and can be defined at any location within the oceans. Changes in RSL are driven, therefore, by processes that produce a height shift in either of these two bounding surfaces. At any given location, RSL changes over a wide range of time scales with characteristic periods ranging from less than a second to millions of years. This broad spectrum of time variation reflects the variety of processes that drive RSL changes, such as atmospheric circulation, ocean circulation, lunar tides, deformation of the solid Earth associated with internal buoyancy forces (mantle convection), and changes in the distribution of mass at the Earth's surface (e.g., growth and melting of ice sheets, erosion and deposition of rocks).

During the Quaternary, the dominant component of the RSL change observed using proxy records was driven by the accumulation and ablation of major ice sheets. There are a number of physical mechanisms associated with this glaciation-induced sea-level change and these are described in the following section. Models that incorporate these mechanisms can be employed to predict RSL changes which can, in turn, be compared to the observations. It is through this modeling procedure that researchers have been able to better understand the physical processes involved in driving the sea-level changes as well as place constraints on key model parameters. The most common applications of this type will be presented in the section titled 'Applications of Isostatic Sea-Level Models.'

Physical Principles of Glaciation-Induced Sea-Level Change

Vertical Deflection of the Solid Earth Surface

The late Quaternary ice sheets reached thicknesses of several kilometers in places. During a typical glacial cycle, the net flux of mass from the oceans to the ice sheets and back again is on the order of 100 m of eustatic sea-level change. Note that this quantity is also referred to as 'ice-volume equivalent sea-level change' (Lambeck and Chappell, 2001). This redistribution of surface water mass is large enough to cause a significant isostatic deformation of the solid Earth. This deformation is one of the key processes that contributes to glaciation-induced sea-level change (e.g., O'Connell, 1971; Chappell, 1974; Peltier and Andrews, 1976; Clark et al., 1978).

Figure 1 shows a model prediction of vertical solid Earth deformation at present associated with the most recent deglaciation (ca. 20–7 cal ka BP). The signal is dominated by crustal uplift in regions where large ice sheets once existed (e.g., North America, Eurasia) or where the extent and thickness of ice was larger than at present (e.g., Antarctica). Even though ice had

disappeared in North America and Eurasia during the early to mid-Holocene, there remains a significant uplift in these regions due to the relatively high viscosity of mantle material, which limits the rate at which the solid Earth can attain isostatic equilibrium following the removal of ice. Note that there are extensive regions of subsidence peripheral to areas of uplift. This deformation is linked to uplift of the solid Earth in regions peripheral to the growing ice sheets; during deglaciation, this process is reversed and these so-called 'peripheral bulges' subside.

The solid Earth deformation associated with changes in ice mass distribution, known as 'glacioisostasy,' is a dominant contributor to RSL in regions near major ice centers (so-called 'near-field' regions). A significant amount of deformation is also caused by the changes in sea level associated with the changes in ice distribution. The solid Earth deformation associated with this ocean loading, known as 'hydroisostasy,' is most apparent in regions far removed from the major ice centers (so-called 'far-field' regions), where the glacioisostatic signal is considerably reduced. For example, the small uplift around the periphery of Australia and Africa in Figure 1 is associated with the positive ocean load applied around these continents during deglaciation. This uplift has been termed 'continental levering' (e.g., Clark et al., 1978).

The surface mass redistribution and subsequent solid Earth deformation associated with Earth glaciation perturbs the Earth's inertia tensor and therefore affects a change in Earth rotation: commonly described as a change in the spin rate (change in length of day) and a change in the orientation of the spin axis relative to the solid Earth (True Polar Wander; e.g., Sabadini et al., 1982; Mitrovica et al., 2005). The latter of these produces a significant global-scale deformation of the Earth due to the corresponding shift in the rotational potential. The dominant geometry of this deformation is shown in Figure 2. This rotation-induced pattern of deformation is relatively small (approximately 1–2 orders of magnitude less than the glacioisostatic signal), and so is difficult to see in Figure 1. The signal is most evident as a subtle and broad uplift near South America and north-west Asia, and as broad regions of subsidence centered on North America and Australia (the North American component of the signal is masked by the much larger glacioisostatic deformation).

Vertical Deflection of the Ocean Surface

As discussed earlier, the vertical position of the ocean surface is perturbed due to a variety of processes (winds, ocean circulation, tides, etc.). Over time scales of 100s to 1000s of years, the mean height of the ocean surface approximates the geodetic surface known as the geoid, which is an equipotential of the Earth's gravitational field. Therefore, for the purposes of modeling secular sea-level changes associated with Earth glaciation,

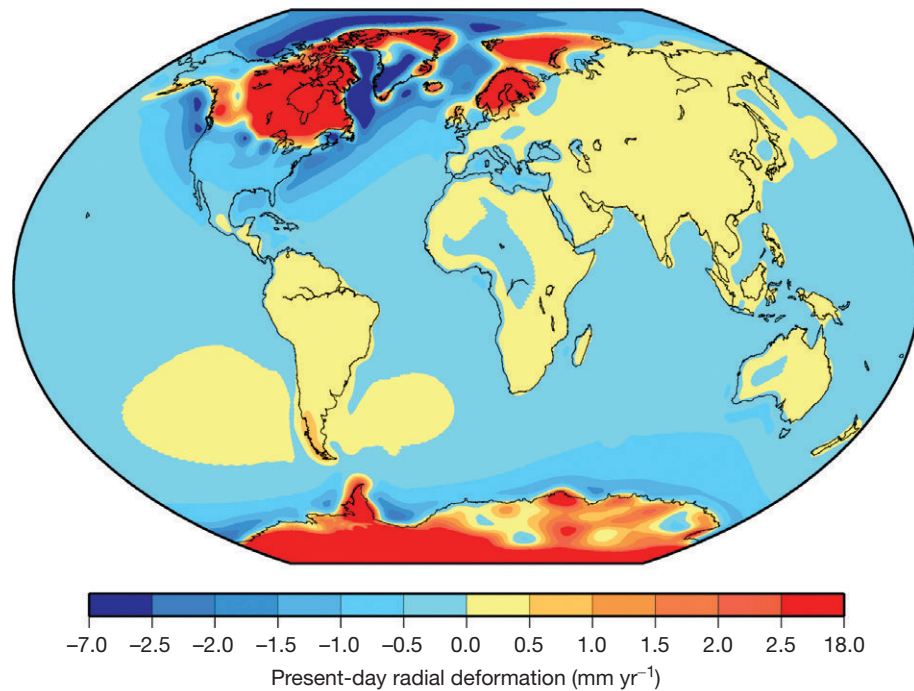


Figure 1 Predicted present-day vertical motion of the solid Earth based on a model of glacial isostatic adjustment. The ice component of the model is adapted from the ICE-5G deglaciation history (Peltier, 2004).

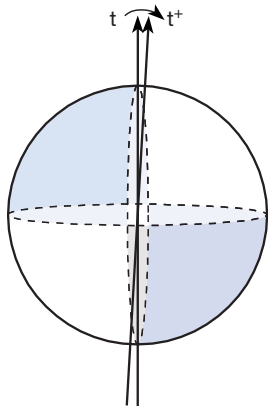


Figure 2 Schematic diagram illustrating the geometry of the vertical solid Earth deformation driven by a clockwise wander of the rotation pole between the times t and t^+ . Shaded areas indicate uplift and nonshaded areas indicate subsidence (due to the associated perturbations in the rotational potential). Reproduced from Mound JE and Mitrovica JX (1998) True polar wander as a mechanism for second-order sea-level variations. *Science* 279: 534–537, with permission.

vertical deflections of the ocean surface are determined by considering changes in the Earth's gravity field (or geopotential).

Any redistribution of mass on or within the Earth will affect the geopotential and therefore perturb the ocean surface. The height shift of the ocean surface caused directly by the flux of mass on the Earth's surface between ice sheets and oceans during glaciations and the changing rotational potential is known as the 'direct effect' (e.g., Clark et al., 1978). The direct influence of the rotational potential on the mean sea surface

exhibits the same geometry as the perturbation to the solid Earth surface (Figure 2). The direct effect associated with an ablating ice sheet is illustrated schematically in Figure 3. Between the two time steps, t_0 and t_1 , the ice sheet loses mass and therefore the gravitational pull of the ice on the ocean water relaxes, causing a lowering of the ocean surface proximal to the ice and a height increase of the ocean surface in the far field. Note that the redistribution of ocean water has a similar effect – for example, regions characterized by a sea-level rise will experience a greater rise due to the increased gravitational pull of a deeper water column.

The perturbation to the geopotential due to the deformation of the solid Earth by the surface ice-ocean loading and changing rotational potential is termed the 'indirect effect' since it is a consequence of the solid Earth response to these changes. The spatial pattern of the present-day sea-surface height change as a consequence of the indirect effect will resemble that shown in Figure 1 (albeit a smoother version of this pattern). For example, the mass influx of solid Earth material to uplifting regions causes an increase in the local geopotential and therefore produces a steady rise in the mean ocean surface. Note, however, that the rates of sea-surface height changes driven by Earth deformation will be approximately an order of magnitude less than those for vertical land motion in Figure 1.

Mass Conservation

An important criterion that must be satisfied when calculating sea-level change associated with the growth and ablation of ice sheets is that the mass lost (gained) by the ice sheets is gained

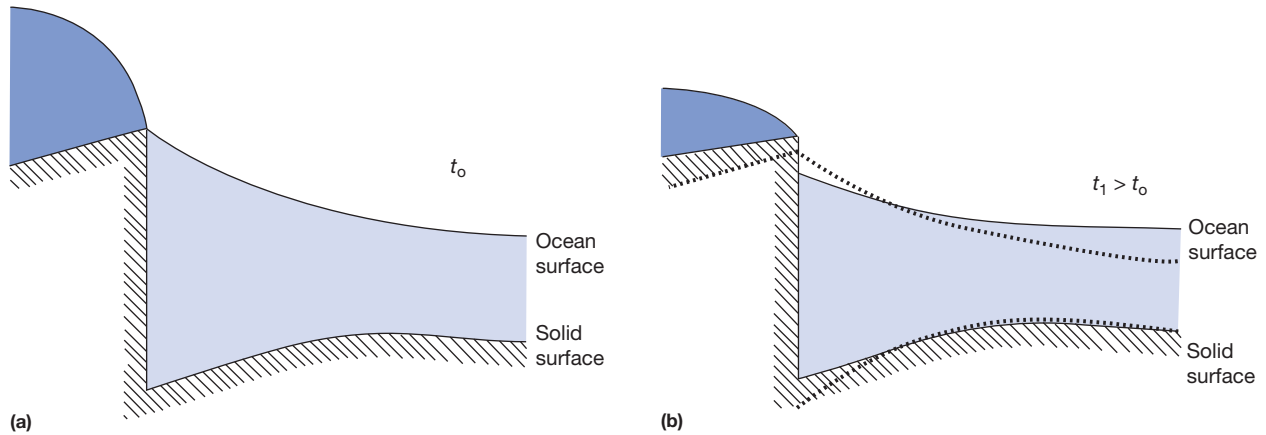


Figure 3 Schematic diagram illustrating the influence of an ablating ice sheet on the vertical displacement of the solid surface and the time-averaged ocean surface (or geoid). A short time interval is assumed (~ 100 years) so that the Earth deformation is largely elastic and of relatively small amplitude. As a result, the spatial variation in ocean surface height change is dominated by the gravity perturbation due to surface mass exchange between the ice sheet and the ocean (i.e., the direct effect). The dotted lines in (b) denote the position of these two surfaces in (a). The difference between the solid and dotted lines in frame (b) illustrates the uplift of the solid Earth immediately under the ablating ice mass and the subsidence of the solid Earth peripheral to the ice mass; it also shows the height shift of the ocean surface. Note that the latter is characterized by a fall near the ice mass and a rise at a greater distance from the ice mass. Reproduced from Tamisiea ME, Mitrovica JX, Davis JL, and Milne GA (2003) Long wavelength sea level and solid surface perturbation driven by polar ice mass variations: Fingerprinting Greenland and Antarctic ice sheet flux. *Space Science Reviews* 108: 81–93, with permission.

(lost) by the oceans to ensure that the total amount of water within the system is conserved. This criteria is satisfied by

$$\iint_{\text{ocean}} S(\theta, \varphi, t) dA = V_{\text{OW}}(t) = -\frac{\rho_i}{\rho_w} V_I(t) \quad [1]$$

in which $S(\theta, \varphi, t)$ is the change in sea level since a reference time (e.g., the onset of ice growth); $V_{\text{OW}}(t)$ and $V_I(t)$ are, respectively, the changes in the volume of ocean water and grounded ice since the reference time; and ρ_i and ρ_w are the densities of ice and water, respectively. Note that the influence of changes in temperature is not considered in eqn [1].

To satisfy eqn [1], the mean ocean surface cannot remain on the same equipotential of the Earth's gravity field. To simulate this in the model, a globally uniform shift in the height of the ocean surface is applied (e.g., Farrell and Clark, 1976; Mitrovica and Peltier, 1991). The amplitude of this shift is governed by two processes: the change in ocean water volume associated with ice sheet growth/melting (known as 'eustatic sea level') and the volumetric change in the height shift between the ocean surface and the ocean floor due to the processes described in the previous two subsections. In the former case, for example, the increase in ocean water volume during deglaciation will cause the ocean surface to rise, on average, and therefore lie on an equipotential further displaced from the Earth's center. The second of these two processes leads to the mechanism known as 'ocean syphoning' (Mitrovica and Peltier, 1991; Mitrovica and Milne, 2002).

Ocean syphoning is driven by two aspects of glaciation-induced sea-level change – these are illustrated in Figure 4. Figure 4(a) shows the contribution to syphoning associated with peripheral bulge subsidence. A dominant proportion of

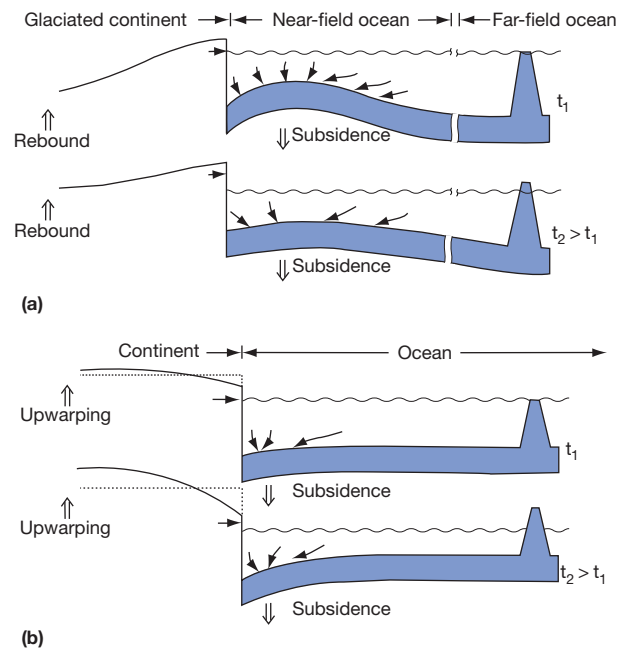


Figure 4 Schematic diagram illustrating the syphoning mechanism. (a) Shows the lowering of the ocean surface caused by the syphoning of ocean water into subsiding peripheral bulge regions adjacent to major glaciation centers during and subsequent to periods of deglaciation. (b) Illustrates the similar effect caused by subsidence around all continental margins due to hydroisostasy during and following periods of deglaciation. Reproduced from Mitrovica JX and Milne GA (2002) On the origin of late Holocene highstands within equatorial ocean basins. *Quaternary Science Reviews* 21: 2179–2190, with permission.

peripheral bulges are located in oceanic regions and so the subsidence of these regions during deglaciation contributes to a significant, spatially uniform fall in sea level. [Figure 4\(b\)](#) illustrates the contribution of continental levering to the syphoning process during deglaciation (again, giving a sea-level fall). In total, the magnitude of the global mean sea-level fall due to syphoning during deglaciation is a few 10s of meters from the Last Glacial Maximum (LGM) to present and is about 0.3 mm year^{-1} at present (the magnitude of the signal depends on the rate of solid earth motion which is governed by the Earth's internal viscosity structure).

Sea-Level Equation

All of the previously described processes can be represented in a single equation that can be solved to predict sea-level change over the world's oceans due to Earth glaciation. [Farrell and Clark \(1976\)](#) were the first to present and solve a version of the sea-level equation. Since their seminal work, a number of extensions have been made to the theory built into the equation as well as the methods applied to solve it. Solving the sea-level equation is nontrivial since the quantity being solved for is a component of the surface load that influences sea level. Most of the methods used involve an iterative procedure and the most popular of these is the so-called 'pseudospectral technique,' which was first applied to the sea-level problem by [Mitrovica and Peltier \(1991\)](#).

The primary theoretical extensions to the original theory of [Farrell and Clark \(1976\)](#) include the incorporation of shoreline migration, the influence of changes in Earth rotation, and a more accurate treatment of sea-level change in regions with ablating marine-based ice (see [Milne, 2002](#) for a review). The most recent form of the sea-level equation that incorporates all of these recent improvements in an accurate, generalized theory was published by [Mitrovica and Milne \(2003\)](#) and can be solved efficiently by an algorithm that is based on the pseudo-spectral method ([Kendall et al., 2005](#)).

Applications of Isostatic Sea-Level Models

Introduction

The previous section describes the processes that govern sea-level change due to mass changes in land ice. These processes are embedded in the sea-level equation, which is solved within the framework of a model of glacial isostatic adjustment (GIA). In essence, a GIA model is defined by two elements: a model of the Earth forcing and a model of solid Earth structure and rheology ([Figure 5](#)). The Earth forcing is specified by defining the surface load (ice-ocean mass redistribution) and the time-varying rotational potential. The ice component of the load

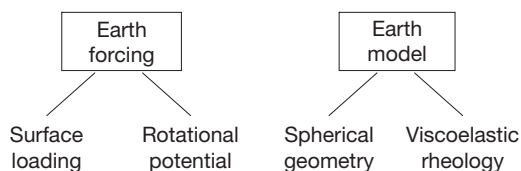


Figure 5 Schematic diagram illustrating the main components comprising a model of glacial isostatic adjustment.

model is commonly developed using constraints from a variety of disciplines, including glacial geology, glaciology, and GIA (see the section 'Inferring Late Quaternary Ice Histories'). The ocean load model is calculated by solving the sea-level equation for a specified ice model and Earth model. In most previous studies, the adopted Earth model has a spherical geometry and internal structure that varies with depth only.

The primary aim of this section is to describe how the Earth and ice components of a GIA model are constructed and then tested by comparing model predictions to sea-level observations, and some of the issues involved in doing this accurately.

Inferring Earth Viscosity Structure

The isostatic response of the solid Earth, for a given load history, is governed by its internal density and rheological structure. To a first approximation, this structure is spherically symmetric (i.e., is depth-dependent only) and so the majority of Earth models previously employed for this purpose have not incorporated lateral variations in these properties. The rheological model applied is most commonly that of a Maxwell viscoelastic material – which simulates both an elastic and (linear) viscous component of the deformation. It is important to include an elastic component given that land ice extent can change significantly within a relatively short time (order 10–100 years).

The elastic and density structure of Earth models applied to problems of glaciation-induced sea-level change is taken from seismic constraints, whereas the viscosity structure is one of the parameter sets that can be inferred when modeling GIA data sets such as RSL change. The viscosity structure of the deep Earth is a primary control on a number of geodynamic processes, including mantle convection, and there are a limited number of methods that can be applied to infer this structure (e.g., [Ranalli, 1995](#)). This application of the model is therefore one of the more important elements in sea-level modeling studies (e.g., [Cathles, 1975](#); [Peltier, 1998](#); [Lambeck and Johnston, 1999](#)).

The viscosity structure in spherically symmetric Earth models is parameterized into a small number of layers. The most common parameterization comprises three layers. The lithosphere, the upper layer, is given a very large viscosity (10^{30} – $10^{50} \text{ Pa s}^{-1}$), so that it behaves as an elastic layer on late Quaternary time-scales – the thickness of this layer is the variable parameter. The second layer extends from the base of the model lithosphere to the 670-km deep seismic velocity discontinuity and the third layer extends from this depth to the core–mantle boundary. These two layers represent the model upper and lower mantle, respectively, and are each assigned a single viscosity value that is a free parameter in the modeling. Note that a number of studies have considered other depth parameterizations (e.g., [Lambeck et al., 1996](#)), and inverse modeling studies commonly parameterize viscosity into a much larger number of layers beneath the lithosphere (e.g., [Forte and Mitrovica, 1996](#); [Peltier, 1998](#)).

As indicated in [Figure 1](#), the largest magnitude of solid Earth deformation is found in areas that have experienced the largest changes in ice extent (e.g., North America, Fennoscandia). For this reason, sea-level observations from regions such as these have been most commonly adopted to infer viscosity

structure. One primary limitation of modeling near-field data is that the model predictions are sensitive to both viscosity structure and the ice history. The accuracy of the viscosity inference will depend on the accuracy of the ice model (see the section 'Inferring late Quaternary ice histories'). This ambiguity can be addressed to some extent by simultaneously modeling for both the ice history and the viscosity structure and performing sensitivity studies to determine the degree of parameter trade-off between these two components of the model (e.g., [Lambeck et al., 1998](#)).

An alternative approach to addressing this issue of non-uniqueness involves identifying a parameterization or subset of the data that is relatively insensitive to one of either the viscosity model or ice model. Two parameterizations of near-field sea-level data that have been shown to reduce the sensitivity of the predictions to the ice history are the 'decay-time' parameterization ([Mitrovia and Peltier, 1993](#)) and the 'relaxation spectrum' derived from mapping out the elevation of ancient shorelines in Fennoscandia ([McConnell, 1968](#)).

A third approach to reduce the sensitivity of the data to uncertainty in the ice component of the model involves the application of far-field sea-level data. By definition, these observations are obtained in areas distant from the major glaciation centers and so the glacioisostatic signal will be relatively small. However, the observations are sensitive to the ice loading, largely through eustatic sea level as well as gravitational and rotational changes. These components of the total sea-level signal have very long wavelengths and so by differencing sea-level observations corresponding to the same time from localities within the same region largely removes the eustatic and glacioisostatic signals. The residual signal is dominated by hydroisostasy. This parameterization of far-field data was pioneered by [Nakada and Lambeck \(1989\)](#) who modeled differential Holocene highstand observations from the Australian region to obtain a robust inference of Earth viscosity structure.

Most inferences of Earth viscosity structure have been based on forward modeling analyses. However, the information obtained from the application of inverse methods has played an important role in partly reconciling the differences in inferences made by various groups ([Mitrovia, 1996](#)). Specifically, the explicit consideration of depth-resolving power in inverse analyses has clearly identified how well a particular set of sea-level data can constrain the depth variation of viscosity. Analyses of this type have shown that the depth-resolving power of a given sea-level data set depends on the spatial extent of the local ice sheet and the viscosity structure within the mantle. The relatively small number of inverse modeling studies performed to date have shown that sublithosphere viscosity structure can be uniquely resolved in only two to three layers for any specific regional data set, with a maximum depth constraint extending to ~ 1800 km for data associated with the deglaciation of the North American ice sheets (e.g., [Mitrovia, 1996](#)).

The limited resolving power of sea-level data has been addressed by inverting these data jointly with other GIA data sets (e.g., [Kaufmann and Lambeck, 2002](#)) and data related to convective flow within the mantle (such as the global free air gravity anomaly, and motion of the tectonic plates) (e.g., [Forte and Mitrovia, 1996](#); [Mitrovia and Forte, 2004](#)). With regard

to the viscosity constraint provided by RSL observations, an optimal resolving power would be obtained by jointly modeling data from regions once covered by ice sheets with a different geographic extent (e.g., the British Isles, Eurasia, and North America).

The real Earth is known to exhibit significant departures from spherical symmetry. With dramatic improvements in computational power, a growing number of groups have begun to model sea-level changes using global three dimensional (3D) Earth models (e.g., [Paulson et al., 2005](#); [Wu et al., 2005](#); [Wang et al., 2008](#)). A primary aim of these studies has been to determine the significance of the 3D structure in predicting sea-level changes in order to assess the extent to which inferences of viscosity and ice histories (see next subsection) based on 1D Earth models might have been biased.

A second important initiative with respect to modeling the isostatic component of the Earth response has been the consideration of rheologies that incorporate transient creep (e.g., [Peltier et al., 1981](#)) or nonlinear viscous flow (e.g., [Wu, 1999](#); [Wu and Wang, 2008](#)). A large number of studies on the deformation of rocks and minerals at high pressures and temperatures in the laboratory have shown that, under a wide range of conditions, mantle rocks will deform in a nonlinear, power law fashion (e.g., [Drury and Fitzgerald, 1998](#)). It is expected that linear and nonlinear viscous flow will dominate in different parts of the sublithospheric mantle. A small number of studies have considered models that include both linear and nonlinear flow (e.g., [van der Wal et al., 2010](#)).

The consideration of lateral variations in Earth model structure and more sophisticated (non-Maxwell) rheologies increases the level of complexity within GIA sea-level models and therefore the degree of nonuniqueness in obtaining model solutions. It remains to be seen just how large an impact these developments will have in this regard.

Inferring Late Quaternary Ice Histories

A second important application of sea-level modeling is the inference of more accurate ice sheet reconstructions – particularly for the period following the LGM during which there is a relative abundance of high-quality RSL data. Constraining the evolution of the Earth's ice sheets during the late Quaternary is addressed within three research disciplines: GIA, Glaciology, and Glacial Geology.

The strong impact of ice sheets on other components of the climate system has become evident through the application of general circulation models of climate change, in which reconstructions of ice sheet evolution are commonly input as a boundary condition. A continued effort to employ RSL observations to better constrain such reconstructions is clearly important to complement information derived from glaciological observations and modeling as well as field-based geological and geophysical studies.

As discussed in the previous subsection, near-field sea-level data can be employed to develop regional and global reconstructions of ice sheets (e.g., [Tushingham and Peltier, 1991](#); [Lambeck, 1993](#); [Lambeck et al., 1998](#)). In the majority of previous studies, little to no glaciological constraints (relating to the physics of ice flow and surface topography) were employed in arriving at the final reconstructions. Some more

recent studies have addressed this issue by using numerical models of ice sheet flow forced by climate (e.g., [Tarasov and Peltier, 2002](#); [Simpson et al., 2009](#)). While the application of these models introduces a larger number of variables to the problem, the resulting reconstructions are more realistic and can be used to infer additional information regarding the mechanisms that governed the evolution of the ice sheet(s).

The veracity of past ice sheet reconstructions depends largely on the quality of the data employed in their construction (sea level and other), the accuracy of the models applied (GIA/glaciological), and the significance of any trade-offs between earth and ice model parameters. As discussed in the previous subsection, an efficient way in which to reduce the ambiguity due to parameter trade-off is to identify parameterizations or subsets of the data that are relatively insensitive to either the Earth or ice component of the model. For example, employing a viscosity model that is based on the decay-time parameterization (which is relatively insensitive to the ice loading model; see previous subsection) to derive an ice sheet reconstruction would be an efficient way to minimize the non-uniqueness problem inherent within near-field sea-level data.

On account of their distance from the major glaciation centers, far-field RSL data exhibit a primary sensitivity to mass flux between the ice sheets and oceans. For this reason, when the data are corrected for the processes that cause the total signal to deviate from the eustatic component signal (due to glacio- and hydroisostasy, syphoning, etc.; see previous section 'Physical Principles of Glaciation-Induced Sea-Level Change'), they provide a useful measure of the total volume of grounded ice on the Earth's surface in the past compared to the present, which can be compared to, or used to calibrate, ice volume estimates based on oxygen isotope records.

Estimates of grounded ice volume at the LGM have converged within the past few years and generally lie within the range 120–135 m of eustatic sea-level change (e.g., [Lambeck et al., 2002](#); [Milne et al., 2002](#); [Peltier, 2004](#)). These estimates serve as an important constraint which global LGM ice sheet reconstructions must adhere to. Note that the most recent estimates of LGM ice volume support the CLIMAP 'minimum model' with 127 m of eustatic sea-level lowering compared to the 'maximum model' with 163 m ([CLIMAP, 1981](#)).

The modeling of post-LGM far-field RSL data has greatly enhanced our understanding of the rate and sources of glacial melting from the LGM to the present. Far-field observations have been interpreted to support the existence of a series of very rapid and short-lived melting events during the most recent deglaciation. For example, from the LGM to the Late Glacial, three such events have been proposed:

- ~19 cal ka BP (10–15 m; [Yokoyama et al., 2000](#));
- ~14 cal ka BP (20–30 m; [Fairbanks, 1989](#); [Bard et al., 1990](#)); and
- ~11.5 cal ka BP (10–15 m; [Fairbanks, 1989](#); [Bard et al., 1990](#)).

For a more detailed discussion of data relating to these proposed rapid events, see elsewhere in the encyclopedia. Determining how these so-called 'meltwater pulses' relate to other large and rapid climate events during this period is central to our understanding of deglacial- and millennial-scale climate change. A necessary first step toward this aim involves

determining which ice reservoirs were the primary contributors to these events. Isostatic sea-level models have provided useful information in this regard.

In far-field locations, the direct gravitational effect of ice sheet ablation (see [Figure 4](#) and related discussion) is the mechanism that dominates the spatial variation in the predicted far-field sea-level signal during rapid melt events. The contribution from solid Earth isostatic deformation is relatively minor and so, given an appropriate spatial distribution of data that capture a specific meltwater pulse, it is possible to obtain relatively robust constraints on the global-scale melt source geometry for that event. The first studies of this kind have shown, for example, that the existing far-field data favor a dominant Antarctic rather than North American source for the ~14 cal ka BP event ([Clark et al., 2002](#); [Bassett et al., 2005](#)). This melt source model can be tested further and spatially refined by modeling near-field sea-level data from near-field regions.

A number of far-field RSL observations display a high stand during the mid-Holocene. Sea-level modeling has shown that this high stand results from a dramatic decrease in the rate of glacial melting around this time, which leads to the subsequent sea-level fall through the operation of ongoing isostatic effects, primarily syphoning and hydroisostasy ([Clark et al., 1978](#); [Mitrovica and Peltier, 1991](#)). The shape (e.g., smooth or sharp), timing, and magnitude of these high stands is controlled by both the isostatic response of the solid Earth and the rate and magnitude of glacial melting. The isostatic component can be effectively isolated and therefore modeled by considering spatial differences in the observed sea-level high stands ([Nakada and Lambeck, 1989](#); see previous subsection). The solid Earth component can then be predicted and removed from the observations, leaving a residual RSL signal for the period during and following the high stand that provides a direct measure of when and how abruptly the most recent deglaciation ended and the history of global-scale melting after this time (e.g., [Fleming et al., 1998](#); [Peltier, 2002](#)). The latter serves as an important reference value with which to compare estimates of global mean sea-level rise during the twentieth century (e.g., [Lambeck, 2002](#)).

Conclusion

Changes in RSL during the Quaternary are dominated by large-scale glaciation. A number of physical mechanisms contribute to glaciation-induced sea-level changes through their influence on the relative vertical height between the ocean floor and the ocean surface. These physical mechanisms provide the theoretical basis of the so-called sea-level equation, which can be solved to generate predictions of glaciation-induced sea-level changes for a prescribed Earth and ice model. These predictions can be compared to observations to test the accuracy of the model and to infer model parameters. With respect to the Earth model, the majority of such studies have focused on inferring the viscosity structure within the mantle. This structure is a primary control on a variety of geodynamic processes that involve the solid-state creep of mantle material. With respect to the ice model, the majority of such studies have focused on inferring the

space–time history of regional-scale ice extent using near-field sea-level data, as well as the time variation in global (grounded) ice volume, during and following the LGM. This information has played a key role in advancing our understanding of Ice Age climate change.

See also: **Paleoceanography, Biological Proxies:** Benthic Foraminifera. **Paleoclimate Modeling:** Last Glacial Maximum GCMs. **Sea Level Studies:** Eustatic Sea-Level Changes – Glacial–Interglacial Cycles; Eustatic Sea-Level Changes Since the Last Glacial Maximum.

References

- Bard E, Hamelin B, Fairbanks RG, and Zindler A (1990) Calibration of the ^{14}C timescale over the past 30,000 years using mass spectrometric U–Th ages from Barbados corals. *Nature* 345: 405–410.
- Bassett SE, Milne GA, Mitrovica JX, and Clark PU (2005) Ice sheet and solid earth influences on far-field sea-level histories. *Science* 309: 925–928.
- Cathles LM (1975) *The Viscosity of the Earth's Mantle*, pp. 386. Princeton, NJ: Princeton University Press.
- Chappell J (1974) Late Quaternary glacio- and hydro-isostasy, on a layered earth. *Quaternary Research* 4: 429–440.
- Clark JA, Farrell WE, and Peltier WR (1978) Global changes in postglacial sea level: A numerical calculation. *Quaternary Research* 9: 265–287.
- Clark PU, Mitrovica JX, Milne GA, and Tamisiea M (2002) Sea-level fingerprinting as a direct test for the source of global meltwater pulse 1A. *Science* 295: 2438–2441.
- CLIMAP (1981) Seasonal reconstruction of the Earth's surface at the last glacial maximum. *Geological Society of America, Map and Chart Series* C36.
- Drury MR and Fitzgerald JD (1998) Mantle rheology: Insights from laboratory studies of deformation and phase transition. In: Jackson I (ed.) *The Earth's Mantle: Composition, Structure and Evolution*, pp. 503–560. Cambridge: Cambridge University Press.
- Fairbanks RG (1989) A 17000-year glacio-eustatic sea level record: Influence of glacial melting rates on the Younger Dryas event and deep-ocean circulation. *Nature* 342: 637–642.
- Farrell WE and Clark JA (1976) On postglacial sea-level. *Geophysical Journal of the Royal Astronomical Society* 46: 647–667.
- Fleming K, Johnston P, Zwart D, Yokoyama Y, Lambeck K, and Chappell J (1998) Refining the eustatic sea-level curve since the Last Glacial Maximum using far- and intermediate-field sites. *Earth and Planetary Science Letters* 163: 327–342.
- Forte AM and Mitrovica JX (1996) A new inference of mantle viscosity based on a joint inversion of post-glacial rebound data and long-wavelength geoid anomalies. *Geophysical Research Letters* 23: 1147–1150.
- Kaufmann G and Lambeck K (2002) Glacial isostatic adjustment and the radial viscosity profile form inverse modelling. *Journal of Geophysical Research* 107: 2280. <http://dx.doi.org/10.1029/2001JB000941>.
- Kendall R, Mitrovica JX, and Milne GA (2005) On post-glacial sea level – II. Numerical formulation and comparative results on spherically symmetric models. *Geophysical Journal International* 161: 679–706.
- Lambeck K (1993) Glacial rebound of the British Isles – II. A high resolution, high-precision model. *Geophysical Journal International* 115: 960–990.
- Lambeck K (2002) Sea level change from mid-Holocene to recent time: An Australian example with global implications. In: Mitrovica JX and Vermeersen LLA (eds.) *Glacial Isostatic Adjustment and the Earth System: Sea Level, Crustal Deformation, Gravity and Rotation, AGU Monograph, Geodynamics Series*, vol. 29, pp. 33–50. Washington DC: American Geophysical Union.
- Lambeck K and Chappell J (2001) Sea level change through the last glacial cycle. *Science* 292: 679–686.
- Lambeck K and Johnston P (1999) The viscosity of the mantle: Evidence from analyses of glacial-rebound phenomenon. In: Jackson I (ed.) *The Earth's Mantle: Composition, Structure and Evolution*, pp. 461–502. Cambridge: Cambridge University Press.
- Lambeck K, Johnston P, Smither C, and Nakada M (1996) Glacial rebound of the British Isles – III. Constraints on mantle viscosity. *Geophysical Journal International* 125: 340–354.
- Lambeck K, Smither C, and Johnston P (1998) Sea-level change, glacial rebound and mantle viscosity for northern Europe. *Geophysical Journal International* 134: 102–144.
- Lambeck K, Yokoyama Y, and Purcell A (2002) Into and out of the Last Glacial Maximum: Sea-level change during Oxygen Isotope Stages 3 and 2. *Quaternary Science Reviews* 21: 343–360.
- McConnell RK (1968) Viscosity of the mantle from relaxation time spectra of isostatic adjustment. *Journal of Geophysical Research* 73: 7089–7105.
- Milne GA (2002) Recent advances in predicting glaciation-induced sea-level changes and their impact on model applications. In: Mitrovica JX and Vermeersen LLA (eds.) *Glacial Isostatic Adjustment and the Earth System: Sea Level, Crustal Deformation, Gravity and Rotation, AGU Monograph, Geodynamics Series*, vol. 29, pp. 157–176. Washington DC: American Geophysical Union.
- Milne GA, Mitrovica JX, and Schrag DP (2002) Estimating past continental ice volume from sea-level data. *Quaternary Science Reviews* 21: 361–376.
- Mitrovica JX (1996) Haskell [1935] revisited. *Journal of Geophysical Research* 101: 555–569.
- Mitrovica JX and Forte AM (2004) A new inference of mantle viscosity based upon joint inversion of convection and glacial isostatic adjustment data. *Earth and Planetary Science Letters* 225: 177–189.
- Mitrovica JX and Milne GA (2002) On the origin of postglacial ocean syphoning. *Quaternary Science Reviews* 21: 2179–2190.
- Mitrovica JX and Milne GA (2003) On post-glacial sea level – I. General theory. *Geophysical Journal International* 154: 253–267.
- Mitrovica JX and Peltier WR (1991) On post-glacial geoid subsidence over the equatorial oceans. *Journal of Geophysical Research* 96: 20053–20071.
- Mitrovica JX and Peltier WR (1993) A new formalism for inferring mantle viscosity based on estimates of post glacial decay times: Application to RSL variations in N.E. Hudson Bay. *Geophysical Research Letters* 20: 2183–2186.
- Mitrovica JX, Wahr J, Matsuyama I, and Paulson A (2005) The rotational stability of an ice-age earth. *Geophysical Journal International* 161: 491–506.
- Nakada M and Lambeck K (1989) Late Pleistocene and Holocene sea-level change in the Australian region and mantle rheology. *Geophysical Journal International* 96: 497–517.
- O'Connell RJ (1971) Pleistocene glaciation and the viscosity of the lower mantle. *Geophysical Journal* 23: 299–327.
- Paulson A, Zhong S, and Wahr J (2005) Modelling post-glacial rebound with lateral viscosity variations. *Geophysical Journal International* 163: 357–371.
- Peltier WR (1998) Postglacial variations in the level of the sea: Implications for climate dynamics and solid-earth geophysics. *Reviews of Geophysics* 36: 603–689.
- Peltier WR (2002) On eustatic sea level history: Last Glacial Maximum to Holocene. *Quaternary Science Reviews* 21: 377–396.
- Peltier WR (2004) Global glacial isostasy and the surface of the ice-age earth: The ICE-5G (VM2) model and GRACE. *Annual Review of Earth and Planetary Sciences* 32: 111.
- Peltier WR and Andrews JT (1976) Glacial isostatic adjustment – I. The forward problem. *Geophysical Journal of the Royal Astronomical Society* 46: 605–646.
- Peltier WR, Wu P, and Yuen DA (1981) The viscosities of the Earth's mantle. In: Stacey FD, Paterson MS, and Nicolas A (eds.) *Anelasticity in the Earth. Geodynamics Series*, vol. 4, pp. 59–77. Boulder, CO: AGU and GSA.
- Ranalli G (1995) *Rheology of the Earth*, 2nd edn., p. 413. London: Chapman and Hall.
- Sabadini R, Yuen DA, and Boschi E (1982) Polar wander and the forced responses of a rotating, multilayered, viscoelastic planet. *Journal of Geophysical Research* 87: 2885–2903.
- Simpson MJR, Milne GA, Huybrechts P, and Long AJ (2009) Calibrating a glaciological model of the Greenland ice sheet from the last glacial maximum to present-day using field observations of relative sea level and ice extent. *Quaternary Science Reviews* 28: 1631–1657.
- Tarasov L and Peltier WR (2002) Greenland glacial history and local geodynamic consequences. *Geophysical Journal International* 150: 198–229.
- Tushingham AM and Peltier WR (1991) ICE-3G: A new global model of late Pleistocene deglaciation based on geophysical predictions of post-glacial relative sea level change. *Journal of Geophysical Research* 96: 4497–4523.
- Van der Wal W, Wu P, Wang H, and Sideris MG (2010) Sea levels and uplift rate from composite rheology in glacial isostatic adjustment modeling. *Journal of Geodynamics* 50: 38–48. <http://dx.doi.org/10.1016/j.jog.2010.01.006>.
- Wang H, Wu P, and Van der Wal W (2008) Using postglacial sea level, crustal velocities and gravity-rate-of-change to constrain the influence of thermal effects on mantle lateral heterogeneities. *Journal of Geodynamics* 46: 104–117. <http://dx.doi.org/10.1016/j.jog.2008.03.003>.

- Wu P (1999) Modelling postglacial sea levels with power-law rheology and a realistic ice model in the absence of ambient tectonic stress. *Geophysical Journal International* 139: 691–702.
- Wu P and Wang HS (2008) Postglacial isostatic adjustment in a self-gravitating spherical Earth with power-law rheology. *Journal of Geodynamics* 46: 118–130. <http://dx.doi.org/10.1016/j.jog.2008.03.008>.
- Wu P, Wang H, and Schotman H (2005) Postglacial induced surface motions, sea-levels and geoid rates on a spherical, self-gravitating laterally heterogeneous earth. *Journal of Geodynamics* 39: 127–142.
- Yokoyama Y, Lambeck K, De Deckker P, Johnston P, and Fifield LK (2000) Timing of the Last Glacial Maximum from observed sea-level minima. *Nature* 406: 713–716.

Use of Cave Data in Sea-Level Reconstructions

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Introduction

The use of radiometrically dated cave deposits offers a generally rare but important contribution to understanding sea-level oscillations during the late Quaternary. Most of these data originate from speleothems recovered in caves that are presently submerged by seawater but can provide direct evidence for periods of time when the cave was either emergent (above sea level), situated coincident with the sea surface, or submerged. In particular, the timing of speleothem growth, phreatic overgrowths, or biogenic encrustations have all been employed to infer periods of time when the cave was above, at, or below local sea level, respectively.

During recent glacial–interglacial cycles, direct observational evidence of sea-level position and timing rests largely on radiometric dates of sedimentary materials that have a definable relationship to the sea surface. In particular, U-series dating of fossil corals that grew near the sea surface have been instrumental in defining the timing and structure of past sea-level change (Edwards et al., 1987, 2003). The sea-level archive contained in submerged speleothems is complementary to that of corals in two significant ways. First, porous, aragonite coral skeletons are susceptible to alteration with time; hence, the utility of this record rapidly degrades prior to the last full glacial cycle. Fortunately, dense spelean calcite is much more resistant to alteration and can more readily be applied to the full-time range accessible by U-series dating (Richards and Dorale, 2003). Second, corals grow at or near the sea surface and therefore provide a minimum estimate of sea-level position. In contrast, speleothems grow in caves above sea level and hence provide an upper bound to sea surface elevation, which helps to bracket the position of sea level when combined with coral data.

The additional value of the submerged speleothem record is that it has the potential to shed light on sea level across the full range of elevations experienced by sea level during the late Quaternary, to the degree that these are represented in sample collections. For purposes of constructing comprehensive sea-level curves, data from glacial periods are essential as glacials dominate the Pleistocene; they comprise about 90% of the time relative to the shorter-lived interglacials. Unfortunately, there is little direct observational evidence for sea-level position during glacial periods because of the difficulty entailed in recovering samples (such as corals or speleothems) that remain submerged by seawater. Hence, the limited submerged speleothem archive has served as a valuable window into the reconstruction of sea-level lowstands, the details of which are generally less well known than their counterpart highstands that are better expressed in the sedimentary record, physically more accessible, and correspondingly better constrained in terms of both elevation and radiometric age.

Sea-level observations from cave data also provide critical benchmarks to test against other sea-level reconstructions. For glacial–interglacial cycles prior to the inception of the last

interglacial sea-level highstand (128 ka; Stirling et al., 1998), sea-level curves are primarily based on the chemical composition of carbonate shells of benthic Foraminifera that reflect the combined influence of ice volume and temperature. Compilations of benthic oxygen isotope ($\delta^{18}\text{O}$) data are often used to infer sea-level variability and corresponding volumes of grounded ice despite the recognition that the relative contributions of the ice volume and water temperature components to this signal are uncertain in terms of both magnitude and timing and that the relationship between these components may not behave in a consistent, predictable manner over multiple glacial–interglacial cycles (e.g., Chappell and Shackleton, 1986; Lisiecki and Raymo, 2005). In light of this, several different approaches have been taken to attempt to strip out the ice volume signal using combinations of proxies or modeling-based reconstructions (Bintanja et al., 2005; Lea et al., 2002; Siddall et al., 2003; Waelbroeck et al., 2002). These sea-level reconstructions can differ from each other significantly and hence, it is critical to provide benchmarks of direct observational evidence for sea level that are well constrained in terms of both elevation and time to test existing sea-level models and reconstructions. Cave data can be employed to test these reconstructions, particularly beyond the last full glacial cycle when well-dated, direct observational evidence of sea-level position becomes increasingly scarce.

Of course cave data, like other archives of sea-level data, must be considered in the context of past and present changes in the Earth's surface that result from glacio-hydro-isostasy. Variations in the loading of volumes of ice and water on Earth's surface on glacial–interglacial timescales impart a signal on sea-level observations – whether they be historical tide gauge measurements, Holocene or Quaternary sea-level data, or paleoshorelines that were formed deeper in geologic time. Even cave data from localities that are generally considered stable platforms may have relative sea-level histories that depart significantly from the global, or eustatic, sea-level curve (Farrell and Clark, 1976) depending, in part, on their distance from former ice sheet margins and, on longer timescales relating to dynamic topography of mantle flow (Moucha et al., 2008). The influence of glacio-hydro-isostasy is particularly relevant for the submerged speleothem record and for biogenic and phreatic overgrowths on speleothems, which have been dominantly reported in two regions (the Bahamas and Mediterranean) that have both been situated on peripheral bulges of former land-based ice sheets in the Northern Hemisphere.

Speleothem Growth: Providing Upper Bounds on Sea-Level Position

Presently submerged caves as well as some emergent caves may contain speleothems – stalagmites, stalactites, or flowstones – some of which may display hiatuses formed during seawater incursions into the cave. Such specimens provide rare and

informative markers of relative sea level. Using precise and accurate U-series dating techniques to establish the age of the spelean calcite defines the period of time when the cave was positioned above the sea surface. While all speleothems must have formed at some elevation above sea level, specimens with growth interruptions related to cave flooding by seawater can be used to establish the timing of sea-level oscillations by dating spelean carbonate on either side of the hiatus marking submarine conditions. These hiatuses are sometimes marked by dissolution features, Fe-rich flocculates or sediment layers, or seawater evaporates such as halite and gypsum (Antonioli et al., 2004; Dutton et al., 2009a; Li et al., 1989; Suric et al., 2009). In rare circumstances, biogenic crusts such as serpulid calcite (worm tubes) encase speleothems during marine submergence and can be preserved within the hiatus of speleothem growth (Antonioli et al., 2004; Bard et al., 2002; Dutton et al., 2009a). Several specimens from the Bahamas and the Mediterranean, including those that contain hiatuses due to sea-level transgressions, have been examined in this manner, and provide important constraints for sea-level oscillations over the last two glacial–interglacial cycles (e.g., Bard et al., 2002; Dutton et al., 2009a; Harmon et al., 1983; Li et al., 1989; Lundberg and Ford, 1994; Richards et al., 1994; Spalding and Mathews, 1972; Suric et al., 2009).

Submerged speleothems (predominantly stalagmites and flowstones) were first exploited as an archive of sea-level change using U-series alpha-dating techniques for specimens recovered in the Bahamas and Bermuda (Gascoyne, 1984; Gascoyne et al., 1979; Harmon et al., 1981, 1983; Spalding and Mathews, 1972). These studies included specimens from up to 16 m above sea level and submerged samples extending down to –20 m. Improvements in precision offered by mass spectrometry allowed for more precise determinations of previous sea-level highstands in the Bahamas based on a submerged flowstone (DWBAH) (Li et al., 1989; Lundberg and Ford, 1994) extending back to marine isotope stage (MIS) 8 (~275 ka). There remains some uncertainty around the accuracy of dates from this flowstone, because this region can have unusual compositions of detrital thorium (Richards and Dorale, 2003). The assumption that is made regarding the composition of the detrital thorium can change the ages significantly (by ~13 ka in one sample from the DWBAH flowstone). However, the number of hiatuses preserved in this flowstone still provides valuable insight into the relative elevation of various sea-level highstands in the past.

Sample collections extending over a range of depths (0 to –60 m) from blue holes in the Bahamas further defined periods of cave emergence and submergence during the last and penultimate glacial period (Richards et al., 1994; Smart et al., 1998; Figure 1). Richards et al. (1994) noted that the submerged speleothem data displayed a structure of gradually lower sea levels approaching the glacial maximum followed by rapid recovery of sea level during the deglaciation that mimicked the pattern long observed in marine oxygen isotope record (Emiliani, 1955). Moreover, Richards et al. (1994) recognized that speleothem growth ceased in some samples even before rising sea levels could have interrupted their growth, implicating another mechanism controlling speleothem growth rates such as reduction of groundwater recharge due to aridity.

Hiatuses observed in submerged stalagmites recovered from various sites along the Mediterranean coastline have also been dated to reconstruct past sea-level oscillations, particularly the sea-level highstands associated with MIS 5 and 7 (Bard et al., 2002; Dutton et al., 2009a; Suric et al., 2009). The work along the Italian coastline in Argentarola Cave (Bard et al., 2002; Dutton et al., 2009a) is notable for providing important constraints on the timing of sea-level highstands during MIS 7, a period of time that has far fewer radiometric ages to constrain sea-level oscillations than those of the last glacial cycle (Figure 2). This work also combined data from specimens across a limited depth range of a few meters, and recognized a different number of highstands recorded in different stalagmites that was used to interpret the elevation of sea-level position during MIS 7.3 (Dutton et al., 2009a). This observation in particular proved useful in testing various reconstructions of sea level during MIS 7 derived from marine cores or models that differed significantly in terms of the relative elevations predicted for the MIS 7 sea-level highstands.

Suric et al. (2009) dated the timing of hiatuses preserved in a submerged stalagmite from Croatia to interpret a double peak in sea level for MIS 5a. This observation provides corroboration for evidence of such a double peak observed at coral-based sea-level reconstruction from Barbados (Potter et al., 2004; Schellmann and Radtke, 2004). Because relative sea-level records can be influenced by local effects such as tectonic or isostatic effects, the establishment of sea-level oscillations such as those during MIS 5a should be recognized as a globally distributed feature before invoking a eustatic origin to the signal, making the discovery of this double peak in Croatia as well as in Barbados particularly important.

Where tectonic movement is a factor, submerged speleothem data can also be invoked in a different manner. Provided a reasonable eustatic sea-level (and ice sheet) history in combination with glacio-hydro-isostatic modeling to characterize the relative sea-level signal, the timing of speleothem growth at different elevations along a tectonically active coastline can be used to constrain the rate of vertical movement. This approach was used to define a range of reasonable uplift rates for the Sicilian coastline, whereby the timing of speleothem growth was limited to the area above the sea-level curve (Dutton et al., 2009b; Figure 3). Suric et al. (2009) also make a similar comparison with their MIS 5a data to constrain uplift rates along the Adriatic coast, but in this case the comparison was made to a eustatic sea-level curve (Lambeck and Chappell, 2001) rather than a model-derived prediction for relative sea level.

In all cases where spelean calcite has been dated to establish the timing of cave emergence, these ages provide minimum estimates of the period of time that the cave was above sea level. Speleothem growth may terminate before sea-level rise due to changing climatic conditions (Bard et al., 2002; Richards et al., 1994) or changes in cave hydrology and it is also not clear how long it takes after sea level has dropped for speleothem growth to resume. In some cases this uncertainty may be within the error of the age determination, but in other instances the delay may be more significant. Despite long periods of seawater submergence, many speleothems show no evidence of dissolution, but in other cases there is some evidence of dissolution that has been attributed to

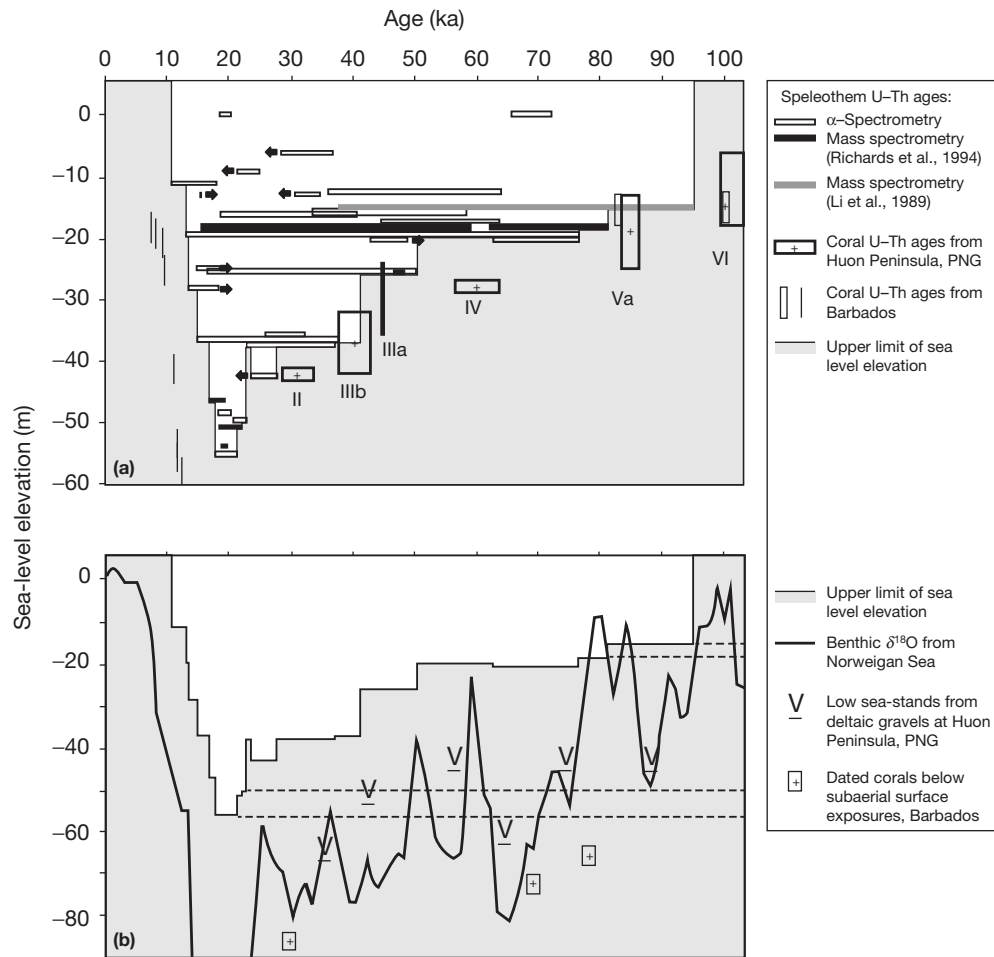


Figure 1 Sea-level data from speleothems in Bahama 'blue holes.' Constrained age/elevation sea-level curve based on speleothems from the Bahamas. (Sea-level elevation defined as distance below present sea-level.) (a) Horizontal bars represent single ages (arrows associated with bars indicate whether top or basal age only), or periods of continuous growth extrapolated from basal and top ages. Both α -spectrometric (white bars) and mass-spectrometric analyses (black bars, and gray bar from Li et al. (1989)) are shown. These bars constrain (± 0.1 m) the upper limit of sea-level elevations to within the shaded portion of the figure (subsidence rates are not taken into account). For comparison, estimates of age and elevation for high sea-stands, based on ^{230}Th ages of coral reef terraces are illustrated: Huon Peninsula terraces II–VI (Aharon and Chappell, 1986) (heavy boxes) and Barbados (Bard et al., 1990; Gallup et al., 1994) (light boxes). The postglacial sea-level curve is shown to highlight the cessation of speleothem growth before inundation by rising sea levels. (b) Comparison with inferred record of ice volume from $\delta^{18}\text{O}$ of benthic Foraminifera from the Norwegian Sea (Labeyrie et al., 1987) and low sea-stands based on deltaic gravels from the Huon Peninsula (Chappell and Shackleton, 1986) (V) and dated corals below subaerial surface exposures, Barbados (light boxes). Periods of time represented by hiatuses in growth for several speleothems denoted by dashed lines. Modified from Richards DA, Smart PL, and Edwards RL (1994) Maximum sea levels for the last glacial period from U-series ages of submerged speleothems. *Nature* 367: 357–360, with permission.

calcite-undersaturated waters of the mixing zone between fresh water and seawater (Dutton et al., 2009a; Richards and Dorale, 2003).

Marine Submergence: Providing Lower Bounds on Sea-Level Position

Organisms that encrust or bore into submerged speleothems when caves become inundated with seawater can also be dated to establish a minimum age for cave flooding. Antonioli and Oliverio (1996) established the radiocarbon age of a mussel that had bored into a submerged speleothem

recovered from -48 m relative to present mean sea level in Palinuro Cave along the west coast of Italy. The age of the mussel represents a minimum estimate for the timing of cave flooding during the last deglaciation. Serpulid calcite crusts on submerged speleothems have also been ^{14}C dated to determine the timing of cave flooding in Palinuro Cave, Argentarola Cave, and near Siracusa, Italy (Antonioli et al., 2001; Dutton et al., 2009b). The timing of cave inundation has been compared in both studies to relative sea-level curves derived from glacio-hydro-isostatic modeling, which show good agreement with the radiocarbon ages. These ages provide important benchmarks to test the models of relative sea-level change in this region.

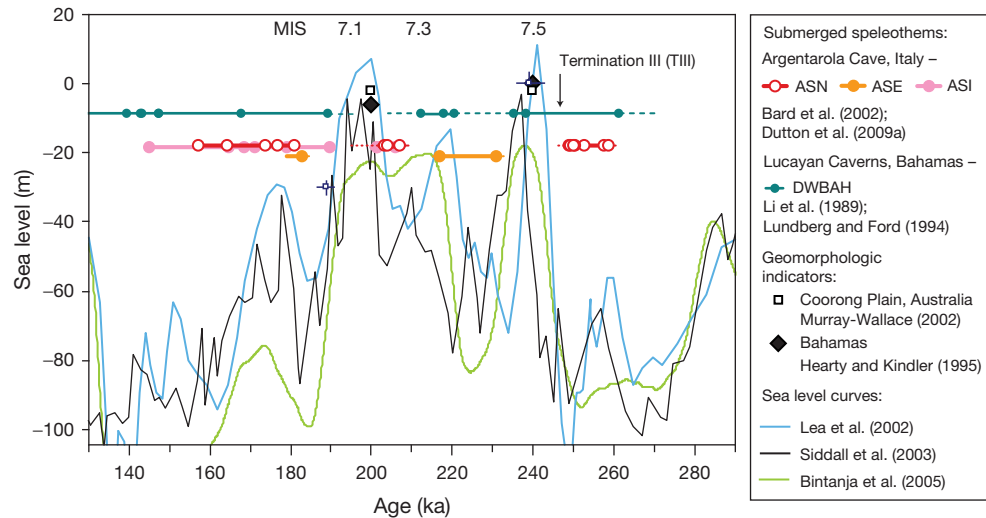


Figure 2 Summary of MIS 7 sea-level highstands with constraints from submerged stalagmites recovered at -18 to -21.5 m below sea level in Argentarola Cave, Italy. Submerged stalagmite U-series data represent growth periods and should be above the sea-level curve. Solid lines connecting speleothem U-series ages represent periods of continuous growth. The error bars for samples next to hiatuses are shown as dashed lines. Peak MIS 7.5 coral data were assumed to sit at modern sea level. Modified from Dutton A, Bard E, Antonioli F, Esat TM, Lambeck K, and McCulloch MT (2009) Phasing and amplitude of sea-level and climate change during the penultimate interglacial. *Nature Geoscience* 2: 355–359, with permission.

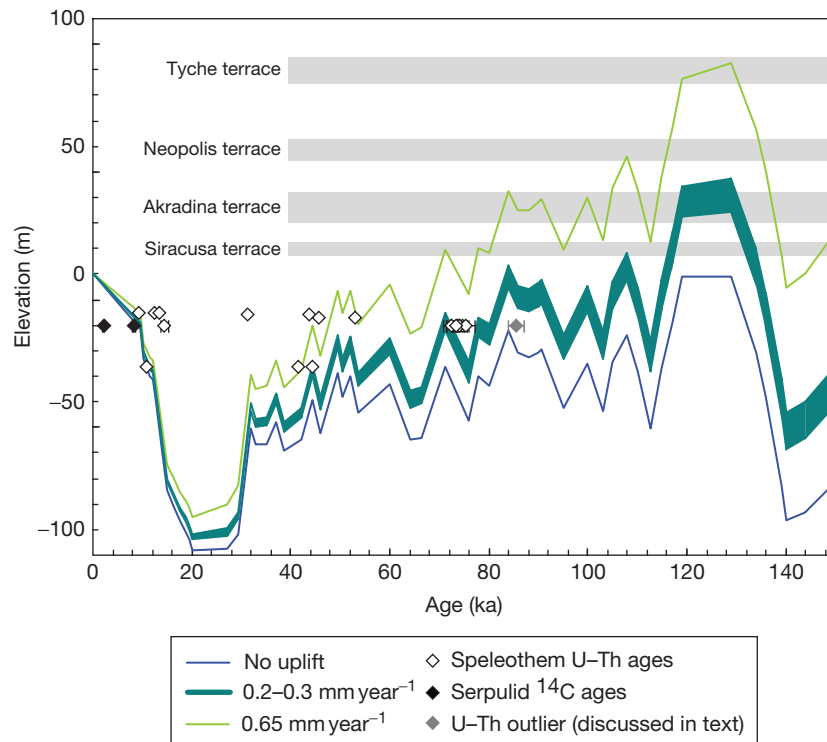


Figure 3 Submerged speleothem data from Sicily to constrain uplift history. U-series and ^{14}C data are plotted with several uplift scenarios for the Siracusa coastline. In many cases, error bars are not visible because the errors are smaller than the size of the symbol. Open symbols represent U-series ages of speleothem calcite growth and hence should sit above the sea-level curve. Solid symbols are ^{14}C ages of serpulid calcite and should be below the sea-level curve as they represent time periods of cave submergence. Scenarios where the open symbols sit below the sea-level curve are not plausible uplift histories. Uplift rates between 0.20 and 0.40 mm a^{-1} appear to be in best agreement with the speleothem data. Reproduced from Dutton A, Scicchitano G, Monaco C, et al. (2009) Uplift rates defined by U-series and ^{14}C ages of serpulid-encrusted speleothems from submerged caves near Siracusa, Sicily (Italy). *Quaternary Geochronology* 4: 2–10, with permission.

Phreatic Overgrowths: Tracking Sea-Level Position

A more unusual, but extremely informative type of cave deposit that has been used to establish the timing and magnitude of sea-level changes is phreatic overgrowths on speleothems, sometimes referred to as POS. These have been described for caves in Mallorca, where encrustations of calcite or aragonite coat the surface of speleothems and cave walls at the air–water interface (Pomar et al., 1979). The overgrowths occur within the range of tides and are thickest at the mean position of sea level (Figure 4). The core speleothem (stalagmite or stalactite) will have an older age compared to the encrustation, which can be dated itself to establish the timing of sea level still stands that appear to be recorded across a wide range of elevations within these cave systems, from +46 to –23 m above and below mean sea level (Gines et al., 2003). Slight variations in elevations of cogenetic phreatic overgrowths along the laterally extensive cave system of the southeast coastline of Mallorca have also been used to establish the rate of tectonic tilting in the region (Fornós et al., 2002).

Up to 30 paleo levels of phreatic speleothems have been recognized in Mallorca (Gines et al., 2003). This archive clearly has the potential to deliver a detailed history of sea-level positions over recent glacial–interglacial cycles. Mass-spectrometric U-series data for these phreatic overgrowths has largely focused on those formed during the

entirety of MIS 5 (Dorale et al., 2010; Gines et al., 2003; Tuccimei et al., 2006), including both the sea-level highstands as well as the lowstands. Interestingly, the elevations of MIS 5a, 5c, and 5e all appear to be similar (+1 to +3 m above present mean sea level) (Gines et al., 2003). The significance of these elevations in terms of global grounded ice volume is not entirely clear in the absence of glacio-hydro-isostatic modeling to understand the relation between relative sea level at Mallorca and the eustatic sea-level history. Even if Mallorca may be currently undergoing only slow vertical movement in response to ongoing relaxation of the Earth's surface to the Last Glacial Maximum (Dorale et al., 2010), because the elevations of these former sea-level highstands are also a function of the ice history preceding the highstand, the difference between present sea level and the elevation of a past sea-level highstand cannot be directly equated to ice volume (Lambeck et al., 2012). This is a generalization that holds true for other sites outside of Mallorca as well, and is important in the context of all of the cave data discussed here.

Additional Considerations

In all cave environments there may be other factors than sea level that can affect the timing of hiatuses in speleothems.

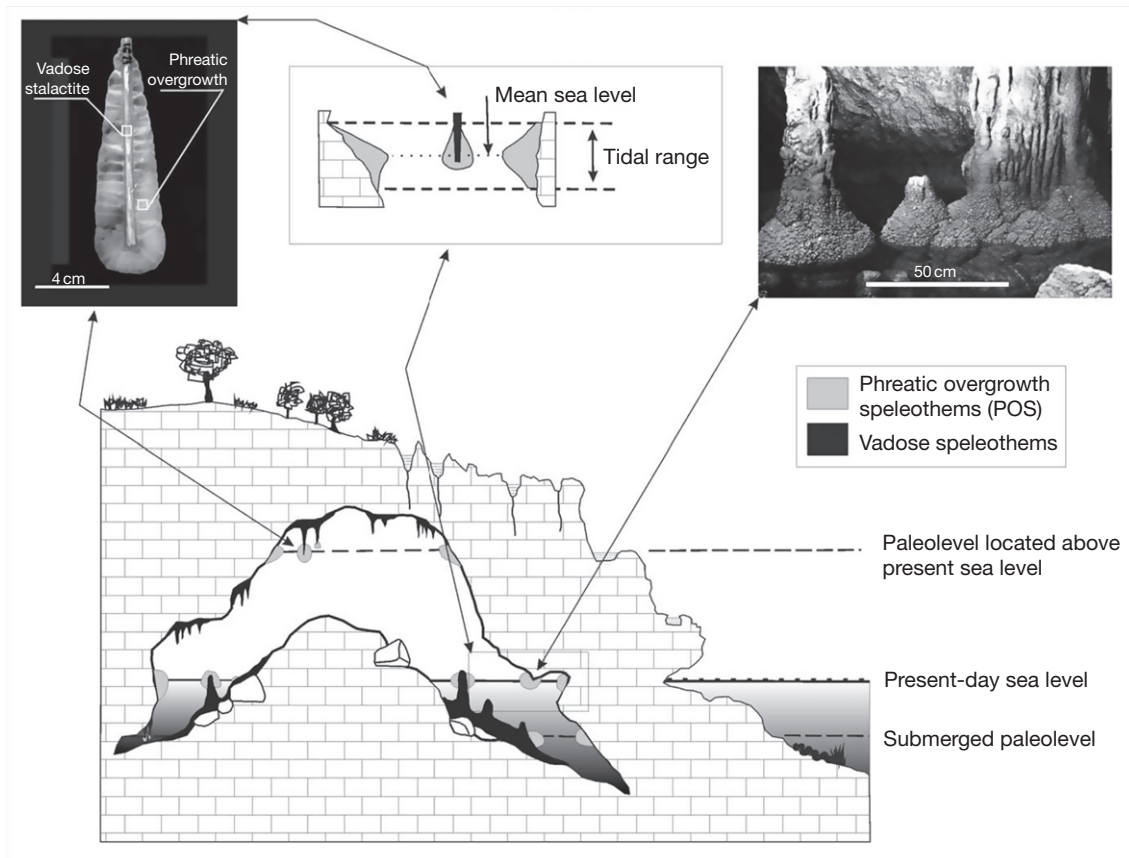


Figure 4 Schematic diagram of littoral karst in Mallorca outlined in an ideal cross-section. Note presence of phreatic overgrowths related to past and present sea levels, as well as the detailed morphology due to the tidal and atmospheric oscillation of the coastal water table. Reproduced from Fornós JJ, Gelabert B, Gines A, Gines J, Tuccimei P, and Vesica P (2002) Phreatic overgrowths on speleothems: A useful tool in structural geology in littoral karstic landscapes. The example of eastern Mallorca (Balearic Islands). *Geodinamica Acta* 15: 113–125, with permission.

Hiatuses can develop in subaerial caves far from the coasts due to changes in precipitation, groundwater recharge, and cave hydrology. For this reason, it is important to establish a line of evidence that indicates that a hiatus was formed due to the incursion of seawater; for example, trace amounts of seawater evaporates (Suric et al., 2009), serpulid crusts (Antonioli et al., 2004; Bard et al., 2002; Dutton et al., 2009a), or Fe-rich flocculates (Li et al., 1989) have all been interpreted to indicate the presence of seawater in the cave.

To serve as an accurate recorder of sea-level position, the cave must be situated along the coastline, for example, a flank margin cave (Myroie and Carew, 2006), because cave systems that extend inland may also be flooded, but the elevations of the fresh water lens and the seawater/freshwater interface will be a function of distance from the coastline and will not be at the same elevation as sea level along the coast. Therefore, for the purposes of reconstructing sea-level position, it is important to focus on caves that were immediately adjacent to the coastline during the time period of interest.

Finally, there remains the issue of how best to interpret radiometric ages and associated elevations of the various cave deposits discussed here given that each site will have a relative sea-level history that may be different from the eustatic sea-level curve (Farrell and Clark, 1976). Particularly in the Bahamas, Bermuda, and the Mediterranean, where many of these studies have been conducted, nearby ice sheets as well as water loading effects are expected to cause significant departures of relative sea level from the eustatic signal (Lambeck and Purcell, 2005; Potter and Lambeck, 2003). There are essentially two effective approaches to handling this: (1) discuss the cave data in the context of a full glacio-hydro-isostatic model in cases where an ice model covering the time interval from before the period of interest up to the present; or (2) understand the limitations of the data and make appropriate interpretations without dismissing the unquantified gravitational and rotational effects related to glacio-hydro-isostasy. For example, if the first scenario is not possible, then one appropriate and meaningful comparison would be to compare the timing of speleothem growth with that of nearby coral growth to understand the timing of sea-level change, and possible time lags between the two archives. Also, simply quantifying the number of sea-level oscillations within a certain time span can provide important information (e.g., Suric et al., 2009). If the age data in question falls during a deglacial event where rates of sea-level change are fast, on the order of ~ 10 m/1000 years, then it is possible that any uncertainty in elevation due to isostatic effects is already accounted for in the analytical error of the age (e.g., Dutton et al., 2009a), particularly in the case of older samples that may have larger error bars associated with measured U-series ages. In this case, the error envelope on the age spans a sufficient period of time during which sea level changed rapidly and will contribute a larger uncertainty than the isostatic correction to the relative sea-level data.

Conclusions

Radiometric dates of stalagmites, stalactites, and flowstones are a valuable tool to establish upper limits of sea-level position through time. Establishing the timing of speleothem growth helps to set bounds on the timing of cave emergence above sea

level. Dates bounding erosional unconformities in speleothems formed during seawater submergence provide further constraints on the timing of sea-level highstands and the elevation of the speleothem in the cave provides an important limit on the position of sea level at that time. Organisms that encrust or bore into cave deposits can be studied to provide a complementary age on the timing of cave submergence. In particular environments, it is also possible to capitalize on the occurrence of phreatic overgrowths that form at the sea surface to establish the timing and position of sea level. In this last case, instead of providing a bracketing age, that is an age that represents a minimum or maximum estimate on the timing of sea level passing through the elevation of the cave, the age of the phreatic overgrowth represents the exact time at which sea level was at that position. In all cases, these age constraints provide important benchmarks to test existing models and reconstructions of sea level through time.

Our understanding of the timing and magnitude of sea-level changes can be greatly enhanced by further study of both emergent and submerged caves. In particular, submerged caves have the potential to deliver rare information on the timing of sea-level lowstands, and the timing of sea-level changes during glacial periods in addition to providing bracketing ages for sea-level highstands. Existing data from submerged speleothems and associated cave deposits are scarce due to practical difficulties entailed in locating and recovering samples from underwater caves. In fact, this raises an ethical question of how this line of research should be pursued in light of the risk involved with cave diving to collect these valuable specimens. Perhaps there is scope to conduct future explorations or even sampling of submerged cave systems with remotely operated underwater vehicles.

Alternatively, one could also turn to uplifted areas to access speleothems in caves that were once submerged. Coral terraces exposed by rapid tectonic uplift have been exploited to access this portion of the record at localities such as Barbados and Huon Peninsula, Papua New Guinea (Chappell and Polach, 1991; Fairbanks, 1989). Unfortunately, there remain few examples of radiometrically dated speleothems from similarly uplifted environments to constrain former sea-level oscillations. Cave exploration in such environments certainly has the potential to yield an important and largely untapped resource for additional constraints on past sea-level change.

See also: U-Series Dating. **Sea Level Studies:** Eustatic Sea-Level Changes – Glacial–Interglacial Cycles; Eustatic Sea-Level Changes Since the Last Glacial Maximum; Isostasy: Glaciation-Induced Sea-Level Change; Overview.

References

- Aharon P and Chappell J (1986) Oxygen isotopes, sea level changes and the temperature history of a coral reef environment in New Guinea over the last 10^5 years. *Palaeogeography, Palaeoclimatology, Palaeoecology* 56: 337–379.
- Antonioli F, Bard E, Potter E-K, Silenzi S, and Imbrota S (2004) 215-ka History of sea-level oscillations from marine and continental layers in Argentario cave speleothems (Italy). *Global and Planetary Change* 43: 57–78.

- Antoninoli F and Oliverio M (1996) Holocene sea-level rise recorded by a radiocarbon-dated mussel in a submerged speleothem beneath the Mediterranean Sea. *Quaternary Research* 45: 241–244.
- Antoninoli F, Silenzi S, and Frisia S (2001) Tyrrhenian Holocene palaeoclimate trends from spelean serpulids. *Quaternary Science Reviews* 20: 1661–1670.
- Bard E, Antoninoli F, and Silenzi S (2002) Sea-level during the penultimate interglacial period based on a submerged stalagmite from Argentarola Cave (Italy). *Earth and Planetary Science Letters* 196: 135–146.
- Bard E, Hamelin B, and Fairbanks RG (1990) U–Th ages obtained by mass spectrometry in corals from Barbados: Sea level during the past 130,000 years. *Nature* 346: 456–458.
- Bintanja R, Van De Wal RSW, and Oerlemans J (2005) Modelled atmospheric temperatures and global sea levels over the past million years. *Nature* 437: 125–128.
- Chappell J and Polach HA (1991) Post-glacial sea-level rise from a coral record at Huon Peninsula, Papua New Guinea. *Nature* 349: 147–149.
- Chappell J and Shackleton NJ (1986) Oxygen isotopes and sea level. *Nature* 324: 137–140.
- Dorale JA, Onac BP, Fornos JJ, et al. (2010) Sea-level highstand 81,000 years ago in Mallorca. *Science* 327: 860–863.
- Dutton A, Bard E, Antoninoli F, Esat TM, Lambeck K, and McCulloch MT (2009a) Phasing and amplitude of sea-level and climate change during the penultimate interglacial. *Nature Geoscience* 2: 355–359.
- Dutton A, Scicchitano G, Monaco C, et al. (2009b) Uplift rates defined by U-series and ^{14}C ages of serpulid-encrusted speleothems from submerged caves near Siracusa, Sicily (Italy). *Quaternary Geochronology* 4: 2–10.
- Edwards RL, Chen JH, Ku T-L, and Wasserburg GJ (1987) Precise timing of the last interglacial period from mass spectrometric determination of thorium-230 in corals. *Science* 236: 1547–1553.
- Edwards RL, Gallup CD, and Cheng H (2003) Uranium-series dating of marine and lacustrine carbonates. In: Bourdon B, Henderson GM, Lundstrom CC, and Turner SP (eds.) *Uranium-series Geochemistry*, vol. 52, pp. 363–406. Washington, DC: Mineralogical Society of America.
- Emiliani C (1955) Pleistocene temperatures. *Journal of Geology* 63: 538–578.
- Fairbanks RG (1989) A 17,000-year glacio-eustatic sea level record: Influence of glacial melting rates on the Younger Dryas event and deep-ocean circulation. *Nature* 342: 637–641.
- Farrell WE and Clark JA (1976) On postglacial sea-level. *Geophysical Journal of the Royal Astronomical Society* 46: 647–667.
- Fornós JJ, Gelabert B, Gines A, Gines J, Tuccimei P, and Vesica P (2002) Phreatic overgrowths on speleothems: A useful tool in structural geology in littoral karstic landscapes. The example of eastern Mallorca (Balearic Islands). *Geodinamica Acta* 15: 113–125.
- Gallup CD, Edwards RL, and Johnson RG (1994) The timing of high sea levels over the past 200,000 years. *Science* 263: 796–800.
- Gascoyne M (1984) Uranium series ages of speleothems from Bahaman Blue Holes and their significance. *Transactions of the British Cave Research Association* 11: 45–49.
- Gascoyne M, Benjamin GJ, Schwarcz HP, and Ford DC (1979) Sea-level lowering during the Illinoian glaciation: Evidence from Bahama “Blue Hole”. *Science* 205: 806–808.
- Gines A, Tuccimei P, Fornos JJ, Gines A, Gracia F, and Vesica P (2003) The upper Pleistocene sea-level history in Mallorca (western Mediterranean) approached from the perspective of coastal phreatic speleothems. In: Ruiz MB, Dorado M, and Valdeolmillos A, et al. (eds.) *Quaternary Climatic Changes and Environmental Crises in the Mediterranean Region*. Madrid: Alcalá de Henares.
- Harmon RS, Land LS, Mitterer RM, Garrett P, Schwarcz HP, and Larson GJ (1981) Bermuda sea levels during the last interglacial. *Nature* 289: 481–483.
- Harmon RS, Mitterer RM, Kriausakul N, et al. (1983) U-series and amino-acid racemization geochronology of Bermuda: Implications for eustatic sea-level fluctuation over the past 250,000 years. *Palaeogeography, Palaeoclimatology, Palaeoecology* 44: 41–70.
- Labeyrie LD, Duplessy JC, and Blanc PL (1987) Variations in mode of formation and temperature of oceanic deep waters over the past 125,000 years. *Nature* 327: 477–482.
- Lambeck K and Chappell J (2001) Sea level change through the last glacial cycle. *Science* 292: 679–686.
- Lambeck K and Purcell T (2005) Sea-level change in the Mediterranean Sea since the LGM: Model predictions for tectonically stable areas. *Quaternary Science Reviews* 24: 1969–1988.
- Lambeck K, Purcell A, and Dutton A (2012) The anatomy of interglacial sea levels: The relationship between sea levels and ice volumes during past interglacials. *Earth and Planetary Science Letters* 315–316: 4–11.
- Lea DW, Martin PA, Pak DK, and Spero HJ (2002) Reconstructing a 350 ky history of sea level using planktonic Mg/Ca and oxygen isotope records from a Cocos Ridge core. *Quaternary Science Reviews* 21: 283–293.
- Li W-X, Lundberg J, Dickin AP, et al. (1989) High-precision mass-spectrometric uranium-series dating of cave deposits and implications for palaeoclimate studies. *Nature* 339: 534–536.
- Lisiecki LE and Raymo ME (2005) A Pliocene–Pleistocene stack of 57 globally distributed benthic $\delta^{18}\text{O}$ records. *Paleoceanography* 20: PA1003.
- Lundberg J and Ford DC (1994) Late Pleistocene sea level change in the Bahamas from mass spectrometric U-series dating of submerged speleothem. *Quaternary Science Reviews* 13: 1–14.
- Moucha R, Forte AM, Mitrovica JX, et al. (2008) Dynamic topography and long-term sea-level variations: There is no such thing as a stable continental platform. *Earth and Planetary Science Letters* 271: 101–108.
- Mylroie JE and Carew JL (2006) The flank margin model for dissolution cave development in carbonate platforms. *Earth Surface Processes and Landforms* 15: 413–424.
- Pomar L, Gines A, and Fines J (1979) Morfología, estructura y origen de los espeleotemas epiaquático. *Endins* 5–6: 3–17.
- Potter E-K, Esat TM, Schellmann G, Radtke U, Lambeck K, and McCulloch MT (2004) Suborbital-period sea-level oscillations during marine isotope substages 5a and 5c. *Earth and Planetary Science Letters* 225: 191–204.
- Potter E-K and Lambeck K (2003) Reconciliation of sea-level observations in the Western North Atlantic during the last glacial cycle. *Earth and Planetary Science Letters* 217: 171–181.
- Richards DA and Dorale JA (2003) U-series chronology and environmental applications of speleothems. In: Bourdon B, Henderson GM, Lundstrom CC, and Turner SP (eds.) *Uranium-series Geochemistry*, vol. 52, pp. 407–460. Washington, DC: Mineralogical Society of America.
- Richards DA, Smart PL, and Edwards RL (1994) Maximum sea levels for the last glacial period from U-series ages of submerged speleothems. *Nature* 367: 357–360.
- Schellmann G and Radtke U (2004) *The Marine Quaternary of Barbados*. Germany: Geographisches Institut der Universität zu Köln.
- Siddall M, Rohling EJ, Almogi-Labin A, et al. (2003) Sea-level fluctuations during the last glacial cycle. *Nature* 423: 853–858.
- Smart PL, Richards DA, and Edwards RL (1998) Uranium-series ages of speleothems from South Andros, Bahamas: Implications for Quaternary sea-level history and palaeoclimate. *Cave and Karst Science* 25: 67–74.
- Spalding RF and Mathews TD (1972) Stalabmites from caves in the Bahamas: Indicators of low sea level stand. *Quaternary Research* 2: 470–472.
- Stirling CH, Esat TM, Lambeck K, and McCulloch MT (1998) Timing and duration of the last interglacial: Evidence for a restricted interval of widespread coral reef growth. *Earth and Planetary Science Letters* 160: 745–762.
- Suric M, Richards DA, Hoffman DL, Tibljas D, and Juracic M (2009) Sea-level change during MIS 5a based on submerged speleothems from the eastern Adriatic Sea (Croatia). *Marine Geology* 262: 62–67.
- Tuccimei P, Gines J, Delitala MC, Gines A, Gracia F, and Fornos J (2006) Last interglacial sea level changes in Mallorca island (Western Mediterranean). High precision U-series data from phreatic overgrowths on speleothems. *Zeitschrift für Geomorphologie* 50: 1–21.
- Waelbroeck C, Labeyrie L, Michel E, et al. (2002) Sea-level and deep water temperature changes derived from benthic foraminifera isotopic records. *Quaternary Science Reviews* 21: 295–305.

SEA-LEVELS, LATE QUATERNARY

Contents

Late Quaternary Relative Sea-Level Changes in High Latitudes

Late Quaternary Sea-Level Changes in Greenland

Late Quaternary Relative Sea-Level Changes at Mid-Latitudes

Late Quaternary Relative Sea-Level Changes in the Tropics

Tectonics and Relative Sea-Level Change

Late Quaternary Relative Sea-Level Changes in High Latitudes

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Introduction

The distribution of glaciers and ice sheets on the Earth's surface today is overwhelmingly concentrated in the high latitudes. During the Late Quaternary, however, ice cover in these areas was even more extensive and ice sheets expanded across continental shelves, depressing the Earth's crust and delivering sediment to the deep ocean basins (Clark and Mix, 2002). Relative sea-level (RSL) histories from these high-latitude regions preserve a record of the nature and pattern of crustal rebound following ice-sheet decay and can be used to construct isobases from which the location of former ice-sheet loading centers can be determined. The aim of this encyclopedia entry is to outline the controls on, and nature of, Late Quaternary RSL change in the high latitudes. These concepts are illustrated by case studies of Late Quaternary RSL change: the first from the Queen Elizabeth Islands (QEI) of the Canadian High Arctic and the second from Antarctica, which discusses a series of RSL records from around the continent. The focus of this entry is predominantly on observational studies of near-field RSL histories; far-field components (cf. Bassett et al., 2005) are not dealt with in detail here.

RSL Change in High Latitudes: Controls and Patterns

The RSL history of a given site is influenced by a combination of eustatic, isostatic, hydroisostatic and tectonic influences. The dominance of eustasy versus isostasy depends on the location of the site with respect to the former ice-sheet loading center. Within former ice-sheet limits ('near-field sites'), RSL history is dominated by the isostatic component and records crustal rebound following ice-sheet decay. Far from the ice-sheet margin ('far-field sites'), the RSL history is dominated by eustatic factors.

Clark et al. (1978) present a classic model which divides the globe into a series of zones based on the characteristics of their RSL curves (Figure 1) (note that later models and data show variants on this). The high latitudes are characterized by glacioisostatically dominated, Late Quaternary RSL histories due to the former presence of ice sheets in these regions. Zone I comprises those areas within the former ice-sheet margin that have experienced continuous, exponential RSL fall since deglaciation, with the magnitude of emergence reflecting the proximity to the ice-sheet loading center. Zone II is characterized by continuous submergence due to the presence of the collapsing proglacial forebulge which, following deglaciation, migrates toward the former ice-sheet center with time. The forebulge is formed of sublithospheric material extruded from beneath the growing ice sheet as it isostatically depresses the crust. During deglaciation, the ice sheet retreats and the resulting isostatic disequilibrium causes this material to flow back toward the former ice-sheet center. Regions near the former ice-sheet margin experience RSL changes that are transitional between Zones I and II. At the ice-sheet margin there is initial emergence followed by later submergence as the forebulge migrates (Figure 2). The relative magnitude of submergence versus emergence depends on the position of a site with respect to the former loading center; greater submergence occurs with increasing distance from the ice sheet and submergence commences at a later time for sites close to the ice sheet than for sites at a greater distance.

In the case of the last North American (Laurentide) ice sheet, the time-transgressive nature of its retreat meant that the forebulge migrated back through Atlantic Canada following the retreating ice edge (Quinlan and Beaumont, 1981). Quinlan and Beaumont (1981) have shown the effect of forebulge migration on RSL history (Figure 2). Sites A, B and C are all on the leading edge of the forebulge. These sites all experience emergence/RSL fall as the bulge crest approaches, and submergence/RSL rise as the bulge crest passes. Site D is

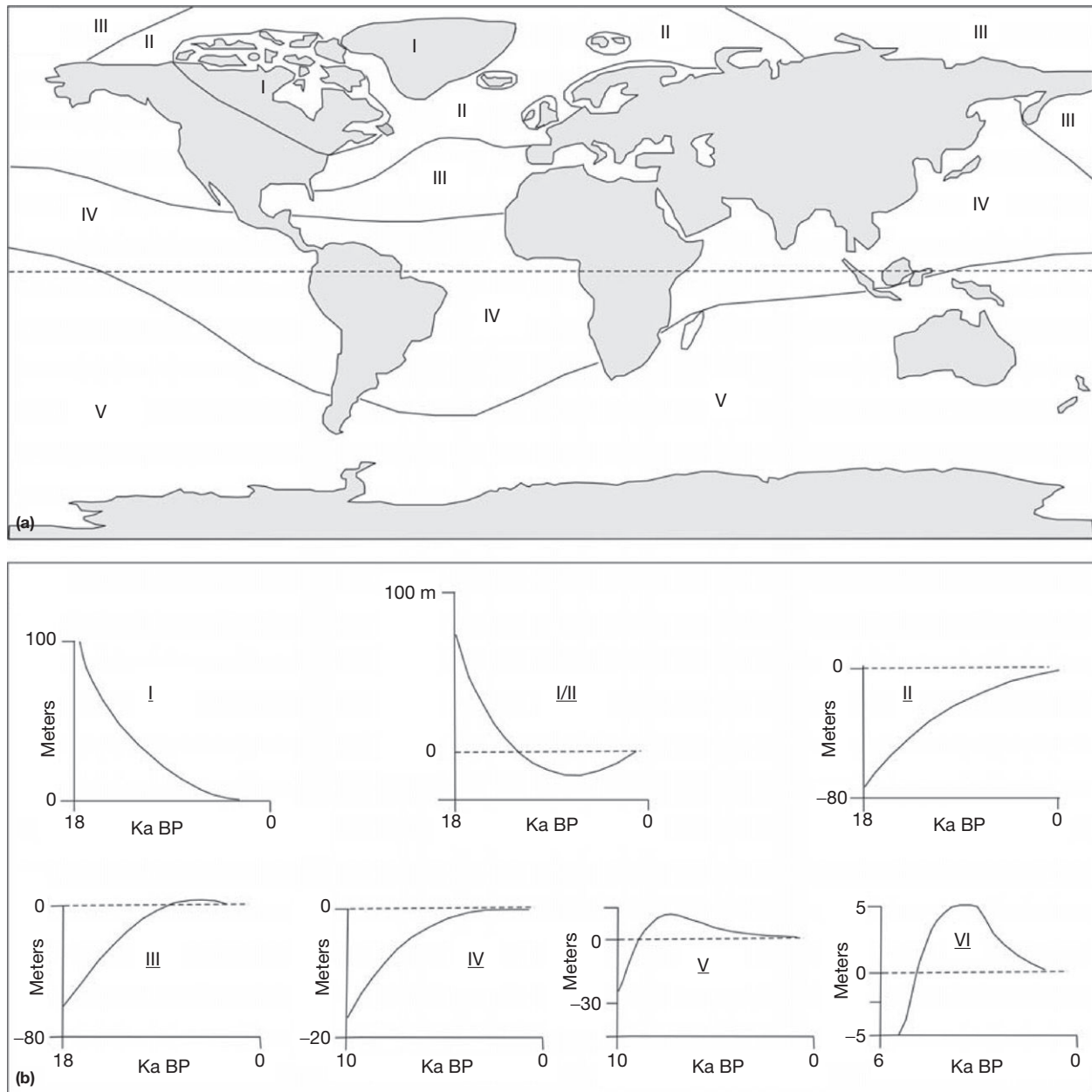


Figure 1 (a) Distribution of sea-level zones and (b) RSL curves typical of each zone. Modified from Clark JA, Farrell WE, and Peltier WR (1978) Global changes in postglacial sea level: A numerical calculation. *Quaternary Research* 9: 265–287.

located beyond the maximum position of the crest and experiences only subsidence/RSL rise. Thus RSL falls continuously at site A with the bulge crest yet to pass this point. At sites B and C there is an initial period of RSL fall as the crest approaches, followed by RSL rise as the crest passes. However, the duration of RSL rise and fall is different for these sites. At site D there is only RSL rise. Sites A and D correspond to Zones I and II of Clark et al. (1978) while sites C and D correspond to the transitional Zone I/II.

The maximum elevation attained by the sea along a glacioisostatically depressed coastline is referred to as the 'marine limit.' Following the seminal pioneering work on the postglacial RSL history of the Canadian Arctic by Andrews

(1970), the elevation of marine limit and form of the subsequent RSL curve are recognized as being a function of distance from the former ice-sheet margin (which is an indication of ice thickness over the site), date of deglaciation and eustatic sea-level rise. In areas that were formerly glacier covered, marine limit is formed time transgressively as the ice sheet retreats. In the Arctic and Antarctica, marine limit is generally defined by a range of criteria: (1) the elevation of the highest raised beach or delta terrace; (2) the lowest elevation of undisturbed till, delicately poised erratic boulders, or felsenmeer (washing limits); and (3) the highest elevation at which marine shells or the remains of marine mammals can be found. These criteria generally provide a minimum estimate for marine limit

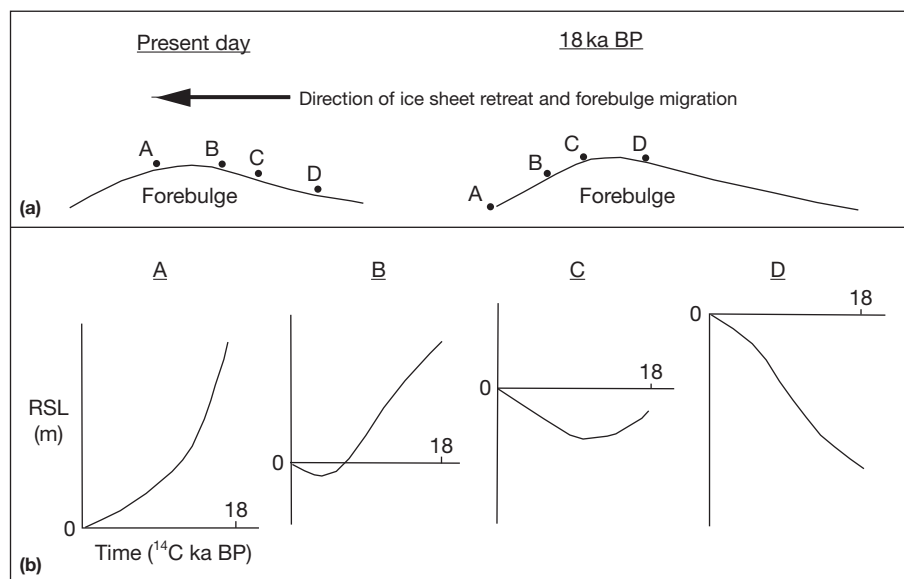


Figure 2 (a) Schematic representation of a peripheral forebulge at two points in time, 18 ka BP and the present day. The forebulge migrates in the direction of ice-sheet retreat thereby affecting RSL at the four sites A, B, C, and D. (b) RSL curves from sites A–D recording the forebulge migration shown in panel A. Modified from [Quinlan G and Beaumont C \(1981\)](#) A comparison of observed and theoretical postglacial relative sea level in Atlantic Canada. *Canadian Journal of Earth Sciences* 18: 1146–1163. The vertical scale on the four RSL curves indicates the nature of the RSL change rather than the precise magnitude of this change, as the latter will differ depending upon a range of variables including ice-sheet thickness, deglacial history, and ice-sheet volume.

elevation. The age of marine limit (and thus the timing of initial deglaciation) is generally constrained through radiocarbon dating of a variety of associated organic material: marine shells in growth position within delta bottomset beds that can be related to the marine limit delta terrace; buried driftwood or whalebone in raised beach ridges; and surface shells on raised beaches or delta foreslopes (these give a minimum estimate on the age of marine limit as they could have been moved downslope subsequent to marine limit formation). In addition, dating of the transition from marine sediment to organic lacustrine material in cores recovered from isolation basins close to marine limit also provide chronological constraints on the formation of marine limit and the timing of deglaciation (e.g., [Bentley et al., 2005](#); [Long et al., 1999](#)).

Marine limit is not a synchronous shoreline and its elevation over an area does not represent the time-dependent form of glacioisostatic rebound. For this, measurement of synchronous shorelines must be made. Such shorelines slope up toward the ice-sheet center because they have been uplifted by differing amounts depending on their proximity to the former loading center. The tilt of such synchronous shorelines can be represented by isobases (contour lines joining points of equal uplift over an equal amount of time). Younger shorelines tend to be tilted less steeply than older shorelines due to the decrease in differential uplift over time.

Case Studies of Late Quaternary RSL Change in High Latitudes

QEI, Canadian High Arctic

The QEI of the Canadian High Arctic are located north of the Parry Channel ([Figure 3](#)) and comprise a geologically and

physiographically complex archipelago. The greatest relief occurs in the fretted mountains of northern and eastern Ellesmere Island, eastern Devon Island, and central Axel Heiberg Island. The smaller islands of the central and western parts of the archipelago are characterized by lowland and plain topography. Modern glacier cover is largely restricted to ice caps and outlet glaciers on Ellesmere and Devon islands with only small, scattered ice masses in the rest of the archipelago.

[Blake \(1970\)](#) first proposed the existence of the Innuitian Ice Sheet over the QEI during the last glaciation (marine isotope stage 2) based mainly on the pattern of differential postglacial rebound since 5 ka BP. He demonstrated that shorelines of this age were highest throughout a broad corridor oriented northeast–southwest extending from northern Eureka Sound to Bathurst Island. This ridge of high Holocene emergence was termed the Innuitian Uplift by [Walcott \(1972\)](#). Blake proposed that this emergence reflected the removal of a regional ice sheet which covered the QEI during the Last Glacial Maximum (LGM), whose axis of maximum thickness corresponded to the corridor of maximum uplift. The Innuitian Ice Sheet was coalescent with the Greenland Ice Sheet to the east and the Laurentide Ice Sheet to the south. In contrast, a markedly different reconstruction of the glacial history was proposed by [England \(1976, 1983\)](#) who argued that the last glaciation was very restricted and that outlet glaciers only advanced a few tens of kilometers beyond their present margins. This alternative interpretation was based on an apparent absence of primary glacial geological evidence for an Innuitian Ice Sheet, evidence that the only extensive glaciation in the QEI predated the LGM, and RSL evidence that the fjords and interisland channels of the eastern QEI were occupied by a full-glacial sea during the last glaciation. These contrasting reconstructions formed the end members in an ensuing debate over glacial

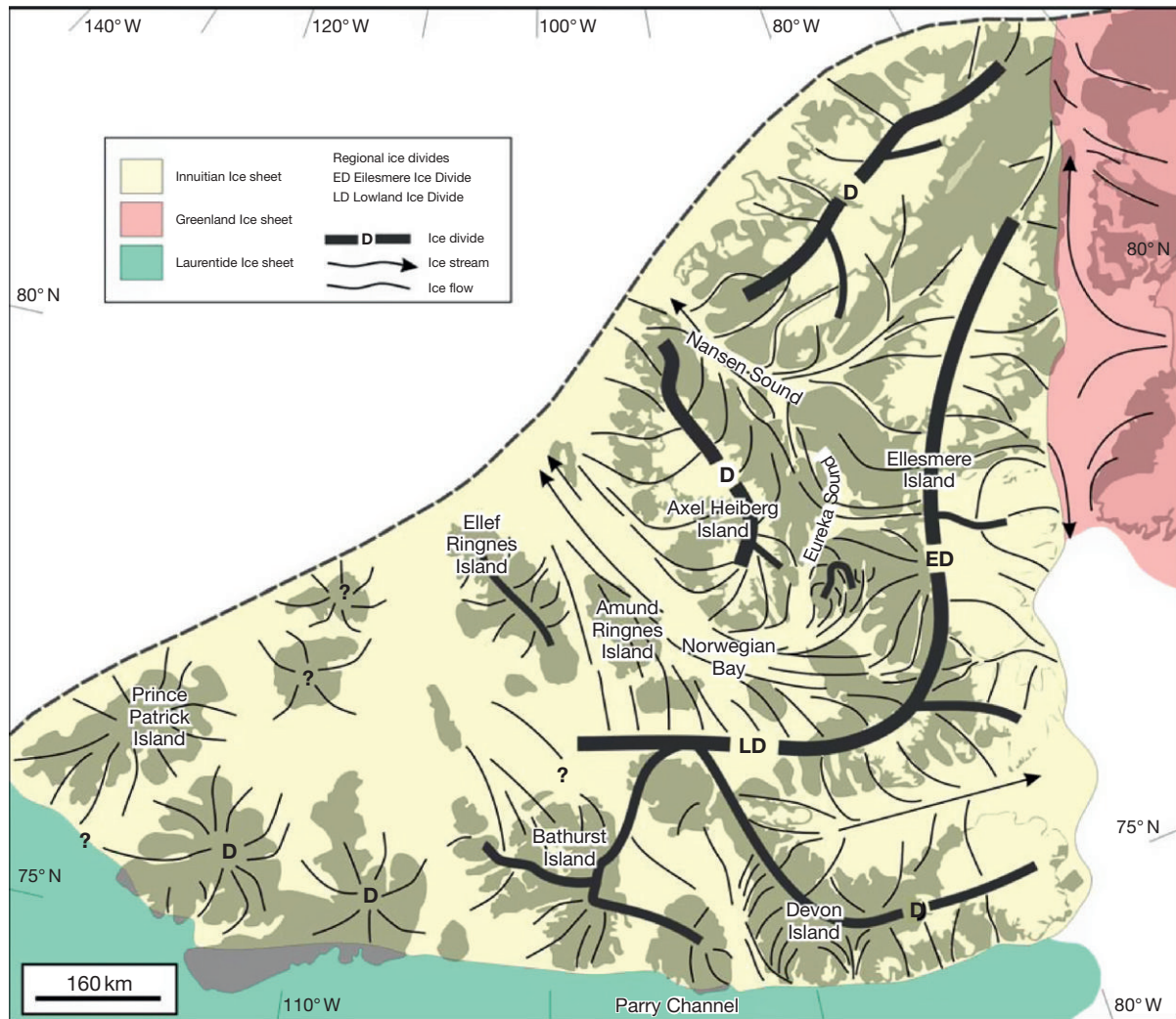


Figure 3 Schematic map of the paleogeography of the Innuitian Ice Sheet, Queen Elizabeth Islands, Canadian High Arctic, at the Last Glacial Maximum (modified from England J, Atkinson N, Bednarski J, Dyke AS, Hodgson DA, and Ó Cofaigh C (2006) The Innuitian Ice Sheet: Configuration, dynamics and chronology. *Quaternary Science Reviews* 25: 689–703). The Innuitian Ice Sheet was coalescent with the Greenland Ice Sheet to the east and the Laurentide Ice Sheet to the south. Primary ice divides in the alpine sector of the ice sheet (thick black lines) are termed the Ellesmere Divide (ED) and Lowland Divide (LD). Adjoining secondary divides (thinner black lines) extend from primary divides and generate areas of strong flow convergence that are recorded by glacial geological and geomorphological evidence.

extent during the LGM in the QEI which lasted over two decades. More recent investigations have rejected evidence for a full-glacial sea and the supposed antiquity of regional glaciation and have reconciled the pattern of postglacial emergence in the QEI with geological evidence for extensive ice cover at the LGM, thereby providing consensus for an Innuitian Ice Sheet (e.g., Atkinson, 2003; Atkinson and England, 2004; Dyke, 1998, 1999; England, 1998, 1999; England et al., 2006; Ó Cofaigh, 1999; Ó Cofaigh et al., 2000). We briefly summarize the salient features of the ice-sheet history here and detail the pattern and form of postglacial rebound associated with removal of the ice load upon deglaciation (see England et al., 2006).

Radiocarbon dates on subfossil organics and ice transported marine shells in till provide maximum ages for the build-up of the Innuitian Ice Sheet. They indicate that the ice sheet developed about 23–19 ka BP. During build-up, ice from eastern Ellesmere Island advanced into fjords and interisland

channels, forming trunk ice at least 1000 m thick. Ice flow from a divide situated over central Ellesmere Island flowed west into Eureka Sound (Figure 3) and coalesced with flow from a divide on Axel Heiberg Island, resulting in a 1200-m thick trunk ice in Eureka Sound, which debouched north into Nansen Sound and south into the Norwegian Bay. Ice thickness over the Ellesmere Island ice divide (Figure 3) did not exceed 200 m thicker than the present day thickness and thus the zone of thickest ice, and hence maximum ice-sheet loading, was situated over the intermontane basin of Eureka Sound rather than over the ice sheet divides. The lowland sector of the ice sheet occupied the central and possibly the western sector of the QEI. Ice flow from an east–west orientated divide in the Norwegian Bay fed ice southward as well as north to the continental shelf. Deglaciation was underway by at least 11.6 ka BP, and by 10 ka BP the ice sheet had started to break up with prominent marine embayments extending into the archipelago from the Arctic Ocean. By 8.5 ka BP, the sea had

penetrated the former ice-sheet center in Eureka Sound and the ice sheet had broken up into a series of residual island ice caps.

The pattern and form of Holocene RSL change in the QEI associated with rebound following deglaciation of the Innuitian Ice Sheet has been well documented through successive field investigations involving surveying and radiocarbon dating of raised marine landforms such as raised marine deltas, beaches, and washing limits (Figure 4). Radiocarbon dates on samples of *in situ* marine shells in growth position, buried driftwood, and whalebone (Figure 4(f)), which can be

associated stratigraphically with these former sea levels, provide the chronology of RSL change.

Because the Innuitian Ice Sheet had broken into a series of residual ice caps by about 8.5 ka BP, shorelines of this age are present across the QEI and have been widely surveyed and dated with the result that isobase maps showing the pattern of unloading associated with this shoreline have been completed for much of the QEI (e.g., Atkinson and England, 2004; Dyke, 1998; England, 1992; England et al., 2006; Ó Cofaigh, 1999). The most recent isobase compilation for

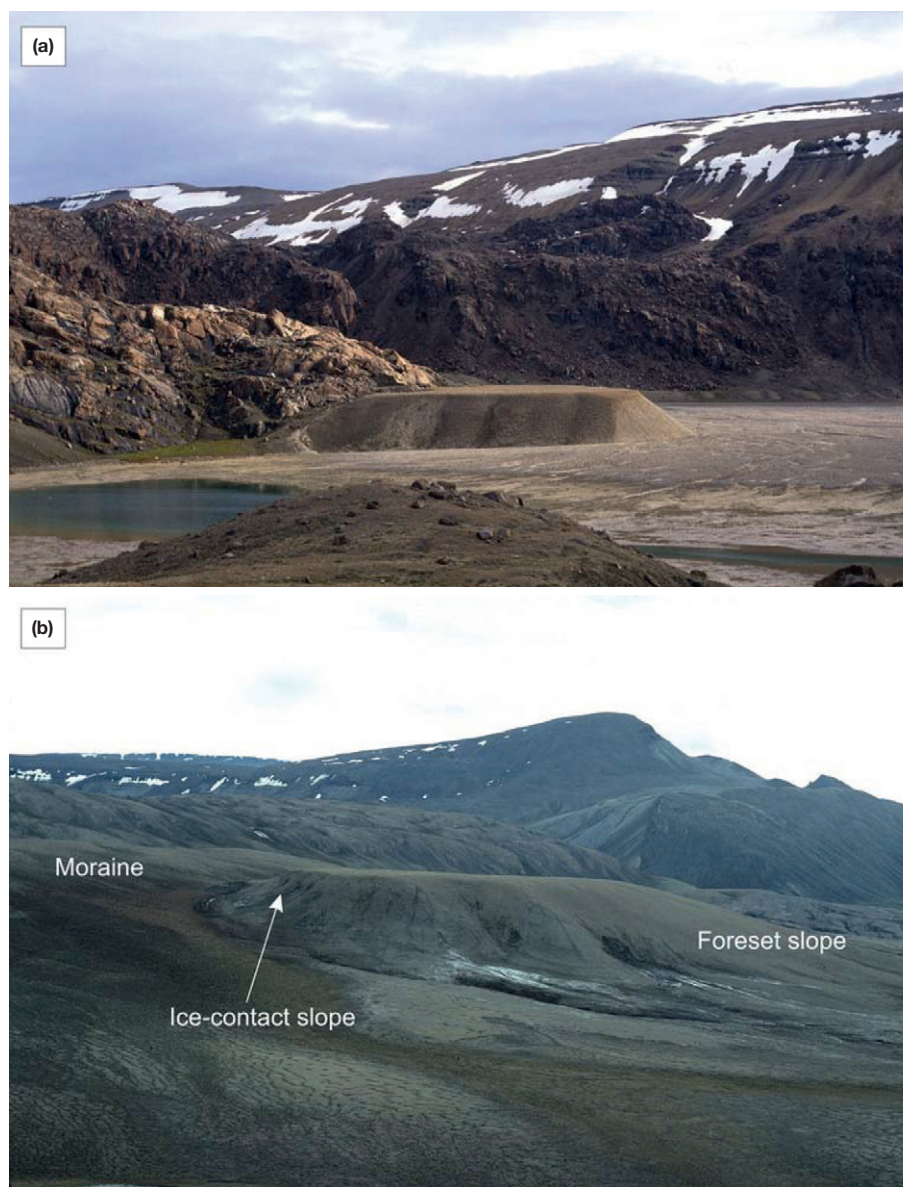


Figure 4 Examples of deglacial and raised marine landforms and sediments marking marine limit, Queen Elizabeth Islands. (a) Ice-contact glaciomarine delta abutting ice-molded bedrock knoll, Devon Island. (b) Ice-contact marine-limit delta (119 m asl) and recessional moraine, western Ellesmere Island. Note steep ice-contact slope and gentler delta foreset slope. (c) Raised beaches extending to marine limit at 116 m asl (arrowed) which trims ice-molded bedrock knolls, west-central Ellesmere Island. (d) Raised beaches extending to 138 m asl, southern Eureka Sound. (e) Prominent washing limit trimming ice-molded bedrock knoll and marking marine limit at 101 m asl, western Ellesmere Island. The area in the foreground contains thick sequences of deglacial raised marine silt. (f) Art Dyke (Geological Survey, Canada) excavating bowhead whale skull from raised beach gravels, southern Ellesmere Island. Radiocarbon dates on organic material such as whalebone and marine bivalves from raised marine sediments provide a chronology for RSL change and the timing of deglaciation.

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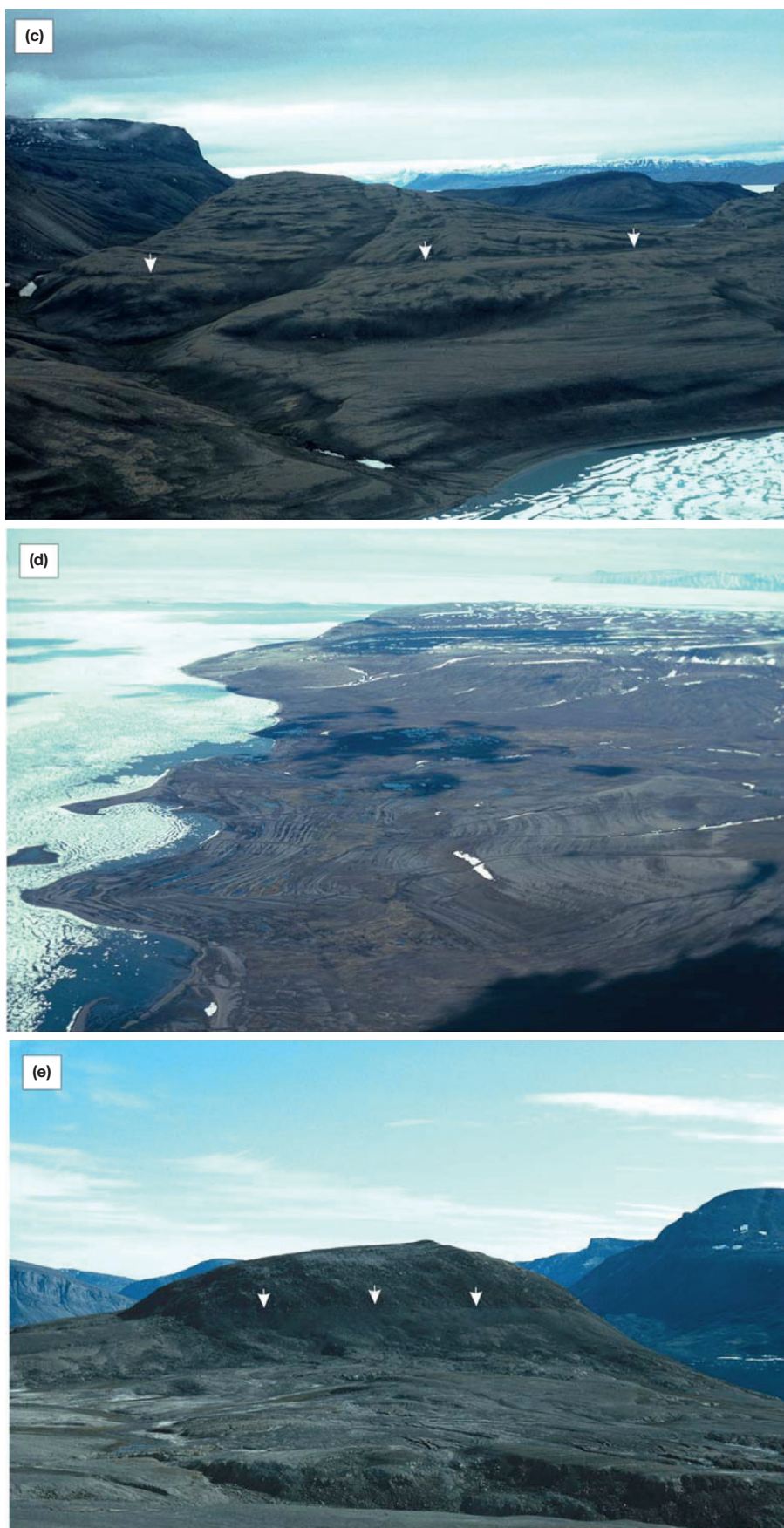


Figure 4 (Continued)



Figure 4 (Continued)

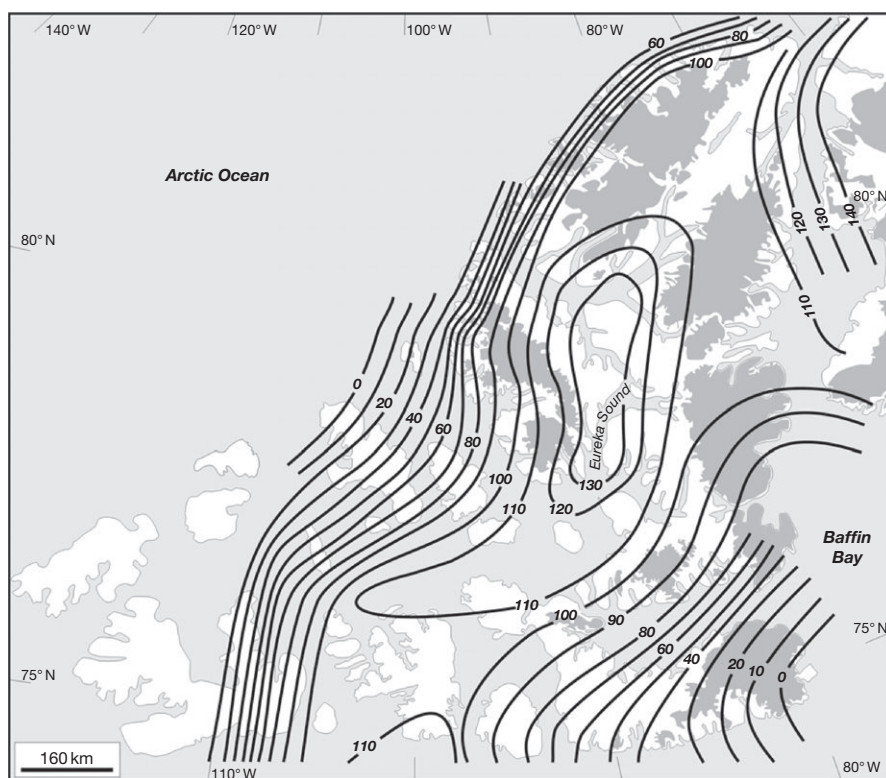


Figure 5 Postglacial isobases drawn on the 8.5 ka BP shoreline in the Queen Elizabeth Islands, Canadian Arctic (modified from England J, Atkinson N, Bednarski J, Dyke AS, Hodgson DA, and Ó Cofaigh C (2006) *The Innuitian Ice Sheet: Configuration, dynamics and chronology. Quaternary Science Reviews* 25: 689–703). The isobase pattern records delevelling associated with removal of the Innuitian Ice Sheet following the Last Glacial Maximum. Note the ridge of maximum postglacial emergence (>130 m asl) which encloses Eureka Sound, the location of former maximum ice-sheet thickness (>1 km). The isobases decline southwestward from Eureka Sound and enclose a prominent northeast–southwest orientated ridge (>110 m asl) in the central archipelago. This ridge marks the location of former maximum ice-sheet thickness in the lowland sector of the Innuitian Ice Sheet.

the 8.5 ka BP shoreline is shown in Figure 5. The isobases form a closed cell of maximum emergence (≥ 130 m asl) over Eureka Sound and the axis of the Innuitian Uplift can be seen to extend southwestward into the Norwegian Bay and then

westward (Figure 5). This demonstrates that the zone of maximum ice-sheet loading at the LGM was centered over the intermontane basin of Eureka Sound and corresponds to the zone of maximum ice thickness (~ 1200 m) rather than

its maximum surface elevation, which occurred over the encircling ice divides on Ellesmere and Axel Heiberg islands. In the lower relief terrain of the central QEI, the zone of maximum uplift corresponds to the former lowland ice divide (Dyke, 1999; England et al., 2006). Isobases on the 8.5 ka shoreline descend both northeast and southwest toward the Arctic Ocean and Baffin Bay respectively and record decreasing ice-sheet thickness toward the former margins. Furthermore, the lack of strong deflection in the isobase pattern across the marine channels of the western part of the QEI suggests that ice-sheet thickness was relatively uniform on, and between, neighboring islands. The pattern of the Innuitian Uplift is asymmetric, with a steeper gradient across the western islands than across those which face Baffin Bay (Figure 5). This may reflect a combination of earlier and more extensive deglaciation of the northwestern QEI, or the proximity of the

southeast QEI to Baffin Bay, which may have facilitated rates of snow accumulation and the generation and maintenance of greater ice thickness (England et al., 2006).

The highest Holocene marine limits in the QEI are found along the coastline of Eureka Sound, where shorelines at 158 m asl are present at the northern end of the channel. Marine limit falls southward down the channel but is still 140 m asl at its southern end. These high marine limits are consistent with the zone of maximum unloading along Eureka Sound as recorded by the pattern of shoreline delevelling drawn on the 8.5 ka BP shoreline (Figure 5; see above). In the eastern and central QEI, in those areas of the archipelago formerly covered by the ice sheet, RSL curves exhibit continuous exponential Holocene RSL fall from marine limit to the present (Blake, 1975; Dyke, 1998; Ó Cofaigh, 1999; Figure 6). The emergence history is thus similar to that of other areas of

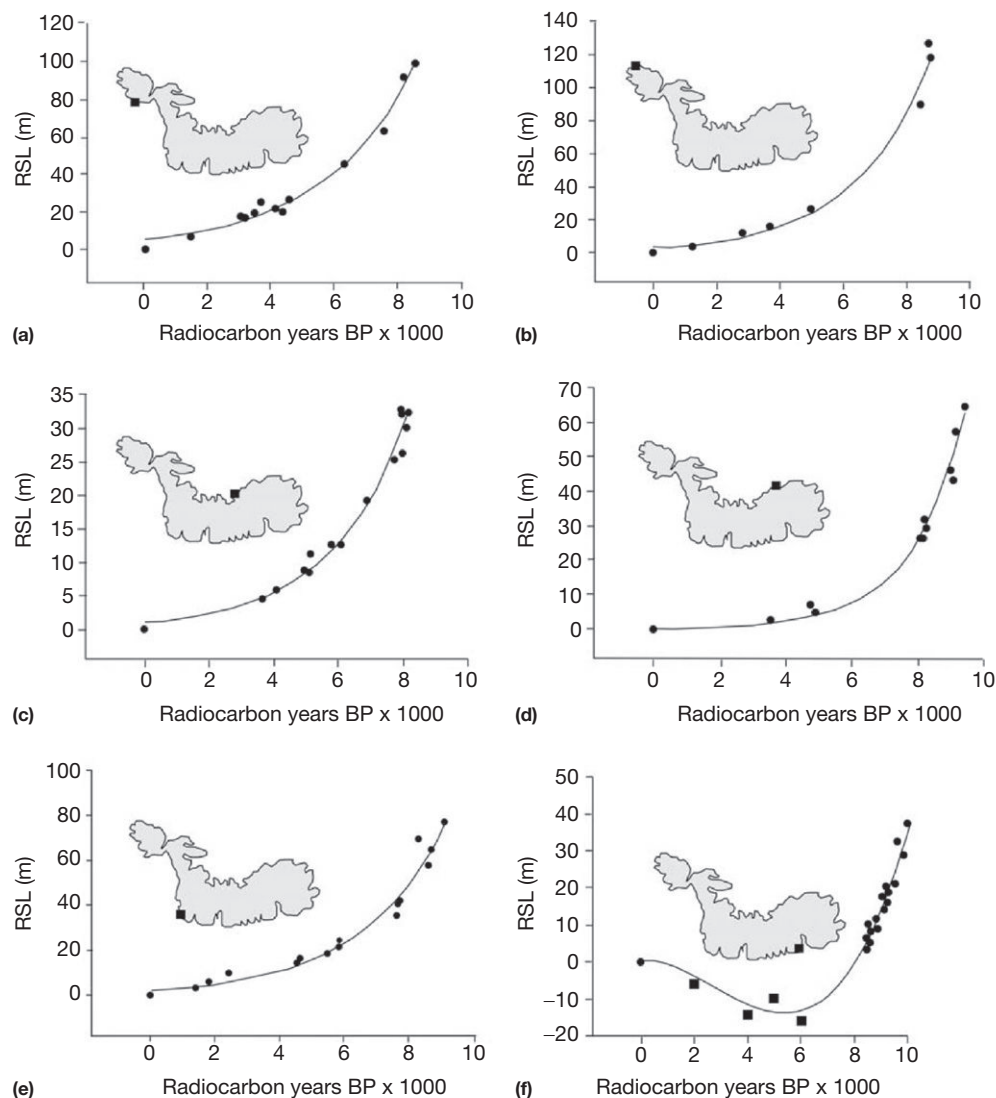


Figure 6 RSL curves for sites on Devon Island that were deglaciated by the Late Quaternary Innuitian Ice Sheet. Sites (a–e) show continuous exponential emergence and RSL fall since deglaciation and the formation of marine limit. Site (f) was located closer to the former margin of the Innuitian Ice Sheet and shows initial emergence and RSL fall followed by a period of submergence and RSL rise to the present day. Modified from Dyke AS (1998) Holocene delevelling of Devon Island, Arctic Canada: Implications for ice sheet geometry and crustal response. *Canadian Journal of Earth Sciences* 35: 885–904.

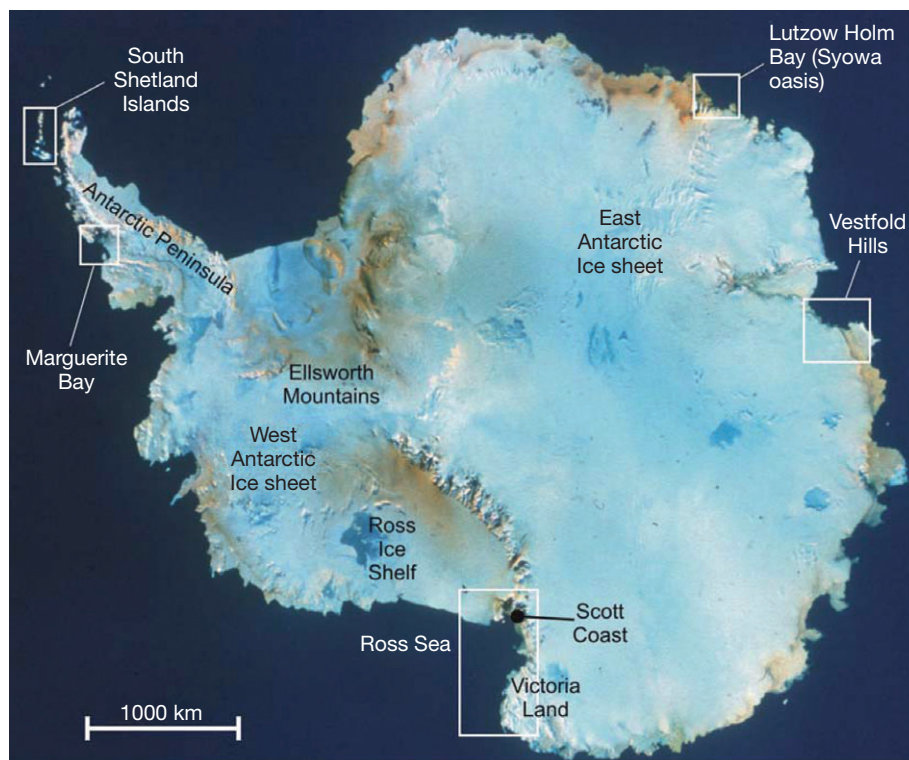


Figure 7 Main areas of RSL data collection in Antarctica – Ross Sea sector, East Antarctic coastal ice-free areas (notably the Vestfold Hills) and the Antarctic Peninsula. There are very few data for other parts of the Antarctic continent, although RSL studies have been carried out on a number of the sub-Antarctic islands.



Figure 8 Examples of raised marine sediments recording former high relative sea levels, Marguerite Bay, Antarctica. (a) Raised beaches on Lagoon Island, Marguerite Bay, Antarctica. Particularly well-developed beaches composed predominantly of cobble-sized material can be seen behind the hut. (b) Flight of raised beaches extending up to 40 m asl adjacent to a small outlet glacier, Horseshoe Island, Marguerite Bay, Antarctica. Note two figures for scale. (Photo: D. Hodgson). (c) Excavation of penguin remains from within sediment, Lagoon Island, Marguerite Bay, Antarctica. The penguin remains from the base of this pit have been radiocarbon dated to over 3000 ^{14}C years BP. (d) Penguin remains (bones, feather quills, guano-rich sediment) on surface of raised beach, Lagoon Island, Marguerite Bay, Antarctica. Material such as this has been excavated from depth in similar beaches (**Figure 8(c)**) and used to constrain the altitude of former sea level.

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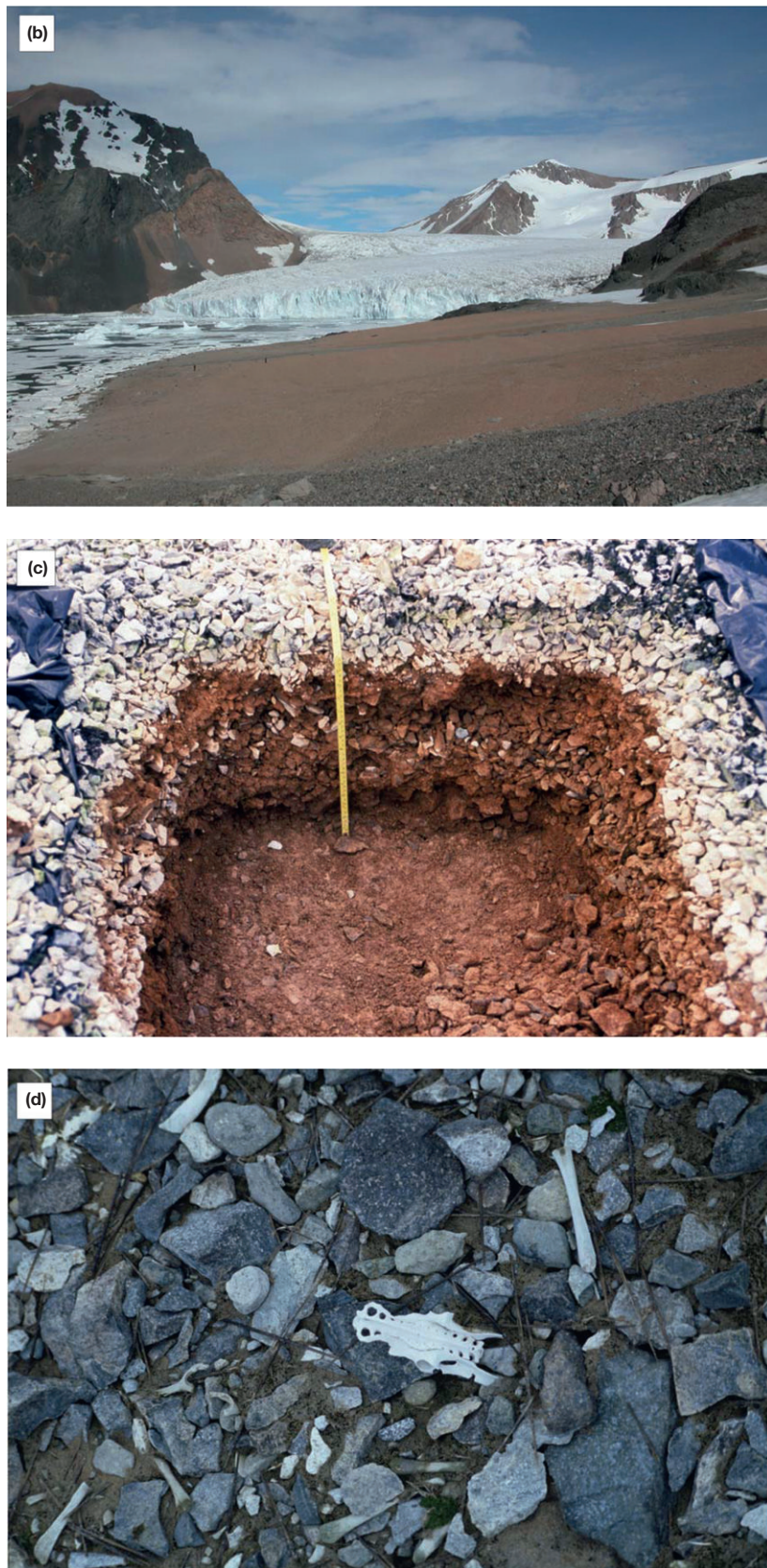


Figure 8 (Continued)

eastern Arctic Canada formerly covered by the Laurentide Ice Sheet (Andrews, 1970) and is consistent with 'Type-A curves' of Quinlan and Beaumont (1981) and the RSL history of sites located within Zone 1 of Clark et al. (1978) (Figures 1 and 2). RSL curves from the northern part of the QEI (Ellef Ringnes and Amund Ringnes islands), from areas behind the former ice-sheet margin, also show continuous exponential crustal relaxation since deglaciation (Atkinson and England, 2004). However, RSL curves from areas close to the ice-sheet margin have a more complex history that comprises an initial period of emergence followed by a late Holocene transgression (Figure 6 (f); Atkinson and England, 2004; Dyke, 1998). On the western islands of the archipelago close to or immediately beyond the ice-sheet limit, there is evidence for a period of late Holocene sea-level rise or a stillstand following RSL fall (Hodgson et al., 1994). The RSL curves toward the former ice-sheet margins and beyond therefore show an initial phase of exponential emergence followed by RSL rise during the Holocene. In some cases, this submergence may be ongoing. The curves from these areas are thus characteristic of 'Type B' RSL curves (Quinlan and Beaumont, 1981) and the Zone I/II transition of Clark et al. (1978) (Figures 1 and 2).

Antarctica

In contrast to the Arctic as illustrated by the case study from the QEI, the development of RSL data from Antarctica is far less comprehensive and is predominantly based on a relatively small number of sites (Figure 7). This is due to several reasons: a large proportion of the coast (>95%) is fronted by glacier ice or ice shelves and so there are few sites where RSL change is recorded; there are logistical difficulties and expense in reaching remote coastal sites; there are difficulties with dating the evidence because of the paucity of organic material. The latter problem is gradually being overcome with the development of innovative ways of dating raised beaches (Figure 8) but the problem of finding suitable sites remains the most intractable and for several sectors of the Antarctic coastline there is little prospect of good quality RSL data.

Most Antarctic RSL data come from one of three regions (Figure 7): (i) the Ross Sea sector, (ii) East Antarctic coastal ice-free areas ('oases'), particularly 75–105° E, and 30–45° E, and (iii) the Antarctic Peninsula. There are very few data for other parts of the Antarctic continent, although RSL studies have been carried out on a number of sub-Antarctic islands.

The Ross Sea arguably has the best quality RSL data in Antarctica. Well-dated RSL curves have recently been developed from the Scott Coast, along the west side of the Ross Sea embayment (Figure 9; Baroni and Hall, 2004; Hall and Denton, 1999; Hall et al., 2004). The development of RSL curves in the Ross Sea region has been aided by innovative approaches to material for radiocarbon dating. While material similar to that used in the Arctic (e.g., *in situ* marine shells) is sometimes found, this is not usually abundant enough to constrain RSL curves in the Antarctic and so other organic material has also been used. For example, subfossil remains from former penguin rookeries (guano-rich soils, bones, feathers) have been used to constrain sea level, based on the assumption that sea level must have been lower than the penguin remains (penguins nest above the current storm

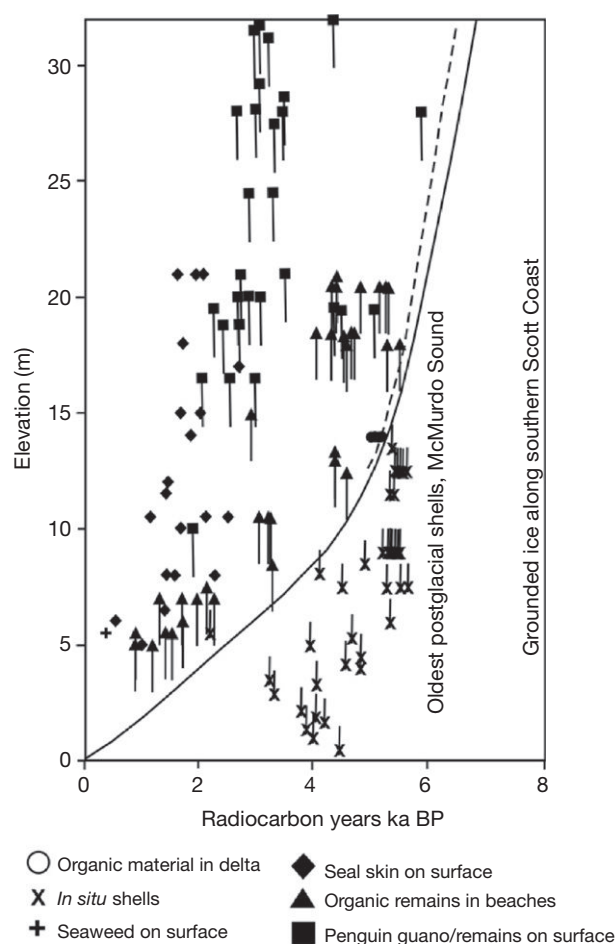


Figure 9 RSL curves for the southern Scott Coast of Victoria Land, Antarctica (see Figure 7 for location). The solid line represents the best fit to the data, whereas the dashed line defines the range of possible RSL change. The samples are from a 60-km stretch of coastline that experienced a relatively uniform ice load and RSL curves from the individual sites all overlap. Dates are corrected for a marine reservoir effect of 1300 years and the precision of sample elevations is estimated at 1 m or better for most samples. Modified from Hall BL, Baroni C, and Denton GH (2004) Holocene relative sea-level history of the southern Victoria Land Coast, Antarctica. *Global and Planetary Change* 42: 241–263.

beach). Similarly, skin fragments from hauled-out elephant seals that have been incorporated into beach material have been used to constrain past sea level (Baroni and Hall, 2004).

The Ross Sea curves show exponential RSL fall since deglaciation of the coastline by the retreating West Antarctic Ice Sheet (Conway et al., 1999). Marine limit in the Ross Sea (~30 m) (Figure 9) is substantially lower than in the Arctic (>150 m in the QEI – see above) because much of the isostatic recovery took place as restrained rebound (i.e., was not recorded as raised marine landforms because of ongoing ice cover of the Ross Sea coastline), and because the Antarctic Ice Sheet continues to provide a substantial regional load on the crust whereas, with the exception of Greenland, the Late Quaternary ice sheets over much of the Arctic have now been largely removed. The Scott Coast is unusual in that the orientation of the former ice-sheet grounding line (east–west) was

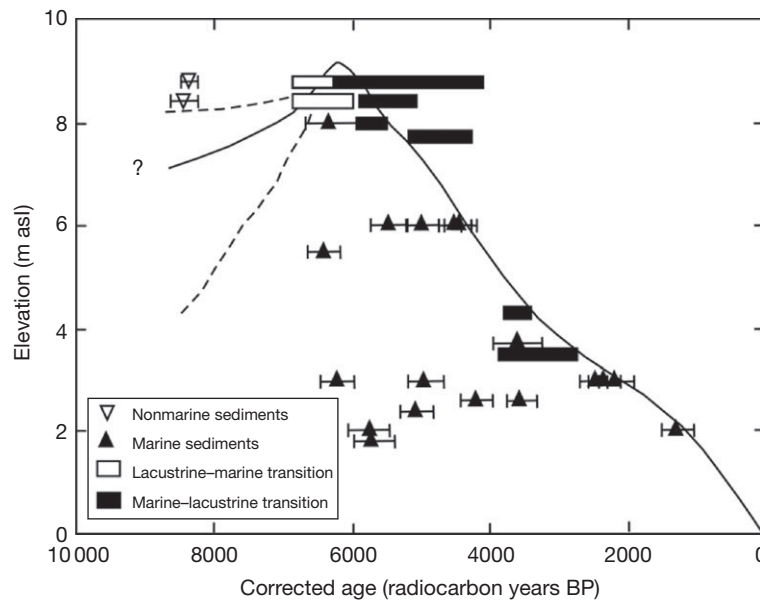


Figure 10 Relative sea-level history of the Vestfold Hills, East Antarctica (see [Figure 7](#) for location) for the last ~8000 ^{14}C years, based on isolation basin observations (modified from [Zwartz D, Bird M, Stone J, and Lambeck K \(1998\)](#) Holocene sea level and ice-sheet history in the Vestfold Hills, East Antarctica. *Earth and Planetary Science Letters* 155: 131–145). The box symbols 'lacustrine–marine transition' and 'marine–lacustrine transition' represent the ages of lacustrine–marine and marine–lacustrine sediment transitions recorded in cores collected from six lakes, plotted against the elevation of the sill connecting each of the six lake basins to the sea. Core lengths required for dating these transitions varied from ~3–6 cm of organic-rich lacustrine sediment to 10–15 cm of marine sediment.

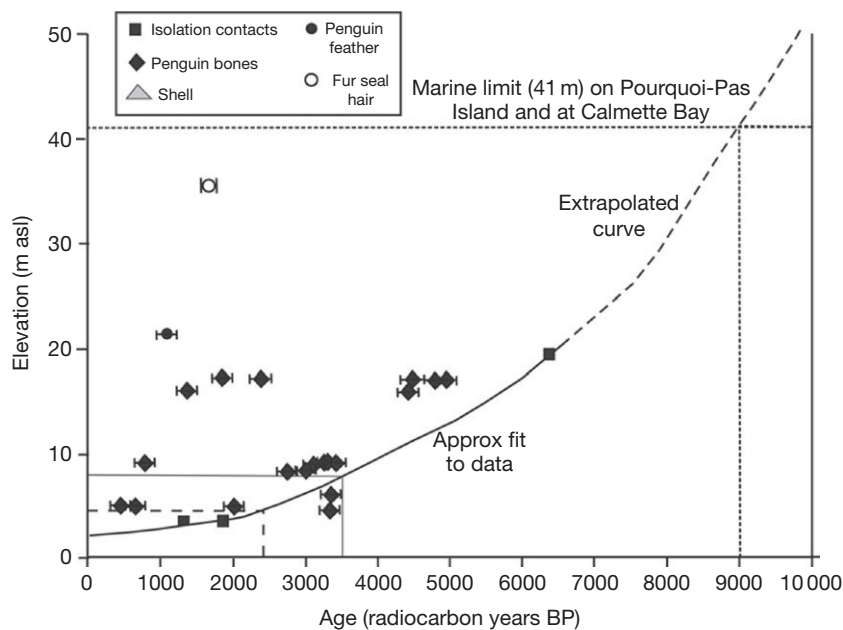


Figure 11 RSL curve for Marguerite Bay, Antarctic Peninsula (see [Figure 7](#) for location). The curve utilizes data from seven sites spread over an approximately 60 km × 30 km area in the northern part of the bay. The data are plotted with 1-sigma error bars for the ^{14}C ages and +0.2 m for altitudes. The thick solid line shows the approximate fit to the data and the dashed line shows an extrapolation of the RSL curve to the marine limit. The altitude of the marine limit is shown by the dotted line. The y-intercept of the curve is at 2 m asl, which is the altitude of the modern beach in this area. Modified from [Bentley MJ, Hodgson DA, Smith JA, and Cox NJ \(2005\)](#) Relative sea level curves for the South Shetland Islands and Marguerite Bay, Antarctic Peninsula. *Quaternary Science Reviews* 24: 1203–1216.

transverse to the coastline (which runs north–south), and the grounding line retreated south, along the coast. Despite this, the RSL curves for a series of sites along the Scott Coast are nearly identical. This implies that retreat of the ice-sheet grounding line along this coastline was rapid. The RSL curves have been used to show that the West Antarctic Ice Sheet grounding line in the Ross Sea retreated past the southern Scott Coast by 6300–6600 ^{14}C years BP.

In East Antarctica, Zwartz et al. (1998) used data from isolation basins in the Vestfold Hills to constrain former RSL (Figure 10). A series of freshwater–marine and marine–freshwater transitions in lake sediment cores were used to determine an RSL curve for the past 8000 ^{14}C years. The RSL curve is characteristic of near-field sites located close to the margin of a retreating ice sheet. During the early Holocene, the continued regional loading of the crust by the ice sheet (located only a few tens of kilometers away) prevented isostatic recovery from keeping pace with eustatic sea-level rise and so there was a net RSL rise. A subsequent switch to RSL fall occurred in the mid-Holocene when the meltwater contribution of the Northern Hemisphere ice sheets had declined to near-zero, and so the slow isostatic recovery of the East Antarctic coastline dominated (Figure 10). This form of RSL curve corresponds to Zone V of Clark et al. (1978) (Figure 1). The shift from RSL rise to fall is marked by an RSL highstand. In the Vestfold Hills, the mid-Holocene highstand ($\sim 9\text{--}9.5\text{ m}$) was reached at 6200 ^{14}C years BP. Similar isolation basin work has been carried out in other parts of the East Antarctic coast and results show a similar pattern of a mid-Holocene highstand and subsequent RSL fall (e.g., Verleyen et al., 2005). There has also been extensive work on raised beaches in Lutzow-Holm Bay (Figure 7; e.g., Miura et al., 1998) but some of the shells used to underpin the chronology have proved challenging to date (Takada et al., 2003).

In the Antarctic Peninsula region, RSL curves reflect a more complex history. Bentley et al. (2005) reviewed published investigations of isolation basins and raised beaches in the South Shetland Islands and showed that the northernmost part of the Peninsula has an RSL history broadly comparable to that for Vestfold Hills, with strong geomorphological and isolation basin evidence for a mid-Holocene highstand, but whose timing (especially the initial RSL rise) is not yet well constrained.

In contrast, further south in Marguerite Bay, the RSL curve (Figure 11) most likely reflects exponential isostatic decay after removal of a thick grounded ice load that covered the bay during the LGM. This curve is reasonably well constrained by dates on penguin remains from raised beaches (Figure 8) and two isolation basins from about 6500 ^{14}C years BP to the present. Marine limit is defined by the uppermost elevation of raised beaches and is between 41 and 55 m in altitude. The shape of the curve prior to 6500 ^{14}C years BP is not well-constrained chronologically but an exponential shape is assumed on the basis of the regional ice-sheet history, which shows that the former ice margin extended well offshore from the Antarctic Peninsula. Assuming exponential uplift, extrapolation of the curve prior to 6500 ^{14}C years BP back to the regional marine limit at 41 m asl gives an estimate of about 9000 ^{14}C years BP as a minimum for deglaciation of inner Marguerite Bay.

One of the most important applications of existing and new RSL curves from Antarctica is to constrain coupled models of ice-sheet volume change and glacioisostatic adjustment (e.g., Bassett et al., 2007). These models attempt to infer reasonable ice-sheet histories by testing which histories can satisfy the constraints provided by far-field and near-field RSL datasets. At present, such models are not well constrained in Antarctica, compared to other continents, which is hampering efforts to understand the history of the Antarctic ice sheet.

See also: **Glaciations: Late Quaternary of Antarctica. Sea Level Studies:** Eustatic Sea-Level Changes Since the Last Glacial Maximum; **Isostasy:** Glaciation-Induced Sea-Level Change.

References

- Andrews JT (1970) *A Geomorphic Study of Postglacial Uplift with Particular Reference to Arctic Canada*. London, England: Institute of British Geographers Special Publication 2.
- Atkinson N (2003) Late Wisconsinan glaciation of Amund and Ellef Ringnes islands, Nunavut: Evidence for the configuration, dynamics and deglacial chronology of the northwest sector of the Innuitian Ice Sheet. *Canadian Journal of Earth Sciences* 40: 351–363.
- Atkinson N and England J (2004) Postglacial emergence of Amund and Ellef Ringnes islands, Nunavut: Implications for the northwest sector of the Innuitian Ice Sheet. *Canadian Journal of Earth Sciences* 41: 271–283.
- Baroni C and Hall BL (2004) A new relative sea-level curve for Terra Nova Bay, Victoria Land, Antarctica. *Journal of Quaternary Science* 19: 377–396.
- Bassett SE, Milne GA, Bentley MJ, and Huybrechts P (2007) Modeling Antarctic sea-level data to explore the possibility of a dominant Antarctic contribution to meltwater pulse 1A. *Quaternary Science Reviews* 26(17–18): 2113–2127. <http://dx.doi.org/10.1016/j.quascirev.2007.06.011>.
- Bassett SE, Milne GA, Mitrovica JX, and Clark PU (2005) Ice sheet and solid earth influences on far-field sea-level histories. *Science* 309: 925–928.
- Bentley MJ, Hodgson DA, Smith JA, and Cox NJ (2005) Relative sea level curves for the South Shetland Islands and Marguerite Bay, Antarctic Peninsula. *Quaternary Science Reviews* 24: 1203–1216.
- Blake W Jr. (1970) Studies of glacial history in Arctic Canada. *Canadian Journal of Earth Sciences* 7: 634–664.
- Blake W Jr. (1975) Radiocarbon age determination and postglacial emergence at Cape Storm, southern Ellesmere Island. *Geografiska Annaler* 57A: 1–71.
- Clark JA, Farrell WE, and Peltier WR (1978) Global changes in postglacial sea level: A numerical calculation. *Quaternary Research* 9: 265–287.
- Clark PU and Mix AC (2002) Ice sheets and sea level of the Last Glacial Maximum. *Quaternary Science Reviews* 21: 1–7.
- Conway H, Hall BL, Denton GH, Gades AM, and Waddington ED (1999) Past and future grounding-line retreat of the West Antarctic Ice Sheet. *Science* 286: 280–283.
- Dyke AS (1998) Holocene delevelling of Devon Island, Arctic Canada: Implications for ice sheet geometry and crustal response. *Canadian Journal of Earth Sciences* 35: 885–904.
- Dyke AS (1999) The last glacial maximum and the deglaciation of Devon Island: Support for an Innuitian Ice Sheet. *Quaternary Science Reviews* 18: 393–420.
- England J (1976) Late Quaternary glaciation of the Queen Elizabeth Islands, Northwest Territories, Canada: Alternative models. *Quaternary Research* 6: 185–202.
- England J (1983) Isostatic adjustments in a full glacial sea. *Canadian Journal of Earth Sciences* 20: 895–917.
- England J (1992) Postglacial emergence in the Canadian High Arctic: Integrating glacioisostasy, eustasy, and late deglaciation. *Canadian Journal of Earth Sciences* 29: 984–999.
- England J (1998) Support for the Innuitian Ice Sheet in the Canadian High Arctic during the Last Glacial Maximum. *Journal of Quaternary Science* 13: 275–280.
- England J (1999) Coalescent Greenland and Innuitian ice during the Last Glacial Maximum: Revising the Quaternary of the Canadian High Arctic. *Quaternary Science Reviews* 18: 421–556.

- England J, Atkinson N, Bednarski J, Dyke AS, Hodgson DA, and Ó Cofaigh C (2006) The Innuitian Ice Sheet: Configuration, dynamics and chronology. *Quaternary Science Reviews* 25: 689–703.
- Hall BL, Baroni C, and Denton GH (2004) Holocene relative sea-level history of the southern Victoria Land Coast, Antarctica. *Global and Planetary Change* 42: 241–263.
- Hall BL and Denton GH (1999) New relative sea-level curves for the southern Scott Coast, Antarctica: Evidence for Holocene deglaciation of the western Ross Sea. *Journal of Quaternary Science* 14: 641–650.
- Hodgson DA, Taylor RB, and Fyles JG (1994) Late Quaternary sea level changes on Brock and Prince Patrick islands, western Canadian Arctic Archipelago. *Géographie physique et Quaternaire* 48: 69–84.
- Long AJ, Roberts DH, and Wright M (1999) Isolation basin stratigraphy and Holocene relative sea-level change on Arveprinsen Ejland, Disko Bugt, West Greenland. *Journal of Quaternary Science* 14: 323–345.
- Miura H, Moriwaki K, Maemoku H, and Hirakawa K (1998) Fluctuations of the East Antarctic ice sheet margin since the last glaciation from the stratigraphy of raised beach deposits along the Soya Coast. *Annals of Glaciology* 27: 297–301.
- Ó Cofaigh C (1999) Holocene emergence and shoreline develevelling, southern Eureka Sound, High Arctic Canada. *Géographie physique et Quaternaire* 53: 235–247.
- Ó Cofaigh C, England J, and Zreda M (2000) Late Wisconsinan glaciation of southern Eureka Sound: Evidence for extensive Innuitian ice in the Canadian High Arctic during the Last Glacial Maximum. *Quaternary Science Reviews* 19: 1319–1341.
- Quinlan G and Beaumont C (1981) A comparison of observed and theoretical postglacial relative sea level in Atlantic Canada. *Canadian Journal of Earth Sciences* 18: 1146–1163.
- Takada M, Tani A, Miura H, Moriwaki K, and Nagatomo T (2003) ESR dating of fossil shells in the Lutzow-Holm Bay region East Antarctica. *Quaternary Science Reviews* 22: 1323–1328.
- Verleyen E, Hodgson DA, Milne GA, Sabbe K, and Vyverman W (2005) Relative sea-level history from the Lambert Glacier region, East Antarctica, and its relation to deglaciation and Holocene glacier readvance. *Quaternary Research* 63: 45–52.
- Walcott RI (1972) Late Quaternary vertical movements in eastern North America: Quantitative evidence of glacio-isostatic rebound. *Reviews of Geophysics and Space Physics* 10: 849–884.
- Zwartz D, Bird M, Stone J, and Lambeck K (1998) Holocene sea level and ice-sheet history in the Vestfold Hills, East Antarctica. *Earth and Planetary Science Letters* 155: 131–145.

Late Quaternary Sea-Level Changes in Greenland

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Overview of the Greenland Ice Sheet

The Greenland Ice Sheet (GrIS) today is ~ 1.7 million km² in area and covers $\sim 80\%$ of the land mass of Greenland. Its maximum thickness is 3367 m, but has an average thickness of 1600 m (Johnsen et al., 2001; Thomas, 2001). The highest surface elevations are in the northern dome (~ 3200 m above sea level) and the southern dome (~ 2850 m above sea level). It is the only Northern Hemisphere ice sheet to have survived the climate warming since the end of the Last Glacial Maximum (LGM) and would contribute up to 7.3 m of globally averaged sea-level rise if it were to melt entirely (Lemke et al., 2007). Studies of relative sea-level (RSL) changes in Greenland record spatial and temporal patterns of glacio-isostatic rebound that provide important information about the size and former history of the GrIS. In particular, there has been a long tradition of using dated shorelines to infer ages of associated moraine deposits, thus providing age constraints on ice margins through time (e.g., Funder and Hansen, 1996; Kelly, 1980; Weidick, 1968, 1996).

In this article, we start by reviewing existing knowledge regarding the Greenland interglacial sea-level history and the types of RSL data used in such reconstructions. Following this, we briefly describe the long-term history of the GrIS from its initiation ~ 3 Ma until the LGM. We then discuss the evidence for RSL changes since the LGM, with a particular focus on southwest Greenland.

RSL Curves and Types of Evidence Found in Greenland During Interglacials

Within and close to the limits of former ice sheets, RSL histories following deglaciation are dominated by glacio-isostasy in the form of crustal rebound following ice sheet decay. In Clark et al. (1978) classic model of zones of RSL change around the globe, Greenland stands between zones I and II. During interglacials Greenland has experienced tens to hundreds of meters of RSL fall following local (Greenland) deglaciation, but it is also affected by forebulge collapse causing subsequent RSL rise. Clark et al. model simplifies the interglacial RSL curve for Greenland because it does not take into account the proximity of the Laurentide ice sheet and its decaying forebulge, or the late Holocene regrowth of the GrIS that happened during the neoglacial period, which followed the mid-Holocene climatic optimum (8–5 k cal yr BP) (Dahl-Jensen et al., 1998; Kelly, 1980). Nevertheless, the Clark et al. (1978) model provides a good indication of typical interglacial RSL changes in Greenland. This shape of curve is often referred to in the literature as a 'J-shaped' RSL curve because of the fall in RSL from above to below present sea level, followed by a late Holocene rise to present (Long et al., 2011; Rasch, 2000).

Evidence for pre-Holocene RSL changes in Greenland is fragmentary (Funder et al., 1991). In east Greenland, raised marine, deltaic, and fluvial sediments from several previous interglacials are preserved in small coastal cliff and river valley sections, notably in the Scoresby Sund area (Funder et al., 1998) (Figure 1(a)). Identifying their age and context can be difficult because sites are widely spaced and each contains only a small part of the overall deglaciation history. Typically, the sequences record sediments associated with a fall in RSL, with progressively shallower water deposits that may be capped with beach deposits as the land finally emerges from the sea. Since the LGM, evidence for RSL changes is much more plentiful. Once the ice margin had retreated on land it is relatively easy to map the height of the local marine limit, which represents the maximum postglacial uplift in that locality. The marine limit is often defined by the lower limit of perched boulders or the upper limit of wave washed bedrock.

After the marine limit has formed, local RSL fall leaves evidence of former marine conditions across the landscape in the form of glacio-marine spreads, raised beaches, and deltas (e.g., Funder, 1989; Funder and Hansen, 1996; Kelly, 1980; Rasch, 2000; Ten Brink and Weidick, 1975; Weidick, 1972). Isolation basins (natural depressions that at different times are either connected to or isolated from the sea by changes in RSL) provide another source of RSL data, useful for tracking millennial-scale RSL change (Long et al., 2011). Researchers are also starting to investigate salt marsh deposits in west Greenland which record small-scale RSL movements over the last few centuries (Long et al., 2010, 2012; Woodroffe and Long, 2009).

Greenland Ice Sheet Through Time

Eocene–Pliocene

During the late Eocene (~ 41 Ma), it is thought Greenland was mostly ice free at a time when glaciation was developing in Antarctica (Edgar et al., 2007). There is evidence that small mountain glaciers were present in east Greenland between ~ 38 and 30 Ma from records of ice-rafted debris in marine sediments from the Norwegian–Greenland sea (Eldrett et al., 2007). However, it was not until much later at the Pliocene/Quaternary boundary (~ 3 –2 Ma) that the Northern Hemisphere entered an 'ice-house' world with extensive glaciation over Greenland (Maslin et al., 1998). Models suggest that during this period the ice sheet grew from a small area in east Greenland to cover most of the land surface, driven by changes in orbital forcing and a decrease in atmospheric CO₂ levels (Lunt et al., 2008). Despite this overall cooling trend, the Kap København and Île de France formations in northern and northeastern Greenland show that climate was mild for at least small periods at this time (Bennike et al., 2002a; Funder

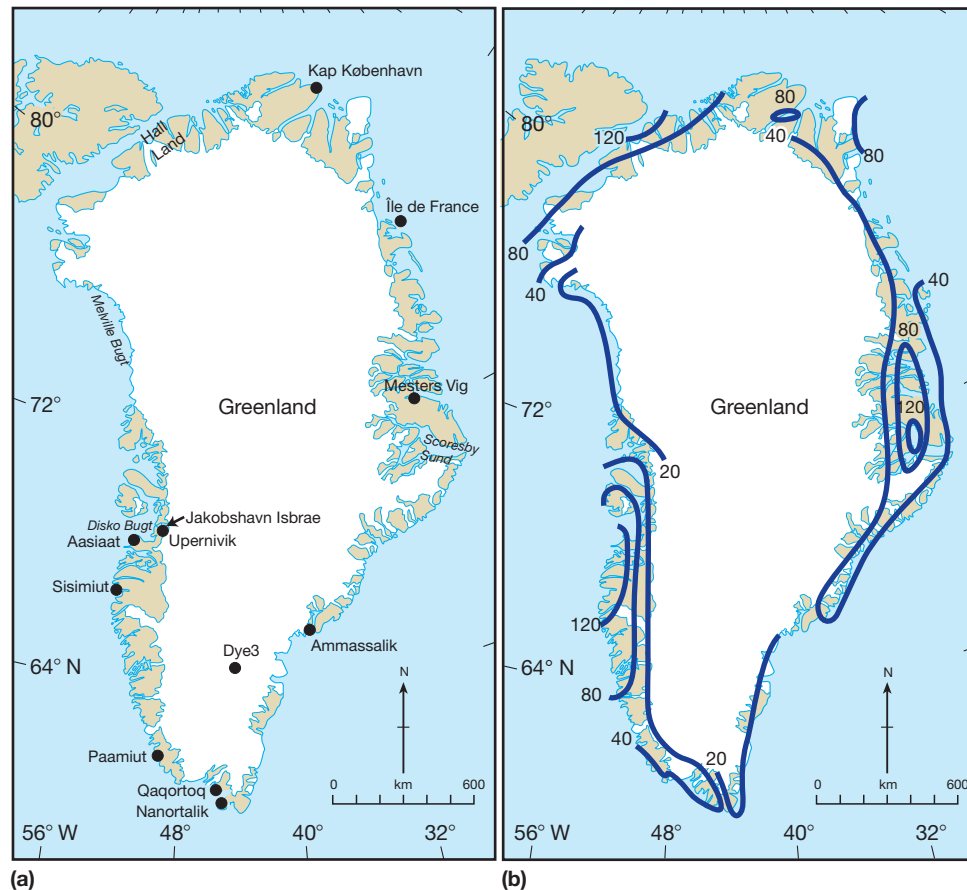


Figure 1 (a) Location map of Greenland showing the locations discussed in the text. (b) Map of the Holocene marine limit from Greenland. Modified from Funder S and Hansen L (1996) The Greenland ice sheet – A model for its culmination and decay during and after the Last Glacial Maximum. *Bulletin of the Geological Society of Denmark* 42: 137–152.

et al., 2001) (Figure 1(a)). The Kap København formation is interpreted as an interglacial marine and then freshwater sequence, deposited under the J-shaped RSL history described above. The formation has yielded plant, insect, and a few vertebrate remains suggesting the presence of local boreal forest with estimated mean temperatures of the warmest month of 10 °C here ~2.4 Ma. Funder et al. (2001) suggest that the rest of Greenland could not have been glaciated at this time if the northern tip experienced mean temperatures of 10 °C in summer, but these climatic optimums were likely intensely warm, short-lived interglacial phases within the overall trend of ice build-up during the late Pliocene/Quaternary transition.

Early to Late Quaternary

Terrestrial evidence for ice sheet size during the early to late Quaternary is limited and most scientific attention has focused on ice-sheet reconstructions during the Marine Isotope Stage (MIS) 11 (420–360 ka) and 5e (130–115 ka) interglacials, which are seen as potentially good analogs for the GrIS under future climate change. The Saalian stage (MIS 6 – 188–135 ka) was probably the largest glacial phase in the Quaternary from sedimentary evidence in east Greenland that the ice sheet

overran high mountain plateaux in this region during this time (Funder et al., 1998, 2011).

Direct evidence for the dimension of the GrIS during MIS 11, the so-called ‘super-interglacial’ (Augustin et al., 2004), is lacking. However, some data from far-field sites suggest that global sea level was up to 21 m above present during MIS 11 (Kaufman and Brigham-Grette, 1993; Olson and Hearty, 2009), and this would have required all of the Greenland, West Antarctic, and part of the East Antarctic ice sheets to melt. Sea-level reconstructions from this time interval are debated with, for example, the Red Sea sea-level record suggesting that the MIS 11 high stand was similar (within uncertainties (± 6.5 m)) to the Holocene (Rohling et al., 2010) (see [Eustatic Sea-Level Changes – Glacial-Interglacial Cycles](#)). This alternative interpretation would rule out catastrophic ice sheet collapse at both poles but the uncertainties in the latter reconstructions still allow for significant melt of the GrIS during this period.

The most compelling evidence of ice sheet retreat during the Quaternary is not from sea-level evidence but from a recent study of ancient DNA preserved in basal silty ice from the Dye 3 ice core in southern Greenland (Figure 1(a)). This study suggests that southern Greenland supported a similar boreal ecosystem to that at Kap København at some point

between 800–450 ka, with summer temperatures around 10 °C (Willerslev et al., 2007). In addition, a sea bed core offshore SW Greenland (ODP site 646) shows increased *Picea* pollen during MIS 11 (de Vernal and Hillaire-Marcel, 2008), adding weight to the hypothesis that southern Greenland was covered by boreal forest during this interglacial.

MIS 5e – The Eemian

Debates surrounding the size of the GrIS during the Eemian interglacial (130–115 ka BP) have been ongoing for over a decade. Paleoclimate data suggest peak summer temperatures ~4 and ~5 °C above present in northwest and east Greenland during this interglacial under higher than present solar insolation (Miller et al., 2006). A recent assessment of global sea-level records suggests that the GrIS contributed ~2.5 m to the global sea-level high stand of at least 6.6 m during this period, requiring significant melt of the ice sheet (Kopp et al., 2009).

Modeling experiments also suggest a smaller than present ice sheet during MIS 5e, although they are not all in agreement. For example, Cuffey and Marshall (2000) reconstruct a small, northern GrIS with almost all of southern Greenland ice free. They suggest an ice-equivalent global sea-level contribution of 4–5.5 m from Greenland. Tarasov and Peltier (2003) predict more conservative ice sheet retreat, with ice just covering the Dye-3 ice core location and a global sea-level contribution of between 2.7 and 4.5 m. Otto-Bliesner et al. (2006) infer a combined Greenland and Canadian Arctic contribution to global sea level of ~3.4 m, with a small southern ice cap separated from the larger northern Greenland ice sheet, and Robinson et al. (2011) agree that northern and southern domes are likely to have survived and estimate the global sea-level contribution at 3.7–4.4 m. Although these models differ in their initial set-up, their climate and their sea-level forcing, they all suggest that the GrIS contributed at least 2.7 m of global sea level during MIS 5e, and all suggest a significantly smaller (or absent) ice sheet in southern Greenland during the warmer than present conditions of this interglacial.

The presence of pre-Eemian plant remains at Dye3 (see above) has not solved the issue of ice sheet retreat from southern Greenland, because there is a significant hiatus between the age of silty basal ice (containing macrofossils and suggested to be ~450–800 ka old) and the overlying clean ice which is late MIS 5e in age (Alley et al., 2010). Alley et al. (2010) argue that the material contained in the basal ice may have been preserved as permafrost during MIS 5e when the GrIS had retreated north of this region. Results from a new deep ice core through the Eemian from northern Greenland (NEEM) should contribute toward our understanding of ice sheet melt and climate conditions during this interglacial period.

The same offshore sediment core that provides evidence for boreal forest during MIS 11 also provides evidence for a significantly smaller than present ice sheet during MIS 5e (de Vernal and Hillaire-Marcel, 2008). A rapid increase in alder pollen at the start of MIS 5e suggests high summer temperatures. A later peak in fern spores (*Osmunda*) is unparalleled in the last million years and indicates the development of dense fern vegetation over southern Greenland with conditions

comparable to that in modern boreal forests. Moreover, Carlson et al. (2008) use variations in Ti and Fe in a sediment core offshore of southwest Greenland as a proxy for terrestrial sediment carried by summer meltwater. They interpret high and constant Ti and Fe between ca. 132 and 120 ka BP as evidence that the GrIS was significantly smaller than present at this time and contributed a substantial proportion of the higher sea level during this interglacial.

RSL studies are relatively scarce during this period, but sediments preserved in and near Scoresby Sund, east Greenland, suggest that RSL fell from a marine limit of 70 m to below present sea level before rising to ~20 m (Funder et al., 1998). The magnitude of the RSL rise has no analog in the Holocene and Funder et al. suggest that it may be due to either glacio-isostatic depression during ice sheet readvance or large-scale eustatic melt from other far-field ice sheets after local melt had slowed or stopped. They favor the second hypothesis because macrofossils suggest the region was warmer than today during the transgressive phase, implying that this region of the GrIS had already deglaciated ahead of other far-field land-based ice masses (Funder et al., 2011).

The Last Deglaciation

The GrIS was at its maximum position during the LGM ca. 22–19 ka BP and in most locations ice extended offshore onto the continental shelf (Bennike and Björck, 2002; Funder and Hansen, 1996; Funder et al., 2011; Simpson et al., 2009). In some locations ice margin retreat across the shelf began by the start of the Late Glacial (~15–11.5 ka BP), and in all locations except for southern Greenland, ice retreat across the present coastline occurred only in the early Holocene (Bennike and Björck, 2002; Simpson et al., 2009). In southern Greenland, dates on the marine limit indicate ice sheet recession on land occurred at Qaqortoq and Nanortalik by ~11.0 and ~13.8 ka respectively (Bennike et al., 2002b; Sparrenbom et al., 2006a,b). In southeast Greenland, two studies report deglaciation in fjords near Ammassalik at ~9.9–9.7 ka BP (Long et al., 2008; Roberts et al., 2008).

After the ice sheet retreated onshore we can track when local ice-free conditions became established by inferring the timing of the local marine limit, by extending upward the local RSL curve to intersect the marine limit (Kelly, 1985; Weidick, 1972). However, this assumes that RSL fell immediately after deglaciation, which may not always have been the case (Funder and Hansen, 1996). Most RSL data from Greenland are from this period of rapid RSL fall, reflecting the availability of material for dating. A map of the marine limit for Greenland (Figure 1(b)) shows three domes of maximum uplift, one over the Sisimiut district in west Greenland (maximum ~140 m), one over the Scoresby Sund region in east Greenland (maximum ~135 m), and one on Hall Land in north Greenland (maximum ~130 m) (Weidick and Bennike, 2007). Low marine limits (~20 m) occur in regions where the ice sheet is currently close to the coast, such as in Melville Bugt in northwest Greenland and in parts of southeast Greenland. In these regions either ice volume has not changed significantly since the LGM or there has been significant build-up of ice during the late Holocene. Where there is a wide ice-free strip of land between the ice sheet and coast, the limit has a domed

appearance, rising from the outer coast to a maximum before decreasing toward the ice sheet margin (Funder and Hansen, 1996, Weidick, 1972).

Late-Glacial/Holocene RSL Changes in Greenland

Following the development of the radiocarbon dating technique, researchers in Holocene RSL in east and west Greenland began to develop local and regional chronologies for shoreline development and hence, ice sheet history. Pioneers of this approach dated marine molluscs and driftwood from Mesters Vig in east Greenland to develop two of the first radiocarbon-dated RSL records from Greenland, which revealed the classic pattern of rapid RSL fall during the early postglacial period, with RSL falling close to present by the mid-Holocene (Lasca, 1966; Washburn and Stuiver, 1962). Anker Weidick's work from the 1950s onward on systematic mapping and dating of marine and associated glacial deposits was the first to develop regional scale patterns of RSL change in west Greenland (Weidick, 1968).

Using similar methods, subsequent studies added new knowledge regarding the RSL and ice sheet history of other parts of Greenland (see reviews by Funder and Hansen, 1996; Kelly, 1985). Disko Bugt, a large marine embayment in central west Greenland, has become a particular focus for study because the ice sheet here is strongly influenced by the world's fastest flowing ice stream – Jakobshavn Isbrae, which drains a large portion of the GrIS and discharges ice into Disko Bugt and Baffin Bay. Since the late 1990s, researchers have developed several isolation basin records from Disko Bugt to understand the Holocene ice sheet history of this area, giving smaller age and altitude errors compared to earlier reconstruction methods (Long and Roberts, 2003; Long et al., 1999, 2003, 2006).

An isolation basin RSL record characteristic of postglacial uplift in west Greenland comes from Upernivik, midway along Jakobshavn Isfjord between the open coast of Disko Bugt and the ice sheet margin (Long et al., 2006) (Figure 2(a)). The marine limit, reflecting the timing of local deglaciation and amount of glacio-isostatic rebound in this area dates from ~8 k cal years BP and is ~41 m above present. The isolation basin staircase shows RSL fell steeply from this level in the early-mid Holocene, falling below present shortly before 4.5 k cal years BP. The lowstand in RSL is reached by ~3 k cal years BP at around –5 m before rising to present in the late Holocene. This pattern reflects initial glacio-isostatic rebound followed by depression during the mid-late Holocene neoglaciation period by glacial readvance and forebulge collapse (Kelly, 1980; Long et al., 2006).

This RSL record is similar to those found using isolation basins elsewhere in western Greenland during the Holocene. Figure 2(b) shows the Upernivik record together with two other isolation basin RSL histories from different regions of Greenland; Sisimiut (central west) and Nanortalik (south). The records from Upernivik and Sisimiut show RSL fell rapidly during the early and mid-Holocene, falling below present at or after ~4 k cal years BP and then rising by up to 5 m in the last 2–3 k cal years. The record from the far south of Greenland starts in the Late Glacial, where earlier ice-free conditions were established compared to areas to the north. Sparrenbom et al. (2006b) show that at Nanortalik RSL fell rapidly to below

present by ~10–8 k cal years BP, reached a lowstand (–6 m to –8 m) sometime in the mid-Holocene and then rose by an equivalent amount to present. The very quick fall in early and mid-Holocene RSL here shows that there was a rapid removal of a large volume of ice during ice margin retreat in this region (Sparrenbom et al., 2006a,b).

RSL Changes During the Neoglaciation and Last Millennium

Steeply falling RSL curves in the early Holocene, and a lack of RSL data from the mid- to late Holocene suggest that in many areas RSL fell below present in the mid-Holocene and has risen to present in the late Holocene. There are two hypotheses, not necessarily mutually exclusive, for the cause of this RSL rise. Rasch and Nielsen (1995) argue that J-shaped RSL curves reflect the collapse of the Greenland Ice Sheet forebulge. In contrast, Kelly (1980) and Weidick (1993, 1996) attribute late Holocene RSL rise to crustal reloading by neoglaciation ice sheet expansion from its mid-Holocene minimum during the Holocene climatic optimum (8–5 k cal years BP; Dahl-Jensen et al., 1998). Kelly (1980) calls the late Holocene RSL rise the 'Sandnes transgression' after a Norse church at Ameralik (west Greenland) that, it is claimed, has been lowered by between 5 and 6 m since AD 1600 by glacio-isostatic subsidence (Roussell, 1941). Evidence for ice sheet readvance during this period includes reworked organic material ripped up by the advancing ice sheet and deposited in recent moraines, and threshold lakes which are close to the ice sheet and record changes in the input of glacial material over time, recording local advance and retreat of the ice sheet margin (e.g., Briner et al., 2010; Kelly, 1980).

RSL data together with ice sheet models are particularly important in constraining the magnitude of ice sheet advance during the neoglaciation, because as an ice sheet re-advances it destroys terrestrial evidence of its former position. RSL data therefore provide an important context for interpreting recent ice sheet margin change (Long, 2009). Defining the Holocene sea-level lowstand is challenging but has been attempted using a variety of methods, including dating subaerially exposed desiccation surfaces in a marine core (Kelly, 1980), dating drowned beaches (Kuijpers et al., 1999), interpreting archaeological evidence (Kelly, 1980; Rasch and Jensen, 1997) as well as coring drowned isolation basins (Long et al., 1999, 2009; Sparrenbom et al., 2006a,b). Despite this range of indicators, we rely largely on submerged evidence from ingressed isolation basins now below high tide level. Ingression contacts in isolation basins are often eroded and measuring sill elevations is challenging when the sill is situated in or below the intertidal zone (Long et al., 2011).

The general pattern of late Holocene RSL change appears to be that the lowstand was reached earliest, and at greatest depth in the far south of Greenland. At Nanortalik, the lowstand was ~–8–~–6 m below present at ca. 7 k cal years BP, whereas at Upernivik in Disko Bugt, it occurred at ~–3–~–4 m between 3 and 2 k cal years BP (Figure 2(b)). The difference in timing and magnitude of the lowstand between these two locations in south and west Greenland reflects differences in the timing of deglaciation during the Holocene, the thickness of the local (Greenland) and non-local (especially North American) ice at the LGM causing glacio-isostatic rebound,

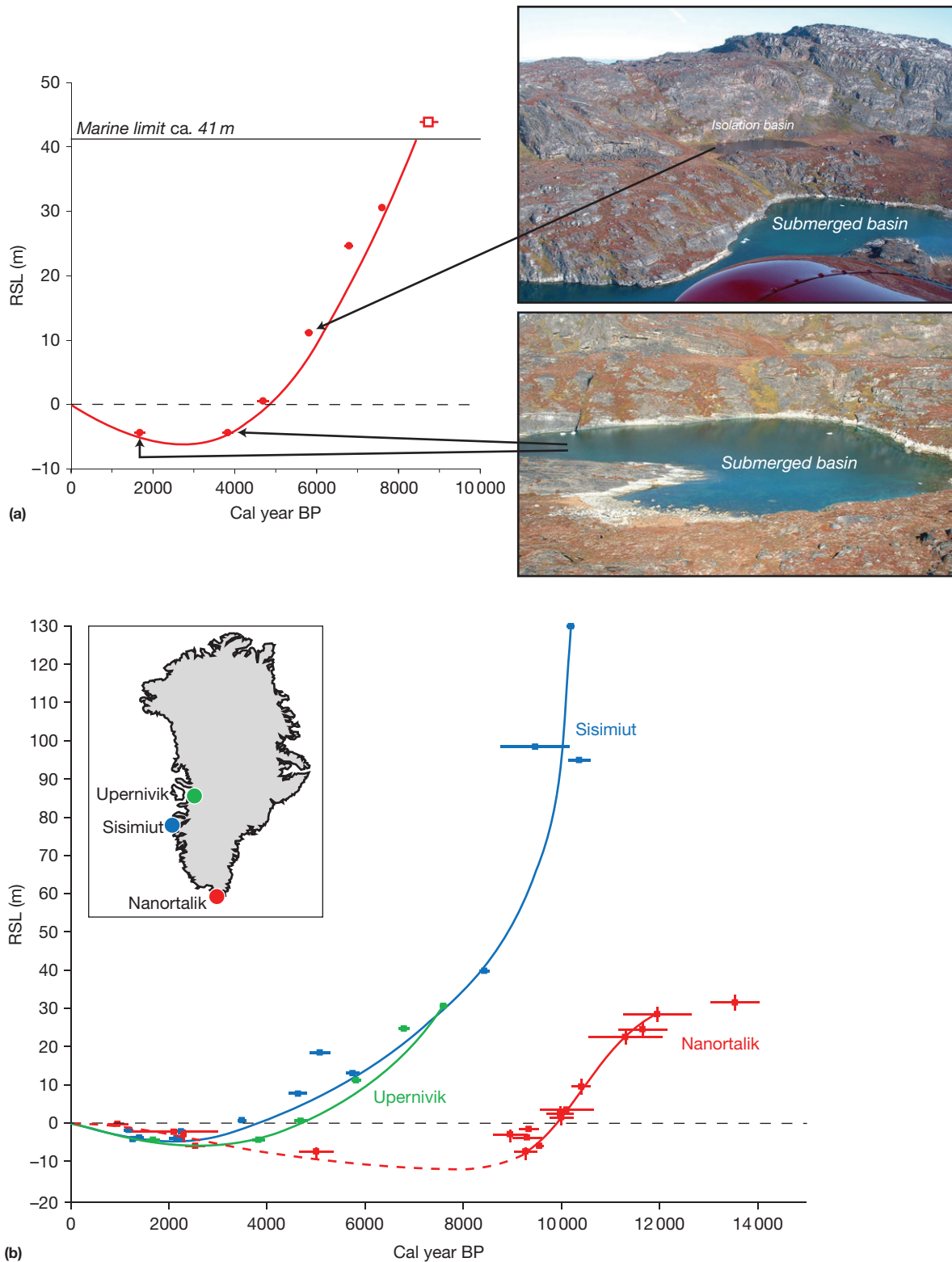


Figure 2 (a) Isolation basin RSL data from Upernivik, Disko Bugt, west Greenland (Long et al., 2006) and photos showing two of the lower basins. The two index points below present sea-level record first basin isolation and then regression during the late Holocene. (b) Isolation basin RSL data from three sites in west and southern Greenland; Upernivik, Disko Bugt (Long et al., 2006), Sisimiut, central west Greenland (Long et al., 2009; Bennike et al., 2011) and Nanortalik, south Greenland (Sparrenbom et al., 2006b). Note the lowest isolation basin of Bennike et al., (2011), at ~18 m, is very shallow and contains an incomplete sedimentary record so is not included here.

and the increase in ice mass and therefore reloading during the neoglacial period between southern and western Greenland. However, the complex interactions of these different factors means that differences in the magnitude and timing of RSL changes across west Greenland should be interpreted cautiously in terms of neoglacial changes in ice load (Long et al., 2009).

An additional approach which has been used to obtain RSL data from the last millennium is to study salt marsh deposits in protected coastal locations in the southern half of Greenland. Marshes exist on the open coast from at least Disko Bugt

southward and record RSL rise over the past millennium in their stratigraphy (e.g., Woodroffe and Long 2009). Although the marshes have low sedimentation rates their sediments have survived under persistent RSL rise and contain precise evidence for RSL changes during the last few centuries. RSL reconstructions using a diatom-based transfer function approach on individual sediment profiles and multiple basal samples show that the RSL rise reconstructed from isolation basins continues until approximately AD 1600 in central west Greenland (Sisimiut), and between AD 1750 and 1950 in southern Greenland (Nanortalik) (Long et al., 2012)

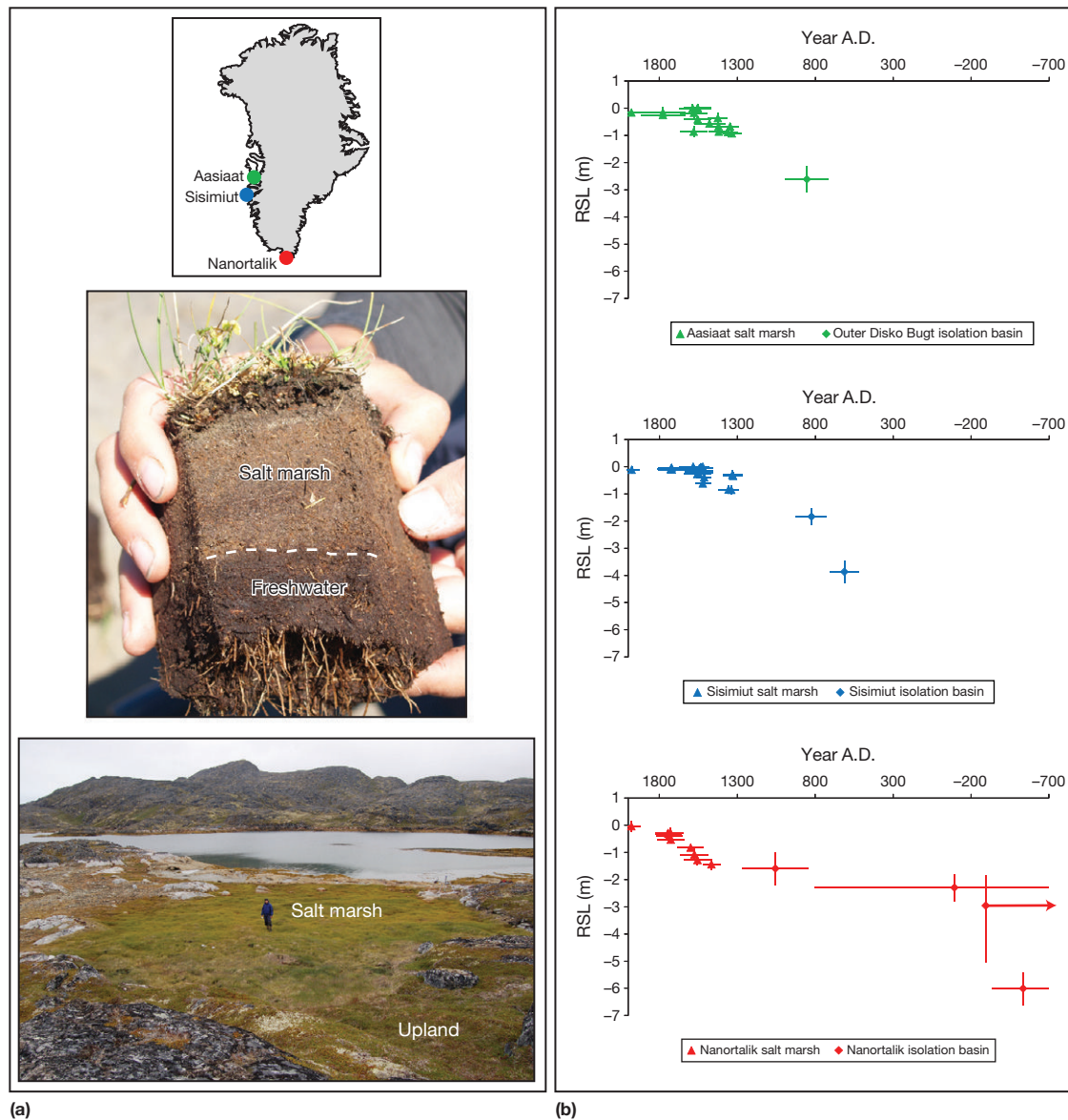


Figure 3 (a) Photograph of a typical thin salt marsh sediment sequence found in west Greenland, showing an upward transition from freshwater to salt marsh sediments formed by RSL rise over the past few hundred years. Also shown is a photograph of a modern salt marsh near Paamiut, in southwest Greenland. (b) RSL reconstructions from salt marshes at Aasiaat, Sisimiut, and Nanortalik together with isolation basin data from each area for the past 2700 years (modified from Long AJ, Woodroffe SA, Milne GA, Bryant CL, Simpson MJR, and Wake LM (2012) Relative sea-level change in Greenland during the last 700 years and ice sheet response to the Little Ice Age. *Earth and Planetary Science Letters* 315: 76–85, including isolation basin data from Long et al. 2003 (Outer Disko Bugt), Long et al., 2009 (Sisimiut) and Sparrenbom et al., 2006b (Nanortalik)).

(Figure 3). Since that time RSL has been stable, which is consistent with local ice mass loss in west Greenland (Wake et al., 2012).

Conclusions

Greenland is an important laboratory for the investigation of RSL change as it is the only location in the Northern Hemisphere where a continental-scale ice sheet which was present at the LGM remains today. Sedimentary evidence from before the LGM is rather limited, mainly because the expansion of the ice sheet onto the continental shelf in many areas during the LGM destroyed evidence of older deposits. However, borehole evidence from the ice sheet itself, notably from the Dye-3 drilling site in south Greenland, together with sediment cores from the offshore of southwest Greenland, strongly indicate a significantly smaller than present ice sheet during MIS 5e, contributing at least 2.4 m to global sea level (Kopp et al., 2009). During the Late Glacial and Holocene, more abundant deposits show that rapid glacio-isostatic rebound dominated in the early and mid-Holocene, followed by RSL fall below present and subsequent rise in the late Holocene. These typical J-shaped RSL curves are observed in many regions. Differences in the detail of the curves point to variations in regional to local-scale ice sheet history.

Long-term trends in Holocene RSL are useful in establishing millennial-scale patterns of ice sheet history and much emphasis to date has been on the early and mid-Holocene, since data from this period is the most abundant. A major scientific challenge is to extend these Holocene records up to the present-day, ideally so they can provide a longer term context for geodetic observations, notably GPS and gravity observations (e.g., Khan et al., 2010; Rignot et al., 2011), that provide so much concern regarding the ice sheet's response to recent climate change. In this area, Quaternary scientists have an important role to play in seeking to identify whether the patterns observed over short-time intervals, measured in years or a few decades, represent significant long-term trends or simply part of the variability that is inherent in a dynamic ice sheet.

See also: Dune Fields: High Latitudes. Plant Macrofossil Records: Greenland. Sea Level Studies: Eustatic Sea-Level Changes – Glacial–Interglacial Cycles; Isostasy: Glaciation-Induced Sea-Level Change.

References

- Alley RB, Andrews JT, Brigham-Grette J, et al. (2010) History of the Greenland Ice Sheet: Paleoclimatic insights. *Quaternary Science Reviews* 29: 1728–1756.
- Augustin L, Barbante C, Barnes PRF, et al. (2004) Eight glacial cycles from an Antarctic ice core. *Nature* 429: 623–628.
- Bennike O, Abrahamsen N, Bak M, et al. (2002a) A multi-proxy study of Pliocene sediments from Île de France, North-East Greenland. *Palaeogeography, Palaeoclimatology, Palaeoecology* 186: 1–23.
- Bennike O and Björck S (2002) Chronology of the last recession of the Greenland Ice Sheet. *Journal of Quaternary Science* 17: 211–219.
- Bennike O, Björck S, and Lambeck K (2002b) Estimates of South Greenland late-glacial ice limits from a new relative sea level curve. *Earth and Planetary Science Letters* 197: 171–186.
- Briner JP, Stewart HAM, Young NE, Philipps W, and Losee S (2010) Using proglacial-threshold lakes to constrain fluctuations of the Jakobshavn Isbrae ice margin, western Greenland, during the Holocene. *Quaternary Science Reviews* 29: 3861–3874.
- Bennike O, Wagner B, and Richter A (2011) Relative sea level changes during the Holocene in the Sisimiut area, south-western Greenland. *Journal of Quaternary Science* 26: 353–361.
- Carlson A, Stoner JS, Donnelly JP, and Hillaire-Marcel C (2008) Response of the southern Greenland Ice Sheet during the last two deglaciations. *Geology* 36: 359–362.
- Clark JA, Farrell WE, and Peltier WR (1978) Global changes in post glacial sea-level: A numerical equation. *Quaternary Research* 9: 265–287.
- Cuffey KM and Marshall SJ (2000) Substantial contribution to sea-level rise during the last interglacial from the Greenland ice sheet. *Nature* 404: 591–594.
- Dahl-Jensen D, Mosegaard K, Gundestrup N, et al. (1998) Past temperatures directly from the Greenland Ice Sheet. *Science* 282: 268–271.
- de Vernal A and Hillaire-Marcel C (2008) Natural variability of Greenland climate, vegetation, and ice volume during the past million years. *Science* 320: 1622–1625.
- Edgar KM, Wilson PA, Sexton PF, and Suganuma Y (2007) No extreme bipolar glaciation during the main Eocene calcite compensation shift. *Nature* 448: 908–911.
- Eldrett JS, Harding IC, Wilson PA, Butler E, and Roberts AP (2007) Continental ice in Greenland during the Eocene and Oligocene. *Nature* 446: 176–179.
- Funder S (1989) Quaternary geology of the ice free areas and adjacent shelves of Greenland. In: Fulton RJ (ed.) *Geology of Canada 1. Quaternary Geology of Canada and Greenland*, pp. 743–792. Ottawa: Geological Society of Canada.
- Funder S and Hansen L (1996) The Greenland ice sheet – A model for its culmination and decay during and after the Last Glacial Maximum. *Bulletin of the Geological Society of Denmark* 42: 137–152.
- Funder S, Hjort C, and Kelly M (1991) Isotope stage 5 (130–74 ka) in Greenland, a review. *Quaternary International* 10–12: 107–122.
- Funder S, Hjort C, Landvik JY, Nam SI, Reeh N, and Stein R (1998) History of a stable ice margin East Greenland during the Middle and Upper Pleistocene. *Quaternary Science Reviews* 17: 77–123.
- Funder S, Kjeldsen KK, Kjaer KH, and O' Cofaigh C (2011) The Greenland Ice Sheet during the last 300,000 years: A review. *Developments in Quaternary Science*, pp. 699–713. Amsterdam: Elsevier. <http://dx.doi.org/10.1016/B1978-1010-1444-53447-53447.00050-53447>.
- Funder S, Bennike O, Bocher J, Israelson C, Petersen KS, and Simonarson LA (2001) Late Pliocene Greenland – The Kap København Formation in North Greenland. *Bulletin of the Geological Society of Denmark* 48: 117–134.
- Johnsen SJ, Dahl-Jensen D, Gundestrup N, et al. (2001) Oxygen isotope and palaeotemperature records from six Greenland ice-core stations: Camp Century, Dye-3, GRIP, GISP2, Renland and NorthGRIP. *Journal of Quaternary Science* 16: 299–307.
- Kaufman DS and Brigham-Grette J (1993) Aminostratigraphic correlations and paleotemperature implications, Pliocene Pleistocene high sea-level deposits, Northwestern Alaska. *Quaternary Science Reviews* 12: 21–33.
- Kelly M (1980) The status of the Neoglacial in Western Greenland. *Rapport Grønlands Geologiske Undersøgelse* 96: 1–24.
- Kelly M (1985) A review of the Quaternary geology of western Greenland. In: Andrews JT (ed.) *Quaternary environments eastern Canadian Arctic, Baffin Bay and western Greenland*, pp. 461–501. Boston: Allen and Unwin.
- Khan SA, Liu L, Wahr J, et al. (2010) GPS measurements of crustal uplift near Jakobshavn Isbrae due to glacial ice mass loss. *Journal of Geophysical Research-Solid Earth* 115: B09405.
- Kopp RE, Simons FJ, Mitrovica JX, Maloof AC, and Oppenheimer M (2009) Probabilistic assessment of sea level during the last interglacial stage. *Nature* 462: 863–867.
- Kuijpers A, Abrahamsen A, Hoffmann G, et al. (1999) Climate change and the Viking age fjord environment of the Eastern Settlement, South Greenland. *Geology of Greenland Survey Bulletin* 183: 61–67.
- Lasca NP (1966) Post-glacial deleveling in Skeldal, northeast Greenland. *Arctic* 19: 349–353.
- Lemke P, Ren J, Alley RB, et al. (2007) Climate Change 2007: The Physical Science Basis. In: Solomon S, Qin D, Manning M, Chen Z, Marquis M, Averyt KB, Tignor M, and Miller HL (eds.) Cambridge and New York: Cambridge University Press Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change.
- Long AJ (2009) Back to the future: Greenland's contribution to sea level change. *Geological Society of America Bulletin* 19: 4–10.
- Long AJ and Roberts DH (2003) Late Weichselian deglacial history of Disko Bugt, West Greenland, and the dynamics of the Jakobshavn Isbrae ice stream. *Boreas* 32: 208–226.

- Long AJ, Roberts DH, and Dawson S (2006) Early Holocene history of the West Greenland Ice Sheet and the GH-8.2 event. *Quaternary Science Reviews* 25: 904–922.
- Long AJ, Roberts DH, and Rasch M (2003) New observations on the relative sea level and deglacial history of Greenland from Innaarsuit, Disko Bugt. *Quaternary Research* 60: 162–171.
- Long AJ, Roberts DH, Simpson MJR, Dawson S, Milne GA, and Huybrechts P (2008) Late Weichselian relative sea-level changes and ice sheet history in southeast Greenland. *Earth and Planetary Science Letters* 272: 8–18.
- Long AJ, Roberts DH, and Wright MR (1999) Isolation basin stratigraphy and Holocene relative sea-level change on Arveprinsen Ejland, Disko Bugt, West Greenland. *Journal of Quaternary Science* 14: 323–345.
- Long AJ, Woodroffe SA, Dawson S, Roberts DH, and Bryant CL (2009) Late Holocene relative sea-level rise and the Neoglacial history of the Greenland ice sheet. *Journal of Quaternary Science* 24: 345–359.
- Long AJ, Woodroffe SA, Milne GA, Bryant CL, Simpson MJR, and Wake LM (2012) Relative sea-level change in Greenland during the last 700 years and ice sheet response to the Little Ice Age. *Earth and Planetary Science Letters* 315: 76–85.
- Long AJ, Woodroffe SA, Milne GA, Bryant CL, and Wake LM (2010) Relative sea-level change in West Greenland during the last millennium. *Quaternary Science Reviews* 29: 367–383.
- Long AJ, Woodroffe SA, Roberts DH, and Dawson S (2011) Isolation basins, sea-level changes and the Holocene history of the Greenland Ice Sheet. *Quaternary Science Reviews in review* 30: 3748–3768.
- Lunt DJ, Foster GL, Haywood AM, and Stone EJ (2008) Late Pliocene Greenland glaciation controlled by a decline in atmospheric CO₂ levels. *Nature* 454: 1102–U1141.
- Maslin MA, Li XS, Loutre MF, and Berger A (1998) The contribution of orbital forcing to the progressive intensification of Northern Hemisphere glaciation. *Quaternary Science Reviews* 17: 411–426.
- Miller G, Anderson P, Bennike O, et al. (2006) Last interglacial Arctic warmth confirms polar amplification of climate change. *Quaternary Science Reviews* 25: 1383–1400.
- Olson SL and Hearty PJ (2009) A sustained +21 m sea-level highstand during MIS 11 (400 ka): Direct fossil and sedimentary evidence from Bermuda. *Quaternary Science Reviews* 28: 271–285.
- Otto-Bliesner BL, Marsha SJ, Overpeck JT, Miller GH, Hu AX, and Mem CLIP (2006) Simulating arctic climate warmth and icefield retreat in the last interglaciation. *Science* 311: 1751–1753.
- Rasch M (2000) Holocene relative sea level changes in Disko Bugt, West Greenland. *Journal of Coastal Research* 16: 306–315.
- Rasch M and Jensen JF (1997) Ancient Eskimo dwelling sites and Holocene relative sea-level changes in southern Disko Bugt, central West Greenland. *Polar Research* 16: 101–115.
- Rasch M and Nielsen N (1995) Coastal morpho-stratigraphy and Holocene relative sea level changes at Tuapaat, southeastern Disko Island, central West Greenland. *Polar Research* 14: 277–289.
- Rignot E, Velicogna I, van den Broeke MR, Monaghan A, and Lenaerts J (2011) Acceleration of the contribution of the Greenland and Antarctic ice sheets to sea level rise. *Geophysical Research Letters* 38: L05503–L05508.
- Roberts DH, Long AJ, Schnabel C, Freeman S, and Simpson MJR (2008) The deglacial history of southeast sector of the Greenland Ice Sheet during the Last Glacial Maximum. *Quaternary Science Reviews* 27: 1505–1516.
- Robinson A, Calov R, and Ganopolski A (2011) Greenland ice sheet model parameters constrained using simulations of the Eemian Interglacial. *Climate of the Past* 7: 381–396.
- Rohling EJ, Braun K, Grant K, et al. (2010) Comparison between Holocene and Marine Isotope Stage-11 sea-level histories. *Earth and Planetary Science Letters* 291: 97–105.
- Roussel A (1941) Farms and churches in the medieval Norse settlements of Greenland. *Meddelelser om Grønland* 89: 342.
- Simpson MJR, Milne GA, Huybrechts P, and Long AJ (2009) Calibrating a glaciological model of the Greenland ice sheet from the Last Glacial Maximum to present-day using field observations of relative sea level and ice extent. *Quaternary Science Reviews* 28: 1631–1657.
- Sparrenbom CJ, Bennike O, Björck S, and Lambeck K (2006a) Relative sea-level changes since 15 000 cal. yr BP in the Nanortalik area, southern Greenland. *Journal of Quaternary Science* 21: 29–48.
- Sparrenbom C, Lambeck K, Bennike O, and Björck S (2006b) Holocene relative sea-level changes in the Qaqortoq area, southern Greenland. *Boreas* 35: 171–187.
- Tarasov L and Peltier WR (2003) Greenland glacial history, borehole constraints, and Eemian extent. *Journal of Geophysical Research – Solid Earth* 108: 2143.
- Ten Brink NW and Weidick A (1975) Holocene history of the Greenland Ice Sheet based on radiocarbon-dated moraines in West Greenland. *Grønlands Geologiske Undersøgelse Bulletin* 114: 1–44.
- Thomas RH (2001) Program for arctic regional climate assessment (PARCA): Goals, key findings, and future directions. *Journal of Geophysical Research-Atmospheres* 106: 33691–33705.
- Wake LM, Milne GA, Long AJ, Woodroffe SA, Simpson MJR, and Huybrechts P (2012) Century-scale relative sea-level changes in west Greenland – A plausibility study to assess contributions from the cryosphere and the ocean. *Earth and Planetary Science Letters* 315: 86–93.
- Washburn AL and Stuiver M (1962) Radiocarbon-dated post-glacial de-leveling in northeast Greenland and its implications. *Arctic* 15: 66–73.
- Weidick A (1968) Observations on some Holocene glacier fluctuations in West Greenland. *Meddelelser om Grønland* 165: 202.
- Weidick A (1972) Holocene shore-lines and glacial stages in Greenland – An attempt at correlation. *Grønlands Geologiske Undersøgelse* 41: 1–39.
- Weidick A (1993) Neoglacial change of ice cover and the related response of the Earth's crust in West Greenland. *Rapport Grønlands Geologiske Undersøgelse* 159: 121–126.
- Weidick A (1996) Neoglacial changes of ice cover and sea level in Greenland – A classical enigma. In: Grønnow B (ed.) *The Palaeo-Eskimo Cultures of Greenland: New Perspectives in Greenlandic Archaeology*, pp. 257–270. Copenhagen: Danish Polar Center.
- Weidick A and Bennike O (2007) Quaternary glaciation history and glaciology of Jakobshavn Isbræ and the Disko Bugt region, West Greenland: A review. *Geological Survey of Denmark and Greenland Bulletin* 14: 1–78.
- Willerslev E, Cappellini E, Boomsma W, et al. (2007) Ancient biomolecules from deep ice cores reveal a forested Southern Greenland. *Science* 317: 111–114.
- Woodroffe SA and Long AJ (2009) Salt marshes as archives of recent relative sea-level change in West Greenland. *Quaternary Science Reviews* 28: 1750–1761.

Late Quaternary Relative Sea-Level Changes at Mid-Latitudes

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Introduction

The dominant influence on relative sea-level (RSL) changes since the Last Glacial Maximum (LGM; 26 ka BP) has been the melting of approximately 50 million km³ of land-based ice as Earth's climate transitioned from glacial to interglacial conditions. Transfer of this mass to global ocean basins increased ocean water volume and caused a large and ongoing isostatic response of the solid Earth ([Figure 1](#)). The largest RSL changes occurred in regions such as Canada and Fennoscandia, which were beneath continental-scale ice sheets at the LGM. These *near-field* regions are characterized by continuous RSL fall, because the rate of glacio-isostatic uplift exceeded the rate of eustatic sea-level rise. Depression of land beneath LGM ice sheets was compensated for by migration of mantle material away from ice-load centers that caused uplift of a forebulge in regions peripheral to ice sheets. These *intermediate-field* regions (e.g., mid-Atlantic coast of the United States) also have deglacial RSL histories influenced by glacio-isostasy. As land-based ice diminished, mantle material returned to the former load centers causing collapse and retreat (glacio-isostatic subsidence) of the forebulge. Thus, isostatic and eustatic effects worked together to cause RSL rise. Reduction in meltwater input after 7 ka BP slowed RSL rise and increased the relative contribution from glacio-isostatic subsidence. Ongoing subsidence in these regions caused late Holocene RSL rise in the absence of a eustatic contribution. At increasing distances from major ice centers, eustatic contributions to RSL change exceeded glacio-isostatic contributions. Regions not significantly influenced by glacio-isostasy are termed *far field*. At tectonically stable far-field locations, RSL changes are an approximation of the eustatic component. However, a number of processes including hydro-isostatic subsidence of ocean basins and perturbations to Earth's rotation driven by mass redistribution cause RSL changes in far-field regions to depart from the eustatic value.

[Clarke et al. \(1978\)](#) seminal paper identified five principal forms of the sea-level curve (plus a transitional form), reflecting variable RSL histories on near-, intermediate-, and far-field coastlines ([Figure 2](#)). These predicted curves provide a general description of deglacial RSL changes but do not reflect the uncertainty associated with geological reconstructions of RSL. The original work of [Clark et al. \(1978\)](#) has been extended to incorporate recent developments in the understanding of the history of eustatic contributions, lateral shifts in shoreline position associated with local sea-level changes, growth and retreat of marine-based ice sheets, feedback from glaciation-induced perturbations in Earth rotation, and gravitational changes in sea level caused by redistribution of mass ([Milne and Mitrovica, 2008](#)).

Late Quaternary Sea Levels

We summarize RSL data from passive coastal margins at mid-latitudes to show patterns of RSL changes since the LGM. We also draw attention to recent advances in the understanding of RSL changes during the past 2 ka. In some instances, processes that may have caused the RSL changes are described. Selected studies from the Atlantic coast of North and South America, Europe, China, South Africa, and Australasia are highlighted as representative examples of regional RSL histories that correspond to [Clark et al. \(1978\)](#) zones I, II, and V. The examples were drawn purposefully from studies published in the last 20 years and are intended as an introduction to Holocene sea-level histories. References to older studies (and ones from other regions) can be found in this modern body of literature. In the examples used, RSL data comprise sea-level index points that estimate the unique position of RSL in space and time ([Engelhart et al., 2009](#)) and are used to describe the form of RSL changes and estimate rates of change. The interpretation of index points was not changed from that in the original studies. Negative values indicate a RSL below the present level. Where necessary, we have converted radiocarbon ages into 1000 calibrated years (ka) before present (BP) where year 0 is conventionally taken to be AD 1950. For clarity, we describe most ages to the nearest 0.1 ka without chronological uncertainty.

Atlantic Coast of North America

The *Atlantic coast of Canada* and *Gulf of Maine* include near- and intermediate-field regions. [Shaw et al. \(2002\)](#) compiled sea-level index points from Atlantic Canada since the LGM. Newfoundland experienced continuous RSL fall because of isostatic uplift caused by ice unloading. RSL reconstructions from Labrador also showed falling RSL throughout the post-glacial period from a maximum marine limit of +120 m at 16 ka BP. The exception is the northern Labrador Peninsula where RSL fell below 0 m at 7.7 ka BP. In Nova Scotia, there was a RSL lowstand of –65 m at 13.5 ka BP, since when RSL has risen at a decreasing rate through the Holocene ([Shaw et al., 2002](#)). A study of sea-level changes during the past 1 ka using Foraminifera preserved in salt-marsh sediment showed that RSL in Nova Scotia rose at 1.6–1.7 mm year^{–1} until the start of the twentieth century when the rate increased to 3.2 mm year^{–1} ([Gehrels et al., 2005](#)).

Isostatic rebound dominated the RSL history of the Gulf of Maine ([Figure 3\(a\)](#)). Retreat of the Laurentide Ice Sheet exposed large areas of isostatically subsided land that were rapidly submerged ([Kelley et al., 2010](#)). Maximum RSL occurred at 17 ka BP and was 70–129 m above the present level.

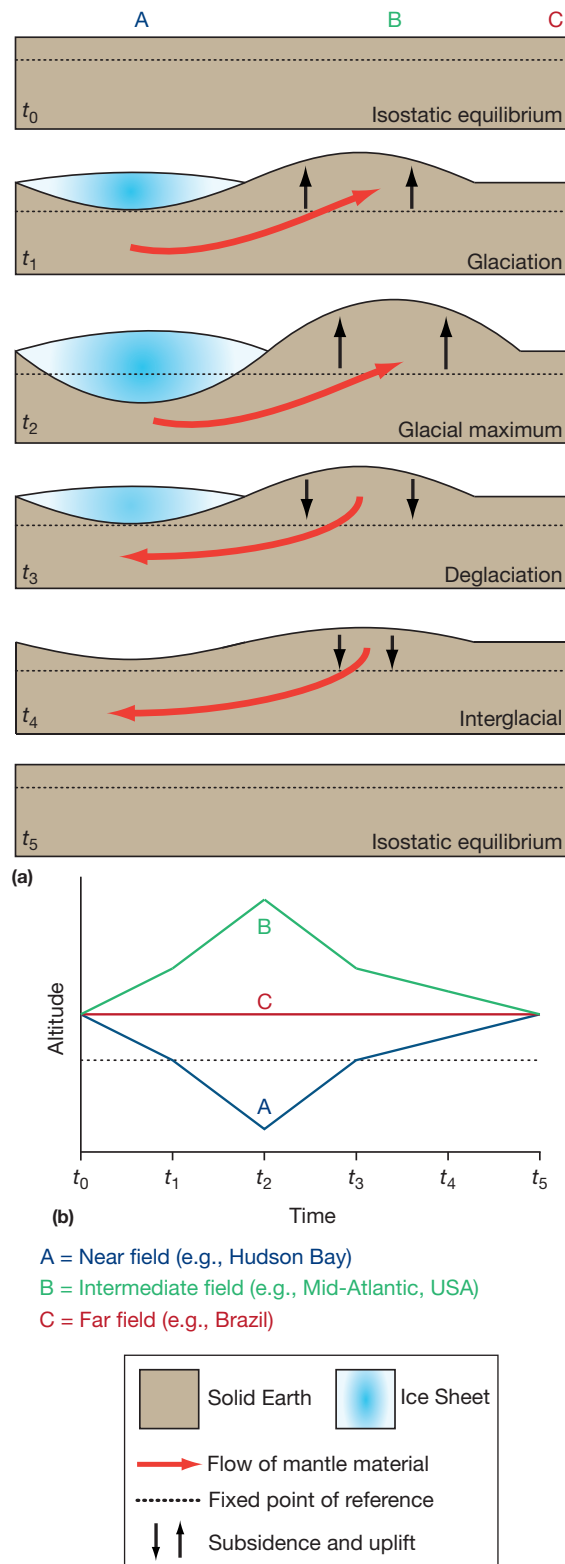


Figure 1 Schematic representation of glacial isostatic adjustment: the response of the solid earth to loading by an ice sheet as observed at near-, intermediate-, and far-field sites. (a) Commencing from isostatic equilibrium, the addition of an ice sheet causes the land below it to subside and pushes a peripheral bulge away from the load. On deglaciation, the mantle material flows back toward the former ice-sheet

Glacio-isostatic uplift caused RSL to fall to a lowstand of -60 to -65 m at 12 ka BP. This lowstand was brief and RSL rose rapidly to -20 m at 11.5 ka BP. RSL remained at this level for 3.5 ka: a period termed the sea-level 'slowstand' (Kelley et al., 2010). Subsequently, the rate of glacio-isostatic rebound slowed and was surpassed by eustatic contributions causing RSL rise. Sea-level rise over the past 4 ka was less than 1 mm year^{-1} (Engelhart et al., 2009). High-resolution reconstructions show that this late Holocene rate of RSL rise, driven solely by isostatic factors, was exceeded in the nineteenth century, since when RSL has risen by 0.3–0.4 m (Gehrels et al., 2002).

The Atlantic coast of the United States from New Hampshire to South Carolina is an intermediate-field region. Glacio-isostatic contributions were much reduced in magnitude compared to those of Atlantic Canada or Maine but remain the dominant influence. The coastline in this region is subsiding in response to collapse and retreat of the Laurentide Ice Sheet forebulge and return flow of mantle material toward formerly glaciated areas. RSL records are characterized by continual rise during the Holocene, with the rate slowing through time because of continued, but decreasing, glacio-isostatic subsidence. In Massachusetts, RSL was -120 m at the LGM and -65 m at 13.5 ka BP. It subsequently rose at approximately 6 mm year^{-1} to -7 m at 4.5 ka BP (Uchupi et al., 2005). In the late Holocene, the rate of RSL rise was reduced to 2 mm year^{-1} . Using salt-marsh sediments, Donnelly (2006) showed RSL to have been -2.6 m at 3.3 ka BP, after which it rose by 0.8 mm year^{-1} until 1 ka BP and since then by 0.5 mm year^{-1} . In Connecticut, a late Holocene rise of 1 mm year^{-1} , which was driven by isostatic subsidence, was punctuated by a sea-level oscillation at 1 ka BP (Medieval Climate Anomaly) when sea level was 25 ± 25 cm higher than AD 1650 (Little Ice Age; van de Plassche, 2000). Donnelly et al. (2004) identified the onset of recent, higher rates of rise (3 mm year^{-1}) at around AD 1850.

Several studies have reconstructed Holocene RSL in New Jersey and Delaware. Miller et al. (2009) showed that RSL in New Jersey rose from -18 m at 8.5 ka BP to -9 m at 5 ka BP. Thereafter, RSL rose at 1.8 mm year^{-1} until 0.5 ka BP. In Delaware, RSL was -27 m at 12 ka BP (Nikitina et al., 2000). RSL rose at more than 3 mm year^{-1} between 7 and 5 ka BP, before slowing down to 2.1 mm year^{-1} from 5 to 2 ka BP and 0.9 mm year^{-1} from 1.3 ka BP to AD 1900. RSL rise increased to 3.3 mm year^{-1} in the twentieth century (Nikitina et al., 2000). Engelhart et al. (2009) compilation of sea-level index points showed that New Jersey and Delaware experienced the highest rates of RSL rise on the Atlantic coast of the United States during the past 4 ka (up to 1.7 mm year^{-1}), because they are located near the center of proglacial forebulge collapse.

Horton et al. (2009) showed that RSL in North Carolina rose from -36 m at 11 ka BP to -4 m at 4 ka BP, which equated to an average rate of rise during the early and mid-Holocene of 5 mm year^{-1} (Figure 3(b)). RSL rise slowed in the late Holocene to 1 mm year^{-1} (Engelhart et al., 2009). A study of RSL changes during the past 0.5 ka showed that the late

load until equilibrium is reached once more. (b) The isostatically induced land movement at near-, intermediate-, and far-field sites associated with the glacial and deglacial cycle in (a).

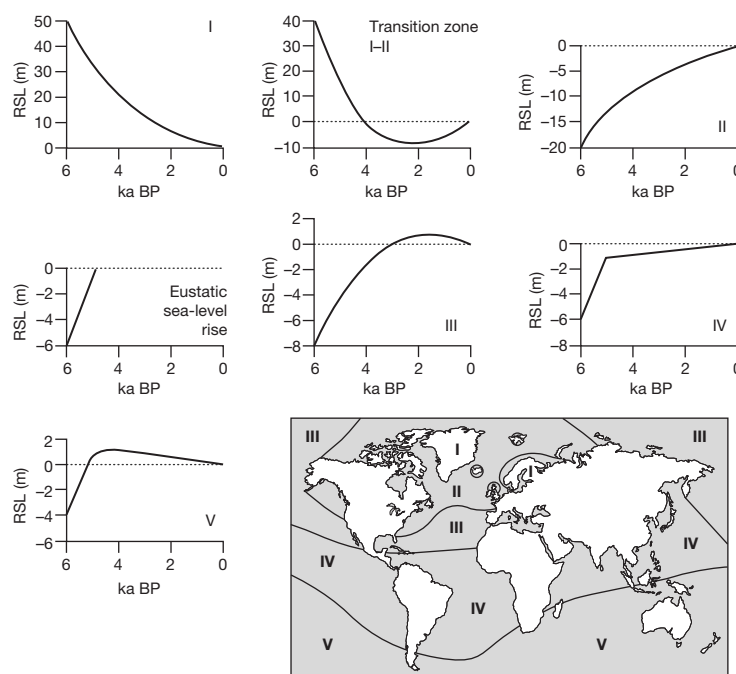


Figure 2 Relative sea-level (RSL) curves predicted by Clark et al. (1978). Under an assumption of eustatic input ending at 4 ka BP, five sea-level zones (I–V) plus a transitional zone (I–II) were recognized. All x-axes are 1000 calibrated years before present (ka BP).

Holocene, isostatically driven, rate of rise has been exceeded since the end of the nineteenth century (Kemp et al., 2009). Further south along the Atlantic coast of the United States, the contribution from glacio-isostatic subsidence is reduced in magnitude. This can be seen in the RSL history of southern *South Carolina*, where the rate of late Holocene sea-level rise was 0.6 mm year^{-1} (Engelhart et al., 2009).

Europe

RSL reconstructions from mid-latitudes in Europe define areas of postglacial isostatic rebound (e.g., Sweden and Scotland) and subsidence (e.g., the Netherlands and Germany). The Baltic Ice Lake (a proto-Baltic Sea) formed around 14 ka BP following ice margin readvance in northern *Estonia* (Raukas, 2000). Retreat of continental ice lowered the Baltic Ice Lake level by 25–30 m at 12.5 ka BP. RSL rose to 42–45 m above the present level at 10 ka BP and subsequently fell by 30–35 m between 9 and 8 ka BP (Raukas, 2000). Eustatic contributions exceeded the rate of glacio-isostatic land uplift from 8.4 to 6.8 ka BP, causing RSL rise (Raukas, 2000). Isolation basins in southeastern Sweden (Berglund et al., 2005) recorded RSL fall at 8.1 ka BP, which was correlated with the 8.2 ka BP climate event. This event was an abrupt, widespread climate instability that Berglund et al. (2005) suggested was associated with a positive phase of North Atlantic Oscillation leading to high atmospheric pressure over the Baltic Sea with strong northern winds and lowered mean sea level. Yu et al. (2007) identified an abrupt (occurring over less than 200 years) RSL rise of 4.5 m at 7.6 ka BP in southeastern Sweden likely caused by the final decay of the Labrador sector of the Laurentide Ice Sheet (Figure 3(c)). Since the mid-Holocene, there was

continuous RSL fall in the southern Baltic due to slow isostatic adjustment (uplift) and limited eustatic input.

Owing to its location at the margin of the former Fennoscandian Ice Sheet, northern *Denmark* experienced RSL fall since the LGM, while areas to the south experienced RSL rise. Clemmensen et al. (2001) reconstructed Holocene RSL for Skagen at the northern tip of Denmark. RSL fell from +12 to +1 m between 5.6 and 1.4 ka BP. In Ho Bugt (western Denmark), Gehrels et al. (2006) used basal salt-marsh peats to show that 4 m of RSL rise has occurred since 4 ka BP. Numerical modeling of glacial isostatic adjustment in this region suggests that this rise can be adequately explained by land subsidence alone. Szkornik et al. (2008) suggested that this rise was punctuated by two periods with greater than average rates of sea-level rise. The first was between 2 and 1.4 ka BP and the second commenced at AD 1500. *Baltic Germany* lies beyond the margins of the former Fennoscandian Ice Sheet. RSL rose rapidly there (20 mm year^{-1}) from 8 to 5.7 ka BP, after which it rose gradually to present (Meyer and Harff, 2005).

Shennan and Horton (2002) provided a synthesis of RSL data and proposed summary RSL curves for 52 areas in the *United Kingdom*. This dataset was recently reviewed by Gehrels (2010). Although small in global terms, the British Ice Sheet was large enough for glacial isostatic adjustment to produce spatially variable RSL histories. At Arisaig (northwest Scotland), which was covered by relatively thick ice (900 m) at the LGM, RSL fell from a marine limit between +36 and +40 m at 16 ka BP to an early Holocene minimum (11 ka BP) of +3 m (Figure 3(d)). RSL reached a mid-Holocene highstand (+6.5 m) that persisted from 8 to 5 ka BP. In the United Kingdom, there were clear north–south variations in RSL. For example, along the east coast of England, at

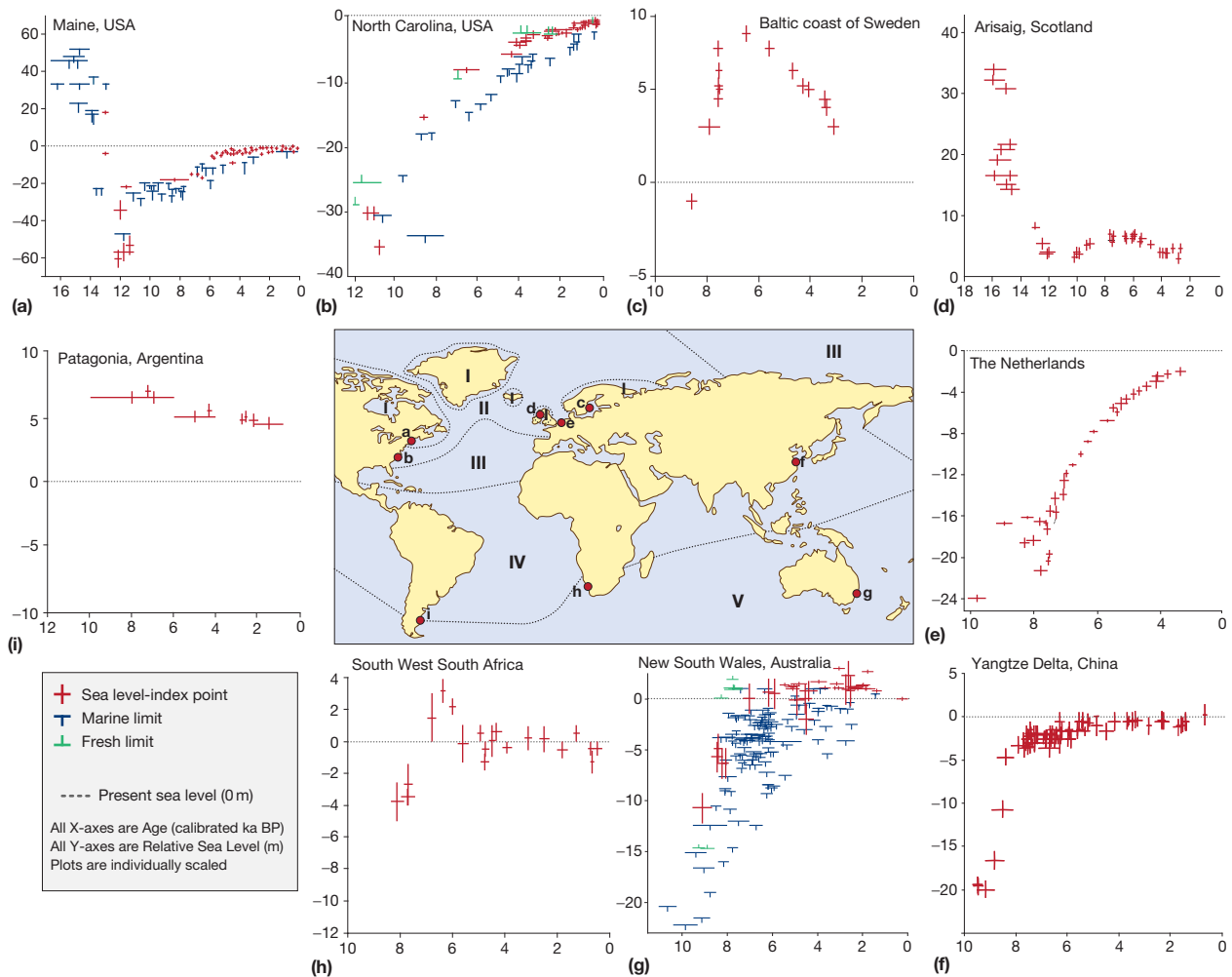


Figure 3 Selected relative sea-level reconstructions from mid-latitudes: (a) Maine, USA (Kelley et al., 2010); (b) North Carolina, USA (Horton et al., 2009); (c) Baltic coast of Sweden (Yu et al., 2007); (d) Arisaig, Scotland (Shennan and Horton, 2002); (e) The Netherlands (Hijma and Cohen, 2010); (f) Yangtze Delta, China (Zong, 2004); (g) New South Wales (NSW), Australia (Sloss et al., 2005); (h) southwest South Africa (Compton, 2001); (i) Patagonia, Argentina (Rostami et al., 2000). The spatial areas of Clarke et al. (1978) five types of 'sea-level curve' (I–V) are shown. Altitudes and interpretation of sea-level index points (including errors) have not been amended from the original publication. All y-axes are relative sea level (meters to present). Radiocarbon ages were converted to calibrated dates where necessary, all x-axes are in 1000 calibrated years before present (ka BP). Index points are represented as + symbols, and in some instances, errors may be plotted larger than the original publication to aid in clear visual presentation (e.g., Maine).

8 ka BP, RSL in Northumberland was 20 m higher than that in Norfolk. Sea-level index points from Northumberland showed rising RSL from -5 m at 8.6 ka BP to a mid-Holocene maximum of $+2.5$ m at 3–4 ka BP. In Norfolk, RSL rose from -13.5 m at 7.5 ka BP to -4.0 m at 5.3 ka BP, which equates to a mean rate of 4 mm year^{-1} . The maximum late Holocene glacio-isostatic uplift (1.7 mm year^{-1}) occurred in central and western Scotland, while the maximum subsidence (1 mm year^{-1}) occurred in the southwest of England (Gehrels, 2010).

Using a similar database approach, Brooks et al. (2008) collated sea-level index points for Ireland. The data displayed a coherent and consistent pattern of RSL change in southern Ireland driven by forebulge collapse. For example, RSL in East Cork rose from -18 m at 9 ka BP to -5 m at 7 ka, followed by a slow rise in RSL of 0.7 mm year^{-1} to present. In northeast

Ireland, RSL history was more complicated because of variable isostatic influence.

In the southern North Sea, RSL has risen continuously since the LGM. RSL was 110–130 m below the present level at the LGM, which exposed much of the North Sea basin as dry land (Streif, 2004). RSL rose from -72 m at 12.5 ka BP to -25 m at 7.5 ka BP; the average rate of RSL rise was 15 mm year^{-1} but was 21 mm year^{-1} between 9.9 and 7.7 ka BP, after which RSL rose at a decreasing rate to present (Streif, 2004).

On the Belgian coastal plain, Denys and Baeteman (1995) reconstructed Holocene RSL rise during deglaciation using basal salt-marsh peats and showed rates of 7 mm year^{-1} from 9.4 to 7.5 ka BP. This caused rapid transgression of the continental shelf toward a position close to the present-day boundary of the coastal plain. After 7.5 ka BP, the rate of rise

decreased to 2.5 mm year^{-1} with a further fall to 0.7 mm year^{-1} since 5.5–5 ka BP (Denys and Baeteman, 1995). Hijma and Cohen (2010) suggested that in the Netherlands, RSL rose at $9\text{--}10 \text{ mm year}^{-1}$ from 9 to 7.5 ka BP, with a jump in RSL of 4 m commencing at 8.45 ka BP and lasting less than 200 years (Figure 3(e)). Catastrophic outflow of proglacial lakes Agassiz and Ojibway through the Hudson Strait was proposed as the cause of this jump (Hijma and Cohen, 2010).

Sea-level reconstructions from Atlantic France showed RSL rise prior to 7.5 ka BP, with no evidence of RSL being above the present level (Allard et al., 2008). In Portugal, RSL was -100 m at 16 ka BP and rose to -40 m between 12 and 11 ka BP (Dias et al., 2000). Boski et al. (2008) indicated that during the Holocene, RSL rose by 8.5 mm year^{-1} to approximately 6.5 ka BP and 3 mm year^{-1} until 5 ka BP, when RSL approached the present level. In Spain, RSL rose at $0.25 \text{ mm year}^{-1}$ after 0.5 ka BP because of isostatic adjustment until a change to modern rates of rise at the start of the twentieth century (Leorri et al., 2008).

China

Reconstructions from China indicate a complicated history of Holocene RSL changes, including debate about the existence of a mid-Holocene highstand (Zong, 2004). Sea-level index points from mid-latitude locations in China (southern edge of the Yangtze Delta plain and the Taihu area) were compiled by Zong (2004). RSL in the Yangtze Delta region rose from -20 m at 9.5 ka BP to -5 m at 8 ka BP, since when the rate of RSL rise has been less (1 mm year^{-1}). Since 3 ka BP, RSL has been close to 0 m (Figure 3(f)).

Australasia

Australia has been an important far-field location for RSL reconstruction, including the first widely proposed eustatic sea-level curve and early debates on hydro-isostasy. RSL reconstructions from mid-latitudes including New South Wales (Figure 3(g)) showed that RSL rose from -120 m at the LGM to $+1 \text{ m}$ at 7.2 ka BP (Sloss et al., 2005). In South Australia and Tasmania, RSL had risen sufficiently by 20 ka BP to enter the Bass Basin and form an estuarine environment that isolated Tasmania from the Australian mainland at 17 ka BP (Lambeck and Chappell, 2001). In Western Australia, RSL was between -50 and -40 m at 11.5–13 ka BP (Lessa and Masselink, 2006).

A prominent feature of Australian Holocene RSL records is the mid-Holocene highstand. Lewis et al. (2008) compilation of RSL data from eastern Australia using coral microatolls and fixed biological indicators (oyster beds and tubeworms) suggested that RSL reached a maximum of $+1.0$ to $+1.5 \text{ m}$ at 7 ka BP. In southeastern Australia, the maximum RSL was $+1.5 \text{ m}$ at 7.5 ka BP (Sloss et al., 2005). Sites on the south coast of Western Australia recorded a maximum RSL of $+1 \text{ m}$ at 7.5 ka BP (Lessa and Masselink, 2006).

RSL fall from the highstand was caused by ocean siphoning (Milne et al., 2005). Along the eastern Australian coast, Lewis et al. (2008) suggested, the fall included two $0.3\text{--}1.0 \text{ m}$ oscillations at 4.6 and 2.9 ka BP. In contrast, Sloss et al. (2005) suggested that RSL remained within 0.5 m of the highstand for 6.5 ka before a gradual fall to 0 m at 2 ka BP (Figure 3(g)).

Similarly, Lessa and Masselink (2006) suggest that RSL maintained its highstand elevation for much of the middle and late Holocene before falling to the present level since 2.5 ka BP.

The most complete Holocene RSL record for New Zealand is based on data from tectonically 'quiet' areas near Dunedin and Auckland (Hayward et al., 2002). Sea level rose from -25 m at 9 ka BP to reach 0 m at 7 ka BP. Since then, RSL has been no more than 1 m above and no less than 0.5 m below the present level. A reconstruction of RSL since 0.5 ka BP using Foraminifera preserved in salt-marsh sediments showed that RSL in New Zealand rose slowly ($0.3 \pm 0.3 \text{ mm year}^{-1}$) until the twentieth century when the rate increased to 2.8 mm year^{-1} (Gehrels et al., 2008).

South Africa

Being remote from Quaternary ice sheets, the South African coast experienced only minor isostatic movements. A Holocene RSL record developed from beachrock showed that RSL rose at 8 mm year^{-1} between 10.5 and 8.5 ka BP, reaching 0 m at 7.4 ka BP (Ramsay, 1996). It then rose to a mid-Holocene highstand of $+3.5 \text{ m}$ at 5.5 ka BP. Ramsay (1996) suggested that RSL fell to -2 m at 3.2 ka BP and subsequently rose to $+1.5 \text{ m}$ at 1.5 ka BP before falling to 0 m at 0.9 ka BP. This oscillation in the late Holocene RSL record may have been due to ocean siphoning and warmer ocean temperatures of the eastern Agulhas Bank (Ramsay, 1996). Compton's (2001) investigation of salt-marsh lagoons on the southwest coast of South Africa supported the presence of a mid-Holocene highstand with similar 1-m-scale RSL fluctuations in the late Holocene (Figure 3(h)).

South America

There have been two recent analyses of RSL data from the Atlantic coast of South America. Peltier (2002) demonstrated that this region was strongly influenced by rotational feedback, causing an elevated mid-Holocene highstand. Using mollusk shells from raised beaches that were dated using uranium-series, electric spin resonance, and radiocarbon methods, Rostami et al. (2000) produced a RSL record for Patagonia that showed a highstand of $+6 \text{ m}$ at 8 ka BP, with a gradual fall to present (Figure 3(i)). Milne et al. (2005) divided a RSL dataset of radiocarbon-dated organic sediments and shells into regions to identify spatial trends in Patagonia. Data from the Strait of Magellan showed RSL to be $+1 \text{ m}$ at 9 ka BP, with limited change ($\pm 1 \text{ m}$) to 5 ka BP. RSL data from the Beagle Channel demonstrated a highstand of $+6 \text{ m}$ at 6 ka BP, followed by a fall to 0 m by 3 ka BP.

Conclusion

RSL data from mid-latitudes reveal spatial and temporal changes caused by the varying dominance of eustatic and isostatic (glacio and hydro) factors since the LGM. Near-field regions (e.g., Atlantic coast of Canada, Sweden, and Scotland) are characterized by a complex sea-level fall from a maximum marine limit, reflecting the interplay between eustatic sea-level rise and glacial-isostatic uplift. In intermediate-field regions

such as mid-Atlantic coast of the United States, isostatic and eustatic effects combined to cause RSL rise. Recent research in northwestern Europe (e.g., the Netherlands) has identified rapid RSL rises and falls during the Holocene. A feature of sea-level records from mid-latitude, far-field locations is the mid-Holocene highstand, which differs in timing and magnitude across sites from Australasia, South Africa, and South America. Detailed RSL reconstructions from salt marshes have led to an improved understanding of sea level during the last 2000 years. A common feature of these records is an increase to modern rates of RSL rise that was likely initiated at the end of the nineteenth or early twentieth century.

See also: [Sea Level Studies: Eustatic Sea-Level Changes Since the Last Glacial Maximum](#); [Isostasy: Glaciation-Induced Sea-Level Change](#).
Sea-Levels, Late Quaternary: [Late Quaternary Relative Sea-Level Changes in High Latitudes](#); [Late Quaternary Relative Sea-Level Changes in the Tropics](#).

References

- Allard J, Chaumillon É, Poirier C, Sauriau P-G, and Weber O (2008) Evidence of former Holocene sea level in the Marennes-Oléron Bay (French Atlantic coast). *Comptes Rendus Geosciences* 340: 306–314.
- Berglund BE, Sandgren P, Barnekow L, et al. (2005) Early Holocene history of the Baltic Sea, as reflected in coastal sediments in Blekinge, southeastern Sweden. *Quaternary International* 130–139: 111.
- Boski T, Camacho S, Moura D, et al. (2008) Chronology of the sedimentary processes during the postglacial sea level rise in two estuaries of the Algarve coast, Southern Portugal. *Estuarine, Coastal and Shelf Science* 77: 230–244.
- Brooks AJ, Bradley SL, Edwards RJ, Milne GA, Horton B, and Shennan I (2008) Postglacial relative sea-level observations from Ireland and their role in glacial rebound modelling. *Journal of Quaternary Science* 23: 175–192.
- Clark JA, Farrell WE, and Peltier WR (1978) Global changes in post-glacial sea-level – Numerical calculation. *Quaternary Research* 9: 265–287.
- Clemmensen LB, Richardt N, and Andersen C (2001) Holocene sea-level variation and spit development: Data from Skagen Odde, Denmark. *The Holocene* 11: 323–331.
- Compton JS (2001) Holocene sea-level fluctuations inferred from the evolution of depositional environments of the southern Langebaan Lagoon salt marsh, South Africa. *The Holocene* 11: 395–405.
- Denys L and Baeteman C (1995) Holocene evolution of relative sea level and local mean high water spring tides in Belgium – A first assessment. *Marine Geology* 124: 1–19.
- Dias JMA, Boski T, Rodrigues A, and Magalhães F (2000) Coast line evolution in Portugal since the Last Glacial Maximum until present – A synthesis. *Marine Geology* 170: 177–186.
- Donnelly JP (2006) A revised late Holocene sea-level record for northern Massachusetts, USA. *Journal of Coastal Research* 22: 1051–1061.
- Donnelly JP, Cleary P, Newby P, and Ettinger R (2004) Coupling instrumental and geological records of sea-level change: Evidence from southern New England of an increase in the rate of sea-level rise in the late 19th century. *Geophysical Research Letters* 31: L05203.
- Engelhart SE, Horton BP, Douglas BC, Peltier WR, and Tornqvist TE (2009) Spatial variability of late Holocene and 20th century sea-level rise along the Atlantic coast of the United States. *Geology* 37: 1115–1118.
- Gehrels WR (2010) Late Holocene land- and sea-level changes in the British Isles: Implications for future sea-level predictions. *Quaternary Science Reviews* 29: 1648–1660.
- Gehrels WR, Belknap DF, Black S, and Newnham RM (2002) Rapid sea-level rise in the Gulf of Maine, USA, since AD 1800. *The Holocene* 12: 383–389.
- Gehrels WR, Hayward B, Newnham RM, and Southall KE (2008) A 20th century acceleration of sea-level rise in New Zealand. *Geophysical Research Letters* 35: L02717.
- Gehrels WR, Kirby JR, Prokoph A, et al. (2005) Onset of recent rapid sea-level rise in the western Atlantic Ocean. *Quaternary Science Reviews* 24: 2083–2100.
- Gehrels WR, Szkornik K, Bartholdy J, et al. (2006) Late Holocene sea-level changes and isostasy in western Denmark. *Quaternary Research* 66: 288–302.
- Hayward B, Grenfell HR, Sandiford A, Shane PAR, Morley M, and Alloway BV (2002) Foraminiferal and molluscan evidence for the Holocene marine history of two breached maar lakes, Auckland, New Zealand. *New Zealand Journal of Geology and Geophysics* 45: 467–479.
- Hijma MP and Cohen KM (2010) Timing and magnitude of the sea-level jump preluding the 8200 yr event. *Geology* 38: 275–278.
- Horton BP, Peltier WR, Culver SJ, et al. (2009) Holocene sea-level changes along the North Carolina Coastline and their implications for glacial isostatic adjustment models. *Quaternary Science Reviews* 28: 1725–1736.
- Kelley JT, Belknap DF, and Claesson S (2010) Drowned coastal deposits with associated archaeological remains from a sea-level slowstand: Northwestern Gulf of Maine, USA. *Geology* 38: 695–698.
- Kemp AC, Horton BP, Culver SJ, et al. (2009) The timing and magnitude of recent accelerated sea-level rise (North Carolina, USA). *Geology* 37: 1035–1038.
- Lambeck K and Chappell J (2001) Sea-level change through the last glacial cycle. *Science* 292: 679–686.
- Leorri E, Cearreta A, and Horton BP (2008) A foraminifera-based transfer function as a tool for sea-level reconstructions in the southern Bay of Biscay. *Geobios* 41: 787–797.
- Lessa G and Masselink G (2006) Evidence of a mid-Holocene sea level highstand from the sedimentary record of a macrotidal barrier and paleoestuary system in northwestern Australia. *Journal of Coastal Research* 22: 100–112.
- Lewis SE, Wust RAJ, Webster JM, and Shields GA (2008) Mid-late Holocene sea-level variability in eastern Australia. *Terra Nova* 20(1): 74–81.
- Meyer M and Harff J (2005) Modelling palaeo coastline changes of the Baltic Sea. *Journal of Coastal Research* 21: 598–609.
- Miller KG, Sugarman PJ, Browning JV, et al. (2009) Sea-level rise in New Jersey over the past 5000 years: Implications to anthropogenic changes. *Global and Planetary Change* 66: 10–18.
- Milne GA, Long AJ, and Bassett SE (2005) Modelling Holocene relative sea-level observations from the Caribbean and South America. *Quaternary Science Reviews* 24: 1183.
- Milne GA and Mitrovica JX (2008) Searching for eustasy in deglacial sea-level histories. *Quaternary Science Reviews* 27: 2292–2302.
- Nikitina DL, Pizzuto JE, Schwimmer RA, and Ramsey KW (2000) An updated Holocene sea-level curve for the Delaware coast. *Marine Geology* 171: 7–20.
- Peltier WR (2002) On eustatic sea level history: Last Glacial Maximum to Holocene. *Quaternary Science Reviews* 21: 377.
- Ramsay PJ (1996) 9000 Years of sea-level change along the southern African coastline. *Quaternary International* 31: 71–75.
- Raukas A (2000) Rapid changes of the Estonian coast during the late glacial and Holocene. *Marine Geology* 170–175: 169.
- Rostami K, Peltier WR, and Mangini A (2000) Quaternary marine terraces, sea-level changes and uplift history of Patagonia, Argentina: Comparisons with predictions of the ICE-4G (VM2) model of the global process of glacial isostatic adjustment. *Quaternary Science Reviews* 19: 1495–1525.
- Shaw J, Gareau P, and Courtney RC (2002) Palaeogeography of Atlantic Canada 13–0 kyr. *Quaternary Science Reviews* 21: 1861–1878.
- Shennan I and Horton B (2002) Holocene land- and sea-level changes in Great Britain. *Journal of Quaternary Science* 17: 511–526.
- Sloss CR, Jones BG, Murray-Wallace CV, and McClennen CE (2005) Holocene sea level fluctuations and the sedimentary evolution of a barrier estuary: Lake Illawarra, New South Wales, Australia. *Journal of Coastal Research* 21: 943–959.
- Streif H (2004) Sedimentary record of Pleistocene and Holocene marine inundations along the North Sea coast of Lower Saxony, Germany. *Quaternary International* 112: 3–28.
- Szkornik K, Gehrels WR, and Murray AS (2008) Aeolian sand movement and relative sea-level rise in Ho Bugt, western Denmark, during the ‘Little Ice Age’. *The Holocene* 18: 951–965.
- Uchupi E, Giese G, Driscoll N, and Aubrey DG (2005) Postglacial geomorphic evolution of a segment of Cape Cod Bay and adjacent Cape Cod, Massachusetts, USA. *Journal of Coastal Research* 21: 1085–1106.
- van de Plassche O (2000) North Atlantic climate–ocean variations and sea level in Long Island Sound, Connecticut, since 500 cal yr A.D. *Quaternary Research* 53: 89–97.
- Yu SY, Berglund BE, Sandgren P, and Lambeck K (2007) Evidence for a rapid sea-level rise 7600 yr ago. *Geology* 35: 891–894.
- Zong YQ (2004) Mid-Holocene sea-level highstand along the southeast coast of China. *Quaternary International* 117: 55–67.

Late Quaternary Relative Sea-Level Changes in the Tropics

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Introduction

The tropical zone is commonly defined by the parallels of Tropic of Cancer at 23°30'N and Tropic of Capricorn at 23°30'S. In recent decades, the tropics have been researched intensively for the history of Late Quaternary relative sea-level (RSL) changes. These studies are important because data from the tropics can be used to reconstruct the history of eustatic sea-level changes as the tropics have not been affected by the glacio-isostatic processes that influenced areas close to former ice sheets in the polar regions.

For some time now, geophysical models have been capable of predicting local and regional RSL for any coastal site on Earth. Because of the varied interplay between eustatic and glacio- and hydro-isostatic processes, RSL records are highly varied; therefore, it is useful to distinguish between records from 'far-', 'intermediate-', and 'near'-field sites (Long, 2001). Far-field sites are where the hydro-isostatic signal outweighs the glacio-isostatic signal because they are areas far from the former ice sheet margins. Far-field sites are defined as the area between about 30°N and 30°S, within which the tropics lie.

Because of the distinctive RSL history in far-field sites, the patterns of coastal evolution are important issues for the tropics. For instance, the directions and paces of shoreline movement in large estuaries and deltas during the Holocene have been controlled to a certain extent by changes of RSL. Looking into the future, many low-lying coastal areas in the tropics, such as the Maldives in the Indian Ocean and many similar atolls in the Pacific Ocean, and the ecosystems therein, will become vulnerable to, or affected by, RSL rise in the near future as a result of global climatic warming.

Sea-Level Indicators

There are a number of sea-level indicators found uniquely in the tropics. These indicators include coral, mangrove, beach rock, and tubeworm. They live or form close to sea level. Their carbonate structures or organic materials enable us to determine the age of each sample by the use of radiometric dating methods. With the age, altitude and indicative meaning, they can be used as sea-level index points for the reconstruction of the Late Quaternary history of RSL changes. The spatial distribution of these indicators is generally confined by nature. For instance, the southern and northern limits of coral and mangrove distributions closely follow the 20 °C isotherm of winter average sea surface temperature (SST; Figure 1). Each type of sea-level indicators has its own strength and limitations. Details of these sea-level indicators are discussed below with special attention given to issues associated with dating and indicative meaning.

Coral

Although corals are common throughout the world's oceans, the reef-building, or hermatypic, corals are restricted to tropical or subtropical latitudes, roughly between the 30°N and 30°S parallels where SST does not fall below 20 °C in the coldest month of the year, and to relatively shallow (<100 m) waters. The basic unit of the reef-building corals is the polyp, which sits within an external skeleton of calcium carbonate secreted by the outer layer of the coral tissue.

Over time as the coral grows, new skeletal material is laid down so that while the structure of the coral may be considerable, only the outermost layer is living tissue. Polyp colonies are normally long-lived, spanning hundreds to thousands of years. Together with a number of interlocking elements, including autochthonous and allochthonous remains of animals and plants, clastic sediment and chemical alteration processes, they produce massive calcareous structures close to sea level (Figure 2). Growth rates of about 10–30 mm years⁻¹ are common (Lowe and Walker, 1997). Gradual rise in sea level has enabled reefs to build vertically, maintaining their shallow-water habitat over time, and thus provide crucial records of sea-level movements.

There are, however, a number of issues that are associated with the application of corals to sea-level studies. The most important one is to establish the indicative meaning of the coral specimen under study, because different coral species live under different water depths. A reef-crest coral *Acropora palmata*, for example, is the best suited to sea-level studies as this species is generally restricted to the upper 5 m of water and dominates the reef-crest community in many locations of the Caribbean. Coral samples can be dated by radiocarbon dating methods and the U–Th method which is significantly more precise than the former (Walker, 2005; Yu and Zhao, 2009).

Mangrove

Mangrove swamps or forests occur in almost the same climate zone as coral, that is, between the 30°N and 30°S parallels (Figure 1). While corals enjoy living in clear water, mangroves prefer coastal sites where there are muddy or fine-grained sediments (Figure 3). Within coastal settings, mangroves' distribution and species composition vary considerably. Although extending into the marine environment, mangrove forest is nevertheless a terrestrial ecosystem, linked to the hinterland landform, vegetation and hydrology. Therefore, zonations of mangroves are peculiarly site specific (Clough, 1982).

The strength of mangrove deposits being used for sea-level studies lies in the fact that the rich organic carbon contained in mangrove deposits can be dated by radiocarbon methods, and the fact that mangrove communities live close to local mean high water levels (Figure 4(a)). Therefore, the indicative meaning for mangrove samples can be determined within a

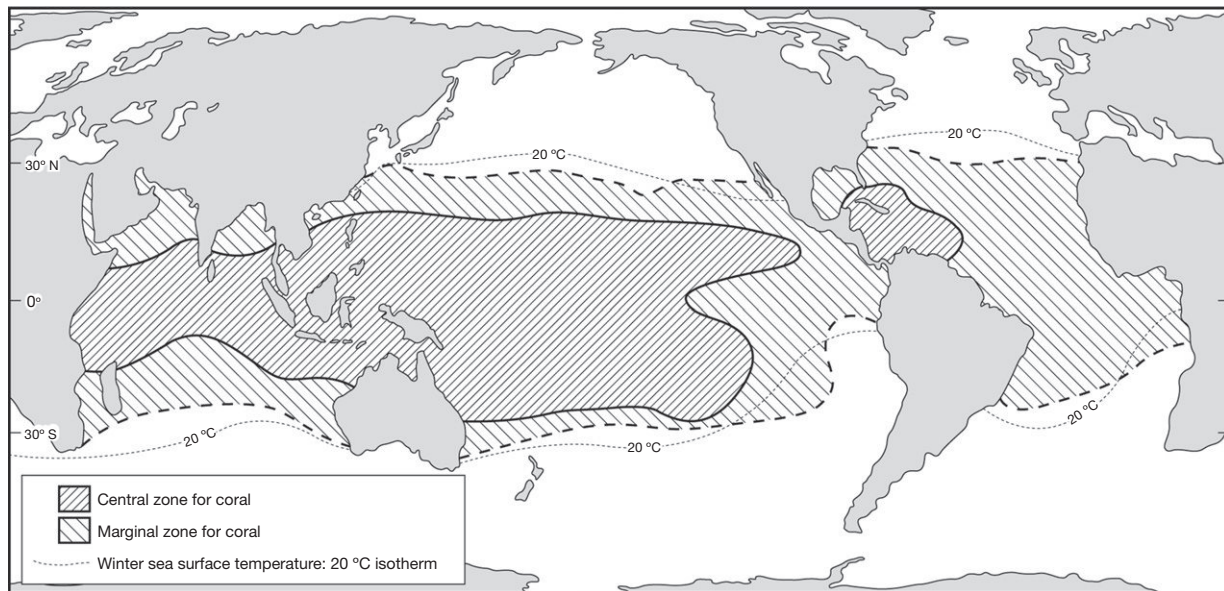


Figure 1 The distribution of reef-building corals and mangroves are generally limited to the 20 °C isotherm of winter average sea surface temperature, which roughly follow the 30° N and 30° S parallels. Most beach rocks are also found within this zone, although tubeworms extend further into mid-latitudes.



Figure 2 A coral reef (*Acropora palmata*) of mid-Holocene in age emerged above local mean sea level in Barbados. ©2005, Photograph by A.J. Long.

narrow altitudinal error band. Furthermore, microfossil pollen, diatom, and Foraminifera are generally well preserved within mangrove deposits (Horton et al., 2005; Zong and Kamaludin, 2004), and microfossil analysis can help determine the indicative meaning for a dated horizon (Horton et al., 2007). As a result, mangrove deposits (Figure 4(b)) have been widely used as sea-level indicators.

On a regional scale, mangrove development and decline can be linked to specific phases of Late Quaternary sea-level change. Data from the north Australian continental shelf indicate that mangrove expansion occurs during marine transgressions, whereas mangrove contraction occurs during marine regressions. The dominant factor controlling the cycle of expansion and decline of mangrove communities is explained in



Figure 3 Mangrove forest (*Rhizophora* sp.) colonized on a tidal mudflat in Hong Kong. ©2010, Photograph by Y. Zong.

terms of contrasting physiographic conditions of coasts during transgressive and regressive phases, in particular by the existence, or lack, of well-developed tidal estuaries (Grindrod et al., 1999).

Beach Rock, Oysters, and Tubeworms

Beach rock, in the form of cemented sandstone, is widely recorded from sandy beaches in the tropics and subtropics as a mixture of sand grains and calcareous marine biofragments that have undergone marine processes of cementation. Beach rock has developed in the kind of climatic and physiographical conditions in which calcium carbonate cements have been precipitated from sea water and come from beach sediments.

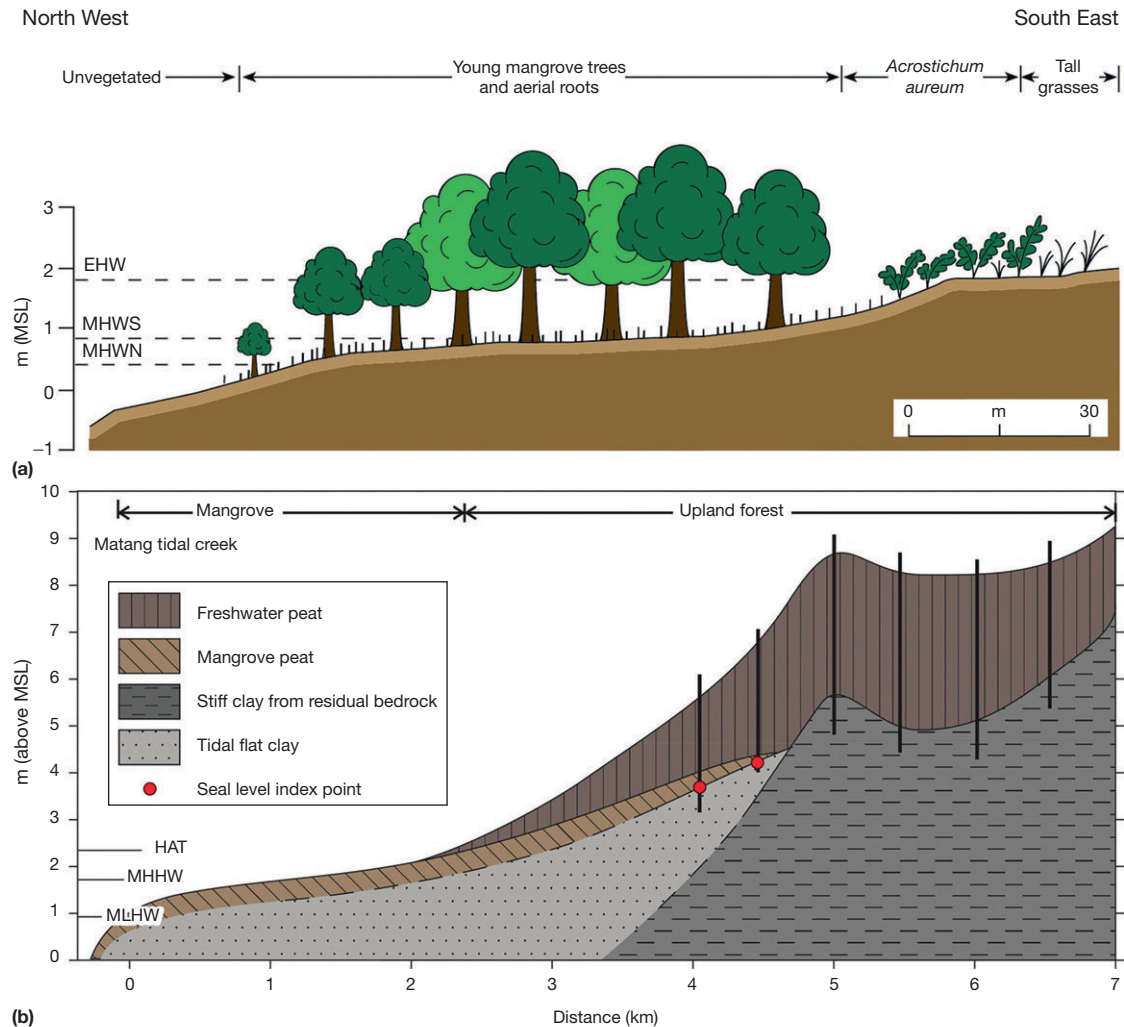


Figure 4 A transect of mangrove forest (a) recorded on the east coast of Peninsular Malaysia shows the pioneer mangrove vegetation is confined within local mean high water of neap tides and mean high water of spring tides. Mature mangrove vegetation grow between local mean high water of spring tides and tidal limit, above which terrestrial vegetation occupies. Sequences of tidal flat and mangrove deposit can be obtained from many coastal sites, and the contact between tidal flat sediment and mangrove peat (b) can be used as sea-level indicator. EHW, Extremely high water; HAT, highest astronomical tide; MHHW, mean higher high water; MHWN, mean high water neaps; MHWS, mean high water springs; MLHW, mean lower high water. Reproduced from Zong Y and Kamaludin BH (2004) Diatom assemblages from two mangrove tidal flats in Peninsular Malaysia. *Diatom Research* 19: 329–344.

Beach-rock forms in the upper part of a beach, generally between mean sea level and mean high water levels (Figure 5). In other words, the indicative meaning and altitudinal error for beach rock can be readily estimated. The main concern for using beach-rock data for sea-level studies lies in the considerable uncertainties associated with the age of a beach-rock sample: the cement materials, mostly calcite and aragonite, may have formed significantly later than the calcareous marine biofragments.

Shells of oyster and tubeworm are also useful proxies of RSL as they are found in coasts across the tropics. Oysters (such as *Saccostrea cucullata*) and tubeworms (such as *Galeolaria caespitosa*) fix their shells on rock faces and normally occur in a tightly constrained altitudinal zone. Therefore, the indicative meaning can be easily determined and their error ranges are as narrow as 0.1 m in the case of certain species (Baker et al., 2001).



Figure 5 Beach rock formed on a sandy beach on the southern coast of Hainan Island, China. ©2005, Photograph by W.W.-S. Yim.

Late Quaternary RSL Records

Eustatic Sea-Level History and Climate Change

The tropics as a far-field site has yielded RSL records that provide a reasonable approximation of eustatic changes linked to glacial cycles, because the combined glacio- and hydro-isostatic effects in far-field sites are usually small. RSL data from the tropics have been used to constrain geophysical models concerning eustatic sea level since the Last Glacial Maximum (LGM) and eustatic contribution of major ice sheets. Observations of RSL in the tropics are therefore crucial to improving the predictive power of geophysical models. In particular, it is important to identify the key reference point for the beginning of deglaciation, periods of rapid sea-level rise, and the early-middle Holocene slowdown of RSL, as well as to identify sites where the RSL curves switch from rising to falling during the mid-Holocene.

Late Quaternary sea-level history comprises the time span of a full glacial cycle, that is, from the last interglacial, through the last glacial, to the present interglacial. Evidence of RSL changes in the tropics can be discerned for four distinctive periods: sea-level highstand of the last interglacial, sea-level lowstand during the LGM, rapid sea-level rise from the Late Glacial to early Holocene and sea-level highstand during middle to late Holocene. To understand the distinctive characteristics of these intervals, the following sections review age/altitude evidence of RSL for each of the four periods.

Sea-Level Highstand for the Last Interglacial

RSL records for the time period prior to the LGM are sparse, despite the shallow marine deposits and coral reefs of the last interglacial (Marine Isotopic Stage 5, MIS 5). The shallow

marine deposits are found at altitudes close to the present-day sea level in several tropical regions including the Red Sea, Peninsular Malaysia, southern Vietnam, and southern China. For example, a shallow marine depositional unit exists consistently at a depth of around 10 m below present sea level across the area of Hong Kong and within the Pearl River delta plain (Zong et al., 2009a). Earlier, the age of this unit was mistakenly attributed to a younger age because of the young age bias of radiocarbon dates, but it was later confirmed to be in the range around the MIS 5 (Yim et al., 1990). In Papua New Guinea (Yokoyama et al., 2001), Barbados (Bard et al., 1990), and Taiti (Dumas et al., 2006), the sea-level highstand is recorded from raised coral reefs or marine terraces.

Sea-Level Lowstand During the LGM

Earlier observations of the LGM sea level in the tropics have shown considerable uncertainty, in part because of the limitation of the preserved indicators and inadequate age determination. Similarly, earlier geophysical models have produced reconstructions of the LGM eustatic sea level with large uncertainties, ranging from -127 m as the upper limit estimate to -163 m as the lower limit estimate. By the end of the 1980s, the first reliable evidence of the LGM sea-level lowstand was reported from Barbados, where the local LGM RSL was recorded to be at approximately -120 m around 18 000 cal years BP, based on a deep core into the coral reef (Fairbanks, 1989). Subsequently, in Sunda Shelf, the local LGM RSL was observed at -115 m for a period between 21 000 and 19 000 cal years BP, based on only a few intertidal sediment samples (Hanebuth et al., 2000). The best record comes from the Bonaparte Gulf, north of Australia (Figure 6(a)), where the local LGM RSL was observed at -125 m, based on a significant

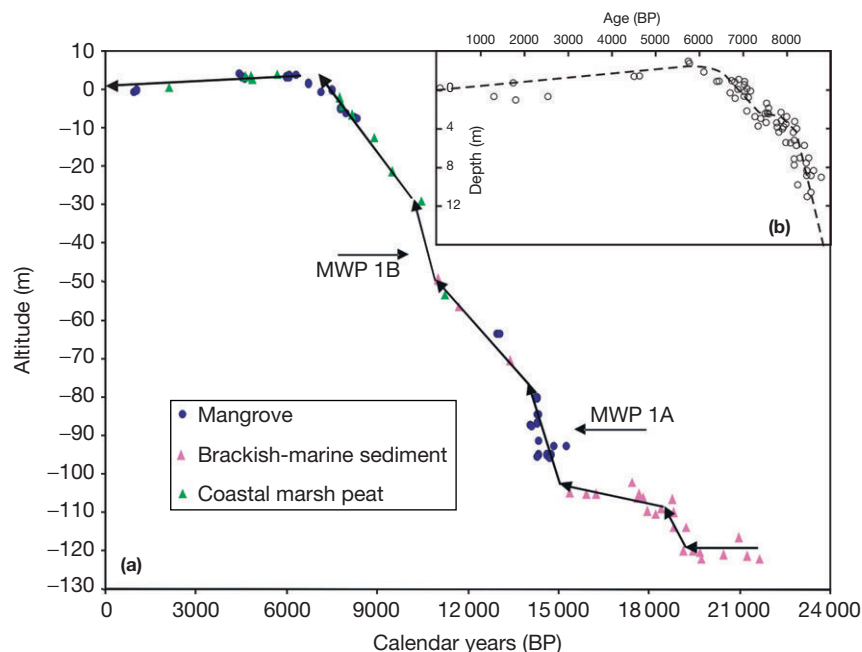


Figure 6 A relative sea-level curve (a) based on sea-level index points compiled from the Bonaparte Gulf, north Australia (Yokoyama et al., 2000), the Sunda Shelf (Hanebuth et al., 2000), and the Strait of Malacca (Geyh et al., 1979). A new record (b) from Singapore (Bird et al., 2010) shows that sea-level rise slowed down around 8000 years BP and accelerated around 7500 years BP.

number of intertidal and shallow marine sediment samples (Yokoyama et al., 2000). This altitude of the LGM RSL low stand is confirmed by new records from New Guinea and Barbados, and the low stand had started 26 000 cal years BP, ca. 5000 years earlier than previously assumed (Clark et al., 2009; Peltier and Fairbanks, 2006).

After the LGM sea-level data are adjusted using a glacio-hydro-isostatic formulation with Northern Hemisphere ice models and an Antarctic ice model that contributes 25 m to global sea level. The ice-volume-equivalent (eustatic) sea level during the LGM is placed between -130 and -135 m, corresponding to a volume of grounded ice in excess of the present volume by $52.5 \times 10^6 \text{ km}^3$ (Yokoyama et al., 2000). This finding is significant because it provides an accurate estimate of the total LGM ice volumes, which in turn can be used to constrain glaciologically based models. Furthermore, the results from the Bonaparte Gulf suggest RSL was relatively stable between 22 000 and 19 000 cal years BP at about -120 m (Figure 6(a)), although the oldest samples may have biased the ages upward by as much as several thousand years (Lambeck et al., 2002). The LGM sea-level lowstand was followed by a rapid rise of 10–15 m for a few hundreds of years, and a much slower rise around 19 000 cal years BP (Figure 6(a); Hanebuth et al., 2009; Yokoyama et al., 2000). It is noted that the data points for the LGM sea-level lowstand scatter around -120 m, possibly reflecting the differential hydro-isostatic loading at the different sites (Bassett et al., 2005).

During the LGM sea-level lowstand, much of the continental shelves in the tropics were exposed (Figure 7). As a result, Australia was linked up with New Guinea. A large land area emerged, joining the Malaysia Peninsula with major islands of Sumatra, Java, and Borneo, and separating the Indian Ocean from the South China Sea. Such an emergence of the extensive continental shelves significantly reduced the size of the maritime area between Australia and Southeast Asia, an important part of the Indo-Pacific Warm Pool. This in turn led to a reduction of moisture supply to the region and to some extent, the global system.

Late Glacial Rapid Sea-Level Rise

The first complete record of Late Glacial rapid sea-level rise comes from Barbados, where more than 30 coral (*A. palmata*) samples were dated and used in the reconstruction of RSL history (Fairbanks, 1989). This and other studies in Barbados and New Guinea have produced good RSL records that help clarify the dynamics of large ice sheets for the postglacial period. However, the reefs in these two locations overlie active subduction zones, where tectonic movements are large and often discontinuous. Thus, the apparent sea-level records from these two sites may be confounded by a complex tectonic component. Subsequently, a more reliable sea-level record was produced from a coral reef in Tahiti, a tectonically more stable site as it is far from plate boundaries (Bard et al., 1996). In Sunda Shelf, a number of sediment cores containing mangrove swamp deposits and intertidal sediment samples were analyzed and a local RSL was reconstructed (Hanebuth et al., 2000).

These sea-level records all point to the fact that the rate of rising sea level during the period between the LGM and the mid-Holocene has not been constant. In addition to the rise in

RSL of 10–15 m around 19 000 cal years BP (Yokoyama et al., 2000), two periods of accelerated rise in sea level have been observed, and named as melt water pulses (MWP) 1A and 1B (Figure 6(a)). According to the records from Sunda Shelf (Hanebuth et al., 2000), the first phase of accelerated sea-level rise started at 14 700 cal years BP and ended at 14 100 cal years BP, as RSL rose from approximately -95 to -80 m, in correspondence to the first cold reversal of the Bølling–Allerød period observed in Greenland ice cores. This accelerated rise in sea level was likely triggered by the rapid melting of the Antarctic ice sheet and ice sheets in Northern Hemisphere (Bassett et al., 2005). The second phase of accelerated sea-level rise is less well recorded, but is believed to have started soon after the Younger Drays cold period that ended at ca. 11 500 cal years BP (Bard et al., 2010). These two periods of accelerated sea-level rise are in full agreement with the ^{18}O records from deep-sea cores. Recently, a further change in the rate of sea-level rise has been recorded from Singapore, where sea-level rise slowed down from ca. 8000 cal years BP and accelerated again from ca. 7500 cal years BP (Figure 6(b); Bird et al., 2010). This punctuated sea-level rise has resulted in widespread marine transgression in many coastal lowlands of east and Southeast Asia (e.g., Hori et al., 2004; Zong et al., 2009b).

As a result of the rapid rise in sea level, the extensive continental shelves were submerged (Figure 7) and the Indo-Pacific Warm Pool was restored, leading to a significant change in the moisture supply and ecosystems across the region of east and southeast Asia (Griffiths et al., 2009). Similar changes also took place in the Gulf of Mexico, albeit to a lesser extent, and resulted in the restoration of moisture supply to southern Mexico and the surrounding areas.

Holocene Sea-Level Highstand

An RSL approximately 3 m higher than present-day RSL has been observed widely across the tropics (Pirazzoli, 1996). The reason for the occurrence of such a sea-level high-stand lies in the fact that the eustatic sea-level rise markedly slowed down in mid-Holocene, because the major ice sheets ceased to release meltwater to the oceans, combined with the hydro-isostatic processes that (1) water migrated away from far-field equatorial ocean basins in order to fill space vacated by collapsing forebulges at the periphery of previously glaciated regions, and/or (2) ocean loading at continental margins induced a levering of continents and a subsidence of offshore regions, resulting in the influx of water into the volume created by the subsidence and a sea-level fall at locations from these margins over major ocean basins (Mitrovica and Milne, 2002). However, differences in the timing and amplitude of the sea-level highstand remain, because it is still not clear whether meltwater release stopped abruptly in the mid-Holocene or whether there was a continued gradual release of meltwater during mid-Holocene and much of the late Holocene (Long, 2001).

The best record of mid-Holocene sea-level highstand comes from the Strait of Malacca, where mangrove and coastal peat deposits were surveyed and dated (Geyh et al., 1979). The results suggest that RSL reached its maximum height of approximately 4 m at about 6000–7000 cal years BP (Table 1) and afterward fell to the present level. The emergent Holocene



Figure 7 Continental shelves that were exposed at sea-level lowstand during the Last Glacial Maximum and submerged during postglacial sea-level rise and the Holocene sea-level highstand.

microatolls in Torres Strait indicate that during the past 6600 years the reefs have prograded seawards while sea level has fallen gradually from about 1 m above present (Woodroffe et al., 2000). On the Great Barrier Reef of Australia, a 1.7 m sea-level highstand is dated to about 6000 years ago (Larcombe et al., 1995). Along the south coast of China, a mid-Holocene sea-level highstand of approximately 1 m above present-day sea level has been recorded (Zong, 2004). Along the east coast of India, evidence from marine terraces of hermatypic corals and estuarine molluscs indicates a mid-Holocene sea-level highstand a few meters above the present-day sea level at ca. 7350 cal years BP (Banerjee, 2000). According to the times and heights of the mid-Holocene sea-level highstand recorded from tectonically stable sites of the tropics listed in

Table 1, a general pattern can be seen: the amplitude of mid-Holocene sea-level highstand decreases from its maximum at the equator toward mid-latitudes or intermediate-field sites where highstands are not recorded.

Records from the Marian Islands, Caroline Islands, Papua New Guinea, and the Taiwan Strait have all suggested mid-Holocene sea-level highstands of 1–3 m. But these records are less reliable because these areas have been subject to subduction tectonics. Furthermore, sea-level records from large estuaries are also less reliable due to local subsidence and sedimentary compaction. For example, no significant sea-level highstand was recorded from the Ganges delta (22–23°N; Islam and Tooley, 1999) and the Pearl and Han river deltas (22–24°N; Zong, 2004), while a mid-Holocene sea-level highstand is

Table 1 Examples of Mid-Holocene sea-level highstands recorded in tectonically stable sites of the tropics

Location	Latitude	Time (cal year BP)	Highstands (above sea level) (m)
South coast of China ^a	21–23° N	6200–6000	0.5–1.0
North coast of Vietnam ^b	20° N	5500	3.25
Thailand ^c	14° N	7000	~3
Southeast coast of India ^d	8–11° N	7430–7190 4750–4200	~3 ~1
Southwest coast of Thailand ^e	6–8° N	ca. 6000 ca. 4000	>1.6 1
Malay–Thai Peninsula ^f	7° N	4850–4450	5
Malacca Strait ^g	2–4° N	6350–4620	3.1–3.5
Torres Strait ^h	10° S	5800–5300	>0.8–1.0
The Great Barrier Reef of Australia ⁱ	19° S	7000 5800	0.7 1.6
The Great Barrier Reef of Australia ^j	20–22° S	6050–4060	1.65
Central New South Wales of Australia ^k	33–34° S	5500–4800 3600–1500	1.4–1.5 1

^aZong (2004).^bBoyd and Lam (2004).^cTanabe et al. (2003).^dBanerjee (2000).^eFujimoto et al. (1999).^fHorton et al. (2005).^gGeyh et al. (1979).^hWoodroffe et al. (2000).ⁱYu and Zhao (2009).^jLarcombe et al. (1995).^kHaworth et al. (2002).

observed on the adjacent coasts. A number of studies have reported a second sea-level highstand occurring in the late Holocene (Table 1). However, the mechanism for this second highstand in far-field sites is neither known nor predicted by geophysical models. Therefore, the occurrence of this second highstand may be a result of local processes.

Conclusions

The Late Quaternary history of eustatic sea-level changes is revealed in multiple kinds of records from the tropics. After the sea-level highstand of the last interglacial, RSL in the tropics fell gradually to its lowest point at about 125 m below present and maintained this level during the period of the LGM between ca. 26 000 and 19 000 cal years BP. In the initial phase of deglaciation, the RSL rose rapidly by 10–15 m within a few hundred years, followed by a slow rise toward 15 000 cal years BP. Between ca. 15 000 and 10 000 cal years BP, RSL rose rapidly in two phases, corresponding to MWP 1A and MWP 1B, when a number of ice sheets on the Northern Hemisphere collapsed. From ca. 10 000 cal years BP, the RSL rose rapidly, but it slowed down at ca. 8000 cal years BP before continuing to rise from ca. 7500 cal years BP. By ca. 7000 cal years BP, the RSL reached to within a few meters above the present-day sea level, since which time it has gradually fallen to the present level.

See also: **Diatom Methods:** $\delta^{18}\text{O}$ Records. **Paleoceanography, Biological Proxies:** Corals, Sclerosponges and Mollusks. **Sea Level Studies:** Eustatic Sea-Level Changes – Glacial–Interglacial Cycles; Eustatic Sea-Level Changes Since the Last Glacial Maximum.

References

- Baker RGV, Haworth RJ, and Flood PG (2001) Inter-tidal fixed indicators of former Holocene sea levels in Australia: A summary of sites and a review of methods and models. *Quaternary International* 83–85: 257–273.
- Banerjee PK (2000) Holocene and Late Pleistocene relative sea level fluctuations along the east coast of India. *Marine Geology* 167: 243–260.
- Bard E, Hamelin B, Arnold M, et al. (1996) Deglacial sea-level record from Tahiti corals and the timing of global meltwater discharge. *Nature* 382: 241–244.
- Bard E, Hamelin B, and Delanghe-Sabatier D (2010) Deglacial meltwater pulse 1B and Younger Dryas sea levels revisited with boreholes at Tahiti. *Science* 327: 1235–1237.
- Bard E, Hamelin B, and Fairbanks RG (1990) U-Th ages obtained by mass spectrometry in corals from Barbados: Sea level during the past 130,000 years. *Nature* 345: 456–458.
- Bassett SE, Milne GA, Mitrovica JX, and Clark PU (2005) Ice sheet and solid earth influences on far-field sea-level histories. *Science* 309: 925–928.
- Bird MI, Austin WEN, Wurster CM, Fifield LK, Mojtabid M, and Sargeant C (2010) Punctuated eustatic sea-level rise in the early mid-Holocene. *Geology* 38: 803–806.
- Boyd WE and Lam DD (2004) Holocene elevated sea levels on the north coast of Vietnam. *Australian Geographical Studies* 42: 77–88.
- Clark PU, Dyke AS, Shakun JD, et al. (2009) The Last Glacial Maximum. *Science* 325: 710–714.
- Clough BF (ed.) (1982) *Mangrove Ecosystems in Australia: Structure, Function and Management*. Australia: Australia Institute of Marine Science.
- Dumas B, Hoang CT, and Raffy J (2006) Record of MIS5 sea-level highstands based on U/Th dated coral terraces of Haiti. *Quaternary International* 145–146: 106–118.
- Fairbanks RG (1989) A 17,000-year glacio-eustatic sea level record: Influence of glacial melting rates on the Younger Dryas event and deep-ocean circulation. *Nature* 342: 637–642.
- Fujimoto K, Miyagi T, Murofushi T, et al. (1999) Mangrove habitat dynamics and Holocene sea-level changes in the southwestern coast of Thailand. *Tropics* 8: 239–255.
- Geyh MA, Kudrass H-R, and Streif H (1979) Sea-level changes during the late Pleistocene and Holocene in the Strait of Malacca. *Nature* 278: 441–443.
- Griffiths ML, Drysdale RN, Gagan MK, et al. (2009) Increasing Australian–Indonesian monsoon rainfall linked to early Holocene sea-level rise. *Nature Geoscience* 2: 636–639.
- Grindrod J, Moss P, and Van de Kaars S (1999) Late Quaternary cycles of mangrove development and decline on the north Australia continental shelf. *Journal of Quaternary Science* 14: 465–470.
- Hanebuth T, Stattegger K, and Bojanowski A (2009) Termination of the Last Glacial Maximum sea-level lowstand: The Sunda Shelf data revisited. *Global and Planetary Change* 66: 76–84.
- Hanebuth T, Stattegger K, and Grootes PM (2000) Rapid flooding of the Sunda Shelf: A late-glacial sea-level record. *Science* 288: 1033–1035.
- Haworth RJ, Baker RGV, and Flood PG (2002) Predicted and observed Holocene sea-levels on the Australian coast: What do they indicate about hydro-isostatic models in far-field sites? *Journal of Quaternary Science* 17: 581–591.
- Hori K, Tanabe S, Saito Y, Haruyama S, Nguyen V, and Kitamura A (2004) Delta initiation and Holocene sea-level change: Example from the Song Hong (Red River) delta, Vietnam. *Sedimentary Geology* 164: 237–249.
- Horton BP, Gibbard PL, Milne GM, Morley RJ, Purintavaragul C, and Stargardt JM (2005) Holocene sea levels and palaeoenvironments, Malay–Thai Peninsula, southeast Asia. *The Holocene* 15: 1199–1213.
- Horton BP, Zong Y, Hillier C, and Engelhart S (2007) Diatoms from Indonesian mangroves and their suitability as sea-level indicators for tropical environments. *Marine Micropaleontology* 63: 155–168.
- Islam MS and Tooley MJ (1999) Coastal and sea-level changes during the Holocene in Bangladesh. *Quaternary International* 55: 61–75.
- Lambeck K, Yokoyama Y, and Purcell T (2002) Into and out of the Last Glacial Maximum: Sea-level change during Oxygen Isotope Stages 3 and 2. *Quaternary Science Reviews* 21: 343–360.
- Larcombe P, Carter RM, Dye J, Gagan MK, and Johnson DP (1995) New evidence for episodic post-glacial sea level rise, central Great Barrier Reef, Australia. *Marine Geology* 127: 1–44.

- Long A (2001) Mid-Holocene sea-level change and coastal evolution. *Progress in Physical Geography* 25: 399–408.
- Lowe JJ and Walker MJC (1997) *Reconstructing Quaternary Environments*, ch. 4. London: Longman.
- Mitrovica JX and Milne GA (2002) On the origin of late Holocene sea-level highstands within equatorial ocean basins. *Quaternary Science Reviews* 21: 2179–2190.
- Peltier WR and Fairbanks RG (2006) Global glacial ice volume and Last Glacial Maximum duration from an extended Barbados sea level record. *Quaternary Science Reviews* 25: 3322–3337.
- Pirazzoli PA (1996) *Sea Level Changes: The Last 20,000 Years*. Chichester: Wiley.
- Tanabe S, Saito Y, Sato Y, et al. (2003) Stratigraphy and Holocene evolution of the mud-dominated Chao Phraya delta, Thailand. *Quaternary Science Reviews* 22: 789–807.
- Walker MJC (2005) *Quaternary Dating Methods*, ch. 3. Chichester: Wiley.
- Woodroffe CD, Kenneday DM, Hopley D, Rasmussen CE, and Smithers SG (2000) Holocene reef growth in Torres Strait. *Marine Geology* 170: 331–346.
- Yim WW-S, Ivanovich M, and Yu K-F (1990) Young age bias of radiocarbon dates in pre-Holocene marine deposits of Hong Kong and implications for Pleistocene stratigraphy. *Geo-Marine Letters* 10: 165–172.
- Yokoyama Y, Esat TM, and Lambeck K (2001) Coupled climate and sea-level changes deduced from Huon Peninsula coral terraces of the last ice age. *Earth and Planetary Science Letters* 193: 579–587.
- Yokoyama Y, Lambeck K, De Deckker P, Johnston P, and Fifield LK (2000) Timing of the Last Glacial Maximum from observed sea-level minima. *Nature* 406: 713–716.
- Yu K and Zhao J (2009) U-series dates of Great Barrier Reef corals suggest at least +0.7 m sea level 7000 years ago. *The Holocene* 20: 161–168.
- Zong Y (2004) Mid-Holocene sea-level highstand along the southeast coast of China. *Quaternary International* 117: 55–67.
- Zong Y, Huang G, Switer AD, Yu F, and Yim WW-S (2009a) An evolutionary model for the Holocene formation of the Pearl River delta, China. *The Holocene* 19: 129–142.
- Zong Y and Kamaludin BH (2004) Diatom assemblages from two mangrove tidal flats in Peninsular Malaysia. *Diatom Research* 19: 329–344.
- Zong Y, Yim WW-S, Yu F, and Huang G (2009b) Late Quaternary environmental changes in the Pearl River mouth region, China. *Quaternary International* 206: 35–45.

Tectonics and Relative Sea-Level Change

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Introduction

Records of relative sea-level (RSL) change archive the response of the Earth's surface to tectonic processes during the Quaternary, the most recent period of Earth history. Landforms or deposits marking former shorelines, termed strandlines, are widespread datums that can be measured more precisely than other types of landforms and deposits produced in response to tectonic processes. Although limited to coasts, sequences of successive strandlines are commonly better preserved over longer periods of time than other tectonic landforms. One third to one half of the Earth's marine coasts lie along or near tectonically active plate boundaries where the potential for large earthquakes threatens many population centers (Lajoie, 1986). Much of our understanding of tectonic processes over hundreds to hundreds of thousands of years has come from study of displacements (vertical components of tectonic deformation caused by faulting and local or regional folding) obtained through mapping and dating sequences of strandlines along tectonically active coasts. This improved understanding of the rates and scales of tectonic processes leads to better forecasting and mitigation of large earthquakes and related hazards – such as tsunamis and landslides – in inland as well as coastal regions worldwide.

RSL changes are a composite of real sea-level changes, which are vertical movements of the ocean's surface, and apparent sea-level changes, which are the inverse of vertical land-level changes along coasts (Lajoie, 1986; van de Plassche, 1986; Figure 1). Although sea level may rise and fall by meters over periods of hours during tidal cycles, storm surges, or tsunamis, most long-term evidence of real changes in mean sea level consists of erosional or depositional strandlines formed over periods of years to millennia. Real changes recorded by strandlines range from local changes of decimeters, for example, caused by changes in the configuration of the mouth of an estuary following a severe storm, to fluctuations of global sea level of 100–150 m caused by expansion and contraction of continental ice sheets over tens of thousands of years (Chappell et al., 1996; Milne and Shennan, 2007). Causes of apparent sea-level changes range from the slow (<0.01 mm year⁻¹) regional epeirogenic uplift of coasts over hundreds of thousands of years (Bloom, 1998; Murray-Wallace, 2007), to the meters of instantaneous, localized uplift of some coasts during the largest subduction-zone (a regional fault where one tectonic plate slides beneath another) earthquakes (Ota and Yamaguchi, 2004; Plafker, 1972). Quantifying the tectonic displacement (uplift or subsidence) of strandlines requires separating the apparent changes from the real changes in strandline records.

The degree to which real sea-level changes are conceptually distinct from apparent (inverse of land-level) changes is scale and time dependent. Over thousands to tens of thousands of years Earth's crust and underlying mantle adjust to isostatic loading from continental ice sheets and global increases in

ocean volume, and over hundreds of thousands to millions of years to lithosphere isostatic loading. Changes in the volume of water in the ocean cause real sea level to rise or fall and over thousands of years ocean volume responds in complex ways to changes in the shape of ocean basins that are driven by slow adjustments in the crust and mantle (Milne and Shennan, 2007; Woodroffe, 2007). From this long-term perspective, all sea-level changes have an apparent component; even strandlines on the most tectonically stable coasts formed partly as a result of apparent sea-level changes.

The areal extent and scale of RSL changes, and the precision with which former sea levels can be measured from strandlines, depend on time span and sea-level history. For this reason, different types of information about tectonic processes are derived from different types of strandlines of differing age. Small apparent sea-level changes of a few meters or less are much more easily identified during the late Holocene (<6 ka, <6000 years old) when real sea-level changes were <1 – 2 m on many coasts than during the early Holocene (12–6 ka) when global sea level was rising rapidly (>10 mm year⁻¹) due to melting of continental ice sheets (Figure 1(b)). Strandlines of the past few hundred years in unusual settings, such as the coralline coasts of Sumatra (Figures 2 and 3), can record centimeters of apparent change in sea level over decades because the amplitude of real sea-level changes over this period has been small. Pleistocene (1.6 My to 12 ka) strandlines record highstands – and less commonly lowstands – of global sea level (Figures 4 and 5), but because they average sea-level position over the hundreds to thousands of years over which they formed, only rarely can they be used to identify sea-level changes of less than a few meters (Chappell et al., 1996; Muhs et al., 2004). Nevertheless, Pleistocene strandlines provide an invaluable measure of the net effect of long-term tectonic processes along tens to hundreds of kilometers of coast over tens of thousands to hundreds of thousands of years.

Strandlines formed on different types of coasts, and at different elevations within the elevational range of tides, have differing degrees of uncertainty, termed indicative meaning, in their relation to mean sea level over their time period of formation (van de Plassche, 1986). For example, the relation of the landward edge of a 120-ka, gravelly, marine terrace to its former mean sea level may have an indicative meaning of several meters, whereas the indicative meaning of the landward edge of a buried tidal marsh to its former mean sea level may be a few tens of centimeters or less. Additional uncertainty in the position of former sea levels depends on the degree of preservation of strandline deposits, landforms, and included fossils, as well as on the level of detail of strandline study. Significant increases in the precision of strandline measurement in the past quarter century have come about through improved understanding of shoreline processes and more rigorous evaluation of errors in strandline indicative meanings.

The most significant advances in understanding tectonic processes through study of strandline records have come

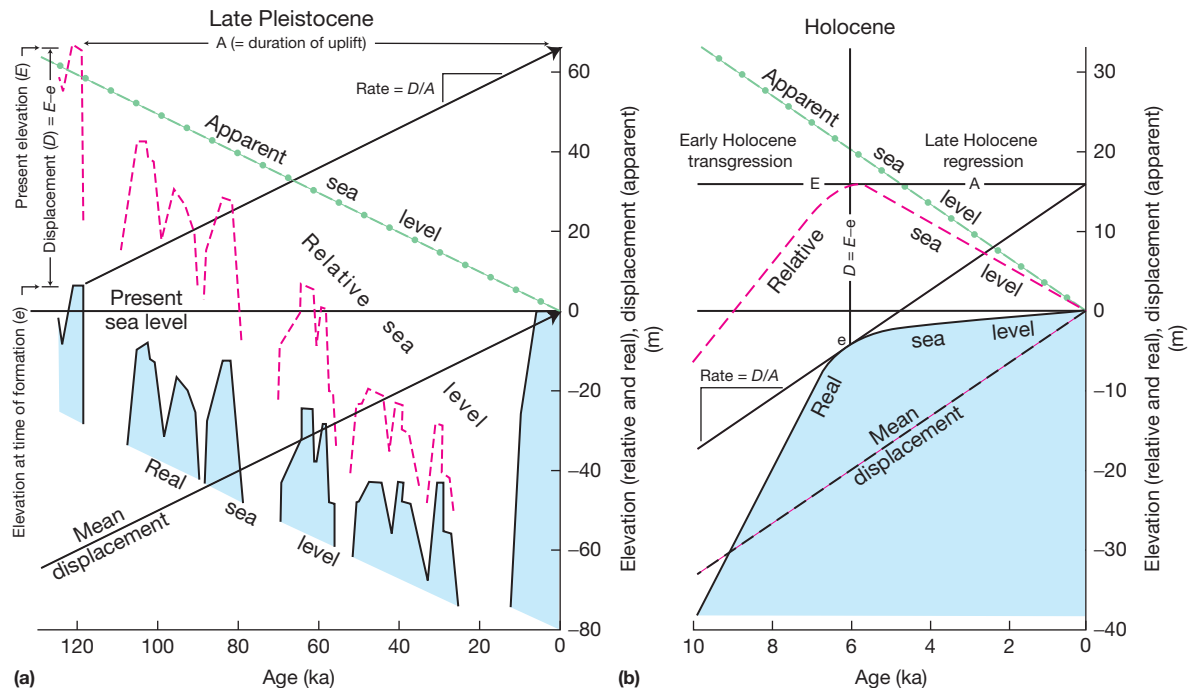


Figure 1 Relative, apparent, and real sea-level changes; (a) late Pleistocene; (b) Holocene. Relative sea-level (RSL) changes are the sum of apparent and real changes. Real sea-level change is the change in the height of the ocean surface adjacent to a shoreline. Apparent sea-level change is the inverse of displacement (D), or the vertical component of tectonic deformation of a shoreline. Displacement is the difference between the present elevation of a strandline (E) and its original elevation (e). The rate of displacement (assumed constant here) is the displacement divided by age (D/A ; ka, 1000 years before AD 1950). Modified from Lajoie KR (1986) Coastal tectonics. In: Wallace RE (ed.) *Active Tectonics*, pp. 95–124. Washington, DC: National Academy of Sciences, National Academy Press. In (b), on many uplifting coasts relative sea level rose until about 6 ka and then fell as uplift rates exceeded real sea-level rise. This produced an early Holocene transgression of the sea followed by a late Holocene regression.

from new and improved methods of dating strandlines (Muhs et al., 2004). Accelerator mass spectrometer ^{14}C dating of samples with less than a thousandth of the carbon formerly needed for dating Holocene strandlines has increased the number of potentially datable strandline samples more than a thousand fold. Mass spectrometric uranium-series dating of well-preserved fossil corals from drill cores on the tectonically stable island of Barbados in the Caribbean has yielded a Pleistocene and Holocene sea-level history that has been used to separate real from apparent sea-level changes worldwide (Bard et al, 1990; Woodroffe, 2007). In the past decade and a half, Holocene as well as Pleistocene strandlines on many coasts have been dated with thermoluminescence and optically stimulated luminescence methods, and more recently with several methods of cosmogenic nuclide dating (^{10}Be , ^{26}Al , ^{36}Cl) using accelerator mass spectrometers (Figure 6). Amino-acid racemization ratios on fossil molluscs have also been an important tool for correlation of Pleistocene strandlines (Figure 7; Muhs et al., 2004; Murray-Wallace, 2007). Some of the highest resolution sea-level studies use uranium-series ages with analytical errors of only a few years to identify centimeter-scale sea-level changes recorded by coral microatolls above the subduction-zone megathrust (plate-boundary fault) of western Sumatra that produced the devastating Indian Ocean tsunami of 2004 (Figures 2 and 3).

Tectonic Processes and Rates

Earthquake Deformation Cycle

For over a century, the accumulation and release of tectonic strain along faults in the shallow (<15 km deep), brittle crust of the Earth has been modeled by the earthquake deformation cycle (Sholtz, 2002; Yeats et al., 1997; Figures 8 and 9). Constant tectonic stress accumulates as strain in rock near a fault until the frictional strength of the fault is exceeded and the fault slips during an earthquake. In the classic model of the cycle, this coseismic slip along the fault is just enough to release all the strain stored in rock on either side of the fault. The instantaneous coseismic deformation (slip on a fault expressed as faulting or folding at the surface) is greatest along the fault plane and in rock adjacent to the fault due to release of accumulated strain, which decreases exponentially away from the fault. Geodetic and tidal observations near active faults in Japan, Chile, and other areas before and after large earthquakes show that the amount and direction of coseismic deformation at various distances from a fault are approximately the opposite of the deformation caused by strain accumulation prior to the earthquake. For example, coasts above subduction-zone megathrusts slowly rise or fall (recording displacement from strain accumulation above the megathrust) for hundreds of years between great (M8-9, magnitude 8 or 9) earthquakes

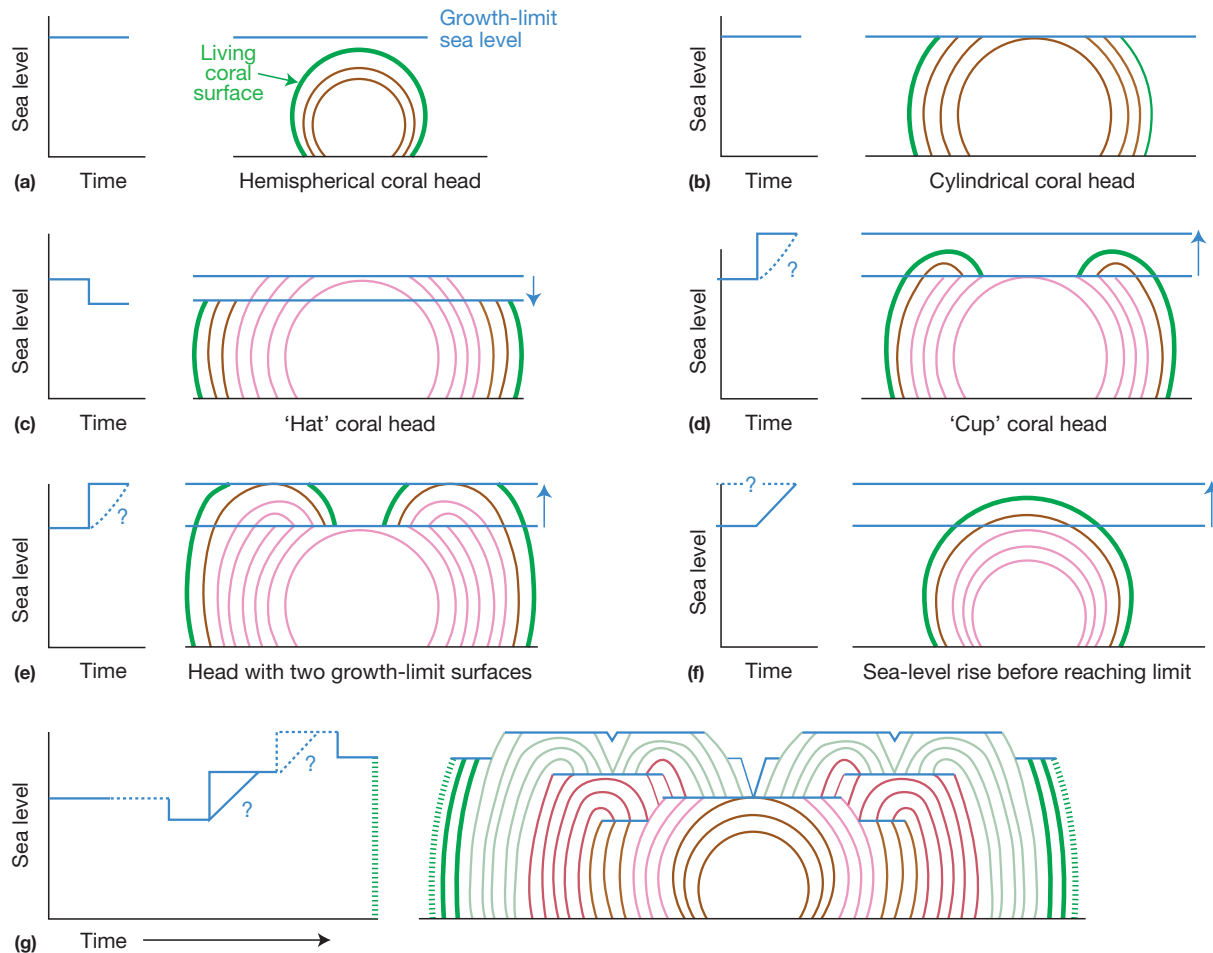


Figure 2 Coral microatoll growth in response to RSL changes above the western Sumatran subduction zone: heavy green rings, living coral bands; brown rings, dead coral bands that grew under present sea level; pink and thin green rings, dead bands that grew under older sea levels; graph at left shows trend of sea level (SL) with time (T). (a) Coral grows unimpeded with hemispherical shape below growth-limit sea level. (b) When coral reaches growth-limit level, coral head grows laterally. (c) If sea level drops to lower level, exposed coral dies, but lateral growth continues below growth-limit level to form coral head with 'hat' morphology. (d) If sea level rises to higher level, coral head grows upward and outward to form 'cup' morphology, but sudden (coseismic) sea-level rise cannot be distinguished from a gradual (interseismic) rise (dashed queried line). (e) If sea level rises to higher level reached by coral head, head grows laterally from both outer and inner edges covering older coral in center of head. (f) If sea level rises more rapidly than coral grows upward, coral will not record sea-level changes. (g) Cross-section through hypothetical coral head that records sea-level changes shown in graph at left; queried lines show alternative sea-level history. Modified from Zachariassen J, Sieh K, Taylor FW, and Hantoro WS (2000) Modern vertical deformation above the Sumatran subduction zone: Paleogeodetic insights from coral microatolls. *Bulletin of the Seismological Society of America* 90: 897–913.

on the megathrust, and then are suddenly jerked downward or upward, respectively, by a similar amount during earthquakes (Figure 9; Satake and Atwater, 2007). Following each earthquake, strain again accumulates during the interseismic period between the previous and the next earthquake on the fault.

Instrumental and tidal observations of the past century near tectonically active coasts show much more complicated patterns of deformation than implied by the classic model of the earthquake cycle (Briggs et al., 2008; Ota and Yamaguchi, 2004; Pirazzoli, 1996; Sholtz, 2002; Wang, 2007). In some cases, strain accumulation accelerates during both the pre-seismic period, just prior to the earthquake, and the postseismic period following it (Figure 9(c)). At some coastal sites, the amount of pre-seismic or postseismic displacement may be as

much as or even exceed the coseismic slip, and on most faults, displacement commonly varies from cycle to cycle. In the case of some plate-boundary faults that extend to depths well below the brittle shallow crust of the Earth, 'slow' (occurring over hours to years) earthquakes may release huge amounts of energy which may be recorded by postseismic or interseismic deformation above deep faults. Beneath coastal south-central Chile, continuing small adjustments of the deep crust and mantle are a response to megathrust slip during the M9.5 earthquake of 1960. Most importantly for studies of tectonically displaced strandlines, all of the pre-faulting strain may not be recovered during large earthquakes. The small but permanent unrecovered strain following successive earthquakes may accumulate in complex patterns that are recorded by uplifted or subsided strandlines.

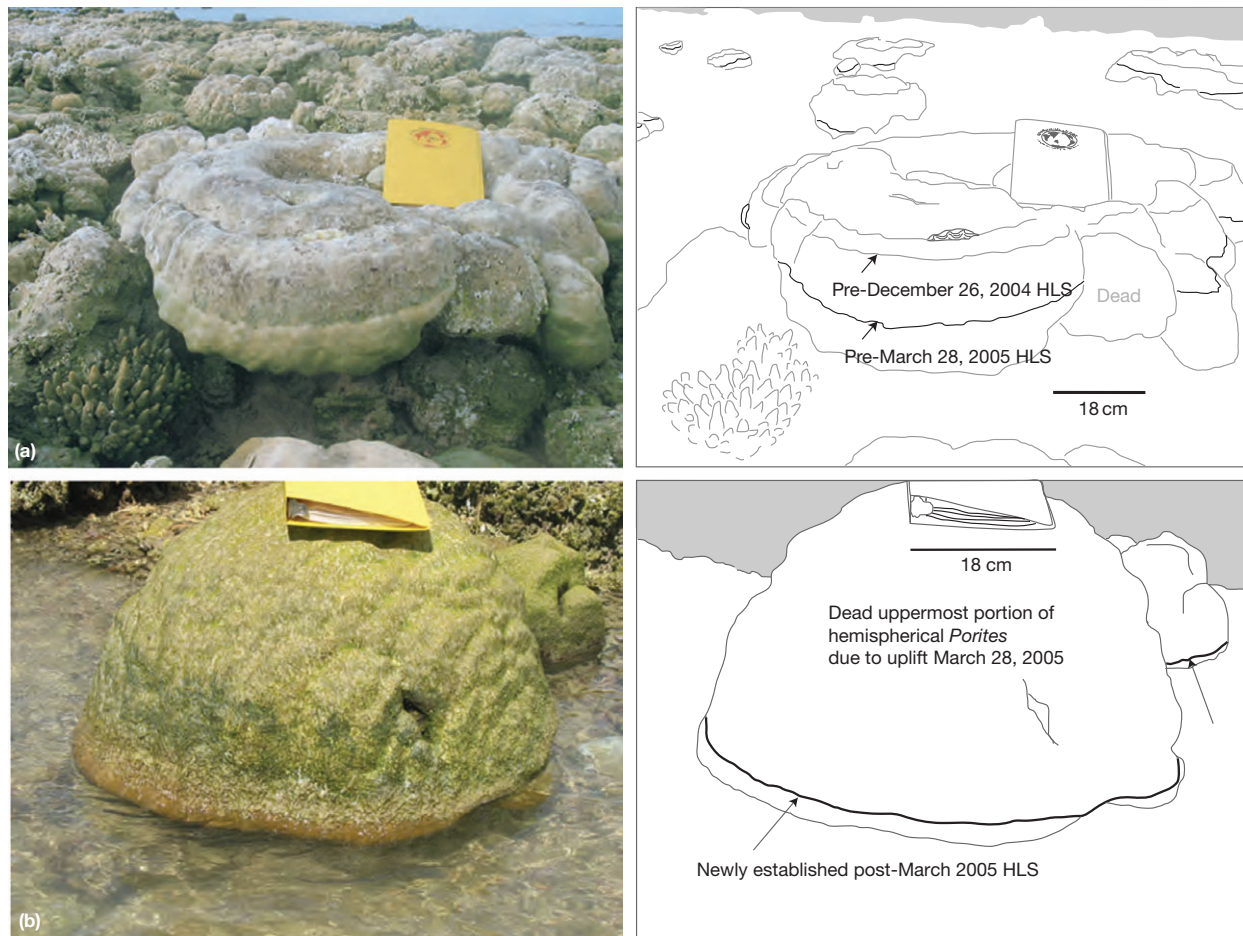


Figure 3 Photographs and corresponding line drawings of uplifted *Porites* coral microatolls on the west coast of Sumatra. (a) An uplifted hemispherical *Porites* head that records the new, postearthquake highest level of survival (HLS) as the top of a thin living strip at its base. The uppermost, exposed portion of the microatoll is dead and covered with algae. (b) A *Porites* microatoll that records multiple uplift events. Uplift of ~11 cm during the December 2004 earthquake resulted in the elevation difference between the uppermost living coral before and after the earthquake. The more brightly colored area beneath the pre-March 2005 HLS is outward growth during the period between the December 2004 and March 2005 earthquakes. During the March earthquake the coral was uplifted another ~65 cm, causing the head to be lifted entirely out of the water into a position too high to record post-March 2005 HLS. Reproduced from Briggs RW, Sieh K, Meltzner AJ, et al. (2006) Deformation and slip along the Sunda megathrust in the great 2005 Nias-Simeulue earthquake. *Science* 311: 1897–1901.

Two important end-member models of the earthquake cycle allow prediction of the time or size of the next earthquake (Figure 8; Burbank and Anderson, 2001; Yeats et al., 1997). In the classic periodic cycle model, the frictional strength of the fault, the stress drop during slip on the fault, and the time till the next earthquake are the same during each cycle. Thus, both the time and magnitude (amount of energy released) of the next earthquake can be predicted. In the time-predictable model, earthquakes occur whenever a constant stress level on the fault is reached, but the amount of stress drop and the slip are unpredictable. If interseismic strain accumulation is constant and the amount of slip in the previous earthquake is known, the time of the next earthquake – but not its slip – can be predicted. In the slip-predictable model, slip on a fault during an earthquake is assumed to stop when the stress has dropped to a constant level. If the time since the previous earthquake is known, the amount of slip during the next earthquake at any given time can be predicted, but not the time of the next earthquake.

The nucleation and propagation of earthquake ruptures along fault planes, especially on large plate-boundary faults, is far more complex than implied by such simple models (Wang, 2007). As shown in many studies of modern earthquakes, deep earth structure, laboratory studies of rock, geologic mapping, and historical studies of preinstrumental large earthquakes, the strength and topography of fault planes varies greatly along fault strike as well as with depth (Figures 10 and 11; Sholtz, 2002). Cumulative slip along fault planes over many earthquake cycles produces changes in fault plane properties over time. Despite the popularity of models of uniform fault slip during successive earthquakes, such changes in fault plane properties imply that the slip during successive earthquakes at the same location on most faults, as well as along the length of faults during a single earthquake, will vary. Furthermore, slip on some large faults alters the stress fields on nearby – and occasionally distant – faults and so affects the timing and magnitude of future earthquakes on the faults. Such complexity in the amounts and rates of fault

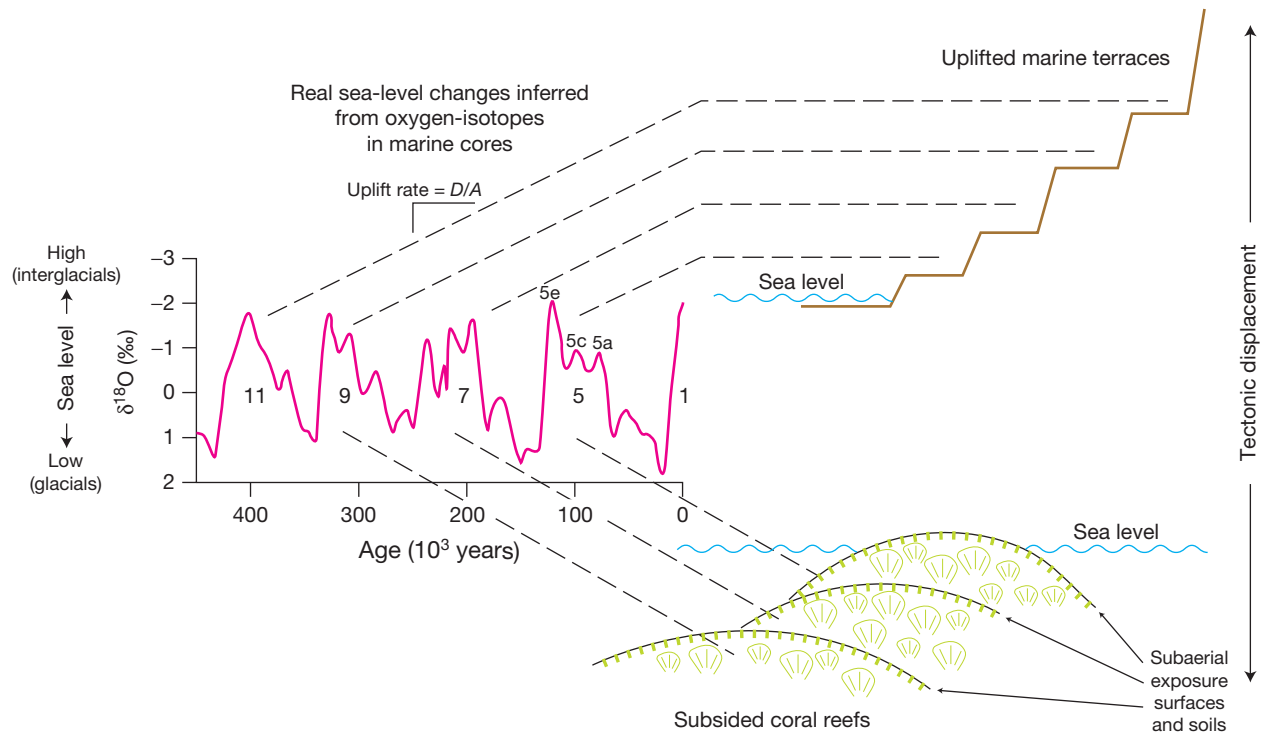


Figure 4 Hypothetical strandlines of uplifted and subsided coasts and their relations with highstands and lowstands of real sea level derived from the oxygen-isotope record from deep-sea foraminifers. Uplifted reef or wave-cut terraces approximately correspond with highstands (interglacial periods); subsided coral reefs correspond with lowstands (glacial periods) on some tropical coasts with high uplift rates. Modified from Muhs DR, Wehmler JF, Simmons KR, and York LL (2004) Quaternary sea-level history of the United States. In: Gillespie AR, Porter SC, and Atwater BF (eds.) *The Quaternary Period in the United States*, pp. 147–183. Amsterdam: Elsevier.

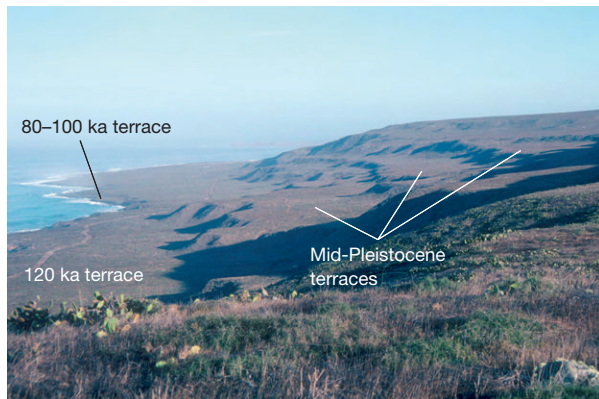


Figure 5 View of marine terraces on the west coast of San Clemente Island, about 80 km northwest of San Diego, California, looking northward. See references in Muhs et al. (2004). Photograph by Dan Muhs.

slip at depth is reflected in the variability in displacement rates recorded by strandlines above faults and folds (Figure 12; Burbank and Anderson, 2001; Merritts, 1996; Ota and Yamaguchi, 2004). In Chile, the M9.5 1960 earthquake was larger than would have been predicted from historical records of Chilean earthquakes and plate convergence rates (Cisternas et al., 2005), whereas modeling of measurements of coastal uplift and subsidence during the M8.8 2010 earthquake farther

north suggest that it released most fault strain accumulated since the previous earthquake. As a result of such complex fault behavior, precise prediction of the time or magnitude of future earthquakes is not a practical goal (Figure 8(d)).

Volcanic Processes

Strandlines record extremely high rates of uplift or subsidence on the flanks of some active volcanoes. For example, on the island of Iwo Jima in the western Pacific, uplift rates vary between 100 and 800 mm year⁻¹ and average about 200 mm year⁻¹ for the past 800 years (Kaizuka, 1992). Vertical rates of volcanic deformation of more than 50 mm year⁻¹ also have been measured for short periods on several coasts, but are highly localized in time (less than a few decades) and space (generally within kilometers of volcanoes; Lajoie, 1986; McCalpin, 2009, ch. 4). Over periods of hundreds to hundreds of thousands of years, displacement rates associated with volcanic and other processes of magma generation – such as migrating hotspots like the Hawaiian Islands and the Yellowstone region of Idaho – are commonly at least an order of magnitude lower (<2–5 mm year⁻¹; Lajoie, 1986; Muhs et al., 2004).

Epeirogenic and Isostatic Tectonic Processes

Bloom (1998) distinguishes epeirogeny, or regional scale, vertical tectonic deformation of low amplitude relative to its

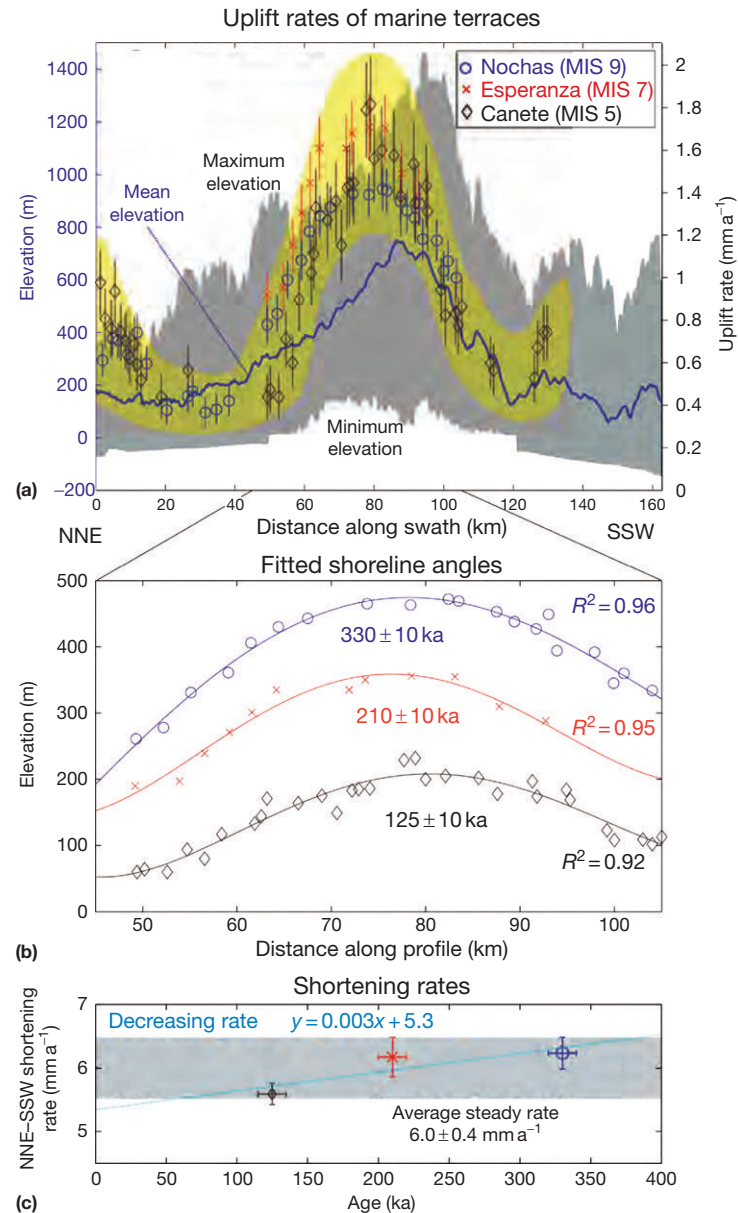


Figure 6 Marine terrace uplift and folding on the Arauco Peninsula of south-central Chile. (a) Surface uplift rates for the Cañete, Esperanza, and Nochas terrace surfaces and a maximum–minimum profile of peninsula topographic elevations parallel to the coast (gray shading). MIS indicates the marine isotope stage with which each terrace is correlated based on cosmogenic nuclide dating. (b) Elevation of shoreline angles (the upper edges of marine terraces) for the three main marine terraces (dated by cosmogenic methods as shown) from the main fold in the center of the peninsula (marked at bottom of (a)) and polynomial curves fitted to the elevations. (c) Shortening (amount of lateral fold compression) rates for the three marine terraces estimated from the curves of (b). Errors on the fold shortening rates are too large to determine whether or not shortening has decreased or remained constant over the past few hundred thousand years. Reproduced from Melnick D, Bookhagen B, Strecker MR, and Echtler HP (2009) Segmentation of megathrust rupture from fore-arc deformation patterns over hundreds to millions of years, Arauco peninsula, Chile. *Journal of Geophysical Research* 114: B01407. <http://dx.doi.org/10.1029/2008JB005788>.

wavelength that cannot be measured in a single outcrop, from orogeny, the tectonic processes that produce deformation of shorter wavelengths in mountain belts. During the seconds to minutes of earthquake faulting and folding in mountain belts, the crust, and mantle deform elastically. Over the decades to thousands of years of complete earthquake cycles or the glacioisostatic subsidence and uplift caused by ice-sheet expansion and retreat, the crust is strong and elastic whereas the

mantle is weak and ductile. Over millions of years, the topographic balance maintained by light, rigid crust floating on much denser, ductile mantle ensures that epeirogenic uplift and subsidence proceed at rates at least several orders of magnitude lower than deformation rates within mountain belts. Epeirogenic and isostatic processes are only briefly mentioned in this overview because they reflect surface and subsurface loading of the crust and flow of mantle material rather than

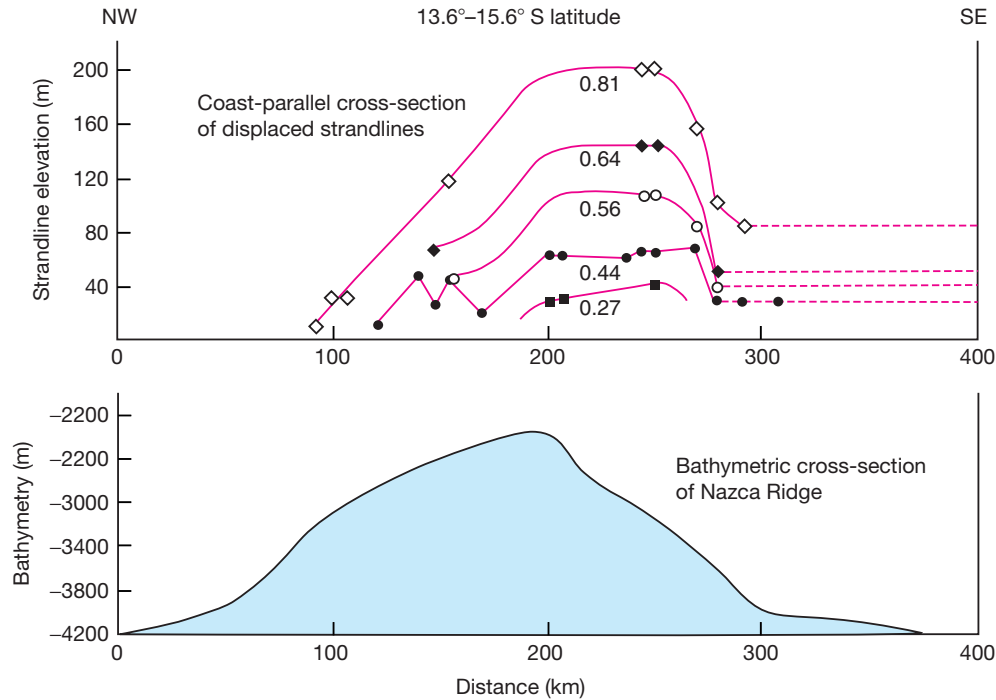


Figure 7 Coast-parallel pattern of displacement of Pleistocene strandlines in relation to the bathymetry of the 1200-m-high Nazca Ridge, which is subducting beneath this part of Peru. Symbols mark locations where elevations of marine terraces of different relative age were measured on coast-perpendicular transects. Solid lines connecting symbols show correlations of mapped terraces; dashed lines are tentative correlations. Amino acid ratios (mean ratio for each terrace shown under terrace correlation lines) on mollusks in terrace deposits show terrace relative ages. Terraces are displaced in proportion to the bathymetry of the ridge being subducted beneath them. Modified from Hsu JT (1992) Quaternary uplift of the Peruvian coast related to the subduction of the Nazca Ridge: 13.5 to 15.6 degrees south latitude. In: Ota Y, Nelson AR, and Berryman KR (eds.) Impacts of Tectonics on Quaternary Coastal Evolution. *Quaternary International* 15/16: 87–97.

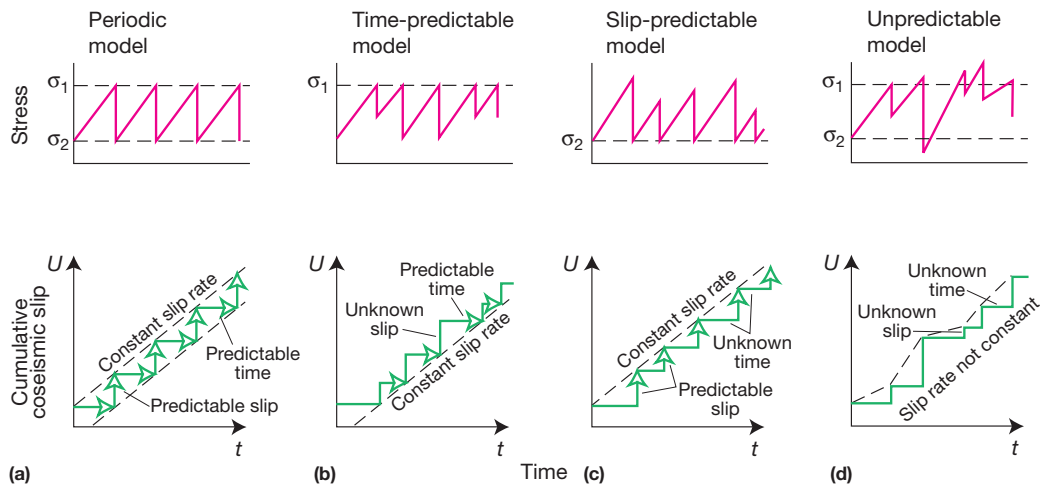


Figure 8 (a) Periodic model of the earthquake deformation cycle in which stress levels before and after an earthquake are known. In the periodic model the time to the next earthquake and the fault slip during the earthquake are predictable. (b) Time-predictable model of cycle where fault slip occurs at a constant level of stress. (c) Slip-predictable model where stress drops to a constant level during an earthquake. (d) Time-and-slip-unpredictable model of the earthquake cycle with changing rates of stress accumulation and release that result in variable slip rates. Modified from Shimazaki K and Nakata T (1980) Time-predictable recurrence model for large earthquakes. *Geophysical Research Letters* 7: 279–282.

sustained, primary tectonic processes, such as subduction and rifting, caused by horizontal motions of tectonic plates.

Glacioisostasy has produced the highest rates of regional strandline displacement (90 mm year^{-1}) yet measured. Strandlines near the centers of continental glaciation in North America

and Europe record hundreds of meters of postglacial uplift over a few thousand years in areas tens to hundreds of kilometers wide. Measurement and dating of the rapid uplift and subsidence of strandlines in glaciated areas yield information about the rigidity of the underlying mantle that helps us understand tectonic

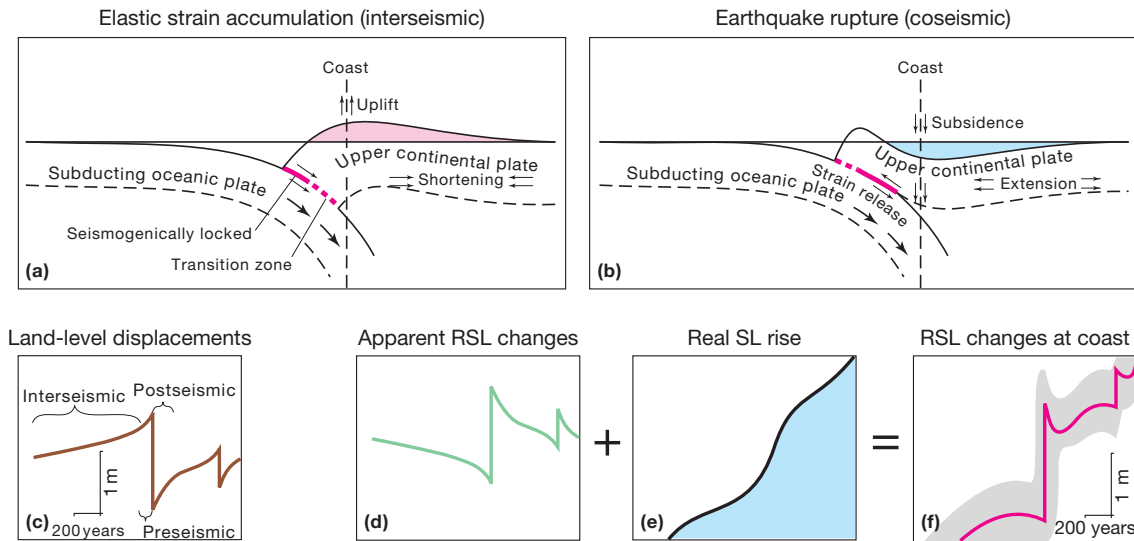


Figure 9 Schematic diagram showing deformation associated with a subduction-zone thrust fault beneath a coast during an earthquake cycle: (a) interseismic strain accumulation; (b) coseismic fault slip and strain release. Components of RSL change during two earthquake cycles: (c) land-level (strandline) displacement; (d) apparent sea-level changes; (e) real sea-level rise; (f) RSL history during two cycles. Preseismic and postseismic parts of the cycle (c) immediately precede and follow instantaneous coseismic fault displacement. In (a) and (b), the red solid and dashed lines mark the fault boundary between the upper and subducting plates; the red dashed line is a transition zone between locked (solid line) and plastically deforming parts of the plate boundary. Shading shows the relative amount of coastal uplift and subsidence of the upper plate during the interseismic and coseismic parts of the earthquake cycle. Shaded band in (f) shows typical uncertainty in some of the more precise sea-level indicators associated with late Holocene strandlines. The typical ± 0.5 -m uncertainty in late Holocene sea-level indicators makes identification of displacements of < 0.5 m difficult. Modified from Nelson AR, Shennan I, Long AJ (1996) Identifying coseismic subsidence in tidal-wetland stratigraphic sequences at the Cascadia subduction zone of western North America. *Journal of Geophysical Research* 101: 6115–6135.

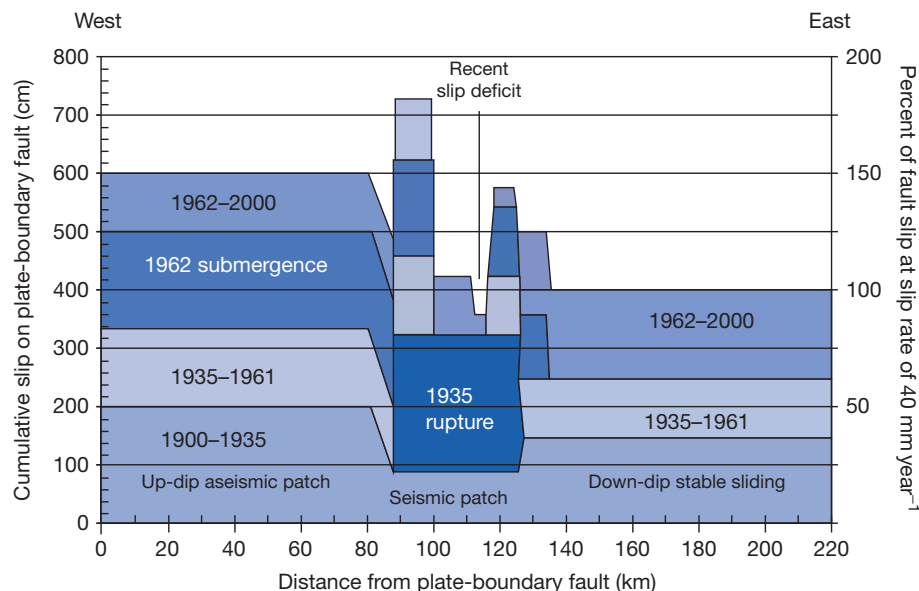


Figure 10 Coseismic and interseismic slip history for a transect perpendicular to the subduction-zone megathrust beneath western Sumatra developed from studies of twentieth-century coral microatolls (Figures 2 and 3). The dark areas represent the modeled amounts of coseismic slip during an earthquake in 1935 and an inferred slow earthquake in 1962. Lighter shading shows interseismic slip for labeled periods between earthquakes. The shallower and deeper parts of the fault have slipped continuously, whereas the fault at intermediate depths sticks and then slips during large earthquakes. Such patterns of coseismic and interseismic slip on the fault hint at the complex nature of slip on the fault planes of plate-boundary faults, even within a single cross-section perpendicular to one fault. Modified from Natawidjaja DH, Sieh K, Ward SN, et al. (2004) Paleogeodetic records of seismic and aseismic subduction from central Sumatran microatolls, Indonesia. *Journal of Geophysical Research* 109: B04306. <http://dx.doi.org/10.1029/2003>.

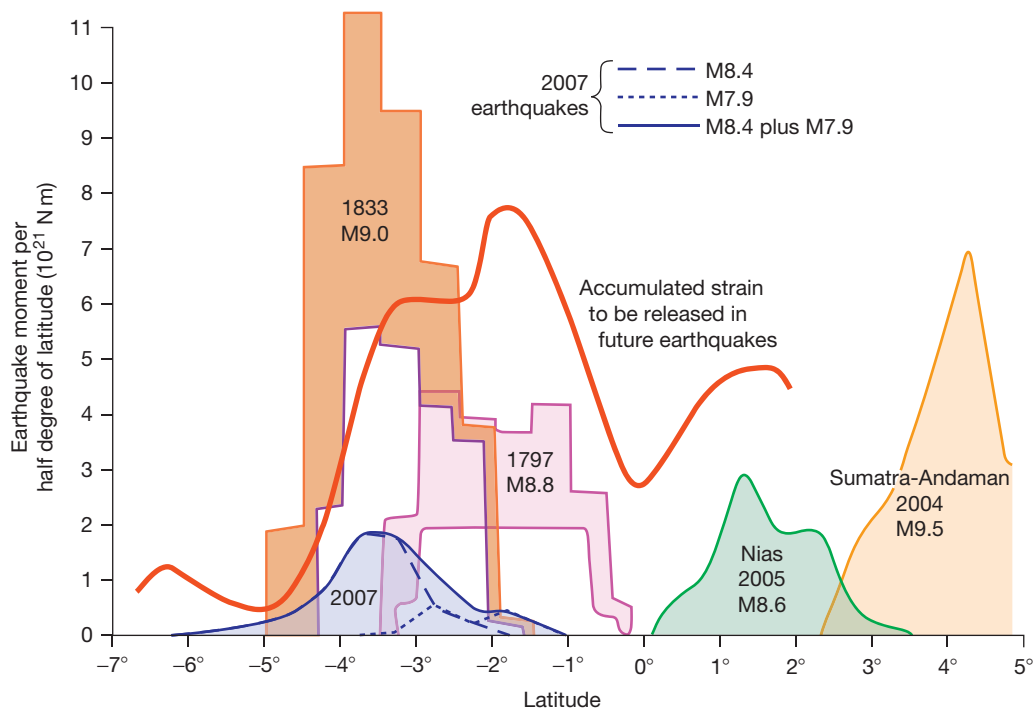


Figure 11 Comparison of the moment (energy released during an earthquake) released by earthquakes in 2007 with the moment deficit (unreleased strain) accumulated since the earthquakes in 1797 and 1833 along the western Sumatran subduction zone. Brown and pink lines are the upper and lower bounds of estimated moment release during the 1833 and 1797 earthquakes based on models using geodetic and coral-microatoll data. The red curve shows the accumulated moment deficit since the last great earthquake derived from modeling strain accumulation between great earthquakes. Modified from Konca AO, Avouac J-P, Sladen A, et al. (2008) Partial rupture of a locked patch of the Sumatra megathrust during the 2007 earthquake sequence. *Nature* 456: 631–635.

processes in unglaciated as well as glaciated regions (Milne and Shennan, 2007). The rapid decrease in vertical crustal stress that accompanies land uplift following deglaciation (ice-sheet retreat) has triggered earthquakes on faults in deglaciated areas of Canada and Scandinavia. As part of the redistribution of mantle material during deglaciation, coasts well beyond uplifting glaciated areas may subside due to a collapsing glacial forebulge.

Other regional epeirogenic processes that displace strandlines over hundreds to thousands of years include the loading of continental shelves with water (hydroisostasy; Milne and Shennan, 2007; Pirazzoli, 1996; Stewart and Vita-Finzi, 1998). Loading during deglaciation is primarily a response to an increase in ocean volume as a result of ice-sheet melting. The loading is accompanied by landward migration of viscous mantle material, which may uplift strandlines without any reduction in ocean volume (Murray-Wallace, 2007). Although commonly more localized (near major river deltas) than hydroisostasy, sediment loading on continental shelves over thousands of years also displaces strandlines.

Strandline Records of Deformation

Strandlines preserved over later Pleistocene (400–12 ka) and Holocene (<12 ka) timeframes record differing types and rates of real and apparent sea-level changes.

Pleistocene

Most Pleistocene standlines consist of emergent (uplifted) shoreline platforms formed during highstands of sea level on tectonically rising coasts when global ice volumes were low (Chappell et al., 1996; Woodroffe, 2007; Figure 1). Given sufficient rates of uplift ($>0.1 \text{ mm year}^{-1}$), successively higher wave-cut platforms, commonly 0.1–10 km wide, form a stair-step pattern along emergent coasts (Figure 4). Platforms may be erosional (high-wave-energy coasts; Figures 5 and 6) or constructional (low-energy coasts, such as reefs in tropical waters; Figure 12). Erosional platforms are typically veneered with several meters of sandy to gravelly, shallow-water sediment, occasionally containing datable fossils such as molluscs or corals. On most coasts, subaerial erosion makes strandlines older than 400 ka indistinct and difficult to date; where uplift rates are very high ($>4 \text{ mm year}^{-1}$) erosion typically removes strandlines older than 120 ka (Lajoie, 1986). Strandlines formed during global highstands are also present on slowly rising ($<0.1 \text{ mm year}^{-1}$) or stable coasts, but are difficult to identify and date (Muhs et al., 2004).

Strandlines marking Pleistocene lowstands of sea level are less well preserved and much less accessible than those marking highstands. Lowstand strandlines were commonly eroded during the transgressions and regressions to and from sea-level highstands, or they lie buried in subsided basins. Thus, most studies of lowstand strandlines are limited to coasts where

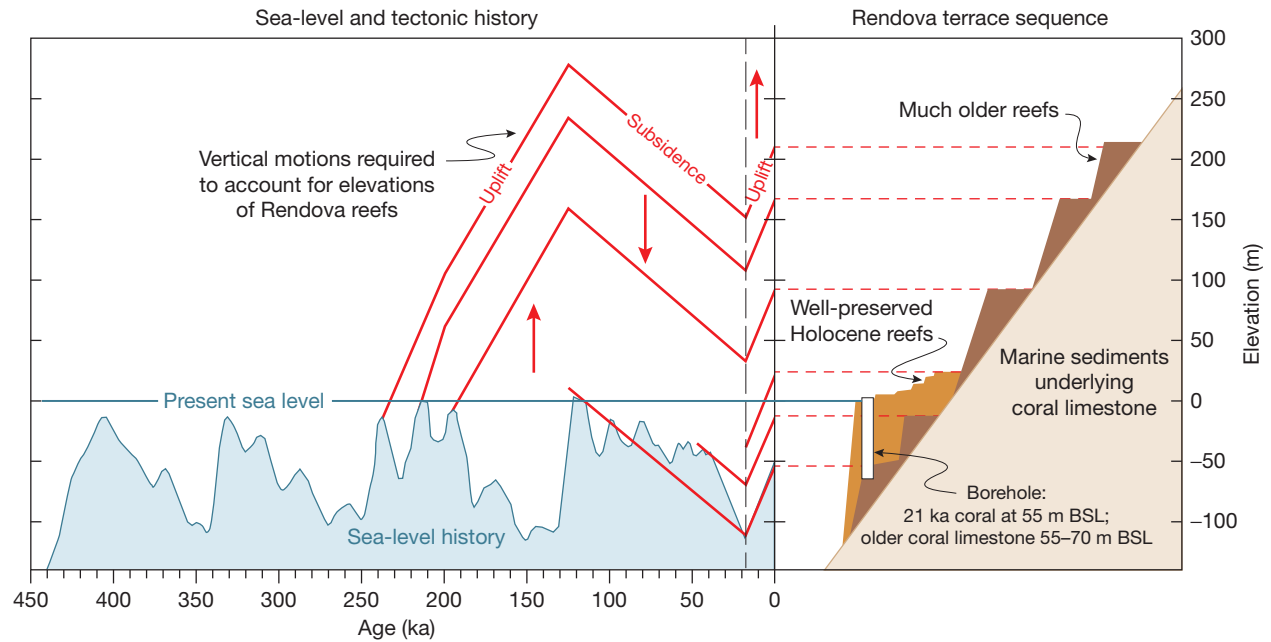


Figure 12 Reconstructed uplift history for central Rendova Island, New Georgia Group, southwestern Pacific, showing highly variable rates of uplift and subsidence for the past 270 000 years. The Holocene reef (6 ka) reaches 20 m above sea level and has uplifted $\sim 3 \text{ mm year}^{-1}$ at this location. Drilling penetrated the Last Glacial Maximum ($\sim 21 \text{ ka}$) at a depth of $\sim 55 \text{ m}$ below sea level (BSL) showing that the 3 mm year^{-1} rate dates from at least 21 ka. The reefs uphill from the 20-m Holocene reef surface are very weathered and are much older than 40–130 ka. Therefore, uplift that raised the Holocene reef to 20 m above sea level and the 21 ka reef to 55 m below sea level began recently, probably after $\sim 50 \text{ ka}$. Prior to the uplift, subsidence is required to account for the older reefs beneath the 21-ka reef and to explain the absence of 40–130-ka reefs uphill from the Holocene reef. The older reefs that are now at 50–300 m above sea level rose during a period of uplift prior to the subsidence. Modified from Taylor FW, Mann P, Bevis MG, et al. (2005) Rapid forearc uplift and subsidence caused by impinging bathymetric features: Examples from the New Hebrides and Solomon arcs. *Tectonics* 24: TC6005. <http://dx.doi.org/1029/2004TC001650>.

uplift rates are high enough to have raised lowstand features above present sea level ($>0.3\text{--}1.0 \text{ mm year}^{-1}$), or where transects of drill cores access coral reefs formed during both high and lowstands of sea level (Figures 4 and 12; Bard et al., 1990). For example, in deep valleys cut into rapidly uplifting terraces of the Huon Peninsula of Papua New Guinea, small deltas marking lowstands of sea level are protected from erosion by coral reefs deposited during higher sea levels (Chappell et al., 1996). Although geophysical methods such as side-scan sonar allow detailed mapping of submerged strandlines (McCalpin, 2009, ch. 2B), such methods are costly and dating the strandlines is difficult. In Hawaii, where rates of subsidence are particularly high, lowstand strandlines as old as 475 ka have been dated by uranium-series methods (Muhs et al., 2004).

Dating of Pleistocene strandlines over the past three decades has been improved through comparisons with the oxygen-isotope record from marine cores and the RSL records of emergent coral-reef strandlines on New Guinea and Barbados (Figure 4; Woodroffe, 2007). Assuming a constant coastal uplift rate, subtraction of the amount of uplift from elevations of strandlines well dated by uranium-series methods on fossil corals yields paleosea-level curves extending back to about 300 ka. On uplifting coasts, where only one or two strandlines have been dated by numerical methods, assumption of an average uplift rate and correlation of the elevations of strandlines of unknown age with the New Guinea and Barbados records yields inferred ages for undated terraces (Figure 12).



Figure 13 Late Holocene strandlines along the central east coast of Isla Mocha, a 12-km-long island 30 km off the coast of Chile above the Chilean subduction zone (latitude 38° S). Some of the strandlines were uplifted suddenly during great earthquakes on the Chilean subduction zone, whereas others were probably built by storm waves during periods of rapid interseismic uplift (Figure 14). Reproduced from Nelson AR and Manley WF (1992) Holocene coseismic and aseismic uplift of Isla Mocha, south-central Chile. In: Ota Y, Nelson AR, and Berryman KR (eds.) *Impacts of Tectonics on Quaternary Coastal Evolution. Quaternary International* 15/16: 61–76.

Recent dating of Pleistocene strandlines in North America and elsewhere shows significant differences in the length and timing of some highstands (Muhs et al., 2004).

Regional patterns of Pleistocene strandline displacement commonly reflect primary tectonic processes related to

horizontal plate motions, whereas local patterns reflect deformation on secondary faults and folds in the shallow crust, which respond to more localized stresses (Lajoie, 1986). Regional patterns are most clearly shown by maps of displacement isobases, which generally mimic topographic contours,

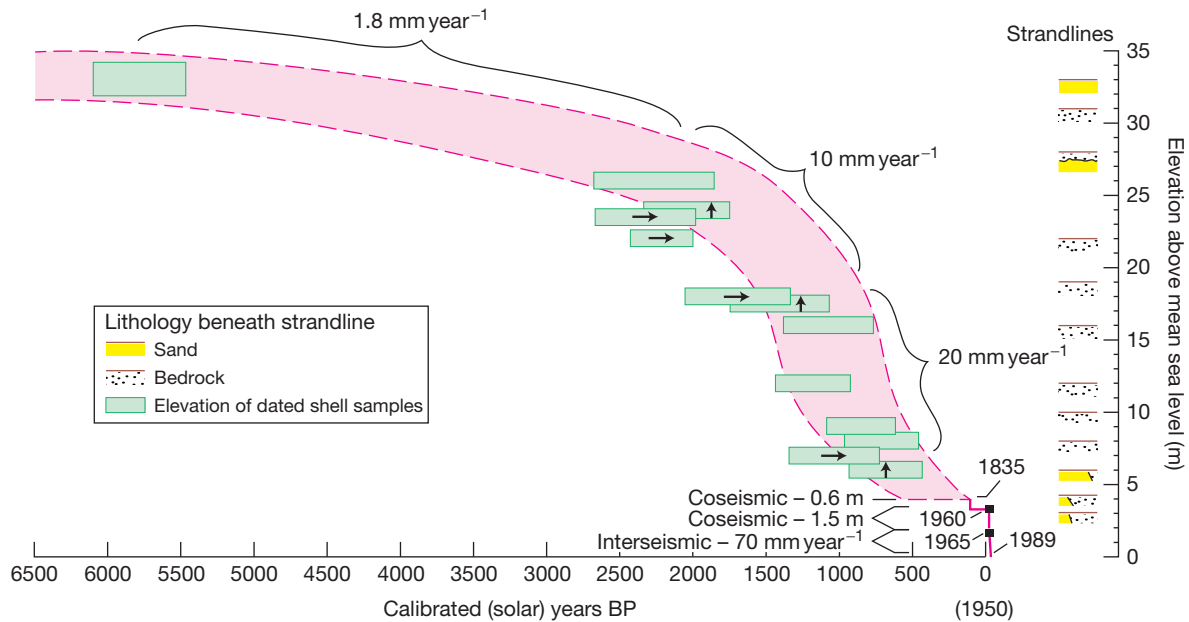


Figure 14 RSL history of Isla Mocha (Figure 13). Radiocarbon ages on shells from uplifted strandlines suggest an increasing rate of uplift in the late Holocene. The dashed line outlines an envelope, encompassing the most reliable ^{14}C ages, of RSL change since 6 ka. Real sea-level history, which is poorly known in this region, must be subtracted from the envelope to determine changes in uplift rate. Coseismic uplift of the island in 1835 and 1960 suggests that late Holocene uplift is the result of similar great earthquakes during the past few thousand years. However, the extremely high rate of interseismic uplift since at least 1965 (70 mm year^{-1}) suggests that some raised strandlines ringing the island rose during interseismic as well as coseismic parts of the earthquake cycle. Horizontal arrows in boxes mark shell samples that are probably reworked from older strandlines. Small vertical arrows show samples that were probably deposited $>1 \text{ m}$ below a contemporaneous shoreline. Lithologies of strandlines suggest that some are erosional and others depositional. Modified from Nelson AR and Manley WF (1992) Holocene coseismic and aseismic uplift of Isla Mocha, south-central Chile. In: Ota Y, Nelson AR, and Berryman KR (eds.) Impacts of Tectonics on Quaternary Coastal Evolution. *Quaternary International* 15/16: 61–76.



Figure 15 Genroku marine terrace at Mera on the Boso Peninsula of Japan uplifted 4.8 and 1.5 m during the great Kanto earthquakes of 1703 and 1923, respectively (Matsuda et al., 1978). Houses in background are on a higher terrace estimated from the average uplift rate to date from about 3.5 ka.

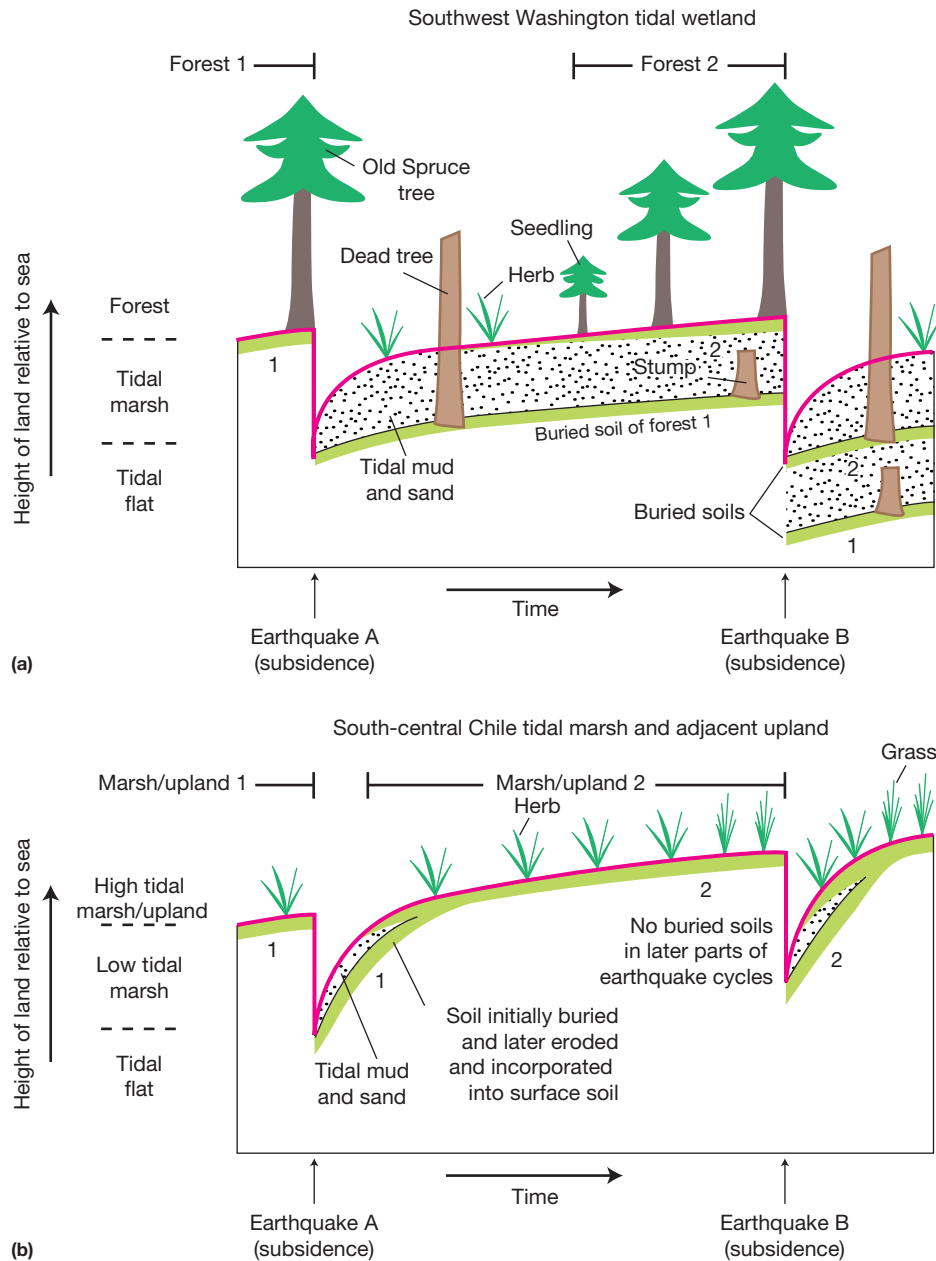


Figure 16 Simplified models of land-level change in estuary tidal wetlands during two earthquake cycles above subduction zones of southwest Washington and south-central Chile. Heavy line shows inferred land-level change relative to sea level. In (a), buried coastal forest and wetland soils (1 and 2) are quickly buried by many decimeters of tidal mud and sand following coseismic subsidence due to modest postseismic recovery and slowly rising sea level. This process creates stacks of well-preserved buried soils that can be dated by radiocarbon. In (b), similar subsidence, largely complete postseismic recovery, and slowly falling sea level result in minimal burial and later erosion of tidal marsh soils; buried soils are not preserved.

but require areally extensive strandline or other paleoelevation measurements. Faults, folds, and broad regional tilting resulting from fault slip near plate boundaries are illustrated by isobase maps of coastal areas of Japan, Taiwan, New Zealand, and Chile. More commonly, sequences of strandlines illustrate progressive folding or faulting along coast-parallel, vertical planes that transect active folds and faults (Keller and Pinter, 2002; Kaizuka, 1992; Stewart and Vita-Finzi, 1998). For example, regionally deformed marine terraces along the coasts of southern Peru and south-central Chile record folding over the

past half million years caused by oblique subduction of the Nazca oceanic ridge beneath Peru and the northward movement of a sliver of Chilean continental crust, respectively (Figures 6 and 7).

Rates of uplift measured from Pleistocene strandlines rarely exceed 2 mm year^{-1} , even near plate boundaries with the highest rates of slip. Regional isostatic adjustment of the crust beneath more rapidly uplifting coasts apparently prevents higher Pleistocene rates (Lajoie, 1986). Graphical comparisons of Pleistocene strandline elevations on many emergent coasts

with the well-dated New Guinea strandlines show that assumptions of constant uplift are reasonable for many sequences in the range 100–500 ka (Lajoie, 1986); direct evidence of constant rates in New Guinea includes converging coast-parallel strandline profiles that show progressive tilting at a constant rate since 125 ka (Burbank and Anderson, 2001). Exceptionally high rates of Pleistocene uplift of 8 mm year^{-1} followed by rapid subsidence measured from reef corals in the Solomon Islands of the southwest Pacific can be explained by subduction of a seamount over the past 200 000 years (Figure 12). The high rates of subsidence as well as uplift measured from some Holocene strandlines, tide gauges, and by global positioning systems are short-term rates that include only parts of earthquake cycles and are averaged over much longer Pleistocene time spans. For example, on the Muroto Peninsula in Shikoku, Japan, uplift rates for the 120- and 6-ka strandlines are about 2 mm year^{-1} , but episodic uplift of the past 300 years was six times as high and tide-gauge data indicate subsidence during part of this period (Ota and Yamaguchi, 2004; Sholtz, 2002). Similarly, parts of the Oregon coast are currently subsiding at rates three times the uplift rates of 0.3 mm year^{-1} recorded by nearby 125-ka marine terraces.

Holocene

The Holocene timeframe includes an early Holocene period of rapidly rising sea level (12 ka to about 6 ka on many coasts) and a following, late Holocene period of slowly rising, stable, or slowly falling sea level, depending on how glacioisostatic adjustments affect real sea level in different parts of the globe (Figure 1; Milne and Shennan, 2007; Murray-Wallace, 2007; Pirazzoli, 1996). Early Holocene strandlines provide little information about tectonic processes because most were eroded and submerged by a rapidly rising sea and are now largely inaccessible. In contrast, late Holocene strandlines provide insight into tectonic processes at a level of detail not possible with Pleistocene strandlines. This is especially true of strandlines formed during the past 2000 years when real sea-level changes on most coasts were $< \pm 2 \text{ m}$. For this reason, RSL changes during the past 2000 years are largely tectonic (apparent) changes (Burbank and Anderson, 2001; Lajoie, 1986).

Rapidly uplifting ($> 1\text{--}2 \text{ mm year}^{-1}$) late Holocene strandline sequences yield some of the most detailed archives of large plate-boundary earthquakes. Historically uplifted marine terraces in Japan, Alaska, and New Zealand are the basis for the time-predictable and slip-predictable models of earthquake recurrence discussed above (Matsuda et al., 1978; Shimazaki and Nakata, 1980). These widely influential models have been used to develop earthquake histories from Holocene marine terraces 1–15 m above sea level on tectonically active coasts and even to forecast future earthquakes at well-dated sites (Figures 13, 14, and 15; Merritts, 1996; Ota and Yamaguchi, 2004; Stewart and Vita-Finzi, 1998).

As attractive as the models of coseismic displacement are, no strandline records fit the models exactly and the complexities of coastal and earthquake-cycle processes caution against using the models to forecast future coseismic strandline displacement. First, most emergent Holocene strandlines record localized displacement related to folding or faulting above thrust faults in the upper few kilometers of crust. For this

reason, most uplifted strandlines record displacement on local rather than regional plate-boundary faults, usually in compressive tectonic settings (Lajoie, 1986; Melnick et al., 2009; Nelson and Manley, 1992; Ota and Yamaguchi, 2004; Plafker, 1972). Because local folds and faults respond only intermittently to regional stresses, it is unlikely that strandlines above a shallow fold or fault will record all large earthquakes on a deeper plate-boundary fault, or that strandline spacing will reflect the relative amount of slip on plate-boundary faults over either Pleistocene or Holocene timeframes (Briggs et al., 2008; Burbank and Anderson, 2001). Secondly, geodetic and tide-gauge data from Japan, Alaska, Chile, Indonesia, and other coasts near plate-boundary faults show that RSL may rise or fall at rapid nonlinear rates during interseismic parts of the earthquake cycle (Wang, 2007). If major storms create distinct but short-lived shorelines on such rapidly rising coasts, storm shorelines could easily be mistaken for strandlines formed by meters of coseismic uplift (Figures 13 and 14).

Finally, even strandlines that are distinct from higher and lower strandlines may span multiple earthquake cycles. For example, on the Boso and Muroto peninsulas of Japan, the lowest marine terrace was uplifted during two or three closely spaced historical earthquakes rather than during a single large earthquake (Figure 15; Ota and Yamaguchi, 2004; Sholtz, 2002). Even above subduction-zone megathrusts

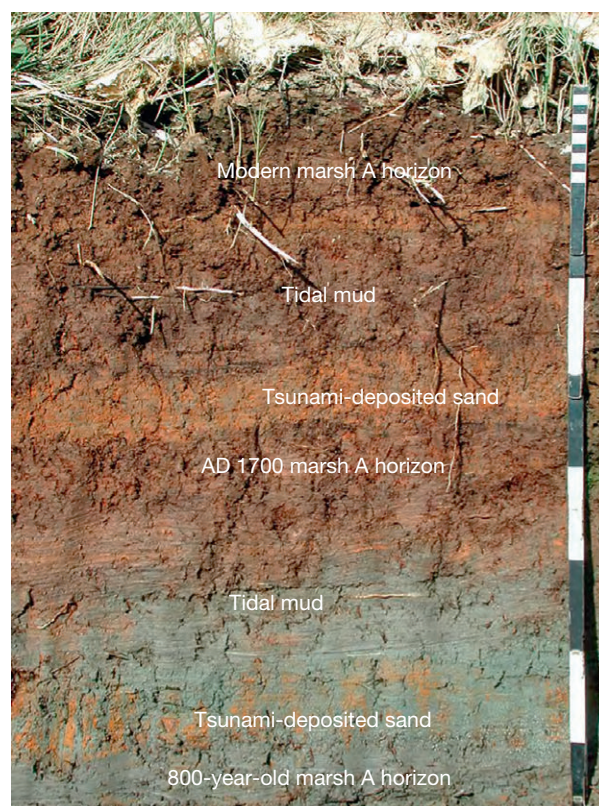


Figure 17 Section of sediment in a tidal channel at Alsea Bay, on the central Oregon coast, showing tidal-marsh A horizons formed on tidal mud prior to coastal subsidence during great earthquakes about 300 (AD 1700) and 800 years ago, and the A horizon of the present marsh. Sandy beds deposited by tsunamis immediately overlie the two older A horizons. Scale in decimeters.

where shallow, localized faults are apparently inactive, such as on Nias Island off western Sumatra, the modest (<11 m) uplift of mid-Holocene terraces suggests that none record uplift during only a single earthquake. The mismatch between the 0–2.9 m of uplift of the island during the M8.7 2005 earthquake and the 1.5 to -0.2 mm year $^{-1}$ net rates of uplift measured from mid-Holocene terraces show that strain accumulation on this part of the megathrust is elastic with largely complete recovery of coseismic uplift during interseismic periods (Briggs et al., 2008).

In contrast to most emergent late Holocene strandlines, the most widely studied submergent strandlines record regional (rather than local) deformation during the largest known earthquakes. In the Pacific Northwest of the U.S. and Canada, Alaska, Chile, and Japan, tidal wetland soil buried by tidal mud and sand testify to repeated coseismic subsidence of wetlands

during great earthquakes on underlying subduction-zone megathrusts many hundreds of kilometers long (Figure 16; Atwater and Hemphill-Haley, 1997; Carver and Plafker, 2008; Cisternas et al., 2005; McCalpin, 2009, ch. 5; Satake and Atwater, 2007; Sawai et al., 2004; Shennan and Hamilton, 2006). The modern analogs for such records of coseismic subsidence are regions of 80 000 and 110 000 km 2 , in Chile and Alaska, respectively, that subsided 1–2 m during M9 earthquakes in the 1960s (Plafker, 1972). Study of such decimeter-to-meter-scale records of regional subsidence requires easily dated strandlines with precise indicative meanings in settings favorable for strandline preservation. For example, the well-preserved stacks of wetland soils interbedded with tidal mud beneath the modern tidal marshes of Oregon (Figure 17) are more the result of slow sea-level rise, tidal sedimentation, and gradual sediment compaction than of unrecovered coseismic

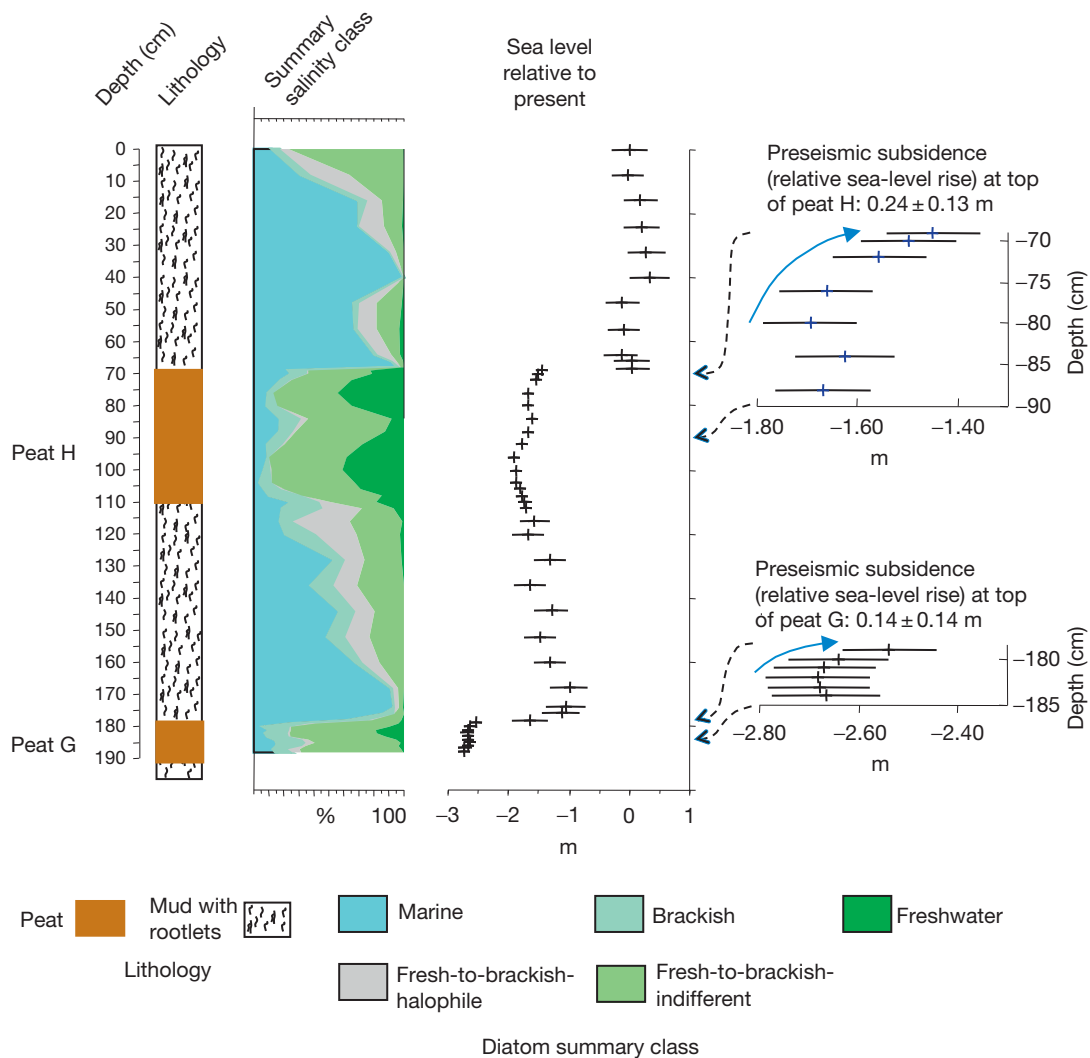


Figure 18 Changes in diatom summary water-salinity classes, which reflect changes in tidal elevations and environments, in a sediment section spanning two subduction-zone earthquake cycles near Girdwood, south-central Alaska. Corresponding RSL changes are reconstructed from diatom species data using transfer function methods. Two short periods of preseismic submergence immediately prior to the substantial coseismic subsidence marked by the contacts at the tops of peat G and peat H are highlighted on the right. Modified from Shennan I and Hamilton SL (2006) Coseismic and preseismic subsidence associated with great earthquakes in Alaska. *Quaternary Science Reviews* 25: 1–8. <http://dx.doi.org/10.1016/j.quascirev.2005.09.002>.

subsidence during great earthquakes. Along much of the coast of south-central Chile, the region of the largest earthquake of the past century (M9.5 in 1960), similar coseismically subsided soils are not widely preserved. This is probably due to largely complete postseismic recovery following many earthquakes, and to shoreline erosion and retreat during the slow fall of RSL since 6 ka (Figure 16). Recent work in Sumatra shows that subsided wetland soils may record past subduction-zone earthquakes even in tropical regions where organic material decays rapidly.

Ongoing studies of late Holocene strandlines with the most precise indicative meanings show promise for unparalleled reconstruction of strandline displacement during past earthquake cycles above subduction-zone megathrusts. Changes in the proportions of tidal microfossils (foraminifers, diatoms,

and pollen) in coastal stratigraphic sequences, first used in the 1930s to infer late Holocene sea-level changes in Europe, show decimeters to meters of coseismic subsidence during great earthquakes at the Cascadia subduction zone of western North America (Atwater and Hemphill-Haley, 1997; Hawkes et al., 2011). Similar biostratigraphic studies in wetlands above the southeastern Alaska megathrust suggest decades of preseismic subsidence just prior to as many as six past great earthquakes, perhaps due to slow slip on the megathrust just prior to great earthquakes (Figure 18). In contrast, diatom biostratigraphy in northern Japan shows about a meter of gradual coastal uplift following great earthquakes; such crustal behavior implies slow, deep, postseismic slip on the megathrust like that used to explain tide-gauge measurements in southern Chile since 1960 (Figure 19; Satake and Atwater, 2007). The

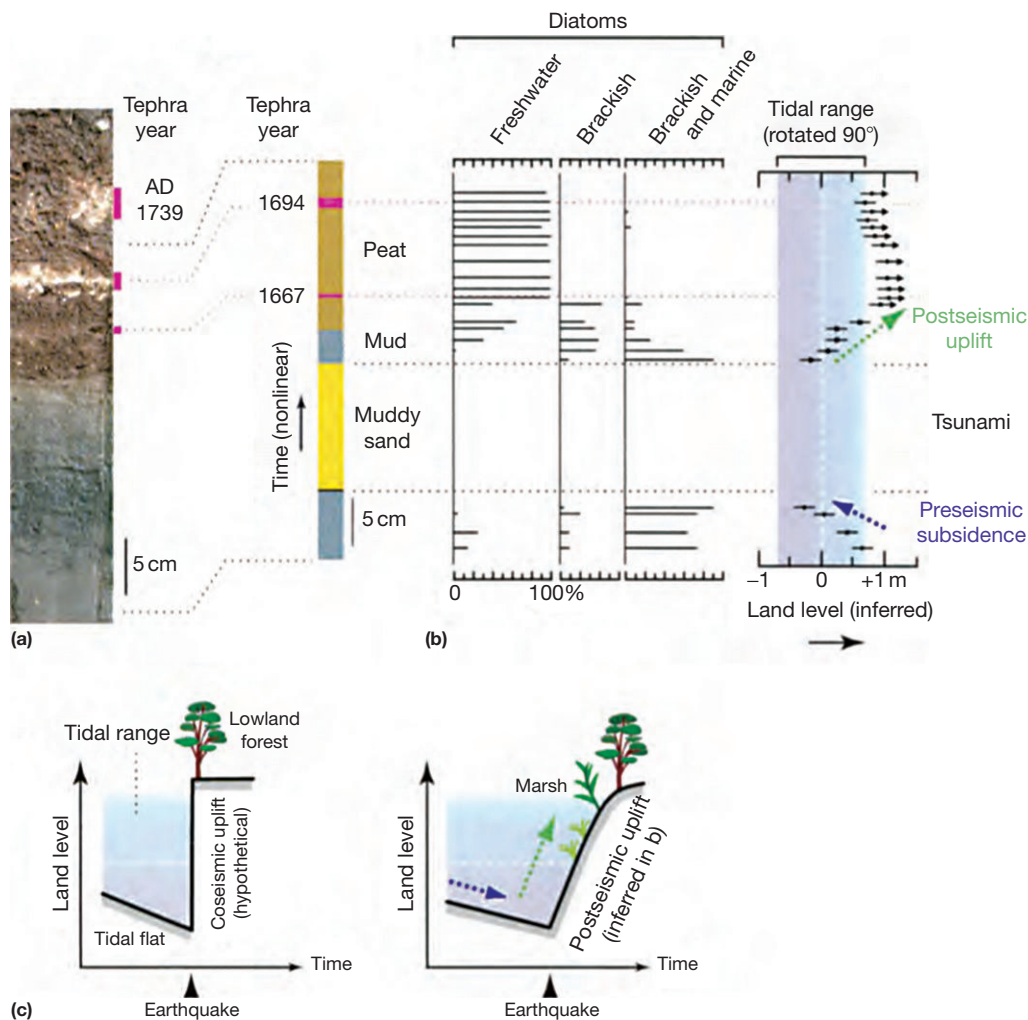


Figure 19 Stratigraphic evidence for postseismic uplift beneath a coastal marsh at Mochirippu, above the eastern Hokkaido (Japan) subduction zone. (a) Photograph and type of sediment in part of a core from the marsh; three tephra (volcanic ashes) of known age date the upper part of the core. (b) Changes in diatom summary water-salinity classes from which gradual preseismic subsidence and postseismic uplift are inferred before and after deposition of a muddy sand by a tsunami. Error bars for height estimates calculated with transfer function methods span two standard deviations. (c) Model of coseismic (sudden) uplift that converts a tide flat into a lowland forest following a great subduction-zone earthquake (Figures 13 and 15); model of gradual postseismic uplift of a tide flat inferred for Mochirippu. Modified from Sawai Y, Satake K, Kamataki T, et al. (2004) Transient uplift after a 17th-century earthquake along the Kuril subduction zone. *Science* 306: 1918–1920.

most precise studies of fossil coral microatolls, above the western Sumatran subduction zone, track complex interseismic as well as coseismic changes in apparent sea level along strike and perpendicular to the subduction zone (Figures 2, 3, 10, and 11). Briggs et al. (2006) used mostly coral-microatoll measurements supplemented with global positioning system data to construct some of the most detailed strandline maps yet compiled (0.25-m contours) of deformation above the megathrust during the M9.2 2004 and M8.7 2005 earthquakes. Corals allow uplift from the two earthquakes only 3 months apart to be measured separately (Figure 3). The high precision and short timescale of such methods is not only bridging the gap between geologic and instrumental studies of sea-level change, but in the case of coral microatolls is supplementing the modern instrumental record (Taylor et al., 2008).

Such high-resolution studies of sea-level changes throughout past earthquake cycles have considerable practical as well as theoretical value. For example, at Cascadia determining whether the coast immediately rebounded or continued to subside following the subsidence that occurred during past great earthquakes will help show which of two competing models of earthquake cycles should be used to assess earthquake hazards in the region (Wang, 2007). In the Solomon Islands, a study of coral microatolls displaced during the 2007 M8.1 earthquake showed for the first time that a single earthquake can rupture two structurally separate subducting plates (Taylor et al., 2008). This finding illustrates that the assumption that earthquakes do not rupture across major structural boundaries, which is widely used in estimating the maximum size of earthquakes in fault zones, may be invalid. Application of high-resolution methods to the study of local faults and folds in suitable coastal environments should advance our understanding of tectonic processes in non-subduction-zone settings.

Acknowledgments

This overview owes much to Lajoie's (1986) review, which has influenced many over the past quarter century. My research on Quaternary strandlines has been supported by the Earthquake Hazards Program of the U.S. Geological Survey, a USGS Gilbert Fellowship, and the U.S. Nuclear Regulatory Commission. Lee-Ann Bradley did many of the graphics. Reviews by Dan Muhs, Rob Witter, and Rich Briggs improved this overview. Muhs provided a photograph.

See also: Amino Acid Dating; U-Series Dating. **Glaciation, Causes:** Tectonic Uplift and Continental Configurations. **Luminescence Dating:** Optical Dating; Thermoluminescence. **Radiocarbon Dating:** AMS Radiocarbon Dating. **Sea Level Studies:** Coral Records of Relative Sea-Level Changes; Eustatic Sea-Level Changes Since the Last Glacial Maximum; Geomorphological Indicators; Isostasy; Glaciation-Induced Sea-Level Change; Microfossil-Based Reconstructions of Holocene Relative Sea-Level Change; Overview. **Sea-Levels, Late Quaternary:** Late Quaternary Relative Sea-Level Changes in High Latitudes.

References

- Atwater BF and Hemphill-Haley E (1997) Recurrence intervals for great earthquakes of the past 3,500 years at northeastern Willapa Bay, p. 108. Washington. U.S. Geological Survey Professional Paper 1576.
- Bard E, Hamelin B, and Fairbanks RG (1990) U-Th ages obtained by mass spectrometry in corals from Barbados: Sea level during the past 130,000 years. *Nature* 346: 456–458.
- Bloom AL (1998) *Geomorphology: A Systematic Analysis of late Cenozoic Landforms*, 3rd edn. Upper Saddle River, NJ: Prentice Hall.
- Briggs RW, Sieh K, Amidon WH, et al. (2008) Persistent elastic behavior above a megathrust rupture patch: Nias Island, west Sumatra. *Journal of Geophysical Research* 113: B12406. <http://dx.doi.org/10.1029/2008JB005684>.
- Briggs RW, Sieh K, Meltzner AJ, et al. (2006) Deformation and slip along the Sunda megathrust in the great 2005 Nias-Simeulue earthquake. *Science* 311: 1897–1901.
- Burbank DW and Anderson RS (2001) *Tectonic Geomorphology*. London: Blackwell Science.
- Carver G and Plafker G (2008) Paleoseismicity and neotectonics of the Aleutian subduction zone – An overview. In: Freymueller JT, Haussler PJ, Wesson RL, and Ekstrom G (eds.) *Active Tectonics and the Seismic Potential of Alaska*. *Geophysical Monograph* 179, pp. 43–63. Washington, DC: American Geophysical Union.
- Chappell J, Omura A, Esat T, et al. (1996) Reconciliation of late Quaternary sea levels derived from coral terraces at Huon Peninsula with deep sea oxygen isotope records. *Earth and Planetary Science Letters* 141: 227–236.
- Cisternas M, Atwater BF, Torrejon F, et al. (2005) Predecessors of the giant 1960 Chile earthquake. *Nature* 437: 404–407.
- Hawkes AD, Horton BP, Nelson AR, Vane Y, and Sawai Y (2011) Coastal subsidence in Oregon, USA, during the giant Cascadia earthquake of AD 1700. *Quaternary Science Reviews* 30: 364–376.
- Kaizuka S (1992) Coastal evolution at a rapidly uplifting volcanic island: Iwo-Jima, western Pacific Ocean. In: Ota Y, Nelson AR, and Berryman KR (eds.) *Impacts of tectonics on Quaternary coastal evolution*. *Quaternary International* 15(16): 7–16.
- Keller EA and Pinter K (2002) *Active Tectonics: Earthquakes, Uplift, and Landscape*, 2nd edn. Upper Saddle River, NJ: Prentice Hall.
- Lajoie KR (1986) Coastal tectonics. In: Wallace RE (ed.) *Active Tectonics*, pp. 95–124. Washington, DC: National Academy of Sciences, National Academy Press.
- Matsuda T, Ota Y, Ando M, and Yonekura N (1978) Fault mechanism and recurrence time of major earthquakes in southern Kanto district, Japan, as deduced from coastal terrace data. *Geological Society of America Bulletin* 89: 1610–1618.
- McCalpin JP (2009) *Paleoseismology*, 2nd edn. San Diego: Elsevier.
- Melnick D, Bookhagen B, Strecker MR, and Echler HP (2009) Segmentation of megathrust rupture from fore-arc deformation patterns over hundreds to millions of years, Arauco peninsula, Chile. *Journal of Geophysical Research* 114: B01407. <http://dx.doi.org/10.1029/2008JB005788>.
- Merritts DJ (1996) The Medocino triple junction: Active faults, episodic coastal emergence, and rapid uplift. *Journal of Geophysical Research* 101: 6051–6070.
- Milne G and Shennan I (2007) Isostasy. In: Shennan I (ed.) *Quaternary sea level*. In: Elias S (ed.) *Encyclopedia of Quaternary Science*, pp. 3043–3051. Amsterdam: Elsevier.
- Muhs DR, Wehmler JF, Simmons KR, and York LL (2004) Quaternary sea-level history of the United States. In: Gillespie AR, Porter SC, and Atwater BF (eds.) *The Quaternary Period in the United States*, pp. 147–183. Amsterdam: Elsevier.
- Murray-Wallace CV (2007) Eustatic sea-level changes since the last glaciation. In: Shennan I (ed.) *Quaternary sea level*. In: Elias S (ed.) *Encyclopedia of Quaternary Science*, pp. 3034–3043. Amsterdam: Elsevier.
- Nelson AR and Manley WF (1992) Holocene coseismic and aseismic uplift of Isla Mocha, south-central Chile. In: Ota Y, Nelson AR, and Berryman KR (eds.) *Impacts of Tectonics on Quaternary Coastal Evolution*, *Quaternary International* 15(16): 61–76.
- Ota Y and Yamaguchi M (2004) Holocene coastal uplift in the western Pacific Rim in the context of late Quaternary uplift. *Quaternary International* 120: 105–117.
- Pirazzoli PA (1996) *Sea-level Changes*. Chichester: Wiley.
- Plafker G (1972) Alaskan earthquake of 1964 and Chilean earthquake of 1960: Implications for arc tectonics. *Journal of Geophysical Research* 77: 901–925.
- Satake K and Atwater BF (2007) Long-term perspectives on giant earthquakes and tsunamis at subduction zones. *Annual Review of Earth and Planetary Sciences* 35: 349–374.
- Sawai Y, Satake K, Kamataki T, et al. (2004) Transient uplift after a 17th-century earthquake along the Kuril subduction zone. *Science* 306: 1918–1920.
- Shennan I and Hamilton SL (2006) Coseismic and preseismic subsidence associated with great earthquakes in Alaska. *Quaternary Science Reviews* 25: 1–8. <http://dx.doi.org/10.1016/j.quascirev.2005.09.002>.
- Shimazaki K and Nakata T (1980) Time-predictable recurrence model for large earthquakes. *Geophysical Research Letters* 7: 279–282.

- Sholtz C (2002) *The Mechanics of Earthquakes and Faulting*, 2nd edn. Cambridge: Cambridge University Press.
- Stewart IS and Vita-Finzi C (eds.) (1998) *Coastal Tectonics*. London: The Geological Society.
- Taylor FW, Briggs RW, Frohlich C, et al. (2008) Rupture across arc segment and plate boundaries in the 1 April 2007 Solomons earthquake. *Geoscience* 1: 253–257. <http://dx.doi.org/10.1038/ngeo159>.
- van de Plassche O (ed.) (1986) *Sea-level Research: A Manual for the Collection and Evaluation of Data*. Norwich, UK: Geo Books.
- Wang K (2007) Elastic and viscoelastic models of crustal deformation in subduction earthquake cycles. In: Dixon T and Moore JC (eds.) *The Seismogenic Zone of Subduction Thrust Faults*, pp. 540–575. New York: Columbia University Press.
- Woodroffe CD (2007) Coral records. In: Shennan I (ed.) *Quaternary sea level*. In: Elias S (ed.) *Encyclopedia of Quaternary Science*, pp. 3006–3015. Amsterdam: Elsevier.
- Yeats RS, Sieh KE, and Allen CR (1997) *The Geology of Earthquakes*. New York: Oxford University Press.

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Sea Surface Temperature Reconstruction *see* **Paleoceanography**: Paleoceanography An Overview; **Paleoceanography, Biological Proxies**: Alkenone Paleothermometry Based on the Haptophyte Algae; Biomarker Indicators of Past Climate; Coccolithophores; Dinoflagellates; Marine Diatoms; Planktic Foraminifera; Temperature Proxies, Census Counts; **Paleoceanography, Physical and Chemical Proxies**: Carbon Cycle Proxies ($\delta^{11}\text{B}$, $\delta^{13}\text{C}_{\text{calcite}}$, $\delta^{13}\text{C}_{\text{organic}}$, Shell Weights, B/Ca, U/Ca, Zn/Ca, Ba/Ca); Mg/Ca and Sr/Ca Paleothermometry from Calcareous Marine Fossils; Oxygen-Isotope Stratigraphy of the Oceans; **Paleoceanography, Records**: Late Pleistocene North Atlantic; Late Pleistocene North Pacific; Late Pleistocene South Atlantic; Postglacial Indian Ocean; Postglacial North Atlantic; Postglacial North Pacific; Postglacial South Pacific

Silicoflagellates *see* **Paleoceanography, Biological Proxies**: Radiolarians and Silicoflagellates

Speleothems *see* **Carbonate Stable Isotopes**: Speleothems

Stable Isotope Studies, Carbonates *see* **Carbonate Stable Isotopes**: Overview; Speleothems; Terrestrial Teeth and Bones; Nonmarine Biogenic Carbonates; Terrestrial Organic Materials; Non-Lacustrine Terrestrial Studies; Lake Sediments

Stable Isotope Studies, Deep Sea Records *see* **Paleoceanography, Physical and Chemical Proxies**: Oxygen-Isotope Stratigraphy of the Oceans; Oxygen Isotope Composition of Seawater; Salinity Proxies $\delta^{18}\text{O}$; **Paleoceanography, Records**: Early Pleistocene; Late Pleistocene North Atlantic; Late Pleistocene South Atlantic; Postglacial Indian Ocean; Postglacial North Atlantic; Postglacial North Pacific; Postglacial South Pacific; **Ice Core Methods**: Stable Isotopes; **Ice Core Records**: Antarctic Stable Isotopes; Greenland Stable Isotopes; Correlations Between Greenland and Antarctica

Stable Isotope Studies, Ice Cores *see* **Ice Core Methods**: Stable Isotopes; **Ice Core Records**: Antarctic Stable Isotopes; Greenland Stable Isotopes; Correlations Between Greenland and Antarctica

Sub-Milankovitch Events *see* **Paleoclimate Reconstruction**: Sub-Milankovitch (DO/Heinrich) Events

T

Talus Slopes *see* **Permafrost and Periglacial Features**: Talus Slopes

Teeth and Bones, Stable Isotope Studies *see* **Carbonate Stable Isotopes**: Terrestrial Teeth and Bones

Tephrochronology *see* **Quaternary Stratigraphy:** Tephrochronology

Thermohaline Circulation of the Oceans *see* **Glacial Climates:** Thermohaline Circulation

Thermoluminescence Dating *see* **Luminescence Dating:** Thermoluminescence

Till, Glacial *see* **Glacial Landforms, Sediments:** Till Fabric Analysis; Tills; Glacial Erratics and Till Dispersal Indicators

Tree Rings *see* Dendrochronology; Dendroclimatology; **Glacial Landforms, Tree Rings:** Dendrogeomorphology; Dendroglaciology; **Plant Macrofossil Methods and Studies:** Dendroarchaeology

Treeline Reconstruction *see* **Pollen Methods and Studies:** Archaeological Applications

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USE OF QUATERNARY PROXIES IN FORENSIC SCIENCE

Contents

Use of Geology in Forensic Science: Search to Locate Burials

Soils and Sediment

The Use of Macroscopic Plant Remains in Forensic Science

Insects

Analytical Techniques in Forensic Palynology

Use of Geology in Forensic Science: Search to Locate Burials

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Introduction

The term 'forensic' may mean 'of or relating to courts of law, crime detection, or to public debate.' Therefore, 'forensic geology' (known also as 'geoforensics') may be regarded as 'the

application of geology to the investigation and detection of crime, which may come before a court of law.' Forensic geologists may assist in some types of criminal investigations to help determine if there may be an association between items and objects, which came into contact at a crime scene.

Alternatively, forensic geologists may assist in the design and implementation of a 'search' to confirm the presence or absence of objects or items buried as part of a criminal act.

Forensic geology may support crime investigations which include murder, rape, kidnapping, fraud, firearms offences, and the search for missing persons. Both 'forensics' and 'geology' have the ability to reach out to the public and to capture their imagination, and increased media coverage, TV documentaries, and film coverage has possibly benefited forensic science and geology in recent years.

The history and early developments in forensic geology may be traced back to the middle part of the 1800s and early 1900s when the collection and analysis of soils was used to determine whether there was an association between an offender, item, and/or victim of crime. This was based on the principle developed by the French medicolegal pioneer Edmund Locard, who suggested that when two objects come into contact there is always a trace left behind. Forensic geologists also contributed to criminal investigations throughout the remainder of the 1900s. For example, geologists have been working with the Federal Bureau of Investigation (FBI) since about 1935 (Murray and Tedrow, 1975).

Objectives

Forensic geology may be conveniently grouped into two categories. Firstly, 'geological evidence' (sometimes called 'trace evidence') and, secondly, 'search.' The objective of this article is to provide a general overview on 'search.' Geological evidence aspects of forensic geology are discussed elsewhere in this encyclopedia.

Forensic geology has a variety of applications, such as in policing, law enforcement, counter terrorism, domestic incidents, humanitarian, geotechnical, civil engineering, mining, minerals and material engineering, environmental pollution and contamination, hydrogeology, and fraud investigations. However, the scope of this article is limited to 'searching the ground' in a policing and law enforcement context.

Some of the information in this article is from other publications by Donnelly, Harrison, and others. Some case details and images have also been purposely changed and made anonymous to enable their publication.

Recent Developments in Forensic Geology

Since 2002, forensic geology has experienced somewhat of a global renaissance. There has been increased awareness and interest in forensic geology by the police, law enforcement, military, consultancy, academia, and the media. There have been several international meetings and conferences, and numerous text books, scientific papers, conference proceedings, and magazine/journal articles have been written (see, e.g., Bergslien, 2012; Donnelly, 2008a,b, 2009a,b, 2010, 2011; Donnelly and Harrison, 2010a,b, 2012; Harrison and Donnelly, 2008, 2009; Murray, 2004, 2005, 2011; Murray and Solebello, 2005; Murray and Tedrow, 1992; Pye and Croft, 2004; Ruffell and McKinley, 2008; Ritz et al., 2009).

A meeting was held in 2002 as part of the All-Party Parliamentary Group for Earth Science, at Westminster Palace, House of Commons, London (Donnelly, 2002a,b, 2003). This event stimulated professional and academic interests in forensic geology within the United Kingdom. Five forensic geology meetings were subsequently held at the Geological Society of London (2003, 2004, 2006, 2008, and 2010) and three at the Geological Society of America in Philadelphia (2006), Denver (2007), and Houston (2008). International Soil Forensic meetings were held in Australia (2006), Edinburgh (2007), and California (2010). The first Ibero-Americano course on forensic geology was held in Colombia (2009), with the next possibly taking place in Brazil (in 2013). Interpol held its 16th International Symposium in Lyon, France, in 2010, which was attended by forensic geologists. The GeoNZ 2010 conference in New Zealand held a session on Forensic and Evidential Geology.

In 2006, The Geological Society of London, Forensic Geoscience Group (FGG) became established to help promote and develop forensic geology. This was followed by the International Union of Geological Sciences (IUGS), Commission on Geosciences for Environmental Management (GEM) forensic geology workshops held in Uruguay (2009) and Namibia (2010). Following the success of IUGS-GEM, the IUGS Initiative on Forensic Geology (IFG) was launched during the 62nd Executive Committee Meeting of IUGS, which was held at UNESCO headquarters in Paris in 2011. The first meeting was held in Rome later in 2011 and training on search was delivered to the Australian Federal Police in Australia in 2012.

Historical Overview of Search

Traditional police methods of finding graves often involved large-scale line searches and trail-and-error excavations. These were inefficient and labor intensive, and potentially destroyed evidence or subtle ground disturbances associated with burials. These were often supported by nonspecialists such as public volunteers and local interest groups. The value of large numbers of public volunteers, or military infantry forming large lines and walking through large areas of open land or woodland, must be carefully considered.

Law enforcement officers may often be inundated with offers of assistance from a variety of specialists, each with a range of search techniques and instruments. This, combined with the absence of any recognized acceptable national or international standard, can often make search deployment decision making difficult for law enforcement officers engaged in homicide searches.

The primary assets of search volunteers are their physical mobility and ability to move over the terrain in order to use their eyes to observe the ground. As with any such groups, there is no set standard of competence as to each individual's mobility, eye sight, or knowledge to identify and interpret ground disturbances such as, for example, anomalous vegetative growth, natural ground movements, and/or animal activity. It could be concluded that such searches offer little or no value to the investigator. These types of searches may also inhibit the investigation through cross contamination of the search area, or destruction of evidence such as offender foot prints and signs of past digging, but such searches may provide the

opportunity for the offender to alibi the presence of their DNA at the crime scene by his/her participation as a 'volunteer' in the search. Despite these limitations in homicide searches, the use of large untrained groups may be the most prevalent method used. This is often driven by large numbers of the public self-deploying to the search area following awareness of the investigation or appeals made through the news media. Law enforcement personnel are then faced with highly motivated, but untrained people, wishing to assist. Images shown on broadcast media then encourage further groups to deploy and offer their services and the investigators can soon become overwhelmed. The management of this, with its associated logistical support required for large groups of searchers, becomes a full-time role. The aim can rapidly become finding more and larger areas to search in order to keep people active and motivated and demonstrating to the victim's family and local community that a large effort is being undertaken to locate the missing person. Little or no assessment is made of the geology and environment prior to the search. Critically, these types of searches can also jeopardize the investigation, as when such large areas have been 'searched' by these large untrained groups for typically 1 or 2 weeks, the investigators may be given a potentially false assurance that no evidence or burial exists. Clearly, these circumstances present challenges for law enforcement, as it would be unwise to reject any offers of assistance in an active homicide investigation with a high media and local community profile. What is important, though, is the recognition that the search actions of large untrained groups may limit or inhibit the investigation. Therefore, careful consideration should be given before allowing such groups to gain access to an area where the victim's remains are likely to be found.

Local public volunteer groups and the victim's family members may offer their services during a large-scale search. These are well-motivated individuals and some may be appropriately equipped to perform a search in their local environment. However, their training may not have been accredited by a national body and they may not necessarily be familiar with techniques to locate persons that go missing in nonsuspicious or noncriminal circumstances. They are unlikely to have any enhanced detecting technologies or equipment other than their own canine pets they have trained.

The use of specialist mountain rescue or military personnel is preferable to untrained, nonspecialist, public volunteers, or even law enforcement officers who have no formal search training. Military and mountain rescue searchers may at least provide consistent searches that conform to their own organizational standards, which can be measured and assessed even if they are not formally trained in any search techniques.

Drivers for Change

Irish republican terrorist activity in the 1970s and the bombing of the Grand Hotel in Brighton, UK, on 12 October 1984, which killed five people, resulted in an inquiry being set up. As a result of this inquiry, the UK Police established a training school, jointly with the UK military, to train specialist police officers in search techniques. These were derived from military and police search units that had received training and practice in places such as Northern Ireland and the Middle East. Each

police force or geographic region within the United Kingdom has officers who receive nationally delivered training, focused on counterterrorist search techniques. This includes finding terrorist resources or hidden improvised explosive devices and other burials (such as graves).

Search

Law enforcement activities may be divided into one of four primary functions: investigation, interview, search, and recovery. Of these, the 'search' function appears to be the least developed and advanced. A search is 'the application and management of systematic procedures and appropriate detection equipment to locate specific targets' (Harrison et al., 2006). It may be regarded as 'the skill of looking and the art of finding.' An 'Offensive/Detective Search' is undertaken when evidence is required after a crime has occurred. The objective of this type of search would be to restrict a suspect or offender's ability to carry out criminal activities, obtain evidence or intelligence, or locate a buried or concealed object/item. For example, an offensive search may be conducted to locate firearms. A 'Protective/Defensive Search' is preventative and proactive and is carried out to confirm the absence of a specified item, object, or person within an area or location and allow freedom of movement and access to the general public. For example, a protective search may be undertaken to ensure that there are no buried explosive devices ahead of a VIP visit to a particular venue (Harrison et al., 2006; Harrison and Donnelly, 2009). Some of the objectives of a search are summarized in Table 1.

A police/law enforcement search may include the searches of people/individuals, missing persons, mass fatalities, crime scenes, roads and highways, railways, buildings and engineered structures, vehicles, and ground and water bodies (e.g., sea, lake, ponds, reservoirs, canals, streams, rivers) (Ruffell, 2007).

Table 1 Some principal objectives of search

<i>Search objective</i>	<i>Details</i>
Locate buried or concealed objects/items	Objects buried in the ground may include unmarked homicide or mass graves, explosives, money, firearms, drugs, jewelry, etc. (known as 'the target')
Locate missing persons	Law enforcement searches are conducted for voluntarily and suspicious missing persons
Deprive criminals of resources and opportunities	Items/objects required by criminals (e.g., firearms) may be searched for, to reduce the ability of the criminals to commit offences
Protect vulnerable targets	A search may be conducted to protect and safeguard a person, group of people, or the event itself
Intelligence gathering	A search may obtain information of use in crime prevention
Evidence gathering	A search may obtain information of use in supporting a prosecution

Ground Searches

The responsibility of a ground search may rest with a Senior Investigative Officer (known as an SIO in the United Kingdom). Usually, the search may be designed, managed, and supervised by a Police Search Advisor (POLSA in the United Kingdom) and a trained Police Search Team (PST). The SIO/POLSA may decide what other specialists and resources may be required to help locate the target buried in the ground.

Geologists may explore and investigate the ground to locate mineral resources, investigate geological hazards, or determine the ground characteristics ahead of civil engineering construction. Many of these exploratory and investigative methods, techniques, and instruments used are applicable to law enforcement ground searches. Geological exploration and ground investigations usually take place in a phased approach, starting with a desk study involving the collation, review, and analysis of preexisting data and information. This may be followed by a reconnaissance site visit (known also as a walk-over survey) to make observations on the geology, geomorphology, and ground conditions, and to determine any logistical, technical, and environmental constraints that may be relevant for any future detailed ground investigations and/or monitoring. This philosophy has been successfully applied to law enforcement searches in recent years (Donnelly, 2010; Donnelly and Harrison, 2010a).

Geological advice on search can be variable, depending on the specific nature of each operational case. This advice may be limited to the provision of information over the telephone, or the design, implementation, and management of a search. The duration of a search may also vary, lasting from hours to days and, in high profile and rarer cases, weeks to months (rarely years). During the design stages of a search, it will be desirable to strike a pragmatic balance between cost and resources. This, however, will be influenced by the nature of the target being sought and the profile of the individual case.

Searches for burials may be grouped into one of five categories, as summarized in Table 2 (Harrison and Donnelly, 2009).

Before the implementation of a ground search for a homicide grave, it will be helpful to gain a basic understanding of the behavioral characteristics associated with the disposal of human remains in an unmarked grave. Often, this may be provided from a clinical psychologist, environmental profiler, or highly trained police search advisor. These are summarized in Table 3.

Conceptual Geological Model

Burials in the ground usually take place in soils or weak or weathered rock. Therefore, it is necessary to understand the geology of ground at and in the immediate vicinity of the burial place. One of the most valuable roles a forensic geologist can provide the police/law enforcement officer is a geological model of the search area and target. For example, in the past a law enforcement officer may have chosen a particular type of instrument because it worked successfully in another part of the country where the geology and ground conditions may

Table 2 Search philosophy

Standard operating procedure (SOP)	A written document is applied to a search to provide a search strategy, instrument types and search assets deployed, assurance of consistency throughout the search, opportunities for critical and peer review, and an official record of the search
Intelligence-led	Searches should not be undertaken speculatively but must be intelligence-led and based on hypotheses. Geological, geochemical, or geophysical techniques must be based on sound scientific justification
Feature-focused	Features of the landscape/ground may be consciously or subconsciously used by offenders to bury objects in the ground (known sometimes as the Winthrop technique). The location may be familiar to the offender and this could facilitate the choice of the grave/excavation location. If the offender's mode of operation can be determined by the investigators, this may significantly reduce the search area
Scenario-based searching	Searches are conducted using intelligence and information from behavioral profiling and/or an assessment of the victims (known also as a victimology assessment). This often may require contributions from other forensic specialists such as a clinical psychologist or an environmental profiler. A hypothesis will be generated by profiling the offender and/or victim to identify possible locations where a burial may have taken place. Comparison with data for other similar crimes may be used to facilitate this process along with factors such as likely distance traveled and suitability of locations for burial. This method provides a more objective approach and removes the subjective 'instinctive,' 'gut-feeling,' or 'hunch' used as a search basis in some historical searches. A range of scenarios may be generated, which can then be considered and prioritized in a proportionate manner
Search and Rescue	Search and Rescue is applied to lost or injured, missing persons, or groups of people who wish to be found (i.e., they are actively trying to be recovered or self-discovered). Estimations are made to identify a search area to provide a search radius from the last known place the person was seen, or where evidence was found to verify that person had been. Since this is based on assumptions and subjective calculations, this may introduce a degree of vulnerability into this particular approach. A hypothesis may be generated based on a correlation between the person's mobility, distance, and time. Where significant time has passed, the search areas may be large and subdivided into search zones, including several search teams, but coordinated by a search manager

Table 3 Attributes of victim homicide disposal in a shallow grave

Attribute	Characteristics
Diggability and principal of least effort	Minimizing the time spent digging and reinstating a grave may be significant for the offender. The offender may choose a location that offers the opportunity to quickly dig an excavation of length, width, and depth for the victim to be concealed. For example, softer soils such as peat and sand may offer better alternatives than clay- and gravel-rich soils
Bulking, excess soil, and scarring	Digging shallow graves may result in an excess of soil caused by bulking (the expansion of the soil following its <i>in situ</i> removal) and the volume taken up by the deceased. Digging and excess soil may produce scarring of the ground but this may become less apparent with time due to weathering and erosion, or vegetation growth
Concealment	The location of the grave may have a low witness potential and is likely to be out of view from the public or passers-by to enable the covert burial to take place
Familiar location	'We go where we know' and therefore the location of a grave may be familiar to the offender. If the offender is seen or compromised, he may have an explanation for his presence at that particular location. Familiar locations will also facilitate navigation (particularly if at night) and escape routes. This will usually be aided by the recognition of physical features of the landscape

have been different, or because it happened to be available at the time the search was being conducted.

Mineral exploration geologists, engineering geologists, and geomorphologists are trained to 'read the ground.' They usually have the luxury of some 'time' to develop a more detailed understanding of the ground conditions ahead of the development of a mine, the building of an engineered structure, or the prediction of risks associated with a geological hazard. This may be facilitated by the acquisition of geological maps, memoirs, reports, and papers, supplemented by, for example, geological mapping, analysis of aerial photographs and satellite images, drilling, trenching, and geophysical and geochemical surveys.

Law enforcement investigations, particularly those involving the search for missing or vulnerable persons, or unmarked homicide graves, may be 'time critical' and therefore the forensic geologist may be required to quickly develop a conceptual geological model (CGM) for the target. This will often be based on the intelligence provided by the police/law enforcement officers on the possible burial site(s).

A CGM provides an estimate of what is likely to be found and the possible conditions of the target (Figure 1). This is a 'model', and as such it must be challenged and tested as more information becomes available. This may be gained from the progressive acquisition of additional geological data/observations from the reconnaissance (walk-over) survey and the main search itself, and then the model can be updated.

As a result of the variability and complexity of the ground, the geological model is unique to each particular crime scene or search area. The more detailed and reliable the original geological information and police intelligence, the more accurate will be the CGM for the search area. Those based on high-quality intelligence and good-quality published geological maps will be the most reliable. However, even where high-quality geological maps have been published, there is still the need to conduct detailed site reconnaissance walk-over surveys to provide the necessary geological observations that will be specific to the individual search area. For instance, published geological maps and memoirs may not necessarily provide information on the detailed stratigraphy, principal lithological (rock) types, groundwater seeps and springs, and areas of ground disturbances (e.g., scarring, scarps, fissures, compression, debris accumulations, changes in slope profile, topography, geomorphology, and landslides and mass wasting, including falls, topples, slides, and slips).

The development of a geological model will be based on the training and experiences of the individual geologist, case-specific intelligence, and the time and resources available. The accuracy of the model will improve as more information is made available and as the search advances. The model will be verified only when the target has been located and recovered.

The strength of a CGM in a police/law enforcement search for a burial is in the provision of the following:

- Determination of the 'detectables' (i.e., what can potentially be found). For example, although human remains may be the desired target, it may be relevant to design a search to locate the 'indicator' or 'associated' target objects such as any clothing, coinage, or jewelry. (This is the adaptation of a typical strategy deployed in mineral exploration, for instance, where some metallic ore minerals, diamonds, or other gemstones are being explored for; the exploration strategy may be designed to locate the minerals associated with the desired mineral.)
- Estimation of the age, geometry, size, depth, and expected duration of preservation.
- Information on the geology in the immediate vicinity of the burial: for example, the physical, chemical, and geotechnical characteristics of the soils, rocks, and 'made ground,' depth to bedrock, degree and intensity of weathering, groundwater, past mining and mineralization (metaliferous minerals in the ground may negate the deployment of certain geophysical search assets).
- Knowledge of the number of interactive, dynamic, geological, and geomorphological processes that may have taken place before, during, and since burial took place. These may have influenced the target and associated detectables. These natural processes are variable and occur over a wide variety of time frames, from days to millennia. They include, for instance, wind, stream and river erosion, landslide, mass wasting, and weathering. The forensic geologists will determine which (if any) of these processes may be significant in affecting the grave's location. The processes may have also been influenced by the activities of the offender, such as digging and soil reinstatement. For example, some peat deposits in parts of the United Kingdom may promote the

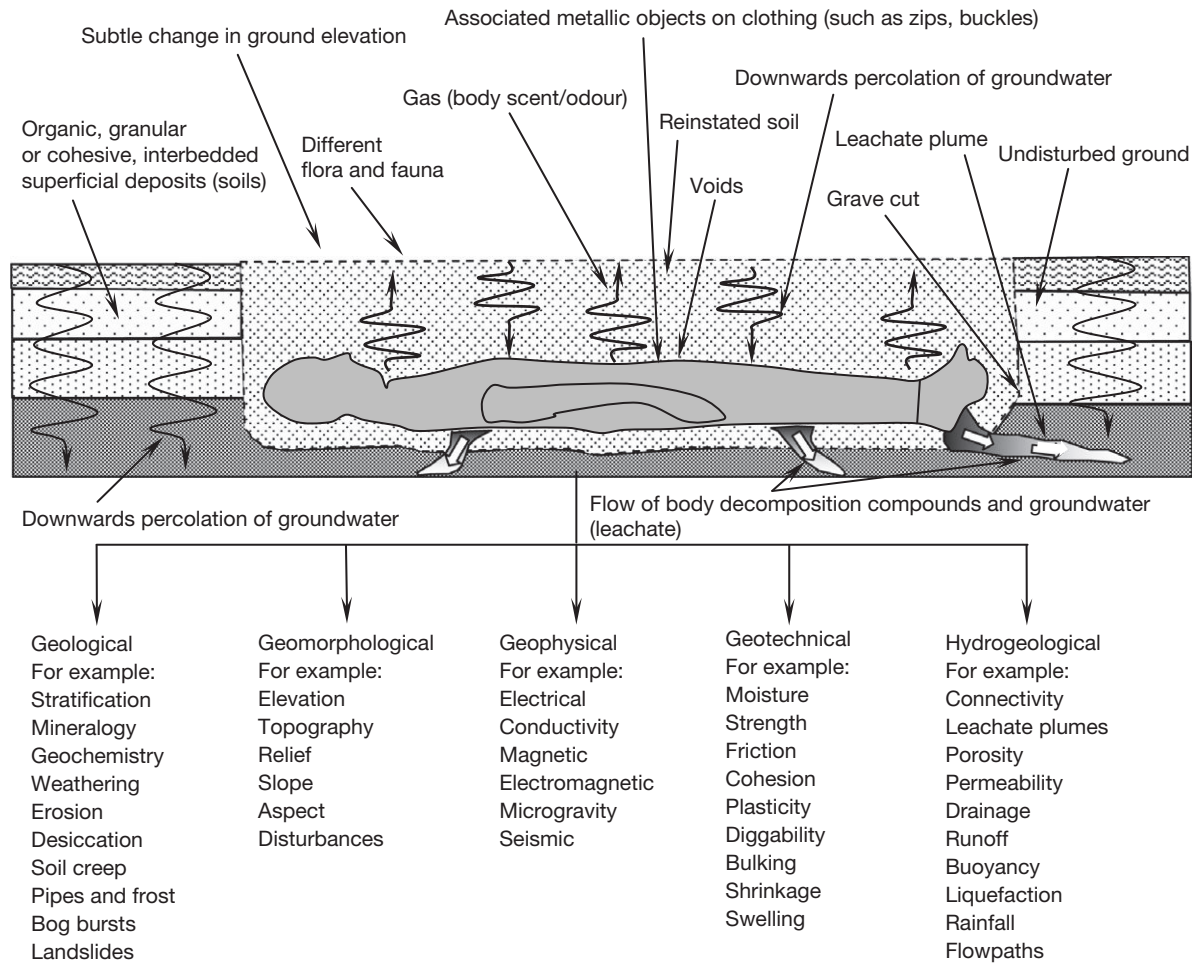


Figure 1 Conceptual geological model (CGM) for a shallow homicide grave (after Donnelly, 2002a, 2008a, in Donnelly, 2010; Harrison and Donnelly, 2008, 2009).

preservation of human tissue, whereas others, such as granular (sand) soils, may promote decomposition. Determination of the hydrogeology and the location of hydrogeological domains (cells) and groundwater flow paths may influence the generation of leachate and any volatile organic compounds (VOCs; known also as gas/scent/odor), which may be relevant for canine searches (Figure 2).

- Predetermination of the most suitable search assets to be used that would provide the greatest opportunity to locate the desired target/burial.
- Identification of the optimum search strategy and methodology for the deployment of search assets and instrumentation.

Therefore, the development of a CGM provides an estimation of the ground conditions and facilitates the optimum choice of search assets most likely to locate the target.

Search Planning and Phases

Ground searches may be divided into three phases:

Presearch Phase

The presearch phase will involve the initial briefing between the forensic geologist and police/law enforcement officers. They will discuss the type/nature of the target desired to be found and determine precisely what expertise the geologist may be able to offer the search team. Terms of reference, deliverables, and contractual agreements may be discussed at this stage. Police intelligence and geological data will be collated, reviewed, and analyzed. This will form the precursory stages of a reconnaissance walk-over survey of the crime scene/search area. The forensic geologist will develop a CGM at this stage, including soil types and thicknesses, principal rock types, hydrogeology, and geomorphological processes active before, during, and after the burial took place. Operational and logistical constraints may also be observed during the initial walk-over survey. For example, the presence of utilities, overhead power lines, or metal fences may inhibit or exclude the deployment of certain types of geophysical techniques. It is also at this stage of the search that operational matters will be considered, such as the management of the public, media, psychics, and family and friends of the victim. Any ground disturbances observed, such as subsidence, settlement, and scarring, can be

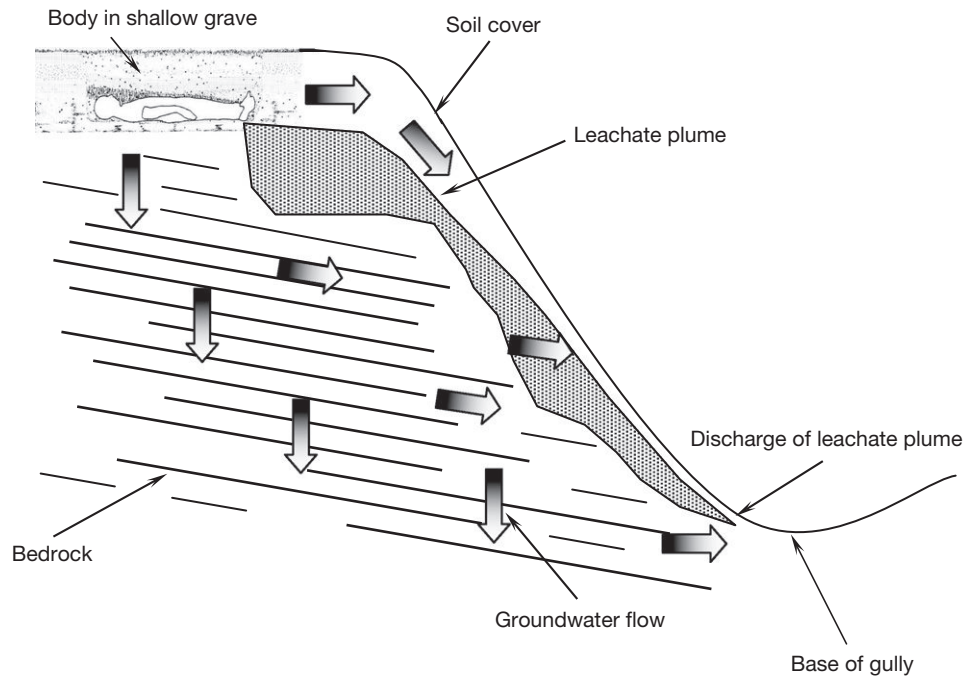


Figure 2 Schematic model to show the generation of a leachate plume and gas/scent/odor and how this may be potentially controlled by the geology and increase the search target area (after Donnelly, 2002a, in Donnelly, 2010; Harrison and Donnelly, 2008, 2009).

assessed at this stage to determine whether these were generated by geomorphological processes, animal burrowing activity, the activities of man (e.g., mining, farming), or potentially associated with the action of the offender (Figure 3).

The determination of ground diggability and production of color-coded RAG (red–amber–green) maps may also be developed during the initial reconnaissance survey to prioritize the search area. Geological mapping and the determination of soil types and depths by probing or the digging of exploratory trial pits may be required at this stage to determine the offender's relative ease or difficulty in digging a shallow excavation for disposal of any human remains or associated objects. In a police/military search, areas shown as 'red' are usually of the highest priority (Donnelly and Harrison, 2010a). The CGM and diggability map will predetermine the most appropriate types of search assets required to deliver a high-assurance search.

The search area should be identified during the presearch stage and, if required, protected by a police cordon. The limits of the search area may vary considerably, from the confines of a small garden at the back of a house to an entire mountain side. Other typical search areas include, for example, parks and other public open spaces in towns and cities, coastlines, forest/woodland areas, caves, abandoned mines and quarries, landfill and other areas of waste disposal/tipping, agricultural land, railway corridors, tracks, and roadside verges.

Presearch reconnaissance visits to the search area/crime scene may also benefit from the presence of the investigative officers or other specialists, such as a forensic botanist, palynologist, or environmental profiler. This may allow hypothesis generation from the recognition of any primary, secondary, or



Figure 3 Settlement of the ground and changes in soil color that may potentially be observed during the initial reconnaissance walk-over inspection (Donnelly and Harrison, 2010b, photograph by Harrison).

tertiary 'Winthrop Markers,' which may have potentially formed part of the offender's mode of operation.

The objective and deliverable of the presearch phase will be the production of a standard operational procedure (SOP) to provide assurance and consistency throughout the search.

Search Phase

The aim of the main search phase is to implement the SOP to provide a high-level assurance search to verify the presence of, or confirm the absence of, the desired target. The search may begin with a briefing to the police search team and associated support personnel. This will provide information on the aims and objectives of the search, nature of the target objects to be found, search asset, strategy, methodology, agreed methods for the recovery, identification and recording of any finds, and the search management and reporting structure. Any operational, logistical, and health and safety matters will also be provided at this stage. All search personnel must be already familiarized with the use of the search assets and instruments, and be appropriately qualified, trained, and experienced (Figure 4).

The forensic geologists will be responsible to assist with the design and implementation of the search to locate the target. This will also require effective communications between the forensic geologist, police/law enforcement officers, and other specialists (Donnelly, 2008a). If possible, searches must initially be noninvasive to minimize any cross contamination and ground disturbances. However, a degree of digging and trial pit excavation or probing will inevitably be required to verify the presence of a target or to investigate any positive anomaly provided by the geophysical instruments or specialist canine. All surface and buried finds will require formal identification, recording, and recovery usually by nongeological specialists. This may include, for example, a forensic anthropologist, archaeologist, crime scene investigator, and/or exhibits officer.

Postsearch Phase

Upon completion of a search, the search area and any finds must be mapped and accurately recorded. The nature of this documentation will be variable and case-specific, but may include photographic records, a descriptive search account, any deviations from the SOP, conventional geological/geomorphological mapping, the geology in the vicinity of any finds, an account of the soil types and depths, the degree and intensity of weathering, groundwater presence, and dimensions and conditions of any finds. The presence of soils and rock fragments from another unknown location associated with the body or target object may require sampling and analysis by a forensic geologist suitably trained and experienced in geological (trace) evidence (Lee et al., 2009).

Surveying, including basic geological and topographic mapping, the deployment of precise leveling, global positioning systems (GPS), total station theodolite, or laser scanning techniques, may be required, and in some forces specialist forensic and surveying units may be deployed to fill this role.

A search debriefing forms an important part of the overall search procedure. This provides the opportunity for each search team member and the investigative team to offer views and opinions on the search strategy, methodology, and assets used. Where possible, any feedback provided and 'lessons learned' should be recorded and if necessary fed into future similar searches and the SOP.

All searches must have a predetermined 'exit strategy' to carefully manage the outcome of the search.

Search Assets

Search assets are techniques, instruments, and methods that may be used by a forensic geologist to help search the ground to locate the desired target. In general, ground searches proceed



Figure 4 Defining the search area and setting up search lanes during the initial stages of a search (Donnelly and Harrison, 2010b, photograph by Donnelly).

from the macrolevel to the microlevel and from the noninvasive to the invasive. Based on this, some of the principal search assets may be grouped into the following categories:

- *Remote sensing* (also known as *Earth observation*). These methods acquire data about the ground without direct contact, obtained from the air (fixed-wing aircraft, helicopter, or a drone) or satellites. Advances in remote sensing technology and improved resolution mean that some of these techniques can be increasingly applied for the investigation of individual crime scenes and search areas. The acquisition and analysis of pre- and postburial conventional aerial photography (vertical and oblique), when used in association with geological mapping, offers a particularly useful technique to help locate graves or to assist with the planning and management of a search strategy. Other remote sensing methods include, for example, infrared photography, photogrammetry, elevation modeling, hyperspectral and multispectral imaging, laser scanning, long-distance light detection and ranging (LiDAR), synthetic aperture radar (SAR) and interferometry, and X-ray imagery. Large or complex search data may be managed and analyzed using geographic information systems (GIS) and geostatistics. Examples of successful applications of remote sensing in forensic geology can be found in [Ruffell and McKinley \(2008\)](#).
- *Geological and geomorphological mapping*. As noted earlier, geological mapping forms an important part of the initial site reconnaissance walk-over survey. Detailed geological maps for crime scenes and search areas may potentially identify higher priority search areas, provide an estimate of diggability, and identify subtle ground disturbances associated with digging and burials ([Donnelly and Harrison, 2010a](#)).
- *Geological analysis of intelligence*. Interview statements, crime scene photographs, body disposal plans, letters from offenders, and other types of intelligence may sometimes be of benefit when reviewed and analyzed from a geological perspective. This could provide information to assist with a search. This may be particularly valuable if geological landmarks or rock exposures occur in the background of photographs, or geomorphological features (e.g., waterfalls, steep slopes, streams, sandstone banks, shale gullies) are mentioned in documents. A good example of this followed the terrorist attack on the World Trade Centre, in New York, on September 11, 2001, when a videotape showed Osama Bin Laden and limestone rocks which cropped out in the background. These were recognized by a geologist who had field experience in Afghanistan where these rocks occurred in a specific region ([Pinsker, 2002](#)).
- *Probing* (known also as *auguring* or *window sampling*). This may sometimes be regarded as an 'old fashioned' search technique. For example, this method was used in the 1960s during a search of Saddleworth Moor, UK, for missing children (later to become infamously known as 'The Moors Murders'), whereby lines of public volunteers would 'line search' and stick a wooden probe into the ground to smell for any decomposition possibly associated with decomposing human remains. However, soil probes have advanced significantly, and there are now a variety of soil probes designed to suit organic (peat), granular (sand), or cohesive (clay) soil or weathered rock conditions, or combinations thereof. A line of highly trained and well-disciplined police officers, probing on a grid of at least 0.2 m, can provide an effective, although time-consuming, method to search for buried objects or recover samples of the target. Probing techniques rely on the ability to detect resistance caused by the target. Probing may also be undertaken to facilitate canine searches, if the probe holes are allowed to vent for a short period ahead of the deployment of cadaver dogs (this is best undertaken in favorable barometric pressure conditions). Before probing takes place, control measures should be established so the search teams can practice this technique on buried items comparable to the target ([Figure 5; Ruffell, 2004](#)).
- *Trenching*. Trenching of the ground takes place by the use of handheld digging implements including spades, picks, and mattocks. Exploratory 'trial pits' or 'inspection pits' may be dug to act on intelligence considered to be reliable or to verify a geophysical anomaly or positive canine indication.
- *Geophysics*. A variety of geophysical surveying techniques are available that may potentially be used to assist with ground searches. However, each method should be deployed only following the careful investigation and characterization of the geology at and in the immediate vicinity of the crime scene/search area. An advantage of geophysical surveys is their ability to be deployed noninvasively (without contact with the ground). This reduces the possibility of any evidential contamination and damage to crime scenes. Common geophysical techniques that may potentially be used to search the ground include ground penetrating radar (GPR), conventional metal detectors, as well as electromagnetic (conductivity), resistivity, and magnetic probes. Less common methods are seismic and microgravity. When required, some geophysical surveys may be deployed remotely on board a fixed-wing aircraft or helicopter. Alternatively, they may be towed behind a 4×4 vehicle or tractor when large tracts of relatively flat land are required to be investigated. The results of geophysical investigations are measurements showing vertical and lateral variation of the physical properties of the ground, often displayed in plan or as a series of cross and longitudinal sections. These must be analyzed and interpreted in association with a detailed understanding of the search area's specific geology and intelligence ([Figure 6](#)). Positive geophysical anomalies will require subsequent investigation by excavation ([Figure 7; Fenning and Donnelly, 2004; Ruffell, 2005](#)).
- *Hydrochemistry and geochemistry*. Objects buried in the ground may be subjected to geological processes of weathering and erosion. For example, the decomposition of human remains in unmarked graves may be influenced by groundwater flows. This may lead to the generation of a 'leachate' plume, which could flow from the grave through the soils and rocks. The generation, geometry, and distance traveled by this plume will be dependent on several inter-related factors including the porosity and permeability of the soils and rocks, geological structure, and local weather conditions. In these circumstances, VOCs (known also as



Figure 5 Different line probing techniques: one method using a line of trained and disciplined police officers and a forensic geologist (left), the other method requiring the deployment of two police officers to search a small area of ground in Eastern Europe (after Donnelly, 2008b; Donnelly and Harrison, 2010b, photographs by Donnelly).



Figure 6 A noninvasive geophysical survey being conducted to search part of a coastline for a missing person (Donnelly and Harrison, 2010b, photograph by Harrison).

gas/scent/odor) may also be generated. The collection and analysis of soil and/or groundwater samples may provide the opportunity for leachate and VOC detection (Figure 8). The isotopic analysis of teeth and bone, using strontium-87 and carbon-13, may also have some application in search or help determine geographical provenance of the deceased. By comparison with better established geophysical methods, geochemistry and hydrochemistry appear to be less well developed in terms of their applications to search,

but nevertheless these may be worthy of consideration in certain circumstances.

- *Cadaver dogs (victim recovery dogs)*. Although the use of highly trained specialist police dogs is not conventionally a geological search technique, their deployment with a highly trained and experienced dog handler and geological search advisor may provide an important search option. The positive reaction of a dog may be influenced by the geological and environmental conditions, and the better



Figure 7 Police search teams digging small excavations to verify positive anomalies identified by geophysical surveys (Donnelly and Harrison, 2010b, photographs by Donnelly).

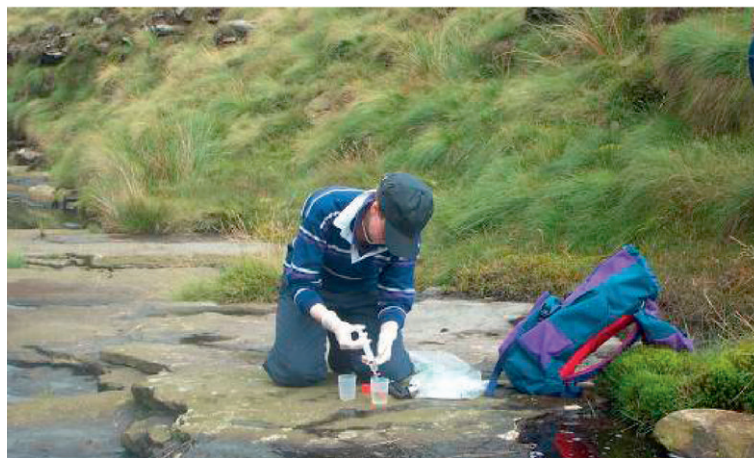


Figure 8 Sampling of soils (upper) and groundwater (lower) for geochemical analysis to detect traces or indicators of the presence of decomposing human remains in Eastern Europe (Donnelly and Harrison, 2010b, photographs by Donnelly).

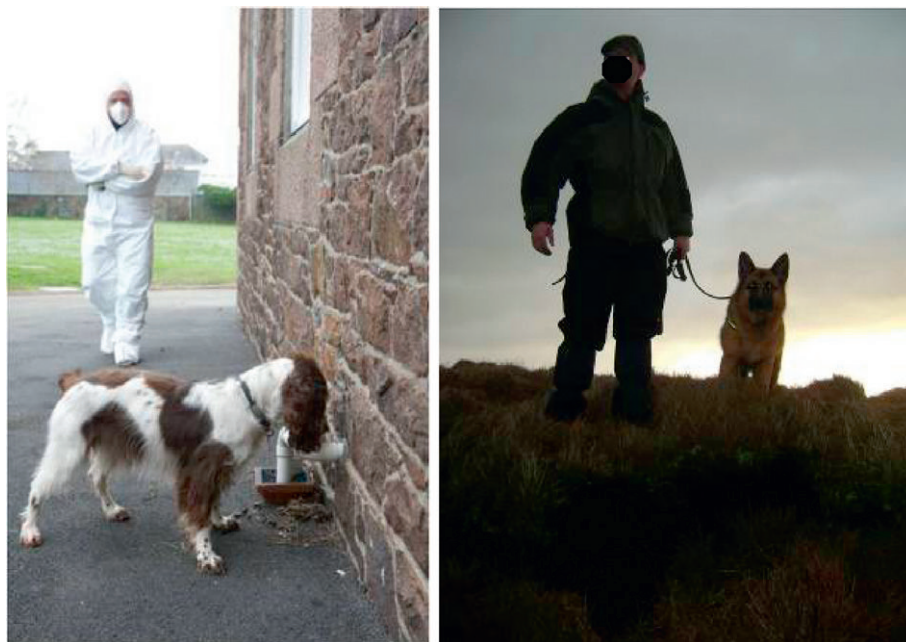


Figure 9 Victim recovery dogs search urban (left) and rural (right) locations as part of a geoforensic search strategy (Donnelly and Harrison, 2010b, photographs by Harrison and Donnelly).

the understanding of these controlling measures, the greater will be the likelihood for the dogs to be deployed to high levels of assurance (Figure 9).

See also: **Use of Quaternary Proxies in Forensic Science: Soils and Sediment.**

References

- Bergslien E (2012) *An Introduction to Forensic Geoscience*. Oxford: Wiley Blackwell.
- Donnelly LJ (2002a) Finding the silent witness. *The Geological Society of London, Geoscientist* 12(5): 16–17.
- Donnelly LJ (2002b) *How forensic geology helps solve crime*. Westminster Palace, House of Commons: All-Party Parliamentary Group for Earth Science.
- Donnelly LJ (2003) The applications of forensic geology to help the police solve crimes. *European Geologist. Journal of the European Federation of Geologists* 16: 8–12.
- Donnelly LJ (2008a) Communication in geology: A personal perspective and lessons from volcanic, mining, exploration, geotechnical, police and geoforensic investigations. In: Liverman DGE, Pereira CP, and Marker B (eds.) *Communicating Environmental Geoscience*, pp. 107–121. London: Geological Society. Special Publication 305.
- Donnelly LJ (2008b) Forensic investigations in engineering geology, mining geology, geomorphology and geohazards. In: Donnelly LJ (ed.) *Geoscientific Equipment and Techniques at Crime Scenes, 2nd FGG Meeting of the Geological Society of London, Forensic Geoscience Group*. Burlington House, London, 35 Programme and Abstracts. 17 December 2008.
- Donnelly LJ (2009a) The geological search for a homicide grave. *The Investigator* 42–49.
- Donnelly LJ (2009b) Operation Colombia. Article on UK and USA geoforensic specialist visit to Bogota, Colombia, April 2009 *Geoscientist* 19(9): 6–7 A. DeWind (reporter).
- Donnelly LJ (2010) *The role of geoforensics in policing and law enforcement*, pp. 19–22. Manchester: Emergency Global Barclay Media Limited.
- Donnelly LJ (2011) The renaissance in forensic geology. *Teaching earth sciences. Magazine of the Earth Science Teachers' Association (ESTA)* 36: 46–52.
- Donnelly LJ and Harrison M (2010a) Geomorphological and geoforensic interpretation of maps, aerial imagery, conditions of diggability and the colour coded RAG prioritisation system in searches for criminal burials. *3rd International Workshop on Criminal & Environmental Soil Forensics*. 2–4 November 2010, Long Beach, CA (pending).
- Donnelly LJ and Harrison M (2010b) *Development of geoforensic strategy & methodology to search the ground for an unmarked burial or concealed object*, pp. 30–35. Manchester: Emergency Global Barclay Media Limited.
- Donnelly LJ and Harrison MH (2012) Geomorphological and geoforensic interpretation of maps, aerial imagery, conditions of diggability and the colour coded RAG prioritisation system in searches for criminal burials. In: Pirrie D, Ruffell A, and Dawson A. (eds.) *Environmental and Criminal Geoforensics*. Geological Society of London Special Publication (in press).
- Fenning PJ and Donnelly LJ (2004) Geophysical techniques for forensic investigations. In: Pye K and Croft D (eds.) *Forensic Geoscience: Principles, Techniques and Applications*, pp. 11–20. London: Geological Society. Special Publications 232.
- Harrison M and Donnelly LJ (2008) Buried homicide victims: Applied geoforensics in search to locate strategies. *The Journal of Homicide and Major Incident Investigations* 4(2): 71–86. Produced on behalf of the Association of Chief Police Officers (ACPO) Homicide Working Group, by the National Policing Improvement Agency (NPIA).
- Harrison M and Donnelly LJ (2009) Locating concealed homicide victims; developing the role of geoforensics. In: Ritz K, Dawson L, and Miller D (eds.) *Criminal and Environmental Soil Forensics*, pp. 197–219. Berlin: Springer.
- Harrison M, Hedges C, and Sims C (eds.) (2006) *Practice Advice on Search Management and Procedures*. UK: National Police Improvement Agency (NPIA).
- Lee S, Smith K, and Oliphant S (eds.) (2009) *Body Recovery Briefing Paper. Policing Issues, Scene and Forensic Specialist Considerations*. Bedfordshire: National Policing Improvement Agency.
- Murray RC (2004) *Evidence from the earth*. Missoula, Montana: Mountain Press Publishing Co.
- Murray RC (2005) Collecting crime evidence from the earth. *Geotimes*.
- Murray RC (2011) *Evidence from the Earth: Forensic Geology and Criminal Investigation*, 2nd edn. Missoula, MT: Mountain Press Publishing Co.
- Murray RC and Solebello LP (2005) Forensic examination of soil. In: Saferstein R (ed.) *Forensic Science Handbook*, vol. 1, 2nd edn. New Jersey: Prentice Hall.
- Murray RC and Tedrow JCF (1975) *Forensic Geology: Earth Sciences and Criminal Investigation*. New Brunswick, NJ: Rutgers University Press.
- Murray RC and Tedrow JCF (1992) *Forensic Geology*. Englewood Cliffs: Prentice Hall.
- Pinsker LM (2002) Geology adventures in Afghanistan. *Geotimes*. Web Feature http://www.agiweb.org/geotimes/fe02/Feature_Shroderside.html (13 March 2003).

- Pye K and Croft D (eds.) (2004) *Forensic Geoscience: Principles, Techniques and Applications*, pp. 11–20. London: Geological Society. Special Publications 232.
- Ritz K, Dawson L, and Miller D (eds.) (2009) *Criminal and Environmental Soil Forensics*, pp. 197–219. Berlin: Springer.
- Ruffell A (2004) Burial location using cheap and reliable quantitative probe measurements. Diversity in forensic anthropology. *Forensic Science International* 151: 207–211.
- Ruffell A (2005) Searching for the I.R.A. Disappeared: Ground-penetrating radar investigation of a churchyard burial site, Northern Ireland. *Journal of Forensic Sciences* 50: 414–424.
- Ruffell A (2007) Freshwater under water scene of crime reconstruction using freshwater ground-penetrating radar (GPR) in the search for an amputated leg and jet-ski, Northern Ireland. *Science and Justice* 46(3): 133–145.
- Ruffell AR and McKinley J (2008) *Geoforensics*. Chichester: Wiley-Blackwell.

Soils and Sediment

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History

As with so many other kinds of physical evidence, forensic examination of soils and sediment, forensic geology, began with the writings of Sir Arthur Conan Doyle, who wrote the Sherlock Holmes series between 1887 and 1927. He was a physician who apparently had two motives: writing saleable literature and using his scientific expertise to encourage the use of scientific methods and scientific thinking as evidence. Holmes could tell where someone had walked around London from the splotches of dirt on their pants.

Gross (1893), an Austrian forensic scientist, wrote the book *Handbook for Examining Magistrates*, in which he suggested that: “perhaps the dirt on someone’s shoes could tell more about where a person had last been than toilsome inquiries.” It was only a matter of time before these ideas from an author of fiction and criminalists’ handbook would appear in a courtroom. In 1908, German chemist Georg Popp undertook the study of soil evidence in the homicide case of Margarethe Filbert. In this case, Popp was able to demonstrate that soil samples from two locations associated with the crime had similar characteristics to those collected from the suspects’ shoes. In addition to using soil from the suspects’ shoes, he was able to show a sequence of events in soil accumulation that was consistent with the theory of the crime; he found no soil evidence that supported the suspect’s alibi.

A century later, the use of geologic materials in criminal and civil cases is commonplace. Many state and private laboratories around the world have trace evidence examiners qualified to examine soil, glass, and related material. See Murray (2004), Pye (2007), and Ruffell and McKinley (2008) for more detail.

Soil and Related Material as Evidence

The value of any type of physical evidence in large part depends on how many different kinds of a given material can be characterized, how many kinds exist on Earth, and how they are distributed. The value of soil evidence rests on the fact that there are an almost unlimited number of rock types, mineral types, and fossils, as well as durable man-made materials such as glass and brick and concrete. In addition, it is not uncommon for soil material to have incorporated man-made particles such as plastics and metals, and chemicals such as fertilizers and pesticides. Each of these constituents can be identified in soils, and are useful to pin down the source of the sample.

Forensic geology studies vary in scope. A common type of investigation involves identifying a material that is key to a case – for example, examining pigments in an oil painting or material in a sculpture when authenticity or value is at issue. Identification is also important in questions of mining, mineral, or gem fraud, to determine if the material is what its sellers claim it to be. The presence of fire-resistant safe

insulation on a person or their property may provide probable cause for further investigation of burglary.

Beyond identification, forensic geologists can also look at the source of particular material. Here the examiner needs a broad knowledge of the geology and the best geologic and soil maps to answer questions. For example, if the soil on a body does not match the location where the body is found, from where was the body moved? Such studies are often called ‘an aid to an investigation.’ Similarly, examiners can compare two samples, one associated with the suspect and the other collected from the crime scene, to see if they had a common source: Does the soil on the suspect’s shoe have similar characteristics to the soil type collected at the crime scene (Junger, 1996)? Forensic geologists are also involved in the search for bodies and other objects buried in soil. More details on the subject are available elsewhere in the encyclopedia.

Another new developing area of forensic geology is its use in intelligence work. A person, for example, may claim to have never been to a particular location, but is then found with rocks from that spot, thus linking the individual to a geographic location. Remember the outcrop behind Osama bin Laden on TV after September 11? What was the location? A geologist who has done field work in the area would be able to locate that outcrop, and that actually happened: geologist John Shroder was able to identify the region where bin Laden had been sighted in Afghanistan in 2001. As with all class evidence, geologic evidence rarely provides a unique solution for which the legal mind cannot imagine another explanation. However, the evidence may be important in establishing guilt or innocence. There are some exceptions, as illustrated by the following three cases.

Murder and the Pond

The murder of John Bruce Dodson produced one of the most interesting cases in the entire history of forensic geology. Here, the geologic evidence is unequivocal in that it tied the suspect directly to the crime and eliminated the suspect’s alibi. Most importantly, the investigator of the crime recognized the potential importance of the geologic evidence and arranged for the examination of that evidence. The testimony of the forensic geologist was critical to the prosecution of the case. The case began on 15 October 1995, when John Dodson was found dead while on a hunting trip with his wife of 3 months, Janice. The scene was high in the Uncompahgre Mountains of western Colorado. At first glance, it appeared to be a hunting accident. However, the autopsy revealed two bullet wounds to the body and one bullet hole through John’s orange vest.

The investigation showed that the Dodsons were camped near other hunters, one of whom was a Texas law enforcement officer. He responded to Janice’s frantic call that her husband had been shot. She was standing about 200 yards from the

camp in a grassy field along a fence line. The officer determined that John was dead and started the process of getting help. Prior to calling for help, Janice had returned to her camp and removed her hunting coveralls, which were covered with mud from the knees down. She later told investigators that she had stepped into a mud bog along the fence near camp. Investigators found a 0.308-caliber shell case approximately 60 yards from the body. In addition, they found a 0.308-caliber bullet in the ground on the other side of the fence, which created a direct line from the location of the case to the body to the bullet.

Janice's ex-husband, J.C. Lee, was also camped three quarters of a mile from the Dodsons. Janice knew the site was his favorite camp location. He naturally came under suspicion. However, Lee was hunting far away from camp with his boss at the time of the shooting. Most importantly, Lee reported to investigators that while he was out hunting, someone had stolen his 0.308 rifle and a box of 0.308 cartridges from his tent. Winter comes early at 9000 ft in the Uncompahgre, and little more could be done at the scene. However, investigators Bill Booth, Dave Martinez, and Wayne Bryant returned during the summers of 1996, 1997, and 1998 and searched for the rifle and other evidence. They tried to search every place a weapon could have been hidden. They combed the entire area, including ponds, with metal detectors in hope of finding the rifle; it has never been found. During the final search of the pond near Janice's ex-husband's camp, Al Bieber of Necro-Search International commented that the mud in and around a cattle pond near Lee's camp was bentonite, a clay that someone brought to the pond to stop the water from seeping out of the bottom. That evening, Booth and Martinez were camped near the crime scene. They were discussing the evidence in the case when investigator Booth said, "The mud." He was referring to the dried mud that was found on Janice Dodson's clothing. If Janice had obtained the rifle from Lee's camp, she would most likely have stepped or fallen into the bentonite clay that drained across the road from the cattle pond (Figure 1). Remembering Janice's statement that she was returning to camp on the morning of the crime and stepped into a mud bog near her camp, Booth and Martinez decided



Figure 1 Pond lined with bentonite clay where Janice Dodson got clay on her shoes and pants when she went to steal the rifle that she used to kill her husband John.

they needed to obtain dried mud samples from the bog near the Dodsons' camp, the area around a pond nearby the camp, and the human-made pond and runoff near Lee's camp.

Booth and Martinez packaged the dried mud from each location and sent the samples along with the dried mud that had been recovered from Janice's coveralls to the laboratory section of the Colorado Bureau of Investigation in Denver, where it was examined by Jacqueline Battles, a forensic scientist and lab agent. She concluded and later testified to the fact that the dried mud found on Janice Dodson's clothing was consistent with the dried mud recovered from the pond near Lee's camp. The dried mud that had been recovered from Janice's overalls was found not to be consistent with the mud bog or the pond near her camp. This was a breaking point in the case that allowed Booth and Martinez to put Janice Dodson in her ex-husband's camp around the time his rifle had been stolen. There are no other bentonite-lined ponds in the area and no bentonite deposits.

Booth and Martinez went to Texas and served an arrest warrant on Janice. She was extradited to Colorado, tried in court, and convicted in the murder of John Bruce Dodson. The jury understood the results that followed Booth's insightful 'mud' exclamation. Janice is now serving a life sentence without the possibility of parole in Colorado's state prison for women. An appreciation of the value of soil evidence and the collection of that evidence was critical in this case.

Oil Slicks and Sands

A case that illustrates many of the issues involved in comparing soil and related material occurred in Canada a few years ago. The body of 8-year-old Gupta Rajesh was found alongside a road outside of Scarborough, Ontario. The back of his shirt had a smear of oily material, and the preliminary conclusion was that he was the victim of a hit and run accident, with the oily material coming from the undercarriage of a vehicle. However, forensic geologist William Graves of the Centre of Forensic Sciences in Toronto examined the oily material and the particles suspended in it; the results told a different story.

Investigators had collected samples of oily material on the floor of an indoor concrete parking garage where a suspect, Sarbjit Minhas, parked her automobile. Analysis of the samples showed that the sand and other particles within the oil from the victim's clothes and the parking garage were similar. Analysis of the oil from the victim's shirt and garage floor showed them both to be similar. It is also important that they were different from oil collected on the floor of ten other garages in the area.

Particles in samples from the victim's clothes and the suspect's parking place provided considerable information. The sand from both samples was sieved, and subsamples produced of the various size grades for the two samples. When compared after the oil had been removed, the color of each pair of subsamples was identical. Additionally, the heavy minerals in both samples were similar, and three distinct kinds of glass were found in the two samples: amber glass, tempered glass, and light bulb glass. Each of the different glasses was identical in refractive index value (the amount a ray of light bends when passing through the glass into another medium). Small particles of yellow paint with attached glass beads were

found in both samples. This type of paint is often found on center stripes of highways and reflects light.

Graves concluded that there was a high probability that the body of Gupta Rajesh had been in contact with the concrete floor of the garage at the place where the suspect parked her car. During that contact there had been a transfer of oil and its contained particles to the back of his shirt. Minhas was tried in the Superior Court of the Province of Ontario in November 1983 and convicted, with help from testimony by Graves.

All Soil Quarries Are Different

A classic aid to an investigation case occurred in southern Australia a few years ago. Neighbors heard a struggle taking place at the home of two women. When police arrived, the women and their car were missing. The car broke down over 100 miles away. When the car was found, the suspect had returned to the car with a mechanic. The trunk of the car contained a shovel with soil attached. There was also a lot of blood. Rob Fitzpatrick, Director of the Centre for Australian Forensic Soil Science, examined the soil and the shovel. The soil on the back of the shovel had been smeared, indicating the shovel had been used to pat down moist soil. Based on his knowledge of the local soils, Fitzpatrick determined that the soil on the shovel came from one of the local residual soil gravel quarries. The minerals on the shovel soil narrowed the search down to a small group of quarries. Samples from these quarries were collected and one produced identical minerals to the soil on the shovel. Rain had obliterated any sign of digging or footprints in that quarry. Police staked out that quarry, and within 3 days a fox arrived, started digging and unearthed the bodies. This led to the arrest and conviction of the man who came back to get the disabled car (Fitzpatrick et al., 2007).

Examination Methods

There are various instruments, methods, and procedures that are used to study minerals, rocks, soils, and related materials for forensic purposes (Murray, 2004). These methods are used to collect data from the various questioned and known samples that are then used to make the judgment of comparison or lack of comparison. All are methods that have long been shown to be reliable in obtaining information about soils. The evidential value of some of these methods is greater than others. Because of the extreme diversity of soils and related material, the examiner may decide that in a particular case some other method will provide information that will assist in reaching a conclusion. Because soils often have added particles such as fibers, hair, and/or paint chips these can be collected for examination by experts in those forensic science fields. One of the reasons microscopic examination is so important is that it is only way that particles of other forms of evidence can be found and unusual mineral particles discovered. We must remember that examination is focused on determining whether two samples have the same properties and thus have the possibility of having a common source. That determination is greatly strengthened when rare and unusual minerals and particles are found. The experienced forensic geologist can express an opinion based on experience and education. The strength of that opinion is often based on

knowledge of the rarity of particular minerals, rocks, particles, and related properties.

Color

Color is one of the most important identifying characteristics of minerals and soils. Minerals form a mosaic of grays, yellows, browns, reds, blacks, and even greens and brilliant purples. Virtually all possible colors of the visible light spectrum are represented. With most geologic materials and soils, the native minerals contribute directly to the soil color. However, the soil-forming processes generally produce coatings and staining on the grains that result in changes in color.

In the 1970s, investigators at the Home Office Forensic Science Service in Great Britain studied the use of color as an examination method and made many contributions to establishing the study of color as an important first step (Dudley, 1975). Sugita and Marumo (1996) of the National Research Institute of Police Science in Tokyo studied the use of color and provided important ideas that firmly established color as an important, useful method in screening samples. It is important first to record the color of the untreated soil sample and then remove the organic particles so that the true color and appearance of the sand grains can be studied.

In studying soil samples for forensic purposes, the sample is normally dried at approximately 100 °C and viewed with natural light (Figure 2). A north-facing window is a good location for such observations. Such studies should be made on samples that have the same general size distribution of particles. The color of samples prepared from the individual sieved-out particle size ranges gives important additional data.

Instrumental determination of color provides more quantitative information and can now be done with a precision that exceeds the human eye. Pye (2007) found that rapid, reliable, and highly reproducible results are achieved using the Minolta™ CM-2002 Photospectrometer. Calibration of the instrument is carried out for every analysis to black, white, and CERAM II color standard materials. Guedes et al. (2009) provided a study of the application of instrumental methods for forensic application.

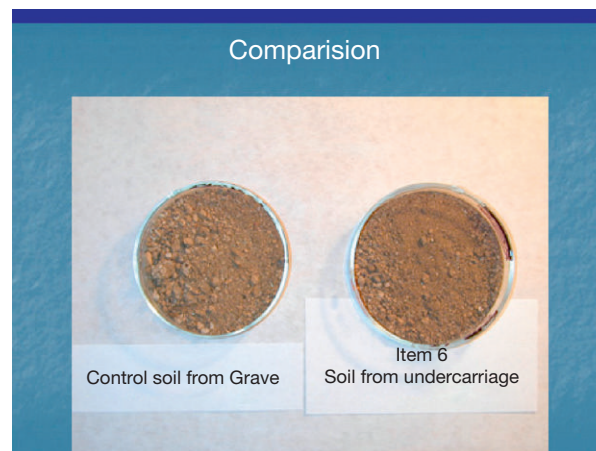


Figure 2 Two soil samples available for examination for color comparison.

Particle Size Distribution

The determination of the distribution of particle sizes in a sample can often provide significant evidence. This determination may be done for a variety of reasons (1) to produce samples for comparison studies that are similar, in which case, the control sample may contain some larger or smaller particles that are not present in the questioned or associated sample and they must be removed; (2) the samples may be broken down into subsamples in which all the particles are in the same size range for mineral or color studies; (3) a determination of the distribution of sizes of particles may be produced as a method of comparison. A diagram showing the distribution of grain sizes can be used as a presentation and in many cases may be of evidential value.

The basic methods used for the separation of sizes are (1) passing the sample through a nest of wire sieves, with the size of the openings decreasing from top to bottom (Figure 3), (2) determining the rate of settling of the grains in a fluid, which is a measure of the size of the particles, and (3) using instruments that measure the size of particles in a microscopic view and record the number of particles of each size. The distribution of particle sizes is then plotted on a diagram.

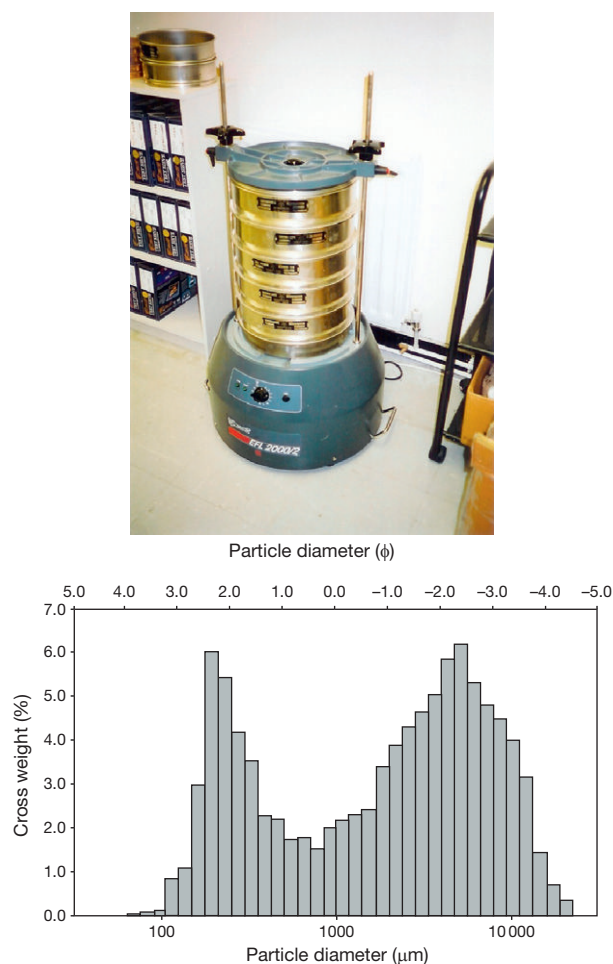


Figure 3 Nest of sieves used to separate particles into different size grades and a diagram showing the size distribution of particles.

When a transfer is made from a soil it is unlikely that the transferred sample is truly representative of the distribution of sizes that existed at that place. For this reason, two samples that are to be compared using grain size distribution may not show exactly the same distribution of grain sizes. The method can be useful in coming to the conclusion of comparison or lack of comparison, but is seldom definitive in itself.

Stereo Binocular Microscope

The stereo binocular microscope can be a most useful tool to the forensic geologist. After an analysis of color has been made, the information that can be obtained by examination of soil samples with this instrument makes it the logical first step in the direct examination of particles (Figure 4).

Some of these microscopes are seated on a base that contains a second light source so that objects can be viewed in both transmitted and reflected light. When the transmitted light is polarized the microscope may be used for both stereo reflected light viewing and low-power transmitted polarizing light studies. Objects as small as $\sim 10 \mu\text{m}$ in diameter can be viewed with the stereo binocular microscope.

In examining a soil sample or similar material, the scientist will commonly first examine the whole sample as it is received and observe the types of grains and particles. Recording a general impression of the material is normally done at this time. It is not uncommon to observe foreign materials in soil samples, such as fibers, metals, paint, glass, and plastics. These objects can have important evidential value. In some situations they can be the most important parts of the sample. Plant particles can also be of great value.

With a clean sample, the experienced scientist can identify the rocks and minerals at sight or by using simple tests. In addition, it is possible to observe the texture and coatings on the surface of the grains. Properties of grains such as shape, degree of rounding, weathering, inclusions, color, and polish can be observed and recorded. However, the samples may be so different on first examination that further work is not useful because a determination of common source is ruled out.



Figure 4 Forensic scientist using a stereo binocular microscope to identify particles in a soil sample.

The Petrographic Microscope

Petrographic microscopes have a rotating stage and a polarizing filter under the stage that transmits light vibrating usually in a N-S direction (front to back). Above the stage a second, removable polarizing filter is placed in the tube of the microscope. It transmits light usually in an E-W (left to right) direction. When the upper filter is inserted, light is blocked from passing through the microscope. However, an anisotropic mineral on the stage will cause the light path to be altered, thus making the mineral visible. Isotropic minerals and glass remain dark. Many methods are available to the examiner that permit the identification of minerals and rocks using this type of microscope. The instrument is most important in the identification of the relatively uncommon heavy minerals, after they have been separated from the common light minerals (Figure 5).

Refractive Index

The index of refraction of a transparent material is the ratio of the velocity of light in a vacuum, normally considered to be 1, to the velocity of light in the material being analyzed. Thus a refractive index of 2.4553 means that light travels 2.4553 times as fast in a vacuum than in the transparent material. The measurement of refractive index is one of the most important methods for the study and comparison of glass. Glass appears in a wide variety of crimes such as automobile accidents, breaking and entering, and assault with broken bottles. Today, most forensic laboratories use the semiautomated refractive index instrument GRIM 11 for glass identification. These instruments can produce refractive indices to 0.00007 (Figure 6).

Cathodoluminescence

The instrument used for cathodoluminescence is a lumino-scope that is attached as a stage on a light or a scanning electron microscope (SEM). The specimen, for example, mineral grains or a thin section, is bombarded with a beam of electrons generated by the instrument. When the electrons strike the surface of the specimen, an optical luminescence is

produced, which is seen as a display of colors. The colors and their intensity depend in large part on very small changes in the concentration of trace impurities, the minerals present, and the location of the trace impurities in the structure of the minerals. Thus, the method has wide application in determining or observing a variety of differences in mineral grains that otherwise appear similar.

Scanning Electron Microscope

The SEM has a wide range of magnifications, generally from $25\times$ to over $650\,000\times$, and can record objects as small as 1.5 nm. The depth of field is very large, and most SEM pictures have an excellent three-dimensional appearance. The magnification settings are easily changed, thus facilitating the study of the appearance of the surface of a grain from very low to very high magnification. The presence of very small fossils that were not previously known in a given soil can now be detected during routine examination. Surface features of individual mineral grains such as quartz can be seen and shown to have many different kinds of scratches, pitting, and mineral growth. Some of these features may be useful in telling us the past history of the individual grain. It is not uncommon to observe other minerals such as clay flakes filling the scratches and thus adding another characteristic that can be useful in comparing the minerals.

When using these powerful instruments in forensic work, one must keep in mind that no two objects are ever exactly the same. No two sand grains are ever exactly alike when studied under the high magnifications of an electron microscope. This is true even if they have been side by side for the past million years. Observations made with these instruments can be very useful for establishing similarity or dissimilarity between samples. SEMs also have the ability to determine the elemental composition of the particles being examined. This is possible because X-rays are produced when the electron beam of the microscope strikes a target. The SEM can be coupled to an X-ray analyzer. The emitted X-rays are sorted by their energy or wavelength values, which are related to specific elements, and the analyzer produces information that identifies the elements present in the material being viewed. The amount of each element present is determined by the intensity of the emitted

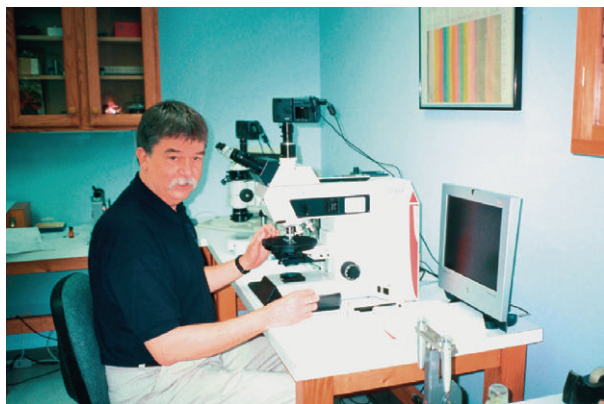


Figure 5 Forensic scientist using a petrographic microscope to identify particles.

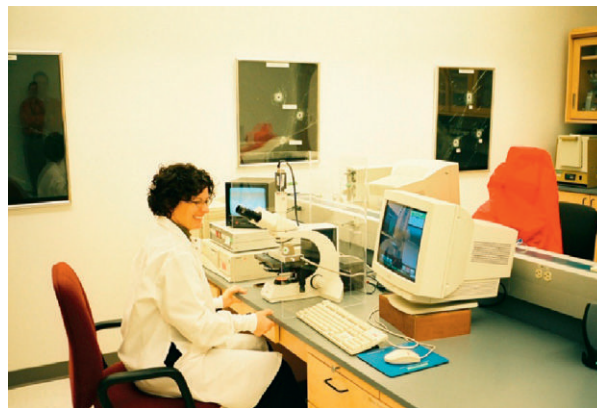


Figure 6 Forensic geologist using GRIM 11 instrumentation to determine the refractive index of a glass.

X-rays. Thus the examiner can determine the chemical composition of the individual particle or particles being viewed.

X-Ray Diffraction

X-ray diffraction is one of the most important and reliable methods of identifying the composition of geologic, soil, and other crystalline substances. The sample is analyzed by passing X-rays through a crystal and measuring their angle of the diffraction as they exit. Each crystalline material has its own distinctive X-ray pattern. The X-ray diffraction pattern of a sample is controlled by the internal structure of the specimen. The diffraction pattern can be collected on film, on an image plate, or by using an electronic detector. The interpretation of X-ray patterns under normal situations is a comparatively simple matter.

Among the strong points favoring the use of X-ray diffraction patterns is that they record the crystal structure. If two substances, diamond and graphite, were analyzed chemically, they would be identical because both are composed of pure carbon, but X-ray diffractograms of the two minerals are quite different. X-ray diffraction is now used as the principal tool in the identification of clay minerals. The chemical composition of clays generally tells us very little as to the nature of clay substances, but the possibilities of identification of clays by X-ray diffraction are almost unlimited (Figure 7).

Chemical Methods

There are a large number of instruments and methods that measure the chemical composition of organic and inorganic materials. There are times where these methods can be very useful and provide valuable information. For example, the chemical composition of glass is often a valuable addition to the optical and physical properties. Identification of organic compounds such as a fertilizer in a soil adds a whole new dimension to the list of properties used in determining comparison.

Methods for determining the amount and kind of elements in a sample may rely on the fact that, as one of its properties, an element selectively emits or adsorbs light. This is the basis for emission spectroscopy and atomic adsorption spectrophotometry. Neutron activation analysis



Figure 7 Forensic geologist using an X-ray diffractometer to identify minerals.

is a nondestructive method. Organic compounds contain carbon and their identification requires different methods. There are several techniques for separating out the various compounds in a mixture. Generally, these methods depend on the relative amounts of the gas and liquid phases of a compound under fixed conditions. That proportion is a characteristic property of each compound. Compounds that more easily enter the gas phase will disappear from a given location more quickly than more stable substances. The rate at which the more volatile substances depart (and hence the distance they move away over a given time interval) allows them to be identified and separated from other compounds. These methods are generally called chromatography. Identification is also done using spectrophotometers, which measure the light-adsorbing properties of a compound. Mass spectrometers have the ability to uniquely identify compounds if analyzed using proper laboratory procedures. This instrument bombards the sample with high-energy electrons, causing the molecules to lose electrons and become positively charged. This is an unstable state and the molecules immediately break up into fragments. The instrument then passes the fragments through an electric or magnetic field where they are separated according to their masses. This permits the specific identification of the atoms in the sample because atomic mass is a unique property of each of element.

The Future of the Science

The last several years have witnessed a tremendous increase in both the quantity and the quality of forensic examination of soil and related material. The future will depend on how well we address a series of issues that include the following:

1. New methods are being developed that take advantage of the discriminating power inherent in earth materials. The opportunities appear to lie primarily in quantitative mineral identification. For this reason, the development of quantitative X-ray diffraction techniques that will provide quantitative data on the mineral composition of a sample would seem to be an important direction for research.
2. Considerable effort must be devoted to defining appropriate sampling methods and the training of those who collect samples. Communicating to law enforcement personnel the potential evidential value of soil and related material is particularly important, because unless the evidence is collected there will never be an opportunity for forensic examination.
3. There is a tremendous need for studies that attempt to demonstrate the diversity of soils. Such studies provide the opportunity for establishing worthwhile information on the distribution of the various soil types. The information should be made available on databases and widely distributed.
4. There must be a continuing effort at all levels to improve the qualifications of examiners in forensic geology.
5. A problem has developed that results from the availability of new instruments that provide increasingly detailed measurements or observations, such as the SEM. These

instruments are capable of discriminating between individual grains. When you discriminate between individual grains, you lose the ability to say that the sample being examined was once part of, and thus had a common source with, another sample. When this is done, the whole concept of comparison is eliminated and the evidential value is lost. The forensic geologist must choose methods that provide the maximum discriminating power between samples without falsely excluding samples that are, in fact, associated.

The real future of the science lies with education. Investigators and evidence collectors must be taught that earth materials can make a major contribution to justice if they are properly collected and studied. The real challenge in all criminal investigations is the production of the best evidence by the most skillful and objective people to serve the cause of justice. Our system of justice is still run by trained advocates (judges and lawyers). Those who the courts honor by soliciting their professional opinions, the expert scientific witnesses, must provide evidence to the highest scientific standards. If they do not do this, the advocates may find ways to remove expert testimony, and then we would return to a judicial system in which the only admissible testimony comes from human witnesses reciting their stories from memory.

See also: [Use of Quaternary Proxies in Forensic Science: Use of Geology in Forensic Science: Search to Locate Burials.](#)

References

- Dudley RJ (1975) The use of color in the discrimination between soils. *Journal of the Forensic Science Society* 15: 209–218.
- Fitzpatrick RW, Raven MD, Heath M, and Rinder G (2007) How soil evidence helped solve a double murder case: A display. *2nd International Conference on Criminal and Environmental Soil Forensics*, Edinburgh UK, 30 October to 1 November 2007. Book of Abstracts.
- Gross H (1893) *Handbuch für Untersuchungsrichter*. Munich.
- Guedes A, Ribeiro H, Valentim B, and Noronha F (2009) Quantitative colour analysis of beach and dune sediments for forensic purposes: A Portuguese example. *Forensic Science International* 190: 42–51.
- Junger EP (1996) Assessing the unique characteristics of close-proximity soil samples: Just how useful is soil evidence? *Journal of Forensic Sciences* 41: 27–34.
- Murray RC (2004) *Evidence from the Earth*, p. 226. Missoula, MT: Mountain Press.
- Pye K (2007) *Geological and Soil Evidence*, p. 335. Boca Raton, FL: CRC Press.
- Ruffell A and McKinley J (2008) *Geoforensics*, p. 332. Chichester: Wiley-Blackwell.
- Sugita R and Marumo Y (1996) Validity of color examination for forensic soil identification. *Forensic Science International* 83: 201–210.

The Use of Macroscopic Plant Remains in Forensic Science

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Introduction

Forensic botany applies the knowledge and techniques of plant science to legal matters. Here, the term macroscopic plant remains is given to those plant materials not included within forensic palynology or microbiology. Research centered on spores, pollen, and certain microorganisms is well developed and will not be discussed here. For decades, these materials have been used successfully by archaeologists, geologists, anthropologists, and botanists to determine the cause of death for prehistoric or modern humans.

Recent articles show the efficacy of forensic mycology as an important tool for forensic scientists (Page, 2011). Medical mycology is a highly developed field, but except for attention to some psychoactive fungi, this subject matter has not been fully used by investigators.

One of the early documented cases of forensic botany connected with macroscopic plant materials was the suicide death of Socrates. Plato described the death of his mentor as he attended the legally imposed suicide of Socrates. He was convicted of corrupting youth and disrespecting the state religion. Because Socrates was of high social standing, he was allowed to choose his own manner of death. He selected a deadly tea made from poison hemlock (*Conium maculatum* L.: Apiaceae). In Plato's *Phaedo* (Plato and Gallop, 2009), we read of Socrates' symptoms after he drank the fatal brew. This narration agrees with contemporary descriptions of poison hemlock's effect on humans (Lewis and Elvin-Lewis, 1977).

From that time to this, in most of the world's societies the knowledge of plants' effects on humans has appeared in courts (Simoons, 1998). Forensic botany became accredited in the courts of the United States in the trial of Bruno Hauptman who was accused of kidnapping and killing Charles and Anne M. Lindbergh's baby son in 1932 (Graham, 1997, 2006). Arthur Koehler, a wood anatomist with the US Forest Service, matched the wood from the ladder used to get into the second floor Lindbergh nursery with wood from Hauptman's attic. Hauptman was convicted of the crime and executed. The US Federal Bureau of Investigation called Koehler's evidence 'critical.' This crime also resulted in kidnapping becoming a federal offense.

Legitimizing Botanical Evidence

Historic Justification

In trials concerning major crimes, a forensic botanist sometimes is asked whether botanical science has the appropriate scientific background to be relevant. Here facts concerning the history of botany can be presented. The ancient illustrations of plants in herbals have served as guides to medicinal properties of plants including poisons for thousands of years. Another answer to such queries is that botanical science is at least as old as the joining of the printing press, invented in the 1400s, with

the appearance of light microscopy (early 1600s). The early books of note include Robert Hooke's *Micrographia* (1665), Marcella Malpighi's *Anatome plantarum* (1675 and 1679), and Nehemiah Grew's *Anatomy of Plants* (1682) that showed the properties of plants and plant cells.

Evidence Tests

When a judge questions the efficacy of botanical evidence, legal tests must be taken and passed by the forensic botanist. These tests were inspired by recognition that 'junk science' had found its way into courtrooms and was being passed off as accurate 'scientific' evidence. This problem is reviewed and clearly attacked in a recent report from the US National Academy of Sciences (USNAS, 2009). Two legal tests are used for the admissibility of scientific evidence. The first and less rigorous one is the Frye test, which originated in 1923 (United States v. Frye, 293 F. 1013. D.C. Cir. 1923). Here a judge must determine whether an expert's methodology is 'generally accepted' by others in the scientific community. Because many arguments arose about the concept of 'generally accepted' and because of concerns from the science community at large, a second, more rigorous test called the Daubert Challenge is now in place (US Supreme Court case, Daubert v. Merrell Dow Pharmaceuticals, Inc., 509 US 579, 1993). The Daubert Challenge is a hearing conducted before the judge of a case where the validity and admissibility of expert testimony is challenged by opposing counsel. The expert scientist is required to demonstrate that his or her credentials, methods, and reasoning are scientifically valid and can be applied to the facts of the case. Forensic botany usually is readily admitted, unlike certain other kinds of evidence.

General Usage of Forensic Botany

Forensic botany tends to be underused in criminal investigations, perhaps because too few botanists are involved in the field and because most other forensic investigators, such as medical examiners and coroners, have educational backgrounds lacking botany.

DNA Evidence

Trends occur in science. For example, in forensics the use of DNA evidence has become especially important because of its power to identify individuals of the species *Homo sapiens*. DNA testing at present is time-consuming and costly, although this is changing. Sometimes unnecessary DNA testing is performed because of its overstated efficacy. DNA patterns have proven invaluable in the identification of individual humans and have shown the innocence of some. However, comparable genetic information is known for a very few of the 300 000 known

species of higher plants. In almost all cases, using DNA to identify a plant species is a waste of time and money. However, Coyle (2005) has summarized well the uses of plant DNA's value in forensics, especially in cases involving suspected drug plants. When forensic questions center on an individual plant, such as a specific plant associated with a particular crime, DNA identifications can be powerful (Graham, 2006). A case in point would be recent attempts to have a piece of an angiosperm identified to order or family, not species, where costs became prohibitive. Funding often is limited, especially for defense experts in cases where the suspect does not have a great deal of money.

Botanical Applications

The applications of botany to forensic investigations can involve almost any aspect of plant science (Lane et al., 1990). The three subdivisions of botany to be discussed here are:

1. plant anatomy, the study of plant cells and tissues that contain them,
2. plant taxonomy, a subdivision of plant systematics dealing with correct identification of a species, and
3. plant ecology, the relationship between plants and their environment.

These will be discussed in order. In the discussions, case examples, without exception, are based upon personal experience.

Plant Anatomy

Plant anatomy owes its origin to the invention of the light microscope. Hooke's (1665) discovery of plant cell walls in his early drawings remains timely and accurate. The value of food plant cells as forensic evidence in stomach contents depends on the nature of cell walls. Cellulose, the structural material, is undigested by the human digestive tract. Even though the living contents of a plant cell can easily be destroyed, the shapes and sizes of the cellulose cell walls remain remarkably constant during their passage through the human digestive system.

Plants in diets

Browsing and grazing materials in animal diets have been studied by zoologists in many ways, leading to better management of domestic livestock and wildlife (Borchichevski, 2009; Perez-Barberia and Jordan, 1998). Human food plants have been largely ignored except for those scientists studying preserved human remains. For example, the diets of the bog people from Europe, Eurasia, and North America are well studied. Here, scientists reconstructed diets of prehistoric people using remnant plant material associated with preserved human remains. Also, palynology's role, as reported elsewhere, has been highly significant in this work.

The initial challenge in identifying food plants in the human digestive tract came with the knowledge that cell sizes and shapes found in ordinary food plants were not cataloged. Sometimes closely related plants show similar cell patterns

such as apricot and peach flesh (Figure 1). However, an examination of these two photomicrographs shows apricot's distinctive crystals that peaches lack.

Case: This case came as a request from a medical doctor who was an assistant coroner and professor of pathology at a medical school. He wanted to know the plant food contents of a victim's last meal. He sent prepared slides with samples of her stomach contents. Based upon experience in preparing laboratories for courses in plant anatomy, some cells were identified as coming from certain foods.

The victim's last known meal differed from some of the food plant cells present in her stomach contents. This meant she had had another meal before her death; and she would not have done so alone. This suggested her fellow diner could have been her killer. He was. He later confessed to her murder (Sachs, 2002).

Plant cells in the stomach

The above case led to an ongoing collaboration with David O. Norris, comparative endocrinologist and expert in vertebrate physiology, including that of humans (Bock and Norris, 1991). Botanical knowledge alone is insufficient in investigations dealing with human digestion. Once food is swallowed, digestion proceeds in the stomach where the food is immersed in an acid bath, primarily HCl, and subjected to shaking through involuntary muscle activity. Stomach digestion varies in duration, varying in time from less than an hour to as long as 4 h. At its conclusion, the chyme passes through the pyloric valve into the small intestine. This pyloric valve or pyloric sphincter closes at death. This means the stomach contents are held in place until the stomach is subjected to further trauma or the stomach wall decays. Stomach contents sometimes yield important information about time of death.

Case: Several years ago, two young women disappeared on the same January night from the vicinity of a ski resort in the Colorado Rockies. They bore superficial resemblance to each other, but were not acquainted. One young woman's remains were discovered 3 days later along a nearby highway. The following June, the second young woman's body was found in a melting snow drift beside a minor road. The authors were sent the stomach contents of these victims. The stomach contents of the two showed they had eaten similar foods shortly before they were killed by gunshots to the upper torso. In this case, the plant foods were associated with Mexican cuisine, a great favorite with people in the area (Figure 2). The distinctive cell shapes and sizes of the tomato skin and chile pepper seed coat are components of such foods.

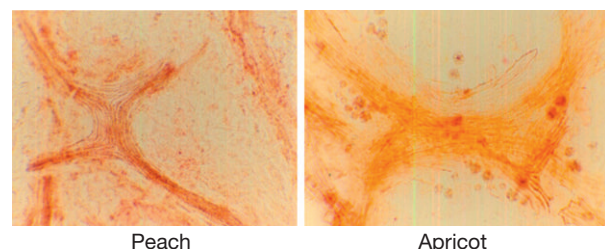


Figure 1 Left picture: peach flesh at 10 $\times$; vascular tissue cells similar to those in apricot flesh (see right picture). Right picture: apricot flesh at 10 $\times$; note crystal inclusions not present in peach flesh (see left picture).

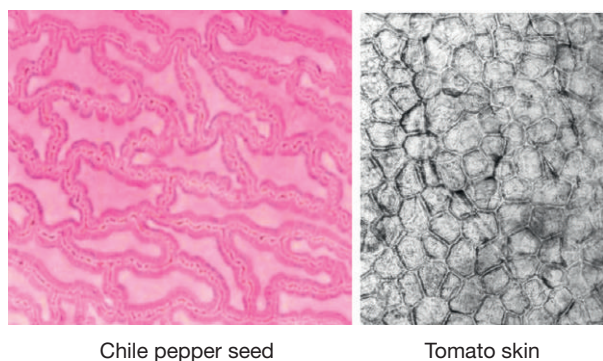


Figure 2 Chile pepper seed coat at 25 $\times$ (left). Tomato fruit epidermis at 10 $\times$ (right).

Research

An investigation was done to determine the effects of stomach digestion on plant food cells. Clumps of food plant cells were suspended in HCl solutions approximating those found in human stomachs. The vessels containing the suspensions were placed in a 37 °C water bath stationed on a shaker and left for up to 24 h. No changes occurred in cellulose wall shape or size. Subsequent work demonstrated that the cellulose of plant cells maintains its shape regardless of other processing treatments including freezing, boiling, and roasting (Bock and Norris, 1997).

Methods based on research

Many different food plant cells were measured and a guide prepared (Bock et al., 1988). Today, digestive tract specimens are identified using low power microscopy and a photographic record is taken of the findings (Figure 3). The photos here are of the pulp of an olive. The scanning electron microscope shows more clearly the oil droplets common to olives, but this more expensive technique is not necessary for food plant cells in our experience. This sort of photographic evidence plus verbal testimony has been used in many homicide investigations. When in doubt about a food plant's identity, samples of suspected foods are purchased, ground, placed on microscope slides, and compared with the unknown (Ruzin, 1999; Sass, 1958).

Case: An incarcerated person was murdered in a prison. The suspects in the case were questioned about when they had last seen the victim alive. Was it before or after the 5 p.m. dinner hour? Dieticians in places of confinement such as hospitals and prisons record menus. The last food plant materials in the victim's stomach were from the mid-day meal, not dinner, showing certain inmates were not telling the truth.

Case: Sometimes food plant cells found in human feces can be used as evidence in court. Plant cells from fecal evidence in a rape-homicide were pivotal in a trial. Fecal matter from the victim and from the victim's and suspect's clothing was compared. These data, when presented in court, agreed in kinds of plants and the relative frequencies of the specific cell types. These data were highly significant in the conviction of the suspect (Norris and Bock, 2000, 2001).

Plant Taxonomy

Plant taxonomy is a subdivision of plant systematics that deals with assigning an accurate scientific name to a plant. In order

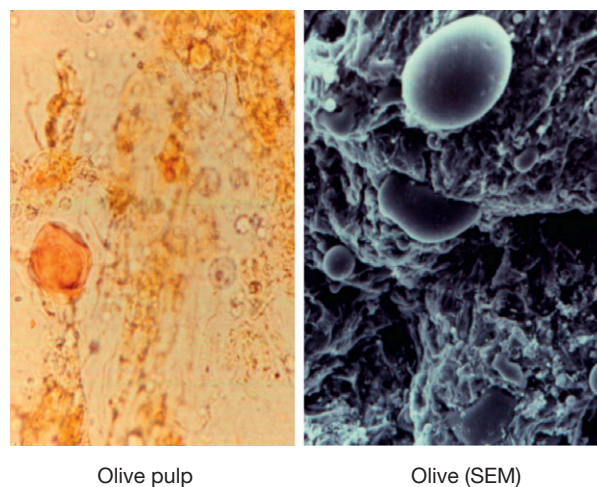


Figure 3 Olive pulp at 25 $\times$ (left); oil droplets and crystals present. At 280 $\times$ magnification (SEM, right) oil droplets are more distinct.

for the name to be current it should adhere to the most recent rulings of the *International Code of Botanical Nomenclature* (McNeill et al., 2006). Such matters are treated each 6 years at International Botanical Congresses (<http://www.abc2011.com>) by systematists and other botanists from all over the world. Their sessions are always interesting.

Plant identification is the most commonly used tool in forensic botany at present. With such evidence, a victim can be linked to a suspect, a suspect to a place, and a vehicle linked to a suspect, a victim, or both.

Collection of evidence

The collection of plant material for use in criminal investigations differs from techniques taught in plant systematic courses. Forensic collections are assumed to be legal evidence. Such materials need to be collected, if possible, either by officers of law enforcement organizations or by a botanist in the presence of officers. Rules surrounding evidence are strict. When significant vegetation is collected, a chain of evidence must be established at once. Notebook records of time and place and case numbers are required. It is wise, but not required, to assign your personal case number that will be linked to the number that will be used in court. This information must always remain attached to the evidence. Each person in possession of evidence must be clearly documented as the evidence passes among those involved in a case.

Plant collections should be placed either in paper or cloth bags unless pollen analysis also is to be undertaken. Bags need to be the smallest size to accommodate the material. Evidence can be stored in laboratories or evidence lockers for long periods of time, even years. Evidence rooms always are short of space, so economy of collection without minimizing the value of the specimens is essential. Plastic bags, glass jars, and tin cans are unacceptable for long-term storage because they encourage decay.

Identification of specimens

Plant evidence comes from whole plants or plant fragments. It is helpful to know where the plant grew so identification can

be narrowed. For example, it is much better to start using a region-specific scientific flora such as *Colorado Flora Eastern Slope* (Weber and Wittman, 2001) than first consulting *Flora North America*, although you might want to check your identification in the latter or in other floras. The more restricted the flora, the easier the identification becomes. Specialty keys are especially helpful, for example, *The Cacti of Arizona* (Benson, 1969) or *Illustrated Keys to the Grasses of Colorado* (Wingate, 1994). Try to avoid floras prepared for the tourist trade because they sometimes lack scientific rigor.

Paleontological guides to leaf fossils also can be a short cut to contemporary tree and shrub identification.

Plant collection at crime scene

In early stages of an investigation, it may be necessary to characterize the general vegetation of a crime scene. This is best done by a botanist accompanied by an officer who is familiar with the site. If the botanist has familiarity with the general area, leaves, flowers, fruits, seeds, or pieces of bark make sufficient floristic samples. For general collections, more than one species can be put in the same collection bag. If plants require further identification in the botanist's laboratory, larger specimens are taken. This often is necessary for vegetative (i.e., specimens not in flower) Poaceae and Cyperaceae. On the collecting bags or on attached evidence tape, record the date and global positioning system (GPS) location of the collection plus the case number. Also, note this information in a bound notebook.

Collecting plant material from clothing

These items should be collected with forceps. Paper coin envelopes are suitable for storing individual samples. Be sure each envelope is labeled with the case number, date, and place. Be as specific as possible. Place only one species in an envelope. Often these sorts of collections will require laboratory work. You may have a few seeds, a fragment of moss, or partial tree leaf that cannot be field identified. Be sure to note which article of clothing and on what place the specimen was found. For example, the label could read the 'right rear pocket of mustard-colored Bermuda shorts' or 'inside sock near toe.'

Case: A rape-murder victim was found on a golf course in the Caribbean. This course, like many in warm climates, was covered in Bermuda grass (*Cynodon dactylon* (L.) Pers.: Poaceae) a plant native to East Africa. The head groundsman took special pride in the strain he had developed for the course's greens. This grass strain appeared almost hair-like. The clothing and shoes of one of two suspects were examined for plant fragments. Pieces of this special strain of Bermuda grass were found in the shoes and socks of one suspect, and none were found on his Bermuda shorts or his shirt. This was used to show that he had been at the golf course on the night in question but he had not participated in the rape and murder because no grass fragments were found on or in the clothing covering the upper portion of his body. Due to this circumstantial evidence he was spared the death penalty, although he confessed to being a witness to the crime.

Collecting plant material from exterior of vehicle

Usually this is done on impounded vehicles. With an investigator, a botanist collects materials from the underside of the car. Places to be collected include the tire treads, wheel wells,

fenders, bumpers, window wiper slots, and the undercarriage itself. For small plant objects, the collection is made with forceps, using paper coin envelopes for storage. Larger pieces of vegetation may be attached to the undercarriage and stuck in the wheel wells, and larger paper bags are needed. See section 'Collection of Evidence' for label instructions.

Collecting plant materials from inside vehicles

Almost all plants and plant fragments will be small. For small objects, the collection is made with forceps and paper coin envelopes. Places where plant materials are found include on and under floor carpets and seat covers, in the grooves on pedals, behind seats, in the boots (trunks), and under the bonnets (hoods). Investigative departments often possess vacuum sweepers for evidence collection from carpets in homes and in vehicles. Try to avoid having these machines used for materials you must examine. They pulverize the plant pieces.

Case: A man was suspected of having killed his wife. Her body was found in the Rocky Mountains of Colorado at an elevation of just over 3000 MASL. The man's truck was impounded and vegetation specimens were collected from it. The suspect claimed that his truck was used only in his work on the eastern Colorado plains and was not taken to elevations above about 1880 MASL. Investigators visited the place where the victim's body was found and sampled a reference collection of vegetation. They also collected the vegetation around the couple's home and surrounding grassland. A comparison of the vegetation of the two collection sites with the plant fragments found on and in the vehicle showed clearly that the truck had been at the higher elevation. This was not conclusive evidence that the truck had been in the exact spot at 3000 m where the body was found because there were nearby canyons that held similar vegetation. This circumstantial evidence showed that the suspect lied about where the vehicle had been. This, plus other evidence, led to a murder conviction.

Plant Ecology

Ecological methods have considerable value in certain kinds of investigation. A plant ecologist must be aware of plant succession patterns in the region related to a crime. The plants indigenous to the region in question should also be known. This knowledge often is used to help locate clandestine graves. Recently buried animals, including humans, enhance growth and greenness of the vegetation above the site and this can be evident once decay has proceeded. The late Bombeck (1976) wrote a book entitled *The grass is always greener over the septic tank*. As this title implies, buried decaying matter supplies fertilizer to the plants growing above, and gives them a distinct appearance in comparison with surrounding vegetation. Depending upon the environmental conditions of a burial site, these sorts of clues can persist for as short a time as a few weeks in tropical and semitropical damp environments, but they can remain in place much longer in cold and dry habitats. Another ground level indicator occurs as decay proceeds. Subsidence of the topsoil takes place due to the reduction in overall volume of the remains. In recent years, this has helped in the search for mass graves of war victims (Brown, 2006).

Colonizing plant species can indicate where an organism was deposited. These are the plant species, often short-lived,

that thrive on bare ground. In time they tend to disappear or be greatly reduced in prominence because they are succeeded by other species that compete well with other plants. Succeeding plants also respond to the slight environmental modifications that colonizers engender. In both tropical and temperate areas, this process of species succession can indicate where bodies were buried for many years, even decades after the burial. Also, knowledge of plant phenology can be helpful. Plant parts associated with crimes may indicate the season when an event took place, in contrast to the time when the grave was discovered.

Research on buried bodies

Aspects of decomposition by buried animals have been a topic of research over the last few decades (Sachs, 2002). The most famous of these is associated with the University of Tennessee's Anthropology Department's Forensic Anthropology Center, commonly referred to as the Body Farm (Bass and Jefferson, 2003). Valuable information about human decay processes has come from here.

Necrosearch International is a volunteer organization whose purpose is to search for clandestine graves (Jackson, 2002). The members have learned a great deal from burying pigs on a law enforcement training ground south of Denver, Colorado. The pigs were buried at various times over the past several years at varying depths, sometimes naked, sometimes wrapped in plastic, and sometimes clothed. This activity was named Project P.I.G. (Pigs In Ground). Now, P.I.G. is used extensively in crime-scene training programs.

Necrosearch has succeeded in finding the graves of missing persons in several parts of the United States and abroad. Forensic botany is an integral part of its program. Plant anatomy, plant taxonomy, and plant ecology methods are used.

Necrosearch investigations usually involve teams of specialists. For example, a team comprising a forensic botanist, a cadaver dog handler, and a ground penetrating radar expert could be used to find a missing person. Some of the other experts in Necrosearch include geologists, forensic anthropologists, forensic entomologists, criminal profilers, DNA analyzers, medical pathologists, and ground and aerial photographers.

Education for Forensic Scientists

General Education

There are no specific programs at present for training a forensic botanist. However, higher education institutions ranging from junior colleges to major research institutions offer certificates and degrees in criminal justice, *sensu lato*. Individual specialization within these programs is advised. Such studies can prepare a student to become a law enforcement officer, a para-legal, or a forensic laboratory technician.

Most, but not all, crime-scene investigation courses do not include forensic botany in the curriculum. Courses with botany included have been given to police academy training courses, FBI seminar courses on new techniques in crime-scene investigation and courses for coroners, medical examiners, and criminal lawyer groups. Without exception, these sorts of groups have embraced the techniques of forensic botany (J. Bock, personal experience).

Advanced Degrees in Forensic Studies

Thesis topics that prepare one for work in forensics can be obtained within an advanced degree program. People with PhDs in botany, entomology, anthropology, geology, paleontology, computer science, and most other sciences, along with individuals holding medical and law degrees, can modify their careers in ways to become forensic specialists. Many careers for forensic scientists are linked to local and state forensic laboratories. Alas, the otherwise outstanding national FBI laboratory at Quantico, Virginia, has no forensic botanist on staff. However, major laboratories in other countries including New Zealand, Australia, England, Canada, and Great Britain are staffed with forensic botanists.

Acknowledgments

I am especially indebted to Dr. David O. Norris, Dorothy C. Sims Esq., and the Hon. Haskell Pitluck for fostering justice through science.

See also: [Diatoms. Use of Quaternary Proxies in Forensic Science: Soils and Sediment.](#)

References

- Bass W and Jefferson J (2003) *Death's Acre*. New York: G.P. Putnam's Sons.
- Benson LD (1969) *The Cacti of Arizona*. Tucson: University of Arizona Press.
- Bock JH, Lane MA, and Norris DO (1988) *Identifying Food Plant Cells in Gastric Contents for Use in Forensic Investigation: A Laboratory Manual*. Washington, DC: US Dept. of Justice, National Institute of Justice.
- Bock JH and Norris DO (1991) A new application of plant anatomy. *Strategies for Success in Anatomy, Physiology, and Life Science*, vol. 6, pp. 6–8. New York: Benjamin/Cummings Publishers.
- Bock JH and Norris DO (1997) Forensic botany: An under-utilized resource. *Journal of Forensic Sciences* 42: 364–367.
- Borchchevski V (2009) The May diet of Capercaillie (*Tetrao urogallus*) in an extensively logged area of NW Russia. *Ornis Fennica* 86: 18–29.
- Brombeck E (1976) *The Grass is Always Greener over the Septic Tank*. New York: McGraw Hill.
- Brown AG (2006) The use of forensic botany and geology in war crimes investigations in NE Bosnia. *Forensic Science International* 163: 204–210.
- Coyle HM (ed.) (2005) *Forensic Botany*. Boca Raton, FL: CRC Press.
- Graham SA (1997) Anatomy of the Lindbergh kidnapping. *Journal of Forensic Sciences* 42: 368–377.
- Graham SA (2006) Crime-solving plants. *Plant Science Bulletin* 52: 78–84.
- Grew N (1682) *Anatomy of Plants*. London: W. Rawlins.
- Hooke R (1665) *Micrographia*. London: Johannis Martyn.
- Jackson S (2002) *No Stone Unturned*. New York: Kensington Books.
- Lane MA, Anderson LC, Barkley TM, et al. (1990) Forensic botany: Plants, perpetrators, pests, poisons, and pot. *Bioscience* 40: 34–90.
- Lewis WH and Elvin-Lewis MPF (1977) *Medical Botany*. New York: Wiley-Interscience.
- Malpighi M (1675/1679) *Anatome Plantarum*. London: Impensis Johannis Martyn.
- McNeill J, Barrie FR, Burdet HM, et al. (2006) *International Code of Botanical Nomenclature (Vienna Code)*. *Regnum Vegetabile* 146. Liechtenstein: A.R.G. Gantner.
- Norris DO and Bock JH (2000) Use of fecal material to associate a suspect with a crime scene: Report of two cases. *Journal of Forensic Sciences* 45: 184–187.
- Norris DO and Bock JH (2001) Method for examination of fecal material from a crime scene using plant fragments. *Journal of Forensic Identification* 51: 367–377.
- Page D (2011) Mycology: Missing weapon in forensic arsenals? *Forensic Magazine* 8: 15–18.
- Perez-Barberia FJ and Jordan IJ (1998) Factors affecting food comminution during chewing in ruminants: A review. *Biological Journal of the Linnean Society* 63: 233–256.

- Plato and Gallop D (2009) *Phaedo*. Oxford: Oxford University Press.
- Ruzin SE (1999) *Plant Microtechnique and Microscopy*. Oxford: Oxford University Press.
- Sachs JS (2002) *Corpse*. Cambridge, MA: Perseus Printing Services.
- Sass JE (1958) *Botanical Technique*. Ames: The Iowa State University Press.
- Simoons FJ (1998) *Plants of Life, Plants of Death*. Madison: University of Wisconsin Press.
- United States National Academy of Sciences (2009) *Strengthening Forensic Science in the United States: A Path Forward*. Committee on Science, Technology and Law and Committee on Applied Theoretical Statistics. Washington, DC: USNAS.
- Weber WA and Wittman RC (2001) *Colorado Flora: Eastern Slope*. Boulder: University Press of Colorado.
- Wingate JL (1994) *Illustrated Keys to the Grasses of Colorado*. Denver: Wingate Consulting.

Introduction

Forensic scientists use insects to address questions of legal relevance. This discipline is called 'forensic entomology.' The approaches taken in forensic entomology and Quaternary science are similar: entomological material is interpreted in the light of ecological and biological information. The morphology of insects and sometimes other arthropods is used to identify species; knowledge of life cycle durations and habitat preferences is used to interpret aspects such as length of colonization and time since death. This interpretation becomes part of the evidence in the court. Amendt et al. (2007) proposed standards and guidelines to ensure good forensic entomological practice.

Forensic entomology can be classified into several areas. Matters relating to time since death of a person or animal, where entomology is used to determine postmortem interval (PMI), and the likelihood of neglect or abuse are considered by criminal courts. Cases utilizing forensic entomology in the civil courts tend to relate to urban entomology, such as insect infestations of structural material or complaints about insects in food and stored products (Gennard, 2012).

In urban environments, insects which are perceived as a nuisance, such as bees, wasps, and flies, can lead to court cases. For example, the stable fly *Stomoxys calcitrans* (Linnaeus) can reduce milk production by a small percentage. Thus, its presence has an economic cost to dairy farmers as well as engendering dislike because this species bites humans, cattle, and even dogs. Bad farm management or poor hygiene, which encourages the presence of such flies, is likely to result in a civil case to force the removal of the nuisance. Interestingly, in the tropics, an increase in populations of *S. calcitrans* has been positively correlated with rises in precipitation and temperature. Their presence can also have indirect forensic relevance. Goff (1991) considers *S. calcitrans* an indoor species on the island of Oahu, Hawaii. Therefore, from a forensic perspective, larvae of *S. calcitrans* on a body found in the open in Hawaii would indicate that the body had been moved.

Evidence of insect infestation of building materials can also have legal implications. For example, purchasers could sue for damages against the estate agent, solicitor, surveyor, and professional timber specialist, on the basis of lack of information, if they found that a house survey failed to record signs of insect infestation, once they had purchased the house. The basis of such a case would be that if the buyers had known about a prior (albeit treated) infestation by, for example, the common furniture beetle (*Anobium punctatum* DeGeer), which the surveyor considered unlikely to recur, they would not have continued with the property purchase. In such instances, the entomological information would be the subject of the case. As in this hypothetical example, provision of entomological advice to solicitors about urban entomological issues constitutes a further, and often major, aspect of the work of the forensic entomologist.

Stored product pest civil cases are usually related to food infested by insects: weevils in biscuits, or the remains of flour moths in a sack of flour. Adult cockroaches are frequently the topic of forensic investigations in buildings such as hotels and hospitals. However, in Iran cockroach nymphs of *Blatta orientalis* (Linnaeus) have also been recorded from excised human viscera (Tüzün et al., 2010).

Any insect, or indeed other arthropod, can be of forensic value as a proxy to determine what happened in the past. It has either to be the subject of evidence in court or provide information that assists in interpreting a legal matter. Some insects can be present by chance and hence are of no legal importance, while others are there because of the presence of favorable food, which may include dead bodies.

Insects for which a corpse provides both food and a habitat are termed necrophages. Insects or other arthropods such as spiders, present because they feed upon the adults, larvae, or eggs of the necrophages, are termed predators. Some species can consume both the corpse and organisms living on the corpse; these are termed omnivores. The insects present, as it were, by chance are termed accidentals. Such carrion-related species are drawn to the body either by decomposition odors alone, or they are attracted by the presence of other insect species inhabiting the corpse.

Research on both scavenger species such as the Turkey Vulture and those insects causing myiasis (such as flies colonizing living organisms) has shown that they are olfactory foragers, attracted by the smell emanating from the food source. The nature of these attractants to the corpse is a matter of ongoing research.

The odor profile of decomposing bodies changes over time, both in terms of the odors present and of their proportions. Different insect assemblages are attracted to a body during the various stages of decomposition. Thus the presence of different insect species on a corpse provides evidence of the length of time since death. The stages of decomposition through which a body passes are based on visual descriptions. They are fresh, bloat, active decay, advanced decay, and the dry stage – skeletal remains (Table 1).

Speed of decomposition varies with environmental conditions.

Work by Statheropoulos et al. (2005) explored the decomposition volatiles of two individuals estimated to have been dead for between 3 and 4 weeks. The greatest contributors to the odor profiles were dimethyl disulfide, toluene, hexane, benzene, 1,2,4,trimethyl-2-propanone, and 3-pentanone. However, little was known about the origins of the deceased and the influence that the cause of death had on the odors revealed.

Initial chemical attractants of *Calliphora* species include dimethyl disulfide and dimethyl trisulfide: chemicals that have been shown to illicit a positive electrophysiological response from species such as *Calliphora vomitoria* (Linnaeus; the Holarctic blue bottle fly). Vass et al. (1992) considered that after 1285 accumulated degree hours (ADH), volatile fatty

Table 1 Stages of decomposition of a body

Visual features of a decomposing body	Decomposition stage
No obvious change from normal; no odor	Fresh
Body is swollen and there is some body fluid seepage; livor mortis (looks like expansive bruising) may be obvious. Later in this stage the skin may split. Some odor is noticeable	Bloat
Strong odor; the body has deflated and often the head has become defleshed in places and the eyes have sunk into their sockets	Active decay
Soft tissue covering the skull has become reduced in quantity, often there is only a small amount of tissue in the abdominal region. The odor is less obvious	Advanced decay
Bones, hair, and dried skin remain. There is no noticeable odor.	Skeletal remains

acids such as propionic, valeric, and butanoic acids fall to background levels, thereby providing an explanation for the reduction in the number of insect species attracted to the body in the last stages of decomposition.

Insect Life Cycles

The key insect species used by forensic entomologists are those that feed directly on the body or food, as they provide a means of determining how long the body has been present at a particular location, or how long the food has been infested. Determination of the life cycle duration and body size (e.g., length) in each of these stages is a key tool in forensic entomology calculations. The majority of relevant research is laboratory-based. The average time spent in each of the life stages at fixed temperatures has been calculated, along with the average length (or width) of the larvae in that stage. This information is then interpreted according to the life cycle stage found at a crime scene, or recovered from a particular situation that the forensic entomologist is being asked to examine. There is a specific physiological energy budget for each stage. Speed of insect growth is related to temperature, so it is possible, based on our knowledge of insect larval growth rates, to determine the contribution towards the total physiological energy budget provided at the crime scene in each hour or day. The total of these hourly or daily contributions (ADH or accumulated degree days) reflects the total amount of time taken to reach a given stage in larval development, and hence time since death or colonization by the specific larval instar on the body. Key morphological features such as spiracle shape, structure, and slit number that relate to identification of the life stage are used to determine species and life cycle stage.

Both beetles (Coleoptera) and flies (Diptera) are insect orders which contain forensically relevant species that have a complete life cycle with morphologically distinct stages: egg stage, larval stage, pupal stage, and adult stage. The egg stage provides some indication of species but cannot easily provide anything other than a broad time frame for the purposes of PMI determination.

Calliphoridae (blowflies) have three larval instars. The first (LI) is the result of hatching from the egg and continues until the next molt. The second instar larva (LII) exists between this

molt and the next. The final instar (LIII) exists between the periods of second molt and pupation, although the cuticle (outer skin) is retained and tanned and becomes the puparium in which the fly pupa subsequently forms.

From the perspective of species relevance as proxies in a Quaternary context, it is likely to be the remnants of larval spiracles on puparia that provide evidence useful for environmental reconstruction. To determine time since death (PMI), the time to reach the oldest larval growth stage found on the body is used as an estimate. Some entomologists, in situations where the temperature is fairly consistent at the scene, use maggot size (length or width) to determine the time since death, assuming that food is not a limiting factor as it will influence the size of the maggot (Figure 1(a) and 1(b)).

The insects of major importance in forensic entomology are discussed under their family headings below. The list is split into two; the majority of families discussed have terrestrial forensic relevance and are flies or beetles (Diptera or Coleoptera). The last two families, one a Diptera and the second from the Trichoptera (caddisflies), have aquatic forensic relevance but have also been shown to have a valuable role in Quaternary science. The specific insect species that visit a body will vary depending upon their habitat and biogeographic range.

The Forensic Value of Members of Specific Insect Families

Diptera (Flies)

Calliphoridae (blowflies)

A number of members of the Calliphoridae are valuable forensic tools. In particular, *Calliphora vicina* Robineau-Desvoidy (the European blue bottle fly) and *C. vomitoria* are species commonly used to determine the time of death.

These species are primary invaders and readily colonize the body after death. *C. vicina* is especially associated with urban environments, although its territory can be expanded into less developed areas close by. *C. vicina* is valuable, as it is found in a range of habitats and has even been noted on a body in a cave where the temperature was consistently 5 °C.

Calliphora vomitoria is also a primary colonizer but tends to be the first species to arrive at bodies found in more rural environments. This species is attracted to decomposition odors such as the sulfides – for example, methyl disulfide, methyl trisulfide, and methyl tetrasulfide (Gennard, unpublished data).

Other *Calliphora* species of note include *Calliphora hilli hilli* Paton (a subspecies of Hill's brown blowfly) and *Calliphora stygia* Fabricius (the common brown blowfly), species that are found in Australasia. Their significance is the result of the studies undertaken because these species are often causal agents of myiasis (parasitic infestation by fly larvae) in sheep. Therefore they have an economic significance.

A number of *Lucilia* species (green bottle flies) also have forensic significance and have been found to colonize bodies. They include *Lucilia sericata* Meigen the common green bottle fly, *Lucilia caesar* (Linnaeus), *Lucilia richardsi* Collin, *Lucilia illustris* Meigen, and *Lucilia cuprina* (Wiedemann). The rate of egg laying (oviposition) in *L. cuprina* has been shown to increase where there is opportunity for the flies to make contact

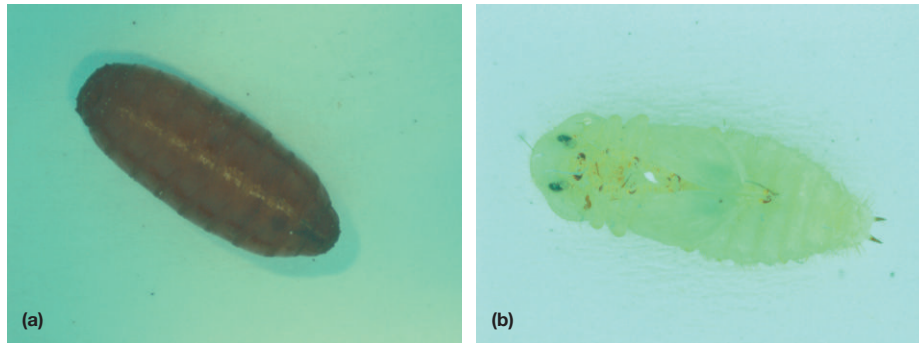


Figure 1 (a) The puparium of the fly. (b) The beetle pupa.

with water (Barton Browne, 1962); so environmental conditions are significant.

Lucilia sericata (called *Phaenicia sericata* in North America) can be an initial colonizer where temperatures are sufficiently warm (Horenstein et al., 2010). It tends to like warm, sunny habitats and prefers to lay its eggs on carcasses in direct sunlight. Therefore, it will be the colonizer of bodies found in exposed conditions such as open fields or on moors, rather than where shade is provided by canopy vegetation.

Protophormia terraenovae Robineau-Desvoidy (the northern blowfly) is found early in spring in the United Kingdom and is cold-tolerant. It is therefore, an early colonizer, and its range has recently expanded into other locations in Europe. Significantly, the pupal cases (puparia) of this species are found in direct association with the body, rather than some distance away from it, because pupation (the process by which the larva becomes a pupa) takes place on the body. *P. terraenovae* was recorded from the Holocene in Greenland, and is very much more significant in Scandinavia as a forensic indicator than in other parts of Europe.

Phormia regina Meigen (the black blowfly) is recorded from bloat stage in forests in western Poland (Matuszewski et al., 2008). The decomposition of pig carcasses is often used by forensic scientists as a proxy for human body decomposition. Field observations of *P. regina* larvae on pig carcasses showed that they lived on the carcasses for a period of 15 days (days 5–20 after death); this is a much shorter period of habitation than that of other blowfly species. Larvae of *L. caesar*, for example, were present from day 3 to day 31, while *C. vomitoria* larvae were present from day 4 to day 38.

The species of *Chrysomya* found on bodies include *Chrysomya rufifacies* Macquart (the hairy maggot blowfly), *Chrysomya megacephala* (Fabricius) (the oriental latrine fly), and *Chrysomya albiceps* (Wiedemann). These species are spreading beyond their historical geographic ranges, presumably as a result of changes in climate. For example, *Ch. megacephala* is now recorded from Spain and Portugal, whilst *Ch. albiceps* has extended its range to colonize Poland, the Ukraine, and Northern Italy. *Ch. albiceps* is considered a secondary invader, where members of the Calliphoridae subfamily are present as primary invaders (Denno and Cochran, 1975). Interestingly, in the warmer months in Argentina, *Ch. albiceps* is thought to be able to limit the number of eggs laid by *L. eximia* (Wiedemann) if the former colonizes the body first (Horenstein et al., 2010).

Cochliomyia macellaria (Fabricius; the secondary screwworm fly) is a primary colonizer. However, its population can be reduced if *Ch. rufifacies* is present on the body at the same time, especially in the later instars. A similar effect seems to occur if *Ch. megacephala* and *Ch. rufifacies* coexist on a body.

***Sarcophagidae* (flesh flies)**

Members of this family are significant in that they will lay live larvae (larviposit) on corpses, rather than lay eggs. Flesh flies are known to have a wide geographic distribution. For example, a study of carrion flies in three areas of the Amazon rainforest by Amat (2010) (flood plain forest, white sand forest, and terrestrial forest) has revealed that Sarcophagidae had the greatest species richness.

In colder seasons of the year, some flesh flies, for example *Sarcophaga africa* Wiedemann, colonize a body after *C. vicina*, but act as the initial colonizer when the weather is warmer. If it has been raining when the body was placed at a location, or where the person died outside, flesh flies may also be the initial colonizer. They can fly in the rain, whereas calliphorids do not (Smith, 1986). So in this sense, sarcophagids, in the absence of calliphorids, are a proxy for a past weather incident.

***Muscidae* (house flies and stable flies)**

Muscidae are flies that are most often associated with feces, urine, and waste material emanating from a body. From a forensic perspective, their value as an indicator is geographically variable. For example, the muscid fly *Synthesiomyia nudiseta* (van der Wulp) was found at corpses during a number of stages of decomposition in experiments in Costa Rica; so it is not specific to a particular stage of decomposition. In experiments conducted in Argentina, it was found only in small numbers in the muscid assemblage and the value of this species as a postmortem indicator has been questioned (Calderón-Arguedas et al., 2005). Muscid flies, for the most part, are species that arrive at a body soon after death when body fluids have been exuded, or later when there has been leakage of decomposition fluid in sufficient quantity to provide them with a suitable food source. In Argentina, Horenstein et al. (2010) found that they were the second most abundant family collected from carcasses. Their presence at crime scenes may also give them a value as an indicator of seasonality.

Flies such as *Musca domestica* Linnaeus (the common house fly) may be equally attracted to decaying vegetable matter and become a pest associated with farming or domestic compost

heaps, which brings them into the realm of forensic entomologists. These flies are also found in restaurants, kitchens, and other warm places.

Members of the genus *Hydrotaea* (dump flies), or its synonym *Ophyra*, have been recorded from corpses by a number of researchers. *Hydrotaea dentipes* (Fabricius) is known from a wide range of habitats including cesspools, decaying plant material, fungi, and carrion, as well as on bodies discovered both inside buildings and outdoors. Larvae of this species (and those considered part of the *H. dentipes* group) hatch from the egg as first instar larvae which are either predators, reducing the number of other larvae present at a location, or saprophagous (feeding on dead or decaying animal matter). This is not true of all species of *Hydrotaea*, since they can also emerge from the egg into the second instar larva. Other species of *Hydrotaea* such as *Hydrotaea rostrata* Robineau-Desvoidy (the black carrion fly) have also been recorded from forensic contexts.

Fanniidae

Forensically relevant Fannid flies, much as muscids, feed on fecal material and areas of the ground or clothing stained by such body fluids. They are not specific to any particular stage of decomposition, although Horenstein et al. (2010) found them in greater numbers in winter and spring than in summer and autumn in central Argentine experiments. Fannid larvae are flattened unlike those of the Calliphoridae, and so are easily distinguishable. They have obvious lateral protuberances.

Fannia species of significance include *Fannia scalaris* Fabricius, the latrine fly. This species is particularly associated with human excrement. The lesser house fly *Fannia canicularis* Linnaeus is particularly recognizable at the egg stage. Its eggs have a modified corion (outer coat), with what appear to be two floats, allowing the eggs to be suspended above the surface of the fluid in which they have been laid. This species too is attracted to excrement and dead bodies of animals and humans, although it has been recorded as feeding on living organisms.

Fannia canicularis is considered to be an agent of urinogenital myiasis and therefore can be an important insect species in cases of suspected neglect. It is recorded as a secondary sheep strike agent in some countries and will infest wool that has been soiled by urine and feces. In forensic contexts, the larvae have been associated with advanced stages of decomposition.

Phoridae (scuttle flies)

This family is called scuttle flies because of the way the adults move across a surface if disturbed. They are extremely small flies, which colonize the body early in decomposition, and have been known from both buried bodies and those that died indoors. *Megaselia scalaris* (Loew) is a scuttle fly species found on bodies indoors as well as outside. It has a recorded lifecycle of 19–22 days at a temperature between 20 and 25 °C and a relative humidity constantly greater than 50% (Leccese, 2004).

Puparia of *Megaselia abdita* Schmitz were recorded as the dominant insect of evidence by Manlove and Disney (2008) in a case where the body was discovered in January. By determining the temperature of the location where the deceased had been found, and relating this to the time taken for the phorid to pass through a complete life cycle at that temperature, it was possible for them to propose a minimum time since death.

The species most commonly found in association with buried bodies is *Conicera tibialis* Schmitz (the coffin fly). This species, once mated, is considered to be able to pass down through the soil to the body, and lay eggs such that the larvae can colonize the body as a source of food (Colyer, 1954). The next generation then re-emerges from the soil.

Members of the Phoridae are also of value as postmortem indicators in winter. For example, *Triphleba autumnalis* (Becker) can be found as larvae on carrion between December and February, although it enters diapause (a period in which growth or development is suspended and physiological activity is diminished) to survive conditions during the rest of the year (Lundt, 1964).

Sphaeroceridae (lesser dung flies)

This family comprises small black flies around 1.5–3.5 mm long, and the basitarsus on their hind leg is short and dilated. Sphaerocerids can be found in habitats ranging from decomposing vegetation and seaweed to organic material in caves or on dead bodies. Smith (1986) considers that they are indicative of the advanced decay stage in response to the presence of fatty acids, but Erzinçlioğlu (1983) recorded them at the start of the decomposition process.

This is an example of the need for specific ecological knowledge about the colonizing sphaerocerid species. Amendt et al. (2010) describe the case of a body sealed into a mine shaft from which they recovered 20 third-instar larvae (LIII) of *Leptocera caenosa* Rondani. This species has a preference for dark, cool habitats such as caves and cellars or similar habitats. It also prefers to colonize bodies at the advanced stage of decomposition. This preference allowed Amendt et al. to estimate a period of 30 days from oviposition to post feeding for this species, which they concluded had colonized the body once it had been placed down the mine shaft.

Piophilidae (cheese flies or skipper flies)

Piophilids of forensic interest are attracted to the active decay and also the dry stage of body decomposition. Members of the family have been found on human cadavers, human excreta, bones, skin, and fur. They are also thought to be present when volatile fatty acids are emitted from the body. Their period of colonization is short, as is their life cycle. Piophilids have been found associated with food materials such as cheese and meat, and one particular species, *Piophilidae casei* (Linnaeus), is called the Cheese Skipper because of this association. Specimens of *P. casei* have even been recovered from Egyptian mummies, so the species has broad forensic relevance.

Russo et al. (2005) determined the duration of each life stage of *P. casei* at a range of constant temperatures (15, 19, 25, 28, and 32 °C). They showed that adult longevity was temperature-dependent and that the net reproductive rate was greatest at 25 °C. Such data is valuable, as it is the basis by which the forensic entomologist determines time since death/colonization.

Sepsidae (black scavenger flies)

Members of this family, commonly called scavenger flies, are associated with bodies at an advanced stage of decomposition. They are small, shiny flies, distinctive because of the constriction in the region of their first abdominal segments. The adults

are distinguished from piophilids or sphaerocerids by their trait of wing wagging as they move. The significance of their presence on the corpse varies because they are also associated with feces. One species, *Orygma luctuosum* Meigen, usually found breeding in seaweed, has been found to be of forensic relevance and its presence on a body provides a link to marine environments.

Stratiomyidae (soldier flies)

These are an indicator of the later stages of decomposition. Larvae of the black soldier fly *Hermetia illucens* Linnaeus, although originally a species native to America, have been recorded on corpses in a number of places and more recently in southern Europe. Larvae of stratiomyids are characterized by having a layer of calcium carbonate over their body surface. This gives the larval body a roughened appearance. It also protects the larvae from drying out. The species has additionally been recovered from drug hauls of cannabis.

Several forensically relevant species with an aquatic life stage also have Quaternary value. Trying to date the time since submergence presents a number of difficulties for the forensic entomologist. In general, a range of species of macroinvertebrates are used, which may include some families of insects. The calculations are based on the proportions of various members of different functional feeding guilds. The main difficulty for the forensic entomologist arises because species that colonize the body will vary both geographically and seasonally. There are some families of insects with aquatic life stages which are particularly relevant. These are discussed below.

Chironomidae (nonbiting midges)

This is one of the families of aquatic insects that have forensic relevance in crime scenes where the body has been recovered from water. These midges are also of value in Quaternary science. Subfossil chironomids have long had a role in paleolimnological research.

One notable forensic case, which characterizes the role of chironomid larvae (Figure 2), is that of the 'red fiber.' Trace-evidence forensic experts were unable to identify a red-colored 'fiber' dried onto a surface of a body recovered from water. In desperation, they showed the material to forensic



Figure 2 Chironomid larva.

biologists. The biologists later confirmed the 'red fiber' as a member of the chironomid family which had become attached to the murder victim whose body had previously been submerged in water (Hawley et al., 1989). The subsequent observation of chironomids on other bodies led to them to be considered a good indication of submergence and of value for the calculation of the submerged interval.

Trichoptera (caddisflies)

A caddisfly assemblage can provide information on the local aquatic habitats, based on their biological and distributional information. As such, they are of use in both Quaternary and forensic science. Caddisflies have been found on submerged bodies (Wallace et al., 2008) and used to determine the duration of time since submergence. The value of the particular families of caddisflies varies with the circumstance (Figure 3).

Net-spinning caddisflies (Hydropsychidae), although of value in Quaternary studies, do not necessarily provide a specific means of determining the postmortem submerged interval. Those species of caddisfly that make their cases from vegetable matter and sand and other fragments can often be more useful. For example, Wallace et al. (2008) used the seasonality and the duration and timing of life stages, including the onset of dormancy, of two species of Limnephilid caddisflies to estimate the time of submergence of a body at a crime scene. These forensic entomologists were able to infer that the body was submerged between late April and late May.

Coleoptera (Beetles)

Histeridae (clown beetles)

Members of this family are often missed in forensic assessments, as they are found beneath the body and tend to be most active at night. They are predators of the fly larvae present on bodies. Therefore, they can potentially reduce the numbers and range of fly species present upon which the PMI can be based.



Figure 3 Caddisfly larva uncased.

The population of Histerids tends to be greatest in the active decay stage of decomposition when fly larvae populations are also large. The distribution of specific species colonizing a body may vary with the stage of decay. Matuszewski et al. (2008), for example, in studies on corpse decomposition in forests in western Poland, recorded species of *Hister* and *Saprinus* as restricted to the active decay stage, while *Margarinotus* species' presence extended into the advanced decay stage as well. In other areas of Europe, the Histerid species, *Hister unicolor* Linnaeus and *Pachylister inaequalis* (Olivier), are most frequently recovered from bodies (Figure 4).

Cleridae (checkered beetles)

Clerid beetles are found on bodies during a range of decomposition stages. They not only consume the remains of cadavers but also prey upon the eggs and larvae of necrophagous flies present there. The family becomes of significance at the later stage of corpse decomposition when members of the genus *Necrobia* can be recovered from the body.

Clerids are also stored-product pests and so may have relevance to forensic entomologists who are involved with stored-product pest cases. They are also the cause of damage to, or found in association with, archaeological finds. *Necrobia rufipes* (DeGeer) the red-legged beetle, for example, was observed on the mummy of Ramses II.

Silphidae (carrion beetles)

Silphids are present on corpses located outdoors. Some will feed on the corpse directly; others will feed on both the corpse and the other insects present on the body, such as the larvae, while others feed solely on larvae of insects such as flies.

There are two Silphidae subfamilies; the Silphinae and the Nicrophorinae. Watson and Carlton (2005) distinguished the importance of these two subfamilies in terms of attraction to size of corpse rather than habitat; Silphinae are found in rural locations and are attracted to larger corpses and hence have a greater value in forensic terms. In part, this may be determined by the food preference of the particular species. Adult *Silpha* species, for example, consume fly larvae, whilst the larvae of this species consume the remains of the body.



Figure 4 Histerid beetle.

Different species of silphid dominate, depending on the habitat type. Nicrophorinae are considered to be attracted to small carcasses and they also vary in the habitat in which they are found. For example, *Nicrophorus humator* Gleditsch, (the sexton beetle) and *Nicrophorus vespillo* Linnaeus tend to be restricted to forests. Michaud et al. (2010) found that *Nicrophorus tomentosus* (Weber), the gold-necked carrion beetle, was present in both fields and forests, although other members of the Nicrophorinae showed a preference for forest habitats (Figure 5).

One of the common silphid species, considered to be a forensic indicator and found on both human and pig corpses in Southeastern Brazil, is *Oxyletrum discicollae* (Brullé; Carvalho et al., 2000). Mise et al. (2008), however, recorded the presence of this species over 126 days, from the fresh stage until the body had entered dry decay stage and fly larvae had left the bodies. In North America, *Necrodes surinamensis* Fabricius and *Oiceoptoma inaequale* Fabricius can be found on the corpse at the same time. They have three larval instars; which specific larval instar is present can most readily be distinguished by using the distance between the dorsal ocelli (simple eyes; Watson and Carlton, 2005).

Dermestidae (skin or larder beetles)

Different species of dermestids are known from corpses in a range of decomposition stages. They are also a cause of damage to museum specimens and an economic pest recognized from dried and preserved food.

Three species are perhaps of particular forensic relevance, and have been recorded from archaeological contexts. These are *Dermestes frischi* Kugelman, *Dermestes maculatus* DeGeer, and *Dermestes lardarius* Linnaeus.

Dermestes frischi is found predominantly in the wild and therefore from outdoor bodies. Drugs present in the tissues of the corpse can be recovered from the larvae and pupa of this species. The advantage of being able to use pupae from some insect species is that they may be recovered many weeks, months, or years after the death or consumption of the material concerned. Therefore, they have long-term value as a forensic indicator. The two other forensically significant species of dermestid, *D. maculatus*, known as the hide beetle, and *D. lardarius*, commonly called the bacon or larder beetle, are also pests of stored products, fur, and other animal products, and are found

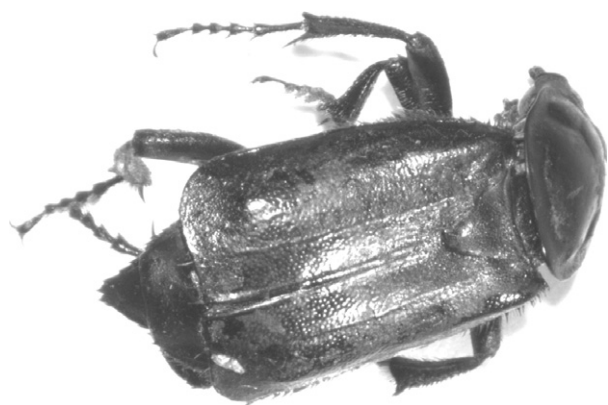


Figure 5 Silphid beetle, *Nicrophorus humator* Gleditsch.

on corpses in certain circumstances. Much of the stored product damage is caused at the larval stage when their larvae burrow into material such as wood, hide, or even concrete, in order to pupate in a location safe from cannibalism by other larvae. The adults and the larvae, which have hairs, will consume meat and carrion as well as other animal products. The adults do not consume wood or books as a food source, however.

D. maculatus has been found to be a pest of stored museum specimens. It has also been found when new museum exhibits were examined. The Bristol Mummy, when it was unwrapped, for example, revealed the presence of *D. maculatus* which had attacked the body of Han-Em-Kem-Esi. It appeared that the beetles had become incarcerated in the wrapping, having been attracted to the dead body as it was prepared during embalming, prior to entombment (Figure 6).

Scarabaeidae (dung beetles)

Members of this family are recorded from carcasses, although they may not be present in large numbers. The majority of specimens of this family found on pigs modeling human corpse decomposition have been from the genera *Onthophagus* and *Aphodius*. Michaud et al. (2010) similarly recorded these genera amongst the scarabaeid species captured from the body of a pig, in a Canadian study. They were able to compare the scarabaeids collected from the bodies of pigs in forests and agricultural fields. There was a greater abundance of Scarabaeidae colonizing bodies left in fields than those in forests. The assemblage species remained stable throughout the year, irrespective of the season.

Ururahy-Rodrigues et al. (2008) made a particularly important point about the taphonomic effect that some Scarabaeidae might have. In their work in Brazil, they showed that *Coprophanæus lancifer* (Linnaeus), a scavenger dung beetle, could have a significant effect on the position of the body. This species is necrophagous and feeds its larvae on fragments of the decomposing corpse. It will lay eggs in chambers located near to a corpse, often tunneling beneath it. This behavior can result in postmortem damage to the body, as recorded by the authors, since adult dung beetles bite through the tissue and can actually move the body physically and/or disturb the disarticulated bones.

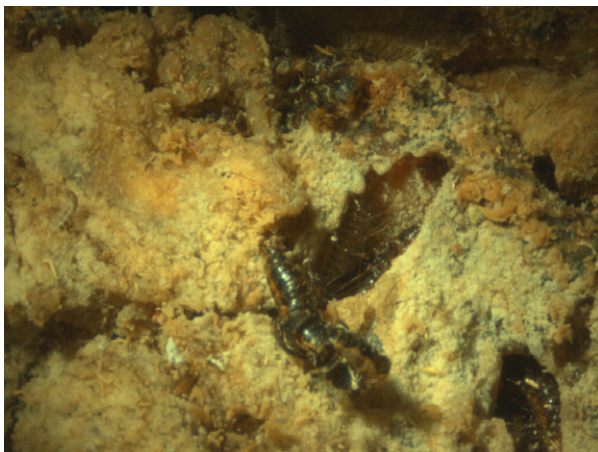


Figure 6 Dermestid larva consuming flesh.

Nitidulidae (sap beetles)

These are minute beetles, smaller than 6 mm in size, found at the drier stages of decomposition. *Omosita discoidea* (Fabricius) is an example of a sap beetle which may be recovered from a corpse or decomposing pig carcasses at the later stages of decomposition, frequently in the dry stages. Another nitidulid beetle, *Glischrochilus quadripunctatus* Linnaeus, has been found on bodies in autumn where the corpse is in mixed woodland. The latter species has also been recovered from rotting fruit (Figure 7).

Staphylinidae (rove beetles)

Members of this family are found in the environment of the body and feed off both the eggs and larvae of necrophagous species. They, like spiders, can reduce the population of indicators of time since death. They confirm that the body has been in place sufficiently long to attract flies and for them to have laid eggs on the body. Only those species that lay eggs in the soil near the body have a forensic significance for calculation of the PMI. They may be the dominant beetle family present on a body, and in this case reduction in the number of eggs or larvae may result, and this should be taken into account in interpreting the results of the PMI. This is true especially if there is evidence of a missing larval instar in the samples of fly larvae collected; for example, there are lots of instar LI and LIII but few larvae in instar LII. Such a disruption in the sequence of larval instars present on a body is most often considered to be the result of removal of a body from one site to another.

Limitations in the Use of Insects in Forensic Entomology as Proxies

The major limitations to the use of insects for forensic purposes are either

- i. a lack of appropriate data recovered from the crime scene, or
- ii. a paucity of ecological knowledge about the specific insect species recovered from the body in relation to that particular crime scene.



Figure 7 A saprophagous nitidulid beetle (*Glischrochilus* sp.).

The majority of calculations of time since death relate to the larval stage and which particular larval instar and species is present. Once the larvae have left the body, it becomes more difficult to interpret the PMI. Several researchers have explored this aspect, particularly in relation to the rate of change of the larva as it shortens and thickens and subsequently changes its color as the resulting puparium (the pupal case) tans. Up to 24 h after pupariation starts (the period of color change from a white-colored puparium to its final tanned color), it may be possible, under fixed conditions, to estimate the age of the puparia. Thereafter, it may be possible to examine the pupa itself to determine stage of development. However, in both instances further research is necessary on the part of the forensic entomologist conducting the case in order to decide the time period concerned.

The time between the start of the postfeeding stage when the larva starts to wander and the point at which there are metabolic changes triggered by the transition into a puparium remains a problem in terms of estimating the time since death. Work by Charabidze et al. (2008) on the crawling speed of *Protophormia terraenovae* indicated that at 23 °C the maximum speed of movement for a 14-mm larva was 27.8 cm min⁻¹. This calculation allowed the time since the larva had left the body to be determined at this temperature, based upon distance away from the body.

A further limitation in the use of forensic entomology is sparsity of knowledge about life cycles of fly parasites, for example, the parasitic wasp *Nasonia vitripennis* (Walker), a jewel wasp. Parasitoids cause the death of the particular specimen they parasitize. However, with further research, it may be possible to determine the time since parasitization of the colonizer. This could provide a measure of time since death using a different organism and under these circumstances makes parasitoids a useful forensic indicator.

The variation in presence of carrion-inhabiting species in different seasons of the year in countries around the world is also a limitation in their use in forensic entomology. Some insects enter diapause in winter and disappear from view, whilst the distribution of others may vary with season and locality. Further study of insect life cycles and the changes in presence and distribution of species as a result of climate change will help to resolve such limitations and refine the use of insects as proxies for past events.

See also: **Beetle Records: Overview.** **Chironomid Records: Chironomid Overview.** **Paleolimnology: Multiproxy Approaches.**

References

- Amat E (2010) Notes on necrophagous flies (Diptera: Calyptratae) associated with fish carrion in the Colombian Amazon. *Acta Amazonica* 40: 397–400.
- Amendt J, Campobasso CP, Gaudry E, Reiter C, LeBlanc HN, and Hall MJR (2007) Best practice in forensic entomology – Standards and guidelines. *International Journal of Legal Medicine* 121: 90–104.
- Amendt J, Campobasso CP, Goff ML, and Grassberger M (eds.) (2010) *Current Concepts in Forensic Entomology*. Dordrecht: Springer Science and Business Media B.V.
- Barton Browne L (1962) The relationship between oviposition in the blowfly *Lucilia cuprina* and the presence of water. *Journal of Insect Physiology* 8: 383–390.
- Calderón-Arguedas O, Troya A, and Solano ME (2005) Larval quantification of *Synthesiomia nudisetia* (Diptera: Muscidae) as a criteria in analysis of the post

- mortem interval in an experimental model. *Parasitología latinoamericana* 60: 18–143.
- Carvalho LML, Tyson PJ, Linares AX, and Palhares FAB (2000) A checklist of arthropods associated with pig carrion and human corpses in southeastern Brazil. *Memórias do Instituto Oswaldo Cruz* 95: 135–138.
- Charabidze D, Bourel B, LeBlanc H, Hedouin V, and Gosset D (2008) Effect of body length and temperature on the crawling speed of *Protophormia terraenovae* larvae (Robineau-Desvoidy) (Diptera: Calliphoridae). *Journal of Insect Physiology* 54: 529–533.
- Colyer CN (1954) More about the 'coffin' fly *Conicera tibialis* Schmitz, (Diptera, Phoridae). *The Entomologist* 87(1094): 129–132.
- Denno RF and Cochran WR (1975) Niche relationships of a guild of necrophagous flies. *Annals of the Entomological Society of America* 68: 741–754.
- Erzinçlioğlu Z (1983) The application of entomology to forensic science. *Medicine, Science, and the Law* 23: 57–63.
- Gennard D (2012) *Forensic Entomology: An Introduction*, 2nd edn. Chichester: Wiley.
- Goff ML (1991) Comparison of insect species associated with decomposing remains recovered inside dwellings and outdoors on the island of Oahu, Hawaii. *Journal of Forensic Sciences* 36: 748–753.
- Hawley D, Haskell NH, McShaffrey DG, Williams RE, and Pless JE (1989) Identification of a red 'fiber': Chironomid larvae. *Journal of Forensic Sciences* 34: 617–621.
- Horenstein BM, Linhares AX, de Ferradas BR, and Garcia D (2010) Decomposition and dipteran succession in pig carrion in central Argentina: Ecological aspects and their importance in forensic science. *Medical and Veterinary Entomology* 24: 16–25.
- Leccese A (2004) Insects as forensic indicators: Methodological aspects. *Aggrawal's Internet Journal of Forensic Medicine and Toxicology* 5: 26–32.
- Lundt H (1964) Ecological observations about the invasion of insects into carcasses buried in soil. *Pedobiologia* 4: 158–180.
- Manlove JD and Disney RHL (2008) The use of *Megaselia abdita* (Diptera: Phoridae) in forensic entomology. *Forensic Science International* 175: 83–84.
- Matuszewski S, Bajerlein D, Konwerski S, and Szpila K (2008) An initial study of insect succession and carrion decomposition in various forest habitats of Central Europe. *Forensic Science International* 180: 61–69.
- Michaud J-P, Majka CG, Privé J-P, and Morceau G (2010) Natural and anthropogenic changes in insect fauna associated with carcasses in the North American Maritime lowlands. *Forensic Science International* 202: 64–70. <http://dx.doi.org/10.1016/j.forsciint.2010.04.028>.
- Mise KM, Martins CBC, Kob EL, and Almeida LM (2008) Long decomposition process and influence on coleoptera fauna associated with carcasses. *Brazilian Journal of Biology* 68: 907–908.
- Russo A, Cocuzza GE, Vasta MC, Simola M, and Virone G (2005) Life fertility tables of *Piophilidae* L. (Diptera: Piophilidae) reared at five different temperatures. *Environmental Entomology* 25: 194–200.
- Smith KGV (1986) *A manual of Forensic Entomology*. London: Trustees of the British Museum (Natural History).
- Statheropoulos M, Spiliopoulou C, and Agapiou A (2005) A study of volatile organic compounds evolved from the decaying human body. *Forensic Science International* 153: 147–155.
- Tüzün A, Dabiri F, and Yüksel S (2010) Preliminary study and identification of insects' species of forensic importance in Urmia, Iran. *African Journal of R N Biotechnology* 9: 3649–3658.
- Ururahy-Rodrigues A, Rafael JA, Wanderley RF, Marques H, and Pujol-Luz JR (2008) *Coprophanaeus lancifer* (Linnaeus 1767) (Coleoptera, Scarabaeidae) activity moves a man-size pig carcass: Relevant data for forensic taxonomy. *Forensic Science International* 18: e19–e22.
- Vass AA, Bass WM, Wolt JD, Foss JE, and Ammons JT (1992) Time since death determinations of human cadavers using soil solution. *Journal of Forensic Sciences* 47: 542–553.
- Wallace J, Merritt R, Kimbirauskas MS, Benbow ME, and McIntosh M (2008) Caddisflies assist with homicide case: Determining a postmortem submersion interval using aquatic insects. *Journal of Forensic Sciences* 53: 219–221.
- Watson EJ and Carlton CE (2005) Succession of forensically significant carrion beetle larvae on large carcasses (Coleoptera: Silphidae). *Southeastern Naturalist* 4: 335–346.

Relevant Websites

- <http://www.eafe.org> – Website of the European Association for Forensic Entomology.
- <http://www.forensicentomologist.org> – Website of the American Board of Forensic Entomology (ABFE). The ABFE is a certification body which is able to award the status of Diplomate to those with a high level of expertise as an expert witness in forensic entomology.

Analytical Techniques in Forensic Palynology

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Introduction

It is uncertain exactly when the first attempt to use pollen in a forensic application might have occurred. The current information suggests that any attempt made prior to the late 1950s, either failed or was never reported. One of the earliest successful cases using forensic palynology occurred in Austria in 1959 (Erdtman, 1969). Here, pollen was used successfully to link a suspect with the crime scene and the shallow grave containing the murdered victim's body. However, for several decades after that murder case there were no reported attempts to use pollen evidence again in a forensic application. An exception was the collection and analysis of pollen from the holy relic called the Shroud of Turin during the 1970s by Max Frei. In that study, Frei used tape pulls and the trapped pollen on the tapes to attempt to prove the Shroud's origin and authenticity (Bryant, 2000). The next application appears to have been in New Zealand during the early 1980s where pollen was used to solve a case involving the illegal poaching and selling of velvet from deer antlers (Mildenhall, 1990). In the United Kingdom, Wiltshire (1993) first brought attention to the use of pollen in helping to solve criminal cases. As a result of that initial work, the United Kingdom now leads the world in the routine application and use of forensic palynology as part of the armory of forensic techniques available to investigators.

In the United States, there was little interest in forensic palynology until after the events of September 11, 2001, when the twin towers of the World Trade Center in New York City were destroyed by terrorists using commercial airplanes. In an effort to determine who had been responsible for the destruction of the World Trade Center, and where the terrorists may have lived, a number of investigations were begun by U.S. government agencies searching for forensic trace evidence that might help identify the terrorists and prevent similar events in the future. Although none of those federal agencies had ever used pollen or spores as forensic tools prior to 9/11/2001, that event triggered the interest of federal agencies to use palynology (the study of pollen and spores) in their investigations. In the years since 9/11, U.S. federal agencies have relied on palynology to resolve a number of issues including the origin and distribution of illegal drugs, the origin of seized contraband goods, the importation and transshipment of illegal goods, the geolocation of aliens and other items seized or captured at border crossings, the potential link between individual criminals and specific geographical locations, the manufactured origin of weapons and explosive devices, where various vehicles seized during criminal investigations had been driven, the geolocation associated with the use of cell phones and laptop computers, and a wide variety of similar applications involving military, criminal, and civil cases. During these examinations questions are sometimes raised by investigators, and by other forensic specialists about the precision, reliability, and accuracy of using what they often call 'soft scientific data' such as soil profiling and the use of pollen or spore evidence.

Analytical techniques in palynology rely on the identification of pollen and spore types and, in most cases, the ratio of each pollen or spore type to the total palynomorph composition recovered in a sample. Thus, it is often the combined identification of both the pollen taxa and the abundance of each taxon that provide the critical data needed for analytical precision. The primary problem, however, has been that light microscopy (LM) is typically limited to the identification of pollen and spore types either to the family (e.g., Cupressaceae, Poaceae, Chenopodiaceae) or to the generic level (e.g., *Picea*, *Quercus*, *Artemisia*, *Pinus*). Unfortunately, one of the drawbacks of using LM analysis is that it is rarely precise enough to identify pollen confidently to the species level. Sometimes, one must rely upon the higher resolution capacity of a scanning electron microscope (SEM), or a transmission electron microscope (TEM) to establish the precise genus and species of pollen grains found in forensic samples (Milne et al., 2005). Examining entire forensic pollen samples using SEM and/or TEM is time consuming, costly, and requires access to expensive equipment and the expertise of competent palynologists familiar with SEM and TEM usage. Although this can, and has been done (Jones and Bryant, 2007), these techniques are rarely feasible based on the funding constraints facing most investigative agencies.

Discussion

In the United States, attempts to increase the identification precision of pollen and spore evidence in forensic applications have explored various new technologies including computer-based imaging programs designed to identify pollen grains (Allen et al., 2007), examination of the DNA of individual pollen grains, the use of isotope values associated with individual pollen grains, and the potential application of Raman microspectroscopy. To date, these techniques are not commonly used in most palynology applications to forensic evidence because of the limitations associated with each technique. Although these techniques can bring some refinements in the levels of pollen and spore identification, none can yet duplicate the full quantitative range of data achieved by using LM pollen analyses.

Light Microscopy Versus Scanning Electron Microscopy

Jones and Bryant (2007) compared the pollen analysis of a forensic sample using both LM and SEM. In that study, they examined the same sample and conducted a complete pollen analysis using each technique and then identified the advantages and limitations of each. They also discussed a number of considerations that should be examined and considered before

determining which technique to use in any type of forensic analyses of pollen.

Extraction and Processing

Initially one needs to determine whether or not the forensic sample should be examined fresh or chemically prepared for analysis. If it is important to know whether or not some, or all, of the pollen in the sample is 'fresh' and was recently dispersed, then special care must be taken not to destroy that type of evidence by chemically removing the cytoplasm. For example, knowing 'when' pollen was trapped in the keyboard of a laptop computer might be a vital and important clue. Knowing whether the pollen was from fossil types that were recycled or from freshly deposited pollen might aid in identifying the time and use location of the laptop. Similarly, recovering fresh *Cannabis* (marijuana) pollen from the clothing of someone suspected of growing those plants illegally could provide stronger evidence for conviction than the argument that the trapped pollen on the clothing came only from previous dispersed pollen that had been recycled. In another instance, a suspect related to a crime committed in a California avocado orchard could not have been linked to the crime scene had the pollen on his clothing been processed using normal techniques. Avocado pollen is extremely fragile and is easily destroyed using even mild oxidizing methods normally associated with the removal of cellulose and other nonpollen organic debris found in forensic samples.

There are a number of reliable extraction procedures that can be used to prepare forensic pollen samples for both LM and/or SEM analysis (Brown, 2008; Faegri and Iversen, 1989; Moore et al., 1991). The primary objective in most cases is to remove as much debris in the samples as possible, thereby freeing the pollen grains for analysis and preventing remaining debris from obscuring part or all of the individual pollen grains during analysis. Nevertheless, one should make an initial observation of samples before chemical extraction to search for and note what types of 'debris,' such as plant cuticle, coniferous tracheids, phytoliths, starch grains, carbon soot, and similar debris might be present. Those items, when present, can often provide additional clues about the origin of a sample. Next, caution should be exercised in selecting laboratory procedures that will remove unwanted debris but which will also prevent the inadvertent destruction of some or all of the pollen. This is especially critical in pollen samples that come from locations where pollen and spores are already partly degraded or might contain suspected pollen types (e.g., Lauraceae) that are easily destroyed using standard chemical treatments, such as acetolysis (a mixture of sulfuric acid and acetic anhydride). For example, in a pretest of the existing pollen and spore preservation in a series of forensic samples collected from a location with high soil alkalinity, different palynomorph recovery rates were found depending on the extraction process that was selected. In that example, the most successful process consisted of using only 5% KOH, HCl, and HF. When acetolysis or other oxidation procedures were added to the extraction process a portion of the poorly preserved pollen disappeared.

Once a forensic sample has been chemically prepared using appropriate laboratory techniques, the resulting residue can be

added to a suitable mounting media for LM analysis or temporarily placed in a sealed vial containing glacial acetic acid for storage or later use for SEM/TEM analysis. One important consideration for LM analysis is to use a mounting media which will allow the analyst to gently roll or move pollen grains mounted under the coverslip on slides. Often positive identification of individual pollen grains can only be confirmed if critical aperture features, wall construction, and surface ornamentation features can be adequately observed. Often, these criteria cannot always be obtained in only one view of a pollen grain, especially if the grain is partly collapsed, broken, folded, or partly obscured by debris. Thus, it is important that the analyst be able to move the pollen grains and spores into different positions during LM analysis to make positive identifications. One of the preferred mounting media used in forensic palynology is glycerine jelly. Once the pollen sample is mixed with glycerine jelly and allowed to dry under a cover slip on a slide, pollen grains are fixed in their orientation and location. However, placing the tip of a heated implement on the coverslip will briefly melt the media enough to move a pollen grain into a slightly different orientation before the media hardens again. The use of this type of mounting media is essential for many types of situations where pollen grains related to criminal or civil investigations must be repeatedly located on the slides for verification by different palynologists working with the prosecution and/or with defense attorneys. Another alternative, which allows for quicker scanning and counting of pollen grains on slides, is the use of liquid glycerine. Slides made with liquid glycerine can be permanently sealed around the coverslip with clear fingernail polish and the position of pollen grains on those sealed slides will not move provided the slides remain sealed and do not leak, and provided the slides are always stored in a perfectly horizontal position. The location of key pollen grains seen on those slides can be marked using the tip of a fine-point, felt-tip marker. In working on forensic pollen samples where information is critical and is time sensitive, the liquid glycerine technique allows for quicker movement of multiple pollen grains into many different orientations, and also permits the rapid scanning of multiple slides while searching for additional, new pollen taxa. This type of rapid analysis is sometimes needed when individuals are being held temporarily until essential information is determined, often through pollen analysis. Nevertheless, when in doubt of which mounting media to use, and if viewing by multiple palynologists over longer periods of time will be needed, the glycerine jelly technique is preferable.

Forensic samples prepared for SEM analysis must also be processed to remove debris and then rinsed twice with 95% ETOH. Next, one drop of the pollen residue should be pipetted from the ETOH mixture onto a sterile SEM aluminum stub that has been previously coated with an adhesive such as TEMPFIX™. After drying, samples must be coated before using. A common technique is to coat the stub with 400 Å of gold palladium applied in a Hummer II sputter coater. Stubs should be identified by data etched onto the sides or bottom with a diamond pen. One or more marks should also be etched on the top edge of each stub to ensure proper reorientation if a stub is removed. This last point is especially critical for two different types of circumstances. First, if a key pollen grain on a stub must be relocated at a later time to check the identification

or if verification, then the marks on the stub surface will allow for correct reinsertion of the stub in the exact original position (Jones and Bryant, 2007). Second, if a complete pollen analysis using SEM is needed, as has been done on occasions, it is time consuming and often takes a number of hours to complete. Because an SEM pollen analysis cannot always be completed within the time available, one must return several times. One does not want to inadvertently recount the pollen on previously viewed transects. Therefore, one must be able to reinsert the SEM stub in the exact position each time. The markings on the top of the stub enable one to reinsert the stub in the exact original orientation.

Equipment, Cost, Time, and Expertise

In most forensic analytical situations cost is a primary concern. Invariably, the length of time to complete a forensic pollen analysis is also an important factor that must be considered. Other aspects that might determine which technique to use are the availability of equipment and the expertise of the palynologist doing the forensic study.

A good quality light microscope with good photographic capabilities, bright and dark field phase contrast, and differential interference contrast is a costly investment and can range in price from \$25 000 to over \$75 000 (United States). However, once it is purchased, it can be used for years without additional costs other than an annual cleaning, checking alignments, and making needed adjustments. The cost of an adequate SEM, the proper setting with stable flooring to prevent vibrations, cooling systems to prevent overheating, and digital photographic systems can easily exceed \$500 000 (United States); an amount many forensic laboratories cannot afford. There are some less costly SEM 'desktop' versions but often these are not adequate for most forensic studies of pollen grains. Since many forensic facilities cannot afford to purchase and maintain SEM equipment, they must rent beam time from some facility that has available SEM equipment and is willing to rent it to others. The rental cost for beam time will vary greatly and must be considered if one wishes to conduct a SEM study of forensic pollen samples. For identification purposes at both the LM and SEM level, one must be very precise in explaining on what basis pollen types were 'lumped' or 'separated' into various taxa. If time is deemed critical, trying to prepare forensic samples for SEM usage and then examining a pollen sample using SEM technology is generally not a viable option. By far, the fastest and least expensive method of forensic pollen analysis remains the use of LM.

One other problem related to SEM analyses is the expertise of the palynologist. Most palynologists can effectively use a wide variety of LM equipment, but few are well versed on the proper use of SEM equipment. The expertise level of the palynologist may determine which examination technique is used regardless of what method, SEM or LM, might be the most desired or available. Past experience has confirmed that hiring a SEM microscopist to examine and identify pollen in forensic samples is inadequate unless that person is also a trained palynologist (Jones and Bryant, 2007). Skilled SEM technicians are not usually familiar with the range of pollen images they might encounter. Pollen grains and spores in a SEM forensic sample might be partly degraded, folded, collapsed, broken, partly obscured by debris, not exposed in a position that would

enable precise identification, or might be confused as being only a piece of nonpolleniferous debris. Even trained palynologists very familiar with the identity of pollen in LM analyses often find it very difficult to identify many pollen types found in a SEM forensic sample. In the SEM study of forensic samples (Jones and Bryant, 2007) the identity was recorded and each pollen was photographed 'only after both of us agreed on the identification' of each piece of debris or pollen grain. In terms of identifications, there are various levels that are sometimes used by those doing either LM or SEM analyses. For example, it was found that at the LM level one must often be 'lumpers' who include all of the pollen in a genus, such as *Tilia* (linden) or all of the pollen in the family Myrtaceae into a single taxon category because one cannot be certain of significant differences that might reflect separate genera or species. However, if the same sample is examined using the added resolution of the SEM, then often clear distinctions can be seen enabling one to separate different types (species) of pollen, such as in the genus *Tilia* or family Myrtaceae.

LM Pollen Counting and Pollen Identification

Most forensic pollen samples contain debris that needs to be removed to enable pollen grains to be observed. Depending on the type of matrix material in samples, one needs to apply various extraction techniques that could include the use of HCl, HF, KOH, acetolysis, and/or heavy liquid separation. Even under the best extraction circumstances there are often bits of debris remaining in samples when all processing techniques are completed. Debris such as small pieces of lignin, tracheids, phytoliths, charcoal, soot, hair, fungal spores, and mycelia, partly decomposed fragments of cellulose, insect chitin, and tiny silica particles will often remain in processed pollen samples. Sometimes these debris fragments provide additional information about a sample's location, such as soot suggesting an urban location, or abundant fungal spores indicating a location with high fungal activity, or diatoms that indicate an aquatic location. In processed samples, however, these bits of remaining debris can be troublesome and a hindrance to speedy and accurate pollen analyses. For some forensic samples, the presence of fungal spores and hyphae can greatly enhance the final interpretation and when possible should routinely be included in forensic studies (Hawksworth and Wiltshire, 2011). Fungal palynomorphs provide another avenue of trace evidence that is often valuable in assessing the similarity or dissimilarity between samples and is especially important in recognizing specific soils and sediments.

Often, the problems of debris in a sample can be overcome for LM studies by using a mounting media that will enable pollen grains and debris to be moved under the cover slip. Mounting forensic pollen samples in some solid mounting medias, such as Permout™ is not advised because it 'freezes' pollen in one orientation, which may not expose vital clues needed for correct identification. Moving palynomorphs once they are sealed in Permout™ is not always impossible, but it is difficult.

Analysis

The final question is whether or not a completed analysis of a forensic pollen sample will be similar using either SEM or LM.

To test this Jones and Bryant (2007) examined four separate 300 grain pollen counts from the same prepared sample. Two of the counts were conducted using only SEM, and two using only LM, thereby providing 600 grain pollen counts of the same sample for each of the two techniques. After completion of the 600 grain pollen counts, additional material mounted on slides or mounted on SEM stubs were scanned for new pollen types not already seen. These additional LM and SEM scans revealed 20 additional pollen types not found during any of the previous counts. In all, 93 separate pollen taxa were identified in the one forensic sample. The LM counts found 56% of the total 93 taxa and the SEM counts found 79%. Neither method found all 93 pollen taxa. The SEM identification percentage was higher mainly because of the higher resolution level enabling identification of multiple species of individual pollen genera or different genera within families, whereas the LM analysis was limited mostly to the family or genus level of identification. However, ANOVA tests applied to both techniques revealed no significant differences between the relative percentages of pollen taxa in each of the two studies (LM vs. SEM). On the other hand, a T-test applied to both sets of data indicated a significant difference between the LM and SEM in terms of the total number of pollen taxa identified during the analyses.

DNA Analysis of Pollen Grains

To date, the identification and classification of pollen associated with forensic materials have relied mostly on LM analysis. Most plant genera contain many species and each species is often best suited to the specialized environments it occupies. Therefore, if pollen identification to the species level could be improved, it should be possible to add precision to assigning a forensic sample to a specific location or a specific crime scene. Under certain conditions, SEM analyses can provide this type of added precision, but can DNA studies do this even better? DNA analysis of haploid sperm cells, plastids, and mitochondria in the cytoplasm of individual pollen grains has already been accomplished (Longhi et al., 2009). Nevertheless, this suggested application to the study of forensic pollen samples is fraught with problems.

Pollen (the male sex cells) contains haploid DNA which is carried to the ovules of the plant where it then completes fertilization. In angiosperm and gymnosperm pollen grains the cytoplasm contains two or three large vegetative cells that enclose one or two generative haploid cells, plastids, and mitochondria, which means that DNA from the plastids and mitochondria is present in multiple copies (Bennett and Parducci, 2006). DNA amplification from the pollen grain's cytoplasmic sources can then be used to identify the plant species. However, two of the biggest problems facing DNA pollen taxonomic identifications are the limited number of plant species for which DNA sequences are now available (Bennett and Leitch, 2010), and the rapid decay of the DNA contents in dispersed pollen grains.

Extraction and Processing

Extracting pollen from matrix material, such as soils, is relatively easy if you want to do a pollen analysis using LM.

Depending on the number of extraction procedures used and how quickly a sample must be analyzed, a critical forensic pollen sample could be completely processed and an examination begun in as little as 1 h. If time is not critical, and depending on the matrix material to be removed, normal pollen extraction of forensic pollen samples includes a number of processing steps that might take anywhere from about 4 h to several days.

Recovery of individual pollen grains from various types of matrix materials and the subsequent DNA extraction and amplification process is much more complicated and more time consuming than collecting and extracting pollen for subsequent LM analysis (Parducci et al., 2005; Suyama et al., 1996). In one report, a team of scientists recovered and then identified individual pollen grains of one species of *Chenopodium* from a surface soil sample (Zhou et al., 2007). In that study, they collected 10 g of a surface soil, washed it through a sieve with sterilized, deionized water to separate pollen from other debris, and then searched only for one size (species) of *Chenopodium* pollen. Fortunately, the pollen species they used was a small pollen grain enabling the team to screen out both larger and smaller particles of debris before their search began. However, if individual pollen grains in a forensic sample included a wide size range of types, then the minimum size openings of any screening device should be no smaller than about 150 μm , which will remove the larger, but not the smaller debris particles. Right away this makes any potential DNA study of pollen types in a forensic sample much more complicated because of the inability to screen out and remove vast amounts of debris that might fall within the same size range of pollen grains. Assuming that one disregards the problems created by using the larger size screen openings, then the liquid portion of sterilized, deionized water passing through the sieve must be centrifuged in a sterile glass or plastic centrifuge tube and kept sealed to prevent atmospheric contamination. To ensure that all pollen in a sample is collected, the centrifuge should run for about 10 min at a speed of 5000 rpm. Even under those rigid circumstances there is a slight danger of losing some pollen that might never sink (Jones and Bryant, 2004). Next, the sediments should be rinsed in sterilized, deionized water and centrifuged again using the same procedure as before. Analysts conducting these types of studies recommend rinsing samples four additional times using sterile, deionized water (Zhou et al., 2007). When this procedure is completed, small drops of the residue can be placed on a sterile microscope slide for observation. The mounting medium should be additional drops of sterile, deionized water and no cover slips should be used.

Using a compound microscope at magnifications ranging from 100 to 400 $\times$, individual pollen grains can be isolated on the slide and extracted using a sterilized glass capillary tube or a sable hair extraction brush; however, attempting this procedure using a dissecting microscope has proven inadequate (Bryant et al., 1991). Each removed pollen grain should then be transferred to another sterilized slide and carefully rinsed four times using sterilized, deionized water. Using a hanging-drop slide for this procedure is ideal. Great care must be taken to remove the rinse water each time and to ensure that the individual collected pollen grain on each slide is not lost. All extraction and observation procedures must be carried out in a

contamination-free area, such as a clean bench, to ensure no additional pollen grains enter the forensic sample being examined. To ensure the proper care of forensic evidence, all forensic laboratories must comply with standards for cleanliness and should make stringent contamination checks a routine practice.

When the rinsing process is complete and most of the rinse water removed from the slide, the pollen grains should be moved to a clean, flat, glass slide and the location of the pollen grain should be marked and noted using a felt-tip marker. Next, a second sterile glass slide should be gently placed on top of the first slide containing the pollen grain and then gently tapped to crush the pollen grain and permit its cytoplasm to be released. The crushed pollen grain, the exine fragments, and released cytoplasm should then be transferred to a sterile, 0.5 ml Eppendorf tube and suspended in deionized, distilled water. Steps can then be followed to isolate the internal transcribed spacer regions, and then amplify them using a nested polymerase chain reaction (PCR) reaction with two pairs of primers (Zhou et al., 2007).

In a previous successful test of individual pollen grains the researchers found that the resulting sequences produced by PCR reactions were not long enough for positive alignment with known DNA sequences in established plant DNA databases. Therefore, the researchers had to take the purified pollen DNA and clone it into a pGEM-T easy vector (Promega) and then transform it into cells of *Escherichia coli* (Zhou et al., 2007).

Equipment, Cost, Time, and Expertise

To conduct DNA analyses of equivalent forensic pollen samples one must have access to special equipment designed for the extraction, amplification, and recovery of DNA fragments needed to make positive identifications. Of even greater concern in most DNA forensic pollen studies is having an available DNA database of plant species against which amplified samples can be matched.

The original Plant DNA C-values Database is a comprehensive catalog listing the nuclear DNA content, or in diploids the genome size, for a small number of land plants and algae. Dr. Michael D. Bennett and Dr. Ilia J. Leitch of the Royal Botanic Gardens, Kew, England, created the original database, which has been continuously updated and now contains C-values for 7058 species. Of these, 6287 species are from angiosperms, 204 gymnosperms, 82 pteridophytes, 232 bryophytes, and 253 algae. By comparison there are about 352 000 angiosperm species, 1050 gymnosperms, 15 000 pteridophytes, and 22 750 bryophytes (Bennett and Leitch, 2010). Therefore, the current known DNA database for plants is very limited and contains only 0.18% of the total, which may not include critical DNA information about many pollen species isolated from some forensic samples. An added problem is that among the known plant types with DNA profiles, the pollen of some species can hybridize easily with other species of the same genus producing an F1 generation with DNA that is not identical to either parent (Schaal et al., 1998; Valbuena-Carabana et al., 2005).

Pollen Counting and Identification

Pollen analysis (quantification of taxa and rapid identification) for a typical forensic sample can often be accomplished

fairly quickly. For DNA pollen studies, similar levels of precision and rapid analysis of individual samples are impossible to obtain using current technology. The procedures needed to recover each pollen grain, prepare each pollen grain for DNA analysis, and then attempt to amplify the recovered DNA fragments so they can be matched against an available plant DNA database are far too time consuming and costly to even be considered for most forensic work. What is even more problematic is that pollen grains, which have lost their cytoplasm for a variety of reasons, will not produce any DNA fragments needed for amplification (Bennett and Parducci, 2006). This means that one might have to isolate a large number of individual grains from a single forensic pollen sample in hopes of finding a few that still contain sufficient cytoplasm for DNA amplification. For those successful few containing DNA, the final problem focuses on the hope that the recovered DNA fragments can then be matched to an available DNA database.

A significant problem of DNA identification studies, as it might be applied to any forensic pollen sample, is the inability to determine the quantitative relationship of each pollen type to all other taxa present. Even if successful, DNA identification studies of any pollen sample could only provide a ubiquity view of the sample. What the resulting DNA data cannot provide is information about how abundant or scarce each pollen type might be in a total sample.

Analysis

The analysis of most forensic pollen samples relies on the identification of individual pollen types, and on the quantitative relationship of each pollen type to all other types encountered in the same sample. A specific forensic sample might contain pine, spruce, grass, and sagebrush pollen grains; but it might be equally important to know whether or not those taxa are represented by a single pollen grain or by a large percentage of the total pollen count within the sample. A single pollen grain from any of those types might reflect nothing more than an unusual case of long distance transport of an airborne pollen grain (Faegri and Iversen, 1989) or possible contamination. However, if a forensic sample contains 5, 10%, or more pollen for any taxon, then that provides significant information about the probable location of the sample or its similarity or dissimilarity to other samples collected from either a crime scene or a possible suspect. A pollen analysis using LM can provide essential quantitative information about each pollen type but a DNA study provides only ubiquity data as to the presence/absence of taxa without any additional quantitative information.

Finally, aside from the other already mentioned limitations created by the data obtained from the DNA of individual pollen grains, the ultimate cost and time needed for these types of studies make them unsuitable for forensic pollen analyses. In one nonforensic related study, which attempted to recover and identify individual pollen grains in a surface soil sample, the researchers found that they could only derive DNA data through amplification of 40% of the total number of pollen grains they isolated and then tested (Zhou et al., 2007). This means that 60% of all the pollen grains that were individually isolated and then tested for DNA amplification proved useless. Any effort where even a small portion of a

sample's pollen is unavailable for analysis does not provide the type of reliability needed for the precision necessary in forensic applications.

Perhaps in the future as new and better techniques are developed for the recovery and amplification of DNA in individual pollen grains, this type of analysis technique might become useful. Nevertheless, even with better techniques of recovery and amplification, DNA studies of pollen grains will still have many limitations because it provides only presence/absence data and it can be used only on pollen types still containing cytoplasm rather than all the pollen present in a forensic sample.

Raman Spectroscopy of Pollen Grains

The concept and use of Raman spectroscopy was named after its discoverer, the Indian scientist Sir C. V. Raman, who observed the Raman Effect of sunlight in 1928 and was recognized for his discovery with a Nobel Prize in Physics in 1930. In Raman spectroscopy, a laser beam is used to excite electron transitions in the molecules of a sample. After excitation, the majority of the electrons return to the ground state by releasing photons of the identical wavelength as that of the laser beam they absorbed. However, a small percentage of the excited electrons in a sample undergo Raman scattering, releasing scattered light of different wavelengths prior to relaxing to the ground state. As a result of the scattering, less energy is emitted in the transition back to ground state than was originally absorbed in the excitation. The difference in these two energies is referred to as the Raman shift. Since various molecular bonds within a molecule undergo Raman shifting of different wavelengths, the unique Raman spectra or patterns of shifts, as a function of wavelength, can be used to identify the molecular composition of samples. Because pollen from different taxa contain different molecular compositions, Raman spectroscopy should prove beneficial in separating various pollen types by family, genus, and even species (Atanu et al., 2005).

Raman spectroscopy has gained a great deal of attention for its use in forensics as an analytical tool. There are a number of advantages of Raman spectroscopy, which makes it ideal for applications in the field of forensics. First, the field of Raman spectroscopy is rapidly changing as improved techniques broaden its potential uses. Second, Raman spectroscopy is a noncontact and nondestructive technique, which means that many types of items and samples of forensic trace evidence do not have to be destroyed or changed in order to enable analysis. Third, Raman spectroscopy does not require any special type of sample preparation prior to examination and analysis. This means that trace evidence can often be examined first using Raman spectroscopy, and then used for other types of analyses, which might need to alter or destroy the material during analysis (such as pollen extraction techniques). Fourth, only minimal amounts of material are required for a complete analysis; sometimes even microscopic amounts of trace evidence can be examined. Fifth, Raman spectroscopy is highly sensitive to slight differences in the molecular and chemical composition of materials, which enables it to detect very small differences in samples. That type of high resolution potential can be essential in the study of forensic samples where minor

differences are essential to distinguish one type of paint from all others, one type of glass from other types, or linking some synthetic fiber to a unique production time at only one manufacturing facility.

Extraction and Processing

One problem in using this technique on pollen grains is that the spectral signatures of carotenoid pigments embedded in the surface and exine of pollen grains can provide signals leading to the misclassification of pollen grains; therefore, those carotenoid signatures should be eliminated prior to using Raman spectroscopy. A common method of removing the carotenoid signals is to photo-destroy the carotenoids using a 633 nm laser light prior to attempting the Raman spectral analysis.

Schulte et al. (2009) suggest that signatures of the carotenoid pigments might provide valuable information that could be used as an additional method to distinguish one pollen species from another. They suggest that it is important to first take a Raman spectrum signature of a pollen grain, then photo-destroy the carotenoids, then take a second Raman spectrum signature. By using the comparisons of the two spectra it would be possible to isolate and catalog the different spectral features of carotenoid pigments in each pollen type examined.

Equipment, Cost, Time, and Expertise

The cost for equipment to conduct Raman spectroscopy varies depending on the type of desired studies. There are three basic types of Raman instruments: bench top spectrometers, fiber optic spectrometers, and Raman microscopes. For pollen studies one would need a Raman microscope.

The time needed to conduct Raman spectroscopy analyses depends on the level of resolution needed, whether or not the carotenoid pigments in the pollen wall and surface need to be removed, and how many pollen grains in a sample need to be examined. Another time-consuming requirement would be determining each of the precise X and Y coordinates (locations of the pollen grains within the matrix materials) in each sample where the Raman laser beam and analysis must be focused. Once pollen data are collected using Raman spectroscopy, then those data must be compared against a previously recorded pollen database of the spectral parameters for each pollen taxon. This becomes a major problem because at present there are very few pollen taxa for which any type of Raman spectral data are known.

Conducting and interpreting the signatures of Raman spectroscopy require a special expertise, which most palynologists may not have. Therefore, these types of analyses would require either a palynologist to work closely with someone familiar with Raman spectroscopy, or it would require the palynologist to learn how to use and interpret the results of Raman spectra produced by Raman spectroscopy.

Pollen Counting and Identification

When trying to identify pollen grains to species using Raman spectroscopy, an important question is whether individual variations in pollen grains reflect different species of the same

genus, or are there expected differences found within pollen signatures of the same species. Ivleva et al. (2005) found that to answer that question required them to examine high resolution spectral maps produced from five to ten individual pollen grains per species. They reported that developing a single high-resolution spectral map of a single pollen grain requires about 72 h. To obtain the five to ten high-resolution spectral maps needed per pollen species, a researcher would need to spend an estimated 360–720 h gathering data for each pollen species being considered. Once completed, the five to ten spectral maps for a pollen species can then be averaged to provide the expected spectral variations expected within the Raman signatures of one pollen species. However, before those individual pollen grain signatures can be obtained or recorded, a researcher must first destroy the spectra produced by the carotenoid pigments in that pollen grain species by using a pretreatment of 633 nm laser light. Failure to do this would obscure the true Raman spectral signatures of each individual pollen species.

Another problem with Raman spectroscopy is linking spectral signatures to individual pollen species. A major problem can be created because of the spheroid, triangular, or oblong shape of most pollen grains, because of the complex variety of surface ornamentations, and because of aperture variation and numbers. These types of variations create a heterogeneous pollen surface area with slightly different molecular signatures depending on where on the pollen grain the Raman laser beam is focused. This means that the focal position and the resulting spectral information of the excitation laser can vary depending on where on the pollen grain surface and which potential ornamentation feature is being examined. Additional problems are created by the pollen exine, which can vary in thickness depending on whether there is an aperture, fissure, or operculum. Some pollen grains have a thin exine while others have a thick or a multilayered exine. Different spectral signatures will be recorded depending on which exine surface is being examined and how deep into the exine the laser beam is focused.

Analysis

The most important question is whether or not Raman spectroscopy can be effectively used in forensic samples to determine the precise pollen species present. Ivleva et al. (2005) note that for many types of pollen the precise species and genus level identification should be possible using Raman spectroscopy, provided one is able to generate the needed signature patterns for each potential pollen type that is expected to occur in a forensic sample. If it requires 360–720 h to gather the data needed to create a reference for each pollen species, and if two or three dozen potential pollen types might be expected in a specific forensic sample, it would take weeks of preparation time obtaining the needed signatures before one could even begin the study of the forensic sample. Furthermore, any unknown signature might reflect some pollen type, which was not expected and therefore, would have no comparative reference signature. One advantage is that often pollen grains belonging to different species of the same genus will have spectral signatures that are similar enough to place all species into the same genus. Likewise, the spectral signatures of

many genera within the same plant family also have signatures that are similar enough to identify them with that family (Sengupta et al., 2005). In both types of spectral signatures, some studies suggest the patterns are similar enough to recognize they are related, but different enough so that individual species and genera can still be identified provided sufficient reference signatures were available for comparison. Yet another problem is that preliminary studies have noted that in some plant taxa there are essentially no differences among similar species in a single genus. Among the oaks, for example, researchers have been unable to identify spectral signatures that are unique to only one species in the genus *Quercus*. This inability is believed to be caused by the frequent hybridization among different species of oaks (Valbuena-Carabana et al., 2005). Although not yet tested for a number of other plant genera, similar types of hybridization are known to occur in many plants and thus, it could present similar problems of signature resolution.

Although a few studies have shown that successful spectral signatures can be used to identify certain types of pollen using Raman spectroscopy (Manoharan et al., 1991), the use of this technology is currently hindered by a number of problems. First, very few pollen grain spectral signatures are available for comparison use. Second, establishing more pollen spectral signatures requires a very large investment of time and effort for each taxon. Therefore, the question remains as to what organization could provide the funding needed to conduct even a single forensic pollen study using this type of technology. Third, similar to the problems of using DNA studies of individual pollen grains in forensic samples, current techniques using Raman spectroscopy do not offer quantitative information about the abundance or rarity of each pollen type in a sample. Raman spectroscopy, similar to DNA, provides only ubiquity information rather than the quantitative data related to each pollen taxon; which is usually considered essential in most forensic pollen studies. Finally, in a study conducted by the author with a few colleagues it was experimented to see if Raman spectroscopy could be used to distinguish the differences between fresh *Celtis* (hackberry) and fresh *Cannabis* (hemp, marijuana) pollen. After mounting the grains from both types it was noted that some of the pollen grains in each group were folded, deflated, broken, or overlapped with other grains or with debris. When setting the laser beam on specific targets within each group it was found that the resulting data was not conclusive enough to provide with certainty the identity of the pollen taxon being examined. In other words, the ranges of variations were too great and somewhat overlapping between both groups. It was suspected that the problem may have resulted from the laser beam reading the surfaces on some grains, the interior of the exine on other grains, the aperture on other grains, and the surface morphology on still others. It is possible that perhaps some of the returning signatures came from a focus in the interior of some grains that may have been folded.

Stable Isotope Analysis of Pollen Grains

The application of stable isotope analysis to various materials has been growing in importance since the early 1970s and its

use has expanded widely. Today, stable isotope analysis is a useful technique applied to a number of disciplines including forensics, archaeology, paleoenvironmental studies, and questions related to geolocation. The principle behind isotope ratios is that organic and inorganic materials retain a record in their molecules of information generated from their immediate surroundings (Hurley et al., 2010; West et al., 2006).

Thousands or millions of years later, even if the original materials move or are buried, those initial environmental isotope signals can be used to identify the environment and the general geographical location where an item originated. Other types of important data generated from stable isotope analyses focus on direct inferences regarding human and other animal diets, nutritional levels, and subsistence patterns (Bousman and Quigg, 2006). In archaeology, for example, isotopic signatures are used to gain insights into prehistoric diets, human migration patterns, trade routes, class status in burials, and the environmental conditions that confronted ancient cultures (Calo and Cortés, 2009).

Although there have been many studies of the stable isotopes found in plant leaves, wood, stems, and roots, it has only been during the last decade that scientists have begun to focus on the use of isotopes recoverable from types of cellulose, lignin, *n*-alkanes, *n*-alkenoic acids, phenylglucosazone, and other materials, including individual fossil pollen grains. The resulting data serve as proxy information for determining paleoenvironmental conditions and climatic variations (Jahren, 2004). Because stable isotopes of pollen grains from different genera and species of plants also carry different signatures, isotope analyses have the potential for providing additional information helpful in identifying pollen types (Descolas-Gros and Schölzel, 2007). Under the right conditions where information from some sample might be deemed critical, isotope studies of pollen grains might provide a complementary technique that could be combined with standard pollen analyses in a forensic study. Nevertheless, trying to conduct such a study would be costly, time consuming, and complicated.

Extraction and Processing

An initial problem in conducting stable isotope studies of pollen is to isolate the pollen grains from the surrounding matrix. Under many circumstances, one might consider using a pollen extraction technique such as acetolysis to remove unwanted organic debris and attempt to concentrate the pollen grains (Faegri and Iversen, 1989). However, as Amundson et al. (1997) and also Loader and Hemming (2000, 2001) point out, the use of the acetolysis treatment will alter the carbon isotope composition in the pollen wall thereby providing inaccurate signals. Other extraction techniques, which have been suggested for use in concentrating pollen for isotope analysis, include the use of Schulze's solution and/or the use of concentrated sulfuric acid. Jahren (2004) found that the sulfuric acid extraction technique has less of an effect on the composition of the pollen wall and appears to do less damage to the isotope composition ratios. In his study, he found that the sulfuric acid extraction technique altered the carbon isotope composition ratios of grass pollen by 2.5%, whereas using the acetolysis treatment altered the composition of the same pollen type by

as much as 5%. However, both techniques significantly alter the original carbon isotope ratios of pollen grains and thus neither is suitable.

A major problem conducting isotope studies of individual pollen grains in various types of samples is that the use of almost any chemical treatment designed to concentrate the pollen and remove nonpollen debris in samples will alter the molecular composition of the pollen wall. Concentrating pollen in samples, removing debris, and then finding individual pollen grains of each specific taxon is very labor intensive and expensive. Jahren (2004) reports that it was necessary to manually isolate up to 2500 individual pollen grains of a single species in order to recover sufficient pollen for using automated combustion methods to identify the isotopes of that pollen species. However, Loader and Hemming (2004) state they achieved these same results for a $\delta^{13}\text{C}$ analysis with as few as 350 spruce pollen grains in one sample. Even at that reduced level, these techniques are very labor intensive and could never be used for a forensic sample that might contain small numbers of pollen grains.

Recently, Nelson et al. (2008) reported that it was no longer necessary to isolate hundreds or thousands of pollen grains to conduct stable isotope analyses of pollen. They demonstrated that they can analyze the stable isotopes of individual pollen grains using a spooling-wire microcombustion device interfaced with an isotope-ratio mass spectrometer. Using this technique, they isolated pollen grains that were initially obtained from a small deposit of sediment. First, they used nano-pure water to produce a thin slurry of the sediment that was then transferred to clean microscope slides with additional nano-pure water. The slurry was gently washed on the slide with additional nano-pure water until individual pollen grains could be located at $200\times$. Individually, each grain was identified, isolated, and then carefully removed and analyzed for $\delta^{13}\text{C}$ (Nelson et al., 2008).

Equipment, Cost, Time, and Expertise

Palynologists are usually not trained in stable isotope analysis, nor do most palynology laboratories have the equipment needed to conduct isotope analyses. The role the palynologist might be able to play in these studies would be to identify and collect samples from forensic sources, use appropriate techniques to isolate, concentrate, identify, and then remove individual pollen grains for analysis by those trained in isotope analyses.

Pollen Counting and Identification

The pollen collecting and identification would need to be done by a trained palynologist. The number of taxa to examine and the number of pollen grains to collect would depend on time, funding, and the expected goal of the project. Even in circumstances where this type of isotope study was attempted, the following analysis and interpretation phases would be even more complicated.

Analysis

During photosynthesis, the carbon isotope intake and later the formation of plant tissues, including the pollen wall, will be

influenced by the environment. Oxygen and hydrogen atoms from leaf water are incorporated into the initial products of photosynthesis and then form a component of the oxygen and hydrogen of other plant compounds. Therefore, studies of stable isotope ratios in plant material have the potential to reveal information about the parent plant's geolocation and environment. During plant formation and growth $\delta^{13}\text{C}$ ratios record the environmental benefits or restraints on photosynthesis, nitrogen isotope ratios ($\delta^{15}\text{N}$) record the nitrogen fixation, and the isotope ratios of hydrogen ($\delta^2\text{H}$) and oxygen ($\delta^{18}\text{O}$) record precipitation and water availability (West et al., 2006). Because of the variety of stable isotope ratios found in plant materials, including pollen, these materials have the potential to reveal important information about the parent plant's location and environment, which might prove especially useful in trying to find the origin of seized illegal drugs (West et al., 2009a,b).

Loader and Hemming (2004) examined the oxygen ($\delta^{18}\text{O}$) and hydrogen (δD) isotope ratios in pollen grains and suggested that these ratios could be used to predict precipitation availability at the time of pollen grains were formed. They also found, during an earlier study, that there was a positive relationship between $\delta^{13}\text{C}$ signatures in the exine of pollen and the climatic temperature of the environment when the pollen-producing plant lived (Loader and Hemming, 2001). As these two researchers found, there seems to be a close correlation between the individual stable isotope ratios found within a single plant's leaves, wood, stems, and roots. They also suggest that this may be true for the exine of the pollen a plant produces. If further testing reveals these aspects to be true, then it means that many of the stable isotope ratios already developed for other tissues in many plant species can also be used as a database for isotope ratios in pollen (Jahren, 2004).

Nelson et al. (2008) examined isotopic ratios from grass pollen collected from a series of lakes to determine if lake-deposited pollen reflected the surrounding grassland environment. Their study focused on two important aspects. First, they wanted to determine if individual grass pollen grains, deposited in nearby lake sediments, could serve as a proxy for the host grass communities (either C_3 or C_4) from which they originated. To test this, they collected surface sediments from the bottom of ten different lakes surrounded by areas dominated by either C_3 or C_4 grass communities. Next, they tested individual pollen grains collected from the top sediment layers at the bottom of the ten lakes against fresh grass pollen grains collected from grasses growing in areas around each of the lakes. Their testing confirmed a strong $\delta^{13}\text{C}$ signature correlation between the individual grass pollen grains deposited in the lake sediments and the surrounding C_3 or C_4 grass community. Second, they confirmed that the examination of individual grass pollen grains provided distinct advantages over studies of the isotopes from bulk sediment samples collected at the bottom of the same lakes. They found that the bulk samples most frequently reflected the aquatic plant habitat rather than the surrounding terrestrial grasses.

The future use of information derived from stable isotope studies for pollen grains associated with forensic applications is questionable because there are a number of problems and limitations. First, more work is needed and confirmation of the relationship between the various environmental constraints

and the stable isotope ratios in pollen grains is essential. Second, one needs to determine if individual pollen genera and even species can be recognized by some unique combination of isotope ratios in the pollen walls, or will continued studies show that isotopes of pollen grains are important only because they offer clues about the environmental surroundings and perhaps their geographical origins. Third, it is already known that the time and especially the costs of trying to isolate individual pollen grains in a forensic sample and then carrying out the isotope studies on each pollen grain can become very expensive. Fourth, in any given forensic sample all of the pollen may not have come from plants in only one location; therefore, the isotope signatures of a suite of pollen in a single sample might generate a variety of interpretations with no verification as to which ones might be the most important. Fifth, similar to some of the other new analytical techniques mentioned, isotope studies are potentially limited to providing ubiquity rather than quantitative studies of individual samples.

Summary

Forensic studies using pollen as trace evidence are in their infancy in terms of usage and acceptance worldwide. In some countries, the use of forensic palynology has become routinely accepted and its use has become court tested while in other countries neither is occurring at present. The United Kingdom is the recognized world leader in the use of forensic palynology yet prior to the 1990s it was virtually unknown in that country. It was not until Wiltshire (2003) began her extensive forensic pollen work and encouraged law enforcement agencies to consider pollen studies in a wide range of criminal cases that it became a routine type of forensic trace evidence in the United Kingdom. In New Zealand, the early work by Mildenhall (1990) issued in the use of this technique in that country. In Australia, the application of pollen studies, as it relates to forensic cases, was little recognized and appreciated until Bruce and Dettmann (1996) published an article pointing out the potentials for its use in that country. Later, Milne (2005) conducted what may be the earliest application of the use of forensic palynology in Australia. In Canada, Mathewes (2006) reported that aside from a single case where pollen evidence was a factor in a 1987–97 era civil lawsuit between the national government and Native American Canadians, there has been little use of this technique in Canada. Similar to Canada, there has been little effort or wide spread appreciation for the use of pollen as trace evidence in forensic applications within the United States (Bryant and Jones, 2006).

Since the beginning of the twenty-first century, the world has become a more dangerous place to live. Today, rapid travel and instant communication across continents, coupled with the worldwide rise in local and regional conflicts as well as crimes associated with the illegal manufacture and use of drugs have created many unsafe regions for travel. Piracy on the high seas combined with acts of international terrorism has placed a high premium on the safety of military and civilian personnel. One effective way to deal with these growing problems is to find new and more effective ways to outwit criminals and catch terrorists before they can commit acts of mass destruction, such as the 2001 bombing of the World Trade Center in New York

City or the 2005 bombings that occurred in the London Underground and also in one of its double-decker buses.

Although there is no 'magic bullet' that will solve all of these problems, the wider use of forensic palynology might help identify and prevent the wider spread of criminal activity and terrorism. The analytical techniques discussed in this article need more development and refinements before they will be of much value as tools in forensic palynology. Nevertheless, with possible future technological developments in examining pollen using DNA, Raman spectroscopy, and stable isotope analysis, and with the hope for a reduction in costs, these techniques may play a wider role in forensic palynology. Likewise, the increased applications of SEM and TEM may be feasible in the future provided costs and time can be reduced. For the present, however, the most reliable and least expensive analysis technique in forensic palynology remains the use of a light microscope by a highly trained palynologist.

New technology is important, but there remains no shortcut for quick results without the adequate and careful training of competent forensic palynologists. In other areas of forensic sampling, examples have been seen of how doubt and suspicion are easily raised when those charged with examining forensic evidence are found to be guilty of using hasty or improper techniques (Fisher, 2008). Because forensic palynology is still a relatively new technology it has rarely been victimized by those doing incompetent or inaccurate analyses, and hopefully, with international efforts, it might remain that way. What is most needed is wider appreciation for the potential results that this discipline can offer.

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See also: Carbonate Stable Isotopes: Lake Sediments; Non-Lacustrine Terrestrial Studies. Paleobotany: Ancient Plant DNA. Use of Quaternary Proxies in Forensic Science: Soils and Sediment; Use of Geology in Forensic Science: Search to Locate Burials.

References

- Allen GP, Hodgson RM, Flenley JR, et al. (2007) An automated pollen recognition and counting system. *Quaternary International* 167–168: 6; Abstracts.
- Amundson R, Evett R, Jahren A, and Bartoleme J (1997) Stable carbon isotope composition of Poaceae pollen and its potential in palaeovegetational reconstructions. *Review of Palaeobotany and Palynology* 99: 17–24.
- Atanu S, Laucks M, and Davis E (2005) Surface-enhanced Raman spectroscopy of bacteria and pollen. *Applied Spectroscopy* 59(8): 1016–1023.
- Bennett DK and Leitch I (2010) Plant DNA C-values database (Release 5.0, Dec. 2010). <http://www.kew.org/cvalues/>.
- Bennett KD and Parnocci L (2006) DNA from pollen: Principles and potential. *The Holocene* 16: 1031–1034.
- Bousman CB and Quigg M (2006) Stable carbon isotopes from Archaic human remains in the Chihuahuan Desert and Central Texas. *Plains Anthropologist* 51(198): 123–139.
- Brown CA (2008) Riding JB and Warny S (eds.) *Palynological Techniques*, 2nd edn., 137 pp. Dallas: American Association of Stratigraphic Palynologists Foundation.
- Bruce RG and Dettmann ME (1996) Palynological analyses of Australian surface soils and their potential in forensic science. *Forensic Science International* 81: 77–94.
- Bryant VM (2000) Does pollen prove the Shroud authentic? *Biblical Archaeology* 26: 18–30.
- Bryant VM and Jones GD (2006) Forensic palynology: Current status of a rarely used technique in the United States of America. *Forensic Science International* 163: 183–197.
- Bryant VM, Pendleton M, Murry R, Lingren P, and Raulston J (1991) Techniques for studying pollen adhering to nectar feeding corn earworm (*Lepidoptera*: Noctuidae) moths using scanning electron microscopy. *Journal of Economic Entomology* 84(1): 237–240.
- Calo C and Cortés L (2009) A contribution to the study of diet of formative societies in northwestern Argentina: Isotopic and archaeological evidence. *International Journal of Osteoarchaeology* 19: 192–203.
- Descolas-Gros C and Schölzel C (2007) Stable isotope ratios of carbon and nitrogen in pollen grains in order to characterize plant functional groups and photosynthetic pathway types. *New Phytologist* 176: 390–401.
- Erdtman G (1969) *Handbook of Palynology*, p. 486. New York: Hafner Publishing Co.
- Fægri K and Iversen J (1989) *Textbook of Pollen Analysis*, 4th edn., p. 328. Caldwell, NJ: Blackburn Press.
- Fisher J (2008) *Forensics Under Fire: Are Bad Science and Dueling Experts Corrupting Criminal Justice?* p. 324. Newark: Rutgers University Press.
- Hawksworth DL and Wiltshire PEJ (2011) Forensic mycology: The use of fungi in criminal investigations. *Forensic Science International* 206: 1–11.
- Hurley JM, West JB, and Ehleringer JR (2010) Stable isotope models to predict geographic origin and cultivation conditions of marijuana. *Science & Justice* 50: 86–93.
- Ivleva N, Niessner R, and Panne U (2005) Characterization and discrimination of pollen by Raman microscopy. *Analytical and Bioanalytical Chemistry* 381: 261–267.
- Jahren AH (2004) The carbon stable isotope composition of pollen. *Review of Palaeobotany and Palynology* 132: 291–313.
- Jones GD and Bryant VM (2004) The use of ETOH for the dilution of honey. *Grana* 43: 174–182.
- Jones GD and Bryant VM (2007) A comparison of pollen counts: Light versus scanning electron microscopy. *Grana* 46: 20–33.
- Loader NJ and Hemming DL (2000) Preparation of pollen for stable carbon isotope analysis. *Chemical Geology* 165: 339–344.
- Loader NJ and Hemming DL (2001) Spatial variation in pollen $\delta^{13}\text{C}$ correlates with temperature and seasonal development timing. *The Holocene* 11: 587–592.
- Loader NJ and Hemming DL (2004) The stable isotope analysis of pollen as an indicator of terrestrial palaeoenvironmental change: A review of progress and recent developments. *Quaternary Science Reviews* 23: 893–900.
- Longhi S, Cristofori A, Gatto P, Cristofolini F, Grando M, and Gottardini E (2009) Biomolecular identification of allergenic pollen: A new perspective for aerobiological monitoring? *Annals of Allergy, Asthma & Immunology* 103(December): 508–514.
- Manoharan R, Ghiamati E, Britton K, Nelson W, and Sperry J (1991) Resonance Raman spectra of aqueous pollen suspensions with 222.5–242.4-nm pulsed laser excitation. *Applied Spectroscopy* 45: 307–311.
- Mathewes RW (2006) Forensic palynology in Canada: An overview with emphasis on archaeology and anthropology. *Forensic Science International* 163: 198–203.
- Mildenhall DC (1990) Forensic palynology in New Zealand. *Review of Palaeobotany and Palynology* 64: 227–234.
- Milne LA (2005) *A Grain of Truth: How Pollen Brought a Murderer to Justice*, p. 175. Sydney: Reed New Holland.
- Milne LA, Bryant VM, and Mildenhall DC (2005) Forensic palynology. In: Coyle H (ed.) *Forensic Botany: Principles and Applications to Criminal Casework*, pp. 217–252. Boca Raton, FL: CRC Press LLC.
- Moore PD, Webb JA, and Collinson ME (1991) *Pollen Analysis*, 2nd edn., 216 pp. London: Blackwell Science Publication.
- Nelson DM, Hu FS, Scholes DR, Joshi N, and Pearson A (2008) Using SPIRAL (single pollen isotope ratio analysis) to estimate C_3 - and C_4 -grass abundance in the paleorecord. *Earth and Planetary Science Letters* 269: 11–16.
- Parnocci L, Suyama Y, Lascoux M, and Bennett D (2005) Ancient DNA from pollen: A genetic record of population history of Scots pine. *Molecular Ecology* 14: 2873–2882.
- Schaal B, Hayworth D, Olsen L, Rauscher J, and Smith W (1998) Phylogeographic studies in plants: Problems and prospects. *Molecular Ecology* 7: 465–474.
- Schulte F, Mader J, Kroh LW, Panne U, and Kneipp J (2009) Characterization of pollen carotenoids with in situ and high-performance thin-layer chromatography supported resonant Raman spectroscopy. *Analytical Chemistry* 81: 8426–8433.

- Sengupta A, Laucks ML, and Davis EJ (2005) Surface-enhanced Raman spectroscopy of bacteria and pollen. *Applied Spectroscopy* 59(8): 1016–1023.
- Suyama Y, Kawamuro K, Kinoshita I, Yoshimura K, Tsumura Y, and Takahara H (1996) DNA sequence from fossil pollen of *Abies* spp. from Pleistocene peat. *Genes & Genetic Systems* 71: 145–149.
- Valbuena-Carabana M, Gonzalez-Martinez SC, Sork VL, et al. (2005) Gene flow and hybridization in a mixed oak forest (*Quercus pyrenaica* Willd. and *Quercus petraea* Matts. (Liebl.)) in central Spain. *Heredity* 95: 457–465.
- West JB, Bowen GJ, Cerling TE, and Ehleringer JR (2006) Stable isotopes as one of nature's ecological recorders. *Trends in Ecology & Evolution* 21: 408–414.
- West JB, Hurley JM, Dudas FO, and Ehleringer JR (2009a) The stable isotope ratios of marijuana. II. Strontium isotopes relate to geographic origin. *Journal of Forest Science* 54: 1261–1269.
- West JB, Hurley JM, and Ehleringer JR (2009b) The stable isotope ratios of marijuana. I. Carbon and nitrogen stable isotopes describe the growth conditions. *Journal of Forest Science* 54: 84–89.
- Wiltshire PEJ (1993) Environmental profiling and forensic palynology: Background and potential value to the criminal investigator. In: *Handbook for the National Police Improvement Agency*, p. 14–21. London: National Police Improvement Agency.
- Wiltshire PEJ (2003) *Current applications of environmental profiling and forensic palynology in the United Kingdom. Program, Challenges & Changes, 17th International Symposium on the Forensic Sciences*, p. 202. Wellington: The Australian and New Zealand Forensic Science Society.
- Zhou LJ, Pei KQ, Zhou B, and Ma KP (2007) A molecular approach to species identification of Chenopodiaceae pollen grains in surface soil. *American Journal of Botany* 94: 477–481.

U-Series Dating

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Introduction

The known decay rate of natural uranium into a series of daughter isotopes provides a radiometric clock that is one of the very few methods for dating natural materials over the last 600 000 years. U–Th dating of carbonates is one of the most accurate and precise methods in geochronology and is the key chronometer for the Quaternary. U-series geochronology has been applied to the dating of climate records in speleothems (cave deposits), the calibration of the radiocarbon timescale, and the reconstruction of sea level from fossil corals. The dating of fossil corals that once grew near the sea surface is one of the premier methods for reconstructing sea level history. Over the last 30 years, technical advancements in analytical capabilities have led to continued and significant improvements in dating precision (Stirling and Andersen, 2009). In addition, better understanding of U-series systematics in corals is leading to more accurate ages (Thompson, submitted). In principle, any material that contains a sufficient amount of uranium that is not in radioactive equilibrium with its daughters is suitable for dating. U-series dating has been applied to a wide variety of materials including volcanic rocks, marine sediments, bone, teeth, mollusks, cave deposits, reef-building corals, and deep-sea corals. The dating of each of these materials has had its successes and difficulties that are documented in a recent and detailed review of U-series geochemistry (Bourdon et al., 2003). In particular, reef coral and speleothem dating has been of crucial importance for calibrating the radiocarbon timescale, establishing sea level variability, testing orbital theories of climate change, and resolving details of Quaternary climate change over a broad range of timescales.

There are three U-series decay chains that start with the naturally occurring radioactive isotopes ^{238}U , ^{235}U , and ^{232}Th and end with stable isotopes of lead: ^{206}Pb , ^{207}Pb , and ^{208}Pb , respectively. While each decay chain includes a dozen or more isotopes, the most widely used U-series isotopes for Quaternary dating are the first two products in the ^{238}U decay chain: ^{234}U and ^{230}Th . In the ^{235}U decay chain, the ^{235}U and ^{231}Pa parent–daughter isotope pair has also been used in recent years. Both ^{238}U and ^{235}U are long-lived isotopes with half-lives of 4.47 and 0.7 billion years, respectively, while the half-lives of their immediate daughters are much shorter: 245 thousand years (ka) for ^{234}U , 76 ka for ^{230}Th , and 33 ka for ^{231}Pa . This large difference between parent and daughter half-lives has an important impact on these U-series systems. Over Quaternary timescales, the change in the parent's activity, or number of decays per unit time, will be negligible, and the activity of the daughter isotope will change until the number of atoms supplied by the decay of the parent equals the number of atoms removed by decay of the daughter. This condition is known as radioactive or secular equilibrium and is characterized by equal activities of parent and daughter or a parent–daughter activity ratio of 1. Because the change in ^{238}U

or ^{235}U activity is negligible on these timescales, the return of a system to secular equilibrium is controlled by the change in daughter activity and therefore its half-life. Left undisturbed, parent–daughter pairs will attain equilibrium in about six half-lives of the daughter isotope.

A U-series system in radioactive equilibrium cannot be used as a chronometer, but dating rather requires an initial state of disequilibrium. Therefore, U-series dating is sometimes referred to as disequilibrium dating. U-series dating is anchored in a few fundamental principles (Edwards et al., 1986). Many carbonates precipitated from natural waters contain significant amounts of uranium and negligible levels of thorium. This is ideal material for U–Th dating, which is based on the ingrowth of ^{230}Th . Uranium is commonly fractionated from its immediate daughters in natural waters due to their contrasting geochemical properties. In the presence of oxygen, the solubility of uranium in natural waters is fairly high, while the solubilities of Th and Pa are quite low. Furthermore, both Th and Pa have a strong tendency to adsorb onto particles and be removed from solution. Therefore, both abiotic and biogenic carbonates precipitated from natural waters frequently contain significant uranium and negligible thorium and protactinium concentrations, an ideal situation for U-series dating. The dating of corals and speleothems has been quite important for establishing Quaternary chronologies as well as for reconstructing sea level and climate change histories, and the dating of these materials provides an instructive illustration of U-series geochronology.

Dating of Reef Corals

The dating of reef corals has been instrumental in extending the radiocarbon calibration past the limits of tree rings and has been used to reconstruct the changing sea levels of the Quaternary (e.g., Thompson and Goldstein, 2005). The most unambiguous record of sea level comes from fossil reef corals, which grow near the sea surface. In principle, it should be possible to date corals extremely accurately by the decay of natural uranium that is incorporated into the coral skeleton during growth (Edwards et al., 1986). In practice, much U/Th coral dating is complicated by significant open-system behavior of U-series nuclides in corals. The aragonitic skeletons of corals initially contain several ppm of uranium and negligible amounts of Pa and Th and are therefore ideal candidates for U-series dating providing the system remains closed. Unfortunately, the delicate and porous structure of corals leaves them susceptible to postdepositional alteration that can seriously degrade the accuracy of U/Th ages. Gross forms of alteration such as partial dissolution, recrystallization, secondary cements, and significant uranium or thorium loss or gain are readily detected by routine screening procedures. However, a more subtle process commonly affects the ages of fossil corals

that appear to be otherwise pristine. Living corals accurately record the uranium isotopic composition of modern seawater (Robinson et al., 2004). However, it has long been known that uranium isotope ratios of most older fossil corals have more ^{234}U than can be explained by closed-system decay from initial seawater uranium. Furthermore, it has been demonstrated repeatedly that corals with excess ^{234}U have excess ^{230}Th and, therefore, older U-series ages (Bender et al., 1979; Gallup et al., 1994; Thompson et al., 2003). These open-system effects are best documented in corals older than 40 ka that have been uplifted into terrestrial environments by tectonic processes (Figures 1 and 2). It is not entirely clear at this time whether U-series ages of younger corals are similarly affected. However, these corals would be expected to have very subtle shifts in their isotope ratios that might be difficult to detect. Older marine aragonite sediments show patterns of open-system behavior that are very similar to older uplifted corals (Henderson et al., 2001), suggesting that the process disturbing U-series nuclides in corals is not restricted to terrestrial environments.

As it became obvious that many older corals showed significant signs of open-system behavior, dating efforts focused on identifying those few corals that satisfied the closed-system requirements of conventional U/Th dating. The most widely used and sensitive test for open-system behavior is the initial

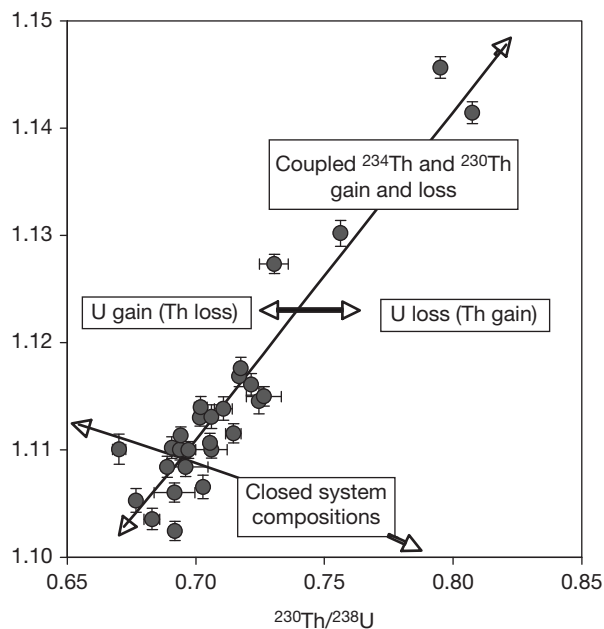


Figure 1 A U/Th activity ratio diagram illustrating the major processes affecting U-series dating of older corals. Closed-system compositions for 95–130 ka corals are expected to lie along the arc near the bottom left of the diagram. The coupled adsorption of ^{234}Th and ^{230}Th produced by radioactive decay of uranium in the reef matrix will produce a linear array of isotopic compositions as indicated. The gain and loss of uranium or thorium will move compositions to the left or right if the isotopic composition of the added uranium is similar to that of the coral. Data are isotopic compositions of corals from the Barbados MIS 5c terrace. Reproduced from Thompson WG (submitted). Why the duration of the Last Interglacial is systematically overestimated. *Earth and Planetary Science Letters*.

$^{234}\text{U}/^{238}\text{U}$ ratio, which can be calculated from a coral's measured $^{234}\text{U}/^{238}\text{U}$ ratio and the U-series age. While criteria vary somewhat among investigators, there is an established consensus that for 'reliable' ages, corals should have initial $^{234}\text{U}/^{238}\text{U}$ ratios within some small range of that expected from the closed-system decay of a modern seawater uranium isotope ratio (e.g., Gallup et al., 1994). However, recent modeling of diagenetic processes suggests that such screening only removes the most egregious age artifacts, without removing a strong and systematic bias toward artificially older ages (Thompson, submitted). Investigators have also tried measuring isotopes of a second U-series chain, $^{231}\text{Pa}/^{235}\text{U}$, as an additional check for closed-system conditions (e.g., Gallup et al., 2002). The agreement of U/Th and U/Pa ages, commonly illustrated on a concordia diagram, is taken as evidence of a closed system. Cheng et al. provide a full discussion of the combined U/Th/Pa system (Cheng et al., 1998). If both initial $^{234}\text{U}/^{238}\text{U}$ and Th/Pa concordance criteria are used, only a few percent of all coral ages measured are considered 'reliable.'

Corrections to U-Series Coral Ages

As the empirical relationship between $^{234}\text{U}/^{238}\text{U}$ and $^{230}\text{Th}/^{238}\text{U}$ or conventional U-series age in older corals became clearer, models were developed to quantify this relationship and correct conventional ages for open-system effects. Although continuing improvements in measurement techniques are desirable and important (Andersen et al., 2004; Potter et al., 2005), it has been clear for some time that the accurate dating of older corals is limited primarily by a lack of understanding of their U-series geochemistry rather than the precision of measurement techniques (Bard et al., 1991; Chen et al., 1991; Hamelin et al., 1991).

The strong positive correlation between excess ^{234}U and older conventional ages in corals was discovered early (Bender et al., 1979), has been confirmed repeatedly (Gallup et al., 1994; Stirling et al., 1998, 2001; Thompson et al., 2003), and has been generally accepted as a screening criteria for almost 20 years (Gallup et al., 1994). This correlation between $^{234}\text{U}/^{238}\text{U}$ and $^{230}\text{Th}/^{238}\text{U}$ ratios for corals from the same terrace demonstrates that corals have experienced the addition of these isotopes in a systematic manner (Gallup et al., 1994). Until very recently, this association of ^{234}U and ^{230}Th seemed to be impossible to explain with any geochemically reasonable scenario, due to the strongly contrasting behavior of uranium and thorium in natural waters. The earliest isotope addition model was based on the empirical correlation of $^{234}\text{U}/^{238}\text{U}$ and $^{230}\text{Th}/^{238}\text{U}$ in corals of approximately the same age (Gallup et al., 1994). A constant addition of ^{234}U and ^{230}Th over time was assumed and used to calculate 'addition lines' that approximated empirical trends. This model was used to establish a criterion for an 'acceptable' range of initial $^{234}\text{U}/^{238}\text{U}$ rather than to correct coral ages for the observed isotope addition.

Shortly thereafter, it was proposed that the isotopic anomalies observed in corals are best explained by the alpha-recoil mobilization during radioactive decay of ^{230}Th and ^{234}Th (which quickly decays to ^{234}U ; Fruijtier et al., 2000). At the same time, it was demonstrated that this process could explain

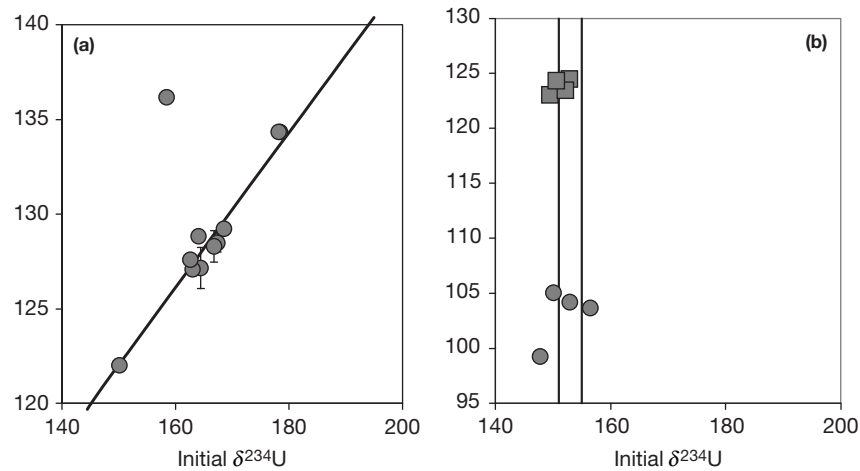


Figure 2 Examples of diagenetic impacts on coral ages. (a) The effects of coupled ^{234}Th and ^{230}Th addition: These 11 subsamples of a large individual coral from the Last Interglacial terrace in Western Australia must all be the same age. However, the ages range from 122 to 136 ka, encompassing the entire range of ages commonly reported for the Last Interglacial. All but one of the samples fall along the theoretically predicted trend for coupled ^{234}Th and ^{230}Th addition, demonstrating that this is the dominant process producing the age artifacts in this coral. The single sample that plots off the addition line has lower ^{238}U and higher ^{232}Th than the others. (b) The effect of uranium addition: Ages and initial $\delta^{234}\text{U}$ for two corals (four subsamples each) from the Last Interglacial terrace in the Yucatan. The vertical lines show the upper limits for 'reliable' (155) and 'strictly reliable' (151) ages according to commonly applied screening criteria. Although these two corals should be of similar ages, and both pass the screening criteria, their age differs by ~ 20 ka. One coral gives a typical Last Interglacial age of 124 ka, while the younger coral yields apparent ages of 99–105 ka. The apparently younger coral has a uranium concentration that is elevated by approximately 25% over living corals of the same species, a clear example of uranium addition. Reproduced from Thompson WG (submitted). Why the duration of the Last Interglacial is systematically overestimated. *Earth and Planetary Science Letters*.

similar anomalies in aragonitic marine sediments, and a set of decay equations was proposed to correct the ages for the added isotopes (Henderson and Slowey, 2000; Henderson et al., 2001). Building on this work, two new types of approaches have been developed to correct U-series coral ages for open-system isotope addition: modified decay equations and isochron techniques. Isochron approaches (Scholz et al., 2004) rely on the observed empirical relationship between initial $^{234}\text{U}/^{238}\text{U}$ and age. While this technique may be useful in some situations, it relies on the fundamental assumption that all corals used to construct the isochron are the same age. This assumption is often difficult to test. Furthermore, it gives an average age for a suite of samples rather than individual ages for each coral. Modified decay equations (Thompson et al., 2003; Villemant and Feuillet, 2003) assume that isotope addition is related to radioactive decay through alpha-recoil processes or geochemical contrasts between parent and daughter isotopes. The more comprehensive approach (Villemant and Feuillet, 2003) also included the possibility of added initial ^{230}Th and ^{232}Th , but these particular equations incorporated an earlier erroneous assumption (Henderson and Slowey, 2000) that the $^{230}\text{Th}/^{234}\text{U}$ addition ratio was related to the $^{234}\text{U}/^{238}\text{U}$ ratio of the coral that had experienced isotope addition and was found to give erroneous results (Andersen et al., 2008; Frank et al., 2006).

Correcting these mathematical and conceptual problems, a similar set of decay equations was developed to correct coral ages for isotope addition (Thompson et al., 2003). The fundamental concept behind this approach is simple: very small amounts of ^{234}Th and ^{230}Th produced by radioactive decay are mobilized by alpha recoil from the carbonate matrix surrounding corals and are added over time to the measured coral.

This coupled addition of two thorium isotopes, with identical chemical properties, explains the observed correlation of ^{234}U and ^{230}Th (and conventional age) in a geochemically consistent way for the first time. The $^{234}\text{Th}/^{230}\text{Th}$ addition ratio at any given time is determined by the $^{234}\text{U}/^{238}\text{U}$ activity ratio of the source of the added ^{234}Th and ^{230}Th because the amount of ^{230}Th and ^{234}Th mobilized is proportional to the number of decays of ^{234}U and ^{238}U (respectively) occurring there. Therefore, this addition process can be described by a modified set of decay equations (Thompson et al., 2003). The $^{234}\text{U}/^{238}\text{U}$ ratio of modern seawater and the fraction of decays mobilized are constraints for the equations that are derived from high-precision measurements and first principles of physics, respectively. The predicted $^{230}\text{Th}/^{234}\text{Th}$ addition ratios arise directly from the hypothesis that these isotopes are mobilized by radioactive decay and are independent of the empirical trends and not 'tuned' to them in any way. Therefore, the excellent match of the theoretically predicted and empirically observed addition ratios (Figure 1) provides strong validation of the equation's underlying principles. These correction equations have been shown to produce robust and reliable results (Andersen et al., 2008, 2009, 2010; Frank et al., 2006; Stirling and Andersen, 2009; Thompson, submitted; Thompson and Goldstein, 2005, 2006; Thompson et al., 2011).

These results have important implications for conventional U-series coral dating. It is quite clear that the precision of uncorrected conventional U-series ages depends directly on the range of initial $^{234}\text{U}/^{238}\text{U}$ deemed 'acceptable' by the investigator. For example, a commonly quoted criterion of $\pm 0.8\%$ from modern seawater for acceptable initial coral $^{234}\text{U}/^{238}\text{U}$ imposes a ± 3200 -year uncertainty on the reported ages. An additional uncertainty of approximately 800 years

arises from uncertainty in the calculated initial $^{234}\text{U}/^{238}\text{U}$ ratio. Furthermore, very small amounts of U gain or loss can change ages significantly without changing the initial $^{234}\text{U}/^{238}\text{U}$ ratio, further reducing the efficacy of the screening criteria. Screened ages retain a strong bias toward artificially older ages and are not as accurate as the label 'reliable' suggests (Thompson, submitted). The application of this screening criterion results in the rejection of greater than 90% of coral ages measured. Therefore, the use of uncorrected conventional ages in sea level reconstructions degrades both their accuracy and resolution (Thompson and Goldstein, 2005). Furthermore, $^{231}\text{Pa}/^{235}\text{U}$ dating does not provide an independent assessment of the accuracy of U/Th ages. Both the $^{230}\text{Th}/^{238}\text{U}$ and $^{231}\text{Pa}/^{235}\text{U}$ systems will be similarly affected by alpha-recoil mobilization of daughter isotopes and strong contrasts in geochemical behavior between parent and daughter isotopes. In the combined Pa/Th system, this process moves isotope ratios along the $^{231}\text{Pa}/^{230}\text{Th}$ concordia so that the U/Th and $^{231}\text{Pa}/^{235}\text{U}$ ages may be in reasonable agreement even though they are both incorrect. Therefore, this concordance test may fail to reject samples that have behaved as an open system. Corals with concordant Pa/Th ages but clearly disturbed initial $^{234}\text{U}/^{238}\text{U}$ (Cutler et al., 2003) support this theoretical prediction. Furthermore, corals that have experienced small amounts of open-system behavior may have $^{231}\text{Pa}/^{235}\text{U}$ and U/Th ages that disagree significantly simply because the geochemistry of Pa and Th is not identical and these isotopes are fractionated, causing rejection of fairly reliable ages. Corals with seawater initial $^{234}\text{U}/^{238}\text{U}$ and different U/Pa and U/Th ages suggest that this is indeed the case (Cutler et al., 2003). Thus, it appears that initial coral $^{234}\text{U}/^{238}\text{U}$ ratios provide a more reliable test for closed-system conditions than $^{231}\text{Pa}/^{235}\text{U}$ ages.

The advantage of age correction equations is the ability to calculate ages for a large suite of samples, regardless of the degree of ^{234}U and ^{230}Th addition. The reproducibility of ages from a single stratigraphic unit can then provide a direct assessment of the uncertainties involved in the correction procedure, and replicate measurement can also reduce age artifacts due to U gain and loss (Thompson et al., 2011). Correcting coral ages for the isotopic addition of ^{234}Th and ^{230}Th provides a quantitative way of solving the isotope addition problem that is a significant improvement over using initial $^{234}\text{U}/^{238}\text{U}$ as screening criteria for uncorrected conventional ages.

Dating of Speleothems

Speleothems are carbonate cave deposits that form as the result of precipitation from flowing or dripping groundwater. Both evaporation and CO_2 degassing are important processes in the deposition of these carbonate minerals. During their growth, speleothems incorporate a complex suite of isotopes and trace elements that record environmental conditions. Although there are many forms of cave deposits, records from stalagmites and flowstones appear to be the most widely used. Because they precipitate from natural waters where the solubility of uranium is high and the solubility of thorium low, speleothems often have significant uranium concentrations and nearly negligible amounts of thorium. This high U/Th ratio makes these environmental archives attractive candidates for

U-series dating. They have been used in a number of ways to infer past environmental conditions (e.g., Bar-Matthews et al., 1999), and U-series dating of such records provides a radiometric timescale that yields important insights into the timing and mechanisms of global climate change (e.g., Wang et al., 2001). Dating of speleothems has also been used to assess sea level change, landscape evolution, earthquake history, and the age of archaeological remains. The best candidates for U-series dating are composed of dense, columnar, and unaltered calcite. Speleothems appear to be much less affected, if affected at all, by the postdepositional isotope addition processes that complicate U/Th coral dating. This is very likely the result of their dense and nonporous structure. However, the U-series dating of speleothems has its own set of challenges. The resolution of speleothem timescales is ultimately limited by the growth rate of the speleothem, its uranium content, and the number of dates obtained. While continuous growth is commonly assumed between dated portions, it is not clear that such an assumption is warranted because hiatuses between dates may escape detection. At some level of resolution, hiatuses that are not visible will not be detectable because a sample for dating that spans such a hiatus will give an intermediate age, falsely supporting the appearance of continuity. Although the ideal candidates for U-series dating have relatively high uranium content (several ppm) and little or no initial ^{230}Th , significant quantities of initial ^{230}Th are often present due to incorporation of detrital material or a 'hydrogenous' component of dissolved or colloiddally scavenged thorium. Initial ^{230}Th can be assessed in most cases by measuring the common isotope of thorium ^{232}Th . Corrections for initial ^{230}Th are often made by assuming a fixed initial $^{230}\text{Th}/^{232}\text{Th}$ ratio. Alternatively, isochron methods may be applied using subsamples along a growth band. Insight into the initial incorporation of ^{230}Th and ^{231}Pa may also be obtained by measurements on recently precipitated cave calcite. Richards and Dorale (2003) gave a detailed and recent review of U-series speleothem dating.

See also: [Cosmogenic Nuclide Dating: Exposure Geochronology](#).
Sea Level Studies: [Coral Records of Relative Sea-Level Changes](#);
[Eustatic Sea-Level Changes – Glacial–Interglacial Cycles](#).
Sea-Levels, Late Quaternary: [Tectonics and Relative Sea-Level Change](#).

References

- Andersen MB, Gallup CD, Scholz D, Stirling CH, and Thompson WG (2009) U-series dating of fossil coral reefs: Consensus and controversy. *PAGES News* 17(2): 54–56.
- Andersen MB, Stirling CH, Potter E-K, and Halliday AN (2004) Toward epsilon levels of measurement precision on $^{234}\text{U}/^{238}\text{U}$ by using MC-ICPMS. *International Journal of Mass Spectrometry* 237(2–3): 107–118.
- Andersen MB, Stirling CH, Potter E-K, et al. (2008) High-precision U-series measurements of more than 500,000 year old fossil corals. *Earth and Planetary Science Letters* 265: 229–245.
- Andersen M, Stirling C, Potter E-K, et al. (2010) The timing of sea level high-stands during Marine Isotope Stages 7.5 and 9: Constraints from the uranium-series dating of fossil corals from Henderson Island. *Geochimica et Cosmochimica Acta* 74: 3598–3620.

- Bard E, Fairbanks RG, Hamelin B, Zindler A, and Hoang CT (1991) Uranium-234 anomalies in corals older than 150,000 years. *Geochimica et Cosmochimica Acta* 55: 2385–2390.
- Bar-Matthews M, Ayalon A, Kaufman A, and Wasserburg GJ (1999) The Eastern Mediterranean paleoclimate as a reflection of regional events: Soreq cave, Israel. *Earth and Planetary Science Letters* 166(1–2): 85–95.
- Bender ML, Fairbanks RG, Taylor FW, Matthews RK, Goddard JG, and Broecker WS (1979) Uranium-series dating of the Pleistocene reef tracts of Barbados, West Indies. *Bulletin of the Geological Society of America* 90(6): 577–594.
- Bourdon B, Henderson GM, Lundstrom CC, and Turner SP (2003) *Uranium Series Geochemistry*. Washington, DC: Mineralogical Society of America.
- Chen JH, Curran HA, White B, and Wasserburg GJ (1991) Precise chronology of the last interglacial period: ^{234}U – ^{230}Th data from fossil coral reefs in the Bahamas. *Bulletin of the Geological Society of America* 103: 82–97.
- Cheng H, Edwards RL, Murrell MT, and Benjamin TM (1998) Uranium–thorium–protactinium dating systematics. *Geochimica et Cosmochimica Acta* 62(21/22): 3437–3452.
- Cutler KB, Edwards RL, Taylor FW, et al. (2003) Rapid sea-level fall and deep-ocean temperature change since the last interglacial period. *Earth and Planetary Science Letters* 206: 253–271.
- Edwards RL, Chen JH, and Wasserburg GJ (1986) ^{238}U – ^{234}U – ^{230}Th – ^{232}Th systematics and the precise measurement of time over the past 500,000 years. *Earth and Planetary Science Letters* 81: 175–192.
- Frank N, Turpin L, Caboich G, et al. (2006) Open system U series ages of corals from a subsiding reef in New Caledonia: Implications for sea level change, and subsidence rate. *Earth and Planetary Science Letters* 249(3–4): 274–289.
- Fruhgtier C, Elliot T, and Schlager W (2000) Mass-spectrometric ^{234}U – ^{230}Th ages from the Key Largo Formation, Florida Keys, United States: Constraints on diagenetic age disturbance. *GSA Bulletin* 112(2): 267–277.
- Gallup CD, Cheng H, Taylor FW, and Edwards RL (2002) Direct determination of the timing of sea level change during Termination II. *Science* 295: 310–313.
- Gallup CD, Edwards RL, and Johnson RG (1994) The timing of high sea levels over the past 200,000 years. *Science* 263: 796–800.
- Hamelin B, Bard E, Zindler A, and Fairbanks RG (1991) $^{234}\text{U}/^{238}\text{U}$ mass spectrometry of corals: How accurate is the U–Th age of the last interglacial period? *Earth and Planetary Science Letters* 106: 169–180.
- Henderson GM and Slowey NC (2000) Evidence from U–Th dating against Northern Hemisphere forcing of the penultimate deglaciation. *Nature* 404: 61–66.
- Henderson GM, Slowey NC, and Fleisher MQ (2001) U–Th dating of carbonate platform and slope sediments. *Geochimica et Cosmochimica Acta* 65(16): 2757–2770.
- Potter E-K, Stirling CH, Andersen MB, and Halliday AN (2005) High precision Faraday collector MC-ICPMS thorium isotope ratio determination. *International Journal of Mass Spectrometry* 247(1–3): 10–17.
- Richards DA and Dorale JA (2003) Uranium-series chronology and environmental applications of speleothems. *Uranium-Series Geochemistry*, vol. 52, pp. 407–460. Washington, DC: Mineralogical Society of America.
- Robinson LF, Belshaw NS, and Henderson GM (2004) U and Th concentrations and isotope ratios in modern carbonates and waters from the Bahamas. *Geochimica et Cosmochimica Acta* 68(8): 1777–1789.
- Scholz D, Mangini A, and Felis T (2004) U-series dating of diagenetically altered fossil reef corals. *Earth and Planetary Science Letters* 218(1–2): 163–178.
- Stirling CH and Andersen MB (2009) Uranium-series dating of fossil coral reefs: Extending the sea-level record beyond the last glacial cycle. *Earth and Planetary Science Letters* 284: 269–283.
- Stirling CH, Esat TM, Lambeck K, and McCulloch MT (1998) Timing and duration of the Last Interglacial: Evidence for a restricted interval of widespread coral reef growth. *Earth and Planetary Science Letters* 160: 115–130.
- Stirling CH, Esat TM, Lambeck K, et al. (2001) Orbital forcing of the marine isotope stage 9 interglacial. *Science* 291: 290–293.
- Thompson WG (submitted). Why the duration of the Last Interglacial is systematically overestimated. *Earth and Planetary Science Letters*.
- Thompson WG, Curran HA, Wilson MA, and White B (2011) Sea level and ice sheet instability during the Last Interglacial: New Bahamas evidence. *Nature Geoscience* 4: 684–687.
- Thompson WG and Goldstein SL (2005) Open-system coral ages reveal persistent suborbital sea-level cycles. *Science* 308(5720): 401–404.
- Thompson WG and Goldstein SL (2006) A radiometric calibration of the SPECMAP timescale. *Quaternary Science Reviews* 25: 3207–3215.
- Thompson WG, Spiegelman MW, Goldstein SL, and Speed RC (2003) An open-system model for the U-series age determinations of fossil corals. *Earth and Planetary Science Letters* 210: 365–381.
- Villemant B and Feuillet N (2003) Dating open systems by the ^{238}U – ^{234}U – ^{230}Th method: Application to Quaternary reef terraces. *Earth and Planetary Science Letters* 210(1–2): 105–118.
- Wang YJ, Cheng H, Edwards RL, et al. (2001) A high-resolution absolute-dated late Pleistocene monsoon record from Hulu Cave, China. *Science* 294: 2345–2348.

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Varved Lake Sediments

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Introduction

Natural archives of past climates and environments are necessary to improve our understanding of climate change during the Holocene and beyond, and to estimate the degree of human impact on past environments. Instrumental and written (historic) information about these conditions are restricted to the last few centuries and millennia, respectively. Therefore, records that archive datasets about natural (climatic) and human environmental change are relevant to extend our knowledge about trends, variabilities, events, and their forcing factors. This information is also essential to develop and verify climate modeling. For this approach, records of decadal or annual resolution are preferred by paleoclimatologists, because only such records provide proxy data on a timescale relevant to society, that is, with a temporal resolution of less than one human generation (<30 yr). There are only a few types of incrementally dated records that provide this high resolution: varved lake sediment is one of them.

The term ‘varve’ is a Swedish word that means cyclic layer. It was first used in 1862 as *Hvarfig lera* by the Swedish Geological Survey to describe rhythmically deposited clays in a proglacial environment on one of the first geological maps of Sweden.

History

The study of varved lake sediments has a history of more than 150 yr. At the same time that Swedish geologists were mapping varved clays for the first time, alternating pale and dark laminated lake sediments were interpreted as a potential source of chronological information by A. Heim in Switzerland and by E. Hitchcock in North America. But another 50 yr passed until this concept was developed into a useful dating tool, completely different from the common method of stratigraphic classification. It was again in Sweden that the geologist Gerard De Geer applied the chronological information stored in varved clays that were deposited near the margin of the retreating Scandinavian ice sheet to determine the time elapsed since the termination of the Pleistocene. De Geer established a network of sites along the Baltic coast of Sweden and linked them using distinct varve thickness variations (De Geer, 1912). Moreover, he defined the term ‘varve’ as being one complete annual cycle consisting of a coarse-grained pale summer layer

and a fine-grained dark winter layer (Figure 1). He also coined the term ‘geochronology’ and for the first time the annual resolution timescale was used in geology (De Geer, 1912). Varves were regarded as typical deposits related to glacial meltwater entering a proglacial lake.

De Geer was convinced that solar forcing controls meltwater discharge and in turn varve thickness variations. Therefore, De Geer and his followers matched Swedish varve thickness records with records from Finland (M. Sauramo), North America (E. Antevs), and Patagonia (C. Caldenius) to provide additional evidence for the idea of global forcing of varve thickness variations. However, his long-distance correlation (teleconnection) was soon found to be unreliable. This discredited varve chronologies for a long time as they were regarded as a subjective and nonreliable dating tool. The attitude against this incremental dating method was intensified due to the advent of a new, seemingly precise dating method discovered in 1949 – radiocarbon dating.

It was not until the 1970s that varved records started to become of interest again. The discussion of human impact on the environment (eutrophication, acidification, pollution, soil erosion, climate change) needed high-resolution information about the past to allow an extension of the short instrumental records of environmental change and to determine natural background levels. In combination with new coring techniques and more sophisticated analytical tools, varved records became one focus of paleoenvironmental research. For the first time nonglacial lakes were investigated and thus it was only fitting that the term ‘varve,’ so far restricted to proglacial environments, was extended to all annually (seasonally) laminated sediments on the continents (O’Sullivan, 1983; Anderson et al., 1985; Saarnisto, 1986) as well as in the oceans (Kemp, 1996). Since then, the terms ‘varved sediments’ and ‘annually laminated sediments’ have been used as synonyms. Occasionally, for proglacial varves, the terms ‘clastic’ or ‘classic varves’ are used to distinguish them from nonglacial organic or biogenic varves.

Additionally, since the 1980s, it has become increasingly obvious that radiocarbon dating does not provide a calendar year chronology but needs to be calibrated (van der Plicht, 2000). Dendrochronology is the method of choice to calibrate Holocene radiocarbon ages. However, for the Late Glacial and beyond trees are rare or absent and thus annually laminated sediments have been used for calibration (van der Plicht, 2000).

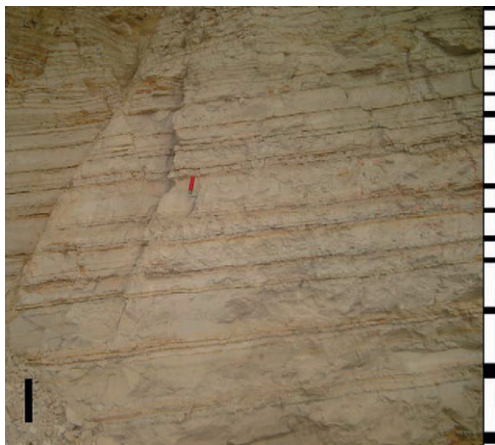


Figure 1 Pleistocene glacial varves cropping out at Eberswalde-Macherslust, Brandenburg, Germany. Note the decreasing varve thickness towards the top as a result of increasing distance from the active ice margin. The scale bar is 30 cm. Photograph by B. Zolitschka.

Varved lake sediments are also ideal to trace past land-use changes if used in combination with archaeological investigations (Hicks et al., 1994). Important societal issues include geological hazards such as earthquakes, volcanic eruptions, or tsunamis, which can be recognized and dated precisely in varved records. More recently, efforts have been undertaken to better understand the processes that cause the formation of varved lake sediments by applying sediment trap studies in lakes (Leemann and Niessen, 1994) in combination with highest-resolution image analysis techniques capable of deciphering the seasonal structure of varves (Francus, 2004).

Formation of Varved Lake Sediments

Annually laminated lake sediments are not typical for the Quaternary and have also been reported for sedimentary sequences as far back as the Paleozoic (Anderson and Dean, 1988). Their variability in space and time, however, is difficult to assess because their formation processes are not well enough understood. This is different for Quaternary and especially for late Holocene records, a time period when predominantly modern processes were active.

The formation of varved lake sediments depends on overall climatic conditions. The main controlling factor is the seasonal variability of temperature and precipitation with amplitudes twice as large as long-term climatic fluctuations. Seasonal climatic variation is responsible for the succession of different life forms in the lake waters and for physical and chemical processes in the lake and its catchment area controlling biological productivity, precipitation of authigenic minerals as well as transport of minerogenic and organic particles from the catchment area to the lake. These natural conditions are dominated by the regional climate and control the prevailing mechanisms of lacustrine sedimentation (Håkanson and Jansson, 1983).

Clastic sediments. Minerogenic particles are carried through fluvial input from the catchment area or by atmospheric deposition to the lake. To a certain degree organic fragments such as

leaves or pieces of wood may additionally be incorporated via runoff.

Biogenic sediments. Biological productivity in the lake generates organic matter that sinks to the lake bottom and accumulates.

Evaporitic sediments. Chemical precipitation of minerals from the water column adds authigenic mineral components.

Sediment constituents are classified into autochthonous (produced in the lake) and allochthonous components (formed in the catchment area). Although there are clear environmental constraints necessary for the formation of the varve types related to these depositional processes, they are mostly the result of a combination of processes. Altogether, the annual rhythm produces an internal clock, which makes varved lake sediments ideal for environmental reconstruction as it provides continuous dating without interpolation, typical for all incremental dating methods. Thus, seasonally changing constituents are deposited, leading to the formation of a characteristic varve structure with at least two laminae that ideally contrast in both color and composition.

Preservation of Varves

Only few lakes are reported to contain long records with annual laminations. This occurs despite a continuous lacustrine sedimentation process with a pronounced seasonal climatic forcing for almost every lake except for equatorial lakes that not necessarily have a clear seasonal signal. The reason for this is that certain internal lake processes can prevent varve preservation:

- burrowing activities of benthic organisms (bioturbation),
- resuspension of bottom sediments through wind- or density-driven lake water circulation, and
- bottom currents that interrupt continuous sedimentation and cause erosional hiatuses.

The latter two processes are excluded if the lake basin is deep with a small surface area and is geomorphologically protected. Such preconditions exclude or reduce strong wind influences and promote varve preservation (O'Sullivan, 1983; Ojala et al., 2000). These morphological features, in combination with eutrophic conditions, favor the formation of anoxic bottom water, which is the result of oxygen depletion related to decomposition of organic matter near the sediment surface. This eventually restricts or excludes the presence of bottom-dwelling organisms and thus eliminates bioturbation. Anoxic conditions can be permanent, as in meromictic lakes (Anderson et al., 1985), or seasonal to preserve annual laminations.

Preconditions for varve preservation are natural or anthropogenic. Cultural eutrophication related to industrialization since the mid-nineteenth century has often generated anoxia in deep lakes initiating the formation of varved lake sediments (Figure 2; Wehrli, 1997; Lüder et al., 2006).

Clastic Varved Lake Sediments

Clastic sediments predominate under cold climatic conditions, such as those found in the Arctic (Figure 3) or in high Alpine regions (Figure 4). Such sediments are typical for proglacial

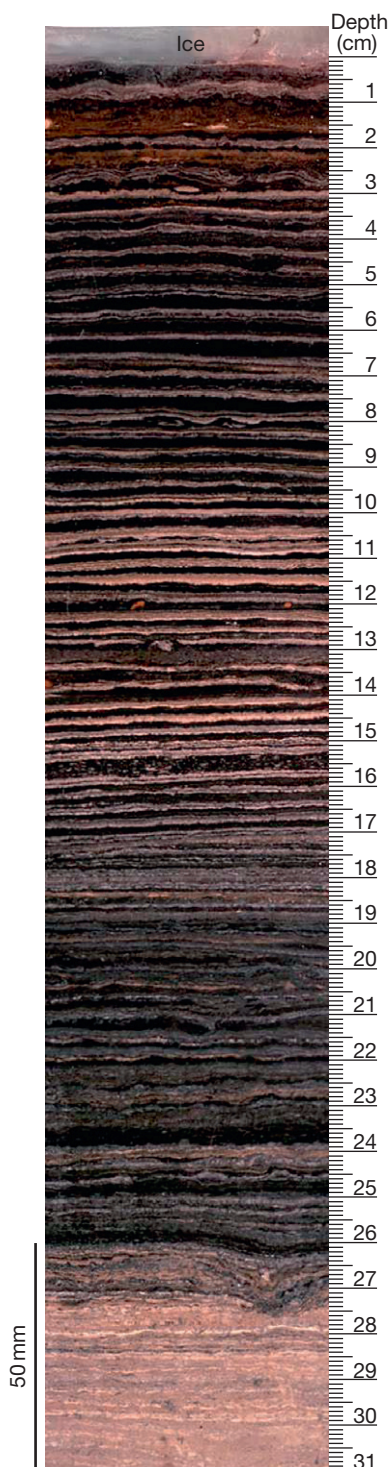


Figure 2 Varved lake sediments from Greifensee, Switzerland. The upper 26 cm of this freeze core displays the history of cultural eutrophication through the formation of carbonaceous organic varves since 1930. Below 26 cm the productivity of the lake was lower; no oxygen depletion occurred near the lake floor and thus organic matter was not preserved. Photograph by M. Sturm.

and periglacial lakes. Intensive physical weathering and the lack of a densely vegetated catchment area provide high amounts of minerogenic detritus, which is easily eroded and

transported into the lake. The sediment transfer is related to the annual freeze–thaw cycle and the amount of runoff. In regions with a continental climate, runoff is governed by melting of snow and ice through solar insolation during summer. Under oceanic climatic conditions, runoff is controlled either by the melting of snow and ice through advective heat transport or by precipitation. Such lakes are poor in nutrients (oligotrophic), which inhibits high organic productivity and the formation of organic varves.

Clastic varves are formed when a discontinuous (highly seasonal) stream loaded with suspended sediment enters a lake of stratified water (Sturm, 1979). According to its density and in relation to the density of the lake water, the inflowing stream water is positioned in the lake as over-, inter-, or under-flow. Over- and interflows cause a wide distribution of suspended matter across the lake. After entering the lake, the flow velocity of the stream water is transformed into turbulence causing a reduction of transport capacity for suspended sediment particles. Accordingly, coarse particles (sand, silt) are deposited immediately, whereas fine particles (fine silt, clay) remain in suspension until runoff has stopped or the lake is ice-covered. Thus, clastic varves are composed of a pale coarse-grained basal lamina and a darker fine-grained (clay) top lamina (Figure 3) or vice versa (Figures 4 and 5; Table 1). Additional coarse-grained laminae are frequently observed and are often related to successive annual runoff events.

Organic Varved Lake Sediments

Under temperate humid climatic conditions the catchment area is vegetated, reducing the availability and transport capability of clastic material into the lake. Chemical weathering is prevalent and releases nutrients from the bedrock that are incorporated into plant organic matter, buffered in soils, or washed out and transported in dissolved form to the lake. Nutrient rich (eutrophic) conditions in the lake are the consequence. The related increase of organic productivity leads to the formation of organic varves reflecting the lacustrine life cycle (Figures 5 and 6).

In late winter or early spring, mixing of the lake water body (holomixis) distributes nutrients to the photic zone where algal blooms develop. Diatoms mainly start reproducing in spring, followed by green and blue-green algae during summer. Sometimes, after the second annual mixis in fall, diatoms bloom again. A layer of chrysophyte cysts often terminates the lacustrine life cycle (Figures 5 and 6; Table 1). While green and blue-green algae decompose during deposition, siliceous diatom frustules are resistant to dissolution and are preserved in the sedimentary record.

Another component of organic varves are authigenic calcites. Carbonaceous laminae are present if limestone or limy particles of unconsolidated till or loess deposits are dissolved and transferred via runoff into the lake (Figures 2, 5, and 7). In such hard-water lakes calcite crystals are precipitated. Ca^{2+} and (HCO_3^-) stay in solution in the lake water until their saturation point is reached. This is partly achieved by the seasonal temperature increase in the surface water during insolation maxima in summer. As CaCO_3 decreases in solubility with temperature, calcite may precipitate due to this effect alone. However, the temperature-related increase of photosynthetic

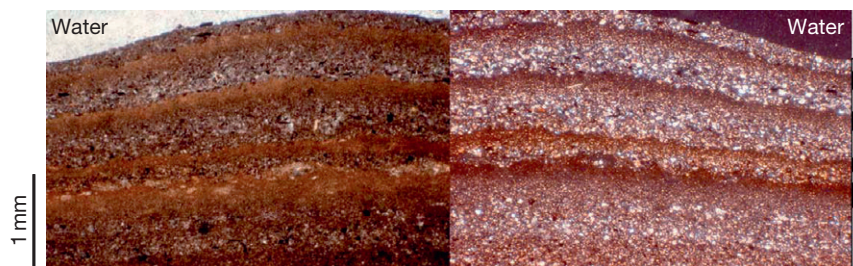


Figure 3 Clastic varves of Lake C2, north coast of Ellesmere Island, Canadian High Arctic. The microscopic photograph of nonglacial clastic varves shows the undulated sediment–water interface in normal (left) and polarized light (right). Pale and coarse-grained laminae are related to snowmelt runoff events during summer, which contrast to the dark fine-grained laminae on top that settled out of suspension during winter under ice cover. Photographs by B. Zolitschka.

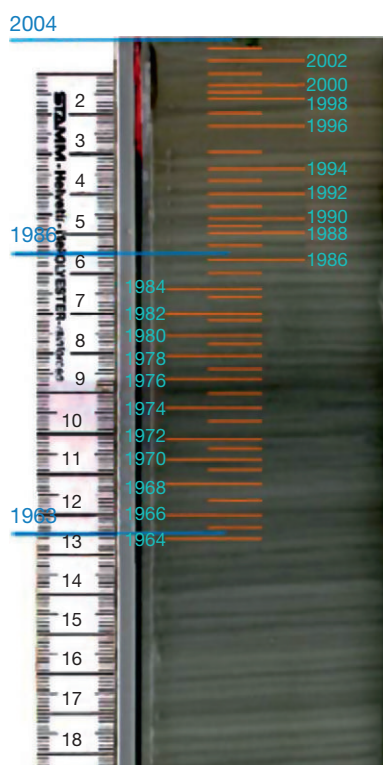


Figure 4 Nonglacial clastic varves of Alpine Brienzersee, Switzerland. Varve counts are marked and labeled; the scale is in cm. The marker years (1963, 1986) determined by radiometric dating (^{137}Cs) are highlighted as well as the sediment surface (2004). Photograph by M. Sturm.

activities in the water column through phytoplankton blooms is more important. Accordingly, CO_2 is withdrawn from the water, pH rises up to 9, and CaCO_3 solubility is reduced. Thus, the combination of temperature increase and phytoplankton activities adds another autochthonous component to organic varves – white carbonate laminae (Figure 5 and Table 1).

In addition to these autochthonous sedimentary components, an allochthonous component to organic varves may also occur during winter. As plant cover is reduced and soils become more susceptible to erosion, runoff increases, especially after fall and winter rain events. This transports minerogenic as well as (littoral and terrestrial) organic debris to the lake (Table 1).

Evaporitic Varved Lake Sediments

Under arid or semiarid climatic conditions evaporitic varves (Figure 8) may be formed if the salinity and pH of the lake water increase through enhanced evaporation. This causes the saturation of specific mineral compounds (salts), which precipitate out of the water column. In addition to calcite, which can precipitate either biogeochemically in mid-latitudes (organic varves) or physicochemically under semiarid climatic conditions, evaporitic varves are composed of aragonite, gypsum, and halite. They are typically composed of a pale–dark couplet (Figure 8 and Table 1). The pale lamina is attributed to salt precipitates, whereas the dark lamina is related to runoff events introducing a mixture of minerogenic and organic detritus. Runoff occurs during the less arid part of the year, usually during winter.

Varve Chronology and Applications

Varved lake sediments are ideal sources of paleoenvironmental and paleoclimatic data because:

- varves provide an internal, continuous time frame in calendar years (Figure 4);
- radiometric dating and sometimes tephrochronology or OS� dating can be applied to the same record (multiple dating approach) to independently verify the varve-based age model (Zolitschka et al., 2000);
- variations in varve thickness (Figure 9) and related microfacies analyses provide a continuous dataset of information with highest possible time resolution if the local processes of varve formation are known;
- quantified flux rates for the entire record can be precisely determined, based on the continuous chronological information and in combination with dry density measurements; and
- precise time control and flux rates improve supplementary analyses carried out on discrete (noncontinuous) subsamples such as sedimentological, micropaleontological (pollen, diatoms, chrysophytes, chironomids, ostracods), geophysical, or inorganic and organic geochemical analyses.

However, each studied proxy parameter is only one piece in a big puzzle with an indeterminable number of jigsaw pieces. The total number of pieces needed to understand the full range

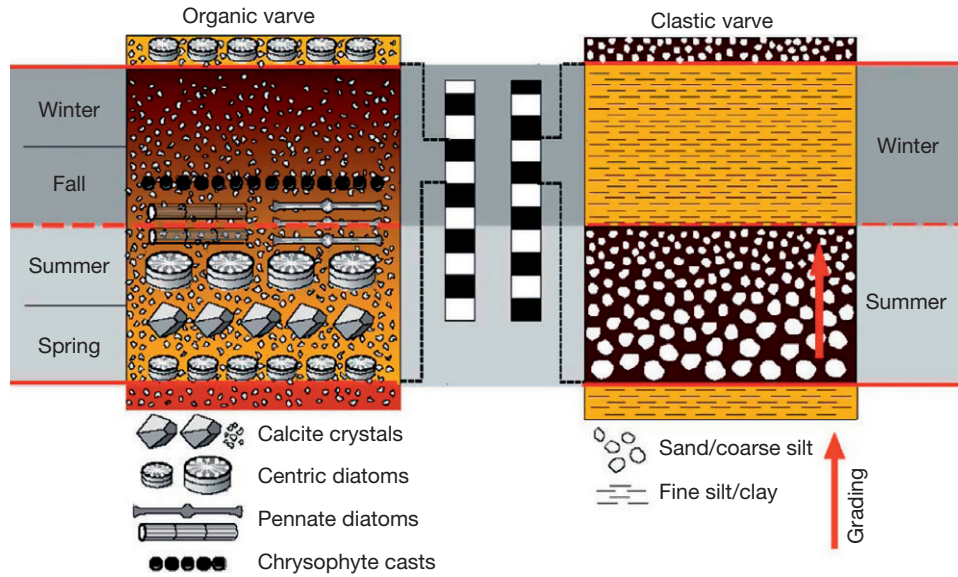


Figure 5 Model of varve formation for carbonaceous organic varves (left) and clastic varves (right). Design: Sturm M and Lotter A (1995). Lake sediments as environmental archives. *EAWAG News* 38E: 6–9.

Table 1 Different varve types with idealized composition, color, and timing of corresponding seasonal laminae

	<i>Clastic varves</i>	<i>Organic varves</i>	<i>Evaporitic varves</i>
Spring		First Diatom bloom (pale)	
Summer	Coarse-grained minerogenic particles via runoff (pale or dark)	Calcite precipitation (white, optional)	Precipitation of calcite, aragonite, gypsum or halite (white)
Late summer		Second diatom bloom, chrysophytes cysts (pale, both optional)	
Fall/winter	Fine-grained minerogenic particles out of suspension (dark or pale)	Organic and minerogenic detritus via runoff (dark)	Organic and minerogenic detritus via runoff (dark)

of complex environmental variability is site-, time- and proxy-specific and thus varies even within the same record.

The annual character of laminations needs to be verified for the study of varved lake sediments and prior to establishing a varve chronology. This includes error estimations. Finally, environmental reconstruction builds upon this temporal framework.

Varve Verification

Before varves are counted and their thickness measured it is necessary for every specific site to develop a conceptual and mostly idealized model about the processes that formed the different varve laminae (Figure 5; Lamoureux, 2001). This varve model depends on limnological conditions as well as on environmental conditions in the catchment area. It should characterize specific features in varve composition, which are later used to determine boundaries between consecutive varves and to identify laminae of no chronological meaning (turbidites, homogenites). This approach is especially important if varves cannot be easily distinguished. The development of such a process-oriented varve model (Figure 5) should ideally be supported by sediment trap studies (Leemann and Niessen, 1994) and micropaleontological methods, for example, by investigating the succession of pollen types or diatom species

during one year (Card, 1997; Hausmann et al., 2002). A varve chronology can only be established if the structure of the varve cycle is clear.

Varve Chronology

After determination of the annual structure, the varve chronology is established through replicate varve counts and thickness measurements along at least three transects (Lamoureux, 2001). This is usually performed either via microstratigraphical (microscopic) analysis of thin sections (Brauer, 2004) or via techniques related to image analysis (Francus, 2004). Both methods have their own characteristic advantages: microstratigraphic investigation allows the analysis of internal structure and composition (microfacies) of every individual varve, whereas image analysis techniques provide more objective and faster results. Based on thickness measurements of every individual varve (Figure 9), the age–depth model (varve chronology) is established with as many data points as varves that have been counted.

Varve chronologies are regarded as absolute chronologies, if they are continuous and extend to the sediment surface. However, in many cases, inconsistencies of the chronology or hiatuses have been detected. Such often-encountered difficulties

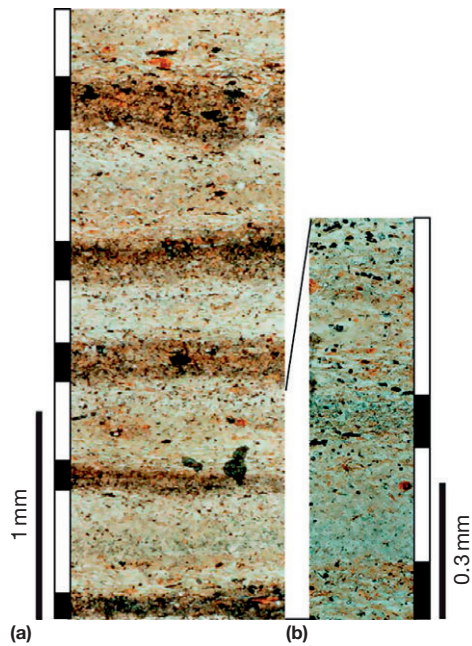


Figure 6 Organic varved lake sediments from Holzmaar, Eifel, Germany. A, The macroscopic photograph shows organic varves composed of diatoms (pale laminae) and organic detritus (dark laminae); B, The microscopic photograph with normal light (right) shows the internal structure of these organic varves with massive planktonic diatom blooms in pale laminae and larger benthic and epiphytic diatoms in dark laminae. Photographs by B. Zolitschka.

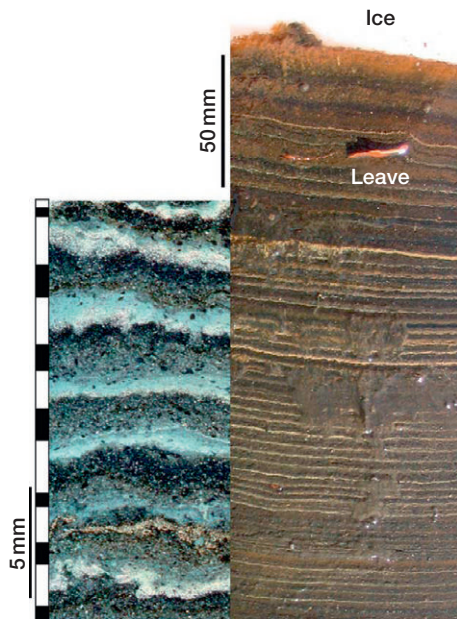


Figure 7 Varved lake sediments from Sacrower See near Potsdam, Brandenburg, Germany. Macroscopic photograph (right): Freeze core of carbonaceous organic varves with well-preserved sediment-water (ice) interface on top and a well-preserved leave at 4 cm depth. Pale laminae are composed of calcite. Microscopic photograph with polarized light (left): Internal structure of carbonaceous organic varves. The annual succession consists of three laminae: spring planktonic diatom blooms (almost black), calcite laminae from summer (white) and detritic wintertime laminae (grayish). Photographs by D. Enters (left), P. Bluszcz (right).

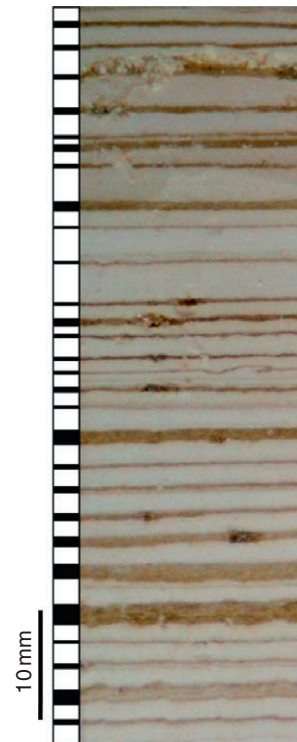


Figure 8 Evaporitic varves from the late Pleistocene Lisan Formation, Dead Sea, Israel. The annual cycle is composed of white aragonite laminae precipitated during arid summer and dark laminae resulting from winter flood events. Photograph by M. Stein.

can only be overcome by applying additional independent chronological methods (multiple dating approach). Usually, these include radiocarbon dating (^{14}C) and, less extensively, OSL dating. In addition, stratigraphic methods are applied to provide another line of evidence in support of time control and to facilitate regional correlation. Most common among these is palynology from which distinct changes in regional patterns of vegetation are derived (biostratigraphy). For a larger, regional scale, paleomagnetic studies are ideal because they reveal global variations of the Earth's magnetic field (magnetostratigraphy). In volcanically active regions, tephra layers provide another means of developing an independent time frame (tephrochronology). Altogether, multiple dating significantly improves the accuracy of dating (Zolitschka et al., 2000).

Sometimes the connection of varved sediment records to the present day can be problematic. This is checked with radiometric dating (^{137}Cs , ^{210}Pb) or by comparing the record with historic data or events (Lüder et al., 2006). Varve chronologies are called floating chronologies, if they cannot be connected to the present day.

Error Estimations

Systematic errors of varve chronologies (Sprowl, 1993) are attributed to different sources.

Technical problems. These errors are caused by incomplete core recovery (sediment gaps as the result of coring artifacts) or incorrect correlation of core segments.

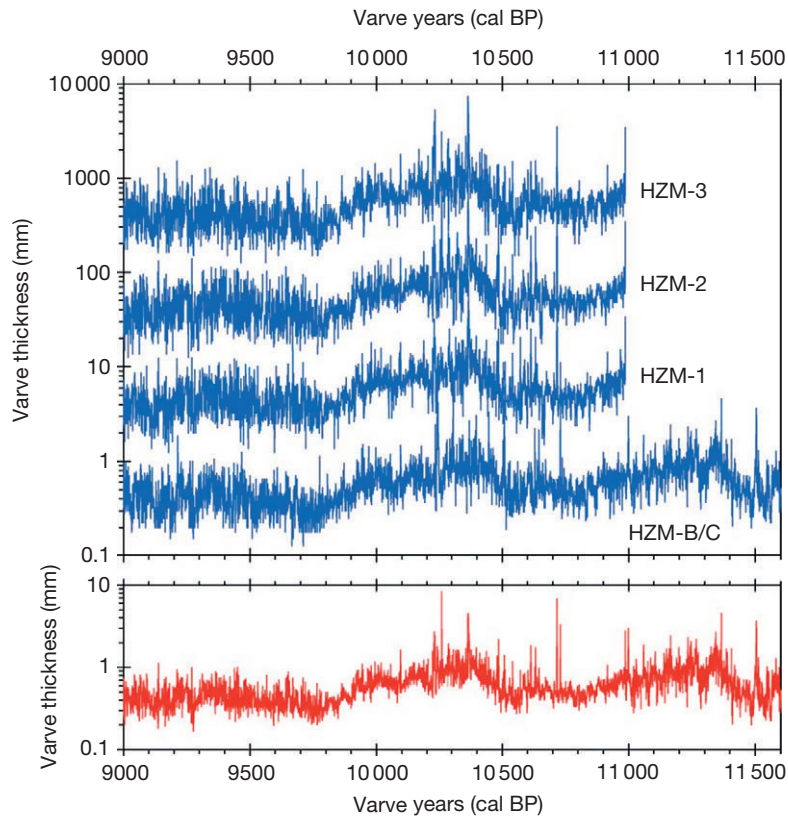


Figure 9 Varve thickness measurements (cf. Fig. 6) of four parallel sediment sequences (blue) from Holzmaar, Eifel, Germany. Each core is offset by a factor of 10 to separate the plots. early Holocene varves were counted and measured using thin sections by two different persons (HLM-B/C by B. Zolitschka; HLM-1, HLM-2, and HLM-3 by B. Rein). In the bottom panel the stacked varve record is displayed in red reflecting the climatic fluctuations of the Preboreal Oscillation (11,400–11,000 cal. BP) and the Boreal Oscillation (10,500–10,200 cal. BP); both fluctuations show markedly thicker varves related to increased minerogenic sediment input via runoff.

Depositional events. Such errors occur as the result of erosion (erosive turbidites, volcanic ash layers), deposition of plant or animal macrofossils (e.g., tree trunks or animal carcasses), or drop stones, and are restricted to certain areas on the lake floor.

Sedimentation rates. The occurrence of very high (>20 mm) or very low (<0.2 mm) varve thickness makes the individual annual layer difficult to distinguish. Thick varves are generally composed of many seasonal laminae (sometimes >10, related to rainfall or runoff events), which are hard to distinguish from the annual cycle. Thin varves show the opposite: laminae within the annual cycle are extremely thin, difficult to determine, or even missing. For these varves, the entire annual cycle is easily mistaken as a lamina of a seemingly thicker varve.

Varve preservation. Changes in lake level or autochthonous productivity have significant effects on anoxic bottom water conditions and thus on varve preservation (Figure 2). Periods with well-preserved varves may thus alternate with periods of no varve preservation.

Technical problems as well as depositional events can often be avoided by a careful selection of coring sites. Moreover, potential errors can be minimized by applying appropriate coring techniques and recovering several parallel and overlapping core series from the same lake (Figure 9) that are matched and cross-dated. The other errors mentioned are difficult to circumvent and often must be regarded as an inherent problem

to varve chronologies. This increases the importance of cross-checking varve chronologies against other independent dating methods following the multiple dating approach.

If varves are well developed and not disrupted, replicate counts show only little variation between different counts and observers (Figure 9; Ojala and Saarnisto, 1999). The counting error will be low. But if varved sediments are disturbed or not very distinct, counting errors increase (Tian et al., 2005). Generally, the determination of errors as variance between different investigators or replicate counts gives only a measure for the studied section. It does not provide any indication about the chronological error of the entire record, which is mostly the result of undetected systematic errors. Such errors can only be determined either by applying the multiple dating approach or by comparing environmental reconstruction with other well-dated records.

Paleoenvironmental Reconstruction Based on Varved Lake Sediments

Recently, multiproxy investigations of varved lake sediments became common because they provide well-dated independent lines of evidence for environmental change as response to climate change and human impact and thus considerably improve the possibilities for interpretation. Several

compilations concerning annually laminated sediments covering methodological aspects as well as interdisciplinary approaches to paleoclimatic and paleoenvironmental interpretation give an overview about possibilities and restrictions and are suggested for further reading. Special issues on individual lakes are available for Elk Lake, Minnesota (Bradbury and Dean, 1993), Lake C2, Canadian High Arctic (Bradley, 1996), Baldeggersee, Switzerland (Wehrli, 1997), and Lake Gosciąz, Poland (Ralska-Jasiewiczowa et al., 1998). Additionally, five volumes are dedicated to methods and applications of varved sediments: an introduction to the formation and composition of clastic varves (Schluchter, 1979), methods and case studies of laminated lacustrine sediments throughout Europe (Saarnisto and Kahra, 1992), the significance of varved lake sediments for the study of human impact on the environment (Hicks et al., 1994), applications of laminated sediments as paleoindicators (Kemp, 1996), and the calibration of radiocarbon dates using varve chronologies (van der Plicht, 2000).

Conclusion

Time control is one of the most important issues for the study of the Quaternary. Without an accurate timescale it is impossible to compare instrumental, historical, and archaeological data as well as other proxy records from natural archives such as lakes, tree rings, ice cores, peat bogs, speleothems, oceans, or corals on a local, regional, or global scale. Of all sedimentary records, varved lake sediments provide the best possibility of solving these chronological concerns. In addition to a precise and reliable high-resolution timescale, varved lake sediments are usually studied in an interdisciplinary way including sedimentological, geochemical, geophysical, and biological methods, combined within the field of paleolimnology (Berglund, 1986; Last and Smol, 2001). Therefore, such records provide a unique opportunity to study the dynamics of past environmental systems on various spatial and temporal scales. This perspective is necessary to realistically estimate future changes, changes of the environmental system that are at least partly the result of human activities and which will definitely influence the future of human societies.

See also: Dating Techniques; Dendrochronology; Diatom Introduction; Geomagnetic Excursions and Secular Variations; Pollen Analysis, Principles; Varved Marine Sediments. **Chironomid Records:** Chironomid Overview. **Diatom Records:** Freshwater Laminated Sequences. **Glacial Climates:** Volcanic and Solar Forcing. **Introduction:** History of Dating Methods; Societal Relevance of Quaternary Research. **Lake Level Studies:** Overview. **Luminescence Dating:** Optical Dating. **Paleobotany:** Overview of Terrestrial Pollen Data. **Paleoceanography, Physical and Chemical Proxies:** Terrigenous Sediments. **Paleoclimate:** Introduction; Timescales of Climate Change. **Paleoclimate Modeling:** Quaternary Environments. **Paleoclimate Reconstruction:** Approaches; The Last Millennium. **Paleolimnology:** Overview of Paleolimnology. **Pollen Methods and Studies:** Use of Pollen as Climate Proxies. **Quaternary Stratigraphy:** Chronostratigraphy; Overview; Tephrochronology. **Radiocarbon Dating:** ^{14}C of Plant Macrofossils; AMS Radiocarbon Dating; Calibration of the ^{14}C Record; Sources of Error.

References

- Anderson RY and Dean WE (1988) Lacustrine varve formation through time. *Palaeogeography, Palaeoclimatology, Palaeoecology* 62: 215–235.
- Anderson RY, Dean WE, Bradbury JP, and Love D (1985) Meromictic lakes and varved sediments in North America. *United States Geological Survey Bulletin* 1607: 1–19.
- Berglund BE (ed.) (1986) *Handbook of Holocene Palaeoecology and Palaeohydrology*, p. 869. Chichester and New York: Wiley.
- Bradbury JP and Dean WE (eds.) (1993) *Elk Lake, Minnesota: Evidence for rapid climate change in the north-central United States*, p. 336. Geological Society of America Special Paper, 276.
- Bradley RS (ed.) (1996) The Taconite Inlet lakes project. *Journal of Paleolimnology* 16: 97–255.
- Brauer A (2004) Annually laminated lake sediments and their palaeoclimatic relevance. In: Fischer H, Kumke T, and Lohmann G, et al. (eds.) *The Climate in Historical Times – Towards a Synthesis of Holocene Proxy Data and Climate Models*, pp. 109–128. Berlin: Springer.
- Card VM (1997) Varve-counting by the annual pattern of diatoms accumulated in the sediment of Big Watab Lake, Minnesota, AD 1837–1990. *Boreas* 26: 103–112.
- De Geer G (1912) A geochronology of the last 12,000 years. *Proceedings of the International Geological Congress Stockholm*, (1910) 1: 241–257.
- Francus P (2004) Image analysis, sediments and paleoenvironments. *Developments in Paleoenvironmental Research*, p. 330. Dordrecht: Springer.
- Håkanson L and Jansson M (1983) *Principles of Lake Sedimentology*, p. 316. Berlin and Heidelberg: Springer.
- Hausmann S, Lotter AF, van Leeuwen JFN, et al. (2002) Interactions of climate and land use documented in the varved sediments of Seeburgsee in the Swiss Alps. *The Holocene* 12: 279–289.
- Hicks S, Miller U, and Saarnisto M (eds.) (1994) Laminated sediments. *Journal of the European Study Group on Physical, Chemical, Biological and Mathematical Techniques Applied to Archaeology* 41: 148.
- Kemp AES (ed.) *Palaeoclimatology and Palaeoceanography from Laminated Sediments*, pp. 258. Geological Society of London Special Publication, 116, The Alden Press, Oxford.
- Lamoureux S (2001) Varve chronology techniques. In: Last WM and Smol JP (eds.) *Tracking Environmental Change Using Lake Sediments: Physical and Geochemical Techniques*, 247–260. *Developments in Paleoenvironmental Research* 1. Dordrecht: Kluwer.
- Last WM and Smol JP (eds.) (2001) Tracking environmental change using lake sediments. *Developments in Paleoenvironmental Research*, vol. 1–4. Dordrecht: Kluwer.
- Leemann A and Niessen F (1994) Varve formation and the climatic record in an Alpine proglacial lake: Calibrating annually laminated sediments against hydrological and meteorological data. *The Holocene* 4: 1–8.
- Lüder B, Kirchner G, Lücke A, and Zolitschka B (2006) Cultural eutrophication processes of the hardwater lake Sacrower See (north-eastern Germany) since the 17th century. *Journal of Paleolimnology* 35: (in press).
- Ojala A and Saarnisto M (1999) Comparative varve counting and magnetic properties of a 8400 years sequence of an annually laminated sediment in Lake Valkiajärvi, Central Finland. *Journal of Paleolimnology* 22: 335–348.
- Ojala A, Saarinen T, and Salonen V-P (2000) Preconditions for the formation of annually laminated lake sediments in southern and central Finland. *Boreal Environment Research* 5: 243–255.
- O'Sullivan PE (1983) Annually laminated lake sediments and the study of Quaternary environmental changes. *Quaternary Science Reviews* 1: 245–313.
- Ralska-Jasiewiczowa M, Goslar T, Madeyska T, and Starkel L (eds.) (1998) *Lake Gosciąz, Central Poland – A Monographic Study, Part 1*, p. 340. Krakow: W. Szafer Institute of Botany.
- Saarnisto M (1986) Annually laminated lake sediments. In: Berglund BE (ed.) *Handbook of Holocene Palaeoecology and Palaeohydrology*, pp. 343–370. Chichester and New York: Wiley.
- Saarnisto M and Kahra A (eds.) (1992) *Laminated Sediments*. Geological Survey of Finland, Special Paper 14, p. 113. Espoo.
- Schluchter C (ed.) (1979) *Moraines and Varves*, p. 441. Rotterdam: A.A. Balkema.
- Sprowl DR (1993) On the precision of the Elk Lake varve chronology. In: Bradbury JP and Dean WE (eds.) *Elk Lake, Minnesota: Evidence for rapid climate change in the North-Central United States*, pp. 69–74. Geological Society of America Special Paper 276.
- Sturm M (1979) Origin and composition of clastic varves. In: Schluchter C (ed.) *Moraines and Varves*, pp. 281–285. Rotterdam: A.A. Balkema.

- Sturm M and Lotter A (1995) Lake sediments as environmental archives. *EAWAG News* 38E: 6–9.
- Tian J, Brown T, and Hu F (2005) Comparison of varve and ^{14}C chronologies from Steel Lake, Minnesota, USA. *The Holocene* 15: 510–517.
- van der Plicht J (ed.) (2000) The 2000 radiocarbon varve-comparison issue. *Radiocarbon* 42: 313–452.
- Wehrli B (ed.) (1997) High resolution varve studies in Baldeggersee. *Aquatic Sciences* 59: 283–375.
- Zolitschka B, Brauer A, Negendank JFW, Stockhausen H, and Lang A (2000) Annually dated late Weichselian continental paleoclimate record from the Eifel, Germany. *Geology* 28: 783–786.

Varved Marine Sediments

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Introduction and History

The study of varved sediments has a long history extending back more than 150 years. The term varve is a Swedish word that means cyclic layer. It was first used in 1862 as 'Hvarfig lera' by the Swedish Geological Survey to describe rhythmically deposited clays in a proglacial environment on one of the first geological maps of Sweden. At the same time when Swedish geologists mapped varved clays for the first time, alternating pale and dark laminated lake sediments were interpreted as a potential source of chronological information by A. Heim in Switzerland and by E. Hitchcock in North America. But another 50 years passed until this concept was developed into the first dating tool, completely distinct from the common stratigraphic classification. It was again in Sweden that the geologist Gerard De Geer applied the chronological information stored in varved clays that were deposited near the margin of the retreating Scandinavian ice sheet to determine the time elapsed since the termination of the Pleistocene. De Geer established a network of sites along the Baltic coast of Sweden and linked them using recognizable variations in varve thickness (De Geer, 1912). Moreover, he defined the term varve as being one complete annual cycle consisting of a coarse-grained pale summer layer and a fine-grained dark winter layer (Figure 1). In addition, De Geer coined the term geochronology and for the first time introduced a temporal unit (the year) to geology (De Geer, 1912). Following his work, varves were regarded as typical deposits related to glacial meltwater entering a proglacial lake.

De Geer was convinced that solar forcing controls meltwater discharge and in turn varve thickness variations. Therefore, De Geer and his followers matched Swedish varve thickness records with records from Finland (M. Sauramo), North America (E. Antevs), and Patagonia (C. Caldenius) to provide additional evidence for the idea of global forcing of varve thickness variations. However, his long-distance correlations (teleconnection) were soon discredited, resulting in a period of reduced trust in varve chronologies, such that they were regarded as a subjective and nonreliable dating tool. The attitude against this incremental dating method was intensified due to the advent of a new seemingly precise dating method discovered in 1949 – radiocarbon dating. It was not before the 1970s that varved records started to become of interest again. The discussion of human impact on the environment (eutrophication, acidification, pollution, soil erosion, climate change) needed high-resolution information about the past to allow an extension of the short instrumental records of environmental change and to determine natural background levels. In combination with new coring techniques and more sophisticated analytical tools, varved records became one focus of paleoenvironmental research. At that time, predominantly nonglacial lakes were being investigated, and it was only subsequent to this that the term varve, so far restricted to proglacial environments, was

extended to all annually (seasonally) laminated sediments on the continents (O'Sullivan, 1983; Saarnisto, 1986) as well as in the oceans (Kemp, 1996). As a result, the terms varved sediments and annually laminated sediments are now used as synonyms.

Varve Formation and Preservation

Varved sediments are created primarily through the interplay of two processes: seasonal variations in the genesis, transport, and deposition of sediments, and the preservation of that signal after burial. A strong seasonal cycle that influences sedimentation is the fundamental process controlling varve deposition. In most settings, this is not a limiting factor because a dominant seasonal climate cycle with large impacts on the environment, particularly the hydrological cycle, can be found nearly everywhere. In temperate climates, for example, surface water runs freely in summer and freezes over in winter, greatly varying the amount of energy available for clastic sediment transport. Similarly, in the tropics, monsoons create dramatically different runoff during wet and dry seasons. Of course, biology also follows seasonal cycles, and this influences sedimentation. Diatoms in freshwater lakes bloom during spring and summer and die out in winter, and production based on marine upwelling may fluctuate seasonally following movements of zonal winds. Such manifestations of seasonality result in different sedimentary regimes, delivering biological or mineralogical material to the sediment surface at different times of year as distinct layers (e.g., Cook et al., 2009; Francus et al., 2008; Ojala et al., 2000).

Seasonal sediment layers are produced and deposited in many environments, but are usually disturbed by bioturbation from benthic organisms, as well as sediment mixing and resuspension due to slumping and current action. In order for varves to be preserved, the fragile seasonal layers must remain intact after deposition. In lakes, morphometry of the basin is one of the most important features controlling laminae preservation, as it affects the potential for sediment disturbance. A relationship exists between lake area and depth in those sites with preservation of laminated sediments (O'Sullivan, 1983). In general, very deep lakes or lakes with small surface area that have great *relative* depth will avoid deep wind mixing and thus resuspension of surface sediments. Basin morphometry is also critical with respect to another important aspect controlling varve preservation – oxygenation of bottom water. Deep lakes allow development of strong thermal stratification, reducing ventilation of bottom water. Oxidation of organic matter falling through the water column then depletes available oxygen and the deep waters become anoxic, preventing the growth of benthic organisms and the influence of bioturbation. Physical isolation of bottom waters can also occur in the ocean, in shallow-silled marine basins formed as deep fjords such as

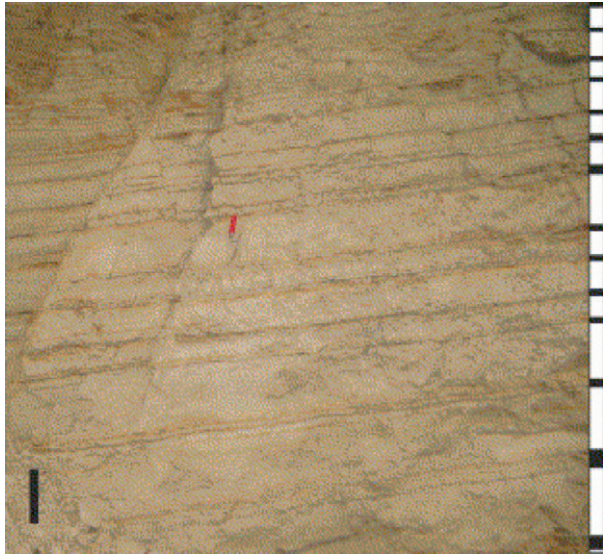


Figure 1 Pleistocene glacial varves cropping out at Eberswalde–Macherslust, Brandenburg (Germany). Note the decreasing varve thickness toward the top of the section as a result of increasing distance from the active ice margin. The scale bar is 30 cm. Photograph by B. Zolitschka.

the Saanich Inlet, British Columbia (Blais-Stevens et al., 1997) or as tectonic basins such as the Santa Barbara Basin, California (Behl and Kennett, 1996), Cariaco Basin, Venezuela (Peterson et al., 2000), and Guaymas Basin, Gulf of California (Calvert, 1966). Water column stratification occurs due to chemical as well as thermal processes, such as the juxtaposition of water masses with different salinity. This situation can be found in coastal tidewater lakes where saltwater introduced at high tides is overlain by a freshwater cap from continual stream input (Hughen et al., 1996a). Dense, hypersaline bottom waters can also form from dissolution of exposed salt formations on the seafloor, as is found in the Orca Basin, Gulf of Mexico (Kennett et al., 1985). Although not absolutely required for varve formation, anoxic bottom waters are one of the best indicators of potentially laminated sediments and are often used to help guide coring expeditions. For example, anoxic bottom waters occur where the oxygen minimum zone intersects the seafloor on the continental slopes, and annually laminated sediments have been located and studied off the coast of Pakistan (Schultz et al., 1998) as well as California (Anderson et al., 1987) and Peru (De Vries and Schrader, 1981).

Classification and Morphology

For convenience, varves can be classified into basic types according to the most prominent seasonal feature or layer components. The earliest recognized and most commonly studied varved sediments are *clastic varves*, composed primarily of allochthonous mineral grains that differ seasonally in grain size (De Geer, 1912; O'Sullivan, 1983; Saarnisto, 1986). Clastic varves are common in glacial marine and lacustrine settings where mineral grain production and deposition rates are high, and biological productivity is low. A magnified image of clastic

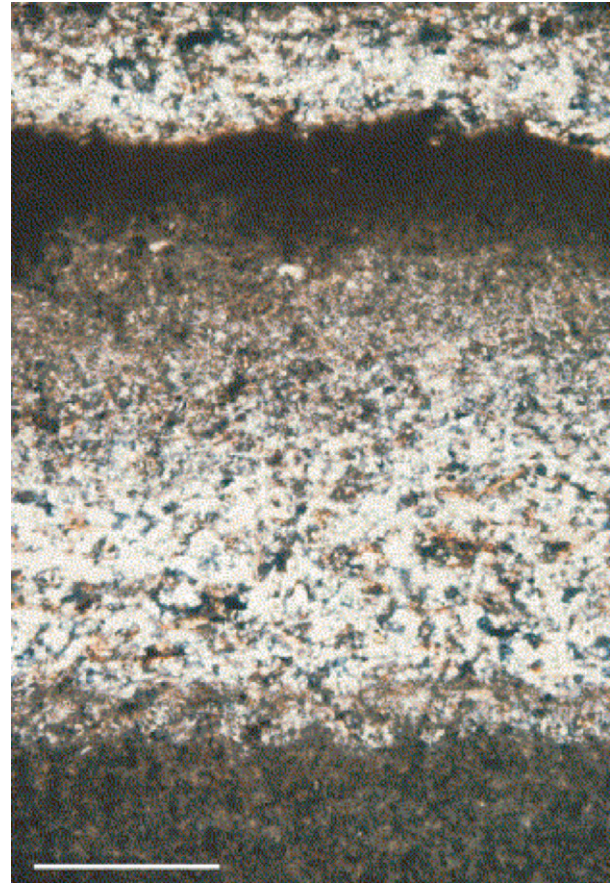


Figure 2 Photomicrograph of clastic varve couplets from glacial Donard Lake, Baffin Island, core 95DON-05, 47 cm. The sediment surface is toward the top of the photo. Scale bar is 1 mm. Photograph by K. Hughen.

varves from an Arctic lake (Figure 2) illustrates typical sedimentary structures and fabric in this type of varve. The sediments consist of couplets of a light-colored, coarser-grained layer covered by a thin dark layer. Grains as large as fine sand mark the bottom of the coarse layer and grade upward into silt-sized particles, whereas a dark layer of fine clays caps the top of each couplet (Moore et al., 2001). The silts grade gradually upward into the fine clays, but the boundary between the clays and the overlying coarse grains is typically sharp (Figure 2).

The succession of fine sands to silts in a thick layer, topped by a thin clay cap, is the classic structure of glacial varves (Cockburn and Lamoureux, 2008; De Geer, 1912; Francus et al., 2008; Leemann and Neissen, 1994; Moore et al., 2001; O'Sullivan, 1983; Retelle and Child, 1996; Saarnisto, 1986; Wohlfarth et al., 1998; Zolitschka, 1996a). The upward gradation from relatively coarse to fine grains in the thick layer results from a decrease in the energy available for sediment transport during the summer months. During the spring meltwater freshet, runoff is at a maximum and sands and silts previously locked up in winter ice are first available to be transported. Glacial runoff continues throughout the summer, but with snow cover mostly melted, the flux of runoff and energy of transport are diminished, and the size of grains washed into the basin is reduced (Hardy, 1996; Retelle and

Child, 1996). During winter in glacial environments, surface water freezes and transport of sediment is greatly reduced or ceases altogether. In the absence of continual mixing, fine, clay-sized particles suspended in the water column settle out and form an independent dark clay layer. This idealized sequence can be interrupted by abrupt events such as intense rainstorms during the late summer or fall, when an increased pulse of runoff from a rainstorm can result in a sublaminae of larger grains, often recognizable as an anomalous sand lens (Hardy, 1996; Lamoureux, 1999; Lamoureux et al., 2006; Moore et al., 2001; Retelle and Child, 1996). Other sublaminae structures may result from ice rafting or other processes (Tomkins et al., 2009). However, the most diagnostic part of the sediment fabric is the clay cap, usually used to distinguish consecutive years' accumulation.

Another widely studied type of varved sediments are termed *biogenic* (O'Sullivan, 1983; Saarnisto, 1986) or *biochemical varves* (Lotter, 1991). Biogenic varves form due to distinct pulses of aquatic biological productivity, including seasonal plankton blooms (Ojala et al., 2000). The biological layer may interrupt and temporarily dilute a more constant rate of background deposition, such as mineral grains from steady stream flow or eolian input, or else alternate with a different distinct pulse of sedimentation, for example, mineral grains from monsoonal river runoff. A magnified image of biogenic varves from a coastal Arctic tidewater lake illustrates an example of these seasonal layers (Figure 3). The biogenic layers form during the ice-free summer months, and during the winter, suspended particles settle out of the frozen, low-energy environment and are deposited as dark-colored, clay-rich layers (Hughen et al., 1996a, 2000a). Biogenic laminae form in a variety of marine and lacustrine environments, and in many cases a succession of distinct sublaminae can be deposited as different diatom species bloom throughout the growth season (Dean et al., 1999).

Varves may also form through seasonal changes in the physical and chemical properties of the water column, and are usually restricted to lakes. *Calcareous varves* may form in lakes with high CO_3^{2-} content, when calcite layers precipitate out of solution during summer warming (Lotter, 1991; O'Sullivan, 1983). *Ferrogenic varves* may form due to changes in redox conditions affecting the solubility of Fe species and the formation of sulfides versus hydroxides throughout the year (O'Sullivan, 1983). *Evaporite varves* may form as interbedded laminae of precipitates in shallow evaporite basins, and can consist of couplets of calcite and anhydrite (e.g., Castile Formation; Anderson and Dean, 1995), or aragonite with eolian grains (e.g., Lake Lisan; Katz and Kolodny, 1989), depending on water chemistry.

Analytical Techniques

Sampling and Image Analysis

The study and use of varves in geochronology requires special techniques for the recovery and analysis of fragile unconsolidated sediments. The sediment–water interface must be recovered intact to provide a baseline for dating, as well as calibrating the annual climate signal. Specialized sediment-coring methods have been developed for the intact recovery

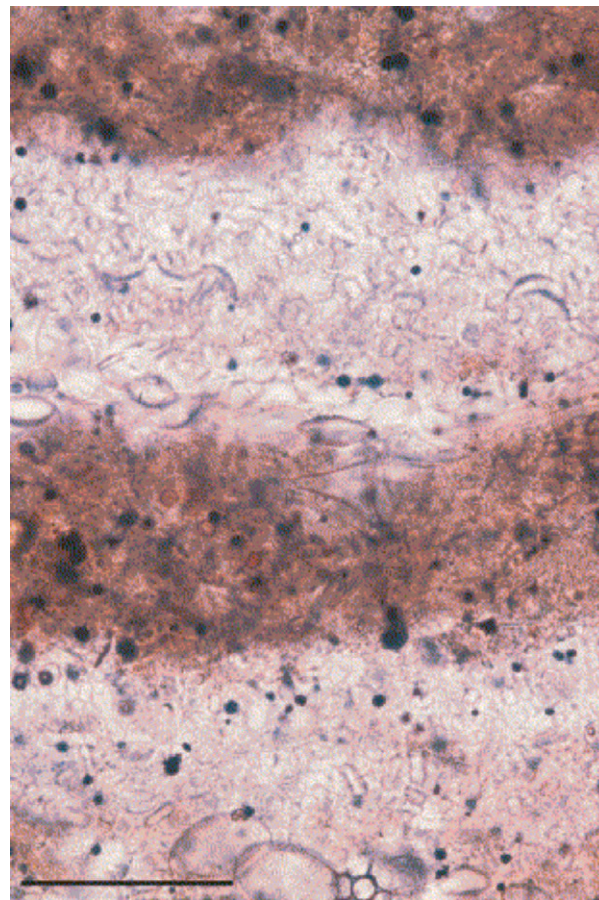


Figure 3 Photomicrograph of biogenic varve couplets from Ogaq Lake, Baffin Island, core 91-2b, 73.5 cm. The sediment surface is toward the top of the photo. Scale bar is 100 μm . Photograph by K. Hughen.

and preservation of fine laminated structures and the sediment–water interface in soupy sediments. One method is freeze coring, in which a hollow tube filled with a dry ice–alcohol slurry is inserted into soft sediments (Shapiro, 1958). The freeze corer is left in place long enough to allow the sediments and interstitial water to freeze onto the outside of the corer, perfectly preserving the sediment–water interface and uppermost, porous ($\geq 99\%$ water) laminae (Hughen et al., 1996a). Freeze coring requires intensive logistics for bringing large quantities of dry ice into the field and maintaining the cores in a frozen state during transport back to the lab. In some sediments, freezing causes ice crystal disturbance of the fine laminae microstructure, rendering the technique unsuitable for high-magnification analyses (Pike and Kemp, 1996).

Another coring technique successfully used to recover the intact sediment–water interface involves the careful recovery and subsampling of dredges or box cores (e.g., Zolitschka, 1996a), or else the use of a closed-top, miniature gravity corer such as the 'Glew' corer (Glew, 1991). These techniques allow the researcher to raise fragile laminated sediments unfrozen to the surface, but require special handling to ensure that the sediments remain undisturbed during sampling or, much more difficult, transport back to the lab. Typically, this requires carefully siphoning off overlying water right down to

the sediment surface. The cores must be left upright and undisturbed for a sufficient time to allow them to dewater, and the siphoning repeated until no further dewatering is observed. Although time intensive, this method offers an additional advantage over freeze coring in that the effects of porosity are largely removed, whereas the freezing of layers *in situ* requires later corrections for decreasing porosity with depth when making layer thickness or mass flux measurements (Hughen et al., 2000a).

Following successful recovery of varved sediment cores, several techniques are available for investigating the annual laminations, including obtaining high-resolution images of laminae and their internal microfabric and composition. Initially, split core surfaces (or varves exposed in outcrops) may be investigated using noninvasive techniques such as photography, digital imaging, or x-radiography (Schaaf and Thurow, 1994; Zolitschka, 1996b). These techniques are often used in combination to overcome limitations inherent to single methods (e.g., blurring in x-rays due to three-dimensional distortion of laminations). Digital imaging, whether digitization of existing photographs and x-ray film or directly from the sediment surface, allows sophisticated image analysis techniques to aid in the identification of fine laminae (Schaaf and Thurow, 1994; Schultz et al., 1998; Zolitschka, 1996b). For an in-depth review of digital imaging techniques for sediment analysis, see Francus (2004) and articles therein.

Microscopic analysis of unconsolidated sediments requires that the sediments be embedded in epoxy resin and prepared into petrographic thin sections. There are numerous embedding methods used to remove water from microscopic pore spaces and replace it with resin, without disturbing sediment microfabric. The two principle techniques involve either using acetone to initially displace the water, and then resin to replace the acetone (fluid displacive – Clark, 1988; Lamoureux, 1994; Pike and Kemp, 1996), or freeze-drying the sediment to remove water, and then adding resin to the dried sediment block (Hughen et al., 1996a). Freeze-drying is a fast and economical way to prepare sediments for embedding. However, depending on sediment type, freeze-drying can allow minor disturbance of grains at the intralaminae scale and disruption of internal laminae fabrics. Fluid displacement is the least disruptive embedding technique and is recommended for analyses involving internal laminae fabric and *in situ* composition. For detailed reviews of sediment preparation and analytical techniques for varved sediment studies, the reader is referred to Lamoureux (1994), Pike and Kemp (1996), and Dean et al. (1999).

Individual laminations down to mm-scale can usually be discerned using unmagnified images. In order to construct reliable varve chronologies, however, care must be taken to avoid misidentification of 'event deposits' such as turbidites or rainstorm sediment pulses as an entire year's deposition. Such event deposits often possess characteristics of composition or fabric that distinguish them from typical varve couplets (or triplets), including homogenous structure (lacking sublaminae), mixed composition of normally distinct laminae types, anomalous grain size and sorting, and many others (Dean et al., 1999; Hughen et al., 1996b; Lamoureux, 1999). Discrimination of fabrics and compositional elements at the intralamina scale requires powerful techniques for high-resolution and high-magnification sediment analysis, primarily

optical and electron microscopy. Transmitted-light optical microscopy can reveal details of fabric such as sublaminae and grading, as well as grain size, sorting, and the nature of laminae boundaries (Hughen et al., 1996b; Lamoureux, 1999; Pike and Kemp, 1996). For the highest-resolution analyses, scanning electron microscopy, and back-scattered electron imagery in particular (Kemp, 1996), provide the best technique for determination of laminae and sublaminae fabric and structure, as well as details of composition at the scale of individual mineral grains. For example, back-scattered electron imagery can be used to identify the seasonal sequence of diatom species as recorded in distinct sublaminae within 1 year's deposition (Pike and Kemp, 1996). Such high-resolution studies provide critical insights aiding in the interpretation of laminated sediments, both for chronological and paleoclimatic applications. Recent advances in sediment analysis have been introduced by Francus et al. (2002) who analyzed individual clastic grains for size and geometry, providing a more precise indicator of energy of runoff, air temperature, etc.

Independent Dating and Age Control

Before laminated sediments can be counted and used to construct calendar-age chronologies, it must be demonstrated using independent dating techniques that the laminae couplets are indeed annually deposited varves. Simply identifying and counting laminae couplets can lead to misinterpretation of event deposits as seasonal layers, or the counting of layers representing seasonal processes, but ones that do not occur on a regular basis from year to year. Numerous studies have used independent dating techniques to demonstrate that supposed 'varves' were not in fact annually deposited (e.g., Crusius and Anderson, 1992). Whenever possible, a combination of several dating methods should be used both to increase confidence and to provide temporal coverage over a broader range of time scales. Lead-210 is a radioactive isotope resulting from U-series decay with a relatively short half-life (22.3 years). Unsupported ^{210}Pb falls out from the atmosphere (in contrast to supported ^{210}Pb that comes from continuous, *in situ* uranium decay within the buried sediments) and is deposited globally in lake and marine sediments (Appleby and Oldfield, 1978). Excess (unsupported) ^{210}Pb concentrations can be measured precisely by decay-counting methods (Hughen et al., 1996a), providing age control over the past ~150 years. Measurements must be made deep enough to reach background, supported ^{210}Pb levels, and the ^{210}Pb activity should show an exponential decay (Figure 4). These criteria reveal a simple depositional regime (i.e., no turbidites, erosional unconformities, or sudden shifts in sedimentation rate) that can be used to calculate sedimentation rates. A comparison of ^{210}Pb and varve-based sedimentation rates should reveal good agreement in slope, supporting the interpretation of the laminations as annual (Figure 4).

In some cases, the ^{210}Pb dating method may lack the precision necessary to adequately constrain annual varve chronologies. Another technique widely used to precisely date sediments with near-annual resolution is to measure nuclides produced by atmospheric nuclear weapons testing and nuclear reactor accidents. Common nuclides used to identify the 'bomb spike' resulting from the peak of weapons testing in

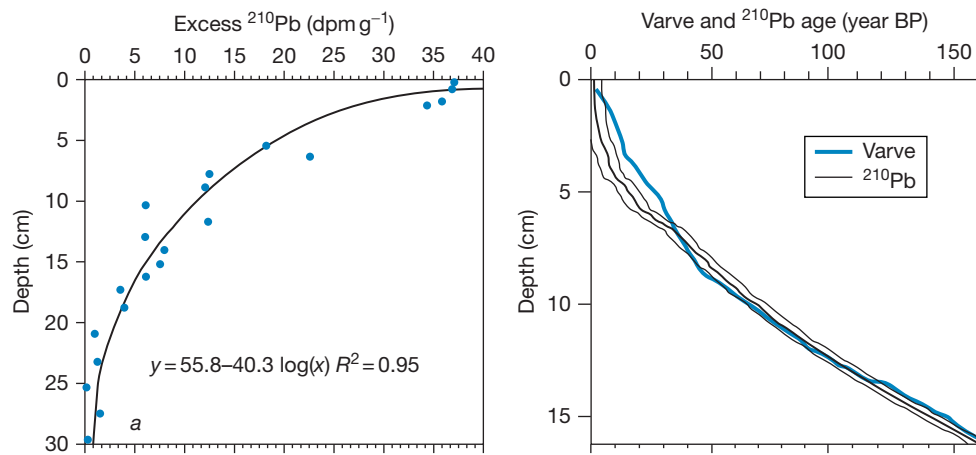


Figure 4 ^{210}Pb dating method used to confirm varve chronology in sediment core 93-7, Upper Soper Lake, Baffin Island. Excess ^{210}Pb shows good agreement with exponential curve and reaches a plateau at background, supported levels, indicating a simple sedimentation regime without hiatuses or large changes in sedimentation rate. ^{210}Pb ages show good agreement with varve ages versus depth, supporting the hypothesis that the laminae couplets are annually deposited varves. Redrawn from Huguen KA, Overpeck JT, and Anderson RF (2000) Recent warming in a 500-year paleotemperature record from varved sediments, Upper Soper Lake, Baffin Island, Canada. *The Holocene* 10: 9–19.

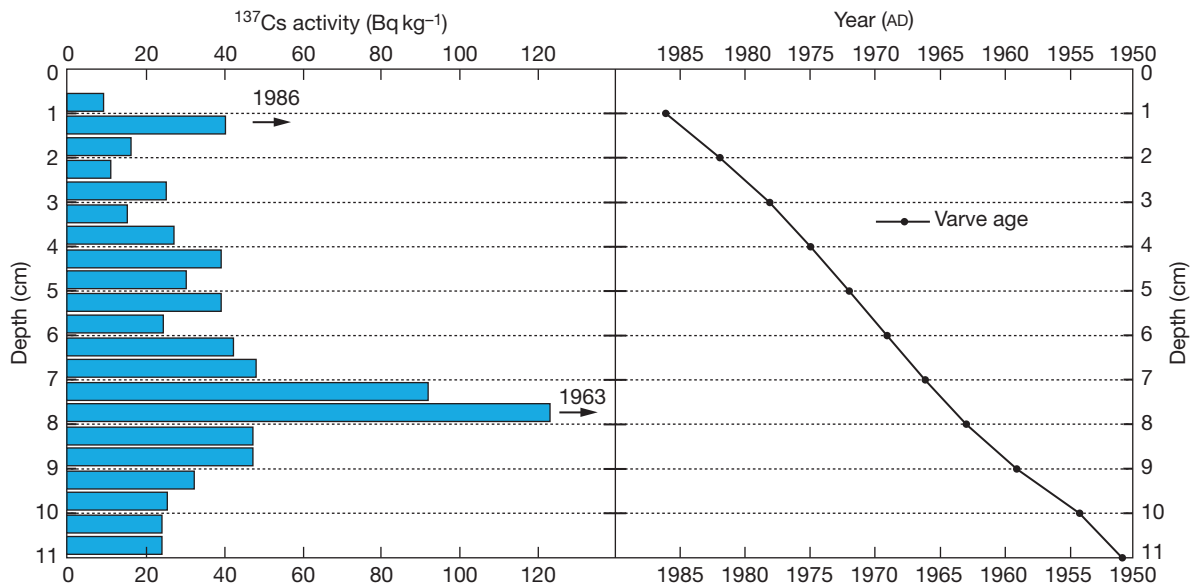


Figure 5 ^{137}Cs dating method used to confirm varve chronology for sediment core NL2, Nicolay Lake, Cornwall Island. Continuous ^{137}Cs measurements show distinct peaks from nuclear bomb testing in 1963 and the Chernobyl reactor accident in 1986. Close agreement between these known age horizons and varve ages with depth supports the interpretation of the laminated sediments as annual varves. Redrawn from Lamoureux SF (1999) Catchment and lake controls over the formation of varves in monomictic Nicolay Lake, Cornwall Island, Nunavut. *Canadian Journal of Earth Sciences* 36: 1533–1546.

1963 are cesium-137 (^{137}Cs) and plutonium ($^{139+140}\text{Pu}$; Bis-caye et al., 1988). In addition to the bomb-spike, ^{137}Cs from the Chernobyl reactor accident in 1986 can be identified in sediments across the globe (Lamoureux, 1999). In order to test a varve chronology, a high-resolution sequence of ^{137}Cs or Pu measurements must be used to identify the known activity peaks in 1963 and 1986, and the depths of these time horizons compared to the corresponding depths in the varve chronology (Figure 5). Although it is possible to calculate sedimentation rates without obtaining the uppermost sediments for ^{210}Pb dating, for this ^{137}Cs (Pu) method, it is critical to recover a

sediment time marker of absolute known age, usually the undisturbed sediment–water interface. Because of the large amplitude of the production spikes, the main limitation to the precision of the ^{137}Cs and Pu methods is the sampling resolution and sample size needed for measurements.

Chronology Construction

By their nature, varves are a result of changing sediment supply to a depositional basin, and should be present and recognizable throughout that basin. Subtle disturbances may disrupt or

distort the varve sequence recovered in single sediment cores, however, and numerous steps are necessary to create a reliable annual chronology from laminated sediments. In constructing a varve chronology, many of the principles used in dendrochronology must be applied. For example, multiple measurements of varves (tree rings) from a single core (tree) must be averaged, followed by cross-correlation and averaging between multiple cores to build a chronology representative of an entire sediment basin (stand of trees). Small fluctuations in varve thickness as seen in thin sections require that each lamination be measured several times and averaged to provide an objective, representative record from each sediment core (Zolitschka, 1996a). In addition, local disturbances such as turbidites and erosional scours, as well as core breaks or gaps arising from the sampling process, may limit the physical integrity of a record from any single sediment core. To minimize these sources of error, it is critical to correlate varve chronologies constructed from multiple different cores and produce a basin-wide average (Hughen et al., 2000a; Zolitschka, 1996a). Downcore patterns of variation in varve thickness are usually quite distinct and can be recognized in cores across a basin (Figure 6). Cross-correlation of these marker patterns allow the identification and bridging of missing or disturbed sections in single cores. A larger number of cores used for cross-correlation increases the

likelihood of avoiding all core gaps and constructing an accurate and reliable varve chronology.

Varve Applications

Geochronology

The geochronological applications of varved sediments are too numerous to provide an exhaustive list. One significant application that should be emphasized is the calibration and development of other geochronological tools, specifically, calibration of the radiocarbon time scale. Varves from lakes and marine basins have been one of the primary tools used to provide calendar-age calibration of ^{14}C ages beyond the limit of tree rings, >12 000 years ago (Hughen et al., 2000b). Annual calendar-age varve chronologies also provide accurate dates for specific sedimentological events, as well as age models for continuous sediment time series. An example of a study investigating discrete events would be paleoseismology. Earthquakes cause turbidity flows and turbidite deposits that in the proper setting can be dated by counting overlying varves (Hughen et al., 1996b). Interbedded sequences of varved sediments and massive turbidites can be used to determine recurrence intervals of earthquakes in a specific region (Blais-Stevens et al., 1997). Varved sediments have been used to date other types of episodic events such as hurricanes (Besonen et al., 2008) and jökulhlaups (Lewis et al., 2009).

Varve chronologies offer an excellent means for constructing age models for high-resolution paleoclimate records (Augustsson et al., 2010; Enters et al., 2010; Rolland et al., 2009) as well as other time series such as paleogeomagnetic variations (Ojala and Tiljander, 2003; Snowball et al., 2007). Varves are based on seasonal cycles, similar to tree rings and ice-core layers, and do not require calibration in order to provide calendar ages. Varves are therefore well-suited to studies comparing the relative timing and lead-lag relationships of rapid climate changes in different regions, based on high-resolution paleoclimate records with calendar-age chronologies from varved sediments, tree rings, and ice cores (Hughen et al., 2000b). In addition, varve chronologies are built up from fine annual layers, and do not need to be interpolated between infrequent datums such as radiocarbon dates. This provides increased precision at the annual scale and is particularly well suited for studies involving spectral analysis at short (subdecadal) periodicities. Several studies have recently used varved sediments to investigate the past periodicities and frequency behavior of natural systems such as the El-Niño Southern Oscillation, and the North Atlantic Oscillation (Godsey et al., 1999; Rittenour et al., 2000).

Applications Other than Dating

In addition to having geochronological applications, varved sediments may themselves contain valuable paleoenvironmental information. For example, hydrological and sedimentological process studies in glacier-fed lakes have shown that the fluxes of runoff and suspended sediment, as well as varve thickness, vary logarithmically as a function of average temperature during the snow or glacier melting season (Hardy, 1996; Leemann and Neissen, 1994; Retelle and Child, 1996).

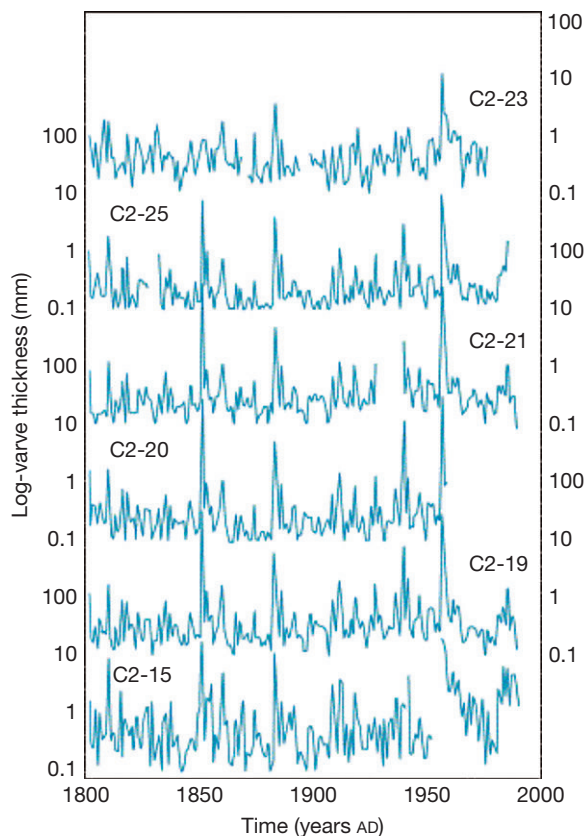


Figure 6 Lamination thickness variations in six cores from Lake C2, Ellesmere Island. Data represent five individual measurements of each lamina from thin sections. Gaps in the records indicate missing or disturbed laminations. Redrawn from Zolitschka B (1996) Recent sedimentation in a high arctic lake, northern Ellesmere Island, Canada. *Journal of Paleolimnology* 16: 169–186.

Downcore records of clastic varve or clastic laminae thickness from both glacial and nonglacial lakes correlate well with average summer or snowmelt season temperatures (Hardy et al., 1996; Hughen et al., 2000a; Leemann and Neissen, 1994; Moore et al., 2001; Wohlfarth et al., 1998) and can be used for annually resolved reconstructions of paleotemperature. Other studies have correlated specific forms of sublaminae with runoff and precipitation (Lamoureux, 1999), providing proxies for long reconstructions of hydrological variability (Lamoureux et al., 2006). Before varve thickness signals in any new location are used as paleotemperature and paleohydrological proxies, it is critical that they are calibrated for the recent period using instrumental data, whether from a nearby meteorological station or temporary measurements from process studies.

Varves have also been used to reconstruct changes in other paleoenvironmental parameters. Biogenic laminae in marine upwelling regimes reflect primary productivity, and therefore the intensity of upwelling and trade winds, in the tropical North Atlantic during the last deglaciation (Hughen et al., 1996c). Laminated sediments, and their absence, have been used to reconstruct the ventilation history of intermediate-depth waters off the coast of California (Behl and Kennett, 1996), Pakistan (Schultz et al., 1998), and northern South America (Peterson et al., 2000). These studies and others have shown that abrupt climatic shifts first seen in the high latitude North Atlantic region were in fact global in scope. Finally, varves can be used to study changes in past glacial activity and evolution of sediment source regions, as well as evolution of sedimentary basins (Lamoureux, 1999). The applications of varved sediments to geochronological and paleoenvironmental research are limitless, and will continue to grow as increasingly sophisticated methods are developed for obtaining and analyzing them.

See also: Varved Lake Sediments. **Ice Core Records:** Correlations Between Greenland and Antarctica. **Paleoclimate Reconstruction:** The Last Millennium. **Radiocarbon Dating:** Calibration of the ^{14}C Record.

References

- Anderson RY and Dean WE (1995) Filling the Delaware Basin: Hydrologic and climatic controls on the Permian Castile Formation varved evaporite. In: Scholle PA, et al. (ed.) *The Permian of Northern Pangea 2*, pp. 61–78. New York: Springer.
- Anderson RY, Hamphill-Halley E, and Gardner JV (1987) Persistent late Pleistocene–Holocene seasonal upwelling and varves of the coast of California. *Quaternary Research* 28: 307–313.
- Appleby PG and Oldfield F (1978) The calculation of lead-210 dates assuming a constant rate of supply of unsupported ^{210}Pb to the sediment. *Catena* 5: 1–8.
- Augustsson A, Peltola P, Bergbäck B, Saarinen T, and Haltia-Hovi E (2010) Trace metal variability during 5500 years in a lake sediment from eastern Finland. *Journal of Paleolimnology* 44: 1025–1038. <http://dx.doi.org/10.1007/s10933-010-9471-z>.
- Behl RJ and Kennett JP (1996) Brief interstadial events in the Santa Barbara Basin, NE Pacific, during the past 60 kyr. *Nature* 379: 243–246.
- Besonen MR, Bradley RS, Mudelsee M, Abbott MB, and Francus P (2008) A 1,000-year, annually-resolved record of hurricane activity from Boston, Massachusetts. *Geophysical Research Letters* 35: L14705. <http://dx.doi.org/10.1029/2008GL033950>.
- Biscaye PE, Anderson RF, and Deck BL (1988) Fluxes of particles and constituents to the eastern U.S. continental slope and rise: SEEP-1. *Continental Shelf Research* 8: 855–904.
- Blais-Stevens A, Clague JJ, Bobrowsky PT, and Patterson RT (1997) Late Holocene sedimentation in Saanich Inlet, British Columbia, and its paleoseismic implications. *Canadian Journal of Earth Sciences* 34: 1345–1357.
- Calvert SE (1966) Origin of diatom-rich, varved sediments from the Gulf of California. *Journal of Geology* 74: 546–565.
- Clark JS (1988) Stratigraphic charcoal analysis on petrographic thin sections: Applications to fire history in northwestern Minnesota. *Quaternary Research* 30: 81–91.
- Cockburn JMH and Lamoureux SF (2008) Inflow and lake controls on short-term mass accumulation and sedimentary particle size in a High Arctic lake: Implications for interpreting varved lacustrine sedimentary records. *Journal of Paleolimnology* 40: 923–942. <http://dx.doi.org/10.1007/s10933-008-9207-5>.
- Cook TL, Bradley RS, Stoner JS, and Francus P (2009) Five thousand years of sediment transfer in a high arctic watershed recorded in annually laminated sediments from Lower Murray Lake, Ellesmere Island, Nunavut, Canada. *Journal of Paleolimnology* 41: 77–94. <http://dx.doi.org/10.1007/s10933-008-9252-0>.
- Crusius J and Anderson RF (1992) Inconsistencies in accumulation rates of Black Sea sediments inferred from records of laminae and ^{210}Pb . *Paleoceanography* 7: 215–227.
- De Geer G (1912) A Geochronology of the past 12,000 years. *Congres de Geologie International, Comptes Rendues 11*, Stockholm, pp. 241–253.
- De Vries TJ and Schrader H (1981) Variation of upwelling/oceanic conditions during the latest Pleistocene through Holocene off the central Peruvian coast: A diatom record. *Marine Micropaleontology* 6: 157–167.
- Dean JM, Kemp AES, Bull D, Pike J, Patterson G, and Zolitschka B (1999) Taking varves to bits: Scanning electron microscopy in the study of laminated sediments and varves. *Journal of Paleolimnology* 22: 121–136.
- Enters D, Kirilova E, Lotter AF, et al. (2010) Climate change and human impact at Sacrower See (NE Germany) during the past 13,000 years: A geochemical record. *Journal of Paleolimnology* 43: 719–737.
- Francus P (2004) *Image Analysis, Sediments and Paleoenvironments. Developments in Paleoenvironmental Research*, vol. 7, p. 330. Dordrecht, Pays-Bas: Springer. ISBN: 1-4020-2061-9.
- Francus P, Bradley RS, Abbott M, Patridge W, and Keimig F (2002) Paleoclimate studies of minerogenic sediments using annually resolved textural parameters. *Geophysical Research Letters* 29: 59–1–59–4.
- Francus P, Bradley R, Lewis T, Abbott M, Retelle M, and Stoner JS (2008) Limnological and sedimentary processes at Sawtooth Lake, Canadian High Arctic, and their influence on varve formation. *Journal of Paleolimnology* 40: 963–985. <http://dx.doi.org/10.1007/s10933-008-9210-x>.
- Glew JR (1991) Miniature gravity corer for recovering short sediment cores. *Journal of Paleolimnology* 5: 285–287.
- Godsey HS, Moore TC Jr., Rea DK, and Shane LCK (1999) Post-Younger Dryas seasonality in the North American Midcontinent region as recorded in Lake Huron varved sediments. *Canadian Journal of Earth Sciences* 36: 533–547.
- Hardy DR (1996) Climatic control of stream flow and sediment flux into Lake C2, northern Ellesmere Island. *Journal of Paleolimnology* 16: 133–149.
- Hardy DR, Bradley RS, and Zolitschka B (1996) The climatic signal in varved sediments from Lake C2, northern Ellesmere Island, Canada. *Journal of Paleolimnology* 16: 227–238.
- Hughen KA, Overpeck JT, and Anderson RF (2000a) Recent warming in a 500-year palaeotemperature record from varved sediments, Upper Soper Lake, Baffin Island, Canada. *The Holocene* 10: 9–19.
- Hughen KA, Overpeck JT, Anderson RF, and Williams KM (1996a) The potential for paleoclimatic records from varved Arctic lake sediments: Baffin Island, Eastern Canadian Arctic. In: Kemp AES (ed.) *Palaeoclimatology and Palaeoceanography from Laminated Sediments*, pp. 57–71. London: Geological Society. Special Publication 116.
- Hughen KA, Overpeck JT, Peterson LC, and Anderson RF (1996b) The nature of varved sedimentation in the Cariaco Basin, Venezuela, and its palaeoclimatic significance. In: Kemp AES (ed.) *Palaeoclimatology and Palaeoceanography from Laminated Sediments*, pp. 171–183. London: Geological Society. Special Publication 116.
- Hughen KA, Overpeck JT, Peterson LC, and Trumbore S (1996c) Rapid climate changes in the tropical Atlantic region during the last deglaciation. *Nature* 380: 51–54.
- Hughen KA, Southon JR, Lehman SJ, and Overpeck JT (2000b) Synchronous radiocarbon and climate shifts during the last deglaciation. *Science* 290: 1951–1954.
- Katz A and Kolodny N (1989) Hypersaline brine diagenesis and evolution in the Dead Sea-Lake Lisan system (Israel). *Geochimica et Cosmochimica Acta* 53: 59–67.
- Kemp AES (ed.) (1996) *Palaeoclimatology and Palaeoceanography from Laminated Sediments*. London: Geological Society. Special Publication 116.

- Kennett JP, Elmstrom KM, and Penrose N (1985) The last deglaciation in Orca Basin, Gulf of Mexico: High-resolution planktonic foraminiferal changes. *Palaeogeography, Palaeoclimatology, Palaeoecology* 50: 189–216.
- Lamoureux SF (1994) Embedding unfrozen lake sediments for thin section preparation. *Journal of Paleolimnology* 10: 141–146.
- Lamoureux SF (1999) Catchment and lake controls over the formation of varves in monomictic Nicolay Lake, Cornwall Island, Nunavut. *Canadian Journal of Earth Sciences* 36: 1533–1546.
- Lamoureux SF, Stewart KA, Forbes AC, and Fortin D (2006) Multidecadal variations and decline in spring discharge in the Canadian middle Arctic since 1550 AD. *Geophysical Research Letters* 33: L02403. <http://dx.doi.org/10.1029/2005GL024942>.
- Leemann A and Neissen F (1994) Varve formation and the climatic record in an Alpine proglacial lake: Calibrating annually-laminated sediments against hydrological and meteorological data. *The Holocene* 4: 1–8.
- Lewis T, Francus P, and Bradley RS (2009) Recent occurrence of large jökulhlaups at Lake Tuborg, Ellesmere Island, Nunavut. *Journal of Paleolimnology* 41: 491–506.
- Lotter AF (1991) Absolute dating of the Late-Glacial period in Switzerland using annually laminated sediments. *Quaternary Research* 35: 321–330.
- Moore JJ, Kughen KA, Miller GH, and Overpeck JT (2001) Little Ice Age recorded in summer temperature reconstruction from varved sediments of Donard Lake, Baffin Island, Canada. *Journal of Paleolimnology* 25: 484–499.
- O'Sullivan PE (1983) Annually-laminated lake sediments and the study of Quaternary environmental changes: A review. *Quaternary Science Reviews* 1: 245–313.
- Ojala AEK, Saarinen T, and Salonen V-P (2000) Preconditions for the formation of annually laminated lake sediments in southern and central Finland. *Boreal Environment Research* 5: 243–255.
- Ojala AEK and Tiliander M (2003) Testing the fidelity of sediment chronology: Comparison of varve and paleomagnetic results from Holocene lake sediments from central Finland. *Quaternary Science Reviews* 22: 1787–1803.
- Peterson LC, Haug GH, Hughen KA, and Roehl U (2000) Rapid changes in the hydrologic cycle of the tropical Atlantic during the last glacial. *Science* 290: 1947–1951.
- Pike J and Kemp AES (1996) Preparation and analysis techniques for studies of laminated sediments. In: Kemp AES (ed.) *Palaeoclimatology and Palaeoceanography from Laminated Sediments*, pp. 37–48. London: Geological Society. Special Publication 116.
- Retelle MJ and Child JK (1996) Suspended sediment transport and deposition in a high arctic meromictic lake. *Journal of Paleolimnology* 16: 151–167.
- Rittenour TM, Brigham-Grette J, and Mann ME (2000) El Niño-like climate teleconnections in New England during the late Pleistocene. *Science* 288: 1039–1042.
- Rolland N, Larocque I, Francus P, Pienitz R, and Laperrière L (2009) Evidence for a warmer period during the 12th and 13th centuries AD from chironomid assemblages in Southampton Island, Nunavut, Canada. *Quaternary Research* 72: 27–37. <http://dx.doi.org/10.1016/j.yqres.2009.03.001>.
- Saarnisto M (1986) Annually laminated lake sediments. In: Berglund BE (ed.) *Handbook of Holocene Palaeoecology and Palaeohydrology*, pp. 343–370. Chichester: John Wiley & Sons.
- Schaaf M and Thurow J (1994) A fast and easy method to derive highest-resolution time series data sets from drill cores and rock samples. *Sedimentary Geology* 94: 1–10.
- Schultz H, von Rad U, and Erlenkeuser H (1998) Correlation between Arabian Sea and Greenland climate oscillations of the past 110,000 years. *Nature* 393: 54–57.
- Shapiro J (1958) The core-freezer – A new sampler for lake sediments. *Ecology* 39: 748.
- Snowball I, Zillen L, Ojala A, Saarinen T, and Sandgren P (2007) FENNOSTACK and FENNORPIS: Varve dated Holocene palaeomagnetic secular variation and relative palaeointensity stacks for Fennoscandia. *Earth and Planetary Science Letters* 255: 106–116.
- Tomkins JD, Lamoureux SF, Antoniades D, and Vincent WF (2009) Sedimentary pellets as an ice cover proxy in a High Arctic ice-covered lake. *Journal of Paleolimnology* 41: 225–242. <http://dx.doi.org/10.1007/s10933-008-9255>.
- Wohlfarth B, Holmquist B, Cato I, and Linderson H (1998) The climatic significance of clastic varves in the Angermanälven Estuary, northern Sweden, AD 1860 to 1950. *The Holocene* 8: 521–534.
- Zolitschka B (1996a) Recent sedimentation in a high arctic lake, northern Ellesmere Island, Canada. *Journal of Paleolimnology* 16: 169–186.
- Zolitschka B (1996b) Image analysis and microscopic investigation of annually laminated lake sediments from Fayetteville Green Lake (NY, USA), Lake C2 (NWT, Canada) and Holzmaar (Germany): A comparison. In: Kemp AES (ed.) *Palaeoclimatology and Palaeoceanography from Laminated Sediments*, pp. 49–55. London: Geological Society. Special Publication 116.

Further Reading

- Ojala AEK, Francus P, Zolitschka B, Besonen M, and Lamoureux SF (2012) The chronological characteristics of sedimentary varves – a review. *Quaternary Science Reviews* 43: 45–60.

Relevant Websites

- <http://users.utu.fi/tijusa/> – Annually Laminated Lake Sediments Page, University of Turku, Finland.
- <http://www.pages-igbp.org/workinggroups/varves-wg> – Past Global Changes (PAGES) Varves Working Group.
- <http://www.geo.umass.edu/climate/> – Taconite Inlet Project, University of Massachusetts, Amherst.

Vertebrate Overview

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Historical Background

The study of Quaternary vertebrates has a long-established and illustrious history and is intimately linked both to the discoveries of some of the earliest Paleolithic (old Stone Age) artifacts and to the eventual recognition of the antiquity of the human lineage. The earliest findings of fossil mammals were frequently interpreted as the remains of mythical beasts, such as dragons' (in fact, cave bear (*Ursus spelaeus*)) teeth from the caverns of central Europe, unicorns (fragments of mammoth tusk), and even the Cyclops (fossil dwarf elephant skulls from the Mediterranean islands). One of the oldest publications of Quaternary mammalian remains is that by an English writer, Mr. William Somner, who in 1669 described the finding of a 'sea monster' in 'Chartham News: A Brief Relation of Some Strange Bones There Lately Digged Up in Some Grounds of Mr. John Somner's of Canterbury.' In reality, the report recounted the discovery of three associated upper molars of a woolly rhinoceros (*Coelodonta antiquitatis*), still preserved today more than 300 years later in the Natural History Museum in London. In 1696, the finding of a fossil elephant during quarrying of travertine deposits at the site of Burgtonna in the Grand Duchy of Gotha in Germany engendered a fierce debate between Wilhelm Ernest Tentzel, historian to the dukes of Saxony, who correctly recognized the remains as 'the real bones of an Elephant, but calcined by subterraneous heat, and in a great measure petrified' and the Medical College at Gotha, which dismissed them as either the skeleton of a unicorn or naturally deposited minerals mimicking the shape of an elephant. Tentzel, despite his inspired observations, otherwise subscribed to the prevailing philosophy of the day, which invoked the biblical Deluge as the mechanism for depositing the remains. Around this time, a flint hand axe associated with the remains of an elephant was found near the Grays Inn Road in central London by a Mr. Conyers. The finds were published in 1715 by the antiquarian John Bagford, thereby furnishing the oldest known description of a Paleolithic artifact, although Bagford envisaged the implement to be the weapon of an ancient Briton, used in a fight against the Romans during the Claudian invasion of AD 43.

By the end of the 18th century, the celebrated French anatomist, Georges Cuvier, had instigated a number of comparative anatomical studies of recent and fossil vertebrates and proposed, on the basis of these, the past occurrence of fossil elephants that were similar to but clearly distinct from extant species. The recognition that extinct animals had once coexisted with ancient humans was critical to the succeeding debate concerning the ancestry of the human species. It was in the village of Hoxne in Suffolk (England) in 1797 that John Frere first established the great antiquity of the human lineage. In his famous account to the Society of Antiquaries in London, Frere reported how he had witnessed the discovery of 'weapons of war, fabricated and used by a people who had not the use of metals' (Frere, 1800). The flint hand axes that Frere had observed in the Hoxne brick

pits were associated with fragments of wood and with 'some extraordinary bones, particularly a jawbone of enormous size, of some unknown animal, with the teeth remaining in it.' The significance of Frere's report lay in his perception of the stratigraphical relationship of the deposits overlying the hand axes to the present-day topography of the Waveney valley, with the inevitable conclusion that their deposition must relate to 'a very remote period indeed; even beyond that of the present world.' Unfortunately, his inspired observations were ignored until the 1850 s, when the debate over the alleged association of flint implements with extinct animals and the question of human antiquity was reopened. Taxonomic studies nevertheless continued with the formal naming by Blumenbach in 1799 of two of the best recognized Pleistocene mammals, the woolly mammoth (*Mammuthus primigenius*) and the woolly rhinoceros.

Serious paleontological studies date back to the 19th century, such as the investigations conducted by the Reverend William Buckland, Reader in Mineralogy at Corpus Christi, Oxford (England), and later Dean of Westminster. Buckland correctly recognized that the assemblage of mammalian bones from Kirkdale Cave in North Yorkshire, which included many extinct forms and species no longer native to Britain, had been accumulated by spotted hyenas (*Crocota crocuta*). In his *Reliquiae Diluvianae* of 1823, Buckland assigned the period when the animals had been present in Britain to a time pre-dating the biblical Flood. Later, Buckland embraced the 'glacial theory' to explain geological phenomena formerly attributed to the Flood but steadfastly refused to believe that early humans had been around at the same time as these fossil mammals. A substantial breakthrough on this front was made in the 1830 s with the commencement of excavations at Kent's Cavern and Brixham Cave, both near Torquay in the southwest of England. The excavators, William Pengelly and colleagues, unequivocally demonstrated that flint artifacts had been found in association with the remains of woolly mammoth, lion (*Panthera leo*), and other animals from beneath an unbroken stalagmitic floor (Pengelly et al., 1873). In 1859, reports of exciting new finds from the Somme valley in northern France led the geologist Joseph Prestwich and the antiquarian Sir John Evans to visit the celebrated sites around Abbeville, from where they returned fully convinced that Paleolithic flint tools were indeed contemporary with an extinct fauna. Nearly 60 years after Frere's observations were so widely ignored, Prestwich and Evans revisited Hoxne to confirm his findings, thereby paving the way for a new era in Quaternary vertebrate paleontological research.

Distribution of Quaternary Vertebrate Remains

One of the reasons that vertebrates are so useful in Quaternary studies is that they make robust fossils and are found in an extremely wide range of depositional environments. Although the soft tissues, such as hair, keratin (which forms horns and

claws), and internal organs, are only found under exceptional circumstances, bones and teeth are both remarkably durable and readily identifiable to species level in the majority of cases.

One of the most common sedimentary contexts for recovering vertebrate fossils is from fluvial and estuarine deposits laid down along former river courses. These represent a combination of autochthonous remains of, for example, fish and

amphibians that would have been living in the water body and allochthonous remains of other animals, principally nonaquatic mammals and birds, that may have died nearby on the floodplain and been incorporated through overbank flooding or that may have become mired at the water's edge, been hunted there, or even drowned in the water (Figure 1). Although the fluvial record is a global one, arguably some of the

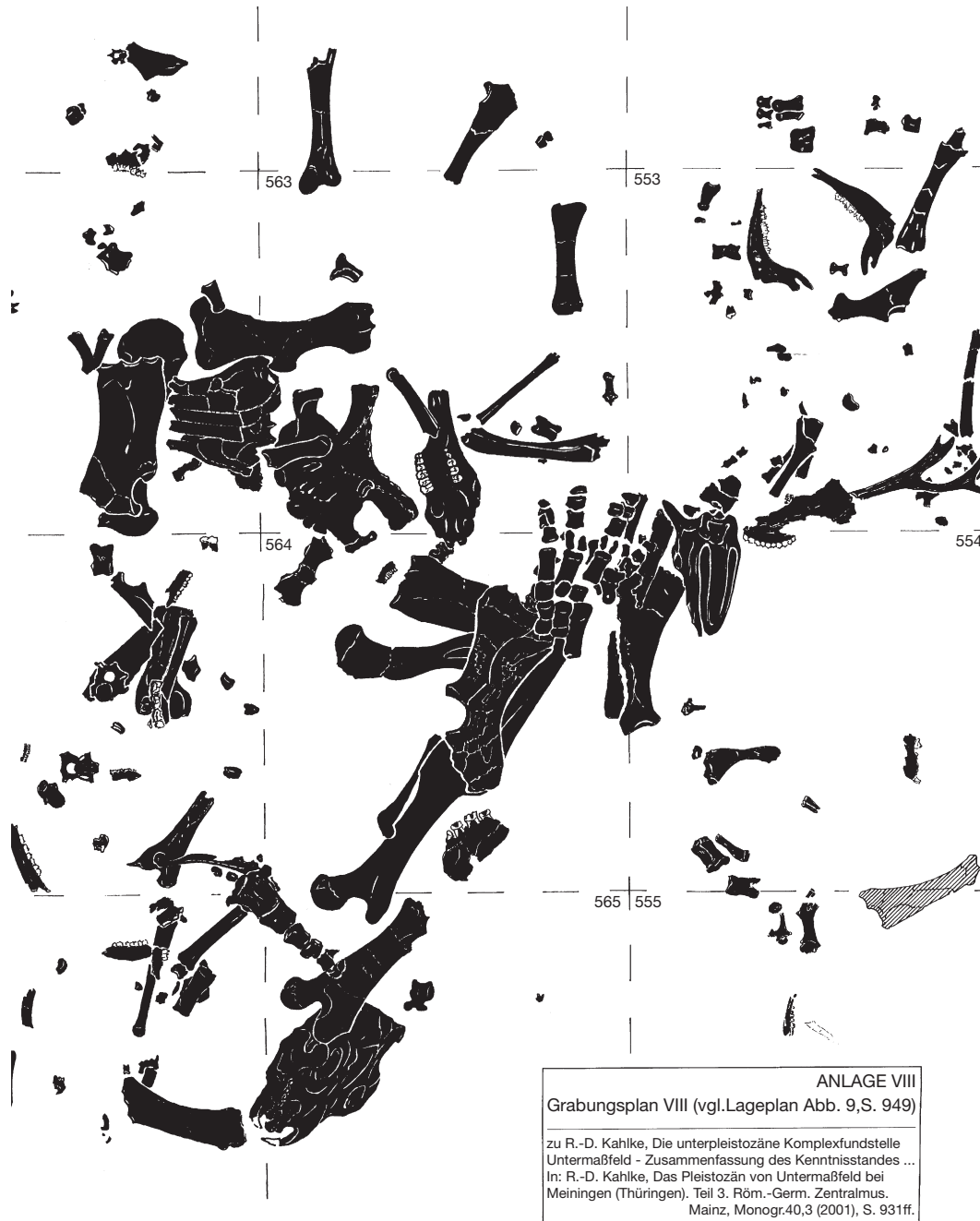


Figure 1 Part of the excavated area of the early Pleistocene Lower Fluvial Sands at Untermaßfeld (Germany) showing a floodwater accumulation including remains of bison (*Bison menneri*), comb-antlered deer (*Eucladoceros giulii*), ancient fallow deer (*Cervus* s.l. *nestii valloinetensis*), hippopotamus (*Hippopotamus amphibius antiquus*), Etruscan rhinoceros (*Stephanorhinus etruscus*), short-faced hyena (*Pachycrocuta brevirostris*), and bear (*Ursus rodei*). From Kahlke R-D (2001) Das Pleistozän von Untermaßfeld bei Meiningen (Thüringen). *Monographien des Römisch-Germanschen Zentralmuseums Mainz* 40: 3.

best sequences come from southern and western Europe, where stratified river terrace sequences preserve a remarkable succession of cold- and temperate-climate assemblages covering much of the Pleistocene. Notable examples include the Tiber (Italy), the Rhine (Germany), the Somme (France), the Maas (The Netherlands), and the Thames (England). Although some fluvial exposures are naturally created through erosion (particularly coastal sections), many fossils came to light during the late 19th and early 20th centuries when quarries for the extraction of sand and gravel were opened up, thereby providing opportunities for quarrymen and interested antiquarians to collect remains. The digging of deep foundations for buildings has also exposed Pleistocene vertebrates in unusual locations—for example, the finding of fossils of last interglacial age (Marine Oxygen Isotope substage (MIS) 5e) hippopotamus (*Hippopotamus amphibius*) in former Thames deposits under Trafalgar Square in central London. Later mechanization of aggregates extraction has made it more difficult to recover the quantity of remains previously obtained, although sand and gravel pits continue to yield spectacular finds (Figure 2) wherever paleontological and archaeological research can be combined with commercial needs. At Boxgrove in West Sussex (England), the site of an extraordinarily well-preserved early middle Pleistocene landscape with a rich vertebrate fauna, Paleolithic archaeology and hominin remains came to light as the result of quarrying activities (Roberts and Parfitt, 1999). Large numbers of vertebrate fossils continue to be dredged from fluvial sites below the water table, for example, in The Netherlands and Germany, although the stratigraphical context of these specimens has evidently been lost. Associated with the fluvial record may be spring-line deposits of subaerially precipitated tufa or travertine, laid down during interglacial periods. Examples from western and central Europe have proved to be the source of very large numbers of vertebrate fossils, including highly unusual finds such as the eggs of the European pond terrapin (*Emys orbicularis*) and bird feathers and nests from Ehringsdorf (Germany).

The deposits of former lakes and sinkholes also constitute one of the major sedimentary contexts in which vertebrate fossils may be found. The importance of lacustrine environments lies in their relatively gentle depositional environment,



Figure 2 Anterior skeleton of a Last Cold Stage (MIS 3) woolly rhinoceros (*Coelodonta antiquitatis*) discovered in 2002 at Whitmoor Haye quarry, Staffordshire, UK. Photo courtesy of G. Coates, Birmingham University Field Archaeology Unit.

thereby favoring good preservation of vertebrate material, including the potential for complete or partial skeletons. Deep lake basins have played a particularly significant role in our understanding of African Pleistocene vertebrate history, as well as providing a rich source of early hominin remains and archaeology. Lakes can also be associated with large numbers of animals that have perished due to accidental death—for example, the site of Hot Springs, South Dakota (USA), where 43 late Pleistocene mammoths, mostly young males, perished.

Caves have proved to be an extremely rich source of vertebrate fossils and excellent examples are known from every continent except Antarctica. The most common type are limestone caves, formed by the percolation of acidic groundwater through cracks and along planes of weakness in the limestone, ultimately leading to the formation of often large and complex cave systems over tens of thousands of years. Other types include volcanic caves (formed during cooling of lava flows on the flanks of volcanoes and during other, more unusual, volcanic processes, such as the circular caverns created by the *nuée ardente* explosion of Mount Fantale in Ethiopia), vertical fissure caves (created by tension in hard rocks, especially in volcanic areas or where dissected hard sedimentary rocks overlie clay), and sea caves (created by wave action). Vertebrate remains commonly found within caves include a wide variety of carnivores depending on location, such as lion, leopard (*Panthera pardus*), wolf (*Canis lupus*), spotted hyena, dire wolf (*Canis dirus*), American lion (*Panthera atrox*), and thylacine (*Thylacinus cynocephalus*). These animals used the caves for denning and rearing their young, but other carnivores occupied them mainly for overwintering (e.g., the many thousands of Last Glaciation cave bears that died during hibernation in western and central Europe). In addition, the remains of carnivore prey are also found. Herbivores otherwise rarely enter caves deliberately, unless to obtain salt (as seen today, for example, on Mount Elgon in Kenya) and occasionally to seek shelter. Notable exceptions are the spectacular Cueva de Ultima Esperanza in Chile, with rich ground sloth (*Mylodon* sp.) remains and thick deposits of ground sloth dung, and Bechan Cave in Utah (USA), where abundant Columbian mammoth (*Mammuthus columbi*) dung has been found.

More commonly, a range of vertebrates including very large mammals may become preserved in caves through accidental death, principally by falling through natural shafts or pitfalls in the roof. This can lead to the buildup of large talus cones of bone beneath former openings, an excellent example of which is known from Joint Mitnor Cave in Devon (England), which includes narrow-nosed rhinoceros (*Stephanorhinus hemitoechus*) and straight-tusked elephant (*paleoloxodon antiquus*). A similar situation is known from the Naracoorte Caves in southern Australia, where extinct sthenurine kangaroos and *Zygomaturus trilobus* (an animal the size and build of a modern pygmy hippopotamus) apparently fell through a shaft in the roof. In addition, mudflows and underground streams may carry remains inside the cave from the surrounding land surface. Small mammal remains from caves include many species of bat (usually rarely recovered from open sites) and a wide range of small rodents and insectivores brought in by birds of prey and deposited in pellets under roosting spots within the cave. Taphonomic studies conducted on these small mammal remains (Andrews, 1990) reveal the presence of patterns of

differential digestion on the molars and incisors from which the predator concerned can be identified. Early humans routinely made use of caves for shelter (presumably in competition with local carnivores), for the burial of their dead, and, from the Upper Paleolithic onwards, as a focal point for cave art.

Former marine deposits now located in onshore areas have yielded remains of whales, walruses, seals, and other marine mammals, as well as turtles and fish. Important examples are known from the 'crag' sands of northwest Europe, which represent an early Pleistocene shallow sea. Occasional terrestrial mammals of contemporary age, including the mammoth *Mammuthus meridionalis* and comb-antlered deer (*Eucladoceros* spp.), are also represented, presumably washed down from the estuaries. Dredging of the sea bed can also yield surprising finds. Many thousands of specimens of Last Glaciation mammals, including woolly mammoth (*Mammuthus primigenius*), reindeer (*Rangifer tarandus*), and bison (*Bison priscus*), have been dredged by trawlers from the floor of the North Sea between The Netherlands and the United Kingdom. This area, commonly known as 'Doggerland,' formed a land bridge between the United Kingdom and the European mainland at times of lowered sea level during the Pleistocene and was inhabited by a wide range of fauna that died and became incorporated into fluvial deposits later flooded by the Holocene sea level rise.

Some of the most unusual and celebrated Pleistocene vertebrate finds are those preserved as 'mummies' in the permafrost zones of the Northern Hemisphere, in arid areas where cold, dry air has 'freeze-dried' the bodies or hot, dry conditions have desiccated them, and in places where natural petroleum seeps have preserved the remains, sometimes 'pickling' them through a combination of oils and salt. Many well-documented examples of woolly mammoth, both juvenile and adult, have come from Siberia, where they are naturally eroded out of the frozen ground by fluvial processes. On the other side of the Bering Straits, the activities of placer mining in Alaska have brought to light a range of fauna, including spectacular remains of bison and small mammals such as black-footed ferret (*Mustela nigripes*) and ground squirrel (*Spermophilus* sp.). Examples of freeze-dried remains include the aforementioned ground sloths from South America, whereas Australia has yielded a number of desiccated specimens, including the marsupial thylacine. Remarkable accumulations of Pleistocene vertebrates have come from the petroleum seeps of Rancho La Brea in downtown Los Angeles and Talara (Peru), and both mammoth and woolly rhinoceros carcasses, pickled in ozocerite, have been recovered from Starunia on the Polish-Ukrainian border. These particular circumstances provide an exceptionally valuable insight into the appearance, diet, and overall health of these animals through investigation of the soft parts of the body, such as the skin, hair, appendages (e.g., mammoths' trunks), the stomach contents, and internal parasites. In addition, they also provide material for ancient DNA analyses, thereby allowing the reconstruction of evolutionary relationships and phylogeography (Höss et al., 1996; Barnes et al., 2002).

It is important to bear in mind that a range of materials other than the physical remains of these animals can equally provide information concerning their paleobiology and

appearance. The presence of fossil trackways (ichnofossils) can reveal the past presence of certain taxa where their actual remains may not have been preserved as well as allow stature and gait to be reconstructed. Similarly, the presence of carnivores can be detected through characteristic gnawing patterns on bones, including puncture marks, striations, and bone breakage. The most dramatic signs of carnivore activity come from the spotted hyena, a notorious 'bone cruncher.' The bone fragments are later passed out of the animal and the droppings fossilize as coprolites, thereby creating another source of evidence for past presence. Signs of rodent gnawing may also be present on bones, as well as wood and nut shells. Paleolithic cave paintings and portable art also provide important information concerning the appearance and behavior of extinct Pleistocene vertebrates that will not be apparent from their fossilized bones and teeth, such as the small ears of the woolly mammoth, the coloration of wild horses (*Equus ferus*), and the apparently maneless state of Pleistocene lions.

Sampling, Excavation, and Analysis Methods

The majority of *in situ* larger vertebrate remains in unconsolidated sediments can be recovered by hand using standard paleontological and archaeological techniques. These primarily involve the excavation of the material and the recording of its position in three dimensions (with an electronic distance meter, theodolite, or similar) and, where appropriate, on a hand-drawn scaled plan to show orientation. On Paleolithic sites, where a high level of detail may be required concerning the relationship of vertebrate remains to the archaeology, it is common practice to instigate a grid system (usually 0.5×0.5 m or larger) for excavation and recording purposes. Each square can then be excavated in 0.1-m deep spits (layers) and a proportion of the excavated spoil dry-sieved (with a 1- or 2-mm mesh) to recover any small remains that may have been missed by the excavator. Most remains can be easily exposed using a standard pointing trowel, although care should be taken when nearing the bone to switch to a wooden or plastic implement (a clay modeling tool is ideal) so as not to scratch or damage the surface. More friable material may require consolidation *in situ* (frequently involving the application of a damp tissue paper 'barrier' followed by quick-hardening plaster strips) and lifting as a block for later careful excavation in the laboratory. Better preserved material can normally be lifted, bagged, and subsequently cleaned with warm water and a soft brush prior to being allowed to dry out slowly (never in direct sunlight). Vertebrate remains found in consolidated cave deposits such as breccias usually require patient and time-consuming exposure by expert preparators using compressed air drills and other equipment. Full records of any conservation steps taken should be kept. Specialist conservation advice should be sought, as required, from appropriate authorities prior to the application of any consolidants or preservatives because these can impact later dating or chemical studies. Much late 19th and early 20th century material has been rendered useless for radiocarbon dating through the routine soaking of bones in a fishbone-derived glue that was in vogue as a consolidant.

In addition to the recovery of larger vertebrate material by hand, it is now standard practice to take bulk samples of

sediment to be wet-sieved (screened) for the extraction of microvertebrate remains. This practice has only become widespread since the 1970 s but has revolutionized paleontological studies by allowing the collection of small vertebrate fossils that are particularly useful for paleoecological and biostratigraphical studies. Samples can be taken from any suitable exposed face and would normally consist of a series of column samples from the base to the top so that any change upward through the sequence can be identified. Additional samples can then be taken from particularly fossiliferous horizons. The size of an individual sample will vary depending on the extent and nature of the sediments under investigation, but a minimum of 10 L of sediment (10–20 kg) is recommended for each. Clay-rich samples may require air-drying to weaken the hydrophilic bond of the clay particles prior to sieving in order to permit easier disaggregation of the sediment. Sediment should be wet-sieved using a 500- μm mesh, either using a handheld sieve and any convenient water source or a bespoke sieving machine, usually consisting of a large tank with a lid on the underside of which are fixed oscillating sprinklers that gently disaggregate the sediment. Once the residue is clean, it can be dried and then sorted under a low-power ($10\times$) binocular microscope in order to extract small vertebrates (teeth, bones, and bone fragments), invertebrates (if rare in the sample), and microdebitage.

Identification of vertebrate remains is carried out using keys, drawings, and comparative material (often museum collections). In addition to morphological analyses, standard metrical analyses are undertaken. These can be applied to a variety of questions, including identification problems, the recognition of sexual dimorphism, and examination of the range of natural variation within and between populations. A relatively new area of analysis is the application of geometric morphometrics to groups as diverse as microtine rodents, pigs, and horses. This involves the digital imaging of specimens and the establishment of 'landmarks' and outlines, which are then subjected to shape analysis. This permits the recognition of discrete differences in shape between populations that would not normally be revealed by standard metrical analyses. This has proved useful in identifying the area of origin of particular species and in tracking movements of different populations through time (Bignon et al., 2005; Larson et al., 2005).

Of critical importance to any fossil study is an understanding of the taphonomy (literally the 'laws of burial') of the assemblage under consideration. It is only through detailed and frequently painstaking paleontological study that the habits and lifestyles of former animal communities can be reconstructed. Understanding the biases inherent in the fossil record is therefore vital. An example of the importance of taphonomic analysis was highlighted by Brain (1981), who demonstrated that numerous South African assemblages formerly attributed to the hunting activities of *Australopithecus africanus* were in fact accumulated by carnivores. Moreover, it became apparent that the australopithecines were frequently featured as prey items (the 'hunted' as opposed to the 'hunters'). In order to understand the formation of an assemblage, it is necessary to understand the cause of death and the sequence of events that has influenced the transition from a living organism to a fossil. In vertebrates, death is usually caused by one of the following: predation, accident, senility,

starvation, or dehydration. These processes will differentially prejudice the amount and condition of material entering the sedimentary records since, for example, the victims of predation are less likely to be preserved as fossils than animals that die by accident (e.g., by falling into a cave). Biases before or during burial include: (1) varying rates of decay, dependent on temperature and invertebrate activity; (2) scavenging activities, particularly by carnivores; (3) butchery and disarticulation by early humans; (4) trampling and dispersal of the material, notably by hooved mammals; (5) the degree of exposure, weathering, and root damage on the bone surface; and (6) the effects of transportation in rivers (rolling, abrasion, and changing the orientation of bones) prior to burial. Potential biasing factors affecting an assemblage after burial include pH, recovery techniques, choice of analytical procedures, and publication decisions. Due consideration of all of these factors will greatly enhance any attempt at paleontological reconstruction.

Paleoecology

Today, most mammalian herbivores are limited to specific ecosystems, but carnivores generally have a wider distribution. During the Pleistocene, species that are today restricted largely to sub-Saharan Africa, such as lions and spotted hyenas, were found in cold and warm stages alike. Nevertheless, since most Pleistocene mammals and birds are still extant or have close relatives living today, extrapolation from the comparative anatomy and habitat preferences of modern species can be used to reconstruct past environments with reasonable confidence. Particular species associations may therefore be indicative, for example, of deciduous woodland, slow-moving water, or steppe grassland, although it is important to remember that the climatic oscillations of the Pleistocene may have disrupted communities, leading to the existence of 'disharmonious' populations that contained a mixture of species that are today native, exotic, or even extinct in a given area. However, as warm-blooded organisms, mammals and birds are more able to maintain a constant temperature against a climatic gradient than cold-blooded organisms, such as reptiles, amphibians, and fish. The latter are thus much more sensitive to temperature for basic survival and, particularly, for breeding. Accordingly, the past appearance in Britain of 'exotic' reptiles such as the European pond terrapin (Figure 3), now found only in southern and eastern Europe, is indicative of higher mean summer temperatures during previous interglacials.

Paleobiogeography

Although today, human activity has disrupted and excluded many vertebrates from existing in certain regions, the repeated patterns of climatic change during the Pleistocene also caused a change in the distributions of many vertebrate taxa. Not only did the growth and decay of glaciers cause species to move in and out of regions affected by these processes but also the changes in global ice volume had a direct effect on sea level by dictating the total land surface available for colonization and alternately opening and closing land bridges between land masses. One of the best known examples of such a land bridge,



Figure 3 Modern European distribution of the European pond terrapin (*Emys orbicularis*). From Arnold EN and Burton JA (1980) *A Field Guide to the Reptiles and Amphibians of Britain and Europe*. London: Collins.

critical to vertebrate migration, was the extensive high-latitude land connection between Asia and North America, known as the Bering land bridge, which permitted large-scale faunal interchange and provided the original route for human expansion into the Americas. Other temporary land bridges to islands allowed the immigration of certain species that then became isolated once sea level rose again. This had a striking effect on some of the largest mammals, causing them to become dwarfed in order to cope with reduced vegetational availability, whereas the smallest animals were able to expand in size because of a lack of predators. Examples of dwarf elephants and hippopotamuses are known from various islands in the Mediterranean, including Cyprus and Malta (Figure 4), whereas small-sized mammoths are known from the California Channel Islands off the west coast of the United States and Wrangel Island in northeastern Siberia. Regarding small mammals, examples of 'giant' dormice, shrews, and hedgehogs are known from the Mediterranean. Another very important factor influencing Quaternary vertebrate distribution has been the periodic expansion, movement, and contraction of areas of moderate to extreme aridity since the presence of water is critical for sustaining life. This is clearly seen in the dramatic dispersal undertaken by the saiga antelope (*Saiga tatarica*), which today inhabits the steppes and semidesert regions of southeastern Europe and Central Asia but during the Late Glacial suddenly expanded its range from southwestern Britain to Alaska, perhaps in response to a one-off pulse of exceptional aridity.

Many species therefore underwent considerable distributional shifts during the Quaternary. This is particularly evident with European fallow deer (*Dama dama*), which experienced

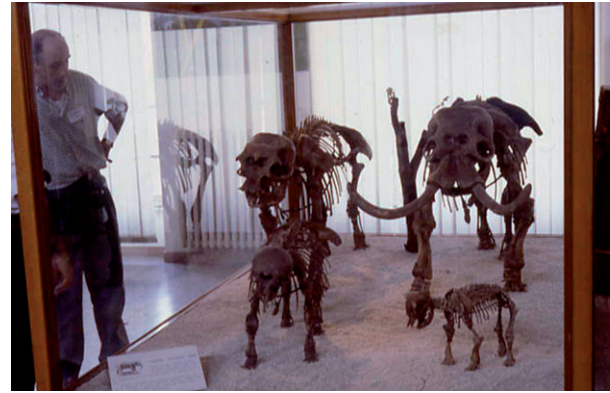


Figure 4 'Family group' of reconstructed individuals from Spinagallo Cave, Sicily, on display at the University of Rome (Dr T. Kotsakis as scale).

extensive expansion and contraction of its range, retreating back to its core area in the Mediterranean during cold stages and expanding throughout northwest Europe during interglacials. By tracking a preferred environment and thus maintaining a constant habitat, a species may avoid the need for evolutionary modification. However, changes in distribution (particularly contraction of the range) will often result in widespread selective mortality. This may act as a trigger for adaptive evolution as animals attempt to cope with new conditions, or it may lead to fragmentation of the population and subsequently allopatric speciation. The potential of mitochondrial DNA studies for identifying areas of glacial refugia and reconstructing phylogeographic patterns of dispersal has been highlighted by Hewitt (2000).

Biostratigraphy

One of the major advantages that vertebrates, notably mammals, have over other biological proxies concerns their applicability for the relative dating of Quaternary deposits through features such as assemblage composition, size change, origination and extinction first appearance and last appearance datums (FADs and LADs), and the significant patterns of morphological evolution, often quantifiable, that are observable in many lineages (Lister, 1992). The polar ice masses are known to have undergone major fluctuations in extent for millions of years, but it is only during the Quaternary that there was a significant mammalian response to these changes. Faunal turnover was therefore greater during the Quaternary because extinctions, immigrations, and evolutionary changes were both precipitated and magnified by these climatic changes. Because mammals can rapidly disperse into new areas and are readily preserved in a wide range of depositional environments, they are ideal for correlating different sorts of deposits over wide geographical areas. Along with these studies must be a rigorous approach to the stratigraphical and paleoenvironmental reconstruction of each site and an understanding of the taphonomy of individual assemblages. However, the replication of particular faunal groupings and identification of age-significant morphotypes at numerous sites can lead to the establishment

of credible and testable biostratigraphical frameworks, into which new assemblages may be slotted.

A number of different biochronological schemes spanning the entirety of the Pleistocene are in operation in Europe, based on land mammal ages or mammal assemblage zones. These are paralleled in North America, South America, and the Far East, although with a substantially lower degree of resolution. Formal biochronological zones in Africa and Australia have yet to be defined. Some of the most powerful evidence for the identification and correlation of different climatic episodes within Europe has come from mammalian assemblages from long fluvial sequences, such as the Tiber in Italy (Gliozzi et al., 1997), the Somme and Seine in France (Auguste, 1995), the Wipper and Neckar in Germany (Schreve and Bridgland, 2002), and the Thames in the United Kingdom (Schreve, 2001; Figure 5). These have enabled the linking of biological and lithostratigraphical evidence to create chronological frameworks that are applicable over wide areas.

The first appearance of key genera or species is of particular significance, with notable origination and dispersal events in the late Pliocene–early Pleistocene of the Canidae (Old World dogs), the genus *Mammuthus* and voles of the genus *Microtus* with unrooted (permanently growing) molars. LADs can be more difficult to establish since finds of younger age can potentially be made in the future but they are most easily recognized at the marginal areas of a species' range (e.g., the British Isles) since taxa will persist longer in the source area from where they originally dispersed. An example from Britain is the replacement of the cave bear by the brown bear (*Ursus arctos*) in the late middle Pleistocene (MIS 9; Schreve, 2001),

even though the former persisted on the continent until the end of the Pleistocene. Nevertheless, it is clear that events such as glaciations, in particular, had the potential to cause a suite of extinctions over a wide area. In addition to the origination of new species, many mammalian lineages display cladogenetic speciation (i.e., change within a lineage, without the need for geographical separation). Well-known case studies exist for the elk (moose) (Lister, 1993), mammoth (Lister and Sher, 2001), and water vole (Maul et al., 1998) lineages, with a succession of 'chronospecies' described for each. In combination, the evidence from evolutionary change within different vertebrate lineages, FADs and LADs, size change, and assemblage composition can provide a robust relative dating framework for time periods that are beyond the reach of radiocarbon dating and for contexts in which suitable materials for other absolute dating techniques may be absent.

Interactions with Hominins and Extinctions

In addition to contributing to a wider paleoenvironmental and dating context at Paleolithic archaeological sites, the final area of significance to Quaternary vertebrate studies relates to their interaction with contemporary early humans. The nature of such interactions, ranging from subsistence-related issues such as identifying predation by hominins on mammals (and vice versa) to the role of vertebrates in the parietal and portable art of *Homo sapiens*, the responsibility of humans in the late Pleistocene extinction of the megafauna, and the subsequent domestication of key mammals, is summarized elsewhere in this encyclopedia.

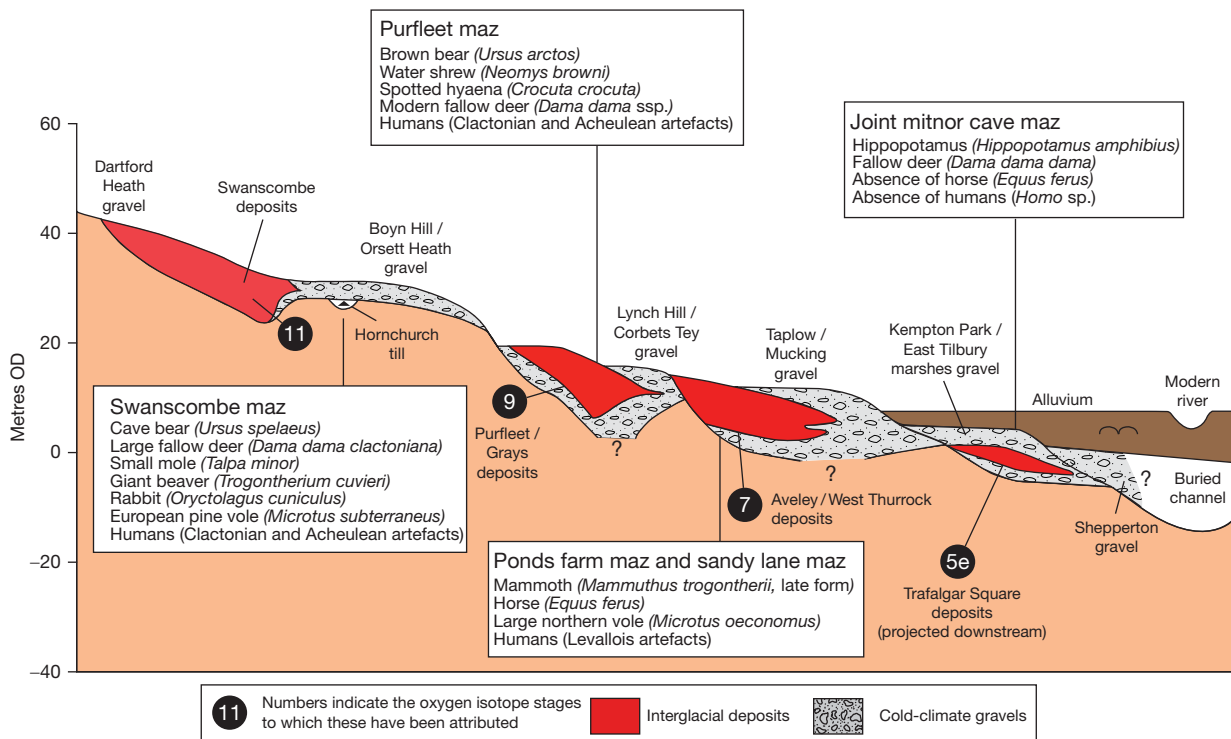


Figure 5 Idealized transverse section through the Thames terrace staircase, with principal features of the Mammalian Assemblage Zones (MAZ) and correlation with the marine oxygen isotope record indicated. Modified from Schreve DC (2001) Differentiation of the British late middle Pleistocene interglacials: The evidence from mammalian biostratigraphy. *Quaternary Science Reviews* 20: 1693–1705.

See also: **Vertebrate Records:** Early Pleistocene; Mid-Pleistocene of Australia; Mid-Pleistocene of Europe; Mid-Pleistocene of North America; Late Pleistocene Megafaunal Extinctions; Late Pleistocene Mummified Mammals; Late Pleistocene of North America; Late Pleistocene of South America. **Vertebrate Studies:** Ancient DNA; Interactions with Hominids.

References

- Andrews P (1990) *Owls, Caves and Fossils*. London: Natural History Museum.
- Antoine P (1993) Le système de terrasses du Bassin de la Somme: modèle d'évolution morphosédimentaire cyclique et cadre paléoenvironnemental pour le Paléolithique. *Quaternaire* 4: 3–16.
- Arnold EN and Burton JA (1980) *A Field Guide to the Reptiles and Amphibians of Britain and Europe*. London: Collins.
- Attenborough D (1987) *The First Eden: The Mediterranean World and Man*. London: Collins.
- Barnes I, Matheus P, Shapiro B, Jensen D, and Cooper A (2002) Dynamics of Pleistocene population extinctions in Beringian brown bears. *Science* 295: 2267–2270.
- Bignon O, Baylac M, Vigne JD, and Eisenmann V (2005) Geometric morphometrics and the population diversity of Late Glacial horses in Western Europe (*Equus caballus arcelini*): Phylogeographic and anthropological implications. *Journal of Archaeological Science* 32: 375–391.
- Brain C (1981) *The Hunters or the Hunted. An Introduction to African Cave Taphonomy*. Chicago: University of Chicago Press.
- Buckland W (1823) *Reliquiae Diluvianae*. London: John Murray.
- Frere J (1800) Account of flint weapons discovered at Hoxne in Suffolk. *Archaeologia* 13: 204–205.
- Gliozzi E, Abbazzi L, Argenti P, et al. (1997) Biochronology of selected mammals, molluscs and ostracods from the middle Pliocene to the Late Pleistocene in Italy. The state of the art. *Rivista Italiana di Paleontologia e Stratigrafia* 103: 369–388.
- Hewitt G (2000) The genetic legacy of the Quaternary ice ages. *Nature* 405: 907–913.
- Höss M, Dilling A, Currant A, and Pääbo S (1996) Molecular phylogeny of the extinct ground sloth. *Myodon darwini*. *Proceedings of the National Academy of Sciences* 93: 181–185.
- Kahlke R-D (2001) Das Pleistozän von Untermassfeld bei Meiningen (Thüringen). *Monographie des Römisch-Germanschen Zentralmuseums Mainz* 40: 3.
- Larson G, Dobney K, Albarella U, et al. (2005) Worldwide phylogeography of wild boar reveals multiple centers of pig domestication. *Science* 307: 1618–1621.
- Lister AM (1992) Mammalian fossils and Quaternary biostratigraphy. *Quaternary Science Reviews* 11: 329–344.
- Lister AM (1993) Patterns of evolution in Quaternary mammal lineages. In: Edwards D (ed.) *Patterns and Processes of Evolution*, pp. 72–79. London: Academic Press.
- Lister AM and Sher AV (2001) The origin and evolution of the woolly mammoth. *Science* 294: 1094–1097.
- Maul L, Masini F, Abbazzi L, and Turner A (1998) The use of different morphometric dates for absolute age calibration of some South- and Middle-European arvicolid populations. *Palaeontographica Italica* 85: 111–151.
- Pengelly W, Busk G, Evans J, et al. (1873) Report on the exploration of Brixham Cavern conducted by a committee of the Geological Society, and under the supervision of Wm. Pengelly, Esq., FRS., aided by a local committee; with descriptions of the animal remains by George Busk, Esq., FRS., and of the flint implements by John Evans, Esq., FRS., Joseph Prestwich, FRS., FGS., &c. *Philosophical Transactions of the Royal Society of London* 163: 471–572.
- Roberts MB and Parfitt SA (1999) *Boxgrove. A Middle Pleistocene Hominid Site at Earham Quarry, Boxgrove, West Sussex*. London: English Heritage Archaeological Report No. 17.
- Schreve DC (2001) Differentiation of the British late Middle Pleistocene interglacials: The evidence from mammalian biostratigraphy. *Quaternary Science Reviews* 20: 1693–1705.
- Schreve DC and Bridgland DR (2002) Correlation of English and German Middle Pleistocene fluvial sequences based on mammalian biostratigraphy. *Netherlands Journal of Geosciences* 81: 357–373.

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VERTEBRATE RECORDS

Contents

Early Pleistocene

Early and Middle Pleistocene of Northern Eurasia

Mid-Pleistocene of Africa

Mid-Pleistocene of Australia

Mid-Pleistocene of Europe

Mid-Pleistocene of North America

Mid-Pleistocene of Southern Asia

Late Pleistocene of Africa

Late Pleistocene of North America

Late Pleistocene of South America

Late Pleistocene of Southeast Asia

Late Pleistocene Megafaunal Extinctions

Late Pleistocene Mummified Mammals

Early Pleistocene

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Introduction

The Plio–Pleistocene Transition and Biochronology

Following recent revision, the Plio–Pleistocene boundary is now dated to 2.6 Ma (Gibbard et al., 2010). Understanding the early Pleistocene vertebrate record is impossible without taking into account the late Pliocene events that influenced and characterized the composition of early Pleistocene land vertebrate communities.

Vertebrate paleontologists are used to subdividing geological time by using biochronological scales based on evolutionary stage and dispersal events. Two different biochronological scales exist for the late Pliocene–early Pleistocene in Europe, one based on micromammals that spans the late Villanyan to the Biharian and the other based on large mammals, which extends from the middle Villafranchian to the early Galerian. These ‘mammal ages’ have been further subdivided into faunal units on the basis of bio-events that produced changes in the composition of the assemblages. Similar biochronologic units are used in different continents (Figure 1): in North America, the Irvingtonian; in South America, the Uquian and the early Ensenadian; in China, the Nihewanian cover the latest Pliocene and early Pleistocene. In Australia the succession of faunal units (although not yet well defined) is outlined in Woodburne et al., (1984), while in Africa formal biochronological scales or units based on mammals have also not yet been defined (Lindsay et al., 1990).

A special mention must be made of the large-mammal age Villafranchian, due to its extensive use, especially in European literature. The term Villafranchian, proposed by Pareto in 1865, was originally considered to represent the youngest part of the continental Pliocene. In 1916, Gignoux proposed a correlation of this stage with his marine Calabrian, assumed to represent the late Pliocene. In 1948, at the 18th International Geological Congress, it was agreed to place the Calabrian at the base of the Pleistocene. Consequently, the Villafranchian was also considered to represent the earliest stage of the continental Pleistocene. Some authors (Azzaroli, 1992) remarked that the many so-called Villafranchian assemblages are neither homogeneous nor strictly contemporary. The chronological sub-division of the Villafranchian was extensively discussed during two meetings in 1975 and 1976 (Azzaroli, 1977; Alberdi and Aguirre, 1977). As a result, a number of faunal units were recognized within the Villafranchian, the oldest three belonging to the middle and late Pliocene and the others to the early Pleistocene. Within these limits the Villafranchian spans from a little over 3.0 Ma to about 1.1 Ma. During this long time interval remarkable faunal changes took place. However, the term Villafranchian has come to lose much of its intrinsic value, its name being maintained for historical reasons and for the sake of stability in nomenclature. It has no real meaning unless it is used with definite qualification, such as early, middle, and late Villafranchian, or better still with the indication of a definite faunal unit (Azzaroli, 1992).

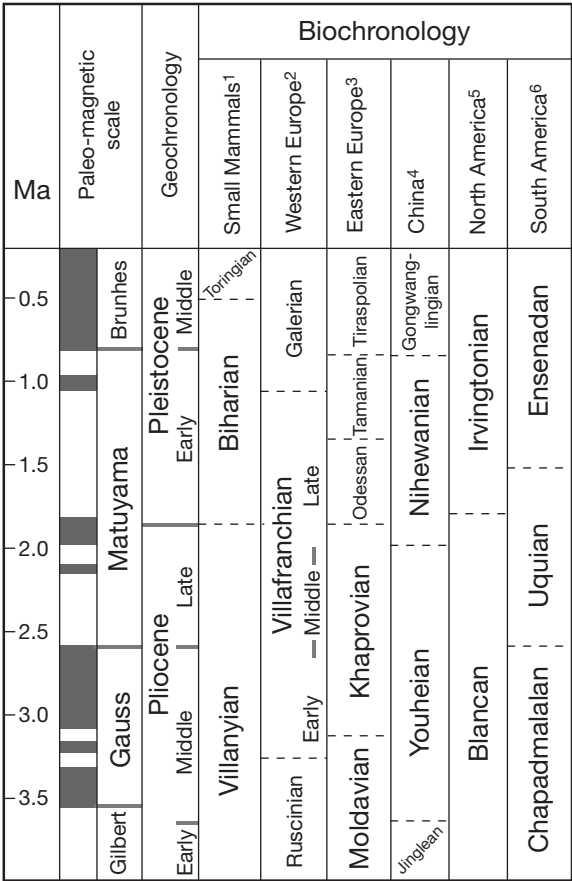


Figure 1 Schematic correlation chart of Geochronology and Mammal Biochronology Units according to: (1) Fejfar & Heinrich (in Lindsay et al., 1990); (2) Gliozzi et al. (1997); (3) Gabunia et al. (2000); (4) Li et al. (1984); (5) Cione and Tonni (1995); (6) Marshall et al. (1984).

Early Pleistocene Case Histories

The Old World Canidae (genus *Canis*)

In a pioneering paper, Azzaroli (1983) defined the arrival in Western Europe of the first representative of the genus *Canis* 'sensu stricto' as the 'wolf event'. This event, in the original view of Azzaroli, defined the beginning of the late Villafranchian (latest Pliocene–early Pleistocene), represented by the so-called Olivola Faunal Unit.

The evolutionary history of the genus *Canis* and its dispersal across North America and the Old World reveal a much more complex pattern of dispersals and evolution of this lineage that is still obscure in several aspects. The center of origin and evolution of dogs is North America from where, across Beringia, dogs repeatedly spread into Eurasia and Africa. The first event of canid dispersal occurred in the late Miocene (Hemphillian mammal age in North America, Turolian mammal age in Europe), with forms attributable to the primitive genus *Eucyon*. Species belonging to this genus are then found, albeit rarely, across the Old World until the late Pliocene.

The earliest known representative of *Canis* 'sensu stricto' occurs in North America, in late Hemphillian deposits, about 5.5 Ma. The initial dispersal across Beringia of the genus *Canis* into the Old World occurred around 3.0 Ma. The occurrence of both

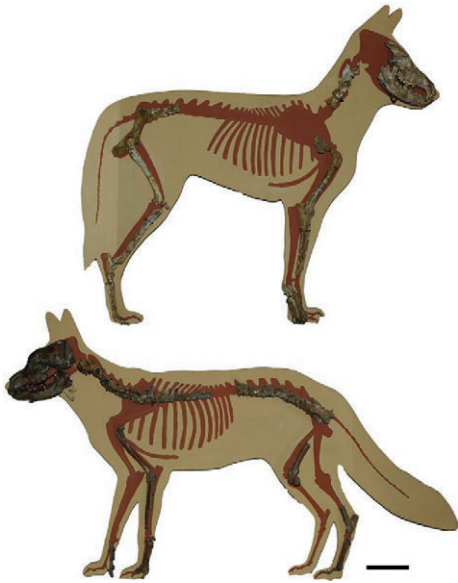


Figure 2 *Canis etruscus* (top), composite skeleton from Olivola (Italy; Olivola F.U., late Villafranchian) and *Canis arnensis* (bottom), composite skeleton from Upper Valdarno (Italy, late Villafranchian). Museo di Storia Naturale (Sezione Geologia e Paleontologia), Università Firenze. Scale bar 20 cm approx. Photo: L Rook.

Canis cf. *etruscus* (the Etruscan wolf) and *Canis* cf. *arnensis* (the Tuscan coyote-like dog) are reported in deposits of this age in China. The radiation of canids in Eastern Europe, the 'wolf event', approximates with a major extinction event among carnivores, such as the raccoon-dog *Nyctereutes megamastoides* and the cursorial hyaenid *Chasmaporthetes*. The 'wolf event', one of the major faunal renewals marking the Plio–Pleistocene transition, is characterized by the appearance *Canis etruscus* (Figure 2). The event is rapidly followed in Western Europe by other arrivals, such as two other large canid species, *Canis arnensis* (Figure 2) and the primitive wild dog *Lycaon falconeri* (Figure 3), as well as by the large scavenging hyaena *Pachycrocuta brevirostris*. This carnivore guild survived over most of Eurasia for more than 1 Ma throughout the entire early Pleistocene (Rook and Torre, 1996; Martínez-Navarro and Rook, 2003).

From the Iberian peninsula to the Caucasus, a number of latest Pliocene and early Pleistocene localities provide an extensive documentation of the canid record, including Spain (Venta Micena), Italy (Poggio Rosso and Pirro Nord), Greece (Apollonia-1 and Gerakarou), Germany (Untermassfeld), and Georgia (Dmanisi).

Some of these localities (Poggio Rosso, Untermassfeld, Dmanisi) are providing an unusually abundant record that will allow a firm base to direct future research in developing studies, such as analysis of population structure and intra- and inter-specific variability.

Villafranchian Proboscideans and the Early Dispersal of Elephants

A considerable turnover in the proboscidean fauna of Eurasia took place during the Pliocene, with the local extinction of the Deinotheriidae (deinotheres survived in Africa until the early



Figure 3 *Lycaon falconeri* from Fan Tsun (Taiku, China). Skull and mandible, cast kept in the Museo di Storia Naturale (Sezione Geologia e Paleontologia), Università di Firenze; original housed in the American Museum of Natural History (New York). Scale bar represents 5 cm. Photo: L Rook.

Pleistocene), Mammutidae (*Mammuth* was present in North America until the late Pleistocene), and Gomphotheriidae (last recorded in South America in the early Holocene) and the rise of stegodonts (Stegodontidae) and modern elephants (Elephantinae). From their center of origin and diversification in Africa during the late Miocene (Shoshani and Tassy, 2005), elephants migrated into Eurasia and, in the case of mammoths, also to North America in a series of dispersal events encompassing the Pliocene and early Pleistocene. The migration of elephants out of Africa, and especially the incursion of species into Europe, have been linked to major climatic changes that had a profound impact on the evolution of Eurasian local faunas.

The earliest record of elephants in Eurasia is in early Villafranchian (middle Pliocene) deposits of the Dacic Basin (Romania), Bethlehem (Israel), the Siwaliks (India), and the Yushe Basin (China), dated between 3.5 and 3.0 Ma (Lister et al., 2004). The material from Romania has been referred to a primitive mammoth species, *Mammuthus rumanus*, to which the Chinese finds may belong, whereas the other Asian specimens represent *Elephas planifrons*, an early form of the extant Asian elephant lineage. Although the precise details of the initial radiation of elephants in Eurasia are not entirely clear, present evidence suggests that the migration of *Mammuthus* and *Elephas* out of Africa could be part of a single middle Pliocene dispersal event. This event broadly coincides with the first Northern Hemisphere glaciation at ca. 3.2 Ma that represents a major climatic change that might have triggered the Ruscinian/Villafranchian faunal transition in Europe.

After its arrival, *Elephas* dispersed throughout southern Asia, including the Indonesian archipelago. A species close to *E. planifrons* entered South Sulawesi in the middle to late Pliocene (Walanae Fauna), giving origin to an endemic small-sized form, *E. celebensis*, characterized by the presence of mandibular tusks in males. During the early/middle Pleistocene transition, *Elephas* dispersed to Java, where it is represented by *E. hysudrindicus*, a form that is dentally very close to the living Asian elephant.

Mammuthus reached a much wider geographic distribution than *Elephas* in Eurasia, which could be related to its adaptation to a more open, grassy landscape. Although *Mammuthus* is already known in Eastern Europe and Asia at around 3.2 Ma, it is not recorded in Western Europe until ca. 2.7–2.3 Ma. The dispersal of *Mammuthus* toward Western Europe marks an important faunal turnover, the so-called Elephant-*Equus*



Figure 4 *Mammuthus* cf. *rumanus* from Montopoli, Lower Valdarno (Italy, early Villafranchian). Partial skull (maxillary and premaxillary) with M3. Scale bar represents 10 cm. Museo di Storia Naturale (Sezione Geologia e Paleontologia), Università di Firenze. Photo: L Rook.

event, characterized by the extinction of warm-loving, forest-dwelling mammalian species and the entry of open habitat taxa such as the mammoth and the horse (Azzaroli, 1983). This faunal event is coeval with the global climatic deterioration that occurred at the middle to late Pliocene transition at ca. 2.6 Ma. Early records of *Mammuthus* in Western Europe are at Montopoli (Italy) and the Red Crag (England), where *M. rumanus*, or a closely similar taxon, is represented (Figure 4). *M. rumanus* is replaced in younger deposits in Eurasia by *M. meridionalis*, one of the best-known Pliocene–Pleistocene elephants (Palombo and Ferretti, 2005; Lister and van Essen, 2005). This species is well represented in the latest Pliocene to early Pleistocene deposits from Central Italy (e.g., Upper Valdarno, Pietrafitta, Scoppito), where several complete skeletons have been found (Figure 5).

Mammuthus was also the sole elephant taxon to reach North America through Beringia. The first occurrence of *Mammuthus* in North America is at the Pliocene–Pleistocene transition, and characterizes the beginning of the Irvingtonian mammal age (Agenbroad, 2005). At present, the specific identity of this first American mammoth remains unclear, though it has commonly been considered close to *M. meridionalis*. Mammoths failed to colonize South America, although it was reached by gomphotheres during the Great American Biotic Interchange. It has been hypothesized that tropical Central America acted as an ecological barrier to the southward dispersal of grassland-adapted mammoths (Prado et al., 2005).

At the end of the late Villafranchian (early Pleistocene) elephants reached their most widespread distribution since their initial radiation in the Pliocene, becoming a characteristic component of the Quaternary fauna.

The Voles (*Microtus s.l.*) Radiation

The latest Pliocene–early Pleistocene period records an important event in the history of Northern Hemisphere mammal assemblages, that is the dispersal of modern voles with continuously growing teeth, referred to the genus *Microtus* (family Arvicolidae, subfamily Arvicolinae). This genus is



Figure 5 *Mammuthus meridionalis* from Borro al Quercio, Upper Valdarno (Italy, late Villafranchian). Skeleton of an old male individual. Museo di Storia Naturale (Sezione Geologia e Paleontologia), Università di Firenze. Photo: L. Rook.

the most diverse within the arvicolids, with many subgenera, species, and subspecies and with a wide Holarctic geographic distribution.

Voles are useful paleoecological, paleoclimatic, and paleogeographic indicators. They are common in fossiliferous deposits, where they usually occur in the form of highly resistant isolated teeth or dentitions, often derived from pellets regurgitated by birds of prey.

Dental morphology is used as the main taxonomic criterion for the classification of fossil species. The most variable, and therefore most useful anatomical structure in vole evolution, is the first lower molar (M_1). Paleontologists have created a classification based entirely on the characters of this tooth, with additional reference to other teeth such as the upper third molar (M^3). The evolution of this group is characterized by a gradual complication of the occlusal surface of the molars, with an increase in the length of the anterior part of the tooth (the so-called *anteroconid complex*) due to the addition of new elements, and also by changes in the relative thickness of the enamel walls. Many studies are focused on the Pleistocene radiation of these rodents, which are ideal subjects through which to examine various hypotheses of evolutionary tempo and mode (Martin, 1998; Chalain et al., 1999).

The primitive forms of rootless voles dispersed around 2 Ma throughout Eurasia and are grouped in the genus *Allophaiomys*, which is considered both as a subgenus of *Microtus*, and as a separate genus. However, recent findings of *Allophaiomys* from China trace the origin of this lineage back to 2.3–2.4 Ma (Zheng and Zhang, 2000). The name *Allophaiomys* was coined by Kormos (1932), who recognized the similarity of these fossil dentitions to those of the recent *Phaiomys* from Southeast Asia.

Allophaiomys is considered as a transitional form between Pliocene voles of the rooted genus *Mimomys* (the species *Mimomys tornensis* being the most likely ancestor to *Allophaiomys*) and recent representatives of *Microtus* 'sensu lato' (e.g., *Terricola*, *Stenocranius*, *Pallasinus*, etc. in Eurasia, *Pedomys* and *Pitymys*, etc. in North America). The change from a Pliocene fauna dominated by rooted voles to Pleistocene faunas dominated by rootless voles was interpreted as the result of increasing aridity during the Pliocene, which modified the vegetation and induced the adaptation of these rodents to feed on more abrasive food.

Allophaiomys deucalion is the first species seen in the fossil record, with a wide geographic range across Eurasia. It is reported from Romania, Poland, Turkey and Russia (see Hir, (1998), and from China (Cai and Li, 2004). *A. deucalion* probably dispersed towards North America through Beringia, as suggested by the occurrence of species with a primitive dental condition, for example, the extinct *Microtus cumberlandensis* (Martin, 1998).

The most successful radiation of *Allophaiomys* included the descendant of *A. deucalion*, *A. pliocaenicus*, which appeared in the early Biharian and quickly developed a Holarctic distribution (Figure 6). In North America, the forms derived from *A. pliocaenicus* evolved independently from the Eurasian species, with cases of parallel evolution with the Eurasian forms, as subgenera *Pitymys* and *Terricola*, in North America and Eurasia respectively.

During the late Biharian, in the late early Pleistocene, many sites document the appearance of more progressive forms of rootless Arvicolini. An important contribution to the knowledge of this phase of vole evolution comes from the assemblage found in the lower-middle levels at the Spanish sites of Atapuerca (Cuenca-Bescós et al., 1999), which record the Bruhnes-Matuyama boundary. Within these deposits is documented the radiation of *Microtus* 'sensu lato', with the presence of early representatives of lineages that ultimately led to modern forms, for example, *Stenocranius gregaloides*, *Iberomys huescarensis*, *Terricola arvalidens*.

Other sites in Europe, with a similar chronological level as Atapuerca, bear diverse *Microtus* 'sensu lato' faunas, which include Rifreddo (Italy), Mahlis (Germany), Unéč (Czech Republic), Kozi Grzbiet (Poland), Kovesvarad (Hungary), Karai Dubina and Shamin (Russia) (see papers in 'Quaternary International', volume 131, 2005).

An Overview on the Early Pleistocene Herpetofauna

The early Pleistocene amphibian and reptile faunas are not uniformly well-known at present. Even disregarding the bias due to the skewed geographic distribution of fossil-bearing sediments of Pleistocene age, striking differences occur

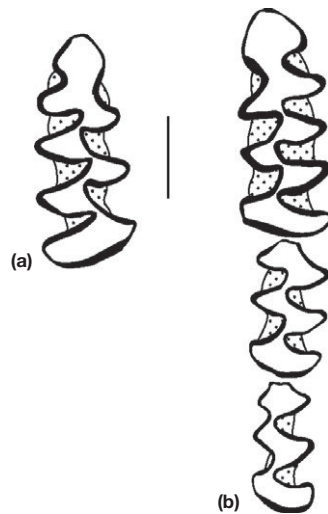


Figure 6 Primitive *Allophaiomys* dentitions from the Betsia complex type locality (Romania). (a) *Allophaiomys deucalion* holotype, occlusal view of right M_1 ; (b) *A. pliocaenicus* holotype, occlusal view of left lower molar series. Redrawn from Hír (1998). Scale bar represents 1 mm.

between the information available for the well-known herpetofaunas of Europe and North America (and to some extent also Australia) and for those largely unexplored from most of Asia and Africa. The former group is represented by a considerable number of localities and taxa corresponding to several 'time slices' of early Pleistocene age, while the latter is usually represented by comparatively few sporadic findings or by localities that have yielded hominoid and hominid remains.

Given such a disparity, definitive general conclusions cannot be built on a solid dataset and only some common trends can be tentatively reported.

The most remarkable characteristic of the early Pleistocene herpetofaunas is their modernity. Although several modern taxa (genera and even species) go back to the Neogene, it is with the onset of the early Pleistocene that the herpetofaunal assemblages become almost entirely composed of modern taxa. Holman (1998 and references therein), reviewing the rich Pleistocene record of North American and European amphibians and reptiles, concluded that Pleistocene extinctions are limited to a few isolated cases (for the rest of the planet, see datasets in 'Handbuch der Paläoherpetologie' volumes).

No families appear to go extinct in North America and only one extinct amphibian family (Paleobatrachidae) and one extinct genus (*Latonia*) inhabited Europe during the early Pleistocene (the only two ascertained herpetological extinctions at supra-specific rank in Europe during the whole of the Quaternary).

Even if a few extinct taxa have not yet been detected in the fossil record of some morphologically homogeneous cryptic taxa (e.g., ranid frogs, lacertid lizards, colubrid snakes), such stability sharply contrasts with the high turnover rate in mammals and birds during the Quaternary. The possible reasons for such disparity have been discussed by Holman (1998), who suggested that the apparent stability of the herpetofaunas could be related to low metabolic rates, reproductive strategies, and food-web relationships. Such stability could also be

related to the ability to bypass climatic stress by hibernation and estivation, to adapt to climatic and vegetation changes, to exploit ephemeral and unstable habitats, and to survive for generations in sub-optimal conditions, despite facing severe genetic bottlenecks.

The early Pleistocene evolutionary history of amphibians and reptiles, as indeed for the whole Quaternary, seems to be mostly characterized by range adjustments due to habitat tracking, resulting from the environmental disruptions triggered by climatic changes. Models of herpetofaunal successions have been developed for North America and Europe, since their relatively good fossil records and the climatic oscillations leading to the cyclic development of ice sheets allow, at least in some cases, patterns of dispersal or withdrawal to be established. Conversely, almost nothing can be said about the range variations of herpetofaunal taxa in tropical forests (e.g., during pluvial periods) and arid areas. This is partly because of lack of research but also because these areas are usually unsuitable for the preservation and retrieval of small vertebrates (most of the herpetofaunal species are small-sized).

As for the early Pleistocene representatives of modern taxa, the presence of large-sized and robust forms is particularly remarkable (e.g., the thick-shelled remains of the tortoise *Testudo hermanni* and the large and massive legless lizard *Pseudopus apodus* – formerly named *P. pannonicus*). In the past these were sometimes referred to as extinct taxa but they can most likely be considered as ecotypes, probably reflecting the development of particularly, but not necessarily, favorable, auto- and sinecological conditions.

The pattern depicted above has a partial but remarkable exception in the Australian herpetofauna. Probably because Australia did not experience the Plio–Pleistocene glacial–interglacial succession that characterized a large part of the world and because of its long term of isolation and peculiar faunal stock (Irrmin, 2005), the early Pleistocene Australian herpetofauna hosted, along with the modern reptilian and amphibian taxa, some peculiar, and in most cases, primitive reptiles that later died out before the end of the Epoch (Vickers-Rich et al., 1991; Molnar, 2004).

The most striking examples of these long-term survivors are represented by the mekosuchine crocodylian *Quinkana fortirostrum*, the giant horned turtle *Ninjemys* (Figure 7) or related meiolanids, the giant monitor lizard *Varanus priscus*, and the giant snake *Wonambi naracoortensis*. Worth mentioning is that *Quinkana* represents the last ziphodont crocodylian (characterized by blade-like serrated teeth) and *Wonambi* the last madtsoiid snake.

Despite some criticism concerning the 'reptilian domination' in the Australian Pleistocene, that according to Wroe (in Irrmin, 2005) underestimates the actual ecological role of marsupial carnivores, the size of *Varanus priscus* (up to about 7 m in total length and to nearly 2 tons body mass), remains unparalleled among all Australian terrestrial vertebrates and fully deserves a top place in the trophic relationships of the continental Pleistocene.

Future directions for improving our knowledge of the earlier stages of the Pleistocene herpetofauna, that is to say of the first steps of the modern herpetofaunal assemblages, should be the careful analysis of localities in Asia and Africa, as well as in Central and South America, although such study is



Figure 7 Skull of the giant horned turtle *Ninjemys oweni* from the undetermined Pleistocene of Darling Downs (Queensland, Australia). Anterodorsal view of the cast 1834–31 in the collections of Museum National d'Histoire Naturelle, Paris. Note the prominent lateral horns (snout pointing downward); maximum width 55 cm approximately. Photo (digitally modified): M Delfino.

hindered by the absence of osteological information for most of the taxa that nowadays inhabit those areas.

Acknowledgments

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See also: Vertebrate Overview. **Beetle Records:** Late Pleistocene of Northern Asia. **Vertebrate Records:** Mid-Pleistocene of Australia; Late Pleistocene of Africa; Late Pleistocene of North America; Late Pleistocene of South America; Late Pleistocene of Southeast Asia; Mid-Pleistocene of Africa; Mid-Pleistocene of Europe; Mid-Pleistocene of Southern Asia; Mid-Pleistocene of North America. **Vertebrate Studies:** Ancient DNA.

References

Agenbroad J (2005) North American Proboscideans: Mammoths: the state of knowledge, 2003. *Quaternary International* 126–128C: 73–92.

Alberdi MT and Aguirre E (1977) Round table on Mastrostratigraphy of the Mediterranean Neogene, Madrid 28 September–1 October 1976. *Trabajos sobre Neógeno-Cuaternario* 7: 1–47.

Azzaroli A (1977) The Villafranchian Stage in Italy and the Plio-Pleistocene boundary. *Giornale di Geologia* 41(1–2): 5–16.

Azzaroli A (1983) Quaternary Mammals and the End-Villafranchian Dispersal event – a turning point in the history of Eurasia. *Palaeogeography, Palaeoecology, Palaeoclimatology* 44: 117–139.

Azzaroli A (1992) The cervid genus *Pseudodama* n.g. in the Villafranchian of Tuscany. *Palaeontographia Italica* 79: 1–41.

Bergh GD and van den (1999) The late Neogene elephantoid-bearing faunas of Indonesia and their paleogeographic implications. *Scripta Geologica* 117: 1–419.

Cai BQ and Li Q (2004) Human remains and the environment of Early Pleistocene in the Nihewan Basin. *Science in China, Series D. Earth Sciences* 47(5): 437–444.

Chaline J, Brunet-Lecomte P, Montuire S, Viriot L, and Courant F (1999) Anatomy of the arvicoline radiation (Rodentia): Palaeogeographical, palaeoecological history and evolutionary data. *Annales Zoologici Fennici* 36: 239–267.

Cione AL and Tonni EP (1995) Chronostratigraphy and “land-mammal ages” in the Cenozoic of South America: Principles, practices, and the “Uquian” problem. *Journal of Paleontology* 69(1): 135–159.

Cuenca-Bescós G, Laplana C, and Canudo IJ (1999) Biochronological implications of the Arvicolidae (Rodentia, Mammalia) from the Lower Pleistocene hominid-bearing level of Trinchera Dolina 6 (TD6, Atapuerca, Spain). *Journal of Human Evolution* 37: 353–373.

Gabunia L, Lordkipanidze D, and Vekua A (2000) The environmental contexts of Early Human occupation of Georgia (Transcaucasia). *Journal of Human Evolution* 38: 785–802.

Gibbard PL, Head MJ, Walker MJC and the Subcommittee on Quaternary Stratigraphy (2010) Formal ratification of the Quaternary System/Period and the Pleistocene Series/Epoch with a base at 2.58 Ma. *Journal of Quaternary Science* 25: 96–102.

Giozzi E, Abbazzi L, Argenti P, et al. (1997) Biochronology of selected Mammals, Molluscs and Ostracods from the Middle Pliocene to the Late Pleistocene in Italy. The state of art. *Rivista Italiana di Paleontologia e Stratigrafia* 103: 369–388.

Hir J (1998) The *Allophaiomys* type-material in the Hungarian collection. *Paludicola* 2(1): 28–36.

Holman JA (1998) Pleistocene amphibians and reptiles in Britain and Europe. *Oxford Monographs on Geology and Geophysics no. 38*. Oxford: Oxford University Press.

Irmán RB (2005) Book review: Dragon in the dust. The paleobiology of the giant monitor lizard *Megalania*. *Journal of Vertebrate Paleontology* 25(2): 479.

Kormos T (1932) Neue wühlmause aus dem Oberpliozän von Püskpöckfürdő. *Neues Jahrbuch für Mineralogie, Geologie und Paläontologie* 69: 323–346.

Li C, Wu W, and Qiu Z (1984) Chinese Neogene Correlations: Subdivision and correlation. *Vertebrata Palasiatica* 22: 163–178.

Lindsay EH, Fahlbusch V, and Mein P (1990) *European Neogene Mammal Chronology*. NATO ASI, A. New York: Plenum Press.

Lister AM and van Essen HE (2003) *Mammuthus rumanus*, the earliest mammoth in Europe. *Advances in Vertebrate Paleontology, 'Hent to Pantha'*, pp. 47–52. Bucharest: Institut of Speleology of the Romanian Academy. Memorial Volume.

Lister AM, Sher AV, van Essen H, and Wie G (2004) The pattern and process of mammoth evolution in Eurasia. *Quaternary International* 126–128C: 49–64.

Marshall LG, Berta A, Hoffstetter R, et al. (1984) Mammals and stratigraphy: Geochronology of the continental mammal-bearing Quaternary formations of South America. *Palaeovertebrata, Mémoire Extraordinaire* 1–76.

Martin RA (1998) Time's arrow and the evolutionary position of *Orthiomys* and *Herpetomys*. *Paludicola* 2(1): 70–73.

Martínez-Navarro B and Rook L (2003) Gradual evolution in the African hunting dog lineage: Systematic implications. *Comptes Rendus Palevol* 2(8): 695–702.

Molnar RE (2004) *Dragon in the Dust. The Paleobiology of the Giant Monitor Lizard Megalania*. Indiana University Press.

Palombo MR and Ferretti MP (2004) Elephant fossil record from Italy: Knowledge, problems, and perspectives. *Quaternary International* (126–128C): 107–136.

Prado JL, Alberdi MT, Azanza B, Sanchez B, and Frassinetti D (2005) The Pleistocene Gomphotheriidae (Proboscidea) from South America. *Quaternary International* (126–128C): 21–30.

Rook L and Torre D (1996) The ‘Wolf event’ in western Europe and the beginning of the Late Villafranchian. *Neues Jahrbuch für Geologie und Paläontologie – Monatshefte* 8: 495–501.

Shoshani J and Tassy P (2005) Advances in proboscidean taxonomy and classification, anatomy and physiology, and ecology and behavior. *Quaternary International* (126–128C): 5–20.

Vickers-Rich P, Monaghan JM, Baird RF, and Rich TH (eds.) (1991) *Vertebrate paleontology of Australasia*. Design Studios with Monash University Publications Committee.

Woodburne MO, Tedford RH, Archer M, Turnbull WD, Plane MB, and Lundelius EL (1984) Biochronology of the continental mammal record of Australia and New Guinea. Special Publication. *South Australia Department of Mines and Energy* 5: 347–363.

Zheng SH and Zhang ZQ (2000) Late Miocene-Early Pleistocene micromammals from Wenwanggou of Lingtai, Gansu, China. *Vertebrata Palasiatica* 38: 58–71.

Early and Middle Pleistocene of Northern Eurasia

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Introduction

The best represented vertebrate records in the early and middle Pleistocene of Northern Eurasia are mammalian and they are summarized in this article. The data from Russia and neighboring countries of the former USSR are very important for the understanding of the history of Late Cenozoic biota. The territory was a prime area of origin and evolution of taxa and of dispersals of some mammals to Western Europe, Africa, North America and China, and vice versa. The configuration of Asia was close to that of the present day but differed through the periodic existence of the Bering land bridge, connecting Siberia and Alaska. The reorganization of the biota went through climatic and environmental changes; its main trends and phases are very pronounced there.

The key stages in the history of Pleistocene mammals in this territory were first recorded by Gromov (1948) who defined the faunal assemblages for Eastern Europe and named them by their type localities or regions. Regional complexes were later defined in the Caucasus, Siberia, Transbaikalia, and Central Asia. The data on the ages of the assemblages and their compositions were considerably enriched thanks to many subsequent researchers and new discoveries. In addition to large mammals, a detailed biochronology based on small mammals has appeared in the last decades. At present, early and middle Pleistocene mammal remains are known from many localities of Northern Eurasia. The age of the localities has been established by a combination of geological and paleontological methods (including the first appearance of index taxa, as well as the evolutionary levels of dominant forms and of some phyletic lineages), paleoclimatic data and correlation. Paleomagnetic data also help to distinguish more precisely the age of early Pleistocene sites, whereas the glacial–interglacial succession provides further important markers for age determination.

The beginning of the Pleistocene is currently set at ~2.6 Ma (the Gauss–Matuyama magnetic reversal), the early–middle Pleistocene boundary at ~0.78 Ma (the Brunhes–Matuyama magnetic reversal), and the middle–late Pleistocene boundary at 0.126 Ma, according to the International Commission on Stratigraphy in 2009. However, the Geological Surveys of Russia and Ukraine have until recently retained the Pliocene–Pleistocene boundary in its former position at 1.8 Ma. In 2012 the Russian Stratigraphic Committee voted to lower the base of the Quaternary to 2.58 Ma. The position of the lower boundary of the Pleistocene and its divisions have thus been repeatedly changed over the last decades and it must be appreciated that the published names of particular stratigraphical units often refer to different time slices.

The early and middle Pleistocene in Northern Eurasia currently include the following mammal assemblages: Khaprovian, 2.6–2.2 Ma; Psekupsian, 2.2–1.2 Ma; Tamanian, 1.2–0.8 Ma; Tiraspolian, 0.8–0.4 Ma; Singilian and Khasarian, 0.4–0.126 Ma.

Mammal Assemblages of Northern Eurasia

Early Pleistocene: Gelasian, Middle Villafranchian, Late Villanyian, MN17

The Khaprovian (Khapry) mammal assemblage and its analogs

This stage is well documented in the northern Black Sea region, peri-Azov area, Siberia, and Central Asia. The mammal communities superficially resembled those of modern Africa or southern Asia by the presence of proboscideans, equids, rhinoceroses, giraffids, and various antelopes, but differed considerably from them in species composition. These assemblages are characterized by the first appearance and subsequent wide distribution of the elephant *Archidiskodon gromovi* (= *A. meridionalis gromovi* in Titov, 2008, = *Mammuthus gromovi* of Lister and Sher, 2001). In contrast to Russian paleontologists, most researchers refer this species to the genus *Mammuthus*, mainly based on the tooth structure (Lister and Sher, 2001), although *Archidiskodon* is retained here. Other features of the fauna include the presence of the bear *Ursus etruscus* and the rhinoceros *Stephanorhinus* sp., the rapid diversification and wide distribution of stenonid horses, the presence of diverse canids (*Eucyon*, *Nyctereutes*, *Canis*), two saber-toothed cats *Homotherium crenatidens* and *Megantereon cultridens*, the camel *Paracamelus*, the large-sized comb-antlered deer *Eucladoceros* and the elk (= moose) *Libralces*, and other extinct forms. In Northern Eurasia, there were three paleogeographic subareas: European-Siberian, Mediterranean, and Central Asian (Vangengeim and Pevzner, 1991; Vislobokova et al., 1995) (Figure 1). The global cooling witnessed at about 2.6 Ma (indicated in the territory by the glaciations of the Caucasus and Pamir, the most ancient loess formation in Tadzhikistan, Uzbekistan, etc.) caused a decrease in humidity in temperate latitudes and in Central Asia (Dodonov, 2002), where animals adapted to open woodlands and grasslands became dominant. Some taxa have North American ancestry, immigrating during periods of low sea level prior to the start of the Pleistocene, such as the camels, *Hipparion* and *Eucyon* in the late Miocene, and *Equus*, *Canis*, and *Megantereon* in the Pliocene (Vislobokova et al., 2003). A number of thermophilous herbivores (mastodons, giraffids, and others), typical of Pliocene times, were gradually replaced by forms adapted to cooler climatic conditions and possessing more advanced herbivorous adaptations. The data on herbivorous taxa (elephants, horses, ruminant Artiodactyla, voles, etc.) are of great value for biochronology and paleoenvironmental reconstruction because their tooth structure yields clearly recognizable signals in wear patterns, hypsodonty, etc. The increase in diversity of herbivore taxa was accompanied by considerable changes in the carnivore guild.

At this time, an extensive European-Siberian part of the Palearctic was inhabited by the elephant *Archidiskodon gromovi*, the rhinoceros *Elasmotherium*, and large *Equus livenzovensis* and small *Equus* sp., together with other animals adapted to

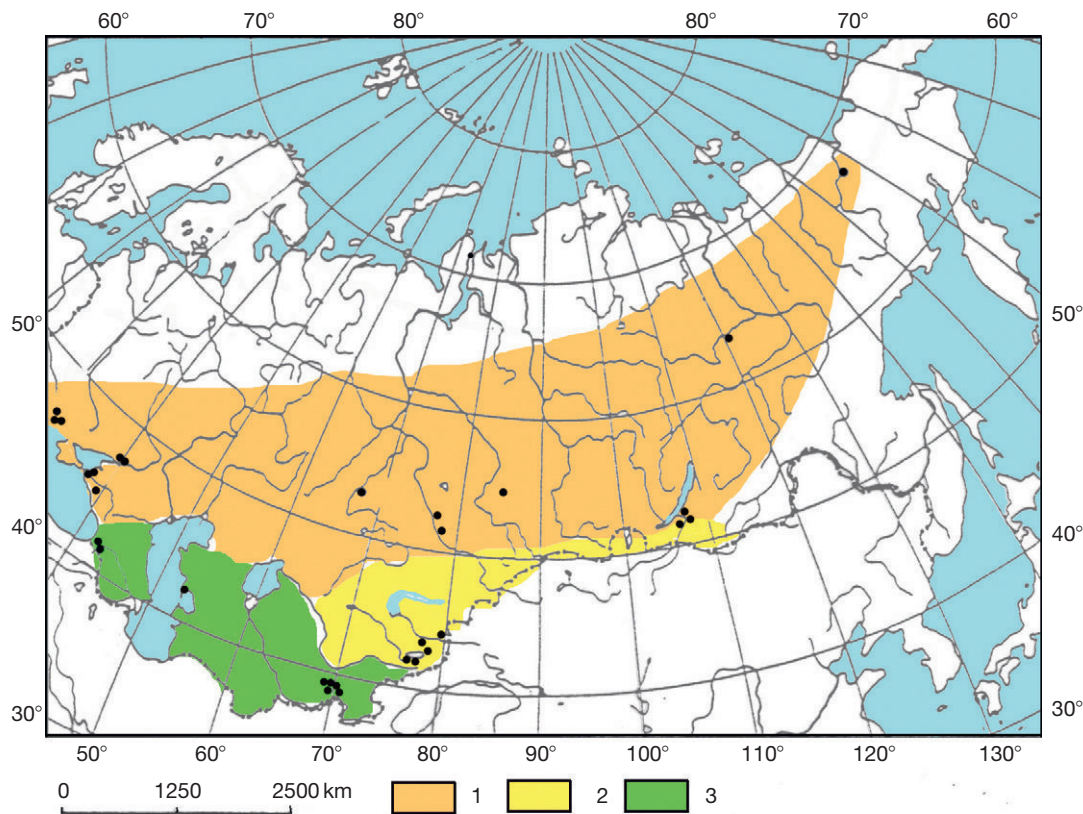


Figure 1 Early Pleistocene paleozoogeographic subareas and main localities: (1) European-Siberian subarea; (2) Central Asian subarea; (3) Mediterranean subarea.

savanna-like conditions (the wolves of the *Canis etruscus* group, the hyena *Pliocrocuta*, the saber-toothed cat *Homotherium crenatidens*, the cheetah *Acinonyx*, the camel *Paracamelus*, the antelope *Gazellospira*, the bison-like *Leptobos*, and others), along with forest and ecotonal animals (the bear *Ursus etruscus* and the deer *Eucladoceros* and *Libralces*). Most of these mammals were found in common with those from the Middle Villafranchian faunas of Western Europe (Kahlke et al., 2011; Palombo et al., 2006). Early Villafranchian elements (the mastodon *Anancus arvernensis*, the archaic horse *Hipparion*, and giraffids) became very rare. Among the small mammals, the bank voles *Clethrionomys* made their first appearance, the water vole *Mimomys polonicus* was replaced by *M. pliocaenicus* in the Northern Black Sea area, and the pikas *Pliolagomys* and *Ochotona* and jerboas *Allactaga*, more typical of Asian faunas, were present.

Fossil remains of *Archidiskodon gromovi*, the main indicator species of the Khaprobian mammal assemblage, have been found in a number of localities in the European part of Russia (Khapry, Liventzovka, and others), in the Ukraine (Zhevakhova Gora), in the south of Western Siberia (Podpusk-Lebyazh'e), and in Central Asia (Adyrgan, Kuruksay) (Bajgusheva, 1971; Garutt and Bajgusheva, 1981; Vangengeim et al., 1988; Vislobokova, 1996, 2005; and others). *A. gromovi* has also been found in Italy (Montopoli) (Gliozzi et al., 1997).

Remains of other members of the Khaprobian assemblage are recorded from many European-Russian sites along the northern coast of the Sea of Azov and the right bank of the

Don River near Rostov-on-Don and Taganrog. They come from the lower part of the Khapry alluvium recorded in sand pits and natural outcrops. The large mammals were represented by *Anancus arvernensis alexeevae*, the horses *Hipparion moritorium*, *Equus* (*Allohippus*) *livenzovensis*, and a small *Equus* sp., the rhinoceroses *Stephanorhinus* ex gr. *megarhinus-kirchbergensis* and *Elasmotherium chaprovicum* (= *E. cf. caucasicum*), the wild boar *Sus strozzi*, the camels *Paracamelus alutensis* and *Paracamelus cf. gigas*, the deer *Cervus* (*Rusa*) *philisi*, *Eucladoceros cf. dicranios*, *Arvernoceros* sp., and *Libralces gallicus*, the giraffid *Palaeotragus* (*Yurlovia*) *priasovicus*, and the bovids *Leptobos* sp., *Gazellospira gromovae*, *Tragelaphini* gen. indet., *Gazella cf. subgutturosa*, and others (Alexeeva, 1977; Bajgusheva, 1971; Bajgusheva et al., 2001; Gromov, 1948; Titov, 2008; and others) (Figure 2). The Khapry carnivores included the raccoon-dog *Nyctereutes megamastoides*, *Canis* sp. (*C. cf. senezensis*: Sotnikova et al., 2002), the mustelids *Lutra* sp. and *Pannonictis nesti*, the medium-sized hyena *Pliocrocuta perrieri* and the dominant larger-sized short-faced hyena *Pachycrocuta brevirostris*, the felids *Lynx issiodorensis*, the saber-toothed cat *Homotherium crenatidens*, and the cheetah *Acinonyx pardinensis*. Most of these animals were inhabitants of open savannah-like landscapes, with a part of the fauna adapted more to forested or ecotonal zones (mustelids, lynx, wild boar, and deer). Among the small mammals, the archaic lagurine *Borsodia praeungarica*, the water voles *Mimomys praepliocaenicus* and *Mimomys* ex gr. *reidi*, and the primitive red-backed voles *Clethrionomys kretzoi* were also present (Tesakov, 2004).



Figure 2 *Anancus arvernensis*, C. Flerov painting exhibited in the Paleontological Museum of the Paleontological Institute, Russian Academy of Sciences (PIN) in Moscow, with PIN permission.

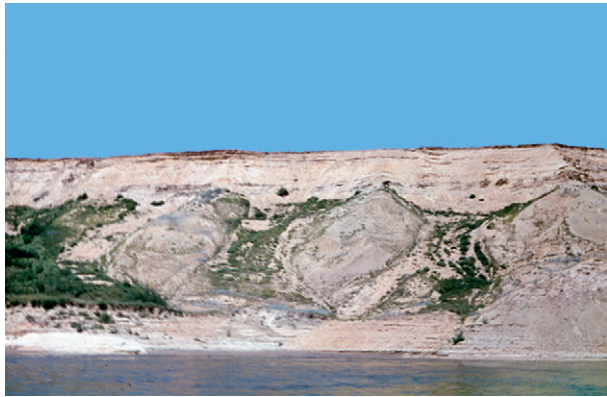


Figure 3 Podpusk–Lebyazh'e locality at the right bank of the Irtysh River, Western Siberia.

In the south of Western Siberia, faunal assemblages described by Vangengeim and Zazhigin have been reported from Podpusk-Lebyazh'e, from the lower part of the thick sandy alluvial deposits exposed on the right bank of the Irtysh River, south of Pavlodar (Vangengeim, 1977; Vislobokova, 1996; Figure 3). Typical members include *A. gromovi*, *S. cf. etruscus*, *Elasmotherium* sp., *Equus livenzovensis*, *Equus* sp. (small), *Paracamelus* cf. *gigas*, *Eucladoceros* sp., *Leptobos*, and the antelopes *Gazella* cf. *sinensis* and *Antilospira* cf. *gracilis* (Vislobokova, 1996). The faunal assemblage consisted mainly of forms widespread in the temperate zone of the Palearctic region (*Ursus* cf. *etruscus*, *Homotherium*, *Pachycrocuta*, and others), with some Central Asian elements (*Ochotonoides*, *Elasmotherium*, *Paracamelus*, *Antilospira* cf. *gracilis*, *Gazella* cf. *sinensis*), some of which (e.g., *Paracamelus*) reached Eastern Europe. Among the small mammals, voles with rooted dentition of the genus *Mimomys* (*M. pliocaenicus* and *M. reidi*) and *Borsodia* (*B. praeungarica-petenyii*) were predominant and *Clethrionomys* appeared for the first time in Siberia (Zazhigin, 1980). The mammalian assemblages indicate the prevalence of steppes with woodland developed along river valleys.

In northeastern Asia, the association of *Synaptomys*-like lemmings *Pliotomys* and mimomyian voles *Cromeromys* ex gr. *irtyshensis-hordijki* and smaller *Mimomys* sp. were found in the lower member of the Kutuyakh Formation at the Krestovka River in the Kolyma Lowland (Sher, 1987; Tesakov and van Kolfshoten, 2011). The assemblage is dominated by lemmings and 'Cromeromys' voles, with *Ochotona* sp. and *Mimomys* sp. being less abundant.

Several faunas of this age have been referred to the Central Asian subarea (Erbaeva and Alexeeva, 2000; Sotnikova et al., 1997; Vangengeim, 1977; Vislobokova et al., 1995). These include the Itanza fauna of Klochnevo I, II (Transbaikalia) with *Ochotona intermedia*, the ground squirrel *Spermophilus itancinicus*, *Allactaga* sp., the hamster *Cricetinus varians*, the voles *Mimomys pseudintermedius*, *Clethrionomys* sp., and 'Villanyia' klochnevi, the archaic mole rat *Prosiphneus* cf. *paringi*, the first representative of the giant deer genus *Praemegaceros* and others, as well as the Kiikbai and Andyrkan faunas (southern Kazakhstan) with *Ochotonoides complicidens*, *Mimomys pliocaenicus*, the gerbil *Meriones* cf. *meridianus*, *Equus stenonis*, the camel *Gigantocamelus longipes*, and other forms.

The fauna of Southern Tadzhikistan has affinities with the Eastern Mediterranean subarea and is characterized by a combination of Mediterranean and Central Asian forms. The rich assemblage from Kuruksay contained, along with *A. gromovi* and *Mastodontoidea* fam. et gen. indet., the baboon *Papio sushkini* (= *Paradolichopitecus sushkini*), the porcupine *Hystrix* sp., diverse carnivores including *Nyctereutes megamastoides*, eucyon 'Canis' kuruksaensis, *Ursus* cf. *etruscus*, the cursorial hyena *Chasmaporthetes lunensis kani*, *Pliocrocuta perrieri*, *Lynx* ex gr. *issidorensis*, *Acinonyx* cf. *pardinensis*, *Megantereon megantereon*, and *Homotherium crenatidens*, the rhinoceros *Stephanorhinus* sp., the horse *Equus stenonis pamirensis* (= *E. s. bactrianus*), the camel *Paracamelus praebactrianus*, the deer *Axis flerovi* and *Elaphurus eleonora*, the first representative of the giant deer genus *Sinomegaceros* (*S. tadzhikistanis*), the elk *Libralces* cf. *gallicus*, the giraffids *Sogdianotherium kuruksaense* and *Sivatherium*, and the antelopes *Gazella parasinensis*, *Protorix paralaticeps*, *Damalops palaeindicus*, *Gazellospira gromovae*, and *Antilospira* sp., amongst others (Dmitrieva, 1977; Vangengeim et al., 1988). In this region, a large number of Asian elements occurred. Animals of dry, open landscapes (*Ellobius*, *Pliocrocuta*, *Chasmaporthetes*, *Paracamelus*, *Equus*) were accompanied by forest animals (bear, lynx, monkey, the deer *Axis* and *Libralces*) and savannah inhabitants (*Sivatherium*, *Protorix*, *Damalops*, *Gazellospira*). Such a mixed composition reflected the mosaic landscapes and vertical zonality typical of mountainous regions. In Tadzhikistan, the local faunas attributed to this assemblage are known also from Obigarm, Karamaidan, Tutak, and Zil'fi localities, while faunas close to them in age were found in Kyrgyzstan (Akterek, Dzylygdykoo) (Dmitrieva and Nesmeyanov, 1982; Sotnikova et al., 1997).

The reversed polarity of the fossiliferous deposits in Khapry indicates that these sediments should be attributed to the lower part of the Matuyama Chron, together with analogous sites in Siberia and Central Asia (Podpusk-Lebyazh'e, Krestovka, Kuruksay) (Sotnikova et al., 1997; Tesakov et al., 2007; Vangengeim and Pevzner, 1991; Vangengeim et al., 1988; Vislobokova, 1996).

Late Early Pleistocene: Late Villafranchian, MN18–MQ19, = Tiglian–Menapian

Psekups mammal assemblage and its analogs

This stage is characterized by significant changes in the diversity of mammal communities (Figure 4). In Northern Eurasia, a cold-climate event around 2.2–1.9 Ma resulted in the appearance of new boreal elements, including some modern genera. In the European-Siberian subarea, *Archidiskodon gromovi* was replaced by the more progressive *A. meridionalis* (= *Mammuthus meridionalis*), close to the type of the species from the Upper Valdarno, Italy. Voles of the genus *Allophaiomys*, the giant deer *Praemegaceros*, the elk *Alces*, the large bovids of the genus *Bos* and the *Bison* (*Eobison*) – *B. (Bison)* phyletic lineage, and the musk-ox *Soergelia* first appeared in this stage.

In Psekups (Northern Ciscaucasia), *Archidiskodon meridionalis*, *Eucladoceros orientalis* (= *Psekupsoceros orientalis*, first recognized as *Cervus pliotarandoides*), *Stephanorhinus etruscus*, and *Equus stenonis* were present (Alexeeva, 1977; Gromov, 1948; Vislobokova, 1990; and others). The small mammals included *Borsodia newtoni-arankoides*, *Pitymimomys pitymyoides*, *Mimomys* cf. *pliocaenicus*, *Mimomys reidi*, and *Clethrionomys kretzoi* (Alexandrova, 1976; Tesakov, 2004). These reversed polarity deposits correspond to the Matuyama Chron, between the Reunion and Olduvai Subchrons (Tesakov, 2004; Vangengeim and Pevzner, 1991). Slightly younger assemblages from the Odessa small mammal complex contain hypsodont *Mimomys* forms (dominant *M. reidi* and *M. ex gr. savini*), the last representatives of *Borsodia*, the first occurrence of lagurines with

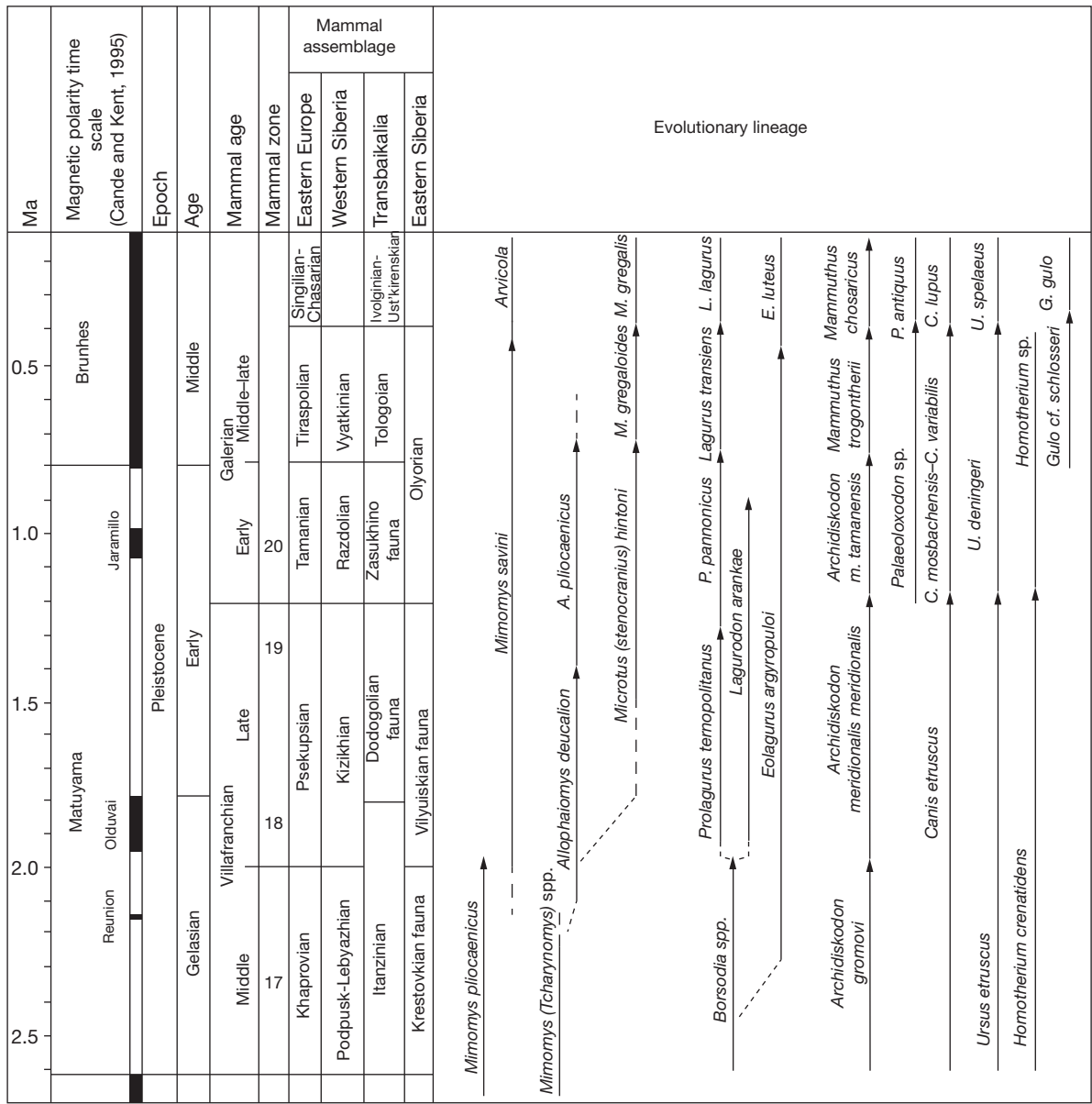


Figure 4 Mammal assemblages and faunas of Northern Eurasia and the main lineages.

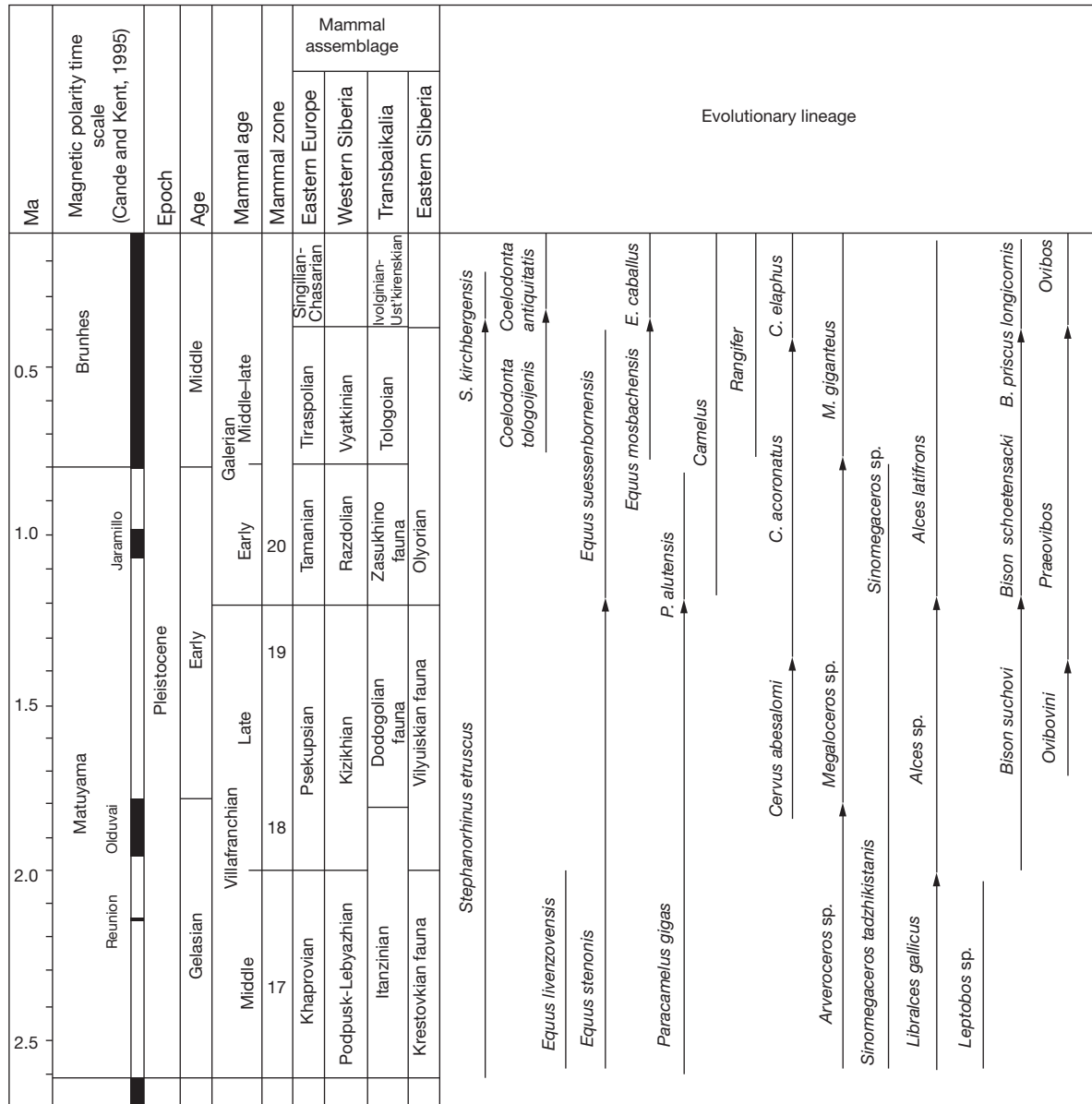


Figure 4 (Continued)

unrooted teeth *Prolagurus* and *Lagurodon*, and *Allophaiomys deucalion* (Alexandrova, 1967; [Rekovets and Nadachowski, 1995](#); [Tesakov, 2004](#)). The Odessa assemblage is subdivided into two phases: (1) an early phase, with the remains of rootless voles *Allophaiomys deucalion* and (2) a late phase, with the first cemented rootless voles of *Prolagurus* (*P. ternopolitanus*) and *Lagurodon* (*L. arankae*) ([Markova, 1982, 2007](#)).

In the Caucasus, the Dmanisi fauna is famous for the presence of the first appearance in Eurasia of *Homo*, following dispersal from Africa. The remains of the Dmanisi hominins are attributed to the *Homo erectus* group and were found in the layers directly overlying a lava flow with a radiometric date of 1.8 ± 0.1 Ma, within the Olduvai Subchron. The fauna comprises inhabitants of dry, open area (the tortoise *Testudo*, the ostrich *Struthio*, the gerbil *Parameriones*, the pika *Ochotona*, the hamster *Cricetulus*, *A. meridionalis*, *Canis etruscus*, *Pliocrocota*

cf. perrieri, *Megantereon cultridens*, *Homotherium crenatidens*, *Palaeotragus*) and forest animals (*Bison* (*Eobison*) *georgicus*, the deer *Cervus abesalomi* and *Dama nestii*, the bear *Ursus etruscus*) ([Gabunia and Vekua, 1995](#); [Hemmer et al., 2010](#); [Vekua, 1995](#); [Vekua and Lordkipanidze, 2008](#)). The first occurrences of the 'jaguar' *Panthera onca georgica* and musk ox *Soergelia* are recorded there. The contemporaneous Palan-Tyukan fauna (Azerbaijan) includes *Canis etruscus*, *Panthera ex gr. onca gom-baszoegensis*, and *Nyctereutes megamastoides*.

In Western Siberia, *A. cf. meridionalis*, *Equus ex gr. stenonis*, *Alces sp.*, and *Gazella sp.* along with *Borsodia ex gr. petenyi-hungaricus*, and *Mimomys coelodus* are known from the Podpusk fauna (from the uppermost part of the sands in the Irtysh River Basin) ([Vislobokova, 1996](#)). In southern Kazakhstan, the Aktogai (Kopaly) locality with *Mimomys* (*Tcharynomys*) *haplo-dentatus* (bed 3) and *Allophaiomys deucalion* (bed 7) also

corresponds to this age (Sotnikova et al., 1997; Tjutkova and Kaipova, 1996).

In the Kuznetsk Depression, *Allophaiomys pliocaenicus*, *Prolagurus* ex gr. *pannonicus-posterius*, *Eolagurus argyropuloi*, *Archidiskodon* cf. *meridionalis*, and *Ovibovini* gen. indet. were found in the Mokhovo Formation, 1.8–1.2 Ma (Foronova, 2001).

In northeastern Asia, a tooth of *Archidiskodon meridionalis* was found in the Vilyuisk district, Yakutia; the find confirmed the spread of the subspecies into North America over the Bering land bridge (Dubrovo, 1990), probably at the end of the 2.9–2 Ma global sea level fall.

In Transbaikalia, the Dodogol and Zasuchino II faunas have been referred to the Central Asian subarea. In Dodogol, *Borsodia laguroformis*, *Allophaiomys* cf. *pliocaenicus*, *Prosiphneus youngi*, the first woolly rhino *Coelodonta* (of Asian origin), *Equus* ex gr. *sanmeniensis*, and others were present (Vangengeim, 1977). The revised data on the Pleistocene small mammals of the region are presented in Alexeeva et al. (2007) and Alexeeva and Erbaeva (2008).

Terminal Early Pleistocene: Early Galerian, 1.2–0.8 Ma

The time range from 1.2 to 1.1 Ma (before the Jaramillo paleomagnetic reversal) is an important boundary in the history of the biota, and is reflected in the nomenclature of stratigraphic schemes (the Villafranchian–Galerian boundary or the Early–Late Eopleistocene boundary in the scheme of the Russian Stratigraphic Committee).

Tamanian mammal assemblage and its analogs (Nogaitsky–Morozovka, = Menapien–Bavelian)

This stage is characterized by the replacement of *Archidiskodon meridionalis meridionalis* by the more advanced *A. meridionalis tamanensis*, the first appearance of the giant deer *Praemegaceros verticornis* and *Megaloceros*, the red deer *Cervus* (*Cervus*) *acoronatus*, the roe deer *Capreolus* cf. *suessenbornensis*, the broad-fronted elk *Alces latifrons*, the bison *Bison* (*Bison*), camels of the genus *Camel*, the first occurrence of the woodland elephant *Palaeoloxodon* and woodland antelopes *Tragelaphus* and *Pontoceros*, and the presence of *Paracamelus* (*P. kujalnensis*) (Alexeeva, 1977; Gromov, 1948; Vekua, 1962; Vereshchagin, 1957; and others). From the end of the early to the middle Pleistocene, true wolves attributed to the *Canis mosbachensis*–*C. variabilis* group and large hunting dogs *Lycaon lycaonoides* (= *Canis* (*Xenocyon*) *lycaonoides*) existed in Eurasia from Western Europe to Transbaikalia and China (Sotnikova and Rook, 2010).

The Tamanian mammal assemblage is based on the fauna of Synaya Balka (type locality) and Tsymbal in the Taman Peninsula. *A. meridionalis tamanensis* coexisted in the North Black Sea area with *Homotherium crenatidens*, the short-faced hyena *Pachycrocuta brevirostris*, the horse *Equus suessenbornensis*, the rhinoceroses *Stephanorhinus etruscus* and *Elasmotherium caucasicum*, and others (Alexeeva, 1977; Sotnikova and Titov, 2009; Vereshchagin, 1959; Figure 5). Vereshchagin (1957) also identified the potential impact of early hominins on some of these assemblages. And recently, stone artifacts of the Oldowan type were reported from the Synyaya Balka site by Shchelinsky et al. (2010). In the Asian part of the territory, the first reliable evidence of human activity comes from the stone tool assemblage of Kuldara (Ranov et al., 1995) dated to

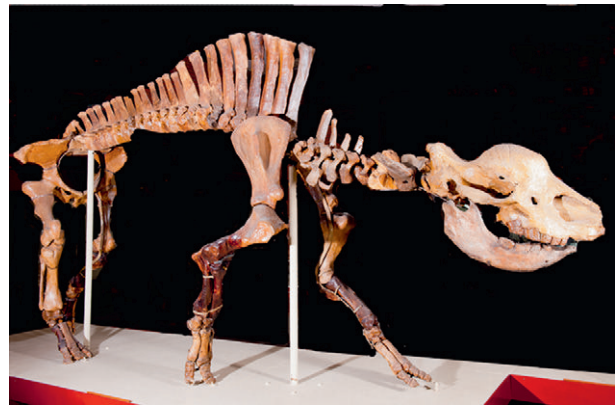


Figure 5 A skeleton of *Elasmotherium* on display in the Paleontological Museum of the Paleontological Institute, the Russian Academy of Sciences (PIN), Moscow, with PIN permission.

0.8 Ma and broadly synchronous with stone tools from the Karama site in the Altai region (Derevyanko and Shunkov, 2005).

Morozovkian small mammal fauna with more derived voles *Microtus* (*Stenocranius*) *hintoni* and the pitiomyoid *Microtus* sp. are correlated with these assemblages (Alexandrova, 1976; Rekovets and Nadachowski, 1995). The small mammal fauna of this stage are also represented in the Dnieper, Dniester, and Don basins (Agadjanian, 2009; Alexandrova, 1976; Markova, 2007; Tesakov et al., 2007; Topachevsky et al., 1987; and others). The assemblages include an advanced meadow vole *Allophaiomys* (*A. pliocaenicus*) and the steppe lemmings *Prolagurus pannonicus*, *Lagurodon arankae*, and *Eolagurus argyropuloi*.

At this time, differentiation of environment sharply increased. In the Caucasus (Akhalkalaki, Georgia), the most thermophilous elements continued to exist, including *Hippopotamus georgicus*, *Equus suessenbornensis*, and *Stephanorhinus hundsheimensis*, the cave or ‘spelaeoid’ bear close to *U. deningeri*, and the first jaguar (close to *P. onca gomboszoegensis*) (Hemmer et al., 2010; Vekua, 1962, 1986). In Siberia, the diversity of musk-oxen increased. The ovibovine *Soergelia* occurred in the Razdolie assemblage (Western Siberia) with *A. meridionalis*, *Palaeoloxodon* sp., *Equus* (*Allohippus*) sp., *Bison* sp., and *Ovibovini* (Vangengeim, 1977). The remains of *Soergelia* and the oldest members of *Praeovibos* were found in the early stage of the Olyorian land mammal age (Beringia, Bering Province) (1.4–0.8 Ma) together with those of the first wolverine *Gulo minor* (= *G. cf. schlosseri*) (Sher, 1971, 1987; Sher et al., 2011; Vangengeim, 1977). These co-occurred with *Archidiskodon* sp., *Lycaon lycaonoides*, *Equus* (*Plesippus*) *verae*, *Alces* sp., *Bison* sp., and others. The northernmost finding of *Homotherium* in Asia is associated with this fauna. In Eastern Siberia, the Aldan fauna included a number of forest inhabitants (*Trogontherium* cf. *cuvieri*, *Palaeoloxodon* ex gr. *namadicus*, *Alces latifrons*, and *Microtus gregaloides*). In Transbaikalia, the Kudun (Kizhing–Kudun depression) and Zasuchino (Itanza River basin) correspond to the youngest part of the early Pleistocene and are correlated with the uppermost Matuyama Chron (Erbaeva and Alexeeva, 2000).

The oldest *Camelus* (*C. cf. knoblochi*) coexisted with the giant deer *Praemegaceros* (close to *P. verticornis*) in southern

Tadjikistan (Lakhuti locality, the uppermost part of the Matuyama) and members of diverse habitats. They included inhabitants of open, dry, and savannah-like landscapes, such as the gerbil *Meriones lacutensis* and mole vole *Ellobius lakhutensis*, *Archidiskodon* sp., *Equus* cf. *namadicus*, and *Canis* cf. *mosbachensis*, in association with a large *Lycaon*, *Pachycrocuta brevirostris*, *Homotherium* sp., the red-backed vole *Clethrionomys*, *Panthera onca gomboszoegensis*, the badger *Meles* ex gr. *meles*, and the giant deer *Sinomegaceros* sp. (Sotnikova, 1989; Sotnikova and Vislobokova, 1990). Small mammals also included white-toothed shrew *Crocidura* sp., a hamster *Cricetulus* sp., the meadow voles *Allophaiomys* sp., and *Microtus* (*Phaiomys*) *lakhutensis* (data from Zazhigin).

Faunas intermediate in age between the Tamanian and Tiraspolian faunas are known from the lower course of the Dnester River (Karai-Dubina) and in the middle Don (Petro-pavlovka) area and have been studied in detail (Agadjanian, 2009; Markova, 2007). The reversed polarity deposits contain the remains of *Prolagurus* and pitomyoid *Microtus*, the last representatives of *Allophaiomys*, and the oldest *Microtus* with five closed enamel triangles in the first lower molar.

The dispersal of mammals out of Northern Eurasia increased from about 1.2 to 1.1 Ma; *Praemegaceros verticornis* and *Soergelia* invaded Western Europe and *Homo erectus* and *Sinomegaceros* first occurred in China (Kahlke et al., 2011; Qiu, 2006).

Early Middle Pleistocene: Middle–Late Galerian, = Cromerian–Elsterian, 0.8–0.4 Ma

The early–middle Pleistocene boundary of the standard scheme corresponds to the Eopleistocene–Neopleistocene boundary of the Russian stratigraphic scheme and the Early–Middle Galerian boundary in Italy.

Tiraspolian mammal assemblages and its analogs

The considerable reorganization in the composition of mammalian communities was related to a drop in temperature at the early–middle Pleistocene boundary, with a notable increase in amplitude of climatic oscillations. Along with progressive cooling, climatic fluctuations changed from 41-ka low-amplitude cycles to a 100-ka high-amplitude cycle (Shackleton, 1995). In the Russian Plain, three periods of glaciation correspond to this stage (Pokrovka, Don, and Oka). The climatic zonation of the northern Eurasian landscape began to resemble that of the present day (Figure 6). *Archidiskodon* was replaced by the mammoth *Mammuthus trogontherii* (= *A. wüsti*) and the proportion of modern vertebrate genera sharply increased.

The representatives of the Tiraspolian mammal assemblage in southeastern Europe existed under a temperate climatic regime in a variety of landscapes, dominated by forest-steppes (Alexeeva, 1977). Kolkotova Balka (near Tiraspol) is the type

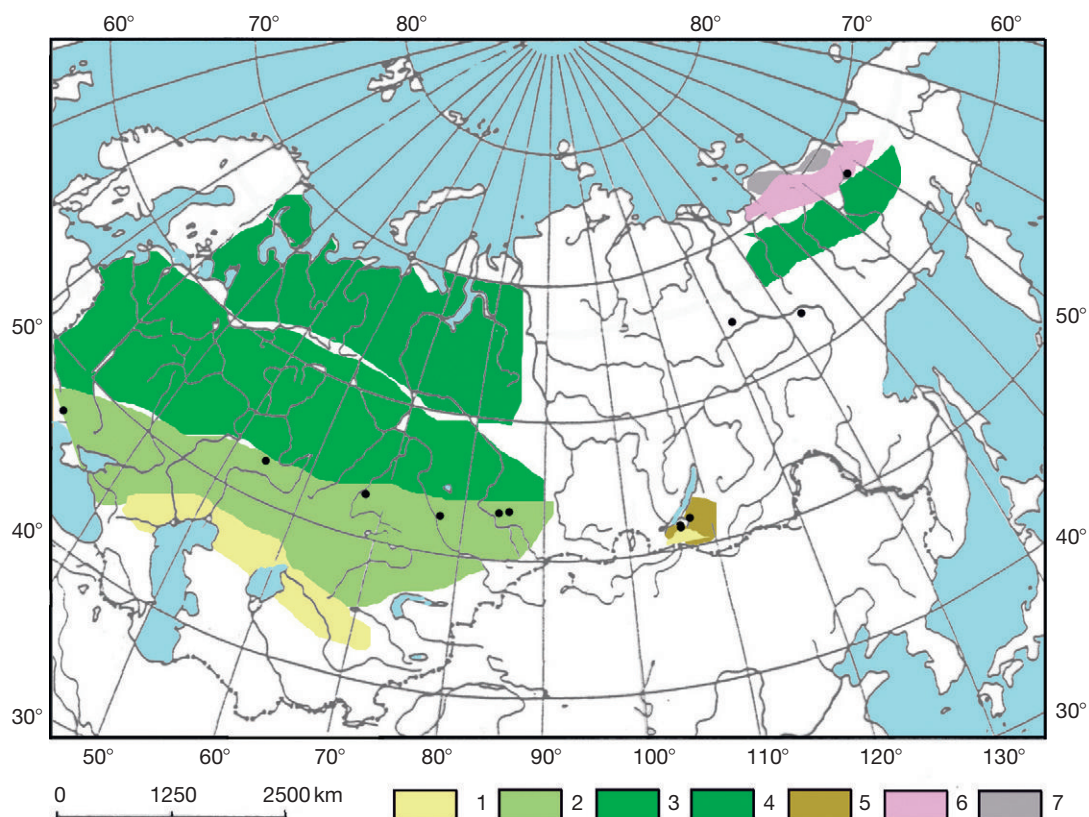


Figure 6 Zonal occurrence of mammals in the early–middle Pliocene, with some data from Markova (2006): (1) steppe; (2) forest-steppe; (3) mixed and broadleaved forests; (4) *Picea-Pinus* and *Betula* forests, with some broadleaved plants; (5) *Picea-Pinus* forests and steppe; (6) forest–tundra; (7) tundra.

locality of this assemblage. The fossil remains were found in the so-called 'Tiraspolian gravels' (fluvial deposits of the Dniester 6th terrace). Horses were represented by both caballoid (*Equus* cf. *mosbachensis*) and stenois (*E. aff. suessenbornensis*) forms. *Mammuthus trogontherii*, *Stephanorhinus etruscus* (late form), the large deer *Cervus acoronatus*, *Praemegaceros verticornis*, the fallow deer *Praedama* and *Alces latifrons*, and a short-horned bison *Bison schoetensacki* are common to both the Tiraspolian fauna and the fauna of Western Europe (Kahlke et al., 2011; Nikiforova, 1971). Two skeletons of *M. trogontherii* were found in the Kagalnik sand pit (Azov). Data on small mammal fauna represent several phases of Tiraspolian mammal assemblages (see Markova, 2007). A considerable cooling has been recognized in the southern part of European Russia at the time of the early middle Pleistocene Don glaciations, highlighted by the presence of subarctic species such as the lemmings *Lemmus* ex. gr. *Sibiricus* and *Dicrostonyx* sp. and the North Siberian vole *Microtus* ex gr. *hyperboreus* (Agadjanian, 2009).

The southward shifting of ranges, with persistence of some forms in refugia (including mountain ones), was usual during continental glaciations. Very diverse mammal faunas are known from the northern Caucasus. Most are associated with Early Paleolithic sites and stone artifacts of Acheulian type. The taxa include the macaque monkey *Macaca* cf. *sylvana*, the spotted hyena *Crocota spelaea*, diverse bears including cave bear *Ursus deningeri* (dominant), black bear *Ursus* ex gr. *thibetanus* probably related to *U. etruscus*, and the small cave bear *U. savini rossicus* (= *U. aff. spelaeus rossicus*), the cave lion *Panthera spelaea* and leopard *P. pardus*, the gray wolf *Canis lupus*, the fox *Vulpes vulpes*, *Stephanorhinus*, *Megaloceros giganteus*, *Capreolus* cf. *suessenbornensis*, *Bison* cf. *schoetensacki*, together with typical montane genera such as the ibex *Rupicapra* and the goat *Capra*.

The fauna of the late stage of the Olyorian faunal assemblage (West Bering Province) are characteristic of tundra and forest-tundra environments and cold climatic conditions similar to modern ones (Sher, 1971, 1987; Vangengeim, 1977). This fauna included the first reindeer *Rangifer*. *Alces* aff. *latifrons* and *Bison* sp. were also present in the artiodactyl community. *Rangifer* is also recorded in the early middle Pleistocene of the Kuznetsk Depression (Foronova, 2001).

In Transbaikalia, the Tologoi assemblage contained the woolly rhino *Coelodonta tologojensis*, *Equus* ex gr. *sanmeniensis*, *Cervus* ex gr. *elaphus*, *Spirocerus* cf. *peii*, and *Bison* sp. (Vangengeim, 1977). Cryogenic disturbance at the base of the deposits is associated with remains from the Zasuhino assemblages (Zasuhino 3, 1–0.78 Ma) and in Tologoi 2 beneath the Brunhes–Matuyama magnetic reversal, above the Jaramillo (Alexeeva, 2005). The fauna is characterized by a large diversity of carnivores and ungulates: *Nyctereutes* sp., *Canis variabilis*, *Lycan* cf. *lycaonoides* (= *Xenocyon lycaonoides*), *Ursus* sp., *Gulo* sp., *Pachycrocuta brevirostris*, *Homotherium* sp., the large-sized tiger *Panthera* ex gr. *tigris*, *Coelodonta tologojensis*, *Equus* ex gr. *sanmeniensis*, *Equus* ex gr. *suessenbornensis-verae*, *Alces* cf. *latifrons*, and others (Vangengeim et al., 1990). Zasuhino 3 locality also has yielded *Cervus* sp., *Megacerini* indet., *Capreolus* cf. *suessenbornensis*, *Ovibovini* indet., and *Bison* sp.

At the boundary of the early and middle Pleistocene, a prominent climatic deterioration was marked over northern and central Eurasia and many immigrants from these regions

are recorded in Western Europe. The Bering land bridge (0.8–0.011 Ma) further facilitated faunal exchanges between Northern Asia and North America.

Late Middle Pleistocene, = Aurelian, = Holsteinian–Saalian, 0.4–0.126 Ma

Singilian and Khasarian faunas and their analogs

The Singilian fauna and its analogs are related to the Likhvin interstadial (= Mindel-Riss, = Holsteinian), one of the most pronounced warm stages in the middle Pleistocene. In Eastern Europe, the Singilian faunas were represented by the straight-tusked elephant *Palaeoloxodon antiquus* and Merck's rhinoceros *Stephanorhinus kirchbergensis*. In the south of Northern Eurasia, gray wolf (*Canis lupus*), *Camelus knoblochi*, *Megaloceros giganteus*, and *Bison priscus longicornis* became widespread after the mid-middle Pleistocene (Alexeeva, 1977; Gromov, 1948). Typical small mammals include the water vole *Arvicola mosbachensis*, the yellow steppe lemming *Eolagurus luteus*, the narrow-skulled vole *Microtus gregalis*, the northern or root vole *M. oeconomus*, *Ellobius*, and the ground squirrel *Spermophilus* (Markova, 2007). In Western Siberia, *Palaeoloxodon* coexisted with *Equus* cf. *steinheimensis*, a small-sized cave bear *Ursus savini rossicus* (= *Ursus spelaeus rossicus*), and the giant deer *Megaloceros giganteus* during the contemporaneous Tobol interstadial (Vangengeim, 1977; Figure 7). From this time onward (~400 ka), the giant deer became widespread in the temperate latitudes of Eurasia, together with modern deer species (red deer *Cervus elaphus* and the reindeer *Rangifer tarandus*).

The Khasarian faunal assemblage (Dnepr glaciation) with its type locality at Cherny Yar (Volga River Basin) contained *Mammuthus trogontherii chosaricus* (ancestor of the woolly mammoth *Mammuthus primigenius*), *Panthera spelaea*, *Elasmotherium sibiricum*, the woolly rhinoceros *Coelodonta antiquitatis*, *Equus chosaricus*, *Camelus knoblochi*, the long-horned bison *Bison priscus longicornis*, *Megaloceros giganteus*, and modern genera and species, including *Cervus elaphus*, *Rangifer tarandus*, the musk-oxen *Ovibos*, the saiga antelope *Saiga*, etc., the water vole *Arvicola chosaricus*, and a diverse assemblage of steppe-adapted small mammals (Alexandrova, 1976; Dubrovo, 1985; Gromov, 1948). The fauna existed in the colder and drier



Figure 7 *Megaloceros giganteus*, C. Flerov's painting exhibited in the Paleontological Museum of the Paleontological Institute, Russian Academy of Sciences (PIN), Moscow, with PIN permission.

climate of the periglacial forest-steppe along the southern margin of the continental Dnieper glaciation (Riss, RI) (Vangengeim, 1977). However, localities rich in large mammals are rather rare. In the south of the Russian Plain, typical subarctic and steppe small mammals occurred (*Dicrostonyx simplicior*, *Lemmus sibiricus*, *Lagurus* ex gr. *lagurus*, *Microtus gregalis*). In the Caucasus region, the tar pit Binagady fauna near Baky belongs to the same stage. Under modern conditions, the range of some animals became completely separated and confined to different biomes.

Conclusions

The data on the early and middle Pleistocene mammals from Northern Eurasia offer important insights into our knowledge of the history of mammals. These assemblages differ considerably in their composition as a result of evolutionary and migration processes. In the Pleistocene, the general trend toward global temperature decrease, changes in landscape and climatic conditions, and differentiated environmental patterns caused important changes in mammal communities within the territory at the suprageneric, generic, and species levels. The faunas demonstrate adaptation from the temperate climatic conditions of the pre-Pleistocene to the cooler and more diverse conditions at the end of the Pleistocene. A gradual disappearance is therefore noted of the Miocene–Pliocene and thermophilous forms (mastodons, giraffids, hippopotamids, hyenids) with their extinction or displacement to southern Asia or Africa, paralleled by the development of Pleistocene and modern genera and species, and the gradual emergence of associations typical for modern biomes (tundra, steppes, forest, etc.).

The main periods of turnover in mammal communities of Northern Eurasia coincided approximately with the maximum drops in global temperature, which were accompanied by large-scale transformations of the environment. The most significant changes occurred at the Pliocene–Pleistocene (2.6 Ma) and early–middle Pleistocene (0.8 Ma) boundaries. Changes in faunal composition were also reported at 1.8, 1.2, and 0.4 Ma. Beginning in the middle Pleistocene, the turnover affected mammal diversity at the generic and species levels. Immigration to and from adjacent territories played an important role in these faunal transformations.

The roots of some modern genera are found in the late Miocene (*Canis*, *Axis*, etc.), the Pliocene (*Nyctereutes*, *Ursus*, *Lynx*, *Equus*), the Gelasian (*Gulo*, *Cervus*, *Alces*, *Capreolus*, *Bison*), and the middle Pleistocene (*Camelus*, *Rangifer*, *Ovibos*, *Saiga*). The Pliocene–Pleistocene boundary is marked by the disappearance of a number of thermophilous forms. Evolutionary trends can be traced equally in the large herbivorous animals (elephants, rhinoceroses, horses, artiodactyls), various carnivores (canids, ursids, hyenids, felids), and small mammals. A clear boreal group, which comprised the ancestral forms of modern Northern Eurasian species, started to form actively at the beginning of the late early Pleistocene, around 1.8 Ma. The increasing amplitude of global cooling was then accompanied by the appearance and widespread distributions of first boreal and then arctic forms in northern Siberia and the mountainous regions of Central Asia. Boreal communities

became more widespread in the early middle Pleistocene (true wolves, tiger *Panthera tigris*) and at the end of late middle Pleistocene (Khasarian), with the first appearance of modern species (*Cervus elaphus*, *Rangifer tarandus*, *Ovibos*, *Saiga*, etc.). Zoogeographic provinciality therefore increased from the Early toward the late middle Pleistocene.

See also: [Vertebrate Records: Early Pleistocene; Late Pleistocene Megafaunal Extinctions; Late Pleistocene Mummified Mammals; Late Pleistocene of Africa; Late Pleistocene of North America; Late Pleistocene of South America; Late Pleistocene of Southeast Asia; Mid-Pleistocene of Africa; Mid-Pleistocene of Europe; Mid-Pleistocene of Southern Asia.](#)

References

- Agadjanian AK (2009) *Small Mammals of the Pliocene–Pleistocene of the Russian Plain*. Moscow: Nauka (in Russian).
- Alexandrova LP (1976) *Rodents of the Anthropogene of the European Part of the USSR*. Moscow: Nauka (in Russian).
- Alexeeva LI (1977) *Early Anthropogene Theriofauna of East Europe*. Moscow: Nauka (in Russian).
- Alexeeva NV (2005) *Environmental Evolution of Late Cenozoic of West Transbaikalia (Based on Small Mammal Fauna)*. Moscow: GEOS (in Russian).
- Alexeeva NV and Erbaeva MA (2008) Diversity of Late Neogene–Pleistocene small mammals of the Baikalian region and implications for paleoenvironment and biostratigraphy: An overview. *Quaternary International* 179: 190–195.
- Alexeeva NV, Karasev VV, and Erbaeva MA (2007) Pleistocene biostratigraphy of the Transbaikalian area (South East Russia). *Courier Forschungsinstitut Senckenberg* 11: 19–26.
- Bajgusheva VS (1971) Fossil theriofauna of Liventzovka sand pit (northeastern of the Sea of Azov). In: Bykhovskii BE (ed.) *Materials on Faunas of the Anthropogene of the USSR*, pp. 5–28. Leningrad: Nauka (in Russian).
- Bajgusheva VS, Titov VV, and Tesakov AS (2001) The sequence of Plio–Pleistocene mammal faunas from the south Russian Plain (the Azov Region). *Bollettino della Società Paleontologica Italiana* 40(2): 133–138.
- Cande SC and Kent DV (1995) Revised calibration of the geomagnetic polarity timescale for the Late Cretaceous and Cenozoic. *Journal of Geophysical Research, Series B* 100(4): 6093–6095.
- Derevyanko AP and Shunkov MV (2005) Early Paleolithic site Karama in Altai Region: First results. *Archaeology, Ethnography and Anthropology of Eurasia* 3(23): 52–69 (in Russian).
- Dmitrieva EL (1977) *Neogene Antelopes of Mongolia and Adjacent Territories*. Moscow: Nauka (in Russian).
- Dmitrieva EL and Nesmeyanov SA (1982) *Mammals and Stratigraphy of Continental Tertiary Deposits of Southeastern Middle Asia*. Moscow: Nauka (in Russian).
- Dodonov AE (2002) *Quaternary of Middle Asia: Stratigraphy, Correlation, Paleogeography*. Moscow: GEOS (in Russian).
- Dubrovo IA (1990) The Pleistocene elephants of Siberia. In: Agenbroad LD, Mead JL, and Nelson LW (eds.) *Megafauna and Man*, vol. 1, pp. 1–8. South Dakota: Hot Springs.
- Erbaeva MA and Alexeeva NV (2000) Pliocene and Pleistocene biostratigraphic succession of Transbaikalia with emphasis on small mammals. *Quaternary International* 68–71: 67–75.
- Foronova IV (2001) *Quaternary Mammals of the South-east of Western Siberia (Kuznetsk Basin): Phylogeny, Biostratigraphy, and Paleoeology*. Novosibirsk: SB RAS GEO (in Russian).
- Gabunia L and Vekua A (1995) A Plio–Pleistocene hominid from Dmanisi, East Georgia, Caucasus. *Nature* 373: 509–512.
- Garutt VE and Bajgusheva VS (1981) *Archidiskodon gromovi* Garutt et Alexeeva – der älteste Elefant der Mammutlinie in Eurasien. *Quartärpaläontologie* 4: 7–18.
- Giozzi E, Abbazzi L, Argenti P, et al. (1997) Biochronology of selected mammals, mollusks and ostracods from the Middle Pliocene to the Late Pleistocene in Italy. The state of the art. *Rivista Italiana di Paleontologia e Stratigrafia* 103(3): 369–388.

- Gromov VI (1948) *Palaeontological and Archaeological Foundation of the Stratigraphy of the Quaternary Continental Deposits in the Territory of the USSR*. Moscow: The USSR Academy of Sciences (in Russian).
- Hemmer H, Kahlke R-D, and Vekua AK (2010) *Panthera onca georgica* ssp. nov. from the Early Pleistocene of Darnisi (Republic of Georgia) and the phylogeography of jaguars (Mammalia, Carnivora, Felidae). *Neues Jahrbuch für Geologie und Paläontologie Abhandlungen* 257(1): 115–127.
- Kahlke R-D, García N, Kostopoulos DS, et al. (2011) Western Palaearctic palaeoenvironmental conditions during the Early and early Middle Pleistocene inferred from large mammal communities, and implications for hominin dispersal in Europe. *Quaternary Science Reviews* 30(11–12): 1368–1395.
- Lister A and Sher A (2001) The origin and evolution of the woolly mammoth. *Science* 294: 1094–1097.
- Markova AK (1982) *Pleistocene Rodents of the Russian Plain*. Moscow: Nauka (in Russian).
- Markova AK (2006) Likhvin interglacial small mammal faunas of Eastern Europe. *Quaternary International* 149(1): 67–79.
- Markova A (2007) Pleistocene mammal faunas of Eastern Europe. *Quaternary International* 160(1): 100–111.
- Nikiforova KV (ed.) (1971) *Pleistocene of Tiraspol*. Kishinev: Shtiintza.
- Palombo MR, Valli AMF, Kostopoulos DS, Alberti MT, Spassov N, and Vislobokova I (2006) Similarity relationships between the Pliocene to Middle Pleistocene large mammal faunas of Southern Europe from Spain to the Balkans and the North Pontic Region. *Courier Forschungsinsitut Senckenberg* 256: 329–347.
- Qiu Zh-X (2006) Quaternary environmental changes and evolution of large mammals in Northern China. *Vertebrata Palasiatica* 44(2): 110–132.
- Ranov VA, Carbonell E, and Rodríguez XP (1995) Kuldara, earliest human occupation in Central Asia in its Afro-Asian context. *Current Anthropology* 36: 337–346.
- Rekovets LI and Nadachowski A (1995) Pleistocene voles (Arvicolidae) of the Ukraine. *Paleontologia i Evolucio* 28–29: 145–245.
- Shackleton NJ (1995) New data on the evolution of Pliocene climatic variability. In: Vrba ES, Denton GH, Partridge TC, and Burke LH (eds.) *Paleoclimate and Evolution with Emphasis on Human Origin*, pp. 242–248. New Haven: Yale University Press.
- Shchelinsky VE, Dodonov AE, Baigusheva VS, et al. (2010) Early Palaeolithic sites on the Taman Peninsula (Southern Azov Sea region, Russia): Bogatyri/Sinyaya Balka and Rodniki. *Quaternary International* 223–224: 28–35.
- Sher AV (1971) *Mammals and Stratigraphy of the Pleistocene of Northeast USSR and North America*. Moscow: Nauka (in Russian). Sher AV (1974) Pleistocene Mammals and Stratigraphy of the Pleistocene of the Far Northeast USSR and North America. *International Geological Review* 16(7–10): 1–284 (in English).
- Sher AV (1987) Olyorian land mammal age of northeastern Siberia. *Palaeontographia Italica* 74: 97–112.
- Sher AV, Weinstock J, Baryshnikov GF, et al. (2011) The first record of 'spelaeoid' bear in Arctic Siberia. *Quaternary Science Reviews* 30(17–18): 2238–2249.
- Sotnikova MV (1989) *Late Pliocene–Early Pleistocene Carnivora: Stratigraphic Significance*. Moscow: Nauka (in Russian).
- Sotnikova MV, Baigusheva VS, and Titov VV (2002) Carnivores of the Khapry faunal assemblage and their stratigraphic implication. *Stratigraphic and Geological Correlation* 10(4): 375–390.
- Sotnikova MV, Dodonov AE, and Pen'kov AV (1997) Upper Cenozoic bio-magnetic stratigraphy of Central Asian mammalian localities. *Palaeogeography, Palaeoclimatology, Palaeoecology* 133: 243–258.
- Sotnikova M and Rook L (2010) Dispersal of the Canini (Mammalia, Canidae: caninae) across Eurasia during the Late Miocene to Early Pleistocene. *Quaternary International* 212: 86–97.
- Sotnikova M and Titov V (2009) Carnivora of the Tamanian faunal unit (the Azov Sea area). *Quaternary International* 201: 43–52.
- Sotnikova MV and Vislobokova IA (1990) Pleistocene mammals from Lakhiti, Southern Tajikistan, USSR. *Quartärpaläontologie* 8: 237–244.
- Tesakov AS (2004) *Biostratigraphy of Middle Pliocene–Eopleistocene of Eastern Europe (Based on Small Mammals)*. Moscow: Nauka (in Russian).
- Tesakov AS, Dodonov AE, Titov VV, and Trubikhin VM (2007) Plio–Pleistocene geological record and small mammal faunas, eastern shore of the Azov Sea, Southern European Russia. *Quaternary International* 160(1): 57–69.
- Tesakov AS and van Kolfschoten T (2011) The Early Pleistocene *Mimomys hordijki* (Arvicolinae, Rodentia) from Europe and the origin of modern nearctic sagebrush voles (*Lemmyscus*). *Palaeontologia Electronica* 14(3): 1–11.
- Titov VV (2008) *Late Pliocene Large Mammal from Northeastern Sea of Azov Region*. Rostov-on-Don: SSC RAS Publishing (in Russian).
- Tjutkova LA and Kaipova GO (1996) Late Pliocene and Eopleistocene micromammal faunas of southeastern Kazakhstan. *Acta Zoologica Cracoviensia* 39: 549–557.
- Topachevsky VA, Scorik AF, and Rekovets LI (1987) *Rodents of the Upper Neogene and Early Anthropogene Deposits of the Khadjibei Lagoon*. Kiev: Naukova Dumka (in Russian).
- Vangengeim EA (1977) *Paleontological Foundation of the Anthropogene Stratigraphy of Northern Asia*. Moscow: Nauka (in Russian).
- Vangengeim EA, Erbaeva MA, and Sotnikova MV (1990) Pleistocene mammals from Zasuino, Western Transbaikalia. *Quartärpaläontologie* 8: 257–264.
- Vangengeim EA and Pevzner MA (1991) The Villafranchian of the USSR: Bio- and magnetostratigraphy. In: Vangengeim EA (ed.) *Pliocene and Anthropogene Palaeogeography and Biostratigraphy*, pp. 124–145. Moscow: Geological Institute (in Russian).
- Vangengeim EA, Sotnikova MV, Alekseeva LI, et al. (1988) *Biostratigraphy of Late Pliocene–Early Pleistocene of Tadzhikistan*. Moscow: Nauka (in Russian).
- Vekua AK (1962) *Akhalkalaki Lower Pleistocene Mammal Fauna*. Tbilisi: The Georgian National Academy of Sciences (in Russian).
- Vekua AK (1986) The Lower Pleistocene mammalian fauna of Akhalkalaki (Southern Georgia, USSR). *Paleontographia Italica* 74: 63–96.
- Vekua AK (1995) Die Wirbeltierfauna des Villafranchian von Dmanisi und ihre biostratigraphische Bedeutung. *Jahrbuch des Romisch-Germanischen Zentralmuseums Mainz* 42: 77–180.
- Vekua A and Lordkipanidze D (2008) The history of vertebrate fauna in Eastern Georgia. *Bulletin of the Georgian National Academy of Sciences* 2(3): 149–154.
- Vereshchagin NK (1957) *Remains of Mammals from Lower Quaternary Deposits of the Taman*. Leningrad: Nauka (in Russian).
- Vereshchagin NK (1959) *Mammals of the Caucasus*. Moscow: The USSR Academy of Sciences (in Russian).
- Vislobokova I (1990) *Fossil Deer of Eurasia*. Moscow: Nauka (in Russian).
- Vislobokova I (1996) The Pliocene Podpusk-Lebyazh'e mammalian faunas and assemblages, Western Siberia. *Palaeontographia Italica* 83: 1–23.
- Vislobokova IA (2005) On Pliocene faunas with Proboscideans in the territory of the former Soviet Union. *Quaternary International* 126–128: 93–105.
- Vislobokova IA, Sotnikova MV, and Dodonov AE (2003) Bio-events and diversity of the Late Miocene–Pliocene mammal faunas of Russia and adjacent areas. *Deinsea* 10: 563–574.
- Vislobokova IA, Sotnikova MV, and Erbaeva MF (1995) The Villafranchian mammalian faunas of the Asiatic part of former USSR. *Il Quaternario* 8(2): 367–376.
- Zazhigin VS (1980) *Late Pliocene and Anthropogene Rodents of the South of Western Siberia*. Moscow: Nauka (in Russian).

Mid-Pleistocene of Africa

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The mammalian fauna of the African continent has undergone many changes during the Neogene. Most of these occurred by the end of the Pliocene and the beginning of the Pleistocene, when the African predator guild attained its modern composition and structure, driven principally by the extinction of saber-toothed cats on the continent and the disappearance of the more archaic hyenas (Turner, 1990, 1995a; Turner and Antón, 1999). Other archaic elements of the African megafauna, such as the deinotheres, gomphotheres, and chalicotheres, also disappeared at approximately the same time. While the hominin genus *Paranthropus* went extinct, ungulates such as pigs and giraffes became less diverse and the three-toed hipparion horses became rare (Turner and Wood, 1993; Turner and Antón, 2004). In contrast, Migrants to Africa in the Pliocene and more recently thrived, adding more variety to the indigenous fauna. African faunas thus appear more modern at a much earlier date, in contrast to the situation in Eurasia and the New World, where archaic forms survived until the Holocene and the appearance of modern faunas is a very recent phenomenon. The living African mammal fauna can thus be considered largely a Pliocene relict.

Nevertheless, at the beginning of the Pleistocene the structure and composition of faunas still differed in several important ways from the pan-African fauna of today (Turner and Antón, 2004). Werdelin and Lewis (2005) suggested that this difference may have been even more marked than previously thought, at least with regard to the Carnivora. They argued that extant taxa such as the lion (*Panthera leo*), leopard (*Panthera pardus*), cheetah (*Acinonyx jubatus*), and spotted hyena (*Crocuta crocuta*) can only be traced back to the Pliocene at the genus level at best, at least in eastern Africa, instead of at the species level as previously argued by Turner (1990). They also cast doubt on previous arguments for the extinction of the machairodont cats at approximately 1.5 Ma as proposed by Turner and Antón (1999, 2004) and for the interpretations of turnover in the guild of larger Carnivora also proposed by these authors. These differences in viewpoint stem in part from differences in the tendency to split rather than lump taxa, and they will always be a part of paleontological debate. However, arguments for the splitting of the genus *Crocuta* in particular are founded in large part on unpublished data (Werdelin and Lewis, 2005, p. 128) and are therefore difficult to address. They also do not take into account the South African record.

However this difference in viewpoints is resolved, it remains true that the nature and pattern of changes within the African Pleistocene have been poorly understood, largely because well-dated deposits such as those of eastern Africa that provide our picture of Pliocene events are relatively rare or information about them is unpublished. For these reasons, relatively little synthesis of Pleistocene faunal evolution has been attempted. Fortunately, this situation is now changing, and in recent years details of several sites have appeared or the

material has been reanalyzed and deposits have been dated. As a result, overviews such as that of O'Regan et al. (2005) show that the early to middle Pleistocene faunas of Africa do not show the major changes across the boundary seen in Eurasia and the Americas at that time. Instead, they concluded that the middle Pleistocene faunas of Africa have been heavily influenced by geographic locality, such that although a small amount of turnover may have taken place, the modern zoogeographic communities in the Levantine and north African areas of the Palearctic and those of the sub-Saharan African region can be traced back as far as the early Pleistocene.

Our aim is to provide a summary of the faunas of the middle Pleistocene of Africa in order to extend previous treatments of the available data (O'Regan et al., 2005). Following a brief survey of the climatic background to the period, the discussion will be regional, focusing on northern Africa, eastern Africa, and southern Africa (Figure 1 and Table 1). West Africa has few, if any, middle Pleistocene sites; thus, it is omitted from this discussion. Many of the sites that we mention were collected from or excavated for the purposes of archaeological rather than paleontological study. Many preserve trace fossils of the genus *Homo* in the form of stone tools, which range in technology from the Acheulean to Middle Stone Age. Therefore, the presence of *Homo* is inferred from behavioral remnants as well as, in some cases, actual hominin fossils. There is thus a potential for collection bias in that the presence of archaeology is the main impetus for the continued exploration of these sites. However, it may be that, by the middle Pleistocene, hominins were such a well-established and pan-African taxon that their actual or inferred presence in these assemblages represents a real feature of their distribution in the past landscape.

The Middle Pleistocene

Climatic Background

We take the early-middle Pleistocene boundary to occur at 0.78 Ma (Berggren et al., 1995). Between 0.92 and 0.65 Ma, the climate of the Earth shifted from the smaller scale oscillations in the Pliocene and early Pleistocene to the major oscillations that characterize the middle and late Pleistocene. The change produced a large increase in global ice volume at the beginning of the interval, followed by a change in the frequency of the Quaternary Ice Ages as the 41-ka glacial cyclicity was replaced by the 100-ka glacial cyclicity by 0.65 Ma (Shackleton and Opdyke, 1976). Between 0.9 and 0.6 Ma, the interglacials were warm and arid and the glacials cool and humid in Africa (Dupont et al., 2001), but the transition to the 100 ka cyclicity brought warm and more humid interglacials and colder and drier glacials. Evidence from deep-sea cores and isolated terrestrial records (Schefuss et al., 2003) suggest that the African vegetation responded to changes in aridity, temperature, and CO₂ concentrations provoked by



Figure 1 Map showing faunal occurrences.

these climatic shifts. Recent work suggests that during the middle Pleistocene the lakes of the East African Rift System, which were focal elements of many of the sedimentary basins from which we know faunal assemblages, were at times very deep; this may correspond to a variable climate following the mid-Pleistocene revolution (Trauth et al., 2005).

North Africa

The faunal record in northern Africa is patchy, but several sites offer a chance to compare this area with other areas of Africa. In Morocco, localities near Casablanca provide the possibility to examine faunal assemblages of the region. The lower levels (L) of the Thomas Quarry sites also may date from the lower middle Pleistocene and preserve human fossils, stone tools, and some poorly preserved animal remains, including *Hippopotamus*, *Equus* (zebra) and gazelle at the lowest levels. All of the potentially middle Pleistocene sites—that is, those that postdate Ternefine (= Tighenif, thought to be early Pleistocene) have yielded remains of fossil bears, genus *Ursus*, which are completely absent in African deposits south of the Sahara. The presence of the suid *Kolpochoerus* suggested to Raynal et al. (2001) that these deposits may predate the middle Pleistocene, but without precise biostratigraphic analysis of the material it is impossible to determine whether it is a late example of the taxon, which survives until the middle or even perhaps the late Pleistocene. The absence of extinct forms from the extensive carnivore fauna contrasts with the micromammal remains, which may suggest that the site is early Pleistocene in age, since specific anatomical details of some rodents appear rather

archaic (Raynal et al., 2001). Thomas Quarry 1, which may have been redeposited secondarily, presents a carnivore-dominated assemblage with an estimated date earlier than 400 Ka.

Oulad Hamida 1 Quarry has several sites, one of which, Rhino Cave, has an abundance of the eponymous white rhino *Ceratotherium* in association with numerous Acheulean artifacts. *Camelus* is unknown in Pleistocene deposits south of the Sahara, so the Rhino Cave example may be the earliest on the continent. Numerous carnivore remains have also been recovered, although all are from extant species (Raynal et al., 2001).

The Moroccan middle Pleistocene sites yield the giant baboon *Theropithecus oswaldi* and *Equus mauritanicus*, but not *Metridiochoerus*, the antelope *Hippotragus gigas*, or the scimitar cat *Homotherium*, which Raynal et al. (2001) used to suggest that the sites postdate the lower Pleistocene as represented at Tighenif. However, it is just as likely that at least some of these apparent absences are due to environmental factors since *Metridiochoerus* is known to survive well into the middle Pleistocene. Differences in the rodent, particularly gerbil, fauna from the Rhino Cave assemblage compared with the lower levels of the Thomas Quarry site may also suggest environmental rather than temporal contrasts. For present purposes, these sites are considered as middle Pleistocene, although the potential conflation of temporal and environmental factors in identifying biostratigraphic markers is noted as problematic.

Eastern Africa

Eastern African sites are, to a certain degree, the most useful when trying to assess the nature and scope of faunal change

Table 1 Faunal occurrences

<i>Locality</i>	<i>Location</i>	<i>Age (Ma)</i>	<i>Dating method</i>	<i>Paleoenvironment</i>	<i>References</i>
Oulad Hamida 1	Morocco	~0.7–0.13	Biostratigraphy	Shoreline	Raynal et al. (2001)
Thomas Quarry 1	Morocco	0.7–0.6	Biostratigraphy	Marine cave	Geraads (2002), Raynal et al. (2001)
Asbole	Lower Awash Valley, Ethiopia	0.8–0.6	Biostratigraphy	Wooded; riverine gallery forest	Geraads et al. (2004)
Wehailu Formation	Middle Awash Valley, Ethiopia	~0.6	$^{40}\text{Ar}/^{39}\text{Ar}$, biostratigraphy	Seasonal river (Wadi)	Kalb et al. (1982a,b)
Bodo	Middle Awash Valley, Ethiopia	~0.6	$^{40}\text{Ar}/^{39}\text{Ar}$, biostratigraphy	Seasonal river (Wadi)	Kalb et al. (1982a,b), Clark et al. (1994)
K3(ii), Kapthurin Formation	Baringo, Kenya	0.542 ± 0.004 to 0.509 ± 0.009	$^{40}\text{Ar}/^{39}\text{Ar}$	Mixed, open and closed environment faunas	McBrearty et al. (1996), Deino and McBrearty (2002), McBrearty and Jablonski (2005)
KN-5, Kanjera North	Homa Peninsula, Kenya	>0.78	Paleomagnetism	Swamp, returning to mudflats	Plummer (1991, personal communication), Ditchfield et al. (1999)
Members 10, 11 Olorgesailie	Kenya	0.662 ± 0.004 ; 0.601 ± 0.003	$^{40}\text{Ar}/^{39}\text{Ar}$	floodplain	Potts et al. (1999), Koch (1986)
Lainyamok	Kenya	0.392 ± 0.004 to 0.33 ± 0.006	$^{40}\text{Ar}/^{39}\text{Ar}$	Lacustrine/riverine	Potts and Deino (1995)
Masek Beds, Olduvai Gorge	Tanzania	0.62–0.40	Paleomagnetism, sedimentation rate	Open, dry grassland, streams	Hay (1994)
Ndutu Beds, Olduvai Gorge	Tanzania	>0.4	Paleomagnetism, sedimentation rate	Open, dry grassland	Hay (1994)
Bed IV, Olduvai	Tanzania	0.8–0.6	Paleomagnetism, sedimentation rate	Open, dry grassland, by a river channel	Hay (1994)
Eyasi Beds	Tanzania	~0.7–0.13	Sedimentation rate	Near lake	Mehlman (1987)
Isimila	Tanzania	~0.26	Uranium series	Shrubby, near stream channels	Cole and Kleindeinst (1974), Coryndon et al. (1972)
Twin Rivers	Zambia	0.27–0.17	Uranium–thorium (speleothem)	Bushland near standing water	Barham (2000), Bishop and Reynolds (2000)
Mumbwa Caves	Zambia	$\sim 0.172 \pm 0.022$ to 0.13 ± 0.06	OSL; TL	Wooded grassland	Klein and Cruz-Uribe (2000), Barham (2000)
Gladysvale	Gauteng, South Africa	0.78–0.578	ESR, paleomagnetism	Open grassland	Lacruz et al. (2003)
Cornelia	Free State, South Africa	~0.7	Biostratigraphy		Butzer et al. (1974), McKee et al. (1995)
Florisbad	Free State, South Africa	~0.7–0.13	Biostratigraphy		McKee et al. (1995), Kuman and Clarke (1986)
Cave of Hearths	Limpopo Province, South Africa	~0.7–0.13	Biostratigraphy		McKee et al. (1995)
Duinefontein 2	Western Cape Province, South Africa	~0.4 to ~0.2	Biostratigraphy; Uranium series	Grass-and-bush mosaic	Klein et al. (1999)
Elandsfontein	Western Cape Province, South Africa	~0.7–0.4	Fauna, archaeology	Moist, with grassland and bush	Klein and Cruz-Uribe (1991)
Namib IV	Namibia	~0.7–0.4	Biostratigraphy		Shackley (1980)

through time. Due to a continuous history of faulting activity on the East African Rift System (EARS) and concomitant volcanic activity, the sites that preserve products of volcanism can be dated using absolute methods, such as potassium–argon or single crystal laser fusion argon–argon dating. Even when such methods are not applicable, biostratigraphic comparisons have a greater prospect of validity since they are made over smaller geographic distances.

Fossils from the Asbole area in the lower Awash Valley of Ethiopia have been dated at 0.80–0.60 Ma using biostratigraphy of the diverse mammalian fauna (Geraads et al., 2004). Extinct forms, such as the pigs *Kolpochoerus majus* and *Metridiochoerus*, are found with an early occurrence of the genus *Bos*, a new species of herpestid (mongoose), and Acheulean artifacts. In the Middle Awash Valley of Ethiopia, the Wehailu Formation has yielded a fauna assigned to the middle Pleistocene (Kalb et al., 1982a,b), whereas the Bodo Member, which has been correlated on the basis of its fauna to Olduvai Bed IV, has several extinct taxa, including the pigs *Kolpochoerus limnetes*, *K. majus*, and *Metridiochoerus*. The overlying Andalee Member may traverse the middle/late Pleistocene boundary but contains the extinct *Giraffa* cf. *G. jumae* and *Kolpochoerus majus*.

Kanjera, in Western Kenya, preserves an extensive middle Pleistocene fauna in the northern exposure members of the Kanjera Formation, which span the Brunhes–Matuyama boundary. There, *Elephas recki* is found in KN-5 in sediments deposited during the Brunhes chron, whereas other extinct taxa, such as *Metridiochoerus compactus* and *M. hopwoodi*, derive from below this, in magnetically normal sediments. The most extensive collection of fauna from the middle Pleistocene beds at Kanjera comes from the GP (Giraffe Phalanx) site, where modern taxa such as the suid genus *Phacochoerus* are found in association with species that subsequently became extinct, such as *Metridiochoerus*, *Syncerus acoelotus* (African buffalo), *Hippopotamus* cf. *H. gorgops*, and *Kolpochoerus* (Plummer, 1991; Ditchfield et al., 1999).

Lainyamok, a site west of Lake Magadi in Kenya, preserves a well-dated fauna that has some interesting characteristics (Shipman et al., 1983; Potts and Deino, 1995). This 392–330 Ka site has a fauna that is notable both for its lack of extinct forms and for the presence of modern forms occurring beyond their modern range, such as the blesbok *Damaliscus dorcus*, currently restricted to southern Africa. In many ways, Lainyamok can be considered one of the earliest well-dated modern faunas in Africa and demonstrates that, at least in some regions, an essentially modern fauna was in place here before the end of the middle Pleistocene. The site further provides a date for the disappearance of several extinct taxa that were common in older sites.

Ologesailie, in Kenya, spans the interval from the latest part of the early Pleistocene through the middle Pleistocene (Deino and Potts, 1990). The Member 10 (0.662 ± 0.004 Ma) and 11 (0.601 ± 0.003 Ma) faunas are the best described and contain a range of modern antelope genera as well as the extinct pigs *K. majus* and *Metridiochoerus* and the extinct *Hippopotamus gorgops* (Koch, 1986). The Kaphurur Formation in the Tugen Hills near Lake Baringo, Kenya, provides an extensive fauna in sediments dated from 0.75–0.23 Ma (Deino and McBrearty, 2002; McBrearty et al., 1996). Although most of the taxa recovered are extant, there are a few exceptions, such as *K. majus*. *Hippopotamus gorgops* may be represented in some of the larger hippo remains,

but *Syncerus acoelotus* is not documented. The modern chimpanzee, *Pan*, has also been reported from the Kaphurur Formation, which falls well outside of its current known geographical range (McBrearty and Jablonski, 2005).

Olduvai Gorge, Tanzania, preserves a middle Pleistocene fauna in three geological units: Olduvai Bed IV (0.83–0.62 Ma), the Masek Beds (0.62–0.40 Ma), and the lower portion of the Ndutu Beds, the oldest date for which is 0.40 Ma (Hay, 1994). Bed IV preserves *H. gorgops*, *S. acoelotus*, *M. compactus*, and *Phacochoerus antiquus*, a primitive relative of the modern warthog (Harris and White, 1979). The Masek Beds have a less diverse fauna than Bed IV and preserve all of these except *S. acoelotus*, due either to its local extinction or to sampling error. Near Lake Eyasi, Tanzania, extinct species have been recovered from the Eyasi Beds, including *Hippopotamus* aff. *H. gorgops* and large giraffids that may be *G. jumae* (Mehlman, 1987). These remains are assigned a tentative middle Pleistocene age based on calculations of sedimentation rates. Isimila, on the Iringa Plateau, may be as old as 0.26 Ma based on an early use of uranium series dating (Cole and Kleindienst, 1974). Several extinct taxa have been recovered from the site, including *M. compactus* and *K. limnetes* (Coryndon et al., 1972).

Southern Africa

The history of faunal change and movement in the southern part of Africa differs from that in the east. Although *K. limnetes*, *M. hopwoodi*, and *H. gorgops* are present in the middle Pleistocene, they all disappear by its end. *Metridiochoerus compactus*, which is locally extinct by the latest part of the middle Pleistocene in eastern Africa, is present until very late in southern Africa, with the youngest dated occurrence of 30–20 Ka from Redcliffe Cave, Zimbabwe (McKee et al., 1995). The extant *Syncerus caffer* is known throughout the middle Pleistocene, but *S. acoelotus*, which is more commonly found in East Africa, is not documented from southern sites on the continent.

Two localities in Zambia, Mumba Falls and Twin Rivers, preserve fossils that date to the later part of the middle Pleistocene (Barham, 2000). Fauna from Twin Rivers is scanty and fragmentary but is associated with Lupemban stone tools. There is no evidence that any extinct species were present (Bishop and Reynolds, 2000). The middle Pleistocene assemblage from Mumbwa Cave is also very small; however, the presence of an antilopine bovid is unusual. Gazelles and springboks have not been known from the Zambezian Zone during historic times (Klein and Cruz-Urbe, 2000).

The majority of South African sites are caves and thus have different geological derivation and depositional history from those in eastern Africa, which are almost exclusively open-air sites. They may be sampling different local environments, which may or may not be representative of wider regions. When comparing sites from more than one region, it is important to note differences that are ecological rather than temporal in nature, although distinguishing between the two can be difficult. McKee et al. (1995) observed that species such as *Hippopotamus amphibius*, which has a long temporal range but a very narrow ecological one, has a very patchy distribution in the cave sites of southern Africa and is not useful for dating sites.

McKee et al. (1995) consider Cornelia to be a middle Pleistocene site notable for the absence of archaic *Homo sapiens*,

which is otherwise present at most of the southern African localities of this age. Elsewhere in southern Africa, archaic *H. sapiens* is usually found with a higher proportion of extant species than have been recovered at Cornelia (McKee et al., 1995). Indeed, more than 60% of the taxa recovered from the site are extinct, which presents a stark contrast to the situation in eastern Africa toward the end of the middle Pleistocene. *Hipparion libycum*, an archaic three-toed horse, has its last appearance in the southern African faunal record there, suggesting that it may be one of the oldest middle Pleistocene sites in the region.

Elandsfontein (= Hopefield = Saldhana) in the Western Cape Province preserves an extensive fauna and has been suggested to date from the first half of the middle Pleistocene (0.7–0.4 Ma) (Klein and Cruz-Urbe, 1991). The majority of the remains recovered are from extant taxa, but a few archaic forms, such as the antelope *Antidorcas recki*, *Metridiochoerus andrewsi*, *Theropithecus oswaldi*, and *Megantereon*, have what may be their last appearances. However, it has long been recognized that the deposits at the site have no clear date; the assemblage probably represents a considerable palimpsest derived from hyena dens and almost certainly includes earliest Pleistocene taxa (Hendey, 1974; Turner, 1990; Turner and Antón, 1999). It is notable that the archaic taxa span a number of biological families and ecological preferences, suggesting that these disappearances are temporal rather than ecological phenomena. The fact that so little of the assemblage was influenced by human agents reinforces this conclusion, allowing hominin activity to be ruled out as a potential taxonomic bias in the composition of the faunal assemblage.

Duinefontein 2, Western Cape Province, is an excavated site that preserves an extensive fauna in association with Acheulean artifacts (Klein et al., 1999). The fossil bone is extensively damaged by carnivores, with little evidence of hominin processing of bone. The ecological preferences of the fauna suggest that the local environment was grass-and-bush mosaic with standing water nearby. Klein et al. (1999) suggested that the fauna is similar in morphology and/or size to examples from nearby Elandsfontein Main. With the exception of *Equus capensis*, there are no definite examples of extinct species, leading to a biostratigraphic estimate of ~0.4 to ~0.2 Ma.

Gladysvale, a cave in Gauteng Province, has yielded assemblages that date from the Pliocene to the middle Pleistocene. The middle Pleistocene external deposits have produced a carnivore fauna that, although scanty, is modern. A few archaic ungulates, including a late occurrence of the South African ovibovine *Makapania* as well as *Pelorovis* and *Equus capensis*, have been recovered, but the remaining ungulates are extant forms.

The fauna from the Cave of Hearths in the Makapansgat Valley, Limpopo Province, suggests that the site was formed later than Elandsfontein and Cornelia. The presence of Acheulean artifacts, which are relatively developed from a technological standpoint, along with remains of *H. sapiens rhodesiensis*, reinforces this conclusion (McKee et al., 1995.) Some extinct forms of antelope, such as *Megalotragus* and *Pelorovis*, were recovered there, but the pig *Phacochoerus* is modern.

Florisbad, in the Free State of South Africa, preserves several extinct taxa in combination with Middle Stone Age artifacts and remains of archaic *H. sapiens*. It may be the most recent site

in South Africa that can be characterized as middle Pleistocene in age. As is the case at the Cave of the Hearths, the extinct large antelopes *Megalotragus* and *Pelorovis* were present at Florisbad, so the relative dating is based as much on the technology and hominin biostratigraphy as on other factors (Kuman and Clarke, 1986).

In southwestern Africa, the site of Namib IV within the central Namib Desert preserves *Elephas recki* along with other faunal and archaeological elements indicative of a middle Pleistocene age (ca. 700–400 Ka) (Shackley, 1980). Although this site adds little to our understanding of the composition of middle Pleistocene faunas, it extends considerably the geographic range over which we can document their presence.

Overview of the African Middle Pleistocene Mammal Fauna

We might expect, based on the evidence from other areas of the world (Azzaroli et al., 1988; Turner, 1995b), that the changes in climate summarized previously and any attendant vegetational responses would have impinged upon African mammals very directly. However, the change to the 100-ka cyclicity does not appear to have greatly affected the faunas because, although there are some changes, the turnover is largely attritional rather than a marked event (O'Regan et al., 2005). Approximately 0.70 Ma, several new species appeared, including the modern *H. amphibius*, the reedbuck, and the black wildebeest. Klein (1984) noted that at the other extreme of this time interval, several large mammalian species became extinct without issue, at or before the middle/late Pleistocene boundary. These include *Theropithecus oswaldi* and the elephantids *Loxodonta atlantica*, *E. recki*, and *E. iolensis*. Several suid species, including *Kolpochoerus limnetes* and *Metridiochoerus modestus*, also disappear by the end of the middle Pleistocene, as do the giraffids *Giraffa jumae*, *G. gracilis*, and *Sivatherium maurusium*. The majority of antelope extinctions occur somewhat earlier, but a species of *Tragelaphus*, along with *Thalcerocerus radiformis*, three species of the genus *Parmularius*, and the alcelaphine *Beatragus*, have their last occurrences in the middle Pleistocene, which also sees the end of the genus *Hipparion* (Klein, 1984). Although the precise timing of their disappearances is difficult to determine, due to both a paucity of samples and the problems of absolute dating during this time interval, there are no known occurrences at late Pleistocene sites.

Conclusions

With some exceptions, African mammalian faunas have remained quite stable throughout the Pleistocene. During the middle Pleistocene, there are a few extinctions, the exact timing of which is not precisely determinable, and a few new species also arise. Changes in the geographical ranges of some modern taxa can also be seen so that occurrences of the blesbok *Damaliscus dorcas* in Lainyamok and the chimpanzee *Pan* in the Kapthurin Formation of Kenya fall outside of their historical ranges. This review of fossil assemblages from the middle Pleistocene supports the suggestion that the development of the African large mammal fauna is characterized by fluctuations in species associations

and geographical ranges that may have been responses to short-term or rapid fluctuations in climate (Potts and Deino, 1995, p. 111).

See also: Vertebrate Overview. **Vertebrate Records:** Early Pleistocene; Late Pleistocene of Africa.

References

- Azzaroli A, De Guili C, Ficcarelli G, and Torre D (1988) Late Pliocene to early mid-Pleistocene mammals in Eurasia: Faunal succession and dispersion events. *Palaeogeography, Palaeoclimatology, Palaeoecology* 66: 77–100.
- Barham L (2000) The palaeobiogeography of central Africa. In: Barham L (ed.) *The Middle Stone Age of Zambia, South Central Africa*, pp. 223–236. Bristol, UK: Western Academic and Specialist Press.
- Berggren WA, Kent DV, Swisher CC III, and Aubry M-PA (1995) Revised Cenozoic geochronology and chronostratigraphy. In: Berggren WA, Kent DV, and Hardenbol J (eds.) *Geochronology, Time Scale and Global Stratigraphic Correlations: A Unified Temporal Framework for a Historical Geology*, pp. 129–212. Bloemfontein, South Africa: Society of Economic Paleontologists and Mineralogists. Special Publication No. 54.
- Bishop LC and Reynolds SC (2000) Fauna from Twin Rivers. In: Barham L (ed.) *The Middle Stone Age of Zambia, South Central Africa*, pp. 217–222. Bristol, UK: Western Academic and Specialist Press.
- Butzer KW, Clark JD, and Cooke HBS (1974) *The Geology, Archaeology and Fossil Mammals of the Cornelia Beds, O.F.S. Memoirs of the National Museum. Bloemfontein, Monograph 9*. Tulsa, Oklahoma, USA: SEPM.
- Clark JD, de Heinzelin J, Schick KD, et al. (1994) African *Homo erectus*: old radiometric ages and young Oldowan assemblages in the Middle Awash Valley, Ethiopia. *Science* 264: 1907–1910.
- Cole GH and Kleindienst M (1974) Further reflections on the Isimila Acheulian. *Quaternary Research* 4: 346–355.
- Cornelissen E, Boven A, Dabi A, et al. (1990) The Kapthurin Formation revisited. *African Archaeological Review* 8: 23–75.
- Coryndon SC, Gentry AW, Harris JM, et al. (1972) Mammalian remains from the Isimila Prehistoric Site, Tanzania. *Nature* 37: 292.
- Deino AL and McBrearty S (2002) $^{40}\text{Ar}/^{39}\text{Ar}$ dating of the Kapthurin Formation, Baringo, Kenya. *Journal of Human Evolution* 42: 185–210.
- Deino A and Potts R (1990) Single-crystal $^{40}\text{Ar}/^{39}\text{Ar}$ dating of the Olorgesailie Formation, Southern Kenya Rift. *Journal of Geophysical Research* 95: 8453–8470.
- Ditchfield P, Hicks J, Plummer T, Bishop LC, and Potts R (1999) Current research on the Late Pliocene and Pleistocene deposits north of Homa mountain, southwestern Kenya. *Journal of Human Evolution* 36: 123–150.
- Dupont LM, Bonner B, Schneider R, and Wefer G (2001) Mid-Pleistocene environmental change in tropical Africa began as early as 1.05Ma. *Geology* 29: 195–198.
- Geraads D (2002) Plio-Pleistocene mammalian biostratigraphy of Atlantic Morocco. *Quaternaire* 13: 43–53.
- Geraads D and Amani F (1997) La fauna du gisement à *Homo erectus* de l'Ain Maarouf, près de El Hajeb (Maroc). *L'Anthropologie* 101: 522–530.
- Geraads D, Alemseged Z, Reed D, Wynn J, and Roman DC (2004) The Pleistocene fauna (other than primates) from Asbole, lower Awash Valley, Ethiopia, and its environmental and biochronological implications. *Geobios* 37: 697–718.
- Harris JM and White TD (1979) *Evolution of the Plio-Pleistocene African Suidae*. Philadelphia: American Philosophical Society.
- Hay RL (1994) Geology and dating of Beds III, IV and the Masek Beds. In: Leakey MD and Roe DA (eds.) *Olduvai Gorge. Excavations in Beds III, IV and the Masek Beds, 1968–1971*, vol. 5, pp. 8–14. Cambridge, UK: Cambridge University Press.
- Hendey QB (1974) The Late Cenozoic Carnivora of the south-western Cape Province. *Annals of the South African Museum* 63: 1–369.
- Kalb JE, Jolly CJ, Tebedge S, et al. (1982a) Vertebrate faunal from the Awash Group, Middle Awash Valley, Afar, Ethiopia. *Journal of Vertebrate Paleontology* 2: 237–258.
- Kalb JE, Jaeger M, Jolly CJ, and Kana B (1982b) Preliminary geology, paleontology and palaeoecology of a Sangoan Site ad Andalee, Middle Awash Valley, Ethiopia. *Journal of Archaeological Science* 9: 349–363.
- Klein R and Cruz-Uribe K (1991) The bovids from Elandsfontein, South Africa, and their implications for the age, palaeoenvironment, and origins of the site. *African Archaeological Review* 9: 21–79.
- Klein R and Cruz-Uribe K (2000) Macromammals and reptiles. In: Barham L (ed.) *The Middle Stone Age of Zambia, South Central Africa*, pp. 51–56. Bristol, UK: Western Academic and Specialist Press.
- Klein RG (1984) Mammalian extinctions and Stone Age people in Africa. In: Martin PS and Klein RG (eds.) *Quaternary Extinctions*, pp. 553–573. Tucson: University of Arizona Press.
- Klein RG, Avery G, Cruz-Uribe K, et al. (1999) Duinefontein 2: An Acheulean site in the Western Cape Province of South Africa. *Journal of Human Evolution* 37: 153–190.
- Koch CP (1986) *The Vertebrate Taphonomy and Palaeoecology of the Olorgesailie Formation (Middle Pleistocene), Kenya*. PhD dissertation, Toronto, Ontario, Canada: University of Toronto.
- Kuman K and Clarke RJ (1986) Florisbad—New investigations at a Middle Stone Age hominid site in South Africa. *Geochronology* 1: 103–125.
- Lacruz RS, Brink JS, Hancox PJ, et al. (2003) Palaeontology and geological context of a Middle Pleistocene faunal assemblage from the Gladysvale Cave, South Africa. *Palaeontologia Africana* 38: 99–114.
- McBrearty S and Jablonski NG (2005) First fossil chimpanzee. *Nature* 437: 105–108.
- McBrearty S, Bishop LC, and Kingston J (1996) Variability in traces of Middle Pleistocene hominid behavior in the Kapthurin Formation, Baringo, Kenya. *Journal of Human Evolution* 30: 563–580.
- McKee JK, Thackeray JF, and Berger LR (1995) Faunal assemblage seriation of southern African Pliocene and Pleistocene deposits. *American Journal of Physical Anthropology* 96: 235–250.
- Mehlman MJ (1987) Provenience, age and associations of archaic *Homo sapiens* from Lake Eyasi, Tanzania. *Journal of Archaeological Science* 14: 133–162.
- O'Regan H, Bishop L, Lamb A, Elton S, and Turner A (2005) Large mammal turnover in Africa and the Near East 1.0–0.5 Ma. In: Head MJ and Gibbard P (eds.) *Early-Middle Pleistocene Transition: The Land–Ocean Evidence*, pp. 231–249. London: Geological Society of London. Special Publication No. 247.
- Plummer TW (1991) *Site Formation and Palaeoecology at the Early to Middle Pleistocene Locality of Kanjera, Kenya*. PhD dissertation, New Haven, CT: Yale University.
- Potts R and Deino A (1995) Mid-Pleistocene change in large mammal faunas of East Africa. *Quaternary Research* 43: 106–113.
- Potts R, Behrensmeier AK, and Ditchfield P (1999) Paleolandscape variation and Early Pleistocene hominid activities: Members 1 and 7, Olorgesailie Formation, Kenya. *Journal of Human Evolution* 37: 747–788.
- Raynal JP, Sbihi Alaoui FZ, Geraads D, Magago L, and Mohi A (2001) The earliest occupation of North-Africa: The Moroccan perspective. *Quaternary International* 75: 65–75.
- Schefuss E, Schouten S, Jansen JHF, and Sinninghe Damsté JS (2003) African vegetation controlled by tropical sea surface temperatures in the Mid-Pleistocene. *Nature* 422: 418–421.
- Shackley M (1980) An Acheulean industry with *Elephas recki* fauna from Namib IV, South West Africa (Namibia). *Nature* 284: 340–341.
- Shackleton NJ and Opdyke ND (1976) Oxygen-isotope and palaeomagnetic stratigraphy of Pacific core V28-239 Late Pliocene to latest Pleistocene. *Geological Society of America* 145: 449–464.
- Shipman P, Potts R, and Pickford M (1983) Lainyamok, a new Middle Pleistocene hominid site. *Nature* 306: 365–368.
- Trauth MH, Maslin MA, Deino A, and Strecker MR (2005) Late Cenozoic moisture history of East Africa. *Science* 309: 2051–2053.
- Turner A (1990) The evolution of the guild of larger terrestrial carnivores during the Plio-Pleistocene in Africa. *Geobios* 23: 349–368.
- Turner A (1995a) Plio-Pleistocene correlations between climatic change and evolution in terrestrial mammals: The 2.5 Ma event in Africa and Europe. *Acta Zoologica Cracoviensia* 38: 45–58.
- Turner A (1995b) The Villafranchian large carnivore guild: Geographic distribution and structural evolution. *Italian Journal of Quaternary Sciences* 8: 349–356.
- Turner A and Antón M (1999) Climate and evolution: Implications of some extinction patterns in African and European machairodontine cats of the Plio-Pleistocene. *Estudios Geológicos* 54: 209–230.
- Turner A and Antón M (2004) *Evolving Eden: An Illustrated Guide to the Evolution of the African Large Mammal Fauna*. New York: Columbia University Press.
- Turner A and Wood B (1993) Taxonomic and geographic diversity in robust australopithecines and other African Plio-Pleistocene larger mammals. *Journal of Human Evolution* 24: 147–168.
- Werdle L and Lewis ME (2005) Plio-Pleistocene Carnivora of eastern Africa: species richness and turnover patterns. *Zoological Journal of the Linnean Society* 144: 121–144.

Mid-Pleistocene of Australia

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Introduction

Australia (including New Guinea) separated from Antarctica in the early Paleogene, with a unique mix of Gondwanan and more cosmopolitan lineages aboard. For most of the Cenozoic, as the continent drifted northward from higher to lower mid-latitudes, intercontinental migration to and from Australia was severely constrained by significant transoceanic dispersal distances. This long isolation resulted in a terrestrial vertebrate radiation shaped largely by shifts in dominant vegetation types driven, in particular, by a general drying trend beginning in the middle Miocene. Much of Australia's Pleistocene vertebrate radiation can be envisaged as the evolutionary end-product of 15 million years of adaptation to increasingly drier conditions (Prideaux, 2004). This chapter reviews the zoogeography of the middle Pleistocene vertebrates of Australia, then briefly compares the fauna with what is known of the preceding and succeeding faunas.

Paleobiology and systematics of Pleistocene vertebrates have previously been reviewed by Murray (1991) and others. Accounts of early discoveries and development of the field of Pleistocene vertebrate paleontology in Australia are provided by Rich and Van Tets (1985) and Ride and Davis (1997). Reconstructions of a range of extinct Pleistocene vertebrates have been presented in Quirk and Archer (1983), Rich and Van Tets (1985) and Murray (1991). The approach taken in this chapter builds upon the work of Hope (1982) and Horton (1984), two pioneering studies in the analysis of Australian Pleistocene vertebrate distributions in relation to climate.

Caveats

Our understanding of the stratigraphy and chronology of Pleistocene vertebrates in Australia is poor compared with that of the continents in the Northern Hemisphere. This results from several factors, including the lack of widespread episodic ice sheets, isolation from intercontinental faunal interchange, a relative paucity of localities yielding sufficient taxonomic diversity, limitations of conventional dating techniques, too few detailed stratigraphic studies, and insufficient basic exploration. Significant improvements have been made over the past decade, facilitated by advances in dating methods and by more rigorous selection and study of sites. A middle Pleistocene (780 000–126 000 years ago) age was not confirmed for any Australian site by absolute dating until the publication of Ayliffe et al. (1998). No middle Pleistocene sites have yet been identified in New Guinea. Perusal of the pre-1998 literature reveals that the default age of 'late Pleistocene' has often been applied to assemblages lacking apparent phylogenetic precursors to later forms or containing any marsupials believed not to have persisted into the Holocene (Prideaux, 2004). In a

sense, 'late Pleistocene' (herein, 126 000–10 000 years ago) and 'Pleistocene' have often been treated as synonyms. So, while Australia presents a unique opportunity to study the responses of a fundamentally different vertebrate fauna to Quaternary environmental change, the necessary temporal framework is only just beginning to take shape.

Taxon occurrences within dated middle Pleistocene deposits are considered here, along with those of undated assemblages of possible or likely middle Pleistocene age. The Australian record is not yet adequately understood to facilitate the differentiation of Middle and earlier late Pleistocene faunas, so it is certain that at least some undated 'middle Pleistocene' assemblages included herein will eventually be found to be of last glacial age. Middle Pleistocene occurrences for all vertebrate species (Tables 1–3) are plotted within provinces based largely on Australia's modern climatic zones (Figures 1–2). Mammal species recorded in more than one province are plotted on the distribution maps (Figures 3–7).

Zoogeography

Frogs

Frogs are the only amphibians represented in the Australian Cenozoic record. Three families (Hylidae, Myobatrachidae, Microhylidae), 10 genera, and 24 species have been recognized from the middle Pleistocene (Table 1), but this assessment is the result of just two locality surveys. Tyler (1977) identified the ilia of five frog species in the Main Fossil Chamber deposit of Victoria Fossil Cave in the Naracoorte Caves World Heritage Area of southeastern South Australia (Figure 8). All those species are still present in the region today. Absence of the now locally common *Litoria raniformis* from Tyler's sample is the only current evidence suggesting change in the Naracoorte frog fauna since the middle Pleistocene. However, much more frog material has been retrieved from Victoria Fossil Cave and other variously-aged cave deposits in the region over the past three decades and should be a source of paleoclimatic information when eventually studied. An appraisal of the middle Pleistocene frog fossils from the Mt. Etna cave-fills near Rockhampton in coastal northeastern Queensland (Figure 9) reveals a fauna indicative of conditions wetter than exist in the area today (Hocknull, 2005). Deposits recognized as early Pliocene by biocorrelation have now been dated and are evidently middle Pleistocene in age (SA Hocknull, pers. comm., June 2005). The giant *Etnabatrachus maximus* is the only frog known from the middle Pleistocene of Australia that is now extinct.

Reptiles

Crocodyles

Strata in northern South Australia documented by Tedford and Wells (1990) have yielded the only dated middle Pleistocene

Table 1 Middle Pleistocene frog and reptile records of Australia by province. Bold letters indicate dated middle Pleistocene records; plain letters indicate possible (undated) middle Pleistocene records. Taxonomic uncertainty is indicated by '?'. Only taxa identified to genus or below are included. Unequivocally extinct species indicated by '†'

TAXON	Southwest	South central	Inland	Southeast	Tasmania	Lower east	Upper east	North
FROGS								
Hylidae								
<i>Litoria caerulea</i>								x
<i>Litoria citropa</i>						x		
<i>Litoria ewingi</i>				x				
<i>Litoria</i> sp. 1								x
<i>Litoria</i> sp. 2								x
<i>Litoria</i> sp. 3								x
<i>Litoria</i> sp. 4								x
<i>Nyctimystes</i> sp. 1								x
<i>Nyctimystes</i> sp. 2								x
<i>Etnabatrachus maximus</i> †								x
Myobatrachidae								
<i>Crinia signifera</i>				x				
<i>Crinia</i> sp.								x
<i>Geocrinia laevis</i>				x				
<i>Kyarranus</i> sp.								x
<i>Lechriodus</i> sp.								x
<i>Limnodynastes dumerili</i>				x				
<i>Limnodynastes peronii</i>						x		
<i>Limnodynastes spenceri</i>								x
<i>Limnodynastes tasmaniensis</i>				x				x
<i>Limnodynastes</i> sp. 1								x
<i>Limnodynastes</i> sp. 2								x
<i>Limnodynastes</i> sp. 3								x
<i>Neobatrachus</i> sp.								x
Microhylidae								
<i>Hylophorbus</i> sp.								?
REPTILES								
Crocodylidae								
<i>Crocodylus johnstoni</i>								x
<i>Crocodylus porosus</i>			?				x	
<i>Pallimnarchus gracilis</i> †								x
<i>Pallimnarchus pollens</i> †			x				x	
<i>Quinkana fortirostrum</i> †								x
<i>Quinkana</i> sp. †							x	
Chelidae								
<i>Chelodina longicollis</i>				x				
<i>Emydura macquarii</i>				x				
Meiolaniidae								
<i>Meiolania platycephala</i> †								?
<i>Ninjemys oweni</i> †							x	
Agamidae								
<i>Amphibolurus</i> sp.								x
<i>Hypsilurus</i> sp.								x
<i>Pogona barbata</i>				x				
<i>Pogona</i> sp.								x
<i>Tympanocryptis cephalus</i>								?
Scincidae								
<i>Cyclodomorphus gerrardii</i>								x
<i>Egernia whitii</i>				x				
<i>Eugongylus</i> sp.								x
<i>Eulamprus tympanum</i>				x				
<i>Lerista bougainvillii</i>				x				
<i>Sphenomorphus</i> sp. 1								x
<i>Sphenomorphus</i> sp. 2								x
<i>Tiliqua nigrolutea</i>				x		x		
<i>Tiliqua rugosa</i>				x				

(Continued)

Table 1 (Continued)

TAXON	Southwest	South central	Inland	Southeast	Tasmania	Lower east	Upper east	North
<i>Tiliqua scincoides</i>							x	x
<i>Tiliqua</i> sp.								x
Varanidae								
<i>Megalania prisca</i> †			x	x		x	x	x
<i>Varanus giganteus</i>						?		
<i>Varanus gouldii</i>				x				
<i>Varanus varius</i>				x				
<i>Varanus</i> sp. 1								x
<i>Varanus</i> sp. 2								x
Madtsoiidae								
<i>Wonambi naracoortensis</i> †	x			x		x		
<i>Yurlunggur</i> sp. †								x
Elapidae								
<i>Notechis scutatus</i>				x				
<i>Pseudechis porphyriacus</i>				x				
<i>Pseudonaja nuchalis</i>				x				

Table 2 Middle Pleistocene bird records of Australia by province. See **Table 1** caption for details

TAXON	Southwest	South central	Inland	Southeast	Tasmania	Lower east	Upper east	North
Casuariidae								
<i>Droaius novaehollandiae</i>	x		x	x	x	x	x	
Dromomithidae								
<i>Genyornis newtoni</i> †			x	x		?		
Podicipedidae								
<i>Podiceps</i> sp.			x					
Pelecanidae								
<i>Pelecanus cadimurka</i> †			x					
<i>Pelecanus conspicillatus</i>			x					
Phalacrocoracidae								
<i>Phalacrocorax melanoleucos</i>				x				
<i>Phalacrocorax carbo</i>			?					
<i>Phalacrocorax varius</i>			?					
Anhingidae								
<i>Anhinga laticeps</i>			x					
<i>Anhinga novaehollandiae</i>			x					x
Ardeidae								
<i>Ardea novaehollandiae</i>							?	
Ciconiidae								
<i>Ephippiorhynchus asiaticus</i>							x	
Anatidae								
<i>Anas castanea</i>			x					
<i>Anas superciliosa</i>								?
<i>Anseranus semipalmata</i>							?	x
<i>Biziura lobata</i>			?					
<i>Cygnus atratus</i>			?					
Accipitridae								
<i>Accipiter</i> sp.				x				
<i>Aquila audax</i>	x		?	x				
Falconidae								
<i>Falco berigora</i>			x					
Megapodiidae				x				
<i>Leipoa ocellata</i>								
<i>Progura gallinacea</i> †						?	x	
<i>Progura naracoortensis</i> †				x				
Phasianidae								
<i>Coturnix pectoralis</i>				x				

(Continued)

Table 2 (Continued)

TAXON	Southwest	South central	Inland	Southeast	Tasmania	Lower east	Upper east	North
<i>Coturnix ypsilophora</i>				x				
<i>Coturnix</i> sp.								x
Turnicidae								
<i>Turnix varia</i>				x				
Pedionomidae								
<i>Pedionomus torquatus</i>				x				
Rallidae								
<i>Gallinula mortierii</i>				x	x		x	?
<i>Gallinula tenebrosa</i>				x			x	
<i>Gallirallus philippensis</i>				x				
Otididae								
<i>Ardeotis australis</i>				?				
Burhinidae								
<i>Burhinus grallarius</i>				x				
Charadriidae								
<i>Charadrius australis</i>				x				
Scolopacidae								
<i>Calidris ruficollis</i>				x				
<i>Gallinago hardwickii</i>				x				
<i>Tringa glareola</i>				x				
Phoenicopteridae								
<i>Xenorhynchopsis minor</i> †			x					
<i>Xenorhynchopsis tibialis</i> †			x					
Palaelodidae								
<i>Palaelodus wilsoni</i> †			x					
Tytonidae								
<i>Tyto alba</i>				x				
<i>Tyto novaehollandiae</i>			?	x				
Strigidae								
<i>Ninox novaeseelandiae</i>				x				
Columbidae								
<i>Phaps chalcoptera</i>				x				
<i>Phaps</i> sp.							x	
Cacatuidae								
<i>Cacatua tenuirostris</i>				x				
<i>Callocephalon fimbriatum</i>				x				
<i>Calyptorhynchus banksii</i>				x				
<i>Calyptorhynchus lathamii</i>				x				
Psittacidae								
<i>Pezoporus wallicus</i>				x				
<i>Platycercus</i> sp. indet.				x				
Cuculidae								
<i>Centropus colossus</i> †				x				
Alcedinidae								
<i>Dacelo novaeguineae</i>				x				
Hirundinidae								
<i>Hirundo neoxena</i>				x				
Orthonychidae								
<i>Orthonyx hypsilophus</i> †				x				
Acanthizidae								
<i>Dasyornis broadbenti</i>				x				
Meliphagidae								
<i>Manorina melanocephala</i>				x				
Dicruridae								
<i>Grallina cyanoleuca</i>				x				
Artamidae								
<i>Gymnorhina tibicen</i>				x				
Corvidae								
<i>Corvus</i> sp.				x				

Table 3 Middle Pleistocene mammal records of Australia by province. See **Table 1** caption for details

TAXON	Southwest	South central	Inland	Southeast	Tasmania	Lower east	Upper east	North
Ornithorhynchidae								
<i>Ornithorhynchus anatinus</i>						x	x	
Tachyglossidae								
<i>Megalibgwilia ramsayi</i> †	x			x	x			
<i>Tachyglossus aculeatus</i>	x			x			x	
<i>Zaglossus Hackett</i> †	x							
Thylacinidae								
<i>Thylacinus cynocephalus</i> †	x	x		x		x	x	x
Dasyuridae								
<i>Antechinus flavipes</i>	x			x		?		x
<i>Antechinus minimus</i>				x				
<i>Antechinus agilis</i>				x				
<i>Antechinus swainsonii</i>				x				x
<i>Antechinus</i> sp. 1								x
<i>Antechinus</i> sp. 2								x
<i>Dasycercus</i> sp.		x						
<i>Dasyurus geoffroyi</i>	x	x						
<i>Dasyurus hallucatus</i>								x
<i>Dasyurus maculatus</i>				x		?		
<i>Dasyurus viverrinus</i>				x		x	?	x
<i>Ningau yvonnae</i>				x				
<i>Phascogale calura</i>				x				
<i>Phascogale tapoatafa</i>	x			x		x		x
<i>Phascogale</i> sp.						x		x
<i>Planigale maculata</i>								x
<i>Sarcophilus harrisii</i>	x			x	x			
<i>Sarcophilus lanarius</i>		x	x	x		x	x	x
<i>Sminthopsis crassicaudata</i>				x				
<i>Sminthopsis leucopus</i>				x				
<i>Sminthopsis macroura</i>								x
<i>Sminthopsis murina</i>	x			x		x		x
Peramelidae								
<i>Chaeropus ecaudatus</i> †		x						x
<i>Isoodon macrourus</i>								x
<i>Isoodon obesulus</i>	x			x		?		x
<i>Isoodon</i> sp.								x
<i>Perameles bougainville</i>	x	x		x		?		x
<i>Perameles gunnii</i>				x	x	x		
<i>Perameles nasuta</i>						?		
<i>Perameles</i> sp. 1							x	x
<i>Perameles</i> sp. 2								x
Thylacomyidae								
<i>Macrotis lagotis</i>		x	x			x		x
Phascolarctidae								
<i>Phascolarctos cinereus</i>	x			x		?		
<i>Phascolarctos stirtoni</i> †			?	x			x	
Diprotodontidae								
<i>Diprotodon minor</i> †							x	
<i>Diprotodon optatum</i> †		x	x	x		x	x	x
<i>Nototherium inerme</i> †							x	x
<i>Zygomaturus trilobus</i> †	x			x	x	x	x	
Palorchestidae								
<i>Palorchestes azael</i> †				x	x	x	x	x
<i>Palorchestes parvus</i> †				?		x	x	?
Vombatidae								
<i>Lasiorninus angustidens</i> †							x	
<i>Lasiorninus krefftii</i>				x		x		
<i>Phascolomys medius</i> †				?		x	x	
<i>Phascolonus gigas</i> †		x	x			x	x	x
<i>Ramsayia magna</i> †						x	?	

(Continued)

Table 3 (Continued)

TAXON	Southwest	South central	Inland	Southeast	Tasmania	Lower east	Upper east	North
<i>Vombatus hacketti</i> †	x							
<i>Vombatus ursinus</i>				x	x	x	x	x
<i>Warendja wakefieldi</i> †				x				
Thylacoleonidae								
<i>Thylacoleo carnifex</i> †	x	x	x	x	x	x	x	?
<i>Thylacoleo hill</i> †				x				x
Burramyidae								
<i>Cercartetus lepidus</i>				x				
<i>Cercartetus nanus</i>				x				
<i>Cercartetus</i> sp.								x
Pseudocheiridae								
<i>Petauroides</i> sp.								x
<i>Pseudocheirops</i> sp. 1								x
<i>Pseudocheirops</i> sp. 2								x
<i>Pseudocheirulus</i> sp. 1								x
<i>Pseudocheirulus</i> sp. 2								x
<i>Pseudocheirulus</i> sp. 3								x
<i>Pseudocheirus peregrinus</i>	x			x		?		
<i>Pseudocheirus</i> sp. 1								x
<i>Pseudocheirus</i> sp. 2								x
<i>Pseudocheirus</i> sp. 3								x
<i>Pseudokoala</i> sp. †								x
Petauridae								
<i>Dactylopsila</i> sp. 1								x
<i>Dactylopsila</i> sp. 2								x
<i>Petaurus breviceps</i>				x				
Acrobatidae								
<i>Acrobates pygmaeus</i>				x				
<i>Acrobates</i> sp.								x
Phalangeridae								
<i>Strigocuscus</i> sp.								x
<i>Trichosurus vulpecula</i>	x	x	x	x				
<i>Trichosurus</i> sp. 1						x		x
<i>Trichosurus</i> sp. 2								x
Hypsiprymnodontidae								
<i>Propleopus chillagoensis</i> †								x
<i>Propleopus oscillans</i> †			?	x			x	
<i>Propleopus wellingtonensis</i> †						x		
Macropodidae								
<i>Aepyprymnus rufescens</i>				x		x	x	
<i>Bettongia gaimardi</i>				x				
<i>Bettongia lesueur</i>	x	x	x	x				
<i>Bettongia penicillata</i>	x			x				
<i>Bettongia pusilla</i> †		x						
<i>Bettongia</i> sp.						x		
<i>Bohra paulae</i> †						x		
<i>Bohra</i> sp. 1 †		x						
<i>Bohra</i> sp. 2 †		x						x
<i>Borongaboodie hatcheri</i> †	x							
<i>Congruus congruus</i> †				x		x		
<i>Congruus kitcheneri</i> †	x							
<i>Congruus</i> sp. †		x		x				
<i>Dendrolagus</i> sp.								
<i>Kurrabi</i> sp. †								x
<i>Lagorchestes hirsutus</i>			x					x
<i>Lagorchestes leporides</i> †				x		x		
<i>Lagostrophus fasciatus</i>				x				
<i>Macropus eugenii</i>	x	x						
<i>Macropus ferragus</i> †		x	x				x	
<i>Macropus fuliginosus/giganteus</i>	x	x		x		x	x	?

(Continued)

Table 3 (Continued)

TAXON	Southwest	South central	Inland	Southeast	Tasmania	Lower east	Upper east	North
<i>Macropus greyi</i> †				X				
<i>Macropus irma</i>	X							
<i>Macropus parryi</i>				?				
<i>Macropus pearsoni</i> †							X	X
<i>Macropus piltonensis</i> †							X	
<i>Macropus robustus</i>						X	X	
<i>Macropus rufogriseus</i>				X	X			
<i>Macropus rufus</i>		X						
<i>Macropus siva</i> †			X			X	X	X
<i>Macropus thor</i> †							X	
<i>Macropus titan</i> †			?	X	X	X	X	
<i>Macropus</i> sp. nov. †	X	X		X				
<i>Macropus</i> sp. 1 †								X
<i>Macropus</i> sp. 2 †						X		
<i>Macropus</i> sp. 3 †						X		
<i>Macropus dryas</i> †							?	
<i>Macropus rama</i> †			X			?	X	
<i>Metasthenurus newtonae</i> †	X	X		X		X	X	
<i>Onychogalea fraenata</i>						?		
<i>Onychogalea lunata</i> †	X	X		X				
<i>Onychogalea unguifera</i>							X	
<i>Petrogale penicillata</i>						?		
<i>Petrogale</i> sp.								X
<i>Potorous gilbertii</i>	X							
<i>Potorous platyops</i> †	X			X				
<i>Potorous tridactylus</i>				X	X			
<i>Procoptodon goliah</i> †		X	X	X			X	
<i>Procoptodon pusio</i> †						X	X	
<i>Procoptodon rapha</i> †			X	X		X	X	
<i>Procoptodon browneorum</i> †	X	X		X			X	
<i>Procoptodon gilli</i> †				X				
<i>Procoptodon mccoysi</i> †				X				
<i>Procoptodon oreas</i> †						X	X	
<i>Procoptodon williamsi</i> †		X	X			X		
<i>Protemnodon anak</i> †			?	?	X	X	X	?
<i>Protemnodon brehus</i> †	?		?	?		X	X	?
<i>Protemnodon devisi</i> †						?		
<i>Protemnodon roechus</i> †		X	?	X		X	X	
<i>Setonix brachyurus</i>	X							
<i>Simosthenurus euryskaphus</i> †							X	
<i>Simosthenurus maddocki</i> †		X		X		X	X	
<i>Simosthenurus occidentalis</i> †	X			X	X	X		
<i>Simosthenurus antiquus</i> †			?					
<i>Simosthenurus baileyi</i> †		X	X	X				
<i>Simosthenurus brachyselenis</i> †						X		
<i>Simosthenurus pales</i> †	X		X	X		X	X	
<i>Sthenurus atlas</i> †			X			X	X	
<i>Sthenurus andersoni</i> †	X	X	X	X		X	X	
<i>Sthenurus stirlingi</i> †			X					
<i>Sthenurus tindalei</i> †		X	X					
<i>Troposodon kenti</i> †						X	X	
<i>Troposodon minor</i> †			X			X	X	
<i>Thylogale billardieri</i>				X	X		X	
<i>Thylogale</i> sp. 1								X
<i>Thylogale</i> sp. 2								X
<i>Wallabia bicolor</i>				X				
<i>Wallabia indra</i> †							X	
<i>Wallabia</i> sp. †						X		
Muridae								
<i>Conilurus albipes</i> †				X				

(Continued)

Table 3 (Continued)

TAXON	Southwest	South central	Inland	Southeast	Tasmania	Lower east	Upper east	North
<i>Conilurus</i> sp. †			?					x
<i>Hydromys chrysogaster</i>				x	x			
<i>Hydromys</i> sp.								x
<i>Leggadina</i> sp.								x
<i>Leporillus conditor</i>		x						
<i>Mastacomys fuscus</i>				x	x	x		
<i>Mesembriomys</i> sp.								x
<i>Notomys mitchellii</i>		?		x				
<i>Notomys</i> sp.								x
<i>Pogonomys</i> sp.								x
<i>Pseudomys albocinereus</i>	x			x				
<i>Pseudomys apodemoides</i>								
<i>Pseudomys auritus</i> †				x				
<i>Pseudomys australis</i>		x		x		x		
<i>Pseudomys fumeus</i>				x				
<i>Pseudomys gouldii</i> †				x				
<i>Pseudomys occidentalis</i>	x							
<i>Pseudomys oralis</i>		x						
<i>Pseudomys shortridgei</i>	x			x				
<i>Pseudomys</i> sp.								x
<i>Rattus fuscipes</i>	x			x				
<i>Rattus lutreolus</i>				x				
<i>Rattus sordidus</i>								?
<i>Rattus tunneyi</i>				x				
<i>Rattus</i> sp.								x
<i>Zyzomys</i> sp.								x
Vespertilionidae								
<i>Miniopterus schreibersii</i>				x				
<i>Nyctophilus geoffroyi</i>		?		x				
Megadermatidae								
<i>Macroderma gigas</i>						x		
<i>Macroderma</i> sp.								x

occurrences of crocodiles in Australia (Figure 10). Material has been referred to the large extinct *Pallimnarchus pollens* and, tentatively, to the extant estuarine crocodile *Crocodylus porosus*. Both taxa also occur in fluvial deposits of southeastern Queensland of possible middle Pleistocene age (Table 1). Remains of the poorly understood ziphodont crocodilian *Quinkana* have been collected from a cave deposit in the same region, while *Q. fortirostrum* is recorded from the Chillagoe Caves of northeastern Queensland (Molnar, 1981). The presence of middle Pleistocene crocodiles in the North Province accords with their current distribution throughout the estuaries and rivers of tropical Australia. Their occurrences in what are now subtropical and arid regions may reflect a more southerly extending summer monsoonal system, a phenomenon which occurred periodically during the Pleistocene (Magee et al. 2004). The effect of this may have been a more consistent flowing of the Diamantina (Warburton) River and Cooper Creek systems, which today carry monsoonal rains from central Queensland into the Lake Eyre Basin of inland Australia.

Turtles

Only four turtles of known or possible middle Pleistocene age have been identified to genus level or below (Table 1), the widespread extant chelids *Chelodina longicollis* and *Emydura macquarii*, which are known from cave deposits in southeastern

South Australia (Reed and Bourne, 2000), and two extinct, giant horned turtles, *Ninjemys oweni* and *Meiolania* cf. *platyceps*. *N. oweni* is recorded with certainty only from King's Creek on the Darling Downs in southeastern Queensland, but a specimen from northeastern New South Wales may also be referable to this species (Gaffney, 1996). No site on King's Creek is known to be of middle Pleistocene age, although it would be extraordinary if *N. oweni* did not occur in the region before the late Pleistocene. The same must also be said of the Lord Howe Island horned turtle, *M. platyceps*, and its Walpole Island neighbor, *M. mackayi*, which both come from undated horizons. The Wyandotte Local Fauna of far northeastern Queensland has yielded *M.* cf. *platyceps* (Gaffney, 1996). The Pleistocene distribution of Australian mainland meiolaniids falls entirely within the North and Upper East Provinces, which today are characterized by high summer rainfall and tropical and subtropical climates, respectively (Figures 1–2).

Snakes

The remains of snakes (especially vertebrae) are common in Pleistocene cave deposits, but have not been studied in detail. Apart from the identification by Smith (1976) of three locally extant elapid species in the middle Pleistocene of the Naracoorte Caves (Table 1), no other members of extant groups have been identified to genus level or below. The main focus

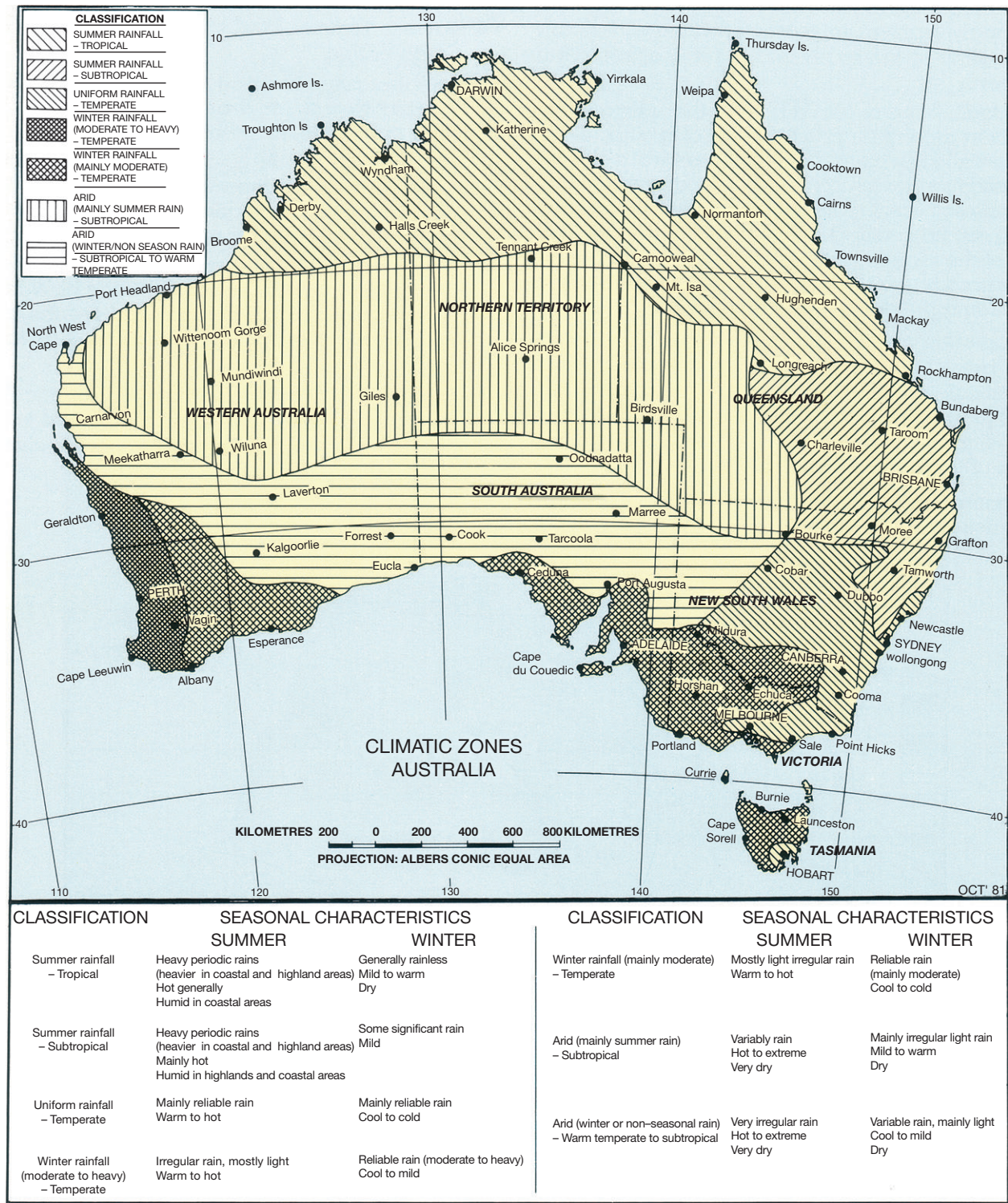


Figure 1 Climatic zones of Australia forming the basis for provinces recognized in this chapter. (Reproduced from Bureau of Meteorology, 1989. Climate of Australia, Australian Government Publishing Service, Canberra, with permission.)

of research has been the systematics of the madtsoiid *Wonambi naracoortensis*. This species is known from south-western and southeastern Australia, and a possible middle Pleistocene deposit in the Wellington Caves of central eastern New South Wales (Scanlon, 2005). The closely related

genus *Yurlunggur* is represented in the Wyandotte Local Fauna of northeastern Queensland by material unidentifiable to species (Scanlon, 2005). Like the meiolaniids, the Australian Pleistocene madtsoiids represent end-members of a Gondwanan radiation.

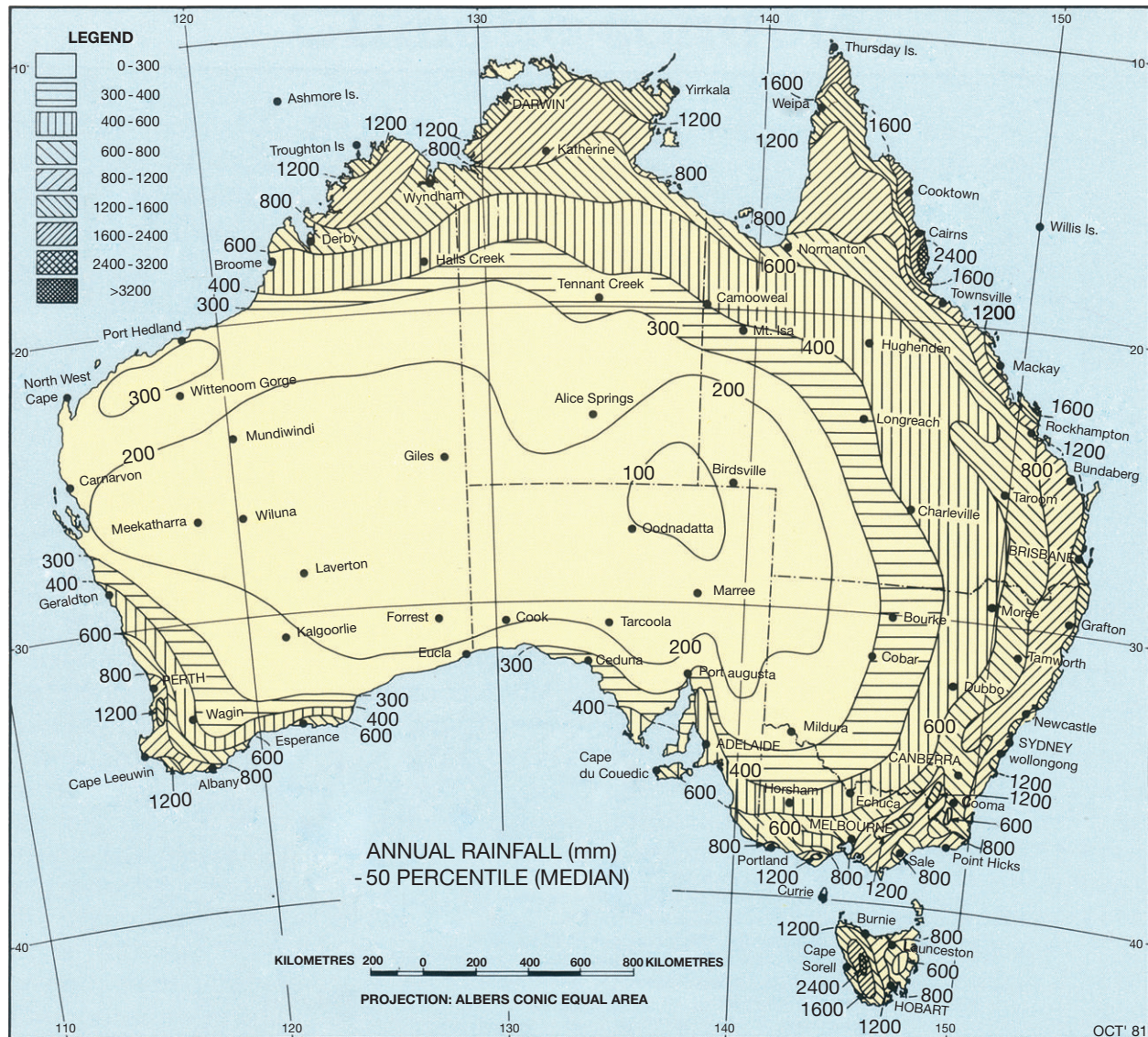


Figure 2 Median annual rainfall distribution of Australia. (Reproduced from Bureau of Meteorology, 1989. *Climate of Australia*, Australian Government Publishing Service, Canberra, with permission.)

Lizards

Australian Pleistocene lizard faunas have been the subject of little study. [Smith \(1976\)](#) remains the only primary study of a middle Pleistocene assemblage from southern Australia (Victoria Fossil Cave, Naracoorte). With the exception of the giant varanid *Megalania prisca*, the Naracoorte middle Pleistocene lizards thus far identified ([Table 1](#)) are all locally extant, which suggests that environmental conditions were at least broadly similar to present. [Hocknull \(2005\)](#) provided a census of the lizard fossils from the Mt Etna cave-fills ([Figure 9](#)) in the North Province, interpreting the presence of a species of the agamid *Hypsilurus* and the relative abundance of the skink *Cyclodomorphus gerrardii* to indicate a wetter environment than today, which agrees with the frog and mammal data. *M. prisca* is the only lizard with a middle Pleistocene record known to have subsequently become extinct. Although an uncommon faunal element wherever found, it was widely distributed through the eastern half of the continent from

the northeast tropics to the temperate southeast ([Reed and Hutchinson, 2005](#)). Modern mean annual rainfall throughout this range varies from ~150 to ~800 mm ([Figure 2](#)).

Birds

[Baird \(1991\)](#) provides a comprehensive review of the Australian Quaternary bird record. At least 36 families of known or possible middle Pleistocene age have been collected from many localities across Australia, but many represent singular finds and are indeterminate even to genus level. The only substantial middle Pleistocene bird fauna (41 species) known comes from the caves of southeastern South Australia ([Table 2](#)). Most of these species have modern records in the region, but four became extinct subsequent to the accumulation of the cave deposits in which they are represented: the giant flightless *Genyornis newtoni* (a dromornithid), the giant mallee fowl, *Progunya naracoortensis*, the giant coucal, *Centropus*

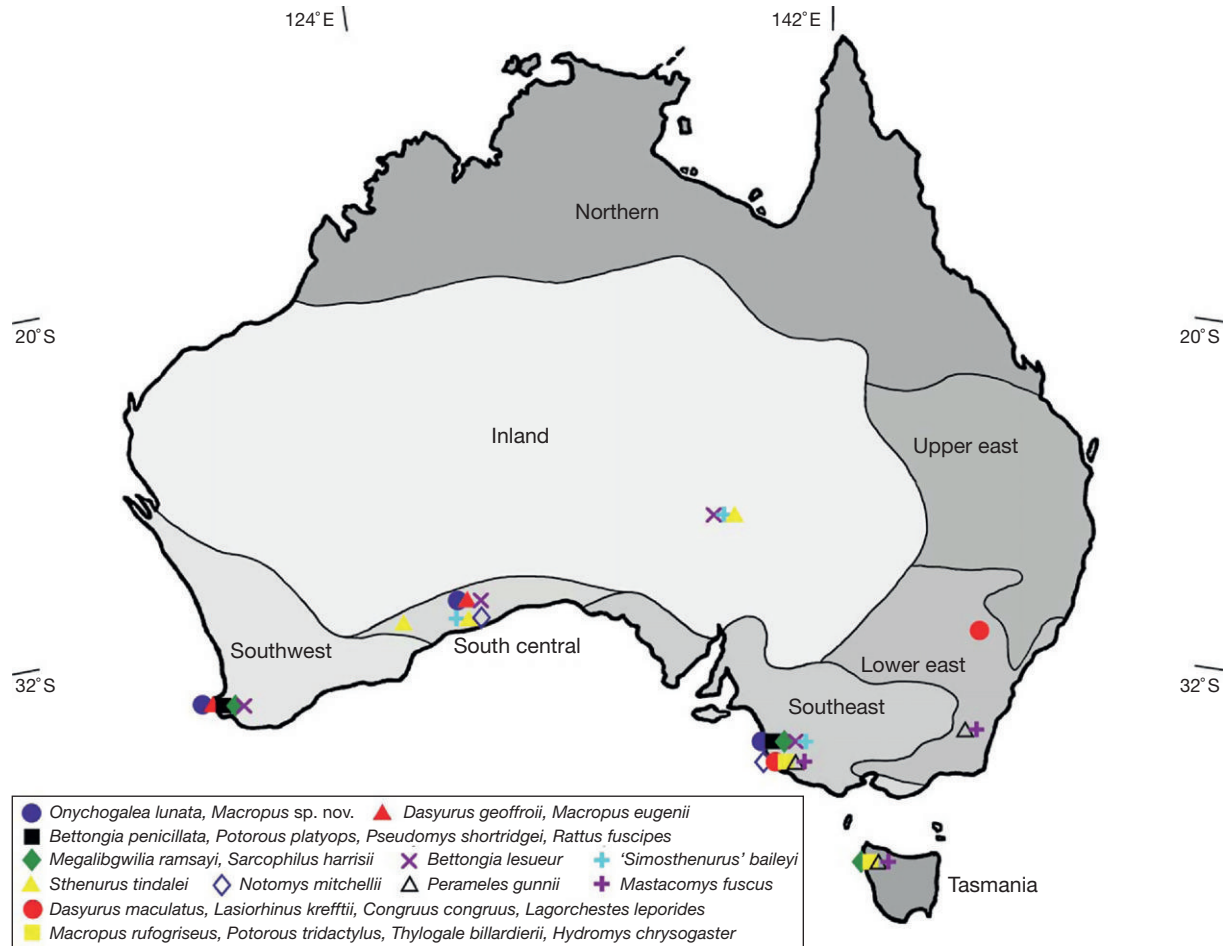


Figure 3 Middle Pleistocene records of mammal species with a mainly southern distribution.

colossus, and the logrunner, *Orthonyx hypsilophus*. Judging from the habitat preferences of their closest living relatives, the latter three species may have been adapted to woodland or forest. Three of four extant species (*Gallinula mortierii*, *Calyptorhynchus lathami*, *Pezoporus wallicus*) that no longer occur in the region, because of human-induced range contractions in the late Holocene, prefer similar habitats. The fourth regionally extinct species is the inland dotterel, *Peltohyas australis*, which is restricted today to arid inland Australia. *Genyornis* was widespread inland during the Pleistocene, and it is possible that, along with several of the mammals with middle Pleistocene records in southeastern South Australia (Southeast Province), *Genyornis* and *Peltohyas* extended southeastward during drier intervals. The presence of now-extinct endemic flamingos and flamingo-like palaelodids (Table 2) suggest more permanent and substantial water bodies in northern South Australia during the middle Pleistocene than exist in the region today.

Mammals

Monotremes (Ornithorhynchidae, Tachyglossidae)

Four monotremes occur in known or likely middle Pleistocene sites in Australia (Table 3), including the extant platypus, *Ornithorhynchus anatinus*, and short-beaked echidna, *Tachyglossus aculeatus*. The presence of *Ornithorhynchus* in fluvial deposits

of the Darling Downs and southeastern New South Wales (Figure 4) is congruent with its modern distribution through the waterways of subtropical and temperate eastern Australia (Strahan, 1995). *Tachyglossus* is currently distributed across the entire continent, so its presence in the Southwest, Southeast and Upper East Provinces during the middle Pleistocene is unsurprising. Remains of two giant extinct echidnas, '*Zaglossus*' *hacketti* and *Megalibgwilia ramsayi*, have been retrieved from caves in southwestern Australia. The latter also occurs in the Naracoorte Caves in the Southeast Province, and in Pleisto Scene Cave (Murray and Goede, 1977), the only dated middle Pleistocene locality in Tasmania (Roberts RG, pers. comm., September 2005; Figure 3). Echidnas are consistently rare in whichever Pleistocene assemblages they are found, so their absence from northern and inland sites should not be taken as evidence of absence from these regions.

Thylacines (Thylacinidae)

The last captive thylacine (*Thylacinus cynocephalus*) died in Tasmania in 1936 and the species is now presumed extinct, having disappeared from the mainland about 3 000 years ago. Middle Pleistocene deposits across the south and east of mainland Australia have yielded remains of *T. cynocephalus* (Table 3, Figure 6). It has not yet been recorded from inland Australia, nor possible middle Pleistocene sites in the north.

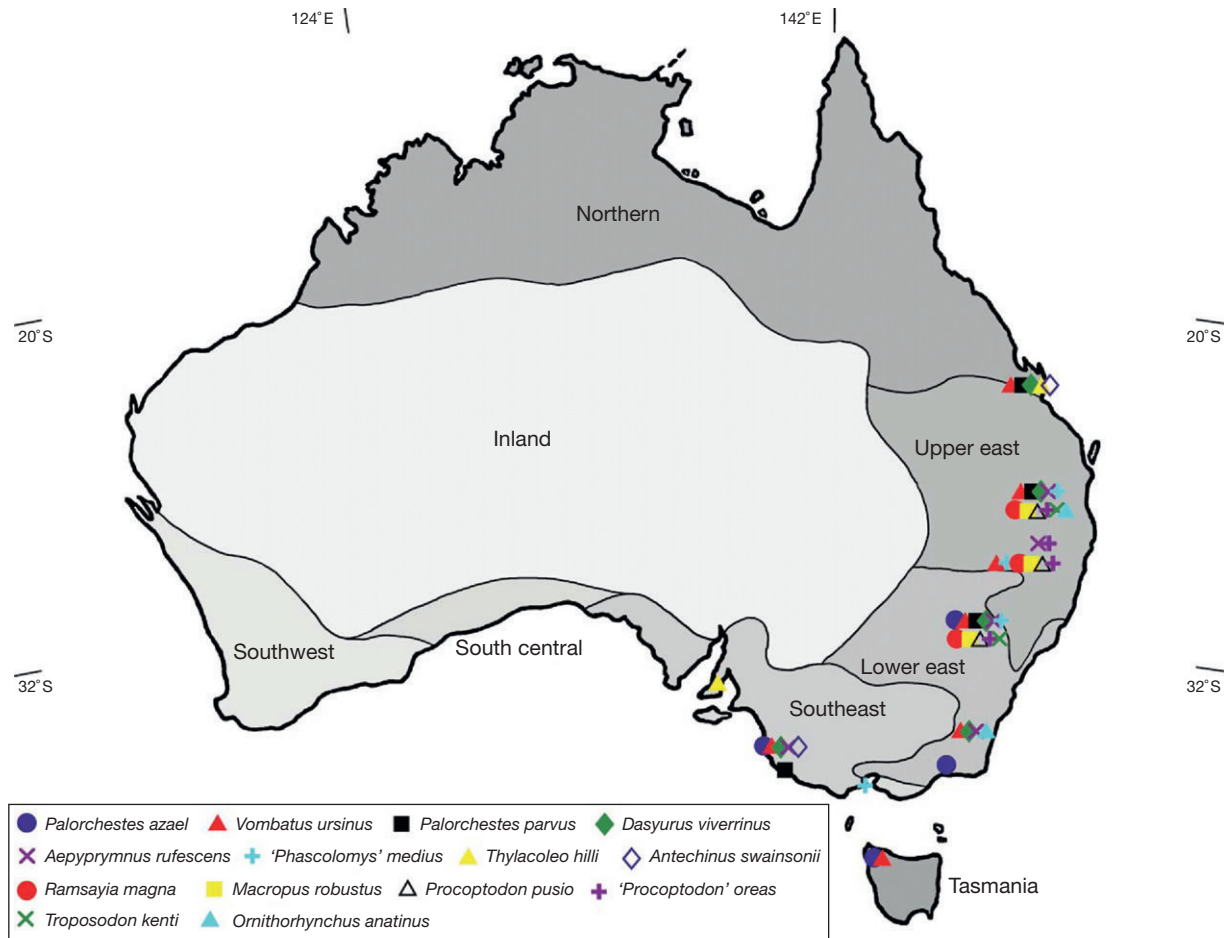


Figure 4 Middle Pleistocene records of mammal species with a mainly eastern distribution.

T. cynocephalus is absent from the Pleisto Scene Cave sample studied by Murray and Goede (1977), although its extensive Quaternary record in Tasmania suggests that it would also have occurred there during the middle Pleistocene.

Dasyurids (Dasyuridae)

Only two middle Pleistocene dasyurid assemblages have been documented, one from the Naracoorte Caves (Reed and Bourne, 2000) and a second from the Mt. Etna cave-fills (Hocknull, 2005). Most other dasyurid records (Table 3) constitute 'incidental' finds made during the collection of mega-faunal remains, and most of these represent the larger taxa (i.e., species of *Sarcophilus* and *Dasyurus*). As with other small vertebrates, there is a significant taphonomic bias against the preservation of small dasyurid remains in fluvial deposits. Among the Naracoorte middle Pleistocene dasyurids no longer inhabiting the local area, the two *Dasyurus* species became locally extinct in the late 1800s, while *Ningaui yvonnae* occurs in early Holocene strata (McDowell, 2001). *Sarcophilus* apparently disappeared from southern mainland Australia during the late Holocene, before European settlement. *Antechinus agilis*, *A. swainsonii* and *Phascogale calura* have no post-middle Pleistocene records at Naracoorte. However, with the exception of *Sarcophilus* (Tasmanian devil), all of these taxa have modern records elsewhere within the Southeast Province (Strahan,

1995). The *Antechinus* and *Dasyurus* species occur within more mesic areas of Victoria (mainly higher rainfall coastal habitats), while *N. yvonnae* and *P. calura* have modern records in semi-arid mallee habitats further to the north (Strahan, 1995). More refined biostratigraphic and paleoecological studies of the Pleistocene assemblages of the Naracoorte Caves are required before such distributional shifts can be interpreted as evidence of climate-driven faunal change in the region.

Bandicoots (Peramelidae, Thylacomyidae)

Most bandicoot species thus far retrieved from deposits of known or possible middle Pleistocene age (Table 3) have latest Pleistocene or Holocene records in the same regions. However, the Mt Etna cave-fills (Figure 9) represent a major exception. These sites include a remarkable 10 different bandicoots, consisting of the eight species listed in Table 3 and two unnamed extinct taxa (Hocknull, 2005). No more than four species are known to locally co-occur anywhere in the late Pleistocene, and only in a tiny area of northeastern Queensland rainforest do three species co-occur today (Strahan, 1995). While this may reflect considerable time-averaging and some marked climate-driven local habitat changes, it is worth noting that each of the four modern bandicoots recognized at Mt. Etna (*Chaeropus ecaudatus*, *Isodon obesulus*, *Perameles bougainville*, *Macrotis lagotis*) have recent records within a wide range of

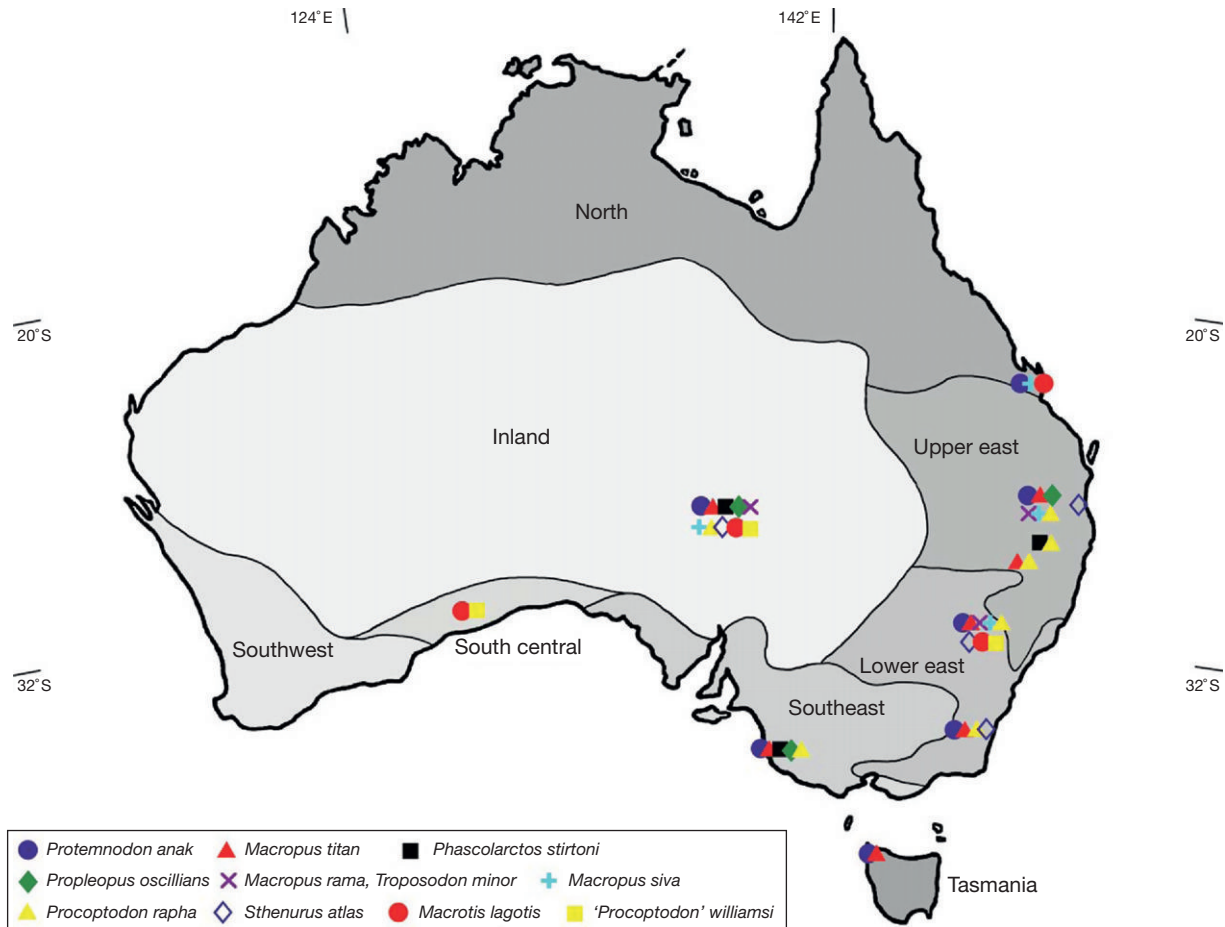


Figure 5 Middle Pleistocene records of mammal species that occur in the inland province and eastern Australia. Two of these species are also known from the South Central province.

habitats and areas differing markedly in total annual rainfall and season-ality of rainfall (Strahan, 1995; Figure 2).

Koalas (*Phascolarctidae*)

Two koala species have been recognized in the middle Pleistocene of Australia (Table 3, Figures 4–5), the modern *Phascolarctos cinereus*, and the 30% larger, but morphologically very similar, *P. stirtoni*. The taxonomic validity of the latter species deserves reinvestigation; it may simply represent a larger morph of *P. cinereus*. This species has a wide modern distribution through the sclerophyll forests and woodlands of mainland eastern Australia. Although present in the Middle and late Pleistocene of the southwest, the species did not persist there into the Holocene.

Wombats (*Vombatidae*)

Wombats were widespread across the continent during the middle Pleistocene (Table 3). Morphological evidence suggests that, like the modern *Vombatus ursinus*, *Lasiorninus krefftii*, and *L. latifrons*, all six extinct Pleistocene wombats were grazers, including the widely distributed, ~200 kg *P. gigas* (Figure 7). The middle Pleistocene records of *V. ursinus* largely conform with its modern distribution through the moderate to higher rainfall areas of southeastern and eastern Australia, including

Tasmania (Figure 4). A close relative, *V. hacketti*, is well-represented in middle Pleistocene cave deposits in the southwest but, as with the koala, it had become extinct by the latest Pleistocene. The small, gracile *Warenda wakefieldi* is known only as a rare element in the cave deposits of southeastern South Australia and southwestern Victoria. Middle Pleistocene wombat diversity appears to have been greatest in the Lower East and Upper East Provinces (Figures 3–4, 7).

Diprotodontoids (*Diprotodontidae*, *Palorchestidae*)

middle Pleistocene remains of at least five diprotodontoids and two palorchestids are known (Table 3). Their large body sizes (ranging in body weight from ~500 to ~2 000 kg) mean that their fossils are among the most conspicuous species in many localities, even though they are rarely abundant in any one assemblage. *Zygomaturus trilobus* was widespread through temperate southern and eastern Australia, extending no further north than southeastern Queensland (Figure 6). *Nototherium inerme* overlaps with the similarly-sized *Z. trilobus* in the Upper East Province, but is otherwise known only from the North Province (Figure 7), including a locality near Kununurra in the northwest (pers. obs.). Several *Diprotodon* species have been named, although their taxonomic validity remains unclear. Nevertheless, most authorities have agreed that at least two

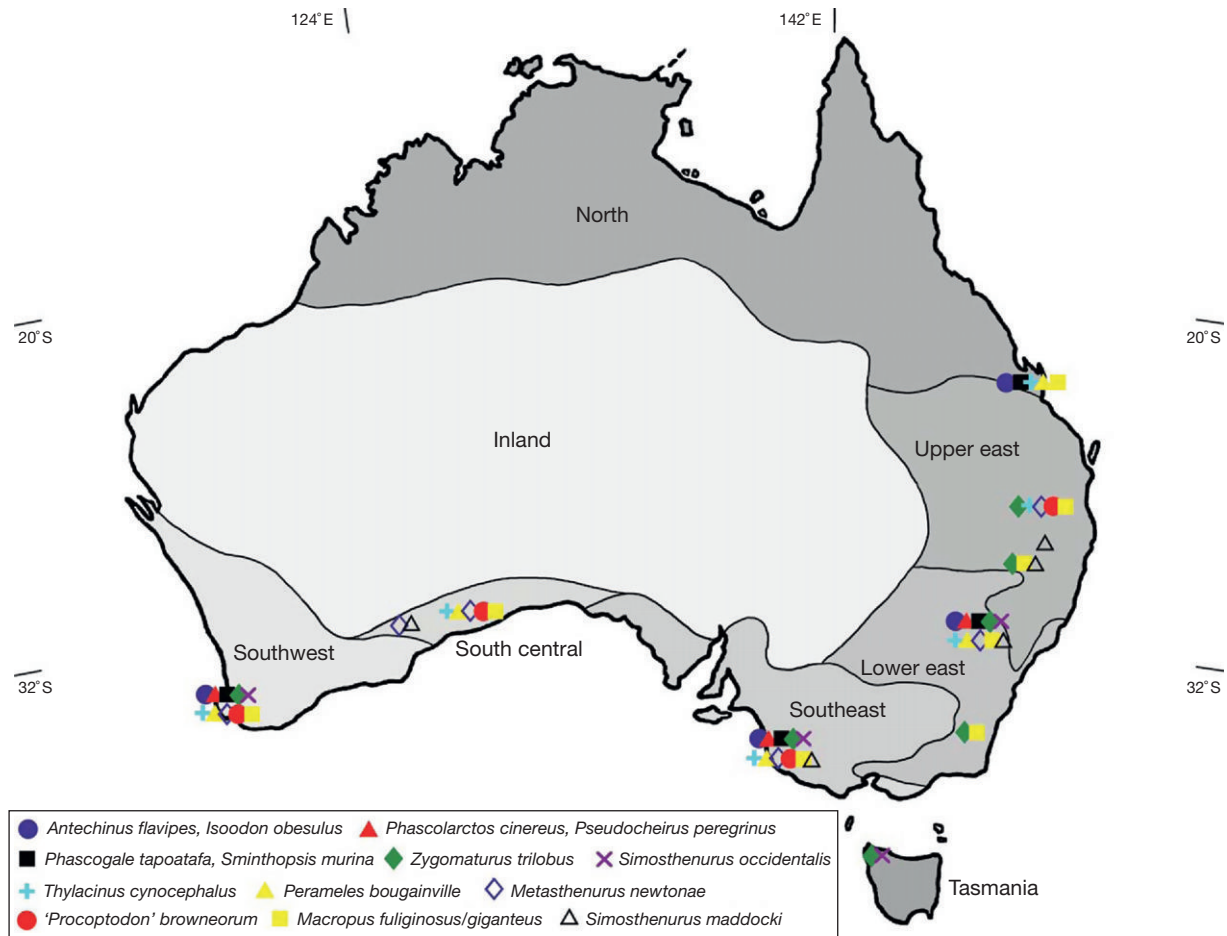


Figure 6 Middle Pleistocene records of mammal species that are widely distributed through southern and eastern Australia.

species existed, including *D. optatum*, the largest marsupial ever to have lived. *D. optatum* was one of the most widespread Pleistocene marsupials, ranging from the tropical north, through arid inland Australia, to the temperate south (Figure 7). There are no middle Pleistocene records of *D. optatum* in the southwest or Tasmania, but it is known from the late Pleistocene of King Island in Bass Strait, off the north coast of Tasmania. Two palorchestid species occur in deposits of known or possible middle Pleistocene age, but they remain unknown outside of mesic eastern Australia (Figure 4).

Marsupial lions (*Thylacoleonidae*)

Thylacoleo carnifex is the most commonly preserved large mammalian carnivore in the middle Pleistocene of Australia (Table 3, Figure 7). There are some subtle variations in body size and morphology across southern Australia, but whether these are taxonomically significant is unclear. Although the specimen of a large *Thylacoleo* species from the Mt. Etna cave-fills (Figure 9) in the North Province preserves insufficient features to definitively identify it to species (Hocknull 2005), it may be tentatively referred to *T. carnifex* since no other similar-sized thylacoleonids are known from the Pleistocene. *T. hilli* is a diminutive species recovered from Mt Etna as well as Curramulka Town Well Cave on Yorke Peninsula in South

Australia (Figure 4), a locality which has also yielded *T. carnifex* (Pledge 1977).

Possums (*Burramyidae*, *Pseudocheiridae*, *Petauridae*, *Acrobatidae*, *Phalangeridae*)

By far the most diverse middle Pleistocene possum assemblage thus far sampled is from the Mt. Etna cave-fills in the North Province (Table 3). Hocknull (2005) has recognized 17 species, including representatives of all five extant families, although none has yet been identified to species level. Such high species richness suggests a highly diverse floristic environment (including rainforest habitats), marked temporal changes in prevailing habitats, or a combination of both. Records for the six southeastern species are derived from the middle Pleistocene deposits of the Naracoorte Caves (Reed and Bourne 2000; Table 3), and each of them still occurs today within woodland or heathland habitats in southeastern South Australia. Similarly, *Pseudocheirus peregrinus* and *Trichosurus vulpecula* persist in the woodlands and forests of the Southwest Province, and the latter has modern records in the South Central Province (Strahan, 1995).

Kangaroos (*Hypsiprymnodontidae*, *Macropodidae*)

Kangaroos are the most conspicuous and diverse group of Late Cenozoic marsupials. In species richness and abundance, they

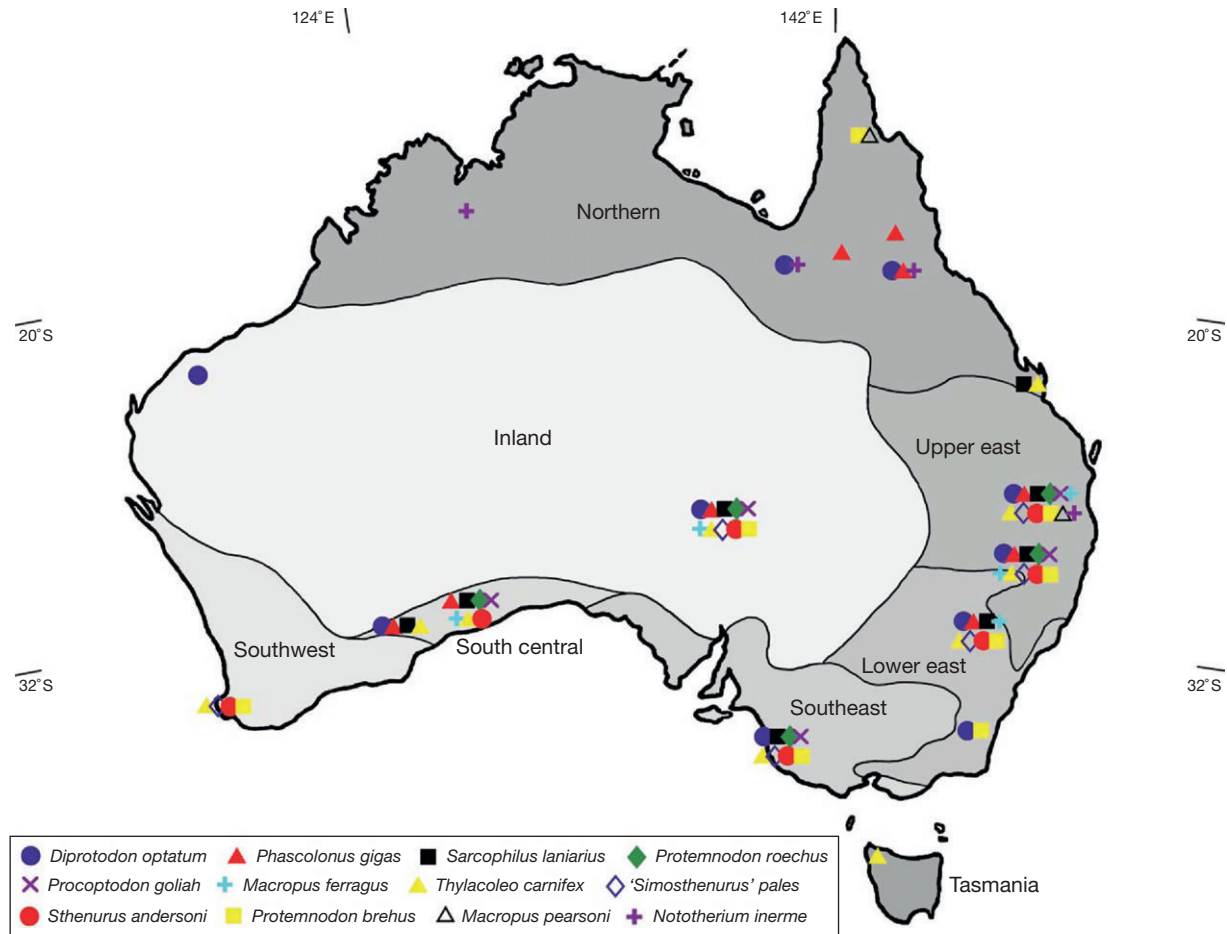


Figure 7 Middle Pleistocene records of mammal species that have either very wide (most species) or predominantly northern distributions (*Nototherium inerme*, *Macropus pearsoni*).

dominate the > 1-kg-body-weight component of most middle Pleistocene faunas. At least 72 species occur in known or possible middle Pleistocene sites (Table 3). The extinct, possibly omnivorous, propleopine hypsiprymnodontids are represented in eastern Australian deposits by three rare and poorly understood *Propleopus* species, but the group is unknown in Tasmania and west of longitude 137° E. Their inland middle Pleistocene record is restricted to one molar fragment referred to *Propleopus oscillans* from the Kutjitarra Formation of northern South Australia (pers. obs.; Figure 5).

At least nine potoroine macropodids are known from the middle Pleistocene (Table 3). *Borungaboodie hatcheri* is an extinct giant bettong represented by a single specimen from ~140ka sediments in Tight Entrance Cave in the Southwest Province (Prideaux, 1999). *Bettongia pusilla* persisted in the Nullarbor region into the Holocene (McNamara, 1997), but was not recorded there in modern times. Six potoroines are known from middle Pleistocene deposits of the Naracoorte Caves in the Southeast Province (Reed and Bourne, 2000); none occurs in the area today, although most have local late Holocene records. Today, *Aepyprymnus rufescens*, *Bettongia gaimardi*, and *Potorous tridactylus* inhabit wooded areas with median annual rainfalls ranging from 600 to 1600 mm, while most modern records of *Bettongia penicillata*, *B. lesueur*,

and *P. platyops* are in areas with an annual rainfall of 600 mm or less (Strahan, 1995; Figure 2). Even slight temporal fluctuations around the present-day 600-mm annual rainfall for the Naracoorte Caves may have been sufficient to facilitate the local occurrence of these six potoroines. The remaining regional middle Pleistocene records for potoroines are consistent with modern records of the species (Figures 3–4).

The extinct sthenurine macropodids (short-faced kangaroos) were browsers of sclerophyllous or xeromorphic vegetation, and were widespread and diverse during the middle Pleistocene. Sthenurine distributions group into several general patterns, that is species principally distributed through:

- southern and eastern Australia;
- south central and inland Australia;
- southern, eastern and inland Australia; and
- eastern and inland Australia (Prideaux 2004; Figures 3–7).

There are no published Pleistocene records of sthenurines north of latitude 26° S in central and eastern Australia, and they are unknown north of 31° S in Western Australia. Given the rarity and low diversity of larger marsupials so far retrieved from the North Province, the absence of evidence for sthenurines should not yet be taken as evidence of absence from tropical and subtropical arid Australia.



Figure 8 The Main Fossil Chamber in Victoria Fossil Cave, Naracoorte Caves World Heritage Area, southeastern South Australia (Southeast Province), preserves the richest and most diverse assemblage of middle Pleistocene vertebrates in Australia. (Photo courtesy of S Bourne.)



Figure 9 Cave-fills exposed on the side of Mt. Etna, near Rockhampton in northeastern Queensland, preserve the only diverse succession of middle Pleistocene vertebrates from northern Australia (North Province). (Photo courtesy of S Hocknull.)

Macropodines ('modern' kangaroos and wallabies) account for more than half of the macropodid species recorded from the middle Pleistocene (Table 3). Subsequent to the accumulation of the deposits in which they are represented, 26 macropodine species have become extinct, three of these since European settlement (*Lagorchestes leporides*, *Macropus greyi*, *Onychogalea lunata*). Four middle Pleistocene genera have no modern representatives: the tree-kangaroo genus *Bohra*, the wallaby genus *Congruus*, and two genera of large browsing kangaroos, *Protemnodon* and *Troposodon*. Species of *Bohra* are known from the Mt. Etna cave-fills (Hocknull, 2005), Wellington Caves in the Lower East Province, and caves on the Nullarbor Plain in the South Central Province (pers. obs.). This suggests that *Bohra* was less restricted to well-wooded habitats



Figure 10 Middle Pleistocene channel-fills intercepted by the Cooper Creek, northern South Australia (Inland Province), yield a diverse array of vertebrate remains. (Photo courtesy of R Wells.)



Figure 11 Three recently discovered caves on the Nullarbor Plain in Western Australia (South Central Province) have produced a remarkable array of well-preserved skeletons of middle Pleistocene vertebrates, including several new kangaroo taxa. (Photo courtesy of L Hatcher.)

than extant tree-kangaroos of the genus *Dendrolagus*, which are limited in distribution to the tropical rainforest of far northeastern Queensland and New Guinea (Dawson, 2004). Although poorly known until recently, species of *Congruus* were distributed across southern Australia and extended into the lower east (Dawson, pers. comm. July 2005; pers. obs.; Figure 3). The large browsing *Protemnodon* species are represented in most Australian middle Pleistocene assemblages (Figures 5, 7). Two *Troposodon* species have been retrieved from known or possible middle Pleistocene deposits (Figures 4–5). *T. minor* occurs in the Kutjitarra Formation of northern South Australia (Tedford and Wells, 1990), and in the Wellington Caves and Darling Downs deposits along with *T. kenti*. Both species are also known from the late Pliocene, apparently browsed tough vegetation, and were convergent upon the sthenurines in some craniodental attributes (Prideaux, 2004). *Macropus* is the most diverse fossil and modern macropodine genus, and at least 17 species have middle Pleistocene records (Table 3), including representatives of the three subgenera, *Macropus*, *Notamacropus*, and *Osphranter*. Only six *Macropus* species are known from the western half of Australia; four have records in the southwest caves and five occur in the Nullarbor caves (Figure 11).

Rodents (Muridae)

Murid rodent fossils have been recorded from known or possible middle Pleistocene localities in all major regions except for the Upper East Province (Table 3). Only *Conilurus* has been tentatively recorded from the inland (Tedford and Wells, 1990). The predominance of fluvial localities in the Upper East and Inland Provinces has resulted in a major taphonomic bias against the preservation of rodent remains. The two most diverse middle Pleistocene rodent assemblages are from the Mt. Etna cave-fills (Hocknull, 2005) in the north and the Naracoorte Caves in the southeast (Reed and Bourne, 2000). None of the Mt. Etna species has yet been identified to species, but each of the 13 Naracoorte species either remain locally extant or persisted there until the late Holocene (McDowell, 2001), probably until the arrival of Europeans.

Bats (Vespertilionidae, Megadermatidae)

As with dasyurids and murid rodents, remains of bats are usually only encountered after the application of wet-screening techniques. At present, there are only two dated and three possible middle Pleistocene bat records (Table 3). Each of the three nominal species is extant, and no record lies outside of current ranges.

Comparison with the Early Pleistocene Record

Only a handful of Australian sites have been recognized as early Pleistocene in age, based largely on the reversed magnetic polarity of the fossiliferous sediments (Whitelaw, 1991; Tedford, 1994). By far the most faunally diverse site is situated near Portland in southwestern Victoria (Flannery and Hann, 1984). Several undescribed species apparently express stages of evolution intermediate between late Pliocene and middle Pleistocene relatives, although the first detailed study of its fauna is only now being undertaken (Piper K, pers. comm., October 2005). All genera presently recognized at Portland persisted into the middle Pleistocene elsewhere, with the exception of an as-yet-undescribed ektopodontid possum (Rich, 1991).

Comparison with the Late Pleistocene Record

Absence or very low diversity of megafaunal species in Australian localities younger than ~50 ka is the only biostratigraphic marker presently useful for differentiating late Pleistocene assemblages from those of middle Pleistocene age, although sound evidence for persistence into the late Pleistocene has been retrieved for only 19 of the roughly 50 megafaunal species. However, given the paucity of dated Australian sites within the 126–50 ka period, and the previous rarity of many of these species in the middle Pleistocene, 19 of 50 species probably represents a substantial underestimate of the number that became extinct during the late Pleistocene (Prideaux, 2004). Whether extinction of the megafauna, and a range of often-overlooked medium-sized taxa, was prolonged or occurred relatively rapidly at the end of this interval, remains to be effectively established.

Ongoing Research

Although several general distribution patterns are evident within the 'time-averaged' middle Pleistocene vertebrate fauna of Australia (Figures 3–7), the current data are extremely limited geographically and temporally. A detailed understanding of the regionally variable influences of climatic fluctuations on vegetation structures and taxon distributions will emerge only after many more years of focused research. Ongoing studies into the middle Pleistocene vertebrate record of Australia are centered on five main topics:

- (1) Discovery and analysis of assemblages from regions for which the existing records are least understood (e.g., Tasmania, North Province);
- (2) Biostratigraphic studies documenting and interpreting shifts in species diversity and relative abundance within well-stratified sequences (e.g., Naracoorte Caves);
- (3) Paleocological analyses combining traditional morphological and relative abundance methods with stable isotopic and enamel microwear techniques;
- (4) Taphonomic studies describing and comparing site accumulation mechanisms and resultant biases in the fossil record.
- (5) Application of multiple geochronologic methods (e.g., U-series, optically stimulated luminescence (OSL)) to the dating of key sites.

See also: **Pollen Records, Late Pleistocene: Australasia.**

References

- Ayliffe LK, Marianelli PC, Moriarty KC, et al. (1998) 500 ka precipitation record from southeastern Australia: Evidence for interglacial relative aridity. *Geology* 26: 147–150.
- Baird RF (1991) Avian fossils from the Quaternary of Australia. In: Vickers-Rich P, Monaghan JM, Baird RF, and Rich TH (eds.) *Vertebrate Palaeontology of Australasia*, pp. 809–870. Melbourne: Monash University.
- Dawson L (2004) A new Pliocene tree kangaroo species (Marsupialia, Macropodinae) from the Chinchilla Local Fauna, southeastern Queensland. *Alcheringa* 28: 267–273.
- Flannery TF and Hann L (1984) A new macropodine genus and species (Marsupialia: Macropodidae) from the Early Pleistocene of southwestern Victoria. *Australian Mammalogy* 7: 193–204.
- Gaffney ES (1996) The postcranial morphology of *Meiolania platyceps* and a review of the Meiolaniidae. *Bulletin of the American Museum of Natural History* 229: 1–166.
- Hocknull SA (2005) Ecological succession during the Late Cainozoic of central eastern Queensland: Extinction of a diverse rainforest community. *Memoirs of the Queensland Museum* 51: 39–122.
- Hope JH (1982) Late Cainozoic vertebrate faunas and the development of aridity in Australia. In: Barker WR and Greenslade PJM (eds.) *Evolution of the Flora and Fauna of Arid Australia*, pp. 85–100. Adelaide: Peacock Press.
- Horton DR (1984) Red kangaroos: last of the Australian mega-fauna. In: Martin PS and Kline RG (eds.) *Quaternary Extinctions: A Prehistoric Revolution*, pp. 639–680. Tucson: University of Arizona Press.
- Magee JW, Miller GH, Spooner NA, and Questiaux D (2004) Continuous 150 k. y. monsoon record from Lake Eyre, Australia: insolation-forcing implications and unexpected Holocene failure. *Geology* 32: 885–888.
- McDowell MC (2001) The analysis of late Quaternary fossil mammal faunas from Robertson Cave (5U17, 18, 19) and Wet Cave (5U10, 11) in the Naracoorte Caves World Heritage Area, South Australia. MSc Thesis. Adelaide: Flinders University.
- McNamara JA (1997) Some smaller macropod fossils of South Australia. *Proceedings of the Linnean Society of New South Wales* 117: 97–105.

- Molnar RE (1981) Pleistocene ziphodont crocodilians of Queensland. *Records of the Australian Museum* 33: 803–834.
- Murray PF (1991) The Pleistocene megafauna of Australia. In: Vickers-Rich P, Monaghan JM, Baird RF, and Rich TH (eds.) *Vertebrate Palaeontology of Australasia*, pp. 1071–1163. Melbourne: Monash University.
- Murray PF and Goede A (1977) Pleistocene vertebrate remains from a cave near Montagu, N.W. Tasmania. *Records of the Queen Victoria Museum* 60: 1–29.
- Pledge NS (1977) A new species of *Thylacoleo* (Marsupialia: Thylacoleonidae) with notes on the occurrences and distribution of Thylacoleonidae in South Australia. *Records of the South Australian Museum* 17: 261–267.
- Prideaux GJ (1999) *Borongaboodie hatcheri* gen. et sp. nov., a very large bettong (Marsupialia: Macropodoidea) from the Pleistocene of southwestern Australia. *Records of the Western Australian Museum Supplement No. 57*: 317–329.
- Prideaux GJ (2004) Systematics and evolution of the sthenurine kangaroos. *University of California Publications in Geological Sciences* 146(i–xvii): 1–623.
- Quirk S and Archer M (eds.) (1983) *Prehistoric Animals of Australia*. Sydney: The Australian Museum.
- Reed EH and Bourne SJ (2000) Pleistocene fossil vertebrate sites of the South East region of South Australia. *Transactions of the Royal Society of South Australia* 124: 61–90.
- Reed EH and Hutchinson MN (2005) First record of a giant varanid (*Megalania*, Squamata) from the Pleistocene of Naracoorte, South Australia. *Memoirs of the Queensland Museum* 51: 203–213.
- Rich PV and Van Tets GF (eds.) (1985) *Kadimakara: Extinct Vertebrates of Australia*. Melbourne: Pioneer Design Studio.
- Rich TH (1991) Monotremes, placentals, and marsupials: Their record in Australia and its biases. In: Vickers-Rich P, Monaghan JM, Baird RF, and Rich TH (eds.) *Vertebrate Palaeontology of Australasia*, pp. 893–1069. Melbourne: Monash University.
- Ride WDL and Davis AC (1997) Origins and setting: Mammal Quaternary palaeontology in the Eastern Highlands of New South Wales. *Proceedings of the Linnean Society of New South Wales* 117: 197–222.
- Scanlon JD (2005) Cranial morphology of the Plio-Pleistocene giant madtsoiid snake *Wonambi naracoortensis*. *Acta Palaeontologica Polonica* 50: 139–180.
- Smith MJ (1976) Small fossil vertebrates from Victoria Cave, Naracoorte, South Australia. IV. Reptiles. *Transactions of the Royal Society of South Australia* 100: 39–51.
- Strahan R (ed.) (1995) *The Mammals of Australia*. Sydney: Reed Books.
- Tedford RH (1994) Succession of Pliocene through medial Pleistocene mammal faunas of southeastern Australia. *Records of the South Australian Museum* 27: 79–93.
- Tedford RH and Wells RT (1990) Pleistocene deposits and fossil vertebrates from the 'Dead Heart of Australia'. *Memoirs of the Queensland Museum* 28: 263–284.
- Tyler MJ (1977) Pleistocene frogs from caves at Naracoorte, South Australia. *Transactions of the Royal Society of South Australia* 101: 85–89.
- Whitelaw MJ (1991) Magnetic polarity stratigraphy of Pliocene and Pleistocene fossil vertebrate localities in southeastern Australia. *Geological Society of America Bulletin* 103: 1493–1503.

Mid-Pleistocene of Europe

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Introduction

The middle Pleistocene in Europe is characterized by multiple faunal exchanges that were mainly controlled by climatic oscillations. The definition of the early middle Pleistocene boundary is still a matter of debate among Quaternary geologists, stratigraphers, and paleontologists. Some authors strongly support the placing of the beginning of the middle Pleistocene at the Matuyama/Bruhnes paleomagnetic reversal (ca. 0.78 Ma) (XII INQUA Congress), whereas other researchers correlate the base of the middle Pleistocene with Marine Oxygen Isotopic Stage (MIS) 25, dated around 0.9 Ma (see Gliozzi et al. (1997)). A wider consensus among Quaternary geoscientists places the middle late Pleistocene boundary at 0.125 Ma, correlated with MIS 5e.

The middle Pleistocene vertebrate record must be considered against a backdrop of the latest early Pleistocene bio-events, since these form the basis for the definition of middle Pleistocene terrestrial vertebrate communities. Vertebrate paleontologists are used to subdividing geological time using biochronological or biostratigraphical scales. These are based on the evolutionary degree of the faunal elements of the assemblages and on dispersal events. Two different biochronological scales are in use for the middle Pleistocene in Europe. The first is based on micromammals and spans from the Biharian to the early middle Toringian; in particular the fossil assemblages are biochronologically divided on the basis of water voles belonging to the *Mimomys savini*–*Arvicola mosbachensis* lineage (Koenigswald and Kolfschoten, 1996; Maul et al., 1998) (Figure 1).

The other biochronological scale is based on large mammals and spans from the Galerian to the middle Aurelian (*sensu* Gliozzi et al. (1997)). The term Galerian was introduced by Ambrosetti et al. (1972) to define a particular faunal assemblage in Europe. An updated and more accurate biochronological definition is given by Gliozzi et al. (1997), who considered Galerian to be a ‘Mammal Age’ marked by the first appearance in Italy of the large-sized megacerine deer *Praemegaceros* (= *Megaceroides* = *Megaloceros*) *verticornis* (Figure 2).

In the literature, the vertebrate faunas referable to the early middle Pleistocene and having a similar composition to the Galerian fauna were often referred to as Cromerian or Tiraspolian but, as stated by Azzaroli et al. (1988) the first term ‘defines a stage based on pollen and of much shorter duration than Galerian,’ whereas the second one, based on a faunal complex from Moldova and used for Russian and Eastern Europe deposits, has to be restricted to the younger part of the Galerian. At present, the data collected and analysed from Dutch and UK deposits support the evidence for as many as four and possibly five ‘Cromerian Complex’ interglacials, although these may not all be of full interglacial status.

On the basis of the Italian fossil record, post-Galerian faunas that are still referable to the middle Pleistocene have been attributed to a different, younger Mammal Age, the Aurelian (Gliozzi et al., 1997). The Aurelian is marked by the first appearance in Italy of the giant deer *M. giganteus*, the ‘modern’ wolf *Canis lupus* and the cave bear *Ursus spelaeus*.

These ‘Mammal Ages’ have subsequently been further subdivided into faunal units on the basis of bio-events that produced changes in the composition of assemblages.

During the last three decades many paleontologists established biochronological frameworks mainly based on the fossil record of their own countries, for example: (1) faunal units (FUs), Italy (Gliozzi et al., 1997); (2) mammal zones (MNQs), Western Europe (Guerin, 1990); and (3) mammal assemblage zones (MAZs), Great Britain (Schreve, 2001). A general comparison of such schemes reveals that, despite the overall regional significance of many of the biochronological units, and the uncertain chronological position of some fossiliferous localities, bio-events of wider significance can be defined and correlated at a European scale (Figure 1).

In this article, the faunal changes and the evolutionary trends witnessed during the late early Pleistocene and middle Pleistocene in the terrestrial ecosystems of Europe are placed into three different ‘episodes’ of general biochronological significance: (1) the latest early Pleistocene (Villafranchian–Galerian transition); (2) the early middle Pleistocene; and (3) the late middle Pleistocene.

The Latest Early Pleistocene (Villafranchian–Galerian Transition)

The faunal renewal recorded in terrestrial vertebrate faunal assemblages of Europe during the latest early Pleistocene is known as the ‘end Villafranchian’ dispersal event (Azzaroli et al. (1988), and references therein). Furthermore, the same authors underlined that the transition between Villafranchian and Galerian did not take place at once. Subsequent further discoveries and new studies continued to provide more data on this transitional phase. Moreover, new terms have since been proposed to define such an informal biochron, referable to faunal assemblages calibrated with the base of the Jaramillo submagnetochron (Protogalerian, Caloi and Palombo (1996) and Epivillafranchian, Kahlke (2001)).

During the latest early Pleistocene, a series of dispersals started in Eurasia, involving mammals of Asiatic and African origin and probably related to climatic change.

Large felids such as *Puma* and the dirk-toothed cat *Megantreon*, together with large-sized deer such as *Eucladoceros*, are recorded for the last time and then become extinct. Among the ungulates *Bison* and *Praemegaceros* occur and their presence

		GPTS MIS Ma $\delta^{18}\text{O}$ BP	Britain	Central Europe	Small mammal ages	Large mammal ages	MAZ	MNQ	FU	European Russia	Selected local faunas		
											Britain	S-W Europe	N-E Europe
Pleistocene	Upper	5e	0.1	Ipswichian	Eemian	<i>A. terrestris</i>	Middle				Mikulino Moskva	Aveley	Terra Amata, Lazaret, Vitinia
		6	0.2	Wolstonian	Saalian	<i>Aurelian</i>							
		7											
		8											
		9	0.3	Hoxnian	Holsteinian	<i>Toringian</i>							
	Middle	10				<i>Arvicola mosbachensis</i>	Late	24	Torre in Pietra	Likhvin	Purfleet	Polledrara, Malagrotta, Castel di Guido, Torre in Pietra	
		11	0.4	Anglian	Elsterian								
		12											
		13											
		14	0.5										
		15											
		16	0.6	Cromerian	Cromerian								
		17											
		18	0.7	Cromerian Complex (Cromerian s.l.)	Cromerian Complex (Cromerian s.l.)								
		19											
	Lower	0.8				<i>Mimomys savini</i>	Middle	21	Isernia	Tiraspol	West Runton, Sidestrand, Sugworth, Little Oakley	Soleilhac, Isernia La Pineta, P. Galeria 2	Bilzingsleben II
		0.9											
		27											
		31	1.0			<i>Mimomys savini</i> - <i>M. pusillus</i>							
		1.1											
							Late Villafranchian						

Figure 1 Schematic correlation chart of Geochronology and Mammal Biochronology Units including selected European localities. Columns GPTS, Britain, Central Europe, Small Mammal Ages and European Russia adapted from [Maul \(2002\)](#); Large Mammal Ages and FU from [Petronio and Sardella \(1999\)](#); MAZ from [Schreve \(2001\)](#); MNQ adapted from [Alberdi et al. \(1997\)](#).

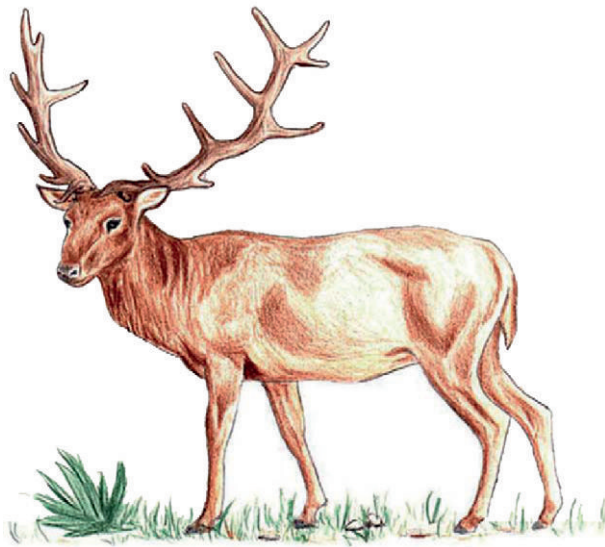


Figure 2 Reconstruction of *Praemegaceros verticornis*. Drawing by G. Di Stefano, modified from Di Stefano, G. Petronio, C. and Sardella, R. 2005 – Large bovid and cervids from latest Villafranchian-Galerian faunas of Italy. QUATERNAIRE (International Journal of the French Quaternary Association) Hors-serie-n. 2: 95–102.

continues to characterize subsequent middle Pleistocene faunal assemblages.

In SW Europe, faunal assemblages referable to the early Galerian, are present at Colle Curti (Central Italy) and Redicicoli (Roma) in Italy ([Glozzi et al., 1997](#); [Milli and Palombo, 2005](#)), Le Vallonet (Alpes Maritimes) in France ([Lumley et al., 1988](#)), Huéscar 1 and Cullar de Baza 1 (Granada), and Sierra de Atapuerca (Burgos) in Spain ([Alberdi et al., 1997](#), and references therein; and Bermudez de Castro et al., 2004, and references therein).

At Colle Curti, together with *P. verticornis*, several Villafranchian taxa, such as the cervid *Axis eurygonos*, the hunting dog *Lycan lycaonoides*, the giant hyaena *Pachycrocuta brevirostris*, the saber-toothed cat *Homotherium latidens*, the mammoth *Mammuthus meridionalis*, a small rhino (*Stephanorhinus* sp.), and the rodents *Pliomys lenki* and *Microtus* (*Allophaiomys* sp.) occurred. In the Vallonet cave deposits, the rich 'fossil assemblage 3' has been referred to the Jaramillo magnetochron and includes 25 taxa, with several carnivores and, amongst the ungulates, *Ammotragus europaeus*. Rodents are represented by *Microtus* (*Allophaiomys*) *nutiensis*, *Ungaromys nanus*, and *Mimomys* cf. *savini* ([Lumley et al., 1988](#)).

In Spain, a large number of mammal species are represented at different sites located in the Sierra de Atapuerca (Burgos) (e.g., Trinchera Dolina, Trinchera Elefante). The mammal assemblages recorded by the entire stratigraphic

sequence at Trinchera Dolina span from the late early to the middle Pleistocene. In particular, Trinchera Dolina level 6 (TD6) has yielded one of the most ancient fossil hominin assemblages recorded thus far from Western Europe, dated to approximately 0.8 Ma (Bermudez de Castro et al., 2004). At Trinchera Elefante the occurrence and the relative abundance of *Hippopotamus* and *Castor* among mammals, and of *Anas* and the sea eagle *Haliaeetus* among birds, provides evidence of the presence of inland water biotopes.

In Northern and Central Europe, the fossiliferous locality of Untermassfeld (Thuringia, Germany) can be considered as one of the most representative of the late early Pleistocene.

Since its discovery, in 1978, on the right slope of the river Werra, this site has been intensively excavated. Until present, more than 9500 determinable large mammals and 3000 microvertebrates have been discovered and 64 vertebrate taxa have been defined (toads and frogs amongst the amphibians; reptiles including skinks, geckos, and pond tortoises; and birds including swans, geese, birds of prey, gallinaceous, and crows) (Kahlke (2001), and references therein).

In the large mammal assemblage, ungulates are represented by an unusual species of bison (the long-legged *Bison menneri*) and the cervids (*Eucladoceros giulii* and *Capreolus cusanoides*). Despite the presence of several Villafranchian elements (e.g., *Megantereon*, *Puma*, *Pachycrocuta*, and *Eucladoceros*), Markova and Kolfschoten (2005) highlight a discrepancy with the biochronological framework proposed by Kahlke (2001), because of the occurrence at Untermassfeld of voles very similar to those recorded at Podumci 1 (Croatia), close to Matuyama–Brunhes boundary (Markova and Kolfschoten (2005), and references therein). For this reason these authors suggest a younger age for the faunal assemblage of Untermassfeld.

The Early Middle Pleistocene

The first occurrence in Europe of the steppe mammoth *Mammuthus trogontherii* and of the red deer *Cervus elaphus acoronatus* can be considered as bio-events characterizing the oldest early middle Pleistocene faunas (middle and late Galerian). The typical Galerian fauna spread over Europe giving rise to new adaptations, in particular among the ungulates. Moreover, at the end of this period, the climatic deterioration referable to the Elsterian (Anglian) glacial stage provided the ecological basis for the appearance of the cold-stage faunal grouping generally known as the ‘*Mammuthus–Coelodonta* faunal complex,’ which characterized the late middle and late Pleistocene of the Northern Hemisphere (Kahlke, 1999).

Among bovids, different large-bodied taxa became a very important element of the middle Galerian faunas such as *Bison schoetensacki*, and the ovibovine *Praeovibos* and *Ovibos*. In late Galerian faunas the aurochs *Bos primigenius* appeared and, together with the water buffalo *Bubalus murrensis*, spread into Northern Europe during temperate climatic stages.

An interesting bubaline bovid is ‘*Bos*’ *galerianus*, known only by a fragmentary skull from Cava di Breccia di Casal Selce (Ponte Galeria area, Rome, Italy; Figures 3(a) and (b) (Petronio and Sardella, 1998), which has been recently referred to the genus *Hemibos* (Martinez Navarro and Palombo,

2004). Even if the taxonomic position of this species is still a matter of debate, its biogeographical significance is important, probably indicating a dispersal from the Levant.

Cervids are quite common and provide detailed data that are very useful for biochronological purposes. The middle and late Galerian are represented by *P. verticornis*, replaced by *Praemegaceros solilhacus* (Figure 4) at approximately 0.6 Ma.

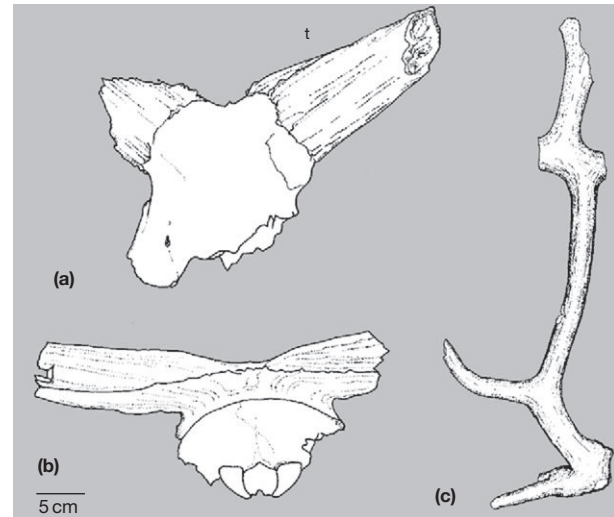


Figure 3 Fossil bones recorded in the Ponte Galeria area (Roma, Italy): (A) ‘*Bos*’ *galerianus* skull frontal view; (B) ‘*Bos*’ *galerianus* skull occipital view; (C) *Praemegaceros verticornis* antler. Drawings by R. Sardella.

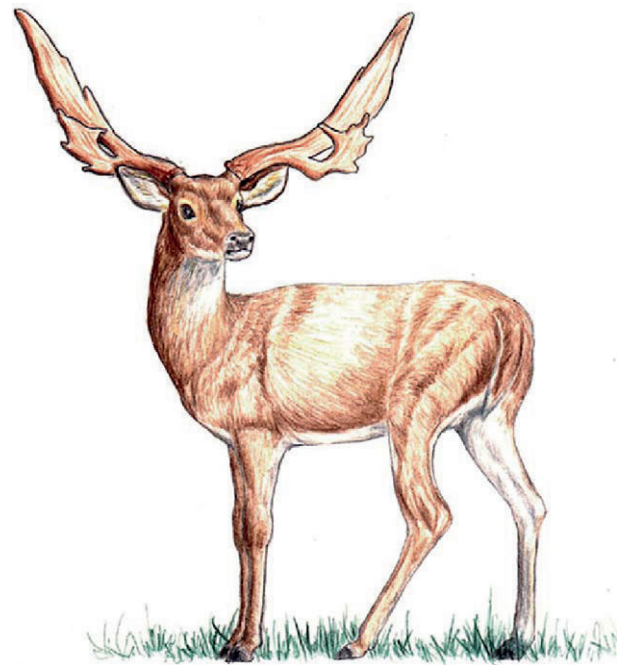


Figure 4 Reconstruction of *Praemegaceros solilhacus*. Drawing by G. Di Stefano, modified from Di Stefano, G. Petronio, C. and Sardella, R. 2005 – Large bovid and cervids from latest Villafranchian–Galerian faunas of Italy. QUATERNAIRE (*International Journal of the French Quaternary Association*) Hors-serie-n. 2: 95–102.

The endemic British megacerine *Praemegaceros* (= *Megaloceros*) *dawkinsi* is closely related to *P. solilhacus*. Another megacerine lineage is represented by *Megaloceros savini*.

The red deer *Cervus elaphus* became quite common during the Galerian and is represented by different subspecies that are considered to be of great biochronological significance: *Cervus elaphus acoronatus* and *Cervus elaphus priscus* (found in the middle Galerian), and *Cervus elaphus eostephanoceros* (found in the late Galerian) (Figure 5) (Di Stefano and Petronio, 1993). In addition, around 0.6 Ma, the large-bodied fallow deer *Dama clactoniana* appears, while the Villafranchian cervid *Axis* (= *Pseudodama*) *eurygonos* survives until the end of early middle Pleistocene (Di Stefano and Petronio (2000–2002), and references therein).

Rhinos are represented by three species: *Stephanorhinus hundsheimensis*, *Stephanorhinus hemitoechus*, and *Stephanorhinus kirchbergensis*; amongst the equids *Equus altidens* and the large-sized *Equus suessenbornensis* occur. They represent different lineages, better adapted to highly seasonal and even very cold climates (Azzaroli et al., 1988). *M. trogontherii* and the straight-tusked elephant *Elephas* (*Palaeoloxodon*) *antiquus* characterize different faunal assemblages of the early middle Pleistocene, surviving until the end of the middle Pleistocene and the end of the late Pleistocene respectively.

Among carnivores the giant short-faced hyaena *P. brevirostris* become extinct at the beginning of middle Pleistocene and

was ecologically replaced by the spotted hyaena *Crocota crocuta praesplaea*, an African immigrant. Lion and leopard are widespread and the large felid *Panthera gombaszoegensis* become extinct during this period. The cave bear lineage is represented by *Ursus deningeri*, but the Himalayan black bear *Ursus thibetanus* and the brown bear *Ursus arctos* are recorded in several localities (Palombo and Valli, 2004). Canids are represented by the middle-sized wolf *Canis mosbachensis*, the hunting dog *L. lycaonoides*, and the dhole *Cuon priscus*.

The evolutionary changes in voles of the genus *Microtus* are stratigraphically important in Western and Southern Europe; furthermore in Central and in Eastern Europe as well as in Asia, the evolution of the lagurids has also yielded additional information about the biochronology of faunal assemblages (Markova and Kolfshoten, 2005).

Middle Galerian mammal faunas are quite well represented in Europe. In Italy, the Slivia local fauna (Trieste) is considered to be representative of the earliest middle Galerian (Slivia Faunal Unit – FU, *sensu* Gliozzi et al. (1997)). From the Slivia karst infill the first occurrences of species that were to spread out widely in the Italian peninsula during the middle Pleistocene are recorded (i.e., *B. schoetensacki* and *S. hemitoechus*) together with the last occurrence of *P. brevirostris* and the water vole *Mimomys savini*.

The most complete middle Pleistocene stratigraphical succession crops out in the Ponte Galeria area (Rome), where at least three different faunal assemblages have been discovered. The earliest faunal assemblage comes from Fontignano, where the ‘cold’ rodents *Prolagus pannonicus* and *Predicrostonyx* sp. have been found (Petronio and Sardella (1999), and references therein; Milli and Palombo (2005)). Petronio and Sardella (1999) defined the Ponte Galeria FU for the faunal assemblage recorded in the area giving its name to the Galerian Mammal Age. This faunal assemblage is dated at ca. 0.75 Ma and includes *M. trogontherii*, *M. savini*, *P. verticornis* (Figure 3(c)), *A. eurygonos*, ‘*B. galerianus*’ (Figures 3(a) and (b)), and *Crocota crocuta praesplaea*.

The second faunal assemblage from this area is referable to the Isernia FU. Together with bison, deer, equids, and small mammals (the hamster *Allocricetus bursae* and the water vole *A. mosbachensis*), birds, reptiles, and amphibians have been found. The avifauna includes the partridge *Perdix palaeoperdix* and the vulture *Gyps melitensis*, together with the ‘Villafranchian survivor’ *Palaeocryptonyx* (Bedetti, 2003). Moreover, the fossil record from this site includes avian species prevalently inhabiting lacustrine and marshy areas (i.e., Anatidae), not commonly recorded in the Galerian localities in the rest of Europe, which are dominated by terrestrial birds, such as Galliforms and Corvids (Bedetti, 2003).

The Clactonian fallow deer first occurs in the more recent Isernia FU (ca. 0.6 Ma, Coltorti et al. (2005)), together with *P. solilhacus*. The age of this well-known archaeological site, particularly rich in remains of *B. schoetensacki* and *S. hundsheimensis*, was for many years, matter of debate among researchers. The K/Ar age of 0.736 Ma, provided by Coltorti et al. (1982), was considered too old in comparison to the paleontological evidences (e.g., occurrence of *A. mosbachensis* and *P. solilhacus*) (Kolfshoten and Koenigswald, 1996; Petronio and Sardella, 1999). The new radiometric data provided by Coltorti et al. (2005) confirm the value of the biochronological framework established for Italy. Coeval



Figure 5 Reconstruction of *Cervus elaphus eostephanoceros*. Drawing by G. Di Stefano, modified from Di Stefano, G. Petronio, C. and Sardella, R. 2005 – Large bovid and cervids from latest Villafranchian-Galerian faunas of Italy. QUATERNAIRE (International Journal of the French Quaternary Association) Hors-serie-n. 2: 95–102.

fossiliferous localities from Italy are Venosa, Valdemino, and Visogliano (Glozzi et al., 1997).

In Spain, the middle Galerian is attested by the faunal assemblage of Sierra de Atapuerca, in particular Trinchera Dolina (TD4). In France fossiliferous localities of comparable age are rather poorly represented. These faunal assemblages are referred to biochrons MNQ21 and MNQ22 and are characterized by the first occurrence of *Panthera pardus*, *U. thibetanus*, *Cervus elaphus acoronatus*, and *U. deningeri*. The association of *P. brevirostris* and *Cervus elaphus acoronatus* is particular to this stage (Palombo and Valli, 2004). According to Abbazzi and Lacombat (2005, and references therein) the local fauna of Soleilhac (Haute Loire), previously calibrated with a submagnetochron of Jaramillo and therefore related to the latest early Pleistocene, must be considered younger than supposed before and referable to the Isernia FU because of the co-occurrence of *Cervus elaphus acoronatus*, *D. clactoniana*, and *P. solilhacus*. Other approximately coeval French localities are Lunel Viel and L'Escale (Palombo and Valli, 2004). In France MNQ22 is characterized by the first occurrence of the modern brown bear *U. arctos*, the thar *Hemitragus bonali*, and the rhino *S. hemitoechus*. Finally, MNQ23 is diagnosed by the occurrence of the badger *Meles thoralis spelaeus*, the wolf *C. mosbachensis* (= *C. lunellensis*), and the cave lion *Panthera leo spelaea*.

In Greece the following Galerian faunal assemblages can be listed: Maratousa, Volos, and Petralona (Kahlke (2001), and references therein).

The large mammal fossil record in Central Europe (Germany, The Netherlands, Belgium, Czech Republic, and Austria) presents a general biochronological framework similar or, in some aspects transitional, between Eastern and Western Europe. The detailed studies on the rich micromammalian record provide a biochronological scheme for Central Europe based mainly on rodents and, in particular on the *Mimomys-Arvicola* lineage (Maul et al., 1998; Koenigswald and Heinrich, 1999). One of the most important European early middle Pleistocene mammal fauna is that from the Main Fossiliferous Horizon ('Hauptfundschrift') of Voigstedt (Thuringia, Germany), referred to as the '*Mimomys savini* faunas,' where 18 micromammals have been found. The site is also the type locality of the ground squirrel *Spermophilus dietrichi* and of the flying squirrel *Petaurium voigstedtensis* (Maul (2002), and references therein). Large mammals include typical middle Galerian species such as *M. trogontherii*, *P. verticornis*, and *B. schoetensacki*. In Thuringia other important fossiliferous localities are Süssenborn, Rastenberg, and Bad Frankenhausen, yielding faunal assemblages from a cold stage earlier than the Elsterian glaciation, as testified by the occurrence of cold climate-adapted elements such as the ovibovine *Soergelia elisabethae* (Süssenborn and Rastenberg), the reindeer *Rangifer tarandus*, and the ovibovines *Praeovibos priscus* and *Ovibos moschatus* (Bad Frankenhausen). The Bad Frankenhausen faunal assemblage represents the first Central Europe occurrence of the *Coelodonta-Mammuthus* complex (Kahlke (1999), and references therein).

Other important Galerian fossil assemblages coming from Central and Eastern Europe include: Karlich C, D, and E, Mauer, Mosbach 2 (Germany), Kozi Grzbiet (Poland), Hundsheim (Austria), Stranska Skala (Moravia, Czech Republic), Koneprusy C718 and Přezletice (Bohemia, Czech

Republic), Gombasek 1 (Slovakia), Kövesvár and Vértesszőlős 1 and 2 (Hungary), Betfia VII and Feldioara Cariera (Romania), and Tiraspol (Moldova) (Kahlke (2001), and references therein).

The younger 'early *Arvicola mosbachensis* (= *A. cantiana*) faunas' (sensu Koenigswald and Heinrich (1999)) are characterized by the first occurrence in Central Europe of *E. (Palaeoloxodon) antiquus* and *A. mosbachensis*, and the widespread dispersal of *Homo* at higher latitudes. This 'faunal zone' includes cold and warm stages. *E. (Palaeoloxodon) antiquus* seems to leave Central Europe during colder phases, while this species, together with *Hippopotamus amphibius*, becomes quite common during warm stages, even in the Rhine valley. This biochronological unit can be represented by the faunas from Mauer and Mosbach 2 in Germany (Koenigswald and Heinrich, 1999).

Steppe adapted taxa, together with northern Palearctic tundra elements, such as reindeer, spread as far as the Pyrenees and characterize the late Galerian faunas recorded at La Caune de l'Arago (Tautavel, France). In this locality *Rangifer* has been found together with *Vulpes* cf. *praeglacialis*, *Equus mosbachensis*, *Bison priscus*, *P. priscus*, *Ovis ammon*, and many other taxa (Kahlke, 1999; Palombo and Valli, 2004). This biochron can be compared to the Italian Fontana Ranuccio FU, although the cold climate-adapted taxa are missing from the latter.

The Brunhes-Matuyama paleomagnetic boundary has not yet been established in Britain, although the mammalian faunal record from the middle Pleistocene onward is very detailed. In Britain six early middle Pleistocene assemblages of temperate-climate affinity have been recognised (Stuart et al., 2004), all of which predate the Anglian glaciation (MIS 12), a major stratigraphical marker for NW Europe. However, more detailed correlations with the marine oxygen-isotope record are not yet well established.

Over almost two centuries the exposures of the Cromer Forest Bed Formation on the Norfolk and Suffolk coasts have provided the most important fossil record for the early middle Pleistocene of Britain. One of the richest localities is West Runton with a large number of different taxa recorded (45 mammals, 17 birds, 3 reptiles, 5 amphibians, and 7 fishes) (Stuart, 1996).

The presence at Pakefield (Suffolk) of the vole *Mimomys pusillus*, a species that disappears from the continental record after MIS 16, may indicate that this site alone predates MIS 16 whereas all others, including the typesite of the Cromerian Complex at West Runton (Norfolk) belong to the latest part of the early middle Pleistocene (MIS 15 and 13 inclusive). The presence at Pakefield of *H. amphibius*, *P. dawkinsi*, *E. (Palaeoloxodon) antiquus*, and *E. altidens* provides evidence for 'Mediterranean' climates that are substantially warmer than southern Britain at the present day, as well as evidence for the oldest human occupation in the country (Stuart et al., 2004; Parfitt et al., 2005).

Other important localities are Sidestrand, Sugworth (Berkshire), and Little Oakley (Essex) characterized by the occurrence of *M. savini*. In contrast, Ostend (Norfolk), Boxgrove (West Sussex), and Westbury-sub-Mendip (Somerset) which contain *A. mosbachensis* and are thus referred to a younger 'Cromerian Complex' interglacial that postdates the Cromerian stratotype at West Runton (Stuart, 1996).

The Late Middle Pleistocene

In the late middle Pleistocene almost all the Villafranchian elements of the mammal faunas became extinct and the distribution of the European terrestrial vertebrates was strongly affected by climatic oscillations. The ‘*Elephas antiquus*–*Stephanorhinus kirchbergensis* faunal complex’ spread into Northern latitudes across Europe during temperate stages; from the Saalian cold stages onward the diffusion of the ‘*Mammuthus*–*Coelodonta* faunal complex’ took place in the Mediterranean region, with more advanced faunal elements (e. g., large-sized true wolf) if compared to the early middle Pleistocene faunas. Such a complex characterizes the glacial late Pleistocene times (Kahlke (1999), and references therein).

A useful tool for establishing long distance correlations and creating an accurate biochronological framework for the late middle Pleistocene vertebrate faunas comes from the detailed analysis of fluvial terrace sequences yielding different faunal assemblages, for example, the river Thames in Britain or the Tiber in Italy.

The Aurelian faunas include a large number of modern taxa, or closely related species, differing from extant associations in the occurrence of large mammals (horses, rhinos, hippos, elephants, wolves, hyaenas, lions, and leopards). The name Aurelian derives from the Via Aurelia, which crosses the area NE of Rome, where a great number of fossiliferous localities have been discovered. In Italy the beginning of the Aurelian Mammal Age is indicated by the first occurrence of the giant deer *Megaloceros giganteus*, *U. spelaeus*, and *C. lupus* and is correlated with MIS 9, ca. 0.3 Ma (Gliozzi et al., 1997). Early Aurelian faunas (MIS 9–8) are included in the Torre in Pietra FU (early Aurelian). In the time span represented by these faunal assemblages *E. (Palaeoloxodon) antiquus* and *B. primigenius* are particularly abundant and widespread in Italy (Milli and Palombo, 2005). Faunal assemblages dated to the interval between MIS 7 and 6 are included by Gliozzi et al. (1997) in the Vitinia FU (middle Aurelian). The appearance of the fallow deer *Dama dama tiberina*, *Equus hydruntinus*, and *Mammuthus* specimens showing advanced skull features have been regarded as peculiar to the Vitinia FU. However, Milli and Palombo (2005) remark that differences in composition of the faunal complexes ascribed to Torre in Pietra and Vitinia FUs are minimal, both from the taxonomical and the structural point of view, thus both of them could be referred to MIS 9.

In France the early Aurelian can be related to the biochron MNQ24, characterized by the first occurrence of *Saiga tatarica*, *U. spelaeus*, *Capra ibex*, and *Coelodonta antiquitatis*.

Rich vertebrate fossil assemblages are recorded in cave deposits with archaeological layers in Southern France (Lazaret and Terra Amata), in sequences referred to MIS 7–6 (Palombo and Valli, 2004).

The sequence for the late middle Pleistocene in Britain is now very comprehensively described and consistent faunal differences have been identified between the various temperate episodes of this period (Schreve, 2001).

The British evidence has one particular advantage if compared to the evidence from NW Europe (France, The Netherlands, and Germany), since, although the last interglacial is already defined at oxygen-isotope substage level, the

mammalian evidence has identified much smaller-scale environmental and climatic oscillations within MIS 11 and 7 (and more tentatively within MIS 9), possibly corresponding to isotopic substages. Three post-Anglian interglacials are represented in the record from the lower Thames valley, as typified by the sites of Swanscombe (MIS 11), Purfleet (MIS 9), and Aveley (MIS 7) in descending relative age, each of which has yielded a diagnostic mammalian assemblage (Schreve, 2001).

The deposits at Swanscombe preserve a fluvial sequence which can be considered as the oldest terrace of the Lower Thames and is related to MIS 11. The interglacial represented at Swanscombe also witnesses the first appearance in Britain of four taxa that are not found in the preceding Cromerian Complex: *S. kirchbergensis*, *S. hemitoechus*, *M. giganteus*, and *B. primigenius*. Furthermore, the large fallow deer *D. clactoniana*, the equid *E. hydruntinus*, and the cave bear *U. spelaeus* occur. Among micromammals the small mole *Talpa minor*, the giant beaver *Trogontherium cuvieri*, the rabbit *Oryctolagus cuniculus*, and the European pine vole, *Microtus (Terricola) subterraneus* have been recorded.

The Purfleet MAZ is based mainly on the terrace correlated to MIS 9 exposed at Purfleet, near London. The rich faunal assemblage includes mammals, birds, amphibians, and fishes. Among mammals, *E. antiquus*, *Macaca sylvanus*, and water voles referable to the transitional *A. mosbachensis/A. terrestris* occur; among fishes the presence of cyprinids and of their associate predator, the pike, clearly indicate temperate climatic conditions.

The mammalian assemblages correlated to MIS 7 have been subdivided into two biochrons, the Ponds Farm MAZ and Sandy Lane MAZ. Both assemblages are present at Aveley (Essex). The Sandy Lane MAZ has a distinct composition, with the coexistence of a particular late morphotype of *M. trogontherii* with *E. antiquus*, *Equus ferus*, and *S. kirchbergensis*.

The comparison of the terrace sequence of Thames with the Wipper River terrace system (Thuringia, Germany) reveals clear affinities. Bilzingsleben II, a rich paleontological and archaeological deposit, has been correlated to MIS 11 and to the fauna compared with the Swanscombe MAZ (Schreve and Bridgland, 2002).

In Central Europe the ‘younger *Arvicola mosbachensis* (= *A. cantiana*) faunas’ (sensu Koenigswald and Heinrich (1999)) reflect a period of strong climatic oscillations and a decrease of biodiversity, followed by a temperate climatic phase, when *B. primigenius* and *B. murrensis* dispersed into Central Europe. Finally, the ‘*Arvicola mosbachensis/terrestris* faunas’ also include *Apodemus maastrichtensis* representing the final part of middle Pleistocene, before the Eemian.

See also: Vertebrate Overview. **Vertebrate Records: Early Pleistocene.**

References

- Abbazzi L and Lacombe F (2005) *Dama clactoniana*. In: Lacombe F (ed.) *Les Grands Mammifères Fossiles du Velay*, p. 122. Annales des Amis du musée Crozatier 13/14 (2004–2005).

- Alberdi MT, Azanza B, Cerdeño E, and Prado JL (1997) Similarity relationship between mammal faunas and biochronology from latest Miocene to Pleistocene in the western Mediterranean area. *Eglogae geologicae Helvetiae* 90: 115–132.
- Ambrosetti P, Azzaroli A, Bonadonna FP, and Follieri M (1972) A scheme of Pleistocene chronology for the Tyrrhenian side of central Italy. *Bollettino della Società Geologica Italiana* 91: 169–184.
- Azzaroli A, De Giulio C, Ficarelli G, and Torre D (1988) Late Pliocene to Early Mid-Pleistocene mammals in Eurasia: Faunal succession and dispersal events. *Palaeogeography, Palaeoecology, Palaeoclimatology* 66: 77–100.
- Bedetti C (2003) *Le Avifaune Fossili del Plio-Pleistocene Italiano: Sistematica, Paleoeologia ed Elementi di Biocronologia*. PhD Thesis. pp. 1–188. Modena: Bologna Roma 'La Sapienza' Universities.
- Bermúdez De Castro JM, Martínón-Torres M, Carbonell E, et al. (2004) The Atapuerca sites and their contribution to the knowledge of human evolution in Europe. *Evolutionary Anthropology* 13: 25–41.
- Caloi L and Palombo MR (1996) Latest Early Pleistocene mammal faunas of Italy, biochronological problems. *Il Quaternario* 8(2): 391–402.
- Coltorti M, Cremaschi M, Delitala MC, et al. (1982) Reversed magnetic polarity at Isernia La Pineta: A new lower Paleolithic site in Central Italy. *Nature* 300: 173–176.
- Coltorti M, Feraud G, Marzoli A, et al. (2005) New $^{40}\text{Ar}/^{39}\text{Ar}$, stratigraphic and palaeoclimatic data on the Isernia La Pineta Lower Palaeolithic site, Molise, Italy. *Quaternary International* 131: 11–22.
- Di Stefano G and Petronio C (1993) New *Cervus elaphus* subspecies of Middle Pleistocene age. *Neues Jahrbuch für Geologie und Paläontologie* 203: 57–75.
- Di Stefano G and Petronio C (2000–2002) Systematics and evolution of the Eurasian Plio-Pleistocene tribe Cervini (Artiodactyla, Mammalia). *Geologica Romana* 36: 311–334.
- Gliozzi E, Abbazzi L, Argenti P, et al. (1997) Biochronology of selected Mammals, Molluscs and Ostracods from the Middle Pliocene to the Late Pleistocene in Italy. The state of art. *Rivista Italiana di Paleontologia e Stratigrafia* 103: 369–388.
- Guerin C (1990) Biozones or mammal units? Methods and limits in Biochronology. In: Lindsay EH, Fahlbusch V, and Mein P (eds.) *European Neogene Mammal Chronology*, pp. 119–130. New York: Plenum.
- Kahlke R-D (1999) The History of the Origin, Evolution and Dispersal of the Late Pleistocene. *Mammuthus-coelodonta Faunal Complex in Eurasia (Large Mammals)*, pp. 1–219. Mammoth Site of Hot Springs.
- Kahlke R-D (2001) Die Unterpleistozäne komplexfundstelle Untermassfeld – zusammenfassung des kenntnisstandes sowie synthetische betrachtungen zu genesemodell, paläoökologie und stratigraphie. In: Kahlke RD (ed.) *Das Pleistozän von Untermassfeld bei Meiningen (Thüringen)*, pp. 931–1030. Bonn: Dr. Rudolf Habelt GMBH.
- Koenigswald W and Kolfschoten T (1996) The *Mimomys-Arvicola* and the enamel thickness quotient (SDQ) of *Arvicola* as a stratigraphic markers in the Middle Pleistocene. In: Turner C (ed.) *The Early Middle Pleistocene in Europe*, pp. 211–226. Rotterdam: Balkema.
- Koenigswald W and Heinrich W-D (1999) Mittelpleistozäne Säugetierfaunen aus Mitteleuropa – der Versuch einer biostratigraphischen Zuordnung. *Kaupia* 9: 53–112.
- Lumley Hde, Kahlke R-D, Moigne AM, and Moule O-E (1988) Les faunes des grands mammifères de la Grotte du Vallonet (Roquebrune-Cap-Martin, Alpes Maritimes). *L'Anthropologie* 92(2): 465–496.
- Markova AK and Kolfschoten T (2005) Response of the European mammalian fauna to the mid-Pleistocene transition. In: Head MJ and Gibbard PL (eds.) *Special Publications 247: Early-Middle Pleistocene Transitions: The Land–Ocean Evidence*, pp. 221–229. Geological Society.
- Martínez-Navarro B and Palombo MR (2004) Occurrence of the Indian genus *Hemibos* (Bovini, Bovidae, Mammalia) at the Early–Middle Pleistocene transition in Italy. *Quaternary Research* 61: 314–317.
- Maul LC (2002) The Quaternary small mammal faunas from Thuringia (Central Germany). In: Meyrich RA and Schreve DC (eds.) *The Quaternary of Central Germany (Thuringia and surroundings): Field Guide*, pp. 79–95. London: Quaternary Research Association.
- Maul LC, Masini F, Abbazzi L, and Turner A (1998) The use of different morphometric data for absolute age calibration of some South- and Middle European arvicolid populations. *Palaeontographia Italica* 85: 111–151.
- Milli S and Palombo MR (2005) The high-resolution sequence stratigraphy and the mammal fossil record: A test in the Middle–Upper Pleistocene deposits of the Roman Basin (Latium, Italy). *Quaternary International* 126–128: 251–270.
- Parfitt SA, Barendregt RW, Breda M, et al. (2005) The earliest record of human activity in northern Europe. *Nature* 438: 1008–1012.
- Petronio C and Sardella R (1998) *Bos galerianus* n. sp. (Bovidae, Mammalia) from Ponte Galeria formation (Rome, Italy). *Neues Jahrbuch für Geologie und Paläontologie, Abhandlungen* 1998(5): 289–300.
- Petronio C and Sardella R (1999) Biochronology of the Pleistocene mammal fauna from Ponte Galeria (Rome) and remarks on the middle Galerian faunas. *Rivista Italiana di Paleontologia e Stratigrafia* 105(1): 155–164.
- Schreve DC (2001) Differentiation of the British Late Middle Pleistocene interglacials: The evidence from mammalian biostratigraphy. *Quaternary Science Reviews* 20: 1693–1705.
- Schreve DC and Bridgland DR (2002) The Middle Pleistocene hominid site of Bilzingsleben. In: Meyrick RA and Schreve DC (eds.) *The Quaternary of Central Germany (Thuringia and surroundings): Field Guide*, pp. 131–144. London: Quaternary Research Association.
- Stuart AJ (1996) Vertebrate faunas from the early Middle Pleistocene of East Anglia. In: Turner C (ed.) *The Early Middle Pleistocene in Europe*, pp. 9–24. Rotterdam: Balkema.
- Stuart AJ, Parfitt SA, and Breda M (2004) An early Middle Pleistocene mammalian fauna from Pakefield-Kessingland, Suffolk. In: Schreve DC (ed.) *The Quaternary Mammals of Southern and Eastern England Germany: Field Guide*, pp. 19–27. London: Quaternary Research Association.

Mid-Pleistocene of North America

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Introduction

Pleistocene vertebrate faunas from North America are abundant, taxonomically diverse, and widespread across much of the continent. These faunas thus provide a reasonably detailed and robust record of faunal change during an interval that witnessed numerous advances and retreats of glacial ice. The terms ‘early Pleistocene’ and ‘middle Pleistocene’ are not formally defined in North America, nor is their sporadic use necessarily consistent with informally designated and recognized units on other continents. The term ‘late Pleistocene’ in North America also is not formally defined, but is commonly used for faunas that fall within the temporal resolution of radiocarbon dating techniques.

Several extensive stratigraphic sequences deposited over the last few million years, combined with numerous faunas associated with reliable external age control, make the Pliocene and Pleistocene vertebrate faunal record of North America among the best in the world. These sequences are situated in the midwestern and western portions of the continent, including the Anza-Borrego Desert and San Timoteo Badlands of southern California, the Ringold Formation in southeastern Washington State, the San Pedro and Gila River valleys in Arizona and New Mexico, The Snake River Plain in Idaho, and the Meade Basin in Kansas.

External age control for these faunas (i.e., age determinations independent of the faunas themselves) are derived from both paleomagnetic analyses of the sediments associated with the faunas, and from radio-isotopic dating techniques. In the absence of these external controls, faunas are often placed in an approximate chronological sequence on the basis of biochronological inference (relying on taxonomic similarity with externally-dated localities in nearby regions). By the mid-1960s, four magnetic polarity chrons were established for the time interval considered here. From oldest to youngest they are the Gilbert, Gauss, Matuyama, and Brunhes. These names, and the subchron designations encompassed by them, are still widely used in the literature for North American faunas. Alternative, numerical designations for magnetic chrons from the Mesozoic through Miocene were recently extended through to the present (Opdyke and Channell, 1996), and these numeric designations are being used more frequently in modern literature.

The mammals are by far the best-studied group of vertebrates from the early and middle Pleistocene of North America. They constitute a taxonomically diverse assemblage, are abundant in many faunas within this time range, are found in fossilized deposits across most of the continent, and are widely used as tools to help establish age relationships among faunas in this interval. Therefore, it is appropriate to begin this summary discussion with a review of the significant mammalian record.

North American Land Mammal Ages

Most vertebrate paleontologists in North America utilize a formal nomenclature centered on North American land mammal ages originally proposed by Wood et al. (1941) and Savage (1951). There are three land mammal ages relevant for this discussion; these are (from oldest to youngest) the Blancan, Irvingtonian, and Rancholabrean. These are faunal terms that are based on, and differentiated by, the taxonomic composition and stage of evolution of the mammalian fauna in North America. They are often (erroneously) used as operational synonyms with the traditional Lyellian epochs, or are otherwise confused with formal and informal time-stratigraphic units. The history, development, structure, operational framework, and current understanding of these mammal ages were addressed in detail by various authors in two useful compendium volumes (Woodburne, 1987, 2004). Traditionally, these mammal ages were conceptualized as being applicable across the entire North American continent, but a recent formal proposal restricting their latitudinal scope to regions south of 55°N was based on an emerging consensus that northern latitude faunas in Canada and Alaska are sufficiently distinct during the Pliocene and Pleistocene to merit independent biochronologies (Bell et al., 2004b). The same is likely also true of faunas in the southern latitudes of central America; presumably a boundary line would be drawn somewhere in Mexico, but there are insufficient data available at this time to warrant any formal proposal for restricting Mexican faunas from the traditional North American mammal ages.

The relatively recent formal recognition of boundaries between the classic late Cenozoic epochs, and the accompanying designation of stratotypes for those boundaries, has led to a refined assessment of the correspondence between mammal age boundaries and epochal boundaries. The base of the Pliocene Series is now interpreted to be 5.335 Ma. The base of the Pleistocene series is now approximately 1.77 Ma. The Holocene is not formally defined, but in North America it is operationally recognized as beginning at the conveniently round number of 10,000 years ago. As currently defined and understood (see below), the Blancan mammal age begins between 5.2 to 4.6 Ma and extends to the base of the Irvingtonian, at approximately 1.35 Ma. Therefore, most of the Blancan is within the Pliocene Epoch but the youngest Blancan faunas (those between 1.77 Ma and 1.35 Ma) are earliest Pleistocene in age. An informal designation of those latter faunas as ‘latest Blancan’ was recently proposed by Bell et al. (2004b). As currently defined and understood (see below), the Irvingtonian begins at approximately 1.35 Ma, and extends to the poorly-constrained beginning of the Rancholabrean, some time between 210 ka and 160 ka. The Irvingtonian is entirely within the Pleistocene. The Rancholabrean is considered to

end with extinction of the large-bodied megafauna in North America, south of latitude 55°N. This extinction event (see [Late Pleistocene Megafaunal Extinctions](#)) was time-transgressive, but its termination is currently interpreted to be at approximately 9,500 years ago. Most of the Rancholabrean is, therefore, Pleistocene in age, extending only a few hundred years into the Holocene. The Rancholabrean is discussed elsewhere in this volume (see [Late Pleistocene of North America](#)).

Blancan

The Blancan is defined by the earliest appearance in North America, south of latitude 55°N, of the arvicoline rodents *Mimomys*, *Ogmodontomys*, and *Ophiomys*. These are extinct vole-like rodents with readily recognizable teeth, being fairly common in Blancan faunas throughout much of North America. *Ogmodontomys* and *Ophiomys* are among many taxa that are found only in Blancan faunas; others include the ground sloths *Megalonyx leptostomus* and *Glossotherium chapadmalense*, the rabbits *Aluralagus*, *Sylvilagus webbi*, *Pewelagus dawsonae*, and *Pratilepus kansasensis*, the beavers *Dipoides rexroadensis* and *Procastoroides*, the volelike rodents *Guildayomys*, *Hibbardomys*, *Nebraskomys*, and *Pliophenacomys*, the lemmings *Mictomys vetus* and *Pliolemmus*, the muskrats *Pliopotamys* and *Ondatra idahoensis*, the cotton rats *Sigmodon curtisi* and *S. minor*, the dogs *Borophagus diversidens* and *Canis lepophagus*, the bear *Ursus abstrusus*, the peccaries *Platygonus bicalcaratus* and *P. pearcei*, and the horse *Nannipus peninsulatus*. All of these taxa are extinct, but living representatives of *Sylvilagus*, *Mictomys*, *Sigmodon*, *Canis*, and *Ursus* are still found in North America. Taxa that first appear in the earlier parts of the Blancan but persist into the Irvingtonian or even younger faunas include the giant xenarthran *Glyptotherium*, shrews of the genus *Blarina*, the rabbit genus *Sylvilagus*, the American cheetah *Miracinonyx inexpectatus*, the saber-tooth cat *Smilodon gracilis*, and the elephant *Stegomastodon*. With the exceptions of *Blarina* and *Sylvilagus*, these taxa now are extinct.

The recently proposed definition of the Irvingtonian ([Bell et al., 2004b](#)) resulted in a somewhat awkward situation in which faunas that were for many decades considered to be earliest Irvingtonian are now considered to be latest Blancan. That new definition is not universally accepted, and other definitions are possible ([Bell et al., 2004b](#)) and are used operationally for some important faunas ([Martin et al., 2003](#)). In recognition of this, a distinction is made for taxa that first appear in these latest Blancan faunas and persist into younger faunas (see [Early Pleistocene](#)). These include the ground sloths *Nothrotherium* and *Paramylodon harlani*, the giant armadillo *Holmesina*, the jackrabbit *Lepus*, the voles *Allophaiomys*, *Microtus*, and *Phenacomys*, the lemmings *Synaptomys*, *Mictomys kansasensis* and *M. meltoni*, the muskrats *Neofiber* and *Ondatra annectens*, the tree squirrel *Sciurus*, the scimitar cat *Homotherium*, the peccary *Platygonus vetus*, and the mastodon *Mammut americanum*. Most of the listed rodent and rabbit genera still have living representatives in the North American fauna.

The relatively long time frame of this mammal age is reflected by the diversity of mammals represented in Blancan

faunas, and the readily apparent changes in the composition of these faunas throughout the mammal age. These changes permit the recognition of shorter intervals of time within the Blancan, and thus a more refined mammalian biochronology. Several authors over the last four decades proposed various formal designations to establish such refinement. The simplest of these proposals entailed a twofold division of the Blancan, with the older faunas united under the term Rexroadian, and the younger ones under the name Senecan ([Schultz et al., 1978](#)). That two-fold system was based on faunas from the Great Plains and the faunal sequence preserved in outcrops along the Snake River Plain. The terms are most commonly applied to faunas in those areas, and there has been minimal effort to expand the conceptual framework to other areas on the continent.

Two other efforts to divide the Blancan were based on changes in the arvicoline rodent fauna through the Blancan. Following their earliest appearance in the fossil record nearly five million years ago, arvicoline rodents rapidly diversified and expanded their geographical range across most of the Holarctic. They also appear to be characterized by a rapid rate of evolutionary change. They are widely utilized for stratigraphic correlation throughout the Holarctic. Arvicoline divisions of the Blancan were discussed extensively by [Martin \(1979\)](#), [Repenning \(1987\)](#), [Repenning et al. \(1990\)](#), and [Bell \(2000a, b\)](#). In current form there are three recognized arvicoline divisions of the Blancan. The first of these spans approximately 4.9 to 4.1 Ma and is defined by the first appearance of *Ogmodontomys* and *Ophiomys*. The second spans approximately 4.1–2.5 Ma and is defined by the first appearance of muskrats in North American faunas. The third interval begins at approximately 2.5 Ma and is defined by the earliest appearance of bog lemmings in North America; it ends with the beginning of the first Irvingtonian arvicoline division (discussed below).

Blancan faunas are found throughout much of the continental United States, but the most important localities are concentrated in the western states, Florida, and the Great Plains regions of Nebraska, Kansas, and northern Texas. Temporally equivalent faunas are also known from northern Alaska, and Michoacan, Guanajuato, Aguascalientes, and Baja California Sur in Mexico. These faunas were reviewed recently by [Bell et al. \(2004b\)](#).

Irvingtonian

There is currently no consensus on an appropriate definition for the Irvingtonian. A committee of Quaternary mammalian paleontologists in North America recently completed a paper in which they proposed to formally and explicitly define the Irvingtonian based on the earliest appearance of mammoths of the genus *Mammuthus* in North America south of latitude 55°N latitude ([Bell et al., 2004b](#)) (see [Early Pleistocene](#)). This definition is concordant with an informal proposal suggested by [Kurten and Anderson \(1980\)](#) in their major summary volume of North American Pleistocene faunas. This became the operational use that was adopted by many mammalian paleontologists in the last several decades, and also has the advantage of

defining the mammal age based on a large, recognizable, and geographically widespread taxon. Unfortunately, *Mammuthus* is absent (or rare) from some of the most important faunas that fall within the Irvingtonian interval. These include the Anza-Borrego Desert sequence in southern California (where it is rare), and the important Meade Basin faunas in Kansas (where it is conspicuously absent). Our current understanding of the time of arrival of *Mammuthus* in North America is complicated by poor external age control at many sites (reviewed by Bell et al., 2004b) that are candidates for earliest, or at least very early, appearance. Based on well-dated faunas that have unambiguous association of *Mammuthus* remains with reliable external age control, *Mammuthus* appears to have arrived and dispersed rapidly across much of the continental United States between approximately 1.3 and 1.4 Ma. Therefore, a date of 1.35 Ma for the beginning of the Irvingtonian seems acceptable.

Alternative proposals to define the Irvingtonian on the basis of various taxa of arvicoline rodents were also reviewed by Bell et al. (2004b), and may be preferable to the *Mammuthus* definition discussed above. These operational definitions are complicated by perceived faunal provinciality, resulting in provincial definitions that are based on first appearance of different taxa in different geographic areas. One consequence of this is that the Irvingtonian could be conceptualized as beginning at different times in different parts of the continent. A time-transgressive, multi-taxon definition system would complicate communication if nothing else, but a small mammal definition is unquestionably useful for the Great Plains faunas, and is used there in preference to *Mammuthus* (Martin et al., 2003). The small mammal definition now used in the Great Plains is based on the earliest appearance of the vole *Allophaiomys pliocaenicus* (also sometimes named *Microtus pliocaenicus* (see Early Pleistocene)). The temporal differences associated with alternative definitions are significant; for example, using the *Allophaiomys* definition results in a starting time of the Irvingtonian at approximately 1.9 Ma (Martin et al., 2003; Bell et al., 2004b), instead of 1.35 Ma.

Using the *Mammuthus* definition, there are only a few taxa known to be restricted to the Irvingtonian. These include the voles *Microtus llanensis*, *M. meadensis*, and *M. paroperarius*, the dog *Canis armbrusteri*, and the four-horned antelope *Tetrameryx irvingtonensis*. All of these taxa are now extinct, but there are many extant species of *Microtus* in North America. Taxa that first appear in Irvingtonian faunas and persist into the younger Rancholabrean or Holocene faunas include the opossum *Didelphis*, the pygmy rabbit *Brachylagus idahoensis*, red-backed voles of the genus *Clethrionomys*, the sagebrush vole *Lemmiscus curtatus*, the living species of muskrat *Ondatra zibethicus*, the marmots or groundhogs *Marmota flaviventris* and *M. monax*, prairie dogs *Cynomys gunnisoni* and *C. ludovicianus*, the jaguar *Panthera onca*, the saber-toothed cat *Smilodon populator*, the ermine *Mustela erminea*, the extinct skunk *Brachyprotoma*, the hog-nosed skunks *Conepatus*, the coyote and wolf *Canis latrans* and *C. lupus*, the extinct short-faced bear *Arctodus simus*, the shrub ox *Eucatherium*, mountain goats of the genus *Oreamnos*, and mammoths in the genus *Mammuthus*.

The Irvingtonian also can be divided into shorter time intervals based on the recognition of mammalian faunal change within the mammal age. The outline of a three-fold division was proposed by Schultz et al. (1978) and was subsequently

modified and expanded by Lundelius et al. (1987). It was centered on changes detected in the extensive faunal sequence preserved in the Great Plains region of the central United States. The Sappan interval spans the latest Blancan and early Irvingtonian as they are defined here. The Cudahyan was erected in recognition of transitional faunas between the earlier Sappan and later Sheridanian. Sheridanian faunas are found within normally magnetized sediments of the Brunhes Chron (Chron 1n). An alternative division is based on arvicoline rodents and was originally proposed in the late 1970s and developed extensively since that time. Traditionally, this system recognized three divisions of the Irvingtonian, numbered Irvingtonian I, II, and III, from oldest to youngest. The history of the development of this system is somewhat complex, and was reviewed extensively by Repenning (1987), Repenning et al. (1990), Bell (2000a), and Bell et al. (2004b). In the most recent revisions, new data on temporal distribution and taxonomic affinity of relevant fossils resulted in proposals that the Irvingtonian III be dropped as an undefined division. The Irvingtonian I is recognized based on the first appearance of different taxa in different regions. In the southern United States west of the Rocky Mountains, it is marked by the earliest appearance of the vole *Microtus* (Repenning, 1992) at approximately 1.4 to 1.6 Ma. In the northern United States west of the Rocky Mountains, it is marked by the earliest appearance of the heather vole *Phenacomys* at approximately 1.72 Ma. East of the Rocky Mountains, it is marked by the earliest appearance of *Allophaiomys pliocaenicus* at approximately 1.9 Ma. The Irvingtonian II is defined based on the earliest appearance, at approximately 0.85 Ma, of the voles *Microtus meadensis*, *Lasiopodomys*, and *Clethrionomys*. The Irvingtonian II extends up to the Rancholabrean, defined by the arrival of *Bison* south of 55°N latitude.

Irvingtonian mammalian faunas are distributed across much of the continental United States. Unlike the Blancan, many important Irvingtonian faunas were recovered from cave deposits. Our only records of faunal dynamics in the Appalachian region during the Irvingtonian are derived from cave deposits in Pennsylvania, Maryland, and West Virginia. The Conard Fissure deposit in Arkansas provides our only data for the Ozark region of the United States (Brown, 1908). One of the most important and taxonomically diverse Irvingtonian faunas was discovered in Porcupine Cave in central Colorado (Barnosky, 2004). The only record from the Dakotas is from Salamander Cave, and the only diverse Irvingtonian biota from the Great Basin was also recovered from a cave (Bell et al. 2004b). There are unique complications in working with Irvingtonian cave biotas, including complicated stratigraphic and depositional histories, taphonomic pathways to preservation of fossils, and problems of establishing reliable external age control. These problems are not eliminated in open-air sites, but for many of the key open-air localities in the Irvingtonian, volcanic ash beds provide a ready means for establishing the age of faunal constituents.

Non-Mammalian Vertebrate Faunal Groups

A relatively recent summary of the North American Pleistocene fish, amphibian, reptile, and bird faunas was provided by

Bell (2000b). No comprehensive synopsis of the Pleistocene fishes of North America has appeared since Miller's (1965) compilation. Literature on the group remains widely scattered, and a modern summary would be a welcome and timely addition to syntheses of Quaternary biotas.

The avian fauna also lacks a modern summary treatment. Terminal Pleistocene extinctions were reviewed by Steadman and Martin (1984), but studies of early and middle Pleistocene avifaunas are not abundant. A remarkably diverse assemblage of birds from Porcupine Cave was recently reported by Emslie (2004), who summarized additional records from the early and middle Pleistocene of the inter-mountain west.

A relatively recent review of North American Pleistocene amphibian and reptile faunas was provided by Holman (1995). He subsequently also published monographic reviews of the fossil anurans (Holman, 2003), and snakes (Holman, 2000) of North America, and these include important discussions of the early and middle Pleistocene records. In sharp contrast to the mammalian fauna during the Pleistocene, the picture that has emerged for amphibians and reptiles is one of general taxonomic and geographical stability throughout the epoch (except for the areas where continental ice sheets covered the landscape). This may in part be a result of methodological approaches used in the identification of fossil amphibian and reptile remains, but this requires further investigation (Bell et al., 2004a; Bever, 2005).

Broader Significance of Irvingtonian Biotas

The extinction of the North American large mammal fauna at the end of the Pleistocene is perhaps one of the best-known and most intensively studied aspects of faunal dynamics in the Pleistocene (MacPhee, 1999; Alroy, 1999, 2001; Barnosky et al., 2004a (see *Early Pleistocene*)). Four decades of intense research have not yielded a consensus on causal mechanisms for the extinction event that eliminated most of the large-bodied mammals on the continent. The difficulty stems from the fact that North American faunas appear to have experienced at least two major perturbations during the late Pleistocene. The first is a general climatic warming following the end of the Last Glacial Maximum. The second is the arrival of humans, with a sophisticated hunting technology. Separating the impacts of the two has proven to be difficult, and both perturbations likely played a role in the extinction (Barnosky et al., 2004a, b).

Much of what we think we know about how current global warming will affect vertebrate communities in the future is derived from studies of the late Pleistocene-Holocene transition. There is a rich record of vertebrate faunas from this transition in North America (FAUNMAP Working Group, 1994), and this record has played an important role in shaping scientific opinions regarding the potential impacts of climatic change on biotas. The strong taxonomic similarity between modern biotas and those fossilized in the Pleistocene through Holocene invites direct assessment of the impacts of climate change on the past populations of species still extant in our modern biota. As paleontologists and paleoecologists explored the fossil record over the last 60 years, it became clear that the geographical distribution and community structure

of modern species could not be taken as an accurate portrayal of their past. The discovery of Pleistocene fossils at more southerly latitudes, of species that today occupy boreal and tundra environments, soon became an accepted signature of past climatic changes. In some cases, geographic distributions in the past included apparent sympatric associations of species that today are entirely allopatric. These assemblages have no modern 'analog' and beginning in the 1940s were used (often in conjunction with stratigraphic evidence of past glacial activity) as proxy data for climatic correlations, and to interpret past glacial-interglacial cycles and their impact on faunas. The reliability, prevalence, and interpretation of these assemblages was recently questioned by Alroy (1999). Alroy's challenge was based in large part on questioning the reliability of temporal associations and possible causal relationships between climate change, the development and disintegration of no-analog assemblages, and the end-Pleistocene extinction event. The controversy here points to a fruitful area for continued research.

The record of early and middle Pleistocene vertebrates does not yet mirror that of the younger intervals of the Pleistocene and Holocene, and synthetic analysis has been hampered by apparent faunal provinciality and a relative paucity of well-dated, taxonomically diverse faunas. Nonetheless, intriguing efforts to unravel local and regional community dynamics for earlier Pleistocene times reveal that the faunal dynamics of the late Pleistocene may not be generally representative of dynamics at early warming events, even within the Pleistocene (Barnosky et al., 1996). Broad-scale comparisons that integrate and compare data sets from different time intervals have been attempted and some generalities about biotic response to climatic warming are beginning to emerge. One recent analysis of the response of mammalian faunas to four separate warming events will serve as an example (Barnosky et al., 2003). That study revealed that population density changes and phenotypic change within populations can be expected within approximately 100 years of the onset of warming. Extinction, alteration of community composition and structure, and reduction in species richness are expected as warming continues for a few thousand years. Extended warm intervals (lasting hundreds of thousands to millions of years) would be expected to lead to almost total faunal turnover. In part this would result from cumulative changes derived from shorter-term processes, and in part as a result of evolutionary responses (speciation) to new environmental conditions.

References

- Alroy J (1999) Putting North America's end-Pleistocene megafaunal extinction in context: Large-scale analyses of spatial patterns, extinction rates, and size distributions. In: MacPhee RDE (ed.) *Extinctions in Near Time: Causes, Contexts, and Consequences*, pp. 105–143. New York: Kluwer Academic/Plenum Publishers.
- Alroy J (2001) A multispecies overkill simulation of the end-Pleistocene megafaunal mass extinction. *Science* 292: 1893–1896.
- Barnosky AD, Rouse TI, Hadly EA, Wood DL, Keesing FL, and Schmidt VA (1996) Comparison of mammalian response to glacial-interglacial transitions in the Middle and Late Pleistocene. In: Stewart KM and Seymour KL (eds.) *Palaeoecology and Palaeoenvironments of Late Cenozoic Mammals: Tributes to the Career of C. S. (Rufus) Churcher*, pp. 16–33. Toronto, Ontario: University of Toronto Press.
- Barnosky AD (ed.) (2004) *Biodiversity Response to Climate Change in the Middle Pleistocene: The Porcupine Cave Fauna from Colorado*, p. 385. Berkeley: University of California Press.

- Barnosky AD, Hadly EA, and Bell CJ (2003) Mammalian response to global warming on varied temporal scales. *Journal of Mammalogy* 84: 354–368.
- Barnosky AD, Koch PL, Feranec RS, Wing SL, and Shabel AB (2004a) Assessing the causes of Late Pleistocene extinctions on the continents. *Science* 306: 70–75.
- Barnosky AD, Bell CB, Emslie SD, Goodwin HT, Mead JI, Repenning CA, Scott E, and Shabel AB (2004b) Exceptional record of mid-Pleistocene vertebrates helps differentiate climatic from anthropogenic ecosystem perturbations. *Proceedings of the National Academy of Sciences of the United States of America* 101: 9297–9302.
- Bell CJ (2000a) Biochronology of North American microtine rodents. In: Noller JS, Sowers JM, and Lettis WR (eds.) *Quaternary Geochronology: Methods and Applications*, pp. 379–406. Washington DC: American Geophysical Union AGU. Reference Shelf 4.
- Bell CJ (2000b) Synopsis of terrestrial and non-marine aquatic faunal groups. In: Noller JS, Sowers JM, and Lettis WR (eds.) *Quaternary Geochronology: Methods and Applications*, pp. 407–411. Washington DC: American Geophysical Union AGU. Reference Shelf 4.
- Bell CJ, Head JJ, and Mead JI (2004a) Synopsis of the herpetofauna from Porcupine Cave. In: Barnosky AD (ed.) *Biodiversity Response to Climate Change in the Middle Pleistocene: The Porcupine Cave Fauna from Colorado*, pp. 117–126. Berkeley, California: University of California Press.
- Bell CJ, Lundelius EL Jr., Barnosky AD, Graham RW, Lindsay EH, Ruez DR Jr., Semken HA Jr., Webb SD, and Zakrzewski RJ (2004b) The Blancan, Irvingtonian, and Rancholabrean mammal ages. In: Woodburne MO (ed.) *Late Cretaceous and Cenozoic Mammals of North America: Biostratigraphy and Geochronology*, pp. 232–314. New York: Columbia University Press.
- Bever GS (2005) Variation in the ilium of North American *Bufo* (Lissamphibia; Anura) and its implications for species-level identification of fragmentary anuran fossils. *Journal of Vertebrate Paleontology* 25: 548–560.
- Brown B (1908) The Conard Fissure, a Pleistocene bone deposit in northern Arkansas: With descriptions of two new genera and twenty new species of mammals. *Memoirs of the American Museum of Natural History* 9: 155–208, Plates 13–24.
- Emslie SD (2004) The Early and Middle Pleistocene avifauna from Porcupine Cave. In: Barnosky AD (ed.) *Biodiversity Response to Climate Change in the Middle Pleistocene: The Porcupine Cave Fauna from Colorado*, pp. 127–140. Berkeley: University of California Press.
- FAUNMAP Working Group (1994) FAUNMAP: A database documenting late Quaternary distributions of mammal species in the United States. *Illinois State Museum Scientific Papers* 25: 1–690.
- Holman JA (1995) *Pleistocene Amphibians and Reptiles in North America*, p. 239. New York: Oxford University Press.
- Holman JA (2000) *Fossil Snakes of North America: Origin, Evolution, Distribution, Paleogeology*, p. 357. Indiana: Indiana University Press. Bloomington.
- Holman JA (2003) *Fossil Frogs and Toads of North America*, p. 246. Indiana: Indiana University Press. Bloomington.
- Kurtén B and Anderson E (1980) *Pleistocene Mammals of North America*, p. 442. New York: Columbia University Press.
- Lundelius EL Jr., Churcher CS, Downs T, Harington CR, Lindsay EH, Schultz GE, Semken HA, Webb SD, and Zakrzewski RJ (1987) The North American Quaternary sequence. In: Woodburne MO (ed.) *Cenozoic Mammals of North America: Geochronology and Biostratigraphy*, pp. 211–235. Berkeley, California: University of California Press.
- MacPhee RDE (ed.) (1999) *Extinctions in Near Time: Causes, Contexts, and Consequences*, p. 394. New York: Kluwer Academic/ Plenum Publishers.
- Martin LD (1979) The biostratigraphy of arvicoline rodents in North America. *Transactions of the Nebraska Academy of Sciences* 7: 91–100.
- Martin RA, Hurt RT, Honey JG, and Peláez-Campomanes P (2003) Late Pliocene and Early Pleistocene rodents from the northern Borchers Badlands (Meade County, Kansas), with comments on the Blancan-Irvingtonian boundary in the Meade Basin. *Journal of Paleontology* 77: 985–1001.
- Miller RR (1965) Quaternary freshwater fishes of North America. In: Wright HE Jr. and Frey DG (eds.) *The Quaternary of the United States*, pp. 569–581. Princeton, New Jersey: Princeton University Press.
- Opdyke ND and Channell JET (1996) *Magnetic Stratigraphy. International Geophysics Series*, vol. 64, p. 346. San Diego, California: Academic Press.
- Repenning CA (1987) Biochronology of the microtine rodents of the United States. In: Woodburne MO (ed.) *Cenozoic Mammals of North America: Geochronology and Biostratigraphy*, pp. 236–268. Berkeley, California: University of California Press.
- Repenning CA (1992) *Allophaiomys* and the age of the Olyor Suite, Krestovka Sections, Yakutia. *US Geological Survey Bulletin* 2037: 1–98.
- Repenning CA, Fejfar O, and Heinrich W-D (1990) Arvicolid rodent biochronology of the northern hemisphere. In: Fejfar O and Heinrich W-D (eds.) *International Symposium: Evolution, Phylogeny and Biostratigraphy of Arvicolids (Rodentia, Mammalia)*, pp. 385–417. Czechoslovakia: Prague Geological Survey.
- Savage DE (1951) Late Cenozoic vertebrates of the San Francisco Bay region. *University of California Publications, Bulletin of the Department of Geological Sciences* 28: 215–314.
- Schultz CB, Martin LD, Tanner LG, and Corner RG (1978) Provincial land mammal ages for the North American Quaternary. *Transactions of the Nebraska Academy of Sciences and Affiliated Societies* 5: 59–64.
- Steadman DW and Martin PS (1984) Extinction of birds in the Late Pleistocene of North America. In: Martin PS and Klein RG (eds.) *Quaternary Extinctions: A Prehistoric Revolution*, pp. 466–477. Tucson, Arizona: The University of Arizona Press.
- Wood HE II, Chaney RW, Clark J, Colbert EH, Jepsen GL, Reeside JB Jr., and Stock C (1941) Nomenclature and correlation of the North American continental Tertiary. *Bulletin of the Geological Society of America* 52: 1–48, Plate 1.
- Woodburne MO (ed.) (1987) *Cenozoic Mammals of North America: Geochronology and Biostratigraphy*. Berkeley, California: University of California Press.
- Woodburne MO (ed.) (2004) *Late Cretaceous and Cenozoic Mammals of North America: Biostratigraphy and Geochronology*, p. 391. New York: Columbia University Press.

Mid-Pleistocene of Southern Asia

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Introduction

This article gives the faunal lists, which are known from the literature, of Pleistocene large mammals from some of the better known sites in southeast Asia. Only the most important sites are discussed here, although there are clearly many more sites and caves known. However, the faunal lists from the latter are generally based on only a few species or even just a few specimens. The first problem encountered is that different authors give different faunal lists from a particular site. The basis of the lists is more or less the same, but some species, such as bovines, are differently interpreted. Furthermore, the faunal lists are in most cases based on collections made at the end of the nineteenth century and the first half of the twentieth century. These old collections were not always made under controlled stratigraphic conditions and consist of relatively a few specimens. Most of the faunal lists need an update, particularly from a taxonomic perspective.

The oldest descriptions of the faunas from a particular time slice within the sites of a particular area show that the sites have species in common, in most cases these are the larger faunal elements, such as proboscideans. However, they also show differences in species composition, mostly in the case of the cervids and smaller carnivores. This is caused principally by the fact that:

- (1) Absence of evidence was taken as evidence of absence.
- (2) There was widespread taxonomic 'splitting' and the creation of new species based on minor characters.
- (3) The material is sometimes fragmented in such a way that sound determinations could not be made, and the material can frequently be identified only to genus or family level.

In this way, the fauna of every site had its own 'set' of species, which could be termed 'local paleontology.' Thus, the presence/absence of particular species and artificial differences in comparable faunal elements in the different sites were interpreted as reflecting a chronological difference between the sites, leading to the establishment of a biostratigraphical framework. Without revisions of the material, an updated list cannot, unfortunately, be given. For this reason, the original lists are presented here, with (as far as possible) the author's reinterpretation. Some species names have been updated, like *Felis tigris* (= *Panthera tigris*), *Palaeoloxodon namadicus* (= *Elephas namadicus*) and *Euarctos kokeni* (= *Ursus thibetanus*).

Absolute dates obtained at some sites are given, but in most cases these data are under discussion. Thus, the sites are attributed broadly to the early, middle, or late Pleistocene. To have a better understanding of the region of southeast Asia, the Indian subcontinent is included.

Early-middle Pleistocene Faunas from Open Sites

Pakistan and India

The Siwalik Hills

Table 1 shows the most complete list of 'early Pleistocene' Siwalik mammals from Pinjor, made by Colbert (1935). 'Pinjor' is supposed to span Olduvai N (1.9 Ma, MIS 73–72) to Jaramillo N (1.18 Ma, MIS 35–34). The collection was assembled by Barnum Brown and most of the specimens are surface finds. This list shows clearly a mixture of (Mio) Plio-Pleistocene species and is therefore not useful for biostratigraphic purposes. In addition, the taxonomy is in clear need of updating, since many species names are synonyms. For example:

- (1) *Simia* cf. *satyrus*, which is based on one specimen. This is probably a *Sivapithecus*, but the specimen is lost. The following other specimens are noted.
- (2) *Crocota sivalensis* = *Crocota felina* = *Crocota colvini* = *Pachycrocota brevirostris* *Hyaenictis bosei* = *Thalassictis indicum* from the Mio/Pliocene.
- (3) *Pentalophodon sivalensis* = *Anancus sivalensis* (early Pleistocene).
- (4) *Archidiskodon planifrons* = *Elephas planifrons* (early Pleistocene).
- (5) *Probubalus triquetricornis* = *Hemibos triquetricornis*.
- (6) *Probubalus acuticornis* = *Hemibos acuticornis*.

The list gives a good insight into the kinds of mammals there were in the (Mio) Plio-early Pleistocene of the Siwaliks, notably *Merycopotamus dissimilis*, *Hexaprotodon sivalensis*, *Stegodon ganesa*, *Elephas hysudricus*, and *Pachycrocota brevirostris*.

Mangla-Samwal anticline (Pakistan)

The only faunal list (Table 2) based on a collection of fossils that were collected under controlled stratigraphic conditions, is from the Mangla-Samwal anticline in the Siwalik Hills, Pakistan (Hussain et al., 1992). The Mangla-Samwal anticline is located in the Jhelum re-entrant molasse basin. The oldest record of *E. planifrons* is ± 3.3 Myr BP in the Mangla-Samwal anticline, and the youngest is 2.7 Myr BP. Cervids and *E. hysudricus* first appear at the same level as the last occurrence of *E. planifrons*. The last record of *E. hysudricus* is ± 2.0 Myr BP, but this taxon may have continued for a longer period (Hussain et al., 1992). The youngest locality in the area is about 1.8 Ma old. Table 2 shows clearly that the fossils from the list of Colbert (1935) are from different levels. However, most of the taxa in the list are identified only to family or genus level, thereby limiting their use.

Table 1 Siwaliks (India, Pakistan), Narmada (India), Irrawady (Cambodia)

<i>Pinjor, Siwalik Hills (Colbert, 1935)</i>	<i>India, Narmada Beds (Soniaka, 1985)</i>	<i>Myanmar, Irrawady (Colbert, 1943)</i>
	<i>Homo</i> sp. (<i>Homo erectus</i>)	
<i>Simia</i> cf. <i>satyrus</i>		
<i>Papio falconeri</i>		
<i>Papio subhimalayanus</i>		
<i>Semnopithecus palaeindicus</i>		
<i>Megantereon palaeindicus</i>		
<i>Megantereon falconeri</i>		
<i>Felis subhimalayana</i>		
<i>Panthera cristata</i>		
<i>Sivafelis potens</i>		
<i>Sivafelis brachygnathus</i>		
<i>Canis cautleyi</i>		
<i>Hyaenictis bosei</i>		
<i>Crocota sivalensis</i>		
<i>Crocota felina</i>		
<i>Crocota colvini</i>		
<i>Agriotherium sivalense</i>		
<i>Melursus</i> (?) <i>theobaldi</i>		
<i>Hexaprotodon sivalensis</i>	<i>Hexaprotodon namadicus</i>	<i>Hexaprotodon iravaticus</i>
<i>Merycopotamus dissimilis</i>		<i>Merycopotamus dissimilis</i>
<i>Merycopotamus nanus</i>		
<i>Rhinoceros sivalensis</i>		<i>Rhinoceros sivalensis</i>
<i>Rhinoceros palaeindicus</i>		
<i>Coelodonta platyrhinus</i>		
<i>Stegolophodon stegodontoides</i>		<i>Stegolophodon latidens</i>
<i>Pentalophodon sivalensis</i>		
<i>Stegodon ganesa</i>	<i>Stegodon ganesa</i>	
<i>Stegodon insignis</i>	<i>Stegodon insignis</i>	<i>Stegodon insignis</i>
<i>Stegodon pinjorensis</i>		<i>Stegodon elephantoides</i>
<i>Archidiskodon planifrons</i>		
<i>Elephas hysudricus</i>	<i>Elephas namadicus</i>	<i>Elephas hysudricus</i>
<i>Platelephas platycephalus</i>		
<i>Equus namadicus</i>		
<i>Equus sivalensis</i>		<i>Equus</i> cf. <i>yunnanensis</i>
		<i>Hipparion</i> cf. <i>antelopinum</i>
<i>Damaliscus palaeindicus</i>		
<i>Taurotragus latidens</i>		
<i>Hippotragus</i> (?) <i>sivalensis</i>		
<i>Cobus</i> (?) <i>patulicornis</i>		
<i>Cobus gyricornis</i>		
<i>Cobus palaeindicus</i>		
<i>Capra sivalensis</i>		Caprinae
	<i>Gazella</i> sp.	<i>Gazella</i> sp./ Hippotraginae
<i>Bucapra daviesi</i>		

(Continued)

Table 1 (Continued)

<i>Pinjor, Siwalik Hills (Colbert, 1935)</i>	<i>India, Narmada Beds (Soniaka, 1985)</i>	<i>Myanmar, Irrawady (Colbert, 1943)</i>
<i>Leptobos falconeri</i>	<i>Leptobos</i> sp.	<i>Proleptobos birmanicus</i>
<i>Hemibos occipitalis</i>		
<i>Probubalus triquetricornis</i>		<i>Hemibos triquetricornis</i>
<i>Probubalus acuticornis</i>		
<i>Probubalus antelopinus</i>		
<i>Bubalus platyceros</i>		
<i>Bubalus palaeindicus</i>	<i>Bubalus palaeindicus</i>	
<i>Bison sivalensis</i>		
<i>Bos acutifrons</i>	<i>Bos namadicus</i>	<i>Bos</i> cf. <i>sondaicus</i>
<i>Bos planifrons</i>		
<i>Bos playrhinus</i>		
<i>Sivatherium giganteum</i>		
<i>Indratherium majori</i>		
<i>Giraffa sivalensis</i>		
<i>Camelopardalis affinis</i>		
<i>Camelus sivalensis</i>		
<i>Cervus sivalensis</i>	<i>Cervus duvauceli</i>	<i>Cervus</i> sp.
<i>Cervus punjabiensis</i>		
<i>Cervus simplicidens</i>		
<i>Cervus triplidens</i>		
<i>Axis shansius</i>		
<i>Axis rugosus</i>		
<i>Metacervulus capreolus</i>		
<i>Muntiacus lacustris</i>		
<i>Paracuvulus attenuatus</i>		
<i>Sus strozzii</i>		
<i>Sivachoerus giganteus</i>		
<i>Tetraconodon magnus</i>		
<i>Tetraconodon mirabilis</i>		
<i>Potamochoerus theobaldi</i>		<i>Potamochoerus</i> sp.
<i>Potamochoerus palaeindicus</i>		
<i>Dicoryphochoerus vagus</i>		
<i>Dicoryphochoerus durandi</i>		
<i>Hippohyus sivalensis</i>		
<i>Sus hysudricus</i>		
<i>Sus bakeri</i>		
<i>Sus falconeri</i>		
<i>Sus cautleyi</i>		

Narmada beds (India)

Table 1 also presents the faunal list of the Narmada (Narbada) Beds (Soniaka, 1985), equally in need of updating. *H. namadicus* is a junior synonym for *H. sivalensis*. The list clearly shows Siwalik elements, such as the *Stegodon* species. The age estimations range from middle to early-late Pleistocene.

Myanmar

The faunal list of the Irrawady beds in Myanmar (**Table 1**) is based on Colbert (1943). The material was not collected under controlled stratigraphic conditions. The fauna contains typical

Table 2 Mammals of the Mangla–Samwal Anticline, Pakistan

Mangla–Samwal anticline	
Elephas planifrons Range-zone	Elephas hysudricus Range-zone
3.3–2.7 Ma	2.7 Ma–?
	<i>Caprolagus sivalensis</i>
	<i>Canidae</i> indet.
	<i>Ursidae</i> indet.
	Mustellidae indet.
	<i>Crocuta sivalensis</i> = <i>Hyaena brevirostris</i>
<i>Hyaenidae</i> indet.	<i>Hyaenidae</i> indet.
<i>Carnivora</i> indet.	
<i>Pentalophodon</i> sp. (= <i>Anancus</i> sp.)	
<i>Stegodon</i> sp.	<i>Stegodon</i> sp.
	<i>Elephas hysudricus</i> (Falconer and Cautley)
<i>Elephas planifrons</i> Falconer & Cautley	
<i>Proboscidea</i> indet.	<i>Proboscidea</i> indet.
	<i>Equus</i> cf. <i>sivalensis</i> (Falconer and Cautley)
	<i>Equus</i> sp. A. small species
<i>Equidae</i> indet.	<i>Equidae</i> indet.
	<i>Rhinoceros sensu lato</i>
<i>Suidae</i> indet	<i>Suidae</i> indet.
	<i>Merycopotamus dissimilis</i> (Falconer and Cautley)
<i>Hexaprotodon sivalensis</i> Falconer and Cautley	<i>Hexaprotodon sivalensis</i> (Falconer and Cautley)
	<i>Cervus triplidens</i>
	<i>Cervidae</i> indet.
	<i>Sivatherium giganteum</i>
Large Giraffid	Large Giraffid
<i>Antilope</i>	<i>Antilope</i>
	<i>Hemibos</i> sp.
cf. <i>Leptobos kashmiricus</i>	
	Large Bovid sp. C
	Large Bovid sp. D
<i>Bovidae</i> indet.	<i>Bovidae</i> indet.

Based on Hussain ST, van den Bergh GD, Steensma KJ, et al. (1992) Biostratigraphy of the Plio-Pleistocene continental sediments (Upper Siwaliks) of the Mangla–Samwal Anticline, Azad Kashmir, Pakistan. *Proceedings of the Koninklijke Nederlandse Akademie van Wetenschappen, Series B* 95(1): 65–80.

Siwalik species, such as *S. insignis* and *E. hysudricus*. The presence of *Hipparion* points to an age within the Pliocene/early-early Pleistocene, whereas the *E. hysudricus* points to the middle Pleistocene. The age of this fauna assemblage should therefore be considered as (Plio?) early–middle Pleistocene.

Thailand

Nakorn Sawan

Lekagul (1949) reported *Hexaprotodon* sp., *S. ganesa*, and *Bubalus* from Nakorn Sawan, in central Thailand, where Siwalik species are again encountered. The age can be considered as early–middle Pleistocene.

Sunda Shelf – Java

The Satir fauna

The oldest fauna in Java is from Satir (Figure 1), where we find the oldest proboscidean, *Sinomastodon* (= *Tetralophodon*) *bumiajuensis*. The only other fossils, besides *S. bumiajuensis*, are a hippo, *H. simplex*, cervids, and a giant tortoise *Geochelone*. This fauna is designated as the *Tetralophodon–Geochelone* fauna. The unbalanced character of the assemblage is suggestive of island isolation. The age is considered to be early Pleistocene (de Vos and Long, 2001).

Ci Saat fauna

A younger fauna is from Ci Saat (Figure 1, in the area of Satir) and consists of *S. trigonocephalus*, *H. sivalensis*, cervids, bovines, possibly a felid *Panthera* sp. and a suid, *Sus stremmi*. The age of the fauna is considered to be late-early Pleistocene (de Vos and Long, 2001).

The Trinil HK (Haupt Knoenschicht), Kedung Brubus, and Ngandong fauna

The faunas from Trinil HK, Kedung Brubus, and Ngandong (see Figure 1 for site locations) are given in Table 3 (de Vos and Long, 2001). The faunal lists from Trinil and Kedung Brubus are based on fossils in the Dubois collection (now housed in Leiden, The Netherlands) and every fossil has been individually described. The oldest site, Trinil, became famous on account of the finding there, between 1891 and 1893, of the first fossils of *Homo erectus*. Almost all the species or subspecies of this fauna are endemic in Java.

The younger Kedung Brubus fauna comes from the locality of Dubois, situated near the village of Kedung Brubus.

The type locality of the Ngandong fauna is a fluvial terrace some 20 m above sea level near the village of Ngandong. There, several *H. erectus soloensis* skulls were found in association with mammalian remains.

The high number of bovine fossils and the low number of primate fossils point to a dry climate and an open woodland environment.

The Ci Saat, Trinil HK, Kedung Brubus, and Ngandong fauna assemblages have been designated the *Stegodon–H. erectus* fauna (de Vos and Long, 2001).

Within this faunal complex, we find the following faunal elements (although all unfortunately lack stratigraphical context): *Macaca nemestrina*, *Hystrix sgigantea*, *Caprolagus* cf. *sivalensis*, *Hemimachairodus zwierzyckii*, *Megantereon* sp., *Homotherium ultimum*, *P. pardus*, *Lutrogale robusta*, *Cuon* sp., *Nestoritherium* cf. *sivalense*, and *M. dissimilis*.

The Satir and *Stegodon–H. erectus* faunas from Java show clear affinities with the early Pleistocene faunal association from the Indian subcontinent (the Siwaliks), Myanmar, and Thailand. Five species have a direct relation with the fauna of the Siwaliks, viz., *H. sivalensis*, *P. brevirostris*, *C. cf. sivalensis*, *N. cf. sivalense*, and *M. megartereon*.

Three species from this Javan fauna are closely related to Siwalik species, viz., *S. trigonocephalus* to *S. ganesa*, *E. husudricus* to *E. hysudricus*, and *D. santeng* to the Bose-laphini. De Vos and Long, (2001) therefore concluded that



Figure 1 Map of southeast Asia indicating the position of some sites mentioned in the text and the so-called Siva-Malayan and Sino-Malayan routes.

the *Stegodon*–*H. erectus* faunal association originated from the Siwaliks and reached Java by filter dispersal via the so-called Siva-Malayan route (Figure 1). Other Asian species, such as horses and camels, in contrast, never reached Java.

This early-middle Pleistocene ecosystem, consisting of an open environment, ranged from the Indian subcontinent via Myanmar and Thailand to Java, and supported populations of *H. erectus*.

Table 3 Trinil HK, Kedung Brubus, and Ngandong

<i>Trinil HK (de Vos and Long, 2001)</i>	<i>Kedung Brubus (de Vos and Long, 2001)</i>	<i>Ngandong (de Vos and Long, 2001)</i>
<i>Homo erectus</i>	<i>Homo erectus</i>	<i>Homo erectus soloensis</i>
<i>Trachypithecus cristatus</i>	<i>Trachypithecus cristatus</i>	
<i>Macaca fascicularis</i>	<i>Macaca fascicularis</i>	<i>Macaca fascicularis</i>
<i>Acanthion brachyurus</i>	<i>Acanthion brachyurus</i>	
<i>Rattus trinilensis</i>	<i>Rattus trinilensis</i>	
<i>Panthera tigris trinilensis</i>	<i>Panthera tigris trinilensis</i>	<i>Panthera tigris soloensis</i>
<i>Prionailurus bengalensis</i>	<i>Prionailurus bengalensis</i>	
<i>Mececyon trinilensis</i>	<i>Mececyon trinilensis</i>	
<i>Stegodon trigonocephalus</i>	<i>Stegodon trigonocephalus</i>	<i>Stegodon trigonocephalus</i>
<i>Rhinoceros sondaicus</i>	<i>Rhinoceros sondaicus</i>	
<i>Muntiacus muntjak</i>	<i>Muntiacus muntjak</i>	
<i>Axis lydekkeri</i>	<i>Axis lydekkeri</i>	Cervids
<i>Duboisia santeng</i>	<i>Duboisia santeng</i>	
<i>Bubalus palaeokerabau</i>	<i>Bubalus palaeokerabau</i>	<i>Bubalus palaeokerabau</i>
<i>Bibos palaesondaicus</i>	<i>Bibos palaesondaicus</i>	<i>Bibos palaesondaicus</i>
<i>Sus brachygnathus</i>	<i>Sus brachygnathus</i>	<i>Sus brachygnathus</i> (?)
	<i>Lutrogale palaeoleptonyx</i>	
	<i>Hyaena brevirostris</i>	
	<i>Rhinoceros unicornis kendengindicus</i>	
	<i>Tapirus indicus</i>	<i>Tapirus indicus</i>
	<i>Rusa sp.</i>	
	<i>Epileptobos groeneveldtii</i>	
	<i>Sus macrognathus</i>	<i>Sus macrognathus</i>
	<i>Manis palaeojavanica</i>	
	<i>Hexaprotodon sivalensis</i>	<i>Hexaprotodon sivalensis</i>
	<i>Elephas hysudrindicus</i>	<i>Elephas hysudrindicus</i>
		Cervids

Early-middle Pleistocene Faunas from Caves or Fissure Fillings

China

South Chinese caves and fissures

Longgupo cave

The Longgupo site is located at Damiao, Wushan County, Chongqing. The cave lies 20 km south of the Yangtze River near the eastern border of Sichuan (=Szechwan) province. More than 120 species have been collected. In the middle zone of the site, early Pleistocene species such as *Sinomastodon* sp., *Equus yunnanensis*, *Nestoritherium* sp., *Ailurapoda microta*, *Miomys peii*, and an early *Homo* sp. are contained. In addition, *Gigantopithecus blacki*, *P. licenti*, *Homotherium*, and *Hystris* are also present. In contrast to the aforementioned sites where there are only dental remains, here long bones are present. The cave is well known for its stone artefacts as well as a mandible fragment and incisor, which are attributed to *Homo* sp. The age is estimated to lie within the Olduvai chron, 1.96–1.78 Myr. (Wanpo et al., 1995).

Yenchinkou

Table 4 shows the fauna list of Yenchinkou, South China (Figure 2). The list is based on fossils from limestone fissures or pits on the tops of high Paleozoic ridges, which parallel the Yangtze River in Szechwan, collected by Walter Granger during the 1920s. Granger made frequent visits along the top of the limestone ridge to purchase from the native diggers the best of the fossils procured by them (Colbert and Hooijer, 1953).

Matsumoto (Colbert and Hooijer, 1953), who described the fossils obtained by T. Sakawa, without having visiting the site, considered the Szechwan (= Sichuan) fossils as belonging to two distinct faunas, an older or ‘*Stegodon* fauna’ and a younger fauna from the ‘cave loam.’ According to Granger, this represented an ‘ecological stratification.’ The fauna was later described by Colbert and Hooijer (1953). The assemblage appears to represent a mixture of early Pleistocene elements, such as *Nestoritherium*, and Middle or late Pleistocene faunal elements, such as *Stegodon orientalis*.

The caves Liucheng, Hoshantung, Ta-hsin, Changyang, Koloshan and Hsing-an

Table 4 shows the faunal lists from Liucheng, Hoshantung, Ta-hsin, Changyang, Koloshan, and Hsing-an (Figure 2). AS.-*Ailuropoda* fauna has been described from these caves (Kahlke, 1961a, b), consisting of a combination of primitive genera (*Mastodon*, *Stegodon*) and some advanced or recent genera (*Elephas*), with animals adapted to woodland or forest environments (*Pongo*), and animals adapted to mixed or more open habitats (*Equus*). The *Stegodon*–*Ailuropoda* faunas are known from different caves, but each appears to represent a discrete time period. The fauna from one of these caves (Hoshantung; faunal composition according to Kahlke (1961b)) is generally considered to be a late part of the *Stegodon*–*Ailuropoda* fauna, dating from the late-middle Pleistocene (Kahlke, 1968) or early-late Pleistocene (Aigner, 1978), based on the presence of *Palaeoloxodon* (= *Elephas*). Within this age we can add the sites of Ta-hsin, Changyang, Koloshan, and

Table 4 South Chinese cave faunas

<i>Liucheng Cave (Pei, 1957; Kahlke, 1961a,b)</i>	<i>Hoshangtung cave (Bien and Chia, 1938; Kahlke, 1961a,b)</i>	<i>South China caves Hei-T'ung Cave, Ta-hsin (Kahlke, 1961a,b)</i>	<i>Changyang (Kahlke, 1961a,b)</i>	<i>Koloshan (Kahlke, 1961a,b)</i>	<i>Hsing-an (Kahlke, 1961a,b)</i>
<i>Gigantopithecus blacki</i>		<i>Gigantopithecus blacki</i>	<i>Homo</i> sp.	<i>Szechuanopithecus yangtsensis</i>	
<i>Pongo</i> sp. (<i>Pongo pygmaeus</i> according to Aigner, 1978)	<i>Pongo pygmaeus weidenreichi</i> (5)	<i>Pongo</i> sp.			<i>Pongo pygmaeus weidenreichi</i>
<i>Primates</i> indet.	<i>Macaca</i> sp. (?)			<i>Macaca robustus</i>	
<i>Hystrix</i> cf. <i>subcristata</i>	<i>Hystrix</i> sp. (100 teeth)				
<i>Ursus</i> n.sp. (cf. <i>Ursus thibetanus kokeni</i>)	<i>Ursus thibetanus kokeni</i> (abundant isolated teeth)		<i>Ursus angustidens</i>	<i>Ursus thibetanus koken</i>	<i>Ursus thibetanus koken</i>
<i>Ailuropoda</i> n. sp.	<i>Ailuropoda melanoleuca fovealis</i> (several isolated teeth)	<i>Ailuropoda</i> sp.	<i>Ailuropoda</i> sp.	<i>Ailuropoda melanoleuca baconi</i>	<i>Ailuropoda melanoleuca baconi</i>
	<i>Ailurus fulgens</i> (based on 2 molars)				
<i>Arctonyx</i> sp.	<i>Arctonyx</i> sp. (?)	<i>Arctonyx collaris rostralis</i>		<i>Parameles simplicidens</i>	<i>Arctonyx collaris rostratus</i>
		<i>Paguma larvata</i> ?	<i>Meles</i> sp.	<i>Mustela siberica/ Martes sinensis</i>	
<i>Cuon</i> sp.	<i>Canidae</i> indet.	<i>Cuon</i> sp.	<i>Cuon antiquus</i>	<i>Cuon simplicidens/ Canis</i> sp.	<i>Canidae</i> indet.
<i>Hyaena brevirostris licenti</i> (= <i>Pachycrocuta brevirostris</i>)	<i>Crocota crocuta ultima</i> (21 premolars and 3 molars)		<i>Hyaena brevirostris sinensis</i>		<i>Crocota crocuta ultima</i>
<i>Felis</i> sp.	<i>Panthera</i> cf. <i>tigris</i> (?)		<i>Panthera tigris</i>	<i>Panthera tigris</i>	<i>Panthera tigris</i>
	<i>Panthera</i> cf. <i>pardus</i> (?)	<i>Felis</i> sp.	<i>Felidae</i> indet.		<i>Felis</i> sp. (lynx-sized)
	<i>Lynx</i> cf. <i>lynx</i> (?)				
<i>Stegodon preorientalis</i> (= <i>Stegodon orientalis</i>) (1)	<i>Stegodon</i> sp. (= <i>Stegodon orientalis</i>) (2)	<i>Stegodon</i> sp. (= <i>Stegodon orientalis</i>)	<i>Stegodon orientalis</i>		<i>Stegodon orientalis</i>
	<i>Palaeoloxodon</i> cf. <i>namadicus</i> (3)				<i>Palaeoloxodon</i> sp.
<i>Tapirus</i> sp. (smaller than <i>Megatapirus</i>)	<i>Megatapirus</i> cf. <i>augustus</i> (5)	<i>Megatapirus</i> sp.	<i>Megatapirus augustus</i>	<i>Megatapirus augustus</i>	<i>Tapirus sinensis</i>
<i>Rhinoceros sinensis</i>	<i>Rhinoceros</i> sp. (probably <i>Rhinoceros sinensis</i>) (a number)	<i>Rhinoceros sinensis</i>	<i>Rhinoceros sinensis</i>	<i>Rhinoceros sinensis</i>	<i>Rhinoceros sinensis</i>
<i>Sus scrofa</i>	<i>Sus</i> sp. (many isolated teeth)	<i>Suidae</i> indet.	<i>Sus</i> sp.	<i>Sus</i> sp.	<i>Sus</i> sp.
<i>Cervidae</i> gen. et sp. indet.	? <i>Rusa</i> sp. I (several hundred isolated teeth)	<i>Cervidae</i> indet.	<i>Cervidae</i> indet.	<i>Rusa</i> sp.	<i>Rusa</i> sp. 1
	? <i>Rusa</i> sp. II (?)				<i>Rusa</i> sp. 2
	<i>Muntiacus</i> sp. (numerous isolated teeth)			<i>Muntiacus szechuanensis</i>	<i>Muntiacus</i> sp.
<i>Caprinae</i> gen. et sp. indet.	<i>Caprinae</i> gen. et sp. indet. I, II (?)			<i>Ovis</i> sp.	
<i>Bovidae</i> gen. et sp. indet.	<i>Bovidae</i> gen. et sp. indet. (?)	<i>Bovidae</i> indet.	<i>Bovidae</i> indet. 1 <i>Bovidae</i> indet. 2	<i>Bubalus brevirostris</i>	<i>Bovinae</i> indet.
<i>Chalicotheriidae</i> gen et sp. indet.					
<i>Equus yunnannensis</i> (one isolated tooth)					
<i>Paguma</i> ? <i>Larvata</i>					
<i>Mastodon</i> sp. (one isolated molar)					

(5) number of specimens on which the species is based.

(?) number of specimens on which the species is based is not given.

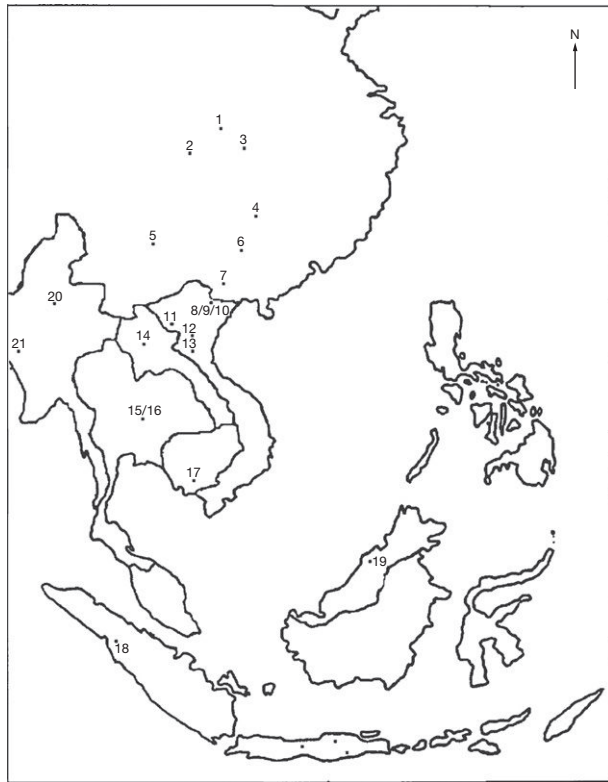


Figure 2 Map of southeast Asia indicating the position of some sites mentioned in the text: 1 = Yenchingkuo, 2 = Koloshan, 3 = Changyang, 4 = Hsing-an, 5 = Hoshantung, 6 = Liucheng, 7 = Ta-hsin, 8 = Keo Leng, 9 = Thâm Khuyen, 10 = Thâm Hai, 11 = Hang Hum, 12 = Lang Trang, 13 = Thâm Om, 14 = Thâm Hang and Thâm P'a Loi, 15 = Thum Wiman Nakin, 16 = Thum Phra Khai Phet, 17 = Phnom Loang, 18 = Sumatran caves (Lida Ajer), 19 = Niah, 20 = Mogok, 21 = Narmada.

Hsing-an. The fauna of Liucheng Cave (see Kahlke (1961a); Table 4), is generally considered to belong to an early part of the *Stegodon*-*Ailuropoda* fauna, ranging in time from the early Pleistocene (Linghong and Lian, 1982) until the early-middle Pleistocene (Li Yanxian, 1981), based on the presence of archaic elements such as *Mastodon* sp. (represented by one isolated lower molar) and *E. yunnanensis* (also represented by one isolated tooth). The exact stratigraphical position of the fossils is not known.

Vietnam

Lang Trang Cave

The Lang Trang Caves are situated about 120 km southwest of Hanoi (Figure 2). Various very cemented 'breccias' were distinguished in the caves, and about 10,000 fossils were collected (de Vos and Long, 2001), consisting only of molars, most of which have the roots gnawed. The faunal list, given in Table 5 (de Vos and Long, 2001), is based on the description of 3250 fossils from one breccia B5 from Cave II. Although Ciochon et al. (1990) mentioned *H. cf. erectus* from this cave, *Homo* specimens have not been found. The age of Lang Trang ranges dates from >350 ka to 180 ka, that is, Middle to late

Pleistocene (Ciochon, personal communication; see Late Pleistocene of Southeast Asia).

The Thâm Khuyen, Thung Lang, Thâm Hai, Thâm Om, Hang Hum, and Keo Leng Caves

In the Thâm Khuyen Cave (Table 5), *H. erectus* apparently co-occurs with *Gigantopithecus*, and *P. pygmaeus* (Cuong, 1971; Ciochon et al., 1996). However, this must be treated with caution on account of the regular confusion of *Pongo* molars with those of hominids. The *H. erectus* molar from the site, which is illustrated by Ciochon et al. (1996), is typical of this. The site is dated 475 ± 125 ka (Ciochon et al., 1996). The fauna is similar to the one from Lang Trang. This is also valuable for the faunas excavated in the 1960s (Table 5) of Thung Lang, Thâm Hai (300–200 ka), Thâm Om, Hang Hum (=Tan Van Caves, consisting of two horizons, I and II). Hang Hum I is considered to be 80–140 ka (Cuong, 1985), Keo Leng 42–37 ka (Ciochon et al., 1996) (see for position of the sites in Figure 2). Although there are small differences in species composition of the different caves, we can be sure that we are dealing with one ecosystem, a tropical rainforest.

Laos

The caves Thâm Hang and Thâm P'a Loi

An update of the late middle Pleistocene Thâm Hang (Figure 2) fauna is given by de Vos et al. (in prep). Thâm P'a-Loi, a dozen kilometers north of Thâm-Hang, has yielded a similar fauna. In both the caves, bones are rare and porcupines have gnawed the roots of the dental elements.

Thailand

The caves Wiman Nakin and Thum Phra Khai Phet

Tougaard (1998) described the larger mammal (368) specimens from Thum Wiman Nakin and Thum Phra Khai Phet (Figure 2) in detail. Only the crowns of teeth, mostly with the roots gnawed, were recovered. The faunal list for Thum Wiman Nakin is given in Table 6. The position of the sites is given in Figure 2. The age is considered to be late-middle Pleistocene. The fauna of Thum Wiman Nakin is similar to the one from Lang Trang, with the notable presence of *Crocota*, which is unknown from the latter site.

Cambodia

The cave Phnom Loang

The faunal list from the site Phnom Loang, province of Kampot (Figure 2), is given in Table 6 (Beden and Guerin, 1973; Tougaard, 1998). The list is composed of only a few specimens, consisting only of molars. In this fauna, *P. pygmaeus* is apparently absent, although it has been reported from neighboring sites along with *Elephas* sp. and *Tapirus indicus intermedius* by Beden and Guerin (1973).

Sunda Shelf

The Sumatran caves

Table 7 shows the faunal list of the Lida Ajer Cave (see for position Figure 1) (de Vos and Long, 2001). The fauna

Table 5 Comparison of the Vietnamese fauna of Lang Trang with those of Tham Khuyen, Tham Hai, Tham Om, Hang Hum, and Keo Leng

Species	Vietnam Lang Trang (<i>de Vos & Long, 2001</i>)	Vietnam Tham Khuyen (<i>Cuong, 1971</i>)	Vietnam Tham Hai (=Lang Son) (<i>Cuong, 1985</i>)	Vietnam Tham Om (<i>Tougaard, 1998</i>)	Vietnam Hang Hum Horizon II (<i>Kahlke, 1972</i>)	Vietnam Hang Hum Horizon I (<i>Kahlke, 1972</i>)	Vietnam Keo Leng (<i>Cuong, 1985</i>)
<i>Homo</i> sp.	0	<i>Homo erectus</i>		<i>Homo erectus</i>			
<i>Pongo pygmaeus</i>	201	<i>Pongo pygmaeus weidenreichi</i>	<i>Pongo pygmaeus weidenreichi</i>		<i>Homo sapiens Pongo pygmaeus pygmaeus</i>	<i>Homo sapiens Pongo</i>	<i>Homo sapiens Pongo pygmaeus weidenreichi</i>
<i>Gigantopithecus blacki</i>	0	<i>Gigantopithecus blacki</i>					
<i>Presbytis/Trachypithecus/Macaca</i> sp.	358	<i>Macaca</i> cf. <i>assamensis/Mac.</i> sp.	<i>Macaca</i> sp.	<i>Macaca</i> sp.	<i>Macaca</i> sp.	<i>Presbytis</i> sp. <i>Macaca</i> sp.	<i>Presbytis</i> sp. <i>Macaca assamensis/Mac. mulatta</i>
<i>Hylobates</i> sp.	7	<i>Hylobates concolor</i>				<i>Hylobates</i> sp.	<i>Hylobates</i> sp.
<i>Arctonyx collaris</i>	33	<i>Arctonyx collaris</i> cf. <i>rosatratus</i>		<i>Arctonyx collaris</i> cf. <i>rosatratus</i>	<i>Arctonyx collaris</i> cf. <i>rosatratus</i>	<i>Arctonyx collaris</i> cf. <i>rosatratus</i>	<i>Arctonyx collaris</i> cf. <i>rosatratus</i>
<i>Panthera pardus</i>	9	<i>Panthera pardus</i>	<i>Panthera pardus</i>		<i>Panthera pardus</i>	<i>Panthera pardus</i>	<i>Panthera pardus?</i>
<i>Neofelis nebulosa</i>	0	<i>Neofelis nebulosa</i>			<i>Neofelis nebulosa</i>		
<i>Panthera tigris</i>	6	<i>Panthera tigris</i>	<i>Panthera tigris</i>	<i>Panthera tigris</i>		<i>Panthera tigris</i>	<i>Panthera tigris</i>
<i>Felis temmincki</i>	6	<i>Felis</i> sp.		<i>Felis</i> sp.			
<i>Ursus thibetanus</i>	33	<i>Ursus thibetanus kokeni</i>	<i>Ursus thibetanus kokeni</i>	<i>Ursus thibetanus kokeni</i>	<i>Ursus thibetanus kokeni</i>	<i>Ursus thibetanus kokeni</i>	<i>Ursus thibetanus kokeni</i>
<i>Ursus malayanus</i>	2	<i>Ursus malayanus</i>					
<i>Ailuropoda melanoleuca</i>	5	<i>Ailuropoda melanoleuca baconi</i>		<i>Ailuropoda melanoleuca baconi</i>			<i>Ailuropoda melanoleuca baconi</i>
<i>Cuon javanicus</i>	1	<i>Cuon javanicus/Canis</i> sp.		<i>Cuon</i> sp.	<i>Cuon</i> sp.		
<i>Nyctereutes</i> sp.	0	<i>Nyctereutes</i> sp.					
<i>Crocota crocuta ultima</i>	0						
<i>Paradoxurus hermaphoditus</i>	11	<i>Paradoxurus hermaphoditus</i>			<i>Paradoxurus</i> sp.	<i>Paradoxurus</i> cf. <i>hermaphoditus</i>	<i>Paradoxurus</i> cf. <i>hermaphoditus</i>

<i>Martes flavigula</i>	0						
<i>Paguma larvata</i>	0	<i>Paguma larvata</i>	<i>Paguma cf. larvata</i>			<i>Paguma larvata</i>	<i>Viverra cf. zibetha</i>
<i>Lutra perspicillata</i>	2						
<i>Elephas sp.</i>	14	<i>Elephas cf. namadicus</i>	<i>Elephas namadicus</i>	<i>Palaeoloxodon cf. namadicus</i>	<i>Palaeoloxodon namadicus</i>		
<i>Stegodon orientalis</i>	11	<i>Stegodon orientalis</i>	<i>Stegodon orientalis</i>	<i>Stegodon orientalis</i>	<i>Stegodon orientalis</i>		<i>Stegodon orientalis</i>
<i>Tapirus indicus</i>	18	<i>Tapirus sp.</i>	<i>Tapirus sp.</i>	<i>Tapirus sp.</i>			
<i>Megatapirus augustus</i>	0	<i>Megatapirus augustus</i>		<i>Megatapirus augustus</i>			<i>Megatapirus augustus</i>
<i>Rhinoceros sp.</i>	190	<i>Rhinoceros sinensis</i>	<i>Rhinoceros sinensis</i>	<i>Rhinoceros sinensis</i>	<i>Rhinoceros sinensis</i>		<i>Rhinoceros sinensis</i>
<i>Sus scrofa</i>	1027	<i>Sus scrofa/Sus cf lydekkeri/Sus sp.</i>	<i>Sus scrofa</i>	<i>Sus scrofa/Sus sp.</i>	<i>Sus scrofa</i>	<i>Sus scrofa</i>	<i>Sus scrofa/Sus cf lydekkeri/Sus sp.</i>
<i>Large bovid</i>	33	<i>Bubalus bubalis/Bos frontalis</i>	large Bovids	<i>Bubalus bubalis/Bos frontalis</i>		<i>Bubalus bubalis/Bibos gaurus cf. grangeri</i>	<i>Bubalus bubalis</i>
<i>Naemorhedus/Capricornis sp.</i>	23	<i>Capricornis sumatraensis</i>		<i>Naemorhedus sumatraensis</i>	Caprinae indet.		<i>Naemorhedus sumatraensis</i>
<i>Muntiacus muntjak</i>	481	<i>Muntiacus muntjak</i>		<i>Muntiacus muntjak margae/M. sp.</i>			<i>Muntiacus muntjak/M. sp.</i>
<i>Rusa sp.</i>	411	<i>Rusa unicolor</i>	<i>Rusa cf. unicolor</i>	<i>Rusa unicolor</i>	<i>Rusa unicolor</i>	<i>Rusa unicolor</i>	<i>Rusa unicolor</i>
<i>Cervus sp.</i>	0	<i>Cervus sp.</i>		<i>Cervus sp./Elaphodus sp.</i>			<i>Cervus sp.</i>
<i>Hystrix brachyura</i>	358	<i>Hystrix subcristata</i>	<i>Hystrix subcristata</i>		<i>Hystrix subcristata</i>	<i>Hystrix subcristata</i>	<i>Hystrix subcristata</i>
<i>Rhizomys sp.</i>	0	<i>Rhizomys cf. troglodytes</i>			<i>Rhizomys cf. troglodytes</i>	<i>Rhizomys cf. troglodytes</i>	<i>Rhizomys cf. troglodytes</i>
<i>Atherurus macrourus</i>	8	<i>Atherurus sp.</i>			<i>Atherurus sp.</i>	<i>Atherurus cf. macrourus</i>	<i>Atherurus cf. macrourus</i>
<i>Rattus sp.</i>	2	<i>Rattus rattus</i>					<i>Rattus sp./Mus. Sp.</i>
	2684						

N = number of specimens.

Table 6 Comparison of Lang Trang (Vietnam), Tam Hang (Laos) yellow tuff, Tam Hang (Laos) (red breccia), Thum Wiman Nankin (Thailand)

Species	Vietnam Lang Trang yellow breccia (de Vos and Long, 2001)	Thailand Thum Wiman Nakin (Tougaard, 1998)	Thailand Thum Phra Khai Phet (Tougaard, 1998)	Laos Tam Hang yellow tuff (de Vos, et al. (in prep))	Laos Tam Hang red breccia (de Vos, et al. (in prep))	Birma Mogok (Tougaard, 1998)	Cambodia Phnom Loang (Beden and Guerin, 1973 Tougaard, 1998)
<i>Homo</i> sp.	0	<i>Homo</i> sp. 1		<i>Homo</i> sp.?	0		
<i>Pongo pygmaeus</i>	201	16		X	0		<i>Pongo pygmaeus</i>
<i>Gigantopithecus blacki</i>	0	0		?	0		
<i>Presbytis/Trachypithecus/maca.</i>	358	42		X	X		
<i>Hylobates</i> sp.	7	0		X	0		
<i>Arctonyx collaris</i>	33	7		X	X		
<i>Panthera pardus</i>	9	0		X	0		
<i>Panthera tigris</i>	6	0		X	0		<i>Panthera tigris</i>
<i>Felis temmincki</i>	6	0		0	0		
<i>Ursus thibetanus</i>	33	18		X	?		
<i>Ursus malayanus</i>	2	0		X	?		
<i>Ailuropoda melanoleuca</i>	5	2		X	0	<i>Ailuropoda melanoleuca baconi</i>	
<i>Cuon javanicus</i>	1	0		X	0		
<i>Crocuta crocuta ultima</i>	0	10	<i>Crocuta crocuta ultima</i>	X	X		<i>Crocuta crocuta ultima</i>
<i>Paradoxurus hermaphoditus</i>	11	3		X	X		
<i>Martes flavigula</i>	0	5		0	0		<i>Martes flavigula</i>
<i>Paguma larvata</i>	0	2		0	0		
<i>Lutra perspicillata</i>	2	0		0	0		
<i>Elephas</i> sp.	14	2		X	X	<i>Elephas namadicus</i>	<i>Elephas</i> sp.
<i>Stegodon orientalis</i>	11	0		X	0	<i>Stegodon orientalis</i>	
<i>Tapirus indicus</i>	18	1		X	X		<i>Tapirus indicus intermedius</i>
<i>Rhinoceros</i> sp.	190	6	<i>Rhinoceros sondaicus</i>	X	X	<i>Rhinoceros</i> sp.	<i>Rhinoceros sondaicus guthi</i>
<i>Sus scrofa</i>	1027	56	<i>Sus</i> cf. <i>barbatus</i>	X	X	<i>Sus</i> sp.	
<i>Large bovid</i>	33	52	<i>Bos sauveli</i>	X	X	Bovinae	<i>Bubalus</i> cf. <i>bubalus</i>
<i>Naemorhedus/Capricornis</i> sp.	23	20	<i>Naemorhedus sumatraensis</i>	X	X		<i>Naemorhedus</i> sp.
<i>Muntiacus muntjak</i>	481	51	<i>Muntiacus muntjak</i>	X	X		
<i>Rusa</i> sp.	411	93	Cervidae indet./ <i>Axis porcinus</i>	X	X	<i>Cervus</i> sp.	<i>Rusa unicolor</i> /R. cf. <i>leptodus</i>
<i>Hystrix brachyura</i>	358	0		X	X		
<i>Rhizomys</i> sp.	0	X		X	X		
<i>Atherurus macrourus</i>	8	0		0	0		
<i>Rattus</i> sp.	2	X		X	X		
	2684	328					

Number of specimens (143), X present, 0 absent.

Table 7 Comparison of the Sumatran Cave fauna, Lang Trang (Vietnam), Punung, Gunung Dawung (Java), Niah (Malaysia)

Species	Vietnam Lang Trang yellow breccia (de Vos and Long, 2001)	Sumatra Sumatran Caves (idem).	Java Punung (Storm et al., 2005)	Java Gunung Dawung (Storm et al., 2005)	Borneo Niah Malaysia (de Vos and Long, 2001)
<i>Homo</i> sp.	0	<i>Homo sapiens</i>	<i>Homo sapiens</i>	<i>Homo sapiens</i>	<i>Homo sapiens</i>
<i>Pongo pygmaeus</i>	201	143	199	X	X
<i>Gigantopithecus</i> <i>blacki</i>	0	0	0	0	0
<i>Presbytis</i> / <i>Trachypithecus</i> / <i>maca</i> .	358	16	16	X	X
<i>Hylobates</i> sp.	7	a few	41	X	X
<i>Arctonyx collaris</i>	33	a few	0	o	X
<i>Panthera pardus</i>	9	a few	0	<i>Neofelis nebulosa</i>	X
<i>Panthera tigris</i>	6	a few	15	X	X
<i>Felis temmincki</i>	6	a few	0	0	0
<i>Ursus thibetanus</i>	33	0	0	0	0
<i>Ursus malayanus</i>	2	6	11	X	0
<i>Ailuropoda</i> <i>melanoleuca</i>	5	0	0	0	X
<i>Cuon javanicus</i>	1	a few	0	0	X
<i>Crocota crocuta</i> <i>ultima</i>	0	0	0	0	0
<i>Paradoxurus</i> <i>hermaphoditus</i>	11	a few	0	0	X
<i>Martes flavigula</i>	0	0	0	0	0
<i>Paguma larvata</i>	0	0	0	0	0
<i>Lutra perspicillata</i>	2	0	0	0	0
<i>Elephas</i> sp.	14	8	4	X	X
<i>Stegodon</i> <i>orientalis</i>	11	0	0	0	0
<i>Tapirus indicus</i>	18	32	21	X	X
<i>Rhinoceros</i> sp.	190	22	37	X	X
<i>Sus scrofa</i>	1027	a few hundred	>100	X	X
Large bovid	33	28	a few	X	X
<i>Naemorhedus</i> / <i>Capricornis</i> sp.	23	3	8	X	X
<i>Muntiacus muntjak</i>	481	167	49	X	X
<i>Rusa</i> sp.	411	43	12	X	X
<i>Hystrix brachyura</i>	358	>100	>100	X	X
<i>Rhizomys</i> sp.	0	0	0	0	X
<i>Atherurus</i> <i>macrourus</i>	8	0	0	0	0
<i>Rattus</i> sp.	2	a few	0	X	0
	2684		157		0

Number of specimens (143), X present, 0 absent.

composition is more or less similar to the one from Lang Trang, with the exception of the absence of *Stegodon* and *Ailuropoda*. As we are dealing with thousands of fossils (again, only molars, most of which have the roots gnawed), these appear to be genuine absences. A right upper central incisor, which is semi-shovel shaped, and a left upper molar, both from Lida Ajer cave (Figure 2), were identified as *Homo sapiens* by Hooijer (1948: 182–187). All species are still extant. The age is considered to be late Pleistocene (de Vos and Long, 2001). The large quantity of orangutan indicates a humid tropical rainforest environment. The age of the Lida Ajer material date from a little more than 80 000 (Drawhorn, 1994).

The Punung Caves and Gunung Dawung

Table 7 (Storm et al., 2005) shows the faunal list of Punung (Figure 1) and Gunung Dawung, which is in the vicinity of Punung. The Punung collection of Von Koenigswald originates from two different fissure deposits and consists only of molars, in which, in most cases, the roots are gnawed. The sites were relocated in 2003 by an Indonesian–Dutch team. A new site, Gunung Dawung, was discovered and excavated. The fauna of this site is not different from the table given here (Storm et al., 2005). In the so-called Punung fauna, *H. sapiens* is present (Storm et al., 2005), because of which this fauna is indicated as the *Pongo–H. sapiens* fauna (de Vos and Long, 2001), which

is of late Pleistocene age. The large quantity of orangutan indicates a humid tropical rainforest environment.

In all the caves mentioned above (south China, Vietnam, Myanmar, Thailand, Sumatra, and Java), the fossils come from a breccia. A similar taphonomy is found. Only crowns of teeth, mostly with the roots gnawed, were recovered. The gnawing was probably caused by porcupines, which bring bones and skulls to their lairs leaving only the tooth crowns.

Borneo: Niah Cave

Medway (1979) gave a faunal list of Niah Cave material, which is dated from 40 ka (Drawhorn, 1994) until Recent. The list (Table 7) is composed of both the Pleistocene and Holocene species found in Niah. Apart from domesticated mammals (like *Canis familiaris* in the upper 6–12 inch level and *S. scrofa* dom. in the 0–24 inch level), *P. pygmaeus* in the deepest level of 102–105 inches and *Hyllobates* in the deepest level of 60–72 inches are also present. Also *H. sapiens* is present in this cave. A remarkable feature of this cave is that the fossils not only consist of molars, but also bones are present.

Based on the presence of orangutan and other elements of the tropical rainforest of the Sumatran caves, Punung, and Niah on one side, and the similar fauna of the caves from China, Vietnam, Laos, and Thailand with a similar fauna on the other side, we can deduce that during the middle and late Pleistocene there was a tropical rainforest ecosystem, running from south China, Vietnam, Laos, Thailand, and Myanmar to the islands of the Sunda Shelf. On the continent, the climate became drier and the tropical rainforest ecosystem disappeared. De Vos and Long (2001) deduced that the fauna from Sumatran Caves, Niah (Borneo), and Punung (Java) came from South China by corridor dispersal, the so-called Sino-Malayan route (Figure 1) via the Malaysian area during the late Pleistocene. One of the species that entered the Sunda region at this stage was *H. sapiens*. However, *Stegodon*, *Crocota*, and *Ailuropoda* never reached the Sunda Shelf.

Unbalanced Endemic Island Faunas

Wallacea

The many islands and seas in between Sunda and Sahul have been given the name Wallacea. In contrast to the islands of the Sunda Shelf, these islands were not connected to the mainland at times of low sea levels and show unbalanced endemic island faunas. The same is true for the Philippines.

Sulawesi

Van den Bergh et al. (2001) described the fossil record and the stratigraphic sequence for Sulawesi (Figure 1). In the late Pliocene sedimentary rocks dating to ca. 2.5 Ma, a pygmy *Stegodon* (*S. sompoensis*), a pygmy *Elephas* ('*E. celebensis*'), a strange pig, *Celebochoerus heekereeni*, and a tortoise are present. During the early Pleistocene, a large-sized *Stegodon* possibly immigrated to south Sulawesi. By the middle or late Pleistocene, both pygmy proboscideans and the tortoise had become extinct, while the large-sized *Stegodon* and *Celebochoerus* continued, or, alternatively, a new immigration of a large *Stegodon* took place. In this period, there is also a new immigration of a highly advanced *Elephas* species.

Flores

Van den Bergh et al. (2001) describe the fossil record and biostratigraphy of Flores (Figure 1). Two members are present in the so-called Ola Bula Formation: the older Member A and the younger Member B. The fossil locality of Tangi Talo in Member A contains an unbalanced endemic island fauna with the pygmy *S. sondaari*, and a tortoise *Geochelone*, which is on average much smaller than the fossil giant tortoises known from the other islands. Besides these two species, remains of *Varanus komodoensis* have been recovered from the same layer. In Member B, in the sites of Mata Menge and Boa Leza, there is another unbalanced endemic island fauna consisting of the relatively large (but still smaller-than-continental forms) *S. florensis*, the murid *H. nusatenggara*, and *V. komodoensis*. This younger fauna is associated with artefacts at the localities Mata Menge and Boa Leza, indicating the co-occurrence with humans.

Paleomagnetic dating results suggest a Late-early Pleistocene age for the Tangi Talo fauna and early-middle Pleistocene for the Mata Menge fauna (Sondaar et al., 1994), indicating an age of at least 0.7 Ma for the artefact-bearing layer. Fission track ages of stone tools by Morwood et al. (1998) date 0.88 ± 0.07 and 0.80 ± 0.07 Ma.

Timor

On Timor (Figure 1), Verhoeven (1964) discovered the first remains of a dwarf *Stegodon* at Weaiwe. The posterior part of an upper third molar of the pygmy *S. timorensis* was described and illustrated by Sartono (1969). Additional molar material from this Timor species was described and figured by Hooijer (1969), who also reported a large-sized *Stegodon* from the island, represented by a slightly worn deciduous upper third from Sadilaun. Hooijer classified this fossil as *S. timorensis* subspecies D. More material of both the dwarf and the large-sized *Stegodon* was collected during a 1970 expedition in which Hooijer participated. Hooijer (1972) described molar remains as well as postcranial elements obtained during this expedition. The stratigraphical context of the fossils, however, is not clear.

Sumba

Sartono (1979) announced the discovery of a pygmy *Stegodon* from Sumba (Figure 1). Though the stratigraphic context is not clear, the presence of a pygmy *Stegodon* suggests that Sumba was an island.

Sangihe

The small island of Sangihe (Figure 1) lies between the northern tip of north Sulawesi and the island of Mindanao of the Philippines. In 1989, F. Aziz and Shibazaki collected some stegodont material from the island, including an upper tusk and some molar remains. All material originates from the Pintareng Formation on the southeast of Sangihe Island. The age of this formation is thought to be Pleistocene (Samodra, 1989). The material was briefly described and figured by Aziz (1990) and attributed to *Stegodon* sp. B cf. *trigonocephalus*.

Philippines

Only a few mammal fossils are known and described from the Philippines (de Vos and Bautista, 2003; Figure 1). Based on the size and morphology of the molars, there is only one species of *Stegodon* present, namely *S. luzonensis* and a large species *Elephas*. *S. luzonensis* is a little smaller than the continental form. Postcranial elements reveal the presence of both a small and a large proboscidean. In addition, there is a relatively small rhino, *Rhinoceros philippinensis*, and remains of a giant tortoise (de Vos and Bautista, 2003). This faunal composition indicates clearly that it is an endemic island fauna but its stratigraphical position is not well known. Archaeological investigations by Fox and Paralta (1974) suggested that the faunal remains are of middle Pleistocene age and that at least some of the tools found are coeval with the fossils. This may reflect a similar succession to that observed in Flores: first a pygmy proboscidean and a giant tortoise, followed by a large proboscidean and artefacts.

See also: Vertebrate Overview. Vertebrate Records: Late Pleistocene of Southeast Asia; Early Pleistocene.

References

- Aigner JS (1978) Pleistocene faunal and cultural stations in south China. In: Ikawa-Smith F (ed.) *Early Paleolithic in South and East Asia*, pp. 129–160. Mouton: The Hague.
- Aziz F (1990) Pleistocene Mammal Faunas of Sulawesi and their Bearings to Paleozoogeography, pp. 1–98. DPhil Dissertation, Kyoto University.
- Badoux DM (1959) *Fossil mammals from two deposits at Punung (Java)*, pp. 1–151. PhD Thesis, Utrecht University.
- Beden M and Guérin C (1973) Le gisement de vertébrés du Phnom Loang (Province de Kampot, Cambodge). Faune Pléistocène moyen terminal (Loangien). *Travaux et Documents de l'Office de la Recherche Scientifique et Technique Outre-Mer Paris* 27: 6–97.
- Bien MN and Chia LP (1938) Cave and rock-shelter deposits in Yunnan. *Bulletin of the Geological Society of China* 18: 325–348.
- Brown P, Sutikna T, Morwood MJ, et al. (2004) A new small-bodied hominin from the Late Pleistocene of Flores, Indonesia. *Nature* 431: 1055–1061.
- Ciochon R, Olson J, and James J (1990) Other Origins: The Search for the Giant Ape in Human Prehistory, pp. 1–262. Iowa: Bantam Books.
- Ciochon R, Long VT, Larick R, et al. (1996) Dated co-occurrence of *Homo erectus* and *Gigantopithecus* from Thâm Khuyen Cave, Vietnam. *Proceedings of the National Academy of Science USA* 93: 3016–3020.
- Colbert EH (1935) Siwalik mammals in the American Museum of Natural History. *Transactions of the American Philosophical Society New Series* 26(i–x): 1–401.
- Colbert EH (1943) Pleistocene vertebrates collected in Burma by the American Southeast Asiatic expedition. *Transactions of the American Philosophical Society New Series* 32(3): 395–429.
- Colbert EH and Hooijer DA (1953) Pleistocene mammals from the limestone fissures of Szechuan, China. *Bulletin of the American Museum of Natural History* 102: 1–134.
- Cuong NL (1971) After the excavations of Hang Hum, Thâm Kuyen and Keo Leng caves. *Archaeology (Hanoi)* 11/12: 7–11 (in Vietnamese).
- Cuong NL (1985) Fossile Menschenfunde aus Nordvietnam. In: Herrmann J and Ullrich H (eds.) *Menschwerdung – Biotischer und Gesellschaftlicher Entwicklungsprozess*, pp. 96–102. Berlin: Akademie Verlag.
- de Vos J and Bautista A (2003) Preliminary notes on the Vertebrate fossils from the Philippines. *Proceedings of the Society of Philippine Archaeologists*, vol. 1, 42–62. Semantic and Systematics, Philippine Archaeology.
- de Vos J and Long Vu (2001) First settlements: Relations between continental and Insular Southeast Asia. In: Sémah F, Falguères C, Grimaud-Hervé D, and Sémah A-M (eds.) *Origine des Peuplements et Chronologie des Cultures Paléolithiques dans le Sud-est Asiatique*, pp. 225–249.
- de Vos J, Demeter F, Tougaard C, and Bacon A-M (in prep). An updating of the fauna from Tam Hang (Laos).
- Drawhorn GM (1994) *The Systematics and Paleodemography of Fossil Orangutans*, pp. 1–232. PhD Thesis, University of California, Davis.
- Fox RB and Peralta JT (1974) Preliminary report on the palaeolithic archaeology of Cagayan Valley, Philippines and the Cabalwanian industry. *Proceedings of the first Regional Seminar on South East Asian Prehistory and Archaeology*, pp. 100–147. Manila: National Museum of the Philippines.
- Hooijer DA (1948) Prehistoric teeth of man and of the orang-utan from central Sumatra, with notes on the fossil orang-utan from Java and Southern China. *Zoologische Mededelingen uitgegeven door het Rijksmuseum van Natuurlijke Historie te Leiden* 29: 175–301.
- Hooijer DA (1969) The *Stegodon* from Timor. *Proceedings of the Koninklijke Nederlandse Akademie van Wetenschappen, Amsterdam, series B* 72(3): 201–210.
- Hooijer DA (1972) *Stegodon trigonocephalus florensis* Hooijer and *Stegodon timorensis* Sartono from the Pleistocene of Flores and Timor. *Proceedings of the Koninklijke Nederlandse Akademie van Wetenschappen, Series B* 75(1): 12–33.
- Hussain ST, van den Bergh GD, Steensma KJ, et al. (1992) Biostratigraphy of the Plio-Pleistocene continental sediments (Upper Siwaliks) of the Mangla-Samwal Anticline, Azad Kashmir, Pakistan. *Proceedings of the Koninklijke Nederlandse Akademie van Wetenschappen, Series B* 95(1): 65–80.
- Kahlke HD (1961a) Zur chronologischen Stellung der südchinesischen *Gigantopithecus*-Funde. *Zeitschrift für Wissenschaftliche Zoologie* 165: 47–80.
- Kahlke HD (1961b) On the Complex of the *Stegodon-Ailuropoda*-Fauna of South China and the chronological position of *Gigantopithecus blacki* Von Koenigswald. *Vertebrata Palasiatica* 2: 83–108.
- Kahlke HD (1968) Zur relativen Chronologie ostasiatischer Mittelpleistozän-Faunen und Hominoidea-Funde. In: Kurth G (ed.) *Evolution und Hominisation*, pp. 91–118. Stuttgart: Fischer Verlag.
- Kahlke HD (1972) A review of the Pleistocene history of the Orang-Utan (Pongo) Lapepe, 1799. *Asian Perspectives* 15: 5–14.
- Lekagul B (1949) The excavation of the fossil *Hippopotamus*, *Stegodon insignis*, at Nakon Sawan. *Science Conference 1949*, pp. 91–118. Thailand.
- Linghong W and Lian O (1982) Subdivision of *Ailuropoda-Stegodon*-fauna by applying cluster analysis. *Vertebrata Palasiatica* 20(3): 263.
- Medway L (1979) The Niah Excavations and Assessment of the Impact of Early Man on mammals in Borneo. *Asian Perspectives* 20(1): 51–69.
- Morwood MJ, O'Sullivan PB, Aziz F, and Raza A (1998) Fission-track ages of stone tools and fossils on the east Indonesian island of Flores. *Nature* 392: 173–176.
- Morwood MJ, Soejono RP, Roberts RG, et al. (2004) Archaeology and age of a new hominin from Flores eastern Indonesia. *Nature* 431: 1087–1091.
- Pei WC (1957) Discovery of lower jaws of Giant Ape in Kwangsi, South China. *Science Records New Series* 1(3): 49–52.
- Samodra H (1989) Kedudukan stratigrafi temuan fosil *Stegodon* di pulau Sangihe (in Indonesian). *Paper presented at the XVIII annual meeting of the A.I.G., Yogyakarta, December 1989*.
- Sartono S (1969) *Stegodon timorensis*: A pygmy species from Timor (Indonesia). *Proceedings of the Koninklijke Nederlandse Akademie van Wetenschappen, Series B* 72: 192–202.
- Sartono S (1979) The discovery of a pygmy *Stegodon* from Sumba, East Indonesia: An announcement. *Modern Quaternary Research in SE Asia* 5: 57–63.
- Sondaar PY, van den Bergh GD, Bondan M, Aziz F, de Vos J, and Unkai LB (1994) Middle Pleistocene faunal turnover and colonization of Flores (Indonesia) by *Homo erectus*. *Comptes Rendus de la Académie des Sciences Paris* 319 Série II, 1255–1262.
- Soniaka A (1985) Early *Homo* from Narmada valley, India. In: Delson E (ed.) *Ancestors: The Hard Evidence*, pp. 334–338. New York: Alan Liss.
- Storm P (2001) The evolution of humans in Australasia from an environmental perspective. In: Dam RAC and van der Kaars S (eds.) *Quaternary Environmental Change in the Indonesian Region, Palaeogeography, Palaeoclimatology, Palaeoecology*, pp. 363–383. Amsterdam: Elsevier Science B.V.
- Storm P, Aziz F, de Vos J, et al. (2005) Late Pleistocene *Homo sapiens* in a tropical rainforest fauna in East Java. *Journal of Human Evolution* 49: 536–545.
- Tougaard C (1998) *Les Faunes de Grands Mammifères du Pléistocène Moyen Terminal de Thaïlande dans leur Cadre Phylogénétique, Paléocécologie et Biochronologique. Thèse de Doctorat*, pp. 1–175. Université de Montpellier II.
- Van den Bergh GD, de Vos J, and Sondaar PY (2001) The late Quaternary palaeogeography of mammal evolution in the Indonesian Archipelago. *Palaeogeography, Palaeoclimatology, Palaeoecology* 171: 385–408.
- Verhoeven Th (1964) *Stegodon*-Fossilien auf der Insel Timor. *Anthropos* 59: 634.
- Wanpo H, Ciochon R, Yumin G, et al. (1995) Early *Homo* and associated artefacts from Asia. *Nature* 378(1): 275–278.
- Yanxian Li (1981) On the subdivisions and evolution of the quaternary mammalian faunas of South China. *Vertebrata Palasiatica* 19(1): 75–76.

Late Pleistocene of Africa

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Introduction

The late Pleistocene vertebrate record of Africa plays a key role in our understanding of human evolution and the development of modern biotic environments. Africa's large landmass and location on the equator helped support prolific radiations of large terrestrial mammals, and this past mammalian diversity has survived to modern times to a greater extent here than on other continents. Furthermore, Africa is truly the cradle of humanity, with many significant steps in human evolution occurring here, including the origins of physically and behaviorally modern humans during the late Pleistocene (see [Global Expansion 300 000–8000 Years Ago, Africa](#)). This article explores the late Pleistocene vertebrate record of Africa, spanning 126–11.5 ka, paying particular attention to large terrestrial mammals, but also considering smaller vertebrates. Because of the significance of humans in the late Pleistocene of Africa, much of what we know about the other vertebrates from this time comes from archaeological assemblages (see [Interactions with Hominids; Overview; Global Expansion 300 000–8000 Years Ago, Africa](#)). [Figure 1](#) provides the geographic location for each fossil sample discussed below, and [Table 1](#) provides the relevant references. [Table 2](#) provides the scientific names for all the species whose common names are used below.

Modern Physical and Environmental Setting

Africa's large landmass, over 30 million km² or one-fifth of the land surface of Earth, and its position straddling the equator means that it contains a diversity of habitats, which support highly diverse vertebrate communities. The continent is characterized by moderately elevated tablelands. In general, the east and south are higher, and elevation decreases moving north and west. The Atlas Mountains in the far northwest are an exception. This physical geography greatly influences the distribution of moisture, and therefore influences plant communities and ultimately the faunal communities. In general, Africa is a warm and dry continent, and patterns of rainfall largely determine vegetation patterns because of the plants' dependence on water availability ([Figure 2](#)). Annual rainfall is distributed broadly in a western pointing horseshoe-shaped pattern centered on the equator, although rainfall diminishes in the eastern highlands. For the most part, rain falls during the summer. The Cape and Mediterranean littorals are exceptions, with dry summers and wet winters creating unique ecosystems. The dominating desert, the large, extremely dry Sahara, lays to the north of this horseshoe, but deserts are also found on the African Horn and the southwestern coast. Savanna and open woodland cover about 40 percent of the continent, mainly south of the Sahara ([Turner and Antón 2004](#)). Heavy rains in

the tropical highlands create extensive river systems that bring life-giving water to otherwise dry areas. In Africa, the availability of surface drinking water is the single most limiting climatic influence for the majority of mammals ([Kingdon 1997: 449](#)).

Africa's physical geography and climate have not always been this way. Major changes caused by movements of continents and the resulting rise of mountains, rifting, and volcanoes have created profound variations in the deep past. However, glacial cycles and the related changes in sea levels and precipitation have had the greatest effect on late Pleistocene African vertebrate communities ([Link: 121, 389](#)). [Kingdon \(1997: 455\)](#) summarizes the pattern of change as occurring along two poles (see also [Adams \(1997\)](#)). During cool, dry times, arid habitats have expanded north and south, contracting again when more moisture is available; during warm, wet periods, tropical forests expanded east and west along the equator, again contracting during cooler, drier times ([Figure 2](#)). These changes have resulted in the repeated fragmentation of habitats, isolating plants and animals but ultimately increasing biodiversity.

The Fossil Records

Southern Africa

For the most part, an essentially modern fauna was in place in southern Africa by about 270 ka. This faunal community persisted until the Pleistocene to Holocene boundary 11.5 ka and is sometimes referred to as the Florisian Land Mammal Age ([Klein, 1984a](#)). However, a more precise dating of the transition from the more archaic middle mid-Pleistocene assemblages (sometimes referred to as the Cornelian Land Mammal Age) to more modern late mid-Pleistocene and late Pleistocene assemblages remains elusive (see [Mid-Pleistocene of Africa](#)), and in eastern Africa, it appears that a modern fauna was in place by 330 ka ([Potts and Deino, 1995](#)). In southern Africa, the deposits at Duinefontein 2 (DFT2) appear to straddle this transition. The upper horizon at DFT2 contains an essentially modern fauna and has been dated to ca. 270 ka using optically stimulated luminescence (OSL) (see [Optical Dating](#)). The lower horizon at the site contains an extinct alcelaphine antelope, characteristic of more archaic assemblages. OSL indicates that the sands between the two horizons accumulated 290 ka, providing a minimum age for the more archaic assemblage. The nearby site of Elandsfontein Main (EFTM) contains a very large assemblage of vertebrate remains, a high percentage of which are extinct. Among the bovids alone, when extinct subspecies are counted, 89% (16/18) of the EFTM taxa are now extinct, while only 20% (4/20) of the species in nearby late Pleistocene faunas are extinct; even when a more conservative approach is taken and only species are compared, 67% (12/18) of the EFTM species are extinct while only 14% (3/21) of the late Pleistocene species are extinct. The EFTM deposits



Figure 1 Map of the sites discussed in the text. **Table 1** provides references.

are not suitable for direct radiometric dating, so their age can be estimated only by comparing species presence and absence with East African deposits, which are more readily radiometrically dated (see [K/Ar](#) and [\$^{40}\text{Ar}/^{39}\text{Ar}\$ Dating](#)). This comparison indicates that the archaic fauna at EFTM accumulated sometime between 600 ka and 1 Ma. These results indicate that an essentially modern faunal community came into place in southern Africa sometime between 270 ka and 400 ka. After 270 ka, in the bovids, only subtle changes occurred. At DFT2, kudu horns were smaller and more tightly spiraled, and black wildebeest horns had smaller frontal bosses and failed to pass downward as they pass forward; by the late Pleistocene, the animals had evolved to more closely resemble today's animals. Once the essentially modern faunal community came into place during the late mid-Pleistocene, no extinctions of large mammals occurred until the very end of the late Pleistocene (see [Late Pleistocene Megafaunal Extinctions](#)).

Throughout southern Africa, it is possible to identify changes in mammalian faunal communities in response to late Pleistocene glacial cycles, but these changes are best documented along the southern and western coasts of South Africa, where good contextual data are available on large, well-documented faunal assemblages ([Klein, 1980](#)). The most notable aspect of the late Pleistocene faunas from this region is the overwhelming abundance of grazing species, especially alcelaphines and equids, compared to the late Holocene and historic fauna, which is dominated by browsers and mixed feeders, indicating that grasses were much more common during the late Pleistocene than at the present. This contrast is most striking at Nelson Bay Cave, on the southern coast of South Africa. The faunal assemblages from the levels spanning 18.5–12 ka are dominated by grazing ungulates, including quagga, alcelaphines, long-horned buffalo, and springbok, as well as ostrich, which also prefers open environments; none of these taxa were present in the vicinity of the site historically.

Table 1 List of sites discussed in the text and references for additional information about each site

Site	Country	References
Bir Tarfawi	Egypt	Gautier (1993)
Boomplaas	South Africa	Klein (1978a)
Border Cave	South Africa	Klein (1977); Avery (1992)
Cave of Hearths	South Africa	Cooke (1963); Latham and Herries (2004)
Duinefontein 2	South Africa	Klein et al. (1999); Feathers (2002); Cruz-Urbe et al. (2003)
Elands Bay	South Africa	Klein and Cruz-Urbe (1987)
Elandsfontein Main	South Africa	Klein and Cruz-Urbe (1991); Klein et al. (in preparation)
Florisbad	South Africa	Cooke (1963); Grün et al. (1996)
Haua Fteah	Libya	Klein and Scott (1986)
Jebel Irhoud	Morocco	Thomas (1981); Grün and Stringer (1991); Amani and Geraads (1993); Amani and Geraads (1998)
Klasies River	South Africa	Klein (1976); Avery (1987)
Lainyamok	Kenya	Potts and Deino (1995)
Lukunya Hill	Kenya	Marean and Gifford-Gonzalez (1991); Marean (1992)
Makhadma 2 and 4	Egypt	Vermeersch (ed.) (2000)
Mbi Crater	Cameroon	Cornelissen (2002)
Mugharet el 'Aliya	Morocco	Wrinn (in preparation); Wrinn and Rink (2003)
Nelson Bay Cave	South Africa	Klein (1972)
Mumbwa Cave	Zambia	Klein and Cruz-Urbe (2000)
Redcliff Cave	Zimbabwe	Klein (1978b); Cruz-Urbe (1983)
Rhafas Cave	Morocco	Michel (1992)
Shuwikhat 1	Egypt	Vermeersch (ed.) (2000)
Sibudu Cave	South Africa	Plug (2004)

Table 2 List of the common names of animals discussed in the text, along with their scientific names

Common name	Scientific name
Mammals	Mammalia
Monkeys, apes, and allies	Primates
Human	<i>Homo sapiens</i>
Baboon	<i>Papio</i> sp.
Lemurs	Lemuridae
Bats	Chiroptera
Insectivores	Insectivora
Tenrecs	Lipotyphla
Hares, pikas, and rabbits	Lagomorpha
Cape hare	<i>Lepus capensis</i>
Scrub hare	<i>Lepus saxatilis</i>
European rabbit	<i>Oryctolagus cuniculus</i>
Rodents	Rodentia
North African crested porcupine	<i>Hystrix cristata</i>
Cape porcupine	<i>Hystrix africaeaustralis</i>
Giant gerbil*	<i>Gerbillus grandis</i>
Cape dune molerat	<i>Bathyergus suillus</i>
Common molerat	<i>Cryptomys hottentotus</i>
Carnivores	Carnivora
Dogs	Canidae
Golden jackal	<i>Canis aureus</i>

Table 2 (Continued)

Common name	Scientific name
Black-backed jackal	<i>Canis mesomelas</i>
Red fox	<i>Vulpes vulpes</i>
Hunting dog	<i>Lycan pictus</i>
Weasels, badgers, and polecats	Mustelidae
European polecat	<i>Mustela putorius</i>
Genets and civets	Viverridae
Mongoose	Herpestidae
Egyptian mongoose	<i>Herpestes ichneumon</i>
Hyenas	Hyaenidae
Striped hyena	<i>Hyaena hyaena</i>
Brown hyena	<i>Parahyaena brunnea</i>
Spotted hyena	<i>Crocuta crocuta</i>
Cats	Felidae
Wild cat	<i>Felis sylvestris</i>
Sand cat	<i>Felis margarita</i>
Serval	<i>Leptailurus serval</i>
Caracal	<i>Caracal caracal</i>
Leopard	<i>Panthera pardus</i>
Lion	<i>Panthera leo</i>
Bears	Ursidae
Brown bear	<i>Ursus arctos</i>
Fur seals and sea lions	Otariidae
Cape fur seal	<i>Arctocephalus pusillus</i>
Hyraxes	Hyracoidea
Hyrax	<i>Procavia</i> sp.
Elephants	Proboscidea
African elephant	<i>Loxodonta africana</i>
Odd-toed ungulates	Perissodactyla
Zebras, horses, and relatives	Equidae
Burchell's (plains) zebra	<i>Equus burchellii</i>
Quagga*	<i>Equus quagga</i>
Mountain zebra	<i>Equus zebra</i>
Grevy's zebra	<i>Equus grevyi</i>
Giant Cape zebra*	<i>Equus capensis</i>
Rhinoceroses	Rhinocerotidae
Black rhinoceros	<i>Diceros bicornis</i>
White rhinoceros	<i>Ceratotherium simum</i>
Merck's rhinoceros*	<i>Dicerorhinus kirchbergensis</i>
Even-toed ungulates	Artiodactyla
Hippopotamuses	Hippopotamidae
Hippopotamus	<i>Hippopotamus amphibius</i>
Madagascan hippopotamus*	<i>Hippopotamus laloumena</i>
Madagascan dwarf hippopotamus*	<i>Hippopotamus lemerlei</i>
Madagascan pygmy hippopotamus*	<i>Hexaprotodon madagascariensis</i>
Pigs and hogs	Suidae
Boar	<i>Sus scrofa</i>
Bushpig	<i>Potamochoerus larvatus</i>
Giant hog	<i>Hylochoerus meinertzhageni</i>
Warthog	<i>Phacochoerus africanus</i>
Giant warthog*	<i>Metridiochoerus/Stylochoerus</i>
Deer	Cervidae
Camels, llamas, and relatives	Camelidae
Wild camel*	<i>Camelus thomasi</i>
Giraffes and okapis	Giraffidae
Giraffe	<i>Giraffa camelopardalis</i>
Bovids, antelopes, gazelles, and relatives	Bovidae
Oxen	Bovini

(Continued)

Table 2 (Continued)

Common name	Scientific name
Cape/African buffalo	<i>Syncerus caffer</i>
Long-horned buffalo*	<i>Pelorovis antiquus</i>
Aurochs	<i>Bos primigenius</i>
Spiral-horned bovids	Tragelaphini
Bushbuck	<i>Tragelaphus scriptus</i>
Kudu	<i>Tragelaphus strepsiceros</i>
Eland	<i>Taurotragus oryx</i>
Duikers	Cephalophini
Bush duiker	<i>Sylvicapra grimmia</i>
Blue duiker	<i>Cephalophus monticola</i>
Dwarf antelopes	Neotragini
Dwarf antelope	<i>Neotragus batesi</i>
Grysbok	<i>Raphicerus melanotis</i>
Steenbok	<i>Raphicerus campestris</i>
Oribi	<i>Ourebia ourebi</i>
Reduncines and kobs	Reduncini
Mountain reedbuck	<i>Redunca fulvorufula</i>
Reedbuck	<i>Redunca redunca</i>
Lechwe	<i>Kobus leche</i>
Waterbuck	<i>Kobus ellipsiprymnus</i>
Gazelline antelopes	Antilopini
Gazelles	<i>Gazella</i> sp.
Thomson's gazelle	<i>Gazella rufifrons</i>
Springbok	<i>Antidorcas marsupialis</i>
Bond's springbok*	<i>Antidorcas bondi</i>
Southern springbok*	<i>Antidorcas australis</i>
Hartebeest, wildebeest and allies	Alcelaphini
Impala	<i>Aepyceros melampus</i>
Blesbok/bontebok	<i>Damaliscus dorcas</i>
Korrigum	<i>Damaliscus lunatus</i>
Hartebeest	<i>Alcelaphus buselaphus</i>
Giant hartebeest*	<i>Megalotragus priscus</i>
Blue wildebeest	<i>Connochaetes taurinus</i>
Black wildebeest	<i>Connochaetes gnou</i>
Horse-like antelopes	Hippotragini
Roan antelope	<i>Hippotragus equinus</i>
Sable antelope	<i>Hippotragus niger</i>
Blue antelope*	<i>Hippotragus leucophaeus</i>
Oryx	<i>Oryx gazella</i>
Sheep and goats	Caprini
Barbary sheep	<i>Ammotragus lervia</i>
Reptiles	Reptilia
Angulate tortoise	<i>Chersina angulata</i>
Spur-thighed tortoise	<i>Testudo graeca</i>
Birds	Aves
Penguin	<i>Spheniscus demersus</i>
Cormorant	<i>Phalacrocorax</i> sp.
Flightless birds	Struthioniformes (ratites)
Ostrich	<i>Struthio camelus</i>
Elephant or marsh birds*	<i>Aepyornis maximus</i>

*Indicates extinct species.

However, the dominant taxa of the younger deposits include bushpig, bushbuck, grysbok, and Cape buffalo, which also were found historically in the region of the site. A similar transition occurred at Melkoutboom Cave, located inland northeast of Nelson Bay Cave, where the terminal Pleistocene fauna was also dominated by equids (either mountain zebra or quagga or both) and alcelaphines, while the Holocene deposits resembled the modern local fauna with bushbuck, kudu, blue

duiker, grysbok, and bushpig. Again, a similar change is seen on the western coast of South Africa in Elands Bay Cave where the terminal Pleistocene fauna contains relatively more large grazers than the late Holocene component. All of these assemblages indicate that the faunal communities underwent a marked transition as environments changed from the terminal Pleistocene to the Holocene.

A similar contrast is found during the more ancient periods of the late Pleistocene at the site of Klasies River, further east along the coast from Nelson Bay Cave. Similar to Nelson Bay Cave, the historic fauna included bushpig, bushbuck, grysbok, and Cape buffalo, and also blue duiker and hartebeest. These taxa are also found in the Holocene levels of the site, as well as in the oldest deposits, which possibly correspond to oxygen isotope stage (OIS) 5e, the Last Interglacial, which began 126 ka (see [Oxygen-Isotope Stratigraphy of the Oceans](#)). As the sequence progresses through OIS 5, the relative abundance of taxa that prefer open habitats (alcelaphines and equids) fluctuates with those that prefer more closed habitats (bushbuck and grysbok). Further to the northeast at Border Cave, again a similar contrast is apparent, but the historic fauna is more characteristic of an open environment, with warthog, Burchell's zebra, and alcelaphines. These grazers are also present during the warmer times between OIS 5d and 3, but during colder times, a fauna that prefers more bush is present with bushpig, Cape buffalo, tragelaphines, and impala. Klasies River and other coastal sites also show evidence of changing sea levels and sea temperatures during the late Pleistocene, as seen in the changing abundances and taxa of marine birds and mammals, especially penguins, cormorants, and Cape fur seals. During warmer times, when the seas were higher, these animals are common, but they are rare during cold times when sea levels were lower, such as during the terminal Pleistocene at Nelson Bay Cave and Elands Bay Cave.

In the northern part of southern Africa (Namibia, Botswana, Zimbabwe, Mozambique, and Angola), less is known about the late Pleistocene faunal community. Contrasts between glacial and interglacial communities are not as clear, because the record is more sparse and because regions closer to the equator likely experienced less climatic and therefore faunal changes than southern South Africa. The late (and mid-) Pleistocene record from Zambia is better known ([Brooks, 2003](#)). Here, fossil assemblages, such as from Mumbwa Cave, closely resemble the modern faunal community with an abundance of Burchell's zebra, warthog, and alcelaphines; any deviation from the modern pattern is in the direction of adding even more grazing ungulates, such as blesbok, springbok, and mountain reedbuck, and bovids that are tied to water, such as lechwe and waterbuck. At Redcliff Cave, Zimbabwe, blesbok, springbok, and mountain reedbuck are more common during what appears to be moister conditions in OIS 2 and 4, but these animals were not documented in the region historically. In eastern South Africa, the historic and late Pleistocene faunas are dominated by grazers, but during the moister times of the late Pleistocene, lechwe, waterbuck, Cape buffalo, impala, and roan antelope appear, suggesting open woodlands were present. Blue wildebeest, sable/roan antelope, waterbuck, impala, and hartebeest in the late Pleistocene deposits of Sibudu Cave, near the coast in KwaZulu-Natal, South Africa, indicate that more open savannas existed there during ancient times than during

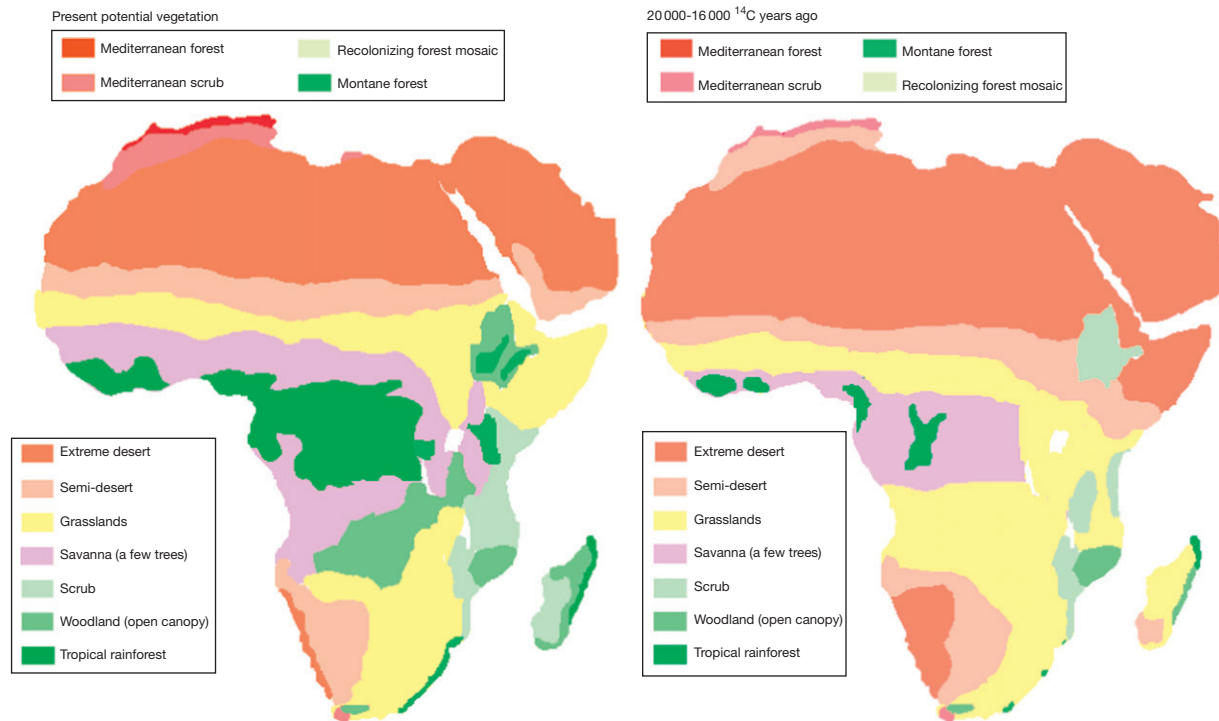


Figure 2 Distribution of the major vegetation zones of Africa in the present day (left) and during the Last Glacial Maximum (right). Note that under the colder, drier glacial conditions, the central forests contract east-west while the deserts expand north-south. Maps from Adams (1997) accessible at <http://www.esd.ornl.gov/ern/qen/nerc.html>.

the present. In sum, the data indicate that open grasslands were much more common in southern Africa during the late Pleistocene than they have been historically.

The above discussion has focused on bovids, equids, and suids, because these taxa are most common in the fossil record and because changes in their distributions and abundances more closely follow changes in vegetation. Carnivores have more widespread distributions than ungulates, reflecting their more generalized habitat requirements (Turpie and Crowe, 1994). Throughout the late Pleistocene of southern Africa, lions, brown and spotted hyenas, black-backed jackals, and wild cats are frequently found in fossil assemblages, along with occasional leopards, servals, caracals, hunting dogs, and a variety of mustelids and viverrids. Of the smaller mammals, typical species include baboons, hyraxes, hares, porcupines, and Cape dune moles, when the local environments are suitable. Hippopotamuses were common throughout southern and eastern Africa wherever permanent water and open grazing were available, both historically and in the late Pleistocene. Both black and white rhinoceroses make regular appearances in late Pleistocene assemblages, along with occasional elephants. Historically, giraffes were found throughout the central and northern parts of southern Africa, especially where savannas and open woodlands were common, but they are rare in the fossil record. A few examples are known, as in Florisbad and Cave of Hearths, but their chronological position of these fossils within the late mid- or late Pleistocene is unclear. The presence of giraffe in Florisbad especially indicates that past environments were very different from the historically treeless local vegetation. An example of late Pleistocene giraffe

more securely dated to OIS 3 was found in Sibudu Cave, KwaZulu-Natal, South Africa, located in an environment that has not supported giraffes in the recent past, but may have prehistorically.

In sum, the mammalian assemblages from the late Pleistocene of southern Africa closely resemble the modern-day fauna of the region, but the distribution of the taxa changed through time in response to climatic influences on vegetation. Although almost all the taxa found historically in southern Africa are seen in the fossil record, there are six exceptions: giant Cape zebras, giant warthogs, long-horned buffalo, giant hartebeest, Bond's springbok, and southern springbok (Figure 3). Although these taxa appeared throughout southern Africa during the late Pleistocene (the southern springbok appeared only on the southern and western coasts of South Africa, while Bond's springbok was everywhere else), the best sites for determining the dates of their extinctions come from the coasts of South Africa. In this region, long-horned buffalo, giant hartebeest, southern springbok, and Cape zebras are last seen between 12 and 9.5 ka (Klein, 1984b). These terminal Pleistocene dates coincide with the major environmental changes described above that occurred with the transition to the Holocene, but hunting by humans may have also played a role (global evidence reviewed in Barnosky et al. (2004)) (see *Late Pleistocene Megafaunal Extinctions*). Over-hunting by humans almost certainly caused the next set of extinctions in southern Africa: the last blue antelope was last seen around 1800 and the last-known quagga died in 1883 (Klein, 1987; Rau, 1986). However, recently studies of DNA preserved in a few of their bones and skins stored in museums have provided some data

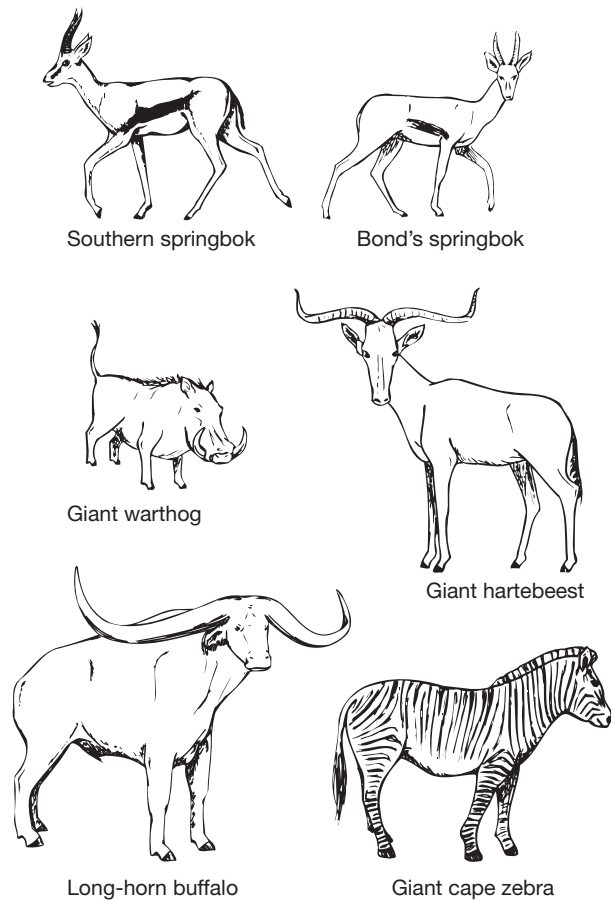


Figure 3 Depictions of African mammals that went extinct at the end of the late Pleistocene. Drawings are courtesy of Richard G. Klein.

about their taxonomic status and phylogenetic relationships (Higuchi et al., 1984; Leonard et al., 2005; Robinson et al., 1996) (see [Ancient DNA](#)).

Compared to other regions of Africa, fossil samples from southern Africa are large enough and frequent enough to examine changes in body size within a species through time and across. This is done for two reasons. First, the average size of individuals in a species may vary in relation to temperature and precipitation, thereby being useful paleoenvironmental reconstruction. Size variation is examined between geographic locations of modern populations to understand the causes of the variation. Then these parameters can be inferred from the size of fossil specimens. This variation in size has been studied in carnivores (Klein, 1986), rock hyrax (Klein and Cruz-Uribe, 1996), Cape dune molerats (Klein, 1991), and common molerats (Avery, 2004). The second reason for studying size variation is because it may provide information about human predation pressure on a species, and from this, it is possible to infer human population densities. This has been examined using the angulate tortoise. Tortoises grow continuously throughout their lives, and if they frequently preyed upon, their median size will decrease. Studies of tortoise size from archaeological assemblages show that late Pleistocene tortoises typically were larger than Holocene tortoises (Klein and Cruz-Uribe, 1983; Steele and Klein, in press).

Northern Africa

Most of our knowledge about late Pleistocene vertebrates from northern Africa comes from coastal regions of the Maghreb, including Morocco, Algeria, Tunisia, and Libya, although Egypt is also well represented. One of the challenges to understanding the vertebrate record from this region is the lack of good chronological control. However, recent efforts to explore the late Pleistocene archaeological record of especially Morocco is changing this, and hopefully this challenge will be overcome soon. The northern African paleontological record is interesting because it contains elements of Eurasian faunas, as well as traditional African species. These immigrants include bears, cervids, and boars. The European component almost certainly entered northern Africa from the northeastern corner through what is today Egypt. An analysis of the zoogeographical affinities of mammals in the Maghreb showed that only bats were able to cross the Straits of Gibraltar and colonize from the north; all other taxa moved in from the east or crossed the Sahara from the south (Dobson and Wright, 2000).

As in southern Africa, the late Pleistocene species from northern Africa closely resemble modern species. However, it is unclear when exactly this fauna came into place, as very few reliable samples exist from the later mid-Pleistocene, when the change is documented in eastern and southern Africa (Geraads, 2002). The site of Jebel Irhoud, Morocco, lacks the later arriving cervids and boars and contains an alcelaphine with simple teeth and a large gerbil, indicating a late mid-Pleistocene age, of perhaps 150 ka (see [Mid-Pleistocene of Africa](#)). Electron spin resonance (ESR) has placed the site between 90 and 190 ka old, and it is possible that the material accumulated throughout this time span (see [Electron Spin Resonance Dating](#)). The abundance of alcelaphines, oryx, and gazelles in the sample suggests that the environment was a dry savanna. In contrast, the terminal Pleistocene and early Holocene (12–4 ka) was especially moist, and most of the mammals present in during the late Pleistocene reappeared during the Holocene, so the clear break between late Pleistocene and Holocene records, which is seen in southern Africa, is not as apparent in northern Africa. However, a few species go extinct during the Holocene, including camels, cervids, and long-horned buffalo (Klein, 1984b).

In general, the faunal remains also indicate the presence of savannas in the Maghreb during the late Pleistocene. In Haua Fteah, Libya, in Mugharet el 'Aliya and in Rhafas Cave, Morocco, and in nearby sites, the most abundant taxa are Barbary sheep, aurochs, and various gazelle, but hartebeests and equids are also present. Rhinoceroses (both white and Merck's) and elephants are also found, along with both the Cape hare and European rabbit. More rare species include porcupine, hippopotamus, deer, long-horned buffalo, eland, and blue wildebeest. Both warthogs and boar have been identified. Unfortunately, the current data available for the Maghreb do not allow us to make as detailed comparisons between glacial and interglacial cycles as in southern Africa. The fauna at Haua Fteah suggests that Barbary sheep were more abundant during drier periods and that aurochs was more abundant during wetter ones.

The carnivores of this time period are equally diverse (Aouraghe, 2000), and include golden jackals, red foxes, lions, leopards, wildcats, sand cats, caracals, spotted hyenas,

Egyptian mongooses, and European polecats. Bears and striped hyenas were also present. The bears, foxes, and polecats provide another link to Eurasian faunas that make the Maghreb's faunal community unique. Among nonmammals, spur-thighed tortoises and ostriches are also common, mirroring the assemblages often seen in southern Africa.

Late Pleistocene vertebrates are best known in northern Africa from the Maghreb, because the geology of this region allows for many deep caves, which permit good bone preservation. Unfortunately, most of the late Pleistocene sites from the eastern part of northern Africa are open-air sites, where bone is poorly preserved. What data are available, from sites such as Bir Tarfawi, Shuwikhat 1, and Makhadma 2 and 4 Egypt, indicate that the faunal communities were roughly similar between the western and eastern regions of northern Africa. Various gazelles are present, along with hartebeest, aurochs and/or long-horned buffalo. Equids, white rhinoceroses, hippopotamuses, and warthogs were also found in the east, along with porcupines, hares, and ostriches. The mix of carnivores is also similar between the two regions. The eastern region was unique because of the presence of wild camels, giraffes, and reduncines. Some assemblage also include rodents whose modern ranges are much further south in much wetter places, indicating that the fossil assemblages accumulated when the region received higher rainfall than today.

Eastern Africa

While eastern Africa has enjoyed a long history of paleontological research, most of this research has focused on the Pliocene and early Pleistocene, because these phases are associated with the hunt for early hominin fossils (see [Vertebrate Overview](#)). Recent attention on modern human origins during the late Pleistocene is changing this, but large, well-preserved, and well-dated late Pleistocene faunal assemblages remain rare.

The oldest-dated fossil assemblage in eastern Africa that contains no extinct species, essentially a modern fauna, is from Lainyamok, Kenya, and dated to about 330 ka old by Ar/Ar (see [K/Ar and \$^{40}\text{Ar}/^{39}\text{Ar}\$ Dating](#)). This date places the assemblage firmly in the mid-Pleistocene and indicates that modern species existed in eastern Africa throughout the late Pleistocene (see [Mid-Pleistocene of Africa](#)). Some nearby late Pleistocene assemblages, such as those found on Lukenya Hill, Kenya, contain the long-horned buffalo, which was also common in the late Pleistocene assemblages from southern and northern Africa, and may contain another extinct species, a small alcelaphine. However, other authors argue that these small alcelaphines are bontebok or blesbok. Blesbok also are common at Redcliff, suggesting that they once extended from South Africa through Central Africa to eastern Africa. The small alcelaphine is not found in local Holocene samples, but they are present in small numbers in Holocene samples in northern Tanzania. Oryx and Grevy's zebra are also found in the Lukenya faunas but not locally historically, suggesting that in the past, the region was more arid than today and their modern ranges were expanded to encompass Lukenya. Other common late Pleistocene species include hartebeest, wildebeest, Thomson's gazelle, and Burchell's zebra,

and mountain reedbuck, oribi, steenbok, bush duiker, and warthog are also present.

Madagascar

Although Madagascar is typically considered part of Africa, its isolation from Africa for 165 Myr and from India for 88 Myr allowed its flora and fauna to evolve along a unique path. Located almost 400 km from the coast of Africa at Mozambique, very limited faunal exchange occurred until the first humans arrived ca. 2 ka. Only ca. 100 species of mammals lived on the island before colonization by humans, and this faunal community is characterized by high endemism (about 80% of its flora and fauna are unique to the island) and taxonomic imbalance. While the mammalian community included primates (lemurs), insectivores (lipotyphlans), rodents, carnivores, hippopotamuses, and bats, very few of the typically African species ever reached the islands: no elephants, rhinoceroses, camels, giraffes, bovids, felids, or canids successfully made the crossing. Recent genetic studies indicate that most of the clades of nonvolant (nonflying) mammals to reach Madagascar did so with single colonization events, either by swimming or by rafting from Africa; lemurs arrived 62–66 Ma, and carnivores arrived later between 18 and 24 Ma ([Yoder et al., 2003, 1996](#)). However, the presence of three species of hippopotamuses, each with different ancestors, suggests that they may have arrived in three separate colonization events ([Faure and Guérin, 1990; Stuenkel, 1989](#)). Knowledge of the late Pleistocene fossil record of Madagascar is limited, but much work has been done on the subfossil record (terminal Pleistocene and Holocene), especially of lemurs, which provide an impressive example of adaptive radiation in primates. The subfossil record adds 17 extinct species to the ca. 30 still-living lemurs ([Godfrey and Jungers, 2002](#)) and allows the study of their ancient DNA, further confirming their single and ancient origin ([Karanth et al., 2005](#)) (see [Ancient DNA](#)). Among nonmammals, Madagascar was also home to giant tortoises and giant ratites (flightless birds), known as elephant birds, marsh birds, or vorompatra. These taxa probably existed on the land-mass before it broke away from Africa. The long persistence of lemurs and the other unique species on an isolated island and then their rapid and recent extinction has caused much discussion about whether their demise is related to climatic change, human hunting, or human habitat disturbance (deforestation, increased erosion, change in fire regime, introduction of new species) ([Burney, 1999; Goodman and Patterson, 1997](#)) (see [Late Pleistocene Megafaunal Extinctions](#)).

Western and Central Africa

Very little is known about the late Pleistocene vertebrate record from western and central Africa, because bone preservation tends to be poor and because few researchers have conducted paleontological and archaeological research in the region. What little evidence that is available for the region, such as found in Mbi Crater, Cameroon, indicates that during the terminal Pleistocene animals tended to be more characteristic of open habitats, with hartebeest, korrigum, roan antelope, waterbuck, warthog, reedbuck, and bush duiker being present.

Forest cover, while consistently present during the late Pleistocene, increased in the Holocene, as seen by increasing numbers of various duikers, bushbuck, dwarf antelope, bush pig, giant hog, and leopards.

Summary

Most of the mammal species found in the late Pleistocene vertebrate records of Africa were also found on the continent historically. However, many changes in the species' ranges, abundance, and associations make late Pleistocene faunal communities look different from today's communities. As the climate changed in the past, so did the vegetation. Many times in the fossil record, species are found in places where today their descendants live hundreds, if not thousands, of kilometers away, indicating that their past ranges were either completely shifted to different location or expanded to encompass new locations. Late Pleistocene climatic and vegetation changes created unique habitats, many of which do not exist today. Each species' unique responses to these dramatic changes meant that mammal communities were formed that do not have analogs today, meaning that species were often found together in fossil assemblages that are not found together historically. Only a handful of African species went extinct during the transition from the late Pleistocene to the Holocene.

See also: K/Ar and $^{40}\text{Ar}/^{39}\text{Ar}$ Dating. **Archaeological Records:** Global Expansion 300 000–8000 Years Ago, Africa; Overview. **Glaciations:** Overview. **Luminescence Dating:** Electron Spin Resonance Dating; Optical Dating. **Paleoceanography, Physical and Chemical Proxies:** Oxygen-Isotope Stratigraphy of the Oceans. **Sea Level Studies:** Overview. **Vertebrate Records:** Late Pleistocene Megafaunal Extinctions; Mid-Pleistocene of Africa. **Vertebrate Studies:** Ancient DNA; Interactions with Hominids.

References

- Amani F and Geraads D (1993) Le gisement moustérien du Djebel Irhoud, Maroc: Précisions sur la faune et la biochronologie, et description d'un nouveau reste humain. *Compte Rendu de l'Académie des Sciences de Paris* 316: 847–852.
- Amani F and Geraads D (1998) Le gisement Moustérien du Djebel Irhoud, Maroc: Précisions sur la faune et la paléocologie. *Bulletin d'Archéologie Marocaine* 18: 11–17.
- Aouraghe H (2000) Les carnivores fossiles d'El Harhoura 1, Temara, Maroc. *L'Anthropologie* 104: 147–171.
- Avery DM (1987) Late Pleistocene coastal environment of the southern Cape Province of South Africa: Micromammals from Klasies River Mouth. *Journal of Archaeological Science* 14: 405–421.
- Avery DM (1992) The environment of early modern humans at Border Cave, South Africa: Micromammalian evidence. *Palaeogeography, Palaeoclimatology, Palaeoecology* 91: 71–87.
- Avery DM (2004) Size variation in the common mole rat *Cryptomys hottentotus* from southern Africa and its potential for palaeoenvironmental reconstruction. *Journal of Archaeological Science* 31: 273–282.
- Barnosky AD, Koch PL, Feranec RS, Wing SL, and Shabel AB (2004) Assessing the causes of Late Pleistocene extinctions on the continents. *Science* 306: 70–75.
- Brooks AS (2003) Sorting out the muddle in the Middle Pleistocene of Zambia. *Journal of Human Evolution* 45: 475–488.
- Burney DA (1999) Rates, patterns, and processes of landscape transformation and extinction in Madagascar. In: MacPhee RDE (ed.) *Extinctions in Near Time: Causes, Contexts, and Consequences*. New York: Kluwer/Plenum.
- Cooke HBS (1963) Pleistocene mammal faunas of Africa, with particular reference to southern Africa. In: Howell FC and Bourlière F (eds.) *African Ecology and Human Evolution*, pp. 65–116. Chicago: Aldine Publishing Company.
- Cornelissen E (2002) Human responses to changing environments in Central Africa between 40,000 and 12,000 B.P. *Journal of World Prehistory* 16: 197–235.
- Cruz-Urbe K (1983) The mammalian fauna from Redcliff Cave, Zimbabwe. *South African Archaeological Bulletin* 38: 7–16.
- Cruz-Urbe K, Klein RG, Avery G, et al. (2003) Excavation of buried late Acheulean (Mid-Quaternary) land surfaces at Duinefontein 2, Western Cape Province, South Africa. *Journal of Archaeological Science* 30: 559–575.
- Dobson M and Wright A (2000) Faunal relationships and zoogeographical affinities of mammals in north-west Africa. *Journal of Biogeography* 27: 417–424.
- Faure M and Guérin C (1990) *Hippopotamus laloumena* nov. sp., la troisième espèce d'hippopotame holocène de Madagascar. *Compte Rendu de l'Académie des Sciences* 310(Série II): 1299–1305.
- Feathers JK (2002) Luminescence dating in less than ideal conditions: Case studies from Klasies river mouth and Duinefontein, South Africa. *Journal of Archaeological Science* 29: 177–194.
- Gautier A (1993) The Middle Paleolithic archaeofaunas from Bir Tarfawi (Western Desert, Egypt). In: Wendorf F, Schild R, and Close AE (eds.) *Egypt during the Last Interglacial: The Middle Paleolithic of Bir Tarfawi and Bir Sahara East*, pp. 121–143. New York: Plenum.
- Geraads D (2002) Plio–Pleistocene mammalian biostratigraphy of Atlantic Morocco. *Quaternaire* 13: 43–53.
- Godfrey LR and Jungers WL (2002) Quaternary fossil lemurs. In: Hartwig WC (ed.) *The Primate Fossil Record*, pp. 97–121. Cambridge: Cambridge University Press.
- Goodman SM and Patterson BD (eds.) (1997) *Natural Change and Human Impact in Madagascar*. Washington, DC: Smithsonian Books.
- Grün R, Brink JS, Spooner NA, et al. (1996) Direct dating of Florisbad hominid. *Nature* 382: 500–501.
- Grün R and Stringer CB (1991) Electron spin resonance dating and the evolution of modern humans. *Archaeometry* 33: 153–199.
- Higuchi RE, Bowman B, Freiburger M, Ryder OA, and Wilson AC (1984) DNA sequences from the quagga, an extinct member of the horse family. *Nature* 312: 282–284.
- Karanth KP, Delefosse T, Rakotosamimanana B, Parsons TJ, and Yoder AD (2005) Ancient DNA from giant extinct lemurs confirms single origin of Malagasy primates. *Proceedings of the National Academy of Sciences* 102: 5090–5095.
- Kingdon J (1997) *The Kingdon Field Guide to African Mammals*. San Diego, California: Academic Press.
- Klein RG (1972) The late Quaternary mammalian fauna of Nelson Bay Cave (Cape Province, South Africa): Its implications for megafaunal extinctions and for cultural and environmental change. *Quaternary Research* 2: 135–142.
- Klein RG (1976) The mammalian fauna of the Klasies River Mouth sites, southern Cape Province, South Africa. *South African Archaeological Bulletin* 31: 75–96.
- Klein RG (1977) The mammalian fauna from the Middle and Later Stone Age (later Pleistocene) levels of Border Cave, Natal Province, South Africa. *South African Archaeological Bulletin* 32: 14–27.
- Klein RG (1978a) A preliminary report on the larger mammals from the Boomplaas Stone Age cave site, Cango Valley, Oudtshoorn District, South Africa. *South African Archaeological Bulletin* 33: 66–75.
- Klein RG (1978b) Preliminary analysis of the mammalian fauna from the Redcliff Stone age Cave Site, Rhodesia. *Occasional Papers of the National Museum of Southern Rhodesia* 4: 343–374.
- Klein RG (1980) Environmental and ecological implications of large mammals from Upper Pleistocene and Holocene sites in southern Africa. *Annals of the South African Museum* 81: 223–283.
- Klein RG (1984a) The large mammals of southern Africa: Late Pliocene to Recent. In: Klein RG (ed.) *Southern African Prehistory and Paleoenvironments*, pp. 107–146. Rotterdam: A. A. Balkema.
- Klein RG (1984b) Mammalian extinctions and stone age people in Africa. In: Martin PS and Klein RG (eds.) *Quaternary Extinctions: A Prehistoric Revolution*, pp. 553–573. Tucson: University of Arizona Press.
- Klein RG (1986) Carnivore size and quaternary climatic change in Southern Africa. *Quaternary Research* 26: 153–170.
- Klein RG (1987) The extinct Blue Antelope. *Sagittarius* 2: 20–23.

- Klein RG (1991) Size variation in the Cape dune mole rat (*Bathyergus suillus*) and late quaternary climatic change in the southwestern Cape Province, South Africa. *Quaternary Research* 36: 243–256.
- Klein RG, Avery G, Cruz-Urbe K, et al. (1999) Duinefontein 2: An Acheulean site in the western Cape Province of South Africa. *Journal of Human Evolution* 37: 153–190.
- Klein RG and Cruz-Urbe K (1983) Stone Age population numbers and average tortoise size at Byneskranskop Cave 1 and Die Kelders Cave 1, Southern Cape Province, South Africa. *South African Archaeological Bulletin* 38: 26–30.
- Klein RG and Cruz-Urbe K (1987) Large mammal and tortoise bones from Elands Bay Cave and nearby sites, Western Cape Province, South Africa. In: Parkington JE and Hall M (eds.) *Papers in the Prehistory of the Western Cape, South Africa*, vol. 332, pp. 132–163. Oxford: British Archaeological Reports International Series.
- Klein RG and Cruz-Urbe K (1991) The bovids from Elandsfontein, South Africa, and their implications for the age, palaeoenvironment, and origins of the site. *African Archaeological Review* 9: 21–79.
- Klein RG and Cruz-Urbe K (1996) Size variation in the rock hyrax (*Procavia capensis*) and late quaternary climatic change in South Africa. *Quaternary Research* 46: 193–207.
- Klein RG and Cruz-Urbe K (2000) Macromammals and reptiles. In: Braham L (ed.) *The Middle Stone Age of Zambia, South Central Africa*, pp. 51–56. Bristol: Western Academic and Specialist Press Limited.
- Klein RG and Scott K (1986) Re-analysis of faunal assemblages from the Haua Fteah and other late Quaternary archaeological sites in Cyrenaica, Libya. *Journal of Archaeological Science* 13: 515–542.
- Latham AG and Herries AIR (2004) The formation and sedimentary infilling of the Cave of Hearths and Historic Cave Complex, Makapansgat, South Africa. *Geoarchaeology* 19: 323–342.
- Leonard JA, Rohland N, Glaberman S, Fleischer RC, Caccone A, and Hofreiter M (2005) A rapid loss of stripes: The evolutionary history of the extinct quagga. *Biology Letters* 1: 291–295.
- Marean CW (1992) Implications of late Quaternary mammalian fauna from Lukenya Hill (south-central Kenya) for paleoenvironmental change and faunal extinctions. *Quaternary Research* 37: 239–255.
- Marean CW and Gifford-Gonzalez D (1991) Late Quaternary extinct ungulates of East Africa and paleoenvironmental implications. *Nature* 350: 418–420.
- Michel P (1992) Pour une meilleure connaissance du Quaternaire Continental Marocain: Les vertébrés fossiles du Maroc Atlantique, Central et Oriental. *L'Anthropologie* 96: 643–656.
- Plug I (2004) Resource exploitation: Animal use during the Middle Stone Age at Sibudu Cave, KwaZulu-Natal. *South African Journal of Science* 100: 151–158.
- Potts R and Deino A (1995) Mid-Pleistocene change in large mammal faunas of East Africa. *Quaternary Research* 43: 106–113.
- Rau R (1986) The Quagga and its kin. *Sagittarius* 1: 8–10.
- Robinson T, Bastos A, Halanich K, and Herzig B (1996) Mitochondrial DNA sequence relationships of the extinct Blue Antelope. *Hippotragus leucophaeus*. *Naturwissenschaften* 83: 178–182.
- Steele TE and Klein RG (2005) Mollusk and tortoise size as proxies for stone age population density in South Africa: Implications for the evolution of human cultural capacity. *Munibe* 57(1): 221–237.
- Stuenkel S (1989) Taxonomy, habits and relationships of the sub-fossil Madagascan hippopotamuses *Hippopotamus lemerlei* and *H. madagascariensis*. *Journal of Vertebrate Paleontology* 9: 241–268.
- Thomas H (1981) La faune de la Grotte a Néandertaliens du Jebel Irhoud (Maroc). *Quaternia* 23: 191–217.
- Turner A and Antón M (2004) *Evolving Eden: An Illustrated Guide to the Evolution of the African Large-Mammal Fauna*. New York: Columbia University Press.
- Turpie J and Crowe T (1994) Patterns of distribution, diversity and endemism of larger African mammals. *South African Journal of Science* 29: 19–32.
- Vermeersch PM (ed.) (2000) Palaeolithic living sites in Upper and Middle Egypt. In: *Egyptian Prehistory Monographs*. Leuven: Leuven University Press.
- Wrinch PJ and Rink WJ (2003) ESR Dating of tooth enamel from Aterian levels at Mugharet el 'Aliya (Tangier, Morocco). *Journal of Archaeological Science* 30: 123–133.
- Yoder AD, Cartmill M, Ruvolo M, Smith K, and Vilgalys R (1996) Ancient single origin for Malagasy primates. *Proceedings of the National Academy of Sciences* 93: 5122–5126.
- Yoder AD, Burns MM, Zehr S, et al. (2003) Single origin of Malagasy carnivora from an African ancestor. *Nature* 421: 734–737.

Late Pleistocene of North America

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Introduction

A synopsis is provided here of the late Pleistocene terrestrial vertebrate faunas of continental North America. The term ‘Pleistocene’ has a complicated early history and was originally equated to the glacial deposits in England (Bell et al., 2004). The base of the Quaternary is now set at 2.58 Ma (Gibbard and Head, 2010). Within chronostratigraphy, this system must be followed; however, overall this scheme does little to help elucidate the variations found within the changing climate (glacial–interglacial) and vertebrate faunas of the North American Pleistocene.

Subsequent to the first usage of the term, temporal relationships among the terrestrial Pleistocene deposits of North America were traditionally based on their affiliations to the classical four major glaciations (Nebraskan, Kansan, Illinoian, and Wisconsin), including interglacials between. Deep-sea sedimentary cores have shown that as many as 22 discrete glacial pulses over the past two million years are recognized (Richmond and Fullerton, 1986), which leaves the glacial and interglacial names basically useless in North America and are abandoned for use with faunal assemblages except for the most recent glaciation, Wisconsin. This glacial event retains its utility because it is within the age constraints of radiocarbon dating and is completely within the Upper or late Pleistocene. Richmond and Fullerton (1986) have the following age limitations for the Wisconsin Glaciation: ‘Eowisconsin’ (122–79 ka BP), early Wisconsin (79–65 ka BP), middle Wisconsin (65–35 ka BP), and late Wisconsin (35–10 ka BP).

A major task of a mammalian stratigrapher is to devise and operate within a framework of data that leads to the development of a succession of temporal intervals that account for all of geologic time (Woodburne, 2004). The North American Land Mammal Ages (LMAs) are constructed as biochronologic units and designed to recognize discrete intervals of time based on the evolution and association of fossil mammals (Woodburne, 2004). These units have been correlated into various geochronometric units including the Geomagnetic Polarity Time Scale (GPTS; see synopsis by Lindsay (2001)). The global standard for interpreting Pleistocene paleoclimate and chronology are the marine isotope stages (MIS), which use oxygen-isotope ratios as recorded in marine sedimentary cores. Through chronological correlation (based on additional radiometric controls and analyses), terrestrial faunas can be related and compared to these observed climatic changes.

Rancholabrean Land Mammal Age

The Pleistocene epoch contains three North American LMAs as defined in Bell et al. (2004, figure 7.3): Blancan (earliest),

Irvingtonian (early, middle, early–late), and Rancholabrean (late). The beginning of the Rancholabrean (RLB), and therefore the end of the Irvingtonian, is still inadequately understood. *Bison* (bison) is the defining taxon for the RLB. Bell et al. (2004) recognized the benefits of a single-taxon boundary definition (Woodburne, 2004), but also acknowledged the complications that seem to arise from such a definition, especially with the use of a large taxon species such as *Bison*.

The arrival time of *Bison* in North America is problematic, and therefore, the end of the Irvingtonian LMA and beginning age for the RLB is not well constrained. Until the chronology of the Irvingtonian–Rancholabrean transition can be better established, we follow Bell et al. (2004) in: (1) restricting the RLB to faunas south of 55° N latitude in North America, (2) defining the RLB by the earliest arrival of *Bison* south of this latitude, and (3) placing the best-corroborated arrival time of *Bison* (therefore the beginning of the RLB) between approximately 210 and 160 ka (i.e., late Pleistocene). Earlier dispersal events of *Bison* have been proposed; these preliminary reports of late Pliocene *Bison* in North America cannot be assessed adequately until more information is presented. Not all late Pleistocene faunas contain *Bison* and therefore their assignment to an RLB age is invariably assumed, often via a correlation with a radiometric time assignment.

Although the appearance of *Bison* defines the RLB, there are a number of taxa that are limited to just this LMA (shaded taxa in Table 1). There are a number of large and small mammals that first appear in the fossil record prior to the RLB (in most cases during the Irvingtonian LMA); some of these continue on through the RLB and some help define the end to this age (see below). Table 1 also lists some of the mammals that first appear in the RLB and continue on into the recent.

Using the definition provided by Bell et al. (2004), the RLB includes that area south of 55° N latitude, incorporating all continental USA, Mexico, southeastern Alaska, and southernmost Canada. Eastern Beringia (most of Alaska and the Yukon) is not included here because the faunas from there are characterized to be more ‘Siberian-like’ (Mammoth Steppe) than similar to the rest of temperate North America (Guthrie, 2001). Temporally, the RLB is equated, conservatively, to approximately 210–160 to 11 ka BP. This time interval would include at a minimum the entire Wisconsin Glaciation (MIS 5d-2) and the preceding interglacial (‘Sangamonian’; MIS 5e; ~130 to 115 ka BP). It remains to be determined whether or not the RLB began earlier than this (based on the arrival of *Bison*) and whether or not the age contained additional glacial and interglacial climate fluctuations and therefore concomitant alterations in the mosaic of each faunal community.

Table 1 Some of the more prominent mammals that help characterize the Rancholabrean Land Mammal Age (RLB)

Taxon	Irv or earlier	RLB	Holocene recent
<i>Megalonyx jeffersoni</i>		X	
<i>Panthera atrox</i>		X	
<i>Miracinonyx trumani</i>		X	
<i>Canis dirus</i>		X	
<i>Platygonus compressus</i>		X	
<i>Platygonus vetus</i>	X	X	
<i>Nothrotheriops</i>	X	X	
<i>Paramylodon harlani</i>	X	X	
<i>Glyptotherium</i>	X	X	
<i>Holmesina</i>	X	X	
<i>Brachyprotoma</i>	X	X	
<i>Smilodon</i>	X	X	
<i>Arctodus simus</i>	X	X	
<i>Euceratherium</i>	X	X	
<i>Oreamnos harringtoni</i>	X	X	
<i>Mammot americanum</i>	X	X	
<i>Mammuthus</i>	X	X	
↖ Extinction			
<i>Didelphis</i>	X	X	X
<i>Brachylagus idahoensis</i>	X	X	X
<i>Sylvilagus</i>	X	X	X
<i>Clethrionomys</i>	X	X	X
<i>Lemmys curtatus</i>	X	X	X
<i>Ondatra zibethicus</i>	X	X	X
<i>Marmota flaviventris</i>	X	X	X
<i>Marmota monax</i>	X	X	X
<i>Cynomys gunnisoni</i>	X	X	X
<i>Cynomys ludovicianus</i>	X	X	X
<i>Panthera onca</i>	X	X	X
<i>Mustela erminea</i>	X	X	X
<i>Canepatus</i>	X	X	X
<i>Canis latrans</i>	X	X	X
<i>Canis lupus</i>	X	X	X
<i>Ursus americanus</i>	X	X	X
<i>Aplodontia rufa</i>		X	X
<i>Puma concolor</i>		X	X
<i>Vulpes velox</i>		X	X
<i>Alces alces</i>		X	X
<i>Bison</i>		X	X
<i>Ovis canadensis</i>		X	X
<i>Oreamnos americanus</i>		X	X
<i>Homo sapiens</i>		X	X

The taxa above and including *Mammuthus* are restricted to the RLB and help define the age. *Bison* is the defining species for the RLB. For more details, see Bell CJ, Lundelius EL, Barnosky AD, et al. (2004) The Blancan, Irvingtonian, and Rancholabrean mammal ages. In: Woodburne MO (ed.) *Late Cretaceous and Cenozoic Mammals of North America*, p. 232. New York: Columbia University Press.

Terrestrial Faunas

Fossil deposits of the RLB contain a diverse record of amphibians, reptiles, birds, and mammals, but certainly the most comprehensively studied are the mammals. Some of the earliest publications about these remains include those by Hay (1899) and subsequent volumes, which are still of importance. Kurtén and Anderson (1980) set the pace to understand the North American late Pleistocene mammals. There has been a recent trend to provide useful overviews and compilations of the Pleistocene vertebrates known from various states and

regions of North America (Webb et al., 2004). Such compilations provide an opportunity to update, synthesize, and simplify a large amount of information that is scattered through the literature, especially for large geographic areas. These compilations typically embrace just the mammals, but can be more geographically comprehensive and incorporate, for instance, southeastern Alaska (Heaton and Grady, 2003), Arizona (Mead et al., 2005), California (Jefferson, 1991), the Great Basin (Intermountain West) region (Grayson, 1994), Mexico (Arroyo-Cabral and Polaco, 2003), Idaho (Jefferson et al., 2002), northern North America (Harrington (2003), also includes much of the Mammoth Steppe faunas (not included here); and New Mexico (Morgan and Lucas, 2005).

The above and below references provide a good overall coverage about the North American RLB faunas but treatment is predominantly of the Wisconsin Glaciation (MIS 4–2). Little is known of the faunistics for the ‘Sangamonian’ interglacial except for what is previewed in (Pinsof 1996).

Mammals

Studies which are taxonomic in approach (e.g., *Mammuthus*, *Pampatherium*, *Desmodus*; Figure 1) are numerous and cannot be listed here; however, these can be easily located via web-based reference searches. Published investigations of morphologic variation in various mammals from the Blancan LMA, Irvingtonian, RLB, and Holocene encompass a broad range of taxa and are listed in Bell et al. (2004). Some authors have used the evolutionary changes and dispersal history of the arvicoline rodents to divide the previous LMAs (Blancan and Irvingtonian) into temporal subdivisions; however, this as yet is not possible within the RLB (Bell et al., 2004). Detailed radiocarbon dates (using accelerator mass spectrometry) on critical mammalian taxa were the key to unraveling the issue about the stratigraphic association of extant species that do not currently live together today. These currently allopatric species were found to be sympatric during certain phases of the Wisconsin Glaciation. The precision radiocarbon dating documented the contemporaneity of nonanalog species (Graham and Mead, 1987).

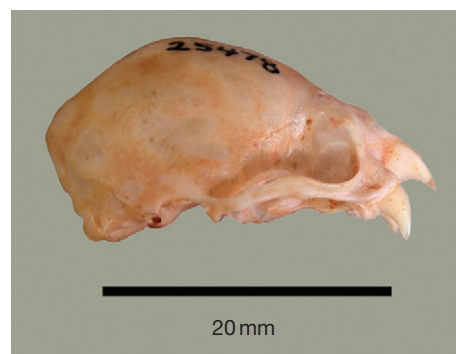


Figure 1 Skull of *Desmodus stocki* (extinct Stock's vampire bat) recovered from the *Nothrotheriops shastensis* (Shasta ground sloth) dung layer in Rampart Cave, western Grand Canyon, Arizona.

Amphibians, Reptiles, and Birds

Compared to research about mammalian faunas, published accounts of the RLB herpetofaunas are less common for North America. Regional compilations are rare, such as the one for the Colorado Plateau and the Great Basin (Mead and Bell, 1994) and Arizona (Mead, 2005). General overviews do exist and provide in-depth reference data sets (Holman, 1995), along with specific summaries of the anurans (Holman, 2003) and the snakes (Holman, 2000). The lizards have yet to be assessed in detail for overall North America.

A glacial climate regime did create nonanalog community scenarios for the amphibians and reptiles, as was found with the mammals. For example, the record of the cold-intolerant lizard, chuckwalla (*Sauromalus*), and the desert tortoise (*Gopherus*) in the Grand Canyon during the Wisconsin Late Glacial was used to illustrate that the climate then was more equable than today; these reptiles were found to be living in a community with juniper (*Juniperus*) and single-leaf ash (*Fraxinus*) (Graham and Mead, 1987).

No in-depth summary exists for the RLB birds, but the best available overview data set is provided by Steadman in Lundelius et al. (1983) and in Webb et al. (2004). Nonanalog scenarios for the birds are also recorded in the literature about the RLB avifauna.

Localities

A map of North America showing the location of each and every RLB locality would be of little use here; critically dated sites can be found, in part, in Graham and Mead (1987) and at the FAUNMAP website (see below). Fossil localities containing faunas of the RLB are found over the entire region. Some physiographic areas are clearly better understood than others. For instance, the intermountain west (including the Great Basin) is far less well understood than the Pacific coastal region or the karstic region of Appalachia in the eastern portion of the continent.

A project entitled FAUNMAP analyzed 2945 localities to demonstrate that the geographic ranges of individual species shifted at different times, in different directions, and at different rates in response to Late Glacial environmental fluctuations in North America (<http://www.museum.state.il.us>). They found that the modern community patterns emerged only in the most recent few thousands of years, and possibly most importantly, they determined that many Late Glacial faunal mosaics do not have a modern analog with which to compare. Much of their data set is online and can be of great use to all researchers.

Many of the RLB faunas from North America are preserved in the numerous limestone caves found throughout the continent (Figure 2), as shown in a recent book by Schubert et al. (2003). Caves typically provide snapshots of depositional history and are often biased toward select physiographic, topographic, and taphonomic settings (Harris, 2005) (Figure 3), yet they also provide extremely detailed faunal accounts and in some cases preserve organic remains often destroyed in open-air, nonpermafrost scenarios. Dry caves in the arid southwest (Figure 4) have preserved a detailed record of organic remains



Figure 2 Smith Creek Cave, Great Basin, Nevada. This classic large-entranced chamber contains millions of bones and is the type locality for Rancholabrean-age animals such a *Oreamnos harringtoni* (Harrington's extinct mountain goat), *Spizaetus willetti* (extinct eagle), and *Teratornis incredibilis* (extinct incredible teratorn).

(dung, hair, body tissues; Figures 5 and 6), which in turn provide dietary reconstructions of the extinct large mammals; examples include the Shasta ground sloth (*Nothrotheriops shastensis*; Xenarthra), Harrington mountain goat (*Oreamnos harringtoni*; Artiodactyla), shrub ox (*Euceratherium collinum*; Artiodactyla), and mammoth (*Mammuthus*; Proboscidea) (see overview in Mead and Agenbroad, 1992).

It would be extremely difficult to tally here the 'best' or 'most important' RLB localities as it would depend on the chosen criteria. For PaleoIndian localities with RLB faunas, the following sites are extremely important: Agate Basin, Blackwater Draw, Bonfire, Colby, Dent, Kimmswick, Lehner, Lubbock Lake, Murray Springs, to name but a few (Grayson and Meltzer, 2002). For dry-preserved organic remains of extinct mammals along with well-associated plant remains, one should understand the following caves: Bechan, Grobott, Rampart, Stanton's (Mead and Agenbroad, 1992). For remains preserved via permafrost, one would examine the Mammoth Steppe faunas (Harrington, 2003). For critical localities containing early *Bison* records, Bell et al. (2004, Figure 7.6) provide a list and critique. Critical RLB localities that should be understood to



Figure 3 One of many limestone caves in the Grand Canyon that preserve nests, egg shells, and feathers of Rancholabrean-age *Gymnogyps californianus* (California condor).

comprehend the full meaning of this LMA include, besides those listed above: Rancho La Brea (the type locality for the LMA), American Falls, Boney Spring, Brynjulfson Caves, Dry Cave, Camp Cady, Cheek Bend Cave, Fossil Lake, Friesenhahn Cave, Haile, Hot Springs Mammoth, Ingleside, Kennewick, Ladds Quarry, Little Box Elder Cave, Meade Basin localities, Medicine Hat, Natural Trap Cave, New Paris No. 4 Cave, Papago Springs Cave, Peccary Cave, Rainbow Beach, Reddick, San Josecito Cave, Silver Creek, Smith Creek Cave, Snake Creek Burial Cave, U-Bar Cave, Welsh Cave (references in Kurtén and Anderson (1980) and Bell et al. (2004)).

Late Pleistocene Extinction

The end of the RLB is marked by the disappearance of at least 32 genera of primarily large mammals (Table 2). The timing and cause of this extinction has been of considerable debate that began active discussion and plateaued in debate in the late 1960s through 1980s (Martin and Klein, 1984), and it still continues with great fervor. Part of the debate is whether all these mammals became extinct at the same time (an event) or if they 'slowly' died out over a period of thousands of years, possibly throughout the late Wisconsin Late Glacial (a process). A detailed examination of radiocarbon analyses directly on the

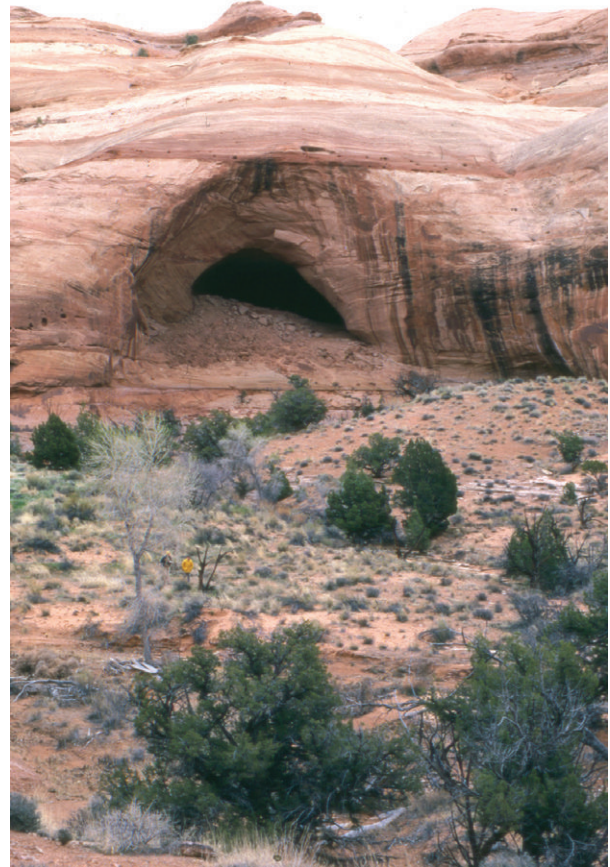


Figure 4 Bechan Cave (Utah) is a large cavern in sandstone that preserved over 300 m³ of dried dung, predominantly of *Mammuthus* (extinct mammoth) and *Eucatherium* (extinct shrubox).

critical mammal taxa has at least determined when the last of the species seems to have existed (e.g., see Mead and Agenbroad, 1992). However, this analysis does not determine when a taxon might have begun the final irreversible decline, just its apparent final demise recorded in the known fossil record.

Although it is easier to determine the time of extinction, the development of the cause is of heated debate. Two major hypotheses are established to answer or attempt to answer the cause for the terminal RLB extinction: (1) overkill at the hands of humans (directly or indirectly), and (2) environmental change.

The arrival of humans to North America is an RLB event, although the timing for this dispersal episode(s) is not precisely established. The culture utilizing the Clovis tool kit is often considered that of the first Americans (PaleoIndians) and may have arrived approximately 11,500 radiocarbon ka (Grayson and Meltzer, 2002). Yet there is still debate as to whether or not people were in South America prior to the Clovis. The Clovis event is a sudden appearance in the late RLB but it is not understood whether this was a rapid spread of a people into a new land or merely a diffusion of a new and highly distinctive stone tool technology across an extant population in the Americas. In any case, the fossil record of humans in the Americas during the RLB (and its equivalent in South America; in part, Lujanian) is extraordinary sparse.



Figure 5 Dung bolus of *Mammuthus* from Bechan Cave, Utah. Dung dates to approximately 11.600 ka BP. Note the large fragments of poorly chewed grass.



Figure 6 A preserved *Bison* (bison) keratinous hoof dating to about 13.000 ka BP from a dry cave in the Grand Canyon of Arizona.

Much has been discussed and likely will continue into the future; many critical references to this subject can be found in [Grayson and Meltzer \(2002\)](#).

The environmental change explanation for extinctions is much more complex than the overkill supposition. Whereas overkill has the direct influence of one predator (*Homo*) upon a diverse fauna occupying a multitude of communities throughout the continent, the environmental hypothesis requires the understanding the critical ecological tolerances of each taxon affected by the overall environmental changes. These climatic changes occurred presumably from the onset of deglaciation (possibly by about 18 ka BP) and lasted to (or culminated at) approximately 11 ka BP. The complexity occurs because each taxon reacts individually to climate/environmental change. Some taxa may have been adversely affected at the onset of change and became extinct early in the Late Glacial; others persisted to the end of the deglaciation. When and if there were early deglaciation extinctions is not understood and not well established, even with the existing robust radiocarbon data set. The late persistence of select large mammals is well established and their apparently synchronous timing for

Table 2 North American (southern Canada, USA, Mexico) genera and species of Rancholabrean age mammals either extinct or extirpated (†) from North America yet found living (at the generic level) on another continent

Small mammals

Chiroptera

Desmodus stocki – Stock's vampire bat

Desmodus draculae – vampire bat

Lagomorpha

Azlanolagus – rabbit

Rodentia

Cratogeomys onerosus – pocket gopher

Neotoma pygmaea – woodrat

Neotoma findleyi – woodrat

Carnivora

Bassariscus ticuti – ringtail

Brachyprotoma – short-faced skunk

?*Martes nobilis* – noble marten (*Martes americana*)

Large mammals

Xenarthra

?*Dasypus bellus* – beautiful armadillo

Eremotherium – giant ground sloth

Glyptotherium – glyptodon

Holmesina – pampathere; giant armadillo

Megalonyx jeffersoni – Jefferson's ground sloth

Nothrotheriops – Shasta ground sloth

Pampatherium – pampathere; giant armadillo

Paramylodon – mylodont ground sloth

Rodentia

Castoroides – giant beaver

† *Hydrochaeris* – capybara

Carnivora

Arctodus – short-face bear

Canis dirus – dire wolf

Homotherium – scimitar cat

Miracinonyx trumani – American cheetah

Panthera atrox – American lion

Smilodon – sabertooth

Tremarctos – bear

Proboscidea

Cuvieronius – gomphothere

Mammuthus – mammoth (many species)

Mammut – mastodon

Perissodactyla

† *Equus* – horse (many species)

† *Tapirus* – tapir

Artiodactyla

Bootherium/Symbos – muskox

Camelops – camel

Capromeryx – diminutive pronghorn

Cervalces – stag-moose

Euceratherium – shrubbox

Hemiauchenia – llama

Mylohyus – woodland peccary

Navahoceros – mountain deer

Oreamnos harringtoni – Harrington's mountain goat

?*Ovis catclawensis* – mountain sheep (*Ovis canadensis*)

Paleolama – llama

Platygonus – peccary

Stockoceros – pronghorn

Tetrameryx – pronghorn

?, there is some question as to whether or not that this is a distinct species from a living form and/or if it survived the terminal RLB extinction and persisted into the Holocene. Most genera have more than one species assigned to it but those with a significant number of species are so indicated.

extinction (in at least the greater southwest) is well established (Mead and Agenbroad, 1992). This data set has been used to support the overkill hypothesis; however, at best, it only illustrates the synchronicity of the extinction process. [Graham and Mead \(1987\)](#) provide an overview about the nature of the RLB disharmonious (nonanalog) biotas of North America that were apparently maintained by equable climates. This concept of climate effects on the biota has played heavily into the discussions about late Pleistocene, RLB, extinctions.

Mexico

Although southern Alaska and Canada, and the continental USA, are fairly well understood for their RLB faunas (especially for the mammals), it is safe to state that most of Mexico, especially northern Mexico, is basically unknown. [Arroyo-Cabral and Polaco \(2003\)](#) provide an in-depth overview of important faunas recovered from caves of Mexico, yet this covers only 15 caves and three fissure deposits for the entire country. Their publication provides the best overview (with references) about the overall RLB faunas (herps, birds, and mammals) and indicates that little is comprehensively known of this southernmost North American region. Preliminary reports occur as abstracts for annual meetings of the Society of Vertebrate Paleontology, but few data-heavy publications exist with a thorough treatment of the entire faunas (many from the central and southern regions of the country). A new locality in northeastern Sonora (bordering Arizona) is producing a wealth of remains (invertebrates, herps ([Figure 7](#)), birds, and mammals), many with a tropical ecological preference. The in-press report indicates that a tropical environment during some portion of the RLB approached the xeric and boreal communities of the southwest; it is not understood if the deposition occurred during a glacial or interglacial climatic regime. The states of Sonora and Chihuahua include the region that was the apparent ecotone between temperate USA and tropical Mexico. The lack of understanding about this region is illustrated by the creation of a faunal province to categorize

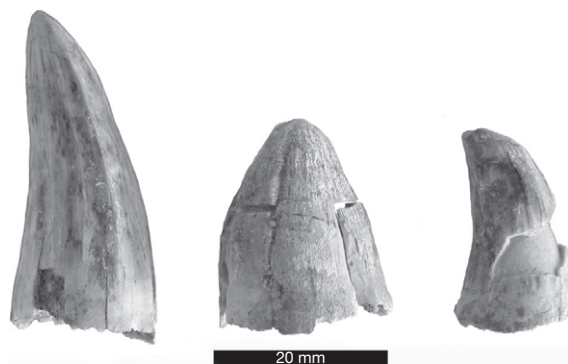


Figure 7 Teeth of *Crocodylus* (crocodile) from a marsh deposit in northeastern Sonora, Mexico, located 350 km upriver from the Gulf of California. These teeth were recovered in association with *Bison* (Rancholabrean) and many other taxa such as *Glyptotherium* (glyptodont), pampathere (extinct giant armadillo), *Tapirus* (tapir), *Cuvieronius* (gomphothere), and some of tropical affinities including *Hydrochoerus* (capibara).

the 'known' faunas. The concept of faunal provinces is discussed in [Bell et al. \(2004, p. 294\)](#) and brings to light the issues around Mexico ('limits of the Mexican faunal region are not well established or characterized'). The tropical fauna of Florida are often included within the Mexican faunal province. The FAUNMAP data confirm that the Floridian faunas were distinct from other North American faunas. Much needs to be learned about this Mexican region. A comprehensive assessment of the RLB of Mexico is sorely needed.

The Next Step

The RLB is quite well-known, yet clearly much remains to be understood, or better understood. The mammals are the best and most comprehensively understood of the terrestrial vertebrates. The amphibians and reptiles are known for some regions, some areas are better than others, while in other physiographic districts there is a total void of information. The record of the avifauna is equally as spotty and deserves more attention. These biases appear to illustrate more of a void in the number of researchers and adequate comprehensive skeleton collections and/or bias in collecting the fossils than they do some taphonomic factor in the fossil record. Screen-washing of fine sediments should be routine during excavations, no matter what the taxonomic or stratigraphic focus. Sieve sizes should be no larger than ~500 μm for screen washing in order to recover these microfossils. If the sieve sizes are on this order and screen-washing (along with sorting with the aid of a microscope) is a consistent routine, then it is certain that more information about amphibians, reptiles, and birds will emerge from fossil localities in the near future.

Vertebrate faunas are in need of better, more precise, direct chronological control. These data need to be applied to vegetational and climatic proxy data sets in order to develop a better understanding of local and regional changes. We have learned that climate is constantly changing and therefore vegetational and faunal communities were also changing, often in mosaic construct. Although constant, the magnitude of change appears to be cyclical, but not necessarily rhythmic. When possible, more dietary, molecular (DNA), and other biochemical analyses need to be conducted on or in association with the RLB taxa. Such studies should help elucidate the individual reactions to environmental change and possibly evaluate the cause for extinctions.

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See also: [Archaeological Records: Global Expansion 300 000–8000 Years Ago, Americas](#). [Quaternary Stratigraphy: Chronostratigraphy; Continental Biostratigraphy](#). [Vertebrate Records: Late Pleistocene Megafaunal Extinctions](#). [Vertebrate Studies: Ancient DNA](#).

References

- Arroyo-Cabrales J and Polaco OJ (2003) Caves and the Pleistocene vertebrate paleontology of Mexico. In: Schubert BW, Mead JI, and Graham RW (eds.) *Ice Age Cave Faunas of North America*, p. 273. Bloomington: Indiana University Press.
- Bell CJ, Lundelius EL, Barnosky AD, et al. (2004) The Blancan, Irvingtonian, and Rancholabrean mammal ages. In: Woodburne MO (ed.) *Late Cretaceous and Cenozoic Mammals of North America*, p. 232. New York: Columbia University Press.
- Berggren WA, Hilgen FJ, Langereis CG, et al. (1995) *Geological Society of America Bulletin* 107: 1272–1287.
- Gibbard PL and Head MJ (2010) The newly-ratified definition of the Quaternary System/Period and redefinition of the Pleistocene Series/Epoch, and comparison of proposals advanced prior to formal ratification. *Episodes* 33: 152–158.
- Graham RW and Mead JI (1987) Environmental fluctuations and evolution of mammalian faunas during the last deglaciation in North America. In: Ruddiman WF and Wright HE (eds.) *North America and Adjacent Oceans during the Last Deglaciation: The Geology of North America*, vol. K-3, p. 371. Boulder: Geological Society of America.
- Grayson DK (1994) The extinct Late Pleistocene mammals of the Great Basin. In: Harper KT, St. Clair LL, Thorne KH, and Hess WM (eds.) *Natural History of the Colorado Plateau and Great Basin*, p. 55. Niwot: University Press of Colorado.
- Grayson DK and Meltzer DJ (2002) Clovis hunting and large mammal extinction: A critical review of the evidence. *Journal of World Prehistory* 16: 313–359.
- Guthrie RD (2001) Origin and causes of the mammoth steppe: A story of cloud cover, woolly mammal tooth pits, buckles, and inside-out Beringia. *Quaternary Science Reviews* 20: 549–574.
- Harington CR (ed.) (2003) *Annotated Bibliography of Quaternary Vertebrates of Northern North America* Toronto: University Toronto Press.
- Harris AH (2005) Caves as unique resources for Pleistocene vertebrate faunas. In: Lucas SG, Morgan GS, and Zeigler KE (eds.) *New Mexico's Ice Ages* 28: p. 249. New Mexico: Museum of Natural History and Science Bulletin.
- Hay OP (1899) A census of the fossil vertebrata of North America. *Science* 10: 681–684.
- Heaton TH and Grady F (2003) The Late Wisconsin vertebrate history of Prince of Wales Island, southeast Alaska. In: Schubert BW, Mead JI, and Graham RW (eds.) *Ice Age Cave Faunas of North America*, p. 17. Bloomington: Indiana University Press.
- Holman JA (1995) *Pleistocene Amphibians and Reptiles in North America*. New York: Oxford University Press.
- Holman JA (2000) *Fossil Snakes of North America*. Bloomington: Indiana University Press.
- Holman JA (2003) *Fossil Frogs and Toads of North America*. Bloomington: Indiana University Press.
- Jefferson GT (1991) A catalogue of Late quaternary vertebrates from California. Part Two: Mammals. In: Technical Reports 7 *Natural History Museum of Los Angeles County* 1–129.
- Jefferson GT, McDonald HG, Akersten WA, and Miller SJ (2002) Catalogue of Late Pleistocene and Holocene fossil vertebrates from Idaho. In: Akersten WA, Thompson ME, Meldrum DJS, Rapp RA, and McDonald HG (eds.) *And Whereas... Papers on the Vertebrate Paleontology of Idaho Honoring John A. White*, vol. 2, p. 157. Idaho Museum of Natural History. Occasional Paper 37.
- Kurtén B and Anderson E (1980) *Pleistocene Mammals of North America*. New York: Columbia University Press.
- Lindsay E (2001) Correlation of mammalian biochronology with the geomagnetic polarity time scale. *Bollettino della Società Paleontologica Italiana* 40: 225–233.
- Lundelius EL, Graham RW, Anderson E, et al. (1983) Terrestrial vertebrate faunas. In: Porter SC (ed.) *Late-Quaternary Environments of the United States, Vol. 1: The Late Pleistocene*, p. 311. Minneapolis: University of Minnesota Press.
- Martin PS and Klein RG (1984) *Quaternary Extinctions. A Prehistoric Revolution*. Tucson: The University of Arizona Press.
- Mead JI (2005) Late Pleistocene (Rancholabrean) amphibians and reptiles of Arizona. In: Heckert AB and Lucas SG (eds.) *Vertebrate Paleontology in Arizona*, vol. 29, p. 137. New Mexico: Museum of Natural History and Science Bulletin.
- Mead JI and Agenbroad LD (1992) Isotope dating of Pleistocene dung deposits from the Colorado Plateau, Arizona, and Utah. *Radiocarbon* 34: 1–19.
- Mead JI and Bell CJ (1994) Late Pleistocene and Holocene herpetofaunas of the Great Basin and Colorado Plateau. In: Harper KT, Clair LL St., Thorne KH, and Hess WM (eds.) *Natural History of the Colorado Plateau and Great Basin*, p. 255. Niwot: University Press of Colorado.
- Mead JI, Czaplewski NJ, and Agenbroad LD (2005) Rancholabrean (Late Pleistocene) mammals and localities of Arizona. In: McCord RD (ed.) *Vertebrate Paleontology of Arizona*. Mesa Southwest Museum Bulletin 11, p. 139.
- Morgan GS and Lucas SG (2005) Pleistocene vertebrate faunas in New Mexico from alluvial, fluvial, and lacustrine deposits. In: Lucas SG, Morgan GS, and Zeigler KE (eds.) *New Mexico's Ice Ages*, vol. 28, p. 185. New Mexico: Museum of Natural History and Science Bulletin.
- Pinsof JD (1996) Current status of North American Sangamonian local faunas and vertebrate taxa. In: Stewart KM and Seymour KL (eds.) *Palaeoecology and Palaeoenvironments of Late Cenozoic Mammals (Tributes to the Career of C. S. (Rufus) Churcher)*, p. 156. Toronto: University of Toronto Press.
- Richmond GM and Fullerton DS (1986) Summation of quaternary glaciations in the United States of America. *Quaternary Science Reviews* 5: 183–196.
- Schubert BW, Mead JI, and Graham RW (eds.) (2003) *Ice Age Cave Faunas of North America*. Bloomington: Indiana University Press.
- Webb SD, Graham RW, and Barnosky AD (2004) Vertebrate Paleontology. In: Gillespie AR, Porter SC, and Atwater BF (eds.) *The Quaternary Period in the United States*, p. 519. Amsterdam: Elsevier Developments in Quaternary Science 1.
- Woodburne MO (2004) *Late Cretaceous and Cenozoic Mammals of North America*. New York: Columbia University Press.

Relevant Website

<http://www.stratigraphy.org> – International Commission on Stratigraphy.
www.museum.state.il.us – Illinois State Museum.

Late Pleistocene of South America

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Glossary

Browser: Feeding on leaves.

Graviportal: Columnar limbs, vertically oriented, and heavily constructed to carry large bodies.

Grazer: Feeding on grass.

Interandean depression: A tectonic depression flanked by the western and eastern Andean cordilleras located ca.

2000–3000 m above sea level that crosses Ecuador from north to south.

Mixed-feeder: Feeding on grass and leaves.

Scratch-digger: Digging with limbs.

Scavenger: Feeding on dead animals.

Biogeographic Origin

The long history of South America (SA) as an island continent during most of the Cenozoic was viewed by Simpson (1980, p. 240), as ‘...a grand experiment in population biology, ecology, biogeography, and other subjects.’ From the late Miocene to the early Pleistocene, SA became progressively connected to North America (NA) by way of Caribbean tectonic processes. In the Pliocene–Pleistocene, the Centroamerican isthmus emerged and the acme of the Great American Biotic Interchange occurred. This terrestrial connection provided an ecologically selective dispersal corridor, where a number of animals changed their geographic distribution. This had remarkable effects on the SA fauna as northern terrestrial mammals successfully spread and diversified (Webb, in Eisenberg and Redford (1999)) (Table 1).

Late Pleistocene sediments have yielded the result of ca. 2 Myr of evolution of native SA vertebrates with the addition of NA immigrants. They include the last representatives of some indigenous and immigrant mammals that went extinct during the late Pleistocene/early Holocene (LP/EH) and diverse taxa, which successfully persist today.

Current SA vertebrate communities do not reflect that Pleistocene diversity, owing to the decimation of a considerable number of taxa (notably large turtles and large to megamammals), reflecting a tragic end to a long and fruitful trip.

Biostratigraphic Framework

The evidence from continental mammals supports the current biostratigraphic scheme, which includes the Lujanian Stage/age (late Pleistocene to early Holocene), based on the standard reference scale established by Carlos and Florentino Ameghino in late nineteenth–early twentieth centuries (Tonni and Cione, 1999). The reference area covers the Buenos Aires province and the Mesopotamian region located between the Paraná and Uruguay rivers.

Although geographic differences related to ecologic/physical processes are potential limiting factors, intracontinental correlations can still be useful. Authors usually correlate the

assemblage being studied with the typical fauna of the late Pleistocene Lujanian Stage, which includes the most typical and odd extinct SA Pleistocene mammals (glyptodonts, toxodonts, ground-sloths, etc.), some of which may have persisted into the early Holocene.

Key Sites

Geographic areas with fossiliferous sites yielding absolute ages (^{14}C : radiocarbon, TL: thermoluminescence, ESR: electron spin resonance, Th–U: uranium-series ages) and paleoenvironment/climatic interpretations are presented and briefly summarized (Figure 1; Table 2). The evidence comes predominantly from the Last Glacial Maximum (LGM) as little is known from earlier phases such as the last interglacial or from the subsequent Late Glacial interstadial.

The following information is drawn from: Hoffstetter (1952), Marshall et al. (1984), Cartelle (1994), Politis et al. (1995), Eriksen and Strauss (1998), Tonni and Cione (1999), Zárate (1999), chapters of Cartelle, Rancy, and Webb in Eisenberg and Redford (1999), Rancy (2000), Cione et al. (2003), Ficarella et al. (2003), Latrubesse (2003), Santos et al. (2003), Salemm and Miotti (2003), Iriondo et al. (2004), Alberdi and Prado (2004), Rossetti et al. (2004), Ubilla (2004) and references therein.

Northern SA (Venezuela and Colombia)

In many cases, fossiliferous sites require further research due to low stratigraphic and taxonomic resolution. Nevertheless, important sites, such as Sabana de Bogotá, Curiú, Tibitó (Colombia), Muaco, Cueva del Guácharo, and Taima-Taima (Venezuela), have yielded vertebrates including evidence of human/megafauna association. The radiocarbon dates range from 21.9 to 10.8 ka BP (Colombia), and from 16.3 to 12.4 ka BP (Venezuela) (Marshall et al., 1984; Cooke, in Eriksen and Strauss, 1998). Megafaunal mammals consumed by humans were found in Tibitó I (11.74 ka BP) and Taima-Taima (13.5–12.43 ka BP).

The climate was drier and cooler than today between 21 and 14 ka BP (6–8 °C below current values); temperature oscillations colder than today occurred at 14–10 ka BP.

Table 1 A: mammals living in SA before the Pliocene/Pleistocene interchange; B: immigrant continental mammals from NA^{a-d}

A	B
Marsupials (opossums)	Carnivores (foxes, bears, raccoons, dogs, cats, weasels, otters, skunks)
Sloths (ground and arboreal)	Mastodonts-like (Gomphotheriids)
Anteaters	Deer
Armadillos and pampatheres	Horses
Glyptodonts	Tapirs
Notoungulates	Peccaries
Litopterns	Camels (guanacos and vicuñas)
Long-tailed monkeys ^e	Squirrels
Caviomorph rodents (cavies, capybaras, agouties, nutrias, spiny rats) ^f	Rabbits
Sigmodontine rodents (field and marsh mice) ^g	More sigmodontine rodents

^aFlynn et al. (2003).

^bPardiñas et al. (2002).

^cPascual et al. (1996).

^dWebb (1991).

^eFirst fossil record in SA in late Oligocene (immigrants likely from Africa).

^fFirst fossil record in SA in late Eocene/early Oligocene (immigrants likely from Africa).

^gFirst fossil record in SA in late Miocene (immigrants from NA).

In northern Venezuela, desert-like conditions prevailed during the pleniglacial.

Andean and Western Ecuador

The open-air sites in the Interandean Depression (Quebrada Cuesaca, Pistud, and Chalán) have yielded mammalian assemblages that are unmodified by early humans. The radiocarbon dates range from 20.9 to 12.3 ka BP (see [Coltorti et al. \(1998\)](#) in [Table 2](#)). Due to an extremely arid/cold climate, the mega-fauna may have migrated to the coast ca. 12 ka BP.

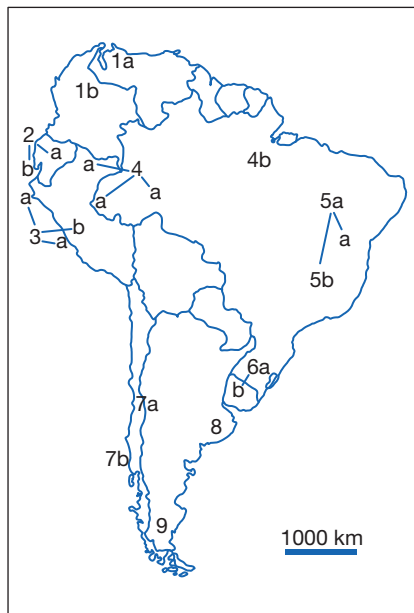


Figure 1 Key sites: 1. Northern SA: a, Venezuela, b, Colombia; 2. Ecuador: a, Andean, b, western; 3. Peru: a, western, b, Andean; 4. Amazonia: a, western, b, central; 5. a, northeastern Brazil, b, central-eastern Brazil; 6. a, southern Brazil, b, Uruguay; 7. a, central Argentina and Chile, b, central-southern Chile; 8. pampean sites; 9. southern cone.

The Ecuadorian coast includes open-air sites such as San José, La Libertad (=La Carolina) and El Cautivo (Santa Elena Península), Machalilla, and Pueblo Nuevo. Amphibians, reptiles, birds, and mammals including forested and open-air taxa were found. The assemblages have been ¹⁴C-dated to 18.4 ka BP (La Carolina) and to 8.68 ka BP (El Cautivo, with mega-fauna associated with lithic artifacts) ([Ficcarelli et al., 2003](#)). The climate shifted ca. 10 ka BP from cold/arid to humid, when densely forested areas developed. Megafauna might have survived until the early Holocene in refugia, becoming extinguished largely by climatic/environmental changes.

Andean and Western Peru

Caves in the central and southern highlands (3.5–4-m altitude) (Huarco, Sanson-Machay, Uchcumachay, Pikimachay) provide good examples of the maximal altitudinal distribution reached by megafauna (glyptodonts, ground-sloths, mastodonts-like, horses). These have been radiocarbon dated from ca. 20 to 13 ka BP ([Marshall et al., 1984](#); Aldenderfer, in [Zárate, 1999](#)).

Coastal outcrops (the rich Talara-La Brea tar-seeps, Cupisnique, La Cumbre, Pampa de los Fósiles, Quebrada El Jahuay) have yielded reptiles, amphibians, birds, and mammals, including megafauna. Some of these sites have been ¹⁴C-dated from 15 to 8 ka BP ([Marshall et al., 1984](#)) and to 26–15 ka BP by uranium-series dating (see [Falgüeres et al. \(1994\)](#) in [Table 2](#)). Most of the younger ¹⁴C ages are extremely controversial (Aldenderfer, in [Zárate, 1999](#)).

As the climate shifted from cool/humid to warm/dry, the development of coastal desert may have affected the fauna. There are few, if any, instances of human/megafauna interactions, suggesting humans had an insignificant effect in local extinction of the megafauna.

Western and Central Amazonia

Western Amazonia has riverside outcrops in the rainforest (Peru, Ecuador, and Brazil) that have been radiocarbon-dated

Table 2 Examples of extinct mammals with absolute ages at late Pleistocene or early Holocene in SA. All are ^{14}C ages unless otherwise indicated

<i>Taxon</i>	<i>Age</i>	<i>Material</i>	<i>Site</i>	<i>Source</i>
<i>Toxodon platensis</i>	6500 ESR	Teeth	Sao Paulo (Brazil)	<i>c</i>
<i>Toxodon platensis</i>	11 590 $\pm$ 90	Bone	Arroyo Seco (Pampean, Argentina)	<i>d</i>
<i>Equus (A.) neogaeus</i>	8890 $\pm$ 90	Bone	Arroyo Seco (Pampean, Argentina)	<i>d</i>
<i>Hippidion saldiasi</i>	10 860 $\pm$ 160	Bone	Cueva del Medio (southern Chile)	<i>e</i>
<i>Megatherium americanum</i>	7320 $\pm$ 50	Bone	Arroyo Seco (Pampean, Argentina)	<i>d</i>
	8390 $\pm$ 240			
<i>Eremotherium laurillardi</i>	11 340 $\pm$ 50	Bone	Pará (Brazil)	<i>f</i>
<i>Glossotherium wegneri</i>	12 350 $\pm$ 70	Bone	Quebrada Pistud (Ecuador)	<i>g</i>
<i>Glossotherium robustum</i>	10 500 $\pm$ 90	Bone	Arroyo Seco (Pampean, Argentina)	<i>d</i>
<i>Glyptodon clavipes</i>	4300 $\pm$ 90 ^a	Bone	Río Luján (Argentina)	<i>h</i>
<i>Doedicurus clavicaudatus</i>	6555 $\pm$ 160 ^b	Bone	La Moderna (Pampean, Argentina)	<i>d</i>
	12 350 $\pm$ 370			
<i>Scelidotherium</i>	16 000 $\pm$ 200 Th–U	Bone	Pampa de los fósiles (Perú)	<i>i</i>
<i>Scelidotherium</i>	9580 $\pm$ 200	Bone	Lapa Vermelha (Brazil)	<i>j</i>
<i>Catonyx cuvieri</i>	9990 $\pm$ 40	Bone	Central Brazil	<i>k</i>
<i>Nothrotherium</i>	12 200 $\pm$ 120	Coprolite	Gruta Brejões (Brazil)	<i>l</i>
<i>Stegomastodon waringi</i>	15 290 $\pm$ 70	Bone	Pará (Brazil)	<i>f</i>
<i>Smilodon populator</i>	9260 $\pm$ 150	Bone	Central Brazil	<i>k</i>

^aRejected by Cione et al. (2003).^bMinimum age.^cBaffa et al. (2000).^dPolitis et al. (1995).^eNami and Nakamura (1995).^fRossetti et al. (2004).^gColtorti et al. (1998).^hRossello et al. (1999).ⁱFalguères et al. (1994).^jProus and Fogaça (1999).^kAraujo et al. (2004).^lCzaplewski and Cartelle (1998).

from 44 ka BP (minimum ages) to 7.9 ka BP (Latrubesse, 2003). Open landscapes, though not necessarily grassland-dominated, have been proposed for the late Pleistocene, which agrees with the opinion that the current Amazonian rainforest is not a long-established ecosystem but one that has significantly expanded during the Holocene. Megamammals indicate the presence of savanna and a dry climate in the middle/early stages of the late Pleniglacial.

Central Amazonia has yielded megafauna dated to ca. 15–11 ka BP (Rossetti et al., 2004), when gradual humidity changes occurred, causing a shift from drier to arboreal savanna ca. 11.3 ka BP and to the current dense forest ca. 4.6 ka BP.

Northeastern and Central Brazil

In this region, there are rich fossiliferous limestone travertine caves, such as Toca da Boa Vista and Gruta dos Brejões, Bahia, Lagoa Santa and Lapa do Sumidouro, Minas Gerais, and Piauí, some with over 100 km of galleries.

Inferred fossil ages and stratigraphic relationships should be treated with caution due to stratigraphic inaccuracies in the cave deposits. The following ^{14}C dates have been provided: 20.06 and 12.2 ka BP (Toca da Boa Vista and Gruta Brejões) (see Czaplewski and Cartelle, (1998) in Table 2), ca. 11 cal ka BP to a minimum age of 8 cal ka BP (Lapa do Sumidouro). While empirical evidence suggests human presence in Brazil

ca. 11–12 ka BP, and sites ca. 10 ka BP show human/megafauna coexistence (see Prous and Fogaça, (1999) in Table 2), older settlements, ca. 45–30 ka BP, are still challenged.

Mammals include extinct and living species and suggest a moister and more forested environment than today, indicating a mosaic of habitats with open and forested areas.

Uruguay and Southern Brazil

Outcrops at river and creek sides in northern Uruguay (Gutiérrez, Malo, and Sopas creeks; Cuareim river) are rich in extant and extinct vertebrates, dominated by mammals and biostratigraphically correlated with the Lujanian Stage. There are radiocarbon dates at 11–12 ka BP and ranging up to >43 ka BP (minimum ages), whereas TL has produced age estimates of 43 and 58 ka BP (Ubilla and Perea in Tonni and Cioni, 1999; Ubilla et al. in Iriondo et al., 2004). The vertebrates are indicative of various habitats and tropical/temperate taxa suggest different climatic conditions compared to the typical Lujanian fauna, which was greatly influenced by the LGM. Sites in southern Uruguay (Santa Lucía river) have yielded vertebrates including megafauna and the radiocarbon dates range from 10.5 to 11.5 ka BP. However, there is no apparent association between megafauna and cultural remains in archaeological sites between 11.2 and 9.5 ka BP (Suárez and López, in Salemme and Miotti, 2003).

In Rio Grande do Sul (southern Brazil), open-air sites alongside rivers and creeks have predominantly yielded mammals, zoogeographically correlated with the Lujanian fauna. These have been dated by ^{14}C and TL from 8 to 30 ka BP (Oliveira in Tonni and Cione, 1999; Ubilla, 2004). Sites with mammals characteristic of tropical areas suggest a temperate/warm climate. In contrast, other localities have produced an austral fauna related to a dry/cold climate.

Grasslands dominated the southern Brazilian landscapes during the late Pleistocene when savanna and dry forest were replaced by grassland and subtropical gallery forest.

Central Argentina and Central-Southern Chile

Open-air and rock-shelter sites (Gruta del Indio, central Argentina; Quereo, Tagua-Tagua, central Chile; Monte Verde, central-southern Chile) provide both remarkable information about the early peopling of SA and important evidence concerning human predation on megafauna (mastodon-like, camelids). The radiocarbon dates range from ca. 30.8 to 8 ka BP and early human settlement could have occurred ca. 13 ka BP (Monte Verde) (Marshall et al., 1984; Borrero in Zárate, 1999; Alberdi and Prado, 2004). The paleozoogeographic differences with pampean Lujanian fauna were created by geographic isolation because of the Andean altitude.

A climatic shift from cold/wet to dry has been observed. This, along with human predation in central Chile, might have triggered the local megafauna extinction.

Pampean Area (Central-Eastern Argentina)

Sites on the flat grasslands have produced typical Lujanian fauna (Paso Otero, Quequén Salado, Tres Arroyos, Lobería, Luján River, etc.). There are many ^{14}C dates ranging from 29 to ca. 4.3 ka BP (Cione et al., 2003, and references therein; Table 2).

A cold and arid climate would have predominated during the late Pleistocene. Nevertheless, ca. 29 ka BP, the mammals indicate the presence of a warm/humid climate. In contrast, colder and more arid conditions prevailed from 18 to 10 ka BP due to the influence of the LGM, as witnessed in the vertebrate record from the cold/arid central Patagonian region. A replacement of dry steppe by humid grassland suggests a change from a subhumid to humid climate at the LP/EH boundary.

Relevant sites (Arroyo Seco, La Moderna, Cerro La China, Cueva Tixi cave) provide valuable information on the megafauna/human relationship. Radiocarbon dating of these sites spans the period 12.35–6.5 ka BP (Politis et al., 1995; Borrero et al. in Eriksen and Strauss, 1998). Despite criticisms of some of the youngest ^{14}C ages (see Table 2), it is likely that extinct mammals (glyptodonts, horses, and perhaps ground-sloths) were consumed by humans also in the early Holocene. Consequently, in the pampean area, early hunter-gatherers could have had more of an impact on megafaunal extinction than in the higher latitudes.

South American Southern Cone (Argentina and Chile)

Caves and open-air sites in southern Patagonia and Tierra del Fuego have yielded many paleontological and archaeological

assemblages: Cueva del Mylodon, Fell's, Palli Aike, De Los Chingues, Lago Sofia, and Del Medio caves and the rock shelter Tres Arroyos (southern Chile), and Las Buitreras Cave, Los Toldos, and Piedra Museo (southern Argentina). These sites have been ^{14}C -dated from 13.2 to 8.5 ka BP (Borrero et al., in Eriksen and Strauss, 1998; Alberdi and Prado, 2004). Birds and mammals, including megafauna, predominate. Despite having latitudinal differences, there are paleozoogeographic similarities between the western southern fauna and the pampean Lujanian one that were likely facilitated by areas of low Andean altitude.

These sites shed light on issues concerning high-latitude SA megafaunal extinction. Nevertheless, more geoarchaeological/taphonomic studies and first appearance datum analyses need to be performed in order to more precisely assess patterns of predation by humans.

There is no conclusive evidence that megafauna persisted into the Holocene, but those confidently dated to 13–12 ka BP are in general unassociated with humans (see Late Pleistocene Megafaunal Extinctions).

Horse and camel bones show evidence of human consumption, but if humans definitively occupied the southern cone by 11 ka BP, the interaction could have been no longer than 500 yr.

Before 11 ka BP, xeric scrub-steppe developed. Later, ca. 11–10 ka BP grassland including swamps, ponds, and flood-plains predominated under an arid/cold to humid/cold climate.

Fauna

A selected set of late Pleistocene SA taxa (predominantly mammals) is presented. Due to taxonomic controversy concerning many taxa, the most commonly used nomenclature is given.

The following information is compiled from: Marshall et al. (1984), Cartelle (1994), Alberdi et al. (1995), Arratia (1996), Tonni and Cione (1999), chapters of Cartelle and Rancy in Eisenberg and Redford (1999), Vrba and Schaller (2000), Bond et al. (2001), Croft (2001), Pardiñas et al. (2002), Bargo (2003), Cione et al. (2003), Alberdi and Prado (2004), Guerin and Faure (2004), Soibelzon et al. (2005), MacFadden (2005), Prado et al. (2005) and references therein. Paleobiological and anatomical reconstructions were created by G. Lecuona.

Chelonia

Large turtles (*Chelonoides*), the equivalent in size to Galapagos tortoises but with a thicker carapace, were widespread in SA and died out in the late Pleistocene. Despite being conspicuous in terrestrial vertebrate communities, unfortunately, their paleobiology and taxonomy are elusive (Figure 2).

Aves

In the Pampean region, avian fossils mostly belong to extant species. Nevertheless, an extinct rhea (*Rhea fossilis*) (Figure 3(a)), more slender than the recent ones, was probably widespread outside the pampas in lowland open country landscapes of SA. These flightless birds inhabited grasslands and also scrub forest and are useful paleoecological indicators.



Figure 2 Large terrestrial extinct turtles: *Chelonoides* (northern Uruguay). Photograph: Lecuona, G.

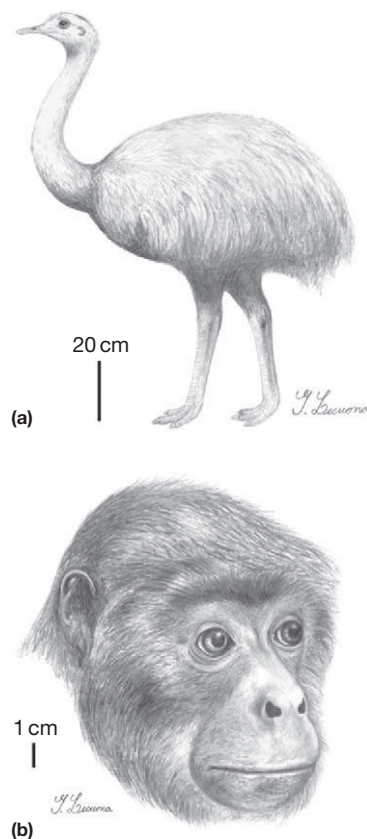


Figure 3 (a) An extinct rheid (flightless bird) from the pampas (*R. fossilis*); (b) large extinct monkey *P. brasiliensis* (Bahia State, Brazil).

In the Talara tar-seeps (coastal Peru), a rich assemblage of birds including tinamous, grebes, cormorants, storks, cranes, ducks, falcons, egrets, and lapwings was found.

Mammalia

Primates

There are remarkable records of two extinct ateline monkeys in Brazilian caves.

Protopithecus brasiliensis, at 25 kg in weight, more than twice the size of living SA monkeys, possessed a unique combination of cranial howler alouattine features (modified for an enlarged vocal sac) with postcranial spider and woolly spider ateline adaptations (for suspensory posture and a brachiating mode of locomotion) (Figures 3(b) and 4). *Caipora bambuiorum*, also bigger than recent atelines (>20 kg), is similar to spider-monkeys (without vocal sac).

Both required closed forest and suggest a more prominent expansion of forest in the late Pleistocene than exists today in northeast Brazil.

Xenarthra

These are a peculiar mammal group, including forms still living in SA, with remarkable features such as extra-articular surfaces in the backbone, few or no incisors, absent canines, cheek-tooth enamel usually absent, and bizarre fossil types. Xenarthrans include anteaters, tree- and ground-sloths, glyptodonts, and armadillos. The LP/EH extinction decimated mostly the strangest forms.

Glyptodonts

These armadillo-like animals were armored and tank-like, with a dorsal bony helmet, three-lobate straight and ever-growing grinding teeth, a bony rigid carapace (made of polygonal plates



Figure 4 Large extinct monkey *P. brasiliensis* (Bahia State, Brazil). Photograph: Cartelle, C. (Universidade Federal Minas Gerais-Brazil). Adapted from Cartelle (1994), page 84, with permission.

with surface designs that are taxon specific and therefore useful diagnostic characters) at the top and sides of the body, and an armored tail usually composed of rings and with a club-like end (Figures 5 and 6). The glyptodonts became extinct in the LP/EH, they were probably open-country grazers, and well-known species ranged in weight from 300 to 1500 kg.

Glyptodon clavipes, with an entirely ringed tail, is a frequent Lujanian megamammal along with *Panochthus tuberculatus*, with a clubbed tail (Figures 5(a), (b), 6(a), and (b)); *Doedicurus clavicaudatus*, one of the largest glyptodonts, with spikes in tail, was exploited by humans in the pampean region.

Armadillos and armadillo-like pampatheres

Successful xenarthrans are represented today by only nine armadillo genera, ranging from small- to large-sized (0.450–60 kg) and inhabiting both open and forested areas. Armadillos are diggers, scratchers, and are generally omnivorous. The carapace is not rigid as in glyptodonts; they have cephalic, shoulder, and hip shields constituted of bony scutes with movable bands in between. *Eutatus seguini*, a large armadillo (ca. 50 kg), disappeared in the LP/EH; *Propraopus*, a digger (ca. 50 kg) with a wide geographic distribution, went extinct in the late Pleistocene.

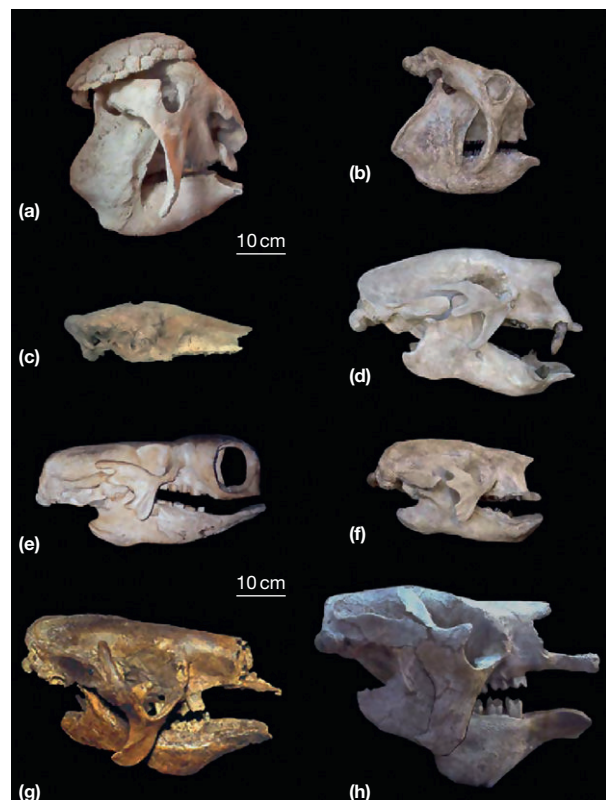


Figure 5 Late Pleistocene xenarthrans: (a) *P. tuberculatus* and (b) *G. clavipes* (southern Uruguay); (c) *P. humboldti* (northern Uruguay); (d) *L. armatus* (southern Uruguay); (e) *M. darwini* (Argentina); (f) *G. robustum* (southern Uruguay); (g) *E. laurillardii* (Minas Gerais, Brazil); (h) *M. americanum* (Argentina). Upper scale: (a–c); lower scale: (d–h) Photographs: a, b, d, f, h: Lecuona, G.; C: Ubilla, M.; E: Rinderknecht, A.; G: Cartelle, C. (UFMG-Brazil).

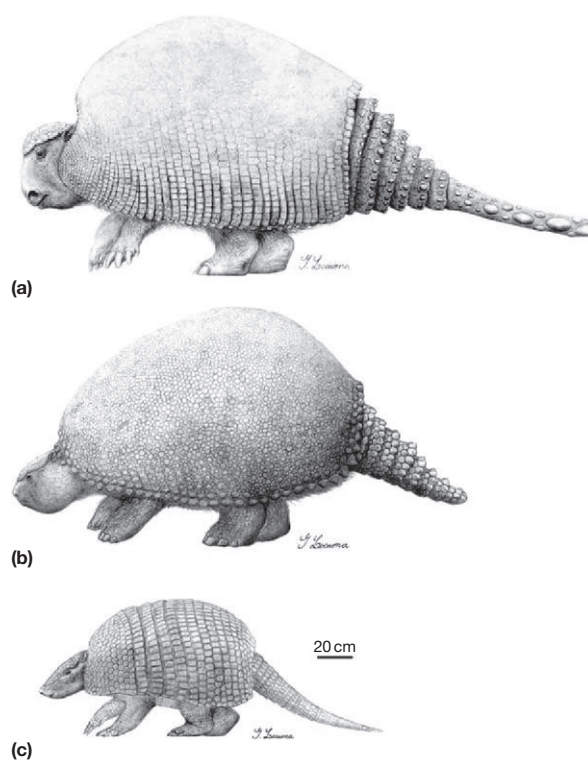


Figure 6 Glyptodonts: (a) *P. tuberculatus*; (b) *G. clavipes*. Pampatheres: (c) *P. humboldti*.



Figure 7 Late Pleistocene xenarthrans: *P. humboldti* (Brazil). Photograph: Ubilla, M.

The Pampatheres may be an intermediate group between armadillos and glyptodonts. They were large herbivorous animals (ca. 200 kg) and were predominantly grazers. Taxa such as *Pampatherium* and *Holmesina* measured more than 1 m in length. They spread across SA and became extinct in the LP/EH (Figures 5(c), 6(c), and 7).

Ground-sloths

Though only tree-sloths (4–9 kg) live today in the tropical forest zones of SA, the curious and intriguing large to mega-ground-sloths, clearly different from each other in their cranial morphologies, were successful in SA: *Nothrotherium maquinense*, *Scelidotherium leptocephalum*, *Glossotherium robustum*, *Myiodon darwini*, *Lestodon armatus*, and the giants *Megatherium*

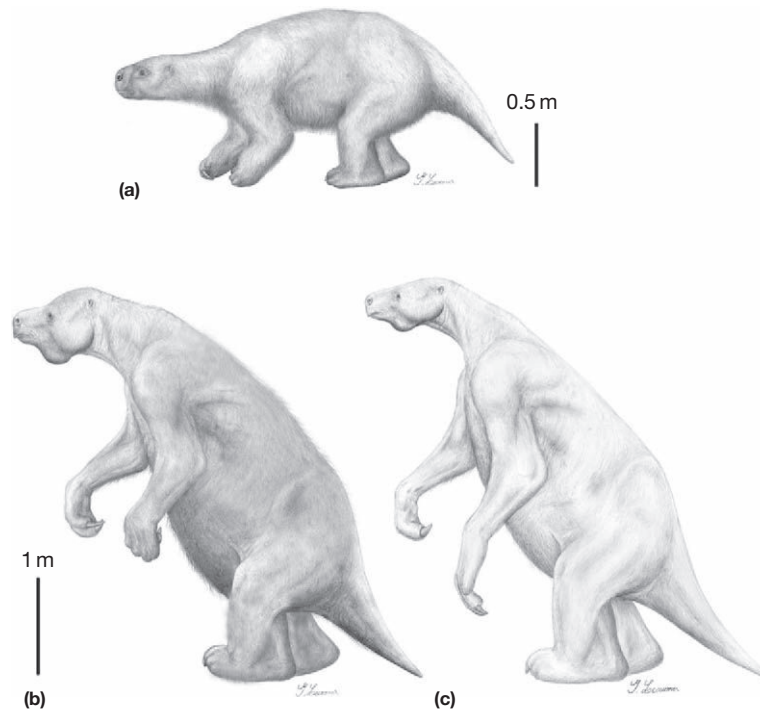


Figure 8 Ground-sloths: (a) *G. robustum*; (b) *M. americanum*; (c) *E. laurillardi*.

americanum and *Eremotherium laurillardi* (Figures 5(d)–(h) and 8). They ranged from 50 to ca. 10 000 kg of body mass and could have been bipedal. Ground-sloths had hair instead of a carapace and some had round ossicles embedded in the skin.

The morphological differentiation suggests a niche partitioning based on size, habitat, or diet. Traditionally viewed as clumsy, mostly browsing mammals, new interpretations suggest that they might have been more active, encompassing diggers (*S. leptocephalum*), grazers (*G. robustum*, *L. armatus*), mixed feeders (*Myiodon darwini*), and browsers (*S. leptocephalum*, *M. americanum*). All ground-sloths were wiped out in the LP/EH and *Megatherium* may have been preyed by humans in the pampean region.

Litopterna

Two families of this ancient and extinct SA ungulate group persisted in the late Pleistocene.

Macrauchenids

The large litopterns, a group of strange herbivores reminiscent of camels, originated in the SA Tertiary. *Macrauchenia patachonica* (ca. 1000 kg), the best-known late Pleistocene macrauchenid, became extinct in the LP/EH. It had relatively low-crowned teeth and stocky limbs with three functional toes; due to its elongated neck, it is usually reconstructed as similar to the extant giraffe. Over its long snout, the nasal bones were vestigial and nasal openings were located far back on the top of skull, which has led to an unresolved discussion over the presence of a proboscis (Figures 9(a), 10(a) and (b)). *Macrauchenia* could have been a mixed feeder. Traditionally thought of as amphibious, it is today considered to be a cursorial terrestrial mammal. It was widespread in the highland and

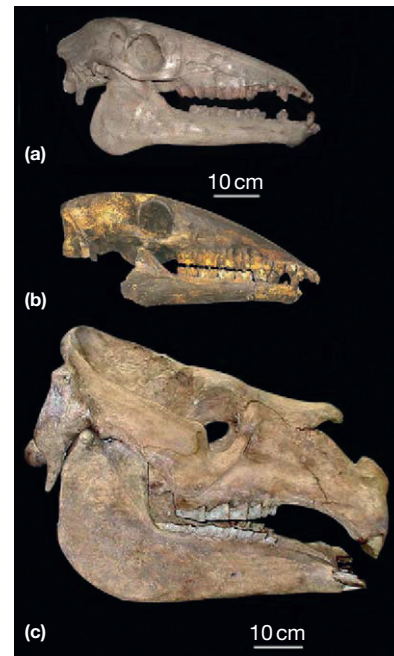


Figure 9 SA ungulates: (a) *M. patachonica* (Argentina). (b) *X. bahiense* (Brazil). Adapted from Cartelle (1994), page 96, with permission. (c) *T. platensis* (southern Uruguay). Photographs: (A) Reguero, M. (Museo de La Plata, Argentina); (B) Cartelle, (C) (UFMG-Brazil); (C) Lecuona, G.

lowland zones of SA and could have occupied diverse biotopes. *Xenorhinotherium bahiense* (Figure 9(b)), a slightly smaller close relative of *M. patachonica*, inhabited the intertropical landscapes of SA.

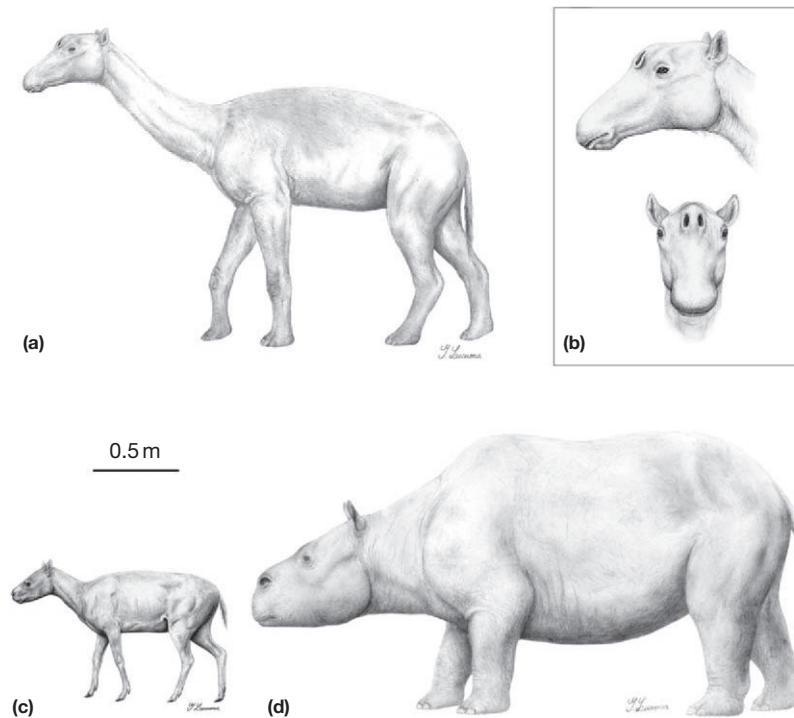


Figure 10 Southern ungulates: (a, b) *M. patachonica*, reconstructed without proboscis. (c) Proterotheres *N. recens*. Notoungulate: (d) *T. platensis*.

Proterotheres

Small- to medium-sized cursorial litopterns range in size from a rabbit to peccaries. They show a horse-like tendency to monodactyly and were commonly used as an example of an inadaptable limb pattern. Although 16 proterotheres genera diversified in the SA Tertiary, only one straggler persisted in the Pleistocene – *Neolicaphrium recens* (Figure 10(c)). The proterotheres were not an iteration of an ecological Holarctic group such as equids, but were a southern ‘evolutionary experiment,’ convergent with horses in their monodactyl condition and probably confined to a few warm forested relict microhabitats (Mesopotamian Argentina and Uruguay). The extinction of *N. recens* was apparently unrelated to the immigration of NA ungulates.

Notoungulata

Notoungulates were a group of highly diverse ungulates in the SA Tertiary, but during the late Pleistocene the diversity reduced, leaving only *Toxodon* and its relative *Mixotoxodon* to persist into early–middle Holocene (see Table 2).

T. platensis, the best-known species in Lujanian age, was a large and heavily built megaherbivore (ca. 2000 kg), a hippo- or rhino-like mammal with short nasal bones in a heavy skull (Figures 9(c) and 10(d)). Its short, stocky, and almost graviportal legs had three-hoofed toes. It possessed chisel-like incisors and rootless, permanently growing and high-crowned molars. It was widespread mainly in the eastern Andean lowland and, on account of its hippo-like aspect, was traditionally considered semiaquatic. Nevertheless, reappraisal of its postcranial morphology suggests a terrestrial open-country mode of life.

Despite having the morphological adaptations of an open-country grazer, stable isotope studies of dental enamel indicate

a broad range of dietary adaptations according to latitudinal local vegetation: forest browser (Amazonian), mixed feeder (Argentina and Brazil, Bahia State), to grazer (Chaco, Bolivia). Consequently, according to their principal food source, *Toxodon* could have inhabited different habitats.

Rodentia

Caviomorph rodents are represented by several taxa (Table 1), mostly extant. It is noticeable that the large extinct capybara *Nechoerus*, which was widespread over most of SA, is closely related but ca. 40% larger than the largest living rodent *Hydrochoerus*, weighing ca. 90 kg. A large and extinct porcupine (*Coendu magnus*) has been found in Brazil, Bolivia, and Uruguay (Figure 11); there are also extinct species of caviines, mainly found in the arid/semiarid landscapes of southern SA.

A considerable number of sigmodontine rodents, useful as ecological indicators, were found in SA including few extinct species (pampean area and northern Argentina; Talara, Perú; Tarija, Bolivia; and Lagoa Santa, Brazil).



Figure 11 Caviomorph rodents: *C. magnus* (Bahia, Brazil). Photograph: Ubilla, M.

Perissodactyla

Two lineages of ungulate mammals with a mesaxonic (i.e., having a central toe much enlarged that bears the weight) foot reached SA: horses and tapirs. Three species of tapir survive today in the tropical/temperate landscapes of SA but horses disappeared at the LP/EH boundary.

Horses

Two genera inhabited SA: *Hippidion* (three species; 250–460 kg), an endemic Pleistocene horse, and *Equus* (*Amerhippus*) (five species; 220–380 kg). The last genus, already well-established, colonized SA during the middle Pleistocene. Both genera of horses were robust and had comparatively large heads relative to their bodies.

Hippidion, aside from a different dental morphology, is characterized by retracted and stylized long nasal passages that clearly differentiate it from *Equus* (*Amerhippus*) (Figures 12, 13(a), and (b)).

Both developed similar ecological morphotypes (small species were presumably adapted to savannas, high-latitude areas, and mountain valleys, and large ones inhabited savannas) but *Equus* (*Amerhippus*) could have been more cursorial. Dental chemistry reveals that *Hippidion* was a mixed feeder and *Equus* (*Amerhippus*) predominantly a grazer to mixed feeder.

The small horse, *H. saldiasi*, frequently found in archaeological sites of southern SA, exhibits clear evidence of human predation.

Tapirs

Little is known about late Pleistocene tapirs in SA and remains are usually identified as *Tapirus terrestris*, the most widespread species in SA. Nevertheless, the diversity is likely underestimated; further fieldwork and better records may improve their taxonomic status and knowledge of their paleoecological adaptations at different latitudes. Today, modern tapirs live up to 28° S but in the late Quaternary they reached ca. 32°–35° S (pampean area and Uruguay).



Figure 12 NA invaders. Horses: (a) *H. principale* and (b) *E. (A.) neogeus* (Argentina). Photographs: Alberdi, M. T. (Museo Nacional Cs. Nat., Madrid). A and B adapted from Alberdi and Prado (2004), pages 109 and 140 respectively, with permission.

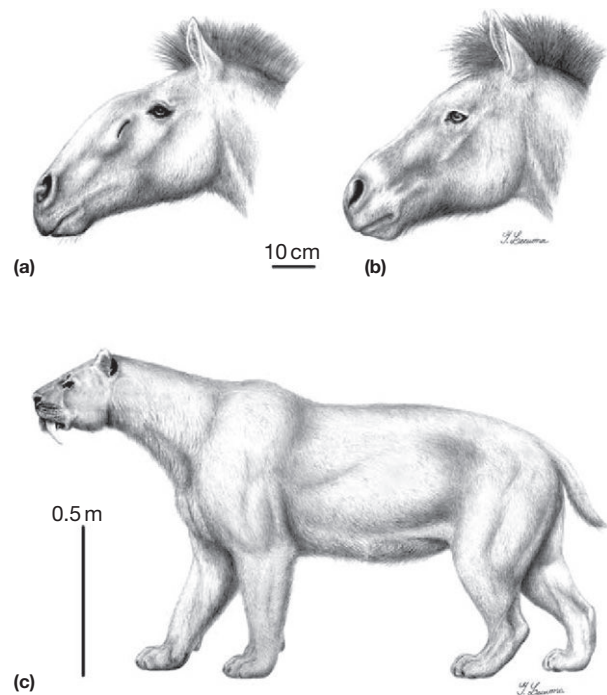


Figure 13 Horses: (a) *Hippidion principale*; (b) *E. (A.) neogeus*. Carnivore: (c) *Smilodon populator*.

Carnivora

Carnivora are represented in SA by a variety of successful small to large-sized immigrant mammals, that were mainly terrestrial but secondarily aquatic. These animals usually bore large canines and ranged in dietary adaptations from predominantly omnivorous to strictly carnivorous. Though the late Pleistocene extinction affected some taxa, they are conspicuous members of current mammal communities (cats, bears, foxes, raccoons, river otters, skunks).

Bears

SA bears differ from other bears in having short faces. The small spectacled bear (*Tremarctos*) only survives today in the northern-central mountains. In contrast, three species of medium-large-sized *Arctotherium* (300–700 kg) were present during the late Pleistocene. They were presumably carnivorous and/or omnivorous, were widespread over SA, and went extinct in the early Holocene.

Cats

Small to large felids, including the mountain lion (*Puma*) and the jaguar (*Panthera*), the largest predator in SA, were widespread in different highland and lowland habitats in the late Pleistocene and in recent times.

An extinct felid (ca. 400 kg), the saber-tooth *Smilodon populator*, possessed large blade-like upper canines in a broad muzzle (Figures 13(c) and 14(a)). It spread over most of SA and is generally considered a predator. Nevertheless, a different opinion claims a scavenging habit based on the premise that a functional constraint of the blade-like canine would not allow the bite of a proper predator.

Artiodactyla

NA immigrant ungulates with paraxonic arrangements of the foot (i.e. with the third and fourth metapodials larger than the others and bearing most of the weight) are represented in late Quaternary sites by three different families (deer, camels, and peccaries), which successfully diversified and became widespread in highland and lowland landscapes. Although some taxa disappeared in the late Pleistocene, they are still part of the modern mammal communities of SA.

Peccaries

The two genera living in SA today have a stocky body, relatively slender limbs, and are omnivorous. *Tayassu* (two species) are widespread over most of SA in diverse habitats; *Catagonus* (one species), the largest, is endemic of the Chaco (Argentina, Paraguay, and Bolivia), and inhabits semiarid xerophytic woodlands and thorn forests in a dry/warm climate. *Catagonus* and related forms have been recorded from Brazil, Uruguay, and Argentina. *Brasiliuchoerus stenocephalus*, from the caves of Brazil, is an extinct peccary close to *Catagonus* with a long and narrow snout (Figures 14(b) and 15(b)).

Camels (guanacos and vicuñas)

These are medium- to large-sized artiodactyls with a long neck and two toes in their lower limbs. Today, they live in arid/semiarid regions. The mixed-feeder guanaco (*Lama guanicoe*) (45–96 kg) is found at elevations ranging from sea level to 5 000 m in southern/western SA, and the grazer vicuña (*Vicugna vicugna*) (30–65 kg) occurs in the rolling grasslands and plains of highland Andean landscapes (3500–5700 m; Peru, Bolivia, and Chile). In the late Pleistocene, they were represented by more taxa and were widespread over most of SA including the intertropical and pampean lowlands, thereby

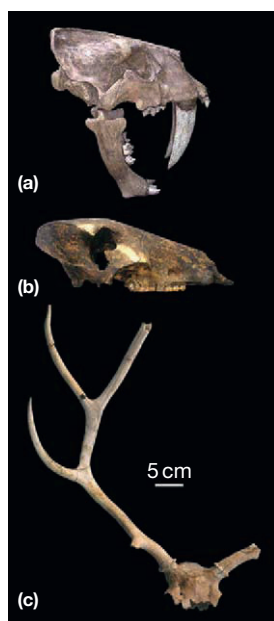


Figure 14 NA invaders. Carnivore: (a) *S. populator* (southern Uruguay). Artiodactyls: (b) *B. stenocephalus* (Brazil); (c) *P. fragilis* (northern Uruguay). Photographs: (a) Lecuona, G.; (b, c) Ubilla, M.

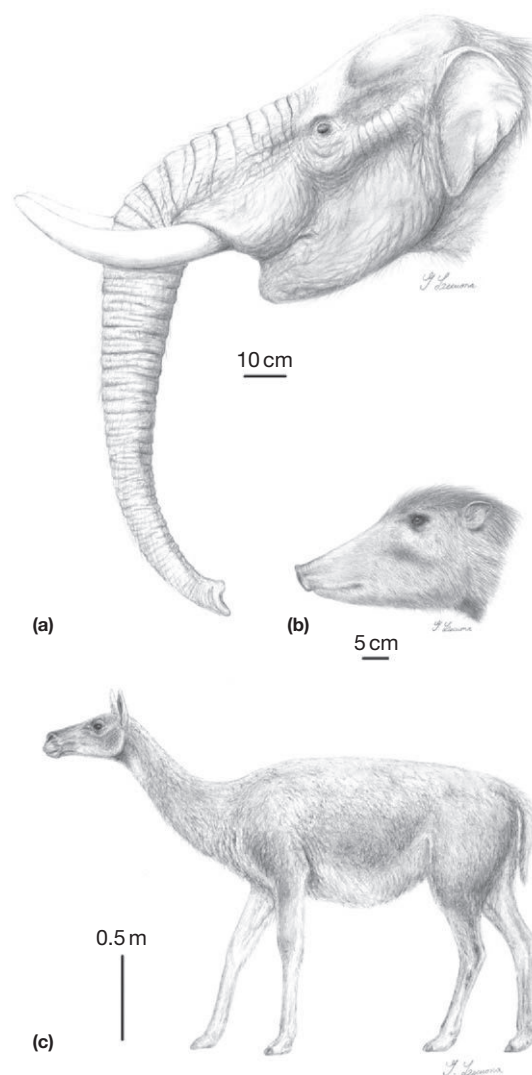


Figure 15 Gomphotheriid: (a) *S. platensis*. Artiodactyls: (b) *B. stenocephalus*; (c) *Hemiauchenia*.

suggesting that the current highland distribution is a secondary adaptation.

Large guanacos were also present during the late Pleistocene: the cursorial grazer *Hemiauchenia* (Figure 15(c)) and the likely browser *Paleolama* (ca. 400 kg) in the highland and lowland zones became extinct in the LP/EH. *Lama gracilis*, an extinct open-country grazer, intermediate in size between the guanaco and the vicuña, was described from pampean area. The presence of the guanaco in Amazonia and northeastern Brazil is remarkable and unexpected.

Extinct and extant SA camels were predated by humans in the LP/EH (Chile and Argentina).

Deer

The deer underwent an explosive radiation in SA occupying a broad range of habitats in the lowlands and highlands. They are mostly represented in sites by antlers and sometimes dominate the record (see sites of northern Uruguay in the



Figure 16 NA invaders. Gomphotheriids: *S. waringi* (Quebrada Pistud, Ecuador). Photograph: Marco Ferretti (Museo di Storia Naturale, Università di Firenze, Italy).

late Pleistocene). Living genera have been reported in the late Pleistocene (*Pudu*, *Odocoileus*, *Mazama*, *Ozotoceros*, *Blastoceros*, *Hippocamelus*), ranging from the dwarf *Pudu* (9 kg) to the large *Blastoceros* (ca. 100 kg). Extinct taxa such as the large *Antifer* and medium-sized *Morenelaphus* and *Paraceros* (Figure 14(c)) contributed to the largest diversity of herbivorous artiodactyls in SA.

Proboscidea

Two gomphother genera, derived from a NA lineage, inhabited SA and became extinct in the late Pleistocene. *Cuvieronius* (one species) was small and distributed in Andean landscapes, whereas *Stegomastodon* (two species) was larger and was widespread, mainly in the lowlands. These megaherbivores had a short skull and mandible, with elongate upper tusks (Figures 15(a) and 16).

Cuvieronius might have been a mixed feeder in cold/temperate climates; *Stegomastodon*, mostly related to warm/temperate climates, could have had a broader range of dietary preferences (browser to grazer) according to its altitudinal/latitudinal distribution.

Humans utilized gomphotheres as a food source in Venezuela, Colombia, and Chile.

LGM Effect

Regardless of the fact that the cone shape of SA might have affected the timing of the LGM cooling, it is possible to confidently demonstrate some vertebrate responses to the following major processes:

1. Lowland landscapes at low/median latitudes underwent climatic/geographic changes in the middle and late pleni-glacial. Despite there being a controversy over the effect of the LGM in the Amazonian forest, there is empirical evidence that open-country mammals (guanacos, glyptodonts, macrauchenids) colonized intertropical areas (Cartelle in Eisenberg and Redford, 1999; Latrubesse, 2003).
2. A severe aridification and cooling occurred both in the highlands and in the lowlands at different latitudes. Megamammals migrated from the Ecuadorian highlands to the lowlands (Ficcarelli et al., 2003). Aridification, cooling, and an expansion of open ground promoted a south–north shifting of cold/arid and open landscape. ‘Southern taxa’

such as the Patagonian rhea *Pterocnemia*, the marsupial *Lestodelphis*, rodents such as *Dolichotis patagonum* and *Microcavia*, and the guanaco *L. guanicoe* were reported in pampean landscapes (Cione et al., 2003).

3. The post-LGM warming caused a modification of a range of megamammals and favored survival or expansion of forest fauna (Cione et al., 2003).

Table 3 Some large and megamammals that became extinct in SA at the late Pleistocene and LP/EH boundary

Order XENARTHRA	Order NOTOUNGULATA
Family Dasypodidae (armadillos)	Family Toxodontidae
<i>Propraopus</i>	<i>Toxodon</i> (AS)
<i>Eutatus</i> (AS)	<i>Mixotoxodon</i>
Family Pampatheriidae	Order LITOPTERNA
(armadillo-like)	Family Macrauchiidae
<i>Pampatherium</i>	<i>Macrauchenia</i> (AS)
<i>Holmesina</i>	<i>Xenorhinotherium</i>
Family Glyptodontidae	Family Proterotheriidae
(glyptodonts)	<i>Neolicaphrium</i>
<i>Doedicurus</i> (AS)	Order PERISSODACTYLA
<i>Glyptodon</i>	Family Equidae (horses)
<i>Neosclerocaliptus</i> (AS)	<i>Hippidion</i> (AS)
<i>Neothoracophorus</i>	<i>Equus</i> (<i>Amerhippus</i>) (AS)
<i>Panochthus</i>	Order ARTIODACTYLA
<i>Parapanochtus</i>	Family Tayassuidae
<i>Neuryurus</i>	(peccaries)
<i>Plaxhaplous</i>	<i>Brasiliochoerus</i>
Family Megatheriidae (ground sloths)	Family Camelidae (camels)
<i>Megatherium</i> (AS)	<i>Hemiauchenia</i> (AS)
<i>Eremotherium</i>	<i>Paleolama</i> (AS)
Family Nothrotheriidae (ground sloths)	<i>Eulamaops</i>
<i>Nothrotherium</i>	Family Cervidae (deer)
<i>Nothropus</i>	<i>Antifer</i>
Family Megalonychidae (ground sloths)	<i>Morenelaphus</i>
<i>Megalonychops</i>	<i>Paraceros</i>
<i>Ocnopus</i>	Order CARNIVORA
<i>Valgipes</i>	Family Ursidae (bears)
Family Mylodontidae (ground sloths)	<i>Arctotherium</i>
<i>Glossotherium</i> (AS)	Family Felidae (sabretooth, jaguar)
<i>Lestodon</i>	<i>Smilodon</i>
<i>Mylodon</i> (AS)	Family Canidae (dogs, foxes)
<i>Mylodonopsis</i>	<i>Procyon</i>
<i>Ocnotherium</i>	<i>Theriodictis</i>
<i>Scelidotherium</i>	Order RODENTIA
<i>Oreomyodon</i>	Family Hydrochoeridae
<i>Catonyx</i>	(capybaras)
Order PROBOSCIDEA	<i>Nechoerus</i>
Family Gomphotheriidae	
(mastodonts)	
<i>Cuvieronius</i> (AS)	
<i>Stegomastodon</i>	
Order PRIMATES	
Family Cebidae	
<i>Protopithecus</i>	
<i>Caipora</i>	

AS. Compiled from Cione et al. (2003) and Cartelle (1999).

South American LP/EH Extinction

In contrast to Africa, where megafauna are still alive, in SA 80% of large mammals weighing over 45 kg and 100% of megamammals (>1 000 kg in weight), giant turtles, and some medium-sized mammals became extinct at the LP/EH boundary (Table 3; see [Late Pleistocene Megafaunal Extinctions](#)). Both immigrant and native mammals (mostly xenarthrans), adapted or presumably adapted to open landscapes, comprising ca. 50 genera of the large Lujanian fauna, died out. The few surviving large mammals (14 species) are predominantly adapted to forest, highlands, and wetlands. The causes of extinction in SA are still debated and many factors are invoked ([Cione et al., 2003](#); [Alberdi and Prado, 2004](#)).

Only the large and megamammal extinctions seem to correlate with climatic or environmental changes, and the larger the animal, the more susceptible it was ([Alberdi and Prado, 2004](#)).

Humans certainly hunted horses, mastodon-like pachyderms, large camelids, some glyptodonts, and ground sloths in some parts of SA. Although it is unlikely that human predation was the sole cause of extinction of all large/megafauna, it may have reduced population sizes of some large/megamammals, particularly in extreme environmental conditions.

See also: [Archaeological Records: Global Expansion 300 000–8000 Years Ago, Africa](#); [Pollen Records, Late Pleistocene: South America](#); [Vertebrate Records: Late Pleistocene Megafaunal Extinctions](#).

References

- Alberdi MT and Prado JL (2004) Caballos fósiles de América del Sur. Una historia de tres millones de años. *Serie Monográfica Instituto Investigaciones Arqueológicas y Paleontológicas Cuaternario Pampeano olavarría Argentina. Facultad de Ciencias Sociales*.
- Alberdi MT, Leone G, and Tonni EP (1995) Evolución biológica y climática de la Región Pampeana durante los últimos cinco millones de años. Un ensayo de correlación con el Mediterráneo occidental. *Monografías Museo Nacional de Ciencias Naturales de Madrid. Madrid, España: Consejo Superior de Investigaciones Científicas*.
- Araújo A, Neves W, and Piló L (2004) Vegetation changes and megafaunal extinction in South America: Comments on de Vivo and Carmignato. *Journal of Biogeography* 31: 2039–2040.
- Arratia G (1996) Contributions of Southern South America to Vertebrate Paleontology. *Münchner Geowissenschaftliche Abhandlungen*, 30: Verlag: München, Germany.
- Baffa O, Brunetti A, Karmann I, and Martins Dias Neto C (2000) ESR dating of a toxodon tooth from a Brazilian karstic cave. *Applied Radiation and Isotopes* 52: 1345–1349.
- Bargo S (2003) Biomechanism and paleobiology of the Xenarthra: The state of the art (Mammalia, Xenarthra). *Senckenbergiana Biologica* 83: 1–8.
- Bond M, Perea D, Ubilla M, and Tauber A (2001) *Neolophoceros recens* Frenguelli, 1921, the only surviving Proterotheriidae (Litopterna, Mammalia) into de South American Pleistocene. *Palaeovertebrata* 30: 37–50.
- Cartelle C (1994) *Tempo Passado. Mamíferos do Pleistoceno em Minas Gerais*. Belo Horizonte, Brazil: Editora Palco.
- Cartelle C (1999) Pleistocene mammals of the Cerrado and Caatinga of Brazil. In: Eisenberg J and Redford K (eds.) *Mammals of the Neotropics. The center Neotropics: Ecuador, Perú, Bolivia, Brazil*, vol. 3, p. 27. Chicago, USA: University of Chicago Press.
- Cione A, Tonni EP, and Soibelzon L (2003) The broken zig-zag: Late Cenozoic large mammals and tortoise extinction in South America. *Revista Museo Argentino Ciencias Naturales* 5: 1–19.
- Coltorti M, Ficcarelli G, Jähren H, Moreno-Espinosa M, Rook L, and Torre D (1998) The last occurrence of Pleistocene megafauna in the Ecuadorian Andes. *Journal of South American Earth Sciences* 11: 581–586.
- Croft D (2001) Cenozoic environmental change in South America as indicated by mammalian body size distributions (cenograms). *Diversity and Distribution* 7: 271–287.
- Czaplewski N and Cartelle C (1998) Pleistocene bats from cave deposits in Bahia, Brazil. *Journal of Mammalogy* 79: 784–803.
- Eisenberg J and Redford K (1999) *Mammals of the Neotropics Vol. 3. The Center Neotropics. 3. Ecuador, Peru, Bolivia, Brazil*. Chicago, USA: University of Chicago Press.
- Eriksen B and Strauss L (1998) As the world warmed: Human adaptations across the Pleistocene/Holocene boundary. *Quaternary International* 49(50): 1–199.
- Falguères C, Fontugne M, Chauchat C, and Guadelli JL (1994) Datations radiométriques de l'extinction des grandes faunes pléistocènes au Pérou. *Compte Rendu Académie Science Paris* 319(II): 261–266.
- Ficcarelli G, Coltorti M, Moreno-Espinosa M, Pieruccini PL, Rook L, and Torre D (2003) A model for the Holocene extinction of the mammal megafauna in Ecuador. *Journal of South American Earth Sciences* 15: 835–845.
- Flynn J, Wyss A, Croft D, and Charrier R (2003) The Tinguiririca fauna, Chile: Biochronology, paleoecology, biogeography, and a new earliest Oligocene South American land mammal age. *Palaeogeography, Palaeoclimatology, Palaeoecology* 195: 229–259.
- Guérin C and Faure M (2004) *Macrauchenia patachonica* Owen (Mammalia, Litopterna) de la région de Sao Raimundo Nonato (Piauí, Nordeste brésilien) et la diversité des Macrauchiinae pléistocènes. *Geobios* 37: 516–535.
- Hoffstetter R (1952) Les mammifères pléistocènes de la République de l'Equateur. *Mémoires Société Géologique de France* 31: 1–391.
- Iriondo M, Kröhlting D, and Stevaux J (2004) Advances in the Quaternary of the De La Plata River Basin, South America. *Quaternary International* 114: 1–181.
- Latrubesse E (2003) The late-Quaternary paleohydrology of large South American fluvial system. In: Gregory K and Benito G (eds.) *Palaeohydrology: Understanding Global Change*, p. 193. New York, USA: Wiley.
- MacFadden B (2005) Diet and habitat of toxodont megaherbivores (Mammalia, Notoungulata) from the late Quaternary of South and Central America. *Quaternary Research* 64: 113–124.
- Marshall LG, Berta A, Hoffstetter R, et al. (1984) Mammals and stratigraphy: Geochronology of the continental mammal-bearing Quaternary of South America. *Palaeovertebrata Mémoire Extraordinaire* 1–76.
- Nami HG and Nakamura T (1995) Cronología radiocarbónica con AMS sobre muestras de hueso procedentes del sitio Cueva del Medio (Ultima Esperanza, Chile). *Anales del Instituto de la Patagonia* 23: 125–133.
- Pardiñas U, D'Elia G, and Ortiz P (2002) Sigmodontinos fósiles (Rodentia, Muroidea, Sigmodontinae) de América del Sur: estado actual de su conocimiento y prospectiva. *Mastozoología Neotropical* 9: 209–252.
- Pascual R, Ortiz-Jaureguizar E, and Prado J (1996) Land Mammals: Paradigm for Cenozoic South American geobiotic evolution. *Münchner Geowiss. Abhandlungen* 30: 265–319.
- Politis GG, Prado JL, and Beukens R (1995) The human impact in Pleistocene–Holocene extinctions in South America. The Pampean case. In: Johnson E (ed.) *Ancient Peoples and Landscapes*, p. 187. Lubbock: Museum of Texas Tech University.
- Prado JL, Alberdi MT, Azanza B, Sánchez B, and Frassinetti D (2005) The Pleistocene Gomphotheriidae (Proboscidea) from South America. *Quaternary International* 126(128): 21–30.
- Prous A and Fogaça E (1999) Archaeology of the Pleistocene–Holocene boundary in Brazil. *Quaternary International* 53(54): 21–41.
- Rancy A (2000) *Paleoecologia da Amazonia. Megafauna do Pleistoceno*. Florianópolis, Brazil: Editora da Universidade Federal Santa Catarina.
- Rossello E, Bor-Ming J, Liu TK, and Petrocelli JL (1999) New 4,300 YR ¹⁴C age of glyptodonts at Luján river (Buenos Aires, Argentina) and its implications. *Actas Segundo Simposio Sudamericano de Geología Isotópica* 1: 105–110.
- Rossetti DF, Man de Toledo P, Moraes-Santos H, and Araújo A (2004) Reconstructing habitats in central Amazonia using megafauna, sedimentology, radiocarbon, and isotope analyses. *Quaternary Research* 61: 289–300.
- Salemm MC and Miotto LL (2003) South America: Long and winding roads for the first Americans at the Pleistocene/Holocene transition. *Quaternary International* 109(110): 1–179.
- Santos G, Bird M, Parenti F, Fifield L, Guidon N, and Hausladen P (2003) A revised chronology of the lowest occupation layer of Pedra Furada Rock Shelter, Piauí, Brazil: The Pleistocene peopling of the Americas. *Quaternary Science Reviews* 22: 2303–2310.
- Simpson GG (1980) Splendid Isolation. *The Curious History of South American Mammals*. Yale University Press.

- Soibelzon L, Tonni EP, and Bond M (2005) The fossil record of South American short-faced bears (Ursidae, Tremarctinae). *Journal of South American Earth Sciences* 20: 105–113.
- Tonni EP and Cione L (1999) Quaternary vertebrate paleontology in South America. *Quaternary South America and Antarctic Peninsula* 12: 1–310.
- Ubilla M (2004) Mammalian biostratigraphy of Pleistocene fluvial deposits in northern Uruguay, South America. *Proceedings of the Geologist's Association* 115: 347–357.
- Vrba E and Schaller G (2000) *Antelopes, Deer, and Relatives. Fossil Record, Behavioral Ecology, Systematics, and Conservation*. Yale University Press.
- Webb SD (1991) Ecogeography and the great American interchange. *Paleobiology* 17: 266–280.
- Zárate M (1999) The Pleistocene/Holocene transition and human occupation in South America. *Quaternary International* 53(54): 1–105.

Late Pleistocene of Southeast Asia

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Introduction

The late Pleistocene period begins 126 ka with the base of the Eemian interglacial stage, corresponding to the base of marine oxygen isotopic substage 5e. It also includes the last glacial episode of the Pleistocene and ends with the beginning of the Holocene, 11.5 ka. Due to high amplitude climatic fluctuations induced by cyclical glaciations alternating with interstadials, this period represents a key for the understanding of the distribution and composition of extant mammalian communities and for the evolution of modern humans, *Homo sapiens*.

Biogeography of Southeast Asia

Today, two biogeographic provinces are distinguished in Southeast Asia. The Indochinese Province or Northern Province includes many mammals that do not extend west into India or north into temperate China, but some extend southward, as far as Malaysia and the Greater Sunda Islands. The Sundaic or Southern Province contains many endemic genera (Corbet and Hill, 1992). The boundary between these two provinces is located at the Isthmus of Kra in Thailand, at about 10° 30' N (Figure 1). This boundary was initially based upon vegetation changes with evergreen rainforest in the south and more seasonal forest in the north. Most animal species' distributions are affected by this boundary, but its importance varies according to the different taxonomic groups. The mammalian boundary is also located slightly north of the Isthmus of Kra (Tougaard, 2001; Woodruff, 2003). A cluster analysis by Tougaard (2001) indicated that large mammal faunas of Thailand north of Kra Isthmus group with those of the Indochinese Province, whereas faunas south of Kra Isthmus group with those of the Sundaic Province. This was initially demonstrated by Chaimanee (1998) for rodents, which show a more significant change at the Kra Isthmus than large mammals.

Southeast Asia has undergone dramatic climatic and geographic changes during the Pleistocene. It is supposed to have been significantly drier and cooler with more seasonal climates during glacial periods (An, 2000). These climatic oscillations are associated with repeated shrinking and expansion of rainforests and correlative spreading and shrinking of savannas. According to Morley (2000), pine woodlands and savanna grasslands expanded and replaced the rainforest over large areas of Southeast Asia during the glacial periods. Such a change corresponds to a downward shift of vegetation zones, estimated of 600–1000 m. Simultaneously, glacioeustatic drops in sea level enlarged the surface of the Thai–Malay Peninsula and connected nearly all the islands together. The sea-level drop is estimated to have been at least 100 m (Vorisi, 2000) for the Last Glacial Maximum (LGM) (20 ka), causing the Sunda Shelf to emerge and the Andaman Sea to retreat.

Such a change is supposed to have had a strong impact on the monsoonal climate which led to the development of a savannah corridor in the central flood plain of the Sunda Shelf. The distribution of teak forest (*Tectona grandis*), which widely occurs today in the Northern Province but with isolated patches in Java, supports this hypothesis. Although most authors agree that during the coldest and driest glacial periods, including the LGM, some rainforest refuges existed, they disagree about their location (Kershaw et al., 2001; Gathorne-Hardy et al., 2002; Meijaard, 2003).

Surprisingly, the late Pleistocene period is probably the less well-known period concerning its climate and vegetation, but also concerning its sedimentary deposits, either from caves or from fluvial terraces. In Thailand, among more than 20 investigated fossiliferous caves and fissure fillings, four of them had been assigned to the late Pliocene or the early Pleistocene and all others to the late middle Pleistocene (see Mid-Pleistocene of Southern Asia) (Figures 2 and 3). Only one archaeological site could be assigned to the late Pleistocene (Chaimanee, 1998). Cave filling is the result of a sedimentary process that occurs during phases of heavy rainfall or of tectonic subsidence. At some other periods, erosion is dominant because of the drop of sea level or tectonic uplift. Therefore, the fossiliferous cave sediments document the most recent episode of infilling. In some rare cases, remnants of older filling events can be found as relictual patches left by erosion on the lateral walls of the caves, but they can only be identified if they yield older fossils. Speleothems (calcite deposits) are rarely contemporaneous to the fossiliferous sediments and are usually more recent than fossils. As cave fillings, they are polycyclic and their dating is of little use to date the fossiliferous cave deposits unless they clearly cover the sediments like stalagmitic floors. Terrace deposits are also dependent on climate and tectonic activity (Figures 4 and 5). During periods of low sea levels, river channels cut the former terraces deposits and develop new terraces of minimal width which may be easily eroded during later Holocene flooding. Therefore, the scarcity of late Pleistocene deposits in Southeast Asia can be explained by a peculiar climatic setting. The climate was colder and drier, and these characteristics have been confirmed by the fossil pollen record in Java and Thailand and by the larger than average size of mammals that have extant representatives in the same area. Palynological data are extremely scarce, and most authors refer to the study of Van der Kaars and Dam (1995) in Java where a core documenting the vegetation of the last 135 ka has been investigated. Nevertheless, despite its great interest, it cannot be simply extrapolated to other areas. Recently, however, new data have been provided concerning Thailand, which confirm the extension of a temperate forest during the cooler and drier LGM (Penny et al., 2001; White et al., 2004).

In the light of these considerations, the late Pleistocene mammalian fossil records of Southeast Asia will be reviewed.



Figure 1 Indochinese and Sundaic Province and the Kra Isthmus.



Figure 2 Late Pleistocene and Late middle Pleistocene vertebrate localities in Southeast Asia mentioned in the text.

Late Pleistocene Mammals in Southeast Asia

Sundaic Province

Java

The rich *Homo erectus* record in Java has stimulated much paleontological and biochronological research. However, a fluctuating sea level has caused island isolation and led to



Figure 3 Mammalian community of the middle and late Pleistocene in Thailand.



Figure 4 Cave deposits in limestone are source of Pleistocene fossils in Southeast Asia.



Figure 5 Another important source for Pleistocene mammalian fossils is terrace deposits.

a high endemism, making it difficult to develop a biochronological timescale. Nevertheless, a consensus exists currently that the Kedung Brubus faunal stratum corresponds to a middle Pleistocene faunal assemblage. It is followed by the highly

controversial Ngandong fauna, then by Punung and finally by Wajak, which represents the last faunal assemblage before the extant fauna comes into existence.

The Punung fauna, Wajak fauna, and perhaps also the older Ngandong fauna are considered to be of late Pleistocene age.

Punung faunal assemblage

De Vos (1983) and Sondaar (1984) defined the Punung faunal strata on the first appearance datum (FAD) of the extant Asian elephant (*Elephas maximus*) and the extant pig species (*Sus vittatus*). The Punung Caves fauna has yielded 18 mammalian species including *H. sapiens*. The composite faunal list, from at least two separate caves, I and II, contains no extinct species. The occurrence of orangutans (*Pongo pygmaeus*), siamang (*Hylobates syndactylus*), and sun bear (*Ursus malayanus*), taxa not known to cross water barriers, suggests that a land connection existed at that time with mainland Southeast Asia and that the community lived in a rainforest environment. As orangutans and modern humans are unrecorded from older faunal strata, one can therefore conclude that they are immigrants from the mainland (Van den Bergh et al., 2001). Therefore, this rainforest faunal association probably corresponds to the end of the last interglacial or the beginning of the last glacial.

According to this interpretation, one could also add orangutans and modern humans to the previous list of late Pleistocene FAD on Java. Another interesting point is that if this fauna is really from the terminal part of the last interglacial, it supports the argument that the sea level was, at that time, at least 60–80 m lower than at present day, to permit land mammal exchanges with the Indochinese Province, where some of these taxa (Asian elephant and orangutans) have been documented from older localities. Tham Wiman Nakin Cave in Thailand, for example, which is older than 169 ka, has yielded Asian elephant. It is unfortunate that no absolute ages are available for Punung I and II localities, which have yielded early *H. sapiens*. The presumed age of the Punung fauna is 80–60 ka. Van den Bergh et al (2001) also mention the discovery of Asian elephant from the Cipeundeuy sand pit in Java, located next to Bandung, dated between 35.5 and 29.6 ka by ^{14}C (Van den Bergh, 1999). The site has also yielded a few modern taxa that are still present on that island.

Wajak faunal assemblage

The Wajak fauna is of early Holocene age and has been dated to about $10.56 \pm .075$ ka BP (Storm, 1995). The fauna is modern with 10 large mammal species including *H. sapiens* and four rodent species. *Elephas* is absent and tapir (*Tapirus indicus*) has its last appearance datum (LAD) on the island. According to De Vos (1983) the fauna indicates an open woodland environment and shows that rainforest must therefore have disappeared before 10 ka BP from that area because it lacks the species most strongly adapted to the humid forest.

Ngandong faunal assemblage

The Ngandong faunal composition appears largely similar to the Kedung Brubus fauna. Of the 15 species of large mammals, 9 are extinct and shared with Kedung Brubus (Sondaar, 1984). However, several species, including *H. erectus*, differ at the subspecific level, suggesting a younger age. Van den Bergh et al. (2001) therefore propose an intermediate age of about 135 ka for that faunal assemblage. In contrast, Swisher et al.

(1996) U/Th series dating on Ngandong fossil teeth gave an age between 53 and 27 ka. One dramatic consequence of that absolute age is the putative contemporaneity and sympatry of late *H. erectus* (Ngandong man) with early *H. sapiens*, which has so far never been observed elsewhere in Asia, with the possible exception of Flores Island (Morwood et al., 2004). However, the U/Th series dating method is not accurate in tropical regions where the weathering of dentine and enamel is significant and diagenesis is considerable (Esposito et al., 2002). For these reasons, the attribution of the fauna to the early late Pleistocene by Van den Bergh (1999) is preferred here.

Sumatra

The material collected by Dubois, as early as 1888, originates from three distinct caves: the Lida Ajer Cave near Pajakombo, the Sibrambang Cave, and the Djamboe (Jambu) Cave near Tapisello. The fauna was subsequently studied by Hooijer (1947), who assigned Holocene age to these faunas. However, Drawhorn (1994) used amino acid racemization to date the Lida Ajer Cave fauna from 80 to 60 ka. The Lida Ajer Cave has yielded two teeth of *H. sapiens*, one upper central incisor and one upper left molar associated with 16 other mammalian taxa, including Asian elephant, orangutan, *Hylobates*, all taxa indicative of a rainforest environment. Almost all of these fossil mammal taxa are also present in Sibrambang Cave and have been described by Hooijer in several publications between 1947 and 1962. De Vos (1983) also mentioned additional taxa such as *Rusa* sp., *Muntiacus muntjak*, *S. vittatus*, and *S. barbatus*.

Borneo

The late Pleistocene of Borneo is famous for its historical discovery of the first Southeast Asian modern human skull and associated rich mammalian fauna at Niah Cave (see *Interactions with Hominids*). Several levels have been excavated in recent years and have yielded abundant mammal fossils, from large mammals to rodents (Medway, 1972), including *H. sapiens* (Figure 6). The sequence possibly dated back to more than 40 ka. One white marker bed, which may correspond to ash from the Toba eruption, has been dated to 32.6 ka BP. From the fossiliferous level, 30 extant species of mammals, including six primates (*H. sapiens*, orangutan, *Hylobates*,



Figure 6 Dry screen method was used to separate fossils from cave sediments.

Macaca, *Semnopithecus*, and *Nycticebus*) have been reported with more than eight rodents and insectivores and more than ten species of bats. Asian elephant is also present and, with orangutan, characterizes the Sundaic late Pleistocene faunal assemblage. Only one extinct species, a pangolin, *Manis paleojavanica*, occurs, and the overall assemblage is characteristic of rainforest. Therefore, it demonstrates that this part of Borneo was a rainforest refuge at that time and that *H. sapiens* were already adapted to rainforest environments.

Flores

Island endemic vertebrates have been recorded for many years from Flores (Sondaar et al., 1994; Van den Bergh et al., 1996; Morwood et al., 1998) including dwarf *Stegodon*, giant rat, *Hooijeromys*, and Komodo dragon. However, the recent discovery, in the Liang Bua Cave of a dwarf human skeleton described under the name *H. floresiensis*, interpreted as an island dwarf endemic *H. erectus* and dated to 18 ka, changes the understanding of the faunal evolution of that island. The dwarf *Homo* is associated with an endemic fauna, with a dwarf *Stegodon*, a Komodo dragon, and endemic rats. According to the different absolute dates provided by Morwood et al. (2004), this dwarf human and associated fauna occurs during a time interval spanning 74–37.7 ka up to 18 ka in the cave. This discovery testifies to the long isolation of middle Pleistocene immigrants including *H. erectus*, and furthermore, stone tools as old as 800 ka have been reported (Morwood et al., 1998). Therefore, the FAD of the mainland mammals recorded from Holocene sediments (Van den Bergh et al., 1996) should be younger, from shortly after the LGM, and modern *H. sapiens* may have immigrated simultaneously when the deep-water barrier isolating flores during most of the middle Pleistocene became shallower.

Indochinese Province

The Indochinese Province fauna of the middle and late Pleistocene is characterized by a peculiar mammalian assemblage comprising of *Stegodon orientalis*, *Ailuropoda melanoleuca baconi* (giant panda), and orangutan. The fossil record of these species dates from the early middle Pleistocene, and they survived until the late Pleistocene or even close to the present day. Some even migrated southward to reach the Sundaic Province at an early time and, in the case of *Stegodon*, this probably occurred in several waves. Orangutan and modern human, in contrast, reached Sundaland only during the end of the last interglacial or the beginning of the last glacial, and giant panda never immigrated there. Biochronologically, the Pleistocene period is subdivided by the FAD and LAD of different species. The early Pleistocene is characterized by a primitive species of panda, *Ailuropoda minuta*, *Gigantopithecus*, and several other species which also survived in the earliest middle Pleistocene. This later period (see Mid-Pleistocene of Southern Asia) is characterized by a short-faced hyena (*Hyaena brevirostris sinensis*) which is replaced after migration or extinction by the spotted hyena (*Crocuta crocuta ultima*) whose origin and FAD is problematic. For Kurten (1956), the spotted hyena had reached Southeast Asia between 480 and 450 ka. It also immigrated to Northern China, where it was found in the upper layer of the Choukoutien *H. erectus* locality. Another

biochronological marker is *H. sapiens* whose appearance seems to correspond to the end of the middle Pleistocene. The late Pleistocene fauna of that province can be therefore characterized by the local or total extinction of some mammal species. Giant panda survived in Southern China whereas orangutans survived in Borneo and Sumatra. A similar history has been elaborated for rodents by Chaimanee (1998). Early Pleistocene and early middle Pleistocene rodent faunas contain several genera that became extinct before the late middle Pleistocene, a period that is characterized by the presence of extant species no longer native to Thailand, such as *Hadromys humei*, *Hapalomys delacouri*, *Pithecheir parvus*, *Nannosciurus melanotis*, and *Iomys horsfieldi*. Finally, the late Pleistocene rodent assemblage is characterized by the progressive local extinction of these 'exotic' rodents and by the immigration of some large-sized flying squirrels originating from South China.

Vietnam

Several early to middle Pleistocene mammal faunas are known, but so far only one rich late Pleistocene locality has been reported, the Lang Trang Cave from North Vietnam with 23 species of large- and medium-sized mammals (Vu et al., 1996; Long, 1996). Its age has been estimated between 80 and 60 ka. Primates are represented by *Macaca*, *Semnopithecus*, and orangutan. *S. orientalis* and *Elephas namadicus* are associated with giant panda and tapir, indicating a forest environment of early late Pleistocene age. No human remains have been collected so far from the cave which represents the LAD for *E. namadicus*. The Ma U'O'i Cave in Northern Vietnam (Bacon et al., 2004) has yielded a small fauna with six species of large mammals including *Elephas* and *Homo* sp. and four species of rodents. The age is uncertain since the fauna has a modern composition with no extinct species. The Keo Leng fauna from Northern Vietnam has an 'estimated' age of 30–20 ka (Kha, 1976; Olsen and Ciochon, 1990). It has yielded a rich mammalian fauna of 27 species including *H. sapiens*. Orangutan, giant panda, *Megatapirus augustus*, and *S. orientalis* are indicative of a late middle Pleistocene assemblage, thereby casting doubts on its estimated age. If confirmed, it would indicate that the previously mentioned taxa, with the exception of modern human, survived until the LGM or even later in the Indochinese Province.

Thailand

No precisely dated late Pleistocene mammal fauna has been reported from Thailand, since most Thai caves and fissure fillings have been dated to the late middle Pleistocene (Chaimanee, 1998). Among them, the fossiliferous layer of Tham Wiman Nakin Cave in Northeastern Thailand (see Mid-Pleistocene of Southern Asia) has been dated as older than 169 ka, using speleothems covering the main fossiliferous layer (Esposito et al., 2002). However, uranium–thorium ages obtained from mammalian teeth gave confusing results due to leaching and diagenesis. *Homo*, orangutan, giant panda, and modern Asian elephant therefore existed in Thailand before the last interglacial. Only one fissure-filling locality, the Crystal Cave near Kanchanaburi in Central West Thailand, has yielded a rodent fauna that may be of late Pleistocene age. It consists of only extant species associated with an unusual large species of flying squirrel, which has been interpreted as a large-sized

species, that shifted its range southward during a cold period, as predicted by Bergmann's rule (Chaimanee and Jaeger, 2000a). Many Pleistocene prehistoric excavations, such as Lang Rongrien rockshelter and Moh Kiew Cave in Peninsular Thailand near Krabi, and Spirit Cave and Tham Lod rock shelter near Mae Hong Son in North Thailand, have yielded mammalian remains which have not been published so far. They consist of only modern species. In addition, recent palynological data obtained from three localities in Thailand confirm the hypothesis of a drier and cooler climate during the LGM (White et al., 2004), with a fagaceous–coniferous forest in northeast Thailand (Penny, 2001).

Cambodia

Only one cave fauna in Cambodia, Phnom Loang (Beden et al., 1972) from Kampot Province shows a composition which could correspond with either a late Pleistocene or a late middle Pleistocene age, no absolute dating being presently available. Here, Javan rhinoceros (*Rhinoceros sondaicus*) is associated with spotted hyena. Nearby caves have shown the presence of orangutan, Asian elephant, and tapir. The presence of a spotted hyena supports a late middle Pleistocene age rather than a late Pleistocene one. In that last case, the *C. crocuta ultima* LAD would be latest middle Pleistocene.

Discussion

Several conclusions can be reached from this review. The most important is the lack of absolute dating and the huge imprecision concerning the age of most of these mammalian faunas. The second is that the complex geographic history of Southeast Asia makes it nearly impossible to construct a regional biochronological timescale. Local faunal evolution is strongly related to the main geographical changes induced by sea-level changes and by neotectonic uplift or subsidence. Only in the more stable Indochinese Province, a reliable biochronological scale potentially could be established, but more accurate dating would be necessary.

Late Pleistocene Local and Global Extinctions

Giant panda is today confined to a small range in Southern China at high altitude, with a temperate climate, where it feeds mostly on temperate bamboos. Tougaard et al. (1996) have suggested that climatic cooling and an associated altitudinal drop of vegetation zones caused the extension of the temperate bamboo range on which the giant pandas fed. This would explain their wide Pleistocene expansion. Simultaneously, the same cause could explain the spread of the spotted hyena, which is clearly an open woodlands and grassland species, and its replacement of the short-faced hyena. Both giant panda and spotted hyena never reached the Sundaic Province, probably because of rainforest barriers. Orangutan was also widespread in the Indochinese Province during the middle Pleistocene and migrated later on to the Sundaic Province, where it still survives until today. No data are available concerning the extinction of orangutan in the Indochinese Province, but Bacon et al. (2001) described an orangutan skeleton

from Hao Bin Province (northern Vietnam) that seems to be of very recent age, suggesting a long survival of that species and supporting an anthropic cause for its extinction. Other species have become extinct in both provinces, such as *Megatapirus* in the Indochinese Province and *Duboisia santeng* in the Sundaic Province, before the late Pleistocene. But again, these events are not yet accurately dated and their causes are unknown. A similar situation occurs for the rodents, with the grassland species *Hadromys humei* becoming extinct before the late Pleistocene (Chaimanee and Jaeger, 2000b).

Changing Biogeography and the Fluctuating Boundary between the Indochinese and Sundaic Faunal Provinces

Both of these provinces are characterized by a small number of taxa during most of the late middle and late Pleistocene. Not only extant forms such as *Macaca assamensis*, *Ursus thibetanus*, giant panda, but also extinct ones such as *Crocota*, *S. orientalis*, *E. namadicus*, and *Megatapirus* can be considered as typical markers of the Indochinese Province. The extant gibbon (*Hyllobates syndactylus*) and the pig (*S. barbatus*) for the living species, and *Stegodon trigonocephalus* and *Duboisia santeng* for the extinct taxa, are also characteristic of the Sundaic Province. But many other species were common to both provinces such as *H. sapiens*, tiger, Malayan bear, wild dog (*Cuon alpinus*), tapir, rhinocerotids (*R. sondaicus* and *Dicerorhinus sumatraensis*), wild pig, buffalo, muntjac, Sumatran goat, and several rats, and squirrels. Progress in taxonomic work may increase this list, especially when more small mammals are taken in account. Their association clearly defines a Southeast Asian mammalian assemblage. But the real question concerns the past location of the 'paleo-Kra Isthmus' and its fluctuations. The fossil record yields little information with the exception of a perceptible southward extension of this boundary during glacial periods. The fauna from Tambun locality next to Kuala Lumpur in Malaysia, the age of which is not well established (Hooijer, 1962), contains *E. namadicus*, indicating a southward expanded Indochinese Province fauna. Some extant grassland rodent species, such as *Hadromys humei*, presently restricted to Northeast India and South China, have been found in a late middle Pleistocene locality close to the Thai–Malay border (Chaimanee and Jaeger, 2000b). On the other hand, some Sundaic rats, presently not living in Thailand, such as *Pithecheir parvus*, extended their range northward. Similar examples occurred among squirrels. However, as very few rodents have been described from the Pleistocene localities of Sundaland, it is difficult to follow the complicated history of the fluctuating boundary between these provinces. Nevertheless, many species of rodents showed a larger latitudinal range than today and many ranges did overlap in a much broader ecotone with high small-mammal biodiversity.

Bergmann's Rule and Size Changes

One common feature of Pleistocene fossil mammal communities concerns their composition and the large size of their constituent species. As in other parts of the world at the same time, large herbivores were more numerous than they are at

present. Despite the lack of updated taxonomic studies, six or more large herbivores, two proboscideans and four bovids, have coexisted in these Southeast Asian faunas. Pleistocene herbivores are usually larger than their living representatives. Many factors are involved in controlling adult size of mammals. The most common is Bergmann's rule, which stipulates that warm-blooded animal size increases according to a decreasing mean temperature gradient. This has been demonstrated for the extant *Macaca fascicularis* by Fooden and Albrecht (1993). Hooijer (1949) uses this principle to explain the progressive size decrease of Sundaland mammals, such as orangutans, Javan rhinoceros, tapirs, tigers, and even rodents (Medway, 1972). However, many other factors could have played an important or additional role including interspecific competition and island isolation.

Conclusions

The late Pleistocene mammalian faunas of Southeast Asia are very incompletely known. Many localities have not yet been dated with modern methods and their inferred ages sometimes correspond only to rough estimations. However, some general conclusions can be drawn from the available data. First, the beginning of this period corresponds to the immigration of modern humans in the area. Some scanty human remains attest to the occurrence of *Homo* before 169 ka in Thailand and in several other places in the Indochinese Province, but the fossils are not sufficiently complete to identify these remains to species level. The faunas were distributed into two distinct provinces but their boundary, located today in the Thai Peninsula at the Kra Isthmus, was situated further south during most of the late Pleistocene. Rodent faunas suggest that the transition area, which is sharp today, was much more latitudinally extended. During these periods, savanna was widely expanded and many species were occupying that niche. Probably, many extant forest herbivores were also living in the savanna during the late Pleistocene as indicated by recent unpublished isotopic studies. Sea-level and climatic changes permitted some exchanges between the two provinces, allowing *H. sapiens*, orangutan, and some other species to reach the biotas of the Sundaic Province. Orangutan's present distribution range in Sumatra and Borneo appears also to be a refuge area and not as a center of origin. In the Sundaic Province, many endemic species, including *H. floresiensis*, differentiated through island evolution as a consequence of isolation. It is expected that in future more evidence of island evolution will be discovered. The history of the late Pleistocene fauna consists of the progressive range reduction and the local to total extinction of several species. Two main periods of climatic changes may be responsible for these events. First, the Late Glacial Maximum, some 20 ka, corresponded to an important cooling and extension of savanna. It was followed, about 10 ka, by the Holocene warming which was associated with a dramatic expansion of the rainforest that led to the present biogeographic situation. The causes of the local extinctions, as those of the spotted hyena, *Stegodon*, giant panda, and orangutan in the Indochinese Province, are not well understood, since the importance of the human factor cannot yet be evaluated (see Late Pleistocene Megafaunal Extinctions). Finally, the modern

distribution of mammals has developed only very recently, after a long period of relative stability during the middle and late Pleistocene.

See also: **Vertebrate Records: Late Pleistocene Megafaunal Extinctions; Mid-Pleistocene of Southern Asia. Vertebrate Studies: Interactions with Hominids.**

References

- An Z (2000) The history and variability of the east Asian palaeomonsoon climate. *Quaternary Science Reviews* 19: 171–188.
- Bacon A-M and Long VT (2001) The first discovery of a complete skeleton of a fossil orang-utan in a cave of the Hoa Binh Province, Vietnam. *Journal of Human Evolution* 41: 227–241.
- Bacon A-M, Demeter F, Schuster M, et al. (2004) The Pleistocene Ma U'O'i cave, northern Vietnam: Palaeontology, sedimentology and palaeoenvironments. *Geobios* 37: 305–317.
- Beden M, Carbonnel JP, and Guerin C (1972) La faune du Phnom Loang (Cambodge) Comparaison avec les faunes Pleistocenes du nord de l'Indochine. *Extrait des archives Geologiques du Viet-Nam* 15: 113–122.
- Chaimanee Y (1998) Plio-Pleistocene rodents of Thailand. *Thai Studies in Biodiversity* 3: 1–303.
- Chaimanee Y and Jaeger J-J (2000a) A new flying squirrel *Belomys thamkaewi* n. sp. (Mammalia: Rodentia) from the Pleistocene of West Thailand and its biogeography. *Mammalia* 64: 307–318.
- Chaimanee Y and Jaeger J-J (2000b) Occurrence of *Hadromys humei* (Rodentia: Muridae) during the Pleistocene in Thailand. *Journal of Mammalogy* 81: 659–665.
- Corbet GB and Hill JE (1992) The Mammals of the Indomalayan Region: A Systematic Review. *Natural History Museum Publications*. Oxford University Press.
- de Vos J (1983) The *Pongo* faunas from Java and Sumatra and their significance for biostratigraphical and paleo-ecological interpretations. *Proceedings Koninklijke Nederlandse Akademie Wetenschappen B* 86: 417–425.
- Drawhorn GM (1994) *The Systematics and Paleodemography of Fossil Orangutans (genus Pongo)*. PhD Thesis, Davis: University of California.
- Esposito M, Reyss J-L, Chaimanee Y, and Jaeger J-J (2002) U-Series dating of fossil teeth and carbonates from Snake Cave, Thailand. *Journal of Archaeological Science* 29: 341–349.
- Fooden J and Albrecht GH (1993) Latitudinal and insular variation of skull size in crab-eating macaques (Primates, Cercopithecidae: *Macaca fascicularis*). *American Journal of Physical Anthropology* 92: 521–538.
- Gathorne-Hardy FJ, Syaekani, Davies RG, Eggleston P, and Jones DT (2002) Quaternary rainforest refugia in south-east Asia: Using termites (Isoptera) as indicators. *Biological Journal of the Linnean Society* 75: 453–466.
- Hooijer DA (1947) On fossil and prehistoric remains of *Tapirus* from Java, Sumatra and China. *Zoologische Mededelingen* XXVII: 253–299.
- Hooijer DA (1949) Mammalian evolution in the Quaternary of Southern and Eastern Asia. *Evolution* 3: 125–128.
- Hooijer DA (1962) Report upon a collection of Pleistocene mammals from tin-bearing deposits in a limestone cave near Ipoh, Kinta valley, Perak. *Federation Museums Journal* 7: 1–5.
- Kershaw AP, Penny D, Kaars S, van der Anshari G, and Thamotherampilai A (2001) Vegetation and climate in lowland Southeast Asia at the Last Glacial Maximum. In: Metcalfe I, Smith JMB, Morwood M, and Davidson I (eds.) *Faunal and Floral Migration and Evolution in SE Asia-Australasia*, pp. 227–236. Lisse: Balkema.
- Kha LT (1976) First remarks on the quaternary fossil fauna of northern Vietnam. *Vietnamese Studies* 46: 107–126.
- Kurten B (1956) The status and affinities of *Hyaena sinensis* Owen and *Hyaena ultima* Matsumoto. *American Museum Novitates* 1–48.
- Long VT (1996) A fossil mammal fauna of Vietnam (Lang Trang caves) compared with fossil and recent mammal faunas of Southeast Asia. *Bulletin of Indo-Pacific Prehistory Association* 14: 101–109.
- Medway L (1972) The Quaternary mammals of Malesia: A review. *Transactions of the Second Aberdeen-Hull Symposium on Malesian Ecology*, pp. 63–98. University of Hull.
- Meijaard E (2003) Mammals of south-east Asian islands and their late Pleistocene environments. *Journal of Biogeography* 30: 1245–1257.
- Morley RJ (2000) *Origin and Evolution of Tropical Rain Forest*. Chichester: Wiley.

- Morwood MJ, O'Sullivan PB, Aziz F, and Raza A (1998) Fission-track ages of stone tools and fossils on the east Indonesian island of Flores. *Nature* 392: 173–176.
- Morwood MJ, Soejono RP, Roberts RG, et al. (2004) Archaeology and age of a new hominin from Flores in eastern Indonesia. *Nature* 431: 1087–1091.
- Olsen JW and Ciochon RL (1990) A review of evidence for postulated middle Pleistocene occupations in Vietnam. *Journal of Human Evolution* 19: 761–788.
- Penny D (2001) A 40,000 year palynological record from north-east Thailand: Implications for biogeography and palaeo-environmental reconstruction. *Palaeogeography, Palaeoclimatology, Palaeoecology* 171: 97–128.
- Sondaar PY (1984) Fauna evolution and mammalian biostratigraphy of Java. *Courier Forschung Institut Senckenberg* 69: 219–235.
- Sondaar PY, Gert D, Bergh VD, Mubroto B, Aziz F, and De Vos J (1994) Middle Pleistocene faunal turnover and colonization of Flores (Indonesia) by *Homo erectus*. *Comptes Rendus de l'Academie des Sciences Serie II (Paris)* 319: 1255–1262.
- Storm P (1995) The evolutionary significance of the Wajak skulls. *Scripta Geologica* 110: 1–247.
- Swisher CC III, Rink WJ, Anton SC, et al. (1996) Latest *Homo erectus* of Java: Potential contemporaneity with *Homo sapiens* in southeast Asia. *Science* 274: 1870–1874.
- Tougaard C (2001) Biogeography and migration routes of large mammal faunas in south-east Asia during the late middle Pleistocene: Focus on the fossil and extant faunas from Thailand. *Paleogeography, Paleoclimatology, Paleoeecology* 168: 337–358.
- Tougaard C, Chaimanee Y, Suteethorn V, Triamwichanon S, and Jaeger J-J (1996) Extension of the geographic distribution of the giant panda (*Ailuropoda*) and search for the reasons for its progressive disappearance in Southeast Asia during the latest middle Pleistocene. *Comptes Rendus de l'Academie des Science Serie II a (Paris)* 323: 973–979.
- Van den Bergh GD (1999) The late Neogene elephantoid-bearing faunas of Indonesia and their palaeozoogeographic implications. *Scripta Geologica* 117: 1–419.
- Van den Bergh GD, de Vos J, and Sondaar PY (2001) The late Quaternary palaeogeography of mammal evolution in the Indonesian Archipelago. *Palaeogeography, Palaeoclimatology, Palaeoecology* 171: 385–408.
- Van den Bergh GD, Mubroto GD, Sondaar PY, and de Vos J (1996) Did *Homo erectus* reach the island of Flores? *Bulletin of Indo-Pacific Prehistory Association* 14: 27–36.
- Van der Kaars WA and Dam MAC (1995) A 135, 000 year record of vegetational and climatic change from the Bandung area, West Java, Indonesia. *Palaeogeography, Palaeoclimatology, Palaeoecology* 117: 55–72.
- Voris HK (2000) Maps of Pleistocene sea levels in South East Asia: Shorelines, river systems, time durations. *Journal of Biogeography* 27: 1153–1167.
- Vu TL, de Vos J, and Ciochon RL (1996) The fossil mammal fauna of the Lang Trang Caves, Vienam, compared with Southeast Asian fossil and recent mammal faunas: The geographical implications. *Bulletin of Indo-Pacific Prehistory Association* 14: 119–128.
- White JC, Penny D, Kealhofe L, and Maloney B (2004) Vegetation changes from the late Pleistocene through the Holocene from three areas of archaeological significance in Thailand. *Quaternary International* 113: 111–132.
- Woodruff DS (2003) Neogene marine transgression, palaeogeography and biogeographic transitions on the Thai–Malay Peninsula. *Journal of Biogeography* 30: 551–567.

Late Pleistocene Megafaunal Extinctions

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Introduction

From the late nineteenth century, the work of paleontologists was beginning to reveal the fact that many of the large animals that had lived during what was then known as ‘The Great Ice Age’ are no longer living today. The great evolutionary biologist, Alfred Russel Wallace, wrote the following memorable comment on this phenomenon in 1876:

We live in a zoologically impoverished world, from which all the hugest, and fiercest, and strangest forms have recently disappeared; and it is, no doubt, a much better world for us now they have gone. Yet it is surely a marvellous fact, and one that has hardly been sufficiently dwelt upon, this sudden dying out of so many large mammalia, not in one place only but over half the land surface of the globe. (1876, p. 150)

Thirty years later, speculating on the causes of this extinction event, Wallace ventured the following:

Looking at the whole subject again, with the much larger body of facts at our command, I am convinced that the rapidity of . . . the extinction of so many large Mammalia is actually due to man’s agency, acting in co-operation with those general causes which at the culmination of each geological era has led to the extinction of the larger, the most specialised, or the most strangely modified forms. (1911, p. 264)

Thus the comments of Wallace set the stage for a modern discussion of this phenomenon. Today we know a great deal more about the timing of these extinctions, the number of species involved, and the geographic patterns of extinction, but much of the mystery remains, and the debate over the cause(s) of the late Pleistocene megafaunal extinction continues to rage on in the current literature.

Definition of Megafauna

The fossil evidence from many continents points to the extinction mainly of large animals at or near the end of the last glaciation. These animals have been termed ‘megafauna.’ The most common definition of megafauna is an animal with an adult body weight in excess of 44 kg (Martin, 1984). These animals were worst affected because they exist at relatively low density in the landscape, mature late, and have few offspring. They are slow to recover numbers after any collapse of population. In many regions of the world, the Pleistocene megafauna was dominated by herbivorous mammals, such as the woolly mammoth and woolly rhinoceros. However, in biologically isolated regions of the world, such as Australia and New Zealand, the Pleistocene megafauna also included other vertebrate groups, such as large flightless birds, and giant turtles and lizards. One of the most curious aspects of late Pleistocene megafaunal extinctions is that they affected some continents far more than others (Table 1). For reasons

which are still being debated, Australia and the Americas were much harder hit than Africa and Asia. In the next section, we discuss the extinction patterns of the various continents.

Extinctions Before the End of the Last Glaciation

The extinction of late Pleistocene megafaunal species was not a singular event, but rather a process that spanned many thousands of years. It was also time-transgressive on the various continents, so it was not tied to a single global climatic change, and even within individual species, with key taxa being lost from certain areas before others. One of the difficulties in developing an understanding of megafaunal extinction is that it is very difficult to pin down an exact time when a given species ceased to exist, just by examining the fossil record. When the fossil bones of a given species fail to appear above a certain stratigraphic horizon in a region, we assume that this species had died out, at least in that region. However, as the old proverb says, “Absence of evidence is not evidence of absence.” The discovery of any younger fossils will, of course, mean that the extinction date has to be revised. One of the most startling examples of this concerns the woolly mammoth, probably the best known Pleistocene megafaunal mammal, based on its appearance on the logos of Quaternary research organizations. Until quite recently, paleontologists believed that the woolly mammoth became extinct at the end of the last glaciation, perhaps 12 800 years ago (Elias, 1999). Then a Russian paleontologist discovered woolly mammoth bones from Wrangel Island, off the northeast coast of Kamchatka. These bones were radiocarbon dated, and found to be only 4079 years old (Long et al., 1994). Thus the date of extinction for this species had to be revised by 8700 years, placing its last appearance well within the Holocene. Since then, other species have similarly been found to survive into the current interglacial (Stuart and Lister, 2007).

The Wrangel Island woolly mammoth story is a cautionary tale, and it also highlights a different aspect of the demise of the megafauna: extinction versus extirpation. Extirpation is the dying out of a species in a given region. For instance, the fossil evidence indicates that woolly mammoths were extirpated from mainland North America by 12 800 years ago. Extinction refers to the worldwide demise of a species. There is a very important difference between these two phenomena, because, during the Pleistocene, extreme environmental changes caused the regional extirpation of countless numbers of species in the middle and high latitudes, time and again. However, as long as populations of a given species were able to survive in other regions (perhaps only a single population in one region), then that species had a chance to survive until environmental conditions in its former range became more favorable again. The fossil record of the Pleistocene is replete with examples of the regional waxing and waning of plant and animal

Table 1 Megafaunal extinctions (genera) during the last 100 ka

Continent	Extinct	Living	Total	Extinct (%)	Landmass (km ²)
Africa	7	42	49	14.3	30.2 × 10 ⁶
Europe	15	9	24	60.0	10.4 × 10 ⁶
North America	33	12	45	73.3	23.7 × 10 ⁶
South America	46	12	58	79.6	17.8 × 10 ⁶
Australia	19	3	22	86.4	7.7 × 10 ⁶

Source: Wroe et al. (2004).

populations, in the face of climate change and glaciation. Extinction, however, is permanent. When a species' unique genotype is lost to the world, it cannot come back. In the next section, we examine the regional history of the timing of megafaunal extinctions (keeping in mind the caveats discussed above about evidence of absence).

Australia

The continent of Australia has been geographically isolated from much of the rest of the world for many millions of years. The only groups of mammals to successfully colonize Australia were the primitive group of marsupials. In marsupial mammals, the female typically has a pouch (marsupium) in which it rears its young through early infancy, thereby differing from placental mammals, which give birth to fully formed young. Placental mammals have dominated the mammalian faunas of the other continents since the Late Tertiary, but the marsupials were the only mammal group in Australia. In the absence of competition from placental mammals, the Australian marsupials radiated to form a rich and diverse Pleistocene megafauna, including giant herbivores, such as *Procoptodon*, a giant kangaroo (Figure 1(c)), and *Diprotodon*, a lumbering rhinoceros-sized marsupial (Figure 1(a)). The megafaunal group that suffered the greatest number of extinctions in the late Pleistocene was the short-faced kangaroos in the subfamily Sthenurinae. Approximately 14 species of sthenurines died out in the late Pleistocene; only one species survives today. The hind feet of these kangaroos had a single toe, with a broad, flattened claw that resembled (and may have acted as) a horse's hoof, reflecting an adaptation to rapid movement through progressively open, arid grasslands.

Giant wombats and wallabies also roamed the grasslands of Pleistocene Australia. One of the most unusual of these was the giant wombat, *Phascolonus gigas* (Figure 1(b)), that weighed up to 200 kg, and was built for burrowing. Flannery (1994) speculated that it was the giant dirt piles adjacent to these burrows that alerted human hunters to the presence of these animals, ultimately leading to their demise.

In the place of placental mammal predatory groups such as wolves, large cats, and bears, the marsupial predators of Australia included *Thylacoleo*, the equivalent of a Pleistocene lion (Figure 1(d)), and *Sarcophilus*, an ancestral Tasmanian devil. Based on its skull and teeth, *Thylacoleo* appears to have been a ferocious predator with arboreal adaptations. It had an enormous set of stabbing front teeth, slicing carnassials, and

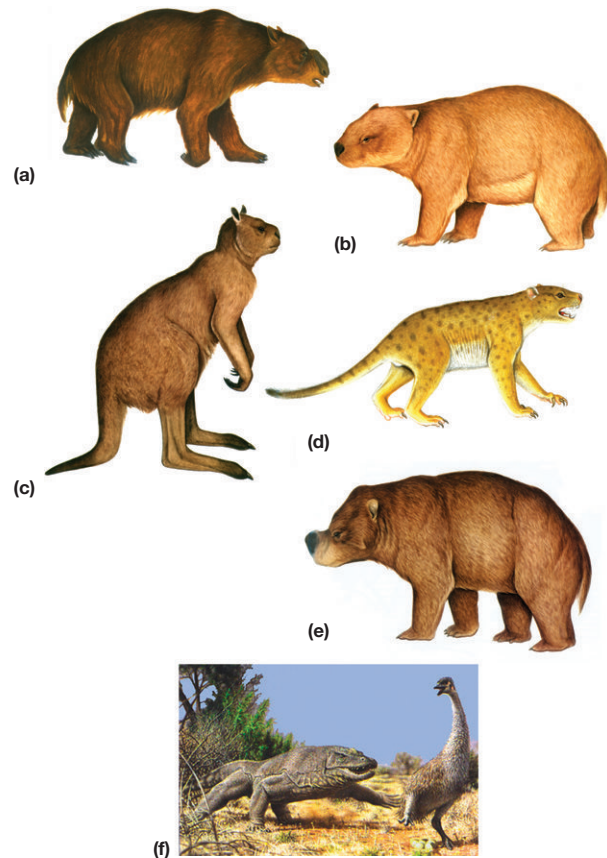


Figure 1 Australian megafaunal species that became extinct in the late Pleistocene. (a) *Diprotodon optatum*, from Long et al. (2002, p. 87). (b) *Phascolonus gigas*, from Long et al. (2002, p. 113); (c) *Procoptodon goliah*, from Long et al. (2002, p. 158); (d) *Thylacoleo carnifex*, from Long et al. (2002, p. 105); (e) *Zygomaturus trilobus*, from Long et al. (2002, p. 99); (f) *Megalania* chasing *Genyornis*, from University of California Berkeley Website: <http://www.geology.ucdavis.edu/cowen/HistoryofLife/CH18images.html>.

powerful claws. Paleontologists have equated it with the modern leopard, killing its prey, and then dragging the carcass up a tree to feed (Long et al., 2002). *Sarcophilus*, the modern Tasmanian devil, had a Pleistocene relative that was about 16% larger than the modern species. It was not technically part of the megafauna, weighing only 10 kg, but as with the modern species, the Pleistocene variety probably 'fought above its weight.'

Other groups of vertebrates flourished in the absence of placental mammals in Australia. These included a large flightless bird, *Genyornis* (Figure 1(f)), crocodiles, the giant madtsoiid snake *Wonambi naracoortensis*, giant horned turtles, and the giant predatory 'ripper lizard,' *Megalania prisca* (Figure 1(f)), a relative of the modern monitor lizard or goanna and the largest terrestrial lizard known. The Australian Pleistocene bestiary could well fit Wallace's description of 'large, fierce, and strange.' Then, beginning about 50 000 years ago, they started dying off (Table 2).

There are a number of important fossil sites in Australia that have yielded abundant, well-preserved megafaunal remains. Perhaps the best known of these is the caves contained in the Naracoorte Caves National Park, South Australia. These are

Table 2 Timing of extinction of the Australian Pleistocene megafauna

<i>Taxon</i>	<i>Common name</i>	<i>Last 100 ka</i>	<i>100–50 ka</i>	<i>50–16 ka</i>	<i>16–11.5 ka</i>	<i>11.5–0 ka</i>	<i>References</i>
Reptilia							
Varanidae							
<i>Megalania</i>	Giant ripper lizard			X			Roberts et al. (2001)
Meiolanidae							
<i>Meiolania</i>	Giant horned turtle			X			Roberts et al. (2001)
<i>Ninjemys</i>	Giant horned turtle	X					Roberts et al. (2001)
Crocodylidae							
<i>Palimnarchus</i>	Pleistocene crocodile						Roberts et al. (2001)
<i>Quinkana</i>	Saw-toothed crocodile						Flannery and Roberts (1999)
Boiidae?							
<i>Wonambi</i>	Giant madtsoiid snake		X				Roberts et al. (2001)
Aves							
<i>Genyornis</i>	Giant ratite bird			X			Roberts et al. (2001)
Mammalia							
Marsupialia							
Diprotodontidae							
<i>Diprotodon</i>	Diprotodon			X			Roberts et al. (2001)
<i>Euryzygoma</i>	Euryzygoma	X					Flannery and Roberts (1999)
<i>Euowenia</i>	Euowenia	X					Flannery and Roberts (1999)
<i>Nototherium</i>	Nototherium	X					Flannery and Roberts (1999)
<i>Zygomaturus</i>	Zygomaturus		X				Roberts et al. (2001)
Palorchestidae							
<i>Palorchestes</i>	Marsupial tapir			X			Roberts et al. (2001)
Vombatidae							
<i>Phascolomys</i>	Hairy-nosed wombat	X					Flannery and Roberts (1999)
<i>Phascolonus</i>	Giant wombat			X			Roberts et al. (2001)
<i>Ramsayia</i>	Pleistocene wombat	X					Flannery and Roberts (1999)
Thylacoleonidae							
<i>Thylacoleo</i>	Marsupial 'lion'			X			Roberts et al. (2001)
Macropodidae							
<i>Protemnodon</i>	Giant wallaby			X			Flannery and Roberts (1999)
<i>Procoptodon</i>	Giant kangaroo	X					Flannery and Roberts (1999)
<i>Simosthenurus</i>	Short-faced kangaroo			X			Roberts et al. (2001)
<i>Sthenurus</i>	Pleistocene kangaroo		X				Roberts et al. (2001)

limestone solution caves that formed in the early Pleistocene and which acted as giant pit-fall traps in the last 500 000 years. To date, 118 species of vertebrate animals have been recorded representing four of the major vertebrate groups: the amphibians, reptiles, birds, and mammals. Many of these qualify as megafauna, based on their estimated size. Mammoth Cave, another limestone cave, lies near the southwest tip of Australia in New South Wales. More than 10 000 late Pleistocene bones have been found in the cave, some greater than 35 000 years old. The fossils include the remains of a Tasmanian devil, the 'marsupial lion' (*Thylacoleo*), the giant echidna (*Zaglossus hacketti*), and a Pleistocene short-faced kangaroo (*Simosthenurus*). Pleistocene fossil localities are dotted across much of southern and eastern Australia (Figure 2). While the cave sites, such as the ones described above, have yielded spectacular bone piles, there are hundreds of open-air sites as well. For more complete inventories of sites, see Horton (1989) and Roberts et al. (2001).

The Australian megafauna contained many large animals (Figure 3), but few that were true giants, compared with the many megafaunal species in Eurasia and the Americas that weighed several tons. Flannery (1994) attributes the paucity of extremely large megafaunal species in Australia to the low level of nutrients available in Australian soils. Also, because of

Australia's geographic isolation since before the evolution of placental mammals on other continents, vertebrate groups other than mammals, such as flightless birds and reptiles, played a larger role in Australian Pleistocene faunas than they did elsewhere. Notable among these were giant snakes, lizards, and crocodiles (although for an opposing view, see Wroe, 2002).

The story of Australian megafaunal extinction is confounded by a relatively poor chronology, and new information is being obtained with great frequency, but the main pulse of extinction came here much earlier than elsewhere in the world. Roberts et al. (2001) dated the youngest sedimentary layers containing megafaunal species, using a variety of methods, including optically stimulated luminescence (OSL) and U/Th. They placed the extinctions all within a window from 51 200 to 39 800 (95% confidence interval), substantially earlier than the main waves of extinction on other continents. Barnowsky et al. (2004) cite the following statistics: of 21 extinct genera of Australian megafauna, 12 persisted to at least 80 000 years ago, and at least 6 persisted to between 51 000 and 40 000 years ago. Wroe et al. (2004) produced a slightly different reckoning of these events, with 19 genera becoming extinct in the Pleistocene, rather than 21. The differences are mostly due to different taxonomic classification schemes, but the bottom

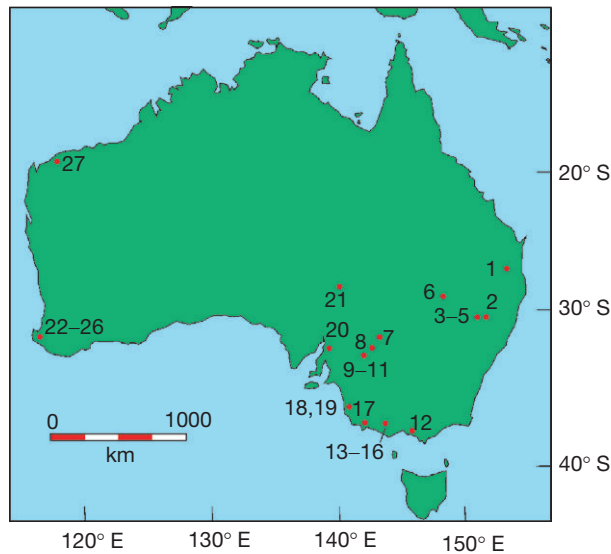


Figure 2 Map of Australia, showing important late Pleistocene fossil localities from which megafaunal remains have been found, and radiometric dates have been analyzed by Roberts et al. (2001). 1, Ned's Gully; 2, Mooki River; 3, Cox's Creek (Bando); 4, Cox's Creek (Kenloi); 5, Tamar Springs; 6, Cuddie Springs; 7, Lake Menidee (Sunset Strip); 8, Willow Point; 9, Lake Victoria (Site 50); 10, Lake Victoria (Site 51); 11, Lake Victoria (Site 73); 12, Montford's Beach; 13, Lake Weering; 14, Lake Corangamite; 15, Lake Weeranganuk; 16, Lake Colongulac; 17, Warrnambool; 18, Victoria Fossil Cave (Grant Hall); 19, Victoria Fossil Cave (Fossil Chamber); 20, Wood Point; 21, Lake Callabonna; 22, Devil's Lair; 23, Kudjal Yolgah Cave; 24, Mammoth Cave; 25, Moondyne Cave; 26, Tight Entrance Cave; 27, Du Boulay Cave.

line remains the same: roughly 86% of the Australian megafauna died out before 16 000 years ago, leaving only three genera of large mammals in the Holocene. In one of the best constrained chronologies of megafaunal remains from Tight Entrance Cave in southwestern Australia, Ayliffe et al. (2008) dated the last appearance of the short-faced kangaroo, *Simosthenurus*, between 50 000 and 48 000 years BP. However, not all fossil faunas are so well dated, making the timing of Australian megafaunal extinctions, and possible associations with human hunting, a matter of some controversy. The interpretation of the fossil record at the Cuddie Springs site (Figure 2, No. 6) is a good example of this.

Cuddie Springs is an ancient lake bed in the semiarid zone of northcentral New South Wales. Pleistocene megafaunal remains have been found there in a clay pan in the center of the lake bed. The fossil bones were discovered in the 1870s when a well was sunk in the center of the ancient lake bed. Paleontological research began there in earnest in the 1930s, and five megafaunal species have been identified from the site. The work there eventually led to the discovery of stone tools and evidence of butchering of animals (cut marks on fossil bones). The most significant archaeological discovery was made in one of the lower levels at the Cuddie Springs site. A small flaked stone tool (possibly an arrowhead or scraper) was found lodged between a *Diprotodon* mandible and a *Genyornis* femur. The tool had traces of dried blood, and the wear patterns were consistent with its use in butchering. But when did these events take place?

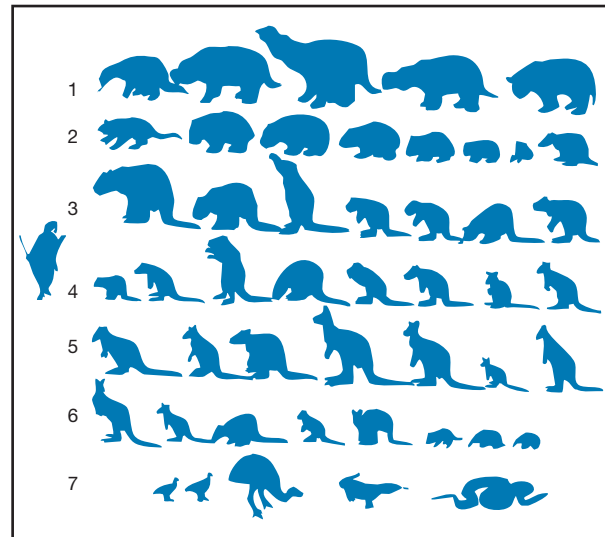


Figure 3 Silhouettes of most of the extinct late Pleistocene Australian vertebrate species drawn to scale (human hunter provides scale). Row 1, right to left: Tree-feller (*Palorchestes azael*), marsupial 'rhino' (*Zygomaturus trilobus*), large and small diprotodons (*Diprotodon optatum*, *D. minor*), Euowenia (*Euowenia grata*). Row 2, Marsupial lion (*Thylacoleo carnifex*), five species of extinct wombats (*Ramsayia curvirostris*, *Phascolonus gigas*, *Phascolomys major*, *P. medius*, *Vombatus hacketti*, *Phascolarctos stirtoni*), giant rat-kangaroo (*Propleopus oscillans*). Row 3, Seven species of giant short-faced kangaroos (*Procoptodon goliath*, *P. rapha*, *P. pusio*, *Sthenurus maddocki*, *S. brownei*, *S. occidentalis*, *S. orientalis*). Row 4, Eight species of extinct kangaroos (*S. gilli*, *S. atlas*, *S. tindalei*, *S. pales*, *S. oreas*, *S. andersoni*, *Troposodon minor*, *Wallabia indra*). Row 5, Seven species of extinct or dwarfed kangaroos (*Protemnodon roechus*, *P. anak*, *P. brehus*, *Macropus ferragus*, *M. (Osphranter) birdselli*, *M. siva*, *M. titan*). Row 6, Extinct or dwarfed marsupials and monotremes (*M. tama*, *M. thor*, *M. piltonensis*, *M. goodi*, *M. stirtoni*, *Sarcophilus laniarius*, *Zaglossus hacketti*, *Z. ramsayi*). Row 7, Extinct birds and reptiles (*Progura naracoortensis*, *P. gallinacea*, *Genyornis newtoni*, *Megalania prisca*, *Wonambi naracoortensis*).

Radiocarbon dating of butchered remains indicates that this activity took place between about 40 000 and 31 000 years ago. However, there are numerous problems to be considered when radiocarbon dating of collagen extracted from bones that have been repeatedly wetted and dried in a hot climate, so these ages are suspect. Roberts et al. (2001) subsequently obtained OSL dates between 30 and 27 ka on sand grains from the megafaunal fossil beds, but they dismiss the apparent age of the site on the grounds that sediment mixing has occurred there and that the remains must have been reworked. Field and Fullager (2001), who have worked at the site, refute this interpretation. In their view, the megafauna disappeared from the Cuddie Springs record around 28 000 years ago, well after the arrival of humans in Australia. The view of the archaeologists who have studied Cuddie Springs is that after the lake dried out, people lived on and around the clay pan, broadening their resource base to include a range of plant foods. Megafauna appear to be just one of a range of food resources exploited during this time. The archaeological evidence suggests no specialized strategies for hunting megafauna. Thus, depending on which interpretation of the Australian fossil record one

chooses to believe, the megafauna either died out very shortly after the arrival of humans, about 50 000 years ago, or it persisted for tens of thousands of years afterwards. However, more recent dating work on this site done by Grün et al. (2010) casts doubt on the dating of the site. Grün and his colleagues analyzed more than 60 fossil bones and teeth from Cuddie Springs by laser ablation inductively coupled plasma mass spectrometry. They also analyzed 29 fossil teeth using electron spin resonance. They concluded that either the sediment layers that yielded the bones at Cuddie Springs are significantly older than the previous OSL estimate of 30–27 ka, or that a large fraction of the faunal remains and charcoal from this layer are reworked from older deposits that came from the sides of the spring.

The Americas

The ancient megafauna of the Americas were very different from that of Australia, but this is not to say that the Americas were not home to some very strange beasts during the Pleistocene. Many forms of Pleistocene mammals remain familiar to us today, even if they are no longer native to the New World. The predators include lions and cheetahs, bears, and wolves, albeit larger sized versions than exist today. Many of the Pleistocene herbivores also have modern counterparts, such as camels, horses, bison and muskoxen, deer, and elephants. Again, their Pleistocene counterparts were generally larger than the Holocene survivors. These include a host of different giant armadillo-like creatures known as Glyptodonts (Figure 4(g)). These were particularly important in the Pleistocene fauna of South America. South America was geographically isolated from the rest of the world during the heyday of mammalian evolution in the Tertiary (65–3 million years ago). With the formation of the Isthmus of Panama, about 3 million years ago, a land connection finally allowed the interchange of mammalian faunas from North and South America. Some large herbivores, such as the mastodon (Figure 4(i)) and some related proboscideans, deer, and bears, were able to colonize South America, but the large variety of ground sloths persisted through this interval, and some species of ground sloths and armadillos spread into North America. Some of the ground sloths were true giants. For instance, *Megatherium* was about 6 m long and weighed 4000 kg.

Of the North American megafauna that existed at the beginning of the last glaciation, only three genera are known to have become extinct before 16 000 years ago. The timing of the extinction of 12 genera remains relatively uncertain (Barnowsky et al., 2004), but the remaining extinctions (20 genera) apparently took place between 16 000 and 11 500 years ago (Table 3).

The South American extinction data are much more vague, because less work has been done here. Of the 48 genera of megafauna that have become extinct in the last 100 000 years, we have more-or-less certain evidence that 17 died out between 16 000 and 11 500 years ago (Table 4). Twenty-seven genera died out sometime in the last 100 000 years, but the timing of their extinction remains uncertain (Barnowsky et al., 2004). Four megafaunal genera are known to have persisted into the early Holocene, and then died out. All told, the North American megafauna suffered a 73% reduction in genera by the end of the Pleistocene, and the South America megafauna suffered almost an 80% reduction in genera. Recent careful assessment of

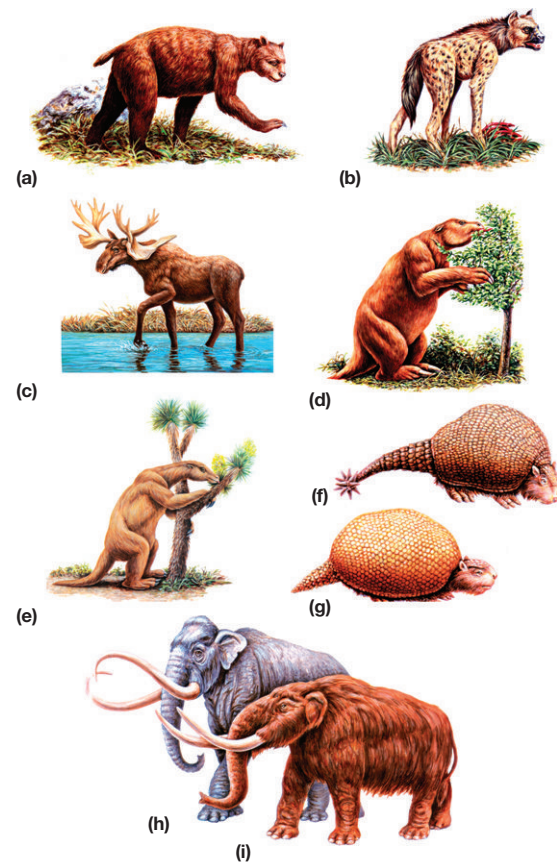


Figure 4 American Megafauna that became extinct in the late Pleistocene. (a) *Arctodus simus*, from Lange (2002, p. 101); (b) *Chasmaporthetes ossifragus*, from Lange (2002, p. 113); (c) *Cervalces scotti*, from Lange (2002, p. 140); (d) *Eremotherium rusconii*, from Lange (2002, p. 84); (e) *Nothrotheriops shastensis*, from Lange (2002, p. 91); (f) *Doedicurus*, from Lange (2002, p. 72); (g) *Glyptotherium arizonensis*, from Lange (2002, p. 72); (h) *Mammuthus columbi*, from Lange (2002, p. 167); (i) *Mammuth americanum*, from Lange (2002, p. 167).

radiocarbon dates on megafaunal remains by Barnosky and Lindsey (2010) indicates that most species persisted until after humans arrived in South America. Most species persisted a thousand years or more after human arrival, and some apparently survived more than 6000 years after this. The geographic pattern of last appearance is variable, but shows the strongest correlation with the arrival of humans in southern and eastern South America, between about 13 500 and 11 200 years ago. This was also a time of rapid environmental change, and the authors conclude that the combination of hunting pressure and environmental change drove most of the South American megafauna to extinction.

Regions Suffering Few Late Pleistocene Megafaunal Extinctions

In contrast, some regions of the world saw few megafaunal extinctions during the late Pleistocene. Some of these regions, such as Africa and southern Asia, preserve the only living examples of the types of animals that died out elsewhere,

Table 3 Timing of extinction of the North American Pleistocene megafauna

<i>Taxon</i>	<i>Common name</i>	<i>Last 100 ka</i>	<i>100–50 ka</i>	<i>50–16 ka</i>	<i>16–11.5 ka</i>	<i>11.5–0 ka</i>	<i>References</i>
Mammalia							
Xenarthra							
Glyptodontidae							
<i>Glyptotherium</i>		X					Martin and Steadman (1999)
Megalonychidae							
<i>Megalonyx</i>	Ground sloth				X		Stuart (1991)
Megatheriidae							
<i>Eremotherium</i>	Giant ground sloth	X					Martin and Steadman (1999)
<i>Nothrotheriops</i>	Shasta ground sloth				X		Stuart (1991)
Mylodontidae							
<i>Glossotherium</i>	Giant ground sloth			X			Stuart (1991)
Pampatheriidae							
<i>Pampatherium</i>	Pampas ground sloth				X		Johnson (1987)
Rodentia							
Castoridae							
<i>Castoroides</i>	Giant beaver				X		Steadman et al. (1997)
Hydrochoeridae							
<i>Hydrochaeris</i>	Capybara	X					Martin and Steadman (1999)
<i>Nechoerus</i>		X					Martin and Steadman (1999)
Carnivora							
Ursidae							
<i>Arctodus</i>	Short-faced bear				X		Stuart (1991)
<i>Tremarctos</i>	Spectacled bear				X		Emslie and Morgan (1995)
Felidae							
<i>Homotherium</i>	Saber-toothed cat			X			FAUNMAP (2003)
<i>Miracinonyx</i>	American cheetah				X		Emslie (1986)
<i>Panthera</i>	Lion				X		Stuart (1991)
<i>Smilodon</i>	Saber-toothed cat				X	?	Stuart (1991)
Proboscidea							
Elephantidae							
<i>Mammuthus</i>	Mammoth				X	?	Stuart (1991)
Gomphotheriidae							
<i>Cuvieronius</i>	Mastodon-like proboscidean			X			Koch et al. (1998)
Mammutidae							
<i>Mammut</i>	Mastodon				X	?	Stuart (1991)
Perissodactyla							
Equidae							
<i>Equus</i>	Horse				X		Stuart (1991)
Tapiridae							
<i>Tapirus</i>	Tapir				X		Beck (1996)
Artiodactyla							
Tayassuidae							
<i>Mylohyus</i>	Peccary				X		Stuart (1991)
<i>Platygonus</i>	Flat-headed peccary				X		Stuart (1991)
Camelidae							
<i>Camelops</i>	Camel				X		Stuart (1991)
<i>Hemiauchenia</i>	Prehistoric llama				X		Stuart (1991)
<i>Paleolama</i>	Pleistocene llama				X		Stuart (1991)
Cervidae							
<i>Bretzia</i>	Pleistocene deer	X					Martin and Steadman (1999)
<i>Cervalces</i>	Stag-moose				X		Stuart (1991)
<i>Navahoceros</i>	Pleistocene mountain deer	X					Martin and Steadman (1999)
<i>Torontoceros</i>	Pleistocene caribou	X					Martin and Steadman (1999)
Antilocapridae							
<i>Stockoceros</i>	Four-homed pronghorn	X					Martin and Steadman (1999)
<i>Tetrameryx</i>	Pronghorn	X					Martin and Steadman (1999)
Bovidae							
<i>Bootherium</i>	Helmeted musk ox	X					Martin and Steadman (1999)
<i>Euceratherium</i>					X		Stuart (1991)
<i>Saiga</i>	Saiga antelope	X					Martin and Steadman (1999)

Table 4 Timing of extinction of the South American Pleistocene megafauna

<i>Taxon</i>	<i>Common name</i>	<i>Last 100 ka</i>	<i>100–50 ka</i>	<i>50–16 ka</i>	<i>16–11.5 ka</i>	<i>11.5–0 ka</i>	<i>References</i>
Mammalia							
Xenarthra							
Dasypodidae							
<i>Eutatus</i>	Pleistocene armadillos				X		Vizcaino et al. (2003)
<i>Propaopus</i>					X		Faure et al. (1999)
Glyptodontidae							
<i>Chlamydotherium</i>	Pleistocene glyptodonts				X		Ficcarelli et al. (2003)
<i>Doedicurus</i>					X		Borrero et al. (1998)
<i>Glyptodon</i>					X		Faure et al. (1999)
<i>Heteroglyptodon</i>		X					McKenna and Bell (1997)
<i>Hoplophorus</i>					X		Faure et al. (1999)
<i>Lomaphorus</i>		X					Martin and Steadman (1999)
<i>Neosclerocalyptus</i>		X					Martin and Steadman (1999)
<i>Neothoracophorus</i>		X					Martin and Steadman (1999)
<i>Parapanochthus</i>		X					Martin and Steadman (1999)
<i>Panochthus</i>		X					Martin and Steadman (1999)
<i>Plaxhaplous</i>		X					Martin and Steadman (1999)
<i>Sclerocalyptus</i>		X					Martin and Steadman (1999)
Megalonychidae							
<i>Valgipes</i>	Two-toed sloths	X					Martin and Steadman (1999)
Megatheriidae							
<i>Eremotherium</i>	Giant ground sloth				X		Ficcarelli et al. (2003)
<i>Megatherium</i>	Giant ground sloth				X		Borrero et al. (1998)
<i>Nothropus</i>	Aquatic sloth	X					Martin and Steadman (1999)
<i>Nothrotherium</i>	Ground sloth	X					Martin and Steadman (1999)
<i>Ocnopus</i>		X					Martin and Steadman (1999)
<i>Perezfontanatherium</i>	Ground sloth	X					McKenna and Bell (1997)
Mylodontidae							
<i>Glossotherium</i>	Giant ground sloth				X		Coltorti et al. (1998)
<i>Lestodon</i>	Giant ground sloth	X					Martin and Steadman (1999)
<i>Mylodon</i>	Giant ground sloth				X		Borrero (2003)
Pampatheriidae							
<i>Pampatherium</i>	Pampas ground sloth	X					Martin and Steadman (1999)
Scelidotheriidae							
<i>Scelidotherium</i>	Sloth				X		Ficcarelli et al. (2003)
Litopterna							
Macraucheniidae							
<i>Macrauchenia</i>		X					Martin and Steadman (1999)
<i>Windhausenia</i>		X					Martin and Steadman (1999)
Notoungulata							
Toxodontidae							
<i>Mixotoxodon</i>		X					Martin and Steadman (1999)
<i>Toxodon</i>					X		Martinez (2001)
Rodentia							
Hydrochoeridae							
<i>Nechoerus</i>	Capybara					X	Ficcarelli et al. (2003)
Octodontidae							
<i>Dicolpomys</i>		X					Martin and Steadman (1999)
Carnivora							
Canidae							
<i>Theriodictis</i>	Wolf-like canid	X					Martin and Steadman (1999)
Felidae							
<i>Smilodon</i>	Saber-toothed cat						Ficcarelli et al. (2003)
Ursidae							
<i>Arctodus</i>	Short-faced bear	X					Martin and Steadman (1999)
Proboscidea							
Gomphotheriidae							
<i>Cuvieronius</i>	Mastodon-like proboscidean				X		Dillehay and Collins (1991)
<i>Haplomastodon</i>	Mastodon-like proboscidean					X	Ficcarelli et al. (2003)
<i>Notiomastodon</i>	Mastodon-like proboscidean	X					Martin and Steadman (1999)

(Continued)

Table 4 (Continued)

<i>Taxon</i>	<i>Common name</i>	<i>Last 100 ka</i>	<i>100–50 ka</i>	<i>50–16 ka</i>	<i>16–11.5 ka</i>	<i>11.5–0 ka</i>	<i>References</i>
<i>Stegomastodon</i>	Mastodon-like proboscidean				X		Núñez et al. (1994)
Perissodactyla							
Equidae							
<i>Equus</i>	Horse					X	Ficcarelli et al. (2003)
<i>Hippidion</i>	Pleistocene horse				X		Alberdi et al. (2001)
<i>Onhippidion</i>	Pleistocene horse						Martin and Steadman (1999)
Artiodactyla							
Camelidae							
<i>Eulamaops</i>		X					Martin and Steadman (1999)
<i>Hemiauchenia</i>	Prehistoric llama				X		Martinez (2001)
<i>Palaeolama</i>	Pleistocene llama					X	Faure et al. (1999)
Cervidae							
<i>Agalmaceros</i>	Pleistocene deer	X					Martin and Steadman (1999)
<i>Antifer</i>	March deer				X		Núñez et al. (1994)
<i>Charitoceros</i>		X					Martin and Steadman (1999)
<i>Morenelaphus</i>	Pampas deer	X					Martin and Steadman (1999)
Tayassuidae							
<i>Platygonus</i>	Flat-headed peccary	X					Martin and Steadman (1999)

Table 5 Timing of extinction of the African Pleistocene megafauna

<i>Taxon</i>	<i>Common name</i>	<i>Last 100 ka</i>	<i>100–50 ka</i>	<i>50–16 ka</i>	<i>16–11.5 ka</i>	<i>11.5–0 ka</i>	<i>References</i>
Mammalia							
Proboscidea							
Elephantidae							
<i>Elephas</i>	Pleistocene elephant	X					Martin (1984)
Perissodactyla							
Equidae							
<i>Hipparion</i>	Hipparion	X					Martin (1984)
Artiodactyla							
Camelidae							
<i>Camelus</i>	Camel	X					Martin (1984)
Cervidae							
<i>Megaceroides</i>	Giant deer						Martin (1984)
Bovidae							
<i>Megalotragus</i>	Hartebeest					X	Lee-Thorp and Beaumont (1995)
<i>Pelorovis</i>	Pleistocene buffalo					X	Klein (1994)
<i>Parmularius</i>	Pleistocene bison	X					Martin (1984)
<i>Bos</i>	Pleistocene bison					X	McKenna and Bell (1997)

such as lions, hyenas, elephants, native horses, and rhinoceroses. Their survival in these parts of the globe begs the question of the cause of their extinction elsewhere.

Africa

Of all the continents, Africa was the least affected by the late Pleistocene megafaunal extinction. The modern African fauna includes 42 genera of large mammals; only seven genera died out during the last 100 000 years, according to Wroe et al. (2004). Barnowsky et al. (2004) listed four genera that became extinct in Africa sometime during the last 100 000 years (Table 5). These include a genus of Pleistocene elephants (*Elephas*) of which the only surviving species is the Asian

elephant (*Elephas maximus*), a genus of Pleistocene three-toed horses, a Pleistocene camel, and a Pleistocene bison. Since about 11 500 years ago, the hartebeest, a genus of African buffalo (*Parmularius*), and the genus of modern cattle (*Bos*) became extinct in Africa. Thus the African megafauna suffered the loss of only about 14% of their genera in the last 100 000 years – the smallest faunal decline of any continent.

Eurasia

Eurasia, like Africa, suffered relatively few megafaunal extinctions in the late Pleistocene (Table 6). Of the ten genera that died out, one became extinct sometime between 100 000 and 50 000 years ago (the straight-tusked elephant, *Palaeoloxodon*

Table 6 Timing of extinction of the Eurasian Pleistocene megafauna

Taxon	Common name	Last 100 ka	100–50 ka	50–16 ka	16–11.5 ka	11.5–0 ka	References
Mammalia							
Carnivora							
Hyaenidae							
<i>Crocuta</i>	Hyena				X		Stuart (1991)
Proboscidea							
<i>Mammuthus</i>	Mammoth				X		Stuart (1991)
<i>Palaeoloxodon</i>	Straight-tusked elephant		X				Stuart (1991)
Perissodactyla							
Rhinocerotidae							
<i>Stephanorhinus</i>	Rhinoceros			X			Stuart (1991)
<i>Coelodonta</i>	Woolly rhinoceros				X		Stuart (1991)
Artiodactyla							
Hippopotamidae							
<i>Hippopotamus</i>	Hippopotamus		X				Stuart (1991)
Camelidae							
<i>Camelus</i>	Camel				?		Stuart (1991)
Cervidae							
<i>Megaloceros</i>	Giant deer					X	Stuart (1991)
Bovidae							
<i>Spirocerus</i>	Antelope			X			Stuart (1991)
<i>Ovibos</i>	Musk ox					X	Stuart (1991)

[*Elephas antiquus*], one was extirpated from Europe (the hippopotamus), and three more became extinct between 50 000 and 16 000 years ago (the interglacial rhinoceros, *Stephanorhinus*; a cave bear, *Ursus spelaeus*; and a heavy-bodied Asian antelope, *Spirocerus*) (Barnowsky et al., 2004). The spotted hyena apparently died out in Asia by about 32 000 years ago, roughly corresponding to the onset of the last glaciation (Stuart and Lister, 2007). Another predator, the cave lion (*Panthera spelaea*), became extinct in Eurasia between 14 500 and 14 000 years ago, as climates warmed and shrubs and trees expanded their ranges at the expense of the open steppe-tundra habitat preferred by this species (Stuart and Lister, 2011). Woolly rhinoceros (Figure 5(b)) was extirpated from Western Europe and probably the rest of Eurasia by 16 000–14 000 years ago (Stuart and Lister, 2007). The woolly mammoth (Figure 5(a)) died out in Eurasia by the early Holocene (~11 000 years ago), except for the Wrangel Island mammoth population discussed above. Finally, giant deer (*Megaloceros*; Figure 5(c)) became extinct in much of Eurasia about 11 500 years ago, but a recent dating project has uncovered the fact that giant deer populations persisted in Western Siberia and the Ural Mountains until about 7800 years ago (Stuart and Lister, 2007). Such revisions for the age of extinction will undoubtedly continue to push extinction dates into younger time horizons in the coming years.

Late Pleistocene archaeological sites at Mezin, Mezhirich, and elsewhere in the Ukraine were originally held up as evidence of large-scale hunting of woolly mammoths, because the huts built by Paleolithic hunters were made almost exclusively of carefully piled mammoth bones. For instance, the Mezhirich site contains the stacked bones of at least 149 individual mammoths. The bones subsequently yielded radiocarbon ages ranging from 22 000 to 14 000 ¹⁴C years BP (Lister and Bahn, 1994). Upon closer inspection, they were found to be in

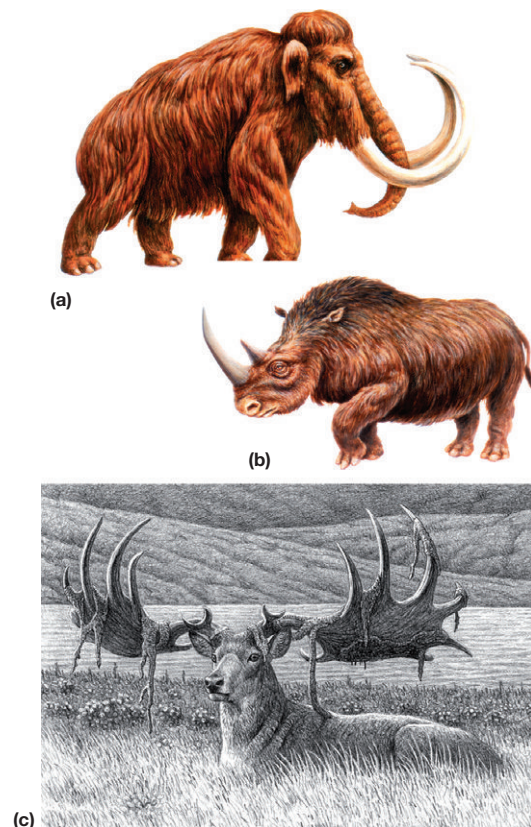


Figure 5 European megafaunal species that became extinct in the late Pleistocene or Holocene. (a) *Mammuthus primigenius*, from Lange (2002, p. 176); (b) *Coelodonta antiquatus*, from Lange (2002, p. 136); (c) *Megaloceros giganteus*, from Poortvliet (1994, p. 173).

various states of preservation, and many had been gnawed by carnivores. It now appears that the humans who made shelters from mammoth bones had not killed the mammoths in large-scale attacks, but had rather collected most, if not all, of the bones of dead mammoths from the treeless landscape for use as building materials. The bone huts thus bear testimony to a huge thriving mammoth population in this region during the last Pleistocene.

Theories on the Causes of Late Pleistocene Extinction

To return to a question posed earlier, why should some regions of the world have been relatively immune from megafaunal extinctions while others suffered losses of up to 80% of their large animal genera? What can we glean from the geographic patterns discussed above? Finally, what role, if any, did human beings play in the demise of these large animals? The last question has driven the study of late Pleistocene megafaunal extinctions in the last 40 years. Researchers who have worked on this topic seem to be divided into two main camps: those who believe that humans played a significant role in these extinctions, and those who lay the blame on natural environmental change. As we shall see, the debate has become quite heated at times, with some of the environmental-change champions (perhaps prematurely?) declaring victory over their opponents in the human-induced extinction camp.

Martin's Pleistocene Overkill Hypothesis

The paleontologist, [Paul Martin \(1967\)](#), wrote a highly influential chapter entitled 'Prehistoric Overkill' for a book entitled *Pleistocene Extinctions, the Search for the Cause*. In this publication, and in his subsequent (1984) book chapter, he propounded the theory that humans were the chief cause of late Pleistocene megafaunal extinctions. Martin's theory rested on several tenets that have since been attacked by other authors. He argued that: (1) the large Pleistocene mammals were decimated by in these extinctions; (2) small mammals (except on islands) were relatively untouched by extinction during the late Pleistocene; (3) large mammals survived best in Africa; (4) the extinctions were often quite sudden; (5) the extinctions took place at different times on different continents; (6) the extinctions occurred without replacement by new taxa; (7) the extinctions followed in man's footsteps; and (8) the archaeology of extinction is obscure. Martin argued that large mammals were the preferred prey of Paleolithic hunters who specialized in big game hunting. Species living on continents not previously inhabited by people would have had little or no fear of humans, making them easier to kill. For the continents of North and South America and Australia, newly colonized by humans during the last Ice Age, Martin envisioned waves of human hunters, fanning out across new lands, killing the large mammals in a 'blitzkrieg,' thus bringing on megafaunal extinction. Such decimation failed to occur in Eurasia and Africa precisely because humans had been hunting on these continents for hundreds of thousands of years. Hence the megafauna of Eurasia and Africa, wary of human predation, were not as easily killed.

Few subsequent authors have really argued with Martin's points 1, 2, 3, 5, 6, and 8. The chief arguments have been over points 4 and 7, which are the most controversial because the evidence concerning them is so difficult to pin down. Let us deal with points 4 and 7 as best we can, in light of recent investigations.

The Environmental-Change Hypothesis

What is clear from the fossil record is that megafaunal extinctions occur equally in areas where there are few or indeed no people, thereby casting doubt on the requirement for human intervention in the process. The suddenness of the extinction events has been called into question by a number of authors. Grayson (1987) began his attack on the megafaunal overkill theory shortly after Martin's major work on the subject was published. Arguing from the North American data, he suggested that numbers of individuals of megafaunal species may already have dwindled before 14 000 years ago, and that many of the genera that have yet to be dated to terminal Wisconsin (last glacial) times may not have survived beyond 14 000 years ago. In the almost 20 years since this article was written, a considerable number of new radiocarbon ages have been established for last-known examples of many North American megafaunal species. By the late 1990s, a suite of 140 AMS ^{14}C dates had been done on proteins extracted from fossil bones of North American megafaunal mammals that died near the end of the last glaciation. The results of that study were quite startling ([Elias, 1999](#)). It appears that the major megafaunal extinction event took place at 13 300 years ago (11 400 ^{14}C years BP). This event included the extinction of camels, horses, giant sloths, Pleistocene bison, and all other genera of megafaunal mammals that did not survive into the Pleistocene, with the exception of the proboscideans. Mammoths and mastodons persisted until about 12 850–12 830 years ago (10 900–10 850 ^{14}C years BP). So it now appears that there were two distinct extinction episodes in North America, with each event taking less than 100 years.

The older event may have taken place before, during, or just after the first appearance of Clovis artifacts in the archaeological record. The Clovis culture is the first widely accepted human occupation in the ice-free regions of North America. The most recent evaluation of the age of Clovis sites in North America ([Taylor et al., 1996](#)) places the beginning of this culture about 13 500 years ago (11 500 ^{14}C years BP). If the people who used Clovis tools arrived in North America at that time, then they overlapped only very briefly with many megafaunal mammal species. Is there a cause and effect or merely a coincidence here? There is good (although sparse) evidence that Clovis hunters killed and butchered mammoth and mastodon, the megafaunal species that died out a few centuries later. Whether human predation was entirely responsible for the proboscidean extinction remains an open question.

In their review of the topic, Grayson and Meltzer (2003) argue that Martin's overkill model does not work for North America, because there is virtually no evidence that supports it, and there is a large body of evidence that argues against it. The lack of archaeological evidence for overkill is something that Martin himself predicted in point 8 of his overkill theory. He argued that the extinction event took place so rapidly that it

would have left little or no trace on the landscape. Paleoindians were apparently nomadic people, and their wanderings left few permanent marks on the landscape in any case. It is difficult, at best, to argue from negative evidence, and Grayson and Meltzer (2003) castigate Martin for so doing, but Fiedel and Haynes (2003), in their response to Grayson and Meltzer's article, point out that large animal bones have rarely if ever been found in any North American archaeological sites, spanning the past 12 500 radiocarbon years. Therefore if no Holocene 'big game hunting' sites can be found, one cannot expect the Pleistocene to be any better represented.

Proponents of environmental change as the instrument of megafaunal extinction in North America often point to the large-scale changes in plant communities that occurred during the Pleistocene–Holocene transition. These changes, mostly noted in fossil pollen records, yielded plant communities for which there are no modern analogs. However, Gill et al. (2009) studied a suite of Late Glacial-age sites in the eastern and central United States, and concluded that it was the loss of megafaunal herbivores that altered plant community structure, not the other way around. They argue that the decline in megafaunal populations released palatable hardwood species from herbivore pressure, bringing about substantial changes in regional plant communities.

Another open-ended question that affects this debate concerns the time of arrival of humans in the Americas. The traditional 'Clovis-First' view that was prevalent when Martin formulated his overkill theory. In the past 20 years, a number of candidate archaeological sites have come to light that indicate a pre-Clovis occupation of the Americas, the strongest of which is the Monte Verde site in Chile (Dillehay, 1997). Evidence from this site unequivocally places people in South America by 15 000 cal years BP. Bona fide pre-Clovis sites have been more difficult to document in North America. Sites such as Meadowcroft, Pennsylvania (Adovasio and Carlisle, 1984), Cactus Hill, Virginia, and Mud Lake, Wisconsin (Falk, 2004), offer tantalizing, if controversial, evidence that people may have been in North America 15 000–20 000 years ago. If these claims can be substantiated, then Martin's theory would be considerably weakened, as one of its principal tenets ('extinction followed in man's footsteps') would be invalidated.

Scott (2010) raised the issue that late Pleistocene megafaunal extinctions in North America may have been driven by competitive pressure from the genus *Bison*, recently arrived from Eurasia across the Bering Land Bridge. In his view, the spread of *Bison antiquus* on the grasslands south of the Laurentide and Cordilleran ice sheets during the last glaciation constituted a major change in regional ecosystems. This species and *Bison bison* formed very large herds and enjoyed competitive advantages over other megafaunal grazers (Guthrie, 1980). Scott proposed that this competitive advantage, in combination with environmental changes at the end of the last glaciation, was the dominant force in driving other the megafauna to extinction in North America. Previous authors had argued that bison had only risen in numbers after the other megafauna became extinct, but radiocarbon dated remains of bison and the other megafauna (Figure 6) show that bison were already increasing in abundance before the demise of the other megafauna.

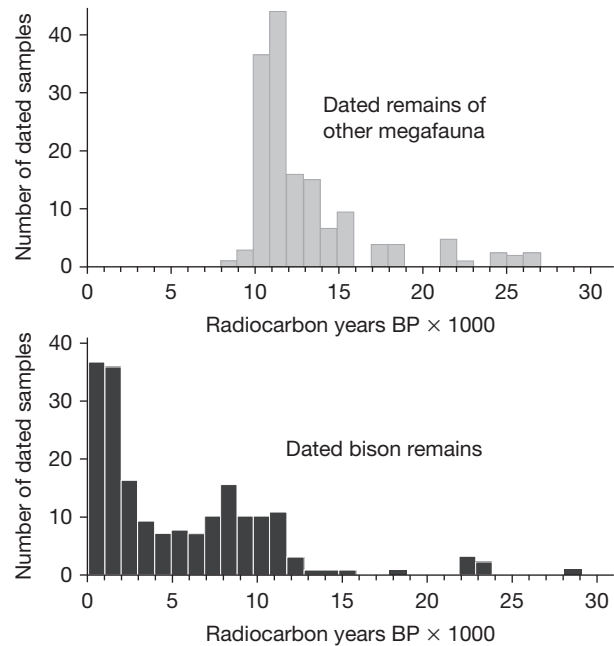


Figure 6 Bar graphs depicting the number of radiocarbon-dated samples of *Bison* and all other megafaunal species in thousand-year increments, during the last 30 000 years in North America. Data from McDonald (1981) and Guthrie (2006).

Meanwhile, in Australia, the debate about Pleistocene overkill likewise rages on. Miller et al. (1999) discussed evidence for the extinction of the large flightless bird, *Genyornis*, dating this extinction to 50 000 years ago, based on ages from fossil eggshells. They argued that people had arrived in Australia by this time, and that humans were either directly responsible for the demise of *Genyornis* and other Australian megafauna, or indirectly responsible. The latter argument asserts that the early human inhabitants of Australia set fire to large tracts of grassland, in order to boost grassland productivity at the expense of trees and shrubs, and that this caused the collapse of long-established elements of the ecosystem, including the demise of browsers that relied on tree and shrub vegetation for food. In the Miller et al. (1999) argument, the demise of megafaunal predators followed the extinction of their large herbivore prey. Roberts et al. (2001) also discussed evidence from 27 sites that a continent-wide extinction event took place in Australia, about 46 000 years ago. Turney et al. (2008) added more evidence for human-induced megafaunal extinction in Australia. They presented refined age estimates for the timing of megafaunal extinction on the island of Tasmania, which was connected to the mainland during intervals of low sea level. They showed that some megafaunal species survived on Tasmania until at least 41 000 years BP – after the time of their extinction on the mainland but coinciding with the timing of the arrival of humans. They also presented evidence showing that no significant environmental changes took place in Tasmania between 43 000 and 37 000 years ago, when a land bridge connected the island to the mainland. They concluded that human hunting pressure drove the Tasmanian megafauna to extinction.

However, the environmental-change camp also has its champions in Australia. Wroe et al. (2004) argued that no megafaunal kill sites have been found in Australia, and that it is probable that early human populations were too small to bring about rapid extinctions of large mammals. They also argued that more recent evidence from the Cuddie Springs site indicates that *Genyornis* did not die out there until 40 000 years ago. Johnson (2005) takes the middle ground between the two extreme views when he argues that, despite evidence for younger ages of extinctions in various parts of Australia, “the evidence supports overkill, not climate change, as the cause of the extinctions.” For obvious reasons, he rejects the ‘blitzkrieg’ model of extinction, but supports humans as the cause, albeit on a slower time scale.

The Hyperdisease Theory

An alternative theory for megafaunal extinction has been proposed by MacPhee and Marx (1997). In their theory, a particularly virulent strain of an unknown, deadly disease was able to make the jump from one mammal species to another, becoming a hyperdisease that wiped out the megafauna. In the hyperdisease theory, humans may have carried the source of infection from one continent to another as they spread across the world. If the pathogen was able to infect immunologically vulnerable populations of mammals, it would have caused rapid population crashes, possibly leading to extinction. Humans were not necessarily the vectors of such a disease. The carriers could have varied from place to place, following in the footsteps of human colonization. MacPhee and Marx include rats or fleas, or parasites of rats and fleas, on their list of potential disease vectors. As of this writing, no firm evidence has been found to support this theory.

The Bolide Impact Theory

Another theory of late Pleistocene megafaunal extinction was published by Firestone et al. (2007). These authors suggested that the cause of this extinction, plus the cause of the demise of the Clovis culture in North America, was a meteoric fireball, or bolide, that detonated over North America at the onset of the Younger Dryas interval. According to this theory, the exploding bolide kindled wildfires across the continent, leading to a catastrophic reduction in the food plants needed by megafaunal herbivores. The authors claim to have found micrometeorite fragments imbedded in fossil megafaunal mammal bones from this interval, and that sediments from a broad region of North America have been found to contain magnetic grains with iridium, magnetic microspherules, charcoal, soot, carbon spherules, and nanodiamonds (Kennett et al., 2009). This phenomenon is supposed to have triggered the Younger Dryas cooling throughout the Northern Hemisphere, as the meteor struck the Laurentide ice sheet, causing catastrophic melting, sending large volumes of cold fresh water into the North Atlantic, and thereby interfering with thermohaline circulation. However, closer inspection of various facets of this theory has led many scientists to reject the theory. Fossil bones that contain the proposed meteorite fragments would all have to date from the same time frame in order to fit the theory, but this appears not to be the case. In fact, some of the bones in

question date to as much as 20 000 years BP (Dalton, 2007). A study by Paquay et al. (2009) found that terrestrial and marine sediments dating to the Bölling–Allerød/Younger Dryas boundary zone did not contain evidence of enriched extraterrestrial elements, such as iridium. Finally, Daulton et al. (2010) examined sediment samples from this time interval, looking for nanodiamonds. What they found instead were graphenes and graphene oxides, which they claim have been misidentified as nanodiamonds by Kennett and co-workers. Graphenes, a natural component of charcoal, occur in carbon spherules that are found throughout late Pleistocene and Holocene sediments. The weight of the evidence thus appears to negate the bolide impact theory of megafaunal extinction.

Can We Find the Smoking Gun?

The indisputable evidence for human-caused extinction of megafauna during the late Pleistocene, the so-called ‘smoking gun’ (or perhaps ‘blood-dripping spear’ would be a better metaphor), will probably never be found. There are several reasons for this. (1) Both the faunal and archaeological evidence are fragmentary in nature. We will never have a complete picture of what happened. This will always leave room for debate. (2) The dating of events will never be exact. Radiometric dating is the only way to date most paleontological or archaeological sites, and all radiometric dates come with uncertainties. In this sense, they are not true ages, as one would get from counting back annually layered sediments or tree rings. They remain age estimates, plus or minus centuries of time (when dealing with ages that are tens of thousands of years before present). (3) The early peoples of all the inhabited continents were hunter–gatherers who were mostly nomadic in lifestyle. They left pitifully little evidence of their occupation in the first place; what remains *in situ* after tens of thousands of years of decomposition, erosion, reworking of sediments, etc., will always be just a tiny fraction of the original material, be it mammal bones or human artifacts.

One scenario that would seem to fit the most facts might be a *coup de grâce* model. In this model, human predation contributes to the demise of the Pleistocene megafauna, to some degree. Indeed, the very nature of the megafauna – slow to mature and reproduce, with limited capacity to recover numbers rapidly following any perturbation – may only have needed a modest increase in human hunting to deliver the final blow. Science has yet to determine the extent of human influence on megafaunal extinction, but it seems most likely that humans played a larger role in some regions than in others. The modeling efforts of Alroy (2001) and Brook and Bowman (2002) were at least more dispassionate attempts to deal with the data in hand. In the coming decades, we may begin to approach a consensus, but it seems very doubtful that the definitive answers will ever be obtained.

See also: Vertebrate Overview. **Vertebrate Records:** Early and Middle Pleistocene of Northern Eurasia; Late Pleistocene of Africa; Late Pleistocene of North America; Late Pleistocene of South America; Late Pleistocene of Southeast Asia.

References

- Adovasio JM and Carlisle RC (1984) An Indian hunters' camp for 20,000 years. *Scientific American* 250: 130–136.
- Alberdi MT, Miotti L, and Prado JL (2001) *Hippidion saldiasi* Roth, 1899 (Equidae, Perissodactyla), at the Piedra Museo Site (Santa Cruz, Argentina): Its implication for the regional economy and environmental reconstruction. *Journal of Archaeological Science* 28: 411–419.
- Alroy J (2001) A multispecies overkill simulation of the end-Pleistocene megafaunal mass extinction. *Science* 292: 1893–1896.
- Ayliffe LK, Prideaux GJ, Bird MI, et al. (2008) Age constraints on Pleistocene megafauna at Tight Entrance Cave in southwestern Australia. *Quaternary Science Reviews* 27: 1784–1788.
- Barnosky AD and Lindsey EL (2010) Timing of Quaternary megafaunal extinction in South America in relation to human arrival and climate change. *Quaternary International* 217: 10–29.
- Barnosky AD, Koch PL, Feranec RS, Wing SL, and Shabel AB (2004) Assessing the causes of Late Pleistocene extinctions on the continents. *Science* 306: 70–75.
- Borrero LA (2003) Taphonomy of the Tres Arroyos 1 Rockshelter, Tierra del Fuego, Chile. *Quaternary International* 109–110: 87–93.
- Borrero LA, Zarate M, Miotti LL, and Massone M (1998) The Pleistocene–Holocene transition and human occupations in the Southern Cone of South America. *Quaternary International* 49–50: 191–199.
- Brook BW and Bowman DM (2002) Explaining the Pleistocene megafaunal extinctions: Models, chronologies, and assumptions. *Proceedings of the National Academy of Sciences USA* 99: 14624–14627.
- Dalton R (2007) *Mammoth tusks show up meteorite shower*. Nature News (online). <http://www.nature.com/news/2007/071212/full/news.2007.372.html>.
- Daulton TL, Pinter N, and Scott AC (2010) No evidence of nanodiamonds in Younger–Dryas sediments to support impact event. *Proceedings of the National Academy of Sciences USA* 107: 16043–16047.
- Dillehay TD (1997) Monte Verde: A Late Pleistocene Settlement in Chile. *The Archaeological Context and Interpretation*, vol. 2. Washington, DC: Smithsonian Press.
- Dillehay TD and Collins MB (1991) Monte Verde Chile: A comment on Lynch. *American Antiquity* 56: 333–341.
- Elias SA (1999) Quaternary paleobiology update: Debate continues over the cause of Pleistocene megafauna extinction. *Quaternary Times* 29: 11.
- Emslie SD (1986) Late Pleistocene vertebrates from Gunnison County, Colorado. *Journal of Paleontology* 60: 170–176.
- Emslie SD and Morgan GS (1995) Taphonomy of a Late Pleistocene carnivore den, Dade County, Florida. In: Steadman DW and Mead JI (eds.) *Late Quaternary Environments and Deep History: A Tribute to Paul S. Martin*, pp. 65–83. South Dakota: The Mammoth Site of Hot Springs. Scientific Papers 3.
- Falk T (2004) Wisconsin dig seeks to confirm Pre-Clovis Americans. *Science* 305: 590. FAUNMAP Working Group (2003) <http://www.museum.state.il.us/research/faunmap/>.
- Faure M, Guerin C, and Parenti F (1999) *Comptes Rendus de l'Académie des Sciences, Serie II: Sciences de la Terre et des Planètes* 329: 443.
- Ficcarelli G, Coltori M, Moreno-Espinosa M, Pieruccini PL, Rook L, and Tone D (2003) A model for the Holocene extinction of the mammal megafauna in Ecuador. *Journal of South American Earth Sciences* 15: 835–845.
- Field J and Fullager R (2001) Archaeology and Australian megafauna. *Science* 294: 7.
- Firestone RB, West A, Kennett JP, et al. (2007) Evidence for an extraterrestrial impact 12,900 years ago that contributed to the megafaunal extinctions and the Younger Dryas cooling. *Proceedings of the National Academy of Sciences USA* 104: 16016–16021.
- Flannery TF (1994) *The Future Eaters*. Sydney: Reed New Holland.
- Flannery TF and Roberts RG (1999) Late Quaternary extinctions in Australasia: An overview. In: MacPhee RDE (ed.) *Extinctions in Near Time: Causes, Contexts, and Consequences*, pp. 239–256. New York: Kluwer Academic/Plenum Press.
- Gonzalez S, Kitchener AC, and Lister AM (2000) Survival of the Irish elk into the Holocene. *Nature* 405: 753–754.
- Grün R, Eggins S, Aubert M, Spooner N, Pike AWG, and Müller W (2010) ESR and U-series analyses of faunal material from Cuddie Springs, NSW, Australia: Implications for the timing of the extinction of the Australian megafauna. *Quaternary Science Reviews* 29: 596–610.
- Guthrie RD (2006) New carbon dates link climatic change with human colonization and Pleistocene extinctions. *Nature* 441: 207–209.
- Horton DR (1989) Red kangaroos: Last of the Australian megafauna. In: Martin PS and Klein RG (eds.) *Quaternary Extinctions: A Prehistoric Revolution*, pp. 639–680. Tucson: University of Arizona Press.
- Johnson E (ed.) (1987) *Lubbock Lake: Late Quaternary Studies on the Southern High Plains*. College Station: Texas A&M Press.
- Johnson CN (2005) What can data on late survival of Australian megafauna tell us about the cause of their extinction? *Quaternary Science Reviews* 24: 2167–2172.
- Kennett DJ, Kennett JP, West A, et al. (2009) Nanodiamonds in the Younger Dryas boundary sediment layer. *Science* 323: 94.
- Klein RG (1994) The long-horned African buffalo (*Pelorovis antiquus*) is an extinct species. *Journal of Archaeological Science* 21: 725–733.
- Lange IM (2002) *Ice Age Mammals of North America. A Guide to the Big, the Hairy, and the Bizarre*. Missoula, MT: Mountain Press.
- Lee-Thorp JA and Beaumont PB (1995) Vegetation and seasonality shifts during the Late Quaternary deduced from $\delta^{13}C/\delta^{12}C$ ratios of grazers at Equus Cave South Africa. *Quaternary Research* 43: 426–432.
- Lister A and Bahn P (1994) *Mammoths*. New York: Macmillan.
- Long J, Archer M, Flannery T, and Hand S (2002) *Prehistoric Mammals of Australia and New Guinea: One Hundred Million Years of Evolution*. Baltimore: Johns Hopkins University Press.
- Long A, Sher A, and Vartanyan S (1994) Holocene mammoth dates. *Nature* 369: 364.
- MacPhee RD and Marx PA (1997) The 40,000-year plague: Humans, hyperdisease, and first-contact extinctions. In: Goodman SM and Patterson BD (eds.) *Natural Change and Human Impact in Madagascar*, pp. 169–217. Washington, DC: Smithsonian Institution Press.
- Martin PS (1967) Prehistoric overkill. In: Martin PS and Wright HE (eds.) *Pleistocene Extinctions, the Search for the Cause*, pp. 75–120. New Haven: Yale University Press.
- Martin PS (1984) Prehistoric overkill: The global model. In: Martin PS and Klein RG (eds.) *Quaternary Extinctions*, pp. 354–403. Tucson: University of Arizona Press.
- Martin PS and Steadman DW (1999) Prehistoric extinctions on islands and continents. In: MacPhee RDE (ed.) *Extinctions in Near Time: Causes, Contexts, and Consequences*, pp. 17–53. New York: Kluwer Academic/Plenum Press.
- Martinez GA (2001) 'Fish-tail' projectile points and megamammals: New evidence from Paso Otero 5 (Argentina). *Antiquity* 75: 523–528.
- McDonald JN (1981) *North American Bison: Their Classification and Evolution*. Berkeley: University of California Press.
- McKenna MC and Bell SK (1997) *Classification of Mammals Above the Species Level*. New York: Columbia University Press.
- Miller GH, Magee JW, Johnson BJ, et al. (1999) Pleistocene extinction of *Genyornis newtoni*: Human impact on Australian megafauna. *Science* 283: 205–208.
- Núñez L, Varela J, Casamiquera R, Schiappacasse V, Niemeyer H, and Villagran C (1994) Cuenca de Taguatagua en Chile: el ambiente del Pleistoceno superior y ocupaciones humanas. *Revista Chilena de Historia Natural* 67: 503–519.
- Paquay FS, Goderis S, Ravizza G, et al. (2009) Absence of geochemical evidence for an impact event at the Bølling–Allerød/Younger Dryas transition. *Proceedings of the National Academy of Sciences USA* 106: 21505–21510.
- Poortvliet J (1994) *Journey to the Ice Age: Mammoths and Other Animals of the Wild*. New York: Harry N. Abrams, Inc.
- Roberts RG, Flannery TF, Ayliffe LK, et al. (2001) New ages for the last Australian megafauna: Continent-wide extinction about 46,000 years ago. *Science* 292: 1888–1892.
- Scott E (2010) Extinctions, scenarios, and assumptions: Changes in latest Pleistocene large herbivore abundance and distribution in western North America. *Quaternary International* 217: 225–239.
- Steadman DW, Stafford TW Jr., and Funk RE (1997) Nonassociation of Paleoindians with AMS-dated Late Pleistocene mammals from the Dutchess Quarry Caves New York. *Quaternary Research* 47: 105–116.
- Stuart AJ (1991) Mammalian extinctions in the late Pleistocene of northern Eurasia and North America. *Biological Reviews* 66: 453–562.
- Stuart AJ and Lister AM (2007) Patterns of Late Quaternary megafaunal extinctions in Europe and northern Asia. *Courier Forschungsinstitut Senckenberg* 259: 287–297.
- Stuart AJ and Lister AM (2011) Extinction chronology of the cave lion *Panthera spelaea*. *Quaternary Science Reviews* 30(17–18): 2329–2340.
- Turney CSM, Flannery TF, Roberts RG, et al. (2008) Late-surviving megafauna in Tasmania, Australia, implicate human involvement in their extinction. *Proceedings of the National Academy of Sciences USA* 105: 12150–12153.
- Vizcaino SF, Milne N, and Bargo MS (2003) Limb reconstruction of *Eutatus seguini* (Mammalia: Xenarthra: Dasypodidae). Paleobiological implications. *Ameghiniana* 40: 89–101.
- Wroe S (2002) A review of terrestrial mammalian and reptilian carnivore ecology in Australian fossil faunas, and factors influencing their diversity: The myth of reptilian domination and its broader ramifications. *Australian Journal of Zoology* 50: 1–24.

Late Pleistocene Mummified Mammals

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Introduction

These well-preserved Ice Age (Pleistocene about 2 Ma to 10 ka) mammal specimens are the ‘crown jewels’ of their kind. Along with a few closely observed, beautifully executed depictions of Paleolithic cave art, they give us the best idea of the appearance of the now-extinct mammals.

The mummies have also yielded important information on: (1) taxonomic relations and dispersal history through the study of ancient DNA; (2) soft tissue structure; (3) paleodiet based on gut contents and associated feces; (4) paleopathology; (5) parasites; (6) predators and scavengers through the study of tooth-puncture marks in the skin and gnaw marks on the bone; and (7) paleoclimatic evidence inferred from plant remains in gut contents, ground squirrel nesting grasses, fecal pellets, and seed caches, as well as from the nature of enclosing sediments.

At least 16 species of Ice Age mammals have been mummified: woolly mammoth, Shasta ground sloth, Jefferson’s ground sloth, Patagonian ground sloth, woolly rhinoceros, Yukon horse, steppe bison, helmeted muskox, Harrington’s mountain goat, caribou, giant moose, black-footed ferret, collared pika, snowshoe hare, arctic ground squirrel, and vole. Among the best of these specimens are: (1) ‘Dima,’ a 40-ka-old baby woolly mammoth from Siberia; (2) a 10-ka-old Shasta ground sloth from New Mexico; (3) ‘Blue Babe’ a 36-ka-old steppe bison from Alaska; (4) an 18-ka-old subadult helmeted muskox also from Alaska; (5) a woolly rhinoceros from Ukraine; and (6) a 40-ka-old black-footed ferret from Yukon.

Most are preserved by a type of freeze-drying in frozen ground (permafrost) of Siberia, Alaska, and Yukon, and date from 50–25 and 15–10 ka. Presumably this is because extra moisture available during those warmer periods of the last glaciation allowed a quicker covering of the carcasses by melted, near-freezing sediments. The sloths and mountain goats were preserved in cave deposits – perhaps dry air and relatively constant temperature in the caves were mainly responsible for such good preservation. Woolly rhino and woolly mammoth carcasses from Starunia, Ukraine seem to have been ‘pickled’ in salty groundwater and coated by a mineral wax.

Where were the mummies found? How complete were they? How old were they individually and geologically? How were they preserved? These are a few of the questions commonly asked, that will be dealt with under the following categories: (1) circumpolar frozen fauna, (2) United States cave fauna, (3) Patagonian cave fauna, and (4) European ‘pickled’ fauna.

Circumpolar Frozen Fauna

Woolly Mammoths (*Mammuthus primigenius*)

The widely known mummies of Siberian woolly mammoths spring to mind – especially the Berezovka mammoth, one of

the earliest iconic finds made about 1900 by Lamut tribesmen on the Berezovka River; and ‘Dima,’ found in 1977.

The Berezovka mammoth was surprisingly well preserved, but had undergone some decomposition, as have most of the specimens (Lister and Bahn, 1994, pp. 44–45). Carnivores had eaten parts of the exposed head and several bones were broken. It had food between its massive molar teeth, and the nature of the muskox-like hair, the relatively short tail, and thick (90 mm) body fat were noted.

‘Dima’ (Figure 1), the frozen mummy of a young male, was found by a placer gold miner using a bulldozer on a terrace of the Kirgiliakh River, a tributary of the Kolyma River north of Magadan, Russia (Vereshchagin and Mikhelson, 1981). The carcass was intact and very little decomposition had occurred before the body was frozen. He may have become mired in a mud sink (Guthrie, 1990, figure 1.7). Judging from his size and degree of tooth eruption, Dima was about 4 months old. Several radiocarbon dates on tissue indicated that he had died during a relatively warm period of the last glaciation some 40 ka. Plant remains found with the carcass indicate that he had occupied a dry, grassy landscape with some trees near the valley bottom. ‘Dima’ is displayed beside the Berezovka mammoth in the Zoological Museum at St. Petersburg, Russia.

Several other Siberian woolly mammoth carcasses are known. One was found on a small tributary of the Khatanga River in 1977 (Vereshchagin and Nikolaev, 1982). It was probably an adult male about 50 yr old according to the degree of wear on its third molar, its large tusks, and body. Some skin remained on the head, including parts of the trunk and entire left ear. Radiocarbon dates indicate that the animal died about 50 ka. The carcass had decomposed and had been heavily scavenged. Another carcass was found on the Shandrin River in 1971. It was dated to between 42 and 32 ka and was notable for its well-preserved gut contents. Plants from the stomach were: 80% herbs (mainly grasses), 15% woody plants, 1% mosses, 1% heather leaves, and 5% unidentified (Gorlova, 1982). The dominance of grasses is similar to the gut contents of other mammoths suggesting an abundance of grasses in the mammoth habitat (Mammoth Steppe). Other woolly mammoth carcasses from Asia include: (1) ‘Mascha’ a young (about 4 months old) female from the Yamal Peninsula dating to about 39 ka (Mol and van Essen, 1992, p. 117; Lister and Bahn, 1994, pp. 46–47); (2) the Late Glacial (about 10 ka old) Yuribei mammoth, a 12-yr-old-female discovered from the Yuribei River in 1979 (Mol and van Essen, 1992, p. 114); (3) the Taimyr mammoth, an adult male (about 45 yr old) found on the Taimyr Peninsula, Siberia in 1948 (Mol and van, 1992, p. 113); (4) the Fishhook mammoth, an adult male 20.6 ka partial carcass from the Upper Taimyra River found in 1990 that had been grazing on moist grassy (98% Poaceae) tracts on the Taimyr Peninsula (Mol and Kahlke, 2004); (5) Adams mammoth near the mouth of the Lena River; (6) a carcass



Figure 1 Left side of the carcass of the baby mammoth (*Mammuthus primigenius*) 'Dima' from frozen ground on a terrace of the Kirgiliakh River northeastern Siberia, Russia (Zoological Institute, St. Petersburg).

from the Liakhov Islands; (7) a carcass from Sanga-Yurakh (see map in [Lister and Bahn \(1994\)](#), pp. 154–155), and (8) a 41 ka carcass of a well preserved baby collected in 2004 at the Olchan Goldmine in Oymyakonsky Ullus, Yakutia (D. Mol, Personal Communication 2006).

In 1948 a partial carcass (head and attached left leg) of a baby mammoth named 'Effie' was found at Fairbanks Creek, Alaska. It was radiocarbon dated to about 21 ka and may have died in its first year, according to its estimated shoulder height of 1 m. A mummified mammoth foot was also discovered at this site in 1940 ([Guthrie, 1990](#), figure 1.26).

There are stories by workers on the gold dredges near Dawson City, Yukon of excavated mammoth carcasses, but the only evidence I have seen during the collection of Pleistocene mammal remains in that area for about 30 years is a mammoth eye socket with attached shreds of dried flesh.

Woolly Rhinoceros (*Coelodonta antiquitatis*)

Woolly rhino mummies are known from frozen deposits in Siberia, as well as from petrochemical seeps in Ukraine. The Churapcha rhino was found in 1972 by a villager digging a cellar in a small settlement on the Lena–Amga interfluvium ([Lazarev, 1977a](#)). In 1971 an entire carcass was found on the Vilyuy River, but only the head and two legs were saved. Another carcass was discovered on the Kahlbuy River (middle Yana drainage) in 1877 ([Sutcliffe, 1985](#); [Guthrie, 1990](#)).

Horse (*Equus lambei* and *Equus* sp.)

Several carcasses of horses are known from the circumpolar regions. Radiocarbon dates on the Selerikan horse, a mature male, indicate a geological age between 39 and 35 ka ([Lazarev, 1977b](#)). It was found in 1968 on the Selerikan River (near the upper part of the Indigirka River), Siberia. The body was well preserved, missing only the head and neck. Perhaps it had been

mired up to its neck, and the head and neck were removed later by predators or scavengers ([Guthrie, 1990](#), figure 1.12). The body was found from below by drift gold miners who unknowingly at first used the lower hind legs to hold cables and lanterns! Pollen found in the horse's intestinal tract suggested a late-summer death. The hair was reddish brown with a dark mane, tail and legs, and the underside was yellowish to white.

This was not the first horse mummy from Siberia. In 1878, a white horse carcass thawed from frozen ground on the Yana River was not collected. Further, a mummified mare containing a large embryo, found on the upper Indigirka River, was excavated in 1950, but only a part of it was saved for study ([Lazarev, 1977b](#)).

A partial carcass of an adult Yukon horse (*Equus lambei*) was found by placer miners at Last Chance Creek near Dawson City, Yukon ([Harington and Eggleston-Stott, 1996](#)). It consisted of a right foreleg with dried flesh, skin and some dark lower-leg hair, the hide (from ear to tail) with tail, mane, body hair, and gut contents. The upper foreleg bone had been gnawed by medium-sized predators or scavengers, possibly wolves. A radiocarbon date of about 26 ka indicates that it had died during a relatively warm period of the last glaciation. Its grayish white body hair, mane, and tail (like the Yana River horse carcass) suggest that it died in winter. Both Last Chance Creek and Selerikan specimens are related to caballine horses.

In nearby Alaska, three partly preserved horse legs were collected from frozen ground in the Fairbanks area ([Guthrie, 1990](#), figure 1.19). During 1983, a horse skull, radius, and forefoot with hoof, all tissues still attached, was recovered from frozen silt along the Titaluk River, Alaska. Hoof keratin yielded a radiocarbon age of about 17 ka. The structure of the hoof suggests that the animal survived on winter range characterized by shallow snow and low forage quality but high quantity. The rate of hoof growth and wear suggests a relatively sedentary existence and little digging through snow for food ([Guthrie and Stoker, 1990](#)).

Steppe Bison (*Bison priscus*)

Two of the best carcasses of this species are from the Fairbanks area of Alaska. A nearly complete mummy, called 'Blue Babe' because of the bluish iron oxide (vivianite) coloring its surface, was discovered by placer miners near Pearl Creek in 1979. Evidently, it was killed by lions (*Panthera leo atrox* or *P. l. spelaea*) about 36 ka. It was restored and is on display at the University of Alaska Museum, Fairbanks (ML [Guthrie, 1988](#); RD [Guthrie, 1990](#), figure 11.5).

A carcass from Dome Creek ([Figure 2](#)), also a large mature male, was discovered in 1952 and was about 7 yr old at death. Radiocarbon dating indicated that it had died about 28 ka ([Guthrie, 1990](#)). A less complete carcass of a female bison, also about 7 yr old, was found on Fairbanks Creek in 1952. Two partial bison legs are known from Cleary and Goldstream creeks near Fairbanks ([Guthrie, 1990](#)).

It is not uncommon to find steppe bison skulls with well-preserved horns sheaths from Yukon Pleistocene deposits in both Dawson City and Old Crow areas. In addition, a well-preserved forefoot covered by dried skin was collected at Hunker Creek, Dawson City area in 1988.



Figure 2 Right side of a partial carcass of an extinct steppe bison (*Bison priscus*) from frozen ground on Dome Creek near Fairbanks, Alaska. The collector of most of the Alaskan Pleistocene carcasses, including this one, the late Otto Geist, is examining a fragment of hide (T. L. Péwé).

A partial carcass of a young female bison found on the Indigirka River is the only one known from Siberia. She died about 30 ka (Flerov, 1977).

Helmeted Muskox (*Bootherium bombifrons*)

A nearly complete carcass of a subadult (about 2.3 yr old according to tooth development) was found in frozen organic silt by gold miners at Fairbanks Creek, Alaska, and collected by Otto Geist (see Figure 2). It is the only known mummy of the helmeted muskox – a native of North America that never reached Eurasia. It lacks the horncores, and the right side of the trunk has been damaged. Radiocarbon dates indicate that the animal lived at a time near the Last Glacial Maximum, about 18 ka. Hair on the lower limbs is darker than that of the living tundra muskox (*Ovibos moschatus*). The limbs of this specimen are longer and relatively more slender and the tail is longer than are those characters in the tundra muskox. Gut contents were preserved but have not been analyzed (McDonald, 1984).

Giant Moose (*Alces latifrons*)

Several partial carcasses of this extinct moose have been collected in Alaska: (1) a yearling male, discovered on Little Eldorado Creek in 1940, had thick body fat and an attached antler suggesting that it had died in fall or early winter; (2) a young female was found on the same creek near Fairbanks in 1942; and (3) fragments of at least two giant moose carcasses were found at Livingood, Alaska in 1980; one of these fragments yielded a radiocarbon age of 32 ka (Guthrie, 1980, figures 1.23–1.25).

Caribou (*Rangifer tarandus*)

Mummified lower legs have been collected from Upper Cleary Creek near Fairbanks, Alaska in 1945 and Tofty, Alaska in 1948 (Guthrie, 1990, figure 1.19).



Figure 3 Left side of the carcass of a 40-ka-old adult male black-footed ferret (*Mustela nigripes*) from Sixtymile, Yukon, Canada (C. R. Harington).

Black-Footed Ferret (*Mustela nigripes*)

The most spectacular of Yukon Ice Age mammal carcasses is a black-footed ferret (Figure 3) – now extinct in Yukon and nearly so in North America. It is of a young adult male, radiocarbon dated to 40 ka, whose fur differed little in color and pattern from recently caught museum specimens. He was originally collected by ‘Molly,’ a placer miner’s dog at Sixtymile, Yukon in 1984! An X-ray image showed that he had remains of his last meal in his intestine – probably a vole (*Microtus*) according to analysis of hair extracted from the gut (Youngman, 1987). A less complete mummy of an adult female black-footed ferret from Hunker Creek near Dawson City, Yukon was collected in 1987. She yielded a radiocarbon age of about 30 ka (Youngman, 1994).

Arctic Ground Squirrel (*Spermophilus parryi*)

This carcass was picked up by a curious placer miner on Glacier Creek in the Sixtymile area, Yukon in 1981. It became a topic of local conversation, and, being round, was called ‘the moon-egg.’ It was finally identified as an Arctic ground squirrel when rust-colored hair began appearing through the micaceous silt that covered it and part of a skull was seen through a crack in the skin. It consisted of a complete individual curled up nose to tail, that probably had died during hibernation. It yielded the oldest finite radiocarbon age known for any Yukon Ice Age mammal – 47 ka (Harington, 1984; Walker, 1984).

Another Arctic ground squirrel carcass was collected in 1935 at Eva Creek in the Fairbanks area of Alaska (Guthrie, 1990, figure 1.26).

Collared Pika (*Ochotona princeps*)

This carcass was collected at Chatanika, Alaska, just north of Fairbanks in 1933. It has some hair around the digits and the body is flecked with blue vivianite. The occurrence of this Alpine species in central Alaska corroborates the idea that Alpine environments extended into drier, steppe-like lowlands during glacial times (Guthrie, 1973, 1990, figure 1.26).

Snowshoe Hare (*Lepus americanus*) and Vole (*Microtus*)

Other Alaskan mummies include a virtually complete snowshoe hare from Fairbanks Creek collected in 1949, and an unspecified vole carcass (Guthrie, 1990, figure 1.26).

United States Cave Fauna

Shasta Ground Sloth (*Nothrotheriops shastensis*)

The best specimen of this species is from Aden Crater, New Mexico (Lull, 1929). It consists of a complete skeleton with ligaments and other dried soft tissue, including patches of skin and hair. In addition, hair, pieces of skin, claws with their sheaths, and even intervertebral disks have been found in other caves such as Rampart Cave, Arizona (Wilson, 1942; Harrington, 1972) and Gypsum Cave, Nevada (Stock, 1931; Harrington, 1933; McDonald, 2003, figure 1.5).

Jefferson's Ground Sloth (*Megalonyx jeffersonii*)

Dry, soft tissue of this species has been recovered from three individuals in Big Bone Cave, Tennessee (Mercer, 1897; McDonald, 2003, figure 1.2). Some of the bones retain cartilage on the articular surfaces, as well as dried ligaments, and one claw retains the keratinous nail.

Harrington's Mountain Goat (*Oreamnos harringtoni*)

Soft parts of this small, extinct mountain goat, such as horn-sheaths and a well-preserved forefoot have been collected from Rampart Cave, Arizona and at other caves in the western United States (Harrington, 1972; Mead and Lawler, 1994).

Patagonian Cave Fauna

Patagonian Ground Sloth (*Mylodon*)

Mummified ground sloth remains have been found in South America. In 1896, Herman Eberhardt, a local farmer, found a piece of dried skin with short yellow-brown hair in the great cave of Ultima Esperanza in Chile (Patagonia). The following year another skin fragment was discovered by F. P. Moreno, hanging on a nearby tree. The small nodules of bone in the skin suggested that it belonged to the extinct ground sloth *Mylodon*, known to have occupied the cave (Sutcliffe, 1985, figure 12.12). The cave also contains thick deposits of ground sloth dung, keratinous claws, and clumps of hair (D. Schreve, Personal Communication 2006).

European 'Pickled' Fauna

Woolly Mammoth (*Mammuthus primigenius*)

At Starunia in southwestern Ukraine in 1907, a woolly mammoth carcass was found at a depth of 43 m in subsoil filled with veins of kerosene. Some decomposition had occurred before the tissues became embalmed, but this was the only mummified mammoth found outside the circumpolar permafrost regions. Like the woolly rhino mummies from the same site, it had been naturally 'pickled' in a petrochemical seep associated with salt deposits and surrounded by the mineral wax ozocerite. The skeleton was well preserved and the skin still supple, but the hair was firmly attached to the surrounding sediment (Lister and Bahn, 1994, p. 57).

Woolly Rhinoceros (*Coelodonta antiquitatis*)

Besides the carcasses reported from frozen ground in Siberia, 'pickled' specimens are known from the Ukraine. The most famous of all woolly rhino mummies are those from Starunia (formerly in Poland, now in Ukraine) that were discovered in 1907 and 1929 (Guthrie, 1990, figure 1.13). Like the mammoth reported above, they were 'pickled' by a saline solution and coated by ozocerite. Plant remains found with the specimens included dwarf birch and small-leaved willows, showing that tundra conditions colder than present prevailed when the rhinos died (Sutcliffe, 1986, p. 40). Guthrie (1990) notes that Siberian woolly rhinos fed mainly on grasses.

Preservation of Frozen Mummies

In 1970, Reginald Harris, a specialist in freeze-drying at the British Museum (Natural History), now The Natural History Museum in London, concluded that the Patagonian ground sloth (*Mylodon*) skin had been naturally freeze-dried at a time of very cold climate (Sutcliffe, 1985, figure 12.13). That too was my conclusion regarding the preservation of Yukon carcasses. Guthrie's (1990, p. 22) hypothesis regarding the preservation of the steppe bison 'Blue Babe' and the Siberian mammoths is more sophisticated. It involves water withdrawal due to the presence of segregated ice surrounding the mummies that dehydrates them in the frozen ground. He further states that underground frost mummification should not be confused with freeze-drying that occurs when a body is frozen and moisture is removed by sublimation, a process accelerated by a partial vacuum.

See also: [Vertebrate Overview](#). [Vertebrate Studies: Ancient DNA](#).

References

- Flerov CC (1977) Bison of northeastern Asia. *Trudy Akademii Nauk SSSR, Zoologicheskii Institut* 73: 39–56.
- Gorlova RN (1982) Macroremains of plants from the stomach of the Shandrin mammoth. In: Vereshchagin NK and Nikolaev AI (eds.) *The Mammoth Fauna of the Asiatic Part of the USSR*, pp. 34–35. Leningrad: Akademia Nauk, Trudy Zoologicheskii Instituta.
- Guthrie ML (1988) *Blue Babe: The Story of a Steppe Bison Mummy from Ice Age Alaska*. Fairbanks: White Mammoth.
- Guthrie RD (1973) Mummified pika (*Ochotona*) carcass and dung pellets from Pleistocene deposits in interior Alaska. *Journal of Mammalogy* 54(4): 970–971.
- Guthrie RD (1990) *Frozen Fauna of the Mammoth Steppe: The Story of Blue Babe*. Chicago and London: University of Chicago Press.
- Guthrie RD and Stoker S (1990) Paleoeological significance of mummified remains of Pleistocene horses from the North Slope of the Brooks Range, Alaska. *Arctic* 43: 267–274.
- Harrington CR (1972) Extinct animals of Rampart Cave. *Canadian Geographical Journal* 85(5): 178–183.
- Harrington CR (1984) Quaternary marine and land mammals and their paleoenvironmental implications – Some examples from northern North America. In: Genoways HH and Dawson MR (eds.) *Special Publication No. 8: Contributions in Quaternary Vertebrate Paleontology: A Volume in Memorial to John E. Guilday*, pp. 511–525. Pittsburgh: Carnegie Museum of Natural History.
- Harrington CR and Eggleston-Stott M (1996) Carcass of a small Pleistocene horse from Last Chance Creek near Dawson City, Yukon. *Current Research in the Pleistocene* 13: 105–107.
- Harrington MR (1933) Gypsum Cave, Nevada. *Southwest Museum Papers* 8: 1–197.

- Kubiak H (1982) Morphological characters of the mammoths: An adaptation to the arctic-steppe environment. In: Hopkins DM, Matthews JV, Schweger CE Jr., and Young SB (eds.) *Paleoecology of Beringia*, pp. 281–290. New York: Academic Press.
- Lazarev PA (1977a) New finds of the skeleton of a woolly rhinoceros in Yakutia. In: Skarlato DA (ed.) *Proceedings of the Zoological Institute 63: The Anthropogene Fauna and Flora of Northeastern Siberia*, pp. 281–285. Leningrad: Academy of Sciences of the USSR.
- Lazarev PA (1977b) History of the discovery of the carcass of the Selerikan horse and its study. In: Skarlato DA (ed.) *Proceedings of the Zoological Institute 63: The Anthropogene Fauna and Flora of Northeastern Siberia*, pp. 56–69. Leningrad: Academy of Sciences of the U.S.S.R.
- Lister A and Bahn P (1994) *Mammoths*. New York: Macmillan.
- Lull RS (1929) A remarkable ground sloth. *Memoirs of the Peabody Museum of Yale University* 3: 1–39.
- McDonald HG (2003) Sloth remains from North American caves and associated karst features. In: Schubert BW and Mead JI (eds.) *Ice Age Cave Faunas of North America*, pp. 1–16. Bloomington and Indianapolis: Indiana University Press.
- McDonald JN (1984) An extinct muskox mummy from near Fairbanks, Alaska: A progress report. In: Klein DR, White RG, and Keller S (eds.) *Special Report 4: Proceedings of the First International Muskox Symposium*, pp. 148–152. Fairbanks: Biological Papers of the University of Alaska.
- Mead JI and Lawler MC (1994) Skull, mandible, and metapodials of the extinct Harrington's mountain goat (*Oreamnos harringtoni*). *Journal of Vertebrate Paleontology* 14(4): 562–576.
- Mercer HC (1897) The finding of the remains of the fossil sloth at Big Bone Cave, Tennessee in 1896. *Proceedings of the American Philosophical Society* 36: 3–70.
- Mol D and Essen H van (1992) *De mammoet: Sporen uit de ijstijd*. s-Gravenhage: Uitgeverij BZZTôH.
- Mol D and Kahlke R-D (2004) The Fishhook Mammoth: A Late Pleistocene (early Sarmatian) individual of *Mammuthus primigenius* with preserved gut contents from the Upper Taimyra River (Taimyr Peninsula, Arctic Siberia). Poster. Frankfurt am Main: Senckenberg forschungsinstitut und Naturmuseum.
- Stock C (1931) Problems of antiquity presented in Gypsum Cave, Nevada. *Scientific Monthly* 32: 22–32.
- Sutcliffe AJ (1985) *On the Track of Ice Age Mammals*. London: British Museum (Natural History).
- Vereshchagin NK and Mikhelson VM (1981) *The Magadan Baby Mammoth, Mammuthus primigenius*. Leningrad: Akademia Nauk, Trudy Zoologicheskii Instituta.
- Vereshchagin NK and Nikolaev AI (1982) Excavations of the Khatanga mammoth. In: Vereshchagin NK and Nikolaev AI (eds.) *The mammoth fauna of the Asiatic part of the USSR*. Leningrad: Akademia Nauk, Zoologicheskii Instituta.
- Walker L (1984) The 'moon-egg' that contained an arctic ground squirrel. *BIOME* 4(1): 3.
- Wilson RW (1942) Preliminary study of the fauna of Rampart Cave, Arizona. *Carnegie Institution of Washington Publication* 530: 169–185.
- Youngman PM (1987) Freeze-dried mummies are hot items with paleontologists. *BIOME* 7(2): 4.
- Youngman PM (1994) Beringian ferrets: Mummies, biogeography, and systematics. *Journal of Mammalogy* 75(2): 454–461.

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VERTEBRATE STUDIES

Contents

Ancient DNA

Speciation and Evolutionary Trends in Quaternary Vertebrates

Dwarfing and Gigantism in Quaternary Vertebrates

Interactions with Hominids

Ancient DNA

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The Early History of Ancient DNA Research

In 1984, [Higuchi et al. \(1984\)](#) realized a landmark in molecular biology: the molecular cloning through bacterial vectors of fragments of the mitochondrial cytochrome *b* gene of the quagga (*Equus burchelli quagga*). The potential of this approach was instantly appreciated, at least by the popular press, as an exciting new way to study the recent past. The enthusiasm stemmed from the fact that the last quagga had died in an Amsterdam zoo in 1883, and the subspecies declared extinct. The source material for the study was a scrap of tissue that had come to light during remounting of the Amsterdam specimen, which had been on display for more than a century. Higuchi et al. had conclusively shown that ‘ancient’ DNA (aDNA) could survive in preserved tissue for appreciable amounts of time post-mortem.

Due to the difficulty of the bacterial cloning techniques employed in these early studies, new findings were published only sporadically between 1984 and 1987, for example, extraction of nucleic acids from an ancient Egyptian mummy ([Paabo, 1989](#)), which supported the idea that human genes could be similarly accessible to molecular techniques.

Polymerase Chain Reaction

The scope of the field of aDNA research was to change permanently with the advent of polymerase chain reaction (PCR) ([Mullis et al., 1986](#)), a new molecular biology technique that allowed for the exponential production of millions of copies of a targeted region of DNA from as little as one template molecule. PCR is an enzymatic reaction in which a series of heating and cooling cycles exponentially increase the number of DNA strands in an extract, resulting in rapid amplification of the

target DNA sequence. Specificity is obtained by the use of pairs of short single-stranded DNA molecules (primers), complementary to the sequences at either end of the region to be copied, which ‘prime’ the action of the polymerase enzyme.

The exceptional sensitivity of PCR is advantageous for aDNA research, as the initial copy number of DNA sequences is often low in ancient samples. However this sensitivity, combined with the minute amounts and degraded nature of surviving DNA, results in an exceptional risk of contamination, as even trace amounts of modern DNA will preferentially amplify and outcompete an ancient template.

The invention of PCR led to a significant increase in the number of aDNA publications, due to the ease with which DNA could be amplified. Long-standing evolutionary questions on the specific relationships between recently extinct creatures and their presumed relatives were among the first to be addressed. Often these studies merely corroborated pre-existing theories, as when preserved tissue from the marsupial thylacine (*Thylacinus cynocephalus*) was used to confirm its relation to the Australian dasyuromorphs ([Thomas et al., 1989](#)) and bone sampled from mammoth showed it to be a member of the Elephantidae ([Hagelberg et al., 1994](#); [Hoss et al., 1994](#)).

Following on from these early studies, a series of increasingly extreme claims for geologically aDNA were made, for example, DNA from Miocene plant fossils ([Golenberg et al., 1990](#)), Oligomocene termites in amber ([Desalle et al., 1992](#)), and Cretaceous dinosaurs ([Woodward et al., 1994](#)). The DNA preservation levels claimed for these studies seemed incompatible with the known properties of the chemical structure of DNA ([Lindahl, 1993](#)) and it was clear that the burgeoning discipline was in need of some critical input. As a result, a series of reviews were published focusing on the necessity of reproducibility in aDNA, as well as the probability of DNA

survival over geological time (e.g., [Lindahl \(1993\)](#) and [Paabo \(1989\)](#)). These reviews eventually led to a broad consensus regarding the burden of proof for publishing aDNA research, and in turn to the formulation of several criteria checklists for aDNA research consisting of various permutations of the following requirements: physically separated work areas (to prevent spread of modern amplified DNA into ancient specimens), sterile equipment, multiple negative controls, evidence of authentic PCR behavior (i.e., inverse relationship between amplified fragment length and successful amplification), quantitation of starting template number, and the demonstration of reproducibility. The appropriateness of each of these procedures, however, should be evaluated on a case-by-case basis (see [Gilbert et al. \(2005\)](#)).

Sources of Ancient DNA

DNA is preserved in a variety of plant and animal remains, including leaf tissue, seeds, bone, hair, teeth, skin, and fecal/urine waste products. While an organism is alive, an array of enzymes serves to maintain DNA by repairing broken and damaged strands. Immediately following the death of the organism, this system stops, and damaged DNA remains unrepaired. As cellular processes cease and bacterial infection spreads, autolytic endonucleases that normally degrade foreign bodies, DNA, and unwanted products are released from control, causing further DNA degradation. Once these enzymes are either destroyed or deactivated, the rate of decay is considerably slowed. However, DNA will continue to decay spontaneously, mainly due to exposure to various forms of radiation, and the biochemical processes of hydrolysis and oxidation ([Hoss et al., 1996](#); [Lindahl, 1993](#)).

Radiation-induced damage inhibits enzyme amplification by causing complete double-strand breakage of the DNA backbone. DNA fragments may then be cross-linked with proteins and/or broken down into small fragments. Hydrolysis causes depurination/depyrimidation of the nucleotide bases and cleavage of base-sugar (N-glycosal) bonds, creating single-strand nicks in the DNA sequence. Deamination of cytosine to uracil, and depurination, or loss of the purines (adenine and guanine), are the most common types of hydrolytic damage. Oxidative damage, caused by the direct interaction of ionizing radicals with the DNA, is most commonly seen as modifications of sugar residues and of the pyrimidines (cytosine and thymine) to hydantoins. Oxidative damage can cause baseless sites (as in hydrolytic damage) and the formation of interand intrastrand cross-links, which can block the processivity of DNA polymerases used in PCR. A small proportion of damage events do not hinder replication, but generate miscoding lesions. These may manifest as base modifications in a PCR-amplified sequence, changing the appearance of the DNA template and potentially misleading haplotype analyses ([Gilbert et al., 2003](#)).

Fortunately, for aDNA research, the processes of DNA decay can be considerably slowed in certain environments. Although a trend toward lower DNA content with burial time has been observed ([Lindahl, 1993](#); [Tuross, 1994](#)), the impact of the environment on DNA preservation is by far the most important factor in determining the amount of retrievable DNA in a

sample (e.g., [Gilbert et al. \(2003\)](#)). The best environments for long-term preservation of DNA involve particularly cold or rapidly desiccating conditions, where the samples are protected from exposure to UV light. DNA has been shown to survive in samples kept in museums for on the order of tens to several hundred years. Cold, dry environments such as the Arctic or high altitude caves have proved to be particularly conducive to long-term preservation of DNA ([Shapiro and Cooper, 2003](#)). Freezing, shortly after death, limits the amount of DNA damage from enzymes intrinsic to the cell as well as subsequent bacterial or fungal degradation ([Lindahl, 1993](#)). However, [Guthrie \(1990\)](#) has observed that an intact carcass generates enough heat during decomposition to limit freezing, and this may explain why the DNA yield from mummified permafrost specimens is generally much lower than from isolated bones. For this reason, it may be possible that the best DNA samples would come from bones that have been rapidly defleshed after death, preventing bacterial growth in close proximity to the bone. Biochemical data suggest, however, that even in optimal conditions, the upper limit for DNA survival will be on the order of only a few hundred thousand years ([Lindahl, 1993](#)).

Mitochondrial DNA (mtDNA), rather than nuclear DNA, is most often the target in aDNA research, as the mitochondrial genome is present in eukaryotic cells at a much higher copy number than is the nuclear genome, resulting in a greater probability of amplifying intact mtDNA fragments. Mitochondria are organelles found within the cytoplasm of a cell that are responsible for the Krebs cycle and the electron-transport chain, and therefore for the cell's energy production. The number of mitochondria in a cell varies depending on the cell's energetic requirements, and can range from several hundred to several thousand copies per cell. Each mitochondrion has one to four circular copies of mtDNA and exhibits high rates of nucleotide substitutions compared with nuclear DNA ([Moritz et al., 1987](#)). Additionally, mitochondria are maternally inherited, and therefore do not undergo recombination. These factors make mtDNA particularly useful in studies of population genetic structure and phylogenetic relationships among species.

Throughout the process of evolution, partial copies of the mitochondrial genome transfer to the nucleus, resulting in pseudocopies of the mitochondrial DNA sequence referred to as nuclear mitochondrial copies or numts. Fortunately for aDNA research, the number of real mitochondrial sequences is large, based on high starting copy number and similar probabilities of DNA damage and decay, compared to the number of amplifiable numts, so, in most instances it is only possible to amplify the mtDNA sequence. Exceptional preservation combined with a disproportionately large number of numts, however, have made distinguishing the real mtDNA sequence difficult in some taxonomic groups. For example, numts not only caused significant problems in the mammoth study mentioned above, but also call into question the only mastodon sequences reported to date ([Yang et al., 1996](#)).

Recent Directions in Ancient DNA

The parallel increase in the complexity of questions addressed by ancient DNA and the sensitivity of molecular techniques

paved the way for many groundbreaking studies in the 1990s. The primary goal of aDNA research has moved away from sample identification and taxonomy toward phylogeography and demography, which we discuss in turn below. Examples are taken primarily from work conducted on late Pleistocene faunal remains.

Although originally mooted as a potential application of the technology, species-level identification of bone specimens that are incomplete or too badly damaged to classify morphologically has seen little application at present. The financial cost of analyses makes this an impractical option for routine use. However, it remains possible that, when an accurate identification is vital in determining the presence/absence of a species from a crucial site or time period, ancient DNA might be employed as an informative tool.

Considerably, more aDNA research has focused on resolving phylogenetic relationships of extinct vertebrates. Recent examples include the dodo (*Raphus cucullatus*), cave lion (*Panthera spelaea*), sabretooth cats (*Smilodon* and *Homotherium*), cave bear (*Ursus spelaeus*), ground sloths (*Nothrotheriops* and *Myiodon*), giant deer (*Megaloceros giganteus*), and woolly rhino (*Coelodonta antiquitatis*), each with molecular results that broadly corroborate previous morphological analyses. Two studies from New Zealand, however, demonstrate that unexpected relationships are sometimes discovered. The kiwi (*Apteryx* spp.) and extinct moa (Aves: Dinornithiformes), for example, were not found to be as closely related as was previously assumed, a discovery which necessitated reinterpretation of the scenario by which these two species colonized New Zealand (Cooper et al., 1992). It was also discovered that the extinct New Zealand giant eagle, *Haarapagornis moorei*, was very closely related to an Australian genus of little eagles, suggesting an unusually rapid increase in body size (Bunce et al., 2005).

One of the most remarkable accomplishments of aDNA research has been the amplification of mitochondrial control region sequences from the Neanderthal-type specimen (Krings et al., 1997) and subsequent analysis of Neanderthal remains from Mezmaiskaya (Ovchinnikov et al., 2000) and Vindija (Paunovic et al., 2001). To date, ten different Neanderthal specimens have had sections of control region amplified (Beauval et al., 2005), all producing sequences that group together to the exclusion of modern humans.

Since the 1980s, there has been a decided increase both in the amount of molecular data that can be amplified from ancient specimens and in the complexity of statistical models capable of examining phylogenetic relationships between extinct organisms. These developments have made it possible for aDNA to address more sophisticated evolutionary questions, for example, at the level of the population rather than the species. Adding ancient data to population genetic analyses makes it possible to directly observe how populations change through time, and potentially to correlate genetic change with changes in climate or habitat.

The majority of population-level aDNA research has focused on the late Pleistocene (<100 ka BP). The widespread abundance of the late Pleistocene megafauna means that traditional paleontological approaches can be applied globally, rather than to a few key localities. Skeletal morphology, however, is more often applied to the differences between species, rather than providing information about within-species changes in population

structure (see, however, Polly (2003)). Additionally, high-resolution dating techniques, such as radiocarbon dating, are widely employed to provide a temporal scale for the analyses. Dating of skeletal remains, however, is generally insufficient to provide a statistical basis for demographic change, as it simply records the presence of a particular taxon at one point of time.

To address these problems, an important complementary approach has been the use of phylogeography, or the study of spatial patterns of genetic diversity within and between species. Phylogeographic reconstruction can be used to infer the geographic origin of populations and subspecies, the course of range expansions, and the presence of genetic bottlenecks. The timing of these can be estimated using a molecular clock, which is normally calibrated by reference to species divergences attested to in the fossil record. However, phylogeography has traditionally been limited by the need to infer past demographic events from data derived from modern samples. Contemporary patterns of genetic variation are the result of a cumulative series of demographic events, making it difficult to resolve past events if histories are complex. Moreover, little or no signal of demographic events can be observed if they pre-date genetic bottlenecks.

Ancient DNA methodology makes it possible to recover incorporating sequence data directly from radiocarbon-dated bones, providing a mechanism to circumvent these problems. A key example is found in one intensively studied group, brown bears, and reveals surprisingly complex patterns of localized extinctions, replacements, and changing biotic interactions in eastern Beringia during the late Pleistocene (Barnes et al., 2002; Leonard et al., 2000). In this instance, a number of genetic events have generated a relatively simple late Holocene phylogeographic distribution, such that studies focusing exclusively on the genetics of modern populations could easily be misconstrued. Despite the dynamic history, an unexpected degree of phylogeographic structure is observed in late Pleistocene mitochondrial haplotypes of brown bears, suggesting that some form of temporal and spatial regionalism has existed within these populations for a considerable time. Changes to this relatively rigid phylogeographic structure enable the timing of significant environmental events to be inferred, and reveal the location and timing of possible dispersal routes between Beringia and the landmasses south of the ice sheets. Importantly, the timing of these late Pleistocene genetic events do not appear to be tightly related to major climatic changes, and it appears, in the bears at least, that environmental effects may be secondarily mediated through interaction with competitors or other unknown factors (Barnes et al., 2002).

A further development in the exploitation of aDNA sequences has been the inference of demographic change through time. In a study of small mammal remains from a single locality in Wyoming, Hadly and colleagues have utilized coalescent simulations to address questions about the relationship between climate, population size, and genetic diversity (Hadly et al., 2004). This work bridges a gap between the use of mtDNA sequences as population markers and the application of models which consider molecular change through time to derive demographic and evolutionary parameters. These types of models, coupled with heterochronous (different age)

sequence data, have been most extensively applied to studies of viral populations sampled over timescales of months to decades (Drummond et al., 2003).

A limitation of this type of analysis is that, in order to have statistical power, a large amount of genetic diversity must be present in the population being analyzed. This is the case where mutation rates are high (e.g., viruses) or where the amount of time over which the samples are collected is long, as with aDNA. Initial studies applying this approach to data from Holocene Adelie penguin colonies allowed estimation of the mutation rate of the mitochondrial first hypervariable region (HVRI), independent of fossil calibration (Lambert et al., 2002). More recently, Shapiro and colleagues have completed a study of Pleistocene bison, generating sequence data for over 600 bison ranging in age from modern to >65 ka BP (Shapiro et al., 2004). The geographical focus of this study is eastern Beringia (Yukon, Canada and Alaska), but samples from Siberia and the continental United States develop the patterns of movement across the Bering land bridge, and south through the ice-free corridor. They move from this description of late Pleistocene population dynamics to the use of a Bayesian Monte Carlo Markov chain (MCMC) method to derive rates of mutation and dates of divergence from the data, and investigate changes in the effective population size through time. This latter analysis shows the power of the technique to resolve a complex pattern of demographic change: population expansion up to around 30 ka BP, decline through the end Pleistocene and Holocene, subsequent expansion at around 3 ka BP, and an additional abrupt population crash in the last few hundred years. As the experimental and statistical techniques for dealing with aDNA data improve, the number of species for which such population-level analyses have been done will continue to rise, moving toward a much better picture of the late Pleistocene ecosystem.

See also: Vertebrate Overview. **Vertebrate Records:** Late Pleistocene Megafaunal Extinctions; Late Pleistocene Mummified Mammals.

References

- Barnes I, Matheus P, Shapiro B, Jensen D, and Cooper A (2002) Dynamics of Pleistocene population extinctions in Beringian brown bears. *Science* 295: 2267–2270.
- Beauval C, Maureille B, Lacrampe-Cuyaubere F, et al. (2005) A late Neanderthal femur from Les Rochers-de-Villeneuve, France. *Proceedings of the National Academy of Sciences of the United States of America* 102: 7085–7090.
- Bunce M, Szulkin M, Lerner HRL, et al. (2005) Ancient DNA provides new insights into the evolutionary history of New Zealand's extinct giant eagle. *PLoS Biology* 3: e9.
- Cooper A, Mourerchaux C, Chambers GK, Vonhaeseler A, Wilson AC, and Paabo S (1992) Independent origins of New Zealand Moas and Kiwis. *Proceedings of the National Academy of Sciences of the United States of America* 89: 8741–8744.
- Desalle R, Gates J, Wheeler W, and Grimaldi D (1992) DNA-sequences from a fossil termite in Oligomocene amber and their phylogenetic implications. *Science* 257: 1933–1936.
- Drummond AJ, Pybus OG, Rambaut A, Forsberg R, and Rodrigo AG (2003) Measurably evolving populations. *Trends in Ecology and Evolution* 18: 481–488.
- Gilbert MTP, Bandelt H-J, Hofreiter M, and Barnes I (2005) Assessing ancient DNA studies. *Trends in Ecology and Evolution*.
- Gilbert MTP, Willerslev E, Hansen AJ, et al. (2003) Distribution patterns of postmortem damage in human mitochondrial DNA. *American Journal of Human Genetics* 72: 32–47.
- Golenberg EM, Giannasi DE, Clegg MT, et al. (1990) Chloroplast DNA-sequence from a Miocene Magnolia species. *Nature* 344: 656–658.
- Guthrie RD (1990) *Frozen Fauna of the Mammoth Steppe*. London: The University of Chicago Press.
- Hadly EA, Ramakrishnan U, Chan YL, et al. (2004) Genetic response to climatic change: Insights from ancient DNA and phylogenetics. *PLoS Biology* 2: e290.
- Hagelberg E, Thomas MG, Cook CE, Sher AV, Baryshnikov GF, and Lister AM (1994) DNA from ancient Mammoth bones. *Nature* 370: 333–334.
- Higuchi R, Bowman B, Freiberger M, Ryder OA, and Wilson AC (1984) DNA-Sequences from the Quagga, an extinct member of the horse family. *Nature* 312: 282–284.
- Hoss M, Jaruga P, Zastawny TH, Dizdareoglu M, and Paabo S (1996) DNA damage and DNA sequence retrieval from ancient tissues. *Nucleic Acids Research* 24: 1304–1307.
- Hoss M, Paabo S, and Vereshchagin NK (1994) Mammoth DNA-sequences. *Nature* 370: 333–333.
- Krings M, Stone A, Schmitz RW, Krainitzki H, Stoneking M, and Paabo S (1997) Neandertal DNA sequences and the origin of modern humans. *Cell* 90: 19–30.
- Lambert DM, Ritchie PA, Millar CD, Holland B, Drummond AJ, and Baroni C (2002) Rates of evolution in ancient DNA from Adelie penguins. *Science* 295: 2272–2273.
- Leonard JA, Wayne RK, and Cooper A (2000) Population genetics of Ice Age brown bears. *Proceedings of the National Academy of Sciences United States of America* 97: 1651–1654.
- Lindahl T (1993) Instability and decay of the primary structure of DNA. *Nature* 362: 709–715.
- Moritz C, Dooling TE, and Brown WM (1987) Evolution of animal mitochondrial DNA: Relevance for population biology and systematics. *Annual Review of Ecology and Systematics* 18: 269–292.
- Mullis K, Faloona F, Scharf S, Saiki R, Horn G, and Erlich H (1986) Specific enzymatic amplification of DNA *in vitro* – The polymerase chain-reaction. *Cold Spring Harbor Symposia on Quantitative Biology* 51: 263–273.
- Ovchinnikov IV, Gotherstrom A, Romanova GP, Kharitonov VM, Liden K, and Goodwin W (2000) Molecular analysis of Neanderthal DNA from the northern Caucasus. *Nature* 404: 490–493.
- Paabo S (1989) Ancient DNA – Extraction, characterization, molecular-cloning, and enzymatic amplification. *Proceedings of the National Academy of Sciences of the United States of America* 86: 1939–1943.
- Paunovic M, Krings M, Capelli C, et al. (2001) The Vindija hominids: A view of Neandertal genetic diversity. *Journal of Human Evolution* 40: A16–A16.
- Polly D (2003) Palaeophylogeography: the tempo of geographic differentiation in marmots (Marmota). *Journal of Mammalogy* 84: 369–384.
- Shapiro B and Cooper A (2003) Beringia as an Ice Age genetic museum. *Quaternary Research* 60: 94–100.
- Shapiro B, Drummond AJ, Rambaut A, et al. (2004) Rise and fall of the Beringian Steppe bison. *Science* 306: 1561–1565.
- Thomas RH, Schaffner W, Wilson AC, and Paabo S (1989) DNA phylogeny of the extinct marsupial wolf. *Nature* 340: 465–467.
- Tuross N (1994) The biochemistry of ancient DNA in bone. *Experientia* 50: 530–535.
- Woodward SR, Weyand NJ, and Bunnell M (1994) DNA-Sequence from Cretaceous Period bone fragments. *Science* 266: 1229–1232.
- Yang H, Golenberg EM, and Shoshani J (1996) Phylogenetic resolution within the Elephantidae using fossil DNA sequence from the American mastodon (*Mammut americanum*). *Proceedings of the National Academy of Sciences of the United States of America* 93: 1190–1194.

Speciation and Evolutionary Trends in Quaternary Vertebrates

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Introduction

The Quaternary is an ideal interval in which to study the origin of vertebrate species and their adaptations in the fossil record. The time intervals available (10^2 – 10^6 years) are of the order of magnitude at which such processes occur; there is a rich fossil record over a wide geographical area; and paleoenvironmental and dating studies allow one to put the evolving vertebrates into a chronological and ecological context. Recent reviews of the subject have been given by Lister (2004) and Blois and Hadly (2009). This article focuses on phenotypic evolution that can be traced through stratified sequences of fossils. In practice, at the current state of knowledge, this restricts available case histories to mammals. Moreover, the present review focuses on the continental landmasses and excludes hominins; evolution in hominins and on islands is treated separately. Molecular data, based on both modern and ancient samples, are increasingly complementing morphology-based studies of Quaternary vertebrate evolution, and some molecular studies relevant to Quaternary speciation processes in mammals and birds are mentioned here.

The principal evolutionary processes that can be traced in Quaternary mammals are: (a) speciation, that is, the multiplication of species by division of an ancestral lineage into two, known as cladogenesis; and (b) phenotypic change within a single lineage, especially when adaptive, known as anagenesis. An evolutionary trend is here taken to mean a net directional phenotypic shift observed across three or more time intervals. It is an accumulation of phenotypic change that may occur within a single species, or across a series of chronospecies.

Phenotypic changes within Quaternary fossil mammals have frequently been presented, often implicitly, as examples of 'gradualistic evolution,' that is, a series of ever-progressing 'transitional forms' linking nominal species. There are also many examples in the literature where the logic is inverted and fossil deposits are dated on the basis of the evolutionary 'level' of a particular mammal species, with an implicit assumption of a gradualistic underlying pattern. However, according to the 'punctuated equilibrium' model of evolution, most adaptive change is thought to occur during the relatively rapid process of species formation. Between speciation events, evolution is minimal. This model implies a strong linkage between phenotypic change and speciation, and downplays the role of anagenesis, long-term trends being driven instead by selection among differently adapted species (species selection: Gould and Eldredge, 1977).

In reality, theoretical considerations and the study of modern species show us that evolutionary rates are highly variable, so that no particular pattern of change has universal application, but has to be determined in each case from fossil samples dated independently of their 'evolutionary level' (Lister, 1992). Indeed, Darwinian natural selection would

predict complex variations of rate and pattern in the constantly changing environments of the Quaternary.

Studies of evolutionary change also need to take account of geographical effects, especially (a) spatial variation of phenotype and (b) dispersal. Most species today exist as a 'metapopulation' – a complex of geographically separated populations linked by restricted gene flow through migration. In the allopatric speciation model of Mayr (1963), one of these populations becomes partially or wholly isolated at the edge of the range of the parent species. The isolated population evolves new features in its local habitat and at the same time, or on subsequent recontact, becomes reproductively isolated from the parent. On invading the range of the latter, it may either displace it or coexist with it. Another model of evolutionary change that emphasizes geographical variation is Wright's 'shifting balance' theory (Levinton, 1988). Here, the parent species is divided into a set of partially isolated populations, each of which may evolve new features. Occasional gene flow between the populations allows these features to spread through the species as a whole. In a further model, the divergence of abutting populations without isolation (parapatric speciation) may occur if the species' ranges are large enough to allow selection to dominate over gene flow (Jiggins and Mallet, 2000). All of these models of speciation have a characteristic in common that novel morphological features appear in one area first and may then spread over a wider range.

Transferred into the fossil record at Quaternary timescales, these processes have predictable signatures that can be used to test between different models of evolution. The gradual transformation of one morphology into another through a chronological series of fossils, coincident in correlated samples across the geographical range, would suggest anagenetic evolution (transformation of a lineage without splitting) over a wide area. On the other hand, if change is found in a small area while elsewhere the ancestor remained little-changed, it would suggest that sampling is being done in the very area where an allopatric isolate is speciating (cladogenetic evolution). Sampling of later deposits over a wider area may then show the process of spread of the new form. A further important line of evidence is the finding that 'ancestral' and 'descendent' forms co-occur at a single time and place, implying that their ranges have come to overlap. This is inconsistent with purely anagenetic change and implies that a cladogenetic event has occurred, presumably outside the sampled area.

The question of whether speciation rates (or evolutionary rates in general) were elevated in the Quaternary, relative to earlier periods, is contentious (Lister, 2004; Blois and Hadly, 2009). Dramatic environmental changes will undoubtedly have exerted elevated and novel selection pressures on mammalian populations, although their cyclicity in the Quaternary may in some cases have checked significant directional change. The repeated expansion–contraction cycle of species' ranges

into and out of refugia would also have had population genetic effects that may in some cases have promoted speciation (Stewart et al., 2009).

The ideal requirements for demonstrating evolutionary processes in Quaternary mammals are statistical samples of fossils through finely stratified sequences, in correlated deposits across the geographical range, with excellent chronological control and paleoenvironmental proxy data. Only in rare cases are all of these conditions met. In the following account, selected examples illustrating aspects of speciation and evolutionary trends are presented.

Species and Subspecies Origins

The origin of familiar species is a topic of enduring fascination. Questions of interest include the location, time, and duration of a species' origin, the adaptive novelties that arose, and the environmental pressures that produced them.

In an approach to assessing the time required for speciation among Quaternary mammals, Lister and Rawson (2003) and Lister (2004) tabulated some published examples of divergence rates. Starting at the 'short' end of possible timescales, northern mammalian species occupying Beringia in the last glaciation provide a test case, as they were separated between Siberia and Alaska at the last flooding of the Bering Strait, about 10 ka (Elias et al., 1996). Many species today are found on both sides of the strait, for example, the arctic fox, *Alopex lagopus*, but no example of sibling species exists among the mammals, and even nominal subspecies pairs across the strait are often barely divergent: the duration of isolation was evidently too short. An interesting example has been provided by Fedorov et al. (2003) in their study of lemmings (*Lemmus*). This group shows strong phylogeographical division into four haplogroups (groups of related individuals based on mitochondrial DNA (mtDNA)) around the Holarctic, but a single haplogroup of one species, *Lemmus trimucronatus*, straddles the Bering Strait.

On a slightly longer timescale, incipient speciation can in some cases be traced within a single 100 ka Milankovitch cycle, as in shrews of the *Sorex araneus* complex in Europe. Two groups identified by Polly (2001) show not only morphological and genetic differentiation but also karyotypic differences, which appear to have accumulated between populations since the last interglacial.

In various other taxa, 'species pairs' occur on either side of the Bering Strait (Hoffmann, 1981), but events much earlier than 10 ka are implicated. Some of these species pairs are Quaternary in origin, for example, the lynx. This genus is represented by *Lynx lynx* in Eurasia and *Lynx canadensis* in North America, but the fossil form *Lynx issiodorensis* was spread across both continents in the early Pleistocene (about 1 Ma) and is a likely progenitor to both (Kurtén and Anderson, 1980), even though mtDNA divergence of 4% suggests a deeper coalescence of haplotypes, at about 3.2 Ma (Johnson and O'Brien, 1997). In this and many other examples, however, neither the precise timing nor the duration of the speciation process can be estimated from available fossils.

For several living species, mtDNA evidence suggests that diverging clades have survived several 100 ka Milankovitch climatic cycles, with extensive range changes, as well as the

formation of hybrid zones when the populations meet, yet without homogenization or losing their momentum toward speciation. Examples include the incipient red deer/wapiti species *Cervus elaphus*/*Cervus canadensis* (Lister, 2004) and various Siberian lemmings, *Lemmus* spp. (Jarrell and Fredga, 1993; Fedorov et al., 2003). Evidently, divergence leading to speciation can accumulate over several glacial-interglacial cycles.

Avise and Walker (1998), reviewing mtDNA divergence between intraspecific phylogroups of birds, similarly found a range of divergence dates from 75 ka to 4 Ma and concluded that Pleistocene conditions have played an active role both in initiating major phylogeographic separations within species and in completing speciations that had been inaugurated earlier. In a detailed study of southern skuas, Ritz et al. (2008) deduced that allopatric speciation on sub-Antarctic islands had been initiated during MIS 6 (about 200–140 ka) as the expanded ice cap reduced nesting and foraging sites on the Antarctic continent.

Turning to morphology-based studies of fossils, the bear family (Ursidae) provides a good example of adaptive speciation in the Quaternary (Figure 1). The bears that first appeared in Europe in the early Pleistocene, about 1.3 Ma (e.g., at Pietrafitta, Italy; Mazza and Rustioni, 1994), gave rise to three lineages. The first was the omnivorous brown/grizzly bear (*Ursus arctos*), now of broad Holarctic distribution. The other two became more specialized: one led to the largely vegetarian cave bear, *Ursus spelaeus*, and the other to the largely carnivorous, semiaquatic polar bear *Ursus maritimus*.

Fossils with cave bear characteristics first appeared in Europe in sites dated to about 1 Ma, such as Atapuerca TD4, Spain, and Vallonet, France (García, 2003), apparently by transformation of a regional population of the brown bear/cave bear stem lineage, while populations in Asia gave rise to the brown bear that later spread more widely (P. Mazza and N. García, personal communication). This is consistent with a recently obtained mtDNA split date between brown and cave

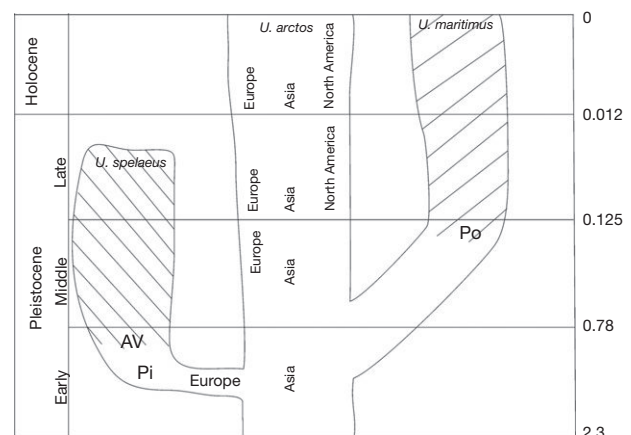


Figure 1 Schematic diagram of speciation in the Pleistocene brown bear lineage. P, Poolepynten; Pi, Pietrafitta; AV, Atapuerca and Vallonet (see text). NW–SE diagonal shading = evidence of 'cave bear' morphology. NE–SW diagonal shading = evidence of polar bear morphology. Approximate timescale in Ma on the right. Speciation date of cave bear based on fossils Pi and AV; speciation date of polar bear based on DNA evidence and known only within broad limits (see text).

bears of about 1–2 Ma (Lindqvist et al., 2010). Until recently, it was assumed that the cave bear lineage was limited to Europe as far as the Urals. Recent finds in the Altai, and even the Yana basin of NE Siberia, indicate a more widespread and complex biogeographic history (Knapp et al., 2009; Sher et al., 2011).

The polar bear, *U. maritimus*, has long been envisaged as having evolved in the middle Pleistocene from a coastal population of brown bears that took to feeding on marine mammals (Kurtén, 1964), but the fossil record is extremely poor. Mitochondrial DNA studies supported Kurtén's supposition, suggesting that the polar bear evolved from within the range of brown/grizzly bear populations. Initial molecular estimates of the date of divergence, based on mtDNA, ranged from 250 to 150 ka (Talbot and Shields, 1996; Barnes et al., 2002; Lindqvist et al., 2010). However, nuclear DNA suggests that the species pre-dates the modern diversity of brown bears, and separated within the Early to early middle Pleistocene, between 0.9 and 0.4 Ma (Edwards et al., 2011; Hailer et al., 2012). Lindqvist et al. (2010) described a mandible from Poolepynten, Svalbard, in the Arctic Ocean north of Norway, from a deposit of estimated age 130 ka, that shows polar bear morphology and an isotopic signature indicating a marine diet, indicating that the species' modern adaptations were in place by at least that date. The phenotypic and ecological separation of the two species has persisted despite evidence of occasional subsequent introgression (Edwards et al., 2011; Hailer et al., 2012).

The mammoth lineage (*Mammuthus*) also shows clear evidence of cladogenetic speciation, but as this was part of a complex process that led to directional evolutionary trend in the lineage, and it is described in the next section.

Evolutionary Trends

Bank Voles (*Clethrionomys*)

Small mammals (principally rodents) have provided many examples of evolutionary trends in the Quaternary, mostly followed in the dentition (Martin, 1984, 1993; Maul et al., 1998). Moreover, certain features are found to have evolved in parallel across many lineages – the most widespread being an increase in crown height, or hypsodonty, generally considered an adaptation to eating increasingly abrasive plant food. A good example is seen in the bank vole lineage *Clethrionomys* (= *Myodes*) in Eurasia, illustrated by Tesakov (1996).

Twelve samples from Eastern Europe and Western Siberia, arranged in chronological order independently of the character in question, show a substantial increase in hypsodonty of the molars across the past 2.5 Myr (Figure 2). The trend is largely unidirectional, but the most significant shifts in mean occur in the early part of the sequence, between 2.5 and 1.5 Ma, with virtual stasis thereafter – an example of varying evolutionary rate. Tesakov (1996) details other morphological changes that occurred through the lineage. The sequence has been codified as a series of chronospecies: *C. kretzoi* – *C. hintonianus* – *C. acrorhiza* – *C. glareolus* (the living species), but it is unclear whether this represents a series of cladogenetic speciation events or a single continuous lineage. There is considerable overlap in hypsodonty ranges between successive samples, consistent with a purely anagenetic mode, but further dated sampling and character analysis, across the species' range, would be required to test this.

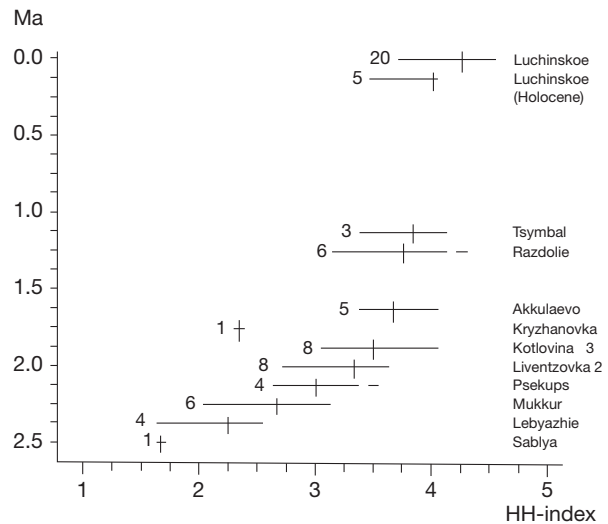


Figure 2 Increase in crown height of lower M1 in the bank vole lineage (*Clethrionomys*) in Eastern Europe and Western Siberia. Redrawn from Tesakov A (1996) Evolution of bank voles (*Clethrionomys*, Arvicolinae) in the late Pliocene and early Pleistocene of eastern Europe. *Acta Zoologica Cracoviensis* 39: 541–547. The variable plotted, HH-index, is the square root of the sum of squares of two posterior dentine tract heights. Mean, range, and number of specimens shown for each sample.

Sagebrush Vole (*Lemmys curtatus*)

Most examples of mammalian evolution in the Quaternary, such as the bank vole example just described, are pieced together from dated samples from a series of different localities.

A long series of samples from a single stratigraphic sequence is rarer and particularly valuable because the relative age of the samples is not in doubt, and the location is constant, allowing clearer identification of trends driven by environmental change. The deposits in Porcupine Cave, Colorado, provide just such a sequence and have been studied by Barnosky and Bell (2003) and Barnosky (2004). The sagebrush vole (*Lemmys curtatus*) was traced through ten stratigraphic levels covering several climatic cycles between about 1.0 and 0.6 Ma (Barnosky and Bell, 2003) and compared to living populations of the species. The average number of discrete enamel loops in the molars increased through time (Figure 3), the mode shifting from four to six, the last 4-loop individuals known as recently as about 15–9 ka, so the morphological change accumulated across several glacial–interglacial cycles.

Water Voles (*Mimomys/Arvicola*)

An intensely studied lineage is that of *Mimomys* and its descendent, the living water vole *Arvicola*. A trend of increase in crown height is reflected by the height and form of the 'linea sinuosa' (the lateral enamel–dentine junction) and can be traced (Figure 4) although *M. davakosi*, *M. hassiacus*, *M. polonicus*, *M. pliocaenicus*, and '*M. ostramosensis*' (possibly synonymous with *M. pliocaenicus*) (Chaline and Laurin, 1986; Neraudeau et al., 1995; revised taxonomy given by Maul (1996) and personal communication). The ontogenetic basis for the crown height increase is a progressively delayed formation of roots (von Koenigswald and van Kolfschoten, 1996). As in the

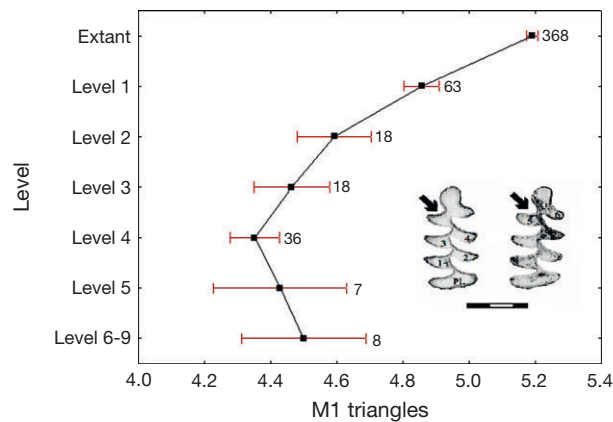


Figure 3 Trend in average number of enamel loops (triangles) in first lower molar of *Lemmiscus* through the Porcupine Cave Pit Sequence, Colorado. Mean $\pm$ SE and sample size is shown from layer 9 (about 1 Ma) to layer 1 (about 0.6 Ma) and for modern *L. curtatus* from across its range (U.S. Great Basin and adjacent areas). Data from Barnosky and Bell (2003, personal communication). Numbers of triangles quantified as follows: 4 triangles = 4, 5 pinched = 4.33, 5 incipient = 4.67; 5 triangles = 5, 6 pinched = 5.33, 6 incipient = 5.67; 6 triangles = 6. Inset: left, the 4-triangle morphology; what would be the 5th triangle is widely confluent (arrow) with the anterior part of the tooth; right, the 5-triangle morphology; the enamel bands touch such that triangle 5 is separated from the front part of the tooth. Reproduced from Barnosky AD and Bell CJ (2003) Evolution, climate change and species boundaries: perspectives from tracing *Lemmiscus curtatus* populations through time and space. *Proceedings of the Royal Society of London Series B* 270: 2585–2590.

bank vole example, it is unclear whether this trend proceeded in a purely anagenetic mode or included speciation. In the replacement of *M. pliocaenicus/ostramosensis* by the next species in Europe, *M. savini*, it has been suggested that rather than a direct lineal transition, *M. savini* was an immigrant, having evolved from *Cromeromys* in Siberia (Neraudeau et al., 1995), although *Cromeromys* is rather poorly defined. In the final transition of the sequence, from *M. savini* to *Arvicola terrestris cantiana*, roots cease to form at all, so the tooth can grow continuously throughout life as it wears (Figure 4). In this respect, it differs from the *Clethrionomys* lineage, where hypsodonty increase ceased but roots continued to form.

The subsequent *A. terrestris* lineage demonstrates a further evolutionary trend, but in a different character (Figure 5). The sequence has been investigated by Heinrich (1987), van Kolfschoten (1992) and others, sampled across more than 20 horizons in northwest continental Europe between about 400 and 10 ka BP. The molars of these voles comprise a series of enamel loops, each loop formed of a concave 'leading' edge and a convex 'trailing' edge. There was a reduction in the thickness of the trailing edge until it became thinner than the leading edge, quantified by a ratio between the two (enamel quotient or SDQ) (Figure 5). A very similar trend is seen in samples from MIS 13, 11, 9, 7, and 5e from England (Parfitt, 1998, who also discusses other character changes in this lineage).

This approximately gradualistic trend of reduction in SDQ in central and northern European voles from MIS 11 to 6 was interrupted by the appearance of more 'primitive' (high-index) animals in the Eemian (MIS 5e), which then resumed the trend

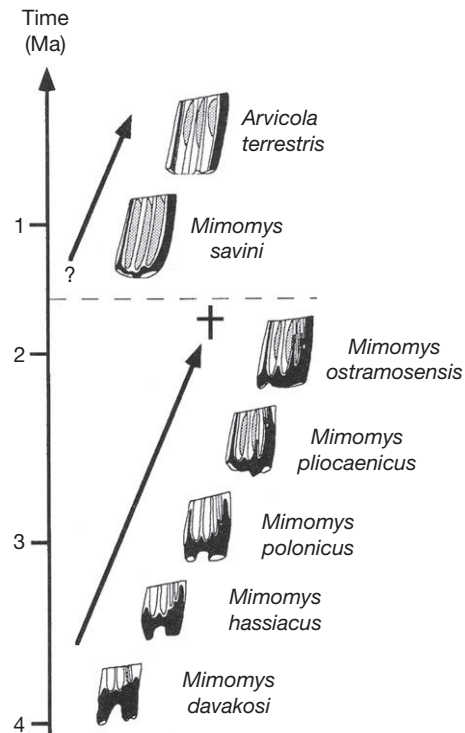


Figure 4 Increase in crown height and change in the form of linea sinuosa in molars of the European *Mimomys-Arvicola* lineage. Modified from von Koenigswald W and van Kolfschoten T (1996) The *Mimomys-Arvicola* boundary and the enamel thickness quotient (SDQ) of *Arvicola* as stratigraphic markers in the middle Pleistocene. In: Turner C (ed.) *The Early middle Pleistocene in Europe*, pp. 211–226. Rotterdam: Balkema. Black, dentine; white, enamel; stippled, cement; linea sinuosa is the boundary between dentine and enamel. Roots are finally lost between *M. savini* and *A. terrestris*. The diagram also illustrates the theory that the lineage was interrupted by the extinction of *M. ostramosensis* and immigration of *M. savini*.

toward a reduced index until the late Weichselian (MIS 4–2; Figure 5). It is likely that the late Saalian (MIS 6) ice advance led to local extinction of *A. terrestris* in northwest Europe, followed in the Eemian by re-immigration of a more primitive form (van Kolfschoten, 1992). A similar pattern of evolutionary reversal is seen in voles from Central Europe (Hungary: Janossy, 1976), but the apparent inversion in southern France and northwestern Italy shown by Desclaux et al. (2000) was not confirmed by larger samples (Escudé et al., 2008), nor is the reversal seen in Eastern Europe (van Kolfschoten, personal communication). Röttger (1987) showed that even today there is a cline in SDQ of water voles across Europe, with those in the south and east showing a higher (more 'primitive') index than those in the north and west. If this pertained in the past, it would suggest repopulation of northern areas from the south or east in the Eemian.

The detection of a morphological reversal in this sequence is a reflection of its relatively high stratigraphic resolution. Such reversals quite likely occurred within longer, apparently unidirectional, sequences of other taxa, but are not perceived because of coarser sampling. In addition, this very rapid event was not a case of rapid evolution but of immigration. Many such apparently instantaneous morphological jumps seen in

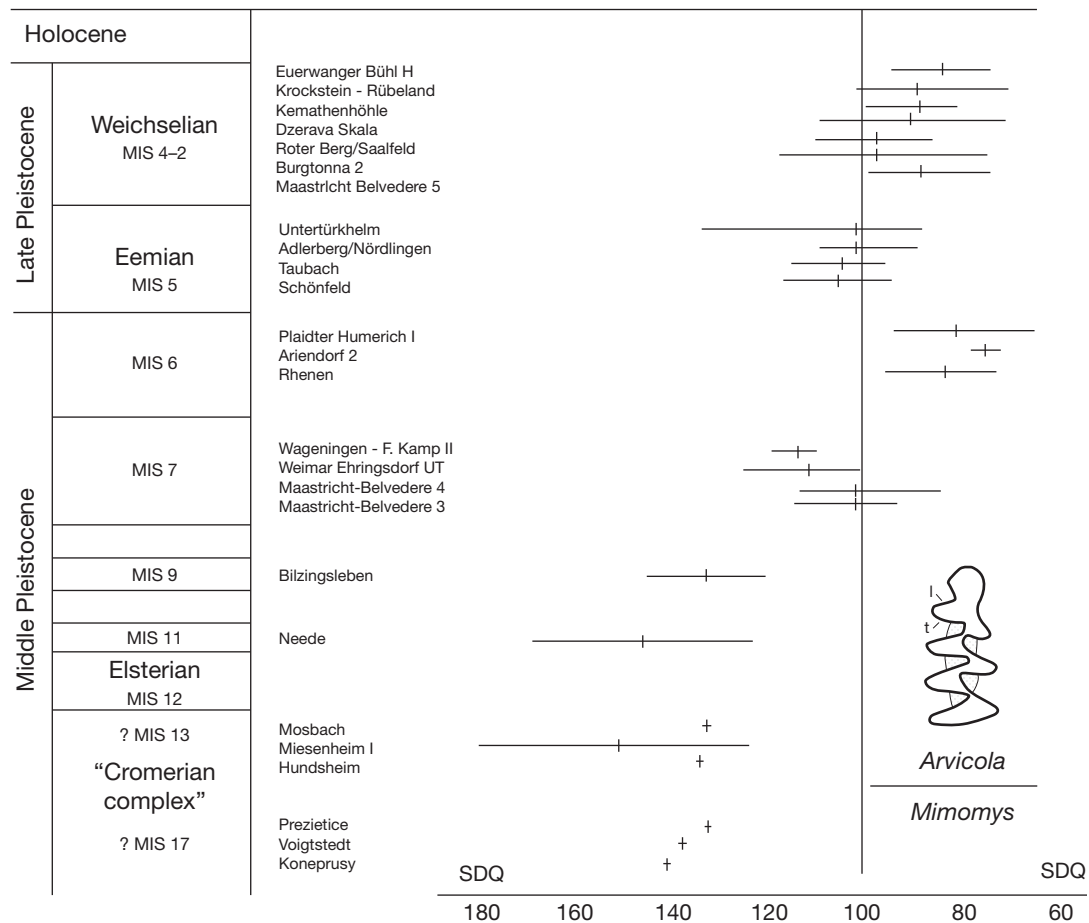


Figure 5 Trend in enamel differentiation index (SDQ) of molars (lower M1 plotted) in the middle to late Pleistocene *Arvicola terrestris* lineage from Western Europe. SDQ is the ratio of enamel widths between the leading (l) and trailing (t) edges of each loop, averaged across each loop in the tooth. Stratigraphy from van Kolfschoten (1992, personal communication). Modified from von Koenigswald W, van Kolfschoten T (1996) The *Mimomys*-*Arvicola* boundary and the enamel thickness quotient (SDQ) of *Arvicola* as stratigraphic markers in the middle Pleistocene. In: Turner C (ed.) The Early middle Pleistocene in Europe, pp. 211–226. Rotterdam: Balkema.

the fossil record of a given region may be the result of migration events of already differentiated populations. The genuine evolutionary changes, which earlier produced that differentiation, could have occurred more slowly.

Escudé et al. (2008) showed a significant reduction in SDQ through the late Middle and late Pleistocene of France, with strongly overlapping ranges between successive samples, consistent with a purely anagenetic mode. A morphometric analysis of overall occlusal outline of the molar, however, failed to show any directional trend, emphasizing that a trend in one character may be accompanied by stasis or fluctuation in another.

Mammoths (*Mammuthus* spp.)

Arguably, the most detailed account of an evolutionary trend in a Quaternary large mammal lineage is provided by the Eurasian representatives of the mammoth genus, *Mammuthus*. A series of fossils from Europe spanning about 3 Myr demonstrates a broadly directional change in molar morphology, with an increase in the number of enamel 'lamellae,' thinning of the individual enamel bands, and heightening of the tooth crown, with concomitant changes in the skull. The transitions are

seen across four nominal chronospecies: *M. rumanus* (broadly late Pliocene), *M. meridionalis* (broadly early Pleistocene), *M. trogontherii* (broadly middle Pleistocene), and *M. primigenius* (broadly late Pleistocene). Viewed coarsely over the whole time-span, the sequence (Figure 6, open histograms) appears approximately 'gradualistic.' On closer examination, however, the pattern resolves into three intervals of stasis (no statistically significant change; colored vertical lines in Figure 6), corresponding broadly to the nominal species, with relatively rapid periods of transition in between.

When samples from other regions are added, a more complex picture emerges. *Mammuthus trogontherii*, appearing in western and central Europe around 800–700 ka, is found in NE China around 1.7 Ma (Wei et al., 2003; green star in Figure 6) and in NE Siberia around 1.2–1.0 Ma (Lister and Sher, 2001; black histogram in Figure 6). The transition from *M. meridionalis* to *M. trogontherii*, with substantial advancement of molar and skull morphology, therefore, appears to have taken place in the early Pleistocene of eastern Asia, probably in China, in response to a continental climate and partly steppic habitat (Wei et al., 2003; Lister et al., 2005). Only much later did it spread into Europe. This hypothesis of allopatric

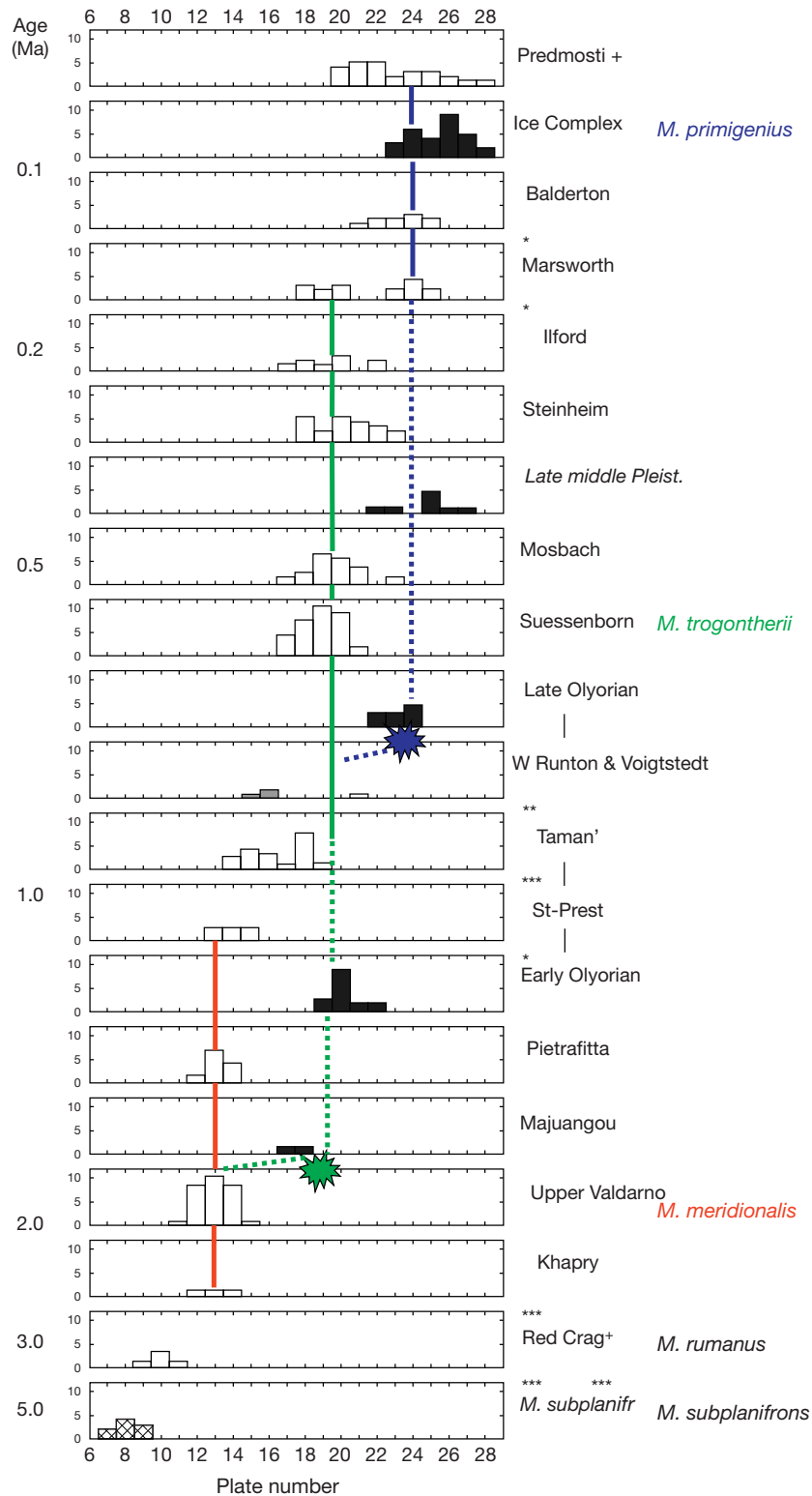


Figure 6 Evolution of the Eurasian mammoth lineage. Modified from Lister AM and Sher AV (2001) The origin and evolution of the woolly mammoth. *Science* 294: 1094–1097. The graph shows the number of enamel plates (lamellae) in upper and lower M3; species identifications are also based on crown height (Lister and Sher, 2001, Figure 3B). Cross-hatch: *Mammuthus subplanifrons*, Africa; white: European samples, except gray, Voigtstedt; black, NE Siberia, plus Majuango, China (earliest *M. trogontherii*). Red: *M. meridionalis*; green, *M. trogontherii*; blue, *M. primigenius*. Dashed lines, Asian sites; solid lines, European sites. Site names linked by vertical lines are of approximately the same age. Site names separated by asterisks are significantly different in mean plate number.

speciation is corroborated by the finding, in Europe, of a period of overlap between indigenous *M. meridionalis* and apparently immigrant *M. trogontherii* in the interval 1.0–0.8 Ma, most notably at the Sinyaya Balka locality in SE Europe (Lister and Sher, 2001; Lister et al., 2005).

Further specialization to produce the woolly mammoth, *M. primigenius*, can be traced through a series of samples from NE Siberia, indicating that the genesis of this cold-adapted, grazing species took place in the Beringian region (Lister and Sher, 2001). The transitions from *M. trogontherii* to early *M. primigenius* in the interval 1.2–0.7 Ma (blue star in Figure 6), and thence to late *M. primigenius* by 0.4 Ma, each show an increase in grazing adaptation of the molars. Northern Siberia as the locus of origin of the woolly mammoth, presumably from semi-isolated populations of *M. trogontherii*, makes sense in terms of strong natural selection in this severe environmental regime (Sher et al., 2005). *M. primigenius* later dispersed into Europe and North America; its time of arrival is uncertain but is currently not definitively demonstrated until MIS 6 (about 180–140 ka).

In this way, the ‘broadly gradualistic’ mammoth sequence in Europe, formerly perceived as *in situ* evolution, can be resolved into a series of immigrations of successively more ‘advanced’ forms, those forms having originated, by cladogenesis, elsewhere. The processes of speciation and directional evolution (trends) are therefore intimately linked. The replacement of each species by its successor, not by anagenesis but by immigration and presumably competition, can even be viewed as a small-scale example of a trend driven by ‘species selection.’

It should be noted that even though the fossil record has pinpointed Early to middle Pleistocene China and northeast Siberia as likely areas of speciation, the duration of the cladogenetic events is known only within broad limits.

Microevolution

The above examples of directional trends (even with occasional reversals) must be viewed in the context of many other cases where there is little change through time (stasis) or, most commonly, fluctuation about a mean but with no net directional trend. The latter may in theory be seen in any character, but it is most commonly encountered in measures of body size or, at least, in the size of a measured dental or skeletal element (Lister, 1992; Blois and Hadly, 2009). van der Made (2010), for example, graphs changes in the size of a dental element in several large mammal species through the European Quaternary. Wolf (*Canis*) shows apparent directional change through the middle Pleistocene, with approximate stasis before and after, but both red deer (*C. elaphus*) and fallow deer (*Dama* spp.) show fluctuating patterns with little or no net trend. In a study of the marmot *Marmota* sp. at Porcupine Cave – from the same stratigraphic sequence demonstrating an evolutionary trend in sagebrush vole (see above) – no net change was seen in either nonmetric or metric traits (Barnosky et al., 2004). This emphasizes the individualistic nature of species responses.

Small-scale or reversible body size changes may often be ecophenotypic (a direct effect of the environment on development). In other cases, they may be genetically based (‘evolutionary’ in a strict sense), but these can be difficult to separate

paleontologically (Lister, 1997, 2004; Blois and Hadly, 2009). In terms of adaptation, size changes have often been regarded as responses to temperature increase, according to the theory that body size responds inversely to ambient temperature as an adaptation for heat regulation by altering the surface area to volume ratio (e.g., Davis, 1981). This model, inspired by ‘Bergmann’s rule’ of alleged increased size with latitude within modern species, has been widely applied to body size differences in Quaternary mammals. Kurtén (1968) found oscillation in body size of several mammalian species, large forms occurring in cold stages smaller ones in interglacials, and attributed the changes to ‘Bergmannian’ effects.

Others, however, have contested this interpretation, as body size is subject to many selective forces in addition to temperature regulation, and in the modern fauna, many species show trends contrary to Bergmann’s rule (McNab, 1971). An example is the American otter, *Lutra canadensis*, whose reduction in size toward the north is believed to be because holes in the ice become smaller in the arctic (Scholander, 1956). Such causal factors would be very difficult if not impossible to deduce in the fossil record. Several recent workers suggest that the duration and abundance of the plant growing season is the primary determinant of herbivore body size, which in turn selects for corresponding changes in carnivore size (McNab, 1971; Geist, 1987; Guthrie, 1990). Langvatn and Albon (1986), studying size clines in living red deer in Norway, attributed the larger size of animals toward the north to clines in the productivity and quality of plants. Under this interpretation, the Pleistocene tundra-steppe is regarded as a richer habitat for many large herbivores (and hence carnivores) than the interglacial forests (Geist, 1987; Guthrie, 1990), and the size reduction found by Davis (1981) at the Pleistocene/Holocene transition in the Middle East could have resulted from the vegetational response to increased aridity.

Such fluctuations must, in any event, be regarded as microevolutionary, in that they are short-term responses to local environmental conditions and do not lead to long-term trends, still less to speciation.

Biostratigraphic Value of Quaternary Mammal Evolution

Evolutionary trends are potentially of biostratigraphic value. In formal biostratigraphic zonation, they form the basis for the erection of ‘lineage zones,’ biostratigraphic zones based on a segment of the morphological trend in an evolving lineage (Hedberg, 1976). An ‘ideal’ evolutionary change from a biostratigraphic point of view would be one that is (a) easily observable (and preferably quantifiable) on the fossil remains, (b) genetically based, and (c) conforms to one of the following two patterns:

- a character that changes gradually, unidirectionally, and synchronously over the geographic range of the species;
- a new character that appears suddenly at a particular time, synchronously over the geographic range of the species.

However, studies of evolving lineages indicate that the rate of change is often very far from constant. There may be long periods of stasis, when little or no change occurs. At other times, gradual transitions may take place over considerable

timespans, but with a fluctuating rate of change. Finally, in certain circumstances, relatively very rapid changes may occur where the transition would be barely if at all resolvable in the geological record (Levinton, 1988). Two key consequences of this for biostratigraphy are as follows:

- (a) Morphological similarity between two samples, while consistent with their being the same age, does not guarantee it. They may be temporally separated by a period of evolutionary fluctuation or stasis. This is particularly likely when only a single character is considered. If many characters are examined, covering different aspects of the animal's anatomy, it becomes less likely that nothing at all will have changed over a significant period of time. Thus Rackham (1981) examined *Bison priscus* fossils from a series of British Devensian (Last Cold Stage) sites and found no significant differences between the samples on an array of biometric and morphological criteria. He therefore suggested that the samples were penecontemporaneous. Conversely, Lister et al. (2010) found significant differences in measurements of deer species across a series of early middle Pleistocene localities within a limited geographical area (southern England). They concluded that even if some of the samples date to the same interglacial stage, they could not be precisely contemporaneous.
- (b) The degree of morphological difference among a series of samples will be directly proportional to the time intervals between them only if the evolution was gradual and at a constant rate. In other cases, a sample of intermediate age between samples A and B may either still show morphology A (stasis) or already have arrived at morphology B (rapid change followed by stasis). The prediction of age based on morphology, assuming a constant rate of morphological change, is therefore not reliable unless there is clear independent evidence that this was the pattern of change. How frequently very rapid changes occur in practice is uncertain, and there must be an upper limit to the amount of change that can occur in a given period of time, which should become clearer as more independently calibrated cases are examined.

As well as lineage trends, the other major category of biostratigraphic data derived from species evolution is the first appearance datum (FAD) of a species in the fossil record, globally or in a given region. However, the distinction between these processes may be blurred in practice. Some of the 'first appearances' that are marked with a new species or even genus name are the result of a genuine speciation process in the sense of the splitting off of a new, reproductively isolated species from the parent (cladogenesis). In other cases, however, they may represent processes within a reproductively continuous lineage (anagenesis).

In either case, there is a further dichotomy: the apparent time of replacement of one form by another in a particular area may represent either the actual origination event of the new form, or else the time of its migration into that area, after origination elsewhere. The latter will be more common, unless there is sufficient geographical coverage to suggest that sampling is being done in the area of allopatric origin (as in the mammoth example, above).

The time of appearance of a species in a particular area may therefore have value as a local or regional biostratigraphic marker, but because of the potentially time-transgressive nature of its spread, FADs are a less secure basis for correlation the greater the geographic distance or the finer the time scale. Moreover, even in a single area, one cannot argue from a significant morphological difference between a species in horizon A, and its replacement in horizon B, that there must be a significant time interval between A and B. If the replacement was due to immigration rather than *in situ* evolution, it can have been effectively instantaneous.

In sum, mammalian evolution has potentially great value in Quaternary biostratigraphy, but because of the multiple potential modes of evolution, the pattern of change has to be established, for each lineage, in independently dated samples, preferably across the geographical range, before it can be confidently applied in biostratigraphy. It is, finally, important that morphological differences among samples be compared statistically, using adequate samples (Lister, 1992; Escudé et al., 2008).

Discussion

Rates of evolution observed in the Quaternary can be viewed in the context of both longer-term processes in the deep fossil record, and present and future observed or potential rates of adaptation. There is an apparent discrepancy between the extremely fast rates of observed historical transitions and laboratory experiments on the one hand, the generally slower Quaternary evolution on the other, and the even slower rates observed over longer geological timescales. Gingerich (1983, 1993) has gone some way to resolving this paradox by analyzing evolutionary rates measured over different timescales. He showed that fast rates are maintained for only a short time, and the longer the observed timespan, the lower the measured rate. Long evolutionary sequences will have included short episodes of change as fast as the modern examples, but over time these were averaged out with periods of slower change or stasis. This effect is enhanced by the reversals of trend, seen in Quaternary examples, which act as 'negative rate' and slow the net progress. Moreover, many of the fast transitions observed in short fossil or historical sequences are probably not evolutionary at all, but result from ecophenotypic or migrational effects.

In general, reviewing Quaternary mammal evolution, there is no clear evidence to support the idea that small mammals evolve faster than large ones (Liow et al., 2008). In terms of cladogenesis, it is clear that small mammals have produced greater numbers of species, for rather obvious demographic and ecological reasons. But in terms of rates of morphological change, there is no very clear difference between the dental sequences of the water vole and the mammoth, for example. This is perhaps unsurprising because neither of these lineages was evolving at anything near the maximal rate implied by laboratory selection experiments. Only closer to this limit might factors such as the faster generation time of small mammals have an effect. In the field, voles and mammoths took the order of 10^5 – 10^8 years to complete significant adaptive morphological changes, and the slowness of these net rates must

have been imposed by other factors such as environmental conditions, developmental constraints, and coadaptation of the animal's total phenotype.

It is not yet possible to confidently assess the relative contribution of gradualistic/anagenetic and punctuated/cladogenetic modes to evolutionary change. In part, this is due to biases in the fossil record or its perception. Biases in favor of 'seeing' punctuations include stratigraphic unconformities and mistaking rapid immigration for rapid origin. Biases in favor of 'seeing' gradualism include the greater ease with which an evolutionary lineage can be recognized the more gradational its links; and the selection for study of sequences showing clear directional trends (Gould and Eldredge, 1977).

The mammoth and water vole lineages, for example, were selected for detailed research because, as well as having a well-sampled sequence, each shows significant directional evolution. Moreover, the characters highlighted here (mammoth plate count and hypsodonty, and vole root formation and enamel differentiation) are those that show clear morphological trends. On the other hand, it would be wrong to ignore the many Quaternary mammalian taxa that 'appear' in the record without clear ancestry or a sequence of chronospecies behind them, such as the reindeer *Rangifer tarandus* (L). Whether this reflects punctuated equilibrium or simply one's ignorance (evolution – at an unknown rate – in an unsampled area) is not known. Also, in many cases, one cannot be sure whether a case of directional evolution, interpreted as an anagenetic trend, included complex patterns of population replacement or even directional sorting among the array of variously modified species (species selection). The mammoth example shows how complex a real process can be, incorporating elements of different evolutionary modes. It also shows that the perception of rate or of 'anagenetic/cladogenetic' mode is dependent on the time-scale over which it is viewed.

The examples quoted here cannot, therefore, yet be used to 'head count' the relative frequency of different modes or to assess whether Quaternary rates of evolution were faster than those of earlier periods (Lister, 2004). They do, nonetheless, provide valuable examples of evolution producing significant modification and demonstrating a variety of modes. The Quaternary is the unique interface where fruitful linkage can be forged between the deeper paleontological record, with its limits to resolution, and the neontological study of population genetics, with its limited time depth. The study of ancient DNA provides a further highly promising avenue of research in this respect (Ramakrishnan and Hadly, 2009).

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See also: Vertebrate Overview. **Vertebrate Studies:** Ancient DNA; Dwarfing and Gigantism in Quaternary Vertebrates; Interactions with Hominids.

References

- Avise JC and Walker D (1998) Pleistocene phylogeographic effects on avian populations and the speciation process. *Proceedings of the Royal Society of London Series B* 265: 457–463.
- Barnes I, Matheus P, Shapiro B, Jensen D, and Cooper A (2002) Dynamics of Pleistocene population extinctions in Beringian brown bears. *Science* 295: 2267–2270.
- Barnosky AD (ed.) (2004) *Biodiversity Response to Climate Change in the Middle Pleistocene: The Porcupine Cave Fauna from Colorado*. Berkeley, CA: University of California Press.
- Barnosky AD and Bell CJ (2003) Evolution, climate change and species boundaries: Perspectives from tracing *Lemmus curtatus* populations through time and space. *Proceedings of the Royal Society of London Series B* 270: 2585–2590.
- Barnosky AD, Kaplan MH, and Carrasco MA (2004) Assessing the effect of Middle Pleistocene climate change on *Marmota* populations from the Pit locality. In: Barnosky AD (ed.) *Biodiversity Response to Climate Change in the Middle Pleistocene: The Porcupine Cave Fauna from Colorado*, pp. 332–340. Berkeley, CA: University of California Press.
- Blois J and Hadly E (2009) Mammalian response to Cenozoic climatic change. *Annual Review of Earth and Planetary Sciences* 37: 181–208.
- Chaline J and Laurin B (1986) Phyletic gradualism in a European Plio-Pleistocene *Miomys* lineage (Arvicolidae, Rodentia). *Paleobiology* 12: 203–216.
- Davis SJM (1981) The effects of temperature change and domestication on the body size of Late Pleistocene to Holocene mammals of Israel. *Paleobiology* 7: 101–114.
- Desclaux E, Abbassi M, Marquet J-C, Chaline J, and van Kolfschoten T (2000) Distribution and evolution of *Arvicola* Lacépède, 1799 (Mammalia, Rodentia) in France and Liguria (Italy) during the Middle and the Upper Pleistocene. *Acta Zoologica Cracoviensis* 43: 107–125.
- Edwards CJ, Suchard MA, Lemey P, et al. (2011) Ancient hybridisation and an Irish origin for the modern polar bear matriline. *Current Biology* 21: 1–8.
- Elias SA, Short SK, Nelson CH, and Birks HH (1996) Life and times of the Bering land bridge. *Nature* 382: 60–63.
- Escudé E, Montuire S, Desclaux E, Quéré J-P, Renvoisé E, and Jeannet M (2008) Reappraisal of 'chronospecies' and the use of *Arvicola* (Rodentia, Mammalia) for biochronology. *Journal of Archaeological Science* 35: 1867–1879.
- Fedorov VB, Goropashnaya AV, Jaarola M, and Cook A (2003) Phylogeography of lemmings (*Lemmus*): No evidence for postglacial colonization of Arctic from the Beringian refugium. *Molecular Ecology* 12: 725–731.
- García NG (2003) *Osos y otros carnívoros de la Sierra de Atapuerca*, Proaza: Fundación Osos de Asturias.
- Geist V (1987) Bergmann's rule is invalid. *Canadian Journal of Zoology* 65: 1035–1038.
- Gingerich PD (1983) Rates of evolution: Effects of time and temporal scaling. *Science* 222: 159–161.
- Gingerich PD (1993) Quantification and comparison of evolutionary rates. *American Journal of Science* 293A: 453–478.
- Gould SJ and Eldredge N (1977) Punctuated equilibria: The tempo and mode of evolution revisited. *Paleobiology* 3: 115–151.
- Guthrie RD (1990) *Frozen Fauna of the Mammoth Steppe: The Story of Blue Babe*. Chicago: University of Chicago Press.
- Hailer F, Kutschera VE, Hallstrom BM, et al. (2012) Nuclear genomic sequences reveal that polar bears are an old and distinct bear lineage. *Science* 336: 344–347.
- Hedberg HD (ed.) (1976) *International Stratigraphic Guide*. New York: Wiley.
- Heinrich W-D (1987) Neue Ergebnisse zur Evolution und Biostratigraphie von *Arvicola* (Rodentia, Mammalia) im Quartär Europas. *Zeitschrift für Geologische Wissenschaften* 15: 389–406.
- Hoffmann RS (1981) Different voles for different holes: Environmental restrictions on refugial survival of mammals. In: Scudder GGE and Reveal JL (eds.) *Evolution Today: Proceedings of the Second International Congress of Systematic and Evolutionary Biology*, pp. 25–45. Pittsburgh: Hunt Institute for Botanical Documentation, Carnegie-Mellon University.
- Janossy D (1976) The influence of the glaciations on the microvertebrate fauna in the periglacial area in Europe. *GFF (Journal of the Geological Society of Sweden)* 98: 291–296.
- Jarrell GH and Fredga K (1993) How many kinds of lemming? A taxonomic overview. In: Stenseth NC and Ims RA (eds.) *The Biology of Lemmings*, pp. 45–57. London: Academic.
- Jiggins CD and Mallet J (2000) Bimodal hybrid zones and speciation. *Trends in Ecology & Evolution* 15: 250–255.
- Johnson WE and O'Brien SJ (1997) Phylogenetic reconstruction of the Felidae using 16S rRNA and NADH-5 mitochondrial genes. *Journal of Molecular Evolution* 44(supplement 1): S098–S116.

- Knapp M, Rohland N, Weinstock J, et al. (2009) First DNA sequences from Asian cave bear fossils reveal deep divergences and complex phylogeographic patterns. *Molecular Ecology* 18: 1225–1238.
- von Koenigswald W and van Kolfschoten T (1996) The *Mimomys–Arvicola* boundary and the enamel thickness quotient (SDQ) of *Arvicola* as stratigraphic markers in the Middle Pleistocene. In: Turner C (ed.) *The Early Middle Pleistocene in Europe*, pp. 211–226. Rotterdam: Balkema.
- Kurtén B (1964) The evolution of the polar bear. *Ursus maritimus* Phipps. *Acta Zoologica Fennica* 108: 1–30.
- Kurtén B (1968) *Pleistocene Mammals of Europe*. London: Weidenfeld and Nicholson.
- Kurtén B and Anderson E (1980) *Pleistocene Mammals of North America*. New York: Columbia University Press.
- Langvatn R and Albon SD (1986) Geographic clines in body weight of Norwegian red deer: A novel explanation of Bergmann's rule? *Holarctic Ecology* 9: 285–293.
- Levinton J (1988) *Genetics, Paleontology and Macroevolution*. Cambridge: Cambridge University Press.
- Lindqvist C, Schuster SC, Sun Y, et al. (2010) Complete mitochondrial genome of a Pleistocene jawbone unveils the origin of polar bear. *PNAS* 107: 5053–5057.
- Liow LH, Fortelius M, Bingham E, et al. (2008) Higher origination and extinction rates in larger mammals. *PNAS* 105: 6097–6102.
- Lister AM (1992) Mammalian fossils and Quaternary biostratigraphy. *Quaternary Science Reviews* 11: 329–344.
- Lister AM (1997) The evolutionary response of vertebrates to Quaternary environmental change. *North Atlantic Treaty Organization Advanced Science Institutes Series I* 47: 287–302.
- Lister AM (2004) The impact of Quaternary ice ages on mammalian evolution. *Philosophical Transactions of the Royal Society of London B* 359: 221–241.
- Lister AM, Parfitt SA, Owen FJ, Collinge SA, and Breda M (2010) Metric analysis of ungulate mammals in the early Middle Pleistocene of Britain, in relation to taxonomy and biostratigraphy. II. Cervidae, Equidae and Suidae. *Quaternary International* 228: 157–179.
- Lister A and Rawson P (2003) Land-sea relations and speciation in the marine and terrestrial realms. In: Rothschild L and Lister A (eds.) *Evolution on Planet Earth: The Impact of the Physical Environment*, pp. 297–315. London: Academic Press.
- Lister AM and Sher AV (2001) The origin and evolution of the woolly mammoth. *Science* 294: 1094–1097.
- Lister AM, Sher AV, van Essen H, and Wei G (2005) The pattern and process of mammoth evolution in Eurasia. *Quaternary International* 126–128: 49–64.
- Martin LD (1984) Phyletic trends and evolutionary rates. In: Genoways HH and Dawson MR (eds.) *Contributions in Quaternary Vertebrate Paleontology: A Volume in Memorial to John E. Guilday*, pp. 526–538. Museum of Natural History Special Publication 8, Pittsburg.
- Martin RA (1993) Patterns of variation and speciation in Quaternary rodents. In: Martin RA and Barnosky AD (eds.) *Morphological Change in Quaternary Mammals of North America*, pp. 226–280. Cambridge: Cambridge University Press.
- Maul L (1996) Biochronological implications of the arvicolid (Mammalia: Rodentia) from the Plio- and Pleistocene faunas of Neuleiningen (Rheinland-Pfalz, SW-Germany). *Acta Zoologica Cracoviensis* 39: 349–356.
- Maul L, Masini F, Abbazzi L, and Turner A (1998) The use of different morphometric data for absolute age calibration of some South- and Middle European arvicolid populations. *Paleontographica Italica* 85: 111–151.
- Mayr E (1963) *Animal Species and Evolution*. Cambridge, MA: Belknap Press.
- Mazza P and Rustioni M (1994) On the phylogeny of Eurasian bears. *Palaeontographica Abteilung A* 230(1–3): 1–38.
- McNab BK (1971) On the ecological significance of Bergmann's rule. *Ecology* 52: 845–854.
- Neraudeau D, Viriot L, Chaline J, Laurin B, and van Kolfschoten T (1995) Discontinuity in the Plio-Pleistocene Eurasian water vole lineage. *Palaeontology* 38: 77–85.
- Parfitt S (1998) The interglacial mammalian fauna from Barnham. In: Ashton NA, Lewis SG, and Parfitt S (eds.) *Excavations at the Lower Palaeolithic site at East Farm, Barnham, Suffolk 1989–94*, pp. 111–147. London: British Museum Press.
- Polly PD (2001) On morphological clocks and paleophylogeography: Towards a timescale for *Sorex* hybrid zones. *Genetica* 112: 339–357.
- Rackham DJ (1981) *Mid-Devensian Mammals in Britain*. Unpublished MSc Thesis, University of Birmingham.
- Ramakrishnan U and Hadly EA (2009) Using phylochronology to reveal cryptic population histories: Review and synthesis of 29 ancient DNA studies. *Molecular Ecology* 18: 1310–1330.
- Ritz MS, Millar C, Miller GD, et al. (2008) Phylogeography of the southern skua complex – Rapid colonisation of the southern hemisphere during a glacial period and reticulate evolution. *Molecular Phylogenetics and Evolution* 49: 292–303.
- Röttger U (1987) Schmelzbandbreiten an Molaren von Schermäusen (*Arvicola Lacépède*, 1799). *Bonner Zoologische Beiträge* 38: 95–105.
- Scholander PF (1956) Climatic rules. *Evolution* 10: 339–340.
- Sher AV, Kuzmina SA, Kuznetsova TV, and Sulerzhitsky LD (2005) New insights into the Weichselian environment and climate of the East Siberian Arctic, derived from fossil insects, plants, and mammals. *Quaternary Science Reviews* 24: 533–569.
- Sher AV, Weinstock J, Baryshnikov GF, et al. (2011) The first record of “spelaeoid” bears in Arctic Siberia. *Quaternary Science Reviews* 30: 2238–2249.
- Shoshani J, Ferretti MP, Lister AM, et al. (2007) Relationships within the Elephantinae using hyoid characters. *Quaternary International* 169–170: 174–185.
- Stewart JR, Lister AM, Barnes I, and Dalén L (2009) Refugia revisited: Individualistic responses of species in space and time. *Proceedings of the Royal Society of London Series B* 277: 661–671.
- Talbot SL and Shields GF (1996) Phylogeography of brown bears (*Ursus arctos*) of Alaska and parapatry within the Ursidae. *Molecular Phylogenetics and Evolution* 5: 477–494.
- Tesakov A (1996) Evolution of bank voles (*Clethrionomys*, Arvicolinae) in the late Pliocene and early Pleistocene of eastern Europe. *Acta Zoologica Cracoviensis* 39: 541–547.
- van der Made J (2010) Biostratigraphy – “large mammals” In: Meller H (ed.) *Elefantenreich – Eine Fossilwelt in Europa*, pp. 83–92. Landesamt für Denkmalpflege und Archäologie Sachsen-Anhalt & Landesmuseum Vorgeschichte.
- van Kolfschoten T (1992) Aspects of the migration of mammals to northwestern Europe during the Pleistocene, in particular the reimmigration of *Arvicola terrestris*. *Courier Forschungsinstitut Senckenberg* 153: 213–220.
- Wei G, Taruno H, Jin C, and Xie F (2003) The earliest specimens of the steppe mammoth, *Mammuthus trogontherii*, from the Early Pleistocene Nihewan Formation, North China. *Earth Science (Japan)* 57: 269–278.

Dwarfing and Gigantism in Quaternary Vertebrates

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Introduction

"I do not deny that there are many and grave difficulties in understanding how several of the inhabitants of the more remote islands, whether still retaining the same specific form or modified since their arrival, could have reached their present homes." (Charles Darwin, 1859: p. 396)

Islands have been regarded by scientists as a living laboratory of evolution, and a prime target for the study of forces influencing evolution and diversification. In response to the special characteristics of insular environments, plants and animals often undergo evolutionary changes which can be observed on islands of varying surface area. These include primary productivity and available resources, diversity, and intensity of ecological interactions, and time and magnitude of their isolation. Two main issues have been attentively scrutinized and debated: the loss of biodiversity of insular faunas and the peculiar changes undergone by island settlers, above all changes in body size shown by endemic reptiles, birds, and mammals. This pattern was first described in insular mammals by Foster (1964) as a tendency for rodents (and possibly marsupials) to increase and artiodactyls, carnivores, and possibly lagomorphs (rabbits and hares) to decrease in body size on islands, with insectivores supposedly showing no consistent trend. Afterward, this pattern was considered as a rule by Van Valen (1973), who postulated gigantism in smaller and dwarfism in larger species of insular mammals. Although the rule has a number of exceptions and departures from the predictions – which can be found among both extinct and extant endemic species of differing body size, trophic habits, and phylogenetic affinities – the 'island rule' continues to be a key topic and hotly debated issue in island biogeography.

Over time, a number of hypotheses have been formulated to explain causal mechanisms behind the graded body size change of insular vertebrates, and even recent overviews have presented conflicting points of view, confirming or rejecting the soundness of the 'rule.' Some authors maintain that it is a general principle (e.g., Benton et al., 2010; Köhler et al., 2008; Lomolino et al., 2006; Palombo, 2009; Welch, 2009 and references in those papers); others reject its generality, saying the 'rule' would be an artifact of comparing distantly related groups, showing clade-specific responses to insularity (Meiri et al., 2008). The role played by different factors in driving changes in body size has been interpreted differently by various authors, who mainly focused at times on the reduced (often absent) predation pressure exerted by large carnivores, host-island surface, limitation in trophic resources, population density and overcrowding, genetic segregation, and endogamy, and so on. Body size evolution of insular vertebrates, especially mammals, actually fluctuated in accordance with the synergistic influences of selective biotic and abiotic forces (geographic, climatic, and ecological characteristics of the islands,

biology of the focal species, structure of insular communities, species interaction, ecological displacement and release), whose relative importance and nature could vary on spatial and temporal scales, giving rise to a different body size evolution, even in species originating from the same ancestor.

Giants and Dwarfs: The Most Intriguing Case Histories

Miniaturized elephants and hippopotamuses, herbaceous plants becoming woody and as high as a tree, giant lizards, turtles and varans, insects and birds losing the power of flight, deer either as large as a dog or more than a moose, short-legged bovids with stereoscopic vision, little owls larger than horned owls, giant scavenger shrew-like insectivores larger than a cat, small rodents that had changed their size and life behavior: these are examples of extreme and fascinating animals whose ancestors are often difficult to know or imagine. During the Quaternary, dramatically specialized vertebrates inhabited a number of islands, from large to small, from close to significantly far from the mainland, scattered all over the world, from the Mediterranean sea to Indonesia, from the Eastern to Western Pacific Ocean, from the Caribbean to Canary islands. Their terrestrial ancestors reached these islands mainly by sweepstake dispersal, crossing either wide or narrow water barriers by swimming, rafting on floating vegetation, maybe drifting with the aid of currents, or else they crossed over on discontinuous terrestrial bridges or strips of emergent continental shelves. These mechanisms acted as a selective filter for dispersal.

Modes of colonization, pioneer species, and their evolution varied on each island as well on the same island over time. As a result, every insular fauna has its own unique structure. This depends on several interacting factors such as the type of islands (continental, oceanic, oceanic-like, either isolated or part of an archipelago), their size, physiography and climate regime, the distance separating the islands from the mainland, the nature of the barrier and its changing over time, the dispersal methods, the variety of ecological niches available on the islands, and the dispersal ability, ethology, and ecology of the nonflying mainland incoming species as well as the characteristics of the species already inhabiting the island (if any). These factors have been interwoven and subject to changes over time, sometimes leading to dramatic faunal turnovers.

Insular faunas, on the other hand, share some particular characteristics, above all the occurrence of 'dwarf' or 'giant' species in unbalanced, disharmonic, impoverished faunas, in which large carnivores, perissodactyls, and ungulates with little or no ability to swim, are rare or absent. This enables us to identify a number of 'fossil' islands, whose inhabitants constitute some of the most impressive examples of insular evolution. The preeminent cases of dwarfism among mammals (pygmy proboscideans) and gigantism (giant rodents and insectivores) are found in the fossil record. Accordingly,

paleontology offers unique insights into the long-term evolutionary processes and causal mechanisms that led to the peculiar insular ecosystems, shedding light, at the same time, on some intriguing paleogeographic issues.

Dwarf Giants: Proboscideans and Hippopotamuses

Proboscideans

Proboscideans were the most characteristic and common taxa in the Pleistocene disharmonic island faunas. Endemic stegodonts (*Sinomastodon*, *Stegoloxodon*, and *Stegodon*) were found in the Pleistocene deposits from Java, Sulawesi, Sumba, Timor, Flores, and Philippine islands. Dwarf *Mammuthus* species have been recorded in the late middle–late Pleistocene of Sardinia (*Mammuthus lamarmorai*), the early Pleistocene of Crete (*Mammuthus creticus*), and the late Pleistocene of Santa Rosa Island, California (*Mammuthus exilis*) (Figure 1). Endemic elephants that originated from southern European populations of straight-tusked *Palaeoloxodon* inhabited a number of Mediterranean islands (e.g., *Palaeoloxodon falconeri* and *Palaeoloxodon mnaidriensis* from Sicily and Malta; *Palaeoloxodon creutzburgi* from Crete, *Palaeoloxodon* spp. from Cyclades Islands, *Palaeoloxodon tiliensis* from Tilos, *Palaeoloxodon* sp. from Rhodos, *Palaeoloxodon cypriotes* and *Palaeoloxodon* sp. from Cyprus) (Figure 2) (see e.g., Agenbroad, 2012; Palombo, 2007, 2010; van der Bergh et al., 2001).

The frequency with which various proboscideans successfully colonized islands can be explained by their excellent swimming capacities, their trunk serving as a snorkel device, and their digestive system producing gasses which aid buoyancy. Moreover, these animals live and swim in herds; therefore, the founder group may have been sufficiently large to establish viable island populations. Sometimes proboscideans colonized a single island more than once, each time giving rise to endemic species differing in body size reduction, although the colonizers all originated from the same continental ancestor. They are therefore especially appropriate subjects for

evaluating causal mechanisms driving body size evolution in insular mammals. The species showing the most pronounced size reduction occurred in faunas where no other large herbivores were present (e.g., in the Sicilian ‘*Elephas falconeri* Faunal Complex [FC], the Cretan *Kritimys kiridus* FC, and in the *Stegodon sondaari* fauna from Flores). Conversely, species showing a less pronounced dwarfism occurred in impoverished, but quite diversified faunas either including large predators (e.g., the Sicilian ‘*Elephas mnaidriensis* FC and the San Teodoro Pianetti FC) or not including large predators (e.g., the Cretan *Mus minotaurus* FC). Proboscideans’ dwarfing was apparently unaffected by island surface area (Figure 3) and only indirectly influenced by the extent of island isolation from the mainland due to the temporal and geographical extent of barriers, acting as dispersal filters. Conversely it was strongly related to the presence/absence of other herbivores against which proboscideans must have competed for resources. When insular proboscideans were the only large mammals on an island, they underwent size reduction to an extreme degree, irrespective of the size of their mainland relatives, as shown for instance by *P. falconeri*, *M. creticus*, and *S. sondaari*, which can be regarded as the most prominent examples of the island rule.

The quite hypsodont (with high-crowned teeth) *S. sondaari*, a dwarfed species with a body weight of about 300 kg, is the only large mammal in the oldest Flores fauna, dated to about 900 ka. The fauna, strongly impoverished and disharmonic, includes the still-extant Komodo dragon, *Varanus komodoensis* (Figure 4), a giant *Geochelone* land tortoise, and a small crocodile. This fauna, disrupted by a volcanic eruption, was replaced at the transition from the early to middle Pleistocene by the *Stegodon florensis* fauna (van der Bergh et al., 2001). The dwarfed stegodon *S. sondaari* likely originated from Java elephants that reached Flores from the Sunda Shelf by crossing quite severe sea barriers.

M. creticus, likely a descendent of *Mammuthus meridionalis*, is known from a few remains, mainly molars, found at Cape Malekas (western Crete) together with the giant rodent

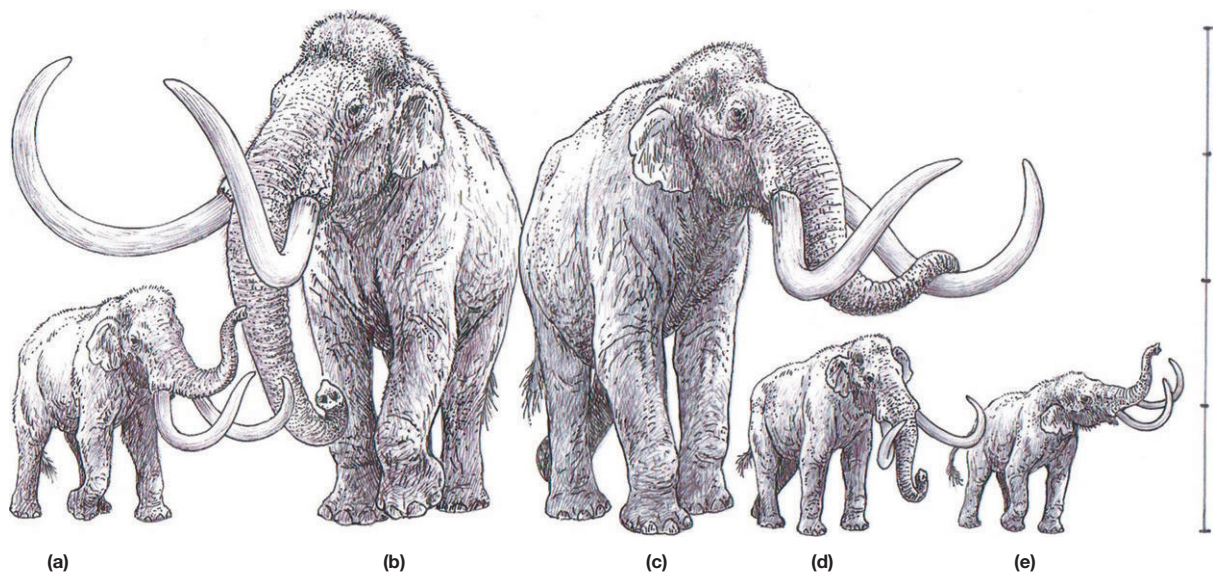


Figure 1 Artistic reconstruction of dwarf insular mammoths *Mammuthus creticus* (e), *Mammuthus lamarmorai* (d), and *Mammuthus exilis* (a) compared with their mainland ancestors *Mammuthus meridionalis* (c) and *Mammuthus columbi* (b). Drawing by S. Maugeri.

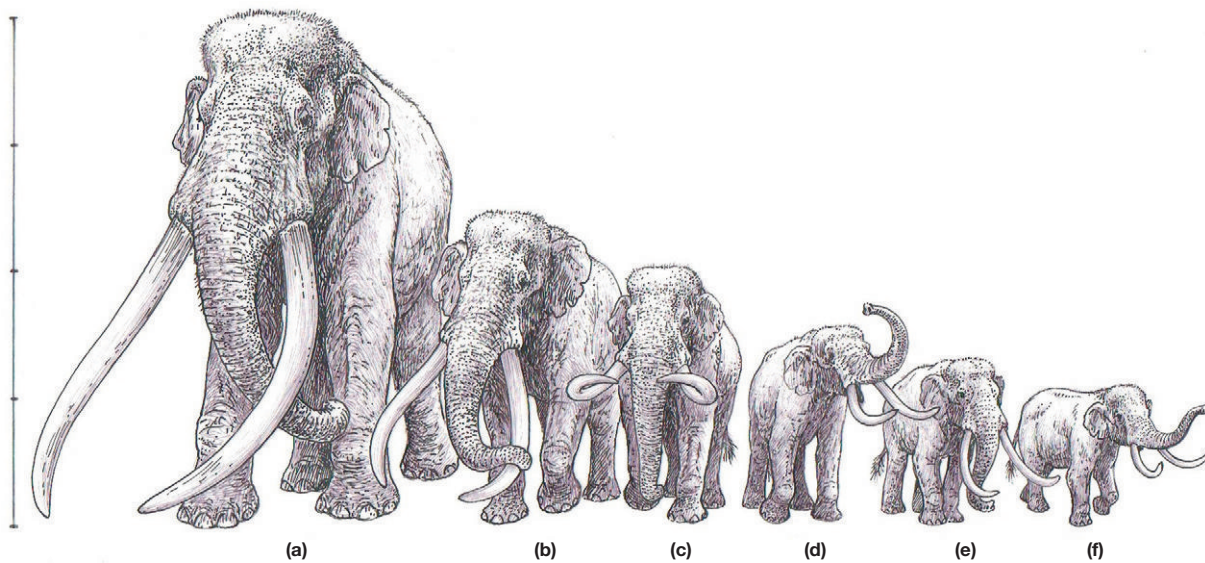


Figure 2 Artistic reconstruction of the dwarf insular *Palaeoloxodon creutzburgi* (b), *Palaeoloxodon mnaidriensis* (c), *Palaeoloxodon tiliensis* (d), *Palaeoloxodon cypriotes* (e), *Palaeoloxodon falconeri* (f) compared with their mainland ancestors *Palaeoloxodon antiquus* (a). Drawing by S. Maugeri.

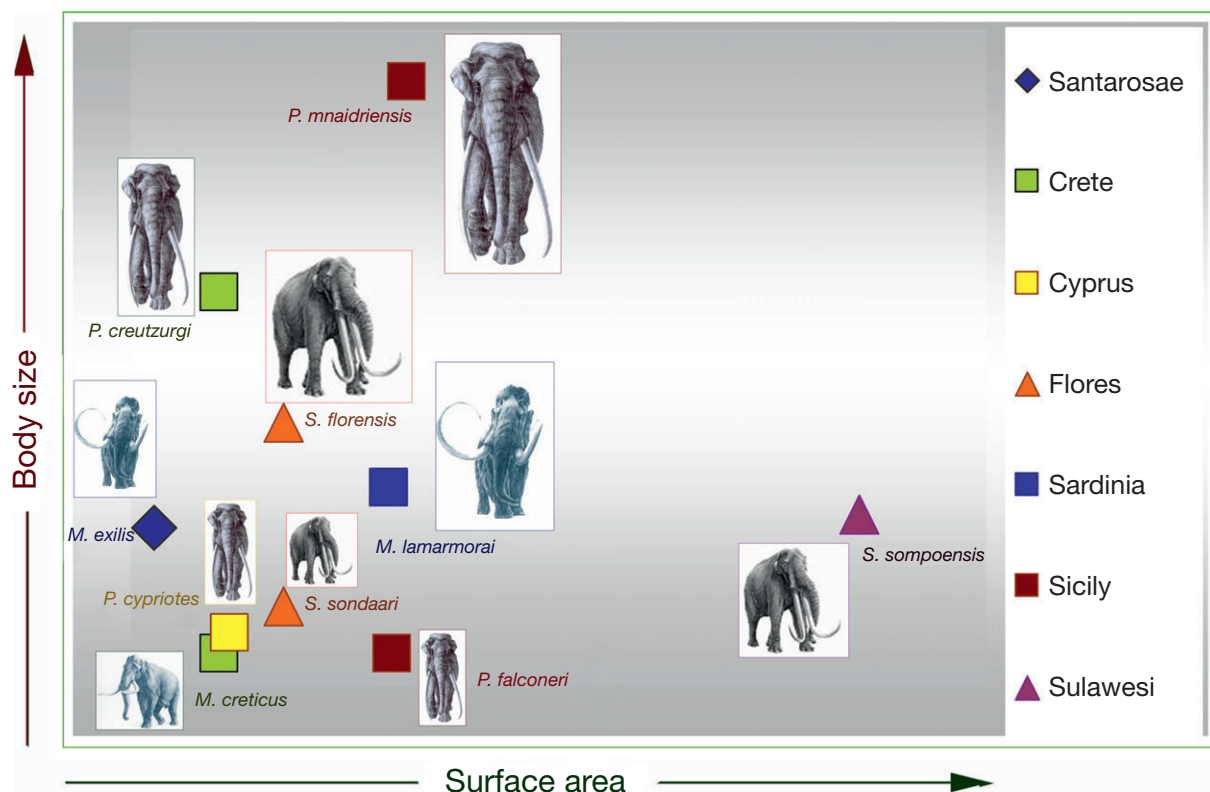


Figure 3 Body mass of endemic proboscideans plotted against the surface area of the islands they inhabited.

K. kiridus in deposits thought to be older than 800 ka. Its size has been estimated as slightly larger than that of *P. falconeri*. The latter is particularly known by the rich Sicilian material from Spinagallo cave (Syracuse, Southeastern Sicily), dated at about 500–450 ka (Figure 5). Its ancestor likely entered Sicily, and then colonized Malta, by the end of the early Pleistocene,

crossing a severe barrier, which prevented any other large mammal, except for the ancestor of the otter *Lutra trinacriae*, from reaching the insular system. As a result, the *E. falconeri* FC is disharmonic, poorly diversified and strongly endemic. Other than the elephant and otter, there were only small mammals that either evolved from endemic taxa that had existed since

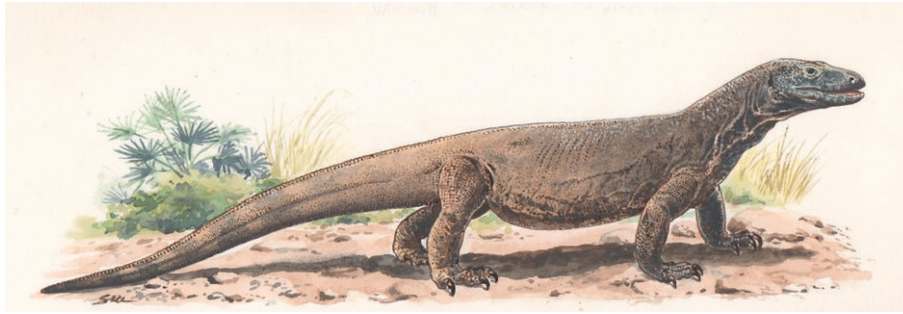


Figure 4 Artistic reconstruction of the giant Komodo dragon, *Varanus komodoensis*. Drawing by S. Maugeri.

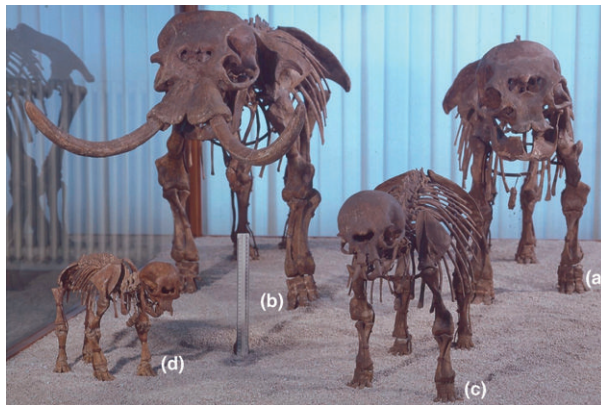


Figure 5 Mounted skeletons of the extinct (middle Pleistocene) *Palaeoloxodon falconeri* from Spinagallo Cave (early middle Pleistocene, Siracusa, Southeastern Sicily). This group (displayed at the Museum of Paleontology, Sapienza University, Rome, Italy) is said to include: (a) an adult female without tusks and with a height at the shoulder of about 80 cm; (b) an adult male high about 93 cm; (c) a young male, high about 35 cm; (d) a cub high about 24 cm. Courtesy of S. Conti, Director of the Museum.

the early Pleistocene Monte Pellegrino FC (*Leithia cartei* and *Leithia melitensis*, and *Maltamys* ex gr. *M. gollcheri-wiedincensis*, a distant relative of *Eliomys*) or from newly arrived species such as the ancestor of the endemic shrew *Crocidura esui*. This FC also included rather rich reptile and bird faunas, including giant strigiform birds (Figure 6). *P. falconeri* is the smallest endemic elephant found to date on the Mediterranean islands: the adult male would have a maximal withers height of about 110 cm and a body mass of about 170 kg, reduced by about 70% in height and 98.4% in weight with respect to its continental ancestor. The species shows precocious stunting of ontogenetic growth, as confirmed by skull features, tusk structure, and the proportions among the cranium, axial skeleton, and limbs. The peculiar shape of the nearly globose skull, similar to those of juvenile elephants, the extensive development of the brain case (the brain weight is about 1/60 of its body weight contra 1/1600 of the continental ancestor) (Figure 7), and some features of long bones indicate the persistence of juvenile traits in adult dwarf elephants (Palombo, 2010).

The smallest as well as medium-sized insular proboscideans – though evolved independently on each island and, therefore, taxonomically different – were characterized by

similar evolutionary traits such as the reduction of cranial bone pneumatization (the formation of air cells or cavities in bone), the decrease in the number of molar plates and the increase in enamel thickness, a relative increase in the brain size, less graviportal structure of limbs (columnar limbs with humerus and femur longer than radius-ulna and tibia-fibula, sometime adapted for slow movement because of large body mass), and a great morphological variability, at least in part associated, with an increased morphological and dimensional differentiation between sexes.

Hippopotamuses

Endemic hippopotamuses are known to be present in many different times and localities. They are found in late Pleistocene faunas of Madagascar. This included *Hippopotamus madagascariensis*, weighing about 400 kg, showing a typical aquatic habit, and *Hippopotamus lemerlei*, a more terrestrial and slightly smaller species weighing about 350 kg. They are also known from late middle to late Pleistocene faunas from Sicily, including *Hippopotamus pentlandi*, about 20% reduced in size with respect to its ancestor *Hippopotamus* ex gr. *H. amphibius*, and from Malta, where *Hippopotamus melitensis* was found. *H. melitensis* was smaller than the Sicilian species from which it likely originated. They are also known from the middle Pleistocene of Crete, where *Hippopotamus creutzburgi* is thought to have originated from *Hippopotamus* ex gr. *H. antiquus* and from the late Pleistocene and early Holocene of Cyprus, where *Phanourios minor* was found. The mainland ancestor and the actual time of dispersal of this species are uncertain (Figure 8).

Each population of dwarf hippo generally presents a wide range of morphological and dimensional variability, to the extent that extreme examples from the same island have sometimes been erroneously interpreted as two different species or subspecies (e.g., on specimens found on Malta and Crete). *Phanourios minor* (Figure 9) is undoubtedly the smallest (70 cm high at the shoulder) and the most specialized endemic hippopotamus, showing marked biometrical and morphological modifications of skull, dentition, and limb bones. Apomorphic characters (evolutionarily advanced character state), such as the lophodont dentition (teeth with cusps elongated to form narrow ridges), the quite stiff limb joints, the unguligrade structure (the animal walks only on its digits) of the autopodium (foot), indicate that the Cypriot hippo was mainly a browser, not a good runner, but able to negotiate rocky terrain (van der Geer et al., 2010). Similar limb modifications, shared by other hippos from the Mediterranean islands and more

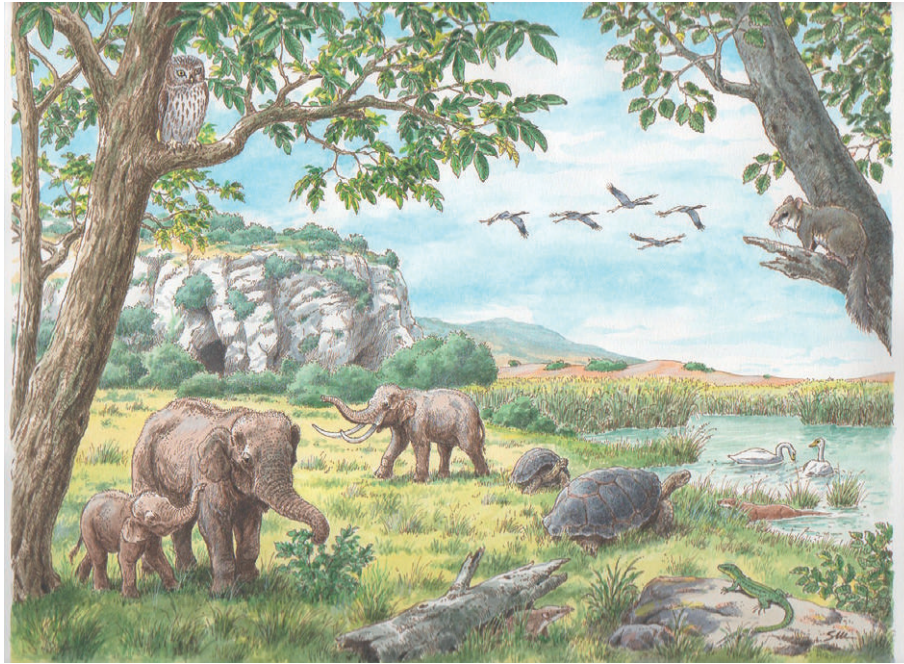


Figure 6 Artistic reconstruction of the environment of Spinagallo Cave (southeastern Sicily, early middle Pleistocene) showing *Palaeoloxodon falconeri* and its accompanying fauna: the dormouse *Leithia melitensis*, the otter *Lutra trinacrae*, the tortoise *Geochelone* sp., the lizard *Lacerta siculomelitensis*, the swan *Cygnus falconeri* (although it is not recorded in this site), the crane *Grus* cf. *G. melitensis*, the owl *Athena trinacrae*. Drawing by S. Maugeri.

evident as size reduction increases, are likely related to a habit of insular hippopotamuses being more terrestrial than that of their mainland (aquatic) relatives.

Anagenetic and Radiative Evolution: The Mediterranean Deer

During the Pleistocene and early Holocene, endemic deer, descendant from a number of different mainland ancestors (Megacerini, *Muntiacus*, *Diceros*, *Capeolus*, *Cervus*, *Dama*, *Axis*, and *Elaphurus*) were a common component of a number of insular faunas all over the world, from Japan to the Philippines, from Jersey to a number of Mediterranean islands. The Mediterranean insular deer, found in Eastern Mediterranean (Crete, Kassos, Karpathos, Amorgos, and Tilos) and Western Mediterranean islands (Sardinia and Corsica, Capri, Pianosa, Sicily, and Malta), belong to different mainland ancestors, and authors remain divided about the phylogenetic relationships, taxonomy, time of dispersal, and evolutionary history of most of them. No species with an entirely comparable adaptation existed on more than one island, therefore it is difficult to identify a dominant factor in their evolution. Evolutionary processes, mainly depending on intra-guild competition and nature of available niches, led to changes in size and adaptive strategy for a more efficient exploitation of resources under the peculiar environmental conditions of each island ecosystem, but causal mechanisms interplayed differently in shaping the evolution of each insular species.

The endemic deer from Sardinia and Crete are of particular interest. Although they are generally regarded as stemming from closely related ancestors (i.e., the Megacerini tribe), they underwent different evolutionary processes, even on islands with similar surface area and physiography. For instance, the

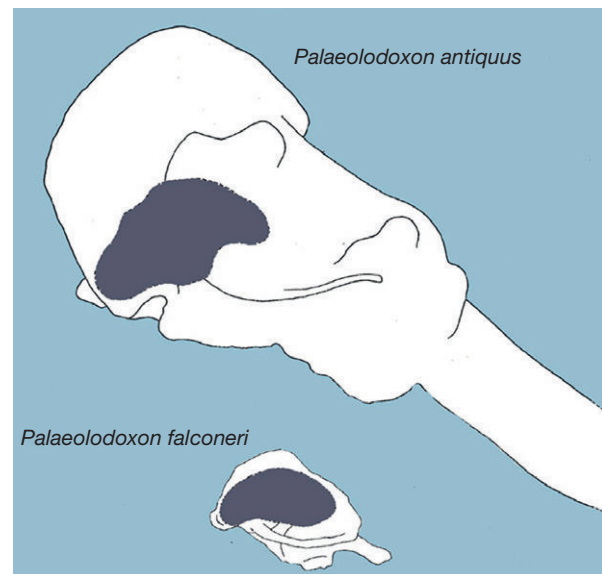


Figure 7 Comparison between the brain case of *Palaeoloxodon antiquus* (Pian dell'Olmo, Central Italy) (above) and *Palaeoloxodon falconeri* (female from Spinagallo Cave, Southeastern Sicily) (below). Drawing by M.R. Palombo.

sympatric speciation and radiation process of Cretan deer led both to 'dwarfism' and 'gigantism,' whereas in Sardinia endemic deer underwent a progressive decrease in size.

The Pleistocene endemic deer from Sardinia belong to the '*Praemegaceros*' group, as supported by some skull, mandible, and teeth features (Figure 10). The forerunner (?*Praemegaceros verticornis*) probably entered Sardinia during the late early

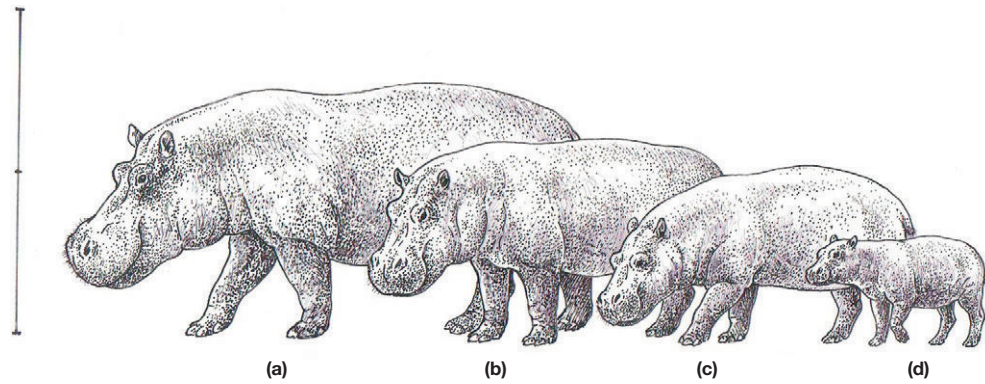


Figure 8 Artistic reconstruction of the dwarf insular hipopotamuses *Hippopotamus pentlandi* (Sicily and Malta) (b), *Hippopotamus creutzburgi* (Crete) (c) and *Phanourios minor* (Cyprus) (d) compared with the extant *Hippopotamus amphibius* (a). Drawing by S. Maugeri.



Figure 9 Mounted skeleton of the Cypriot dwarf hippopotamus *Phanourios minor*. Height at the shoulder is about 50 cm. Museum of Geology and Paleontology, University of Athens, Greece. Courtesy of G. Lyras.

Pleistocene low sea level episode (marine isotope stage (MIS) ?24–22) and survived on the island until 7000 years ago. During its anagenetic evolution (gradual evolution of an ancestral species to a descendant species with no branching or splitting of the lineage), Sardinian ‘*Praemegaceros*’ underwent a progressive size reduction, from the largest *Praemegaceros sardus* to the smallest latest Pleistocene population of *Praemegaceros cazioti*, found in Corbeddu cave. Metapodials (elongated bones in the feet) became shorter, but the Sardinian megacerines acquired a more agile gait on both hard and uneven ground and increased their consumption of dry grass as compared to their possible ancestor (Palombo, 2005).

On Crete endemic deer of different sizes, feeding behavior and locomotion (Caloi and Palombo, 1995; de Vos, 1996) were recorded from the late Pleistocene *Mus catreus* and *Mus minotaurus* FCs. Their forerunners probably reached the island by swimming during the late middle Pleistocene glacio-eustatic low sea level stand. No direct competitors were present when these deer arrived on the island. Consequently, many habitats and empty niches were available. This fact allowed deer to undergo an adaptive radiation. Some decreased in size, as

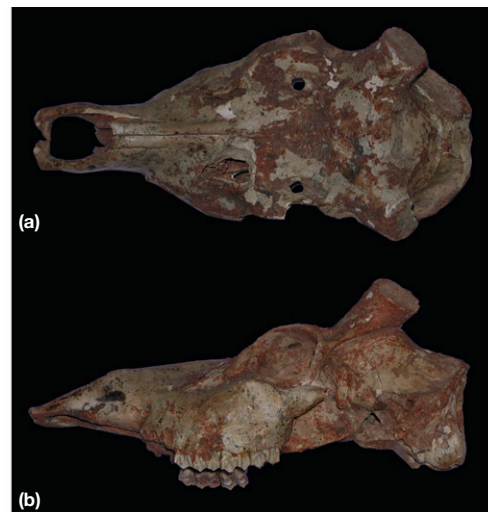
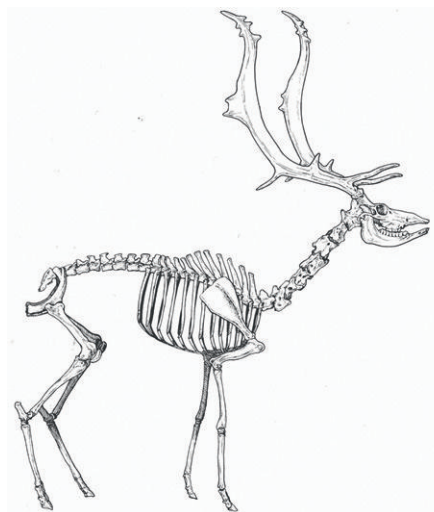


Figure 10 *Praemegaceros cazioti* from Dragonara Cave (late Pleistocene, Alghero, Northwestern Sardinia): reconstructed skeleton (on the left, modified from Caloi L and Malatesta A (1974) Il cervo pleistocenico di Sardegna. *Memorie dell'Istituto Italiano di Paleontologia Umana* 2: 163–246), and skull in dorsal (a) and lateral view (b). Photograph by M.R. Palombo.

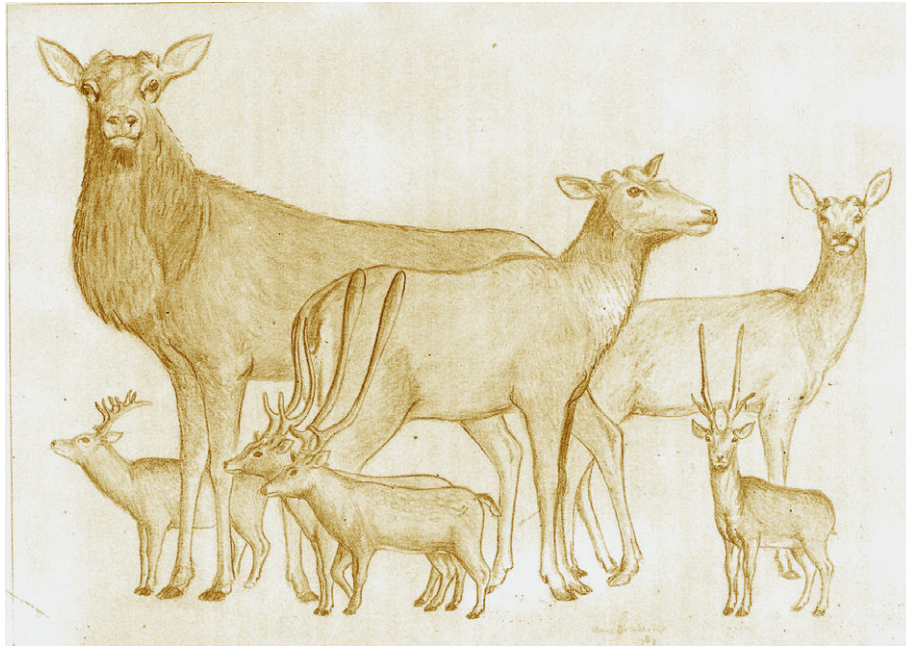


Figure 11 Artistic reconstruction of the endemic deer, which inhabited Crete during the late Pleistocene. Large species, from the left to the right: *Candiacervus majori*, *Candiacervus dorothenensis*, *Candiacervus rethymnensis*, *Candiacervus* sp. IIc, and *Candiacervus* sp. IIb, *Candiacervus ropalophorus*. Modified from de Vos J (1996) Taxonomy, ancestry and speciation of the endemic Pleistocene deer of Crete compared with the taxonomy, ancestry and speciation of Darwin's Finches. In: Reese DS (ed.) *Pleistocene and Holocene Fauna of Crete and Its First Settlers*, *Monographs in World Archaeology* 28: 111–124.



Figure 12 Mounted skeleton of the Cretan dwarf deer *Candiacervus* sp. II. Height at the shoulder is about 50 cm. Museum of Geology and Paleontology, University of Athens, Greece. Courtesy of G. Lyras.

expected by the island rule, but others become larger than their mainland ancestors. Cranial and limb morphology, dimension, and proportion allow us to recognize six size classes (Figure 11): *Candiacervus ropalophorus* had a height of about 40 cm at the shoulder; *Candiacervus* spp. II (three morphotypes) was about 50 cm at the shoulder (Figure 12); *Candiacervus cretensis* was about 65 cm at the shoulder; *Candiacervus rethymnensis* whose size was comparable to the mainland species *Cervus elaphus*; the larger *Candiacervus dorothenensis* (= *Candiacervus* sp. V) was high about 150 cm at the shoulder, and the giant *Candiacervus major* (= *Candiacervus* sp. VI) about

200 cm. The smallest Cretan deer *C. ropalophorus* shows a modified locomotory system (lowering of centre of gravity, greater stability, low-gear locomotion, and stiffer joints), but brain/body mass proportion provides evidence that this endemic deer underwent only minor changes in the relative size of brain after its geographic isolation.

Shortened and Long Metapodials: The Strange Case of Balearic and Sardinian Bovids

Fossil bovids are unusual components of Quaternary insular faunas, only recorded in some Asian islands and in the Western Mediterranean, where the most intriguing species, showing highly endemic features, inhabited the Eastern Balearic Islands and Sardinia.

On the Balearic Islands, the only large mammals recorded in highly impoverished, disharmonic faunas, belong to six chronospecies of the *Myotragus* phylogenetic lineage, which evolved in isolation for more than 5 Ma before becoming extinct somewhere between 3700 and 2040 years BC (Bover et al., 2010). Evolutionary patterns within *Myotragus* include body size reduction, reduction in number of incisors and premolars, and changes in skull and postcranial morphology (Figures 13 and 14). The most advanced species, *Myotragus balearicus*, the Balearic Islands Cave Goat, was no more than 50 cm tall at the shoulder; its eye sockets almost faced the front of the face, granting stereoscopic vision; there was only one perennial-growth incisor in its mandible and its feet bones were dramatically shortened. Indeed, Balearic bovids acquired a slow, powerful walking gait by reducing limb length in a predator-free environment. This was done through a



Figure 13 Skulls of archaic and advanced representatives of the *Myotragus* phylogenetic lineage: *M. peponellae* (early/middle Pliocene) (on the left) and *M. balearicus* (middle Pleistocene to Holocene) (on the right). IMEDEA (CSIC-UIB), Institut Mediterrani d'Estudis Avançats (Illes Balears, Spain). Photograph by P. Bover.

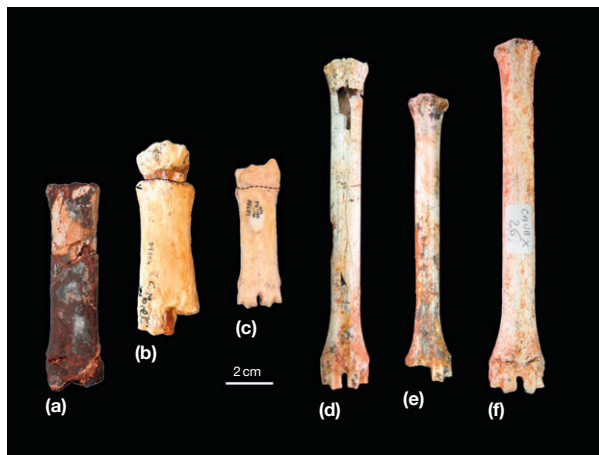


Figure 14 Comparison among metatarsals of *Myotragus* and *Nesogoral*: (a) *Myotragus palomboi*, IMEDEA 90140 (holotype) (photograph by P. Bover); (b) *Myotragus peponellae* MNIB 59204 (photograph by P. Bover); (c) *Myotragus balearicus* IMEDEA 85651 (dashed lines indicate the anatomical separation of metatarsal and cubonavicular, fused in the more modern species of *Myotragus*) (photograph by P. Bover); (d–e) *Nesogoral* sp. ORXB1064 and ORXB1065 from the quarry X Ghiro, Monte Tuttavista, Orosei, Eastern Sardinia (photograph by R. Rozzi); (f) *Nesogoral* sp. CAVAX26 from the quarry X Mele, Monte Tuttavista, Orosei, Eastern Sardinia. Photograph by R. Rozzi.

marked shortening of metapodials and phalanges (the bones of the hind and front feet, respectively). Some remarkable similarities can be traced between *Myotragus* and both Miocene archaic 'goats' *Aragoral* (late Vallesian, MN10) and *Norbertia* (Turolian–Ruscinian boundary, MN13/14), which had robust metapodials, weak upper premolar styles and reduced second lower premolar, and therefore, eligible as putative ancestors of the Balearic genus. Since no bovid remains are known from the middle Miocene of the Balearic Islands, the ancestor of *Myotragus* presumably reached Mallorca sometime during the Messinian Salinity Crisis (MSC) by a temporary and selective emerged path. During the MSC, started at 5.96 Ma, parts of the Mediterranean basin became dry land because of a large fall of about 1500–2500 in the Mediterranean water level due to the isolation from the Atlantic Ocean.

On Sardinia, an endemic bovid, '*Antilope* (?*Nemorhaedus*)' *melonii* (= *Nesogoral melonii*) was first described, at the beginning of the last century, from breccias deposits cropping out at Capo Figari (northeastern Sardinia) (Figure 15), a locality whose karstic fissures yielded a number of bovid remains, one of which dated at about 1.8 Ma. The richest sample of bovids of the so-called '*Nesogoral* group' comes from Monte Tuttavista (Orosei, western Sardinia) (morphotype A = *Nesogoral* sp. 1 aff. *N. melonii* and morphotype B = *Nesogoral* sp. 2) (Figure 15), from a few sites ranging in age from the late Pliocene (Mandriola, *Nesogoral* sp.) to early Pleistocene (Capo Mannu, *Nesogoral* sp.), and from an unknown locality in the Campidano area (*Nesogoral cenisae*) (Palombo, 2009).

For decades, a close phylogenetic relationship between *Nesogoral* and *Myotragus* has been widely accepted. Nonetheless, although *Nesogoral* and archaic *Myotragus* share some cranial features (Palombo et al., 2006), differences in their limb bones cast doubt on this hypothesis. In particular, the metapodials of *Nesogoral*, despite their morphological and dimensional variability (Figure 14), are elongated and more slender than in archaic continental caprines such as *Aragoral* and *Norbertia*. On the other hand, their morphology and proportion are close to some representatives of the *Protoryx*/*Pachytragus* group, known from the late Miocene of Tunisia, Turkey, and Greece (Rozzi, 2010). In contrast to *Myotragus*, the Sardinian bovids likely retained the characteristics of long leg bones and the ability to run, due to the presence of a top predator, the hyaena *Chasmaporthetes melei* (Figure 16). Neither hyaenids nor bovids can easily swim or float, so it is rational to suppose that their ancestors did not cross water to reach Sardinia, nor was these sweepstake dispersals but, more likely, they reached the island by a selective path sometime during the MSC. Possibly at the same time, Sardinia was colonized by the unknown ancestor of *Asoletragus gentryi*, the strange bovid with stout, straight, almost conical and strongly backwards slanting horn-cores, known from a single incomplete skull found at Monte Tuttavista (Figure 15).

Bovids of a relatively reduced size have also been recorded in late middle Pleistocene to late Pleistocene deposits of Sicily (*Bos primigenius siciliae* and *Bison priscus siciliae*) and have been claimed on Pianosa (*Bos primigenius bubaloides*) and Malta (*Bos* sp.). Outside the Mediterranean, endemic bovids have been reported from the early–middle Pleistocene of Java



Figure 15 Skulls of early Pleistocene endemic bovids from Sardinia: (a) incomplete skull still embedded in the bone breccias block, holotype of the species *Nesogoral melonii* (Capo Figari, Northeastern Sardinia) (photograph by M. Pavia); (b) *Nesogoral* aff. *N. melonii* (quarry XI Antilope, Monte Tuttavista, Orosei, Eastern Sardinia) (photograph by M.R. Palombo); (c) *Nesogoral* sp.1 (quarry X Ghiro, Monte Tuttavista, Orosei, Eastern Sardinia) (photograph by M.R. Palombo); (d) *Asoletragus gentryi* (quarry VI-3, Monte Tuttavista, Orosei, Eastern Sardinia). (b–d) Specimens kept in the National Archeological Museum of Nuoro (Italy) (photograph by M.R. Palombo).

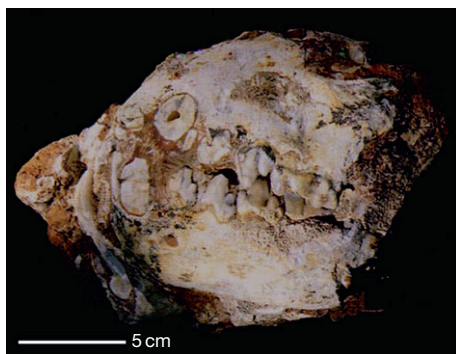


Figure 16 Incomplete skull and mandible still embedded in the bone breccias block of the early Pleistocene Sardinian hyaena *Chasmaporthetes melei* from Monte Tuttavista (Orosei) (photograph by M.R. Palombo). Courtesy of the National Archeological Museum of Nuoro (Italy).

(*Duboisia santeng*, **Figure 17**; *Bubalus paleokerabau*, *Bibos palae-sondaicus*, *Epileptobos groeneveldtii*), from the late Pleistocene-Holocene of Sulawesi (*Bubalus (Anoa)* sp.), from the middle and late Pleistocene of Japan (*Bubalus* sp. and *Capricornis crispus/Capricornis* sp.), from the middle Pleistocene of Taiwan (*Bubalus* sp.), from the Holocene of Madagascar (*Bos* sp.),



Figure 17 Skull of the early Pleistocene endemic bovid *Duboisia santeng* from Java, lectotype of the species in frontal (on the left) and lateral view (on the right) (Trinil, Java). Courtesy of J. De Vos, NCB Naturalis (Leiden, The Netherlands). Photograph by R. Rozzi.

and from the Pleistocene-Holocene of the Philippines (*Bubalus cebuensis*, with a body mass of 150–165 kg, and *Bubalus* sp.).

Uncommon Island Settlers: Carnivora and Primates Including Hominins

Carnivora

Although the absence/scarcity of predators on islands has been assumed by some authors as a causal mechanism underlying ‘dwarfism’ and ‘gigantism’ of insular mammals, extant and fossil terrestrial and semiaquatic Carnivora are known worldwide also from a number of oceanic and oceanic-like islands, inhabited by impoverished, disharmonic endemic faunas.

Large predators have been reported from a few impoverished but balanced insular faunas (e.g., the middle–late Pleistocene *Panthera tigris* from Java and Japan, *Panthera spelaea* and *Crocota crocuta* from Sicily), but the only fossil endemic top predator thus far known is the early Pleistocene Sardinian hunting hyaena *C. melei*, known from a single partial skull from Monte Tuttavista (**Figure 16**). It remains unknown whether the ancestor was *Chasmaporthetes lunensis*, which might have entered Sardinia at the beginning of the Quaternary, or a Miocene species, dispersing to the island together with bovids and suids during the MSC (Palombo, 2009). *C. melei* is supposed to have been slightly reduced in size, consistent with the small size of its prey (the small monkey *Macaca majori*, the dwarf pig *Sus sondaari*, **Figure 18**, and the goat-sized bovids), and the lack of competitive pressure exerted by the two small mustelids, *Pannonictis*, and *Mustela* aff. *M. putorius*, present at that time on Sardinia.

However, the reduction in size of fossil insular carnivores was not a rule. Among canids, for instance, Java *Megacyon merriami* retained roughly the size of its ancestor *Xenocyon lycaonoides*, while its descendent *Mececyon trinilensis* was dwarfed. A reduction in size and significant morphological apomorphies (derived or advanced character states) characterize the fossil ‘wild dog’ *Cynotherium*, known from the late Early to the latest Pleistocene (11.35 ka BP) of Sardinia and the middle to late Pleistocene of Corsica (**Figure 19**). Although the actual phylogenetic significance of the presence of both hypercarnivorous (dental) and hypocarnivorous (cranial) features is still a matter of debate, *Cynotherium* possibly originated from an advanced European *Xenocyon*. Once on the island and consistently with its specialized hunting on small prey, *Cynotherium*

progressively reduced its size from the about 17 kg of *Cynotherium* sp. found in deposits older than 400 ka at Grotta dei Fiori cave (Carbonia), to the about 12 kg of the ~11.350 ka old *Cynotherium sardous* from Corbeddu cave (Oliena). *C. sardous* was likely a stalking hunter of the ochotonid *Prolagus*, as suggested by the high lateral mobility of the head and flexibility of its fore limb joints, as well as by the enamel structure and microwear pattern of its teeth (Novelli and Palombo, 2007; van der Geer et al., 2010).

All in all, patterns of size change of terrestrial carnivores, albeit highly variable, mainly depended on the size of the most readily available prey, while a less clear pattern characterized

the semiaquatic species, such as otters, whose remains, though scarce, have been found in many disharmonic faunas from the Mediterranean islands. Seven otter species have been described, three from the late Pleistocene of Sardinia: *Sardolutra ichnusae*, *Algalolutra major*, and *Megalenhydrys barbaricina*, the only giant otter possibly descended from the Corsican species *Lutra castiglioni*. In addition, *Lutra euxema* have been found from Malta, *Lutra trinacrae* from Sicily, and *Lutrogale cretensis* from Crete. The morphological variations and different types of adaptation of the insular otters seem to emphasize the differential characteristics already present in their ancestors. *Lutrogale cretensis* (Figure 20), for instance, accentuated the terrestrial adaptation already present in *Lutrogale*, while *L. euxema*, *L. trinacrae*, and the Sardinian otters exhibited a definitely aquatic habit, although they were adapted to different environments and had a quite different feeding behaviour (Willemsen, 2006).

Primates

Few fossil Quaternary primates have been recorded from islands elsewhere. They include the well-known late Pleistocene–early Holocene strepsirrhine primates endemic to Madagascar, among which very large extinct fossil lemurs (*Hadropithecus*, *Archaeolemur*, and *Babakotia*) as well as giant species (*Megalapsis*, *Archaeoindris*, and *Palaeopropithecus*) have been recorded. Outside of Madagascar, we can mention the dwarf macaque *Macaca majori* reported from few early Pleistocene localities of Sardinia (Figure 18) and the Platyrrhine monkeys reported from Cuba (*Paralouatta marianae* and *Paralouatta varonai*), Hispaniola (*Anthillotrix bernensis*), and Jamaica (*Xenothrix mcgregori*).

On the other hand, insular primates include one of the most amazing and debated insular species: *Homo floresiensis*. Since its discovery, in a disharmonic, impoverished fauna including a dwarf proboscidean (*Stegodon florensis insularis*), endemic murids (among others *Papagomys armandvillei*, *Papagomys theodorverhoeveni*, *Spelaeomys florensis*), giant varans (*Varanus komodoensis* and *V. hooijeri*), and a giant marabou, *Leptoptilos robustus* (Figure 21), this so-called ‘hobbit’



Figure 18 Skulls of the Sardinian dwarf macaque *Macaca* aff. *M. majori* (a) and of a juvenile individual belonging to the pygmy pig *Sus sondaari* (b) from the early Pleistocene of Monte Tuttavista, quarry VI-3 (Orosei, Eastern Sardinia) (photograph by M.R. Palombo). Courtesy of the National Archeological Museum of Nuoro (Italy).



Figure 19 *Cynotherium sardous* from Dragonara cave (late Pleistocene, Alghero, Northwestern Sardinia): (a) skull in lateral (above) and dorsal (below) view. (b) Mounted skeleton (above) of the dwarf Sardinian ‘wild dog’ *Cynotherium sardous*. Paleontological Museum, ISPRA, Rome, Italy (courtesy of F. Angelelli); artistic reconstruction of the dwarf Sardinian ‘wild dog’ (below). Drawing by S. Maugeri.



Figure 20 Complete skeleton of the Cretan otter *Lutrogale cretensis* from Liko Cave (late Pleistocene). Head-body length is 72.5 cm. Museum of Geology and Paleontology, University of Athens, Greece. Courtesy of G. Lyras.

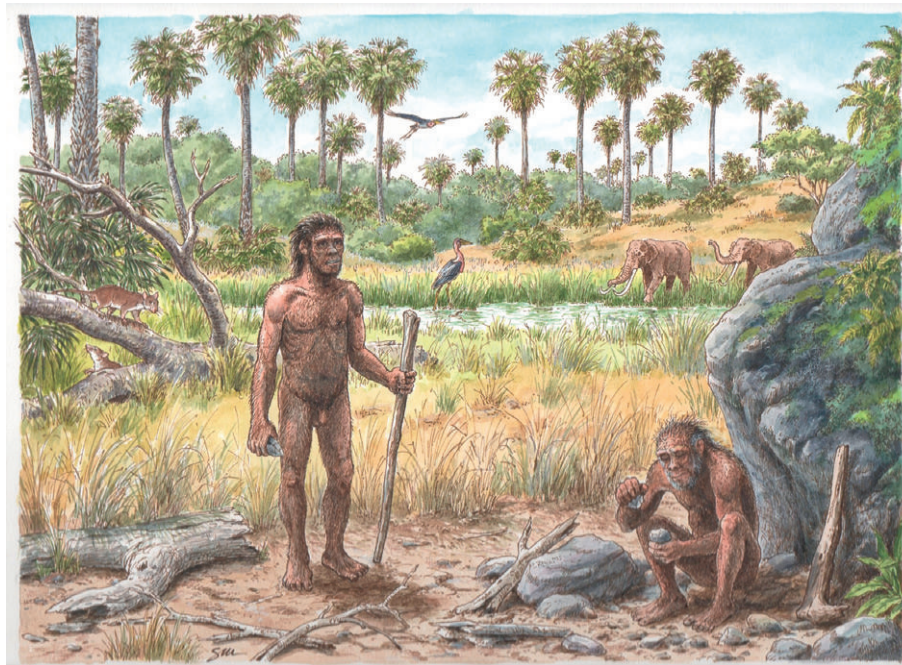


Figure 21 Artistic reconstruction of the environment of Flores in the late Pleistocene showing the dwarf hominin *Homo floresiensis*, the giant marabou *Leptoptilos robustus*, the giant rodent *Papagomys armandvillei*, and the dwarf proboscidean *Stegodon florensis insularis*. Drawing by S. Maugeri.

represents one of the most substantial and controversial paleo-anthropological discoveries of the last decade. The Flores endemic hominin, with its mixture of archaic and advanced features, large orbits, small cranial capacity (Figure 22), small body and short limbs, was alternatively regarded as phylogenetically related to a very archaic hominin stock, or as a small-brained dwarf species descendant from *Homo erectus*, or a pathologic, microcephalic modern *H. sapiens* (Aiello, 2010 and references therein). Recently published detailed descriptions of the skull and the rest of the skeleton, as well as data on the associated fauna and lithic implements, besides the environmental and geomorphological context, support the hypothesis that *H. floresiensis* was an insular dwarf species, although whether it was derived from a large-bodied *H. erectus* or from a different, smaller bodied early *Homo* spp., is still an unanswered question (Aiello, 2010; van der Geer et al., 2010).

A Glance at the Insular Small Mammals

Endemic Quaternary rodents are reported from all over the world (van der Geer et al., 2010). Murine murids (Old World rats and mice) were the most widespread, known from several Asian and Mediterranean islands, Madagascar, the Canary, Channel, and Solomon islands, while cricetine murids (New World mice and rats, voles, hamsters, and relatives) are described from the Galapagos and Antilles Islands. Glirids (dormice) and arviculids (voles, lemmings, and muskrats) were mainly present in the Mediterranean islands, ctenodactylids (African guinea pigs) only in the early Pleistocene of Sicily (*Pellegrinia panormensis*, Figure 23). Endemic hystricognath rodents were present during the late Pleistocene, if not earlier, on several small and large islands of the West Indies. Endemic lagomorphs (rabbits, hares, and pikas) were less common: during the Pleistocene ochotonids (pikas) were only present



Figure 22 Cast of the skull of *Homo floresiensis* from Liang Bua cave (late Pleistocene, Flores). Courtesy of J. De Vos, NCB Naturalis (Leiden, The Netherlands). Photograph by E. Kruidenier.

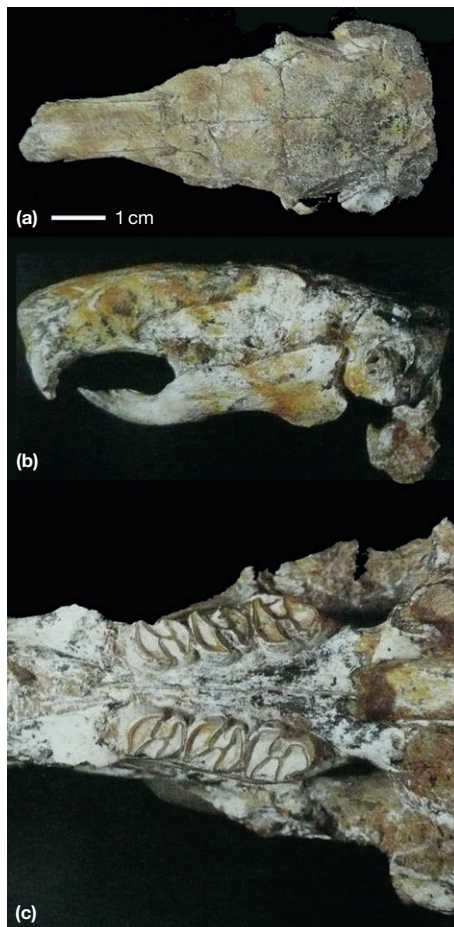


Figure 23 Incomplete skull in dorsal (a), lateral (b), and ventral (c) views of the Sicilian ctenodactylid *Pellegrinia panormensis* (early Pleistocene, Monte Pellegrino, Palermo).

on the Corso-Sardinian massif (*Prolagus figaro*, *Prolagus sardus*, **Figure 24**), Leporidae (rabbits and hares) are reported from the early Pleistocene of Sicily (*Hypolagus peregrinus*) and Sardinia (a new undescribed genus), the late Pleistocene of Ryukyu islands (*Pentalagus furnessi*), and perhaps Sumatra.

Insectivores are less common in Quaternary insular fossil assemblages. More-or-less distant relatives of the shrew *Asoriculus* are reported in Sicily (*Asoriculus burgioi*), Sardinia (*'Asoriculus' similis*, *'Asoriculus' sp. II*), Corsica (*'Asoriculus' corsicanus*) and the Balearic islands (*Nesiotites ponsi*, *Nesiotites hidalgo*). White-toothed shrews are known in Sicily and Malta (*Crocidura esui*, *Crocidura* aff. *Crocidura sicula*), Crete (*Crocidura zimmermanni*), the central Ryukyu Islands (*Crocidura orii*, *Crocidura watesi*), and possibly Flores. Finally, shrew-like insectivores have been recorded from the West Indies. Quaternary endemic moles seem to be restricted to Sardinia (*Talpa tyrrhenica*), solenodons have been restricted to Cuba (*Soledon arredondo*) and Hispaniola (*Solenodon marcanoi*), while a subfossil tenrec has been reported from Madagascar (*Microgale macphee*).

Consistent with the island rule's prediction, a number of fossil rodents significantly increased their size on islands, albeit a similar pattern can be found among some lagomorphs and insectivores. Among the largest rodents, the Sicilian *Leithia melitensis*, (**Figure 25**), in a predator-free environment, reached a size about double the size of recent dormouse, while giant rats from Flores (the still-extant *Papagomys armandvillei* at least twice the average size of *Rattus norvegicus*, and the slightly smaller extinct *P. theodorverhoeveni* weighed more than 1 kg) increased their size despite the presence of powerful predators such as *Homo floresiensis*, the Komodo dragon, and the giant marabou stork (*Leptoptilos robustus*) (**Figure 21**) (Meijer and Due, 2010). Although giant species have also been reported from the Galapagos (*Megaoryzomys curioi*) and the Lesser Antilles (*Megalomys audreyae*, *Amblyrhiza inundata*, **Figure 26**), this evolutionary pattern was not a 'rule,' and some rodents did not undergo significant size changes. On Crete, for instance, although during its anagenetic evolution (a pattern of evolution that results in a linear descent with no branching or splitting) the endemic Old World mouse (*Mus*) lineage underwent an increase in size from *Mus bataei* to *Mus minotaurus*, the latter was only slightly larger than the mainland *Mus* representatives.



Figure 24 Mounted skeleton of *Prolagus sardus* from Garroppu shelter (Carbonia, Southwestern Sardinia). 'Museo dei Palaeoambienti Sulcitani PAS/E.A. Martel' (Carbonia). Photograph by G.L. Pillola.



Figure 25 Mounted skeleton of the Sicilian dormouse *Leithia melitensis* from Spinagallo cave (early middle Pleistocene, Siracusa, southeastern Sicily). Museum of Paleontology, Sapienza University of Rome (Italy). Photograph by M.R. Palombo.



Figure 26 Fragment of skull in ventral view of the giant hutia *Amblyrhiza inundata* from the late Pleistocene of St Martin (West Indies). Courtesy of J. De Vos, NCB Naturalis (Leiden, The Netherlands). Photograph by E. Kruidenier.

Anagenetic evolutionary patterns are not uncommon among insular small mammals. However, such a pattern characterized three distinct Sardinian lineages: *Rhagapodemus/Rhagamys*, *Microtus (Tyrrhenicola)*, and *Prolagus sardus*. As a matter of fact, from the late Pliocene *Rhagapodemus azzarolii* to the most advanced late Pleistocene–Holocene *Rhagamys orthodon*, Sardinian mice show an increase in body size, increased hypsodonty, progressively advanced molar features, and the appearance of accessory tooth roots (Abbazzi et al., 2004). The large pika *Prolagus sardus* (Figure 24), present on the island from the late middle Pleistocene to Roman historical times, shows a clinal morphological variation of dental occlusal surface (the surface of the teeth that comes in contact with those of the opposite jaw during chewing) and an asymptotic increase in size (Angelone et al., 2008), despite the fact that it was the

preferred prey of *Cynotherium* in the Pleistocene and then became an important food source for humans from Upper Paleolithic to Neolithic times.

Giant Turtles, Dragons, and Crocodiles

It has generally been accepted that ecological release, accompanied by low predation and competition pressures in highly disharmonic assemblages, could provide an explanation for the gigantism of most of endemic reptiles on islands. Although the increase in size of insular reptiles is not actually a rule and their change of size, irrespective of the size of the mainland ancestors, depends on the interplay of several factors such as resource availability and, as regard to predators, the size of prey, it is undoubted that a number of impressive cases of gigantism can be found among fossil reptiles.

During the Quaternary period, islands have harbored several giant taxa, among which are lizards and tortoises. One of the most famous examples is the Komodo Dragon (*Varanus komodoensis*) (Figure 4), with a total length of over 3 m and a weight of over 150 kg. This species occurred on Australia during the early Pliocene to middle Pleistocene, on Flores from the early Pleistocene through the Holocene, and is still living on the Indonesian islands of Rinca, Flores, Gili Motang, Gili Dasami, and Komodo. Other fossil insular representatives of this genus have been described from Java (*Varanus* cf. *V. komodoensis* from the middle Pleistocene; *Varanus salvator* from the early Pleistocene) and Timor (a new unnamed species from the middle Pleistocene) (Hocknull et al., 2009). The feeding ecology of this versatile predator has been a subject of long-standing interest and considerable conjecture. The most recent studies suggest that the combination of highly and very specifically optimized cranial and dental architecture, together with a capacity to deliver a range of powerful toxins, minimizes prey contact time and allows *Varanus* to kill a wide range of prey including large taxa, such as the 40 kg Sunda Deer (Fry et al., 2009).

Many other lizards, iguanas, and geckos are good examples of insular gigantism. These include endemic species that are either extant or fossil, such as *Hoplodactylus* and *Oligosoma* (= *Cyclodina*) *northlandi* from New Zealand, *Gallotia* from the Canary Islands, *Aristelliger titan* from Jamaica, *Anolis roosevelti* from Culebra Island, Puerto Rico *Conolophus* from the Galapagos Islands, *Lacerta siculimelitensis* from Sicily and Malta, *Brachylophus* sp. from Lifuka (Tonga), and *Cyclura* from the West Indies.

Extant chelonians (tortoises and turtles) span almost four orders of magnitude of body size, including the famous insular giants *Geochelone gigantea* (from Aldabra in the Seychelles) and *Geochelone elephantopus* (from the Galapagos Islands). Although the evolutionary determinants of size diversity and gigantism in chelonians are still poorly understood, a pronounced relationship between habitat and optimal body size has recently been found in this group (Jaffe et al., 2011). In addition to the living forms, several insular giant tortoises have been reported from Pleistocene and Holocene fossil material from all over the world (e.g., Cuba, Madagascar, the Seychelles, Bahamas, Moro Island, Curaçao, Sicily, Malta, Mallorca, New Caledonia, Timor, Java, Sulawesi, Flores).



Figure 27 Mounted skeleton of the middle Pleistocene crocodile *Toyotamaphimeia machikanensis* from Japan. Moscow Paleontological Museum (Moscow, Russia). Photograph by D. Vasilyan.

Finally, we might mention among large insular crocodiles (more than 3 m long) the terrestrial *Mekosuchus inexpectatus* from the Holocene of New Caledonia, *Mekosuchus kalpokasi* from the Holocene of Vanuatu, *Volia athollandersoni* from the Pleistocene of Fiji islands and the fresh water *Toyotamaphimeia machikanensis* from the middle Pleistocene of Central Japan (Kobayashi et al., 2006) (Figure 27).

Wingless Birds and Giant Owls

The same forces promoting gigantism in reptiles have been invoked to explain the evolution of large size in birds, mainly through the evolution of flightlessness. This trend is not restricted to predatory birds, because herbivorous birds, in the absence of herbivorous mammals, may also develop gigantism. The most famous examples are the eleven species of moa (Dinornithiformes) of New Zealand (Figure 28), the larger representatives of which reached about 3.7 m in height and weighed about 230 kg (Bunce et al., 2003), the elephant birds (Aepyornithiformes) from the Quaternary of Madagascar and the dodo (*Raphus cucullatus*; Columbiformes) from Mauritius. Other giant and in some cases flightless pigeons have been reported from the Quaternary of Fiji (*Natunaornis gigoura*; *Ducula lakeba*), Rodriguez Island in the Indian Ocean (*Pezophaps solitaria*), Polynesia, Wallis Island in the southwest Pacific (*Ducula david*), New Caledonia (*Gallicolumba longitarsus* and *Caloenas canacorum*), Henderson Island in the Pitcairns (*Ducula harrisoni*) and many other islands.

Typically, the most dramatic increase in body size is shown by birds of prey (Accipitriformes) and owls (Strigiformes). The largest eagle known to have existed was the New Zealand endemic Haast's eagle, *Harpagornis moorei*, that reached 10–15 kg of weight and had a wingspan of 2–3 m (Bunce et al., 2005). Other Accipitriformes of increased size have been reported, including species from the Quaternary of the West Indies (Suárez and Olson, 2007). In addition, this region harbored a number of large-sized owls such as the titanic *Ornimegalonyx oteroi* (Cuba), *Bubo osvaldoi* (Cuba), and several species of *Tyto* (Puerto Rico, Cuba, Hispaniola, Barbuda, Antigua). During the Quaternary, other endemic representatives of this order occurred in

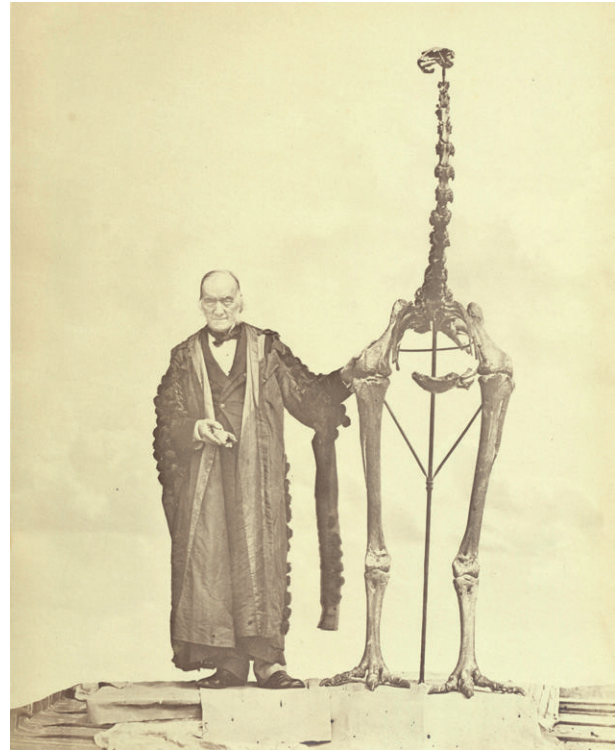


Figure 28 Sir Richard Owen standing next to the giant moa *Dinornis giganteus*. Modified from Owen R (1879) *Memoirs on the extinct wingless birds of New Zealand*. London: John van Voorst.

Sicily (*Tyto mouerchauvireae*, *Aegolius martaie*, *Athene trinacriae*) (Pavia, 2008), the Balearic Islands (*Tyto balearica*), Corsica and Sardinia (*Athene angelis*, *Tyto balearica cyrneichmusae*, *Bubo insularis*), Flores (*Tyto* sp.), Madeira (*Tyto* sp.), Hawaii (*Grallistrix*), Crete and Armathia near Greece (*Athene cretensis*).

Cases of gigantism can also be found in a number of different orders of birds such as Anseriformes (ducks, geese, and swans) (including *Cnemiornis* and *Cygnus sumnerensis* from New Zealand and *Cygnus falconeri* from Sicily), Gruiformes (coots, cranes, and rails) (such as *Aptornis* from New Zealand and *Grus cubensis* from Cuba), Galliformes (chickens, quail, ptarmigan, pheasants) (such as *Sylviornis neocaledoniae* from New Caledonia and *Megavitiornis altirostris* from Fiji), and Ciconiiformes (such as the flightless ibis, Threskiornithidae, of Jamaica and the giant marabou stork, *Leptoptilos*, of Flores and Java).

All in all, giant insular bird species are scattered over a highly variable set of islands, including continental shelf, volcanic, and continental plate islands. Although it is highly likely that their release from mammalian competition and predation has promoted gigantism associated with flightlessness in these forms, the extent to which this explanation works at all taxonomic levels is still debated (Meiri, 2007).

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See also: [Vertebrate Overview](#). [Vertebrate Studies: Speciation and Evolutionary Trends in Quaternary Vertebrates](#).

References

- Abbazzi L, Angelone C, Arca M, et al. (2004) Plio-Pleistocene fossil vertebrates of Monte Tuttavista (Orosei, eastern Sardinia, Italy), an overview. *Rivista Italiana di Paleontologia e Stratigrafia* 110: 603–628.
- Agenbroad LD (2012) Giants and pygmies: Mammoths of Santa Rosa island, California (USA). *Quaternary International* 255: 2–8.
- Aiello LC (2010) Five years of *Homo floresiensis*. *American Journal of Physical Anthropology* 142: 167–179.
- Angelone C, Tuveri C, Arca M, López Martínez N, and Kotsakis T (2008) Evolution of *Prolagus sardus* (Ochotonidae, Lagomorpha) in the Quaternary of Sardinia Island (Italy). *Quaternary International* 182: 109–115.
- Benton MJ, Csiki Z, Grigorescu D, et al. (2010) Dinosaurs and the island rule: The dwarfed dinosaurs from Hateg Island. *Palaeogeography, Palaeoclimatology, Palaeoecology* 293: 438–454.
- Bover P, Quintana J, and Alcover JA (2010) A new species of *Myotragus* Bate, 1909 (Artiodactyla, Caprinae) from the Early Pliocene of Mallorca (Balearic Islands, western Mediterranean). *Geological Magazine* 147: 871–885.
- Bunce M, Szulkin M, Lerner HRL, et al. (2005) Ancient DNA provides new insights into the evolutionary history of New Zealand's extinct giant eagle. *PLoS Biology* 3(1): e9. <http://dx.doi.org/10.1371/journal.pbio.0030009>.
- Bunce M, Worthy TH, Ford T, et al. (2003) Extreme reversed sexual size dimorphism in the extinct New Zealand moa *Dinornis*. *Nature* 425: 172–174.
- Caloi L and Palombo MR (1995) Functional aspects and ecological implications in Pleistocene endemic cervids of Sardinia, Sicily and Crete. *Geobios* 28: 247–258.
- Darwin C (1859) *On the Origin of Species by Means of Natural Selection, or the Preservation of Favoured Races in the Struggle for Life*. London: John Murray.
- de Vos J (1996) Taxonomy, ancestry and speciation of the endemic Pleistocene deer of Crete compared with the taxonomy, ancestry and speciation of Darwin's Finches. In: Reese DS (ed.) *Pleistocene and Holocene Fauna of Crete and Its First Settlers*, *Monographs in World Archaeology*, vol. 28, 111–124.
- Foster JB (1964) Evolution of mammals on islands. *Nature* 202: 234–236.
- Fry BG, Wroe S, Teeuwisse W, et al. (2009) A central role for venom in predation by *Varanus komodoensis* (Komodo Dragon) and the extinct giant *Varanus (Megalania) priscus*. *Proceedings of the National Academy of Sciences* 106: 8969–8974.
- Hocknull SA, Piper PJ, van den Bergh GD, Awe Due R, Morwood MJ, and Kurniawan I (2009) Dragon's paradise lost: Palaeobiogeography, evolution and extinction of the largest-ever terrestrial lizards (Varanidae). *PLoS One* 4: 1–15.
- Jaffe AL, Slater GJ, and Alfaro ME (2011) The evolution of island gigantism and body size variation in tortoises and turtles. *Biology Letters* 7: 558–561.
- Kobayashi Y, Tomida Y, Kamei T, and Eguchi T (2006) Anatomy of a Japanese tomistomine crocodylian, *Toyotamaphimeia machikanensis* (Kamei et Matsumoto, 1965) from the Middle Pleistocene of Osaka prefecture: The reassessment of its phylogenetic status within Crocodylia. *National Science Museum Monographs* 35: 121.
- Köhler M, Moyà-Solà S, and Wrangham RW (2008) Island rule cannot be broken. *Trends in Ecology & Evolution* 23: 6–7.
- Lomolino MV, Sax DF, Riddle BR, and Brown JH (2006) The island rule and a research agenda for studying ecogeographical patterns. *Journal of Biogeography* 33: 1503–1510.
- Meijer HJM and Due RA (2010) A new species of giant marabou stork (Aves: Ciconiiformes) from the Pleistocene of Liang Bua, Flores (Indonesia). *Zoological Journal of the Linnean Society* 160: 707–724.
- Meiri S (2007) Size evolution in island lizards. *Global Ecology and Biogeography* 16: 702–708.
- Meiri S, Cooper N, and Purvis A (2008) The island rule: Made to be broken? *Proceedings of the Royal Society B* 275: 141–148.
- Novelli M and Palombo MR (2007) "Hunter Schreger Bands" in *Cynotherium sardous* Studati, 1857 from Dragonara cave (Late Pleistocene, North-Western Sardinia). In: Coccioni R and Marsili A (eds.) *Proceedings of the Giornate di Paleontologia 2005*, *Grzybowski Foundation Special Publication*, vol. 12, pp. 61–71.
- Palombo MR (2005) Food habit of "*Praemegaceros*" cazioti (Dehaut, 1897) from Dragonara cave (NW Sardinia, Italy) inferred by cranial morphology and dental wears. *Monografies de la Societat d'Història Natural de les Balears* 12: 233–244.
- Palombo MR (2007) How can endemic proboscideans help us understand "island rule"? A case study of Mediterranean islands. *Quaternary International* 169–170: 105–124.
- Palombo MR (2009) Body size structure of the Pleistocene Mammalian communities from Mediterranean Islands: What support for the "Island Rule"? *Integrative Zoology* 4: 341–356.
- Palombo MR (2010) Elephants in miniature. In: Meller H (ed.) *Elefantenreich-Eine Fossilwelt in Europa. Katalog zur Sonderausstellung im Landesmuseum für Vorgeschichte*, pp. 275–292. Landesmuseum Sachsen-Anhalt: Halle.
- Palombo MR, Bover P, Valli AFM, and Alcover JA (2006) The Plio-Pleistocene endemic bovids from the Western Mediterranean islands: Knowledge, problems and perspectives. *Hellenic Journal of Geosciences* 41: 153–162.
- Pavia M (2008) The evolution dynamics of the Strigiformes in the Mediterranean islands with the description of *Aegolius martaie* n. sp. (Aves, Strigidae). *Quaternary International* 182: 80–89.
- Rozzi R (2010) I bovidi endemici del Pleistocene inferiore della Sardegna: plesiomorfie ed apomorfie nello scheletro postcraniale. Unpublished dissertation. Sapienza Università di Roma.
- Suárez W and Olson SL (2007) The Cuban fossil eagle *Aquila borrasii* Arredondo: A scaled-up version of the great black-hawk *Buteogallus urubitinga* (Gmelin). *Journal of Raptor Research* 41: 288–298.
- van der Bergh GD, de Vos J, and Sondaar PY (2001) The Late Quaternary palaeogeography of mammal evolution in the Indonesian Archipelago. *Palaeogeography, Palaeoclimatology, Palaeoecology* 171: 385–408.
- van der Geer A, Lyras G, de Vos J, and Dermitzakis M (2010) *Evolution of Island Mammals: Adaptation and Extinction of Placental Mammals on Islands*. Oxford: Wiley-Blackwell.
- Van Valen LM (1973) Pattern and the balance of nature. *Evolutionary Theory* 1: 31–49.
- Welch JJ (2009) Testing the island rule: Primates as a case study. *Proceedings of the Royal Society B* 276: 675–682.
- Willemsen GF (2006) *Megalenhydrys* and its relationship to *Lutra* reconsidered. *Hellenic Journal of Geosciences* 41: 83–87.

Interactions with Hominids

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Glossary

Aurignacian: First Upper Paleolithic industry extending over much of Europe; dates range from ca. 39 000 to 28 000 BC. Characterized by blade technology and the development of a significant formal bone tool industry, it is named after the site of Aurignac (Haute Garonne, France).

Bone-Isotope Analysis: Analysis of stable isotopes such as ^2H , ^{13}C , and ^{15}N in bone from archaeological sites, in an attempt to determine the nature of the diet of the human or animal concerned.

Magdalenian: An Upper Paleolithic industry found in Western Europe. Named after the type-site of La Madeleine in the Dordogne département in France, it is characterized by blade industries, geometric microliths, and an elaborate bone tool technology, including harpoons, needles and spear points. Dated to between approximately 18 000 to

11 000 BP, the most frequently occurring large game species represented at sites are reindeer and horse, with ibex and red deer in upland areas.

Solutrean: An Upper Paleolithic industry found in France and Spain. Named after the type-site of Solutré in the Saône-et-Loire département in France it is characterised by finely flaked, symmetrical, unifacial and bifacial leaf points which reflect a high level of craftsmanship. Dated to between approximately 17 000 to 21 000 BP, the most frequently occurring large game species represented at sites are reindeer and horse.

Scavenging: Procurement of animal products (usually meat or bone marrow and grease) from carcasses of animals already dead, that is, not killed by the agent scavenging.

Taphonomy: The study of the processes through which organic material becomes fossilized.

Introduction

As recently as the 1970s, early humans were assumed to have been efficient hunters. They were regarded as hunter-gatherers concentrating largely on the taking of large game, spending little time on smaller animals and plants. The very idea of humans as scavengers, for example, let alone prey as opposed to predators, was almost anathema to the archaeological world. The situation has changed, however, as we now realize that the human–environment relationship is a complex and ever-changing one.

Interactions between ‘early humans’ (*Homo habilis* to *Homo sapiens*) and animals are today recognized as taking several forms, both economic and social/ideological. Within the economic realm, humans may be predator, prey, and competitors with the local carnivore population for faunal resources and shelter. Within a nonsubsistence framework, animals featured in belief systems. The occurrence of animal forms in both the rock and mobiliary art of later Paleolithic Europe, Asia, and Africa tells us that the relationship between hunter and prey was a complex one operating at an ideological as well as economic level.

Diet and Subsistence

Pre-Holocene human populations depended on wild resources, practicing a hunting, scavenging, fishing, and gathering lifestyle, interacting constantly with the faunal resource base around them. Bone-isotope analysis has firmly established that Neanderthals were top-level carnivores, consuming such a significant amount of meat in their diet (comparable or

even in excess of contemporary carnivores such as wolf) that it is extremely unlikely this was acquired through scavenging (Bocherens et al., 1991). In the majority of cases, Paleolithic hominins were ‘obligate’ hunters. Thus, their economic/dietary relationship with animals was a close one, which was usually unidirectional. Only the earliest hominins are regarded as having regularly been the prey or ‘the hunted’ (Brain, 1981) rather than the predator or ‘hunter,’ although later they were competing for resources, and it is here that we will start.

The Lower Paleolithic

During the Lower Paleolithic/early Stone Age (see 2.7 Myr–300 000 Years Ago in Africa, 1.9 Myr–300 000 Years Ago in Europe, 2.7 Myr–300 000 Years Ago in Asia), the picture reconstructed is often one of humans hunting only on a sporadic or opportunistic chance-encounter basis. Game was killed as and when encountered and little planning was involved. Hunting usually supplemented or was supplemented by frequent scavenging of large game. These ‘early humans’ of the Lower Paleolithic are frequently considered not to have been hunters but instead to have scavenged from carcasses left by other large carnivores (e.g., wolf, lion, and hyena) or those of animals which died ‘naturally.’ For example, at sites such as FLK Zinj in the Olduvai Gorge of Tanzania, Binford concluded that early humans scavenged rather than hunted (Binford, 1988). Instead of having access to complete carcasses, they picked over what was left behind by hyenas or leopards. Bunn and Kroll (1986), meanwhile, describe the same site as one to which hominids transported carcasses or bits of carcasses from elsewhere – irrespective of whether the latter had been obtained by hunting or scavenging. Once carcass segments had been brought to the

site, human activity was quite intense, for there is evidence of skinning and carcass dismemberment as well as bone marrow processing. According to [Blumenshine and Marean \(1993\)](#), however, the picture is more complex, being an amalgam of human scavenging and butchery, plus removal of selected body parts from the site by carnivores.

Similarly, [Echassoux \(2004\)](#), in her study of the fauna at the Lower Pleistocene site of Le Vallonet (Alpes-Maritimes, France), identifies evidence of human activity as minimal and sporadic; the site is one at which more evidence is seen of carnivore and rodent activity. This is clearly a site at which humans and carnivores each played a role in assemblage formation. However, the absence of butchery marks on carnivore remains (other than the bear, which may have been hibernating at the site) suggests that human and other carnivores did not 'interact' as predator and prey ([Echassoux, 2004](#), p. 48). Instead, both the hominids and carnivores took advantage of both scavenging and hunting opportunities which were presented to them. Meanwhile in the Levant, Ulbeidiya formation assemblages I-15LF/1-16 and II-23 are attributed by [Gaudzinski \(2004\)](#) to hunting rather than scavenging – illustrating the remarkable diversity in subsistence strategies available to very early humans.

Similarly, at Iserna-la-Pineta (Molise, South Italy), it has been suggested that the mixed faunal assemblage, with cervids, hippopotamus, rhinoceros, elephant, and bear, results from chance-encounter hunting supplemented by human scavenging of large game. The presence of marrow bone fractures and cut marks on long bones links the assemblage to human activity ([Anconetani, 1999](#)), while the fact that more than half the material is bison (*Bison schoetensacki*) has been attributed to early specialization. Fauna from Caune de l'Arago in southern France (Pyrénées-Orientales) is also diverse and dates to between 700 and 100 ka ([de Lumley et al., 2004](#)). The site has been identified as a long-term home base, a temporary seasonal site, a hunting stand, and a 'bivouac' or short-term shelter. Furthermore, hunting at the site during each of these occupation phases is slightly different. For example, [de Lumley et al. \(2004\)](#) ascribe specialist hunting to the short-term occupation phases of the site, while longer-term and seasonal occupations see a more generalized strategy. In Britain, the early middle Pleistocene site of Boxgrove (West Sussex), dated to ca. 500 ka, has yielded diverse evidence of exploitation of Etruscan rhinoceros, horse, and red deer ([Roberts and Parfitt, 1999](#)). The rhinoceros appears to have been dispatched on the edge of a lagoon at the site and then systematically butchered, with full access to the prime parts of the carcass, whereas the horse bears evidence of an impact mark in the scapula, assumed to be from a spear. Such spears are rarely preserved in the archaeological record but notable examples are known from Clacton-on-Sea in England ([Oakley et al., 1977](#)) and Schöningen in Germany ([Thieme, 1997](#)), both ca. 400 ka ([Roberts, 1999](#)).

Finally, it should also be remembered that the Lower Paleolithic is a time when hominids were frequently prey as well as predator. For example, at locality 1 Zhoukoudian (China), of the 42 nondental hominid skeletal elements, 28 (67%) show hyenid tooth marks – gnawing, biting, chewing, punctures, even regurgitation (see [Figure 1](#)) ([Boaz et al., 2004](#)). Despite the carnivore marks observed, no such marks are recorded on the rhinoceros bones at the site. Instead, these display cut marks ([Tong, 2000](#)), implying that here early

humans are both predator and prey – evidence of the complexity of early human–animal interactions.

The Middle Paleolithic

Moving from the Lower to Middle Paleolithic, the nature of the early human–animal relationship becomes more complex as the faunal record changes (see [Late Pleistocene of Africa](#)).

It has frequently been suggested that during the Middle Paleolithic, humans could not hunt large game – at least on a regular basis ([Binford, 1981, 1988](#)); instead, meat from species such as horse, bison, mammoth, and aurochs was obtained only by scavenging. At the Grotte Vaufray (Dordogne, France), where deposits date back to 300 ka BP, [Binford \(1988\)](#) sees at least three scenarios that explain the structure of the faunal record. Small or medium game found at the site results from carnivore rather than human activity. Large species such as horse result from scavenging by humans. Red deer, however, he attributes to a mixture of human and carnivore activity, the main carnivore being the dhole – a scavenger, sharing, with humans, access to game already killed by larger predators such as the wolf. Even an alternative viewpoint in which Neanderthals and dhole occupied the site sporadically at different times of year (Neanderthals in autumn, dhole in spring and summer) shows how complex, if perhaps less direct, the relationship remains ([Burgett, 1999](#)). The faunal material recovered at La Quina, however, results from both natural and human processes/activities. Accumulation of material by humans is indicated by marks resulting from butchery observed on the bones (see [Figure 2](#)). Subsequent damage and destruction by carnivores at the site point to scavenging of human kills by species such as hyena and wolf.

Chase invokes a picture of reindeer killed at or near the site, with carcasses prepared for transportation elsewhere and marrow consumed at the site ([Chase, 1999](#), p. 169). Occurrence of more adult than juvenile carcasses, abundance of medium adult or large carcasses, relative rarity of the 'less meaty' cranial/axial skeleton, and greater importance of cut marks on skull and distal long bones ([Chase et al., 1994](#), p. 298) can be attributed to hunting rather than scavenging at this Middle Paleolithic site. The scene in which the human forager is actively hunting is one which is repeated across much of Eurasia, most particularly at sites such as Salzgitter Lebenstedt (Germany) where systematic reindeer hunting is observed ([Gaudzinski and Roebroeks, 2000](#)).

When Curtis Marean looked in detail at the faunal assemblages associated with Neanderthals and early modern humans from Kober Cave (Zagros Mountains, Iran) and layer 10 at Die Kelders Cave I in South Africa, respectively, he found no reliable evidence for scavenging. Indeed, the patterns seen at both sites indicate hunting by the occupants of the sites in question ([Marean, 1998](#), pp. 132–133), followed, on occasion, by scavenging by carnivores – as indicated by carnivore tooth marks superimposed on cut marks at Die Kelders I (see [Figure 3](#)).

Mousterian humans in west-central Italy scavenged to a much greater degree than did their Upper Paleolithic counterparts, despite the fact that they were efficient hunters. [Stiner \(1994\)](#) sees a change in the importance of scavenging about 55 ka, before which many faunal assemblages, she

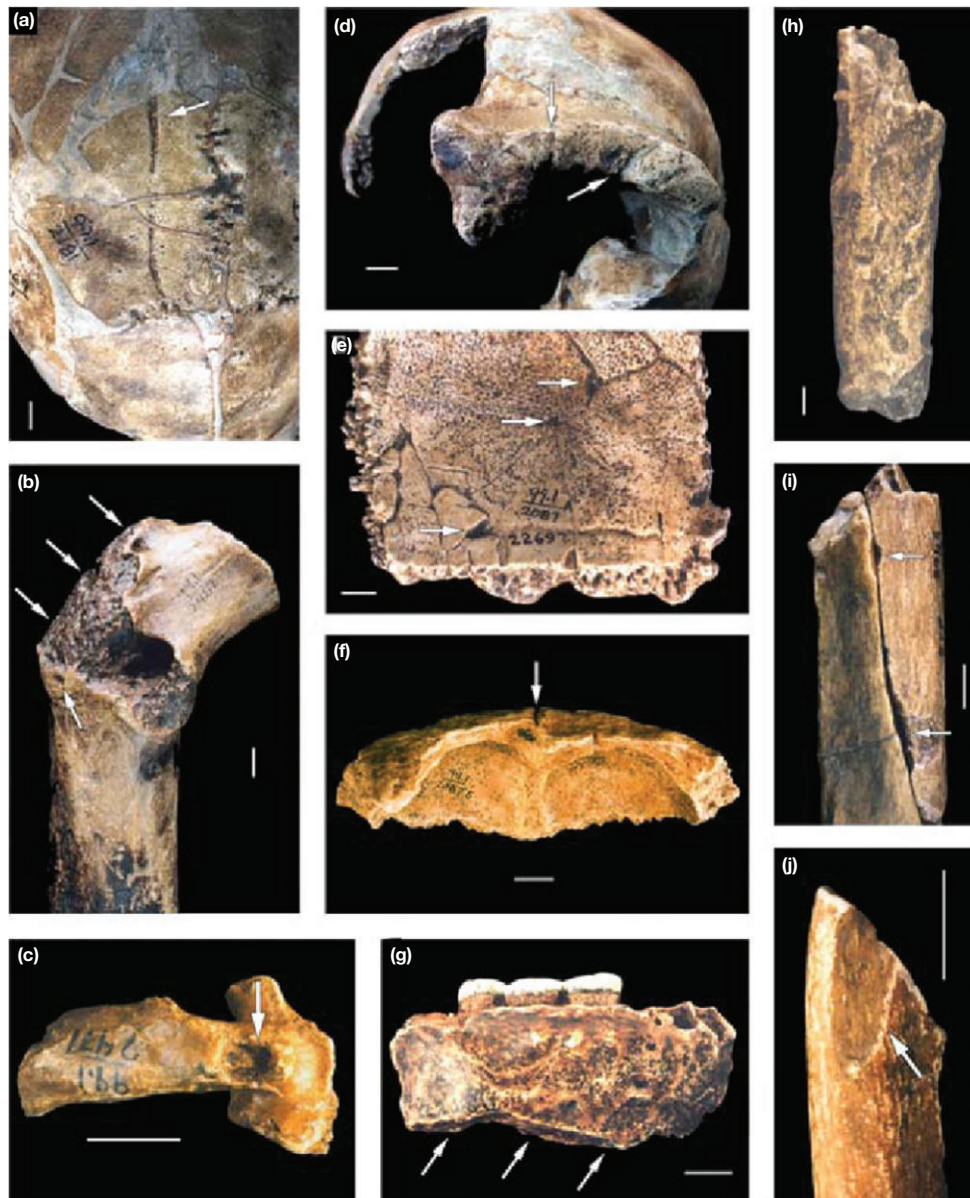


Figure 1 Carnivore (hyenid) damage on bones of *Homo erectus* from locality 1 Zhoukoudian. © Russell L. Ciochon, University of Iowa.

claims, resulted from scavenging. After this, hunting was the norm. Further west in southern France, this change is seen at least 15–20 ka earlier (Boyle, 1998), with scavenging only a secondary mode of subsistence during much of the Middle Paleolithic in Western Europe. Thus taphonomic study has shown us that although the difference between Middle and Upper Paleolithic hunters is small, they did not occupy the same resource niches.

The Upper Paleolithic

During the Upper Paleolithic (Eurasia), late Stone Age (Africa), and Paleoindian periods (America), an increasing degree of organization in, rather than complexity or type of, subsistence strategy is evident, especially during the second half of the period. Some equate this increased organization with a growing

degree of specialization, which Enloe (1999, p. 502) defines as ‘when a particular resource offers a better solution to the problems of year-round, long-term food acquisition than does a general mix of other potential food resources.’ This does not need to imply that only one resource was exploited – we are not, after all, talking about a form of monoculture, and other resources were surely also taken, if only to avoid the health problems, if not death, which result from dietary overspecialization. What it does suggest is that when it was necessary to choose between available resources, the advantages of exploiting just one or two probably outweighed those of taking others: the chosen resource(s) provided a yield greater than/or more appropriate to the needs of the hunter-gatherer at the time of selection. Hunter-gatherers are therefore rarely observed taking food in quantities which directly reflect natural abundance of food in the environment. Instead, obtaining food involves a number of

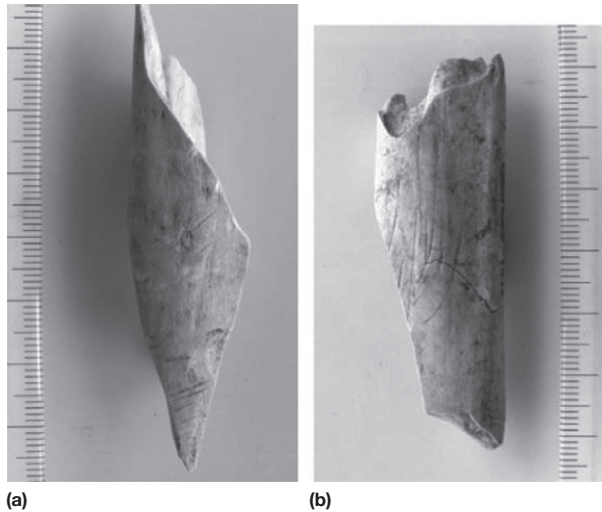


Figure 2 (a) Cut-marks on reindeer tibia from La Quina bed 8; (b) reindeer tibia shaft fragment used as a *retouche* – with two sets of marks. © P. Chase.



Figure 3 Photograph showing carnivore tooth marks superimposed on cut marks at Die Kelders I, indicative of initial human hunting and carcass processing followed by carnivore scavenging. Scale = 1 cm. © C. Marean.

factors, of which natural abundance is but one. Decision-making processes involve choosing between available resources, disregarding some, and perhaps concentrating on others for specific reasons at specific times. Choices depend not only on availability but also on variables such as:

1. labor costs (it may be necessary to disregard a herd of bison or even larger prey such as mammoth due to lack of manpower at the time of herd observation);
2. the type of yield sought (when seeking large skins, there is little point in hunting small or very small game such as rabbit and even less point in spending time at the riverbank catching salmon); and
3. the quality rather than quantity of game available.

Hence it is highly likely that hunters became expert at determining the condition of prey. In particular, procurement and use of fat from animal carcasses provide a way of coping with harsh conditions, especially at times of periodic hardship. Indeed, some species (e.g., reindeer and gazelle, typical resources of the period in Western Europe and the Middle East) are poorer in fat than others, so expanding the diet to include alternative

resources such as fish and birds increases fat intake and improves the overall quality of the diet. As every hunter-gatherer knows, fat meat tastes better, providing more energy than does lean meat, and so is more efficiently metabolized, storing both vitamins and essential acids (Fagan, 2004, p. 64).

At Upper Paleolithic sites in southwest France, this widening of the resource base is seen even when reindeer dominates faunal assemblages, in some cases almost to the exclusion of anything else. It becomes apparent because carcass processing of primary versus secondary species is clearly different. Secondary species are much more likely to be butchered intensively or very intensively – in order to increase carcass yield and dietary variety, especially when the species are rare or very rare. At Magdalenian sites such as La Madeleine, Limeuil, and Reignac, in the classic Vézère valley (Dordogne, France), reindeer dominates assemblages, and red deer and horse form distant seconds. However, the carcass-processing strategy adopted in exploiting these species appears to have been a much more intensive one than in the case of reindeer, with greater effort expended in order to procure greater yield (Boyle, 1990) when they were available. These resources ranked much higher in the Paleolithic diet in terms of desirability than did reindeer, and thus an element of ‘luxury’ featured in man–animal subsistence relationships. The hunter’s relationship with primary and secondary species is different. The predominant species, abundant either consistently or periodically in the local environment, forms the backbone of the economy but the relationship between humans and secondary species will probably be more intense.

Raw Material

Food is not the only thing which animals provide and which therefore features in man–animal interactions during the Paleolithic.

The existence of some fauna at sites may be attributed to their use as raw material, such as the mammoth bones and tusks at Milovice (Czech Republic), where two bone deposits have been interpreted as Gravettian dwellings probably using skins/pelts to form coverings (Svoboda et al., 1996; see Late Pleistocene Megafaunal Extinctions).

Skins and pelts were also used for clothing. Female figurines found, for example, at the Siberian site of Malta appear to be wearing hooded fur overalls (Poikalainen, 2001, p. 53; White, 2003). The presence of cut marks of the sort observed in Figure 4, in which we see marks close to the base of antlers from Combe Grenal (Dordogne, France), is indicative of skinning activity. The skins may have been used for shelter/roofing, clothing, or even containers for short-term storage and transport. One of the earliest-suggested European examples of the use of pelts in shelter construction is that of the Acheulian structure reconstructed at the Grotte du Lazaret and dating back some 400–500 ka (de Lumley, 1969), although it has also been suggested that the stone/bone configuration at the site results instead from natural taphonomic processes (see Villa 2004 and references therein), while trapping of Arctic fox, beaver, and other fur-bearing animals for pelts for the (multi)layered clothing is also possible.



Figure 4 Cut marks close to the base of antlers from Combe Grenal, possibly indicative of skinning activity. © P. Chase.

Bone and antler were also important raw materials throughout the Upper Paleolithic more so than during earlier periods despite their frequent use as soft hammers during knapping. Several bone tools are characteristic of the Aurignacian, including split-base bone points; eyed needles provide a tantalizing hint of stitched clothing during the Solutrean; Magdalenian Stages IV–VI are differentiated on the basis of barbed bone/antler harpoons, perhaps used for fishing. Analysis of material from the terminal Pleistocene (ca. 12.5–10.5 ka) at Niah Cave (Borneo) has shown that particular attention appears to have been paid to disarticulating Cercopithecidae (Old World monkeys) pelvic limbs (Figure 5) into upper and lower elements, a procedure which, although almost certainly reflecting preparation and consumption practices, may also reflect their use as raw material in bone tool production – a noted feature from this period at Niah. Analysis of the butchery further suggests that treatment of the bear cat (*Arctictis binturong*) at the site (see Figure 6) separates it from other species. This distinction points toward a hunter–prey interaction which is systematic to the point of taxonomic classification, so that procurement and carcasses-treatment processes distinguished it from other arboreal taxa at the site and even other civet cats. These studies, part of a forthcoming monograph on Niah Cave (Rabett and Piper, 2012), show that there are hitherto unseen components in early human–animal relationships of which we are only now beginning to catch a sight.

On a perhaps less functional level, bone flutes are also known – such as those from the Grotte d'Isturitz in the

Pyrenées Atlantiques of Southern France (White, 2003, p. 81) and from the Aurignacian level of Geissenklösterle (d'Errico et al., 2003, p. 41). Jewellery may also not have been essential in terms of subsistence and diet. Many of the personal ornaments of the Upper Paleolithic of Western Europe are made of animal material – bone, tooth, or ivory. The existence, in the Magdalenian burial at St.-Germain-la-Rivière (Gironde, France), (according to Vanhaeren and d'Errico), of numerous perforated red deer canines, a species almost completely absent from faunal assemblages in the surrounding region, suggests, according to Vanhaeren and d'Errico (2005), that they were prestige items of exotic origin resulting from long-distance trade. The presence of these, and of necklaces and pendants from sites such as Rocher de la Peine (Dordogne, France) (Figure 7) and Arcy-sur-Cure (Figure 8) emphasize an important aspect of the early human–animal relationship which may not only have been based on the tangible yield obtained, but may also have had a ideological or religious significance.

Ideology and Religion

As already suggested, the relationship between humans and secondary game species was frequently more intense and complex than that with the ever-available primary resources. This is particularly interesting when we consider that it is frequently these secondary species that form the bulk of the images of animals observed in Upper Paleolithic cave art. The main focus of the art is large and small ungulates and carnivores, with some portrayals of fish and birds; medium-sized ungulates (e.g., deer) are less frequently portrayed, despite comprising the larger portion of the faunal assemblages excavated.

In the three areas that form the main concentration of Paleolithic parietal and mobiliary art, Périgord-Quercy, French Pyrenees, and Franco-Cantabria, the relationship between faunal assemblage composition and artistic portrayal of animals does not, on initial inspection, appear to be a positive one.

At La Madeleine (Dordogne, France), for example, reindeer dominates the fauna but is secondary in the art; horse, on the other hand, dominates the art but is second in the fauna. At La Vache (Ariège, France), horse again dominates the art while ibex is the main faunal species, second in art. Meanwhile, the ibex, which dominates several Magdalenian faunal assemblages in Cantabria, is third in the art in the region (Straus, 1992). Altuna (1983) has also observed that at Cantabrian sites such as Ekain and Tito Bustillo, where horse and red deer dominate the art, faunal sample structure does not mirror artistic impressions. Instead, the importance of the large game species becomes apparent: Straus (1992, p. 180) views the larger bison (a species which is 4 times the size of red deer and frequently much more important in the art) as 'impressive beasts and sources of food bonanzas when procured,' while red deer was the main staple resource in the Altamira region – less spectacular but critical perhaps to day-to-day subsistence and survival. Following a similar argument, the greater size of horse over ibex may mean that the frequent rank reversal of these species, and horse over the reindeer, is more apparent than real.

But what does the art actually tell us about the nature of early human–animal relationships? Mithen has suggested that the purpose of Paleolithic art was one of information gathering,

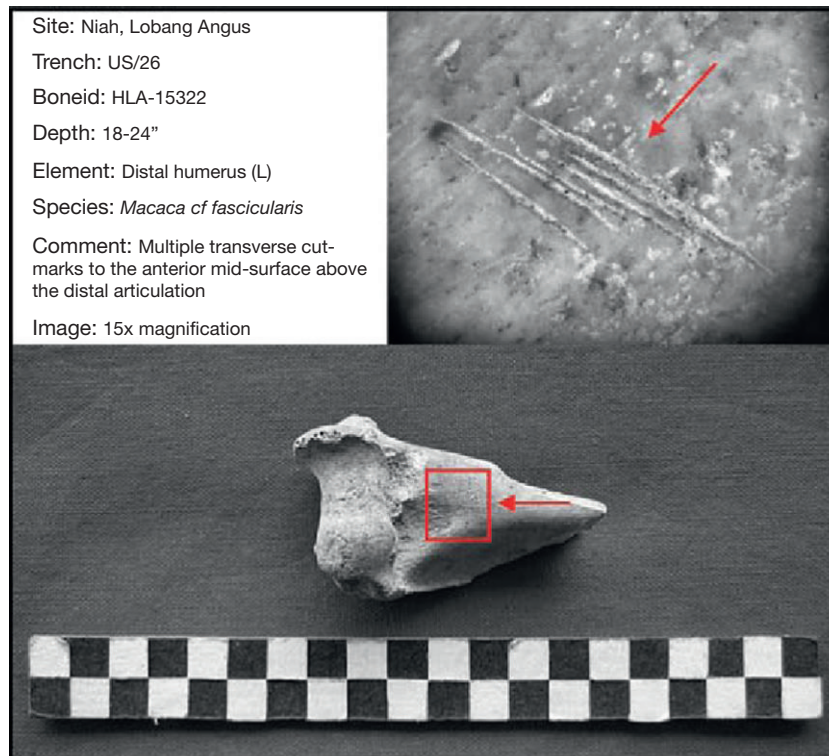


Figure 5 Cut marks resulting from the disarticulation of a Cercopithecidae pelvic limb into upper and lower elements from Lobang Aangus, Niah Cave, Sarawak. © R. Rabett (Harrisson Archive, Sarawak Museum).

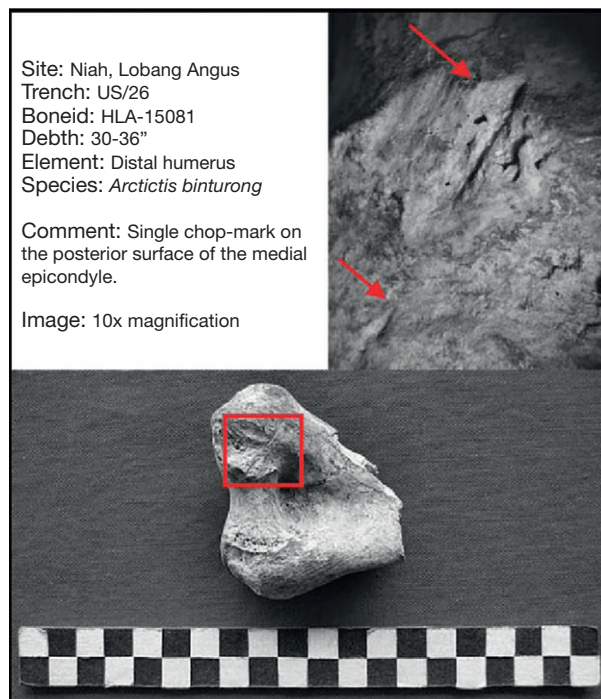


Figure 6 Cut marks on bones of bear cat (*Arctictis binturong*) from Lobang Aangus, Niah Cave, Sarawak. © R. Rabett (Harrisson Archive, Sarawak Museum).



Figure 7 A late Magdalenian necklace from the site of Rocher-de-la-Peine (Dordgoen, France), made of the teeth of bear, lion, fox and deer, and *Dentalium* shells (Photo, Randall White).

this information concerning the natural environment and available large game species in the local environment, more particularly the location and the condition of game (Mithen, 1990). He claims that the animals dominant in cave art (bovids and horse) were tracked individually, acting as backup species at times of economic hardship or at least when Cantabrian red deer or Périgord reindeer levels were low, perhaps seasonally. The way in which horse and bovids (in Cantabria at least) are portrayed provides, according to Mithen, information concerning methods of hunting the less frequently taken species – species which might have been taken so infrequently that even knowledge of hunting practices would not have been passed on readily from one generation to another. Thus, the pictures provide intergenerational information available at times of low game yield or economic stress. They are an indication of game species either available for exploitation when the painting was completed or possibly available in the future – an indication of potential resources.

According to González Sainz (1988), art ‘tracks’ hunting practices on both temporal and geographical axes. He relates the fall in frequencies of stags and hinds in mobiliary art, as well as an increase in ibex through the Magdalenian, to decreasing red deer hunting and the growing importance of ibex hunting as revealed by faunal assemblages.

Different hunting techniques are also portrayed. Kehoe (1990, 1999) has identified corrals, ‘extended corrals,’ and drive lines among scenes featured in Paleolithic cave art. For example, he identifies a horse drive on the south wall of the Axial Gallery at Lascaux (Dordogne) on the basis of a series of black dots leading to, and eventually touching, a rectangle, which he interprets as a corral and line of rocks around which horses are running. At the Grotte Villars, he sees a hunter facing a charging bison. At Trois Frères, in the Ariège, he has identified two images of bison drives (Kehoe, 1999, p. 253). At Altamira, meanwhile, based on the shape of the chamber itself, Kehoe considers the principal chamber to represent a corral. He considers the portrayal of both dead bison in the middle of the ‘painted ceiling’ and other bison facing outward, toward portrayed hunters, to be corroborative evidence (Kehoe, 1990). At the Turkish cave site of Öküzini, a limestone pebble, described by Marcel Otte and colleagues (Otte et al., 1995), is incised on both sides (see Figure 9), one of which bears an engraving which could indeed represent a corral and drive line. Such corrals and drives have frequently been observed and discussed in the ethnohistoric literature, most recently by Frison (2004), who, among several examples, describes a caribou ‘procurement complex,’ which included a line of stone cairns over a distance of 1 km which led the animals to a lake where they were killed.

Irrespective of anything else, the fact remains that the people who painted or engraved the images of Paleolithic art were hunters, or hunter-gatherers, dependent in a large measure on the wild game resources available in the surrounding environment (Straus, 1992, p. 187). The art reflects hunting practices, although the relationship between hunters and prey is often less direct than we might initially imagine, adding one more component to the complex set of interrelationships observed between early humans and the vertebrate faunas of the Paleolithic. Of course, whether these human hunters contributed greatly to the extinction of several Pleistocene faunal species, large and small, remains a hotly debated subject

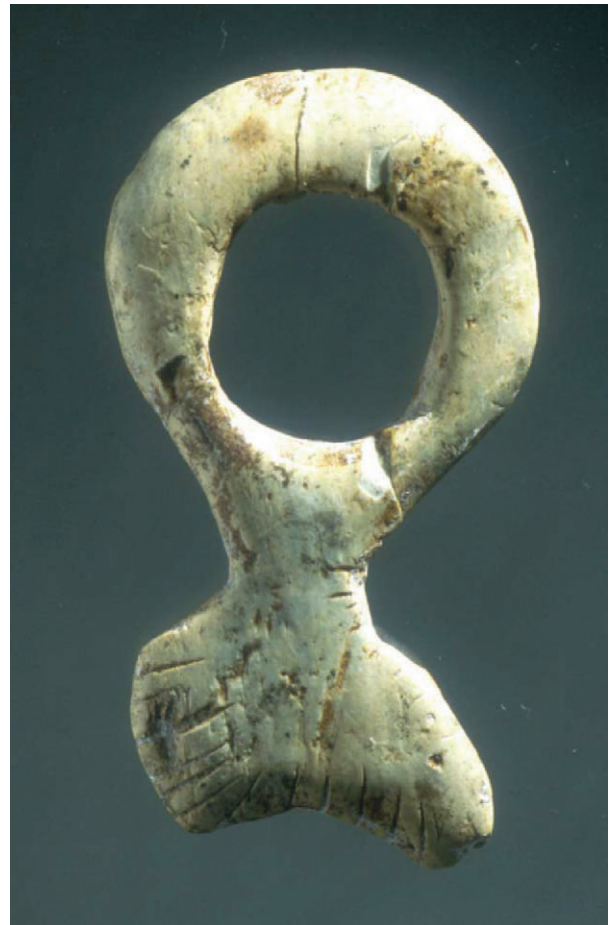


Figure 8 Aurignacian pendant made from mammoth ivory from Arcy-sur-Cure (Photo, Randall White).

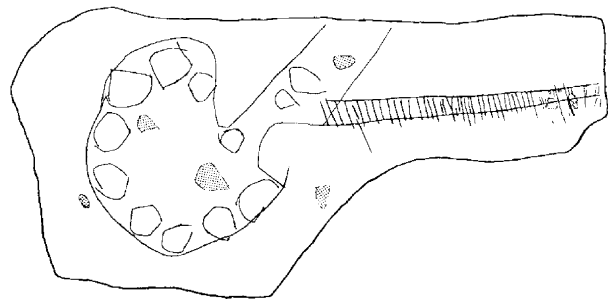


Figure 9 A limestone pebble from the Turkish cave site of Öküzini, incised on both sides. On one side there is a possible engraving of a corral and drive line. © M. Otte.

and is treated in more detail elsewhere (late Pleistocene Megafaunal Extinctions).

A Sign of Things to Come

Finally, recent analysis of dog mtDNA has suggested that the ancestral wolf was first domesticated (the first mammal to be so) as long ago as 15 ka BP in East Asia, with possible archaeological evidence known from Germany just 1 ka later and from Israel at 12 ka BP (Savolainen et al., 2002). It provides a

tantalizing hint of the ever-changing nature of the human-animal relationship of the final Pleistocene and offers a glimpse of the future as humans become steadily more dependent on a relatively small number of domestic animal resources. Of course, the degree to which all of the above can be said depends almost entirely on the archaeozoologist's ability to retrieve and understand useful data – such as the distribution of cut marks on bones – subjecting the data collected to a truly taphonomic approach, a methodology which has thus far raised as many questions as it has answered and will no doubt continue to do the same.

Acknowledgments

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See also: **Archaeological Records:** 1.9 Myr–300 000 Years Ago in Europe; 2.7 Myr–300 000 Years Ago in Africa; 2.7 Myr–300 000 Years Ago in Asia. **Vertebrate Records:** Late Pleistocene of Africa.

References

- Altuna J (1983) On the relationship between archaeofaunas and parietal art in the caves of the Cantabrian region. In: Clutton-Brock J and Grigson C (eds.) *Animals and Archaeology I*, pp. 227–238. Oxford: British Archaeological Reports S163.
- Anconetani P (1999) L'Assemblage faunique du gisement Paléolithique inférieur d'Isernia-La-Pineta (Molise, Italie) et l'exploitation du bison. In: Brugal JPH, David F, Enloe JG, and Jaubert J (eds.) *Le Bison: Gibier et Moyen de Subsistance des Hommes du Paléolithique aux Paléindiens des Grandes Plaines*, pp. 105–120. Antibes: Editions APDCA.
- Binford LR (1981) *Bones: Ancient Men and Modern Myths*. London: Academic Press.
- Binford LR (1988) Etude taphonomique des restes fauniques de la Grotte Vaufray, couche VIII. In: Rigaud JP (ed.) *La Grotte Vaufray: Paléoenvironnement, Chronologie, Activités Humaines*, pp. 535–564. Paris: Société Préhistorique Française.
- Blumenschine RJ and Marean CW (1993) A carnivore's view of archaeological bone assemblages. In: Hudson J (ed.) *From Bones to Behavior: Ethnoarchaeological and Experimental Contributions to the Interpretation of Faunal Remains*, pp. 273–300. Carbondale: Center for Archaeological Investigations, University of Southern Illinois Press.
- Boaz NT, Ciochon RL, Xu Q, and Liu J (2004) Mapping and taphonomic analysis of the *Homo erectus* loci at locality 1 Zhoukoudian, China. *Journal of Human Evolution* 46: 519–549.
- Bocherens H, Fizet C, Mariotte A, et al. (1991) Isotope biogeochemistry (^{13}C , ^{15}N) of fossil vertebrate collagen: Application to the study of a past food web including Neandertal man. *Journal of Human Evolution* 20: 481–492.
- Boyle KV (1990) *Upper Palaeolithic Faunas from South West France: A Zoogeographic Perspective*. Oxford: British Archaeological Reports.
- Boyle KV (1998) *The Middle Palaeolithic Geography of Southern France: Resources and Site Location*. Oxford: British Archaeological Reports.
- Brain CK (1981) *The Hunters or the Hunted? An Introduction to Africa Cave Taphonomy*. Chicago: University of Chicago Press.
- Bunn HT and Kroll EM (1986) Systematic butchery by Plio-Pleistocene hominids at Olduvai Gorge, Tanzania. *Current Anthropology* 27: 431–452.
- Burgett GR (1999) The bones of the beast: Actualistic taphonomic research and its role in understanding hominid exploitation of bison and other large mammals. In: Brugal JPH, David F, Enloe JG, and Jaubert J (eds.) *Le Bison: Gibier et Moyen de Subsistance des Hommes du Paléolithique aux Paléindiens des Grandes Plaines*, pp. 11–28. Antibes: Editions APDCA.
- Chase PG (1999) Bison in the context of complex utilization of fauna resources: A preliminary report on the Mousterian zooarchaeology of La Quina (Charente, France). In: Brugal JPH, David F, Enloe JG, and Jaubert J (eds.) *Le Bison: Gibier et Moyen de Subsistance des Hommes du Paléolithique aux Paléindiens des Grandes Plaines*, pp. 159–184. Antibes: Editions APDCA.
- Chase PG, Armand D, Debénath A, Dibble H, and Jelinek AJ (1994) Taphonomy and zooarchaeology of a Mousterian faunal assemblage from La Quina, Charente, France. *Journal of Field Archaeology* 21: 289–305.
- de Lumley H, Grégoire S, Barsky D, et al. (2004) Habitat et mode de vie des chasseurs paléolithiques de la Caune de l'Arago (600 000–400 000 ans). *L'Anthropologie* 108: 159–184, Paris.
- D'Errico F, Hensilwood C, Lawson G, et al. (2003) Archaeological Evidence for the Emergence of Language, Symbolism, and Music: An Alternative Multidisciplinary Perspective. *Journal of World Prehistory* 17: 1–70.
- Echassoux A (2004) Etude taphonomique, paléocécologique et archéozoologique des faunes de grands mammifères de la seconde moitié du Pléistocène inférieur de la Grotte du Vallonet (Roquebrune-Cap-Martin, Alpes-Maritimes, France). *L'Anthropologie* 108: 11–53.
- Enloe JG (1999) Hunting specialization: Single-species focus and human adaptation. In: Brugal JPH, David F, Enloe J, and Jaubert J (eds.) *Le Bison: Gibier et Moyen de Subsistance des Hommes du Paléolithique aux Paléindiens des Grandes Plaines*, pp. 501–509. Antibes: Editions APDCA.
- Fagan B (2004) *The Long Summer. How Climate Changed Civilization*. London: Granta Publications.
- Frison GC (2004) *Survival by Hunting. Prehistoric Human Predators and Animal Prey*. Berkeley: University of California Press.
- Gaudzinski S (2004) Subsistence patterns of Early Pleistocene hominids in the Levant – Taphonomic evidence from the 'Ubeidiya Formation (Israel). *Journal of Archaeological Science* 31: 65–75.
- Gaudzinski S and Roebroeks W (2000) Adults only. Reindeer hunting at the Middle Palaeolithic site of Salzitter Lebenstedt, Northern Germany. *Journal of Human Evolution* 33: 497–521.
- González Sainz C (1988) Le fait artistique à la fin du paléolithique. *Bulletin de la Société Préhistorique de l'Ariège* 43: 35–62.
- Kehoe T (1990) Corraling: Evidence from Upper Paleolithic cave art. In: Davis LB and Reeves BOK (eds.) *Hunters of the Recent Past*, pp. 34–46. London: Unwin Hyman.
- Kehoe T (1999) Subsistence and beyond in Upper Paleolithic bison economy. In: Brugal JPH, David F, Enloe JG, and Jaubert J (eds.) *Le Bison: Gibier et Moyen de Subsistance des Hommes du Paléolithique aux Paléindiens des Grandes Plaines*, pp. 249–260. Antibes: Editions APDCA.
- Marean C (1998) A critique of the evidence for scavenging by Neandertals and early modern humans: New data from Kobeh Cave (Zagros Mountains, Iran) and Die Kelders Cave I Layer 10 (South Africa). *Journal Of Human Evolution* 35: 111–136.
- Mithen S (1990) *Thoughtful Foragers. A Study of Prehistoric Decision Making*. Cambridge: Cambridge University Press.
- Oakley K, Andrews P, Keeley LH, and Clark JD (1977) A reappraisal of the Clacton spearpoint. *Proceedings of the Prehistoric Society* 43: 13–30.
- Otte M, Yalcinkaya I, Leotard JM, et al. (1995) The epi-Palaeolithic of Öküzini cave (SW Anatolia) and its mobiliary art. *Antiquity* 69: 931–944.
- Poikalainen V (2001) Palaeolithic art from the Danube to Lake Baikal. *Folklore* 18/19: 6–16. <http://haldjas.folklore.ee/folklore/vol18/paleoart.pdf>.
- Rabett RJ and Piper PJ (2012) The emergence of bone technologies at the end of the Pleistocene in Southeast Asia: regional and evolutionary implications. *Cambridge Archaeological Journal* 22: 37–56.
- Roberts MB and Parfitt SA (1999) *Boxgrove, a Middle Pleistocene hominid site at Eartham Quarry, Boxgrove, West Sussex*. London: English Heritage Archaeological Report no. 17.
- Savolainen P, Zhang YA-P, Luo J, Lundberg J, and Leitner T (2002) Genetic evidence for an East Asian origin of domestic dogs. *Science* 298: 1610–1613.
- Stiner M (1994) *Honor Among Thieves: A Zooarchaeological Study of Neanderthal Ecology*. Princeton, NJ: Princeton University Press.
- Straus L (1992) *Iberia before the Iberians: The Stone Age Prehistory of Cantabrian Spain*. Albuquerque: University of New Mexico Press.
- Svoboda J, Lozek V, and Vlcek E (1996) *Hunters between East and West. The Paleolithic of Moravia*. New York: Plenum.
- Thieme H (1997) Lower Palaeolithic hunting spears from Germany. *Nature* 385: 807.
- Tong H (2000) Les rhinocéros des sites à fossiles humains de Chine. *L'Anthropologie* 104: 523–529.
- Vanhaeren M and D'Errico F (2005) Grave goods from the Saint-Germain-la-Rivière burial: Evidence for social inequality in the Upper Palaeolithic. *Journal of Anthropological Archaeology* 24: 117–134.
- Villa P (2004) Taphonomie et stratigraphie dans la préhistoire européenne. *Les Nouvelles de l'Archéologie* 95: 13–18.
- White R (2003) *Prehistoric Art, the Symbolic Journey of Humankind*. New York: Harry N Abrams.

Volcanic Ash *see* **Quaternary Stratigraphy:** Tephrochronology

Vulcanism *see* **Glacial Climates:** Volcanic and Solar Forcing; **Glacial Landforms:** Quaternary Vulcanism: Subglacial Landforms; **Quaternary Stratigraphy:** Tephrochronology

W

Wind-Blown Sediments (Loess, Sand Dunes) *see* Eolian Records, Deep-Sea Sediments; Loess Deposits: Origins and Properties; **Dune Fields:** High Latitudes; Mid-Latitudes; Low Latitudes; **Loess Records:** Central Asia; China; Europe; North America; South America

Y

Younger Dryas Oscillation *see* **Paleoclimate Reconstruction:** Younger Dryas Oscillation, Global Evidence

INDEX

This index is in letter-by-letter order, whereby hyphens and spaces within index headings are ignored in the alphabetization, and it is arranged in set-out style, with a maximum of four levels of heading. Location references refer to the volume number, in bold, followed by the page number. Page numbers suffixed by *f* indicate figures, *t* indicate tables, *b* indicate boxes, and *ge* indicate glossary entries. **Bold locators** refers to Articles.

0–9, and Symbols

‰C_{37.4} salinity proxy, 2:778, 2:778*t*
 3–0.1 ka neoglaciation, 2:220
 8.2 ka event, Europe, 2:551–552, 2:553*f*
¹³C *see* carbon-13
¹⁴C *see* carbon-14; radiocarbon...

A

AAA *see* acid–alkali–acid pretreatment
 AABW *see* Antarctic Bottom Water
 AAR *see* accumulation area ratio; amino acid racemization
Abies (fir), 4:185–186
 ablation, 1:909–917
 ablation zone, 1:909*ge*
 Aborigines, Australia, 1:113, 1:114, 4:112
 ABOX *see* acid-base oxidation pretreatment
 abrasion
 micro/macroscale erosional forms, 1:865–867, 1:868*f*, 1:869, 1:871–872
 particle forms, sediments, 2:3
 sea level indicators, 4:377
 abrasion notches, 4:377
 abrupt climate change/events
 Andean glaciers, 2:392
 fluvial environment responses, 1:676–683
 ice core studies, 2:284, 2:308, 2:392
 New England vegetation history, 4:121
 North Atlantic, 3:49
 thermal diffusion paleotemperature records, 2:433
 varved sediment applications, 4:588
 see also rapid environmental change;
 Younger Dryas climate event
 absolute age, amino-acid dating, 1:45–46
 absorption nuclear interaction rates, 1:429–430
 absorption spectra, photosynthetic pigments, 3:328*f*
 abuse cases, forensics, 4:548
 Acari *see* oribatid mites
 accelerator mass spectrometry (AMS), 4:316–323

0.5–1 MV AMS, 4:321–322, 4:322*f*
 3 MV AMS, 4:317–320, 4:317*f*
 200kV machines, 4:322
 accelerator tank, 4:318, 4:319*f*
 blank corrections, 4:321
 chemical sample preparation, 4:320–321
 components of 3 MV AMS, 4:317–320
 contamination removal, 4:320–321
 dating techniques, 1:448–449, 1:449*f*
 error calculations, 4:321
 injection magnet, 4:318, 4:319*f*
 ion source operation, 4:317, 4:318*f*
 new trends, 4:321–323
 operation examples, 4:316–319
 results calculations, 4:321
 ‘single stage’ (0.3 MV) AMS, 4:322–323, 4:322*f*, 4:323*f*
 smaller machines, 4:321–323
 accommodation space, sedimentary indicators, 4:389
 acculturation, 1:151
 see also cultural...
 accumulation
 cosmogenic radionuclides, 2:358
 definition, 1:909*ge*
 ELA, glacial landforms, 1:909–917
 rate reconstructions, 2:358
 accumulation area ratio (AAR), 1:909*ge*, 1:914–916
 accumulation zone, 1:909*ge*
 accuracy, radiocarbon dating, 4:325, 4:325*f*
 Acheulean Industry, 1:59, 1:61*f*, 1:62–63, 1:63–65
 Asia after 300 ka, 1:98
 definition, 1:83*ge*
 Hunsgi, India, 1:77*f*
 Isampur site, Asia, 1:81*f*
 ‘Ubeidiya, Israel, 1:73*f*
 acid-alkali-acid (AAA) pretreatment, charcoal, 4:356–357
 acid-base oxidation (ABOX) pretreatment, charcoal, 4:357
 acid decomposition, diatom silica, 2:735
 acid rain, 3:320, 3:339–340
 acidification of lakes
 chemistry, 3:292–293

chironomid effects, 1:359
 paleolimnological multiproxy approaches, 3:339–341, 3:341–342
 pigment studies, 3:334–335
 ACR *see* Antarctic cold reversal
Acroperus harpae (cladocera), 3:273*f*
Acropora palmata coral records, 4:413–414, 4:414–416
 active-form frost mounds, 3:472–480
 active layer
 blockfield origin, 3:528
 climate change, 3:426–427, 3:470
 definition, 3:422
 detachments, 3:484–485, 3:579, 3:580*f*
 permafrost, 3:421–429, 3:464, 3:574, 3:577*f*
 climate change, 3:470
 detachments, 3:484–485, 3:579, 3:580*f*
 physical processes, 3:422–424
 processes, 3:421–429
 thickness, 3:424–426, 3:427
 transition zone, 3:426
 variability, 3:425–426
 active-layer thickness (ALT), 3:424–426, 3:427
 active rock glaciers, 3:536*f*, 3:538–539
 active temperature glacial landsystems, 1:813–814, 1:817–820
 active transport, sediments, 2:3–5
 adaptations, human, 1:51–53, 1:154–159, 3:630, 3:633–634, 3:634–636
 adaptive radiations, 1:51–53
 additive-dose method, OSL dating, 2:656
 additive dose with thermal-transfer correction (ADTT), 2:656–657, 2:657*f*, 2:662, 3:321
 adipose tissue, 1:319–320
 Adirondack Park, New York, 3:320
 aDNA *see* ancient DNA
 ADTT *see* additive dose with thermal-transfer correction
 advance of rock glaciers, 3:537–538
 advanced degree programs, forensics, 4:546
 advection, 2:365, 2:918
 adventitious tree roots, 2:99–100, 2:100*f*
 aeolian... *see* eolian...
 aerophilic/aerophilous diatoms, 1:475, 1:524

- aerosol effects, atmospheric dust, 1:729, 1:730f, 1:734–735
- Aerosol Index, 3:415f
- aerosolic dust *see* long-range-transported dust
- aerosols
- biogeochemical role in climate cycles, 3:415f
 - heavy metal pollution, 2:344
 - ice cores, 2:326–327, 2:328–329, 2:329–330, 2:390
 - microparticles and, 2:342–343
 - tephrochronology, 4:281–283
 - see also* dust
- Africa
- 2.7 MYR–300,000 years ago, 1:59–66
 - 300,000–8000 years ago, 1:91–97
 - chironomid paleoecology, 1:361–372
 - Early Quaternary, 2:168
 - environmental change, 1:362–368
 - Great Lakes, 2:501–502, 2:502f
 - history of glaciation, 2:149
 - hominin movement
 - Asia, 1:67, 1:70–71
 - Australia, 1:111–112, 1:112–113
 - global expansion, 1:91–97
 - human evolution, 3:192f
 - ice core records, 2:373–378
 - lakes
 - diatom records, 1:546–547, 1:551–552
 - Late Quaternary lake levels, 2:499–505
 - Late Pleistocene
 - pollen records, 4:9–17
 - vertebrate studies, 4:664–672
 - megafaunal extinctions, 4:707, 4:707t
 - Middle Pleistocene vertebrate studies, 4:615–619
 - monsoon data-model comparisons, 3:139f, 3:142, 3:144f, 3:145–146
 - Pliocene–Pleistocene Hominids, 1:310
 - Pollen Database, 3:832, 3:833
 - pollen records
 - Late Pleistocene, 4:9–17
 - postglacial, 4:85–103
 - sand seas, 1:624–626, 1:627–628, 1:627f, 1:629f
 - vertebrate studies
 - Early Pleistocene, 4:599
 - Late Pleistocene, 4:664–672
 - Middle Pleistocene, 4:615–619
 - Younger Dryas climate event, 3:127, 3:128–129
 - see also* individual countries; North Africa
- African Humid Period, 2:375–376, 4:88–89, 4:88f, 4:98–100, 4:100f
- African Pollen Database (APD), 3:832, 3:833
- Afromontane region, Africa, 4:12–13
- Agassiz Lake, North America, 2:537–538
- Agassiz, Louis, 1:3, 1:3f, 3:180, 3:182
- AGCMs *see* atmospheric general circulation models
- age-altitude analysis, SLIPs, 4:401–404, 4:402f
- age attributes
- ^{14}C of plant macrofossils, 4:361ge
 - amino-acid dating, 1:45–46
 - continental biostratigraphy, 4:206
 - glacial troughs, 1:857–861
 - rock glaciers, 3:539
 - sea-level index points, 4:372
 - soil morphology, 3:405
 - see also* dating techniques
- age control, varved sediments, 4:585–586
- age diagnosis, diatoms, 1:585f
- age distribution, carbonate-precipitating lakes, 1:333
- age equation, argon-40/argon-39 dating, 2:478
- age scale synchronization, ice cores, 2:314–315
- aggradation
- fluvial, 1:676–678, 1:679f, 1:684, 1:685f, 1:691
 - paleosols, 3:370f, 3:372–373
 - permafrost, 3:464–465
 - terrace sequences, 1:684, 1:685f, 1:691
- aggradational pedogenesis, 3:370f, 3:372–373
- aggregate tree growth principle, 1:454–455
- aging, definition, 4:361ge
- see also* age attributes
- agriculture
- impact on beetles, 1:274, 1:278
 - origins, 3:716f
 - pollen records, 3:880–903
 - spread of, 3:706–708, 3:716f
 - see also* crop...
- agricutans, 3:383–386
- Agulhas leakage, 3:22, 3:23f
- ahermatypic corals, 2:795
- air bubbles, ice cores, CO_2 , 2:311–318
- air composition, ice-cores, 2:431
- air extraction methods, ice cores, 2:468–469
- air mass systems, North America, 4:142–143
- air temperature, Arctic paleoclimate, 3:116
- ^{26}Al , erosion-island plots, 1:433, 1:434f, 1:435
- Alaska
- dune fields, 1:599, 1:599f, 1:602
 - insect fossil studies, 1:174–176, 1:177f, 1:178f
 - lake level studies, 2:543–544
 - Late Pleistocene events, 2:191–192, 2:192–193, 2:194, 2:195–197, 2:198–199
 - loess records, 2:620, 2:623f, 2:624, 2:624f, 2:626
 - mummified mammals, 4:713
 - collared pika, 4:715
 - giant moose, 4:715
 - helmeted muskox, 4:715
 - horses, 4:714
 - snowshoe hare/vole, 4:715
 - steppe bison, 4:714–715
 - woolly mammoth, 4:714
 - neoglacal chronology, 2:272
 - pollen records, 4:39–44, 4:47, 4:124–127
 - Yedoma, 3:543, 3:545–546, 3:548, 3:549–550, 3:551t
 - see also* Beringia
- alasses, 1:256, 3:578
- albedo, 1:879
- Alces latifrons* *see* giant moose
- algae
- ^{14}C of plant macrofossils, 4:363
 - calcareous, 1:346–348
 - fresh-water, 4:88–89
 - paleolimnology, 3:264–265
 - sea level indicators, 4:382–383
 - see also* diatom...
- algal blooms, 3:294f, 3:296–297, 3:345–346
- algal rim indicators, 4:382–383
- alkalinity, deep-sea, 2:860–861
- alkenone paleothermometry, 2:755–764
- biological origin, 2:757–762
 - biomarker proxies, 2:775–776
 - calibration to SST, 2:757–759, 2:758f
 - chemical structures, 2:756–757, 2:756f
 - depth/seasonality of production, 2:761
 - haptophyte algae, 2:755–764
 - postglacial North Atlantic, 3:55–56
 - quantification/analytical challenges, 2:755–756
 - sedimentary environments, 2:759–760, 2:759t, 2:760t
 - SST, 2:757–759, 2:758f, 2:759–760, 2:760–761, 2:762
 - stratigraphic offsets between proxies, 2:762
 - suspended particulates and SST, 2:759–760
 - unsaturation index, 2:757–758, 2:759–760, 2:760–761, 2:760t, 2:761–762
 - validity in past, 2:760–761
 - water column transport, 2:759–760
- alkenone proxies, marine carbon cycle, 2:855
- alkenone unsaturation index (U_{37}^K), 2:757–758, 2:759–760, 2:760–761, 2:760t, 2:761–762
- allogenic drivers, fluvial paleohydrology, 3:253, 3:256–257
- allogenic forcing, 1:676
- allopatric speciation model, 4:723, 4:724
- Allophaiomys* genus, 4:602
- allostratigraphy, 4:243–249
- fluvial sediments, 4:244–248
 - glacial sediments, 4:248
 - glacial sequences, 2:86–89
 - lake-level fluctuations, 4:245–247
 - in practice, 4:248
 - see also* lithostratigraphy; morphostratigraphy
- alluvial fans, 1:435–436, 3:554–556, 3:559–560
- alluvial land systems, 3:559–561, 3:561–562
- see also* alluvial sediments
- alluvial pollen, 4:143
- alluvial rivers
- channel patterns, 1:670, 1:670f
 - Holocene records, 1:672f
 - rapid environmental change response, 1:676–683
- alluvial sediments
- plant macrofossils, 3:594
 - streams, 3:768–784
 - see also* alluvial land systems
- alluvial sequences, pedostratigraphy, 4:256–258
- alluvial talus, 3:570–571
- alluvial terraces, 3:404f, 3:405, 3:408f

- Alnus* macrofossils, 3:642–646, 3:644f
 alpha-diversity, definition, 1:361ge
 Alps
 Austrian speleothem records, 1:299
 Eurasian glaciations, 2:231–232
 Middle-Pleistocene glaciations, 2:177–179
 neoglaciation in Europe, 2:258f,
 2:259–260, 2:260f, 2:263, 2:264f
 plant macrofossils, 3:605f, 3:606f
 ALT *see* active-layer thickness
 Altai/Altay Mountains, 2:240–241, 2:242f,
 2:384–385
 Altiplano
 ice core studies, 2:387–388, 2:392
 lake level studies, 2:531–532, 2:532–533,
 2:534–535
 altitudes in ELA, 2:241
 altitude-varied production rates, nuclides,
 1:428–429
 altitude-varied target exposure, nuclear
 interactions, 1:427–428
 altitudinal distribution
 glacial cirques, 1:847–849
 South American vegetation, 4:56f
 aluminum-26 from quartz, 1:413
 aluminum coatings, soil, 3:383
 Amazon basin
 climate records, 2:390
 lake level studies, 2:531
 physical/meteorological setting,
 2:388–389
 precipitation, 2:389
 stable isotopes, 2:390
 see also Amazon river
 Amazon rainforest
 pollen records
 coastal mangroves, 4:156–157
 postglacial, 4:156–157, 4:157–159,
 4:157f, 4:158f, 4:159f
 savannas, 4:157–159, 4:159f
 stability/dynamism, 4:59–61
 vertebrate fossils, 4:681–682
 Amazon river, 3:22–24
 see also Amazon basin
 Ambrosia (ragweed) pollen records, 4:116f,
 4:119–120
 American Code of Stratigraphic
 Nomenclature, 4:222
 American cordilleras neoglaciation,
 2:269–276
 the Americas
 300,000–8,000 years ago, 1:119–134
 see also Central America...; North America;
 South America; United States of
 America
 Amersfoort, Netherlands, 4:1
 Amery Ice Shelf/Oasis, East Antarctica, 2:220
 AMH *see* anatomically modern humans
 amino-acid dating, 1:37–48
 absolute age, 1:45–46
 aminostratigraphy, 1:44–45, 1:47–48
 applications, 1:44
 biomolecules over time, 1:37
 chiral forms, 1:37, 1:39
 D- and L-amino acids, 1:38, 1:39, 1:39f,
 1:43f, 1:44, 1:46f, 1:47–48
 environmental pH, 1:41
 leaching, 1:43–44
 optimizing samples, 1:42
 paleothermometry, 1:45–46
 polypeptide chain, 1:41–42
 preservation of amino acids, 1:38
 taxonomic effects, 1:41–43
 temperature, 1:39–41
 time-averaging, 1:46–47
 amino acid racemization (AAR), 1:37, 1:39,
 1:44–45, 1:45–46, 1:47f, 1:450
 history/introduction, 1:15, 1:15f
 stratigraphy overview, 4:192–193
 see also racemization
 aminostratigraphy, 1:44–45, 1:47–48
 aminozones, 1:44–45, 1:46f
 ammonium (NH₄⁺), ice cores, 2:331
 AMO *see* Atlantic Multidecadal Oscillation
 AMOC *see* Atlantic meridional overturning
 circulation
 amorphous coatings, soil, 3:383, 3:385f
 amphibians, 4:621, 4:622t, 4:675
 amphi-North Atlantic Younger Dryas,
 3:222–223, 3:224–225
 AMS *see* accelerator mass spectrometry
 Anadara trapezia mollusk shells, 4:194, 4:194f
 anagenesis, vertebrates, 4:730
 analog approaches in paleoclimatology,
 3:102–112
 analog technique
 climate reconstructions, 3:787
 glacial landsystems, 1:813–817, 1:817–822
 pollen–climate relation, 3:808–809
 see also modern analog technique
 Anatolian human expansion, 1:99
 anatomical changes
 humans in Africa, 1:92
 tree roots, 2:100
 see also plant anatomy; wood anatomy
 anatomically modern humans (AMH)
 expansion, 1:98
 indigenous population replacement, 1:100
 potential routes, Asia, 1:104f
 ancient bears, teeth/bones, 1:310
 see also bears
 ancient DNA (aDNA)
 American human colonization, 1:119,
 1:126
 beetle records, 1:170–171, 1:171f, 1:171t
 plants, 2:705–715
 age/survival, 2:710
 archeological/historical collections,
 2:706
 bulk specimen analyses, 2:707–708
 contamination, 2:710–711, 2:712f
 credibility of early studies, 2:709
 deep cloning, 2:711
 degradation, 2:709–710
 diet, 2:708–709
 domestication, 2:706–707
 environmental reconstruction,
 2:707–708
 extracting amplifiable DNA, 2:709
 forensic/archeological uses, 2:709
 future directions, 2:711–713
 genetic region analysis, 2:708–709
 herbarium collections, 2:706
 naturally-occurring sources, 2:706
 NGS, 2:711–713
 non-DNA biomolecules, 2:713
 population paleogenetics, 2:707
 postmortem degradation, 2:709–710
 problems with studies, 2:709–711
 protein, 2:713
 recovery/analysis, 2:705–706
 resurrected protein, 2:713
 RNA, 2:713
 scope of, 2:706–709
 shotgun sequencing, 2:711–712
 source material analysis, 2:706
 targeted capture, 2:712–713
 recent directions, 4:720–722
 research history, 4:719
 sources of, 4:720
 vertebrate studies, 4:668–669, 4:670,
 4:719–721
 ancient ungulates, teeth/bones, 1:311
 see also ungulates
 Andes
 central, chronology, 2:273
 glacier climate records, 2:390–393
 ice core studies, 2:387–394
 drilling sites, 2:388f
 history of drilling, 2:387
 stable isotopes, 2:389–390
 insect fossil studies, 1:235, 1:236–237
 lake level studies, 2:532–533, 2:534–535
 Middle-Pleistocene glaciations, 2:187,
 2:188f
 neoglaciation chronology, 2:273–274
 northern
 dynamics, 4:54–58
 neoglaciation chronology, 2:273
 postglacial pollen records, 4:156, 4:162
 southern, chronology, 2:273–274
 vertebrate fossils, 4:681
 see also South American glaciations
 Andorra, pollen records, 4:183–185
 Andrews, JT, 3:103–104
 Andromeda polifolia macrofossils, 3:638,
 3:640f
 angiosperms, leaf stomata, 3:615
 angle of repose, talus slopes, 3:568–569
 Anglian glaciation, 1:186–187
 anhysteretic remanent magnetization (ARM),
 3:377, 3:380
 animals
 human interactions, 4:748–755
 nutrient/contamination vectors,
 3:346–347
 see also individual animals/animal groupings
 anions, 2:326
 anisotropic fabrics, loess records, 2:607–608
 ANNs *see* artificial neural networks
 annual ablation, 1:909ge
 annual accumulation, 1:909ge
 annual balance, 1:909ge
 annual layers
 dating methods, 2:323, 4:573–581
 ice-core dating, 2:323
 thinning, ice sheets, 2:443, 2:444f
 varved sediments, 4:573–581, 4:582–589

- annual varves, marine diatoms, 1:554, 1:559
anomalous dates, definition, 3:634
anoxia, 1:361*ge*
Antarctica, 2:217*f*
 borehole temperature records, 2:299*f*, 2:301
 CO₂ ice core studies, 2:311–318
 conductivity studies, 2:322*f*
 dune fields, 1:599, 1:600*f*
 Early Quaternary, 2:167
 glaciations, 2:216–223
 history of, 2:148
 onset, 2:217
 ice cores, 2:282*f*, 2:283, 2:284, 2:284*f*, 2:285*f*
 CO₂ studies, 2:311–318
 conductivity studies, 2:322*f*
 correlation with Greenland ice core, 2:410–415
 history of research, 2:435–438
 microparticle fingerprinting, 2:343–344
 ice sheet, 2:435–438
 areal ice extent, 1:891
 diatom studies, 1:535–536
 dynamics of, 2:448–455, 2:456–462
 glacier recession, 1:804
 glaciochemical archive, 2:342
 Late Quaternary, 2:216–217, 2:218
 margin sites, 2:416–430
 properties, 2:319
 last glacial maximum, 2:218–219
 Late Quaternary glaciations, 2:216–223
 marine diatom studies/methods, 1:527–539
 methane studies, 2:334–335, 2:335–337, 2:336*f*, 2:337*f*, 2:338, 2:339
 microparticle fingerprinting, 2:343–344
 pre-Quaternary glacial history, 2:217–218
 RSL in high latitudes, 4:477–479
 stable isotopes, 2:348, 2:349, 2:350, 2:351, 2:395–402
 sub-Milankovitch events, 3:205–206, 3:207
 water stable isotope records, 2:419–420
 Younger Dryas oscillation, 3:225–226
 see also Antarctic Peninsula
Antarctic Bottom Water (AABW)
 diatom tracers, 1:535
 salinity proxies, 2:935*f*
 THC, 1:737, 1:741*f*, 1:745–746, 1:746*f*
Antarctic cold reversal (ACR), 1:241, 1:242, 3:225–226
Antarctic dust, South America, 2:639
Antarctic Intermediate-water, 3:129–130
Antarctic Peninsula
 deglaciation, 2:219–220
 ice cap characteristics, 2:216
 postglacial records, 3:80–81, 3:81*f*, 3:82–84, 3:84*f*
 RSL in high latitudes, 4:475*f*, 4:477, 4:478*f*, 4:479
anthracology, 2:716
 see also charcoal records
anthropogenic emissions, 2:313, 2:332, 2:379, 2:384, 2:463, 2:465
anthropogenic features, soil, 3:383–386, 3:385*f*
anthropogenic impact
 American expansion, 1:119–134
 Asian lake level studies, 2:520–521
 atmospheric carbon-14, 4:333–334
 Australia/New Zealand pollen records, 4:112
 beech/lime tree palynology, 3:850–852
 climate change, 3:244–248
 ecosystem dynamics, 3:250
 fingerprint analysis, 3:244–245, 3:245*f*
 ice cores, 3:246–248, 3:248*f*
 instrumental record, 3:245, 3:247*f*
 last millennium, 3:233–234, 3:245–246, 3:247*f*
 paleoclimate relevance to global warming, 3:244
 paleoclimatic record, 3:244–245
 radiative forcing, 3:233–234, 3:244
 scientific consensus, 3:244
ice core studies, 2:281, 2:284–286, 2:332
 carbon monoxide, 2:463
 carbonyl sulfide, 2:465
 China, 2:379, 2:384
 CO₂, 2:313
methane studies, 2:335–337
pollen
 Australian/New Zealand postglacial records, 4:112
 data overview, 2:701, 2:703
 indicators, 3:881–887, 3:884*f*, 3:886*t*
 Southern European postglacial records, 4:185–186, 4:187
 surface sampling/trapping, 3:843–844
 radiative forcing, 3:233–234, 3:244
 see also human impact
anthropogenic pollen indicators, 3:881–887, 3:884*f*, 3:886*t*
anthropology, 1:111–113, 1:125–126
 see also anthropogenic impact; human impact
anticoincidence shielding, radiocarbon dating, 4:308, 4:312
antidunes, 2:10
 see also dunes
antiferromagnetic material, 3:375*ge*
antisynthetic ice wedges, 3:438–439, 3:468
antlers as raw materials, 4:751, 4:752, 4:752*f*
Antonsson, K, 3:103
AO *see* Arctic Oscillation
APD *see* African Pollen Database
aquatic food web changes, 3:333–334
aquatic habitats
 African chironomid records, 1:362–365
 dinoflagellates, 2:800–815
 pigment preservation, 3:330
 see also lake...; sea...
aquatic insect species, 4:549, 4:552
aquatic moss macrofossils, 4:363
aquatic oribatid mites *see* *Hydrozetes*
aquatic plants/vegetation
 Arctic Eurasia, 3:742
 macrofossils, 3:600*f*, 3:601*f*, 3:602*f*, 3:603*f*, 3:657–673
nonmarine biogenic carbonates, 1:346–348, 1:347*f*
paleolimnological applications, 3:657–673
seed bank studies, 3:686
aquatic systems *see* aquatic habitats
Ar *see* argon
Arabian Desert sand seas, 1:629–631
Arabian Sea
 deglaciation, 3:48*f*
 foraminifera, 3:51*f*
 magnetic susceptibility, 2:889
 sea surface temperatures, 3:50*f*
aragonite, corals, 2:880
Araucaria forest, Brazil, 4:156, 4:160–161, 4:161*f*
arboreal pollen, North America, 3:773*f*, 3:774*f*, 3:775
archeoanthracology, 2:716, 2:724
archeobotany, 3:699–724
 see also environmental archeology
archeoglaciology, 2:259–260
archeological charcoal analysis, 2:716, 2:724
archeology
 ¹⁴C of plant macrofossils, 4:361
Africa
 2.7 MYR-300,000 years ago, 1:59–66
 300,000-8000 years ago, 1:91–97
 Late Pleistocene, 4:664
Americas 300,000–8,000 years ago, 1:119–134
applications, 3:880–903
Asia
 2.7 MYR-300,000 years ago, 1:67–82
 300,000-8000 years ago, 1:98–107
Australia 300,000-8000 years ago, 1:108–118
dating sediments, 2:549–551
dendroarcheology, 3:630–636
diatom analysis, 1:516–521
environmental, 1:169
Europe 1.9 MYR-300,000 years ago, 1:83–90
geomagnetic PSV, 1:713–714, 1:714*f*
lake-level reconstruction, 2:549–551
Neanderthal demise, 1:146–153
North America, 1:285–286
oribatid mite research, 2:683–684, 2:685, 2:695
overview, 1:49–58
Paleolithic
 beetle fauna, 1:187
 vertebrate studies, 4:590, 4:593
phytolith research, 3:589–590
plant macrofossil studies, 3:699–724, 4:361
postglacial, 1:154–159, 1:285–286
proxy data interpretation, 3:705
tephrochronology, 4:298–299
see also marine archeology
arches (sea arches), 4:379, 4:380*f*
Arctic
 beetle studies, 1:222
 coastal Canada
 ice wedges, 3:575*f*

- retrogressive thaw slumps, 3:579f
 themokarst lakes, 3:577f, 3:578f
 diatom studies/records, 1:562–570, 1:588–596
- Eurasian
 ice sheet growth/decay, 1:880f
 key sites, 3:735–737, 3:736f, 3:737t
 plant macrofossil records, 3:733–745
 vegetation history, 3:738–742
 fossil site map, 1:179f
 insect fossil studies
 Late Pleistocene, 1:255
 Late Tertiary/Early Quaternary, 1:173–183
 megafossils, 3:621–622, 3:622–623, 3:626–627
 mummified ground squirrel, 4:715
 North America, 1:222, 3:746–759
 paleoclimate history, 3:113–125
 air temperature, 3:116
 Cenozoic, 3:113, 3:122–123
 Early Quaternary, 3:114–115
 greenhouse gases, 3:113, 3:116, 3:122–123
 Holocene, 3:118–121, 3:122–123
 ice-age cycles, 3:113–114
 ice volume, 3:116
 interglaciations, 3:116–117
 LCM, 3:118
 Marine Isotopic Stages, 3:116–117, 3:117–118, 3:118–121
 Mid-Pleistocene Transition, 3:115–116
 Pliocene warmth, 3:113–114
 rapid climate oscillations, 3:117–118
 twentieth century warming, 3:121–122
 plant macrofossils, 3:605f, 3:733–745, 3:761, 3:762
 sea ice, 1:588–596
 throughflow, 2:158f, 2:162, 2:164f
 treelines, 3:621–622, 3:622–623, 3:626–627, 3:867–868
 warming, 3:113–114, 3:121–122, 3:497–498
- Arctic Oscillation (AO)
 last interglacial, 3:159–160
 paleo-ENSO, 3:172–174, 3:173f
 volcanic/solar forcing, 1:750
- Ardleigh sites, England, 1:185, 1:186
- Arduino, Giovanni, 1:2f
- areal ice extent, 1:884–891
 former glaciers, 1:884–890
 geomorphological features, 1:884–890
 glacier/icesheet extent, 1:884–891
 LGM, 1:884
 sediment record, 1:884
 uncertain areas, 1:890–891
- areal scouring, glacial landscapes, 1:862–863
- Argentina
 dune fields, 1:613–614, 1:614f
 glaciations, 2:251
 postglacial pollen records, 4:161, 4:161f
 vertebrate fossils, 4:683
- argon (Ar)
 argon-38, 1:413
 argon-39 radioactive dating, 2:306
 argon-40/argon-39 dating, 2:477–482
 age calculation, 2:478–480
 historical development, 2:477–478
 instrumentation, 2:481
 K/Ar dating, 2:477–482
 methods, 2:478–480
 new perspectives, 2:480–481
 strengths/weaknesses, 2:480
 tephrochronology, 4:287–288
 argon–argon dating method, 1:10–11
 thermal diffusion paleotemperature records, 2:431–433
- arid/semiarid biomes, Africa, 4:14
- arid climates/regions
 Arctic Eurasia, 3:740f
 cosmogenic nuclide dating, 1:437
 eolian sand, 3:357
 low-latitude sand seas, 1:623–636
 Northern Hemisphere 6000 ¹⁴C years BP, 3:869
 pollen records, 4:14
 rodent middens, 3:674–683
 soil morphology, 3:363f, 3:365
 southwestern North America, 3:675
- aridity
 indicator plants, 3:740f
 Northern Hemisphere 6000 ¹⁴C years BP, 3:869
 teeth/bones, 1:308
- arid-zone lakes, Australia, 2:526, 2:529
- Arizona, North America, 3:678t
- ARM *see* anhysteretic remanent magnetization
- armadillos, 4:685
- Arrhenius equation, 1:39, 1:40f
- arroyos, American Southwest, 1:691
- art
 European hominins, 1:89
 fraud case evidence, 4:535
 hominid–animal interactions, 4:752–754
 postglacial human adaptations, 1:158–159
- Artemisia* pollen records, 4:39, 4:48
- arthropods *see* oribatid mites
- artifacts
 definition, 1:108ge
 diatom analysis, 1:516–518
- artificial neural networks (ANNs), 2:843f, 2:844t, 2:845–846, 2:846–847, 3:810
- artiodactyla, 4:689–690
- Arvicola–Mimomys* lineage, 4:639, 4:643, 4:725–727
- ash *see* tephra
- Asia
 2.7 MYR–300,000 years ago, 1:67–82
 300,000–8000 years ago, 1:98–107
 glaciations in Eurasia, 2:224–235
 hominin movement, 1:67, 1:70–71
 ice core studies, 2:379–386
 insect fossil studies, 1:255–273
 lake level studies, 2:506–523
 bases of constructions, 2:506–508
 causes of level changes, 2:511
 Central Asia, 2:511
 climate simulations, 2:517–518, 2:518f, 2:519, 2:519f
 closed/overflowing lakes, 2:506, 2:512f
 deep-lake drilling, 2:508–511, 2:519
 East Asia, 2:512
 historical review, 2:506–511
 late MIS-3, 2:515, 2:519
 Late Quaternary changes, 2:511–512
 LGM, 2:514–515
 Middle East, 2:511
 Mid-Holocene, 2:513, 2:520
 overview of changes, 2:511–512
 paleoclimate modeling, 2:511, 2:516–519, 2:518f, 2:519f
 recent/future aspects, 2:520t, 2:521, 2:521f
 records, 2:506
 regional climate change, 2:511
 spatial patterns, 2:512–515, 2:517f
 Tibetan plateau, 2:511–512
 loess records, 2:585–594
 megafaunal extinctions, 4:707–708
 Northern Hemisphere Younger Dryas, 3:127–128, 3:128f
 pollen records, 4:27–38, 4:164–172
 Southeast Asia, Late Pleistocene, 4:693–699
 vertebrate studies, 4:605–614, 4:651–663
see also Australasia; Central Asia; China; Eurasia; Highland Asia; India; *see also individual countries*
- aspect ratio
 definition, 1:781
 volcanic landforms, 1:794–795, 1:795f
- assemblage, definition, 1:83ge
- assemblage type, definition, 1:83ge
- astronomical theory
 paleoclimates, 2:136–141
 future climates, 2:139–140
 glaciation causes, 2:136–141
 historical viewpoint, 2:136–137
 Ice Ages, 2:136
 long-term parameter variations, 2:137–138, 2:137f
 Milankovitch, 2:136–137, 2:138–139
 modeling past/future climates, 2:139–140
 past climates, 2:136, 2:138–139, 2:139–140
 Quaternary climatic change, 1:26–27
see also climate forcing mechanisms
- astronomical time scales, 4:220
- astronomical tuning, marine oxygen isotopes, 2:910–912
- Aterian unit, archeological evidence, 1:94–95
- Athabasca Sand Dunes, Canada, 1:598, 1:599f, 1:602
- Atherton Tableland, Australia, 4:106–107
- Atlantic coast
 North American sea-levels, 4:489–491, 4:492f
 South American sea-levels, 4:493
see also North Atlantic...
- Atlantic meridional overturning circulation (AMOC)
 last interglacial, 3:162–163
 Late Neogene/Early Quaternary transition, 2:154, 2:160–162
 Younger Dryas climate event, 3:126, 3:127, 3:129–130, 3:131

- Atlantic Multidecadal Oscillation (AMO), 3:233
- Atlantic Ocean
- abrupt climate change, 3:49
 - $\delta^{18}\text{O}$ interannual variations, 2:390
 - diatom studies/records, 1:562–570
 - paleoceanography, Late Pleistocene, 3:9–17
 - radiolarians, 2:832f, 2:834f, 2:835f
 - THC, 1:738f
 - see also* North Atlantic...
- Atlantic rainforest, Brazil, 4:160
- Atlantic sediment proxies, 2:927–929, 2:928f, 2:929–930, 2:929f
- Atlantic Westerly Jet, 2:549, 2:551–552, 2:553
- atmosphere
- cosmic radiation, 1:422–427
 - dynamics, last interglacial, 3:159–160
 - nuclear disintegration rates, 1:424–425
 - nuclear interactions at different altitudes, 1:427–428
 - nuclide production rates, 1:428–429
- atmospheric aerosols *see* aerosols
- atmospheric carbon-14
- anthropogenic era, 4:333–334
 - deglaciation, 4:331–332
 - Holocene, 4:329–331
 - INTCAL09 calibration curve, 4:329–330, 4:330f, 4:331, 4:332, 4:332f
 - Middle/Upper Paleolithic, 4:332–333
 - production of, 4:329
 - radiocarbon dating, 4:329–335
 - variations, 4:329–335
 - see also* carbon-14
- atmospheric carbon dioxide
- deep-sea carbonate dissolution, 2:868–869
 - Early Pleistocene paleoceanographic records, 3:4f, 3:6
 - growth/decay of ice sheets, 1:879–881
 - paleoceanography overview, 2:749–750
 - Pliocene climate warmth, 3:186–187, 3:186f, 3:188, 3:190
 - reconstruction from fossil leaves, 3:613–620
 - see also* carbon dioxide
- atmospheric CH_4 , 2:335–337, 2:336f
- see also* methane...
- atmospheric chemistry biosphere feedback, 1:727
- atmospheric circulation
- African pollen records, 4:85–88
 - Asian lake level studies, 2:519
 - Cenozoic records, 1:640–641
 - cosmogenic radionuclides, 2:358
 - Indian Ocean, 3:47f
 - last millennium climate, 3:231–233
 - major ions, 2:326–333
 - see also* atmospheric vapor transport
- atmospheric deposition
- contaminants, 3:334–335
 - dust, 1:733, 1:733f
- atmospheric dust, 1:729–736
- climate and, 1:729–730, 1:730f
 - climate–dust linkages diagram, 1:730f
 - deposition, 1:733, 1:733f
 - effects of, 1:730f, 1:734–735
 - glacial climates, 1:729–736
 - global deposition, 1:733, 1:733f
 - global sources, 1:731f
 - indirect effects, 1:730f, 1:734–735
 - measurements, 1:730–731
 - models of dust cycle, 1:732–733
 - ocean biogeochemistry, 1:735
 - particle properties, 1:733–734
 - removal from atmosphere, 1:729–730
 - semi-direct effects, 1:734–735
 - sources, 1:731f
 - variability, 1:731–732
 - see also* dust
- atmospheric gases
- ice cores, 2:410, 2:463–470
 - see also* individual gases
- atmospheric general circulation models (AGCMs)
- data-model comparisons, 3:137, 3:144f, 3:145–146
 - North American loess records, 2:627
 - see also* general circulation models
- atmospheric methane *see* atmospheric CH_4
- atmospheric pollution research, 3:319–321
- atmospheric reconstruction, firn air, 2:366–367
- atmospheric vapor transport
- oxygen isotopes, 2:916–917
 - see also* atmospheric circulation
- Attelabidae beetles, 1:216
- Auckland City, New Zealand, 4:291–293, 4:292f
- Auckland Island glaciers, 2:213
- Aulacomnium palustre* macrofossils, 3:649, 3:650f
- Aurelian biozone, 4:210, 4:211f
- Aurelian stage mammal assemblages, 4:612–613, 4:639
- Aurignac, France, 1:142
- Aurignacian industry, 4:748ge, 1:149–150, 4:754f
- Australasia
- Early Quaternary, 2:167–168
 - history of glaciation, 2:148–149
 - pollen records, 4:18–26
 - relative sea level changes, 4:493
- Australia
- 300,000–8000 years ago, 1:108–118
 - arid-zone lakes, 2:526, 2:529
 - beetle records, 1:168, 1:191–199
 - Chironomid records, 1:398–399, 1:399f, 1:403
 - desert sand seas, 1:628–629, 1:630f
 - Early Pleistocene vertebrate studies, 4:599
 - Eastern Highlands, 2:524–525
 - Eyre Lake Basin, 2:525f, 2:526, 2:527–528, 2:527f, 2:528–529
 - Gregory Lake, 2:525f, 2:526, 2:528
 - groundwater playa systems, 2:528
 - lake level studies, 2:524–530
 - Late Pleistocene
 - beetle records, 1:191–199
 - glaciers, 2:202–206
 - pollen records, 4:21–23, 4:23–25
 - Murray–Darling Basin, 2:526–527, 2:529
 - MCR reconstruction, beetle records, 1:168
 - megafauna
 - extinctions, 4:701–704, 4:710
 - fossil localities, 4:703f
 - stratigraphy overview, 4:194–195
 - mid-Pleistocene vertebrate records, 4:621–638
 - monsoon-fed terminal lakes, 2:528–529
 - paleoclimate, 1:168, 2:528–529
 - paleotempestology, 3:213, 3:214f, 3:215
 - pollen records, 4:104–114
 - rodent middens, 3:674
 - sea-level changes, Holocene, 4:444
 - soil morphology, 3:407
 - South Pacific ocean records, 3:77–80
 - tectonic basins, 2:524–525
 - vertebrate studies, 4:599, 4:621–638
 - volcanic lakes, 2:525–526
 - Western Victoria, 2:525–526, 2:529
 - Woods Lake, 2:525f, 2:526, 2:528
- Australian slide method, OSL dating, 2:658f
- Austrian Alps, 1:299, 2:231
- autoecology, definition, 1:361ge
- autogenic processes, fluvial paleohydrology, 3:253, 3:256–257
- autotrophic dinoflagellates, 2:800–801
- avalanches, talus slopes, 3:570, 3:570f, 3:571f
- Aveley, England beetle assemblages, 1:189
- aves, 4:683–684
- see also* birds
- avulsion deposits, 1:664–665, 1:668–670, 1:672
- Awatere Valley, New Zealand, 1:248–249, 1:250–251

B

- Ba/Ca foraminiferal proxy, 2:854
- back-barrier lakes, 3:209–211
- back-barrier wetlands, 3:211–213
- backscattered electron imagery (BSEI), 2:820
- bacteria
- cloning techniques, 4:719
 - Vostok ice core, 2:292f, 2:293, 2:294–295, 2:296
- Badain Jaran Desert, China, 1:606, 1:607f, 1:609, 1:610f, 1:611
- Baerreis, DA, 3:109
- Baffin Bay diatom records, 1:567
- Baffin Island, Canada, 1:893, 4:40f, 4:46f, 4:47
- Baikal Lake, Russia, 1:546–547, 1:547–550
- balance ratio (BR) method, ELA, 1:916
- balance velocities, East Antarctic Ice Sheet, 2:461f
- balance year, definition, 1:909ge
- Balderton Sand/Gravel Formation, England, 1:189

- Balearic Islands
 bovids, 4:739–741
 pollen records, 4:179–183, 4:180*t*
- Balkan peninsula pollen records, 4:181*t*, 4:186–187
- bananas, 3:589–590
- bandicoots, 4:632–633
- banks, morainal, 1:776, 2:30–32, 2:34*f*
- Banks Peninsula, New Zealand, 1:247
- bank vole evolution, 4:725
- Barbados
 coral records, 2:672–673, 2:674*f*, 4:409, 4:411
 ESR coral dating, 2:672–673, 2:674*f*
 sea-level change, 4:430*f*, 4:432, 4:433–434, 4:433*f*, 4:441
- barchanoid dunes, 1:599
- Barents Sea, 1:891, 3:59, 3:60, 3:60*f*
- barium
 North Pacific Ocean, 3:42–43, 3:44*f*
 nutrient proxies, 2:901–903
- Barling, England beetle assemblages, 1:188
- Barnes ice cap, 2:424
- barrier coasts, 3:563
- barrier overstepping model, 4:389–390
- barrier sequences, sedimentary indicators, 4:391*f*
- bars
 braided rivers, 2:10–11, 2:11*f*
 sea-level index points, 4:375*f*
 subaerial mesoforms, 2:10–11, 2:12
- Bartington bridge, 3:376
- Bartington MS2 system, 2:886–887
- Bartlein, PJ, 3:106–107, 3:109
- basal melting, 2:439–440, 2:443, 2:445, 2:446
- basal peat beds, Netherlands, 4:240
- basal processes, permafrost–glacier interactions, 3:508
- basal sliding
 East Antarctic Ice Sheet, 2:456–457, 2:460*f*
 Greenland ice sheet, 2:439–440
 West Antarctic Ice Sheet, 2:450–451
- basal stress, 2:449–450
- basal temperature, Greenland ice sheet, 2:439–440
- basal topography, 2:439
- base level
 changes, 1:684–685
 fluvial terrace formation, 1:684–685, 1:688–689
- basin characteristics, macrofossil taphonomy, 3:688
- basin size, vegetation patterning effects, 3:873–874
- bathymetric maps, Africa, 2:504*f*
- bat records, 4:637
- Bavelian stage mammal assemblages, 4:610–611
- Bayesian modeling, 4:349
- B/Ca marine carbon cycle proxy, 2:851–853, 2:851*f*, 2:852*f*
- ¹⁰Be concentrations
 Australian glaciers, 2:202–204
 dust-borne ¹⁰Be, 2:355
- erosion-island plots, 1:433, 1:434*f*, 1:435
- ice cores, 2:353–360, 2:410, 2:411–412
see also beryllium-10
- beach ridge paleohurricane proxy, 3:213
- beachrocks, sea level changes, 4:383–384, 4:384*f*, 4:388–389, 4:389*f*, 4:392, 4:496–497, 4:497*f*
- bears
 South America, 4:688
 species origins, 4:724–725, 4:724*f*
 teeth/bones, 1:310, 1:311*f*
- Bedfordshire, England beetle assemblages, 1:188
- bedforms of ice sheets, 1:895–896, 1:902–903, 1:903–904
- bedrock
 cosmogenic nuclide dating, 1:435, 1:437
 Greenland ice sheet, 2:439
 outcrops, 1:921–923, 1:922*f*, 1:926*f*, 1:927*f*
 periglacial slope deposits, 3:486–487
 relative-age dating, 1:921–923, 1:922*f*, 1:926*f*, 1:927*f*
- beds in lithostratigraphy, 4:228, 4:238–240
- beech (*Fagus*)
 distribution, 3:857, 3:858, 3:858*f*
Nothofagus (southern beech), 4:109–110, 4:112
- pollen records
 New Zealand postglacial, 4:109–110, 4:112
 northeastern North America, 4:116*f*, 4:119
- stand-scale palynology, 3:850–852, 3:851*f*
- Beeston, England beetle assemblages, 1:184
- beetle fauna (Coleoptera) studies/records
 ancient DNA, 1:170–171, 1:171*f*, 1:171*t*
 Arctic regions, 1:173–183
 Australia, 1:191–199
 emerging research areas, 1:169–170
 environmental archeology, 1:169
 Europe
 Late Pleistocene, 1:200–206
 Middle Pleistocene, 1:184–190
 postglacial, 1:274–281
 exoskeletal parts, 1:165–166, 1:166*f*, 1:169
 forensic science use, 4:549, 4:552–554
 fossil overview, 1:161–172
 Greenland, 3:761–762
 Japan, 1:207–220
 Late Pleistocene
 Australia, 1:191–199
 Europe, 1:200–206
 Japan, 1:207–220
 New Zealand, 1:244–254
 North America, 1:221–234
 Northern Asia, 1:255–273
 South America, 1:235–243
 Late Tertiary/Early Quaternary, 1:173–183
 location of fossils, 1:162
 MCR reconstructions, 1:167–168, 1:168*f*, 1:180–181
 Middle Pleistocene, 1:184–190
 New Zealand, 1:244–254
 North America, 1:282–290
 Northern Asia, 1:255–273
- origins/longevity of, 1:180
- paleoecological assumptions, 1:166
- paleoenvironmental reconstructions, 1:168–169
- plant reconciliation, 1:197–198
- postglacial
 Europe, 1:274–281
 North America, 1:282–290
 preservation/abundance/collection, 1:255
 as proxies, 1:161–162, 1:166–168
 sampling/extraction, 1:162–165, 1:165*f*
 South America, 1:235–243
 species constancy, 1:166
 thermal tolerance, 1:166–167
 Younger Dryas–Preboreal transition, 1:275
 zoogeography, 1:169
- before present (BP) measure, 1:108*ge*
- behavioral changes, humans
 Africa, 1:92, 1:93
 Asia, 1:101
 dendroarcheological samples, 3:634–636
 Europe, 1:88–89
 symbolism, 1:146–147
see also hominin behavior
- Belgium, sea level changes, 4:492–493
- bench erosion marks, 4:378–379, 4:379*f*, 4:382–383, 4:383*f*
- benchmarking, paleoclimate models, 3:141–142, 3:144*f*, 3:145–146
- Benguela System, South Atlantic, 3:18, 3:21
- Bennett's broken stick model, 1:491
- benthic $\delta^{18}\text{O}$ pollen records, 4:5–6, 4:5*f*
- benthic diatoms, 1:474–475, 1:510, 1:546
- benthic foraminifera, 2:765–774
 applications, 2:767–772
 Arabian Sea, 3:51*f*
 biology, 2:765
 carbon-13 nutrient proxies, 2:899, 2:901*f*, 2:902, 2:903, 2:903*f*
 as carbon cycle proxies, 2:852–853, 2:852*f*, 2:853–854
 distribution factors, 2:766*f*
 ecology, 2:765–767
 epifaunal/infaunal forms, 2:767–768, 2:770
 fossil formation, 2:767
 growth stages, 2:766*f*
 high latitude glaciations, 2:768–769
 Indian Ocean, 3:46, 3:53*f*
 intermediate water oxygenation, 2:767–768
 marine archeology, 2:772
 Mg/Ca paleothermometry, 2:871, 2:875–876, 2:877*f*
 monsoons, 2:768
 North Pacific Ocean, 3:34–35, 3:35*f*, 3:38*f*
 oxygen isotopes, 2:920*f*
 pollution monitoring, 2:772
 as recorders, 2:767
 sea levels and, 2:769–770
 shell shapes, 2:766*f*
 storms/tsunamis, 2:770–772
 surface productivity, 2:768
 thermohaline circulation, 2:770
- benthic quality index (BQI), chironomids, 1:389*f*

- benthos, definition, 1:361*ge*
 benzene, 4:309–310, 4:310*f*
 Bergmann's rule, mammal size, 4:697–698
 Beringia
 beetle studies, 1:228–233, 1:229*t*, 1:231*f*, 1:232*f*
 insect fossil studies, 1:174–176
 Late Pleistocene events, 2:191–201
 plant macrofossil records, 3:734–735
 Yedoma, 3:542–552
 Bering Land Bridge (BLB)
 Early–Middle Pleistocene, 3:753
 human colonization, Americas, 1:119, 1:121
 insect fossil studies, 1:174
 Late Pleistocene events, 2:191, 2:192*f*, 2:197
 Bering Sea diatom assemblages, 1:574, 1:575
 Bermuda High hypothesis, 3:215, 3:216*f*
 beryllium-10 (^{10}Be)
 ice-core chronologies, 2:306
 TCN methods, 1:410, 1:411*f*, 1:413, 2:202–204
 see also ^{10}Be concentrations
 Beschel, R, 2:565, 2:567–568, 2:568–569
 beta-diversity, definition, 1:361*ge*
Betula macrofossils, 3:598*f*, 3:608, 3:642–646, 3:644*f*
 BIAs *see* blue ice areas
 bicarbonate (HCO_3), 1:293
 Big Bend, Texas beetle records, 1:225–227
 Bignell Loess, North America, 2:620–621
 Biharian biozone, 4:210, 4:211*f*
 binge–purge model, 1:881–882, 1:882*f*, 2:40–41
 bioarcheology, 3:699–724
 see also environmental archeology
 biobate phytoliths, 3:583*f*, 3:586*f*, 3:587*t*
 biochemical varves, 4:584, 4:584*f*
 biochronology
 coccolithophores, 2:792
 vertebrate studies
 Middle Pleistocene, 4:639, 4:640*f*, 4:643
 Plio-Pleistocene transition, 4:599
 bioclimatic profiles, beetle fauna, 1:192
 BIOCLIM program, 1:192
 biodiversity
 fossil record bias, 3:816–818
 interpreting records, 3:818–820
 landscape biodiversity, 3:819
 measurement, 3:816
 paleolimnological research, 3:321
 palynological richness, 3:818, 3:819–820, 3:819*f*
 pollen studies, 3:816–820
 reconstruction, 3:816–820
 taxonomic precision, 3:816
 bioeroded hole sea level indicators, 4:380–381
 biofacies analysis, oribatid mites, 2:688*t*, 2:689*f*, 2:691–692
 bioforensic diatom analysis, 1:522
 biogenic barium record, 3:42–43, 3:44*f*
 biogenic carbonates, 1:292–293, 1:341–353
 biogenic emissions, ice cores, 2:330–331
 biogenic gases, ice cores, 2:280
 biogenic ooze, 1:637*ge*, 1:638
 biogenic silica (BSi), 1:549*f*, 3:294, 3:344
 see also diatom silica records
 biogenic varves, 4:584, 4:584*f*
 biogeochemical cycles, diatom silica, 2:734
 biogeochemical processes
 ice core studies, 2:326
 postglacial Indian Ocean, 3:46
 biogeochemical records, North Pacific Ocean, 3:42–44
 biogeochemical role of dust, 3:412–420
 biogeography
 atmospheric pollution, 3:319–321
 beetle assemblages
 Japan, 1:219
 North America, 1:286–289
 South America, 1:241–242
 Western Europe, 1:184–190
 biodiversity and, 3:321
 chironomid records, Africa, 1:371
 diatom studies, Antarctica, 1:530–533
 dinocyst distribution, 2:805–808
 disturbance, 3:317–319
 evolution, 3:322–323
 extinctions, 3:322–323
 invasive species, 3:321–322
 paleoclimate, 3:314–317
 paleolimnology, 3:313–325
 questions addressed by, 3:313
 radiolarian patterns, 2:831*f*
 Southeast Asia, 4:693, 4:697
 succession, 3:317–319
 terrestrial-aquatic links, 3:317–319
 vegetation reconstruction, 3:866–867
 vertebrate studies, South America, 4:680
 water quality, 3:319–321
 biogeophysical biosphere feedbacks, 1:725–726, 1:727*f*
 biological communities, beetles, 1:180
 biological dating techniques, lichenometry, 2:565–572
 biological evidence, economic application, 3:708–715
 biological foundations
 chironomid studies, 1:355–356
 dendrochronology, 1:453–454
 biological materials, 2:288–297
 evolutionary potential, 2:294–296
 Greenland ice cores, 2:291–292
 history of, 2:289*f*
 ice cores, 2:283, 2:288–297
 ice sheets, 2:289
 measurement in ice cores, 2:293–294
 metabolic activity, 2:294–296
 temperate mountain glaciers, 2:289, 2:292–293
 Vostok ice core, 2:289–291, 2:290*f*, 2:291*f*, 2:292*f*, 2:293*f*, 2:295*f*, 2:296*f*
 biological proxies
 alkenone paleothermometry, 2:755–764
 benthic foraminifera, 2:765–774
 biomarkers and, 2:775–782
 coccolithophores, 2:783–794
 corals/sclerosponges/mollusks, 2:795–799
 dinoflagellates, 2:800–815
 marine diatoms, 2:816–824
 planktic foraminifera, 2:825–829
 radiolarians/silicoflagellates, 2:830–840
 temperature proxies, 2:841–848
 biological weathering processes, 3:502
 biomarker proxies
 $\%C_{37.4}$ proxy, 2:778, 2:778*t*
 alkenone paleothermometry, 2:775–776
 climate, 2:775–782
 HBI sea ice proxy, 2:779–780
 lipid δD proxy, 2:778–779
 MBT/CBT proxy, 2:776–778
 paleoceanography, 2:775–782
 salinity, 2:778–779
 temperature, 2:775–778
 TEX₈₆ proxy, 2:776–778
 biomarkers
 Arctic sea ice diatoms, 1:588–596
 pigments as, 3:326
 biomass burning, 2:716–729, 4:18–19, 4:25
 BIOME 6000 project, 3:862–863
 data set comparisons, 3:139*f*, 3:141*f*, 3:142, 3:142*f*, 3:145–146
 biomes/biomization
 arid/semiarid biomes, Africa, 4:14
 evolution, South America, 4:52–54
 LGM pollen records, 4:78–79
 northeastern North America, 4:120
 pollen studies/records, 3:836, 3:862–863, 3:863*f*, 4:14, 4:78–79
 temporal composition changes, 3:866–867
 vegetation reconstructions, 3:861–870
 biophysics, 1:722, 1:722*f*
 biosphere feedbacks
 forcing mechanisms, 1:721–728
 Glacial Inception, 1:727–728
 Holocene, 1:723–725
 hot spots, 1:723–728
 land-surface biophysics, 1:722, 1:722*f*
 Last Interglacial, 1:727–728
 LGM, 1:725–727
 long-term carbon cycle processes, 1:722–723
 main mechanisms, 1:721–723
 numerical modeling, 1:723
 paleoclimatology, 1:721–728
 photosynthesis, 1:721
 biosphere history
 ecosystems, 1:23
 megafaunal extinctions, 1:23–24
 research history, 1:23–24
 biosphere responses, treelines, 3:690, 3:697
 biostratigraphy
 coccolithophores, 2:792, 2:793*f*
 continental, 4:206–214
 diatoms
 Lake Baikal, 1:549*f*
 North Atlantic/Arctic, 1:562–563
 Pacific Ocean, 1:584*f*, 1:585
 Southern Ocean, 1:536
 dinocysts, 2:805
 lithostratigraphy and, 4:227, 4:228, 4:229–232
 mammal evolution, 4:729–730
 microfossil-based Holocene reconstructions, 4:423
 planktic foraminifera, 2:827–828, 2:827*f*

- radiolarians, 2:831–833
 RSL change indicators, 4:400–401
 subdivisions of Quaternary, 4:209–213
 vertebrate fossils, 4:595–596, 4:680
 biotic response, climate change, 1:200–206, 3:790–792
 biotite, tephrochronology, 4:283–284, 4:286
 bioturbation, mammals, 2:632, 2:634f
 biozonations, mammals, 4:210–213, 4:213f
 bipolar seesaw
 sub-Milankovitch events, 3:206, 3:207
 Younger Dryas climate event, 3:129
 birds
 Australian vertebrate records, 4:623t, 4:630–631
 North America, 4:675
 of prey, 4:746
 South America, 4:683–684
 birefringence, glacial sediment, 2:58f, 2:59, 2:60
 Birmingham, England beetle assemblages, 1:187
 Bishop Tuff, California, 2:480f, 2:481
 bison
 mummified, Alaska, 4:714–715
 RLB land mammal age, 4:673, 4:677f
 Villafranchian–Galerian transition, 4:639, 4:641
Bison priscus (steppe bison), 4:714–715
 bivalves
 freshwater mollusks, 3:281, 3:282f, 3:283t
 unionid biogenic carbonates, 1:343–346, 1:343f, 1:344f, 1:345f
 black carbon, charcoal, 4:354, 4:355f, 4:359
 black-footed ferret, 4:715
 black scavenger flies, 4:551–552
 blade technology
 social brain evolution, 1:55
 Upper Paleolithic, 1:147
 BLAG hypothesis, tectonic uplift, 2:128
 Blancan age, North America, 4:646–647
 von Blanckenburg F, 2:131–132, 2:132f
 BLB *see* Bering Land Bridge
 blockfields, 3:523–534
 age/origin, 3:528–530
 characteristics, 3:523–528
 clast angularity, 3:528
 dating, 3:532
 model unification, 3:530
 particle size, 3:524–526
 profile characteristics, 3:523–524
 surface characteristics, 3:523
 trimlines/paleoununataks, 1:918–919, 1:922f, 1:923, 1:925, 1:926f, 1:927f
 block slumping, periglacial landforms, 3:493f, 3:494
 block streams, 3:514–522
 active streams, 3:517–518, 3:519–521
 age of relict streams, 3:521–522
 characteristics, 3:514–517
 climatic significance, 3:521, 3:521f
 dimensions, 3:514
 formation processes, 3:518–519
 geographical distribution, 3:514
 macroglaciation, 3:518
 morphology, 3:514–515
 movement rates, 3:519–521
 plan form, 3:514, 3:515f
 six-stage formation process, 3:519
 structure/composition, 3:515–517
 thermal regimes, 3:517–518
 Blombos Cave, Africa, 1:94
 blow flies, 4:549–550
 blowouts, dune fields, 1:599, 1:600–601
 blue bands, Barnes ice cap, 2:424
 blue ice areas (BIAs), Antarctica, 2:416, 2:418f
 meteorites, 2:416–417
 volcanic tephra layers, 2:417, 2:419f, 2:420f
 blue thermoluminescence dating, 2:651
 Blytt–Sernander scheme, Holocene, 4:173–174, 4:174f, 4:174t
 boats, 1:520, 1:520f
 see also ship...
 body size changes, vertebrate records, 4:669
 bog bodies, 3:713
 Bogoria Lake, Kenya, 2:494f
 bogs
 oribatid mite deposits, 2:693–694
 peat bog plant macrofossils, 3:593–594, 3:594f
 bolide impact theory, megafaunal extinctions, 4:711
 Bolivia
 glaciations, 2:252, 2:253
 ice core studies, 2:387–388, 2:388t
 lake level studies, 2:531–532, 2:534–535
 Bolivian Altiplano, 2:387–388, 2:531–532, 2:534–535
 bomb carbon, plant macrofossils, 4:361ge
 Bonaparte Gulf sea-level change, 4:443
 bone-isotope analysis
 definition, 4:748ge
 hominid diet, 4:748
 bones
 cut marks, 1:62, 4:749, 4:751f, 4:753f
 EST dating, 2:676
 as raw materials, 4:751–752
 stable isotopes, 1:292
 stone tools and, 1:59, 1:60, 1:62, 1:64–65
 see also teeth and bones
 bone tool production, 1:65, 4:752
 Bonneville Lake, North America, 2:539–540
Bootherium bombifrons *see* helmeted muskox
 Boothia-type erratic trains, 2:81–83, 2:83f
 bootstrapping methods, till fabric analysis, 2:78
 boreal environment stable isotopes, 1:308–309
 boreal forests
 forest fires, 3:576f
 North America, 3:746, 3:750–753
 themokarst lakes, 3:577–578
 boreal summer, Indian Ocean, 3:46
 Boreal Transgression, Eurasia, 2:224
 boreal tree line, Arctic Eurasia, 3:734
 borehole optical loggers, ice cores, 2:294
 borehole temperature records, 2:298–302
 Antarctica, 2:299f, 2:301
 general concepts, 2:298
 Greenland, 2:299f, 2:300–301, 2:301f
 heat transfer in ice, 2:298–299
 inverse problem, 2:299–300
 locations off main ice sheets, 2:301
 major results, 2:300–301
 profile characteristics, 2:298–299
 quantitative paleoclimate reconstructions, 3:182, 3:183
 specific strategies, 2:300
 temperature-depth profile, 2:299
 West Antarctica, 2:301
 Borneo
 Early Middle Pleistocene vertebrate records, 4:662
 Late Pleistocene mammals, 4:695–696
 boron isotope-pH marine carbon cycle proxy, 2:849–851, 2:850f
 botanical evidence
 environmental archeology use, 3:699–724
 historical justification, 4:542
 legitimizing, 4:542
 botany, 4:542–547
 see also paleobotany
 bottom water proxies, 2:851–853
 bottom water temperature (BWT), calcareous fossils, 2:871–883
 bottom water tracers, Antarctic, 1:535
 boulder beach sea-level indicators, 4:388
 Boulton, GS, 1:755–756, 1:756f
 Bouma sequence, glaciolacustrine, 2:44–45, 2:47f
 bovids
 cave art depictions, 4:752–754
 dwarfing/gigantism, 4:739–741
 southern African Late Pleistocene, 4:664–665, 4:668
 Bowen's reaction series, 2:942f, 2:943
 Box–Cox transformation analysis, 2:724–725, 2:726f
 BP measure *see* before present measure
 BQI *see* benthic quality index
 brachypylus oribatid mites, 2:682–683
 Brahmaputra river delta, 1:700, 1:702f
 braided river bars, 2:10–11, 2:11f
 brain size, human evolution, 1:55–56
 Brazil
 vegetation pollen records, 4:156, 4:159, 4:160–161, 4:160f
 vertebrate fossils, 4:682–683
 Brazos river delta, 1:700, 1:701f
 Breiðamerkurjökull Glacier, Iceland, 1:805–807, 1:806f, 1:807f
 Breiðamerkurjökull, Iceland, Iceland, 1:760f
 brick production, diatomites, 1:504–505
 Britain/British Isles
 ice sheet, 2:230–231
 Middle Pleistocene beetle assemblages, 1:184–189
 see also England; Ireland; Scotland; United Kingdom
 British Columbia Coast Mountains, 2:273
 British Royal Navy logbooks, 3:238, 3:239f, 3:239t
 BR method *see* balance ratio method
 Brooksby, England beetle assemblages, 1:186–187

Brooks Range, North America, 2:271–272, 2:272f
 Broomfield, England beetle assemblages, 1:184–185
 Brørup Interstadial (MIS 5c) beetle assemblages, 1:200–202, 1:201f
 brown bears, 4:724–725, 4:724f
 browsers, definition, 4:680ge
 Brückner, Eduard, 1:4
 Brunhes Normal Chron, North America, 2:180, 2:182–185
 Brunhes–Matuyama boundary, 4:639, 4:641, 4:643
 Brunhes–Matuyama reversal, 2:909, 2:910f
 Bruun Rule, sea-level change, 4:389–390
 Bryozoa macrofossils, 3:652, 3:653f
 Bryson, RA, 3:109
 BSEI *see* backscattered electron imagery
 BSi *see* biogenic silica
 Buckinghamshire, England beetle assemblages, 1:188
 building industry, diatomites, 1:504–505
 buildup methods, cosmogenic nuclide dating, 1:432
 Bulgarian postglacial pollen records, 4:186–187
 bulk density, lake sediment, 3:306–310, 3:307f
 Bunker Hills, East Antarctica, 2:221
 buoyancy-driven convection, 3:457–458, 3:457f
 burial sites/buried bodies, 1:158–159, 4:521–534, 4:546
 buried paleosols, 3:364f, 3:365–366, 3:365f
 buried soils
 amorphous coatings, 3:385f
 anthropogenic features, 3:385f
 cryogenic features, 3:386f
 crystalline features, 3:384f
 identification of, 3:381
 micromorphology, 3:381–383
 microstratigraphic relations, 3:387f
 morphology studies, 3:408
 paleoenvironmental concerns, 3:381–383
 see also paleosols
 Burma *see* Myanmar
 burning biomass, 2:716–729, 4:18–19, 4:25
 Buzzsaw Hypothesis, 1:861
 BWT *see* bottom water temperature
 Byrd ice core, Antarctica, 2:398
 Byrrhidae beetles, 1:214
 Bysjön Lake, Sweden, 2:562, 2:563f

C

¹³C *see* carbon-13
¹⁴C *see* carbon-14
 ‰C_{37.4} salinity proxy, 2:778, 2:778t
 Ca/CA *see* calcium; correspondence analysis
 CAA *see* Canadian Arctic Archipelago
 Caatinga vegetation, Brazil, 4:156, 4:159
 CaCO₃ *see* calcium carbonate
 cadaver dog searches, 4:531, 4:533f
 caddisflies (Trichoptera), 3:652, 3:653f, 4:552
 cadmium nutrient proxy, 2:901–903, 2:902f
 calcareous algae, 1:346–348, 1:347f
 calcareous marine fossils, 2:871–883
 calcareous skeletons, 2:795–799
 calcareous varves, 4:584
 calcic-horizon-bearing paleosols, 2:580f
 calcification
 corals, 2:795
 sclerosponges, 2:797
 calcite
 coatings, soil, 3:383, 3:384f
 lake sediments
 environmental effects, 1:338
 global distribution, 1:333, 1:334f
 varved, 4:575–576, 4:577f, 4:578f
 Mg/Sr coprecipitation, 2:871
 non-lacustrine terrestrial studies, 1:322, 1:325f, 1:327–330, 1:329f
 speleothem formation, 1:294–295, 1:295–296, 1:300–301
 TCNs, 1:413
 varved lake sediments, 4:575–576, 4:577f, 4:578f
 calcium (Ca)
 ice core studies, China, 2:379, 2:382f, 2:383
 lake water/cladocera, 3:277
 marine carbon cycle proxies, 2:851–853, 2:851f, 2:852f, 2:853–854
 marine fossils, 2:871–883
 paleothermometry, 2:871–883
 TCNs, 1:413
 see also calcareous...
 calcium carbonate (CaCO₃)
 biogenic carbonates, 1:292
 coccolithophore production, 2:788f
 concepts, 1:291–292
 lake sediment, 1:293
 speleothems, 1:294–303
 tests, planktic foraminifera, 2:825, 2:828
 calcium carbonate compensation depth (CCD), marine diatoms, 1:528–529
 calcrete crusts, loess, 2:630–631, 2:633f
 Caledonia Fen, Southeastern Australia, 1:196
 calendar year (cal yr) measure, 3:1ge
 calibration
 carbon-14 record, 4:345–352
 data sets, 1:361ge
 error sources, radiocarbon dating, 4:326–327
 glacial troughs, 1:851–852
 pollen data, land cover, 3:892–895
 pollen–climate relation methods, 3:809–810
 sediment properties, 3:351–352
 stomatal frequency analysis, 3:615–616
 temperature
 Mg/Ca, 2:871–873, 2:875–876
 Sr/Ca, 2:878, 2:879t
 calibration data sets, definition, 1:361ge
 California, North America
 coastal diatom assemblages, 1:574
 packrat middens, 3:678t
 pollen records, 4:142–143, 4:144–145, 4:147

Callidina angusticollis macrofossils, 3:652, 3:653f
Calliergon giganteum macrofossils, 3:651f, 3:652
 Calliphoridae (blow flies), 4:549–550
Calluna vulgaris macrofossils, 3:638, 3:639f
 calving, definition, 1:909ge
 calving glaciers, 2:272, 2:272f
 cal yr *see* calendar year measure
 CAM *see* central age model
 cambering, periglacial, 3:487–488
 cambium layer, tree rings, 2:92f, 2:97
 Cambodia, 4:652t, 4:657, 4:697
 Cambridgeshire, England, beetle assemblages, 1:187
 camels, 4:689
 Campbell Island, Southwest Pacific, 2:213–214
 Camp Century ice core, 2:435–436
 Campos vegetation, Brazil, 4:156, 4:160–161, 4:161f
 Canada
 boreal forests, 3:576f
 cordillera lake level studies, 2:540–543
 dune fields, 1:598–599, 1:599f
 age determination, 1:602–603
 origins, 1:602
 types, 1:600–601, 1:600f, 1:601f
 Eugenie Glacier, 1:761f
 glacial landsystems, 1:818f
 ice wedges, 3:575f
 insect fossil studies, 1:174–176, 1:177f, 1:178f, 1:286–289
 lichenometry, 2:570f
 neoglaciation chronology, 2:273
 pollen records, 4:44–47
 retrogressive thaw slumps, 3:579f
 thermokarst lakes, 3:468f, 3:575f, 3:577f, 3:578f
 see also Baffin Island, Canada; North America; Yukon Territory, Canada
 Canadian Arctic Archipelago (CAA), 1:592, 1:593–594, 1:594f
Canis genus records, 4:600
 cannibalism, hominins, 1:65
 canonical correspondence analysis (CCA), 1:368f, 1:361ge, 2:842
C. anthracinus records, 1:393
 caprock, definition, 1:780ge
 Carabidae (ground beetles)
 Japan, 1:211, 1:213f, 1:218
 North America, 1:233, 1:288f
 South America, 1:235, 1:236–237, 1:239f
 carapaces, 3:272, 3:276f, 3:277–278, 3:271ge, 3:278–279, 3:278f, 3:271ge
 carbon
 concentration in soil charcoal, 2:722
 emissions, charcoal records, 2:727
 lake chemistry, 3:295–296
 see also carbonate...; radiocarbon dating
 carbon-13 (¹³C)
 nutrient proxies, 2:899–901, 2:900f
 paleoceanographic reconstructions, 2:899–901
 speleothems, 1:294–295
 systematics, 2:899

- carbon-14 (^{14}C)
 anthropogenic variations, 4:336, 4:342–343
 calibration, 4:345–352
 atmospheric regional effects, 4:351
 Bayesian modeling, 4:349
 comparison records, 4:345–347
 corals, 4:345–346
 curves, 4:346f, 4:347–348
 examples, 4:347f
 freshwater reservoir corrections, 4:350
 future trends, 4:351
 marine reservoir corrections, 4:349–350, 4:351
 marine/terrestrial samples, 4:350–351
 methods, 4:347–349
 nonvarved marine sediments, 4:346–347
 problems/uncertainties, 4:349–351
 radiocarbon error estimates, 4:349
 single age calibration, 4:347–348
 speleothems, 4:346
 tree rings, 4:345, 4:346f, 4:351
 varved sediments, 4:346, 4:351
 wiggle-matching, 4:348–349, 4:349f, 4:351
 carbon cycle and, 4:340–342, 4:341f
 continental aridity, 3:869
 definitions, 4:336–337
 delta-carbon-14, 4:337, 4:339–340, 4:339f, 4:340f, 4:341f, 4:342f, 4:343, 4:343f
 Earth's geomagnetic field, 4:338
 erosion-island plots, 1:433, 1:434f
 fossil fuels, 4:342
 glaciers, Australia, 2:202–204
 global vegetation patterns, 3:864–865
 ice core dating, 2:306, 2:357
 ice-margin sites, 2:424–426
 isotopic fractionation, 4:337
 natural variations, 4:336, 4:338–342
 nuclear power/weapons testing, 4:342f, 4:343
 plant macrofossils, 4:361–367
 algae, 4:363
 aquatic moss, 4:363
 archeological contexts, 4:361
 bioturbation/mechanic disturbance, 4:362
 chemical pretreatments, 4:366
 lacustrine records, 4:363
 'old wood' effect, 4:361–362
 palustrine records, 4:363
 phytoliths, 4:362
 pollen, 4:362–363
 potamogeton seeds, 4:364
 small samples, 4:365–366
 soil/dry sediment, 4:362
 specific-compound dating, 4:364–365
 sphagnum, 4:364
 storage/preparation, 4:365
 superior macrofossils, 4:364
 terrestrial macrofossils, 4:363–364
 volcanic areas, 4:362
 radioactive dating, ice cores, 2:306
 in situ production, 1:413–414, 1:414f, 2:357
 solar activity, 4:338–340, 4:340f
 speleothem dating, 1:296
 temporal variations, 4:336–344
 West African monsoon, 3:868–869
 see also atmospheric carbon-14;
 radiocarbon dating
 carbon cycle
 anaerobic decomposition, 1:723
 biosphere feedbacks, 1:722–723
 carbon-14 temporal variations, 4:340–342, 4:341f
 carbonate sedimentation, 1:722–723
 climate system interaction, 3:613, 3:619
 glacial-interglacial changes, 3:152–153
 Holocene ice core studies, 2:312, 2:314
 long-term, 1:722–723
 phytoplankton, 2:734
 see also marine carbon cycle proxies
 carbon dioxide (CO_2)
 biogeochemical role, 3:414f, 3:415, 3:416–417
 Cenozoic cooling, 2:129–130, 2:130–131
 climate effects, 1:31–32, 1:33–34, 3:414f, 3:415, 3:416–417
 deep-sea carbonate dissolution, 2:869
 firm air dating, 2:367–370
 global feedback mechanisms, 1:31–32
 ice cores, 2:311–318, 3:246–248, 3:248f
 analytical methods, 2:316–317
 reliability of data, 2:317
 Late Neogene/Early Quaternary transition, 2:154, 2:160–162
 Late Pleistocene South Atlantic, 3:30, 3:30f
 oxygen isotope measurement, 2:347
 paleoceanography overview, 2:749–750
 preparation for proportional counting, 4:313
 reconstruction, fossil leaves, 3:613–620
 terrestrial carbonates, 1:293
 terrestrial feedback, 1:726
 see also atmospheric CO_2 ; greenhouse gases
 carbon isotopes
 benthic foraminifera, 4:36
 fractionation, 2:855
 freshwater mollusks, 3:286–287
 soil carbonates, 1:329–330
 speleothems, 1:296, 1:300–301
 surficial meteoric carbonates, 1:323–325
 carbon monoxide (CO), 2:463–470
 carbonate compensation depth (CCD)
 benthic foraminifera, 2:765
 marine diatoms, 1:528–529
 North Pacific Ocean, 3:33
 carbonate-precipitating lakes
 closed-basin lakes, 1:334–335
 global spatial/age distribution, 1:333
 hard-water, 1:334
 saline lakes, 1:334–335
 carbonate stable isotopes
 lake sediments, 1:333–340
 non-lacustrine studies, 1:322–332
 nonmarine biogenic carbonates, 1:341–353
 research overview, 1:291–293
 speleothems, 1:294–303
 teeth/bones, 1:304–314
 carbonates
 accumulation
 loess, 2:630–631, 2:633f
 soils, 3:406–407, 3:406f, 3:407–408, 3:408f
 concretions, lake-levels, 2:549, 2:550f
 coprecipitation, 2:871
 deep-sea dissolution, 2:859–870
 fossils, 2:919
 fractionation, 2:907–908
 lake-level reconstruction, 2:549, 2:550f
 research overview, 1:291–293
 sediments, fresh water lakes, 2:492f, 2:493
 soil
 accumulation, 3:406–407, 3:406f, 3:407–408, 3:408f
 non-lacustrine studies, 1:322, 1:323f, 1:324f, 1:325–326, 1:326–327, 1:328–329, 1:328f, 1:329–330
 speleothems, 1:294–303
 surficial meteoric, 1:322–323, 1:323–325, 1:325f, 1:326f
 carbonation, weathering profiles, 3:394
 carbonised plant remains, 3:707f
 see also charred plant remains
 carbonyl sulfide (COS), 2:465–466
 carcase processing, hominid–animal interactions, 4:751
Carex limosa macrofossils, 3:641f, 3:642
Carex rostrata macrofossils, 3:641f, 3:642
 Caribbean documentary evidence, 3:217–218, 3:218f
 see also individual islands
 caribou, 4:715
 carnivores (Carnivora)
 dwarfing/gigantism, 4:741–742
 Early Stone Age, 1:59
 Lower Paleolithic, 4:749, 4:750f
 Pleistocene
 Africa, 4:668, 4:669–670
 distribution of remains, 4:593
 Europe, 4:642
 South America, 4:688, 4:689f
 see also individual animals/animal groupings
 carotenoids
 absorption spectra, 3:328f
 as biomarkers, 3:326
 chemical structure, 3:328f
 light-absorbing properties, 3:327–328
 long-term records, 3:337f
 structure/nature of, 3:326–327
Carpinus (hornbeam), 4:1–2, 4:7, 4:116f, 4:119
 carrion beetles
 forensic science use, 4:553
 fossil records, 1:205f, 1:212, 1:218
 Japanese sites, 1:212, 1:218
 CART pollen analysis, 3:824
 Caryophyllaceae macrofossils, 3:606f, 3:608
 CAS *see* Central American seaways
 Cascade Range, North America, 2:271f, 2:273
 catastrophic floods, 1:825, 1:830–831, 1:832f
 see also flood...
 catastrophic rock slope failure, 3:556
 catchment events, diatom records, 1:543

- catchment vegetation
lake-level changes, 3:672
pigment changes, 3:335–336
- cathodoluminescence, soil/sediment, 4:539
- cations, ice cores, 2:326
- cats, 4:688
- Caucasus region hominin archives, 1:76
- cave bears, 4:724–725
- caves
art, hominid–animal interactions, 4:752–754
cosmogenic nuclides/landscape evolution, 1:443–444
data, sea-level reconstructions, 4:460–466
fauna
Patagonia, 4:716
Southeast Asia, 4:693, 4:694f, 4:696–697
Southern Africa, 4:619
USA, 4:716
hominin remains, Middle Pleistocene, 1:74f, 1:77
oribatid mite deposits, 2:695
pollen records, North America, 4:143
speleothems, 1:294–303
vertebrate remains, 4:592–593, 4:724–725
North America, 4:675–676, 4:675f, 4:676f
Southern Asia, 4:655–662
see also sea caves
- caviomorph rodents, 4:687
- cavity filling, till, 2:64, 2:71
- CBD extraction *see* citrate–bicarbonate–dithionite extraction
- CCA *see* canonical correspondence analysis
- CCD *see* calcium carbonate compensation depth; carbonate compensation depth
- CDIs *see* composite indices
- cells of plants, stomach contents, 4:543
- Cenozoic
Arctic paleoclimate history, 3:113, 3:122–123
atmospheric circulation, 1:640–641
chronostratigraphy, 4:216, 4:216f
cooling, 2:129–132
definition, 2:216ge
erosion rates, glacial landforms, 1:860
glaciation
Europe, 2:147t
North America, 2:148t
worldwide, 2:146f, 2:149t
pseudomorphs, ice wedges/casts, 3:443–446, 3:448f
- census count temperature proxies, 2:841–848
- centennial-scale
diatom records, 1:559–560
dynamics, African lakes, 2:502–504
West Antarctic Ice Sheet, 2:453
see also high-frequency dynamics
- central Africa, vertebrate records, 4:670–671
- central age model (CAM), 2:661
- central Amazonian vertebrate fossils, 4:681–682
- Central America
lake level studies, 2:531–536
Younger Dryas climate event, 3:128
see also individual countries
- Central American seaways (CAS)
closure, 2:151, 2:152–154, 2:154–156, 2:156–164, 2:159f, 2:160f, 2:161f, 2:163f
Late Neogene/Early Quaternary transition, 2:151, 2:152–154, 2:154–156, 2:156–164, 2:159f, 2:160f, 2:161f, 2:163f
onset of Northern Hemisphere glaciation, 2:156–159
sediment records, 2:156–159
trigger for Quaternary climate, 2:154–156
- central Andes, 2:273
see also Andes
- central Argentinean vertebrate fossils, 4:683
- Central Asia
dune fields, 1:611–614, 1:613f
Early Pleistocene, 1:70–71, 4:606f
history of glaciation, 2:148
hominin settlement, 1:70–71
ice core studies/records, 2:379–386
lake level studies, 2:511
loess records, 2:585–594
Middle Paleolithic lithic production, 1:98
modern human colonization, 1:103–104, 1:106
see also Asia
- central Brazilian vertebrate fossils, 4:682
- central Europe, 2:549–557
see also Europe; *see also* individual countries
- central Greenland
heavy metal pollution, 2:344–345
microparticle fingerprinting, 2:343–344
see also Greenland
- central highlands, Mexico, 2:533
- Central Lowland, North America, 2:620, 2:625–626, 2:627
- central North America, 1:221–225, 1:222f, 1:283t, 1:285f
see also North America; United States of America
- central-southern Chilean vertebrate fossils, 4:683
- central water masses, South Atlantic, 3:28
- century-scale climate variations, Andean glaciers, 2:392
- Cerambycidae* beetles, 1:215
- Ceratophyllum* macrofossils, 3:645f, 3:646
- cereals, 3:699, 3:703f, 3:708–713, 3:713f, 3:715t
see also chaff
- Cerrado vegetation, Brazil, 4:156, 4:160
- Cervus elaphus* red deer, 4:642, 4:642f, 4:643
- cesium-137 (¹³⁷Cs) dating, 4:585–586, 4:586f
- CFA *see* continuous flow analysis
- CGMs *see* conceptual geological models
- CH₃Br *see* methyl bromide
- CH₃Cl *see* methyl chloride
- CH₄ *see* greenhouse gases; methane...
- Chad, Lake, North Africa, 2:489, 2:489f, 2:561
- chaff in environmental archeology, 3:699, 3:703–704, 3:708–713
- change drivers
forensic geology search methods, 4:524
paleohydrology, 3:256–257
- change rates, pollen records, North America, 4:120
- Changyang cave, China, 4:655–657
- channel-bar deposits, 1:663
- channel deposits, fluvial environments, 1:663–664
- channel-fill deposits, 1:663–664
- channel patterns, fluvial environments, 1:670
allogenic forcing response, 1:676
glacial–interglacial cycles, 2:120f
Pleistocene–Holocene transition, 1:679–680, 1:680f
- Channeled Scabland, United States, 1:825–826, 1:826–828, 1:826f
- Chaoborus* remains, 3:342
- CHAR *see* charcoal accumulation rate
- Characeae* algae, 1:346–348, 1:347f, 3:646, 3:647f
- characteristic radius, pollen source area, 3:872, 3:873f
- characteristic remanent magnetization (ChRM), 1:712
- charcoal
AAA pretreatment, 4:356–357
ABOX pretreatment, 4:357
Australasia, Late Pleistocene, 4:23–24, 4:24–25
charred particle analyses, 2:716–729
contamination, 4:355–356
decomposition, 2:722f, 2:723f
definitions, 4:353
‘fine charcoal’ dating, 4:357–358
hypy pretreatment, 4:357
interpreting radiocarbon dates, 4:358–359
macrofossil stand-scale palynology, 3:851f, 3:852–853, 3:852f
plasma ashing, 4:357
pollen data, 4:147–149, 4:147f
pretreatments, 4:356–358
radiocarbon dating, 4:353–360
sampling for dating, 4:356
stability, 4:353–355
stepped combustion, 4:357
structure/chemistry, 4:353
- charcoal accumulation rate (CHAR), 2:720–722, 2:721f
- charred particle analyses, 2:716–729
- charred plant remains, 3:699–702, 3:708t
see also carbonised plant remains
- Châtelperronian technocomplex, 1:147, 1:149, 1:150, 1:152
- checkered beetles, 4:553
- cheese flies, 4:551
- chelation, 3:394
- Chelonoides* turtles, 4:683, 4:684f
- chemical dating
amino-acid dating, 1:37–48
tephrochronology, 4:277–304
- chemical deposits, weathering profiles, 3:392
- chemical examination methods, soil/sediment, 4:540
- chemical impurities, ice cores, 2:319–320
- chemical index of alteration (CIA), weathering, 2:944, 3:394

- chemical measurements
ice cores, 2:326–333
photosynthetic pigments, 3:328f
- chemical proxies
marine carbon cycle, 2:849–858
marine oxygen isotopes, 2:907–914
nutrient proxies, 2:899–906
paleoceanography, 2:871–883, 2:884–898, 2:915–922, 2:923–931
- chemical remanent magnetization (CRM), 1:712
- chemical weathering
blockfields, 3:526
profiles, 3:392, 3:393f, 3:394
rocks, 3:502
saprolite, 3:393f
soil properties, 3:402, 3:408
- chemical weathering index (CWI), 3:394
- Chesapeake Bay, southeastern North America, 4:134t, 4:139f, 4:140
- Chihuahuan Desert
beetle studies, 1:225–227, 1:226t, 1:228f
insect fossil studies, 1:288–289
rodent middens, 3:676–677
- Chile
chironomid records, 1:401, 1:402
glaciations, 2:252, 2:254
lakes, 1:235, 2:531–532
habitats, 1:237f
insect fossils, 1:235, 1:236–240, 1:237f, 1:242
landscape, 1:235
Late Pleistocene, 1:235, 1:237f, 1:240–241, 4:683
South Pacific ocean records, 3:74–76, 3:75f, 3:76–77, 3:77f, 3:78f, 3:82–84, 3:84f
vertebrate fossils, 4:683
- Chilean Altiplano, 2:531–532
- Chilean channels, 1:235, 1:237f, 1:240–241
- Chillagoe, Australia, 3:214f
- China
dune fields, 1:606–611
historical climatology, 3:237, 3:238
hominin archives, 1:71f, 1:76–77
ice core studies, 2:379–386
lake level studies, 2:512f, 2:514–515, 2:517–518, 2:519, 2:520–521
loess, 2:595–605
accretionary paleosols, 2:595
chronology of, 2:596
downwind transport/deposition, 2:598–599
dust record variations, 2:600–601
endemic diseases, 2:604
environmental hazards, 2:604
erosion, 2:603–604
formation, 2:601–603
grain variations, 2:596–597
landslides, 2:604
late Quaternary record, 2:599–600
magnetic susceptibility, 3:378–379
marine isotope records, 2:596
micromorphology, 3:371
monsoon system, 2:595, 2:597f, 2:602f
paleopedology, 3:369f
paleosol stratigraphy, 2:595–596, 2:596f, 2:599f
pedostratigraphy, 4:254–255, 4:255f
rates of dust influx, 2:599, 2:601f
Red Clay, 2:601–603, 2:604
simplified dust/paleosol model, 2:600
stratigraphy overview, 4:199–200, 4:200f
topsoil loss, 2:603–604
upwind sources, 2:597–598
Loess Plateau, 4:254–255, 4:255f
modern human expansion, 1:106
paleotempestology records, 3:218–219, 3:220f
sea-level changes, 4:444, 4:493
speleothem records, 1:298, 1:300f
stone artifacts, 1:79f, 1:80f
transitional industries, 1:98
vertebrate records, 4:655–657
Younger Dryas oscillation, 3:224–225
- chironomids (Chironomidae) (nonbiting midges)
Africa, 1:361–372
Australia, 1:398–399, 1:399f, 1:403
biogeography, 1:371, 3:313–314, 3:314–315, 3:315f
biology, 1:355–356
classification, 1:356
climate change response, 1:370–371
distribution, 1:358
ecology, 1:358, 1:403
environmental change, 1:362–368
Europe, 1:386–397
causative environmental cues, 1:392–395
future developments, 1:395–396
lake ontogeny, 1:386
lake pollution, 1:386–387
Late Pleistocene, 1:373–385
summer temperatures, 1:387–392, 1:395
water quality, 1:386–387, 1:396
forensic science use, 4:552
lake habitat multiproxy approaches, 3:344
lake-level reconstructions, 2:493
Late Pleistocene Europe, 1:373–385
methodological advances, 1:368–371
mire/peat macrofossils, 3:652, 3:653f
morphology, 1:357
New Zealand, 1:399–400, 1:399f, 1:402–403, 1:403–404
overview, 1:355–360
paleoecology, Africa, 1:361–372
paleolimnology, 3:268, 3:313–314, 3:314–315, 3:315f, 3:344
postglacial Europe, 1:386–397
postglacial Southern Hemisphere, 1:398–405
sample preparation, 1:357–358
South America, 1:399f, 1:400–402, 1:402–403, 1:403–404
Southern Hemisphere, 1:398–405
future research, 1:403
hemispheric/interhemispheric comparisons, 1:402–403
modern taxonomy/ecology, 1:403
paleoclimate reconstruction, 1:400–402
paleoproductivity reconstructions, 1:402
temperature inference models, 1:403–404
taphonomy, 1:356–357
taxonomy, Africa, 1:371
temperature inference models, 1:403–404
- chitinous exoskeletons, insect fossils, 1:173
- chlorine-36, 1:414, 2:306
- chlorophylls (Chls)
absorption spectra, 3:328f
algal blooms, 3:345–346
as biomarkers, 3:326
chemical structure, 3:328f
light-absorbing properties, 3:327–328
structure/nature of, 3:326
- ChRM *see* characteristic remanent magnetization
- chromatography
HPLC, amino-acid dating, 1:44, 1:44f
ion concentrations, 2:327–328
soil/sediment in forensics, 4:540
- chronofunction, soil, 3:403–404, 3:404f
- chronology controls, pollen databases, 3:834
- chronology establishment
dendroarcheology use, 3:632–633, 3:632f
Holocene RSL, 4:426–427
human arrivals
Americas, 1:122–125
Australia, 1:110–111, 1:115f
Neanderthal demise, 1:150–152
neoglaciation, 2:270–271
varved sediments, 4:586–587
see also dendrochronology
- chronosequences
definition, 3:375ge
soil properties, 3:403–404, 3:404f, 3:405, 3:406–407
- chronostratigraphic units, definition, 4:223–225
- chronostratigraphy, 4:215–221, 4:223–225
astronomical time scale, 4:220
defining the Quaternary, 4:215–216
Gelasian Stage, 4:216
GSSPs, 4:215, 4:216–217
Holocene Series, 4:218
lithostratigraphy and, 4:227, 4:228, 4:229–232
local units/time scales, 4:218, 4:219f
marine isotope stages, 4:220
markers/horizons, 4:218–220
Pleistocene Series, 4:216–217, 4:217–218
- Chrysomelidae* beetles, 1:215
- chrysophytes, 3:340, 3:340f
- ³⁶Cl, ice core studies, 2:355
- CIA *see* chemical index of alteration
- cichlid fishes, 2:504
- CIE *see* controlled isotope exchange
- cienega pollen, 4:143
- circles, patterned ground, 3:453f, 3:454–457, 3:454f, 3:457f, 3:459–460
- circulation
modern analog approaches, 3:107–109, 3:110
North Atlantic Ocean, 3:9, 3:10f, 3:12
soil, sorted patterned ground, 3:457–458

- circulation (*Continued*)
 stable isotopes, 2:349
see also atmospheric circulation; general circulation models; meridional overturning circulation; ocean circulation; thermohaline circulation
- circulation-to-environment approaches, 3:107–109, 3:110
- circumpolar frozen fauna, 4:713–715
- cirque floor altitudes, ELA, 1:916
- cirque glaciers, 1:847–851
 distribution/development, 1:849–851
 ELA, 1:911–912, 1:912f, 1:915f
 neoglacial chronology, 2:272f, 2:275f
 processes, 1:847–849
 types, 1:848f
 vector plots, 1:850f
- cirques
 loess, Central Asia, 2:585, 2:586f
 South Island, New Zealand, 2:213f
- Ci Saat, Java, 4:653
- citrate–bicarbonate–dithionite (CBD)
 extraction, 3:377, 3:378–379
- ³⁶Cl
 bedrock surface dating, 1:437
 weathered rock dating, 1:432
- Cladium mariscus* macrofossils, 3:641f, 3:642
- Cladocera, 3:271–280
 aquatic research applications, 3:276–279
 assemblage of remains, 3:276–277, 3:276f
 calcium in lake water, 3:277
 chemical composition of remains, 3:278–279
 ehippia, 3:272, 3:276f, 3:277–278, 3:278f, 3:271ge
 eutrophication, 3:277–278, 3:271ge
 food web past changes, 3:278–279
 habitats, 3:274, 3:275t
 indicator value, 3:272–276
 lake distribution, 3:275t
 lake-level reconstructions, 2:493
 living animal to fossil, 3:272
 melanin, 3:276f, 3:278f, 3:279, 3:271ge
 optical environments, 3:279
 parthenogenetic life cycle, 3:275b, 3:271ge
 planktivorous fish indicator, 3:277, 3:271ge
 prehistoric human settlements, 3:276
 sample preparation, 3:273b
 size of remains, 3:276f, 3:277
 stable isotope analysis, 3:276f, 3:278–279
 temperature, 3:275, 3:277
 watershed disturbances, 3:276–277
- cladogenetic speciation, vertebrates, 4:724, 4:725, 4:730
- clandestine graves, 4:546
- Clark, CD, 1:896, 1:902
- Clark, JA, 4:455, 4:467, 4:468f
- classification
 chironomids, 1:356
 coccolithophores, 2:785t
 cryoturbation structures, 3:430, 3:431f
 ice wedges, 3:437f
 lithostratigraphy, 4:227–234
 micro/macroscale erosional form size, 1:866t
 phytoliths, 3:584–585
- planktic foraminifera, 2:825
- pollen data, biome approach, 3:862–863
- salinity of diatoms, 1:507
- sediments, 2:495–496, 4:583–584
 varved sediments, 4:583–584
- clast angularity, blockfields, 3:528
- clast form analysis, glacial landforms, 2:1–5
- clastic sediments, 2:489–493, 4:574–575, 4:576f, 4:577f
- clastic varves, 4:574–575, 4:576f, 4:577f, 4:583–584, 4:583f
- clathrate analysis, ice cores, 2:316–317
- clay bedrock, 3:486–487
- clay coatings, soil, 3:381, 3:386
- clay deposits
 deep-sea, 1:637ge
 periglacial slopes, 3:486–487
- clay minerals
 blockfields, 3:527, 3:527t, 3:528–529, 3:532
 terrigenous sediments, 2:943, 2:948, 2:950f
 trimlines/paleonunataks, 1:923
- cleavers, 1:59
- Cleridae* (checkered beetles), 4:553
- Clethrionomys* (bank voles), 4:725
- Cleveland Bay, Australia, 4:444
- CLIMAP project, 3:37, 3:38
 North Atlantic Ocean, 3:11–12
 paleoceanography overview, 2:749, 2:749t
 Quaternary environments, 3:149–150
 temperature proxies, 2:841, 2:844–845, 2:846
- climate
 Arctic warming, 3:497–498
 atmospheric dust, 1:729–730, 1:730f
 beech/lime tree stand-scale palynology, 3:850–852
 beetle indicators, 1:249–251, 1:275–276
 biogeochemical role of dust, 3:412–420
 biomarker proxies, 2:775–782
 block stream significance, 3:521, 3:521f
 carbon cycle interaction, 3:613, 3:619
 continental records, 1:640–641
 controls
 megafossils, 3:621–622
 vegetation, 4:120–121
 deltas and, 1:699
 dynamics, ice core studies, 2:326, 2:329
 East Asian monsoon, 2:595, 2:597f, 2:602f
 erosion/landscape evolution, 1:442–443
 Greenland
 present day, 3:760
 stable isotopes, 2:403–405, 2:405–408
 Holocene development
 Antarctic Peninsula, 2:219–220
 beetle indicators, 1:275–276
 East Antarctica, 2:220–221
 flood response, 3:255–256
 ice core studies, 2:326, 2:329, 2:353, 2:356, 2:358–359, 2:375–377, 2:387–390, 2:390–393
 introduction to, 3:87–92
 lake water oxygen isotope relation, 1:336–337
- Late Pleistocene
 Southeast Asia, 4:693
 Southern Europe, 4:66–67, 4:67–68
- Late Quaternary Pacific region, 2:202
- Middle Pleistocene
 Africa, 4:615–616
 Southern Europe, 4:66–67, 4:67–68
- modeling, 3:249–250
 biomized data, 3:867–869
 calibration, 3:249–250
 evaluations, 3:867–869
 pollen databases, 3:831
 validation, 3:249
 vegetation reconstruction, 3:861, 3:867–869
 volcanic/solar forcing, 1:749–750
- non-lacustrine studies, 1:325–326
- North Atlantic, 2:615–617, 2:749, 3:11–12
- northeastern North America, 4:115
- northern landscapes, 3:746
- pace of change, 1:17
- patterns, pollen records, 4:124
- plant functional types, 3:861–862
- pre-Quaternary, 2:28
- processes
 dendroclimatic analysis, 3:634
 fluvial terraces, 1:684, 1:689–691
 insect fossil studies, 1:274
- proxies
 biomarkers, 2:775–782
 lake-level changes, 3:672
 marine diatoms as, 2:823
 pollen as, 3:805–815
- quantification, New Zealand, 1:249–251
- reconstructions
 beetle records, 1:249–251
 continental lakes, 1:507–515
 glacier mass balance, 2:109
 ice core studies, 2:353, 2:356, 2:358–359
 plant macrofossil records, 3:785, 3:787–789
 pollen numerical analysis, 3:826
 pollen records, 3:805–815
 pollen surface sampling/trapping, 3:844
 vegetation, 3:861, 3:867–869
- research history, 1:17
- semiarid dune fields, 1:606, 1:611
- sensitivity, North Atlantic, 2:749, 3:11–12
- soil morphology relation, 3:407–408
- South Pacific postglacial records, 3:73–74
- southwestern North America, 4:142–155
- systems
 coccolithophores, 2:791–792
 Highland Asia, 2:236
- variability/variables/variations
 calcareous skeleton studies, 2:795–799
 flood response, Holocene, 3:255–256
 lake level studies
 Latin America, 2:533
 North America, 2:544–545
 west-central Europe, 2:552–553, 2:553–556
 paleoclimatic reconstruction, 1:249, 1:250f
 stable isotopes, Antarctica, 2:398–400
 timescales of climate change, 3:93–94

- vegetation changes, South America, 4:56f
- vegetation
 - controls, 4:120–121
 - linkages, 4:115
 - reconstructions, 3:861, 3:867–869
 - South America, 4:56f
- Vostok ice core records, 2:289–291, 2:290f
- weathering profiles and, 3:397–398
- see also* dendroclimatology; paleoclimate...
- climate change
 - abrupt
 - Andean glaciers, 2:392
 - fluvial environment responses, 1:676–683
 - ice core studies, 2:284, 2:308, 2:392
 - New England vegetation history, 4:121
 - North Atlantic, 3:49
 - thermal diffusion paleotemperature records, 2:433
 - varved sediment applications, 4:588
 - active layer, permafrost, 3:426–427
 - Africa, 1:370–371, 2:499–502, 4:664
 - Andean glaciers, 2:392–393
 - Antarctica, 2:220
 - fossil groups, 1:527
 - ice core, 2:410–415
 - sea-surface conditions, 1:530
 - anthropogenic, 3:244–248, 3:880–881
 - beetle fauna, Late Pleistocene, 1:200–206, 1:241, 1:242
 - calcareous skeleton studies, 2:795–799
 - chironomid records
 - Africa, 1:370–371
 - Late Pleistocene, 1:373–385
 - overview, 1:358, 1:359–360
 - continental configurations, 2:127–128, 2:128–129, 2:129–132, 2:131f
 - deep-sea sediments, 1:640
 - diatom-inferred, 1:543–544, 1:575–580
 - diatom silica proxies, 2:741f
 - eustatic sea-levels, 4:498
 - fire/vegetation interaction, 2:717, 2:718f, 2:727
 - glacial–interglacial fluvial responses, 2:112–125
 - global paleotempestology, 3:215–217
 - Greenland, 2:410–415, 2:439, 3:761
 - history of, 1:17
 - human evolution and, 1:52–53, 1:54
 - human impact, 3:244–248, 3:880–881
 - ice cores
 - China, 2:379, 2:380–382, 2:385
 - Greenland/Antarctica, 2:410–415
 - research history, 2:435–438
 - ice-margin sites, 2:416–430
 - indices of change in past 50Ma., 2:131f
 - lake level studies, 2:558
 - Africa, 2:499–502
 - Asia, 2:511, 2:512–515, 2:516–519, 2:520–521
 - Latin America, 2:531, 2:535
 - Lateglacial/Holocene Greenland, 3:761
 - loess deposits and, 2:573
 - low-latitude sand seas, 1:623, 1:625f, 1:628f, 1:635f
 - migration/dispersal of species, 4:208–209
 - modeling
 - millennial climate change, 3:206–207
 - sub-Milankovitch events, 3:206–207
 - Neanderthal demise, 1:150
 - non-lacustrine studies, 1:327–330
 - nonmarine biogenic carbonates, 1:349
 - North Atlantic Ocean, 3:49
 - pace of, 1:17
 - paleoceanography overview, 2:749–750
 - paleophytogeography, 2:730
 - permafrost and, 3:426–427, 3:469–470
 - pigment studies, 3:336
 - plant macrofossil records, 3:785–793
 - pollen records
 - Late Pleistocene Australasia, 4:18–26
 - northeastern North America, 4:116, 4:117, 4:121
 - overview, 2:700, 2:701
 - Western North America, 4:72–83
 - qualitative indicators, 1:373–374
 - quantitative indicators, 1:374–383
 - reconstruction, thermal diffusion records, 2:433, 2:433f
 - regional, last millennium, 3:229–231
 - rock glaciers, 3:539–540
 - Sahul landmass, 1:108–110
 - soil morphology, 3:404, 3:405
 - Southeast Asia Late Pleistocene, 4:693
 - tectonic uplift, 2:127–128, 2:128–129, 2:129–132, 2:131f
 - THC role, 1:744–745
 - thermokarst and, 3:580
 - timescales, 3:93–101
 - treeline studies, 3:690, 3:693, 3:695–697
 - understanding change, 3:87–88
 - varved sediments, 4:582, 4:587, 4:588
 - see also* environmental change; global warming; paleoclimate...
- climate cycles, dust, 3:412–420
- climate forcing
 - biomass burning, 2:725
 - biosphere feedbacks, 1:721–728
 - data-model comparisons, 3:137, 3:138t
 - glacial climates
 - THC, 1:737–747
 - volcanic/solar forcing, 1:748–753
 - human evolution, 1:51–53
 - Last Glacial Maximum GCMs, 3:166–167, 3:166t
 - neoglaciation in Europe, 2:265–266
 - paleodroughts, 3:196–197
 - paleo-ENSO, 3:174–176
 - pollen records, Australasia, 4:25
- Climate Long-Range Investigation, Mapping and Prediction *see* CLIMAP project
- climatic change *see* climate change
- climatic equilibrium line, definition, 1:910–911, 1:909ge
- Climatic Optimum
 - Antarctic Peninsula, 2:220
 - Greenland borehole temperature records, 2:301
 - Victoria Land, Antarctica, 2:221
- climatostratigraphy, 4:222–226
 - boundaries, 4:223
 - global correlation, 4:225–226
 - nomenclature, 4:222–223, 4:225
 - terminology, 4:222–223
 - units, 4:223–225
- climosequences, soil, 3:404–405, 3:407
- clinoform morphology, deltas, 1:697, 1:698f
- cloning ancient plant DNA, 2:711
- CLORPT approach, soil morphology, 3:403–405
- closed-basin lakes
 - Asia, 2:506, 2:512f
 - carbonate precipitation, 1:334–335
 - climate change effects, 2:558
- closed-system pingos, 3:472–474, 3:473f, 3:476f
- clothing in forensic science, 4:545
- clouds, atmospheric dust, 1:729, 1:735
- Clovis peoples, America, 1:119–121, 1:122, 1:124–125, 1:127–128
- Clovis points, 1:119–121, 1:123f
- clown beetles, 1:212, 1:215f, 1:218, 4:552–553
- CMB *see* core-mantle boundary
- CO *see* carbon monoxide
- CO₂ *see* atmospheric CO₂; carbon dioxide
- coalescent subaqueous fans, moraines, 1:776
- Coast Mountains, British Columbia, 2:273
- coasts
 - Antarctic Peninsula deglaciation, 2:219–220
 - erosion, 1:18–19, 1:19f
 - hazards, tectonic, 4:503–519
 - high-energy, 4:385–395
 - neoglacial chronology, 2:273
 - North American pollen records, 4:142, 4:144–145
 - northern Africa, insularity, 1:96–97
 - paleolimnological applications, 3:658–659
 - paraglacial land systems, 3:562–563
 - postglacial
 - human adaptations, 1:154, 1:156–157, 1:158
 - pollen records, 4:142, 4:144–145
 - proxy records, paleotempestology, 3:211–213
 - upwelling, South Atlantic, 3:21–27
- Coccinellidae beetles, 1:215
- coccolithophores, 2:783–794
 - accumulation mechanisms, 2:788–789
 - biochronology, 2:792
 - biology, 2:783
 - biostratigraphy, 2:792, 2:793f
 - classification, 2:785t
 - climate systems, 2:791–792
 - distribution in water column, 2:786f
 - ecology, 2:784–788
 - Late Pleistocene South Atlantic, 3:20, 3:21f
 - latitude, 2:784–786, 2:789–790
 - life cycle diagram, 2:785f
 - light effects, 2:784–786
 - major types, 2:783
 - morphological types, 2:783
 - nutrients, 2:786–788, 2:787f
 - as paleoceanographic indicators, 2:789–791

- coccolithophores (*Continued*)
 paleoproductivity reconstruction,
 2:790–791
 Pleistocene, 2:791, 2:792, 3:20, 3:21f
 production, 2:788–789, 2:788f
 as proxies, 2:783–794
 seasonality, 2:784
 stability of environment effect, 2:788
 temperature effects, 2:784–786,
 2:789–790, 2:790f, 2:791f
 coccoliths, 2:756–757, 2:757f, 2:758,
 2:760–761
see also haptophyte algae
Coelodonta antiquitatis *see* woolly rhinoceros
Coelodonta–Mammuthus faunal complex,
 4:641, 4:644
 coexistence approach, paleoclimatic, 1:191–192
 cognition, hominins, 1:89
 coherent occurrences, climate change, 3:49
 COHMAP *see* Cooperative Holocene
 Mapping Project
 cold-adapted Chironomidae, 1:373
 cold-based glaciers, 1:833
 cold-based ice
 glacier flow directions, 1:898–899
 permafrost–glacier interactions,
 3:507–508, 3:510
 cold-climate environments, definition,
 1:597ge
 Cold Desert pollen records, 4:130
 cold-region landforms, thermokarst,
 3:574–581
 Cold Stage correlations, beetle assemblages,
 1:189
 Cold tongue, North Pacific, 3:62–64
 cold-water diatoms, Pacific Ocean, 1:573f,
 1:574f
 Coleoptera *see* beetle fauna studies/records
 collaborative trials, radiocarbon dating,
 4:326, 4:327f
 collapsibility
 loess, 2:594
 marine-based ice sheets, 2:449
 palsas/lithalsas, 3:477–478
 pingos, 3:475–476, 3:476f
 collared pika, 4:715
 Colombia
 pollen records, 4:58–59, 4:157–158,
 4:159f, 4:162
 savanna area, 4:58–59
 vertebrate fossils, 4:680–681
 colonization
 Australia 300,000–8000 years ago,
 1:108–118
 Europe
 MPT, 1:84–87
 post-MPT, 1:87
 pre-MPT, 1:84
 new territories, modern humans,
 1:103–104
 colonizing plants, Greenland, 3:765–766
 color of soil, 3:402–403, 3:404f, 3:405–406,
 3:406f, 4:537
 Colorado Desert sand seas, 1:634
 Colorado Plateau rodent middens,
 3:680–681
 Colorado river
 delta site case, 1:700, 1:701f
 glacial–interglacial scale responses, 2:124f
 sediments, 1:671f, 1:673f, 1:674f
 Columbia Plateau, North America, 2:620,
 2:622–624, 2:623f, 2:625f, 2:626, 2:627
 combustion continuum, charcoal, 4:353,
 4:354f, 4:356, 4:358
 combustion reaction, vegetation, 2:716
 comminution till, 2:68–69
 common chronology comparisons, ice cores,
 2:412–414
 common source, soil/sediment, 4:537, 4:538
 comparative approach, vegetation
 reconstruction, 3:891–892, 3:894f
 complex response, fluvial environments,
 1:678–679
 complex spatial structures, vegetation,
 3:871–872
 complex systems, soils, 3:410–411
 composite indices (CDIs), carbonate
 dissolution, 2:867
 composite paleosols, 3:367
 composite ridge structures/landforms,
 1:840–843
 composite tools, definition, 1:83ge
 composite veins, ice wedges/casts, 3:440
 composite wedges, ice, 3:440, 3:443–444,
 3:443f, 3:446, 3:447t, 3:450f
 compression wood, 2:96–97, 2:96f, 2:98f
 computer numerical analysis, pollen, 3:827
 conceptual geological models (CGMs),
 4:525–527, 4:527f
 Conceptual Linear Aggregate Model, 1:460
 conceptual models
 dendroclimatology, 1:460
 forensic geology, 4:525–527, 4:527f
 lakes, 2:558
 conductivity
 definition, 1:361ge
 ice core studies, 2:319–325
 confrontational replacement theory, 1:150
 Congolian forests pollen records, 4:91–95
 Congo river, Late Pleistocene, 3:22–24
 conifer tree rings, 3:631f
 conservation
 human impact and, 3:881
 of mass, 4:453–455
 constrained incremental sum-of-squares
 clustering (CONISS) method, 1:491
 consumption activities, archeology,
 3:713–714
 contamination
 animal multiproxy approaches, 3:346–347
 atmospheric deposition, 3:334–335
 see also pollution
 continental aridity, Northern Hemisphere,
 6000 ¹⁴C years BP, 3:869
 continental biostratigraphy, 4:206–214
 continental climate, Cenozoic, 1:640–641
 continental configurations, tectonic uplift,
 2:127–135
 continental ice sheets, 2:908–909
 continental interiors
 eroding, 1:670–671
 glacial–interglacial cycles, 2:115–119
 continental lakes, 1:507–515
 continental margins
 deltas, 1:697
 dinocysts, 2:812
 glacial–interglacial cycles, 2:119–125
 subsiding, 1:671–674
 see also continental shelf sites; plate
 margins
 continental shelf basins, 1:555, 1:556f,
 1:557–559
 continental shelf sites
 deltas, 1:697
 glacimarine sedimentation, 2:33–35, 2:36f,
 2:37f, 2:38f
 high-latitude, 2:32f, 2:33–35, 2:36f, 2:37f,
 2:38f
 iceberg scour, 2:32f
 laminated sediments, 1:555, 1:556f,
 1:557–559
 relative sea level changes, 4:496, 4:499,
 4:500f
 subaqueous glacial depositional systems,
 1:763
 Continuity Model, human colonization,
 1:113
 continuous flow analysis (CFA), 2:327–328
 continuous studies, pollen–climate relation,
 3:813–814
 contraction fans, glacier flow, 1:897
 contraction response, wet forest biota, 1:198
 controlled isotope exchange (CIE), 1:482
 controlled moraines, 1:774–775, 3:509
 control samples, soil/sediment, 4:538
 convection, firn air, 2:365
 conventional method, radiocarbon dating,
 4:305–315
 convergent oceanic frontal zone, 1:556f
 cooling
 Antarctic Peninsula, 2:220
 Cenozoic, 2:129–132
 Last Glacial Maximum GCMs, 3:167,
 3:168
 Cooperative Holocene Mapping Project
 (COHMAP)
 data-model comparisons, 3:136–137
 paleoclimate relevance, 3:249, 3:250
 pollen databases, 3:831–832
 Coorong coastal plain, Australia, 4:432,
 4:434, 4:435f
 coprecipitation, carbonates, 2:871
 coprolite, 2:708f
 coralline sponges *see* sclerosponges
 corallite, definition, 2:795
 corals, 2:795–799
 carbon-14 record calibration, 4:345–346
 conglomerates, 4:383–384, 4:383f
 ESR dating, 2:668, 2:672–673, 2:674f
 extension rate, 2:880
 reefs
 relative sea level changes, 4:495, 4:496f
 U-series dating, 4:567–568
 sea-level change
 conglomerates as indicators, 4:383–384,
 4:383f
 response to, 4:505f, 4:506f
 RSL records, 4:409–418, 4:495, 4:496f

- secular variability, 2:880–881
 Sr/Ca paleothermometry, 2:871, 2:878–881, 2:879*t*
- Corbicula* biogenic carbonate, 1:343, 1:345–346, 1:348*f*
- Cordilleran ice sheet, 2:245, 2:246*f*, 2:247–248, 2:248*f*
- cordilleras, America, 2:269–276
- core-mantle boundary (CMB), 1:706–707, 1:709*f*
- cores/coring
 Acheulean Industry, 1:59, 1:64*f*
 definitions, 1:83*ge*
 Developed Oldowan, 1:62, 1:63*f*
 Early Pleistocene Asia, 1:67–68
 Middle Pleistocene Asia, 1:77
 Oldowan Industry, 1:59, 1:60–61, 1:61–62, 1:62*f*
 paleolimnology, 3:262–264, 3:262*f*, 3:263*f*
 sea ice, Arctic, 1:589*f*
 technocomplex groupings, 1:147
 tree-ring dating problems, 2:107*f*, 2:108*f*
 varved sediments, 1:540, 4:584–585
see also ice core studies/records
- Coriolis effect, glaciolacustrine, 2:43–44
- cornices as sea level indicators, 4:382–383
- correlation
 ice core records, 2:357–358, 2:410–415
 stratigraphy overview, 4:190–192, 4:195–201, 4:202*f*
- correspondence analysis (CA), 1:361*ge*
 fossil diatom assemblages, 1:491
- Corsican pollen records, 4:180*t*, 4:183–185
- COS *see* carbonyl sulfide
- cosmic rays/radiation
 arrival direction definition, 1:423*f*
 composition, 1:419
 energy spectrum, 1:419, 1:426*f*
 flux
 definition, 1:429–430
 modulation, 1:430
 mineral interactions, 1:418–431
 nuclear disintegration rates, 1:425–426
 omnipresent energetic nuclear radiation, 1:419
 physics, units used in, 1:429
 secondary particle development, 1:422–427
 sun–climate relation, 2:358
see also galactic cosmic rays
- cosmogenic exposure history, blockfields, 3:532, 3:533*t*
- cosmogenic isotopes
 atmospheric carbon-14, 4:333, 4:334
 dating techniques
 neoglaciation, 2:271
 trimlines/paleonunataks, 1:924–925
 ice-core research, 2:283
 ice-margin sites, 2:424–426
 surface exposure dating, 1:924–925
see also carbon-14
- cosmogenic nuclides
 buildup/decay, 1:440–441
 caves, 1:443–444
 dating techniques, 1:410–417, 1:449
 Cronus calculator, 1:409
 exposure dating, 1:432–439
 Australian glaciers, 2:202–204, 2:204–205, 2:206
 blockfields, 3:532–533
 cosmic ray–mineral interactions, 1:418–431
 fluvial terraces, 1:684, 1:686, 1:686*f*
 North American glaciations, 2:185
 Southern Hemisphere glaciation, 2:187, 2:188
 glacier/icesheet extent, 1:892
 methods, 1:14–15, 1:15*f*
 overview, 1:407–409
 radiometric, 1:440–445
 TCN applications, 1:408–409
 definition, 1:440*ge*
 erosion, 1:441–443
 landscape evolution, 1:440–445
 radiometric dating, 1:440–445
 river incision, 1:443–444
 valley lowering rates, 1:443–444
see also cosmic rays/radiation; exposure geochronology; landscape evolution; terrestrial cosmogenic nuclides
- cosmogenic proxies, volcanic/solar forcing, 1:748–749
- cosmogenic radionuclides (CRNs)
 applications, 2:356–358
 basics, 2:353–356
 dating techniques, 1:449
 soil morphology, 3:410–411
 South American glaciations, 2:253
 deposition processes, 2:354–355, 2:358
 in ice cores, 2:353–360
 postdepositional processes, 2:355–356
 production, 2:353–354, 2:356, 2:357
 transport processes, 2:354–355
- counterterrorism search methods, 4:524
- counting method, leaf stomata, 3:614*f*, 3:615
- coup de grâce* model, megafaunal extinctions, 4:711
- Cowley Beach, Australia, 3:214*f*
- crag and tails erosional form, 1:873, 1:875*f*
- crannogs, 1:518–519
- creep, rock glaciers, 3:537–538
- Crescent Island Crater, Kenya, 1:363, 1:364*f*
- Cretan otters, 4:743*f*
- Creten pollen records, 4:181*t*, 4:186–187
- crevasse-channel sediments, 1:666–667
- crevasse-splay sediments, 1:666–667, 1:667*f*
- crevasses, West Antarctic Ice Sheet, 2:450–451
- crime scene investigations, 4:545
- criminal investigations, 4:521–534
- CRM *see* chemical remanent magnetization
- CRNs *see* cosmogenic radionuclides
- crocodiles (*Crocodylus*)
 Australian vertebrate records, 4:621–628
 dwarfing/gigantism, 4:745–746
 Late Pleistocene North America, 4:678*f*
- Croll, James, 1:4–5, 1:5*f*, 2:307
- Croll's orbital theory, glaciation, 1:4–5
- Cromerian complexes, vertebrate studies, 4:639
- Cromerian–Elsterian stage mammal assemblages, 4:611–612
- Cronus calculator, 1:409
- CRONUS-Earth project, 1:410, 1:414–415, 1:414*f*, 1:415–416
- CRONUS programs, cosmogenic nuclide dating, 1:408
- crops
 environmental archeology, 3:714–715
 pollen records, 3:887–889
 processing
 economics, 3:708–713, 3:713–714
 stages, 3:718*f*, 3:720*f*
see also cereals
- cross-cutting ice flow patterns, 1:896–902, 1:898*f*, 1:905*f*
- cross-dating principle, dendrochronology, 1:455, 1:456, 1:456*f*, 2:93*f*
 archeological applications, 3:630
 glacial landforms, 2:104, 2:106*f*
- cross-profiles
 glacial troughs, 1:852, 1:855*f*
 rock basins, 1:858*f*
- cross-sections
 nuclear interactions, 1:430
 tree-ring dating, 1:454*f*
- crossing fluxes, cosmic rays, 1:429–430
- Crotone Basin, Italy, 4:271, 4:272*f*, 4:273*f*
- crustacea, *see also* cladocera crusts, soil, 3:383–386
- cryogenic features, soil, 3:381–383, 3:386*f*
- cryohydrostatic flow structures, 3:432–433, 3:432*f*, 3:433*f*
- cryolithology, 3:546–548, 3:546*f*, 3:548–549
- cryolithosphere, definition, 1:780*ge*
- cryosphere history, 1:20
- cryostatic heave structures, 3:433–434
- cryosuction processes, 3:423–424
- cryoturbation
 active layer, permafrost, 3:424
 European loess records, 2:607–608, 2:613
 Pleistocene periglacial environments, 3:495, 3:496, 3:497*f*
 structures, 3:430–435
 classification of, 3:430, 3:431*f*
 origin of, 3:430–434
- cryptotephra, 4:277*ge*, 4:280–281, 4:299–301
- crystalline features
 buried soils, 3:384*f*
 Greenland ice sheet, 2:442–443
see also calcite coatings; ice crystals
- crystallinity, deep-sea carbonate dissolution, 2:867
- ¹³⁷Cs dating *see* cesium-137 dating
- Cuddie Springs, Australia, 1:114, 1:114*f*, 4:703–704
- Cudmore Grove, Essex, England beetle assemblages, 1:188
- cultivation of crops
 environmental archeology, 3:714–715
 pollen records, 3:887–889
see also plant growing
- cultural landscape history, 3:880–903
- cultures
 changing, 1:57, 2:535
 human evolution, 1:57
 Neanderthal vs. modern human, 1:147–150

cumulative mass balance, 1:909*ge*
 cupola hills, 1:843–844
 Curculionidae beetles, 1:216
 current systems, North Pacific Ocean, 1:572*f*
 cut marks
 antlers, 4:751, 4:752*f*
 bones
 hominid activity, 4:749, 4:751*f*, 4:753*f*
 stone tools and, 1:62
 cut terraces
 aggradation, 1:684, 1:685*f*
 formation, 1:688*f*, 1:690*f*
 cutbanks, plant macrofossils, 3:768, 3:775
 cutoff rigidity, cosmic ray flux, 1:430
 cutting dates, dendroarcheology, 3:630
 Cuvier, Georges, 1:2–3, 1:2*f*
 CWI *see* chemical weathering index
 cyanobacterial blooms, 3:332–333
 cyber-infrastructure, pollen databases, 3:838
Cycladophora davisiana radiolarian, 2:831*f*, 2:832, 2:834*f*
 cyclones, tropical, 3:209–221
 Cyperaceae macrofossils, 3:604*f*, 3:608
 Cypriot dwarf hippopotamus, 4:736–737, 4:738*f*
 cysts, dinoflagellates, 2:800, 2:801–803, 2:803–805, 2:805–811, 2:811–812

D

Dahomey Gap, West Africa, 4:11
 daily timescale, West Antarctic Ice Sheet, 2:454
 Dalmatian Islands pollen records, 4:181*t*, 4:186–187
 d-amino acids, 1:38, 1:39, 1:39*f*, 1:43*f*, 1:44, 1:46*f*, 1:47–48
 Dansgaard, Willi, 2:403, 2:405, 2:407
 Dansgaard–Oeschger (D–O) cycles
 eustatic sea-level change, 4:432
 magnetic susceptibility, 2:891–893
 marine diatoms as proxies, 2:822
 sub-Milankovitch events, 3:200–208
 Dansgaard–Oeschger (D–O) events, 2:410, 2:750–751
 Arctic paleoclimate history, 3:117–118
 common chronology comparisons, 2:412, 2:413, 2:414
 Greenland stable isotopes, 2:403, 2:407
 methane studies, 2:338–339
 nitrous oxide history, 2:474–475, 2:474*f*
 North Atlantic Ocean, 3:13–14, 3:14*f*
 THC, 1:744, 1:745*f*
 thermal diffusion paleotemperature records, 2:433
 see also abrupt climate change/events
 Dansgaard–Oeschger style forcing, fluvial environments, 1:680–681
 DAR *see* diatom accumulation rate
 dasyurid vertebrate records, 4:632
 databases
 charred particle analyses, 2:716–717, 2:719*f*, 2:724–725
 plant macrofossils/pollen, 3:731–732
 pollen methods/studies, 3:731–732, 3:831–838
 applications, 3:836–837
 history, 3:831–833
 nomenclature/taxonomy, 3:834–836
 quality control, 3:834
 structure, 3:833–834
 data collection
 deep-sea sediments, 1:638
 till fabric, 2:76
 data expression, plant macrofossils/pollen, 3:727–729
 data interpretation, 1:489–500
 deep-sea sediments, 1:638–641
 data-model comparisons, 3:135–146
 climate forcing, 3:137, 3:138*t*
 controls on northern tree-line, 3:136*f*
 data syntheses requirements, 3:137–141
 history of, 3:136–137
 iterative processes, 3:145*f*, 3:146
 large scale/global data syntheses, 3:140*t*
 mid-Holocene African monsoon, 3:139*f*, 3:142, 3:144*f*, 3:145–146
 model benchmarking requirements, 3:141–142, 3:144*f*, 3:145–146
 paleoclimate modeling, 3:135–146
 PMIP project boundary conditions, 3:138*t*
 requirements, 3:137–141
 data presentation, till fabric, 2:76
 data reliability, ice core studies, 2:317
 data sets
 analysis, paleoclimate, 3:89–90
 calibration, 1:361*ge*
 existing paleoenvironmental sets, 3:142–146
 Global Charcoal Database, 2:724–725
 see also pollen, data
 data tracking, plant macrofossils/pollen, 3:729–731
 date attributes, dendroarcheology, 3:630–631
 date clusters, definition, 3:632–633
 dating methods/techniques, 1:9–16
 amino-acid dating, 1:37–48
 amino acid racemization, 1:15, 1:15*f*, 1:450
 annual layer methods
 ice cores, 2:323
 varved sediments, 4:573–581, 4:582–589
 argon–argon dating, 1:10–11
 Australian glaciers, 2:202–204
 biological methods, lichenometry, 2:565–572
 blockfields, 3:532
 block streams, 3:521–522
 burials, 1:443–444
 chemical methods, amino-acid dating, 1:37–48
 comparative methods, 1:451
 continental biostratigraphy, 4:206–214
 coral records, 4:409–411, 4:412–413, 4:413–414, 4:414–416, 4:414*f*
 dendroarcheology, 3:630–636
 dendrogeomorphology, 2:91–103
 early developments, 1:9
 environmental archeology, 3:699
 ESR dating, 1:13, 1:14*f*
 firn air, 2:366–367
 fission track dating, 1:11–12
 fluvial terraces, 1:684, 1:686, 1:686*f*, 1:690–691
 geomagnetic excursions, 1:705–720
 history of, 1:6–7, 1:9–16
 human arrivals, Australia, 1:110
 ice cores, 2:323–324
 annual layer methods, 2:323
 cosmogenic radionuclides, 2:357–358
 radioactive isotopes, 2:305–306, 2:305*t*
 laminated sediments, 4:585–586
 lichenometry, biological methods, 2:565–572
 loess records, 2:573, 2:577
 Central Asia, 2:589–590
 Europe, 2:609–610
 luminescence dating, 1:12–13, 1:12*f*
 marine oxygen isotope stratigraphy, 2:909–912
 Middle-Pleistocene glaciation, 2:182–185
 Neanderthals
 extinction, 1:150–152
 vs. modern human technocomplexes, 1:147
 neoglaciation
 chronology establishment, 2:270–271
 Europe, 2:257–259, 2:259*f*
 nonradiometric, 1:450*t*
 overview, 1:447–452
 paleoclimate introduction, 3:88–89, 3:90
 paleolimnological analyses, 3:264, 3:265–266
 paleosols, 3:368–370
 past 100 years, 1:14
 plant macrofossils, Greenland, 3:760
 potassium–argon dating, 1:10–11, 1:10*f*
 qualitative methods, 1:451
 radiative dosimetry methods, 1:449–450
 radioactive decay, 4:316
 radioactive isotopes, ice cores, 2:305–306, 2:305*t*
 relative-age dating
 lichenometry, 2:566
 trimlines/paleonunataks, 1:921–924
 rock glaciers, 3:539
 saline lake reconstructions, 2:496
 sand seas, 1:627, 1:634*f*
 scale forms, glacial troughs, 1:857–859
 sea level studies, 4:370, 4:372
 coral records, 4:409–411, 4:412–413, 4:413–414, 4:414–416, 4:414*f*
 tropics, 4:495
 secular variation, 1:705–720
 sediment analyses, lake-levels, 2:549–551
 soil morphology, 3:410–411
 South American glaciations, 2:253
 Southern Hemisphere glaciations, 2:187, 2:188

- speleothems, 1:296–297
 surface exposure, trimlines/paleonunataks, 1:924–925
 teprochronology, 4:277–304
 time scale of methods, 1:448f
 uranium series methods, 1:13–14, 4:567–571
 weathering profiles, 3:399–400
see also chronologies; cosmogenic nuclides, dating techniques; cosmogenic radionuclides, dating techniques; dendrochronology; luminescence dating; radioactive nuclide dating methods; radiocarbon dating; radiometric dating
- datum levels, biostratigraphy, 1:584f, 1:585
 Davis, MB, 2:700–701
 Davis Strait, Arctic, 1:567
 DCA *see* detrended correspondence analysis
 D/C conductivity, ice cores, 2:320–321
 DCFT *see* diameter-correction fission-track dating
- debris cones, 3:554–556, 3:559–560
 debris flow
 high-latitude continental shelves, 2:38f
 slope deposits/forms, 3:485
 subaqueous glaciolacustrine, 2:44–45
 tree growth reactions, 2:95f, 2:100, 2:100f, 2:101f
 see also ice-rafted debris
- debris slopes *see* talus slopes
- decadal timescale, West Antarctic Ice Sheet, 2:453–454
 decadal variability, coral Sr/Ca, 2:880–881
- decay
 cosmogenic nuclide dating, 1:432
 counting in delta-carbon-14 calculation, 4:337
 ice sheets, 1:877–883
 radioactive decay, 1:447, 4:316
 radioisotope proxies, 2:923, 2:924t, 2:925, 2:927
- decomposition
 bodies in forensic entomology, 4:548, 4:549t
 charcoal, 2:722f, 2:723f
- deep circulation, North Atlantic Ocean, 3:9
 deep ice cores, Asia, 2:380–385, 2:381f
 Deep Lake, Adirondack Mountains, 3:340–341, 3:341f
 deep-lake drilling, Asia, 2:508–511, 2:519
- deep-sea
 carbonates
 anthropogenic CO₂, 2:869
 atmospheric CO₂, 2:868–869
 budgets, 2:868
 CDIs, 2:867
 changes through time, 2:867–869
 chemical indices, 2:867
 compensation, 2:868
 content decrease, 2:865
 crystallinity, 2:867
 DIC, 2:859–860, 2:861, 2:862f
 dissolution, 2:859–870
 extent of, 2:864–867
 kinetics, 2:863
 mechanisms, 2:862–863
 proxies, 2:864–867
 effect on other proxies, 2:867
 foraminifers, 2:866, 2:866t
 fragmentation indices, 2:865
 future aspects, 2:869
 glacial-interglacial changes, 2:868–869
 global distribution, 2:859, 2:860f
 grain-size indices, 2:865–866
 observational indices, 2:866–867
 parameters, 2:862
 sedimentary lysocline, 2:863–864, 2:865f, 2:869f
 species composition, 2:865
 states of preservation, 2:867f
 supralysocline dissolution, 2:864
 surface production, 2:859
 TA, 2:860–861, 2:862f
 thermodynamics, 2:859–862, 2:862–863
 ice-rafted debris, Heinrich layers, 2:37, 2:39f, 2:40f
 oxygen isotopes, 2:920f
 sediments
 eolian records, 1:637–642
 interpretation of data, 1:638–641
 nature of data, 1:638
 North Pacific Ocean, 3:33
 deep water records
 Indian Ocean, 2:935f
 North Pacific Ocean, 2:935f, 3:34–37
 South Atlantic, 3:28–29
 upwelling, 1:737
 see also North Atlantic Deep Water; thermohaline circulation
- deep weathering profiles, 3:399–400
- deer
 Mediterranean, 4:737–739, 4:738f, 4:739f
 South America, 4:689–690
 teeth/bones, 1:307f, 1:309f
- defensive searches, forensic geology, 4:524
- deformations
 earthquake cycle, 4:504–507, 4:509f, 4:510f, 4:511f, 4:514f, 4:516f
 East Antarctic Ice Sheet, 2:456–457
 Greenland ice sheet, 2:441–443
 sedimentary
 glacial micromorphology, 2:54–56, 2:58f, 2:60
 periglacial, 3:430–435
 strandline records, 4:518
- deformation till
 fabric analysis, 2:79
 glacial environments, 2:19–20
 subglacial, 2:21–22, 2:24f, 2:62–64, 2:63f, 2:64f, 2:65–66, 2:65f, 2:69–71
- deformation–lodgment continuum, till, 2:65–66
- De Geer, Gerard, 1:776–777, 4:582
- De Geer moraines, 1:776–777, 2:46, 2:49–50
- deglaciation
 Antarctic Peninsula, Holocene, 2:219–220
 Arabian Sea, 3:48f
 atmospheric carbon-14, 4:331–332
 definition, 2:216ge
 East Antarctica, Holocene, 2:220–221
- Eurasian glaciations, 2:232
 Greenland, 3:760–761, 4:483–484
 ice sheet growth/decay, 1:878–879
 Indian Ocean, 3:46, 3:47, 3:49–52
 North Atlantic Ocean, 1:567, 3:15–17
 Pacific Ocean, Holocene, 1:574–575
 paleoceanography overview, 2:751–753, 2:753f
 paraglacial geomorphology, 3:553
 postglacial pollen records, 4:137, 4:138f
 Ross Sea, Holocene, 2:221
 THC, 1:745
 Victoria Land, Antarctica, 2:221
- degradation
 fluvial terraces, 1:684, 1:685f, 1:691
 palsas/lithalsas, 3:477–478
 permafrost, 3:427, 3:464–465, 3:576–577
 pigments, aquatic environments, 3:330, 3:331f
 pingos, 3:475–476
 see also incision
- degree programs, forensic botany, 4:546
- delta age
 firm air transport, 2:367–368, 2:368f
 gas record models, 2:410–411, 2:413–414, 2:415
- $\delta^{13}\text{C}$
 definition, 3:768ge
 plant macrofossils, 3:768–769
- $\delta^{14}\text{C}$ temporal variations, 4:337, 4:339–340, 4:339f, 4:340f, 4:341f, 4:342f, 4:343, 4:343f
- $\delta^{18}\text{O}$
 Atlantic Ocean influence, 2:390
 biogenic carbonates, 1:292
 corals, 2:796
 diatom silica records, 1:481–488, 1:548
 foraminifer-based, 2:919–921, 3:48f
 fossil carbonate measurements, 2:919
 global seawater distribution, 2:918
 Greenland ice sheet margin, 2:422f, 2:423–424, 2:423f, 2:427f
 ice cores
 Andes, 2:389–390
 Antarctica, 2:395
 interannual variations, 2:390
 marine record, 2:216ge
 North America, 2:180–186
 sea-level change, 4:369–370
 paleoclimatic studies, 1:298–300
 paleowaters, 1:298
 sclerosponges, 2:797
 speleothems, 1:292, 1:298
- $\delta^{18}\text{O}_{\text{civ}}$, 1:298–300
- $\delta^{18}\text{O}_{\text{diatom}}$ (diatom silica) records, 1:481–488
 analytical considerations, 1:481–482
 Lake Baikal, 1:548
 methods of analysis, 1:481–482
- $\delta^{18}\text{O}_{\text{foram}}$ studies, 1:486–487
- $\delta^{18}\text{O}_{\text{water}}$, 1:485–486
- δ notation
 concepts, 1:292
 definition, 2:795ge
 ice core studies, 2:347
- deltas, 1:693–703
 climate influence, 1:699

- deltas (*Continued*)
- controls/processes, 1:693–695
 - definition, 1:693
 - glacial–interglacial scale responses, 2:123f
 - glaciolacustrine, 2:48f, 2:49f, 2:50
 - high-energy coasts, 4:392–393, 4:393f
 - ice-proximal settings, 2:33, 2:34f
 - importance/significance, 1:693
 - lava-fed, 1:780ge, 1:781–782, 1:782–783, 1:791
 - modeling studies, 1:700–702
 - morphology/morphodynamics, 1:695–699
 - research history, 1:693
 - sediments, 1:695–699
 - site cases, 1:699–702
- demography, ancient DNA, 4:720–721, 4:721–722
- dendroarchaeology, 3:630–636
- behavioral information, 3:634–636
 - chronological information, 3:632–633
 - date attributes, 3:630–631
 - environmental information, 3:633–634
 - sampling, 3:631, 3:632–633, 3:633–634, 3:634–636
- dendrochronology, 1:453–458
- archeological applications, 3:630–636
 - atmospheric carbon-14 variations, 4:329–330, 4:330f, 4:331, 4:332f
 - biological foundations, 1:453–454
 - glacial landforms, 2:104–111
 - measurement techniques, 1:455
 - methodologies, 1:10, 1:455–457
 - neoglaciation chronology establishment, 2:270
 - origins/history of, 1:453
 - principles of, 1:454–455, 2:93f, 4:586–587
 - tephra dating, 4:288–289
 - varved sediment application, 4:586–587
- dendroclimatology, 1:459–470
- analysis methods, 3:634
 - applications, 1:461–463
 - basic principles/methods, 1:459
 - changing signals, 1:467
 - chronology development, 1:459–460
 - climate/climate-related reconstructions, 1:460, 1:466
 - historical development, 1:460–461
 - single-site reconstructions, 1:461–463
 - tree-ring networks, 1:463–466
 - tropics, 1:466–467
- dendrogeomorphology, 2:91–103
- dendroglaciology, 2:104–111
- applications, 2:104–109
 - methods, 2:105t
 - neoglaciation in Europe, 2:259–260
- dendrophilic species, 1:255ge
- Denisovians, 1:141, 1:142
- Denisov's principle, 2:594
- denitrification, OMZ variability, 3:52–54
- Denmark
- ice sheet/glaciations, 2:228–229
 - relative sea level changes, 4:491
 - stand-scale palynology, 3:847f, 3:850–852, 3:852f
- denticulates, definition, 1:83ge
- dentition *see* teeth
- deoxidation *see* reduction
- DEP *see* dielectric profiling
- depocenters, glaciolacustrine, 2:43, 2:46–50, 2:48f
- deposition processes
- cosmogenic radionuclides, 2:354–355, 2:358
 - deep-sea sediments, 1:637–642
 - environmental archeology, 3:699
 - indicators
 - erosion marks, 4:381–384
 - sea-level change, 4:386f, 4:394f
 - pollen, 3:871–872
 - slope deposits, 3:481–489
- depositional architecture, ice-proximal settings, 2:30–33
- depositional environments, pollen, 3:725–727, 4:1–3
- depositional landforms, glacier flow, 1:895–896
- depositional remanent magnetization (DRM), 1:710–712, 1:713
- depositional structures, glacial sediment, 2:54–56
- depositional systems
- aerosol deposition, 2:326–327
 - deltas, 1:695–699
 - glacial
 - facies models/associations, 2:20–21
 - marine, 2:24–28
 - terrestrial, 2:21–24
 - pigment studies, 3:330–336
 - soil morphology and, 3:408
 - subglacial, 2:21–24
 - supraglacial, 2:24, 2:26f
- depth dependence, ice sheet annual layers, 2:443, 2:444f
- depth-profile dating, cosmogenic nuclides, 1:432, 1:436f
- Dermeistidae (skin/larder beetles), 4:553–554
- desert pavements, 3:409f
- deserts
- Central Asian loess records, 2:588f
 - dunes, China, 1:609–610, 1:610–611
 - eolian sand, 3:357–359
 - insect fossil studies, 1:288–289
 - loess
 - Central Asia, 2:588f
 - silt particle formation, 2:574–575, 2:576f
 - low-latitude sand seas, 1:623–636
 - pollen records, 4:142–143, 4:152
 - rodent middens, 3:674–683
 - soil morphology, 3:409f
 - see also* Sahara Desert
- Desert Southwest, North America, 4:142–143, 4:152
- desiccation, plant macrofossils, 3:702
- detachments, active-layer, permafrost, 3:579, 3:580f
- detachment slides, permafrost, 3:484–485
- deterministic lake models, 2:558
- detrended canonical correspondence analysis (DCCA), 1:492, 1:493, 1:493f, 1:498f
- detrended correspondence analysis (DCA), 3:848–849, 3:849–850, 3:850f
- detrended time series, climate change, 3:97–98
- detrital remains
- trees, 2:104, 2:105f
 - see also* debris...
- deuterium, ice cores, 2:395
- Antarctica, 2:396, 2:397, 2:399, 2:399f, 2:400
 - excess, 2:350–351, 2:396, 2:397, 2:399, 2:399f, 2:400
- deuterium/hydrogen (D/H) ratio, 1:291–292
- Deva-Deva swamp, Tanzania, 4:12–13, 4:13f
- Developed Oldowan, 1:59, 1:60f, 1:62–63
- Devensian glaciation, 2:230
- Devils Hole, southern Nevada, 1:322, 1:327–328, 1:329f
- D/H ratio *see* deuterium/hydrogen ratio
- diabetes mellitus (DM), 1:315–321
- diachronic units, climatostratigraphy, 4:225
- diagenesis
- alkenone unsaturation index, 2:761–762
 - amino-acid dating, 1:38
 - lake sediment, 3:310
 - MS postdepositional records, 2:895–897
 - teeth/bones, 1:307
 - U-series dating of corals, 4:569f
 - see also* effective diagenetic temperature
- diagnosis of drowning, forensics, 1:522–523
- diamagnetism, 3:375ge, 2:884, 2:887t, 3:375–376
- diameter-correction fission-track (DCFT) dating, 1:650, 1:653f, 1:653t, 1:656
- diamicts/diamictos, 2:18–20, 2:21f, 2:52, 2:54f
- diapirism, sorted patterned ground, 3:458, 3:459–460
- Diaptomus castor* macrofossils, 3:652, 3:653f
- diaries, historical, 3:238, 3:239, 3:240f
- diasporas, human, 1:53–54
- diatom accumulation rate (DAR), marine proxies, 2:817, 2:820, 2:821f
- diatomites, 1:501–506, 1:524
- age, 1:502–503
 - ancient uses, 1:504
 - commercial use, 1:503–505
 - diatom ooze, 1:501
 - distribution, 1:502–503
 - environmental science, 1:505
 - extraction, 1:503–504
 - formation, 1:501–502, 1:502f
 - preservation, 1:502
 - Toome Bridge, UK, 1:503–504, 1:505f
- diatom lipids, 1:590–591
- diatom mat laminations, marine sediments, 1:555, 1:556–557, 1:560
- diatom ooze
- accumulation sites, 1:554–555
 - continental shelf basins, 1:555, 1:556f, 1:557–559
 - enclosed/semienclosed seas, 1:556–557
 - high-latitude fjords, 1:559
- diatom silica ($\delta^{18}\text{O}$) records, 1:481–488
- analysis
 - considerations, 1:481–482

- methods, 1:481–482
 procedures, 2:734–735
 cell walls, 1:471
 $\delta^{18}\text{O}_{\text{foram}}$ combined studies, 1:486–487
 dissolution, 2:734, 2:737, 2:738f
 growth, 2:735–737
 Lake Baikal, 1:548
 marine sediment applications, 2:738–739
 seawater paleothermometry, 2:876–877
 silicon isotope analysis, 2:734–743
 diatoms, 1:471–480, 1:481–488, 1:489–500, 1:522–525
 aerophilic, 1:475
 Antarctic marine waters, 1:527–539
 archeological analysis, 1:516–521
 artifacts, 1:516–518
 boats, 1:520, 1:520f
 buildings, 1:516–518
 crannogs, 1:518–519
 forensic techniques, 1:520
 Inuit whalers, 1:516, 1:517f
 moats/medieval fish ponds, 1:518, 1:518f
 Ptolemaic Temple, Dimai, Egypt, 1:517–518, 1:517f
 salt extraction sites, 1:518
 site environments, 1:516
 waterfronts, 1:519–520, 1:519f
 Arctic, 1:562–570, 1:588–596
 assemblage shifts, 1:491f
 Bennett's broken stick model, 1:491
 benthic, 1:474–475
 biogeography, 1:475, 3:313–314, 3:314f, 3:319f
 acid rain, 3:320
 biodiversity and, 3:318f, 3:321
 evolutions, 3:323
 extinctions, 3:322
 invasive species, 3:322
 paleoclimate, 3:314, 3:315–316, 3:316–317
 terrestrial-aquatic links, 3:319
 water quality, 3:320f
 biology, 1:471
 calibration model, 1:494, 1:495f
 centennial- to orbital-scale, 1:559–560
 classes, 1:471, 1:472f
 cleaning, 1:476–477
 collection of, 1:475
 concentrations and accumulation rates, 1:489–490
 CONISS method, 1:491
 detrended canonical correspondence analysis (DCCA), 1:492, 1:493, 1:493f, 1:498f
 diatomites, 1:501–506, 1:524
 diversity, 1:494f
 ecological and habitat classification, 1:490–491
 endemism, 1:475
 enumeration, 1:477
 freshwater laminated sequences, 1:540–545
 habitats, 1:472–475
 freshwater, 1:473–475
 marine, 1:472–473
 Holocene RSL reconstruction, 4:419–420, 4:424–425, 4:425f
 identification, 1:477
 immersion period, 1:524–525
 lake chemistry, 3:293, 3:294, 3:295, 3:296, 3:297f
 lake levels
 Latin America, 2:533
 reconstructions, 2:492–493, 2:496–497, 2:497f
 lake salinity/climate reconstructions, 1:507–515
 large lakes, 1:546–553
 lifecycles/life form, 1:471–472
 marine laminated sediments, 1:554–561
 marine proxies, 2:817, 2:820, 2:821f
 North Atlantic/Arctic, 1:562–570
 North Pacific, 3:68–69
 objective assessment, 1:491–494
 oxygen systematics, 1:482–487
 Pacific Ocean, 1:571–587, 3:68–69
 paleodroughts, 3:195–196
 paleoenvironmental studies, 1:477, 1:546
 paleoindicators, 1:489
 paleolimnology, 3:266–267, 3:268, 3:340, 3:340f, 3:342, 3:343f, 3:345–346
 parameter value estimation, 1:494–496
 planktonic, 1:473–474
 preservation, 1:477
 productivity events, marine varves, 1:556–559
 reconstruction evaluation, 1:496–499
 record from Lake Victoria's Damba Channel, 1:491, 1:492f
 response to site-specific air temperature, 1:497f
 river diatoms, 1:475
 sample preparation, 1:476–477, 1:476f
 sea ice, Arctic, 1:588–596
 silicon isotopes, 2:734–743
 species-level, 1:544
 structure, 1:471
 surface sediments, 1:566–567
 taphonomy, 1:477
 total planktonic diatom flux record, 1:489, 1:490f
 transfer functions, 1:477–479
 valves, 1:471
 variety of, 1:473f
 vegetative reproduction, 1:474f
 water quality, 1:477–479, 3:320f
see also marine diatoms
 Diaz, H, 3:103–104
 DIC *see* dissolved inorganic carbon
 dielectric profiling (DEP), ice cores, 2:319, 2:321–322, 2:322–323, 2:323–324, 2:323f
 diet
 hominids, 4:748–751
 stone tools and, 1:59, 1:62, 1:64, 1:65
 dietary plants, 4:543
 differential frost heave, 3:452, 3:454–457, 3:461
 diffusion
 firn air, 2:364–365, 2:366, 2:369–370
 Greenland stable isotopes, 2:403–405
 oxygen isotopes, seawater, 2:918
 soil diffusion model, 1:326–327
 thermal, 2:431–434
 Digerfeldt approach, lake-level reconstructions, 2:490, 2:491f, 2:492, 2:493
 diggability, forensic searches, 4:526t, 4:528
 dimethylsulfide (DMS), 2:330–331
 dinocysts, 2:800, 2:803
 biogeographic distribution, 2:805–808
 biostratigraphy, 2:805
 marine waters, 2:805–811, 2:811–812
 taxonomy/morphology, 2:803–805
 dinoflagellates, 2:800–815
 ecology of, 2:800–801, 2:808–811
 life cycle, 2:801–803, 2:803f
 taxonomy/morphology, 2:803–805
 Diogo interdunal depression, Senegal, 4:90, 4:90f
 dipole mode, Indian Ocean, 3:52
 diprotodontoid vertebrate records, 4:633–634
 Diptera (flies)
 forensic science use, 4:548, 4:549–552
 Late Pleistocene records, 1:373–385
 overview, 1:355–360
see also chironomids
 direct current, ice cores, 2:320–321
 direct measurements, fossil waters, 2:921
 direct ordination, definition, 1:361ge
 direct percussion process, 1:83ge
 discharge
 allogenic forcing response, 1:676
 ice into ocean, Greenland, 2:441
 lake-level modeling, 2:560
 periglacial fluvial forms, 3:492
 sediment, river deltas, 1:693–694, 1:694–695, 1:696, 1:702f
 disease
 loess records, China, 2:604
 megafaunal extinctions causation theory, 4:711
 disequilibrium effects, diatom silica records, 1:481
see also vital effects
 disequilibrium offset, corals, 2:878–880
 disintegration ridges, 1:809
see also ridges
 dispersal distance, macrofossils, 3:687–688
 dispersal events *see* dispersal of species
 dispersal models, pollen, 3:871–872
 dispersal of species
 archeology overview, 1:53–54
 human expansion, Americas, 1:119, 1:121–122, 1:124–125
 mammals, 4:208–209, 4:600–601, 4:639
 Neanderthal demise, 1:146–153
 Villafranchian stage, 4:639
see also migration of species
 dispersive mixing, firn air, 2:366
 dissolution
 deep-sea carbonates, 2:859–870
 diatoms
 large lakes, 1:548–549
 silicic acid, 2:734

- dissolution (*Continued*)
 silicon isotope fractionation, 2:737, 2:738f
 foraminifera Mg/Ca, 2:874–876
 dissolved inorganic carbon (DIC)
 biogenic carbonates, 1:292
 deep-sea carbonate dissolution, 2:859–860, 2:861, 2:862f
 marine carbon cycle proxies, 2:854–855, 2:855f
 nutrient proxies, 2:899, 2:900f
 speleothems, 1:294–295, 1:300–301
 dissolved organic carbon (DOC), 3:292, 3:295–296, 3:296f, 3:334–335, 3:335–336
 disturbance biogeography, 3:317–319
 disturbance biology, pollen data, 2:703–704
 divergence of vertebrate species, 4:724
 DLS *see* downlap surface
 Dmanisi site, 1:67
 hominin remains, 1:67–68, 1:69t
 human remains, 1:138–139
 lithic technology, 1:70f
 DMS *see* dimethylsulfide
 DNA
 American human colonization, 1:119, 1:126
 in forensics, 4:542–543, 4:559–561
 insect fossil samples, 1:173–174
 plants, 2:705–715
 vertebrate studies, 4:668–669, 4:670, 4:719–721
see also ancient DNA
 D–O *see* Dansgaard–Oeschger...
 DOC *see* dissolved organic carbon
 Dodson, John Bruce, 4:535–537
 dogs
 domestication of, 4:754–755
 forensic geology use, 4:531, 4:533f
 Dolgoe Lake, east Siberia, 4:168
 Dome C, Antarctica
 research history, 2:437
 stable isotopes, 2:395, 2:398, 2:398f, 2:399f
 Dome F, Antarctica, 2:398, 2:398f, 2:400
 domes, felsic, 1:796, 1:799f
 domesticated animals, 4:754–755
 domesticated plants in archeology, 3:705–706, 3:713f
 Dömnitz/Wacken substage, Middle-Pleistocene, 2:173–174
 Donian glaciation, Russia, 2:172
 Douglass, Andrew Ellicott, 1:453, 1:460–461, 3:630, 3:631f
 downlap surface (DLS), sequence stratigraphy, 4:261, 4:263
 downstream controls, climate change, 1:676
 Drake Passage Effect, THC, 1:740–742
Dreissena biogenic carbonates, 1:343, 1:345–346, 1:349f
Drepanocladus macrofossils, 3:651f, 3:652
 drifting continent configurations, 2:132–134
 drift-mantled slopes, 3:559
 drillholes, lithostratigraphy, 4:228–229
 drilling projects
 Andes ice cores, 2:387, 2:388f
 Antarctica, 1:531f, 1:535–536
 DRM *see* depositional remanent magnetization
Drosera macrofossils, 3:638, 3:640f
 drought
 Africa
 ice core records, 2:377
 lake level studies, 2:503, 2:504
 Andean glaciers, 2:392
 indices, PDSI, 3:89f
 lake level studies, 2:503, 2:504, 2:535
 last millennium climate, 3:230–231, 3:232f
 tree-ring networks, 1:463–464, 1:465–466, 1:465f, 1:467–468
 drowning diagnosis, 1:522–523
 drumlins
 glacier flow directions, 1:895–896
 icesheet/glacier extent, 1:887, 1:888f, 1:889f
 meltwater erosion hypothesis, 1:835–836
 recession of glaciers, 1:807–808
see also rock drumlins
 dry-bed deglaciation fans, 1:896
 dry deposition, aerosols, 2:326–327
 dry extraction methods, ice cores, 2:468–469
 dry periods
 Latin America, 2:533
see also drought
 dry preservation
 ancient DNA, 4:720
 mammal remains, 4:675–676
 dual-aliquot TL dating technique, 2:650–651
 Dubawnt-type erratic trains, 2:81–83, 2:83f
 dump moraines, 1:769–770, 1:772f, 1:773f
 dune fields
 age determination, 1:602–604
 future studies, 1:604
 high latitudes, 1:597–605
 low latitudes, 1:623–636
 midlatitude, 1:606–622
 origins, 1:601–602
 overview, 3:357–359, 3:358f, 3:359f
 paleoenvironmental significance, 1:604
 primary, 1:598–599
 types, 1:599–601
 dune-forming winds, 1:600f, 3:360f
 dunes, gravel bedforms, 2:9
 dung beetles
 forensic science use, 4:554
 Japanese sites, 1:213–214, 1:218–219
 Northern Asia, 1:256f
 dung flies, 4:551
 dust
 abrupt climate events, Africa, 2:376f, 2:377
 atmosphere/land/sea links, 3:412, 3:415–416
 biogeochemical role, 3:412–420
 biosphere feedbacks, 1:726
 climate cycles, 3:412–420
 cycle changes, 3:417
 deep-sea sediments, 1:637–642
 deposition rates, 3:414f
 ice core studies, 2:281–283, 2:282f, 2:329
 Africa, 2:376f, 2:377
¹⁰Be, 2:355
 China, 2:379, 2:380–382, 2:383, 2:384
 microparticles, 2:342–343, 2:343–344, 2:344f
 land-dust emissions, 3:412–415
 links with atmosphere/land/sea, 3:412, 3:413f
 loess records, China, 2:599, 2:600–601, 2:601f
 magnetic susceptibility, 2:889, 2:892f
 polar ice content, 1:31–32
 radiative impact, 3:415–416, 3:417, 3:418
 sea-dust as fertilizer, 3:416–417
see also atmospheric dust
 dust-borne ¹⁰Be, 2:355
 dust storms
 Central Asia, 2:587, 2:592
 North America, 2:624f
 Dutch Elm disease, 4:176
 dwarfing/gigantism
 anagenetic/radiative evolution, 4:737–739
 Balearic bovids, 4:739–741
 Carnivora, 4:741–742
 case histories, 4:733–747
 crocodiles, 4:745–746
 hippopotamuses, 4:736–737
 hominins, 4:741–743
Homo floresiensis, 1:143
 island surface area, 4:734, 4:735f
 Mediterranean deer, 4:737–739, 4:738f, 4:739f
 owls, 4:746
 primates, 4:741–742
 proboscideans, 4:734–736, 4:742–743
 Sardinian bovids, 4:739–741
 shortened/long metapodials, 4:739–741
 small mammals, 4:743–745
 turtles, 4:745–746
 uncommon island settlers, 4:741–743
 vertebrate studies, 4:733–747
 wingless birds, 4:746
 dwarf shrubs, 3:599f, 3:762
 dwellings, lake-shore, Europe, 2:549–551
 Dye-3 ice core, Greenland, 2:436
 stable isotopes, 2:403, 2:404f, 2:405, 2:406f, 2:407
 dynamism hypothesis
 Amazonian rainforest, 4:59–61
 lake level studies, 2:499–502, 2:502–504
 treelines
 Holocene, 3:693–695
 late glacial change, 3:692–693
 vegetation, 4:115–123
see also ice dynamics
 dystrophic lakes, 3:657, 3:660f
 Dytiscidae beetles, 1:211, 1:214f
- ## E
- EAIS *see* East Antarctic Ice Sheet
 early American human expansion, 1:119–134
 Early Galerian mammal assemblages, 4:610–611
 Early Glacial stages, North American ice sheets, 2:245–247

- Early Holocene
 African ice core records, 2:376
 Late Pleistocene boundary extinctions, 4:691
 southeastern North American pollen records, 4:137
 early *Homo*, Africa, 1:65
see also Homo...
 Early–Late Glacial human dispersal, 1:124–125
 Early Mesolithic human adaptations, 1:156–157
 Early Middle Pleistocene
 cave/fissure filling faunas, 4:655–662
 vertebrate studies, 4:639, 4:641–643, 4:651–654, 4:655–662
 Early Pleistocene
 Arctic North America, 3:753
 beetle records
 New Zealand, 1:244–247
 Stony Creek Basin, Australia, 1:193–194
 Western Europe, 1:187
 glaciation history, 2:167–171
 hominins, Asia, 1:67–71
 mammal assemblages, Northern Eurasia, 4:608–610
 paleoceanographic records, 3:1–8
 plant macrofossil records, 3:739
 vertebrate studies, 4:599–604
 case histories, 4:604
 Europe, 4:639–641
 Northern Eurasia, 4:605–614
 Early Pliocene climate, 3:185–186
 Early Quaternary, 2:167–171
 Africa, 2:168
 Antarctica, 2:167
 Arctic paleoclimate history, 3:114–115
 Australasia, 2:167–168
 Europe, 2:168–170
 glaciation history, 2:167–171
 Iceland, 2:169
 insect fossil studies, 1:173–183
 maps of glaciations, 2:169f
 North America, 2:168
 Scandinavian ice sheet, 2:168–169
 South America, 2:167
 Southern Europe, 2:169–170
 transition from Late Neogene, 2:151–166
 Early Stone Age (ESA), 1:59–66
 Early Weichselian beetle assemblages, 1:201f, 1:271–272
 Early Wisconsin
 glaciation, Beringia, 2:194–197, 2:197f, 2:198f
 ice sheets, 2:247
 Earth
 geomagnetic field, 4:338
 models
 last interglacial, 3:155, 3:158–159, 3:160–162
 magnetism, 3:375, 3:376f
 surface, isostasy, 4:452, 4:453f, 4:454f
see also atmosphere
 earth hummocks, 3:460–462
 Earth system models of intermediate complexity (EMICs), 3:155, 3:158–159, 3:160–162
 earthquakes, 1:22–23, 1:23t
 deformation cycle, 4:504–507, 4:509f, 4:510f, 4:511f, 4:514f, 4:516f
 major historical, 1:23t
 sea-level change hazards, 4:503–519
 subduction zone thrust, 4:504–505, 4:510f, 4:511f, 4:514f, 4:516f
 East Africa
 fauna, Middle Pleistocene, 4:616–618
 ice core records, 2:373, 2:374f, 2:377
 lake dynamics, 2:501–502, 2:503f
 vertebrate fossil records, 4:670
see also individual countries
 East Antarctica
 borehole temperature records, 2:301
 deglaciation, Holocene, 2:220–221
 ice core microparticle fingerprinting, 2:343–344
 RSL in high latitudes, 4:475f, 4:477, 4:478f, 4:479
 East Antarctic Ice Sheet (EAIS)
 characteristics, 2:216
 definition, 2:216ge
 dynamics, 2:456–462
 Pleistocene, 2:218
 East Asia
 lake level studies, 2:512
 modern human expansion, 1:100
 monsoon system, 2:595, 2:597f, 2:602f
see also individual countries
 East Atlantic climate, 3:128
 eastern Argentinean vertebrates, 4:683
 Eastern Beringia studies, 1:228–233, 1:229t, 1:231f, 1:232f
 eastern equatorial Pacific (EEP), 3:62–64
 Eastern Europe
 loess, 4:254
 vertebrate studies, 4:605–614
see also individual countries
 Eastern Highlands, Australia, 2:524–525
 Eastern North America
 biomes, 3:867
 insect fossil research, 1:282–286, 1:283t, 1:285f
 Late Pleistocene beetle studies, 1:221–225, 1:222f
see also southeastern North America
 Eastern Siberia
 areal ice extent, 1:891
 palynological records, 4:27, 4:29f, 4:30f, 4:31f, 4:36–37
 East Greenland current, 2:157–159, 2:162–164
 East Guangdong, China, 4:444
 EBMs *see* energy balance models
 Ebro frontier, Neanderthal demise, 1:151
 ecdysis interval, tree-ring dating, 2:108
 ECM *see* electrical conductivity measurement
 ecofacts, 3:699
see also plant macrofossil...
 ecological amplitude principle, 1:455, 1:455f
 ecological forensics, 1:524, 4:543, 4:545–546
 ecological grouping, insects, 1:261–263, 1:267t
 ecological processes
 benthic foraminifera, 2:765–767
 chironomids, 1:358
 dinoflagellates, 2:800–801, 2:808–811
 pollen databases, 3:831
 economy
 early Americans, 1:127–129
 environmental archeology, 3:699, 3:705, 3:708–715
 relations, hominid–animals, 4:748
 Sahul landmass, 1:115–117
 ecosystems
 Australia, 1:111, 1:113
 biosphere history, 1:23
 dynamics, reconstructions, 3:250, 3:251f
 lake chemistry, 3:292
 ocean history, 1:19–20
 ecotone, definition, 1:361ge
 Ecuador, vertebrate fossils, 4:681
 eddy bars, 2:12
 education programs, forensic botany, 4:546
 Edwards, ME, 3:103
 Eem Formation, Netherlands, 4:229–234, 4:231f
 Eemian Interglacial
 beetle assemblages, 1:200, 1:201f
 CO₂ reconstruction, 3:618–619
 Eurasian glaciations, 2:224, 2:225f
 European pollen records, 4:1–2, 4:3, 4:3t, 4:6–7
 Greenland
 sea-level changes, 4:483
 stable isotopes, 2:406f, 2:408
see also last interglacial
 effective diagenetic temperature, amino-acid dating, 1:40–41, 1:41f, 1:46
 effective diffusivity, firn air, 2:366
 effusive processes, definition, 1:780ge
 EFTM site *see* Elandsfontein Main site
 eggshells, amino-acid dating, 1:42, 1:42f, 1:43, 1:44, 1:44f
 eigenvalue method, till fabric analysis, 2:76–77, 2:77f, 2:78, 2:78f
 ELA *see* equilibrium-line altitude
 Elandsfontein Main (EFTM) site, Africa, 4:664–665, 4:666t
 Elateridae beetles, 1:214
 electrical conductivity measurement (ECM), 2:319, 2:320–321, 2:322f, 2:323–324
 electrical logging, ice cores, 2:319
 electrical measurements, ice cores, 2:319–325
 electrical properties of ice, 2:319–320
 electron spin resonance dating (ESR), 2:667–677
 applications, 2:672–676
 basic principles, 2:667
 bones, 2:676
 Border Cave, 2:675f, 2:676
 corals, 2:668, 2:672–673, 2:674f
 dose rate, 2:669–672
 dose values, 2:667–668
 errors, 2:672
 measurement, 2:668–669
 methodology, 1:13, 1:14f
 mollusk shells, 2:668, 2:670–672, 2:673–676, 2:674f
 quartz, 2:676
 radioactive elements, 2:672

- electron spin resonance dating (ESR)
(*Continued*)
saturation limits, 2:668
spectrometer diagram, 2:670f
thermal stability, 2:668
tooth enamel, 2:668, 2:668f, 2:670–672, 2:676
- elephants, 4:600–601
see also proboscideans
- elevational gradients, pollen studies, 3:731
- elevation attributes, sea-level index points, 4:372
- eLGM *see* extended last glacial maximum
- El'gygytyn Lake, Beringia, 2:195, 2:195f, 2:196f
- Elm tree decline, 4:175–176
- El Niño
climate effects
paleotempestology, 3:215–217, 3:216f
South America, 2:387, 2:392
SST patterns, Pliocene, 3:186–187, 3:187–188, 3:189f
teleconnection, Indian Ocean, 3:51
- El Niño–Southern Oscillation (ENSO), 2:393, 3:150
abrupt climate events, 2:392
Asian monsoons, 2:379
Australia, effects, 1:108–109
dendroclimatology, 1:465–466, 1:467–468
global warming relevance, 3:248–249
Indian Ocean, 3:51
last millennium climate, 3:231–233, 3:234, 3:235
Latin American lake level studies, 2:531, 2:535
North Pacific, postglacial, 3:62, 3:63–64, 3:65, 3:67, 3:68, 3:69–70
South Pacific, postglacial, 3:73–74, 3:77, 3:79f, 3:80, 3:82–84, 3:83f
stable isotopes, 2:390
volcanic/solar forcing, 1:750
see also paleo-ENSO
- Elsterian stage, Eurasia, 2:172–173, 4:611–612
- emerged sedimentary indicators, sea-level change, 4:393–395
- EMICs *see* Earth system models of intermediate complexity
- Emiliana huxleyi* coccoliths, 2:756–757, 2:757f, 2:758, 2:760–761
- Emiliani, Cesare, 2:908
- Emiran technocomplex, 1:149
- Empetrum nigrum* macrofossils, 3:638, 3:639f
- encephalization, 1:55, 1:56
- enclosed seas, diatoms, 1:556–557
- endemic islands, Southern Asia, 4:662–663
- endemism, beetles, 1:235
- energetic nuclear radiation, 1:419, 1:424f
- energy balance models (EBMs)
lake levels, 2:559
Quaternary environments, 3:147
volcanic/solar forcing, 1:749–750, 1:750–751
- energy spectrum, cosmic rays, 1:419, 1:426f
- England
beetle assemblages, Middle Pleistocene, 1:184–189
Lake District landsystems, 1:819f
see also Britain/British Isles; United Kingdom
- ENSO *see* El Niño–Southern Oscillation
- entomology
Australia, 1:191–199
forensic, 4:548–555
Japanese, 1:219–220
see also insect fossil studies
- environmental archeology
beetle records, 1:169
plant macrofossil studies, 3:699–724
postglacial North America, 1:285–286
results examples, 3:705–708
- environmental change
African chironomid records, 1:362–368
China, Tibetan Plateau, 2:379, 2:384, 2:385
chironomid records, 1:362–368
Colombian savannas, 4:60f
diatom records, 1:541–544, 2:741f
human adaptation, postglacial, 1:154–159
human evolution, 1:54
megafaunal extinction hypothesis, 4:709–711
ocean history, 1:19–20
paleolimnological approaches, 3:339, 3:342–344
plant macrofossil records, 3:785–793
rapid change
biotic response, 3:790
fluvial environment responses, 1:676–683
rock glaciers, 3:539–540
see also climate change
- environmental characteristics
Africa, 4:664
macrofossil taphonomy, 3:688
Western North America, 4:72
- environmental complexity, glacial landforms, 2:18–20
- environmental development, Holocene
Antarctica, 2:221
- environmental effects
diatomites, 1:505
humans in America, 1:126–127
oxygen isotopes, 1:338
sorted patterned ground, 3:458
- environmental information,
dendroarcheological samples, 3:633–634
- environmental interpretation, beetle records, 1:197–198
- environmental magnetism, 1:923–924, 2:884
- environmental modeling, climate change, 1:32–33
- environmental pH, amino-acid dating, 1:41
- environmental reconstructions, plant macrofossils, 3:785–787
- environmental variability, calcareous skeletons, 2:795–799
- environment-to-circulation approaches, 3:102–107, 3:109–110
- Enzel, Y, 3:105–106
- enzymes, ancient DNA, 4:720
- Eocene
definition, 2:216ge
Greenland sea-level change, 4:481–482
- eolian hypothesis, loess origins, 2:585–587
- eolian inputs, climate cycles, 3:416–417
- eolian records
deep-sea sediments, 1:637–642
paleodroughts, 3:196
- eolian sand, 3:357–366
Alaskan dune fields, 1:599
Argentina, 1:613–614, 1:614f
Central Asia, 1:613f
distribution, 1:607f, 2:575f, 3:358f, 3:361f
low-latitude sand seas, 1:623
midlatitude dune fields, 1:606, 1:607f, 1:613–614, 1:613f, 1:614f, 1:615f, 1:618–620, 1:619f, 1:620f
North America, 1:615f, 1:619f, 1:620, 1:620f
Taklimakan Desert, 1:618–620
see also windblown sediments
- EPD *see* European Pollen Database
- epeirogenic tectonic processes, 4:507–511
- ephippia, cladocera, 3:272, 3:276f, 3:277–278, 3:278f, 3:271ge
- EPICA *see* European projects for ice coring in Antarctica
- epicontinental environments, dinocysts, 2:812
- epicontinental seas, 1:763–764
- epifaunal benthic foraminifera, 2:767–768, 2:770
- epigenetic ice wedges, 3:438, 3:438f, 3:439f, 3:443f, 3:468
- Epipaleolithic archeology, 1:95–96, 1:106f
- EPS *see* expressed population signal
- Epstein, S, 2:907–908
- equatorial forest pollen records, 4:98–100
- equatorial oceans, 4:495–502
- equatorial trough (ET), South America, 2:389
- equids, 4:665–667, 4:668
see also horses
- equilibration in modeling, 3:150, 3:152f
- equilibrium line
definition, 1:909ge
Greenland ice sheet, 2:441
- equilibrium-line altitude (ELA)
calculation approaches, 1:913–916
definition, 1:909ge
glacial landforms, 1:847–849, 1:909–917
Highland Asia glaciations, 2:238–239, 2:241, 2:242f
- ice sheet growth/decay, 1:877–878, 1:881–882
- precipitation–temperature relationship, 1:911–912, 1:912–913, 1:912f
- equilibrium time, definition, 1:678–679
- Equisetum* macrofossils, 3:646, 3:647f
- Equus* sp *see* horses
- Erdtman, Gunnar, 2:699–700
- ergs (sand seas), 3:357–359
distribution of, 1:624f
dune distribution in, 1:627t
low latitudes, 1:623–636
- Erhia Lake, Yunnan, 2:486f

- Erica tetralix* macrofossils, 3:638, 3:639f
Eriophorum angustifolium macrofossils, 3:638–642, 3:641f
Eriophorum vaginatum macrofossils, 3:637, 3:638, 3:640f, 3:653f
 Ermakovski Stages, Northern Asia, 4:36–37
 erosion
 blockfields, 3:530–532
 climate influence, 1:442–443
 continental interiors, 1:670–671
 glacial major scale landforms, 1:847–864
 glacifluvial landforms, 1:825–838
 loess records, China, 2:603–604
 micro/macroscale forms, 1:865–876
 rates
 determination of, 1:441–443
 glacial landforms, 1:859–860
 soil production function, 1:442
 RSE, sequence stratigraphy, 4:261, 4:263
 sedimentary
 clast form analysis, 2:1
 indicators of, 4:389–390
 tectonics influence, 1:442–443
 TSE, sequence stratigraphy, 4:260–261, 4:262, 4:265, 4:273
 erosion-island plots, 1:433, 1:434f, 1:435
 erosion marks, 4:377–384
 benches/platforms/strand flats, 4:378–379, 4:379f
 bioeroded holes, 4:380–381
 depositional indicators, 4:381–384
 honeycombs/tafoni/pools/potholes, 4:379
 notches, 4:377, 4:378f
 sea caves/arches, 4:379, 4:380f
 spurs/grooves/furrows/surge channels, 4:380
 erosional forms
 glacier flow directions, 1:895–896
 glacifluvial, 1:825–838
 icesheet/glacier extent, 1:885, 1:885f
 major scale landforms, 1:847–864
 micro/macroscale, 1:865–876
 erosional response model, 4:389–390
 erratics
 glacial, 2:81–84
 icesheet/glacier extent, 1:884
 erratic trains, 2:81–83, 2:82f
 errors/error sources
 radiocarbon dating, 4:324–328
 accuracy, 4:325, 4:325f
 calibration, 4:326–327
 estimations, 4:325–326
 interlaboratory collaborative trials, 4:326, 4:327f
 measurement uncertainty, 4:324
 μ and σ estimation, 4:325
 precision, 4:325, 4:325f
 sampling uncertainty, 4:325
 stochastic models, 4:325
 traceability, 4:324
 uncertainties, 4:324, 4:325–326
 sea-level changes, 4:372–373, 4:375f
 tree-ring dating, 2:107t, 2:108f
 eruptions *see* volcanic eruptions
 ERV *see* extended R value
 ESA *see* Early Stone Age
 escorias, South America, 2:632, 2:634f, 2:634t
 eskers
 glacier recession, 1:807, 1:808f
 ice-contact landforms, 2:13–14, 2:14f
 icesheet/glacier extent, 1:887–888, 1:888f, 1:889f, 1:890f
 ESL *see* eustatic sea-level
 ESR *see* electron spin resonance dating
 Essex, England beetle assemblages, 1:185, 1:186, 1:188, 1:189
 Estonian sea level changes, 4:491
 estuary records, North American pollen, 4:136–140
 ET *see* equatorial trough
 Eugenie Glacier, Canada, 1:761f
 euhedral zircons, 1:651
 Eurasia
 Arctic region
 key sites, 3:735–737, 3:736f, 3:737t
 plant macrofossil records, 3:733–745
 vegetation history, 3:738–742
 deglaciation, 2:232
 glaciations, 2:224–235
 Alps, 2:231–232
 British/Irish ice sheet, 2:230–231
 Eemian interglacial stage, 2:224, 2:225f
 Late Pleistocene, 2:224–235
 Middle-Pleistocene, 2:172–179
 mountain ranges, 2:232
 Northern Russia, 2:224–228
 Scandinavian ice sheet, 2:228–230, 2:228f
 Weichselian cold stage, 2:224–232, 2:226f, 2:227f
 ice sheet
 British/Irish, 2:230–231
 growth/decay, 1:879f, 1:880f, 1:882
 Scandinavian, 2:228–230, 2:228f
 loess distribution, 2:574f
 megafaunal extinctions, 4:707–709, 4:708t
 Middle-Pleistocene glaciations, 2:172–179
 Neanderthal demise, 1:146–153
 vertebrate studies, 4:600–601, 4:605–614
 Europe
 1.9 MYR-300,000 years ago, 1:83–90
 beetle fauna
 Late Pleistocene, 1:200–206
 MCR studies, 1:167–168
 Middle Pleistocene, 1:184–190
 postglacial, 1:274–281
 stenothermic beetle distribution, 1:163f
 biogeography, 3:318, 3:319f
 chironomid records, 1:373–385, 1:386–397
 dune fields, 1:599, 1:600f
 age determination, 1:603–604
 origins, 1:602
 types, 1:601, 1:602f
 Early Quaternary, 2:168–170
 glaciations, 2:224–235
 history of, 2:144, 2:147t
 Middle-Pleistocene, 2:172
 hominin colonization, 1:84–87
 MPT, 1:84–87
 post-MPT, 1:87
 pre-MPT, 1:84
 human adaptations, postglacial, 1:155–159
 human recolonization, 1:55f
 hydrological changes, 2:551–552, 2:553f
 ice age history, 1:4, 1:5f
 lake level studies, 2:549–557
 Last Interglacial pollen records, 4:1–8
 lithostratigraphy, 4:227–242
 loess records, 2:606–619
 cryoturbation, 2:607–608, 2:613
 dating techniques, 2:609–610
 geochemistry, 2:612–613
 geochronology, 2:609–610
 ice wedges, 2:613, 2:614f
 mineralogy, 2:606
 morphology, 2:606–608, 2:608–609
 North Atlantic climatic variability, 2:615–617
 origins, 2:608–609
 paleoclimatic proxies, 2:612–615
 paleosol significance, 2:609
 paleozoogeography, 2:613–615
 particle size distribution, 2:606–608, 2:615, 2:616f
 sedimentation variability, 2:610–612
 sedimentology/grain size, 2:615, 2:616f
 stratigraphy, 2:609, 2:611f, 2:612f, 2:617f
 terrestrial mollusks, 2:613–615, 2:615f
 Western Europe, 2:607–608, 2:610–612
 MCR studies, beetle records, 1:167–168
 megafaunal extinctions, 4:707–708, 4:708f
 Middle-Pleistocene
 glaciations, 2:172
 vertebrate studies, 4:639–645
 Neanderthal demise, 1:146–153
 neoglaciation, 2:257–268
 Northwestern mire/peat macrofossils, 3:638–652
 oribatid mite studies, 2:683–684
 postglacial
 beetle fauna, 1:274–281
 chironomid records, 1:386–397
 human adaptations, 1:155–159
 pollen records, 4:179–188
 relative sea level changes, 4:491–493
 vertebrate studies
 distribution of remains, 4:595f
 Early–Middle Pleistocene, 4:605–614
 Middle Pleistocene, 4:639–645
 ‘pickled’ fauna, 4:716
 Younger Dryas, 3:126, 3:127f
 see also Britain/British Isles; eastern Europe; United Kingdom
 European Pollen Database (EPD), 3:831–832
 European projects for ice coring in Antarctica (EPICA), 2:437
 European settlement in Australia/New Zealand, 4:112
 European–Siberian subarea, Early Pleistocene, 4:606f
 eustasy, sequence stratigraphy, 4:263–264, 4:271

- eustatic sea-level (ESL)
 changes, 4:370, 4:498
 Barbados, 4:432, 4:433–434, 4:433f
 Coorong coastal plain, Australia, 4:432, 4:434, 4:435f
 glacial–interglacial cycles, 4:429–438
 Heinrich events, 4:432
 high-frequency changes, 4:432
 highstand magnitude, 4:434–435
 Holocene, 4:447–449
 Huon Peninsula, 4:430f, 4:432–433, 4:433f, 4:436
 since LGM, 4:439–451
 LGM, 4:429, 4:431, 4:432–434, 4:436
 localities, reconstructions, 4:432–434
 modeling, 4:375
 prior to LGM, 4:432–434
 timing of, 4:434
 relative sea-level distinction, 4:465
 speleothems, 4:461
see also relative sea-level change; sea-level change
- eutrophication
 chironomid effects, 1:359
 cladocera, 3:277–278, 3:271ge
 dinocyst relationship, 2:806–808, 2:812
 lakes, 3:657, 3:660f
 chemistry, 3:293–295
 ecosystems, 3:659–665
 sediment diatom records, 1:541–543
 pigment studies, 3:330–331, 3:332–333
- evaporation
 lake-level modeling
 deterministic models, 2:558
 energy balance, 2:559
 water balance, 2:558–559
 oxygen isotope ratios, seawater, 2:916, 2:917f
 precipitation, Asia, 2:506, 2:511, 2:512f, 2:517–518
 soil water, 1:323
 evaporite varves, 4:584
 evapotranspiration, 2:559, 2:560f
 event-related disturbances, tree growth, 2:93–94, 2:94t
- Everest, Mount, Late Quaternary glaciation, 2:238–239, 2:239f
- Everglades, Florida pollen records, 4:133–134, 4:134f, 4:134t, 4:140
- evidence
 botanical, 4:542
 collection, 4:523, 4:544–545
 forensic materials, 4:542, 4:544–545
 geological, 4:523
 soil/sediment, 4:535
 tests, 4:542
- evolutionary spectral analysis, climate change, 3:96, 3:97f
- evolution of species, 4:206–207
 biogeography/paleolimnology, 3:322–323
 biostratigraphic value, 4:729–730
 cichlid fishes, 2:504
 diatoms, Pacific Ocean, 1:586, 1:586f
 Early–Middle Pleistocene mammals, 4:605–614
- Homo sapiens*, 4:693
 mammoth, 4:207
 North Atlantic Pleistocene, 3:9–10
 vertebrates, 4:723–732
 voles, 4:207–208, 4:209f
see also human evolution
- examination methods, forensic soil/sediment, 4:537–540
- excavation methods, vertebrate remains, 4:593–594
- exhaustion models, paraglacial geomorphology, 3:555f, 3:556
- exhumation rates, tectonics, 1:442–443
- exhumed paleosols, 3:367, 3:368
- exoskeletons, 1:165–166, 1:166f, 1:169, 1:173
- expanding glaciers, Antarctic, 2:220
- expansion bars, 2:11
- expansion fans, 1:898
- exposure dating *see* cosmogenic nuclide exposure dating
- exposure geochronology, 1:432–439
- expressed population signal (EPS), dendroclimatology, 1:460
- extended last glacial maximum (eLGM)
 beetle assemblages, 1:244, 1:247–248, 1:248–249, 1:249–251, 1:251–253
 climate reconstruction, 1:249–251
- extended R value (ERV), pollen analysis, 3:796
- extension rate, coral Mg/Ca, 2:880
- extinctions
 beetles
 New Zealand, 1:245–246, 1:246–247, 1:246f
 South America, 1:241–242
 biogeography/paleolimnology, 3:322–323
 biosphere history, 1:23–24
 climate change and, 1:54
 LP/EH boundary, 4:691
 mammals, 4:208
 Africa, 4:664–665, 4:668–669
 hominin interactions, 4:596
 Late Pleistocene, 4:697
 North America, 4:676–678
 South America, 4:682t, 4:690t
 megafaunal
 causation theories, 4:709–711
 human impact
 Americas, 1:126–127
 Australia, 1:113–115
 Late Pleistocene, 4:700–712
 Neanderthals, 1:146, 1:150–152
 plant species, 2:730
 teeth/bones of species, 1:310–311, 1:311–312
 vegetation, Southern Europe, 4:67
 extirpation (regional extinctions), 1:241–242
- extrapolation approach, paleoceanography, 2:812
- extraterrestrial material, ice-core impacts, 2:283, 2:309
- Eyjafjöll volcano, Iceland, 4:279, 4:280–281
- Eyre Lake, Australia, 2:483–484, 2:490f, 2:525f, 2:526, 2:527–528, 2:527f, 2:528–529
- ## F
- fabric analysis, till, 2:76–80
- facies
 analysis, high-energy coasts, 4:386–387, 4:387–388, 4:392
 associations, glacial depositional systems, 2:20–21
 definition, 2:18
 models
 glacial depositional systems, 2:20–21
 glaciolacustrine depositional system, 2:25f
 glaciomarine depositional system, 2:27f
 subglacial depositional system, 2:24f
 supraglacial depositional systems, 2:26f
 preservation potential, 4:391
- facilitation in climate change, active layer, 3:426–427
- factor loadings, diatom assemblages, 1:577f
- FAD *see* first appearance datum
- Fagus* *see* beech
- Fagus sylvatica* beech, 3:850–852, 3:851f
- Fairbridge, RW, 4:413
- Falkland Islands, 3:514, 3:516f, 3:519, 3:520f, 3:521–522
- fall dump, diatoms, 1:554, 1:556–557
- FALL sampling method, lichenometry, 2:570–571
- False Cougar Cave, Montana beetle studies, 1:225, 1:226t
- Fannid flies, 4:551
- fans
 coalescent subaqueous fans, 1:776
 glacial erratics/till dispersal indicators, 2:81–83
 glacier flow directions, 1:896–902, 1:899f, 1:900f
 high-latitude continental shelves, 2:33–35, 2:36f, 2:37f, 2:38f
 ice-proximal settings, 2:30, 2:32f, 2:33f
 latero-frontal fans, 1:770–772, 1:772f
 paraglacial geomorphology, 3:554–556, 3:559–560
 pedostratigraphy, 4:257–258
 subaerial outwash, 2:14–15, 2:14f
 subaqueous outwash, 2:46–48, 2:48f
 surfaces, lobe abandonment, 1:435–436
- far-field regions
 eustatic sea-level changes, 4:441–444, 4:444–449
 relative sea-level changes, 4:489, 4:495, 4:498
- Farrell, WE, 4:455
- far-traveled dust *see* long-range-transported dust
- fast ion chromatography (FIC), 2:327–328
- fast nucleon interactions, 1:427
- fatty acids, diatoms, 1:590
- faulting tectonics, 4:503, 4:504–505, 4:506–507, 4:507–509
- fault trenches, pedostratigraphy, 4:258
- fauna
 climate change response, 3:790
 Late Pleistocene South America, 4:683–690
 Middle Pleistocene

- Africa, 4:615–619
 Southern Asia, 4:651–663
 Quaternary changes, 4:206–209
see also individual types; mammals; megafauna; vertebrate studies/records
 FCCFs *see* frost coated clastic flows
 feature-focused forensic searches, 4:525*t*
 feedback mechanisms
 biosphere feedbacks, 1:721–728
 global climate change, 1:31–32
 tectonic uplift, 2:129*f*
 vegetation, last interglacial, 3:158–159, 3:160–162
 feldspar
 OSL dating, 2:662, 2:663*f*, 2:664
 TCNs, 1:414
 TL dating, 2:646–648, 2:649, 2:650–651
 felsenmeer *see* blockfields
 felsic compositions, 1:792–793, 1:780*ge*, 1:796
 felsic domes, 1:796, 1:799*f*
 felsic lobes, 1:796
 felsic tuyas, 1:792–793, 1:795*f*, 1:796*f*, 1:800*f*
 fen macrofossils, 3:602*f*, 3:603*f*, 3:637–656
 Fennoscandia
 chironomid records, 1:389–392, 1:391*f*
 ice sheet, 1:899–900, 1:900*f*, 1:904*f*
 megafossils, 3:625–626, 3:627*f*
 ferrets, 4:715
 ferrimagnetism, 3:375*ge*, 2:884–885, 2:886*f*, 2:887*t*
 ferrimanganiferous coatings, soil, 3:385*f*
 ferrogenic varves, 4:584
 ferromagnesian assemblages, 4:283–284
 ferromagnetism, 3:375*ge*, 2:884–885, 2:886*f*, 2:887*t*, 3:375, 3:376
 Fe/Ti-oxides
 minerals, 1:410–411
 tephrochronology, 4:285–286
 FFA *see* free fatty acids
 FIBS *see* Functional Interpretation of Botanical Surveys
 FIC *see* fast ion chromatography
 field measurements, clast forms, 2:2–3
 fill terraces
 aggradation, 1:684
 formation, 1:688, 1:688*f*, 1:690*f*
 fine-grained matrix geochemistry, blockfields, 3:526–528
 fine-grained sediments, 3:360
 fingerprinting
 ice core microparticle provenance, 2:343–344
 paleoclimate record, 3:244–245, 3:245*f*
 finite mixture model (FMM), OSL dating, 2:661
 Finland
 ice sheet/glaciations, 2:228, 2:230
 Fiordland, South Island, New Zealand, 2:211, 2:213*f*
 fir (*Abies*), 4:185–186
 fires
 Australian/New Zealand pollen records, 4:112
 charred particle analyses, 2:716–729
 dendroclimatology, 1:466, 1:467*f*
 NWN pollen records, 4:128–129, 4:128*f*, 4:129–130
 see also biomass burning; forest fires
 firn
 definition, 1:909*ge*
 gas records, 2:463, 2:464*f*
 Greenland ice sheet, 2:442
 firn air, 2:361–372
 age distributions, 2:366–367
 atmospheric reconstruction, 2:366–367
 dating, 2:366–367
 density, 2:361–362, 2:363–364, 2:364*f*
 diffusion, 2:364–365, 2:366, 2:369–370
 oxygen isotopic CO₂ composition, 2:367, 2:367*f*
 porosity, 2:361–362, 2:362–363, 2:364*f*
 sampling, 2:361, 2:362*f*, 2:363*t*
 structure, 2:361–364
 transport
 bubble-trapping, 2:369–370
 delta age, 2:367–368, 2:368*f*
 density layering, 2:363–364
 diffusion, 2:364–365, 2:369–370
 gas records, 2:367–370
 gas thermometry, 2:368
 gravitational fractionation, 2:367–368
 ice core record, 2:367–370
 isotopic diffusive fractionation, 2:369, 2:370*f*
 mass transfer mechanisms, 2:364–366
 molecular size close-off fractionation, 2:368–369, 2:369*f*
 open porosity air flow, 2:362–363
 properties, 2:361–364
 smoothing by diffusion, 2:369–370
 thermal fractionation, 2:365*f*, 2:368
 trace gases, 2:364–366
 first appearance datum (FAD), 4:730
 fish
 evolution, Africa, 2:504
 lake acidification, 3:341–342, 3:342*f*, 3:346
 otoliths, 1:348–350, 1:349*f*, 1:350*f*, 1:351*f*
 fishtail points, 1:123*f*, 1:128–129
 fission products, ice cores, 2:305*t*, 2:306
 fission-track (FT) dating, 1:11–12, 1:643–662
 age equation, 1:644–645
 applications, 1:652–656
 geochronology, 1:645–646, 1:646–647
 glass methods, 1:644, 1:644*f*, 1:646–649, 1:649–650, 1:650–651
 Indonesia, 1:655–656
 New Zealand, 1:652–655
 PTF correction methods, 1:649–650
 tephrochronology, 4:287
 fissure fillings
 fossils, Southeast Asia, 4:693, 4:696–697
 vertebrate remains, 4:655–662
 fitness indicators, definition, 1:83*ge*
 fjords, 1:854–857, 1:858*f*, 1:859*f*
 glacier/icesheet extent, 1:885–886, 1:887*f*
 Greenland, 3:760, 3:761
 high latitudes, 1:559
 paraglacial coasts, 3:562–563
 South Island, New Zealand, 2:211, 2:213*f*
 subaquatic glacial depositional systems, 1:763
 flaked flakes, definition, 1:83*ge*
 flakes
 Acheulean Industry, 1:59, 1:64*f*
 definition, 1:83*ge*
 Early Pleistocene Asia, 1:67–68
 Middle Pleistocene Asia, 1:77
 Oldowan Industry, 1:59, 1:61–62
 technocomplex groupings, 1:147
 flesh flies, 4:550
 flies
 forensic science use, 4:548, 4:549–552
 Late Pleistocene records, 1:373–385
 overview, 1:355–360
 see also chironomids
 flightless birds, 4:746
 floating superior plant macrofossils, 4:364
 flood-basin sediments, 1:667–668
 floodplains
 incision/aggradation, 1:677, 1:679*f*
 sediments, 1:663, 1:667*f*
 Mississippi River, 1:664*f*, 1:665*f*, 1:668*f*, 1:669*f*
 Missouri River, 1:666*f*
 stratigraphy, 4:248
 floods
 catastrophic floods, 1:825, 1:830–831, 1:832*f*
 GLOFs, 2:43
 modern analog approaches, 3:105–106
 response
 Holocene climate, 3:255–256
 LGM, Pacific Ocean, 1:581–585
 flora
 climate change response, 3:790
 Greenland, present day, 3:760
 Quaternary changes, 4:206–209
 see also plant...; vegetation...
 Flores Island, Southern Asia
 human evolution, 1:142–143
 Late Pleistocene mammals, 4:696
 vertebrate records, 4:662
 Florida, US
 pollen records, 4:133–134, 4:134*f*, 4:134*t*, 4:140
 Tulane Lake, 2:545, 2:545*f*
 see also southeastern North America
 flow cytometers, ice cores, 2:294, 2:295*f*
 flow law, ice, 2:443, 2:456
 flow-sets, glaciers, 1:896–902, 1:897*f*, 1:903*f*, 1:905*f*
 flowstones, 1:294, 1:295*f*, 4:460–461
 see also speleothems
 flowstripes, West Antarctic Ice Sheet, 2:453
 flow switching, glaciers, 1:901
 flow traces, West Antarctic Ice Sheet, 2:453
 Flower, RJ, 1:517–518
 fluorination, diatom silica, 2:735
 fluorosis, loess records, 2:604
 flutes, glacier recession, 1:807–808
 fluvial channels *see* channel...
 fluvial-dominated deltas, 1:696, 1:697*f*
 fluvial environments
 deltas, 1:693–703

- fluvial environments (*Continued*)
 glacial–interglacial scale responses, 2:112–125
 paleohydrology, 3:253–258
 plant macrofossil/pollen studies, 3:725, 3:726–727
 rapid environmental change
 case studies, 1:679–682
 responses to, 1:676–683
 sediments, 1:663–675
 terrace sequences, 1:684–692, 4:644
 vertebrate studies, 4:644
 fluvial incision/aggradation, 1:676–678, 1:679f
 fluvial paleohydrology, 3:253–258
 fluvial sediments, 1:663–675
 allostratigraphy, 4:246f, 4:248
 morphostratigraphy, 4:243
 plant macrofossil/pollen studies, 3:726–727
 vertebrate remains, 4:591–592
 see also periglacial fluvial sediments and forms
 fluvial terraces, 1:684–692
 formation, 1:684–685, 1:685–689
 terminology, 1:684
 types, 1:684
 vertebrate studies, Europe, 4:644
 fluviodeltaic systems
 climate influence, 1:699
 Gulf of Mexico, 1:701f, 2:123f
 modeling studies, 1:700–702
 see also deltas
 flux/fluxes
 cosmic rays, 1:429–430
 terrigenous sediments, 2:947–948
 FMM *see* finite mixture model
 folding
 Late Quaternary tectonics, 4:503, 4:507–509, 4:508f
 volcanic tephra layers, 2:418–419
 food industry, diatoms, 1:504
 food plants, forensics, 4:543
 food production archeology, 3:705, 3:708–713, 3:713–714
 food-provision stages, archeology, 3:700t
 food web changes, aquatic studies, 3:333–334
 foraminifera
 alkenone paleothermometry, 2:762
 amino-acid dating, 1:38, 1:43, 1:44, 1:44f, 1:46
 Arabian Sea, 3:51f
 carbon-13 reconstruction, 2:899, 2:901f, 2:902, 2:903, 2:903f
 deep-sea carbonate dissolution, 2:866
 $\delta^{18}\text{O}$
 combined studies, 1:486–487
 paleoceanography, 2:919–921, 3:48f
 eustatic sea-level change, 4:436
 Holocene RSL reconstruction, 4:419, 4:420f, 4:421f, 4:424–425, 4:424f
 Indian Ocean, 3:46–47, 3:48f, 3:53f
 inter-/intra-species variability, 2:873–874
 marine carbon cycle proxies, 2:853, 2:854
 Mg/Ca paleothermometry, 2:871–877
 mire/peat macrofossils, 3:652, 3:653f
 North Pacific Ocean, 3:33, 3:34–35, 3:35–36, 3:35f, 3:38f, 3:40–42, 3:42f
 oxygen isotope ratios, 3:28, 3:28f, 3:29f
 paleoceanography, 2:745–746, 2:746f, 2:747f, 2:919–921, 3:48f
 Pliocene environments, 3:186f
 salinity proxies, 2:932–934, 2:936–937, 2:937f, 2:939f
 shell thickness, 3:35–36
 see also benthic foraminifera; planktic foraminifera; planktonic foraminifera
 forcing factors/mechanisms
 lake level studies, 2:553–556
 last millennium climate, 3:233–235
 monsoons, 3:49–52
 South Pacific ocean postglacial records, 3:82–84, 3:83f
 see also climate forcing; radiative forcing
 forensic archeology, diatom analysis, 1:520
 forensic botany
 education programs, 4:546
 general use of, 4:542–543
 macroscopic plant remains, 4:542–547
 forensic ecology, 1:524, 4:543, 4:545–546
 forensic entomology, 4:548–555
 forensic geology
 burial location searches, 4:521–534
 future of, 4:540–541
 history, 4:535
 objectives, 4:523, 4:524t
 recent developments, 4:523
 soil/sediment examination, 4:535–541
 forensic palynology, 4:556–566
 analytical techniques, 4:556–566
 DNA analysis, 4:559–561
 equipment/cost/time/expertise, 4:560
 extraction/processing, 4:559–560
 pollen counting/identification, 4:560
 light microscopy vs. SEM, 4:556–559
 Raman spectroscopy, 4:561–562
 SEM vs. light microscopy, 4:556–559
 stable isotope analysis, 4:562–564
 training of palynologists, 4:565
 forensic science
 advanced degree programs, 4:546
 botanic education programs, 4:546
 general education, 4:546
 Quaternary proxies use in, 1:522–525, 4:521–534, 4:535–541, 4:542–547, 4:548–555
 forest fires
 boreal forests, 3:576f
 NWA pollen records, 4:128–129, 4:128f, 4:129–130
 permafrost regions, 3:465–466
 forest hollows, 3:847, 3:847f, 3:848, 3:852f
 forest line/limit *see* timberlines
 forest-prairie border, United States, 3:768–769, 3:773f
 forest-prairie ecotone, United States, 3:769f, 3:773f, 3:777, 3:783
 forest refugia, New Zealand, 1:251–253
 forests
 Africa
 expansion, 4:95–98
 pollen records, 4:10–12, 4:12f, 4:89–91, 4:91–95, 4:95–98, 4:98–100
 tropical/equatorial, 4:89–91, 4:98–100
 biota contraction, Australia, 1:198
 clearance impact on beetles, 1:274, 1:276–277
 Eastern Africa expansion, 4:95–98
 European pollen records, 4:1–2, 4:2–3, 4:4–5, 4:5–6, 4:175–176
 grazing impact, 3:890
 Guineo–Congolian forests, 4:91–95
 habitats, Holocene North America, 3:768, 3:770t
 marine west coast pollen records, 4:127–129
 Midwestern United States, 3:768–769, 3:769f, 3:773f, 3:777, 3:783
 New Zealand eLGM, 1:251–253
 postglacial pollen records
 Northern Europe, 4:175–176
 southeastern North America, 4:133–135
 stand-scale palynology, 3:847, 3:847f, 3:848, 3:852f
 tropical rain forests, Africa, 4:10–12, 4:12f
 see also forest fires; pollen, southeastern North America; rainforests
 forest tundra
 Arctic Eurasia, 3:743
 Northern Asia, 4:27, 4:37
 formations
 definition, 1:83ge
 lithostratigraphy, 4:227–228, 4:235
 slope forms, 3:481–489
 forward modeling, POLLSCAPE, 3:877–878
 fossil fuels, 4:333, 4:342
 fossils
 Africa
 chironomids, 1:361–372
 Middle Pleistocene, 4:615, 4:616, 4:618
 vertebrates, 4:664–671
 Antarctic climate change, 1:527
 Arctic site map, 1:179f
 beetles
 Late Pleistocene, 1:191–199, 1:200–206
 Japan, 1:207–220
 New Zealand, 1:244–254
 Northern Asia, 1:255–273
 South America, 1:235–243
 Late Tertiary/Early Quaternary, 1:173–183
 Middle Pleistocene, 1:184–190
 postglacial
 Europe, 1:274–281
 North America, 1:282–290
 calcareous marine fossils, 2:871–883
 carbonates, $\delta^{18}\text{O}$ inferred, 2:919
 chironomids
 Africa, 1:361–372
 Late Pleistocene, 1:373–385
 overview, 1:355–360
 dating, continental biostratigraphy, 4:206–214
 dinoflagellates, 2:800–815
 freshwater laminated sequences, 1:540–545

- leaves
 CO₂ reconstruction from, 3:613–620
 identification for analysis, 3:615
 marine laminated sediments, 1:554–561
 megafaunal extinctions, 4:700–712
 pollen
 climate proxies, 3:805–815
 Late Pleistocene
 Australasia, 4:23–24
 Western North America, 4:72–83
 paleodroughts, 3:195
 preservation of, 1:173–174, 3:706f
 rarity of locations, 1:174
 site distribution
 Japan, 1:207–211, 1:208t
 North America, 3:675
 Southeast Asian Late Pleistocene, 4:693–699
 treeline evidence/models, 3:697
 vertebrates, 4:590–597
 Africa, 4:664–671
 ancient DNA, 4:719–721
 North America, 4:673–679
 South America, 2:631–632, 4:680–692
 Southern Asia, 4:651–663
 speciation/evolution, 4:723–732
 Yedoma of Beringia, 3:548
see also beetle fauna; diatom...; insect fossils; macrofossils
- fossil water measurements, 2:921
- Fourier transform infrared spectroscopy (FTIRS), 3:351, 3:353–354, 3:354f
- fractionation
 firn air, 2:365f, 2:367–368, 2:368–369, 2:369f, 2:370f
 silicon isotopes, diatoms, 2:735–737, 2:738f
- fracture erosional forms, 1:867–869
- France
 Alps, glaciations, 2:231
 beetle studies, 1:189
 pollen records
 Middle/Late Pleistocene, 4:65–66
 postglacial, 4:180t, 4:183–185
 Pre-Alps lake level studies, 2:549
 relative sea level changes, 4:493
 fraud case evidence, 4:535
 Frederic, Hurricane, paleotempestology, 3:209–210
 freeze corers/coring
 laminated sediments, 4:584–585
 paleolimnology, 3:262f, 3:263–264
 freeze-drying varved sediments, 4:585
 freeze-thaw processes
 active layer, 3:424, 3:426f
 blockfields, 3:525
 sea ice oxygen isotopes, 2:918
 slope deposits/forms, 3:481, 3:483f
 freezing front propagation, active layer, 3:422
 French... *see* France
 frequency-dependent susceptibility
 definition, 3:375ge
 soils, 3:379–380
 freshwater algae records, Africa, 4:88–89
 freshwater beds, West Runton, 1:185
 freshwater biogenic carbonates, 1:341–353
 freshwater diatoms, 1:473–475
 Antarctica, 1:536–538
 laminated sediments, 1:540–545
 large lakes, 1:546–553
 freshwater fluxes, climate change modeling, 3:207
 freshwater input, South Atlantic, 3:22–24
 freshwater lake-level reconstructions
 carbonate sediments, 2:492f, 2:493
 chironomids, 2:493
 Cladocera, 2:493
 clastic sediments, 2:489–493
 diatoms, 2:492–493
 site selection, 2:487
 freshwater mollusks, 3:281–291
 bivalves, 3:281, 3:282f, 3:283t
 gastropods, 3:281–283, 3:282f, 3:283t
 nonmarine biogenic carbonates, 1:343–346
 paleolimnological reconstructions, 3:284–285
 quantitative fidelity, 3:289
 sample size effects, 3:287–288
 sclerochronology, 3:288–289
 spatial fidelity, 3:289–290
 stable isotopes, 3:285–289
 taphonomy, 3:289–290
 taxonomy, 3:281–283, 3:283t
 freshwater routing, Laurentide Ice Sheet, 3:131
 freshwater sponge macrofossils, 3:652, 3:653f
 freshwater transport, Late Neogene/Early Quaternary transition, 2:162
 frictional processes, block streams, 3:518
 friction cracks, 1:870–871, 1:872f, 1:873–874
 Frog Hall Pit, England beetle assemblages, 1:188
 frog vertebrate records, 4:621, 4:622t
 frost boils, patterned ground, 3:459–460
 frost coated clastic flows (FCCFs), 3:570, 3:572
 frost creep, 3:481–482, 3:484f
 frost heave
 active layer, permafrost, 3:424
 differential, 3:452, 3:454–457, 3:461
 frost mounds, 3:472–480
 frost polygons, patterned ground, 3:458–459
 frost shattering *see* macrogelivation
 frozen fauna
 circumpolar, 4:713–715
 preservation, 4:716
 frozen plant remains
 economics, 3:711f
 preservation, 3:702
 frozen sediments, insect preservation, 1:174
 fruit remains, environmental archeology, 3:699
 FT *see* fission-track dating
 FTIRS *see* Fourier transform infrared spectroscopy
 Functional Interpretation of Botanical Surveys (FIBS), 3:1ge
 fungi macrofossils, 3:652, 3:653f
 furrow erosion marks, 4:380
 Furtwängler Glacier (FWG), Tanzania, 2:373, 2:374–375, 2:374f
- G**
- GAD *see* geocentric axial dipole
 galactic cosmic rays (GCRs)
 cosmogenic radionuclide production, 2:353–354
 energy spectra, 1:419, 1:420
 geomagnetic modulation, 2:353
 intensities, time histories, 1:421f
 proton spectra, 1:423f
 solar modulation, 2:353
 variations, 2:353–354
 Galerian biozone, 4:210, 4:211f
 Galerian stage mammal assemblages
 Europe, 4:639–641
 Northern Eurasia, 4:610–611, 4:611–612
 gamma-ray spectrometers, 2:648f
 Ganges–Brahmaputra river delta, 1:700, 1:702f
 Gangotri Glacier, India, 1:805f
 garnet, TCNs, 1:410–411
 gas chromatography
 alkenone paleothermometry, 2:756, 2:757f
 methane studies, 2:340
 gas chromatography with flame ionization detector (GC-FID), 2:756, 2:757f
 gas chromatography with mass spectrometry (GC-MS), 2:756
 gas proportional counting (GPC), 4:311–313
 activity calculations, 4:313
 anticoincidence shielding, 4:312
 CO₂ preparation, 4:313
 counters, 4:311–312, 4:313f, 4:314t
 electronics, 4:312
 Groningen counters, 4:313f, 4:314t
 instrumentation, 4:311–312
 passive shielding, 4:312, 4:312f
 procedures, 4:313
 radiocarbon dating, 4:311–313
 sample analysis, 4:312–313
 schematic representations, 4:311f, 4:312f
 gas records
 ice cores
 correlation techniques, 2:410–412, 2:413–414, 2:415
 history of, 2:463–470
 measuring gases, 2:463–470
 ice-margin sites, 2:424–427
 gas thermometry, firn air, 2:368
 gastropods, 3:281–283, 3:282f, 3:283t
 gates, ice sheets, 2:461f
 Gaussian models, radiocarbon dating, 4:324, 4:326–327, 4:326f
 gazettes on climatology, 3:240–241, 3:241f
 GCD *see* Global Charcoal Database
 GC-FID *see* gas chromatography with flame ionization detector
 GCM *see* global circulation model
 GC-MS *see* gas chromatography with mass spectrometry
 GCMs *see* general circulation models
 GCRs *see* galactic cosmic rays

- GDGTs *see* glycerol dialkyl glycerol tetraethers
- Geikie, James, 1:3, 1:4, 1:4f
- Gelasian Stage
- chronostratigraphy, 4:216
 - mammal assemblages, 4:605–607
- gelatinous layer (G-layer), tension wood, 2:97
- gelifluction, 3:481–482, 3:482–483, 3:484f
- gem fraud evidence, 4:535
- general circulation models (GCMs)
- Asian lake level studies, 2:516–517, 2:520–521
 - isotopic composition, 2:395, 2:396
 - last interglacial, 3:155, 3:156–158, 3:158–159, 3:160, 3:162
 - modern analog approaches, 3:102, 3:108–109, 3:110
 - numerical environmental change modeling, 1:32–33, 1:33–34
 - paleoclimate introduction, 3:87, 3:88–89, 3:90
 - paleodroughts, 3:196
 - tectonic uplift/continental configurations, 2:134–135
 - volcanic/solar forcing, 1:749–750, 1:750–751, 1:751f, 1:752, 1:752f
 - see also* Last Glacial Maximum GCMs
- general geochronological age equation, argon dating, 2:478
- genetic diversity, paleophytogeography, 2:730–731
- genetic evidence, out-of-Africa model, 1:91–92
- genetic techniques, plant distributions, 3:855, 3:856–857, 3:858
- genome recycling, ice cores, 2:295–296, 2:296f
- genotypes as temperature proxies, 2:842–843
- Genyornis* flightless bird eggshell, 4:194–195, 4:195f
- geoarchaeology, 4:258
- see also* archaeology
- geocentric axial dipole (GAD), 1:706–707
- geochemical signals as indicators, 2:83
- geochemistry
- blockfields, 3:526–528
 - calcareous skeletons, 2:795, 2:796, 2:798
 - forensic searches, 4:530, 4:532f
 - glacial erratics/till dispersal indicators, 2:83
 - ice cores, 2:326
 - lakewater trace metal analysis, 3:296–297
 - loess, 2:574, 2:612–613, 2:634–635, 2:637
 - proxy data, paleosols, 3:373–374
 - stable isotopes, 1:291–292
 - tephrochronology, 4:283–286
 - terrigenous sediments, 2:948
- geochronology
- Europe
 - loess records, 2:609–610
 - vertebrate studies, 4:640f - exposure geochronology, 1:432–439
 - FT dating, 1:645–646, 1:646–647
 - loess records, 2:609–610, 2:637, 2:638t
 - paleolimnology, 3:264
 - pollen databases, 3:834
 - South American loess records, 2:637, 2:638t
 - varved sediments, 4:582, 4:584, 4:587
 - vertebrate studies, 4:640f
- geodynamo, secular variation, 1:705, 1:706f
- geoforensic diatom analysis, 1:524
- geographical models
- modern human origins, 1:91–94
 - vertebrate evolution, 4:723
- geographic settings
- Africa, modern, 4:664
 - dinocysts, 2:805–808
 - West Antarctic Ice Sheet, 2:448–449
- geographic variability, active layer, 3:425–426
- geological analysis, forensic searches, 4:530
- geological evidence, forensics, 4:523
- geological mapping, 4:530
- geologic hazards
- earthquakes, 1:22–23, 1:23t
 - research history, 1:21–23
 - tsunamis, 1:22–23
 - volcanic eruptions, 1:21–22, 1:21t
- geology
- diatoms, varved sediments, 1:543
 - forensic
 - burial location searches, 4:521–534
 - soil/sediment examination, 4:535–541 - Late Pleistocene beetle records, 1:255–257, 1:262t
 - West Antarctic Ice Sheet, 2:448–449
- geomagnetic excursions, 1:705–720
- geomagnetic field, 1:706–709
- carbon-14 temporal variations, 4:338
 - cosmic ray flux modulation, 1:430
 - cosmogenic radionuclides
 - production, 2:353, 2:354f
 - reconstructions, 2:357 - definitions, 1:707f
 - global plots at Earth surface, 1:708f
 - historical records, 1:707–709, 1:710f
 - ice-core chronologies, 2:308
 - intensity determination, 1:712–713
 - models of, 1:707
 - at poles, 1:706–707
 - present-day field, 1:706–707
 - reconstructions, 2:357
 - secular variation, 1:705–720
- geomagnetic modulation, cosmogenic radionuclides, 2:353
- geomagnetic reversals, 1:705, 1:716–718, 1:718–719, 2:477–478
- geomagnetic variation, TCNs, 1:415
- geometry
- blockfields, 3:530–532
 - nuclear interaction rates, 1:429–430
- geomorphic processes
- process–event–response, 2:93–94, 2:94t
 - rapid environmental change, 1:678–679
 - reference chronologies, 2:92–93
 - tree-ring dating, 2:91, 2:92–93, 2:93–94, 2:94t
- geomorphic response time, fluvial environments, 1:678–679
- geomorphology
- archives, lake-level reconstruction, 2:488–489
 - areal ice extent, 1:884–890
 - glacier recession, 1:805–807
 - indicators, 4:377–384
 - alluvial rivers, 1:676
 - depositional features, 4:381–384
 - erosion marks, 4:377, 4:381–384 - isostasy, 4:452–459
 - lake-level archives, 2:488–489
 - mapping
 - forensic searches, 4:530
 - South Island, New Zealand, 2:211f - morphostratigraphy, 4:243, 4:248
 - paleohydrology and, 3:253
 - paraglacial, 3:553–565
 - permafrost–glacier interaction evidence, 3:508–510
 - southwest Pacific region, 2:202–215
 - tephrochronology, 4:293–294
 - tree-ring dating, 2:91–103
 - trimline/paleonunatak evidence, 1:925
 - see also* geomorphic processes
- geophysics, forensic searches, 4:530, 4:531f, 4:532f
- George Lake, Australia, 2:524–525, 2:525f
- geosols, 3:368, 3:368f
- aggradational pedogenesis, 3:372, 3:373f
 - pedostratigraphy, 4:254
 - stratigraphic markers, 3:369
- geothermal inputs, Vostok lake, 2:291
- Geotrupidae beetles, 1:213, 1:216f
- Germany
- ice sheet/glaciations, 2:228–229, 2:232
 - relative sea level changes, 4:491
- germination studies, aquatic seeds, 3:686
- Gesher Benot Ya'aqov, Israel, 1:71–72
- gestational diabetes mellitus, 1:315
- GIA *see* glacial isostatic adjustment
- giant fluvial bedforms, 2:9–10
- gravel antidunes, 2:10
 - gravel dunes, 2:9
 - ice-block obstacle marks, 2:10
 - see also* proglacial landforms
- giant moose, 4:715
- giant turtles, 4:745–746
- Gibraltar Strait, human expansion, 1:96
- gigantism *see* dwarfing/gigantism
- Gilbert-type deltas, 2:48f, 2:50
- GISP *see* Greenland ice sheet project
- GISP2 *see* Greenland ice sheet project 2
- GISS *see* Goddard Institute for Space Studies
- glacial activity *see* glacial processes
- glacial cirques *see* cirque glaciers
- glacial climates
- atmospheric dust, 1:729–736
 - biosphere feedbacks, 1:721–728
 - THC, 1:737–747
 - volcanic/solar forcing, 1:748–753
- glacial cycles
- CO₂ records, 2:315–316
 - last glacial cycle, Antarctica, 2:218–219
 - loess deposits, 2:573–584
 - Pleistocene Antarctica, 2:218
 - see also* glacial–interglacial cycles/stages/transitions
- glacial environments
- complexity, 2:18–20

- diamicts versus till, 2:18–20, 2:21f
 fluvial systems, 2:115–125
- glacial erosion
 Buzzsaw Hypothesis, 1:861
 landscapes of, 1:861–863
 rates of, 1:859–860
see also erosion...
- glacial erratics, 2:81–84
- glacial fluvial systems, 2:115–125
- glacial history
 Late Quaternary Antarctica, 2:222
 pre-Quaternary Antarctica, 2:217–218
- glacial ice volume, oxygen isotopes, 2:917–918
- glacial inception, 1:727–728
- glacial–interglacial cycles/stages/transitions
 Antarctic isotopic records, 2:398–400
 astronomical theories, 2:137, 2:139–140
 definition, 2:216g
 diatom records, Pacific Ocean, 1:580–581
 early Pleistocene records, 3:3, 3:5
 eustatic sea-level change, 4:429–438
 fluvial responses, 2:112–125
 fluvial terrace formation, 1:690–691
 history of ideas, 2:112
 ice-core research, 2:283
 magnetic susceptibility, 2:892f
 nitrous oxide history, 2:475
 North Atlantic Ocean, 3:10–12
 North Pacific Ocean, 3:34–36, 3:42–43
 paleoceanography overview, 2:748–750
 pollen records, 4:79–82
 Quaternary environments, 3:152–153
see also Holocene; interglacial–glacial cycles/changes/transitions
- glacial inversion, 1:895–896, 1:899, 1:901, 1:902
- glacial isostatic adjustment (GIA), 4:455, 4:455f, 4:456
 modeling, 4:373, 4:439, 4:440–441, 4:443–444, 4:447
 sea-levels, 4:369, 4:373, 4:439, 4:490f
- glacial lake outburst floods (GLOFs), 2:43
- glacial lake outbursts, 1:825–828, 1:828t
 erosional landforms, 1:826, 1:828t
 glaciolacustrine, 2:43
 hydrologic characteristics, 1:826–828
 overview, 1:825
 Pleistocene, 1:825–826, 1:826–828
 reconstructions of lake levels, 2:488f, 2:489
 spillways, 1:826–827, 1:827f, 1:828f
see also glaciifluvial landforms
- glacial landforms, 1:755–768, 3:553–565
 active temperate margins, 1:759
 cold regions, 3:574–581
 cosmogenic nuclide dating, 1:435
 dendrogeomorphology, 2:91–103
 dendroglaciology, 2:104–111
 ELA, 1:909–917
 erosional features, 1:865–876
 erosion rates, 1:859–860
 glaciated valleys, 1:764–765, 1:814, 1:815f, 1:819f, 1:820–822
 glacier recession, 1:803–812
 ice sheets
 ELA, 1:909–917
- extent, 1:884–894
 glacier flow directions, 1:895–908
 growth/decay, 1:877–883
 marginal associations, 1:759–761
 trimlines/paleoununataks, 1:918–928
- major scale forms, 1:847–864
 moraine forms/genesis, 1:769–779
 mountain glaciation, 1:764–765
- sediments
 associations, 1:756–757, 1:759–761, 1:764–765
 clast form analysis, 2:1–5
 glacial erratics indicators, 2:81–84
 glaciomarine, 2:30–42
 glaciolacustrine, 2:43–51
 ice-rafted debris, 2:30–42
 lithofacies, 2:18–29
 micromorphology, 2:52–61
 sequence stratigraphy, 2:85–90
 till, 2:62–75
 dispersal indicators, 2:81–84
 fabric analysis, 2:76–80
 subaquatic depositional systems, 1:761–764
 subglacial environment, 1:755–759
 volcanic, 1:780–802
see also fjords; periglacial landforms
- glacial landsystems, 1:813–824
 active temperature, 1:813–814, 1:817–820
 modern analogs, 1:813–817
- glacial loess, 2:574–575, 2:576f, 3:359
- glacially influenced marine depositional system, 2:26–28
- glacially killed trees, 2:104
- glacially modified trees, 2:108–109
- glacial prelude, human adaptations, 1:155–156
- glacial processes
 Antarctic diatom studies, 1:533–535, 1:535–536
 depositional systems
 facies models/associations, 2:20–21
 marine, 2:24–28
 terrestrial, 2:21–24
 large lake formation, 1:546–553
- glacial sediments
 allostratigraphy, 4:248
 micromorphology, 2:52–61
 developments, 2:60–61
 microfabric analysis, 2:53f, 2:56–58
 plasmic fabric analysis, 2:53f, 2:54f, 2:58–60, 2:58f
 structural analysis, 2:54–56
 textural analysis, 2:53–54
 thin sections, 2:52–60, 2:56f
 morphostratigraphy, 4:243
see also glacial landforms, sediments
- glacial sequence stratigraphy, 2:85–90
 idealized sequence diagrams, 2:86f, 2:87f
 Jurby Formation, Isle of Man, 2:89f
 Mona moraine, Norway, 2:88f
 nomenclature, 2:86–89
 sea-level change, 2:85
 terrestrial glaciogenic deposits, 2:85–86
- glacial timescale, West Antarctic Ice Sheet, 2:451–452
- glacial troughs *see* troughs
- glacial varves, 4:583–584
- glaciated valleys, 1:764–765, 1:814, 1:815f, 1:819f, 1:820–822
- glaciations
 active-layer processes, 3:421–429
 American cordilleras, 2:269–276
 beetle fauna
 Anglian, 1:186–187
 New Zealand, 1:247
 blockfields, 3:523–534
 causes
 astronomical theory, 2:136–141
 tectonic uplift/continental configurations, 2:127–135
 cryoturbation structures, 3:430–435
 definition, 4:222
 dune field age determination, 1:602–604
 frost mounds, 3:472–480
 history
 climatic change, 1:30
 overview, 2:143–150
 Holocene development
 Antarctic Peninsula, 2:219–220
 East Antarctica, 2:220–221
 landscape correlation, North America, 2:180–182
 landsystems, 1:813–824
 Late Pleistocene, 4:673
 Beringia, 2:191–201
 Japan, 1:219
 South America, 1:236, 1:241–242
 Southeast Asia, 4:693
 Late Quaternary
 Antarctica, 2:216–223
 Highland Asia, 2:236–244
 North America, 2:245–248
 southwest Pacific region, 2:202–215
 megafaunal extinctions, 4:700–704
 Middle-Pleistocene
 Eurasia, 2:172–179
 North America, 2:180–186
 Southern Hemisphere, 2:187–190
 mid-Quaternary North America, 2:180–186
 NHG, Pliocene environments, 3:185, 3:188, 3:189–191, 3:192
 periglacial
 fluvial sediments/forms, 3:490–499
 rock weathering, 3:500–506
 permafrost, 3:464–471
 features, 3:436–451, 3:452–463, 3:514–522
 rock glaciers, 3:535–541
 slope deposits/forms, 3:481–489
 Pleistocene, 1:3–4
 rock glaciers, 3:535–541
 sea-level change induced, 4:452–459
 theories, 1:4–6, 2:136–141
see also glacial landforms; late-glacial period
- glaciation threshold, definition, 1:909g
- glacier-fed lakes, 3:307–308, 3:308f
- glacier ice
 dating, 4:294–296
 microparticle sources, 2:343

- glaciers
 active temperate margins, 1:759
 beds, 1:755–756
 China, Tibetan Plateau, 2:379–386
 dating ice, 4:294–296
 East Africa, 2:373, 2:374–375, 2:374f, 2:377
 East Antarctic Ice Sheet, 2:457, 2:457f, 2:459–460
 extent, 1:884–894
 areal ice, 1:884–891
 consolidation measurements, 1:892
 controversial areas, 1:893
 cosmogenic dating, 1:892
 detailed maps, 1:884
 ice-sheet modeling, 1:892
 isostatic rebound, 1:892
 landmarks, 1:892
 past ice extent, 1:884
 relative–hemispheric contrasts, 2:275
 sea-level change, 1:892
 vertical extent, 1:892–893
 flow directions, 1:895–908
 cold-based ice, 1:898–899
 cross-cutting flow patterns, 1:896–902, 1:898f, 1:905f
 fans, 1:896–902, 1:899f, 1:900f
 flow-sets, 1:896–902, 1:897f, 1:903f, 1:905f
 glacial inversion, 1:895–896, 1:899, 1:901, 1:902
 ice streams, 1:896–902, 1:904f
 indicators, contemporary ice sheets, 1:902–904
 multitemporal evidence, 1:896
 paleo-ice sheets, 1:895–896
 paleo-ice streaming, 1:899–902, 1:904f
 pre-LGM ice flow, 1:898–899
 significance, 1:895
 lake sediment physical properties, 3:307–308, 3:308f
 mass balance
 growth/decay of ice sheets, 1:877–878
 reconstructions, 2:109
 microparticle/trace element studies, 2:342–346
 neoglaciation in Europe, 2:257–259, 2:259–260, 2:264–265
 oscillations, mid–late Holocene, 2:221
 permafrost interactions, 3:507–513
 polar-continental margins, 1:761
 polythermal margins, 1:759–761
 recession/retreat, 1:803–812
 Andes, 2:392–393
 estimated rates of, 1:809–810
 evidence of, 1:803–812
 geomorphology, 1:805–807
 glacial landforms, 1:803–812
 ice-marginal stagnation, 1:808–809
 implications of evidence, 1:810–811
 Late Holocene, 1:803–805
 Late Pleistocene, 1:807–808, 1:809–810
 Laurentide ice sheet, 1:808–809
 modern recession, 1:805–807
 South America, 2:387–394
 southwest Pacific region, 2:202–215
 temperate mountain biogenic matter, 2:289, 2:292–293
 thermal regime, 1:798–800
 tree destruction/modification, 2:104, 2:108–109
 variations, temporal pattern, 2:274–275
 West Antarctic Ice Sheet, 2:448, 2:449
 glaciifluvial landforms
 deposition, 2:6–17
 glaciifluvial systems, 2:6
 ice-contact landforms, 2:12–15
 links to other systems, 2:15
 proglacial landforms, 2:6–12, 2:8f, 2:14f
 erosion, 1:825–838
 glacial lake outbursts, 1:825–828, 1:828t
 ice-marginal meltwater channels, 1:832–833
 meltwater hypothesis, 1:833–836
 subglacial sheetfloods, 1:833–836
 tunnel valleys, 1:828–832, 1:833f, 1:834f, 1:835–836
 glaciifluvial stratigraphy, Europe, 2:260–262, 2:261f
 glacial deposits, sequence stratigraphy, 2:85–86, 2:89f
 glacial lithofacies, 2:18–29
 glacial origins, North American loess records, 2:620, 2:624–625
 glacial marine depositional system, 2:24–26, 2:27f
 glacial marine environment, definition, 2:30
 glacial marine sediments, 2:30–42
 high-latitude continental shelves, 2:33–35, 2:36f, 2:37f, 2:38f
 micromorphology, 2:54, 2:55f
 in Quaternary records, 2:30–35
 glaciochemical archive, ice cores, 2:342
 glaciochemistry
 background/measurement, 2:326–328
 ice cores, 2:326–333, 2:385
 spatial/temporal variability, 2:328–332
 glacioeustasy
 definition, 2:26–27
 deltas, influence on, 1:699
 sea-level changes, 4:439–451
 sequence stratigraphy, 4:263–264, 4:271
 glaciofluvial... *see* glaciifluvial...
 glaciogenic... *see* glacial...
 glacio-hydro-isostasy, sea-levels, 4:460, 4:462, 4:465
 glacioisostasy
 definition, 2:26–27
 sea-level change, 2:85
 glaciolacustrine, 2:43–51
 depocenters, 2:43, 2:46–50, 2:48f
 depositional system facies model, 2:25f
 processes, 2:43, 2:44f
 sediments, 2:43–46
 stratigraphy, 2:260–262, 2:261f
 glaciomarine... *see* glacial marine
 glaciotectionite, 2:68, 2:70f, 2:71–73
 see also glacial tectonic structures/landforms
 glaciovolcanic interactions, 1:780–781, 1:780ge
 glacial tectonic structures/landforms, 1:839–845
 composite ridges, 1:840–843
 cupola hills, 1:843–844
 glacial tectonic processes, 1:839
 hill-hole pairs, 1:843, 1:844f
 main types, 1:843f
 megablocks, 1:844–845
 rafts, 1:844–845
 thrust-block moraines, 1:840–843, 1:843f
 see also glaciotectionite
 glass
 chemical analysis, 4:540
 forensic science, 4:539, 4:540
 FT dating, 1:644, 1:644f, 1:646–649, 1:649–650, 1:650–651
 accuracy of ages, 1:650–651, 1:654t
 glass-DCFT, 1:650, 1:653f, 1:653t, 1:656
 glass-IIPFT, 1:650, 1:653–654, 1:653f, 1:653t, 1:655, 1:655t, 1:656, 1:659f
 population method, 1:647–649
 population-subtraction method, 1:649
 PTF correction methods, 1:649–650
 refractive index, 4:539
 tephrochronology, 4:284–285, 4:285f, 4:286f
 G-layer, tension wood, 2:97
 GLIMS *see* Global Land Ice Measurements from Space
 Global Charcoal Database (GCD), 2:716–717, 2:724–725
 global circulation model (GCM), stable isotopes, 2:349
 global climate changes, 3:215–217
 see also climate change
 global distribution, carbonate-precipitating lakes, 1:333
 global expansion
 Africa 300,000–8000 years ago, 1:91–97
 Americas 300,000–8,000 years ago, 1:119–134
 Asia 300,000–8000 years ago, 1:98–107
 Australia 300,000–8000 years ago, 1:108–118
 global extinctions, Late Pleistocene, 4:697
 see also extinctions
 global feedback mechanisms, climate change, 1:31–32
 Global Land Ice Measurements from Space (GLIMS), 1:804
 global patterns, vegetation 18,000/6000 years BP, 3:864–865
 Global Pollen Database (GPD), 3:831, 3:832–833
 global sea-level change, 4:371f, 4:372–375
 see also sea-level change
 global seawater, $\delta^{18}\text{O}$ distribution, 2:918
 see also seawater
 global sedimentation rates, Cenozoic cooling, 2:131–132
 Global Stratotype Sections and Points (GSSPs)
 chronostratigraphy, 4:215, 4:216–217
 stratigraphy overview, 4:191–192
 global warming
 African pollen records, 4:89–98
 ice core studies, 2:382, 2:385
 paleoclimate relevance, 3:244–252

- anthropogenic climate change, 3:244–248, 3:250
 climate modeling, 3:249–250
 climate processes, 3:248–249
 ecosystem dynamics, 3:250
 paleoclimate record, 3:244–245, 3:249–250
 research history, 1:17, 1:18
 THC, 1:745–746
see also climate change
- globally averaged temperatures, last interglacial, 3:160–162
- Globigerina bulloides* foraminifera, 2:745–746, 2:747f
- Gloeotrichia* macrofossils, 3:646, 3:647f
- GLOFs *see* glacial lake outburst floods
- Grovers Pond, New Jersey, 1:344–345, 1:347f
- glucose metabolism
 altered mechanisms, 1:317–318
 impaired, 1:315–321
- glycerol dialkyl glycerol tetraethers (GDGTs), 2:776–778, 2:776f
- glyptodonts, 4:684–685
- gnawing marks, carnivores, 4:749
- goats, 4:716
- Goddard Institute for Space Studies (GISS), NASA, 1:752, 1:752f
- Godwin, H, 2:700
- 'Golden Age', Late Neogene/Early Quaternary transition, 2:152f, 2:153f
- golden spikes *see* Global Stratotype Sections and Points
- gold-mining, Yedoma, 3:542–543
- Goldstream Formation, Beringia, 3:550
- gomphotheriidae, 4:689f, 4:690, 4:690f
- Gonyaulacales dinoflagellates, 2:804, 2:805f
 applications, 2:809f, 2:810f
 life cycle, 2:801f
- Goody, R, 3:147–148, 3:148f
- government records, climatology, 3:239–240
- GPC *see* gas proportional counting
- GPD *see* Global Pollen Database
- grab samplers, surface studies, 3:685, 3:685f
- graded streams
 definition, 1:685
 steady-state profiles, 1:685
- grain size
 deep-sea sediments, 1:638, 1:639, 1:640, 1:640f, 1:641, 1:641f
 ice-rafted debris, 2:35–41
 lake sediment, 3:305–306, 3:306f, 3:307–308, 3:307f
 last glacial cycle, South America, 2:635
- loess records
 China, 2:596–597
 Europe, 2:615, 2:616f
 South America, 2:635
 overwash sand layers, 3:210–211
 terrigenous sediments, 2:945, 2:951
- grain storage, 3:708–713
see also cereals
- grain thickness, loess, 2:596–597
- graminoid tundra, 3:746
- granular disintegration, blockfields, 3:525
- granular periglacial slope deposits, 3:486
- grasses/grasslands
 forest clearance impact, 1:276–277
 phytoliths, 3:582, 3:583f, 3:588–589, 3:589–590, 3:590f
 savanna grasslands, 4:13–14
 grassland beetles, 1:276–277
 gravel antidunes, 2:10
 gravel beach sea-level indicators, 4:388
 gravel dunes, 2:9
 graves
 remains, forensics, 4:546
 searches, forensic geology, 4:521–534
- Gravettian technocomplex, 1:149–150
- graviportal vertebrates, 4:680ge
- gravitational fractionation, firm air, 2:367–368
- gravitational slopes
 deformations, 3:556–558
 displacement, 3:556–558
- gravity corers, paleolimnology, 3:262–263, 3:262f, 3:263f
- grazers
 definition, 4:680ge
 South America, 4:685, 4:686, 4:689
- grazing land pollen records, 3:890
- Great Barrier Reef coral records, 4:416
- Great Basin, North America
 lake level studies, 2:539–540, 2:542f
 pollen records, 4:142–143, 4:147, 4:149f, 4:151
 rodent middens, 3:678f, 3:679–680
- Great Ice Age theory, 1:3–4
- Great Kobuk Sand Dunes, Alaska, 1:599, 1:599f
- Great Lakes, Africa, 2:501–502, 2:502f
- Great Plains, North America
 dune fields, 1:614–615, 1:615–616, 1:616–617, 1:617f, 1:620–621, 1:620f
 lake level studies, 2:537–539
 loess records, 2:620, 2:625–626, 2:627
- Great Salt Lake, Utah, 2:495f, 2:496
- Greece, pollen records
 Middle/Late Pleistocene, 4:63
 postglacial, 4:186–187
- greenhouse gases
 Arctic, 3:113, 3:116, 3:122–123
 astronomical theory, 2:140
 glacial cycles and, 2:315–316
 ice-core research, 2:280, 2:285f
 last interglacial, 3:155–156
see also carbon dioxide; methane; nitrous oxide
- Greenland
 accumulation map, 2:440f
 borehole temperature records, 2:299f, 2:300–301, 2:301f
 colonizing plants, 3:765–766
 East Greenland current, 2:157–159, 2:162–164
 heavy metal pollution, 2:344–345
- ice cores
 biological materials, 2:291–292
 chironomid records, 1:374, 1:378–379, 1:379f
 CO₂ studies, 2:311, 2:313, 2:314–315, 2:317
 correlation with Antarctic ice core, 2:410–415
 loess records, 2:616, 2:617, 2:617f
 microparticle fingerprinting, 2:343–344
 research/research history, 2:282f, 2:284, 2:285f, 2:435–438
 stable isotopes, 2:403–409
 stratigraphy overview, 4:201
 Younger Dryas, 3:126, 3:127f
- ice sheet, 2:435–438
 current mass balance, 2:446
 dynamics of, 2:439–447
 glacier recession, 1:804, 1:804f
 glaciochemical archive, 2:342
 growth/decay, 1:883
 history of, 2:445–446
 Late Neogene/Early Quaternary transition, 2:151, 2:155f, 2:161, 2:164
 Late Quaternary, 2:246f, 2:247, 2:249
 LSW, last interglacial, 3:163
 margin sites, 2:416–430
 properties, 2:319
 sea-level changes, 4:481–487
- insect fossil studies, 1:177, 1:181f
- methane studies, 2:334–335, 2:335–338, 2:337f, 2:338–339, 2:338f
- plant macrofossil records, 3:760–767
- sea-level changes
 Early to Late Quaternary, 4:482–483
 Eemian interglacial, 4:483
 Eocene to Pliocene, 4:481–482
 Holocene, 4:484–487
 ice sheet, 4:481–487
 interglacials, 4:481, 4:483
 last deglaciation, 4:483–484
 Late Quaternary, 4:481–488
 MIS 5e, 4:483
 neogacial/last millennium, 4:484–487
 RSL, 4:481, 4:483–484, 4:484–487
- stable isotopes, 2:348, 2:349, 2:351, 2:403–409
 deep ice cores, 2:405–408
 diffusion, 2:403–405
 early work, 2:403
 Eemian, 2:406f, 2:408
 Holocene, 2:405, 2:406–407, 2:406f, 2:407f
 ice core records, 2:403–409
 last glaciation, 2:407–408
 location of ice cores, 2:404f
 Oxygen-18, 2:403–405, 2:404f, 2:405–408, 2:405f, 2:405t, 2:406f, 2:407f, 2:408t
 shallow ice cores, 2:403–405
 water records, 2:420–424
- stadial
 GS 1, 1:205
 GS 2, 1:203–204
 sub-Milankovitch events, 3:200–202, 3:201f, 3:205–206, 3:207
 water stable isotope records, 2:420–424
- Greenland ice core projects (GRIP)
 climate change, 1:27–29, 1:28f
 correlation techniques, 2:410
 isotope thermometer, 2:350

- Greenland ice core projects (GRIP)
(Continued)
 research history, 2:436, 2:437
 stable isotopes, 2:404f, 2:405, 2:406f, 2:407, 2:408
see also Greenland, ice cores; North Greenland Icecore Project
- Greenland ice sheet project (GISP) research history, 2:436
- Greenland ice sheet project 2 (GISP2)
 biological materials, 2:291–292
 correlation techniques, 2:410, 2:411f
 firm air, 2:368
 isotope thermometer, 2:350
 postglacial North Atlantic, 3:56f, 3:57f
 research history, 2:436
 stable isotope records, 2:351f
 tephrochronology, 4:281–283
see also Greenland, ice sheet
- 'green' Sahara biosphere feedbacks, 1:723–724, 1:723f, 1:724f
- Gregory Lake, Australia, 2:525f, 2:526, 2:528
- Greifensee, Switzerland, 4:575f
- grêzes *litées* talus slopes, 3:570
- GRIP *see* Greenland ice core projects
- Groningen gas proportional counters, 4:313f, 4:314t
- groove erosion marks, 4:380, 4:380f
- ground beetles, 1:166f, 1:167f, 1:171f
 Japan, 1:211, 1:213f, 1:218
 North America, 1:233, 1:288f
 South America, 1:235, 1:236–237, 1:239f
- ground freezing, slope deposits/forms, 3:481, 3:483f
- ground ice, permafrost, 3:421–429, 3:467, 3:574–575
 frost mounds, 3:472
 rock glaciers, 3:535, 3:539
 slumps, 3:485
- ground searches, forensics, 4:525, 4:527–529
- ground sloths, 4:685–686, 4:716
- ground squirrels, 4:715
- ground surface temperature, permafrost, 3:465–466, 3:470
- grounding line
 definition, 2:216ge
 fans, 2:30, 2:32f, 2:33f
 retreat, Antarctica, 2:221
- grounding-zone wedges, 2:33, 2:36f
- groundwater lakes, Australia, 2:528
- growth curves, lichenometry, 2:567, 2:571
- growth/decay of ice sheets, 1:877–883
 atmospheric CO₂, 1:879–881
 ELAs, 1:877–878, 1:881–882
 elevation, 1:881
 feedback processes, 1:877, 1:878f, 1:881
 fluctuations, 1:879–881, 1:881–883
 marine-based, 1:878–879
 mass balance, 1:877–878, 1:879f, 1:881
 ocean circulation, 1:881, 1:881f
 process complexity, 1:881–883
 sea-level, 1:878–879
 solar radiation, 1:879
 surface albedo, 1:879
 terrestrial, 1:877–878
- growth increments, calcareous skeletons, 2:795, 2:798
- growth reactions, trees, 2:94–100
- growth release, trees, 2:94–95, 2:94f
- growth rings, roots, 2:100
see also tree-ring...
- growth suppression, trees, 2:94, 2:94f
- grus
 blockfields, 3:526, 3:526f
 rock weathering, 3:503–504, 3:503f
- GSSPs *see* Global Stratotype Sections and Points
- guanacos, 4:689
- Guangdong typhoon time series, 3:218–219, 3:220f
- Guineo-Congolian forests, 4:91–95
- Gulf air mass system, 4:142–143
- Gulf of Mexico
 benthic foraminifera, 2:769f
 deltas, 1:700, 1:701f, 2:123f
 fluvial sediments, 1:673f
- Gunung Dawung, Sunda Shelf, 4:661–662, 4:661t
- Gurbantunggut Desert, China, 1:606, 1:607f, 1:608f, 1:610f
- Gymnodiniales dinoflagellates, 2:804, 2:811f
- gymnosperms
 leaf stomata counting method, 3:615
 megafossils, 3:623–624
- Gyrinidae beetles, 1:211
- H**
- HA *see* hydroxyapatite
- habitat groups, oribatid mites, 2:683–684
- Hackney, England beetle assemblages, 1:188
- halobiont, definition, 1:361ge
- hammer stones, 1:83ge
- handaxes
 Acheulean Industry, 1:63, 1:64f
 assemblages, 1:83ge
 definition, 1:83ge, 1:83ge
 ideational realm, 1:89–90
 Western Europe post-MPT, 1:87
- Hang Hum Cave, Vietnam, 4:657, 4:658t
- Happisburgh sites, England beetle assemblages, 1:185
- haptophyte algae, 2:755–764
- hard hammers, definition, 1:83ge
- hard water
¹⁴C, plant macrofossils, 4:361ge
 lakes, 1:334
- hares, 4:715
- Harrington's mountain goat, 4:716
- harvesting methods, weed seeds, 3:714
- Hathnora, India, 1:75, 1:76f
- Haua Fteah site, 1:94, 1:95
- Haupt Knochenschicht (HK), Java, 4:653–654, 4:655t
- hay meadow pollen records, 3:890–891
- HBI *see* highly branched isoprenoid
- HCO₃ *see* bicarbonate
- ³He
 arid region dating, 1:437
 as TCN, 1:410–411
- head deposits, granular periglacial slopes, 3:486
- headwall collapse, cirque glaciers, 1:849
- heat flux adjustments, 3:148–149, 3:150f
- heat transfer, borehole temperatures, 2:298–299
- heat transport
 Late Neogene/Early Quaternary transition, 2:160–162
 oceanic, Pliocene climate, 3:186–187, 3:190f
- heave, cryostatic, 3:433–434
- heavy metal pollution, 2:344–345
- Heimsath AM cosmogenic nuclides, 1:442
- Heinrich events (HEs), 2:750–751
 benthic foraminifera, 2:768
 eustatic sea-level change, 4:432
 ice-rafted debris, 2:37, 2:40–41, 2:245–247
 Laurentide Ice Sheet, 2:247
 North Atlantic Ocean, 3:14–15
 paleoclimate reconstruction, 3:200–208
 THC, 1:744, 1:745f
- Heinrich layers
 diatom studies, 1:564–565
 ice-rafted debris, 2:35, 2:37, 2:39f, 2:40f
 internal characteristics, 2:37–40
 mechanisms/controls, 2:40–41, 2:41f
 Quaternary climatic change, 1:29
- Helan Mountains, China, 1:608, 1:608f, 1:610–611
- helium-3
 arid region dating, 1:437
 as TCN, 1:410–411
- helmeted muskox, 4:715
- hematite, 3:376
- hemipelagic mud, 1:637ge, 1:638
- hemispheric contrasts, relative glacier extent, 2:275
- hemispheric scale studies, paleoclimatology, 3:102–104, 3:110
- herbarium collections, DNA, 2:706
- herbs, Holocene North America, 3:770t, 3:773f
- hermatypic corals, 2:795, 2:796
- Herning Stadial (MIS 5d) beetle assemblages, 1:200
- herpetofauna, 4:602–604
- HEs *see* Heinrich events
- heterotrophic dinoflagellates, 2:800–801
- hiatuses in speleothems, 4:461, 4:464–465
- Hidden Cave, North America, 4:149–150, 4:150f, 4:151f
- Hidden Lake, Colorado, 2:541–543, 2:541f
- hierarchical structure
 lithostratigraphy, 4:227–228, 4:235–238
 pollen databases, 3:834
- high-energy coasts
 definition, 4:385
 sea-level change, 4:385–395
- high-frequency dynamics, lakes, 2:502–504
see also centennial-scale dynamics
- Highland Asia, 2:236–244
 Altai Mountains, 2:240–241, 2:242f
 ELA, 2:238–239, 2:241
 Himalayas/Trans Himalayas, 2:237–240, 2:238f

- Late Quaternary glaciations, 2:236–244
 Pamir, Tajikistan, 2:241
 Tibet, 2:238f, 2:240, 2:242f
 Tien Shan, 2:240–241
 highlands of Mexico, lake level studies, 2:533
 high latitudes
 continental shelves, 2:32f, 2:33–35, 2:36f, 2:37f, 2:38f
 dune fields, 1:597–605
 age determination, 1:602–604
 definition, 1:597ge
 future studies, 1:604
 origins, 1:601–602
 paleoenvironmental significance, 1:604
 primary, 1:598–599
 types, 1:599–601
 fjords, 1:559
 hard-water lakes, 1:334
 relative sea levels, 4:467–480
 High lodge site, England beetle assemblages, 1:185
 highly branched isoprenoid (HBI)
 alkene diatom biomarker, 1:590–591, 1:591f–595, 1:591f
 sea ice proxy, 2:779–780
 see also IP₂₅ HBI alkene
 high-performance liquid chromatography (HPLC), 3:328–329, 3:330
 high polar ice cap margin sites, 2:416
 high-pressure liquid chromatography (HPLC), 1:44, 1:44f
 high-resolution diatom records, 1:541–544
 highstands, sea-levels
 cave data, 4:460, 4:461, 4:463f, 4:464
 coral records, 4:409–411, 4:410f
 eustatic change, 4:434–435
 Holocene, 4:447, 4:499–501, 4:501t
 last interglacial, 4:498
 hill-hole pairs, glaciectonics, 1:843, 1:844f
 hillslopes
 ice wedges, 3:468
 permafrost, 3:468, 3:469
 Himalayas
 ice core studies, 2:379, 2:382f, 2:383–384
 Late Quaternary, 2:237–240
 hippopotamuses, 4:736–737
 Histeridae (clown beetles), 1:212, 1:215f, 1:218, 4:552–553
 historical aspects
 biogeographical vegetation reconstruction, 3:866–867
 botanical evidence justification, 4:542
 climatology, 3:237–243
 diaries/journals, 3:238, 3:239, 3:240f
 documents
 potential/limitations, 3:237–238
 sources, 3:238
 types, 3:238–241
 future research, 3:242
 gazettes, 3:240–241, 3:241f
 government records, 3:239–240
 newspapers, 3:240–241
 official records, 3:239–240
 research methods, 3:241–242
 ships' logbooks, 3:237–238, 3:239f, 3:239t
 dating methods, 1:6–7, 1:9–16
 glaciations
 Africa, 2:149
 Antarctica, 2:148
 Australasia, 2:148–149
 Cenozoic, 2:146f, 2:147t, 2:148t, 2:149t
 Central Asia, 2:148
 distribution of, 2:145f
 Early Quaternary, 2:151–166, 2:167–171
 Eurasia, 2:224–235
 Europe, 2:144, 2:147t, 2:257–268
 Highland Asia, 2:236–244
 Late Neogene/Early Quaternary transition, 2:151–166
 Late Pleistocene, 2:224–235, 2:250–256
 Late Quaternary, 2:236–244
 North America, 2:144–148, 2:148t
 overview, 2:143–150
 Pleistocene, 1:3–4
 South America, 2:148, 2:250–256
 theories, 1:4–6
 Quaternary science, 1:1–8
 dating methods, 1:6–7
 geologic framework, 1:1–2
 mammals, 1:2–3
 nineteenth century, 1:1
 science research social relevance, 1:17–25
 Hjulström-type deltas, 2:48f, 2:50
 HK (Haupt Knochenschicht), Java, 4:653–654, 4:655t
 Hofmann, Wolfgang, 1:373–374
 Holebuddalen Lake, Norway, 1:389, 1:392f
 Holocene, 3:87–92
 Antarctic Peninsula, 2:219–220
 Arctic paleoclimate, 3:118–121, 3:122–123
 HTM, 3:118–120
 neoglaciation, 3:120
 past 2ka, 3:120–121
 atmospheric carbon-14, 4:329–331
 Australia
 human colonization, 1:112f, 1:116f
 postglacial pollen records, 4:106–109
 beetle records, 1:164f, 1:282–285
 biosphere feedbacks, 1:723–725
 charred particle analyses, 2:720, 2:724, 2:725
 chironomid records
 Europe, 1:374, 1:377f, 1:378f, 1:381f, 1:382f
 South America, 1:398, 1:399, 1:401, 1:403–404
 chronostratigraphy, 4:218
 climate forcing responses, 1:681–682
 climatic optimum
 Antarctic Peninsula, 2:220
 Victoria Land, Antarctica, 2:221
 climatic variability
 Andean glaciers, 2:390–391, 2:392
 flood response, 3:255–256
 CO₂
 ice core studies, 2:312–313, 2:313–314
 reconstruction, fossil leaves, 3:617–618, 3:617f, 3:619
 COHMAP, 3:136–137, 3:249, 3:250
 coral records, RSL changes, 4:413–416
 definition, 2:216ge
 deglaciation
 Pacific Ocean, 1:574–575
 Wisconsin–Holocene transition, 2:301
 dendroglaciology, 2:104, 2:109
 diatom records
 North Atlantic/Nordic Seas, 1:567–569
 Pacific Ocean, 1:574–575
 salinity/climate reconstructions, 1:510–512
 East Antarctica climate development, 2:220–221
 erosion rates, glacial landforms, 1:859–860
 ESL changes, 4:447–449
 fluvial paleohydrology, 3:253–255, 3:255–256
 fluvial sediments
 alluvial rivers, 1:672f
 Colorado river, 1:673f
 Mississippi River, 1:668f, 1:669f
 glacier oscillations, 2:221
 Greenland
 climate change, 3:761
 sea-level changes, 4:484–487
 stable isotopes, 2:405, 2:406–407, 2:406f, 2:407f
 'green' Sahara, 1:723–724, 1:723f, 1:724f
 historical Quaternary science, 1:1
 ice-core research, 2:284, 2:312–313, 2:313–314, 2:375, 2:376–377
 insect fossil studies
 Europe, 1:274–281
 South America, 1:235
 Kilimanjaro ice core records, 2:375, 2:376–377
 lake levels
 Latin America, 2:532–533, 2:534
 modeling
 climate change effects, 2:558
 simulation analysis, 2:560–561, 2:562f
 solar radiation, 2:561, 2:561t
 reconstructions, 2:485f, 2:487
 west-central Europe, 2:549–557
 land-cover changes, 1:725, 1:726f
 last glacial epoch transition to, 2:313–314
 LP/EH extinctions, 4:691
 macrofossil records
 mire/peat, 3:637–656
 North America, 3:756–757, 3:768–784
 megafossils, 3:623f, 3:625–626, 3:628f
 methane studies, 2:335–338, 2:337f
 microfossil-based reconstruction, RSL, 4:419–428
 mire/peat macrofossils, 3:637–656
 neoglaciation
 Arctic, 3:120
 Europe, 2:257, 2:258f, 2:259, 2:261f, 2:262, 2:262f, 2:263f, 2:264, 2:264f
 New Zealand pollen records, 4:109–112
 nitrous oxide history, 2:472, 2:473–474, 2:473f

Holocene (*Continued*)

North America

- Arctic macrofossil records, 3:756–757
- lake level studies, 2:538–539, 2:540, 2:541–543, 2:544–545, 2:545–546
- loess records, 2:625–626, 2:626f
- Mid-Pleistocene vertebrate records, 4:649
- plant macrofossil records, 3:768–784
- pollen records, 4:47–49
- northern treeline dynamics, 1:724–725, 1:725f
- NWNA pollen records, 4:125–126, 4:126–127, 4:128–129, 4:129–130
- optimum, China, 1:610–611, 1:612f
- paleolimnology, 3:658, 3:662f, 3:672t
- paleophytogeography, 2:731–732
- paraglacial geomorphology, 3:553, 3:563
- plant distributions, 3:855–856, 3:857–858, 3:858f
- plant macrofossil records
 - Arctic Eurasia, 3:741–742
 - North America, 3:768–784
- Pleistocene, transition from
 - beetle fossil record, 1:282–285
 - fluvial environment responses, 1:679–680
- pollen records, 3:763–765
 - Alaska and Yukon, 4:40, 4:43f
 - Australasia, 4:18–19
 - Australia, 4:106–109
 - Canada, 4:45f, 4:46f, 4:47
 - continuous studies, 3:813–814
 - New Zealand, 4:109–112
 - North America, 4:47–49
 - northwestern, 4:125–126, 4:126–127, 4:128–129, 4:129–130
 - southwestern, 4:142–155
- postglacial
 - Australia, 4:106–109
 - New Zealand pollen records, 4:109–112
 - North American pollen records, 4:137–140, 4:139f
 - North Atlantic, 3:55–56, 3:56f, 3:59, 3:59f, 3:60, 3:61
 - Northern European pollen records, 4:173–174
 - South America, 1:398, 1:399, 1:401, 1:403–404
 - South Pacific ocean, 3:75f, 3:76, 3:77f, 3:80f, 3:82, 3:83f, 3:84f
- PSV records, 1:715f
- QUI RSL, 4:471, 4:474–477
- radioisotope proxies, 2:929f
- rodent middens, 3:679t
- Ross Sea environmental development, 2:221
- salinity/climate reconstructions, 1:510–512
- sea-level changes, 4:504f, 4:515–518
- coral records, 4:413–416
- ESL, 4:444–449, 4:449–450
- Greenland, 4:484–487
- microfossil-based reconstructions, 4:419–428
- QUI RSL, 4:471, 4:474–477

- sea-level highstand, 4:447, 4:499–501, 4:501t
- sea-surface temperature development, 1:568f
- sediment budgets, 3:253–255, 3:254f
- South Pacific ocean records, 3:75f, 3:76, 3:77f, 3:80f, 3:82, 3:83f, 3:84f
- southwest Pacific region, 2:202, 2:203t
- speleothem records, 1:299
- stable isotopes
 - Antarctica, 2:396–398
 - Greenland, 2:405, 2:406–407, 2:406f, 2:407f
- strandlines, 4:503–504, 4:512f, 4:515–518
- treeline dynamics, 1:724–725, 1:725f, 3:693–695
- vegetation
 - development, 3:763–765
 - patterns, 3:864–865
- Victoria Land, Antarctica
 - climatic optimum, 2:221
 - environmental development, 2:221
- volcanic/solar forcing, 1:750
- Wisconsin–Holocene deglacial transition, 2:301
- Younger Dryas oscillation, 3:226–227
- see also* Early Holocene; Late Holocene; Mid-Holocene
- Holocene Series chronostratigraphy, 4:218
- Holocene Thermal Maximum (HTM)
 - Arctic paleoclimate, 3:118–120
 - North American pollen records, 4:47–49
- holomictic lakes, 1:361ge
- holostratotypes, 4:229–232
- Holsteinian stage
 - mammal assemblages, 4:612–613
 - Middle-Pleistocene glaciation, 2:173–174
- homicide cases
 - burial location searches, 4:525, 4:526t, 4:527f
 - plant remains, 4:543, 4:544
 - soil/sediment as evidence, 4:535–537
- hominids
 - adaptive radiation, 1:52t
 - definition, 1:50t
 - evolution, 4:298
 - interaction studies, 4:748–755
 - overview, 1:49–58
 - teeth/bones, 1:310, 1:311f
 - see also* Homo...; human...
- hominins
 - adaptive radiation, 1:53f
 - behavior
 - Africa
 - 2.7 MYR–300,000 years ago, 1:59–66
 - 300,000–8000 years ago, 1:91–97
 - Asia
 - 2.7 MYR–300,000 years ago, 1:67–82
 - 300,000–8000 years ago, 1:98–107
 - Australia 300,000–8000 years ago, 1:108–118
 - Europe
 - 1.9 MYR–300,000 years ago, 1:83–90
 - Neanderthal demise, 1:146–153
 - definition, 1:50t, 1:83ge

- dwarfing/gigantism, 4:741–742, 4:742–743, 4:743f, 4:744f
- interaction studies, 4:596
- overview, 1:49–58
- see also* human...
- Homo* genus fossils, Africa, 4:615
- Homo antecessor*
 - cognition, 1:89
 - human evolution, 1:140
 - MPT, 1:86–87
 - subsistence behavior, 1:88
- Homo erectus*, 1:135–136, 1:137–139, 1:142
 - Africa, 1:62, 1:63f, 1:64, 1:65
 - Asia, 1:67, 1:69t, 1:70
 - lithic technology, 1:67–68
 - Maliang site, 1:76–77
 - modern humans replacing, 1:100–103
 - speciation events/isolation, 1:80
- human evolution, 1:137–139
- Java, 4:694–695
- key features, 1:137t
- modern humans replacing, 1:100–103
- Sahul landmass, 1:108
- stone tools, 1:62, 1:63f, 1:64, 1:65
- Homo ergaster*
 - African stone tools, 1:62–63, 1:63f, 1:64
 - Europe pre-MPT, 1:84
 - human evolution, 1:138
- Homo floresiensis*
 - dwarfing/gigantism, 4:742–743, 4:743f, 4:744f
 - human evolution, 1:142–143
 - modern humans replacing, 1:100–103
 - Sahul landmass, 1:108, 1:112–113
- Homo georgicus*
 - Europe pre-MPT, 1:84
 - human evolution, 1:138–139
- Homo habilis*, 1:135–137, 1:138
- Homo heidelbergensis*
 - African stone tools, 1:64
 - Asia, 1:80
 - Europe
 - cognition, 1:89
 - post-MPT, 1:87
 - subsistence behavior, 1:88
 - human evolution, 1:139–140
- Homo neanderthalensis* *see* Neanderthals
- Homo rudolfensis*, 1:135–136, 1:136–137
- Homo sapiens*, 1:141–143
 - cranial features, 1:141–142
 - geographical origin models, 1:91
 - human evolution, 1:141–143
 - Neanderthals and, 1:140–141, 1:142
 - research history, 1:24
 - Southeast Asia, 4:693, 4:695–696
 - see also* anthropomorphic...; human...; modern humans
- honeycomb erosion marks, 4:379
- hop hornbeam (*Ostrya*), 4:116f, 4:119
- horizon designations, paleosols, 3:367
- horizon development, soils, 3:402, 3:403f
- hornbeam (*Carpinus/Ostrya*), 4:1–2, 4:7, 4:116f, 4:119
- horses
 - cave art depictions, 4:752–754

- mummified, 4:714
 South American Late Pleistocene, 4:688
 teeth/bones, 1:306f
see also equids
 Hoshantung cave, China, 4:655–657
 Hoton–Nur Lake, Mongolia, 4:170, 4:170f
 house flies, 4:550–551
 Howard Valley, New Zealand, 1:252–253
 Hoxne, England beetle assemblages, 1:187
 HPLC *see* high-performance liquid chromatography; high-pressure liquid chromatography
 Hsing-an cave, China, 4:655–657
 HTM *see* Holocene Thermal Maximum
 Huascarán site, Andes, 2:387, 2:390, 2:391f
 Huelmo, Chile, 1:402, 1:402f
 Hugin So Interstade plant macrofossils, 3:763
 Hulu Cave, China, 1:298, 1:300f
 human, definition, 1:50t
 human adaptations
 dendroarcheology, 3:630, 3:633–634, 3:634–636
 postglacial, 1:154–159
 human colonization
 Americas, 1:119–134
 Australia 300,000–8000 years ago, 1:108–118
 chronology establishment, 1:122–125
 Europe, 1:84–87
 MPT, 1:84–87
 post-MPT, 1:87
 pre-MPT, 1:84
 see also human settlements
 human evolution, 1:135–145
 Africa
 archeological records, 1:91–97
 vertebrate studies, 4:664
 Asia 300,000–8000 years ago, 1:98–107
 Dmanisi remains, 1:138–139
 emergence of humans, 1:135–137
 European Neanderthal demise, 1:146–153
 first expansions, 1:137–139
 Homo erectus and, 1:135–136, 1:137–139, 1:137t, 1:142
 Homo sapiens and, 1:140–141, 1:141–143
 NHG, Pliocene environments, 3:192
 overview, 1:49–58
 skulls, 1:138f
 transitional hominins, 1:139–141
 see also hominids; hominins
 human impact
 Africa 2.7 MYR–300,000 years ago, 1:59–66
 Asia 2.7 MYR–300,000 years ago, 1:67–82
 atmospheric dust, 1:735
 beetle fauna, 1:274, 1:276–277, 1:278
 biogeography, 3:317–318, 3:318–319, 3:319–320, 3:321–322, 3:322–323
 chironomid studies, 1:355–356, 1:394–395
 environmental effects
 Americas, 1:126–127
 Australia, 1:113–115
 Europe
 1.9 MYR–300,000 years ago, 1:83–90
 beetle fauna, 1:274, 1:276–277, 1:278
 pollen records, 4:176–177
 global expansion, 1:119–134
 ice cores, 2:313, 2:332
 lake ecosystems
 eutrophication, 3:659
 hydroseres, 3:657
 lake sediment diatom records, 1:541–543
 megafaunal extinctions, 4:711
 paleohydrology, 3:253
 postglacial Europe
 beetle fauna, 1:274, 1:276–277, 1:278
 pollen records, 4:176–177
 Quaternary climate change, 1:33–34
 reconstructions, pollen data, 3:881–901
 vegetation, 3:880–903
 Australasia, 4:24–25
 environmental archeology, 3:699–724
 see also anthropogenic...; biomass burning
 human interactions, vertebrate studies, 4:748–755
 human revolution concept, 1:57
 human settlements
 lakes/river, 1:158
 prehistoric, 2:549–551, 3:276
 see also human colonization; settlement
 humid climates
 African pollen records, 4:88–89, 4:88f, 4:98–100, 4:100f
 soil morphology, 3:363f, 3:365
 hummocks
 glacier recession, 1:809, 1:810f
 patterned ground, 3:460–462
 hummocky moraines, 1:774–775, 3:509
 hummocky terrain, meltwater hypothesis, 1:835–836
 Hunsgi, India, 1:75, 1:76f, 1:77f
 hunter-gatherers
 plant remains, 3:705–706, 3:712f
 Upper Paleolithic, 4:750–751
 hunting, hominids, 4:748–751, 4:752–754
 Huon Peninsula, Papua New Guinea
 coral records, RSL changes, 4:409, 4:411, 4:411f, 4:412–413
 eustatic sea-level change, 4:430f, 4:432–433, 4:433f, 4:436
 late glacial sea-level change, 4:441–442
 Huron River, Michigan, 1:343–346, 1:345f
 hurricanes
 Atlantic, 3:215–217
 environmental impact diagram, 3:210f
 paleotempestology, 3:209–221
 hydration, weathering, 3:394, 3:501
 hydration shattering, 3:501
 hydrochemistry in forensics, 4:530
 hydrofractures, rock weathering, 3:500
 hydrogen isotopes
 beetle records, 1:169–170
 concepts, 1:291
 ice cores, 2:347
 sclerochronology, 3:289
 hydrogen pyrolysis (hypy) charcoal
 pretreatment, 4:357
 hydrography
 reconstructions, dinocysts, 2:812
 sub-Milankovitch events, 3:204–205, 3:206–207
 surface water records, North Pacific, 3:37–42
 hydrological cycle
 lake water oxygen isotope relation, 1:336–337
 marine oxygen isotope stratigraphy, 2:908f
 seawater oxygen isotope ratios, 2:915–916, 2:916f
 hydrology
 changes
 contrasting patterns of, 2:551–552, 2:553f
 pigment studies, 3:335
 lake water oxygen isotope relation, 1:336–337
 paleohydrology, 3:253–258
 periglacial rivers, 3:491–492
 terrestrial systems, 2:737–738
 hydrolysis, weathering profiles, 3:394
 Hydrophilidae beetles, 1:211, 1:217f
 hydrosere distribution, lakes, 3:657
 hydroxyapatite (HA), 1:292
Hydrozetes, 2:681–682, 2:681f, 2:682f, 2:685f, 2:689f, 2:690f, 2:691, 2:692f, 2:694f, 2:695
 hyperdisease theory, megafaunal extinctions, 4:711
 hypersalin, definition, 1:361ge
 hypocoatings, soil, 3:383, 3:385f
 hypolimnetic anoxia, lakes, 3:294–295
 hypolimnion region, definition, 1:361ge
 hypothalamo-pituitary-adrenal (HPA) system, 1:319
 Hypsithermal interval neoglaciation, 2:269
 hypsodonts
 proboscideans, 4:734
 rodent evolution, 4:725–727
 hypy (hydrogen pyrolysis) charcoal
 pretreatment, 4:357
 hyrax middens, 4:15
 hysteresis curves, THC, 1:742–743, 1:743f
 hysteresis loops, millennial climate change, 3:207
- I**
 Iberia/Iberian Peninsula
 Neanderthals in, 1:151–152
 postglacial pollen records, 4:179–183, 4:180t
 Iberian model, American human colonization, 1:119
 IC *see* ion chromatography
 ice..., *see also* sea ice...
 Ice Ages
 European, 1:4, 1:5f
 Last Glacial Maximum GCMs, 3:167
 paleophytogeography, 2:730–731
 pollen records, 4:66–67
 theory
 paleoclimates, 2:136
 Pleistocene epoch, 1:3–4
 ice augen, 3:510, 3:511f
 icebergs
 glacimarine sediments, 2:30, 2:32f
 ice-rafted debris distinction, 2:37, 2:39f

- ice-block obstacle marks, proglacial landforms, 2:10
- ice caps
 Antarctic Peninsula, 2:216
 Barnes ice cap, 2:424
 high polar margin sites, 2:416
 microparticle/trace element studies, 2:342–346
 volcanic landforms, 1:781
- ice casts, 3:436–451
see also ice wedges
- ice centers, North America, 2:185
- Ice Complex *see* Yedoma...
- ice-contact landforms, 2:12–15
 eskers, 2:13–14, 2:14f
 kames, 2:15
 kame terraces, 2:13
 linear features, 2:13–14
 nonlinear features, 2:14–15
 subaerial outwash fans, 2:14–15, 2:14f
see also glaciifluvial landforms, deposition
- ice cores
 abrupt climate events, 2:284, 2:308
 Africa, 2:373–378
 equatorial East Africa, 2:373, 2:374f
 glacier retreat, 2:377
 Kilimanjaro records, 2:373, 2:374f, 2:375–377
- air
 age of occluded air, 2:309
 extraction methods, 2:468–469
 isotopic composition, 2:431
 Andes, drilling in, 2:387, 2:388f
 anthropogenic impact, 2:281, 2:284–286, 3:246–248, 3:248f
 Arctic diatom biomarkers, 1:589f
 basics, 2:278
¹⁰Be in, 2:353–360
 biogenic gases, 2:280
 biological materials, 2:283, 2:288–297
 borehole temperatures, 2:281, 2:298–302
 carbon monoxide, 2:463–470
 chemistry of, 2:319–320
 China, Tibetan mountains, 2:379–386
 chironomids, Europe, 1:374, 1:378–379, 1:379f
 chronologies, 2:278–279, 2:303–310
 abrupt climate events, 2:308
 age horizons/synchronization, 2:306–309
 extraterrestrial object impact, 2:309
 fission products, 2:305t, 2:306
 flow models, 2:303–305, 2:304f
 geomagnetic field, 2:308
 layer counting, 2:303, 2:304f
 Milankovitch cycles, 2:307–308
 occluded air age, 2:309
 optimized combined modeling, 2:305
 past accumulation rates, 2:305
 radioactive dating, 2:305–306, 2:305t
 Raymond bumps, 2:305, 2:305f
 research, 2:278–279
 rheology of ice, 2:303–304
 solar activity, 2:308–309
 volcanic eruptions, 2:307, 2:307f
- CO₂ studies, 2:311–318
- conductivity studies, 2:319–325
- correlation techniques
 cosmogenic radionuclides, 2:357–358
 Greenland/Antarctica, 2:410–415
- cosmogenic isotopes, 2:283
- cosmogenic radionuclides, 2:353–360
- dating
 conductivity studies, 2:323–324
 radioactive dating, 2:305–306, 2:305t
- dust, atmospheric, 1:730–731
- early projects, 2:435–436
- East Antarctic Ice Sheet, 2:456–462
- electrical properties, 2:319–320
- extraterrestrial material, 2:283, 2:309
- firm air, 2:361–372
- future research, 2:286, 2:437–438, 2:469
- gas measurements, 2:463–470
- glacial-interglacial cycles, 2:283
- glaciochemistry, 2:281, 2:326–333
- greenhouse gases, 2:280, 2:285f
- Greenland
 biological material, 2:289–291
 correlation techniques, 2:410–415
 ice sheet, 2:439–447
 stable isotopes, 2:403–409
 Younger Dryas, 3:126, 3:127f
- Holocene, 2:284
- insoluble particles, 2:281–283
- isotopic stratigraphy, 2:351
- last ice age, 2:284
- major highlights, 2:283–286
- margin ice, 2:416–430
- measurements, 2:279–283, 2:463–470
- methane studies, 2:334–341
- microparticles, 2:342–346
- NGRIP project, 1:27–29, 1:28f
- nitrous oxide history, 2:471–476
- noble gases, 2:280–281
- overview, 2:277–287
- polar climatic change, 1:26–27, 1:27–29, 1:29–31
- recent research, 2:284–286, 2:436–437
- research history, 2:435–438
- South America, 2:387–394
- stable isotopes, 2:279–280, 2:347–352, 2:395–402
- stratigraphy overview, 4:201
- thermal diffusion paleotemperature records, 2:431–434
- trace elements, 2:342–346
- ultra-trace gases, 2:463–470
- visible/visual stratigraphy, 2:279, 2:443–444, 2:444f
- Vostok biological material, 2:289–291
- West Antarctic Ice Sheet, 2:448–455
see also coring; drilling projects
- ice coverage
 Australia before LGM, 4:104–106
 rivers, 3:494
 sea ice
 Antarctica, 1:530–531, 1:532t, 1:533–535, 1:533f, 1:534f
 Arctic diatom biomarkers, 1:588–596
 volcanic landforms, 1:781
- ice crystals
 formation process, 2:319
 Greenland ice sheet, 2:442–443
- ice dams, 3:494
- ice deformation
 East Antarctic Ice Sheet, 2:456–457
 Greenland ice sheet, 2:441–443
- ice discharge into ocean, Greenland, 2:441
- ice divide, 2:442f
- ice dynamics
 blockfields, 3:530–532
 East Antarctic Ice Sheet, 2:456–462
 Greenland ice sheet, 2:439–447
 West Antarctic Ice Sheet, 2:448–455
- ice-equivalent sea-level *see* eustatic sea-level
- ice fields
 Andes, 2:393
 Kibo, Tanzania, 2:373, 2:374–375, 2:374f, 2:377f
 landsystems, 1:814–817, 1:819f, 1:822
- ice flows, Greenland, 2:439, 2:441–443
- ice-free coastal areas, Antarctic Peninsula, 2:219–220
- ice geometry, blockfields, 3:530–532
- icehouse times, drifting continents, 2:132–134
- Iceland
 Arctic paleoclimate history, 3:114
 Early Quaternary, 2:169
 mafic tuya, 1:791f
 sheet-like sequences, 1:797f
 tephra mounds/ridges, 1:795–796
- ice lenses, 3:467, 3:575
- Iceman finds, plant remains, 3:702, 3:709f
- ice-margins, 2:416–430
 apron forms, 1:769–770, 1:773f
- glaciers
 extent, 1:888–890
 recession, 1:808–809
- landforms
 erosion, 1:832–833
 permafrost-glacier interactions, 3:508–509
 meltwater channels, 1:832–833
 stagnation, 1:808–809
- ice models, lake-levels, 2:559
- ice-proximal settings, depositional architecture, 2:30–33
- ice-rafted debris/detritus (IRD), 2:30–42
 Arctic paleoclimate history, 3:114
 deep-sea case study, 2:37, 2:39f, 2:40f
 glaciolacustrine, 2:45, 2:46
 grain size, 2:35–41
 Late Neogene/Early Quaternary transition, 2:157–159, 2:161–162
 magnetic susceptibility, 2:889, 2:891–893, 2:893–895, 2:893f
 North America, 2:245–247
 North Atlantic Ocean, 3:9–10, 3:11f
 petrography, 2:948, 2:949f
 Pliocene climate warmth, 3:188, 3:190f
 South Atlantic Late Pleistocene, 3:24, 3:27f
 sub-Milankovitch events, 3:200, 3:202–203, 3:205f

- ice-rich zones, permafrost, 3:421–429, 3:574–575
- ice-scour lakes, 1:546
- ice segregation
- active layer, permafrost, 3:423–424
 - ground ice, 3:467, 3:472
 - rock weathering, 3:500–501, 3:501f, 3:504f
- ice sheets
- Antarctica, 2:435–438
 - areal ice extent, 1:891
 - diatom studies, 1:535–536
 - dynamics of, 2:448–455, 2:456–462
 - glacier recession, 1:804
 - glaciochemical archive, 2:342
 - Late Quaternary, 2:216–217, 2:218
 - margin ice, 2:416–430
 - properties, 2:319
 - biogenic matter, 2:289
 - borehole temperature records, 2:301
 - continental, 2:128f, 2:908–909
 - cross-section, 2:417f
 - cryosphere history, 1:20
 - decay/growth, 1:877–883
 - dune activity, 1:603, 1:603f
 - durations of, 2:128f
 - dynamics
 - Antarctica, 2:448–455, 2:456–462
 - definition, 2:456
 - East Antarctic Ice Sheet, 2:456–462
 - Greenland, 2:439–447
 - Eurasian glaciations, 2:224–232
 - extent of, 1:884–894
 - glacier flow directions, 1:895–908
 - glacier recession, 1:804, 1:804f, 1:807, 1:808–809
 - Greenland, 2:435–438
 - current mass balance, 2:446
 - dynamics of, 2:439–447
 - glacier recession, 1:804, 1:804f
 - glaciochemical archive, 2:342
 - growth/decay, 1:883
 - history of, 2:445–446
 - Late Neogene/Early Quaternary transition, 2:151, 2:155f, 2:161, 2:164
 - Late Quaternary, 2:246f, 2:247, 2:249
 - LSW, last interglacial, 3:163
 - margin ice, 2:416–430
 - properties, 2:319
 - sea-level changes, 4:481–487
 - growth/decay, 1:877–883
 - instability, 1:30
 - local to global sea-level change, 4:373
 - marginal associations, 1:759–761
 - microparticle/trace element studies, 2:342–346
 - millennial climate change modeling, 3:206–207
 - North America
 - Late Quaternary, 2:245–248
 - Middle-Pleistocene, 2:184f
 - paleo-ELAs, 1:909–917
 - Patagonia, 1:236f, 1:241
 - QEI RSL, 4:469–470, 4:470–471, 4:470f, 4:471–474, 4:474f
 - snowfall crystal formation, 2:319
 - subglacial landforms, 1:755–756, 1:756f, 1:758–759
 - trimlines/paleonunataks, 1:918–928
 - Younger Dryas, 3:130–131
 - see also* Greenland ice sheet project 2; Innuitian ice sheet; isostasy; Laurentide Ice Sheet
- ice shelves
- East Antarctic Ice Sheet, 2:460, 2:460f
 - West Antarctic Ice Sheet, 2:448
- ice slumps, permafrost, 3:485
- ice streams
- East Antarctic Ice Sheet, 2:457, 2:457f
 - glacier flow directions, 1:896–902, 1:904f
 - Greenland ice sheet, 2:441
 - landsystems, 1:816f, 1:817, 1:821f, 1:822
 - West Antarctic Ice Sheet, 2:449, 2:450–451, 2:452–453
- ice surge events, 3:205f
- ice veins, 3:436–439
- see also* veins
- ice volume
- Arctic paleoclimate history, 3:116
 - Early Pleistocene, 3:3
 - glacial, 2:917–918
- ice wedges, 3:436–451
- anti-syngenetic, 3:438–439
 - casts, 3:496–497, 3:497f
 - Cenozoic age pseudomorphs, 3:443–446, 3:448f
 - classification, 3:437f
 - composite wedges, 3:440, 3:443–444, 3:443f, 3:446, 3:447t, 3:450f
 - development of, 3:440
 - epigenetic wedges, 3:438, 3:438f, 3:439f, 3:443f
 - European loess records, 2:613, 2:614f
 - paleoenvironmental significance, 3:447–449
 - periglacial features, 3:436–451, 3:496–497, 3:497f
 - permafrost, 3:426, 3:467–468, 3:468f, 3:574–575, 3:575f
 - polygons, 3:458–459
 - primary sand wedges, 3:448–449
 - pseudomorphs, 3:436, 3:443–446, 3:444f, 3:446f, 3:447–448, 3:447t, 3:448f, 3:450f
 - rejuvenated wedges, 3:438f, 3:439
 - relict wedges, 3:447, 3:448–449
 - selective preservation, 3:449
 - soil veins/wedges, 3:440, 3:443–446, 3:449
 - syngenetic, 3:438, 3:438f, 3:439f, 3:543f, 3:545f, 3:547f
 - terminology, 3:437t
 - thermal contraction cracking, 3:436–440, 3:440–446, 3:441f
 - veins, 3:436–439
- icings, rivers, 3:494
- ICP *see* inductively coupled plasma
- ideational realm, hominins, 1:89–90
- ideological relations, hominids–animals, 4:748, 4:752–754
- IGCP *see* International Geological Correlation Programme
- iGFD *see* inert gas flow dehydration
- iHTR *see* inductive high-temperature carbon reduction
- IKM *see* Imbrie–Kipp Transfer Function Method
- Illinoian MIS glaciation, 2:180
- image analysis, varved sediments, 4:584–585
- Imbrie–Kipp techniques
- quantitative paleoclimate reconstructions, 3:181, 3:182
 - temperature proxies, 2:841, 2:844–845, 2:846–847
- Imbrie–Kipp Transfer Function Method (IKM), 2:841, 2:844–845, 2:846–847
- immersion period, diatom analysis, 1:524–525
- immigration of plants, Greenland, 3:764, 3:765–766
- impact glasses, loess, 2:632, 2:634f, 2:634t
- Impagidinium* dinocysts, 2:810f
- impaired glucose metabolism, OSA, 1:315–321
- inactive rock glaciers, 3:536f, 3:539
- incision, fluvial, 1:676–678, 1:679f, 1:684, 1:686–688, 2:115–119
- see also* degradation
- increments *see* growth increments
- independent dating, varved sediments, 4:585–586
- independent tests, isotope thermometer, 2:349–350
- index points, sea level, 4:371–372, 4:374f, 4:375f
- India
- Acheulean Industry, 1:77f, 1:81f, 1:98
 - Early Middle Pleistocene vertebrate studies, 4:651–652, 4:652t
 - Middle Pleistocene hominin archives, 1:75, 1:76f, 1:77f
 - monsoon lake level studies, 2:509–511
 - sand seas, 1:631–632
- Indiana, USA plant macrofossil records, 3:779–782, 3:781f
- Indian Ocean
- postglacial, 3:46–54
 - pre-glacial changes, 3:46–47
- Indian summer monsoon (ISM), 2:509–511
- indicative meaning, RSL changes, 4:372, 4:495–496, 4:497
- indicator erratics, 2:81–84
- indicator species
- chironomid paleoecology, 1:368–369
 - climate reconstructions, 3:787
 - definition, 1:361ge
 - human impact reconstructions, 3:881–891
- indigenous hominins, modern humans replacing, 1:100–103
- indirect ordination, definition, 1:361ge
- Indochinese Province, 4:693, 4:694f
- Late Pleistocene mammals, 4:696–697
 - Sundaic fluctuating boundary, 4:697
- Indonesia
- FT dating applications, 1:655–656
 - gateway systems, 2:151–152, 2:153f

- Indo-Pacific warm pool (IPWP), 3:62, 3:64–67
inductive high-temperature carbon reduction (iHTR), 1:482
inductively coupled plasma (ICP), 2:735
industrial emissions, 3:319–320
see also anthropogenic emissions
inert gas flow dehydration (iGFD), 1:482
infaunal benthic foraminifera, 2:767–768, 2:770
inference models
 chironomid paleoecology, 1:362, 1:369–370
 definition, 1:361*ge*
 see also qualitative inferences; quantitative inferences
infield-outland/outfield system, pollen records, 3:890–891
infillings, calcite, 3:383
information gathering, hominid–animal interactions, 4:752–754
infrared (IR) spectroscopy
 paleolimnological applications, 3:349–355
 statistical approaches, 3:351–352
 theoretical background, 3:349–350
infrared-stimulated luminescent (IRSL) dating, 1:13
injection ice, frost mounds, 3:472
inland waters *see* paleolimnology
inlet-associated facies, high-energy coasts, 4:387–388
inner shelf facies, high-energy coasts, 4:386–387
Innuition ice sheet
 areal ice extent, 1:891
 Canadian pollen records, 4:44
 Late Quaternary, 2:245, 2:247–248, 2:248*f*
 RSL, 4:469–470, 4:470–471, 4:470*f*, 4:471–474, 4:474*f*
Inuit whalers, 1:516, 1:517*f*
INQUA project, 2:144, 2:146*f*, 2:147*t*
insectivore dwarfing/gigantism, 4:744
insects
 forensic science use, 4:548–555
 fossil studies
 beetles
 Late Pleistocene, 1:191–199, 1:200–206, 1:244–254
 Japan, 1:207–220
 North America, 1:221–234
 Northern Asia, 1:255–273
 South America, 1:235–243
 Late Tertiary/Early Quaternary, 1:173–183
 Middle Pleistocene, 1:184–190
 overview, 1:161–172
 postglacial
 Europe, 1:274–281
 North America, 1:282–290
 chironomid records
 Africa, 1:361–372
 Late Pleistocene, 1:373–385
 overview, 1:355–360
 postglacial
 Europe, 1:386–397
 Southern Hemisphere, 1:398–405
 frozen sediments, 1:174
 history of, 1:173
 longevity of species, 1:178–179
 MCR reconstructions, 1:180–181
 oribatid mites, 2:679–698
 paleoclimate indications, 1:177–178
 paleoecological indications, 1:179–180
 preservation of fossils, 1:173–174
 life cycles, 4:549
 in situ carbon-14, 1:413–414, 1:414*f*, 2:357, 2:424–426
 in situ cosmogenic radionuclides, 2:356, 2:357
 insolation
 last interglacial, 3:155, 3:156–159, 3:158*f*
 Milankovitch theory, 2:136–137, 2:140*f*
 neoglaciation, 2:275, 2:275*f*
 North American lake level studies, 2:543–544, 2:545–546
 Quaternary climatic change, 1:27
 insoluble microparticles
 aerosols, 2:342–343
 heavy metal pollution, 2:344
 insulin resistance, 1:315, 1:317, 1:319
 intelligence work
 geological analysis, 4:530
 search philosophies, 4:525*t*
 soil/sediment as evidence, 4:535
 interandean depression, definition, 4:680*ge*
 interannual diatom productivity, marine varves, 1:556–559
 interannual variations, $\delta^{18}\text{O}$, 2:390
 interflows, glaciolacustrine, 2:43–44, 2:44*f*
 interglacial climatostratigraphy, 4:222–223
 interglacial cycles *see* interglacial–glacial cycles/changes/transitions
 interglacial deposits, insect fossils, 1:187–188, 1:188–189
 interglacial Eemian beetle assemblages, 1:200, 1:201*f*
 interglacial fluvial systems, 2:115–125
 interglacial–glacial cycles/changes/transitions
 Antarctic isotopic records, 2:398–400
 definition, 2:216*ge*
 diatom records, Pacific Ocean, 1:580–581
 fluvial responses, 1:690–691, 2:112–125
 history of ideas, 2:112
 loess deposits, 2:573–584
 North Atlantic Ocean, 3:10–12
 North Pacific Ocean, 3:34–36, 3:42–43
 pollen records, 2:704, 4:79–82
 see also glacial–interglacial cycles/stages/transitions; Holocene
 interglacial loess, 3:359
 interglacial stages, Northern Asia, 1:272, 4:36–37
 interior transect, WNA, 4:82
 INTERMAGNET geomagnetic observatories, 1:707–709
 intermediate-field regions, RSL changes, 4:489
 intermediate water
 North Pacific, 3:33, 3:34–37
 South Atlantic, 3:28–29
 internal accumulation, definition, 1:909*ge*
 internal stratigraphy, Greenland ice sheet, 2:443–445, 2:444*f*
 International Geological Correlation Programme (IGCP), 4:370, 4:371
 interpolated comparisons, ice cores, 2:414
 interspecies variability, foraminifera, 2:873–874
 interstadial climatostratigraphy, 4:222–223
 intertidal environments, RSL changes, 4:396–397, 4:397–398
 intertidal notch indicators, 4:388
 intertropical convergence zone (ITCZ)
 carbon-14 record calibration, 4:351
 eolian sediments, 1:639–640
 lake level studies, 2:531, 2:532*f*, 2:533–534
 postglacial North Pacific, 3:62–63, 3:64–65, 3:65–66, 3:69
 precipitation, South America, 2:389
 THC, 1:743, 1:746
 intra-annual diatom productivity, marine varves, 1:556–559
 intrapolygon sediment columns, Yedoma, 3:545*f*, 3:549
 intraspecies variability, foraminifera, 2:873–874
 invasive species biogeography, 3:321–322
 inverse modeling, POLLSCAPE, 3:877–878
 involution, cryoturbation structures, 3:430
 ion balance, seawater, 2:862*f*
 ion chromatography (IC), 2:327–328
 ion concentrations
 ice cores, 2:326–333
 salinity/climate reconstructions, 1:507–509
 see also major ions
 ionic budget, glaciochemistry, 2:326, 2:329–330
 Iowa, USA plant macrofossil records, 3:772–775, 3:776*f*
 IP₂₅ HBI alkene, 1:591–595
 flux profiles, 1:595*f*
 future directions, 1:594–595
 refining, 1:594–595
 relative abundances, Arctic, 1:593*f*
 structure/properties, 1:591–592
 see also highly branched isoprenoid
 IPWP *see* Indo-Pacific warm pool
 Iranian late glacial studies, 3:786–787, 3:788*f*
 IRD *see* ice-rafted debris/detritus
 Ireland
 ice sheet, 2:230–231
 relative sea level changes, 4:492
 IRM *see* isothermal remanent magnetization
 IRMS *see* isotope-ratio MS
 iron (Fe)
 coatings, soil, 3:383
 fertilization, climate cycles, 3:416–417, 3:417*f*
 hypothesis, deep-sea sediments, 1:637
 magnetic susceptibility, 2:888
 Irrawaddy, Cambodia, 4:652*t*

- IRSL *see* infrared-stimulated luminescent dating
- IR spectroscopy *see* infrared spectroscopy
- Irvingtonian age, North America, 4:646–647, 4:647–648, 4:649
- Isampur site, Asia
Acheulean Industry, 1:81*f*
- island fauna, 4:662–663
see also dwarfing/gigantism
- Isle of Man glacial sequence stratigraphy, 2:89*f*
- ISM *see* Indian summer monsoon
- isocon technique, blockfields, 3:527–528
- isolation, hominins, 1:80
- isolation basins
benthic foraminifera, 2:770
sea-level changes
Greenland, 4:484, 4:485*f*
microfossil-based reconstructions, 4:424, 4:424*f*
sedimentary indicators, 4:401–403, 4:403*f*, 4:404*f*
- isopollen maps, plant distributions, 3:854–855, 3:855–856
- isoprenoid alkene diatom biomarkers, 1:590–591, 1:591–595, 1:591*f*
- isostasy, 4:452–459
Earth model viscosity structures, 4:455–456
glacier/icesheet extent, 1:892
Late Quaternary ice histories, 4:456–457
mass conservation, 4:453–455
model applications, 4:455–457
ocean surface deflection, 4:452–453
sea-level change, 4:370
cave data, 4:460, 4:461, 4:462
GLA modeling, 4:439, 4:440–441, 4:443–444, 4:447
glaciation-induced, 4:452–459
sea-level equation, 4:455
solid earth surface deflection, 4:452, 4:453*f*, 4:454*f*
tectonic processes, 4:507–511
- isostatic rebound, glacier/icesheet extent, 1:892
- isothermal plateau fission-track (ITPFT)
dating, 1:12
glass, 1:650, 1:653–654, 1:653*f*, 1:653*t*, 1:655, 1:655*t*, 1:656, 1:659*f*
tephrochronology, 4:287, 4:299–301
- isothermal remanent magnetization (IRM), 3:377, 3:380
- isotope paleothermometers, 2:395–396
- isotope-ratio MS (IRMS), 3:330
- isotopes
bone-isotope analysis, 4:748*ge*
chironomid records
Late Pleistocene, 1:374
postglacial Europe, 1:395*f*, 1:396
definition, 3:768*ge*, 2:915
dendrochronology, 1:456
diffusive fractionation, 2:369, 2:370*f*
equilibrium, carbonate-precipitating lakes, 1:334
fractionation
carbon-14 temporal variation, 4:337
firn air, 2:369, 2:370*f*
glacial–interglacial cycles, 2:398–400
Greenland stable isotopes, 2:403–409
Holocene Antarctica, 2:396–398
ice-margin sites, 2:419–424, 2:421*f*
lake levels, Latin America, 2:533–534
lake sediments, 1:333–340
loess composition, 2:637
marine diatom proxies, 2:819
paleolimnology, 3:265
paleothermometers, 2:395–396
prairie-forest border, 3:768–769, 3:773*f*
precipitation composition, 2:395–396
ratios
dust, ice cores, 2:343–344
ice cores, 2:343–344, 2:347–352, 2:389–390
Indian Ocean, 3:46–47, 3:49*f*
pigment analysis, 3:330
research overview, 1:291–293
seawater, 2:915–918
terrigenous sediments, 2:951
stratigraphy, ice cores, 2:351
surface exposure dating, 1:924–925
see also carbonate stable isotopes; cosmogenic isotopes; hydrogen isotopes; Marine Isotope Stage...; marine oxygen isotope records; oxygen isotopes; radioisotopes; silicon isotopes; stable isotopes
- isotope thermometers
independent tests, 2:349–350
reliability, 2:348–349
- isotopic proxies *see* diatom silica records
- Israel
Early Pleistocene, 1:68
Middle Pleistocene, 1:71–72
see also 'Ubeidiya, Israel
- Italy
Middle/Late Pleistocene pollen records, 4:64–65
postglacial pollen records, 4:181*t*, 4:185–186
- ITCZ *see* intertropical convergence zone
- iterative processes, paleoclimate modeling, 3:145*f*, 3:146
- ITPFT *see* isothermal plateau fission-track dating
- IVGTT *see* intravenous glucose tolerance test
- J**
- Jakobshavn Isbrae, ice sheet, 2:441
- Jamaica, sea-level changes, 4:446–447
- Japan
entomology, 1:219–220
Late Pleistocene insect fossils, 1:207–220
pollen database, 3:833
Sea of, 2:896, 2:896*f*
Tohoku area diatom records, 1:580–581
Younger Dryas oscillation, 3:224–225
- Japanese Islands, Northwest Pacific, 1:571, 1:574–575, 1:575–580
- Japanese Pollen Database (JPD), 3:833
- Japan Sea diatom records, 1:581
- Java
Early Middle Pleistocene vertebrate studies, 4:653–654, 4:655*t*, 4:661*t*
Early Pleistocene hominin remains, 1:67, 1:69*t*
Late Pleistocene mammals, 4:694–695
Jefferson's ground sloth, 4:716
Jeffery rotation, till fabric, 2:78–79, 2:79*f*
Johnsen, SJ, 2:404, 2:405, 2:407, 2:408
jökulhlaups, 2:488*f*, 2:489
see also glacial lake outbursts
- journals, climatology, 3:238, 3:239
- JPD *see* Japanese Pollen Database
- Juncus* macrofossils, 3:642, 3:643*f*
- Jura Mountains lake level studies, 2:549, 2:550*f*, 2:552*f*, 2:553, 2:554*f*
- Jurby Formation, Isle of Man, 2:89*f*
- K**
- Kaharoa Tephra, 4:298–299
- Kalahari Desert, 1:627–628
- kames, 2:15
- kame terraces, 2:13
- kangaroo records, 4:634–636
- Kap København Formation, Greenland, 3:114–115, 3:761–762
- Karaginsky interstadial, Beringia, 2:197
- Kara Sea diatoms, 1:567
- K/Ar dating, 1:10–11, 1:10*f*, 1:449, 2:477–482
age calculation, 2:478–480
argon-40/argon-39 dating, 2:477–482
historical development, 2:477–478
instrumentation, 2:481
methods, 2:478–480
new perspectives, 2:480–481
strengths/weaknesses, 2:480
- Karginski interstade, Northern Asia, 4:34–36
- karstic caves, 1:294–303
- Kaschin-Beck disease, loess, 2:604
- Kazakhstan pollen records, 4:166
- Kazantsevskoye Stages, Northern Asia, 4:36–37
- Kedung Brubus, Java, 4:653–654, 4:655*t*
- Kenyan chironomid records, 1:363, 1:366*f*
- Keo Leng Cave, Vietnam, 4:657, 4:658*t*
- Keshan disease, loess, 2:604
- Kessler, MA, 3:454–455, 3:455*f*, 3:457
- Khaprovian (Khapry) mammal assemblage, 4:605–607
- Khasarian fauna, 4:612–613
- Khomustakh Lake, Siberia, 4:168, 4:169*f*
- Khonako, Central Asia, 2:586*f*
- Kibo ice fields, Tanzania, 2:373, 2:374–375, 2:374*f*, 2:377*f*
- Kilimanjaro ice core records, 2:373, 2:374*f*, 2:375–377
abrupt climate events, 2:376–377
African Humid Period, 2:375–376
Early Holocene, 2:376
meteorology, 2:373
Middle Holocene, 2:376–377

kinematics, rock glaciers, 3:537–538
 kinetic energy, cosmic rays, 1:419, 1:422f, 1:426f, 1:430
 King Island, Tasmania, 1:194–195
 Kipp... *see* Imbrie–Kipp...
 Klasies River Mouth (KRM), Africa, 1:91, 1:92–93
 Kleman, J, 1:898–899, 1:900–901, 1:902, 1:903f
 knapping, 1:83ge, 1:88–89
 knickpoint, fluvial terraces, 1:685
 koala records, 4:633
 Kola Peninsula, Russia
 chironomid records, 1:388f
 megafossils, 3:623f, 3:625–626, 3:625f, 3:627f
 Koloshan cave, China, 4:655–657
 Komodo Dragons, 4:736f, 4:745
 Kotokel Lake, Siberia, 4:168–170, 4:169f
 Kra Isthmus, Southeast Asia, 4:693, 4:694f
 Kråkenes, Western Norway, 3:788–789
 Krantz, WB, 3:459–460
 KRM *see* Klasies River Mouth
 krummholz, treelines, 3:621, 3:622f, 3:690
 krypton-81, ice-core dating, 2:306
 krypton-85, ice-core dating, 2:306
 Kul dara, Tajikistan, 1:75–76, 1:78f
 Kumtag Desert, China, 1:607f, 1:608f, 1:610f
 Kuroshio Current, North Pacific, 3:62, 3:64–67
 Kutzback JE, 3:104
 kyr measures *see* thousand year measures

L

Labrador, Canada, pollen records, 4:47, 4:49
 Labrador Sea
 last interglacial, 3:162–163
 postglacial, 3:57–59, 3:58f, 3:60, 3:61
 Labrador Sea Water (LSW), 3:162–163
 lacustrine... *see* lake...
 Lago Grande di Monticchio, Italy, 4:291, 4:291f
 Lagriidae beetles, 1:214
 Lahontan Lake, North America, 2:539–540, 2:562
 LA-ICP-MS *see* laser ablation inductively coupled plasma mass spectrometry
 Lake District, England, 1:819f
 lake levels
 African Late Quaternary, 2:499–505
 allostratigraphy, 4:245–247
 Asia, 2:506–523
 Australia, 2:524–530
 changes/fluctuations
 chironomid records, 1:360
 cichlid fishes, Africa, 2:504
 macrofossil analysis contribution, 3:665–669
 west-central Europe, 2:551, 2:552–553, 2:553–556
 chironomid records, 1:360
 data-model comparisons, paleoclimates, 3:137f
 glacial sequence stratigraphy, 2:87f

Latin America, 2:531–536
 modeling, 2:558–563
 basis of, 2:558
 discharge, 2:560
 energy balance, 2:559
 ice models, 2:559
 input data, 2:560
 morphometry effects, 2:563
 paleoclimates, 3:137f
 results, 2:561
 soil moisture balance, 2:559–560
 strategies, 2:560–561
 surface energy balance, 2:559
 water balance, 2:558–559, 2:559f
 morphostratigraphy, 4:243
 North America, 2:537–548
 Alaska, 2:543–544
 continental-scale climate variations, 2:545–546
 East, 2:540f
 Great Basin, 2:539–540, 2:542f
 Midwest, 2:537–539
 Northeast, 2:544–545
 Northern Great Plains, 2:537–539
 regional histories, 2:537–545
 Rocky Mountains, 2:540–543, 2:543f
 Southeastern, 2:545
 Southern Canadian Cordillera, 2:540–543
 Yukon territory, 2:543–544
 overview, 2:483–498
 reconstructions, 2:483–498
 background, 2:484–487
 core recovery, 2:495–496
 diatoms, 2:492–493, 2:496–497, 2:497f
 freshwater lakes
 carbonate sediments, 2:492f, 2:493
 clastic sediments, 2:489–493
 site selection, 2:487
 geomorphic archives, 2:488–489
 lithological archives, 2:489–495
 North America, 2:539f, 2:544f
 ostracods, 2:496
 oxygen-isotope analysis, 2:496–497
 past change, 2:484–488
 saline lakes, 2:493–495, 2:494f
 diatoms, 2:496–497
 ostracods, 2:496
 oxygen-isotope analysis, 2:496–497
 sediment classification, 2:495–496
 site selection, 2:487
 sediment classification, 2:495–496
 site selection, 2:487
 site survey, 2:487–488
 techniques, 2:488–495
 west-central Europe, 2:549–551
 west-central Europe, 2:549–557
 see also lakes; lake water
 lake sediments, 3:302–303
 annual layer dating methods, 4:573–581, 4:582–589
 basin-wide properties, 3:300–302, 3:302f
 beetle records, 1:162, 1:168–169
 bulk density, 3:306–310, 3:307f
 carbonate stable isotopes, 1:293, 1:333–340

charred particle analyses, 2:720f
 chironomid records, Africa, 1:365, 1:370–371
 classification, saline lakes, 2:495–496
 color assessment, 3:302–303
 diatom records, 2:739–741
 human impact, 1:541–543
 laminated, 1:540–545
 large lakes, 1:546–553
 early diagenesis, 3:310
 freshwater lakes, overview, 2:489–493, 2:495–496
 future research, 3:310–311
 geomagnetic PSV, 1:714
 glacial depositional systems, 2:22–24, 2:25f
 grain size, 3:305–306, 3:306f, 3:307–308, 3:307f
 human impact diatom records, 1:541–543
 laminated diatom records, 1:540–545
 large lake diatom records, 1:546–553
 magnetic properties, 3:303–305, 3:304f
 physical properties, 3:300–312
 plant macrofossils, 3:593–594, 3:594f, 3:609f, 3:726
 sampling strategy, 1:333
 shear strength, 3:309–310, 3:310f
 stomatal frequency analysis, 3:614, 3:618
 tephrochronology, 4:296–298, 4:296f
 vertebrate remains, 4:592
 visible/infrared spectroscopy, 3:349
 water content, 3:307–308, 3:308f, 3:309f
see also lakes; varved sediments, lakes
 lake water
 balance equation, 2:558–559, 2:559f
 oxygen isotopes
 climate/hydrology relation, 1:336–337
 environmental effects, 1:338
 temperature calculations, 1:337–338
 surface energy balance, 2:559
see also lake levels
 lakes
 acidification, 1:359, 3:292–293, 3:334–335
 back-barrier, overwash sand-layers, 3:209–211
 bivalves, 1:344–345, 1:346f
 carbonate stable isotopes, 1:333–340
 chemistry, 3:292–299
 acidification, 3:292–293
 DOC, 3:292, 3:295–296, 3:296f
 ecosystems, 3:292
 eutrophication, 3:293–295
 POPs, 3:297, 3:298f
 salinity, 3:295
 trace metals, 3:296–297
 Chilean insect fossils, 1:235, 1:236–240, 1:237f, 1:242
 chironomid records
 acidification, 1:359
 Africa, 1:361–372
 ontogeny, 1:358
 postglacial Europe, 1:386–387, 1:387f, 1:392–393, 1:393f
 deposits
 allostratigraphy, 4:247, 4:247f
 paraglacial land systems, 3:561–562

- subaquatic glacial systems, 1:763–764
- diatom silica in, 1:481, 1:484–486, 2:739–741
- dynamics, Africa, 2:499–502
- ecosystems
- chemistry, 3:292
 - chironomid records, Europe, 1:386–387, 1:387f, 1:392–393, 1:393f
 - eutrophication, 3:659–665
 - macrophytes role, 3:657
- energy balance, 2:559
- glacier/icesheet extent, 1:886f, 1:888–890, 1:888f
- GLOFs, glaciolacustrine, 2:43
- habitat multiproxy approaches, 3:339–341, 3:341–342, 3:344, 3:345–346
- human settlements, 1:158
- ice models, 2:559
- macrophyte distribution, 3:657
- morphometry effects, 2:563
- ontogeny, chironomid studies, 1:358
- oribatid mite deposits, 2:693–694
- paleodroughts, 3:195
- paleolimnological multiproxy approaches, 3:339–341, 3:341–342, 3:344, 3:345–346
- pigment studies, 3:327f, 3:329–330
- acidification, 3:334–335
 - degradation, 3:330
 - food web changes, 3:333–334
 - hydrological change, 3:335
- pingo formation, 3:472–474, 3:476f
- plant macrofossils
- ^{14}C , 4:363
 - pollen studies, 3:725, 3:726, 3:728f, 3:729
- postglacial
- European chironomid records, 1:386–387, 1:387f, 1:392–393, 1:393f
 - North American pollen records, 4:136–140
- proglacial
- ice sheet growth/decay, 1:882
 - subaquatic glacial depositional systems, 1:763–764
- reconstructions
- Antarctica, 1:536–538
 - diatom oxygen systematics, 1:484–486
 - west-central Europe, 2:549–551
- salinity, 1:507–515, 3:295
- stratigraphic tephra dating, 4:288–289
- trap efficiency, 3:301–302
- trophy, 3:657
- vegetation
- Greenland, 3:765
 - surface samples/taphonomy/representation, 3:684–689
- see also* freshwater lakes; *see also* individual lakes; lake levels; lake sediments; paleolimnology; saline lakes; thermokarst, lakes; tundra lakes
- Lakhuti, Central Asia, 2:587f
- Lama Lake, Siberia, 4:167, 4:168f
- Lamb Spring, Colorado, 1:225, 1:226t
- laminated sediments
- analytical techniques, 4:584–585, 4:585–586, 4:587f
 - applications, 4:588
 - classification, 4:584
 - diatom productivity events, 1:556–559
 - freshwater diatoms, 1:540–545
 - marine diatoms, 1:554–561
 - accumulation sites, 1:554–555
 - historical perspective, 1:554 - millennial-scale changes, North Pacific, 3:37
 - preservation, 4:582–583
- l-amino acids, dating, 1:38, 1:39, 1:39f, 1:43f, 1:44, 1:46f, 1:47–48
- Lampsilis* bivalves, 1:343–344, 1:344f, 1:345f
- land cover
- Holocene biosphere feedbacks, 1:725, 1:726f
 - pollen data calibration, 3:892–895
 - pollen representation simulation approach, 3:871–879
- landforms
- cold region thermokarst topography, 3:574–581
 - cosmogenic nuclide dating, 1:432–434, 1:435–436
 - glacitectonic, 1:839–845
 - landform–sediment suites, 1:813–814
 - morphostratigraphy, 4:244, 4:244f, 4:247f
 - permafrost, 3:469
 - blockfields, 3:523–534
 - frost mounds, 3:472–480 - solifluction rates, 3:483–484
 - see also* glacial landforms; glacialfluvial landforms; periglacial landforms; strandlines
- landform–sediment suites, 1:813–814
- Land Mammal Ages (LMAs), North America, 4:673–679
- landnam theory, 3:880, 3:881f
- Landscape Reconstruction Algorithm (LRA), 3:892
- applications, 3:895–901, 3:900f, 3:901f
 - pollen analysis, 3:798–799
 - POLLSCAPE model, 3:874–875, 3:877–878, 3:878f
- landscapes
- biodiversity reconstruction, 3:819
 - climate, North America, 4:142–143
 - cultural history, 3:880–903
 - erosion, glacial, 1:861–863
 - evolution
 - cosmogenic nuclide dating, 1:440–445
 - micromorphological reconstructions, 3:388–389
 - northern climate, 3:746 - protection, blockfields, 3:532–533
 - stability, weathering profiles, 3:392, 3:398–399, 3:400
 - tephrochronology reconstruction, 4:293–294
 - see also* glaciated landscapes
- landslides
- cosmogenic nuclide dating, 1:437
 - loess records, China, 2:604
- land snails, 3:286f, 3:287f
- land-surface biophysics, 1:722, 1:722f
- landsystems
- glacial, 1:813–824
 - modern analogs, 1:813–817
 - see also* paraglacial land systems
- land use
- Holocene aboriginal Australians, 4:112
 - pollen surface sampling/trapping, 3:843–844
- Lang Trang Cave, Vietnam, 4:657, 4:658t, 4:660t, 4:661t
- La Niña events, 3:171, 3:175–176
- LAPD *see* Latin American Pollen Database
- Lapland
- chironomid records, 1:389–392
 - Finnish, ice sheet, 2:228
- Laptev Sea
- pollen records, 4:34
 - surface sediment diatoms, 1:567
- larder beetles, 4:553–554
- large lake diatom records, 1:546–553
- see also* lakes
- large mammals
- biozonations, 4:210, 4:213f
 - Middle-Pleistocene Southern Asia, 4:651–663
 - teeth/bones, 1:308, 1:309f
 - see also* mammals
- large-scale flow, Greenland ice sheet, 2:441–443
- large-scale mapping, vegetation, 3:861
- Larix* wood megafossils, 3:621, 3:622f, 3:623f, 3:627t
- Larsemann Hills, East Antarctica, 2:220–221
- Laschamp event, ice cores, 2:308
- laser ablation inductively coupled plasma mass spectrometry (LA-ICP-MS), 4:284–285, 4:299–301
- last cold stage plant macrofossil records, 3:740–741
- last deglaciation
- African pollen records, 4:89–98
 - Greenland, 3:760–761
 - North Atlantic Ocean, 3:15–17
 - paleoceanography, 2:751–753, 2:753f
- last glacial cycle
- nitrous oxide history, 2:473–474, 2:473f, 2:474–475
 - South American loess records, 2:635–638, 2:636f, 2:638t
- Last Glacial Maximum (LGM), 1:725–727
- African pollen records, 4:10, 4:11–12, 4:11f, 4:12–13, 4:13–14, 4:14–15
 - American human colonization, 1:122–124
 - Antarctica, 2:218–219
 - sea ice diatom studies, 1:533–535, 1:534f
 - stable isotopes, 2:398
- Arctic
- North America, 3:755
 - paleoclimate history, 3:118
- Asia
- lake level studies, 2:514–515
 - Late Quaternary glaciations, 2:239, 2:240, 2:241, 2:242f

- Last Glacial Maximum (LGM) (*Continued*)
 modern humans, 1:104–106
 atmospheric dust, 1:732–733, 1:732t
 Australia/New Zealand pollen records, 4:106
 beetle studies, 1:203
 New Zealand, 1:244, 1:247–248, 1:248–249, 1:249–251
 North America, 1:233
 biogeophysical feedbacks, 1:725–726, 1:727f
 biomes
 steppe/tundra, 3:866–867
 WNA pollen records, 4:78–79
 biosphere feedbacks, 1:725–727
 data-model comparisons, paleoclimates, 3:137, 3:139f, 3:142–145, 3:142f, 3:143f
 definition, 2:216ge
 dune fields, China, 1:610, 1:612f
 dust
 atmospheric, 1:732–733, 1:732t
 biogeochemical role, climate cycles, 3:413–415, 3:416, 3:416t
 feedback, 1:726
 eustatic sea-level change
 glacial–interglacial cycles, 4:429, 4:431, 4:432–434, 4:436
 since LGM, 4:439–451
 GCMs, 3:165–170
 air-sea interactions, 3:169
 cooling, 3:167, 3:168
 experiment design, 3:166–167
 future directions, 3:169
 MOC, 3:168–169
 monsoons, 3:168
 Northern Hemisphere jet stream, 3:167–168
 results, 3:167–169
 tropical climate, 3:168, 3:169
 types, 3:165–166
 glacier extent, 1:884, 1:891, 1:892, 1:893
 glacier flow directions, 1:898–899
 Greenland
 plant macrofossils, 3:763
 sea-level changes, 4:483–484
 ice sheet
 dimensions, sea-level change, 4:373
 extent, 1:884, 1:891, 1:892, 1:893
 isostasy, 4:457
 lake level studies
 Asia, 2:514–515
 Latin America, 2:531–532, 2:533–534
 Late Pleistocene
 Beringia, 2:192, 2:197–198
 South Atlantic, 3:19, 3:24–26
 Late Quaternary glaciations, 2:239, 2:240, 2:241, 2:242f
 magnetic susceptibility, 2:895, 2:895f
 marine biota feedback, 1:726–727
 mega-floods, Northeast Pacific Ocean, 1:581–585
 methane/atmospheric chemistry feedback, 1:727
 North America
 Arctic, 3:755
 beetle studies, 1:233
 loess records, 2:627
 pollen records, 4:47–49, 4:78–79, 4:124–125, 4:127–128, 4:129, 4:136–137
 North Atlantic Ocean, 3:10
 North Pacific Ocean
 mega-floods, 1:581–585
 surface water, 3:38, 3:39f, 3:40f
 nutrient proxies
 barium in Atlantic, 2:903, 2:903f
 cadmium in Atlantic, 2:902f
 nitrogen-15, 2:903f
 NWNNA pollen records, 4:124–125, 4:127–128, 4:129
 paleophytogeography, 2:731
 plant distributions, 3:856–857, 3:858
 plant macrofossils, Greenland, 3:763
 postglacial pollen records
 Australia/New Zealand, 4:106
 southeastern North America, 4:136–137
 regional pollen studies, 3:800–801
 relative sea level changes, 4:489
 lowstand, 4:498–499, 4:500f
 QEI RSL, 4:469–470
 Sahul landmass, 1:108–109
 salinity proxies, 2:935–936, 2:936f
 silicon leakage, 2:739
 South America
 glaciations, 2:250, 2:252–253, 2:253–254
 ice core studies, 2:387, 2:390
 vertebrate responses, 4:690
 terrestrial CO₂ feedback, 1:726
 trimlines/paleonunataks, 1:918–919, 1:919f, 1:920–921, 1:921f, 1:922f, 1:924, 1:926–927
 vegetation patterns, 3:864
 weathering feedback, 1:727
 last glaciation
 beetle studies
 Japan, 1:219
 North America, 1:221–225
 South America, 1:236, 1:242
 diatom studies, North Atlantic/Arctic, 1:564–565
 Greenland stable isotopes, 2:407–408
 Holocene transition from, 2:313–314
 ice core studies, 2:313–314, 2:314–315
 megafaunal extinctions before, 4:700–704
 southwest Pacific region, 2:202, 2:204, 2:204f, 2:205f
 last ice age
 ice-core research, 2:284
 plant macrofossils, Greenland, 3:763
 last interglacial (LIG), 3:155–164
 annual mean SAT response, 3:160–162
 Arctic
 paleoclimate history, 3:117
 plant macrofossil records, 3:739–740, 3:754–755
 atmosphere dynamics, 3:159–160
 beetle distributions, New Zealand, 1:247
 biosphere feedbacks, 1:727–728
 boundary conditions, 3:155–156, 3:158t
 coral records, RSL changes, 4:409–411, 4:414
 European pollen records, 4:1–8
 globally averaged temperatures, 3:160–162
 insect assemblages, 1:272
 insolation, 3:155, 3:156–159, 3:158f
 northern summer response, 3:156–159
 northern winter response, 3:159–160
 ocean circulation changes, 3:162–163
 paleoclimate modeling, 3:155–164
 plant macrofossil records
 Arctic Eurasia, 3:739–740
 Arctic North America, 3:754–755
 Greenland, 3:762–763
 sea-level highstand, 4:498
 seasonal responses, 3:156–159
 stratigraphy overview, 4:193–194, 4:193f, 4:200
 vegetation feedbacks, 3:160–162
 last millennium climate, 3:229–236
 atmospheric circulation, 3:231–233
 drought, 3:230–231, 3:232f
 ENSO, 3:231–233, 3:234, 3:235
 large-scale temperature trends, 3:229
 NAO, 3:231, 3:232f, 3:233, 3:234, 3:235
 oceanic circulation, 3:231–233
 paleoclimate reconstruction, 3:229–236
 precipitation, 3:230–231
 regional climate changes, 3:229–231
 surface temperature, 3:229–230
 theoretical modeling, 3:233–235
 last occurrence (LO) event, *Proboscidea curvirostris*, 1:562–563, 1:563–564, 1:565f
 Latamne site, Asia, 1:72, 1:75f
 Late Cenozoic glacial landforms, 1:860
 Late Early Pleistocene mammal assemblages, 4:608–610
 Late Galerian mammal assemblages, 4:611–612
 lateglacial interstadial beetle assemblages, 1:204–205
 see also lateglacial stadial beetle assemblages
 Lateglacial period/change
 Alaska/Yukon pollen records, 4:40, 4:43f
 beetle assemblages, 1:203–205
 Canadian pollen records, 4:45f
 chironomid records
 Europe, 1:373–374, 1:375, 1:376–378, 1:377f, 1:378f, 1:380–383, 1:381f, 1:382f, 1:383f
 Southern Hemisphere postglacial, 1:400–401, 1:400f, 1:402–403, 1:403–404
 climate change, Greenland, 3:761
 CO₂ fossil leaves reconstruction, 3:617–618
 eustatic sea-level, 4:441–444
 human adaptations, 1:156
 human dispersal, Americas, 1:124–125
 insect fossil studies, 1:274
 plant distributions, 3:857–858
 plant macrofossils, 3:741–742
 Greenland, 3:763
 multidisciplinary studies, 3:785–793

- paleolimnology, 3:657–658, 3:662f
 rapid sea-level rise, 4:499
 rodent middens, 3:679t
 salinity/climate reconstruction, 1:510–512
 South American glaciations, 2:250, 2:252, 2:253, 2:254–255
 treeline dynamics, 3:692–693
 lateglacial stadial beetle assemblages
 GS 1, 1:205
 GS 2, 1:203–204, 1:205
 see also lateglacial interstadial beetle assemblages
 Late Holocene
 glacier oscillations, 2:221
 glacier recession, 1:803–805
 postglacial pollen records, 4:137–140
 rodent middens, 3:679t
 salinity proxies, 2:934–935
 sea-level change, 4:375
 ESL, 4:449–450
 midlatitudes, 4:489, 4:490–491
 standlines, 4:512f
 see also Holocene
 Late Mesolithic human adaptations, 1:157–159
 Late Middle Pleistocene vertebrate studies, 4:644
 Late Neogene/Early Quaternary transition, 2:151–166
 AMOC, 2:154, 2:160–162
 atmospheric CO₂ decline, 2:154
 CAS, 2:151, 2:152–154, 2:154–156, 2:156–164, 2:159f, 2:160f, 2:161f, 2:163f
 CO₂ drawdown in ocean, 2:160–162
 Indonesian gateway systems, 2:151–152, 2:153f
 models of NHG onset, 2:151–154
 North Pacific oceanography, 2:162–164
 orbital forcing, 2:151
 Panama hypothesis, 2:152–154
 Plio-Pleistocene boundary, 2:151
 poleward freshwater transport, 2:162
 poleward heat transport, 2:160–162
 sediment records, 2:156–159, 2:160–162
 Late Paleolithic *see* Upper Paleolithic
 Late Pleistocene
 Africa
 pollen records, 4:9–17
 vertebrate studies, 4:664–672
 archeological sites, Australia, 1:109f
 beetle fauna
 Australia, 1:191–199
 climate change, 1:200–206, 1:241, 1:242
 Europe, 1:200–206
 New Zealand, 1:244–254
 North America, 1:282–285
 Beringia events, 2:191–201
 Early Wisconsin/Zyryan glaciation, 2:194–197, 2:197f, 2:198f
 eastern/western Beringia, 2:192, 2:192f, 2:195f, 2:198f, 2:199f
 Late Wisconsin/Sartan glaciation, 2:197–199, 2:198f, 2:199f
 Middle Wisconsin/Karaginsky interstadial, 2:197
 MIS 4, 2:193, 2:194–197
 MIS 5, 2:193–194
 paleoclimate contrasts, 2:192
 sea-level history, 2:192–193, 2:193f
 timing of major cold events, 2:193t
 Yedoma, 3:542–552
 chironomid records, Europe, 1:373–385
 dynamics, northern Andes, 4:54–58
 extinctions
 Early Holocene boundary, 4:691
 mammals, 4:676–678
 megafaunal, 4:700–712
 glaciation
 Eurasia, 2:224–235
 North America, 2:180
 South America, 2:250–256
 southwest Pacific region, 2:202–215
 glacier recession, 1:807–808, 1:809–810
 insect fossil studies
 Japan, 1:207–220
 Northern Asia, 1:255–273
 South America, 1:235–243
 mammal assemblages
 mummies, 4:713–717
 Northern Eurasia, 4:612–613
 Southeast Asia, 4:693–699
 megafaunal extinctions, 4:700–712
 regional disparities, 4:704–709
 theories on causes of, 4:709–711
 mummified mammals, 4:713–717
 North America
 beetle studies, 1:221–234, 1:282–285
 pollen records, 4:39–51, 4:72–83
 vertebrate studies, 4:673–679
 paleoceanography
 North Atlantic, 3:9–17
 North Pacific, 3:33–45
 pollen records
 Africa, 4:9–17
 Australasia, 4:18–26
 North America, 4:39–51, 4:72–83
 Northern Asia, 4:27–38
 South America, 4:52–62
 Southern Europe, 4:63–71
 rodent middens, 3:674–683
 sea-level changes, 4:504f
 South America
 glaciations, 2:250–256
 insect fossil studies, 1:235–243
 pollen records, 4:52–62
 vertebrate studies, 4:680–692
 South Atlantic, 3:18–32
 Southern European Middle/Late Pleistocene pollen records, 4:63–71
 teeth/bones, carbonate stable isotopes, 1:311–312
 vertebrate studies
 Africa, 4:664–672
 ancient DNA, 4:720–721, 4:722
 North America, 4:673–679
 South America, 4:680–692
 Yedoma of Beringia, 3:542–552
 see also Pleistocene
 Late Pleistocene/Early Holocene (LP/EH)
 boundary extinctions, 4:691
 Late Pliocene
 Arctic North America, 3:749–753
 vertebrate studies, 4:599
 Late Quaternary
 African lake level history, 2:499–505
 Antarctic glaciations, 2:216–223
 Chinese loess records, 2:599–600
 European beetle fauna, 1:200–206
 Greenland sea-level changes, 4:481–488
 Highland Asia glaciations, 2:236–244
 isostasy, 4:456–457
 North America
 ice sheets, 2:245–248
 vegetation reconstruction, 3:867
 paleoclimatic plant–beetle reconciliation, 1:197–198
 plant distributions, 3:855–857
 sea-level change, 4:489–494
 Greenland, 4:481–488
 RSL
 high latitudes, 4:467–480
 midlatitudes, 4:489–494
 tropics, 4:495–502
 tectonics, 4:503–519
 southwest Pacific region, 2:202–215
 vegetation reconstruction
 large spatial scales, 3:861
 North America, 3:867
 Late Stone Age, 1:95
 see also Upper Paleolithic
 Late Tertiary insect fossil studies, 1:173–183
 Late Villafranchian mammal assemblages, 4:608–610
 Late Villányian mammal assemblages, 4:605–607
 Late Weichselian (LW) insect assemblages, 1:263–271, 1:269t
 Late Wisconsin
 Colorado Plateau rodent middens, 3:680f
 glaciation, Beringia, 2:197–199, 2:198f, 2:199f
 Great Basin rodent middens, 3:679t
 ice sheets, North America, 2:247–249
 lateral meltwater channels, 1:832–833
 laterite profiles, weathering zones, 3:395–396
 latero-frontal fans/ramps, 1:770–772, 1:772f
 Latest Early Pleistocene vertebrate studies, 4:639–641
 Latin America
 lake level studies, 2:531–536
 pollen studies, 3:831, 3:832, 3:833
 see also South America
 Latin American Pollen Database (LAPD), 3:831, 3:832, 3:833
 latitude
 coccolithophores, 2:784–786, 2:789–790
 radiolarians, 2:832f
 RSL in high latitudes, 4:467–480
 see also high latitudes; low latitudes
 Laurentide Ice Sheet (LIS)
 Arctic paleoclimate history, 3:115–116
 areal ice extent, 1:884, 1:890, 1:891, 1:891f
 Canadian pollen records, 4:44, 4:49
 glacial lake outbursts, 1:827f, 1:828
 glacier flow directions, 1:897f, 1:899, 1:902f, 1:905f
 glacier recession, 1:807, 1:808–809

- Laurentide Ice Sheet (LIS) (*Continued*)
 ice sheet growth/decay, 1:881–882
 Late Quaternary, 2:245, 2:246f, 2:247–248, 2:248f, 2:249
 North American lake level studies, 2:543–544, 2:545–546
 postglacial North Atlantic, 3:60
 Younger Dryas climate event, 3:130–131
- lava-fed deltas
 definition, 1:780ge
 mafic tuyas and, 1:781–782, 1:782–783, 1:791
- lava flows
 felsic tuyas dominated by, 1:792, 1:796f
 secular variation, 1:711f, 1:713–714
- law enforcement
 geology use, 4:521–534
 palynology use, 4:556–566
- Law of Superposition, stratigraphy, 4:190
- LCM *see* Last Glacial Maximum
- leaching, amino-acid dating, 1:43–44
- lead-210 dating
 ice cores, 2:306
 laminated sediments, 4:585–586, 4:586f
- lead (Pb)
 dating techniques, 2:306, 4:585–586, 4:586f
 Greenland
 D5 ice core, 2:281, 2:282f
 pollution, 2:344, 2:345
 isotopes, U-series dating, 4:567
 microparticles, ice cores, 2:343
- leaves
 CO₂ reconstruction from fossils, 3:613–620
 identification for analysis, 3:615
 macrofossils, 3:597f
 preparation for analysis, 3:615
see also plant...; vegetation...
- lee-side cavity filling, till, 2:64, 2:71
- Leicestershire, England beetle assemblages, 1:186–187
- lemmings, 4:724
- Lemmus curtatus* (sagebrush voles), 4:725
- lenses *see* ice lenses
- Lepus americanus* *see* snowshoe hare
- Lesotho, block streams, 3:517f
- lesser dung flies, 4:551
- Levallois core technique, 1:59, 1:86t, 1:147, 1:83ge
- Levant region
 hominin archives, 1:71–74, 1:74f
 Middle Paleolithic periodization, 1:99t
 modern human expansion, 1:98–99, 1:105f
 technocomplexes, 1:149–150
 transitional industries after 300 ka, 1:98
- Levinson-Lessing Lake, Siberia, 4:167, 4:167f
- Lewis Glacier, Kenya, 2:373, 2:377
- LFS *see* local flooding surface
- LGM *see* Last Glacial Maximum
- LIA *see* Little Ice Age
- Liang Luar stalagmite, Indonesia, 3:65–66, 3:65f
- Libby, Willard F, 1:7, 1:7f
- lichenometry, 2:565–572
 accuracy/precision aspects, 2:566, 2:571
 approaches to, 2:566, 2:567–570, 2:570–571
 assumptions, 2:565
 future directions, 2:571
 growth curves, 2:567, 2:571
 growth rate calibration, 2:566–567
 neoglacial chronology establishment, 2:270–271
 Northwest Territories, Canada, 2:570f
 origins, 2:565
 rationale, 2:565
 relative age dating, 2:566
Rhizocarpon geographicum, 2:565, 2:568–569
 statistical approaches, 2:570–571
 suggestions/concerns, 2:571
 traditional approach, 2:567–570
Xanthoria elegans, 2:570f
- lichen trimlines, 2:566
- life cycle stages
 dinoflagellates, 2:801–803, 2:803f
 insects, 4:549
- LIG *see* last interglacial
- light
 absorption, photosynthetic organisms, 3:327–328
 effects on coccolithophores, 2:784–786
 microscopy, palynology, 4:556–559
- light microscopy (LM), 4:556–559
- limestone records, RSL changes, 4:411, 4:412f
- lime (*Tilia*)
 distributions, 3:855, 3:856f
 stand-scale palynology, 3:850–852, 3:851f
- limiting dates, sea-level change, 4:372
- limiting factor principle, dendrochronology, 1:454
- Limnos samplers, pollen, 3:839–840, 3:841f
- linear dunes, 1:629f, 1:630f
- linear erosion, selective
 blockfields, 3:530–532
 glacial landscapes, 1:862
- linear modulation (LM), OSL dating, 2:660, 2:660f, 2:664
- lipids
 δ D salinity proxy, 2:778–779
 diatoms, Arctic, 1:590–591
- lipolysis effects, 1:320
- liquid scintillation counting (LSC), 1:9, 4:306–311
 anticoincidence shielding, 4:308
 background reduction feature, 4:309t
 benzene synthesis, 4:309–310, 4:310f
 counter setup, 4:310
 enhanced passive shielding, 4:307–308
 instrumentation, 4:306–309
 procedures, 4:310–311
 pulse discrimination electronics, 4:308
 quasi-active guard detector, 4:308–309
 radiocarbon dating, 4:306–311
 sample analysis, 4:309–311
 sample quenching, 4:310
- LIS *see* Laurentide Ice Sheet
- listerine weevils, 1:235, 1:236
- lithalsas, 3:477–478
- lithic technology, Asia
 after 300 ka, 1:98, 1:101f
 Early Pleistocene, 1:67–68, 1:70f
 Middle Pleistocene, 1:72f, 1:73f
see also stone tools
- lithified till fabric analysis, 2:76
- lithofacies, glacial, 2:18–29
- lithology
 archives, lake levels, 2:489–495
 changes
 lake-level reconstruction, 2:549
 particle forms, sediments, 2:3
 lithostratigraphy and, 4:227–228
 pollen databases, 3:834
- lithosphere
 nuclear disintegration rates, 1:426–427
 nuclide production rates at different altitudes, 1:428–429
 depths, 1:428
 research history, 1:20–21
 soil studies, 1:20–21
- lithostratigraphic units
 defining, 4:228–229, 4:230f
 description of, 4:233t, 4:234
 requirements, 4:233t
 splitting/lumping, 4:235–238
- lithostratigraphy, 4:227–242
 classification, 4:227–234
 definition, 4:227
 hierarchic structure, 4:227–228, 4:235–238
 nomenclature, 4:228–234, 4:231f
 Quaternary record examples, 4:235–240
 rules of application, 4:228–234
 sedimentary indicators, RSL change, 4:398, 4:400, 4:400f
- litopterna, South America, 4:686–687
- Little Ice Age (LIA)
 African ice core records, 2:374–375
 Arctic paleoclimate history, 3:120, 3:121, 3:122
 beetle records, 1:276
 climatic variability, Andes, 2:392
 definition, 2:216ge
 dendroglaciology, 2:105f, 2:106f
 glacier recession, 1:804, 1:807
 glacier variation patterns, 2:274
 last millennium climate, 3:229
 modern analog approaches,
 paleoclimatology, 3:103, 3:104, 3:109
 neoglaciation in Europe, 2:257, 2:262–263
 recognition of, 2:269
 volcanic/solar forcing, 1:752
- Little Molas Lake, North America, 2:541–543
- littoral pioneer plants, 3:742
- littoral zone, definition, 1:361ge
- Liucheng cave, China, 4:655–657
- lizards, 4:630
see also Komodo Dragons
- LM *see* light microscopy; linear modulation
- LMAs *see* Land Mammal Ages
- loading, periglacial, 3:430–432, 3:431f, 3:432f, 3:433f
- Loas, Early Middle Pleistocene vertebrate records, 4:657, 4:660t

- lobes
 abandonment, fan surfaces, 1:435–436
 felsic, 1:796
 solifluction, 3:483–484, 3:485f
 local extinctions, Late Pleistocene, 4:697
 local flooding surface (LFS), sequence stratigraphy, 4:262–263
 local hydrology, lake water, 1:336–337
 local scale studies, paleoclimatology, 3:104–107
 local sea-level change, 4:371f, 4:372–375
 Local Vegetation Estimates (LOVE) model, 3:892
 applications, 3:895–901
 pollen analysis, 3:798–799
 POLLSCAPE, 3:874–875, 3:877–878
 location attributes, sea-level index points, 4:371
 lodgment, till, 2:62, 2:63f, 2:65–66, 2:66–67, 2:78
 lodgment–deformation continuum, till, 2:65–66
 loess, 3:359, 3:362f
 Central Asia, 1:613f, 2:585–594
 China, 1:607f, 1:609–610, 1:610f, 2:595–605
 pedostratigraphy, 4:254–255, 4:255f
 stratigraphy overview, 4:199–200, 4:200f
 collapsibility, 2:594
 deep-sea sediments, 1:637–642
 definition, 2:573, 3:375ge
 dune fields, 1:607f, 1:609–610, 1:610f, 1:613f, 1:614f, 1:615f
 Europe, 2:606–619
 magnetic susceptibility, 3:378–379
 micromorphology, 3:371
 mineralogy/geochemistry, 2:574
 North America, 1:615f, 2:620–628
 origins, 2:573–584, 2:585–587
 overview, 3:357
 paleoclimatic studies, 3:359, 3:366
 paleopedology, 3:369f
 pedostratigraphy, 4:254–255, 4:255–256
 properties, 2:573–584
 sequences
 paleoenvironmental information from, 2:580–582
 paleosol sequences, 3:367
 South America, 1:614f, 2:629–635
 spatial distribution, 2:573–574, 2:574f, 2:575f
 stratigraphy, 2:577–579, 2:577f, 2:582f, 4:199–200, 4:200f
 world distribution, 3:361f
 Yedoma of Beringia, 3:548, 3:549–550
 loessification hypothesis, 2:585–587
 loess-like deposits, South America, 2:629–635
 loess–paleosol sequences, 3:367
 Loess Plateau, China, 4:254–255, 4:255f
 LO event *see* last occurrence event
 London, England beetle assemblages, 1:188
 Longgupo cave, China, 4:655
 longitudinal dunes, 1:599
 longitudinal profiles, fluvial environments
 allogenic forcing response, 1:676
 case studies, 1:679–682
 incision/aggradation, 1:677–678
 long pollen records, South America, 4:52–54
 long profiles, glacial landforms, 1:852–854, 1:855f, 1:856f, 1:859f
 long-range-transported (LRT) dust, 3:357, 3:360
 long-term exhumation, glacial landforms, 1:860
 long-term incision, strath terraces, 1:686–688, 1:688–689
 long-term rates, erosion, glacial landforms, 1:860–861
 long-term records, pigment studies, 3:336, 3:337f
 long-term sediment transport, 3:253–255
 long-term storage, oxygen isotopes, 2:917–918
 long-term variability, coral Sr/Ca, 2:880–881
 loss of plant macrofossils, 3:687
 Loveland Loess, North America, 2:620–621, 2:624f
 LOVE model *see* Local Vegetation Estimates model
 low-energy coast sea-level indicators, 4:385
 Lower Ermakovski Stage, Northern Asia, 4:36–37
 Lower Mystic Lake, Massachusetts, 3:213
 Lower Paleolithic
 archeological beetle fauna, 1:187
 hominid–animal interactions, 4:748–749, 4:750f
 Low lake, Canada, 3:345–346
 low latitudes
 ice core cosmogenic radionuclides, 2:358
 sand seas, 1:623–636
 vegetation patterns
 LGM, 3:864
 mid-Holocene, 3:864–865
 lowstand, sea-levels, 4:464, 4:498–499, 4:500f
 LP/EH boundary *see* Late Pleistocene/Early Holocene boundary
 LRA *see* Landscape Reconstruction Algorithm
 LRT dust *see* long-range-transported dust
 LSC *see* liquid scintillation counting
 LSW *see* Labrador Sea Water
 Lucanidae beetles, 1:213
 luminescence dating, 1:7, 1:108ge
 Australian lake level studies, 2:527–528
 definition, 1:108ge
 eolian sand dating, 3:357
 ESR dating, 2:667–677
 loess deposit dating, 2:573, 2:577
 methodology, 1:12–13, 1:12f
 OSL dating, 2:653–666
 radiative dosimetry dating methods, 1:449–450
 sand sea dating, 1:627, 1:634f
 tephrochronology, 4:289
 thermoluminescence, 2:643–652
 lumping, lithostratigraphic units, 4:235–238
 lunettes, lake level studies, 2:526–527
 LW *see* Late Weichselian
 Lyadhej-To Lake, Ural Mountains, 4:164, 4:165f
 Lyall, Charles, 1:1, 1:2f, 1:4–5
 Lyndon Stream, New Zealand
 beetle forest refugia, 1:251–252
 eLGM climate reconstruction, 1:249–250, 1:250f
 lysocline, 2:863–864, 2:865f, 2:869f
- ## M
- MAAT *see* mean annual air temperature
 McCall Glacier, Alaska, 1:804f
 McGregor Lake, Canada, 1:818f
 Macquarie Island glaciers, 2:213, 2:214
 macrauchenids, 4:686
 MacRobertson Land, East Antarctica, 2:220
 macrofabric analysis, till, 2:76–80
 macrofossils
 Holocene record, 3:756–757, 3:768–784
 lake-levels
 changes, 3:665–669
 reconstructions, 2:491, 2:493
 North America
 Arctic, 3:756–757
 Holocene record, 3:756–757, 3:768–784
 site summary, 3:746–757
 paleolimnological change causation, 3:670–672
 production, 3:686–687
 site summary, North America, 3:746–757
 stand-scale palynology, 3:851f, 3:852–853, 3:852f
 treeline reconstructions, 3:690–692
see also plant macrofossils
 macrogelivation
 block streams, 3:515, 3:518
 sediment particle forms, 2:3
 macrophytes
 distribution in lakes, 2:491, 3:657
 lake ecosystem role, 3:657
 lake-level reconstructions, 2:491
 long-term development, 3:666f
 macropodids (kangaroos), 4:634–636
 macroscopic charcoal, 2:720
 macroscopic plant remains, 4:542–547
 Madagascar
 hippopotamuses, 4:736, 4:742
 vertebrate fossil records, 4:670
 Maebashi Peat Formation, Japan, 1:207, 1:211f
 mafic compositions, 1:781–792, 1:780ge
 mafic tuyas, 1:781–792, 1:793f, 1:800f
Magalonyx jeffersonii *see* Jefferson's ground sloth
 Magdalenian industry
 definition, 4:748ge
 hominid–animal interactions, 4:753f
 maghemite, 3:376, 3:377–378
 Maghreb, modern human expansion, 1:96–97
 magnesium (Mg)/calcium (Ca)
 paleothermometry, 2:871–883
 salinity proxies, 2:936–937, 2:937–938, 2:937f, 2:938–939

- magnetic poles, 1:706–707
- magnetic susceptibility (MS), 2:884–898
- common materials, 2:887*t*
 - definition, 3:375*ge*, 2:884
 - environmental magnetism, 2:884
 - environmental susceptibility records, 2:888
 - lake sediment physical properties, 3:303–304, 3:305, 3:305*f*
 - loess records
 - China, 2:595, 2:598*f*, 2:599–600, 2:602*f*, 2:603*f*
 - South America, 2:632–634
 - measurement, 2:885–886, 2:886–888, 3:376
 - minerals, 2:887*t*
 - paleoceanography, 2:884–898
 - parameters, 2:885*t*
 - postdepositional records, 2:895–897
 - as proxy
 - paleoceanographic, 2:888–897
 - sediment redistribution, 2:891–893
 - terrigenous flux, 2:889–891
 - relative behaviour strengths, 2:886*f*
 - soils, 3:377–379
 - source variation, 2:893–895
 - values, influences on, 2:884–885
- magnetism
- definition, 3:375–376
 - environmental, 1:923–924
 - interpretation of properties, 3:377–379
 - lake sediment properties, 3:303–305, 3:304*f*, 3:305*f*
 - loess deposits, 2:581
 - measuring, 3:376–377
 - mineral analysis, 3:375–380
 - North American glaciations, 2:182–185
 - paleosol proxy data, 3:373–374
 - see also* paleomagnetism
- magnetite, 3:376, 3:377–378
- magnetostratigraphy
- dating methods, 2:182, 2:185
 - marine oxygen isotopes, 2:909–910
 - overview, 4:190
 - South American loess records, 2:632, 2:634–635, 2:635*t*
- magnolia DNA, 2:710*f*
- Mahoma Lake, Africa, 4:97–98, 4:99*f*
- maize DNA, 2:708*f*
- major ions, ice cores, 2:326–333, 2:379, 2:383, 2:384, 2:385
- major scale glacial landforms, 1:847–864
- Majuangou site, Asia, 1:67–68
- Malan Loess, China, 2:595–596, 2:597–598, 2:600*f*
- Malawi, Lake, Africa, 1:546–547, 1:552
- Malaysian sea-level changes, 4:444
- Maliang site, Asia, 1:76–77
- MAM *see* minimum age model
- mammals
- Australia, 4:625*t*, 4:631–637, 4:631*f*, 4:632*f*, 4:633*f*, 4:634*f*, 4:635*f*
 - bioturbation, 2:632, 2:634*f*
 - biozonations, 4:210–213, 4:213*f*
 - continental biostratigraphy, 4:206–214
 - Early Middle Pleistocene, 4:605–614
 - Early Pleistocene records, 4:599–604
 - evolution of species, 4:206–207, 4:207–208, 4:723–732
 - extinctions, 4:208
 - Africa, 4:664–665, 4:668–669
 - hominin interactions, 4:596
 - Late Pleistocene, 4:697
 - North America, 4:676–678
 - South America, 4:682*t*, 4:690*t*
 - fossil records, 4:590–597
 - Late Pleistocene
 - Africa, 4:664–672
 - mummified mammals, 4:713–717
 - North America, 4:673–679
 - South America, 4:681*t*, 4:682*t*, 4:684–690, 4:690*t*
 - Southeast Asia, 4:693–699
 - Middle Pleistocene
 - Africa, 4:615–619
 - Australia, 4:625*t*, 4:631–637, 4:631*f*, 4:632*f*, 4:633*f*, 4:634*f*, 4:635*f*
 - Early Middle, 4:605–614
 - Europe, 4:639–645
 - Southern Asia, 4:651–663
 - migration/dispersal, 4:208–209
 - mummified mammals, 4:713–717
 - North America
 - extinctions, 4:676–678
 - Late Pleistocene, 4:673–679
 - Mid-Pleistocene vertebrate records, 4:646–647
 - size change rule, 4:697–698
 - small mammals, dwarfing/gigantism, 4:743–745
 - South America
 - extinctions, 4:682*t*, 4:690*t*
 - Late Pleistocene, 4:681*t*, 4:682*t*, 4:684–690, 4:690*t*
 - loess records, 2:632, 2:634*f*
 - speciation trends, 4:723–732
 - teeth/bones, carbonate stable isotopes, 1:308, 1:309*f*
 - see also individual mammals*
- mammoth (*Mammuthus*)
- Arctic Eurasia, 3:741–742
 - Coelodonta* faunal complex, 4:641, 4:644
 - dwarfing/gigantism, 4:734–736, 4:734*f*
 - evolution of, 4:207, 4:727–729, 4:731
 - Irvingtonian age, North America, 4:647–648
 - Pleistocene, 4:207
 - Eurasia, 4:600–601
 - extinctions, 4:700–704, 4:707–708, 4:708–709
 - North America, 4:677*f*
 - vertebrate records, 4:647–648
 - species/subspecies origins, 4:725
 - see also* woolly mammoth
- Mammuthus-Coelodonta* faunal complex, 4:641, 4:644
- Mammuthus primigenius* *see* woolly mammoth
- manganese coatings, soil, 3:383
- Mangla-Samwal anticline, Pakistan, 4:651, 4:653*t*
- mangroves/mangrove swamps
- Amazon Rainforest, 4:156–157
 - relative sea level changes, 4:397, 4:399*f*, 4:495–496, 4:496*f*, 4:497*f*
 - sedimentary RSL change indicators, 4:397, 4:399*f*
- Maninjau volcanic edifice, Indonesia, 1:656
- Mann, ME, 3:245–246, 3:248–249
- MAP *see* mean annual precipitation
- mapped networks, plant macrofossils, 3:731
- mapping
- forensic geology searches, 4:530
 - lithostratigraphic, 4:227, 4:229–232, 4:235–238
 - plant macrofossil networks, 3:731
 - pollen data, 3:836–837, 3:837*f*, 3:862–863
 - South Island, New Zealand, 2:211*f*
 - vegetation reconstruction, 3:861
- MAR *see* mass accumulation rate
- March rotation, till fabric, 2:78–79, 2:79*f*
- margin ice, 2:416–430
- Marguerite Bay, Antarctica, 4:475*f*, 4:478*f*, 4:479
- marine archeology, benthic foraminifera, 2:772
- marine-based ice sheets
- collapse theories, 2:449
 - growth/decay, 1:878–879
 - West Antarctic, 2:448–455
- marine biota feedback, LGM, 1:726–727
- marine biozonation, 4:213
- marine carbon cycle proxies, 2:849–858
- alkenones, 2:855
 - Ba/Ca proxy, 2:854
 - B/Ca proxy, 2:851–853, 2:851*f*, 2:852*f*, 2:854
 - boron isotope-pH proxy, 2:849–851, 2:850*f*
 - carbon isotope fractionation, 2:855
 - foraminiferal Ba/Ca, 2:854
 - foraminiferal U/Ca, 2:852*f*, 2:853
 - pCO₂, 2:849, 2:851*f*, 2:855, 2:856*f*
 - SNWs, 2:855–857
 - stable isotopes, DIC in seawater, 2:854–855, 2:855*f*
 - U/Ca proxy, 2:852*f*, 2:853
 - Zn/Ca in benthic foraminifera, 2:853–854, 2:853*f*
- marine $\delta^{18}\text{O}$ record, definition, 2:216*ge*
- see also* diatom silica
- marine depositional systems, 2:24–28
- marine diatoms, 2:216*ge*, 2:816–824
- Antarctica, 1:527–539
 - applications, 1:528–536
 - biological proxies, 2:816–824
 - biostratigraphy, 2:824
 - as climate tracers, 2:823
 - colony types, 2:817*f*
 - habitats, 1:472–473
 - laminated sediments, 1:554–561
 - accumulation sites, 1:554–555
 - historical perspective, 1:554
 - oceanic processes, 2:820–823
 - centennial/decadal variability, 2:822
 - general circulation, 2:823
 - glacial/interglacial variability, 2:820–822
 - lateral transport/displacement, 2:823

- millennial variability, 2:822
 monospecific mats, 2:822–823, 2:822f
 productivity, 2:820–823
 sea ice dynamics, 2:823
 paleoecology, 2:816–823
 paleoenvironmental reconstruction, 2:817–820
 BSEI, 2:820
 isotopic approaches, 2:819
 molecular marker techniques, 2:819–820
 quantitative methods, 2:817–819
 trace metals, 2:820
 valve size, 2:819
 marine dinocysts, 2:805–811, 2:811–812
 marine ecosystem history, 1:19–20
 marine fossil paleothermometry, 2:871–883
 marine habitats, diatoms, 1:472–473
 marine isotope stages (MIS)
 2, 1:203
 Alaska/Yukon pollen records, 4:39–40, 4:42f
 Arctic, 3:118, 3:755–756
 New Zealand eLGM, 1:248–249
 North America
 Arctic, 3:755–756
 ice sheets, 2:247–249
 pollen records, 4:77–82
 Sartanskaya Stade, 4:27–34
 3, 1:202–203, 1:271, 4:34–36
 Alaska/Yukon pollen records, 4:42f
 Arctic, 3:117–118, 3:755
 Asia, 1:271, 2:515, 2:519
 Beringia, 2:192
 New Zealand, 1:247–248
 North America
 Arctic, 3:755
 ice sheets, 2:247
 pollen records, 4:75–77
 Northern Asia, 1:271
 4, 1:202
 Alaska/Yukon pollen records, 4:39
 Beringia, 2:193, 2:194–197
 North America
 ice sheets, 2:247
 pollen records, 4:75–77
 5
 Alaska/Yukon pollen records, 4:39, 4:41f
 Arctic Eurasia, 3:739–740
 Beringia, 2:193–194
 New Zealand, 1:247
 Northern Asia, 4:36–37
 North Pacific Ocean, 3:38–40
 WNA pollen records, 4:74–75, 4:78f
 5a, Odderade Interstadial, 1:202
 5b, 1:202, 2:245–247
 5c, Brørup Interstadial, 1:200–202
 5d, 1:200, 2:245–247
 5e, 1:200, 3:117, 4:483
 6
 Cold Stage correlation with, 1:189
 New Zealand, 1:247
 7, interglacial correlations, 1:188–189
 9, interglacial correlations, 1:188
- 11
 Arctic, 3:116–117
 interglacial correlations, 1:187–188
 12, Anglian glaciation, 1:186–187
 25, Middle Pleistocene, 4:639
 Anglian glaciation, 1:186–187
 Brørup Interstadial, 1:200–202
 chronostratigraphy, 4:220
 Cold Stage correlation, MIS 6, 1:189
 Eemian Interglacial, 1:200
 eLGM, New Zealand, 1:248–249
 Eurasian glaciations, 2:224
 European pollen records, 4:3–4, 4:4–5, 4:5–6
 Greenland sea-level changes, 4:483
 Herning Stadial, 1:200
 insect fossil correlations, 1:184, 1:186–187, 1:187–188, 1:188–189, 1:200–205, 1:247–249
 interglacial correlations
 MIS 7, 1:188–189
 MIS 9, 1:188
 MIS 11, 1:187–188
 last glacial maximum, 1:203
 last interglacial, 1:247, 3:739–740
 magnetic susceptibility, 2:896f
 Middle-Pleistocene glaciation
 Eurasia, 2:172, 2:173
 North America, 2:180–186
 Southern Hemisphere, 2:187
 Middle Weichselian
 MIS 3, 1:202–203, 1:271
 Stadials/Interstadials MIS 4, 1:202
 North American beetle studies, 1:221–225, 1:231, 1:232
 Odderade Interstadial, 1:202
 oxygen-isotope stages, 2:216ge, 2:596
 penultimate glaciation, New Zealand, 1:247
 pre-eLGM late, New Zealand, 1:247–248
 Rederstal Stadial, 1:202
 stratigraphy overview, 4:195–201
 marine limit
 deglaciation, Antarctica, 2:219
 RSL, high latitudes, 4:468–469, 4:471f
 marine notch sea level indicators, 4:388
 marine oxygen isotopes, sea-level change, 4:431–432, 4:434–435, 4:436, 4:436f
 marine oxygen-isotope stages
 definition, 2:216ge
 loess records, 2:596
 marine pollen sequences, Africa, 4:85–88, 4:88–89
 marine reconstructions, $\delta^{18}\text{O}_{\text{diatom}}$ records, 1:486–487
 marine records
 calcareous fossils, 2:871–883
 eustatic sea-level change, 4:431–432, 4:434–435, 4:436, 4:436f
 last glacial maximum Antarctica, 2:218
 mid-Quaternary North America, 2:180–186
 sea-level change overview, 4:369–370
 marine resources, human adaptations, 1:156–157
- marine sediments
 annual layer varve dating methods, 4:582–589
 carbon-14 record calibration, 4:346–347
 diatom silica applications, 2:738–739
 geomagnetic PSV, 1:714–716
 pollen analysis, 4:200–201
 protactinium-231 radioisotope proxies, 2:927–929, 2:929–930
 thorium-230 radioisotope proxies, 2:925–927, 2:927–929, 2:929–930
 see also glacial marine sediments; marine diatoms
 marine submergence, lower bounds, 4:462–463
 marine terraces
 loess deposits, 2:580f
 sea level indicators, 4:379f, 4:381–382, 4:381f, 4:507f, 4:509f, 4:513f
 marine-terrestrial stratigraphic correlation
 Chinese loess, 4:199–200, 4:200f
 Matuyama/Brunhes boundary, 4:197–198, 4:198f
 overview, 4:195–201
 pollen analysis, 4:200–201
 Rangitawa Tephra, 4:196–197, 4:197f
 tektites, 4:198–199, 4:199f
 Wanganui Basin, 4:192, 4:195–196, 4:196f
 marine varves
 diatom records, 1:554–561
 intra- to interannual diatom productivity, 1:556–559
 types, 1:557f
 maritime human adaptations, 1:156–157
 markers
 chronostratigraphy, 4:218–220
 European loess records, 2:609, 2:612
 see also biomarkers; stratigraphic markers
 Murray–Darling Basin, Australia, 2:526–527, 2:529
 MARs *see* mass accumulation rates
 Mars, volcanic landforms, 1:781
 marshes
 elevation diagrams, 4:405, 4:406f
 paleotempestology, 3:211–213
 marsupial lions, 4:634
 Marsworth, England beetle assemblages, 1:188
 Martin, Paul, 4:709
 mass accumulation rates (MARs)
 deep-sea sediments, 1:638, 1:640
 thorium-230, marine sediments, 2:926–927, 2:926f
 mass balance
 amplitude, definition, 1:909ge
 cumulative, 1:909ge
 definition, 1:909ge
 glacial landforms, 1:910, 2:109
 glaciers
 growth/decay, ice sheets, 1:877–878, 1:879f, 1:881
 recession, 1:803, 1:804
 gradients, definition, 1:909ge
 Greenland ice sheet, 2:440–441, 2:446
 ratio, definition, 1:909ge
 mass conservation, 4:453–455

- massive ice, permafrost, 3:467, 3:467f, 3:574–575
- mass movement
- active layer, permafrost, 3:424
 - slope deposits/forms, 3:481–484, 3:484–485
 - thermokarst topography, 3:580
- mass spectrometry (MS)
- alkenone paleothermometry, 2:756
 - delta-carbon-14 calculation, 4:337
 - diatom silica analysis, 2:735
 - pigment analysis, 3:330
 - soil/sediment in forensics, 4:540
- mass transport, rock glaciers, 3:537–538
- master chronology, dendrochronology, 1:457
- master horizons, soils, 3:402
- MAT *see* modern analog technique
- Matayama Chron, Early Quaternary, 2:168, 2:169–170, 2:169f
- mat-forming diatoms, 1:555, 1:556–557, 1:560, 2:822–823, 2:822f
- Mather, JR, 3:102–103
- Mathon, England beetle assemblages, 1:186
- matrix geochemistry, blockfields, 3:526–528
- mats, marine diatoms, 1:555, 1:556–557, 1:560, 2:822–823, 2:822f
- Matterhorn, Switzerland, 1:887f
- Matthes, Francois, 2:270f
- Matuyama–Brunhes boundary (MBB), 4:639, 4:643
- stratigraphy overview, 4:197–198, 4:198f
 - vole assemblage, 4:641
- Matuyama–Brunhes reversal, geomagnetic excursions, 1:716–718
- Maunder Minimum, volcanic/solar forcing, 1:748–749, 1:752, 1:752f
- Mauritanian sand seas, 1:625–626, 1:628f
- maximum elevation of lateral moraines (MELM), 1:913, 1:915f
- maximum latewood density (MXD), 1:459, 1:461, 1:464f
- maximum likelihood, definition, 1:361ge
- maximum likelihood envelope (MLE)
- method, paleoclimate reconstruction, 1:191–192, 1:249–250, 1:251f
- Mayotte, sea-level changes, 4:445–446
- Mazama, Mount, eruption, 4:129
- MBB *see* Matuyama–Brunhes boundary
- MBT/CBT temperature proxy, biomarkers, 2:776–778
- MC-ICP-MS *see* multicollector inductively-coupled plasma mass spectrometry
- MCR *see* mutual climatic range
- meadows
- Arctic Eurasia, 3:742
 - pollen records, 3:888f, 3:890–891
- mean annual air temperature (MAAT), ice-wedge pseudomorphs, 3:447–448, 3:449
- mean annual precipitation (MAP), North America, 1:227
- measurement year, mass balance, 1:909ge
- mechanical weathering
- blockfields, 3:525
 - profiles, 3:392
- medial moraines, 1:772–774
- median elevation of glaciers (MEG), 1:914
- medicolegal aspects, forensics, 4:548–555
- medieval fish ponds, diatom analysis, 1:518
- Medieval Warm Period
- Arctic paleoclimate history, 3:120–121, 3:121–122
 - Greenland borehole temperature records, 2:300–301
 - last millennium climate, 3:229–230
- Mediterranean
- climate, pollen records, 4:14–15, 4:66–67, 4:179, 4:183, 4:186
 - deer, dwarfing/gigantism, 4:737–739, 4:738f, 4:739f
 - Early Pleistocene, 4:606f
 - mountains, beetle fauna, 1:278–280
 - sub-Milankovitch events, 3:205
 - vegetation, 4:183, 4:185–186, 4:187–188
- mediterranean outflow water (MOW), 3:205
- MEG *see* median elevation of glaciers
- megablock structures/landforms, 1:844–845
- megafauna
- Australia, 1:113–115, 4:112
 - definition, 4:700
 - extinctions
 - Americas, 1:126–127
 - Australia, 1:113–115
 - biosphere history, 1:23–24
 - Late Pleistocene, 4:690t, 4:700–712
- mega-floods, LGM, 1:581–585
- megafutes, meltwater hypothesis, 1:835–836
- megafossils, 3:621–629
- climatic controls, 3:621–622
 - Fennoscandia-Kola Peninsula, 3:623f, 3:625–626, 3:625f, 3:627f
 - Holocene treeline dynamics, 3:693–694
 - macrofossil comparisons, 3:690, 3:746
 - Northern Europe, 3:626–629
 - Northern Quebec, 3:624–625
 - regional studies, 3:624–629
 - treelines, 3:621–622, 3:622–624, 3:628f, 3:693–694
 - Tuktoyaktuk Peninsula, 3:624, 3:626f, 3:626t
 - vegetation, 3:621–622
 - wood megafossils, 3:622–624
- megalakes, Australia, 2:528
- Meiendorf Interstadial deglaciation, 2:232
- Melanesia, human chronology, 1:111
- melanin, 3:276f, 3:278f, 3:279, 3:271ge
- MELM *see* maximum elevation of lateral moraines
- melting processes *see* thaw processes
- melt-out of ice, rock glaciers, 3:537–538
- melt-out tills
- fabric analysis, 2:78
 - subglacial, 2:62, 2:63f, 2:67, 2:67f, 2:73
- meltwater
- active temperate glacier margins, 1:759
 - channels
 - glacifluvial landforms, 1:832–833
 - permafrost–glacier interactions, 3:509 - $\delta^{18}\text{O}_{\text{diatom}}$ records, 1:487
 - Late Pleistocene South Atlantic, 3:24
- meltwater hypothesis, 1:833–836
- bedrock erosional forms, 1:833–835
- drumlins/rogens/hummocky terrain/megafutes/tunnel valleys, 1:835–836
- overview, 1:833
- see also* glacifluvial landforms
- meltwater pulses (MWPs), sea-level change, 4:373–375
- members in lithostratigraphy, definition, 4:228
- Menapian stage mammal assemblages, 4:608–610, 4:610–611
- Menyanthes trifoliata* macrofossils, 3:642, 3:643f
- meridional overturning circulation (MOC)
- environment modeling, 3:150–152
 - Last Glacial Maximum GCMs, 3:168–169
 - North Atlantic Ocean, 3:9, 3:12
 - paleoceanography overview, 2:749
 - radioisotope proxies, 2:927, 2:929f
 - THC, 1:737, 1:739f, 1:740–742, 1:742f, 1:743, 1:745
- meromictic lakes, 1:361ge
- Mersea Island, England beetle assemblages, 1:188
- mesic tundra, 1:233
- Mesolithic archeology, 1:106f, 1:154–159
- see also* Early Stone Age; Late Stone Age; Middle Stone Age; Neolithic; Paleolithic...
- mesophilous plants, 4:183, 4:185, 4:186
- mesosaline, definition, 1:361ge
- mesotrophic lakes, 3:657
- metabolic activity, ice-bound
- microorganisms, 2:294–296
- metabolism, glucose, 1:315–321
- metals
- pollution, 2:344–345, 3:266, 3:268
 - trace metals
 - lake chemistry, 3:296–297
 - marine diatom proxies, 2:820
- meteoric carbonates, *see also* surficial meteoric carbonates
- meteoric water, speleothem formation, 1:294
- meteorites
- chronostratigraphy, 4:218–220
 - ice-margin sites, 2:416–417
- meteorology
- equatorial East Africa, 2:373
 - tropical South America, 2:387–390
 - see also* climate...
- methane (CH_4), 2:334–341
- analytical techniques, 2:340
 - atmospheric, 2:335–337, 2:336f
 - biosphere feedbacks, LGM, 1:727
 - cycle, 2:334
 - glacial–interglacial cycles, 2:339–340
 - Holocene, 2:335–338, 2:337f
 - ice cores, 2:334–341
 - correlation techniques, 2:410–411, 2:411f, 2:412, 2:413, 2:413f - ice-margin sites, 2:426
 - last glaciation/deglaciation, 2:338–339
 - last millennium, 2:335
 - mixing ratios, 2:340, 2:340f
 - reconstruction of levels, 2:334–335
- methyl bromide (CH_3Br), 2:467–468

- methyl chloride (CH₃Cl), 2:466–467
 Meuse river, Netherlands, 4:235–236, 4:236–237, 4:236f
 Mexico
 central highland lake level studies, 2:533
 dendroclimatology, 1:461–462, 1:462f
 Gulf of
 deltas, 1:700, 1:701f, 2:123f
 fluvial sediments, 1:673f
 RLB fauna, 4:678
 Mg/Ca *see* magnesium/calcium
 microatolls, 4:382–383, 4:382f, 4:383f, 4:415f, 4:416–417, 4:417f, 4:505f, 4:506f
 microcephaly, *Homo floresiensis*, 1:143
 microcharcoal, 2:720
 microcracks, weathering profiles, 3:392
 microevolution, vertebrate studies, 4:729
 microfabric analysis, glacial sediments, 2:53f, 2:56–58
 microfauna, definition, 1:83ge
 microfossils
 diatom records
 freshwater laminated sequences, 1:540–545
 Pacific Ocean, 1:571–587
 dinoflagellates, 2:800–815
 Holocene RSL reconstructions, 4:419–428
 phytoliths, 3:582–592
 temperature proxies, 2:841, 2:843–844
 microgelivation, sediment particle forms, 2:3
 microlithic industries, South Asia, 1:99–100, 1:102f
 microliths, 1:158
 micromorphology
 glacial sediments, 2:52–61
 paleosols, 3:370, 3:371, 3:372
 reconstructions, landscape evolution, 3:388–389, 3:390f
 soil, 3:381–391
 microorganisms, ice cores, 2:289–291, 2:290f, 2:294–296, 2:295t
 micropaleontological evidence, North Pacific Ocean, 3:36
 microparticles
 aerosols and, 2:342–343
 fingerprinting, 2:343–344
 glacier ice, 2:343
 ice core studies, 2:342–346
 microremains, plants, 3:699
 microsampling nonmarine biogenic carbonates, 1:342–343, 1:342f, 1:349f
 micro/macroscale erosional forms, 1:865–876
 abrasion, 1:865–867, 1:868f, 1:869, 1:871–872
 applications in science, 1:873–876
 crag and tails, 1:873, 1:875f
 fracture, 1:867–869
 friction cracks, 1:870–871, 1:872f, 1:873–874
 fundamentals, 1:865–869
 p-forms, 1:871, 1:873f
 plucking/entrainment, 1:867, 1:869
 polished appearance, 1:871, 1:873f
 quarrying, 1:865, 1:867, 1:868f, 1:869, 1:870f, 1:871–872
 roches moutonnées, 1:870f, 1:871–872, 1:874–875
 rock drumlins, 1:872–873, 1:874f
 size classifications, 1:866t
 striae, 1:865–867, 1:866f, 1:870, 1:871–872, 1:871f, 1:873–874, 1:875f
 whalebacks, 1:872–873, 1:874–875, 1:874f
 microscopic studies
 glacial sediment micromorphology, 2:60–61
 soil/sediment in forensics, 4:538, 4:539–540
 varved sediments, 4:585
 microstratigraphy
 diatom analysis, 1:540–541
 relations, paleosols, 3:386–388
 microtomography, 2:60, 2:60f, 2:61
Microtus *see* voles
 microzooplanktonic remains, 2:812
 middens, rodents, 3:674–683, 4:143, 4:153f, 4:154
 Middle East
 lake level studies, 2:511
 Younger Dryas, 3:127
 see also Asia
 Middle Ermakovski *see* Mid-Ermakovski...
 Middle Galerian mammal assemblages, 4:611–612
 Middle Holocene
 African ice core records, 2:376–377
 Asian lake level studies, 2:513, 2:520
 glacier oscillations, 2:221
 paleoclimate data-model comparisons, 3:137, 3:139f, 3:141f, 3:142, 3:144f, 3:145–146
 vegetation patterns, 3:864–865
 Middle Paleolithic
 atmospheric carbon-14, 4:332–333
 hominid-animal interactions, 4:749–750
 human expansion, Africa, 1:94–95
 Levantine periodization, 1:99t
 lithic production after 300 ka, 1:98, 1:101f
 technocomplexes, 1:147, 1:148t
 see also Middle Stone Age
 Middle Pleistocene
 Arctic North America, 3:753
 Australian vertebrate records, 4:621–638
 beetle assemblages
 New Zealand, 1:244–247
 Western Europe, 1:184–190
 climatic background, 4:615–616
 glaciation
 Eurasia, 2:172–179
 North America, 2:180–186
 Southern Hemisphere, 2:187–190
 hominin archives, Asia, 1:71–77, 1:80
 North America
 Arctic, 3:753
 glaciation, 2:180–186
 vertebrate records, 4:646–650
 plant macrofossil records, 3:739
 Southern European pollen records, 4:63–71
 vertebrate studies
 Africa, 4:615–619, 4:664–665
 Australia, 4:621–638
 Europe, 4:639–645
 North America, 4:646–650
 Northern Eurasia, 4:605–614
 Southern Asia, 4:651–663
 Middle Pleistocene transition (MPT)
 Arctic, 1:563, 3:115–116
 definition, 1:83ge
 diatom studies, 1:563
 hominin colonization, 1:84–87
 North Atlantic Ocean, 1:563, 3:10
 Middle Quaternary *see* mid-Quaternary...
 Middle Saalian glaciation, Eurasia, 2:176
 Middle Stone Age (MSA)
 African archeological records, 1:91, 1:92–94, 1:94–95
 substage proposal, 1:93t
 see also Middle Paleolithic
 Middle Villafranchian mammal assemblages, 4:605–607
 Middle Weichselian beetle assemblages
 MIS 3, 1:202–203, 1:271
 MIS 4 Stadials/Interstadials, 1:202
 Middle Wisconsin cold interval/interstadial beetle studies, 1:221
 Beringia, 2:197
 ice sheets, 2:247
 Mid-Ermakovski Stage pollen records, 4:36–37
 midges
 forensic science use, 4:552
 Late Pleistocene Europe, 1:373–385
 overview, 1:355–360
 postglacial records
 Europe, 1:386–397
 Southern Hemisphere, 1:398–405
 see also *Chaoboridae*; chironomids
 mid-Holocene *see* Middle Holocene
 mid-infrared (MIR) region, 3:349
 statistical approaches, 3:351
 theoretical background, 3:349–350, 3:350f
 mid-infrared spectroscopy (MIRS), 3:353–354
 Midlands, England beetle assemblages, 1:187
 mid-late Holocene *see* Late Holocene; Middle Holocene
 midlatitude dune fields, 1:606–622
 Central Asia, 1:611–614, 1:613f
 China, 1:606–611
 desert dunes, 1:609–610, 1:610–611
 Holocene optimum, 1:610–611, 1:612f
 LGM, 1:610, 1:612f
 loess deposits, 1:607f, 1:609–610, 1:610f
 origins, 1:609
 types, 1:606–608
 North America, 1:614–621
 ages, 1:616–617
 forms of, 1:615
 origins, 1:615–616
 paleoclimatic significance, 1:618–621
 South America, 1:611–614, 1:614f

- midlatitude hard-water lakes, 1:334
midlatitude sea-level changes, Late Quaternary, 4:489–494
Mid-Pleistocene... *see* Middle Pleistocene...
mid-Quaternary glaciation, North America, 2:180–186
Midwestern US plant macrofossil records, 3:768–784
Mid-Wisconsin *see* Middle Wisconsin
migration of species
 beetles, South America, 1:241
 human expansion, Americas, 1:121–122
 mammals, 4:208–209
 vegetation, 3:866–867, 4:116, 4:119
 see also dispersal of species
Milankovitch cycles, 2:307–308
 see also sub-Milankovitch...
Milankovitch, Milutin, 1:5–6, 1:6f, 1:26–27, 2:136–137, 2:138–139
Milankovitch theory
 climatic change, 1:26–27
 glaciation, 1:5–6
 sub-Milankovitch variations, 1:27–29
 see also Milankovitch cycles
Milankovitch variability, definition, 1:637*ge*
millennial-scale changes
 climate change, sub-Milankovitch events, 3:206–207
 lake levels, Africa, 2:499–502
 ventilation, North Pacific Ocean, 3:36–37
 dynamics, paleoceanographic records, 3:6–7
 variability
 Andean glacier climate records, 2:390–391
 Antarctica, 2:398–400
 North Atlantic Ocean, 3:12–15
 North Pacific Ocean, 3:40–42, 3:43–44
 West Antarctic Ice Sheet timescale, 2:452–453
 see also last millennium climate
million years (My) measure, speleothems, 1:292
Miomys–Arvicola lineage
 biochronology, 4:639, 4:643
 evolutionary trends, 4:725–727
mineralization, plant macrofossils, 3:702, 3:708f
mineralogy/minerals
 atmospheric dust composition, 1:734
 clay, trimlines/paleonunataks, 1:923
 cosmic ray interactions, 1:418–431
 last glacial cycle, 2:635–637, 2:638*t*
 loess, 2:574
 European records, 2:606
 South American records, 2:635–637, 2:638*t*
 magnetic analysis, 3:375–380
 palsas, 3:477–478
 tephrochronology, 4:283–286
 terrigenous sediments, 2:943, 2:948, 2:950f
 TL dating, 2:643–652
 trimlines/paleonunataks, 1:923
 mineral palsas, 3:477–478
 minimum age model (MAM), OSL dating, 2:661
 mining fraud evidence, 4:535
 Minnesota, US plant macrofossil records, 3:775–777, 3:778f
 Miocene epoch, definition, 2:216*ge*
 Miomote Site Group, Japan, beetle fossils, 1:207–210
 mire/peat macrofossils, 3:609, 3:637–656
 Acari, 3:652, 3:653f
 applications, 3:637
 Bryozoa, 3:652, 3:653f
 Characeae, 3:646, 3:647f
 chironomids, 3:652, 3:653f
 foraminifera, 3:652, 3:653f
 freshwater sponges, 3:652, 3:653f
 fungi, 3:652, 3:653f
 indicator value, 3:638–652
 methods, 3:637–638
 Northwest Europe, 3:638–652
 Nymphaeaceae, 3:645f, 3:646
 ostracods, 3:652, 3:653f
 tree macrofossils, 3:642–646
 Trichoptera, 3:652, 3:653f
 MIR region *see* mid-infrared region
 MIRS *see* mid-infrared spectroscopy
 MIS *see* marine isotope stages
 Mississippi River, 1:666f
 avulsion deposits, 1:668–670
 floodplain, 1:664f, 1:665f, 1:668f, 1:669f
 natural-levee sediments, 1:665, 1:665f, 1:669f
 Missouri River, floodplain, 1:666f
 Missouri, US plant macrofossil records, 3:782f, 3:783
 Mitchell, JM Jr, 3:94
 mites *see* oribatid mites
 mitochondrial DNA (mtDNA), 4:720
 mitochondrial Eve, human evolution, 1:141
 mixed-feeders
 definition, 4:680*ge*
 Late Pleistocene South America, 4:686, 4:689
 Miyanomae Site, Japan beetle fossils, 1:211
 Mizozono Formation, Japan, 1:211, 1:212f
 MLE method *see* maximum likelihood
 envelope method
 MN17 stage mammal assemblages, Northern Eurasia, 4:605–607
 MN18–MQ19 (Tiglian–Menapian) stage mammal assemblages, 4:608–610
 moats, diatom analysis, 1:518, 1:518f
 MOC *see* meridional overturning circulation
 Mock, CJ, 3:106–107, 3:109
 modeling
 conceptual geological models, 4:525–527, 4:527f
 delta studies, 1:700–702
 fossil evidence, 3:697
 geological models, 4:525–527, 4:527f
 ice core studies, 2:329
 lithostratigraphic, 4:235, 4:240
 modern human origins, 1:91–94, 1:146
 pollen representation, 3:871–879
 precipitation isotopic composition, 2:395–396
 sea-level change, 4:369, 4:370, 4:373, 4:374f, 4:375, 4:375f, 4:439, 4:440–441, 4:443–444, 4:447
 strategies, 2:560–561
 vegetation
 biome approach, 3:862
 human impact, 3:892–895
 vertebrate evolution, 4:723
 see also climate, modeling; *see also* individual models; inference models; lake levels, modeling; Quaternary environments, modeling; vegetation, modeling
modern analog approaches, paleoclimatology, 3:102–112
 circulation-to-environment, 3:107–109, 3:110
 environment-to-circulation, 3:102–107, 3:109–110
 flood analogs, 3:105–106
 local to regional scale studies, 3:104–107
 regional to hemispheric scale studies, 3:102–104
 synoptic analogs, 3:102, 3:103, 3:105, 3:106–107, 3:108–109, 3:109–110
modern analog technique (MAT)
 definition, 1:361*ge*
 diatoms, 1:496
 glacial landsystems, 1:813–817, 1:817–822
 planktic foraminifera, 2:828
 pollen–climate relation, 3:808–809
 temperature proxies, 2:841–842, 2:844*t*, 2:845, 2:846–847
 vegetation construction, 3:826–827
 see also analog technique
modern humans
 anatomical changes, 1:92
 behavioral changes, 1:92, 1:93
 colonization of new territories, 1:103–104
 expansion of, 1:94–97, 1:98–100, 1:100–103, 1:103–104, 1:104–106
 indigenous population replacement, 1:100–103
 Neanderthal demise and, 1:146, 1:150–152
 origins, 1:146
 geographical models, 1:91–94
 human revolution concept, 1:57
 vertebrate studies, 4:664
 species definition, 1:146–152
 technologies, 1:147–150, 1:148*t*
 see also *Homo sapiens*
modified bone, hominin behavior, 1:60, 1:62, 1:64–65
moisture proxies, oribatid mites, 2:693–694, 2:695
Mojave Desert
 postglacial pollen records, 4:142, 4:153f
 rodent middens, 3:678–679
 sand seas, 1:633–634
molecular markers, marine diatoms, 2:819–820
molecular-size fractionation, firn air, 2:368–369, 2:369f

- molecular studies, vertebrate evolution, 4:723
- Molinia caerulea* macrofossils, 3:642, 3:644f
- mollusks, 2:795–799
- amino-acid dating, 1:38, 1:42, 1:42f, 1:43–44, 1:45, 1:46, 1:47
- ESR dating, 2:668, 2:670–672, 2:673–676, 2:674f
- European loess records, 2:613–615, 2:615f
- freshwater, 1:343–346, 3:281–291
- paleolimnology, 3:264–265
- shells, 2:668, 2:670–672, 2:673–676, 2:674f, 4:194, 4:194f
- Mona moraine, Norway, 2:88f
- Money Creek, Minnesota, US, 3:775–777, 3:778f
- Mongolian pollen records, 4:170
- monogenetic formations, 1:791–792, 1:780ge, 1:793f
- monsoonal forcing, 1:680, 3:49–52
- monsoons
- Australian terminal lakes, 2:528–529
 - benthic foraminifera, 2:768
 - Central Asia, 2:379, 2:380, 2:383–384
 - changes, Indian Ocean, 3:49
 - deglaciation effects, 3:49–52
 - East Asian climate, 2:595, 2:597f, 2:602f
 - equatorial East Africa, 2:373
 - fluvial environment responses, 1:680
 - Indian Ocean, 3:46–54
 - lake level studies
 - Africa, 2:500f, 2:501, 2:502–503
 - Asia, 2:509–511, 2:514, 2:517
 - Latin America, 2:531
 - Last Glacial Maximum GCMs, 3:168
 - last interglacial, 3:158–159
 - postglacial
 - African pollen records, 4:85, 4:89–98
 - Indian Ocean, 3:46–54
 - North Pacific, 3:64–65
 - variability, Indian Ocean, 3:47–52
 - West Africa 6000 ¹⁴C years BP, 3:868–869
- Montane, Southeastern Australia, 1:196
- montremes, 4:631
- moose, 4:715
- morainal banks, 1:776, 2:30–32, 2:34f
- moraine ridges, 2:257–259, 2:258f, 2:567–568, 2:570
- moraines
- Australia, 2:204, 2:204f
 - cosmogenic nuclide dating, 1:435
 - dendroglaciology, 2:104, 2:105–106, 2:105f, 2:106f, 2:108f
 - ELA, MELM approach, 1:913, 1:915f
 - forms/genesis, 1:769–779
 - coalescent subaqueous fans, 1:776
 - controlled moraine, 1:774–775
 - De Geer moraines, 1:776–777
 - dump moraines, 1:769–770, 1:772f, 1:773f
 - hummocky moraine, 1:774–775
 - ice-marginal aprons, 1:769–770, 1:773f
 - ice-shelf moraines, 1:777–778
 - latero-frontal fans/ramps, 1:770–772, 1:772f
 - medial moraines, 1:772–774
 - morainal banks, 1:776
 - push moraines, 1:769, 1:771f, 1:772f
 - squeeze moraines, 1:769
 - glacial landsystems, 1:818f
 - glacier/icesheet extent, 1:886–887, 1:886f, 1:888f
 - ice-proximal settings, 2:30–32, 2:34f
 - lichenometry, 2:567–568, 2:570
 - New Zealand, Late Pleistocene, 2:210, 2:210f
 - permafrost–glacier interactions, 3:508–509, 3:510f
 - Saalian glaciation, 2:174, 2:175f
 - Scandinavian ice sheet, 2:230
 - South American glaciations, 2:252, 2:253, 2:254
 - stratigraphy, neoglaciation, 2:257–259, 2:258f
 - thrust-block, 1:840–843, 1:843f
- Morozovka stage mammal assemblages, 4:610–611
- morphodynamics, deltas, 1:695–699
- morpho-lithostratigraphy, 4:244
- morphology
- block streams, 3:514–515
 - deltas, 1:695–699
 - dinoflagellates/dinocysts, 2:803–805
 - European loess records, 2:606–608, 2:608–609
 - fossil records, chironomids, 1:357
 - glacial sediment, 2:52–61
 - modern human origins, 1:91
 - paleosols, 3:370–372, 3:402–411
 - periglacial landforms, 3:490–491
 - phytoliths, 3:584–585, 3:586f
 - rockfall/talus, 3:568–569
 - rock glaciers, 3:538–539
 - soil, 3:402–411
 - arid climates, 3:363f, 3:365
 - examples, 3:364f
 - humid climates, 3:363f, 3:365
 - stratigraphy, 3:365
 - tunnel valleys, 1:828–830
 - varved sediments, 4:583–584
- morphometry
- definition, 1:780ge
 - lake-level modeling, 2:563
 - subglacial volcanic landforms, 1:784t, 1:794f, 1:795f
- morphostratigraphy, 4:243–249
- fluvial sediments, 4:243
 - glacial sediments, 4:243
 - lake-level fluctuations, 4:243
 - in practice, 4:248
 - sea-level change, 4:243–244
 - see also* allostratigraphy; lithostratigraphy
- morphotypes
- definition, 1:361ge
 - temperature proxies, 2:842–843
- moss macrofossils ¹⁴C, 4:363, 4:364
- moss polsters, 3:839, 3:840, 3:840f
- Motu Ihupuku *see* Campbell Island
- mountain goats, 4:716
- mountains
- Australia Late Pleistocene, 2:202–204
 - China, Tibetan Plateau, 2:379–386
 - glaciation
 - cirques, 1:847–849
 - glacial landforms, 1:764–765
 - Highland Asia Late Quaternary, 2:238–239, 2:239f
 - high-relief environments, 1:765
 - low-relief environments, 1:764–765
 - rock glaciers, 3:535–541
 - Highland Asia Late Quaternary, 2:238–239, 2:239f
 - Iran, 3:786–787, 3:788f
 - Mediterranean beetle records, 1:278–280
 - New Zealand Late Pleistocene, 2:207–212
 - permafrost, 3:535
 - plant macrofossil/pollen studies, 3:731
 - South American ice core studies, 2:387–394
 - temperate mountain glaciers, 2:289, 2:292–293
 - Western Cordillera, NWNA, 4:129
 - see also* individual mountains/mountain regions
 - Mousterian unit archeological evidence, 1:94–95, 1:98, 1:147
 - movement rates, block streams, 3:519–521
 - Movius line, Asia, 1:77–80
 - MOW *see* mediterranean outflow water
 - mowing system pollen records, 3:890–891
 - MPT *see* Middle Pleistocene transition
 - MS *see* magnetic susceptibility; mass spectrometry
 - MSA *see* Middle Stone Age
 - mtDNA *see* mitochondrial DNA
 - muck *see* Yedoma...
 - mud boils, nonsorted patterned ground, 3:459–460
 - multicollector inductively-coupled plasma mass spectrometry (MC-ICP-MS)
 - dating methods, 1:14, 1:14f
 - diatom silica, 2:735
 - multidisciplinary studies, plant macrofossils, 3:785–793
 - multiple-aliquot techniques, OSL, 2:656–657
 - multiple collector... *see* multicollector inductively-coupled plasma mass spectrometry
 - multiple dispersal model, human evolution, 1:54
 - multiproxy studies
 - chironomid records, Europe, 1:386–387, 1:393–394
 - environmental change evidence, 3:788–789, 3:788f
 - paleolimnology, 3:339–348, 3:670–672
 - plant macrofossils, 3:610
 - pollen records, North America, 4:117
 - multiregional model, modern human origins, 1:57, 1:91, 1:146
 - multivariate techniques
 - climate–plant macrofossil reconstructions, 3:787
 - pollen climate proxies, 3:806
 - sediment analysis, 3:351–352
 - surface sample studies, 3:685–686
 - visible-NIR spectroscopy, 3:349
 - mummified remains

mummified remains (*Continued*)
 mammals, Late Pleistocene, 4:713–717
 vertebrates, Pleistocene, 4:593
 muons
 ice-margin sites, 2:425
 interactions at different altitudes, 1:427
 landscape evolution, 1:440*ge*, 1:440–441
 murder cases
 plant remains, 4:543, 4:544
 soil/sediment as evidence, 4:535–537
 Muscidae (house/stable flies), 4:548, 4:550–551
 muskox, 4:715
Mustela nigripes *see* black-footed ferret
 mutual climatic range (MCR)
 beetle records, 1:167–168, 1:168*f*
 North America, 1:221, 1:225, 1:227, 1:231, 1:232
 insect fossils, 1:200, 1:204–205
 Middle Pleistocene, 1:184, 1:189
 postglacial North America, 1:282
 reconstructions of, 1:180–181, 1:191–192
 pollen records, 3:807–808
 Mu Us Desert, China, 1:607*f*, 1:609–610, 1:610*f*
 MWPs *see* meltwater pulses
 MXD *see* maximum latewood density
 Myanmar vertebrate studies, 4:652–653
 mycology, forensic science, 4:542
Myiodon (Patagonian ground sloth), 4:716
 My measure *see* million years measure
Myrica gale macrofossils, 3:642, 3:643*f*

N

N₂O *see* nitrous oxide
 NADW *see* North Atlantic Deep Water
 Naivasha Lake, Kenya, 1:363, 1:364*f*
 Nakorn Sawan, Thailand, 4:653
 NAM *see* Northern Annular Mode
 Namib Desert sand seas, 1:628
 Namibia, African pollen records, 4:14
 naming conventions, pollen databases, 3:834–836
 NAO *see* North Atlantic Oscillation
 NAP *see* nonarborescent pollen
 NAPD *see* North American Pollen Database
 Naracoorte Caves, Australia, 4:621, 4:628–629, 4:634, 4:636*f*
 Nariokotome Boy, human evolution, 1:137–138
 Narmada beds, India, 4:651–652
 NASA, GISS, 1:752, 1:752*f*
 Native American ancient DNA, 1:119, 1:126
 natural abundance, oxygen isotopes, 2:915
 natural-levee sediments, 1:665–666, 1:669*f*
 natural remanent magnetization (NRM), 3:377, 3:380
 natural selection, 4:206
 nature conservation, human impact, 3:881
²¹Ne
 depth-profile dating, 1:436*f*
 erosion-island plots, 1:433, 1:434*f*
 volcanic terrain dating, 1:435
 Neanderthals
 demise of, 1:146–153
 diet, teeth/bones, 1:310
 extinctions, 1:146, 1:150–152
 human evolution, 1:138*f*, 1:139–140, 1:140–141, 1:142, 1:143
 Levant region, 1:98–99
 lithic production, Asia, 1:98
 modern humans replacing, 1:100–103, 1:150–152
 species definition, 1:146
 technologies, 1:147–150, 1:148*t*
 near-field regions, relative sea level changes, 4:489
 near-infrared (NIR) region, 3:349, 3:350*f*
see also visible-NIR spectroscopy
 near-infrared spectroscopy (NIRS), 3:293, 3:296
 near-shore areas, Antarctic Peninsula, 2:219
 near-surface ground ice, 3:574–575
 Nebraska Sand Hills, USA, 1:616, 1:616*f*, 1:619*f*
 Nebraska, USA plant macrofossil records, 3:777–779, 3:780*f*
 Nechells, Birmingham, England beetle assemblages, 1:187
 necrophages, 4:548
 Necrosearch International, 4:546
 needle ice, 3:455–456, 3:456*f*, 3:458
 NEEEM project *see* North Greenland Eemian ice drilling project
 negative muon interactions at different altitudes, 1:427
 neglect cases, forensic entomology, 4:548
 Nemaha River, Nebraska state, 3:777–779, 3:780*f*
 NENA *see* northeastern North America
 Neogene studies, 1:328–330, 1:330*f*, 1:331*f*
see also Late Neogene...
 neoglaciation
 American cordilleras, 2:269–276
 Antarctic Peninsula, 2:220
 chronology
 Americas, 2:271–274
 establishment, 2:270–271
 concept development, 2:269
 definition, 2:216*ge*, 2:269
 Europe, 2:257–268
 Alps, 2:258*f*, 2:259–260, 2:260*f*, 2:263, 2:264*f*
 climatic forcing, 2:265–266
 continuous records, 2:260–262
 events, 2:265–266, 2:266*f*
 glaciers
 expansion dating, 2:257–259
 retreat dating, 2:259–260
 variations, 2:257, 2:264–265
 models, 2:264–265
 moraine-ridge stratigraphy, 2:257–259, 2:258*f*
 pattern/timing, 2:262–263
 reconstructions, 2:257–262
 Scandinavia, 2:262–263, 2:263*f*, 2:265*f*
 Greenland sea-level changes, 4:484–487
 onset of, 2:274
 Neolithic

archeological evidence, 1:95–96, 3:715*t*
 forest clearance impact, 1:274, 1:277
see also Early Stone Age; Late Stone Age; Mesolithic...; Middle Stone Age; Paleolithic...
 neon-21, TCNs, 1:412–413
 Neotoma Paleoecology Database, 3:831, 3:832–833, 3:838
 structure, 3:833–834, 3:834*f*
 synonymy/naming conventions, 3:834–836
 NEP *see* North East Pacific
 nervous system *see* sympathetic nervous system
 net balance, definition, 1:909*ge*
 Netherlands
 basal peat beds, 4:240
 lithostratigraphy, 4:229–234, 4:231*f*, 4:235–238, 4:238–240
 relative sea level changes, 4:492–493
 net mass balance, glacial landforms, 1:910
 networks
 neural, 2:845–846, 3:810
 plant macrofossil/pollen studies, 3:731
 neural networks
 ANN temperature proxies, 2:845–846
 artificial, pollen-climate relation, 3:810
 neutrons
 activation analysis, soil/sediment, 4:540
 ice-margin sites ¹⁴C, 2:425
 interactions at different altitudes, 1:428
 Newell, JP, 3:106
 New England, North America, 4:120–121
 New Guinea pollen records, 4:18, 4:23–24
 newspapers, climatology, 3:240–241
 New Zealand
 chironomid records, 1:399–400, 1:399*f*, 1:402–403, 1:403–404
 FT dating applications, 1:652–655, 1:655*t*
 Late Pleistocene
 beetle fauna, 1:244–254
 glaciers, 2:202, 2:206–212
 pollen records, 4:18, 4:19–21
 MCR paleoclimate reconstruction, 1:168
 pedostratigraphy, 4:252*f*, 4:253*f*, 4:255–256, 4:257*f*
 pollen records
 Late Pleistocene, 4:18, 4:19–21
 postglacial period, 4:104–114
 relative sea level changes, 4:493
 South Pacific ocean postglacial records, 3:77–80, 3:80*f*
 next-generation sequencing (NGS), ancient plant DNA, 2:711–713
 Ngandong, Java
 faunal assemblage, 4:695
 vertebrate fossils, 4:653–654, 4:655*t*
 NGRIP *see* North Greenland Icecore Project
 NGS *see* next-generation sequencing
 NH₄⁺ *see* ammonium
 NHG *see* Northern Hemisphere glaciation
 Niah Cave, Borneo, 4:662
 Nicholson, SE, 3:103
 Nipkow, Friedrich, 1:540, 1:541*f*, 1:542*f*
 NIR region *see* near-infrared region
 NIRS *see* near-infrared spectroscopy

- Nitidulidae (sap beetles), 4:554
 nitrate (NO_3^-), ice core studies, 2:331, 2:332f
 nitrogen-15
 denitrification, 2:905
 nitrate utilization, 2:903–905
 nutrient proxies, 2:903–905, 2:904f
 nitrogen (N)
 assimilation, diatom silica, 2:738–739
 cycling, ice cores, 2:331
 isotope concepts, 1:291
 teeth/bones, 1:305, 1:305f
 thermal diffusion paleotemperature records, 2:431–433
 nitrous oxide (N_2O), ice cores, 2:471–476
 analytical techniques, 2:471
 glacial-interglacial cycles, 2:475
 Holocene, 2:472, 2:473–474, 2:473f
 last glacial, 2:473–474, 2:473f, 2:474–475
 last two millennia, 2:472, 2:472f
 polar cores, 2:471–472
 reliability of records, 2:471–472
 see also greenhouse gases
 nival discharge, periglacial rivers, 3:492
 NMDS *see* non-metric multidimensional scaling
 NO_3^- (nitrate), ice core studies, 2:331, 2:332f
 no-analog plant associations, North America, 4:120
 noble gases, ice cores, 2:280–281
 nodules
 amorphous coatings, 3:383, 3:385f
 buried soils, 3:381
 calcite coatings, 3:383
 Nogaisky–Morozovka stage mammal assemblages, 4:610–611
 Nojiri-ko insect fauna, 1:207, 1:217–219
 nonanalog assemblages, tundra-steppe, 1:259–261
 nonanalog vegetation communities, South America, 4:52–54
 nonarborescent pollen (NAP), 3:880–881, 3:881–882, 3:883f, 3:892, 3:895f
 nonbiting midges *see* chironomids...
 noncutting dates, dendroarcheology, 3:631
 nonglaciogenic origins, loess, 2:620, 2:624–625
 non-handax assemblages, definition, 1:83ge
 non-lacustrine terrestrial studies
 calcites, 1:322, 1:325f, 1:327–330, 1:329f
 carbonates overview, 1:293
 climate, 1:325–326, 1:327–330
 Devils Hole, 1:322, 1:327–330, 1:329f
 paleosol records, 1:328–330
 precipitation, 1:323
 soil carbonates, 1:322, 1:323f, 1:324f, 1:325–326, 1:326–327, 1:328–329, 1:328f, 1:329–330
 soil diffusion model, 1:326–327
 surficial meteoric carbonates, 1:322–323, 1:323–325, 1:325f, 1:326f
 vegetation, 1:325–326
 nonmarine biogenic carbonates, 1:341–353
 aquatic plants, 1:346–348, 1:347f
 calcareous algae, 1:346–348, 1:347f
 calibration, 1:341–342
 climate change records, 1:349
 equilibrium precipitation, 1:341
 fish otoliths, 1:348–350, 1:349f, 1:350f, 1:351f
 freshwater mollusks, 1:343–346
 micro-sampling techniques, 1:342–343, 1:342f, 1:349f
 unionid bivalves, 1:343–346, 1:343f, 1:344f, 1:345f
 ‘vital effects’, 1:341–342
 non-metric multidimensional scaling (NMDS), pollen, 3:823–824
 nonsea salt (nss), ice cores, 2:327f, 2:328f, 2:329
 nonsorted patterned ground, 3:458–462
 nontemperature effects, Mg/Ca
 corals, 2:878–881
 foraminifera, 2:874
 Nordic Seas
 diatom studies/methods, 1:566–567
 magnetic susceptibility variation, 2:893–895, 2:894f
 paleoceanographic conditions, 1:568f
 SSTs, 1:567–569
 Norfolk, England beetle assemblages, 1:184–186
 North Africa
 fauna, Middle Pleistocene, 4:616
 Younger Dryas, 3:127, 3:128f
 see also Africa
 North America
 Arctic, 3:746–759
 Atlantic coast sea-level change, 4:489–491, 4:492f
 biogeography, 3:318–319
 drought, last millennium, 3:231
 dune fields, 1:598–599, 1:598f, 1:614–621
 age determination, 1:602–603
 origins, 1:602
 sand seas, 1:633–634
 types, 1:600–601
 Early Pleistocene vertebrate studies, 4:599
 Early Quaternary, 2:168
 glaciation history, 2:144–148, 2:148t, 2:180–186
 ground beetle distribution, 1:167f
 Holocene plant macrofossil records, 3:768–784
 human global expansion, 1:119–134
 lake level studies, 2:537–548
 Late Pleistocene
 beetle studies, 1:221–234
 vertebrate studies, 4:673–679
 Late Quaternary
 ice sheets, 2:245–248
 vegetation reconstruction, 3:867, 3:868f
 loess records, 2:620–628
 chronology, 2:620–624
 distribution, 2:575f, 2:620, 2:622f
 paleoclimatic implications, 2:626–627
 sources of loess, 2:624–626
 stratigraphy, 2:620–624, 2:626
 thickness, 2:620, 2:622f
 timing of deposition, 2:624–626
 MCR studies, beetle records, 1:167–168
 megafaunal extinctions, 4:704
 environmental change hypothesis, 4:710, 4:710f
 timing of, 4:705t
 Mid-Pleistocene vertebrate records, 4:646–650
 mid-Quaternary glaciation, 2:180–186
 neoglacial chronology, 2:271–274, 2:274f
 northeastern pollen records, 4:115–123
 oribatid mite studies, 2:684–685
 packrat middens, 3:674–675, 3:676–681
 passive plate margins, 1:688–689
 PDSI data, dendroclimatology, 1:465f
 pollen records, 4:72–83
 NAPD, 3:831, 3:832
 northeastern, 4:115–123
 postglacial, 4:133–141, 4:142–155
 postglacial
 beetle records, 1:282–290
 pollen records, 4:133–141
 Southwest arroyos aggradation–degradation cycles, 1:691
 speleothem records, 1:299
 stratigraphic code, NACSN, 4:244, 4:245, 4:248
 vegetation reconstruction, Late Quaternary, 3:867, 3:868f
 Younger Dryas, 3:126–127, 3:128f
 see also Canada; northern North America; northwestern North America; Yukon Territory
 North American Pollen Database (NAPD), 3:831, 3:832
 North American Stratigraphic Code, NACSN, 4:244, 4:245, 4:248
 North Atlantic
 climate
 abrupt change, 3:49
 loess records, 2:615–617
 sensitivity, 2:749
 Younger Dryas event, 3:126, 3:127f, 3:129–130, 3:130f
 diatom studies/records, 1:562–570
 evolution, 3:9–10
 glacial-interglacial cycles, 3:10–12
 last deglaciation, 3:15–17
 Late Neogene/Early Quaternary transition, 2:156–159, 2:158f
 Late Pleistocene paleoceanography, 3:9–17
 magnetic susceptibility, 2:889, 2:895, 2:895f
 millennial-scale variability, 3:12–15
 modern circulation, 3:9, 3:10f
 postglacial period, 3:55–61
 circulation variations, 3:57–61
 SST variations, 3:55–57, 3:58f, 3:60
 sediment records, 2:156–159
 SSTs, 1:567–569, 1:569f, 3:55–57, 3:58f, 3:60
 sub-Milankovitch events, 3:202, 3:203, 3:203f, 3:204f, 3:205, 3:205f, 3:206
 surface circulation, 1:563f
 Younger Dryas
 climate event, 3:126, 3:127f, 3:129–130, 3:130f
 oscillation, 3:222–223, 3:223–224, 3:224–225

- see also* Atlantic...; North Atlantic Oscillation
- North Atlantic Deep Water (NADW), 3:9, 3:12, 3:13f, 3:15
- carbon-13 reconstruction, 2:899–900, 2:902
- deglaciation effect, 3:15–16, 3:17
- Heinrich events, 3:15
- Late Pleistocene, 3:18, 3:24, 3:26f, 3:28
- magnetic susceptibility, 2:891–893
- NHG causes, 3:189
- paleoceanography overview, 2:750f
- postglacial North Atlantic, 3:57–59, 3:57f
- radioisotope proxies, 2:927–928, 2:929f
- salinity proxies, 2:935f
- sub-Milankovitch events, 3:205
- THC, 1:737, 1:739f, 1:743–744, 1:744f, 1:745–746, 1:745f, 1:746f
- Younger Dryas oscillation, 3:222, 3:223
- North Atlantic Oscillation (NAO), 2:379, 2:384
- Greenland stable isotopes, 2:404–405, 2:405f
- last millennium climate, 3:231, 3:232f, 3:233, 3:234, 3:235
- paleo-ENSO, 3:172, 3:173f, 3:174, 3:175f
- volcanic/solar forcing, 1:750
- see also* North Atlantic
- northeastern Brazil sites, vertebrate fossils, 4:682
- northeastern Iowa sites, plant macrofossil records, 3:772–775, 3:776f
- northeastern North America (NENA)
- climate, 4:115
- pollen records, 4:115–123
- see also* northern North America
- North East Pacific (NEP), 3:63–64, 3:67–69
- see also* North Pacific
- Northeast Siberia
- insect fossil studies, 1:256–257, 1:258f, 1:259t, 1:262t, 1:264t, 1:270t
- palynological records, 4:27, 4:28f, 4:29f, 4:32f, 4:33f, 4:36–37
- northern Africa
- archeological evidence, modern human expansion, 1:94–96
- coastal insularity, 1:96–97
- routes out of/into, modern human expansion, 1:96
- vertebrate fossil records, 4:669–670
- northern Andes
- Late Pleistocene pollen records, 4:54–58
- neoglacial chronology, 2:273
- see also* Andes
- Northern Annular Mode (NAM), volcanic/solar forcing, 1:750
- Northern Asia
- insect fossil studies, 1:255–273
- pollen records, 4:27–38, 4:164–172
- vertebrate studies, 4:605–614
- northern Australia pollen records, 4:23–25
- Northern Eurasia vertebrate studies, 4:605–614
- Northern Europe pollen records, 4:173–178
- see also* Northwest Europe
- Northern Hemisphere
- climate/landscape, 3:746
- continental aridity, 3:869
- glaciation, 2:151–154, 2:156–159, 2:275, 3:185, 3:188, 3:189–191, 3:192
- jet stream, 3:167–168
- LGM vegetation patterns, 3:864, 3:865f
- mid-Holocene vegetation patterns, 3:864, 3:865f
- permafrost distribution, 3:466f
- plant macrofossil records, 3:734f
- Younger Dryas, 3:126–128
- Asia, 3:127–128, 3:128f
- Europe, 3:126, 3:127f
- Greenland ice cores, 3:126, 3:127f
- Middle East, 3:127
- North Africa, 3:127, 3:128f
- North America, 3:126–127, 3:128f
- North Atlantic records, 3:126, 3:127f
- oscillation, 3:222–225, 3:226–227
- Northern Hemisphere glaciation (NHG), 2:151–154, 2:156–159, 3:185, 3:188, 3:189–191, 3:192
- Northern Laurentide ice sheet, 1:891
- northern North America
- pollen records, 4:39–51
- see also* northeastern North America; northwestern North America
- northern South American sites, vertebrate fossils, 4:680–681
- North Greenland Eemian ice drilling (NEEM) project, 2:437
- North Greenland Icecore Project (NGRIP)
- biological materials, 2:291–292, 2:291f, 2:294f
- Quaternary climatic change, 1:27–29, 1:28f
- research overview, 2:437
- stable isotopes, 2:404f, 2:405, 2:406f, 2:407, 2:408
- see also* Greenland Icecore Project
- North Island, New Zealand, 2:207
- North London, England, beetle assemblages, 1:188
- North Pacific
- diatom studies/records, 1:571–587, 2:737f
- olian sediments, 1:638, 1:639–640
- Late Neogene/Early Quaternary transition, 2:158f, 2:162–164
- Late Pleistocene, 3:33–45
- postglacial records, 3:62–72
- Cold tongue, 3:62–64
- EEP, 3:62–64
- ENSO, 3:62, 3:63–64, 3:65, 3:67, 3:68, 3:69–70
- equatorial Pacific Ocean, 3:62–64
- IPWP, 3:62, 3:64–67
- Kuroshio Current, 3:62, 3:64–67
- map of North Pacific, 3:63f
- millennial timescales, 3:69
- NEP, 3:63–64, 3:67–69
- paleoceanographic records, 3:62–72
- Subarctic gyre, 3:64–67
- salinity proxies, 2:935f
- sub-Milankovitch events, 3:203
- Younger Dryas oscillation, 3:224–225
- see also* Pacific
- North Pacific/Indian Ocean deep water (NPDW/IODW), 2:935f
- North Pacific intermediate water (NPIW), 3:33
- North Sea relative sea level changes, 4:492
- Northwest Crater, Tower Hill, Southeastern Australia, 1:196–197
- northwestern North America (NWNNA)
- pollen records, 4:124–132
- see also* northern North America; Western North America
- Northwest Europe
- lithostratigraphy, 4:227–242
- Middle-Pleistocene glaciations, 2:172
- mire/peat macrofossils, 3:638–652
- see also* Northern Europe
- Northwest Territories, Canada, 2:570f
- Norway
- ice sheet/glaciations, 2:229
- Mona moraine glacial sequence
- stratigraphy, 2:88f
- temperature reconstructions, 3:788–789
- notches
- definition, 1:83ge
- sea level indicators, 4:377, 4:378f, 4:388
- Nothofagus* (southern beech), 4:109–110, 4:112
- Nothotheriops shastensis* *see* Shasta ground sloth
- notoungulata, South America, 4:687
- novel biomes, vegetation reconstruction, 3:866–867
- NPDW/IODW *see* North Pacific/Indian Ocean deep water
- NPIW *see* North Pacific intermediate water
- NRM *see* natural remanent magnetization
- nss *see* nonsea salt
- nuclear disintegration rates
- atmospheric, 1:424–425
- cosmic ray temporal variations, 1:425–426
- upper lithosphere, 1:426–427
- nuclear explosions, atmospheric carbon-14 variations, 4:333
- nuclear interactions
- altitudes in atmosphere, 1:427–428
- cross-sections, 1:430
- rates, definition, 1:429–430
- nuclear power/weapons testing, 4:342f, 4:343
- nuclear radiation, energetic, 1:419, 1:424f
- nucleons
- interactions at different altitudes, 1:427
- landscape evolution, 1:440ge, 1:440–441
- nuclides
- production rates, 1:428–429
- radioactive, dating techniques, 1:447–449
- varve sediment dating, 4:585–586
- see also* cosmogenic nuclides; cosmogenic radionuclides
- Nukui Peat Bed, Japan, 1:207
- null hypotheses, beetle fossils, 1:239–240
- numerical dating, paleosols, 3:368–370
- numerical modeling
- biosphere feedbacks, 1:723
- environmental/climate change, 1:32–33
- nunataks *see* paleonunataks

- Nussloch sequence, European loess records, 2:609–610, 2:610f, 2:614f, 2:616, 2:616f, 2:617, 2:617f, 2:618
- nutrient proxies, 2:899–906
barium, 2:901–903
cadmium, 2:901–903, 2:902f
carbon-13, 2:899–901
nitrogen-15, 2:903–905
trace metals, 2:901–903
zinc, 2:901–903
- nutrients
coccolithophores, 2:786–788, 2:787f
transfer functions, 1:538
- NWNA *see* northwestern North America
- Nymphaeacea macrofossils, 3:645f, 3:646
- ## O
- ^{18}O , *see* Oxygen-18
- oak (*Quercus*)
plant distributions, 3:855, 3:856f, 3:857f
postglacial pollen records, 4:179, 4:182, 4:185
- objective vegetation mapping, 3:861
- obliquity
astronomical theory, paleoclimates, 2:138
changes, Holocene Antarctica, 2:397
forcing, Early Pleistocene records, 3:6
- obliterative overlap, glaciations, 2:180
- obsidian
Australian human colonization, 1:112f, 1:116f
definition, 1:108ge
FT dating, 1:649–650, 1:651f
- obstructive sleep apnea (OSA), 1:315–321
- Ocean Drilling Program (ODP), 2:889, 2:890f, 2:891f, 2:893f, 4:73–74, 4:78f
- oceans
biogeochemistry, atmospheric dust, 1:735
- circulation
ice sheet growth/decay, 1:881, 1:881f
last interglacial, 3:162–163
last millennium climate, 3:231–233
North Pacific, 3:33
tectonic uplift/continental configurations, 2:127–135
Younger Dryas climate event, 2:129–130
see also Atlantic meridional overturning circulation
- coastal erosion, 1:18–19, 1:19f
- conveyors, climate change, 1:29–31
- ecosystems, 1:19–20
- environmental change, 1:19–20
- frontal migration, Antarctica, 1:528f, 1:533–535
- frontal zone laminated sediments, 1:556f, 1:560
- gateways, 2:132–134
- global circulation pathways, 1:19f
- heat transport, 3:186–187, 3:190f
- hydrography, 3:204–205, 3:206–207
- ice discharge, Greenland, 2:441
- magnetic susceptibility, 2:889, 2:890f, 2:891f, 2:893f
- pollen records, Late Pleistocene, 4:73–74, 4:78f
- research history, 1:18–20
- role in Quaternary climatic change, 1:29–31
- sea-level change, 1:18
- setting, North Pacific, 3:33
- surface, isostasy, 4:452–453
- syphoning, isostasy, 4:454–455, 4:454f
- water
 $\delta^{18}\text{O}$ record, 4:369–370
diatom studies/records, 1:562–570, 1:571–587, 4:369–370
see also individual oceans;
paleoceanography; sea...
- Ochotona princeps* *see* collared pika
- Odderade Interstadial beetle assemblages, 1:202
- ODP *see* Ocean Drilling Program
- Oeschger events *see* Dansgaard–Oeschger events
- Oeschger, Hans, 2:403
- offensive searches, forensic geology, 4:524
- Ogallala Formation dunes, North America, 1:615–616
- OGIT *see* oral glucose tolerance test
- oil slicks, 4:536–537
- Okhotsk Sea
diatom assemblages, 1:571–574, 1:575
Late Pleistocene South Atlantic, 3:20–21
- Older Saalian glaciation, 2:174–176
- Oldowan Industry, 1:59–62
- 'old wood' effect, plant macrofossil ^{14}C , 4:361–362
- old world origins
Canis genus, 4:600
human colonization, Americas, 1:119–121
- Oligocene epoch, definition, 2:216ge
- oligosaline, definition, 1:361ge
- oligotrophic lakes, 3:657, 3:660f
- olivine, TCNs, 1:410–411, 1:412–413, 1:413–414
- OLLSDB *see* Oxford Lake Level Data Base
- Olson's ecosystem phytolith classes, 3:590f
- Oman speleothem records, 1:299
- omnidirectional fluxes, cosmic rays, 1:429–430
- omnipresent energetic nuclear radiation, 1:419
- OMZ *see* oxygen minimum zone
- Ono Peat Bed, Japan, 1:207
- on-site data, environmental archeology, 3:705
- ontogeny, lakes, 1:358
- opal
magnetic susceptibility, 2:889, 2:891f
phytoliths, 3:585
- open-basin lakes, 2:558
- open land–NAP relationship, 3:895f
- open sites, Southern Asia, 4:651–654
- open-system pingos, 3:473f, 3:474–475, 3:475f
- optical dating, TL dating, 2:650–651
- optically stimulated luminescence (OSL)
dating, 2:653–666
environmental dose rates, 2:655
- equivalent dose, 2:655–656, 2:660–661
- European loess records, 2:609–610, 2:610f
- feldspar, 2:662, 2:663f, 2:664
- fluvial terraces, 1:684, 1:690–691
- generating OSL, 2:655
- history of, 1:7
- inaccuracy causes, 2:664
- instrumentation, 2:654f
- Late Quaternary glaciations, 2:236, 2:238–239
- linear modulation, 2:660, 2:660f, 2:664
- measuring OSL, 2:655
- mechanism responsible for, 2:653–655
- methodology, 1:12–13
- multiple-aliquot techniques, 2:656–657, 2:657f, 2:662
- pulsed stimulation, 2:660–661
- quartz, 2:653, 2:662–664, 2:663f
- sample collection/preparation, 2:662
- single-aliquot techniques, 2:657–660, 2:659f, 2:660–661, 2:662
- suitable environments, 2:662–664
- tephrochronology, 4:289
- optical thickness, atmospheric dust, 1:729, 1:730, 1:732f
- optimum productivity phase, vegetation, 3:763, 3:764
- oral glucose tolerance test (OGIT), 1:315
- orbital forcing
Late Neogene/Early Quaternary transition, 2:151
Southern European Middle/Late Pleistocene pollen records, 4:67–68
- orbital oscillations, marine oxygen isotopes, 2:912
- orbital-scale diatom records, 1:559–560
- ordination techniques
definition, 1:361ge
diatoms, 1:491, 1:492
- Oreamnos harringtoni* *see* Harrington's mountain goat
- Oregon Coast, Northeast Pacific, 1:574
- organic materials
carbonate stable isotopes, 1:293
redistribution effects, 3:434
- organic-walled cysts, dinoflagellates, 2:800, 2:803, 2:805–811
see also dinocysts
- oribatid mites (Acari), 2:679–698
archeological research, 2:683–684, 2:685, 2:695
biofacies analysis, 2:688t, 2:689f, 2:691–692
biostratigraphic applications, 2:690–692
biozonation by species acme zones, 2:691
definition, 2:681–682
development of studies, 2:683–685
applications era, 2:683–685
descriptive era, 2:683
examination of, 2:688–690
fossilization of, 2:682–683
Glovers Pond, NJ, 2:689f, 2:691f, 2:692f, 2:693, 2:694f, 2:695
Hiscock Site, NY, 2:686f, 2:687f, 2:688t, 2:689f

- oribatid mites (Acari) (*Continued*)
 hydrophilic/aquatic mites, 2:685f
 identification, 2:688–690
 laboratory protocols, 2:684f
 mire/peat macrofossils, 3:652, 3:653f
 moisture proxies, 2:693–694, 2:695
 paleoclimate proxies, 2:692–695
 sectors of application, 2:696
 sediment sampling, 2:685–688
 fossil extraction, 2:686–688
 weight vs. volume, 2:686
 socio-economic patterns, 2:695
 spring deposits, 2:695
 taphonomic analysis, 2:695, 2:696f
 techniques/methods, 2:685–690
 temperature proxies, 2:693–694, 2:695
 orogenesis, tectonic uplift, 2:127–135
 OSL *see* optically stimulated luminescence
 dating
 osmotic stress, 1:361ge
 Ostend, England beetle assemblages, 1:186
 ostracods
 lake-level reconstructions, 2:496
 lake salinity, 3:295
 Mg/Ca paleothermometry, 2:871, 2:877–878
 microfossil-based reconstruction, Holocene RSL, 4:421–422
 mire/peat macrofossils, 3:652, 3:653f
 salinity proxies, 2:936–937
Ostrya (hop hornbeam), 4:116f, 4:119
 otoliths, fish, 1:348–350, 1:349f, 1:350f, 1:351f
 Out of Africa hypothesis *see* out-of-Africa model
 outbursts *see* glacial lake outbursts
 outcrops, sequence stratigraphy, 4:261–262, 4:262f, 4:263t, 4:267f, 4:268f, 4:269–271
 outflows, Indian Ocean, 3:47
 outlet glaciers
 East Antarctic Ice Sheet, 2:457, 2:457f, 2:459–460
 West Antarctic Ice Sheet, 2:448, 2:449
 out-of-Africa model
 Australian colonization, 1:111–112, 1:112–113
 human evolution, 1:142
 modern human origins, 1:91
 Pleistocene legacy, 1:53
 outwash fans, 2:14–15
 outwash plains/channels, New Zealand, 2:210, 2:210f
 overbank deposits, fluvial environments, 1:664–670, 1:684
 overflowing lakes, Asia, 2:506, 2:512f
 overflows, glaciolacustrine, 2:43–44, 2:44f
 overkill hypothesis, Pleistocene extinctions, 4:709
 overwash sand layers, 3:209–211, 3:212f, 3:213f, 3:215f
 Owatari II Site, Japan, beetle fossils, 1:207
 owls, 4:746
 Oxford Lake Level Data Base (OLLSDB), 2:511
 Oxfordshire, England beetle assemblages, 1:188
 oxidation
 dinocyst techniques, 2:811
 iron-bearing minerals, 3:502
 weathering profiles, 3:392–394, 3:393f, 3:395f, 3:399f
 oxycline layer, definition, 1:361ge
Oxycoccus palustris macrofossils, 3:638
see also *Vaccinium oxycoccus*
 oxygen
 levels, chironomid effects, 1:359
 reactivity, OSA and, 1:320
 systematics, $\delta^{18}\text{O}_{\text{diatom}}$ records, 1:482–487
 teeth/bones, carbonate stable isotopes, 1:305–306, 1:305f
see also oxidation; oxygen isotopes; oxygen minimum zone
 Oxygen-18 ($\delta^{18}\text{O}$), 1:481, 1:483f, 1:485f, 1:486, 1:486f
 Greenland stable isotopes, 2:403–405, 2:404f, 2:405–408, 2:405f, 2:405t, 2:406f, 2:407f, 2:408t
 North American records, 2:180–186
 sea-level change, 4:369–370
 speleothems, 1:294–295
 oxygen isotopes
 beetle records, 1:169–170
 benthic foraminifera, 3:28, 3:28f, 3:29f, 3:46
 carbonate composition, 1:338
 chironomid records, 1:374, 1:378–379, 1:379f
 CO₂ composition, firm air, 2:367, 2:367f
 concepts, 1:291
 deglaciation, Arabian Sea, 3:48f
 $\delta^{18}\text{O}$ records, 1:481, 1:483f, 1:485f, 1:486, 1:486f
 Greenland, 2:403–405, 2:404f, 2:405–408, 2:405f, 2:405t, 2:406f, 2:407f, 2:408t
 North America, 2:180–186
 sea-level change, 4:369–370
 speleothems, 1:294–295
 freshwater mollusks, 3:286
 ice cores
 Andes, 2:389–390
 Antarctica, 2:395
 correlation techniques, 2:410, 2:411f, 2:412
 distribution, 2:347–348
 standard notation/measurements, 2:347
 lake-level reconstructions, 2:496–497
 lake water composition, 1:336–337, 1:337–338
 magnetic susceptibility, 2:889, 2:891f, 2:893f
 measurement, 2:915, 2:919, 2:921
 natural abundance, 2:915
 physicochemical behavior, 2:915
 Quaternary lacustrine records, 1:338
 ratios, Late Pleistocene South Atlantic, 3:28, 3:28f, 3:29f
 salinity
 Indian Ocean, 3:46–47
 proxies, 2:932–940
 seawater composition, 2:915–922
 soil carbonates, 1:328–329
 speleothems, 1:294–296, 1:297–298, 1:301f
 stratigraphy
 paleoceanography, 2:907–914
 absolute dating, 2:909
 astronomical tuning, 2:910–912
 carbonate fractionation, 2:907–908
 continental ice sheet, 2:908–909
 delta notation, 2:907
 global reference curve, 2:912
 global signal, 2:908–909
 hydrological cycle, 2:908f
 improvements, 2:912
 limits, 2:912–914
 magnetostratigraphy, 2:909–910
 marine carbonate, 2:907–908
 overview, 2:745
 Pleistocene, 2:908–909, 2:909f
 primary orbital oscillations, 2:912
 ‘Rosetta Stone’ of ice ages, 2:909–910, 2:909f
 SPECMAP approach, 2:910–912, 2:911f, 2:913f
 timescale development, 2:909–912
 Quaternary climatic change, 1:27–29, 1:28f
 surficial meteoric carbonates, 1:322–323
 temperature calculations, 1:337–338
 variability, Indian Ocean, 3:49f
see also marine oxygen isotope records
 oxygen minimum zone (OMZ)
 denitrification, 2:905
 laminated sediment diatom records, 1:555, 1:556–557, 1:556f
 variability, Indian Ocean, 3:52–54
 oxygen stable isotopes, 3:286
see also oxygen isotopes
 oxyhydroxides, 3:375ge
 Oyo depression, Sudan, 4:90, 4:91f
 oyster beds, 4:496–497
 Ozerki swamp, Kazakhstan, 4:166, 4:166f
 ozone chemistry, atmospheric dust, 1:735

P

- Pacific air mass system, 4:142–143
 Pacific Decadal Oscillation (PDO), 2:384, 3:172–174, 3:233
 Pacific North American (PNA) patterns, paleo-ENSO, 3:172
 Pacific Ocean
 diatom silica concentrations, 2:737f
 diatom studies/records, 1:571–587, 2:737f
 eolian sediments, 1:638, 1:639–640
 Last Glacial Maximum GCMs, 3:169
 Late Neogene/Early Quaternary transition, 2:162–164
 Late Pleistocene, 3:33–45
 pollen records, 4:72, 4:73f, 4:75f
 salinity proxies, 2:935f, 2:938–939, 2:938f, 2:939f
 sediments
 eolian, 1:638, 1:639–640
 radioisotope proxies, 2:929–930, 2:930f
 tropical, Younger Dryas, 3:129

- see also eastern equatorial Pacific; North East Pacific; North Pacific; South Pacific Ocean
 Pacific Ocean Islands, 2:189
 Pacific region, southwest, 2:202–215
 Pacific transect, 4:80–82
 packrat middens
 analysis methods, 3:675
 beetle studies, 1:162, 1:164f, 1:225, 1:226t
 fossil site distribution, 3:675
 implications of studies, 3:681–682
 North America, 3:674–675
 beetle studies, 1:225, 1:226t
 regional studies, 3:676–681
 PAD *see* Peace-Athabasca Delta
 Päijänne Lake, Finland, 1:386–387, 1:389f
 paired terraces, climatic processes, 1:689–690
 Pakefield, England beetle assemblages, 1:185
 Pakistan, vertebrate studies, 4:651–652, 4:652t, 4:653t
 Pakitsiq ice-margin site, Greenland, 2:417f, 2:423–424, 2:425f, 2:426–427, 2:428f
 palaeo... *see* paleo...
 Paleoamerican physical anthropology, 1:126
 paleoanthropology
 development, 1:49–51
 overview, 1:49–58
 paleobiogeography
 beetle assemblages, 1:184–190
 vertebrate fossils, 4:594–595
 paleobiology, 2:795–799
 paleobotany
 ancient plant DNA, 2:705–715
 Antarctic marine waters, 1:527–539
 archeological applications, 1:516–521, 3:699–724, 3:880–903
 biome approach, vegetation reconstruction, 3:861–870
 charred particle analyses, 2:716–729
 CO₂ reconstruction, fossil leaves, 3:613–620
 diatomites, 1:501–506
 diatoms
 Antarctic marine waters, 1:527–539
 archeological analysis, 1:516–521
 $\delta^{18}\text{O}_{\text{diatom}}$ records, 1:481–488
 freshwater laminated sequences, 1:540–545
 introduction, 1:471–480
 large lakes, 1:546–553
 marine laminated sediments, 1:554–561
 North Atlantic/Arctic, 1:562–570
 Pacific Ocean, 1:571–587
 salinity/climate reconstruction, 1:507–515
 freshwater laminated sequences, 1:540–545
 large lake diatom records, 1:546–553
 Late Pleistocene
 Africa, 4:9–17
 Australasia, 4:18–26
 North America, 4:39–51
 northern Asia, 4:27–38
 South America, 4:52–62
 southern Europe, 4:63–71
 Late Quaternary European pollen records, 4:1–8
 marine laminated sediments, 1:554–561
 paleophytogeography, 2:730–733
 phytoliths, 3:582–592
 plant macrofossils, 3:593–612
 Arctic
 Eurasia, 3:733–745
 North America, 3:746–759
 environmental archeology use, 3:699–724
 Greenland, 3:760–767
 Holocene North America, 3:768–784
 megafossils, 3:621–629
 mire/peat macrofossils, 3:637–656
 multidisciplinary studies, 3:785–793
 paleolimnology, 3:657–673
 pollen studies, 3:725–732
 rodent middens, 3:674–683
 surface samples/taphonomy/representation, 3:684–689
 treeline studies, 3:690–698
 pollen
 analysis principles, 3:794–804
 climate proxies, 3:805–815
 databases, 3:831–838
 data overview, 2:699–704
 Late Pleistocene
 Africa, 4:9–17
 North America, 4:39–51
 Southern Europe, 4:63–71
 Late Quaternary records, 4:1–8
 northeastern North America, 4:115–123
 numerical analysis, 3:821–830
 plant distributions, 3:854–860
 POLLSCAPE model, 3:871–879
 postglacial
 Africa, 4:85–103
 Australia, 4:104–114
 New Zealand, 4:104–114
 North America, 4:124–132, 4:133–141
 northern Asia, 4:164–172
 northern Europe, 4:173–178
 South America, 4:156–163
 southern Europe, 4:179–188
 reconstructing biodiversity, 3:816–820
 southern Europe, 4:63–71, 4:179–188
 southwestern North America, 4:142–155
 stand-scale palynology, 3:846–853
 surface samples, 3:684–689, 3:839–845
 trapping, 3:839–845
 validation, 3:725–732
 Western North America, 4:72–83
 salinity/climate reconstruction, 1:507–515
 paleoceanography, 2:745–754
 Antarctic sea ice, 1:535
 atmospheric carbon dioxide, 2:749–750, 3:4f, 3:6
 biological proxies
 alkenone paleothermometry, 2:755–764
 benthic foraminifera, 2:765–774
 biomarkers, 2:775–782
 coccolithophores, 2:783–794
 marine diatoms, 2:816–824
 planktic foraminifera, 2:825–829
 radiolarians/silicoflagellates, 2:830–840
 temperature proxies, 2:841–848
 carbon-13 reconstructions, 2:899–901
 chemical proxies, 2:849–858, 2:884–898
 deep-sea carbonate dissolution, 2:859–870
 nutrient proxies, 2:899–906
 oxygen isotope stratigraphy, 2:907–914
 climate change aspects, 2:749–750
 corals/sclerosponges/mollusks, 2:795–799
 diatom proxies, Pacific Ocean, 1:571
 dinocyst reconstructions, 2:812
 dinoflagellates, 2:800–815
 Early Pleistocene, 3:1–8
 atmospheric CO₂, 3:4f, 3:6
 chaotic response, obliquity forcing, 3:6
 glacial–interglacial changes, 3:3, 3:5
 ice volume, 3:3
 Late Pleistocene boundary, 3:1–3
 long-term temperature trends, 3:3–5
 Mid-Pleistocene transition, 3:1–3, 3:5–6
 millennial-scale dynamics, 3:6–7
 paleotemperature changes, 3:3–5
 Pliocene–Pleistocene boundary, 3:1
 regolith hypothesis, 3:6
 sea-ice switch hypothesis, 3:6
 sea-level changes, 3:3
 stratigraphy, 3:1–3
 thermohaline circulation, 3:5, 3:6
 Walker circulation hypothesis, 3:6
 evolution, 2:746–748
 glacial–interglacial cycles, 2:748–750
 last deglaciation, 2:751–753, 2:753f
 Late Pleistocene
 Early Pleistocene boundary, 3:1–3
 North Atlantic, 3:9–17
 North Pacific Ocean, 3:33–45
 South Atlantic, 3:18–32
 magnetic susceptibility, 2:884–898
 Mg/Ca and Sr/Ca paleothermometry, 2:871–883
 Mid-Pleistocene transition, 2:747–748, 3:1–3, 3:5–6
 millennial-scale variability, 2:750–751, 2:752f
 North Atlantic, 1:563–565, 1:568f
 climate sensitivity, 2:749
 Late Pleistocene, 3:9–17
 postglacial, 3:55–61
 North Pacific Ocean, 3:33–45, 3:62–72
 overview, 2:745–754
 oxygen isotopes
 seawater proxies, 2:915–922
 stratigraphy, 2:745, 2:907–914
 physical proxies, 2:884–898
 deep-sea carbonate dissolution, 2:859–870
 nutrient proxies, 2:899–906
 oxygen isotope stratigraphy, 2:907–914
 terrigenous sediments, 2:941–952
 postglacial, 2:753–754
 Indian Ocean, 3:46–54
 North Atlantic, 3:55–61
 North Pacific, 3:62–72
 South Pacific, 3:73–85

- paleoceanography (*Continued*)
 proxies overview, 2:745–746, 2:746t, 2:753
 radioisotope proxies, 2:923–931
 salinity proxies, 2:932–940
 silicon isotopes in diatoms, 2:734–743
 Southern Ocean, 1:527
 SST reconstructions, 2:746–747
 zonal/meridional gradients, 2:746–747
 Paleoclimate Modeling Intercomparison Project (PMIP), 3:137, 3:138t, 3:142, 3:144f, 3:145–146
 paleoclimate/paleoclimatology
 Arctic history, 3:113–125
 astronomical theory, 2:136–141
 Australia
 beetle records, 1:191–192
 lake-level studies, 2:528–529
 beetle records
 Australia, 1:191–192
 New Zealand, 1:244–254
 North America, 1:221–234, 1:282–290
 plant reconciliation, 1:197–198
 significance, 1:274
 biogeography, 3:314–317
 block streams significance, 3:519–521
 changes
 chironomid records, 1:359–360
 loess records, 2:589, 2:591
 oxygen isotopes, 1:297–298
 quantified estimates, 1:200
 chironomids
 Africa, 1:369
 changes, 1:359–360
 Southern Hemisphere, 1:400–402
 contrasts, Late Pleistocene events, 2:192
 corals/sclerosponges/mollusks, 2:795–799
 cryoturbation structures significance, 3:434
 deep-sea sediments, 1:638, 1:640
 dinoflagellates, 2:800–815
 dunes significance, 1:612–613, 1:618–621
 encyclopedia article organization, 3:90–91
 forcing mechanisms, glacial climates, 1:721–728, 1:737–747, 1:748–753
 fundamental research approaches, 3:88–89, 3:88f
 glacial landforms, ELAs, 1:912–913
 insect indicators, 1:177–178
 introduction to, 3:87–92
 lake sediments
 environmental effects, 1:338
 sampling, 1:333–334
 large lake diatom records, 1:546–547, 1:548, 1:550–551
 Late Pleistocene events in Beringia, 2:192
 loess
 Central Asian changes, 2:589, 2:591
 characteristics, 2:573, 2:582–583, 3:359, 3:366
 North American records, 2:626–627
 proxies, European records, 2:612–615
 South American records, 2:638–639
 modeling
 Asian lake level studies, 2:511, 2:516–519, 2:518f, 2:519f
 data-model comparisons, 3:135–146
 Last Glacial Maximum GCMs, 3:165–170
 last interglacial, 3:155–164
 paleo-ENSO, 3:171–177
 PMIP, 3:137, 3:138t, 3:142, 3:144f, 3:145–146
 Quaternary environments, 3:147–154
 modern analog approaches, 3:102–112
 North America
 beetle studies, 1:221–234, 1:282–290
 dunes significance, 1:618–621
 loess records, 2:626–627
 Western, 4:72–83
 oribatid mite studies, 2:692–695, 2:693f
 overview/approaches, 3:87–92
 pedostratigraphy, 4:254–256
 planktic foraminifera, 2:828
 plant macrofossils
 Greenland, 3:760
 paleolimnological applications, 3:657–673
 proxies
 European loess, 2:612–615
 reconstructions, 3:179–180
 reasons to study, 3:87–88
 recent developments, 3:88–90
 reconstructions, 3:179–184
 Australian beetle records, 1:191–192
 biome approach, 3:861
 chironomid records, Southern Hemisphere, 1:400–402
 coexistence approach, 1:191–192
 global warming, 3:244–252
 historical climatology, 3:237–243
 lacustrine carbonate, 1:336–337, 1:337–338
 last millennium climate, 3:229–236
 methodological context, 1:191
 New Zealand beetle records, 1:244–254
 paleodroughts/society, 3:194–199
 paleotempestology, 3:209–221
 Pliocene environments, 3:185–193
 pollen data, 3:810–814
 proxies, 3:179–180
 qualitative approaches, 3:180–181
 quantitative approaches, 3:181–183
 sub-Milankovitch events, 3:200–208
 Younger Dryas oscillation, 3:222–228
 records introduction, 3:87, 3:88
 speleothems
 applications, 1:298–300
 oxygen isotope changes, 1:297–298
 timescales of climate change, 3:93–101
 varved sediment use, 4:587–588
 Western North America, 4:72–83
 Younger Dryas, 3:126–134, 3:222–228
 paleodroughts, 3:194–199
 climatic forcing, 3:196–197
 diatoms, 3:195–196
 eolian records, 3:196
 fossil pollen, 3:195
 historical documents, 3:194
 impact on past societies, 3:197–198
 lake records, 3:195
 past drought proxies, 3:194–196
 and society, 3:194–199
 tree rings, 3:194–195
 paleoecology
 Arctic North America, 3:746–759
 beetle studies, 1:166
 New Zealand, 1:244–254
 North America, 1:282–290
 Western Europe, 1:189
 chironomids
 Africa, 1:361–372
 overview, 1:358–360
 diatom records, 1:540
 Greenland, 3:760–767
 Holocene North America, 3:768
 indicators
 dinocysts, 2:808–811
 insects, 1:179–180
 plant macrofossils, 3:593, 3:608
 pollen studies, 3:725–732
 Northern Asia, 4:28–29, 4:34–35
 South America, 4:52
 vertebrate fossils, 4:594
 Western North America, 4:72–83
 paleoeconomic studies, 3:699
 paleo-ELAs, 1:909–917
 calculation approaches, 1:913–916
 reconstruction approaches, 1:916–917
 paleo-ENSO, 3:171–177
 climatic forcing mechanisms, 3:174–176
 ENSO, 3:171, 3:174
 regional El Niño responses, 3:174–176
 teleconnections, 3:171–174
 paleoentomology, chironomid overview, 1:355–360
 paleoenvironments
 American human colonization, 1:121–122
 archeobotany contrast, 3:699
 assessments, paleolimnology, 3:266–268
 data sets, paleoclimate modeling, 3:142–146
 $\delta^{18}\text{O}_{\text{diatom}}$ records, 1:481, 1:482
 high-latitude dune fields, 1:604
 ice-margin sites, 2:416, 2:428
 indicators
 chironomid records, postglacial
 Europe, 1:386, 1:395, 1:396
 Southern Hemisphere, 1:398
 diatoms, 1:477
 insect fossil studies, 1:255–273, 1:274
 loess records, 2:580–582, 2:591, 2:593f
 loess sequences, 2:580–582
 low-latitude sand seas, 1:623
 micromorphology, buried soils, 3:381–383
 North American beetle studies, 1:221–225
 paleosol morphology, 3:370–372
 proxy diatoms, 1:546
 reconstructions
 beetle records, 1:168–169, 1:244–254
 chironomid records
 Africa, 1:362
 Europe, 1:373–385
 dendrochronology, 1:453, 1:454, 1:457
 IP₂₅ HBI for Arctic sea ice, 1:592–594
 phytoliths, 3:588–589
 varved lake sediments, 4:579–580
 Sahul human colonization, 1:108–110

- talus slopes significance, 3:572
 Yedoma archives, Beringia, 3:548
- paleo-erosion rates, landscape evolution, 1:444
- paleoethnobotany *see* archeobotany
- paleoflood hydrology, 3:255–256
- paleoglaciological considerations, trimlines/paleonunataks, 1:925–926
- paleohydrology, 3:253–258
 chironomid records, 1:360
 definition, 3:253
- paleo-ice sheets, 1:895–896
- paleo-ice streaming, 1:899–902, 1:904f
 see also ice stream landsystems
- Paleoindians
 environmental impact, 1:127
 technology, 1:128–129
- paleoindicators, diatoms as, 1:489
- paleolake simulation analysis, 2:561, 2:562
 see also lakes
- paleolimnology, 3:259–270
 Antarctic diatom studies, 1:538
 applications, 3:266–268, 3:657–673
 biogeography, 3:313–325
 change causation hypotheses, 3:339, 3:342–344, 3:670–672
 chironomid records, Africa, 1:362
 cladocera, 3:271–280
 coring, 3:262–264, 3:262f, 3:263f
 dating analyses, 3:264, 3:265–266
 diatom studies
 Antarctica, 1:538
 silicon isotope analysis, 2:734–743
 varved sediments, 1:540, 1:541
 freshwater mollusks, 3:281–291
 geochronology, 3:264
 IR spectroscopy applications, 3:349–355
- lakes
 as archives, 3:260
 chemistry, 3:292–299
 multiproxy approaches, 3:339–341, 3:341–342, 3:344, 3:345–346
 postglacial event progression, 3:260f
 sediment physical properties, 3:300–312
 methodology, 3:260–268, 3:262f
 multiproxy approaches, 3:339–348
 algal blooms in Low lake, 3:345–346
 animals as nutrient/contaminant vectors, 3:346–347
 environmental change hypotheses, 3:339, 3:342–344
 environmental stressors, 3:344–347
 holistic approach, 3:339–341
 lake habitats, 3:339–341, 3:341–342, 3:344, 3:345–346
 stressors/effects, 3:341–342, 3:344–347
 new developments, 3:268
 overview, 3:259–270
 paleoenvironmental assessment, 3:266–268
 pigment studies, 3:326–338
 postglacial lake event progression, 3:260f
 proxy indicator analysis, 3:264–265
 biological indicators, 3:261t, 3:264–265
 chemical indicators, 3:261t, 3:265
 physical indicators, 3:261t, 3:265
- quantification, 3:265–266
 subsectioning, 3:264
 VIS spectroscopy applications, 3:349–355
- Paleolithic archeology
 beetle fauna, 1:187
 hominin archives, 1:78f
 modern human expansion, 1:94–95, 1:98–100, 1:102f, 1:103f
 Neanderthal demise, 1:146–153
 technological modes, 1:56t
 vertebrate studies, 4:590, 4:593
 see also Middle Paleolithic
- Paleolithic hominid–animal interactions, 4:748–749, 4:749–750, 4:750–751
- paleomagnetic dating
 lake sediment physical properties, 3:304–305
 North American glaciations, 2:182–185
- paleomagnetic fields, 4:333
- paleomagnetism, 1:709–713
 data reliability, 1:712
 rocks, 1:709–712
 secular variation, 1:709–713
 stratigraphy, 1:718–719
 tephrochronology, 4:289
- paleontology
 eastern Africa, 4:670
 northern Africa, 4:669
 western/central Africa, 4:670–671
- paleonunataks, 1:918–928
 see also trimlines
- paleopedology
 loess, 3:369f
 paleosols, 3:367
- paleophytogeography, 2:730–733
 future trends, 2:732
 Holocene, 2:731–732
 ice age impact, 2:730–731
 last glacial phase, 2:731
 present interglacial, 2:731–732
- paleoproductivity, chironomids, 1:398, 1:402
- paleosalinity, chironomid records, 1:360
- paleosecular variation (PSV), 1:713–714, 1:715f, 1:718, 1:719
- paleoseismic rockfalls, 2:570–571
- paleoseismic sea-level change, 4:503–519
- paleoshorelines
 cosmogenic nuclide dating, 1:437
 morphostratigraphy, 4:243, 4:246f
- paleosols
 deep-sea sediments, 1:637–642
 definitions, 3:365–366
 dune fields
 high latitudes, 1:597–605
 low latitudes, 1:623–636
 midlatitude, 1:606–622
 future research directions, 3:374
 loess records
 Central Asia, 2:585–594
 China, 2:595–605
 deposit origins/properties, 2:573–584
 Europe, 2:606–619
 North America, 2:620–628
 South America, 2:629–641
- microstratigraphic relations, 3:386–388
- mineral magnetic analysis, 3:375–380
- morphology, 3:370–372, 3:402–411
- nature of, 3:367–374
- non-lacustrine terrestrial studies, 1:324f, 1:328–330
- numerical dating, 3:368–370
- overview, 3:357–366
- pedostratigraphic nomenclature, 4:253–254
- postburial alteration, 3:372–373
- proxy data from, 3:373–374
- in Quaternary records, 3:360–366
- recognition of, 3:367–368
- role in climate cycles, 3:412–420
- soil micromorphology, 3:381–391
- terminology issues, 3:367
- weathering profiles, 3:392–401
- Yedoma of Beringia, 3:547–548, 3:547f, 3:550
 see also windblown sediments
- paleotemperature
 chironomid records, 1:359–360
 estimating with radiolarians, 2:834, 2:835f
 proxies, Mg/Ca and Sr/Ca, 2:871–883
 speleochem records, 1:300
 thermal diffusion, 2:431–434
- paleotempestology, 3:209–221
 back-barrier lakes, 3:209–211
 beach ridges as proxies, 3:213
 British archives, 3:217–218
 Chinese documentary records, 3:218–219, 3:220f
 coastal site proxy records, 3:211–213
 geological proxy records, 3:209–217
 global climate changes, 3:215–217
 Guangdong typhoon time series, 3:218–219, 3:220f
 historical documentary records, 3:217–219
 overwash sand layers, 3:209–211, 3:212f, 3:213f, 3:215f
 research needs/future directions, 3:219–220
 Spanish archives, 3:217–218
 stable isotopes, tree rings/stalagmites, 3:214–215
 US data sources/reconstructions, 3:217
- paleothermometry
 amino-acid dating, 1:45–46
 borehole temperature records, 2:299
 isotopic paleothermometer, 2:395–396
 Mg/Ca, 2:871–883, 2:936
 see also alkenone paleothermometry
- paleovegetation interpretations, Northern Asia, 4:35
- paleovegetation maps, ecosystem dynamics, 3:251f
- paleowater $\delta^{18}\text{O}$ determination, 1:298
- paleowind loess deposits, 3:359
- paleozoogeography, European loess records, 2:613–615
- Palliser Bay, New Zealand, 1:247, 1:248f
- Palmer Deep, Antarctic Peninsula, 3:80–81, 3:81f, 3:82–84
- Palmer Drought Indices, 1:463–464, 1:465f, 1:467–468, 3:89f, 3:196f, 3:197

- Palmer Drought Severity Indices (PDSI), 1:463–464, 1:465*f*, 1:467–468, 3:89*f*
- Palouse loess, 2:625*f*, 2:626, 4:255
- palsas, 3:477–478, 3:477*f*
- Paludella squarrosa* macrofossils, 3:649–651, 3:651*f*
- palustrine records, 4:363
- palynology
- Australasian Late Pleistocene, 4:18–26
 - biodiversity richness, 3:818, 3:819–820, 3:819*f*
 - databases, 3:831–838
 - dinoflagellates, 2:800, 2:803, 2:805, 2:811
 - Latin America, 2:531, 2:533–534
 - northeastern North America, 4:115–117
 - Northern Asian Late Pleistocene, 4:27–38
 - Northern European records, 4:177
 - plant macrofossils combined, 3:768
 - Southern European records, 4:185
 - Western North America, 4:72–73
 - see also* forensic palynology; pollen...
- palynomorphs, dinoflagellates, 2:800–815
- Pamirs, Central Asia
- ice core studies, 2:384–385
 - loess records, 2:586*f*, 2:587–589, 2:590–591
- Pamir, Tajikistan glaciations, 2:241
- Pampa, Argentina, pollen records, 4:161, 4:161*f*
- pamphers, South America, 4:685
- pampean area, South America, 4:683
- Pampean Loess, South America, 2:629–635, 2:631*f*, 2:631*t*, 2:633*t*, 2:635*t*, 2:637, 2:637*f*, 2:638–639, 2:640
- Panama hypothesis, Late Neogene/Early Quaternary transition, 2:152–154
- Panamaian seaways, 2:151, 2:156*f*, 2:157*f*, 2:162
- see also* Central American seaways
- PAR *see* pollen accumulation rate
- parabolic dunes
- definition, 1:597*ge*
 - high latitudes, 1:599
 - Canada, 1:600–601, 1:600*f*, 1:601*f*
 - Europe, 1:601, 1:602*f* - low latitudes, 1:632*f*
- paraglacial coasts
- barrier coasts, 3:563
 - fjords, 3:562–563
 - shelf deposits, 3:563
- paraglacial geomorphology, 3:553–565
- definition, 3:553
 - land systems, 3:556–563
 - modification implications, 3:563–564
 - paraglacial succession, 3:556, 3:557*f*
 - sediments
 - cascades, 3:553, 3:554*f*
 - storage modeling, 3:554–556, 3:555*f*
 - yield models, 3:553–554 - theory, 3:553–556
- paraglacial landscapes, 3:492–493
- paraglacial land systems, 3:556–563
- alluvial systems, 3:559–561
 - coasts, 3:562–563
 - drift-mantled slopes, 3:559
 - glacier forelands, 3:559
 - lacustrine deposits, 3:561–562
 - rock slopes, 3:556–559
- paraglacial succession, 3:556, 3:557*f*
- paramagnetism, 2:884, 2:887*t*, 3:375*ge*, 3:375–376
- parasitoids, 4:555
- parent materials
- definition, 3:375*ge*
 - soils, 3:378, 3:402, 3:410*f*
- parthenogenetic life cycle, cladocera, 3:275*b*, 3:271*ge*
- partial least squares (PLS) analysis, pollen–climate relation, 3:809–810
- partial track fading (PTF), glass-FT dating, 1:649–650
- glass-DCFT, 1:650
 - glass-ITPFT, 1:650
 - plateau correction methods, 1:649–650, 1:653*t*
 - track-size correction methods, 1:649
- particle analyses, charred particles, 2:716–729
- particle forms
- cosmic radiation, 1:422–427
 - loess origins, 2:574–576, 2:576*f*
 - sediments
 - clast form analysis, 2:1–2, 2:3–5
 - defining, 2:1–2, 2:3*t*, 2:4*f*
 - influences on, 2:3–5
- particle scavenging, marine sediments, 2:925–926, 2:926*f*, 2:928*f*
- particle size
- blockfields, 3:524–526
 - rock glaciers, 3:538*f*
 - soil/sediment, 4:538
- passive shielding, radiocarbon dating, 4:307–308, 4:312
- passive transport, sediment particle forms, 2:3–5
- pasture land pollen records, 3:888*f*, 3:890
- Patagonia
- beetle fauna, 1:235, 1:242
 - cave fauna, 4:716
 - glaciations, 2:254, 2:255
 - ice sheet, 1:236*f*, 1:241
 - landscape, 1:235, 1:242
- patterning of vegetation, pollen assemblage, 3:873–874
- patterned ground, 3:452–463
- active ground, 3:452
 - circles, 3:453*f*, 3:454–457, 3:454*f*, 3:457*f*, 3:459–460
 - differential frost heave, 3:452, 3:454–457, 3:461
 - earth hummocks, 3:460–462
 - environmental significance, 3:458
 - formation of sorted ground, 3:454–458
 - frost boils, 3:459–460
 - ice-wedge polygons, 3:458–459
 - needle ice, 3:455–456, 3:456*f*, 3:458
 - nonsorted, 3:458–462
 - porewater movements, 3:458
 - soil circulation, 3:457–458
 - sorted, 3:452–458
 - stripes, 3:453*f*, 3:455–456, 3:456*f*, 3:459–460, 3:461*f*
- vegetation-defined nonsorted circles, 3:459–460
- Pb *see* lead
- PBO *see* Preboreal oscillation
- PCA *see* principal component analysis
- PCI *see* Polar Circulation Index
- pCO₂ proxies, marine carbon cycle, 2:849, 2:851*f*, 2:855, 2:856*f*
- PCR *see* polymerase chain reaction
- PCT *see* prepared core technology
- PDA methods *see* photodiode array methods
- PDD approach *see* positive degree day approach
- PDI *see* profile development index
- PDO *see* Pacific Decadal Oscillation
- PDRM *see* postdepositional remanent magnetization
- PDSI *see* Palmer Drought Severity Indices
- PDSRFE *see* Pollen Database for Siberia and the Russian Far East
- P–E *see* precipitation–evaporation
- Peace-Athabasca Delta (PAD), 3:335, 3:335*f*
- peat
- bed lithostratigraphy, 4:240
 - bogs, 3:593–594, 3:594*f*
 - fossil leaves from, 3:616–617
 - plant macrofossils, 3:593–594, 3:594*f*, 3:725, 3:726, 3:729
 - see also* mire/peat macrofossils
- peatland plant macrofossils, 3:725, 3:726, 3:729
- peccaries, South America, 4:689
- pedoanthracology *see* soil charcoal analysis
- pedocomplexes, 3:367
- pedogenesis, aggradational, 3:370*f*, 3:372–373
- pedosedimentary evolution, 3:388–389, 3:390*f*
- pedostratigraphy, 4:250–259
- alluvial sequences, 4:256–258
 - correlation, 4:254–256
 - definitions, 4:250
 - fans, 4:257–258
 - fault trenches, 4:258
 - features of, 4:251*t*
 - geoarcheology, 4:258
 - geosols, 4:254
 - laboratory studies, 4:256
 - loess, 4:254–255, 4:255–256
 - China, 4:254–255, 4:255*f*
 - Europe, 2:611*f*, 2:612*f*, 2:617*f*, 4:254
 - Palouse, 4:255 - milestones, 4:250–252
 - New Zealand, 4:252*f*, 4:253*f*, 4:255–256, 4:257*f*
 - nomenclature problems, 4:253–254
 - paleoclimatology, 4:254–256
 - tephra sequences, 4:256
- pelagic clay, 1:637*ge*
- pelts as raw materials, 4:751
- Penck, Albrecht, 1:4
- pendant bar subaerial mesoforms, 2:12
- penguin colonies, Antarctica, 2:219, 2:220*f*
- penultimate glaciation, New Zealand, 1:247

- Peoria Loess, North America, 2:620–621, 2:621–622, 2:622*f*, 2:624–625, 2:624*f*, 2:627
- peptide chains, amino-acid dating, 1:41–42
- percentage relationships, pollen–vegetation, 3:874
- percussors, definition, 1:83*ge*
- perennially frozen ground, rock glaciers, 3:535–541
- Peridinales dinoflagellates, 2:804, 2:805*f*, 2:809*f*
- periglacial features, 3:464–471, 3:535–541
 - active-layer processes, 3:421–429
 - blockfields, 3:523–534
 - cryoturbation structures, 3:430–435
 - frost mounds, 3:472–480
 - ice casts, 3:436–451
 - ice wedges, 3:436–451
 - rock glaciers, 3:535–541
 - slope deposits/forms, 3:481–489
 - thermokarst topography, 3:574–581*see also* permafrost
- periglacial fluvial sediments/forms, 3:490–499
 - future perspectives, 3:497–498
 - landforms/processes, 3:490–495
 - Pleistocene environments, 3:495–497
- periglacial landforms, 3:490–499
 - hydrology, 3:491–492
 - ice, 3:494
 - morphology, 3:490–491
 - paraglacial geomorphology, 3:553–565
 - permafrost, 3:494–495
 - processes, 3:490–495
 - rock weathering, 3:500–506
 - sediment, 3:492–493
 - talus slopes, 3:566–573
- periglacial landscape dating, 1:435
- periglacial loading, cryoturbation structures, 3:430–432, 3:431*f*, 3:432*f*, 3:433*f*
- perissodactyla, South America, 4:688
- permafrost, 3:464–471
 - active layer, 3:421–429, 3:574, 3:577*f*
 - aggradation, 3:464–465
 - ancient plant DNA, 2:707*f*
 - blockfields, 3:523–534
 - block streams, 3:514–522
 - climate change, 3:469–470
 - cryoturbation structures, 3:430–435
 - definition, 3:535
 - degradation, 3:427, 3:464–465, 3:576–577
 - distribution, 3:466
 - disturbances to, 3:575–576
 - frost mounds, 3:472–480
 - ground surface temperature, 3:465–466, 3:470
 - hillslopes, 3:468, 3:469
 - ice casts, 3:436–451
 - ice wedges, 3:436–451
 - insect fossil studies, 1:173–174
 - interactions, 3:507–513
 - basal processes, 3:508
 - cold-based ice, 3:507–508, 3:510
 - further research, 3:512
 - geological evidence, 3:510–512, 3:511*f*
 - geomorphological evidence, 3:508–510
 - glacial environments, 3:507–508
 - ice-marginal landforms, 3:508–509
 - proglacial landforms, 3:508–509
 - subglacial landforms, 3:509–510
 - landforms, 3:469, 3:508–509, 3:509–510
 - patterned ground, 3:452–463
 - periglacial fluvial sediments/forms, 3:494–495, 3:498
 - plant macrofossil records, 3:733, 3:734*f*, 3:735–736, 3:736*f*
 - processes, active layer, 3:421–429
 - rock glaciers, 3:535–541
 - slope deposits/forms, 3:481–489
 - thermal regime, 3:464, 3:465*f*
 - thermokarst topography, 3:574–581
 - Yedoma deposits, Beringia, 3:542–552
- persistent organic pollutants (POPs), 3:266, 3:297, 3:298*f*
- Peru
 - glaciations, 2:252, 2:253
 - ice core studies, 2:387–388, 2:388*t*
 - South Pacific ocean postglacial records, 3:76–77, 3:79*f*, 3:84*f*
 - vertebrate fossils, 4:681
- Peterborough, England beetle assemblages, 1:187
- Peterson, RA, 3:459–460
- petrographic microscopes, 4:539, 4:539*f*
- petrography, 2:948, 4:539, 4:539*f*
- Peyto Glacier, Canada, 1:466, 1:466*f*
- P forms
 - meltwater hypothesis, 1:833–834
 - micro/macroscale erosional forms, 1:871
- PFTs *see* plant functional types
- PGMs *see* platinum group metals
- pH
 - amino-acid dating, 1:41
 - B/Ca proxy, marine carbon cycle, 2:851–853
 - boron isotope-pH proxy, marine carbon cycle, 2:849–851
 - foraminifera Mg/Ca, 2:874
 - lake study multiproxy approaches, 3:340–341
 - oxygen isotope carbonate composition, 1:338
- phase relationships, methane studies, 2:339
- phenotypic change, vertebrates, 4:723
- Philippine vertebrate records, 4:663
- Phnom Loang cave, Cambodia, 4:657
- Phoridae (scuttle flies), 4:551
- phosphorus (P)
 - biogeochemical role in climate cycles, 3:416–417
 - biogeography, 3:320*f*
 - lake eutrophication, 3:293–294
- photodiode array (PDA) methods, pigment analysis, 3:328–329
- photoprotective compounds, pigment studies, 3:327
- photosynthesis feedbacks, 1:721
- photosynthetic organisms
 - light-absorbing properties, 3:327–328
 - pigments, 3:326–338
- Phragmites australis* macrofossils, 3:642, 3:643*f*
- phreatic overgrowths on speleothems (POS), 4:464
- phylogenetic analysis, ice cores, 2:292
- phylogenetic relationships, vertebrates, 4:721
- phylogenic evolution, diatoms, 1:586, 1:586*f*
- phylogeography, ancient DNA, 4:720–721
- physical anthropology
 - American human colonization, 1:125–126
 - Australian human colonization, 1:111–113
- physical geography, Africa, 4:664
- physical processes, active layer, 3:422–424
- physical proxies
 - paleoceanography, 2:871–883, 2:884–898, 2:915–922, 2:923–931
 - terrigenous sediments, 2:941–952
- physical setting, tropical South America, 2:387–390
- physical weathering, 3:402
 - see also* mechanical weathering
- physicochemical behavior, oxygen isotopes, 2:915
- physics, cosmic ray units, 1:429
- physiography, northeastern North America, 4:115
- physiology of plants, 3:613–614
- phytoliths, 3:582–592
 - analytical techniques, 3:585
 - archeological research, 3:589–590
 - ¹⁴C of plant macrofossils, 4:362
 - characteristics, 3:583–584
 - chemical analysis, 3:587
 - chronology, 3:585–588
 - classification, 3:584–585
 - composition, 3:583–584
 - formation, 3:582–583
 - function, 3:583–584
 - international naming, 3:587*t*
 - morphology, 3:584–585, 3:586*f*
 - paleoenvironmental reconstruction, 3:588–589
 - preservation, 3:584
 - radiocarbon dating, 3:585–586
 - representativeness, 3:584
 - stable isotope analysis, 3:587–588
 - stratigraphic frequency diagram, 3:589*f*
 - terminology, 3:584–585, 3:587*t*
 - TL dating, 3:587
 - West African grassland, 3:590*f*
- phytoplankton
 - carbon cycle diatom silica, 2:734
 - dinocysts and, 2:812
 - lake diatom silica, 2:740
 - productivity, North Pacific Ocean, 3:34, 3:40
- Picea abies* (spruce)
 - paleophytogeography, 2:732
 - plant distributions, 3:856, 3:857*f*
 - pollen records, North America, 4:116*f*, 4:117–118
 - wood, megafossils, 3:621, 3:622*f*, 3:625–626, 3:626*f*, 3:626*t*, 3:627*t*
- Pickering, KT, 2:132–134
- 'pickled' fauna, Europe, 4:716
- pigments
 - analysis of, 3:328–330

- pigments (*Continued*)
 as biomarkers, 3:326
 depositional systems, 3:330–336
 long-term records, 3:336, 3:337f
 photosynthetic organisms, 3:326–338
 preservation, aquatic environments, 3:330
 sedimentary, paleolimnology, 3:268
 structure/nature of, 3:326–328
- pillow mounds, 1:793, 1:795–796
- pine *see* *Pinus*
- pingos, 3:467, 3:469, 3:470f
 collapse, 3:475–476, 3:476f
 degradation, 3:475–476
 formation, 3:472–476
- Pinjor, India, 4:651
- Pinus* (pine)
 plant macrofossils, 3:600f, 3:608,
 3:642–646, 3:645f
 pollen records
 North America, 4:116f, 4:118–119
 Southern Europe, 4:179–182, 4:183,
 4:185
 wood, megafossils, 3:621, 3:623f,
 3:625–626, 3:625f, 3:627f, 3:627t
- Pinus sylvestris* macrofossils, 3:642–646,
 3:645f
- pioneer vegetation
 Arctic Eurasia, 3:742, 3:743
 Greenland, 3:763, 3:764
- Piophilidae (cheese/skipper flies), 4:551
- Pipe Clay Lagoon, Tasmania, 1:197
- Piscicola geometra* macrofossils, 3:652, 3:653f
- piston corer paleolimnology, 3:262f, 3:264
- Pithecanthropus erectus*, 1:137
- planktic/planktonic foraminifera, 2:825–829
 applications, 2:828
 biostratigraphy, 2:827–828, 2:827f
 classification, 2:825
 common species, 2:826f
 description, 2:825
 distribution of living foraminifera,
 2:825–827
 historical background, 2:825
 Indian Ocean, 3:46–47, 3:48f
 Mg/Ca paleothermometry, 2:871–873
 North Pacific Ocean, 3:34–35, 3:40–42,
 3:42f
 seafloor test distribution, 2:827
 stratigraphic distribution, 2:827–828
see also temperature proxies
- planktonic diatoms, 1:473–474
 Lake Baikal, 1:548, 1:549, 1:549f
 Lake Titicaca, 1:550
 North Atlantic/Arctic, 1:565f
 as paleoenvironmental proxies, 1:546
 salinity/climate reconstruction,
 1:509–510, 1:509f
- plant anatomy, forensics, 4:543–544
- plant associations, no-analog, 4:120
- plant–beetle reconciliation, Late Quaternary,
 1:197–198
- plant distributions, 3:854–860
 abundance, 3:854–860
 gaining knowledge, 3:854–855
 influential factors, 3:859
 Late Quaternary spreading, 3:855–857
- models, 3:858f
 Northern Hemisphere, 3:859
 speed/patterns, 3:857–859
 tree spread, 3:859
- Plant DNA C-values Database, 4:560
- plant ecology, forensics, 4:543, 4:545–546
- plant functional types (PFTs)
 biome relationships, 3:869
 climate–biome approach, 3:861–862
 modern analog technique, 3:808
 pollen databases, 3:836
- plant growing, 3:706–708
see also cultivation of crops
- plant macrofossils, 3:593–612
 alpine plants, 3:605f, 3:606f
 analysis, 3:595–596, 3:609–610
 aquatic plants, 3:600f, 3:601f, 3:602f,
 3:603f
 Arctic, 3:605f
 Eurasia, 3:733–745
 North America, 3:746–759
 basics, 3:699–705
¹⁴C dating, 4:361–367
 Caryophyllaceae, 3:606f, 3:608
 CO₂ reconstruction, 3:613–620
 Cyperaceae, 3:604f, 3:608
 data interpretation, 3:596–607
 definition, 3:593
 dendroarcheology, 3:630–636
 environmental archeology use, 3:699–724
 fen plants, 3:602f, 3:603f
 future research, 3:610
 Greenland, 3:760–767
 Holocene North America, 3:768–784
 identification, 3:596
 late glacial change, 3:785–793
 leaves, 3:597f, 3:613–620
 locations, 3:593–594
 megafossils, 3:621–629
 mire/peat macrofossils, 3:637–656
 multidisciplinary studies, 3:785–793
 paleoecology, 3:593, 3:608
 paleolimnological applications,
 3:657–673
- pollen
 study validation, 3:725–732
 uninformative situations, 3:608–609
- processes in studies, 3:595f
 range extensions, 3:757
 reasons for studying, 3:593
 representation, 3:684–689
 rodent middens, 3:674–683
Salix herbacea, 3:597f, 3:610, 3:610f
 sampling, 3:594–595
 Saxifragaceae, 3:606f, 3:608
 shrubs, 3:598f, 3:599f
 surface samples, 3:684–689
 taphonomy, 3:684–689
 treeline studies, 3:690–698
 trees, 3:598f, 3:608
- types
 of assemblage, 3:702–703
 of remains, 3:699
 vegetation history, 3:608
 weeds, 3:607f
see also plant remains
- plant physiology, CO₂ reconstruction,
 3:613–614
- plant remains
 analysis/identification, 3:703–704
 macroscopic forensic use, 4:542–547
 recovery of, 3:703–704
see also plant macrofossils
- plant taxonomy
 forensics, 4:543, 4:544
 pollen databases, 3:834–836
- plasma ashing, charcoal, 4:357
- plasmic fabric analysis, glacial sediment,
 2:53f, 2:54f, 2:58–60, 2:58f
- plateau glaciers, 1:911–912, 1:912f
- plateau ice field landsystems, 1:814–817,
 1:819f, 1:822
- plateau uplift, 2:128–129
- plate margins
 passive, 1:688–689
 tectonically active, 1:686–688
see also continental margins; continental
 shelf...
- plate tectonics, 1:686–688, 4:371
see also tectonic processes
- platform erosion marks, 4:378–379
- platinum group metals (PGMs), 2:345
- playa lakes, Australia, 2:528
- Pleistocene
 African hominid teeth/bones, 1:310
 Antarctic glaciation, 2:217–218
- Arctic
 diatom record, 1:562–565
 Eurasian plant macrofossil records,
 3:733, 3:739
 North America, 3:753
- beetle fauna
 Australia, 1:191–199
 Europe, 1:184–190, 1:200–206
 fossils, 1:162, 1:166
 New Zealand, 1:244–254
- blockfield origin, 3:529–530
 chironomid records, Europe, 1:373–385
 coccolithophores, 2:791, 2:792
 definition, 2:216ge
 diatom record, North Atlantic/Arctic,
 1:562–565
 flora/fauna changes, 4:206, 4:207, 4:209
 glacial lake outbursts, 1:825–826,
 1:826–828
- glaciations
 Antarctic, 2:217–218
 Eurasia, 2:172–179
 Quaternary science history, 1:3–4
 Southern Hemisphere, 2:187–190
- GSSP, 4:191–192
 history of Quaternary science, 1:1–2, 1:3–4
 Holocene transition, 1:679–680
 hominin framework, 1:50t
 human dispersals/diasporas, 1:53–54
 ice-margin sites, 2:420–423
 insect fossil studies, 1:173, 1:174–176
 Japan, 1:207–220
 Northern Asia, 1:255–273
 South America, 1:235–243
 loess records, Central Asia, 2:589–590,
 2:590–591, 2:592

- mammals
 history of, 1:2–3
 mummified, 4:713–717
 marine oxygen isotope stratigraphy, 2:908–909, 2:909f
 megafaunal extinctions, 1:23–24, 4:700–712
 non-lacustrine terrestrial studies, 1:328–330
 North Atlantic
 diatom record, 1:562–565
 evolution, 3:9–10
 NWN pollen records, 4:125–126, 4:128, 4:129
 overkill hypothesis, 4:709
 paleoceanography overview, 2:747–748
 periglacial fluvial sediments/forms, 3:495–497
 permafrost–glacial interactions, 3:511
 plant macrofossil records, Arctic Eurasia, 3:733, 3:739
 Pliocene transition to, 2:217–218
 relative sea level changes, 4:504f
 sequence stratigraphy, 4:269–271, 4:271f, 4:272f
 strandlines, sea-level change, 4:503–504, 4:511–515
 terminology use, 4:673
 varved sediments, 4:574f, 4:583f
 vertebrate studies, 4:599–604
 Africa, 4:615–619, 4:664–672
 distribution of remains, 4:591–592, 4:591f, 4:593
 Europe, 4:639–645
 mummified mammals, 4:713–717
 North America, 4:673–679
 Northern Eurasia, 4:605–614
 South America, 4:680–692
 Southeast Asia, 4:693–699
 Southern Asia, 4:651–663
see also Early Pleistocene; Late Pleistocene; Middle Pleistocene; Pleistocene Series
 Pleistocene Series, 4:216–217, 4:217–218
 chronostratigraphy, 4:216–217, 4:217–218
 Lower Pleistocene, 4:217
 Middle Pleistocene, 4:217–218
 Upper Pleistocene, 4:218
see also Pleistocene
 Pleniglacial change, plant macrofossils, 3:786–787
 Pliocene, 3:185–193
 African hominid teeth/bones, 1:310
 Antarctic glacial history, 2:217–218
 Arctic
 Eurasian plant macrofossil records, 3:738–739
 North America, 3:749–753
 paleoclimate history, 3:113–114
 chronostratigraphy, 4:216–217
 definition, 2:216ge
 Greenland sea-level changes, 4:481–482
 insect fossil studies, 1:174–176
 Late Neogene/Early Quaternary transition, 2:151
 mechanisms for warmth, 3:186–188
 NHG
 causes, 3:189–191
 human evolution, 3:192
 intensification, 3:188
 paleoceanographic records, 3:1
 paleoclimates
 history of, 3:113–114
 introduction to, 3:87–92
 reconstruction, 3:185–193
 Pleistocene boundary/transition, 2:151, 2:217–218, 4:599
 PRISM project, 3:185–186, 3:188
 vertebrate studies, 4:599
 warm climates, 3:185–186, 3:186–188
 Pliocene Research, Interpretation and Synoptic Mapping (PRISM) project, 3:185–186, 3:188
 Pliocene Series, 4:216–217
see also Pliocene
 ploughing till, 2:65
 PLS analysis *see* partial least squares analysis
 plucking/entrainment, micro/macro scale erosional forms, 1:867, 1:869
 PMI *see* postmortem interval
 PMIP *see* Paleoclimate Modeling Intercomparison Project
 PNA *see* Pacific North American patterns
 Poland, ice sheet/glaciations, 2:229
 Polar air mass system, 4:142–143
 polar bear species origins, 4:724, 4:725
 Polar Circulation Index (PCI), lake levels, 2:555–556, 2:555f
 polar ice cap landforms, 1:781
 polar ice cores
 Asian records, 2:379–380
 CH₃Br measurements, 2:468f
 CH₃Cl measurements, 2:467f
 CO₂ studies, 2:311–318
 CO measurements, 2:464f
 COS measurements, 2:466f
 future of research, 2:437–438
 Quaternary climatic change, 1:26–27, 1:27–29, 1:29–31
see also ice cores
 polarity reversals, 1:705, 1:716–718, 1:718–719
 polar jet stream, 3:167–168
 polar region stable isotope distribution, 2:347–348
 polar seas *see* Antarctic...; Arctic...
 poleward heat transport (PW), NHG causes, 3:190f
 police forensic geology use, 4:521–534
 polish, micro/macro scale erosional forms, 1:871, 1:873f
 pollarding, 3:891
 pollen
 Africa, 4:9–17
 Afromontane region, 4:12–13
 arid/semiarid biomes, 4:14
 atmospheric circulation, 4:85–88
 ‘cool’ tree species, 4:91–95, 4:97f
 Eastern Africa, 4:95–98, 4:96f
 freshwater algae records, 4:88–89
 global warming, 4:89–98
 Guineo–Congolian forests, 4:91–95
 Humid Period, 4:88–89, 4:88f, 4:98–100, 4:100f
 Last Deglaciation onset, 4:89–98
 Late Pleistocene, 4:9–17
 LGM vegetation map, 4:11f
 marine pollen sequences, 4:85–88, 4:88–89
 Mediterranean-climate regions, 4:14–15
 monsoon activity, 4:89–98
 postglacial period, 4:85–103
 rainfall variations, 4:88–89
 Saharan Desert, 4:89–91, 4:92t, 4:100
 savanna woodlands/grasslands, 4:13–14
 tree pollen types, 4:92t
 tropical forests, 4:10–12, 4:12f, 4:89–91, 4:98–100
 upwards forest expansion, 4:95–98
 vegetation change, 4:89–98, 4:89f
 vegetation map, 4:10f, 4:11f
 West Africa, 4:85–88, 4:88–89, 4:88f, 4:89–91, 4:89f, 4:95f, 4:97f, 4:100f
 analysis, 3:794–804
 applications, 3:799–802, 3:880–903
 archeological applications, 3:880–903
 climate reconstructions, 3:806
 continental-scale studies, 3:801–802
 dispersal, 3:796–799
 future directions, 3:802–803
 Greenland, 3:760, 3:761, 3:763–765, 3:765–766
 lake level studies, 2:531, 2:533–534
 late glacial multidisciplinary studies, 3:785
 local-scale studies, 3:799–800
 long pollen records, 3:801–802, 3:803f
 macrofossil representation, 3:688–689
 marine sediment overview, 4:200–201
 model-based correction approaches, 3:796–799
 plant distributions, 3:854–855
 plant macrofossils combined, 3:768, 3:773f, 3:774f, 3:775
 plant–pollen representation, 3:795–796
 POLLSCAPE model, 3:871–879
 preservation, 3:795
 principles, 3:794–804
 prospects, 3:901–902
 regional-scale studies, 3:800–801, 3:803f
 sample sizes, 3:797f
 sedimentation, 3:795
 spores, 3:794–795, 3:795f
 stand-scale records, 3:798t, 3:802f, 3:803f
see also pollen, numerical analysis
 archeological applications, 3:880–903
 Australia
 anthropogenic impacts, 4:112
 beetle reconciliation, 1:197–198
 climate events, 4:112
 Holocene, 4:106–109, 4:107f, 4:109f
 postglacial period, 4:104–114
 recovery from LGM, 4:106
 biodiversity reconstruction, 3:816–820
 biome vegetation reconstruction, 3:861–870

- pollen (*Continued*)
- ¹⁴C of plant macrofossils, 4:362–363
 - chironomid records
 - postglacial Southern Hemisphere, 1:401, 1:402
 - treeline altitudes and, 1:393–394
 - climate proxies, 3:805–815
 - Colombian savannas, 4:58–59
 - data, 2:699–704
 - calibration, land cover, 3:892–895
 - classifying/mapping biome approach, 3:862–863
 - climate reconstructions
 - advantages/disadvantages, 3:806–807
 - examples, 3:810–814
 - Fennoscandia, 2:703f
 - future trends, 2:704
 - global databases, 2:704
 - history of, 2:699–700
 - Holocene
 - southwestern North America, 4:143–144
 - treeline dynamics, 3:695f, 3:696f
 - human impact reconstructions, 3:881–901
 - key insights, 2:701–704
 - lake-level fluctuations, 2:552–553
 - new/exotic species, 2:703
 - nonanalog plant communities, 2:701
 - overview, 2:699–704
 - paleoclimatic reconstruction, 3:810–814
 - recent developments, 2:700–701
 - tree distribution dynamics, 2:701–703
 - variable scales, 2:701, 2:701f
 - vegetation reconstruction biome
 - approach, 3:861–870
 - vegetation–climate equilibrium, 2:703
 - see also* pollen, numerical analysis
 - databases, 2:704, 3:731–732, 3:831–838
 - applications, 3:836–837
 - history, 3:831–833
 - nomenclature/taxonomy, 3:834–836
 - quality control, 3:834
 - structure, 3:833–834
 - dispersal/deposition models, POLLSCAPE, 3:871–872
 - fossils, paleodroughts, 3:195
 - grains, plant distributions, 3:854
 - Last Interglacial of Europe, 4:1–8
 - depositional environments, 4:1–3
 - duration/timing of events, 4:7
 - Holocene comparisons, 4:6–7
 - pollen zonation schemes, 4:3t
 - sea-level highstands/climatic optima, 4:7f
 - vegetation character, 4:6–7
 - vegetation development, 4:1–3
 - Late Pleistocene
 - Australasia, 4:18–26
 - North America, 4:39–51
 - Northern Asia, 4:27–38
 - South America, 4:52–62
 - Southern Europe, 4:63–71
 - loading, POLLSCAPE model, 3:871, 3:873–874, 3:875f
 - megafossils, 3:624
 - microfossil-based reconstruction,
 - Holocene RSL, 4:422, 4:423f
 - New Zealand
 - anthropogenic impacts, 4:112
 - climate events, 4:112
 - Durham Rd, North Island, 4:107f
 - Holocene, 4:109–112, 4:109f, 4:111f
 - postglacial period, 4:104–114
 - recovery from LGM, 4:106
 - Northern Europe postglacial period, 4:173–178
 - northern North America, 4:39–51, 4:115–123
 - Alaska/Yukon, 4:39–44, 4:47
 - Canada, 4:44–47
 - HTM vegetation, 4:47–49
 - Late Pleistocene, 4:39–51
 - LGM, 4:47–49
 - map of localities, 4:40f
 - summary of records, 4:39–47
 - see also* pollen, NWNA
 - numerical analysis, 3:821–830
 - causative factors, 3:827
 - climate reconstructions, 3:826
 - data, 3:825–826
 - collection/assessment, 3:822–823
 - geographic sets, 3:824
 - interpretation, 3:826–827
 - paleoenvironmental books, 3:828t
 - single data sets, 3:823–824
 - summarization, 3:823–824
 - error estimation, 3:822–823
 - future developments, 3:827–828
 - geographic data sets, 3:824
 - identification, 3:822
 - paleoenvironmental data books, 3:828t
 - palynological richness, 3:825
 - population analysis, 3:825
 - rate of change analysis, 3:825
 - reference texts, 3:822t
 - single data sets, 3:823–824
 - software/computing, 3:827, 3:827t
 - stratigraphic sequences, 3:823, 3:824
 - time series analysis, 3:825–826
 - vegetation reconstruction, 3:826–827
 - NWNA
 - Alaska, 4:124–127
 - climate patterns, 4:124
 - Cold Desert, 4:130
 - Holocene, 4:125–126, 4:126–127, 4:128–129, 4:129–130
 - LGM, 4:124–125, 4:127–128, 4:129
 - map of, 4:125f
 - marine west coast forest, 4:127–129, 4:128f
 - Pleistocene, 4:125–126, 4:128, 4:129
 - postglacial period, 4:124–132
 - vegetation patterns, 4:124
 - Western Cordillera, 4:129–130
 - percentage vegetation relationships, 3:874
 - plant distributions, 3:854–860
 - POLLSCAPE model, 3:871–879
 - postglacial
 - North America, 4:115–123, 4:124–132, 4:133–141, 4:142–155
 - northern Asia, 4:164–172
 - Northern Europe, 4:173–178
 - North Pacific, 3:68
 - South America, 4:156–163
 - Southern Europe, 4:179–188
 - productivity estimates, POLLSCAPE model, 3:875–877
 - representation
 - stand-scale palynology, 3:846–847
 - vegetation/land cover, 3:871–879
 - slide approach, charred particles, 2:717–719, 2:720
 - source area, POLLSCAPE model, 3:872–873
 - southeastern North America
 - deglaciation, 4:137, 4:138f
 - Early Holocene, 4:137
 - future research, 4:140
 - glacial/deglacial records, 4:138f
 - Holocene, 4:137–140, 4:139f
 - lakes/estuaries/wetland records, 4:136–140
 - Late Holocene, 4:137–140
 - LGM, 4:136–137
 - map of sites/site names, 4:134f, 4:134t
 - modern forest distribution, 4:133–135
 - modern vegetation pollen signatures, 4:135–136
 - physiographic regions, 4:135f
 - postglacial period, 4:133–141
 - Southern Europe
 - Andorra, 4:183–185
 - Balearic Islands, 4:179–183, 4:180t
 - Balkan peninsula, 4:181t, 4:186–187
 - climate cycles, 4:67–68
 - Corsica, 4:180t, 4:183–185
 - Crete, 4:181t, 4:186–187
 - current knowledge, 4:179
 - Dalmatian Islands, 4:181t, 4:186–187
 - France, 4:65–66, 4:180t, 4:183–185
 - Greece, 4:63
 - Iberian Peninsula, 4:179–183, 4:180t
 - ice age beginning, 4:66–67
 - Italy, 4:64–65, 4:181t, 4:185–186
 - Late Pleistocene, 4:63–71
 - major sites map, 4:64f
 - Mediterranean climate, 4:66–67
 - orbital forcing, 4:67–68
 - Pleistocene extinctions, 4:67
 - postglacial period, 4:179–188
 - refuges for vegetation, 4:67
 - research history, 4:63–66
 - Sicily, 4:181t, 4:185–186
 - Spain, 4:65
 - suborbital forcing, 4:68–69
 - vegetational events, 4:184t
 - vegetation trends, 4:66–69
 - southwestern North America, 4:142–155
 - stand-scale palynology, 3:846–853
 - surface samples, 3:839–845
 - anthropogenic disturbance, 3:843–844
 - archiving, 3:842–843
 - climate reconstructions, 3:844
 - collection strategy, 3:842–843
 - data applications, 3:843–844
 - databases, 3:843

- dispersal/deposition models, 3:844
land use in past, 3:843–844
need for, 3:839
vegetation history, 3:843
- trapping, 3:839–845
anthropogenic disturbance, 3:843–844
archiving, 3:842–843
climate reconstructions, 3:844
collection strategy, 3:842–843
data applications, 3:843–844
databases, 3:843
dispersal/deposition models, 3:844
land use in past, 3:843–844
need for, 3:839
vegetation history, 3:843
- uninformative plant macrofossil situations, 3:608–609
- validation, 3:725–732
- vegetation relationships
northeastern North America, 4:116–117
percentage relationships, 3:874
POLLSCAPE model, 3:871–879
Western North America, 4:72–83
see also palynology
- pollen accumulation rates (PAR)
analysis, 3:799, 3:801f
data overview, 2:702f
numerical analysis, 3:821, 3:823, 3:825
surface sampling/trapping, 3:842, 3:842f, 3:843t
- Pollen Database for Siberia and the Russian Far East (PDSRFE), 3:831, 3:832, 3:833
- POLLSCAPE model, 3:846–847, 3:871–879
- pollution
atmospheric, biogeography, 3:319–321
glaciochemistry, ice core studies, 2:332
heavy metals, 2:344–345
lakes
chemistry, 3:297, 3:298f
chironomids and, 1:386–387
monitoring, benthic foraminifera, 2:772
POPs, lake chemistry, 3:297, 3:298f
see also contamination
- polygenetic formations, 1:780ge
- Polykrikos* dinocysts, 2:811f
- polymerase chain reaction (PCR), 4:719–720
- Polynesian settlements, 4:112
- polysaline water, 1:361ge
- polythermal glacial landsystems, 1:817–820
- polythermal glaciers, 3:508
- Polytrichum* macrofossils, 3:649, 3:650f, 3:651f
- pool erosion marks, sea level indicators, 4:379
- Pool's Farm Pit, England beetle assemblages, 1:186
- POPs *see* persistent organic pollutants
- population dispersal
archeology overview, 1:53–54
Neanderthal demise, 1:146–153
see also dispersal of species
- population extinction *see* extinctions
- population genetics, alkenone paleothermometry, 2:761
- population-level aDNA research, 4:721, 4:722
- population replacements, human evolution, 1:146, 1:150–152
- Populus tremula* macrofossils, 3:642–646, 3:644f
- pore pressure solifluction, 3:482f
- porewater movements, 3:457f, 3:458
- porosity, firm air, 2:361–362, 2:362–363, 2:364f
- Portugal, sea level changes, 4:493
- Portuguese margin diatoms, 1:567
- POS *see* phreatic overgrowths on speleothems
- positive degree day (PDD) approach, surface mass balance, 2:441
- possums, 4:634
- post-Anglian Middle Pleistocene beetle assemblages, 1:187
- postburial alteration, paleosols, 3:372–373
- postdepositional processes
cosmogenic radionuclides, 2:355–356
varved sediment diatom records, 1:544
- postdepositional remanent magnetization (PDRM), 1:710–712, 1:713
- postglacial
African pollen records, 4:85–103
Australian pollen records, 4:104–114
beetle records
Europe, 1:274–281
North America, 1:282–290
Europe
beetle records, 1:274–281
chironomid records, 1:386–397
human adaptations, 1:155–159
pollen records, 4:173–178, 4:179–188
human adaptations, 1:154–159
Europe, 1:155–159
world overview, 1:154
Indian Ocean, 3:46–54
New Zealand pollen records, 4:104–114
North Atlantic, 3:55–61
North Pacific paleoceanographic records, 3:62–72
NWN pollen records, 4:124–132
paleoceanography overview, 2:753–754
pollen records
Africa, 4:85–103
Australia, 4:104–114
New Zealand, 4:104–114
northeastern North America, 4:115–123
northern Asia, 4:164–172
Northern Europe, 4:173–178
NWN, 4:124–132
South America, 4:156–163
southeastern North America, 4:133–141
Southern Europe, 4:179–188
southwestern North America, 4:142–155
Southern Hemisphere chironomid records, 1:398–405
South Pacific ocean records, 3:73–85
- postmortem interval (PMI) entomology use, 4:548, 4:549, 4:555
- postnuclear tracers, radioactive nuclides, 1:449
- postsearch phase, forensic searches, 4:529
- Potamogeton*
aquatic plant biogenic carbonates, 1:346–347, 1:347f
mire/peat macrofossils, 3:645f, 3:646
seeds, ¹⁴C, 4:364
- potassium–argon dating *see* K–Ar dating
- Potentilla palustris* macrofossils, 3:642, 3:643f
- pothole erosion marks, sea level indicators, 4:379
- Praemegaceros*
biochronology, 4:639
Early Middle Pleistocene, 4:641–642, 4:641f
reconstruction, 4:640f, 4:641f
Villafranchian–Galerian transition, 4:639
- prairie-forest border
Midwestern United States, 3:768–769, 3:773f
speleothem isotopes, 3:768–769, 3:773f
- prairie-forest ecotone, 3:769f, 3:773f, 3:777, 3:783
- prairie plants, Holocene, 3:768, 3:771t, 3:776f, 3:778f
- pre-Anglian Middle Pleistocene beetle assemblages, 1:184–186
- Preboreal oscillation (PBO)
CO₂ reconstruction, 3:617
hydrological patterns, 2:551–552, 2:553f
- Preboreal–Younger Dryas transition, 1:275
- precipitation
active-layer thickness, 3:425
Asian lake level studies, 2:506, 2:511, 2:512f, 2:517–518
carbonate stable isotopes, 1:333, 1:334–335
deltas, 1:699
isotopic composition, 2:395–396
isotopic ratios, seawater, 2:916–917
last millennium climate, 3:230–231
Late Pleistocene Atlantic, 3:22–24
local water budget, 2:506, 2:511, 2:512f, 2:517–518
North American beetle studies, 1:227, 1:228
salinity proxies, 2:934f
seasonality, United States, 3:104–105, 3:104f, 3:105f
temperature relationship, 1:911–912, 1:912–913, 1:912f
Tibetan Plateau, 2:379, 2:383–384
tropical South America, 2:389, 2:390–391
see also monsoons; rainfall
- precipitation–evaporation (P–E), Asian lake level studies, 2:506, 2:511, 2:512f, 2:517–518
- precision, radiocarbon dating, 4:325, 4:325f
- predators, forensic entomology, 4:548
- pre-eLGM beetle fauna, New Zealand, 1:247–248
- pre-glacial changes, Indian Ocean, 3:46–47
- prehistoric human settlements, 2:549–551, 3:276
- preindustrial atmospheric trace gases, 2:463, 2:465–466

- pre-LGM human colonization, Americas, 1:122–124
- pre-Little Ice Age glacier variations, 2:274
- Prentice–Sugita pollen dispersal model, 3:796–798
- prepared core technology (PCT)
definition, 1:83ge
Europe, 1:86t
see also Levallois core technique
- Pre-Pleistocene blockfields, 3:530
- pre-Quaternary
climate, glacial lithofacies, 2:28
glacial history, Antarctica, 2:217–218
- presearch phase, forensic searches, 4:527–529
- Preservation Index, oribatid mites, 2:695, 2:696f
- preservation methods/potential
ancient DNA, 4:720
facies, 4:391
fossil beetles, 1:255
frozen mummies, 4:716
pigments, 3:330
plant macrofossils, 3:684, 3:687, 3:699–702, 3:702f, 3:706f, 3:707t
varved sediments, 4:582–583
vertebrate remains, 4:675–676
- primary dune fields, 1:598–599
- primary energy spectra, GCR, 1:420
- primary orbital oscillations, marine oxygen isotopes, 2:912
- primary productivity indicators, diatoms, 1:530–533
- primary species
carcase processing, 4:751
cave art, 4:752
- primates
dwarfing/gigantism, 4:741–742
South American Late Pleistocene, 4:684
- principal component analysis (PCA), sediment, 3:351–352
- Principle of Uniformitarianism, freshwater mollusks, 3:284
- PRISM *see* Pliocene Research, Interpretation and Synoptic Mapping project
- probing, forensic searches, 4:530, 4:531f
- proboscideans
dwarfing/gigantism, 4:734–736, 4:742–743
Late Pleistocene South America, 4:690
Proboscia curvirostris, 1:562–563, 1:563–564, 1:565f
Villafranchian stage, Eurasia, 4:600–601
- processing techniques, crops
economics, 3:708–713, 3:713–714
stages of, 3:718f, 3:720f
- process modeling, paleoclimate
reconstruction, 3:183
- process–event–response, geomorphic processes, 2:93–94, 2:94t
- Procladius* chironomid records, 1:393
- production
of macrofossils, 3:686–687
see also food production
- productivity
estimates, pollen, 3:875–877
- indicators
diatom studies, 1:530–533, 1:556–559
dinoflagellates, 2:800, 2:812
paradox, Late Pleistocene, 4:28–29
variations, Indian Ocean, 3:52
- profile development index (PDI), soils, 3:409–410
- profiles
blockfields
characteristics, 3:523–524
weathering, 3:527t
weathering, 3:392–401, 3:527t
- profundal habitats, 1:362ge
- proglacial glaciectonics, 1:839, 1:840f
- proglacial lakes, 1:882
- proglacial landforms, 2:6–12, 2:8f, 2:14f
giant fluvial bedforms, 2:9–10
permafrost–glacier interactions, 3:508–509, 3:510f
sandar, 2:6–9, 2:8f, 2:14f
subaerial mesoforms, 2:10–12
see also glaciifluvial landforms
- proglacial sites, ELA calculations, 1:916
- proluvial hypothesis, loess origins, 2:585–587
- propagation, thawing/freezing fronts, active layer, 3:422
- protactinium-231, 2:927–929, 2:928f, 2:929–930, 2:930f
- protactinium (Pa)
radioisotope proxies, 2:927–929, 2:928f, 2:929–930, 2:930f
U-series dating, 4:567–568, 4:569–570
- protalus forms, 3:535–541
- protalus ramparts, talus slopes, 3:571
- protein, ancient plant DNA, 2:713
- protherotheres, South America, 4:687
- Protista organisms, 2:803–804
see also dinoflagellates
- proton spectra
GCRs, 1:423f
SCRs, 1:423f
solar modulation intensities, 1:421f
- provenance questions, diatom samples, 1:523–524
- proxies
beetle fossils, 1:161–162, 1:166–168
biomarker proxies, 2:775–782
coastal sites, 3:211–213
corals/sclerosponges/mollusk data, 2:795–799
data, definition, 1:362ge
deep-sea carbonate dissolution, 2:859–870
diatoms
Pacific Ocean, 1:571
silicon isotopes, 2:734–743
dinoflagellates, 2:800–815
environmental archeology interpretation, 3:705
forensics
palynology, 4:556–566
use in forensic science, 1:522–525, 4:521–534, 4:535–541, 4:542–547, 4:548–555
- indicators
analysis, paleolimnology, 3:264–265
chironomids, postglacial Europe, 1:386, 1:395, 1:396
IP₂₅ HBI, Arctic sea ice, 1:591–595
lake-level changes, 3:672
Late Pleistocene chironomids, 1:373–385
magnetic susceptibility, 2:884–898
marine carbon cycle, 2:849–858
Mg/Ca and Sr/Ca paleothermometry, 2:871–883
moisture, oribatid mites, 2:693–694, 2:695
paleoceanography, 2:745–746, 2:746t, 2:753, 2:884–898
paleoclimate
European loess records, 2:612–615
introduction to, 3:87–92
modern analog approaches, 3:110
reconstruction approaches, 3:179–180, 3:182
paleodroughts, 3:194–196
paleosols, 3:373–374
paleotempestology, 3:209–217
plant macrofossils, 3:610
pollen, 3:805–815
tree rings/stalagmites, 3:214–215
volcanic/solar forcing, 1:748–749, 1:750–753, 1:751f
see also biological proxies; chemical proxies; nutrient proxies; physical proxies; radioisotope proxies; salinity proxies; temperature, proxies
- Psekups mammal assemblage, Eurasia, 4:608–610
- pseudomorph ice wedges/casts, 3:436, 3:443–446, 3:444f, 3:446f, 3:447–448, 3:447t, 3:448f, 3:450f
- PSV *see* paleosecular variation
- PTF *see* partial track fading
- Ptolemaic Temple, Dimai, Egypt, 1:517–518, 1:517f
- Puerto Rico, paleotempestology, 3:215–217, 3:216f
- Pulbeena Swamp, Tasmania, 1:196
- pulse discrimination electronics, 4:308
- pulsed stimulation, OSL dating, 2:660–661
- pumiceous grains, FT dating, 1:645–646, 1:645f
- punctuated equilibrium model, evolution, 4:723
- Punung caves, Sunda Shelf, 4:661–662, 4:661t
- Punung, Java, faunal assemblage, 4:695
- puparia
blow flies, 4:550
forensic use, 4:550, 4:555
- push moraines, 1:769, 1:771f, 1:772f
glacier recession, 1:805–806, 1:807–808, 1:807f, 1:808f
permafrost–glacier interactions, 3:508–509
Saalian glaciation, 2:174, 2:175f
see also moraines
- PW *see* poleward heat transport
- Pyganodon lacustris* bivalves, 1:344–345, 1:346f

- Pyrenees, Southern Europe, 4:183, 4:184–185
- pyroclasts, tephrochronology, 4:278
- pyrolysis, charcoal, 4:353, 4:354–355, 4:354f, 4:357
- pyroxene, TCNs, 1:410–411, 1:412–413
- ## Q
- Qaidam Desert, China, 1:607f, 1:608, 1:608f, 1:610f, 1:611
- QEI *see* Queen Elizabeth Islands
- qualitative approaches, diatoms, 1:490–491
- qualitative environmental reconstructions, 3:785–787
- qualitative indicators, climate change, 1:373–374
- qualitative inferences
- indicator-species approach, 1:368–369
 - sediment
 - constituents, 3:353–354
 - properties, 3:351f, 3:352–353
- qualitative paleoclimate reconstructions, 3:180–181
- continued utility, 3:180
 - current techniques, 3:180–181
 - history of, 3:180
 - limitations, 3:181
- quality control, pollen databases, 3:834
- Quantalus liquid scintillation counting, 4:308f, 4:309t
- quantification, plant remains, 3:703–704
- quantitative estimates
- climatic variables, 2:552–553
 - paleoclimatic, insect fossils, 1:200
 - soil morphology, 3:409–410
- quantitative fidelity, freshwater mollusks, 3:289
- quantitative indicators, climate change, 1:374–383
- quantitative inferences
- models, chironomid paleoecology, 1:369–370
 - sediment
 - constituents, 3:353–354
 - properties, 3:351f, 3:352–353
- quantitative reconstructions
- climate, plant macrofossils, 3:785, 3:787–789
 - paleoclimate, 3:181–183
 - calibrating proxies, 3:182
 - equation-based, 3:182–183
 - general concepts, 3:181–182
 - geochemical techniques, 3:181–182
 - problems, 3:182–183
 - process modeling, 3:183 - vegetation, POLLSCAPE model, 3:871, 3:874–878
- quarries/quarrying
- micro/macroscale erosional forms, 1:865, 1:867, 1:868f, 1:869, 1:870f, 1:871–872
 - soil/sediment examination, 4:537
- quartz
- ESR dating, 2:676
 - OSL dating, 2:653, 2:662–664, 2:663f
- TCNs, 1:410, 1:412–413, 1:413–414
- TL dating, 2:649, 2:650–651
- quasi-active guard detector, LSC, 4:308–309
- Quaternary climatic change, 1:26–35
- astronomical rhythm, 1:26–27
 - global feedback mechanisms, 1:31–32
 - humans, 1:33–34
 - numerical environmental modeling, 1:32–33
 - role of oceans, 1:29–31
 - short-term variation, 1:27–29
 - sub-Milankovitch variations, 1:27–29
 - see also* climate change
- Quaternary entomology, 1:191–199
- see also* entomology
- Quaternary environments, 3:147–154
- boundary conditions, 3:149–150
 - carbon cycle glacial-interglacial changes, 3:152–153
 - equilibration, 3:150, 3:152f
 - heat flux adjustments, 3:148–149, 3:150f
 - MOC, 3:150–152
 - modeling
 - applicability, 3:147–148
 - complexity, 3:147–148
 - design, 3:148–153
 - examples, 3:150–153
 - paleoclimate, 3:147–154
 - uncertainties, 3:148 - see also* environmental...
- Quaternary Environments of the Eurasian North (QUEEN) programme, 2:225
- Quaternary period
- definition, 2:216ge
 - history of Quaternary science, 1:1–8
- Quaternary proxies, 1:522–525, 4:521–534, 4:535–541, 4:542–547, 4:548–555
- see also* proxies
- Quaternary stasis, New Zealand, 1:244–247
- Quebec, Canada
- megafossils, 3:624–625
 - pollen records, 4:47, 4:49
- QUEEN *see* Quaternary Environments of the Eurasian North programme
- Queen Elizabeth Islands (QEI), Canada, 4:469–477, 4:470f, 4:471f, 4:473f, 4:474f
- Queen Maud Land, East Antarctica, 2:220
- Quelccaya site, Andes, ice cores, 2:387, 2:388f, 2:390, 2:391, 2:392–393, 2:392f, 2:393f
- Quercus* (oak)
- plant distributions, 3:855, 3:856f, 3:857f
 - postglacial pollen records, 4:179, 4:182, 4:185
- questioned samples
- diatom analysis, 1:523–524
 - soil/sediment, 4:538
- Quinn, WF, 3:174–175
- Quinton, England beetle assemblages, 1:187
- ## R
- racemization, amino-acid dating, 1:37, 1:38–39, 1:38f, 1:39f, 1:40, 1:40f, 1:41–42, 1:41f, 1:43, 1:44–45, 1:45–46, 1:45f, 1:47–48, 1:47f
- radar observations, ice cores, 2:445
- radiation
- damage to ancient DNA, 4:720
 - impact of dust, climate cycles, 3:415–416, 3:417, 3:418
 - see also* cosmic radiation; nuclear radiation
- radiative dosimetry dating methods, 1:449–450, 2:667–668, 2:669–672, 2:671f
- radiative forcing, global warming, 3:244, 3:250
- radioactive decay
- dating methods, AMS, 4:316
 - nuclides, 1:447
- radioactive isotope dating, ice cores, 2:305–306, 2:305t
- radioactive mean life, cosmogenic nuclides, 1:440ge
- radioactive nuclide dating methods, 1:447–449
- cosmogenic radionuclides, 1:449
 - in-growth, 1:447–448
 - K–Ar method, 1:449
 - postnuclear tracers, 1:449
 - radioactive decay, 1:447
 - radiocarbon dating, 1:448–449
 - uranium-series methods, 1:449
- radiocarbon age determinations ($^{14}\text{C}_{\text{AMS}}$), 2:795ge
- radiocarbon calibration, varved sediment, 4:587
- radiocarbon dating
- alkenone paleothermometry, 2:762
 - AMS, 4:316–323
 - carbon-14
 - atmospheric variations, 4:329–335
 - plant macrofossils, 4:361–367
 - record calibration, 4:345–352 - charcoal, 4:353–360
 - conventional method, 4:305–315
 - coral records of RSL, 4:413–414, 4:414–416, 4:414f
 - error sources, 4:324–328
 - exposure geochronology, 1:432
 - history of, 1:6–7
 - methodology, 1:9–10, 1:10f
 - microfossil-based reconstruction, Holocene RSL, 4:426
 - neoglaciation, 2:257, 2:260f, 2:270
 - phytoliths, 3:585–586
 - plant macrofossils, 3:609–610
 - pollen records, North America, 4:116, 4:117
 - radioactive nuclides, 1:448–449
- sea level studies, 4:370
- cave data, 4:462
 - coral records of RSL, 4:413–414, 4:414–416, 4:414f
 - microfossil-based reconstruction of RSL, 4:426
- sediment analyses, lake levels, 2:549–551
- speleothems, 1:296
- stratigraphy overview, 4:192
- temporal variation causes, 4:336–344

- radiocarbon dating (*Continued*)
 tephrochronology, 4:288
 varved lake sediments, 4:573
 radiocarbon ($\Delta^{14}\text{C}$), 2:795*ge*, 2:796
 radiocarbon time series, carbon-14
 variations, 4:336
 radio-echo sounding (RES), ice cores, 2:445,
 2:445*f*
 radiography, lake sediment properties, 3:303
 radioisotope dating, K/Ar and argon-40/
 argon-39, 2:477
 radioisotope proxies, 2:923–931
 definitions, 2:923
 general applications, 2:923–925
 protactinium-231, 2:927–929, 2:928*f*,
 2:929–930, 2:930*f*
 thorium-230, 2:925–927, 2:927–929,
 2:928*f*, 2:929–930, 2:930*f*
 tracers in paleoceanography, 2:925
 radiolarians, 2:830–840
 biogeographic patterns, 2:831*f*
 biostratigraphy, 2:831–833
 marine biological proxies, 2:830–840
 modern ecology, 2:830
 outlook for use of, 2:838
 paleoproductivity estimation, 2:834–838
 paleotemperature estimation, 2:834,
 2:835*f*
 preparation methods, 2:830
 preservation, 2:830
 proxy applications, 2:830–838
 shells, 2:830, 2:831*f*
 water mass boundaries, 2:830–831
 radiometric dating
 cave data sea-level reconstructions,
 4:460–466
 ESR dating, 2:667–677
 FT dating, 1:643–662
 K/Ar and argon-40/argon-39 dating,
 2:477–482
 landscape evolution, 1:440–445
 methodology, 1:410–417
 OSL dating, 2:653–666
 overview, 1:407–409
 see also cosmogenic nuclides; luminescence
 dating
 radiometric methods
 cosmic ray interactions, 1:418–431
 exposure geochronology, 1:432–439
 radionuclides *see* cosmogenic radionuclides
 rafts, glaciectonic, 1:844–845
 ragweed *see* *Ambrosia*
 rainfall
 Africa
 patterns, 4:664
 pollen records, 4:88–89
 patterns, 4:664
 see also precipitation
 rainforests
 African pollen records, 4:10–12, 4:12*f*
 Amazon pollen records, 4:59–61,
 4:156–157, 4:157–159, 4:157*f*,
 4:158*f*, 4:159*f*
 Atlantic pollen records, 4:160
 Australasian Late Pleistocene, 4:18, 4:22,
 4:23–24, 4:24–25
 raised beaches, RSL, 4:468–469, 4:475*f*,
 4:477
 Raman spectroscopy, 4:561–562
 RAM method, temperature proxies, 2:843*f*,
 2:844*t*, 2:845, 2:846–847
 ramparted depressions, 3:478, 3:478*f*
 ramps
 moraine forms/genesis, 1:770–772,
 1:772*f*
 rock glacier types, 3:536*f*
 Rancholabrean (RLB), 4:673–679
 localities, 4:675–676
 vertebrate records, 4:646–647
 range extensions, plant macrofossils, 3:757
 Rangifer tarandus see caribou
 Rangitawa Tephra, New Zealand
 stratigraphy overview, 4:196–197, 4:197*f*
 tephrochronology, 4:290*f*
 rapid environmental change
 biotic response, 3:790
 case studies, 1:679–682
 fluvial environment responses, 1:676–683
 see also abrupt climate change
 rapid mass movement, slope deposits/forms,
 3:484–485
 rapid sea-level rise, 4:499
 rarefaction
 analysis, 1:494, 1:494*f*
 biodiversity, 3:816–817, 3:817–818,
 3:819*f*
 rat middens *see* rodent middens
 ravinement surface (RS), stratigraphy,
 4:260–261, 4:262
 raw materials, hominid–animal interactions,
 4:751–752
 Raymond bumps, ice cores, 2:305, 2:305*f*
 reaction wood, tree growth, 2:95–96
 reactive oxygen species (ROS), 1:320
 recent warming, Andes, 2:392
 see also global warming
 recessional phase, vegetation development,
 3:763
 recolonization of Europe, humans, 1:55*f*
 reconstruction methods
 CO₂
 from fossil leaves, 3:613–620
 overview, 3:616–619
 lakes
 Antarctica, 1:536–538
 diatom oxygen systematics, 1:484–486
 west-central Europe, 2:549–551
 paleo-ELAs, 1:916–917
 relative sea level, 4:371–372, 4:419–428
 sea-surface conditions, 1:530–535
 see also climate, reconstructions; lake-levels,
 reconstructions; Landscape
 Reconstruction Algorithm; marine
 reconstructions; paleoclimate/
 paleoclimatology, reconstructions;
 paleoenvironments, reconstructions;
 qualitative...; quantitative
 reconstructions; salinity/climate
 reconstruction; vegetation,
 reconstruction
 record-keeping role, active layer/climate
 change, 3:427
 record monitoring, sedimentary pigments,
 3:331
 recovery of plant remains, 3:703–704
 Red Clay, 1:637*ge*, 2:601–603, 2:604
 red deer, 4:642, 4:642*f*, 4:643
 Rederstad Stadal (MIS 5b) beetle
 assemblages, 1:202
 redistributed organic materials,
 cryoturbation, 3:434
 Red Sea method, sea-level reconstruction,
 2:921*f*
 red thermoluminescence dating, 2:651
 reduction, weathering profiles, 3:394–395
 reef flat sea level indicators, 4:383–384,
 4:383*f*
 reefs, coral
 RSL change record, 4:409–418, 4:495,
 4:496*f*
 U-series dating, 4:567–568, 4:568–570
 reference chronologies, geomorphic
 processes, 2:92–93
 reforestation, 4:187–188
 see also forests
 refractive index, soil/sediment, 4:539
 refugia
 beetles, New Zealand forests, 1:251–253
 hypothesis, Amazonian rainforest, 4:59–60
 mapping, pollen databases, 3:836–837
 theories, pollen data, 2:703
 vegetation reconstruction, 3:866–867
 regeneration methods, OSL dating,
 2:656–657
 regional climate
 changes, last millennium, 3:229–231
 volcanic/solar forcing, 1:752
 Regional Continuity model, human
 colonization, 1:111–112, 1:112–113
 Regional Estimates of Vegetation Abundance
 from Large Sites (REVEALS), 3:892–895
 applications, 3:895–901, 3:896*f*, 3:898*f*,
 3:899*f*
 pollen analysis, 3:798–799
 POLLSCAPE model, 3:874–875
 regional extinctions, beetle fauna, 1:241–242
 regional patterns
 Holocene lake-level fluctuations, 2:551
 low-latitude sand seas, 1:624–634
 rodent middens, 3:676–681
 regional scale studies
 paleoclimatology approaches, 3:102–104,
 3:104–107
 plant macrofossils/pollen, 3:731
 regolith
 hypothesis, paleoceanographic records, 3:6
 trimlines/paleonunataks, 1:918–919,
 1:921*f*, 1:922*f*
 weathering profiles, 3:392
 regression, sedimentary indicators,
 4:390–391
 regressive sequences, microfossils, 4:423–424
 regressive surface of erosion (RSE), 4:261,
 4:263
 rejuvenation
 ¹⁴C of plant macrofossils, 4:361*ge*
 ice wedge periglacial features, 3:438*f*, 3:439
 relational databases, pollen, 3:831, 3:832

- relative-age dating
 lichenometry, 2:566
 trimlines/paleonunataks, 1:921–924
- relative geomagnetic paleointensity, 1:714–716, 1:717f, 1:718
- relative glacier extent, hemispheric contrasts, 2:275
- relative humidity (RH), teeth/bones, 1:292
- relative sea-level (RSL) changes, 4:369
 coral records, 4:409–418
 highstands, 4:409–411, 4:410f
 Holocene variations, 4:413–416
 interannual variations, 4:416–417
 microatolls, 4:415f, 4:416–417, 4:417f
 postglacial rise, 4:412–413
 present reef distribution, 4:410f
 eustatic sea-level distinction, 4:465
- general principles, 4:370
- high-energy coasts, 4:385–395
- high latitudes, 4:467–480
 Antarctica case study, 4:477–479
 Clark's model, 4:467, 4:468f
 controls and patterns, 4:467–469
 QEI case study, 4:469–477, 4:470f, 4:471f, 4:473f, 4:474f
- isostasy, 4:452–459
- Late Quaternary
 Greenland, 4:481, 4:483–484, 4:484–487
 midlatitudes, 4:489–494
 tropics, 4:495–502
- model predictions, 4:375f
- radiocarbon dating, 4:462
- reconstructions, 4:371–372
 microfossil-based, Holocene, 4:419–428
 chronologies, 4:426–427
 indicative meaning, 4:425
 isolation basins, 4:424
 methods, 4:422–425
 regressive sequences, 4:423–424
 seismic vertical motion, 4:424–425
 SLIPs, 4:422–423
 transfer functions, 4:425
 transgressive sequences, 4:423–424
 sedimentary indicators, 4:398–405, 4:407f
- sedimentary indicators, 4:396–408
 age-altitude analysis, 4:401–404, 4:402f
 data sources, 4:396
 high-energy coasts, 4:385–395
 ideal indicators, 4:405–407
 intertidal environments, 4:396–397, 4:397–398
 low-energy, 4:396–408
 mangroves, 4:397, 4:399f
 marsh elevation diagrams, 4:405, 4:406f
 saltmarshes, 4:396–397, 4:397f, 4:398f, 4:399f
 sea-level tendencies, 4:404–405
 SLIPs, 4:399–404, 4:402f
- tectonics and, 4:461, 4:503–519
see also eustatic sea-level change
- relative water depth, lake sediment, 3:306f
- relevant source area of pollen (RSAP), 3:798, 3:799, 3:872–873, 3:876t
- reliability
 data, CO₂ ice core studies, 2:317
 methods, isotope thermometer, 2:348–349
- relict-form frost mounds, 3:472–480
- relict landsurfaces, 3:530–532
- relict paleosols, 3:368, 3:371
- relict periglacial slope deposits, 3:485–487
- relict rock glaciers, 3:536f, 3:539
- religion, hominid–animal interactions, 4:752–754
- remanent magnetization, 3:376, 3:380
 definition, 3:375*ge*
 measuring, 3:377
 secular variation, 1:709–710, 1:710–712, 1:711f
- remote sensing
 data, glacier recession, 1:803, 1:804, 1:804f
 forensic geology searches, 4:530
 lake sediment physical properties, 3:300–301
- Renland ice caps, Greenland, 2:404f, 2:406–407, 2:406f, 2:408
- repeatability conditions, radiocarbon dating, 4:326
- replication principle
 coral Sr/Ca variability, 2:880–881
 dendrochronology, 1:455
see also reproducibility conditions
- representation, plant macrofossils, 3:684–689
- reproducibility conditions, radiocarbon dating, 4:326
see also replication principle
- reptiles
 Australian Mid-Pleistocene, 4:621–630, 4:622t
 dwarfing/gigantism, 4:745–746
 RLB land mammal age, 4:674, 4:675
- Rerewhakaaitu Tephra, New Zealand, 4:294, 4:294f
- RES *see* radio-echo sounding
- research developments/directions
 frost mounds, 3:478–479
 insect fossil studies, 1:177–181
 paleosols, 3:374
 soil morphology studies, 3:410–411
- research history
 ancient DNA, 4:719
 Antarctic ice core, 2:435–438
 biosphere history, 1:23–24
 climate history, 1:17
 cryosphere history, 1:20
 deltas, 1:693
 geologic hazards, 1:21–23
 Greenland ice core, 2:435–438
Homo Sapiens, 1:24
 lithosphere history, 1:20–21
 ocean history, 1:18–20
 palynology, North America, 4:115–117
 plant remains forensics, 4:544, 4:546
 societal relevance, 1:17–25
- residuum, weathering profiles, 3:392
- resin ducts, tree growth, 2:97–99, 2:99f
- response surfaces, pollen–climate relation, 3:809
- response time, fluvial environments, 1:678–679
- retouched tools, definition, 1:83*ge*
- retrogressive thaw slumps, 3:469, 3:469f, 3:578–579
- REVEALS *see* Regional Estimates of Vegetation Abundance from Large Sites
- revolution concept, human origins, 1:57
- RH *see* relative humidity
- rheids, South America, 4:683, 4:684f
- rheology, ice cores, 2:303–304
- rhinoceros *see* woolly rhinoceros
- Rhizocarpon geographicum* lichenometry, 2:565, 2:568–569
- Rhynchospora alba* macrofossils, 3:637, 3:641f, 3:642
- rhyolite
 FT dating, 1:652–653, 1:653–654, 1:654–655, 1:658f
 tephrochronology, 4:286f, 4:287
- rhythmites, glaciolacustrine, 2:44–45, 2:44f, 2:48f
- ribbed moraines, 3:509
- rice, phytoliths, 3:589–590
- ridges
 beach ridges, 3:213
 composite, 1:840–843
 disintegration ridges, 1:809
 moraine ridges, 2:257–259, 2:258f, 2:567–568, 2:570
 subglacial volcanic, 1:793–796, 1:798f
 tephra ridges, 1:793–796, 1:798f
- Rift Valley, Kenya, 1:363, 1:364f
- ring forms, glacier recession, 1:809, 1:810f
- rinnentaler *see* tunnel valleys
- riparian ruderals, 3:742
- rituals, archeobotanical research, 3:717
- river deltas, 1:693–699
 controls, 1:693–695
 high-energy coasts, 4:392–393, 4:393f
 processes, 1:693–695
 Quaternary site cases, 1:699–702, 1:701f
see also deltas
- river settlements, human adaptations, 1:158
- river terraces, 1:684–692
 eroding continental interiors, 1:670, 1:671f
 glacial–interglacial cycles, 2:115–119
 incision/aggradation, 1:677, 1:679f
- rivers
 diatoms, 1:475
 glacial–interglacial scale responses, 2:112–125
 incision measurement, 1:443–444
 management, paleohydrology, 3:253–258
 periglacial sediments, 3:490–491, 3:491–492, 3:492–493, 3:494–495, 3:495–497
 rapid environmental change responses, 1:676–683
 runoff, tropical Atlantic, 3:22–24
 sediments, 1:663–675
 periglacial, 3:490–491, 3:491–492, 3:492–493, 3:494–495, 3:495–497
see also individual rivers; river deltas
- Riwat site, Asian hominin archives, 1:76f
- RLB *see* Rancholabrean land mammal age
- RLQ analysis, pollen, 3:828

- RNA, ancient plants, 2:713
- Roberts Creek, Iowa, US, 3:772–775, 3:776f
- Robe, South Australia, 2:649, 2:650f
- roches moutonnées, 1:870f, 1:871–872, 1:874–875
- rock art, 1:158–159
- rock basins, 1:854–857, 1:857f, 1:858f, 1:862f
- rock coast sea-level indicators, 4:388
- rock dating
- cosmogenic nuclides, 1:432–434
 - TCNs, 1:414
- rock drumlins, 1:872–873, 1:874f
- rockfall/talus, 3:566–570
- hazards, 3:568
 - materials, 3:569–570
 - morphology, 3:568–569
 - movement, 3:567–568
 - surface processes, 3:569
 - triggers, 3:566–567
- rock forms, 3:566–573
- see also* rock slopes
- rock glaciers, 3:535–541
- activity/age, 3:539
 - advance of, 3:537–538
 - composition, 3:535–537
 - definition/distribution, 3:535
 - kinematics, 3:537–538
 - shape/surface morphology, 3:538–539
 - talus slopes, 3:571
 - thermal conditions, 3:535
- rock paleomagnetism, 1:709–712, 1:711f
- rock slopes, 3:556–559
- catastrophic failure, 3:556
 - deep-seated gravitational deformation, 3:556–558
 - rockfall/talus accumulation, 3:558–559
 - weathering, 3:481
 - see also* paraglacial land systems; rock forms
- rock streams *see* block streams
- rock weathering, 3:500–506
- biological processes, 3:502
 - chemical processes, 3:502
 - mechanical processes, 3:500–502
 - products, 3:503–505
 - profiles, 3:392, 3:504f, 3:505
 - rates of, 3:502–503, 3:503t
 - rinds, 3:504
 - slope deposits/forms, 3:481
- Rocky Mountains, Canada/US
- insect fossil studies, 1:286–289
 - lake level studies, 2:540–543, 2:543f
 - neoglacial chronology, 2:273
- rodent middens, 3:674–683
- analysis methods, 3:675
 - beetle fossils, 1:162, 1:164f
 - implications of studies, 3:681–682
 - pollen records, 4:143, 4:153f, 4:154
- rodents
- Australian Mid-Pleistocene, 4:637
 - dwarfing/gigantism, 4:736, 4:743–744
 - evolutionary trends, 4:725–727, 4:731
 - South American Late Pleistocene, 4:687
- rogens, meltwater hypothesis, 1:835–836
- rollover scenario, sedimentary indicators, 4:389–390
- Roman remains
- archeobotanical research, 3:715–717
 - mineralization, 3:708f
 - preservation, 3:702f
 - sediments, 3:704f
 - taphonomy, 3:699
- root exposure, trees
- debris-flow events, 2:100, 2:101f
 - growth rings, 2:100
- root growth, tree reactions, 2:99–100, 2:100f
- ROS *see* reactive oxygen species
- Ross Sea
- deglaciation, 2:221
 - RSL, 4:475f, 4:477–479
- roundness properties, sediment particles, 2:1, 2:2, 2:2f, 2:3t, 2:4f
- rove beetles, 1:169
- forensic science use, 4:554
 - Japanese sites, 1:212
 - North American studies, 1:225
- Royal Navy ship's logbooks, 3:238, 3:239f, 3:239t
- RS *see* ravinement surface
- RSAP *see* relevant source area of pollen
- RSE *see* regressive surface of erosion
- RSL *see* relative sea-level changes
- Rub al Khali, Arabia, 1:629, 1:630, 1:631f
- Ruddiman, WF, 2:129, 2:130–131, 2:131f, 3:248
- ruderals, Arctic Eurasia, 3:742
- running means, time series, climate change, 3:98
- runoff
- lake-level modeling, 2:558–559
 - rivers, Late Pleistocene, 3:22–24
- runout zones, rockfalls, 3:568
- Russia
- dune fields, 1:599
 - age determination, 1:603–604
 - origins, 1:602
 - types, 1:601
 - glaciations, 2:224–228
 - Donian, 2:172
 - Holsteinian stage, 2:174
 - Middle Pleistocene, 2:172, 2:174, 2:175, 2:177
 - Saalian, 2:175, 2:177
 - Lake Baikal diatom records, 1:546–547, 1:547–550
 - megafossils, 3:625–626, 3:626–629, 3:628f
- R value, pollen analysis, 3:795, 3:796
- ## S
- SA *see* South America
- SAAD *see* single-aliquot additive-dose
- Saalian stage, Eurasia, 2:173–174, 2:174–177, 4:612–613
- Sacred Lake, Kenya, 4:97, 4:99f
- Sac River, Missouri, US, 3:782f, 3:783
- sagebrush voles, 4:725
- Saginaw Lobe, US, 1:829f, 1:831f, 1:832f
- Sahara Desert
- atmospheric dust, 1:729, 1:731–732
 - 'green' Sahara, 1:723–724, 1:723f, 1:724f
 - lake level history, 2:499–501
 - modern human expansion, 1:94, 1:95–96
 - postglacial pollen records, 4:89–91, 4:92t, 4:100
 - sand seas, 1:624–626, 1:627f
- Sahel region
- lake dynamics, 2:501
 - sand seas, 1:625, 1:627f
- Sahul landmass, 1:108–118
- St. Charles, Iowa, North America, 1:221–222
- St. Saufflieu soil loess records, 2:610–611, 2:612f
- Sajama site, Andes, 2:387, 2:390–391
- saline lakes
- Asia, 2:506–508, 2:509f
 - carbonate precipitation, 1:334–335
 - chemistry, 3:295
 - continental lakes, 1:507–515
 - levels, reconstruction of, 2:493–495, 2:494f
 - diatoms, 2:496–497
 - ostracods, 2:496
 - oxygen-isotope analysis, 2:496–497
 - sediment classification, 2:495–496
 - site selection, 2:487
 - South America, 1:509–510, 1:510f
 - see also* salinity/climate reconstruction
- salinity
- Asian lake studies, 2:506–508, 2:509f
 - changes
 - chironomid records, 1:360
 - diatoms as indicators, 1:538
 - foraminifera Mg/Ca, 2:874, 2:876–877 - contrast, Indian Ocean, 3:46–47
 - definition, 2:932
 - levels
 - North Atlantic Ocean, 3:10f
 - sea-surface, 2:800, 2:806–808, 2:812, 2:813f - THC, 1:742–743
 - Western Atlantic, 2:933f
 - see also* salinity proxies
- salinity/climate reconstruction, 1:507–515
- continental lakes, 1:507–515
 - diatoms
 - classification of, 1:507
 - environmental history from, 1:509–512
 - preservation of, 1:509
 - as salinity indicators, 1:507–509 - Lake Titicaca, South America, 1:509–510, 1:510f
 - Late-Glacial/Holocene climate change, 1:510–512
 - orbital/millennial time scales, 1:509–510
 - quantitative approaches, 1:507–509
 - see also* saline lakes
- salinity proxies, 2:932–940
- $\delta^{18}\text{O}$, 2:932–940
 - biomarker proxies, 2:778–779
 - current reconstruction methods, 2:932–938
 - estimating errors, 2:937–938
 - Mg/Ca, 2:936, 2:937–938, 2:938–939
 - paleothermometry, 2:936

- temperature calibration equations, 2:936–937, 2:937f
- Pacific water variations, 2:938–939, 2:938f, 2:939f
- past variations, 2:932
- secondary effects, 2:937
- Western Atlantic salinity, 2:933f
- Salix herbacea* macrofossils, 3:597f, 3:610, 3:610f
- salmon
- life cycles, paleolimnology, 3:266
 - nutrient/contaminant biovectors, 3:346
- Salmon Lake, Alaska, 2:195, 2:197f
- salt dissolution, soil, 3:402
- salt extraction sites, diatom analysis, 1:518
- saltmarshes
- deposits, Greenland, 4:486–487, 4:486f
 - sedimentary indicators, RSL change, 4:396–397, 4:397f, 4:398f, 4:399f
- sampling
- dendroarcheology, 3:631, 3:632–633, 3:633–634, 3:634–636
 - diatoms
 - analysis, 1:523–524
 - microstratigraphy, 1:540–541 - lacustrine sediments, 1:333
 - location factors, macrofossils, 3:688
 - micro-sampling techniques, 1:342–343
 - plant macrofossils, 3:684–689, 3:702–703, 3:702f, 3:710f, 3:711f
 - pollen databases, 3:833–834
 - preparation of samples
 - chironomid fossils, 1:357–358
 - dendroarcheology, 3:631
 - dendrochronology, 1:455
 - stomatal frequency analysis, 3:614
 - VIS/IR spectroscopy, 3:351–352 - soil/sediment in forensics, 4:538
 - stomatal frequency analysis, 3:614
 - surface samples, plant macrofossils, 3:684–689
 - till fabric analysis, 2:78
 - uncertainty, radiocarbon dating, 4:325
 - varved sediments, 4:584–585
 - vertebrate remains, 4:593–594
- Sanborn Cove marsh, Maine, USA, 2:771f
- Sanchez, WA, 3:104
- sand
- forensic science use, 4:536–537
 - layers, overwash, 3:209–211, 3:212f, 3:213f, 3:215f
 - see also* eolian sand; sand dunes; sand seas; sand veins; sandy deserts
- sandar, 2:6–9, 2:8f, 2:14f
- confined, 2:8–9
 - unconfined, 2:7–8
 - see also* proglacial landforms
- sand dunes
- definition, 1:597ge
 - high latitudes, 1:597–605
 - low latitudes, 1:623–636
 - see also* dune...
- sand seas (ergs), 3:357–359
- distribution of, 1:624f
 - dune distribution in, 1:627t
 - low latitudes, 1:623–636
- sand sheets
- Alaska, 1:599
 - definition, 1:597ge
- Sandström, Johan, 1:737, 1:740
- sand veins, ice wedges/casts, 3:440, 3:441f, 3:442t
- sand wedges, ice wedges/casts, 3:440, 3:442t, 3:448–449
- sandy deserts, Central Asia, 2:588f
- Sangamon Geosol, 3:368, 3:368f, 3:369, 3:372, 3:373f
- Sangihe island, Southern Asia, 4:662
- Sangiran hominin remains, 1:67
- Sangoan tools, Africa, 1:65
- sanidine, TCNs, 1:412–413
- sap beetles, 4:554
- Sappan interval, North America, 4:648
- saprolite
- landscape evolution, 1:440ge
 - weathering
 - chemical weathering, 3:393f
 - profiles, 3:392, 3:393f
- sapropels, 1:67, 1:68f
- saproxylophagous beetles, 1:276–277
- SAR *see* single-aliquot regenerative-dose
- SARA *see* single-aliquot regeneration and added dose
- Sarcophagidae (flesh flies), 4:550
- Sardinian bovids, 4:739–741
- Sarenes Glacier, Europe, 2:265, 2:265f
- Sartan glaciation, Beringia, 2:197–199, 2:199f
- Sartanskaya Stade, Northern Asia, 4:27–34
- SAT *see* surface-air temperature
- Satir, Java, 4:653
- saturation isothermal remanent magnetization (SIRM), 3:380
- savanna
- Africa
 - hominids, 1:310
 - pollen records, 4:13–14 - Amazon Rainforest pollen records, 4:156, 4:157–159
 - Colombian pollen records, 4:58–59
 - woodlands/grasslands, 4:13–14
- Saxifragaceae macrofossils, 3:606f, 3:608
- SBs *see* sequence boundaries
- scale in lithostratigraphy, 4:229f, 4:235–238
- Scandes Mountains, Sweden, 3:625–626, 3:627t
- Scandinavia
- ice sheet
 - Early Quaternary, 2:168–169
 - glaciations, 2:228–230, 2:228f
 - neoglaciation, 2:261f, 2:262–263, 2:262f, 2:263f, 2:265f
 - stand-scale palynology, 3:846, 3:847f, 3:848–853
 - see also* individual countries - scanning electron microscopy (SEM)
 - forensics
 - palynology, 4:556–559
 - soil/sediment in, 4:539–540
 - oribatid mites, 2:685f, 2:690
- Scaphidiidae beetles, 1:212, 1:215f
- Scarabaeidae (dung beetles)
- forensic science use, 4:554
 - Japanese sites, 1:213–214, 1:218–219
- scarps, loess, 2:585, 2:586f
- scars
- glacially modified trees, 2:108
 - tree growth reactions, 2:97
- scavenger flies, 4:551–552
- scavenger vertebrates, definition, 4:680ge
- scavenging
- definition, 4:748ge
 - hominids, 4:748–751
- scenario-based searches, forensic geology, 4:525t
- SCEs *see* steryl chlorin esters
- Scheuchzeria palustris* macrofossils, 3:642, 3:643f
- Schmidt-hammer exposure-age dating (SHD), 2:257–259, 2:259f
- Schmidt hammer, trimlines/paleonunataks, 1:921–923, 1:924f
- Scirpus cespitosus* macrofossils, 3:641f, 3:642
- see also* *Trichophorum cespitosum*
- scleractinian corals, 2:795
- sclerochronology, 2:798, 2:799, 3:288–289
- sclerophyll vegetation, 4:18, 4:22–23, 4:23–24, 4:24–25
- sclerosponges, 2:795–799
- SCLF *see* single-crystal laser fusion
- Scotia sea cores, 2:889, 2:892f
- Scotland, glaciations, 2:230–231
- scrapers, definition, 1:83ge
- scratch-diggers, 4:680ge
- scree *see* talus...
- SCRs *see* solar cosmic rays
- scuttle flies, 4:551
- SD *see* stomatal density
- SE... *see* South East...
- sea..., *see also* marine... sea arch indicators, 4:379, 4:380f
- seabed mapping, 4:392–393, 4:393–395
- seabirds, nutrient/contaminant biovectors, 3:347
- sea cave indicators, 4:379, 4:380f, 4:388
- sea-dust role, climate cycles, 3:416–417
- sea ice
- albedo-SST feedback, 3:156–158, 3:159, 3:160–162
 - biota, 1:588–589
 - climate sensitivity, 3:11–12
 - diatom biomarkers, Arctic, 1:588–596
 - diatom distribution
 - Antarctica, 1:530–531, 1:532t, 1:533–535, 1:533f, 1:534f
 - Pacific Ocean, 1:571–573, 1:573–574, 1:576f - dinoflagellates, 2:800, 2:806–808, 2:808f, 2:812, 2:813f
 - dynamics proxies, 2:823
 - freezing/melting, 2:918
 - glaciochemistry, 2:326–333
 - habitat types, 1:588–589
 - increase in Antarctic Peninsula, 2:220
 - Late Pleistocene South Atlantic, 3:24–26
 - reconstruction, IP₂₅ HBI, 1:592–594
 - switch hypothesis, 3:6

- Sea of Japan, 2:896, 2:896f
 sea-level index points (SLIPs), 4:371–372, 4:374f
 age attributes, 4:372
 elevation attributes, 4:372
 error bars, 4:375f
 location attributes, 4:371
 microfossil-based reconstruction, Holocene RSL, 4:422–423
 sedimentary indicators, RSL change, 4:399–404, 4:402f
 tendency attributes, 4:372
 sea levels
 cave data use, 4:460–466
 changes, 4:369–376
 cave data, 4:460–466
 coral records of RSL, 4:409–418
 as downstream control, 1:676
 Early Pleistocene records, 3:3
 glacial sequence stratigraphy, 2:85
 glacial–interglacial scale, 2:112–114, 2:114f, 2:117f, 2:121–122, 2:121f
 glacier/icesheet extent, 1:892
 glacier recession, 1:803–804, 1:805
 global to local, 4:371f, 4:372–375
 Greenland
 ice sheet relation, 2:446
 Late Quaternary, 4:481–488
 high-energy coasts, 4:385–395
 highstand
 Holocene, 4:447, 4:499–501, 4:501t
 last interglacial, 4:498
 since last glacial maximum, 4:439–451
 lowstand, LGM, 4:498–499, 4:500f
 midlatitudes, Late Quaternary, 4:489–494
 morphostratigraphy, 4:243–244
 ocean history, 1:18, 1:19f
 postglacial human adaptations, 1:154
 rapid rise, 4:499
 sedimentary indicators, 4:385–395
 tectonics and, 4:503–519
 tropics, Late Quaternary, 4:495–502
 West Antarctic Ice Sheet, 2:448, 2:449
 coral records, RSL changes, 4:409–418
 curves, ESL change, 4:429, 4:430f
 cyclicity orders, 4:263
 East Antarctic Ice Sheet, 2:456
 equation, 4:439–440, 4:455
 general principles, 4:370–371
 glacial–interglacial cycles, 4:429–438
 growth/decay of ice sheets, 1:878–879
 high-energy coasts, 4:385–395
 indicators
 cave data, 4:460
 geomorphological, 4:377–384
 high-energy coasts, 4:385–395
 sedimentary, 4:385–395, 4:396–408
 tectonic shoreline, 4:505f, 4:506f, 4:507f, 4:509f, 4:513f
 tropics, 4:495–497
 isostasy, 4:452–459
 Late Pleistocene events in Beringia, 2:192–193, 2:193f
 Late Quaternary
 Greenland, 4:481–488
 midlatitudes, 4:489–494
 tectonics, 4:503–519
 tropics, 4:495–502
 overview, 4:369–376
 reconstructions
 cave data in, 4:460–466
 microfossil-based, Holocene RSL, 4:419–428
 Red Sea method, 2:921f
 rise
 glacier recession, 1:803–804, 1:805
 rapid rise, 4:499
see also sea levels, changes
 sedimentary indicators, 4:385–395, 4:396–408
 tendencies, sedimentary indicators, 4:404–405
 variability, marine diatoms, 2:823
see also eustatic sea-level; relative sea-level changes
 Sea of Okhotsk, Late Pleistocene, 3:20–21
 search assets, forensic geology, 4:529–533
 search methods, forensics, 4:521–534
 historical overview, 4:523–524
 principle objectives, 4:524t
 search phases, forensic geology, 4:527–529
 search planning, forensics, 4:527–529
 search and rescue philosophy, 4:525t
 sea salt aerosols, ice cores, 2:328–329
 seasonal cycles, varved sediments, 4:582, 4:587
 seasonal frost mounds, 3:476–477
 seasonality, coccolithophores, 2:784
 seasonal Oxygen-18, Greenland, 2:403–405
 seasonal records/responses
 laminated sediment marine diatoms, 1:554–561
 last interglacial, 3:156–159
 sea-surface conditions, reconstructions, 1:530–535
 sea-surface salinity, 2:800, 2:806–808, 2:812, 2:813f
 sea-surface temperatures (SSTs)
 alkenone paleothermometry, 2:757–759, 2:758f, 2:760–761, 2:762
 Antarctica, 1:528–529, 1:530–531, 1:532t, 1:533–535
 Arabian Sea, 3:50f
 biomarker proxies, 2:775–776, 2:776f
 calcareous marine fossil proxies, 2:871–883
 diatom-inferred climate change, 1:575–580
 dinoflagellates, 2:800, 2:806–808, 2:812, 2:813f
 Holocene development, 1:568f
 Indian Ocean, 3:46, 3:47
 last interglacial, 3:156–158, 3:157f, 3:159, 3:160–162
 Late Pleistocene South Atlantic, 3:18–19, 3:20f, 3:21f, 3:22f, 3:25f
 New Zealand/Australia, 3:77–78, 3:80f
 Nordic Seas, 1:567–569
 North Atlantic, 1:567–569, 1:569f, 3:9–10, 3:11f, 3:12f
 D–O events, 3:13–14
 postglacial, 3:55–57, 3:58f, 3:60
 Pacific Ocean
 diatom records, 1:571, 1:575–580, 1:575f, 1:579f, 1:580f, 1:581f, 1:583f, 1:584f
 North, 3:37–42, 3:43f
 South, postglacial records, 3:75f, 3:76, 3:77–78, 3:78f, 3:80f, 3:81f, 3:82–84
 Pliocene environments, 3:185, 3:187, 3:188f, 3:190–191
 postglacial
 North Atlantic, 3:55–57, 3:58f, 3:60
 South Pacific records, 3:75f, 3:76, 3:78f, 3:81f
 Quaternary environments, 3:149–150
 sea ice
 albedo–SST feedback, 3:156–158, 3:159, 3:160–162
 Antarctica, 1:530–531, 1:532t, 1:533–535
 South Pacific postglacial records, 3:75f, 3:76, 3:77–78, 3:78f, 3:80f, 3:81f, 3:82–84
 temperature proxies, 2:842, 2:843f
 zonal/meridional gradients, 2:746–747, 2:747–748, 2:747f, 2:749t, 2:751f
 seawater
 B/Ca proxy for pH, 2:851–853
 boron isotope–pH proxy, 2:849–851
 diatom silica
 global distribution, 2:918
 Mg/Ca paleothermometry, 2:876–877
 oxygen isotope composition, 2:915–922
see also sea levels
 SEB *see* surface energy balance
 secant analysis, glacial sediment, 2:56–58, 2:57f
 secondary calcite, soil, 3:383
 secondary ion mass spectrometry (SIMS), 2:735
 secondary nontemperature effects, Mg/Ca corals, 2:878–881
 foraminifera, 2:874
 secondary particle development, cosmic radiation, 1:422–427
 secondary species
 carcass processing, 4:751
 cave art, 4:752
 secular variation, 1:705–720
 coral Sr/Ca, 2:880–881
 geomagnetic
 excursions, 1:705–720
 field intensity, 1:712–713
 relative paleointensity, 1:714–716, 1:717f, 1:718
 reversals, 1:705, 1:716–718, 1:718–719
 geomagnetism, 1:706–709
 magnetization of rocks, 1:709–712, 1:711f
 paleomagnetism, 1:709–713, 1:718–719
 PSV, 1:713–714, 1:715f, 1:718, 1:719
 stratigraphy, 1:718–719
 SED *see* surface exposure dating
 sedentarism/sedentism
 environmental archeology, 3:699
 post-LGM humans, 1:104–106

- sedimentary basin tephrochronology, 4:296–298
- sedimentary facies analysis, 4:264
- sedimentary structures, 3:430–435
see also sediments
- sedimentation
- biosphere feedbacks, 1:722–723
 - long-term carbon cycle processes, 1:722–723
 - pollen analysis, 3:795
 - rates, Cenozoic cooling, 2:131–132
see also sediments
- sediment budgets, Holocene, 3:253–255, 3:254f
- sediment cascades, 3:553, 3:554f
- sediment fabric, lake physical properties, 3:303, 3:304f
- sediment limit, lake-level changes, 3:668f
- sedimentology
- European loess records, 2:610–612, 2:615, 2:616f
 - fluvial environments
 - Holocene climate forcing, 1:682
 - incision, 1:676–677
 - ice-proximal settings, 2:30–33
- sediments
- analyses
 - charcoal records, 2:720–722
 - lake-level reconstruction, 2:549–551
 - areal ice extent, 1:884
 - charcoal records, 2:717f
 - analysis of, 2:720–722
 - decomposition, 2:723f
 - chironomid records, Africa, 1:365, 1:370
 - consolidation, sea-level change, 4:372
 - constituents, qualitative/quantitative inference, 3:353–354
 - cosmogenic nuclides
 - dating methods, 1:432–434
 - landscape evolution, 1:440–441, 1:441–442, 1:443–444
 - deep-sea, 1:637–642, 2:863–864, 2:865f, 2:869f
 - deformations, periglacial conditions, 3:430–435
 - deltas, 1:693–694, 1:694–695, 1:695–699, 1:702f
 - deposits
 - beetle correlation, 1:186
 - lithostratigraphy, 4:227
 - vertebrate remains, 4:591–592
 - diatoms
 - diatom silica, 2:739–741
 - freshwater laminated sequences, 1:540–545
 - surface distribution, 1:532f, 1:566–567, 1:572f, 1:575f, 1:576f
 - dinocyst significance, 2:808–811
 - discharge, deltas, 1:693–694, 1:694–695, 1:696, 1:702f
 - dry sediment ¹⁴C, 4:362
 - environmental alkenone
 - paleothermometry, 2:759–760, 2:759t, 2:760t
 - fluvial
 - allostratigraphy, 4:246f, 4:248
 - environments, 1:663–675
 - plant macrofossil/pollen studies, 3:726–727
 - Quaternary records, 1:670–674
 - morphostratigraphy, 4:243
 - systems, Quaternary, 1:670–674
 - forensic science use, 4:535–541
 - freshwater laminated sequences, 1:540–545
 - frozen, insect preservation, 1:174
 - glacial
 - allostratigraphy, 4:248
 - erratics indicators, 2:81–84
 - landforms
 - clast form analysis, 2:1–5
 - ice-rafted debris, 2:30–42
 - lithofacies, 2:18–29
 - till, 2:62–75, 2:76–80
 - morphostratigraphy, 4:243
 - permafrost interactions, 3:510–511, 3:511–512
 - sequence stratigraphy, 2:85–90
 - glacimarine, 2:30–42
 - glaciolacustrine, 2:43–51, 2:45f
 - hiatuses, lake-level reconstruction, 2:549
 - Holocene budgets, 3:253–255, 3:254f
 - indicators
 - glacial erratics, 2:81–84
 - sea-level change
 - high-energy coasts, 4:385–395
 - RSL, 4:396–408
 - landform–sediment
 - associations, 1:756–757, 1:759–761, 1:764–765
 - suites, 1:813–814
 - Late Neogene/Early Quaternary transition, 2:156–159, 2:160–162
 - limits, lake sediment, 3:668f
 - lysocline, 2:863–864, 2:865f, 2:869f
 - North Pacific Ocean, 3:33–34, 3:37
 - paraglacial geomorphology, 3:553–554, 3:554f
 - particle forms
 - defining, 2:1–2, 2:3t, 2:4f
 - influences on, 2:3–5
 - permafrost–glacial interactions, 3:510–511, 3:511–512
 - physical properties, lake sediment, 3:303, 3:304f
 - pigment-monitoring records, 3:331
 - plant macrofossils, 3:593–594, 3:594f, 3:609f
 - ¹⁴C, 4:362
 - fluvial environments, 3:726–727
 - plant remains analysis/identification, 3:704f
 - properties
 - principal component analysis, 3:351–352
 - qualitative/quantitative inference, 3:351f, 3:352–353
 - redistribution, magnetic proxies, 2:891–893
 - sampling
 - oribatid mite studies, 2:685–688
 - surface sediment, 3:684–689
- screening, insect fossil studies, 1:257f
- sequences
 - freshwater laminated, 1:540–545
 - stratigraphy, 4:263–264
- South American glaciations, 2:251–252, 2:253
- Southern Ocean diatom distribution, 1:532f
- storage
 - modeling, 3:554–556, 3:555f
 - paleohydrology, 3:253–255
- streams, North America, 3:768–784
- structural changes, lake levels, 2:549
- surface
 - North Atlantic/Arctic, 1:566–567
 - Pacific Ocean, 1:572f, 1:575f, 1:576f
- sample studies, 3:684–689
- Southern Ocean, 1:532f
- tephrochronology, 4:296–298
- terrigenous, 2:132f, 2:941–952
- thermokarst topography, 3:579–580
- transport
 - clast form analysis, 2:1, 2:2–3, 2:3–5
 - deep-sea, 1:637–642
 - paleohydrology, 3:253–255
 - sand sea formation, 1:623
 - solifluction rates, 3:483
 - weathering profiles, 3:392
- treeline reconstructions, 3:691–692
- visible/infrared spectroscopy, 3:349
- wetlands, plant macrofossil/pollen studies, 3:726
- windblown
 - deep-sea, 1:637–642
 - dune fields
 - high latitudes, 1:597–605
 - low latitudes, 1:623–636
 - ice core studies, China, 2:379
 - loess
 - Central Asia, 2:585–594
 - origins/properties, 2:573–584
 - micromorphology, 3:381–391
 - mineral magnetic analysis, 3:375–380
 - morphology, 3:402–411
 - overview, 3:357–366
 - weathering profiles, 3:392–401
- yield models, 3:553–554
- see also* diatomites; lake sediments; laminated sediments; marine sediments; paleolimnology; paleosols; periglacial fluvial sediments/forms; soils; varved sediments
- seed bank studies, 3:686
- seed remains, environmental archeology, 3:699, 3:703–704, 3:708–713, 3:714–715
- seesaw models, ice core comparisons, 2:412–413, 2:414–415
- segregated ice *see* ice segregation
- seismic vertical motion, Holocene RSL, 4:424–425, 4:425f
- selective linear erosion
 - blockfields, 3:530–532
 - glacial landscapes, 1:862
- selective preservation, ice wedges, 3:449
- SEM *see* scanning electron microscopy

- semiarid climates
 American Southwest, 1:691
 eolian sand, 3:357
 midlatitude dune fields, 1:606, 1:611
 soils, 3:365
- semi-deciduous forests, 4:156, 4:160
see also deciduous forest
- semienclined seas, 1:556–557
- SEMs *see* scanning electron microscopes
- sensitivity analysis, lake-level modeling, 2:561, 2:562–563
- Sepsidae (black scavenger flies), 4:551–552
- sequence boundaries (SBs), 4:260, 4:262
- sequence motifs, 4:266–269
 basin margin, 4:266–267
 inner-shelf, 4:267–269
 mid-shelf, 4:269
 outer-shelf, 4:269
 slope, 4:269
- sequence stratigraphy, 4:260–276
 astronomical calibration, 4:271
 Crotone Basin case study, 4:271, 4:272f, 4:273f
 DLS, 4:261, 4:263
 eustasy and sediment, 4:263–264
 glacial, 2:85–90
 LFS, 4:262–263
 orders of sea-level cyclicity, 4:261f, 4:263, 4:269–271
 outcrops, 4:261–262, 4:262f, 4:263t, 4:267f, 4:268f, 4:269–271
 parasequences, 4:263
 Quaternary sequences, 4:264–271
 RS, 4:260–261, 4:262
 RSE, 4:261, 4:263
 sedimentary facies analysis, 4:264
 sediment types, 4:263–264
 sequence boundaries, 4:262
 sequence motifs, 4:266–269
 sequence surfaces, 4:261–263
 shellbeds, 4:264, 4:265f, 4:266f
 SSMs, 4:260–261, 4:263, 4:264–265, 4:265–266
 study results, 4:273
 TSE, 4:260–261, 4:262, 4:265, 4:273
 Wanganui Basin case study, 4:267f, 4:268f, 4:269–271, 4:270f, 4:271f, 4:274f
- sequence stratigraphy models (SSMs), 4:260–261, 4:263, 4:264–265, 4:265–266
- sequestration, anthropogenic CO₂, 2:869
- settlement
 Early Pleistocene hominins, 1:69–71
 Europeans in Australia/New Zealand, 4:112
 Polynesian settlements, 4:112
 postglacial human adaptations, 1:158
 post-LGM humans, 1:104–106
see also human colonization; human settlements
- S forms, meltwater hypothesis, 1:833–834, 1:835, 1:836f
- SGI *see* standardized growth indices
- Shackleton, Nicholas, 4:201, 4:201f
- shallow grave searches, 4:526t, 4:527f
- shallow-marine sequences, 4:269–271, 4:271f, 4:272f
- shape properties, sediment particle forms, 2:1–2, 2:2f, 2:4f
- Shasta ground sloth, 4:716
- shears, periglacial slope deposits, 3:486–487, 3:486f
- shear strength, lake sediment, 3:309–310, 3:310f
- sheet-like sequences, subglacial, 1:796–798
- sheet sands *see* sand sheets
- sheets, subglacial volcanic, 1:793
- shelf areas
 Antarctic Peninsula, 2:219
see also continental shelf sites
- shelf configuration, deltas, 1:697
- shelf deposits, paraglacial coasts, 3:563
- shellbed sequence stratigraphy, 4:264, 4:265f, 4:266f
- shell mounds, 1:158
- shells
 freshwater mollusks, 3:285, 3:285f, 3:286–287, 3:287–288, 3:288–289, 3:288f, 3:290f
 mollusk types, 2:668, 2:670–672, 2:673–676, 2:674f, 4:194, 4:194f
 postglacial human adaptations, 1:158
 radiolarians, 2:830, 2:831f
 sequence stratigraphy, 4:264, 4:265f, 4:266f
 thickness, foraminifera, 3:35–36
 weights, marine carbon cycle proxies, 2:855–857
- shield volcanoes, 1:781–782, 1:791–792
- shifting balance theory, evolution, 4:723
- shine-down curves, OSL dating, 2:653, 2:656
- Shinker, JJ, 3:106
- ships' logbooks, 3:237–238, 3:239f, 3:239t
- shoreface-inner shelf facies, 4:386–387
- shoreline indicators, sea-level change, 4:391–395, 4:393f
- short-term incision, strath terraces, 1:688–689
- shotgun sequencing, ancient plant DNA, 2:711–712
- shrubs
 Arctic Eurasia, 3:743
 Greenland, 3:762
 Northern Asia, 4:37
 plant macrofossils, 3:598f, 3:599f
- shrub tundra, 3:743, 4:37
- SI *see* stomatal index
- Siberia
 areal ice extent, 1:891
 Early Pleistocene vertebrate studies, 4:606f
 glaciations, 2:224–228, 2:226f
 Middle-Pleistocene, 2:172, 2:174, 2:176, 2:177
 insect fossil studies, 1:173, 1:174, 1:176–177, 1:180f, 1:181f, 1:255, 1:256–257, 1:258f, 1:259t, 1:262t, 1:264t, 1:270t
 Late Pleistocene
 Beringia events, 2:191
 pollen records, 4:27–38
 MCR paleoclimate reconstruction, 1:168
- megafossils, 3:626–629
- Middle-Pleistocene glaciations, 2:172, 2:174, 2:176, 2:177
- modern human colonization, 1:103–104, 1:105–106
- mummified mammals, 4:713
 horses, 4:714
 woolly mammoth, 4:713–714
 woolly rhinoceros, 4:714
- periglacial landforms, 3:493f, 3:494f
- 'pickled' fauna, 4:716
- plant macrofossil records, 3:733–745
- postglacial pollen records, 4:164–166, 4:167, 4:168–170
- Yedoma of Beringia, 3:544–545, 3:545–546, 3:547–548, 3:548–549, 3:551t
- Sicily
 dwarfing/gigantism, 4:734–736, 4:735f, 4:736f, 4:737, 4:737f, 4:745f
 postglacial pollen records, 4:181t, 4:185–186
- side shear, West Antarctic Ice Sheet, 2:449
- Sidestrand, England beetle assemblages, 1:185
- Siebenhengste–Hohgant cave system, Switzerland, 1:444
- Sierra Nevada neoglaciation chronology, 2:273
- sieving approach, charred particle analyses, 2:717–719, 2:720–721
- significance tests, till fabric, 2:77–78
- silica
 biogeochemical role in climate cycles, 3:416–417, 3:417f
 cell wall, diatoms, 1:471
 lake eutrophication, 3:294
 maturation, diatom oxygen, 1:483–487
 phytoliths, 3:582, 3:583–584
see also biogenic silica; diatom silica
- silicate minerals, tephrochronology, 4:285–286
- siliceous ooze belt, Southern Ocean, 1:527–528, 1:529f
- silicic acid (Si[OH]₄), 2:734
 analytical procedures, 2:734–735, 2:736–737, 2:736f
 leakage hypothesis, 2:739
 marine concentrations, 2:739
 Southern Ocean uptake, 2:739f
- silicoflagellates, 2:830–840, 2:838f
- silicon-32 radioactive dating, ice cores, 2:306
- silicon isotopes
 analytical procedures, 2:734–735
 in diatoms, 2:734–743
 fractionation, 2:735–737, 2:738f
 terrestrial hydrological systems, 2:737–738
- silicon leakage hypothesis, 2:739
- silicon phytoliths, 3:582, 3:583–584
- silled basins, laminated sediments, 1:555, 1:556f, 1:557–558
- Silphidae *see* carrion beetles
- silt
 Central Asia, 2:587f, 2:589
 coatings, soil, 3:386
 eolian, 3:357, 3:359
 particle formation, 2:574–576, 2:576f

- Silvaplana Lake, Switzerland, 1:387–388, 1:388–389, 1:390f
- similarity methods, pollen–climate relation, 3:807–809
- SIMMAX method, temperature proxies, 2:843f, 2:845, 2:846–847
- SIMS *see* secondary ion mass spectrometry simulations
- lake-level modeling, 2:560–561, 2:561–562
- pollen representation, 3:871–879
- Singilian fauna, Northern Eurasia, 4:612–613
- single-aliquot additive-dose (SAAD) dating, 2:657–658, 2:658–659, 2:662–664
- single-aliquot regeneration and added dose (SARA) dating, 2:657–658
- single-aliquot regenerative-dose (SAR) dating, 2:658–659, 2:659f, 2:662
- single-aliquot techniques, OSL, 2:657–660, 2:659f, 2:660–661, 2:662
- single-crystal laser fusion (SCLF), 1:11, 4:287–288
- single-origin models, human evolution, 1:146
- single site studies, pollen–climate relation, 3:811–812
- ‘single stage’ (0.3 MV) AMS, 4:322–323, 4:323f
- Sino-Malayan route, Southern Asia, 4:654f
- Si[OH]₄ *see* silicic acid
- Siple Dome, Antarctica, 2:436f, 2:437
- SIRM *see* saturation isothermal remanent magnetization
- site selection principle, dendrochronology, 1:455, 1:455f
- site transects, plant macrofossil/pollen studies, 3:731
- Siva-Malayan route, Southern Asia, 4:654f
- Siwalik Hills, India, 4:651, 4:652t
- size change rule, mammals, 4:697–698
- size-normalized shell weights (SNWs), 2:855–857
- skeletal associations, technocomplexes, 1:150t
- skeletal morphology, modern human origins, 1:91
- skin beetles, 4:553–554
- skins, hominid raw materials, 4:751
- skipper flies, 4:551
- skulls, human evolution, 1:138f
- slab oceans, 3:150, 3:151f
- slackwater flood deposits, 3:255
- slash and burn practices, 3:889–890
- sliding bed deposits, subglacial till, 2:62, 2:63f, 2:67–68
- SLIPs *see* sea-level index points
- slope collapse, cirque glaciers, 1:849–850
- slope deposits, 3:481–489
- slope evolution, periglacial, 3:488
- slope forms, 3:481–489
- slope hummocks, 3:460, 3:461f, 3:462
- slope land systems, 3:556–559
- sloths
- ground sloths, 4:685–686, 4:716
- South American Late Pleistocene, 4:685–686
- slow mass movement, slope deposits/forms, 3:481–484
- slow negative muon interactions, 1:427
- small mammals, biozonations, 4:210
- Smith AG, 2:132–134
- ‘smoking gun’ theory, megafaunal extinctions, 4:711
- Smorodinovoye Lake, northeastern Asia, 4:170–171, 4:171f
- snails, 3:285, 3:286f, 3:287f
- Snake River Plain, North America, 2:622–624, 2:626
- snakes, Australia, 4:628–629
- snow avalanches, 3:570, 3:570f, 3:571f
- snow-bed vegetation, Arctic Eurasia, 3:743
- snowfall, ice sheets, 2:319, 2:435
- snow–ice transformation, Greenland, 2:442
- snowshoe hares, 4:715
- Snowy Mountains, Australia, 2:202–204
- SNWs *see* size-normalized shell weights
- SO *see* Southern Oscillation
- social brain evolution, 1:55–56
- social change, human evolution, 1:54, 1:57
- social construction, human archeology, 1:87–88
- social relations, hominids–animals, 4:748
- social relevance
- paleodroughts and, 3:194–199
- research, 1:17–25
- socio-economic patterns, oribatid mites, 2:695
- Socotra Island, Indian Ocean, 3:202f
- soft hammers, definition, 1:83ge
- software, pollen numerical analysis, 3:827, 3:827t
- soils
- age–morphology relation, 3:405
- anthropogenic features, 3:383–386, 3:385f
- buried soils
- identification, 3:381
- paleoenvironmental concerns, 3:381–383
- ¹⁴C of plant macrofossils, 4:362
- calcite coatings/infillings, 3:383
- carbonates
- accumulation, 3:406–407, 3:406f, 3:407–408, 3:408f
- non-lacustrine studies, 1:322, 1:323f, 1:324f, 1:325–326, 1:326–327, 1:328–329, 1:328f, 1:329–330
- charcoal analysis, 2:716, 2:722–723, 2:724t
- chemistry, 3:377–378
- circulation, sorted patterned ground, 3:457–458
- clay/silt coatings, 3:386
- color properties, 3:402–403, 3:404f, 3:405–406, 3:406f, 4:537
- composite paleosols, 3:367
- cryoturbation process, 3:424
- definitions, 3:367
- degradation statistics, 1:21t
- diatom analysis, 1:524
- diffusion model, 1:326–327
- eluvial theory, loess origins, 2:585–587
- forensic science use, 4:535–541
- forming processes–properties relation, 3:402–403
- frequency-dependent susceptibility, 3:379–380
- ice wedges, 3:440, 3:441f, 3:443–446, 3:448–449
- iron/manganese/aluminum coatings, 3:383
- lithosphere history, 1:20–21
- magnetic properties, 3:377–379, 3:379–380
- micromorphology, 3:381–391
- microstratigraphic relations, paleosols, 3:386–388
- moisture
- active-layer thickness, 3:425
- balance, lake-level modeling, 2:559–560
- morphology, 3:402–411
- age of soil relation, 3:405
- arid climates, 3:363f, 3:365
- climate relation, 3:407–408
- depositional systems and, 3:408
- examples, 3:364f
- future of studies, 3:410–411
- humid climates, 3:363f, 3:365
- quantitative evaluation, 3:409–410
- state factor, 3:403–405
- stratigraphy, 3:365
- time-dependent changes, 3:405–407
- overview, 3:357
- palynology, 3:846
- production of, 1:442
- profile indices, morphology, 3:409–410
- properties–forming processes relation, 3:402–403
- quarries, 4:537
- in Quaternary records, 3:360–366, 3:363f
- reconstructions, pedosedimentary/landscape evolution, 3:388–389, 3:390f
- stand-scale palynology, 3:846
- thermal properties, 3:425
- water evaporation, 1:323
- weathering
- profiles, 3:392
- trimlines/paleounataks, 1:923
- see also* paleosols; pedostratigraphy; sediment...
- soil veins, ice wedges, 3:440, 3:441f, 3:443–446, 3:449
- soil wedges, ice wedges, 3:440, 3:443–446, 3:448–449
- solar activity
- carbon-14 temporal variations, 4:338–340, 4:340f, 4:341f
- cosmogenic radionuclide reconstructions, 2:356–357
- ice-core chronologies, 2:303–304
- lake-level studies, west-central Europe, 2:553–555, 2:555–556
- postglacial Indian Ocean, 3:52
- see also* solar forcing; solar radiation
- solar cosmic rays (SCRs)
- energy spectra, 1:420–422
- proton spectra, 1:423f
- solar cycles, 1:748

- solar flares, 4:339–340
 solar forcing, 1:748–753, 3:234–235
 solar modulation
 cosmogenic radionuclides, 2:353, 2:354f
 TCN production, 1:415
 solar plasma, 1:420
 solar proton events (SPEs), 2:354
 solar radiation
 Arctic paleoclimate history, 3:117
 atmospheric dust, 1:734–735
 HTM, North American pollen records, 4:48
 ice sheet growth/decay, 1:879
 lake-level modeling, 2:559, 2:561
 neoglacal Europe, 2:265–266
 see also insolation; solar activity
 soldier flies, 4:552
 solid-source counting, radiocarbon dating, 4:306
 solifluction
 landforms, 3:483–484
 lobes, 3:483–484, 3:485f
 pore pressures, 3:482f
 sediment transport rates, 3:483
 slope deposits/forms, 3:481–482, 3:482–483
 terraces, 3:483–484
 soluble major ions, ice cores, 2:326–333
 soluble microparticles, aerosols, 2:343
 Solutrean origins model, human colonization, 1:119
 Sonoran Desert
 insect fossil studies, 1:288–289
 pollen records, 4:142
 rodent middens, 3:677–678
 sand seas, 1:633f, 1:634
 SOP *see* standard operating procedure
 sorted patterned ground, 3:452–458
 source area, pollen, 3:872–873
 South Africa
 fauna, Middle Pleistocene, 4:618, 4:619
 relative sea-level changes, 4:493
 South America (SA)
 chironomid records, 1:399f, 1:400–402, 1:402–403, 1:403–404
 dune fields, 1:611–614, 1:614f
 Early Quaternary, 2:167
 glaciations, 2:250–256
 Central Andes, 2:250–253, 2:254, 2:255
 history of, 2:148
 Late Glacial, 2:250, 2:252, 2:253, 2:254–255
 Late Pleistocene, 2:250–256
 LGM, 2:250, 2:252–253
 Middle Pleistocene, 2:187–189
 Northern Andes, 2:250, 2:254
 pre-LGM, 2:252
 Southern Andes, 2:253–255
 human global expansion, 1:119, 1:121–122, 1:123, 1:125, 1:128–129
 ice core studies, 2:387–394
 insect fossil studies, 1:235–243
 lake level studies, 2:531–536
 Lake Titicaca
 diatom records, 1:546–547, 1:550–551
 glaciations, 2:250–251, 2:252
 salinity/climate reconstruction, 1:509–510, 1:510f
 loess, 2:629–641
 Antarctic dust, 2:639
 carbonate accumulations, 2:630–631, 2:633f
 distribution, 2:575f, 2:629, 2:630f, 2:635, 2:638t
 impact glasses, 2:632, 2:634f, 2:634t
 last glacial cycle, 2:635–638, 2:636f, 2:638t
 magnetic susceptibility, 2:632–634
 magnetostratigraphy, 2:632, 2:634–635, 2:635t
 paleoclimatic significance, 2:638–639
 paleosols, 2:629–630, 2:632f, 2:634–635, 2:636t
 Pampean Loess, 2:629–635, 2:631f, 2:631t, 2:633t, 2:635t, 2:637, 2:637f, 2:638–639, 2:640
 sources of, 2:639f
 Tucumán loess, 2:631t, 2:634–635, 2:635f, 2:636t, 2:637, 2:638t, 2:640
 vertebrate fossil content, 2:631–632
 wind systems, 2:638–639, 2:638t
 megafaunal extinctions, 4:704
 environmental change hypothesis, 4:710
 timing of, 4:706t
 neoglacal chronology, 2:271–274, 2:274f
 paleoecological studies, 4:52
 pollen records
 Late Pleistocene, 4:52–62
 postglacial, 4:156–163
 relative sea level changes, 4:493
 sand seas, 1:634
 Southern Cone, 4:683
 South Pacific postglacial records, 3:74–76, 3:75f, 3:76–77, 3:77f, 3:78f, 3:79f, 3:81f, 3:82–84, 3:84f
 tropical, 2:387–394
 physical/meteorological setting, 2:387–390
 precipitation over, 2:389, 2:390–391
 vertebrate studies, 4:680–692
 Younger Dryas oscillation, 3:225
 see also individual countries
 South Asia
 Middle-Pleistocene vertebrate studies, 4:651–663
 modern human expansion, 1:99–100, 1:102f
 see also Asia; Indonesia; Southeast Asia
 South Atlantic, Late Pleistocene, 3:18–32
 Agulhas leakage, 3:22, 3:23f
 central water masses, 3:28
 coastal upwelling, 3:21–27
 deep water records, 3:28–29
 freshwater input, 3:22–24
 intermediate water records, 3:28–29
 IRD, 3:24, 3:27f
 meltwater, 3:24
 present-day circulation, 3:18, 3:19f
 sea-ice, 3:24–26
 SSTs, 3:18–19, 3:20f, 3:21f, 3:22f, 3:25f
 surface-water records, 3:18–19
 thermocline depth, 3:19–21, 3:28
 upper ocean stratification, 3:19–21
 South China
 cave vertebrate records, 4:655–657, 4:656t
 sea-level changes, 4:444
 Southeast Asia
 biogeography, 4:693, 4:697
 Late Pleistocene mammal assemblages, 4:693–699
 Middle Pleistocene hominin archives, 1:76–77
 Southeast China, sea-level changes, 4:444
 southeastern Australia
 beetle records, 1:195–196, 1:196–197
 pollen records, 4:21–23
 southeastern North America
 Minnesota site plant macrofossil records, 3:775–777, 3:778f
 Nebraska site plant macrofossil records, 3:777–779, 3:780f
 postglacial pollen records, 4:133–141
 see also Eastern North America
 South East (SE) trade winds, 3:18, 3:23–24
 southern Africa
 archeological evidence, modern human origins, 1:92–94
 fauna, Middle Pleistocene, 4:618–619
 lake dynamics, Late Quaternary, 2:501–502
 vertebrate fossil records, 4:664–669
 see also Africa; *individual countries*; South Africa
 Southern Alaska, neoglacal chronology, 2:272
 Southern Alps, New Zealand, 2:207, 2:209–210
 southern Andes, 2:273–274
 see also Andes
 southern Brazil vertebrate fossil sites, 4:682–683
 Southern Canadian Cordillera, 2:540–543
 southern Chile, 1:235
 see also Chile
 Southern Cone, South America, 4:683
 Southern Europe
 Middle/Late Pleistocene pollen records, 4:63–71
 postglacial pollen records, 4:179–188
 see also Europe; *individual countries*
 Southern Hemisphere
 glaciation
 Pleistocene, 2:187–190
 relative glacier extent, 2:275
 LGM vegetation patterns, 3:864
 mid-Holocene vegetation patterns, 3:864–865
 postglacial chironomid records, 1:398–405
 Younger Dryas, 3:129, 3:225–226, 3:227
 see also individual countries/features
 southern North Sea, RSL changes, 4:492
 Southern Ocean
 deep-water, 3:129–130
 diatom biostratigraphy, 1:536
 drilling projects, 1:531f
 magnetic susceptibility, 2:889, 2:893–895, 2:894f

- ocean frontal migration, 1:528f
 sea-surface temperature, 1:528–529, 1:530
 siliceous ooze belt, 1:527–528, 1:529f
 silicic acid uptake, 2:739f
 study foci, 1:529–530
 surface sediment diatom distribution, 1:532f
 Southern Oscillation (SO), 2:379, 2:384
 see also El Niño–Southern Oscillation
 Southern Oscillation Index (SOI), 3:231, 3:232f
 Southern Westerlies wind patterns, 3:73, 3:76, 3:82–84
 South Island, New Zealand, 2:206f, 2:207–212
 South Pacific Ocean, postglacial records, 3:73–85
 Antarctic Peninsula, 3:80–81, 3:81f, 3:82–84, 3:84f
 Australia, 3:77–80
 Chile, 3:74–76, 3:75f, 3:76–77, 3:77f, 3:78f, 3:82–84, 3:84f
 continental margin records, 3:74–81
 ENSO, 3:73–74, 3:77, 3:79f, 3:80, 3:82–84, 3:83f
 forcing mechanisms, 3:82–84
 general ocean circulation, 3:73
 maps of South Pacific, 3:74f
 millennial–centennial forcing timescales, 3:82–84
 New Zealand, 3:77–80, 3:80f
 oceanographic/climatic setting, 3:73–74
 open ocean records, 3:82
 orbital forcing timescales, 3:82, 3:83f
 Palmer Deep, 3:80–81, 3:81f, 3:82–84
 Peru, 3:76–77, 3:79f, 3:84f
 South American margin, 3:74–76, 3:75f, 3:76–77, 3:77f, 3:78f, 3:79f, 3:81f, 3:82–84, 3:84f
 wind pattern, 3:73, 3:76, 3:82–84
 South Pacific Ocean Islands, glaciations, 2:189
 southwestern North America
 fossil site distribution, 3:675
 Indiana site plant macrofossil records, 3:779–782, 3:781f
 Missouri site plant macrofossil records, 3:782f, 3:783
 postglacial pollen records, 4:142–155
 see also Western North America
 southwest Pacific region, Late Quaternary, 2:202–215
 Spain
 archives, paleotempestology, 3:217–218
 Middle/Late Pleistocene pollen records, 4:65
 relative sea level changes, 4:493
 spallation, landscape evolution, 1:440, 1:440ge
 spatial change
 active temperature glacial landsystems, 1:817–820
 glaciated valley landsystems, 1:819f, 1:820–822
 ice stream landsystems, 1:821f, 1:822
 spatial climate heterogeneity, 3:110
 spatial distribution, carbonate-precipitating lakes, 1:333
 spatial fidelity, freshwater mollusks, 3:289–290
 spatial patterns
 loess distribution, 2:573–574, 2:574f, 2:575f
 weathering profiles, 3:397–399
 spatial relations, pollen–climate relation, 3:812–813
 spatial sampling, plant macrofossils, 3:710f
 spatial-scale vegetation reconstruction, 3:861
 spatial structures, vegetation, 3:871–872
 spatial variability
 diatom records, 1:578f, 1:582f, 1:583f, 1:584f
 glaciochemical sources, 2:328–332
 specialization, hunting, hominids, 4:749, 4:750–751
 speciation events
 dispersals/diasporas, 1:54
 Middle Pleistocene hominins, 1:80
 vertebrates, 4:723–732
 species-level diatom studies, 1:544
 species origin vertebrate studies, 4:724–725
 species pairs, vertebrate evolution, 4:724
 species richness, biodiversity, 3:816
 specific-compound dating, plant macrofossils, 4:364–365
 specimen identification, plant remains, 4:544–545
 SPECMAP approach, marine oxygen isotopes, 2:910–912, 2:911f, 2:913f
 spectral analysis
 definition, 1:637ge
 eolian sediments, 1:639
 spectrometers, TL dating, 2:648f
 spectrophotometers, 4:540
 spectroscopy
 forensic palynology, 4:561–562
 paleolimnological applications, 3:349–355
 spectrum of climate variability, 3:94–98
 changing variability, 3:96
 spurious periodicity, 3:97–98
 timescales of climate change, 3:94–98
 variance spectrum, 3:94–96, 3:96f
 speleothems, 1:294–303
 carbon-14 record calibration, 4:346
 dating techniques, 1:296–297, 2:672, 4:463f, 4:570
 definition, 3:768ge
 ESR dating, 2:672
 future directions/problems, 1:301
 hiatuses in, 4:461, 4:464–465
 lower bounds, sea-level position, 4:462–463
 paleoclimate inferred from, 1:297–298
 phreatic overgrowths on, 4:464
 prairie-forest border, 3:768–769, 3:773f
 research overview, 1:292
 sea levels
 indicators, 4:383–384, 4:460
 lower bound position, 4:462–463
 upper bound position, 4:460–462
 systematics, 1:294–296
 upper bounds, sea-level position, 4:460–462
 U-series dating, 4:463f, 4:570
 see also flowstones; stalactites; stalagmites
Spermophilus parryi *see* arctic ground squirrel
 SPEs *see* solar proton events
 Sphaeroceridae (lesser dung flies), 4:551
Sphagnum
 ¹⁴C of plant macrofossils, 4:364
 mire/peat macrofossils, 3:637, 3:638, 3:646–648, 3:647f, 3:648f, 3:649, 3:650f
 oribatid mite studies, 2:686f, 2:690f, 2:693–694
 spillways, glacial lake outbursts, 1:826–827, 1:827f, 1:828f
 Spingallo cave, Sicily, 4:734–736, 4:736f, 4:737f, 4:745f
Spiniferites dinocysts, 2:810f, 2:812f
 split-flow thin fractionation (SPLIT), diatoms, 1:484f
 splitting lithostratigraphic units, 4:235–238
 sponges, freshwater, 3:652, 3:653f
 sponge spicules, diatom silica, 2:739
 spores, pollen analysis, 3:794–795, 3:795f
 sporophyte generation, plant macrofossils, 3:725
 spring bloom, diatoms, 1:554, 1:558, 1:559
 Spring Creek, Southeastern Australia, 1:195–196
 spring deposits, oribatid mites, 2:695
 spruce (*Picea abies*)
 paleophytogeography, 2:732
 plant distributions, 3:856, 3:857f
 pollen records, North America, 4:116f, 4:117–118
 wood, megafossils, 3:621, 3:622f, 3:625–626, 3:626f, 3:626t, 3:627t
 spur erosion marks, sea level indicators, 4:380, 4:380f
 spurious periodicity, climate change, 3:97–98, 3:98f
 square-wave model, glacial–interglacial cycles, 2:112, 2:113f
 squeeze moraines, 1:769
 squirrels, 4:715
 Sr/Ca *see* strontium/calcium
 SSMS *see* sequence stratigraphic models
 SSTs *see* sea-surface temperatures
 stability
 Greenland ice sheet, 2:446
 hypothesis, Amazonian pollen records, 4:59–61
 stable flies, 4:548, 4:550–551
 stable isotopes
 analysis
 cladocera, 3:276f, 3:278–279
 forensic palynology, 4:562–564
 phytoliths, 3:587–588
 pigment analysis, 3:330
 chironomid records, Late Pleistocene, 1:374
 dendrochronology, 1:456
 geochemistry concepts, 1:291–292
 ice core studies, 2:279–280, 2:347–352
 Andes, 2:389–390

- stable isotopes (*Continued*)
 Antarctica, 2:395–402
 China, 2:379–380
 Greenland, 2:403–409
 ice-margin sites, 2:419–424, 2:421f
 lake level studies, Latin America, 2:533–534
 lake sediments, 1:333–340
 mollusks, 2:798–799, 3:285–289
 non-lacustrine terrestrial studies, 1:322–332
 nonmarine biogenic carbonates, 1:341–353
 pigment analysis, 3:330
 polar region distribution, 2:347–348
 research overview, 1:291–293
 salinity proxies, 2:932–940
 seawater content, 2:915–922
 speleothem research, 1:294–303
 standard notation/measurements, 2:347
 systematics in speleothems, 1:294–296
 teeth/bones, 1:304–314
 tree rings/stalagmites, 3:214–215
see also carbonate stable isotopes
- stadials
 in climatostratigraphy, 4:222–223
 definition, 2:202
- stalactites, 1:294, 1:295f, 4:460–461
see also speleothems
- stalagmites, 1:294, 1:295f, 1:296f
 layer analysis, 1:297f
 oribatid mites, 2:695
 paleotempestology, 3:214–215
 sea-level position
 highstand, 4:463f
 upper bounds, 4:460–461
see also speleothems
- standardization, tree-ring dating, 1:456
- standardized growth indices (SGI), mollusks, 2:798
- standard operating procedure (SOP), forensic searches, 4:525t, 4:529
- stand-scale palynology, 3:846–853
 case study insights, 3:848–853
 charcoal/macrofossils, 3:851f, 3:852–853, 3:852f
 closed-canopy sites, 3:847–848, 3:847t
 compositional trends, 3:848–850
Fagus sylvatica, 3:850–852, 3:851f
 history, 3:846
 macrofossils/charcoal, 3:851f, 3:852–853, 3:852f
 pollen representation theory, 3:846–847
 rate of vegetational change, 3:850, 3:850f
 Scandinavia, 3:846, 3:847f, 3:848–853
 site types, 3:847–848
 Sylvania, USA, 3:848
T. cordata, 3:850–852, 3:851f
- Stanton Harcourt, England, beetle assemblages, 1:188
- Staphylinidae (rove beetles), 1:169, 1:212, 1:225, 4:554
- stasis *see* Quaternary stasis
- state factor, soil morphology, 3:403–405
- statistical analysis/approaches
 ice core comparisons, 2:414
 lichenometry, 2:570–571
 till fabric data, 2:76–78
 VIS/IR spectroscopy, 3:351–352
- steady-state equilibrium line
 definition, 1:909ge
 glacial landforms, 1:910–911, 1:911f
 steady-state profiles, graded streams, 1:685
- Stefan Solution, active-layer thickness, 3:424–425
- stegodonts, 4:734, 4:742–743
- stem-and-leaf technique, dendroarcheology, 3:632–633
- stenothermic beetle species, 1:162, 1:163f
- steppe
 tundra, 1:259–261
 Arctic North America, 3:755
 North American beetle studies, 1:233
 plant macrofossil records, 3:739, 3:740–741, 3:740f, 3:741–742
 Yedoma of Beringia, 3:550–551
 vegetation reconstruction, 3:742–743, 3:866–867
- steppe bison, 4:714–715
- stepped combustion charcoal pretreatment, 4:357
- stepwise fluorination (SWF), 1:482
- stereo binocular microscopes, 4:538
- stereology, glacial sediments, 2:56–58
- stereonets, till fabric analysis, 2:76, 2:77f
- sterol diatom biomarkers, 1:590
- steryl chlorin esters (SCEs), 3:326, 3:336, 3:337f
- Stewart Island, New Zealand, 2:212
- stick-nest rats, 3:674
- stick-slip events, West Antarctic Ice Sheet, 2:454
- 'stiffness' of ice, 2:456
- stochastic models
 lakes, 2:558
 radiocarbon dating, 4:325
- Stoke Goldington, England beetle assemblages, 1:188
- stomach contents, plant cell forensics, 4:543
- stomata
 counting method, 3:614f, 3:615
 fossil leaf CO₂ reconstruction, 3:613–620
- stomatal density (SD), fossil leaves, 3:614, 3:615, 3:617
- stomatal frequency analysis
 basis of method, 3:613–614
 calibration, 3:615–616
 fossil leaf CO₂ reconstruction, 3:613–620
 limitations of method, 3:619
 outline of method, 3:614–616
 overview, 3:616–619
 potential of method, 3:619
 presentation of results, 3:616
 site selection, 3:614
- stomatal index (SI), fossil leaves, 3:614, 3:615, 3:617f
- Stommel's salinity model, 1:742, 1:742f
- Stomoxys calcitrans* (stable fly), 4:548, 4:550–551
- Stone Age *see* Early Stone Age; Late Stone Age; Mesolithic...; Middle Stone Age; Neolithic; Paleolithic...
- stone runs *see* block streams
- stone tools
 adaptive radiation and, 1:51
 Africa 2.7 MYR-300,000 years ago, 1:59–66
 Asia, 1:79f, 1:80f
 Australia, 1:115
 Europe 1.9 MYR-300,000 years ago, 1:83–90
 social brain evolution, 1:55–56
 technocomplex groupings, 1:147
see also lithic technology
- Stony Creek Basin, Australia, 1:193–194
- storage of food, economics, 3:708–713
- stored product cases, forensic entomology, 4:548
- storm events, 2:770–772
see also paleotempestology
- strain, till fabric analysis, 2:79
- Stramproy Formation, Netherlands, 4:238
- strand flat erosion marks, 4:378–379
- strandlines, 4:503–519
- strath terraces
 dating, 1:686, 1:686f
 formation, 1:685, 1:686–688, 1:687f, 1:688–689
 incision, 1:684, 1:686–688, 1:688–689
- stratification
 models, lake water surface, 2:559
 upper ocean, South Atlantic, 3:19–21
- stratified scree, 3:570
- stratigraphy
 AAR, 4:192–193
 alkenone paleothermometry, 2:762
 allostratigraphy, 4:243–249
 aminostratigraphy, 1:44–45
 architecture, subaquatic glacial, 1:761–763
 Australian megafauna, 4:194–195
 biostratigraphy
 continental, 4:206–214
 oribatid mite applications, 2:690–692
 RSL, 4:400–401, 4:423
 carbon-14 record calibration, 4:350f
 charcoal records, 2:717–720
 chronostratigraphy, 4:215–221
 climatostratigraphy, 4:222–226
 continental biostratigraphy, 4:206–214
 correlation, 4:190–192
 basic principles, 4:190
 commonly used methods, 4:190–191
 global chart, 4:202f
 marine-terrestrial, 4:195–201
- data
 definition, 2:216ge
 Greenland ice sheet, 2:443–445, 2:444f
 Neanderthal demise, 1:150–151
- distribution, planktic foraminifera, 2:827–828
- Early Pleistocene paleoceanographic records, 3:1–3
- fluvial environments, 1:676–678, 1:682
- frequency diagram, phytoliths, 3:589f
- glaciofluvial, 2:260–262, 2:261f
- glaciolacustrine, 2:260–262, 2:261f
- global correlation chart, 4:202f
- GSSP for Pleistocene, 4:191–192
- ice-core research, 2:279, 2:351

- insect fossil studies, 1:256–257
isotopic, ice cores, 2:351
lake sediment physical properties, 3:302
Last Interglacial, 4:193–194, 4:193f, 4:200
lithostratigraphy, 4:227–242, 4:398, 4:400, 4:400f
loess, 2:577–579, 2:577f, 2:582f
 China, 2:595–596, 2:596f, 2:599f
 European records, 2:609, 2:611f, 2:612f, 2:617f
 North American records, 2:620–624, 2:626
marine-terrestrial correlation, 4:195–201
markers
 paleosols, 3:366, 3:368–370
 soils, 3:365
moraine-ridges, 2:257–259, 2:258f
morphostratigraphy, 4:243–249
numerical age estimates, 4:192–195
observations, permafrost–glacial interactions, 3:508
ordering carbon-14, 4:350f
oribatid mite applications, 2:690–692
overview, 4:189–205
oxygen isotope paleoceanography, 2:745, 2:907–914
paleomagnetic dating, 1:718–719
pedostratigraphy, 4:250–259
permafrost–glacial interactions, 3:508
phytolith frequency diagram, 3:589f
planktic foraminifera distribution, 2:827–828
pollen
 analysis, 3:801, 3:802
 Late Pleistocene records, 4:27, 4:28t
 sequence numerical analysis, 3:823, 3:824
radiocarbon dating, 4:192
sections, North America, 2:183f
sequences, 3:823, 3:824, 4:260–276
subaquatic glacial depositional systems, 1:761–763
tephrochronology, 4:277–304
vertebrate fossils, 4:595–596
 see also magnetostratigraphy
Stratiomyidae (soldier flies), 4:552
stratosphere, 1:752
stratotypes
 definition, 4:228–229, 4:229–232
 nomenclature, 4:229–232, 4:232t, 4:233–234, 4:233t
stratovolcano construction, 1:780ge
streaklines *see* flowstripes
stream sediments
 alluvial, 3:768–784
 cosmogenic nuclides, 1:441–442
strepsirrhine primates, 4:742
stresses
 micro/macroscale erosional forms, 1:865–867, 1:867–869, 1:867f, 1:868f, 1:869f
 thermally-induced rock weathering, 3:501–502
striae erosional forms, 1:865–867, 1:866f, 1:870, 1:871–872, 1:871f, 1:873–874, 1:875f
stripes, patterned ground, 3:453f, 3:455–456, 3:456f, 3:459–460, 3:461f
strontium (Sr)/calcium (Ca)
 coral aragonite monthly oscillations, 2:880
 paleothermometry, 2:871–883
strontium stable isotopes, freshwater mollusks, 3:287
structural evidence, environmental archeology, 3:699
structural notch sea level indicators, 4:377
STT *see* subglacial traction till
Stylatractus universus radiolarians, 2:833f
subaerial mesoforms
 bars, 2:10–11, 2:12
 eddy bars, 2:12
 expansion bars, 2:11
 pendant bars, 2:12
 see also proglacial landforms
subaerial outwash fans, 2:14–15, 2:14f
sub-Alpine area, west-central Europe, 2:549–551
sub-Antarctic island glaciers, 2:213–214
subaquatic glacial depositional systems, 1:761–764
 continental shelves, 1:763
 epicontinental seas, 1:763–764
 fjords, 1:763
 stratigraphic architecture, 1:761–763
subaqueous debris flows, 2:44–45, 2:47f
subaqueous moraines *see* De Geer moraines
subaqueous outwash fans, 2:44f, 2:46–48, 2:48f, 2:49–50
sub-Arctic
 gyre, North Pacific, 3:64–67
 North American sites, 3:749
 water masses, Pacific Ocean, 1:571, 1:576f
subduction zone, earthquake deformation cycle, 4:504–505, 4:510f, 4:511f, 4:514f, 4:516f
subfossils
 oribatid mites, 2:680–681
 preservation, 3:706f
subglacial depositional system, 2:21–24
subglacial landforms
 cross-cutting lineations, 1:758–759
 glacier/ice-sheet beds, 1:755–756
 ice-sheet dynamics, 1:758–759
 permafrost–glacier interactions, 3:509–510
 regional distribution, 1:757–758
 sediment associations, 1:756–757
 soft/hard beds, 1:756–757
 volcanic, 1:780–802
subglacial processes
 deformation, 2:62–64, 2:63f, 2:64f, 2:65–66, 2:65f, 2:69–71
 lodgment, 2:62, 2:63f, 2:65–66
 melt-out, 2:62, 2:63f, 2:67, 2:67f, 2:73
 ploughing, 2:65
 sliding, 2:62, 2:63f, 2:67–68
 till production, 2:62–75
subglacial sheetfloods, 1:833–836
subglacial sheet-like sequences, 1:796–798
subglacial tills
 fabric analysis, 2:78–79
 nomenclature, 2:71–73
 types, 2:66–71, 2:72f, 2:73
subglacial traction till (STT), 2:72f, 2:73
subglacial water, West Antarctic Ice Sheet, 2:454
submerged sedimentary indicators, sea-level change, 4:393–395
submerged speleothem record, 4:460–462, 4:462–463, 4:463f
sub-Milankovitch events, 3:200–208
 DO events, 3:200–208
 HEs, 3:200–208
 high-latitude signals, 3:200–203
 low-latitude signals, 3:204
 millennial climate change modeling, 3:206–207
 north-south linkage, 3:205–206
 ocean hydrography, 3:204–205, 3:206–207
 oscillations, 4:225
 variations, 1:27–29
suborbital forcing, Southern Europe, 4:68–69
subsidence
 continental margin fluvial records, 1:671–674
 Late Quaternary tectonics, 4:503, 4:507f, 4:514f
subsistence
 Early Pleistocene hominins, 1:69–71
 European hominins, 1:88
 hominids, 4:748–751
 postglacial human adaptations, 1:154
subspecies origin studies, vertebrates, 4:724–725
subsurface ocean circulation, North Pacific, 3:33
subtidal notch sea level indicators, 4:388
succession
 biogeographical/paleolimnological research, 3:317–319
 tree growth reactions, 2:94
successional replacement theory, Neanderthal demise, 1:150
‘Suess’ effect, carbon-14 variations, 4:333, 4:334f
‘Suess’ wiggles, 4:330
Suffolk, England beetle assemblages, 1:187
Sugita *see* Prentice–Sugita
Sugworth, England, beetle assemblages, 1:185
Suigetsu Lake, Japan, 3:224–225
Sulawesi island, Southern Asia, 4:662
sulfate aerosols, ice cores, 2:329–330
sulfur (S)
 cycle, ice cores, 2:326, 2:330–331
 lake acidification, 3:292, 3:293
 magnetic susceptibility, 2:896f
Sumatra, Late Pleistocene mammals, 4:695
Sumatran caves, Sunda Shelf, 4:657–661, 4:661t
Sumba island, Southern Asia, 4:662
summary statistics, till fabric, 2:76–78
summer balance, definition, 1:909ge
summer surface, definition, 1:909ge
summer temperatures, chironomid records, 1:387–392
Sun–climate link, 2:356, 2:358
 see also insolation; solar...; sunspots

- Sundaic Province/Sundaland, 4:693, 4:694f
Indochinese fluctuating boundary, 4:697
Late Pleistocene mammals, 4:694–696
Sunda Shelf
 sea-level change, 4:443
 vertebrate fossils, 4:653–654, 4:657–662, 4:661t
sunspots
 atmospheric carbon-14 variations, 4:331, 4:331f
 carbon-14 temporal variations, 4:338–339
 volcanic/solar forcing, 1:748
Superior Lobe, United States, 1:829f
superior plant macrofossil ^{14}C , 4:364
supernovae, 4:340, 4:341f
Superposition law, stratigraphy, 4:190
suppression of tree growth, 2:94, 2:94f
supraglacial depositional system
 facies model, 2:26f
 terrestrial, 2:24
supralysoclinal dissolution, deep-sea
 carbonates, 2:864
supratidal notch sea level indicators, 4:388
surface-air temperature (SAT), last
 interglacial, 3:155, 3:156–159, 3:160–162, 3:161f, 3:163f
surface albedo, ice sheets, 1:879
surface biophysics, land-surface, 1:722, 1:722f
surface characteristics, blockfields, 3:523
surface circulation, North Atlantic Ocean, 3:9
surface corers, plant macrofossils, 3:685, 3:685f
surface currents, North Pacific Ocean, 1:572f
surface dating, dendroglaciology, 2:104–108
surface displacement vectors, rock glaciers, 3:537f
surface energy balance (SEB), lake levels, 2:559
surface exposure dating (SED)
 cosmogenic nuclides, 1:410
 trimlines/paleonunataks, 1:924–925
surface mass balance, Greenland ice sheet, 2:440–441
surface morphology, rock glaciers, 3:538–539
surface samples
 methodology, 3:685–686
 plant macrofossils, 3:684–689
 pollen studies, 3:839–845
 review of, 3:684–685
 significance, 3:684
surface sediment diatom distribution
 North Atlantic/Arctic, 1:566–567
 Pacific Ocean, 1:572f, 1:575f, 1:576f
 Southern Ocean, 1:532f
surface sequence stratigraphy, 4:261–263
surface temperature
 air-surface, last interglacial, 3:155, 3:156–159, 3:160–162, 3:161f, 3:163f
 Greenland ice sheet, 2:439–440
 ground, permafrost, 3:465–466, 3:470
 see also sea-surface temperatures
surface texture, sediment particle forms, 2:1, 2:2
surface topography, Greenland ice sheet, 2:439, 2:440f
surface velocities, ice sheets, 2:443
surface water
 biogeochemical record, North Pacific Ocean, 3:42–44
 $\delta^{18}\text{O}$ distribution, seawater, 2:918f
 hydrographic record, North Pacific Ocean, 3:37–42
 Late Quaternary South Atlantic, 3:18–19
 see also lake water; sea-surface temperatures; seawater
surf bench sea level indicators, 4:382–383, 4:383f
surficial meteoric carbonates, 1:322–323, 1:323–325, 1:325f, 1:326f
 carbon-isotopic composition, 1:323–325
 oxygen-isotopic composition, 1:322–323
 precipitation, 1:323
 soil water evaporation, 1:323
 temperature effects, 1:322–323
surge channel erosion marks, sea level
 indicators, 4:380, 4:380f
surge fans, 1:897
susceptibility (SUS) *see* magnetic susceptibility
swamp paleotempestology, 3:211–213
Sweden
 ice sheet/glaciations, 2:229
 lakes, 2:562, 2:563f
 relative sea level changes, 4:491
 stand-scale palynology, 3:849–850, 3:850–852, 3:850f
SWF *see* stepwise fluorination
Swiss Plateau
 lake level studies, 2:549, 2:553, 2:554f
 plant macrofossil records, 3:785–786
Switzerland
 Alps, glaciations, 2:231
 late glacial plant macrofossil records, 3:785–786
 Siebenhengste–Hohgant cave system, 1:444
Sylvania, Upper Michigan, USA, 3:848
sylvinetic cycles, pollen records, 4:11
symbolism
 modern humans, 1:146–147
 Neanderthals, 1:146, 1:147
sympathetic nervous system, 1:318–319
synchronized age scales, ice cores, 2:314–315
synchronous fans, 1:897
syngenetic ice wedges, 3:438, 3:438f, 3:439f, 3:468, 3:543f, 3:545f, 3:547f
syngenetic permafrost deposits (Yedoma), 3:542–552
synonymy, pollen databases, 3:834–836
synoptic climatology, 3:102, 3:103
synoptic-scale studies, plant macrofossils/pollen, 3:731
syntax, pollen databases, 3:836
systematics, speleothems, 1:294–296
Szafer, W, 3:854–855
T
TA *see* total alkalinity
tafoni erosion marks, 4:379
Taguro Peat Bed, Japan, 1:207
Tahiti, sea-level change, 4:442–443
Ta-hsin cave, China, 4:655–657
taiga, Arctic Eurasia, 3:734, 3:738–739
Tajikistan
 Middle Pleistocene hominin archives, 1:75–76
 Paleolithic sites, 1:78f
 Pamir glaciations, 2:241
 stone tools, 1:79f
Taklimaken Desert, China, 1:606, 1:607f, 1:608f, 1:609, 1:610f, 1:611, 1:618–620
taliks, permafrost regions, 3:495
talus, 3:558–559, 3:566–573
talus cones, 3:568f, 3:571f
talus slopes, 3:566–573
 debris flows, 3:570–571
 development rates, 3:571–572
 form modification, 3:570–571
 paleoenvironmental significance, 3:572
 protalus ramparts, 3:571
 rockfall and, 3:566–570
 rock glaciers, 3:571
 snow avalanches, 3:570, 3:570f, 3:571f
 talus creep, 3:569
Tamanian mammal assemblage, Eurasia, 4:610–611
Tambora volcano, Indonesia, 4:287
tandem applications, plant macrofossil/pollen studies, 3:729–731
Tanganyika Lake, Africa
 bathymetric map, 2:504f
 chironomid records, 1:367–368
 postglacial pollen records, 4:95–97, 4:98f
Tanytarsus records, 1:398, 1:404f
taphonomy
 analysis, oribatid mites, 2:695, 2:696f
 chironomid fossils, 1:356–357
 definition, 4:748ge, 1:108ge
 diatom studies, 1:477, 1:546
 freshwater mollusks, 3:289–290
 human arrivals, Australia, 1:110
 loss, 3:687
 plant macrofossils, 3:594, 3:594f, 3:596, 3:608f, 3:684–689, 3:699, 3:729, 3:731–732
 processes
 charcoal particles, 2:717
 diatom proxies, 1:546
 plant macrofossil/pollen studies, 3:729, 3:731–732
 vertebrate fossils, 4:594
tapirs, South America, 4:688
Tasman Glacier, New Zealand, 2:212f
Tasmania
 human arrival chronology, 1:111, 1:111f, 1:115f
 Late Pleistocene
 archeological sites, 1:109f
 beetle records, 1:194–195, 1:196, 1:197
 glaciers, 2:202, 2:204–206
 postglacial
 chironomid records, 1:398–399, 1:402–403, 1:403–404
 pollen records, 4:104, 4:106, 4:107f
 tool technology, 1:117

- Tategahana Site, Japan, 1:207, 1:212f
 Tauber-type pollen trap, 3:840, 3:841f
 Taupo Volcanic Zone (TVZ), New Zealand
 FT dating, 1:652–653
 stratigraphy overview, 4:196
 tephrochronology, 4:293–294
 Taurus-Zagros Mountains, 1:99
 taxonomy
 ancient DNA, 4:720–721
 beetle fauna, 1:235–236
 chironomids, 1:371
 dinoflagellates/dinocysts, 2:803–805
 plants
 forensic botany, 4:543, 4:544
 pollen databases, 3:834–836
 Taylor Dome, Antarctica
 ice-margin sites, 2:419–420, 2:421f
 research history, 2:437
 seesaw comparison models, 2:413
 Taylor Glacier, Antarctica, 2:418f, 2:419–420, 2:421f
 TCN *see* terrestrial cosmogenic nuclides
T. cordata (small-leaved lime tree), 3:850–852, 3:851f
 technocomplexes, 1:147–150, 1:148t
 technology
 early Americans, 1:127–129
 lichenometry, 2:571
 modes, social brain evolution, 1:55–56
 Neanderthal vs. modern human, 1:147–150, 1:148t
 Sahul landmass, 1:115–117
 see also lithic technology; tool technology
 tectonic basins, Australia, 2:525–526
 tectonic climate change, 1:676
 tectonic erosion, cosmogenic nuclides, 1:442–443
 tectonic lakes, formation, 1:546–553
 tectonic processes
 cosmogenic nuclides, 1:436–437, 1:442–443
 epeirogenic/isostatic, 4:507–511
 fluvial terraces, 1:684, 1:685–686, 1:686–688
 sea-level change, 4:371
 cave data, 4:461
 Late Quaternary, 4:503–519
 tectonic shorelines, 4:503–519
 tectonic structures/landforms, 1:839–845
 tectonic uplift, 2:127–135
 BLAG hypothesis, 2:128
 Cenozoic cooling, 2:129–132
 climate change, 2:127–128, 2:128–129, 2:129–132, 2:131f
 continental ice-sheet durations, 2:128f
 deep-sea oxygen/carbon isotope records, 2:130f
 drifting continents/oceanic gateways, 2:132–134
 feedbacks between uplift and climate, 2:129f
 GCMs, 2:134–135
 plateau uplift, 2:128–129
 teeth and bones
 African Pliocene–Pleistocene hominids, 1:310
 aridity, 1:308
 bears, 1:310, 1:311f
 boreal environments, 1:308–309
 C₄ biomes, 1:308
 carbon, 1:304, 1:305f
 carbonate stable isotopes, 1:304–314
 composition, 1:305f
 deer, 1:307f, 1:309f
 diagenetic alteration, 1:307
 enamel ESR dating, 2:668, 2:668f, 2:670–672, 2:676
 extinct species, 1:310–311, 1:311–312
 hominids, 1:310, 1:311f
 African Pliocene–Pleistocene, 1:310
 ancient, 1:310, 1:311f
 horse diet, 1:306f
 large mammal ecological flexibility, 1:308
 Late Pleistocene extinctions, 1:311–312
 mineral fraction, 1:307
 Neanderthal diet, 1:310
 nitrogen, 1:305, 1:305f
 organic fraction, 1:307
 oxygen, 1:305–306, 1:305f
 paleobiological tracking, 1:304–307
 paleoenvironment reconstruction, 1:307–310
 physiological effects, 1:306–307
 rodents, 4:725–727
 seasonality, 1:310
 skeletal tissue chronology, 1:306
 stable isotopes, 1:292
 temperate environments, 1:308–309
 terrestrial, 1:304–314
 tropical environments, 1:307–308
 ungulates, 1:311
 see also bones
 tektites, 4:198–199, 4:199f, 4:218–220
 teleconnections
 El Niño, Indian Ocean, 3:51
 paleo-ENSO, 3:171–174
 temperate environments, teeth/bones, 1:308–309
 temperate ice, volcanic landforms, 1:781
 temperate mountain glaciers, 2:289, 2:292–293
 temperature
 air, Arctic, 3:116
 amino-acid dating, 1:39–41
 Antarctic sea ice, 1:532t
 beetle studies
 New Zealand, 1:249–251
 thermal tolerance, 1:166–167
 boreholes, 2:298–302, 3:182, 3:183
 calculations, oxygen isotopes, 1:337–338
 calibrations
 Mg/Ca
 benthic foraminifera, 2:875–876
 planktonic foraminifera, 2:871–873
 Sr/Ca, corals, 2:878, 2:879t
 changes
 Early Pleistocene, 3:3–5
 stable isotopes, 2:347–352, 2:398
 sub-Milankovitch events, 3:204
 chironomid records
 Late Pleistocene, 1:373–385
 postglacial Southern Hemisphere, 1:398–399, 1:403–404
 summer temperatures, 1:387–392, 1:393f, 1:395
 cladocera, 3:275, 3:277
 coccolithophores, effects on, 2:784–786, 2:789–790, 2:790f, 2:791f
 diatoms
 Antarctic sea ice, 1:532t
 oxygen systematics, 1:482–483, 1:484–485
 glacial landforms, 1:911–912, 1:912–913, 1:912f
 glacial landsystems, 1:813–814, 1:817–820
 globally averaged, last interglacial, 3:160–162
 Greenland ice sheet, 2:439–440
 ground surface, permafrost, 3:465–466, 3:470
 inference
 chironomid records, 1:373–385
 Mg/Ca and Sr/Ca paleothermometry, 2:871–883
 Pliocene environments, 3:185, 3:187, 3:188f, 3:190–191
 prediction, beetle records, 1:249–251
 proxies, 2:841–848
 ANNs, 2:843f, 2:844t, 2:845–846, 2:846–847
 approaches, 2:842
 assumption validity, 2:841–842
 biomarker proxies, 2:775–778
 case studies, 2:846
 census counts, 2:841–848
 controversies, 2:846–847
 ecological basis, 2:841
 genotypes vs. morphotypes, 2:842–843
 IKM, 2:841, 2:844–845, 2:846–847
 MAT, 2:842, 2:844t, 2:845, 2:846–847
 Mg/Ca, 2:936–937, 2:937–938, 2:937f, 2:938–939
 microfossil-based, 2:841, 2:843–844
 modern analog methods, 2:845
 morphotypes vs. genotypes, 2:842–843
 oribatid mites, 2:693–694, 2:695
 Oxygen-18, 2:403–405, 2:405–406, 2:405t
 RAM method, 2:843f, 2:844t, 2:845, 2:846–847
 SIMMAX method, 2:843f, 2:845, 2:846–847
 statistically significant data, 2:842
 Quaternary environments, 3:149–150
 reconstructions
 dendroclimatology, 1:463f
 late glacial change, 3:788–789
 quantitative paleoclimate approaches, 3:182, 3:183
 surficial meteoric carbonate composition, 1:322–323
 transfer functions, 1:387–388, 1:389–392, 1:393f, 1:395
 trends, last millennium, 3:229
 see also global warming;
 paleotemperature...; sea-surface
 temperature; surface temperature;
 thermal...

- temperature–precipitation ELA (TP-ELA), 1:911–912, 1:912–913, 1:912f
- tempests, 3:209–221
see also paleotempestology; storm events
- temporal changes/variations
 active temperature glacial landsystems, 1:817–820
 biome composition, 3:866–867
 cosmic ray nuclear disintegration rates, 1:425–426
 diatom records, 1:578f, 1:582f, 1:583f, 1:584f
 glaciated valley landsystems, 1:819f, 1:820–822
 glaciochemical sources, 2:328–332
 ice stream landsystems, 1:821f, 1:822
see also time...
- temporal pattern, glaciers, 2:274–275
- temporary equilibrium-line altitude, definition, 1:909ge
- Tenaghi Philippon basin, 4:63, 4:64f, 4:67, 4:68
- tendency attributes, sea-level index points, 4:372
- Tengger Desert, China, 1:606–608, 1:607f, 1:609, 1:610f
- tension wood, 2:97, 2:97f
- tephra
 definition, 4:277ge
 felsic tuya domination, 1:792, 1:795f
 FT dating
 glass, 1:646–647, 1:647f, 1:648t
 Gold Run, 1:648t, 1:650f
 Huckleberry Ridge, 1:646, 1:646f, 1:647f, 1:650f
 Indonesia, 1:655–656
 New Zealand, 1:652–653, 1:654–655, 1:655t, 1:658f, 1:659f
 ZFT, 1:651, 1:655f
- layers
 dating, 2:271, 2:271f
 volcanic, 1:780ge, 2:417–419, 2:419f, 2:420f
- mounds, 1:793–796, 1:798f
- Rangitawa Tephra, New Zealand, 4:196–197, 4:197f
- ridges, 1:793–796, 1:798f
- sequence pedostratigraphy, 4:256
- tephrochronology, 4:277–304
 applications, 4:289–299
 archeology, 4:298–299
 argon-39/argon-40 dating, 4:287–288
 case studies, 4:289–299
 chemical traces, 4:281–283
 chronology overview, 4:287–289
 dating techniques, 4:287–288, 4:288–289, 4:294–296
 definition, 4:277ge
 direct dating, 4:287–288
 dispersal, 4:278–279, 4:280f, 4:281f
 early studies, 4:278
 enveloping/entrained materials, 4:288–289
 eruption frequencies, 4:289–293
 ferromagnesian assemblages, 4:283–284
 Fe-Ti oxides, 4:285–286
 field characteristics, 4:279–283
 FT dating, 4:287
 geochemical characteristics, 4:283–286
 geomorphic reconstruction, 4:293–294
 glacier ice dating, 4:294–296
 glass, 4:284–285, 4:285f, 4:286f
 historical accounts of eruptions, 4:287
 hominid evolution, 4:298
 landscape reconstruction, 4:293–294
 long tephra records, 4:287
 luminescence dating, 4:289
 macroscopic aspect, 4:279–280
 magnitude, 4:278–279
 microscopic aspect, 4:280–281
 mineralogy, 4:283–286
 neoglaciation chronologies, 2:271, 2:271f
 paleomagnetism, 4:289
 radiocarbon dating, 4:288
 sedimentary basin history, 4:296–298
 silicate minerals, 4:285–286
 tephra characteristics, 4:278–286
 Thorarinsson, 4:278–286
 volcanic hazard assessment, 4:289–293
 wiggle-match dating, 4:288–289
see also tephra
- tephrochronometry, definition, 4:277ge
- tephrostratigraphy, definition, 4:277ge
- Terminal Early Pleistocene mammal assemblages, 4:610–611
- terrace riser
 fluvial terraces, 1:684
 solifluction terraces, 3:483–484
- terraces
 coral records, RSL changes, 4:409–411
 deposits, Late Pleistocene mammals, 4:694f
 proglacial landforms, 2:7, 2:13
 sediments, 3:497f
 sequences, fluvial, 1:684–692, 2:115–119, 4:644
 solifluction, 3:483–484
see also alluvial terraces; marine terraces; river terraces
- terrestrial carbonates, research overview, 1:293
- terrestrial cosmogenic nuclides (TCNs), 1:407–408, 1:408–409
 atmospheric shielding, 1:415
 geomagnetic variation, 1:415
 glaciers, Australia, 2:202–204, 2:204–205, 2:206
 neoglaciation, Europe, 2:257–259, 2:259f
 production
 rates, 1:412t
 in rocks, 1:414–415
 recent developments, 1:410–414
 in situ methods, 1:413–414, 1:414f, 1:418
 solar modulation, 1:415
 surface exposure dating, 2:236–237, 2:237–238, 2:238–239, 2:240–241
- terrestrial depositional glacial systems, 2:21–24
- terrestrial dust, ice cores, 2:329
- terrestrial fauna, 4:674–675
see also individual animals/animal groupings
- terrestrial glacial deposits, sequence stratigraphy, 2:85–86, 2:89f
- terrestrial hydrological system silicon isotopes, 2:737–738
- terrestrial landforms *see* landforms
- terrestrial mollusks, Europe, 2:613–615, 2:615f
- terrestrial non-lacustrine studies, 1:322–332
- terrestrial organic materials, carbonate stable isotopes, 1:293
- terrestrial plant macrofossils *see* plant macrofossils
- terrigenous flux magnetic proxies, 2:889–891
- terrigenous sediments, 2:941–952
 applications, 2:945–948
 clay minerals, 2:943, 2:948, 2:950f
 geochemistry, 2:948
 grain size, 2:945, 2:951
 major/trace element concentrations, 2:948–951
 mineralogy, 2:943, 2:948, 2:950f
 mixing proportion changes, 2:947–948
 oceans, tectonic uplift/continental configurations, 2:132f
 petrography, 2:948
 processes, 2:943–945, 2:945–946
 provenance, 2:941–942, 2:945–946
 solution techniques, 2:948
 sorting aspects, 2:945
 terrain types/components, 2:946–947
 tools, 2:948–951
 weathering, 2:942, 2:943, 2:944–945, 2:946
 x-Ray diffraction, 2:948
 x-Ray fluorescence, 2:948–951
- terrorist activities, forensic searches, 4:524
- Tertiary period
 definition, 2:216ge
 insect fossil studies, 1:173–183
- testate amoebae, 4:420–421, 4:421f
- TEX₈₆ temperature proxy, 2:776–778
- Texas Gulf *see* Gulf of Mexico
- textural properties, sediment particle forms, 2:1, 2:2
- Thailand
 Early Middle Pleistocene vertebrate studies, 4:653, 4:657, 4:660t
 Late Pleistocene mammals, 4:694f, 4:696–697
- Thames terrace staircase vertebrate fossils, 4:596f
- Thâm Hai Cave, Vietnam, 4:657, 4:658t
- Thâm Hang cave, Laos, 4:657, 4:660t
- Thâm Khuyen Cave, Vietnam
 Middle Pleistocene hominins, 1:74f, 1:77
 vertebrate records, 4:657, 4:658t
- Thâm Om Cave, Vietnam, 4:657, 4:658t
- Thâm P'a Loi cave, Laos, 4:657
- THAR *see* toe-to-headwall altitude ratio
- Thar Desert sand seas, 1:631–632
- thaw consolidation, permafrost, 3:481, 3:574, 3:577f, 3:579
- thaw depth, active layer, 3:424, 3:425
- thawing front propagation, active layer, 3:422

- thaw lakes *see* thermokarst lakes
- thaw processes
 sea ice oxygen isotopes, 2:918
 slope deposits/forms, 3:481, 3:483f
 see also freeze–thaw processes
- thaw settlement, active layer, 3:424
- thaw slumps, 3:469, 3:469f, 3:578–579
- thaw subsidence, active layer, 3:424
- THC *see* thermohaline circulation
- Thelypteris palustris* macrofossils, 3:646, 3:647f
- Thera eruption, Santorini, 4:298
- thermal conditions, rock glaciers, 3:535
- thermal contraction cracks
 ice veins, 3:436–440
 ice wedges, 3:436–440, 3:440–446, 3:441f
 primary infilling, 3:436–440, 3:441f
 secondary infilling, 3:440–446
- thermal diffusion paleotemperature records, 2:431–434
 direct climate change reconstruction, 2:433
 isotopic composition of ice-core air, 2:431
 trace interpretation, 2:433
 trace measurement, 2:431–433, 2:432f
- thermal fractionation, firn air, 2:365f, 2:368
- thermally-induced stress, rock weathering, 3:501–502
- thermal neutron interactions at different altitudes, 1:428
- thermal offset, active layer, 3:425
- thermal properties
 active layer, 3:422, 3:422f, 3:424–425
 see also temperature
- thermal regimes
 block streams, 3:517–518
 glaciers, 1:798–800
 permafrost, 3:464, 3:465f
- thermocline depth, South Atlantic, 3:19–21, 3:28
- thermocline-surface radiolarian index (TSRI), 2:836–837
- thermodynamics
 deep-sea carbonate dissolution, 2:859–862, 2:862–863
 Mg and Sr coprecipitation, carbonates, 2:871
- thermohaline circulation (THC), 1:737–747
 benthic foraminifera, 2:770
 climate forcing mechanisms, 1:737–747
 definition, 1:737
 drivers, 1:739–742
 Early Pleistocene paleoceanographic records, 3:5, 3:6
 effect on climate, 1:743–744
 future of, 1:745–746
 glacial climates, 1:737–747
 Greenland/Antarctica ice core comparisons, 2:414
 key features, 1:737
 nonlinear behavior, 1:742–743
 observational data, 1:737–739
 Pliocene climate warmth, 3:187
 postglacial North Atlantic, 3:57–59, 3:60–61
 role in climate change, 1:744–745
- Younger Dryas oscillation, 3:223–224, 3:227
- thermokarst
 climate change and, 3:470, 3:580
 lakes, 3:468–469, 3:574, 3:577–578
 Canada, 3:468f, 3:575f, 3:577f, 3:578f
 climate change and, 3:470
 Yedoma of Beringia, 3:544
 mounds, 3:544
 topography, 3:574–581
- thermoluminescence (TL) dating, 2:643–652
 apparatus diagram, 2:645f
 determining dose rate, 2:646–649
 environments suitable for, 2:649
 equivalent dose estimation, 2:644–646
 additive dose method, 2:644, 2:646f
 partial bleach method, 2:644–645, 2:647f
 regeneration method, 2:645–646
 total bleach method, 2:644, 2:647f
 factors leading to inaccuracy, 2:649–650
 gamma-ray spectrometer, 2:648f
 history of, 1:7
 loess records, Central Asia, 2:589–590
 mechanism responsible for, 2:643–644
 methodology, 1:12
 OSL dating, 2:653, 2:655, 2:656–657, 2:662
 phytoliths, 3:587
 recent advances, 2:650–651
 sample collection, 2:649
 tephrochronology, 4:289, 4:289f
 see also optically stimulated luminescence dating
- thermophilic species, 1:255ge
- thermophilous plants, Southern Europe, 4:183, 4:185, 4:186
- thermoremanent magnetization (TRM), 1:709–710, 1:712, 1:713
- thickness, active layer, permafrost, 3:424–426, 3:427
- thinning of ice sheet annual layers, Greenland, 2:443, 2:444f
- thin sections
 buried soils
 amorphous coatings, 3:385f
 anthropogenic features, 3:385f
 cryogenic features, 3:386f
 crystalline features, 3:384f
 microstratigraphic relations, 3:387f
 glacial sediment micromorphology, 2:52–60, 2:56f
 sediment sample microstratigraphy, 1:541
- Thorarinsson, Sigurdur, 4:278
- thorium-230, 2:925–927, 2:927–929, 2:928f, 2:929–930, 2:930f
- thorium (Th)
 radioisotope proxies, 2:925–927, 2:927–929, 2:928f, 2:929–930, 2:930f
 see also U-series dating
- thousand year (kyr) measures, isotope thermometer, 2:350
- thrust-block moraines, 1:840–843, 1:843f
- Thum Phra Khai Phet cave, Thailand, 4:657
- Thung Lang Cave, Vietnam, 4:657
- thylacines, 4:631–632
- Tibetan glaciations, 2:238f, 2:240, 2:242f
- Tibetan Plateau
 ice core studies/records, 2:379–386
 lake level studies, 2:511–512, 2:514–515, 2:517, 2:519
 tectonic uplift/continental configurations, 2:128, 2:129, 2:134f
- tidal notch sea level indicators, 4:377, 4:378f
- tide-dominated deltas, 1:696
- Tien Shan, Central Asia
 glaciations, 2:240–241
 ice core studies, 2:384–385
- Tierra del Fuego, South America, 2:254
- Tigalmamine Lake, Morocco, 4:14–15
- Tiglian–Menapian stage mammal assemblages, 4:608–610
- Tilia* *see* lime
- till, 2:62–75
 Canadian prairies, 1:757f
 continuum, lodgment–deformation, 2:65–66
 dispersal indicators, glacial erratics, 2:81–84
 fabric analysis, 2:76–80
 data collection, 2:76
 data presentation, 2:76
 statistical analysis, 2:77–78
 summary statistics, 2:76–78
 formation, 2:78
 glacial environment, 2:18–20, 2:21f
 glacier/icesheet extent, 1:890–891, 1:893f
 ice-margin sites, 2:428
 lithology, 2:83
 North American glaciations, 2:182, 2:183f
 plains, 2:21–22, 2:22–24
 subglacial
 depositional system, 2:21–22, 2:22–24
 nomenclature, 2:71–73
 production processes, 2:62–75
 types, 2:66–71
 West Antarctic Ice Sheet, 2:449, 2:450
- timberlines, 3:690, 3:691f
 see also treelines
- time-dependent changes, soil morphology, 3:405–407
- timescales
 climate change, 3:93–101
 features of time series, 3:98–100
 hierarchy of controls/responses, 3:93–94
 rates, plant macrofossil records, 3:789
 representative time series, 3:94, 3:95f
 spectrum of variability, 3:94–98
 spurious periodicity, 3:97–98, 3:98f
 variability/variations, 3:93–94, 3:94–98, 3:98–100, 3:99t
 variance spectrum, 3:94–96, 3:96f
 see also paleoclimatology
- West Antarctic Ice Sheet, 2:451–454
- time series
 Asian lake level studies, 2:520–521
 carbon-14 temporal variations, 4:336, 4:338
 climate change, 3:94, 3:98–100
 detrended series, 3:97–98
 events, 3:100
 features of series, 3:98–100

- time series (*Continued*)
 fluctuations, 3:100
 oscillations, 3:100
 representative series, 3:94, 3:95f
 running means, 3:98
 steps, 3:100
 trends, 3:99–100
 magnetic susceptibility, 2:890f
 pollen–climate relation, 3:811–812
 pollen numerical analysis, 3:825–826
 sedimentary charcoal records, 2:721
 time-slice studies, pollen–climate relation, 3:812–813
 Timor, vertebrate records, 4:662
 tindar, 1:780ge, 1:793–794
 Tiraspolian mammal assemblage, 4:611–612
 Titicaca, Lake, South America
 diatom records, 1:546–547, 1:550–551
 glaciations, 2:250–251, 2:252
 salinity/climate reconstruction, 1:509–510, 1:510f
 Titusville, Pennsylvania, North America, 1:221
 TL dating *see* thermoluminescence dating
 Toba caldera complex, Indonesia, 1:656
 Todd Lake, Canada, 3:578f
 toe-to-headwall altitude ratio (THAR), ELA calculations, 1:914
 toe-to-summit altitude method (TSAM), ELA calculations, 1:914
 Tohoku area, Japan, 1:580–581
 Tomizawa Site, Japan, 1:207
 tool technology
 adaptive radiation and, 1:51
 Africa 2.7 MYR–300,000 years ago, 1:59–66
 Asia
 after 300 ka, 1:98, 1:101f
 Early Pleistocene, 1:67–68, 1:70f
 Middle Pleistocene, 1:72f, 1:73f
 Australia, 1:115–117
 Europe 1.9 MYR–300,000 years ago, 1:83–90
 social brain evolution, 1:55–56
 see also bone tool production; lithic technology; stone tools
 tooth... *see* teeth and bones
 Top End, Australia, 4:107
 topography
 Greenland ice sheet, 2:439, 2:440f
 thermokarst, 3:574–581
 topsoil loss, loess, 2:603–604
 Toringian biozone, 4:210, 4:211f
 tors
 blockfields, 3:523, 3:530, 3:533t
 trimlines/paleonunataks, 1:925
 tosca *see* carbonate accumulations
 total alkalinity (TA), deep-sea carbonate dissolution, 2:860–861, 2:862f
 total glacier area, ELA calculations, 1:914–916
 Tower Hill, Southeastern Australia, 1:196–197
 TP-ELA *see* temperature-precipitation ELA
 trace elements, ice cores, 2:342–346
 trace evidence, forensics, 4:523
 trace gases
 firn air, 2:364–366
 ice core history, 2:463–470
 ice-margin sites, 2:426–427
 preindustrial atmosphere, 2:463, 2:465–466
 trace metals
 lake chemistry, 3:296–297
 marine diatoms as proxies, 2:820
 track fading *see* partial track fading
 traction till, 2:72f, 2:73
 traditional lichenometry, 2:567–570
 transects
 surface sample studies, 3:685
 Western North America pollen records, 4:80–82
 transfer functions
 biogeography, 3:313–314, 3:317f, 3:319
 chironomid records/temperature, 1:387–388, 1:389–392, 1:393f, 1:395, 1:399–400
 diatoms, 1:494, 1:496
 human impact, 1:541
 nutrient indicators, 1:538
 salinity and, 1:507–508, 1:538
 water quality, 1:477–479
 microfossil-based reconstruction, Holocene RSL, 4:425
 nutrient indicators, 1:538
 planktic foraminifera, 2:828
 pollen–climate relation, 3:806, 3:806f
 salinity
 climate reconstruction, 1:507–508
 diatoms as indicators, 1:538
 transgression indicators, sea-level change, 4:390–391, 4:392f
 transgressive sequences, microfossils, 4:423–424
 transgressive surface of erosion (TSE), 4:260–261, 4:262, 4:265, 4:273
 transitional industries
 Early Stone Age, 1:62–63
 Middle Paleolithic, 1:98, 1:101f
 transition forces, West Antarctic Ice Sheet, 2:450–451
 transition zone, active layer, 3:426
 translation role, active layer, 3:426
 translocation, soils, 3:402
 transported cosmogenic radionuclides, 2:354–355
 transported sediments
 clast form analysis, 2:1, 2:2–3, 2:3–5
 deep-sea, 1:637–642
 paleohydrology, 3:253–255
 sand sea formation, 1:623
 solifluction rates, 3:483
 weathering profiles, 3:392
 see also paleosols; windblown sediments
 transverse dunes, 1:599, 1:600f, 1:601f
Trapa natans macrofossils, 3:646, 3:647f
 trapped-charge dating *see* electron spin resonance dating
 trapped electron dating *see* thermoluminescence dating
 trapped gases, ice cores, 2:410, 2:463–470
 trapping pollen, 3:839–845
 tree-death dates *see* cutting dates
 tree limit *see* treelines
 treelines, 3:690–698
 altitudes, chironomid records, 1:393–394, 1:394f
 Arctic
 climate model evaluations, 3:867–868
 Early Holocene paleoclimate history, 3:119
 Eurasia, 3:734–735, 3:735f, 3:742
 North America, 3:754, 3:754f
 dynamics, Holocene, 1:724–725, 1:725f
 megafossils, 3:621–622, 3:622–624, 3:628f
 northern climates, 3:746
 reconstructions, macrofossils, 3:690–692
 see also trees
 tree rings
 carbon-14 record calibration, 4:345, 4:346f, 4:351
 dating, 1:453–458
 archeological applications, 3:630–636
 errors/correction factors, 2:107t, 2:108f
 geomorphology, 2:91–103
 glacial landforms, 2:104–111
 methodology, 1:10
 requirements, 3:630
 sediment analyses, 2:549–551
 dendroclimatology, 1:459–470
 development of, 2:91–92
 paleodroughts, 3:194–195
 paleotempestology, 3:214–215
 quantitative paleoclimate reconstruction approaches, 3:181
 radiocarbon time series, 4:336
 trees
 Australian lack, 1:197–198
 cover, northeastern North America, 4:120
 defining, 3:690
 glacially killed, 2:104
 glacially modified, 2:108–109
 growth reactions, 2:94–100
 Holocene North America, 3:770t, 3:773f
 macrofossils, 3:598f, 3:608, 3:642–646
 pollen records
 African postglacial, 4:91–95, 4:92t, 4:97f, 4:100
 Northern Asia, 4:34
 spread/abundance, 3:859
 see also forests; treelines
 trenching, forensic searches, 4:530
 tributaries, West Antarctic Ice Sheet, 2:450–451
Trichophorum cespitosum macrofossils, 3:642
 see also *Scirpus cespitosus*
 Trichoptera (caddisflies), 3:652, 3:653f, 4:552
 trimlines, 1:918–928
 blockfields, 3:528, 3:529–530, 3:532, 3:532f, 3:533t
 cosmogenic isotopes, 1:924–925
 environmental magnetism, 1:923–924
 formation hypotheses, 1:918–919, 1:920f, 1:921f
 geomorphology, 1:919–921, 1:925
 paleoglaciological considerations, 1:925–926

- relative-age dating, 1:921–924
 soil weathering, 1:923
 surface exposure dating, 1:924–925
 Trinil HK, Java, 4:653–654, 4:655t
 Trinity river
 delta, 1:700, 1:701f
 sediments, 1:666f
 TRM *see* thermoremanent magnetization
 Trogossitidae beetles, 1:214
 trophic status, lakes, 3:657
 tropics/tropical
 areal ice extent, 1:891
 Atlantic river runoff/precipitation, 3:22–24
 chironomid paleoecology, 1:362
 climate
 Last Glacial Maximum GCMs, 3:168, 3:169
 Younger Dryas event, 3:128–129, 3:128f
 cyclone activity, 3:209–221
 dendroclimatology, 1:466–467
 forests, African pollen records, 4:10–12, 4:12f, 4:89–91, 4:98–100
 ice core studies/records, 2:387–394
 rainforests, 4:10–12, 4:12f
 relative sea level changes, 4:495–502
 South America, 2:387–394
 teeth/bone carbonate stable isotopes, 1:307–308
 trottoir sea level indicators, 4:382–383
 trough-mouth fans, 2:33–35, 2:36f, 2:37f, 2:38f
 troughs
 ages/rates, 1:857–861
 calibration, 1:851–852
 cross-profiles, 1:852, 1:855f
 glacial landforms, 1:851–854, 1:855f, 1:857–861, 1:862f
 selective linear erosion, 1:862f
 trough-side cirques, 1:850f, 1:851f
 true flies *see* Diptera
 TSAM *see* toe-to-summit altitude method
 TSE *see* transgressive surface of erosion
 TSRI *see* thermocline-surface radiolarian index
Tsuga canadensis (eastern hemlock), 3:848, 3:849f
 Tsukumogawa Peat Bed, Japan, 1:207
 tsunamis
 benthic foraminifera, 2:770–772
 Late Quaternary tectonics, 4:503
 paleolimnology, 3:267
 research history, 1:22–23
 Tsushima Warm Current (TWC), 1:571
 tubeworms, 4:496–497
 Tucumán loess, South America, 2:631t, 2:634–635, 2:635f, 2:636t, 2:637, 2:638t, 2:640
 tuff cones, 1:780ge, 1:795–796, 1:799f
 Tuktoyaktuk Peninsula, Canada
 megafossils, 3:624, 3:626f, 3:626t
 pollen records, 4:44, 4:48
 Tulane Lake, Florida, 2:545, 2:545f
 tundra
 Arctic
 Eurasia, 3:733–734, 3:734–735, 3:739, 3:740–741, 3:741–742, 3:743
 North America, 3:746, 3:754–755
 lakes, 3:577–578, 3:577f, 3:578–579
 LGM vegetation reconstruction, 3:866–867
 North America
 Arctic, 3:746, 3:754–755
 beetle studies, 1:233
 Northern Asia Late Pleistocene, 4:27, 4:31f, 4:37
 steppe, 1:259–261
 Arctic North America, 3:755
 North American beetle studies, 1:233
 plant macrofossil records, 3:739, 3:740–741, 3:740f, 3:741–742
 Yedoma of Beringia, 3:550–551
 tunnel channels
 definitions, 1:828, 1:830
 formation of, 1:831f, 1:833f
 glacier/icesheet extent, 1:887–888, 1:888f
 Saginaw Lobe, 1:829f, 1:831f, 1:832f
 Superior Lobe, 1:829f
 tunnel valleys, 1:828–832, 1:833f, 1:834f
 definitions, 1:828, 1:830
 glacifluvial, 1:828–832, 1:833f, 1:834f, 1:835–836
 landform associations, 1:830
 meltwater hypothesis, 1:835–836
 morphology, 1:828–830
 origin, 1:830–832
 overview, 1:828
 turbate structures, glacial sediment, 2:56, 2:57f
 turbulence, coccolithophores, 2:786–787, 2:787–788, 2:787f
 turnover pulse model, human evolution, 1:52–53
 turtles
 Australia, 4:628
 dwarfing/gigantism, 4:745–746
 South America, 4:683, 4:684f
 tuyas, 1:780ge
 felsic, 1:792–793, 1:795f, 1:796f, 1:800f
 mafic, 1:781–792, 1:793f, 1:800f
 TVZ *see* Taupo Volcanic Zone
 TWC *see* Tsushima Warm Current
 two-sided freezing, active layer, 3:422, 3:423f
 type 2 diabetes, 1:315–321
Typha macrofossils, 3:641f, 3:642
 typhoons, 3:209–221
- ## U
- U *see* uranium...
 'Ubeidiya, Israel
 Early Pleistocene, 1:68, 1:69–70, 1:72f, 1:73f
 Movius line, 1:77
 ubiquity hypothesis, biodiversity, 3:321
 U/Ca proxy, marine carbon cycle, 2:852f, 2:853
 U₃₇ *see* alkenone unsaturation index
 UK *see* United Kingdom
 Ukraine
 mummified mammals, 4:713, 4:714
 'pickled' fauna, 4:716
 ultra-trace gases, ice cores, 2:463–470
 ultraviolet (UV) absorbing compounds, pigment studies, 3:327
 ultraviolet (UV) radiation
 dissolved organic carbon, 3:335–336
 lake chemistry, 3:296, 3:296f
 unbalanced endemic islands, Southern Asia, 4:662–663
 uncertainty, radiocarbon dating, 4:324, 4:325–326
 underflows, glaciolacustrine, 2:44–45, 2:44f
 ungulates
 South American Late Pleistocene, 4:686–687
 teeth/bone carbonate stable isotopes, 1:311
 uniformitarian methods/principle
 dendrochronology, 1:454
 freshwater mollusks, 3:284
 unionid bivalves, 1:343–346, 1:343f, 1:344f, 1:345f
 United Kingdom (UK)
 archives, paleotempestology, 3:217–218
 relative sea level changes, 4:491–492
 vertebrate fossils, 4:596f
 see also Britain/British Isles; England; Ireland; Scotland
 United States of America (USA)
 cave fauna, 4:716
 Channeled Scabland, 1:825–826, 1:826–828, 1:826f
 data sources/reconstructions, paleotempestology, 3:217
 lake level studies, 2:545
 modern analog approaches, paleoclimatology, 3:104–105, 3:104f, 3:105f, 3:106–107, 3:107f, 3:108f, 3:109–110
 Younger Dryas, 3:126–127, 3:128f
 see also American...; *see also* Americas; North America
 University of Arizona, US, AMS operations, 4:316–319, 4:317f
 unlithified tills, 2:76
 uplift
 Late Quaternary, 4:503, 4:506f, 4:507f, 4:508f, 4:511–512, 4:512f, 4:517f
 submerged speleothem data, 4:463f
 tectonic uplift/continental configurations, 2:127–135
 upper lithosphere nuclear disintegration rates, 1:426–427
 upper ocean circulation, North Pacific, 3:33
 Upper Paleolithic
 atmospheric carbon-14, 4:332–333
 hominid–animal interactions, 4:750–751
 modern human expansion, 1:95, 1:98–100, 1:102f, 1:103f
 technocomplexes, 1:147, 1:148t, 1:149
 Upper Pleistocene *see* Late Pleistocene
 Upper Strensham, England beetle assemblages, 1:188
 upstream controls, climate change, 1:676
 upwelling radiolarian index (URI), 2:836f
 upwellings
 coastal South Atlantic, 3:21–27
 deep water, THC, 1:737

- upwellings (*Continued*)
 dinocyst relationship, 2:812
 Indian Ocean, 3:47
 Pacific Ocean diatom records, 1:574, 1:577f
 radiolarians, 2:836f
 Ural Mountains, pollen records, 4:164
 uranium-234 radioisotope proxies, 2:925, 2:925f, 2:927
 uranium marine carbon cycle proxy, 2:852f, 2:853
 uranium-series dating *see* U-series dating
 urban entomology, 4:548
 Urey, H, 2:907–908
 URI *see* upwelling radiolarian index
 Uruguay, vertebrate fossils, 4:682–683
 USA *see* United States of America
 U-series dating, 1:449, 4:567–571
 carbon-14 record calibration, 4:345–346
 cave data, 4:460–461, 4:463f, 4:464
 coral
 age corrections, 4:568–570
 processes for dating, 4:568f
 ESR dating, 2:670–672, 2:673f, 2:676
 FT dating, 1:644–645, 1:646, 1:651
 history of, 1:6
 methods, 1:13–14
 radioactive dating, ice cores, 2:306
 reef corals, 4:567–568
 speleothems, 1:296–297, 4:570
 stratigraphy overview, 4:193–194
 see also uranium...
 U-uptake models, ESR dating, 2:670–672, 2:673f
 UV *see* ultraviolet radiation
 U–Th dating *see* U-series dating
- V**
- Vaccinium oxycoccus* macrofossils, 3:638, 3:640f
see also *Oxycoccus palustris*
 validation
 models, 2:560
 paleo-ELA reconstruction approaches, 1:916–917
 pollen studies, 3:725–732
 valley bulge structures, 3:487–488
 valley-fills
 alluvial land systems, 3:561–562
 complexes, fluvial sediments, 1:672, 1:674f
 valley glaciers
 Antarctica, 2:416
 icesheet extent and, 1:885–886, 1:886f, 1:887f
 neoglaciation chronology, 2:275f
 valley incision, continental interiors, 2:115–119
 valley landforms, glaciated, 1:764–765
 valley landsystems, glaciated, 1:814, 1:815f, 1:819f, 1:820–822
 valley lowering rates, cosmogenic nuclides, 1:443–444
 valves, diatoms, 1:471
 vapor transport, 2:916–917
 see also water vapor
- variables, concept of, pollen databases, 3:833–834
 variance decomposition, 1:492–493
 variance spectrum, climate change timescales, 3:94–96, 3:96f
 varved sediments
 analytical techniques, 4:584–587
 applications, 4:576–580, 4:587–588
 carbon-14 record calibration, 4:346, 4:351
 chronologies, 1:451, 4:576–580
 classification, 4:583–584
 formation, 4:574–576, 4:582–583
 freshwater diatoms, 1:540–545
 glaciolacustrine, 2:43–44, 2:46f, 2:49f
 history, 4:573–574, 4:582
 lakes, 4:573–581
 annual layer dating methods, 4:573–581
 applications, 4:576–580
 biogenic sediments, 4:574
 chronology, 4:576–580
 clastic sediments, 4:574–575, 4:576f, 4:577f
 as data sources, 4:576
 error estimations, 4:578–579
 evaporitic sediments, 4:574, 4:576, 4:578f
 formation, 4:574–576
 history, 4:573–574
 organic sediments, 4:575–576, 4:578f
 paleoenvironmental reconstruction, 4:579–580
 preservation, 4:574
 thickness measurements, 4:579f
 types, 4:577t
 verification, 4:577
 marine diatoms, 1:554–561
 morphology, 4:583–584
 overview, 4:582–589
 preservation, 4:574, 4:582–583
 subglacial depositional system, 2:22–24
 Vatnajökull ice cap, Iceland, 1:805f
 vector plots, cirque glaciers, 1:850f
 Vedde Ash tephrochronology, 4:280–281, 4:284f
 vegetation
 Australia/New Zealand pollen records, 4:105f, 4:113f
 change
 Africa, 4:89–98, 4:89f, 4:668f
 charred particle analyses, 2:716, 2:717, 2:718f, 2:727
 human adaptations, 1:154
 pigment studies, 3:335–336
 Western North America pollen records, 4:72–83
 communities
 altitudinal distribution, South America, 4:56f
 Arctic Eurasia, 3:733–745
 climatic variables, South America, 4:56f
 Europe, 1:274
 nonanalog, South America, 4:52–54
 Quaternary changes, 4:206–209
 South America, 1:121–122, 4:52–54, 4:56f
- South Island, New Zealand, 2:208f
 surface samples/taphonomy/
 representation, 3:684–689
 cycles, Southern Europe, 4:67f
 dynamics, North America, postglacial, 4:115–123
 European pollen records, 4:1–3, 4:6–7, 4:67, 4:67f, 4:183, 4:185–186, 4:187–188
 extinctions, Southern Europe, 4:67
 feedbacks, last interglacial, 3:158–159, 3:160–162
 history
 environmental archeology, 3:699–724
 Greenland, 3:760–767
 North America
 northeastern, 4:117–120, 4:120–121
 pollen records, 4:142–155
 rodent middens, 3:674–683
 treeline studies, 3:691–692
 Mediterranean postglacial pollen records, 4:183, 4:185–186, 4:187–188
 megafossils, 3:621–622
 modeling
 biome approach, 3:862
 pollen representation, 3:871–879
 treeline studies, 3:695–697
 nonsorted circles, patterned ground, 3:459–460
 parts, plant macrofossils, 3:746
 pattering effect, 3:873–874
 patterns, NWNA pollen records, 4:124, 4:125t, 4:126f
 postglacial pollen records
 Africa, 4:89–98, 4:89f
 Australia/New Zealand, 4:105f, 4:113f
 southeastern North America, 4:135–136
 Southern Europe, 4:183, 4:185–186, 4:187–188
 reconstruction
 Arctic Eurasia, 3:742–743
 Arctic North America, 3:757
 Australasian Late Pleistocene, 4:18–26
 biome approach, 3:861–870
 current developments, 3:869
 future directions, 3:869
 human impact, 3:880–903
 pollen numerical analysis, 3:826–827
 POLLSCAPE model, 3:871, 3:874–878
 treeline studies, 3:690–698
 response
 climate change, 3:790
 combustion reaction, 2:716
 sensing properties, plant macrofossils, 3:725–732
 soil carbonates, 1:325–326
 trends, Southern Europe, 4:66–69
see also plant macrofossils; pollen
 vegetation-defined nonsorted circles, 3:459–460
 vegetation-sensing properties, plant macrofossils, 3:725–732
 vegetative parts, plant macrofossils, 3:746
 vehicles
 exterior plant collection, 4:545
 interior plant collection, 4:545

- veins, ice wedges/casts, 3:436–440, 3:437t, 3:441f, 3:442t, 3:443–446, 3:447–449
- velocity field, rock glaciers, 3:537, 3:539, 3:540
- Venezuela
- glaciations, 2:250
 - vertebrate fossils, 4:680–681
- ventilation, North Pacific waters, 3:34–36, 3:36–37
- vertebrate studies/records
- analysis methods, 4:593–594
 - ancient DNA, 4:719–721
 - Australia, 4:621–638
 - birds, 4:623t, 4:630–631
 - climatic zones map, 4:629f
 - Early Pleistocene, 4:637
 - frogs, 4:621, 4:622t
 - Late Pleistocene, 4:637
 - mammals, 4:625t, 4:631–637, 4:631f, 4:632f, 4:633f, 4:634f, 4:635f
 - mid-Pleistocene, 4:621–638
 - Naracoorte Caves, 4:621, 4:628–629, 4:634, 4:636f
 - rainfall distribution map, 4:630f
 - reptiles, 4:621–630, 4:622t
 - research topics, 4:637
 - Victoria Fossil Cave, 4:621, 4:636f
 - zoogeography, 4:621–637
 - distribution of remains, 4:590–593, 4:595f
 - dwarfing/gigantism, 4:733–747
 - Early Pleistocene, 4:599–604
 - Australia, 4:637
 - Northern Eurasia, 4:605–614
 - evolutionary trends, 4:723–732
 - excavation, 4:593–594
 - historical background, 4:590
 - hominid interactions, 4:748–755
 - Late Pleistocene
 - Africa, 4:664–672
 - Australia, 4:637
 - megafaunal extinctions, 4:700–712
 - mummified mammals, 4:713–717
 - North America, 4:673–679
 - South America, 4:680–692
 - Southeast Asia, 4:693–699
 - microevolution, 4:729
 - Middle Pleistocene
 - Africa, 4:615–619
 - Australia, 4:621–638
 - Europe, 4:639–645
 - North America, 4:646–650
 - Northern Eurasia, 4:605–614
 - Southern Asia, 4:651–663
 - North America, 4:646–650
 - biota change significance, 4:649
 - Blancan age, 4:646–647
 - Irvingtonian age, 4:646–647, 4:647–648, 4:649
 - land mammal ages, 4:646–647
 - Late Pleistocene, 4:673–679
 - Mid-Pleistocene, 4:646–650
 - non-mammalian faunal groups, 4:648–649
 - Rancholabrean age, 4:646–647
 - overview, 4:590–597
 - sampling, 4:593–594
 - South America
 - Late Pleistocene, 4:680–692
 - loess records, 2:631–632
 - speciation, 4:723–732
 - vertical profiles, glacial depositional systems, 2:21, 2:23f
 - Vestfold Hills, East Antarctica, 2:220–221
 - vibrational spectroscopy, 3:349
 - victim recovery dogs, forensic searches, 4:531, 4:533f
 - Victoria Fossil Cave, Australia, 4:621, 4:636f
 - Victoria, Lake, Africa
 - chironomid records, 1:365–367, 1:367, 1:367f
 - diatom records, 1:546–547, 1:551–552
 - Victoria Land, Antarctica, 2:221
 - vicuñas, South America, 4:689
 - Vienna Standard Mean Ocean Water (V-SMOW), 2:347
 - Vietnam
 - Late Pleistocene mammals, 4:696
 - Middle Pleistocene
 - hominin remains, 1:74f, 1:77
 - vertebrate records, 4:657, 4:658t, 4:660t, 4:661t
 - Villafranchian biozone, 4:210, 4:211f, 4:213f
 - Villafranchian–Galerian transition, 4:639–641
 - Villafranchian stage mammal assemblages, 4:599, 4:639–641
 - Europe, 4:639–641
 - Northern Eurasia, 4:605–607, 4:608–610
 - proboscideans, 4:600–601
 - Villányian biozone, 4:210, 4:211f
 - Villányian stage mammal assemblages, 4:605–607
 - virtual geomagnetic poles (VGPs), 1:707, 1:707f, 1:716
 - viruses, ice cores, 2:294–295, 2:296, 2:296f
 - VIS *see* visible spectroscopy; visible stratigraphy
 - viscosity structures, Earth models, 4:455–456
 - viscous flow, block stream formation, 3:518
 - visible-NIR spectroscopy (VNIRS), 3:349
 - applications, 3:352–353
 - sample preparation, 3:351
 - visible (VIS) spectroscopy
 - paleolimnological applications, 3:349–355
 - statistical approaches, 3:351–352
 - theoretical background, 3:349–350
 - visible (VIS) stratigraphy, ice cores, 2:443–444, 2:444f
 - vital effects
 - nonmarine biogenic carbonates, 1:341–342
 - oxygen isotope carbonate composition, 1:338
 - see also* disequilibrium effects
 - VNIRS *see* visible-NIR spectroscopy
 - volcanic aerosols, ice cores, 2:329–330
 - volcanic areas *see* volcanic terrains
 - volcanic ash, lake sediment, 3:305
 - volcanic eruptions
 - frequency, 4:289–293
 - geologic hazards, 1:21–22, 1:21t
 - ice-core chronologies, 2:307, 2:307f
 - major historical eruptions, 1:21t
 - South American loess records, 2:637–638
 - tree-ring networks, 1:463, 1:464f
 - volcanic forcing, 1:748–753
 - climate modeling, 1:749–750
 - last millennium climate, 3:234–235
 - model/data comparisons, 1:750–753
 - proxy data, 1:748–749, 1:750–753, 1:751f
 - solar variability, 1:748–749
 - volcanic glass dating, 1:11–12
 - volcanic hazard assessments, 4:289–293
 - volcanic lakes, Australia, 2:525–526
 - volcanic processes
 - strandline records, 4:507
 - see also* tectonic...; volcanic forcing
 - volcanic tephra layers, 2:417–419, 2:419f, 2:420f
 - see also* tephra; *see also* tephro...
 - volcanic terrains, 1:780–802
 - cosmogenic nuclide dating, 1:435
 - morphometric parameters, 1:784t, 1:794f, 1:795f
 - plant macrofossil ¹⁴C, 4:362
 - volcanism/vulcanism, 1:749, 1:780–802
 - voles (*Microtus*)
 - Early Pleistocene, 4:601–602
 - evolution of, 4:207–208, 4:209f, 4:725–727, 4:731
 - Matuyama–Brunhes boundary, 4:641
 - mummified, 4:715
 - radiation and, 4:601–602
 - von Blanckenburg, F, 2:131–132, 2:132f
 - von Post, L, 2:699–700, 3:854–855
 - Vostok ice core, Antarctica
 - biological material, 2:289–291, 2:290f, 2:291f, 2:292f, 2:293f, 2:295f, 2:296f
 - correlation techniques, 2:410
 - methane studies, 2:339–340
 - research history, 2:437
 - stable isotopes, 2:396, 2:398f, 2:400
 - Vostok lake biological materials, 2:291, 2:291f
 - Vrica GSSP, stratigraphy, 4:191–192
 - V-SMOW *see* Vienna Standard Mean Ocean Water
 - vulcanism/volcanism, 1:749, 1:780–802
 - see also* volcanic...
- ## W
- Waalre Formation, Netherlands, 4:238
- Wacken/Dömnitz substage, Eurasia, 2:173–174
- WADE *see* Water Depth Ecology methods
- WAIS *see* West Antarctic Ice Sheet
- WAIS-D *see* West Antarctic Ice Sheet Divide
- Waitotara Valley, New Zealand, 1:247–248, 1:249f
- Wajak, Java, 4:695
- Walker circulation hypothesis, paleoceanography, 3:6
- wallabies, 4:636
- Wallacea islands, Southern Asia, 4:662
- Walther's law, diamicts, 2:18–19

- Wanganui Basin, New Zealand
 sequence stratigraphy, 4:267f, 4:268f,
 4:269–271, 4:270f, 4:271f,
 4:274f
 stratigraphy overview, 4:192, 4:195–196,
 4:196f
 warm climates, Pliocene, 3:185–186,
 3:186–188
 warming processes
 Andes, 2:392
 Antarctica, 2:220
 Pliocene, 3:185–186, 3:186–188
see also global warming; Medieval Warm
 Period
 warmth indicator plants, Arctic Eurasia,
 3:739f
 warm-water diatom species, 1:573f, 1:574f,
 1:580f
 Warwickshire, England, beetle assemblages,
 1:188
 water balance, lake-level modeling,
 2:558–559, 2:559f
 water budgets, Asian lakes, 2:506, 2:511,
 2:512f, 2:517–518
 water column
 photosynthetic pigments in, 3:327t
 transport, alkenone paleothermometry,
 2:759–760
 water depth
 radiolarians, 2:836–837, 2:837–838,
 2:837f
 relative, lake sediment, 3:306f
see also lake levels; sea levels
 Water Depth Ecology (WADE) methods,
 2:836–837, 2:837–838, 2:837f
 waterfronts, diatom analysis, 1:519–520,
 1:519f
 waterlogging, plant macrofossils, 3:702
 water loss, $\delta^{18}\text{O}_{\text{water}}$, 1:485–486
 water masses, Late Pleistocene South Atlantic,
 3:28
 water oxygenation, benthic foraminifera,
 2:767–768
 water quality
 biogeography, 3:319–321
 chironomid records, postglacial Europe,
 1:386–387, 1:396
 diatom use, 1:477–479
 watershed disturbances, cladocera,
 3:276–277
 water stable isotopes, 2:419–424, 2:421f
see also stable isotopes
 water vapor, stable isotopes, 2:347–348,
 2:349, 2:916–917
 water voles, 4:725–727, 4:731
 wave-dominated deltas, 1:696, 1:697f
 wavelet analysis, diatom records, 1:581,
 1:582f
 Waverley Wood, England beetle assemblages,
 1:185–186
 weapons testing, nuclear power, 4:342f,
 4:343
 weathered rock, 3:500–506
 cosmogenic nuclide dating, 1:432
 slope deposits/forms, 3:481
 weathering profiles, 3:392
 weathering
 climate/tectonics, 1:442–443
 feedback, biosphere, 1:727
 front, profiles and, 3:392–394, 3:393f
 limits, trimlines/paleonunataks, 1:918
 processes
 biological, 3:502
 blockfields, 3:523, 3:524–525,
 3:526–528, 3:527f, 3:529
 chemical, 3:502
 diatom silica proxies, 2:734, 2:737–738,
 2:739, 2:740–741
 hydration shattering, 3:501
 ice segregation, 3:500–501, 3:501f,
 3:504f
 mechanical, 3:500–502
 profiles and, 3:392–395
 rocks, 3:500–502, 3:501f, 3:504f
 sediments, 2:1, 2:3
 soil properties, 3:402, 3:407, 3:408
 thermally induced stress, 3:501–502
 volumetric expansion, 3:500
 profiles, 3:392–401
 applications, 3:400
 blockfields, 3:527t
 dating, 3:399–400
 processes and, 3:392–395
 spatial patterns, 3:397–399
 zones, 3:395–397, 3:398t
 rock weathering, 1:432, 3:392, 3:481,
 3:500–506
 terrigenous sediments, 2:942, 2:943,
 2:944–945, 2:946
 zones, 3:395–397, 3:398t
 weather records *see* historical climatology
 Weaver Drain, Michigan, North America,
 1:222
 wedges *see* ice wedges
 weeds
 plant macrofossils, 3:607f
 seeds, environmental archeology,
 3:714–715
 weevils
 North America, 1:225, 1:233
 Northern Asia, 1:256f
 South America, 1:235, 1:236–237, 1:238f
 Weichselian cold stages
 Alps, glaciations, 2:231–232
 beetle assemblages, 1:201f, 1:202–203,
 1:263–271, 1:269t, 1:271–272
 British/Irish ice sheet, 2:230–231
 Eurasian glaciations, 2:224–232, 2:226f,
 2:227f
 Northern Russia, glaciations, 2:224–228
 Scandinavian ice sheet, 2:228–230, 2:228f
see also last cold stage
 weighted-averaging
 chironomid-based inference models,
 1:370f
 definition, 1:362ge
 diatoms, 1:494, 1:496
 partial least squares
 definition, 1:362ge
 pollen–climate relation, 3:809–810
 pollen–climate relation, 3:809–810
 Werowrap Lake, Australia, 1:398
 West Africa
 monsoon 6000 ^{14}C years BP, 3:868–869
 postglacial pollen records, 4:85–88,
 4:88–89, 4:88f, 4:89–91, 4:89f, 4:95f,
 4:97f, 4:100f
 vertebrate fossil records, 4:670–671
see also Africa; *see also* individual countries
 West Antarctic borehole temperature records,
 2:301
 West Antarctic Ice Sheet Divide (WAIS-D),
 2:438
 West Antarctic Ice Sheet (WAIS)
 characteristics, 2:216
 definition, 2:216ge
 dynamics of, 2:448–455
 Pleistocene, 2:218
 research overview, 2:437, 2:438
 West Atlantic Younger Dryas climate event,
 3:128–129
 west-central European lake level studies,
 2:549–557
 Westerlies
 Asian lake level studies, 2:511, 2:517–518
 Australian lake level studies, 2:529
 Latin American lake level studies,
 2:531–532
 Southern wind patterns, 3:73, 3:76, 3:82–84
 western Amazonian vertebrate fossil sites,
 4:681–682
 Western Cordillera pollen records, 4:129–130
 western Ecuador vertebrate fossil sites, 4:681
 western Eurasia, Neanderthal demise,
 1:146–153
 Western Europe
 beetle assemblages, 1:184–190
 hominin colonization, 1:84–87
 MPT, 1:84–87
 post-MPT, 1:87
 pre-MPT, 1:84
 lake level studies, 2:549–557
 loess records, 2:607–608, 2:610–612
see also Europe; *see also* individual countries
 Western North America (WNA)
 beetle studies
 fossil record, 1:284t, 1:286–289, 1:286f
 Late Pleistocene, 1:225–228, 1:226t,
 1:227f
 environments of, 4:72
 pollen records, 4:72–83
 vegetation/climate history, 4:144–154
see also northwestern North America;
 southwestern North America
 Western Norway temperature
 reconstructions, 3:788–789
 western Peruvian vertebrate fossil sites, 4:681
 Western Siberian palynological records, 4:27,
 4:29f, 4:36–37
 Western Victoria, Australia, 2:525–526, 2:529
 West Midlands, England beetle assemblages,
 1:187
 Westport area, New Zealand, 1:252
 West Runton freshwater bed, England, 1:185
 wet-based deglaciation fans, 1:896
 wet deposition, aerosols, 2:326–327
 wet extraction methods, ice cores, 2:468
 wet forest biota, Australia, 1:198

- wetlands
 back-barrier paleotempestology, 3:211–213
 postglacial North American pollen records, 4:133–134, 4:136–140
 sediments, plant macrofossil studies, 3:726
 vegetation, Arctic Eurasia, 3:743
 whaleback erosional forms, 1:872–873, 1:874–875, 1:874f
 whaling, nutrient/contaminant biovectors, 3:347
 wiggle-matching
 atmospheric carbon-14 variations, 4:329, 4:330
 carbon-14 record calibration, 4:348–349, 4:349f, 4:351
 tephrochronology, 4:288–289
 wild plants, archeology use, 3:705–706, 3:713f, 3:717f
 WildWood, Northern Europe, 4:175
 Willandra Lakes, Australia, 2:525f, 2:526–527, 2:527f
 Willenbring, JK, 2:131–132, 2:132f
 Wiman Nakin cave, Thailand, 4:657, 4:660t
 windblown sediments
 deep-sea, 1:637–642
 dune fields
 high latitudes, 1:597–605
 low latitudes, 1:623–636
 midlatitude, 1:606–622
 ice core studies, China, 2:379
 loess
 Central Asia, 2:585–594
 origins/properties, 2:573–584
 micromorphology, 3:381–391
 mineral magnetic analysis, 3:375–380
 morphology, 3:402–411
 overview, 3:357–366
 weathering profiles, 3:392–401
see also eolian...; paleosols; transported sediments
 wind direction, dunes, North America, 1:620–621
 wind field maps, Africa, 4:87f
 Windmill Islands, East Antarctica, 2:221
 wind patterns, South Pacific Ocean, 3:73, 3:76, 3:82–84
 wind systems, South America, 2:638–639, 2:638t
 wingless birds, 4:746
 winter balance, definition, 1:909ge
 Winter Gulf, New York, 1:222
 Wisconsin glaciations
 age limitations, 4:673
 Beringia events, 2:194–197, 2:197–199, 2:197f, 2:198f, 2:199f
 Wisconsin–Holocene deglacial transition, 2:301
 WNA *see* Western North America
 Wolfe, AP, 3:342–344
 wombats, 4:633
 wood anatomy, 1:453–454
 future trends, 2:101
 glacially modified trees, 2:109f
 root changes, 2:100
see also dendrochronology; tree...
 woodland pollen records, Africa, 4:13–14
 wood megafossils, 3:622–624
see also wood remains
 wood pasture pollen records, 3:890
 wood rat middens, 4:143, 4:153f, 4:154
 wood remains
 analysis/identification, 3:703–704, 3:705f
 archeobotanical research, 3:715–717
 domesticated plants, 3:708
 preservation, 3:699–702
see also wood megafossils
 Woods Lake, Australia, 2:525f, 2:526, 2:528
 Woodston Beds, England beetle assemblages, 1:187
 wood use behaviors, dendroarcheology, 3:633f
 woody cover *see* tree...
 woolly mammoth
 extinction, 4:700–704, 4:707–708, 4:708–709
 mummified, 4:713–714
 ‘pickled’, 4:716
 woolly rhinoceros
 mummified, 4:714
 ‘pickled’, 4:716
 Worcestershire, England beetle assemblages, 1:188
 world overview, postglacial human adaptations, 1:154
 wounding scars, tree growth reactions, 2:97
 Würmian glaciation, 2:231–232, 2:231f
- X**
- Xaiochangliang site, Asia, 1:67–68
Xanthoria elegans lichenometry, 2:570f
 xenarthra, South America, 4:684–686, 4:685f
 xerophilic species, 1:255ge, 1:269t
 xerophilous upland vegetation, Arctic, 3:743, 3:743f
 X-ray analyzers, SEMs and, 4:539–540
 X-ray densitometry, 1:453, 1:456
 X-ray diffraction (XRD)
 soil/sediment in forensics, 4:540
 terrigenous sediments, 2:948
 trimlines/paleonunataks, 1:923
 X-ray fluorescence (XRF), terrigenous sediments, 2:948–951
 XRD *see* X-ray diffraction
 XRF *see* X-ray fluorescence
 xylophages, 1:255ge
- Y**
- Yabrudian industry, Asia, 1:98
 Yakutia plant macrofossil records, 3:739–740, 3:740–741, 3:741f
 Yarra Creek, King Island, Tasmania, 1:194–195
 Yasuda Formation, Japan, 1:210–211
 YD *see* Younger Dryas
 Yedoma (permafrost deposits), 1:255–256, 1:260f, 3:542–552
 Alaska/Yukon, 3:543, 3:545–546, 3:547f, 3:548, 3:549–550, 3:551t
 cryolithology, 3:546–548, 3:546f, 3:548–549
 distribution, 3:543f, 3:544–546
 formation processes, 3:548–550, 3:549f
 fossils, 3:548
 grain-size distribution curves, 3:546f
 Late Pleistocene, 3:542–552
 northeast Siberia, 3:544–545, 3:548–549, 3:551t
 paleoenvironmental archives, 3:548
 terminology, 3:542–544
 typical features, 3:551t
 Yellow river delta site case, 1:700, 1:702f
 Yenchinkou, South China, 4:655
 Yoa Lake, Chad, 4:100, 4:101f
 Younger Dryas (YD), 3:126–134, 3:222–228
 African postglacial pollen records, 4:85–86, 4:87–88, 4:88–89
 Alaska/Yukon pollen records, 4:44
 Arctic paleoclimate history, 3:118
 atmospheric carbon-14 variations, 4:331
 Australia/New Zealand time signal, 4:106
 beetle assemblages, 1:205
 Europe, 1:275
 South America, 1:235, 1:241, 1:242
 Beringia events, 2:198–199
 biogeography, 3:314–315, 3:317f
 causes of, 3:130–131
 chironomid records
 climate change response, 1:375f, 1:376–378, 1:376f, 1:378–379, 1:380
 postglacial Southern Hemisphere, 1:400–401, 1:401–402, 1:402–403
 deglaciation
 Eurasia, 2:232, 2:233f
 diatom records, 1:551, 1:551f
 geographic response summary, 3:129
 glacial landforms, 1:913, 1:917
 global evidence, 3:222–228
 ice sheets
 Laurentide Ice Sheet, 3:130–131
 Indian Ocean, 3:46
 Northern Hemisphere, 3:126–128, 3:222–225, 3:226–227
 NWNA pollen records, 4:126, 4:128
 ocean circulation, 3:129–130
 outside amphi-North Atlantic region, 3:224–225
 paleoceanography overview, 2:751–752, 2:752–753
 paleoclimate reconstruction, 3:222–228
 Preboreal transition, 1:275
 research history, 3:222
 Southern Hemisphere, 3:225–226, 3:227
 chironomid records, 1:400–401, 1:401–402, 1:402–403
 climate, 3:129
 THC, 1:744, 1:745
 timing of, 3:223–224, 4:106
 triggering mechanisms, 3:223
 tropical climate, 3:128–129
 Younger Dryas Chronozone (YDC) pollen records, 4:126, 4:128

Younger Saalian glaciation, 2:176–177
 Youngest Toba Tephra (TYY), 1:656
 Yucatan peninsula, lake level studies,
 2:531–532, 2:532–533, 2:533–534, 2:535
 Yukon Territory, Canada/North America
 boreal forests, 3:576*f*
 insect fossil studies, 1:174–176, 1:178*f*
 lake level studies, 2:543–544
 Late Pleistocene events, 2:191, 2:197,
 2:198
 mummified mammals, 4:713
 arctic ground squirrel, 4:715
 black-footed ferret, 4:715
 horses, 4:714
 woolly mammoth, 4:714
 neoglacial chronology, 2:272
 pollen records, 4:39–44, 4:47

retrogressive thaw slumps, 3:579*f*
 themokarst lakes, 3:575*f*
 Yedoma, 3:543, 3:545–546, 3:547*f*, 3:548,
 3:549–550, 3:551*t*
see also Eastern Beringia

Z

Zagros Mountains, Iran, 3:786–787, 3:788*f*
 Zebra Mussel, nonmarine biogenic
 carbonates, 1:346
 Zeribar, Lake, Iran, 3:786–787, 3:788*f*
 zero-curtain effect, permafrost, 3:423
 zeta bay shoreline sea level indicator,
 4:393*f*
 ZFT *see* zircon fission-track dating

zinc
 marine carbon cycle proxy, 2:853–854,
 2:853*f*
 nutrient proxy, 2:901–903
 zircon fission-track (ZFT) dating, 1:11–12,
 1:644, 1:644*f*, 1:651–652, 1:655, 1:655*f*,
 4:287
 Zn/Ca marine carbon cycle proxy,
 2:853–854, 2:853*f*
 see also zinc
 zones, weathering profiles, 3:395–397, 3:398*t*
 zoobenthos, definition, 1:362*ge*
 zoogeography
 Australian vertebrate records, 4:621–637
 beetle records, 1:169
 zooplankton, 3:264–265, 3:267
 Zyryan glaciation, 2:194–197, 2:195*f*



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